Geohydrology of the Artificial-Recharge Site at Bay Park, Long Island, New York

GEOLOGICAL SURVEY PROFESSIONAL PAPER 751-C

Prepared in cooperation with the Nassau County Department of Public Works

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By JOHN VECCHIOLI, G. D. BENNETT, F. J. PEARSON, JR.,
and L. A. CERRILLO

DEEP-WELL ARTIFICIAL-RECHARGE EXPERIMENTS AT BAY PARK, LONG ISLAND, NEW YORK

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GEOHYDROLOGY OF THE ARTIFICIAL-RECHARGE SITE AT BAY PARK, LONG ISLAND, NEW YORK

By JOHN VECCHIOLI, GORDON D. BENNETT, FREDERICK J. PEARSON, JR., and LAWRENCE A. CERRILLO

ABSTRACT

Reclaimed water (highly treated sewage-plant effluent) is being injected into a 480-foot-deep well at Bay Park, Long Island, N.Y., as part of a cooperative experimental study by the U.S. Geological Survey and the Nassau County Department of Public Works. Before the recharge experiments were started, the local geology, well and aquifer hydraulics, and chemistry of the native water were studied to define the background against which to measure the effects of injecting the reclaimed water. Results of these studies are presented in this report.

The recharge well is screened in the lower part of the Magothy aquifer of Late Cretaceous age. Because the Magothy, the principal source of ground water in most of Long Island, is hydraulically interconnected with bodies of salty surface water, salt-water encroachment is a great concern. A search for methods with which to manage this problem prompted the cooperative study of recharging the aquifer with reclaimed water. At Bay Park, the Magothy aquifer is confined below by the clay member of the Raritan Formation, also of Late Cretaceous age. The aquifer is largely unconfined above, owing to the generally coarse-grained texture of the overlying Pleistocene deposits. The injection zone is a 60-foot interval of slightly silty fine to medium sand beds that lie between Magothy beds of lower hydraulic conductivity. Average lateral hydraulic conductivity of the stratified 60-foot injection zone is estimated to be 940 gallons per day per square foot (126 feet per day), but flowmeter surveys indicate considerable variation in lateral hydraulic conductivity within this interval. Vertical hydraulic conductivity of the material between the water table and the top of the injection zone ranges from 2 to 20 gallons per day per square foot (0.27 to 2.7 feet per day). The hydraulic characteristics of the hydrologic system were determined by standard aquifer-test methods and were later verified by electric-analog studies.

Water from the Magothy aquifer has an unusually low dissolved-solids content because of the lack of readily soluble minerals in the aquifer deposits. At Bay Park, the dissolved-solids content is about 25 milligrams per liter. Most of the specific chemical components are those present in precipitation, the source of natural recharge. Dissolved silica, the single most abundant dissolved constituent in water from the Magothy aquifer, results from the solution of quartz, the dominant mineral in the sand that constitutes most of the aquifer.

INTRODUCTION

PURPOSE AND SCOPE

The U.S. Geological Survey in cooperation with the Nassau County Department of Public Works is continuing a study of hydrologic and geochemical problems connected with injection of reclaimed water into a deep well at Bay Park, Long Island, N.Y. Although this work is part of a larger study to evaluate the feasibility of the use of barrier recharge wells to prevent salt-water encroachment, the work is not concerned with the effect of recharge on the salt-water front. Rather, it is restricted to a study of the mechanics of the recharge process. Pilot-plant experimentation is underway at the Nassau County sewage-treatment plant in Bay Park, Long Island, N.Y. Bay Park, on the northern shore of Hewlett Bay in southwestern Nassau County (fig. 1), is about 3 miles north of the southern shore of Long Island.

Early work in the project included construction of the experimental recharge well and a series of observation wells, installation of injection equipment, and instrumentation of observation wells and injection equipment. Lithologic core samples collected during drilling of the wells and geophysical logs of the wells were studied to define the hydrogeologic framework at the recharge site. Aquifer tests were made to determine the hydraulics of the wells and the aquifer. These tests were later supplemented with electric-analog studies. Hydrochemical studies were made to define the geochemical environment of the aquifer. This report presents the results of geologic, hydrologic, and hydrochemical studies that

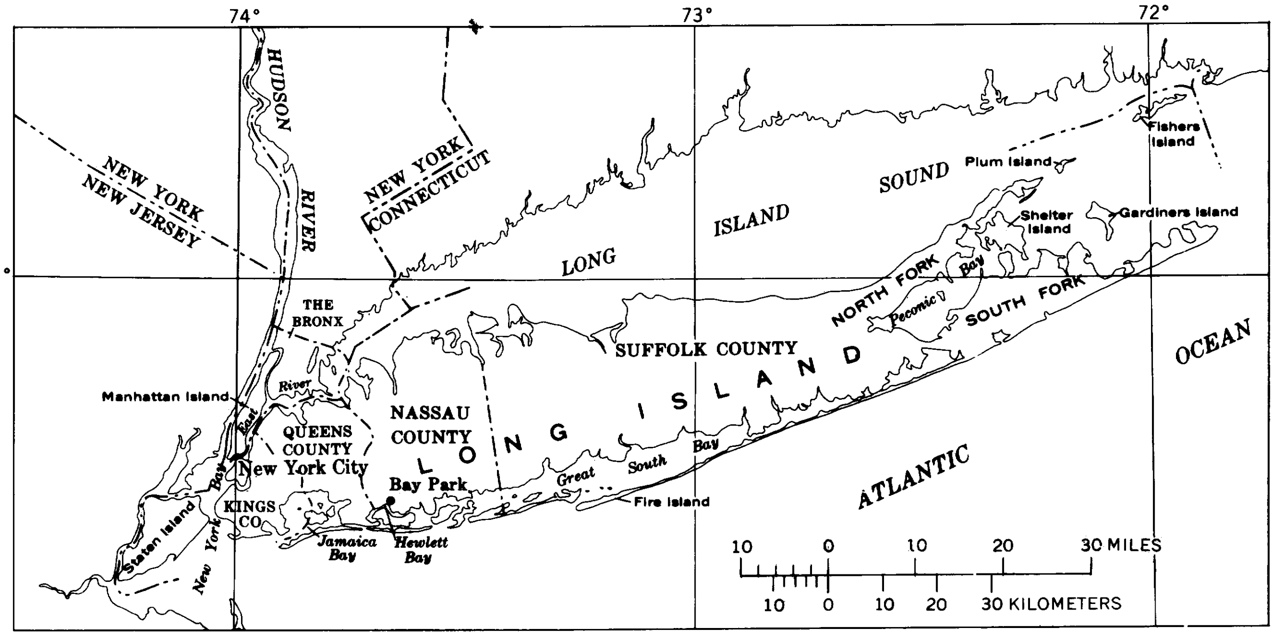

FIGURE 1.—Location of the Bay Park artificial-recharge site.

were done to define conditions prevailing before the recharge tests were started.

SYNOPSIS OF WATER-RECLAMATION AND INJECTION FACILITIES

Reclaimed water for recharge is obtained by tertiary-stage treatment (Peters and Rose, 1968; Peters, 1968) of about 0.6 mgd (million gallons per day) of the effluent from an activated-sludge type 60 mgd sewage-treatment plant. This treatment consists of coagulation and sedimentation, primary filtration through a dual-media sand-anthracite filter followed by secondary filtration through one to four activated carbon columns, and, finally, chlorination. The reclaimed water produced by this process meets commonly accepted drinking water standards (U.S. Public Health Service, 1962). Additional treatment includes degasification, pH adjustment, and dechlorination, which can be applied at the recharge facility.

Major components of the recharge facility include (1) a 50,000-gallon storage tank into which either public-supply water or reclaimed water can be delivered, (2) a vacuum-operated degasifier tank, (3) injection and redevelopment pumps with automatic-flow controls, (4) a recharge well consisting of an 18-inch fiberglass casing above a 16-inch stainless-steel well screen set at 418 to 480 feet below land surface, (5) numerous observation wells equipped with water-level recording equipment, and (6) equipment for monitoring some of the chemical and the physical characteristics of the water.

Locations of observation wells in reference to the recharge well are shown in figure 2. Pertinent construction details are given in table 1.

PREVIOUS WORK

The first report on regional geology and hydrology of Long Island was prepared by Veatch, Slichter, Bowman, Crosby, and Horton (1906). Fuller (1914) prepared a more detailed geologic report. An extensive compilation of well records and a detailed description of the subsurface geology of Long Island were given in a report by Suter, de Laguna, and Perlmutter (1949). Perlmutter and Geraghty (1963) discussed the geology and the ground-water conditions in southern Nassau and southeastern Queens Counties, including the Bay Park area. Lusczynski and Swarzenski (1966) gave a detailed description of the hydrologic environment in southeastern Nassau County. General discussions of recent regional hydrologic conditions are given by Cohen, Franke, and Foxworthy (1968).

The water-treatment and the recharge facilities at Bay Park have been described in several reports. Cohen and Durfor (1966) presented a detailed description of the recharge well. Cohen and Durfor (1967) discussed the objectives of the recharge study and described the injection equipment. Peters and Rose (1968) reported extensively on all aspects of the project, with particular emphasis on the water-treatment facilities. Peters (1968) commented on the progress of the project and briefly dicussed early operations of treatment and injection plants. The geology and the hydrology of the Bay Park site, the recharge facility, and the nature of the recharge experimentation were discussed briefly by Perlmutter, Pearson, and Bennett (1968). Preliminary results of the early injection experiments were presented by Vecchioli and Ku (1972). A detailed description of the design and operation of the Bay Park artificial-recharge plant was given by Koch, Giaimo, and Sulam (1973).

ACKNOWLEDGMENTS

The recharge facility was designed mainly by personnel of the Geological Survey. Certain engineering specifications were developed by personnel of the Nassau County Department of Public Works and their engineering consultants, Burns and Roe, Inc., and some of the construction was supervised by this department. Individuals who had a major role in the design and the construction of the recharge facility include John H. Peters, Commissioner of the Nassau County Department of Public Works; John L. Rose, of Burns and Roe, Inc.; and Philip Cohen, Charles N. Durfor, and Bruce L. Foxworthy, of the Geological Survey. Particular recognition is given Messrs. Cohen and Durfor, for without their efforts early in the project the detailed testing of the hydraulics and the hydrochemistry of the aquifer system would not have been possible. Installation and early testing of the wells was done under the joint direction of Messrs. Cohen and Durfor. Most of the credit for the design of operating controls and monitoring instrumentation of the recharge facility belongs to Mr. Durfor, currently (1972) with the Environmental Protection Agency. He was also instrumental in the collection of some of the water-quality data on which part of this report is based.

Thanks are given also to Dr. John K. Adams, of Temple University, for providing X-ray diffraction analyses of clay samples. William R. Miller, of the Geological Survey, compiled lithologic logs of the wells. Anthony A. Giaimo, Lillian B. Maclin, George J. Medick, and Thomas K. Mueller, all of the Geological Survey, assisted in the collection and the analysis of much of the data on which this report is based. The authors are grateful for all their assistance.

GEOLOGIC ENVIRONMENT

By LAWRENCE A. CERRILLO and JOHN VECCHIOLI

REGIONAL GEOLOGY

Long Island lies at the extreme north end of the Atlantic Coastal Plain physiographic province. The island is underlain by unconsolidated deposits of sand, gravel, and clay of Quaternary and Late Cretaceous ages, which in turn overlie bedrock of schist, gneiss, and granitic rocks of Precambrian age. (See fig. 3, pl. 1, and table 2.)

Bedrock crops out in northwestern Queens County, along the East River, and dips in a generally southeastward direction to a depth of more

TABLE 1.—*Selected construction details of test wells and observation wells at the Bay Park artificial-recharge site*

Well	Screened interval (ft below land surface)	Nominal screen size (in.)	Screen material	Casing ID (in.)	Casing material	Approximate distance and direction from recharge well
N7884	418–480	16	Stainless steel [1]	17.5	Fiberglass	0 ft.
N7885	418–480	5	---do [1]	4.4	---do	Annular space.
N7886	417–480	6	---do [1]	6.4	---do	20 ft SW.
N7887	535–545	5	---do [1]	4.4	---do	20 ft S.
N7888	307–317	5	---do [1]	4.4	---do	20 ft S.
N7889	535–545	5	---do [1]	4.4	---do	100 ft SW.
N7890	452–462	5	---do [1]	4.4	---do	100 ft SW.
N7891	307–317	5	---do [1]	4.4	---do	100 ft SW.
N8022	420–480	6	Fiberglass	6.4	---do	200 ft NE.
N2790	538–560	4.5	Steel	3	PVC [2] lined steel	200 ft NE.
N6928	716–726	4	---do	6 and 2	Steel	200 ft NE.
N7676	7–10	4	---do	4	---do	200 ft NE.
N7678	65–80	4	---do	4	---do	100 ft N.
N7675	28–34	4	---do	4	---do	200 ft NE.
N7677	84–89	4	---do	4	---do	200 ft NE.
N7679	81–86	2	Stainless steel [1]	2	Fiberglass	100 ft SW.
N7680	39–44	2	---do [1]	2	---do	100 ft SW.
N7681	41–44	2	---do [1]	2	---do	100 ft SW.
N7682	9–12	2	---do [1]	2	---do	100 ft SW.

[1] Type 304.
[2] PVC is abbreviation for polyvinyl chloride.

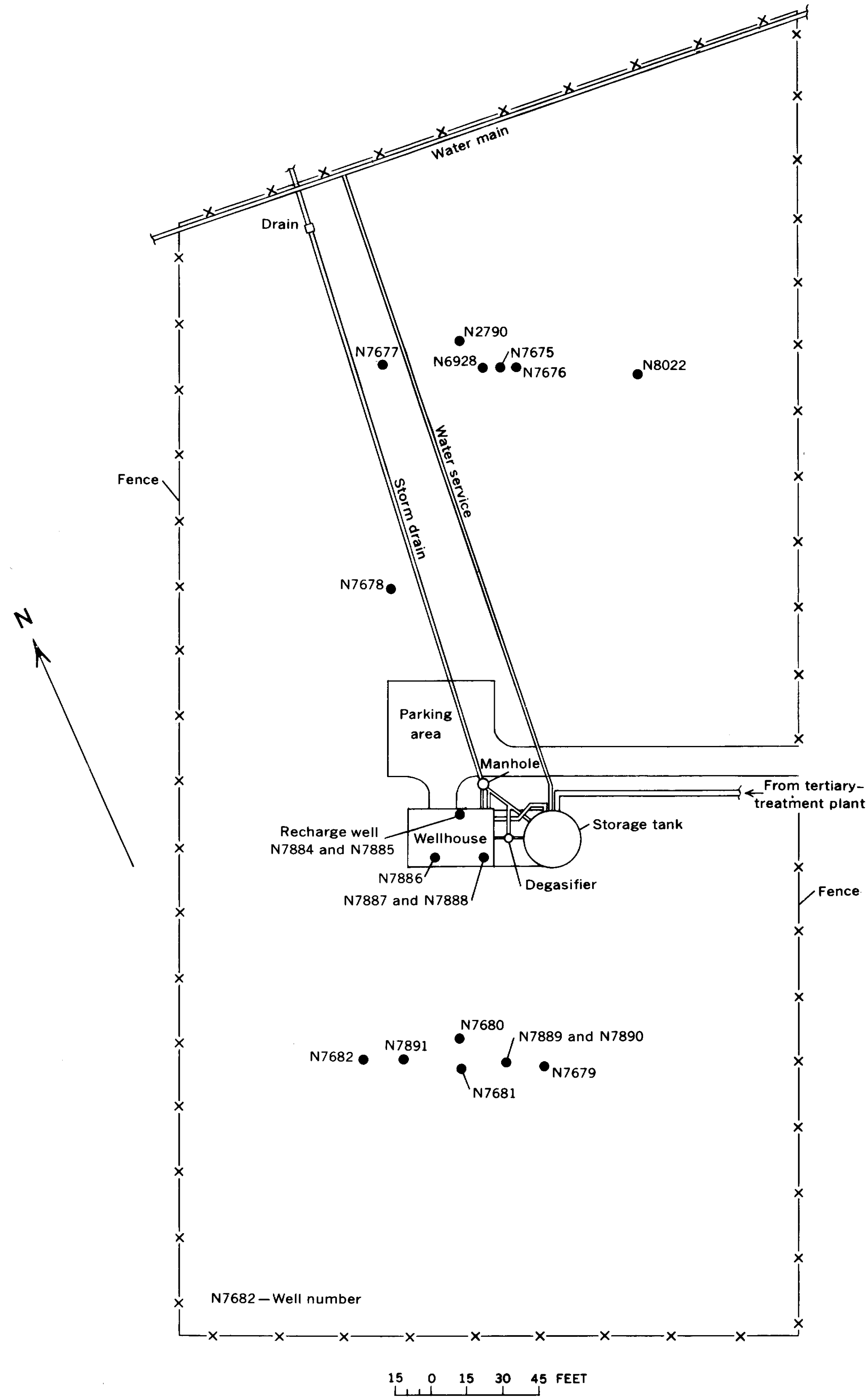

FIGURE 2.—Distribution of wells at the Bay Park artificial-recharge site.

than 2,000 feet below sea level beneath Fire Island, on the south shore of Suffolk County (Suter and others, 1949, pls. 8 and 9).

