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

SIMULATION OF THE EFFECTS OF DEVELOPMENT OF THE GROUND-WATER FLOW SYSTEM OF LONG ISLAND, NEW YORK

Water-Resources Investigations Report 98-4069

Prepared in cooperation with the
NASSAU COUNTY DEPARTMENT OF PUBLIC WORKS,
SUFFOLK COUNTY DEPARTMENT OF HEALTH SERVICES,
SUFFOLK COUNTY WATER AUTHORITY, and the
NEW YORK CITY DEPARTMENT OF ENVIRONMENTAL PROTECTION

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hydraulic conductivity of any aquifer on Long Island. The hydraulic conductivity of the Magothy aquifer varies with depth; values for the upper part range from 35 ft/d to 90 ft/d; values for the coarser, basal zone were estimated to be about 50 percent higher. Hydraulic conductivity of the Lloyd aquifer ranges from 30 ft/d to 80 ft/d and generally is greatest in Nassau County. The anisotropy of these aquifers is estimated to be 100:1 because of their highly stratified character.

Although data on hydraulic conductivity of the confining units are scant, the high clay and silt content indicates values several orders of magnitude lower than those of adjacent aquifers. Franke and Cohen (1972) estimated the average vertical hydraulic conductivity of the confining units to be 0.001 ft/d; Reilly and others (1983) estimated a value of 0.0029 ft/d for the Gardiners Clay. The vertical hydraulic conductivity values of the major confining units used in this analysis are Gardiners Clay, 0.004 ft/d, Port Washington confining unit, 0.0015 ft/d, and Raritan confining unit, 0.0012 ft/d.

Estimates of specific yield for the glacial outwash deposits are 0.18 (Getzen, 1977), 0.22 (Reilly and Buxton, 1985), 0.24 (Warren and others, 1968), 0.24 (Perlmutter and Geraghty, 1963), and 0.30 (Franke and Cohen, 1972). Estimates as low as 0.10 have been proposed for morainal deposits (Getzen, 1977), and estimates for unconfined parts of the Magothy aquifer have been as low as 0.10 (Getzen, 1977; Reilly and Buxton, 1985). Specific yield values for the water-table model layer are shown in figure 10. Specific yield of the upper glacial outwash is 0.30; of the moraine deposits is 0.25; and of the Magothy deposits is 0.15. Storage coefficients for confined aquifers were calculated from aquifer thickness and a specific storage of 6.0×10^{-7}/ft (Getzen, 1977). This value of specific storage is at the minimum extreme; the authors suggest that future analyses use values close to 1.3×10^{-6}/ft, as calculated by Jacob (1941).

PREDEVELOPMENT HYDROLOGIC CONDITIONS (PRE-1900)

Before development, the Long Island ground-water system was in a state of dynamic equilibrium. Ground-water levels and rates of discharge to the ocean, streams, and springs, underwent natural fluctuations in response to natural fluctuations in recharge from precipitation. Despite short-term fluctuations in recharge and discharge, these budget components were in balance over the long term.

This section describes an average predevelopment (pre-1900) hydrologic condition that forms a basis for comparison with subsequent conditions. The predevelopment condition is based on the earliest available hydrologic data, and on results of a steady-state simulation made with the islandwide model. This section also describes (1) the natural hydrologic boundaries and their operation; (2) the system's ground-water budget, as estimated from field measurements and model-generated flow rates, and (3) general patterns of ground-water movement, as indicated by measured and simulated ground-water levels.

Hydrologic Boundaries

The body of fresh ground water beneath Long Island is enclosed by natural hydrologic boundaries (fig. 11). The upper boundary is the water table and the many surface water bodies that intersect it. The lower boundary is consolidated bedrock. The lateral boundaries consist of the saline ground water and saline surface-water bodies that surround the island. Under natural (non-pumping) conditions, all water enters and leaves the system through these boundaries; therefore, the system's water budget and, ultimately, the amount of ground water available for development, is affected by the characteristics of these boundaries.

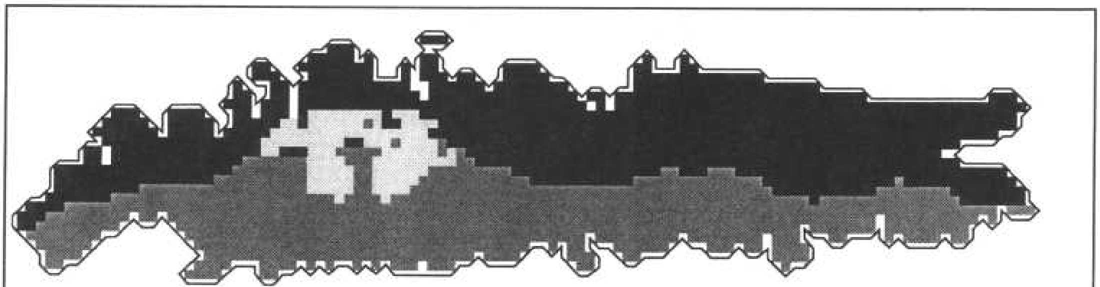


Figure 10. Specific yield and extent of unconfined areas of major hydrogeologic units.

