Quantitative Assessment of the Shallow Ground-Water Flow System Associated With Connetquot Brook, Long Island, New York

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Prepared in cooperation with the Nassau County Department of Public Works, Suffolk County Department of Health Services, and Suffolk County Water Authority

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By KEITH R. PRINCE, O. LEHN FRANKE, and
THOMAS E. REILLY

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Quantitative Assessment of the Shallow Ground-Water Flow System Associated with Connetquot Brook, Long Island, New York

By Keith R. Prince, O. Lehn Franke, *and* Thomas E. Reilly

Abstract

Streamflow on Long Island is derived principally from shallow ground water that flows above the deeper regional flow system. The movement of shallow ground water was studied during 1975–82 at Connetquot Brook, an undisturbed stream in Connetquot River State Park, in south-central Long Island, New York. The investigation encompassed (1) field studies of streamflow, ground-water levels, and age of water as indicated by tritium and dissolved-oxygen concentrations and (2) numerical simulation of the shallow flow system to evaluate the hydraulic factors that govern the direction of shallow ground-water flow near and beneath the stream.

Analysis of water-level data indicates that ground water flows essentially horizontally throughout the drainage basin except near and beneath the stream, where it moves vertically upward toward the stream discharge boundary. Water levels in wells driven directly into the streambed and into the streambank at three sites were 1 to 2 feet higher than stream stage in the first 5 feet of penetration. Increases in head, which were detected to depths of 30 feet beneath the streambed, indicate upward movement of water above that depth. Hydraulic conductivity of the streambed sediments was calculated from head gradients in the streambed and from measured stream seepage; values were between 11 and 15 feet per day.

Water samples from selected wells were analyzed for dissolved-oxygen and tritium concentrations to determine the relative age of the water in an attempt to locate the bottom boundary of the shallow flow system. Dissolved oxygen showed no pattern, but tritium concentrations about 1,000 feet from the stream were lower than those near the stream. The tritium concentrations indicate that the lower flow boundary was between 45 and 100 feet below the water table.

A two-dimensional cross-sectional flow model of the shallow flow system was developed. The near-stream model response compared well with field data when the streambed discharge boundary was simulated as a uniform leaky bed. A systematic sensitivity analysis was done to determine which factors have the greatest influence on hydraulic head in the system. Ten dimensional parameters that describe the important aspects of the flow system were combined into a series of dimensionless parameters to simplify analysis. Results indicate that (1) streambed factors (width and hydraulic conductivity) are most influential upon heads near the stream, (2) factors representing thickness of the shallow flow system influence heads distant from the stream but have a negligible effect near the stream, and (3) factors that represent the quantity of water entering the system (recharge) influence the heads throughout the area.

Field measurements of hydraulic head indicate that the thickness of the shallow flow system below the stream channel is about 30 feet. However, results of the sensitivity analysis indicate that the shallow system's thickness has a negligible effect on head distribution beneath the stream.

INTRODUCTION

Long Island, N.Y., is underlain by a southward-dipping sequence of unconsolidated gravel, sand, silt, and clay that overlies Precambrian bedrock. This sequence forms the large fresh ground-water reservoir that supplies water for the 3 million inhabitants of Nassau and Suffolk Counties (fig. 1). A significant feature of the Long Island hydrologic system is that, because its many streams act as drains for the water table, the streams are sustained principally by ground-water seepage. About 95 percent of the total flow under natural (predevelopment) conditions is derived from ground water (Pluhowski and Kantrowitz, 1964). Although the flow of individual streams is relatively small, their combined base flow represents a significant percentage of the total freshwater outflow from the island's hydrologic system.

Construction of an extensive sanitary sewer system in southwestern Nassau County was begun in the 1950's to help protect the ground-water system from septic-waste contamination, which had increased with population growth. Since then, sewer construction has been extended into southeastern Nassau and southwestern Suffolk Counties. In recent years, water managers and the public have become concerned about the detrimental effects that the sewer systems may have on the ground-water system. One major concern is a predicted decline in the water table and

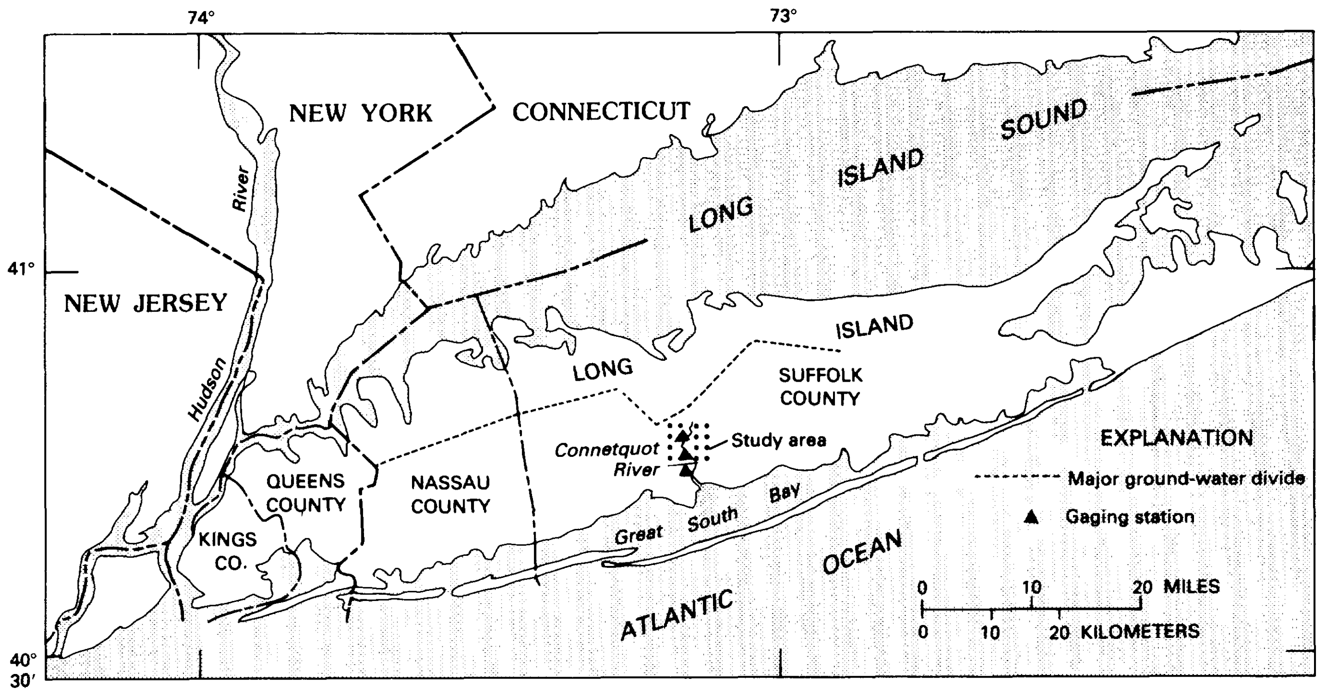


Figure 1. Major geographic features of Long Island, N.Y., and location of Connetquot River State Park study area.

in streamflow throughout central Long Island because waste water is piped to the sea instead of being returned to the ground-water system through cesspools and septic tanks (Kimmel and others, 1977).

The predicted lowering of the water table and the attendant reduction in streamflow have led to concern over the future of many of Long Island's wetlands and parks where surface water is essential for recreation and wildlife habitat. A reduction in streamflow will result also in an increase in bay salinity, which will adversely affect the shellfish industry. As a first step toward safeguarding the island's surface-water resources, the magnitude of the expected streamflow reductions must be assessed. Contamination of the shallow ground-water system, not only by cesspools and septic tanks but also from point sources (for example, landfills, chemical spills, and illegal disposal of chemical wastes), also has become a major concern. Continued protection of the island's surface-water resources from contamination by these sources requires a knowledge of the source and flow paths of the ground water discharging to the streams.

This study was one of several by the U.S. Geological Survey that address the interaction between shallow ground water and streams on Long Island. The study focused on Connetquot Brook, which flows through Connetquot River State Park, a limited-use preserve in Suffolk County (fig. 1). The park is an appropriate site for the study of ground water and surface water because it has been isolated and protected from the effects of urbanization—an uncommon situation on Long Island. The study area encompasses 7.5 mi^2 of the Connetquot Brook drainage area along a 5-mi reach of stream in Connetquot Park. The field phase of this investigation was limited to Connetquot Brook; however, results of the study are applicable to other ground-water-fed streams in similar geologic settings.

Purpose and Scope

This report is a companion to a preliminary report by Prince (1980), which described the local hydrologic system and the data-collection network and interpreted some of the early field data. This report presents and interprets more recent field data on shallow ground-water flow near and beneath Connetquot Brook and describes results of a sensitivity analysis using a numerical model to evaluate factors that govern the flow of shallow ground water near Long Island's streams.

Previous Investigations

Franke and Cohen (1972) gave the first detailed qualitative description of shallow ground-water flow systems associated with streams and included a discussion of the system boundaries. In a summary of the hydrologic situation on Long Island, Franke and McClymonds (1972, p. F23) discussed the predominantly two-dimensional flow of the regional system and provided a highly generalized cross section of the southern half of the Long Island

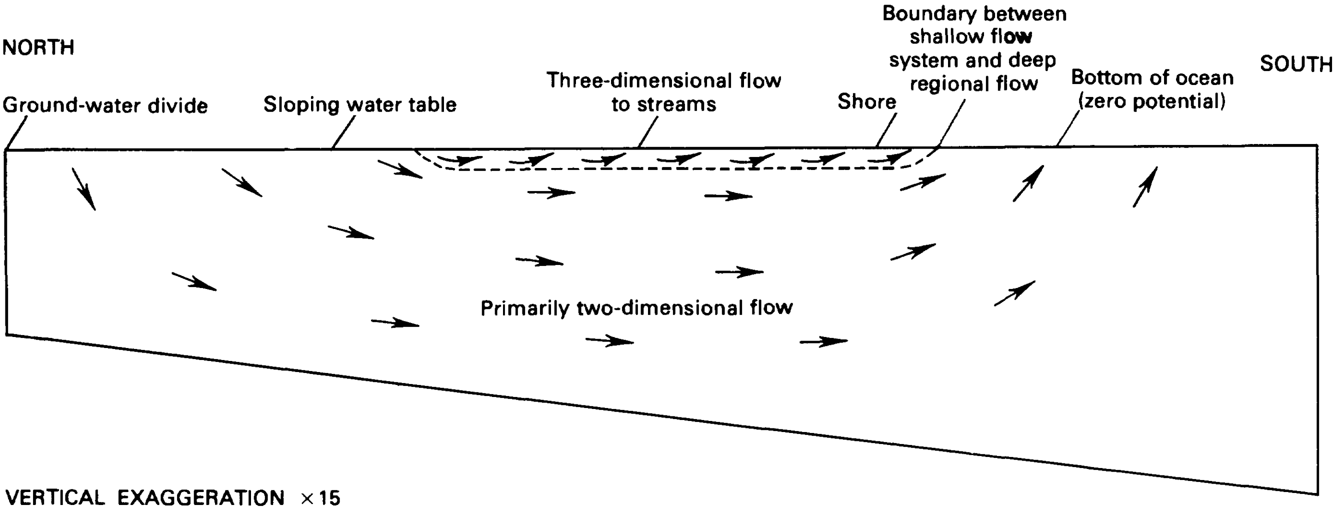


Figure 2. Idealized cross section of southern part of Long Island ground-water reservoir showing relative size and location of shallow flow system. (Modified from Franke and McClymonds, 1972, fig. 19.)

ground-water reservoir. The cross section depicted the general pattern of ground-water flow and the relative size of the shallow ground-water system associated with streams in this area (fig. 2). Harbaugh and Getzen (1977) were the first to describe the three-dimensional nature of flow in the shallow flow system (fig. 3). Prince (1980) discussed the shallow flow system further and also depicted the system's three-dimensional nature (fig. 4).

An important question suggested by figures 2 through 4 is the relationship between shallow flow systems of differing depths and scales to the total ground-water flow system. Toth (1963) discussed concepts related to this question at some length and depicted the relative position of a series of small flow systems in a hypothetical cross section (fig. 5).

Harbaugh and Getzen (1977) modeled Long Island streams and their associated shallow flow systems on a regional electric-analog model. Reilly and others (1983, p. 23–26) extended this work to include digital-numerical models.

Pluhowski and Kantrowitz (1964, p. 50) reported measurable vertical gradients beneath the bed of a flowing stream. Prince (1980, p. 20) described a series of detailed head measurements near and beneath Connetquot Brook; these and additional measurements are discussed in the section "Field Studies."

Acknowledgments

The authors thank the Long Island State Park Commission for permitting access to Connetquot River State Park during the field investigations of 1975–82, and especially Park Superintendent Gilbert Bergen and his staff for their cooperation during the field operations. We also thank Donald E. Vaupel of the U.S. Geological Survey in Trenton, N.J., for the original inspiration that resulted in this study, and Theodore A. Wyerman of the U.S. Geological Survey in Reston, Va., for the tritium analyses reported herein.

HYDROGEOLOGIC SETTING

Regional Conditions

Long Island is underlain by a sequence of unconsolidated sedimentary deposits resting on bedrock. The bedrock surface, which crops out in northeastern Queens County, dips gently to the southeast to a maximum depth of about 2,000 ft below sea level at the south shore in central Suffolk County. The deposits overlying the bedrock surface have been classified into three major aquifers—the Lloyd aquifer (in the Lloyd Sand Member of the Cretaceous Raritan Formation), which overlies the bedrock surface; the Magothy aquifer (in the Cretaceous Matawan Group and Magothy Formation), which is separated from the Lloyd by the Raritan clay (in the clay member of the Raritan Formation); and the upper glacial (water-table) aquifer (in upper Pleistocene deposits), which overlies the Magothy and may be locally separated from it by confining units. The regional hydrogeology of Long Island is described in several reports (Cohen, Franke, and Foxworthy, 1968; McClymonds and Franke, 1972). A detailed hydrogeologic description of the area surrounding Connetquot Brook is given in Pluhowski and Kantrowitz (1964) and Prince (1980).

