THE REGIONAL AQUIFER SYSTEM UNDERLYING THE NORTHERN ATLANTIC COASTAL PLAIN IN PARTS OF NORTH CAROLINA, VIRGINIA, MARYLAND, DELAWARE, NEW JERSEY, AND NEW YORK—SUMMARY

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1404-A

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By HENRY TRAPP, JR., *and* HAROLD MEISLER

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) Program was started in 1978 following a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which in aggregate underlie much of the country and which represent an important component of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and accordingly transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system and the changes brought about in it by human activities, and to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number, and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400 and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

Dallas L. Peck
Director

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For those readers who prefer to use metric (International System) units rather than the inch-pound system of units in this report, the conversion factors are listed below.

Multiply	By	To obtain
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	259.0	hectare (ha)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
million gallon per day (Mgal/d)	3,785	cubic meter per day (m^3/d)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
foot per day per foot (ft/d)/ft	1	meter per day per meter (m/d)/m

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

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By HENRY TRAPP, JR., and HAROLD MEISLER

ABSTRACT

The northern Atlantic Coastal Plain aquifer system extends along the Atlantic Coastal province from North Carolina through Long Island, New York. Its western limit is the landward edge of water-bearing Coastal Plain sediments. This study focuses on the emergent Coastal Plain and adjoining estuarine bodies, an area of about 50,000 square miles.

The northern Atlantic Coastal Plain contains a multilayered aquifer system composed of sedimentary deposits capable of large yields to wells. The sediments were deposited on a basement surface that slopes gently toward the Atlantic Ocean. The thickness of the emergent Coastal Plain sediments ranges from a featheredge near the Fall Line to about 10,000 feet at Cape Hatteras, North Carolina. Offshore from New Jersey and the Delmarva Peninsula, the thickness of sediments in the Baltimore Canyon Trough exceeds 7.5 miles.

The Coastal Plain sediments range in age from Jurassic to Holocene. Younger deposits tend to progressively overlap older ones in a landward direction. In the Cretaceous section, there is a general upward transition from fluvial and fluviodeltaic to marginal marine to marine deposits. The marine deposits consist primarily of glauconitic sand, silt, clay, and limestone beds, which are traceable over longer distances than are the more lenticular nonmarine deposits. The Tertiary sediments are predominantly marine except for the Upper Miocene and Pliocene beds, which are partly nonmarine. The Pleistocene section includes glacial drift on Long Island and marine, dune, and terrace deposits elsewhere.

The Coastal Plain sediments have been subdivided into 11 regional aquifers and 9 confining units on the basis of continuity of permeability. Transmissivity of most of the aquifers ranges from about 500 to 10,000 ft^2/day (feet squared per day), but the transmissivity of parts of the Castle Hayne–Piney Point aquifer in North Carolina exceeds 100,000 ft^2/day and that of the Magothy and upper glacial aquifers exceeds 20,000 ft^2/day on Long Island.

Ground-water withdrawals (not including most withdrawals for irrigation) in the northern Atlantic Coastal Plain have increased from about 100 Mgal/day (million gallons per day) in 1900 to about 1,200 Mgal/day in 1980.

A multilayer digital computer model of the northern Coastal Plain simulates ground-water flow from Long Island, New York, through North Carolina for prepumping and pumping conditions (1900–80). According to the model results, pumpage increased the area of recharge to confined aquifers from about 26 to 45 percent of the total area by 1980, and locally, recharge rates increased by as much as 10 in./yr. Simulation suggests that most of the pumped water has been derived from reduced ground-water discharge to streams in updip areas, and only a small part of the water pumped is derived from storage. In most areas, the modeled system approaches equilibrium in less than 5 years after a change in pumpage.

Oxidation and reduction reactions predominate in the continental deposits, which have a higher percentage of feldspars, micas, and unstable clay and heavy minerals than do marine deposits. Ion exchange, carbonate dissolution-precipitation, incongruent dissolution of silicate minerals, dissolution of amorphous silica, and mass transfer of phosphatic material are the more important geochemical processes controlling water chemistry in marine sediments. These sediments tend to be rich in glauconite, phosphate minerals, calcite, and silica.

Ground water in the northern Atlantic Coastal Plain is classified into four hydrochemical facies: (1) variable composition, (2) calcium plus magnesium bicarbonate, (3) sodium bicarbonate, and (4) sodium chloride. The hydrochemical facies in each aquifer follow a general coastward, or downdip, sequence from variable composition through sodium chloride. The same sequence generally appears in moving downward in the section.

Salty ground water underlies freshwater in the eastern part of the northern Atlantic Coastal Plain from New Jersey to North Carolina. Chloride concentrations generally increase with depth. The transition zone between concentrations of 250 and 18,000 mg/L (milligrams per liter) of chloride ranges in thickness from about 400 to 2,200 feet. No ground water containing 10,000 mg/L or more of chloride was found north of southeastern New Jersey. Ground water containing less than 5,000 mg/L of chloride extends 55 miles offshore from New Jersey and lesser distances offshore farther south.

The position of the saltwater-freshwater transition zone probably reflects the effects of past sea levels. Comparison of sharp interfaces simulated by a cross-sectional digital model extending from southern New Jersey to the Continental Shelf to observed chloride concentrations indicates that the location of the transition zone in this area is

adjusted to an average sea level about 50–100 feet below present sea level. This position may approximate the average sea level during the late Tertiary and the Pleistocene.

Evidence for bacteria and for methanogenic and sulfate-reducing activity was found in deeply buried Coastal Plain sediments. Bacteria can produce carbon dioxide by decomposing lignite fragments, which could account for the apparent excess of dissolved inorganic carbon in Atlantic Coastal Plain ground water.

INTRODUCTION

During 1979–87, the U.S. Geological Survey conducted a regional analysis of the northern Atlantic Coastal Plain as part of the Regional Aquifer-System Analysis (RASA) program (see Foreword). The major focus of the study was to develop an understanding of the ground-water flow system and the way that it responds to pumping. The objectives of the northern Atlantic Coastal Plain RASA study were to (1) define the hydrogeologic framework of regional aquifers and confining units and describe their geologic and hydrologic properties; (2) construct multilayer finite-difference models for simulating predevelopment and transient-flow conditions; (3) define the predevelopment flow system and the changes caused by development and analyze these through use of the models; (4) describe the water chemistry and relate it to ground-water flow patterns, saltwater intrusion, and mineralogy of the sediments; and (5) define the regional saltwater-freshwater transition zone and evaluate its location and movement relative to the hydrogeology of the Coastal Plain.

PURPOSE AND SCOPE

Professional Paper 1404 describes the hydrologic framework, hydrology, and geochemistry of the northern Atlantic Coastal Plain. The professional paper consists of the following 12 chapters. (This report, chapter A, summarizes the findings reported in more detail in the other chapters.)

- A—The Regional Aquifer System Underlying the Northern Atlantic Coastal Plain in Parts of North Carolina, Virginia, Maryland, Delaware, New Jersey, and New York—Summary, by Henry Trapp, Jr., and Harold Meisler
- B—Hydrogeologic Framework of the New Jersey Coastal Plain, by O.S. Zapecza
- C—Hydrogeologic Framework of the Virginia Coastal Plain, by A.A. Meng III and J.F. Harsh
- D—The Occurrence and Geochemistry of Salty Ground Water in the Northern Atlantic Coastal Plain, by Harold Meisler
- E—Hydrogeologic Framework of the Coastal Plain in Maryland, Delaware, and the District of Columbia, by D.A. Vroblesky and W.B. Fleck
- F—Conceptualization and Analysis of Ground-Water Flow System in the Coastal Plain of Virginia and Adjacent Parts of Maryland and North Carolina, by J.F. Harsh and R.J. Laczniak
- G—Hydrogeologic Framework of the Northern Atlantic Coastal Plain in Parts of North Carolina, Virginia, Maryland, Delaware, New Jersey, and New York, by Henry Trapp, Jr.
- H—Ground-Water Flow in the New Jersey Coastal Plain, by Mary Martin
- I—Hydrogeologic Framework of the North Carolina Coastal Plain Aquifer System, by M.D. Winner, Jr., and R.W. Coble
- J—Simulation of the Ground-Water Flow System of the Coastal Plain Sediments: Maryland, Delaware, and the District of Columbia, by W.B. Fleck and D.A. Vroblesky
- K—Geohydrology and Simulation of Ground-Water Flow in the Northern Atlantic Coastal Plain Aquifer System, by P.P. Leahy and Mary Martin
- M—Simulation of Ground-Water Flow in the Coastal Plain Aquifer System of North Carolina, by G.L. Giese, J.L. Eimers, and R.W. Coble

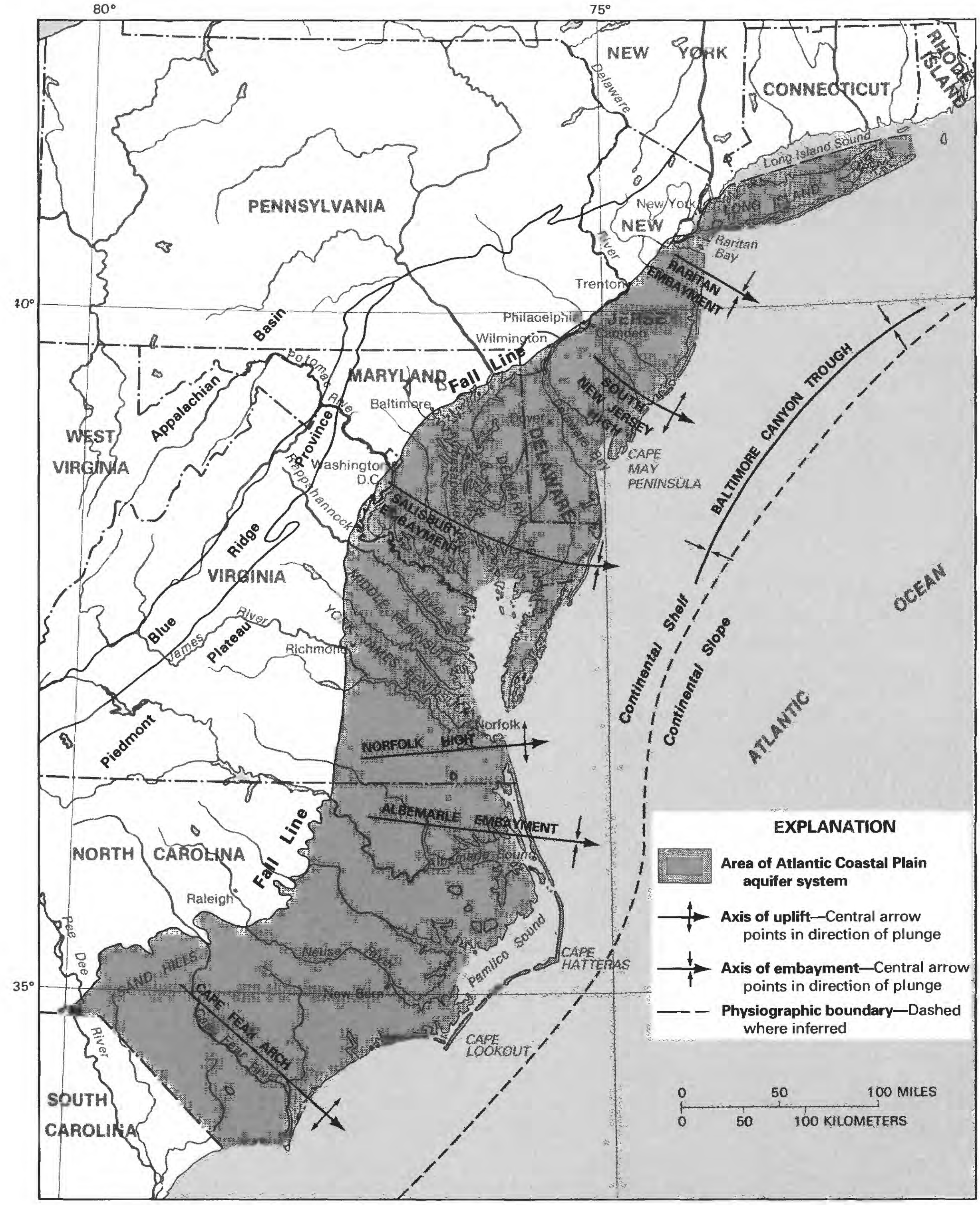
Complete bibliographic references for all chapters except this one are given under "Selected References." These and other reports that were prepared wholly or in part through the Northern Atlantic Coastal Plain RASA program are indicated by asterisks in "Selected References."

STUDY AREA

The northern Atlantic Coastal Plain extends from North Carolina to Long Island, N.Y. (fig. 1), and is underlain by predominantly unconsolidated sediments. The western boundary is the landward edge of water-bearing strata of Cretaceous through Pleistocene age, which approximates the Fall Line. The Coastal Plain extends eastward beneath the Continental Shelf, which underlies the Atlantic Ocean. This study focuses primarily on the emerged part of the Coastal Plain, an area of about 50,000 mi^2. However, the occurrence and movement of salty ground water within the submerged part have received considerable study.

SUMMARY OF PREVIOUS WORK

A general discussion follows of the increase in understanding the geologic history, extent of geologic formations and aquifers, ground-water flow, and distribution of waters of varying chemical composition in the northern Atlantic Coastal Plain. Included are references to



Base enlarged from U.S. Geological Survey National Atlas, 1970, 1:7,500,000

Positions of structural features adapted from Uchupi (1968), Maher (1971), and Owens and Gohn (1985).

FIGURE 1.—Location of the Atlantic Coastal Plain and major structural features (from Trapp, in press).

some of the papers produced over more than two centuries of investigations.

Among those who first recognized the Atlantic Coastal Plain as a geologic entity was J.D. Schöpf (1787), a German physician who accompanied Hessian troops during the Revolutionary War. He noted the rough parallelism of the Appalachian ranges with the coastline, described the Fall Line, and recognized the Coastal Plain basement as an extension of the harder rocks inland. He inferred that the Coastal Plain sediments, because of their unconsolidated nature, were the youngest formations in the area and that they were formed under water like the offshore banks of New England. He concluded that the water left the continent because of deepening of the ocean basin, which also included subsidence of the coast, and that the Gulf Stream helped shape the Coastal Plain and probably impinged on it in the past.

William Maclure (1809, 1817) prepared geologic maps of the United States up to the frontier of that period, with explanatory texts that included a description of the Fall Line. He mapped the Coastal Plain formations as "the Alluvial."

John Finch, of the University of Birmingham (in England), recognized that the Coastal Plain was more than a single alluvial formation. Finch wrote (1824, p. 32–33): "*** in America, an immense tract of country, extending from Long Island to the sea of Mexico, and from thirty to two hundred miles in width, is called an alluvial formation, by most of the geologists who have written upon the subject, and by some it appears to be considered as an exception to the general arrangement and position of strata, which are found to occur in other countries ***. I wish to suggest that what is termed the alluvial formation in the geologic maps of Messrs. Maclure and Cleaveland is identical and contemporaneous with the newer secondary and tertiary formations of France, England, Spain, Germany, Italy, Hungary, Poland, Iceland, Egypt, and Hindoostan ***. There are no rivers on the coast which could have deposited such an accumulation of sand and marle, and the hills of limestone."

Organized geologic exploration of the Coastal Plain began with State geologic surveys. Initially, they emphasized the location of natural resources such as clay and lime and the relation of geology to soil fertility. Understanding of the geology of the Coastal Plain grew with the systematic tracing and naming of its strata. The application of stratigraphic names based on type localities began with Ruffin (1843, p. 24–27), who named the Upper Cretaceous Peedee Formation (which he called the Peedee bed) for its exposures along the "Peedee" River (now spelled Pee Dee) in South Carolina. It was later traced into North Carolina (Stephenson, 1912, p. 145, 170). Summaries of the stratigraphy of the Coastal Plain by State, with formation and age designations approaching present usage, were published as early as 1906 (Veatch and others, 1906) for Long Island, N.Y.; 1905 (Weller, 1905, 1907) for New Jersey; 1884 (Chester, 1884) for Delaware; 1901 (Clark, Martin, and others, 1901; Clark, Shattuck, and Dall, 1904; Clark, Bibbins, and others, 1911; Clark, Berry, and Gardner, 1916; Clark, Goldman, and others, 1916; Shattuck and others, 1906) for Maryland; 1906 (Clark and Miller, 1906, 1912) for Virginia; and 1912 (Clark, Miller, and others, 1912) for North Carolina.