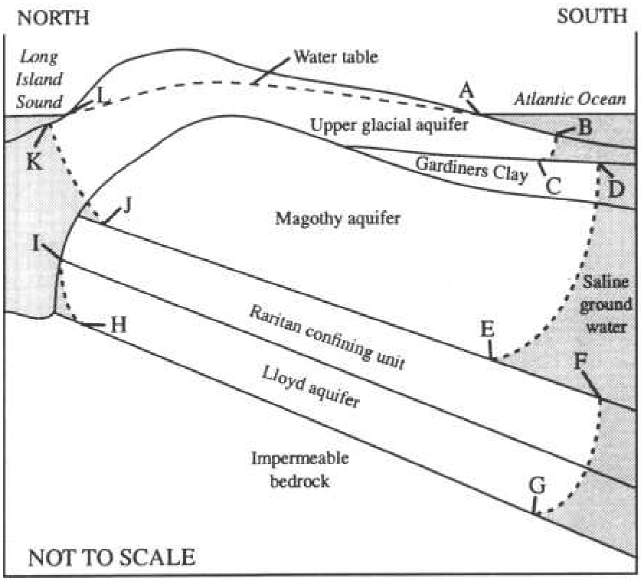
Water Table

The water table, which is a free surface, rises and falls with changing hydrologic conditions, determining the saturated thickness of the water-table aquifer (fig. 11, segment LA). The upper layer of the islandwide model is represented as a water-table layer, in which the saturated thickness in each cell is calculated as the difference between the simulated head and the altitude of the bottom of the layer in that cell (fig. 7A). Recharge enters the ground-water system at the water table. Under predevelopment conditions, recharge was derived solely from precipitation (fig. 12), which averages about 45 in/yr (Peterson, 1987). About 52 percent of the annual precipitation recharged the ground-water system (fig. 12); only about 1 percent of precipitation was lost as overland flow because the topography is relatively flat, and the highly permeable unconsolidated deposits at land surface allowed nearly all water to infiltrate. The remaining 47 percent was lost through evapotranspiration largely before recharging the system.

Precipitation is not uniform across Long Island. The long-term average distribution of precipitation has been estimated by Miller and Frederick (1969), Bailey and others (1985), and Peterson (1987). The corresponding distribution of recharge under predevelopment conditions (fig. 13) was estimated from the above sources and adjusted slightly during model calibration. Recharge values range from 22 to 26 in/yr across the island; highest recharge rates are in the center of the island.

Bedrock

The bedrock surface that underlies Long Island is considered the bottom boundary of the ground-water system (fig. 11, segment GH). The hydraulic conductivity of these poorly fractured igneous and metamorphic rocks probably is at least as low as the vertical hydraulic conductivity of the major confining units (Freeze and Cherry, 1979, p. 29). Furthermore, no underlying water-bearing unit is known that would induce vertical flow across this boundary. For these reasons, the bottom boundary of the ground-water system is considered impermeable (no-flow).



BOUNDARY SEGMENT	HYDROGEOLOGIC FEATURE	MATHEMATICAL REPRESENTATION
LA	Water table and streams	Specified flow (free surface) Specified flow and head-dependent flow[1]
HG	Consolidated bedrock	No flow (streamline)
AB,KL	Shore discharge	Constant head
BC, DE ,FG, HI, JK	Saltwater-freshwater interface	No flow (streamline)
CD, EF, IJ	Subsea discharge	Specified head

[1]Stream boundaries are specified differently in different simulations.

Figure 11. Generalized hydrogeologic section showing major hydrologic boundaries and their mathematical representation.

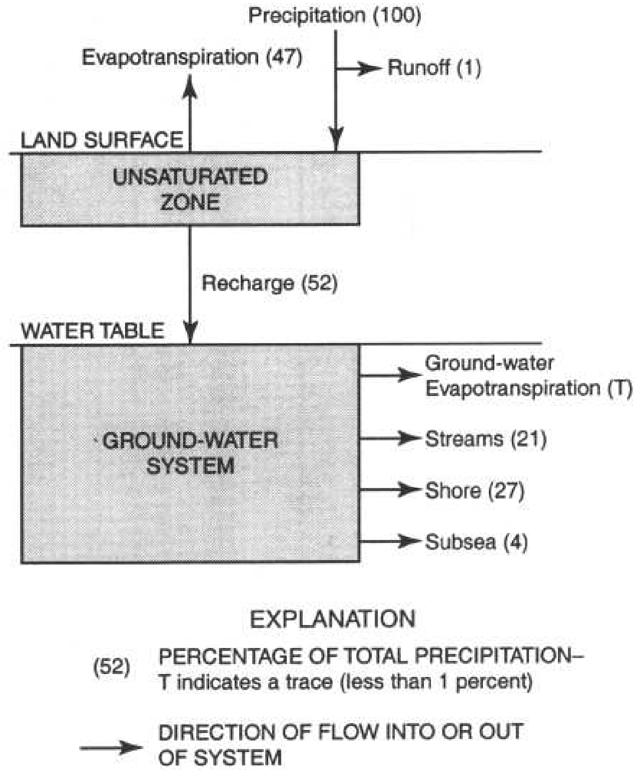
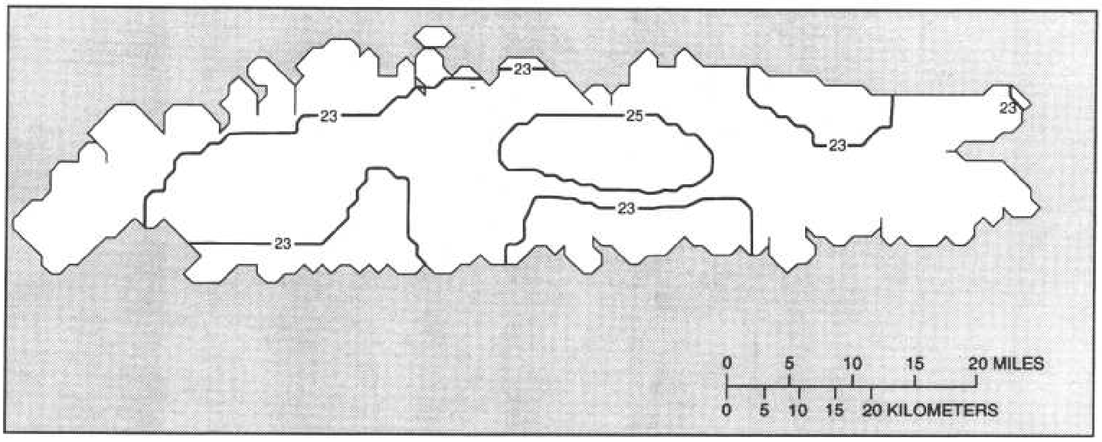


Figure 12. Predevelopment fate of precipitation in Long Island, N.Y.

