Hydrology of Brookhaven National Laboratory and Vicinity Suffolk County, New York

GEOLOGICAL SURVEY BULLETIN 1156-C

This report concerns work done on behalf of the U.S. Atomic Energy Commission

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By M. A. WARREN, WALLACE DE LAGUNA and N. J. LUSCZYNSKI

STUDIES OF SITES FOR NUCLEAR ENERGY FACILITIES—BROOKHAVEN NATIONAL LABORATORY

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STUDIES OF SITES FOR NUCLEAR ENERGY FACILITIES—BROOKHAVEN NATIONAL LABORATORY

HYDROLOGY OF BROOKHAVEN NATIONAL LABORATORY AND VICINITY, SUFFOLK COUNTY, NEW YORK

By M. A. Warren, Wallace de Laguna, and N. J. Lusczynski

ABSTRACT

The Brookhaven Natonal Laboratory is in central Suffolk County, Long Island, New York. The area studied surrounds and includes the Laboratory and is referred to herein as the Upton area. It extends across the island in a band about 13 miles wide from the Atlantic Ocean to Long Island Sound between longitudes 72°45′ and 73°00′. Its climate is characterized by mild winters and relatively cool summers. Precipitation averages about 45 inches a year evenly distributed throughout the year. The soil and the immediately underlying sediments are generally sandy and highly permeable. Water penetrates them readily and except in periods of intense precipitation there is very little direct overland runoff to streams.

Permeable Pleistocene deposits, 100–200 feet thick, constitute the uppermost aquifer. It receives recharge from precipitation (the only source of fresh water on the island) and discharges mainly into streams, the ocean, and the sound and to a some lesser extent into lower aquifers. The lower aquifers, several hundred feet in total thickness, transmit water under artesian pressure from the high central part of the island toward its edges where it is discharged into streams or into bodies of salt water. Streamflow is supported throughout the year very largely by ground-water discharge.

Within this broad pattern the details of the movement and behavior of water are determined by the geology, the topography, and the seasonal and local distribution of precipitation. Tests at the Laboratory site indicated that under favorable conditions water may move from the land surface to the water table at a rate of about 30 feet per day. Under less favorable conditions it may move 1 foot a day or less.

The topography of the water table conforms only generally to that of the land surface. Ground-water divides between the small streams in the area differ significantly from topographic divides and explain apparent differences in the rates of discharge per square mile. At the Laboratory site most of the ground-water movement is southward toward the Atlantic Ocean, but part of it is eastward to Peconic Bay. Ground-water movement in a part of the Laboratory area is either to the south or to the east, depending upon the stage of the water table, and is controlled by the presence of relatively impermeable beds near the surface.

A detailed pumping test, using one of the Laboratory supply wells, showed that the 45-foot saturated thickness of the upper Pleistocene aquifer has a transmissibility of about 190,000 gallons per day per foot, a field coefficient of permeability of 1,300 gallons per day per square foot, and a specific yield of 0.24. Flow-net analysis of the water-table contours and measurements of the streamflow of the middle reach of the Carmans River showed that in this area, about 4 miles southwest of the Laboratory, the transmissibility of the 170-foot thick saturated upper Pleistocene sand is 240,000 gallons per day per foot and the field coefficient of permeability is about 1,400 gallons per day per square foot. From these data, several detailed water-table maps, and a knowledge of the geology of the area, a map was constructed showing the direction and rate of movement of the ground water in the upper Pleistocene aquifer at times of low and high water table. The average rate of horizontal movement of the water is about 0.5 foot per day.

Three streams drain the Laboratory site; the Carmans and Forge Rivers flow southward to the Atlantic Ocean and the Peconic River flows eastward to Peconic Bay. A system of ditches constructed during World War II when the Laboratory area was Camp Upton may tend to extend the drainage of the Peconic at the expense of the Carmans River under some circumstances.

LOCATION AND EXTENT OF THE AREA

The Brookhaven National Laboratory is in central Suffolk County in eastern Long Island. The area surrounding the Laboratory is sometimes referred to as the Upton area after Camp Upton which occupied the site during World War II. The area primarily considered in this report and occasionally referred to herein as the Upton area is a north-south strip across the island about 13 miles wide between longitude 72°45′ and 73°00′ and includes all land and water between Long Island Sound on the north and the Atlantic Ocean on the south. The Brookhaven National Laboratory occupies about 5¾ square miles in the central part of this area.

PURPOSE AND SCOPE

The purpose of this report is to describe the hydrology of the area. Accordingly, the report describes the climate of the area with emphasis on precipitation, the source of all fresh water in the area. It estimates the proportion of the precipitation that is returned to the atmosphere by evapotranspiration and describes the recharge, movement, and discharge of the ground water in the area. Finally it describes the source (largely ground-water discharge) and movement of water in the stream systems of the area under varying conditions of precipitation and ground-water levels. All these descriptions are as quantitative as possible after several years of intensive investigation. Together they provide not only an understanding of the behavior of water in the area but also a basis for predicting its movement in the future under any given set of antecedent circumstances—not always in detail but with considerable confidence as to the general pattern.

CLIMATE

GENERAL FEATURES

The climate of Long Island is influenced more by the ocean than by the adjacent mainland. It is characterized by mild winters and relatively cool summers and is comparatively free from sudden or extreme changes of temperature. Information about the climate of central Long Island may be obtained from several sources, including: "The Climate of New York State," by R. A. Mordoff (1949); U.S. Weather Bureau, climatological data; and U.S. Department of Agriculture 1941 Yearbook; "Climate and Man," by G. R. Williams. These publications show that the average annual temperature is about 51°F, the average January temperature about 30°F, and the average July temperature about 70°F. The average maximum annual temperature is 95°F, and the average minimum annual temperature is 0°F. The maximum and minimum observed temperatures are 102°F and −20°F. The growing season on Long Island is about 180–200 days, from the last of April to the last of October. During an average year, the percentage of possible sunshine ranges from about 50 percent in January to about 65 percent in July and averages about 62 percent during the growing season.

The winds on Long Island are predominantly westerly, shifting from southwest in summer to northwest in winter. During 3 years of record at the Brookhaven National Laboratory meteorology observation tower, the wind direction at a level 37 feet above the ground was south-southwest to north-northwest about 50 percent of the time. The wind had a velocity between 0 and 5 mph (miles per hour) 42 percent of the time, between 5 to 12 mph 46 percent of the time, and 12 to 24 mph 12 percent of the time. Wind velocities of more than 25 mph occurred less than 1 percent of the time. The percentage of the time that the wind velocities were in the above categories did not vary appreciably with the season. The average wind velocity in New York City is about 11½ miles per hour and is presumably about the same in the Upton area.

Long Island is infrequently visited by tropical hurricanes which have wind velocities exceeding 75 mph. The 1938 hurricane, one of the most severe in the 20th century, caused storm waves 3–16 feet above mean high tide along the south shore. In the Great South Bay area the waves ranged from 3 to 7 feet above mean high tide, but in the Moriches Bay area they were as high as 16 feet above mean high tide. Along the north shore of the island, the height of the storm waves ranged from

5 to 10 feet above mean high tide. Long Island was also visited by two hurricanes in 1954. Unconfined ground water in low-lying areas near the shore is salted by sea water blown inland during hurricanes.

The maximum depth of freezing in the soil zone is 15 inches; the average is much less. Because the soil is not frozen during most of the winter season, recharge to the water table is possible during the winter, and because evapotranspiration is low, most of the ground-water recharge does, in fact, take place during the colder months, from December to May.

PRECIPITATION

Precipitation, the only source of fresh water for the streams and ground water in the Upton area, is used here as the starting point of the hydrologic cycle. The average precipitation ranges from about 42 inches in the western part to about 46 inches in the eastern part of Long Island. In an average year, about 120 days have 0.01 inch or more of precipitation. Long Island is supplied with moisture from the Gulf of Mexico and from the Atlantic Ocean through the action of winds of cyclonic storms. The general current of the prevailing westerlies plays only a small part in producing precipitation in Long Island. Natural variations in precipitation are largely due to physiographic and storm-pattern factors.

The Upton area of Long Island has little relief and thus monthly, and especially yearly, precipitation does not differ much from one locality to another within the area. Such differences as do occur are due largely to local summer storms or to differences in the local details of the rain gage or its exposure. But, though geographic variations are not large, a careful study of cummulative records shows some variation in rainfall within the Upton area.

RECORDS AVAILABLE

Precipitation records for eight stations within a 13-mile radius of the center of the Brookhaven National Laboratory are used in this report. Three of these stations are on the Laboratory grounds; no two stations are more than 20 miles apart (fig. 1). The length of record at the end of 1953 ranges from 5 complete years (at two gages within the Laboratory area) to nearly 69 complete years at Setauket (tables 1 and 2). The earliest records are for 1864–82 at the village of Brookhaven. The record at Setauket began in 1885.

The rainfall records and the values for average, minimum, and maximum precipitation proved satisfactory for correlating precipitation with surface-water stages and flows and with ground-water levels. Precipitation data for periods of less than a month are discussed briefly, because they have some bearing on the problems of ground-water contamination (de Laguna, 1966).

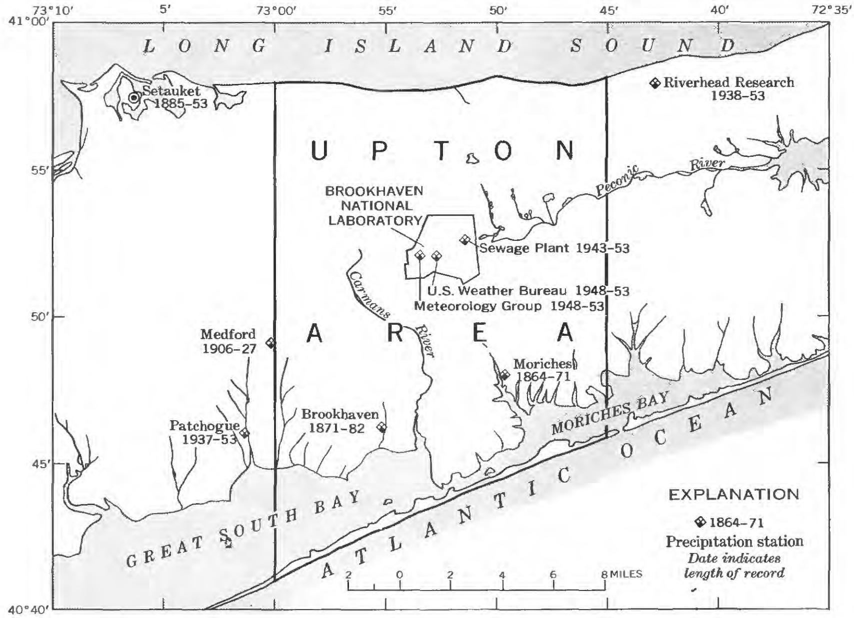

Figure 1.—Location of study area and precipitation stations.

The precipitation data for the 1864–71 period, listed for the village of Brookhaven, were actually collected at Moriches about 5 miles to the east. From 1871 to 1882 the data were collected at the village of Brookhaven, about 7 miles south of the present Laboratory area. This record, started under the sponsorship of the Smithsonian Institute (tables 1 and 2) before the establishment of the U.S. Weather Bureau, show that the average annual precipitation from 1864 to 1882 was 46.20 inches. This precipitation record includes the maximum and minimum yearly rainfalls for the Upton area, a high of 71.38 inches in 1869 (a year of a hurricane) and a low of 27.65 inches in 1881. The 2-year average for 1868–69 was 65.51 inches; the 3-year average for 1867–69 was 62.05 inches; and the 5-year average from 1865–69 was 59.61. These are all records and are considerably in excess of any recent data.

These data, especially those for 1865–69, are accepted with some reservation because they are much greater than those recorded at other stations along the northeastern seaboard. For example, precipitation in the city of New York, about 57 miles to the west, averaged 48.45 inches during this period, or about 11.16 inches less than that at Brookhaven. The present-day average at New York City is only 2–4 inches less than that for the Brookhaven area. Furthermore, the average precipitation reported for 1865–69 at Brookhaven was 0.35 inch higher

TABLE 1.—*Average maximum and minimum precipitation, in inches, for the eight stations, in central Suffolk County, Long Island, N.Y.*

Station [1]	Lat	Long	Altitude (feet)	Period of record	Complete years of record	January	February	March	April	May	June	July	August	September	October	November	December	
						Average monthly												Average yearly
1. Brookhaven	40°48′	72°39′	13	[2] 1864–82	19 (1864–82)	3.87	4.04	5.03	3.84	3.95	2.90	3.66	3.46	3.36	3.67	4.22	4.20	46.20
2. Brookhaven National Laboratory, Meteorology Group	40°52′	72°53′	[3] 80	1948–53	5 (1949–53)	5.07	4.34	5.54	3.73	4.64	2.03	2.42	3.96	1.56	2.97	4.64	5.35	46.25
3. Brookhaven National Laboratory, sewage plant	40°53′	72°51′	[3] 50	1943–53	11 (1945–53)	4.19	3.15	4.12	3.76	4.31	2.56	2.75	4.12	2.15	2.84	5.09	4.17	43.21
4. Brookhaven National Laboratory, U.S. Weather Bureau	40°52′	72°53′	75	1948–53	5 (1949–53)	5.28	4.45	5.16	3.76	4.50	1.92	2.33	4.46	1.48	2.65	4.77	4.59	45.35
5. Medford	40°45′	73°00′	89	1906–27	22 (1907–26) (1938–46)	4.13	3.52	3.89	4.12	3.77	3.45	3.93	4.44	2.92	3.69	3.59	4.48	45.94
6. Patchogue	40°46′	73°01′	25	1937–53	15 (1948–53)	3.92	3.23	4.34	3.32	3.50	3.63	3.46	4.48	2.83	2.90	4.62	3.81	44.04
7. Riverhead Research	40°56′	72°43′	100	1938–53	15 (1939–53)	3.95	3.21	4.23	3.51	3.66	3.02	3.14	4.18	2.58	3.06	4.75	3.82	43.11
8. Setauket	40°57′	73°06′	40	1885–1953	68 (1886–1953)	3.97	3.61	4.15	3.72	3.57	3.20	3.85	4.14	3.46	3.62	3.86	3.82	44.97
						Weighted monthly average [4]												Weighted yearly average
						3.99	3.54	4.25	3.74	3.70	3.18	3.71	4.14	3.12	3.44	4.13	3.99	4.93
						Maximum monthly												Maximum yearly
				1864–1882 1885–1953		11.14 (5) 1915	7.88 (1) 1869	10.36 (2) 1953	9.34 (1) 1868	8.91 (1) 1868	10.10 (8) 1943	18.18 (8) 1897	13.05 (8) 1927	13.16 (8) 1938	12.86 (1) 1869	8.96 (5) 1911	9.43 (8) 1900	71.38 1 1869
						Minimum monthly												Minimum yearly
				1864–1882 1885–1953		1.16 (1) 1876	0.69 (8) 1901	0.41 (8) 1915	1.02 (8) 1896	0.10 (8) 1887	0.00 (4) 1949	0.34 (5) 1924	0.41 (1) 1881	0.23 (8) 1914	0.23 (8) 1924	0.43 (8) 1917	0.31 (8) 1877	22.65 1 1881

[1] Referred to by number (in parentheses) in data for maximum and minimum monthlies.
[2] Record for 1864–71 at Moriches.
[3] Estimated.
[4] For stations 1, 3, 5, 6, 7, and 8 combined; for weighted average records at 2 and 4 combined with 3 for single record see table 2.

TABLE 2.—*Comparison of average yearly precipitation for different periods of record at six stations and comparison of 1-, 2-, 3-, 5-, and 10-year average maximum and minimum records, in inches per year*

Station	Average					Average maximum for number of years indicated					Average minimum for number of years indicated				
	1907-26	1939-46	1939-46 1948 1950-52	1939-46 1948-53	1943-46 1948-53	1	2	3	5	10	1	2	3	5	10
Brookhaven						71.38 1869	65.51 1868-69	62.05 1867-69	59.61 1865-69	52.15 1864-73	27.65 1881	31.33 1880-81	34.11 1879-81	36.83 1877-81	40.17 1873-82
Brookhaven National Laboratory, sewage plant					43.16	51.40 1953	48.61 1952-53	47.99 1951-53	44.57 1949-53	44.18 1944-53	33.46 1943	37.85 1943-44	38.89 1943-45	40.39 1943-47	42.39 1943-52
Medford	45.95					56.37 1919	55.88 1919-20	52.55 1919-21	51.12 1919-23	48.00 1911-20	37.87 1908	40.83 1924-25	41.01 1924-26	43.92 1914-18	44.77 1908-17
Patchogue		44.05	44.68	44.06	43.85	54.94 1948	49.97 1947-48	47.55 1948-48	47.00 1944-48	45.21 1939-48	36.02 1949	37.04 1949-50	39.37 1949-51	41.95 1949-53	43.28 1943-52
Riverhead Research		41.01	42.90	42.41	42.37	50.83 1953	47.52 1952-53	48.18 1951-53	43.77 1949-53	43.31 1944-53	32.27 1941	37.15 1949-50	38.33 1941-43	39.53 1941-45	40.35 1941-50
Setauket	42.67	42.07	44.03	43.37	43.55	59.26 1897	58.63 1897-98	53.11 1898-90	51.39 1886-90	48.97 1897-1906	33.65 1910	37.27 1916-17	37.50 1908-10	40.07 1908-12	40.05 1908-17

than any annual precipitation since 1885 at Setauket, which is hard to believe even though the periods compared are not concurrent. The highest annual precipitation recorded at any other gage in the area is 59.26 inches at Setauket in 1897, the lowest 32.47 inches at Riverhead Research in 1941. These records are probably somewhat more representative of the extremes to be found in the Upton area.

The Setauket station, near the north shore about 13 miles west-northwest of Brookhaven National Laboratory, has the longest record of any station on Long Island (1885–1953). The annual precipitation for the period of record through 1953 averaged 44.97 inches; the highest was 59.26 inches in 1897, and the lowest was 33.65 inches in 1910. The two highest observed 5-year average annual precipitations are 51.39 inches (1886–90) and 51.24 inches (1897–1901).

Of the five gages outside the Laboratory area, the nearest to the Laboratory was the former gage at Medford, about 7 miles to the west-southwest. Because Medford is inland in Suffolk County, this gage had a setting somewhat physiographically similar to that of the Laboratory area. The average annual precipitation for the 1907–26 period was 45.95 inches; during the same period the average annual precipitation at Setauket, 10½ miles to the north-northwest, was less by 3.28 inches. During the 20 complete years of record common to both stations, the annual precipitation at Medford exceeded that at Setauket by amounts ranging from 0.11 inch to 8.30 inches; it was greater at the Setauket gage only in 1920 and then only by 0.39 inch. The persistently higher precipitation at Medford was probably due to the exposure of the gages rather than to a true difference in rainfall.

The Patchogue station, 10½ miles southwest of the Laboratory, has continuous records from 1937 to 1953, except for 4 months in 1947, when the site was changed. The average annual precipitation is 44.04 inches. The Riverhead Research Station, at the Long Island Vegetable Research Farm, 10 miles northeast of Brookhaven National Laboratory, had an average annual precipitation for the 15-year period (1939–53) of 43.11 inches.

The Patchogue and Riverhead gages were both in operation for the 8-year period (1939–46). The average annual precipitation for Patchogue was 44.05 inches, and for Riverhead 41.01 inches. For the 13 years of record, 1939–46, 1948, and 1950–53, common to the gages at Setauket, Patchogue, and the Riverhead Research Station, the annual precipitation averaged 44.03, 44.68, and 42.90 inches, respectively.

Records are available from three gages within the boundaries of Brookhaven National Laboratory (table 3). The Meteorology Group

TABLE 3.—*Monthly, yearly, and mean yearly precipitation, in inches, for the three gages at the Brookhaven National Laboratory*

Year	Station	January	February	March	April	May	June	July	August	September	October	November	December	Annual
1943	Sewage plant	2.24	2.37	3.35	2.76	2.42	3.89	3.24	1.25	2.24	4.85	2.93	1.92	33.46
1944	do	4.23	1.72	5.67	4.76	1.30	1.78	.65	1.53	6.98	2.20	8.43	2.99	42.24
1945	do	1.98	3.82	2.03	3.68	5.16	1.43	3.79	2.27	1.54	3.38	6.00	5.90	40.98
1946	do	3.65	2.34	2.23	2.09	4.48	3.94	4.35	11.95	1.88	.32	1.13	3.19	41.55
1947	do	3.47	.39	2.20	6.21	3.97	3.15	2.85	3.77	1.70	3.62	9.23	3.18	43.74
1948	Sewage plant	5.60	2.23	4.52	4.09	6.97	4.35	3.61	1.42	1.26	4.60	5.70	6.13	50.48
	Meteorology Group								1.23	1.25	5.42	5.92	7.05	
	U.S. Weather Bureau									1.28	4.49	5.21		
	Mean								1.32	1.26	4.84	5.61	6.59	50.99
1949	Sewage plant	5.44	4.95	2.46	3.61	3.56	.05	2.80	5.03	3.02	1.69	3.24	2.99	38.84
	Meteorology Group	5.55	4.71	2.88	3.63	3.32	.01	3.07	5.21	3.49	1.74	2.96	3.66	40.23
	U.S. Weather Bureau	5.70	4.39	2.86	3.57	3.21	.00	2.78	5.08	2.93	1.56	3.09	2.82	37.99
	Mean	5.56	4.68	2.73	3.60	3.36	.02	2.88	5.11	3.15	1.66	3.10	3.16	39.02
1950	Sewage plant	2.53	4.38	3.61	2.78	5.26	2.57	3.51	4.10	1.31	2.04	4.04	3.91	40.04
	Meteorology Group	2.80	4.28	3.98	2.41	5.23	2.72	3.22	4.26	1.31	1.69	4.34	4.36	40.60
	U.S. Weather Bureau	2.90	4.83	3.22	2.89	5.33	2.82	3.44	4.46	1.35	1.57	5.53	4.33	42.67
	Mean	2.74	4.50	3.60	2.69	5.27	2.70	3.39	4.27	1.32	1.77	4.64	4.20	41.10
1951	Sewage plant	4.29	4.65	4.49	2.33	3.96	2.33	2.11	3.67	1.36	4.40	7.26	5.90	46.75
	Meteorology Group	3.75	4.99	5.02	3.42	3.68	2.64	2.08	3.51	1.07	5.48	6.01	6.17	47.82
	U.S. Weather Bureau	5.23	5.48	5.01	3.50	3.60	2.06	2.04	3.26	1.35	4.45	6.90	6.02	48.90
	Mean	4.42	5.04	4.83	3.08	3.75	2.34	2.08	3.48	1.26	4.78	6.72	6.03	47.82
1952	Sewage plant	6.21	3.66	4.81	3.76	6.73	2.66	.71	7.07	1.54	.70	3.05	4.01	45.81
	Meteorology Group	6.54	3.54	5.44	3.61	7.62	2.78	1.00	7.15	1.35	.31	3.56	4.45	47.35
	U.S. Weather Bureau	6.20	3.50	4.95	3.87	6.85	2.89	.77	7.09	1.17	.59	3.02	3.88	44.78
	Mean	6.32	3.57	5.07	3.75	7.07	2.78	.83	7.40	1.35	.53	3.21	4.11	45.98
1953	Sewage plant	6.48	4.09	9.92	5.31	3.65	1.94	2.60	2.35	.80	3.41	4.97	5.88	51.40
	Meteorology Group	6.73	4.16	10.36	5.59	3.34	1.98	2.76	2.40	.90	3.17	5.03	6.43	52.85
	U.S. Weather Bureau	6.39	4.07	9.77	4.95	3.51	1.85	2.62	2.42	.80	3.22	4.87	5.92	50.39
	Mean	6.53	4.11	10.02	5.28	3.50	1.92	2.66	2.39	.83	3.27	4.96	6.08	51.55

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gage is near building T-51 on Brookhaven Avenue, about 1,350 feet west of Upton Road. The U.S. Weather Bureau gage was about 350 feet north of Brookhaven Avenue and about 140 feet east of Rochester Avenue, 3,800 feet east of the Meteorology Group gage. The sewage plant gage is near the pumphouse, about 6,400 feet northeast of the U.S. Weather Bureau gage and about 9,700 feet east-northeast of the Meteorology Group gage. Records of the Meteorology Group gage start in August 1948. Data for the U.S. Weather Bureau gage, which start in September 1948, were supplied by the local office of the Bureau at the Brookhaven National Laboratory. The sewage plant gage records, which start in January 1943 when the Laboratory area was still called Camp Upton, were furnished through the courtesy of E. W. Jeffries, sewage plant operator, who was also the gage observer.

COMPARISON OF RECORDS

Monthly and yearly precipitation for the three gages within the boundaries of the Laboratory are compared in figure 2 and table 3. Although the gages are less than 2 miles part, in many months the differences in precipitation are as much as 1–1½ inches and nearly 3 inches in certain years.

As shown in figure 3, a comparison of monthly precipitation at the Riverhead Research Station, Patchogue, and the sewage plant shows plus and minus differences for some month of more than 2 inches. In general, precipitation at the Laboratory was intermediate between that at the Riverhead Research Station and that at Patchogue, but monthly records at the sewage plant most closely resembled those at Riverhead Research.

The average annual precipitation for the Patchogue and Riverhead Research gages for the 14 complete calendar years common to both gages (1939–46, 1948–53) is 44.06 and 42.41 inches, respectively. For the 10 complete calendar years of record (1943–46, 1948–53) common to the three gages at Patchogue, Laboratory sewage plant, and Riverhead Research, the annual precipitation averaged 43.85, 43.16, 42.37 inches, respectively.

The average monthly precipitation for each of the eight stations in table 2 differ appreciably, in part because the averages are for different periods and are also for different lengths of records. The longer records produce better monthly averages because differences in the average monthly precipitation at a gage become smaller as the length of record increases. For example, the average monthly precipitation at the Weather Bureau gage at the Laboratory, with a little more than 5 years of record, ranges from 5.28 inches in January to 1.48 inches in September, or by 3.80 inches, whereas at Setauket, with more than 68 years of record, the average monthly precipitation

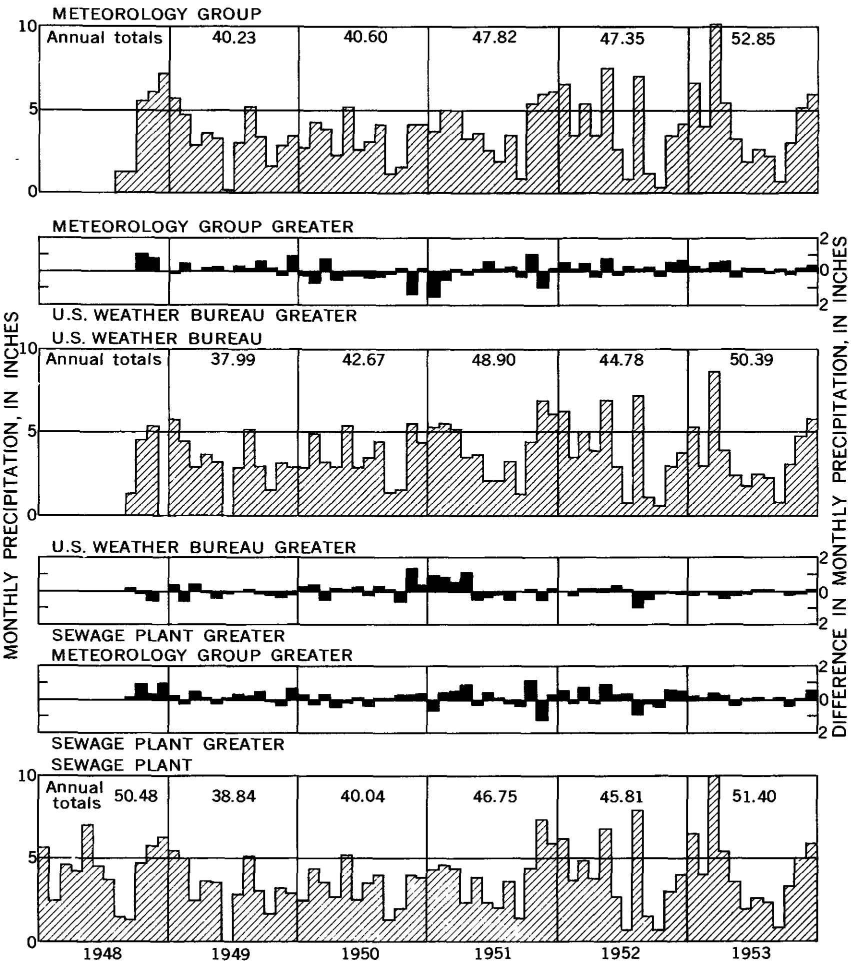

FIGURE 2.—Comparison of monthly and yearly precipitation at sites within the Brookhaven National Laboratory.

varies from 4.15 inches in March to 3.20 inches in June, or by 0.95 inch.

The pattern of monthly precipitation found from long-term records is rarely duplicated in any one year. An analysis of data for Brookhaven (1864–81), Medford (1906–27), Patchogue (1938–51), Riverhead Research (1939–51), and Setauket (1886–1951) shows that in most years there are about 2 months with precipitation of less than 2 inches and 1–2 months with precipitation exceeding 6 inches.

MONTHLY AVERAGE PRECIPITATION

The average monthly precipitation at the three gages within the Laboratory area for 1948–53 is shown in table 3. The weighted average monthly precipitation for the 150 years of combined record at the six

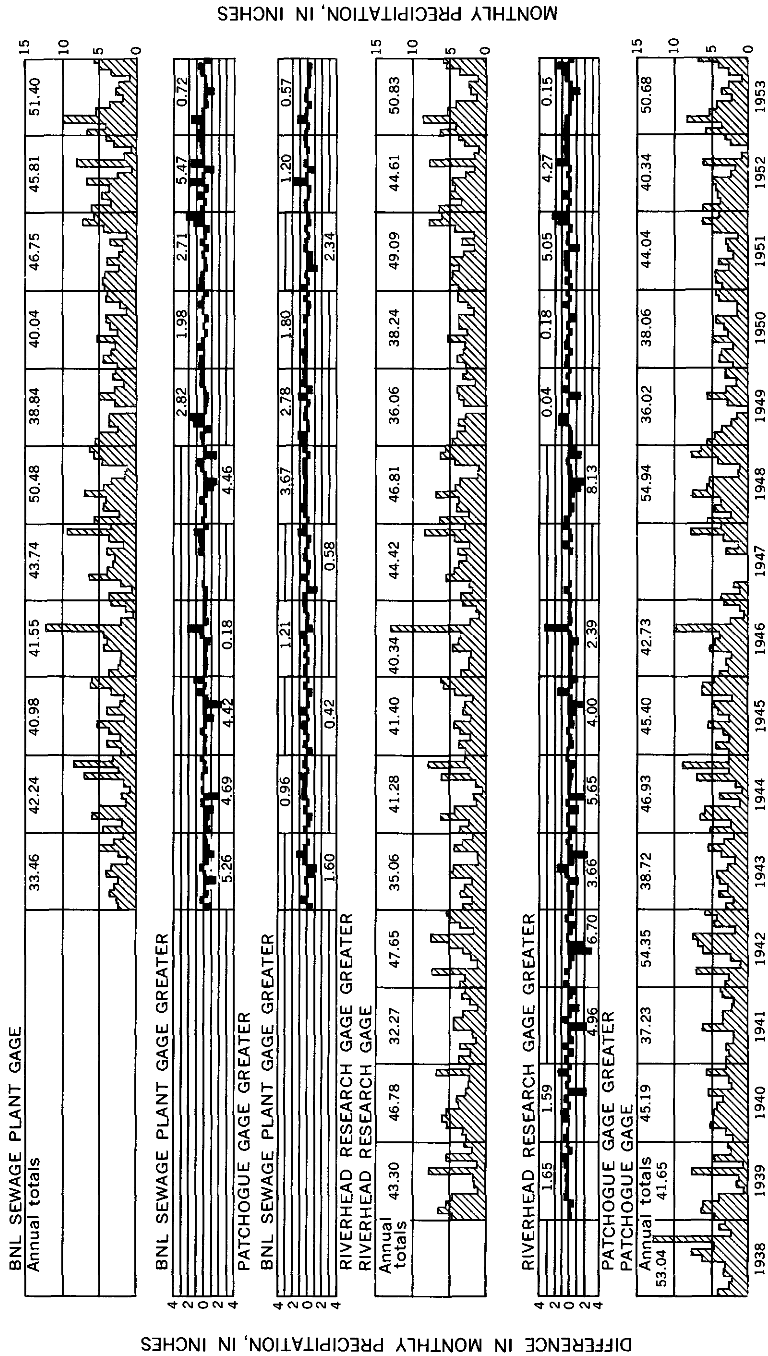

FIGURE 3.—Comparison of monthly and yearly precipitation at the sewage plant of the Brookhaven National Laboratory, at Patchogue, and at the Riverhead Research Station.

stations—(Brookhaven, Brookhaven National Laboratory (composite), Medford, Patchogue, Riverhead Research, and Setauket—is listed in table 1. The weight assigned was directly proportional to the length of record; for example, Setauket with 68 complete years of record had 68/11 times as much weight as the composite Laboratory gage with only 11 complete years of record. The weighted monthly average for the six stations varies from 4.25 inches in March (with 4.13 inches for November only slightly less) to 3.12 inches in September (with 3.18 inches for June only slightly more). The average monthly weighted precipitation is 3.74 inches.

DISTRIBUTION OF RAINFALL

In the 150-year composite record, about 56 percent of the months, with precipitation less than the monthly average of 3.74 inches, received about 35 percent of the total precipitation, whereas 44 percent of the months, with precipitation greater than the monthly average, received about 65 percent of the total precipitation. The most frequent monthly rates are from 2 to 4 inches in a month, which account for 34 percent of the precipitation and 43 percent of the time, and rates between 3 to 5 inches, which account for 39 percent of the precipitation and 37 percent of the time. Only during 18 of the 1,800 individual months did precipitation exceed 9 inches. The highest monthly rainfall was 18.18 inches at Setauket in July 1897; the lowest was zero at the U.S. Weather Bureau gage in June 1949.

During an average year in the city of New York, the total precipitation of 42 inches falls in amounts of 0.01 inch or more on about 120 days; the average is 0.35 inch for each day of measurable precipitation. The rainfall in the Upton area is similarly distributed. The maximum rate of precipitation in the Upton area is about 3 inches per hour, 10 inches per day, 18 inches per month, and 70 inches per year; these are unusual occurrences.

The difference in average yearly precipitation at the eight stations in table 1 is a little more than 3 inches, from 46.25 inches for the Meteorology Group gage at the Laboratory (5 complete years of record), to 43.11 inches at the Riverhead Research gage (15 complete years of record); the average annual precipitation for the Setauket gage (nearly 69 years of record) is 44.97 inches. Agreement is closer, in general, when precipitation for the same periods is compared (table 2).

Because the mean weighted average annual precipitation in table 1 is 44.93 inches, the long-term average precipitation in the Upton area and the Laboratory area may be taken as 45 inches. This figure includes an average snowfall of about 30 inches per year, or an equivalent of about 3 inches of rain. A snow cover of 1 inch or more is on the ground for about 40 days during an average year.

Because the weighted average monthly and yearly precipitation for the three Laboratory gages corresponds closely to that at Setauket, earlier precipitation data at this station may be representative of the Upton area, even though the gage is outside the area.

The preceding discussion dealt with variations with time and with locality and with the frequency of monthly and yearly rates. For periods less than a month and for individual storms, the percentage variation in the precipitation rate at individual stations and the percentage differences in the catch between neighboring stations are greater, but these short-term differences do not produce significant differences at the water table, where the water table is deep.

The precipitation records used in this report to show the correlation between precipitation, surface-water flow, and the height of the water table, are the mean of the Patchogue and Riverhead gages from October 1941 to December 1942, the sewage plant gage from January 1943 to December 1948, and the mean of the three Laboratory gages from December 1948 to September 1953. The average annual rainfall determined from this record is 43.64 inches, about an inch less than the indicated long-time average.

GROUND WATER

GENERAL DEFINITIONS

According to Meinzer (1923), the zone between the land surface and the water table is the zone of aeration, and water in this zone is vadose water, or suspended water. The zone of aeration is divided into three parts: (1) A belt of soil in which are found the roots of plants; water in this belt was referred to by Meinzer as soil water. The term "soil zone" is used more commonly now for this belt, and the contained water is referred to as soil moisture. (2) The intermediate belt lies between the soil zone and the capillary fringe of the water table; water in this zone was referred to by Meinzer as intermediate vadose water, or vadose water in the intermediate zone. (3) The capillary fringe is a moist belt just above the water table; its water content in general increases downward as progressively larger interstices are filled with water.

The water table is defined as that surface where pressure in an unconfined aquifer is equal to the atmospheric pressure; it is mapped as the surface represented by the static water level in wells ending in an unconfined aquifer. The height above the water table to which the capillary interstices are filled with water is largely a function of the size of the interstices, the temperature of the water, and the rate of rise or fall of the water table.

Ground water in the Upton area moves in all the unconsolidated formations from land surface to bedrock. The residual water from precipitation (after evapotranspiration and overland runoff) percolates first downward through the zone of aeration to the water table. It then moves vertically through and horizontally along layers of gravel, sand, silt, and clay in the zone of saturation. Most of the ground water advances to discharge points under water-table conditions; that is, the pressure at any point in the system is equal to the head of water vertically above that point. The principal water-table aquifer (water-bearing formation) in the area is 100–200 feet of upper Pleistocene deposits that rest either upon the Gardiners Clay or upon the Magothy(?) Formation.

Some ground water percolates downward from the upper Pleistocene deposits and moves under artesian conditions through several hundred feet of deeper unconsolidated formations. Artesian conditions require that pressure be transmitted in part laterally through the system because of more or less horizontal confining layers. Eventually, and for the most part at some distance from the Upton area, this water rises again toward the land surface and is discharged into streams on Long Island or into bodies of salt water surrounding the Island.

Ground water moves north, south, or east in all the unconsolidated formations in the Upton area along more or less definite paths or flow lines. In general, the movement is away from areas west of the central Upton area, where the water table is the highest, to the most readily accessible discharge points, such as the Carmans, Forge, Peconic, and other nearby streams or more distant streams or salt-water bodies. The pressure of a higher water table west of the Upton area generally inhibits movement toward the west.

The general pattern of flow lines in a north-south section through the center of the Upton area is shown in figure 4. The flow-line diagram was constructed from: (1) the water-table contours, (2) a few water-level measurements in the Magothy(?) Formation and in the Lloyd Sand member of the Raritan Formation, (3) the relations, where known, of salt water to fresh water in the formations, and (4) the depth, thickness, and extent of the several formations. The positions of the salt-water front on the north and south shore are only approximate; in the artesian formations on the south shore it may well be a mile or two further seaward than is shown in figure 4. By far the largest quantity of water moves through the Pleistocene deposits.

