Hydrogeology and Analysis of Ground-Water-Flow System, Sagamore Marsh Area, Southeastern Massachusetts

U.S. Geological Survey
Water-Resources Investigations Report 96-4200

Prepared in cooperation with the
U.S. ARMY CORPS OF ENGINEERS

U.S. DEPARTMENT OF THE INTERIOR
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By DONALD A. WALTER, JOHN P. MASTERSON, and PAUL M. BARLOW

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Marlborough, Massachusetts
1996

U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, Secretary

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CONVERSION FACTORS, VERTICAL DATUM, AND WATER-QUALITY UNITS

CONVERSION FACTORS

Multiply	by	To obtain
acre	4,047	square meter
cubic foot per day (ft^3/s)	0.02832	cubic meter per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter square per day
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	2.54	centimeter
inch per year (in/yr)	2.54	centimeter per year
mile (mi)	1.609	kilometer
million gallons (Mgal)	3.785	million liters
million gallons per day (Mgal/d)	0.04381	cubic meter per second
square foot per day (ft^2/d)	0.09290	square meter per day
square mile (mi^2)	2.590	square kilometer

Air temperature is given in degrees Fahrenheit (˚F), which can be converted to degrees Celsius (˚C) by the following equation:

˚C = (˚F-32)/1.8.

VERTICAL DATUM

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic vertical datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level of 1929.

WATER-QUALITY UNITS

Chemical concentration is given in units of milligrams per liter (mg/L) or micrograms per liter (µg/L). Milligrams and micrograms per liter are units expressing the mass of the solute per unit volume (liter) of water. One thousand micrograms per liter is equivalent to 1 milligram per liter. Micrograms per liter is equivalent to "parts per billion." Milligrams per liter is equivalent to "parts per million." Chemical concentration is also given in units of milligrams per kilograms (mg/kg). One milligram per kilogram is equivalent to 1 microgram per gram. Milligram per kilogram is equivalent to "parts per million."

Hydrogeology and Analysis of Ground-Water-Flow System, Sagamore Marsh Area, Southeastern Massachusetts

By Donald A. Walter, John P. Masterson, *and* Paul M. Barlow

Abstract

A study of the hydrogeology and an analysis of the ground-water-flow system near Sagamore Marsh, southeastern Massachusetts, was undertaken to improve the understanding of the current (1994-95) hydrogeologic conditions near the marsh and how the ground-water system might respond to proposed changes in the tidal-stage regime of streams that flood and drain the marsh. Sagamore Marsh is in a coastal area that is bounded to the east by Cape Cod Bay and to the south by the Cape Cod Canal. The regional geology is characterized by deltaic and glaciolacustrine sediments. The sediments consist of gravel, sand, silt, and clay and are part of the Plymouth-Carver regional aquifer system. The glacial sediments are bounded laterally by marine sand, silt, and clay along the coast. The principal aquifer in the area consists of fine to coarse glacial sand and is locally confined by fine-grained glaciolacustrine deposits consisting of silt and sandy clay and fine-grained salt-marsh sediments consisting of peat and clay. The aquifer is underlain by finer grained glaciolacustrine sediments in upland areas and by marine clay along the coast.

Shallow ground water discharges primarily along the edge of the marsh, whereas deeper ground water flows beneath the marsh and discharges to Cape Cod Bay. Tidal pulses originating from Cape Cod Bay and from tidal channels in the marsh are rapidly attenuated in the subsurface. Tidal ranges in Cape Cod Bay and in the tidal channels were on the order of 9 and 1.5 feet, respectively, whereas tidal ranges in the ground-water levels were less than 0.2 foot. Tidal pulses measured in the water table beneath a barrier beach between the marsh and Cape Cod Bay were more in phase with tidal pulses from Cape Cod Bay than with tidal pulses from the tidal channels in Sagamore Marsh, whereas tidal pulses in the regional aquifer were more in phase with tidal pulses from the tidal channels.

A 5-day aquifer test at a public-supply well adjacent to the marsh gave a transmissivity of the regional aquifer of 9,300 to 10,900 feet squared per day and a hydraulic conductivity of 181 to 213 feet per day, assuming a saturated thickness of the aquifer of 51.3 feet. The regional aquifer became unconfined near the pumped well during the test. The ratio of tidal ranges in the tidal channel to the ranges in the underlying aquifer at two sites (the lower and upper marsh) indicated aquifer diffusivities for the marsh sediments of 380 and 170 feet squared per day; these values correspond to hydraulic conductivities of 2.5×10^{-3} and 1.7×10^{-3} feet per day, respectively. The maximum distances from the tidal channel at the lower and upper marsh sites where tidal ranges would exceed 0.01 foot, as calculated from aquifer diffusivities and current (1995) tidal ranges in the tidal channels, were 24.4 and 26.7 feet, respectively. The maximum distances from the tidal channel where tidal pulses in the ground water would exceed 0.01 foot, using potential increased tidal stages resulting from proposed tidal-stage modifications and predicted by the U.S. Army Corps of Engineers, were 37.1 and 42.0 feet, respectively.

A numerical model of the marsh and surrounding aquifer system indicated that the contributing area for the supply well adjacent to the marsh, for current (1994) pumping conditions, extends toward Great Herring Pond, about 2 miles northwest (upgradient) of the well, and does not extend beneath the marsh. The model also indicates that the predicted increases in tidal stages in the marsh will have a negligible effect on local ground-water levels.

INTRODUCTION

Sagamore Marsh is a salt marsh in the coastal area of southeastern Massachusetts that is bounded to the east by Cape Cod Bay and to the south by the Cape Cod Canal (fig. 1). The construction and maintenance of the Cape Cod Canal from 1883 until the late 1930's greatly altered the geography, hydrology, and ecology of the marsh; the primary result of canal construction and maintenance was a reversal in tidal-flow directions in the marsh and a decrease in the amount of saltwater flowing into the marsh (Carlisle, 1994). Predevelopment drainage in the marsh was to the east into Cape Cod Bay through a tidal channel known as the Scusset River. Sediments from canal dredging and natural long-shore sediment transport effectively closed the Scusset River in 1932, and inputs of fresh surface water and precipitation resulted in the formation of a large body of ponded freshwater in the marsh. A trench cut between the marsh and the canal in the early 1930's caused the ponded water to drain and allowed saltwater to enter the marsh from the south. At present, saltwater enters the marsh through a 4-foot diameter culvert at the canal. The present-day ecology of the marsh was investigated by Oliver and Swain (1994). Carlisle (1994) discusses in detail the history of development in and around Sagamore Marsh as well as the current ecology, geography, and hydrology of the marsh. The hydraulics of salt-marsh sediments are discussed by Knott and others (1987). The ecology and evolution of New England salt marshes are discussed in detail by Nixon (1982) and Teal (1986).

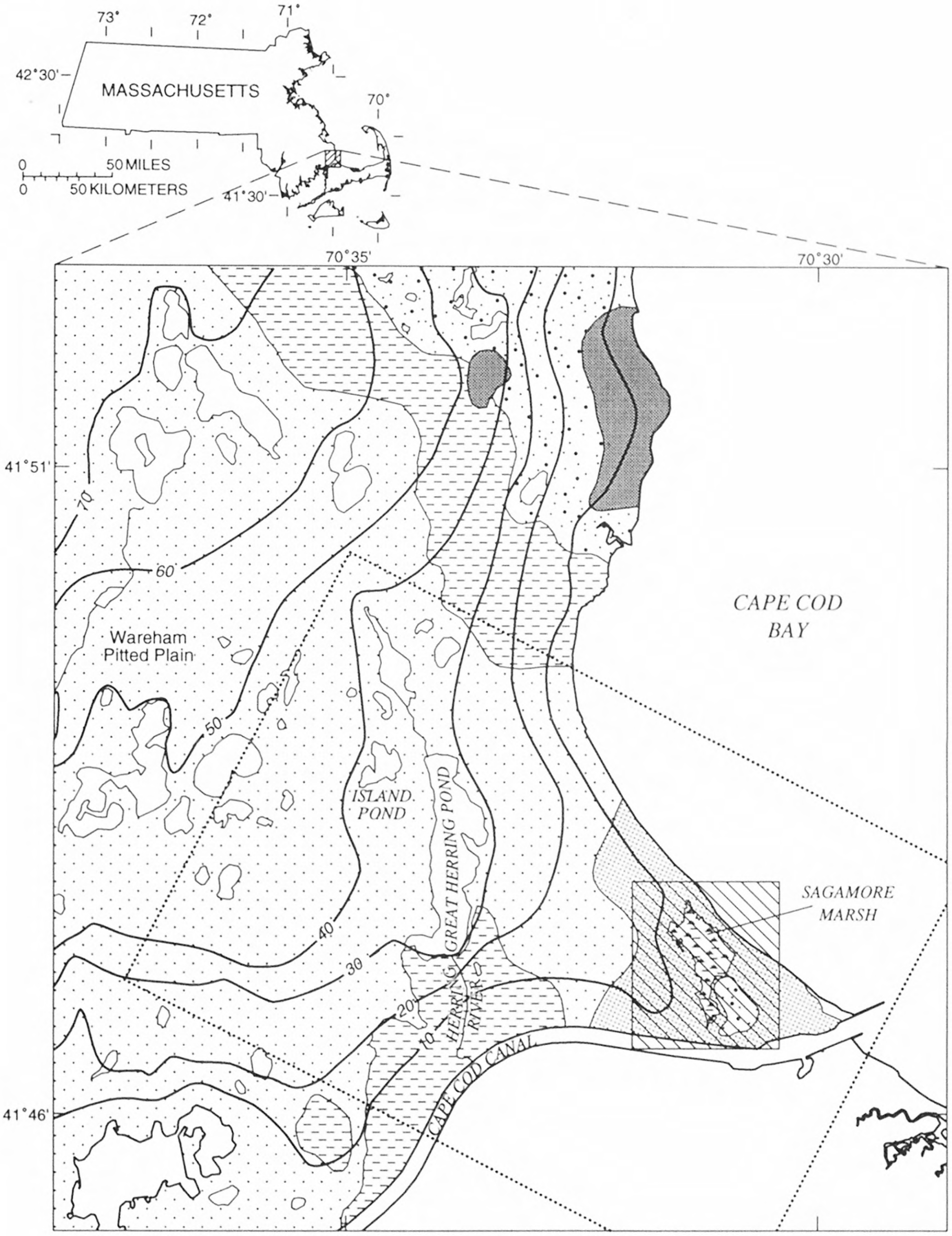
Sagamore Marsh, which currently encompasses an area of about 300 acres, is predominantly a brackish-water marsh that is dominated by common reed (*Phragmites australis*); lower parts of the marsh regularly receive overbank flooding and are dominated by salt-marsh cord grass (*Spartina alterniflora*). *Phragmites* vegetation can tolerate salinities as much as 20,000 mg/L and is commonly associated with brackish-water marshes. This species is typically found along the edges of healthy salt marshes and can invade marsh interiors only when salinities are decreased.

In an effort to restore salt-marsh and estuarine habitat in Sagamore Marsh, the U.S. Army Corps of Engineers (USACE) has proposed to increase the amount of saltwater that enters the marsh by increasing the size of the culvert between the Cape Cod Canal and the marsh. The increase in salt-water inflow will increase the range of tidal fluctuations in the marsh channels and will result in increased overbank flooding at high tide in parts of the marsh; overbank flooding will increase salinities sufficiently to allow salt-marsh vegetation to replace the *Phragmites*. The proposed salt-marsh restoration is anticipated to restore about 50 acres of degraded marsh (Matthew Walsh, U.S. Army Corps of Engineers, oral commun., 1995).

Existing data are limited regarding the hydrogeologic framework and the ground-water-flow system in the area around Sagamore Marsh. In April 1995, the U.S. Geological Survey (USGS), in cooperation with the U.S. Army Corps of Engineers, began an investigation into the geology and ground-water hydrology of the Sagamore Marsh area to improve the understanding of the local hydrogeologic framework and ground-water-flow system and to address issues regarding the potential effects of the proposed marsh-restoration project on the local ground-water system. The investigation was done in part to determine whether the proposed increase in the amount of saltwater entering the marsh at high tide would be likely to increase hydraulic heads in the underlying aquifer, which could affect septic systems near the marsh. The investigation also was done to determine whether the proposed restoration of the marsh would be likely to cause saltwater intrusion into the aquifer, which is the source of water for a large-capacity public-supply well adjacent to the marsh. The extent of the study area is shown in figure 1 and includes a total area of about 27.2 mi^2; the study area corresponds to the part of the regional aquifer that was assessed using a ground-water-flow model. Data were collected in a 2.0 mi^2 area in and near the marsh, which is shown as the field study area in figure 1.

Purpose and Scope

The purpose of this report is to describe the hydrogeology and ground-water-flow system of the Sagamore Marsh area and to evaluate possible changes in the flow system caused by increased tidal stages in the tidal channels. Specifically, the report discusses (1) the stratigraphy of hydrogeologic units in the Sagamore Marsh area; (2) the patterns of ground-water flow and the effects of ground-water pumping and tidal fluctuations on the ground-water-flow system near the marsh; (3) the delineation of the zone of contribution and source of water for a public-supply well adjacent to the marsh; and (4) how the ground-water-flow system may respond to increased tidal stages in the marsh. The report also includes the results of analytical- and numerical-modeling analyses of the ground-water-flow system.



Base from U.S. Geological Survey digital data
1:24,000, 1991
Universal Transverse Mercator projection
Zone 19

0 1 2 MILES
0 1 2 KILOMETERS

EXPLANATION

SURFICIAL GEOLOGY

- Outwash
- Moraine
- Kame
- Till
- Glaciolacustrine deposits

GEOLOGIC CONTACT

AREA OF FIELD STUDY-- A detailed map is shown in figure 2

BOUNDARY OF STUDY AREA--Boundary is coincident with the boundary of the finite-difference model grid shown in figure 13

—*10*— WATER-TABLE CONTOUR--Shows altitude of water table (from Hansen and Lapham, 1992). Contour interval is 10 feet. Datum is sea level

Figure 1. Location of Sagamore Marsh study area, regional water table, and surficial geology of Plymouth-Carver Aquifer, southeastern Massachusetts.

Regional Physiographic, Geologic, and Hydrologic Setting

The regional geology and hydrology of southeastern Massachusetts have been investigated previously by Frimpter (1973), Williams and Tasker (1974), and Hansen and Lapham (1992). Southeastern Massachusetts is in the Coastal Lowlands physiographic province of New England. Land surface is characterized by knob and kettle topography associated with glacial moraines and flat, gently sloping topography associated with glacial outwash plains. Land-surface altitudes range from 250 ft above sea level in the northwestern part of the Plymouth-Carver area to sea level along the coast.

The study area is at the southeastern edge of a glacial aquifer system known as the Plymouth-Carver aquifer (fig. 1). The aquifer system, which encompasses an area of about 140 mi^2 in southeastern Massachusetts, includes several aquifers and confining units. A detailed analysis of the ground-water-flow system in the Plymouth-Carver aquifer is reported in Hansen and Lapham (1992). The aquifer system consists of unconsolidated glacial sand, silt, and clay that was deposited during the Wisconsin Stage of the Pleistocene Epoch (Mather, 1952). Depositional environments for the glacial sediments include moraines, fluvial outwash plains, and glacial lakes. The marsh is near the southeastern edge the Wareham Pitted Plain. This area is a gently sloping plain with numerous kettle holes and ponds. The Wareham Pitted Plain represents a large fluvial outwash plain that formed to the south of the retreating Laurentide ice sheet, about 15,000 years ago. The sediment source lies to the northwest of the plain near the Hog Rock and Ellisville moraines. The glacial outwash deposits consist primarily of stratified sand and gravel and frequently show a fining-downward sequence. The surficial outwash deposits grade laterally to finer grained glaciolacustrine deposits in the area around Sagamore Marsh; these sediments were deposited in a proglacial lake that may be associated with the Wareham Pitted Plain outwash deposits (B.D. Stone, U.S. Geological Survey, oral commun., 1995).

Ground-water levels in the Plymouth-Carver aquifer range from about 125 ft above sea level at the top of the water-table mound to the northwest of the Sagamore Marsh area to sea level at the coast (Hansen and Lapham, 1992). Near Sagamore Marsh, ground-water levels range from about 10 ft above sea level near the western edge of the marsh to sea level at the coast. The regional ground-water-flow system receives about 24 to 27 in. of recharge annually from direct infiltration of precipitation (Hansen and Lapham, 1992). Ground-water flows radially outward from the center of the aquifer system (at the top of the water-table mound) and discharges to Cape Cod Bay, Cape Cod Canal, and coastal embayments and tidal creeks, including those of Sagamore Marsh. Regional ground-water flow in the study area is to the southeast; water-table altitudes in the southeastern part of the aquifer are shown in figure 1.

The regional aquifer system generally is unconfined; however, confined and semiconfined conditions do occur locally. Where the aquifer is confined (or semiconfined), the vertical movement of water is restricted by overlying deposits of low vertical and horizontal hydraulic conductivity. Ground-water flow in the southeastern part of the aquifer system is greatly affected by Great Herring Pond (fig. 1). The 376-acre pond, which is about 2 mi northwest of Sagamore Marsh, has an average depth of about 20 ft, a maximum depth of about 40 ft, and a surface altitude of about 34 ft (Massachusetts Division of Fisheries and Wildlife, 1993).

Estimates of average horizontal hydraulic conductivity of stratified sand and gravel deposits of the Plymouth-Carver aquifer system, based on the analysis of 33 aquifer tests done for public and industrial supplies and lithologic data from test well sites, ranged from 55 to 313 ft/d, with a mean of 188 ft/d (Hansen and Lapham, 1992). Hansen and Lapham (1992) reported that horizontal hydraulic conductivities of till deposits in the Plymouth-Carver aquifer system ranged from 10 to 100 ft/d. Specific yield of the unconfined aquifer system, estimated from the results of 22 aquifer tests, ranged from 0.02 to 0.35, with a mean of 0.16.

The Plymouth-Carver aquifer system is underlain by crystalline bedrock. The bedrock surface ranges from an altitude of 100 ft above sea level in the western part of the aquifer to 200 ft below sea level in the southeastern part of the aquifer. A narrow, northwest-southeast trending bedrock valley transects the southeastern part of the aquifer; bedrock-surface altitudes in the valley are more than 200 ft below sea level (Hansen and Lapham, 1992).

Acknowledgments

The authors appreciate the assistance of several property owners near the Sagamore Marsh who provided access to the marsh and allowed the installation of drive points and (or) observation wells on their property. The assistance of Peter Phippen and William Salomma, Massachusetts Department of Environmental Management and of Paul Gibbs and Sean Anderson, North Sagamore Water District during the aquifer test, also are appreciated.

METHODS OF INVESTIGATION

A network of 12 observation wells, 9 drive points, and 3 tidal-channel stilling wells was installed in and around Sagamore Marsh (fig. 2). The wells were installed near an existing public supply well, which was included in the monitoring network. Physical data for all sites used in the study are presented in table 1. The 2.5-inch PVC (polyvinylchloride) observation wells were installed using a hollow-stem auger drill rig. The 0.75-inch steel drive points were installed using a hand-operated sliding hammer, and the 2.5-inch PVC stilling wells were installed using a hand-operated steel hammer. Field methods used in the installation of drive points and the collection of marsh-sediment cores are those described by Weiskel and others (1995).

A 5-day aquifer test was done in June 1995 to determine the response of the ground-water-flow system near the marsh to pumping of the large-capacity public-supply well BHW013 (known as the North Sagamore Water District Beach Well) and to determine the hydraulic properties of the regional aquifer near the marsh. Ten of the 12 observation wells and the supply well were instrumented with Drucks[1] pressure transducers to continuously measure water levels during the 5-day pumping phase and the 5-day recovery phase of the test.