Streams

More than 100 stream channels, typically less than 5 mi long, flow to the tidewater that surrounds Long Island (fig. 3A). The channels were formed by glacial meltwater and therefore are more abundant along the southern shore than along the northern shore. Ground-water discharge to streams has a major effect on flow patterns within the ground-water system. Under predevelopment conditions, about 21 percent of precipitation, equivalent to more than 40 percent of the ground water leaving the system, discharged to streams (fig. 12). Very little precipitation (1 percent or less) flowed to streams as runoff. Base flow in these streams is maintained year round by ground-water discharge, and analysis of continuous hydrographs of streams in undeveloped parts of Suffolk County indicate that, under predevelopment conditions, base flow constituted 95 percent of total streamflow (Pluhowski and Spinello, 1978; Reynolds, 1982).

Streams flow continually where their channels intersect the water table and collect ground-water discharge (fig. 14A); in most streams this intersection is continuous from the start of flow to the mouth (fig. 14B). The rate of seepage is controlled by (1) the difference between the head in the aquifer and the stream stage, (2) channel geometry, and (3) water-transmitting properties of the aquifer and streambed material. The length of the flowing stream channel and the amount of base flow vary with seasonal and other water-table fluctu-



EXPLANATION

—23— LINE OF EQUAL RECHARGE FROM PRECIPITATION--
Inches per year. Interval 2 inches

Figure 13. Estimated distribution of ground-water recharge from precipitation on Long Isalnd, New York.

ations. Seepage stops and the channel becomes dry when the water table falls below the channel (fig. 14).

In the steady-state analyses of predevelopment conditions, ground-water discharge to streams was estimated from streamflow measurements made largely before development, during 1851-1907 (Spear, 1912; Burr, Hering, and Freeman, 1904; Veatch and others, 1906; Kirkwood, 1867; McAlpine, 1852; Stoddard, 1854) and during the 1940's and 1950's in undeveloped parts of eastern Suffolk County. The average base flow for major streams (flow exceeding 5.0 cubic feet per second) under predevelopment conditions is listed in table 3. The length of each flowing stream channel (fig. 3) was estimated from early maps given in Veatch and others (1906) and Spear (1912).

The discharge specified for each model cell is proportional to the length of channel in that cell. Ground-water discharge to ungaged streams was estimated from seepage rates in nearby gaged streams of similar morphology. The model representation of the stream-channel network is illustrated in figure 3B. In all, 108 streams are represented in the simulation of steady-state predevelopment conditions. Stream representation for transient conditions, in which base flow changes in response to water-table fluctuations, is discussed in later sections.

Shoreline Discharge Boundaries

Long Island is surrounded by tidal saltwater bodies to which ground water discharges. This zone of discharge is associated with the saltwater-freshwater interface, and its width (fig. 11, segments AB and KL) is controlled by the hydraulic conductivity and anisotropy of the local deposits. Discharge is greatest near the shore, where gradients are largest, and decreases rapidly offshore as gradients decrease. The discharge is controlled

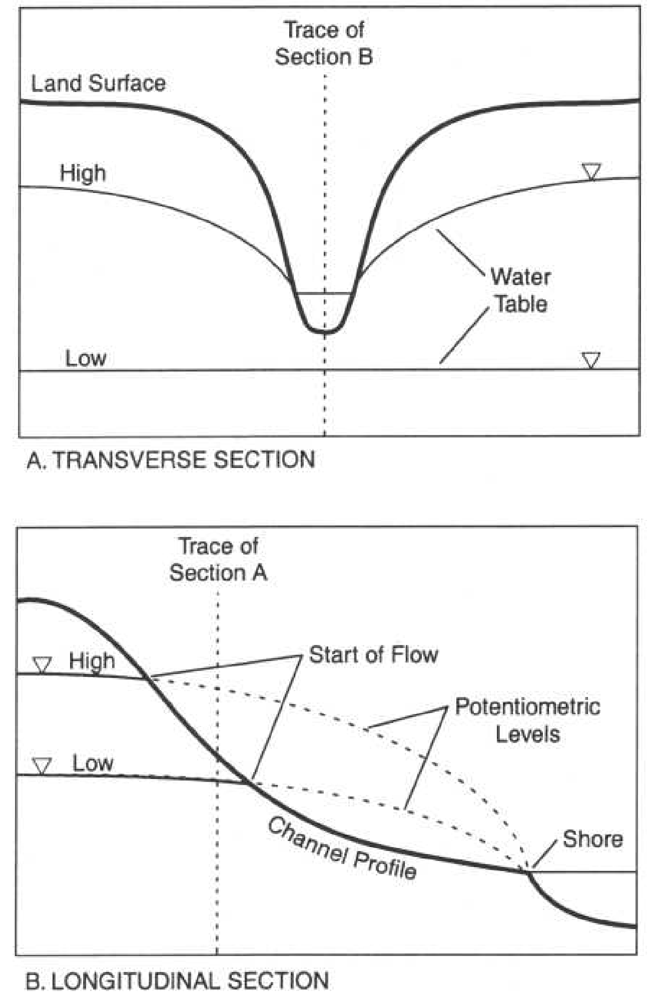


Figure 14. Generalized relation between water table and stream channel during seasonal high- and low-flow periods. A. Transverse section. B. Longitudinal section

by sea level. Cells in the model that correspond to tidewater were assigned constant-head values equal to mean sea level; the outline of constant-head nodes representing the shore is shown in figure 3B.