RECHARGE TO GROUND WATER—INFILTRATION BELOW SOIL ZONE

APPLICABLE HYDROLOGIC EQUATION

The only source of natural fresh water on Long Island is precipitation (rain, snow, and sleet) on the land area of the island. Replenishment, as recharge to the ground water, is the residual of the precipitation after losses due to overland runoff and to evapotranspiration.

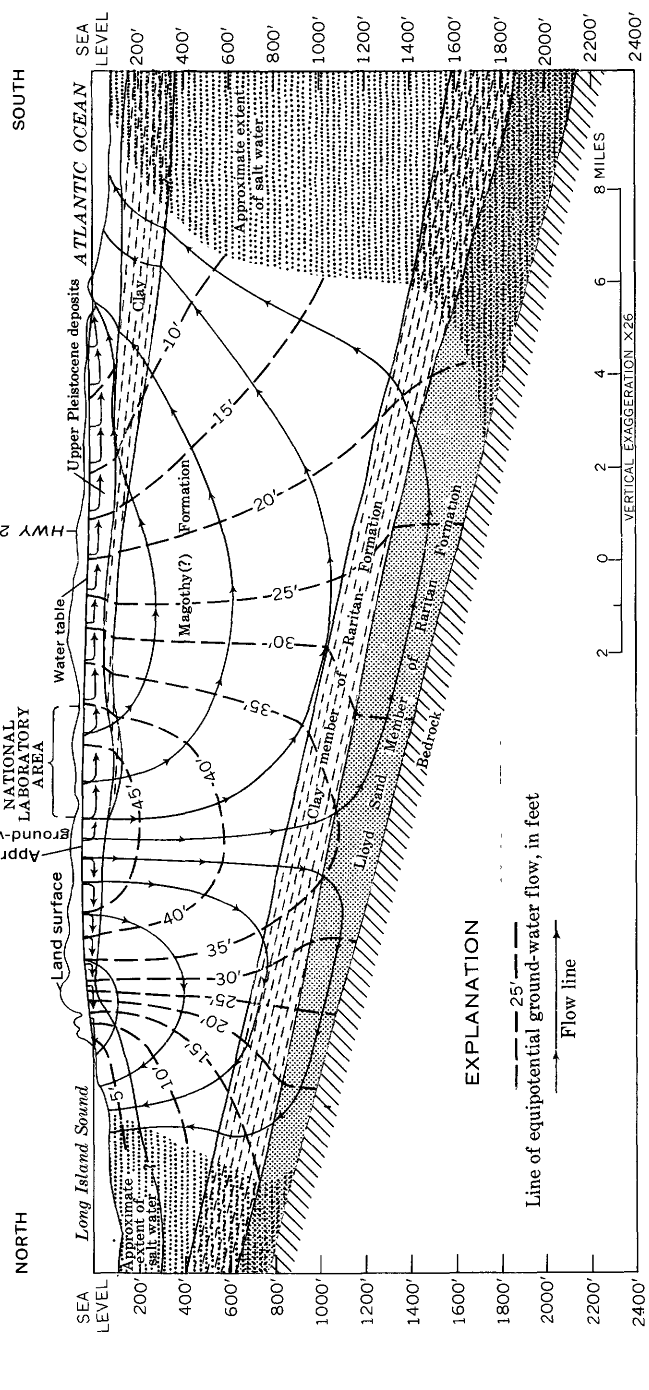

FIGURE 4.—Schematic ground-water flow lines, central Upton area.

The hydrologic equation applicable to the Upton area in Long Island, for any selected water-budgeting period, is as follows: Recharge to ground water (infiltration below soil zone and below reach of plants) plus overland runoff equals precipitation minus evapotranspiration.

The equation, the precipitation data presented above, and values of the overland runoff and evapotranspiration presented in the following pages are used below to compute monthly and annual recharge to ground water for 1941–53. It will be shown that overland runoff is small, because the soil and the underlying material are permeable, and that evapotranspiration is appreciable.

OVERLAND RUNOFF

Overland runoff is the part of the precipitation that moves directly over the land surface to ditches, creeks, streams, or other routes of surface flow, and which, therefore, leaves the area solely by an overland route. For the purposes of this report, overland runoff must be carefully distinguished from surface runoff because, in Long Island, most of the flow in ditches, creeks, and streams is seepage from the ground-water reservoir system, and is part of the final stage of the ground-water flow.

The following factors affect the amount of overland runoff in the Upton area:

1. The distribution, frequency, and intensity of precipitation.
2. The temperature and humidity of the air.
3. The topography, terrain, and vegetation.
4. The characteristics of the soil zone and of the underlying material, particularly if they are temporarily frozen or saturated.
5. The works of man, insofar as they make the land surface more or less permeable.

The topography of Suffolk County has already been described (de Laguna, 1963). The land surface in the Upton area is somewhat irregular; in part it is a hilly moraine; in part it is a flat or gently undulating outwash plain and intermorainal area.

The unconsolidated deposits at and below the land surface are porous and rather permeable in most localities and permit rapid infiltration of precipitation to the water table. The amount of water that moves down through the soil zone in the Upton area varies somewhat with the soil type, but in general it is relatively large because the soil is mostly sand or sandy silt and the underlying formations, to depths of as much as 100 feet and more, are sand and gravel. Clayey soils occur at and near the land surface in only a few localities in the Upton area. These soils are less permeable than the sand, particularly when wet because then they swell and both pores and cracks close.

Sandy soils are much less subject to such swelling. But even in clayey areas the precipitation eventually reaches the water table by circuitous routes. In the soil zone the intake rate is high when the soil is dry but decreases as the soil takes up moisture. The uncleared lands in the inland parts of the Upton area are covered by brush and trees, mostly small oak and pine. Because the soil zone in these wooded areas contains much organic matter, it is porous and permeable and water can pass downward easily; there is little evidence of surface flow and ponding of water in depressions in these areas, even during and after heavy rainstorms.

Where the trees have been cleared, the original soil structure is changed, and the infiltration capacity is generally reduced. In many of these localities there is some evidence of erosion and of ponding of water in the depressions.

In built-up areas in the western and central parts of the island, runoff from roofs and road surfaces is usually concentrated to produce overland flow, which is led to open or wooded areas or to shallow drainage pits, from which it percolates down to the water table. Near the margins of the island, some small streams do receive overland runoff, especially where there are large roof areas nearby, or where there are cleared, packed areas, such as duck pens. In built-up communities, such as Patchogue, gutters and sewers feed overland flow to streams and eventually to the sea, but because there is nothing of this sort within several miles of the Laboratory, overland flow caused by the works of man is inappreciable in the Upton area. In Long Island there is, most of the time and in most localities, a much easier passage for water vertically downward to underground reservoirs than overland to surface outlets.

The detailed hydrograph of a stream can sometimes be used to determine the approximate overland runoff from the area drained by that stream. The seasonal variation of ground-water seepage makes up the base pattern of the hydrograph, on which is superimposed the sharp peaks due to direct overland flow. The quantity of additional flow represented by a peak or peaks is an indication of the amount of overland flow for the period under consideration. At the gaging stations on the Carmans and Peconic Rivers the amount of water over and above base flow is, on an average, about 3 percent of the precipitation.

The amount of overland runoff for a particularly heavy storm was determined from daily discharge hydrographs of the Carmans and Peconic Rivers (fig. 5). On August 7, 1946, about 6½ inches of rain fell in the vicinity of Brookhaven National Laboratory. The recording rain gage at Riverhead Research Station indicated that as much as 4.72 inches fell during a 3-hour period from 9:00 a.m. to 12:00 noon on this date. The moisture in the soil zone before this storm was some-

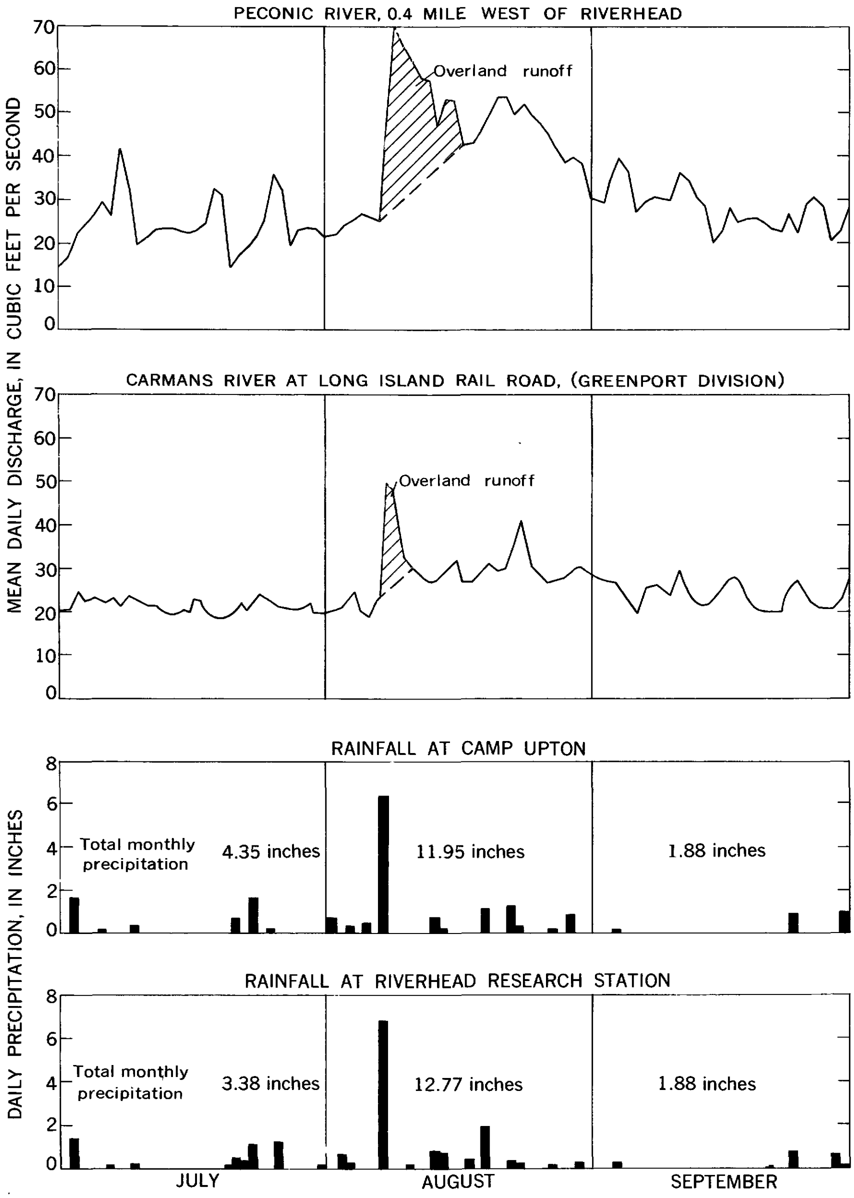

Figure 5.—Relation of overland runoff to precipitation for Carmans and Peconic Rivers after rainstorm of August 7, 1946.

what above normal for this time of year; in the 2-week period prior to the heavy storm, 2.91 inches of rain had fallen at Camp Upton and 3.00 inches at the Riverhead Research Station. The mean daily discharge for the Carmans River for July, August, and September 1946 was computed from the continuous gage-height records obtained at the stream-gaging station on the Carmans River at the Long Island Rail Road (Greenport Division) at Yaphank, and for the Peconic River from similar records from the gaging station near Riverhead. The overland runoff to the Carmans River after the storm of August 7 (represented by that portion of the hydrograph above the dashed line in fig. 5) was about 52 second-foot days. The Carmans River is a perennial stream for about 4 miles upstream from the gaging station, and in this reach it loses about 26 feet in altitude, about 14 feet of which is in drops across the spillways of two ponds. The surface area of these two ponds is 40–45 acres. The assumption that the 6 inches of rain which fell on the pond surfaces drained off into the stream channels would account for about 22 acre-feet or 11 second-foot days. The remaining 41 second-foot days is the overland runoff of about 71 square miles, the topographic drainage area of the Carmans River above the gage. It amounts to 0.022 inch, or about 0.3 percent of the storm rainfall.

Similarly, the overland runoff represented by the area above the dashed line in the daily discharge hydrograph (fig. 5) for the Peconic River for August 1946 is about 220 second-feet. For the topographic drainage of the Peconic River above Riverhead, about 75 square miles, this amount is equivalent to an overland runoff of about 0.1 inch, or 1.6 percent of the storm rainfall. The storm runoff for the Peconic River is less reliable than that for the Carmans River because of complications resulting from streamflow regulation and bank and channel storage in the complex drainage of the Peconic River basin.

Leggette (1940) made similar analyses of the overland runoff from three rainstorms which occurred in Long Island in the summer of 1938. From his study he concluded that, in southern Nassau County, where the geology and topography are similar to those in the southern part of the Upton area, overland runoff was 0.05 inch (4.1 percent) for the 3.65-inch storm of June 26–28, 0.15 inch (2.0 percent) for the 7.45-inch storm of July 18–24, and 0.25 inch (2.3 percent) for the 11.0-inch storm during the hurricane period of September 17–21.

Overland runoff to streams in the Upton area is small, even after unusually high precipitation, and averages less than 5 percent of the precipitation, or less than about 2 inches a year.

EVAPOTRANSPIRATION

Evapotranspiration is the combination of evaporation from land and water surfaces and transpiration by plants. It is a biological as well as a physical process. The rate of evapotranspiration depends on the temperature, humidity, and movement of the air, and also on soil moisture, plant cover, and land management. It can be changed by the various works of man.

Transpiration is the biological process by which water is lost to the atmosphere from the leaves of plants. Water is absorbed from the soil by the plant roots, circulated through the plant and discharged through the leaves in part to cool the plant. Transpiration rates are largely controlled by the rates of plant growth, by the temperature, and by the movement and humidity of the air.

When roots do not reach the capillary zone just above the water table, the amount of soil moisture available for plant growth is the quantity between field capacity and the wilting point; for sandy loam this is about 1 inch of water per foot of depth. Field capacity is that quantity of water retained against gravity drainage by capillary forces. The wilting point is the moisture content of the soil below which plants cannot draw water from the soil.

Thornthwaite (1948) introduced the concept of potential evapotranspiration in an effort to set up a better standard for the measurement of evapotranspiration. He recognized that evapotranspiration is restricted when precipitation is deficient but that it is largely independent of precipitation at other times. Because there are likely to be some months of deficient precipitation, he suggested the need to distinguish between potential and actual evapotranspiration. Potential evapotranspiration is the water loss that would occur if there never were any moisture deficiency in the soil. When there is a deficiency in the soil zone, actual evapotranspiration is less than the potential evapotranspiration.

No instrument measures accurately either actual or potential evapotranspiartion. However, evapotranspiration has been measured approximately in the field and can be computed indirectly by empirical formulas.

Evaporation pans, of various designs, can be used for direct measurement of evaporation in the field, and buried tanks (lysimeters) can be used to measure infiltration through the soil zone. The results of all these methods, however, must be adjusted empirically before they can be applied to large areas, and there are many sources of error.

Evapotranspiration from a large area, such as a drainage basin, can be estimated from the difference between precipitation and surface runoff, provided that a sufficiently long time is covered so that changes in water storage in the soil may be neglected, and provided that allowance can be made for any flow of ground water out from the area. In general this last requirement can only be met in areas where the ground-water flow is so small that it can be neglected.

Empirical formulas for computing evapotranspiration from precipitation, temperature, humidity, and wind velocity have been developed by Meyer (1915, 1944) and Thornthwaite (1948). In the Meyer formula, the variables are time of year, temperature, and precipitation. Thornthwaite's formula uses two basic variables—temperature and hours of sunshine.

Evapotranspiration for the Upton area has been calculated (fig. 6) from (1) rainfall-runoff relationships, (2) the Meyer and Thornthwaite formulas (fig. 6), and (3) data from evaporation pans in New York City and Mineola, N.Y. The vapor transfer and similar methods were not used because not all of the neceessary data were available. An attempt was also made to measure ground-water recharge, and hence evapotranspiration, directly by means of a lysimeter.

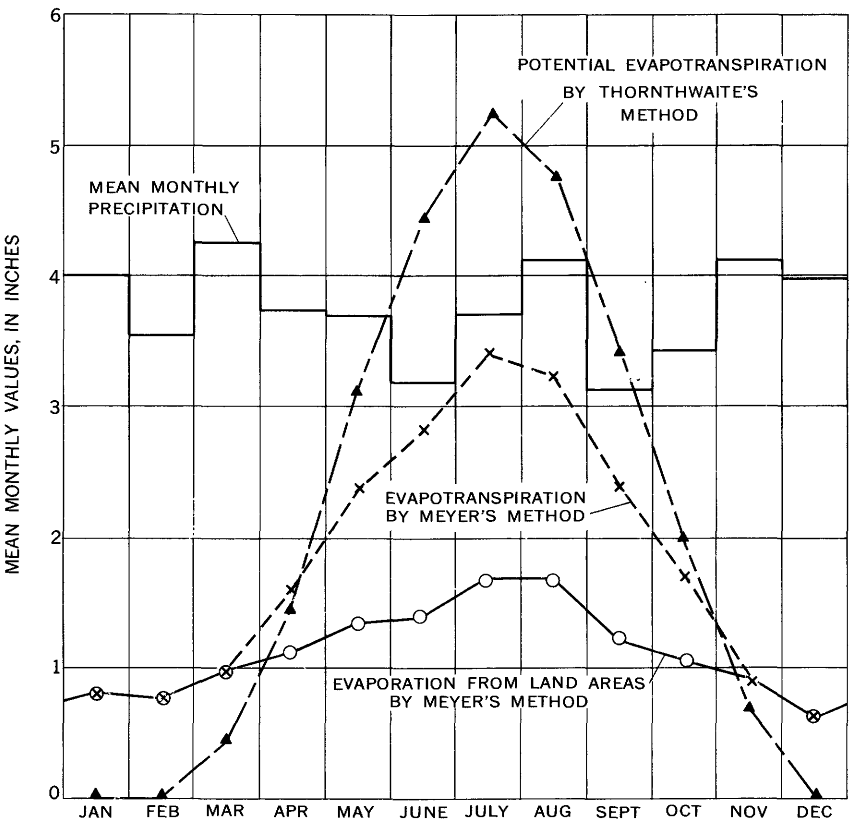

FIGURE 6.—Calculated average evapotranspiration, Brookhaven National Laboratory area.

EVAPOTRANSPIRATION FROM PRECIPITATION-RUNOFF RELATIONSHIP

Evapotranspiration cannot be determined from rainfall-runoff relationships in Suffolk County because the base flow of the streams, which is derived from ground-water seepage, does not come from the apparent topographic drainage area of the streams and because large but unkown amounts of ground water reach Long Island Sound and the ocean as seepage along the shoreline, not as streamflow. However, the evapotranspiration for Suffolk County can be estimated by comparison with areas of similar climate in nearby States where differences in geology make it possible to use the relation between precipitation and runoff. A study of the data presented by Williams and others (1940) for New Jersey, Pennsylvania, Massachusetts, and Connecticut, shows that such areas in these States have average rates of annual evapotranspiration of about 25 inches, with extremes of 21–29 inches per year, depending on the geology and topography. Because infiltration is rapid on Long Island and overland flow is small, the proper comparison is with the lower average rates and suggests that the average annual rate of evapotranspiration for Suffolk County is about 22–23 inches.

EVAPOTRANSPIRATION BY THE MEYER METHOD

Meyer's empirical formula provides a means for the separate computation of monthly values for evaporation from land areas and for transpiration by plants using two sets of graphs, after suitable coefficients for the locality under study have been selected. The graph for evaporation from land areas was originally constructed for conditions in the Midwestern United States. It allows a range in coefficient of from 0.6 for very sandy soil, from which evaporation is low because filtration is rapid, to 1.2 for areas where evaporation is high because of flat topography, clay soil, and low humidity. A comparison of the topography, soil, vegetation, and climate of Suffolk County with the criteria given by Meyer suggests that for this area the coefficient should be 0.8. The mean monthly temperature used was the arithmetic average of the following records: Medford (1906–27), Patchogue (1938–48), Riverhead (1939–48), and Setauket (1886–1945). The monthly precipitation used was the weighted average listed for the six stations in table 2.

For transpiration alone, Meyer gives an upper limit of 8 inches for annual transpiration by small trees and brush in north-central United States. This value was selected as the average for the Upton area after consideration of the controlling factors. This figure may be compared with Meyer's low value of 4 inches per year for the small, sparse, dwarfed pine on very sandy soil, and with his high value of

12 inches for swampy areas supporting a vigorous growth of trees and underbrush. The selection of 8 inches as the average yearly transpiration requires the application of a coefficient of 0.674 to Meyer's transpiration curve; a coefficient of 1.00 is the equivalent of 11.87 inches of transpiration.

With these two coefficients and the average monthly values for temperature and rainfall, a value of 21.82 inches was obtained for the annual evapotranspiration in Suffolk County in a year of average rainfall (table 4). Dry periods tend to reduce evapotranspiration more than wet periods tend to increase it, so that the average annual evapotranspiration suggested by Meyer's method is about 21.5 inches. This value is somewhat subjective because the investigator has a certain freedom in his choice of coefficients, guided though he is by Meyer's descriptions and criteria. The value given, however, is apparently correct to about an inch, if the method is valid.

POTENTIAL EVAPOTRANSPIRATION BY THE THORNTHWAITE METHOD

Thornthwaite's method for computing potential evapotranspiration in areas where plant growth is not restricted by a lack of soil moisture is based on temperature and hours of sunshine (Thornthwaite, 1948). Potential evapotranspiration thus computed for the Upton area, lat 41° N., is listed by months in table 3 and plotted in figure 6; it totals nearly 26 inches a year.

One of Thornthwaite's basic assumptions is that evapotranspiration is nearly independent of type of plant, vegetation, or soil, if there is always adequate soil moisture for plant growth. This assumption is probably an oversimplification, and the evidence now suggests that differences in soil and plant types cause variations in evapotranspiration even under conditions of adequate soil moisture. Mather (1952) reported fair agreement between values for annual potential evapotranspiration obtained experimentally and values computed by the Thornthwaite formula, but he suggested that the Thornthwaite formula underestimates water use for the winter months and overestimates it for the summer months. As shown in figure 6, Meyer's method also gives figures that are lower in the summer and higher in the winter than Thornthwaite's evapotranspiration.

EVAPOTRANSPIRATION FROM EVAPORATION DATA FOR FLOATING PAN ON FREE-WATER SURFACE

Various investigators, including Blaney (1952, p. 954), have suggested that evapotranspiration from a land surface that has vegetation growing on well-drained soil is less than evaporation from a large free-water surface under similar climatic conditions, provided that there is adequate soil moisture at all times and that the water table

TABLE 4.—*Evapotranspiration and residuals from precipitation on the basis of available date on average monthly temperatures and precipitation in central Suffolk County*

[Temperatures in degrees Fahrenheit, all other figures in inches of water]

	January	February	March	April	May	June	July	August	September	October	November	December	Annual
Average mean monthly temperature [1]	30.0	29.7	37.6	46.8	57.2	66.0	71.7	70.3	64.2	54.3	42.9	32.8	50.3
Weighted average monthly precipitation [2]	3.99	3.54	4.25	3.74	3.70	3.18	3.71	4.14	3.12	3.44	4.13	3.99	44.93
Evaporation from land areas (Myer method)	.82	.77	1.00	1.13	1.36	1.41	1.71	1.71	1.23	1.08	.95	.65	13.82
Transpiration (on basis of 8 in. per average year)	.00	.00	.00	.49	1.03	1.43	1.70	1.53	1.18	.64	.00	.00	8.00
Evapotranspiration (Meyer method)	.82	.77	1.00	1.62	2.39	2.84	3.41	3.24	2.41	1.72	.95	.65	21.82
Potential evapotranspiration (Thornthwaite method)	.00	.00	.43	1.45	3.11	4.45	5.24	4.76	3.42	2.02	.71	.00	25.59
Precipitation minus evapotranspiration (Meyer method)	3.17	2.77	3.25	2.12	1.31	.34	.30	.90	.71	1.72	3.18	3.34	23.11
Precipitation minus potential evapotranspiration (Thornthwaite method)	3.99	3.54	3.82	2.29	0.59	−1.27	−1.53	−0.62	−.30	1.42	3.42	3.99	19.34

[1] From arithmetic average for Medford (1906-27), Patchogue (1938-48), Riverhead Research (1939-48), and Setauket (1886-1945).
[2] Weighted average for six stations listed in table 2.

is beyond the reach of plant roots. These conditions are met in the Upton area. The soil is sandy and porous; the water table is within 5 feet of the land surface in less than 7 percent of the area (Spear, 1912); and less than 1 percent of the area is fresh-water ponds, lakes, and streams. Therefore, evaporation-pan data may be used to set an approximate upper limit for evapotranspiration.

Data on evaporation from pans are available for Central Park in New York City and for a storm-water recharge basin in Garden City, Long Island, N.Y. Evaporation records for the New York City pan for the period May through October are available for the years 1944 through 1953; observed values ranged from 29 to 36 inches and averaged 32.27 inches. This value must first be multiplied by 1.33 to give the equivalent evaporation for the full year (Horton, 1943, p. 747), and then multiplied by 0.7, the generally accepted coefficient for this type of pan. The corrected annual rate of evaporation from a large free-water surface in this area is, therefore, about 30 inches. Similar data collected at Garden City gives a corrected annual rate of evaporation of 23 inches. A comparison with evaporation rates in other parts of the country suggests that the value of 30 inches is probably the more nearly correct.

SUMMARY OF COMPUTED VALUES FOR EVAPOTRANSPIRATION

Annual evapotranspiration in the Upton area was found to be 22–23 inches from rainfall-runoff relationships, about 21–22 inches by the Meyer method, less than 26 inches by the Thornthwaite method, and less than 30 inches from evaporation-pan data. There is fair agreement between these values. For the purposes of this report the Meyer method is preferred, principally because it takes into consideration the change in evaporation opportunity with variation in precipitation. The average evapotranspiration in the Upton area is, therefore, taken to be about, and no more than, 22 inches per year. Annual variations in temperature and precipitation are believed to produce corresponding variations in evapotranspiration of 19–24 inches (table 5). Variations in topography, geology, soil cover, and vegetation are estimated to produce corresponding local variations in evapotranspiration of 15–30 inches. The average figure of 22 inches, and the range of from 15 to 30 inches, are approximate, inasmuch as they have been determined by indirect, empirical methods. The data obtained at Brookhaven by use of a lysimeter are in agreement with this conclusion.

QUANTITATIVE ESTIMATES OF RECHARGE

MONTHLY AND YEARLY FIGURES 1941–53

Monthly evapotranspiration for the Upton area, computed by the Meyer method for the 12 water years, October 1, 1941, to September

30, 1953, using the same coefficients as on the preceding pages, is listed in table 5. The monthly residual is almost all recharge to ground water because recharge to ground water (below soil zone and below reach of plants), plus overland runoff, equals precipitation minus evapotranspiration. This method assumes that any accumulated moisture deficiency in the soil zone will have first claim on any water later entering the soil. It was, therefore, assumed in the computations that there can be no recharge to the ground water unless the soil is at or above field capacity, and that any soil moisture deficiency below that point has to be made up prior to any ground-water recharge, when precipitation occurs. In the Upton area, moisture storage between field capacity and permanent wilting, in the top 3 feet of soil, was assumed to be 3 inches; this amount is also taken to be the maximum possible soil moisture deficiency. Figures for deficiencies, which can accumulate from month to month, are indicated by a minus sign.

Table 5 shows for each month, from October 1941 to September 1953, the mean temperature, the total precipitation, the evapotranspiration as calculated by Meyer's method, and the difference between precipitation and evapotranspiration. This difference is regarded as recharge to the water table and overland runoff. As we have seen, overland runoff is so small in central Suffolk County that it is almost negligible.

The largest monthly recharge found was 8.50 inches for March of 1953, the result of a total rainfall that month of more than 10 inches. Evapotranspiration exceeded precipitation in 35 months of these 12 years. The result was a soil-moisture deficiency and no recharge to the water table. The largest deficiency was built up during the months of May, June, July, and August 1944 and totaled 3.00 inches, and theoretical maximum deficiency. The soil-moisture deficiency at the end of the water year (September 30) varied from zero in most years to 1.68 inches in 1953, and because of these variations the sum of the evapotranspiration plus the infiltration and overland runoff is not exactly equal every year to the precipitation. The largest annual recharge to the water table (by water years) was 31.99 inches in the 1951–52 period, and the smallest was 11.80 inches in the 1946–47 period. These differences are due largely to differences in precipitation, which was 55.66 inches in 1951–52 and only 32.25 inches in 1946–47, not to differences in evapotranspiration, which was 23.36 inches in 1951–52 and 20.86 inches in 1946–47. (See also fig. 7.) For a range in precipitation from 55½ inches to 32¼ inches, the coresponding recharge ranged from 32 to less than 12 inches. In other words, for a 70-percent increase in precipitation there is almost a 300-percent increase in recharge to ground water.

TABLE 5.—*Calculated monthly and yearly recharge to ground water (plus overland runoff), October 1941 to September 1953*

Temperatures in degrees Fahrenheit, all other figures in inches of water.
Evapotranspiration computed from temperature and precipitation by the Meyer method.
Soil-moisture deficiency is indicated in any month if the evapotranspiration (plus accumulated soil-moisture deficiency from preceding month) is greater than precipitation.
Ground-water recharge (plus overland runoff) is the difference between precipitation and evapotranspiration (plus accumulated soil-moisture deficiency from preceding month).
Zero recharge is assumed in months having indicated soil-moisture deficiency.
Overland runoff averages less than 5 percent of recharge plus runoff.
Monthly temperature listed is the mean monthly temperature at Patchogue from October 1941 to March 1947 and August 1947 to July 1948, at Riverhead Research from April to July 1947, and at the Brookhaven National Laboratory (Meteorology Group) from August 1948 to September 1953.
Monthly precipitation listed is the average at Patchogue and Riverhead Research from October 1941 to December 1942, the monthly at sewage plant gage at Brookhaven National Laboratory from January 1943 to July 1948, and the average for the three gages at the Brookhaven National Laboratory from August 1948 to September 1953.
Figures preceded by *e* are partly estimated.

	October	November	December	January	February	March	April	May	June	July	August	September	Annual
1941–42													
Mean temperature	56.6	45.8	37.6	29.3	29.6	40.6	48.1	60.6	67.2	72.2	70.2	64.6	51.9
Total precipitation	2.10	3.36	3.27	3.37	2.83	7.27	1.12	2.19	4.61	5.83	7.51	1.78	45.24
Calculated evapotranspiration	1.50	1.01	.64	.76	.68	1.63	1.09	2.14	3.36	4.15	4.25	2.01	23.22
Soil-moisture deficiency	[1] 1.38											.23	
Recharge (plus overland runoff)	.22	2.35	2.63	2.61	2.15	5.64	.03	.05	1.25	1.68	3.26	.00	21.87
1942–43													
Mean temperature	55.3	44.2	30.4	29.4	32.0	37.3	43.0	57.8	71.4	72.4	71.6	63.8	50.7
Total precipitation	4.20	4.62	5.67	2.24	2.37	3.35	2.76	2.42	3.89	3.24	1.25	2.24	38.25
Calculated evapotranspiration	1.95	1.13	.96	.62	.68	.88	1.08	2.08	3.45	3.31	2.28	2.08	20.50
Soil-moisture deficiency										.07	1.10	.94	
Recharge (plus overland runoff)	2.02	3.49	4.71	1.62	1.69	2.47	1.68	.34	.44	.00	.00	.00	18.46
1943–44													
Mean temperature	53.2	41.4	30.2	31.8	30.8	36.0	45.2	60.8	67.9	73.4	73.5	65.4	50.8
Total precipitation	4.85	2.93	1.92	4.23	1.72	5.67	4.76	1.30	1.78	.65	1.53	6.98	38.32
Calculated evapotranspiration	1.97	.66	.32	.90	.58	1.16	1.69	1.89	2.50	2.30	1.57	3.62	19.16
Soil-moisture deficiency								.59	1.31	2.96	3.00		
Recharge (plus overland runoff)	1.94	2.27	1.60	3.33	1.14	4.51	3.07	.00	.00	.00	.00	.36	18.22
1944–45													
Mean temperature	53.0	43.4	30.8	25.2	31.3	44.4	52.4	54.6	66.6	71.6	*e* 71.4	*e* 69.7	51.2
Total precipitation	2.20	8.43	2.99	1.98	3.82	2.03	3.68	5.16	1.43	3.79	2.27	1.54	39.32
Calculated evapotranspiration	1.31	1.59	.46	.54	.83	1.04	2.09	2.60	2.24	3.48	2.66	2.25	21.09
Soil-moisture deficiency									.81	.50	.89	1.60	
Recharge (plus overland runoff)	.89	6.84	2.53	1.44	2.99	.99	1.59	2.56	.00	.00	.00	.00	19.83

1945–46													
Mean temperature	52.6	45.7	28.3	31.5	30.7	44.2	46.0	55.8	64.9	70.3	68.0	65.8	50.3
Total precipitation	3.38	6.00	5.90	3.65	2.34	2.23	2.09	4.48	3.94	4.35	11.95	1.88	52.19
Calculated evapotranspiration	1.57	1.71	.71	.82	.66	1.06	1.18	2.51	3.00	3.53	4.90	2.12	23.77
Soil-moisture deficiency												.24	
Recharge (plus overland runoff)	.21	4.29	5.19	2.83	1.68	1.17	.91	1.97	.94	.82	7.05	.00	27.06
1946–47													
Mean temperature	57.6	48.2	36.4	36.0	29.5	36.0	e48.0	e57.6	e66.0	e73.3	73.2	68.0	52.5
Total precipitation	.32	1.13	3.19	3.47	.39	2.20	6.21	3.97	3.15	2.85	3.77	1.70	32.25
Calculated evapotranspiration	1.00	.66	.68	.88	.32	.68	2.46	2.50	2.83	3.27	3.33	2.25	20.86
Soil-moisture deficiency	.92	.45								.42		.55	
Recharge (plus overland runoff)	.00	.00	2.06	2.59	.07	1.52	3.75	1.47	.32	.00	.02	.00	11.80
1947–48													
Mean temperature	58.0	41.4	30.6	23.6	27.0	37.4	46.2	55.1	64.1	72.6	70.9	62.5	49.1
Total precipitation	3.62	9.23	3.18	5.60	2.23	4.52	4.09	6.97	4.35	3.61	1.32	1.26	49.98
Calculated evapotranspiration	2.04	1.40	.52	.63	.60	1.04	1.67	3.05	3.08	3.49	2.23	1.66	21.41
Soil-moisture deficiency											.91	1.31	
Recharge (plus overland runoff)	1.03	7.83	2.66	4.97	1.63	3.48	2.42	3.92	1.27	.12	.00	.00	29.33
1948–49													
Mean temperature	52.3	48.1	34.0	36.3	35.5	38.4	49.1	57.7	68.1	73.9	71.4	61.2	52.2
Total precipitation	4.84	5.61	6.59	5.66	4.68	2.73	3.60	3.36	.02	2.88	5.11	3.15	48.23
Calculated evapotranspiration	1.90	1.70	.96	1.18	1.04	.76	1.82	2.36	1.69	3.30	3.66	2.20	22.57
Soil-moisture deficiency									1.67	2.09	.64		
Recharge (plus overland runoff)	1.63	3.91	5.63	4.48	3.64	1.97	1.78	1.00	.00	.00	.00	.31	24.35
1949–50													
Mean temperature	57.4	41.7	34.2	38.0	29.0	33.1	43.3	53.4	64.2	70.0	67.4	59.8	49.3
Total precipitation	1.66	3.10	3.06	2.74	4.50	3.60	2.69	5.27	2.70	3.39	4.27	1.32	38.30
Calculated evapotranspiration	1.45	.72	.56	.76	.84	.84	1.11	2.55	2.64	3.20	3.03	1.52	19.22
Soil-moisture deficiency												.20	
Recharge (plus overland runoff)	.21	2.38	2.50	1.98	3.66	2.76	1.58	2.72	.06	.19	1.24	.00	19.28
1950–51													
Mean temperature	53.8	43.4	32.0	32.2	31.8	32.8	46.8	57.0	63.7	70.7	68.6	62.4	49.6
Total precipitation	1.77	4.64	4.20	4.42	5.04	4.84	3.08	3.75	2.34	2.08	3.48	1.26	40.90
Calculated evapotranspiration	1.26	1.07	.68	.92	.96	.91	1.48	2.38	2.43	2.79	2.87	1.66	19.41
Soil-moisture deficiency									.09	.80	.19	.59	
Recharge (plus overland runoff)	.31	3.57	3.52	3.50	4.08	3.93	1.60	1.37	.00	.00	.00	.00	21.88

See footnote at end of table.

TABLE 5.—*Calculated monthly and yearly recharge to ground water (plus overland runoff), October 1941 to September 1953*—Continued

	October	November	December	January	February	March	April	May	June	July	August	September	Annual
1951–52													
Mean temperature	53.4	40.6	34.9	32.4	32.0	36.8	48.9	55.6	67.6	74.3	70.1	62.0	50.7
Total precipitation	4.77	6.72	6.03	6.32	3.57	5.07	3.75	7.07	2.78	.83	7.40	1.35	55.66
Calculated evapotranspiration	1.98	1.28	1.00	1.09	.81	1.12	1.81	3.13	2.84	2.46	4.20	1.64	23.36
Soil-moisture deficiency									.07	1.70		.29	
Recharge (plus overland runoff)	2.20	5.44	5.03	5.23	2.76	3.95	1.94	3.94	.00	.00	1.50	.00	31.99
1952–53													
Mean temperature	49.5	43.0	35.0	33.2	35.0	39.1	47.1	58.3	65.0	70.7	68.3	63.3	50.6
Total precipitation	.53	3.21	4.11	6.53	4.11	10.02	5.28	3.50	1.92	2.66	2.39	.83	45.09
Calculated evapotranspiration	.59	.76	.74	1.12	.94	1.52	1.95	2.45	2.35	2.99	2.53	1.61	19.55
Soil-moisture deficiency	.35								.43	.76	.90	1.68	
Recharge (plus overland runoff)	.00	2.10	3.37	5.41	3.17	8.50	3.33	1.05	.00	.00	.00	.00	26.93

[1] Soil-moisture deficiency at end of September 1941.