Long Island's ground-water system, under natural conditions, is recharged solely by precipitation, which averages about 44 in/yr; recharge to the ground-water reservoir is estimated to be about half the precipitation, or 22 to 24 in/yr (Franke and McClymonds, 1972).

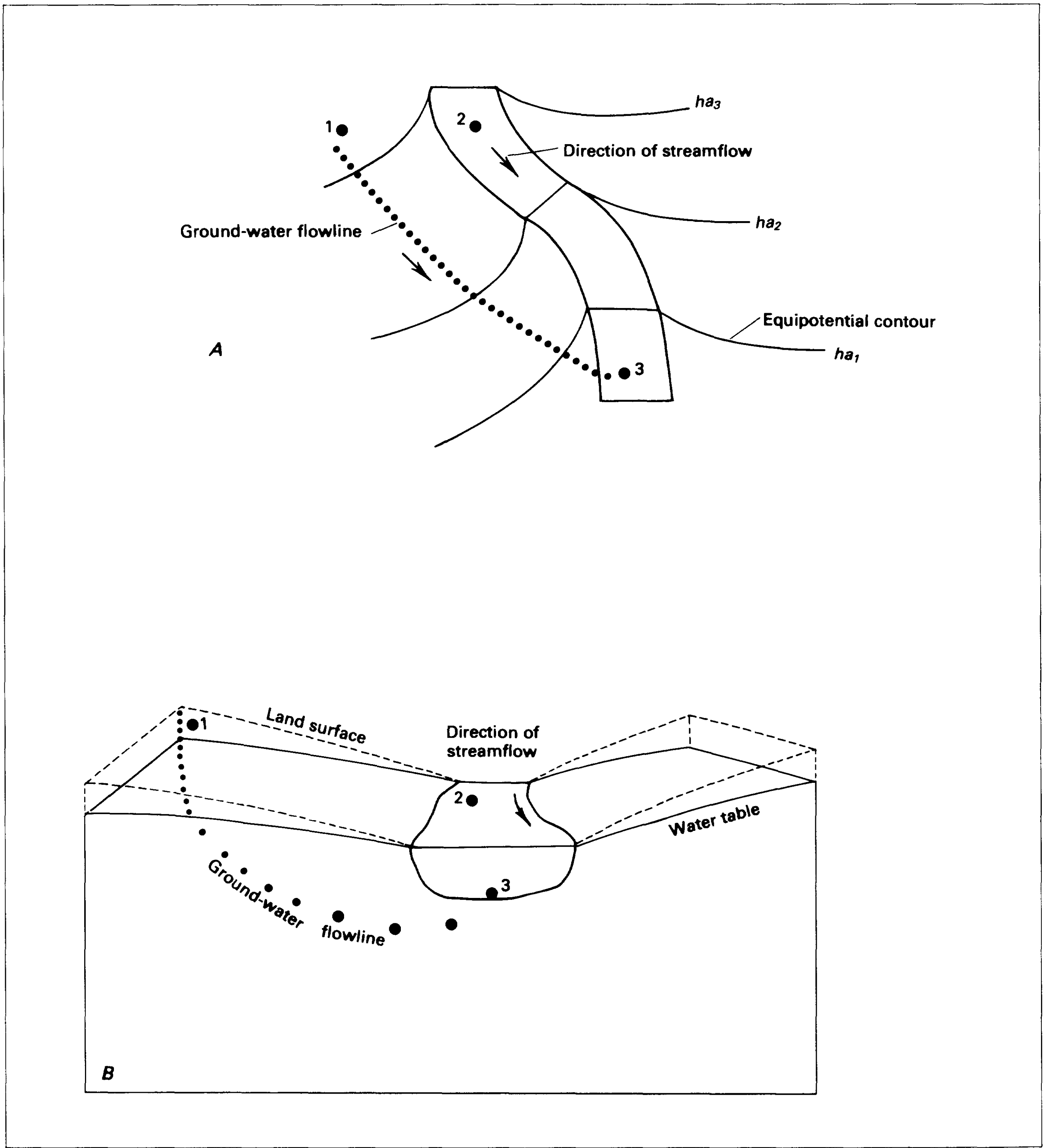


Figure 3. Direction of ground-water movement near a gaining stream. *A*, Plan view. *B*, View in three dimensions. Water moves along a path from point 1 to point 3 rather than directly toward the stream to point 2. The equipotential contours (ha_1, ha_2, ha_3) shown in *A* are omitted from *B* because in three dimensions the contours would be complex surfaces. (Modified from Harbaugh and Getzen, 1977, fig. 1.)

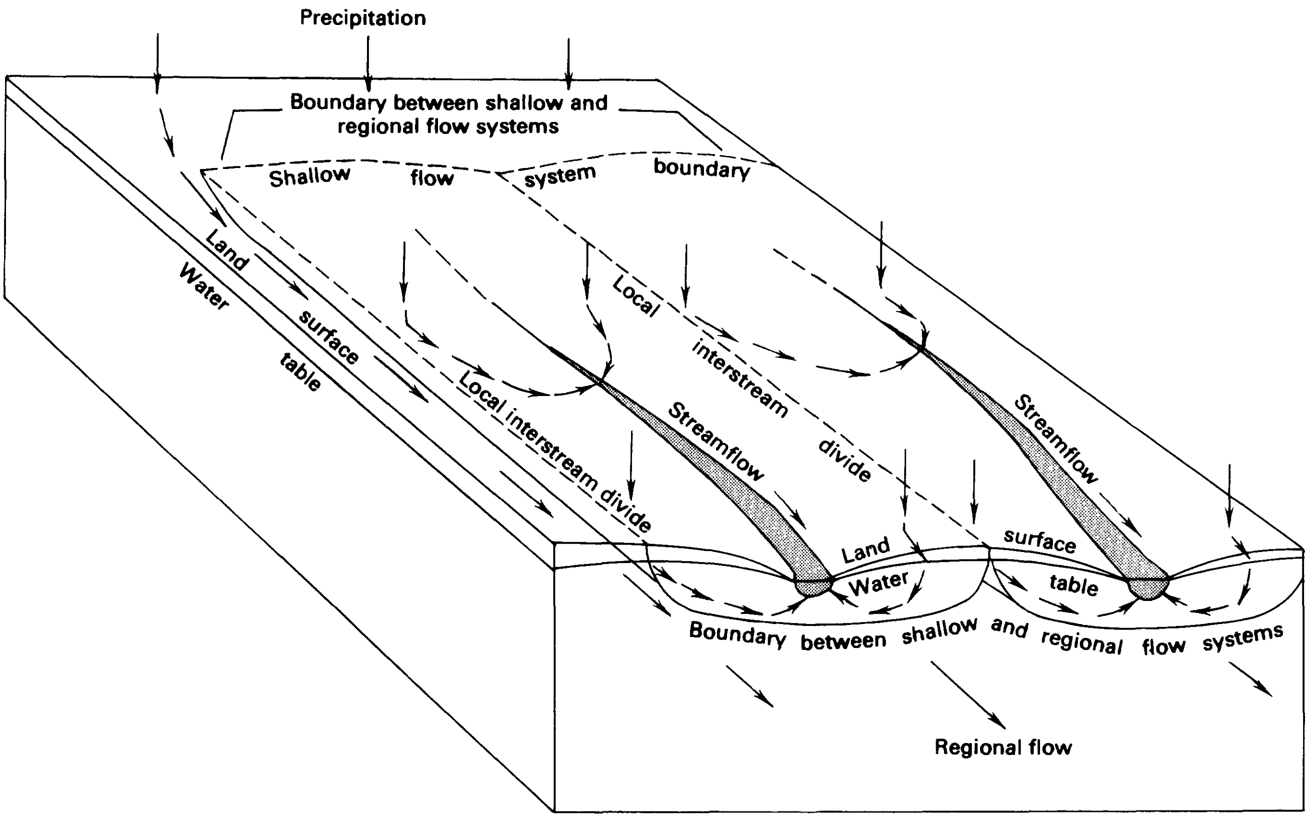


Figure 4. Shallow flow systems associated with two adjacent stream basins. A surface consisting of flow lines beneath each stream separates shallow ground-water flow from regional ground-water flow. (From Prince, 1980, fig. 9.)

Discharge from the ground-water reservoir, under natural conditions, occurs as (1) seepage to streams, (2) evapotranspiration, and (3) subsurface discharge (under flow) to the surrounding saltwater bodies. The regional ground-water flow system can be visualized as flowing laterally north and south from the major ground-water divide (fig. 1) and discharging to streams and to the south-shore bays, Long Island Sound, and the Atlantic Ocean.

Streams on Long Island drain the water table and are fed by local shallow ground-water systems that flow above the regional flow system. All water that enters a shallow flow system as precipitation either discharges as evapotranspiration or flows toward and eventually seeps into a stream channel and then discharges to tidewater as surface water.

The depth to which the shallow flow systems on Long Island extend is unknown, but Prince (1980) estimated that water in the shallow flow system associated with Connetquot Brook probably extends no deeper than 30 ft below the water table. Therefore, the following discussion of geohydrology concerns only the upper glacial aquifer, the formation in which this shallow flow system occurs.

Connetquot Brook Drainage Basin

The upper glacial aquifer in the Connetquot Brook area consists mainly of glacial outwash deposits composed of a heterogeneous mixture of fine to very coarse quartzose sand, gravel, pebbles, and boulders. In some places, the outwash is interlayered with lenses of tightly packed clay, silt, sand, and gravel. The outwash deposits are moderately to highly permeable and have potentially high rates of infiltration. In contrast, the lenses have low permeability and locally impede infiltration. In general, the upper glacial aquifer ranges from 50 to 150 ft thick in the study area (Jensen and Soren, 1974).

Stream systems on Long Island are either relict glacial outwash channels or are recent erosional features that act as drains to the ground-water system. As a stream channel is cut into the parent material and sediment is annually reworked or removed during high flow and redeposited during low flow, the stream develops a bed of reworked material that is commonly of different character than the parent material. This streambed material is a controlling feature in the seepage of water into or out of the stream.

Connetquot Brook is about 37,000 ft long; its average discharge during 1944–84 was 38.5 ft^3/s. The

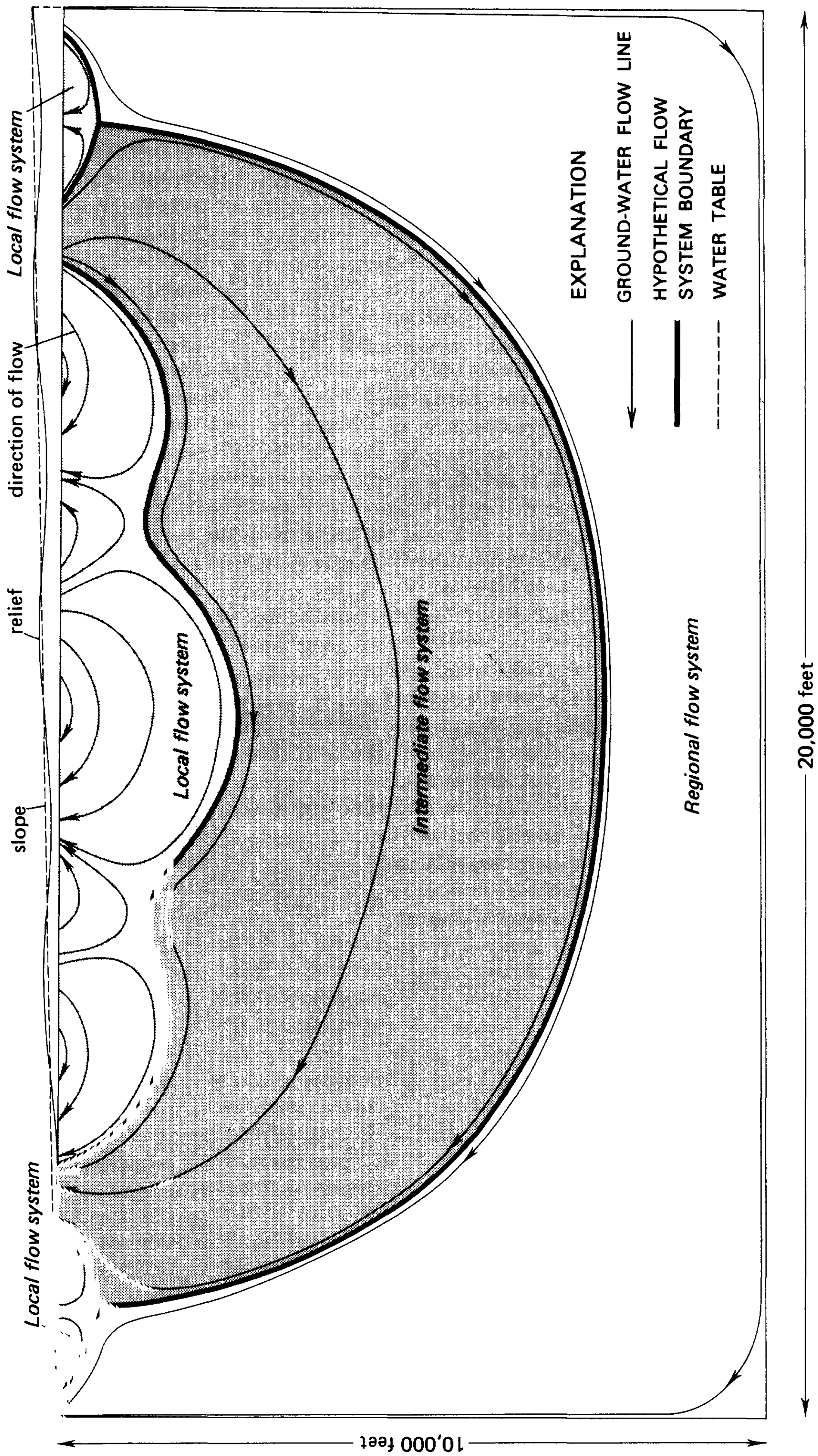


Figure 5. Generalized cross section showing local, intermediate, and regional ground-water flow systems. (Modified from Toth, 1963.)

topographic and ground-water drainage areas of the stream encompass 24 mi^2 and approximately 28 mi^2, respectively, but the boundaries of the two drainage areas differ significantly, as depicted in Prince (1980, fig. 4). The stream-channel width ranges from a few feet to 80 ft, and water depth ranges from a few inches to 3 ft or more in pools. Topographic relief in the Connetquot Brook surface-drainage basin is about 200 ft.