Immediately above the bedrock is the Raritan Formation, which has been divided into the lower Lloyd Sand Member and the upper clay member. The Lloyd overlies the bedrock everywhere beneath Long Island, except where it has been eroded locally in Kings, Queens, and northern Nassau Counties. It ranges in thickness from 20 feet in northwestern Queens County to 300 feet or more in southeastern Suffolk County. The Lloyd consists chiefly of beds of gray and white sand and gravel and commonly some interstitial clay; interbedded in it also are lenses of sandy clay and nearly pure clay. It is the lowermost aquifer on Long Island. The Raritan clay, which is formally referred to as the clay member of the Raritan Formation, ranges in thickness from 30 feet in northern Kings County to about 300 feet along the south shore of Long Island. It consists typically of light- to dark-gray laminated silty clay or nearly pure clay beds. Beds of red, white, yellow, and mottled clay are less common. Sand layers occur locally, as do layers of lignite and pyrite interbedded with carbonaceous clay. The Raritan clay acts as a confining unit separating the Lloyd aquifer from the aquifers above.

The Magothy Formation-Matawan Group, undifferentiated, overlies the Raritan clay. Hydrologically, this unit is known as the Magothy aquifer; it is the principal aquifer on Long Island. The Magothy aquifer consists of as much as 1,000 feet or more of mostly fine to medium gray quartzose sand interbedded with gray clay and silt. Subordinate gravelly beds are commonly near the base of the unit. Lignite is a common constituent, occurring as disseminated flakes throughout the formation and locally as thin seams. Pyrite and marcasite are commonly associated with the lignite. According to N. M. Perlmutter (oral commun., 1970), lignite and pyrite are more abundant in the Magothy along the south shore of Long Island, particularly in the middle and the upper beds, than they are in the central part of the island. The upper surface of the Magothy aquifer is an erosional surface everywhere on Long Island except in south-central Suffolk County, where the Monmouth Group overlies the Magothy. There the Monmouth Group consists of as much as 200 feet of dark-gray and black silty and sandy micaceous clay and greenish-gray glauconitic sandy clay (Perlmutter and Todd, 1965).

The Upper Cretaceous beds are overlain by Pleistocene deposits, which are, from oldest to youngest, the Jameco Gravel, Gardiners Clay, and upper Pleistocene deposits, undifferentiated. The Jameco Gravel consists predominately of quartzose coarse sand and gravel with diabase and red sandstone fragments common locally. Thickness of the Jameco ranges from a few feet in Nassau County to as much as 250 feet in central Queens County (Soren, 1971). The Jameco is an important aquifer in Kings and Queens Counties; however, heavy pumpage there has resulted in local encroachment of salty water and abandonment of many wells. The Gardiners Clay is a gray silty clay, locally greenish gray, overlying the Jameco Gravel where

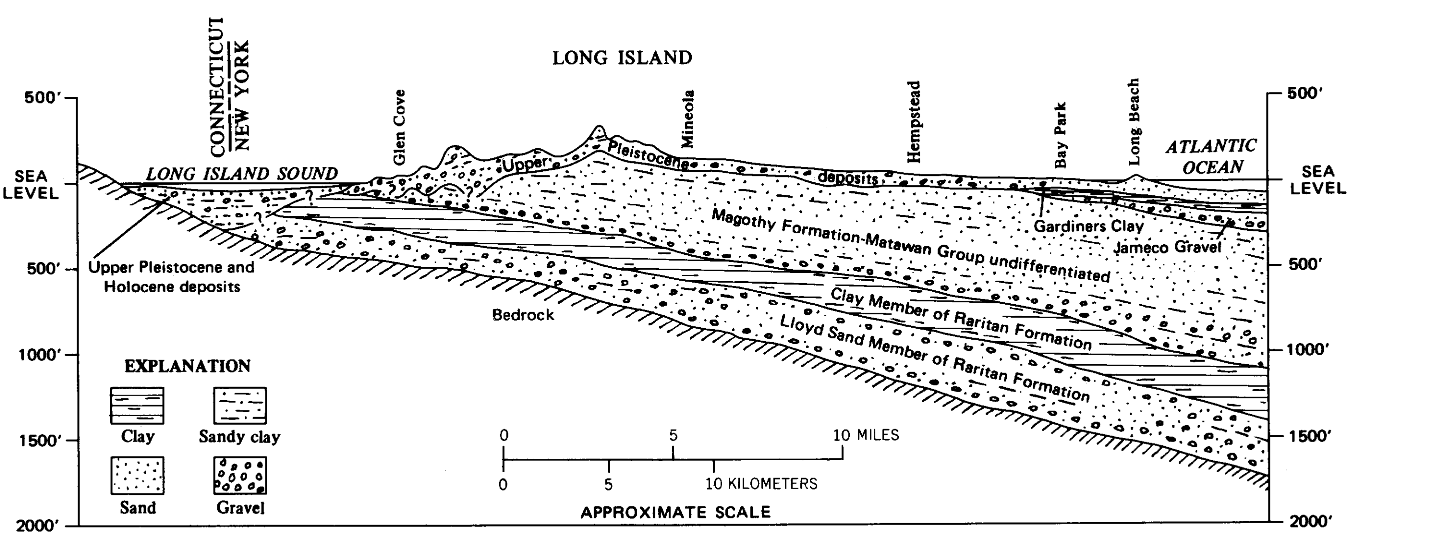

FIGURE 3.—Generalized geologic section showing stratigraphic units in central Nassau County, Long Island, N.Y. (after Perlmutter and Geraghty, 1963, p. A13).

the Jameco is present and lapping onto the upper Cretaceous deposits where the Jameco is missing. The Gardiners Clay, which ranges from a few feet to 150 feet or more in thickness, occurs nearly everywhere along the south shore of Long Island. There it provides a more or less impervious barrier to the movement of water between the upper glacial aquifer and the Jameco or the Magothy aquifers, depending on which unit it overlies. The upper Pleistocene deposits are composed of glacial till, lacustrine silt and clay, and outwash sand and gravel. These deposits are generally less than 100 feet thick, but they are much thicker where they fill buried valleys or occur as morainal deposits. In recent reports, these deposits are referred to as the upper glacial aquifer (Cohen and others, 1968).

Holocene deposits of swamp bogs, stream alluvium deposits, lagoonal sediments, and beach and dune sand occur, generally, in beds less than 20 feet thick along the margins of Long Island.

STRATIGRAPHIC SECTION AT BAY PARK

Well N2790 is the deepest test hole in the Bay Park area, but well N7884, the recharge well, has the most carefully studied lithologic log. Table 2 presents the combined lithologic section extending into the upper part of the Lloyd Sand Member of the Raritan Formation. Plate 1 illustrates the gamma-ray log and lithologic log of well N7884. Cross-sectional representations of the subsurface geology in the vicinity of Bay Park are shown on plate 1. The locations of these sections are shown on plate 1.

TABLE 2.—*Lithologic log of Bay Park recharge well (N7884)*

	Thickness (ft)	*Depth to base of unit (ft)*
Upper Pleistocene and Holocene deposits, undifferentiated:		
Artificial fill; includes thin layers of sludge from sewage-treatment plant	5	5
Sand, brown, fine to coarse, slightly silty; some gravel; biotite abundant	13	18
Clay, solid and silty, gray-green; traces of fine sand (known locally as the "20-foot clay")	6	24
Sand, orange-brown to light-brown, fine to very coarse; layers of fine to coarse gravel	18	42
Pleistocene deposits:		
Gardiners Clay:		
Clay and silt, blue-gray; layers of gray clayey sand; some rock fragments	16	58
Jameco Gravel:		
Sand, light-brown and gray, fine to very coarse, gravelly; layers of gray clay near top and coarse gravel near bottom	46	104

TABLE 2.—*Lithologic log of Bay Park recharge well (N7884)*—Continued

	Thickness (ft)	*Depth to base of unit (ft)*
Cretaceous deposits:		
Magothy Formation–Matawan Group, undifferentiated:		
Sand, gray, very fine to fine, silty, very lignitic, pyritic, and micaceous; contains layers of gray pure clay interbedded with micaceous sand layers	34	138
Sand. gray, very fine to medium, lignitic, micaceous; thin layers of gray clay	18	156
Sand, gray, fine to coarse, silty, lignitic, micaceous; thin layers of gray clay	18	174
Sand, gray, clayey and silty, lignitic, micaceous	11	185
Sand, gray, fine to coarse, lignitic, micaceous, pyritic; layers of gray coarse silty sand and layers of gray silty clay	27	212
Sand, gray, very fine, silty, lignitic, micaceous, and pyritic; layers of gray silty clay	10	222
Sand, gray, very fine to coarse, slightly silty, lignitic, pyritic, and micaceous; layers of gray clay and silt, and layers of gray fine to medium clean sand; contains thin layers of lignite	71	293
Sand, gray, very fine to fine, silty and clayey, lignitic	7	300
Sand, gray, very fine to medium, clean to slightly silty and clayey, lignitic, micaceous; contains layer of pyrite at base	56	356
Sand, gray, very fine to medium, slightly silty, lignitic, micaceous	36	392
Sand, gray, very fine to fine, and gray silt, lignitic, micaceous; lignite layer at 402 ft; pyrite concentration at 435 ft	44	436
Sand, gray, medium to coarse, clean to slightly silty, micaceous, lignitic, slightly pyritic	39	475
Sand, gray, coarse to very coarse, silty, micaceous, slightly lignitic	7	482
Sand, gray, very fine, silty, micaceous, lignitic	8	490
Clay, sandy, gray, lignitic, micaceous	15	505
Sand, gray, very fine to medium, and gray silt, micaceous, lignitic; some layers of lignite; some layers of gray clay; layers of gray coarse clean sand in upper half and coarse sand and gravel at base	88	593
Raritan Formation:		
Clay member:		
Clay, variegated white to light-gray with red streaks; contains abundant siderite grains	42	635
Sand, white to light-gray, fine; gray clay interbedded at top	32	667
Sand, white to light-gray, fine to medium, slightly clayey	21	688
Silt, clayey and sandy, white to light-		

TABLE 2.—*Lithologic log of Bay Park recharge well (N7884)*—Continued

	Thickness (ft)	Depth to base of unit (ft)
Cretaceous deposits—Continued		
Raritan Formation—Continued		
Clay member—Continued		
gray; lignitic and micaceous at 702 ft	17	705
Sand, light-gray, very fine to fine, clayey, micaceous, lignitic in part; thin bed of gray silty clay, micaceous and lignitic, at 727 ft	38	743
[Remainder of log is modified from that of well N2790, after Perlmutter and Geraghty (1963, p. A139).]		
Clay, solid and silty, dark-gray; some layers of silty and fine sandy clay	114	857
Lloyd Sand Member:		
Sand, gray, medium	13	870

BEDROCK

Bedrock in the Bay Park area is about 1,250 feet beneath mean sea level. Depth was estimated from other wells in southern Nassau County that reach bedrock. In well N1927 (pl. 1), 3 miles southeast of Bay Park, the bedrock has been described as probably granite gneiss. The same type of material most likely underlies Bay Park. The hydraulic conductivity of the bedrock is generally low, and bedrock is considered the base for the groundwater system on Long Island. It has little hydrologic significance in the Bay Park recharge study.

RARITAN FORMATION

The Lloyd Sand was not sufficiently penetrated by any of the Bay Park test wells to describe in detail. Apparently the top of the Lloyd Sand Member of the Raritan Formation is about 860 feet below land surface at Bay Park; if so, the sand is nearly 400 feet thick there. In well N1927, at Long Beach, the Lloyd is described in the driller's log as a white coarse-grained sand and gravel with interbedded layers of tough, gray clay. The Lloyd is an artesian aquifer in the Bay Park area. It is used extensively for public water supplies on the barrier beach, such as at Long Beach, where shallower aquifers contain salty water. The Lloyd Sand Member is also of little hydrologic significance in the Bay Park study.

The clay member of the Raritan Formation, about 270 feet thick under Bay Park, consists of about 120 feet of dark-gray clay at the base; about 110 feet of white to light-gray clayey and silty sand in the middle; and about 40 feet of white to light-gray clay with red streaks and abundant siderite granules, at the top. Previous investigators placed the Magothy-Raritan contact 100 feet lower (Perlmutter and Geraghty, 1963, p. A139). The contact was redefined in the present report because (1) the red and white clay at the top is typical of the clay in the Raritan Formation in New Jersey, (2) varicolored clay is described in the Raritan at several places on Long Island, and (3) the unit above the varicolored clay contains coarse sand and fine gravel, which commonly occur at the base of the Magothy on Long Island.

The Raritan clay is a highly effective barrier to the vertical movement of water, and, as such, it is the lowermost hydrologically significant unit in the Bay Park recharge study. For this reason, the upper 40-foot layer of Raritan clay was examined in some detail. Two samples were analyzed by X-ray diffraction. The sample collected 9 feet below the top of the Raritan clay (depth 602 ft) was found to be a mixed kaolinite-illite clay, and the sample collected 19 feet below the top (depth 612 ft) was found to be only kaolinite (J. K. Adams, written commun., 1969). Siderite granules are abundant, particularly in the uppermost 10 feet, and muscovite and tourmaline are scattered throughout the 40-foot layer. Lignite occurs in a few seams.

MAGOTHY FORMATION-MATAWAN GROUP, UNDIFFERENTIATED

The thickness of the Magothy-Matawan unit (the Magothy aquifer) under Bay Park is about 490 feet. Contact between the Magothy and the overlying Jameco Gravel is an erosional unconformity. The Magothy aquifer is the principal source of public and industrial water supply in the Bay Park area.

The Magothy at Bay Park is composed of gray very fine to medium sand and some silt. Clay layers occur throughout the aquifer. X-ray diffraction analysis of a clay sample from 66 feet above the base of the Magothy (at depth 527 ft) showed equal amounts of kaolinite and illite (J. K. Adams, written commun., 1969). Lignite occurs in layers and as disseminated particles throughout the aquifer. Pyrite is also present throughout the aquifer, particularly in association with lignite. Accessory minerals include muscovite, pyrite-marcasite, tourmaline, and, in minor amounts, garnet, zircon, andalusite, and sillimanite.

