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EVALUATION OF HYDROLOGIC PROPERTIES OF THE LONG ISLAND GROUND-WATER RESERVOIR USING CROSS-SECTIONAL ELECTRIC ANALOG MODELS

GEOLOGICAL SURVEY Open-File Report 75-679

Prepared in cooperation with the Nassau County Department of Public Works, the New York State Department of Environmental Conservation, the Suffolk County Department of Environmental Control, and the Suffolk County Water Authority

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CONTENTS

ILLUSTRATIONS

TABLES

CONVERSION FACTORS

Factors for converting English units to metric units are shown to four significant figures. However, in the text, the metric equivalents are shown only to the number of significant figures consistent with the values for the English units.

English	Multiply by	Metric
feet (ft)	3.048×10^{-1}	metres (m)
square feet (ft^2)	9.290×10^{-2}	square metres (m^2)
feet per day (ft/d)	3.048×10^{-1}	metres per day (m/d)
million gallons per day (M gal/d)	3.785×10^{3}	cubic metres per day (m^3/d)
million gallons per day per square mile [(M gal/day)/mi^2]	5.354×10^{1}	cubic centimetres per square centimetre per year (cm/yr)

EVALUATION OF HYDROLOGIC PROPERTIES OF THE LONG ISLAND

GROUND-WATER RESERVOIR USING CROSS-SECTIONAL

ELECTRIC-ANALOG MODELS

By O. L. Franke[1] and R. T. Getzen, Mineola, N.Y.

ABSTRACT

Variables necessary for constructing a quantitative model to predict the response of ground-water reservoirs to hydrologic stresses are in four classes: (a) external and internal geometry of the reservoir, (b) nature of the reservoir boundaries, (c) areal distribution of horizontal (P_x) and vertical (P_z) hydraulic conductivities, and (d) areal distribution of storage capability within the reservoir. In this report, the sensitivity of the Long Island ground-water reservoir's reponse to the first three of these variables was evaluated using analog models.

Long Island is underlain by a wedge-shaped mass of unconsolidated deposits ranging from near zero thickness in northeast Queens County to a maximum thickness of 2,000 ft (600m) in south-central Suffolk County. The bedrock beneath the unconsolidated deposits is crystalline rock of Precambrian (?) age. The principal hydrogeologic units within the unconsolidated deposits are the upper glacial aquifer, Gardiners Clay, Magothy aquifer, Raritan clay, and Lloyd aquifer. The boundaries of the fresh ground-water reservoir are the water table, the fresh-salt water interfaces in the hydrogeologic units, the bedrock surface, and the surfaces of contact between the ground-water reservoir and the bodies of salty surface water surrounding Long Island. In addition, the streams and their associated shallow ground-water flow systems significantly affect the flow within and discharge from the ground-water reservoir. Both the water-table and the interfaces are dynamic (movable) boundaries.

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Four analog models of the same approximately north-south hydrogeologic cross-section, located along the border between Nassau and Suffolk Counties, were built and studied. Variations in hydraulic conductivity and anisotropy in the upper glacial and Magothy aquifers were simulated in the models. Tests with each model varied, in addition, the vertical hydraulic conductivity of two important confining units (Gardiners Clay and Raritan clay), steady-state recharge to the ground-water reservoir, and stream seepage from the ground-water reservoir. Two different boundary conditions for the seaward boundary were also studied.

The most important results of this study are: (1) The vertical hydraulic conductivity of the Gardiners Clay, a thin clay of low hydraulic conductivity that occurs near the south shoreline at the base of the upper glacial aquifer, proved to be a significant influence on model heads and on the distribution of model outflow to the south shore. Although this study may provide a final answer, a vertical hydraulic conductivity of less than 10^{-2} ft/d (3×10^{-3} m/d) for the Gardiners Clay gives model results that compare reasonably well with existing prototype (field) data.

(2) Accurate assessment of the hydrologic effects of the streams in any regional modeling of the Long Island ground-water reservoir is important. Simulation of streams in the models greatly modified the (a) water-table profile, (b) position of the water-table divide, (c) heads within the ground-water reservoir, (d) amount and location of ground-water discharge into adjacent salt-water bodies, and (d) equilibrium position of the interfaces. (3) Heads measured in the 2 upper aquifers between the north and south shorelines were lower in models that simulated the Raritan clay and Lloyd aquifer than in models that did not. (4) Fresh-water heads in the prototype indicate that the interface in the Magothy aquifer south of western Suffolk County was probably in a transient condition before the advent of man's influence. The equilibrium position of the toe of the interface corresponding to present-day fresh-water heads would probably lie near the south shoreline whereas it actually is located some distance, but at least several miles, south of the south shoreline. (5) In the model tests considerable quantities of fresh ground water remain in simulated storage near the study cross section between the north and south shores, assuming steady-state equilibrium positions of the interfaces, for average areal recharge rates as low as 0.10 (M gal/d)/mi^2 [(5.4 cm/yr)]. (6) The set of model values of the hydrologic parameters that resulted in a model response which compared reasonably well with the prototype probably provides an adequate representation of the prototype for most purposes. No combination of parameter values provided an exact representation of the prototype, but combinations of nonhomogeneous aquifers and nonuniform recharge were not considered. The best combination of parameters that was tried is:

INTRODUCTION

Reliable predictions of the response of the Long Island ground-water reservoir to present and future man-induced stresses by means of models, either analog or digital, are of critical importance to individuals and agencies responsible for managing the water resources of the island. The information necessary for constructing unsteady-state models of any ground-water reservoir includes: (a) External and internal geometry of the reservoir, (b) nature of the reservoir boundaries, (c) areal distribution of horizontal and vertical hydraulic conductivities and thicknesses of the hydrogeologic units, and (d) areal distribution of values of the storage coefficient.

Determination of hydraulic conductivities for hydrogeologic units being modeled always involves some uncertainty. This uncertainty is generally greater in assigning meaningful data to vertical than to horizontal hydraulic conductivity and is greatest for confining units.

This report describes the results of tests on four steady-state, electric-analog, cross-sectional models simulating a single north-south cross section of the Long Island ground-water reservoir (section A-A', fig. 1). In these tests, model system parameters including absolute values of hydraulic conductivity, ratios of horizontal to vertical hydraulic conductivity, conductivity of confining beds, rate of recharge, and effect of seaward boundary were varied in a consistent manner. The basic approach was that of sensitivity analysis--modifying model parameters, measuring model response, and comparing model response with prototype (real world) hydrologic data.

The purpose of the investigation was threefold: (1) To determine the effect of the different hydraulic-conductivity combinations and other parameters on the head distribution within the model ground-water reservoir, the steady-state position in the model of the interface[1/] between fresh ground water and salty ground water, and the distribution of model outflow; (2) to compare model heads and other model data with corresponding prototype data under natural conditions; and (3) to discuss the implications of the model results on future modeling efforts and on gradually developing concepts of how the Long Island ground-water reservoir functions.

Of the four categories of information necessary to construct a ground-water model, item c is the principal focus of this report. Items a and b are discussed briefly under "Hydrogeologic Setting," and item d is not a concern here because this investigation considered only steady-state conditions.

1/ The word "interface," where used by itself, will for the remainder of this report refer to the interface between fresh ground water and salty ground water.

HYDROGEOLOGIC SETTING

Long Island is underlain by a wedge-shaped mass of unconsolidated deposits ranging from near zero thickness in northeast Queens County to a maximum thickness of 2,000 ft (600 m) in south-central Suffolk County (fig. 1).

