GEOHYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE NORTHERN ATLANTIC COASTAL PLAIN AQUIFER SYSTEM

REGIONAL AQUIFER-SYSTEM ANALYSIS

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Geohydrology and Simulation of Ground-Water Flow in the Northern Atlantic Coastal Plain Aquifer System

By P. PATRICK LEAHY *and* MARY MARTIN

REGIONAL AQUIFER-SYSTEM ANALYSIS—
NORTHERN ATLANTIC COASTAL PLAIN

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1404-K

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON: 1993

U.S. DEPARTMENT OF THE INTERIOR

BRUCE BABBITT, ***Secretary***

U.S. GEOLOGICAL SURVEY

Robert M. Hirsch, ***Acting Director***

Library of Congress Cataloging in Publication Data

Leahy, P. Patrick.
Geohydrology and simulation of ground-water flow in the northern Atlantic Coastal Plain aquifer system / by P. Patrick Leahy and Mary Martin.

p. cm. — (Regional aquifer-system analysis) (U.S. Geological Survey professional paper ; 1404–K)
Includes bibliographical references.
Supt. of Docs. no.: I 19.16:1404–K
1. Groundwater flow—Atlantic Coast (U.S.)—Simulation methods. 2. Hydrogeology—Atlantic Coast (U.S.) I. Martin, M.M. II. Title. III. Series. IV. Series: U.S. Geological Survey professional paper ; 1404–K.
GB1197.7.L43 1993
551.49′0974—dc20 90–3419
CIP

For sale by U.S. Geological Survey, Map Distribution
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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.

Robert M. Hirsch

Robert M. Hirsch
Acting Director

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[Plates are in separate case]

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CONVERSION FACTORS AND VERTICAL DATUM

For readers who wish to convert measurements from the inch-pound system of units to the metric system of units, the conversion factors are listed below:

Multiply inch-pound units	By	To obtain metric units
inch (in)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
billion gallons per day (Ggal/d)	43.81	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
foot per day per foot [(ft/d)/ft]	1.000	meter per day per meter [(m/d)/m]
foot per day (ft/d)	0.3048	meter per day (m/d)
inch per year (in/y)	0.06954	millimeter per day (mm/d)
inch per hour (in/hr)	25.4	millimeter per hour (mm/hr)

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.

GEOHYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE NORTHERN ATLANTIC COASTAL PLAIN AQUIFER SYSTEM

By P. PATRICK LEAHY and MARY MARTIN

ABSTRACT

A ground-water flow model of the northern Atlantic Coastal Plain was designed and calibrated to increase understanding of the regional flow regime and changes in flow brought about by human activities. A multilayer flow model simulated ground-water flow in an aquifer system covering a 95,000-mi^2 (square miles) area that included Long Island, N.Y., and the Coastal Plain of New Jersey, Delaware, Maryland, Virginia, North Carolina, and parts of the adjacent Continental Shelf. The aquifer system consists of a seaward-thickening sedimentary wedge of 10 regional aquifers (predominantly sand) separated by 9 confining units (silt and clay).

Prepumping and pumping (1900 to 1980) conditions were calibrated through trial-and-error adjustment of hydraulic characteristics until model-simulated heads and flows matched those measured or estimated. Hydrographs from 74 wells were used to calibrate the flow model through time. Model calibration also was done at a local scale using subregional models for North Carolina, Virginia, Maryland-Delaware, and New Jersey. The hydraulic characteristics for the regional and subregional models were compatible.

Areal distributions of aquifer transmissivity and confining-unit leakance were refined through calibration. Transmissivity was highest, more than 100,000 ft^2/d (feet squared per day), in parts of the Castle Hayne aquifer in North Carolina. Transmissivity generally ranged from 500 to about 10,000 ft^2/d for most aquifers in the system. Confining-unit leakance ranged from about 1.0×10^{-8} to 1.0 foot per day per foot. Lower values typically were found in the deeper units. Updip leakance values tended to be higher because of a decrease in thickness or an increase in vertical hydraulic conductivity of the confining unit.

Sensitivity of the model response to hydraulic characteristics, confining-unit transient leakage, and the nature and location of the seaward model boundary was determined. Values of transmissivity, storage coefficient, withdrawals, and confining-unit leakance were changed by varying amounts dependent on a subjective estimate of their uncertainty. Model heads were most sensitive to an order of magnitude change in confining-unit leakance and to confining-unit transient leakage. Additional data are needed to determine the importance of transient leakage in the hydraulics of the aquifer system. A variable-density flow model evaluated the sensitivity of the model to the seaward boundary. This model simulated flow in parts of the aquifer system that contained water having chloride concentrations greater than 10,000 mg/L (milligrams per liter). Along the 10,000-mg/L-chloride-concentration line, head differences between the calibrated constant-density flow model and the variable-density flow model generally were less than 10 ft (feet). Thus, probably only a small percentage of the water pumped from the system was derived from areas containing salty water.

Characteristics of the regional flow system described in the report are based on an analysis of the model derived: (1) water budgets, (2) potentiometric surfaces, (3) vertical leakage between aquifers, and (4) lateral flow patterns and velocities within aquifers. Average areal ground-water recharge to the surficial aquifer of the Coastal Plain was estimated to be about 40,000 Mgal/d (million gallons per day), or 15.4 in/yr (inches per year). The majority of this recharge discharges to the nearest surface-water body. Model results indicate that under prepumping conditions, 592 Mgal/d, or about 0.5 in/yr, of the average areal recharge moved downward to the underlying confined system. This recharge occurred over approximately 25,000 mi^2, or 26 percent of the total area. Discharge from the deeper aquifers occurred over the remaining 74 percent of the study area. In areas where the surficial aquifer is in direct contact with an underlying confined aquifer (no intervening confining unit), the maximum simulated prepumping rate of recharge from the surficial aquifer to the underlying confined system was about 16 in/yr, and the maximum rate of discharge from the underlying confined system to the surficial aquifer was about 20 in/yr.

Interpretation of simulated prepumping potentiometric surfaces indicates that (1) recharge to the confined aquifers occurred in areas of downward hydraulic gradient, generally along or near the Fall Line, and (2) discharge from the deeper confined aquifers occurred by upward leakage through confining units into the ocean or coastal estuaries and bays. The simulated prepumping potentiometric surfaces of the shallower aquifers show relatively local flow patterns. The influence of major rivers, estuaries, and embayments on the flow system is apparent. In contrast, the potentiometric surfaces of the deeper aquifers show a regional flow pattern. Although the streams and rivers affected flow in updip areas near outcrops, generally flow was not influenced by overlying surface-water bodies throughout the areal extent of the deeper aquifers. Notable exceptions included Raritan, Delaware, and Chesapeake Bays. These large surface-water bodies affected flow patterns in all aquifers and had significant influence on the location of the 10,000-mg/L chloride concentration. In the deeper aquifers, ground-water flow paths typically were several tens of miles long. Small lateral hydraulic head gradients and low values of hydraulic conductivity indicate that computed Darcy flow velocities were slow (less than 1 ft per year) along regional flow lines.

By 1980, withdrawals had caused regional heads in several aquifers to decline to more than 100 ft below sea level in areas of North Carolina, Virginia, Delaware, and New Jersey, and to more than 50 ft below sea level in areas of Maryland. The size and shape of the pumping cones depended on the quantity of water being withdrawn, the location of the pumping center relative to the aquifer outcrop, and the hydraulic characteristics of the aquifers and confining units.

Withdrawals in 1980, primarily from the confined system, were estimated to be about 1,210 Mgal/d, or about 3 percent of the estimated average annual ground-water recharge (40,000 Mgal/d) to the surficial aquifer. However, 1980 withdrawals were about twice the simulated recharge from the surficial aquifer to the confined-flow system prior to development (592 Mgal/d).

Pumpage of water resulted in (1) a reduction in aquifer storage, (2) an increase in recharge from the surficial aquifer to the confined system and (or) a decrease in discharge to or direct recharge from streams, and (3) a reduction in discharge from the confined system through overlying confining units to large surface-water bodies. Reduction in aquifer storage was negligible (less than 2 percent of the withdrawals). The system approached equilibrium in less than 5 years after each simulated change in withdrawals. An increase in recharge to the confined system derived from reduced discharge to streams was the principal source of pumped water. In 1980, the recharge to the confined system was 1,330 Mgal/d, and the area of recharge to the confined system from the surficial system was approximately 45 percent of the study area. The 1980 recharge represents an increase in recharge area of 19 percent over prepumping conditions. A smaller source of the pumped water was water that formerly discharged upward through overlying confining units in coastal areas.

INTRODUCTION

This chapter of U.S. Geological Survey Professional Paper 1404 (chapter K) describes the digital simulation of ground-water flow in the northern Atlantic Coastal Plain aquifer system. Computer simulation has been used for many years to evaluate the response of an aquifer system to changing hydrologic stresses. However, unlike many previous studies, which emphasized the predictive capabilities of ground-water flow models, the regional studies of all projects in the Regional Aquifer-System Analysis (RASA) program (described below) used digital simulation primarily for analysis of the regional ground-water flow system (Bennett, 1979, p. 39).

The northern Atlantic Coastal Plain aquifer system study is part of the U.S. Geological Survey's RASA program initiated in 1978. The purposes of the RASA program are to define the regional geohydrology and geochemistry of the major regional ground-water flow systems of the United States and to establish a framework of background information that can be used for regional assessment of ground-water resources (Sun, 1986, p. 6). As part of the RASA program, the northern Atlantic Coastal Plain RASA project defined the geohydrology and geochemistry of the northern Atlantic Coastal Plain aquifer system. Other results of the northern Atlantic Coastal Plain RASA project are described in a number of chapters in Professional Paper 1404.

Chapters of Professional Paper 1404 are designated by letters A through M. These chapters describe (1) the hydrogeologic framework of the Coastal Plain (G) and more detailed frameworks for North Carolina (I), Virginia (C), Maryland and Delaware (E), and New Jersey (B); (2) the distribution of saline water in the Coastal Plain sediments (D); (3) the geochemistry of the Coastal Plain aquifer system (L); (4) the regional ground-water flow and hydraulic properties of aquifers and confining units, as studied through digital simulation, in the northern Atlantic Coastal Plain (K) and in North Carolina (M), Virginia (F), Maryland and Delaware (J), and New Jersey (H). Professional Paper 1404 also contains a summary chapter (A). Complete bibliographic references for these reports are given in the "Selected References" at the end of this report. These and other reports that were prepared wholly or in part during the RASA program are indicated by asterisks in the "Selected References."

The northern Atlantic Coastal Plain aquifer system is a major source of ground water throughout its extent. The northern Atlantic Coastal Plain extends from Long Island, N.Y., to the North Carolina-South Carolina State boundary and covers approximately 55,000 mi^2 (square miles). The Coastal Plain and the adjacent offshore Continental Shelf are underlain by an eastward-thickening sedimentary wedge consisting primarily of consolidated and unconsolidated sand, silt, and clay of Jurassic to Holocene age. Limestone is present locally but is an important aquifer only in North Carolina. In chapter G of Professional Paper 1404, which discusses the regional hydrogeologic framework (Trapp, in press), the sedimentary wedge is described as a complex series of aquifers and confining units. In 1980, the aquifers were the major source of fresh ground water for the area.

The study area includes the Coastal Plain from Long Island, N.Y., southward through North Carolina and a short distance into northeastern South Carolina. The study area is bounded on the west by the Fall Line and on the east by a line approximately representing the interface between water containing chloride concentrations of less than 10,000 mg/L (milligrams per liter) and water containing chloride concentrations of greater than 10,000 mg/L (fig. 1). The study covers an area of about 95,000 mi^2.

PURPOSE AND SCOPE

A purpose of the northern Atlantic Coastal Plain RASA study, as defined by Meisler (1980, p. 10), was to "simulate both the predevelopment and present flow conditions in the aquifer system by digital computer models." One objective of simulating ground-water flow in this study was to improve understanding of both the natural flow system that existed prior to the development of ground water and the flow system as modified through development of ground water from 1900 to 1980.

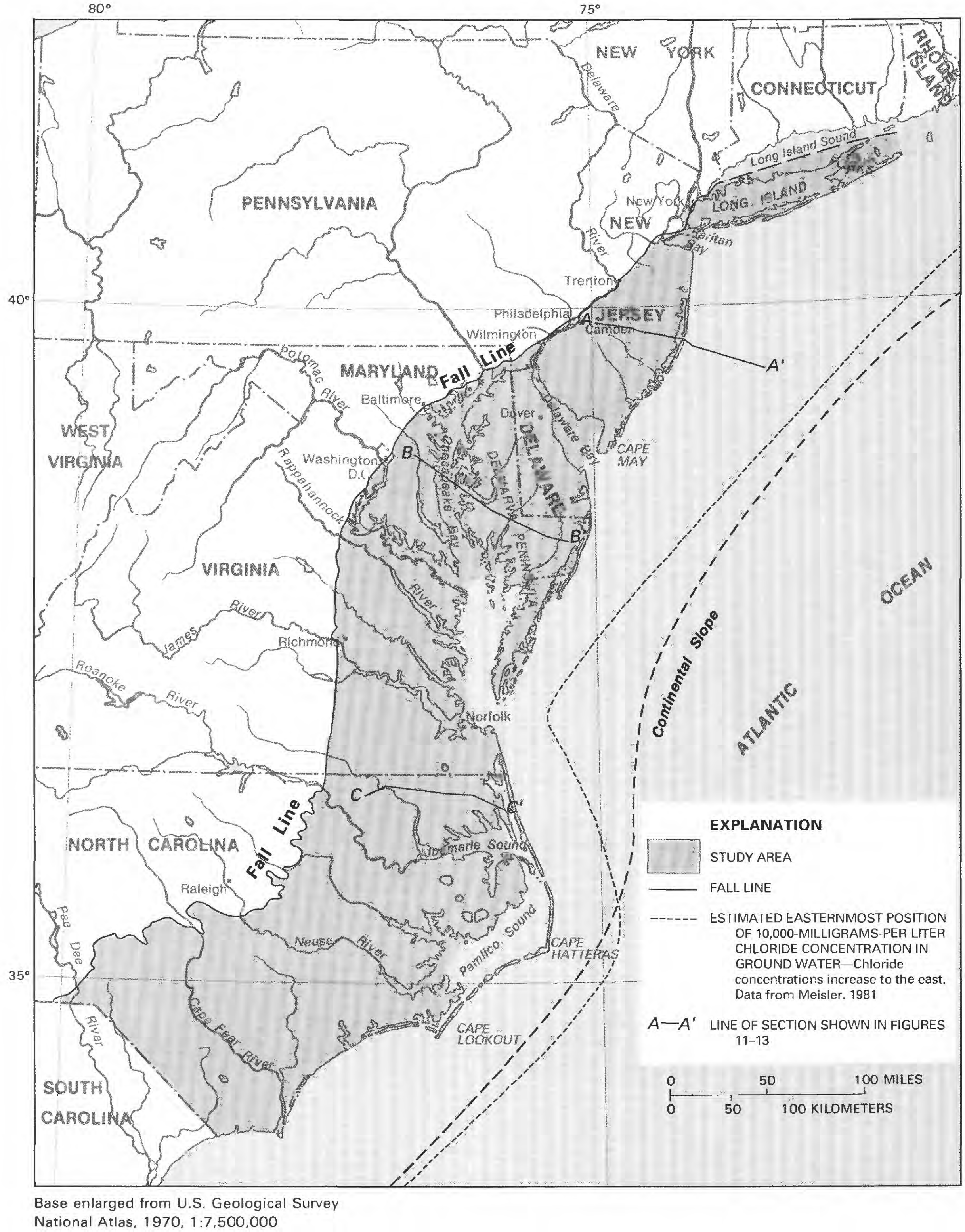


FIGURE 1.—Location of study area and lines of hydrogeologic sections.

Another major objective was to simulate flow conditions deduced from the hydrogeologic framework and from basic hydrologic principles. This report summarizes (1) the ground-water hydrology of the Coastal Plain, (2) the physical and mathematical conceptualizations of both the predevelopment and present ground-water flow systems, (3) the design of the regional digital model, (4) the strategy for calibrating the model, (5) the sensitivity of the flow model to selected hydrologic characteristics and boundary conditions, and (6) the regional flow system as defined by the results of the calibrated flow model.

The process of digitally simulating ground-water flow can generally be divided into five major tasks: (1) system conceptualization, (2) model design, (3) model calibration, (4) determination of parameter sensitivity, and (5) prediction or forecasting. The principal objectives of this study were to improve understanding of the natural flow system and to assess the effects of ground-water development on this system. Therefore, only the first four tasks are discussed in this report.

The flow of ground water in the northern Atlantic Coastal Plain aquifer system was simulated for about 95,000 mi^2. This area includes the emerged Coastal Plain, about 55,000 mi^2, and adjacent offshore areas (Continental Shelf). Simulations were done both for the entire study area (regional study) and for four subregional study areas (New Jersey, Delaware and Maryland, Virginia, and North Carolina). Data were collected and compiled by subregional study teams and supplied to a regional study team. Calibration was completed at both the regional and subregional scales. This report presents only the results of the regional study and the methodology of incorporating the subregional modeling efforts.

PREVIOUS SIMULATIONS OF GROUND-WATER FLOW IN THE COASTAL PLAIN

Simulation has been used to analyze regional flow of ground water in all of the States in the northern Atlantic Coastal Plain. Simulation has been used to define ground-water flow in the following States, in descending order of frequency: New York (Long Island), New Jersey, Maryland, Delaware, Virginia, and North Carolina. A brief summary of the modeling efforts for each State follows.

On Long Island, several studies used either electrical-analog or digital-modeling techniques to evaluate the ground-water flow system. Franke and Getzen (1976) used a cross-sectional electrical-analog model to initially define the degree of vertical hydraulic connection between the Lloyd, Magothy, and upper glacial aquifers. The results of the cross-sectional model provided a basis for development of an areal multilayer-electrical-analog model of the Magothy and upper glacial aquifers (Getzen, 1977). The analog model was later replaced by a multilayer-finite-difference digital model having the same spatial discretization as the analog model (Reilly and Harbaugh, 1980). Other studies of the Long Island aquifers have used simulation to analyze local hydrologic conditions or to document simulation techniques for specialized hydraulic situations. These studies include Harbaugh and Reilly (1976, 1977), Harbaugh and Getzen (1977), Kimmel and Harbaugh (1975, 1976), and Kimmel and others (1977).

Simulation of ground-water flow in New Jersey began in the mid-1970's with the design and calibration of a single-layer two-dimensional model of the Englishtown aquifer in the northern Coastal Plain (Nichols, 1977). Following this initial study, additional single-layer models were designed and calibrated for the Wenonah-Mount Laurel aquifer (Nemickas, 1976), the Potomac-Raritan-Magothy aquifer system in the southern Coastal Plain (Luzier, 1980), the Farrington aquifer (in the Farrington Sand Member of the Raritan Formation) in the northern Coastal Plain (Farlekas, 1979), and the Cohansey aquifer system in the Pine Barrens (Harbaugh and Tilley, 1984). Studies also have been completed that used a calibrated model for predictive purposes (Harbaugh and others, 1980). Local ground-water flow systems have also been modeled for use in simulating contaminant transport (Gray and Hoffman, 1983).

Ground-water flow has been modeled extensively in Maryland, with finite-difference single-layer and multilayer areal models. Single-layer models have been designed and calibrated for the Magothy aquifer in southern Maryland (Mack, 1974; Mack and Mandle, 1977), the Piney Point aquifer in southern Maryland and on the Eastern Shore (Williams, 1979), and the Aquia aquifer in the southern part of the State (Kapple and Hansen, 1976). A single-layer flow and solute transport model was calibrated for the Lower Cretaceous aquifer system near Baltimore (Chapelle, 1985). Studies using multilayer models include studies of the Piney Point-Nanjemoy and Aquia aquifer system in southern Maryland (Chapelle and Drummond, 1983), the Pocomoke, Ocean City, and Manokin aquifer system in the Ocean City, Md., area (Achmad and Weigle, 1979), and the Cretaceous aquifers throughout the Coastal Plain of southern Maryland (Fleck, 1983). Also, a series of uncalibrated single-layer models was designed to evaluate the water-supply potential of the Magothy, Patapsco, and Patuxent aquifers in the Washington, D.C., metropolitan area in southern Maryland (Papadopulos and others, 1974).

In the Coastal Plain of Delaware, ground-water flow has been simulated using finite-difference single-layer models for the Piney Point aquifer in the central part of the State (Leahy, 1979), the unconfined Columbia aquifer in the central and southeastern parts (Johnston, 1977), and the Pocomoke and Manokin aquifers in the southeastern part (Hodges, 1984). Multilayer models have been designed and calibrated for the Piney Point and Cheswold aquifers in central Delaware (Leahy, 1982a) and for the upper, middle, and lower aquifers within the Potomac Formation in northern Delaware (Martin, 1984).

In Virginia, Cosner (1975) was the first to use ground-water flow modeling in the Coastal Plain. He used a single-layer finite-difference model of the Cretaceous aquifer to evaluate the effects of increased ground-water withdrawals in the Franklin, Va., area. Bal (1977) designed and calibrated a single-layer model of the aquifers of the Yorktown Formation on the Eastern Shore of Virginia. Bal (1978) also developed a multilayer model to simulate flow in the unconfined upper artesian aquifer in Tertiary sediments, and in the principal artesian aquifer in Cretaceous sediments of the York-James and Middle Peninsulas. More recently, a multilayer analog model was developed for the aquifer system in Cretaceous sediments south of the James River (Layne Atlantic Company, 1983).

In the Coastal Plain of North Carolina, Sherwani (1973) used both single-layer electrical-analog and finite-difference models to simulate ground-water flow in the Castle Hayne aquifer.

GEOHYDROLOGY

PHYSIOGRAPHY

The onshore part of the study area, although composed entirely of Coastal Plain sediments, varies greatly in physiography from Long Island to North Carolina. The physiography is related to the type of geologic deposits and to the geologic processes that have acted on the deposits. The regional ground-water flow system under prepumping conditions was influenced strongly by the physiography (Bredehoeft and others, 1982, p. 297). It is therefore appropriate to describe briefly the physiography of the study area. The physiographic areas or subprovinces discussed below, and the locations referred to throughout the report, are shown on plate 1.

The topography within the study area is moderately flat, with elevations ranging from sea level along major estuaries and the Atlantic coastline to about 700 ft (feet) above sea level along the Fall Line in North Carolina. The Fall Line is defined as a narrow zone characterized by rapids and falls where streams leave the bedrock area of the Piedmont physiographic province and enter the Coastal Plain province.

On Long Island, unlike the rest of the study area, the major physiographic features were formed by geologic processes of the Wisconsin glacial age. Cohen, Franke, and Foxworthy (1968) reported that the major landforms are (1) hills corresponding to glacial moraine deposits (Harbor Hill and Ronkonkoma moraines) that trend east-west along the center of the island, (2) a gently sloping outwash plain to the south of the hills, (3) deeply eroded headlands along the northern shore, and (4) barrier beaches along the southern shore. The elevation of the hills along the center of the island is typically 100 to 150 ft above sea level. Unlike other parts of the study area, the contact between Coastal Plain sediments and Piedmont rocks in New York is covered by water (Long Island Sound).

In New Jersey, Wolfe (1977) has divided the Coastal Plain into three physiographic subprovinces: the central upland, the outer lowland, and the inner lowland. The outer lowland, located along the coastal margin and Delaware Bay, is characterized by low relief and shallow streams in open valleys that flow to marsh-lined estuaries. The inner lowland, located along the Delaware River from Delaware to Trenton, N.J., and from there northeast to Raritan Bay, is characterized by highly dissected sediments of Cretaceous age that crop out in mappable bands. The central upland is formed primarily by outcrops of upper Cretaceous and lower Tertiary deposits. In the area of the Atlantic Ocean-Delaware River watershed divide it is characterized by a gently rolling terrain; the remaining area of the central upland is a plain of low relief. The elevation of the central upland typically is less than 100 ft but locally may exceed 200 ft above sea level.

The Coastal Plain of Maryland can be divided into two major subprovinces: the western-shore type and the eastern-shore type (Overbeck and Slaughter, 1958, p. 10). These subprovinces also categorize the Coastal Plain of Delaware. The western-shore type subprovince is characterized by rolling uplands highly dissected by steep-sided streams. The elevation of bluffs adjacent to the streams exceeds 200 ft above sea level at some locations. The Maryland Coastal Plain west of Chesapeake Bay and the Delaware and Maryland Coastal Plain north of the Chesapeake and Delaware Canal is predominantly the western-shore type subprovince (pl. 1).

The eastern-shore type physiographic subprovince covers the remaining Coastal Plain on the Delmarva Peninsula, including the Virginia parts. This subprovince consists generally of several plains of minor elevation differences. The plains tend to be broadly rolling to flat, with a typical elevation range of 45 to 80 ft above sea level. Sluggish streams having very low gradients and widely spaced divides dissect the plains.

In Virginia, the Coastal Plain is characterized by a series of relatively flat terraces that rise in steps toward the Fall Line (DeBuchananne, 1968, sheet 1). The topographic relief is about 250 ft from the Fall Line to the coastal areas along Chesapeake Bay, the Atlantic Ocean, and the estuaries of the York, James, Potomac, and Rappahannock Rivers. The relatively flat terraces are separated by steep beach-cut escarpments parallel to Chesapeake Bay. The inner Coastal Plain is located between the Fall Line and the Suffolk Scarp (pl. 1). The outer Coastal Plain is located east of the scarp (Oaks and Coch, 1973, p. 4). The stream morphology varies from broad flood plains in the upper reaches of the York, James, and Rappahannock Rivers to drowned river valleys in estuaries of these rivers (Newton and Siudyla, 1979).

The elevation of the North Carolina Coastal Plain ranges from about 700 ft above sea level along the Fall Line to sea level along the numerous bays and estuaries. Heath (1980) divided this area into two major physiographic subprovinces: the tidewater region and the inner Coastal Plain (pl. 1). The tidewater region is affected by tides and oceanic influences, whereas the inner Coastal Plain is not. In general, the stream morphology is the same as in Virginia, with incised sluggish streams and broad flood plains in the inner Coastal Plain and drowned river valleys in the tidewater subprovince. The major estuaries and bays include Pamlico and Albemarle Sounds, and the Roanoke, Cape Fear, Neuse, Tar, and Chowan Rivers. The same physiography applies to the small part of the study area near the Pee Dee River in South Carolina.

HYDROGEOLOGIC FRAMEWORK

The hydrogeologic framework used for simulation of ground-water flow in this study is discussed in detail in chapter G of this Professional Paper series (Trapp, in press). The aquifer system in the northern Atlantic Coastal Plain consists of interbedded sand, silt, clay, and limestone. The wedge-shaped body of sediments has been divided into geologic formations and chronostratigraphic units ranging in age from Jurassic to Holocene. A notable example of a regional correlation of stratigraphic units appears in Brown and others (1972).

Numerous studies of the hydrogeology of the northern Atlantic Coastal Plain have divided Coastal Plain sediments into hydrogeologic units and assigned aquifers names. These units are lithologic units and do not necessarily correspond to chronostratigraphic units. An aquifer is by definition a formation, a group of formations, or part of a formation that easily transmits water (Lohman and others, 1972), whereas confining units impede its flow. The hydrogeologic framework by Trapp (in press) is divided into regionally continuous aquifers and confining units. Examples of hydrogeologic sections by Trapp (in press) are shown in figures 11–13. Cushing and others (1973) provided a notable example of a more local framework. In that study, the major aquifers underlying the Delmarva Peninsula were identified and mapped.

As part of this study, detailed hydrogeologic frameworks have been developed for New Jersey (Zapecza, 1989), Maryland and Delaware (Vroblesky and Fleck, in press), Virginia (Meng and Harsh, 1988), and North Carolina (Winner and Coble, in press). Trapp (in press) has adapted these hydrogeologic frameworks, together with Getzen's (1977) framework and Garber's (1986) mapping of the Lloyd aquifer for Long Island, into a regional hydrogeologic framework. This regional hydrogeologic framework maintains the continuity of the hydrologic units across the study area and provides the basic framework on which the flow model is based. Table 1 shows the relation between regional and local aquifer names, and modeled aquifer layers. The hydrogeologic framework defines the altitude and areal extent of aquifers and the thickness of confining units in the Coastal Plain. Some of the 12 regional aquifers described in chapter G (Trapp, in press) are combined in model layers 3, 6, and 7 on the basis of hydrogeologic considerations. The layer notation used in this report is adapted so that the reader can correlate regional and subregional aquifer names with model layer numbers. In general, regional aquifer names will be used when discussing aspects of the regional system. However, subregional aquifer names will be used as appropriate when discussing local aspects of the flow system.

In general, the aquifers consist primarily of sand and limestone, and the confining units consist of clay and silt. In some instances, especially in the sediments of the Lower Cretaceous, a distinct division between aquifers and confining units could not be identified because of the lithologic nature of these sediments. The modeled aquifers for these sediments include all the substantial water-bearing sand zones. The confining units consist predominantly of clay and silt, with minor amounts of sand. Ideally, many model layers would be used to represent these sediments in a flow model. However, considering the regional nature of the analysis, vertical head distribution, data availability, and the distribution of withdrawals, the number of layers used in this study seemed the most practical representation of the system.

The aquifers and confining units, although regionally extensive, are not present throughout the entire study area. For simulation purposes, all aquifers and confining units were assumed to terminate downdip toward the east at the line representing the 10,000-mg/L chloride concentration and updip toward the west by the outcrops of the units. The units do not actually cease to exist at the 10,000-mg/L isochlor. However, some aquifers are truncated downdip by a facies change into finer sediments. These aquifers include (1) the Castle Hayne-Piney Point aquifer (layer 7) in New Jersey and Delaware, (2) the Beaufort-Aquia aquifer (layer 6) in New Jersey, Maryland, Delaware, and Virginia, (3) the Peedee-Severn aquifer (layer 5) in New Jersey, and (4) the Black

TABLE 1.—*Relation of regional aquifer names, subregional aquifer names, and model layer numbers used in flow model*

Model layer	Regional aquifer	Subregional aquifer[1]				
		North Carolina	Virginia	Maryland Delaware	New Jersey	New York (Long Island)
10	Surficial	Surficial	Columbia	Surficial	Holly Beach	
9	Upper Chesapeake	Yorktown	Yorktown-Eastover	Upper Chesapeake[2]	Upper Kirkwood-Cohansey	
8	Lower Chesapeake	Pungo River	St. Marys-Choptank	Lower Chesapeake[3]	Lower Kirkwood-Cohansey and Confined Kirkwood	
7	Castle Hayne-Piney Point	Castle Hayne	Chickahominy-Piney Point	Piney Point-Nanjemoy	Piney Point	Upper part of upper glacial
6	Beaufort-Aquia	Beaufort	Aquia	Aquia-Rancocas	Vincentown	Lower part of upper glacial
5	Peedee-Severn	Peedee		Severn	Wenonah-Mount Laurel	Upper part of Magothy
4	Black Creek-Matawan	Black Creek		Matawan	Englishtown	Middle part of Magothy
3	Upper Potomac and Magothy	Upper Cape Fear	Brightseat-upper Potomac	Brightseat and Magothy	Upper Potomac-Raritan-Magothy	Lower part of Magothy
2	Middle Potomac	Lower Cape Fear	Middle Potomac	Patapsco	Middle Potomac-Raritan-Magothy	Lloyd
1	Lower Potomac	Lower Cretaceous	Lower Potomac	Patuxent	Lower Potomac-Raritan-Magothy	

[1]The subregional aquifer names in the six States were derived from the following geologic units, which are generally listed in descending order: Quarternary deposits—Columbia Group, glacial deposits; Tertiary units—Yorktown, Eastover, St. Marys, and Choptank Formations of the Chesapeake Group, Kirkwood Formation, Cohansey Sand, Pungo River, Castle Hayne, Chickahominy, Piney Point, Nanjemoy, Beaufort, Aquia, Vincentown, and Brightseat Formations, and Rancocas Group; Cretaceous units—Peedee, Severn, and Wenonah Formations, Mount Laurel Sand, Black Creek, Matawan, Englishtown, Cape Fear, Magothy, and Raritan Formations, Patapsco and Patuxent Formations of the Potomac Group, and Lloyd Sand Member of the Raritan Formation. The regional aquifer names were derived from subregional aquifer names and combinations of the names.

[2]Contains the Pocomoke, Ocean City, and Manokin aquifers on the Delmarva Peninsula.

[3]Contains the Frederica, Federalsburg, and Cheswold aquifers on the Delmarva Peninsula.

Creek-Matawan aquifer (layer 4) in New Jersey, Maryland, and Delaware.

Updip in the ground-water system, some aquifers and confining units are absent locally. Direct hydraulic connection between aquifers occurs where a confining unit is absent, whereas confining units are directly in contact where aquifers are absent. Areas where confining units are absent are (1) on Long Island, where the upper and lower parts of the upper glacial aquifer (layers 6 and 7) are in contact with the upper, middle, and lower parts of the Magothy aquifer (layers 3, 4, and 5), (2) in New Jersey, where the Cohansey aquifer (layer 9) is directly underlain by the Kirkwood aquifer (layer 8), (3) in Virginia, where the Columbia aquifer (layer 10) is in direct contact with the Yorktown-Eastover aquifer (layer 9), and (4) in North Carolina, where the subcrop of the Castle Hayne aquifer (layer 7) is overlain by the surficial aquifer (layer 10). The most notable example of an absent aquifer is in the area updip of the limit of the Piney Point aquifer (layer 7). The Piney Point aquifer does not outcrop in Virginia, Maryland-Delaware, and New Jersey. The overlying and underlying confining units merge to control vertical ground-water flow between the lower Chesapeake (layer 8) and Beaufort-Aquia (layer 6) aquifers in areas where the Piney Point aquifer (layer 7) is absent.