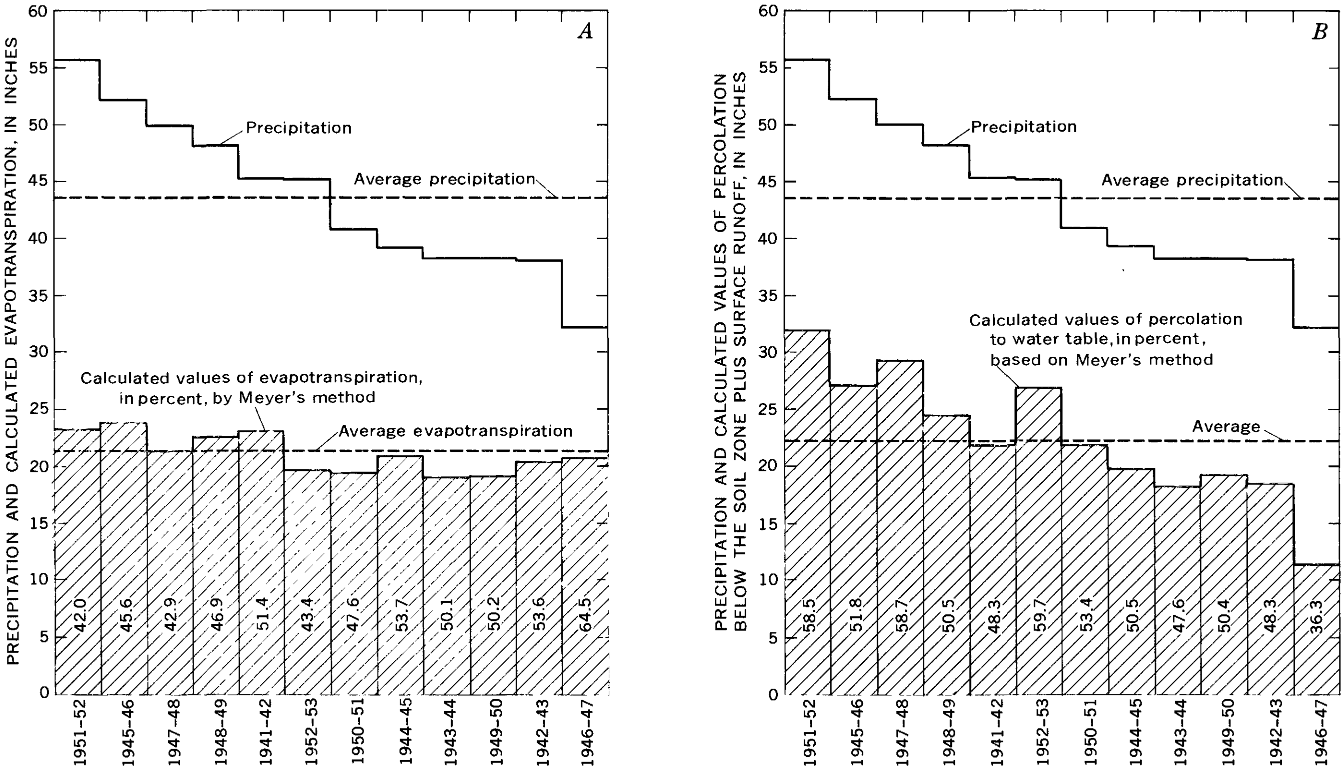

FIGURE 7.—Relation of precipitation, evapotranspiration, and ground-water recharge, Brookhaven National Laboratory area. *A*. Comparison by water years of precipitation and evapotranspiration; arranged in order of decreasing precipitation, from October 1941 to September 1953. *B*. Comparison by water years of precipitation and recharge to ground-water (plus overland runoff); arranged in order of decreasing precipitation from October 1941 to September 1953.

The average annual infiltration plus overland runoff for the 12 years was 22.59 inches. This value may also be computed from the average mean monthly temperatures and average precipitation for each of the calendar months, from which one may calculate average monthly evapotranspiration. From these 12 monthly averages, an average yearly rate of infiltration plus overland runoff of 22.06 inches may be calculated; it is 0.53 inch less than the average annual value found by computing by individual months (table 5), a difference of less than 5 percent.

SUMMARY OF COMPUTED RECHARGE

During the 12 water years from October 1941 to September 1953, the precipitation averaged 43.64 inches, evapotranspiration averaged 21–22 inches, and the residual (mostly recharge to ground water) averaged about 22 inches. During this period, the residual varied appreciably from month to month and from year to year. It was over 7 inches on 3 different months and was zero for about 2–3 months in an average year. The annual rate of infiltration (plus overland runoff) was as much as 31.99 inches in 1951–52, 29.33 inches in 1947–48, 26.93 inches in 1952–53, and as little as 11.70 inches in 1946–47.

Over a 50- to 100-year period, precipitation in the Upton area varies from a minimum of perhaps less than 30 inches per year to a maximum of perhaps more than 60 inches per year. The average annual evapotranspiration, over a similar period, will range from a minimum of 15 inches per year where the soil is very sandy to a maximum of 30 inches per year, and perhaps more, in swampy areas. Replenishment to ground water in the Upton area may, therefore, be as low as 10 inches in some areas in dry years and as much as 35 inches in other areas in wet years. Locally, recharge to ground water may even vary from practically nothing in some swampy localities, when precipitation is extremely low, to as much as 45 inches in sandy localities, when precipitation is extremely high.

GROUND WATER IN UPPER PLEISTOCENE DEPOSITS

OCCURRENCE

The 200 feet of upper Pleistocene deposits in the Upton area consists of sand and gravel, some silt and clay layers, and also some till in the two morainal areas. Water first enters through the soil zone. The zone of aeration, about 50–60 feet in average depth, serves both as a sizable underground reservoir and also as the conduit for water moving downward to the zone of saturation. Locally within the zone of aeration are bodies of perched and semiperched water, held up by layers of relatively impermeable material, one each in the northern, northwestern, and eastern sections of the Laboratory tract, and one east of the Laboratory tract beyond the peconic River. A few small areas of this kind occur in the extreme west-central section of the Upton area. The major areas underlain by relatively impermeable layers above the zone of saturation are shown on plates 1–4.

The zone of saturation in the upper Pleistocene deposits averages about 140–150 feet in thickness. This zone serves both as an immense storage reservoir and also as the principal conduit for water moving from points of recharge to points of discharge.

THE WATER TABLE

MAPS OF THE WATER TABLE

The water table in the Upton area is defined by the position of the static water level in wells ending in the zone of saturation in the upper Pleistocene and Recent deposits. Plates 1 and 2 show the position of the water table on August 29–31, 1951, and July 28–30, 1952. The water-level contours are based on readings in about 120 wells, 50 of them inside the Laboratory area, and also on the altitudes of the water surface in streams, ditches, ponds, and lakes at about 35 additional points. Only a few of the wells are plotted on plates 1 and 2. Plates 3 and 4 show the position of the water table on October 1–3, 1952, and April 25, 1953, and also the locations of all the observation wells within the Laboratory area.

NETWORK OF OBSERVATION WELLS

A table giving complete information on the location, owner, use, depth, method of construction, size of casing, screen setting, altitude of measuring point, and height above land surface for all wells used in this study is on file with the U.S. Geological Survey and State and Laboratory authorities. The well numbers, assigned by the New York State Water Power and Control Commission in chronological order, have no particular geographical significance. The letter S preceding the number signifies Suffolk County. The code numbers of the points used in determining surface-water stages were assigned by the Survey staff at Brookhaven National Laboratory. Letters C and P preceding the number are for measuring points on or near the Carmans and Peconic Rivers, respectively. Some points on the larger lakes or ponds are identified only by their names. The tables on file also give information on the location of all measuring points other than wells, and also their descriptions, altitudes, and the altitude of the accompanying bench marks.

Third-order accuracy (or better) was maintained in the leveling used to determine the altitudes of the measuring points at wells, of the surface-water observation stations, and of bench marks; that is, the error of closure of the level circuit, in feet, did not exceed the length of circuit, in miles, divided by 0.5. For short runs the allowable error of closure, in feet, did not exceed the number of setups divided by 0.008. All levels are referred to the 1929 mean sea-level datum of the U.S. Coast and Geodetic Survey. Observed water levels are accurate within at least 0.1 foot.

RELATION OF WATER TABLE TO SHALLOW, PARTLY CONFINING LAYERS

In some areas (see pls. 1–4) of low permeability, beds of silt or clay occur in the zone of aeration. In these areas, where shallow water is perched or semiperched, the water table is defined by water levels in wells screened below this material. The maximum depth of this retarding zone below land surface is about 30 feet; only at well S9123 east of the Laboratory was the bottom of the less permeable material found to be deeper, at about 50 feet below land surface. The water surface, mapped in plates 1–4 will be referred to as the water table, even though the water is confined to some degree part of the time in localities where less permeable material occurs at shallow depths.

In the Peconic River valley east of the Laboratory, from about Manorville to Riverland, an intersubstage (de Laguna, 1963, p. 32) occurs at about middepth in the glacial sands. In this locality the water-table map is based on levels in wells ending above this clay.

SIGNIFICANT FEATURES OF THE WATER TABLE

The shape of the water table reflects the location of areas of recharge, areas of discharge, and of the ground-water divides. (See pls. 1–4; fig. 34A.) The water table in the Upton area suggests the cross section of a bullet, flattened at the tip and pointing eastward; the south side is somewhat irregular. The depressions and troughs in the contour pattern are ground-water discharge areas.

In the Upton area, the main ground-water divides lies about 3–5 miles south of Long Island Sound and roughly parallel to it. East of the eastern boundary of the Laboratory tract a second ground-water divide appears, which defines the southern boundary of the area contributing ground water to the Peconic. The north branch of the divide extends beyond the Upton area into the North Fork of Suffolk County, and the south branch extends into the South Fork. There are not enough water-level data to define the south branch accurately.

North of the divide, ground water moves northward to Long Island Sound. South of the divide, the ground water moves southward to Great South Bay and Moriches Bay, either directly or by way of streams. In general, the ground water from the area between the two branches of the divide moves out eastward to the Peconic River and

Peconic Bay. Details of the movement vary with the stage and slope of the water table.

The highest part of the water table in the Upton area is the west-central section where it is about 55 feet above sea level; the lowest is along the shoreline, where it stands at about mean sea level. A few miles west of the Upton area (fig. 34A), the water table is about 60 feet above sea level (Lusczynski and Johnson, 1951). The slope of the water table ranges from more than 10 feet per mile to less than 2 feet per mile; in the Laboratory tract, the slope averages about 5 feet per mile.

DEPTH TO WATER TABLE

The depth to the water table in the Upton area ranges from less than 0.1 foot along the shorelines to more than 200 feet under the higher hills on the north shore and averages about 50–60 feet. North of the ground-water divide, and along the south branch of the divide, the average depth to the water table is about 80 feet; between the divides and to the south it is about 40 feet. Figure 8 gives five north-south profiles (pls. 1, 2) showing the water-table altitudes as of July 28–30, 1952, when the water table was slightly below the average stage for 1941–53. As the sections show, from the north shore the land surface rises abruptly about 150 feet or more to a line of hills, part of the Harbor Hill moraine. Here the depths to water are from 75 to 150 feet and locally even 200 feet. Just south of the Laboratory area, the water table is also relatively deep beneath another line of east-west hills known as the Ronkonkoma moraine. Profiles showing the approximate altitudes of the land surface and the water table are shown in figure 8. In the low land between the two moraines the water table is at somewhat shallower depths, and because this wide valley slopes gently eastward, in the eastern part of the Laboratory area and in the Manorville area the water table is even shallower, within 5–10 feet of the land surface. The Peconic River originates in this valley and flows eastward between the two moraines. The headwaters of the Carmans River also lie in this intermoraine belt. South of the Ronkonkoma moraine, the land slopes gently toward the south, and the depth to water decreases southward, so that the land surface and the water table converge.

Figure 9 shows the depth from the land surface to the water table in the Laboratory tract. The depths vary from less than 10 feet along streams in the eastern and northern parts of the Laboratory, to more than 80 feet in a belt extending from the center of the Laboratory tract, near the reactor, to the hospital in the southwest corner. The average depth to the water table is about 45 feet. Land-surface altitudes for this depth-to-water map were taken from the 10-foot con-

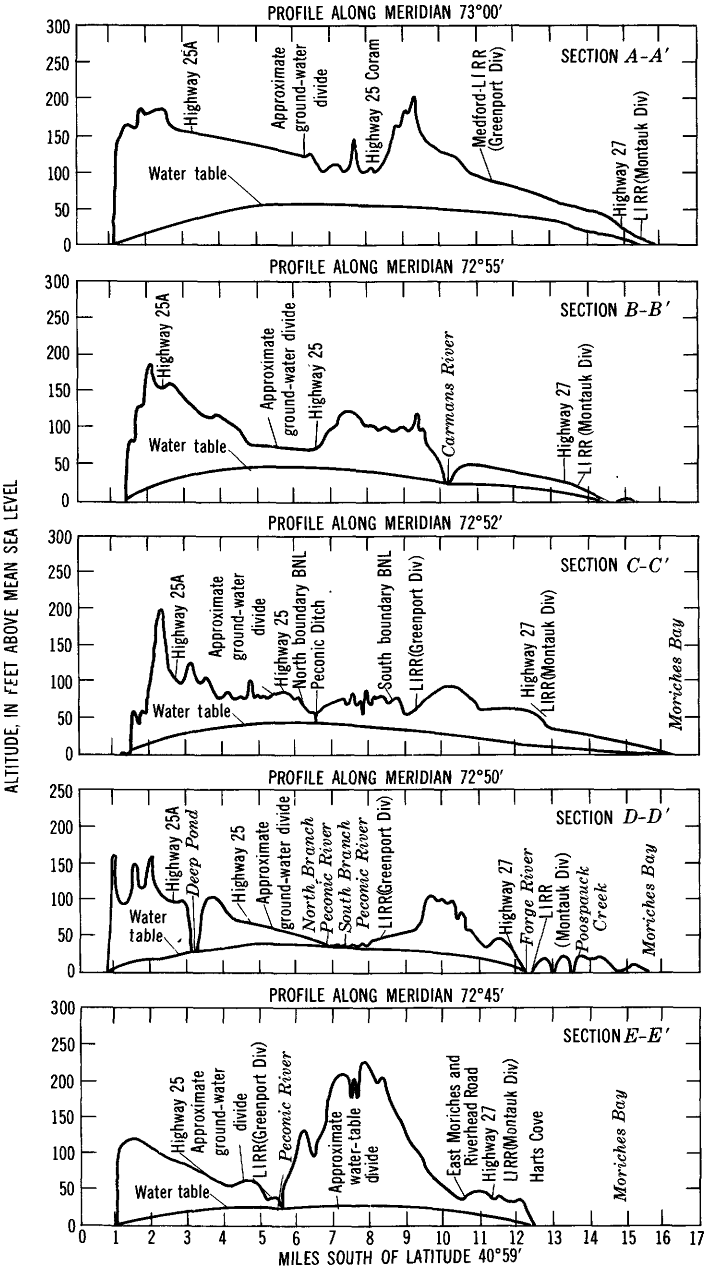

FIGURE 8.—Land surface and water table, July 29–30, 1951, along five meridians in the Upton area; positions are indicated on plates 1 and 2.

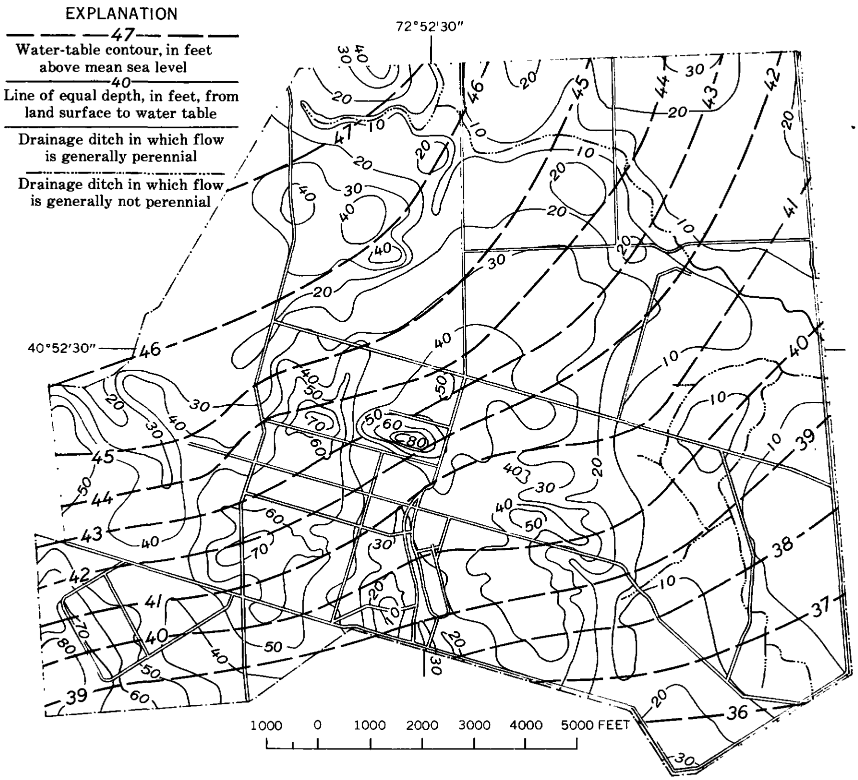

Figure 9.—Depths from land surface to water table, June 4–6, 1951, Brookhaven National Laboratory area.

tour topographic map entitled "Map of the South Tract A.M. 1.3101–1" dated February 1950, and from altitudes established during leveling by the U.S. Geological Survey. The water-table data used was that for June 4–6, 1951, from which a 1-foot interval contour map was prepared. The water table on this date was at about an average stage for the 1941–53 period. The depths to the perched and semiperched water are generally much less than those on the main water table.

FLUCTUATIONS OF THE WATER TABLE

Daily, monthly, seasonal, yearly, and long-term fluctuations of the water table are closely related to variations in recharge and discharge of ground water. When recharge is greater than discharge, water levels rise, the hydraulic gradients steepen, and subsequently the discharge increases. When recharge is less than discharge, the water levels drop, the hydraulic gradients lessen, and subsequently the discharge decreases. Rise and fall of the water table maintains the natural balance between recharge and discharge.

At any point on the water table, the amount of rise and fall and the time of occurrence of the seasonal highs and lows shows the interrelated effect of the following three principal factors:
1. The amount and rate of recharge at the water table.
2. The depth to the water table.
3. The relation of the point to the general recharge-discharge pattern.

These factors are illustrated by the hydrographs of water-table wells and one water-table pond, in or near the Laboratory area (fig. 10). Below the hydrographs, for easy comparison, are plotted the rainfall, the evapotranspiration, and the difference between these, which is mostly recharge. The rise and fall of water levels in the hydrographs correlates better with this value for recharge than it does with the precipitation.

The hydrographs are all very similar to one another, the principal differences being in range; there are also minor differences in the times of the highs and lows and in the smoothness of the fluctuations. The range of vertical fluctuations at any point on the water table is in general greater where the distance from the point, along the line of flow to the point of discharge, is greater, and where the average elevation of the point above sea level is greater. The times of seasonal highs and lows at such points also come a little later. These times are also somewhat retarded where the depth to the water table from the land surface is greater. A more pronounced result of greater depth to water is a smoothing out of the minor fluctuations, particularly those due to individual storms.

Well S3869 is located about 6 miles west of the center of the Laboratory area, near the main ground-water divide, where the average height of the water table is about 55 feet above sea level, and where the depth to the water table from the land surface is about 30 feet. The distance along the ground-water flow lines to the point of discharge into the Carmans River is about 2½–3 miles. The range of fluctuations during the period shown in figure 10 is about 5.7 feet. Because of the relatively great depth to the water table from the land surface in this area, the hydrograph is very smooth and shows only the major seasonal fluctuations. The highs and lows of these fluctuations come comparatively late in the year, largely because the point is in the center of the island near the top of the water-table divide and relatively distant from the main areas of discharge.

Well S3533, about 3 miles west of the center of the Laboratory (pl. 1), is similarly located where the average height of the water table is high, about 47 feet above sea level, but the depth to the water table from the land surface in this area is somewhat less, only 15 feet at the well. The distance to the Carmans River, along the lines of ground-

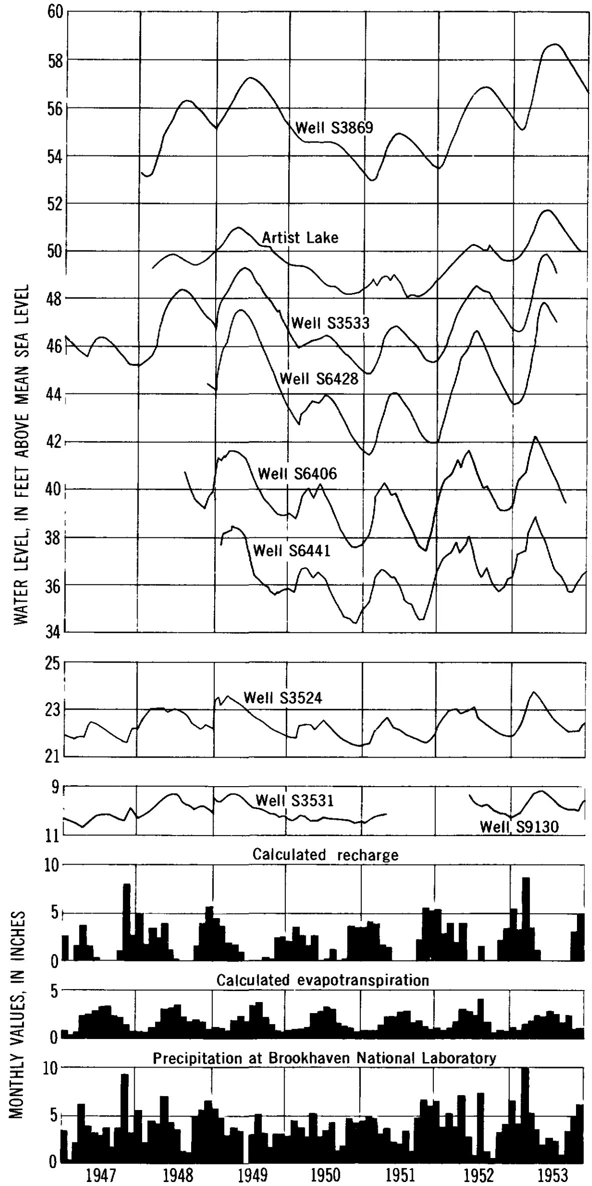

FIGURE 10.—Hydrographs of eight water-table wells and one water-table pond, central and northern Upton area.

water flow, is only about 2 miles. The range of water-table fluctuations is about 4.8 feet, less than at well S3869, in part because the average height of the water table is less and in part because the distance to the point of discharge is less. For these same reasons, and also because the water table is shallower, the seasonal highs and lows come a little earlier.

Well S6428 is in the middle of the Laboratory area (pl. 3). The depth to water, 37 feet, is comparable to that of well S3869. However, the distance to the point of discharge is greater, 5 miles as compared to 2½–3 miles, and the average height of the water table is less, 44 feet as compared to 55 feet. In consequence the range of fluctuations is greater, 6.4 feet as compared to 5.7 feet, and the seasonal highs and lows come somewhat earlier in the year.

Well S6406 is near the middle of the east boundary of the Laboratory area (pl. 3). It differs from the previously discussed wells in that the depth to water here is only about 7 feet. Consequently, the seasonal highs and lows come even earlier than at well S6428, and the fluctuations are not as smooth, because the water level here responds to many individual rainstorms. The well is near the Peconic River, but this part of the river is frequently dry in the fall and does not serve as a continuous point of ground-water discharge. Therefore, despite its proximity to the river, the water level in the well has a range of 4.6 feet.

Well S6441 is about a mile east of well S6406 (pl. 3). It shows fluctuations that are almost identical in character, timing, and amplitude because it is near the river, and because of the shallow depth of the water table, here only 10 feet below the land surface. However, well S6441 is screened below a shallow layer of poorly permeable silt and clay which produces artesian conditions. But despite the confining layer the depth is sufficiently shallow so that variations in pressure are transmitted from the surface quickly and with little loss, and the hydrograph resembles that of a well screened in unconfined permeable material just below the water table. The hydrograph of a well screened in one of the deeper artesian aquifers would, however, show characteristic differences.

The hydrographs of wells S3524 and S3521 show the effects of low general elevation of the water table, proximity to discharge areas, and a shallow depth to a water table, if not at the well itself, at least in much of the immediate area. Well S3524 is near the Peconic River about 3½ miles southwest of the center of the Laboratory area. Water levels here commonly rise after heavy winter rains at a rate of a foot a week. The average height of the water table above sea level is about 21 feet, and the range of water-table fluctuations is about 2 feet. Well

S3531, destroyed and later replaced by a new well, S9130, shows many of the same features. The average height of the water table here is only about 10 feet above sea level, and the range of water-table fluctuations is a little more than a foot.

Artist Lake is a water-table pond about 3½ miles west of the center of the Laboratory, and roughly a mile north of well S3533. The pond has no perennial surface-water outlet and no well-defined stream flowing into it, although it does receive some surface drainage. The pond bed is covered with a thin layer of clay, enough to impede but not to prevent the movement of ground water. The level of the pond rises and falls with fluctuations of the water table, although with characteristic differences. The range of fluctuations was 3.8 feet, as compared with a range of 4.8 feet at well S3533. The range in levels is less, in part because as the pond level rises with the water table in summer the pond area increases from about 15 to about 30 acres, and evaporation losses increase. During this period the ground water feeds the pond, so that the pond level is lower. In winter, when the water table in this area is low, the pond collects rainwater and tends to feed the ground water, and the pond level is higher. As would be expected, the pond level, unlike the water table, responds to individual heavy rainstorms.

The relations between the depth to the water table, the recharge-discharge pattern, and the range in stage of the water table are illustrated in figures 11–13. Figure 11 shows the maximum range in stage between the highest and lowest ground-water levels between 1941 and 1952, as determined from the many observation points in the Upton area. The range varied from 1 foot or less at the shoreline and along the lower reaches of the rivers to about 5–6 feet near the ground-water divides. The map shows clearly the marked difference between the Carmans River, which flows only slightly above sea level and receives an abundant flow of effluent ground water, and the upper Peconic River, which in this area flows well above sea level and receives but little ground water.

The line pattern, as drawn on figure 11 is necessarily generalized, especially in the northern third of the Upton area where comparatively little information was available. The lines here were drawn more or less parallel to the north shore and are admittedly approximate. Figure 11 is based on too short a record to show the maximum range possible for the water table, and indeed in the spring and summer of 1953, water levels reached stages 5–20 percent higher than those observed during the spring and summer of 1949. In 1953, however, not enough observations were made to draw a new contour map.

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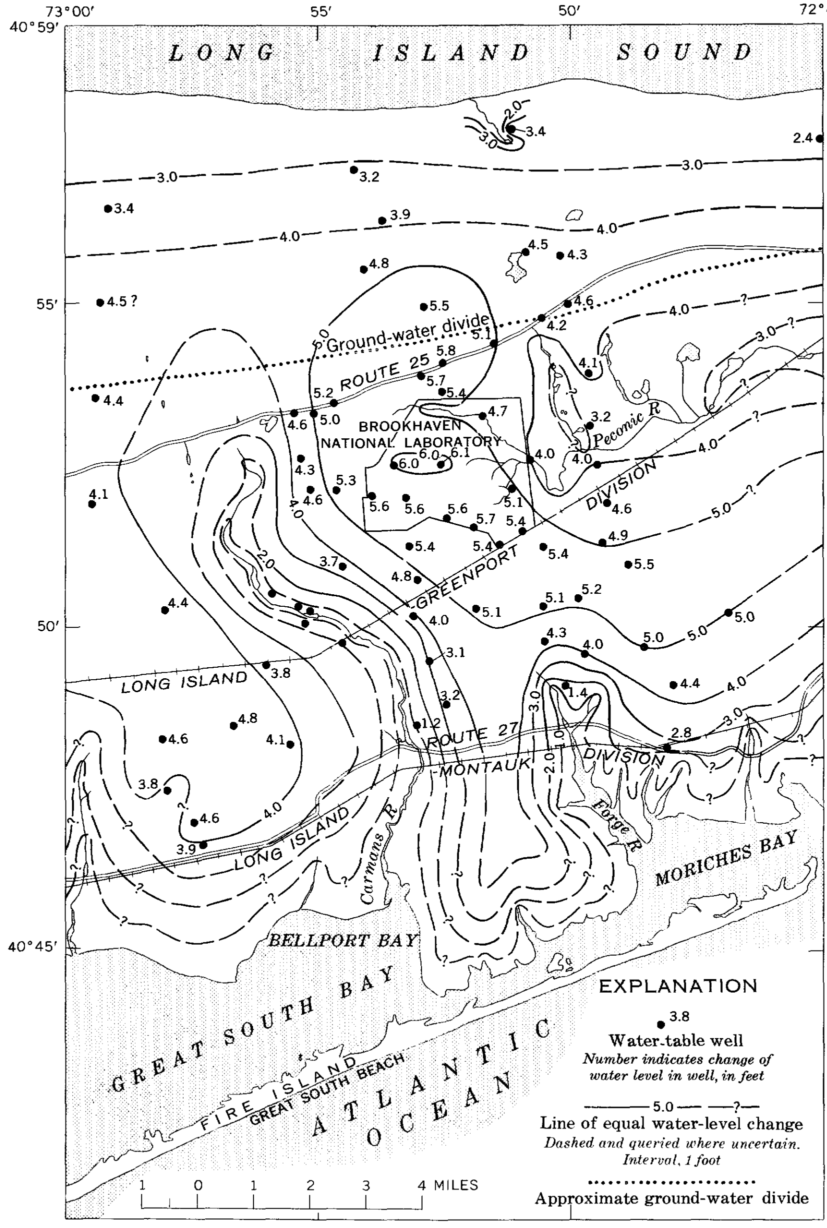

Figure 11.—Range in water table between highest levels observed during the spring and summer 1949 and the lowest levels observed during the winter of 1950–51, Upton area.

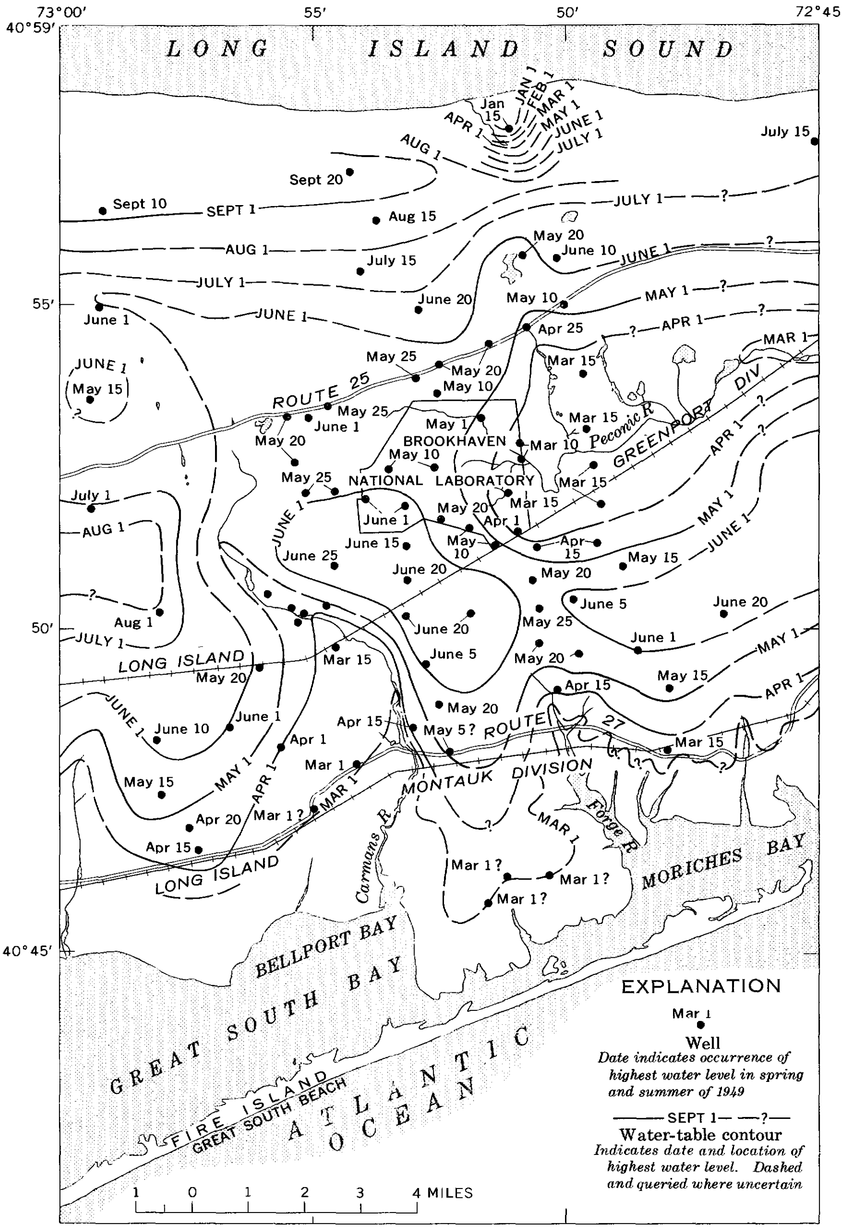

FIGURE 12.—Dates of highest ground-water levels during the spring and summer of 1949, Upton area.

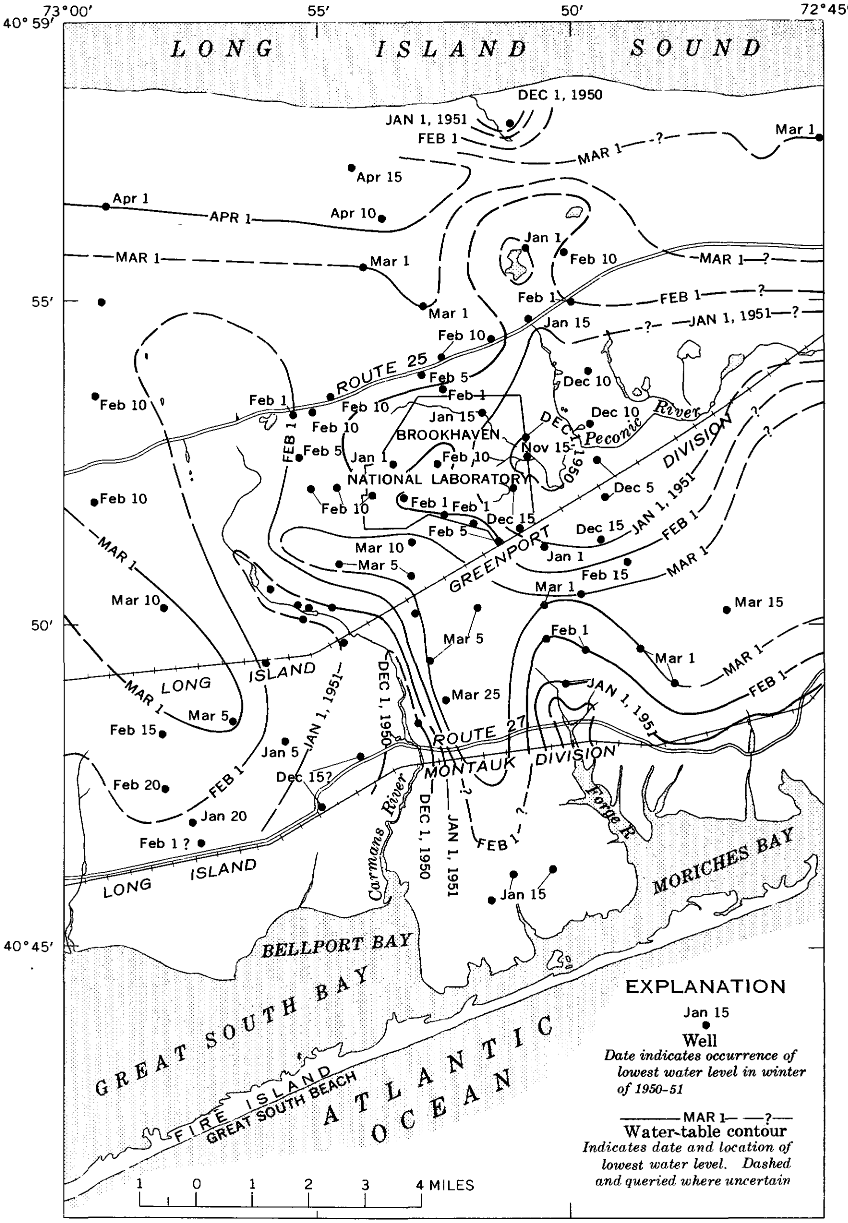

Figure 13.—Dates of lowest ground-water levels during the winter of 1950–51, Upton area.

Some evidence suggests that the long-time extreme range of fluctuation of the water table in the Upton area may be about twice that shown in figure 11. For example, Jacob (1945b, fig. 3) showed that the extreme range (from 1826 to 1941) in a 25-year average of effective rates of precipitation was about twice that in the period from 1912 to 1941. The range in water levels during the period 1941–53 in the Upton area is known to have been about the same as that from 1912 to 1941. Accordingly, in the years before 1912, the extreme range in water levels in the Upton area may possibly have been as much as twice that in 1941–53, as discussed by Lusczynski (written commun. 1953).

The lines showing dates of occurrence of the highest and lowest ground-water levels from 1941 to 1953 are shown in figures 12 and 13. These figures illustrate in a general way the effects of height of the water table above sea level and of distance to point of discharge on the times of maximum and minimum stage of the water table. These times come earlier where these factors are all small, as along the coast, and are as much as several months later in the interior of the island, where, in general, all these factors are greater. The relative importance of the individual factors appears to vary with the local conditions in some complex manner.

The time of the highest levels in 1949 in different parts of the Upton area ranged from about March to September, and of the lowest levels ranged from about December 1950 to April 1951. The lowest levels occurred from 1 year and 7 months to 1 year and 10 months later than the highest levels; the average lag was about 1 year and 9 months. The time interval between the 1950–51 low and 1953 high is a little more than 2 years. Although the period of detailed record is short, these relatively brief time initervals between the highest and lowest observed water levels show that the water table recedes and recovers rapidly in the Upton area.

QUANTITATIVE SIGNIFICANCE OF WATER-TABLE FLUCTUATIONS

Figure 11 shows lines of equal change of altitude of the water table from spring and summer of 1949 to the winter of 1950–51. The table below gives, for the Upton area south of the main branch of the ground-water divide, the approximate areas between lines of equal range, and the approximate average range for each interval. The approximate volume of sediments dewatered by the decline of the water table, from high level in the spring and summer of 1949 to the low level in the winter of 1950–51, is also estimated.

Selected range (ft)	Area (sq mi)	Approximate average range (ft)	Volume of de-watered sedi-ments (sq mi-ft)
6.0–6.2	0. 3	6. 1	1. 8
5.0–6.0	21. 3	5. 4	115. 0
4.0–5.0	41. 4	4. 4	182. 1
3.0–4.0	29. 0	3. 5	101. 5
2.0–3.0	17. 0	2. 5	42. 5
1.0–2.0	12. 7	1. 5	19. 0
0.0–1.0	13. 0	0. 7	9. 1
Total	134. 7		471. 0

Dividing the volume of sediments dewatered (471.0 square mile-feet) by the area affected (134.7 square miles) gives the average change in water level as 3.5 feet. The specific yield of these sediments is 0.24, so that the change is equivalent to a layer of water 10 inches thick. Because the highest stages of the water table do not occur everywhere simultaneously and because the same is also true of the lowest stages, it is apparent that the quantity dewatered from the zone of saturation must be a little less than 10 inches, perhaps about 8–9 inches. The change in water-table stage, equivalent to a depth of water of 8–9 inches, took place during an average period of 21 months. The corresponding change in storage during 12 months of this period is about four-sevenths of 8–9 inches, or about 5 inches. Thus, during a 21-month interval, the recharge to the ground-water system was 8–9 inches less than discharge, and during a 12-month period it was about 5 inches less.

