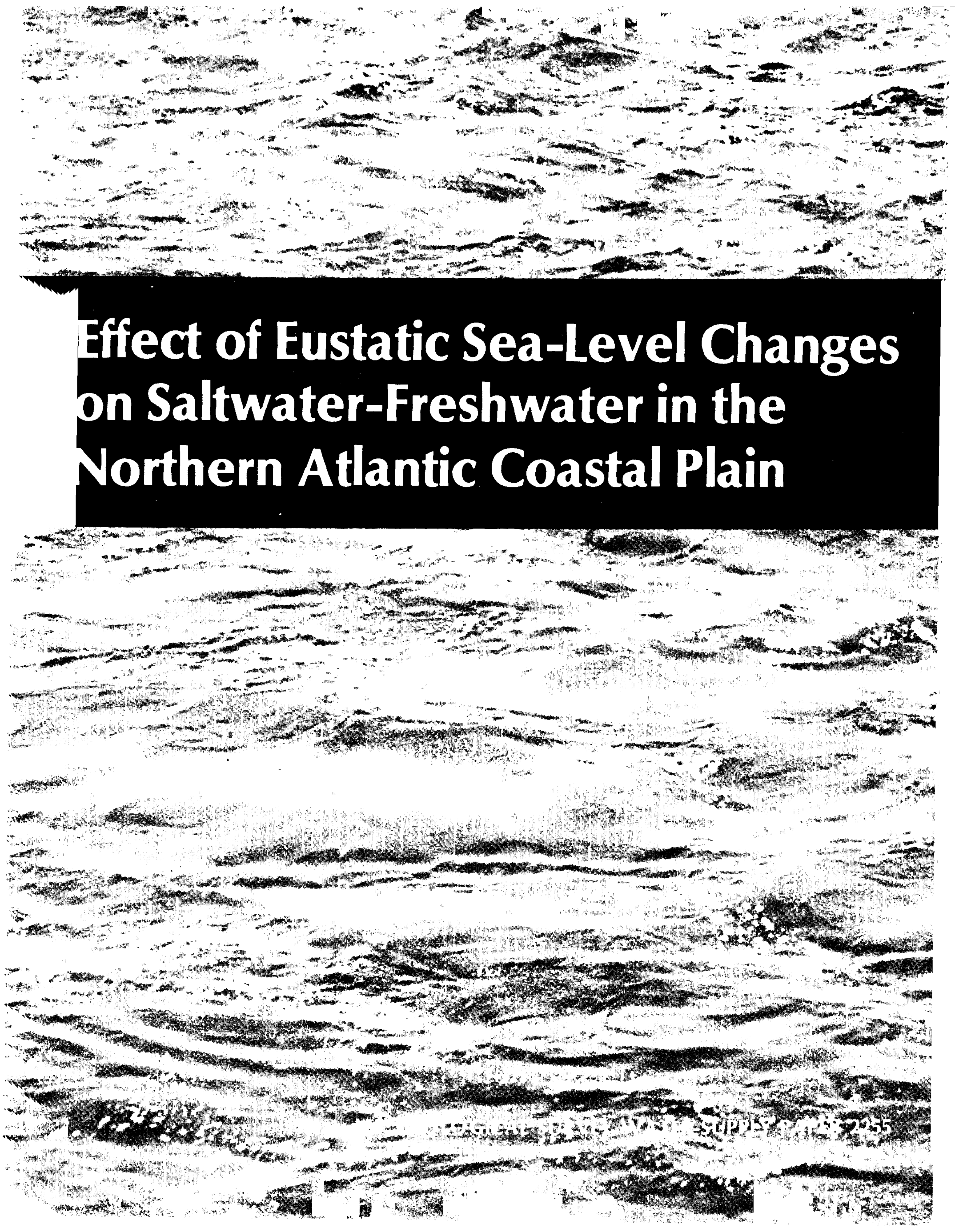

Effect of Eustatic Sea-Level Changes on Saltwater-Freshwater in the Northern Atlantic Coastal Plain

OGICAL SURVEY WATER-SUPPLY PAPER 2255

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By HAROLD MEISLER, P. PATRICK LEAHY,
and LEROY L. KNOBEL

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FIGURES

TABLES

Metric Conversion Factors

For those readers who prefer to use metric units rather than inch-pound units, the conversion factors for terms used in this report are listed below:

Multiply	By	To obtain SI metric unit
inch (in.)	25.40	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)

National Geodetic Vertical Datum of 1929 (NGVD of 1959): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called mean sea level. NGVD of 1929 is referred to as sea level or present sea level in this report.

Effect of Eustatic Sea-Level Changes on Saltwater-Freshwater Relations in the Northern Atlantic Coastal Plain

By Harold Meisler, P. Patrick Leahy, *and* LeRoy L. Knobel

Abstract

A transition zone between fresh ground water and underlying saltwater, 1,000 to 2,000 feet thick, underlies part of the northern Atlantic Coastal Plain and adjacent Continental Shelf. The zone is known to extend 60 miles off the New Jersey coast. The effect of eustatic sea-level fluctuations on the development and location of the zone was analyzed by means of a finite-difference computer model. The model simulates, in cross section, the sedimentary wedge from the Delaware River estuary to the Continental Slope as a highly anisotropic porous medium. Simulated steady-state freshwater flow is separated from static saltwater by a sharp interface. Freshwater pressure at the simulated interface equals the pressure exerted by a column of static seawater.

The model was used to test the sensitivity of the simulated interface position to anisotropy (the ratio of lateral to vertical hydraulic conductivity) as well as to sea-level elevation. Increasing the anisotropy causes the interface to be shallower and to extend farther offshore. Lowering sea level causes the interface to be deeper and to extend farther offshore. Simulations that use hydraulic conductivities based on available data suggest that the transition zone in southern New Jersey and beneath the adjacent Continental Shelf is not in equilibrium with present sea level. The position of the transition zone probably reflects a long-term average sea level of between 50 and 100 feet below present sea level. Because of the most recent rise in sea level, however, the transition zone is probably moving slowly landward and upward towards equilibrium with present sea level.

The cyclic movement of salty ground water in response to eustatic sea-level fluctuations during the Quaternary and late Tertiary caused the saltwater to mix with freshwater; this produced the broad transition zone. A much thinner transition zone immediately below the sea floor overlies the freshwater that extends offshore. This zone was probably produced by smaller sea-level fluctions from storms and tides.

The interpretation that waters in the broad transition zone are mixtures of freshwater and saltwater is supported by a geochemical study of the waters. The freshwater is predominantly sodium bicarbonate in character. The saltwater in North Carolina is primarily seawater. From New Jersey to Virginia, it is a sodium calcium chloride brine which has a chlorinity several times that of seawater.

INTRODUCTION

Salty ground water underlies freshwater in the eastern part of the Atlantic Coastal Plain (fig. 1). There is no sharp interface between the deepest known freshwater and the underlying saltwater; rather there is a zone of transition through which chlorinity of the water increases gradually with increasing depth. The transition zone between chloride concentrations of 1,000 mg/L (milligrams per liter), the lowest considered in this report, and 18,000 mg/L, the approximate chloride concentration of seawater, is generally 1,000 to 2,000 ft thick vertically (Meisler, 1981, fig. 1). The thickness of the transition zone at three well sites is indicated by the relation of chloride concentration to sampling depth below sea level shown in figure 2.

The recent discovery (Hathaway and others, 1976, 1979) of ground water considerably fresher (salinity less than 6 parts per thousand) than seawater 60 mi off the New Jersey coast has prompted speculation that freshwater was "trapped in shelf sediments during the Pleistocene glacial maximum" (Hathaway and others, 1979, p. 523). Indeed, many sea-level changes have occurred during the Quaternary as well as earlier throughout the Tertiary Period. These have caused large areas of the Continental Shelf to be repeatedly covered and uncovered by seawater. Such large sea-level fluctuations must have had impact on the direction and magnitude of flow of both fresh and salty ground water and on the development of a transition zone between them.

Purpose and Scope

This paper explains the effect of eustatic sea-level changes on the physical and chemical relations between fresh and salty ground water in the northern Atlantic Coastal Plain and adjacent Continental Shelf. It is part of a study by the U.S. Geological Survey entitled "Regional aquifer system analysis of the northern Atlantic Coastal Plain."

A finite-difference computer model was used to simulate the effect of eustatic sea-level changes on the development and location of the transition zone between fresh and salty ground water. Geochemical studies of ground water in the area from North Carolina to New Jersey were made to relate the composition of ground water in the transition zone to the composition of fresh ground water and seawater.

Previous Investigations

The relation of fresh ground water in the northern Atlantic Coastal Plain to the underlying saltwater has been studied for many years. Several studies have attempted to delineate the boundary between freshwater and saltwater (usually defined as 250 or 350 mg/L of chloride) and to relate the boundary to water levels and freshwater flow patterns. Examples of such studies are by Barksdale and others (1958) in New Jersey, Back (1966), from New Jersey to Virginia, and Upson (1966), from New York to Maryland.

Contour maps depicting the depth to water of specified chemical quality appear in several reports. Cushing, Kantrowitz, and Taylor (1973) contoured the altitude of ground water containing 1,000 mg/L of dissolved solids on the Delmarva Peninsula. Heath, Thomas, and Dubach (1975, fig. 8.20) contoured the depth below land surface to ground water containing 250 mg/L of chloride in North Carolina. In Virginia, the altitude of water containing 250 mg/L of chloride has been contoured by the Commonwealth of Virginia, State Water Control Board (1978, p. 44 and 1979, p. 57) and by Larson (1981, plate 2). Meisler (1981) contoured the

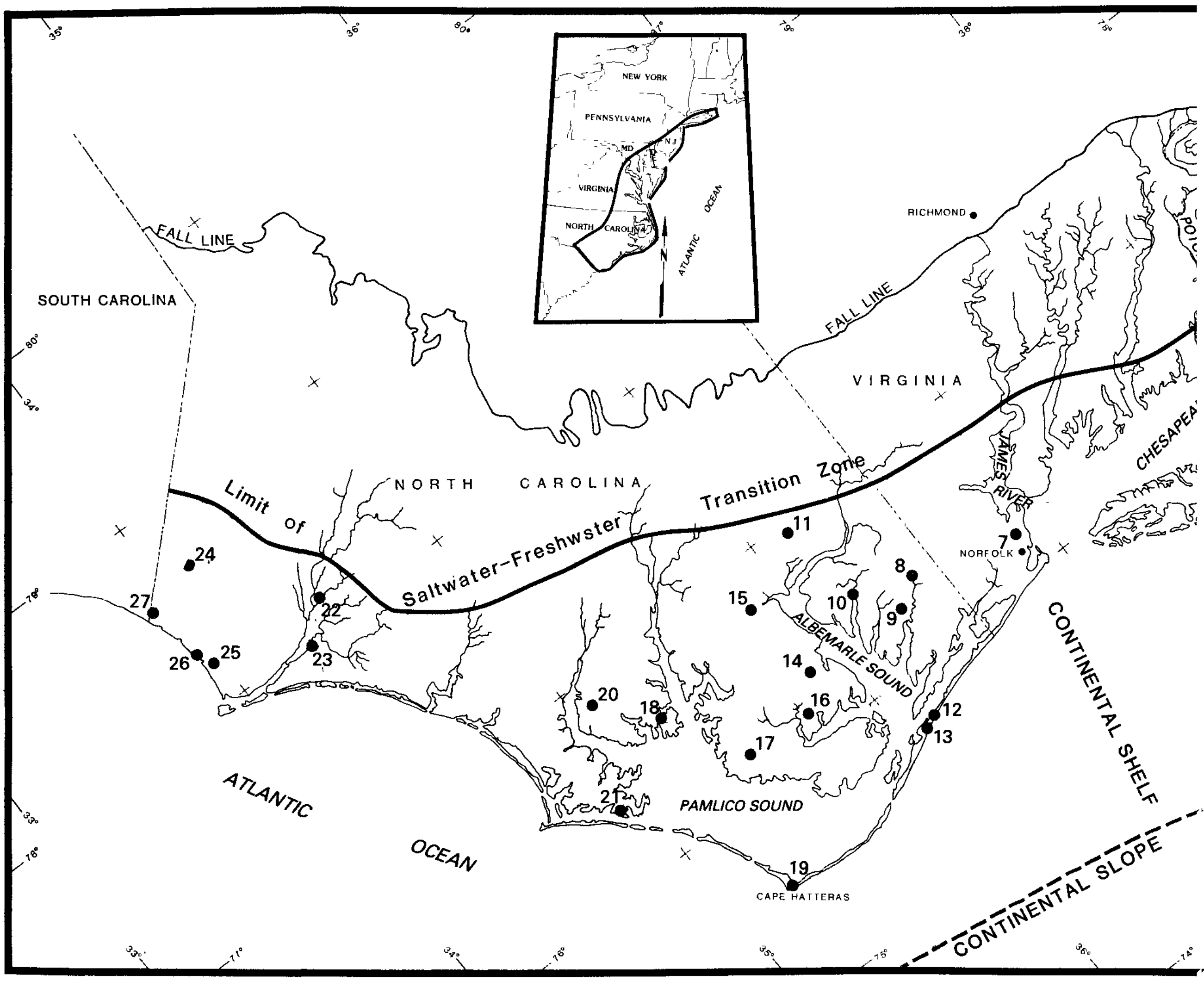


Figure 1. The northern Atlantic Coastal Plain.

depth below sea level to chloride concentrations of 250, 1,000, 10,000, and 18,000 mg/L in the Coastal Plain sediments from North Carolina to New Jersey and included part of the adjacent Continental Shelf. Chloride data were obtained from chemical analyses and interpretation of geophysical logs. Data from the Continental Shelf were obtained from water squeezed from clay cores collected during the Atlantic Margin Coring (AMCOR) Project (Hathaway and others, 1976, and Frank T. Manheim, U.S. Geological Survey, written commun., 1980).

None of these studies examined the chemistry of the saltwater. Manheim and Horn (1968), however, showed the distribution of salinity in a series of deep wells along the Atlantic Coast from Long Island, N.Y., to Key West, Fla., and described the general water chemistry. They discussed several possible sources for the salty water and mechanisms to account for development of brines.

Methods of Study

The present study relates the position and development of the saltwater-freshwater transition zone to the hydrodynamics of a freshwater flow system that has been altered periodically by large-scale sea-level fluctuations. The geochemistry of the transition zone is then examined.