HYDROLOGIC CYCLE

Numerous authors have discussed and described in detail the hydrologic cycle in the Atlantic Coastal Plain (Cohen and others, 1968, p. 56; Cushing and others, 1973, p. 7; Getzen, 1977, p. 4; Heath, 1984, p. 52). A generalized schematic, illustrating the various components of the hydrologic cycle in the northern Atlantic Coastal Plain, is shown in figure 2.

The principal source of freshwater for the northern Atlantic Coastal Plain is precipitation. It falls directly on the Coastal Plain and on the adjacent Piedmont, where it provides a component of the surface-water flow in streams and rivers that cross onto the Coastal Plain. It occurs as rain, snow, and hail. Long-term average precipitation rates range from 40 in/yr (inches per year) along the Potomac River south of 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). A reasonable estimate for the entire northern Atlantic Coastal Plain is 47 in/yr. Precipitation

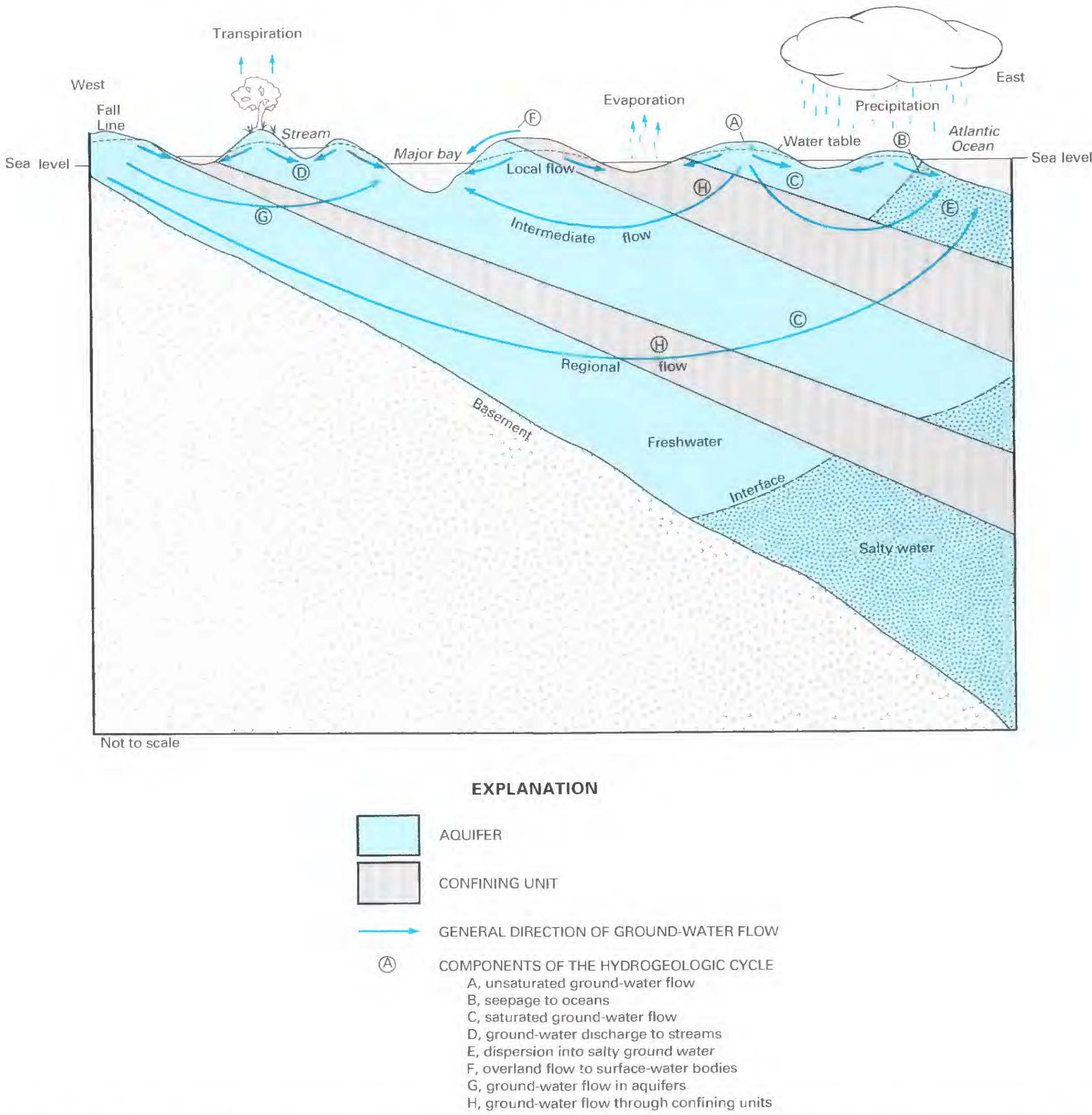


FIGURE 2.—Generalized hydrogeologic section of the northern Atlantic Coastal Plain showing the major components of the hydrologic cycle and an idealized prepumping flow system. Modified from Getzen (1977, fig. 2).

on the Coastal Plain is a source of evapotranspiration, overland flow to surface-water bodies, and infiltration to the soils and the water table. A large percentage of precipitation is either evaporated or transpired by vegetation and is returned to the atmosphere. Estimates of average annual evapotranspiration rates within the northern Atlantic Coastal Plain vary from a maximum of about 34 in/yr in northeastern North Carolina (Wilder and others, 1978, p. 19) to a minimum of nearly 21 in/yr on Long Island (Cohen and others, 1968, p. 37). Within the study area, the annual evapotranspiration rate decreases to the north, reflecting regional climatic trends. A reasonable regional estimate of the annual evapotranspiration rate for the study area is 24 in/yr.

A part of precipitation flows overland to surface-water bodies, including streams, lakes, and the ocean. The long-term average overland flow in the northern Atlantic Coastal Plain is areally variable. For example, Cushing and others (1973, p. 1) reported that the average streamflow for 15 long-term gaging stations on the Delmarva Peninsula is 15 in/yr, of which about 6.5 in/yr is overland flow. Heath (1975, p. 48) estimated the overland flow for the Albemarle-Pamlico area of the tidewater region (pl. 1) of North Carolina to be 14.5 in/yr. In contrast, Cohen and others (1968, p. 38) reported an average overland flow for streams on Long Island, N.Y., of about 0.5 in/yr, or about 5 percent of the total long-term streamflow. An estimate of this annual overland flow for the entire Coastal Plain is 8 in/yr.

Another part of precipitation infiltrates into the ground, if the soil is permeable and unsaturated. Initially, as the recharge enters the ground, any deficiency in soil moisture resulting from transpiration by plants or direct-surface evaporation is replenished. Thereafter, excess recharge percolates through the unsaturated zone crossing the water table to the saturated ground-water reservoir (aquifers). Areal infiltration rates vary widely, depending on such factors as topography, land use, intensity and duration of rainfall, existing soil-moisture conditions, and the type of soil and geologic material through which the water flows.

Winter base flow of streams approximates the amount of ground-water recharging the unconfined aquifer system, if the effects of ground-water withdrawals are negligible (a steady-state condition). Average base flow ranges from about 40 to 95 percent of total streamflow, or 6.7 to 15.9 Ggal/d (billion gallons per day), for 20,000 mi^2 in the Coastal Plain of the Mid-Atlantic region (Sinnott and Cushing, 1978, p. 15). This estimated range in base flow, or ground-water discharge, is approximately 7 to 16.7 in/yr, which is in general agreement with the ground-water recharge of 11 in/yr estimated for northeastern North Carolina by Wilder and others (1978, p. 18) and 14 in/yr estimated for coastal and southeastern Delaware (Johnston, 1977, p. 7). In contrast, a 26-year average of infiltration, or ground-water recharge, for Long Island, N.Y., is nearly 23 in/yr (Cohen and others, 1968, p. 44). Assuming that ground-water discharge to ungaged, salty surface-water bodies occurs at about the same rate as that measured for streams, these estimates are applicable to the entire study area. These values vary because of estimation techniques as well as areal trends. On a more local scale, infiltration rates vary from about 1 in/hr (inch per hour) in forests on sandy soil to zero in paved areas (Heath, 1983, p. 5). For purposes of estimating a regional water budget, 15 in/yr for ground-water recharge was assumed.

Some water infiltrating to the water table moves relatively quickly along local hydraulic gradients to nearby streams and other surface-water bodies. Other water moves more slowly, both vertically and laterally, down the more regional hydraulic gradient. This ground water eventually discharges into streams, lakes, bays, and the Atlantic Ocean. The hydrologic cycle is completed by evaporation of this discharge into the atmosphere.

Since the beginning of the 20th century, ground water has been used increasingly as a source of freshwater in the study area. A water-use inventory estimated ground-water withdrawals for 1980 at about 1,210 Mgal/d (million gallons per day), or an average rate of about 0.5 in/yr over the entire emerged northern Atlantic Coastal Plain. These withdrawals primarily represent major public supply and industrial uses and do not include individual minor withdrawals for irrigation, rural, and domestic uses, which were assumed to have an insignificant effect on the regional ground-water flow system. Although the individual minor withdrawals do not significantly affect the regional flow system, their total is significant. For example, in the Maryland and Delaware parts of the Delmarva Peninsula during 1985–86, withdrawals from the unconfined aquifer were about 52 percent of the total withdrawal (158 Mgal/d) from the aquifer system (Hamilton and others, 1989, p. 12). Major ground-water withdrawals account for less than 1 percent of the average 47 in/yr of precipitation on the Coastal Plain. However, these withdrawals do have an effect on the regional ground-water flow system.

REGIONAL WATER BUDGET

In defining the ground-water flow system, it is necessary to develop a regional water budget. An estimate of an areally averaged hydrologic budget is presented in the following equation:

Precipitation = Evapotranspiration + Overland flow + Base flow (ground-water discharge, which is assumed to be equal to ground-water recharge under steady-state conditions)

47 in/yr = 24 in/yr + 8 in/yr + 15 in/yr

Pumpage, estimated at 0.5 in/yr, is a small fraction of base flow and therefore is not included in the regional water budget.

This estimate provides the preliminary basis for developing a conceptual model of the geohydrology of the northern Atlantic Coastal Plain. The numbers in the equation are estimated long-term areal averages. Local budget estimates vary significantly from area to area in the Coastal Plain.

GENERALIZED GROUND-WATER FLOW SYSTEM

Qualitative definition of the ground-water flow system provides practical insights about the areal distribution of recharge to and discharge from the aquifer system. Bredehoeft and others (1982, p. 297) summarized six concepts that provide a basic understanding of regional ground-water flow. Two concepts that are important in defining flow in the northern Atlantic Coastal Plain are as follows: (1) Differences in hydraulic head produced by topographic relief at the boundaries of the flow system are the most significant driving force for ground-water flow, and (2) Earth materials that have finite permeability allow water to move laterally and vertically. The first concept suggests that topography is an important factor in defining the areal distribution of recharge to and discharge from the northern Atlantic Coastal Plain aquifer system. The second concept suggests that flow through confining units is an important factor to analyze in multilayer aquifer systems such as the northern Atlantic Coastal Plain.

Under prepumping conditions, the water table closely paralleled trends in topography. A map of the long-term average water table under prepumping conditions was constructed for the entire study area (fig. *A*, pl. 2). Ground-water recharge to the shallow parts of the unconfined aquifer occurred over the entire emerged Coastal Plain. Recharge to shallow confined aquifers and deep parts of the unconfined aquifer generally occurred at the topographic highs, and discharge at the topographic lows. The following description refers to the regional flow system associated with the shallow confined and deep unconfined parts of the aquifer system.

On Long Island, the prepumping aquifer system was recharged mostly along the axis of the island. Discharge took place along the northern and southern shores, to Long Island Sound and the ocean. Discharge along the northern shore was constrained to a narrow band because of the relatively high topographic relief and the less permeable character of the sediments. In contrast, discharge along the southern shore probably occurred in a much wider band.

In New Jersey, the topography suggests that prepumping recharge was concentrated along the Fall Line between Trenton and Raritan Bay and at two topographic highs in the southwestern and northeastern Coastal Plain of New Jersey. Discharge was offshore to the Atlantic Ocean and to topographically low areas along estuaries, rivers, and Raritan and Delaware Bays. The Delaware River along the Fall Line between Trenton, N.J., and Wilmington, Del., is an elongate topographically low area and was a discharge area. It coincides with the updip outcrop area of the Potomac-Raritan-Magothy aquifer system.

In Maryland and Delaware, topography is highest along the Fall Line and lowest near the ocean, Delaware and Chesapeake Bays, the Potomac and Patuxent Rivers, and numerous smaller rivers. Throughout the western shore of Maryland and the northern Delmarva Peninsula, prepumping ground-water recharge was primarily in a band along most of the Fall Line, except where surface-water bodies crossed the Fall Line. These surface-water bodies, Delaware and Chesapeake Bays, and the Potomac River were discharge areas. Recharge also occurred along the north-south-trending axis of the Delmarva Peninsula, near the topographic high that separates the Delaware Bay-Atlantic Ocean surface-water drainage from the Chesapeake Bay drainage. Discharge areas were predominantly located along rivers, streams, and marshes that fringe Delaware and Chesapeake Bays and the Atlantic Ocean.

On the western shore of Maryland, ground water discharged from the aquifer system to the Potomac River valley. This situation was similar to the Delaware River discharge area in southern New Jersey, except that a much shorter length of the Potomac River flows along the Fall Line. Ground water also discharged along the Patuxent, Patapsco, Gun Powder, and Susquehanna Rivers (pl. 1). The western shore rivers are topographically low and are characterized by high-relief bluffs. Ground water recharged along the bluffs, with a large percentage discharging to adjoining river valleys.

The most significant surface-water feature in the Maryland-Delaware part of the Coastal Plain is Chesapeake Bay. The bay trends nearly north-south across the Coastal Plain, overlying the outcrops of most of the aquifers. The bay is several miles wide in many places; under prepumping conditions, it was a major discharge area, which separated the shallow freshwater ground-water system of the Delmarva Peninsula from that of the western shore of Maryland.

On the inner coastal plain of North Carolina and the area west of the Suffolk Scarp in Virginia (pl. 1), the topography is similar to the topography of the western shore of Maryland. Numerous rivers flow across the outcrops of aquifers, including the Potomac, Rappahannock, York, James, Roanoke, Chowan, Tar, Cape Fear, Neuse, and Pamlico Rivers (pl. 1). In many areas, high bluffs adjoin the river valleys. Prepumping ground-water recharge occurred on the bluffs, whereas the valleys were discharge areas. In the tidewater region in North Carolina and the area east of the Suffolk Scarp in Virginia, the many bays, including Chesapeake Bay and Albemarle and Pamlico Sounds, and the ocean were also discharge areas.

In the preceding discussion, only the topography was considered to define qualitatively the areal distribution of ground-water recharge and discharge. That analysis applied mainly to the shallower parts of the confined aquifers and to deep parts of the unconfined aquifer. Although heads reflect areal changes in topography that are related to the distribution of recharge and discharge areas, heads also are affected by the hydraulic properties of aquifers and confining units. Generally, local topography has a greater effect on flow in shallow confined aquifers and in the deep parts of the unconfined aquifer than on deeper confined flow.

Simple analytical solutions to cross-sectional flow problems provide qualitative insights into the different flow patterns. Toth's (1963) theoretical analysis of flow in drainage basins defined the ground-water flow regime in a hypothetical hydrologic setting consisting of an isotropic aquifer bounded by topographic highs and lows, the lows representing bodies of surface water. His study showed three types of flow systems in such settings: local, intermediate, and regional. Flow systems of the same type are separated by nearly vertical boundaries or divides, while flow systems of different types are separated by almost horizontal boundaries. The higher the topographic relief, the more important the local system. Ground-water flow velocity is significantly slower in the regional flow system than in the local flow system because of differences in head gradients.

Although Toth's conclusions were derived for small drainage basins, they can be extended to the ground-water flow regime of aquifer systems as large as the northern Atlantic Coastal Plain aquifer system. Back (1960, 1966) and Geraghty (1960) showed idealized cross sections through the Coastal Plain in southern Maryland and Long Island that illustrate the presence of the three types of flow systems. A generalized cross section from the Fall Line to the limit of the freshwater aquifer system and the three types of flow systems are shown in figure 2.

Prepumping flow in the confined parts of the deep freshwater aquifers was primarily intermediate or regional. Most prepumping recharge to the unconfined part of the ground-water system rapidly discharged to the nearest streams or depression along local and shorter flow paths. Only a small percentage of the available water recharged the intermediate and regional flow system in the confined aquifers under prepumping conditions. However, shallow local flow systems include parts of deeper aquifers where unconfined aquifers are in direct hydraulic contact with an underlying aquifer, with no intervening confining unit. In these areas, the amount of prepumping ground-water recharge to and discharge from the aquifers was significant.

The unstressed flow system can be defined if sufficient prepumping head data are available to construct potentiometric surfaces for the various aquifers. These data generally were inadequate to develop a comprehensive flow system of the entire northern Atlantic Coastal Plain aquifer system. However, considerable data were available for some areas, and potentiometric surfaces for individual aquifers in these areas have been constructed.

Getzen (1977) presented prepumping potentiometric surfaces for the lower part of the Magothy (layer 3) and for the water table in the upper glacial aquifer (layer 7), and Garber (1986) showed a prepumping potentiometric surface for the Lloyd aquifer (layer 2) on Long Island. In New Jersey, Zapecza and others (1987, figs. 4–11) presented maps of the prepumping potentiometric surfaces for the Holly Beach (layer 10), Kirkwood-Cohansey (layers 8 and 9), confined Kirkwood (layer 8), Piney Point (layer 7), Vincentown (layer 6), Wenonah-Mount Laurel (layer 5), Englishtown (layer 4), and upper Potomac-Raritan-Magothy (layer 3) aquifers.

In Maryland and Delaware, prepumping potentiometric surfaces were constructed from historic heads and published sources for the following aquifers: Cheswold (locally part of the Chesapeake Group) (layer 8) (Leahy, 1982a, p. 23), Piney Point-Nanjemoy (layer 7) (Williams, 1979, p. 20; Chapelle and Drummond, 1983, pl. 6), Aquia-Rancocas (layer 6) (Chapelle and Drummond, 1983, pl. 3), Brightseat and Magothy (layer 3) (Mack, 1974, fig. 17), and the Patuxent and Patapsco (layers 1 and 2) (Martin, 1984, p. 29–31; Otton and Mandle, 1984, p. 27; Chapelle, 1985, p. 14). One notable feature of these potentiometric surfaces was the low prepumping heads in the deeper aquifers underlying the Delmarva Peninsula. Vertical head gradients beneath the Chesapeake Bay suggested that the bay was a significant discharge area not only for the shallow local ground-water flow system but also for the deeper regional flow system. The presence of this regional drain caused the low heads in deep aquifers beneath the Delmarva Peninsula. A consequence of these low heads was the presence of shallow saline ground water beneath a large area of the Delmarva Peninsula at depths of about 500 ft below sea level (Cushing and others, 1973, fig. 26); in contrast, under the rest of the Coastal Plain, fresh ground water extends to basement because saline water was kept farther seaward by higher freshwater heads.

In Virginia, prepumping head measurements for the confined aquifers were sparse. Thus, regional maps showing prepumping potentiometric surfaces for individual aquifers were not constructed. Some data were available from Sanford's (1913) report. Prepumping potentiometric surfaces were constructed for the upper artesian aquifer (layers 7 and 8) in the Northern Neck Peninsula (Newton and Siudyla, 1979, p. 33) and for the principal artesian aquifer (layers 1, 2, 3, and 6) in the Middle Peninsula (Siudyla and others, 1977, p. 55). In addition, Cosner (1975, p. 28) simulated a combined prepumping potentiometric surface for the Cretaceous aquifers (layers 1, 2, and 3) in the Franklin area of southeastern Virginia.

In North Carolina, Giese (in press) presented prepumping head measurements compiled from data given by Clark and others (1912) for the surficial (layer 10), Yorktown (layer 9), Pungo River (layer 8), Castle Hayne (layer 7), Beaufort (layer 6), Peedee (layer 5), Black

Creek (layer 4), upper Cape Fear (layer 3), and lower Cape Fear (layer 2) aquifers.

Estimated prepumping heads derived from a series of observation wells drilled during the 1960's along the North Carolina-South Carolina State boundary were used to determine the flow in the aquifer system at the southern edge of the study area (Aucott and Speiran, 1985). The conclusion was that ground water generally was flowing from North Carolina to South Carolina, with the Pee Dee River acting as a major discharge point to the flow system. The exception was in the lower Cape Fear aquifer (layer 2), the basal aquifer in the southern North Carolina Coastal Plain. Flow in this aquifer was from South Carolina northward into North Carolina (LeGrand, 1955). Heads greater than 100 ft above sea level have been measured in the lower Cape Fear aquifer in southern North Carolina. Peek and Register (1975, p. 13) suggested that these apparently anomalous heads might have resulted from compression of the sediments, principally the clay, by past tectonic forces; the compression of the overlying clay might have isolated flow in the lower Cape Fear aquifer from the remainder of the aquifer system. Examination of the South Carolina prepumping head data, however, strongly suggests that the high heads in southeastern North Carolina were not anomalous, but were consistent with northward-trending flow originating in South Carolina and Georgia (Wait and others, 1986, fig. 137).

The development of ground water as a source of supply in the study area has caused a significant change in the flow system. Potentiometric surfaces have declined throughout the study area. Large cones of depression with head declines of greater than 50 ft have developed in many aquifers. In New Jersey, heads in the confined Kirkwood aquifer (layer 8), a primary source of supply for the shore communities of Ocean, Atlantic, and Cape May Counties, had declined to as much as 80 ft below sea level in 1978. Heads in the Englishtown and Wenonah-Mount Laurel aquifers (layers 4 and 5) had declined to more than 200 ft below sea level in Ocean and Monmouth Counties in 1978. Heads in the Potomac-Raritan-Magothy aquifer system (layers 1, 2, and 3) had declined to more than 80 ft below sea level in the Camden area and more than 70 ft below sea level in the Middlesex County area in 1978.

In Maryland and Delaware, a deep cone in the Cheswold aquifer (part of layer 8), with heads about 90 ft below sea level in 1975, is centered about Dover, Del. The Piney Point aquifer (layer 7) is heavily pumped in Kent County, Del., and Dorchester County, Md. This pumping has caused the coalescing of two regional cones of depression. The Aquia and Rancocas aquifer (layer 6) have major cones in southern Maryland near Patuxent Naval Air Station and in central Delaware near Smyrna and Odessa. Also, major cones occur in the Patuxent and Patapsco aquifers (layers 1 and 2) in the Baltimore industrial area and in New Castle County, Del. Heads in the Magothy aquifer (layer 3) have declined significantly in southern Maryland, in Anne Arundel County.

In Virginia, heads in the Potomac aquifers (layers 1, 2, and 3) have declined to as much as 200 ft below sea level near the city of Franklin and 60 ft below sea level near the town of West Point owing to industrial pumping. The two cones centered about these areas have coalesced, forming the most areally extensive cone in the study area. This regional cone extends over an area of 5,000 mi^2. Also, a localized cone has formed in the Alexandria area. Regional declines in excess of 50 ft have not been measured elsewhere in Virginia.

In North Carolina, pumping from the Castle Hayne aquifer has produced head declines of less than 50 ft in the adjacent Beaufort and Pungo River aquifers (layers 6 and 8). Dewatering associated with phosphate mining in Beaufort County has caused a significant decline in heads, to more than 150 ft below sea level, in the Castle Hayne aquifer (layer 7). The lower Cape Fear, upper Cape Fear, and Black Creek aquifers (layers 2, 3, and 4) have measured drawdowns of about 100 ft in Lenoir and Craven Counties. Also, the cone centered about Franklin, Va., extends into the northern part of North Carolina.

Theis (1940) defined the source of pumped water as a combination of (1) a decrease in discharge, (2) an increase in recharge, and (3) a reduction in aquifer storage. A reduction in aquifer storage, and possibly in confining-unit storage, has occurred in the Coastal Plain aquifer system, and is shown by declines from prepumping levels. Establishment of new equilibrium conditions in the deeper aquifer system after pumping will cause reduced stream base flow, reduced evapotranspiration, reduced discharge to the ocean and bays, or any combination of these. A reduction in discharge of ground water to a stream causes a reduction in base flow, or possibly a reversal of the hydraulic gradient between the aquifer and stream. A reversal of the hydraulic gradient results in a loss of streamflow to the ground-water flow system. Also, a reduction in the natural discharge of ground water may cause the freshwater-saltwater transition zone to move landward.

Maps showing potentiometric surfaces (pls. 15–17) and head data were compiled for all aquifers (layers 1–9). These maps were based on head data collected from 1978 to 1980. The potentiometric surfaces for 1980 are discussed in detail in a later section on simulation of the transient flow system. Only sparse data were available, and contouring a continuous potentiometric surface was not feasible. Discrete data points are shown for Maryland, Delaware, Virginia, and North Carolina. In New

York, control for contouring the potentiometric surfaces of the Magothy aquifer (layers 3, 4, and 5) and the upper glacial aquifer (layers 6 and 7) was obtained from Getzen (1977). Head data for the Lloyd aquifer (layer 2) were obtained from Garber (1986). The potentiometric surfaces for aquifers in the New Jersey Coastal Plain were derived primarily from a 1978 synoptic head measurement (Walker, 1983). In Maryland and Delaware, head data were collected to update previously published potentiometric surface maps by Leahy (1979, 1982a), Martin (1984), Mack and Mandle (1977), Chapelle and Drummond (1983), and Williams (1979). In Virginia, only measured heads collected during 1979–80 are shown (Harsh and Laczniak, 1986). In North Carolina, heads were from measurements made during 1980 and 1981 (G. Geise, U.S. Geological Survey, written commun., 1982).

SUBSIDENCE CAUSED BY GROUND-WATER WITHDRAWALS

The withdrawal of ground water may cause land subsidence under certain conditions. Land subsidence is a result of compaction of unconsolidated sediments. Poland (1981) and Helm (1982) provided an overview of the theoretical aspects and a summary of the occurrence of land subsidence due to ground-water withdrawal in the United States.

Little information concerning subsidence in the study area is available. Davis (1987) reported that subsidence due to ground-water withdrawal is widespread in the Coastal Plain. He noted that the ratio of the amount of subsidence to the amount of head decline is quite consistent, ranging from 0.0069 in the Franklin area of Virginia to 0.0018 at Dover, Del., and Atlantic City, N.J. Based on repetitive leveling of benchmarks, land subsidence has been identified in several areas of the Coastal Plain, including the Franklin and West Point areas of Virginia, the Dover, Del., area, the Atlantic City, N.J., area, and the coastal Monmouth County, N.J., area (Davis, 1987, p. 69).

Land subsidence, or compaction of sediments due to withdrawals, can be measured using compaction recorders or extensometers (Poland and others, 1975, p. H48). In 1979 and 1980, extensometers were installed near Franklin, Va., and Atlantic City, N.J., areas of significant head decline. Figure 3 shows the measured compaction of sediments at these sites and head declines in nearby observation wells. In Virginia, the recorder measured the compaction occurring throughout all the sedimentary deposits above basement and the observation well measured head in the lower Potomac aquifer (layer 1). Data from the Franklin site show a compaction of 0.04 ft from January 1980 to November 1985. The average rate of compaction during this period was about 0.007 ft/yr (foot per year), or 0.08 in/yr. Heads in the lower Potomac aquifer (layer 1) declined 8 ft during this same period, but fluctuated as much as 10 ft annually owing to seasonal variations in withdrawals. Correlation between periods of head recovery and compaction recovery is good. However, because the compaction never fully recovered, it is apparent that there is an irreversible component of compaction that causes permanent subsidence of the land surface.

The New Jersey recorder measured compaction of the sediments overlying the Piney Point aquifer (layer 7). An observation well located 2.4 mi (miles) from the recorder (extensometer) measured head in the Kirkwood aquifer (layer 8). In New Jersey, compaction data have been collected since 1980. For the 34 months from September 1980 through June 1983, a compaction of 0.014 ft was measured at Atlantic City. The average rate of compaction was approximately 0.005 ft/yr, or 0.06 in/yr. During this period, heads in the observation well showed no long-term decline but a seasonal variation of 20 ft. As with the Virginia data, during periods of head recovery, an elastic component of compaction recovery was measured. During periods of head decline, irreversible (inelastic) compaction was also occurring.

A technique described by Poland (1981) was used to determine the storage coefficient related to the elastic deformation of the aquifers and confining units. Depth to water was plotted against measured compaction for the New Jersey data (fig. 4). This plot produced a typical series of stress-compaction loops. The reciprocal of the average slope of the trend of these loops is the component of the storage coefficient attributed to elastic deformation of the aquifers and confining units above the depth of the compaction recorder. The calculated elastic storage coefficient was 1.5×10^{-4} using the New Jersey data and 5.5×10^{-4} using the Virginia data. The Virginia data generally show that the transition from elastic to inelastic behavior (the elastic limit) occurs at lower heads as compaction increases. For example, the behavior changed at 7.5 ft below the initial head during the first compaction episode and at approximately 10 ft below the initial head during the second episode. In contrast, the New Jersey data showed that during all the stress-strain loops the transition occurred at a head of approximately 85 ft below land surface. These results are consistent with the head data at both sites. In Virginia, heads declined during the period of compaction, causing a rapidly increasing elastic limit (the point at which irreversible compaction begins). In New Jersey, heads showed no long-term decline, and the elastic limit changed very slowly. Although only heads for the period of the compaction data are shown, head data for Atlantic City were available from 1949. These data show that the heads in Atlantic City from 1980 to 1984 were slightly

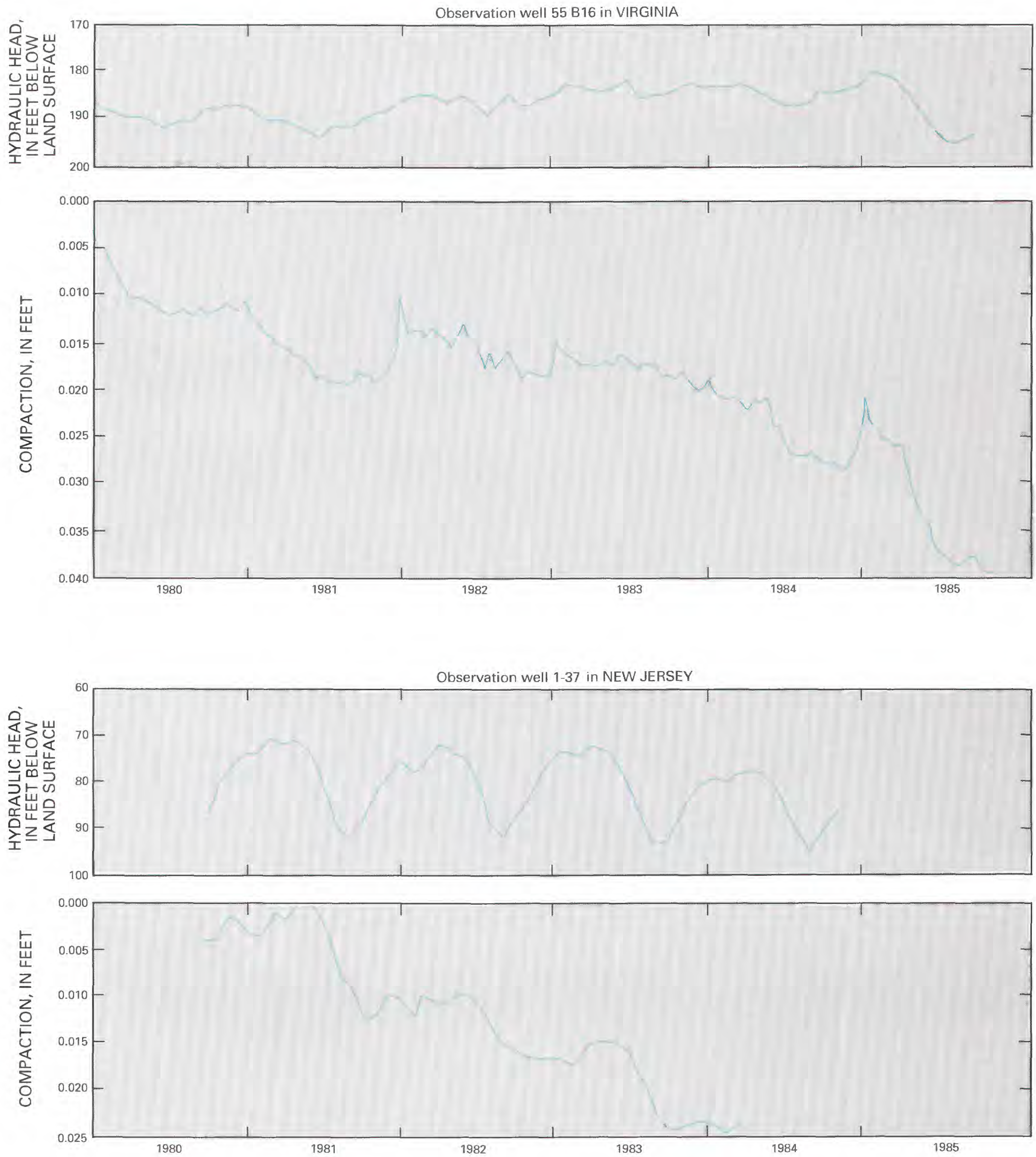


FIGURE 3.—Measured hydraulic heads and compaction at observation wells in Virginia and New Jersey.

above the minimum head of 87 ft below sea level that occurred during 1970.