Boundary Conditions of the Shallow Ground-Water Flow Systems

The top of the shallow system is the water table, which is a free surface at atmospheric pressure and which fluctuates in response to intermittent recharge from precipitation. During periods of recharge, the water table may be regarded as a flux boundary—that is, some quantity of water enters the saturated ground-water system per unit area per unit time. After an extended period without recharge, however, the water table functions as a stream-line surface—that is, the ground-water flow lines move along and parallel to the water-table surface.

The lateral and basal boundaries of the shallow flow system form a surface (fig. 4) whose configuration is complex. The local interstream divides and the upstream boundary of the shallow flow system (fig. 4) represent the intersection of this surface with the water table. The approximate location of the local interstream divides can be estimated from the water-table contours on detailed water-table maps. However, the configuration of the upstream boundary of the shallow flow system is not known and may be more complicated than the nearly straight line indicated in figure 4. The shape and position of this boundary are not fixed but shift in response to changes in recharge and discharge in (1) the stream basin, (2) adjacent stream basins, and (3) the underlying regional ground-water flow system. These changes in position cause the local interstream divides and upstream shallow flow-system boundaries to shift and may also affect the altitude of the bottom boundary.

The discharge boundary of the shallow ground-water system during base flow is the streambed. The stream-surface altitude (stage) above the streambed is a direct measure of the head acting on the streambed boundary. This head at the streambed, in turn, is a function of position (head increases upstream) and time (changes in stream stage and discharge). In Long Island streams, the fluctuation in stage during base flow is small, generally a few tenths of a foot at most locations. The formal name for the streambed boundary under unstressed conditions is "specified-head boundary."

The point in the stream channel at which the water starts to flow represents the intersection of the streambed and the water table. As the water table fluctuates in response to changes in recharge and discharge, the location of the start of streamflow shifts significantly because the gradient of the channel bottom is small—that is, a 1-ft change in the water table might cause the point of start of flow to move several hundred feet.

One of the difficulties in analyzing or modeling the shallow flow systems is that only the altitude and position of the streambed remain fixed through time; the upper and lower boundaries of the shallow flow system shift continuously. These moving or potentially moving boundaries result in a nonlinear relationship between stress and the response of the shallow subsystem, as evidenced by changes in ground-water levels and streamflow. Another aspect of this nonlinear relationship between stress and system response is the change in length of the flowing parts of the stream. Analysis of the shallow flow system requires using simplifying assumptions that allow the investigator to examine the system under more ideal conditions. These assumptions are discussed in the section "Model Studies of the Shallow Ground-Water Flow System."

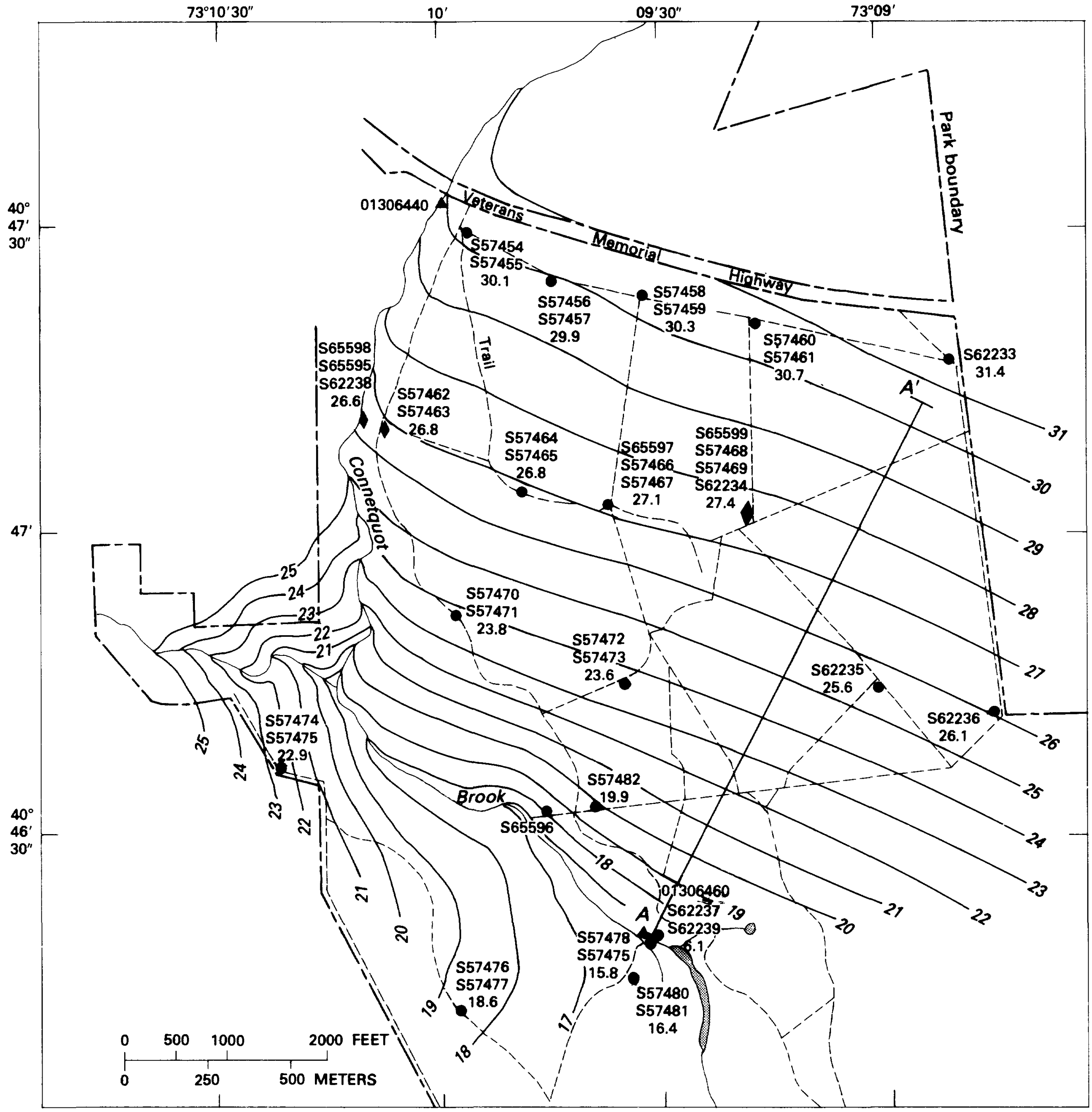
FIELD STUDIES

Field studies were conducted to obtain data on the dimensions and dynamics of the shallow ground-water flow system that drains to Connetquot Brook. The field data consisted of (1) extensive ground-water-level measurements, which were used to prepare a water-table map and evaluate flow paths and gradients, and (2) measurements of tritium and dissolved-oxygen concentrations to discern age, and thus the depth, of the shallow flow system, as described in the section "Concentration of Tritium and Dissolved Oxygen."

Water-Table Configuration

A network of 41 observation wells was installed along an undisturbed 950-ft reach of Connetquot Brook and its drainage basin, as shown in figure 6. Streamflow into and out of this reach was monitored by stream-gaging stations at the north and south ends of the reach to relate changes in seepage rates to changes in ground-water levels.

Ground-water levels were measured periodically from January 1976 through September 1978; these measurements and well-completion data are given in Prince (1980). In addition, two water-table contour maps were drawn, one for September 1977 (fig. 6), the other for March 1978 (Prince, 1980). Comparison of the two maps shows that water levels in the spring are significantly higher and gradients somewhat steeper than in the fall. As a result, the base flow of Connetquot Brook is generally greatest in the spring and lowest in the fall.



Base from U.S. Geological Survey, 1977

Hydrology by K.R. Prince

EXPLANATION

—25— WATER-TABLE CONTOUR—Shows altitude of water table. Contour interval 1 foot. Datum is sea level

A—A' TRACE OF MODELED CROSS SECTION

WELL AND NUMBER—Letter "S" designates well in Suffolk County. Number below well number is altitude of water surface, in feet above sea level

● S57455 30.1 Observation well

♦ S62238 22.6 Observation well with continuous recorder

▲ 01306460 SURFACE-WATER STATION AND NUMBER

Figure 6. Water-table altitude in September 1977 and location of selected observation wells and surface-water data-collection stations on Connetquot Brook. (Modified from Prince, 1980, fig. 7.)

The regional water-table contours (fig. 6) generally are perpendicular to the stream channel in the northern part of the basin and bend upstream as they approach the stream channel, as is true for most of Long Island's south-shore streams (see also fig. 3*A*). However, the unusual channel configuration of Connetquot Brook from the confluence of the unnamed tributary north of wells S57474 and S57475 to gaging station 01306460 (fig. 6) is essentially parallel to the upgradient water-table contours. This water-table configuration lends itself to use as a calibration site for vertical cross-sectional modeling, as described in the section "Model Studies of the Shallow Ground-Water Flow System."

An additional characteristic that is evident in figure 6 is the tendency for the water-level contours to become more closely spaced with proximity to the stream. This may be attributed to two causes:

1. Cross-sectional areas perpendicular to flow in the shallow system that discharges to the stream probably vary by a factor of two to three as a result of the increase in water-table altitude with distance from the stream (fig. 6). In addition, the quantity of flow through successive downstream cross sections increases as a result of intermittent areal recharge, which replenishes the water table and causes a greater cumulative accretion toward the stream. Jacob (1943) showed how this resulted in the classic parabolic head profile in a one-dimensional system having distributed recharge.
2. As pointed out by Rorabaugh (1960), significant head losses may occur near and particularly beneath partially penetrating streams as a result of vertical head gradients, even in homogeneous aquifers. The lower hydraulic conductivity of the streambed sediments in comparison to aquifer material some distance from the stream tends to further increase head losses. These head losses and vertical flow components are not depicted on areal water-table maps.

Field measurements and model results (discussed in the section "Model Studies of the Shallow Ground-Water Flow System") suggest that the first cause (described above) is probably more important. The closer spacing of water-table contours near the stream could not be verified from most regional water-table maps because the maps contain larger contour intervals and fewer control points.

At sites where more than one well was installed, the wells were screened at different depths to detect vertical differences in hydraulic head. At sites having three wells, one well was screened near the water table, one at approximately 100 ft below the water table, and one at an intermediate depth. At sites more than 50 ft from the stream, a lack of measurable difference in simultaneous water-level measurements at different depths indicates that ground-water flow in the shallow flow system is essentially horizontal. Substantial differences in water levels with depth have been observed directly beneath the stream and a few feet to either side of it, however, as described in the next section.

Ground-Water Levels Near and Beneath Stream

Detailed ground-water-level measurements were made at three sites as shown in figure 7. Measurements were made in wells (piezometers) directly beneath and alongside the stream to obtain a detailed profile of head changes with depth. At all three sites (A, B, and C), one well was driven into the streambed, another into the streambank. At sites B and C, a third well was driven about 50 ft from the stream bank. In addition, at site B, a set of measurements across the stream channel was made at shallow depths beneath the streambed.

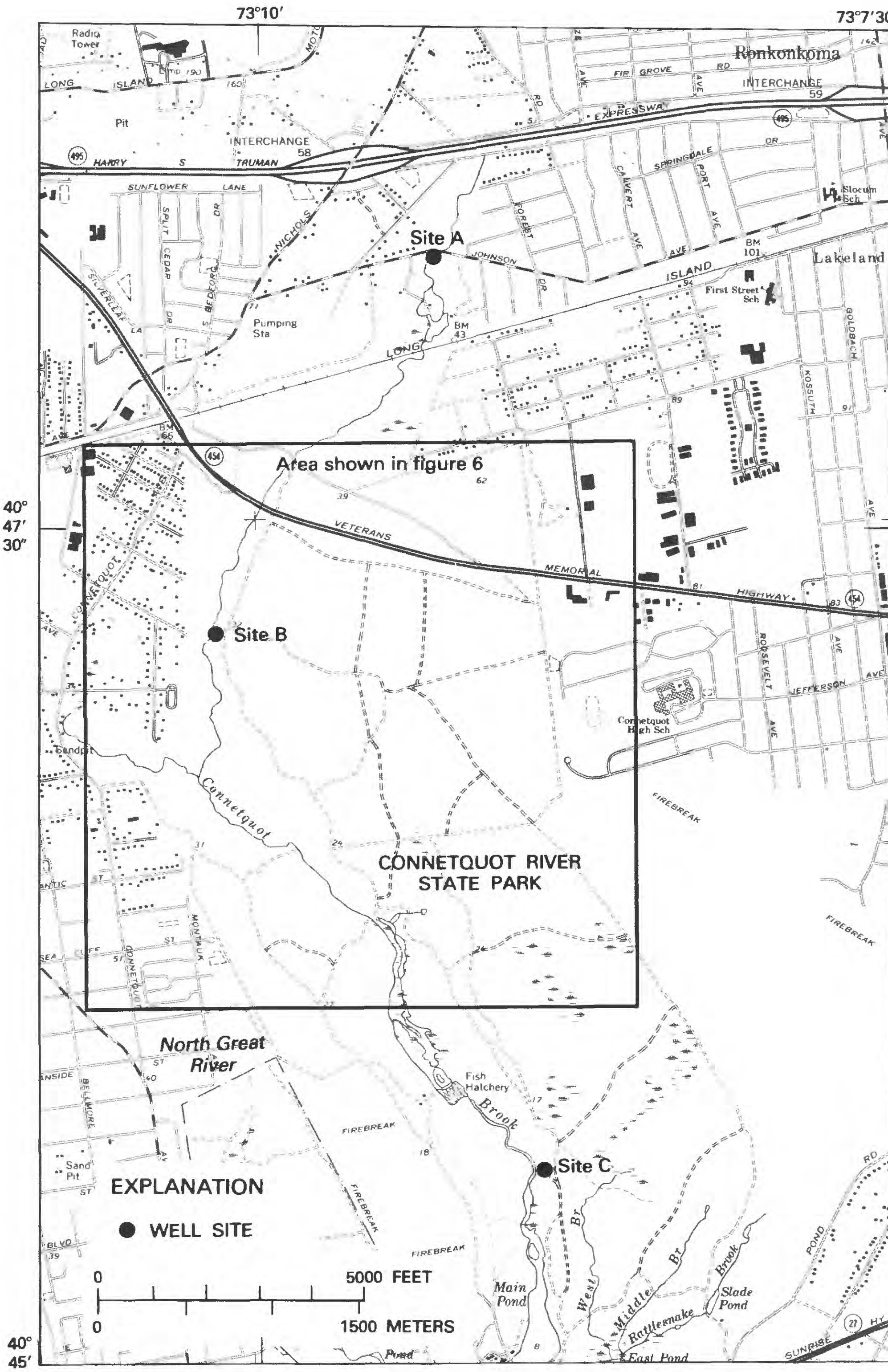
Methods of Measurement

A generalized diagram of a measuring site and the equipment used to measure ground-water levels beneath and adjacent to Connetquot Brook is presented in figure 8. A leveling instrument on the streambank permitted measurement of absolute elevations and relative altitudes by reference to a nearby benchmark as desired. A 5-ft length of well casing was attached to a well screen (length 0.875 ft), and the distance from the center of the screen to the top of the casing was measured. The well was driven by a falling hammer until the center of the well screen reached the desired depth. The well was pumped briefly, generally less than a minute, to ensure hydraulic connection with the surrounding aquifer material. After pumping, the water level in the well was measured periodically until it had fully recovered to its static level, after which the water level in the well and the stream-surface altitude were measured and recorded to 0.01-ft accuracy.