A cross-sectional computer model was developed to simulate the freshwater flow system. The location of the section studied is from the Delaware River estuary near the common border of New Jersey, Delaware, and Pennsylvania, east southeastward through Atlantic City, N.J., to a point on the Continental Slope 90 miles from

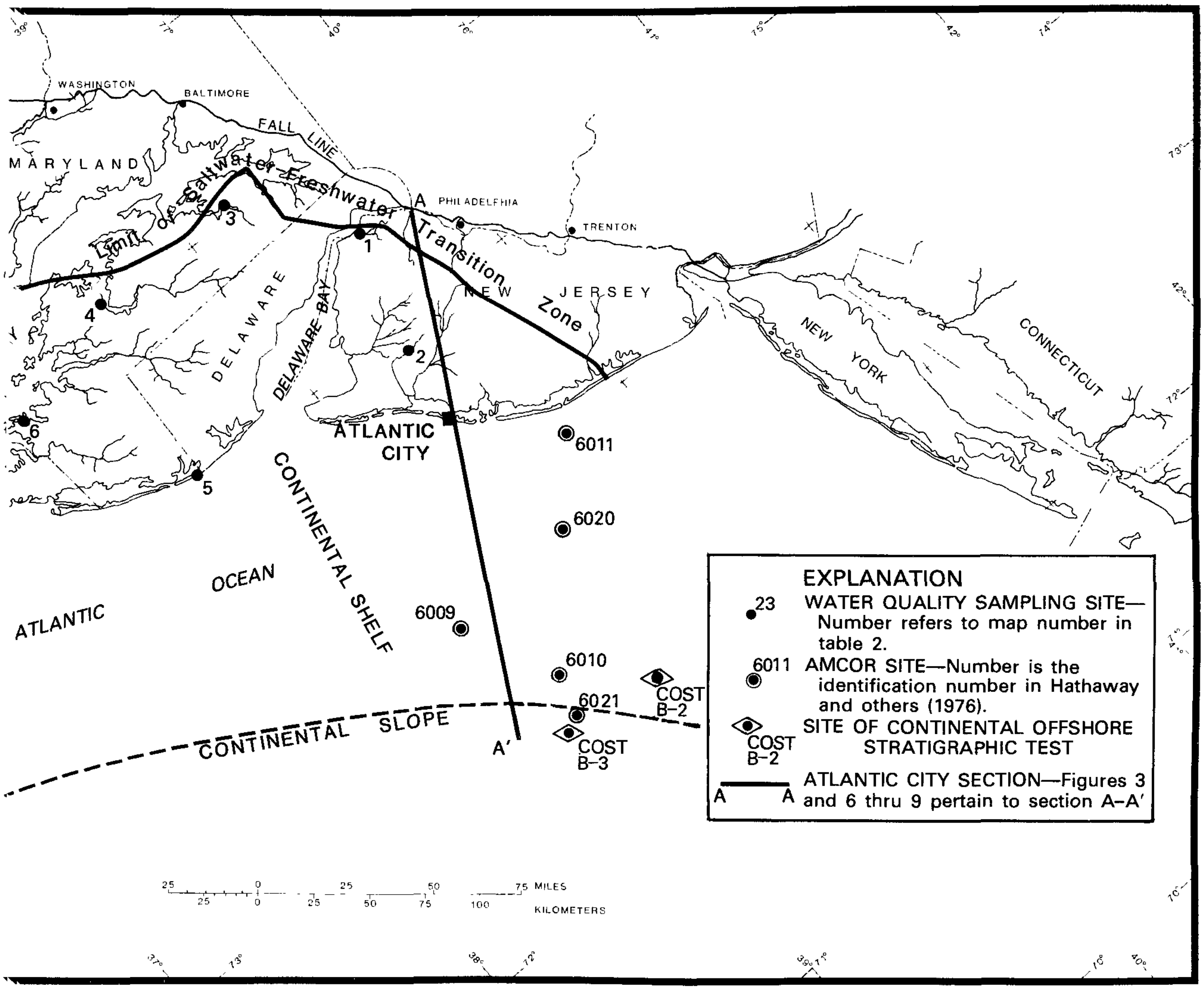


Figure 1. Continued

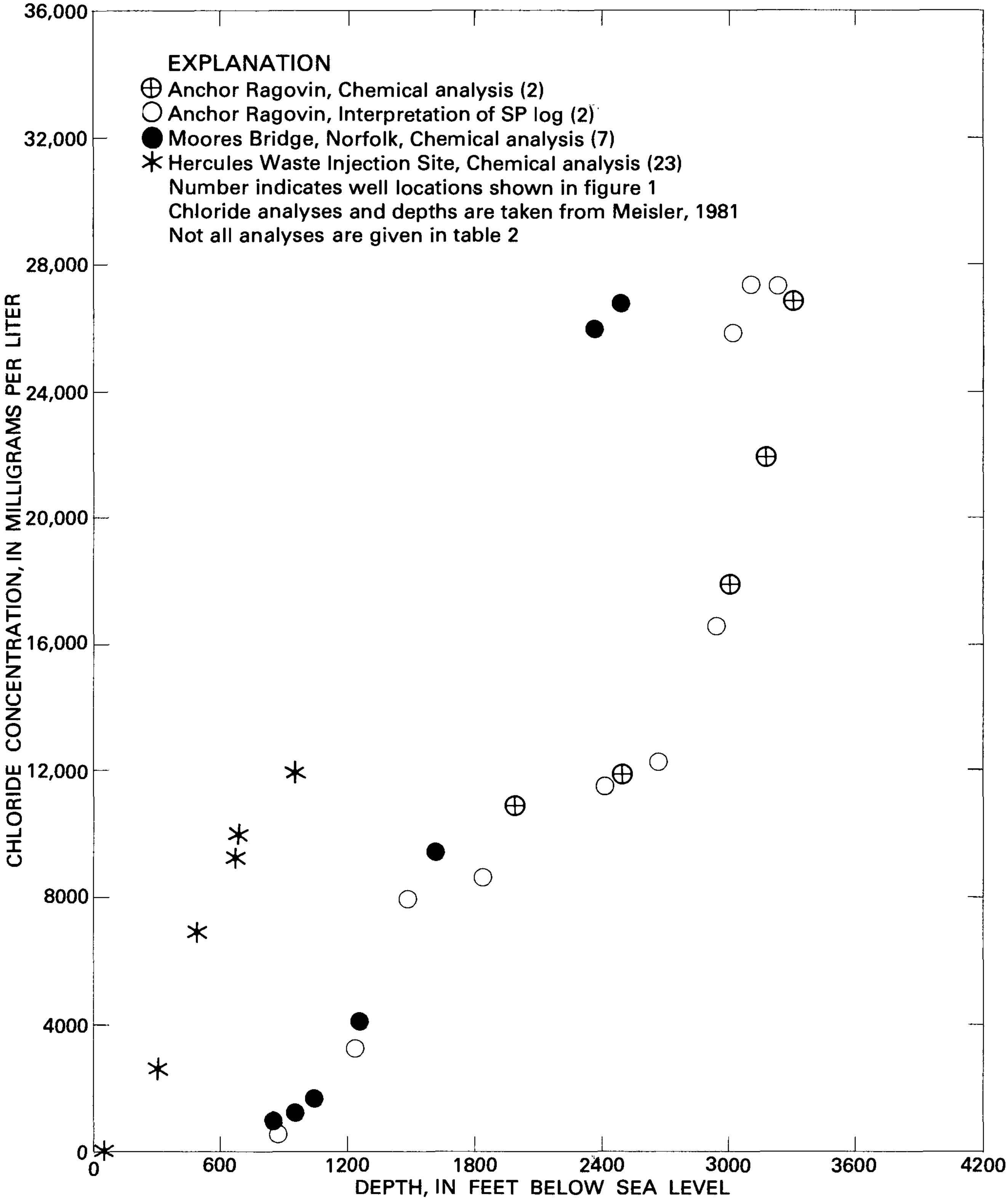


Figure 2. Relation of chloride concentration to sampling depth in selected wells.

the coast (A–A′ in fig. 1). The section, called the Atlantic City section in this report, traverses the area where the AMCOR data indicate relatively fresh water offshore. The model simulates steady-state flow within the freshwater system and treats the boundary between the freshwater and the static saltwater as a sharp interface. The model was used to determine the equilibrium position of the interface for several sea-level positions and for different distributions of lateral and vertical hydraulic conductivity. The computed interfaces were compared with the distribution of chloride in the zone of transition.

The distribution of the principal ions within the freshwater-saltwater transition zone was studied in order to define the chemical character of waters in the zone and to determine the geochemical processes occurring there. As the transition zone is defined as a zone in which chlorinity increases with depth, the study was based largely on graphs that show the relation of ion concentration to chloride concentration. Chloride was assumed to be a conservative ion.

HYDROGEOLOGY

Regional Setting

The Atlantic Coastal Plain sediments form a wedge that thickens from its western edge, the Fall Line, to 8,000 ft along the coast of Maryland, 10,000 ft at Cape Hatteras, N.C., and to more than 45,000 ft beneath the Continental Shelf and Slope. The beds generally dip southeastward and range in age from Jurassic to Holocene. On the emerged Coastal Plain the sediments consist predominantly of unconsolidated sand, silt, and clay of Cretaceous age or younger. Limestone occurs in North Carolina. Eastward, beneath the Continental Shelf and Slope, sediments of Jurassic and Cretaceous age are predominantly consolidated. Thick deposits of Jurassic age include carbonate banks and reefs, salt, and possibly other evaporite deposits.

Atlantic City Section

The Atlantic City section is about 150 miles long, from the Delaware River estuary on the west to a point on the Continental Slope on the east (A–A′ in fig. 1). The generalized stratigraphy of the New Jersey part of the section, shown in figure 3, is adapted from Brown, Miller, and Swain (1972). The offshore part is based partly on seismic interpretation by Grow and Klitgord (1980). In the COST (Continental Offshore Stratigraphic Test) B–3 well on the Continental Slope, located near the eastern end of the Atlantic City section, the Jurassic-Cretaceous contact has been placed variously at 12,200 ft (Poag, 1980) and 13,400 ft below sea level (Valentine, 1980). Depth to oceanic basement here is about 45,000 ft below sea level (Grow, 1980, p. 119, fig. 55).

Thickness of the Coastal Plain sediments in the vicinity of the Delaware River estuary ranges from a featheredge west of the estuary to about 150 ft on the east side. The channel of the Delaware River in this area is about 120 ft deep below sea level. About 100 ft of the channel is filled with late Wisconsinan and Holocene sediments (Owens and Minard, 1979, p. D35, fig. 30). During low stands of sea level in the Quaternary, this channel probably contained much less sediment; during high stands of sea level, however, it was probably filled as at present.

Configuration of the land surface and the ocean floor is a significant feature of the computer simulation model. Altitudes of the section range from sea level at the Delaware River estuary and the coast to more than 125 ft above sea level near the drainage divide between the Delaware River and the Atlantic Ocean. Southeast of the coastline, the Continental Shelf slopes gently at about 3 ft per mile for about 80 miles. On the Continental Slope, east of the Continental Shelf, the ocean bottom slopes southeastward at about 400 ft per mile (Maher, 1971, pl. 1).

Configuration of the saltwater-freshwater zone of transition is shown on figure 3. The isochlors are adapted from maps shown in Meisler (1981) and from chloride analyses from the AMCOR project (Hathaway and others, 1976 and Frank T. Manheim, U.S. Geological Survey, written commun., 1980). The configuration of the isochlors offshore is similar to that of the salinity contours prepared by F. A. Kohout (Hathaway and others, 1976, appendix C, p. 215).

Hydraulic conductivities of the Coastal Plain aquifers in New Jersey generally range from less than 10 to 100 ft/d. Vertical hydraulic conductivities of confining beds generally range from 5×10^{-3} to 1×10^{-5} ft/d (M.M. Martin, U.S. Geological Survey, written commun., 1982). The hydraulic properties of the sediments under the Continental Shelf and Slope are not well known. That the sediments tend to become less permeable with both depth and increased distance from the coast is suggested by intrinsic lateral permeabilities of cores from the COST B–2 well located 90 miles east of Atlantic City. The lateral hydraulic conductivity, computed from the intrinsic permeability of a sample obtained at a depth of about 2,000 ft below sea level is about 2.5 ft/d. Intrinsic permeabilities of samples from depths below sea level of 2,000 to 7,000 ft are about 10 times greater than those from 10,000 to 11,000 ft and more than 200 times greater than those from 12,000 to 14,000 ft (Scholle, 1980, p. 81, fig. 60).

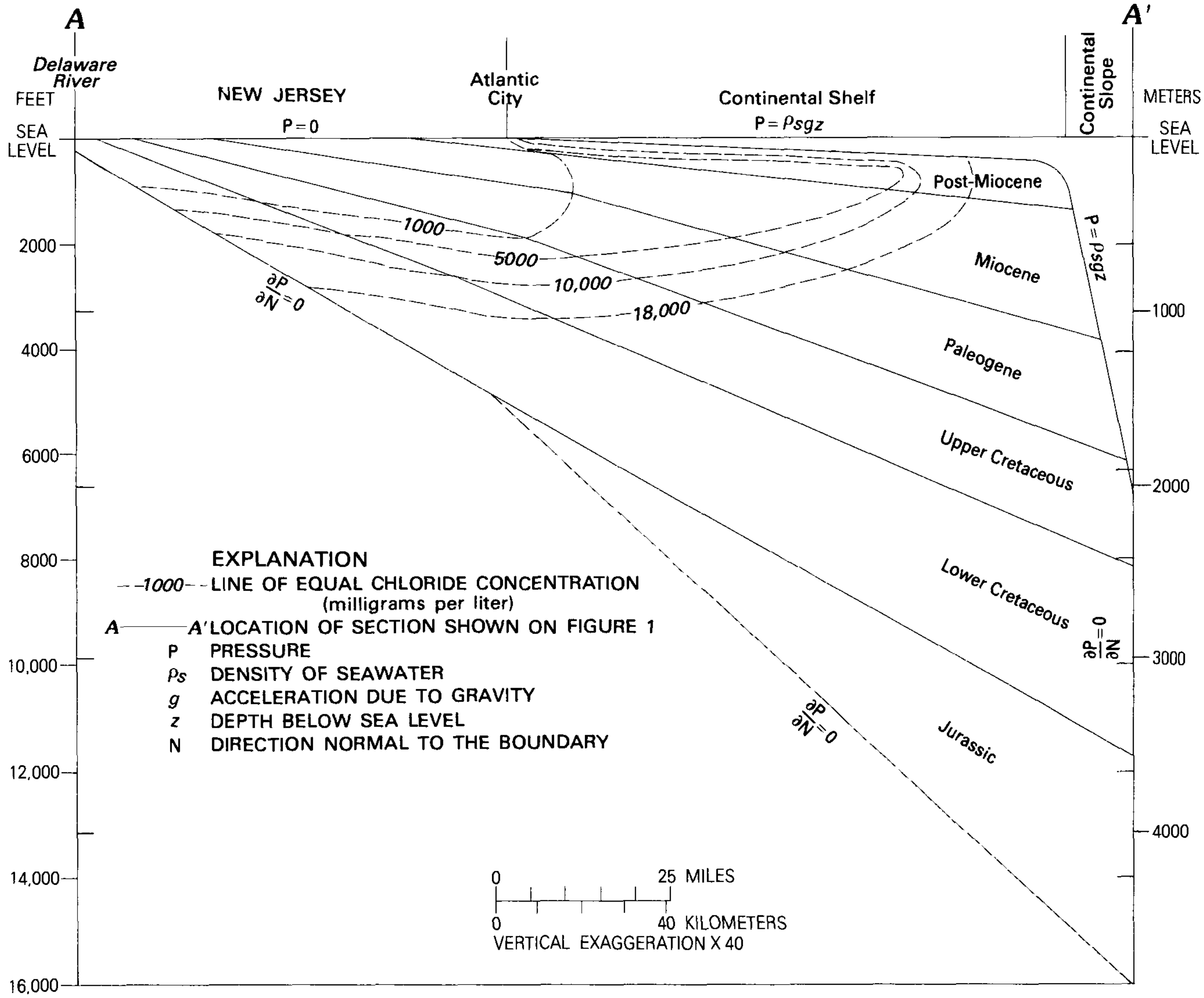


Figure 3. Generalized hydrogeologic section and conceptual model, southern New Jersey to the Continental Slope.

SEA-LEVEL HISTORY

Studies of global changes in sea level have been based on a variety of criteria. Vail and others (1977) developed curves that show changes in sea level from the Cambrian Period through the Pleistocene Epoch. These curves were developed from studies of onlap of coastal deposits in maritime sequences. A sea-level curve for the Tertiary (Vail and others, 1977, p. 87, fig. 3) is shown in figure 4*A*. The curve indicates that sea level during the early Tertiary was generally much higher than during the late Tertiary. Sea level during the middle Miocene and part of the Pliocene was higher than present sea level, whereas it was lower during the late Miocene and most of the Quaternary.

More detailed sea-level curves have been developed for the Quaternary Period using oxygen-isotope data. A eustatic sea-level curve for the past 130,000 years was develped by Shackleton and Opdyke (1973, p. 45, fig. 7). They correlated the volume of glacial ice with oxygen-isotope data from a deep-sea core from the Pacific Ocean, and hence calculated worldwide sea levels for the past 130,000 years. Their curves show a sea-level stand of 390 ft below present sea level during the Wisconsinan glacial maximum of 16,000 to 17,000 years ago.