MOVEMENT OF GROUND WATER IN ZONE OF AERATION

PURPOSE OF STUDY

Knowledge of how, when, and at what rates ground water moves through the zone of aeration, under both normal and extreme hydrologic conditions, may shed some light on the possible underground movement of contaminated water. Such water, if spilled on the land surface, would in all probability percolate downward through the zone of aeration to the zone of saturation, and then move vertically and laterally within the zone of saturation below the water table. It would bypass the zone of aeration only if it reached the zone of saturation through wells or pits deep enough to intersect the water table.

The manner and rate of movement of natural ground water through the zone of aeration is an indication of the probable manner and rate of movement of contaminated water, except that there may be differences in the physical, chemical, and biological properties of the two waters. Conclusions from the study of movement of natural ground water must be applied with care.

GENERAL DISCUSSION

Ground water moves in the zone of aeration as the result of the interaction of the force of gravity and the forces of molecular attraction which exist between the formation and the liquid. Under natural conditions, one or more virtually separate bodies of water may exist in the zone of aeration, and their movement is commonly intermittent and complex. Basically, the movement depends on new water entering the zone of aeration and mixing with, displacing, or bypassing the water already there. Variations in rates of recharge are a common reason, but not the only one, for the irregularity of this movement.

In general, the time it would take a drop of water to move from the surface to the water table, if the flow were steady, is less than that which is required under the common conditions of variable and interrupted flow. Steady movement of ground water in the zone of aeration can be expressed by Darcy's equation as adapted to unsaturated flow:

$$V=\frac{P'\,I}{7.48p}$$

where V is the vertical velocity in feet per day, P' is the effective vertical permeability in Meinzer units, and changes with changing saturation of the medium, I is the hydraulic gradient, and p is the porosity.

An attempt is made below to determine approximate values for these parameters for the upper Pleistocene outwash sands, which form the zone of aeration in the Brookhaven area, and then to calculate the probable range of velocities for water moving down vertically through this zone. Only order-of-magnitude answers are possible.

Veatch (Veatch and others, 1906) made many laboratory determinations of the porosity of sand samples from the upper Pliestocene of Long Island, and an average of his values was used. The vertical permeability is estimated from the results of a pumping test and an infiltration test. The percentage of liquid saturation of the sand under natural field conditions may be estimated from mechanical analyses of the sand. The results of the pumping test, and of tests on a lysimeter installed in 1953 in the Laboratory area, give values for the specific yield of the formation and for the specific retention. The greatest natural variation, and consequently the greatest uncertainty, is in the percentage of saturation of the medium under field conditions. Because the effective permeability is very sensitive to even small changes in percentage saturation, of any estimated value for the velocity of movement of water downward through the zone of aeration must be highly uncertain, and the actual velocity downward must vary between wide limits.

The permeability, P', ranges from zero, when the medium is less than about 20 percent saturated, to the maximum value, when the medium is fully saturated. The experimental studies by Wyckoff and Botset (1936) on the relation between permeability and liquid saturation of the medium are summarized in figure 14. These experimenters studied the simultaneous flow of water and carbon dioxide through sands packed in a tube 2 inches in diameter and 10 feet long and so constructed that the pressure and degree of liquid saturation could be measured at 1-foot intervals along its length. The k_l curve shows the relation between liquid saturation and liquid permeability, and the k_g curve the relation between liquid saturation and gas permeability, for all degrees of liquid saturation, under steady-state flow. The specific liquid permeability, as defined by Wyckoff and Botset, is the ratio of the permeability for liquid at any degree of liquid saturation to that for 100-percent liquid saturation. The specific gas permeability is

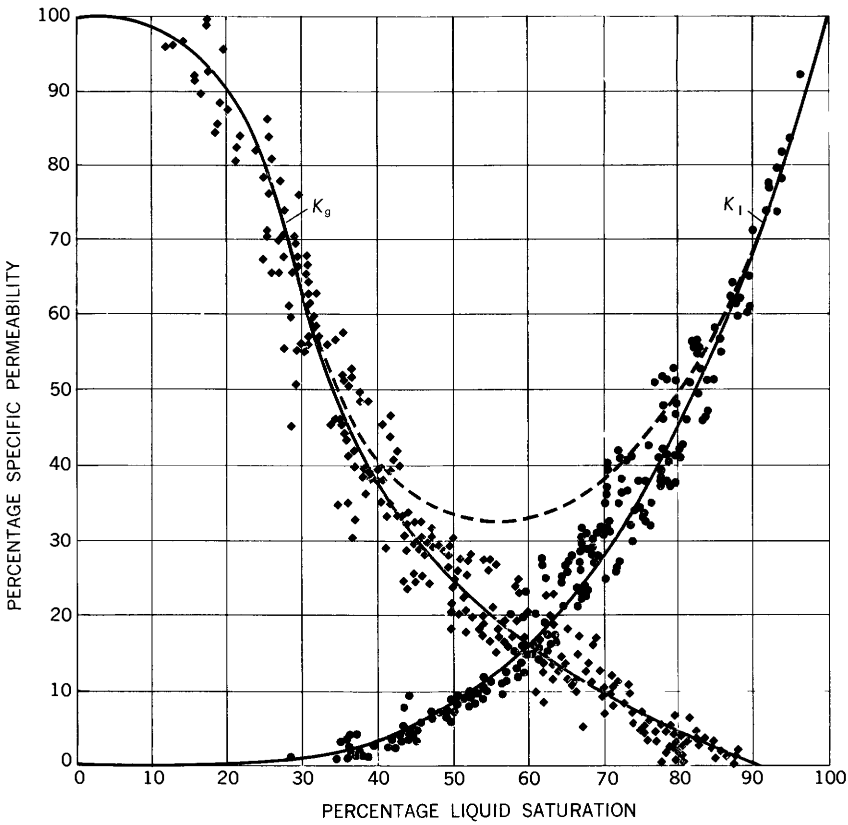

Figure 14.—Relations of permeability of a sand to water and to gas, with changes in liquid saturation of the sand (After Wyckoff and Botset, 1936.)

similarly the ratio of the permeability for gas at any degree of liquid saturation to the permeability of the sand, when dry, for gas. The four unconsolidated sands tested were (1) a sand having a grain size ranging from 0.21 to 0.124 mm (70–120 mesh) and a permeability of 17.8 darcys, (2) a heterogeneous sand with a permeability of 44.3 darcys, (3) an Ottawa sand with a grain size ranging from 0.28 to 0.19 mm (40–50 mesh) and a permeability of 44.3 darcys, and (4) an Ottawa sand having a grain size ranging from 0.56 to 0.42 mm and a permeability of 262 darcys. The curve shown in figure 14 is an average of values obtained from tests on all four of these sands. There is some scatter to the points defining the average curve, and if one were drawn for a fine sand alone it would fall below and to the right of the k_l curve for coarse sand, but the difference would be small.

A permeability of 1 darcy is equal to a permeability of 20.5 Meinzer's units. Meinzer (1923) defined the coefficient of permeability as the rate of flow of water at 60° F., in gallons per day, through a cross section of 1 square foot, under a hydraulic gradient of 100 percent; conditions of laminar flow and 100-percent saturation of pore space are assumed in this definition. Therefore, the permeability of the four sands in Meinzer's units, the unit used by the U.S. Geological Survey, ranges from 365 to 5,370 gpd (gallons per day) per square foot. The upper Pleistocene outwash sand at Brookhaven has a horizontal permeability of 1,400 gpd per square foot.

According to the curves in figure 14, when the pore space of the sand is half filled with water and half with gas, the specific liquid permeability is about 9 percent of the saturated liquid permeability, and the specific gas permeability is about 25 percent of the gas permeability of the dry sand. At 30 percent liquid saturation, the permeability to water is about 1 percent of the saturated permeability and the permeability to the gas about 65 percent of the dry permeability.

Some idea of the average degree of saturation in the zone of aeration can be obtained by a comparison of mechanical analyses of well samples of these sands with the curve in figure 15, after Terzaghi (1940). This curve expresses the relation between the effective grain size (D_{10}) and the average degree of saturation commonly found in a humid climate in the zone between the soil zone and the capillary fringe of the water table. The shaded area represents the probable range of seasonal variations. The term "effective grain size" (D_{10}) was first used by Hazen (1893) and by definition is chosen so that 10 percent by weight of the material is of smaller grains and 90 percent is of larger grains. Hazen's experimental work indicated that effective size was a good working index of the permeability of a filter sand, because he found that the permeability of an unsorted sand was roughly equal

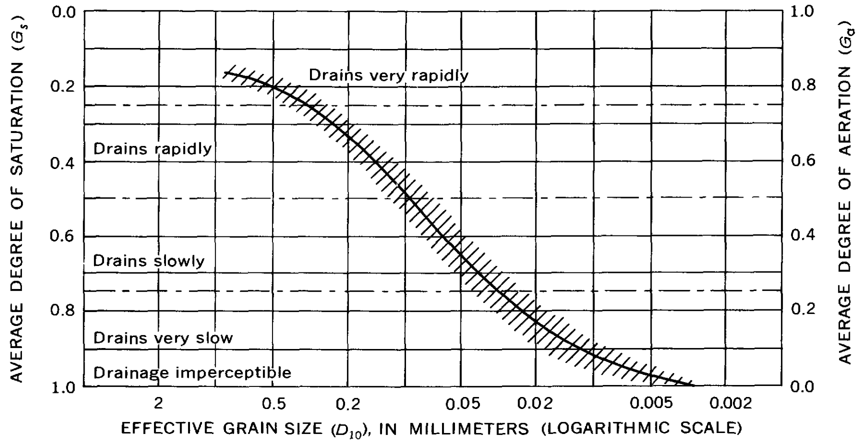

FIGURE 15.—Relation of effective grain size to average degree of liquid saturation in pores of unconsolidated formations (from field observations after Terzaghi, 1949). Diagonal lines represent probable range of seasonal variations.

to that of a sand composed entirely of grains of the effective size. The uniformity coefficient, also defined by Hazen, is the ratio of D_{60}/D_{10}, or the ratio of that grain size chosen so that 60 percent of the sample by weight is of a smaller grain size, to the effective size.

The effective size of nine samples from the upper 135 feet of well S6456 (table 6) near the center of the Laboratory area averaged 0.134 mm; the uniformity coefficient was 4.7. Samples from three wells, S6456, S6458, and S4660, selected by visual inspection as typical glacial outwash sand, were somewhat coarser grained, having effective sizes of 0.25, 0.17, and 0.30 mm and uniformity coefficients of 2.0, 2.4, and 1.8. Figure 15 shows that for a sand having an effective size of 0.20 mm, the percentage of liquid saturation ranges seasonally from 0.28 to 0.38.

TABLE 6.—*Effective size and uniformity coefficient of samples of sand, silt, and clay from well S6456*

Depth, in feet below land surface	Type of sample	Effective size, millimeter	Uniformity coefficient 60 percent size to 10 percent size
0–10	Auger	0. 23	2. 3
10–20	Core	. 35	15. 4
20–30	Bailer	. 16	2. 5
30–40	Bailer	. 18	2. 9
40–50	Bailer	. 088	4. 3
83	Bailer	. 096	4. 9
104	Bailer	. 15	3. 0
118	Bailer	. 085	5. 3
134	Bailer	. 19	2. 0
159	Bailer	. 14	2. 3
168	Bailer	. 20	2. 0
177	Bailer	. 092	2. 0
215	Core	. 13	3. 2

Such values appear reasonable for the glacial outwash sand in the Upton area. Both the porosity and the degree of liquid saturation of the glacial sand in the Upton area vary between wide limits under natural hydrologic conditions. Locally, under certain artificial conditions, the percent saturation has approached 100.

Veatch (Veatch and others, 1906) made many laboratory determinations of the porosity of the upper Pleistocene of Long Island, and the approximate average of these, 0.33, is used here. Specific yield and specific retention were determined from field tests; no attempt was made to determine these values in the laboratory from samples. The specific yield of the outwash sand in the Laboratory area was determined, from a 7-day pumping test, to be 0.24. The specific yield, found by filling and draining the pore space in a lysimeter built by de Laguna in 1953, was 0.26. This lysimeter, installed in the southeastern part of the Laboratory area where the average depth to the water table is 13 feet from land surface, is a vertical metal cylinder 12 feet deep and 5 feet in diameter and open at the top. It was set about 7 feet below land surface so that the bottom was 6 feet in the zone of saturation. In excavating and backfilling, care was taken to keep the material in approximately its original sequence and to compact it as nearly as possible to its original degree of compaction. However, the value of 0.24 from the pumping test is preferred because a much larger volume of sediments was involved.

A porosity of 0.33 and a specific yield of 0.24 gives a specific retention of $0.33-0.24$, or 0.09. On the assumption that 0.28, the low value in the range of liquid saturation in figure 15, is approximately the fraction of the void space filled by specific retention, then specific retention is computed to be 0.28×0.33, or 0.092, which is in good agreement.

The flow-line pattern (fig. 19) in the vicinity of the well pumped during an aquifer test in December 1950 in the Laboratory area suggests that the vertical permeability of the outwash sand in the zone of saturation is about a fourth that of the horizontal permeability, or about 350 gpd per square foot. Results of an infiltration test, discussed in the following section, indicate that the vertical permeability may be as low as 75 gpd per square foot, or about one-eighteenth of the horizontal permeability.

RATE OF MOVEMENT IN THE LABORATORY AREA

High rates

If the sand is saturated with water, if the vertical permeability is 350 gpd per square foot, and if the porosity is one-third, then water will move downward in the zone of aeration at a rate of 140 feet a day.

At this velocity, ground water would travel 45 feet from the land surface to the water table in about 8 hours.

On two occasions during pumping tests at the Laboratory, water was run out on the ground surface in such quantities that it may have saturated the sand in the zone of aeration as it made its way to the water table. In December 1950, during a 7-day test at well S3197 (Laboratory supply well 2), the discharge of 462 gpm (gallons per minute) infiltrated into the ground in a small area about 1,460 feet from the pumped well.

The approximate area of the land surface flooded was about 0.1 acre. The average discharge of 462 gpm is equivalent to about 2 acre-feet per day, so that about 20 cubic feet or 150 gpd of water infiltrated into every square foot of the 0.1-acre area. If the porosity is one-third, then the velocity must have been 60 feet per day, and the time required to reach the water, at a depth of 30 feet, was 12 hours. The vertical permeability in this case would then have been about 150 gpd per square foot. However, these calculations are based on a zone of infiltration only 5 feet wide, which was true at the surface, but the wetted zone probably increased in area with depth and may well have been much larger at the water table than at the land surface. If the effective average infiltration area was 0.2 acre, then the average vertical permeability was about 75 gpd per square foot, and the average time of travel of water from land surface, 30 feet to the water table, was about 1 day.

On June 1, 1949, an artesian well, S6434, screened in the Lloyd Sand Member of the Raritan Formation between 1,312 and 1,392 feet below land surface, was pumped for 24 hours at an average rate of 465 gpm. The water was discharged into a shallow ditch which first runs about 220 feet to the north and then swings 120 feet to the northwest to the place where most of the flow sank into the ground. The water table here is about 38 feet below land surface. An observation well, S6431, located 130 feet west of the pumped well and about 250 feet southwest of the infiltration area. It is screened between 120 and 125 feet below the land surface and about 90 feet below the water table. Water-level changes in this well are plotted continuously by a float-operated recorder until about 4 hours after pumping had stopped. Within about 2 hours after pumping had started the water level in this well started to rise, and it had risen 0.11 foot after 20 hours and 0.13 foot after 28 hours (4 hours after pumping had ceased). The interpretation of this record is uncertain because the loading of the surface by the weight of water pumped from depth, the movement and transmission of pressure by air trapped between the water table and the infiltrating water, and possibly even deformation of the thousand feet of overlying sediment by the removal of water in depth, can all produce minute but measurable changes in the water table near a deep artesian well when it is pumped. It appeared, however, that water began to arrive in large amounts at the water table after about 20 hours; a rate of 38 feet in 20 hours is equivalent to 45 feet in a day.

Low rates

From June to October 1949, rainfall was low, no recharge passed through the soil zone, and water levels steadily declined. Substantial recharge began with the rains in October and early November and by February 15, 1950, amounted to nearly 10 inches. However, the water levels in observation wells in the central and southwestern part of the Laboratory area, where the depth to water averages 45 feet, continued to decline even after these rains had started and did not begin to recover until about February 15, 1950. The sand between the land surface and the water table had drained for 5 months and about 10 inches of recharge during the subsequent 4-month period was required before the downward trend of the water levels in the zone of saturation was reversed.

Similarly, starting in the fall of 1950, water levels in these same wells declined until about Februrary 1, 1951, and then recovered for the next 3 months at an average rate of 0.8 foot per month. From November 1, 1950, until about February 1, 1951, about 10½ inches of recharge moved downward below the soil zone. Again in the fall of 1951, water levels in this area declined at an average rate of 0.43 foot per month until the middle of December, when they began to rise at an average rate of about 0.7 foot per month.

These three examples suggest that when the 45-foot column of material representing the zone of aeration is allowed to drain from 4 to 6 months, and then water is added at the top at an average rate of 1 inch per week, about 10 weeks is needed before there is a marked increase in discharge at the lower end of the column. The apparent time required for a drop of water to travel the length of the column under these conditions is about 10 weeks, and the approximate rate of travel is not more than about 5 feet per week. The rate, of course, varies with the degree of saturation in the column, and if the rate of accretion were increased to 2 inches per week, less time would be required. For example, if we accept a value of vertical permeability of 75 gpd per square foot under saturated conditions, figure 14 shows that the vertical permeability will be about 2 gpd per square foot if the sand is one-third saturated. This condition might prevail after several weeks of drainage. The Darcy equation then shows that the downward rate of movement would be only about 1 foot a day.

These rates are only apparent rates of movement, however, because it is not true that the water which falls as rain in November is the same water that reaches the water table in February. Even after several months of draining the sand still retains about 10 percent by volume of water, or in other words, its specific retention is about a third of its porosity. For a 45-foot-thick sand column this amounts to about 4.5 feet, or 54 inches of water, which is roughly 2½ times the estimated annual average recharge. The actual movement probably takes place in several ways. In a laboratory column packed with well-drained moist sand, water added at the top forces out or releases an equal volume at the bottom, but it is very largely a pistonlike displacement, with very little mixing. If this is true in the field, then, when the depth to the water table is 45 feet, no one drop completes the journey until about 54 inches of recharge have followed in behind it, which would require about 2½ years. However, under actual field conditions, recharge is very likely to be largely concentrated along certain channels and so most drops will make the journey much more quickly. The implications of this have been discussed at greater length by de Leguna (1966).

General conclusions

There is clearly a very wide range in the possible rates of downward travel of ground water in the zone of aeration under the various natural and artificial hydrologic conditions which have been observed in the Laboratory area. More specifically,

1. Rates may be as high as 140 feet per day under saturated conditions. This rate might apply after a sudden accidental spill of a large quantity of contaminated water, where several feet of liquid infiltrated per square foot of land surface.
2. Smaller leaks might result in rates of about 1–3 feet a day; this rate would be applicable where infiltration rates amounted to about an inch a week per square foot of horizontal surface.

MOVEMENT OF GROUND WATER IN THE ZONE OF SATURATION

DIRECTION OF MOVEMENT

Ground water in the zone of saturation accumulates and builds up the head in areas of recharge until, at equilibrium, the flow to the areas of discharge just balances infiltration. The direction of flow in an isotropic medium is always normal to the equipotential surface which expresses the head relationships in the hydraulic system.

Ideally, it would be possible to construct an accurate three-dimensional flow-line pattern for the Upton area, but not enough data are available on the artesian formations, which make up about 90 percent of the entire hydraulic system. Much can be learned, however, from the somewhat generalized representation of a two-dimensional flow-line pattern of a north-south vertical section in the Upton area shown in figure 4. In this diagram the vertical scale is exaggerated about 26 times. The flow lines show the basic pattern: (1) the flow, which is at right angles to the equipotentials, has a downward component in the central third of the Upton area, where the ground water moves into the Magothy(?) and Raritan Formations, and (2) the flow has an upward component in the southern (shoreline) third of the Upton area, where most of the water returns to the upper Pleistocene deposits from the deeper formations. In plan, the general direction of movement of ground water in the Upton area is northeastward in areas north of the ground-water divide, and southeastward and eastward in areas south of the ground-water divide (pl. 1). East of the Upton area the contour map of the water table shows the ground water moving eastward. (See fig. 34.)

The complex hydrology of the Magothy(?) Formation and the Lloyd Sand member is discussed in a later section. The concern here is with the lateral movement of ground water in the zone of saturation in the upper Pleistocene deposits.

Nearly all the water in the upper Pleistocene deposits moves in a lateral direction almost at right angles to the water-table contours (pls. 1–4), although actually the contours indicate the direction of movement of ground water only at small depths below the water table. In the deeper parts of the upper Pleistocene, well below the water table, the direction of flow may be somewhat different, depending on the amount of water moving vertically into and out of the formation and on the variations in permeability in the deposits. In general, the differences between the direction of flow in the upper and lower parts of the upper Pleistocene deposits are believed to be comparatively small. Probably the greatest variations are in areas where ground water discharges into the streams. The inferred direction of movement of ground water in and south of the Laboratory area is shown in figure 29. From the western 40 percent of the Laboratory area, most of the ground water moves underground toward the Carmans River, reaching it somewhere between the railroad crossing at Yaphank and Route 27. The ground water from a northwest-southeast central strip, which includes about 30 percent of the Laboratory area, moves first to the southeast and then to the south to discharge into the Forge River. Ground-water movement from the northeastern triangular section, which comprises the remaining 30 percent of the Laboratory area, is more complex. At times this ground water moves into the Peconic River, but at other times, depending on hydrologic conditions, the direction of underground flow changes and the water reaches the

Forge River. This is the result of perched water, as described in detail below.

RATE OF MOVEMENT

General discussion

The natural rate of movement of ground water in the area is a matter of obvious importance. The thick, simple, and relatively homogeneous aquifers of Long Island are well suited for direct measurement or indirect computation of ground water velocities. Several approaches to the problem are possible.

We have already seen that the average annual rate of recharge to the ground water is about 22 inches, that the porosity of the upper Pleistocene sand deposits is about one-third, and that the saturated part of this acquifer is about 150 feet thick. If the ground-water flow lines were parallel, and if no water makes its way into deeper acquifers, a 1-foot-wide vertical section of the aquifer, 1 mile from the water-table divide and normal to the direction of flow, would have passing through it each year the recharge from 5,280 square feet of surface area, or about 9,700 cubic feet of water. The saturated area of this cross section would be 150 square feet, but the porosity reduces the area through which the water moves to 50 square feet; thus the annual linear flow of the ground water would be 194 feet, and the rate of flow 0.53 foot per day. At the center of the Laboratory area, about 2½ miles from the water-table divide, the rate of flow would be about 1.3 feet per day. This figure is a maximum for the average rate, because part of the water, and as we shall see later a substantial part, does move through the deeper acquifers, and some shallow ground water is deflected into streams.

Tracer measurements

It is possible to measure the rate of movement of ground water directly by the means of tracers, but the method has serious limitations. Slichter (Veatch and others, 1906) in 1903, measured ground water velocities at 22 locations along a 6-mile section parallel to the shore in southern Nassau County and about 8–9 miles from the water-table divide. This results ranged from no measurable flow to 96 feet a day, although unusual circumstances were involved at these particular locations. At most of the locations the observed velocities ranged from about 2 to about 6 feet per day and averaged about 3–4. This velocity is in rough agreement with the suggestion that the maximum velocity in feet per day should be equal to half the distance of the point in miles from the water-table divide.

In 1948 de Laguna used tracers to measure the rate of movement of the shallow ground water at two points in the southeast corner of the Laboratory area. The depth to the water table here is about 6 feet. The first installation consisted of three wells, each about 13 feet deep, arranged in one straight line. The distance from the first well to the second was 5.37 feet and from the second to the third was 5.06 feet. The alinement was in the direction of ground water movement, determined in advance of the drilling of the second and third wells, by carefully comparing the water level in the first well with those in two other shallow wells which formed with the first a triangle about 200 feet on a side. This direction was S. 60°45′ E.

About 4 gallons of a tracer mixture, consisting of 8 l of water, 5 l of methanol, and 2 kg of ammonium chloride, was poured into the bottom of the first well through a tube, and a gallon of water was then poured into the well itself to force the tracer out into the formation. The proportions of the tracer mixture were so chosen that the solution had a density equal to that of the ground water, an important consideration because a solution of ammonium chloride lacking the alcohol is denser that the ground water and will sink through it. Figure 16

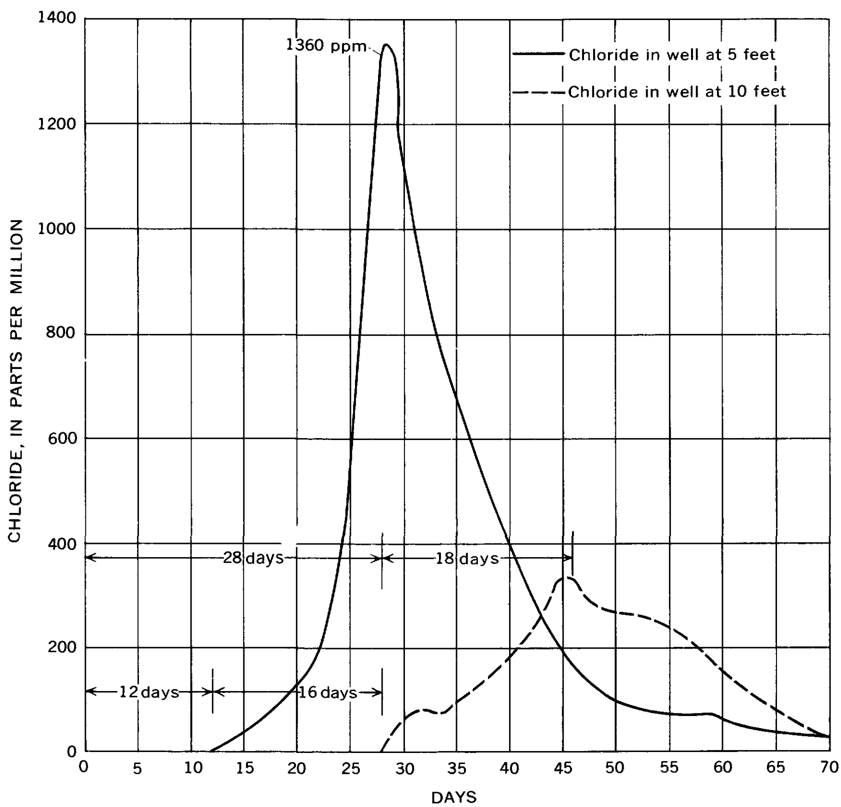

FIGURE 16.—Ground-water velocity, 7 feet below the water table, as measured by the arrival of ammonium chloride tracer at two wells, 5 feet and 10 feet downgradient from the "salted" well.

shows the chloride content of the two downstreams wells as a function of time, determined by the daily removal of small samples from each well. The rate of travel is best determined from the interval between the time of peak concentration in the second and third wells because the tracer may continue to feed slowly from the first or "salted" well, even though this is flushed. (See Jacob, 1938.) This factor, and also a tendency of the tracer to become trapped in stagnant water above the screen in the observation well, is probably responsible for the asymmetric time-concentration curve and makes average arrival time later than the peak. The same effect can result from adsorption and elution of the tracer (Kaufman and Orlob, 1956), but laboratory tests did not show much adsorption of chloride by Brookhaven upper Pleistocene sand. The travel time from the second well to the third well was 18 days and the distance 5.37 feet, or a rate of 0.3 foot per day.

These same wells were then driven to a depth of 33 feet and the test was repeated. No chloride was found in either of the sampled wells. About 25 feet way, six wells were each driven to a depth of 33 feet. These were arranged one at the center, to receive the tracer, and five on the arc of a circle roughly 8 feet in radius to intercept the tracer even if the direction of movement were different at this depth, as was suspected. Measurements in this test were made with an alternating-current resistance bridge, and readings were taken both on a cell lowered into the wells and between the wells themselves. The tracer hit the center well of the arc showing that the direction of movement was S. 20° E., a 40° difference from the direction 20 feet above.

Figure 17 shows the resistance between the "salted" or injection well (No. 0) and the five observation wells as a function of time. The resistance to all the wells dropped slightly when the tracer was introduced. About 6 days later the resistance to well 3 started to fall below that to the others, and after 11 days had reached a minimum. This minimum resistance represents the tracer distribution giving the maximum electrical interconnection between well 0 and well 3, a condition which would exist before the center of the tracer "cloud" reached well 3. The average time of travel is, therefore, something more than 11 days.

Figure 17 also shows the resistances between the several observation wells, taking adjacent wells in pairs. The resistances between wells 1 and 2 and between 4 and 5 did not change. However, as the tracer approached well 3, the resistance from this well to well 2 and to well 4 dropped, and the similarity of the changes suggests strongly that well 3 was in the center of the path of flow of the tracer. These drops began after 7 days and reached a minimum after about 15 days, although the shape of the curve does not permit an exact determination.

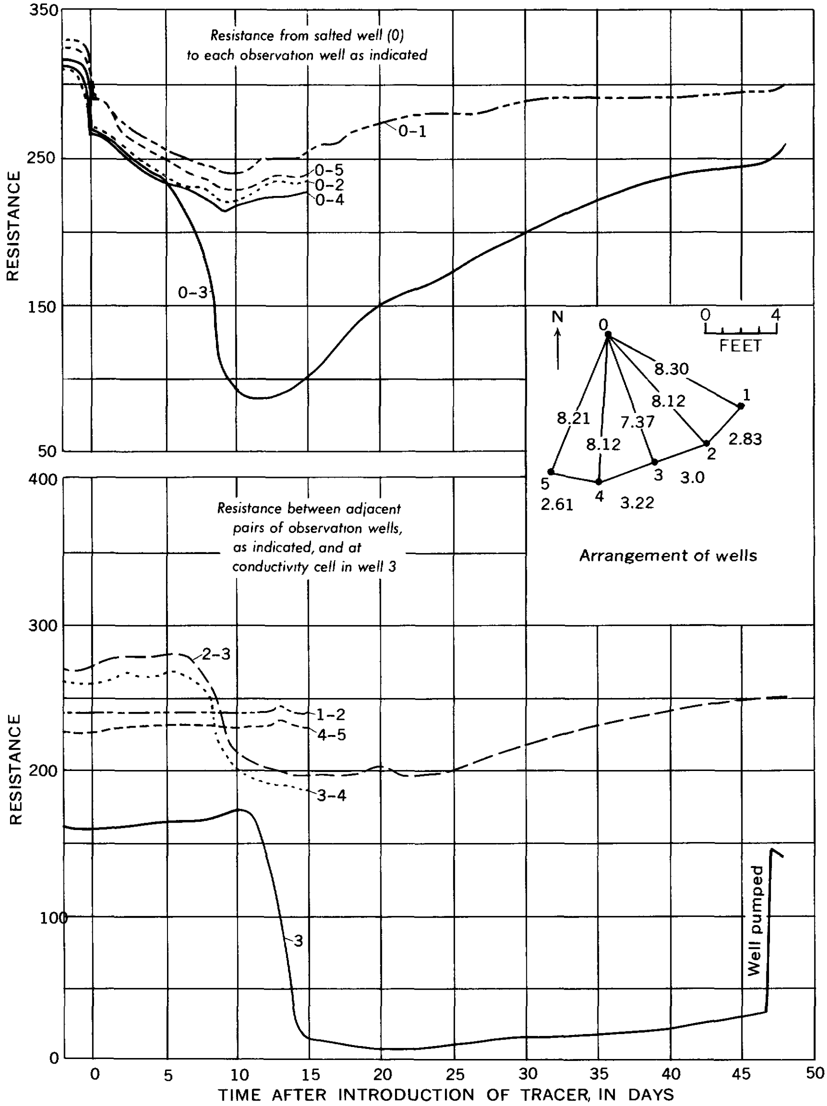

FIGURE 17.—Ground-water velocity measurements, 27 feet below the water table, as measured by arrival times of ammonium chloride tracer solutions. Locations of wells are 50 feet east of three shallow wells used in obtaining data shown in figure 16.

The resistance of the cell in well 3 started to drop after 10 days, and dropped rapidly until about 5 days later, after which little change was noted. Apparently this constant value was due to the tracer mixing with the stagnant water in the well above the screen, from where it would only be slowly rinsed after the tracer had passed by on the outside. After 47 days the resistance of the water in the well was still relatively low, although other measurements suggested that the tracer itself had gone beyond. Well 3 was then pumped briefly, after which the resistance of the water in it returned to its original value, substantiating the explanation. The apparent travel time was about 15 days for a distance of 8 feet, a rate of 0.53 foot per day.

The use of tracers to measure the direction and rate of ground-water flow appears to be simple and straightforward. It is not. There are many complications, due in part to fundamental problems from which the method suffers, such as differences in density, viscosity, and rate of adsorption between the tracer and natural water, and to incidental but almost unavoidable technical problems, such as the tracer mixing with stagnant water in the observation well. Tracer measurements give rate and direction of movement only over very short distances, and many measurements must be made if average values are required. These tracer measurements would, however, show something of the variations from the average, which the methods described below do not provide.

Several attempts to measure ground-water flow with tracers were made in other parts of the Laboratory area but, except for certain work in connection with a pumping test, without success.

The porosity of the upper Pleistocene sands, as determined by Slichter (in Veatch and others, 1906) is very close to 0.33. The thickness of the saturated zone was determined from the drilling records. The hydraulic gradient is derived from the water-table maps. The coefficient of transmissibility is the most difficult factor to measure. Laboratory determinations with unconsolidated materials are in general unsatisfactory because slight differences in packing and arrangement of the grains can make big differences in the observed rate of flow and because the collection of undisturbed samples is virtually impossible. The coefficient of transmissibility can be determined by test-pumping wells screened in the aquifer and from the flow of streams which derive their water from ground water. The application of these methods to the present problem is described below. The results are presented in table 7.

TABLE 7.—*Coefficients of transmissibility and storage from distance-drawdown plots for aquifer test in December 1950*

Time, in minutes since pumping started	Using wells at 30-, 100-, and 200-foot distances		Using wells at 100-, 200-, and 300-foot distances		Using average of two wells at 200-foot distance, and well at 350-foot distance	
	T gpd per ft	*S*	*T* gpd per ft	*S*	*T* gpd per ft	*S*
1,000	212,000	0.160	192,000	0.183	166,000	0.205
2,000	211,000	.183	190,000	.214	160,000	.264
5,000	214,000	.183	182,000	.263	162,000	.321
10,000	213,000	.166	188,000	.239	164,000	.333
Average	212,500		188,000		163,000	

AQUIFER PERFORMANCE TESTS

The theory and practice of aquifer performance tests, made by test-pumping a well screened in an aquifer, have been adequately described in several reports (Brown, 1953; Bruin and Hudson, 1955). The description below of the aquifer performance test (or pumping test as they are commonly called) run at Brookhaven in December of 1950 is intended to show to what extent the conditions of the test met the requirements for the use of the method and to record the observations, data, and calculations.

A pumping test is made on a pumped well, usually of large capacity, which should be under the experimenter's control. In the preliminary stage this well is shut down so that conditions in the aquifer to be tested may come to equilibrium. Changes in water level are measured carefully during this period, both at the test site and at one or more observation wells nearby but outside the area to be affected by the pumping. In this way, normal slow seasonal changes in water level may be determined and extrapolated through the test period, and corrections may be made in the observed data; with this information the net water level changes due solely to pumping can be used in the calculations.

The pump must discharge steadily at a uniform rate (unless the test is intentionally run at several pumping rates, which was not the case here) and accurate and frequent water-level observations must be made in several observation wells located at various distances from the pumped well. The water pumped during the test should be discharged at a distance such that its return to the test aquifer shall not interfere with the readings, or else it must be possible to make corrections for this factor as in the December 1950 test.

The December 1950 test used one of the three Laboratory supply wells, S3197, which has a nominal capacity of 500 gpm. The well was shut down for 6 days prior to the test, except for 36,000 gallons pumped only 3 days before the test started. Corrections for this antecedent pumping were made as necessary. Observations on many wells in the area showed a steady average rate of decline of the water table of 0.011 foot per day from August through December 1950; corrections for these normal seasonal changes could be made accurately. The Laboratory's other supply wells were far enough away so that no significant influence from them was felt in the test area. The hydrologic background against which the test was run was therefore exceptionally favorable.

The aquifer tested was the upper Pleistocene sand and gravel; in the test area it is about 190 feet thick, and about 145 feet is saturated. Below the sand and gravel is the nearly impermeable Gardiners

Clay, here 18 feet thick, and below this, in turn, the clayey sands of the Magothy Formation. The geologic section at the test site was explored by a 4-inch cable-tool test well, S6456, 10.9 feet east of the pumped well. S6456 was originally drilled through the Gardiners Clay and a short distance into the Magothy(?) to a total depth of 217 feet; it was then plugged back to 173 feet and slotted from 173 to 169 feet to serve as an observation well in the lower part of the aquifer during the pumping test. The log of this well and the results of a mechanical analysis of samples from the well are given below:

Log of well S6456

	Thickness (feet)	Depth (feet)
Upper Pleistocene:		
Sand, fine to coarse, white to tan and fine to coarse gravel. Some boulders from 0 to 10 ft. Thin streak of clay, gray-brown at 31, 41, and 83 ft. Erratics numerous throughout	145	145
Sand, fine to medium, clayey, gray. Wet samples have a greenish cast due to small amount of green clay in the formation	47	192
Gardiners Clay:		
Clay, silty, dark-greenish-gray. Bailer samples just beneath core contained clay lumps and medium to coarse gravel	4	196
Sand, white, medium to coarse, some gravel	4	200
Clay, dark-gray, brown at bottom, fossiliferous at bottom of clay	10	210
Magothy(?) Formation:		
Sand, medium to coarse, dark-gray, micaceous. Traces of fine gravel throughout	7	217

The small variations in lithology with depth in the upper Pleistocene deposits at well S6456 is suggested by the similarity of the effective sizes and uniformity coefficients listed in the following table.