The screened interval in the recharge well is between about 420 and 480 feet below land surface (115 to 175 feet above the base of the Magothy). A comparison of the percentage of fines (fine sand

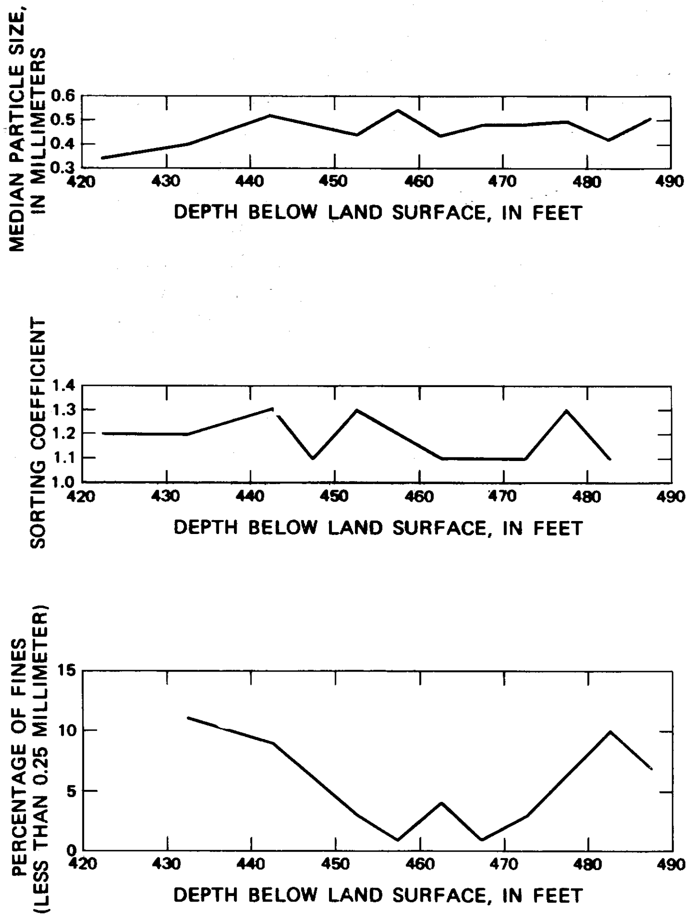

FIGURE 4.—Percentage of fines, sorting coefficient, and median particle size of beds in the screened zone of well N7884.

or smaller), sorting coefficient, and median particle size of material in the screened interval of the recharge well is shown in figure 4. The finest material occurs in the upper 10 to 20 feet of the screened interval. The sorting coefficients are high and range from 1.1 to 1.3. Percentage of fines ranges from 1 to 11 percent. Median particle sizes of samples of the sand range from 0.33 to 0.54 mm.

PLEISTOCENE DEPOSITS

JAMECO GRAVEL

Sand and gravel deposits from 58 to 104 feet below land surface lie on the erosional unconformity at the top of the Magothy aquifer and probably correlate with the Jameco aquifer of central Queens County. These deposits differ somewhat from the Jameco Gravel, as described by Suter, de Laguna, and Perlmutter (1949, p. 21), in that they consist almost entirely of quartzose grains and rock fragments, including a few granite fragments but excluding diabase and red sandstone. Feldspar, biotite, amphibole, and garnet are present in minor quantities. Stratigraphically the Jameco Gravel is continuous along the south shore of Kings and Queens Counties, but it thins to a featheredge to the north.

GARDINERS CLAY

As in central Queens County, the Jameco Gravel is overlain by the Gardiners Clay in the Bay Park area. Although at Bay Park the top of the clay is slightly higher (35 feet below mean sea level) than that reported elsewhere (Suter and others, 1949, p. 22; Perlmutter and Geraghty, 1963, p. 32), the general stratigraphic position and composition correlate with the Gardiners elsewhere. The clay is 16 feet thick at Bay Park and occurs between 42 and 58 feet beneath land surface. The beds consist of blue-gray clay and silt, and a few layers consist of gray clayey sand. X-ray diffraction analysis of the clay fraction showed it to be "illite or glauconitelike clay and a little poorly crystallized kaolinite" (J. K. Adams, written commun., 1969). Lignitic fragments, many with the original wood structure preserved, occur in the Gardiners Clay; other constituents include small amounts of chlorite, biotite, amphibole, and glauconite.

UPPER PLEISTOCENE AND HOLOCENE DEPOSITS

Materials overlying the Gardiners Clay are collectively described as upper Pleistocene and Holocene deposits. Their thickness at the Bay Park site is 42 feet. These deposits consist mostly of fine to very coarse brown quartz sand containing some feldspar and biotite. Within these deposits in the southern part of Nassau County, there commonly exists a clay bed that is known locally as the "20-foot clay," owing to its occurrence at about 20 feet below sea level. The clay is only 6 feet thick at Bay Park.

Fill material consisting of thin layers of sludge from a sewage-treatment plant and fine sand cap the stratigraphic section.

HYDROLOGIC ENVIRONMENT

By GORDON D. BENNETT

GENERAL FEATURES

Hydrologic conditions in the vicinity of the Bay Park recharge site are changing slowly as a result of intensive ground-water development during recent decades. However, data collected in the vicinity of Bay Park since 1961 indicate only minor changes in the local ground-water situation since the completion of the work of Lusczynski and Swarzenski (1966).

Before the beginning of major ground-water development on Long Island (about 1900), the flow pattern along the south shore undoubtedly involved seaward flow of fresh water in all zones and an upward flow component that increased toward the sea. The fresh-water system extended seaward within the aquifer to a transitional zone of mixed fresh water and sea water. The flow system was presumably at equilibrium, and the position of the transitional salt-water front was determined by the total fresh-water discharge to the sea.

The natural hydrologic balance was disrupted by the intensive development of ground water on Long Island from the start of the 20th century. The amount of fresh ground water discharging to the sea was reduced, and the transition zone from fresh ground water to sea water began shifting landward toward a new equilibrium position compatible with the reduced discharge of fresh water. Because of almost continual increases in pumpage from year to year, this new equilibrium position had not been attained up to the time of the work (1958–61) of Lusczynski and Swarzenski (1966). The transition zone from fresh water to sea water was still moving slowly landward in 1961, and in at least one depth interval, the leading edge of the front seemed to be just south of the Bay Park recharge site at that time. The prevailing direction of ground-water flow seems to have been landward (north), in all zones, a short distance to the south of the site. A short distance north of the recharge site, however, the flow of fresh water was still seaward (south) in all zones. Thus, in 1961, the recharge site was located in a hydraulic trough, or discharge area, characterized by lateral ground-water inflow from both north and south and by upward flow throughout the section, which discharged into the bays and the marshes along the shore. In the immediate vicinity of the test site, lateral head gradients were low at all depths, the directions of lateral movement were difficult to define and the directions oscillated in response to local changes in pumping.

Lusczynski and Swarzenski (1966) described three salients, or wedges, in the saline-water front within the Magothy and the Pleistocene deposits of southwestern Nassau County. These include a shallow wedge in the deposits immediately underlying the bays; an intermediate wedge at depths ranging from 100 to 400 feet below mean sea level; and a deep wedge at depths ranging from 150 to 1,000 feet below mean sea level. In 1961, only the deep salt-water wedge had penetrated to the immediate vicinity of the recharge site. According to Lusczynski and Swarzenski (1966), the toe of the deep wedge is approximately 700 feet below mean sea level at Bay Park. Thus, the leading edge of the saline front at Bay Park occurred within the clay member of the Raritan Formation, about 200 feet below the injection zone and separated from this zone by the relatively impermeable clay between 593 and 635 feet below land surface. (Lusczynski and Swarzenski (1966) refer to the clay member as the basal Magothy (?).)

WATER-LEVEL FLUCTUATIONS AND ASSOCIATED CORRECTIONS OF TEST DATA

TIDAL OSCILLATIONS

Various transient hydrologic features are superposed on the general hydrologic conditions described previously. The most noticeable of these is the regular oscillation of ground-water levels in response to tidal oscillations in the bays. The water levels in all the wells screened below the "20 foot" clay at the recharge site exhibit such reciprocative fluctuations. The tidal efficiencies—that is, the ratio of the amplitude of water-level fluctuation in a well to the amplitude of the corresponding tidal oscillation in the bay—range from 0.067 to 0.098 for those wells deeper than 300 feet. Tide-response features in the records of these wells seem to lag the corresponding features in the tidal record of the bay by 3 to 7 minutes.

A detailed discussion of the mechanics of the tide-induced motion is beyond the scope of this paper, but Jacob (1950, p. 331 and 364) describes two types of tidal stress: (1) that in which a confined aquifer is in direct hydraulic contact wih a tidewater body at a submarine outcrop and (2) that in which the tide acts on a perfectly confining layer through which a mechanical stress is transmitted vertically to the confined aquifer. Effects of both types of tidal response are undoubtedly present in the water-level fluctuations of these wells. However, the hydraulic connection between the tidewater and the deep horizons is vertical, through semiconfining beds of low vertical hydraulic conductivity, rather than direct, and therefore the pressure changes are not fully transmitted. Consequently, the amplitude of the hydraulic component is severely attenuated in the vertical transmission from the bay to the deep formations and is further attenuated in lateral transmission to the area of the observation wells, approximately 2,000 feet north of the shore. The component due to mechanical loading of the semiconfining beds is similarly attenuated laterally between some point beneath the bay and the area of the observation wells. A quantitative analysis of the tidal response would require consideration of both types of tidal effect and of both lateral and vertical attenuation, and it would be complex.

For this study, an empirical approach to the tidal disturbance was adopted. Average tidal efficiencies for each observation well were calculated from the record of a tide gage at the Nassau County boat dock, approximately 2,000 feet south of the recharge well, and from water-level records of each of the wells. Several calculations were made for each well, and the results were averaged to remove the effects of random water-level changes not associated with the tidal oscillation. The tidal efficiencies calculated in this way were multiplied by the tidal changes monitored in the bay to obtain corrections for the tidal effect in the record of each observation well during each pumping test.

BAROMETRIC FLUCTUATIONS

Wells more than 300 feet deep at Bay Park show water-level fluctuations in response to changes in local atmospheric pressure. As in tidal oscillation, the mechanics of the barometric effect seem to be more complex than is generally assumed in quantitative analysis of such phenomena. Because the deep zones are not perfectly confined, changes in atmospheric pressure may be transmitted hydraulically, with considerable reduction in amplitude, from the water table to the deep zones. However, the compressive stress that the atmosphere is assumed to exert on a perfectly confined aquifer is at least partly effective in the semiconfined system at Bay Park. Theoretical analysis of the barometric effect was not attempted in this study; barometric corrections of the test data were made empirically by a technique similar to that for the tidal correction.

Barometric efficiencies, the ratio of the change in water in a well to the change in atmospheric pressure assumed to have caused the water-level change, were calculated for each of the 10 deep wells. Water-level and barometric records from nonpumping periods were used in the calculations. The water-level record was corrected for the tidal effect before the calculation of barometric efficiency. The calculations of barometric efficiency were subject to much greater uncertainty than the calculations of tidal efficiency, because water-level fluctuations associated with barometric change were usually smaller than water-level disturbances caused by changes in local pumping. In making the calculations, the authors assumed that the fluctuations associated with pumping were random; several calculations of barometric efficiency were made for each well, and the results were averaged. Individual values varied widely. But the average for any given well was between 0.25 and 0.40 and between 0.31 and 0.40 in nine of the 10 wells for which calculations were made. A value of 0.36 was used to correct the test records for the barometric effect. Changes in atmospheric pressure, in feet of water, occurring during each aquifer test were multiplied by 0.36 to correct the water level for the barometric effect.

WATER-LEVEL FLUCTUATIONS ASSOCIATED WITH CHANGES IN LOCAL PUMPING

The Bay Park area lies in a region of intensive ground-water development, and, thus, hydrologic "noise," or water-level fluctuations caused by random variations in pumping from neighboring wells, is a prominent feature of the local hydrology. Fluctuations of this type were noted in the records of many of the aquifer tests. In some tests, reductions in local pumping caused rising trends in water-level records collected during the test; in others, increases in local pumping caused falling trends. Development of adequate methods for correcting these effects was a principal concern in test analysis. The number of pumping wells in the area and the lack of detailed pumpage records for many of them precluded any analytical evaluation of the effects of pumping.

Accordingly, corrections were determined through a somewhat arbitrary "noise well" procedure. An observation well (N6707) about 11,000 feet from the recharge site and screened between depths 496 and 506 feet was selected as a "noise well." Preliminary testing showed that pumping the recharge well at rates up to 400 gpm produced no measurable effect in this observation well. Water-level records collected for the "noise well" during the various aquifer tests were corrected for tidal and barometric effects. The resulting plots of water level against time were compared with those representing the wells at the recharge site. In certain tests, departures of the observation-well data of Bay Park from normal test behavior could be correlated with trends observed in the noise-well data. In these tests the authors assumed that departures were caused by some change in local pumping. A correction equal to and the opposite of the trend in the noise-well data was applied to the water levels of the Bay Park wells.

AQUIFER CHARACTERISTICS AND WELL HYDRAULICS

By Gordon D. Bennett

GENERAL CHARACTERISTICS OF THE RECHARGE-WELL FLOW SYSTEM

The recharge well is screened from about 420 to 480 feet below land surface in an anisotropic, stratified, but virtually unconfined aquifer. Similar wells have been described in the literature as tapping semiconfined or "leaky artesian" aquifers. The distinction is largely a matter of opinion. The screened zone can properly be described as semiconfined. Within this zone, the response to pumping or recharge is similar to that predicted by various theoretical models for semiconfined aquifers. However, the effect of recharging or pumping the well extends through many zones above and below the screened interval, and an appraisal of this total effect is probably best achieved by viewing the entire aquifer as a stratified, anisotropic unit.

Several aquifer tests were made on the Bay Park wells before the beginning of the recharge experiments. Figures 5 through 8 show the composite time-drawdown curves for tests on October 5, 1966, January 11, 1967, and February 7–9, 1968. Drawdown divided by discharge is plotted on the arithmetic scale in these figures against time on the logarithmic scale. Methods previously described were used to correct the data for tidal, barometric, and random fluctuation effects. Figure 5 shows the response of the recharge well, N7884, and the observation well in the gravel pack, N7885. Figure 6 shows the response of the three observation wells at a radial distance of 20 feet from the recharge well. Figure 7 shows the response of the three observation wells approximately 100 feet from the recharge well. Figure 8 shows the response of observation wells approximately 200 feet from the recharge well; in wells at this distance that were shallower than 300 feet, the responses were very slight and were subject to question in all the tests. Hence, they are not shown.

The recharge well and the observation wells screened within the 420- to 480-foot depth interval show the type of response normally associated with a semiconfined aquifer. The initial part of the time-drawdown curve resembles that of an "ideal" confined aquifer; the slope of the semilogarithmic plot approaches a straight line during the initial moments of discharge. During the first few minutes, the hydraulic effects of pumping are largely restricted to the 420- to 480-foot zone; the drawdowns of the shallower and the deeper wells actually show a slight rise in water level, or reverse trend, during this period, apparently due to structural deformation of the aquifer. The primary significance of their response during this period is that the drawdown, and, therefore, the release of water from storage, is negligible above and below the screened zone during the first few minutes of pumping. The discharge of the well is sustained during this period by release of water from artesian storage within the screened interval, and the initial response of observation wells within this zone is artesian in character.