Zellmer (1979, p. 61, fig. 16) recalibrated Shackleton and Opdyke's 130,000-year sea-level curve. She matched the Wisconsinan low-stand value of 305 ft below present sea level from Dillon and Oldale (1978, p. 59, fig. 5) to the corresponding point on the oxygen-isotope curve. The resulting relation of oxygen-isotope composition to sea-level changes was applied to Shackleton and Opdyke's (1973, p. 48, fig. 9) oxygen-isotope data for 900,000 years to devise a sea-level curve (Zellmer, 1979,

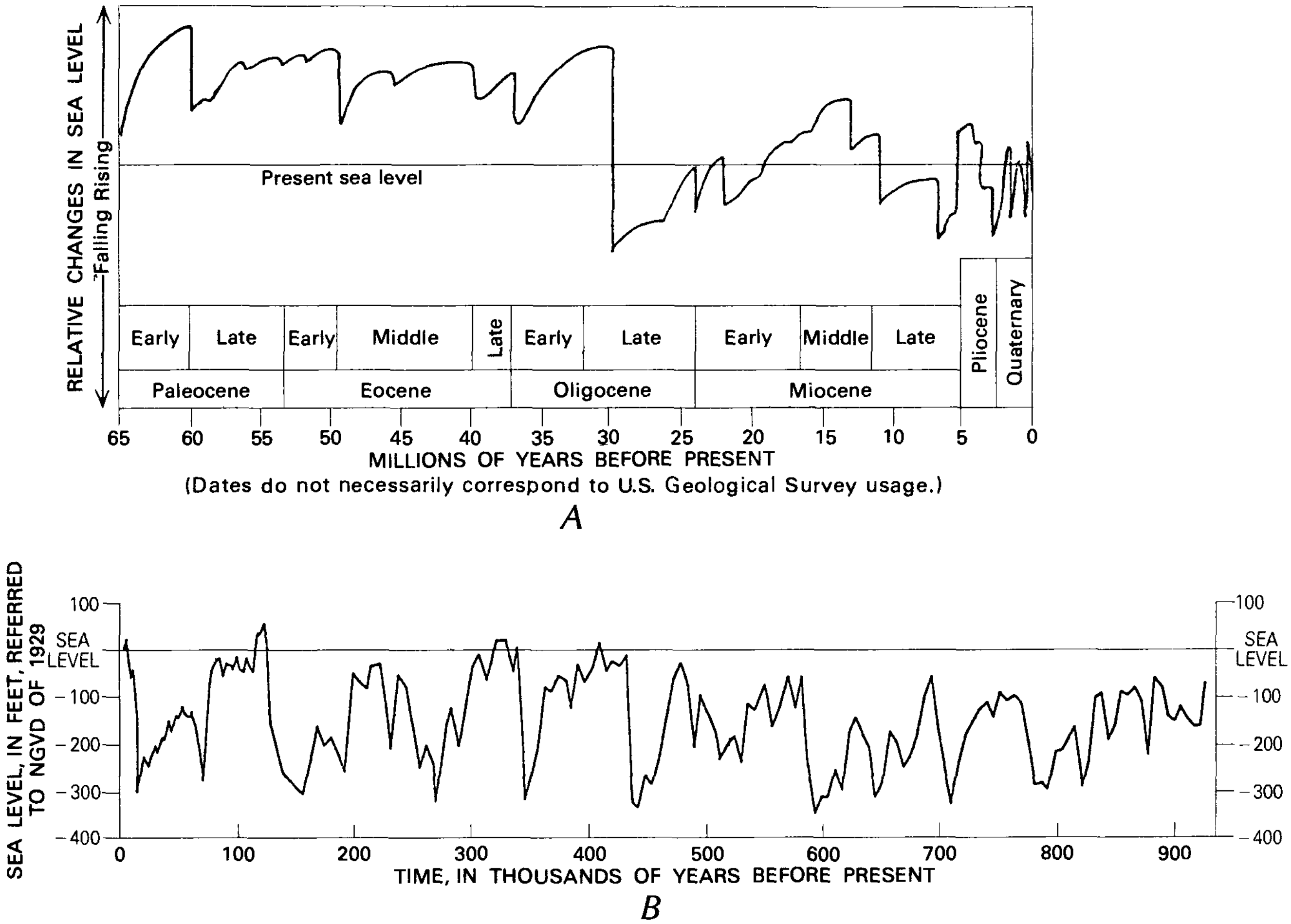


Figure 4. Sea-level curves. *A*, Modified from Vail, Mitchum, and Thompson, 1977, figure 3. *B*, Modified from Zellmer, 1979, figure 17.

p. 63, fig. 17) for the past 900,000 years. This curve (fig. 4*B*) indicates that eight large-scale sea-level fluctuations of more than 200 ft in amplitude took place during this 900,000-year period. An additional eight sea-level fluctuations had an amplitude on the order of 100 to 200 ft.

Zellmer's (1979, fig. 17) curve indicates that the average sea level for the last 900,000 years was approximately 150 ft below present sea level. However, Van Donk (1976) presented oxygen-isotope data showing that ocean-water temperatures were generally higher during the period 1 to 2.3 million years ago than during the past 1 million years. Hence, glacial-ice volumes were generally smaller and sea levels were generally higher during the earlier period. Prell (1980) estimated that the average sea level of the early Quaternary (older than 690,000 years) was about 115 ft higher than that of the late Quaternary. Therefore, the average Quaternary sea level was probably between 50 and 100 ft below present sea level.

COMPUTER SIMULATION

Several computer simulation approaches can be used to analyze the relation of saltwater to freshwater in Coastal Plain sediments. These approaches range in complexity from simulation of the transient movement of freshwater and saltwater with a model that couples ground-water flow and solute transport, to an interface-locating approach using a cross-sectional steady-state flow model. In this study, the latter approach was used because it avoids numerical problems commonly associated with transport simulations having large spatial and temporal domains.

Theoretical Background

The saltwater-freshwater transition zone is not simulated in the interface-locating approach. Instead, the ground-water system is assumed to consist of a steady-state freshwater flow field separated by a sharp interface from static saltwater that has the density of seawater. In reality, however, the presence of a transition zone implies that the saltwater is not static, as saltwater circulation is needed to maintain the transition zone (Cooper, 1964). Several investigators have examined the relation of mathematically computed sharp interfaces to both theoretical and observed dispersion zones. Cooper (1964, figs. 1 and 2) suggested that a computed sharp interface is located on the freshwater side of the zone. Henry (1964b) compared computed sharp interfaces with mathematically computed salt concentrations in an idealized seawater-freshwater system. Except near the model boundaries, the sharp interface tends to coincide with a concentration 0.4 times that of seawater (Henry, 1964b, p. C80, fig. 34B). Hence, the interface-locating approach provides a way of determining equilibrium positions of the sharp interface and, by inference, of the transition zone for a variety of sea-level conditions. It also demonstrates the sensitivity of the interface position to variations in hydraulic properties.

The sharp interface is located so that, at the interface, the pressure resulting from a column of freshwater is equal to the pressure exerted by a column of static saltwater (Hubbert, 1940),

$$P_s = P_f \tag{1}$$

Hydraulic head, h, is defined as

$$h = \frac{P}{\rho g} - z,$$

and saltwater and freshwater pressures are equated in terms of hydraulic head as

$$\rho_s g(h_s + z) = \rho_f g(h_f + z) \tag{2}$$

Because h_s is zero or mean sea level for static saltwater the following relationship is valid along the interface,

$$z = \frac{\rho_f}{\rho_s - \rho_f} \cdot h_f \tag{3}$$

where

ρ_f is density of freshwater,
ρ_s is density of seawater,
z is the depth of the sharp interface below the prevailing sea-level datum, and
h_f is the freshwater head measured at z.

The sharp interface between fresh and salty ground water is a free-surface boundary condition of the freshwater flow system. The free surface is the limiting streamline of the freshwater flow system and, therefore, constitutes a no-flow boundary with the salty water. As with other free surfaces, pressure conditions are known along the surface but the position of the surface is unknown. The purpose of the interface-locating approach is to improve successively an estimate of the position of the sharp interface until computed pressures along the simulated interface agree with the known pressure conditions. The final simulated position of the interface closely approximates the location of a theoretical sharp interface that separates freshwater and static saltwater.

The interface-locating approach is shown schematically in figure 5. A trial interface position, located more seaward and deeper than the final simulated interface, is specified as a no-flow boundary in the simulation of the flow system. Freshwater heads along the trial interface are computed by the flow model. Because both freshwater heads and the depth to the trial interface are known, equation 3 can be used to determine if the sharp interface is located in the correct position. If the depth z computed from equation 3 using the model-derived freshwater head is less than the depth of the trial interface, the position of the trial interface or no-flow boundary is moved to reduce the size of the freshwater flow system. This procedure is repeated until equation 3 is satisfied along the seaward and lower boundaries of the flow model and the final position of the sharp interface is determined.

This approach has been successfully applied in several ground-water studies. Bennett and Giusti (1971) used it with an analog model of a coastal aquifer near Ponce, P.R. More recently, Guswa and LeBlanc (1980) used the technique in an analysis of ground-water flow in the unconfined aquifer of Cape Cod, Mass. In the Cape Cod study, modifications to the digital flow model (Trescott, 1975) were made to allow the model to evaluate and reposition the trial interface during the simulation.

In the present study, the flow model did not automatically position the trial interface. Rather, model results were evaluated and a new trial position of the sharp interface or no-flow boundary was selected and then used in a subsequent simulation of the freshwater flow system.

Conceptualization and Simulation of the System

A cross-sectional finite-difference model was designed to simulate fresh ground-water flow in the Atlantic Coastal Plain in southern New Jersey and the adjacent Continental Shelf. The model, referred to as the INTERA model, is capable of simulating variable-density ground-water flow, solute transport, and thermal-energy

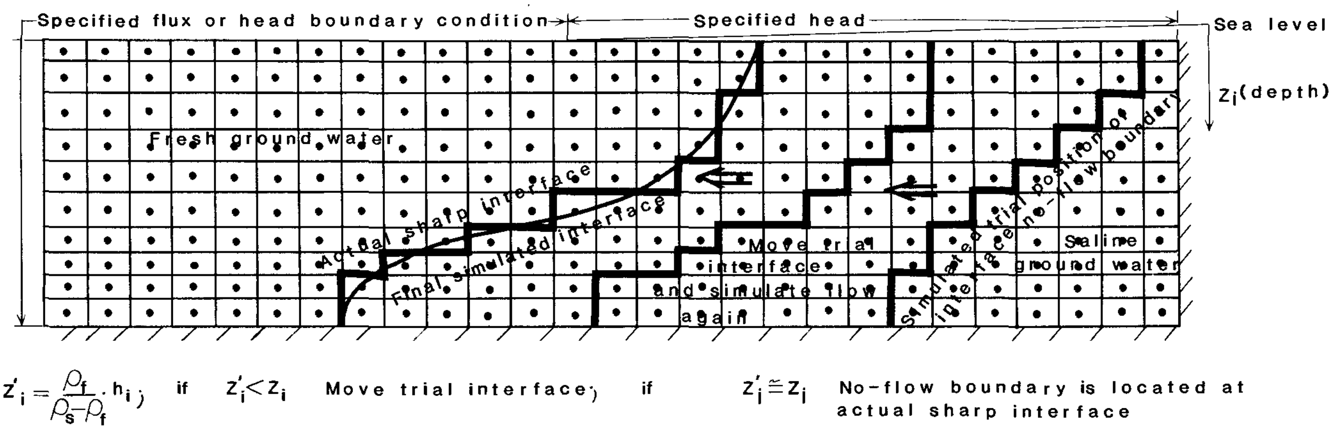


Figure 5. Generalized cross-sectional model showing the interface-locating approach.

transport (INTERCOMP, 1976, and INTERA, 1979). The INTERA model was used to simulate the freshwater flow system. It was chosen primarily because the shape of the hydrogeologic system, a seaward-thickening wedge, can be readily simulated by the model. Also, the input structure of the model permits the interface-locating approach to be applied easily.

The simulated hydrogeologic section, including boundary conditions, is shown in figure 3. The ground-water system consists of a wedge of sediments bounded on the west by the Delaware River estuary and on the east by the Continental Slope. The sediments are underlain by crystalline basement and consolidated rocks of Jurassic age. The wedge, approximately 150 miles long, thickens from 150 ft at the western boundary to more than 16,000 ft at the eastern boundary. The location of the hydrogeologic section was chosen to correspond to a flow line in the present (unstressed) flow system which was estimated from historic water-level data.

The simulated flow system is bounded by an impermeable or no-flow boundary at the base of the sedimentary wedge. At the surface and along the Delaware River estuary and Continental Slope, the system is bounded by specified hydraulic heads. Because the INTERA model can be used to simulate the flow of variable density fluids, the flow equation and boundary conditions are formulated in terms of pressure rather than hydraulic head. Hence, the specified head boundary condition in New Jersey is the elevation of the water table which, by definition, has a pressure of zero. Offshore, on the Continental Shelf and Slope the specified boundary condition is a saltwater head of zero which is designated as the hydrostatic pressure exerted on the system by a seawater column of the prevailing depth. Because the estuary in this cross section is considerably fresher than seawater, the hydrostatic pressure used as a boundary condition was computed for a freshwater column.

Analysis of geophysical logs from the COST B-2 test well (Woodruff, K. D., written commun., 1981) and chemical analyses of water from AMCOR test hole 6021 (fig. 1) show that the sediments contain saline water at the seaward boundary of the cross section. Use of this boundary as an initial sharp interface assures that the position of the final simulated interface will be located landward of the initial simulated interface. This makes the interface-locating procedure less cumbersome, because successive trial interfaces are located within the limits of the initial finite-difference grid.

The hydrogeologic section was discretized by a finite-difference mesh consisting of 82 columns and 10 rows. The vertical grid spacing is variable, ranging from 1 to 3,000 ft. The horizontal grid spacing is constant at 10,000 ft. Unlike many finite-difference models, the INTERA model has the capability of accommodating a wedge-shaped finite-difference grid. This is done by reducing the thickness and depth of finite-difference blocks in the updip direction. The wedge-shaped finite-difference grid provides hydraulic continuity along bedding planes and the appropriate simulation of aquifer outcrops across the top of the cross section.

Because of the paucity of hydrogeologic data on the Continental Shelf and Slope, the hydrogeologic framework of the section was simulated as a highly anisotropic porous medium rather than as a layered system of aquifers and intervening confining beds. The substitution of an equivalent anisotropic medium for a layered aquifer system is discussed by Bear (1979, p. 33).

Aquifers and confining beds are not identical to time-stratigraphic units. However, for the purposes of this model analysis and because of the scale of physical phenomena being simulated, the maximum and minimum values of the anisotropy are parallel and normal, respectively, to the contacts of the time-stratigraphic units shown in figure 3. Variable vertical (K_v) and lateral

(k_l) hydraulic conductivities and anisotropy ratios (K_l/k_v) were used in the simulations. The ranges of values simulated are 1 to 20 ft/d for lateral conductivity and 1×10^{-6} to 1×10^{-3} ft/d for vertical conductivity.

Discussion of Model Results

Three series of simulations were conducted to determine the position of the sharp interface for different sea levels, hydraulic conductivities, and anisotropy ratios. Table 1 summarizes the simulations performed. The simulation results show the sensitivity of the interface position to these parameters. They were used also for comparison with the observed saltwater distribution shown in figure 3.

The first series of simulations (1–8) tested the sensitivity of the position of the interface to hydraulic conductivities and anisotropy ratios. These simulations used present sea level (fig. 6). For simulations 1 to 4, the porous medium was considered homogeneous, as shown in table 1 and figure 6*A*. In simulation 1, the anisotropy ratio of the conductivities (K_l/K_v) is 10,000, which is at least an order of magnitude less than in simulations 3 and 4. The interface is located closer to the present shoreline and is deeper onshore than in other simulations. Simulation 2, using an anisotropy ratio of 50,000, has an interface that is shallower onshore and extends much farther offshore than the simulation 1 interface. Ratios of 100,000 and 200,000 (simulations 3 and 4) produce slightly shallower interfaces that extend farther offshore. Thus, increasing the anisotropy ratio has the effect of making the freshwater flow system shallower and extending it farther offshore.