Water-level data from the network of wells, drive points, and stilling wells were used to map the water table and the potentiometric surface in the Sagamore Marsh area and to assess the effects of tidal fluctuations in the marsh tidal channels and Cape Cod Bay on ground-water levels in and around the marsh. Ground-water levels at all sites were measured during a 3-hour period (within 1.5 hours of high tide) in June 1995 using a hand-operated electric tape. Six of the nine drive points and the three tidal-channel stilling wells were instrumented with Drucks pressure transducers to continuously measure tidal-related water-level fluctuations in the marsh tidal channels and in the ground water from June 29 through July 18, 1995.

Lithologic data collected during the study and the results of the aquifer test were used to develop a steady-state, finite-difference ground-water-flow model of the Sagamore Marsh area. The USGS numerical model MODFLOW (McDonald and Harbaugh, 1988) and the particle-tracking program MODPATH (Pollock, 1994) were used in the analysis. The model was used to determine the contributing area of well BHW013 and to estimate marsh-restoration-related changes in high tide ground-water levels around the marsh. An existing finite-difference model of the regional Plymouth-Carver aquifer system (Hansen and Lapham, 1992) was used to estimate boundary conditions and regional hydrogeologic conditions for the model developed for the Sagamore Marsh area.

[1]Any use of trade, product, or firm names in this publication is for descriptive purposes only and does not imply endorsement by the U.S. Geological Survey.

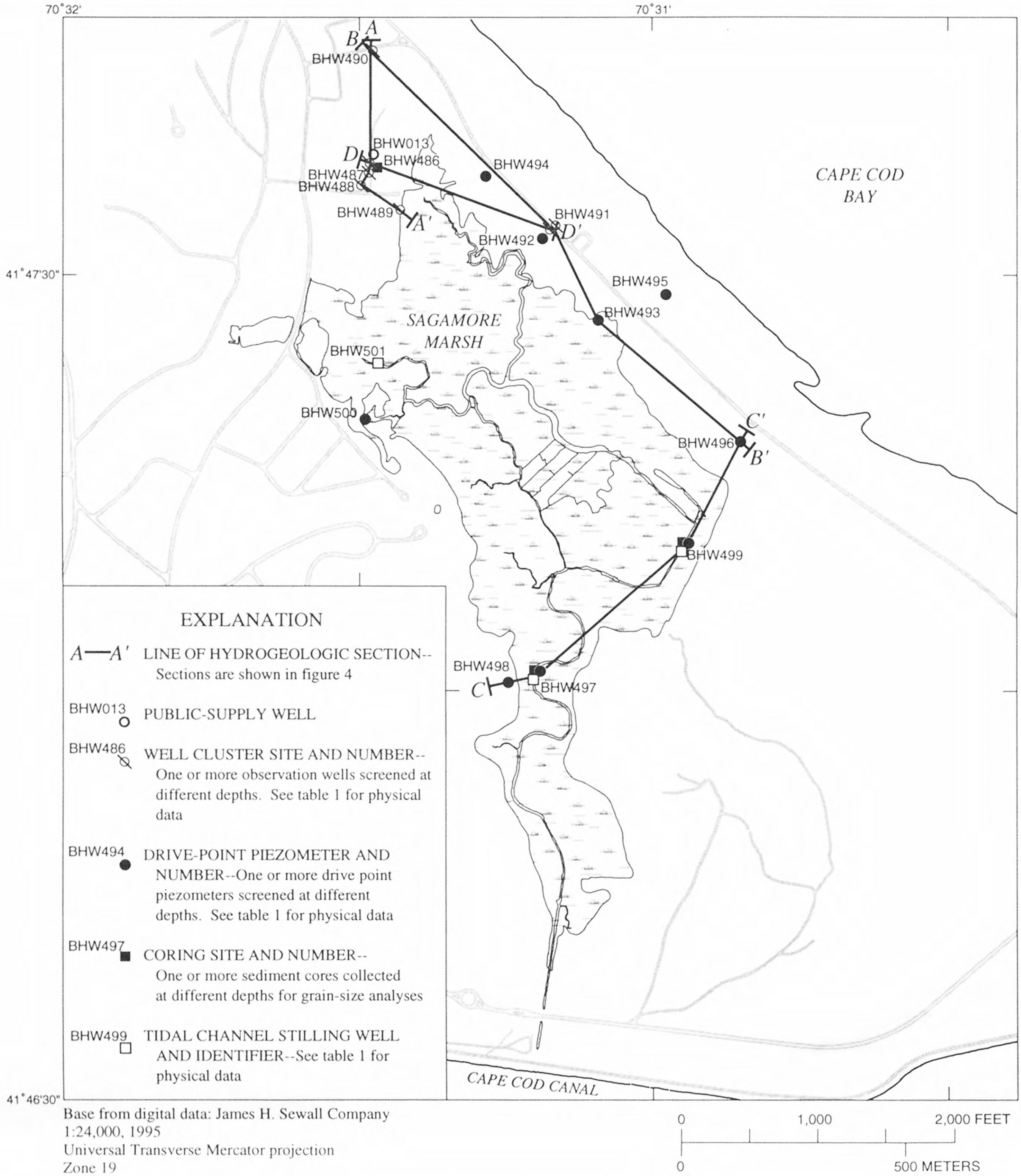


Figure 2. Location of wells, drive points, coring sites, and stilling wells used for data collection and location of geologic sections near Sagamore Marsh, southeastern Massachusetts.

Table 1. Physical data for observation wells, drive points, and stilling wells used for collection of hydrologic data, near Sagamore Marsh, southeastern Massachusetts

[Site identifier: Location of sites shown in figure 2. Type: Supply, 10.0-inch diameter steel well; well, 2.5-inch diameter PVC polyvinylchloride well; DP, 0.75-inch diameter steel drive point; tidal, 2.5-inch diameter PVC tidal channel stilling well. Latitude/Longitude given in degrees, minutes, and seconds. Altitudes are in feet above or below (-) sea level. ft, foot]

Site identifier	Well depth (ft)	Type	Latitude	Longitude	Altitude of land surface	Altitude of top of screen	Altitude of bottom of screen	Remarks
BHW013	54	Supply	414735	0703131	12.29	-32.71	-40.71	--
BHW486	15	Well	414734	0703131	16.26	8.81	3.81	Split-spoon cores taken at depths of 175 ft below land surface. Natural gamma and electromagnetic induction logs collected.
	55	Well	414734	0703131	16.18	-33.18	-35.18	
	175	Well	414734	0703131	16.23	-152.47	-154.47	
BHW487	15	Well	414733	0703131	17.81	9.49	4.49	Natural gamma and electromagnetic induction logs collected.
	52	Well	414733	0703131	18.12	-32.22	-34.22	
BHW488	13	Well	414733	0703132	17.59	10.15	5.15	Do.
	53	Well	414732	0703132	17.50	-30.87	-32.87	
BHW489	10	Well	414731	0703128	10.28	6.48	1.48	Do.
	49	Well	414731	0703128	10.45	-37.05	-39.05	
BHW490	150	Well	414743	0703131	10.18	-19.62	-21.62	Do.
BHW491	7	Well	414729	0703113	6.48	6.45	1.45	Do.
	55	Well	414729	0703113	6.29	-44.69	-46.69	
BHW492	5	DP	414729	0703114	4.78	1.14	.14	Continuous core samples collected at a depth of 4-8 ft below land surface. Hand-operated sand-coring tool was used.
BHW493	5	DP	414723	0703109	3.69	-.22	-1.22	Do.
BHW494	5	DP	414733	0703121	5.24	1.71	.71	Do.
BHW495	5	DP	414724	0703103	11.22	8.41	7.41	Do.
BHW496	5	DP	414714	0703056	4.45	1.42	.42	Do.
BHW497	3	Tidal	414657	0703116	--	1.90	-.10	Continuous core samples collected at a depth of 20 ft below land surface. Hand-operated peat-coring tool was used.
	19	DP	414657	0703116	2.88	-16.48	-17.48	
BHW498	5	DP	414656	0703119	4.63	.51	-.49	Continuous core samples collected at a depth of 4-8 ft below land surface. Hand-operated sand-coring tool was used.
BHW499	3	Tidal	414706	0703101	--	2.84	.84	Continuous core samples collected at a depth of 20 ft below land surface. Hand-operated peat-coring tool was used.
	21	DP	414706	0703101	2.81	-22.46	-23.46	
BHW500	5	DP	414715	0703133	4.50	2.50	1.50	Continuous core samples collected at a depth of 4-8 ft below land surface. Hand-operated sand-coring tool was used.
BHW501	3	Tidal	414720	0703131	--	1.01	-.99	--

HYDROGEOLOGY

Sagamore Marsh is at the edge of a large, complex glacial aquifer system that is laterally bounded by marine sediments along the coast. Ground-water flow, which is locally confined, generally is to the southeast and water levels in the aquifer are affected by nearby Cape Cod Bay and the Cape Cod Canal. Lithologic and hydrologic data collected during this investigation were used to develop a better understanding of the local ground-water-flow system near Sagamore Marsh.

Hydrogeologic Units

Sagamore Marsh is bounded to the west and southeast by glacial upland terrain, to the east by artificial fill related to construction of the Cape Cod Canal, and to the northeast by a barrier beach. Lithologic data from core samples and drillers logs and natural gamma logs were used to determine the hydrogeologic framework of the glacial, marine, and marsh sediments in the Sagamore Marsh area.

Glacial Sediments

Geophysical and lithologic data indicate that glacial sediments at site BHW486 consist of four major lithologic stratigraphic units. The split-spoon core samples and the natural gamma log show a fine-grained confining unit that extends from land surface (18 ft above sea level) to an altitude of about 2 ft above sea level (fig. 3). The lithologic unit consists of brown silt and sandy clay that probably represent Lake Cape Cod glaciolacustrine deposits (Oldale and Barlow, 1986); these sediments were deposited in a large pro-glacial lake that formed north of present-day Cape Cod. The lithologic unit underlying the silt and clay consists of fine to coarse brown sand with some gravel and extends to about 45 ft below sea level. This sediment probably represents coarser grained deltaic deposits related to deposition in Lake Cape Cod (B.D. Stone, oral commun., 1995). The lithology of the unit underlying the sand aquifer is highly variable from an altitude of 45 to 102 ft below sea level. The unit is characterized by interbedded gravel, fine to coarse sand, silt, and clay. The clay and silt layers, which are brown or gray in color, commonly contain large cobbles. Sandy layers, which typically are dense and compact, range in color from red and brown to black. This unit probably represents glaciolacustrine deposits associated with deltaic deposits of the Wareham Pitted Plain (B.D. Stone, oral commun., 1995), which lies to the northwest of the study area. The highly variable unit is underlain by very fine to fine gray sand and silt that extends to the bottom of the borehole at 160 ft below sea level. This unit is glaciolacustrine in origin and probably also is associated with the deltaic deposits of the Wareham Pitted Plain. The Sagamore Marsh area is within a southeast-trending valley in the underlying bedrock surface.

The stratigraphic units at site BHW486 correspond to four major hydrogeologic units. The first hydrogeologic unit, which is the brown silt and sandy clay associated with Cape Cod Lake deposits, is a confining unit near land surface; the water table is within this unit at site BHW486. This confining unit is underlain by fine to coarse brown deltaic sand that comprises the second hydrogeologic unit. This sand unit has the most favorable water-transmitting properties in the stratigraphic column at the site and comprises the major aquifer in the area around the marsh. A nearby public-supply well, BHW013, is screened in the sand unit at a depth of 33 to 41 ft below sea level. The third hydrogeologic unit is the interbedded gravel, sand, silt, and clay unit and the fourth hydrogeologic unit is the fine gray silty sand unit; both of these lower units are associated with Wareham Pitted Plain glaciolacustrine deposits. Although the interbedded sand, silt, and clay unit probably has greater water-transmitting properties than the fine sand and silt unit, both have relatively poor water-transmitting properties and probably are not important aquifers in the area.

The stratigraphy of the glacial sediments along the northwestern side of the marsh is shown in figure 4*A*. Lithologic data and natural gamma logs indicate that the brown silt and sandy clay were near the surface at sites BHW486, BHW487, and BHW488, but not at site BHW489, suggesting that the confining unit may be discontinuous; sandy clay also was at the surface along the western edge of the marsh at sites BHW498 and BHW500 (fig. 2) to the south of section *A-A'*. Natural gamma logs indicate that the sand aquifer underlying the confining unit contains some finer grained material at sites BHW489 and BHW490. A contact between marine and glacial sediments was at an altitude of about 25 ft below land surface at site BHW490 near Cape Cod Bay. The marine sediments consisted of fine to medium gray sand with interbedded gray clayey peat.

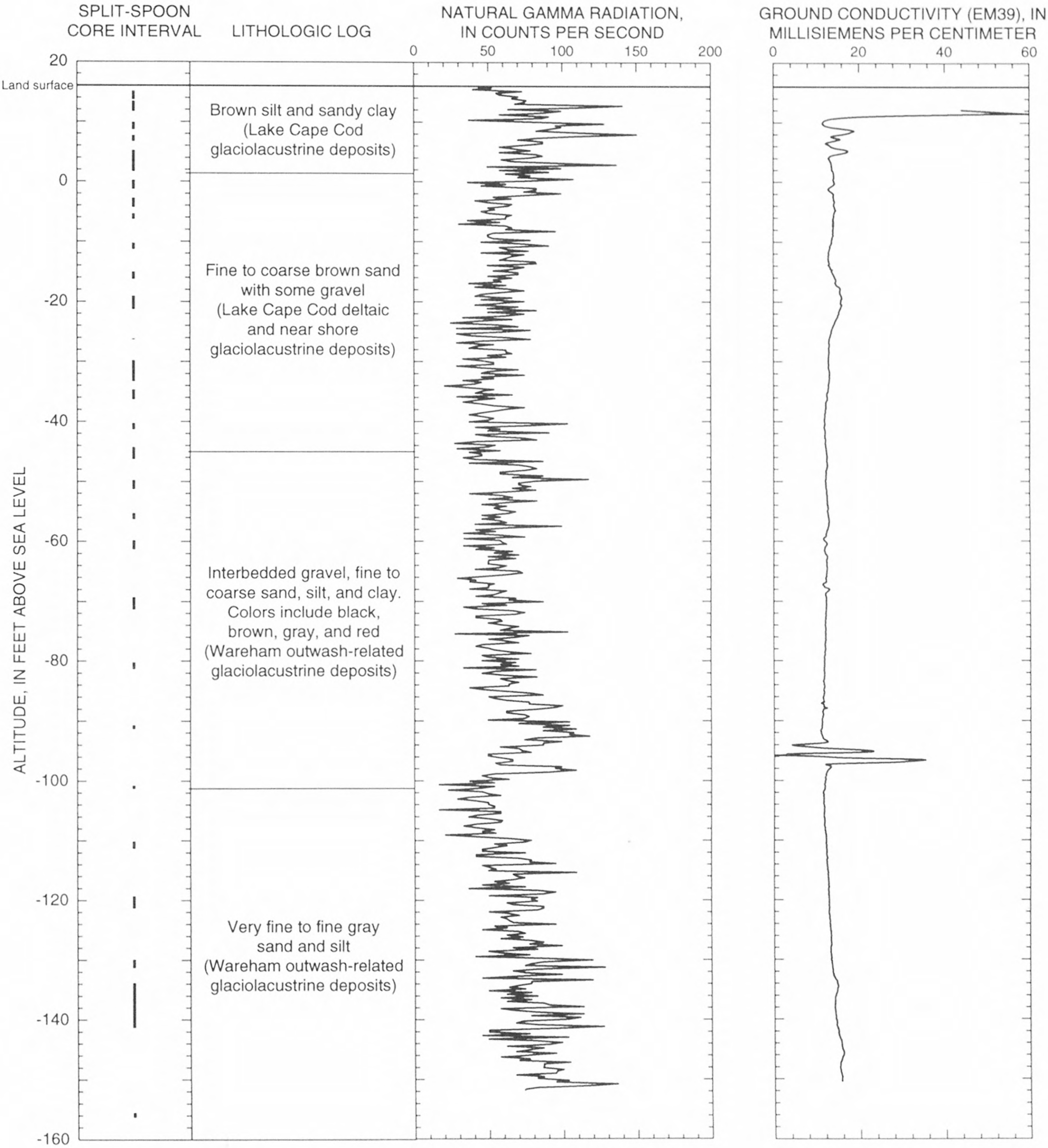
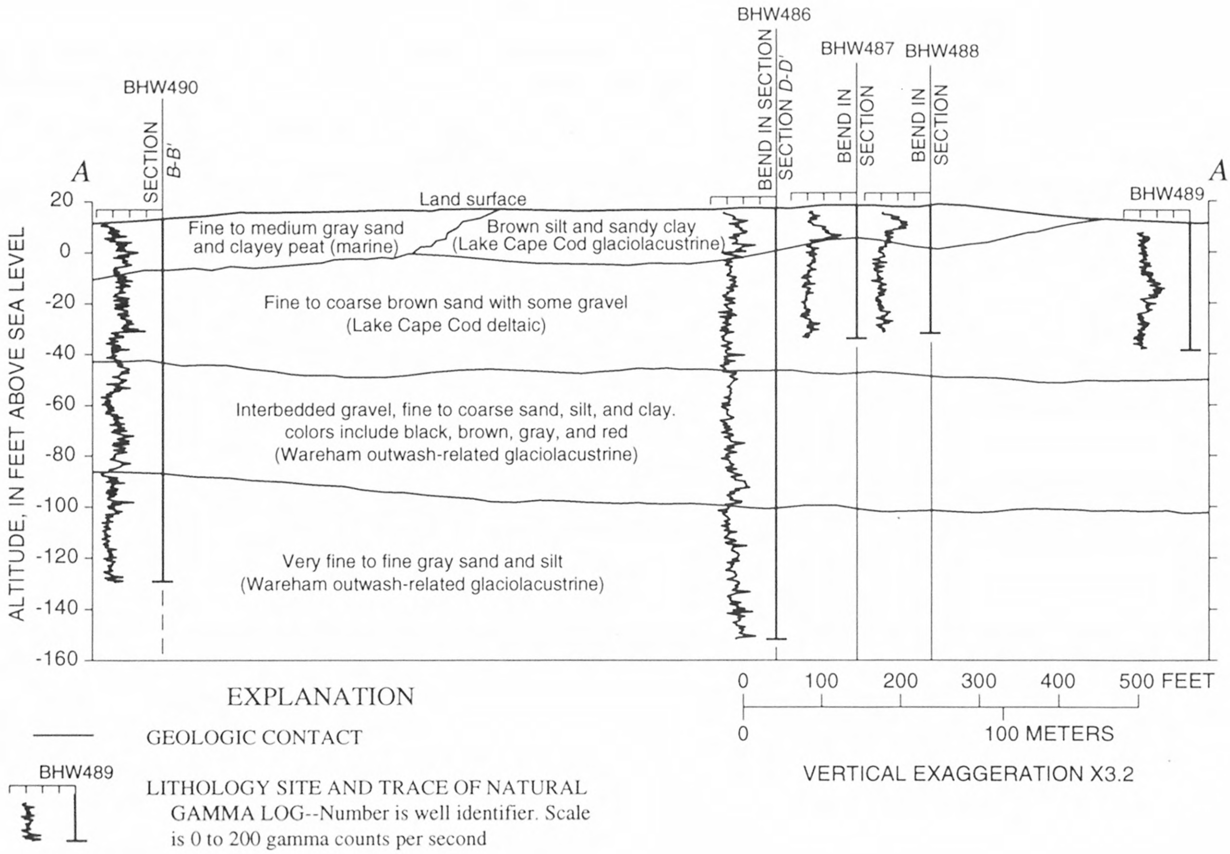


Figure 3. Natural gamma, ground conductivity (EM39), and lithologic logs for site BHW486 near Sagamore Marsh, southeastern Massachusetts.