When new casing was added to a well, the altitude of the top of the well was measured both before and after the addition of casing; the difference in altitude represented the length of additional pipe. At depths where water-level changes were relatively large between successive measurements, the well was driven approximately 1 ft between measurements; where changes in water levels were relatively small, the well was driven to 5 ft or more between measurements.

Vertical Head Distribution

The results of water-level measurements at sites A, B, and C are plotted in figure 9 as three graphs showing water level with respect to depth below the streambed surface. The graphs show a generally similar shape at corresponding depths, as well as several noteworthy individual features.



Base from U.S. Geological Survey 1:24,000
Central Islip, 1967

Figure 7. Location of observation-well sites for measurement of hydraulic head with depth beneath stream, at streambank, and 50 ft from stream.

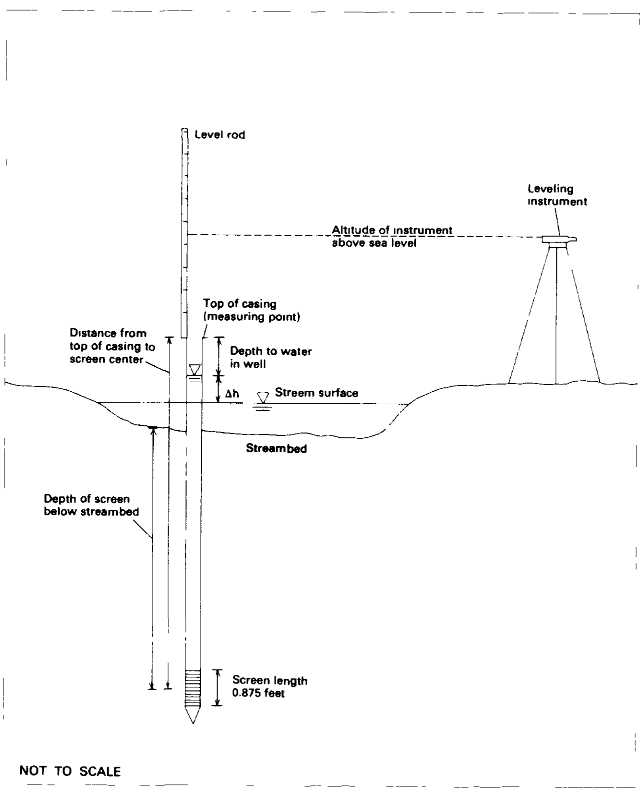


Figure 8. Generalized observation-well site and equipment used to measure ground-water levels beneath and adjacent to Connetquot Brook. (Location of sites shown in fig. 7.)

In general, the largest increases in head with depth were at the streambed wells, followed closely by the streambank wells. At sites A and B, the maximum head increase in water levels among the wells was about 1 ft, and at site C about 2 ft. Most of the head increase was within 5 ft of penetration into the streambed or below the water table (for the streambank wells) at sites A and B and within 10 ft at site C. However, measurable increases in head continued to a depth of about 30 ft beneath the streambed. Water levels in the well about 50 ft from the streambank at site B showed no discernible head change within 25 ft of penetration, and water levels in the well 50 ft from the streambank at site C showed no change within 40 ft.

The sharp changes in the slopes of the individual graphs in figure 9, which are drawn as a series of straight-line segments (labeled 1, 2, or 3), are assumed to result from (1) differences in area as ground-water flow lines converge radially toward the streambed and (2) differences in vertical hydraulic conductivity caused by local vertical variations in the grain-size distribution of the sediments. Gamma logs of the top 100 ft of the upper glacial aquifer in wells S65598, S67559, and S67561, each near one of the sites (fig. 6), clearly show variability in grain size throughout the aquifer. The vertical intervals associated with the straight-line segments of the graphs presumably correspond to local "beds" or lenses having similar permeability and ranging from <1 to 3 ft in thickness.

Table 1. Differences between stream-surface altitude and ground-water level in streambed piezometers driven 3 ft below the streambed along a traverse across the stream at site B, August 17, 1981

[Location of site B shown in fig. 7]

Location	Distance from west edge of stream (ft)	Δh (height of ground-water level above stream-surface altitude, in ft)
West edge of stream	0	0.71
Halfway between west edge and stream center	6.8	.57
Stream center	13.6	.58
Halfway between east edge and stream center	20.4	.66
East edge of stream	27.2	.65
	Average	.63

In addition to water-level measurements at selected depths at each site (fig. 9), a five-point traverse across the stream was made at site B (fig. 7) to measure the difference between stream-surface altitude and the water level in piezometers screened 3 ft below the streambed, referred to as Δh. Results are given in table 1. The procedure for measuring water levels was the same as described in the section "Methods of Measurement." The average of the five Δh measurements was 0.63 ft, and the range was from 0.57 to 0.71 ft. No consistent spatial pattern of Δh is discernible. The small differences in Δh values are attributed largely to lateral variations in the vertical hydraulic conductivity of the streambed. Measurements taken at the well at the west edge of the stream provide indirect evidence in this interpretation. A zone of low permeability at this well is suggested because the largest Δh was recorded in this well (table 1), the greatest number of hammer blows were required to penetrate the top foot of the streambed, and the well had the lowest water yield when pumped.

Calculation of Vertical Hydraulic Conductivity

Measurements of head increases with depth below the streambed were used to estimate the vertical hydraulic conductivity of the material directly beneath the streambed, the depth of the shallow ground-water flow system associated with Connetquot Brook, and the flow patterns within that system.

If constant vertical flow per unit area beneath the streambed wells is assumed, the head changes and bed thickness associated with the straight-line segments of curves in figure 9 can be used to estimate the ratio of vertical hydraulic conductivity between any two beds. For

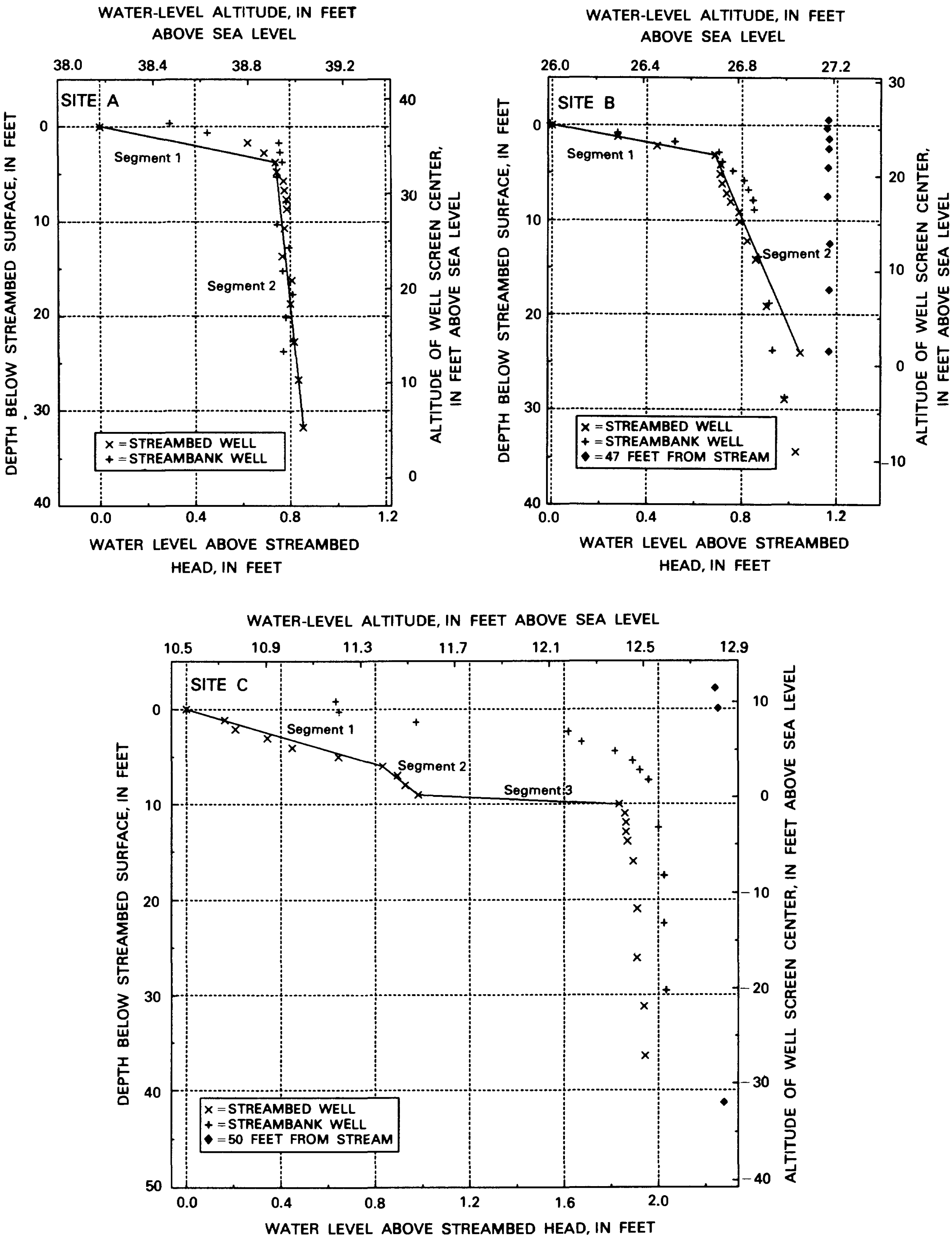


Figure 9. Head increase with depth beneath stream, at streambank, and 50 ft from streambank, at three sites. (Site locations shown in fig. 7.)

Table 2. Sediment thickness, head loss per unit thickness, and estimated ratio of vertical hydraulic conductivity of lower sediment to that of the streambed of Connetquot Brook
[Values based on measured vertical hydraulic gradients. —, no data]

Site	Graph segment in figure 9	Sediment thickness (L) represented by graph segment (ft)	Head loss per unit sediment thickness ($\Delta h/L$)	Ratio of streambed hydraulic conductivity $(K_z)_1$ to that of lower unit $(K_z)_n$*
A....	1	3.8	0.1964	—
	2	28.0	.0041	47.90
B....	1	3.1	.2188	—
	2	20.9	.0173	12.65
C....	1	6.0	.1375	—
	2	3.0	.0501	2.57
	3	1.0	.8471	.16

*The ratio is determined by the relationship $\frac{(K_z)_n}{(K_z)_1} = \frac{\left(\frac{\Delta h}{L}\right)_1}{\left(\frac{\Delta h}{L}\right)_n}$.

this special case of assumed one-dimensional vertical flow, Darcy's law gives:

$$\frac{(K_z)_a}{(K_z)_b} = \frac{\left(\frac{\Delta h}{L}\right)_b}{\left(\frac{\Delta h}{L}\right)_a}$$

where

K_z = vertical hydraulic conductivity of a single lens or bed,
Δh = head loss across an individual lens or bed,
L = thickness of an individual lens or bed, and
a and b = arbitrary designations for any two lenses or beds.

Therefore, the ratio of vertical hydraulic conductivity values for any two beds, a and b, is inversely proportional to the vertical hydraulic gradient across those beds. The relevant data from figure 9 are listed in table 2. In the last column of this table, the vertical hydraulic conductivity of lower units is related to the vertical hydraulic conductivity of the upper unit (graph segment 1), where the upper unit represents the streambed material as delineated by the water-level measurements.

Commonly, ground-water movement between the stream and aquifer is modeled as a uniform leaky bed; this model requires an estimate of both the thickness and the vertical hydraulic conductivity of the streambed. The vertical hydraulic conductivity of the streambed material is generally assumed to be lower than the hydraulic conductivity of the adjacent aquifer. (The curves in figure 9 and the hydraulic-conductivity ratios in table 2 support this assumption.) The hydraulic-head data in figure 9 and in table 2 and the stream-seepage data in table 3 provide a means of estimating the streambed thickness and hydraulic conductivity. The streambed thickness was estimated as being the uppermost zone of low hydraulic conductivity (segment 1) in the curves shown in figure 9. If vertical flow beneath the streambed is assumed, then estimates of the vertical hydraulic conductivity of the streambed can be calculated by applying Darcy's law to the head and stream-seepage data. The resulting values for sites A and B (see fig. 7 for location) range from 11 to 15 ft/d; pertinent data are listed in table 3. Note that the vertical hydraulic conductivity values in table 3 are only order-of-magnitude estimates and that the individual data are assumed to represent "average" values for long reaches of the stream, particularly site B. Thus, a "true" average value for the individual data might differ considerably from the estimated value given in table 3. Nevertheless, the values given provide a reasonable range for this parameter.