Figure 1 (caption on next page) belongs near here.

Pertinent characteristics of these deposits are listed in table 1. Thickness and areal extend of the different hydrogeologic units are shown in a series of maps prepared by McClymonds and Franke (1972).

Before wells and sewers were used on Long Island, precipitation was the source of all recharge to the ground-water reservoir. Previous investigators estimated average annual recharge to the ground-water reservoir to be about half the average annual precipitation, or about 1 (M gal/d)/mi^2 [(50 cm/yr)] (Cohen and others, 1968, p. 44). This estimate of recharge was obtained by subtracting estimates of average annual evapotranspiration and other water-budget components, including streamflow and subsurface outflow, from average annual precipitation. Considerable uncertainty is inherent in the estimates of evapotranspiration and subsurface outflow (Franke and McClymonds, 1972, p. 20 and 33) and in the estimates of average annual recharge.

Figure 1.--Index map of Long Island, N.Y., showing the regional ground-water divide and location of section A-A'.

Table 1.--Major hydrogeologic units on Long Island, N.Y.

Hydrogeologic[1/] unit	Approximate maximum thickness (feet)	Description
Quaternary		
Upper glacial aquifer	400	Mainly sand and gravel; some thin beds of clayey material.
Gardiners Clay	150	Clay, silty clay, and a little fine sand.
Jameco aquifer	200	Mainly medium to coarse sand. Not found along section A-A' (fig. 1).
Cretaceous		
Magothy aquifer	1,000	Mainly very fine to medium sand interbedded with silty sand and silty clay; some coarse to fine sand; locally contains gravel.
Raritan clay	300	Clay; some silt and fine sand.
Lloyd aquifer	300	Sand and gravel; some clayey material.
Precambrian(?)		
Bedrock	...	Crystalline rock of very low interstitial hydraulic conductivity.

1/ Nomenclature after Cohen and others (1968).

Under natural conditions, ground water is discharged by (1) seepage to streams; (2) subsurface outflow to the bays, the Atlantic Ocean, and Long Island Sound; and (3) to a small extent, evapotranspiration from areas near the shorelines. Comparison of water budgets for parts of Nassau and Suffolk Counties (Franke and McClymonds, 1972, p. 33) indicates that more than five times as much streamflow discharges into the south-shore bays as into Long Island Sound. Thus, the ratio of stream discharge to total ground-water outflow is much smaller for north-shore streams than for south-shore streams.

The ground-water system of Long Island can be considered to include an islandwide, regional flow system and a series of shallow subsystems associated with the island's streams. The regional flow system consists of two major subsystems separated by an approximately east-west trending, roughly vertical surface, which is termed the "regional ground-water divide" (fig. 1). North of this divide, ground-water movement is generally toward Long Island Sound; south of this divide, ground-water movement is generally toward the southern bays and the ocean.

Ground water in the shallow subsystems discharges mainly into Long Island's streams, which flow to tidewater. The flow in the shallow subsystems is three dimensional. However, as noted by Franke and McClymonds (1972, p. 22-23) and Franke and Cohen (1972), ground-water flow in the deeper aquifers roughly approximates two-dimensional flow in north-south vertical planes for a considerable part of Nassau and Suffolk Counties.

The boundaries of the fresh ground-water reservoir are the water table, the interfaces in the hydrogeologic units, the bedrock surface, and the surfaces of contact between the ground-water reservoir and the bodies of salty surface water surrounding Long Island (Franke and McClymonds, 1972, p. 11-15). In addition, the streams and their associated shallow subsystems collectively modify the flow within and the discharge from the ground-water reservoir.

Both the water table and the interfaces are dynamic (movable) boundaries. The water table (the upper boundary of the ground-water reservoir) is neither an equipotential nor a flow-line surface, but is a boundary across which water is intermittently recharged. The interfaces are assumed to be sharply bounding flow-line surfaces, but this assumption is only an approximation because it ignores the zone of diffusion. The bedrock surface is assumed to be the lower bounding flow-line surface of the reservoir. Finally, the bottoms of the bodies of salty surface water surrounding the island are assumed to be constant-potential surfaces, which is correct only as a first approximation. This assumption is discussed in the sections on model tests.

MODEL TESTS THAT DO NOT INCLUDE SIMULATION OF THE LLOYD AQUIFER

Description

The geologic section simulated by the four cross-section models (A-A', fig. 1) is in an area where the regional flow in all except the upper 30 feet of the saturated zone is approximately two-dimensional in the vertical plane roughly perpendicular to the regional ground-water divide. This geologic section is simulated by a square resistor network, as shown in figure 2. The network does not include a simulation of the

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Figure 2 (caption on next page) belongs near here.

Raritan clay or the Lloyd aquifer. In omitting the Lloyd aquifer, the authors assumed that this aquifer has only a slight effect on the head distribution in the upper part of the ground-water reservoir. The effect of including a simulaiton of the Lloyd aquifer in the cross-sectional models is discussed in the section "Model tests that include simulation of the Lloyd aquifer."

Near the south shore, the geologic boundaries for the model depicted in figure 2 are good representations of the prototype boundaries. The model cross sections were extended as much as 58,000 ft (17,500 m) south of the shoreline on the basis of an assumed constant slope of the bed-rock surface and a continuous and constant thickness of the Gardiners Clay. Geologic and physical boundaries near the north shore are highly generalized in figure 2 because of the variability of geology and topography in this area.

Figure 2.--Finite-difference network and electrical representation of boundaries for the cross-section analog models.

Procedures outlined by Walton and Prickett (1963) and Karplus (1958) were used in dimensioning and scaling the analog models. Walton and Prickett's procedure is based on studies by Bermes (1960) and Skibitzke (1961). The intersections of perpendicular gridlines in figure 2 represent model nodes (junctions of four resistors). Horizontal resistors simulate the horizontal hydraulic conductivity of a specified volume of aquifer material, and vertical resistors simulate the vertical hydraulic conductivity of an equal-sized volume of aquifer material. One resistor in this cross section simulates a rectangular prism of reservoir material whose dimensions are 2,000 ft (horizontal) by 40 ft (vertical) by 1 ft (perpendicular to section) (610 x 12 x 0.3 m). The one minor exception is the simulation of the Gardiners Clay, whose entire thickness is represented by one continuous line of horizontal and vertical resistors. Thickness of Gardiners Clay on the mainland of Long Island and the barrier beaches is variable but is less than 40 ft (12 m) in the immediate area of the study cross section. Appropriate adjustments are made in the vertical resistance for these changes in thickness. The Magothy and the upper glacial aquifers are simulated as anisotropic and homogeneous throughout.

The bottoms of surface-water bodies surrounding Long Island are simulated by bus (low-resistance) wires to represent surfaces of equal head (heavy, solid lines in cross section, fig. 2). These bus wires are connected to an external reference (zero) voltage. The bottom flow-line surface in these models, at the base of the Magothy aquifer, is represented by the bounding resistors. The upper boundary of the models between the shorelines receives almost uniformly distributed current input at each of the 40 model nodes (maximum variation about 10 percent), which simulates uniform constant recharge. The approximately equal current input is achieved by large input resistors (several megohms) in series with each model node (fig. 2).

Between the shorelines in the model network, the uppermost row of resistance elements consists of variable resistors. These variable resistors permit a trial-and-error modification of the voltage along the upper model boundary when required by the different types of tests that are described subsequently.