***A*. Section *A-A* '.**

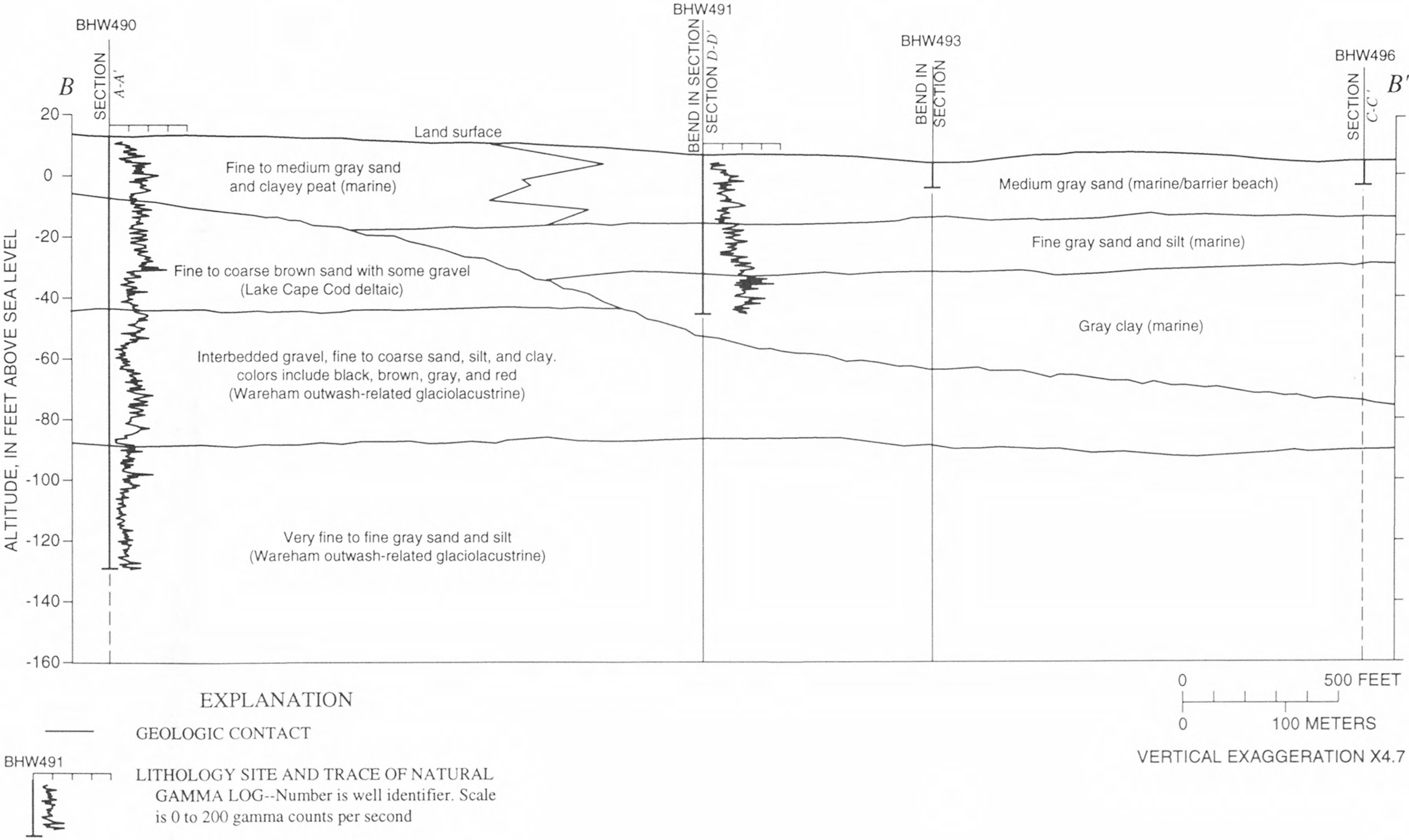
Figure 4. Stratigraphy of major hydrogeologic units in the Sagamore Marsh area along (*A*) section *A-A*', (*B*) section *B-B* ', (*C*) section *C-C* ', and (*D*) section *D-D* ', southeastern Massachusetts. (Lines of sections are shown in figure 2.)

Marine Sediments

Stratigraphy along the northeastern edge of the marsh is shown in figure 4*B*; this section lies along the marsh side of the barrier beach. Gray marine sand was observed near the surface at all sites along the section. Gamma and drillers logs at site BHW491 show that the medium sand extends to a depth of about 20 ft below land surface and is underlain by fine gray sand and gray marine clay. The fine gray sand, silt, and peat is about 15 ft thick at site BHW491. The gray marine clay, which contains numerous mollusk shells, is at least 15 ft thick and probably represents the lower boundary of the beach aquifer. Glacial sediments underly the marine sand, silt, and clay; the erosional contact between marine and glacial sediments is at a depth of about 25 ft below land surface at site BHW490 and greater than 45 ft below land surface at site BHW491. Beach deposits along the coast of the barrier beach at site BHW495 (fig. 2) consist of coarse brown sand and gravel.

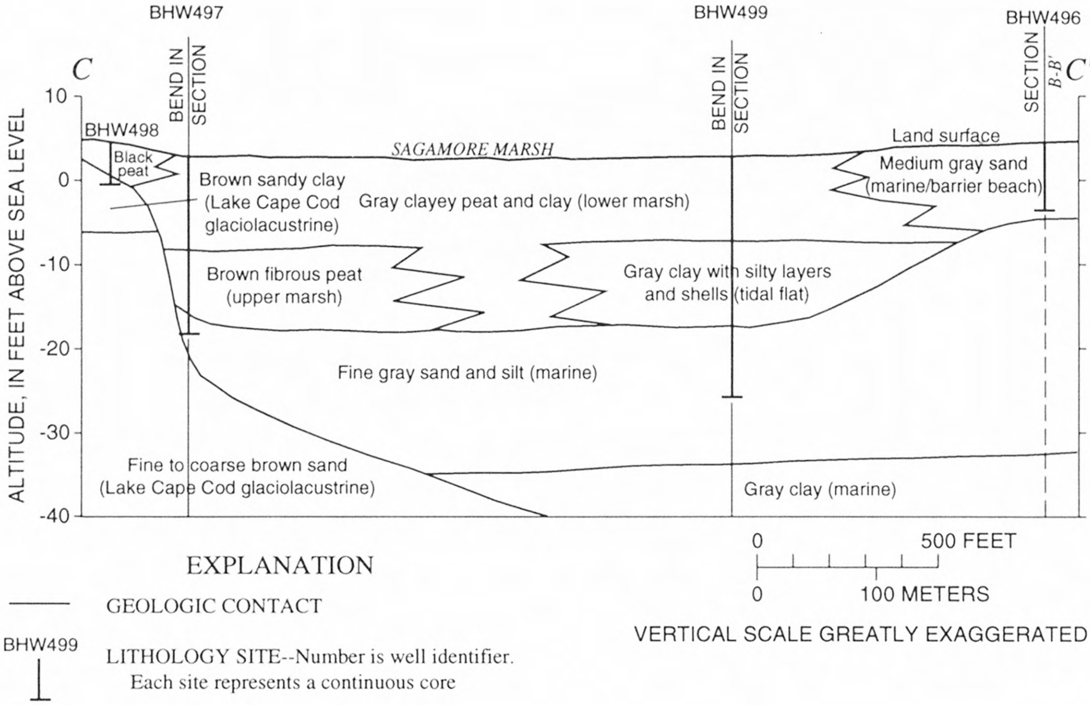
Marsh Sediments

Salt-marsh sediments in Sagamore Marsh range in thickness from 0 ft along the edge of the marsh to more than 20 ft near the center of the marsh (fig. 4*C*). Marsh sediments at site BHW497 include about 12 ft of gray clayey peat and clay underlain by about 8 ft of brown fibrous peat; the total thickness of peat and clay at the site was about 20 ft. Fine gray sand and silt underlies the peat. The brown fibrous peat at the site is consistent with "high-marsh" deposits (Nixon, 1982). Although site BHW497 is close to the present day entrance of the marsh at the Cape Cod Canal, historically the marsh drained to the east into Cape Cod Bay, which indicates this location is analogous to a "high-marsh" environment. Sediments at site BHW498, at the western edge of the marsh consisted of about 4 ft of black and gray peat underlain by brown sandy clay consistent with the glaciolacustrine deposits seen near the surface in section *A-A*'.



B. Section *B-B'*.

Figure 4.—*Continued.*

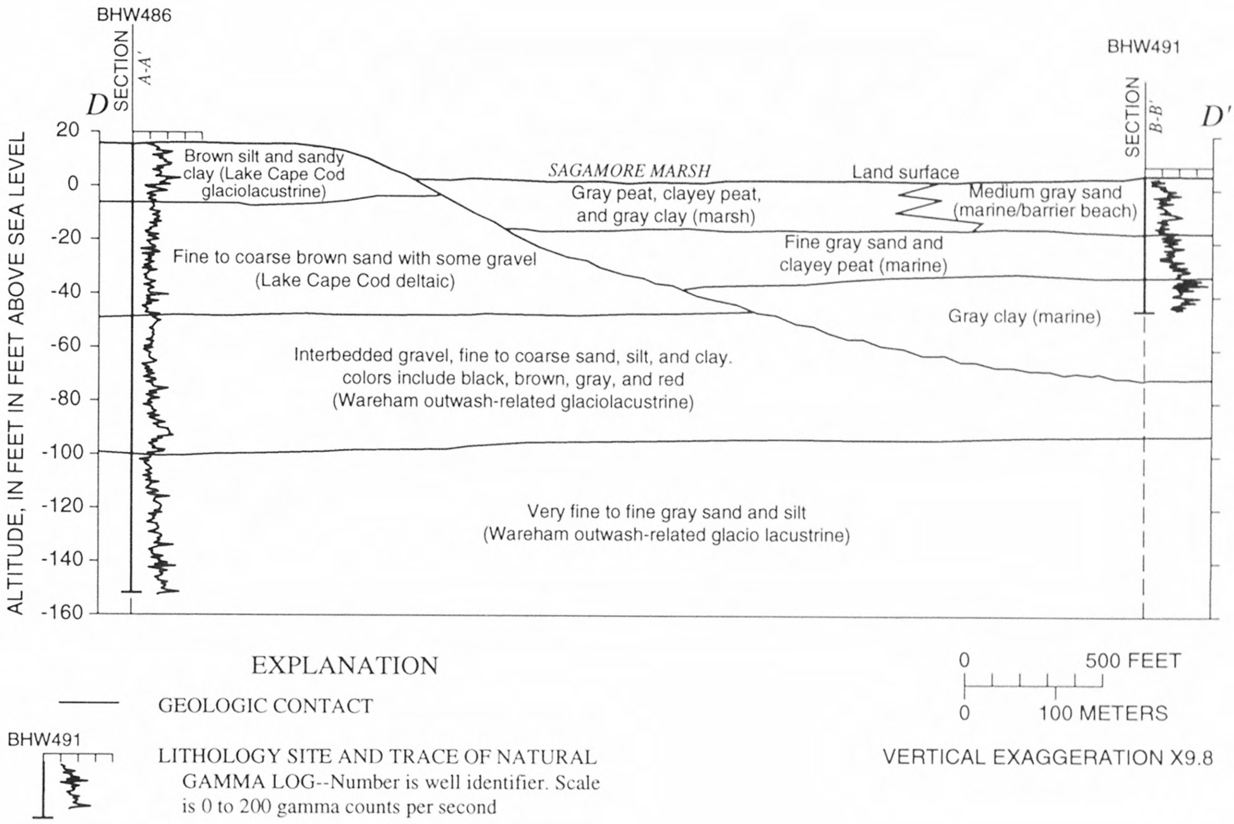


***C*. Section *C-C'*.**

Figure 4.—*Continued.*

The fine-grained marsh sediments (peat and clay) at site BHW497 decrease rapidly in thickness toward the edge of the marsh—from about 20 ft to about 4 ft over a distance of 212 ft. Marsh sediments at site BHW499, which are finer grained than at site BHW497, show about 10 ft of gray clayey peat and clay underlain by about 8 ft of gray marine clay; the total thickness of fine-grained marsh sediments is about 18 ft. The gray clay contains numerous mollusk shells and probably represents a tidal flat depositional environment (Peter Weiskel, U.S. Geological Survey, oral commun., 1995). Site BHW499, which is in the present-day upper marsh, lies closer to the original marsh entrance into Cape Cod Bay and is analogous to a "low-marsh" environment (Teal, 1986). Sediments from site BHW496 show medium gray sand to a depth of at least 8 ft. This sand is underlain by fine sand and silt; the two units are part of the unconfined beach aquifer that underlies the barrier beach between the northeast part of the marsh and Cape Cod Bay. This aquifer, which is about 20 ft thick at site BHW491 (fig. 4*B*), is underlain by marine clay and grades laterally into clayey salt-marsh peat toward the marsh. The stratigraphy shown at site BHW496 is inferred from the stratigraphy at site BHW491 (fig. 4*B*) as determined from geophysical and lithologic logs.

The stratigraphic position of the fine-grained marsh deposits in relation to the glacial and marine stratigraphy in the Sagamore Marsh area is shown in figure 4*D*. The marsh deposits, which consist of peat, clayey peat, and clay, are about 20 ft thick beneath the center of the marsh and are underlain by fine to coarse brown glacial sand and fine gray marine sand and silt. The glacial sand is part of the Lake Cape Cod deltaic hydrogeologic unit at site BHW486, and the fine gray sand is part of the fine sand unit at site BHW491. The two sand units probably are separated by an erosional



***D*. Section *D-D'*.**

Figure 4.—*Continued.*

contact that underlies the marsh and form a single, confined aquifer beneath the marsh. The confining conditions are caused by the overlying marsh sediments, which have low vertical and horizontal hydraulic conductivity. The sand aquifer underlying the northern part of the marsh probably is underlain by gray marine clay. The fine-grained marsh deposits grade laterally to medium gray marine sand toward Cape Cod Bay and are laterally bounded by glaciolacustrine silt and sandy clay along the western edge of the marsh.

Water Table and Potentiometric Surface

Sagamore Marsh is at the downgradient edge of the Plymouth-Carver aquifer system—a large, regional ground-water-flow system. Ground water that enters the flow system as areal recharge generally flows toward the southeast in the area of Sagamore Marsh and discharges to the marsh, Cape Cod Bay, and the Cape Cod Canal (fig. 1). Though most of the regional flow system is unconfined, the low hydraulic conductivity of the fine-grained glaciolacustrine sediments along the western edge of the marsh and of the fine-grained marsh sediments causes confining conditions beneath the marsh; for this reason, the flow system is described as semiconfined near the marsh. Unconfined (or water-table) conditions prevail in the marsh sediments beyond the western and northwestern extent of the confining deposits and along the barrier beach on the northeastern side of the marsh (fig. 5); semiconfined or confined conditions prevail beneath the marsh sediments and along at least the western edge of the marsh (fig. 6).

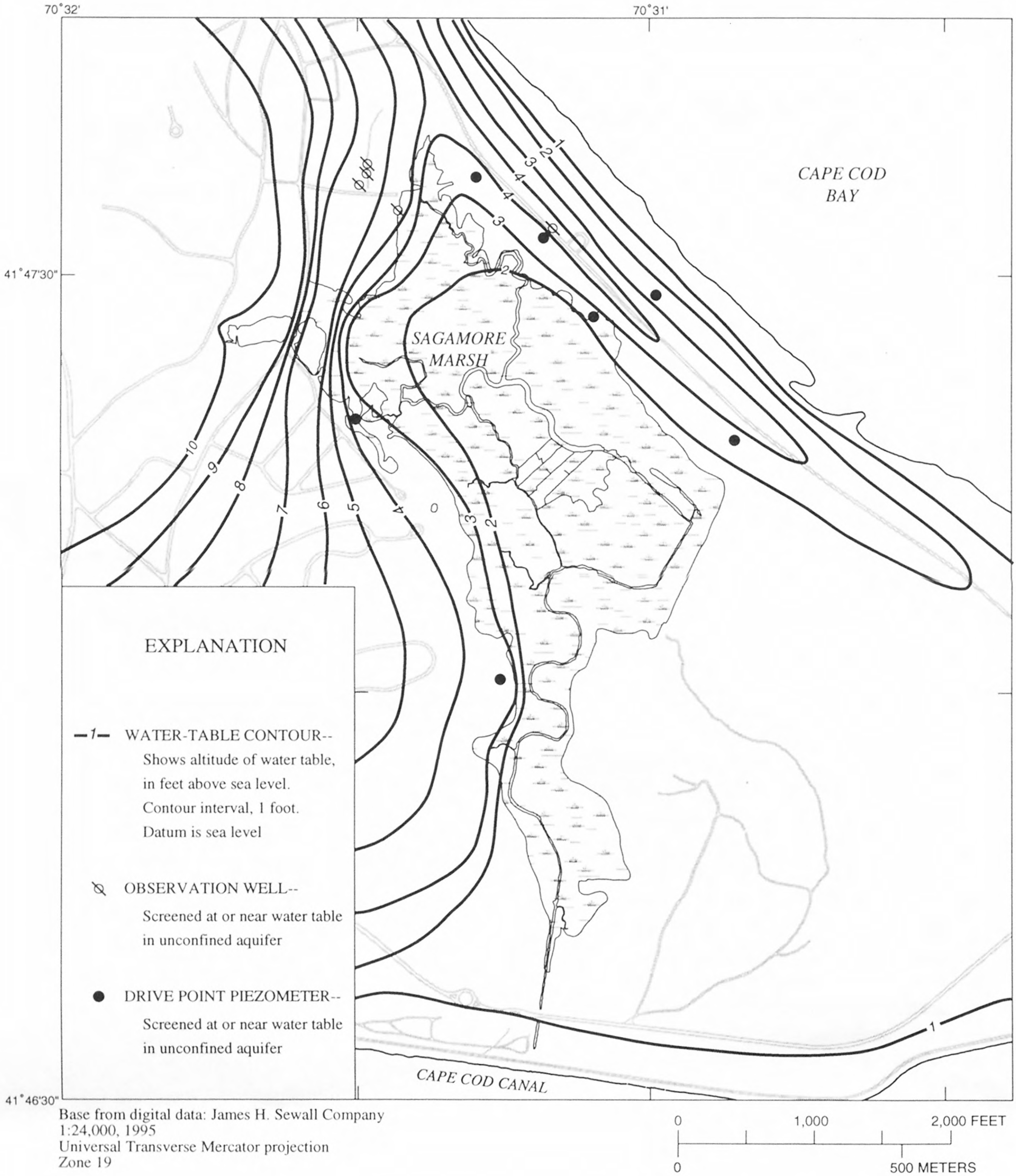


Figure 5. Altitude of water table near Sagamore Marsh, southeastern Massachusetts, June 1995.