Early geophysical work (Ewing and others, 1937, 1939, 1940, 1950; Miller, 1937; Ewing, 1940), largely refraction-seismic and gravity surveys, indicated that the sedimentary layers of the emergent Coastal Plain were a continuation of similar layers underlying the Continental Shelf. Major offshore trends of geophysical anomalies, interpreted as belts of rock of contrasting densities and seismic velocities, were shown to run roughly parallel to Appalachian structural trends. The Atlantic and Gulf Coastal province was described as a geosyncline, separated into two segments by the Ocala arch of Florida and central Georgia (Murray, 1961, p. 21–47, 79–166).

The development of plate-tectonic theory (Wilson, 1968) provided new insights into the deposition of sediments along continental margins, which were applicable to the Atlantic and Gulf Coastal province. Rifting (the first stage of continental breakup) began between North America, Eurasia, and Africa in the Triassic, with wrench and transform faults associated with a thinned continental crust. In North America, the major faults generally ran parallel to old Appalachian lineaments. Sediments accumulated in the rift basins along the faults during Triassic and Early Jurassic time under conditions of great crustal instability, with tilting, folding, igneous intrusion, and widespread volcanism. After the opening of the early Atlantic Ocean and the beginning of seafloor spreading, the environment of deposition was characterized by gentle subsidence of the continental margin and marine incursions (Manspeizer and others, 1978). The post-rifting Coastal Plain and Continental Shelf sediments were deposited during this second phase, which persists through the present.

Brown and others (1972) divided the northern Atlantic Coastal Plain sedimentary section into 17 chronostratigraphic units and mapped thicknesses, lithofacies, and relative intrinsic permeabilities. They also proposed recurrently reversing vertical movement along wrench faults during deposition as a hypothesis to account for variations in thickness and facies of the chronostratigraphic units. Movement along the wrench faults was interpreted as the near-surface expression of the displacement of basement blocks.

Darton (1896) and Fuller (1905) published the first comprehensive reports, consisting primarily of well data, on the ground-water resources of the area of this study.

Freeman (1900), together with Crosby (1900), studied the water resources of the western part of Long Island but saw no possibility of continuing large-scale supplies from deeper aquifers because of the lack of connection to their supposed recharge areas (outcrops) on the mainland. Thompson and others (1937) pointed out that heads in the Lloyd aquifer were highest in the areas where Crosby had thought no recharge was possible. This finding indicated that vertical recharge reached the Lloyd aquifer through the confining units. On the basis of personal observations on Long Island, Jacob (1939, 1940, 1941, 1943) published papers on the behavior of water in artesian aquifers. During an investigation of saltwater intrusion on Long Island, Lusczynski (1961) developed concepts of head and flow in ground water of variable density.

LeGrand (1964) described the hydrogeologic framework of the Atlantic and Gulf Coastal Plains. He noted that the interlayering of relatively permeable material with less permeable material has resulted characteristically in several distinct aquifers in most of the Coastal Plain and that most ground-water recharge is short-circuited to effluent stream valleys through near-surface aquifers, except in the semiarid part of Texas. Cederstrom and others (1971, 1979) and Sinnott and Cushing (1978) summarized information on the ground-water resources of the Coastal Plain and adjacent areas. Brown and Reid (1976) and Lloyd and others (1985) studied the saltwater-saturated part of the Coastal Plain aquifer system with respect to its potential for storing wastes.

The aquifer system of Long Island was the first to be simulated by a flow model encompassing the greater part of the natural system. Simulation was by a five-layer electric-analog model that covered all but the extreme eastern end of the island but that omitted the lowermost (locally the Lloyd) aquifer (Getzen, 1977). The Lloyd aquifer was not included in the simulation because it was considered to be largely isolated from the remainder of the aquifer system by its overlying confining unit, it was used in only a few places, and its hydraulic properties were poorly known (Getzen, 1977, p. 23).

Sanford (1911) initiated the description of the distribution of waters of various chemical compositions in the Coastal Plain aquifer system by summarizing data on saltwater and saltwater aquifers. Cederstrom (1946a, b) mapped the distribution of the principal chemical constituents in the Coastal Plain of Virginia and noted changes in the predominant ions as the distance from the aquifer outcrops increased, especially an increase in bicarbonate. He concluded that carbonate dissolution and ion exchange were major chemical processes in determining the distribution of the principal constituents and that the apparent excess of bicarbonate could be explained by the generation of carbon dioxide within the aquifers in association with sulfate reduction. Foster (1950) maintained that oxidation of lignitic material in the aquifers was the source of the carbon dioxide. On the basis of work in the Coastal Plain, Back (1961) developed a methodology for mapping hydrochemical facies and mapped the northern Coastal Plain (Back, 1966), relating facies patterns to ground-water flow.

Barksdale and others (1958, p. 110–111) delineated the saltwater-freshwater boundary in the Raritan and Magothy Formations of New Jersey and Delaware, largely on theoretical considerations. Upson (1966) studied the relation of freshwater to salty ground water from Long Island to Maryland. Manheim and Horn (1968) showed the salinity distribution in 28 wells along the Atlantic Coast from Long Island, N.Y., to Key West, Fla. Cushing and others (1973) delineated the base of water containing less than 1,000 mg/L dissolved solids on the Delmarva Peninsula. Larson (1981, pl. 2) and Heath and others (1975, fig. 8.20) mapped the base of water containing less than 250 mg/L in the Coastal Plain of Virginia and North Carolina, respectively.

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The authors acknowledge the cooperation of State agencies that provided data on wells and geology. The agencies include the New Jersey Department of Environmental Protection, the Delaware Geological Survey, the Maryland Geological Survey, the Virginia Division of Mineral Resources, the Virginia Water Control Board, and the North Carolina Department of Natural Resources and Community Development (NRCD). P.M. Brown, Chief of the Geologic Section of NRCD, provided copies of many logs of North Carolina and Virginia wells. Drilling contractors, consultants, local water-supply agencies, and industries also contributed data used in developing the hydrogeologic framework. John K. Costain, of the Virginia Polytechnic Institute and State University, provided copies of reports prepared for the U.S. Department of Energy on geothermal resources in the study area. The authors also thank the individuals and agencies (Trapp and others, 1984, p. 1–2) that assisted in drilling a test well at Cambridge, Md., and in acquiring and interpreting geologic and hydrologic data.

PHYSICAL SETTING

PHYSIOGRAPHY

The northern Atlantic Coastal Plain is part of the Atlantic and Gulf Coastal province, which extends from north of Newfoundland to Honduras and from the landward edge of Cretaceous and younger strata eastward to the Continental Slope and Rise. The submerged Coastal province is the Continental Shelf. North of Cape Cod, the Coastal province is entirely submerged (Fenneman, 1938, p. 1–3; Murray, 1961, p. 1–2). The emergent Coastal province constitutes the Coastal Plain in the study area. Southward, the Coastal Plain broadens to as much as 140 mi in North Carolina, and the Continental Shelf correspondingly narrows.

The Fall Line, which marks the boundary between the Coastal Plain and the Piedmont Plateau province (fig. 1), is so named because of the prevalence of falls and rapids in streams crossing the contact between the indurated rocks of the Piedmont Plateau and the unconsolidated and weakly consolidated sediments of the Coastal Plain. The increase in stream gradient near the Fall Line provided favorable locations for water power, and on most major rivers, it coincided with the head of ocean navigation. Thus, major cities grew up along it, including Trenton, N.J., Philadelphia, Pa., Wilmington, Del., Baltimore, Md., Washington, D.C., and Richmond, Va.

The Fall Line is sharply defined at Washington, where the buried basement surface has a slope of 100 ft/mi, but becomes less distinct southward, especially in North Carolina. In North Carolina, the basement surface slopes more gently, and in places, the easternmost rocks of the Piedmont Plateau are Triassic strata that are little more resistant to erosion than are the adjacent and overlying Coastal Plain Cretaceous strata. Under these conditions, zones of rapids are not necessarily related to the landward limit of Coastal Plain sediments. In North Carolina, the Piedmont-Coastal Plain contact in the beds of some of the major rivers is offset 20 mi or more downstream from the contact in the divide areas. These rivers, although walled by banks of Cretaceous strata, are incised into crystalline rocks (Fenneman, 1938, p. 126–129).

Most of the study area from Cape Lookout, N.C., northward is in the embayed section of the Coastal Plain, "so indented by branching bays or estuaries that it is little more than a fringe of peninsulas, narrowing to zero at New York and represented beyond that by islands *** the edge of the continent has here been depressed *** the amount of depression increases toward the north. The rivers of this section as far south as the James and Appomattox are drowned to the Fall Line" (Fenneman, 1938, p. 13). Barrier islands fringe the coast. From New Jersey to the Rappahannock River, a feature of the embayed section is a band of highly dissected Cretaceous outcrops along the Fall Line. On Long Island, the Cretaceous sediments are mantled by glacial deposits, including two terminal moraines (Veatch and others, 1906; Fuller, 1914; Fleming, 1935). They are mostly buried by Tertiary sediments south of the Rappahannock.

The area of investigation south of Cape Lookout, N.C., is part of what Fenneman (1938, p. 38–40, 45–46) called the Sea Island section of the Coastal Plain. According to Fenneman's interpretation, the offshore islands in this section are remnants of the mainland, cut off by enlarged tidal channels. More recent studies have attributed the origin of barrier islands along the coast from New England to Texas to the reworking of deltaic and beach sands by the rising ocean after the Pleistocene (Dolan and Lins, 1986, p. 11–14). Nevertheless, Fenneman noticed that drowning of the rivers is less pronounced in the Sea Island section than in the embayed section and, as previously stated, that the Fall Line west of this section is less distinct.

In both the embayed and Sea Island sections of the Coastal Plain, as many as eight terrace levels have been identified (Fenneman, 1938, fig. 10). The scarps of the terraces are roughly parallel to the present shoreline, major embayments, and rivers. Their typical altitudes range from 12 to 270 ft, but surfaces as high as 500 ft have been correlated with the uppermost (Brandywine) terrace. Their number, correlation, and origin are controversial.

Land-surface altitudes in the study area range from 0 ft to as much as 715 ft in the Sand Hills of the southwestern part of the North Carolina Coastal Plain. The Sand Hills also have the greatest local relief (as much as 350 ft) (Fenneman, 1938, p. 39).

PRECIPITATION

The source of freshwater in the northern Atlantic Coastal Plain is precipitation (rain, snow, and hail) on the Coastal Plain and on the upper drainage basins of Coastal Plain rivers whose headwaters originate in the adjoining Piedmont Plateau and Appalachian Mountains.

Average precipitation on the northern Atlantic Coastal Plain ranges from 40 in/yr near Washington, D.C. (Sinnott and Cushing, 1978, p. 6), to more than 56 in/yr on the Outer Banks of North Carolina (Cederstrom and others, 1979, p. 4). Most of the Coastal Plain receives 44–48 in/yr, but much of the North Carolina coast and part of the southern Delmarva Peninsula receive more than 48 in/yr (Cederstrom and others, 1979, fig. 2). An estimate of the average rate of precipitation for the entire study area is 47 in/yr (Leahy and Martin, in press).

Runoff is that part of precipitation that appears in surface streams. It may be subdivided into surface runoff, storm seepage, and ground-water runoff, which is ground water that has been discharged into a stream channel as spring or seepage water (Langbein and Iseri, 1960, p. 13, 17). Except for irrigation pumpage, most of the water pumped from either ground-water or surface-water sources reappears as runoff. Therefore, runoff in inches approximates precipitation minus evapotranspiration. In the northern part of the study area, annual runoff is generally in the range of 15–20 in/yr. Farther south, it is generally less than 15 in/yr, and it is less than 10 in/yr along the coast of southern Virginia and northeastern North Carolina (U.S. Geological Survey, 1970, p. 118–119). The lower values in these areas may be explained by increased rates of evapotranspiration related to higher average temperatures and the water table nearer the surface.

HYDROGEOLOGY

GEOLOGIC SETTING

The northern Atlantic Coastal Plain is underlain by a wedge of unconsolidated to semiconsolidated, predominantly clastic sedimentary rocks. These sediments overlie older metamorphic, igneous, and consolidated sedimentary rocks of the Piedmont physiographic province. The Coastal Plain sediments range in age from Jurassic to Holocene and consist mostly of sand, silt, clay, and limestone but also contain minor amounts of gravel. Most of the limestone occurs in North Carolina.

The Coastal Plain sediments generally strike northeast-southwest and dip gently seaward (pls. 1A, 1B); however, broad folds, arches, and embayments are superimposed on this relatively simple structure (pl. 1C). The sediments thicken from a featheredge at the western limit of the Coastal Plain, near the Fall Line, to about 8,000 ft at the western edge of the Delmarva Peninsula and 10,000 ft at Cape Hatteras, N.C. Offshore from New Jersey and the Delmarva Peninsula, the thickness of sediments greatly increases (pl. 1A). It exceeds 7.5 mi in the Baltimore Canyon Trough.

Plate 2A shows areal variations in the thickness of the Coastal Plain sediments. The sedimentary wedge thickens eastward beneath the Continental Shelf to more than 40,000 ft in the Baltimore Canyon Trough. Jurassic sediments compose as much as 30,000 ft of this section, whereas the total thickness of Cretaceous and Tertiary sediments generally does not exceed 13,000 ft. On the emerged Coastal Plain, Jurassic sediments compose only a small part of the stratigraphic section, probably not exceeding 1,000 ft at Cape Hatteras and only a few hundred feet farther north along the coast.

The generalized stratigraphic correlation of the principal geologic units in the northern Atlantic Coastal Plain is shown in plate 2B.

GEOLOGIC HISTORY

Deposition of the sediments of the Coastal Plain and Continental Shelf began with the inception of the Atlantic Ocean during the Jurassic Period as the African and North American continents separated. Initial accumulation of sediments occurred in both nonmarine and marine environments east of the present shoreline. Earliest deposition (principally during Early Cretaceous time) on what is now the emerged Coastal Plain was largely in a nonmarine environment; the sediments, derived from adjacent upland areas, were transported by streams and deposited as alluvial and deltaic sediments.

During Late Cretaceous time, a widespread marine transgression caused most of the Atlantic Coastal Plain to be covered by a shallow sea, and marine sediments were deposited on the nonmarine sediments. This general transgressive trend, interrupted by minor regressions, continued through the early Tertiary. The result was the accumulation of sediments in which clayey to silty glauconitic sands were deposited in a marine environment, and quartzitic sands and lenticular clays were deposited in a near-shore or coastal environment. Marine limestones were deposited in North Carolina during the early Tertiary.

The widespread marine transgression was interrupted in parts of the Coastal Plain during the late Tertiary. Miocene sediments are shallow marine to nonmarine in origin in Delaware, New Jersey, and parts of Maryland; however, these sediments are more generally of marine origin farther south. Post-Miocene sediments are chiefly nonmarine clastics except in part of North Carolina, where they are marginal marine. Pleistocene ice sheets advanced as far south as Long Island, leaving the island covered by glacial till and outwash deposits.

HYDROGEOLOGIC UNITS

The sediments of the northern Atlantic Coastal Plain aquifer system have been subdivided into 11 regional aquifers and 9 regional confining units on the basis of sediment geometry and the permeability contrasts between the sediments (Trapp, in press). Table 1 shows the hydrogeologic nomenclature for aquifers and confining units used in this and the other chapters of Professional Paper 1404. For simulation of flow, the aquifers were assigned to layers in the digital computer model. The correspondence of aquifers and model layers is also shown in the table.