Table 3. Vertical hydraulic conductivity of Connetquot Brook streambed at two sites based on field measurements of stream seepage and head loss across the streambed
[Site locations shown in fig. 7]

Term	Site A	Site B	
	8–23–79	10–10–78	8–17–81
Distance between streamflow-measurement sites (ft)......	2,000	9,500	9,500
Increase in discharge between sites (ft^3/s)........	1.90	19.5	12.9
Ground-water seepage per unit stream length [(ft^3/s)/ft].....	0.950×10^{-3}	2.05×10^{-3}	1.36×10^{-3}
Average stream width (ft).	34	53	53
Depth of screen center below streambed (ft).......	3.75	3.13	3.0
Head difference between streambed piezometer and streambed (ft).	.74	.68	.63
Vertical hydraulic conductivity of streambed (ft/d).........	12	15	11

Flow Patterns

Water-level measurements near Connetquot Brook were used to define the flow patterns and location of the bottom boundary of the shallow ground-water system. A vertical section showing the water levels measured at site B is given in figure 10.

The data in figures 9 (sites A, B, and C) and 10 (site B) indicate that water levels are virtually constant with depth at the wells about 50 ft from the streambank at sites B and C. Thus, at this distance from the stream, shallow ground-water flow is essentially horizontal. In the streambank and streambed wells, water levels increase with depth, but the increases at depths between 20 and 30 ft below the streambed become so small as to be within the error of measurement. The depth below the streambed at which vertical (upward) gradients are no longer measurable is assumed to represent the local bottom of the shallow ground-water system that discharges to Connetquot Brook. Water below this depth is therefore part of the deeper regional flow system.

The head contours in figure 10 (site B) indicate that flow in the shallow ground-water system is horizontal to within a few feet of the stream and that the head dissipated by vertical flow is only about 1 ft, most of which occurs within 3 ft of the streambed.

Concentration of Tritium and Dissolved Oxygen

Tritium and dissolved-oxygen concentrations were measured to help delineate the shape and flow patterns of the shallow ground-water flow system near Connetquot Brook. Also, most standard constituents and characteristics were measured during the study.

Tritium

Sixteen samples of ground water and one sample of stream water were collected and analyzed for tritium concentrations by the U.S. Geological Survey Tritium Laboratory in Reston, Va. Results are given in table 4. All samples were predistilled, and samples having a tritium concentration greater than 2.7 T.U. (tritium units) were enriched by electrolysis from 100 mL and counted by a Packard 3255 Liquid Scintillation Counter.[1] Samples having tritium concentrations of 2.7 T.U. or less were enriched from 470 mL and counted by a gas proportional counter. Tritium values were corrected for decay to the collection date on the basis of a tritium half-life of 12.361 years (T.A. Wyerman, U.S. Geological Survey, written commun., 1979).

Interpretation of tritium to establish the age of water is difficult because of the effects of (1) variations in tritium input levels through time, (2) evapotranspiration of precipitation, and (3) hydrodynamic dispersion. Ferronsky and Polyakov (1982) discussed several models of hydrodynamic dispersion and their effects upon tritium concentrations in ground water. For the purposes of this study, a less rigorous interpretation of the tritium data is sufficient. Samples of water having high concentrations of tritium (>10 T.U.) can be assumed to contain mostly post-1952 water, and samples having low tritium concentrations (<1 T.U.) are assumed to contain pre-1952 water. Tritium concentrations in the 1- to 10-T.U. range probably indicate that a small quantity of post-1952 precipitation containing high tritium has mixed with a large quantity of pre-1952 water containing little or no tritium.

The tritium data in table 4 may be interpreted on the basis of (1) the half-life of tritium, (2) the beginning of atmospheric testing of nuclear devices in 1952, which resulted in a marked increase in the tritium content of precipitation, and (3) estimated pre-1952 tritium concentrations of about 8 T.U. for precipitation (Ferronsky and Polyakov, 1982). From these criteria, the data in table 4 may be grouped into four tentative categories: (1) water having a concentration of 0.3 T.U. or less and predating atmospheric testing (well S57479 and October sampling at wells S57478, S62237, and S65598); (2) water having concentrations of 0.5 to 0.6 T.U. and largely predating 1952 (wells S65599 and S65597); (3) water having a concentration of 2.7 T.U. and containing a mixture of pre- and post-1952 water (well S65595); and (4) water having concentrations greater than 25 T.U. and that entered the system after 1952.

Interpretation of the data in table 4 requires consideration of the depth of the well screen below the water table and of the location of the well in relation to the stream and other local system boundaries, such as an interstream divide. The first two groups of wells are adjacent to the stream; the second two groups are several thousand feet from it (see also fig. 6).

The velocity of shallow ground-water flow in the area is useful in interpreting the data in table 4. By using average values of hydraulic conductivity for the upper glacial aquifer (McClymonds and Franke, 1972) and water-table gradients, the average velocity of shallow ground water (the velocity, or specific discharge, calculated from Darcy's law divided by an average value of aquifer porosity) was calculated to be about 500 ft/yr. Thus, ground water whose tritium concentration indicated an age older than 1952 must have entered the ground-water system 10,000 to 15,000 ft or more upgradient of its present location.

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[1]Use of trade names in this report is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey.

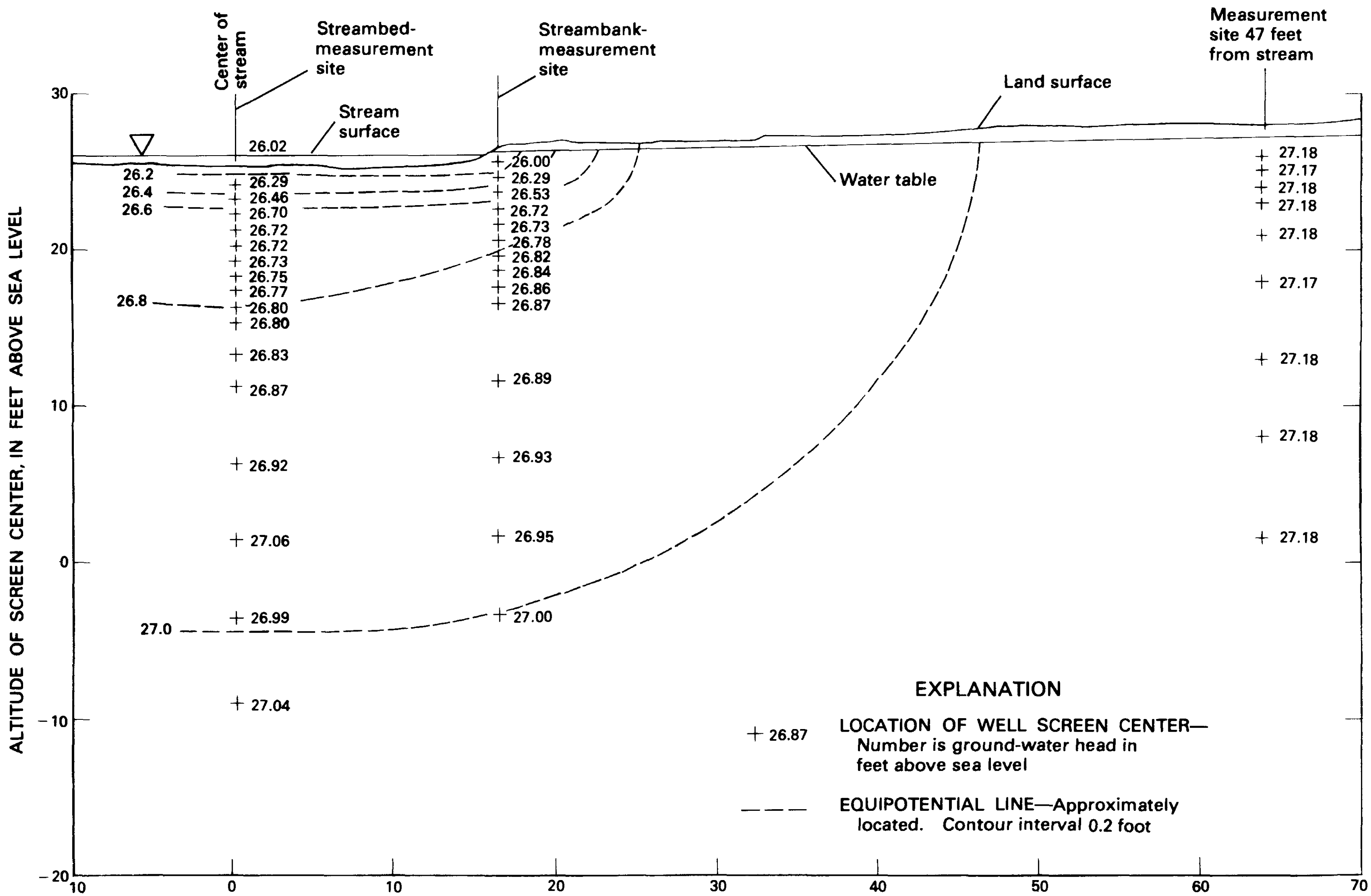

Figure 10. Approximate location of equipotential lines in the shallow ground-water flow system near site B on Connetquot Brook, October 1978. (Site location shown in fig. 7.)

Table 4. Tritium and dissolved-oxygen concentration in ground water and stream water from selected wells near Connetquot Brook

[Well and site locations shown in fig. 6. —, no sample collected on that date. T.U., tritium unit]

Well number	Lateral distance of sampling point from stream (ft)	Depth of center of well screen below water table (ft)	Altitude of center of well screen (ft above sea level)	Tritium (T.U.)		Dissolved oxygen (mg/L)	
				January 1978	October 1978	October 1979	February 1980
S57478	10	4	12	0.9	0.1	8.4	9.6
S57479	10	9	8	.3	—	7.1	9.4
S62237	10	41	−25	.2	0	4.4	9.1
S65595	10	2	25	—	2.7	—	—
S62238	10	40	−13	—	—	1.6	.25
S65598	10	99	−73	—	.1	1.9	1.25
S57468	4,000	1	28	—	31.9	10.1	9.4
S57469	4,000	22	7	—	26.8	—	—
S62234	4,000	41	−12	—	30.8	7.2	6.7
S65599	4,000	97	−71	—	.6	—	—
S57466	2,500	5	24	31.0	35.2	6.7	5.6
S57467	2,500	45	−17	130.	166.	6.7	8.9
S65597	2,500	98	−70	—	.5	6.7	8.3
Stream-flow gaging station 01306460........				—	32.0	—	—

Wells several thousand feet from the stream differed markedly in tritium concentration with depth. Those screened 95 to 100 ft below the water table contained less than 1 T.U., whereas those screened 40 to 45 ft below the water table exceeded 25 T.U. This difference suggests that the boundary between the shallow and deeper flow systems at these distances from the stream is between these two depths. At both locations, water near the water table contained more than 30 T.U., a concentration that is strongly indicative of water originating after 1952. Although the three shallower samples from the site 4,000 ft from the stream had virtually uniform tritium concentrations (25 to 30 T.U.), the well screened at an intermediate depth at the site 2,500 ft from the stream site (S57467) had a significantly greater tritium concentration (166 T.U.) than S57466, the well screened near the water table (about 35 T.U.). The high tritium concentration of water at intermediate depth suggests that this water was derived from precipitation that fell from the late 1950's to mid-1960's, when the tritium concentration of precipitation was at its maximum from atmospheric testing of atomic devices. These measurements also suggest that a "slug" of recharge from precipitation can retain its identity without extensive vertical mixing for 15 years or more in the shallow ground-water flow system.

All tritium concentrations in water from the two sites near the stream (table 4) indicate that the water there either originated before 1952 or contained only small amounts of post-1952 water (well S57478 in January 1978 and well S65595). Because all shallow ground water converges at the streambed (the discharge boundary of the shallow ground-water flow system), waters of differing ages were found close to one another in the streambed vicinity. The pattern and location of flow lines adjacent to the streambed do not remain fixed in time, however, because fluctuations in the water-table altitude cause the local flow lines to shift, which causes mixing. The difference in tritium concentrations of the two samples from well S57478 (0.1 and 0.9 T.U.) was attributed to such mixing and (or) to small shifts in the local pattern of ground-water flow.

Water entering the stream is a mixture of young and old water; this mixing would be expected to give average tritium concentrations. The low values of tritium in well S57478 (0.1 and 0.9 T.U.) were unexpected because very shallow ground water would seem to be of local origin and therefore of young age. This well is only a few feet from the main stream channel. A few feet to the east and extending for several thousand feet up- and downstream is a swamp area that is ponded during wet periods. If local recharge having short, shallow flow lines discharges into the swamp, the ground water in the shallow streambank well originates upgradient of the swamp and is therefore older water.

The single sample of stream water had a tritium concentration of 32 T.U. (table 4), which was assumed to be representative of the tritium concentration of precipitation at the time of sampling.

Dissolved Oxygen

As ground water moves along flow lines, the concentration of dissolved oxygen would be expected to slowly decrease through chemical and biological reactions within the surrounding earth material (Freeze and Cherry, 1979, p. 245). Thus, the concentration would indicate the relative distance traveled in the system. Also, a marked decrease in dissolved-oxygen concentration within a certain depth interval might indicate the boundary between the shallow and deeper flow systems. On the basis of this hypothesis, dissolved-oxygen concentration (table 4) was measured in wells in the Connetquot basin.

The procedure for measuring dissolved-oxygen concentration was to first lower the dissolved-oxygen probe to a depth opposite or slightly above the well screen. The hose of a centrifugal pump was then inserted 1 to 2 ft below the water surface in the well, and the well was pumped continuously at a low rate. As pumping continued, the probe was kept in continuous motion as readings of dissolved oxygen were made. When the readings stabilized to a constant value for several minutes, this value was recorded as representative of the aquifer water opposite the well screen.

All samples except those from wells S62238 and S65598 in table 4 had high concentrations of dissolved oxygen (4.4 to 10.1 mg/L). These high concentrations are attributed to (1) the sandy soil, which permits rapid infiltration of rainfall, (2) a shallow water table (less than 15 ft below land surface in most places), and (3) little or no organic matter in the surficial material. No clear pattern of depth or location is discernible from these measurements. The low dissolved-oxygen concentration at wells S62238 and S65598 is unexplained.

In conclusion, the failure of the dissolved-oxygen measurements to reveal the depth or flow patterns of the shallow ground-water flow system are tentatively attributed to the small thickness of the shallow system, the possible mixing of shallow water with slightly deeper water, and the lack of minerals and organic matter that remove dissolved oxygen through oxidation.