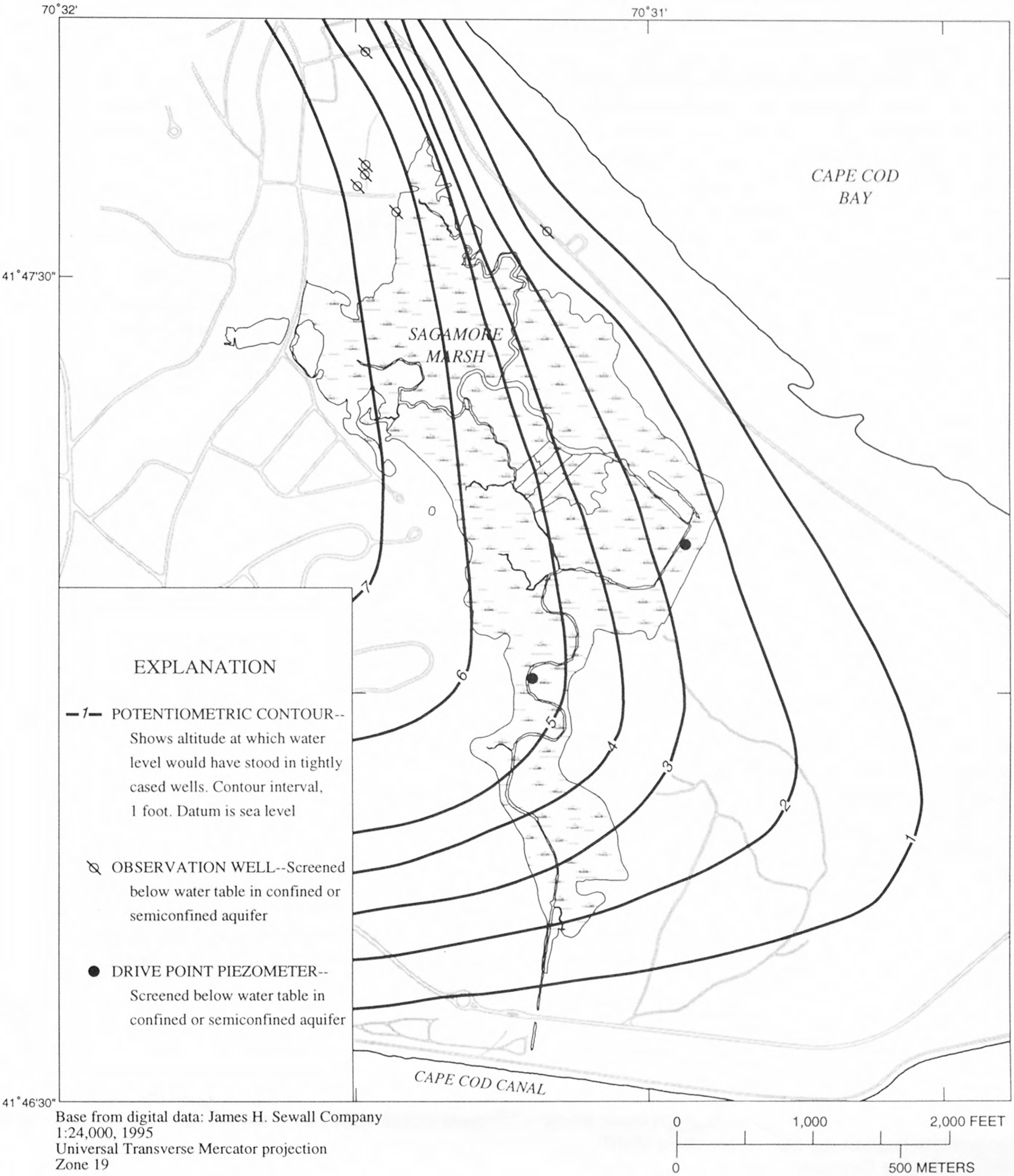


Figure 6. Altitude of potentiometric surface near Sagamore Marsh, southeastern Massachusetts, June 1995.

In June 1995, water-table altitudes near the marsh decreased from about 10 ft along the western edge of the marsh and about 2 ft in the marsh (fig. 5) to sea level near the coast. Maximum water-table altitudes in the fine to medium sand beneath the barrier beach are about 4.5 ft. A water-table mound appears to exist beneath the barrier beach; ground-water flow on the eastern side of the barrier beach is toward Cape Cod Bay and ground-water flow on the western side of the barrier beach is toward the marsh. The water table along the marsh side of the barrier beach was close to land surface during June 1995; depth to the water table from land surface ranged from 1.73 ft (site BHW496) to 0.81 ft (site BHW493). Depth to the water table from land surface along the western boundary of the marsh in June 1995 ranged from 9.98 ft at site BHW486 to 0.82 ft at site BHW498. The direction of water-table gradients along the western, northwestern, and northeastern boundaries of the marsh are toward the marsh, indicating that shallow ground water discharges to the marsh from these directions (fig. 5).

Altitude of the potentiometric surface in the confined or semiconfined aquifer in June 1995 (fig. 6) ranged from about 10 ft above sea level along the western edge of the marsh to near sea level at the coast; hydraulic head may be above sea level and ground water may discharge beneath Cape Cod Bay for some distance out from the shoreline. The sand aquifer that underlies the marsh is confined by the fine-grained marsh sediments and upward head gradients beneath the marsh were large. In June 1995, the confined water level in the lower marsh, site BHW497, was 2.62 ft above land surface and the confined water level in the upper marsh, site BHW499, was about 0.1 ft below land surface. The water table at both of these locations was about 1 ft below land surface at the time of measurement. The upward gradient between the confined aquifer and the marsh surface suggests that there is little if any downward flow of salty water from the marsh to the underlying sand aquifer.

Ground water discharges to springs and seeps near and in the marsh. Springs and seeps occur along the western edge of the marsh (Matthew Walsh, U.S. Army Corps of Engineers, oral commun., 1995); a spring was observed during the investigation near the public-supply well (site BHW013). Ground-water seeps also were observed along the northeastern edge of the marsh. In addition, ground water discharges to a small pond along the western edge of the marsh; freshwater outflow from the pond that was measured at about 0.1 ft^3/s in June 1995 forms a small freshwater tributary to the marsh. Ground water in the confined system beneath the marsh flows to the southeast and likely discharges into Cape Cod Bay and the Cape Cod Canal.

Tidal Effects on Ground-Water Levels

Tidal fluctuations in Cape Cod Bay, Cape Cod Canal, and in tidal channels in Sagamore Marsh affect ground-water levels in the unconfined and confined ground-water systems; tidal ranges in Cape Cod Bay and the Cape Cod Canal were on the order of 9 and 6 ft, respectively, during the study. The tidal cycle in the tidal channels is a function of the tidal cycle in Cape Cod Canal, which had a range of about 6 ft during the study period. The mean range in tidal stage measured in stilling wells in the upper marsh (site BHW499) and lower marsh (site BHW497) were 1.30 and 1.35 ft, respectively, from June 29 to July 18, 1995 (fig. 7). A third stilling well at site BHW501 in a small freshwater tributary at the western edge of the marsh also was tidally influenced and had a range of mean tidal stage of 0.65 ft over the same period. The mean baseflow altitude of the stream at site BHW501 was about 1.9 ft above sea level; during high tide, the stage increased to within 0.1 ft of high-tide stages in the lower marsh. High tide in the upper marsh (site BHW499) and at the marsh edge (site BHW501) occurred after high tide in the lower marsh (site BHW497) and were offset by +22.5 and +47.4 minutes, respectively; a positive offset indicates a later high and low tide than at the lower marsh site. The monthly or lunar tidal cycle at site BHW497 showed a range in mid-tidal stages of 0.45 ft between the neap tide (July 7, 1995) and the spring (full moon) tide (July 15, 1995).

Tidal fluctuations in the water table were measured at two sites along the northeastern edge of the marsh (sites BHW492 and BHW496) and at one site along the beach (site BHW495) (fig. 7*B*). The tidal cycle at site BHW495, which is about 100 ft from Cape Cod Bay, is a function of the tidal cycle in the bay. The range in mid-tidal stages between the neap tide (July 7, 1995) and the spring tide (July 15, 1995) at site BHW495 was 1.05 ft, which was significantly higher than the range in the marsh tidal channels (fig. 7*B*). The range of mean daily tidal stage at site

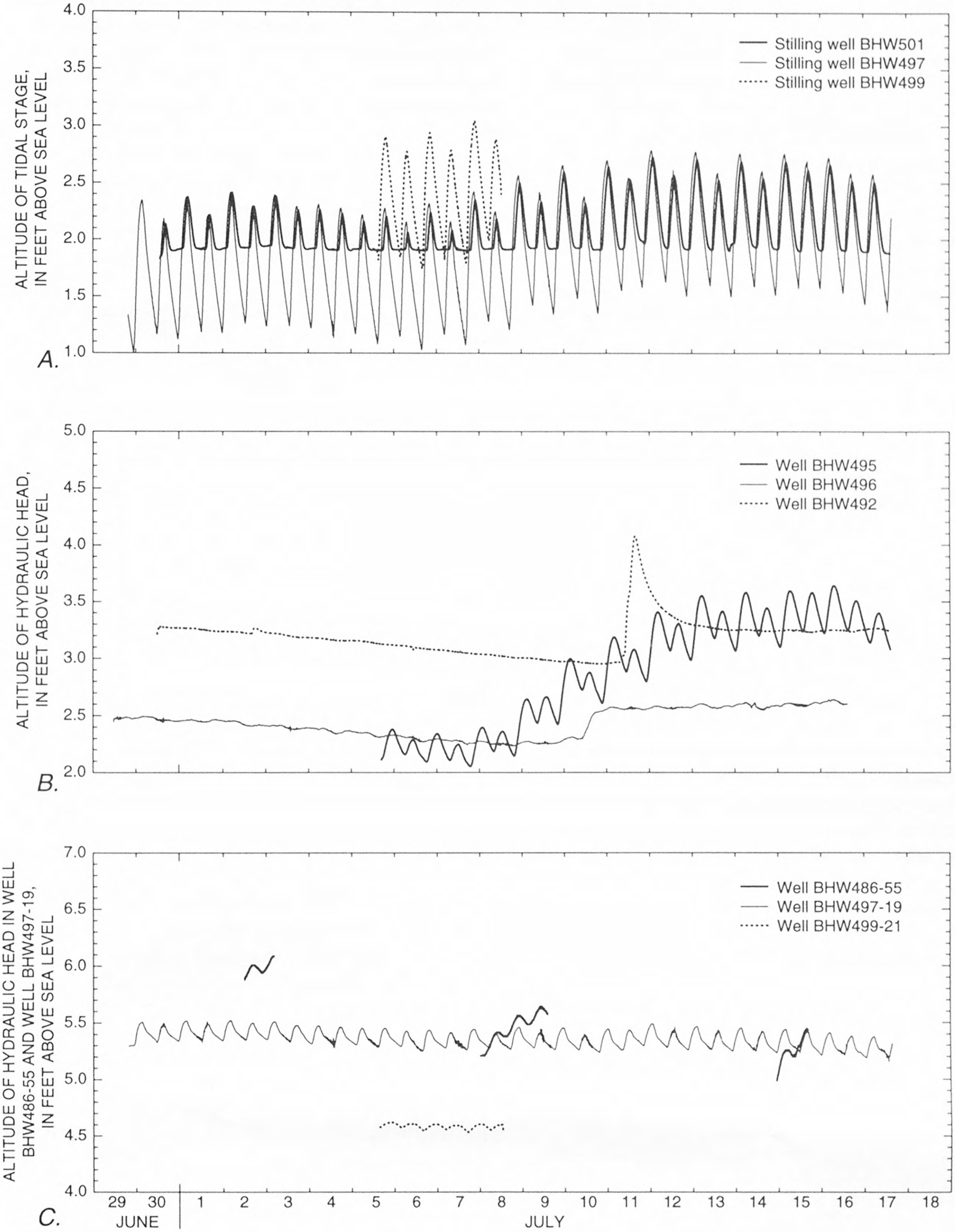


Figure 7. Tidal fluctuations in and around Sagamore Marsh at (*A*) three points in the main tidal channel, (*B*) three points in the unconfined aquifer, and (*C*) three points in the confined aquifer, southeastern Massachusetts, June 29 to July 18, 1995.

BHW495 was 0.5 ft and the range of mean daily tidal stage in Cape Cod Bay was about 9 ft. The large attenuation of the tidal cycle between Cape Cod Bay and site BHW495 probably results from the relatively low diffusivity of the beach deposits—aquifer diffusivity is defined as the ratio of transmissivity (T) to storativity or specific yield (S or Sy). There was no discernible tidal cycle at site BHW492; this site is about 125 ft from the main tidal channel in the marsh and about 300 ft from Cape Cod Bay. The range of mean tidal stage at site BHW496 was about 0.03 ft; the site is about 690 ft from the upper-marsh tidal channel and about 560 ft from Cape Cod Bay. Ground-water level rises at sites BHW492 and BHW496 from July 10-12, 1995, are due to precipitation events (fig. 7*B*).

The tidal cycle at site BHW496 was offset from the tidal cycle in the ground water at site BHW499 by about +317.5 minutes and was offset from the tidal cycle in ground water at the beach (site BHW495) by only +72.5 minutes suggesting that the tidal cycle was more in phase with the tidal pulse originating from Cape Cod Bay than the tidal pulse originating from Sagamore Marsh. Data from the June 1995 aquifer test showed no discernible tidal influence in the water table at site BHW486 near the northwestern boundary of the marsh. The effects of the tidal-channel stage fluctuations on ground-water levels in the unconfined (water-table) aquifer along the barrier beach are negligible suggesting that Cape Cod Bay probably has more affect on ground-water levels in that area than does Sagamore Marsh.

Tidal cycles and fluctuations were measured at two sites in the confined aquifer beneath the marsh (sites BHW497 and BHW499). The mean range of tidal fluctuations in ground water at site BHW497 was 0.15 ft. The range of tidal fluctuations in the tidal channel was about 1.35 ft. The range of tidal fluctuations in ground water was almost one-tenth of those in the tidal channel and probably is due to the low diffusivity of the marsh sediments. The tidal cycle in ground water was offset from the tidal cycle in the channel by about -20.0 minutes—high tide in the ground water occurred about 20 minutes before the high tide in the channel. The presence of a more conductive hydraulic connection between the tidal channel and the aquifer near the present day entrance to the marsh in combination with confined conditions may be the cause of the earlier high tide in ground water. The tidal channel at a downgradient location, which would have an earlier high tide than site BHW497, probably would be underlain by glacial sediments that are coarser grained than the marsh sediments. The earlier high tide in ground water also may indicate that tidal fluctuations in Cape Cod Canal are transmitted more rapidly through the confined glacial aquifer than through the tidal channel and that the ground-water tidal cycle at site BHW497 is more in phase with tidal fluctuations in the canal.

The mean range of tidal fluctuations in ground water at site BHW499 was 0.07 ft, which was about one-twentieth of the range of tidal fluctuations in the tidal channel, indicating that the tidal pulse in ground water at the upper marsh site was more attenuated than the tidal pulse in ground water at the lower marsh site. This may be a function of the different lithologies at the two sites. The marsh sediments at the upper marsh site, which consist of clayey peat and clay, were finer grained than the clayey peat and fibrous peat at the lower marsh site; the lithology indicates that marsh sediments at the upper site probably are less hydraulically connected to the underlying aquifer. The tidal cycle in ground water at site BHW499 was offset from the tidal cycle in the tidal channel by about +123.5 minutes.

The mean range of tidal fluctuations in ground water at site BHW486 at the northwestern edge of the marsh and near the public-supply well was 0.17 ft. The tidal cycle at site BHW486 was offset from the tidal cycle in the channel at site BHW501 by about +69.8 minutes and was offset from the tidal cycle in the channel at site BHW495, adjacent to Cape Cod Bay by about -175.2 minutes. The tidal cycle at site BHW486 is more in phase with tidal cycles in the marsh tidal channels suggesting that the tidal pulse in the confined aquifer originates in the marsh. Data collected during an aquifer test in June 1995 (fig. 9*A*) showed no discernible tidal cycle in the water table at site BHW486-15. Tidal cycles were in phase with the tidal cycle at site BHW486 in the semiconfined wells at sites BHW487, BHW488, and BHW489. The tidal range was largest at site BHW489, which is close to the northeastern edge of the marsh.

Tidal fluctuations at sites BHW487, BHW488, and BHW489 in the confined aquifer are small—less than 0.2 ft. Tidal pulses in the confined aquifer at these sites probably are affected more by tidal pulses originating in the marsh than by tidal pulses originating from Cape Cod Bay.

The use of tidal lags to identify the origin of the tidal fluctuations in ground-water levels can be complicated by extremely long tidal lags, particularly for sites at large distances from the tidal channels and from Cape Cod Bay. In the case of site BHW486, the offset of -175.2 minutes from the drive point near Cape Cod Bay (site BHW495) indicates that the pulse is more in phase with tidal pulses from the tidal channels in the marsh, however, the tidal pulse could originate from Cape Cod Bay and the offset could represent a +579.8 minute lag from the previous tidal pulse in the bay. Large tidal pulse attenuation and large time lags between Cape Cod Bay and site BHW495 may indicate that the beach sediments have a low aquifer diffusivity. The presence of peat interbeds beneath the barrier beach may limit the effective thickness of the aquifer and, therefore, the effective transmissivity of the beach aquifer. The low effective transmissivity and the high specific yield of the beach sediments, which consist of sand and gravel, would cause a low aquifer diffusivity and could result in large time lags in tidal pulses. The presence of a barrier-beach, water-table mound with altitudes of greater than 4 ft above sea level also may indicate that the vertical diffusivity of the beach deposits is low.

Response of Flow System to Pumping

An aquifer test was done near the northwest edge of the marsh at site BHW013 during June 1995. The existing public-supply well at the site was pumped continuously for 5 days at an average rate of 473 gal/min and drawdowns were measured in 10 nearby observation wells. The public-supply well is screened from 32.71 to 40.71 ft below sea level in the fine to coarse brown glacial sand (Lake Cape Cod deltaic sediments) shown in figure 4*A*. After 5 days of pumping, drawdown measured in the pumped well was 17.5 ft. Drawdown was measured in observation wells screened at the same depth as the supply well.

Supply well No.	Distance of observation well from supply well (feet)	Measured drawdown in observation well (feet)
BHW486	100	4.92
BHW487	183	4.05
BHW488	280	3.52
BHW489	491	3.18
BHW490	725	1.67
BHW491	1,450	.37

Drawdown data plotted against time for the four closest observation wells (BHW486-55, BHW487-52, BHW488-53, and BHW489-49) show S-shaped curves that are characteristic of the response of an unconfined aquifer to pumping (fig. 8), as do the drawdown data from observation well BHW486-15 that is screened at the water table (fig. 9). Late-time data also indicate that ground-water levels were affected by tidal cycles in either Cape Cod Bay or the marsh channels, as indicated by the sinusoidal fluctuations of drawdown at times greater than about 400 minutes.

At the start of the aquifer test, the water table near the pumped well was in the glaciolacustrine silt and clay hydrogeologic unit. Shortly after pumping began, the water table measured at the pumped well (BHW013) declined below this silt and clay unit, which is shown in hydrogeologic sections in figures 3, 4*A*, and 9. At well BHW486-15, at a distance of 100 ft from the pumped well, the water table declined below the bottom of the screened interval of the observation well (which is 1.8 ft above the contact of the glaciolacustrine and glacial sand units) after nearly 2 days of pumping; it is uncertain whether or not the water table declined below the bottom of the glaciolacustrine unit at this site during the course of the aquifer test. The glaciolacustrine unit probably contributes to the confining conditions that exist near the marsh for nonpumping conditions. During the course of the aquifer test, the water table was drawn down below the confining unit and the flow system became completely unconfined in the immediate vicinity of the pumped well. The distance from the pumped well at which strictly unconfined flow conditions prevail is unknown. Drawdown was not measurable during the course of the aquifer test in well BHW486-175, which is screened in the bottom glaciolacustrine hydrogeologic unit at a depth of 169 to 171 ft below land surface, indicating that the depth of influence of the pumped well was less than 169 ft.