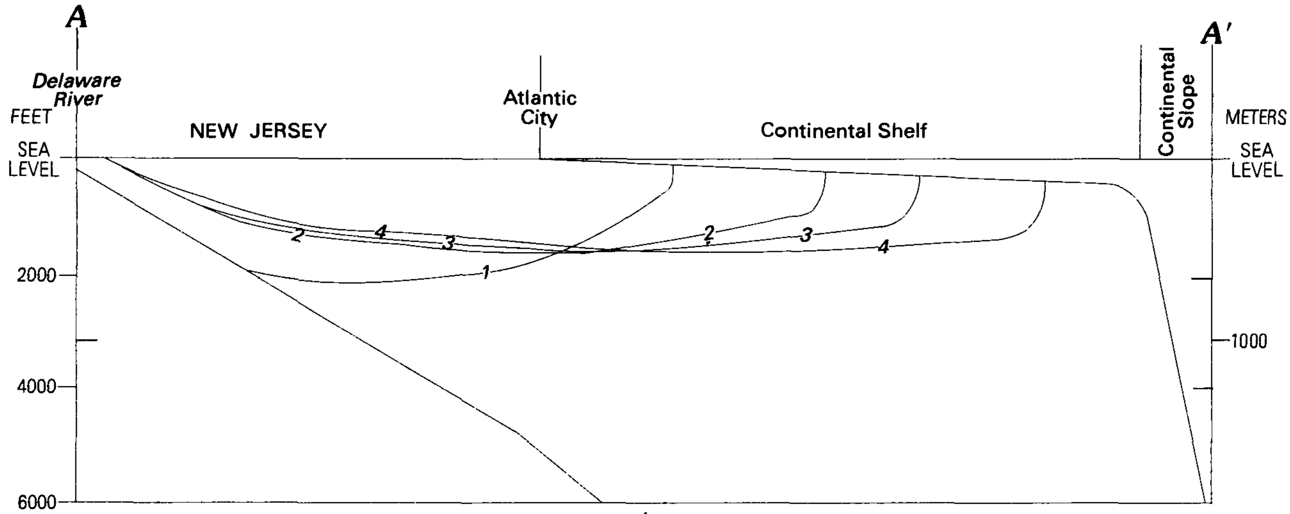
Simulations 5 and 6 tested the effects of a nonhomogeneous lateral hydraulic conductivity. Figure 6*B* shows that the decreased lateral hydraulic conductivities in zones B and C, and hence the decreased anisotropy ratio, cause the interface to be located closer to the shoreline than that of simulation 3.

Simulations 7 and 8, using the same distribution of lateral hydraulic conductivity as that of simulation 6, tested the effects of a nonhomogeneous vertical hydraulic conductivity. Vertical hydraulic conductivities used in

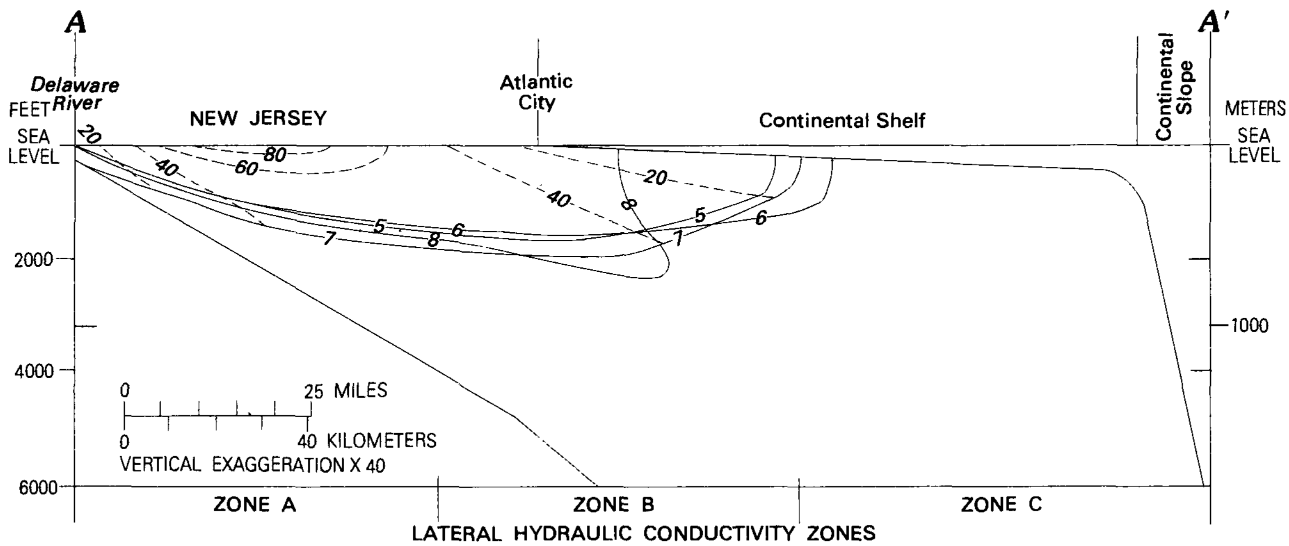
Table 1. Summary of simulations

Simulation no.	Sea level, in feet, relative to present sea level	Lateral hydraulic conductivity[1] K_l (ft/d)			Vertical hydraulic conductivity K_v (ft/d)	Anisotropy ratio K_l/K_v
		Zone A	Zone B	Zone C		
1	0	1	1	1	0.0001	10,000
2		5	5	5	.0001	50,000
3		10	10	10	.0001	100,000
4		20	20	20	.0001	200,000
5		10	3	1	.0001	100,000 to 10,000
6		10	5	2.5	.0001	100,000 to 25,000
7		10	5	2.5	.00033 to .000083	30,000
8		10	5	2.5	.001 to .000001	5,000,000 to 10,000
9	−50	1	1	1	.0001	10,000
10		5	5	5	.0001	50,000
11		10	10	10	.0001	100,000
12		20	20	20	.0001	200,000
13		10	3	1	.0001	100,000 to 10,000
14		10	5	2.5	.0001	100,000 to 25,000
15		10	5	2.5	.00033 to .000083	30,000
16		10	5	2.5	.001 to .000001	5,000,000 to 10,000
17	−100	10	5	2.5	.0001	100,000 to 25,000
18	−100	10	5	2.5	.00033 to .000083	30,000
19	−100	10	5	2.5	.001 to .000001	5,000,000 to 10,000
20	−150	10	5	2.5	.0001	100,000 to 25,000
21	−150	10	5	2.5	.00033 to .000083	30,000

[1]Zones shown in figures 6, 7, 8, and 9.



A



B

EXPLANATION

—*5*— SIMULATED INTERFACE—Number is simulation number

- -*20*- -LINE OF EQUAL POTENTIOMETRIC HEAD FOR SIMULATION 7—Interval 20 feet

A——*A'* LOCATION OF SECTION SHOWN ON FIGURE 1

Simulation number	Lateral hydraulic conductivity[1] K_1 (ft/d)			Vertical hydraulic conductivity K_v (ft/d)	Anisotropy ratio K_1/K_v
	ZONE A	ZONE B	ZONE C		
1	1	1	1	0.0001	10,000
2	5	5	5	.0001	50,000
3	10	10	10	.0001	100,000
4	20	20	20	.0001	200,000
5	10	3	1	.0001	100,000 to 10,000
6	10	5	2.5	.0001	100,000 to 25,000
7	10	5	2.5	.00033 to .000083	30,000
8	10	5	2.5	.001 to .000001	5,000,000 to 10,000

C

Figure 6. Position of sharp interfaces for present sea level. *A*, Simulations 1–4. *B*, Simulations 5–8. *C*, Summary of simulations 1–8.

simulation 7 were selected so that the anisotropy ratio was kept constant. Figure 6*B* shows that the interface is several hundred feet deeper onshore and is located a few miles closer to the shoreline than that of simulation 6. The distribution of vertical hydraulic conductivities used in simulation 8 is based on both age and depth of the sediment. For example, values used for upper Miocene and post-Miocene sediments range from 1×10^{-3} ft/d for depths shallower than 500 ft to 1×10^{-4} ft/d for depths greater than 1,000 ft. Values used for Cretaceous sediments range from 1×10^{-4} ft/d for depths less than 1,000 ft to 1×10^{-6} ft/d for depths greater than 5,000 ft. Figure *6B* shows that the interface is slightly deeper onshore and is located much closer to the shoreline than that of simulation 6. Also, the interface curves back towards the shore, forming a wedge of freshwater overlain by saltwater. During the interface-locating process, the part of the interface that separates freshwater from the overlying saltwater was not treated as a no-flow boundary. Instead, a freshwater density was specified in the freshwater wedge and a seawater density was specified in the overlying saltwater.

Freshwater potentials for simulation 7 are also shown in figure *6B*. Recharge takes place onshore at a ground-water high between the Delaware estuary and the coast. Discharge is to the ocean and the estuary. The maximum simulated ground-water recharge to the freshwater system is 0.43 in./yr.

The distribution of lateral hydraulic conductivity used in simulations 6 to 8, although generalized, is probably the most realistic representation of the inhomogeneity of the system. The value of 2.5 ft/d for the offshore section (Zone C) is based partly on lateral intrinsic permeability analyses of cores taken at the COST B-2 well (Scholle, 1980, p. 81). The onshore (Zone A) lateral conductivity of 10 ft/d is a reasonable representative value for the sediments of the Coastal Plain of New Jersey. Reported values of lateral hydraulic conductivity for the Coastal Plain aquifers are generally higher (Nichols, 1977, and Rhodehamel, 1973). However, the ground-water flow system is simulated as a highly anisotropic porous medium rather than as a series of interbedded aquifers and confining beds having a wide range in lateral hydraulic conductivities (Bear, 1979, p. 33). Hence, the lateral hydraulic conductivity used in the simulations must be less than the reported lateral hydraulic conductivity of the aquifers in order to attain the appropriate hydraulic conductivity and ground-water flow from each stratigraphic unit. The value of 5 ft/d used in zone B is the geometric mean of the values used in zones A and C.

The vertical hydraulic conductivity values of 0.0001 ft/d, used in most simulations, and 0.00033 used for zone A in simulation 7, are typical of Coastal-Plain confining beds. However, in simulation 7 the vertical hydraulic conductivity was reduced in zones B and C in order to maintain a constant anisotropy ratio of 30,000. The distribution of vertical hydraulic conductivity used in simulation 8 is modified from that compiled by M.M. Martin (U.S. Geological Survey, written commun., 1982).

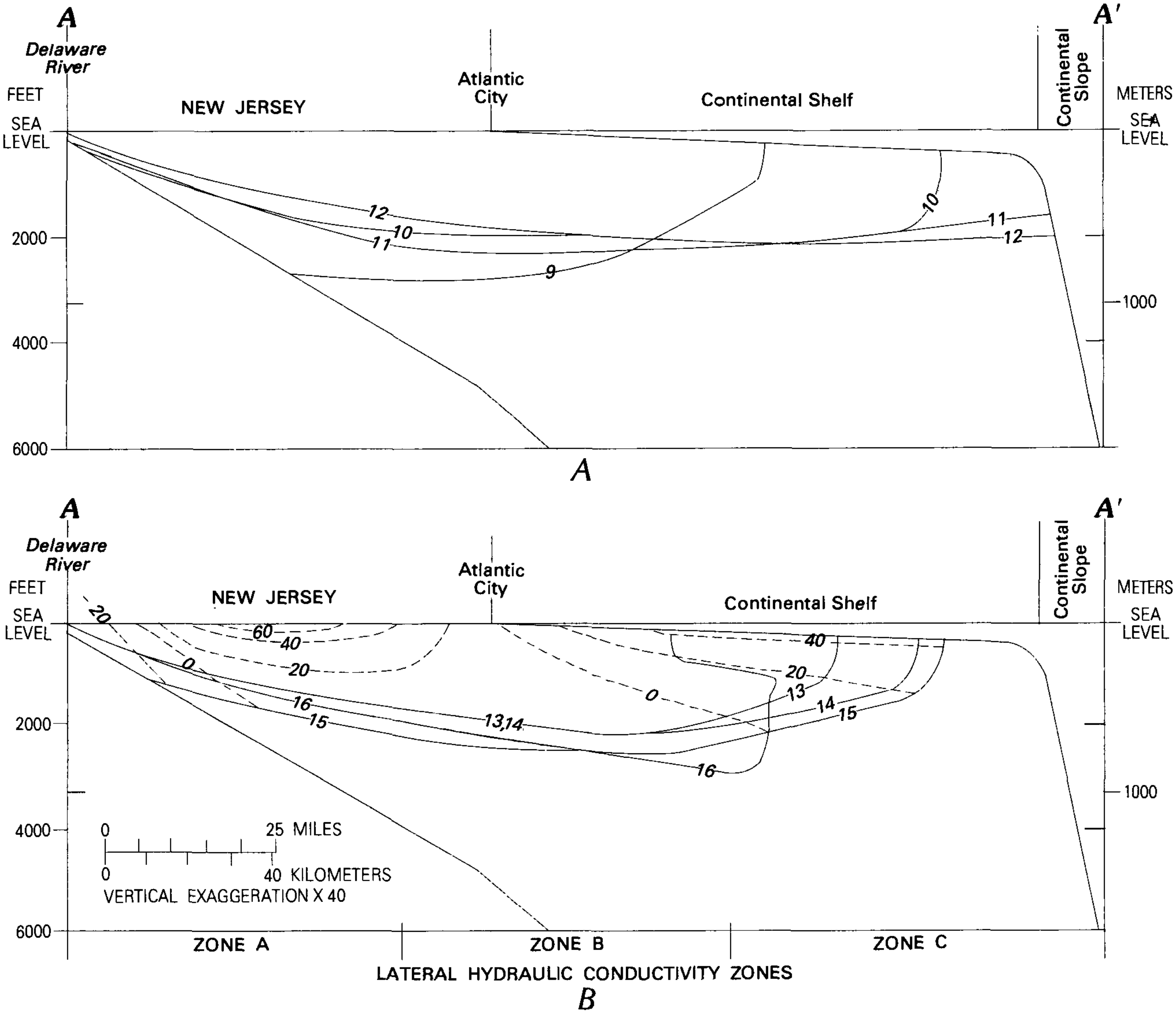
The second series of simulations (9–16, table 1) tested the sensitivity of the interface position to hydraulic conductivities and anisotropy ratios. These simulations used a sea level of 50 ft below present sea level (fig. 7). This sea-level position approximates the upper limit of the average sea level for the Quaternary. The simulated coast for this sea level is 17 miles seaward of the present coast. The ground-water divide between the Delaware River estuary and the Atlantic Ocean was shifted seaward approximately 5 miles for these simulations. This was done in order to maintain both recharge rates and water-table gradients that are similar to those of simulations 1–8.

The effects on the interface position caused by varying the anisotropy ratio and lateral hydraulic conductivity are similar to those obtained in the first series of simulations (1–8) that used present sea level; that is, increasing the anisotropy ratio caused the freshwater flow system to be shallower and to extend farther offshore. Compared with simulations 1–8, the freshwater flow in simulations 9–16 generally extends several hundred feet deeper and about 15 to 30 miles farther from the present coast.

Freshwater potentials for simulation 15, which used the most realistic distribution of lateral hydraulic conductivities and a uniform anisotropy ratio of 30,000, are shown on figure *7B*. As with simulation 7 (fig. *6B*), recharge occurs onshore between the present coast and the Delaware estuary. Maximum simulated recharge is 0.46 in./yr. The area of recharge is slightly larger in simulation 15 than in simulation 7.

The third series of simulations (17–21, table 1) tested the effects of nonhomogeneous lateral and vertical hydraulic conductivities. These simulations used sea levels of 100 and 150 ft below present sea level (fig. 8). The 150 ft below present sea level approximates the average sea level for the past 900,000 years. The 100 ft below present sea level is the estimated lower limit of the average Quaternary sea level. For all simulations except simulation 19, freshwater discharges through the face of the Continental Slope. Simulations 18 and 21, which used an anisotropy ratio of 30,000, produced significantly deeper interfaces than did simulations 17 and 20, which used anisotropy ratios of 100,000 to 25,000. Comparison of the results of simulation 19 (fig. 8) with those of simulations 8 (fig. *6B*) and 16 (fig. *7B*), using the identical hydraulic properties, shows that these interfaces have similar shapes, with saltwater overlying a freshwater wedge. However, lowering sea level caused the interface to be deeper and to extend farther offshore.