In the following discussion, the values for vertical hydraulic conductivity of the confining units and, in part, transmissivity of the aquifers were derived from model calibration. Model-derived values for areas where the confining units are absent (sand-on-sand contacts) and

TABLE 1.—*Regional and subregional hydrogeologic units and their relation to layers used in simulation models*

Layer number used in model	Regional units (this report)	Winner and Coble (1989) North Carolina	Meng and Harsh (1988) Virginia	Vroblesky and Fleck (in press) Maryland-Delaware	Zapecza (1989) New Jersey	Adapted from Getzen (1977) Long Island, NY
10.	Surficial aquifer	Surficial aquifer	Columbia aquifer	Surficial aquifer	Holly Beach aquifer	Upper glacial aquifer[1]
	Confining unit	Confining unit	Yorktown confining unit	Upper Chesapeake confining unit	Cape May confining unit	
9.	Upper Chesapeake aquifer	Yorktown aquifer	Yorktown-Eastover aquifer	Upper Chesapeake aquifer	Cape May confining unit	
	Confining unit	Confining unit	St. Marys confining unit	St. Marys confining unit	Confining unit overlying the Rio Grande water-bearing zone	
8.	Lower Chesapeake aquifer	Pungo River aquifer	St. Marys–Choptank aquifer	Lower Chesapeake aquifer	Lower Kirkwood–Cohansey and confined Kirkwood aquifers	
	Confining unit	Confining unit	Calvert confining unit	Lower Chesapeake confining unit	Basal Kirkwood confining unit	
7.	Castle Hayne–Piney Point aquifer	Castle Hayne aquifer	Chickahominy–Piney Point aquifer	Piney Point–Nanjemoy aquifer	Piney Point aquifer	
	Confining unit	Confining unit	Nanjemoy-Marlboro Clay confining unit	Nanjemoy-Marlboro confining unit	Vincentown-Manasquan confining unit	
6.	Beaufort-Aquia aquifer	Beaufort aquifer	Aquia aquifer	Aquia-Rancocas aquifer	Vincentown aquifer	
	Confining unit	Confining unit		Lower Brightseat confining unit	Navesink-Hornerstown confining unit	Gardiners Clay confining unit
5.	Peedee-Severn aquifer	Peedee aquifer		Severn aquifer	Wenonah–Mount Laurel aquifer	Magothy aquifer, upper model layer
	Confining unit	Confining unit		Severn confining unit	Marshalltown-Wenonah confining unit	
4.	Black Creek–Matawan aquifer	Black Creek aquifer		Matawan aquifer	Englishtown aquifer system	Magothy aquifer, middle model layer
3. NORTH:						
	Confining unit			Matawan confining unit	Merchantville-Woodbury confining unit	
	Magothy aquifer			Magothy aquifer	Upper Potomac-Raritan-Magothy aquifer	Magothy aquifer, lower model layer
SOUTH:						
	Confining unit	Confining unit	Brightseat and upper Potomac confining units	Upper Brightseat confining unit		
	Upper Potomac aquifer	Upper Cape Fear aquifer	Brightseat and upper Potomac aquifers	Brightseat aquifer		
	Confining unit	Confining unit	Middle Potomac confining unit	Patapsco confining unit	Confining unit	Raritan Clay confining unit
2.	Middle Potomac aquifer	Lower Cape Fear aquifer	Middle Potomac aquifer	Patapsco aquifer	Middle Potomac-Raritan-Magothy aquifer	Lloyd aquifer
	Confining unit	Confining unit	Lower Potomac confining unit	Potomac confining unit	Confining unit	
1.	Lower Potomac aquifer	Lower Cretaceous aquifer	Lower Potomac aquifer	Patuxent aquifer	Lower Potomac-Raritan-Magothy aquifer	

[1]Upper glacial aquifer of Long Island is represented by model layers 7 and 6.

values for the offshore extensions of the aquifers and confining units have been excluded.

SURFICIAL AQUIFER

The surficial aquifer is the uppermost aquifer in the northern Atlantic Coastal Plain. It is unconfined, at least in its upper part. Its extent is shown in plate 3A, together with the altitude of the long-term averaged water table. The water table in plate 3A is mapped beyond the extent of the regional surficial aquifer to show its configuration in areas where other regional aquifers are exposed at the surface and are unconfined.

The surficial aquifer consists of unconsolidated sand and gravel and is heterogeneous in sorting and mineralogy. Its saturated thickness averages about 50 ft, except on Long Island (where it averages about 250 ft) and in buried channels on the Delmarva Peninsula (where it reaches as much as 220 ft) (Mack and Thomas, 1972; Johnston, 1973; Bachman, 1984, pl. 5). The sediments of the surficial aquifer range in age from Miocene (Owens and Denny, 1979) to Holocene. On Long Island, the surficial aquifer is composed of glacial sediments (the upper glacial aquifer of local use), consisting of moraines, outwash, and glaciolacustrine deposits, which cover most of the surface (McClymonds and Franke, 1972, p. E13–E15, table 1, pl. 1). The upper glacial aquifer has no lateral connection with the Coastal Plain aquifers of New Jersey.

On Long Island, the average transmissivity of the surficial aquifer is about 27,000 ft^2/d (McClymonds and Franke, 1972, table 6). Outside of Long Island, the transmissivity of the surficial aquifer generally is less than 1,000 ft^2/d, except on the Delmarva Peninsula and the southern tip of New Jersey, where it is commonly on the order of 8,000 ft^2/d and is as much as 53,000 ft^2/d in buried channels on the Delmarva Peninsula (Weigle, 1972, p. 86).

UPPER CHESAPEAKE AQUIFER AND ITS OVERLYING CONFINING UNIT

The upper Chesapeake aquifer consists of permeable beds in the upper part of the Chesapeake Group of Miocene and Pliocene age and their approximate stratigraphic equivalents. Over most of its extent, it is separated from the overlying surficial aquifer by a confining unit. From North Carolina through Delaware, the confining unit consists primarily of clay and silt of Miocene and Pliocene age. In New Jersey, it is restricted to the Cape May Peninsula and consists of estuarine clay of Pleistocene age. Its extent and thickness were shown by Trapp (in press). Its leakance, estimated by model simulation, ranges from 1×10^{-6} to 1×10^{-1} (ft/d)/ft (Leahy and Martin, in press).

The upper Chesapeake aquifer consists of sand and some gravel. In New Jersey, it includes the Cohansey Sand and sand beds in the upper part of the Kirkwood Formation, both of Miocene age, along with terrace gravels that have been mapped as Pleistocene but are considered Miocene by Owens and Denny (1979). The extent of the aquifer, the altitude of its top, and the distribution of its transmissivity are shown in plate 3B.

The thickness of the aquifer penetrated by wells averages about 75 ft in North Carolina, 140 ft in Virginia, 400 ft in Maryland and Delaware, and 190 ft in New Jersey. Transmissivity of the aquifer ranges up to about 6,000 ft^2/d in North Carolina, 24,000 ft^2/d in Maryland and Delaware, and 10,000 ft^2/d in New Jersey.

LOWER CHESAPEAKE AQUIFER AND ITS OVERLYING CONFINING UNIT

The lower Chesapeake aquifer consists principally of shelly sand beds in the lower part of the Chesapeake Group of Oligocene, Miocene, and Pliocene age from Virginia to Delaware, phosphatic sand beds in the Pungo River Formation of North Carolina, and their approximate equivalents in New Jersey. Over most of its extent, it is overlain by a confining unit consisting primarily of clay of Miocene age. The extent and thickness of the confining unit were shown by Trapp (in press). Its leakance, estimated by model simulation, ranges from 1×10^{-6} to 1×10^{-3} (ft/d)/ft (Leahy and Martin, in press). The higher values of leakance are found where the unit is thin.

Locally, the aquifer includes beds of limestone in eastern North Carolina and some gravel in Maryland and Delaware. In New Jersey, it includes sand interbedded with gravel in the lower, or confined, part of the Kirkwood Formation of Miocene age. The aquifer is absent in Long Island and the western shore of Virginia and Maryland. Its extent, the altitude of its top, and its distribution of transmissivity are shown in plate 4A.

The thickness of the aquifer penetrated by wells averages about 50 ft in North Carolina, 275 ft on the Delmarva Peninsula, and 200 ft in New Jersey. Its transmissivity ranges up to about 8,000 ft^2/d in North Carolina and 10,000 ft^2/d in New Jersey.

CASTLE HAYNE–PINEY POINT AQUIFER AND ITS OVERLYING CONFINING UNIT

The Castle Hayne–Piney Point aquifer is mostly Eocene to Oligocene in age. Locally, it includes sands of early Miocene age. It extends from North Carolina to New Jersey. Throughout most of its extent, it is overlain by a confining unit consisting primarily of clay of Miocene age. The extent and thickness of the confining unit were shown by Trapp (in press). Its leakance, estimated by model simulation, generally ranges from 1×10^{-7} to 1×10^{-4} (ft/d)/ft (Leahy and Martin, in press).

The aquifer consists of limestone, sandy marl, and limey sand in North Carolina and glauconitic sand with shells from Virginia to New Jersey. It is absent on Long Island, along the western margin of the Coastal Plain, and in the southwestern part of the North Carolina Coastal Plain. Its extent, the altitude of its top, and the distribution of its transmissivity are shown in plate 4B.

The thickness of the aquifer penetrated by wells averages about 185 ft in North Carolina, 60 ft in Virginia, 150 ft in Maryland and Delaware, and 125 ft in New Jersey. Transmissivity generally ranges up to as much as 100,000 ft^2/d in North Carolina and 5,000 ft^2/d from Virginia to New Jersey.

BEAUFORT-AQUIA AQUIFER AND ITS OVERLYING CONFINING UNIT

The Beaufort-Aquia aquifer is made up of permeable beds of Paleocene age. It is separated from the overlying Castle Hayne–Piney Point aquifer by a confining unit consisting primarily of silt, clay, and sandy clay of Paleocene and Eocene age. The extent and thickness of the confining unit were shown by Trapp (in press). Its leakance, estimated by model simulation, generally ranges from 1×10^{-8} to 1×10^{-5} (ft/d)/ft (Leahy and Martin, in press).

The Beaufort-Aquia aquifer consists primarily of fine- to coarse-grained glauconitic sand with shell beds. It extends northward from the central North Carolina Coastal Plain but covers only a narrow strip in Delaware and New Jersey and is absent on Long Island, N.Y. The extent of the aquifer, the altitude of its top, and the distribution of its transmissivity are shown in plate 5A.

The thickness of the aquifer penetrated by wells averages about 90 ft in North Carolina, 45 ft in Virginia, 120 ft in Maryland and Delaware, and 70 ft in New Jersey. Its transmissivity ranges up to about 5,000 ft^2/d but is typically about 1,000 ft^2/d.

PEEDEE-SEVERN AQUIFER AND ITS OVERLYING CONFINING UNIT

The Peedee-Severn aquifer constitutes the uppermost regional aquifer of Cretaceous age in the area. It is restricted principally to New Jersey, the northern part of the Delmarva Peninsula, and the eastern part of the North Carolina Coastal Plain. The aquifer is overlain by a confining unit consisting mainly of clay, silt, and glauconitic sand of Late Cretaceous and Tertiary age. Model-derived leakance generally ranges from 1×10^{-6} to 1×10^{-3} (ft/d)/ft (Leahy and Martin, in press).

The extent and thickness of the confining unit were shown by Trapp (in press), who extended it to include the confining unit overlying the Magothy aquifer on Long Island, N.Y. This segment is discussed further in the section on the Magothy aquifer.

The Peedee-Severn aquifer consists primarily of fine- to medium-grained glauconitic quartz sand. Its extent, the altitude of its top, and the distribution of its transmissivity are shown in plate 5B. On Long Island, beds that are stratigraphically equivalent to the Peedee-Severn aquifer of New Jersey are part of the Magothy aquifer.

The thickness of the aquifer penetrated by wells averages about 95 ft in North Carolina, 80 ft in both Maryland and New Jersey, and 100 ft in Delaware. Transmissivity of the freshwater part of the aquifer ranges up to about 10,000 ft^2/d in North Carolina and less than 2,000 ft^2/d from New Jersey to Maryland.

BLACK CREEK–MATAWAN AQUIFER AND ITS OVERLYING CONFINING UNIT

The Black Creek–Matawan aquifer is Late Cretaceous in age and present principally in New Jersey and North Carolina. It is overlain by a confining unit consisting mainly of clay and silt of Late Cretaceous age. The extent and thickness of the confining unit were shown by Trapp (in press). Its leakance, estimated by model simulation, generally ranges from 1×10^{-7} to 1×10^{-4} (ft/d)/ft and typically is 1×10^{-6} to 1×10^{-5} (ft/d)/ft (Leahy and Martin, in press).

The Black Creek–Matawan aquifer consists primarily of fine- to medium-grained sand. The sand is commonly glauconitic and interbedded with clay. On Long Island, beds that are stratigraphically equivalent to the Black Creek–Matawan form part of the Magothy aquifer. The extent of the aquifer, the altitude of its top, and the distribution of its transmissivity are shown in plate 6A.

The thickness of the aquifer penetrated by wells averages 180 ft in North Carolina and 55 ft in New Jersey. The aquifer is thin in Delaware and Maryland; data are insufficient to determine the average thickness. Transmissivity of its freshwater part ranges up to about 10,000 ft^2/d in North Carolina but is generally less than 2,000 ft^2/d elsewhere in the study area.

MAGOTHY AQUIFER AND ITS OVERLYING CONFINING UNIT

The Magothy aquifer is of Late Cretaceous age and extends from Maryland to Long Island, N.Y. It is overlain by a confining unit consisting mainly of silt and clay of Late Cretaceous age, except on Long Island, where it consists mainly of the Pleistocene Gardiners Clay and undifferentiated clay, silt, and till. The extent and thickness of the overlying confining unit were shown by Trapp (in press), where it was treated from Maryland through New Jersey as the northern part of a regional hydrogeologic unit that includes the confining unit overlying the upper Potomac aquifer to the south. The confining unit that separates the Magothy aquifer from the overlying unconfined upper glacial aquifer on Long Island was treated by Trapp (in press) as an extension of the regional confining unit overlying the Peedee-Severn aquifer.

South of Long Island, the leakance of the confining unit overlying the Magothy aquifer, estimated by model simulation, generally ranges from 1×10^{-7} to 1×10^{-4} (ft/d)/ft but is typically on the order of 1×10^{-6} (ft/d)/ft (Leahy and Martin, in press). On Long Island, the leakance generally ranges from 1×10^{-4} to 1×10^{-2} (ft/d)/ft.

The Magothy aquifer consists principally of well-stratified to cross-bedded, very fine- to coarse-grained quartz sand and gravel. The extent of the aquifer, the altitude of its top, and the distribution of its transmissivity are shown in plate 6B.

The thickness of the aquifer penetrated by wells averages about 75 ft in Maryland and Delaware, 100 ft in New Jersey, and 460 ft on Long Island. Transmissivity of the freshwater section ranges up to about 6,000 ft^2/d in Maryland, 10,000 ft^2/d in New Jersey, and 56,000 ft^2/d on Long Island.

UPPER POTOMAC AQUIFER AND ITS OVERLYING CONFINING UNIT

The upper Potomac aquifer is of Late Cretaceous age in North Carolina but includes beds of Cretaceous and possibly Paleocene age in Virginia and Maryland.[1] It is overlain by a confining unit consisting mainly of clay beds of Late Cretaceous age in North Carolina and southern Virginia and of micaceous silty clay of Cretaceous and Paleocene age in northern Virginia and southern Maryland. The extent and thickness of the confining unit were shown by Trapp (in press), where it was treated as the southern part of a regional hydrogeologic unit that included the confining unit overlying the Magothy aquifer to the north.

The leakance of the confining unit, estimated by model simulation, generally ranges from 1×10^{-7} to 1×10^{-3} (ft/d)/ft but is typically 1×10^{-6} to 1×10^{-5} (ft/d)/ft (Leahy and Martin, in press).