Conclusions from Field Studies

The principal conclusions from the field studies are summarized briefly as follows:

1. The shallow flow system adjacent to and immediately beneath the stream is 20 to 30 ft thick at sites where detailed water-level measurements were made.
2. The thickness of the shallow flow system several thousand feet from the stream may be somewhat greater than the thickness beneath the stream, as evidenced by tritium concentration of ground water with depth at two sites.
3. Flow in the shallow system is almost horizontal at distances beyond 50 ft from the stream.
4. Total head dissipation immediately adjacent to and beneath the stream, due primarily to vertical flow beneath the streambed, does not exceed 3 ft.
5. The vertical hydraulic conductivity of the streambed material (3 to 6 ft thick) at the three measurement sites (A, B, and C) was 0.02 to 0.36 times the vertical hydraulic conductivity of the sediments immediately beneath it (table 2). Estimates for a spatially averaged vertical hydraulic conductivity of the streambed ranged from 11 to 15 ft/d (table 3). Measured head loss per unit thickness through the uppermost 3 to 6 ft of the streambed ranged from 0.1375 ft to 0.2188 ft (table 2).

MODEL STUDIES OF THE SHALLOW GROUND-WATER FLOW SYSTEM

The Connetquot Brook model studies were conducted to investigate the interaction between ground water and streams on Long Island. As a part of the study, a calibrated two-dimensional cross-sectional model was developed, and a sensitivity analysis was done to evaluate the various parameters that control flow within this system. Although the simulations in this study relate specifically to the system near the middle reach of Connetquot Brook, many of the conclusions regarding the role of individual parameters within this shallow system can also be applied to other shallow systems on Long Island.

Only two-dimensional steady-state flow in the shallow system was simulated in this study because (1) the field studies indicated vertical flow to be the critical factor in this system and (2) the time and effort needed for a three-dimensional transient-state model were beyond the project resources.

The numerical simulations were carried out by using a finite-difference modular-format computer code developed by McDonald and Harbaugh (1984). This code is capable of meeting all requirements relating to variable-grid spacing and simulation of appropriate boundary conditions.

Description of Model

In a two-dimensional vertical cross-sectional model, the cross section must follow a flow line (flow surface in vertical dimension) to eliminate the need to account for flow orthogonal to the cross section. Two important simplifying assumptions that were made in the selection of this cross section were that (1) the shallow system could be represented as independent of the regional flow system and (2) the vertical section through the shallow system, which followed a flow line defined by the water table, had no flow orthogonal to it below the water table.

The first assumption is considered valid because, even though heads and flow within the shallow system are related to and dependent on the regional system, the two systems are separated by a flow line (flow surface in three dimensions). If the recharge and discharge boundaries of the shallow flow subsystem are accurately represented in the model, the bounding flow line at the bottom of the shallow flow system can be treated as a no-flow boundary. As long as the system is in equilibrium, the depth of this

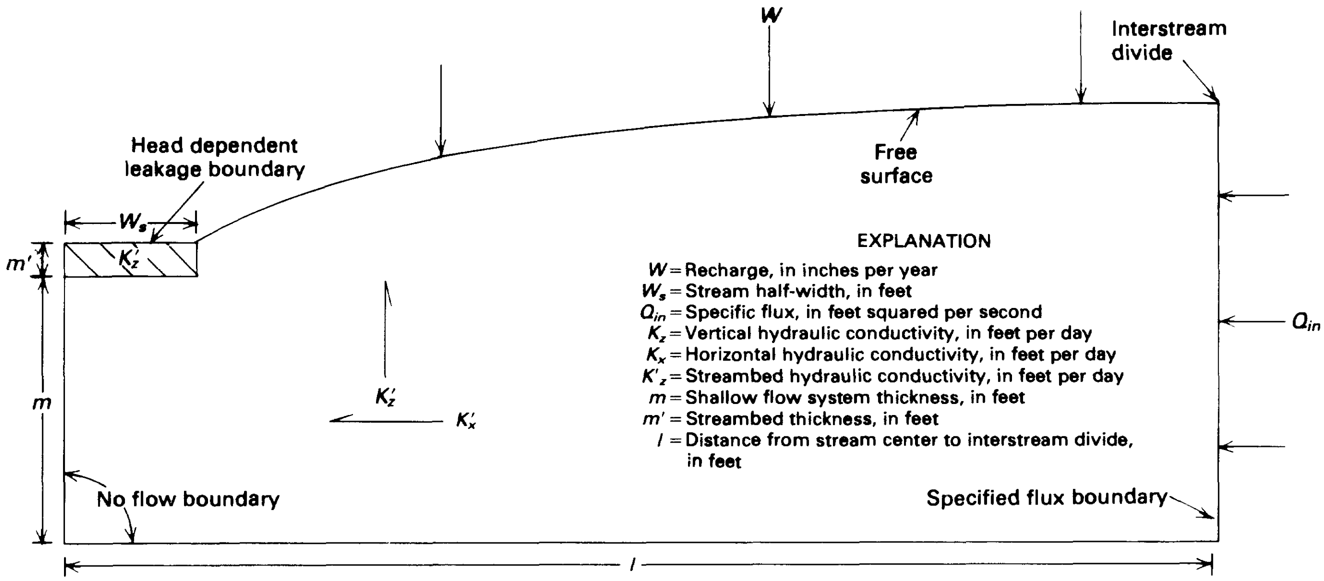

Figure 11. Schematic model cross section of the shallow ground-water flow system associated with a stream and the assumed boundaries and parameters that govern flow.

no-flow boundary will not move, and a model, therefore, can represent only the shallow system.

The second assumption means that the direction of horizontal flow is the same at all depths within the shallow system. This assumption allows the system to be represented by a two-dimensional model that is defined along a stream line determined from a water-table map. To corroborate or refute this assumption with field data would be virtually impossible; however, for the purposes of a sensitivity analysis under equilibrium conditions, this assumption is probably reasonable for a vertical section perpendicular to the longitudinal axis of the stream.

Boundary Conditions

As in any model study, development of the model required a series of simplifying assumptions about the flow system and its boundaries. The flow system and boundaries for the model cross section as conceptualized for this study are depicted in figure 11. The modeled ground-water system is assumed to be in equilibrium, where inflow equals outflow with no change in storage (that is, all model simulations are steady state). If the natural flow system is in equilibrium, the flow-line boundaries at the interstream divide and between the shallow flow system and the regional flow system are stationary. Therefore, if the modeled system dimensions are selected so that the interstream ground-water divide and the stream-line boundary between the shallow ground-water flow system and the regional flow system are correctly located, these boundaries can be simulated as impermeable boundaries.

Field investigations have provided an estimate of the average thickness (m) of the shallow flow system and hence the depth of its bottom boundary, but the distance (l) to the interstream divide is uncertain. If the location of the interstream divide in the model is correct, the divide can be established as an impermeable boundary. If the model cross section is shorter than the distance from the stream center to the true interstream divide, however, the quantity of water applied to the model as recharge will be less than the quantity that enters the natural system. To simulate the flow system accurately, this additional quantity of water must be represented either (1) by expanding the model grid to the proper distance to coincide with the true interstream divide or (2) by introducing a specified-flux boundary (Q_{in}) to provide the proper quantity of water. Because using Q_{in} is more flexible and easier than extending the model, it was selected as the more favorable alternative.

In an idealized aquifer system having straight, regularly spaced stream channels, ground-water flow is symmetric about the central axis of the streambed, and ground-water flow lines from either side of the stream do not cross an imaginary vertical plane beneath the streambed center. Therefore, the model was designed to simulate flow on only one side of the stream—from the center of the stream to one interstream divide. The vertical boundary at the center of the stream was simulated as an impermeable boundary.

The upper boundary of the model, except for the stream, is a freely moving water table that receives uniform areal recharge (W). Simulating the upper boundary as a free surface is inherently correct, but uniform areal recharge is a necessary simplification of the natural system. Uniform areal recharge ignores (1) the intermittent nature of natural recharge, (2) the effects of direct

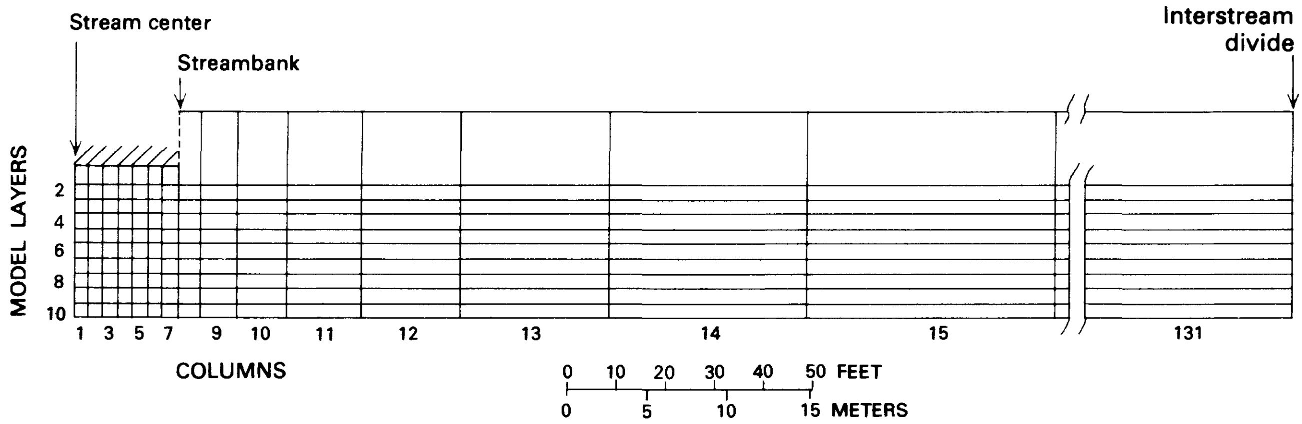

Figure 12. Finite-difference grid of cross-sectional model for simulation of the shallow flow system associated with a stream. Hachures over grid blocks 1–6 in layer 1 indicate streambed head-dependent leakage boundary.

runoff to the stream, (3) the flow in the unsaturated zone, and (4) the variations in evapotranspiration rate as the depth to the water table changes locally and through time. Steady-state flow is assumed because average recharge rates would satisfactorily simulate uniform flow in the ground-water system. No other complicating factors were simulated because they were judged to have little or no effect upon the system under the conditions being investigated.

The discharge boundary at the stream is affected by streambed geometry, thickness, and permeability and by the difference in hydraulic head between the stream and the aquifer. The discharge boundary was simulated as a head-dependent leakage boundary for which the thickness of the streambed, the area of the discharge boundary (streambed area), the vertical hydraulic conductivity of the streambed, and the head in the overlying stream are specified. Thus, the discharge through the streambed boundary is dependent upon the hydraulic gradient between the aquifer and the stream.

Model Grid

A variable rectangular grid (fig. 12) was used to represent the shallow flow system as indicated by the field studies. The grid represents an area of aquifer 6,000 ft long from the center of the stream to the assumed location of the interstream divide. The initial thickness from the bottom of the shallow flow system to the streambed is 30 ft. The vertical dimension is divided into 10 layers; the lower nine are 3 ft thick, and the uppermost is 3 ft thick beneath the stream but elsewhere is dependent upon water-table altitude above the streambed (fig. 12). The grid also is divided into 131 columns that are each 3 ft wide beneath the stream (cols. 1–6) but gradually increase to 50 ft wide at a distance of 150 ft from the stream center. From 150 ft from the stream center to 6,000 ft from the stream, the columns remain at 50-ft widths. The streambed half-width was initially set at 18 ft (fig. 12).

Sensitivity Tests

Ground-water movement in the shallow flow system is dependent upon several parameters that can be specified in the numerical model. Not all the parameters have equal effect upon the head distribution or on the flow patterns in the model; large changes in some may have little effect, while small changes in others may have considerable influence.

A systematic sensitivity analysis requires that pertinent parameters be identified and their effect on each other evaluated.

Range in Numerical Values of Parameters Studied

Head in the shallow flow system can be defined as a function of 10 dimensional parameters (see fig. 11):

$$h = f(W_s, m', m, l, W, K_z, K_x, K'_z, Q_{in})$$

where

h = head (L);
W_s = stream half-width (L);
m' = streambed thickness (L);
m = thickness of shallow flow system (L);
l = distance from stream center to interstream divide (L);
W = constant areal recharge (L/T);
K_z = vertical hydraulic conductivity (L/T);
K_x = horizontal hydraulic conductivity (L/T);
K'_z = streambed vertical hydraulic conductivity (L/T);
Q_{in} = flow into system from right-side lateral boundary (L^2/T); and
f = some function.

For the purposes of sensitivity analysis, head (h) is a measure of system response to changes in other variables; W_s, m', m, and l define the idealized configuration of the shallow system; W and Q_{in} define values of flux at two boundaries; and K_x, K_z, and K'_z define a distribution of hydraulic-conductivity values in the model cross section.

To simplify the process of analyzing the effects of these parameters upon the system, the 10 dimensional parameters were combined into eight dimensionless parameters through the π theorem (Bridgman, 1978, p. 40), as follows:

$$\frac{h}{l} = f\left\{\frac{W_s}{l}, \frac{m'}{l}, \frac{m}{l}, \frac{W}{K_x}, \frac{K'_z}{K_x}, \frac{K_z}{K_x}, \frac{Q_{in}}{K_x l}\right\},$$

where h/l is the dependent variable. Detailed systematic sensitivity tests were conducted on six of the remaining seven dimensionless parameters. (Streambed thickness, m'/l, was assumed to be a known constant derived from field measurements.) Values for each of the independent parameters were varied individually between anticipated extremes, and successive model runs were made to obtain head distributions for each independent hydrologic condition. Only one parameter was changed at a time; all others were set at values used in an initial reference model run. Data from the 12 runs plus the reference run were evaluated through inspection of h/l values beneath the streambed and at the water table near the interstream divide and by comparing the simulated values with measured h/l values from the same locations.