Saltwater-Freshwater Interface

Interfaces between fresh and saline ground water form lateral boundaries of the fresh ground-water system (fig. 11, segments BC, DE, FG, HI, JK). Because fresh ground water generally moves parallel to this interface and does not cross it, the interface is represented in the model as an impermeable (no-flow) boundary. Minor mixing along this interface creates a zone of diffusion that is characterized by a gradual transition from low to high salinity. Analyses of chloride concentration in pore fluid from core samples, and electric borehole logs taken from nearshore wells in eastern Suffolk County (U.S. Geological Survey records), indicate that the zone of diffusion is a few tens of feet thick. These data are discussed in a later section.

The saltwater-freshwater interface is a free surface that, like the water table, moves in response to head changes within the ground-water system. Under steady-state conditions, the location of the interface is the point along which the pressure in the freshwater system balances pressure in the saltwater system. Ground-water levels measured for more than 50 years in confined aquifers along the southern shore have always indicated that pressures within the freshwater system were inadequate to balance saltwater. This imbalance indicates that the interface is not in an equilibrium position and must be moving landward slowly over the long term. An explanation for a similar imbalance in aquifers of the New Jersey coastal plain is provided by Meisler and others (1984); water levels in confined aquifers have not fully adjusted from low stands of sea level during the last glaciation, more than 10,000 years ago. As a result, water levels offshore throughout the freshwater and saltwater systems are lower than expected. Ground-water velocities near the interface under predevelopment conditions are estimated to have been very low—probably not more than a few tens of feet per year; therefore, the interface is represented in the model as a stationary no-flow boundary. The configuration of the interface in the Magothy and Lloyd aquifers under predevelopment conditions is shown in figure 17 (later in this report).

Table 3. Average base flow of major streams on Long Island, under predevelopment conditions

Map number (fig. 3)	Stream name	Flow	Map number (fig. 3)	Stream name	Flow
1	Jamaica Creek	17.9	17	Sampawams Creek	9.9
2	Springfield Stream	7.9	18	Penataquit Creek	6.8
3	Simonsons (Brookfield) Stream	9.6	19	Pardees and Orowoc Creeks	10.3
4	Valley Stream	14.3	20	Rattlesnake Brook	9.2
5	Motts Creek	6.4	21	Connetquot River	36.0
6	Pines Brook	13.0	22	Green Creek	6.5
7	South Pond	20.0	23	Patchogue River	18.9
8	Parsonage Creek	8.1	24	Swan River	13.3
9	Milburn Creek	13.0	25	Carmans River	24.9
10	East Meadow Brook	15.3	26	Forge River	9.6
11	Cedar Swamp Creek	9.5	27	Little River	7.4
12	Bellmore Creek	14.6	28	Peconic River	37.4
13	Massapequa Creek	12.0	29	Nissequogue River	41.7
14	Carman Creek	6.8	30	Mill Neck Creek	7.0
15	Santapogue Creek	10.0	31	Glen Cove Creek	8.7
16	Carlls River	27.3	32	Flushing Creek	21.5

Subsea-Discharge Boundaries

Ground water that discharges to subsea boundaries flows upward through a confining unit and mixes with overlying saline ground water. As a result, the head beneath the confining unit is elevated, and the saltwater-freshwater interface beneath the confining unit is displaced seaward. The areas in which this occurs (fig. 11, boundary segments CD, EF, and IJ) are referred to as subsea-discharge boundaries.

The rate of ground-water discharge to subsea boundaries varies with hydrologic conditions within the ground-water system. In the model, these boundaries are represented by a constant head along the upper surface of the confining unit; this representation allows the rate of ground-water discharge to change as head within the system responds to natural or human-induced stresses. The constant head (H) at these boundaries can be calculated directly from the following equation if the overlying saline ground water is assumed hydrostatic.

$$H = Z\frac{(\rho_s - \rho_f)}{\rho_f}, \qquad (1)$$

where Z = depth to upper surface of confining unit,
ρ_f = density of saline ground water, and
ρ_s = density of fresh ground water.

Saline ground water on Long Island is not hydrostatic, but is moving gradually landward. In addition, the continuous discharge of fresh ground water through subsea boundaries probably has diluted the receiving waters; therefore, the constant-head value that controls discharge from these boundaries was calculated from an adjusted saltwater density of 1.017 g/cm^3, slightly less than the density of seawater, 1.025 g/cm^3. This approximation enabled accurate representation of the observed heads in the confined aquifers. This representation does not consider the slow landward movement of the saltwater interface, and the associated small amount of freshwater derived from saltwater forcing freshwater from pore spaces (storage). These factors probably would have only a small effect very close to the interface, and are assumed negligible for the purpose of this analysis.