Four models were constructed. The major aquifers in each model have hydraulic conductivities that are different from those used in the other three models. Absolute values and the ratio of horizontal to vertical hydraulic conductivity for each model are listed in table 2. Variables investigated in the initial series of tests were (a) vertical hydraulic conductivity of the Gardiners Clay--four different values used in each model--and (b) average areal steady-state recharge--four to six different values ranging from 0.10 to 1.25 $(M\ gal/d)/mi^2$ [(5.4 to 67 cm/yr)] for each model. All existing estimates of average areal steady-state recharge before the advent of man's influence are somewhat less than the maximum value of model recharge studied and several times greater than the minimum value. The Lloyd aquifer was not simulated in this series of tests.

Table 2.--~~Values of~~ hydraulic ~~conductivity~~ conductivities for the hydrogeologic units[1] modeled in the four electric analog cross-section models

Aquifer	Model A			Model B			Model C			Model D		
	P_x[2]	P_z[3]	P_x/P_z	P_x	P_z	P_x/P_z	P_x	P_z	P_x/P_z	P_x	P_z	P_x/P_z
	(feet per day)[4]			(feet per day)			(feet per day)			(feet per day)		
Upper glacial	400	80	5	270	27	10	160	6.7	24	110	2.7	40
Magothy	80	8	10	54	1.8	30	40	6.7×10^{-1}	60	27	2.7×10^{-1}	100

1/ The vertical hydraulic conductivity of the Gardiners Clay was simulated by four different values (1.34×10^{-1}, 6.70×10^{-3}, 6.70×10^{-4}, and 6.70×10^{-5} ft/d) in each of the four models.

2/ P_x = horizontal hydraulic conductivity.

3/ P_z = vertical hydraulic conductivity.

4/ One ft/d is approximately 0.305 m/d.

Four different types of model tests are evaluated in this report:

1) Subsea flow limited by the interface; no simulation of streams [tests in which (a) the steady-state position of the model interface in the Magothy aquifer is determined and (b) the model is not modified to include the three-dimensional flow to streams in the shallow subsystems].

2) Subsea flow beyond the interface; no simulation of streams [tests in which (a) the position of the model interface is not determined and (b) the model is not modified to include the hydrologic effect of the streams].

3) Subsea flow limited by the interface; simulation of streams [tests in which (a) the steady-state position of the interface in the Magothy aquifer is determined and (b) the model is modified to include the hydrologic effect of the streams].

4) Subsea flow beyond the interface; simulation of streams [tests in which (a) the position of the interface is not determined and (b) the model is modified to include the hydrologic effect of the streams].

The rationale for tests of types 2 and 4 (subsea flow beyond the interface) is discussed subsequently. In these tests, the seaward boundary of the model is a streamline boundary at the extreme south edge of the network shown in figure 2.

The basic measurement in all types of tests is voltage (corresponding to head in the prototype) at selected nodes in the models. Before voltage is measured in a specific test, the models are modified to improve simulation of the boundaries and other conditions. The three principal model adjustments are (1) adjusting the variable resistors in the uppermost row of the resistance network to simulate the saturated thickness of aquifer above model sea level, (2) disconnecting appropriate resistors in the model Magothy aquifer to simulate the steady-state position of the interface, and (3) adjusting additional variable resistors that modify the voltages at the uppermost nodes near the model south shore to simulate the hydrologic effect of the streams.

The first adjustment was done for all tests. The variable horizontal resistors in the uppermost row of the network were adjusted (fig. 2) to represent transmissivity of the aquifer material lying between the model position of sea level and the water-table profile of the model.

The second adjustment, determining the steady-state position of the interface, was done for tests of types 1 and 3 (subsea flow limited by interface), as required by their definition. Enough resistors in the model of the Magothy aquifer were disconnected to equalize (a) the model head at the farthest seaward node at each depth in the model, and (b) the model fresh-water head equivalent to a static column of sea water above the point simulated by that node. Bennett and Giusti (1971) provide a thorough discussion of the theory, assumptions, and procedures of this method of modeling interfaces.

Modification to simulate the approximate steady-state position of the interface entailed inclusion of only the model Magothy aquifer and the model Gardiners Clay. Resistors simulating the upper glacial aquifer above the Gardiners Clay were not disconnected, partly because of the uncertainty as to whether the prototype interface is within the Gardiners Clay or at its upper contact, and partly because the voltage changes resulting from disconnecting the resistors simulating the upper glacial aquifer translate into insignificant head changes of 0.01 ft (3 mm) or less. As a practical matter, modification of the model to simulate an interface within the model Gardiners Clay was not possible because the clay's thickness is simulated by only one horizontal and one vertical resistor.

The third adjustment--partly adjusting the response of the model to account for the hydrologic effect of the streams--was made for tests of types 3 and 4 (simulation of streams) by adjusting the variable stream resistors northward from the model position of the south shoreline (fig. 2). By the adjustments, the voltage readings at these uppermost model nodes--corresponding to water-table elevations in the prototype--were lowered to voltages corresponding to water-table elevations slightly less than the minimum that would permit the streams adjacent to the study profile to flow. North-south water-table profiles on Long Island differ slightly in altitude near the south shore, depending on whether the profile is drawn along the local interstream divide (maximum possible altitude of water-table profile in the neighborhood) or immediately adjacent to the flowing part of a stream (minimum possible altitude of water-table profile in the neighborhood). The altitude of the streambed is only slightly lower than the minimum possible water-table altitude near a flowing stream because most south-shore streams are shallow, 1-2 ft deep (0.3-0.6 m) or less, and these streams are gaining streams. A generalized lowest topographic profile, based on streambed altitudes near the study cross section from available topographic maps (fig. 3), provided the basis for this voltage modification.

In tests of types 3 and 4 (simulation of streams), variable resistors were used to modify the water table by lowering it wherever the model water table was greater than the generalized lowest topographic profile. Thus, simulation of the hydrologic effects of the streams consists of superimposing a local one-dimensional resistance network above the existing two-dimensional network. The net effect of this trial-and-error adjustment is the interception of virtually all the input current along the stream's route. The intercepted input current is prevented from reaching the interior parts of the model network. Because of their short length and small discharge, the north-shore streams are not considered. No modification is made to the water-table profile near the north shore.

Voltage modification assured a reasonable lower limit of the water-table profile and a maximum current flow through the variable resistors for a given rate of recharge and a given set of hydraulic conductivities. Current flow through the variable resistor adjacent to the shoreline, translated into discharge, represents an upper limit of possible average streamflow per unit length of shoreline for the existing model conditions. In contrast to types 3 and 4 (simulation of streams), model measurements in tests of types 1 and 2 (no simulation of streams) represent an upper limit of the water-table profile for the existing model conditions because there is no adjustment for the stream's depressing effect on the ground-water level.

Voltage readings were recorded at selected nodes, and current was calculated to determine distribution and streamflow after the various model modifications were completed.

The first step in evaluating the test results was to determine the position of the voltage divide at the upper boundary of the model, which corresponds to the water-table divide in the prototype. Currents flowing to the north and south of the divide were calculated across resistors of the model network that were connected directly to ground. The sum of input current on each side of the divide provided a check of the calculations. Streamflow to the south shore is defined by the calculated current across the variable resistor nearest the south shore. The corresponding value of discharge in the prototype probably includes increased evapotranspiration in areas near the shorelines. Model discharge through the upper glacial aquifer at the south shoreline is defined by the current flow through the two horizontal resistors that simulate the upper glacial aquifer immediately landward of the south shoreline. Discharge through the Gardiners Clay is defined by the current flowing through all vertical resistors that simulate the Gardiners Clay south of the south shoreline. The resistor at the shoreline is included. The test results relating to discharge of current through different parts of the models are reported as a series of discharge ratios. The numerator of the ratio is the appropriate model current; the denominator is the total model current for a particular test.