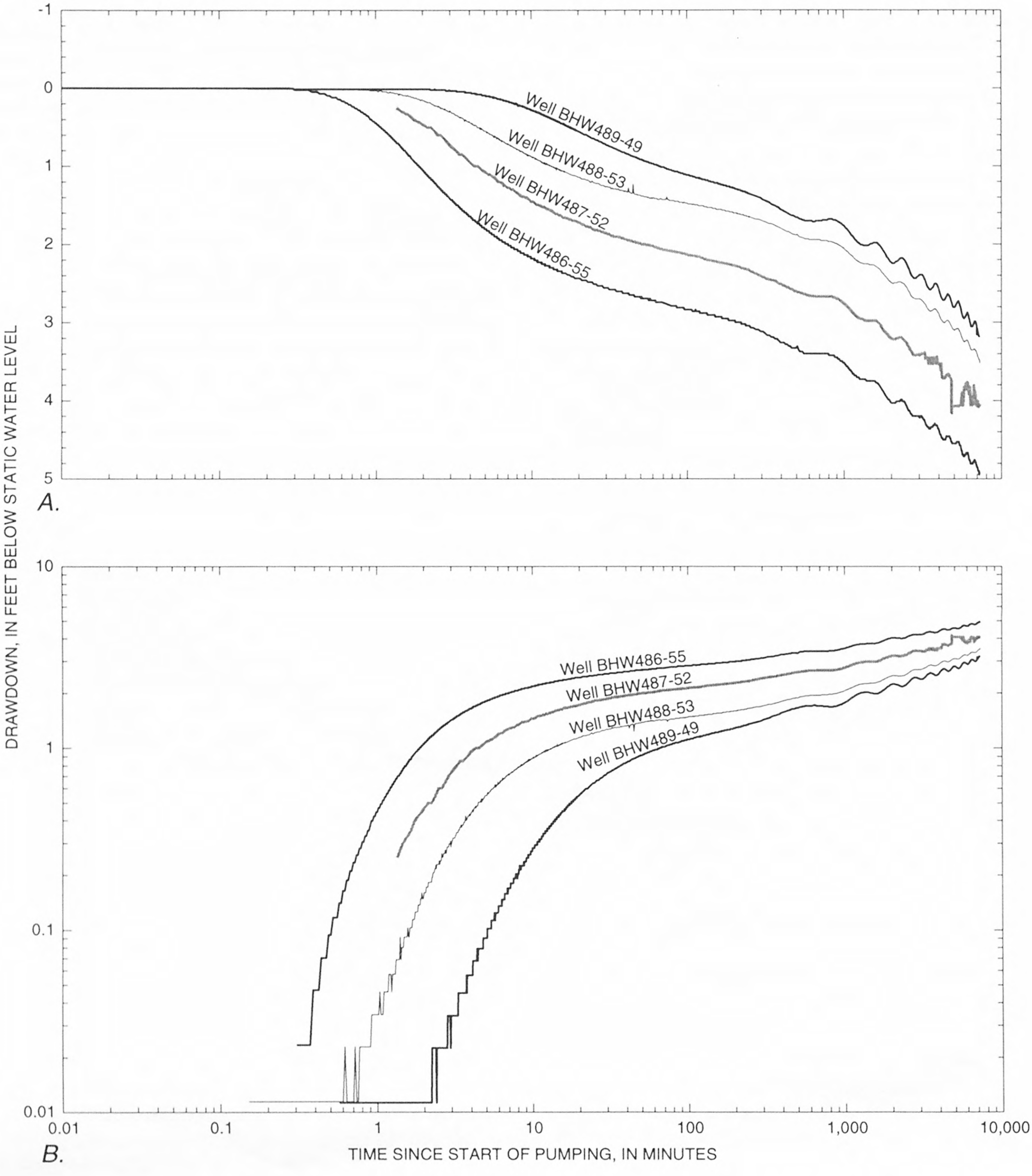


Figure 8. Time-drawdown curves for wells BHW486-55, BHW487-52, BHW488-53, and BHW489-49 near Sagamore Marsh, southeastern Massachusetts, during June 1995 aquifer test using (*A*) a semilog scale and (*B*) a log-log scale.

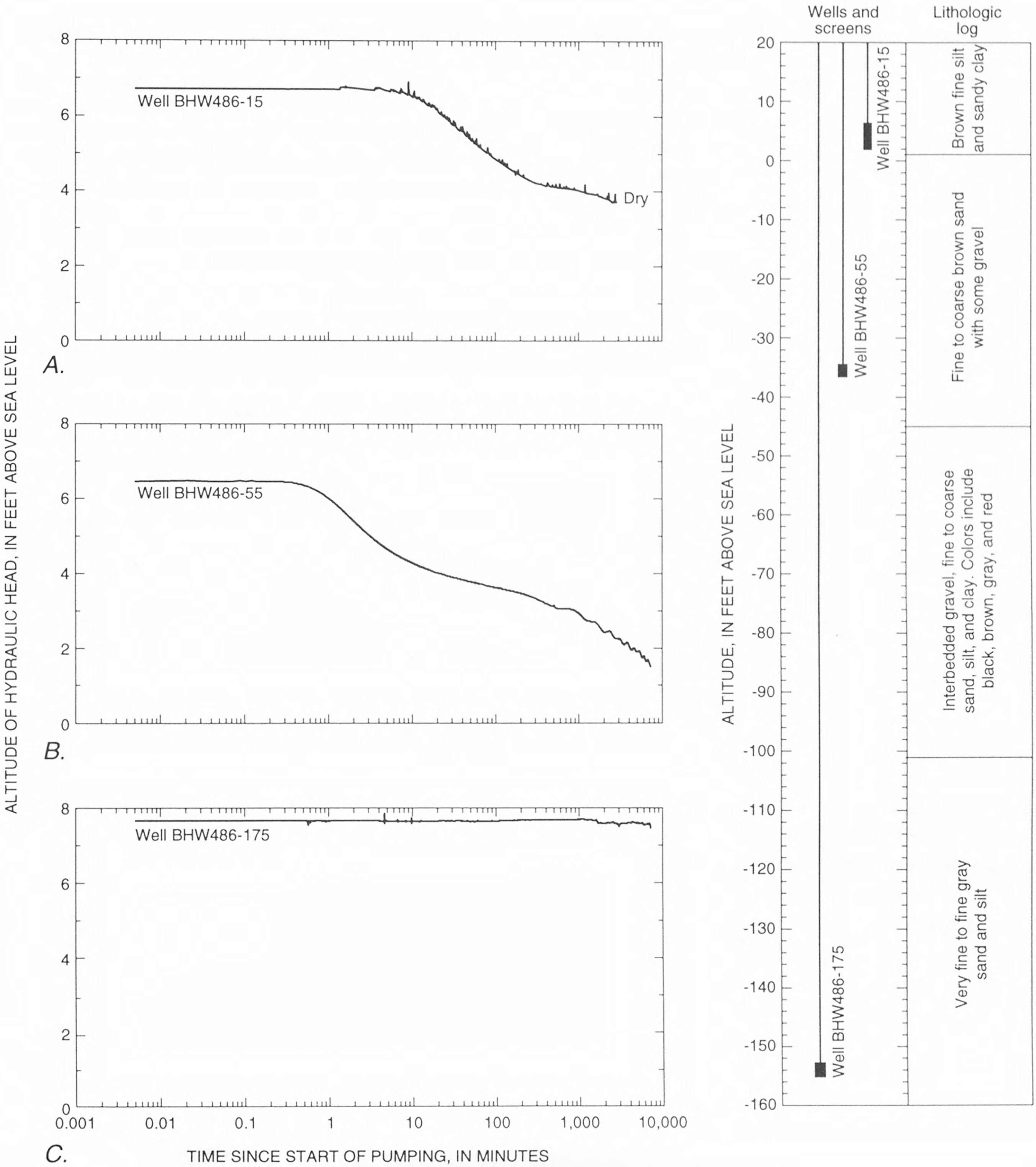


Figure 9. Drawdown in (*A*) water table, (*B*) intermediate confined aquifer, and (*C*) deep aquifer during aquifer test and lithologic log of site BHW486 near Sagamore Marsh, southeastern Massachusetts, June 1995.

Drawdown was 0.37 ft at the water-table well at site BHW491 on the barrier beach at the end of the 5 days of pumping. Site BHW491 is about 1,450 ft from the supply well and is separated from the supply well by the northwestern part of the marsh. These results suggest that the fine to coarse brown sand (Lake Cape Cod deltaic) sediments along the northwestern edge of the marsh are hydraulically connected to the fine gray marine sand underlying the barrier beach; this hydraulic connection, which is illustrated on the hydrogeologic section *D-D'* (fig. 4*D*), is through the sandy sediments underlying the marsh.

Measurements of specific conductance of water discharged at the pumped well and at observation well BHW490, about 725 ft from the pumped well (fig. 2), were made at 4-hour intervals during the aquifer test to monitor the possible saltwater intrusion from Cape Cod Bay into the flow system or the possible upconing of saltwater from deeper intervals of the aquifer into the pumped well. These measurements indicated no significant change and no discernible trend in the specific conductance measurements of ground water at the pumped or observation wells. Specific conductances at the pumped well ranged from 120 to 160 μS/cm and averaged about 150 μS/cm. Electromagnetic induction logs also were collected at sites BHW486 and BHW490 before and after the aquifer test to determine if any change in the conductivity of the aquifer had occurred during the aquifer test; an increase in the conductivity of the aquifer at these deep observation wells would indicate saltwater intrusion or upconing. No changes in the electromagnetic induction logs from pre-pumping conditions were observed. It cannot be concluded from these results, however, that there was no response of the freshwater-saltwater interface to pumping anywhere in the system during the aquifer test; vertical ground-water movement would be slow in fine-grained sediment. The results simply indicate that there were no measurable changes at the observation wells during the aquifer test.

Hydraulic Properties of Hydrogeologic Units

Hydraulic properties of the hydrogeologic units near Sagamore Marsh have been estimated from data collected as part of this investigation. These properties include the aquifer transmissivity, ratio of vertical to horizontal hydraulic conductivity, storage properties of the glacial sand aquifer, and the hydraulic diffusivity of the marsh sediments. Hydraulic properties of the hydrogeologic units are needed for a quantitative analysis of the ground-water-flow system.

Hydraulic Properties of Regionally Significant Hydrogeologic Units Near Sagamore Marsh

Results of the 5-day aquifer test at the public-supply well (BHW013) were analyzed by use of two techniques developed by Neuman (1972, 1974, and 1975). Both methods are based on the assumption that water is withdrawn from an unconfined, homogeneous, and anisotropic aquifer that is of infinite lateral extent and bounded at depth by an impermeable horizontal barrier. Neuman's methods were selected because the results of the 5-day aquifer test indicate unconfined conditions during at least part of the test. The difference between the two methods is that the first assumes that the pumped and observation wells fully penetrate the aquifer, whereas the second assumes that the pumped and observation wells partially penetrate the aquifer. The major limitation of Neuman's methods, as with most analytical methods, is that it is necessary to assume that the aquifer is homogeneous, which is not completely consistent with field data collected at the site. However, the assumption of homogeneity is considered to be sufficient for the intended use of the results, which is a first-level determination of the hydraulic properties of the aquifer near the well.

The first analysis assumes that both the pumped and observation wells fully penetrate the aquifer (Neuman, 1975). This is a common assumption made in aquifer-test analyses and simplifies the analysis by allowing the determination of transmissivity from a semilogarithmic plot of time versus measured drawdown at each of the observation wells (fig. 10).

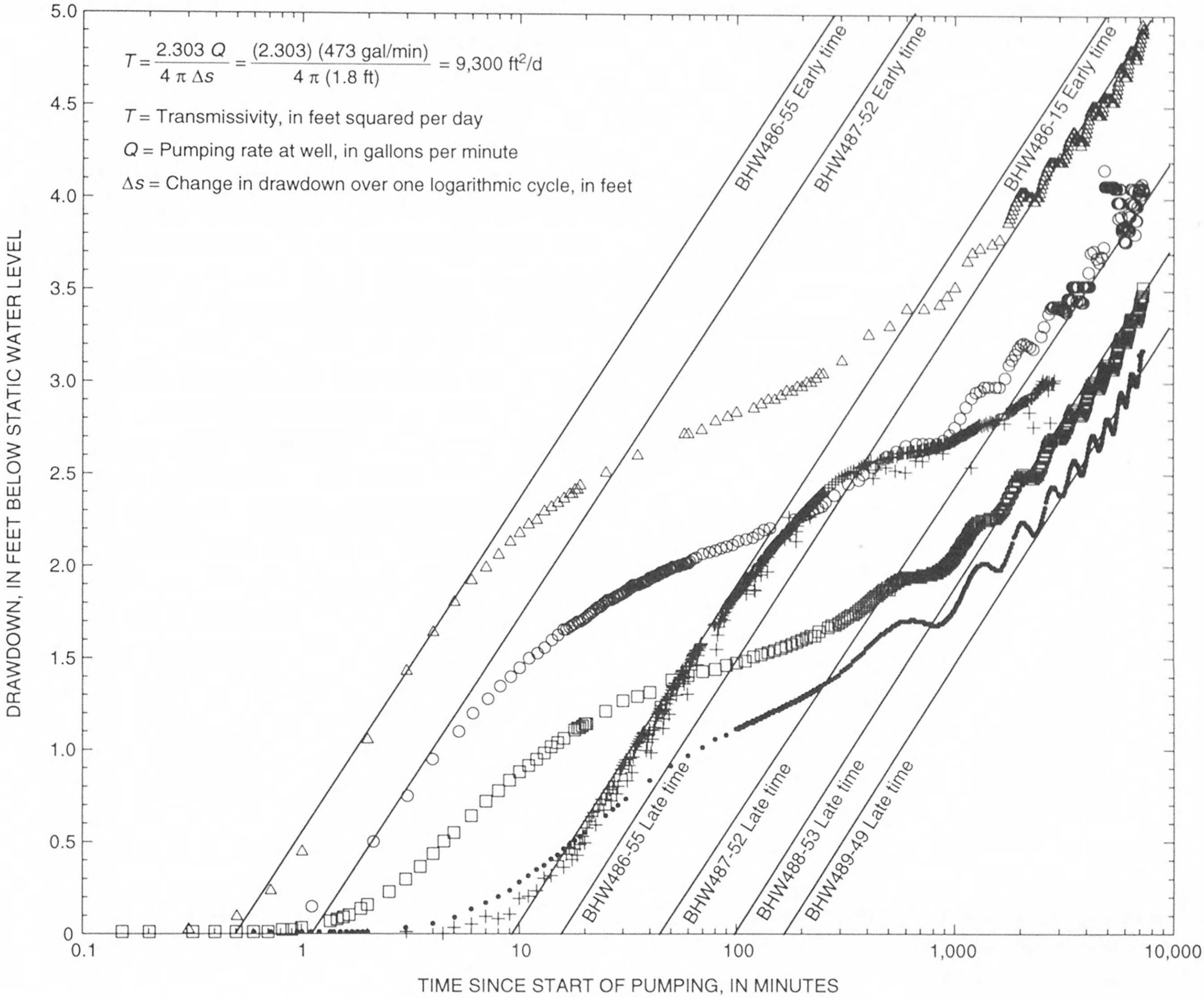


Figure 10. Measured drawdown at selected wells and determination of transmissivity near Sagamore Marsh, southeastern Massachusetts.

As shown in figure 10, parallel lines were drawn through early- and late-time intervals of most of the time-drawdown curves. The slope of each line was the same for early- and late-time intervals, as it should be for conditions of unconfined flow (Neuman, 1975, p. 332) and a homogeneous flow system. A transmissivity of 9,300 ft²/d was determined from the slope of the lines. The saturated thickness of that part of the aquifer that is assumed to supply water to the well is 51.3 ft, which is based on the thickness of the aquifer from the water table to the bottom of the glacial sand hydrogeologic unit (Lake Cape Cod deltaic sediments; fig. 3). Therefore, the calculated transmissivity of 9,300 ft²/d corresponds to a horizontal hydraulic conductivity of 181 ft/d.

Storage coefficient and specific yield also were determined from the semilogarithmic analysis, following Neuman (1975). Early-time data were used to determine storage coefficient and late-time data were used to determine specific yield (Neuman, 1975). Calculated storage coefficient of the aquifer ranged from 4.4×10^{-4} (well BHW486-55) to 5.0×10^{-4} (well BHW487-52); specific yield ranged from 0.01 (well BHW489-49) to 0.02 (wells BHW488-53, BHW487-52, and BHW486-55).

The second method assumes that the pumped and observation wells partially penetrate the aquifer, which is consistent with the conditions at site BHW013. The second method was applied to determine the effect of the assumption of fully penetrating pumped and observation wells on the calculated transmissivity. Theoretical time-drawdown curves (also called type curves) were generated using the computer program WTAQ1 (Moench, 1993), which is based on the theory of Neuman (1972 and 1974). Dimensionless variables for each pumped well-observation well pair are required to generate individual theoretical curves; these data are shown in table 2. An initial, prepumping saturated thickness of the aquifer of 51.3 ft was used in calculating the dimensionless variables shown in table 2. No single (or universal) set of hydraulic property values could be found in which the measured drawdown data and theoretical curves overlapped simultaneously for all five time-drawdown curves, probably because of the heterogeneity of the aquifer deposits near the pumped well. However, good matches could be found when individual drawdown curves were matched to theoretical curves separately. The best match between measured time-drawdown data and the type curves at well BHW486-55 was obtained for a value of transmissivity (T) of 10,900 ft^2/d, a ratio of vertical to horizontal hydraulic conductivity (K_Z/K_H) of 1:44, a ratio of storage coefficient to specific yield (σ) of 0.09, a storage coefficient (S) of 0.0026 and a specific yield (S_y) of 0.03 (fig. 11). Two theoretical curves determined for transmissivity equal to 75 and 125 percent of the value determined from the best match to the measured data also are shown in figure 11. The low values of K_Z/K_H and of S_y determined from the analysis probably result from the presence of the glaciolacustrine silt and clay hydrogeologic unit at the top of the stratigraphic column. The water table at the test site was in this unit at the start of the aquifer test and moved downward through the unit during the test. A horizontal hydraulic conductivity of the glacial sand hydrogeologic unit of 213 ft/d is calculated from the estimated transmissivity on the basis of an assumed saturated thickness of the aquifer of 51.3 ft. The transmissivity and hydraulic conductivity calculated for this observation well are higher than, but very similar to those calculated in the previous analysis, which suggests that partial penetration of the pumped and observation wells is not an important variable in the analysis of hydraulic properties at the site. Transmissivities calculated using the two methods are consistent with those determined for similar deposits of Cape Cod by Barlow and Hess (1993), Barlow (1994), and Masterson and Barlow (1994).

Table 2. Data required for each pumped well and observation well pair for the analysis of hydraulic properties near well BHW013, near Sagamore Marsh, southeastern Massachusetts

[**Site identifier:** Location of sites shown in figure 2. **Dimensionless variables:** l_D is from the initial water table to the bottom of the screened interval of the pumped well to the initial saturated thickness. d_D is from the initial water table to the top of the screened interval of the pumped well to the initial saturated thickness. z_{D1} is from the bottom of the aquifer to the bottom of the screened interval of the observation well to the initial saturated thickness. z_{D2} is from bottom of the aquifer to the top of the screened interval of the observation well to the initial saturated thickness. Dimensionless = foot per foot. ft, foot]

Well identifier	Radial distance from pumped well (ft)	Dimensionless variable			
		l_D	d_D	z_{D1}	z_{D2}
BHW486-15	100	0.92	0.76	0.95	1.00
BHW486-55	100	.92	.76	.19	.23
BHW487-52	183	.92	.76	.21	.25
BHW488-53	280	.92	.76	.24	.27
BHW489-49	491	.92	.76	.12	.15

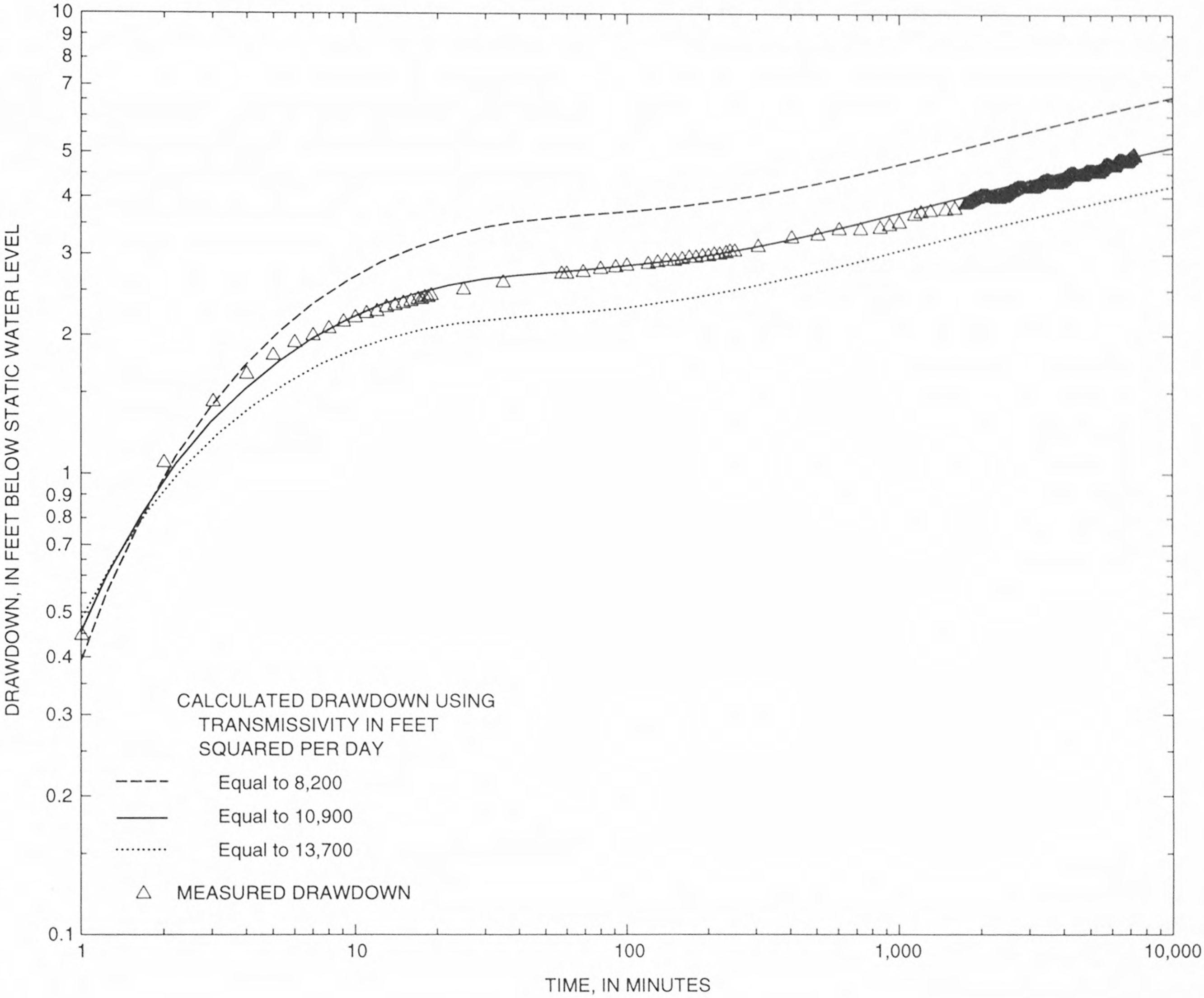


Figure 11. Measured and calculated drawdown at well BHW486-55 near Sagamore Marsh, southeastern Massachusetts.