The elastic storage coefficients based on compaction data are comparable to typical confined-aquifer storage. Data on the inelastic (nonrecoverable) storage coefficient are sparse for confining units in the study area. The specific storage of confining units based on both aquifer-test analysis and consolidation tests at a few locations

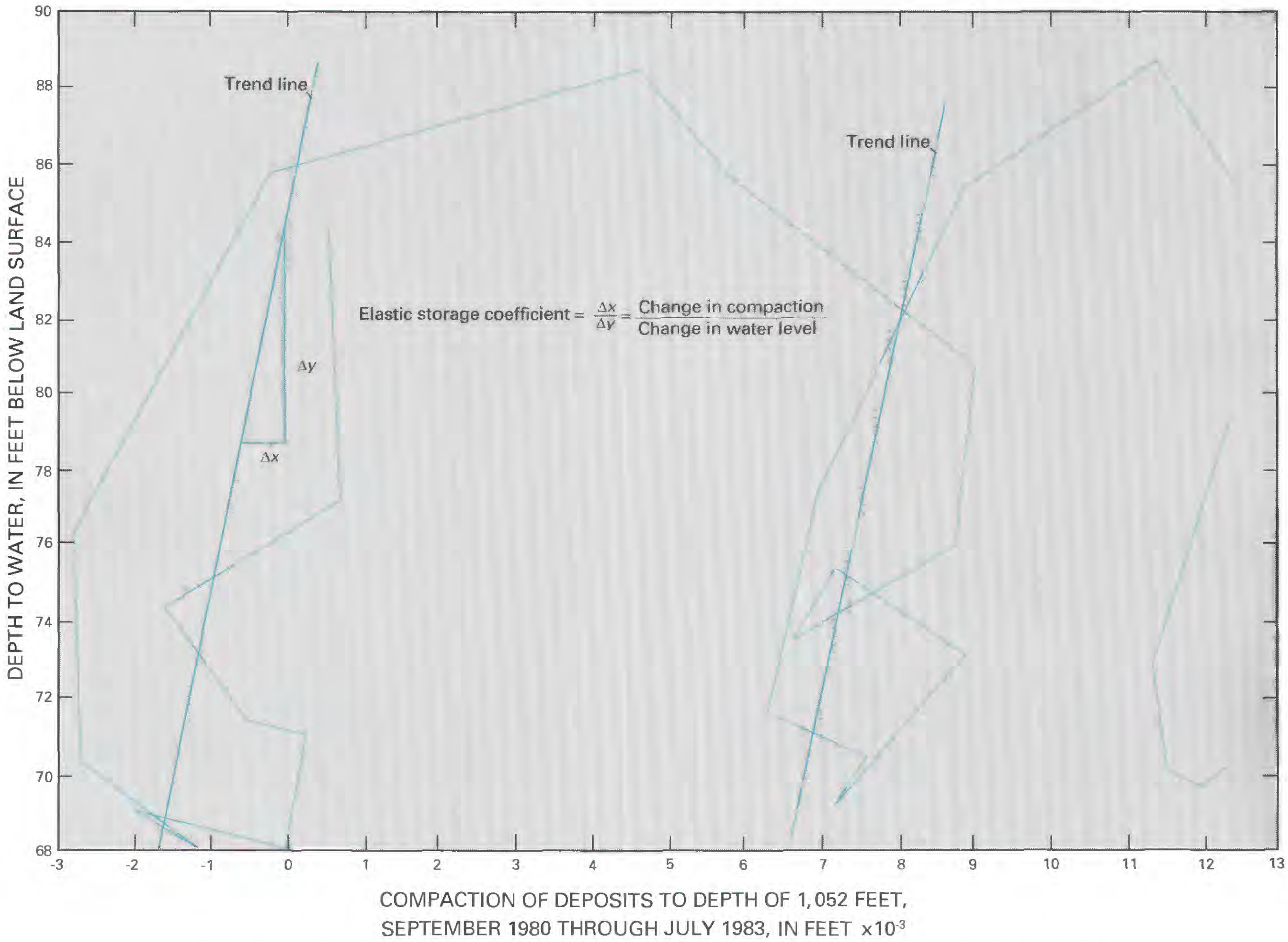


FIGURE 4.—Depth to water and compaction of deposits for observation wells near Atlantic City, N.J.

was 1.0×10^{-6} per foot. Based on analysis of compaction data for the Santa Clara and San Joaquin Valleys (Calif.), inelastic storage coefficients are much greater than the elastic storage coefficient. Green (1964) and Riley (1969) reported that the inelastic storage coefficients in the Santa Clara Valley were 40 and 80 times larger than the respective elastic storage coefficients. Ireland and others (1984, p. 49) reported nonrecoverable confining-unit storage coefficients for the San Joaquin Valley of 0.05 to 0.40, which are about 1,000 times greater than the storage for confined aquifers in the northern Atlantic Coastal Plain.

Compaction generally is the result of a one-time realignment of clay particles in confining units caused by a reduction in heads in the aquifer system. During the compaction phase, the higher inelastic specific storage may be the appropriate hydraulic characteristic of the system. After all possible compaction has occurred, the lower elastic specific storage is the appropriate hydraulic characteristic of the system. Because compaction can be affected both by declining heads due to pumping and by long-term geologic processes, it is uncertain which specific storage is appropriate for the individual confining units throughout the system. Data for Miocene sediments in New Jersey (fig. 3) and Cretaceous sediments in Virginia show that only minor compaction occurred in these areas during the short period of record. These sparse data suggest that the specific storage of the confining units is relatively small. Therefore, confining-unit storage was not simulated in the calibrated flow model, but a sensitivity analysis of confining-unit specific storage was made using values of 6.0×10^{-4}, 6.0×10^{-5}, and 6.0×10^{-6} per foot. This sensitivity analysis is discussed later in the section on "Model Reliability."

SALINE GROUND WATER

A major hydrologic feature of the northern Atlantic Coastal Plain aquifer system is the presence of saline ground water along the seaward margin of the system. Ground water containing chloride concentrations of less than 250 mg/L (freshwater) is separated from ground water containing chloride concentrations equal to seawater (18,000 mg/L) by a transition zone of varying chloride concentrations. The northern Atlantic Coastal Plain RASA study investigated the occurrence and distribution of the salty water (Meisler, 1981) as well as the regional effect of eustatic sea-level changes on the current position of the transition zone in the New Jersey Coastal Plain (Meisler and others, 1985). Lennon and others (1986) investigated the effect of rising sea level due to climatic changes on local ground-water flow conditions.

The regional transition zone where chloride concentrations are between 250 and 18,000 mg/L ranges in thickness from 400 to 2,200 ft. It has been discussed in detail by Meisler (1981, 1989). The zone is shallowest in North Carolina and deepens northward, attaining depths exceeding 3,400 ft in Maryland and New Jersey. Five large areas where the upper part of the transition zone (250 to 1,000 mg/L chloride concentration) is at depths of less than 400 ft have been delineated. These areas are (1) along Delaware Bay and adjacent southwestern New Jersey and northeastern Delaware, (2) along the eastern coast of the Delmarva Peninsula of Delaware, Maryland, and Virginia, (3) beneath lower Chesapeake Bay and adjacent parts of the York-James and Middle Peninsulas, Va., (4) beneath Albemarle Sound to eastern Pamlico Sound, N.C., and (5) along the Cape Fear River, N.C.

On Long Island, saline ground water containing chloride concentrations exceeding 1,000 mg/L is present in (1) the Magothy aquifer in southern Nassau and southeastern Queens Counties and (2) parts of the upper glacial aquifer and Magothy aquifer in eastern Suffolk County. These bodies of saline ground water, although areally extensive, result from local hydrologic conditions and do not reflect the areal configuration of the regional transition zone.

Ground water containing chloride concentrations of less than 5,000 mg/L extends as far as 55 mi off the New Jersey coast from Atlantic City. Two wells drilled to the confined Kirkwood aquifer 1.2 and 5.5 mi offshore in 1985 contained ground water containing chloride concentrations of 15 and 77 mg/L, respectively, at a depth of about 1,000 ft below sea level. The transition zone is closer to the shore farther south, as the depth to the zone decreases. Although the distances are unknown, saline ground water containing chloride concentrations greater than 5,000 mg/L is probably present no more than a few miles offshore in southern Virginia and North Carolina.

The location and configuration of the transition zone is controlled in part by the freshwater flow system. Typically, areas of major ground-water discharge (low heads) coincide with areas where the transition zone is shallow—Jamaica Bay, Delaware Bay, lower Chesapeake Bay, and the Cape Fear River. In contrast, areas of higher heads in the freshwater system cause the transition zone to be deeper.

Eustatic sea-level fluctuations also affect the position and movement of the transition zone. Meisler and others (1985) analyzed the effect of eustatic sea-level fluctuations on the fresh ground-water flow system using a ground-water flow model that simulated steady-state freshwater flow and the location of an idealized sharp saltwater-freshwater interface. The model was used to test the sensitivity of the interface position to anisotropy (the ratio of lateral to vertical hydraulic conductivity) and sea-level altitudes. Simulations that used hydraulic properties based on available data suggested that the transition zone in southern New Jersey and beneath the adjacent Continental Shelf is not in equilibrium with present sea level. The position of the transition zone reflects equilibrium conditions for a long-term average sea level of between 50 and 100 ft below present sea level. Therefore, the transition zone is moving landward in response to current sea-level conditions. The transition zone approximated by the sharp interface model was computed to be moving landward at a lateral velocity of about 0.2 mi per 10,000 years (Meisler and others, 1985).

The withdrawal of ground water has caused localized intrusion of saline water throughout the aquifer system. On Long Island, Cohen and Kimmel (1970) attributed increasing chloride concentrations in several wells tapping the Magothy aquifer and Jameco aquifer (in the Pleistocene Jameco Gravel) (layers 3, 4, and 5) in southeastern Queens and southern Nassau Counties to heavy withdrawals near the transition zone. In addition, in western Suffolk County, increased withdrawals from the upper glacial aquifer (layers 6 and 7) contain increased chloride concentrations (U.S. Geological Survey, 1984, p. 178).

In New Jersey, Schaefer (1983) noted that saltwater intrusion into the middle Potomac-Raritan-Magothy aquifer (layer 2) in Middlesex County and into the upper Kirkwood-Cohansey aquifer (layer 9) in Cape May County has been reported for about 40 years. Production wells in both areas have been abandoned because of increasing chloride concentrations. Schaefer also noted that several other areas, including parts of Monmouth, Gloucester, Salem, and Ocean Counties, are slightly affected by saltwater intrusion. Some wells in these areas contain increasing chloride concentrations owing to withdrawals from the Potomac-Raritan-Magothy aquifer system (layers 1, 2, and 3).

In Maryland and Delaware, Cushing and others (1973, p. 49) discussed saline-water intrusion on the Delmarva Peninsula. The potential for intrusion exists in several areas, including along the Chesapeake and Delaware Canal, Delaware and Chesapeake Bays, and the Atlantic Ocean. No evidence of intrusion has been noted at any pumping centers, with few exceptions. Saline water has moved into wells tapping the surficial aquifer (layer 10) at Lewes, Del., and into wells in the upper Chesapeake aquifer (layer 9) in several coastal communities in Sussex County, Del., and Worcester County, Md. (Hodges, 1984, p. 23). In the Baltimore area, saline-water intrusion has been noted in both the Patuxent and Patapsco aquifers (layers 1 and 2) (Chapelle, 1985, p. 15–18). The intrusion is caused by ground-water withdrawals near the aquifer subcrops beneath the Patapsco River.

In Virginia, Larson (1981, p. 20) reported that a few wells in the Potomac aquifers (layers 1, 2, and 3) on the York-James, Middle Neck, and Northern Neck Peninsulas showed significant increases in chloride concentrations. The source of this saline water may be the overlying bays and estuaries.

In North Carolina, saline-water intrusion due to ground-water withdrawals is not yet a serious problem. However, because of large withdrawals from the Castle Hayne aquifer (layer 7) in northeastern North Carolina, the potential exists for saltwater intrusion into the surficial, Yorktown, Pungo River, and Castle Hayne aquifers (layers 10, 9, 8, and 7) (Wilder and others, 1978, p. 71). Some agricultural land in Tyrrell, Dare, Hyde, Beaufort, Pamlico, and Carteret Counties has been affected by intrusion of saltwater from drainage canals into the surficial aquifer (layer 10) (U.S. Geological Survey, 1984).

In this study, the regional aspects of the ground-water flow system were investigated. A detailed analysis of local saltwater intrusion was beyond the scope of the study. The importance of the location of the transition zone in the definition of the ground-water flow system, however, is discussed in a later section of this report in terms of how the simulated seaward limit of the freshwater flow system affected the model results.

SIMULATION OF GROUND-WATER FLOW

COMPUTER CODE

Ground-water flow in the northern Atlantic Coastal Plain aquifer system was simulated using a multilayer finite-difference model. The model program, written in FORTRAN computer language, has been described by Leahy (1982b). It is a modification of a computer program for simulating ground-water flow in three dimensions by Trescott (1975), who described in detail the theoretical development of the finite-difference approximation to the ground-water flow equation. The modification permitted better simulation of aquifer and confining-unit pinchouts and reduced computer-memory requirements for the input data. The model program also included an option for simulating transient leakage from confining units described by Posson and others (1980).

APPROACH

The approach used in this study was to simulate ground-water flow at both a regional and a subregional scale. The reasons for modeling the system on more than one scale were to (1) provide greater resolution in model results with the smaller scale of the subregional models, (2) better calibrate the regional flow model by incorporating data from the calibrated subregional models, and (3) provide more realistic lateral boundary conditions for the subregional models. In addition to the regional model, four areally smaller, or subregional, models were developed for North Carolina, Virginia, Maryland-Delaware, and New Jersey.

The regional hydrogeologic framework of Trapp (in press) defined a layered sequence of 12 major aquifers and 9 confining units that formed the basis for the 10 layers of the quasi-three-dimensional flow model. The quasi-three-dimensional approach assumed that (1) within aquifers, flow is essentially parallel to the structural trends of the aquifer (that is, flow is nearly horizontal), (2) the vertical components of flow between aquifers are controlled by the hydraulic characteristics of the confining unit, and (3) there is no horizontal flow within confining units.

The model was designed to simulate both steady-state and transient conditions. Prior to pumping, long-term average head conditions prevailed in the system, and the system was in the state of equilibrium. Therefore, a steady-state simulation was used for prepumping conditions. After pumping began and at present (1980's), heads have declined and are declining in the aquifer system. The modeling approach involved simulating the transient response of the aquifer system from the beginning of significant pumping, about 1900, through 1980.

The modeling approach involved the use of two different types of overlying boundary conditions. These boundary conditions were used to most accurately define the prepumping and transient behavior of the system and are discussed in detail in the section on "Boundary Conditions." The model was calibrated using measured heads and estimated flows to define the areal distribution of hydraulic characteristics, particularly aquifer transmissivity and confining-unit leakance. Sensitivity analysis was performed to assess the effect of ranges of hydraulic characteristics on model behavior and, thus, to determine the accuracy of the calibration.

MODEL DESIGN

Model design involved defining boundary conditions, defining spatial and temporal discretization, assigning values for the hydraulic characteristics, and defining the relation between the subregional and regional models.

The sources of data for the model were evaluated during model design, and an analysis of these data determined the spatial discretization and initial estimates of hydraulic characteristics of the aquifer system.

SPATIAL DISCRETIZATION

The finite-difference technique requires that the area be divided into discrete blocks, or cells. The grid used in this study, shown in figure *B*, plate 2, was designed with consideration of the hydrogeologic framework and the conceptualized ground-water flow regime of the aquifer system.

The spatial discretization has 85 rows and 32 columns (fig. *B*, pl. 2). The typical block is 7 mi in each direction, giving a nodal area of 49 mi^2. In general, the grid spacing was uniform for the emerged parts of the Coastal Plain, reflecting a uniform areal distribution of withdrawals and hydrogeologic data within the study area. The grid spacing increased in the seaward parts of the study area and toward the northeastern and southwestern boundaries of the finite-difference grid. The increase in grid spacing reflects a lack of data on which to calibrate the model or a lack of need for resolution in these areas.

The conceptual framework of 10 regional aquifer layers separated by 9 intervening confining units constituted the vertical discretization used in the model. None of the aquifers or confining units is continuous over the entire study area. Appropriate hydraulic characteristics were specified to ensure that the model framework corresponded to the hydrogeologic framework in areas where aquifers and confining units pinch out. The discretized areal distribution of aquifer outcrop areas is shown in figure 5. Blocks that represent areas where more than one aquifer crops out were simulated as the outcrop of only the uppermost aquifer. In this report, this areal distribution of aquifer outcrops is referred to as the "unconfined system."

BOUNDARY CONDITIONS

The modeled aquifer system was bounded laterally (with one exception) and below by no-flow boundaries. The bottom boundary corresponded to the sloping contact between Coastal Plain sediments and the underlying crystalline rocks or low-permeability sedimentary rocks of Paleozoic or early Mesozoic age. The Fall Line was represented by an irregular no-flow boundary to the north and west. The no-flow model boundary to the east and south represented the downdip truncation of aquifers by a facies change or by the location of an idealized saltwater-freshwater interface at the 10,000-mg/L chloride concentration line as determined by Meisler (1981). Use of the 10,000-mg/L chloride concentration line as a no-flow boundary assumes an idealized saltwater-freshwater interface. For modeling purposes, water containing a chloride concentration of less than 10,000 mg/L is considered freshwater of equal density and nonmixing with higher density saltwater. Actually, a transition zone of varying chloride concentration separates freshwater and saltwater. This transition zone moves in response to head changes in the freshwater flow system; however, a stationary, sharp freshwater-saltwater interface, assumed to be a no-flow boundary, probably is a reasonable approximation for both prepumping and transient conditions. The sensitivity of simulated heads to the position of this interface was tested and is described later in the section on model sensitivity.

The southwestern no-flow boundary coincided with a major ground-water divide beneath the Pee Dee River in South Carolina. However, in the basal aquifer in this area, the lower Cape Fear aquifer (layer 2), ground water appears to be unaffected by this divide, and flow is northeastward toward North Carolina from South Carolina (Aucott and Speiran, 1985, p. 742). Therefore, the southwestern model boundary used to represent the lower Cape Fear aquifer was simulated as a specified-head condition. The northeastern no-flow boundary of the model was the submerged limit of the Coastal Plain aquifer system and the estimated position of the 10,000-mg/L chloride concentration northeast of Long Island.

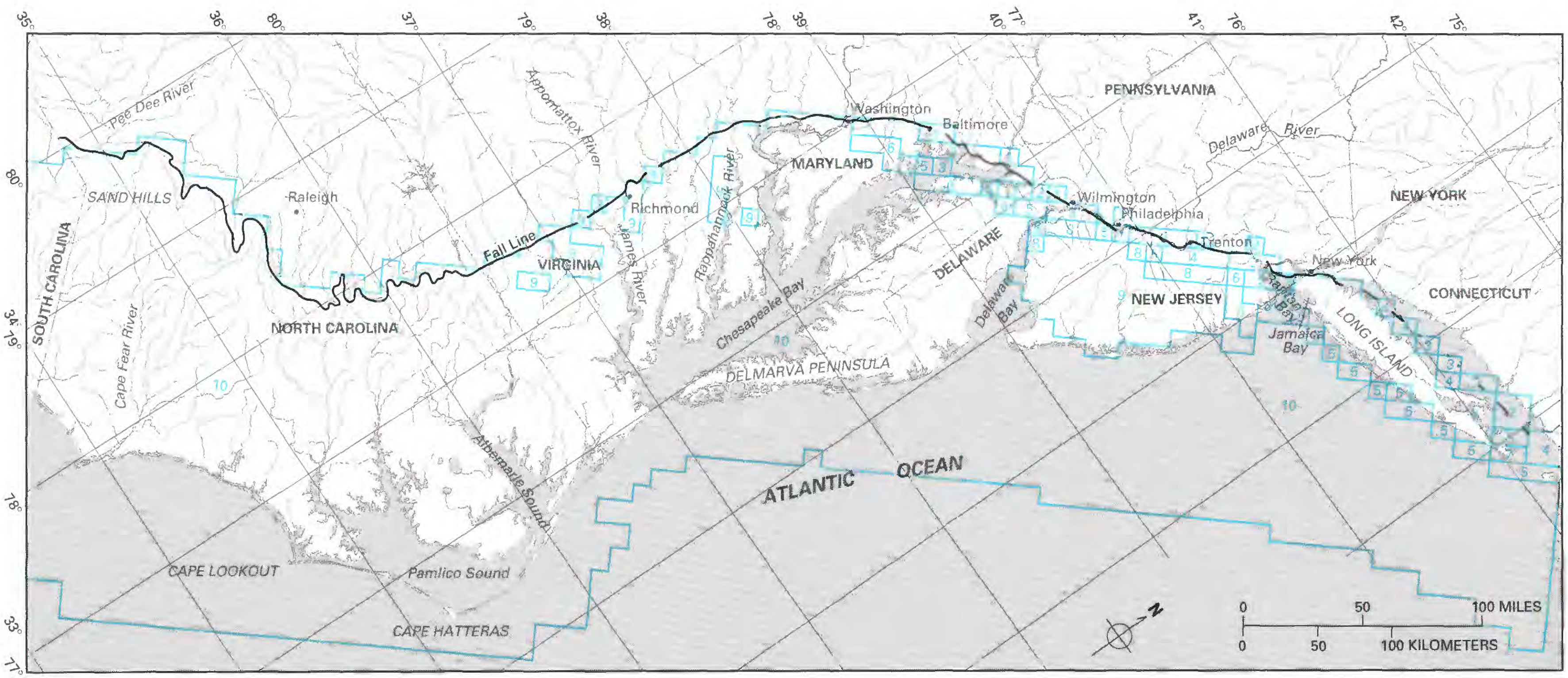
The top boundary was specified-head representing water-table altitudes in the 10-layer simulations used for prepumping conditions, and was specified-recharge with specified-head streams in subsequent 11-layer simulations used for pumping conditions. The latter boundary type is referred to in this report as a "modified specified-flux boundary condition." The modified specified-flux boundary condition was needed to accurately simulate the effects of ground-water withdrawals in the unconfined system. In the 10-layer simulations, the water-table layer was not actively simulated but acted as a source or sink layer. The model calculated the amount of water that moved downward to or upward from the confined aquifers as recharge or discharge. In the 11-layer simulations, the water-table layer was modeled as an active layer; thus, the effects of pumping in the unconfined aquifers could be simulated.

Figure 6 shows the generalized physical system and a conceptual model of the system as simulated with an overlying specified-head boundary. This model conceptualization was used initially to represent the prepumping flow system. An areally distributed specified-head boundary represented the water table onshore. Offshore and in tidal areas, the aquifer system was bounded at the bottom of the ocean and major bays by overlying specified heads.

Simulation of the prepumping flow system with the water table as a specified-head boundary yielded com-

EXPLANATION

9 SIMULATED OUTCROP AREA, MODEL LAYER NUMBER AND AQUIFER NAME
- 10, surficial
- 9, upper Chesapeake
- 8, lower Chesapeake
- 7, Castle Hayne-Piney Point (represents upper part of upper glacial aquifer on Long Island, N.Y.)
- 6, Beaufort-Aquia (represents lower part of upper glacial aquifer on Long Island, N.Y.)
- 5, Peedee-Severn (represents upper part of Magothy aquifer on Long Island, N.Y.)
- 4, Black Creek-Matawan (represents middle part of Magothy aquifer on Long Island, N.Y.)
- 3, upper Potomac and Magothy (represents lower part of Magothy aquifer on Long Island, N.Y.)
- 2, middle Potomac
- 1, lower Potomac



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

FIGURE 5.—Discretization of aquifer outcrops.

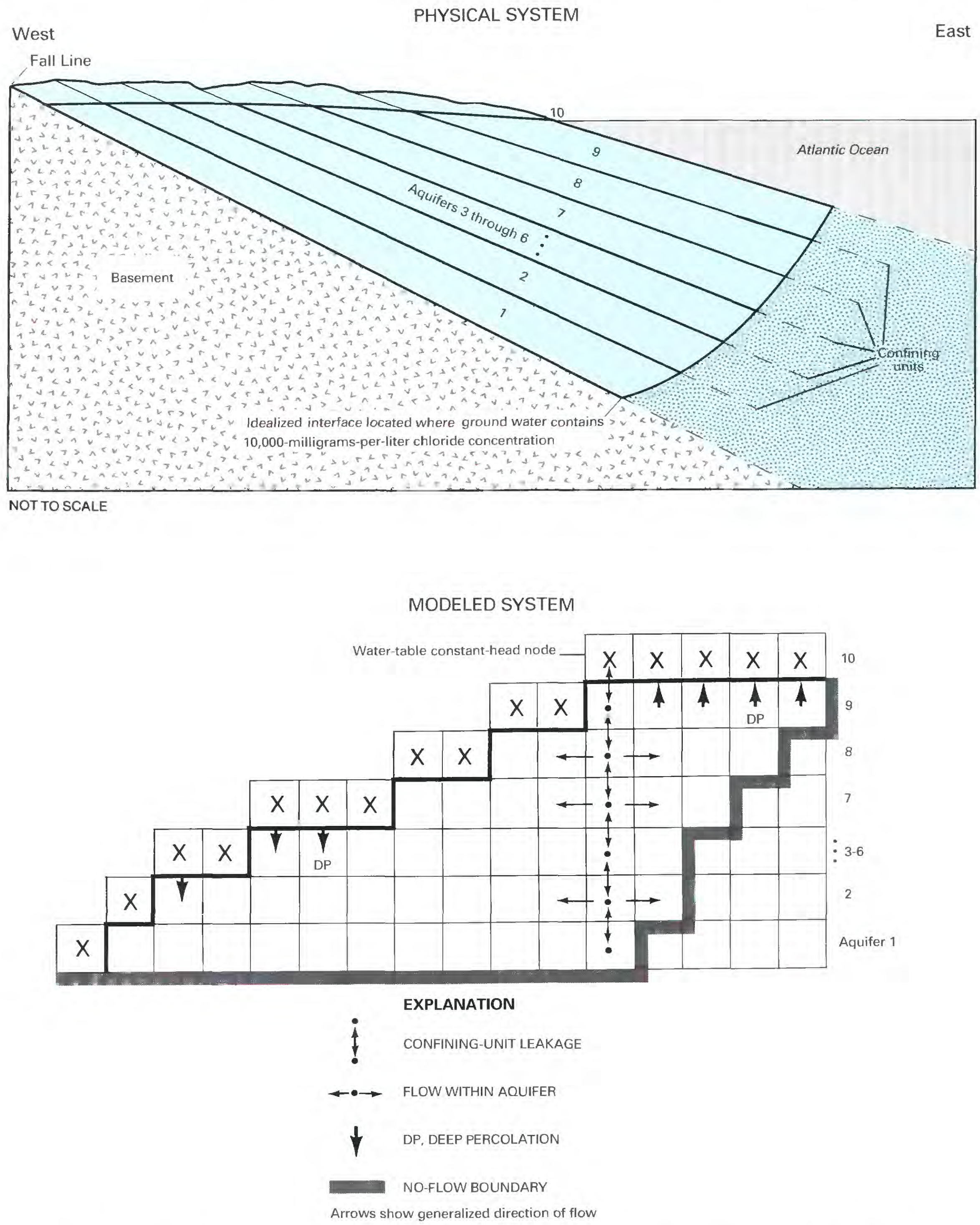


FIGURE 6.—Model conceptualization of water-table specified-head boundary used for prepumping conditions.

puted heads in every cell in the confined freshwater aquifer system and a value of flow to or from the specified-head cells. This flow represented the component of flow to or from the deeper aquifer system and is referred to in this report as "deep percolation." Simulated deep percolation was used as a known flux in the development of the second conceptualization of simulation of the flow system using the modified specified-flux boundary.

The modified specified-flux boundary permitted simulation of transient heads in and near the outcrop areas of the unconfined part of the aquifer system during pump-

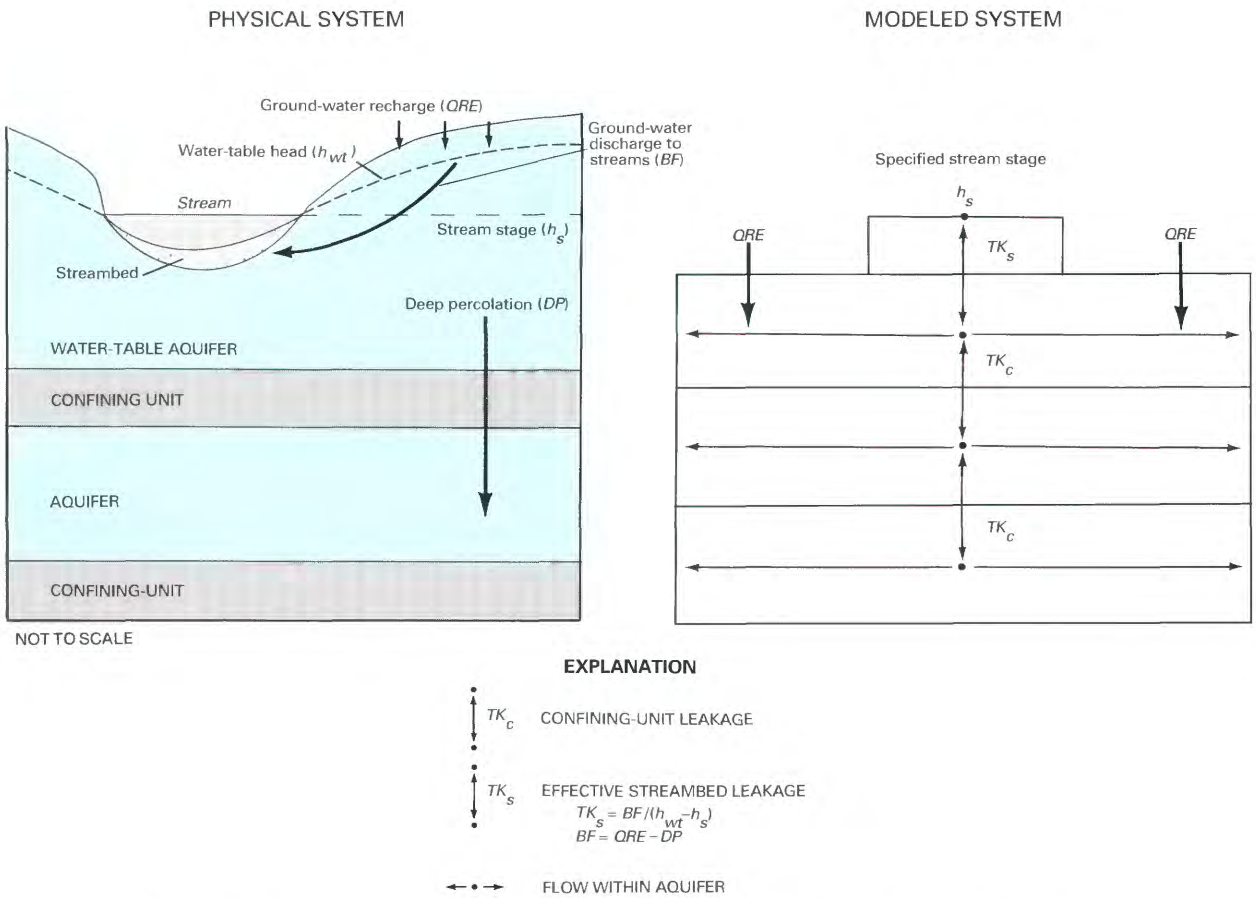


FIGURE 7.—Model conceptualization of modified specified-flux boundary used for transient conditions.

ing. In contrast, the specified-head boundary in the first conceptualization maintained a constant-head condition in the unconfined system throughout the simulation. This assumption is justified because water levels, in response to seasonal climate changes, fluctuate about a long-term average level.

Figure 7 shows cross sections of the physical system and modeled system with the modified specified-flux boundary. This conceptualization required specifying ground-water recharge to the water-table aquifer and simulating streams as specified heads to drain part of the recharge that discharges to the streams as base flow. Streambed leakance (vertical hydraulic conductivity divided by streambed thickness), together with the hydraulic gradient between the water table and stream stage, controls the amount of ground water that discharges to streams as base flow.

The modified specified-flux boundary, in contrast to the specified-head boundary, allowed simulation of lateral flow in the unconfined part of the aquifer system. However, heads in the unconfined part and values of deep percolation simulated for prepumping conditions should be the same using either of the two upper boundary conditions. Simulations using the modified specified-flux boundary, which included stream-stage elevations, deep percolation, and discharge to streams, allowed heads in the unconfined system to change with time in response to changing withdrawals.

In the unconfined part of the aquifer system, the simulation should consider the saturated thickness and its permeability during transient simulations. However, the head declines in this part of the aquifer system were assumed to be small relative to the saturated thickness. Owing to the high specific yield of the unconfined aquifer, therefore, it was reasonable to use transmissivity values instead of saturated thickness and permeability to simulate the unconfined part of the aquifer system on a regional scale.