Values for the reference model parameters were the best estimates obtained from McClymonds and Franke (1972), Cohen and others (1968), the investigations mentioned in the "Field Studies" section, and field experience. The initial values used in the model were:

constant areal recharge	W	= 20 in/yr
horizontal hydraulic conductivity	K_x	= 300 ft/d
vertical hydraulic conductivity	K_z	= 30 ft/d
streambed vertical hydraulic conductivity	K'_z	= 6.5 ft/d
constant flux at lateral boundary	Q_{in}	= 0 ft²/s
half-width of stream	W_s	= 18 ft
length of cross section	l	= 6,000 ft
thickness of shallow flow system beneath stream	m	= 30 ft
streambed thickness	m'	= 3 ft

In all tests, head or stream-surface altitude at the head-dependent discharge boundary was zero; thus, all predicted model heads were referenced to this zero altitude.

Results of Sensitivity Tests

The values of the six parameters that were investigated in the sensitivity analysis are shown in table 5; the corresponding plots of simulated and measured h/l values beneath the stream are shown in figure 13, and those at the interstream divide are shown in figure 14. The model results pertaining to each term are discussed individually in the following paragraphs.

Table 5. Values of dimensional parameters and of combined dimensionless parameters investigated in sensitivity analysis

Dimensional parameters		Combined dimensionless parameters	
Name and symbol	Values tested	Symbol	Values tested
Stream half-width (W_s) (ft)	9 18 27	W_s/l [1]	1.5×10^{3} 3.0×10^{-3} 4.5×10^{-3}
Shallow system thickness (m) (ft)	18 30 39	m/l	3.0×10^{-3} 5.0×10^{-3} 6.5×10^{-3}
Recharge (W) (in/yr)	16 20 24	W/K_x [2]	1.2×10^{-5} 1.5×10^{-5} 1.8×10^{-5}
Vertical hydraulic conductivity (K_z) (ft/d)	300 30 3	K_z/K_x	1 1.0×10^{-1} 1.0×10^{-2}
Streambed vertical hydraulic conductivity (K'_z) (ft/d)	65 6.5 .65	K'_z/K_x	2.16×10^{-1} 2.16×10^{-2} 2.16×10^{-3}
Boundary inflow (Q_{in}) (ft²/s)	0 2.0×10^{-4} 4.0×10^{-4}	$Q_{in}/K_x l$	0 9.60×10^{-6} 1.92×10^{-5}

[1] l = system length (ft).
[2] K_x = horizontal hydraulic conductivity (ft/d).

Stream half-width divided by cross-sectional length, W_s/l (figs. 13A and 14A).—The cross sectional length (l) was held constant; therefore the only variable in this dimensionless parameter was the half-width of the streambed (W_s).

The flow system would be expected to be sensitive to the width of the streambed because the streambed is the only route (other than evapotranspiration, which was ignored in this study) through which water can discharge from the system. The stream half-width (since only half the system was modeled) was set at 18 ft in the reference model run and tested at 9 ft and 27 ft. The model grid was changed to accommodate the changing stream half-width by adding or removing blocks from the grid. This method was preferable to changing the dimensions of a fixed number of blocks beneath the stream because adding or removing blocks avoided errors that might arise from changing the size of the grid blocks. As expected, values of h/l beneath the stream were sensitive to changes in the width of this boundary (fig. 13A), but those near the divide were not (fig. 14A). The simulated values of h/l beneath the stream increased by 137 percent as the stream width (W_s) was decreased from 27 ft to 9 ft.

System depth divided by cross-sectional length, m/l (figs. 13B and 14B).—Again, because the cross-sectional length of the model (l) was constant, the only variable in

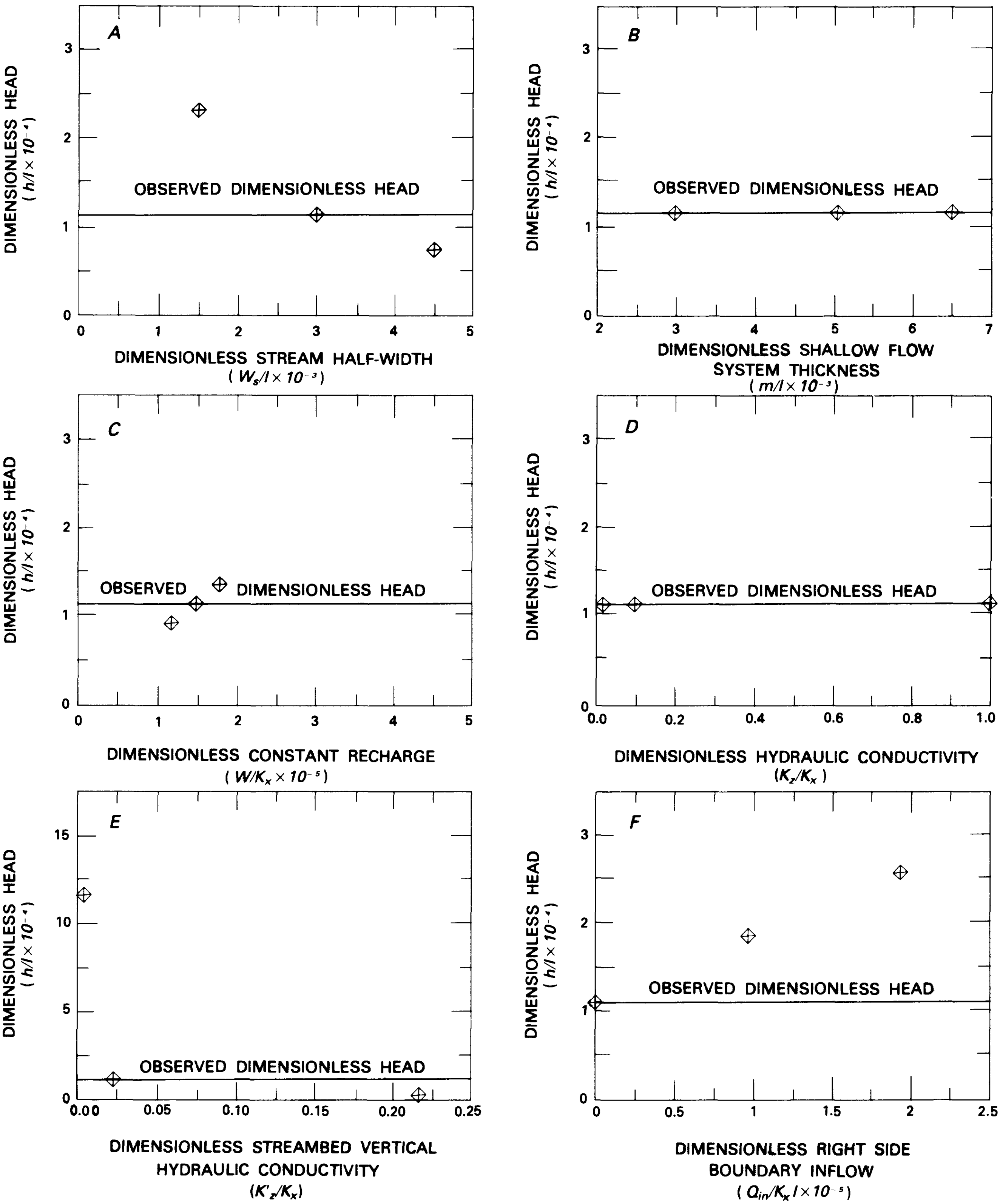

Figure 13. Simulated and observed values of *h/l* versus dimensionless flow-system parameters for a location 4.5 ft beneath streambed.

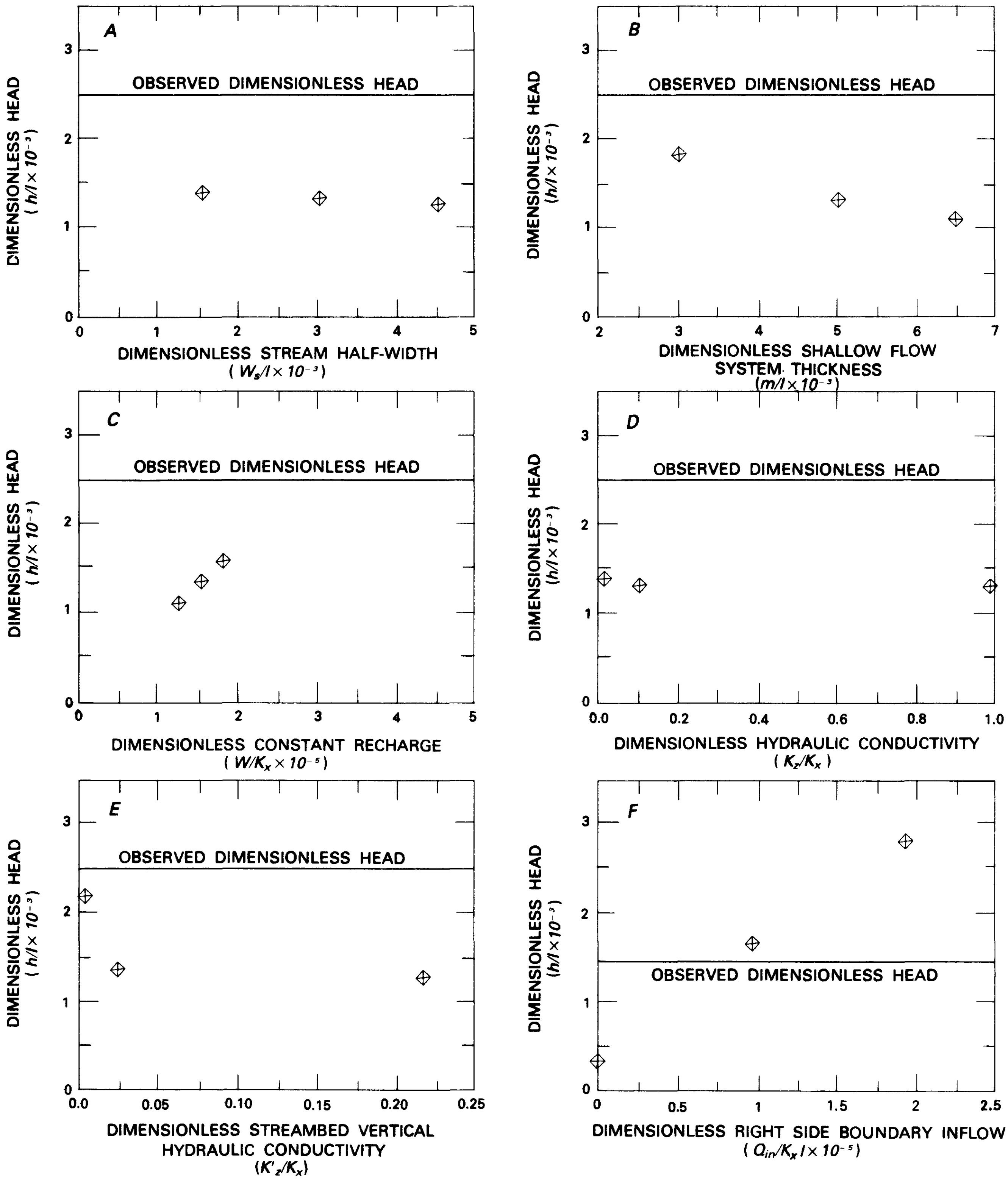

Figure 14. Simulated and observed values of h/l versus dimensionless flow-system parameters near the shallow flow-system divide, 6,000 ft west of stream center.

this dimensionless parameter was the depth of the shallow flow system (m). System depth was one of the parameters that was extensively investigated in the field studies because of its perceived influence upon the movement of ground water through both the regional and shallow flow systems. In this sensitivity analysis, 3-ft layers were added or eliminated from the model to change the system depth from 18 ft to 39 ft. Again, the size of the grid blocks was held constant to avoid errors due to a change in block size. Only the values of h/l near the interstream divide were sensitive to this term (fig. 14*B*). The simulated value of h/l at the divide increased by 51 percent as the system depth (m) was decreased from 39 ft to 18 ft.

*Recharge divided by horizontal hydraulic conductivity, W/K_x (figs. 13*C* and 14*C*).*—The dimensionless head parameter h/l was sensitive to the changes in W/K_x both beneath the streambed and near the interstream divide (figs. 13*C* and 14*C*). Because the parameters that make up the dimensionless parameter are both variable, their individual effects on the system cannot be isolated at this stage of analysis. However, only recharge (W) was varied in these runs. Simulated h/l increased 40 percent beneath the stream and 37 percent at the divide, as recharge (W) was increased from 16 to 24 in/yr.

Aquifer-anisotropy ratio, K_z/K_x (figs. 13D and 14D).—The rate of water movement in the ground-water flow system is dependent upon the hydraulic gradient and hydraulic conductivity. Moreover, the direction of flow in an isotropic aquifer (one having uniform characteristics in all directions) is perpendicular to equipotential lines, but if the aquifer is anisotropic (hydraulic conductivity is higher in one plane or direction than in another), the direction of ground-water flow will not necessarily be orthogonal to equipotential lines. To investigate the effects of anisotropy upon the shallow flow system, the ratio of vertical to horizontal hydraulic conductivity (K_z/K_x) was decreased from 1:1 to 1:100 by reducing the vertical hydraulic conductivity from 300 ft/d to 3 ft/d. This decrease had little or no effect on h/l values either beneath the streambed or near the interstream divide (figs. 13*D* and 14*D*). The maximum change in h/l was less than 6 percent at the interstream divide as K_z/K_x was increased from 1:1 to 1:100.

Streambed vertical hydraulic conductivity divided by horizontal hydraulic conductivity, K'_z/K_x (figs. 13E and 14E).—A reasonable range of values for this dimensionless parameter was difficult to estimate. All other parameters that were represented in the dimensionless parameters were either investigated in the field or could be estimated from data in the literature. The horizontal hydraulic conductivity value used in the reference model run was consistent with many past simulation studies on Long Island. Also, the streambed vertical hydraulic conductivity probably did not exceed the areal vertical hydraulic conductivity of the aquifer. The value of K'_z used in the reference model run was chosen such that the head decrease through the streambed was similar to that observed in the field. The K'_z values tested ranged from 0.65 ft/d to 65 ft/d, two orders of magnitude. The h/l values obtained from these simulations showed that the head near the divide is relatively insensitive to this parameter within the broad range investigated (fig. 14*E*). In contrast, the values of h/l beneath the streambed were quite sensitive to this parameter (fig. 13*E*), as might be expected. Values of h/l increased about 71 percent near the divide and from 0.10 to 11.60, or more than 2 orders of magnitude, beneath the stream as K'_z was increased from 0.65 ft/d to 65 ft/d.