The increase in interface depth and distance offshore with decline in sea level is shown also in figure 9. Interfaces depicted are the results of simulations 7, 15, 18, and 21, which used the most realistic lateral hydraulic conductivity distribution and a constant anisotropy ratio of 30,000. The exceptionally deep onshore interface for simulation 21 is related to the conceptualization of the



EXPLANATION

—11— SIMULATED INTERFACE—Number is simulation number

- -40- - LINE OF EQUAL POTENTIOMETRIC HEAD FOR SIMULATION 15—Interval 20 feet

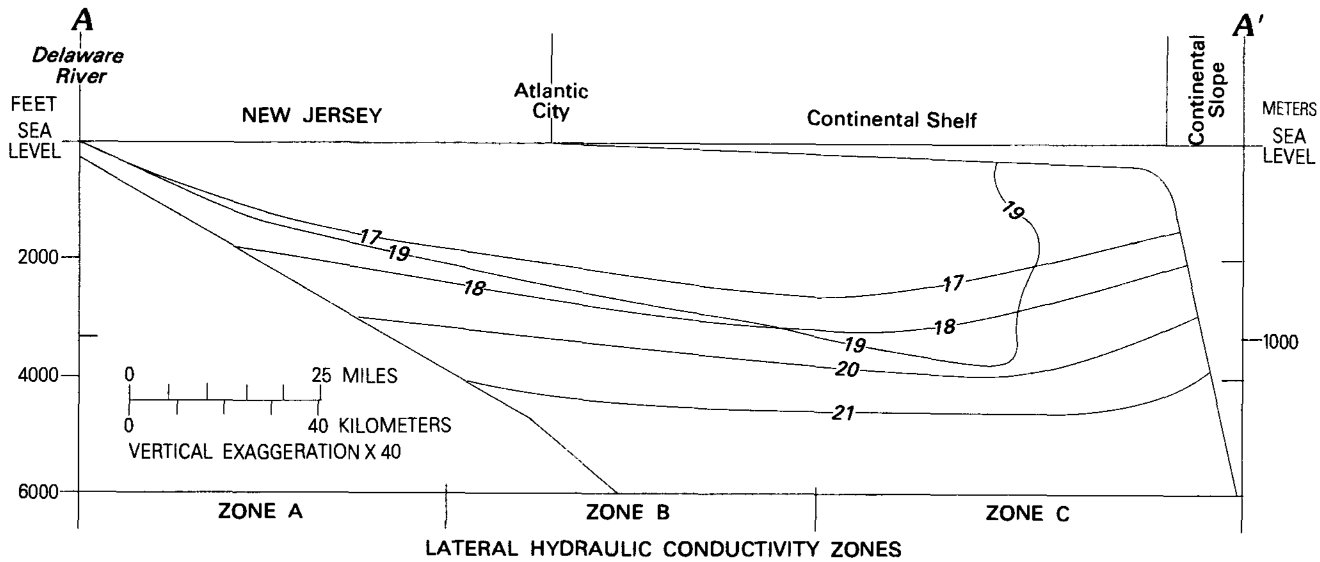
A——A' LOCATION OF SECTION SHOWN ON FIGURE 1

Simulation number	Lateral hydraulic conductivity[1] K_1 (ft/d) ZONE A	ZONE B	ZONE C	Vertical hydraulic conductivity K_v (ft/d)	Anisotropy ratio K_1/K_v
9	1	1	1	.0001	10,000
10	5	5	5	.0001	50,000
11	10	10	10	.0001	100,000
12	20	20	20	.0001	200,000
13	10	3	1	.0001	100,000 to 10,000
14	10	5	2.5	.0001	100,000 to 25,000
15	10	5	2.5	.00033 to .000083	30,000
16	10	5	2.5	.001 to .000001	5,000,000 to 10,000

C

Figure 7. Position of sharp interfaces for a sea level 50 ft below present sea level. *A*, Simulations 9–12. *B*, Simulations 13–16. *C*, Summary of simulations 9–16.

Delaware River estuary in the model. For most simulations, the stage of the Delaware is identical to sea level. However, based on the depth of the channel below the Delaware, it is likely that the stage during the Quaternary did not fall below 100 ft below present sea level. Thus, in simulation 21, which used a sea level of 150 ft below present sea level, the stage in the Delaware River was constrained to remain at 100 ft below present sea level. This



A

EXPLANATION

—21— SIMULATED INTERFACE—Number is simulation number

A——A' LOCATION OF SECTION SHOWN ON FIGURE 1

Simulation number	Sea level, in feet relative to present sea level	Lateral hydraulic conductivity[1] K_1 (ft/d)			Vertical hydraulic conductivity K_v (ft/d)	Anisotropy ratio K_1/K_v
		ZONE A	ZONE B	ZONE C		
17	−100	10	5	2.5	.0001	100,000 to 25,000
18	−100	10	5	2.5	.00033 to .000083	30,000
19	−100	10	5	2.5	.001 to .000001	5,000,000 to 10,000
20	−150	10	5	2.5	.0001	100,000 to 25,000
21	−150	10	5	2.5	.00033 to .000083	30,000

B

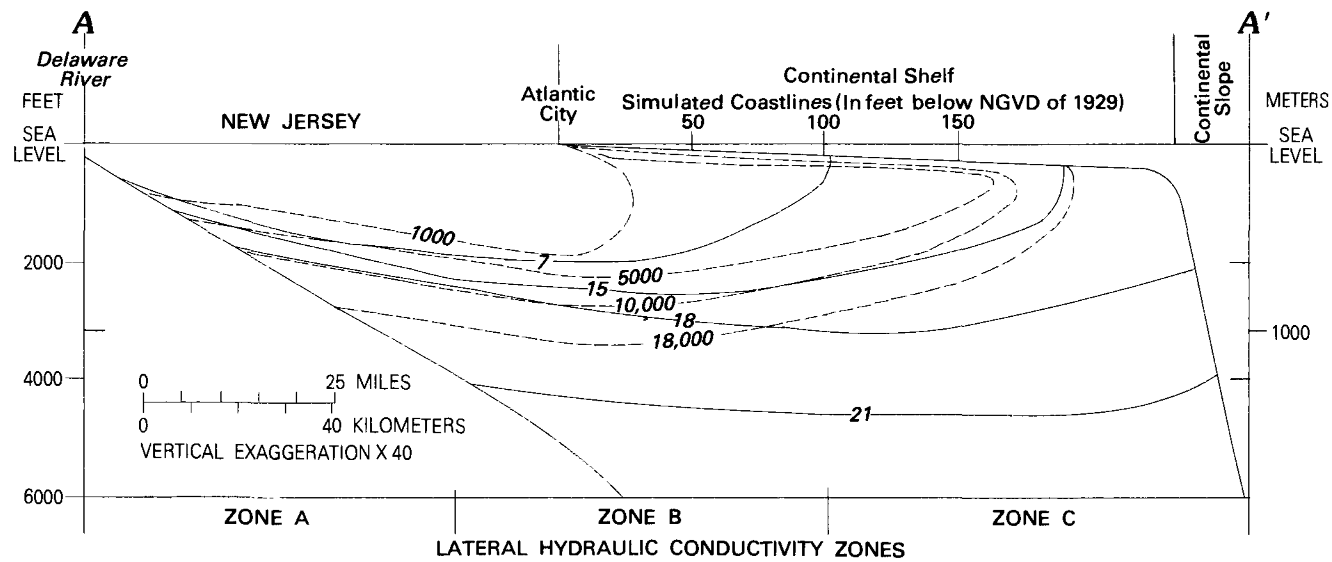
Figure 8. Position of sharp interfaces for sea levels 100 and 150 ft below present sea level. *A*, Simulations 17–21. *B*, Summary of simulations 17–21.

resulted in the interface being positioned considerably deeper than in the other simulations.

Comparison of Model Results with Observed Distribution of Saltwater and Freshwater

The observed distribution of saltwater and freshwater was compared with several simulated interfaces. Comparison of the isochlors in figure 9 with the sharp interfaces computed for present sea level in figure 6 shows that the farthest offshore position of the saltwater-freshwater transition zone is enclosed by the interfaces of simulations 3 and 4. Significantly the center of the transition zone, represented by the 10,000 mg/L isochlor, extends 22 to 51 miles seaward beyond the interfaces that were simulated using the most probable estimates of lateral hydraulic conductivities (simulations 6, 7, and 8). Onshore, the interfaces of simulations 2, 3, 4, 5, 6, and 8 coincide approximately with the 1,000 mg/L isochlor. The interface of simulation 7 more nearly coincides with the 5,000 mg/L isochlor. These computer results strongly suggest that the observed saltwater-freshwater transition zone is not in equilibrium with present sea-level conditions, but reflects lower sea levels.

A comparison of the isochlors with interfaces computed for different sea levels is shown in figure 9. The most realistic distribution of lateral hydraulic conductivities and a constant anisotropic ratio of 30,000 were used. The offshore position of the center of the transition zone (10,000 mg/L isochlor) coincides approximately with the interface of simulation 15, computed for a sea level of 50 ft below present sea level. The sharp interfaces computed for the other sea levels are located either 26 miles landward of the center of the transition zone or from 20 to 25 miles seaward of the center. Onshore, the sharp interface computed for a sea level of 50 ft below present sea level lies between the 5,000 mg/L and the 10,000 mg/L isochlors. The interface computed for a sea level of 100 ft



A

EXPLANATION

—21— SIMULATED INTERFACE—Number is simulation number

--1000-- LINE OF EQUAL CHLORIDE CONCENTRATION. (milligrams per liter)

A——A′ LOCATION OF SECTION SHOWN ON FIGURE 1

Simulation number	Sea level, in feet relative to present sea level	Lateral hydraulic conductivity[1] K_1 (ft/d)			Vertical hydraulic conductivity K_v (ft/d)	Anisotropy ratio K_1/K_v
		ZONE A	ZONE B	ZONE C		
7	0	10	5	2.5	.00033 to .000083	30,000
15	−50	10	5	2.5	.00033 to .000083	30,000
18	−100	10	5	2.5	.00033 to .000083	30,000
21	−150	10	5	2.5	.00033 to .000083	30,000

B

Figure 9. Position of sharp interfaces for sea levels 0, 50, 100, and 150 ft below present sea level, compared with observed distribution of saltwater. *A*, Simulations 7, 15, 18, and 21. *B*, Summary of simulations 7, 15, 18, and 21.

below present sea level coincides with the 10,000 mg/L isochlor. These computer results suggest that the observed saltwater-freshwater transition zone reflects sea levels that were 50 to possibly 100 ft below present sea level.

Because of frequent sea-level fluctuations, it is unlikely that the zone of transition has been in equilibrium during the past 900,000 years. The transition zone presently may be moving landward and upward in response to the most recent sea-level rise of 300 ft during the past 16,000 years. As the interface-locating approach only considers steady-state conditions, it is impossible to use this approach to predict the time required for the system to adjust to this new equilibrium. The rate of movement of the saltwater-freshwater interface, however, can be estimated from the ground-water flow velocities at the interface if it is assumed that the interface moves at the same speed as the water.

In order to do this, the sharp interface (no-flow boundary) that had been determined using the interface-locating method for a sea level of 50 ft below present sea level (simulation 14) was removed from the model and replaced by waters of different densities. A freshwater density was specified on the landward side of the interface and a seawater density on the seaward side. Hence, the model included both freshwater and saltwater systems. Also, pressures along the eastern boundary of the model, beneath the Continental Slope, were defined as the hydrostatic pressures exerted by a column of seawater. The freshwater flow system that was simulated using the two densities was identical to the flow system simulated using the interface-locating approach; the saltwater system was essentially static. Boundary pressures were then modified to simulate an instantaneous 50 ft sea-level rise and the transient response of the system was observed. The particle velocity or estimated speed of the offshore saltwater-freshwater interface was computed from the simulated hydraulic gradients, using a porosity of 0.2, a lateral hydraulic conductivity of 2.5 ft/d, and a vertical hydraulic conductivity of 0.0001 ft/d at the simulated interface. The computed lateral particle velocity at the moving interface suggests that the offshore interface is moving landward at a very slow rate, about 0.2 miles per 10,000 years.

Development of the Saltwater-Freshwater Transition Zone

The development of saltwater-freshwater transition zones has been attributed to the flow of salty water into the freshwater flow system (Cooper, 1964). For example, in the Biscayne aquifer near Miami, Fla., saltwater flow and the resulting transition zone have been attributed to variations in recharge on the freshwater part of the system (Kohout, 1964) and to tidal fluctuations in sea level (Cooper, 1964, and Henry, 1964a). The transition zone there is less than 100 ft thick vertically and is about 2,000 ft wide at its seaward limit (Kohout, 1964, fig. 8). In southern New Jersey, the transition zone is about 1,500 ft thick vertically, and has a width (from the 5,000 mg/L isochlor to the 18,000 mg/L isochlor) of about 10 to 15 miles at its seaward limit. Consequently, a much larger scale mechanism than tidal fluctuations is needed to account for the much broader transition zone.

Large-scale eustatic sea-level fluctuations and the resulting back-and-forth movement of the shoreline provide a mechanism for the circulation of saltwater and the development and maintenance of a transition zone. The potential magnitude of the effect of sea-level changes on the system has been shown by the steady-state interface simulations (fig. 9). Because of the slow response of the interface to sea-level change, the individual sea-level fluctuations of the past 900,000 years (fig. 4*B*) may have been of insufficient duration to develop the broad transition zone. The interface may have moved laterally no more than a mile or two in response to each of these fluctuations. Longer periods, during which the interface could move laterally much longer distances and hence develop the broad transition zone, occurred during the late Tertiary and Quaternary (fig. 4*A*). Initial emplacement of saltwater in Miocene sediments took place during the middle Miocene when sea levels were much higher than at present. Eastward movement of the interface predominated during the late Miocene, followed by a predominant westward movement during the Pliocene. Sea levels were generally lower during the Quaternary than the Pliocene and hence the predominant movement of the interface was again eastward.

The isochlors in figure 9 show the presence of a thin transition zone (less than 250 ft) from seawater, at the ocean floor, to underlying fresher water (5,000 mg/L isochlor). The sharp interface model, however, was not designed to simulate an interface between saltwater and underlying freshwater. Rather, it simulates upward discharge of freshwater into the ocean. The observed distribution of saltwater in this zone may be the result of short-term cyclic fluctuations of sea level caused by storms and tides. The amplitude of these fluctuations is small compared with the eustatic sea level changes discussed earlier. However, at the sea-floor discharge boundary, where upward head gradients are small, these sea-level fluctuations can temporarily reverse the hydraulic gradient and move saltwater into the aquifer system. As suggested by Henry (1964a, p. C61), a transition zone would be enclosed by the hypothetical sharp interfaces resulting from the extremes of these small-scale sea-level fluctuations.

Several authors have investigated the effect of confining beds on the offshore occurrence of freshwater beneath saltwater. Mualem and Bear (1974) have demonstrated analytically and with a Hele-Shaw model that, in equilibrium, freshwater may discharge upward through a leaky confining bed into overlying salty ground water. Ten Hoorn (1981) expanded on this work by including vertical and horizontal flow in both aquifers and confining beds, in order to explain the occurrence of saltwater overlying freshwater along the coast of Holland.

In order to examine the effect of the thin zone of transition on underlying freshwater flow, an additional simulation was made. In this simulation (a modification of simulation 14), water having a density of seawater was specified at the model nodes corresponding to the observed zone of transition below the ocean floor. The resulting freshwater flow system matches, with only minor differences, the flow system simulated without the overlying saltwater. The simulation shows also that freshwater discharges to the ocean through the overlying saltwater. Although the system was modeled as an anisotropic medium, the results are in agreement with the theoretical distribution of saltwater and freshwater in the layered aquifer system analyzed by Mualem and Bear (1974).