Effective size and uniformity coefficient of sand, silt, and clay samples from well S6456

Depth, in feet below land surface	Type of sample	Effective size, millimeter	Uniformity coefficient 60 percent size to 10 percent size
-10	Auger	0. 23	2. 3
10–20	Core	. 35±	15. 4±
20–30	Bailer	. 16	2. 5
30–40	do	. 18	2. 9
40–50	do	. 088	4. 3
83	do	. 096	4. 9
104	do	. 15	3. 0
118	do	. 085	5. 3
134	do	. 19	2. 0
159	do	. 14	2. 3
168	do	. 20	2. 0
177	do	. 092	2. 0
215	Core	. 13	3. 2

OBSERVATION WELLS

Twelve principal observation wells (fig. 18) were used to measure changes in the elevation of the water table produced by the pumping; they are listed in table 8. All these wells were carefully developed and tested, and all had excellent hydrologic connection to the aquifer. For only one, S6456, were small corrections in the observed readings required because of a lag in response of the water level in the well to changes in the water table, and then only for the first 200 minutes of the test. The conditions of the test, therefore, provided excellent opportunity for obtaining the required water-level measurements, and ample, accurate data was obtained.

PROBLEM OF PARTIAL PENETRATION

The test conditions were not ideal because the screen of the pumped well did not fully penetrate the aquifer. The aquifer in the test area is about 145 feet thick between the water table and the nearly impermeable Gardiners Clay which underlies it, and the screen of well S3197,

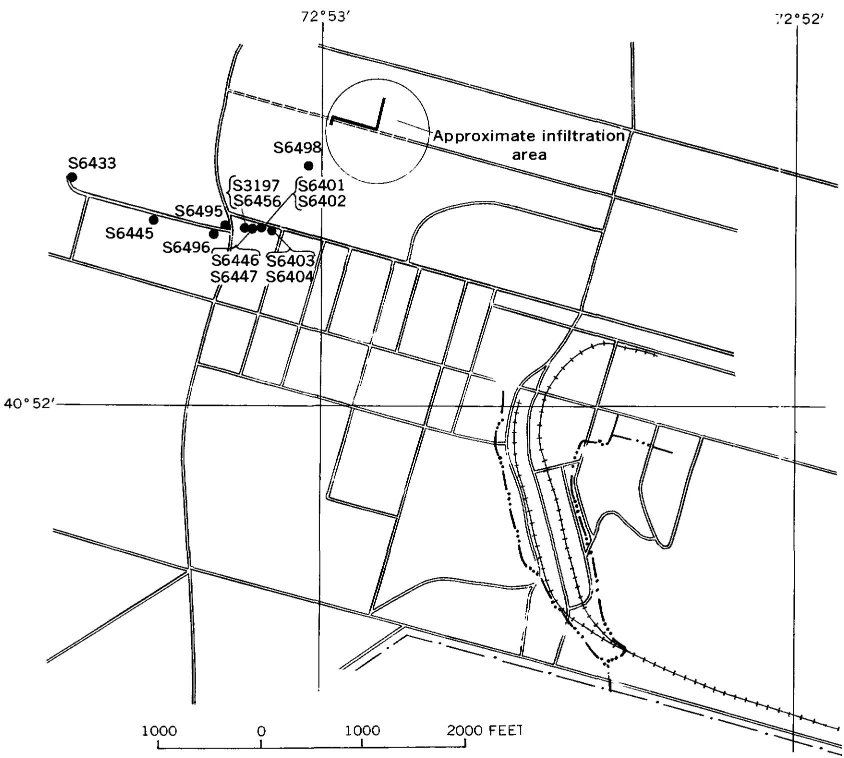

FIGURE 18.—Locations of observation wells in the vicinity of test well S3197, December 1950.

TABLE 8.—*Data for wells considered in connection with pumping test, December 1950*

[All measurements in feet, except diameter of well which is in inches. S(hallow) and D(eep) at same site; E(ast) and W(est) of pumped well. Static water-level readings are for December 11, 1950, except S6498 (Apr. 30, 1951), S6445 (Dec. 5, 1950), and S6443 (Nov. 24, 1950)]

Well	Distance from pumped well	Land surface (msl)	Diameter	Depth below land surface	Screen or perforations: Below land surface	Screen or perforations: msl	Measuring point (msl)	Static water level (msl)
S3197	0	89.65	12	135	115 to 135	−25 to −45	85.65	42.21
S6456	10.9E	90.2	4	173	169 to 173	−78 to −42	92.37	42.17
S6447	S 29.6E	90.8	2	62.8	59 to 64	+32 to +27	91.93	42.18
S6446	D 30.3E	90.8	2	122.7	119 to 124	−28 to −33	91.93	42.22
S6402	S 100.3E	91.9	2½	60.6	56 to 61	+36 to +31	92.34	42.18
S6401	D 100.9E	92.0	2½	115.8	111 to 116	−19 to −24	92.35	42.20
S6404	S 200.5E	96.6	2½	56.2	51 to 56	+45 to +40	97.05	42.02
S6403	D 200.6E	96.7	2½	103.5	99 to 104	−2 to −7	97.39	42.12
S6498	828E	97.0	2	65.0	Open 64	Open 32	97.70	44.35
S6495	200.1W	82.7	1¼	57.5	55 to 57.5	+27.7 to +25.2	82.80	42.30
S6496	349.8W	79.1	1¼	42.0	39.5 to 42	+39.6 to +37.1	79.62	42.35
S6445	902W	79.3	2½	78.8	74 to 79	+5.3 to +0.3	80.54	42.63
S6433	1750W	54.4	1¼	30.9	28 to 31	+26.4 to +23.4	56.56	39.76

the pumped well, is 20 feet long and is set very nearly in the middle of the saturated zone. This placement means that the aquifer may be divided by an imaginary horizontal plane into two equal and equivalent parts, each 75 feet thick and each furnishing water at half the measured pumping rate to a 10-foot section of screen at the very upper or lower surface of the half-aquifer. Jacob (1945a) has shown that the problem of partial penetration for this geometry may be solved approximately, if the aquifer is homogeneous and anisotropic, by the assumption that the drawdown that would be obtained by full penetration of the aquifer is the average of the drawdowns that are observed at the upper and lower surfaces of the aquifer. With the possibility of using this solution in mind, six of the observation wells were installed in pairs, at distances of 30, 100, and 200 feet from the pumping well, and to depths so that one of each pair was screened at the center of the aquifer, opposite the screen of the pumped well, and the other not far below the water table. It was not possible to achieve this geometry exactly, and figure 19 shows the actual pattern used and the changes in head in the aquifer believed to have been produced by the pumping, as determined from water-level observations in the observation wells. Figure 20 shows the difference in observed drawdown in wells at the same radial distance from the pumped well, but screened at different depths, as a function of time after the test pumping was stared. The method of resolving the data, to obtain values for the drawdowns which would have been observed with radial flow to a fully penetrating pumped well, is shown in figure 21. In this figure the observed drawdown for each well, after 10,000 minutes pumping, was plotted as the abscissa, and the depth of screen, referred to sea level, as the ordinate. Straight lines were drawn through the two wells of each pair, and from the intersections of these lines with the water table and the

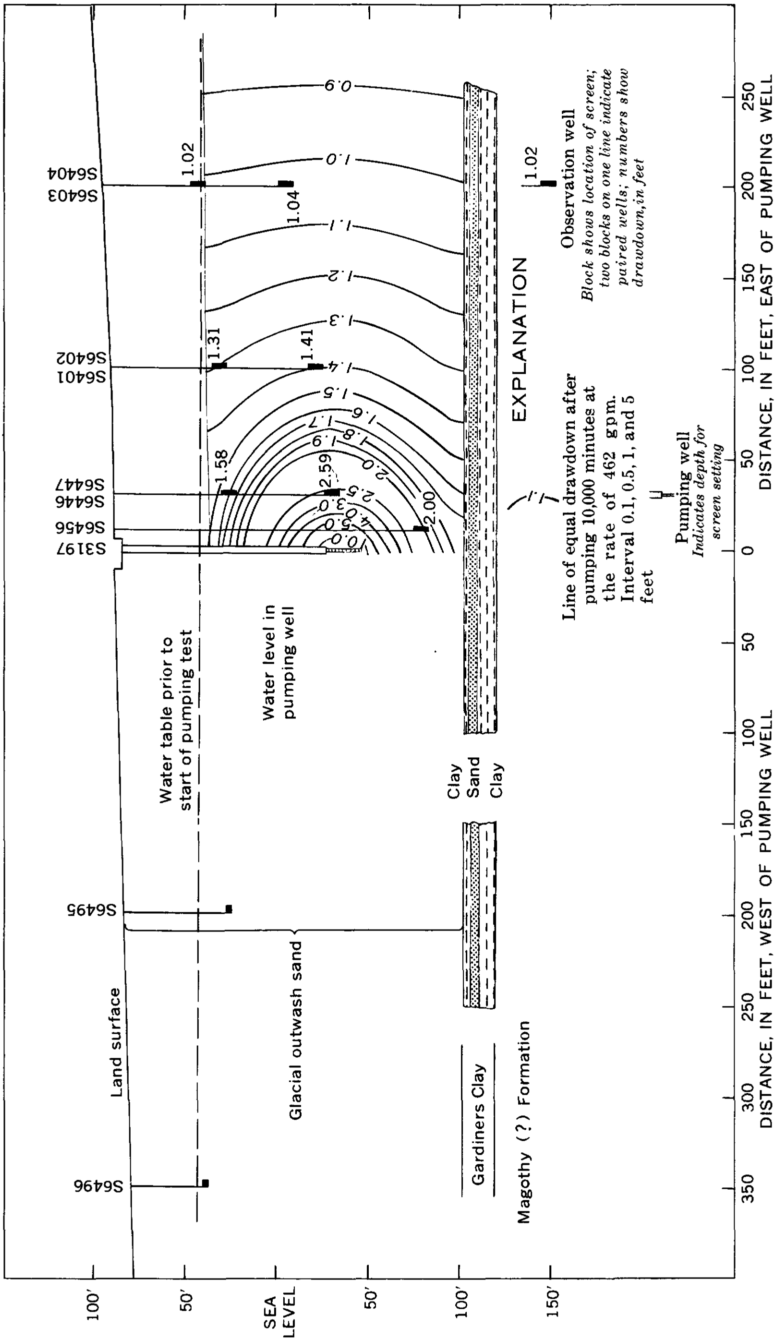

FIGURE 19.—Drawdown pattern east of pumped well S3197, December 1950.

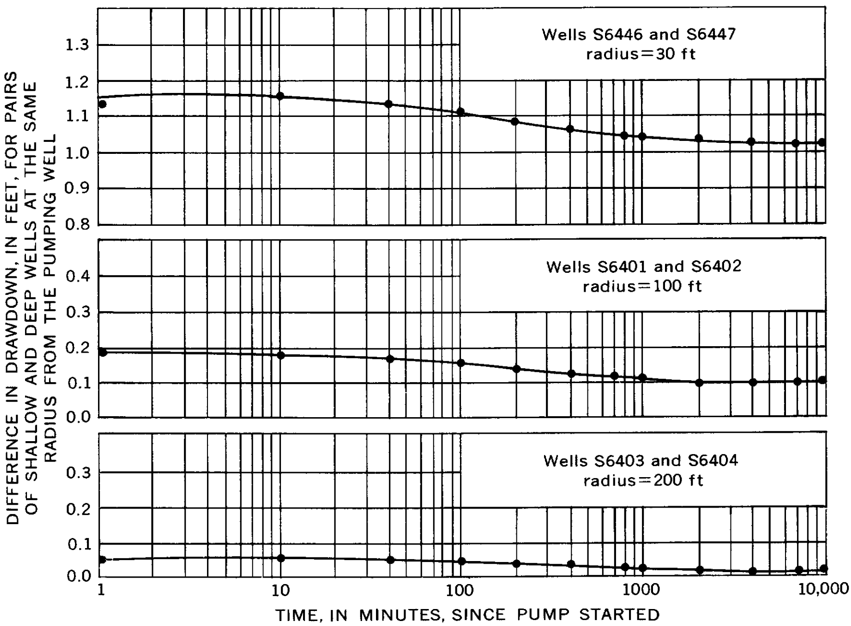

FIGURE 20.—Differences in drawdown at three pairs of observation wells, December 1950.

medial plane of the aquifer, the corrected drawdowns were computed, as shown in table 9.

TABLE 9.—*Corrections for effect of partial penetration of well S3197, test pumped December 1950*

[E(ast) and W(est) of pumped well]

Well	Distance from pumped well (feet)	Depth to center of perforations (feet)	Corrections from figure 20 (feet)
S6456	10. 9E	170	* * *
S6447	29. 6E	61	+0. 43
S6446	30. 3E	121	—. 59
S6402	100. 3E	59	+. 052
S6401	100. 9E	114	—. 046
S6404	200. 5E	55	+. 018
S6403	200. 6E	102	—. 005
S6495	200.1W	56	+. 01
S6496	349.8W	41	. 00

There is a further complication because of the difference between the vertical and horizontal permeability of the test aquifer. This difference could not be quantitatively evaluated because as yet there is no satisfactory workable formula for the problem of three-dimensional nonequilibrium flow in an anisotropic medium. Muskat (1949, p. 264) has described a semiquantitative graphic method for getting an ap-

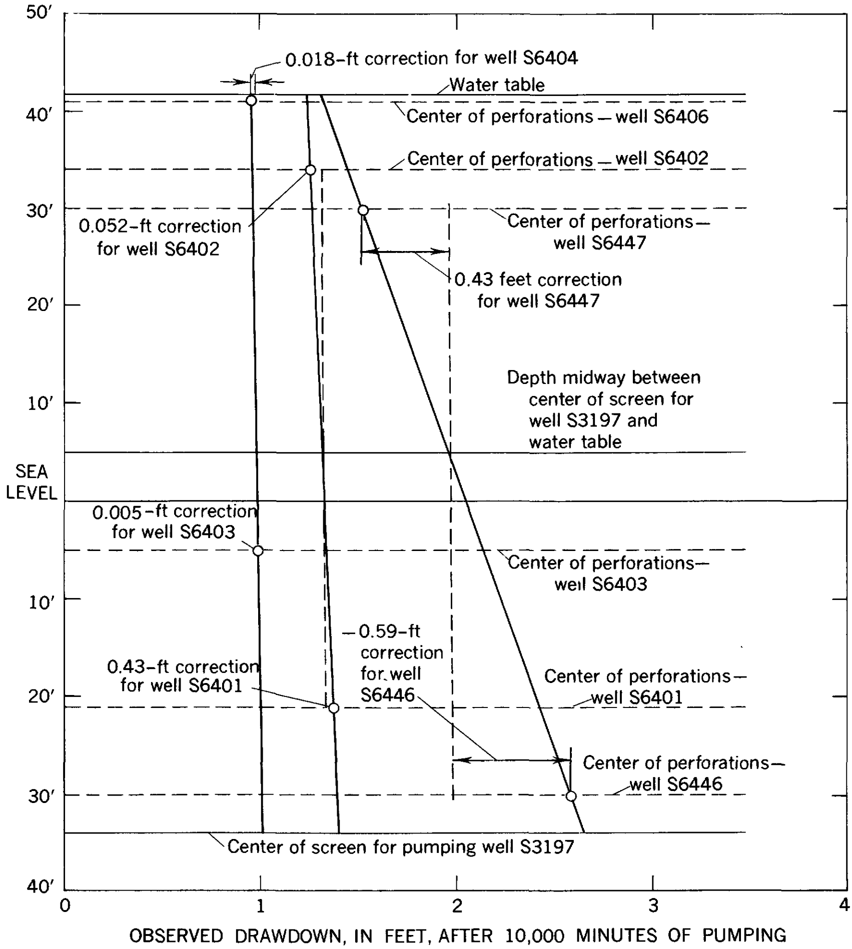

Figure 21.—Method of correction of drawdown for effect of partial penetration by pumped well S3197, December 1950.

proximate solution to the problem by trial and error, as illustrated in figure 22. Figure 22*A* shows the flow net as it would exist adjacent to a gallery with a height equal to the screen of the test well, and extending at right angles to the line of observation wells; the assumption is made that the vertical and horizontal permeabilities are equal. Figure 22*B* is plotted on the basis of the same assumptions, except that the horizontal permeability is taken to be four times the vertical. The transformation is made by doubling the horizontal dimensions of figure 22*A*, but keeping the vertical unchanged. Because the curvature and pattern of the equipotential and flow lines in figure 22*B* are somewhat similar to those in figure 19, particularly for the zone between

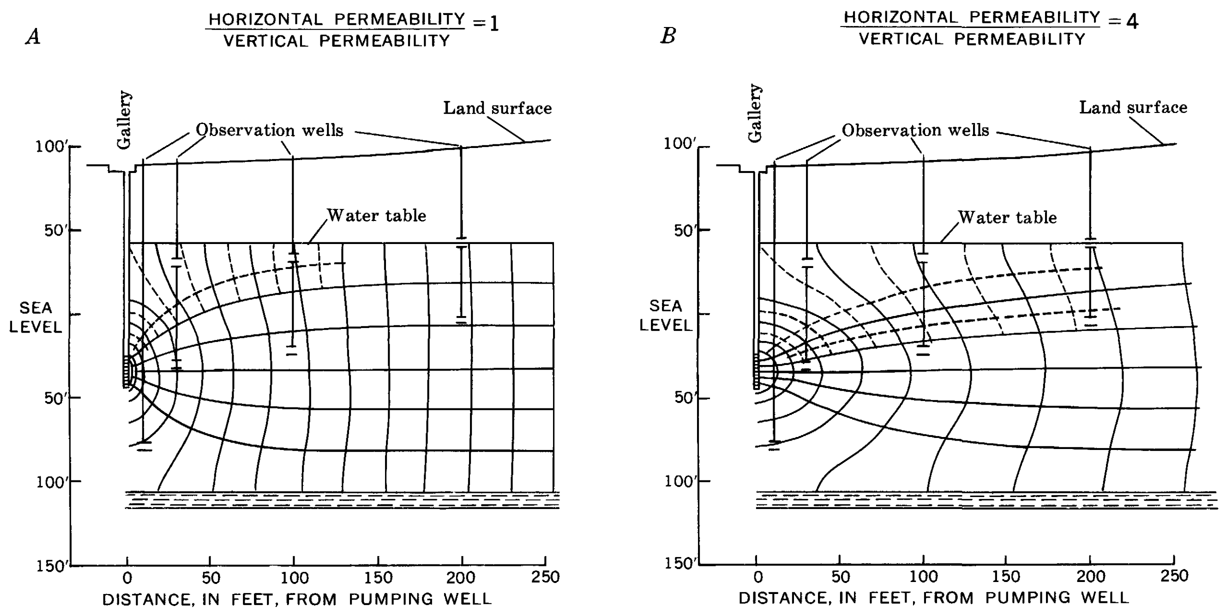

FIGURE 22.—**Flowlines for two-dimensional flow into a vertical gallery. *A*, In an isotropic formation; *B*, In an anisotropic formation.**

30 and 200 feet from the pumped well, where the best control is available, the inference is drawn that the horizontal permeability at well S3197 is approximately four times the vertical. The method and the data do not permit a more precise determination, and little reliance can be placed on the results, particularly because a study of infiltration rates suggests that the horizontal permeability may be as much as 9–18 times the vertical. Because of this uncertainty, no attempt was made to adjust the corrections for partial penetration to allow for differences between the vertical and horizontal permeability, and the corrections used, as shown in figure 21, are plotted on the assumption that the formation is isotropic. These corrections are, therefore, only approximate, although the discrepancy is quantitatively important only for the two wells, S6446 and S6447, located 30 feet from the pumped well. The observed drawdowns in these wells also depart from the ideal or theoretical because of the slow drainage of the sand, a factor for which it is also impossible to correct the observed data. It would, therefore, be pointless to attempt an exact analysis of the problems of partial penetration and of the differences between horizontal and vertical permeability because the best test results come from observation wells some 200–300 feet from the pumped well, and at these distances the ground-water movement is virtually a two-dimensional radial flow toward the pumped well.

OTHER FACTORS INFLUENCING THE RESULTS

The pumping test lowered the water table and thereby reduced the effective thickness of the aquifer during the period of the test. This deviation from the assumed conditions under which the pumping-test theory developed is inherent in tests of water-table aquifers. However, the maximum drawdown was only 10 feet at the pumped well, and was less than 3 feet at the observation wells 30 feet away, amounts which are negligible in comparison with the 146-foot saturated thickness of the aquifer, and determinations of transmissibility may be considered reliable. Direct determinations of the coefficient of storage are more difficult because of the slow drainage of the sand.

Extensive test drilling in the laboratory area shows that the aquifer is sufficiently uniform and extensive that there are no lateral boundaries for which corrections must be made, nor are there any streams or other sources of surface water which could serve as sources of recharge when the water table was lowered by the pumping.

Under normal conditions in the test area water moves slowly from the upper Pleistocene through the Gardiners Clay down into the Magothy(?) Formation; there is a head loss of about 0.5 foot across the Gardiners Clay. During the pumping test the head in the bottom of the upper Pleistocene aquifer immediately below the pumped well was reduced about 1.7 feet, and about 0.9 foot 250 feet away (fig. 19), so that during the test the direction of flow was reversed, and water moved from the Magothy(?) into the upper Pleistocene in the immediate vicinity of the pumped well. As shown by the log of well S6456, the Gardiners Clay in this area is about 18 feet thick. About 14 feet is solid clay, which would appear to offer a very considerable barrier to the movement of water. In December 1949, 1 year prior to the test described, when well S6456 was temporarily screened in the upper part of the Magothy(?) just below the Gardiners Clay, well S3197 was pumped for 14 hours at a rate of 410 gpm. During this pumping the water level in S6456 dropped 0.06 foot, but between a third and a half of this drop was due to a change in barometric pressure. This evidence supports the opinion that little water moves through the clay in the test area but does not permit an exact quantitative determination. This flow probably amounted at most to only a few gallons a minute, perhaps 1 percent of the pumping rate.

DETERMINATION OF FORMATION CONSTANTS

The corrected drawdowns for each of the observation wells used in the test were plotted against time. (figs. 23, 24). These plots were then compared with type curves developed from the Theis (1935) nonequilibrium formula, as described in the references on aquifer performance tests given above, and values for T, the coefficient of transmissibility and S, the storage coefficient, were calculated. The Theis formula and, therefore, the type curves derived from it are based on the assumption that the coefficients do not change during the course of the test, either with time or with distance from the pumped well. Unfortunately, the actual conditions depart from these assumptions, particularly for the coefficient of storage. In the early part of a test, near the pumped well, the changes in water level are relatively large and rapid, and the unwatered sand does not have an opportunity to drain completely. The observed drop in water level in observation wells near the pumped well will, therefore, during the early part of the test, be somewhat larger than they would be if the sand drained promptly, as the formula assumes. Conversely, during the latter part of the test when conditions near the pumping well have somewhat stabilized and the water level is falling slowly, the change in water-level readings in the same observation wells will be somewhat less than predicted by the formula because the water table will receive drainage that would have arrived earlier if the sand had all drained promptly. Farther out from the pumped well the water table is not only lowered less, it is also lowered more slowly, so that the sand has a better opportunity to drain. In any case, because the coefficient of storage is less

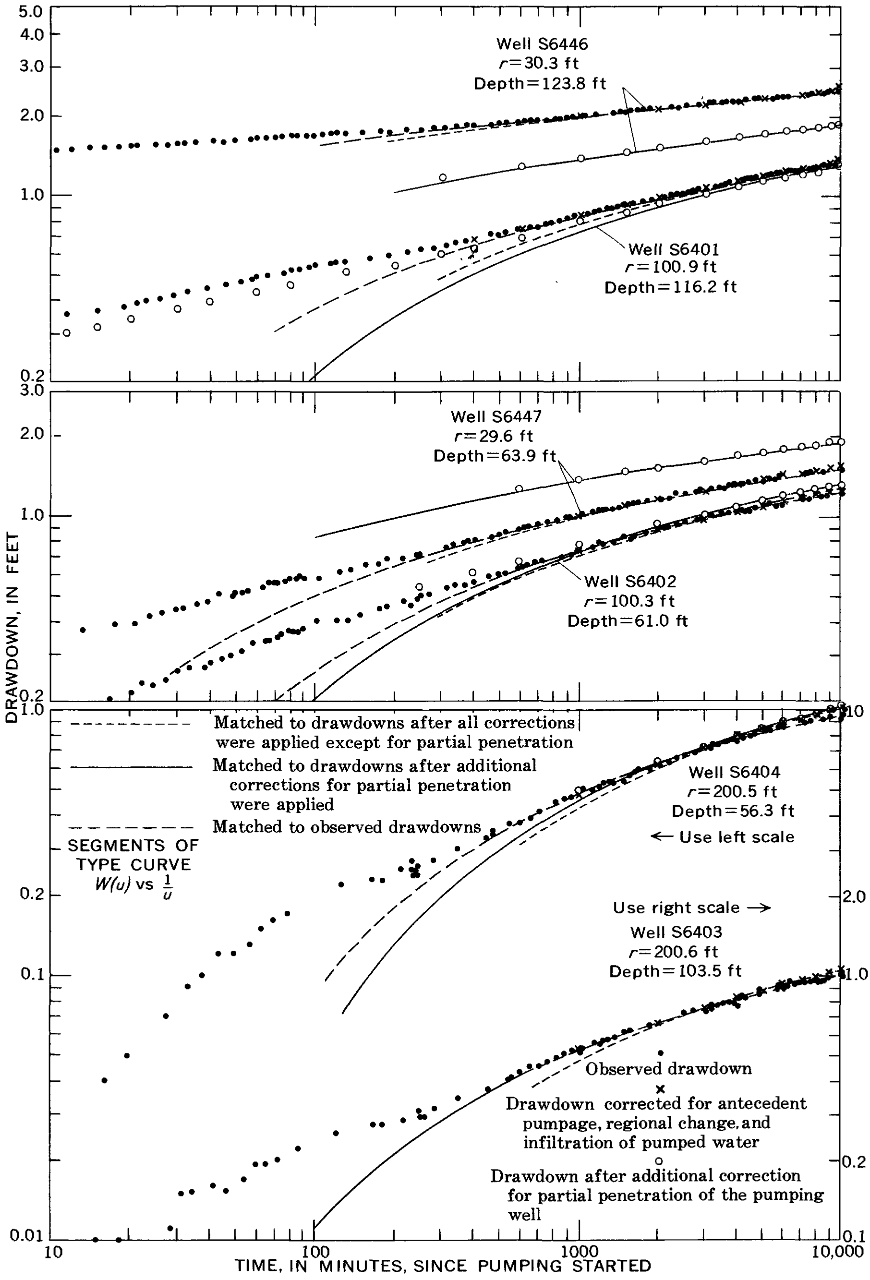

Figure 23.—Time-drawdown plots for six observations wells east of pumped well S3197, December 1950.

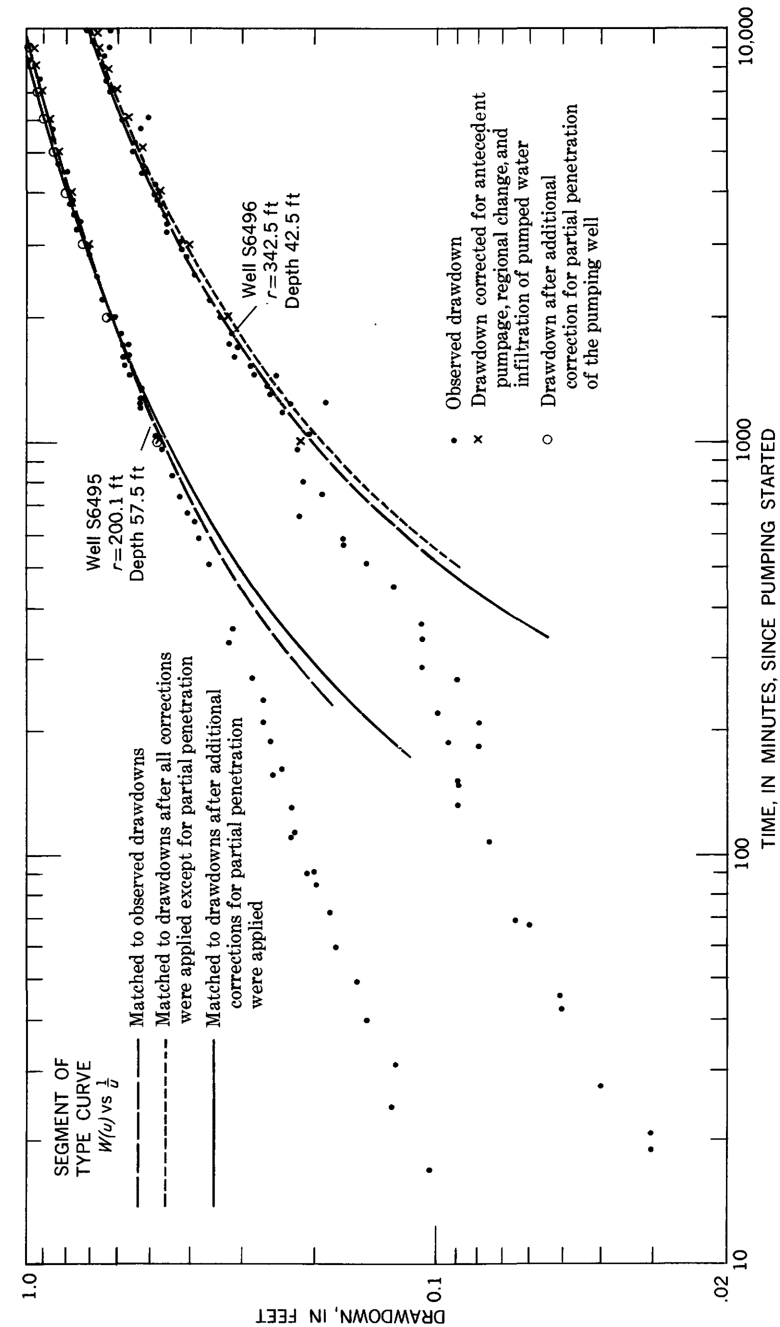

FIGURE 24.—Time-drawdown plots for two observation wells west of pumped well S3197, December 1950.

constant during the early part of the test, data obtained 3,000–10,000 minutes after the start of the test are believed to give the best values. This is well illustrated by the curves shown in figures 23 and 24; the data obtained during roughly the first 1,000 minutes of the test clearly plots off the slope of the type curves which are matched to the data obtained during the last part of the test. Figure 23 illustrates clearly the magnitude of the correction for partial penetration that must be applied to the two 30-foot wells and indicates the magnitude of the other corrections required.

Once the type curve has been matched to the drawdown curve, the computation of T and S is almost automatic and follows fixed rules and procedures. Table 10 shows the values for T and S obtained by matching type curves to the corrected drawdowns of the eight principal observation wells, using the period from 3,000 to 10,000 minutes after the pumping started. The average of the eight values of T obtained is 205,000 gpd per ft and the average value of S is 0.203. However, if the values obtained from the two 30-foot wells are omitted, the average value for T is 199,500 gpd per ft and for S is 0.223.

The same data, similarly corrected, was also plotted as a series of distance-drawdown curves for times 1,000, 2,000, 5,000, and 10,000 minutes after the start of the test. It is impossible, however, to match any single type curve to any of the distance-drawdown plots, and an acceptable match can be made only to segments of each of these plots. As shown in figure 25, type curves were matched in turn to the first three points on each distance-drawdown curve (30, 100 and 200 feet from the pumped well), to the second, third, and fourth points (100, 200, and 350 feet from the pumping well), and to the third and fourth points. These matchings to each of four plots of distance-drawdown data, gave 8 values each for T and S, as shown in table 10.

TABLE 10.—*Coefficients of transmissibility and storage from time-drawdown curves*

[Time interval from 3,000 to 10,000 minutes after start of aquifer test in December 1950. E(ast) and W(est) of pumped well]

Well	Distance from pumped well (feet)	Depth (feet)	T (gpd per ft.)	S
S6447	29.6E	61	210,000	0.192
S6446	30.3E	121	233,000	.096
S6402	100.3E	59	192,000	.270
S6401	100.9E	114	195,000	.247
S6404	200.5E	54	197,000	.206
S6403	200.6E	102	199,000	.191
S6495	200.1W	56	206,000	.193
S6496	349.8W	41	208,000	.230
Average (all wells)			205,000	.203
Average (excluding wells S6446 and S6447)			199,500	.223

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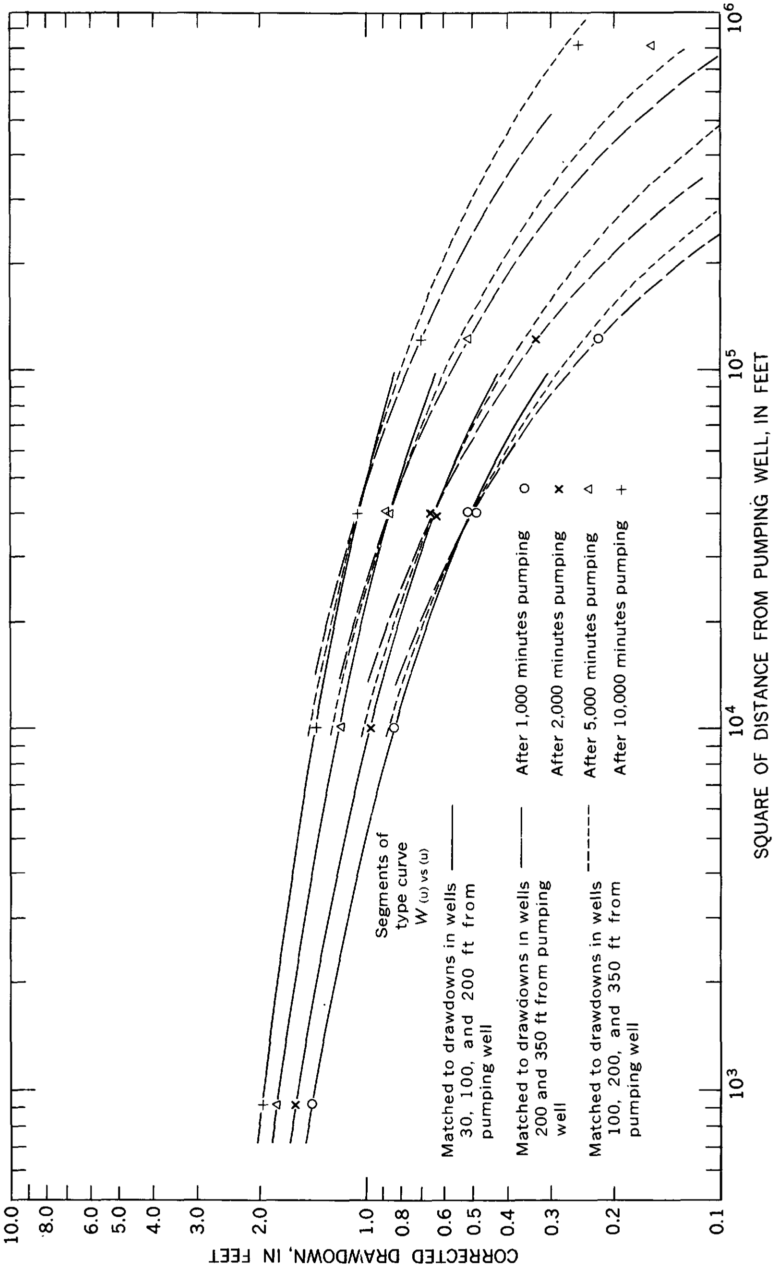


FIGURE 25.—Distance-drawdown plots and corresponding type curves for observation wells used in pumping test, December 1950.

The values for T show little, if any, systematic variation with time, but they show a marked dependence on the distance of the observation well from the pumping well. Data from the 30-, 100- and 200-foot wells give an average value of T of 212,500 gpd per ft, the 100-, 200-, and 300-foot wells give an average value of 188,000 gpd per ft, and the 200- and 350-foot wells give an average value of 163,000 gpd per ft. These values mean that the observed drawdowns at the nearest and most distant observation wells are too small, as compared to those observed at intermediate distances. For this reason a single type curve cannot be passed through all the points. As we have seen, incomplete drainage would tend to increase drawdowns in wells near the pumping well, and decrease them, relatively, for more distant wells, so that this factor cannot be the sole explanation of the failure to fit a single type curve. The effects of partial penetration and of difference between the vertical and horizontal permeabilities are, in general, to increase the observed drawdowns for nearby wells, screened at the same depth as the pumping well, and decrease the observed drawdowns in nearby shallow wells. It is possible that these factors, in combination with partial drainage, are responsible for the poor fit of the type curves in figure 25.

The coefficients of storage (table 10) determined from the several distance-drawdown plots (fig. 25) are noticeably larger for the more distant wells. For all but the nearer observation wells, the coefficient of storage also increases with time, and clearly there is a significant variation of the coefficient of storage both with time and with location in the test area during the period of the pumping test. The Theis nonequilibrium formula is based on the assumption that the coefficient of storage is constant; this discrepancy is one way in which the actual conditions of the test differed from those required for an exact determination of the hydrologic properties of the aquifer.

SUMMARY

The average value for T (coefficient of transmissibility) as determined from the several distance-squared-drawdown curves, is 188,000 gpd per ft, and the average value from the time-drawdown curves is 200,000 gpd per ft, so that the most probable value from a consideration of all the data is 194,000 gpd per ft, or 190,000 gpd per ft rounded off to two significant figures. The saturated thickness of the aquifer in the test area is 145 feet, and the coefficient of horizontal permeability, is 1,300 gpd per sq ft. The average value of S, the coefficient of storage, obtained from the more reliable distance-squared-drawdown curves, is 0.25; from the time-drawdown curve S is 0.223. The value accepted as the result of the test is 0.24.

Once the coefficients of transmissibility and storage have been determined, theoretical drawdowns can be computed for the conditions of the pumping test and compared with the observed drawdowns. The results (fig. 26) may serve as a summary of the pumping test.