The upper Potomac aquifer consists primarily of fine- to medium-grained quartz sand and interbedded silty clay. The sand is partially glauconitic. Southward from its northeastern limit, the aquifer extends throughout all but the western fringe of the Coastal Plain. Its extent, the altitude of its top, and the distribution of its transmissivity are shown in plate 6B.

The thickness of the aquifer penetrated by wells averages about 160 ft in North Carolina and 95 ft in Virginia. Transmissivity of the freshwater section ranges up to about 6,000 ft^2/d in North Carolina and 2,000 ft^2/d in Virginia.

MIDDLE POTOMAC AQUIFER AND ITS OVERLYING CONFINING UNIT

The middle Potomac aquifer includes beds of both Early and Late Cretaceous age. The aquifer extends over most of the northern Atlantic Coastal Plain; it is absent in the southwestern part of the Coastal Plain and along its western border. The middle Potomac aquifer is overlain by a confining unit consisting mainly of clayey sediments of Late Cretaceous age. Locally, however, the aquifer is overlain by beds ranging in age from Early Cretaceous to Pleistocene. The extent and thickness of the confining unit were shown by Trapp (in press). Its leakance, estimated by model simulation, generally ranges from 1×10^{-7} to 1×10^{-4} (ft/d)/ft (Leahy and Martin, in press). The higher values are found in updip areas, where it generally is thin.

The middle Potomac aquifer consists primarily of fine- to coarse-grained sand and some gravel interbedded with silt and clay. The extent of the aquifer, the altitude of its top, and the distribution of its transmissivity are shown in plate 7A.

The thickness of the aquifer penetrated by wells averages 285 ft in North Carolina, 350 ft in Virginia, 770 ft in Maryland and Delaware, 245 ft in New Jersey, and 225 ft on Long Island. Transmissivity of the freshwater section ranges up to about 8,000 ft^2/d in North Carolina; 16,000 ft^2/d in Virginia, Maryland, Delaware, and New York; and 21,000 ft^2/d in New Jersey.

LOWER POTOMAC AQUIFER AND ITS OVERLYING CONFINING UNIT

The lower Potomac aquifer is the lowermost aquifer delineated in the northern Atlantic Coastal Plain. It is overlain by a confining unit consisting primarily of clay, sandy clay, and silt of Early Cretaceous and early Late Cretaceous age. The extent and thickness of the confining unit were shown by Trapp (in press). Its leakance,

[1]The local Brightseat aquifer, which is part of the regional upper Potomac aquifer in northern Virginia and southern Maryland, has been shown in the chapters of Professional Paper 1404 to be correlative with the Paleocene Brightseat Formation. Recent work on cores from two test holes, one in northern Virginia and the other in southern Maryland, has identified fossil pollen and spores of late Early Cretaceous (Albian) age (D.J. Nichols, U.S. Geological Survey, written commun., 1985; Ronald Litwin, U.S. Geological Survey, written commun., 1987) in deposits designated the Brightseat aquifer in this report. This finding indicates that the Brightseat aquifer does not correlate with the Brightseat Formation and therefore that the regional upper Potomac aquifer does not include Paleocene sediments.

estimated by model simulation, ranges from 1×10^{-8} to 1×10^{-4} (ft/d)/ft (Leahy and Martin, in press).

The aquifer consists principally of lenses of medium- to coarse-grained quartz sand with some gravel interbedded with lenses of clay and silt. It is predominantly Early Cretaceous in age but may include beds of Jurassic age. It extends from New Jersey to east-central North Carolina but is absent from Long Island and from the southern and western parts of the Coastal Plain of North Carolina. The extent of the aquifer, the altitude of its top, and the distribution of its transmissivity are shown in plate 7B.

The thickness of the aquifer penetrated by wells averages 285 ft in North Carolina, 525 ft in Virginia, 935 ft in Maryland and Delaware, and 345 ft in New Jersey. Transmissivity of its freshwater part ranges up to about 8,000 ft^2/d in Virginia, 6,000 ft^2/d in Maryland, 5,000 ft^2/d in Delaware, and 10,000 ft^2/d in New Jersey.

REGIONAL GROUND-WATER FLOW

This study focused primarily on the regional ground-water component of the hydrologic cycle, that is, ground water entering confined aquifers and flowing tens of miles prior to discharging to bodies of surface water or wells or into other aquifers. This restriction resulted from the original objectives of the study (Meisler, 1980a, p. 10) and from the coarse grid size (7 × 7 mi) of the digital flow model used for analysis of the regional ground-water flow system. The model simulations did not consider movement of shallow ground water along short flow paths to nearby streams or losses of ground water to evapotranspiration.

PREPUMPING CONDITIONS

Under prepumping conditions, most recharge of ground water to the confined aquifers occurred at topographically high areas and discharge at lows. Ground-water recharge entered the upper, unconfined part of the system, which includes the outcrops of the deeper aquifers as well as surficial aquifers that are everywhere unconfined. Under prepumping conditions, most of the recharge moved a few miles or less to discharge at streams. Over most of the Coastal Plain, only a small percentage, probably less than 10 percent, of the total recharge entered the regional flow system. Larger percentages of the recharge reached deeper aquifers where they underlie unconfined aquifers, for example, the Magothy aquifer on Long Island where it directly underlies the upper glacial aquifer, the Kirkwood aquifer in New Jersey where it underlies the unconfined Cohansey aquifer, and the Castle Hayne aquifer in North Carolina where it directly underlies the surficial aquifer.

The effects of topography on the flow system are greatest where topographic relief is greatest. In these areas, most ground water discharges a short distance from its recharge area (fig. 2). Figure 2 shows patterns of ground-water flow prior to pumping from the Fall Line to the limit of the freshwater flow system offshore along a generalized section. The velocity of flow is greatest along local flow paths and least in regional flow. The effects of topography on flow diminish with depth. Back (1966, figs. 4, 3) published diagrammatic sections showing local, intermediate, and regional patterns of vertical flow in the northern Atlantic Coastal Plain and also horizontal flow paths in the Cretaceous sediments. Back derived the latter from a conductive-paper electric-analog model.

Although recharge to the water table is widespread, flow from the unconfined to the confined aquifers or vice versa (referred to in this study as deep percolation) tends to be concentrated in certain areas. The locations of these areas may be inferred from the configuration of potentiometric surfaces of the confined aquifers, especially from closed contours delineating highs or lows. The mapping of potentiometric surfaces may be refined, as in this study, by simulation of the flow system.

At the northeastern end of the study area, on Long Island, the major flow pattern of the regional aquifer system under natural conditions involved recharge around the topographic divide along the middle of the island and discharge along the northern and southern shores and into rivers, Long Island Sound, and the Atlantic Ocean. In New Jersey, recharge to confined aquifers under prepumping conditions was concentrated in an area just southeast of the Fall Line between Trenton and Raritan Bay and at two topographic highs in southern and central New Jersey. However, the Delaware River Valley south of Trenton was a major natural discharge area, although it is close to the Fall Line. The aquifer system also discharged to other rivers and estuaries and offshore into the Atlantic Ocean and its bays.

In Maryland west of Chesapeake Bay and on the northern Delmarva Peninsula, prepumping recharge to confined aquifers was concentrated in a band just southeast of the Fall Line, interrupted by discharge areas where rivers (typically with steep bluffs) cross the Fall Line and also by the Delaware and Chesapeake Bays. There the flow lines were short, with recharge on the bluff tops and discharge to the bodies of water below. On the central and southern Delmarva Peninsula, prepumping recharge to confined aquifers was concentrated along the Chesapeake-Atlantic drainage divide, and discharge took place in rivers and creeks, through the floors of the ocean and bays, and in marshes fringing the ocean and bays.

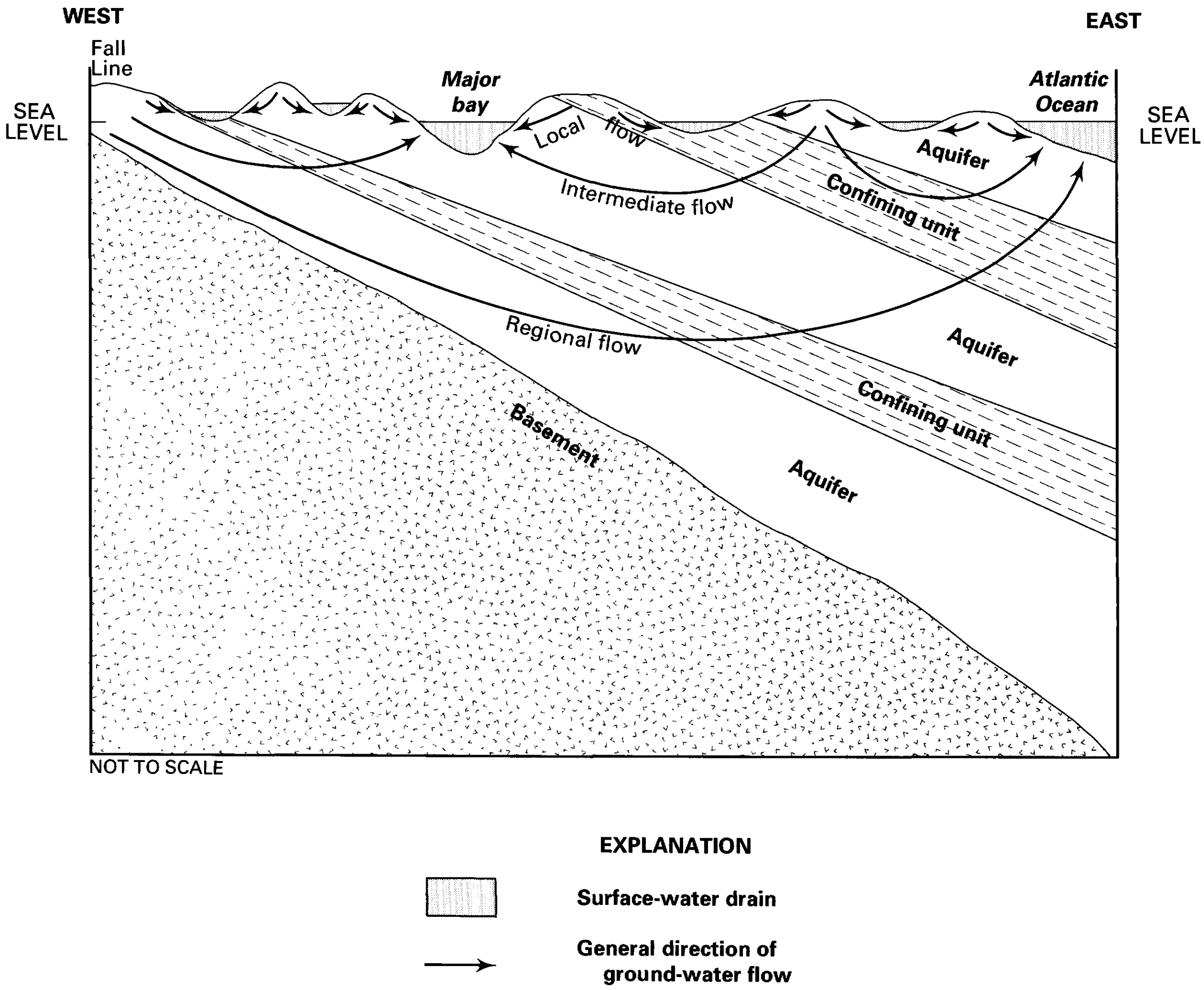


FIGURE 2.—Generalized hydrogeologic section of the northern Atlantic Coastal Plain showing an idealized prepumping flow system (adapted from Leahy and Martin, in press).

In Virginia and North Carolina, topography exerted a similar effect on the prepumping ground-water flow system as that in the Maryland Coastal Plain west of Chesapeake Bay. Numerous rivers flow across the aquifer outcrops and are, in many cases, bordered by high bluffs, especially near the Fall Line. The aquifer system was recharged on the bluffs, while the river valleys were major discharge areas. The ocean and its bays, such as Albemarle and Pamlico Sounds, also were areas of ground-water discharge.

Ground-water flow is controlled by the distribution of the rock materials that comprise the framework of the flow system and their hydraulic properties, as well as by topography. In a sedimentary wedge, such as the Atlantic Coastal Plain, in which the rock materials are unmetamorphosed and only slightly deformed structurally, the principal direction of permeability tends to follow beds of sand, gravel, or limestone, which in turn run approximately parallel to the upper and lower boundaries of formations. The formations and therefore the aquifers and confining units occur in gently dipping layers, as described in the "Hydrogeologic Units" section.

Prepumping water-level measurements are sparse. However, measurements are available of water levels that approach prepumping levels, including measurements made (1) early in the period of ground-water development and (2) at some distance from a pumping center, as well as measurements (3) of levels in deeper aquifers when first tapped and (4) of recovered water levels. These data were used as controls for calibrating the flow models for steady-state prepumping conditions. Prepumping potentiometric surfaces are shown in plate 3A for the surficial water-table aquifer and the outcropping parts of other aquifers. The simulated prepumping potentiometric surfaces of the Castle Hayne–Piney Point (fig. 4) and upper Potomac and Magothy aquifers (fig. 5) are representative of the upper and lower parts, respectively, of the confined aquifer system.

EFFECTS OF GROUND-WATER DEVELOPMENT ON WATER LEVELS

Withdrawal of ground water began with hand-dug wells in the earliest period of settlement. The drilling of artesian wells in the Middle Atlantic States, probably including the Coastal Plain, began about 1827 (Trapp, in press). Ground-water withdrawals (excluding individual domestic systems and most irrigation) in the northern Atlantic Coastal Plain have increased from about 100 Mgal/d in 1900 to about 1,200 Mgal/d in 1980. More than half of the pumping in 1900 was on Long Island, N.Y. (mostly from the western part of the island). In 1980, more than one-third of the pumping was on Long Island. Although pumping was substantially curtailed in the western part of Long Island because of saltwater intrusion, eastward suburban development on the island created an increasing demand for water beginning in the 1950's. Industrial and suburban growth has led to a rapid increase in withdrawals in New Jersey from the 1940's to the present. Pumpage in Delaware, Maryland, and Virginia has increased fairly steadily from the 1930's. A major increase in pumpage began in the 1960's in North Carolina because of large-scale phosphate mining (Leahy and Martin, in press).

The pumpage figures used in this study are more reliable for 1965 through 1980 than are figures for earlier years, which are partly estimated. Since 1965, most States within the study area have mandated the collection of pumpage data.

For modeling purposes, the period 1900–80 has been divided into 10 pumping periods. Average simulated pumping rates for each period are shown in figure 3. The simulated pumping does not include rural domestic supply or most irrigation. Although locally large, as on the Delmarva Peninsula, withdrawals for irrigation were not compiled consistently for this study. (Some irrigation pumpage in New Jersey and Virginia was compiled for model input.) It was assumed that most irrigation withdrawals were from the surficial aquifer or from unconfined parts of other aquifers, that most of the water pumped was recharged into the ground-water flow system as irrigation return flow, and that the pumping had little effect on heads in the underlying confined aquifers.

According to the data compiled for this study (Leahy and Martin, in press), the most heavily pumped regional aquifers as of 1980 in the northern half of the area from Long Island through Maryland were the lowermost three (the lower and middle Potomac and Magothy aquifers) and the surficial aquifer (the upper glacial aquifer) of Long Island. In Virginia, the lower, middle, and upper Potomac aquifers are the major aquifers. The Castle Hayne–Piney Point aquifer is most heavily pumped in North Carolina, followed by the Black Creek–Matawan aquifer.

Major cones of depression, resulting from large ground-water withdrawals for industrial and public supply, occur in New Jersey, on the Delmarva Peninsula, and in southern Virginia. Coalescing cones of depression underlie virtually the entire 4,000-mi^2 Coastal Plain of New Jersey, especially in the lower Chesapeake aquifer (locally, the Kirkwood aquifer), Peedee-Severn aquifer (locally, the Englishtown aquifer), and Black Creek–Matawan aquifer (locally, the Wenonah–Mount Laurel aquifer), and the Magothy and middle and lower Potomac aquifers (locally, the Potomac-Raritan-Magothy aquifer system). On the northern Delmarva Peninsula, coalescing cones of depression underlie 3,000 mi^2, especially in the lower Chesapeake aquifer, the Castle Hayne–Piney Point aquifer, the Beaufort-Aquia aquifer, and the middle and lower Potomac aquifers. Water-level declines are typically 80–200 ft at the centers of these cones. In southern Virginia and adjacent North Carolina, water levels have declined as much as 200 ft near the center of a 5,000-mi^2 cone of depression as a result of industrial pumping from the upper, middle, and lower Potomac aquifers.