Small drawdowns of water level at the water table are characteristic of wells in this type of stratified anisotropic aquifer. The effect of pumping spreads rapidly in the screened zone during the initial "artesian" response; thus, the influence of the well on the water table is distributed over a wide area, and the rates of water-table drawdown required to sustain the well discharge are correspondingly small. At Bay Park, there seems to be ample opportunity for lateral recharge of the Pleistocene deposits through inlets, streams, and boat slips crossing the area. This lateral infiltration may have been an additional factor in causing water-table drawdowns to remain small. Nonetheless, compressive or artesian storage seems to contribute very little to the well discharge after about 100 minutes of pumping. Beyond that time, the flow is nearly at equilibrium and seems to be sustained by surface storage.

For the 420- to 480-foot interval, figure 9 shows the relation of drawdown divided by discharge to the distance from the recharge well for 10, 100, 1,000, and 2,000 minutes after pumping began. The plots are based on the composite time-drawdown curves shown in figures 5 through 8. The linear character of the distance-drawdown graphs for distances of less than 200 feet indicates that within this radius the flow occurs in a nearly horizontal-radial pattern that is largely confined to the 420- to 480-foot interval.

After 5 to 10 minutes of discharge, the observation wells screened in the zones overlying and underlying the screened interval begin to show appreciable drawdown. Their declining water levels indicate that withdrawal from storage in these zones is beginning to contribute to the well discharge. At the same time, the time-drawdown curves of the observation wells screened in the 420- to 480-foot interval begin to flatten in the manner normally associated with semiconfined aquifers. In effect, the zone of influence of the discharging well

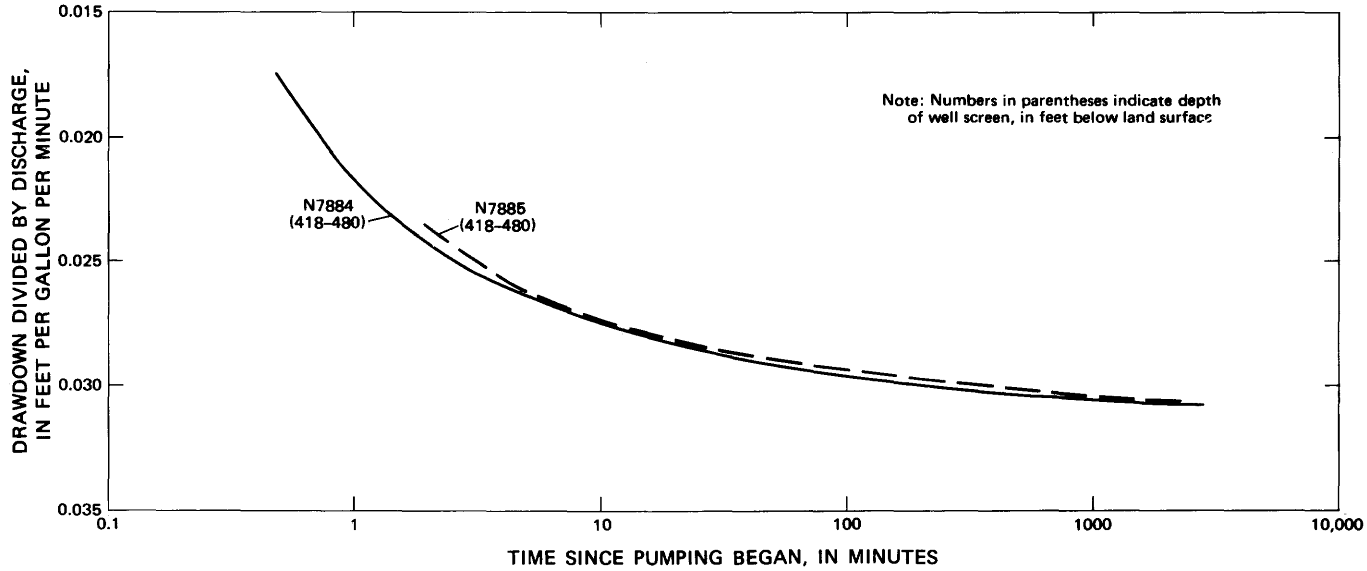

FIGURE 5.—Response of recharge well (N7884) and annular-space observation well (N7885) to pumping of the recharge well.

is spreading both laterally and vertically. As contributions from storage from other zones increase, withdrawals from storage in the 420- to 480-foot zone diminish, and the slopes of the drawdown plots for that zone diminish accordingly. Ultimately, all the drawdown curves flatten. This stabilization in drawdown suggests that the effect of pumping has extended to the surface, where reduction of head induces recharge from the bays and inlets or diverts discharge from these bodies and probably also draws on water-table storage in the upper Pleistocene aquifer. Regardless of the extent of such withdrawal from unconfined storage, the response of the shallow observation wells during the tests showed that the associated water-table drawdowns were very small. Thus, the drawdown at the upper surface of the

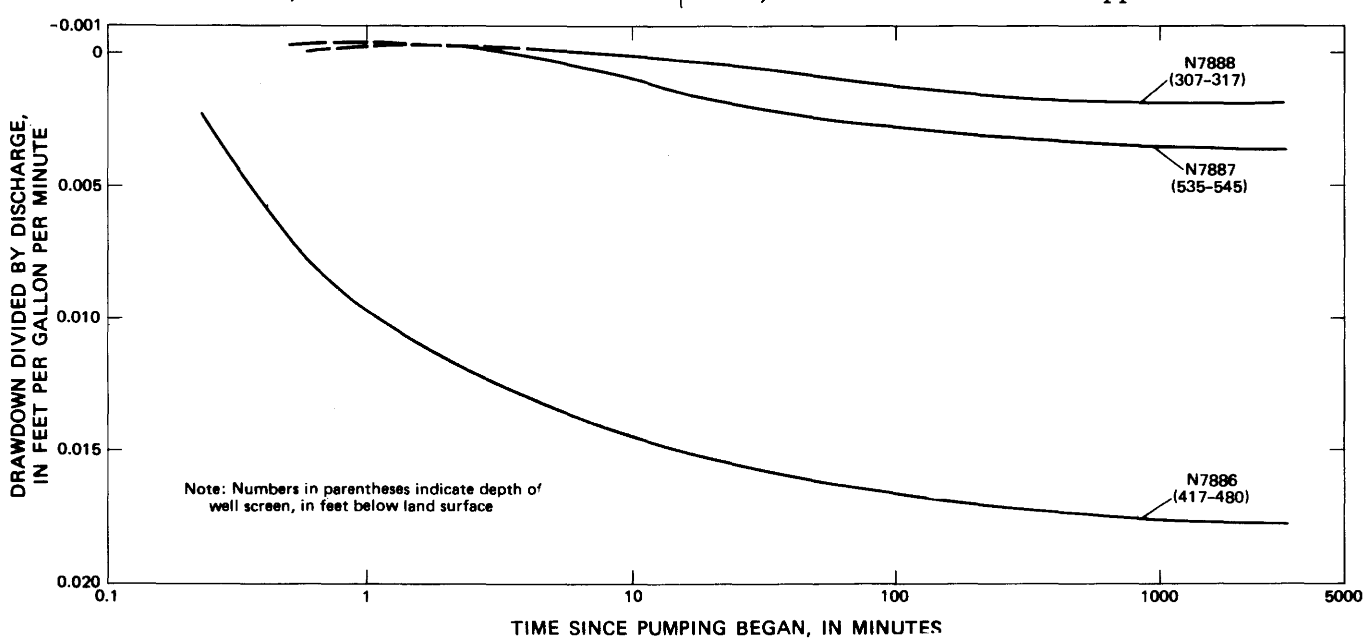

FIGURE 6.—Response of observation wells N7886, N7887, and N7888, 20 feet from the recharge well, to pumping of the recharge well.

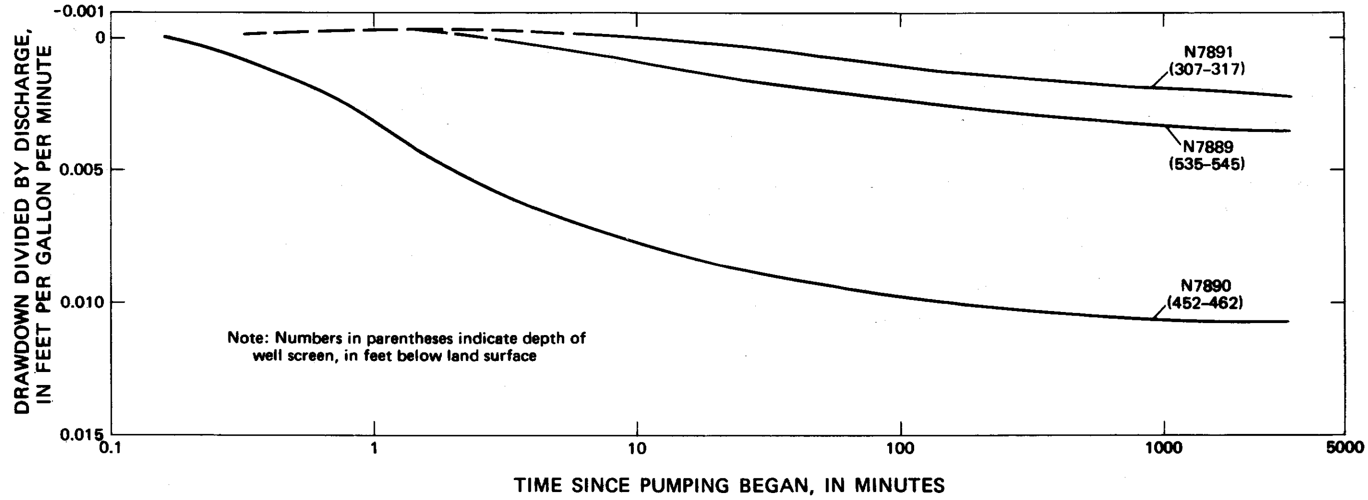

FIGURE 7.—Response of observation wells N7889, N7890, and N7891, 100 feet from the recharge well, to pumping of the recharge well.

aquifer, which includes areas of both the bay floor and the free ground-water surface in the Pleistocene aquifer, was zero, or very nearly so, at all points.

Additional aquifer tests were made in which wells N8022 and N7886 were used as the discharging wells. The results of these tests confirmed those of the tests on the recharge well. The same time-drawdown response was observed in both the discharging well and the observation wells. The results suggest a brief initial artesian condition followed by a transition toward an equilibrium cone of depression sustained by flow from the surface.

SYMMETRY AND EXTENT OF THE EFFECTS OF PUMPAGE

Except for the "noise well" described in the section "Water-level Fluctuations Associated With Changes in Local Pumpage," no observation well was beyond a distance of 200 feet from the recharge well. Thus, it was difficult to estimate the lateral extent of the effect of pumping this well, and no data were available to indicate whether this effect is symmetrical at large distances from the well. However, some information on the extent and symmetry of the cone of depression can be derived indirectly.

At Bay Park, the principal factor that might distort the radial symmetry normally associated with flow to a well is the distribution of potential recharge sources. Hewlett Bay, which would presumably be the major source of recharge, is 2,000 feet south of the well field. If the equilibrium flow pattern involved a heavy proportion of direct flow from the bay to the well, considerable asymmetry

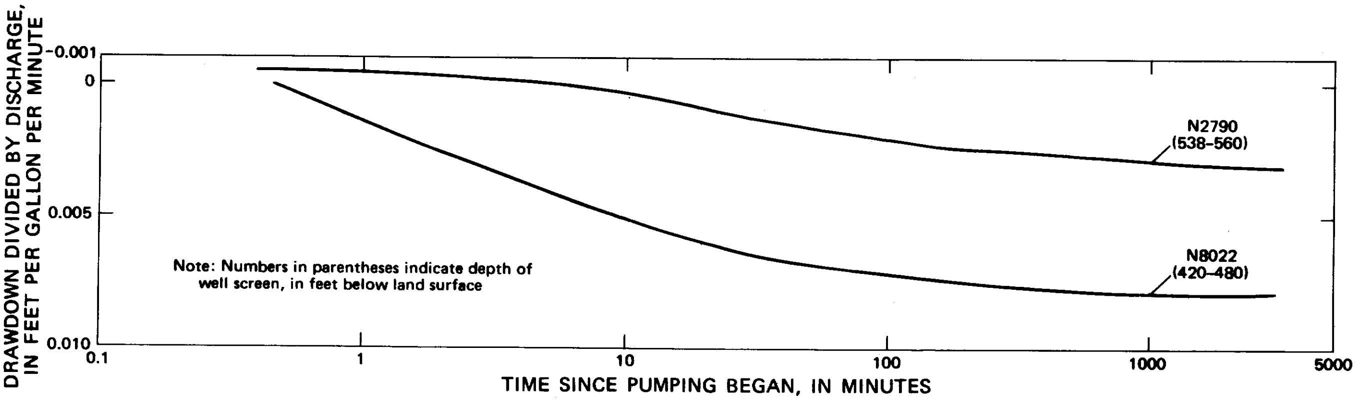

FIGURE 8.—Response of observation well N2790 and N8022, 200 feet from the recharge well, to pumping of the recharge well.

could be introduced. However, such an asymmetrical pattern could develop only if water-table drawdown in the upper Pleistocene deposits were substantial and if the floor of the bay remained a surface of no drawdown.

The response of the shallow observation wells indicated that water-table drawdown was very small during the tests. Thus, there was little difference, hydraulically, between the aquifer surface beneath the bay and the free surface in the upper Pleistocene deposits. Each represented an aquifer boundary on which drawdown was virtually zero, and both could be expected to exert approximately the same control on the flow pattern in deeper zones. Accordingly, there was probably no strong tendency for the well to draw water from the bay, in preference to the landward area, during the aquifer tests; therefore, in the analysis of these tests, the conventional radial symmetry was assumed.

Because radial symmetry is assumed, the lateral extent of the effect of pumping may be discussed in terms of a radius of influence. The test data may be used to set limits on the final, or equilibrium, radius of influence, but close estimation of this radius is not possible. If the recharge to the system is assumed to be proportional to the area of influence of the well, a dependence between the discharge and the equilibrium radius of influence is introduced; and, even if no such assumption is made, a practical dependence between these parameters usually arises as a result of limitations of measurement. For example, if the drawdown at a certain radius is just below measurable drawdown, the usual conclusion is that the influence of the well does not extend to this radius. However, a sufficient increase in discharge will increase the drawdown at this radius to measurable proportions. It is difficult to discuss radius of influence without reference to discharge regardless of the assumptions concerning the flow system. The distribution of observation wells at Bay Park was inadequate to establish a relation between discharge and radius of influence. The preceding remarks on radius of influence applies to discharges in the range from 100 to 400 gpm (gallons per minute); their extension to discharges outside this range is not warranted.

No influence of pumping was noted in the "noise" well, N6707, in any of the tests. The distance of this well from the recharge site, 11,000 feet, may therefore be taken as the maximum radius of influence. A minimum radius may be obtained from the zero-drawdown intercept of distance on the distance-drawdown plot obtained in a pumping test. This intercept is invariably much less than the actual radius of influence. In the test on January 11, 1967, in which the discharge was 345 gpm, this intercept was 1,200 feet after 380 minutes pumping; in the test of February 7–9, in which the discharge was 195 gpm, the intercept was 1,500 feet after 2,000 minutes pumping. In each of these tests, the system seemed to be approaching equilibrium; thus, these intercept figures should be considerably less than the final, or equilibrium, values of the radius of influence. At discharge rates ranging from 100 to 400 gpm, the measurable equilibrium radius of influence is probably between 2,000 and 11,000 feet from the pumped well.