TIME DISCRETIZATION

Simulation of the transient behavior of the aquifer system required discretization of time and time-dependent stresses, such as ground-water withdrawals. Withdrawals were discretized into pumping periods, whereas time was discretized with time steps. Withdrawals were changed abruptly at the start of each pumping period and were held constant for the duration of the pumping period. For the regional model, 10 pumping periods of varying duration were used and each pumping period was divided into 10 time steps of varying duration. The total simulation of 81 years was completed with 100 time steps. The first time step of each pumping period was the shortest, with a maximum of 45 days for the initial pumping period (of 21 years) and a minimum of 6.5 days for the final period (of 3 years). Each succeeding time step during a pumping period was 1.5 times longer than the preceding time step.

RELATION AMONG REGIONAL AND SUBREGIONAL MODELS

Four subregional models were calibrated as part of this study—New Jersey (Martin, in press), Maryland-Delaware (Fleck and Vroblesky, in press), Virginia (Harsh and Laczniak, 1986), and North Carolina (Giese and others, in press). The subregional and regional models were related in that the same hydrogeologic framework and hydraulic characteristics were used in the simulation. The major difference was that the subregional models represented the aquifer system using a finer finite-difference grid. The subregional model boundaries are shown in figure *B*, plate 2. Three of the subregional model grids (Maryland-Delaware, Virginia, and North Carolina) were aligned with the regional model grid. In these three models, four subregional grid blocks typically made up one regional grid block. The typical grid block area of the subregional models was 12.25 mi^2. The New Jersey model grid was rotated 8.5 degrees to the northwest relative to the regional model grid and had a typical block area of 6.25 mi^2. The subregional grid was aligned with the Fall Line in New Jersey to better represent the Delaware River and the aquifer outcrop areas.

The average value of four typical blocks of the subregional models provided hydraulic characteristics, including aquifer transmissivity, storage, confining-unit leakance, effective stream-stage elevations, and water-table altitudes, to a respective block of the regional model in their respective study areas. In areas where subregional models overlap, hydraulic characteristics were equivalent in each of the subregional models. Hydraulic characteristics for each regional grid block were calculated as the arithmetic mean of the subregional grid block values. The arithmetic mean is the correct approximation for most hydraulic characteristics; for values of transmissivity along a flow path, the harmonic mean should be used. A better approximation than the arithmetic mean for the regional value of transmissivity involves using the arithmetic and harmonic means in the row and column directions (along and perpendicular to flow paths). However, such averaging results in two directionally dependent values of transmissivity for each regional cell, thereby complicating the simulation computations. A comparison of the arithmetic and harmonic means of typical transmissivity values in the northern Atlantic Coastal Plain showed that the two means did not vary significantly because the transmissivity varied continuously and smoothly, and that the arithmetic mean was a reasonable approximation.

Subregional lateral boundary fluxes were computed from heads simulated by the regional model. Lateral fluxes were computed across grid block faces in the regional model that corresponded to the boundaries of the subregional models. The flux was computed from the simulated head gradient and the transmissivity in adjacent grid blocks across the appropriate lines. Each flux value represented the flow through the side of a grid block in the regional model. As noted earlier, the edges of grid blocks in the subregional models, except the New Jersey model, corresponded to the edge of each regional grid block.

Boundary fluxes calculated from the regional model were provided to the subregional models for prepumping conditions and for each pumping period during transient conditions. Ideally, fluxes should be continuously updated throughout the transient simulation. Because of operational constraints, however, fluxes were provided only for each pumping period. This procedure was assumed reasonable because pumping periods corresponded to major changes in withdrawal.

INPUT REQUIREMENTS

The input requirements for the simulation of flow in the northern Atlantic Coastal Plain aquifer system depended on the modeling approach and the conceptualization of the system. Table 2 summarizes the input requirements for the models with the specified-head and the modified specified-flux upper boundary conditions. The requirements for the prepumping simulation are given for both conceptualizations, and the requirements for transient conditions are listed for the modified specified-flux conceptualization.

Knowledge of the hydraulic characteristics of aquifers and confining units was necessary to describe the hydrogeology of the aquifer system. Transmissivity, which is the hydraulic conductivity of the aquifer multiplied by the thickness of the aquifer, describes the water-transmitting characteristic of aquifers. Leakance, which is the vertical hydraulic conductivity of the confining unit divided by its thickness, defines the water-transmitting characteristic of confining units. If the transient (time-dependent) behavior of the ground-water flow system is of interest, storage characteristics must be known. The storage coefficient, the specific storage of the aquifer or confining unit multiplied by the unit's thickness, describes the storage characteristics of aquifers and confining units.

Data on the hydraulic characteristics listed in table 2 came from analysis of point data for initial estimates of these hydraulic characteristics in the subregional models. The areal distributions of the estimated hydraulic characteristics were examined, and any discrepancies in overlap areas of the subregional models were reinterpreted to ensure hydrologic consistency throughout the regional model and among subregional models.

Additional data were required to analyze the effects of confining-unit storage and the seaward boundary of the system on model calibration. The effects of confining-unit storage or transient leakage were investigated, requiring data on confining-unit specific storage and confining-unit thickness. Also assessed was the treatment of the seaward boundary of the aquifer system as a static no-flow boundary at the 10,000-mg/L chloride concentration. The effect of this boundary condition was analyzed by means of a variable-density flow model. This model, which is discussed later in the report, required additional data—(1) the areal distribution of total dissolved solids in water, and (2) a temperature-depth profile for the aquifer system.

TABLE 2.—*Summary of input requirements for finite-difference model of the northern Atlantic Coastal Plain aquifer system*

Specified-Head Conceptualization (Prepumping Steady-State Conditions)

1. Water-table altitudes
2. Location of specified-head boundary
3. Transmissivity
4. Confining-unit leakance
5. Grid spacing

Modified Specified-Flux Conceptualization (Prepumping Steady-State Conditions)

1. Requirements 3, 4, and 5 of specified-head conceptualization
2. Stream altitudes
3. Location of specified constant stage to represent surface-water bodies
4. Streambed leakance
5. Ground-water recharge

Modified Specified-Flux Conceptualization (Transient Conditions)

1. All requirements of modified specified-flux conceptualization for prepumping conditions
2. Storage coefficient for aquifer
3. Time-discretization information
4. Location of wells and rate of pumping through time

HYDRAULIC CHARACTERISTICS

Table 3 shows the range in initial estimates of transmissivity for aquifers and of vertical hydraulic conductivity for confining units, by subregion, prior to model calibration. Transmissivities were based on aquifer-test analysis, specific-capacity tests, analysis of geophysical and lithologic logs, and the results of model calibration from previous studies in the northern Atlantic Coastal Plain. Transmissivity that was estimated from geophysical or lithologic logs was calculated by summing the products of the net thickness of each lithology and an appropriate hydraulic conductivity for each lithology. Hydraulic conductivity determined from specific-capacity tests was multiplied by total sand thickness to obtain aquifer transmissivity.

The transmissivity of the Coastal Plain aquifers ranged from about 98,000 ft^2/d (foot squared per day) for the Castle Hayne aquifer (layer 7), a limestone unit in North Carolina, to very small values where aquifers thin or pinch out. Transmissivity as high as about 22,000 ft^2/d occurred on Long Island for the upper and lower parts of the upper glacial aquifer (layers 6 and 7). In general, the transmissivity of most aquifers decreased southward from the northern limit of the study area into Virginia.

The vertical hydraulic conductivities for confining units shown in table 3 were based on sparse data. Thus, ranges of vertical hydraulic conductivity are not shown for all confining units. The confining-unit conductivities were based on (1) core analysis of small sample intervals within the confining unit, (2) aquifer-test analysis, and (3) values derived from previous modeling studies in the study area.

Values developed from previous modeling studies best describe regional confining-unit characteristics. The least reliable data for deriving regional characteristics were those derived from core analysis. Data from core analysis represent only a small sample interval which may or may not be representative of the entire confining unit. In general, the confining-unit vertical hydraulic conductivities decreased with depth and age. For the deeper units (that is, the confining units overlying layers 1 through 5), conductivities typically ranged from 10^{-6} to 10^{-4} ft/d (foot per day). In contrast, conductivities for the shallower units (that is, confining units overlying layers 6 through 9), ranged from 10^{-5} to 10^{-3} ft/d.

Storage coefficients have been determined from numerous aquifer tests throughout the study area. A typical storage coefficient for the confined aquifers was 1.0×10^{-4}, and a typical specific yield for the unconfined aquifers was 0.15.

WATER TABLE

The water table (fig. *A*, pl. 2) was a composite of heads from the unconfined portion of individual layers within

TABLE 3.—*Estimates of aquifer transmissivities and confining-unit vertical hydraulic conductivities for the northern Atlantic Coastal Plain prior to model calibration*

Model aquifer layer[1]	Aquifer transmissivity, in feet squared per day: North Carolina	Virginia	Maryland-Delaware	New Jersey	Long Island
10	100– 1,500	250– 1,000	[2]100–10,000	5,200– 7,800	Absent
9	30– 8,000	200– 3,000	500–15,000	860–25,900	Absent
8	100– 9,100	200– 4,000	350– 7,400	860–19,900	Absent
7	100–98,000	150– 2,000	25– 7,400	860– 5,200	100–22,000
6	100– 3,900	125– 1,000	250– 5,100	860– 3,500	100–22,000
5	100–12,700	300– 1,200	500– 2,200	90– 2,300	100–17,000
4	100–12,500	200– 3,300	500– 1,500	90– 5,400	100–17,000
3	100– 9,000	150–13,000	500– 3,000	860–19,900	100–17,000
2	100– 9,500	2,000–19,000	400– 9,200	860–21,600	100–13,000
1	100–11,000	6,000–55,000	400– 9,200	860–17,300	Absent
Confining unit above model aquifer layer	**Confining-unit vertical hydraulic conductivity, in feet per day: North Carolina**	**Virginia**	**Maryland-Delaware**	**New Jersey**	**Long Island**
9	2.5×10^{-3}	6.0×10^{-4}–4.0×10^{-3}	2.0×10^{-3}	4.0×10^{-4}	Absent
8	1.5×10^{-3}	3.0×10^{-6}–3.0×10^{-5}	6.0×10^{-5}–3.0×10^{-2}	4.0×10^{-4}	Absent
7	1.0×10^{-4}	9.0×10^{-6}	9.0×10^{-5}–7.0×10^{-3}	3.5×10^{-5}	Absent
6	5.0×10^{-3}	2.0×10^{-6}–4.0×10^{-3}	6.0×10^{-5}–3.0×10^{-4}	5.0×10^{-5}	Absent
5	9.0×10^{-5}	8.5×10^{-6}	6.0×10^{-5}	5.0×10^{-4}–1.0×10^{-1}	1.0×10^{-3}
4	8.0×10^{-5}	3.0×10^{-5}	2.0×10^{-4}	6.0×10^{-6}–1.0×10^{-1}	Absent
3	7.0×10^{-5}	8.5×10^{-4}	2.0×10^{-5}–4.0×10^{-3}	9.0×10^{-7}–3.0×10^{-2}	Absent
2	9.0×10^{-6}	3.5×10^{-6}	7.0×10^{-6}	9.0×10^{-6}–4.0×10^{-2}	1.0×10^{-3}
1	9.0×10^{-6}	2.0×10^{-6}	9.0×10^{-5}	9.0×10^{-6}–4.0×10^{-2}	Absent

[1]Regional aquifer names for the model layers are as follows: 10, Surficia; 9, Upper Chesapeake; 8, Lower Chesapeake; 7, Castle Hayne-Piney Point; 6, Beaufort-Aquia; 5, Peedee-Severn; 4, Black Creek-Matawan; 3, Upper Potomac and Magothy; 2, Middle Potomac; 1, Lower Potomac.

[2]Reported transmissivity values from aquifer-test analysis have been as high as 53,000 feet squared per day in buried channels on the Delmarva Peninsula. However, on a regional scale, the maximum value is about 10,000 feet squared per day.

the hydrogeologic framework. The unconfined portion consists of the surficial aquifer (layer 10) over most of the study area (fig. 5). Notable exceptions are on Long Island and in New Jersey, where the unconfined system consists mostly of the upper glacial aquifer (layer 7) and the unconfined Kirkwood-Cohansey aquifer (layers 8 and 9), respectively.

Because of the high storage capacity of the unconfined system, and the typically small withdrawals compared with recharge rates, it was assumed that no significant regional change in the water table had occurred and that the 1980 water table was similar to the prepumping surface. A water-table matrix was constructed for the finite-difference grid designed for the flow model. The grid was superimposed on topographic maps of the area. In the onshore area, the water-table altitude was estimated from the altitude of ponds, lakes, and streams within individual grid blocks, to supplement the long-term average heads from water-table observation wells. In the offshore area, the water table was assumed to be the freshwater-equivalent head computed from the ocean or estuary depth. In general, there was a freshwater-equivalent head of 1 ft for every 40 ft of depth in the ocean or estuaries. Bathymetric maps were used to define seawater depths.

The water table was contoured from the discretized water-table altitudes for the onshore section of the study area, and from the freshwater-equivalent heads for the offshore section (fig. *A*, pl. 2). The water table was generally very similar to the topography, with the exception of areas in the western part of the study area. In these areas, the topography has considerable relief, as much as 100 ft within an individual grid block, which is a result of deeply incised streams and rivers. The average water-table altitude in these grid blocks was considerably higher than the altitude of streams and rivers within the blocks. Water-table altitudes decreased from highs of about 550 ft above sea level in the Sand Hills area along the Fall Line of North Carolina, to sea level along the shore of the ocean, bays, and estuaries. The altitude of the freshwater-equivalent heads increased offshore as ocean depth increased.

GROUND-WATER RECHARGE

The ground-water recharge rate ranged from 15.0 to 22.5 in/yr for the simulations of prepumping and transient conditions when the water table was treated as a modified specified-flux boundary. The specified regional recharge varied areally by subregion, as follows: Long Island, 22.5 in/yr; New Jersey, 20.0 in/yr; Maryland-Delaware, 15.0 in/yr; Virginia, 15.0 in/yr; and North Carolina, 15.0 in/yr. Although recharge varied locally and some subregional models included this variation, it was assumed that a broadly regionalized specified recharge was adequate for regional analyses of the northern Atlantic Coastal Plain aquifer system.

Stream elevations

The elevations of streams were required for simulations that treated the water table as an active recharge layer. A stream existed in every onshore grid block in the regional model. Figure *C*, plate 2, shows estimated stream elevations. Although the stream elevations represented elevations only at the location of the streams, they were contoured to illustrate the regional trends of these streams. The stream elevations were derived by averaging corresponding grid block values specified in the subregional models. The subregional values were estimated from variously scaled topographic maps. Under prepumping conditions, the water-table altitude in each grid block was estimated as higher than the altitude of the stream; therefore, all streams were discharge points for the ground-water flow system. In general, stream elevations on the extensive flat seaward margin of the study area were less than 10 ft below water-table altitudes. In contrast, altitude differences were substantial, on the order of tens of feet, in the uplands along the inner margin of the study area, where major rivers and streams are deeply incised into Coastal Plain deposits.

Effective streambed leakance

A water budget for each grid block was used to compute an effective streambed leakance for each onshore block. Ground-water discharge to a stream block was calculated as the difference between the specified recharge and the simulated deep percolation from the 10-layer prepumping simulation (with the water table as a specified-head boundary). Using discharge, the specified altitude of the water table, and estimated areal and long-term average stream stage, an effective streambed leakance was calculated (fig. 7). Because of the large block size used in the grid, streams existed in every onshore grid block. One effective stream stage was chosen for each block to represent an areal and long-term average stream stage. The effect of regional lateral flow in the water-table aquifer was ignored in the calculation of streambed leakance.

For updip areas, calculated effective streambed leakance values were generally several orders of magnitude higher than confining-unit leakance values. Effective streambed leakance values ranged from about 1 to 1,000 (ft/d)/ft ((feet per day) per foot). Streambed leakance values were recalculated between calibration simulations whenever changes in the hydraulic characteristics caused changes in prepumping deep percolation.

Ground-water withdrawals

Ground-water withdrawals increased dramatically from the late 1800's to 1980. Figure 8 shows estimated ground-water withdrawals in the entire northern Atlantic Coastal Plain, and in individual subregions, from 1900 to 1980. The estimates were based on a recent inventory of historical and current users. The inventory primarily included major public-supply and industrial users, but in New Jersey and Virginia also included major irrigation users. Withdrawals for rural, minor domestic, and minor irrigation use were assumed to have had an insignificant effect on the regional ground-water flow system and were not included in inventory because they (1) tend to be from unconfined parts of the aquifers, (2) are usually small, are often seasonal, and are distributed widely over the study area, and (3) are generally not totally consumptive (rural and domestic water is returned to the unconfined aquifer through onsite wastewater-treatment systems (septic tanks), and irrigation water is seasonally applied directly to the unconfined parts of the aquifers and is returned to the ground-water flow system, with some evapotranspiration loss). Estimates from sparse data were used to construct the history of withdrawal prior to 1965. Estimates of withdrawals after 1965 are more reliable because most States in the study area have mandated the reporting of major withdrawals.

Ground-water withdrawals in 1900 totaled approximately 100 Mgal/d. Most of the withdrawals were from the Brooklyn and Queens areas of New York. Total ground-water withdrawals in Brooklyn and Queens in 1904 were estimated at about 56 Mgal/d (Buxton and others, 1981, p. 25).

Withdrawals in the northern Atlantic Coastal Plain have increased steadily, to more than 1,200 Mgal/d in 1980. Increases on Long Island occurred in the 1950's and 1960's with the suburban development of Nassau and Suffolk Counties. Withdrawals in New Jersey increased significantly from the 1940's to the 1970's. This increase corresponded to a period of industrial and suburban growth that began during World War II. In North Carolina, significant increases occurred in the 1960's, coinciding with development of phosphate mining in the Coastal Plain. In Virginia, Maryland, and Delaware, steady increases in withdrawals occurred during the

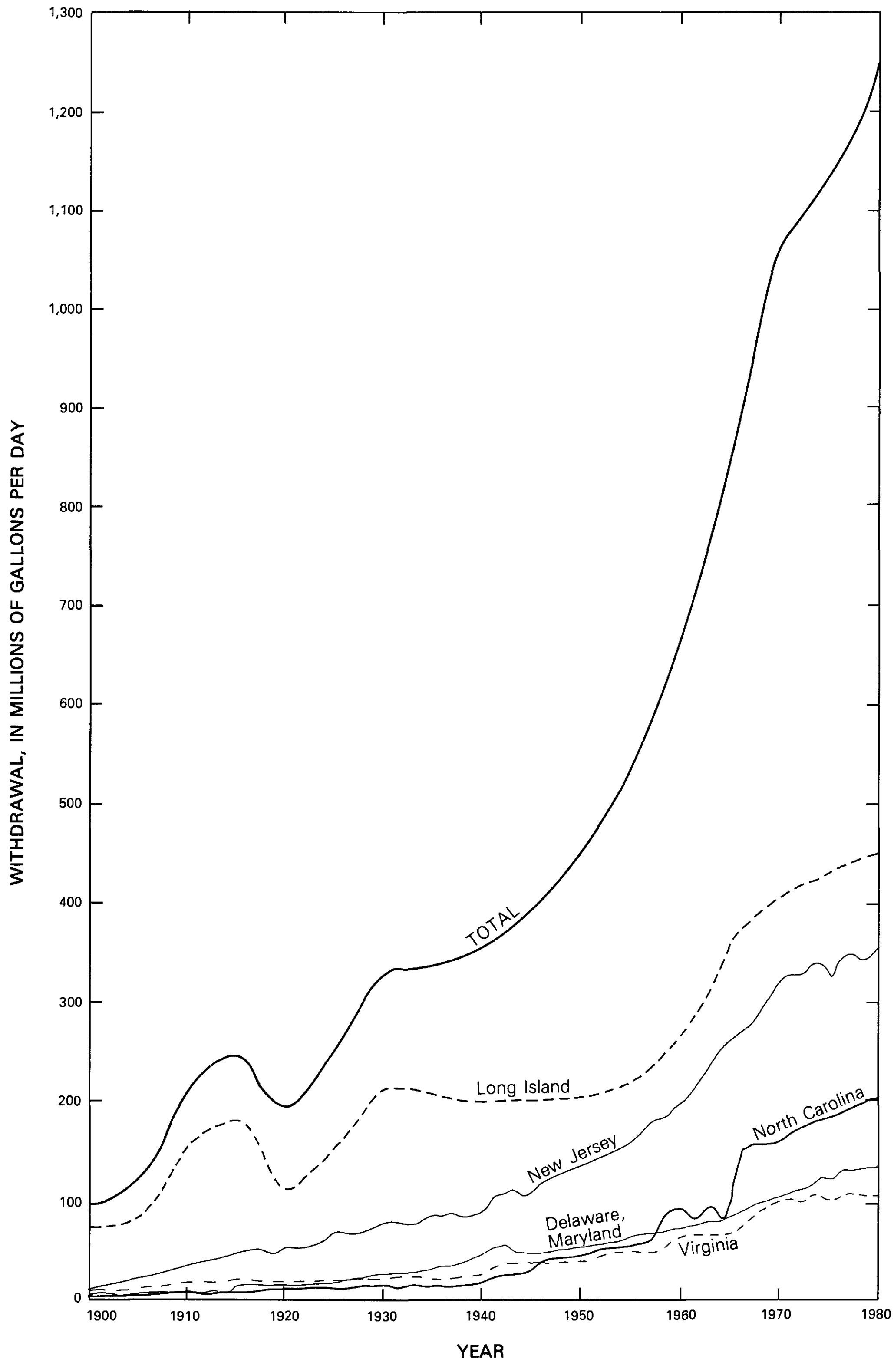


FIGURE 8.—Ground-water withdrawals in the northern Atlantic Coastal Plain and in Long Island, New Jersey, Maryland-Delaware, Virginia, and North Carolina, 1900–80. Figures for all subregions include major public-supply and industrial withdrawals; figures for New Jersey and Virginia also include major irrigation withdrawals.

TABLE 4.—*Simulated withdrawals from the northern Atlantic Coastal Plain by aquifer and subregion, 1978-80*
[In million gallons per day]

Subregion	Model layer[1]										Total[2]
	1	2	3	4	5	6	7	8	9	10	
Long Island	0.0	19.3	153.5	153.5	0.0	103.0	0.0	0.0	0.0	0.0	429.3
New Jersey	65.0	84.8	81.8	11.3	4.6	1.2	2.0	38.0	48.5	.2	337.4
Maryland-Delaware	33.2	37.2	12.7	.0	.7	5.9	6.5	6.1	9.1	25.1	136.5
Virginia	13.6	52.6	17.1	.0	.0	1.9	3.4	.0	5.4	.0	94.0
North Carolina	.0	3.2	12.7	34.8	3.1	.1	134.5	1.2	5.1	.3	195.0
Total	111.8	[3]197.4	[4]283.1	[5]211.0	[6]8.7	112.1	146.4	45.3	68.1	25.6	[7]1,209.5

[1]Regional aquifer names for the model layers are as follows: 10, Surficial; 9, Upper Chesapeake; 8, Lower Chesapeake; 7, Castle Hayne-Piney Point; 6, Beaufort-Aquia; 5, Peedee-Severn; 4, Black Creek-Matawan; 3, Upper Potomac and Magothy; 2, Middle Potomac; 1, Lower Potomac.
[2]Total withdrawals include major public supply and industrial withdrawals, and in New Jersey and Virginia, also include major irrigation withdrawals.
[3]An additional 0.3 Mgal/d was included in the model for South Carolina.
[4]An additional 5.3 Mgal/d was included in the model for South Carolina.
[5]An additional 11.4 Mgal/d was included in the model for South Carolina.
[6]An additional 0.3 Mgal/d was included in the model for South Carolina.
[7]Includes South Carolina withdrawals that were in the model.

1930's and 1940's. In Virginia prior to 1940, flowing wells tapping aquifers in Eocene and Cretaceous sediments were common in the major river valleys. Harsh and Laczniak (1986) estimated withdrawals during the period 1891 to 1940 to range from 4 to 10 Mgal/d.

Ground-water withdrawal data were also compiled for the small portion of the study area in South Carolina. This area included the Coastal Plain part of the following South Carolina Counties: Marlboro, Dillon, Marion, Horry, Georgetown, Williamsburg, Florence, Darlington, and Chesterfield. The South Carolina data included withdrawals in the Myrtle Beach area, estimated at 8.7 Mgal/d in 1980.

Table 4 shows the simulated withdrawals for 1978–80 by aquifer and subregion. The lower four aquifers (layers 1–4) in Cretaceous sediments, along with the Beaufort-Aquia (layer 6) and the Castle Hayne-Piney Point (layer 7) aquifers, sustained the largest withdrawals. Total withdrawals in 1978–80 were greatest for Long Island, at 429.3 Mgal/d, followed closely by New Jersey, with 337.4 Mgal/d, and North Carolina, with 195.0 Mgal/d. The smallest simulated withdrawals were for Maryland-Delaware and Virginia, at about 136.5 and 94.0 Mgal/d, respectively.

In the model, ground-water withdrawals from 1900 to 1980 were broken into 10 pumping periods, ranging from 3 to 21 years. The pumping periods were chosen to best represent major changes in withdrawal rates and to end at times at which head data had been collected over large areas or for which previous hydrologic analysis had been made. In general, the longer pumping periods were at the beginning of the simulation period and the shorter ones near the end, reflecting the reliability of the withdrawal data. The duration of each pumping period, together with the total withdrawals simulated, are shown in figure 9. Table 5 lists the simulated ground-water withdrawals by aquifer and pumping period.

MODEL CALIBRATION

Model calibration refined estimates of the hydrologic characteristics until the model behavior matched, within specified criteria, the measured and estimated behavior of the physical system. The procedure by which the model behavior was made to match the system behavior was similar to that described by Konikow (1978). The ability of the model results to match measured or estimated field data demonstrates that the model can adequately represent the hydrologic system. Although mathematically sophisticated parameter-estimation techniques have been used to calibrate ground-water flow models, in practice calibration is more frequently accomplished through a trial-and-error adjustment of model input data. Parameter-estimation techniques have not yet been tested satisfactorily for the solution of large three-dimensional problems, and therefore were not used in this study.

The four smaller subregional models were the basis for calibration of the regional model. Calibration of the regional model consisted of trial-and-error adjustment of hydraulic characteristics in the subregional models. The subregional models provided a more accurate simulation because their grid blocks generally were four times smaller than those of the regional model. After calibration of the subregional models, spatially averaged hydraulic characteristics were transferred to the regional model for further calibration. Hydraulic characteristics for areas not included in any subregional model were adjusted during regional calibration.

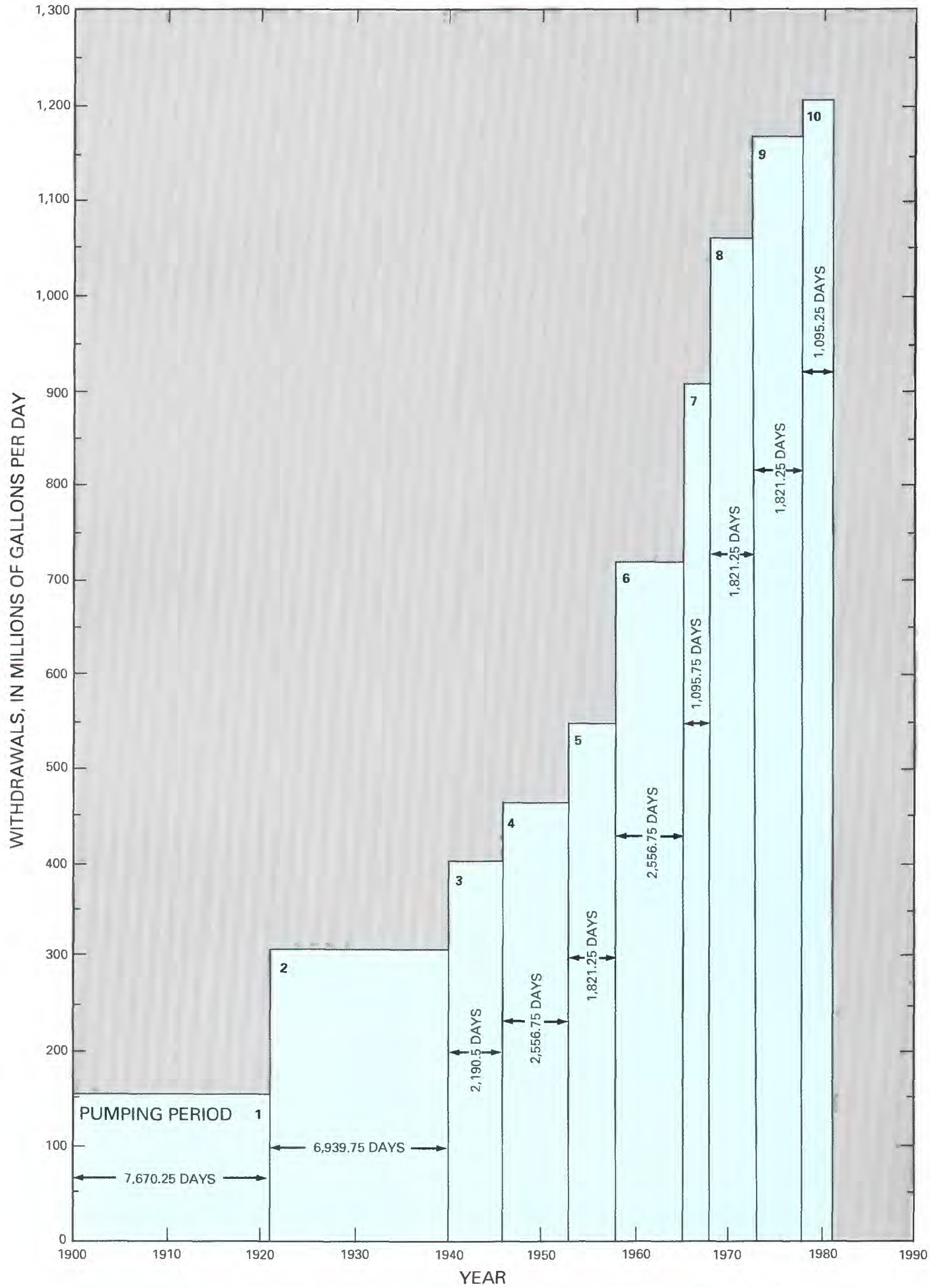


FIGURE 9.—Simulated ground-water withdrawals for each of ten pumping periods.

Calibration of prepumping conditions was achieved by matching, within specified criteria, simulated heads with measured and estimated prepumping heads. Both models, one with the specified-head conceptualization and the other with the modified-specified-flux upper boundary conceptualization, were calibrated. Steady-state calibra-

TABLE 5.—*Simulated ground-water withdrawals by aquifer and by pumping period*
[In million gallons per day]

Pumping period	Model layer[1] 1	2	3	4	5	6	7	8	9	10	Total[2]
1	15.0	12.0	16.5	7.0	0.8	92.8	2.4	2.1	3.7	0.3	152.5
2	37.7	38.3	37.1	26.9	1.6	137.0	5.6	8.9	11.1	2.7	306.7
3	57.0	58.2	66.4	38.4	1.7	132.4	18.8	13.6	13.6	4.3	404.3
4	63.7	76.7	84.7	40.7	2.0	123.1	37.6	15.9	14.5	5.8	464.8
5	67.7	86.4	119.8	70.3	3.4	111.3	42.1	19.9	19.3	8.1	548.3
6	83.3	116.8	169.6	104.3	4.8	110.7	66.1	24.6	29.0	8.9	718.0
7	86.4	149.3	206.9	135.2	6.3	129.8	112.5	32.6	37.1	11.3	907.4
8	104.6	174.3	244.2	164.4	6.8	134.1	128.0	37.8	53.2	13.5	1,061.1
9	111.5	191.0	269.9	198.8	7.5	125.3	141.5	44.1	63.3	18.9	1,171.7
10	111.8	197.4	283.1	211.0	8.7	112.1	146.4	45.3	68.1	25.6	1,209.5

[1]Regional aquifer names for the model layers are as follows: 10, Surficial; 9, Upper Chesapeake; 8, Lower Chesapeake; 7, Castle Hayne-Piney Point; 6, Beaufort-Aquia; 5, Peedee-Severn; 4, Black Creek-Matawan; 3, Upper Potomac and Magothy; 2, Middle Potomac; 1, Lower Potomac.

[2]Total withdrawals include major public supply and industrial withdrawals, but in New Jersey and Virginia, include major irrigation withdrawals. Total withdrawal does not equal sum by layer in every instance because of rounding errors.

tion provided initial estimates of transmissivity and confining-unit leakance, as well as initial heads, for the simulation of transient conditions.

Calibration of transient conditions in the aquifer system was achieved by matching simulated heads with measured or estimated heads for 1980. In addition, the model response through time was compared with 74 selected well hydrographs whose periods of record generally were longer than 20 years. Model behavior relative to measured well hydrographs was evaluated for the latter part of the transient simulation period because most measured data were for this period.

Calibration of transient conditions further refined estimates of aquifer transmissivity and confining-unit leakance. The simulation provided values of ground-water discharge to streams. These values were compared with estimates of ground-water base flow in the study area. The streambed-leakance values calculated from the previously discussed water-budget analysis were in good agreement with estimates of actual streambed leakance.