Dewatering associated with phosphate mining has created a cone of depression extending over an area of more than 2,000 mi^2 in the Castle Hayne–Piney Point aquifer in northeastern North Carolina. Water levels have declined as much as 150 ft at the center of the cone.

SIMULATION OF GROUND-WATER FLOW

DIGITAL FLOW MODELS

Ground-water flow in the northern Atlantic Coastal Plain was simulated using a multilayer finite-difference model. The model is a modification (Leahy, 1982) of the Trescott (1975) computer code. The objective of the modeling analysis is to improve the understanding of the flow system and to assess the effects of ground-water development on the system. The model also may be used to evaluate the effects of projected region-scale withdrawals on the flow system.

In addition to a regional flow model of the area from North Carolina to Long Island, four subregional flow models were developed. Each of these covers the Coastal Plain of one or more regions: (1) North Carolina, (2) Virginia, (3) New Jersey, and (4) Delaware, Maryland, and the District of Columbia. The boundaries of the regional and subregional models are shown in plate 8.

The subregional models use a finer grid than that used in the regional model. In all the models except the New Jersey subregional model, the grid directions are aligned with those of the regional model, and generally, four subregional grid blocks (not shown in plate 8) correspond to one regional grid block. A typical block size is 49 mi^2

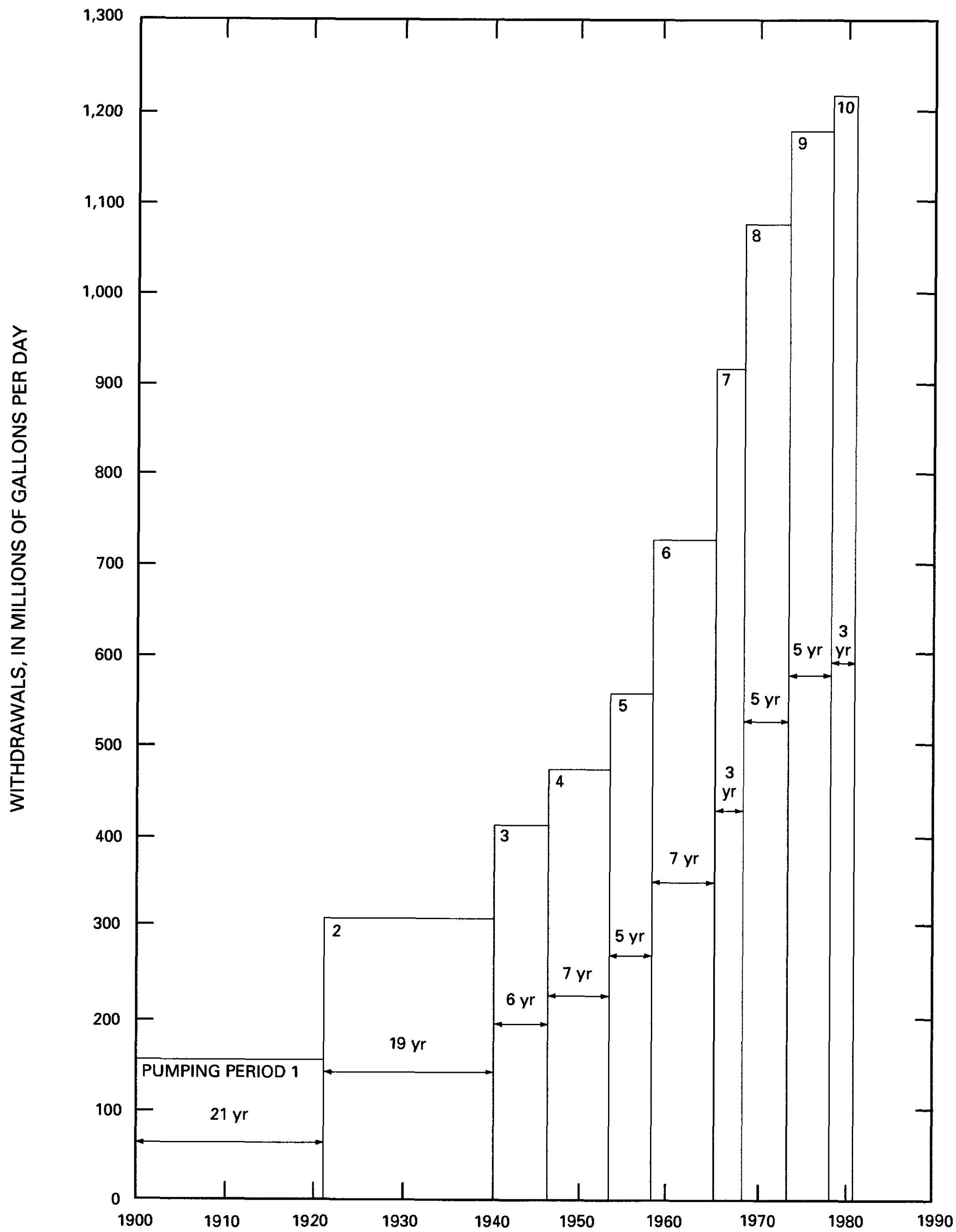


FIGURE 3.—Average simulated ground-water withdrawals for each of 10 pumping periods (from Leahy and Martin, in press).

for the regional model and 12.25 mi^2 for the subregional models. The New Jersey subregional-model grid is aligned with the Delaware estuary to simulate the interaction between the estuary and the ground-water system. Its typical model-block size is 6.25 mi^2.

MODEL BOUNDARIES

The boundaries of the computer models were selected to approximate the hydrogeologic boundaries of the Coastal Plain aquifer system. Hence, the contact of the Coastal Plain sediments with the underlying older rocks of much lower permeability and the updip limit of the sediments were simulated by no-flow boundaries. Downdip, the 10,000-mg/L chloride concentration, mapped and interpreted by Meisler (1989) as a saltwater-freshwater interface, was assumed to be another no-flow boundary. The southwestern boundary of the model coincides with a major ground-water sink, the Pee Dee River in South Carolina. There the boundary of the upper layer was treated as a series of river nodes. All the aquifers underlying the river, except for the lowermost (the middle Potomac aquifer), were terminated at a no-flow boundary. The middle Potomac aquifer was simulated with a specified-head boundary because of subsurface inflow from the south.

The northeastern no-flow boundary was set at the interface of 10,000-mg/L chloride concentration in ground water in the submerged Coastal Plain aquifer system northeast of Long Island, N.Y.

The upper boundary of the model initially was simulated as a specified-head boundary representing average water-table altitudes for prepumping steady-state conditions. Vertical fluxes were computed from this simulation and used to calculate the leakance of streambeds. To verify the computed streambed leakances, a second conceptualization of steady-state conditions was simulated that consisted of specified areal recharge (10–22.5 in/yr, depending on the area) to the water-table aquifer (specified flux) and of streams with the computed streambed leakances as specified-head drains. The simulated water table from this second conceptualization was compared with average long-term water-table altitudes. Finally, with the computed streambed and known areal recharges, the effect on water-table conditions by pumping can be simulated.

The construction and calibration of the regional and subregional models, input data, and boundary conditions are discussed in detail in chapters of Professional Paper 1404 by Leahy and Martin (in press), Harsh and Laczniak (1990), Martin (1990), Fleck and Vroblesky (in press), and Giese and others (in press).

RESULTS OF STEADY-STATE PREPUMPING SIMULATION

The prepumping potentiometric surfaces for the confined regional aquifers were simulated by the calibrated model with no pumping or change in storage (Leahy and Martin, in press). Observed prepumping water-level values are too sparse and unevenly distributed over the area for reliable contouring; however, in general, the simulation agrees with the available observed water levels and with the conceptualized prepumping potentiometric surface. The most complete water-level data are for Long Island and New Jersey.

The potentiometric surface for the Castle Hayne–Piney Point aquifer has been chosen to represent the upper part of the regional aquifer system (figs. 4,5). Figure 5 shows the potentiometric surfaces for the upper Potomac and Magothy aquifers and for the lowermost part of the Magothy aquifer on Long Island, which represent flow conditions in the lower part of the regional aquifer system.

The potentiometric surfaces of the upper part of the regional aquifer system are more greatly influenced by local flow systems than are the potentiometric surfaces of the lower part. The influence of major rivers, estuaries, and embayments is apparent in the upper part of the system. In contrast, the potentiometric surfaces of the lower part of the regional aquifer system show the predominance of regional flow. Although streams and rivers affect flow patterns near the outcrops of the lower aquifers, only the larger surface-water bodies have a discernible influence on flow in the same aquifers downdip. Surface-water bodies that affect flow in the deeper aquifers include the Raritan, Delaware, and Chesapeake Bays. In the deeper aquifers, ground-water flow paths are several tens of miles long and commonly cross State boundaries. The low hydraulic gradients indicate that prepumping flow velocities were sluggish (less than 1 ft/yr) in these aquifers.

Comparisons among the simulated prepumping potentiometric surfaces of the various aquifers show that the confined aquifers were recharged principally near the Fall Line and that water was discharged from the deeper aquifers by upward leakage through confining beds into the ocean, bays, and estuaries.

Simulated flow from water-table aquifers into confined aquifers ranges up to 20 in/yr in the recharge areas, and upward discharge into the water-table aquifers is as much as 16 in/yr when the water table is represented by a specified-head boundary. Infiltration rates are high in areas where the water-table aquifer is underlain by another aquifer with no intervening confining unit, for example, the surficial aquifer (local upper glacial aquifer) underlain by the Magothy aquifer on Long Island, the Cohansey aquifer underlain by the Kirkwood aquifer in New Jersey, and the surficial aquifer underlain by the Castle Hayne aquifer in North Carolina.

Under prepumping conditions, simulated base flow in streams typically ranges from 10 to 15 in/yr. Notable exceptions are New Jersey and Long Island, where simulated base flow tends to be higher (15–25 in/yr). These figures are consistent with estimates of areal water budgets by Wilder and others (1978, p. 18), Cushing and others (1973, p. 35), and Sinnott and Cushing (1978, p. I15).

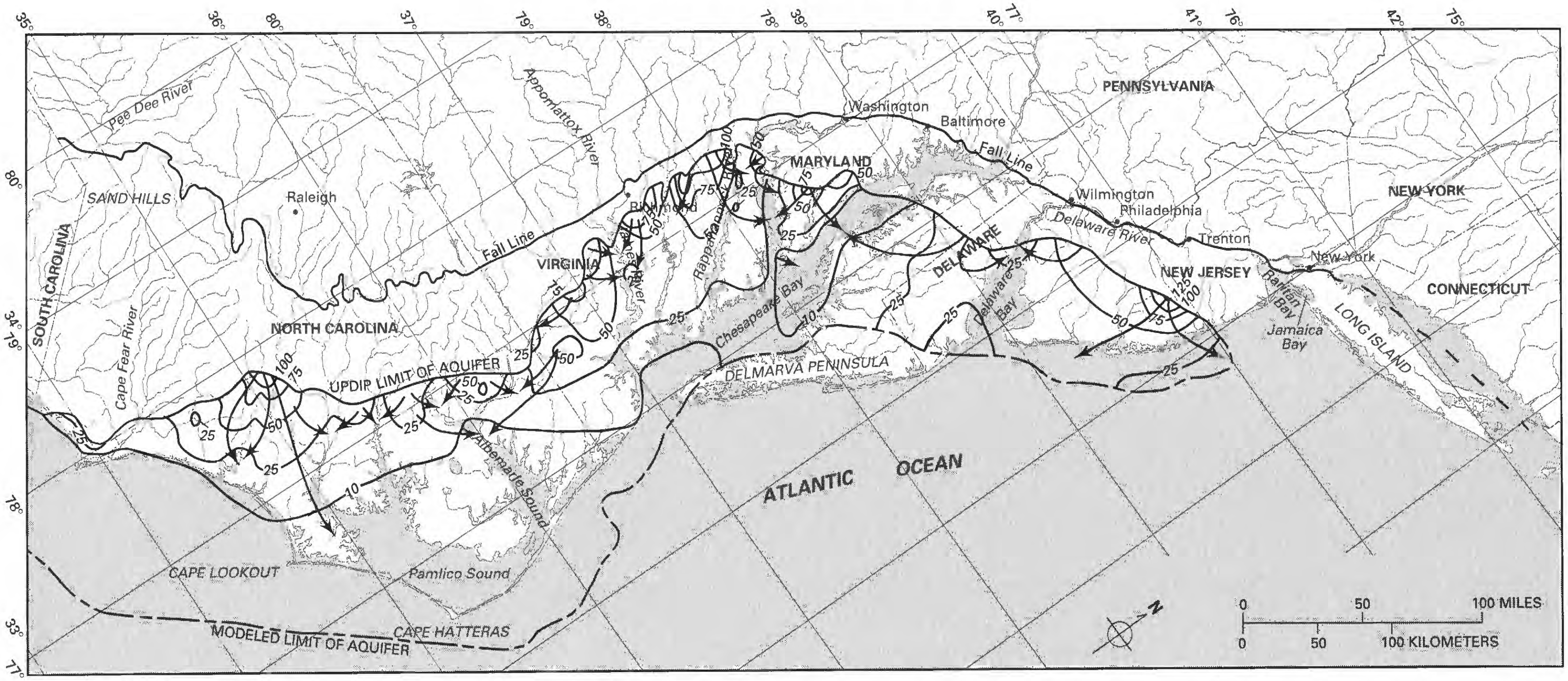
RESULTS OF TRANSIENT SIMULATION

Pumpage caused a major change in the ground-water flow system. By 1980, withdrawals (not including most

EXPLANATION

—25— **Potentiometric contour—**Shows altitude of simulated prepumping potentiometric surface. Contour interval, in feet, is variable. Datum is sea level

←— **Generalized direction of ground-water flow**



Base enlarged from U.S. Geological Survey National Atlas, 1970, 1:7,500,000

FIGURE 4.—Simulated prepumping potentiometric surface for the Castle Hayne–Piney Point aquifer (from Leahy and others, 1987, fig. 8).