Ground-Water Levels and Flow Patterns

A discussion of the patterns and vertical distribution of ground-water flow among the aquifers on Long Island is provided by Buxton and Modica (1992); as part of this analysis a cross section model near the Nassau-Suffolk County border was used to construct a flow net that defines the paths ground-water takes through the system from recharge to discharge (fig. 15). Knowledge of the 3-dimensional patterns of ground-water flow can be inferred from potentiometric maps of the major aquifers. The first comprehensive map of the water-table configuration on Long Island (fig. 16A) was constructed from water levels measured in 1903 (Veatch and others, 1906). At that time, the water table reached a maximum altitude of more than 100 ft. Precipitation for several years after the turn of the century was above average, however, indicating that water levels in 1903 also were above average for predevelopment conditions. Furthermore, ground water was already being used in Kings and Queens Counties for public supply and industry, and pumpage probably exceeded 60 Mgal/d by the turn of the century; therefore, the ground-water levels in western Long Island at that time are not truly indicative of predevelopment conditions. Franke and McClymonds (1972), considering these factors, estimated the average predevelopment water-table configuration (fig. 16B).

Horizontal components of flow in the shallow aquifer generally trend perpendicular to the water-table contours (fig. 16). Upon reaching the water table, ground water flows downward and laterally toward the shore and stream boundaries (figs. 15 and 16). Water-table depressions form where the water table intersects stream channels, and a shallow ground-water flow subsystem develops that discharges to each stream. The three-dimensional nature of these shallow flow systems is described in detail in Prince and others (1989), Harbaugh and Getzen (1977), and Franke and Cohen (1972).

The water-table configuration as depicted in figures 15 and 16 is asymmetrical; the major ground-water divide is closer to the northern shore than to the southern shore. Therefore, more than half the water within the system discharges to the south. This asymmetry is due to three major reasons: (1) the unconsolidated deposits that form the Long Island ground-water system thicken southward (fig. 4); (2) glacial deposits on the southern half of the island have higher permeability than those in the north (fig. 9A); and (3) more numerous streams and greater base flow exists on the southern shore than in the north.

Local areas along the northern shore show anomalously high water-table altitudes that are attributed to zones of very low permeability within the moraine deposits, and the pinch-out of aquifer units near the shore. The distribution of these water-table highs in Queens County is described in detail by Buxton and Shernoff (1995).

The model simulation of predevelopment conditions yields an approximation of the head distribution in the ground-water system (fig. 17). The simulated water-table configuration (fig. 17A) closely matches those based on predevelopment measurements (fig. 15A, 15B) and reproduces the asymmetric water-table shape, the local highs along the northern shore, and convergent flow patterns near stream channels.

Head measurements are insufficient to enable accurate mapping of the predevelopment potentiometric surfaces in the Magothy and Lloyd aquifers; although Kimmel (1973) inferred the potentiometric surface in the Lloyd