Hydraulic Properties of Marsh Sediments

The hydraulic conductivity of the marsh sediments varies with lithology. Knott and others (1987) reported that the hydraulic conductivity of most fibrous salt-marsh peat ranges from 0.03 to 280 ft/d, with a median of about 3 ft/d; the upper part of this range is similar to values reported for medium sand (Freeze and Cherry, p. 29, 1979). Hydraulic conductivities reported for marine clay are much lower and range from 10^{-7} to 10^{-4} ft/d (Freeze and Cherry, 1979). Clayey peat, which is observed at both marsh coring sites, probably has an intermediate hydraulic conductivity that is closer to the hydraulic conductivity of marine clay. Marsh sediments from site BHW497 consisted primarily of gray clayey peat underlain by brown fibrous peat. These sediments probably have a higher hydraulic conductivity than marsh sediments at site BHW499, which consisted of clayey peat underlain by gray clay. Although ranges in tidal stage in the tidal channel at sites BHW497 (lower marsh) and BHW499 (upper marsh) were similar, ranges of tidally influenced ground-water-level fluctuations at site BHW497 (0.15 ft) were significantly greater than at site BHW499 (0.07 ft) suggesting that sediments at the lower marsh site are more conductive than those at the upper marsh site.

Ferris (1963) developed an analytical method to determine the diffusivity (α) of an aquifer that is based on the ratio of cyclic water-level fluctuations in tidal water bodies to the corresponding cyclic water levels in an adjacent aquifer. This ratio is called the tidal-range ratio. As stated previously, the diffusivity of an aquifer is defined as the transmissivity (T) divided by storage coefficient (S). The method developed by Ferris assumes that ground-water flow near the tidal channel is one dimensional; that is, that the hydraulic pulse propagated at the channel moves in one direction outward from the channel. The equation relating the tidal-range ratio to diffusivity is:

$$\log\frac{Rgw}{Rsw} = -0.77x\sqrt{\frac{1}{\alpha t}}\ , \qquad (1)$$

where Rgw and Rsw are the tidally influenced ranges in ground-water and surface-water levels, respectively, in feet; x is the distance of the observation well from the tidal channel, in feet; α is aquifer diffusivity, in square feet per day; and t is the period of the tidal cycle, in days.

Equation 1 was used to estimate the diffusivity of the marsh sediments at sites BHW497 and BHW499. In using equation 1, which is a one-dimensional analysis, the assumption was made that the tidal pulse propagated vertically from the tidal channel through the marsh sediments and that the attenuation observed at the contact between the fine sand aquifer and the marsh sediments was a function of the diffusivity of the marsh sediments. Tidal cycles at both sites for July 5-8, 1995 are shown in figure 12. In the lower marsh (site BHW497), ranges of tidally influenced ground-water and surface-water levels were 0.17 and 1.16 ft, respectively. The vertical distance from the bottom of the tidal channel to the observation well was 15 ft and the period of the tidal cycle was 0.51 days. Using equation 1, a diffusivity of 380 ft^2/d was calculated for sediments at the site. The hydraulic conductivity of the marsh sediments at site BHW497 is 2.5×10^{-3} ft/d assuming a storage coefficient of 0.0001 for the marsh sediments—determined from published values of aquifer compressibility (Freeze and Cherry, p. 5, 1979)—and an effective thickness of 15 ft, which is the assumed width of the tidal channel. In the upper marsh (site BHW499), ranges of tidally influenced ground-water and surface-water levels were 0.06 and 1.06 ft, respectively. The vertical distance from the bottom of the tidal channel to the observation well and the period of the tidal cycle were the same as for the lower marsh site—15 ft and 0.51 days, respectively. The value of diffusivity at the upper marsh site, as determined from equation 1, was 170 ft^2/d. Assuming a storage coefficient of 0.0001 and an effective thickness of 10 ft, which is the assumed width of the tidal channel, the hydraulic conductivity of the marsh sediments at site BHW499 (upper marsh) is 1.7×10^{-3} ft/d. The lower conductivity seen at the upper marsh site (site BHW499) is due to the presence of finer grained sediments than at site BHW497. Estimated hydraulic conductivities at both sites were intermediate between the reported values for fibrous peat and marine clay, but are closer to values reported for marine clay.

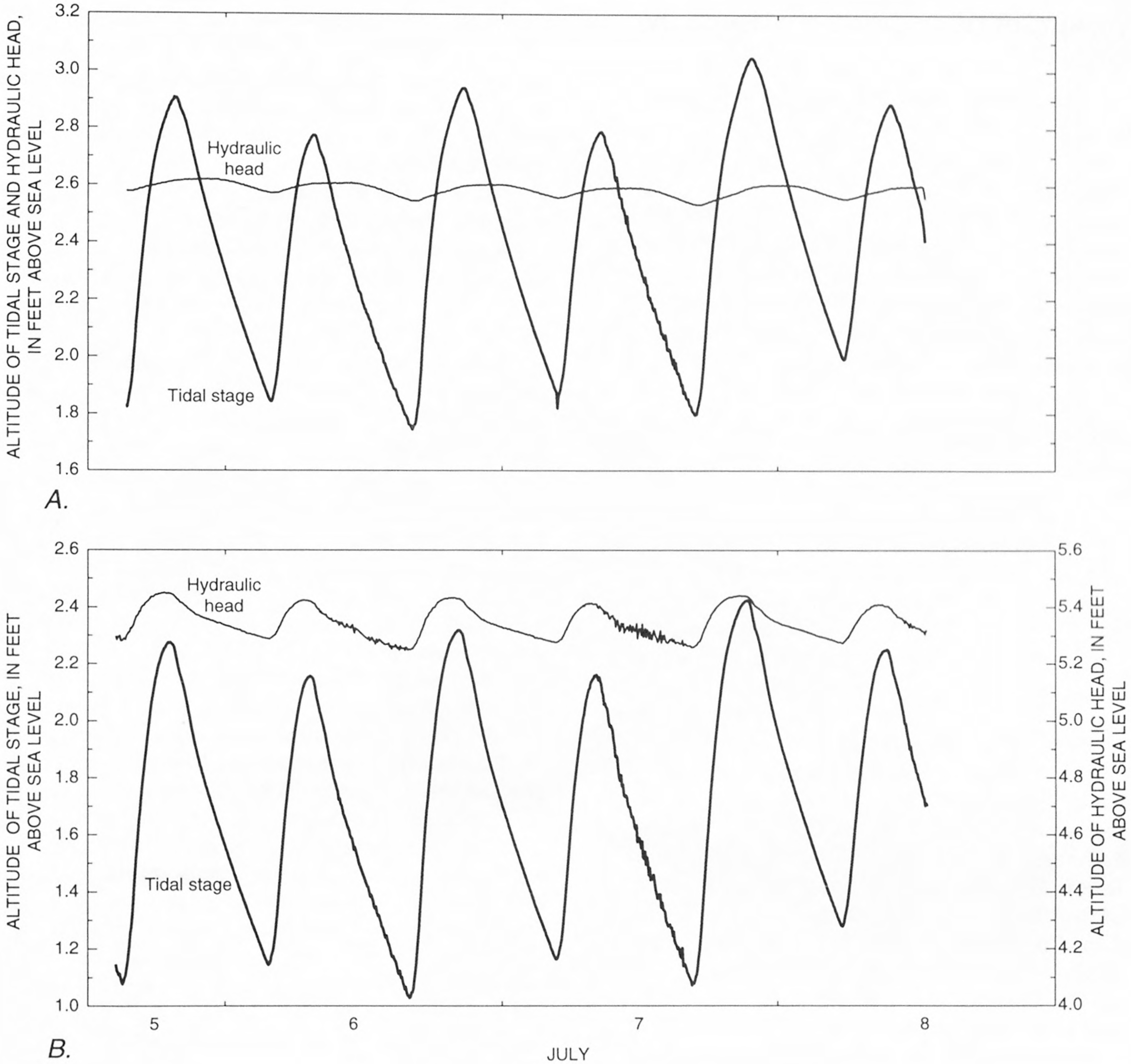


Figure 12. Tidal-stage fluctuations in the main tidal channel and hydraulic head fluctuations in adjacent wells at two locations in Sagamore Marsh, southeastern Massachusetts, July 5-8, 1995.

ANALYSIS OF GROUND-WATER-FLOW SYSTEM

Analytical- and numerical-modeling methods were used in the analysis of the ground-water-flow system near Sagamore Marsh. Specifically, the analysis addresses (1) how ground-water levels in and near the marsh might be affected by increases in tidal-channel stages following increases in the amount of salt water allowed to flow into the marsh, and (2) how the contributing area and source of water to the public-supply well (BHW013) might be affected by the higher tidal stages.

Analytical-Modeling Analysis of the Response of the Ground-Water-Flow System to Increased Tidal Stage

The relation between tidal-pulse attenuation and the diffusivity of marsh and aquifer sediments, discussed previously, was used to predict the effect of higher tidal stages on ground-water levels near the marsh. Projected high tidal stages in the lower and upper parts of the marsh were determined by the U.S. Army Corps of Engineers using a surface-water-flow model; rises in high tidal stages at sites BHW497 (lower marsh) and BHW499 (upper marsh) are anticipated to be less than +1.0 and +0.6 ft, respectively (Matthew Walsh, U.S. Army Corps of Engineers, oral commun., 1995). These anticipated rises in tidal stage are based on an assumed installation of two 10×20 foot culverts at the marsh entrance to the Cape Cod Canal; the actual size of the culvert that is being proposed is only 6×12 ft, which would result in a smaller amount of saltwater flow into the marsh and lower tidal-stage rises than those evaluated here. Tidal pulses originating from the tidal channels could be propagated through the marsh sediments—the water table occurs within the marsh sediments—or through the underlying fine sand aquifer.

Equation 1 was used to calculate the theoretical maximum distance from the tidal channel at which the tidally influenced ranges in ground-water and surface-water levels in the marsh sediments (Rgw) would be less than 0.01 ft. The theoretical distance for current and predicted tidal stages was determined using the calculated value of marsh-sediment diffusivity at the two marsh sites, and substituting present and predicted values for the tidally influenced ranges in ground-water and surface-water levels in the channel (Rsw). The analysis assumes that marsh sediments are nearly isotropic (Knott and others, 1987) and that the calculated diffusivities, which assumed vertical propagation of the tidal pulse, are the same in the horizontal direction. Based on current (1995) conditions, the estimated maximum distance at which the range of tidal fluctuations in the aquifer exceed 0.01 ft is 37.1 ft at the lower marsh site (BHW947) and 24.3 ft at the upper marsh site (BHW499). The maximum distances of tidal effects exceeding 0.01 ft after tidal stages (Rsw) increased to the levels predicted by the U.S. Army Corps of Engineers—2.16 at site BHW497 (lower marsh) and 1.66 at site BHW499 (upper marsh)—were 42.0 and 26.7 ft, respectively.

Equation 1 also was used to determine theoretical distances of tidal pulse propagation in the underlying fine sand aquifer. The hydraulic conductivity of the fine sand and silt underlying the marsh was estimated to be about 30 ft/d; this estimate was based on cores collected in the marsh and published values of hydraulic conductivity from a similar hydrogeologic environment (Masterson and Barlow, 1994). An assumed storativity of 0.0026 from the 5-day aquifer test was used to determine aquifer diffusivities. An aquifer thickness of 15 ft was determined from lithologic data; owing to the probable high anisotropy of the fine sand and silt, the effective aquifer thickness may be significantly smaller and the use of 20 ft as an aquifer thickness probably would overestimate tidal-pulse propagation distances. The tidal pulses measured in the aquifer beneath and adjacent to the tidal channels, which was 0.17 at site BHW497 (lower marsh) and 0.06 at site BHW499 (upper marsh), were used as the initial tidal pulses (Rgw) in equation 1. The initial tidal pulses used in estimating propagation distances after tidal stages increase was determined by multiplying the tidally influenced ranges in ground-water and surface-water levels in the aquifer by the ratio of proposed to current tidally influenced ranges in ground-water and surface-water levels in the tidal channel—this assumes a linear propagation of tidal pulses from the tidal channel to the aquifer. Under current tidal stage conditions, the estimated maximum distances where the tidal pulse in the fine sand aquifer exceeds 0.01 ft was 541.3 ft at site BHW497 (lower marsh) and 342.5 ft at site BHW499 (upper marsh). At site BHW497, the estimated maximum distances where the tidal pulse exceeds 0.05 and 0.1 ft were 233.8 and 101.4 ft, respectively. The tidal pulse at site BHW499 would exceed 0.05 ft for a distance of 34.8 ft. Following an increase in tidal stage to the values projected by the U.S. Army Corps of Engineers, the tidal pulse at site BHW497 would exceed 0.01, 0.05, and 0.1 ft at maximum distances of 662.2, 354.7, and 222.3 ft, respectively. The tidal pulse at site BHW499 would exceed 0.01 ft for a maximum distance of 419.8 ft and exceed 0.05 ft for a distance of 112.3 ft following the projected increases in marsh-channel tidal stages.

Propagation distances projected for the underlying sand aquifer were much higher than those projected for the marsh sediments. The larger propagation distance is a function of the higher aquifer diffusivity of the sand aquifer as compared to the marsh sediments. Aquifer diffusivities of the marsh sediments at the two marsh sites averaged 275 ft^2/d—the estimated aquifer diffusivity in the underlying sand and silt aquifer was about 225,000 ft^2/d.

Whether a marsh-generated tidal pulse is transmitted through the fine-grained marsh sediments or through the underlying aquifer is difficult to determine. The tidal pulse originating at the tidal channel for present and anticipated tidal-stage fluctuations does not propagate very far in the marsh sediments. Propagation distances are significantly higher in the underlying sand aquifer, but are still small compared to the size of the marsh; the distance between the tidal channel and the marsh edge is more than 600 ft along most of the northeastern edge of the marsh. Magnitudes of the tidal pulses evaluated in the calculations—0.01, 0.05, and 0.1 ft—are small as compared to natural fluctuations in the water table beneath the barrier island that arise from tidal fluctuations in Cape Cod Bay and from precipitation events (fig. 7).

The results of the analytical solutions are consistent with observations of ground-water level fluctuations along the northeastern edge of the marsh. Because tidal fluctuation in the water table was not discernible at site BHW492 (fig. 6), any marsh-generated fluctuations in the water table may be transmitted primarily through the marsh sediments; the distance of 250 ft between site BHW492 and the tidal channel is less than the projected propagation distance of a 0.01 ft tidal pulse in the fine sand aquifer of 342.5 ft. The tidal pulse originating in the upper marsh tidal channel has no discernible effect on water levels at sites BHW492 and BHW496 about 250 and 660 ft from the nearest tidal channel, respectively. The projected maximum rise in high tidal stages would not substantially increase ground-water-level fluctuations near the edge of the marsh. Given the small tidal fluctuations in the aquifer and the rapid attenuation of tidal pulses in the marsh sediments, the magnitude of any marsh-generated tidal pulses beneath the barrier island would be significantly smaller than tidal pulses originating from Cape Cod Bay and fluctuations arising from precipitation events. Overbank flooding would not be expected to result in a rise in ground-water levels. The low permeability of the marsh sediments and the fact that overbank water probably would not be in direct hydraulic connection to the underlying aquifer would attenuate the effects on water levels in the aquifer.

Numerical Analysis of Ground-Water-Flow System

A numerical analysis of the ground-water-flow system was done to evaluate the zone of contribution to the public-supply well at site BHW013 (the North Sagamore Water District's Beach Well) for current (1994) pumping conditions and to evaluate the response of the ground-water-flow system to proposed changes in tidal stage in Sagamore Marsh. A steady-state, three-dimensional model of ground-water flow was developed using the finite-difference computer code MODFLOW developed by McDonald and Harbaugh (1988). The finite-difference grid developed for the modeled area (fig. 13) is aligned with the finite-difference grid of the regional flow model of the Plymouth-Carver aquifer developed by Hansen and Lapham (1992). The two grids were aligned so that boundary conditions for the model of the Sagamore Marsh area (a subregional model) could be obtained from the results of the regional-scale model of the Plymouth-Carver aquifer.