EXPLANATION

—50— **Potentiometric contour**—Shows altitude of simulated prepumping potentiometric surface. Contour interval, in feet, is variable. Datum is sea level

←— **Generalized direction of ground-water flow**

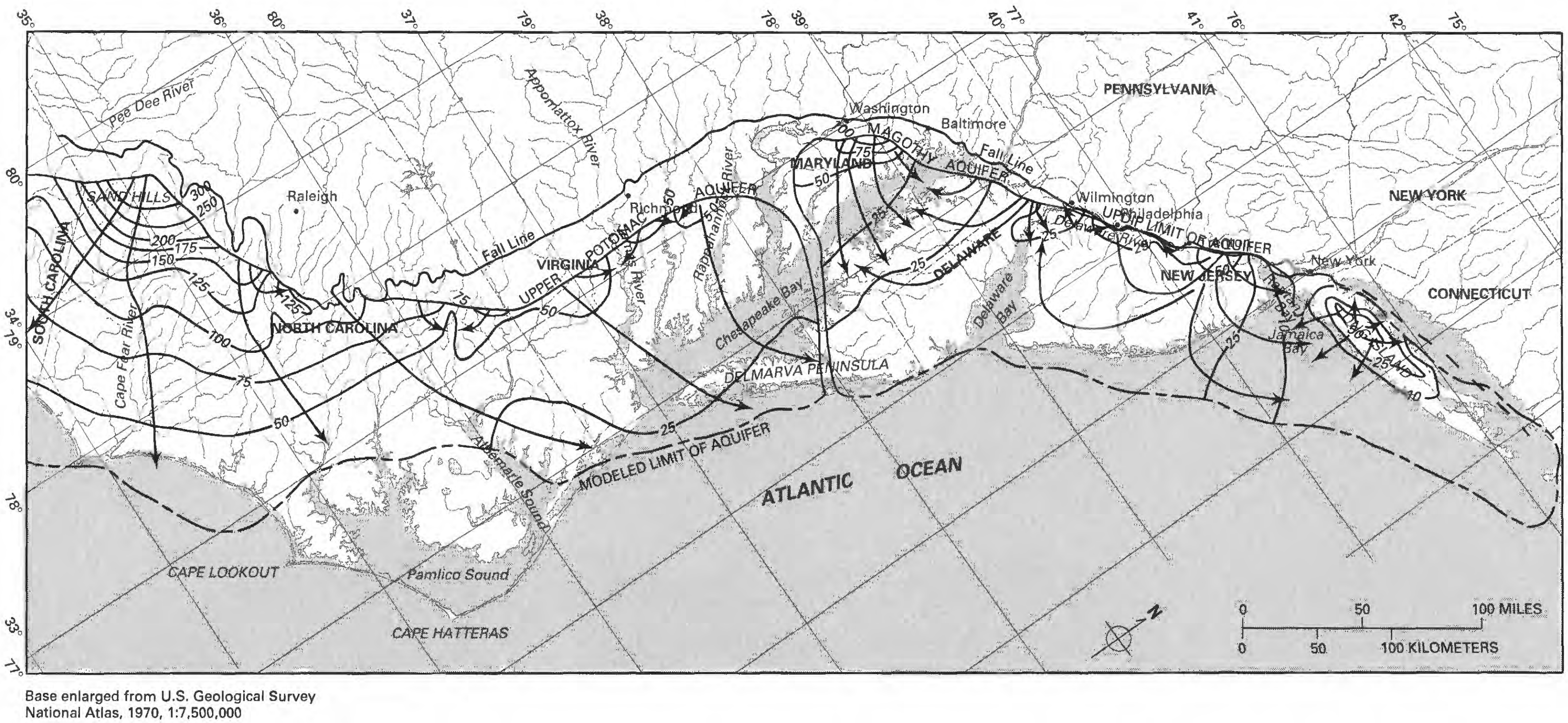


FIGURE 5.—Simulated prepumping potentiometric surface for the upper Potomac and Magothy aquifers and for the lowermost part of the Magothy aquifer on Long Island, N.Y. (from Leahy and others, 1987, fig. 9).

irrigation pumpage, as discussed under "Effects of Ground-Water Development on Water Levels," and rural domestic pumpage) were about 1,200 Mgal/d, or about 3 percent of the recharge to the unconfined parts of the aquifer system. Plates 9A and 10A show the simulated potentiometric surfaces of the Castle Hayne–Piney Point and upper Potomac and Magothy aquifers, respectively, under 1980 pumping conditions. Plates 9B and 10B show measured water levels for 1978–80 in the same aquifers. These aquifers represent the upper and lower parts of the regional aquifer system. Because of its coarser grid, the regional model does not show potentiometric surfaces in the same detail as the subregional models do. However, the regional model is based on the same data as the subregional models.

Sources of ground water contributing to pumpage were evaluated by simulating flow into and out of 10 large cones of depression, which are represented by hydrologic budget areas on plate 8. Their locations and associated aquifers are listed here.

Budget area (pl. 8)	Location	Aquifer	Layer used in model
1	Virginia–North Carolina border area	Lower Potomac	1
2	Virginia–North Carolina border area	Middle Potomac	2
3	Southern New Jersey	Middle Potomac	2
4	Southern Maryland and part of the Delmarva Peninsula	Upper Potomac and Magothy	3
5	Southern New Jersey	Magothy	3
6	Central New Jersey	Black Creek–Matawan	4
7	Parts of northern Virginia, southern Maryland, and the Delmarva Peninsula	Beaufort–Aquia	6
8	Central Delaware and adjoining parts of Maryland and New Jersey	Castle Hayne–Piney Point	7
9	Central North Carolina Coastal Plain	Castle Hayne–Piney Point	7
10	Southern New Jersey	Lower Chesapeake	8

The simulation results for 1980 are given in table 2 and suggest that (1) the volume of water coming from aquifer storage is small, (2) lateral flow into the cones of depression is a major source of the pumped water (primarily from the outcrop areas), and (3) vertical leakage constitutes the greatest volume of the water pumped. Although the rates of vertical leakage generally are much smaller than lateral flow rates, the areas of the top and bottom surfaces of aquifers in the cones of depression are much larger than the cross-sectional areas through which lateral flow occurs. Therefore, the total volume of vertical flow is much greater than that of lateral flow, indicating that vertical flow is an important factor for the evaluation of ground-water development, even though the rate of vertical leakage is small.

INTERPRETATION OF FLOW SYSTEM FROM SIMULATION

Under prepumping conditions, long-term areal ground-water recharge in the Coastal Plain was about 40,000 Mgal/d (15.4 in/yr), most of which discharged to the nearest surface-water body. About 592 Mgal/d, or 0.5 in/yr, flowed from the water-table aquifers into the underlying confined aquifers (fig. 6). This recharge occurred over about 25,000 mi^2, or 26 percent of the area of the model. The remaining 74 percent, which included part of the Continental Shelf, was an area of discharge for the deeper aquifers. Flow to the deeper aquifers accounted for only about 2 percent of the total recharge to the water-table aquifers.

Sources of the water pumped from confined aquifers in the Coastal Plain are (1) an increase in recharge from water-table aquifers to confined aquifers that reduces ground-water discharge from water-table aquifers to streams, (2) a slight reduction in water in storage, and (3) a reduction in upward discharge from the confined aquifers through overlying confining units to open surface-water bodies (fig. 6). The water-level declines caused by pumping have spread updip toward outcrop areas, and the cones of depression have diverted some of the water toward pumping centers that would otherwise have discharged to streams. Simulation results indicate that the aquifer system approaches steady-state conditions over most of its area within about 5 years at a constant pumping rate. Therefore, substantial quantities of water are released from storage, with an accompanying water-level decline, for only a relatively short period after an increase in pumpage. The area of recharge to deeper aquifers increased from about 26 percent of the model area under prepumping conditions to 45 percent under 1980 pumping conditions.

Ground-water flow has not only been redirected as a result of pumping, but its rate has also increased. The flow rate has increased especially in areas of cones of depression near the outcrop or subcrop areas of the stressed aquifers.

GROUND-WATER GEOCHEMISTRY

The solute content of ground water is determined by the minerals and gases with which the water has come into contact. Water entering the ground-water body contains oxygen and carbon dioxide, chemically active gases. Calcite, heavy minerals (especially iron compounds such as pyrite and glauconite), micas, the feld-

TABLE 2.—*Source of ground-water withdrawals for selected areas under 1980 pumping conditions (adapted from Leahy and Martin, in press)*

Budget area[1]	Model layer[2]	Percentage of withdrawals from: Change in aquifer storage 1978–1980	Lateral flow[3] through sides of areas: NW	NE	SW	SE	Total lateral flow[3]	Vertical flow[3]: from above	from below	Total vertical flow[3]	Withdrawals (Mgal/d)	Percent error
1	1	0.6	17.2	5.5	0.0	0.0	22.7	77.2	0.0	77.2	10.7	+0.5
2	2	0.3	39.6	2.3	9.8	0.9	52.6	68.2	−20.9	47.3	39.6	+0.2
3	2	1.3	65.8	1.8	−1.1	2.2	67.7	39.5	−7.5	32.0	16.5	+1.0
4	3	1.2	13.6	−2.9	−0.3	−12.9	−2.5	118.1	−15.3	102.8	4.31	+1.5
5	3	0.7	34.2	2.5	1.3	1.5	39.5	84.1	−24.0	60.1	27.1	+0.3
6	4	1.5	16.5	5.9	−0.1	0.8	23.1	117.8	−41.0	76.8	9.17	+1.4
7	6	1.9	14.5	−3.8	9.3	−1.0	19.0	110.0	−29.5	80.5	3.20	+1.4
8	7	3.6	0.7	30.8	−2.6	0.2	29.1	57.7	11.4	69.1	3.90	+1.8
9	7	0.0	9.2	−0.3	11.7	0.4	21.0	76.5	2.5	79.0	88.0	0.0
10	8	0.7	34.7	−12.2	4.7	2.97	30.2	70.5	−1.0	69.5	20.0	+0.4

[1]Location of area shown in plate 8.
[2]Regional aquifer names for the model layers are as follows:
8. Lower Chesapeake
7. Castle Hayne–Piney Point
6. Beaufort-Aquia
4. Black Creek–Matawan
3. Upper Potomac and Magothy
2. Middle Potomac
1. Lower Potomac

[3]Positive values represent flow into the area; negative values represent flow out of the area.

spars, unstable clay minerals, and lignitic material are among the more reactive solid substances in Coastal Plain sediments under the prevailing conditions of temperature and pressure.

EFFECTS OF DEPOSITIONAL ENVIRONMENT

Sediments derived directly from the erosion of crystalline rocks contain more minerals that are unstable in the sedimentary environment than sediments that have been reworked (Owens and others, 1983, p. F3). Thus, continental deposits derived from the erosion of crystalline Piedmont rocks have a higher percentage of feldspars, micas, and unstable clay and heavy minerals than do marine deposits. Oxidation and reduction reactions predominate. Marine sediments tend to be depleted in the above types of minerals but enriched in glauconite, phosphatic minerals, calcite, and silica. The more important geochemical processes controlling water chemistry in the marine sediments are ion exchange, carbonate dissolution-precipitation, incongruent dissolution of silicate minerals (where the silicate framework is restructured to a more stable clay mineral with excess ions carried off in solution), dissolution of amorphous silica, and mass transfer of phosphatic material.

CHEMICAL PROCESSES IN NONMARINE SEDIMENTS

Geochemical processes in a nonmarine Coastal Plain aquifer are exemplified in the lower Potomac aquifer around Baltimore, Md. The changes in chemical composition of the water flowing from a recharge area toward a discharge area result from the following chemical reactions:

1. If the pH of the water is below 5.5, lignite is oxidized to form carbon dioxide and water. Undisassociated carbon dioxide is the predominant carbon species.
2. If the pH of the water is above 5.5, lignite is oxidized to form hydrogen and bicarbonate. The bicarbonate ion is the predominant carbon species.
3. Lignite reacts with sulfate to form hydrogen sulfide and bicarbonate (normally catalyzed by microorganisms). The bicarbonate ion is the predominant carbon species.

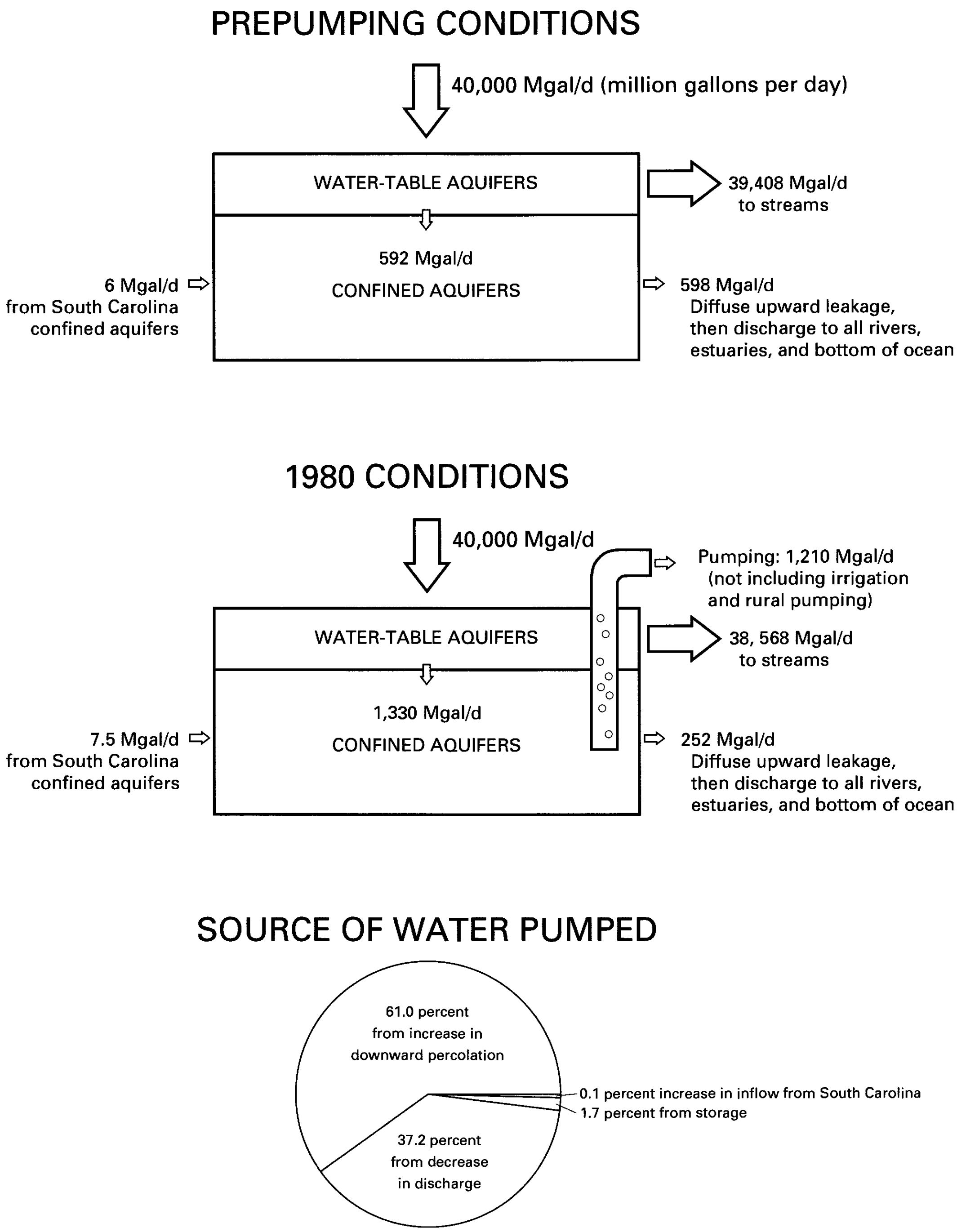


FIGURE 6.—Comparison of prepumping and 1980 ground-water budgets for the northern Atlantic Coastal Plain aquifer system (data from Leahy and Martin, in press).

4. Pyrite reacts with oxygen and water to yield ferrous iron, hydrogen, and sulfate ions.
5. Ferrous ions react with oxygen and water to yield ferric hydroxide and hydrogen ions.

As water moves downgradient in the aquifer, the free oxygen that entered with the water in the recharge area is progressively consumed, and chemical reactions that depend on the presence of free oxygen cease. The oxidation of pyrite to ferric hydroxide is a principal source of iron-cemented beds of sandstone that form conspicuous ledges in the outcrops of the Potomac Group and are commonly penetrated by wells drilled within a few miles of the outcrop. The distribution of the precipitated ferric hydroxide is strongly influenced by microenvironmental and lithologic changes, for example, at the contact between silty material that contains finely disseminated lignite under reducing conditions and coarse gravel that lacks lignitic material and contains oxygenated water (Chapelle, 1985).

Application of the U.S. Geological Survey's WATEQF chemical-equilibrium digital model (Plummer and others, 1978) to chemical analyses of water from the lower Potomac aquifer indicates that the ferric hydroxyl ion ($Fe(OH)_2^+$) is the predominant form of dissolved iron in oxic water, and the ferrous ion (Fe^{+2}) predominates in anoxic water.

The silicate minerals are highly variable in composition and may undergo a wide range of chemical reactions. Reactions representative of the dissolution of these minerals follow:

1. The reaction of plagioclase with hydrogen ions and water to form kaolinite, sodium and calcium ions, and silicon hydroxide.
2. The reaction of illite with hydrogen ions and water to yield kaolinite, potassium and magnesium ions, and silicon hydroxide. (Kaolinite is a stable weathering product of silicate minerals in the lower Potomac aquifer.)