In addition to the distribution of model heads and discharges, the model area occupied by fresh ground water between simulated steady-state positions of the interfaces was carefully determined for each test. These results are reported subsequently as a ratio called the storage ratio. The numerator of the ratio is the model area occupied by fresh ground water; the denominator is approximately the model area enclosed between sea level, the base of the Magothy aquifer, and two vertical lines, one at the north shore and one at the south shore. This area is outlined by a dotted line in figure 2 and is designated "limit of fresh-water storage reference area."

Results

Pertinent data obtained from tests with the four cross-section analog models are listed in tables 3 to 12. These data include (1) model heads and model head differences at key locations, (2) location of the toe of the model steady-state interface in the Magothy aquifer and related volume of the model ground-water reservoir saturated with freshwater, and (3) distribution of model current outflow to boundaries, particularly near the south shoreline, reported as a series of ratios. One purpose in presenting the model data is to allow comparison with observations in the real system. Where no prototype data are available or where these data are highly uncertain, the variation in model data indicates the effect of varying the model hydraulic conductivity on the model parameter under consideration.

The range in maximum model head for tests of types 3 and 4 (simulation of streams) is considerably less than for tests of types 1 and 2 (no simulation of streams). At a recharge rate of 1.0 (M gal/d)/mi^2 [(54 cm/yr)], the maximum head among the different models ranges from 61 to 131 ft (19 to 39 m), as the hydraulic conductivity values assigned to the upper glacial and Magothy aquifers are varied from maximum to minimum; changing the vertical hydraulic conductivity of the Gardiners Clay in different runs of the same model causes little variation in maximum head. Each increment of reduction of the vertical hydraulic conductivity of the Gardiners Clay was accompanied by adjustment of the variable resistors at the top of the models; greater voltage reductions near the south shore than those inland were required to maintain the voltages at the top of the model below the low-voltage profile. These adjustments tended to compensate for the reduction in vertical hydraulic conductivity of the Gardiners Clay, so that approximately the same maximum head was maintained throughout each series of tests.

Four types of tests were made on the cross-sectional analog models. Tests of types 1 and 2 (no simulation of streams) provide an upper limit for the position of the water table and heads within the ground-water reservoir for a given set of model parameters. In these tests, the applied boundary conditions cause the water-table slope to be greatest near the shorelines. Although this theoretical condition approximately describes the water-table profile near the north shoreline (fig. 3), it

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Figure 3.--(caption on next page) belongs near here.

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does not apply near the south shoreline. Model data from tests of types 3 and 4 (simulation of streams) compare more closely with prototype data than data from model tests without this modification, even though the modification overcompensates for the effect of the streams. The prototyp low-topography profile and two model water-table profiles exhibiting relationships just described are also shown in figure 3.

Figure 3.--Generalized prototype water-table profile based on ground-water levels in October 1961 and selected model water-table profiles along section A-A'.

In Model A (highest model values of hydraulic conductivity) for tests of types 3 and 4 (simulation of streams), the maximum head (table 3) is definitely too low for any reasonable average recharge in the prototype. In this context, a reasonable average recharge is between 0.75 and 1.25 (M gal/d)/mi^2 [(40 and 67 cm/yr)]. In Model D (lowest model values of hyraulic conductivity), for recharge of 1.0 (M gal/d)mi^2 [(54 cm/yr)], the maximum head in corresponding tests is about 130 ft (39 m). Even if a value of average recharge considerably less than 1.0 (M gal/d)/mi^2 [(54 cm/yr)] is used in Model D, the maximum head would be considerably greater than that in the prototype. For example, the maximum head would be about 100 ft (30 m) for an average recharge of about 0.75 (M gal/d)/mi^2 [(40 cm/yr)]. In conclusion, according to this one criterion, the model tests suggest that prototype hydraulic conductivities are less than those of Model A but greater than those of Model. D.

The nonlinearity in the response of maximum head to changes in recharge is shown in figure 4, using data from table 3. This nonlinear-

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Figure 4 (caption on next page) belongs near here.

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ity is due to a change in the position of two model boundaries in response to a decrease in model recharge--(1) the interface, which moves landward as recharge decreases, and (2) the water table, which declines in altitude as recharge decreases. The decline in altitude reduces the conducting capacity of the upper part of the model.

Figure 4.--The nonlinear relation between average areal recharge and maximum model head for a selected series of model tests.

The model water-table profiles in figure 3 are considerably lower than the prototype profile near the north shoreline, although the elevation of the model ground-water divide is considerably higher than that of the prototype divide. This condition exists for other model tests not shown in figure 3, except tests on Model D (lowest hydraulic conductivities) of type 2 (no simulation of streams), in which recharge is 1.0 (M gal/d)/mi^2 [(54 cm/yr)]. This probably indicates that the model vertical and horizontal hydraulic conductivities of the upper glacial aquifer near the north shoreline are too large compared to prototype values.

The general location of the head along the principal ground-water divide on Long Island is well known. The head at the ground-water divide in the area of the model cross section is between 80 and 90 ft (24 and 27 m), as given in published water-table maps of the last 30 years. See Kimmel (1971). Location of the ground-water divide can vary in time, but it is definitely north of the center of the island. An average location for comparison with model data is 30,000 to 32,000 ft (9,100 to 9,800 m) south of the northern shoreline on the model cross section.

The model data in tables 3 and 4 show considerable ranges in both maximum head and location of the divide.

As the vertical conductivity of the Gardiners Clay was changed through several orders of magnitude and all other parameters were held constant, each model in tests of types 1 and 2 (no simulation of streams) showed a chnage in maximum head that varied through tens of feet. Moreover, in tests of types 1 and 2 in which the horizontal and vertical hydraulic conductivities of the upper glacial and Magothy aquifers were varied while model recharge and the vertical hydraulic conductivity of the Gardiners Clay were held constant, variations in maximum head were even greater than in tests in which the vertical hydraulic conductivity of the Gardiners Clay was varied.

Locations of the water-table divide range from 28,000 to 44,000 ft (8,500 to 13,400 m) south of the north shoreline in the model tests according to data in table 4. Positions of the divide significantly north of the center of the island, as in the prototype, were obtained only in tests of types 3 and 4 (simulation of streams). In considering tests in which the average recharge is constant at 1.0 (M gal/d)/mi^2 [(54 cm/yr)], the divide clearly moves south in tests of types 1 and 2 (no simulation of streams) in response to a decrease in the vertical hydraulic conductivity of the Gardiners Clay. The divide clearly moves north in tests of types 3 and 4 (simulation of streams) for successive tests in Models A through D in which the vertical hydraulic conductivity of the Gardiners Clay is held constant.

Difference between the maximum head and the head at the base of the Magothy aquifer vertically downward can only be estimated in the prototype from existing data because this difference is small and because only a few wells are available for measurement, particularly in the basal Magothy. Kimmel (1971) provided the first published maps of the water table and the potentiometric surface in the basal Magothy that are based on simultaneous head measurements. Kimmel's map (1971, fig. 3), based on water-level measurements in 1959 and a review of the basic data from which the map was prepared, suggests a head difference of approximately 1 ft (0.3 m) at the divide in the area of the model cross section. A head difference of 2 ft (0.6 m) is probably the upper limit for the prototype.