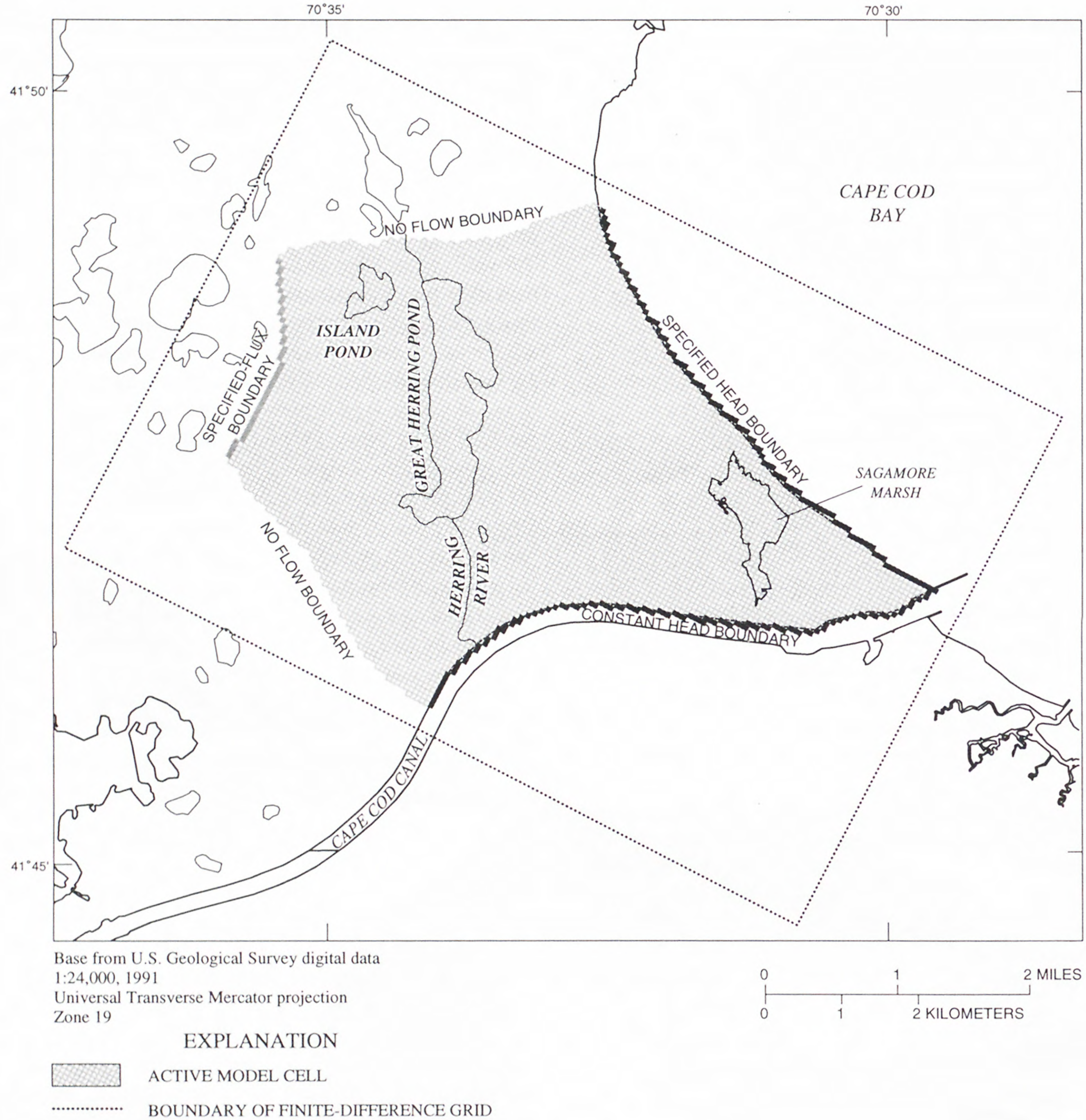


Figure 13. Extent of finite-difference model grid and lateral boundary conditions for the Sagamore Marsh area, southeastern Massachusetts.

The model grid covers an area of about 27.2 mi^2, but the area of the ground-water-flow system that is actually simulated (the active modeled area) is only about 11 mi^2 (fig. 13). The lateral extent of the model was selected to include an area large enough to minimize the effect of model boundary conditions on ground-water heads calculated near Sagamore Marsh and on the zone of contribution delineated for supply well BHW013.

Grid

The finite-difference grid consists of 115 rows and 165 columns of uniformly-spaced cells each 200×200 ft on a side (fig. 13). The grid cells are one-twenty-fifth the size of those of the regional model of the Plymouth-Carver aquifer (Hansen and Lapham, 1992), which had a discretization of 1,000×1,000 ft. The model consists of four layers that extend from the water table to the bedrock surface. The vertical spacing was chosen to coincide with the hydrogeologic units in the study area (fig. 4).

The bottom altitude of cells in layer 1 (top layer) that are coincident with Island and Great Herring Ponds was set equal to the altitude of the bottom of each pond; pond bathymetry was obtained from Massachusetts Division of Fisheries and Wildlife (1993). The bottom altitude of cells in layer 1 of the model that contain marsh sediments coincides with the altitude of the bottom of these sediments, as determined from lithology data collected in the marsh (fig. 4). The bottom altitude of the remaining model cells in layer 1 were set equal to 0 ft above sea level. The bottom altitudes of layers 2 and 3 were set at 50 and 110 ft below sea level, respectively. This vertical spacing was based on the changes in lithology as described in the previous section "Hydrogeologic Units."

In the northwest part of the modeled area, the bottom altitude of layer 3 was truncated at 50 ft below sea level to coincide with the underlying bedrock surface, as reported by Hansen and Lapham (1992). The bottom altitude of layer 4 also coincides with the contact of the bedrock surface and glacial deposits, and extends from 110 to 185 ft below sea level.

Boundary Conditions and Stresses

Hydraulic boundaries of the simulated area include ground-water inflow from streams, ponds, wells, coastal saltwater bodies, the water table, and adjacent areas of the aquifer. The contact between unconsolidated, glacial sediments, and the underlying crystalline bedrock is a hydraulic boundary across which there is no ground-water flow. The northwestern part of the modeled area is an area of ground-water inflow from adjacent, upgradient parts of the aquifer (fig. 13). A specified-flux boundary condition was used to simulate the inflow of ground water across this boundary. The rate of inflow across the boundary was calculated by three methods. First, an inflow rate of 2.9 ft^3/s was obtained by multiplying the area between the northwestern boundary of the model and the water-table divide northwest of the model boundary by the areal recharge rate for the study area (about 27 in/yr) used by Hansen and Lapham (1992). The location of the water-table divide was obtained from the water-table map of the Plymouth-Carver aquifer published by Hansen and Lapham (1992). Second, an inflow rate of 2.5 ft^3/s was obtained from Darcy's law by use of measured hydraulic gradients along the northwestern boundary, estimates of transmissivity, and the cross-sectional area of the northwestern boundary of the model. Finally, an inflow rate of 3.2 ft^3/s was obtained from the results of the regional-scale flow model (Hansen and Lapham, 1992) across the cross-sectional area coincident with the northwestern boundary of the subregional model. The average of these three estimated inflow rates (2.9 ft^3/s) was distributed to the 118 active model cells along the northwestern boundary in proportion to the cross-sectional area and horizontal hydraulic conductivity of each cell.

When short-term (for example, tidal) fluctuations are averaged, coastal saltwater bodies along the southern and eastern boundaries of the model, such as Cape Cod Bay and the Cape Cod Canal (fig. 13) can be considered areas of constant water levels. These boundaries were simulated by means of a specified-head boundary condition in layer 1 of the model similar to previous studies on Cape Cod (Masterson and Barlow, 1994). Because these boundaries consist of saltwater, the saltwater heads were converted to equivalent freshwater heads by dividing the thickness of the saltwater body in each specified-head cell by 40.0, the ratio of the specific weight of freshwater (1.000 g/cm^3) to the difference between the specific weights of saltwater (1.025 g/cm^3) and freshwater. The thickness of the saltwater body for each cell was obtained from the bathymetric contours reported on the U.S. Geological Survey 1: 24,000 Sagamore, Massachusetts quadrangle map.

The northern and western hydraulic boundaries of the active modeled area are ground-water drainage divides. The location of the drainage divides were based on ground-water-flow lines drawn perpendicular to water-table contours in the modeled area for ground-water conditions of November 30 through December 2, 1984 (Hansen and Lapham, 1992) (fig. 1). The ground-water drainage divides were simulated as no-flow boundaries, across which ground water cannot flow. In the natural system, however, ground-water drainage divides may shift in response to seasonal, tidal, and other induced changes in ground-water levels.

The lower boundary of the model was identified as the contact between the unconsolidated glacial deposits and the underlying crystalline bedrock. Flow across this contact was assumed to be insignificant because of the low permeability of the crystalline bedrock relative to overlying glacial sediments. The lower boundary was therefore simulated as a no-flow boundary condition. The location of this no-flow boundary was obtained from a map showing the altitude of the bedrock surface for the study area (Hansen and Lapham, 1992).

The upper boundary of the model is defined by three boundary conditions that simulate the water table, streamflow, and ground-water seepage at Sagamore Marsh. The water table was simulated as a free-surface boundary condition that receives spatially variable rates of recharge. The altitude of the water table is calculated by the model for layer 1. The rate of recharge from precipitation specified for the steady-state model simulations, 27 in/yr, was obtained from the regional investigation of Hansen and Lapham (1992). At Island and Great Herring Ponds, the steady-state rate of recharge was reduced to account for evaporation from the pond surfaces; a specified value of 20 in/yr was obtained by subtracting the estimated rate of free-water-surface potential evaporation from the ponds, 28 in/yr (Farnsworth and others, 1982), from the average rate of precipitation in the study area (48 in/yr).

Recharge rates were decreased in some areas to account for changes in geology. In areas underlain by glaciolacustrine sediments, which have a lower hydraulic conductivity than glacial sands, the recharge rate was decreased to 6.8 in/yr, as was simulated in the regional-scale model (Hansen and Lapham, 1992). This lower recharge rate (6.8 in/yr) is based on the mean-annual ground-water runoff calculations of Morrissey (1983) for fine-grained sediments. This value may be a slight underestimate of the actual recharge due to the absence of surface runoff; however, the limited areal extent of the reduced recharge as compared to the overall modeled area would make these slight differences in recharge negligible in terms of the overall hydrologic budget and model calibration. In areas underlain by Sagamore Marsh, recharge was set at zero to account for the high rates of evapotranspiration and low rates of infiltration typical of salt marshes (Peter Weiskel, oral commun., 1995). Also, recharge was not specified for the marsh because it is assumed to be an area of ground-water discharge. The average recharge rate over the entire modeled area was about 25 in/yr.

Recharge was increased at the surface of Great Herring Pond to balance the surface-water inflows and outflows at the upgradient and downgradient ends of the pond from the Herring River. Measurements of streamflow entering and leaving Great Herring Pond indicated a net inflow to the aquifer from the pond of 2.72 ft^3/s in July 1986 (Hansen and Lapham, 1992). This increase in inflow to the aquifer was distributed uniformly as spatial recharge to the model cells that are coincident with Great Herring Pond and was combined with the pond recharge from precipitation previously corrected for evaporation.

The Herring River, which flows through Great Herring Pond, constitutes the only streamflow in the study area. Measurements of the Herring River indicate that the river is a losing stream along the reach from the outlet of Great Herring Pond to the gage near the Cape Cod Canal. Along this reach, the streamflow decreased from 7.28 to 6.35 ft^3/s in July 1986 (Hansen and Lapham, 1992). This losing reach of stream was simulated as a specified-flux boundary condition in the model. For this investigation, all streamflow loss was assumed to occur in the upper 1,000 ft of the stream, and, therefore, 0.93 ft^3/s of inflow was distributed uniformly to the first five model cells coincident with the Herring River, downgradient from Great Herring Pond.

The final boundary condition in layer 1 of the model is the ground-water seepage occurring at the Sagamore Marsh. Visual inspections in the marsh indicate ground-water seepage along the boundary between the low-permeability marsh sediments and the adjacent glacial sands; however, the actual rate of this seepage was not determined during this investigation. Ground-water seepage at the aquifer-marsh boundary was simulated in the flow model as a head-dependent-flux boundary that could only receive ground-water discharge. This boundary condition does not allow for flow from the marsh sediments to the underlying aquifer. Heads specified at the ground-water seeps were estimated from topographic contours shown on the U.S. Geological Survey 1: 24,000 Sagamore, Massachusetts quadrangle map. The tidally affected marsh channel also was simulated as a head-dependent flux boundary. This allows for ground water to discharge to the channel when the model-calculated head in the aquifer is higher than the stage in the channel, and allows for the possibility of surface-water discharge from the channel to the underlying aquifer when the simulated stage in the channel is higher than the model-calculated head in the underlying aquifer. The simulated stage in the marsh channel was based on the results of a surface-water model developed by the U.S. Army Corps of Engineers (Matthew Walsh, written commun., 1995).

The only stress on the simulated ground-water-flow system is ground-water pumping for public supply at the Black Pond and Beach Wells (BHW013) in the North Sagamore Water District. The Black Pond Well is the primary source of drinking water for the residents of the North Sagamore Water District and was pumped at an average rate of 0.28 Mgal/d in 1994 (Paul Gibbs, North Sagamore Water District, written commun., 1995). The Beach Well (BHW013) is used primarily as an auxiliary supply to meet increased demand during the summer season. The average pumping rate for the Beach Well for 1994 was 0.05 Mgal/d (Paul Gibbs, written commun., 1995). Ground-water withdrawal from these public-supply wells was simulated in the model as a specified flux boundary condition.

Hydraulic Properties

The hydraulic properties required for the ground-water modeling in this investigation are horizontal hydraulic conductivity and vertical hydraulic conductivity. The initial hydraulic conductivities used in the numerical model were obtained from the analysis of the aquifer test described in previous sections, and by comparing hydrogeologic sections (fig. 4) to hydraulic conductivities generalized for individual grain sizes from the results of previous investigations in the Plymouth-Carver and nearby Cape Cod aquifers (Guswa and LeBlanc, 1985; Hansen and Lapham, 1992; Barlow and Hess, 1993; Barlow, 1994; Masterson and Barlow, 1994; Masterson and others, 1996).

Generalized hydraulic conductivities were assigned for each of the hydrogeologic units described earlier and are shown in table 3. Hydraulic conductivity estimates initially were assigned to each model layer based on the lithologic boundaries shown in the hydrogeologic sections (fig. 4) and the conceptual-depositional model of glacial deposits for the study area previously discussed. Where lithologic changes occurred in a model layer, hydraulic conductivities were approximated based on thickness-weighted averages of hydraulic conductivity in a given model layer.

Initial hydraulic conductivities of the glacial and marine sediments in the flow model ranged from 1.0 to 350 ft/d (table 3). The coarser grained glacial sediments in the deltaic, moraine, and kame deposits were assigned hydraulic conductivities from 150 to 230 ft/d in the top two layers of the model and values as low as 30 to 70 ft/d in layers 3 and 4 of the model. The highest hydraulic conductivity of 350 ft/d was assigned to the beach deposits on the Cape Cod Bay side of the Sagamore Marsh in layer 1 of the model. The glacial sediments in the upper part of the flow system constitute the major aquifer in the area around the Sagamore Marsh and were assigned a hydraulic conductivity of 230 ft/d, which is consistent with the results of the aquifer-test analysis described in the previous section. The highest hydraulic conductivity used in the model was 50,000 ft/d and was assigned to model cells coincident with Island and Great Herring Ponds. The high hydraulic conductivity caused calculated hydraulic gradients in the ponds to be nearly zero, which is consistent with pond surfaces.

In general, the lowest hydraulic conductivities in the model were assigned to the glacio-lacustrine, marine, and marsh deposits. These deposits consist of fine sand, silt, and clay and constitute the confining

Table 3. Initial and calibrated hydraulic conductivity of hydrogeologic units for each layer of a ground-water-flow model for Sagamore Marsh, southeastern Massachusetts

[--, not present in model layer]

Hydrogeologic units	Hydraulic conductivity in feet per day for model layers							
	1		2		3		4	
	Initial	Cali-brated	Initial	Cali-brated	Initial	Cali-brated	Initial	Cali-brated
Glacial								
Deltaic deposits	230	230	230	230	70	70	30	30
Moraine	150	150	150	150	30	70	30	30
Kame	150	150	150	150	150	70	70	30
Glaciolacustrine deposits	30	50	--	--	--	--	--	--
Marine								
Beach deposits	350	150	--	--	--	--	--	--
Marsh deposits	1.0	.03	--	--	--	--	--	--
Marine deposits	--	--	10	50	--	--	--	--

units in the area around the Sagamore Marsh. Hydraulic conductivities specified in the model ranged from 1.0 ft/d in the marsh deposits to as high as 30 ft/d for the fine sand and silt of the glaciolacustrine deposits (table 3). The marsh hydraulic conductivity of 1.0 ft/d is significantly higher than values estimated from the tidal-response analytical method. The use of overestimates of hydraulic conductivity in the model allows for "worst case" model simulations; high marsh-sediment conductivities would allow for more drawdown of salty water from the tidal channels to the underlying aquifer and for more extensive propagation of marsh-generated tidal effects through the aquifer than would actually occur.

The vertical conductance, which is a measure of the vertical hydraulic conductivity used in MODFLOW, was specified between vertically adjacent model cells based on the relation discussed in McDonald and Harbaugh (1988). The vertical hydraulic conductivities required to calculate vertical conductances were based on ratios of horizontal to vertical hydraulic conductivity estimates for glacial sediments from the previous investigations in the Plymouth-Carver and Cape Cod aquifers. The ratios of horizontal to vertical hydraulic conductivity assumed in this investigation were 3:1 for hydraulic conductivities ranging from 225 to 350 ft/d; 5:1 for hydraulic conductivities ranging from 175 to 225 ft/d; 10:1 for hydraulic conductivities ranging from 125 to 175 ft/d; 30:1 for hydraulic conductivities ranging from 50 to 125 ft/d; 100:1 for hydraulic conductivities ranging from 1.0 to 50 ft/d; and 1:1 for hydraulic conductivities of 1.0 ft/d.

Calibration and Sensitivity

The numerical model developed for this investigation was calibrated by adjusting horizontal and vertical hydraulic conductivities in the model to provide the best match between measured and model-calculated water levels and pond levels (tables 3 and 4). Only hydraulic conductivity values were adjusted during the model-calibration process because these values were assumed to have the greatest uncertainty of all model input parameters. Recharge values were not varied in the calibration process because the values used in the model simulations are assumed to be more certain on the basis of results of the previous studies in the Plymouth-Carver and adjacent Cape Cod aquifers.

Most of the water-level data used for the model calibration process were collected in a synoptic water-level measurement during November 30 through December 2, 1984, and reported in Hansen and Lapham (1992). The data collected during this period was assumed to represent long-term average conditions in the modeled area (Hansen and Lapham, 1992).

Table 4. Measured heads for selected observation wells and pond sites, and model-calculated heads for average current conditions, Sagamore Marsh, southeastern Massachusetts

[Data from Hansen and Lapham (1992) unless otherwise noted. ft, foot]

Well identifier or pond site	Model cell location			Measured head (ft above sea level)	Model-calculated head (ft above sea level)	Difference (ft)
	Layer	Row	Column			
BHW013[1]	1	39	109	6.3	5.8	0.5
BHW206	1	24	87	3.0	11.0	-8.0
BHW286	2	38	109	6.6	5.9	.7
BHW290	1	73	73	19.0	27.4	-8.4
BHW295	1	46	85	17.1	19.8	-2.7
BHW296	1	83	78	9.9	19.2	-9.3
BHW297	1	89	78	8.4	15.7	-7.3
BHW298	1	88	86	3.6	5.1	-1.5
BHW300	1	66	76	20.8	26.2	-5.4
BHW488-53[1]	2	40	109	6.5	6.0	.5
BHW491-55[1]	2	38	116	1.5	3.8	-2.3
BHW492-5[1]	1	38	116	3.9	3.8	.1
BHW493-5[1]	1	40	119	2.9	3.5	-.6
BHW494-5[1]	1	38	113	3.5	4.0	-.5
BHW496-5[1]	1	42	126	2.7	2.5	.2
BHW497-19[1]	2	54	122	5.5	3.9	1.6
BHW498-5[1]	1	54	122	3.8	3.9	-.1
BHW499-21[1]	2	46	126	2.7	3.0	-.3
BHW500[1]	1	47	113	3.5	4.0	-.5
PWW220	1	60	45	37.1	37.4	-.3
PWW250	1	46	17	42.7	42.4	.3
PWW325	2	23	71	20.0	20.2	-.2
PWW327	2	25	75	19.0	19.0	.0
PWW437	1	24	61	23.3	25.7	-2.4
PWW485	1	84	40	38.3	39.0	-.7
Great Herring Pond	1	50	50	34.0	34.4	-.4
Island Pond	1	40	32	39.0	38.8	.2

[1] Analysis results obtained from current study.

Additional water-level information also was obtained in the area around Sagamore Marsh from observation wells installed as part of the aquifer test conducted during this investigation.

A total of 25 water-level measurements, 2 pond-level measurements, and the 45 ft water-table contour at the northwestern boundary were used in the model calibration. Of the 25 water-level measurements, 14 were from the November 30 through December 2, 1984, synoptic measurements reported in Hansen and Lapham (1992), and the remaining 11 water-level measurements were obtained during this investigation. The two pond levels were obtained from USGS topographic maps.

Generally, agreement between model-calculated and measured water levels at observation wells and ponds is close (table 4). The mean absolute error was 3.6 ft, which is about an 8-percent error over the entire slope of the water table (45 ft). The model-calculated water levels at the northwestern boundary were within 1 ft of the measured 45-foot water level, and, therefore, the specified-flux boundary determined in the previous section was not adjusted during model calibration.