CHEMICAL PROCESSES IN MARINE SEDIMENTS

The interaction of water and aquifer material in the Beaufort-Aquia aquifer in southern Maryland (the aquifer's regional extent shown in pl. 5A) exemplifies the geochemical processes in a Coastal Plain aquifer consisting of marine sediments. The sediments are rich in glauconite and shell debris. The glauconite is the only abundant efficient medium of cation exchange. Cation exchange and the dissolution and redeposition of carbonate minerals are the predominant chemical processes in the Beaufort-Aquia aquifer. Calcium and magnesium ions are adsorbed on glauconite, thereby freeing sodium ions. The concentration of calcium and magnesium decreases and that of sodium increases downgradient, and the pH increases to a range of 8.0–8.9. Downgradient increases in concentrations of dissolved cations in the water correspond to increases in exchangeable cations on the glauconite of the aquifer (Chapelle, 1983; Chapelle and Drummond, 1983; Chapelle and Knobel, 1983).

In the weathering of a glauconitic aquifer, such as the Beaufort-Aquia in Maryland, glauconite may be oxidized to yield potassium and magnesium ions, dissolved silica, goethite (ferric hydroxide), and kaolinite. The process, which consumes hydrogen ions, is an example of incongruent dissolution of a silicate mineral; it reduces the concentration of exchangeable cations on glauconite and raises the pH of the water passing through. However, the goethite produced by the reaction is also active in cation exchange.

The North Carolina segment of the regional Castle Hayne–Piney Point aquifer is composed of dolomitic molluscan and bryozoan limestone and limey sand. It is the only major carbonate aquifer in the northern Atlantic Coastal Plain. Processes of dissolution-precipitation involving calcite, aragonite, and magnesium calcite result in the release of calcium, bicarbonate, and magnesium ions into Castle Hayne–Piney Point water, much as occurs in the Maryland segment of the regional Beaufort-Aquia aquifer just described.

Diatomaceous sediments in the lower Miocene of the Atlantic Coastal Plain, particularly on the Delmarva Peninsula, are a major source of silica to ground water. The evidence for cation exchange in the diatomaceous lower Chesapeake aquifer suggests that amorphous silica is acting as an exchange medium (L.L. Knobel, F.H. Chapelle, and Harold Meisler, written commun., 1988).

SOURCES OF CARBON SPECIES

The quantity of dissolved inorganic carbon in Atlantic Coastal Plain ground water cannot be accounted for entirely by contributions of atmospheric and soil gases, minerals, and organic material in the aquifers, as has been pointed out by Cederstrom (1946a, b) and Foster (1950). An additional source is needed to account for the excess.

The geochemistry of the Beaufort-Aquia aquifer in southern Maryland was studied with respect to the carbon dioxide problem. In the downgradient part of the aquifer, the assumption of a closed system for carbon dioxide does not explain the ion and isotope chemistry of the water. Additional carbon dioxide having an isotopic composition heavier than that entering the outcrop must be accounted for. In addition to the isotopic composition of the carbon dioxide, its occurrence in the deeply buried part of the aquifer must be explained. The most reasonable conclusion is that carbon dioxide is being generated in the aquifer.

The region of the Beaufort-Aquia aquifer in which excess carbon dioxide is evident is one in which dissolved oxygen is absent. Therefore, the carbon dioxide must be generated under anoxic conditions. The action of anaerobic bacteria on lignocellulose material, by which both methane and carbon dioxide are generated, involves mechanisms that could explain the observed increase in carbon dioxide. Isotopically light carbon would accumulate in the methane and heavy carbon in the carbon dioxide. Excess methane could leak out of the aquifer system at a greater rate than carbon dioxide. Chapelle and others (1987) documented the presence of bacteria and of methanogenic and sulfate-reducing activity in deeply buried Coastal Plain sediments.

TABLE 3.—*Description and occurrence of hydrochemical facies*

Hydrochemical facies	Description	Occurrence	Typical range of dissolved-solids concentrations
Variable composition	No dominant anion or cation, or too variable to be mapped.	Near updip limit of aquifers in New Jersey, Maryland, and North Carolina.	<250 mg/L; 250-500 mg/L in southwestern New Jersey.
Ca, Mg, HCO_3^-	$Ca^{+2}+Mg^{+2}\geq 50$ percent of the cations; $HCO_3^- \geq 50$ percent of the anions.	Along updip limit of aquifers or downdip from water of variable composition.	>250 mg/L; 250–500 mg/L common in North Carolina.
$NaHCO_3$	$Na^+ \geq 50$ percent of cations; $HCO_3^- \geq 50$ percent of anions.	Downdip from Ca, Mg, HCO_3^- water.	Up to 1,000 mg/L; <500 mg/L in New Jersey.
NaCl	$Na^+ \geq 50$ percent of cations; $Cl^- \geq 50$ percent of anions.	Downdip from HCO_3^- water; dominant in deepest part of aquifer system.	>1,000 mg/L.

The lower and middle Potomac and Magothy aquifers also contain lignite. Along a flow-path segment in the Magothy and middle Potomac aquifers in southeast Maryland, where the velocity of ground-water flow is assumed to range from about 4 to 40 ft/yr, the rate of generation of carbon dioxide is estimated to be 1.2×10^{-3} to 1.2×10^{-4} millimoles per liter per year (L.L. Knobel, F.H. Chapelle, and Harold Meisler, written commun., 1988). Winograd and Farlekas (1974, p. 86–89) noted a downgradient excess of carbon dioxide in the undifferentiated lower and middle Potomac and Magothy aquifers of New Jersey and attributed its source to the decomposition of lignite under anoxic conditions.

CHEMICAL QUALITY RELATED TO SOURCES OF RECHARGE AND FLOW PATHS

The chemistry of ground water in the northern Atlantic Coastal Plain aquifer system depends on the chemistry of recharge water, minerals along the flow paths, chemical-reaction environment, and residence time, which is a function of flow velocity. For example, the Magothy aquifer is recharged in two principal areas in Maryland. One is along its outcrop belt, where precipitation directly recharges the aquifer. Ground water in this part of the aquifer is characterized by elevated concentrations of iron and sulfate. Its type of water is typical of the upgradient parts of aquifers composed of sediments of nonmarine origin. The processes determining its chemistry are oxidation-reduction and silicate-hydrolysis reactions under low-pH conditions.

The other principal recharge area for the Magothy aquifer in Maryland is a potentiometric high where water flows downward from the overlying Beaufort-Aquia aquifer, which consists of marine deposits. The water in this part of the Magothy aquifer is typical of water in the upgradient parts of aquifers composed of marine deposits. The alkalinity of the water is controlled by dissolution-precipitation reactions involving carbonate material (Chapelle, 1983). Because of cation exchange, the calcium concentration decreases downgradient with increases in sodium concentration. The most likely exchanging medium is montmorillonite (Knobel and Phillips, 1988).

FACTORS AFFECTING WATER QUALITY IN THE SURFICIAL AQUIFER

The surficial aquifer is heterogeneous, and the quality of its water is also heterogeneous. The aquifer is characterized by local flow systems. Short flow paths and residence time tend to restrict the solution of minerals (Cushing and others, 1973, p. 47). Water in the surficial aquifer is especially susceptible to the effects of human activity because of its proximity to land surface. The primary anthropogenic influences on the chemistry of the water in the surficial aquifer have been crop and livestock production. The greatest effect of crop production is the addition of nitrogen and lime to the soil. Livestock production and septic-tank fields also add nitrogen, the stable end-product of which is nitrate in ground water (Denver, 1986).

GEOGRAPHIC DISTRIBUTION OF DISSOLVED CONSTITUENTS

Ground water in the northern Atlantic Coastal Plain is classified into four hydrochemical facies (L.L. Knobel, F.H. Chapelle, and Harold Meisler, written commun., 1988): (1) variable composition, (2) calcium plus magnesium bicarbonate, (3) sodium bicarbonate, and (4) sodium chloride (table 3). The hydrochemical facies in each aquifer follow a general coastward, or downdip, sequence from variable composition through sodium chloride. The same succession of facies generally appears downward in the vertical section. The distribution of hydrochemical facies with respect to outcrop areas and to the coast and the seaward increase in dissolved-solids concentration are largely the result of natural (nonpumping) flow patterns in the aquifer system.

Knobel and others (L.L. Knobel, F.H. Chapelle, and Harold Meisler, written commun., 1988) mapped the ground water for all aquifers except the surficial aquifer with respect to five ranges of dissolved-solids concentration: (1) 0–250, (2) 250–500, (3) 500–1,000, (4) 1,000–2,000, and (5) greater than 2,000 mg/L. Because of large local variations in the chemistry of its water and insufficient information for interpretation, the surficial aquifer was not mapped. The distributions of both the hydrochemical facies and concentrations shown in plates 11A and 11B are for water in the upper Chesapeake aquifer (the shallowest regional aquifer mapped by Knobel and others) and in the lower Potomac aquifer (the lowermost regional aquifer), respectively. The hydrochemical facies and dissolved-solids concentrations are generalized; as mapped, they represent the predominant character of the water in the regional aquifer system. Hydrochemical facies are also superimposed on the hydrogeologic sections (pls. 1A–1C).

The hydrochemical facies in water in each aquifer follow a general coastward, or downdip, sequence from variable composition to calcium plus magnesium carbonate to sodium bicarbonate to sodium chloride, although the first two facies may be absent in places (pls. 1A, 1B, 11A, and 11B). The same succession of facies generally appears downward in the vertical section, as shown in plates 1A–1C, but the sequence can be reversed locally. The general distribution of hydrochemical facies is largely the result of natural (nonpumping) flow patterns in the aquifer system.

OCCURRENCE OF SALTWATER

DISTRIBUTION

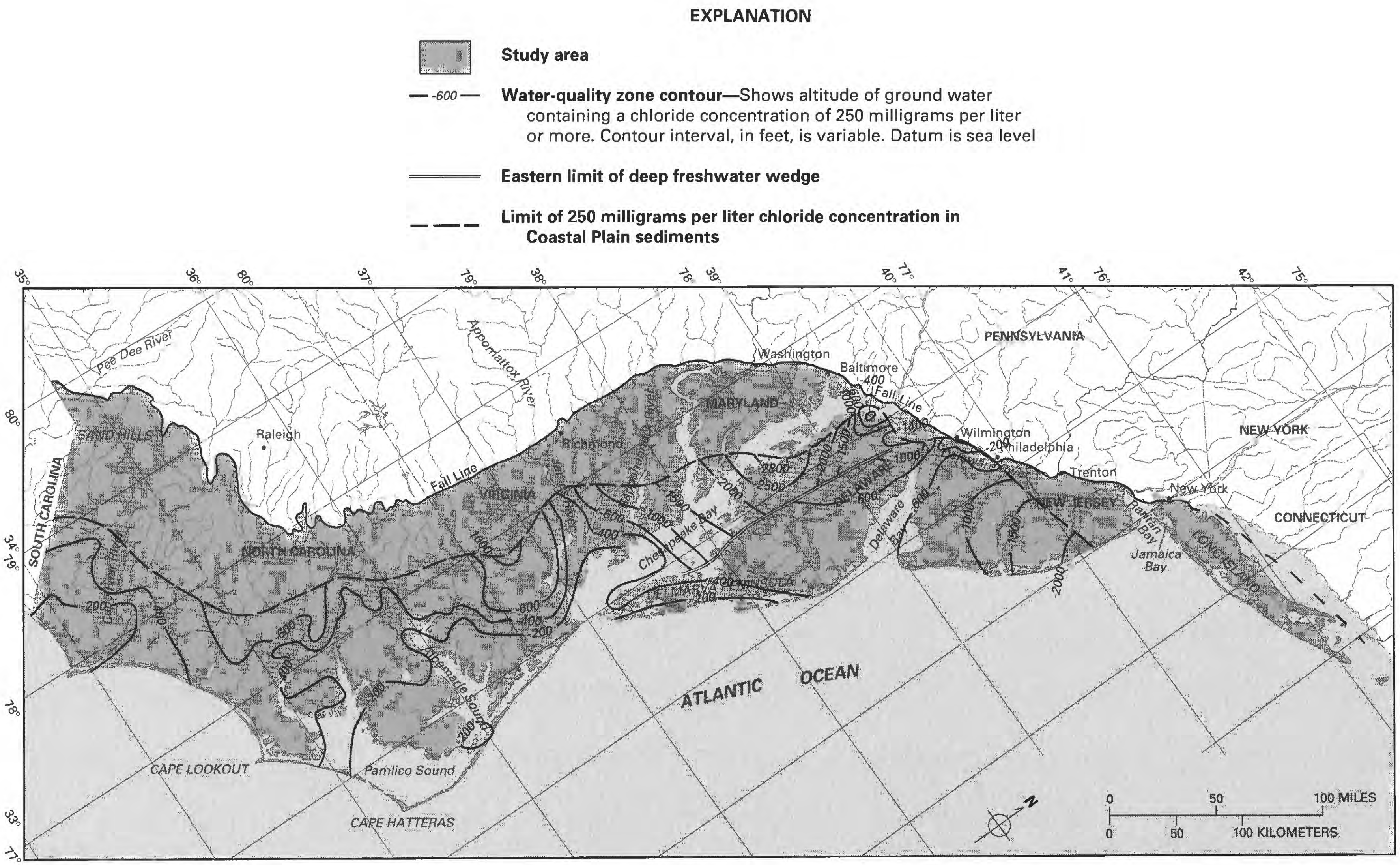
Salty ground water underlies freshwater in the eastern part of the northern Atlantic Coastal Plain from New Jersey to North Carolina. Chloride concentrations generally increase with depth. Meisler (1989, figs. 4–8) mapped depths to water with chloride concentrations of 250, 1,000, 5,000, 10,000, and 18,000 mg/L. The lowest chloride concentration mapped, 250 mg/L, is the drinking-water standard recommended by the U.S. Environmental Protection Agency (1979). The highest chloride concentration, 18,000 mg/L, is approximately that of seawater. The chloride-concentration data came primarily from chemical analyses of water from wells. These data were supplemented by analyses of pore fluids squeezed from clayey cores and by interpretation of borehole spontaneous-potential and resistivity logs. The depth to the 250-mg/L chloride concentration (fig. 7) portrays the top of the transition zone between freshwater and saltwater and the approximate base of the zone of potable water. The transition zone, which is defined as the region between concentrations of 250 and 18,000 mg/L of chloride, ranges in thickness from about 400 to 2,200 ft (Meisler, 1989).

According to Meisler (1989, figs. 7, 8), there is no ground water containing 10,000 mg/L or more of chloride in the Coastal Plain aquifer system north of southeastern New Jersey; the deepest aquifers in east-central New Jersey and western and central Long Island contain freshwater. Saltwater in the Magothy aquifer in southwestern Long Island is underlain by freshwater in the Lloyd aquifer (Perlmutter and others, 1959). The wedge shape of this saltwater body and the distribution of chloride concentrations indicate that the saltwater in the Magothy on Long Island is hydraulically connected to Jamaica Bay and is unrelated to the regional saltwater-freshwater transition zone.

The 10,000-mg/L chloride concentration was assumed to approximate the midpoint of the transition zone (as previously defined). The surface formed by the 10,000-mg/L chloride concentration line was also assumed to be a sharp interface between freshwater and seawater and to represent the seaward boundary of the freshwater flow system. This assumption seems valid as an approximation according to principles discussed by Henry (1964, p. C80, fig. 34b). Figure 8 delineates the altitude of the top of the zone of ground water containing 10,000 mg/L or more of chloride.