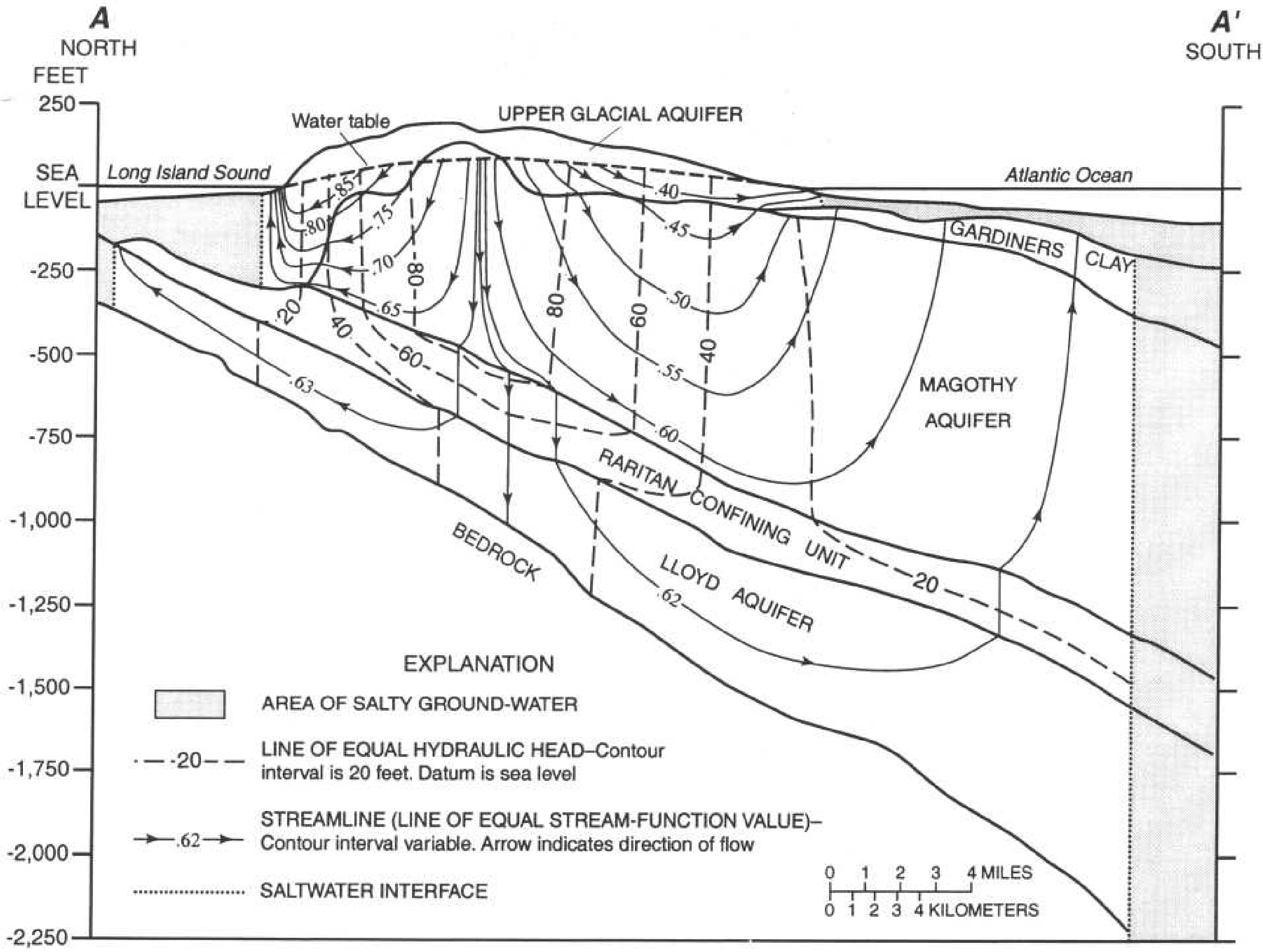
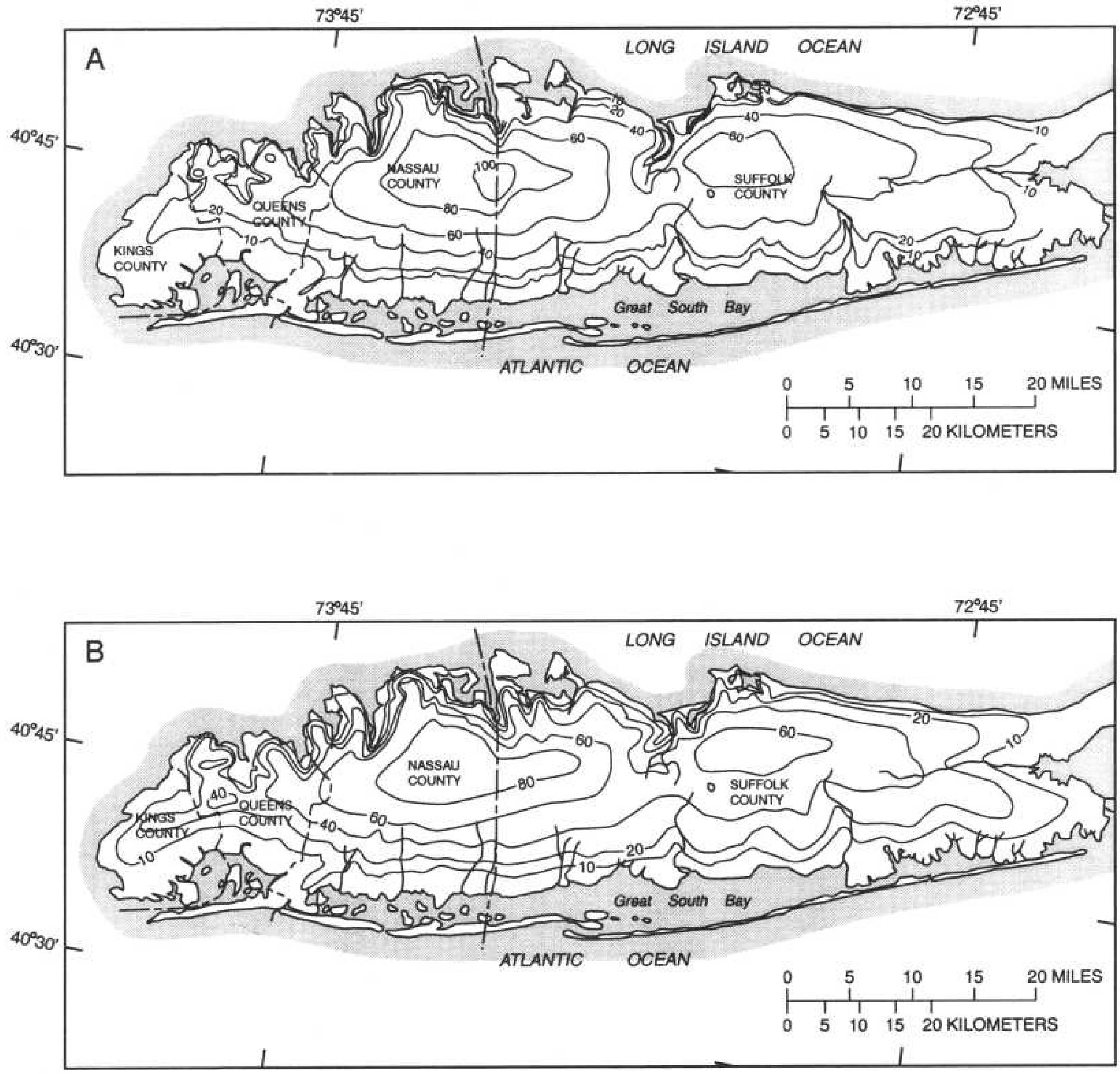


Figure 15. Generalized ground-water flow patterns near the Nassau-Suffolk County border, Long Island, N.Y. (From Buxton and Modica, 1992, fig. 6.)

aquifer in 1900 from water-level measurements made during 1923-70.

The simulated potentiometric surface of the Magothy aquifer (model layer 3) is a subdued replica of the water table (fig. 17B). However, highs along the ground-water divide are several feet lower than the water table. The subdued effects of large streams also are evident, especially at Connetquot and Nissequogue Rivers at Carmans River, and at the Peconic River. (Stream locations are shown in figure 3.) Offshore, beneath the Gardiners Clay, large vertical gradients drive water upward to subsea discharge.