Differences between model-calculated and measured water levels were largest near the Great Herring Pond/Herring River system (BHW290, BHW296, BHW297, BHW300), and near the coast (BHW206). The large differences between model-calculated and measured heads (greater than 5 ft) in the area of Great Herring Pond may be attributed to an oversimplification of the simulated ground-water/ surface-water interaction, which cannot account for the measured head loss occurring between the pond and stream, and the underlying aquifer, resulting in higher model-calculated heads.

The large difference in model-calculated and measured head (about 8 ft) near the coast may be the result of the model discretization being too coarse to accurately simulate steep hydraulic gradients near areas of coastal ground-water discharge. Also, the time periods in which the measurements were made may not be representative of mean water levels due to tidal fluctuations. In addition, ground-water discharge may occur at some distance offshore—the simulation of the salt-water interface as being at or near the coast may focus simulated ground-water discharge over a smaller area and would result in artificially high heads at the coast.

The initial estimates of horizontal hydraulic conductivity and vertical conductance were adjusted within reasonable limits during model calibration (table 3). The largest changes in hydraulic conductivities were made for the beach deposits on the eastern side of the Sagamore Marsh in layer 1 and the underlying marine clay in layer 2. Water levels showed large changes in response to changes in the horizontal hydraulic conductivities and the corresponding vertical conductances in this area of the model. Therefore, the vertical connection between these two deposits apparently has a significant control on the water discharging to the coast.

Water levels near Great Herring Pond and Herring River showed little change in response to large changes in the vertical and horizontal connections of the simulated Great Herring Pond and the surrounding aquifer. These connections were decreased by as much as four orders of magnitude from initial estimates to create a larger head loss between the pond and the underlying aquifer, and yet water levels in this vicinity declined by less than 1 ft. Aquifer recharge also was decreased by 5 percent to improve the match between measured and model-calculated water levels.

Once the flow model was considered calibrated, the hydraulic budget was calculated for current (1994) conditions. Components of the calculated hydrologic budget of the modeled area are given in table 5. Total inflow to the modeled area is 23.8 ft^3/s, of which 84 percent is recharge from precipitation and the remaining 16 percent is inflow from ground-water flow across the northwestern boundary and beneath the Herring River. Nearly all the water leaving the modeled area (93 percent) discharges to the coast. Ground-water discharge to seeps along the perimeter of Sagamore Marsh only represents about 5 percent of the total water leaving the simulated ground-water system, and ground-water pumping in the North Sagamore Water District for current (1994) conditions represents the remaining 2 percent. There was no simulated ground-water seepage directly to the marsh channel for current conditions.

Table 5. Hydrologic budget for the aquifer system calculated for average current conditions, Sagamore Marsh, southeastern Massachusetts

Budget component	Rate of flow	
	Cubic feet per second	Million gallons per day
Inflow		
Recharge	20.0	12.9
Northwestern boundary	2.9	1.9
Herring River	.9	.6
Total inflow	23.8	15.4
Outflow		
Pumping from wells	0.5	0.3
Discharge to marsh	1.1	.8
Discharge to coast	22.2	14.3
Total outflow	23.8	15.4

Simulation of Ground-Water-Flow System

Ground-water flow was simulated near Sagamore Marsh for current conditions and for increased tidal-stage conditions. The analysis of the ground-water-flow system included a comparison of water levels before and after the proposed marsh restoration to determine possible effects of increased tidal stage on ground-water levels in and around the marsh. The zone of contribution to North Sagamore Water District's Beach Well also was determined by use of the flow model for current and increased tidal-stage conditions.

The zone of contribution for the Beach Well was determined by particle tracking, which allows for the delineation of the zone of contribution for selected ground-water sinks such as pumping wells. By use of particle-tracking simulations, the particles of water can be tracked from areas of ground-water recharge to areas of ground-water discharge, thereby identifying the area at the water table that is the source of water for the ground-water sink. A particle-tracking program, MODPATH (Pollock, 1994), was used to calculate flow paths from the results of the ground-water-flow model. The only parameter in addition to those used in the numerical model that is required for the particle-tracking analysis is porosity, which is required for time-of-travel estimates and does not otherwise affect the size, shape, or location of the zone of contribution to the well. Because traveltimes are not the focus of this study, and no information is available on the porosity of the regional aquifer system near Sagamore Marsh, a uniform value of 0.30, which is based on published values of porosity for sand, was selected for this investigation (Freeze and Cherry, 1979).

Current (1994) Conditions

The model-calculated zone of contribution for the Beach Well, which is defined as the area at the water table through which recharge entering the ground-water-flow system supplies water to a pumping well at a given rate, was calculated by tracking 125 particles backward from the simulated location of the well screen to the model-calculated water table for 1994 pumping conditions (fig. 14). The number of particles used in this simulation (125) was chosen arbitrarily to adequately represent the water particles in the model node containing the public supply well.

The results of the model simulations indicate that the source of water for the Beach Well pumping at the current rate is ground-water recharge that occurs primarily between the pumped well and Great Herring Pond. The results also indicate that the zone of contribution for the pumping well extends to Great Herring Pond and may include areas adjacent to, and downgradient of the pumping well. However, the results do not indicate that the zone of contribution for the pumping well will extend into the area of Sagamore Marsh for current pumping conditions.

The zone of contribution calculated for the Beach Well, which has been used primarily as an auxiliary supply to provide additional water to meet increased demand during the summer, was based on an average daily pumping rate of 8,380 ft^3/d (or 0.063 Mgal/d) for the entire 1994 calendar year. The total amount of water pumped from the well in 1994 was 22.88 Mgal (Paul Gibbs, written commun., 1995). Although the Beach Well was not pumped every day in 1994, the pumping rate was substantially higher than 0.063 Mgal/d on those days in which the well was actually pumped. A particle-tracking analysis in which pumping and recharge rates could be varied with time would require a transient particle-tracking analysis, which was beyond the scope of the current investigation.

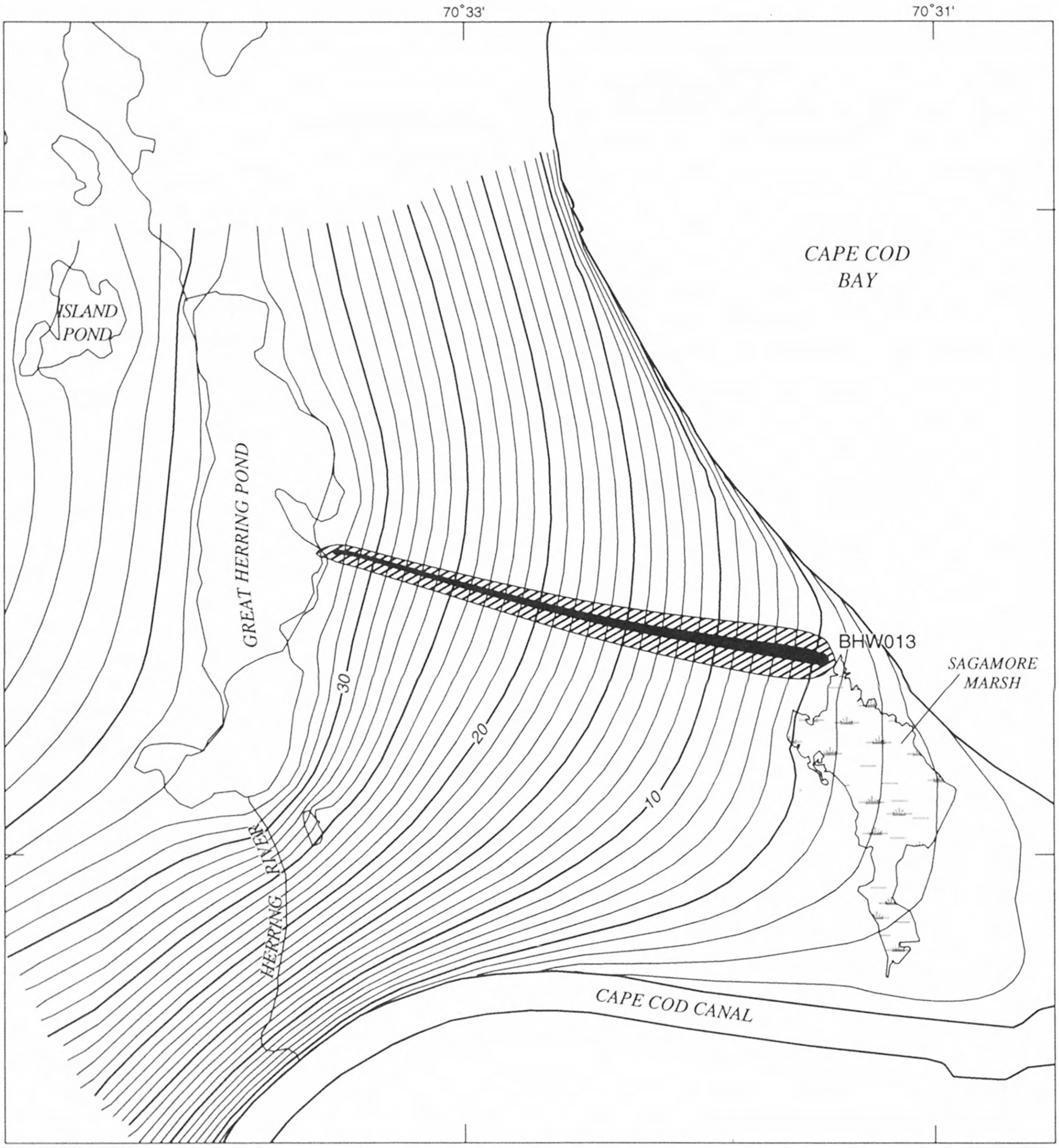
A second simulation was made using an increased pumping rate at the well. All public-supply demands for the North Sagamore Water District were assumed to be met by pumping at the Beach Well; the simulated pumping rate at the Beach Well was increased from an average daily pumping rate of 8,380 to 43,124 ft^3/d (Paul Gibbs, written commun., 1995). This nearly seven-fold increase in the average daily pumping rate at the Beach Well resulted in a substantially larger zone of contribution to the well than that determined for average 1994 pumping conditions, yet the zone of contribution still extends northwestward of the well toward Great Herring Pond (fig. 14). Ground-water flow from the marsh to the well was not induced by pumping in either of the model simulations.

Increased Tidal Stage in Marsh

As part of the proposed changes in saltwater flow to the marsh, the U.S. Army Corps of Engineers is considering the possibility of increasing the culvert size at the entrance to the marsh to increase tidal flow into the marsh. The maximum proposed increase in culvert size is from the current single, 4-foot diameter culvert to two culverts with the dimensions of 10×20 ft. This increase in culvert size is projected to increase the tidal exchange in the marsh such that the high-tide stage in the marsh channel increases by about 1 foot in the lower marsh and about 0.6 ft in the upper marsh (Matthew Walsh, written commun., 1995).

The numerical model was used to evaluate the effect of the increased stage in the marsh channel on ground-water flow near Sagamore Marsh. Model-calculated ground-water levels indicated that no change in ground-water levels in and beneath the marsh would result from increased tidal stages.

Changes in model-calculated ground-water levels between current and increased tidal-stage conditions did not occur and may be a result of the large model discretization (200×200 ft) relative to the marsh-channel width (about 10 ft). The large model discretization may result in an under-projection of water levels immediately adjacent to the marsh channel, and, therefore, this model discretization may not be adequate to project water-level changes immediately adjacent to the marsh channels. Model results do indicate that the effect of the increased tidal-channel stage on the ground-water table would dissipate rapidly over a short distance (in one model cell) and does not affect ground-water levels near the beach at Cape Cod Bay or the Beach Well. Also, the results of the numerical model are consistent with the analytical solution described in previous sections, where the effect of post-restoration tidal-channel stage fluctuations of greater than 0.01 ft in the marsh sediments dissipated in less than 50 ft from the marsh channel.



Base from U.S. Geological Survey digital data
1:24,000, 1991
Universal Transverse Mercator projection
Zone 19

0 1 2 MILES
0 1 2 KILOMETERS

EXPLANATION

ZONE OF CONTRIBUTION FOR WELL BHW013

Current pumping rate (1995)

Increased pumping rate

—10— MODEL-CALCULATED POTENTIOMETRIC CONTOUR IN MODEL LAYER 2-- Contour interval, 1 foot. Datum is sea level

BHW013 ● PUMPING WELL AND NUMBER

Figure 14. Zone of contribution to site BHW013 near Sagamore Marsh for current and increased pumping conditions and model-calculated heads in model layer 2 for current pumping conditions, southeastern Massachusetts.

SUMMARY AND CONCLUSIONS

Sagamore Marsh is a salt marsh in a coastal area of southeastern Massachusetts that is bounded to the east by Cape Cod Bay and to the south by the Cape Cod Canal. Saltwater inflow from the Cape Cod Canal to the marsh is constricted by a 4-foot diameter culvert. In an effort to restore salt-marsh and estuarine habitat in the marsh, the U.S. Army Corps of Engineers has proposed to increase the amount of saltwater that enters the marsh by widening the culvert between the Cape Cod Canal and the marsh. A numerical model by the U.S. Army Corps of Engineers estimates that high-tide stages in tidal channels will increase by less than 1.0 ft in the lower part of the marsh and by less than 0.6 ft in the upper part of the marsh.

The investigation was done to determine whether the proposed increase in the amount of saltwater entering the marsh at high tide would be likely to increase hydraulic heads in the underlying aquifer and whether the proposed restoration of the marsh would be likely to cause saltwater intrusion into an adjacent large-capacity public-supply well. In April 1995, the U.S. Geological Survey, in cooperation with the Army Corps of Engineers, began an investigation into the geology and ground-water hydrology of the Sagamore Marsh area to improve the understanding of the local hydrogeologic framework and ground-water-flow system and to address some of the concerns regarding how proposed changes in tidal-channel stage might affect the ground-water-flow system.

The regional geology is characterized by sequences of coarse-grained glaciofluvial outwash and finer grained glaciolacustrine, moraine, and till deposits that are part of the regional aquifer system known as the Plymouth-Carver aquifer system. In the area around Sagamore Marsh, these deposits consist of glaciolacustrine and deltaic sediments and extend to a depth of greater than 175 ft; the bedrock surface is estimated to be about 200 ft below land surface. Four major hydrogeologic units were defined in the glacial sediments: (1) shallow glaciolacustrine sediments consisting of brown silt and sandy clay, (2) underlying deltaic sediments consisting of fine to coarse brown sand, (3) glaciolacustrine sediments consisting of interbedded fine to coarse sand, silt, and clay; and (4) glaciolacustrine sediments consisting of fine gray sand and silt. The fine to coarse deltaic sand constitutes an important aquifer in the Sagamore Marsh area.

Sagamore Marsh influences shallow ground-water flow near the marsh and shallow ground water discharges along the edge of the marsh. Although most of the regional flow system is unconfined, the low vertical and horizontal hydraulic conductivity of the fine-grained glaciolacustrine sediments along the western edge of the marsh and of the fine-grained marsh sediments cause confining conditions beneath the marsh; for this reason, the regional flow system is referred to as a semiconfined flow system near the marsh. Unconfined (or water table) conditions prevail in the surficial marsh sediments, beyond the western and northwestern extent of the confining deposits, and along a barrier beach on the northeastern side of the marsh; confined conditions prevail just beneath the marsh sediments and along at least the northwestern edge of the marsh near public-supply well BHW013.

Tidal ranges in the marsh tidal channels were between 1.0 and 1.5 ft and tidal ranges in Cape Cod Bay were on the order of 9 ft during the period of study. Tidal pulses from tidal channels and Cape Cod Bay were rapidly attenuated in the ground-water system. Tidal ranges in the water table beneath the barrier beach were less than 0.1 ft along the northeastern edge of the marsh and about 0.5 ft near Cape Cod Bay. Tidal ranges in the regional aquifer were between 0.1 and 0.2 ft. Tidal pulses beneath the barrier beach were more in phase with tidal pulses originating from Cape Cod Bay, whereas tidal pulses in the regional aquifer were more in phase with tidal pulses originating from tidal channels in Sagamore Marsh.

A 5-day aquifer test was done as part of the study to determine the response of the ground-water-flow system near the marsh to pumping at a large-capacity public-supply well near the marsh, and to determine the hydraulic properties of the regional aquifer near the marsh. After 5 days of pumping, drawdown at the pumped well was 17.5 ft below the nonpumping (static) water level and drawdown in wells screened at the same horizon as the supply well at distances of from 100 to 725 ft from the supply well ranged from 4.92 to 1.67 ft. At the observation well closest to the marsh, drawdown at the end of the 5-day test was 3.18 ft. A drawdown of 0.37 ft was measured at a well about 1,450 ft from the pumping well. Results of the aquifer test were used to estimate hydraulic properties of the aquifer near the marsh. The estimates are: a transmissivity (T) of from 9,300 to 10,900 ft^2/d, a horizontal hydraulic conductivity (K_H) of from 181 to 213 ft/d, a ratio of vertical to horizontal hydraulic conductivity (K_Z/K_H) of 1:44, a ratio of storage coefficient to specific yield (S/S_y) of 0.09, a storage coefficient (S) of 4.4×10^{-4} to 2.6×10^{-3} and a specific yield (S_y) ranging from 0.01 to 0.03. The low values of K_Z/K_H and of S_y probably result from the presence of a hydrogeologic unit of glaciolacustrine silt and clay at the top of the stratigraphic column near the supply well.

The diffusivity (T/S) of the marsh deposits at two sites in the marsh was determined from the ratio of tidal ranges in tidal channels to those in adjacent wells. In the lower marsh, where the marsh sediments were coarser grained, aquifer diffusivity was estimated to be 380 ft^2/d; in the upper marsh, diffusivity was estimated to be 170 ft^2/d. The two values correspond to hydraulic conductivities of 2.5×10^{-3} and 1.7×10^{-3} ft/d, respectively. These values are between the ranges of values reported for marsh peat and marine clay, respectively, and are consistent with the lithology at the sites.

The calculated aquifer diffusivities were used to estimate the maximum distances from the tidal channels where tidal pulses in the ground water would exceed 0.01 ft, using an analytical-modeling technique. Estimates were made using current high-tide stages and increased high-tide stages predicted following marsh restoration. In the upper marsh, the maximum distance where tidal ranges in the aquifer would exceed 0.01 ft for current conditions is 24.4 ft. When the increased tidal stages were used, the maximum distance was 26.7 ft. Maximum distances estimated for the lower marsh site for present and maximum predicted high-tide stages were 37.1 and 42.0 ft, respectively. The data indicate that tidal pulses are rapidly attenuated in the aquifer and that changes in tidal stages in the channels and in flooded areas of the marsh will have little effect on ground-water levels near the marsh. The analytical model also was used to predict maximum distances where tidal pulses would propagate in the underlying fine sand aquifer. Diffusivities in the aquifer were estimated from lithology and aquifer test results. Maximum distances where tidal pulses exceeded 0.01, 0.05, and 0.1 ft in the sand aquifer beneath the lower marsh were 662.2, 354.7, and 222.3, respectively. Tidal pulses in the sand aquifer beneath the upper marsh exceeded 0.01 ft for a distances as much as 419.8 ft and exceeded 0.05 ft for a distance as much as 112.3 ft. Fluctuations of this magnitude are much smaller than fluctuations caused by tidal fluctuations in Cape Cod Bay and by natural precipitation events.

A numerical ground-water-flow model was developed to represent the Sagamore Marsh area. Results of the model simulations indicate that the zone of contribution to the public-supply well near the marsh extends from the well northwestward toward Great Herring Pond and that simulated increases in marsh-channel tidal stage have a negligible effect on the location of the zone of contribution to the well. Model results are consistent with field data and with the analytical-modeling results, which showed that tidal pulses originating from the tidal channels extend only about 42 ft laterally into the aquifer, which is less than the model discretization of 200 ft. Ground-water flow from the marsh to the well was not induced by pumping in any of the model simulations.

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