Head differences obtained from model tests (table 5) show minor inconsistencies that can be attributed to instrument drift and other experimental errors. However, trends and ranges in values associated with specific models are clear and consistent. For a given model, the data usually indicate a nonlinear relation between average recharge and the tabulated head difference. Entries in table 5 for any one model and identical average recharge show little effect when the vertical hydraulic conductivity of the Gardiners Clay is varied. (See particularly tests for which average recharge is equal to 1.0 (M gal/d)/mi^2 [(54 cm/yr)]). For average model recharge of, and also considerably less than, 1.0 (M gal/d)/mi^2 [(54 cm/yr)], the head differences in table 5 for Model A are apparently too small (approximately 0.2 ft or 6 cm) and for Model D are too large (approximately 4 ft or 1.2 m) compared with the prototype. Hydraulic conductivities intermediate to those in Models A and D, at least in the vertical section near the ground-water divide, seem appropriate to simulate this prototype head difference.

Under natural conditions, 15 ft (4.6 m) is probably a reasonable value for the prototype head in the basal Magothy aquifer at the south shoreline with which to compare model data in table 6 (Kimmel, 1971, fig. 3). This value, which was obtained from Kimmel's map representing the potentiometric surface in the basal Magothy in 1959, may have been affected by pumping. The effect is probably small because man's net withdrawals of ground water in the adjacent area have not been appreciable. Moreover, a comparison of Magothy heads at this particular location in 1959 and 1970 (Kimmel, 1971) reveals no marked change, although this was a period of rapid urban development in nearby eastern Nassau and southwestern Suffolk Counties.

The relationship between the head in the basal Magothy at the south shoreline and the position of the interface poses a difficult hydrologic question. Glover (1959) describes the freshwater flow pattern near a salt water boundary if the salt water is assumed to be static, and Cooper (1959) presents a flow pattern in which the salt water circulates. Both references indicate that a stable interface cannot exist unless the toe of the salt-water wedge is inland of the shoreline. This is also true of the layered-aquifer solution presented by Rumer and Shiau (1968).

A study of heads in the Magothy aquifer indicates that if the interface were now in an approximately steady-state position that agrees with present-day heads, the toe of the saltwater wedge in the Magothy aquifer would be near and probably inland of the south shoreline of Long Island at the model cross section (fig. 5). Available information indicates that the toe of the Magothy saltwater wedge is south of the barrier beaches along most of the south shore of Suffolk County. The interface seems to be transient and to be moving landward without man's influence. This general conclusion agrees with that of Perlmutter and Geraghty (1963, p. 96-105) in their studies of southern Nassau and southeastern Queens Counties. However, Collins and Gelhar (1971) were unable to resolve observation with theory in their determination of the position of the toe of the salt-water wedge in a Hele-Shaw model; theory predicted a more inland position of the saltwater than that indicated by the Long Island prototype.

An estimate of the steady-state position of the interface at the model cross section is shown in figure 5. The estimate is based on

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Figure 5 (caption on next page) belongs near here.

Magothy heads near the south shoreline in 1959 (Kimmel, 1971, fig. 3) and on the assumption that the Long Island aquifer materials, particularly clays, are chemically inert to the flow of either fresh or salty ground water through them; that is, the steady-state position of the interface is dependent solely on prototype hydraulic conductivities and geometry and prototype heads. Because the effects of man seem insufficient to have caused the marked unbalance between prototype freshwater heads and the present seaward position of the interface in Suffolk County, the authors attribute this unbalance to a rise in sea level caused by the melting of Pleistocene glaciers. The head can adjust to the changing boundary conditions (rising sea level) in a matter of decades, but diffusion of enough salt water into the confined aquifer to balance the head there might require thousands of years.

Figure 5.--Estimated steady-state position of the interface between fresh ground water and salty ground water in the Magothy aquifer at the south shore near cross section A-A', as inferred from freshwater heads in 1959.

Freshwater circulation beyond the theoretical steady-state interface is possible; this type of circulation provides the rationale for tests of types 2 and 4 (subsea flow beyond the interface). By not disconnecting part of the model network to locate the theoretical steady-state interface it was possible to make head measurements in the model seaward from the south shoreline. These measurements may provide a clue to heads, or at least the distribution of heads, at similar locations in the prototype. If the model parameters matched the corresponding prototype paramaters, the heads measured seaward of the south shoreline in a model test in which the interface was not formed would be larger than the corresponding prototype heads under present transient conditions.

In table 6, no freshwater heads at the shoreline are recorded for tests of type 3 (subsea flow limited by interface and simulation of streams) because the toe of the interface is landward of the shoreline. In tests of types 1 and 2 (no simulation of streams), most of the head values are markedly higher than those for the prototype. The heads are higher not only in the model near the shoreline, but also throughout the model.

Magothy heads at the south shoreline (table 6) in tests of type 4 (subsea flow beyond interface and simulation of streams), which probably best simulate other prototype conditions, are less than 15 ft (4.6 m) and in most cases are less than 10 ft (3 m). Several factors probably influence this situation. Voltage modification at the upper boundary of the model, which compensates for the hydrologic effect of the streams, is extreme. As a result, heads throughout the model are less than those for an exact model simulation. The assumption that the Magothy aquifer is anisotropic and homogeneous may be a significant cause of model inaccuracy. In general, geologic logs indicate more silt and clay in sections near the south shoreline than in sections near the center of the island. This condition would tend to decrease the vertical hydraulic conductivity of the Magothy at the shoreline relative to the center of the island and thus to increase heads in the basal Magothy near the shoreline. In addition, presence of the Lloyd aquifer in the model tends to increase heads slightly in the basal Magothy near the shoreline relative to heads at the same location without the Lloyd aquifer. These factors affect all measures of model response, not just the specific data under discussion.

Because the prototype interface is probably in a transient state, there are no prototype data on the location of the toe of the saltwater wedge under steady-state conditions near the south shore. Thus, there is no measure of the size of the prototype ground-water reservoir that can be compared with the model data in tables 7 and 8. However, these model data provide valuable insight into the effect of model hydraulic conductivities on the model interface. Also, these model data indicate possible prototype response to different steady-state conditions of average recharge. For tests of type 3 (simulation of streams) and average model recharge of 1.0 (M gal/d)/mi^2 [(54 cm/yr)], location of the toe of the interface ranges from only -12,000 to -15,000 ft (-3,700 to -4,600 m) from the south shoreline for a wide range of values of model hydraulic conductivity in the principal aquifers and in the Gardiners Clay. Two conclusions are suggested by these observations: (1) Boundary configuration (island shape and size, aquifer thickness, location of confining units, and streams) and recharge rate influence the steady-state position of the interface more than aquifer conductivity does; and (2) boundary configuration is a more important influence than recharge rate.

Storage ratios in table 8, particularly for Model A (highest model values of hydraulic conductivity), have potential application in water management. Storage ratios are obtained by dividing the model area occupied by fresh ground water in a given test by the area between model shorelines. (See dotted line on fig. 2). For Model A, the ratios are the minimum values of all four models for a given average model recharge. More important, however, assuming that the hydraulic conductivities in Model A are considerably greater than prototype conductivities, sotrage ratios in the prototype would certainly be greater than those in Model A for the same values of average recharge. The lower conductivities in the prototype result in higher heads and, consequently, a more seaward position of the interface than in Model A. This means that for a very low average net recharge in the prototype, a significant volume of fresh ground water would remain in storage in the area of the model cross section after the interface attained equilibrium. Glover (1959) reached the same general conclusion for a hypothetical coastal aquifer. For example, for a net average prototype recharge of 0.10 $(\mathrm{M\ gal/d})/\mathrm{mi}^2$ [(5.4 cm/yr)], the prototype storage ratio in the neighborhood of the model cross section would be greater tahn 0.61 (table 8) and possibly much greater. Of course, an unknown but certainly large volume of fresh ground water exists between the present transient position of the interface in the prototype and any steady-state interface position that even remotely corresponds to present recharge and head conditions.