The following sections discuss regional transmissivity, confining-unit leakance, and prepumping and transient flow conditions. Definition of the flow system was accomplished by examining and interpreting (1) the water budget, including areal distribution of recharge and discharge, (2) the potentiometric surfaces, (3) the vertical leakage through confining units, and (4) the flow patterns and velocities in the aquifers. The prepumping steady-state condition as well as the changes caused by pumping are described. The calibration simulation for 1980 is compared with the prepumping simulation used to assess or define the changes relative to prepumping conditions and the effect of withdrawals on the flow system. In addition, local flow budgets for selected major withdrawal areas are presented, and the source of the water pumped at these locations is discussed.

TRANSMISSIVITY

Plates 3 through 6 show the regional transmissivity of each aquifer derived from model calibration. Transmissivity (pl. 3) of the surficial aquifer (layer 10) exceeds 1,000 ft^2/d in few places; an exception is the Delmarva Peninsula, where it exceeds 10,000 ft^2/d. The upper Chesapeake aquifer (layer 9) is one of the most transmissive aquifers in the Coastal Plain (fig. *A*, pl. 4). In New Jersey, where the layer represents the upper Kirkwood-Cohansey aquifer, transmissivity is as much as 10,000 ft^2/d. Another area of high transmissivity is near Ocean City, Md., on the Delmarva Peninsula, where transmissivity is as much as 22,000 ft^2/d. In North Carolina and Virginia, the aquifer is less transmissive, and transmissivity exceeds 4,000 ft^2/d in few places.

In the lower Chesapeake aquifer (layer 8), transmissivity exceeds 10,000 ft^2/d in two areas: in the east-central New Jersey Coastal Plain and near Cape Hatteras, N.C. (fig. *B*, pl. 4). Generally, the aquifer's transmissivity is less than 4,000 ft^2/d on the Delmarva Peninsula, about 6,000 ft^2/d in New Jersey, and between 1,000 and 6,000 ft^2/d in North Carolina. The Castle Hayne-Piney Point aquifer (layer 7) has a wide range of transmissivity (fig. *C*, pl. 4). In North Carolina, where the aquifer consists of highly permeable limestone, transmissivity is about 100,000 ft^2/d—the highest of any aquifer in the northern Atlantic Coastal Plain. Farther north, where the aquifer consists of unconsolidated sand, the transmissivity ranges from about 1,000 to 4,000 ft^2/d. On Long Island, where layer 7 represents the upper part of the upper glacial aquifer, transmissivity ranges from 2,000 to 8,000 ft^2/d. The transmissivity of the Beaufort-Aquia aquifer (layer 6) (fig. *A*, pl. 5) typically is less than 2,000 ft^2/d, but it reaches a maximum of more than 4,000 ft^2/d on the Eastern Shore of Maryland. On Long Island, where layer 6 represents the lower part of the upper glacial aquifer, transmissivity generally exceeds 10,000 ft^2/d and reaches a maximum of 35,000 ft^2/d.

The Peedee-Severn aquifer (layer 5) is minor over much of the study area. In North Carolina, its transmissivity locally exceeds 10,000 ft^2/d but generally is between 2,000 and 8,000 ft^2/d (fig. *B*, pl. 5). In New Jersey, Delaware, and Maryland, transmissivity of the aquifer is typically about 500 ft^2/d. On Long Island, layer 5 corresponds to the upper part of the Magothy aquifer, and its transmissivity is about 16,000 ft^2/d. In North Carolina, transmissivity of the Black Creek-Matawan aquifer (layer 4) is typically 2,000 to 4,000 ft^2/d (fig. *C*, pl. 5). In New Jersey, Delaware, and Maryland, transmissivity generally is less than 2,000 ft^2/d. On Long Island, layer 4 is the middle part of the Magothy aquifer, and its transmissivity exceeds 22,000 ft^2/d. The transmissivity of the upper Potomac and Magothy aquifers (layer 3) generally is less than 6,000 ft^2/d (fig. *A*, pl. 6). Notable exceptions are on Long Island and in New Jersey, where transmissivity is greater than 8,000 ft^2/d.

The middle Potomac aquifer (layer 2) has a highly variable transmissivity distribution (fig. *B*, pl. 6). Transmissivity exceeds 10,000 ft^2/d in much of the central part of the study area (Maryland and Virginia) and locally, in southern Delaware, east-central New Jersey, and Nassau County, Long Island, exceeds 10,000 ft^2/d. In the lower Potomac aquifer (layer 1), transmissivity (fig. *C*, pl. 6) has a maximum value of more than 10,000 ft^2/d in eastern New Jersey and the Northern Neck of Virginia. Typically, the aquifer has a transmissivity of 4,000 to 8,000 ft^2/d over much of its extent.

CONFINING-UNIT LEAKANCE

Plates 7 through 9 show the regional values of confining-unit leakance (vertical hydraulic conductivity divided by thickness of the confining unit) resulting from model calibration.

The leakance of the confining unit overlying the upper Chesapeake aquifer (layer 9) is among the highest in the study area (fig. *A*, pl. 7). Leakance ranges from 1.0×10^{-6} to 1.0 (ft/d)/ft. High values of leakance (1.0×10^{-2} to 1.0 (ft/d)/ft) occur over most of the Coastal Plain, and indicate areas where the confining unit overlying the aquifer is absent and vertical flow between aquifers is through sandy sediments. Leakance values decrease as the confining unit thickens to the east.

The lower Chesapeake aquifer (layer 8) is overlain by a confining unit that is more permeable than deeper confining units (fig. *B*, pl. 7). Leakance ranges from 1.0×10^{-7} to 1.0 (ft/d)/ft. Similar to the confining unit overlying the upper Chesapeake aquifer (layer 9), the higher leakance values indicate areas where clayey confining units are absent. The highest leakance is in North Carolina, whereas the lowest leakance is in New Jersey. A typical leakance for the Delmarva Peninsula is between 1.0×10^{-6} and 1.0×10^{-5} (ft/d)/ft.

Leakance of the confining unit that overlies the Castle Hayne-Piney Point aquifer (layer 7) is highly variable (fig. *C*, pl. 7). In areas of North Carolina where this confining unit is relatively thin, leakance is high, ranging from 1.0×10^{-5} to 1.0×10^{-1} (ft/d)/ft. In Virginia, Maryland, Delaware, and New Jersey, where the confining unit is underlain by the Piney Point aquifer, leakances of 1.0×10^{-7} to 1.0×10^{-5} (ft/d)/ft are typical. Values decrease to the north.

Leakance of the confining unit that overlies the Beaufort-Aquia aquifer (layer 6) also is highly variable (fig. *A*, pl. 8). Leakance values range from 1.0×10^{-8} to 1.0 (ft/d)/ft. The high leakance values on Long Island represent a sand contact between highly permeable upper and lower parts of the upper glacial aquifer. High leakance values also are found in North Carolina, whereas low leakance values are found in New Jersey. A regional trend is apparent, with leakance decreasing downdip in all States and decreasing northward from North Carolina to New Jersey.

The Peedee-Severn aquifer (layer 5) is overlain by a confining unit that has leakance values ranging from 1.0×10^{-6} to 1.0×10^{-2} (ft/d)/ft (fig. *B*, pl. 8). The highest leakance is in North Carolina and on Long Island, where the confining unit is thin or absent. The lowest leakance is in eastern New Jersey and on the Delmarva Peninsula.

Leakance of the confining unit that overlies the Black Creek-Matawan aquifer (layer 4) is highest on Long Island, and in general is higher along the Fall Line throughout the remainder of the study area (fig. *C*, pl. 8). Leakance is considerably lower (1.0×10^{-6} (ft/d)/ft) in east-central New Jersey. This is the lowest value for the study area. Similar low values are found in North Carolina and southern Maryland.

The confining unit that overlies the upper Potomac and Magothy aquifers (layer 3) has a highly variable leakance ranging from 1.0×10^{-8} to about 1.0×10^{-1} (ft/d)/ft (fig. *A*, pl. 9). The highest values are in areas where the confining unit is absent or pinches out. Therefore, the highest values used in the model represent the vertical leakance of the permeable sands of the lower and middle parts of the Magothy aquifer on Long Island and are several orders of magnitude higher than typical values for this confining unit. These high values were related to the simulation scheme of the model as opposed to the mappable regional confining unit. Areas of high leakance generally occur along the Fall Line. Values average about 1.0×10^{-6} (ft/d)/ft over most of the Coastal Plain. An area of east-central New Jersey has the lowest leakance for this unit, 1.0×10^{-8} (ft/d)/ft.

In the confining unit overlying the middle Potomac aquifer (layer 2), leakance ranges from about 1.0×10^{-7} to 1.0×10^{-3} (ft/d)/ft (fig. *B*, pl. 9). The higher values occur in updip areas where the confining unit is generally thin. For example, leakance exceeds 1.0×10^{-3} (ft/d)/ft in areas along the Fall Line in Virginia. Similarly, on Long Island, leakance is relatively high, about 1.0×10^{-4} (ft/d)/ft. Downdip, along the simulated saltwater-freshwater interface, leakance is about 1.0×10^{-7} (ft/d)/ft. The leakance of the confining unit shows a regional trend—it decreases downdip and in a southerly direction across the study area from an average of about 1.0×10^{-4} (ft/d)/ft on Long Island to about 1.0×10^{-7} (ft/d)/ft in North Carolina.

The confining unit overlying the lower Potomac aquifer (layer 1) has a maximum leakance of about 1.0×10^{-4} (ft/d)/ft (fig. *C*, pl. 9), along the Delaware River in southern New Jersey. Leakance is as much as four orders of magnitude less, 1.0×10^{-8} (ft/d)/ft, in southern Maryland. Values ranges from 1.0×10^{-7} to 1.0×10^{-6} (ft/d)/ft in Virginia.

CHARACTERISTICS OF THE REGIONAL FLOW SYSTEM BASED ON SIMULATION

PREPUMPING STEADY-STATE CONDITIONS

Potentiometric surfaces

The simulated prepumping potentiometric surfaces for modeled aquifer layers 1 through 9 are shown on plates 10 through 12. Calibration was based primarily on a few measured prepumping heads and on the interpreted potentiometric surfaces that also are shown on these plates.

The measured and interpreted head data represent three levels of reliability and are presented as both discrete control points and interpreted potentiometric surfaces. In some areas, principally Long Island and New Jersey, sufficient head data existed to produce reliable prepumping potentiometric surfaces. In areas where only data points are shown with the simulated potentiometric surfaces (pls. 10–12), the data were too sparse to produce representative potentiometric surfaces and it was necessary to compare the model calibrated heads with only the measured data points. Measured prepumping data points were used primarily in North Carolina and Virginia, and for some aquifer layers in Maryland and Delaware. The least reliable data were simulated, generalized, or conceptualized prepumping surfaces obtained from the literature. These data covered only small parts of the regional aquifers and were available for the Magothy aquifer (layer 3) in Maryland, the middle Potomac (layer 2) and lower Chesapeake (Cheswold of local usage) (layer 8) aquifers in Delaware, and a composite Potomac aquifer (layers 1, 2, and 3) in Virginia.

Based on the reliability of these data, prepumping conditions were best known for Long Island and New Jersey and least known for Virginia and North Carolina. In general, the heads in the updip, shallow areas of the aquifers were better defined than the heads in the downdip, deeper part of the system. Model calibration was as rigorous as the data allowed.

Measured head data were available for the upper Chesapeake aquifer (layer 9) for North Carolina and the Eastern Shore of Virginia (fig. *A*, pl. 10). In North Carolina, measured heads ranged from more than 100 ft above sea level near the Fall Line to less than 1 ft above sea level near the shore. The simulated potentiometric surface generally agreed with the measured or interpreted data. Flow was from potentiometric highs near the Fall Line southeastward toward the ocean. The simulated potentiometric surface was characterized by local flow to overlying streams and estuaries. In Virginia, measured heads corresponded to a simulated ground-water high situated along the axis of the Eastern Shore of Virginia. In New Jersey, the aquifer is generally unconfined, and simulated heads reflected the interpreted water-table configuration.

Measured head data and interpreted potentiometric surfaces for the lower Chesapeake aquifer (layer 8) were available only for part of North Carolina, Delaware, and New Jersey (fig. *B*, pl. 10). In North Carolina, measured heads tended to be low, ranging from 3 to 17 ft above sea level. The simulated surface reasonably matched these data. The simulated potentiometric surface in Delaware showed the same regional flow pattern as the prepumping potentiometric surface for the Cheswold aquifer (local usage) described by Leahy (1982a, p. 23). Flow was from a potentiometric high in the middle of the Delmarva Peninisula toward areas of low head along Delaware Bay. In New Jersey, model layer 8 was equivalent to the lower part of the unconfined Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer. Simulated heads in the lower part of the aquifer were quite similar to the simulated and interpreted water table in the overlying aquifer (fig. *A*, pl. 10). However, the influence of Great Bay and the Mullica River (pl. 1) was more pronounced on the simulated surface and less apparent on the interpreted surface in both confined and unconfined areas.

The prepumping potentiometric surface of the Castle Hayne-Piney Point aquifer (layer 7) is shown in figure *C*, plate 10. In North Carolina, measured prepumping heads were less than 50 ft above sea level, and most measured head data were located between the simulated 10- and 25-ft potentiometric contours. Many of the lowest measured heads, including the heads of 10 and 17 ft near the Pamlico River and 13 ft near the Neuse River, were located along major streams or their tributaries and represent local features that could not be simulated because of the large grid blocks. In Virginia, most head data were for the Northern Neck Peninsula and were interpreted by Newton and Siudyla (1979, pl. 9). Measured heads typically ranged from 10 to 30 ft above sea level. The simulated potentiometric surface showed the steep gradients along the aquifer outcrop and the major influence of the Rappahannock, York, and James Rivers on the flow system. The measured data showed these same gradients, particularly as evidenced by the low heads along the Rappahannock River.

Measured prepumping data for the subregional Piney Point-Nanjemoy aquifer (layer 7) in Maryland were limited to the southern part of the State and the Delmarva Peninsula. The data suggest that flow in southern Maryland was from the updip limit toward the Potomac River and Chesapeake Bay. The simulated prepumping surface closely matched the measured data. On the Delmarva Peninsula, measured heads ranged from 21 to 35 ft above sea level. Williams (1979, fig. 7) interpreted sparse prepumping head data. His interpretation suggested that heads downdip in the aquifer were less than 5 ft above sea level. The simulated prepumping heads did not match this interpretation, but rather showed a ground-water divide along the center of the peninsula and suggested flow toward Chesapeake and Delaware Bays. Based on theoretical development by Hubbert (1940, p. 886–870), the saltwater-freshwater interface would not have been in equilibrium with current sea level conditions and the interface would be moving landward, if prepumping heads were below 10 to 15 ft in the downdip areas of layer 7, where the top of the aquifer was greater than 500 ft below sea level. In contrast, if the downdip prepumping heads were higher than 10 to 15 ft, the prepumping position of the saltwater-freshwater interface would have been in equilibrium. The regionally simulated heads were typically 20 to 25 ft above sea level in areas where the top of the aquifer was 500 to 1,000 ft below sea level and generally matched measured heads. The simulated potentiometric surface suggests that the position of the saltwater-freshwater interface was in equilibrium with prepumping hydraulic conditions.

In New Jersey, the simulated and interpreted potentiometric surfaces for the Piney Point aquifer (layer 7) matched closely. Updip potentiometric highs caused flow seaward and toward Delaware Bay. This regional flow pattern was evident in both surfaces. On Long Island, layer 7 consisted of the upper part of the upper glacial aquifer and is discussed with layer 6.

The prepumping potentiometic surfaces for the regional Beaufort-Aquia aquifer (layer 6) are shown in figure *A*, plate 11. Layer 6 included the lower part of the upper glacial aquifer on Long Island, which has no lateral connection with the Vincentown aquifer in New Jersey. The upper glacial aquifer is the water-table aquifer. Because water-table altitudes were specified in the regional model for some simulations, a comparison of simulated and interpreted potentiometric surfaces for layer 6 and layer 7 on Long Island has no significant meaning. In New Jersey, the aquifer is minor from a regional perspective, and the simulated and interpreted surfaces also were not compared. The Beaufort-Aquia aquifer is generally most important in North Carolina, Virginia, Maryland, and Delaware. Measured prepumping head data for North Carolina and Virginia were sparse. Measured heads in these States were less than 50 ft above sea level and matched the simulated surface well. The 50-ft simulated contours were located near the outcrop of the aquifer, and the hydraulic gradient was gently seaward in both the simulated and measured heads in North Carolina and Virginia.

In Maryland, the measured data for the Beaufort-Aquia aquifer (layer 6) showed a steep prepumping hydraulic gradient from the outcrop area to discharge points along Chesapeake Bay and the Potomac River (fig. *A*, pl. 11). The simulated prepumping surface showed a similar pattern of flow. Chapelle and Drummond (1983) contoured measured prepumping head data and found a similar flow pattern. However, the influence of the Patuxent River was more evident in their interpretation. As with other aquifer layers, local details of this flow system were not simulated in the regional model. Prepumping water-level data for Delaware were lacking.

The simulated prepumping potentiometric surface for the Peedee-Severn aquifer (layer 5) is shown in figure *B*, plate 11. The Peedee-Severn aquifer is mainly present in North Carolina and New Jersey. In North Carolina, measured heads ranged from 14 to 106 ft above sea level. These data generally corresponded to simulated contours. Some measured heads were lower than simulated heads because the wells were located in areas where the model did not simulate local stream elevations. Visual inspection shows the similarity between interpreted and simulated potentiometric surfaces for the subregional Wenonah-Mount Laurel aquifer in New Jersey. Simulation results for the upper part of the Magothy aquifer on Long Island are discussed with results for the upper Potomac and Magothy aquifers (layer 3).

The Black Creek-Matawan aquifer (layer 4) is present mainly in North Carolina, New Jersey, and Long Island. In North Carolina, the simulated potentiometric surface (fig. *C*, pl. 11) agreed with measured heads. Values of head ranged from 40 to 70 ft and generally were bounded by the simulated 50- and 75-ft contours. The simulated flow pattern is consistent with the measured data. Flow was eastward from areas of high head in the southeastern part of the State toward the Cape Fear area and toward Albemarle Sound.

The simulated potentiometric surface for the subregional Englishtown aquifer (layer 4) in New Jersey showed the same regional trend as the interpreted surface, but the model lacked the areal resolution to duplicate the details of the local flow system. The discharge area adjacent to the Delaware River was evident in both the simulated and interpreted surfaces. Also, the 25-, 50-, and 75-ft simulated contours were within 10 to 15 ft of measured heads. Flow from a potentiometric high in the central part of the New Jersey Coastal Plain was toward Raritan Bay to the northeast and the Delaware River to the southwest. Results for the middle part of the Magothy aquifer on Long Island are discussed with results for the upper Potomac and Magothy aquifers (layer 3).

The simulated contours for the upper Potomac and Magothy aquifers (layer 3) in North Carolina and Virginia reasonably matched the measured heads (fig. *A*, pl. 12). The subregional Brightseat-upper Potomac aquifer in the southern part of the study area and the Magothy aquifer to the north were not continuous, and the gap between them was represented in the model by a band of zero-transmissivity nodes in layer 3 through southern Maryland.

In southern Maryland, the interpreted potentiometric surface of the Magothy aquifer showed the same range of heads as the simulated surface. However, the interpreted direction of flow was from the potentiometric high toward the Severn and Magothy rivers. In contrast, heads in the simulated system were not greatly affected by these rivers, and the simulated flows were toward the submerged subcrop area beneath Chesapeake Bay. Although the model simulated regional trends, local features of the flow system were not simulated because of the large size of the grid blocks.

As in southern Maryland, the simulated potentiometric surface in New Jersey showed the same regional trends as the interpreted surface, but with less detail. The difference was most apparent in the updip area, but it also was present in the downdip areas. For example, the simulated 25-ft contour generally followed the interpreted 20-ft contour, but with less resolution. The difference between these two contours was caused by the previously mentioned discretization as well as by the uncertainty associated with developing the interpreted potentiometric surface from sparse measured data. In this case the discretization problem was not related to the size of geographic features, but rather to averaged hydraulic characteristics used to represent properties that differ areally within a grid block. In either instance, local details were lost in the regional simulation.

On Long Island, model layer 3 represented approximately the lower third of the Magothy aquifer, and model layers 4 and 5 represented the middle and upper thirds, respectively. Getzen (1977, p. 35) presented the 1971 potentiometric surface for the lower part of the Magothy aquifer (layer 3) in the eastern half of Long Island. This area had not been significantly affected by pumping; therefore, the 1971 potentiometric surface was assumed to represent prepumping heads. The simulated potentiometric surface showed the same regional trends as the interpreted surface. Potentiometric highs were in the center of the island. The hydraulic gradient on the northern shore was steep toward Long Island Sound and much gentler seaward toward the Atlantic Ocean on the south. Getzen (1977, fig. 33) demonstrated that a vertical hydraulic gradient of less than 2 ft prevailed within the Magothy aquifer (layers 5, 4, and 3) under prepumping conditions. The simulated regional heads also showed little (less than 1 ft) vertical hydraulic gradient between these layers (figs. *B*, *C*, pl. 11, and fig. *A*, pl. 12).

Prepumping conditions for the middle Potomac aquifer (layer 2) were interpreted from measured heads in North Carolina and Virginia, and from potentiometric surfaces constructed for the Potomac aquifers in Delaware and the Lloyd aquifer on Long Island (fig. *B*, pl. 12). In New Jersey, the prepumping potentiometric surface was assumed to be very similar to the prepumping potentiometric surface available for the overlying subregional upper Potomac-Raritan-Magothy aquifer (layer 3). The simulated surface in North Carolina matched the measured heads and indicated regional flow from South Carolina toward North Carolina. In Virginia, measured heads decreased seaward. Most measured values of less than 50 ft were located seaward of the simulated 50-ft contour. Measured values ranging from 50 to 75 ft fell primarily between the simulated 50- and 75-ft contours. In northern Delaware and New Jersey, the simulated potentiometric surface showed the same regional configuration as the interpreted potentiometric surface and measured data. In general, flow was from potentiometric highs along the Fall Line in northern Delaware and in the central New Jersey Coastal Plain toward potentiometric lows along Raritan Bay, the Delaware River, and Chesapeake Bay. On Long Island, the simulated and interpreted surfaces were similar; flow was from potentiometric highs trending east-west across the island toward Long Island Sound and the Atlantic Ocean. The simulated hydraulic gradient corresponded to the interpreted gradient, in that the steepest gradient was toward Long Island Sound and a much gentler gradient was to the south toward the Atlantic Ocean.

Prepumping conditions for the lower Potomac aquifer (layer 1) were the most poorly defined. As shown in figure *C*, plate 12, the aquifer is present principally in

Virginia, Maryland, and Delaware. The interpreted potentiometric surface in Virginia is a composite map of heads in the subregional Brightseat-upper Potomac and middle and lower Potomac aquifers (layers 3, 2, and 1). Simulated prepumping heads generally agreed with these data. Of particular note is the location of the 50-ft contour in Virginia and the area of flow from the potentiometric high in southern New Jersey and along the Fall Line in northern Delaware toward the Delaware River.

Regional ground-water budget

The components of the prepumping ground-water budget were recharge to the aquifer system, flow into the modeled area from the south, and ground-water discharge to the ocean, bays, and streams. The simulated regional ground-water flow budget for prepumping conditions is shown in figure 10. Simulated ground-water recharge to the emerged Coastal Plain was varied areally from 15 to 22.5 in/yr in the model, and total recharge was estimated to be about 40,000 Mgal/d. This was an average annual recharge of 15.4 in/yr over 54,900 mi^2 (58 percent of the modeled area).

Simulation results suggest that under prepumping conditions the majority of this recharge, an average of 39,408 Mgal/d, discharged to the nearest surface-water body through shallow unconfined aquifers. A small amount of the water (592 Mgal/d) recharged the deeper confined aquifers from the shallow unconfined aquifers in an area of 25,000 mi^2, about 26 percent of the total modeled area. Therefore, the average areal recharge from shallow unconfined aquifers to deeper confined aquifers was 0.5 in/yr, about 2 percent of the total areal recharge.

An additional small volume of water (6.0 Mgal/d) flowed into the confined aquifer system from South Carolina in the middle Potomac aquifer (layer 2). Although this flow was small relative to the deep percolation, it caused heads to be high (as much as 100 ft above sea level) along the North Carolina-South Carolina border in the deeper confined aquifers.

Approximately 598 Mgal/d of water was discharged from the deeper confined aquifers. This discharge occurred over approximately 74 percent of the total modeled area. The discharge from the deep confined aquifers was combined with the local discharge from shallow unconfined aquifers to account for a total ground-water discharge of about 40,000 Mgal/d, which was equal to the total areal recharge under steady-state (prepumping) conditions.

Deep percolation and ground-water discharge to surface-water bodies

The 10-layer simulation with an overlying specified-head boundary at water-table altitudes provided the areal distribution of rates of flow across the base of the unconfined system from or to the underlying confined aquifers (in this report referred to as "deep percolation"). Figure *A*, plate 13, shows the areal distribution of simulated prepumping deep percolation in inches per year.

Typically, prepumping deep percolation was less than 0.5 in/yr over the study area. The maximum rate of recharge to the deeper aquifers was about 20 in/yr along the center of Long Island. The highest simulated discharge, 16 in/yr, occurred along the northern shore of Long Island. Little confinement exists between the two model layers (7 and 6) representing the unconfined upper glacial aquifer. Therefore, vertical flow between these model layers was relatively high. These high values of recharge to the confined aquifers agreed with the estimate by Franke and Cohen (1972, p. C272) of about 21 in/yr of ground-water recharge.

The influence of surface-water bodies (streams, rivers, bays, estuaries, and the ocean) on the areal distribution of prepumping deep percolation was evident. The Atlantic Ocean, major estuaries, and major rivers were discharge areas. For example, Long Island Sound, the New York Bight, Raritan, Delaware, and Chesapeake Bays, and the Potomac, Delaware, Rappahannock, York, and James Rivers were all areas of discharge for the confined aquifer system. In North Carolina, Albemarle and Pamlico Sounds and the Roanoke, Chowan, Tar, Neuse, and Cape Fear Rivers were all regional drains. Rates of discharge to the ocean typically were less than 0.1 in/yr. In general, areas where discharge was less than 0.5 in/yr were large and areas where discharge was greater than 0.5 in/yr were much smaller.

In North Carolina, typical simulated recharge was 0.5 in/yr or less in interstream areas. This agreed with estimates of 0.5 to 1 in/yr of recharge to deeper aquifers (Heath, 1975, fig. 42; Wilder and others, 1978, fig. 11). The areas of highest discharge tended to be adjacent to areas of high recharge. Areas deeply incised by streams and underlain by thin or comparatively permeable confining units exhibited significant local variation in the amount and distribution of recharge and discharge. A band of high recharge, typically greater than 1 in/yr and as much as 5 in/yr, was associated with high-discharge areas. This area is underlain by the subcrop of the Castle Hayne aquifer (layer 7), a highly permeable unit. The headwater areas of many Coastal Plain streams, such as near the Fall Line along the Potomac River in Virginia and along the Delaware River in southern New Jersey, also were areas of high recharge and discharge.

On the Delmarva Peninsula, deep recharge of as much as 1 in/yr occurred along the surface drainage divides, and discharge from the confined system was to Delaware

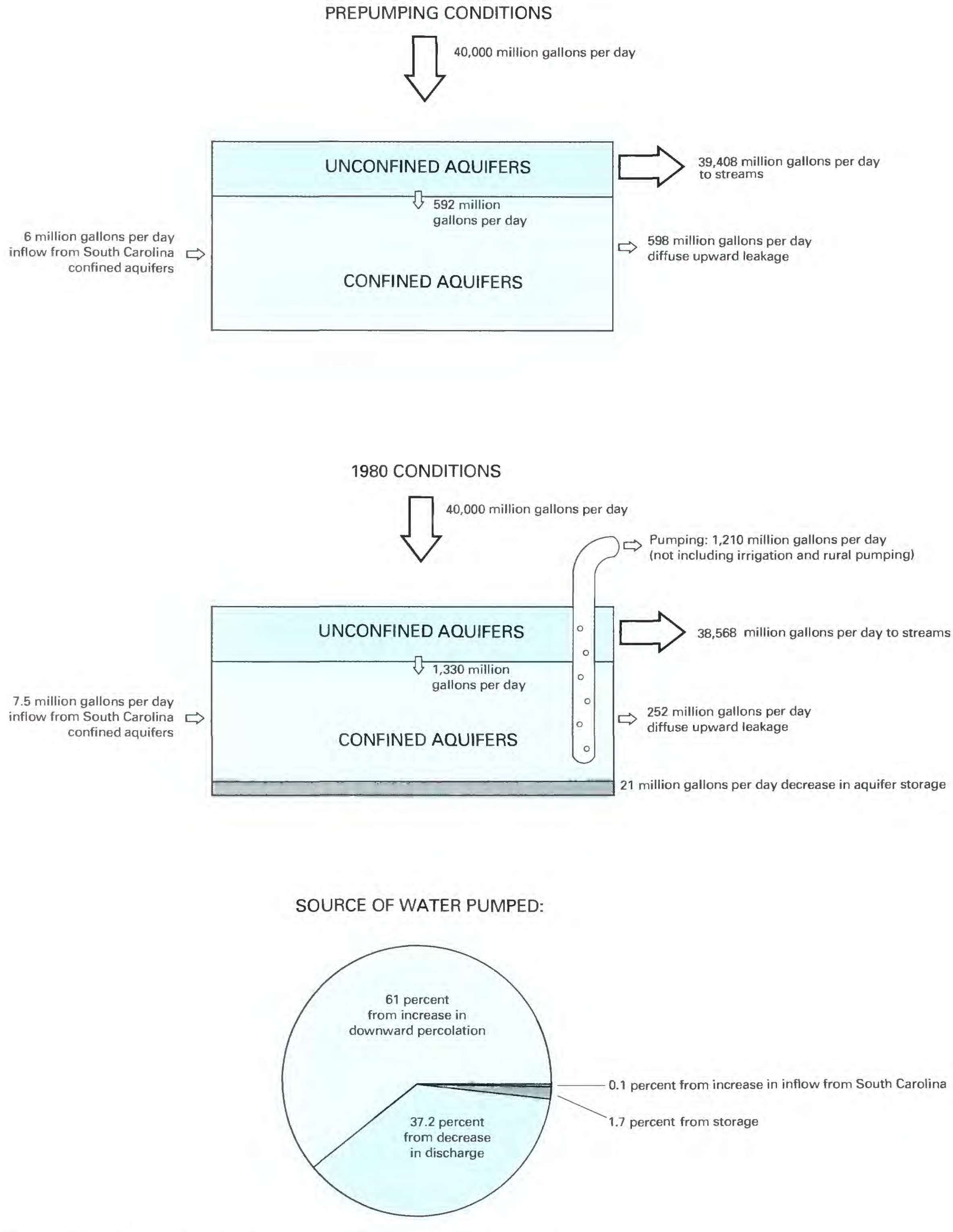


FIGURE 10.—Simulated regional ground-water flow budget for the northern Atlantic Coastal Plain, prepumping and 1980 flow conditions.

and Chesapeake Bays and the Atlantic Ocean. In New Jersey, the Delaware River was a major discharge (greater than 4 in/yr) area. Other major surface-water bodies in New Jersey also acted as discharge areas. Topographically high areas in the central part of the New Jersey Coastal Plain and the area along the Fall Line from Trenton north toward Raritan Bay were recharge areas.

In the second model conceptualization, in which the water table was simulated as a recharge boundary with specified fluxes, streams were simulated as specified heads for every onshore model block. Figure *B*, plate 13, shows the simulated discharge of ground water from the water table to streams. This value approximated the long-term average base flow under prepumping conditions. Simulated ground-water discharge to streams (base flow) averaged 10 to 15 in/yr over most of the study area and was consistent with estimates in areal water budgets of 8.5 to 11 in/yr (Cushing and others, 1973, p. 35; Sinnott and Cushing, 1978, p. I15; Wilder and others, 1978, fig. 11). Over most of the study area the discharge rate averaged 10 to 15 in/yr, whereas in the northern part of the study area (New Jersey and Long Island) the rate tended to be about 15 to 35 in/yr. This regional trend reflects the areal variability of ground-water recharge specified in the model. Areas of high ground-water discharge to streams were along the northern and southern shores of Long Island, along the Delaware River and major coastal streams (Great Egg Harbor, Maurice, Mullica, and Toms Rivers) in New Jersey, along Delaware Bay, and along the Indian and Nanicoke Rivers on the Delmarva Peninsula. In Maryland and Virginia, ground-water discharge was highest along the major streams and rivers, particularly along upstream reaches near the Fall Line. A similar trend was apparent in North Carolina. However, areas of highest discharge to major streams were located more seaward than those in Virginia and Maryland. This is because in North Carolina, streams are not deeply incised near the Fall Line, and major discharge areas (the bays and estuaries) are not as close to the Fall Line as they are in Virginia and Maryland.

Horizontal flow and velocities

The potentiometric surfaces and calculated Darcy velocities were used to describe the regional prepumping ground-water flow system for three aquifers: a shallow confined aquifer, an intermediate-depth aquifer, and a deep aquifer (the upper Chesapeake aquifer (layer 9), the Castle Hayne-Piney Point aquifer (layer 7), and the upper Potomac and Magothy aquifers (layer 3), respectively). Prepumping ground-water velocities ranged from 0.01 to about 100 ft/yr. Generally, most prepumping ground-water flow was local, and ground-water velocities were highest in the updip parts of the aquifers. Ground-water flow patterns were more regional, and flow velocities were less than 1 ft/yr in the deep, confined parts of the aquifers.