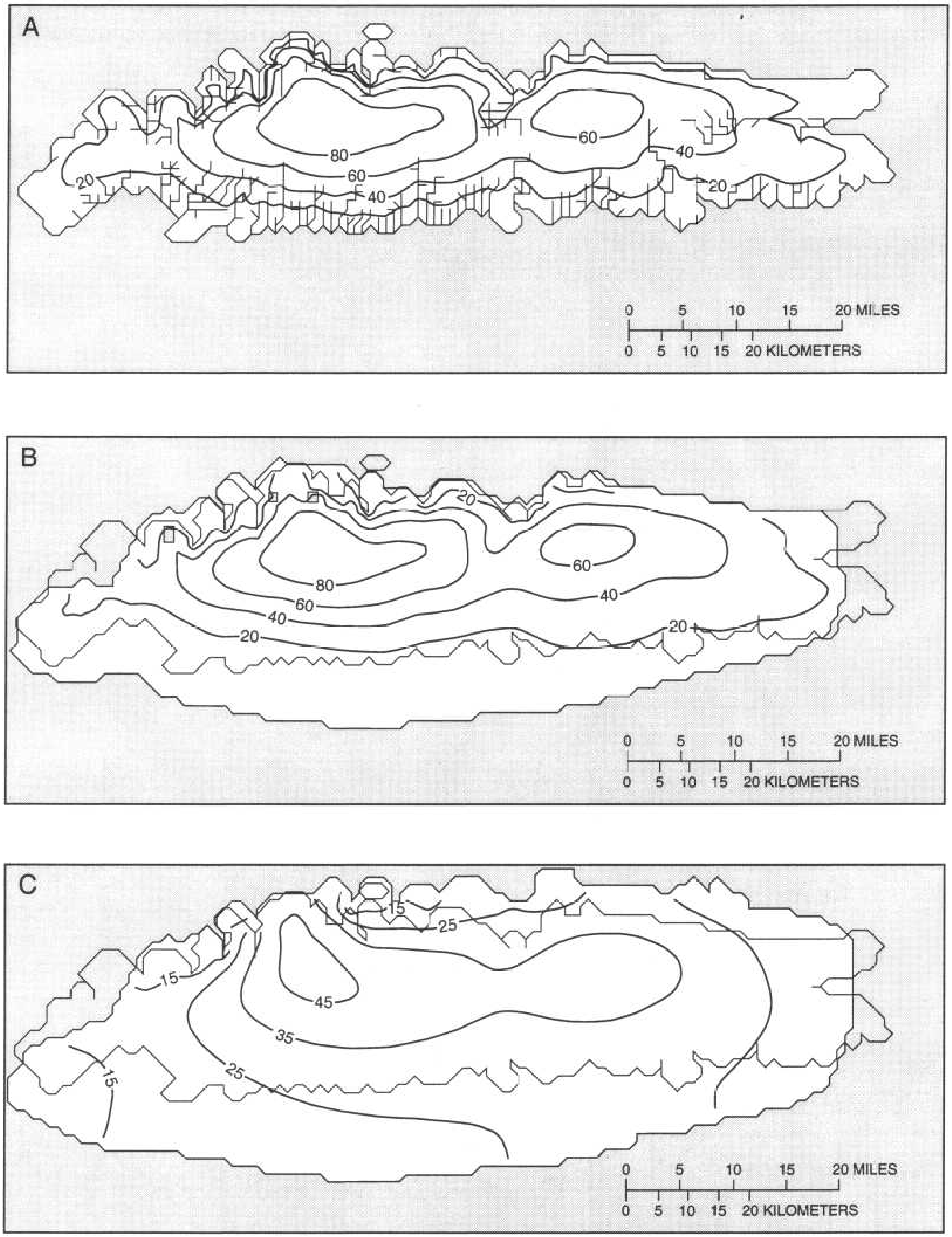
The simulated potentiometric surface in the Lloyd aquifer (fig. 17C) is considerably lower than that in the Magothy aquifer (fig. 17B) because the Raritan confining unit separates the aquifers throughout most of the island. Vertical head differences across the Raritan confining unit are as much as 50 ft in Nassau County. Water in the Lloyd aquifer flows seaward (fig. 17C). Vertical flow downward into the Lloyd is greatest at the ground-water divide but decreases shoreward until flow reverses direction and either reenters the Magothy aquifer or discharges to the Lloyd's subsea-discharge boundary (fig. 15).



EXPLANATION

——20—— WATER TABLE CONTOUR--Shows estimated altitude of water table under natural conditions, Contour intervals 10 and 20 feet. Datum is sea level

Figure 16. Predevelopment water-table configuration. A., 1903 (Modified from Veatch and others, 1906, plate 12). B., Estimated by Franke and McClmonds (1972, fig. 9).



EXPLANATION

AREA OUTSIDE EXTENT OF FRESH GROUND-WATER SYSTEM

—— 20 —— POTENTIOMETRIC CONTOUR--Shows altitude at which water level would rise in a piezometer. Contour interval, in feet, is variable. Datum is sea level

Figure 17. Simulated predevelopment distribution of hydraulic head. A., Water-table aquifer (model layer 1). B., Magothy aquifer (model layer 3). C., Lloyd aquifer (model layer 4).

The flow patterns in the Lloyd aquifer are affected significantly by three holes in the confining units that separate it from the Magothy aquifer—the eroded channel through central Queens County, and two gaps between the northern limit of the Raritan confining unit and the Port Washington confining unit in northern Nassau County (figs. 6B and 6E). The effects are greatest in northern Nassau County, where water enters the Lloyd through one of these holes at model cell (row 37, column 7). The highest part of the potentiometric surface of the Lloyd is centered at this point (fig. 17C), which is much closer to the northern shore than would be expected if the Lloyd aquifer were recharged solely by diffuse leakage through the overlying confining unit. The potentiometric surface indicates flow away from this source area (hole) in all directions.