Lateral inflow into the cross section divided by horizontal hydraulic conductivity times length, $Q_{in}/K_x l$ (figs. 13F and 14F).—This combined dimensionless parameter contains two dimensional parameters that may have an important influence on the flow field. Model response to variation in parameter $Q_{in}/K_x l$ is an indirect measure of the system's sensitivity to the distance from the center of the stream to the interstream divide because the boundary flux accounts for recharge entering the system beyond the model stream-to-divide distance (l) used in the analysis. The sensitivity analysis indicated that h/l is sensitive to this parameter beneath the streambed and also near the interstream divide (figs. 13*F* and 14*F*). Values of h/l increased by 77 percent beneath the stream and by 92 percent at the divide when a Q_{in} value of 4.0×10^{-4} ft^2/s was applied to the model's interstream divide boundary.

The sensitivity analysis for each of the six dimensionless parameters, which included changes in boundary conditions, was revealing. The graphs of h/l beneath the streambed and at the interstream divide for each parameter (figs. 13 and 14) are grouped into two broad categories—sensitive and insensitive; results are summarized in table 6. Insensitivity to a parameter is characterized by little or no change in h/l over the range of values tested, and insensitivity plots as a horizontal or nearly horizontal line in figures 13 and 14.

Correct interpretation of figures 13 and 14 requires an explanation of experimental bias. One value always lies on the observed h/l line in figure 13 (stream vicinity) because the choice of reference-model parameters was based on field measurements of head beneath the stream. The graphs for h/l near the interstream divide (fig. 14) do not indicate this bias, however. All values of h/l for five of the six dimensionless parameters are below the observed h/l values; the exception is the $Q_{in}/K_x l$ term, whose h/l values plot both above and below the observed h/l line (fig. 14*F*).

Simulation of the Shallow Flow System Associated with Connetquot Brook

After the sensitivity analysis had been completed, the model was calibrated to simulate the shallow flow

Table 6. Sensitivity of head (h/l) beneath the streambed and near the interstream divide to variations in dimensionless parameters

[Terms are explained on p. 20]

Dimensionless parameter	Percent varied	Percent change in h/l Beneath streambed	Near divide
(W_s/l)	100	[1]137	9
(m/l)	70	0	[1]51
(W/K_x)	40	[1]40	[1]37
(K_z/K_x)	990	2	6
(K'_z/K_x)	990	[1]1,013	71
$(Q_{in}/K_x l)$	(2)	[1]77	[1]92

[1]Heads were sensitive to this parameter.

[2]Q_{in} varied from 0 to 4.0×10^{-4} ft^2/s; percent change is not applicable.

system near Connetquot Brook. Data used for calibration included the water-table profile along cross section A–A' from the water-table map (fig. 6), a field-measured potentiometric profile beneath the stream (fig. 10), and results of the sensitivity analyses. The September 1977 water levels were chosen because long-term ground-water level records with which to calibrate the steady-state cross-section model were unavailable. The ground-water gradients, direction, and water-level altitudes during that period were similar to those expected under average conditions.

When the initial values for the model variables were chosen, streambed vertical hydraulic conductivity (K'_z) was adjusted to obtain a distribution of head values beneath the stream that were similar in gradient and magnitude to field data obtained earlier in the study. After a close match was achieved, the results were compared to the water-table profile drawn from the September 1977 water-table map (fig. 6). The simulated water-table altitudes became increasingly too low with distance from the stream. The sensitivity graphs (figs. 13 and 14) and data in table 6 indicated that $Q_{in}/K_x l$ was the only dimensionless parameter that could be changed within acceptable limits that would increase the simulated h/l values near the divide enough to equal the observed values. All other parameters either had little effect or would have to be changed to unrealistic values.

The next step was to estimate a value of Q_{in} that would yield an accurate result. Q_{in} near the divide was estimated directly from the sensitivity graph (fig. 14*F*) by fitting a line through the three simulated values of h/l. The value of $Q_{in}/K_x l$ on the x-axis where the fitted line crossed the observed h/l line was used with an assumed value for horizontal conductivity (K_x) to estimate an initial Q_{in}. Subsequently, Q_{in} was adjusted further by trial and error until an acceptable h/l near the divide was obtained. However, the acceptable h/l value near the divide gave an incorrect h/l beneath the stream; thus, further calibration was necessary. Table 6 shows that h/l is sensitive to both W_s/l and K'_z/K_x beneath the streambed but not near the divide. Because the selected stream width was already close to the measured field value, streambed vertical hydraulic conductivity, K'_z, was adjusted through trial and error until the simulated h/l values beneath the stream were acceptably close to observed values and those near the divide remained essentially unchanged. When this was achieved, the model was considered calibrated.

The final value used for K'_z was 9.7 ft/d, which is very close to the estimated 11 to 15 ft/d range given in table 3; the final value for Q_{in} was 1.67×10^{-4} ft^2/s. All other terms and boundary conditions remained the same as those used in the reference model run for the sensitivity testing (see section "Range in Numerical Values of Parameters Studied"). The final calibrated value of Q_{in} is equivalent to areal recharge over a total distance of 9,000 ft from the stream center to the interstream divide, which is 3,000 ft further than the distance initially used in the model study. Locating the interstream divide is difficult and uncertain, and the model studies indicated that the initial estimate of the distance from the stream to the divide was underestimated. The 9,000-ft distance between the stream and the interstream divide obtained in model calibration is a reasonable approximation of field conditions at Connetquot Brook.

Comparison of the simulated water-table profile with the observed water-table profile for September 1977 (fig. 6) shows a fairly close match with only minor differences. Simulated ground-water seepage to the stream was also compared with seepage calculated from observed measurements at two continuous-record gaging stations and to a set of instantaneous measurements above and below the modeled cross section (table 3). The simulated value was within 50 percent of the seepage at the continuous gages and within 15 percent of the value computed from instantaneous measurements. These deviations can be attributed to (1) lack of symmetry in Connetquot Brook's drainage basin, (2) field data from stream locations that are outside the modeled cross section, and (3) deviations in natural recharge during field measurements from the long-term average value used in the model.

Conclusions from Model Studies

The principal conclusions from the model simulations may be summarized as follows:

1. The near-stream model response compared well with field data when the streambed discharge boundary was simulated as a uniform leaky bed.
2. The dimensionless parameters that represent stream width and streambed hydraulic conductivity strongly influenced heads near the stream but did not appreciably affect the system at greater distance.

3. The two dimensionless parameters, recharge (W/l) and the interstream-boundary flux (Q_{in}), that regulate the quantity of water entering the shallow flow system had an appreciable influence on heads both near the stream and at distance.
4. The dimensionless parameter that represents the thickness of the shallow flow system (m/l) had a marked effect at distance but a negligible effect near the stream. The initial estimate of thickness of the shallow flow system (about 30 ft) was derived from vertical head differences near the stream and results of tritium studies. The sensitivity-test results indicate that the head distribution immediately beneath the stream would be the same for a considerable range in thickness of the shallow flow system. Furthermore, in the natural flow system, heads in the shallow flow system are influenced by heads in the underlying regional system and vice versa. These effects were not considered in the model studies described here; thus, a precise estimate of the thickness of the shallow flow system is not possible from the information available. However, the model results indicate that a value of about 30 ft is reasonable; this value is also consistent with present knowledge about the shallow flow system.

SUMMARY AND CONCLUSIONS

Several factors that control the interaction between ground water and surface water on Long Island were investigated during 1975–82 at Connetquot Brook, an undisturbed stream in central Long Island. The first part of the two-part study involved detailed field measurements of ground-water levels and collection of ground-water samples for dissolved-oxygen and tritium analysis to determine the geometry of the shallow flow system. The second part consisted of sensitivity analyses of the factors that govern flow in the shallow ground-water system through a two-dimensional numerical flow model in cross section.

During the field study, water levels were measured beneath and within a few feet of the stream channel to determine head changes with depth and within a few thousand feet of the stream to provide data for construction of detailed water-table maps. Wells were driven directly into the streambed and alongside the stream at three sites along the 5-mi stream reach studied, and head measurements were made at successive depth intervals that ranged from about 1 to 5 ft. In general, the largest increases in head with depth were in the streambed wells, followed by the streambank wells. Water levels in streambed wells rose between 1 and 2 ft above stream level, and most of the change was within the first 5 ft of penetration into the streambed. Head increases were noted to depths of about 30 ft beneath the streambed. Water-level measurements in two wells about 50 ft from the streambank showed no head increase to a depth of 40 ft. Elsewhere in the drainage basin, comparison of water levels in wells screened near the water table with those in wells screened several tens of feet below showed no measurable differences; this similarity indicates essentially horizontal ground-water flow except within 50 ft of the stream.

A traverse of five head measurements across the stream was made to measure head differences between the stream surface and water levels in streambed wells screened 3 ft below the streambed. The average change in head was 0.63 ft, and the range was from 0.57 to 0.71 ft; however, no consistent spatial pattern for these measurements was observed.

The vertical hydraulic conductivity of the streambed material, as calculated from measured head losses, was estimated to be 0.02 to 0.36 times the vertical hydraulic conductivity of the sediments immediately below. Vertical hydraulic conductivity values between 11 and 15 ft/d were calculated from the vertical ground-water gradient beneath the streambed and from measured streamflow increase through the study area.

Relative age of the ground water was used to determine the depth of the shallow flow system. Sixteen samples of ground water and one sample of stream water were collected and analyzed for tritium concentration as an indicator of the age of the water. Concentrations ranged from 0.1 to 166 T.U. The boundary between the shallow and deeper ground-water flow systems is somewhere between 45 and 95 ft below the water table, on the basis of the relative age of the water. Water from wells screened 95 to 100 ft below the water table was significantly older than water from wells screened between 40 and 45 ft below the water table. Tritium concentrations in shallow wells near the stream showed no obvious pattern with depth or spatially; this lack of pattern was attributed to the convergence of flow lines and the mixing of water of several ages near the stream.

The field data provided specific information on ground-water head with depth just below and adjacent to the stream; this information indicates the depth of the shallow flow system, but the data at greater distances from the stream were less definitive. The shallow ground-water system is probably 20 to 30 ft thick beneath and adjacent to the stream and becomes somewhat thicker several thousand feet from the stream. Ground-water flow throughout the shallow system is predominantly horizontal except within a few feet of the stream, where flow lines bend upward toward the stream.

A two-dimensional cross-sectional flow model under steady-state conditions was used to investigate the effects of several hydraulic and geometric factors that control flow in the shallow ground-water system. The shallow flow system was isolated in the model studies on the assumption that interaction between the shallow and the deep regional ground-water flow systems is negligible. In this study, the shallow system was divided into two parts—one part beneath the streambed and within a few feet of the stream channel and the other part at a greater distance from the stream and extending to the interstream divide.

The factors that control flow in the shallow system associated with the stream were classified into three categories: (1) geometric parameters (stream width, streambed thickness, thickness of shallow flow system, and lateral extent of flow system), (2) permeability parameters (horizontal and vertical hydraulic conductivity of the shallow flow system and streambed hydraulic conductivity), and (3) water-source parameters (areal recharge and lateral inflow). These dimensional parameters were combined into a series of dimensionless parameters to simplify analysis of model results. The model studies consisted of a sensitivity analysis of the effect of some of these parameters on model response—specifically, on calculated values of head—in both the near-stream area and more distant parts of the shallow flow system. To simplify this investigation, all model tests were steady state; thus, storage and time-dependent parameters were not considered.

Results of the sensitivity analysis can be classified into three categories, depending on where model heads were sensitive: near the stream, beyond the stream, or both. Model heads in the near-stream area were sensitive to dimensionless parameters containing vertical hydraulic conductivity and width of the streambed; heads in the more distant area were sensitive to dimensionless parameters containing the modeled thickness of the shallow flow system; and heads in both the near-stream and the more distant area were sensitive to dimensionless parameters containing recharge, horizontal hydraulic conductivity, and the interstream-divide flux term, which represents the additional recharge that would occur between the modeled interstream divide and the true interstream divide. Heads in the model were insensitive to the ratio between the vertical and horizontal hydraulic conductivity.

The sensitivity analysis aided in the final calibration of a steady-state model of this system. The calibrated-model response compared well with heads measured in the field both near and away from the stream; the favorable results in the near-stream area indicated that simulating the streambed discharge boundary as a uniform leaky bed was valid.

The initial estimate of the thickness of the shallow flow system (about 30 ft) was derived from head measurements near the stream. The model-sensitivity results indicated that thickness had little effect on the head distribution immediately beneath the stream but had a marked effect on heads some distance from the stream. Furthermore, heads beyond the stream's immediate area in the natural flow system are influenced by heads in the regional system and vice versa. These effects were not considered in the model studies described here. Although the estimate of 30-ft thickness cannot be verified at present, the model results indicated that this estimate was reasonable.

The sensitivity analysis also indicated which factors in the shallow flow system associated with the stream exert the greatest influence on flow patterns and where this influence is greatest. Further investigation of another stream having a more regular channel configuration than Connetquot Brook would provide information about the effects of channel geometry. This modeling study assumed flow in two dimensions along a flow line and ignored the effects of the regional flow system. Future studies using a fully three-dimensional model would eliminate errors related to limiting flow to a perceived flow line and could include the effects of the underlying regional flow system. Transient-state flow modeling would include the effects of storage and temporal variations in recharge and in stream stage, which is the controlling head at the streambed-discharge boundary.

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Conversion Factors and Abbreviations

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

Multiply inch-pound unit	By	To obtain SI unit
inch (in)	2.54	centimeter (cm)
foot (ft)	.3048	meter (m)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
cubic foot per second (ft^2/s)	28.32	liter per second (L/s)
	.02832	cubic meter per second (m^3/s)
gallon per minute per foot [(gal/min)/ft]	.01923	liter per second per meter [(L/s)/m]
foot per day (ft/d)	.3048	meter per day (m/d)
ounce (oz)	.0338	milliliter (mL)

Other Abbreviations

mg/L = milligram per liter
T.U. = Tritium unit

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