Figures 26*A* represents profiles after 10 minutes of pumping. The profile *A–B* is calculated from the Theis nonequilibrium formula using a value for *T* of 190,000 gpd per ft and for *S* of 0.24. Four observed or estimated profiles are compared with the curve *A–B*. The first, *J–D*, represents the estimated actual drawdown of the water

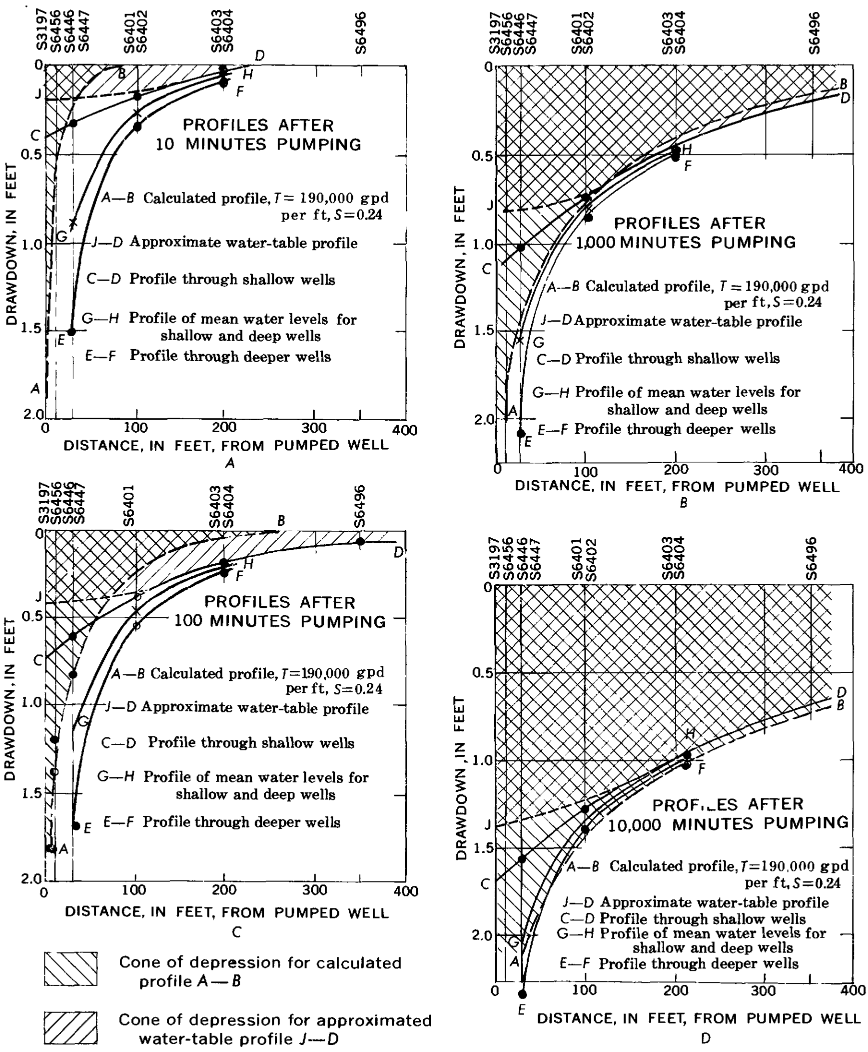


FIGURE 26.—Comparison of observed and theoretical profiles for pumping test, December 1950.

table resulting from the pumping. Near the pumped well the drawdown is much less than that calculated, but the actual drawdown extends much farther from the pumping well than does the theoretical drawdown. The actual drawdown near the pumping well is smaller than the theoretical drawdown, because the water table there is at least 60–65 feet from the screen of the pumping well; with the full-length screen the formula assumes this distance would have been very small. The difference between the calculated and actual drawdowns for these points is due to the pressure drop across 60–65 feet of sand, a drop which is somewhat increased, because the direction is normal to bedding, and the permeability is less than it would be in a horizontal direction.

Because the drawdown of the water table near the well is less than the theoretical, the actual cone of depression must be more extensive because the water pumped must result in a dewatering of the sand somewhere. The volume actually dewatered, however, is about 2.5 times the calculated volume, a point not immediately apparent on inspection, but it must be remembered that the shaded areas in figure 26 are sections of cones of revolution, and the volumes vary as the radii squared. Because the volume unwatered is larger, the coefficient of storage must be smaller; the value indicated by the actual drop of the water table 10 minutes after the pumping started was 0.10.

The curve *C–D* is a plot of the drawdown observed in the shallower of the two wells at 30, 100, and 200 feet, and the curve *E–F* is the observed drawdown in the deeper well of each pair. The difference in drawdown between these two curves is, of course, due to the partial penetration of the aquifer by the screen of the pumping well, so that the drawdown in the deeper observation well, screened at approximately the same depth as the pumping well, is greater than the theoretical drawdown. Similarly, the drawdown in the shallower wells is less than calculated, but farther out it is greater, a result of the extended cone of depression at the water table. The curve *G–H* is the drawdown that would have been observed in observation wells of intermediate depth so chosen that the complications resulting from partial penetration are virtually eliminated. Differences between this curve and the values calculated from the Theis formula should be due almost entirely to incomplete drainage of the sand. The large differences after only 10 minutes of pumping suggest that early data in any test in a water-table aquifer must be used with caution because of slow drainage.

The curves showing similar observed and calculated values after 100, 1,000, and 10,000 minutes indicate that near the pumping well, and out to a distance of about 200 feet, even after extensive pumping there is still a significant departure between the calculated and observed drawdown at the water table. For this reason, it is customary in pumping tests in water-table aquifers, when partial penetration is a problem, to locate the observation wells no closer than twice the saturated thickness of the aquifer to the pumping well. In this test that would mean no closer than 300 feet, which these curves show would be more than adequate.

At the end of 10,000 minutes, the curve *G–H*, the estimated actual drawdown in observation wells of intermediate depth, closely approximates the shape of the theoretical curve, but lies slightly above it. The shapes and the gradients agree; that is, the calculated value for *T* agrees with the observed data, but because the volume unwatered after even long continued drainage is somewhat less than the calculated volume, it is apparent that the value of *S* used was too small. The value for *S* clearly increased with time, and the value used, 0.24, is an average of the values determined for the later parts of the test, and should, therefore, be smaller than that indicated by the observed final drawdowns.

TRACER TEST

During the pumping test on well S3197, a tracer was used to measure the time of travel to the pumped well from the shallower well (S6447) of the pair of wells 30 feet from the pumped well. This well, 2 inches in diameter, is screened from 119 to 124 feet below land surface, whereas the pumped well is screened from 116 to 135 feet below land surface. The horizontal straight-line distance traveled by the tracer was about 30 feet. The tracer (about 15 gallons), which had nearly the same density as ground water was prepared by mixing 8 kg of ammonium chloride (NH_4Cl), 20 l of methyl alcohol (specific gravity about 0.8), and 32 l of water.

The tracer was introduced at a point a foot or two above the bottom of well S6446 by means of a ⅝-inch rubber hose. The time was 1:20–1:30 p.m., December 12, 1950, or about 1,580 minutes after start of the test. To help force the tracer out of the well into the formation, about 10 gallons of water was poured into the observation well immediately after the tracer liquid was placed in it. The tracer and water are believed to have occupied about 10 cubic feet of sand around the perforations of the observation well. The volume would have been a cylinder about 5 feet high and 1.6 feet in diameter if the mixture was forced out uniformly through the 5-foot perforated section of the casing.

The arrival of the tracer at the pumping well was detected by an increase in the conductivity of the water pumped. Calibration of the conductivity cell made it possible to convert this data into chloride content as parts per million (fig. 27). Integration of the area under

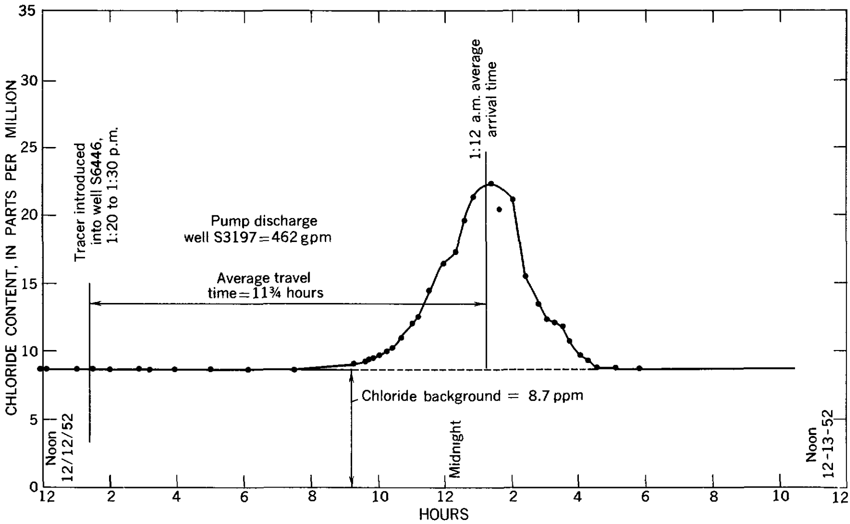


FIGURE 27.—Chloride content of water and time of travel to well S3197 after introduction of tracer in well S6446, 30 feet away.

the curve and above background gives a total of 5,300 g of chloride, or 83 percent of the chloride in the tracer. The first appearance of the tracer was noted 7 hours after it was introduced into the observation well, and the last of the tracer was noted 8 hours later. The average travel time of the tracer was 11.75 hours, during which 325,700 gallons of water was pumped, equivalent to the volume of water in the pores of the sand through which the water moved into the well during 11.75 hours. The shape this volume of water occupied can be estimated because one point on its outer surface was the screen of the well into which the tracer was introduced.

If the screen had been fully penetrating and the flow a radial two-dimensional flow into the screen, the shape would have been a cylinder having a radius of 30 feet and height of 145 feet. If the porosity were 33 percent, the amount would have been 1,025,000 gallons. Because of partial penetration the source of the water has a spherical shape centered around the screen. The difference between 325,700 gallons and 1,025,000 gallons illustrates the problem arising from partial penetration.

Figure 22 shows that we should consider, instead of a cylinder, a spheroid of revolution and a sphere. The spheroid, having an axis of 90 feet and a radius of 30 feet, is the shape through which water would move if the vertical and horizontal permeabilities were equal. (Scale this from fig. 22.) The sphere, having a radius of 30 feet, is the shape that corresponds to a horizontal permeability four times the vertical.

The spheroid has a volume of 170,000 cubic feet and, if the porosity were 0.33, would hold 424,000 gallons of water; the sphere has a volume of 113,000 cubic feet, and it would hold 283,000 gallons. The tracer test suggests that the actual shape, which holds 325,700 gallons, is intermediate between the two and the corresponding horizontal permeability is about three times the vertical. The method is not sufficiently accurate, however, to give much confidence in this numerical result, although it does substantiate the earlier conclusion as to the general relation in the test area between the horizontal and vertical permeabilities.

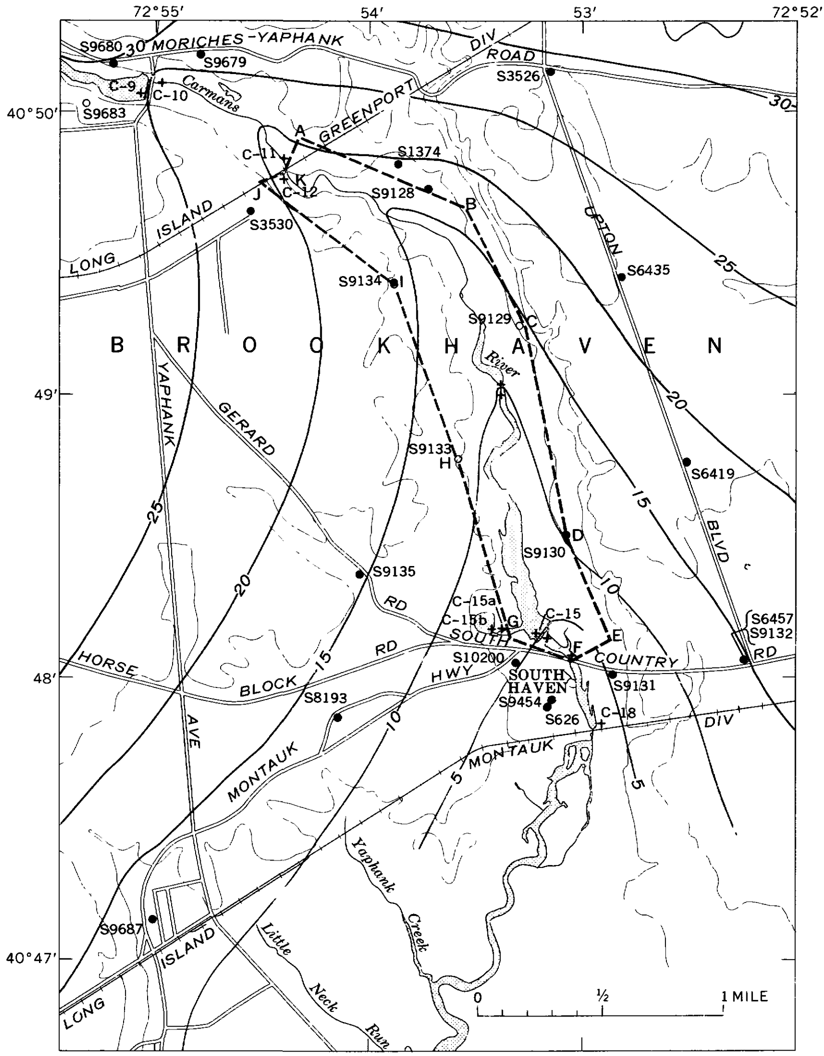
A second similar tracer test was attempted using the shallower of the two 30-foot osbervation wells, but the test pumping had to be stopped 5 days later, before the tracer had appeared. Calculations suggest that even under the most favorable geometry the tracer could not have made the 70-foot diagonal distance between the screens in this short time.

DETERMINATION OF HYDROLOGIC COEFFICIENTS FROM STREAMFLOW DATA

Under conditions of equilibrium the coefficients of transmissibility and permeability can be determined by flow-net analysis for any area when the total flow and the geometry of the aquifer can be measured. The required information can sometimes be obtained from data on streamflow for a selected reach of a stream, from water levels and gradients in the vicinity of the stream, and from other hydrologic and geologic data for the aquifer under consideration. The unknowns—the coefficient of transmissibility, T, and the coefficient of field permeability, P—can then be calculated. The middle reach of the Carmans River between the Long Island Rail Road at Yaphank Station and the Montauk Highway, Route 27, was chosen for such a determination, chiefly because of the large increase in streamflow, about 32–33 cfs in 2¾ miles, due to ground water that has joined the streamflow.

The part of the aquifer considered is shown in figure 28 by the polygon ABCD . . . . K, an area of 0.91 square mile, on both sides of the Carmans River. The polygon is determined by the positions of 11 observation wells, 7 of which were constructed for this purpose. The two longer sides of the polygon are at least 600 feet from the river, far enough to avoid the effects of the vertical convergence of the ground-water flow lines, and also distant from certain local variations in the water table found near the stream.

The water budget for this area, on July 29, 1952, was computed as follows: Stream inflow at the upper end, where the Carmans River flows under the Long Island Rail Road at the Yaphank Station, was 21.8 cfs (cubic feet per second) as measured by a gaging station oper-



EXPLANATION

• S9687
Observation well

+ C-18
Observation point
Number indicates surface-water stage

——— 5 ———
Water-table contour, showing altitude of water table
Contour interval 5 feet. Datum is mean sea level

D— — — — — — —E
Perimeter of selected area along which water-table gradients were determined

FIGURE 28.—Water-table contours, July 29, 1952, lower reach of the Carmans River.

ated by the U.S. Geological Survey. In addition, there was a single small tributary near the lower end of the river segment which contributed 0.6 cfs.

The ground-water inflow into the area, largely through the upper Pleistocene aquifer, is the quantity which we are trying to determine, but it cannot be measured directly; it must be estimated from the balance left when the other factors have been considered. A small amount of ground water from the Magothy(?) Formation also makes its way up through the Gardiners Clay. The amount that thus enters the area from below may be estimated from the permeability of the Gardiners Clay, about 0.3 gpd per square foot, and from the gradient across the clay, about 0.2–0.3 foot per foot. The area is 0.91 square mile, or about 25 million square feet, which multiplied by 0.3 gpd and 0.25 foot per foot, gives about 2 million gpd, or 3.1 cfs.

No rain had fallen for 3 weeks prior to July 29, 1962, and there was thus no direct contribution of water from this source; however, the water table was falling during this period at a rate of 0.003 foot per day and, from the 25-million-square-foot area, contributed 0.21 cfs to the streamflow.

Water was lost by evapotranspiration in about half the area where the depth of the water table was less than 4 feet. Evapotranspiration in July averages between 3.4 and 5.3 inches and may have been somewhat more in July 1952. Evaportation from a class A pan in New York City for the month was 8.40 inches. The value chosen was 5.9 inches, or 0.189 inch per day, which, over an area of 0.45 square mile, is equal to 2.3 cfs.

Streamflow out of the area at the Montauk Highway (Route 27) is difficult to measure because the flow here is affected by the tides. The flow as measured on July 29 was 54.4 cfs.

The inflow into the area, except for the flow in the upper Pleistocene, was 21.8 cfs+0.6 cfs+3.1 cfs, to which should be added change in storage of 0.21 cfs, a total of 25.71 cfs. The outflow was 2.3 cfs+54.4 cfs, or a total of 56.7 cfs. The difference, very nearly 31.0 cfs or 20 mgd, represents the ground-water flow into the area through the upper Pleistocene aquifer.

On the west side of the area, the total length of the sides G–H–I–J is roughly 2.25 miles. Taking into account that the flow is normal to an angle of about 70°, the effective width of aquifer on the west is 2.1 miles, and the average gradient is about 14.3 feet to the mile, a total of 30.0 foot-miles per mile. Similarly, on the east the effective width of the aquifer normal to the direction of flow is 2.5 miles, and the average gradient is 21 feet per mile, a total of 52.5 foot-miles per mile. The flow of 20 mgd is, therefore, being supplied by a total of

82.5 foot-miles per mile, the equivalent of a mile of aquifer having a gradient of 82.5 feet to the mile, or of 82.5 miles of aquifer having a gradient of 1 foot to the mile. In either case, the transmissibility is 20 mgd divided by 82.5 foot-miles per mile, or 240,000 gpd per ft. The aquifer in this area has an average saturated thickness of 170 feet, so that the permeability is about 1,400 gpd per square foot, slightly larger than the value of 1,300 gpd per square foot obtained from the pumping test. The difference is well within the limits imposed by the accuracy of the stream gaging and of the basic assumptions used in the pumping test. It is probably also within the natural limits of variation of the coefficient of permeability. The most important conclusion is that the permeability of the aquifer as measured within the Laboratory area is very similar to that to the south outside the Laboratory because this assumption must be made in computing the velocity of travel of ground water in these areas.

If the 20 mgd of estimated ground-water flow into the area comes in from the east and west in proportion to the effective width of aquifer and the average water-table gradient on each side, then the east side contributes 12.7 mgd and the west side 7.3 mgd. Tracing back the ground-water flow lines to the main water-table divide shows that the area contributing water on the west is 3.9 square miles and that on the east is 10.1 square miles. From the west the contribution is at a rate of 1.87 mgd per square mile, and from the east 1.16 mgd per square mile. The flow per square mile for the west side seems somewhat high. The average annual recharge is about 22 inches a year, about 51 million cubic feet of water per year per square mile, or 1,050,000 gpd per square mile. There are several possible explanations, one being that the drainage area on the west is larger than estimated As may be seen on plate 2, rather small changes in the shape of the water-table contours would substantially increase the long, narrow drainage area. Also, near the ground-water divide is a large area underlain at shallow depth by material of low permeability which may at times deflect water and increase the effective drainage area contributing water to the south or contribute water from perched lenses above the water table. There may be more than the average flow through the Gardiners Clay in some part of the ground-water drainage area. These matters are all speculative, and although of interest as examples of the probable types of hydrologic complexities of this part of Long Island, they do not affect the validity of the determinations of transmissibility and permeability which are the principal concern of this section.

DIRECTION AND RATE OF MOVEMENT OF GROUND WATER UNDER NATURAL CONDITIONS

Plates 1 and 2 show water-table contours for August 29–31, 1951, when the water table was about a half a foot below average, and for July 28–30, 1952, when the water table was 12 feet above average. The direction of ground-water flow may be taken as normal to these contours because the formation is almost isotropic. The rate of flow may be approximately determined by either of two independent methods, one of which is based on consideration of the quantities of water involved, and the other on the relation between transmissibility and the ground-water gradient.

The transmissibility of the upper Pleistocene aquifer is very close to 200,000 gpd at unit gradient. The water-table gradient is about 5 feet to the mile, so that in the Laboratory area each 1-foot width of the aquifer is carrying about 200 gpd, or 26.7 cubic feet per day, which represents a ground-water velocity of about 0.535 foot per day, or about one-third the velocity derived from consideration of the volume of recharge. Thus, in the belt between the Laboratory and the water-table divide, a large proportion of the ground-water recharge, perhaps two-thirds of the total, apparently moves into the deeper Cretaceous aquifers, and only the smaller part moves laterally through the upper Pleistocene aquifer.

A more detailed study of the direction and rate of movement of the ground water in the upper Pleistocene may be based on the map shown in figure 29. The solid flow lines in this figure are based on the water-table map for August 29–31, 1951, and the dashed flow lines on the map for July 28–30, 1952. In general, these lines follow much the same pattern, but, the slight changes in the contours of lines C–D and C'–D' produced a marked difference in the ultimate destination of the water.

The average annual recharge to the water table is about 22 inches. A strip of land 1 foot wide extending from the water-table divide for a distance of 1 mile in the direction of ground-water flow would contribute annually a volume of about 9,700 cubic feet. The water would flow from the lower end of the strip through the saturated part of the aquifer, about 150 feet thick, which has a porosity of about 0.33. The rate of movement is the same as if 9,700 cubic feet of water a year flowed through an opening 50 feet high and 1 foot wide, or about 195 feet per year or 0.535 foot per day. According to this method of analysis, the rate of movement at any point is directly proportional to the flow-line distance from the water-table divide; thus, under the center of the Laboratory tract, 2.5 miles from the divide, the rate of movement of the ground water would be about 1.6 feet per day.

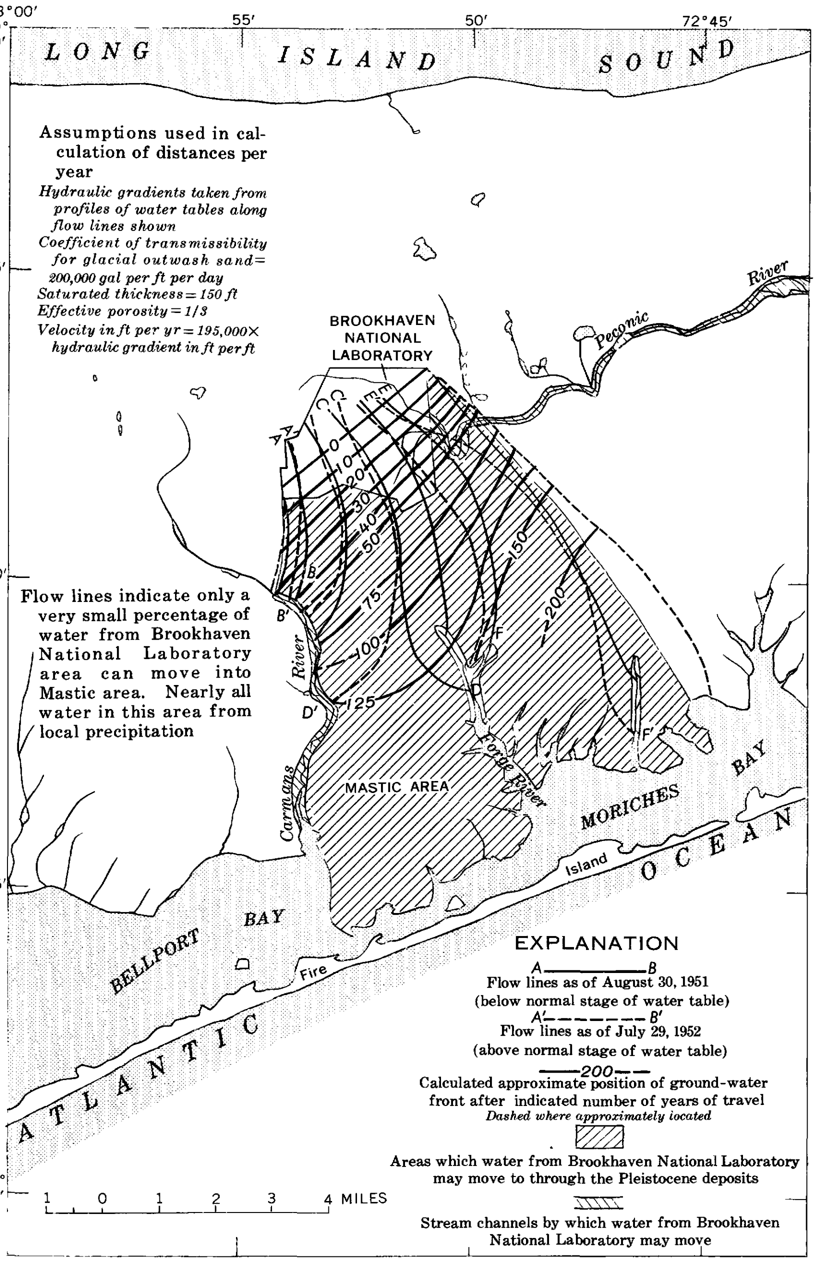


FIGURE 29.—Direction and time of travel of ground water laterally in upper Pleistocene deposits from the Brookhaven National Laboratory area to points of discharge.

These figures are maxima because near the ground-water divide some of the ground water moves downward into deeper aquifers; near the shoreline, where much of this artesian water moves up into the water-table aquifer, streamflow will carry away some of the water which would otherwise move through the upper Pleistocene deposits. Movement of water along these alternate ground-water and surface-water paths will reduce the volume of flow through the upper Pleistocene and, in consequence, reduce also the rate of flow.

In general, ground water from the Laboratory area goes to one of four principal destinations, although the distinctions among these are not always clear. Ground water from the western third of the Laboratory area flows south and then west into the Carmans River; ground water from the central part of the area flows south and east to the Forge River; and water from the eastern part of the area flows southeast, partly into Moriches Bay and partly into small streams which drain into the bay. In addition, some shallow ground water finds its way into the upper section of the Peconic and is carried eastward as surface-water flow, as described in the following section.

The rate of lateral movement, in feet per day, as any point on a flow line, is given by the relation $V=535I$, where I is the water-table gradient. Three sets of plots (fig. 30) show that the average rate of flow in the Laboratory areas is about a half a foot a day and that this rate only increases to 2–3 feet a day near the discharge points. At a rate of a foot a day, the ground water would take 14 years to travel 1 mile.

Figure 30 shows the time of travel from the hospital area, from the waste storage tanks, and from the filter bed of the sewage-treatment plant to the nearest surface-water discharge point. The travel-time from any point where the ground water might become contaminated to the area boundary is 15–50 years. From the boundary to the Carmans River it is at least 25 years, and to the Forge River at least 60 years.

Travel time calculated from the average of the two water-table profiles (fig. 30) represents present-day normal conditions, but even extremely high water-table conditions would not materially increase the rate of flow or shorten the travel time. The rates and times represent averages, and locally, where beds of more than average permeability are present, these rates may be appreciably exceeded. The importance of this factor is difficult to estimate. The excellent exposures of the upper Pleistocene in the wave-cut bluffs along the north shore of the island show that the deposits are lenticular and that any continuous layer of high permeability is of very limited extent. These sections, however, are in the moraine; in the outwash to the south of the Laboratory area, buried stream channels may extend several miles

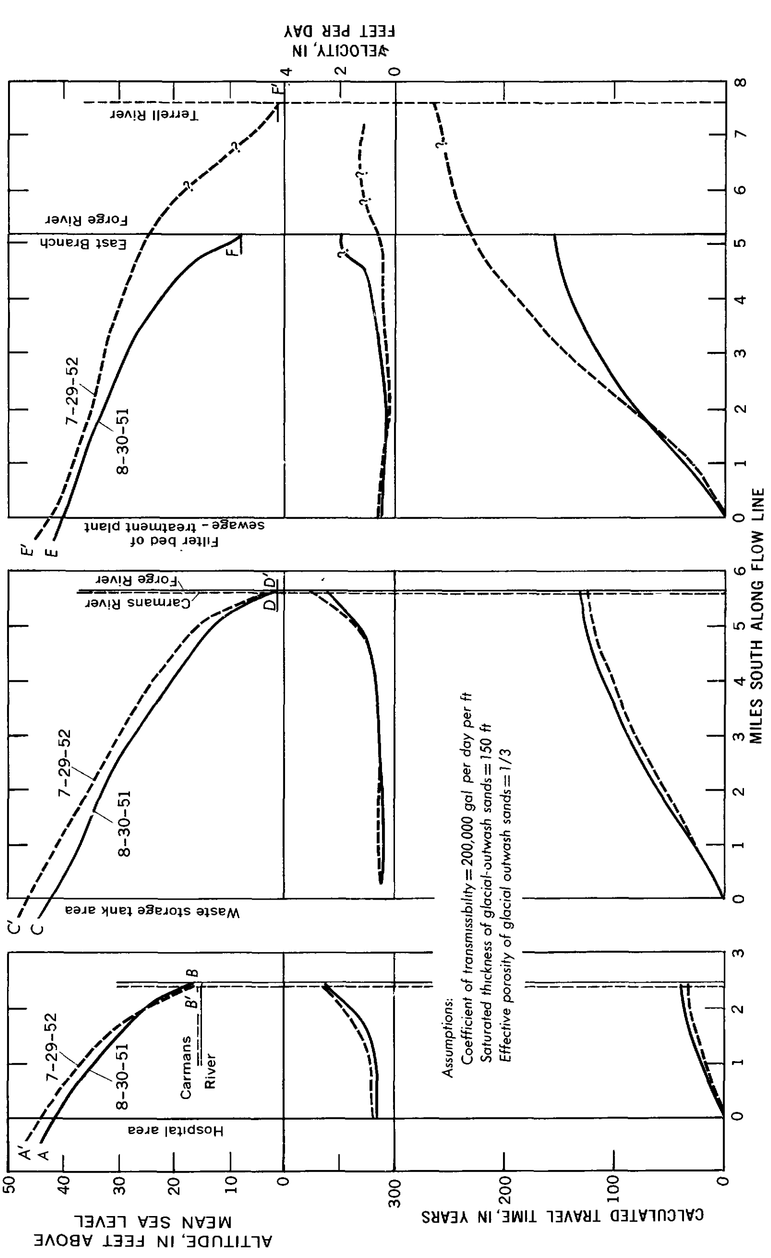


FIGURE 30.—Rate of movement and time of travel of ground water in upper Pleistocene deposits from Brookhaven National Laboratory area along three sets of flow lines, shown on figure 29.

in the general direction of ground-water flow. If any such channel existed, test drilling might well fail to find it, but if the channel carried any appreciable part of the ground-water flow, the shape of the contours drawn on the water table would show the water flowing into it much as they are influenced by the flow of water into a surface stream. Because the water table shows no such irregularities, probably no important buried channels exist. The effect of a number of minor channels, however, is difficult to assess. Locally, the ground water would advance along them more rapidly than through the adjacent sands, but statistically, over any appreciable distance, their net effect would probably be small. The problems raised by the inhomogeneity of the aquifer, however, cannot be answered by such generalizations, and the rates of travel and the travel times shown in figures 29 and 30 must be taken with many reservations. Even in a homogeneous and isotropic aquifer the rates of travel as computed would be average rates, and some part of the water would travel somewhat more rapidly. The purely hydrologic causes of longitudinal dispersion are not important, however, except over short distances of a few feet or less. Over distances of a mile or so they would be completely subordinate to variations in velocity produced by inhomogeneities in the aquifer.

The rates of movement indicated in figure 30 are for the ground water itself and not for a contaminant. The movement of any fission-product isotopes will be but a small fraction of that of the water itself because of adsorption and ion exchange. To be sure, the exchange capacity of the upper Pleistocene sands is relatively low, but the texture of the formation is such that there will be intimate contact between the potential adsorbents and the solution. The minimum travel time for the ground water from points of potential contamination to points of possible outside consumption in streams or wells is measured in decades. The minimum travel time for any radioactive contaminant, except perhaps tritium, is probably to be measured in centuries, although this conclusion is speculative. The absence of overland surface runoff and the very slow and relatively predictable movement of the ground water, however, make the Brookhaven area an exceptionally safe site for a reactor, in terms at least of water-resources contamination. The one possible exception is the Peconic River, which drains the eastern part of the Laboratory area and might possibly receive small amounts of contamination via the sanitary sewer system.

GROUND WATER IN THE MAGOTHY(?) FORMATION

OCCURRENCE

The thickness of the Magothy(?) Formation in the Upton area ranges form a few hundred to 1,200 feet. The position of the upper surface of the formation ranges from about 100 feet below sea level at the shore of Long Island Sound north of the Laboratory to 200–300 feet below sea level at the ocean shoreline south of the Laboratory. The lower surface ranges from 450–650 feet below sea level on the north to about 1,400 to 1,500 feet below sea level on the south. Under the Laboratory area the formation is 800–900 feet thick.

The Magothy (?) is largely composed of sandy clay and clayey sand of moderate to low permeability, but it contains some beds of sand and gravel that yield water abundantly to properly constructed wells. It also contains some beds of pure hard clay through which water can move only with the greatest difficulty. Neither the gravel layers nor the clay layers appear to be continuous over any appreciable distance; the inability to correlate stratigraphic units in the Magothy (?) between the two deep wells at Brookhaven is typical of the experience with even more closely spaced wells in the western part of the island. An exception may be made in the case of the coarse sands commonly found at or near the base of the formation inasmuch as these constitute, if not a stratigraphic unit, then at least a consistently coarse textured zone.

The Magothy (?) Formation is not properly a single hydrologic unit, but it will have to be treated as such here because not enough data are available to permit subdividing it into smaller units.

WATER LEVELS IN THE MAGOTHY(?) FORMATION

The general pattern of movement of ground water through the Cretaceous formations is shown in figure 4. North of the Laboratory near the ground-water divide, a large proportion of the ground-water recharge moves down into the Magothy (?), in no small part because the Gardiners Clay is missing in this area. The Gardiners Clay is present south of a line extending east and west approximately along the north boundary of the main Laboratory tract. This clay layer largely confines the water in the upper Pleistocene but, as discussed elsewhere, it is not uniform in composition and thickness and may locally be somewhat sandy and relatively permeable; it even may be absent over small areas.

Four pairs of wells were drilled to explore the Gardiners Clay and the upper part of the Magothy (?) in the general vicinity of the Laboratory. Wells S9116 and S6458 were installed a few feet apart near the northeast corner of the main Laboratory tract (fig. 31). Well S6458 penetrated about 1 foot of clay at a depth of 220 feet, presumed to be the northern edge of the Gardiners, and was screened at 257 feet in the upper part of the Magothy (?). Well S9116 was screened just below the water table in the upper part of the upper Pleistocene. The water

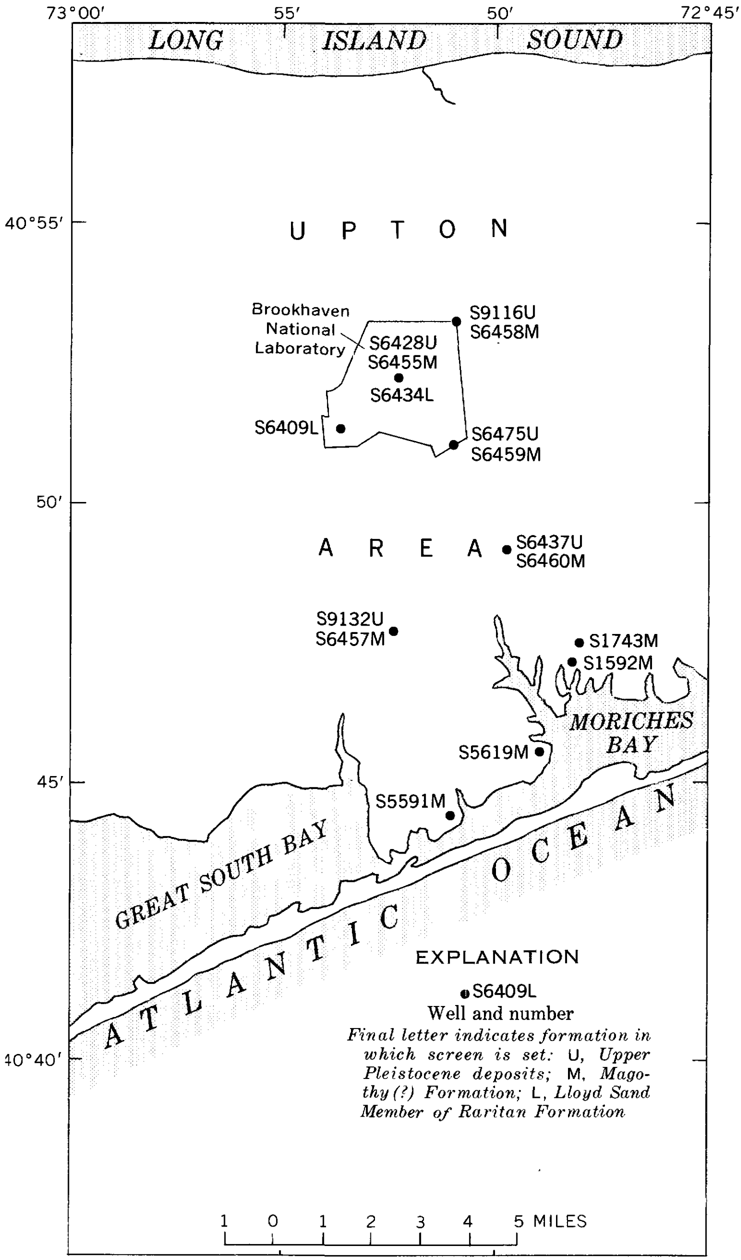


FIGURE 31.—Location of artesian observation wells and selected ground-water wells used in comparison of water levels in the Upton area.

level in S9116 has been consistently higher than the level in S6458, the difference ranging from a small fraction of a foot in September 1952 to over a half a foot in March 1953 and averaging about 0.4 foot (fig. 32).

A second pair of wells, S6459 and S6475, are located 14 feet apart in the southeast corner of the Laboratory area. Well S6459 penetrated 9 feet of sandy Gardiners Clay between 136 and 145 feet and was screened in the upper part of the Magothy(?) at 160 feet. The water level in S6475 (a shallow water-table well) ranged from 1.2 to 0.5 foot above the water level in S6459 and average about 0.6 foot higher.

A third pair of wells, S6460 and S6437, are located about 60 feet apart, 2.4 miles southeast of the southeast corner of the Laboratory area. Well S6460 penetrated 16 feet of Gardiners Clay between 179 and 195 feet and was screened in the upper part of the Magothy(?) at a depth of 200 feet. Well S6437 is a shallow well screened in the upper Pleistocene. The water level in S6437 was consistently higher than the level in the Magothy(?) well, the difference ranging from 0.3 to 0.6 foot and averaging about 0.4 foot. However, if allowance had been made for the distance between the wells and for the hydraulic gradients in the two aquifers, these differences would have been about 0.1 foot less.