GEOCHEMISTRY

Development of a broad saltwater-freshwater transition zone in the northern Atlantic Coastal Plain (depicted in fig. 3) is interpreted, in this report, to be caused by saltwater circulation responding to eustatic sea-level fluctuations. As sea level rises, saltwater invades the sediments and mixes with fresher water. As sea level declines, the fresher water advances seaward and the process of mixing continues. Repeated advance and retreat of the saltwater has thus produced a broad zone of mixed waters in which saltwater predominates in the deeper and seaward parts, and freshwater predominates in the shallower and landward parts.

Chemical analyses of water samples from the transition zone in the northern Atlantic Coastal Plain (table 2) were studied in order to: (1) describe the chemical character of the water; (2) determine if the water chemistry is consistent with the interpretation that the transition zone was produced by the mixing of saltwater and freshwater; and (3) determine if other geochemical

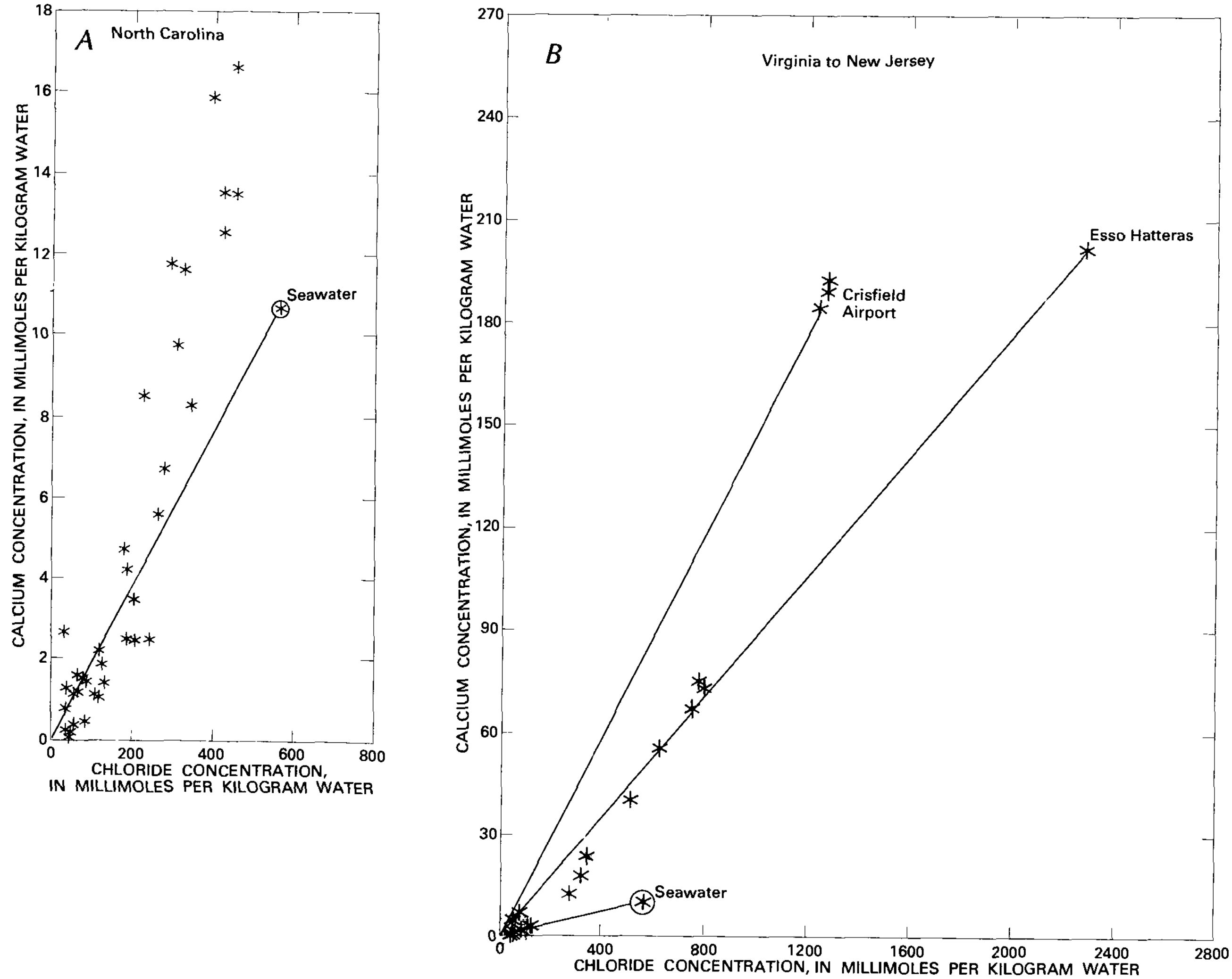


Figure 10. Relation of calcium concentration to chloride concentration. Lines denote conservative mixing curves. *A*, North Carolina. *B*, Virginia to New Jersey.

processes are occurring. Concentrations shown in table 2 and figures 10–17 are expressed as millimoles per kilogram (mmol/kg) of water rather than milligrams per liter of solution in order to facilitate the comparison of constituents that have different molecular weights.

Description of the Transition Zone

The distribution of several major chemical constituents within the saltwater-freshwater transition zone is presented in a series of graphs showing the relation of ion concentration to chloride concentration (figs. 10–15). Chemical data from North Carolina are presented separately from those from Virginia to New Jersey because the chemistry of the transition zone in the two areas differs substantially. The graphs indicate that ion concentrations of calcium, magnesium, sodium, potassium, and sulfate generally increase as chloride concentration increases (figs. 10–14). Bicarbonate tends to decrease as chloride concentration increases (fig. 15).

The order of abundance of cations from Virginia to New Jersey is generally Na > Ca > Mg > K. In North Carolina, the order is generally Na > Mg > Ca > K where chloride exceeds 250 mmol/kg. Potassium commonly exceeds calcium at lower chloride concentrations.

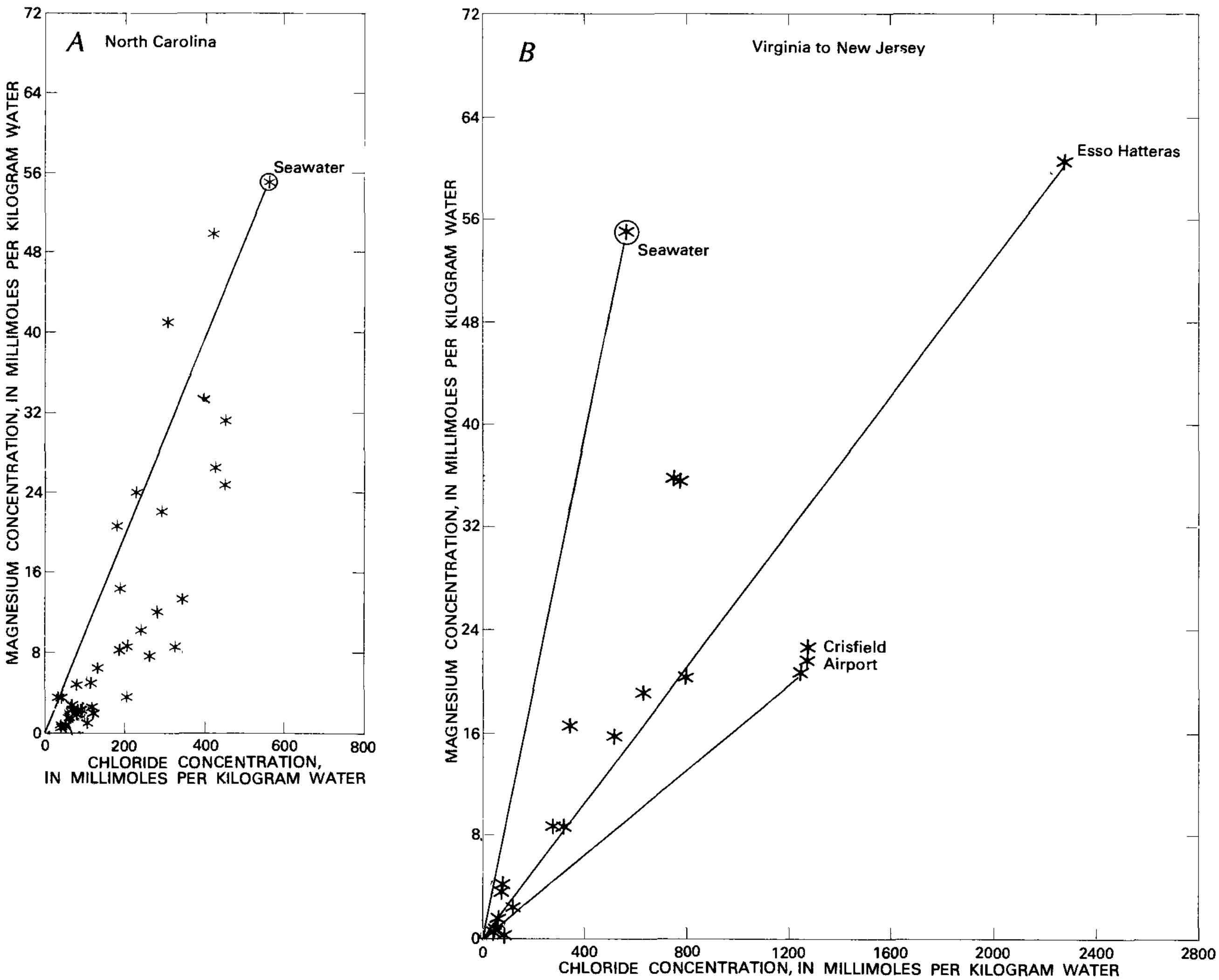


Figure 11. Relation of magnesium concentration to chloride concentration. Lines denote conservative mixing curves. *A*, North Carolina. *B*, Virginia to New Jersey.

Chloride is the most abundant anion. From Virginia to New Jersey, sulfate generally exceeds bicarbonate, where chloride is greater than 700 mmol/kg. In North Carolina, sulfate generally exceeds bicarbonate where chloride is greater than 250 mmol/kg. In the less mineralized waters, bicarbonate generally exceeds sulfate.

Interpretation of Variation in Ion Concentration

The graphs showing the relation of concentrations of selected ions to chloride concentrations can be used to determine if the water chemistry is consistent with the interpretation that the transition zone was produced by mixing of saltwater and freshwater. If the transition zone was produced by mixing, the distribution of ions in the zone can be described by simple mixing theory. When two waters of known chemical composition are mixed in various proportions, in the absence of chemical reactions, the chemical composition of the resulting waters can be predicted precisely. If ion concentrations of several such mixed waters are plotted against chloride on an arithmetic scale, a linear plot results. Hence, concentrations of individual ions from the transition zone plotted against chloride should fall on a straight line, or

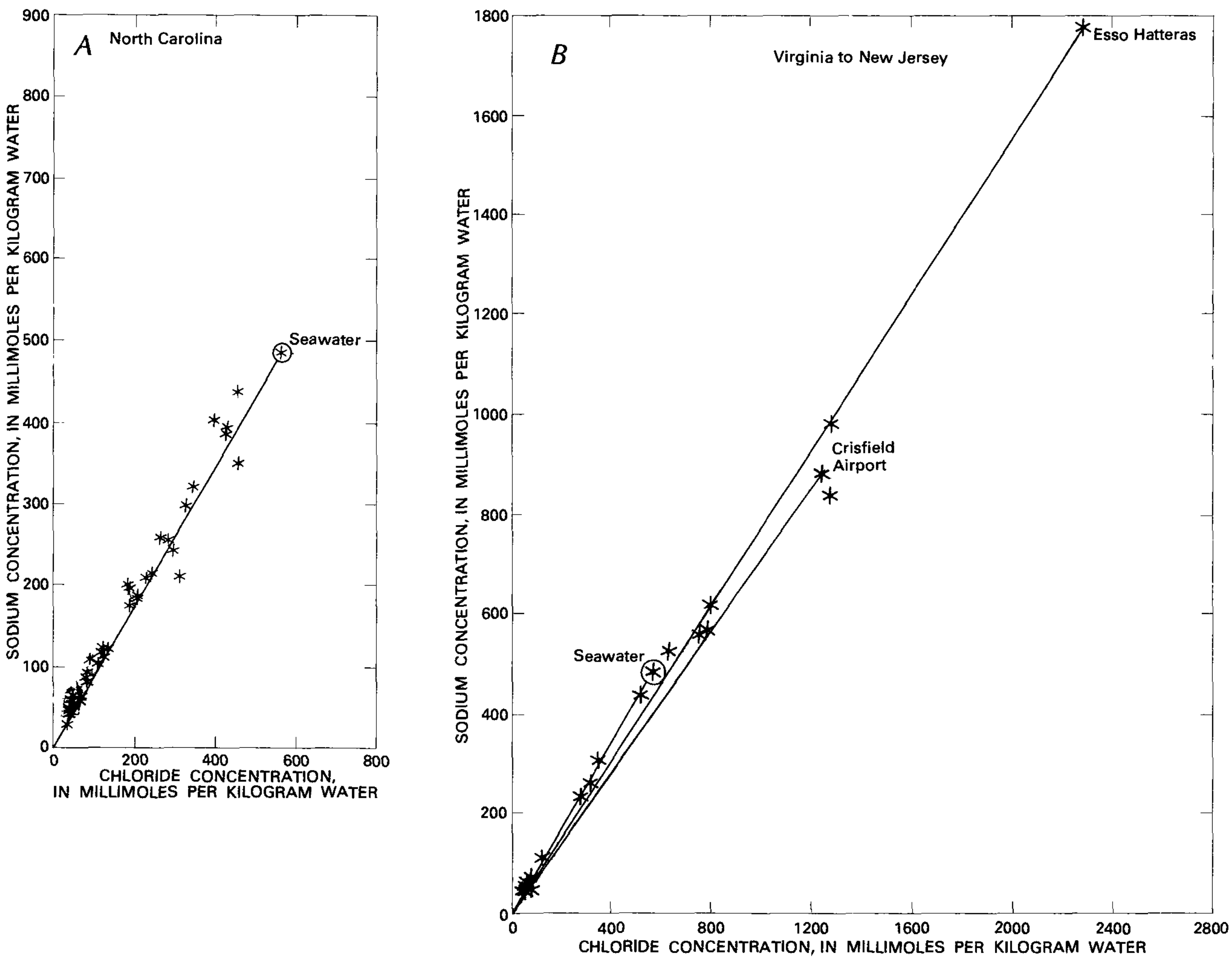


Figure 12. Relation of sodium concentration to chloride concentration. Lines denote conservative mixing curves. *A*, North Carolina. *B*, Virginia to New Jersey.

conservative mixing line. Any deviation from the mixing line suggests that other chemical processes are occurring. In this report, chloride is assumed to be a conservative ion. Hence the degree of mixing between two end members can be represented by the chloride concentrations of the sampled waters.