Simulated heads for the upper Chesapeake aquifer (layer 9), a shallow confined aquifer, ranged from more than 150 ft above sea level along the Fall Line in North Carolina to less than 10 ft above sea level in the seaward parts of the aquifer (fig. *A*, pl. 10). Onshore ground-water flow was characterized by local flow from ground-water highs to nearby discharge areas and relatively high flow velocities (greater than 1 ft/yr). Lateral ground-water velocities in the upper Chesapeake aquifer (layer 9) (fig. *A*, pl. 14) ranged from less than 0.01 ft/yr to more than 100 ft/yr. Separate local flow systems existed within each State. However, in the deeper areas of the aquifer along the coastal areas, velocities generally were lower (less than 1 ft/yr) and flow patterns were more regional. In some offshore areas, relatively high flow velocities occurred in small submerged outcrop areas.

Simulated heads for the Castle Hayne-Piney Point aquifer (layer 7) ranged from about 125 ft above sea level in central North Carolina to less than 10 ft above sea level offshore of Virginia and North Carolina (fig. *C*, pl. 10). On Long Island, this model layer represents the upper part of the unconfined glacial aquifer; therefore, the simulated heads represent the water-table surface. Ground-water flow was from potentiometric highs in the outcrop or subcrop areas down the hydraulic gradient toward the seaward boundary of the aquifer. Horizontal ground-water velocities in the Castle Hayne-Piney Point aquifer (layer 7) (fig. *B*, pl. 14) ranged from less than 0.1 to more than 100 ft/yr (feet per year). Flow rates were greatest along the outcrop of the aquifer. Relatively large lateral flow rates (greater than 10 ft/yr) were computed for the aquifer outcrop area in North Carolina where the Castle Hayne aquifer is a highly permeable limestone. Although the aquifer does not outcrop in New Jersey, simulated velocities were also relatively high. However, over the remainder of its extent, where the aquifer is confined, flow rates were typically 1 ft/yr or less.

Upward vertical leakage dissipates the head along the regional flow paths. Major discharge points in North Carolina, Virginia, and Maryland were evident from the shape of the potentiometric contours (fig. *C*, pl. 10) and the converging flow directions (fig. *B*, pl. 14) at the major surface-water bodies. Converging flow lines were not evident beneath Delaware Bay. This is probably because the Piney Point is entirely confined in Maryland, Delaware, and New Jersey.

Simulated prepumping heads for the upper Potomac and Magothy aquifers (layer 3) ranged from 300 ft above sea level in the Sand Hills area of North Carolina to less than 10 ft above sea level in the New York Bight area (fig. *A*, pl. 12). Lateral ground-water flow velocities in layer 3 (fig. *C*, pl. 14) ranged from more than 100 ft/yr in the updip part of the aquifer to less than 0.01 ft/yr in downdip areas near the seaward boundary of the aquifer. As in the shallower Castle Hayne-Piney Point aquifer (layer 7), ground water typically flowed from potentiometric highs along the Fall Line seaward toward potentiometric lows along the seaward boundary of the aquifer. Flow patterns tended to be regional and to cross State boundaries. Upward vertical leakage to overlying aquifers occurred over most of the downdip parts of the confined-flow system. As the water moved downdip and discharged to overlying aquifers, heads decreased. In turn, the lateral hydraulic gradient and accompanying flow velocity decreased. The effects of major rivers as regional discharge areas were evident by the direction of flow in the updip parts of layer 3. Potentiometric contours also showed this, particularly the 100-ft contour in North Carolina and the 50-ft contour in Virginia (fig. *A*, pl. 12). Chesapeake Bay was a major regional discharge area, as shown by converging flow lines beneath the Maryland part of the Delmarva Peninsula. In some areas, as along the Delaware River in New Jersey, major regional discharge areas caused the flow to be directed back updip toward the aquifer outcrop. Simulation results suggest that increased hydraulic gradient caused lateral flow velocities to increase as water moved updip to where the regional drains incise the aquifer. Localized lateral flow from the outcrop moves quickly to nearby discharge areas, as seen along most of the updip aquifer limit (fig. *C*, pl. 14).

Vertical flow and leakage

Three hydrogeologic sections, shown in figures 11 through 13, illustrate the vertical flow relations in the prepumping system with hydraulic heads and generalized flow vectors. The locations of the sections are shown in figure 1. The northernmost hydrogeologic section, *A–A′* (fig. 11), extends from the Delaware River in the west, across south-central New Jersey, to the Atlantic Ocean in the east. Flow was downward beneath the potentiometric high located about 25 mi east of the Delaware River. Flow was then either updip toward the Delaware River or downdip toward the Atlantic Ocean. A major ground-water divide located in the central part of the section and apparently extending to great depth affected ground-water flow directions in all the aquifers.

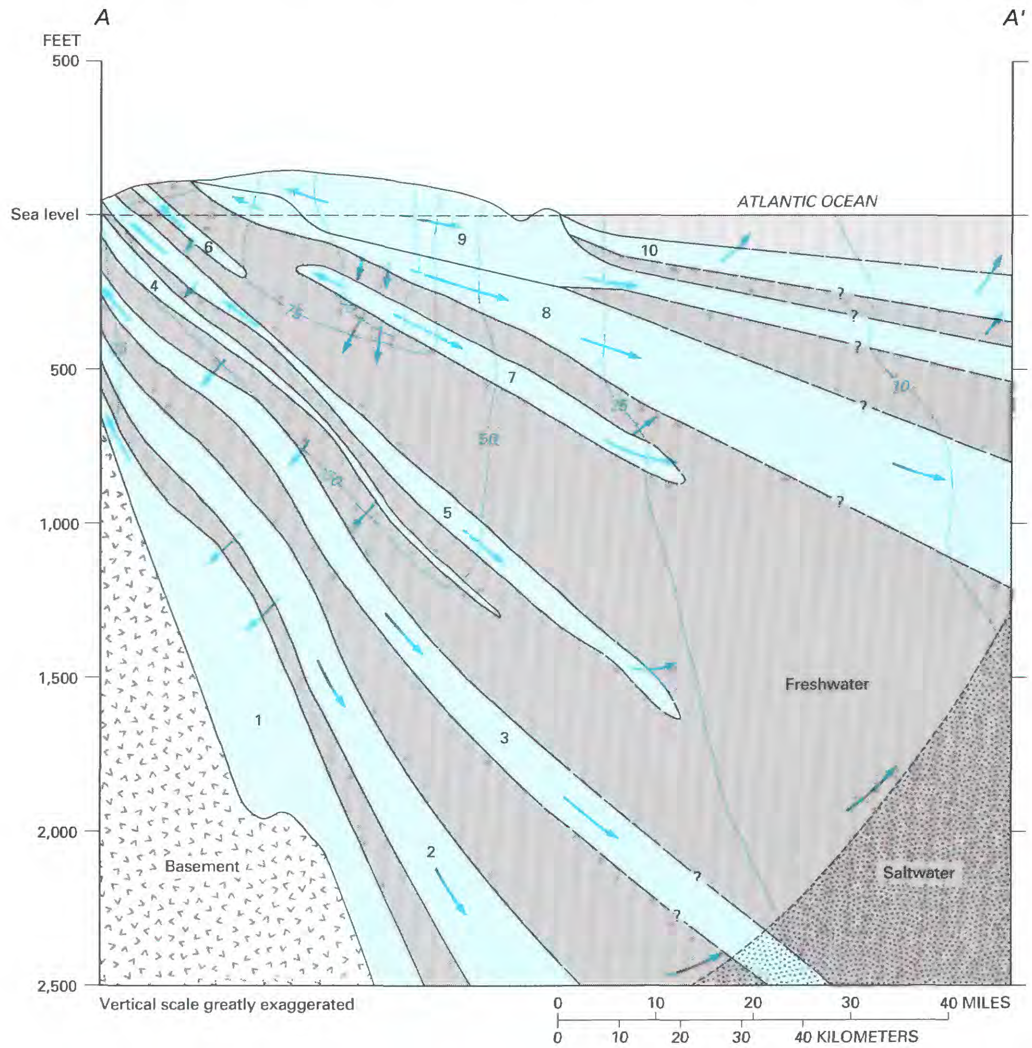
Regional flow paths extended from the potentiometric high in the upper Chesapeake aquifer (layer 9), downward to the Potomac aquifers (layers 3, 2, and 1), and then upward to the Atlantic Ocean or upward to the Delaware River. More intermediate flow paths extended from the potentiometric high to the outcrop areas of the Beaufort-Aquia, Peedee-Severn, and Black Creek-Matawan aquifers (layers 6, 5, and 4) or through the lower Chesapeake and Castle Hayne-Piney Point aquifers (layers 8 and 7) to the Atlantic Ocean.

Hydrogeologic section *B–B′* (fig. 12) extends eastward from the Fall Line in Maryland and crosses Chesapeake Bay and the Delmarva Peninsula (fig. 1). Chesapeake Bay was a regional discharge area, as shown by the generalized flow directions and hydraulic heads. Flow was upward beneath the bay from all of the aquifers. There was flow toward the bay from the west and updip from a potentiometric high on the Delmarva Peninsula. Flow from this high was also eastward toward the ocean. In general, flow was downward near the Fall Line along the western side of the hydrogeologic section, and upward from deeper aquifers throughout the remainder of the section.

Regional flow paths in the Potomac aquifers (layers 3, 2, and 1) extended from the Fall Line downdip to the Atlantic Ocean beneath the Delmarva Peninsula. More intermediate flow paths extended from near the Fall Line to Chesapeake Bay in all the aquifers, and from the potentiometric high on the Delmarva Peninsula to Chesapeake Bay and the Atlantic Ocean in the surficial, upper Chesapeake, lower Chesapeake, and Castle Hayne-Piney Point aquifers (layers 10, 9, 8, and 7).

The southernmost hydrogeologic section, *C–C′* (fig. 13), is located near the North Carolina-Virginia State line and extends eastward toward the ocean (fig. 1). In contrast to the other hydrogeologic sections, flow to major surface-water bodies other than the Atlantic Ocean was not evident. Local recharge and downward vertical flow occurred in the updip parts of the aquifer system (not shown on the section). Lateral flow toward the sea occurred with little vertical flow between aquifers, as shown by the potentiometric contours. Upward vertical flow to the overlying ocean occurred along the eastern side of the section. There was upward flow (note apparent horizontal flow between layers shown in figure 13 due to the vertical exaggeration of the figure) between the 25-ft and 10-ft potentiometric contour in the deep aquifers (layers 2 and 3) and seaward of the 10-ft contour in shallower aquifers (layers 6 and 7). As the section continues seaward, vertical flow in the shallow aquifers (layers 9 and 10) becomes more pronounced, as shown by the more horizontal configuration of the 10- and 25-ft potentiometric contours.

Prepumping vertical leakage between adjacent aquifers was computed using the calibrated model. Vertical flow rates generally were small, less than 1 in/yr. Vertical flows were highest in or near the outcrop areas of aquifers and much less in the deeper, confined parts of the system. For example, downward vertical leakage to the upper Chesapeake aquifer (layer 9) from the overlying surficial aquifer (layer 10) occurred over much of the onshore area of the upper Chesapeake aquifer. The only exceptions were areas of upward flow to major streams



EXPLANATION

9 AQUIFER MODEL LAYER NUMBER AND NAME—Boundaries dashed where approximately located. Queried where doubtful
- 10, surficial
- 9, upper Chesapeake
- 8, lower Chesapeake
- 7, Castle Hayne-Piney Point
- 6, Beaufort-Aquia
- 5, Peedee-Severn
- 4, Black Creek-Matawan
- 3, upper Potomac and Magothy
- 2, middle Potomac
- 1, lower Potomac

CONFINING UNIT

25 SIMULATED POTENTIOMETRIC CONTOUR—Shows altitude of simulated prepumping water levels. Contour interval, in feet, is variable. Datum is sea level

GENERALIZED DIRECTION OF GROUND-WATER FLOW

A — A' LOCATION OF HYDROGEOLOGIC SECTION SHOWN IN FIGURE 1

FIGURE 11.—Hydrogeologic section *A–A'* showing generalized flow in New Jersey, prepumping conditions.

and rivers. The largest downward vertical flow rates occurred in areas of high water-table altitudes. The largest upward flow rates, 2 to 14 in/yr, tended to be beneath major streams. In the downdip, confined parts of each aquifer, vertical flow rates were less than 0.2 in/yr and were almost always upward. Distribution of vertical flow was most variable in the shallower aquifers in the system.

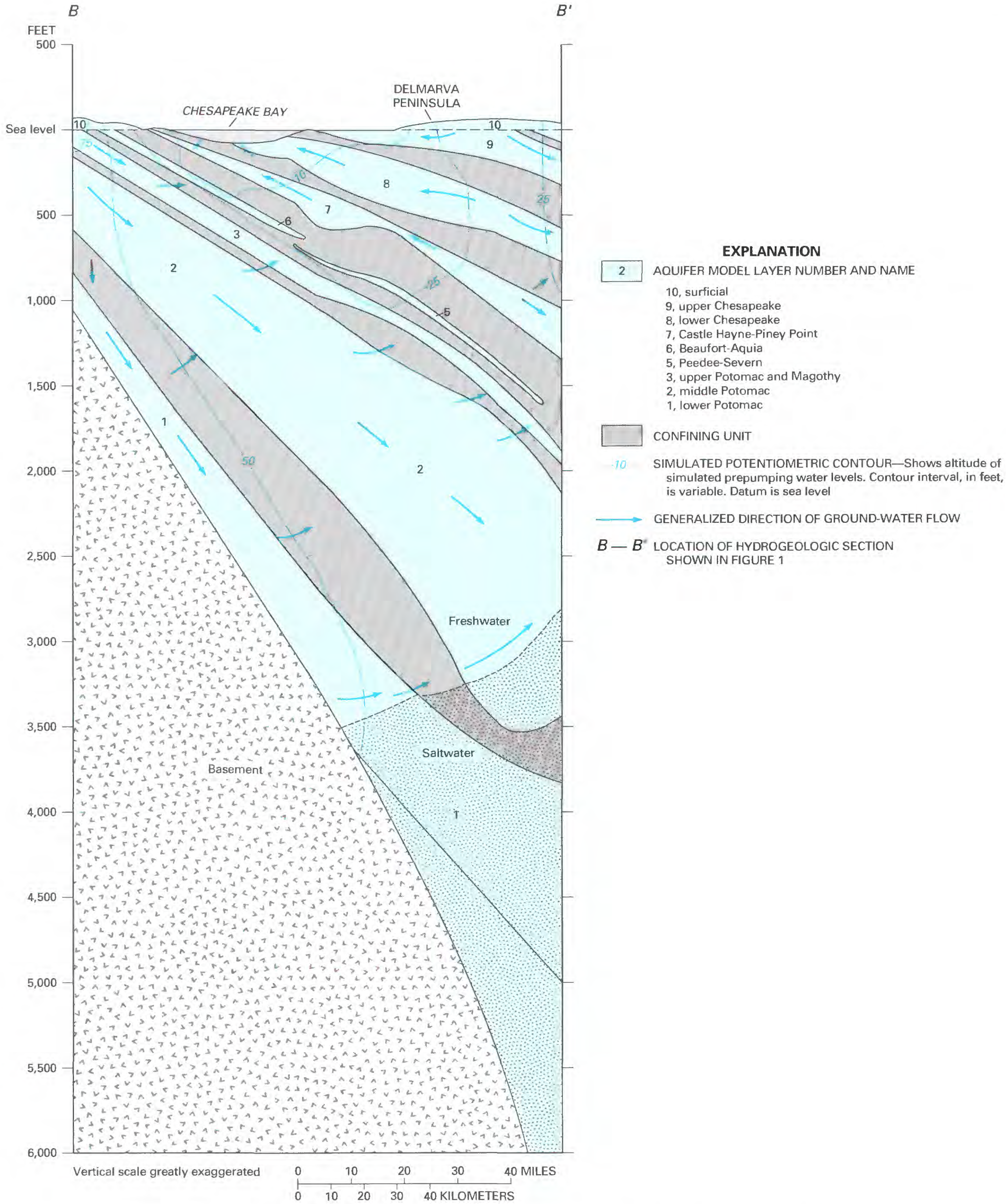


FIGURE 12.—Hydrogeologic section *B–B'* showing generalized flow in Maryland, prepumping conditions.

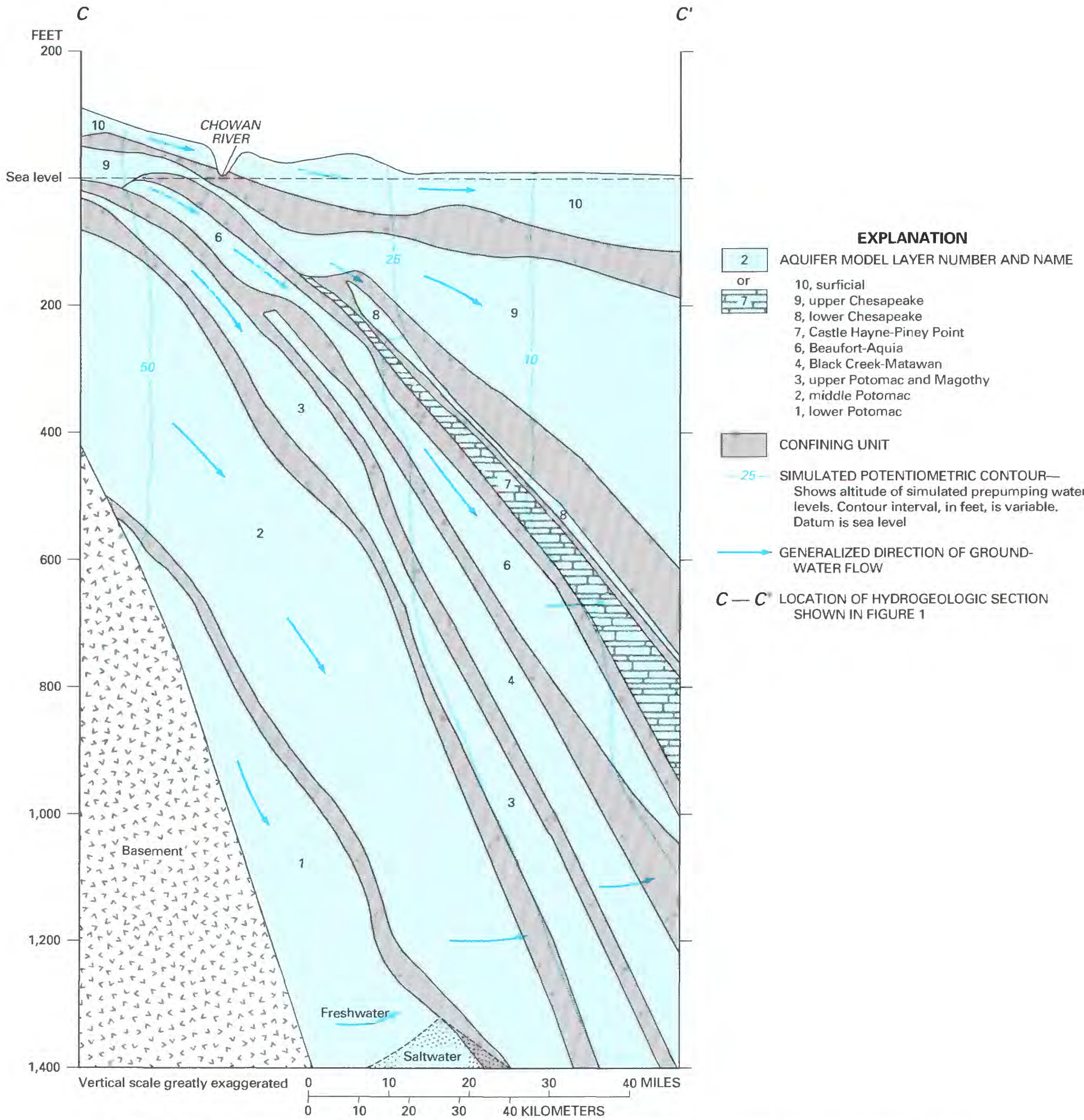


FIGURE 13.—Hydrogeologic section *C–C'* showing generalized flow in North Carolina, prepumping conditions.

In the deeper confining units of the system, downward vertical flow occurred in narrow bands located progressively closer to the Fall Line. Upward vertical flow occurred through the remainder of the deeper confining units. In areas where confining units are absent, vertical flow rates were highest, and the vertical hydraulic conductivity of the aquifers controlled the vertical flow. An example of this type of situation occurred between the upper and lower parts of the upper glacial aquifer (layers 7 and 6) on Long Island, where vertical flows were as much as 14 in/yr. The vertical flow rate normally was small; however, the area of aquifers in many cases was very large. Therefore, the amount of water moving vertically was very large and also very important in maintaining the regional flow system.

PUMPING CONDITIONS DURING 1980

Potentiometric surfaces

Potentiometric maps for 1980, based on measurements of heads from 1978 to 1981, were constructed to calibrate the transient model. Heads were contoured in areas having sufficient data for the subregional models. These interpreted potentiometric surfaces were used in calibration of the regional model. In areas of sparse control, only the data points were used to calibrate the model. Plates 15 through 17 show the simulated 1980 potentiometric surface for each aquifer and include the interpreted 1978–81 potentiometric surfaces or measured heads where available.

Measured head data were available for aquifers in North Carolina except the lower Potomac (layer 1) and the middle Potomac (layer 2) aquifers. Head data also were available for the lower Potomac aquifer (layer 1) in Maryland and New Jersey, mostly in the outcrop areas; for the middle Potomac aquifer (layer 2) in all States except North Carolina; for the upper Potomac and Magothy aquifers (layer 3) in all States; for the Black Creek-Matawan (layer 4), Peedee-Severn (layer 5), and lower Chesapeake (layer 8) aquifers in New Jersey; and for the upper Chesapeake aquifer (layer 9) in Delaware. Otherwise, head data were sparse. Also, head data were better defined for the shallow, updip parts of the aquifer system than for deeper, downdip parts. Thus, model calibration for the 1980 flow conditions was less rigorous in downdip areas.

The simulated 1980 potentiometric surface for the upper Chesapeake aquifer (layer 9, fig. *A*, pl. 15) was similar to the simulated prepumping potentiometric surface (fig. *A*, pl. 10), except for a small cone of depression in New Jersey. The simulated and measured heads for 1980 matched closely except at data points where heads were below sea level. These heads generally occurred within local cones of depression that could not be simulated because of the coarse grid of the model. A shallow cone of depression was simulated in the confined part of the aquifer in Cape May County, N.J., (pl. 1) with heads about 10 ft below sea level. In other parts of New Jersey, layer 9 represents the unconfined Kirkwood-Cohansey aquifer. The simulated potentiometric contours in these parts of the aquifer reflected land-surface elevations.

Few measured heads were available for comparison with the simulated 1980 potentiometric surface of the lower Chesapeake aquifer (layer 8, fig. *B*, pl. 15), except for New Jersey. On the Delmarva Peninsula and in North Carolina, simulated heads corresponded to measured heads, except for the data point of 6 ft in Maryland, where the simulated heads did not reflect the local influence of surface waters. Beneath Chesapeake Bay and near Dover, Del., two small areas with heads slightly below sea level were simulated, although no measured data existed in these areas for 1978–81. In the Dover area, heads deeper than 80 ft below sea level were measured in a local cone of depression during 1977 (Leahy, 1982a, p. 26). The deep heads in the center of the cone could not be simulated because of the coarse grid of the regional model. However, the shallower, more regional aspects (0 and −10-ft contours) of the cone corresponded to the measured 1977 potentiometric surface. Along the coast of New Jersey, a cone of depression was simulated in the subregional confined Kirkwood aquifer (layer 8). Simulated heads of about 55 ft below sea level at its center were 15 ft higher than measured levels. Updip to the northeast in New Jersey, layer 8 represents the lower part of the unconfined Kirkwood-Cohansey aquifer. In this area and in updip areas in Delaware, Maryland, and North Carolina, the simulated potentiometric surface was similar to topography. Heads in these areas were relatively unchanged from prepumping conditions (fig. *B*, pl. 10).

The simulated 1980 potentiometric surface for the Castle Hayne-Piney Point aquifer (layer 7) is shown in figure *C*, plate 15. A large number of measured head data were available for comparison with the simulated potentiometric surface in North Carolina. Only sparse measured head data for 1978–81 were available for New Jersey, Delaware, Maryland, and Virginia. The location and depth of a simulated cone of depression centered along the Pamlico River in North Carolina was consistent with the measured data. Other large cones of depression were simulated in Delaware, Maryland, and Virginia; however, the locations and depths of the cones did not match closely the few heads measured in 1978–81. Potentiometric surfaces are available for 1976 and 1977 (Williams, 1979, pl. 3; Leahy, 1982a, p. 21). These surfaces show major cones of depression centered around Dover, Del., and Cambridge, Md., that are connected by a trough of lowered head. The simulated 1980 potentiometric surface (fig. *C*, pl. 15) reflected the regional trend apparent in the measured potentiometric surfaces for 1976 and 1977. No cones were simulated on Long Island, where layer 7 represents the upper part of the unconfined upper glacial aquifer, and in New Jersey, where withdrawals were minor. In New Jersey and Delaware, the aquifer does not crop out, and simulated heads were not influenced by the topography or the location of surface-water bodies, as they were in the overlying aquifers. In contrast, the simulated 1980 potentiometric surface in updip areas of Maryland, Virginia, and North Carolina strongly reflected the topography and the location of streams. Withdrawals from this aquifer in North Carolina were relatively high (36 Mgal/d in 1980) near and in the aquifer's outcrop area. These withdrawals caused only slight changes in the potentiometric surfaces from prepumping to 1980 conditions. However, farther north, the aquifer does not crop out, and moderate withdrawals (about 10 Mgal/d) caused development of regional cones of depression in Delaware and Maryland.

Only sparse measured head data were available to compare with the simulated 1980 potentiometric surface for the Beaufort-Aquia aquifer (layer 6). The simulated potentiometric surface and measured data are shown in figure *A*, plate 16. In Maryland and Virginia, shallow cones with heads about 50 ft below sea level were simulated along the downdip limit of the modeled aquifer. Shallow cones with heads about 10 ft below sea level were also simulated along the downdip limit of the aquifer in Delaware and along the coast of North Carolina. Simulated 1980 heads and flow patterns were relatively unchanged from prepumping conditions (fig. *A*, pl. 11) in the updip areas near the aquifer outcrop. This includes Long Island, where layer 6 represents the lower part of the unconfined upper glacial aquifer, and New Jersey, where the aquifer is present only in a narrow band adjacent to its outcrop area. The influence of topography on 1980 heads in the updip areas of the Beaufort-Aquia aquifer was evident in the simulated potentiometric surface throughout the study area.

The simulated 1980 potentiometric surface for the Peedee-Severn aquifer (layer 5) generally matched the measured 1980 heads in New Jersey and North Carolina within 20 ft (fig. *B*, pl. 16). A major cone of depression with heads about 150 ft below sea level was simulated in New Jersey. Heads in 1980 were simulated to be about 0 to 10 ft below sea level in southern Delaware and on the western shore of Maryland, although no measured heads were available for these areas. In North Carolina and on Long Island, heads did not change significantly from prepumping conditions (fig. *B*, pl. 11).

The simulated 1980 potentiometric surface for the Black Creek-Matawan aquifer (layer 4) is shown in figure *C*, plate 16, with interpreted contours in New Jersey and measured head data in North Carolina. In New Jersey, a steep cone of depression of about 200 ft below sea level was simulated. In North Carolina, a cone was simulated to a depth of about 50 ft below sea level. Both of these cones closely matched interpreted and measured heads. The high ground-water altitudes in western North Carolina and along the Fall Line in New Jersey were simulated closely. In New Jersey and western North Carolina, the simulated 1980 potentiometric surfaces indicated flow toward the two major cones of depression, rather than toward the downdip limit of the aquifer, as for prepumping conditions (fig. *C*, pl. 11).

The simulated and interpreted potentiometric surfaces and measured heads for the upper Potomac and Magothy aquifers (layer 3) are shown in figure *A*, plate 17. Major potentiometric features, both simulated and interpreted, included (1) ground-water highs near the Fall Line in North Carolina, Virginia, Maryland, and New Jersey, and along the central axis of Long Island, (2) major cones of depression with heads lower than 50 ft below sea level in Virginia and New Jersey, and (3) a smaller regional cone with heads about 25 ft below sea level in central North Carolina. Smaller regional cones were simulated on western Long Island and in southern Maryland, although no measured heads for these areas were available. Simulated heads were less than 25 ft above sea level throughout much of the aquifer area; exceptions were Long Island and areas along the Fall Line in New Jersey, Maryland, and Virginia. The local ground-water high along the central axis of Long Island was simulated about 30 ft lower than the interpreted potentiometric surface, and simulated contours lacked the detail of the interpreted contours. In Maryland, Virginia, and North Carolina, several scattered points with head data below sea level were not matched and are assumed to be affected by local pumping. However, these low heads are generally simulated by the subregional flow models.

Simulated 1980 heads and ground-water flow directions in the upper Potomac and Magothy aquifers (layer 3) in southwestern North Carolina and on Long Island were similar to the simulated prepumping heads (fig. *A*, pl. 12) and flow directions. In New Jersey, Virginia, and North Carolina, the difference between the simulated prepumping and 1980 potentiometric surface was the development of regional cones of depression caused by withdrawals from the upper Potomac and Magothy aquifers and other overlying and underlying aquifers. In Maryland and Delaware, minor cones of depression altered prepumping flow directions locally.

A large amount of measured and interpreted head data was available for the middle Potomac aquifer (layer 2). The simulated potentiometric contours matched the regional features shown by the interpreted potentiometric surface and the measured head data (fig. *B*, pl. 17). High heads were simulated in southeastern North Carolina, near the Fall Line in Maryland and New Jersey, and along the central axis of Long Island. However, simulated heads in some places on Long Island were about 30 ft lower than the interpreted potentiometric surface. Measured heads along the Potomac River in Maryland were 10 to 60 ft lower than simulated heads. These differences were assumed to represent local rather than regional flow patterns. The major cones of depression centered in New Jersey and Virginia, and a shallower cone on Long Island, were in good agreement with measured data. The regional discharge areas shown in the 1980 simulated potentiometric surface for the middle Potomac aquifer (layer 2) coincided with the major cones of depression. Ground-water flow was from the ground-water highs near the Fall Line to these large cones. Generally, simulated heads within the recharge areas

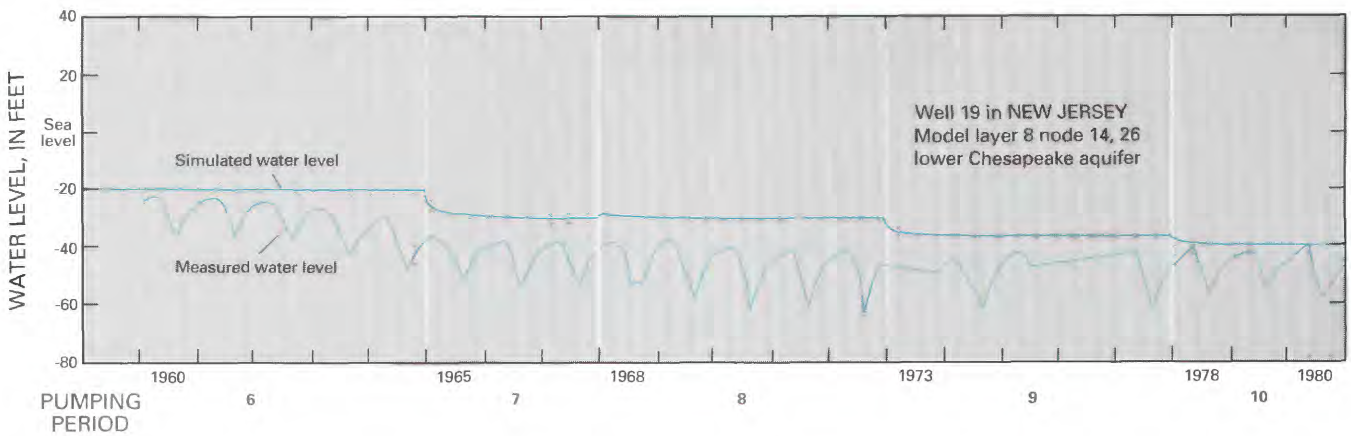


FIGURE 14.—Hydrographs of simulated and measured heads for the lower Chesapeake aquifer (layer 8) in New Jersey. Well location shown on plate 18*A*.

near the Fall Line, along the North Carolina-South Carolina border, and along the central axis of Long Island were about 15 ft lower than measured prepumping heads in these areas (fig. *B*, pl. 12).

Simulated heads for the lower Potomac aquifer (layer 1) were in good agreement with measured values (fig. *C*, pl. 17). Large cones of depressions observed in New Jersey, Delaware, Maryland, and Virginia were simulated. However, the inability of the model to match some measured heads, such as the head of 178 ft below sea level in Delaware, indicates that local conditions could not be simulated. Simulated heads beneath Chesapeake Bay generally were below sea level in Virginia and Maryland. The simulated 1980 potentiometric surface for the lower Potomac aquifer differed from the simulated prepumping potentiometric surface (fig. *C*, pl. 12) in that (1) the major discharge areas for the 1980 flow system were the major cones of depression rather than the major estuaries and bays, and (2) in 1980 almost all offshore heads were below sea level.