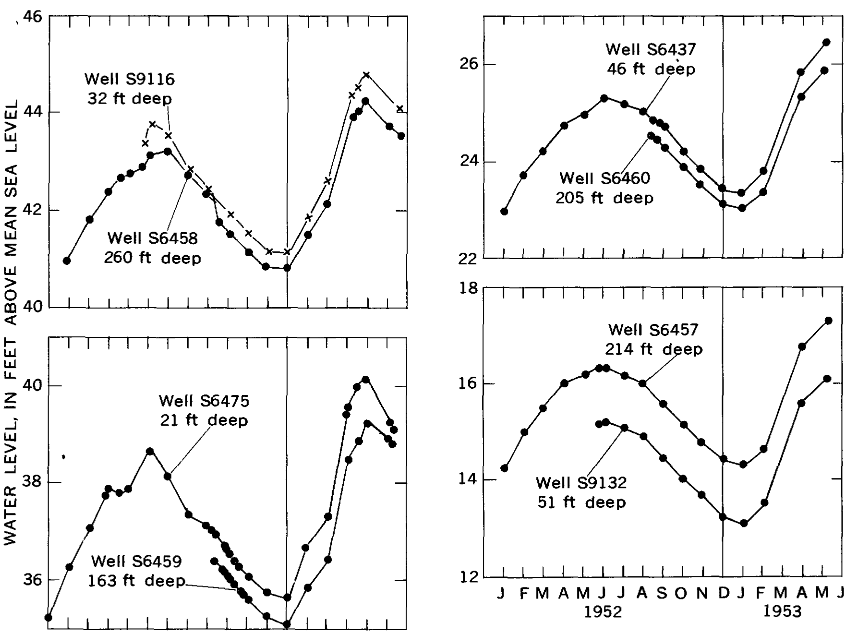


FIGURE 32.—Comparisons of water levels in the upper Pleistocene and in the upper part of the Magothy(?) Formation at four sites in the Upton area. For well locations, see figure 31.

At all these locations the water levels in the upper Pleistocene aquifer stand higher than those in the Magothy(?), and water must be moving down to the Magothy(?). The amount of recharge is difficult to estimate. It is probably greater at S6558 than at S6459 because, although the head difference is less, the Gardiners Clay is much thinner. The recharge at S6460 is probably small because not only is the head difference small but the clay layer is relatively thick.

The water levels are reversed in the fourth pair of wells, S6457 and S9132, located at the intersection of Upton Road and Route 27, about 4 miles south of the Laboratory. Well S6457 penetrated 8 feet of clay and sandy clay of the Gardiners and is screened at 211 feet in the underlying Magothy(?). The water level in S6457 averages 1.16 feet higher than the water level in the upper Pleistocene well. In this area water from the Magothy(?) is moving back up into the upper Pleistocene. The dividing line between areas of predominantly upward movement through the Gardiners Clay and areas of predominantly downward movement probably extends roughly east and west parallel to Route 27 and probably lies betwen half a mile and 1 mile north of it.

In addition to the four pairs of wells installed for study purposes, four privately drilled wells near the south shore of the island were used to study water levels in the upper part of the Magothy(?). Near the south shore and on the barrier beach, the ground water in the upper Pleistocene is commonly brackish, in part because of salt water driven inland by storms, and it is necessary to screen wells below the Gardiners Clay in order to get potable water. Wells S1592 and S1743 are about 300 feet apart, 5½ miles south-southeast of the Laboratory. Well S1592, 220 feet deep, and well S1743, 329 feet deep, both penetrated about 80 feet of clay and sand of the Gardiners, and in both wells the water level is about 11.5 feet above the water table, which here is approximately at the high-tide line.

Well S5619 is about 2 miles to the south-southwest of wells S1743 and S1592 and about 6 miles south of the Laboratory. It is 272 feet deep and penetrated about 80 feet of sand and clay of the Gardiners. The water level in S5619 is about 12.5 feet above sea level. The fourth well in the Magothy(?), S5591, 300 feet deep, about 2 miles southwest of S5619 and 6.5 miles south of the Laboratory, penetrated more than 100 feet of tight clay. The water level in the well stands about 13.5 feet above sea level. All four of these south-shore wells in the Magothy(?) are on low-lying land and will flow freely if permitted to do so.

The measured or estimated water levels in the top part of the Magothy(?) formation, as shown in figure 4, are a foot or so below those in the upper Pleistocene near the ground-water divide where there is a maximum rate of downward movement. Southward in the Laboratory area, the rate of downward movement is less because of the presence of the Gardiners Clay and a smaller difference in head. About 3 miles south of the Laboratory, the head difference has dropped to less than half a foot, and 4 miles south the difference in head and, of course, the direction of flow are reversed. In the next mile southward, the height of the water table drops rapidly as it approaches the shoreline, but the piezometric surface in the upper part of the Magothy(?) declines gradually, so that the difference in head increases markedly to a maximum value of 12–15 feet at the edge of the bay. On the barrier beach the piezometric surface in the upper part of the Magothy(?) stands 10–12 feet above sea level. This relatively large drop in head across the Gardiners Clay near the shoreline is the result of a concentration of flow upward in this area, not only from the upper part of the Magothy(?) but also from still deeper zones, and of the greater thickness and low permeability of the Gardiners Clay and associated clays.

Some water-level data are available for the intermediate and deep zones in the Magothy(?) Formation. On January 25, 1949, the drilling of well S6434 was interrupted and a temporary 20-foot screen was set at 675 feet for a pumping test. At this time the water level in the well stood 39 feet above sea level, or about 6 feet below the water table, and an estimated 5 feet below the water level in the upper part of the Magothy(?). Removing the casing and screen after this test was so difficult that instead of setting another temporary screen at the bottom of the Magothy(?), as had been originally planned, the drilling rig was moved 20 feet, and a new small-diameter well, S6455, was drilled. The new well was screened in the coarse sand of the basal Magothy(?) between 952 and 962 feet. The water level in S6455 varied between 5.5 and 7.5 feet below the water table. The greater part of the head lost by water penetrating down to the central and lower part of the Magothy(?), therefore, is due to the longer path the water must take to reach these depths and to the resistance of the fine sands and clays within the Magothy(?) itself; the loss is not due to the Gardiners Clay, which is, in any case, probably missing in the area north of the Laboratory where most of the recharge to the deeper aquifiers is believed to originate.

Seasonal variations in the water levels in the three adjacent wells, S6428, S6455, and S6434, screened, respectively, in the upper Pleistocene, the basal Magothy(?), and the Lloyd Sand Member of the Raritan Formation, are shown in figure 33. The variations in the pressure in the basal Magothy(?) are similar to the variations in the water table but have only about 70 percent of the amplitude. Because there is little or no lag in the response of the changes in head in the deeper formations to changes in elevation in the water table, the response is probably due to changes in pressure transmitted by a slight compaction of the beds, not to an increase or decrease in the rate of recharge. The rate of recharge to the deeper formations must change, however, because the difference in head between the water in the upper Pleistocene and in the lower part of the Magothy (?) changed by 30 percent during a brief period (fig. 33).

DIRECTION OF FLOW IN THE MAGOTHY(?) FORMATION

Figure 34 shows maps of the water table and of the piezometric surface in the upper Magothy(?), the basal Magothy(?), and in the Lloyd Sand Member of the Raritan Formation. The water-table map is drawn from observations, but the other three are interpretations based on a few direct observations and on inferences from what is known of the geology of the area. The slope of the piezometric surface in the upper part of the Magothy(?) near the Laboratory and to the south has already been described. On the maps, the height and slope of the surface to the east, under the valley of the Peconic River and to the north, near Long Island Sound, are based on the assumption that a somewhat similar relation to the water table

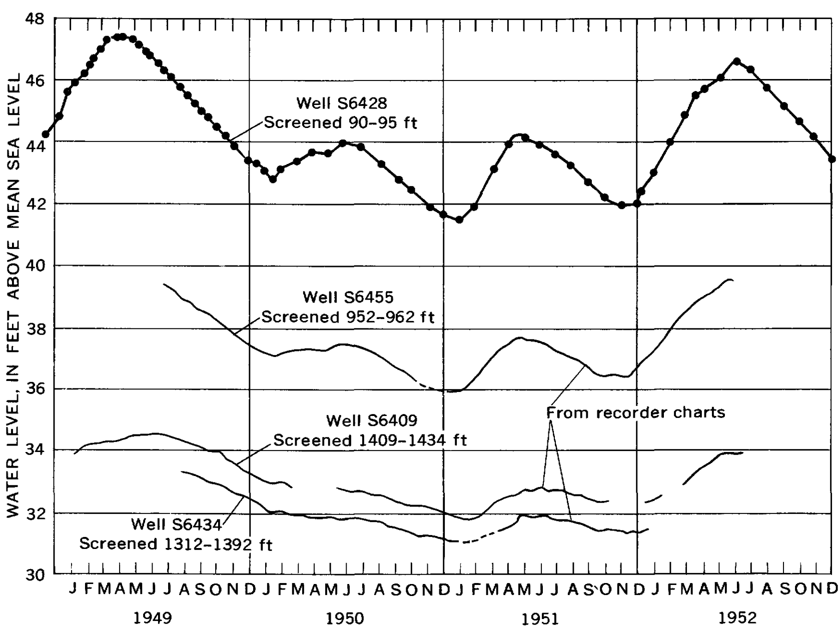


FIGURE 33.—Comparison of water levels in three water-bearing formations, Upton area. Well locations shown in figure 31.

prevails in these areas, but so little is known of the thickness and extent of the several formations that the actual conditions may differ appreciably from those shown. These maps serve, however, to illustrate the expected conditions; they indicate the probable slope of the water table and of the several piezometric surfaces; and they may be used to help visualize the complex pattern of ground-water movement.

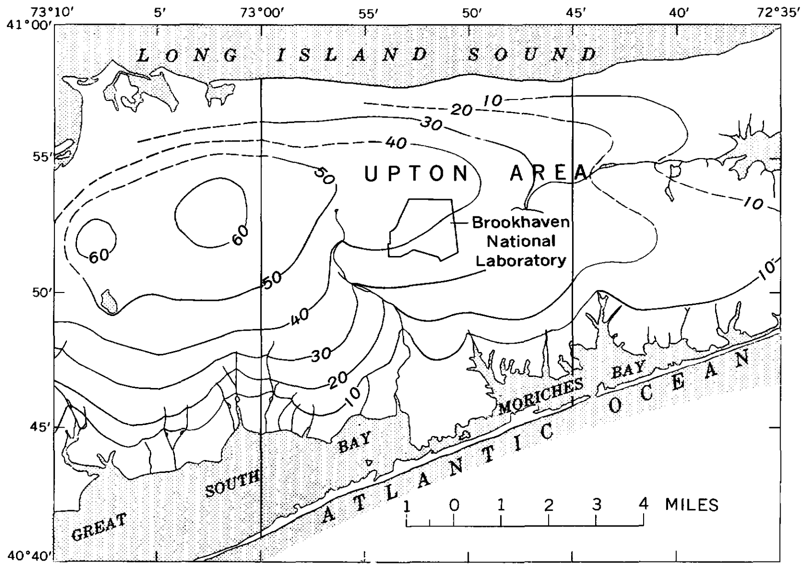
A comparison of the water-table map (fig. 34*A*) with the map of the water levels in the upper beds of the Magothy(?) Formation (fig. 34*B*) shows little difference. The general pattern is much the same, but near such streams as the Carmans and Peconic Rivers, the water table is locally lowered more than is the piezometric surface in the upper part of the Magothy(?); there is a correspondingly greater deflection of the flow lines in the water table toward the flowing streams. In the lower part of the Magothy(?) the pattern is different, not only in detail but also in its general shape. Recharge to these beds is largely concentrated in, although not confined to, the high land to the east of the Laboratory. The streams, and in particular the Peconic River, exert little influence on this deep piezometric surface, and the combined result of these changes is that the direction of ground-water flow in the lower part of the Magothy(?) under the Laboratory area is more to the east than is the flow in the shallow aquifers.

RATE OF FLOW IN THE MAGOTHY(?) FORMATION

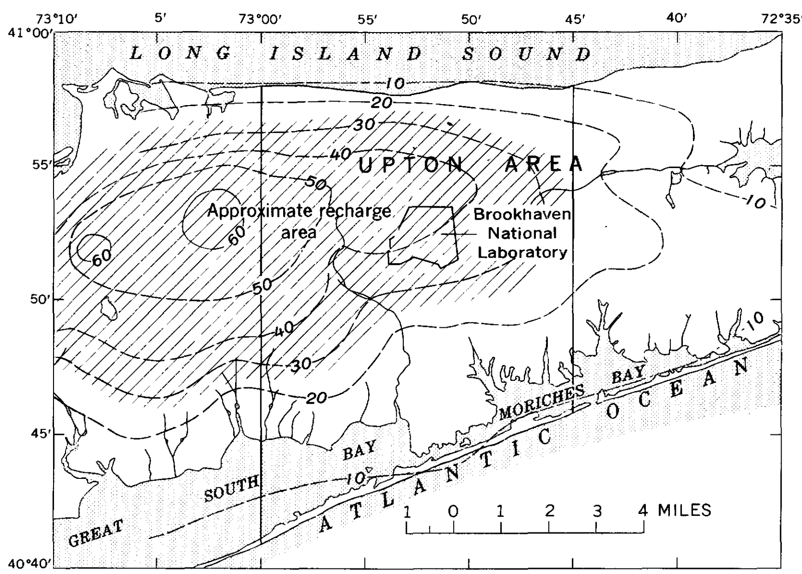
The rate of ground-water flow in the Magothy(?) formation may be roughly estimated either from a consideration of the volume of flow, the thickness of the aquifer and its porosity, or from a consideration of the transmissibility of the aquifer and the prevailing pressure gradients. Unfortunately, data are lacking for a precise determination by either method. The available data show that the rate of ground-water movement in the Magothy(?) is less than that in the upper Pleistocene aquifer and suggest that it may be one-half to one-fifth as great.

A comparison of total ground-water recharge to the rate of ground-water movement in the upper Pleistocene shows that perhaps two-thirds of the ground-water recharge north of the center of the Laboratory is moving down into the deeper aquifers. This amount probably represents the bulk of the recharge to the deeper aquifers because their recharge is much impeded south of the Laboratory by the increased thickness of the Gardiners Clay. The estimate of the volume of water penetrating below the upper Pleistocene may be used to calculate the maximum rate of flow in the Magothy(?).

UPTON AREA



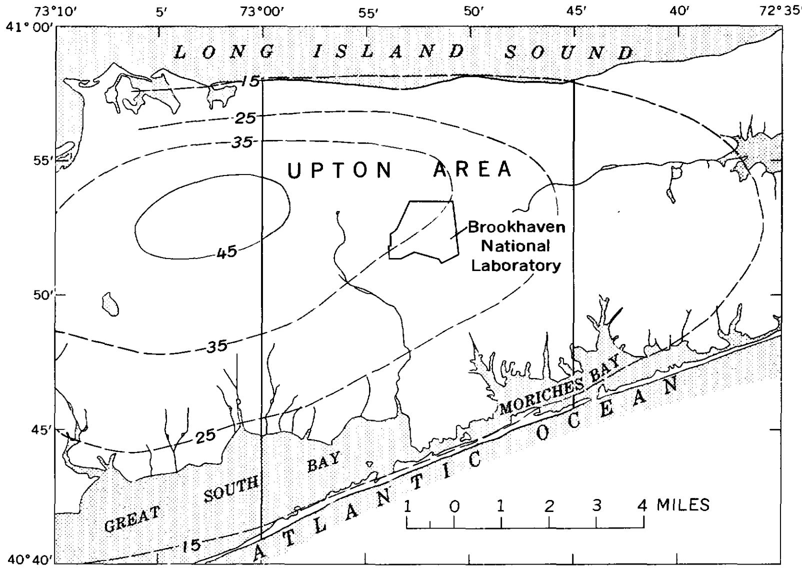
A. Water-table contours, upper Pleistocene deposits, as of January 1951



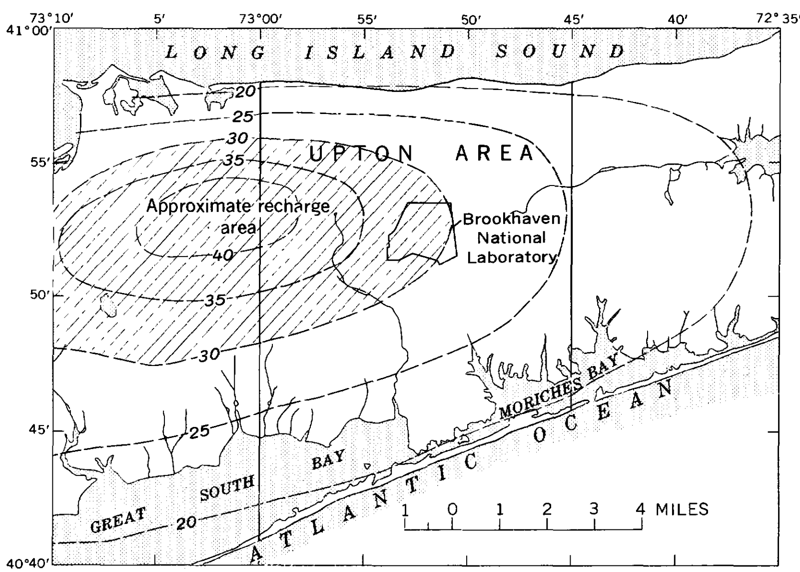
B. Approximate piezometric contours, upper beds of the Magothy(?) Formation

FIGURE 34.—Comparison of water levels in three water-bearing formations in central Suffolk County. Contours dashed where approximately located: contour interval is 10 feet in *A–C*, 5 feet in *D*. Datum is mean sea level. After Lusczynski and Johnson (1951).

UPTON AREA



C. Approximate piezometric contours, basal beds of the Magothy(?) Formation



D. Approximate piezometric contours, Loyd Sand Member of the Raritan Formation

FIGURE 34.—Continued.

As was shown in the discussion of water in the upper Pleistocene, a strip of land 1 foot wide, extending back from the center of the Laboratory 2.5 miles to the ground-water divide would receive ground-water recharge at an average rate of 24,250 cubic feet a year. The transmissibility of the upper Pleistocene and the water-table gradient are such that each 1-foot width of the upper Pleistocene aquifer normal to the direction of ground-water flow is carrying, on the average, about 9,750 cubic feet a year, and 14,500 cubic feet a year of recharge is left to enter the Magothy(?) Formation. The Magothy(?) Formation under the Laboratory is about 900 feet thick, and the porosity is estimated to be about 25 percent. The water in the Magothy(?) at the southeast end of the hypothetical strip is moving through the equivalent of an open channel with a cross-sectional area of 225 square feet; the average velocity, therefore, is about 65 feet a year, or roughly 0.2 foot per day.

Not much reliance can be placed in this estimated velocity because although the thickness of the formation is known the porosity is not. Possibly much of the flow is confined to more permeable beds; this confinement would reduce the effective cross section through which the water is moving and the velocity would accordingly be greater. No allowance is made for water which moves down into and through the Lloyd Sand Member; and although the volume of this flow is probably small, it would reduce the rate of flow in the Magothy(?).

A study of the flow of the Carmans River at the Yaphank gaging station, which is described in the following section on streamflow, furnishes some information on the volume of recharge to the Magothy(?). The average annual discharge of the river at this station is 21.76 second-feet, and the area contributing ground-water flow to the stream above the gage is believed to be 21.5 square miles. The exact size of the contributing area is somewhat in doubt. It is certainly not the same as the apparent topographic drainage area, and the water-table map does not provide a complete picture because there are locally some beds of relatively impermeable material in the upper Pleistocene above the water table which may produce perched water and deflect the infiltration. If the area of 21.5 square miles is correct, however, then the average flow of 21.76 second-feet represents an annual loss of only 13.5 inches of water, 8.5 inches less than the known recharge of 22 inches. This difference of 8.5 inches probably represents the recharge to the Magothy(?) Formation, which is about 40 percent of the total recharge, not 66 percent as estimated in the preceding section, and would suggest a rate of flow in the Magothy(?) of 0.12 foot per day rather than 0.2 foot. Assuming that the combined thickness of the Gardiners Clay and the clay lenses in the upper part of the Magothy(?) is 10 feet and that the pressure gradient across them is half a foot, the permeability of the clays is 0.3 gpd per square foot.

The permeability of certain beds near the middle of the Magothy(?) Formation can be estimated from data collected during an aquifer performance test made on well S6434 when it was temporarily screened between 656 and 676 feet below land surface. Twenty feet of 40-slot Everdur screen, 6 inches in diameter, was set in a coarse-sand zone and test pumped on January 25, 1949. The thickness of the aquifer tested, that is the vertical part of the Magothy(?) Formation affected by the pumping, cannot be ascertained with any accuracy. The upper limit is very probably a clay layer 606 feet below the surface. The lower limit is uncertain; it may be a clay-rich zone about 700 feet below the surface, or it may be the bottom of the Magothy(?) Formation at a depth of 1,000 feet. In the first case, the thickness of aquifer under test was about 100 feet; in the second case, the thickness was 400 feet.

The well was developed by pumping for a total of about 40 hours. The test itself consisted in pumping the well at a rate of 410 gpm for 100 minutes, then pumping at a rate of 254 gpm for about 150 minutes, and finally at 660 gpm for about 40 minutes. In was impossible to hold the pumping rate constant, and these are average figures. Water levels were measured with an air-line gage and are correct to about 0.3 foot. The static water level at the start of the test was 48.2 feet below land surface, or about 36.4 feet above sea level. The specific capacity, in gallons per minute per foot, was 15.5 at 254 gpm, 13.6 at 410 gpm, and 9.9 at 660 gpm.

The time-drawdown data from the first 100 minutes of pumping were plotted on semilog paper (fig. 35), and a coefficient of transmissibility of about 42,500 gpd per ft was computed by a method suggested by Jacob (1944). This value for the coefficient is necessarily approximate because the pumping rate was not constant, the well was not fully developed, and only part of the aquifer, a zone of unknown and probably variable thickness, contributed water to the well during the test. If this zone was roughly 100 feet thick, as seems probable, then the permeability of the Magothy(?) Formation in this somewhat coarse-textured zone was about 400 gpd per square foot, a value similar to those found for this formation in western Long Island. If the thickness of the contributing zone was as much as 400 feet, then the permeability was about 100 gpd per square foot.

The hydraulic gradient in the Magothy(?) Formation in the vicinity of the Laboratory is about 3 feet to the mile. If the permeability is taken as 400 gpd per square foot, then each square foot of the aquifer taken normal to the gradient will transmit about 0.23 gpd per square foot, which, assuming a porosity of 0.25, is equivalent to a velocity of about 0.12 foot per day. If the thickness of the zone contributing to the

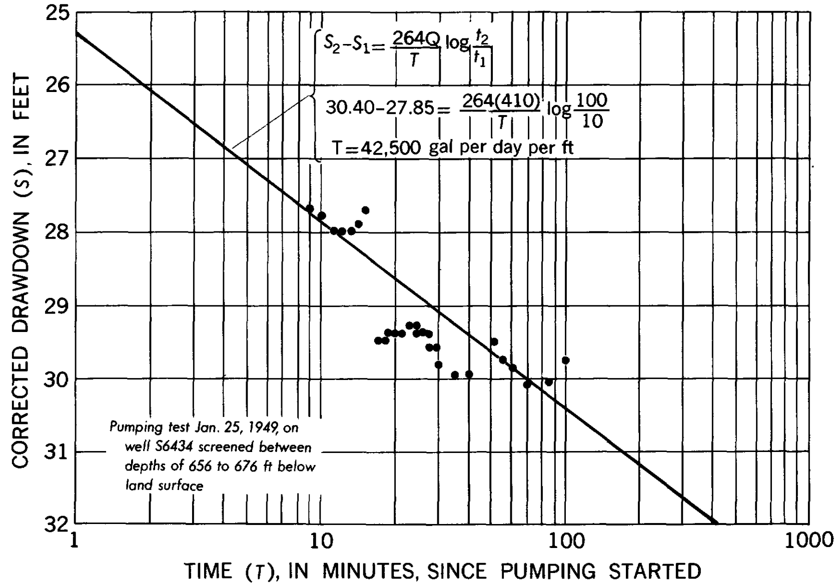


FIGURE 35.—Time-drawdown plot for well S6434 screened middepth in Magothy(?) Formation.

measured transmissibility was as much as 400 feet, the permeability must be correspondingly reduced, and the rate of flow would be about 0.03 foot per day. These values are similar to, but somewhat smaller than, those obtained by the previous method.

In conclusion we may say, despite a lack of detailed knowledge, that an incident at the Laboratory would not result in the contamination of the lower part of the Magothy(?) or Lloyd Sand member aquifers, for they receive almost all their recharge from areas to the north and west of the Laboratory area. The upper Magothy(?) might be contaminated if radioactive materials were able to make their way in solution down through the upper Pleistocene and the Gardiners Clay; but the average rate of ground-water movement in the Magothy(?) is less than in the upper Pleistocene and probably is in the range of 0.2–0.1 foot per day even in the more permeable beds. Movement of the contaminated water to points of natural discharge would therefore take a long time.

GROUND WATER IN THE LLOYD SAND MEMBER OF THE RARITAN FORMATION

OCCURRENCE

The Lloyd Sand Member of the Raritan Formation is the bottom of the stratigraphic column in the Upton area and rests on the deeply weathered surface of the peneplained schists and gneisses of the bedrock. North of the Laboratory, below the shore of Long Island Sound, the Lloyd is believed to be about 200 feet thick, its upper surface being about 900 feet below sea level. Near the ocean shoreline south of the Laboratory, it is believed that the Lloyd is more than 300 feet thick; its upper surface is estimated to be 1,600 feet below sea level. Two deep test wells, S6409 and S6434, showed that under the main Laboratory area the Lloyd is about 300 feet thick and that its upper surface is about 1,100 feet below sea level.

The Lloyd is largely composed of quartz pebbles, coarse to fine sand, and silt and clay. The most characteristic beds in the Lloyd are composed of coarse sand and pebbles and soft clay that fills much of the potential void space and much reduces the porosity. Some beds of sand and clay, however, contain little if any coarse-textured material. Because of marked differences in detail in the sequence of beds in the Lloyd where it was penetrated by the two test wells at the Laboratory, a specific description of this aquifer may be made only for a particular location, although in general it is composed of much the same material over nearly all Long Island.

WATER LEVELS IN THE LLOYD SAND MEMBER

Specific information on water levels in the Lloyd Sand Member in the Upton area is restricted to observations made in the two deep test wells drilled on the Laboratory grounds. These wells are about 8,000 feet apart and the first well drilled, S6409, lies S. 51° W. of S6434, the second.

The water level in S6409 ranged from 31.6 to 34.5 feet above sea level from 1949 to 1952. During this period the water level in well S6434 ranged from 30.9 to 33.4 feet above sea level, which is 11–14 feet below the water table and 5–7 feet below the piezometric surface in the lower part of the Magothy(?). The water table is about 3 feet lower at S6409 than it is at S6434, but the relation is reversed in the piezometric surface of the artesian water in the Lloyd; the water level in S6409 during the period of record stood 0.8–1.0 foot higher than the level in well S6434 to the northeast. Figure 34 shows that this apparent anomaly may be explained if the water in the Lloyd under the Laboratory is moving to the east. The data, although scant, indicates that the movement is in this general direction rather than to the southeast, as is the movement in the upper Pleistocene aquifer. This eastward direction of movement is in agreement with what is known about the probable areas of recharge to the Lloyd, as shown in figure 34. The scant data available would not alone warrant the drawing of these maps. The aquifer, however, has been studied in detail in western Long Island, and this study permits inferences to be drawn about the conditions in the Upton area which local information would not otherwise justify.

RATE OF MOVEMENT IN THE LLOYD SAND MEMBER

It is impossible to compute directly the volume of recharge to the Lloyd, and so this approach to the question of the rate of movement of water in the Lloyd can not be used. The gradient-permeability approach, however, can be used to estimate the order of magnitude of the rate of movement. The transmissibility of the Lloyd can be determined approximately from the results of a pumping test made on well S 6434, on June 1–2, 1949, after it had been screened permanently with 80 feet of 40-slot Everdur screen 6 inches in diameter set in a gravel pack at a depth of 1,227–1,307 feet below sea level. The screened zone was not well chosen because the upper part from 1,226 to 1,261 feet was dominantly clay; below the screen from 1,307 to 1,325 feet there was much better material. Two or three weeks were spent in developing the well, first by bailing, surging with compressed air and pumping, and then by treating the well alternately with Calgon solution and buffered sulfuric acid. The yield and specific capacity of the well were noticeably improved, chiefly by the chemical treatment; further improvement would have been possible, but the expense did not appear to be justified for a test well.

During the test the well was pumped for 24 hours at a rate of 465 gpm, as measured with an orifice. The pumping rate was reasonably constant, except for a 9-minute shutdown 3 hours and 40 minutes after the test started. A correction of 0.3 hour, twice the period of shutdown, was added to the time of all subsequent readings, and it is believed that the short interruption had no effect on the results. The data for the last 20 hours of pumping were plotted on semilog paper (fig. 36), and a value for the coefficient of transmissibility of 12,500 gpd per ft was derived, using the method of Jacob (1944), as was done in the case of the pumping test in the Magothy(?) Formation. The zone contributing water to the well probably extended from 1,260 to 1,325 feet, a total thickness of 64 feet. This thickness gives a permeability of 195 gpd per square foot. Because the test was made without the use of any observation wells, the coefficient of storage could not be determined. The pressure gradient in the Lloyd underlying the Laboratory area is about 1 foot to the mile, and the relatively permeable zone in the Lloyd is transmitting water at a rate of 0.37 gpd per square foot. The porosity of this material is not known, but it is estimated to be 0.20. On the basis of this estimate the rate of movement of the water in the Lloyd is about 0.025 foot per day, substantially less than in either of the principal overlying aquifers. Not only is the water in the Lloyd moving very slowly, but it also must follow a long route before it again emerges, probably seaward of the east end of Moriches Bay (fig. 34), a distance of 15 miles. At the rate of movement calculated, this journey would require nearly 10,000 years.

The possibility of pumping contaminated ground water from the upper Pleistocene down into the Lloyd as a last resort in the unlikely event of serious contamination of the upper Pleistocene aquifer has already been discussed (de Laguna, 1966). The Lloyd east of the Laboratory is not used as a source of water nor is it likely ever to be used. Although it would be most unwise to deliberately contaminate a potential source of water supply, as an alternative to the contamination of an aquifer that is now being used, the idea might have merit under certain circumstances.

STREAMFLOW

INTRODUCTION

The three principal rivers in the Upton area, the Carmans, the Forge, and the Peconic, are fed almost entirely by ground water; that is, they receive very little direct overland runoff. For this reason, and because the apparent topographic drainage areas of these streams do not correspond with the ground-water contributing areas, the flow of these streams bears little relation to the area apparently drained by their valleys. Instead it is determined by the configuration of the water table. The differences between apparent topographic drainage areas and the actual ground-water drainage areas are shown in plate 8.

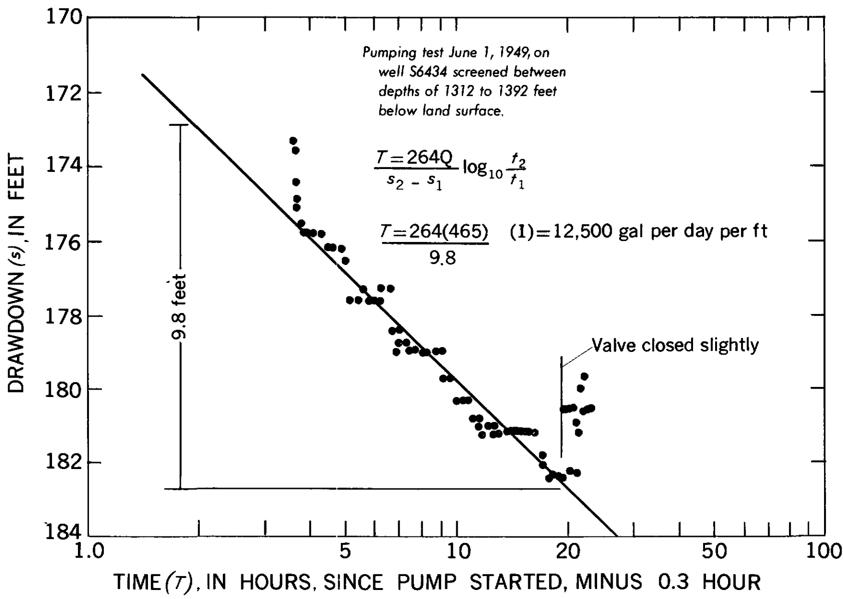


FIGURE 36.—Time-drawndown plot for well S6436 screened in Lloyd Sand Member of Raritan Formation.

Plate 5 shows profiles of Carmans River from Route 25 to Bellport Bay. Plate 6 shows profiles of channel bottom, stream level, and water table for the Peconic River. Figure 37 shows topographic drainage areas for streams in the Upton area. Plate 7 shows water-table contours, ground-water flow lines, and areas contributing ground-water flow to selected gaging stations. The maps on plate 8 show a comparison of topographic and ground-water drainage area for the Carmans, Forge, and Peconic Rivers.

The valleys of both the Carmans and Forge Rivers may be divided into upper, middle, and lower sections. In the upper sections the streambeds are normally dry because they are above the water table; they carry water only on rare occasions when heavy rain has fallen on soaked or frozen ground. All other precipitation not lost by evapotranspiration soaks into the ground to join the water table. The ground-water flow in these areas may be in quite a different direction from the slopes of the streambeds, as in the upper valley of the Carmans River where the ground water flows north although the valley slopes to the south. Because the water table stands high above sea level in these areas, some of the ground water moves downward to recharge the deep aquifers in the Magothy (?) and the Lloyd.

Somewhat similar conditions exist in the upper section of the Peconic River but are more complex because of shallow layers of relatively impermeable sandy clay and silt. These layers produce temporary or permanent areas of perched water which feed or are fed by the streamflow as their relative elevations change with variations in rainfall and in evapotranspiration. In general the upper reaches of the Peconic River skim some of the shallow ground water and direct it as streamflow to the east; the deeper ground water flows to the south and southeast. The Laboratory tract is entirely within the upper topographic drainage area of the Peconic River, but contributes little water to the flow of this river.

In the middle and lower sections of the Carmans and Forge Rivers and in the lower half of the Peconic, the flow is perennial and is fed by ground water that moves in laterally and also upward from the lower part of the upper Pleistocene aquifer. The relation of ground-water flow to streamflow was used to calculate the transmissibility of the upper Pleistocene aquifer adjacent to the middle section of the Carmans River.

In the lower sections of the three rivers, streamflow is sluggish and is periodically slowed or even reversed by tidal fluctuations in the bays. In these areas the ground-water inflow from the water-table aquifer is augmented by water moving upward form the deeper artesian aquifers. Because of these complexities, quantitative streamflow measurements in the lower reaches of the streams are difficult to make and difficult to interpret and were attempted only for the Carmans River.

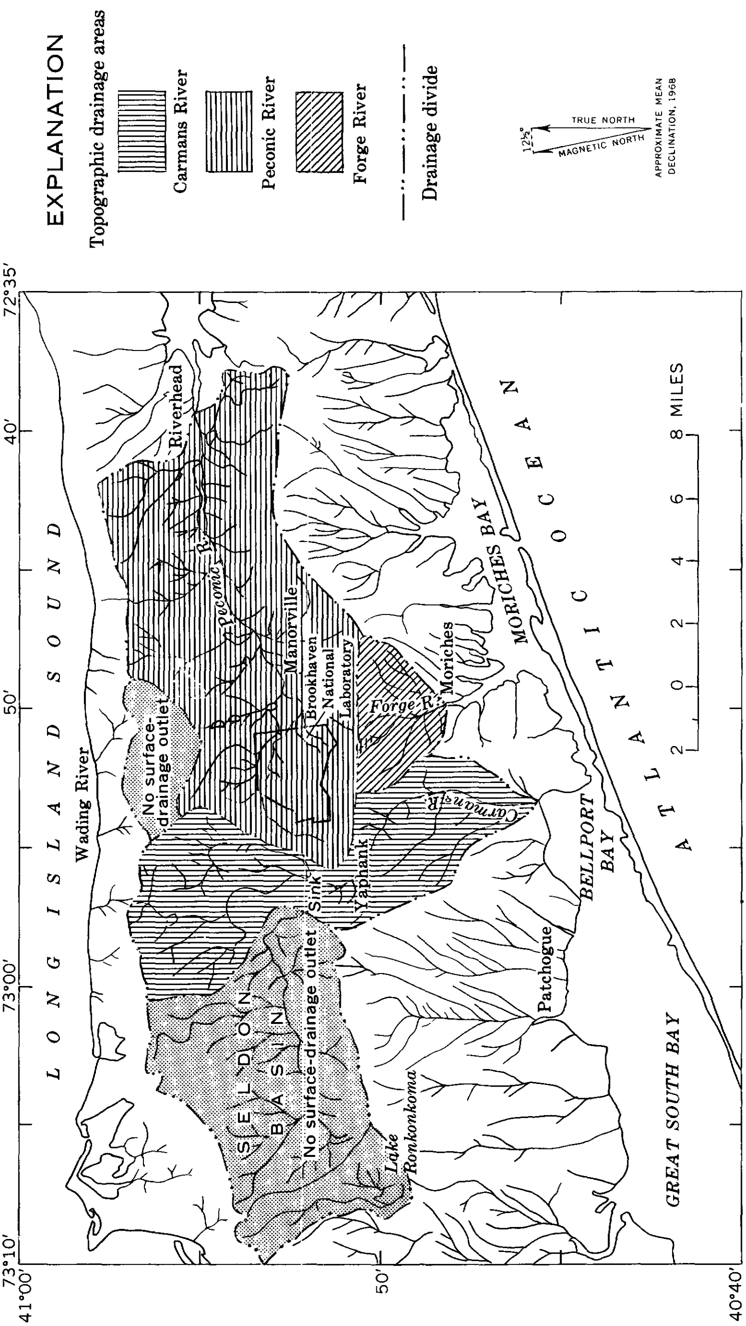


FIGURE 37.—Map showing topographic drainage areas for streams in central Suffolk County.

In the following sections, the Carmans, Forge, and Peconic Rivers are described with the aid of maps (fig. 37 and pls. 7–9), profiles (pls. 5, 6) and tables of streamflow measurements.

CARMANS RIVER

DISCHARGE

The topographic drainage area of the Carmans River is about 100 square miles, but this figure is deceptive because only about half this area contributes water to the river. Some 34 square miles of the apparent drainage area lies north of the ground-water divide, and precipitation falling in this area infiltrates to the water table and then flows north to Long Island Sound. South of the ground-water divide on the west, there is a second area of some 15 square miles that contributes ground-water flow, not to the Carmans River, but to several small south-flowing streams which lie to the west of the river. If the terrain of Long Island were not so very permeable, there would be surface runoff from these areas to the Carmans River, but under the existing circumstances they are drained entirely by ground-water flow. The total area contributing to the flow of the Carmans River is, in fact, only 48 square miles, about one half its apparent drainage area.

Perennial flow of the Carmans begins at Artist Lake, where the river is crossed by Route 25, 3 miles west of the north entrance to the Laboratory, and ends 8 miles to the south at Bellport Bay. The distance, some 12 miles in length as the stream flows, will be divided into three main segments for the purposes of this discussion (pls. 5, 7).