North Carolina

Plots of concentrations of sodium, potassium, and, to a lesser extent, sulfate from North Carolina (except for the Esso Hatteras well to be discussed subsequently) generally follow conservative mixing curves of seawater and freshwater shown on figures 12*A*, 13*A*, and 14*A*. Calcium concentrations (fig. 10*A*), in contrast, are significantly higher than the conservative mixing values for those analyses in which chloride exceeds 250 mmol/kg and slightly lower for those analyses in which chloride is less than 150 mmol/kg. Magnesium concentrations (fig. 11*A*) tend to be lower than the conservative mixing values throughout the chloride range shown. Concentrations of calcium plus magnesium (fig. 16*A*) follow the mixing curve much more closely than do either calcium or magnesium concentrations. However, concentrations of calcium plus magnesium tend to be lower than the con-

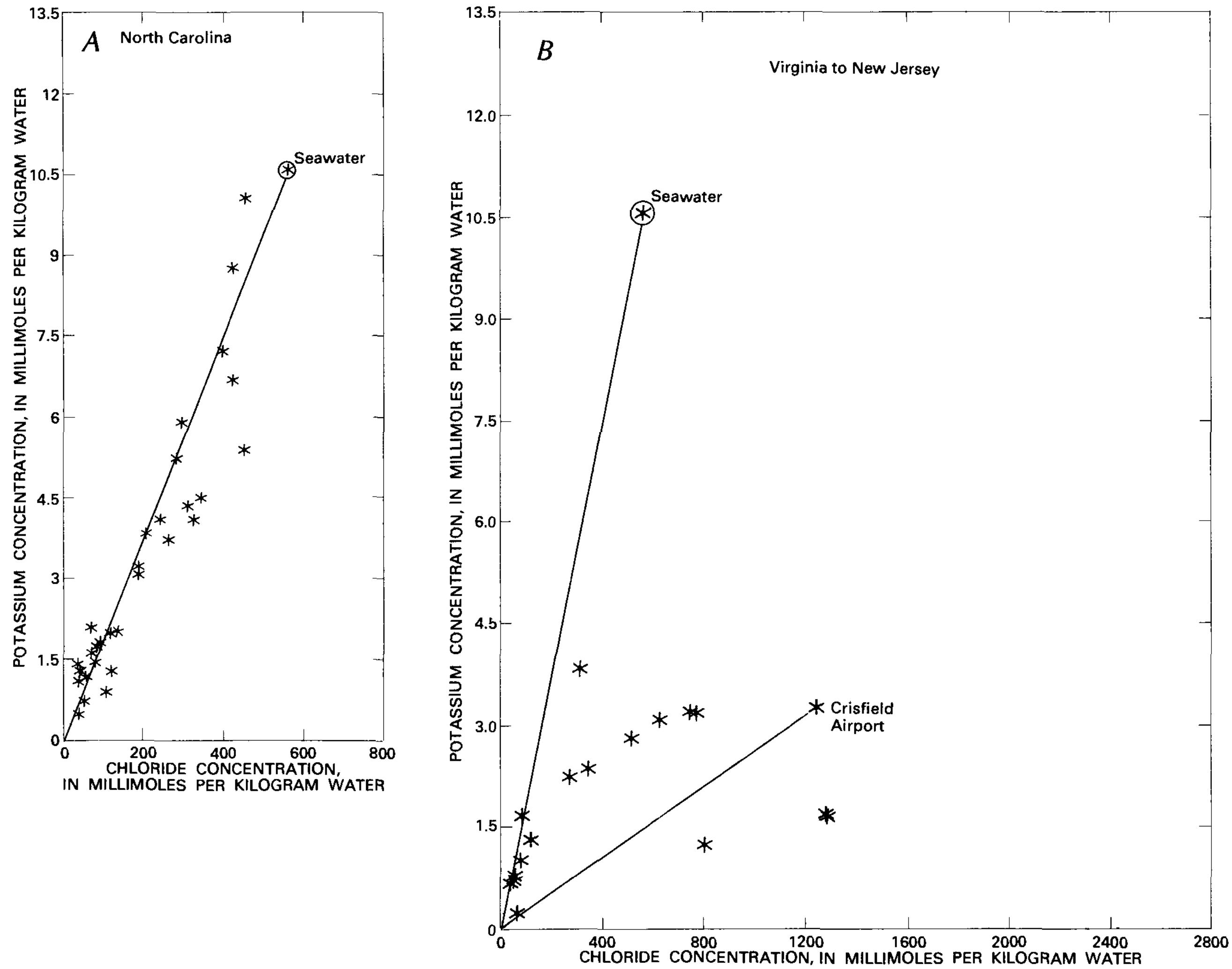


Figure 13. Relation of potassium concentration to chloride concentration. Lines denote conservative mixing curves. *A*, North Carolina. *B*, Virginia to New Jersey.

servative mixing values for those analyses in which chloride is less than 300 mmol/kg.

These comparisions of ion concentrations with the conservative mixing curves suggest that the saltwater-freshwater transition zone in North Carolina is primarily a mixture of seawater and freshwater. The freshwater appears to be a sodium bicarbonate water. This is suggested by the trends in calcium, magnesium, sodium, and bicarbonate concentrations on the dilute side of figures 10*A*, 11*A*, 12*A*, and 15*A*. Numerous reported freshwater analyses (not included in table 2) have sodium and bicarbonate concentrations ranging up to 20 and 12 mmol/kg, respectively. Calcium and magnesium concentrations in the same analyses generally range up to 1 and 2.5 mmol/kg, respectively.

The most significant deviations from the mixing lines—the higher concentrations of calcium and lower concentrations of magnesium—may be the result of ion exchange of magnesium for calcium on exchange sites. However, dolomitization of calcite could also account for a loss of magnesium from solution accompanied by a gain in calcium. The lower concentrations of calcium plus magnesium for those analyses in which chloride concentrations are less than 300 mmol/kg may be the result of ion exchange of calcium and magnesium for sodium on exchange sites.

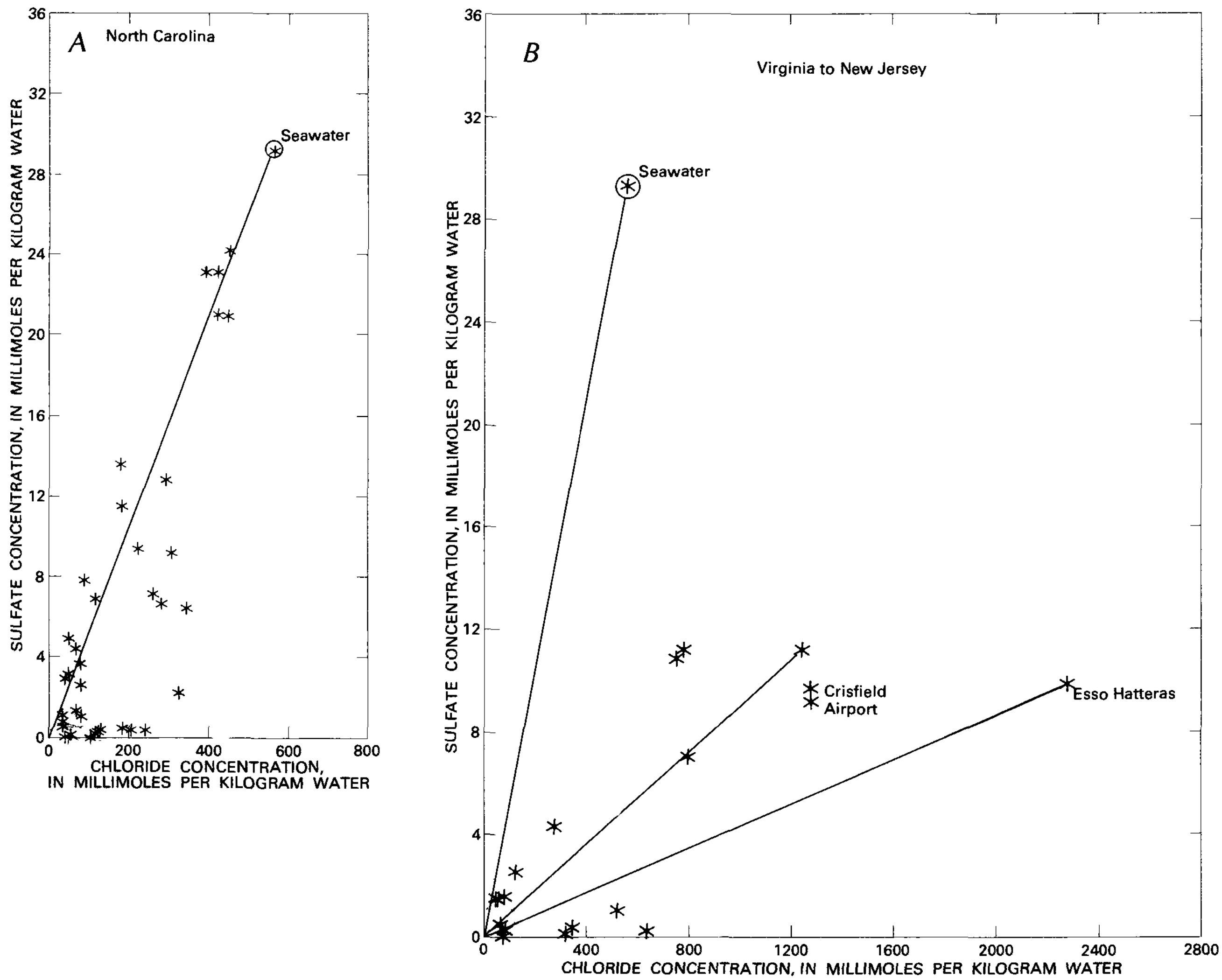


Figure 14. Relation of sulfate concentration to chloride concentration. Lines denote conservative mixing curves. *A*, North Carolina. *B*, Virginia to New Jersey.

The occurrence in North Carolina of a brine containing chloride concentrations greater than that of seawater is indicated by the interpretation of geophysical logs from 22 wells (Meisler, 1981). Nine of the wells have estimated chloride concentrations of at least twice that of seawater. The only chemical analysis available to confirm the occurrence of this brine is from the Esso Hatteras well (table 2, No. 19). The chloride concentration in this analysis is almost four times that of seawater. This analysis more closely resembles analyses from Virginia to New Jersey and has been plotted with the latter.

Virginia to New Jersey

In the area from Virginia to New Jersey, comparison of the plots of ion concentrations with conservative mixing curves indicates that concentrations of most constituents cannot be explained by the conservative mixing of seawater and freshwater. Several analyses (including Esso Hatteras from North Carolina) have chloride concentrations greater than that of seawater. Calcium concentrations (fig. 10*B*) are considerably higher than the seawater-freshwater mixing values whereas magnesium, potassium, and sulfate (figs. 11*B*, 13*B*, and 14*B*) are considerably lower. On the contrary, sodium concentrations

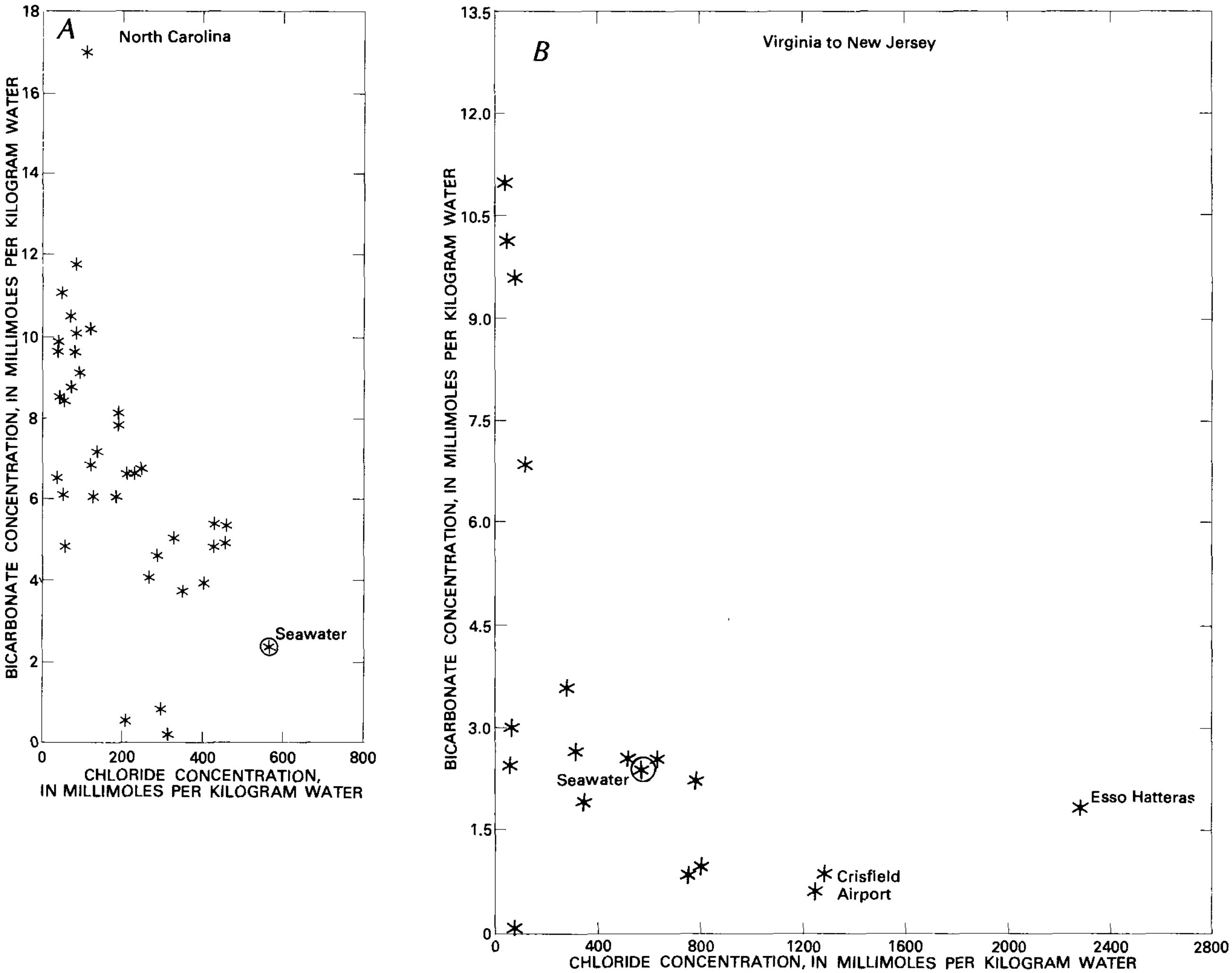


Figure 15. Relation of bicarbonate concentration to chloride concentration. *A*, North Carolina. *B*, Virginia to New Jersey.

(fig. 12*B*) follow the seawater-freshwater mixing curve. Concentrations of calcium plus magnesium (fig. 16*B*) appear to be only slightly lower for those analyses in which chloride concentrations are less than that of seawater.

The plots do suggest, however, that ion concentrations can be explained largely by mixing of a fresh sodium bicarbonate water with a sodium calcium chloride brine. The nature of the freshwater is suggested by the trend of ions on figures 10*B*–15*B* in which all constituents except sodium and bicarbonate approach the origin. Analyses of fresh sodium bicarbonate waters reported by Back (1966) indicate that sodium and bicarbonate concentrations range up to 14 and 12 mmol/kg, respectively. Calcium and magnesium concentrations in the same analyses range up to 0.4 mmol/kg. The exact chemical nature of the brine is not known and undoubtedly varies geographically. However, analyses of waters containing the highest chloride concentrations in the study area (Crisfield Airport and Esso Hatteras) indicate that the brine differs significantly from seawater in having higher concentrations of calcium, sodium, and chloride (figs. 10*B* and 12*B*) and lower concentrations of potassium, sulfate, and bicarbonate (figs. 13*B*, 14*B*, and 15*B*). Conservative mixing curves of freshwater and brine as represented by analyses from both Crisfield Airport and Esso Hatteras are shown on the graphs. The

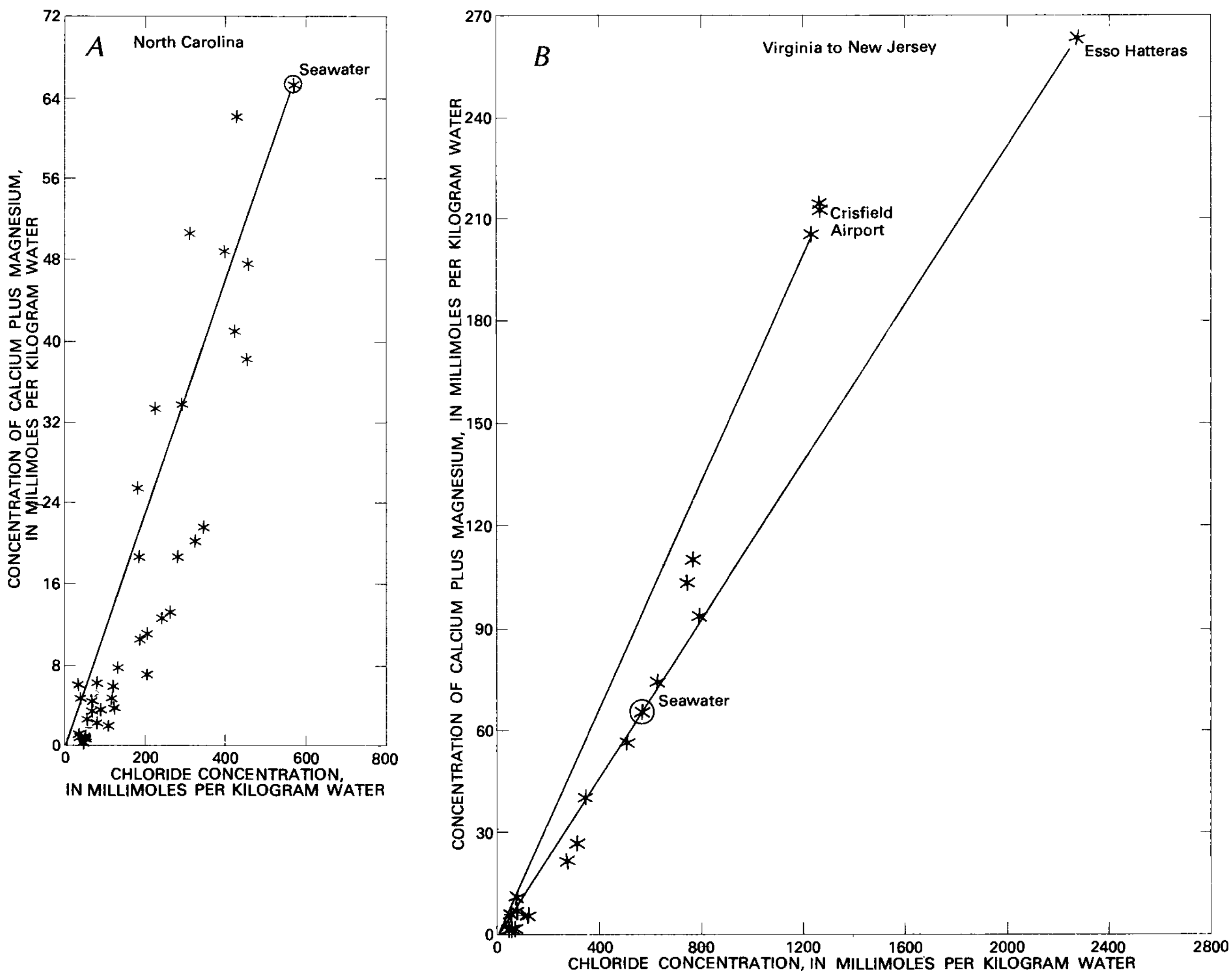


Figure 16. Relation of the concentration of calcium plus magnesium to chloride concentration. Lines denote conservative mixing curves. *A*, North Carolina. *B*, Virginia to New Jersey.