By 1980, withdrawals had caused significant changes in the potentiometric surfaces of aquifers and in the flow directions. In deep, confined aquifers, pumping has caused large major regional cones of depression. In many instances, these regional cones cover several States. Heads were below sea level throughout large parts of these aquifers, in many cases to the simulated seaward boundary of the system. In some instances, the simulated seaward no-flow boundary may have caused excessive drawdown in the cones of depression. The effect of the artificial seaward no-flow boundary at the 10,000-mg/L chloride concentration on simulated heads was evaluated, and is discussed later in this report. Steep lateral hydraulic gradients at the major cones located updip in the system suggest that water was diverted from the outcrop areas and caused a significant decrease in the discharge to streams. In cones located farther downdip in the system, vertical hydraulic gradients suggest that vertical leakage was an important source of the pumped water. It appears that the size and shape of regional cones is dependent on the hydraulic characteristics of the system, the quantity being pumped, and the location of the pumping relative to the surficial aquifer and streams and to the seaward limit of the aquifer. Cones are generally absent in the 1980 potentiometric surfaces for the shallow aquifers (for example, layer 9), and the potentiometric surfaces reflect the local nature of the shallow flow. Any water pumped from the shallow aquifer was supplied by an adjustment in local recharge and discharge to streams. This caused cones of depression to be small and shallow, seldom having any regional significance.

HYDROGRAPHS

Well hydrographs were used to assess the calibration of the model through time because data were too sparse to construct potentiometric surfaces for each pumping period. Several hundred hydrographs were used to calibrate the subregional models. Representative hydrographs from 74 wells were used to evaluate the transient performance of the regional model. Figure *A*, plate 18, shows for each well the assigned well number used in this report, the local identifier, the model layer in which each well was screened, and the latitude and longitude of the well. Hydrographs of 19 of the 74 wells are shown in figures 14 through 20. These hydrographs provide a good areal and temporal representation of head throughout the study area.

Hydrographs were simulated using one head at each time step; the head was a spatially averaged combination of the heads in the three adjacent model grid blocks nearest the location of each observation well. The simulated heads in the hydrographs were a linear interpolation of the three spatially averaged grid block heads

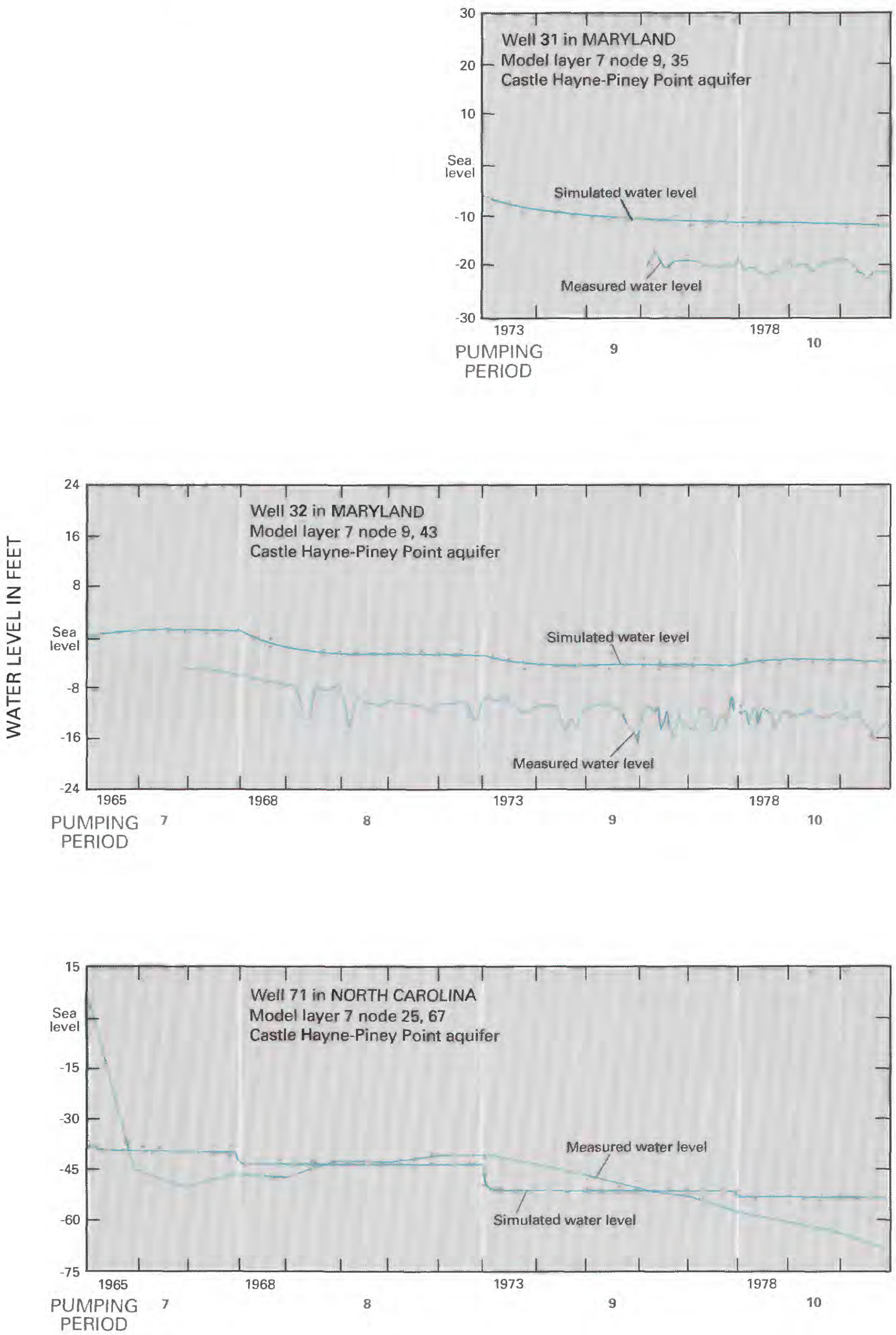


FIGURE 15.—Hydrographs of simulated and measured heads for the Castle Hayne-Piney Point aquifer (layer 7) in Maryland and North Carolina. Well locations shown on plate 18*A*.

based on the distance of the well from the center of each grid block. This procedure yielded a more accurate means of comparing measured and simulated heads at observation wells than just using the head at the nearest node. The approach is based on the concept that the head at a point is located in a triangular plane defined by heads at the three nearest nodes. However, limitations still were present in this analysis. Factors causing those limitations included (1) local variations in withdrawals within a regional grid block, (2) interpolation based on fitting a plane rather than a curved surface to the heads between adjacent nodes, and (3) the scale dependence inherent in model computed heads. A detailed discussion of the effect of these factors is presented in a subsequent section of this report.

The transient response of the model was accepted as calibrated based on the comparison of simulated and measured heads in the 74 hydrographs. Although some simulated hydrographs did not match measured water-level declines, most were within 10 ft of the measured data. In areas that were affected by nearby withdrawals, simulated heads were as much as 25 to 50 ft higher. However, the simulated long-term declines generally depict the regional trends.

A measured hydrograph for well 19 screened in the lower Chesapeake aquifer (layer 8) in New Jersey showed seasonal head fluctuations that were not simulated by the regional model (fig. 14). Such differences between simulated and measured hydrographs were common for wells in or near unconfined parts of the aquifers. The simulated and measured hydrographs showed similar long-term (1960–80) declines of 20 ft, although the simulated heads were about 10 ft higher than the measured heads.

Simulated and measured head declines for two wells in the Castle Hayne-Piney Point aquifer (layer 7) in Maryland were similar, about 5 ft for well 31 and 6 to 8 ft for well 32 (fig. 15). However, simulated heads were 5 to 12 ft higher than measured heads for these wells. Simulated and measured heads for well 71 in the same aquifer in North Carolina correlated closely for the period 1965 to 1978 (fig. 15), but from 1978 through 1980, the simulated heads declined less than the measured heads. In 1980, measured heads were about 15 ft lower than simulated heads. The most notable similarity in the two hydrographs for the North Carolina well was the rapid decline of about 60 ft in 1965.

Simulated heads for well 52 in the Beaufort-Aquia aquifer (layer 6) in Virginia closely matched measured heads from 1964 through 1980 (fig. 16). In contrast, simulated heads for well 44 in the same aquifer were about 10 to 15 ft higher than measured heads but showed the same 30-ft decline from 1943 through 1980. Measured and simulated hydrographs for well 65 in the Peedee-Severn aquifer (layer 5) in North Carolina were within 10 ft of each other from 1963 through 1980 (fig. 16). A similar decline of 9 ft occurred on both hydrographs.

The measured and simulated heads for well 12 in the Black Creek-Matawan aquifer (layer 4) in New Jersey declined about 80 ft from 1965 through 1980 (fig. 17). However, simulated heads were about 20 ft higher than measured heads. The measured and simulated heads for well 61 in the same aquifer in North Carolina were similar and showed similar declines of about 80 ft from 1946 through 1980 (fig. 17).

The simulated heads in well 2 in the lower part of the Magothy aquifer (layer 3) on Long Island (fig. 17) were typically 4 to 6 ft above the measured heads. Both simulated and measured hydrographs declined about 8 ft from 1936 through 1980. The agreement between the simulated and measured heads in the upper Potomac and Magothy aquifers (layer 3) for well 54 in Virginia and well 11 in New Jersey was particularly good (fig. 18). The simulated heads were generally within 10 to 15 ft of the measured heads but were higher. The measured and simulated heads showed the same decline through time. In contrast, simulated and measured hydrographs for well 27 in the same aquifer in Maryland did not correlate as closely as the other hydrographs (fig. 18). From 1963 through 1971, measured and simulated heads were within about 10 to 20 ft of each other, but from 1972 through 1980 simulated heads declined less than measured heads. In 1980, simulated heads were about 40 ft too high. The difference between the simulated and measured hydrographs suggest that well 27 was located in an area of heavy withdrawals. Because of the coarse grid of the model, a better match of the measured and simulated hydrographs was not achieved.

Simulated and measured heads for well 56 in the middle Potomac aquifer (layer 2) in North Carolina were within 15 ft of each other from 1964 through 1980 (fig. 19). The simulated heads for well 7 in the same aquifer in New Jersey were higher than the measured heads from 1934 through 1980. However, the simulated and measured head declines were similar, about 60 ft. Measured head data were available for well 39 in Virginia for the 74-year period of record from 1907 through 1980, the longest period of record for the Coastal Plain hydrographs. Simulated and measured declines of about 75 ft occurred during this period, and simulated heads closely matched the measured data (fig. 19).

The hydrograph for well 45 in the lower Potomac aquifer (layer 1) in Virginia closely matched the measured heads for the 12-year period 1969–80 (fig. 20). The simulated hydrograph for well 3 in the same aquifer in New Jersey showed heads and a regional downward trend that were similar to the measured heads for the 16-year period of record from 1965 through 1980 (fig. 20). The measured head decline in well 24 in northern Delaware from 1955 through 1980 was about 140 ft (fig. 20). This compared favorably with the simulated decline of 145 ft. The major difference between the hydrographs for this well was an initial head of about 10 ft above sea level for the measured hydrograph compared with an

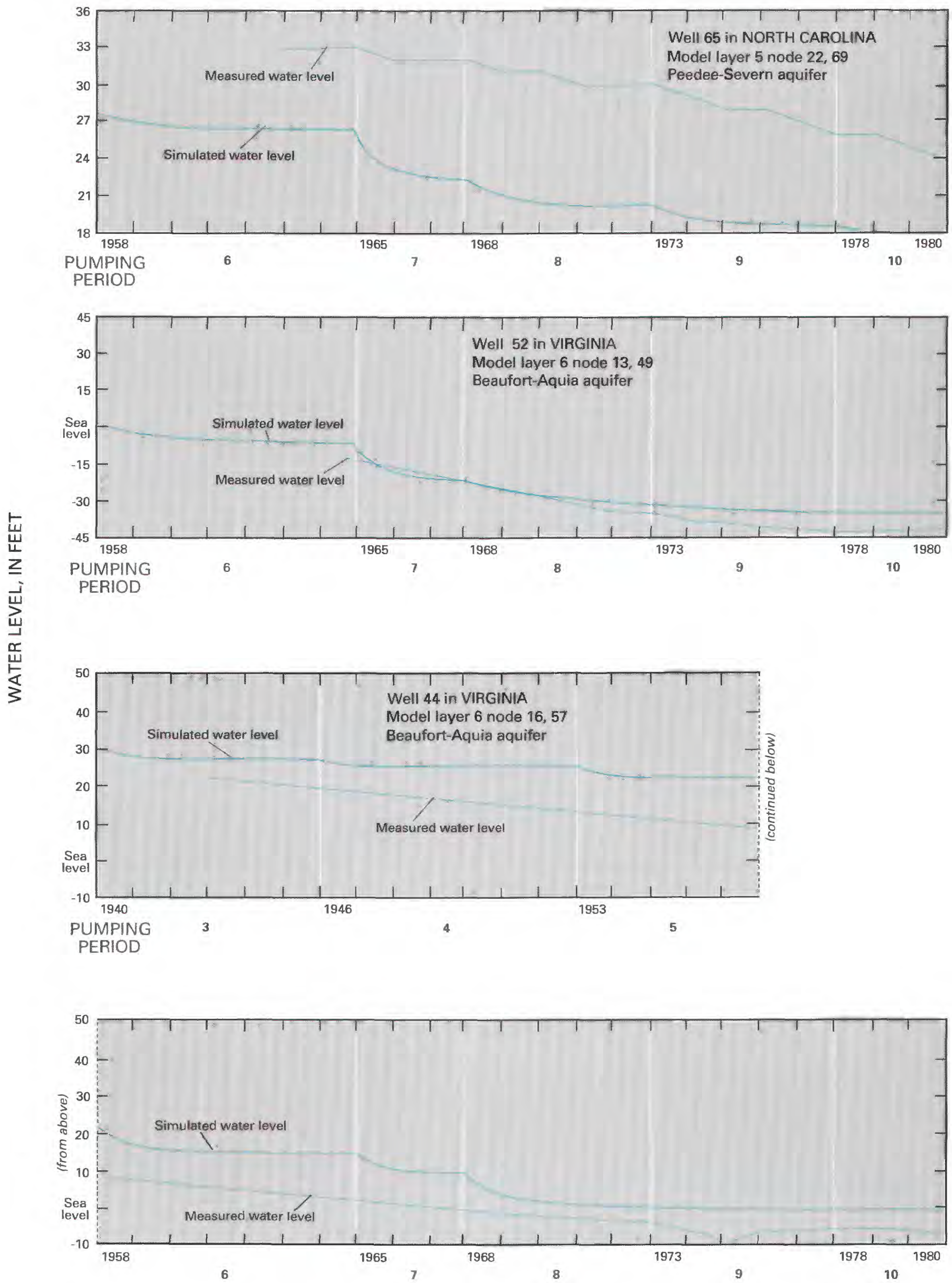


FIGURE 16.—Hydrographs of simulated and measured heads for the Peedee-Severn aquifer (layer 5) in North Carolina and the Beaufort-Aquia aquifer (layer 6) in Virginia. Well locations shown on plate 18*A*.

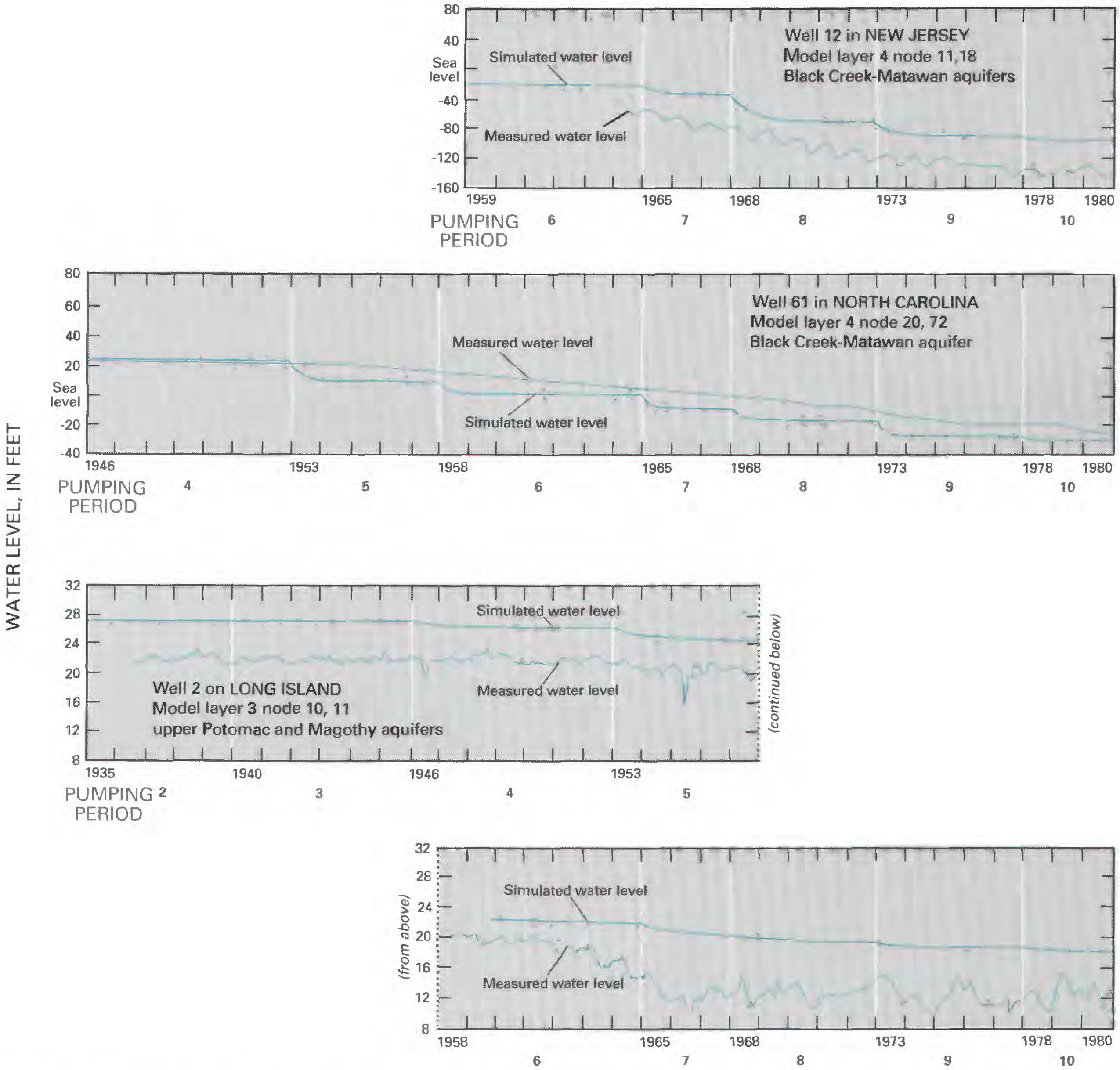


FIGURE 17.—Hydrographs of simulated and measured heads for the upper Potomac and Magothy aquifers (layer 3) on Long Island and the Black Creek-Matawan aquifer (layer 4) in New Jersey and North Carolina. Well locations shown on plate 18*A*.

initial head of about 30 ft below sea level for the simulated hydrograph. This difference was due in part to inaccurate estimates of the areal and temporal distribution of the pre-1958 withdrawals. Of particular note were similarities in the simulated and measured declines early in pumping period 5 and the slight recovery of head in pumping period 6.

Regional ground-water budget

The components of the ground-water budget for 1980 were ground-water recharge, flow into the system from the south, withdrawals from wells, ground-water discharge to the ocean and other surface-water bodies, and water released from aquifer storage. The simulated regional budget for 1980 flow conditions is shown with the prepumping budget in figure 10. The specified areal recharge was about 40,000 Mgal/d, as in the prepumping simulation. The simulated rate at which water was released from aquifer storage was about 21 Mgal/d. Flow into the modeled area from South Carolina in the middle Potomac aquifer (layer 2) was about 7.5 Mgal/d in 1980. This is about a 23-percent increase over prepumping conditions. Thus, withdrawals have induced additional flow from South Carolina into North Carolina in the middle Potomac aquifer. However, decreases in aquifer

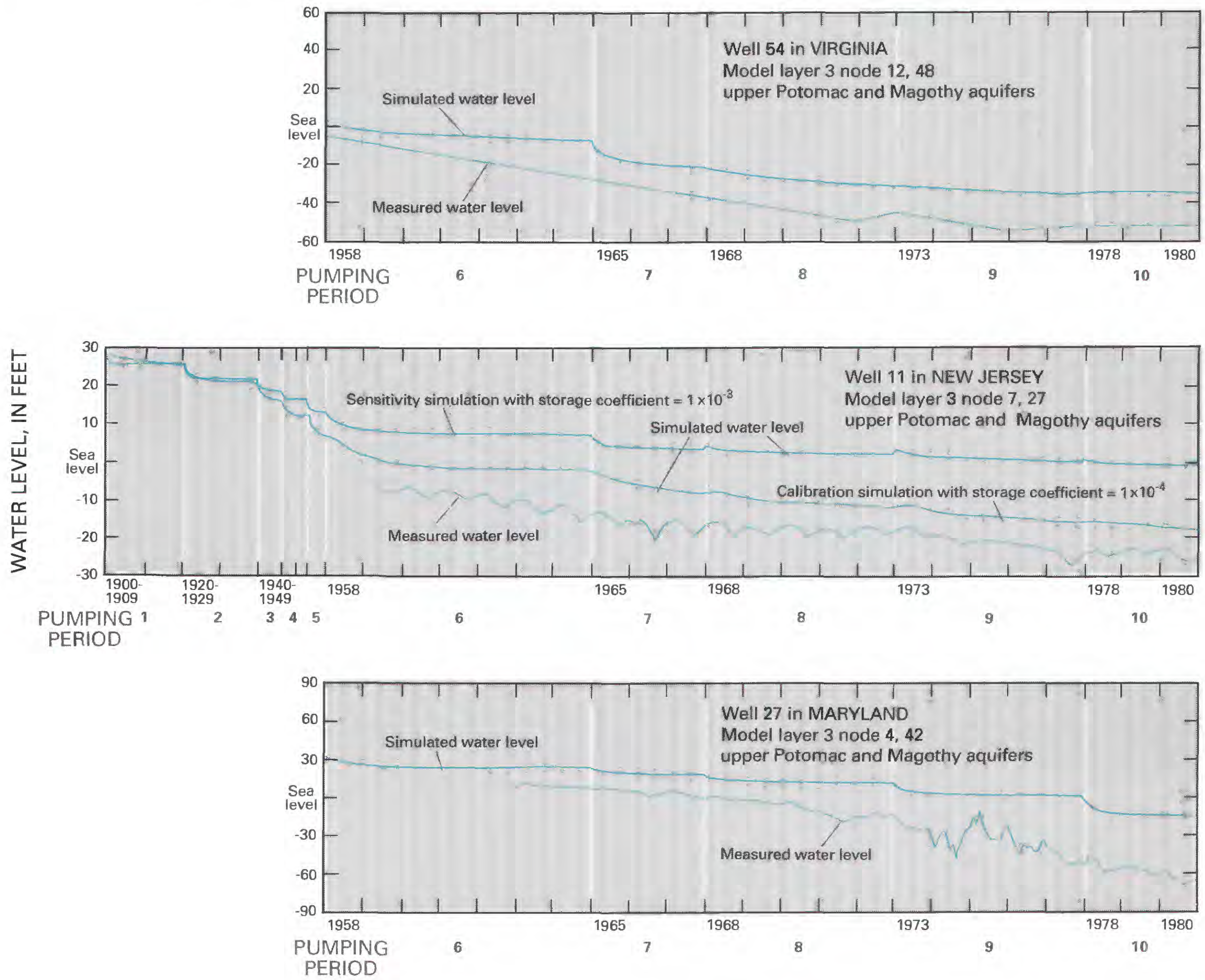


FIGURE 18.—Hydrographs of simulated and measured heads for the upper Potomac and Magothy aquifers (layer 3) in Virginia, New Jersey, and Maryland. Well locations shown on plate 18*A*.

storage and inflow from South Carolina are negligible compared with total areal recharge.

Withdrawals (not including irrigation, rural, and domestic uses) from the system in 1980 were 1,210 Mgal/d, or about 3 percent of the recharge that entered the ground-water flow system. The remaining available ground water, about 38,820 Mgal/d, was discharged to the ocean and surface-water bodies; discharge from the unconfined aquifers was 38,568 Mgal/d and from the confined aquifers was 252 Mgal/d. By far, shallow ground-water flow accounts for the majority of flow through the aquifer system. However, withdrawals in 1980 were about twice the prepumping deep percolation of 592 Mgal/d, and because withdrawals were not distributed evenly over the system, the local effects on the system, including water released from aquifer storage, were significant. Deep percolation into the confined aquifers was increased by pumping to about 1,330 Mgal/d in 1980, as compared with 592 Mgal/d under prepumping conditions.

The regional water budget indicates that the system rapidly approaches a steady-state condition because the amount of water removed from storage (21 Mgal/d in 1980) was negligible compared with the amount of water withdrawn. Hydrographs of simulated heads (figs. 14–20) showed that water was initially removed from storage as withdrawals increased. Generally, the hydrographs rapidly flatten within each pumping period, showing a reduction in the amount of water released from storage and an increase in the amount of lateral and vertical flow to balance the withdrawals. Simulated withdrawals were averaged over discrete pumping periods. As a new increase in withdrawals was simulated, hydrographs showed a rapid decline in heads for a short period of time, followed by a decrease in the decline. At the end of each pumping period, the regional water

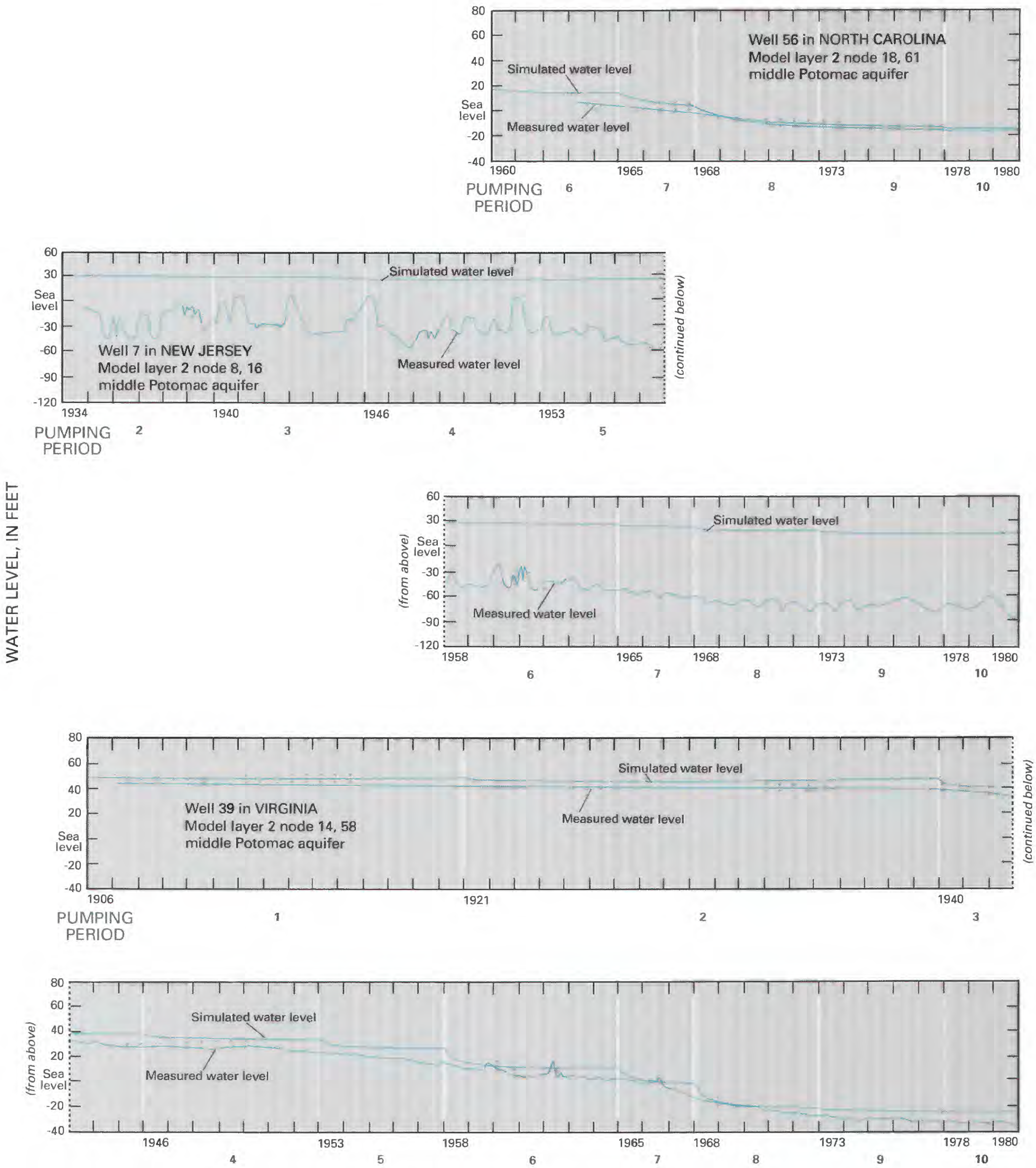


FIGURE 19.—Hydrographs of simulated and measured heads for the middle Potomac aquifer (layer 2) in North Carolina, New Jersey, and Virginia. Well locations shown on plate 18A.

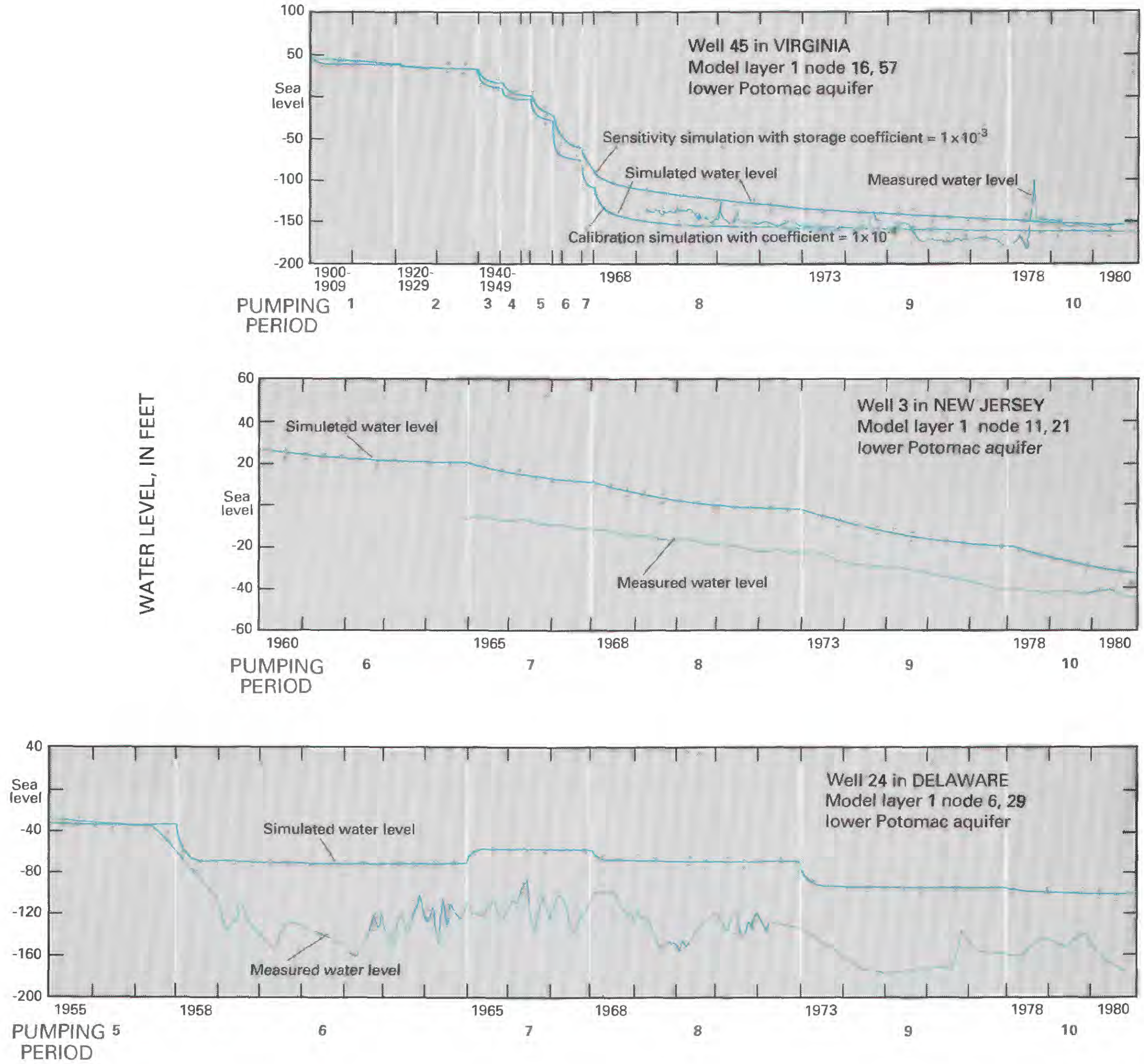


FIGURE 20.—Hydrographs of simulated and measured heads for the lower Potomac aquifer (layer 1) in Virginia, New Jersey, and Delaware. Well locations shown on plate 18A.

budget showed only a small volume of water being released from aquifer storage. Thus, a new steady-state condition was approached. The system adjusts rapidly (in less than 3 years) to simulated changes in withdrawals, and essentially the simulated system progresses from one steady-state condition to another as withdrawals change through time. The actual physical system does not achieve steady-state conditions as they were simulated because withdrawals do not change in a stepwise manner, but change continuously through time.