In a series of steady-state analog-model experiments, described in a later section, models of the well-aquifer system incorporating various radii of influence and anisotropies were constructed. A close simulation of the drawdown distribution observed in the field was obtained by a model having a radius of influence of 8,192 feet. This value cannot be taken as conclusive; but, together with the observation-well data, it indicates the order of magnitude of the radius of influence.

AQUIFER COEFFICIENTS

LATERAL HYDRAULIC CONDUCTIVITY

The average lateral hydraulic conductivity of the 420- to 480-foot screened interval was calculated from the formula

$$K_L = \frac{2.3 \times 1{,}440}{2\pi D \left[\dfrac{\partial (s/Q)}{\partial (\log r)} \right]} \qquad (1)$$

where K_L is the average lateral hydraulic conductivity of the screened interval, in gallons per day per square foot; D is the length of the recharge well screen, in feet; and $\dfrac{\partial (s/Q)}{\partial (\log r)}$ is the slope, in feet per minute per log cycle, of a semilog plot, such as those shown in figure 9, between radii of 0.84 and 200 feet. This technique is equivalent to the Thiem method or to the Theis distance-drawdown procedure (Cooper and Jacob, 1946). The 60-foot thickness of the screened interval is considered to be the effective thickness of aquifer. The method is based on the assumption that within a radius of 200 feet from the discharging well most of the flow occurs in a horizontal radial pattern within the 420- to

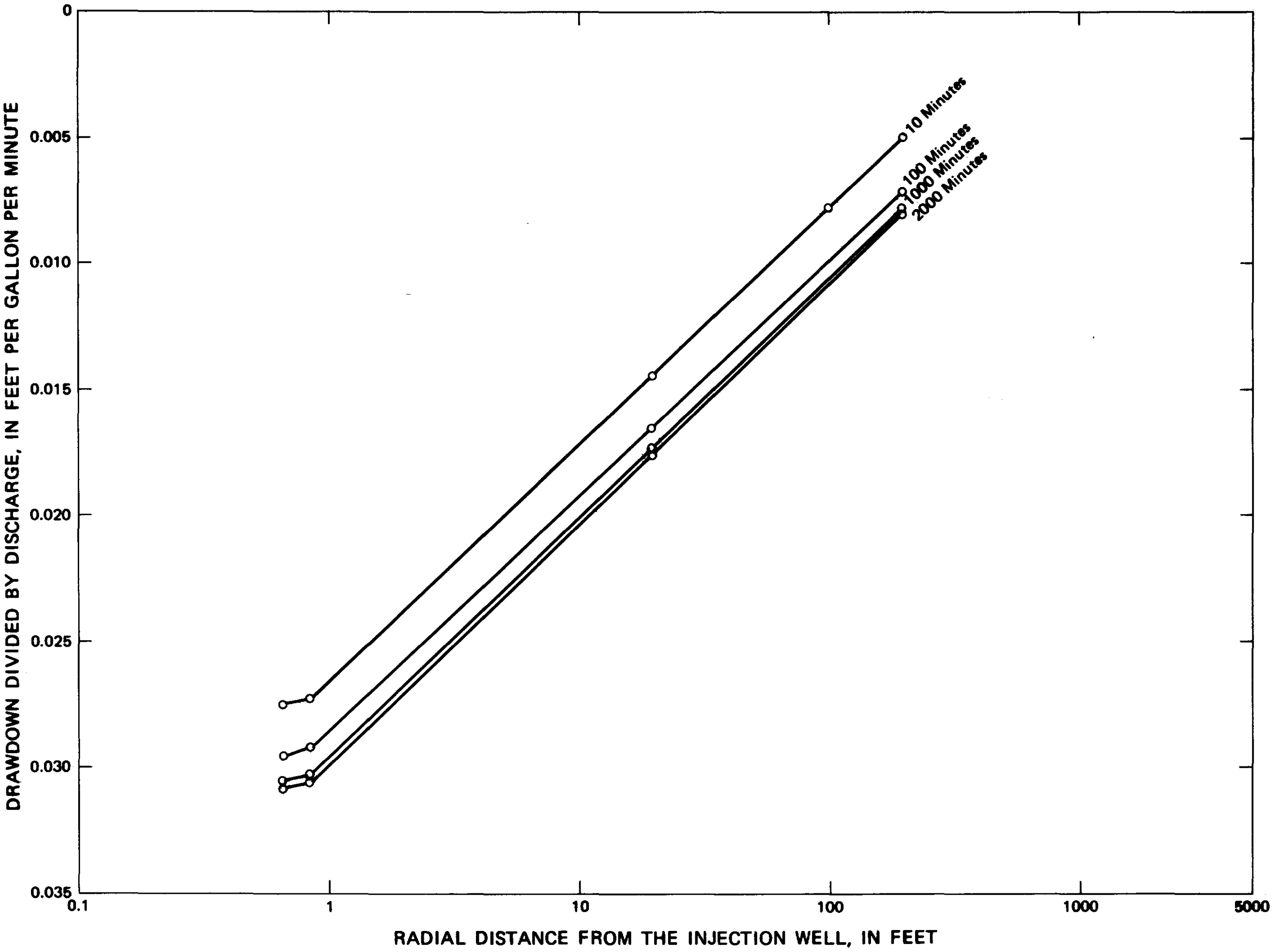

FIGURE 9.—Relation of drawdown divided by discharge versus radial distance from the recharge well, for the 420- to 480-foot interval, for various times during pumping of the recharge well.

480-foot interval. Both linearity of the plots of figure 9 and the results of electric-analog-model experiments representing the well-aquifer system indicate that this assumption was met during the various pumping tests. (See section, "Analog-Model Studies.") The average lateral hydraulic conductivity determined by this method, using data for 2,000 minutes discharge of the recharge well, was 940 gpd per sq ft (gallons per day per square foot), or 126 ft per day. Flowmeter surveys, described in a later section, indicate that the lateral hydraulic conductivity varies considerably within the screened interval; hence, the value of 940 gpd per sq ft (126 ft per day) represents an average for the various strata within the screened interval.

Additional values for the average lateral hydraulic conductivity of the 420- to 480-foot interval were obtained from aquifer tests on wells N7886 and N8022. The results of these tests are probably less reliable than those of the tests on the recharge well. The lack of confidence in the results is due partly to the fact that there were measurable pipe losses in the casings of these wells, which are only 6 inches in diameter, and partly to the fact that the spacing of the observation wells was less favorable for testing wells N7886 and N8022 than for testing the recharge well. Despite these shortcomings, the results show reasonable agreement with the average value obtained from the tests on the recharge well. The test of well N7886 gave a value of 880 gpd per sq ft (118 ft per day), whereas that of well N8022 gave a value of 960 gpd per sq ft (128 ft per day) for the average lateral hydraulic conductivity of the 420- to 480-foot interval.

Specific-capacity tests of several hours duration were made of wells N7887, N7888, N7889, N7890, and N7891. The specific capacities of these wells are an indication of the lateral hydraulic conductivities of their respective screened intervals. The specific capacities and screened intervals of the wells are as follows:

Well and screened interval (ft)	*Specific capacity (gpm per ft)*
N7887 (535–545)	8
N7888 (307–317)	8
N7889 (535–545)	6
N7890 (452–462)	12
N7891 (307–317)	8

Each of the wells screened outside the 420- to 480-foot interval has a lower specific capacity than N7890. Lithologic data collected during test drilling suggest that the material in the screened interval, particularly that between 440 and 480 feet, generally has a higher hydraulic conductivity than the material in other parts of the section. The preceding data support this interpretation.

STORAGE COEFFICIENT

An artesian storage coefficient can be computed from the early test data, but the significance of such a coefficient is difficult to evaluate. Applying the semilog time-drawdown technique (Cooper and Jacob, 1946) to the response of well N7886 during pumping of the recharge well, for the time interval from 5 to 10 minutes, an artesian storage coefficient of roughly 1.5×10^{-4} is obtained. However, as noted in the discussion of general features of the flow field, the zone of influence of the well expands vertically as well as laterally during the initial moments of pumping. Whether the computed artesian storage coefficient refers only to compressive storage in the 420- to 480-foot interval or whether it is a composite value describing release from storage over a wider and continually expanding interval is debatable. Without doubt, an artesian or compressive storage coefficient computed from initial data is inadequate to describe the behavior of the system over a long period. As the flow field expands into zones above and below the screened interval and ultimately extends to areas of recharge or water-table storage, the response of the system will deviate from that predicted on the basis of artesian storage alone. Thus, the utility, as well as the significance, of the artesian storage coefficient obtained is a matter of some conjecture.

VERTICAL HYDRAULIC CONDUCTIVITY

The average vertical hydraulic conductivity of the material between the screened interval and the water table was calculated by the Bessel function analysis of Jacob (1946), as described by Ferris, Knowles, Brown, and Stallman (1962, p. 110–118). Jacob considers the problem in which the discharge of a well tapping a semiconfined aquifer is sustained at equilibrium by vertical leakage through a single semiconfining layer. In applying the methods to the Bay Park data, the 420- to 480-foot interval was considered to be the yielding aquifer; and the interval between 420 feet and the water table was taken as the thickness of the semiconfining layer. The assumptions underlying Jacob's method represent no more than a very crude approximation of the geohydrologic environment at Bay Park; thus, results obtained by this method have only an order-of-magnitude significance. Using data for 2,000 minutes pumping in the test of February 7–9, 1968, the average vertical hydraulic conductivity computed by this method was 12.5 gpd per sq ft (1.67 ft per day).

In the analog-model experiments, the closest match between model results and field data was obtained by a network in which the material between the water table and 420 feet, except for the interval 300 to 360 feet, was modeled as having an average vertical hydraulic conductivity of 4 gpd per sq ft (0.54 ft per day). The interval 300 to 360 feet was modeled as having an average vertical hydraulic conductivity of 2 gpd per sq ft (0.27 ft per day).

Because the analog is only an approximation of the actual flow system, results obtained from analog experiments are subject to much of the same uncertainty as results obtained from mathematical analysis. Hydraulic conductivity may, of course, vary through an extremely wide range, and vertical hydraulic conductivity, in particular, is a difficult parameter to measure in the field. Significantly, however, the value of the vertical hydraulic conductivity obtained from the analog analysis and that obtained from Jacob's Bessel-function procedure show closer agreement than might be expected, and the average vertical hydraulic conductivity of the material between the water table and 420 feet is probably in the 2 to 20 gpd per sq ft (0.27 to 2.7 ft per day) range.

WELL LOSSES

The aquifer tests indicated that head losses through the screen of the recharge well and pipe losses within the 18-inch casing were negligible at discharges up to 400 gpm before injection. As seen in figure 5, the differences in head between N7885 and the recharge well were extremely small during the various aquifer tests. The difference in water level was on the order of 0.05 foot during discharge of approximately 200 gpm.

FLOWMETER SURVEYS

Flowmeter surveys were made in the recharge well, N7884, and in observation wells N7885, N7886, and N8022. All are screened in the 420- to 480-foot interval. These surveys were made with a 3-inch Au current meter as each well was pumped. Because the meter was not equipped with centering guides and because depth control may have been faulty, the results are questionable.

Flowmeter curves for the various wells are presented in figure 10. The curves show percentage of the well discharge against depth. The principal causes for the considerable disagreement among the curves are, presumably, errors of measurement, differences in well development, and local differences in geology. Despite the disagreement, however, certain conclusions about the differences in hydraulic conductivity caused by stratification within the screened interval can be drawn.

The lowermost part of the screen seems to be an interval of high yield, as shown by the curves for N7884, N7886, and N8022; the uppermost part seems to be an interval of low yield in these wells. In the curves for N7884 and N8022, the zone between 440 and 450 feet shows relatively high yield. These features, common to more than one well, probably reflect stratification in the aquifer. Moreover, the general tendency of higher yield toward the lower part of the screen correlates with geologic data collected during drilling of the wells. The extremely low yield recorded in the 420- to 430-foot interval in the recharge well may not be accurate, as it was not recorded in the other wells; and errors caused by improper centering of the current meter were probably most severe in the recharge well.

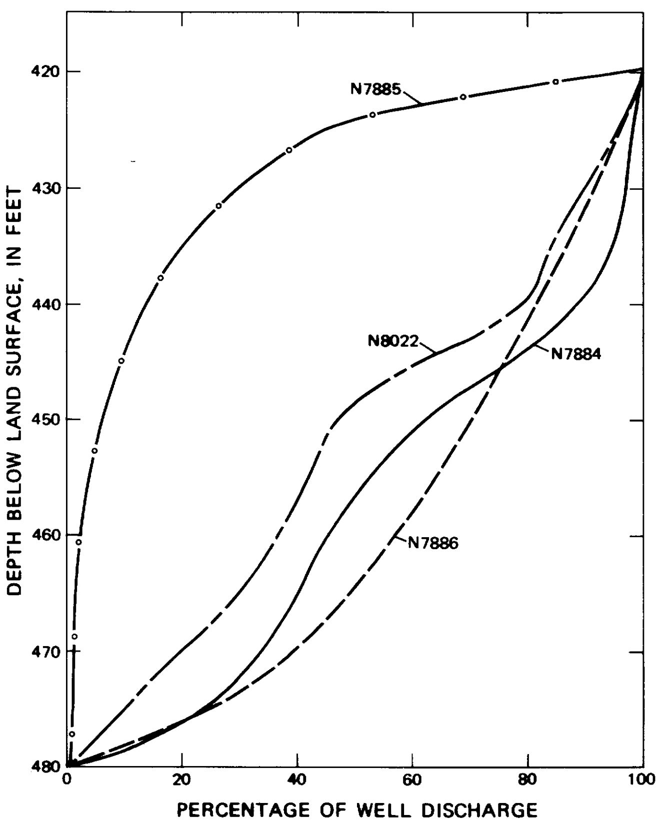

FIGURE 10.—Flowmeter logs of wells N7884, N7885, N7886, and N8022 during pumping of each well. Logs made before any recharge.

The flowmeter curve of the observation well (N7885) immediately adjacent to the recharge well is totally different from that of the other three wells; it shows high inflow in the upper part of the screen and very little in the lower part. This well was not developed after construction, so the lower part of the screen may be choked with drilling mud; or, possibly, when N7885 is pumped, water from the lower permeable zones flows into the recharge well and transfers to N7885 through the gravel pack in the upper part of the screen. If this interpretation is correct, water-level measurements in N7885 during operation of the recharge well reflect the head in the upper part of the gravel pack. In future recharge operations, this head might differ significantly from the head lower in the gravel pack if certain sections of the screen or gravel pack should become severely clogged.

ANALOG-MODEL STUDIES

In further study of the flow pattern around the recharge well, a steady-state, radially symmetrical, electric-analog model of the type described by Stallman (1963) was constructed for the system. In this type of model, a two-dimensional network of electrical resistances is used to represent the r-z plane in the aquifer; r and z are the radial and the vertical coordinates of the cylindrical system, and z is also the axis of the well. Thus, cylindrical symmetry is built into the model, which represents only the flow into or from a single well coaxial with the z-axis. No attempt is made to simulate the interaction of the well with other discharging wells or with any sort of regional flow. The model approach uses the familiar analogy between steady-state flow of liquid through a porous medium, as described by Darcy's Law, and steady flow of electrical charge through a conductor, as described by Ohm's Law.

The general procedure of the analog study was as follows: A series of experiments was made with models representing different ratios of anisotropy and radii of influence. In certain experiments, changes in other parameters were introduced—for example, in the effective well radius (r_w) and in the resistance used to represent certain layers of high or low hydraulic conductivity. Each of the experiments simulated steady discharge of the recharge well sustained by recharge from the aquifer surface, which was assumed to be a surface of zero drawdown. Measurements of the electrical potential were made at significant points in each model, and the potential distribution observed was compared with the drawdown distribution observed at the recharge site during pumping of the recharge well. The model yielding the best agreement with the field data was taken as an approximate representation of the well-aquifer system at Bay Park. A complete analysis was then made with this model, and a flow net representing equilibrium operation of the recharge well was constructed.