Figure 6 shows the most landward interface positions for each value

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Figure 6.--(caption on next page) belongs near here.

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of average recharge. These interface positions are those corresponding to the steady-state interfaces in tests with Model A in which the vertical hydraulic conductivity of the Gardiners Clay is 1.34×10^{-1} ft/d (4.03×10^{-1} m/d).

Evaluation of the model discharge fractions (ratio of model discharge at specified locations to total model discharge) in tables 9 to 12 cannot be accurately evaluated in the prototype. These fractions should be interpreted as high or low extremes compared with the prototype if they are derived from a model whose hydraulic conductivities and average recharges match those of the prototype reasonably well. The one possible exception to this statement is the north-shore-discharge fraction in table 9 (ratio of model discharge to the north shore to total model discharge). Assuming areally uniform recharge, this ratio is directly related to the position of the water-table divide. If the water-table divide is 31,000 ft (9,400 m) south of the north shore line (fig. 3), a reasonable ratio in the prototype is 0.35 to 0.40. Neither assumption used in calculating this ratio is strictly valid for the prototype.

Figure 6.--Steady-state position of the interface between fresh ground water and salty ground water in a series of tests with highest model values of hydraulic conductivity.

As indicated previously, only in tests of types 3 and 4 (simulation of streams) is the position of the model water-table divide close to the position of the prototype divide. Consequently, the north-shore discharge fractions in table 9 will only compare reasonably well with the prototype for tests in which streams are simulated. In tests of types 1 and 2 (no simulation of streams), discharge fractions in table 9 are considerably larger than a reasonable prototype fraction. This discharge fraction is probably not sensitive enough, and knowledge of this fraction in the prototype is too uncertain to warrant selecting certain tests of types 3 and 4 as possibly better simulating the prototype than other tests.

South-shore-shallow-subsurface-discharge fractions (ratio of model discharge through the upper glacial aquifer at the south shoreline to total model discharge) in table 10 are much too large compared with the prototype in tests of types 1 and 2 (no simulation of streams) and somewhat too small in tests of types 3 and 4 (simulation of streams). This situation results because the voltage drops across resistors at the top of the model glacial aquifer near the shoreline are considerably larger in tests of types 1 and 2 and smaller in types 3 and 4 than the corresponding head drops in the prototype. The glacial-subsurface-discharge fraction in the prototype near the model cross section on the south shore would probably be lower than average for the south shore of Nassau and Suffolk counties because the upper glacial aquifer near the shoreline thickens east of the model cross section. Based on water-budget figures provided by Franke and McClymonds (1972, tables 10 and 11), which are consistent with an average recharge of about 1 (M gal/d)/mi^2 [(50 cm/yr)], this fraction is about 0.05. Because of inadequate field data, however, this estimate of the south-shore-shallow-subsurface-~~glacial~~-discharge fraction for the prototype is uncertain. The fraction in tests of types 3 and 4 (table 10) is generally less than 0.05. The glacial-subsurface-discharge fraction of the south shore is not a sufficient criterion for selecting tests that best simulate the prototype.

The Gardiners Clay-discharge fractions in table 11 (ratio of model discharge through the Gardiners Clay seaward of the south shoreline to total model discharge) represent the maximum value for each model condition of recharge, heads beneath the clay, and the model vertical hydraulic conductivity of the clay. These model fractions would correspond reasonably well with prototype discharge for the same head and aquifer conditions only if the fresh ground water in the prototype discharged vertically through the Gardiners Clay. Any other pattern of flow through the clay would result in a flow path of greater length and resistance and, therefore, a smaller discharge fraction than the pattern just described.

A second reason the discharge fractions in table 11 represent maximum values for the specific model conditions is that the model does not recognize that the surface water overlying the seabed is more dense than freshwater. The entire seabed is modeled as a surface at zero head; in other words, the seabed is assumed to be overlain by freshwater. The head at the seabed in the prototype is indeed zero at the shoreline, but the head increases with increasing depth of saltwater. Where the bay and ocean are shallow, the saltwater head has negligible effect on the discharge of water through the Gardiners Clay. Farther seaward, where the saltwater is deeper, the saltwater head may have considerable effect on the quantity of water being discharged through the Gardiners Clay.

Offsetting the factors that make the model discharge through the Gardiners Clay too large with respect to the prototype is the fact that, in tests of types 3 and 4 (simulation of streams), heads at the top of the model are too small compared with those of the prototype. As a result, heads throughout the model, including heads at the base of the Gardiners Clay, are similarly affected to a slight degree. Reduced heads at the base of the Gardiners Clay result in a reduced model discharge through the Gardiners Clay.

No prototype data on discharge through the Gardiners Clay are available for comparison with model data in table 11. The necessary hydraulic head data have not been collected south of the shoreline, and no data from which to derive the vertical hydraulic conductivity of the Gardiners Clay at even one location are available. Although the model discharge through the clay is probably too large, the data in table 11 indicate that this discharge is small relative to the total model discharge. Probably less than 10 percent of the total input to the ground-water reservoir near the study cross section discharges through the Gardiners Clay directly from the Magothy aquifer if (1) the Gardiners Clay is continuous for several miles south of the shoreline, (2) a reasonable value of vertical hydraulic conductivity for a marine clay is chosen for the Gardiners Clay (probably less than 1.0×10^{-2} ft/d or 3×10^{-3} m/d), and (3) only hydraulic gradients are moving fresh ground water through the Gardiners Clay. Furthermore, for tests of types 3 and 4 (simulation of streams), the model head measurements indicate that 90 percent of the model discharge through the Gardiners Clay is within 2 to 3 mi (3 to 5 km) of the south shoreline.

Stream-discharge fractions (ratios of model discharge through south-shore streams to total model discharge) in table 12 are upper limits for the given model parameters. As described previously, model heads near the south shoreline were modified by potentiometers to values that are slightly lower than corresponding prototype heads that would permit a stream to flow. This modification slightly increases the model discharge through the stream potentiometers and decreases, correspondingly, the model discharge through the upper glacial aquifer and the Gardiners Clay compared with what these discharges would be without the head modification.

The prototype stream-discharge fraction, considering only streamflow to the south shore, is approximately 0.35. This determination is based on water-budget data provided by Franke and McClymonds (1972, tables 10 and 11), which are consistent with average recharge of 1 (M gal/d)/mi^2 [(50 cm/yr)]. As the average recharge decreases to less than 1 (M gal/d)/mi^2 [(50 cm/yr)], the fraction also decreases. Values of the fraction in table 12 are greater than 0.48 in all models with average recharge greater than 0.5 (M gal/d)/mi^2 [(27 cm/yr)]. Furthermore, for constant average recharge and hydraulic conductivity of Gardiners Clay, the fractions in table 12 increase from Model A through Model D. This increase shows the increasing head adjustments required to match the minimum-head profile in models with successively lower hydraulic conductivities. An increased head adjustment by the potentiometers must result in increased simulated streamflow.