Ground-Water Budget

The ground-water budget defines the amount of water entering and leaving the system through each of its natural boundaries. Each budget component is represented by an average flow rate, and inflow is balanced by outflow. Rates of recharge from precipitation and ground-water discharge to streams were estimated from field measurements, as described previously; discharge to the shore and subsea boundaries were calculated with the islandwide model. Therefore, the uncertainty associated with total values for water-budget components is low, but increases for model estimates of the spatial distribution of each component for areas of the island.

Recharge exceeding 1.1 billion gal/d entered the Long Island ground-water system under predevelopment conditions (table 4). The greatest outflow was to the shore (585 Mgal/d, or 52 percent), the next greatest was to streams (460 Mgal/d or 41 percent), and the least was to the subsea boundaries (81 Mgal/d, or 7 percent). Discharge to the stream and shore boundaries constituted more than 90 percent of total discharge because both occur in the water-table aquifer, which is nearest the recharge and has a high hydraulic conductivity.

The model approximation of the actual ground-water system introduces some error in the estimation of the system water budget. The model does not represent some features of the island such as narrow peninsulas and barrier islands. It also does not include recharge that enters model cells that represent the constant head shoreline boundary. Buxton and Shernoff, 1995, estimate the total recharge to Kings and Queens counties by applying an average recharge rate of about 1.1 Mgal/d/mi^2 to the entire land area of these counties (189 mi^2), yielding a total recharge from precipitation of 209 Mgal/d or about 30 percent higher than the estimate in this analysis. This discrepancy is attributed to the significant land areas in Kings and Queens near the shore that do not act as part of the main ground-water system; and loss of accounting of recharge to shoreline constant head cells.

The water budget (table 4) is divided into four geographic areas to indicate the spatial variation in the distribution of ground-water flow. Although inflow precisely balances outflow for the entire system, each of the four geographic areas contains imbalances between inflow and outflow that are balanced by flow between adjacent areas. In Kings and Queens Counties, for example, discharge exceeds recharge from precipitation but is balanced by inflow of about 4 Mgal/d from Nassau County. The percentage of flow that discharges to each boundary also differs from area to area. The percentage discharged to streams is less in Kings and Queens Counties (where streams are relatively few and base flow constitutes 36 percent of the water budget) than in Nassau and western Suffolk Counties (where streams are numerous, and base flow constitutes half of the water budget).

Table 4. Ground-water budget for predevelopment conditions on Long Island

County	Recharge	Discharge		
	Precipitation	Stream	Shore	Subsea
Kings and Queens	160	58	96	10
Nassau	257	125	94	24
West Suffolk	273	140	137	28
East Suffolk	436	137	258	19
Total	1,126	460	585	81

Ground-water discharge decreases sharply with depth, as indicated by the small amount of subsea discharge in relation to stream and shore discharge. Progressively smaller amounts enter each successive model layer (aquifer) (table 5); only about 20 percent of the flow in the system enters the basal zone of the Magothy and Jameco aquifer (layer 3), and only about 3 percent enters the Lloyd aquifer (layer 4). A disproportionate amount of water enters the Lloyd in Nassau County (table 5), where the two holes in the confining units, (each represented by only a single model cell), together allow 2.2 Mgal/d to flow to the Lloyd aquifer. Much of the downward flow to each of layers 2, 3, and 4 (table 5) returns to the overlying aquifer, however, and continues flowing through the system. (See fig. 15.)

Findings that most ground-water flows in the shallowest part of the aquifer system and that progressively less water flows to each aquifer with depth suggests that water moves more slowly and has greater residence time in the deep confined aquifers. Results of Buxton and Modica (1992) indicate that under predevelopment conditions, ground-water travel-times in the water-table aquifer are on the scale of tens of years; in the Magothy aquifer are on the scale of hundreds of years; and in the Lloyd aquifer are on the scale of thousands of years.

Table 5. Distribution of ground-water flow with depth under predevelopment conditions as represented in model

County	Model layer[1]			
	1	2	3	4
	(water table)	(Magothy and Jameco)		(Lloyd)
Kings and Queens	160	28	16	3
Nassau	257	116	62	16
West Suffolk	273	141	75	9
East Suffolk	436	177	82	8
Total	1,126	462	235	36

[1]Flow into layer 1 is recharge from precipitation; flow into layers 2, 3, and 4 is leakage from the overlying layer.

EFFECTS OF DEVELOPMENT ON THE GROUND-WATER SYSTEM

Human activities affected the ground-water system on Long Island as early as the mid-17th century, when early European settlers withdrew water from streams or from shallow dug wells that intersected the water table. Most wastewater infiltrated back to the water table and affected water quality locally, but had negligible effect on the quantity or patterns of ground-water flow. Over the next 2 centuries, the population increased significantly, mainly in western Long Island. By the 19th century, local dug wells were being replaced by large-capacity but shallow public-supply wells that served population centers. The increased water use and attendant onsite wastewater disposal posed a major threat to the quality of shallow ground water. To minimize further contamination, the City of Brooklyn, in the mid-19th century, began construction of a combined storm- and sanitary-sewer system to carry wastewater to tidewater. Although these sewers slowed the rate of ground-water contamination, they also diverted a large quantity of water that would have recharged the ground-water system. From the earliest development of Long Island, diversion of recharge to tide water via increased runoff over developed land and storm