Because of decreasing cross-sectional areas of flow, vertical and lateral hydraulic resistance in the aquifer increase as the distance from the well decreases. The resistance to lateral (radial) flow increases logarithmically with decreasing radius; for example, a section between the radii 64 and 32 feet has a lateral hydraulic resistance equal to that of a section between 32 and 16 feet. A logarithmic scale of distance is used along the radial axis of the model, so that resistances of equal value may be used in the lateral rows. As shown by Stallman (1963), the vertical resistances required at successive junctions along the r-axis then increase in a geometric progression as r decreases.

In the model used in this study, the value of r at each electrical junction, or nodal point, differed from that at the next junction along the r-axis by a factor of $\sqrt{2}$. Thus, the junctions along the r-axis represented radii in the series 1, 1.41, 2, 2.83, 4, 5.66, 8, ***n. At large values of r, the intervals represented between successive junctions in this series become very large. Along the vertical axis, successive nodes were assumed to represent depths (in feet) in the sequence 0, 30, 60, 120*** 300, 360, 405, 435, 465, 495, 540, 600, 660, 720. The change in vertical spacing in the vicinity of the screened zone was introduced to provide greater detail in the flow pattern in this interval. The base of the aquifer—that is, the lower terminus of the network—was the line of junctions representing the depth of 720 feet. This depth for termination of the electrical network was selected to agree with the clay indicated at this depth on the lithologic log.

The outer limit of the network, along the radial axis, actually simulates an impervious cylindrical boundary or flow divide coaxial with the well. In practice, if this outer limit of the network is made to coincide with the radius of measurable influence of the well, the error involved in limiting the network in this way is negligible. In the series of analog experiments, different radii of influence were simulated by varying the length of the network along the radial axis.

The stratified, anisotropic character of the aquifer was represented in two ways in the analog network. A general anisotropy that gave a smaller vertical hydraulic conductivity than the lateral hydraulic conductivity at any given point was built into the model. In addition, an effort was made to simulate the more significant aspects of aquifer heterogeneity by using high or low resistances to represent certain intervals. In particular, low resistances were used in the lower part of the screened interval. Both lithologic and flowmeter data indicated this to be a zone of fairly high hydraulic conductivity. High resistances were used to represent a few intervals in which prominent clay layers were observed. In general, assumptions made in the modeling of aquifer heterogeneity were arbitrary. Hydraulic-conductivity values were assigned on the basis of qualitative lithologic data or approximate flowmeter data, rather than on hydraulic-conductivity measurement. During the series of analog experiments, a few variations were made in the resistances representing layers that showed high or low hydraulic conductivity; values in closest agreement with the aquifer-test data were retained. The amount of this trial-and-error variation was not sufficient to alter the basically arbitrary nature of this aspect of the model design.

The effect of clay layers other than those represented by increased resistance was simulated by the general anisotropy of the model. This general anisotropy also contributed to the simulation of those clay layers for which higher resistance was used. In the aquifer, thin clay or silt layers reduce the average vertical hydraulic conductivity much more than they reduce the average lateral hydraulic conductivity. In constructing the model, it was not feasible to represent each individual clay or silt bed in the section; rather, the effect of these thin clay and silt layers was simulated by representing the average vertical hydraulic conductivity as everywhere lower than the average lateral hydraulic conductivity. The effect of a thick or a tight clay or of an interval containing a high proportion of thin clay layers was then simulated by increasing the resistance of the section that included the clay to some specified multiple of the resistance used for other sections. Thus, in an experiment in which the vertical resistance representing the 600- to 660-foot interval was 16 times that used for other 60-foot intervals, the clay layers within this interval were assumed to reduce its average vertical hydraulic conductivity to one-sixteenth that of other 60-foot intervals. This assumption was not an affirmation, however, that the vertical hydraulic conductivity of clay was one-sixteenth that of sand or silt but was rather a rough approximation to facilitate modeling; some clay layers present in the other 60-foot sections were represented by the anisotropy built into the model.

The values given as general anisotropy in this report refer to the ratio of vertical to lateral hydraulic conductivity represented by those sections of the analog network that were not altered to simulate aquifer heterogeneity—for example, by the section representing the interval from land surface to a depth of 300 feet. The analog design used in these studies has the advantage that both the general anisotropy and the radius of influence can be varied through a wide range by means of minor adjustments in the network. The adjustment by which anisotropy is varied allows the ratio of vertical to lateral hydraulic conductivity at every point to be multiplied or divided by the same factor. Thus, if the general anisotropy of a given model is one sixty-fourth but an individual section containing a prominent clay is represented as having a vertical hydraulic conductivity one one hundred twenty-eighth of its lateral hydraulic conductivity, an adjustment to multiply the general anisotropy by four will automatically multiply the anisotropy of the section containing the clay by four.

The adjustment is made by shifting the radial scale of distance along the model network. The model is designed so that each junction represents a radius of $\sqrt{2}$, or 1.41, times that represented by the next junction inward (toward the well, r_w) on the radial axis. If the scale of radial distances is shifted by a single network junction, the radius represented by any given junction is multipled or divided by $\sqrt{2}$, depending on the direction of the shift. Hence, the ratio of vertical to lateral hydraulic conductivity simulated by the model as a whole is divided by 2, and, by shifting the scale of the model a few junctions in either direction, a wide range of general anisotropy can be achieved.

The details of this procedure of varying the general anisotropy are described by Mundorff, Bennett, and Ahmad (1972). The technique is virtually equivalent to the familiar procedure of adjusting for anisotropy by scale distortion. The resulting changes in general anisotropy are effected without disturbing the relative heterogeneity built into the model to simulate sections of particularly high or low hydraulic conductivity.

The radius of influence simulated by the model was varied simply by extending or reducing the network along the radial axis. Addition or removal of a single row of resistors along the outer radial limit of the network alters the radius of influence by a factor of 1.41. With the addition or removal of relatively few rows, a variation of several fold can be achieved.

The techniques outlined in preceding paragraphs were utilized to vary the radius of influence and the general anisotropy during the series of preliminary analog experiments. The general anisotropy was varied in multiples of 2, from 19.5 to 312, whereas the radius of influence was varied in multiples of 2, from 1,024 to 32,768 feet. In the first experiment, increased resistances were used to simulate the Pleistocene clay beds between the surface and a depth of 60 feet; in subsequent experiments, closer agreement with field data was obtained without increasing the resistance in this interval. After the first experiment, the Pleistocene clay beds were simulated only through the general anisotropy of the network. In early experiments, a well radius of 0.71 foot was simulated; in later experiments, the simulated well radius was increased to 0.84 foot. Agreement between the model results and the field data were closer in the later expriments than in the early experiments. For well radii greater than 0.84 foot, the experimental results differed from the field data.

The experiments in the analog-model studies were made with a steady-state analog-model analyzer of the type developed by the Geological Survey. The well screen was simulated by a wire at the inner radial limit of the network that joined the two horizontal rows representing the 420- to 450-foot and 450- to 480-foot intervals. This wire was connected, in turn, to the ground terminal of the analyzer. The junctions along the upper surface of the model were connected by a second wire, which, in turn, was connected to the positive terminal of the analyzer. Because of the radial symmetry built into the model, this wire represented a circular horizontal plane. The potential along the wire was constant. Thus, the upper surface of the aquifer was represented in the model as a horizontal plane of zero drawdown. As noted in the section "General Characteristics of the Recharge-well Flow System," the upper surface of the aquifer at Bay Park includes areas of the water table and areas of the bay floor; strictly speaking, the upper surface is not a horizontal plane. However, although its shape could not be represented accurately, the simulation of zero drawdowns along this surface was consistent with the aquifer-test results.

Measurements at network junctions were made with the potentiometer of the analog analyzer. This instrument indicates the potential at any point as a fraction of the total voltage drop across the model. When the terminal representing the screen is held at zero voltage, this potential fraction at a given point in the network is equivalent to a drawdown fraction at the corresponding point in the aquifer; this fraction is

$$\frac{s_w - s}{s_w}$$

where s is the drawdown or head buildup at the given point and s_w is the drawdown or head buildup of the discharging or receiving well in response to pumping or injection.

The results of each analog experiment were compared with the aquifer-test results, by direct comparison of the potential fractions with the corresponding drawdown fractions and by comparison of certain ratios of vertical to lateral potential difference with the corresponding ratios of vertical to lateral drawdown difference. Approximate agreement between the potential distribution and the drawdown distribution was obtained with those networks in which the general anisotropy was approximately one one hundred fifty-sixth, and the radius of influence was 8,192 or 11,584 feet. As the general anisotropy was increased or decreased by multiples of 2, the potential distribution diverged from the observed drawdown distribution; and as the radius of influence was increased beyond 11,584 feet or decreased below 8,192 feet, by multiples of 1.41, the potential distribution again diverged progressively from the drawdown distribution.

The experimental results in closest agreement with the aquifer-test data were obtained with a network in which the general anisotropy was one one hundred fifty-sixth, the radius of influence was 8,192 feet, the well radius was 0.84 foot, and the representations of aquifer heterogeneity were as indicated in figure 11. This illustration shows the values of lateral and vertical hydraulic conductivities simulated by the analog model network in various intervals. The analog model actually simulates only relative hydraulic conductivity values; the values on the illustration were obtained by assuming that the lower half of the screened interval had a lateral hydraulic conductivity of 1,248 gpd per sq ft (167 ft per day) and the upper half a lateral hydraulic conductivity of 624 gpd per sq ft (83 ft per day). The remaining values were obtained from the relative hydraulic-conductivity distribution represented by the model. Assumed values for the screened interval give an average lateral hydraulic

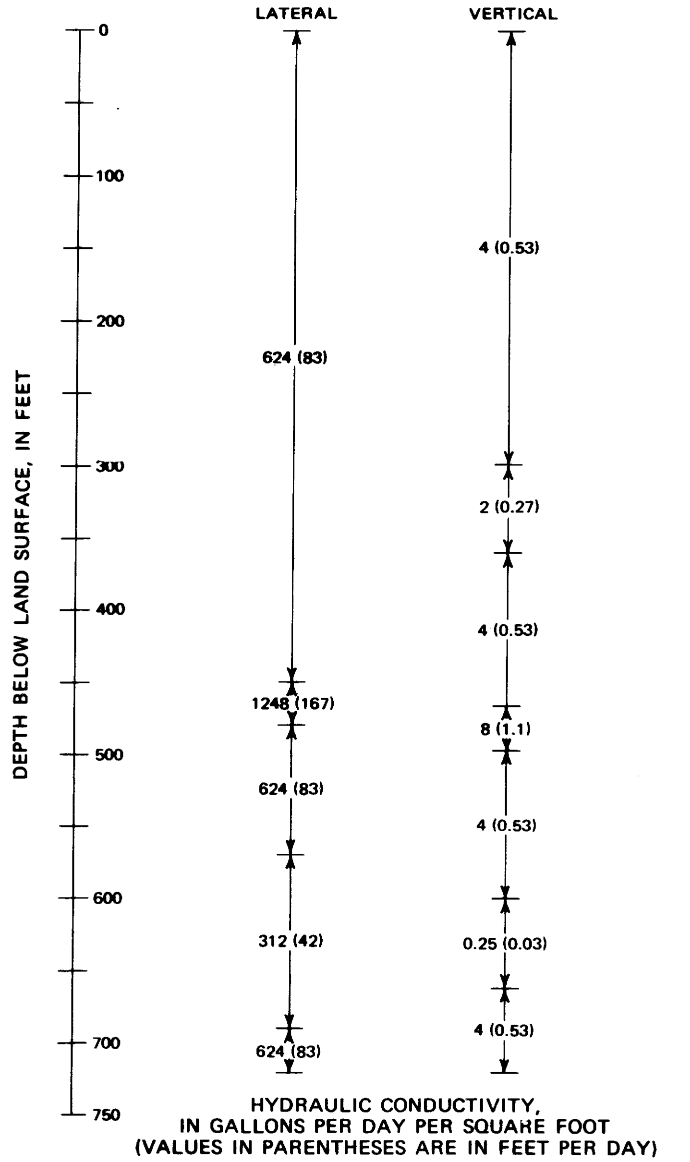

FIGURE 11.—Distribution of lateral and vertical hydraulic conductivities assumed in construction of analog network used in the analysis of the flow net.

conductivity of 940 gpd per sq ft (126 ft per day) for this interval, a value which agrees with the aquifer-test results. The assumption that the hydraulic conductivity of the lower half of the screened interval is twice that of the upper half was a simplification of the relations to achieve at least partial agreement with the flowmeter data. Those data indicated that in comparison with the material at other depths in the screened interval, the hydraulic conductivity of the material between a depth of 420 and 440 feet is relatively low.

Except for the interval between 570 and 690 feet, lateral hydraulic conductivity above and below the screen was assumed to be equal to that in the upper half of the screened interval. This assumption was based on lithologic and specific-capacity data for shallow and deep observation wells; these data suggest the material beyond the screened interval more closely resembles the material opposite the upper part of the screen than that in the more hydraulically conductive lower section. For the 570- to 690-foot zone, lateral hydraulic conductivity was assumed to be 312 gpd per sq ft (42 ft per day) or half the value used for the other material above and below the screened interval. This ratio was based on lithologic data, which indicate that roughly half the total thickness of this interval is clay.

The simulated vertical hydraulic conductivity for most of the section was 4 gpd per sq ft (0.53 ft per day). The simulated vertical hydraulic conductivity between 300 and 360 feet was 2 gpd per sq ft (0.27 ft per day), which represents the effects of the thin clay beds at the top and the bottom of this interval. Between 600 and 660 feet, the simulated vertical hydraulic conductivity was taken as 0.25 gpd per sq ft (0.03 ft per day), which represents the clay in this interval. For the section between 465 and 495 feet, a simulated vertical hydraulic conductivity of 8 gpd per sq ft (1.1 ft per day) was used to agree with the indication of the data that the lower part of the screened interval was generally higher in hydraulic conductivity.

Any representation of an aquifer by an analog network is an approximation, because the assumptions made in constructing the network never fully match field conditions. Moreover, the network is basically a finite-difference grid, representing a continuous medium. The model described is, at best, a rough approximation of the well-aquifer system at Bay Park. Simplifications were introduced in simulating the hydraulic conductivity distribution. As noted previously in this section, the representation of the upper surface of the aquifer as a horizontal plane is inaccurate. In addition, the network is terminated at a radius of influence representing 8,192 feet; if the equilibrium radius of influence of the recharge well and its rate of pumping are mutually dependent, the radius of influence simulated by the network will be accurate only at one rate of flow. Finally, the analog analysis does not simulate the effects of variable fluid density. If the flow system of the recharge well extends into saline or brackish water or if it is bounded by static saline interfaces, the flow pattern will be affected by density differences, but such alteration of the pattern will not be reflected in the analog results. Deficiencies in the analog approach place limitations on the significance of the results.