MODEL TESTS THAT INCLUDE SIMULATION OF THE LLOYD AQUIFER

In addition to the tests from which the data in tables 3-12 were derived, 12 tests, divided evenly between Models B and C, included a simulation of the Raritan clay and the Lloyd aquifer. The Raritan clay was simulated by a single line of vertical resistors connected to nodes at the bottom of the Magothy aquifer. The thickness of the clay in the model was 200 ft (60 m) for the entire cross section except the southernmost 40,000 ft (12,000 m), where its thickness was 400 ft (120 m).

The vertical hydraulic conductivity of the Raritan clay is virtually unknown. In a recent paper, Franke and Cohen (1972, table 1) assumed it to be 1×10^{-3} ft/d (3×10^{-4} m/d). Because of the uncertainty of the prototype vertical conductivity, a large range of model values, from 1.34×10^{-1} to 1.34×10^{3} ft/d (4×10^{-2} to 4×10^{-4} m/d) was used to investigate the effect of varying the vertical hydraulic conductivity of the Raritan clay on model heads above the Raritan clay.

A single row of horizontal resistors connected to the bottom ends of the vertical resistors representing the Raritan clay simulated the Lloyd aquifer. The model transmissivity assumed for the Lloyd aquifer was a uniform 1.6×10^4 ft^2/d (1.5×10^3 m^2/d) for the entire cross section in all tests. This transmissivity approximates, or perhaps slightly exceeds, the value shown for the area of the model cross section on an islandwide map showing estimated transmissivity of the Lloyd aquifer prepared by McClymonds and Franke (1972, plate 3).

The only reliable prototype data with which to compare the model results are head values in the Lloyd aquifer and head drops across the Raritan clay. A value of particular interest is the head drop across the Raritan clay near the ground-water divide. The best available information indicates that this head drop is about 40 ft (12 m) in the area of the model cross section.

Another item of interest is the proportion of ground-water recharge that moves through the Raritan-Lloyd hydrogeologic units. This proportion is about 5 percent, according to water-budget data in Franke and McClymonds (1972, tables 10 and 11), for the entire water-budget area and is consistent with a recharge of 1 (M gal/d)/mi^2 [(50 cm/yr)]. However, the water-budget numbers from which this proportion is estimated may be questionable because of uncertainties in estimated values of recharge to and subsurface discharge from the Long Island ground-water reservoir.

Table 13 lists data from tests that included a simulation of the Lloyd aquifer and comparison data from tests in which the Lloyd aquifer was not modeled. All 12 tests that included the Lloyd aquifer are of types 2 and 4 (subsea flow beyond the interface), but tests of type 4 (subsea flow beyond the interface and simulation of streams) probably best simulate the prototype. In all tests that included the Lloyd aquifer, model recharge was 1.0 (M gal/d)/mi^2 [(54 cm/yr)].

The Lloyd aquifer in either the model or the prototype decreases the resistance of any model or geologic section to the flow of electricity or water, respectively, compared to a cross section without the Lloyd aquifer. Comparison of tests with and without the Lloyd aquifer for the same average recharge and the same hydraulic conductivity should show lower maximum head as well as lower heads at other locations in the Magothy and upper glacial aquifers when the Lloyd aquifer is included. Differences in maximum model head in tests with and without the Lloyd aquifer (table 13) support this simple principle.

When the vertical hydraulic conductivity of the Raritan clay is 1.34×10^{-1} ft/d (4×10^{-2} m/d), inclusion of the Lloyd aquifer reduces the maximum head in tests of type 4 by about 20 ft (6 m) compared with model tests in which the Lloyd was absent. In contrast, when the vertical hydraulic conductivity of the Raritan clay is 1.34×10^{-3} ft/d (4×10^{-4} m/d), the corresponding reduction is 2 to 3 ft (0.6 to 1 m).

The proportion of total model recharge that moves through the Lloyd aquifer ranges from about 0.02 for lowest to 0.24 for highest model vertical hydraulic conductivities of the Raritan clay. Those model tests in the lower third of table 13, in which the addition of the Lloyd aquifer has the smallest effect on the water table and for which flow through the Lloyd aquifer is small relative to total recharge, also result in a maximum model head drop of 40 to 45 ft (12 to 14 m) across the Raritan clay, which compares well with head drops near the location of the model cross section observed in the prototype.

Addition of the Lloyd aquifer reduces model heads to some extent in most of the overlying ground-water reservoir. Seaward of a transition node near the south shoreline, model heads at the bottom of the Magothy aquifer are slightly higher when the Lloyd aquifer is present than when it is absent. The position of this transition node varies from test to test. The maximum absolute value of these head increases ranges from several tenths of a foot to almost 4 ft (0.1 to 1.3 m) in tests of type 4. In several tests, increases in head that are small in absolute value represent increases of more than 100 percent. Slightly higher model heads in the basal Magothy aquifer south of the shoreline mean higher model heads throughout the Magothy aquifer in this area and, thus, a slight increase in the model freshwater discharge through the Gardiners Clay. Clearly, the Raritan clay functions as any confining layer by tending to conserve or very slowly dissipate the head in the Lloyd aquifer. Flow through the Lloyd aquifer creates higher heads at the top of the confining layer (base of Magothy aquifer) south of the shoreline than would exist without the combination of the confining layer and the lower aquifer.

SUMMARY AND CONCLUSIONS

Most of the conclusions of this report are based on the model-test hydraulic conductivities and ranges of hydraulic conductivity that seemed to correspond best with prototype data for the principal hydrogeologic units on Long Island. This information was a starting point in the design of a large islandwide five-layer electric-analog model of the Long Island ground-water reservoir built by the Geological Survey and must be considered in future modeling on Long Island.

(1) The range of changes in the simulated hydraulic conductivities for the upper glacial and Magothy aquifers produced marked changes in the water-table profiles and heads within the model ground-water reservoir. Average values of hydraulic conductivity for most of the prototype reservoir are probably close to the range of values in the two intermediate models (table 2, Models B and C). The horizontal hydraulic conductivity of the upper glacial aquifer was 270 and 160 ft/d (82 and 49 m/d) and of the Magothy aquifer 54 and 40 ft/d (16 and 12 m/d), in models B and C, respectively. In general, comparison of the response of the different models suggests that changes in average hydraulic conductivity of the upper glacial and Magothy aquifers of as much as 20 to 25 percent in any given model would not greatly affect the response of that model. Furthermore, the response of the models indicated that modeling the two uppermost aquifers as anisotopic and homogeneous was not an adequate representation, particularly near and seaward of the shorelines.

Vertical hydraulic conductivity of the Gardiners Clay, a thin clay of low hydraulic conductivity at the base of the upper glacial aquifer that occurs near the south shoreline, also proved to be a significant influence on model heads and on the distribution of model outflow to the south shore. Although this study cannot provide a final answer, a vertical hydraulic conductivity of less than 10^{-2} ft/d (3×10^{-3} m/d) for the Gardiners Clay gives reasonable model results. Despite the uncertainty in assigning a vertical hydraulic conductivity to the Gardiners Clay, the model results clearly indicated that probably less than 10 percent of recharge to the prototype hydrologic system in the area of the study cross section discharges from the Magothy aquifer through the Gardiners Clay seaward of the south shoreline.

(2) Accurate assessment of the hydrologic effect of the streams in any regional modeling of the Long Island ground-water reservoir is important. Simulation of streams in the models greatly modifies (a) water-table profile, (b) position of the water-table divide, (c) heads within the ground-water reservoir, (d) amount and location of ground-water discharge into adjacent salt-water bodies, and (e) equilibrium position of the interfaces.