The first section, roughly that part of the river valley which lies north of and which traverses the Ronkonkoma moraine, extends from an indefinite point north of Artist Lake to Bartlett Road. After a series of wet years, resulting in a high water table, the flow of the river probably begins in a small lake just north of Artist Lake, but after several dry years the flow in late fall, when the water table is low, probably begins near Bartlett Road, more than a mile and a half to the south. Five and a half miles south of Artist Lake at Yaphank, where the river is crossed by the Greenport Division of the Long Island Rail Road, a stream-gaging station has been operated by the U.S. Geological Survey since July 1942. Monthly and yearly records of streamflow for this station from 1942 to 1953 are listed in table 11, and more recent records are available in the publications of the U.S. Geological Survey. The flow of the river at this point is in some measure controlled by the two small artificial ponds upstream. The ground-water contributing area upstream from the gaging station is 21.5 square miles, and the average flow for the period of record at the station is 21.76 second-feet, or about 1 second-foot per square mile, which represents an average annual recharge of 13.5 inches. For other points on the stream there are only scattered measurements (table 12).

TABLE 11.—*Monthly and yearly discharge, in cubic feet per second, Carmans River at Yaphank, N.Y.*

[Average yearly (October 1942 to September 1953)=21.76]

Water year	October	November	December	January	February	March	April	May	June	July	August	September	Year average
1941-42										19.13	21.03	19.37	
1942-43	19.20	18.27	19.58	22.91	24.06	24.58	24.11	23.58	21.80	22.07	19.21	21.06	21.69
1943-44	20.31	18.81	16.91	20.33	19.65	21.34	21.56	20.76	19.44	17.08	16.07	18.37	19.21
1944-45	16.77	18.34	19.99	19.37	19.89	20.58	19.54	20.88	19.84	18.67	18.45	16.50	19.07
1945-46	17.46	18.23	19.79	21.34	21.55	21.21	19.19	20.00	22.02	21.22	29.67	24.03	21.31
1946-47	21.79	20.38	20.77	29.81	19.01	18.80	22.54	22.49	20.35	19.50	17.85	16.47	19.99
1947-48	16.40	20.73	19.04	19.16	21.48	27.11	25.48	27.25	27.96	27.08	24.91	22.45	23.29
1948-49	22.59	23.99	23.47	32.01	31.02	30.82	30.07	29.31	26.86	25.53	25.43	25.29	27.18
1949-50	23.47	22.15	21.39	19.87	21.68	21.99	21.01	20.02	20.33	18.45	17.91	15.37	20.30
1950-51	15.94	15.80	17.26	16.25	19.70	21.61	21.92	20.77	18.59	17.36	16.36	16.04	18.12
1951-52	17.24	19.14	19.86	22.32	24.27	26.40	25.90	26.67	26.18	23.06	24.74	22.41	23.17
1952-53	21.05	21.39	21.91	23.35	24.44	30.04	39.40	33.44	29.05	25.89	24.99	22.70	26.06

TABLE 12.—*Miscellaneous discharge measurements, in cubic feet per second, at three sites on the Carmans River compared to the mean daily discharge at the gaging station at Yaphank, N.Y.*

[Discharge measurements made by personnel of Surface-Water Branch, except those on January 18 and 30, February 28, and April 26, 1952, made by M. A. Warren of the Ground Water Branch by applying coefficient of 0.85 to surface velocities]

	Near Bartlett Road	Coram-Yaphank Road	Gaging station at Yaphank	At Route 27
1947				
Aug. 2	0.25		16.3	
1948				
Feb. 25	.77		22.6	
Mar. 26	1.68	15.3	25.6	
Apr. 17	2.05		23.7	
Oct. 7	.85		21.8	
15		5.84	19.4	
1949				
Jan. 11	2.98		31.0	
Apr. 19	3.50		30.5	
Aug. 11	1.58		23.8	
1950				
Apr. 25	.55		15.6	
Aug. 23		4.12	16.8	
1951				
Feb. 24	.46		19.2	
June 13	.80		17.8	
July 27	.29		12.8	
30	.39		18.9	
1952				
Jan. 18			23.3	[1] 58
30			23.2	[1] 75
Feb. 28			23.0	[1] 53
Apr. 26			26.5	[1] 65
July 29			22.0	[2] 54.4
Aug. 19	1.90	10.5	24.6	
Oct. 2			21.8	[3] 54.8
Dec. 10	1.89	9.35	21.9	
1953				
Feb. 25			27.4	[4] 51.7
Mar. 11	1.96	10.7	24.9	
Apr. 22			33.5	[4] 75.0
June 11	3.92	15.2	32.0	
24			26.5	[4] 53.7
July 28			25.5	[4] 47.6
Aug. 28			23.6	[4] 55.9
Sept. 18			22.8	[4] 46.9
22	1.33	9.65	22.6	[4] 56.1
Oct. 3			21.4	
Nov. 14			22.1	[4] 72.7
24	1.90		30.8	
Dec. 12			24.5	[4] 75.5

[1] Measurement not corrected for changing stage or pondage upstream to C–11; was corrected for diversion of 6 cfs.
[2] On the basis of two measurements—one on falling, and the other on rising stage—as corrected for diversion and pondage upstream to C–11.
[3] On the basis of one measurement corrected for changing stage, diversion, and for pondage upstream to C–11.
[4] On the basis of one measurement corrected for changing stage and for diversion, but not for pondage upstream to C–11.

The second section of the stream lies between Bartlett Road and the dam just above Route 27, 6.2 miles downstream, on the outwash plain south of the Ronkonkoma moraine, which it has slightly dissected. The streamflow at Route 27, just below the dam, was measured on July 29, 1952, at a time when the flow was probably close to average. At this point the stage and flow of the stream are somewhat affected by the tides in Bellport Bay, and corrections for this and other factors were required. From 9:13 a.m. to 10:34 a.m., when the stage was falling because of the falling tide in the bay, the discharge was 52.8 cfs. From 12:54 p.m. until 2:13 p.m., when the stage was rising, the discharge was 36.5 cfs (fig. 38). After corrections for changes in pond and bank storage, the average discharge during this period of normal streamflow was computed to be 47.8 cfs. Partial measurements on one or two other occasions tended to confirm this figure.

The apparent gain in streamflow between the gaging station at Yaphank and the bridge on Route 27 is, therefore, the difference between 47.8 and 21.76 cfs, or about 26 cfs, but two small corrections must be made. About 6.1 cfs was being diverted through the Carmans River Duck Farm upstream from the highway bridge, and the rising change in stage of several small ponds in the Suffolk County Game Preserve represented the holding back of about 0.55 cfs, a total ungaged flow of about 6.65 cfs. The corrected flow at the bridge, therefore, is about 54.5 cfs, and the gain in flow in the 2¾ miles of river upstream is between 32 and 34 cfs.

The third section of the river, from the highway bridge to the mouth of the river at Sandy Point, a distance of about 3.15 miles, crosses the outwash plain south of the Ronkonkoma moraine. In this section, how-

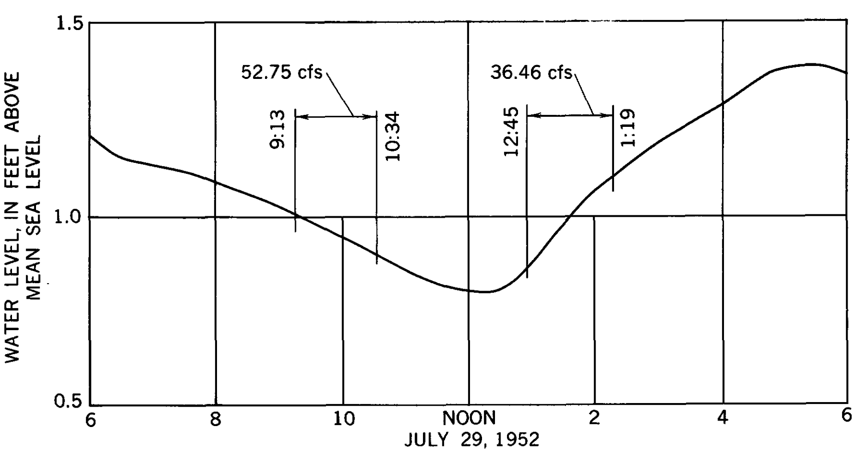


FIGURE 38.—Fluctuation of Carmans River at Route 27 on July 29, 1952.

ever, the river is a tidal estuary; it has been aggrading its bottom and its small flood plain. Tidal fluctuations in Bellport Bay and Great South Bay are the main cause of variations in stage of this part of the river which, at the bridge on Route 27, varies from a maximum of 3.82 feet above sea level, to a minimum of 0.34 foot above sea level. The average daily range at this point is 0.83 foot.

The discharge of the stream at its mouth could not be measured, but the flow at this point has been estimated to be 72 cfs.

RELATION OF STREAMFLOW TO GROUND-WATER CONTRIBUTING AREA

The average discharge of the Carmans River at the gaging station in Yaphank for the period of record through September 1953 is 21.8 cfs, and the area contributing ground water to the stream, as determined from the water-table map, is 21.5 square miles. The average runoff thus is the equivalent of 13.5 inches of water. During these years the rainfall averaged 43.5 inches; because about 22 inches was lost by evapotranspiration, the recharge to the water table must have averaged 21.5 inches. The difference between this recharge and the 13.5 inches of streamflow, or about 8 inches, probably represents recharge to the deeper aquifers, the Magothy(?) Formation and the Lloyd Sand Member of the Raritan Formation.

The measured discharge of the river at the bridge at Route 27, at a time of probable near average flow, was between 54 and 55 cfs. The water-table map shows a contributing area of 36.5 square miles, which represents an average annual runoff of about 20 inches, or 6.5 inches more than at Yaphank, and an amount only slightly less than the average annual recharge. The increase in flow in the 2.75 miles of stream between Yaphank and the bridge is 32 cfs and the contributing area is about 15 square miles; runoff for this area is therefore about 30 inches, or somewhat more than the recharge. The excess over the recharge is due to upward leakage from the aquifers below the Gardiners Clay, but, because the average flow for the total area of some 36.5 square miles contributing to the flow at the highway bridge is only about 20 inches, it is apparent that not all the deeper recharge has come back to the water-table aquifer at this point. There must be considerable additional upward leakage into the area south of Route 27, and the estimated discharge of 72 cfs for the mouth of the Carmans River at Sandy Point is based on the estimate that the 48 square miles of area furnishing this flow contributes an average 22 inches of runoff.

RELATION OF STREAMFLOW TO GROUND-WATER LEVELS

Because the streams are almost entirely supplied by ground water discharge, a close correlation exists between the height and slope of the water table and streamflow (pl. 7). The relation is, however, not always simple or direct, and the seasonal high or low in a particular observation well may come earlier than, at the same time as, or later than the seasonal high or low discharge at a point on a neighboring stream. The more important factors influencing this relation are (1) the depth to the water table at the well site, (2) the position of the well in the pattern of ground-water flow, (3) the distance of the well from the stream, (4) the hydrologic characteristics of the aquifer in question, (5) the hydraulic gradient, and (6) the variations in the amount of water in storage in the aquifer.

Well S3533 (pl. 1) is about 1 mile east of the nearest point on the Carmans River, and about 3.2 miles N. 16° W. of a point on the river at Bartlett Road, on the ground-water flow line passing through the well. By trial and error a good correlation was found between the water-level stage in this observation well and the flow of the river at the gaging station in Yaphank, although the high and low stages in the well lag about 6 weeks behind the corresponding high and low flow in the river. This relation is shown in figure 39, and is used in figure 40 to calculate streamflow from the well hydrograph.

The derived value for streamflow is within a few percent of the gaged value, except for periods such as August 1947 and November 1952 when there was long, intense rainfall. A somewhat similar empirical relation can be worked out between streamflow and the water level in almost any nearby observation well, whether it is in the area contributing ground-water flow to the stream or not, because both are strongly influenced or even controlled by the cumulative recharge

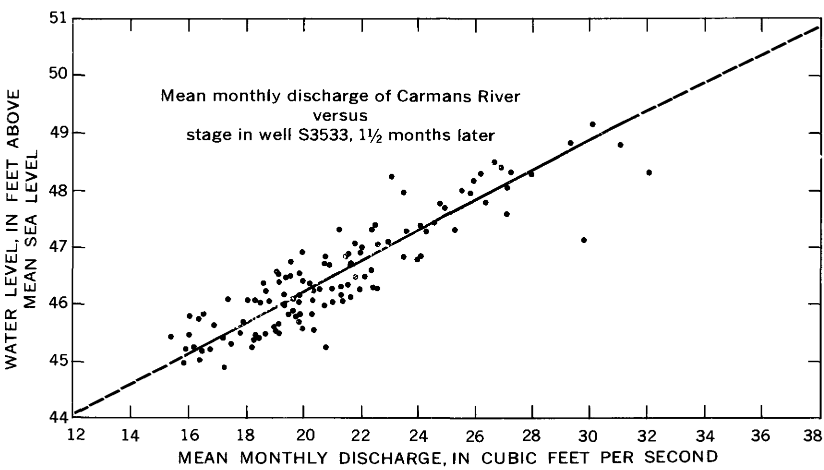


FIGURE 39.—Relation of streamflow in Carmans River to water level at well S3533 from October 1942 to September 1952.

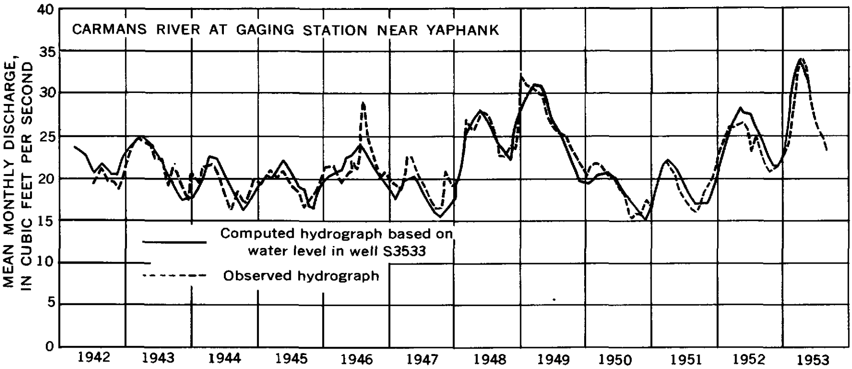


FIGURE 40.—Comparison of observed and computed monthly streamflow of the Carmans River at Yaphank.

to the water table. For example, the water level in well S3532, about 3.5 miles northeast of Artist Lake and 0.5 mile north of the ground-water divide, can be closely correlated to the flow of the Carmans River, although in this case the stage of the well lags about 2.5 months behind the streamflow.

FORGE RIVER

The Forge River is a somewhat smaller stream, immediately to the east of the Carmans and south and southeast of the Laboratory tract. Its topographic drainage area of 16 square miles lies entirely south of the Ronkonkoma moraine. The river may conveniently be divided into three parts; an east branch, about 0.7 mile long, a west branch, 1.1 miles long, and, south of Route 27 where these two branches join, a section of tidal estuary about 0.8 mile long. Both branches have been deepened, widened, extended, and ponded for duck farming, and the measured minimum and maximum flows of the branches, and of their confluence just south of Route 27, are in a large part determined by artificial control. The average slope of the water surface in the two branches is about 15 feet per mile; below the confluence, the water surface in the estuary is nearly horizontal and rises and falls with the tide. Just north of Route 27, culverts with stop-logs form a pond of about 15 acres in the west branch and a pond of about 5 acres in the east branch. These ponds were in existence at least as far back as 1907.

Miscellaneous discharge measurements in the east branch and in the west branch just north of the highway, and of their combined flow just south of the highway, were made in 1907 (Spear, 1912) and by the U.S. Geological Survey from 1947 to 1953 (tables 13, 14). The combined flow ranged from about 7 to 13 cfs, the west branch contributing more than half the discharge. The average flow was 8 cfs.

TABLE 13.—*Discharge data for 1907, in cubic feet per second, west branch of the Forge River at Highway 27 in Moriches, N.Y.*

Date	Discharge	Date	Discharge
Miscellaneous measurements			
January 11	6.8	March 12	7.0
17	5.9	25	5.4
February 18	7.3	30	6.5
March 1	7.0		
Discharges obtained by water-stage recorder			
[Highest, mean, and lowest discharge (in that order) shown for each month. Highest and lowest discharges for July appear to be the result of regulating outlet controls for pond]			
July	9.9	October	6.2
	6.5		5.3
	2.5		4.6
August	7.0	November	9.3
	6.0		5.9
	5.4		5.1
September	7.0	December	7.3
	5.9		5.7
	5.0		5.0

TABLE 14.—*Miscellaneous discharge measurements for 1947–53, in cubic feet per second, Forge River and its two branches at Highway 27 in Moriches, N.Y.*

Date	West branch	East branch	East and west branches
1947			
Nov. 22			7.28
Dec. 20			6.92
1948			
Jan. 28			7.45
Feb. 27			9.67
Mar 26			9.24
Apr. 17			10.9
1949			
Jan. 11			10.2
Mar. 29			11.4
1950			
Apr. 6			7.72
1952			
Jan. 18	[1] 5.5	[1] 3.6	9.1
Mar. 26	[1] 6.4	[1] 4.1	10.5
Apr. 11	[1] 6.2	[1] 4.0	10.2
29	[1] 7.1	[1] 4.1	11.2
June 2	[1] 7.1	[1] 4.3	11.4
July 29	[1] 5.8	[1] 3.8	9.6
Aug. 20			13.1
Oct. 2	[1] 5.6	[1] 3.4	9.0
30	[1] 4.9		
Nov. 26	8.17	4.18	12.4
Dec. 9	[1] 4.1		
1953			
Mar. 12	7.21	4.09	11.2
June 24	7.88	4.28	12.2
Sept. 18	5.90	3.36	9.26
Nov. 21	5.43	2.89	8.32

[1] Obtained by determining head on spillway weir and computing discharge by the Francis formula.

The topographic drainage area above Route 27 is 8.5 square miles, and the ground-water contributing area, as determined from Spear's map 1912 of the water table, was about 7 square miles (fig. 37). However, only a small part of the two areas coincide; the topographic drainage area extends about equally east, west, and north of the confluence of the two branches, whereas the ground-water contributing area, the true source of the water in the river, forms a long, narrow teardrop-shaped area extending to the north as far as the ground-water divide. (See pl. 7.)

The average discharge of 8 cfs for the 7 square miles of ground-water contributing area represents a runoff of about 16 inches. This amount is about 6 inches less than the ground-water recharge, although Route 27, at this point, lies south of the line of reversal of leakage downward to leakage upward into the upper Pleistocene deposits. To the north, therefore, in the central part of the Laboratory tract, the downward leakage to the aquifers below the Gardiners Clay, must be more than 6 inches per year.

PECONIC RIVER

In a general way, the Peconic River may be said to drain the area lying between the Harbor Hill moraine on the north, the Ronkonkoma moraine on the south, the Carmans River watershed on the west, and Peconic Bay, into which it flows, on the east (pl. 6). In detail the drainage area of the Peconic is more complex than these boundaries would suggest, not only because the ground-water contributing area rather than the topographic drainage area must be considered, but also because of other factors peculiar to the upper Peconic. In certain areas around its headwater, which include parts of the northern and eastern sections of the Laboratory tract and an area immediately to the east, the Peconic flows eastward although the gradient of the water table is to the southeast. This difference in direction of flow is the result of, or at least is accentuated by, the presence in these areas in the upper 30 feet or so of the upper Pleistocene deposits of thin discontinuous lenses of relatively impermeable silt and clay. These fine-grained beds were formed at the close of the glacial period by mud and silt deposited in swamps or lakes in the low-lying land between the two moraines, or by the reworking of loessal deposits in this area. The beds were penetrated by some of the test wells, but their extent is largely inferred from hydrologic data. Locally the beds form small areas of perched water when the water table is low, and some of this perched water drains into the Peconic even while the main body of water below the water table is draining southeast into the Forge River or into Moriches Bay. When the water table is high, the fine-grained beds are largely below the water table, but they still deflect a substantial part of the shallow ground-water flow into the Peconic even while the deeper flow is moving southeast.

The topographic drainage area of the Peconic River may be divided into four parts. The first lies north of the main ground-water divide and is drained entirely by ground-water flow north to Long Island Sound. The second, which comprises the western part of the Laboratory tract and a narrow area just west of the Laboratory, is largely drained by ground-water flow to the Carmans River or to Bellport Bay, although heavy springs rains, when the water table is high, may cause some surface runoff into the upper Peconic. The third area includes the central and eastern parts of the Laboratory tract and some adjacent land lying south and east of the Laboratory. Drainage from this area is largely ground-water flow south to the Forge River and Moriches Bay, although, except for the summer months when the water table is low, some shallow ground water makes its way into the Peconic River.

East of the Laboratory tract the main ground-water divide extends eastward to Orient Point and a subsidiary divide develops and extends first to the southeast and then eastward to Montauk Point (pl. 7). The complexity of the hydrologic conditions in the area precludes an exact determination of the point where the subsidiary ground-water divide begins. It probably varies somewhat with changes in the height of the water table. The fourth area lies east of the Laboratory tract and between these two ground-water divides. Virtually all drainage from this area finds its way into the Peconic River.

The channel of the Peconic River may be divided into two main segments. The lower segment, which extends about 9 miles from the Wading River-Manorville Road to the mouth of the river, is marked by a series of artificial ponds and lakes, a few of which were originally built to supply water power for small mills but now retained for esthetic or recreational reasons. This segment of the river lies within and traverses the area which contributes all its drainage to the Peconic River; that is, the area in which the water table always slopes toward and contributes water to the river. Indeed, it is only under such conditions of flow that it would be practical to artifically pond the river.

The flow, as measured at the Wading River-Manorville Road, the dividing point between the upper and lower Peconic, varied from 27.6 to 1.0 cfs, and averaged about 3 cfs. The area contributing ground water to the upper Peconic upstream from this point is variable and difficult to judge, but the maximum potential area is about 8–10 square miles. The average drainage from 1 square mile on Long Island is about 1.4 cfs (1 mgd) so that the river at this point actually receives only a small part of the potential drainage, the major part going by way of ground water flow to the southeast. In other words, the area of the upper Peconic is largely drained by ground-water flow, and the streamflow varies widely depending on the height of the water table and the amount and intensity of the recent rainfall.

Since 1942 the U.S. Geological Survey has operated a stream-gaging station on the lower Peconic River, about 0.4 mile west of Riverhead, just upstream from the point where the effects of the tide would make gaging virtually impossible. The average monthly flow at the station has varied between a maximum of 84.0 cfs and a minimum of 12.7 cfs, and has averaged 30.5 cfs, or about 20 mgd (fig. 41). The ground-water contributing area upstream from the gage is estimated to vary between 22 and 23 square miles, but may reach as much as 28 square miles for brief periods. The apparent topographic drainage area upstream from the gage is about 75 square miles. The Peconic River, like the Carmans and the Forge, is shown by a study of the water-table contour map to carry the runoff from an area much smaller than that suggested by the topographic map.

The upper segment of the river begins rather indefinitely in a poorly drained area just west of the northwest corner of the Laboratory and flows intermittently and irregularly a distance of 6.5 miles to the upper end of the lower segment (pls. 3, 9, 10). This part of the river has largely been ditched to improve its flow as a mosquito-control measure, but the flow is still irregular and, in detail, quite complex. Indeed to describe this segment of the upper Peconic it must be subdivided into six parts. Of these, three represent subdivisions of the main channel of the upper Peconic, and three are small streams tributary to the upper Peconic (pl. 6).

The first of these six parts is the initial 2 miles of the main channel starting at a variable and uncertain point just west of the Laboratory and extending across the Laboratory tract just south of its northern boundary to the upper of two Health Physics weirs established for monitoring purposes in March of 1951. This section of the river, if such it may be called, is a ditch, largely dug by hand during World Wars I and II when the present Laboratory was a military camp. The land on either side of the ditch is still poorly drained and contains, even in dry weather, scattered small swamps and ponds. The ditch, however, is dry much or even most of the time because, as the few scattered observation wells nearby show, the main water table is below the level of the bottom of the ditch virtually all the time. However, in wet weather after a heavy rain the swampy area immediately west of the Laboratory contributes a substantial flow to the ditch, which reached as much as 6 cfs in March of 1953. At that time, because the water table was still low, much of the flow in the ditch was

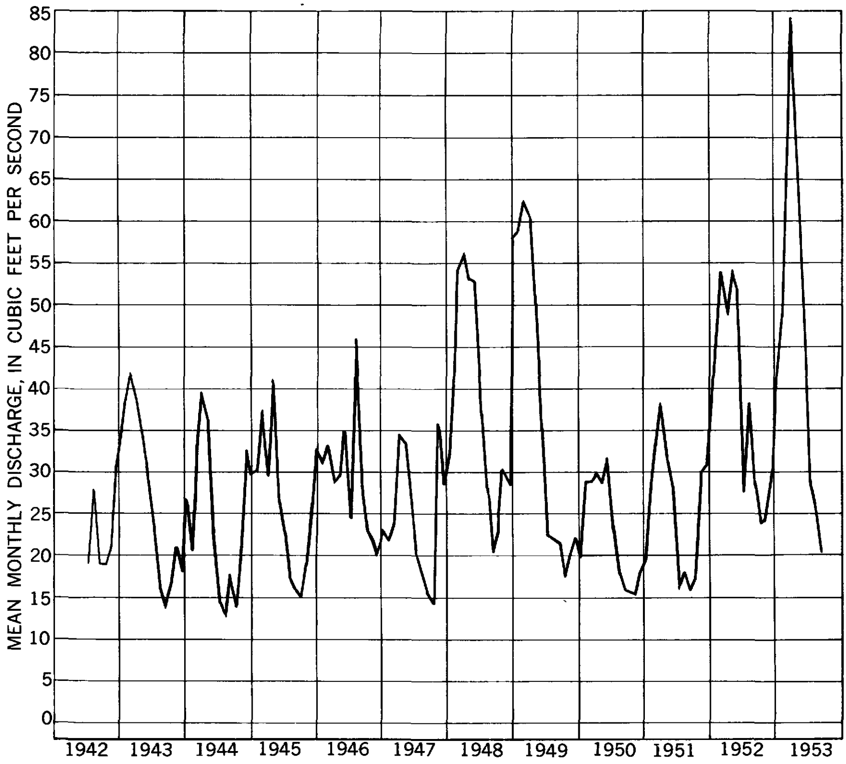


FIGURE 41.—Mean monthly discharge of the Peconic River near Riverhead.

lost to the water table, and the flow over the weir at the lower end of this 2-mile stretch of the river was only 0.67 cfs. In May 1953 when the water table was unusually high, about 3 cfs entered the upper end of the ditch at the west boundary of the Laboratory, little water was lost to the water table, and the average flow over the weir was about 3 cfs. In general even when the swampy area just west of the Laboratory, formed by a local area of perched water table, is feeding water into the upper end of the ditch, much or even all of this flow is lost to the water table from the mile and a half of ditch within the Laboratory tract. Only during periods of exceptionally high water table is this flow maintained without loss down to the upper weir.

During the period of record at the upper weir, the weir was dry only for relatively short periods in late summer or early fall of each year. This statement would seem to contradict the one just made; however, an exceptional short stretch of ditch and swamp immediately above this weir was dug deeper than most and is commonly at or below the main water table. Consequently, it provides some local flow over the weir even when most of the rest of the ditch is dry.

The second subdivision of the upper Peconic extends south-southeast from the upper weir to a point near North Street a half mile east of the east boundary of the Laboratory (pl. 9). Downstream from this point the river makes a sharp bend first to the north and then to the east. This part of the river has also been dug out to improve the drainage, but it appears less artificial than the ditch upstream from it. At the upper end of this stretch, just below the upper weir, the river receives the discharge from the Laboratory sewage-treatment plant, which averages around 0.2–0.3 cfs, except after periods of heavy rain and high water table, when it may receive more than 0.5 cfs. This artificial discharge into the river is sufficient to maintain at all times some flow to and beyond the east boundary of the Laboratory. Just inside the east boundary is the second (lower) Health Physics weir, and during the 3 years of record available, the minimum monthly average flow (September 1951) was 0.17 cfs. However, during many months there is a loss in flow in the river between the point where it receives the discharge from the sewage-treatment plant and the east boundary of the Laboratory, and during the late summer and early fall of many years this stretch of the river would be dry were it not for the water artificially added to it.

East of the Laboratory, as mentioned above, the Peconic River swings south for a distance of about three-quarters of a mile to a point near North Street (the North Road), where it turns north. Although little quantitative information is available as to gain or loss of flow in this part of the river, qualitatively it is in general a losing stretch because near North Street the river is frequently dry in late summer and early fall. The height of the water table and the continuity of flow in the Peconic southeast of the Laboratory are related. If the water table, as measured in observation wells at the east boundary of the Laboratory, is less than a foot below the stream level, then the flow will be continuous around the bend at North Street and on down to the mouth of the river. If the difference in levels is between 1 foot and 1½ feet, the flow may be continuous, or it may be interrupted by a short section of dry channel near North Street. If the difference in levels is more than 1½ feet, the flow of the river is almost certain to be interrupted.

The third section of the upper Peconic extends from North Street to the Wading River-Manorville Road, a distance of about 2 miles. Even in the driest seasons flow reappears in this part of the channel at some point between North Street and Schultz Road. Records of flow at Schultz Road vary from 0.19 to 12.3 cfs, although only a relatively few scattered readings are available. At the Wading River-Manorville Road, a little over a mile downstream from Schultz Road, the recorded values for the flow vary from 1.0 to 20.0 cfs and are in general about twice the flow recorded at Schultz Avenue. The third section of the upper Peconic, therefore, between North Street and the Wading River-Manorville Road, unlike the upper two sections, is in general a gaining segment of the river.

The three remaining subdivisions of the upper Peconic are three small unnamed tributaries, which for the purposes of this report, may be called the south branch, the north branch, and the southwest branch. All three tributaries are dry in the late summer or early fall of many years, although even then there is some standing water locally in their channels or in adjacent small lakes or swamps.

The south branch originates at a point just southeast of the bend in the main channel of the Peconic near North Street, flows southeast for about half a mile, then turns northeast, and joins the main channel of the Peconic just west of the Wading River-Manorville Road. In part the south branch consists of large meanders far too big for any present flow of this tributary; it probably represents an abandoned former channel of the Peconic. No measurements of flow have been made, but the south branch is known to be stagnant much of the time and the flow probably never exceeds 1–2 cfs, except for rare brief periods.

The north branch originates near Route 25, a little more than a mile northeast of the Laboratory, and flows south to join the Peconic just west of Schultz Road. Its course is marked by numerous small ponds, probably kettle holes. Just before joining the Peconic it divides into three channels. During average conditions there is, apparently, no flow in the west or east channels and at times even the central channel is dry or stagnant. The average flow of the north branch is estimated to be about 1 cfs, and the maxium measured flow of the central and west channels combined was about 8.9 cfs.

The southwest branch originates near the southwest corner of the Laboratory tract and flows north and northeast to join the Peconic just east of the eastern boundary of the Laboratory. Its channel is complex and consists largely of ditches which loop and branch as the result of attempts to drain the area. The ditches are dry much of the time or locally hold a little standing water. They flow after a heavy rain if the water table is high, but in some sections, and particularly in one large loop, the flow may be in either direction. This ambiguity in direction of flow results from the very low gradient of the ditches and from the large proportion of their water which they lose to the water table, sometimes in one area, sometimes in another. On the basis of several miscellaneous measurements the average flow of the Southwest Branch is estimated to be about 0.2 cfs, or about 130,000 gpd.

In considering the possible movement of a contaminant which might find its way into the upper Peconic River, it is of considerable importance that much of this stream much of the time is losing water to the main water table. The flow of the upper Peconic is derived largely from smaller or larger areas of perched water, from storm rainfall, and from the sewage-plant effluent. Perhaps 25 percent of the time the river channel is dry or at least stagnant where it rounds the big bend near North Street, a half mile or so east of the east boundary of the Laboratory. During such periods it would obviously be impossible for the stream to carry contamination into the lower Peconic and into Peconic Bay because all the flow goes to join the ground water and then moves southwest into Forge River and Moriches Bay. During much of the time, although some of the streamflow is lost to the ground water within the Laboratory area and near the North Street bend, some flow continues down to the mouth of the river. Under such conditions a contaminant would be divided between the surface-water flow and the ground-water flow. In times of high water table and large streamflow, relatively little water is lost to the water table and virtually all the water goes to Peconic Bay. Despite the ponding of the lower Peconic, the rate of movement of surface water in the Peconic is very much more rapid than the movement of the ground water. The travel time of the surface water to Peconic Bay is probably a few weeks; the travel time of the ground water to Moriches Bay is 50–100 years.

LOCAL SURFACE DRAINAGE IN THE LABORATORY TRACT

Except for certain relatively small areas adjacent to the upper Peconic and its southwest branch, surface drainage is relatively unimportant in the Laboratory tract, or in general in eastern Long Island (pl. 9). Most precipitation infiltrates where it falls and joins the ground water. However, in two additional small areas within the Laboratory tract there is some movement of storm rainwater over the land surface for short distances when the ground is relatively impermeable, due to frost, or is briefly saturated. This overland flow is of little hydrologic significance, but it might affect the distribution of contamination from a local incident.

The first is the general hospital and medical area in the southwest corner of the Laboratory tract where drainage from roofs, roads, and parking areas is concentrated into dry wells or pits, or into certain sodded or wooded areas by drains or small ditches. This drainage does not leave the Laboratory tract by surface flow, but the subsurface recharge is localized by these artificial structures.

In the south-central part of the Laboratory tract, the shop and warehouse area in the general vicinity of Cornell and Bell Avenues and Railroad and Seventh Streets has a concentration of roof and road surfaces that, at times, is responsible for local surface runoff. A system of drainage ditches and dry wells was constructed in this area in 1943. The flow is to the south along two parallel main ditches which end near the center of the south boundary of the Laboratory. From here the water spreads out over an area extending a few hundred yards south of the Laboratory where it infiltrates. The area drained is about 300 acres, and the flow crossing the Laboratory boundary after a 2-inch rain under favorable conditions may total as much as 100,000 cubic feet.

LABORATORY WATER SUPPLY AND SEWAGE DISPOSAL SYSTEM

Water supply for the Laboratory comes from three wells screened in the outwash of the upper Pleistocene deposits. Wells 1 and 2 were installed during the early years of World War II when the area was reactivated as a military camp, and well 3 was installed by the Laboratory in 1948. All three are equipped with deep well turbine pumps and are operated alternately one or two at a time to maintain the water level in an elevated storage tank.

Well 1 (S2476), on the east side of Railroad Street about 200 feet north of Brookhaven Avenue, is 12 inches in diameter and has 15 feet of 30-slot 11½-inch screen set between 86 and 101 feet below land surface. When completed in February 1941, it pumped 562 gpm at a 15-foot drawdown (specific capacity 36.4). When tested in 1948 it pumped 526 gpm at a drawdown of 24.5 feet (specific capacity 21.5).

Well 2 (S3197), at the southeast corner of Upton Road and Cornell Avenue, is 12 inches in diameter and has 20 feet of 40-slot 8-inch screen set between 115 and 135 feet below land surface. When tested in 1950 at 462 gpm it had a specific capacity of 9.4.

Well 3 (S6697), in the boiler house about 75 feet north of Temple Place and about 700 feet east of Railroad Avenue, is 12 inches in diameter and has 26 feet of 11¼-inch screen set between 75 and 101 feet. In 1951 this well pumped 700 gpm at a drawdown of 24 feet (specific capacity 29).

In 1951 the water pumped averaged 544,000 gpd; in 1952 it averaged 627,000 gpd. About 28 percent came from well 1, 25 percent from well 2, and 48 percent from well 3. About two-fifths of this water is discharged to the upper Peconic after it passes through the Laboratory sewage disposal system. The balance (1) is lost to the ground through leaks in the sewer system, (2) is lost to the ground through cesspools or septic tanks that serve several small isolated buildings, (4) is used to water the lawns or for irrigation by the Biology Division, (5) or is used for makeup at the evaporation towers which cool the air from the graphite reactor. The sewer system comprises some 12–15 miles of vitrified pipe, 8–30 inches in diameter, with cemented joints. The high point, in the hospital area, is about 100 feet above sea level, and the low point, at the disposal plant, is about 60 feet above sea level. Even the low end is well above the main water table so that there is considerable leakage to the water table through the joints. Only about 41 percent of the water pumped by the supply wells reaches the disposal plant. Because the other losses are small, most of the loss is from the sewer pipes. However, after heavy rains, considerable perched water leaks into the mains of the sewage-disposal system.

The disposal plant consists of a grit chamber, an Imhoff tank, dosing tanks, six principal slow sand-filter beds, and two spare filter beds. The principal filter beds are underlain by tile drains, about 8 feet above the water table. Just below the tile drains is one of the thin relatively impermeable layers of silt and clay which characterize and complicate the hydrology of the upper Peconic. This bed impedes downward flow below the tile drains, so that the drains intercept and carry to the Peconic 79 percent of the flow which reaches the treatment plant. In all, therefore, about 200,000 gpd of the 600,000—700,000 gallons pumped, reaches the upper Peconic River. As we have seen above, much or even all of this 200,000 gallons may infiltrate down to join the ground water between the outlet from the drains from the sand filters and the point where the Peconic is crossed by Schultz Road; about 75 percent of the time, however, some of this effluent will travel all the way down to Peconic Bay.

It is almost certain that from time to time very small amounts of radioactive contaminants (much too small to represent any possible hazard) will inadvertently be introduced into the sewer system in one of the laboratory or hospital areas. About half of this will find its way to the sewage-disposal plant and about half will leak out of the sewer pipes at various points along the way. Separation of solids in the Imhoff tank and in the sand filters, and adsorption on the filters will remove a substantial proportion of the radioactive contaminants reaching the disposal plant, and the remainder, upon reaching the upper Peconic, will be further reduced by dispersion to the ground water and by dilution in the ponded reaches of the lower Peconic. In the event, therefore, of contamination of the Laboratory sewage-disposal system it is certain that dispersion, dilution, and retention within the system will very greatly reduce any possible contamination of the ground water and surface water outside the Laboratory. The complexities of the system afford an important factor of safety, but in the highly speculative event of massive high-level contamination of the system, they would also greatly complicate problems of monitoring and remedial action.

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[Letters designate the separately published chapters]

(A) Geology of Brookhaven National Laboratory and Vicinity, Suffolk County, New York, by Wallace de Laguna.

(B) Physical Properties and Mineralogy of Selected Samples of the Sediments From the Vicinity of the Brookhaven National Laboratory, Long Island, New York, by George T. Faust.

(C) Hydrology of Brookhaven National Laboratory and Vicinity, Suffolk County, New York, by M. A. Warren, Wallace de Laguna, and N. J. Lusczynski.

(D) Chemical Quality of Water, Brookhaven National Laboratory and Vicinity, Suffolk County, New York, by Wallace de Laguna.

(E) A Hydrologic Analysis of Postulated Liquid-Waste Releases, Brookhaven National Laboratory, Suffolk County, New York, by Wallace de Laguna.