Measurements of potential were taken throughout the analog network. The technique described by Mundorff, Bennett, and Ahmad (1972) was used to construct a flow net for the recharge well. This flow net is shown on plate 2. Intervals of low hydraulic conductivity, as indicated by lithologic data, are shown as shaded areas on this illustration; flow lines are shown as solid lines and lines of equal drawdown, or equal head buildup, as dashed lines. The flow lines represent the intersections of three-dimensional stream surfaces with the *r-z* plane, whereas the lines of equal drawdown are the intersections of three-dimensional surfaces of equal drawdown with the *r-z* plane. The numbers identifying the flow lines are stream functions and indicate the fraction of the total well discharge enclosed by the stream surface; the numbers identifying the lines of equal drawdown are values of the drawdown fraction.

$$\frac{s_w - s.}{s_w}$$

The lines corresponding to values of this fraction less than 0.5 occurred in a logarithmic distribution between the 0.5 line and the well screen. The drawdown fraction is 1.0 along the upper surface of the flow net, which represents the water table and which was maintained as a surface of zero drawdown in the experiment. The stream function is 1.0 along the vertical direction beneath the well screen, along the lower boundary of the flow net, and along the vertical direction at the outer radial boundary of the flow net.

The refraction of the flow lines and lines of equal drawdown in crossing layers of low hydraulic conductivity was inserted arbitrarily in constructing the flow net; simulation of these layers by the anisotropy of the network and by increases in resistance within certain intervals was too generalized to yield this degree of detail directly in the experimental results. The flow net should be considered an interpretation of the analog results constructed to be consistent with those results and with general principles of flow in stratified media. It is an approximate representation of the system at Bay Park for rates of well operation in the range from 200 to 400 gpm.

The flow net of plate 2 indicates changes from an initial condition, rather than the actual distribution of head or pattern of flow in the aquifer. The analog experiments simulated the effects of operating the recharge well alone, whereas the flow pattern observed in the field is a superposition of the flow field of the recharge well on a variety of background features, such as the upward discharge from deeper aquifers toward the surface, the tidal oscillation, and the effects of pumping elsewhere in the aquifer. Changes from the initial condition are represented by lines of equal drawdown; drawdown, by definition, indicates the change in head caused by pumping. Flow lines, similarly, indicate changes in velocity caused by pumping. That is, if a velocity-change vector, Δv, defined in the r-z plane, gives the change in apparent ground-water velocity caused by operation of the recharge well, each stream line of plate 2 is constructed so that the vector Δv is tangent to the stream line at every point along its length. The stream function, ψ, is then defined so that the relations

$$\frac{1}{2\pi r}\cdot\frac{\partial\psi}{\partial r} = -\Delta v_z$$

and

$$\frac{1}{2\pi r}\cdot\frac{\partial\psi}{\partial z} = \Delta v_r \qquad (2)$$

hold true, where Δv_r and Δv_z are the r and the z components of the vector Δv. With the stream lines and lines of equal drawdown understood in this sense, the flow net of plate 2 represents the departure from original conditions caused by pumping the recharge well.

The actual flow pattern for given conditions may be approximated by superposition of the flow net on the conditions assumed to exist before operation of the recharge well. The operational flow pattern is obtained by (1) algebraically adding drawdowns indicated on the flow net to the preexisting heads and (2) by vectorially adding velocity changes indicated on the flow net to the preexisting velocity field. Changes in flow or head obtained from the flow net differ only in sign for recharge or discharge at the same rate; however, the resultant flow patterns obtained by superposing the flow net upon a given initial condition may differ considerably for the two types of flow.

Close to the discharging well, the head changes and the velocities associated with operation of the well tend to be much greater than those in the aquifer when the well is not pumped. Near the well, therefore, the original flow pattern can be ignored, and the flow net of plate 2 can be considered to be an approximation of actual flow conditions in the aquifer during operation of the well. At greater distances from the well, as velocities and head differences associated with well operation decrease to the order of magnitude of the background pattern, this approximation loses significance. Of the various background factors at Bay Park, the most significant for long-term operation is probably the upward movement of water from deeper aquifers. The straightforward combination of this effect with the flow net of plate 2 involves only the addition of a uniform upward velocity vector to the velocity field described by the flow net. For discharge from the recharge well, the resultant flow field would show the reorientation of flow lines from greater depths toward the well screen and a correspondingly lower intensity of flow toward the surface; whereas for recharge, the resultant field would show deflection of the regional flow pattern away from the well screen and a higher intensity of flow toward the surface. Finer subdivisions of the flow pattern of plate 2—for example, by the construction of flow lines representing increments of 0.01 in the stream function—is probably unwarranted because of the approximate nature of the analog simulation.

Internal hydraulics of observation well N7886.—The head distribution in the flow net of plate 2 indicates that vertical head differences appear within the 420- to 480-foot interval at relatively short distances (less than 20 ft) from the recharge well. The lines of equal-head buildup curve as they close around the screen, so that during recharge the head in the upper and the lower extremities of the screened interval must be slightly less than that in the central part of this zone at a given radial distance from the well. The anisotropy of the aquifer causes this effect to be more pronounced than it would be in an isotropic system, as the head gradients associated with vertical divergence of the flow are correspondingly greater. These vertical differences in head could cause vertical flow through the screen of observation well N7886, from the intervals of high head to those of lower head. During recharge, this flow would be directed from the central part of the screened interval toward the extremes (fig. 12) and would represent a short circuiting of the diverging flow pattern within the aquifer. In constructing figure 12, the limits of the contributing zone were set arbitrarily at 430 and 460 feet and do not necessarily represent conditions as they exist in the well. During pumping of the recharge well, the head in the central part of the screened interval would be lower than the heads at the extremes; and the directions of flow shown in figure 12 would be reversed. Knowledge of internal flow in the observation well could be important in the evaluation of the chemical analyses of the water obtained from the well.

Internal flow within a well screen depends not only on the head differences between various zones penetrated by the screen but also on the hydraulic conductivities of those zones as well. The low hydraulic conductivity in the upper part of the screened interval would tend to retard flow into this zone, whereas the higher hydraulic conductivity in the lower part of the screened interval would permit a somewhat greater flow into this zone. The flow shown in figure 12, from the middle of the screen toward the bottom, would probably be greater than the flow toward the upper zone. The magnitude of the flows involved is a matter of conjecture, as no measurements were made in the observation well screen; and, in general, the comments made here on the subject of flow in well N7886 represent only interpretation based on the general characteristics of the flow field of the recharge well.

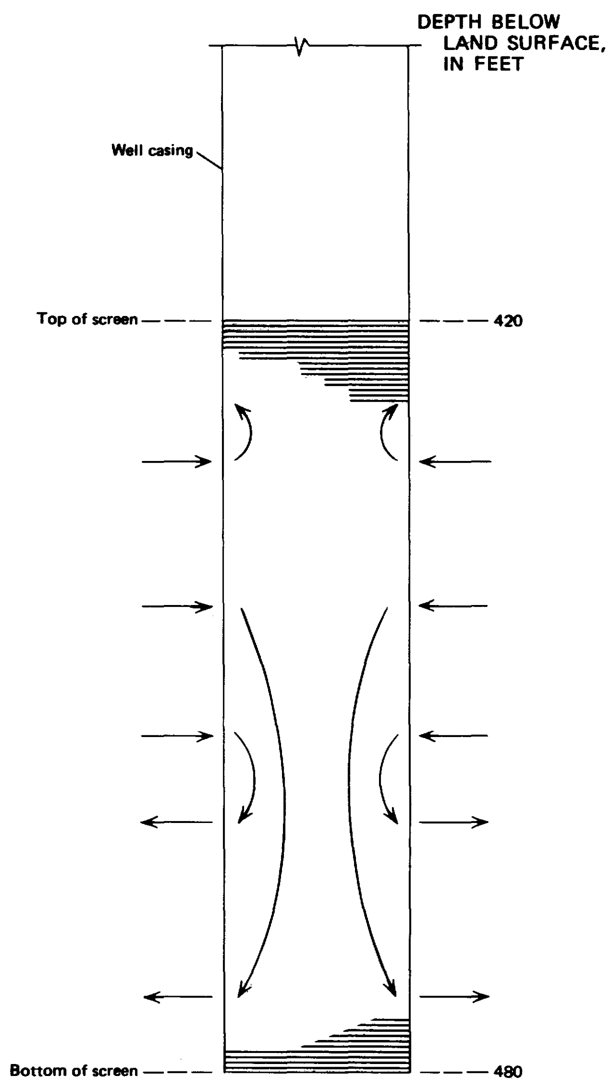

FIGURE 12.—Possible directions of internal flow (arrows) in well N7886 during recharge in well N7884.

HYDROCHEMICAL ENVIRONMENT

By F. J. PEARSON, JR.

GENERAL FEATURES

Water from the Magothy aquifer has an unusually low dissolved-solids content, average of 25 mg/l (milligrams per liter) because of the lack of readily soluble or reactive minerals in the aquifer. Most of the dissolved material in the ground water is derived from dissolved material in the water that recharges the aquifer. Natural recharge water contains material that has been dissolved in precipitation, concentrated by evapotranspiration, and modified by reactions with soluble minerals in the soil zone. For this reason, the concentrations of ions in soil water and in the water infiltrating into the aquifer are greater than those in the precipitation. The chemical composition of present-day precipitation has been measured in samples collected at several locations on Long Island (Pearson and Fisher, 1971). The chemical composition of precipitation corrected for evaporation is shown in table 3. Values for precipitation and evapotranspiration used in the calculation were those given by Cohen, Franke, and Foxworthy (1968, p. 58) for the water-budget area described by them. The average composition of water from the Magothy aquifer at Bay Park is also shown in table 3. The chloride contents of precipitation and ground water are identical, but concentrations of other ions in water from the two sources differ. Most of these differences are caused by soil-zone interactions and by changes in the chemistry of precipitation with time. The behavior of only a few ions need to be accounted for by intra-aquifer reactions.

Concentrations of calcium, magnesium, sodium, and potassium in the precipitation and in the ground water differ; however, the approximate agreements in the sums of their chemical-equivalent concentrations suggests a process in which the calcium and the magnesium in precipitaion are exchanged for

TABLE 3.—*Average chemical composition of atmospheric precipitation on Long Island and of water from wells screened in the Magothy aquifer at the Bay Park artificial-recharge site*

[Analyses by U.S. Geological Survey: Data from "Water Resources Data for New York, 1965—Part 2, Water Quality Records," available from U.S. Geol. Survey, Albany, N.Y., and U.S. Geol. Survey, unpublished data. Data expressed in milligrams per liter and milliequivalents per liter]

	Precipitation [1]		Well water	
	mg/l	meq/l	mg/l	meq/l
Silica (SiO_2)	0	----	7.3	----
Calcium (Ca)	1.7	0.08	.7	0.04
Magnesium (Mg)	.3	.02	.2	.02
Sodium (Na)	2.2	.10	3.8	.17
Potassium (K)	.3	.01	.6	.02
Ammonium (NH_4)	.5	.01	0	0
Ferrous Iron (Fe)	0	0	.2	.01
Hydrogen ion (H)	----	.10	----	0
Sum cations	----	.32	----	.26
Bicarbonate (HCO_3)	0	0	5	.08
Sulfate (SO_4)	10	.21	4.2	.09
Chloride (Cl)	3.8	.11	3.8	.11
Nitrate (NO_3)	.6	.01	0	0
Sum anions	----	.33	----	.28
Dissolved solids (sum)	20	----	26	----

[1] Adjusted for losses by evapotranspiration. See text discussion.

equivalent amounts of sodium and potassium. Clay minerals promote such exchange reactions, and ion exchange on clay minerals in the soil zone and in the aquifer is probably the reason for the differences in concentrations.

Nitrogen-containing ions are generally not naturally present in Long Island ground water, although they are present in significant quantities in precipitation. These ions may be depleted in passing through the soil zone because of their use by plants. Sulfate is considerably higher in the precipitation than in the ground water. Loss of sulfate from solution may be partly caused by the reduction of sulfate and its precipitation as sulfide minerals as discussed in the section, "Iron and Sulfur." More likely, though, the difference in concentration is due to the fact that recent atmospheric precipitation contains more sulfate than the original precipitation which was the source of the present water in the Magothy aquifer at Bay Park. In particular, burning of fossil fuels has increased the content of sulfurous gases and aerosols in the atmosphere (Pearson and Fisher, 1971).

The single most abundant dissolved constituent in water from the Magothy aquifer is silica. This species results from the simple solution of quartz, the dominant aquifer mineral. At 15°C (Celsius), the average temperature of water from the Magothy aquifer at Bay Park, 7.9 mg/l SiO_2 can result from quartz saturation. Water from the Bay Park wells is slightly undersaturated with respect to quartz, as the average silica content is only 7.3 mg/l.

Chemical analyses of water from the test wells at Bay Park are given in table 4. Temperature, pH, ferrous iron, total iron, and Eh were measured in the field. The other parameters were determined by the Geological Survey laboratory in Albany, N.Y.

OXIDATION-REDUCTION REACTIONS

The relative concentrations of dissolved oxygen, the carbonate species (HCO_3^{-1} and H_2CO_3, the latter used here to include dissolved CO_2), iron, and the sulfur-bearing species (SO_4^{-2} and H_2S) in water from the Magothy aquifer are the result of oxidation-reduction reactions. The solid substances in the aquifer involved in these reactions are an organic carbonaceous substance, possibly lignite here idealized as cellulose $(C_6H_{10}O_5)_n$, pyrite (FeS_2), and a ferric oxide or hydroxide mineral, probably hematite (Fe_2O_3).

OXYGEN AND CARBON

The dissolved-oxygen concentration of water in and near the area of recharge to the Magothy aquifer in north-central Long Island is about 10 mg/l or $3\times10^{-4}M$ (molar) (Nassau County Department of Health, unpub. data). The amount absorbed corresponds to an oxygen partial pressure of 0.19 atmospheres at 11°C, which is the temperature of the water in the recharge area and approximately the mean-annual air temperature. The partial pressure of oxygen in the free atmosphere is 0.21 atmospheres; but, in the soil atmosphere, the oxygen is somewhat depleted by organic decay process in the soil. Hence, the concentration of dissolved oxygen in water in contact last with the soil zone is less than oxygen concentration levels in the free atmosphere.

The concentration of all the dissolved carbonate species in water in the recharge area is about $3\times10^{-4}M$, which corresponds to about 5 mg/l bicarbonate (HCO_3^{-1}) plus 13 mg/l carbonic acid (H_2CO_3). In a system free of carbonate minerals at 11°C, this concentration would result from a CO_2 (carbon dioxide) partial pressure of about 3.5×10^{-3} atmospheres, which is considerably higher than the partial pressure of CO_2 in the free atmosphere (3×10^{-4} atm). In the soil zone, however, the CO_2 content may be several orders of magnitude higher than that in the free atmosphere because of plant decay and respiration; so water in contact with the soil atmosphere should reflect high partial pressures of CO_2 (Ingerson and Pearson, 1964, p. 268). Thus, it would seem that interactions between soil atmosphere and natural recharge water control the oxy-

TABLE 4.—*Chemical composition of water from test wells at the Bay Park artificial-recharge site*

[Chemical analyses by U.S. Geological Survey; results expressed in milligrams per liter. Number of significant figures shown reflects use of special methods of analyses]

Well / Screened interval (feet below land surface)	N2790 535–560	N7886 417–480	N7887 535–545	N7888 307–317	N7889 535–545	N7890 452–462	N7891 307–317	N8022 420–480
Silica (SiO_2)	7.4	7.2	7.5	7.0	7.5	7.5	7.1	7.4
Ferrous iron (Fe)	----	.30	.18	.28	.25	.14	.22	.14
Total iron (Fe)	.6	.30	.18	.28	.25	.14	.22	.14
Total manganese (Mn)	.01	.03	.01	.02	.03	.03	.05	.03
Calcium (Ca)	.34	.37	1.08	.68	.39	1.39	.45	.87
Magnesium (Mg)	.17	.22	.24	.24	.30	.21	.24	.24
Sodium (Na)	3.73	3.90	3.90	4.55	4.00	3.82	3.51	3.68
Potassium (K)	.60	.53	.68	1.24	.50	.62	.52	.44
Bicarbonate (HCO_3)	6.0	4.5	7.5	5.2	5.0	7.5	4.0	6.0
Sulfate (SO_4)	4.10	3.55	4.00	5.0	3.90	3.95	4.15	3.75
Chloride (Cl)	3.75	3.90	3.78	4.8	3.75	3.90	3.95	3.85
Fluoride (F)	.01	.13	.10	.0	.11	.13	.10	.13
Nitrate (NO_3)	.00	.00	.00	.15	.00	.00	.00	.00
Phosphate (PO_4)	.015	.02	.02	.03	.12	.005	.00	.015
Dissolved solids (calculated from sum of determined constituents)	23	22	24	26	23	25	22	23
Specific conductance (micromhos per cm at 25°C)	27	28	32	35	28	32	28	28
pH	5.6	5.22	5.58	5.42	5.25	5.72	5.15	5.45
Oxidation-reduction potential (volts)	----	—.03	—.08	—.12	—.05	—.10	—.06	—.06
Hydrogen sulfide (H_2S)	----	.4	.0	.5	.0	.0	.3	.0
Temperature (°C)	15	15	15	15	15	15	15	15

gen and the carbonate contents of the water just entering the saturated zone.