(3) Heads measured in the upper aquifers between the north and south shorelines were lower in models that simulated the Lloyd aquifer than in models that did not. The maximum head differences between tests with and without the model Lloyd aquifer, which is at the divide, varied with the simulated vertical hydraulic conductivity of the Raritan clay. For models in which this vertical conductivity was simulated as approximately 10^{-3} ft/d (3×10^{-4} m/d), this maximum head difference was about 2 ft (0.6 m); for models in which the vertical conductivity was simulated as approximately 10^{-1} ft/d (3×10^{-2} m/d), the maximum head difference was about 20 ft (6 m). When the model vertical hydraulic conductivity of the Raritan clay was about 10^{-3} ft/d (3×10^{-4} m/d), the difference in model head between the base of the Magothy aquifer and the base of the Raritan clay was close to the corresponding head difference in the prototype (about 40 ft or 12 m). In contrast to model heads between the shorelines, model heads seaward of the south shoreline are slightly higher in models containing the Lloyd aquifer than in corresponding models without the Lloyd aquifer.

(4) Fresh-water heads in the prototype indicate that the interface in the Magothy aquifer south of western Suffolk County was probably in a transient condition before the advent of man's influence. This conclusion is based on the assumption that man's influence on fresh-water heads near the study cross section has been very small. Water-level records support this assumption. In fact, head changes during the period of pumping have been small. The equilibrium position of the toe of the interface corresponding to present-day freshwater heads would probably lie near the south shoreline whereas it actually is located several miles south of the south shoreline.

(5) A series of tests in which model hydraulic conductivities for the principal aquifers and for the Gardiners Clay are much higher than corresponding prototype conductivities, indicates that considerable quantities of fresh ground water remain in storage near the study cross section between the north and south shores, assuming steady-state equilibrium positions of the interfaces, for average areal recharge rates as low as 0.10 (M gal/d)/mi^2 [(5.4 cm/yr)]. Because all previous estimates of average areal recharge for Long Island before the advent of man's influence are several times greater than 0.10 (M gal/d)mi^2 [(5.4 cm/yr)], this result indicates that a considerable quantity of fresh ground water will remain in storage after the ground-water system has achieved equilibrium in response to a sizeable man-induced stress.

(6) The set of model values of the hydrologic parameters that resulted in a model response which compared reasonably well with the prototype probably provides an adequate representation of the prototype for most purposes. No combination of parameter values provided an exact representation of the prototype, but combinations of nonhomogeneous aquifers and nonuniform recharge were not considered. The best combination of parameters that was tried is:

Upper glacial aquifer--P_x = 160 to 270 ft/d (49 to 82 m/d)

P_z = 6.7 to 27 ft/d (2 to 8 m/d)

Magothy aquifer--P_x = 40 to 54 ft/d (12 to 16 m/d)

P_z = 0.67 to 1.8 ft/d (0.2 to 0.55 m/d)

Gardiners Clay--P_z = 10^{-2} ft/d (3×10^{-3} m/d)

Raritan clay--P_z = 10^{-3} ft/d (3×10^{-4} m/d)

Average recharge rate--0.75 to 1.0 (Mgal/d)/mi^2 [(40 to 54 cm/yr)]

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Figure 1.--Index map of Long Island, N.Y., showing the regional ground-water divide and location of section A-A'.

Fig 1
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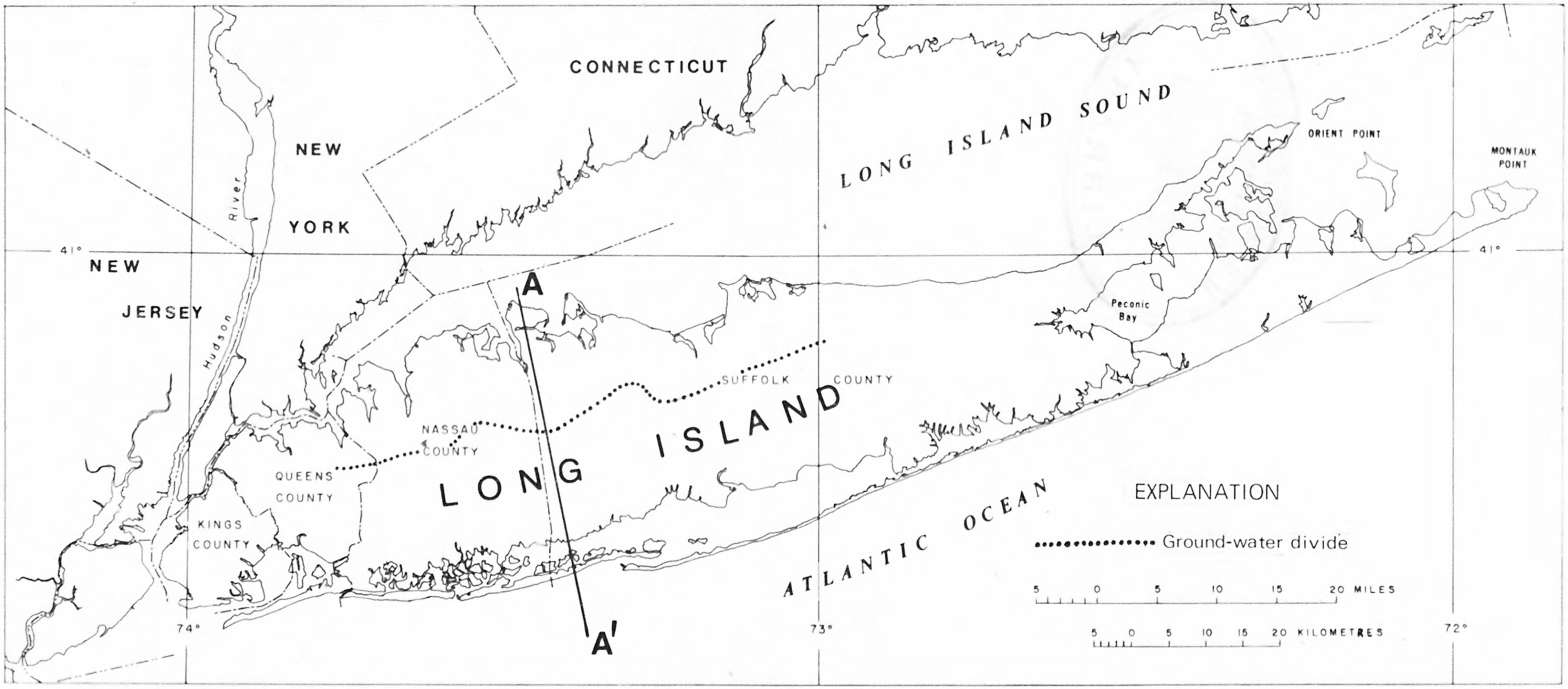

Figure 2.--Finite-difference network and electrical representation of boundaries for the cross-section analog models.

Figure 3.--Generalized prototype water-table profile based on ground-water levels in October 1961 and selected model water-table profiles along section A-A'.

Figure 4.--The nonlinear relation between average areal recharge and maximum model head for a selected series of model tests.

Figure 5.--Estimated steady-state position of the interface between fresh ground water and salty ground water in the Magothy aquifer at the south shore near cross section A-A', as inferred from freshwater heads in 1959.

Figure 6.--Steady-state position of the interface between fresh ground water and salty ground water in a series of tests with highest model values of hydraulic conductivity.