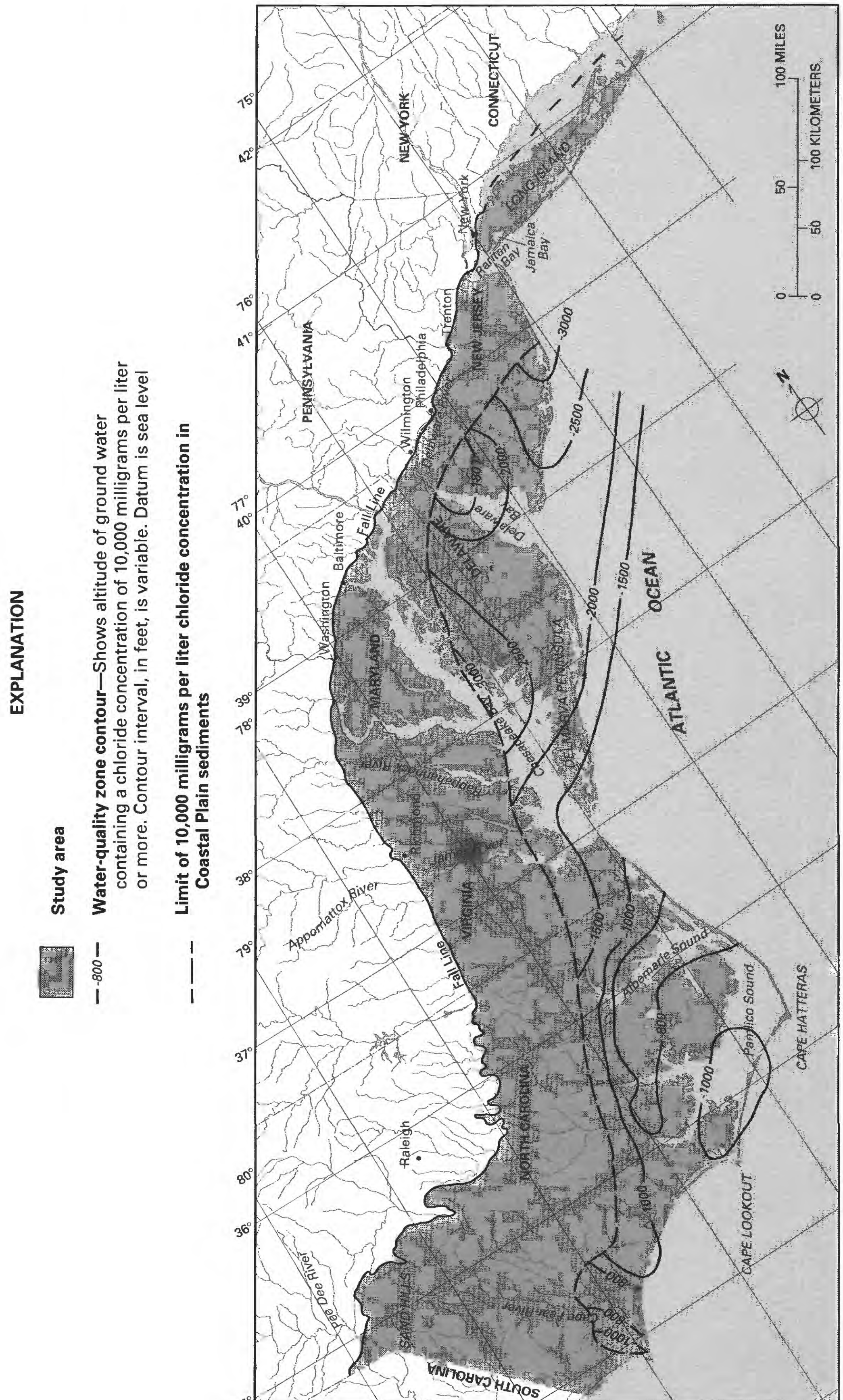
The saltwater-freshwater transition zone generally is shallowest in North Carolina and deepens northward, attaining its greatest depths in Maryland and New Jersey. It deepens inland from the coast, except locally in North Carolina and New Jersey, where it becomes shallower away from the coast toward Albemarle Sound and Delaware Bay, respectively. Ground water containing less than 5,000 mg/L of chloride (Meisler, 1989, fig. 9, section A–A′) extends 55 mi offshore from New Jersey but extends lesser distances offshore farther south (probably no more than a few miles from southeastern Virginia and from North Carolina) as the depth to saltwater decreases.

Areas where the upper part of the transition zone (250 mg/L of chloride, see fig. 7) is relatively shallow coincide with several areas of major fresh ground-water discharge, such as Delaware Bay, the lower Chesapeake Bay, Albemarle Sound, and the Cape Fear River. According to Upson (1966, p. C242), "Theoretically, equilibrium between fresh water and sea water in a coastal region requires that the hydraulic head of the fresh water be at least high enough to balance the head of salt water in the vicinity, considering the difference in



Base enlarged from U.S. Geological Survey
National Atlas, 1970, 1:7,500,000

FIGURE 7.—Altitude of the top of the zone of ground water containing a chloride concentration of 250 mg/L or more (from Meisler, 1986b, fig. 116).



Base enlarged from U.S. Geological Survey National Atlas, 1970, 1:7,500,000

FIGURE 8.—Altitude of the middle of the freshwater-saltwater transition zone (10,000-mg/L chloride concentration) (from Meisler, 1986b, fig. 117).

density ***." Hence, higher freshwater heads maintain the saltwater at greater depths in recharge areas, and lower freshwater heads maintain the saltwater at lesser depths in discharge areas.

Meisler (1989, figs. 4–6) also mapped chloride concentrations at the tops of three hydrogeologic units: (1) in predominantly nonmarine Cretaceous sediments (top of the middle Potomac regional aquifer), (2) in predominantly marine and marginal marine Cretaceous sediments (top of the Peedee-Severn aquifer where present, otherwise top of Black Creek–Matawan or upper Potomac and Magothy aquifers), and (3) in lower Tertiary sediments (top of the Castle Hayne–Piney Point aquifer). Chloride concentrations were shown to increase seaward in each of the mapped hydrogeologic units and to increase with depth from the shallowest of these units (in lower Tertiary sediments) to the deepest of these units (in predominantly nonmarine Cretaceous sediments).

SOURCE OF SALTWATER

Chloride concentrations generally increase with depth within a transition zone between the deepest freshwater and the underlying saltwater. For this study, Meisler (1989) defined the transition zone as that containing chloride concentrations of 250–18,000 mg/L, with 18,000 mg/L being the approximate chloride concentration of seawater. Meisler and others (1985) concluded that the transition zone was probably produced by the mixing of fresh ground water with salty ground water of either brine or seawater origin. Evidence for mixing is based on graphs that show the relations among concentrations of various ions and concentrations of chloride of waters in the transition zone (Meisler and others, 1985, p. 18–19, figs. 10–16; Meisler, 1989, figs. 7, 8). Concentrations of the ions tend to fall along straight lines when plotted against chloride, an ion associated with seawater that is assumed not to be entering or leaving solution.

The transition zone in North Carolina largely contains a mixture of seawater and a fresh sodium bicarbonate water, but brine underlies the transition zone in the vicinity of Pamlico and Albemarle Sounds (Meisler and others, 1985). Analyses of water from the transition zone within the Magothy aquifer on Long Island (Lusczynski and Swarzenski, 1966, table 2) indicate that the water is a simple mixture of seawater and freshwater.

In the area from Virginia to New Jersey, the presence of ion concentrations greater than those in seawater and the deviation of the ion concentrations from the seawater-freshwater mixing curves suggest that the transition zone contains essentially a mixture of a fresh sodium bicarbonate water with brine. Processes other than simple mixing of brine and freshwater that also may determine the ground-water composition in this area include (1) mixing with seawater, particularly in the upper part of the transition zone, and (2) ion exchange, by which the solution becomes enriched with calcium in the more concentrated waters and with sodium and potassium in the more dilute waters.

The origin of brines in the sediments of the northern Atlantic Coastal Plain is controversial. The brines have chloride concentrations several times that of seawater; they also differ from seawater in having higher concentrations of calcium and sodium and lower concentrations of magnesium, potassium, sulfate, and bicarbonate. Two possible sources of brine are (1) the concentration of dissolved solids resulting from reverse chemical osmosis and membrane filtration and (2) the leaching of evaporitic strata. Mechanisms that could cause reverse chemical osmosis, such as rapid sedimentation of fine-grained materials, tectonic compression, or abnormally high geothermal gradients (Bredehoeft and Hanshaw, 1968; Graf, 1982), do not appear to be active in the northern Atlantic Coastal Plain (Meisler, 1989).

According to Meisler (1989), the most likely source of brine in the Atlantic Coastal Plain is the leaching of evaporitic strata of probable Early and Middle Jurassic age (Mattick and Bayer, 1980, p. 8) beneath the Continental Shelf and Slope. The brine thus formed advanced landward, initially because of basinal compaction and subsequently because of major sea-level rises. Diffusion of the salt vertically through the sediments above the evaporites, as postulated by Manheim and Hall (1976, p. 699), could have provided a source from which the brine could migrate laterally.

EFFECT OF SEA-LEVEL CHANGES ON DISTRIBUTION OF SALTWATER

The natural flow pattern of fresh ground water appears to have exerted only partial control of the location of the saltwater-freshwater transition zone. Predevelopment freshwater heads probably were not high enough to account for the relatively great depths to the transition zone, especially depths to chloride concentrations as high as 18,000 mg/L (equivalent to seawater), in Maryland and along the coast of New Jersey nor do the heads appear to account for the wedge of relatively fresh ground water that extends 55 mi from the New Jersey coast.

A cross-sectional, finite-difference flow model was used to evaluate the effect of eustatic sea-level changes on the development and location of the saltwater-freshwater transition zone. The cross-sectional simulation was from the Delaware River in southern New Jersey to the Continental Slope (fig. 9), and the boundary between the freshwater and seawater systems is represented by a sharp interface (Meisler and others, 1985).

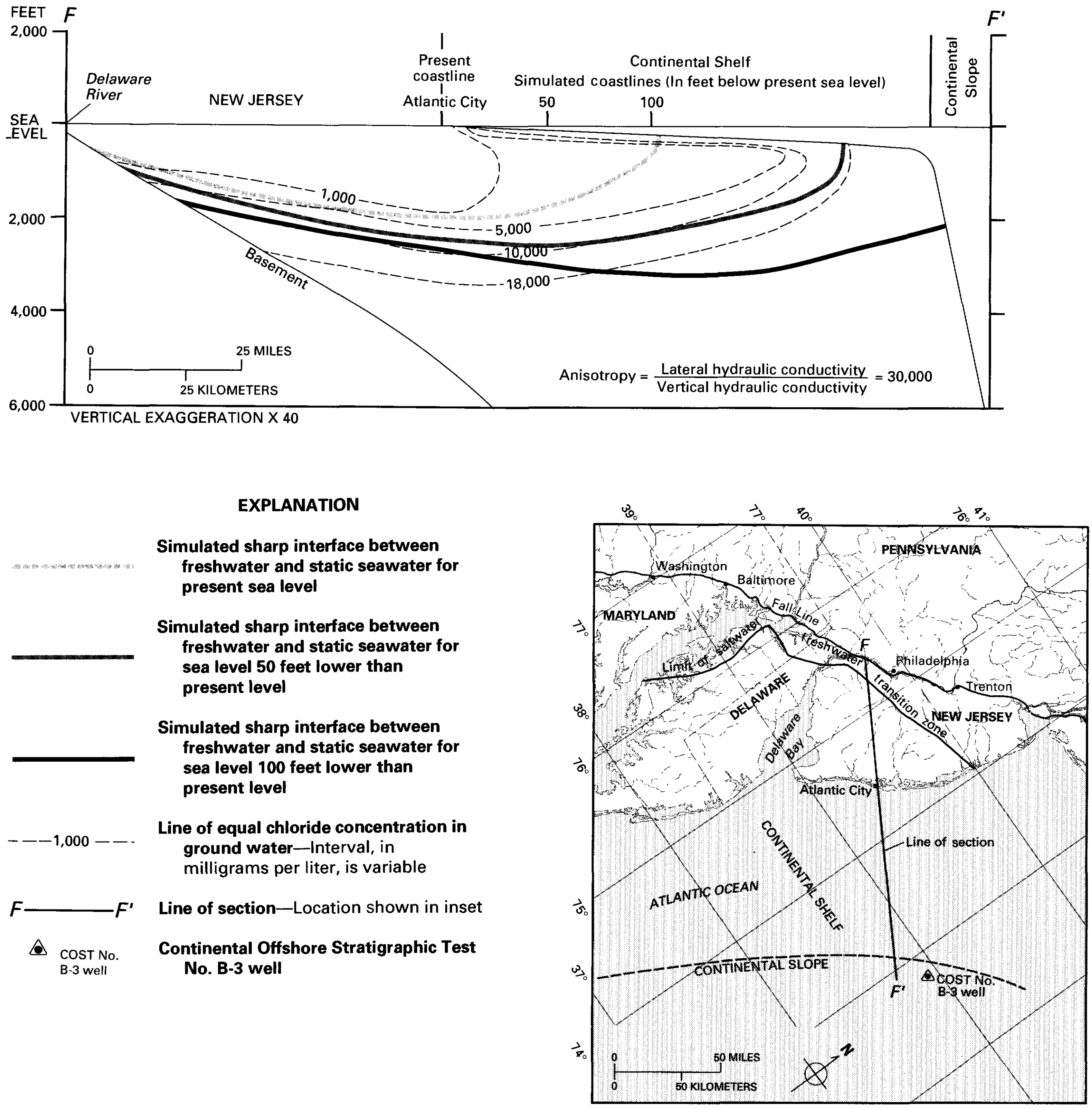


FIGURE 9.—Positions of simulated sharp interfaces between saltwater and freshwater from New Jersey offshore for sea levels 0, 50, and 100 ft below present sea level in comparison with observed chloride concentrations (adapted from Meisler and others, 1985, fig. 9).

The model tested the sensitivity of the location of a simulated saltwater-freshwater interface to anisotropy and changes in sea level. Simulation demonstrates that increasing anisotropy causes the simulated interface to be shallower and to extend farther offshore. Lowering of sea level deepens the simulated interface and also extends it farther offshore.

Comparison of the simulated positions of the sharp freshwater-saltwater interfaces with observed chloride concentrations (fig. 9) suggests that the location of the saltwater-freshwater transition zone (represented by 10,000-mg/L chloride concentration) in southern New Jersey and beneath the Continental Shelf is the result of a sea level that was about 50–100 ft below the present level. This position may approximate the average sea level during the late Tertiary and the Quaternary (Meisler and others, 1985). Because of extremely slow velocities of ground-water flow, the location of the transition zone probably has not adjusted to the present sea level and head distributions in the coastal aquifers.

Farther south, in southeastern Virginia and North Carolina, the general shallowness of the saltwater-freshwater transition zone and reduced offshore extent of ground water fresher than seawater suggest greater submergence in this region during the late Tertiary and the Quaternary than existed in New Jersey.

In the southern part of the North Carolina Coastal Plain, particularly near the Cape Fear River, water at relatively shallow depths (about 400–1,200 ft below sea level) contains chloride concentrations of 5,000 mg/L and above. According to the Ghyben-Herzberg relation, the high freshwater heads in this area (locally exceeding 100 ft above sea level) suggest that the freshwater-saltwater interface should be about 4,000 ft deep, below the base of the Coastal Plain section. Saltwater at shallower depths is probably related to past geologic times (late Miocene to early(?) Pleistocene) when sea level in this area was significantly higher than it is at present and there was not sufficient time for it to be flushed out (Meisler, 1989).

DEVELOPMENT OF THE SALTWATER-FRESHWATER TRANSITION ZONE

The breadth of the saltwater-freshwater transition zone is attributed to cyclic movement of saltwater responding to sea-level fluctuations with amplitudes of hundreds of feet (Vail and others, 1977; L.R. Zellmer, 1979, College of William and Mary, School of Marine Science, unpublished thesis, 85 p.) that occurred during the late Tertiary and the Quaternary Period. Repeated advance and retreat of the salty ground water caused the mixing of saltwater and freshwater in a transition zone in which saltier water predominates in the deeper and seaward parts and fresher water predominates in the shallower and landward parts (Meisler and others, 1985). In some areas, such as southeastern North Carolina, seawater that infiltrated during marine transgressions and that has not been flushed out completely probably contributed to the development of the transition zone.

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