Downgradient from the recharge area, the dissolved-oxygen content of the ground water decreases, and, correspondingly, the total carbon concentration increases. At distances 8 or 9 miles downgradient from the recharge area, the dissolved-oxygen content of the ground water has decreased to zero; and the total carbon has leveled off at 6 to $7 \times 10^{-4} M$. These changes are probably partly caused by the oxidation of organic material, such as lignite, by the dissolved oxygen through a reaction that can be idealized as

$$(C_6H_{10}O_5)_n + nH_2O + 6nO_2 \rightarrow 6n\ H_2CO_3 \qquad (3)$$

The stoichiometry of this reaction is such that if all the initially present dissolved oxygen is converted to H_2CO_3, about $3 \times 10^{-4} M$ of H_2CO_3 will be produced. Adding this amount to the initial soil-atmosphere-derived carbonate would give a carbon content total of about $6 \times 10^{-4} M$, approximately the measured value.

The energetics of reaction 3 are favorable for the production of H_2CO_3. The value of the free energy of formation of cellulose is not readily available, but the free energy of reaction 3 can be approximated several ways. Edsall and Wyman (1958, p. 236) give the free energy of formation, $\Delta G_f°$, of glucose ($C_6H_{12}O_6$) as —217.56 kcal (kilocalories) and that of glycogen $(C_6H_{10}O_5)_n$ as $-158.3n$ kcal. Although glycogen is not structurally identical with cellulose, it is similar; so the free energy of its reaction with oxygen should be close to that of cellulose reaction. The free energy of a reaction like 3 involving glycogen is $-679n$ kcal. Another method of approximation is to assume that glucose, rather than cellulose, is the oxidized species of lignite. Its oxidation can be written

$$C_6H_{12}O_6 + 6O_2 \rightarrow 6H_2CO_3 \qquad (4)$$

The free energy of this reaction is —676 kcal. The near identity of the free energies of these reactions involving compounds structurally related to cellulose makes it reasonable to assign a free energy of about $-680n$ kcal to reaction 3. From this value, the relation between the species O_2 and H_2CO_3 becomes

$$\log P_{O_2} = \log H_2CO_3 - 83$$

At H_2CO_3 concentrations of the magnitude of $10^{-3} M$, which are those found at Bay Park and elsewhere downgradient from the recharge area, the equivalent P_{O_2} is of the order of 10^{-86} atmospheres. This extremely low value is below the stability limit of water and certainly does not represent the natural conditions. It does show, however, that oxidation of lignite can readily occur.

The carbon-isotope composition of the carbonate species in the water from the Magothy aquifer corroborates the preceding interpretations (Pearson and Friedman, 1970). The ratios of the stable-carbon isotopes, carbon-12 and carbon-13, can be used to distinguish carbonate from various sources—for example, marine limestone, free atmospheric CO_2, and plant sources. Determinations of the carbon-isotope ratios from the water of the Magothy aquifer show that near the recharge area the dissolved carbonate is of plant origin. Its biological derivation confirms the influence of the soil atmosphere rather than the free atmosphere on the water and also confirms the petrographically observed lack of carbonate minerals in this part of the aquifer. In contrast, carbon-isotope data show that about half the additional dissolved carbonate appearing deeper in the aquifer in the Bay Park area is derived from plants and half from carbonate minerals. The derivation is, thus, by oxidation of organic matter or by simple solution of carbonate minerals, as in the reaction

$$H_2CO_3+CaCO_3 \rightarrow Ca^{+2}+2HCO_3^- \qquad (5)$$

The reaction is displaced to the right in such acidic dilute water, and the fact that so little dissolved carbonate results from it suggests the presence of only very small amounts of carbonate minerals in the aquifer. This lack is consonant with the observed petrography of the aquifer.

IRON AND SULFUR

Iron-bearing minerals are present in the Magothy aquifer, and, under proper conditions, they react with the water so that ferrous iron goes into solution. In and near the recharge area, the iron in the aquifer occurs as stains and coatings on the quartz grains or is associated with the dispersed lignite masses. In this part of the aquifer, the water contains measurable amounts of dissolved oxygen, and, thus, is strongly oxidizing. Because the iron is in the ferric oxidation state (Fe^{+3}) as solid oxides or hydroxides, most of the compounds are highly insoluble in water. Thus, there is little or no iron in solution in this region. Deeper in the aquifer, where dissolved oxygen is lacking, ferrous iron appears in solution in the water (Nassau County Department of Health, unpub. data). At some distance from the recharge area, pyrite is found associated with lignite, and the water in some wells has a noticeable odor of hydrogen sulfide. The ferrous iron concentration of water from the Magothy test wells at Bay Park is 0.2 mg/l. Water from at least one of the wells contains small but distinctly odoriferous amounts of hydrogen sulfide. The water resembles that from many other deep Magothy wells, and the conditions prevailing at the Bay Park site can be taken as representative of regional conditions.

Oxidation-reduction reactions are conveniently shown on Eh-pH diagrams. Construction and use of the diagrams are described in detail by Garrels and Christ (1965, chap. 7). Data obtained from Bay Park wells are plotted on the Eh-pH diagram shown in figure 13. The water has an ionic strength of 5×10^{-4} moles per liter and a total sulfur-species content of $4.4\times10^{-5}M$ (equivalent to a sulfate content of 4.2 mg/l). The boundaries of fields containing dissolved ferrous iron were drawn for two Fe^{+2} concentrations—$1.8\times10^{-6}M$ and $1.8\times10^{-5}M$—equal to 0.1 mg/l and 1.0 mg/l iron, respectively. The two boundaries bracket the values of iron concentration in water from the Bay Park wells and also indicate the magnitude of changes in dissolved-iron concentration with Eh and pH changes.

The system shown in figure 13 is for a temperature of 15°C, the average temperature of the Magothy water at Bay Park. The Van't Hoff equation was used to convert the standard free energies at 25°C to the free energies at 15°C. In the calculation, dH/dt was assumed equal to zero. Though not rigorously true, the assumption should introduce only negligible errors when the departure from the standard temperature is only 10°C. The values of ΔH° and ΔG° used in these calculations are given by Latimer (1952).

Of several Eh values measured for water from each well, the most negative one is probably the most accurate one. The ranges of all measurements are indicated by the length of vertical lines on figure 13, as is the probable precision of the lowest individual measurement. Figure 13 shows that the Magothy water at Bay Park is below the stability field of hematite (Fe_2O_3). Hematite is the most stable of the various ferric oxide and hydroxide species likely in the oxidizing part of the aquifer near the recharge area. The fact that even hematite is unstable in the Bay Park water means that no solid ferric species are likely in the aquifer there. Unlike the iron content of many ground waters (Oborn and Hem, 1962; Barnes and Back, 1964), the iron content of the ground water at Bay Park is not controlled by the solubility of such solids.

Except for well N7888, the measured Eh values fall on or approach the boundary defined by the reaction

$$FeS_2 + 8H_2O \rightarrow Fe^{+2} + 2SO_4^{-2} + 16H^+ + 14e^- \qquad (6)$$

This condition suggests that FeS_2 (pyrite) is the solid phase controlling iron concentration. In this part of the aquifer, the iron dissolved from oxides and hydroxides upgradient should be precipitating as pyrite. This observation agrees with the petrographic observation that at Bay Park pyrite occurs generally in association with lignite.

Water in well N7888 has an Eh value significantly lower than would be indicated by reaction 6, and it is close to the boundary representing equimolar concentrations of H_2S and SO_4^{-2} for the reaction

$$H_2S + 4H_2O \rightarrow SO_4^{-2} + 10H^+ + 8e^- \qquad (7)$$

Water in well N7888 has a significant H_2S content, so the Eh of the water in terms of reaction 7 is reasonable. The iron content of water in well N7888 far exceeds that indicated by reaction 6 at this Eh. Berner (1967 a, b) discusses the behavior of iron sulfides precipitating from aqueous solutions and states that the initial precipitate is commonly a noncrystalline phase of composition FeS (ferrous sulfide), which has a standard free energy of formation of —21.3 kcal at 25°C. This phase then slowly transforms to the stable mineral, in the Magothy system, pyrite. The iron content of the water from well N7888 is about $10^{-6}M$, which is between the iron content that would be expected in this water from solutions of noncrystalline FeS, $10^{-3}M$, and pyrite, $10^{-11}M$. Possibly the initial iron sulfide phase precipitating is noncrystalline FeS.

There are also discrepancies between the measured Eh values and the chemistry of water in wells N7886 and N7891. Both these wells contain H_2S, yet their Eh values are well within the stability field of SO_4^{-2}. These difficulties may be due to the analytical procedures used to determine the sulfur species, or, more likely, they are due to erroneously high Eh measurements.

In some aquifer systems, the solubility of siderite ($FeCO_3$) seems to control the dissolved iron content of the water (Langmuir, 1969). Solution of siderite in water having pH and total carbon values of the water at Bay Park would permit a ferrous iron content of about $5\times10^{-2}M$. This solubility limit is well above the actual concentration of iron, so siderite cannot be a controlling solid phase here.

Various manganese-bearing species also undergo oxidation-reduction reactions in natural-water systems. The manganese content of Magothy water is low (0.05 mg/l or less), and because stability diagrams of the manganese system are given elsewhere (Bricker, 1965), no calculations were made to describe manganese here. In the oxidizing part of the aquifer near the recharge area, manganese is probably present as one of several oxide or hydroxide minerals. With decreasing values of Eh, but above those necessary to dissolve Fe_2O_3, MnO_2 (manganese dioxide) is reduced to Mn^{+2} (aq), the stable phase.

In the discussion of carbon and oxygen, the oxidation of lignite was proposed to account for loss of dissolved oxygen in the aquifer. Such oxidation can also occur in the absence of free oxygen by the reaction:

$$(C_6H_{10}O_5)_n + 13nH_2O \rightarrow 6nH_2CO_3 + 24nH^+ + 22ne^- \qquad (8)$$

Assuming the H_2CO_3 concentration observed at Bay Park, a pH of 6, and a temperature of 25°C, the Eh for reaction 8 is about —0.4 v. This Eh is below the stability field of water and probably nowhere does the water reach this theoretical lower limit, but this reaction certainly provides a likely mechanism for lowering the Eh of water in the Magothy aquifer and allowing the various iron and sulfur reactions to occur.

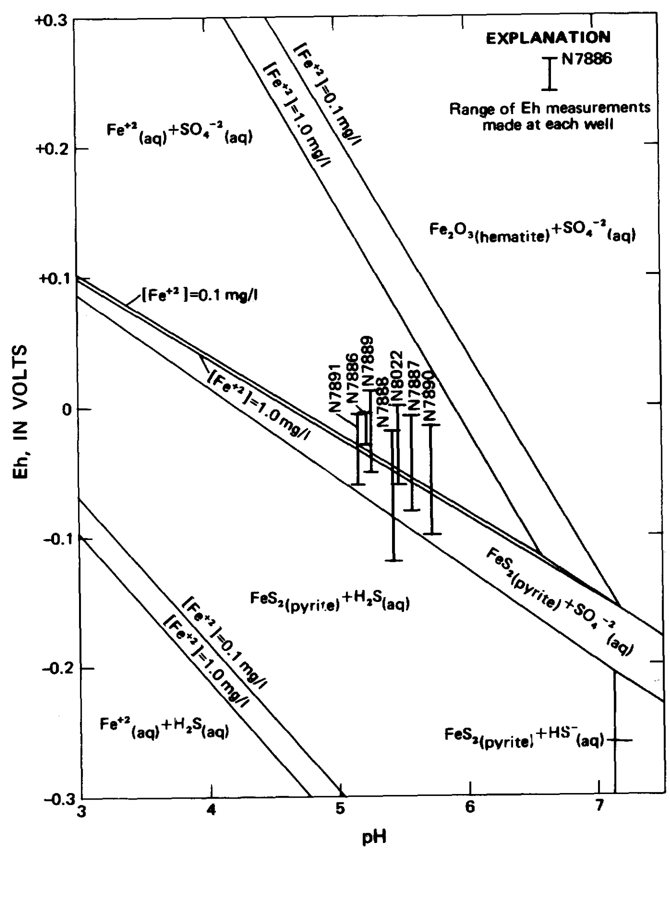

FIGURE 13.—Eh-pH diagram for the system Fe-S of water from the Magothy aquifer at Bay Park, at 15°C, 1 atmosphere of pressure and sulfur species activity of 4.2 mg/l as SO_4^{-2}.

SUMMARY

In order to define conditions prevailing before injection of reclaimed water (highly treated sewage-plant effluent) at Bay Park the geology, the hydrology, and the hydrochemistry of the recharge site were studied in detail before the recharge tests were started.

The recharge site is in the Atlantic Coastal Plain and is underlain by about 1,250 feet of unconsolidated deposits of Pleistocene and Late Cretaceous age, which, in turn, overlie crystalline bedrock of Precambrian age. The recharge well is screened within the lower part of the Magothy aquifer of Late Cretaceous age, at a depth of 420 to 480 feet below land surface. The 60-foot injection zone is mostly a stratified fine to medium slightly silty quartz sand containing lignite, pyrite, and muscovite as common accessory minerals.

The semiconfined injection zone lies between beds having lower hydraulic conductivity. The Magothy aquifer at Bay Park, as a whole, can be considered to be an anisotropic, stratified, but virtually unconfined aquifer. Average lateral hydraulic conductivity of the injection zone is estimated to be 920 gpd per sq ft (126 ft per day), but flowmeter surveys indicate considerable variation within this 60-foot interval. Vertical hydraulic conductivity of the material between the water table and the top of the injection zone is in the range of 2 to 20 gpd per sq ft (0.27 to 2.7 ft per day). The flow system around the injection well has been studied extensively through the use of standard aquifer-test techniques, as well as through electrical analog-model studies.

The very low dissolved-solids content of water from the Magothy aquifer (25 mg/l at Bay Park) is due to a lack of readily soluble minerals in the aquifer. Most of the dissolved-solids content is derived from material in water that recharges the aquifer. Silica, the principal constituent of Magothy water, results from the solution of quartz, the dominant aquifer mineral. Oxidation of lignite probably accounts for some of the carbonate content of Magothy water in the Bay Park area. Pyrite is considered the solid phase controlling dissolved iron concentration of a few tenths of a milligram per liter. Water from each of the test wells has a slightly negative Eh and a pH within the range from 5.2 to 5.7.

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