Esso Hatteras mixing curve is a closer match to the concentration data for calcium, magnesium, and sodium. The Chrisfield Airport mixing curve matches the sulfate data more closely. No potassium analysis was available from the Hatteras well.

Deviation of the ion concentrations from these mixing curves indicates that processes other than the simple mixing of freshwater and brine, as exemplified by the analyses from Esso Hatteras or Crisfield Airport, are taking place. Mixing with a third end member, seawater, is one possibility. As already mentioned, sodium follows the seawater mixing line for all analyses that are fresher than seawater. Plots of other constituents, such as calcium, potassium, and possibly magnesium and sulfate may also be influenced by the addition of seawater for waters containing less than 400 mmol/kg of chloride.

Some of these deviations from the brine-freshwater mixing curves can be explained by another process—ion exchange. Because exchange is dependent upon the nature of the exchanging media and the cation concentrations of the solution, the solution becomes enriched with calcium in the more concentrated waters and with sodium and potassium in the more dilute waters (Drever, 1982, p. 84; Nadler and others, 1980). The converse is true of the solid phase. The enrichment of calcium in solution in the more concentrated waters and sodium in

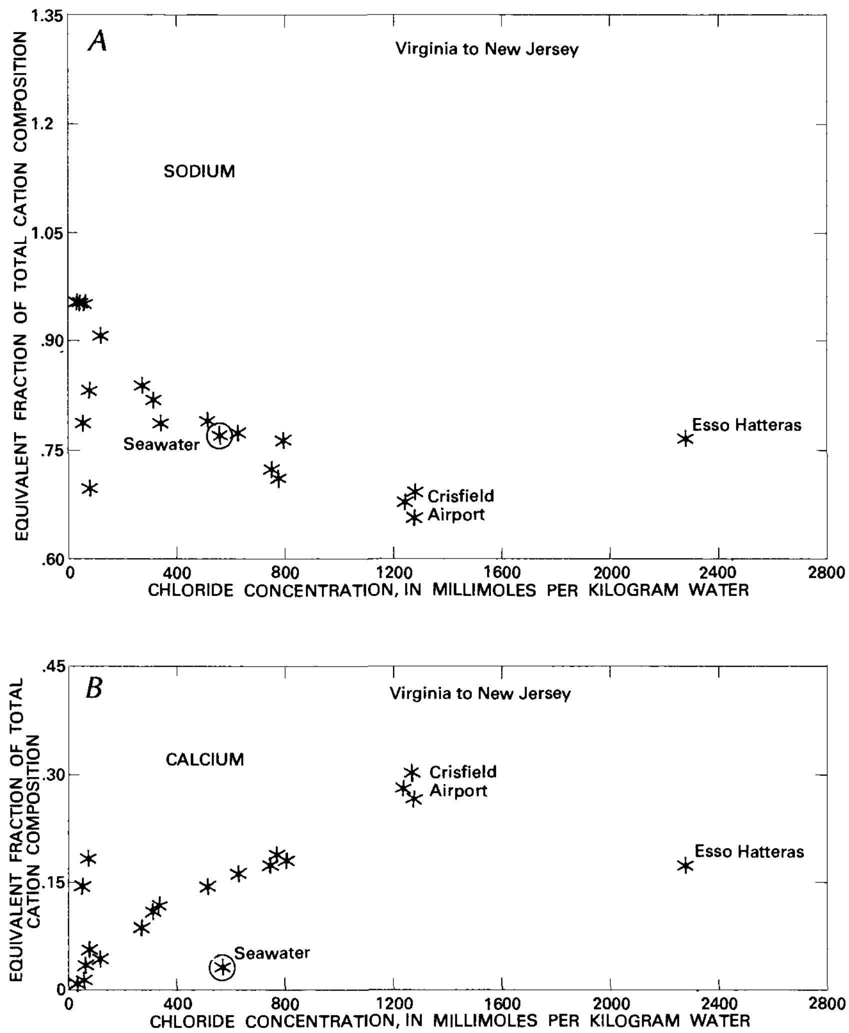


Figure 17. Relation of calcium and sodium (expressed as an equivalent fraction of total cations) to chloride concentration. *A*, sodium. *B*, calcium.

Table 2. Chemical analyses of water from selected wells, northern Atlantic Coastal Plain

Map No.[1]	Well name	Concentrations, in millimoles per kilogram water								
		Calcium	Magnesium	Sodium	Potassium	Chloride	Sulfate	Bicarbonate	Silica	pH[2]
1	Salem Observation	4.00	1.44	43.6	0.74	53.7	0.02	2.46	0.12	[3]6.9
2	Anchor Ragovin	55.56	19.15	528.3	3.11	628.1	.22	2.54	.18	[3]6.6
	do	40.34	15.79	439.5	2.84	513.0	1.05	2.55	.20	[3]6.5
	do	23.20	16.63	307.7	2.38	342.1	.32	1.90	.17	[3]6.7
	do	17.59	8.70	258.5	3.86	312.5	.13	2.66	.18	[3]6.9
3	QA-BE 15	6.79	3.54	50.5	1.00	72.9	.27	.07	.14	[3]6.8
4	DO-CE 88 (Cambridge test)	1.10	.32	61.0	.23	62.1	.40	3.00	.32	[3]7.85
	do	73.46	20.46	618.2	1.25	796.1	7.03	.97	.24	[3]7.60
5	WO-AH 6	2.50	4.04	73.6	1.66	76.6	1.55	9.58	.85	[3]8.4
6	SO-DD 47 (Crisfield Airport)	184.82	20.75	887.3	3.30	1241.6	11.18	.62	.27	[3]6.7
	do	190.80	22.59	983.8	1.65	1276.0	9.19	.88	.30	[3]6.3
	do	192.89	21.64	840.7	1.70	1272.0	9.72	.88	.29	[3]6.7
7	Moores Bridge, Norfolk	2.63	2.39	114.3	1.33	118.8	2.53	6.84	.40	[3]7.42
	do	12.12	8.70	232.5	2.25	273.0	4.32	3.60	.39	[3]6.93
	do	67.69	35.83	559.0	3.23	748.0	10.83	.84	.92	[3]7.13
	do	74.93	35.54	568.7	3.21	775.0	11.16	2.23	.20	[3]6.4
	do	.38	.41	48.4	.67	39.0	1.43	10.97	.22	[3]7.82
	do	.50	.54	56.2	.69	47.5	1.46	10.11	.23	[3]7.79
8	Morgan's Corner Research Station[4]	.42	.62	61.0	.74	50.9	3.13	8.42	--	8.8
	do	2.20	2.56	122.2	1.28	118.9	6.89	6.84	.13	[3]7.85
9	Elizabeth City Research Station[4]	1.38	2.27	91.5	1.74	82.0	2.61	11.76	.60	8.0
10	Parkville Research Station[4]	1.62	2.88	61.0	2.08	67.8	1.36	10.48	.23	8.1
11	Cremo Research Station[4]	1.43	2.27	109.1	1.82	90.6	7.83	9.12	.32	8.1
12	Duck Research Station[4]	1.08	4.94	117.7	1.97	118.7	.21	10.20	.38	8.2
	do	2.50	10.32	213.9	4.11	243.4	.42	6.76	.05	7.6
	do	2.48	8.66	187.5	3.85	206.4	.54	6.62	.04	7.6
13	Duckwood Research Station[4]	1.18	1.52	54.9	1.20	55.7	.10	4.83	.04	7.7
	do	2.48	8.16	173.6	3.23	187.8	.48	7.82	.10	7.9
	do	1.45	6.39	123.8	2.00	133.2	.36	7.16	.20	7.7
14	Scuppernong Research Station[4]	13.56	24.84	350.3	5.41	454.3	20.96	4.95	0.08	[3]7.48
15	Plymouth Research Station[4]	4.77	20.68	201.2	--	181.5	13.61	6.06	.09	7.8
	do	1.20	2.43	65.4	1.61	67.8	4.38	8.77	.11	7.7
16	Gum Neck Research Station[4]	1.50	4.94	82.8	--	81.9	1.04	10.11	--	7.8
17	Hydelands Research Station[4]	16.62	31.14	439.1	10.07	455.6	24.17	5.36	.34	7.5
	do	4.26	14.46	196.6	3.08	187.0	11.50	8.15	.54	7.0
18	Hobucken Research Station[4]	13.58	26.52	385.7	6.70	426.3	20.98	4.86	.16	7.2
	do	1.30	3.54	43.6	1.28	39.5	2.92	8.52	.67	7.4
19	Esso Hatteras	202.84	60.64	1777.9	--	2278.3	9.90	1.84	--	--
20	Arapahoe Research Station[4]	11.78	22.06	242.5	5.91	294.7	12.86	.86	.06	6.6
	do	2.70	3.46	30.0	1.41	33.9	1.09	6.53	.63	7.5
	do	9.77	40.89	209.7	4.37	311.6	9.20	.23	.01	6.0
21	Atlantic Research Station[4]	8.53	23.98	209.8	--	226.8	9.42	6.64	.10	7.6
	do	15.87	33.22	404.0	7.23	398.6	23.12	3.97	.06	7.9
22	Moores Creek Nat. Milit. Park	.30	.54	45.3	1.10	37.3	.55	9.68	.15	7.0
	do	6.72	12.02	255.6	5.24	283.4	6.52	4.61	.15	8.0
23	Hercules Waste Injection Site	8.28	13.36	321.4	4.50	345.9	6.44	3.76	.16	7.7
	do	11.66	8.64	297.2	4.11	325.9	2.25	5.06	.18	7.0
	do	.48	1.90	87.2	1.49	78.6	3.65	9.64	.16	8.1
	do	5.59	7.69	257.8	3.73	263.6	7.11	4.07	.10	7.8
24	Nakina Research Station[4]	.80	.45	47.9	.49	36.7	.70	9.86	.08	8.0
	do	1.90	1.94	113.3	--	124.3	.26	6.07	.05	7.8
25	Sunset Harbor Research Station[4]	.12	.70	65.4	--	50.9	4.90	6.11	--	9.2
	do	.10	.32	52.3	--	45.2	--	11.05	--	9.2
	do	3.50	3.71	183.0	--	206.3	.44	.55	--	9.1
26	Holden Beach Research Station[4]	12.58	49.78	394.9	8.77	426.8	23.10	5.42	.01	7.9
27	Calabash Research Station[4]	1.15	1.01	104.2	.90	107.4	--	17.02	--	8.0
	Seawater[5]	10.66	55.07	485.4	10.58	565.8	29.26	2.41	.07	8.22

[1]Location shown in figure 1.
[2]pH units.
[3]Field measurement.
[4]North Carolina Department of Natural Resources and Community Development. The cooperation of this agency is gratefully acknowledged.
[5]Seawater concentrations from Plummer and others, 1978.

the more dilute waters is depicted in figure 17 which shows the relation of both calcium and sodium (expressed as an equivalent fraction of the total cation composition) to chloride concentration. The data do not clearly indicate a progressive enrichment or depletion of magnesium. The loss of potassium from the more concentrated waters may be due to its tendency to be incorporated into the structure of the clay minerals (Hem, 1970, p. 150–151).

CONCLUSIONS

The effect of eustatic sea-level changes on the distribution of saltwater and freshwater in the sediments of the Atlantic Coastal Plain and Continental Shelf has been analyzed using a finite-difference model. The model simulates, in cross section, a sharp interface between flowing freshwater and static saltwater. Freshwater pressure at the simulated interface equals the pressure exerted by a column of static sea water.

Steady-state positions of the interface were simulated for several hypothetical sea levels and for different distributions of vertical and lateral hydraulic conductivity. The position of the interface is sensitive to sea level and to anisotropy, that is, the ratio of lateral to vertical hydraulic conductivity. Lowering sea level causes the freshwater flow system to be deeper and to extend farther offshore. Increasing the anisotropy also causes the simulated freshwater flow system to extend farther offshore, but causes it to be shallower.

Simulations that used hydraulic conductivities based on available data, in which lateral conductivity decreases offshore, suggest that the transition zone in southern New Jersey and beneath the adjacent Continental Shelf is not in equilibrium with present sea level. The position of the transition zone appears to reflect sea levels that were 50 to 100 ft below present sea level. Although the average sea level for the Quaternary is probably in this range, it is likely that the transition zone reflects sea level conditions for a longer period, including the Pliocene and possibly part of the Miocene. The saltwater-freshwater transition zone is presently moving slowly landward and upward in response to the most recent rise in sea level. Estimates of lateral water particle velocity suggest that the offshore interface is moving landward at a rate of about 0.2 mile per 10,000 years.

Cyclic movement of saltwater, responding to large-scale eustatic sea-level fluctuations during the Quaternary, Pliocene, and possibly the Miocene, caused the mixing of saltwater and freshwater and produced the observed broad transition zone. A much thinner transition zone, from the ocean floor to the underlying fresher water, probably results from short-term reversals in head gradients beneath the ocean floor due to sea-level fluctuations from storms and tides. Although an interface between saltwater and underlying freshwater could not be simulated, the model was able to simulate freshwater discharge to the ocean through the overlying transition zone.

Study of chemical analyses of water samples from the northern Atlantic Coastal Plain supports the interpretation that the broad transition zone was produced by the mixing of freshwater and saltwater. The freshwater is predominantly sodium bicarbonate in character. The saltwater from New Jersey to Virginia is a sodium calcium chloride brine which has a chlorinity several times that of seawater. The exact chemical nature of the brine is not known and undoubtedly varies geographically. It differs also from seawater in having higher concentrations of calcium and sodium and lower concentrations of potassium, sulfate, and bicarbonate. The saltwater in North Carolina is close in composition to seawater. Other data from North Carolina, including geophysical logs, indicate the occurrence of brine beneath the transition zone.

Concentrations of several ions indicate deviations from conservative mixing in the transition zone. As a probable result of ion exchange, calcium enrichment has occurred in the more concentrated waters and sodium enrichment has occurred in the more dilute waters.

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