A forced steady-state condition with 1980 withdrawals was simulated to determine the additional head decline that would occur if the rate of pumping remained constant. The results of this simulation showed additional head declines of more than 5 ft only in part of New Jersey. If the pumping rate were to remain constant indefinitely, heads would continue to decline, at most, an additional 30 ft in the middle Potomac aquifer (layer 2) along the simulated seaward boundary of the aquifer. Similar declines would occur in overlying and underlying aquifers (layers 1 and 3). Heads would decline less than 20 ft onshore, and would decrease in noncoastal areas. The time to reach a steady-state condition was determined by simulating additional time until heads throughout the system ceased to decline. This simulation showed that the volume of water coming from aquifer storage would decrease with time and eventually would reach zero. The time required to reach this steady-state condition with 1980 withdrawals was an additional 40 years. It is worth emphasizing that most of the water from storage would come into the system rapidly (in 3 years). However, minor amounts of water would continue to be released from storage for much longer periods of time, especially in the deep confined aquifers, until the system reached equilibrium.

Changes in deep percolation and ground-water discharge to surface-water bodies

According to Theis (1940), the water pumped from wells comes from an increase in recharge, a decrease in discharge, a depletion in aquifer storage, or a combination of these sources. The volume of water depleted from storage in 1980 was very small, less than 2 percent of the water pumped. Therefore, most of the 1980 pumpage came from an increase in recharge and a decrease in discharge. The water pumped from the confined aquifers was supplied by a redistribution of flow in the shallow surficial aquifer (layer 10) and the outcrop areas of the confined aquifers. Ground-water recharge that under prepumping conditions discharged as base flow to nearby streams or to the ocean, major bays, and estuaries was diverted into the confined aquifers and eventually to pumping wells.

Figure *B*, plate 18, shows the simulated change in discharge to surface-water bodies from prepumping conditions to 1980 flow conditions. In offshore areas, upward discharge from underlying aquifers to the ocean has decreased. However, the change was less than 0.1 in/yr, except along the southern shore of Long Island and locally along the shoreline of Delaware, Maryland, and North Carolina. In those areas, discharge decreased as much as 4 in/yr offshore of Long Island and a few tenths of an inch per year offshore of Delaware, Maryland, and North Carolina. Regional ground-water discharge to streams and rivers decreased significantly in many areas. In parts of Nassau County, Long Island, discharge decreased by more than 25 in/yr. This was expected because Long Island simulated withdrawals (fig. 8) were the highest in the study area.

Withdrawals in New Jersey were the second highest in the study area. Significant withdrawals were made from in or near the outcrop of the Magothy, middle Potomac, and lower Potomac aquifers (layers 3, 2, and 1) along the Delaware River in southern New Jersey and near Raritan Bay. Discharge to streams has decreased significantly in these areas, as much as 15 and 12 in/yr, respectively. Smaller decreases of as much as 4 in/yr occurred in localized areas along the ocean.

In Maryland and Delaware, shallow withdrawals caused a reduction in discharge to streams. Discharge in areas along the Fall Line decreased by as much as 4 in/yr. In the Baltimore area, stream discharge decreased locally by 1 in/yr. Similarly, areas on the Delmarva Peninsula where the shallow upper and lower Chesapeake aquifers (layers 9 and 8) were pumped showed a reduction in discharge to streams of as much as 2 in/yr.

In Virginia, reductions in discharge to streams were less than 1 in/yr, but were highest along the Fall Line. It is noteworthy that large withdrawals (about 50 Mgal/d) near Franklin, Va., did not have a dramatic local effect on ground-water discharge to streams. The withdrawals in Franklin were from deep confined aquifers, and vertical leakage through overlying confining units provided much of the pumped water. Ground-water discharge to streams decreased by only a few tenths of an inch per year, but this leakage was spread over a large area of the surficial aquifer. In North Carolina, the Castle Hayne aquifer (layer 7), which outcrops in a narrow band along the coast, is heavily pumped. The effect of this large withdrawal was relatively large decreases in discharge, as much as 8 in/yr (fig. *B*, pl. 18). Detailed analysis of stream base flow in this area may support the simulated areal variation in ground-water discharge to streams. However, because of the large size of the model blocks, any comparison of measured base flow with simulated ground-water discharge should be made with a model having smaller block sizes.

Under prepumping conditions, recharge from the surficial aquifer to deeper aquifers (deep percolation) occurred over 26 percent of the study area and averaged about 0.5 in/yr. Pumpage has caused a major change in the areal distribution of the deep percolation. In 1980, 45 percent of the study area was a recharge area for deep confined aquifers; this is an increase of 19 percent of the study area from the prepumping conditions.

Horizontal flow and velocities

Lateral Darcy ground-water flow velocities increased in cone-of-depression areas where head gradients increased the most. The most dramatic velocity changes between prepumping and 1980 flow conditions were in updip areas where the slope of the cones was steepest. Withdrawals in downdip parts of the confined aquifers caused lateral velocities to increase moderately (less than 5 ft/yr) over very large areas. However, because the prepumping flow velocities in these downdip areas were very small, pumping in 1980 has caused relatively large changes in flow velocities. In contrast, the shallow and largely unconfined aquifers had large prepumping velocities (greater than 10 ft/yr), and pumping in 1980 has produced only subtle changes in lateral velocities. As with the prepumping flow system, the potentiometric surfaces and calculated ground-water velocities are used to describe the regional 1980 flow system for three aquifers: the upper Chesapeake aquifer (layer 9), the

Castle Hayne-Piney Point aquifer (layer 7), and the upper Potomac and Magothy aquifers (layer 3).

Simulated lateral flow velocities for 1980 flow conditions in the upper Chesapeake aquifer (layer 9) (fig. *A*, pl. 19) ranged from about 0.01 to 100 ft/yr. Because the aquifer typically has high lateral velocity and consists of many localized flow systems, changes in lateral flow velocities due to local pumping were not apparent on a regional scale. The only location where velocity has increased noticeably is in Cape May County, N.J. Lateral velocities increased from about 1 ft/yr under prepumping conditions to about 5 ft/yr in 1980. Lateral flow directions in the offshore area may not be accurate. The hydraulic gradients and Darcy velocity are low, and the simulated onshore direction of flow may reflect the regional trend of the specified freshwater equivalent head that was used for the offshore top boundary of the model.

Lateral flow velocities for the 1980 flow directions in the Castle Hayne-Piney Point aquifer (layer 7) (fig. *B*, pl. 19) ranged from about 100 ft/yr in outcrop areas to about 0.01 ft/yr along the seaward boundary of the aquifer. Pumping from the aquifer increased the lateral flow velocities and changed the flow direction toward the major withdrawal centers. In central Delaware, lateral flow velocities increased from about 1 ft/yr under prepumping conditions to more than 10 ft/yr in 1980. Near Cambridge, Md., lateral velocities increased from 0.1 ft/yr to about 1 ft/yr. In North Carolina, lateral velocities increased from 10 to 100 ft/yr in the cone of depression of the Castle Hayne aquifer.

In the upper Potomac and Magothy aquifers (layer 3), lateral flow velocities in 1980, shown in figure *C*, plate 19, ranged from about 100 ft/yr in the updip part of the aquifer to less than 0.01 ft/yr near the seaward boundary of the aquifer. Withdrawals increased lateral flow velocities near major cones. For example, in 1980 flow velocities exceeded 10 ft/yr, and locally were greater than 100 ft/yr, along the Delaware River in southern New Jersey. Prepumping flow velocities in this area were 5 to 10 ft/yr (fig. *C*, pl. 14). Similar velocities were common along the outcrop paralleling the Fall Line, in northern Delaware, Virginia, and North Carolina. In the Franklin, Va., area, Darcy flow velocities in 1980 were about 5 ft/yr within the cone of depression. Moderate flow velocities toward the center of the cone suggest that the pumped water was derived from lateral flow over a large area (the cone includes an area greater than 5,000 mi^2) and that vertical leakage was a major source of the pumped water.

Luzier (1980, p. 63) computed lateral particle velocities for 1973 in two areas of southern New Jersey for the Potomac-Raritan-Magothy aquifer system. In the present study, Luzier's aquifer system was divided into three aquifers (equivalent to the upper Potomac and Magothy, middle Potomac, and lower Potomac aquifers (layers 3, 2, and 1, respectively)), and his particle velocities were converted to Darcy velocities (by multiplying the particle velocity by a porosity of 0.25), which were then compared with the lateral velocities resulting from the regional model. In the vicinity of the Delaware River and near Clayton, N.J. (about 17 mi southeast of the river), the lateral Darcy velocities estimated by Luzier were about 74 ft/yr and 8 ft/yr, respectively. Lateral Darcy velocities computed from the regional model, 50 to 100 ft/yr along the Delaware River and 5 to 10 ft/yr in the Clayton area, agreed closely with Luzier's findings.

The simulated change in base flow to streams (fig. *B*, pl. 18) showed large decreases along the outcrops. This suggests that the majority of water being pumped was either induced recharge from streams or reduced ground-water discharge in the outcrop areas. Indeed, the hydraulic gradient of the potentiometric surfaces near regional cones (pls. 15–17) steepened toward the Fall Line. It is noteworthy that the updip hydraulic gradient for the regional cone in Virginia was not as steep as the regional cone in New Jersey. This difference suggests that in Virginia, the water pumped was derived not only from horizontal flow in the aquifer, but also from vertical flow from adjacent aquifers. In contrast, the southern New Jersey cone had a very steep hydraulic gradient toward the Delaware River, suggesting that the river was a major source of the water being pumped. The simulated change in discharge to streams (fig. *B*, pl. 18) supports this conclusion.

Changes in vertical leakage

Changes in vertical leakage between adjacent aquifers were computed using the simulated results for prepumping and 1980 conditions. Areas and amounts of change in vertical leakage are shown on plates 20 through 22. The change in vertical leakage is discussed only for the confining units overlying the upper Chesapeake aquifer (layer 9), the Castle Hayne-Piney Point aquifer (layer 7), and the upper Potomac and Magothy aquifers (layer 3).

Figure *A*, plate 20, shows the change in vertical flow through the confining unit overlying the upper Chesapeake aquifer (layer 9). Most onshore areas remained areas of downward flow. Some nearshore areas changed from upward to downward flow. Large offshore areas of the aquifer were unchanged from prepumping conditions and continued to have upward vertical flow. In Virginia and North Carolina, flow changed to a downward direction in areas that coincide with major streams. Areas most affected by withdrawals had changes in vertical flow of as much as 2 in/yr. These large changes in vertical flow may reflect withdrawals from underlying confined aquifers; an example is the large area in North Carolina that coincides with the cone of depression in the Castle Hayne aquifer. Pumping from the upper Chesapeake aquifer (layer 9) also caused a change in the vertical leakage distribution. Changes in vertical leakage of more than 0.4 in/yr were computed for the coastal communities of Rehoboth Beach, Del., and Ocean City, Md., which withdraw from this aquifer. Prior to pumping, flow through the confining unit was upward, but by 1980, pumping had reversed this flow.

One point should be noted: In certain offshore areas, where the direction of the flow seems to have been changed by pumping, the flow directions may not be accurate. The calculated flow rate in these areas is small (less than 0.1 in/yr), and the change in flow direction may be attributable to boundary conditions or numerical errors.

Figure *C*, plate 20, shows the changes in vertical leakage for the confining unit overlying the Castle Hayne-Piney Point aquifer (layer 7). The vertical flow direction has been reversed from upward to downward over a large area of the modeled confining unit. However, in 1980 flow was still upward over a large area of coastal North Carolina and New Jersey. The change in vertical flow was greatest in North Carolina (0.2 to 6 in/yr) over a large regional cone of depression. Small areas of relatively large changes in vertical flow (0.2 to 0.4 in/yr) occurred in areas of less extensive cones of depression in Delaware, Maryland, Virginia, and elsewhere in North Carolina.

The change in vertical leakage through the confining unit overlying the upper Potomac and Magothy aquifers (layer 3) is shown in figure *A*, plate 22. Significant changes in vertical leakage occurred throughout the confining unit. Large areas that under prepumping conditions were areas of upward leakage were, in 1980, areas of downward leakage. The only significant areas where vertical leakage remained upward were in the coastal area of North Carolina, in central Delaware, and off the southern coast of eastern Long Island. The only significant area where leakage reversed from downward to upward was in and offshore of the east-central New Jersey Coastal Plain. Withdrawals from the overlying Englishtown aquifer (layer 4) caused this upward leakage.

Throughout most of the aquifer system, changes in vertical leakage were less than 0.2 in/yr. Changes in vertical leakage of 6 in/yr occurred on western Long Island, and changes of 15 in/yr occurred along the Delaware River in southern New Jersey. Other relatively large changes in vertical leakage occurred near Raritan Bay in New Jersey (greater than 4 in/yr) and near Annapolis, Md. (greater than 0.8 in/yr). These relatively large increases in vertical leakage generally occurred along the Fall Line. In these areas, pumping was concentrated near the outcrop, where the confining unit is thin and leakage from overlying aquifers occurs readily. On Long Island, the Magothy aquifer (layers 2, 3, and 4) is directly overlain by the upper glacial aquifer (layers 5 and 6). Withdrawals from the Magothy aquifer caused an increase in downward leakage from the upper glacial aquifer. In Virginia, changes in vertical leakage (greater than 0.2 in/yr) occurred both along the Fall Line and farther downdip. Withdrawals near Franklin, Va., caused an increase in downward flow both in updip areas where confining units are thin and in a large area of the aquifer coincident with the deepest part of the cone. This suggests that vertical leakage over a large area of Virginia was providing water to the pumping wells near Franklin.

In summary, the largest increase in vertical leakage has been to pumping centers in areas where confining units are thin or absent. These areas were defined by large changes in vertical leakage generally greater than 1 in/yr. Smaller changes in vertical leakage, generally less than 1 in/yr, occurred where there are large regional cones in aquifers overlain by tight confining units. These cones tend to cover hundreds of square miles.

In addition, large areas offshore that once were areas of upward flow have been changed to areas of downward flow. Water that prior to pumping discharged to overlying aquifers and ultimately to the ocean was in 1980 being diverted toward wells. Changes in flow in offshore areas generally were small, less than 0.1 in/yr. Therefore, changes in direction of small flows may have been induced artificially by boundary conditions, and they should be interpreted carefully.

Sources of ground-water withdrawals

To define the effects of ground-water withdrawals on the aquifer system, it was necessary to examine not only head changes, but also the source of the pumped water. By examining areal ground-water budgets, it was possible to determine the consequences of withdrawals at various locations throughout the aquifer system. The purpose of computing areal water budgets in this study was to determine the source of withdrawals for several large regional cones of depression and to answer some fundamental questions, including the following: Is the water released from aquifer storage an important source of water? Is lateral flow of water from the seaward side significant to the cone of depression in Franklin, Va.? Is vertical leakage an important source of water in the Englishtown aquifer (layer 4) in New Jersey? Are withdrawals from the Castle Hayne aquifer (layer 7) in North

TABLE 6.—*Sources of ground-water withdrawals for selected budget areas, 1980 flow conditions*

[NW., northwest; NE., northeast; SW., southwest; SE., southeast. Mgal/d, million gallons per day; - indicates outflow; no sign indicates inflow]

		Percentage of withdrawals from—										
			Lateral flow					Vertical flow				
			Through budget area sides					From				
Budget area[1]	Model layer[2]	Aquifer storage[3]	NW.	NE.	SW.	SE.	Total	Above	Below	Total	Withdrawals (Mgal/d)	Error[4]
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	.3	39.6	2.3	9.8	.9	52.6	68.2	–20.9	47.3	39.6	.2
3	2	1.3	65.8	1.8	–1.1	2.2	68.7	39.5	–7.5	32.0	16.5	2.0
4	3	1.2	13.6	–2.9	–.3	–12.9	–2.5	118.1	–15.3	102.8	4.31	1.5
5	3	.7	34.2	2.5	1.3	1.5	39.5	84.1	–24.0	60.1	27.1	.3
6	4	1.5	16.5	5.9	–.1	.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	.7	30.8	–2.6	.2	29.1	57.7	11.4	69.1	3.90	1.8
9	7	.0	9.2	–.3	11.7	.4	21.0	76.5	2.5	79.0	88.0	.0
10	8	.7	34.7	–12.2	4.7	3.0	30.2	70.5	–1.0	69.5	20.0	.4

[1]See figure *B*, plate 2, for location of budget area.

[2]Regional aquifer names for the model layers are as follows: 10, Surficial; 9, Upper Chesapeake; 8, Lower Chesapeake; 7, Castle Hayne-Piney Point; 6, Beaufort-Aquia; 5, Peedee-Severn; 4, Black Creek-Matawan; 3, Upper Potomac and Magothy; 2, Middle Potomac; 1, Lower Potomac.

[3]Average rate of change in storage from 1978 to 1980.

[4]The sum of aquifer storage, lateral flow, and vertical flow does not always equal 100 percent in every instance because of numerical and rounding errors.

Carolina derived primarily from an increase in ground-water flow from the outcrop area of the aquifer?

Ground-water budgets were computed for 10 selected areas of the aquifer system where large regional cones of depression were present in 1980. Locations of these areas are shown in figure *B*, plate 2. The selected budget areas were in the lower Chesapeake aquifer (layer 8) in southern New Jersey, the Castle Hayne-Piney Point aquifer (layer 7) in central Delaware and central North Carolina, the Beaufort-Aquia aquifer (layer 6) in southern Maryland, the Black Creek-Matawan aquifer (layer 4) in central New Jersey, the upper Potomac and Magothy aquifers (layer 3) in southern New Jersey and southern Maryland, the middle Potomac aquifer (layer 2) in southern New Jersey and near Franklin, Va., and the lower Potomac aquifer (layer 1) near Franklin.

Sources of the ground-water withdrawals based on simulation of 1980 flow conditions are shown in table 6. The table indicates that (1) the percentage of pumped water coming from aquifer storage was negligible (less than 4 percent), (2) lateral flow was a major source of pumped water in some areas, and (3) vertical leakage was the major source of pumped water in most areas.

In area 1 in southeastern Virginia, lateral flow provided 22.7 percent and vertical flow provided 77.2 percent of the 10.7 Mgal/d withdrawn from the lower Potomac aquifer (layer 1). Table 6 also shows that 17.2 percent was flow from the northwest, that is, from the aquifer outcrop, whereas only 5.5 percent came from the northeast, that is, from the seaward part of the aquifer. Also, all vertical flow to the aquifer was from the overlying aquifer.

In the overlying aquifer (area 2), 52.6 and 47.3 percent of the 39.6 Mgal/d withdrawn from the middle Potomac aquifer (layer 2) came from lateral and vertical flow, respectively. As in area 1, the majority (more than 40 percent) of the flow supplying withdrawals came from the updip direction (northwest and southwest sides of area) and less (less than 4 percent) came from the seaward part of the aquifer (northeast and southeast sides of area). Although 68.2 percent of the water flowed into layer 2 from overlying aquifers, 20.9 percent of the water flowed out of the aquifer to the underlying aquifer. Therefore, the net vertical leakage contributed to the pumpage was only 47.3 percent of the withdrawals.

In area 3 in New Jersey, 68.7 percent of the 16.5 Mgal/d pumped in 1980 was derived from lateral flow primarily from the northwestern side of the area along the Delaware River. Thirty-two percent was derived from vertical flow from above.

In area 4 in Maryland, for the upper Potomac and Magothy aquifers (layer 3), the principal source of the 4.31 Mgal/d withdrawn was vertical leakage from above. Flow equivalent to 118.1 percent of the withdrawals entered the aquifer from above, and 15.3 percent flowed to deeper aquifers. In contrast to other budget areas, lateral flow was not a significant source of the pumped water. Because the aquifer does not outcrop in the budget area, vertical leakage was the principal source of the pumped water.

The budget for area 5 in New Jersey for the upper Potomac and Magothy aquifers (layer 3) is similar to the budget for area 3. In budget area 5, 60.1 percent and 39.5 percent of the 27.1 Mgal/d withdrawn were supplied by vertical and lateral flow, respectively. As with area 3, most lateral flow was from the northwest along the Delaware River.

Area 6 is for the Black Creek-Matawan aquifer (layer 4) in New Jersey. Of the 9.17 Mgal/d withdrawn in 1980, 76.8 percent and 23.1 percent was derived from vertical and lateral flow, respectively. It is noteworthy that flow equivalent to 117.8 percent of the withdrawals came from overlying aquifers and flow equivalent to 41 percent of the withdrawals moved from the aquifer to the underlying aquifers.

Area 7 is in southern Maryland in the Beaufort-Aquia aquifer (layer 6). Like many other areas, vertical flow supplied the majority (80.5 percent) of the 3.20 Mgal/d withdrawn. Also, most lateral flow was from the updip, or westerly, direction. The seaward side of the budget area was a discharge area.

Area 8, primarily in Delaware, is for the Castle Hayne-Piney Point aquifer (layer 7). The principal source of the 3.90 Mgal/d withdrawn was vertical flow (69.1 percent), primarily from the overlying aquifer. Unlike most other budget areas, which showed significant lateral flow from the northwest, this budget area had most lateral flow from the northeast. Because the aquifer pinches out both updip (northwest) and downdip toward the seaward side of the model, lateral flow must come through the northeastern or southwestern sides of the budget area. The budget showed that water was leaving New Jersey and flowing beneath Delaware Bay toward the regional cone of depression in central Delaware.

Area 9, in North Carolina, is for the Castle Hayne-Piney Point aquifer (layer 7). The principal source of the 88.0 Mgal/d withdrawn was flow from overlying aquifers (76.5 percent). Only 21.0 percent of the withdrawal was from lateral flow, almost all from the updip direction.

Area 10 is for the lower Chesapeake aquifer (layer 8) in New Jersey. Of the 20.0 Mgal/d withdrawn in 1980, 69.5 percent and 30.2 percent was supplied by vertical and lateral flow, respectively. The vertical flow component was entirely from the overlying aquifer, and the lateral flow component was principally from the northwest, where the aquifer crops out.

In general, the water budgets showed that if the withdrawals are located downdip from outcrops, vertical flow is the dominant source of water. In almost all pumped aquifers, vertical flow was mostly from overlying aquifers and not from underlying aquifers. In most areas, there still was vertical flow from the pumped aquifer to the underlying aquifers. If withdrawals are located near outcrops (areas 2, 3, and 5), lateral flow is an important source of the withdrawn water. In almost all cases, lateral flow was principally from the updip side of the cones. In general, the seaward side of the cones was a discharge area or supplied less than 6 percent of the water withdrawn.

MODEL RELIABILITY

The reliability of model results depends on the accuracy of the model data used to describe the hydraulic characteristics, the distribution of data used for calibration, and the degree to which the model design represents the physical system. Because model reliability is not easily quantified, the following sections include a descriptive discussion of the model sensitivity and the possible sources of error in estimating the hydraulic characteristics. The distribution of data used in calibration and its effects on the accuracy of model calibration were discussed earlier in the sections on prepumping and transient potentiometric surfaces. A discussion of the sensitivity simulations evaluates the model's sensitivity to both hydraulic characteristics and model design. The effect of model design on model reliability is also discussed in terms of the scale dependence of the results.

EVALUATION OF MODEL INPUT DATA

A major source of error in estimating regional hydraulic characteristics from point data may have been the distribution of control points, which may not represent the regional hydrogeologic system. For example, most wells were drilled in the most productive parts of an aquifer; therefore, point data represented this bias. Furthermore, because of the sparsity of data, a statistical analysis of the data was not done.

Another significant source of error was the vertical hydraulic conductivity values for confining units, which were, in part, estimated from hydraulic analysis of core samples. In many instances, these values were assumed to represent the conductivity of the total confining-unit thickness, which in most cases was many times greater than the core length and may have included different types of sediments. Also, many core analyses were done on disturbed cores, and therefore the vertical hydraulic conductivity values estimated on the basis of analysis of core samples may not represent in situ values.

Another source of error was the use of different analytical methods, with different accuracies, to define the hydraulic characteristics of aquifers. For example, transmissivity was estimated from aquifer-test data, specific-capacity data, and lithologies and sand thicknesses derived from interpretation of geophysical and lithologic logs. Therefore, estimated hydraulic characteristics vary in quality depending on available data and the analytical methods. For example, transmissivities for New Jersey were based on specific-capacity data, whereas transmissivities for North Carolina were estimated mostly on the basis of lithologic-description and sand-thickness data.

Several potential sources of error may affect the determination of representative water-table altitudes or stream elevations. Measurement error, which is generally on the order of tenths of a foot, is usually insignificant regionally. The estimation of land-surface altitudes may be a major source of error in the conversion of measured depth to water to head. Although the error from altitude estimates was highly location dependent, errors of several feet probably were common. Another source of error in estimating water-table and stream altitudes involved model discretization. In making these estimates, topographic maps of varying scales were used on the subregional scale. Effective values for the finite-difference block were picked from a subjective analysis of the topography and the elevation of the stream channel. Therefore, the modeled gradient between the water table and stream was dependent on the spatial discretization, resulting in an inherent random error.

The following estimates of the uncertainty in the initial estimates of hydraulic characteristics are subjective. The magnitude of the error or the uncertainty in the initial estimates of hydraulic characteristics was based on hydrologic experience, judgment, and knowledge of the methods used to develop the hydrologic data base. The uncertainty in the initial estimates of transmissivity may range from 50 to 100 percent of the probable values. In areas where transmissivity was estimated from aquifer-test data corrected for sand thickness in aquifers consisting of several sand beds, the range of uncertainty was probably smaller. Storage coefficient, because it was derived from aquifer-test data, had an estimated range of uncertainty similar to that for transmissivity. Confining-unit leakance was the least accurately estimated characteristic. The range of uncertainty associated with leakance may easily be two or three orders of magnitude. Ground-water recharge values were estimated from subregional hydrologic budgets. Although local variability may be great, uncertainty of the regional estimates was believed to be about 10 percent. Withdrawals are a time-dependent variable; therefore, the uncertainty is also time dependent. The greatest uncertainty was for the earlier withdrawals. The uncertainty of withdrawals from individual wells was high. However, the uncertainty of the total withdrawal for all wells for a specific time period was low, because the random errors associated with individual estimates are reduced when calculating total withdrawals. A reasonable estimate of uncertainty for total withdrawals at a specific time was 5 to 10 percent.

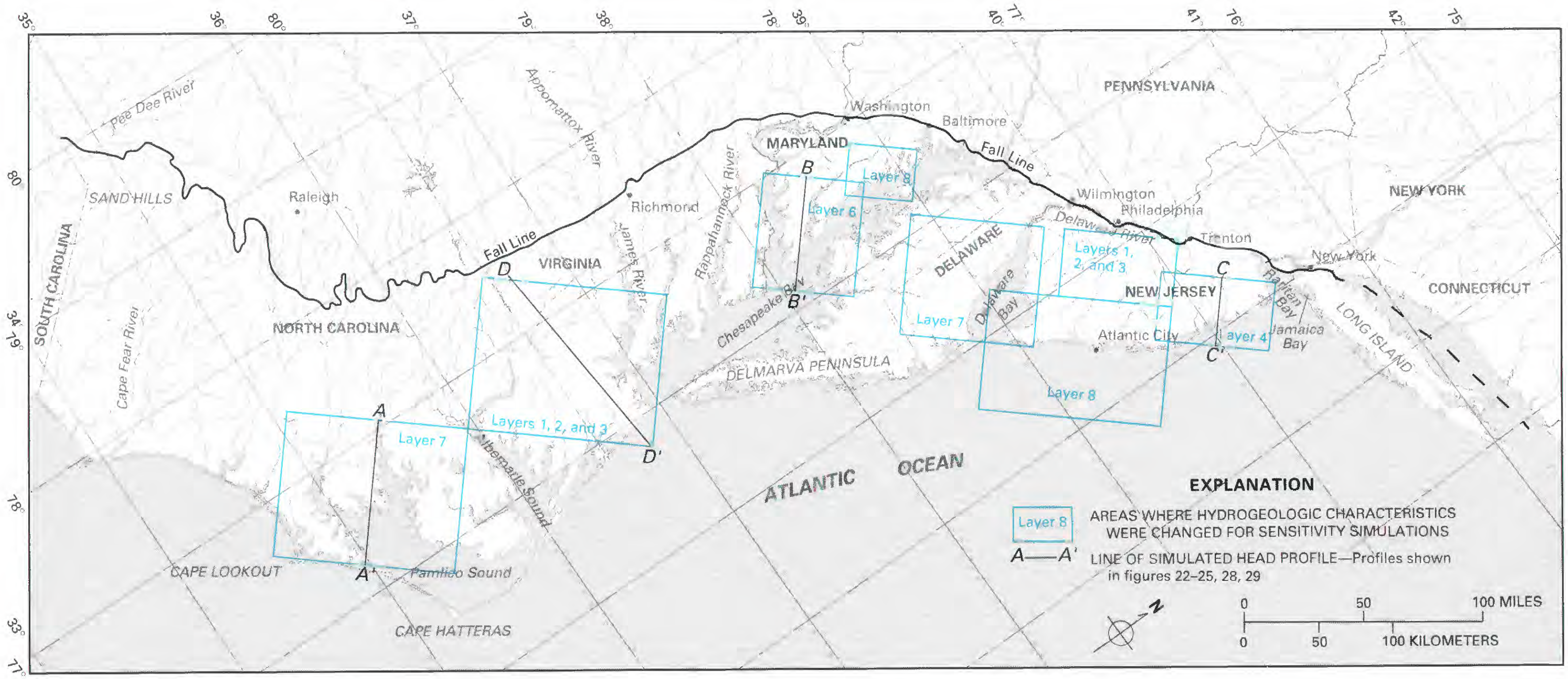
The magnitude of error in estimates of water-table altitudes and stream elevations was expressed in feet, rather than as a percentage. In areas of high topographic relief, such as the deeply incised river valleys in Virginia and North Carolina, the altitude and elevation estimates may be in error as much as 50 ft. In contrast, in areas of low topographic relief, for instance the Delmarva Peninsula, the error associated with the estimates was probably less than 10 ft.

SENSITIVITY

The sensitivity analysis for the regional model consisted of examining the effect of six factors on the ground-water flow system: (1) aquifer transmissivity, (2) aquifer storage coefficient, (3) confining-unit leakance, (4) ground-water withdrawals, (5) transient leakage, and (6) the saltwater-freshwater interface assigned as the boundary of the aquifer system along the seaward side of the model. Factors 1–4 analyzed model sensitivity to the magnitude of selected hydraulic characteristics that were calibrated during model simulation. Factors 5 and 6 analyzed model sensitivity to the effect of hydraulic characteristics that were not included in the calibrated model. The sensitivity analysis identified the factors that were most important in controlling ground-water flow and assessed the reliability of the model by demonstrating the effect of a given range of uncertainty in a hydraulic factor on the simulated heads and flow in the ground-water flow system.

Because of the complexity of simulating 10 aquifer layers, it was impractical to analyze the effects of varying hydraulic characteristics in individual aquifers by different amounts. Rather, selected areas and aquifers assumed most sensitive to changes in hydraulic factors were analyzed. These areas, which coincided with areas of significant regional drawdown, are shown in figure 21.

The area around Atlantic City, N.J., was chosen for the confined Kirkwood aquifer, which is part of the lower Chesapeake aquifer (layer 8), and areas in Delaware and North Carolina were chosen for the Piney Point and Castle Hayne aquifers, respectively, as parts of the Castle Hayne-Piney Point aquifer (layer 7). An area in southern Maryland was chosen for the Aquia aquifer, as part of the Beaufort-Aquia aquifer (layer 6), and an area in New Jersey was chosen for the Englishtown aquifer, as part of the Black Creek-Matawan aquifer (layer 4). Areas selected for the upper Potomac and Magothy aquifers (layer 3) and the middle and lower Potomac aquifers (layers 2 and 1) were centered about Franklin, Va., and in southern New Jersey adjacent to the Delaware River. Hydraulic characteristics in the upper Chesapeake and Peedee-Severn aquifers (layers 9 and 5) were not changed during the sensitivity analysis because major regional cones of depression are not present in these aquifers.



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

FIGURE 21.—Locations of areas where tests of sensitivity to hydrogeologic characteristics were made, and locations of cross sections showing heads from sensitivity simulations.

The sensitivity of simulated heads to a particular hydraulic characteristic was determined by keeping all other characteristics the same as in the calibration simulation and changing a specific characteristic in all selected areas. This procedure resulted in seven sensitivity simulations, as follows: transmissivity at half (1) and twice (2) the calibrated values; (3) aquifer storage coefficient increased by an order of magnitude; confining-unit leakance increased (4) and decreased (5) by an order of magnitude; and ground-water withdrawals decreased (6) and increased (7) by 10 percent for all wells throughout the study area. These ranges in values reflected a subjective estimate of uncertainty for each hydrologic characteristic.

The effect of changing each hydraulic characteristic in the selected sensitivity-test areas was determined by comparing the simulated 1980 potentiometric surfaces from the calibration simulation with surfaces from simulations that included the sensitivity changes. The simulated heads from the sensitivity and calibration simulations for selected cross sections are shown in figures 22 through 25. Hydraulic characteristics were changed for the upper Potomac and Magothy aquifers (layer 3) and the middle and lower Potomac aquifers (layers 2 and 1) in the selected sensitivity-test areas in New Jersey and Virginia (fig. 21). However, only the results for the middle Potomac aquifer (layer 2) are shown (fig. 25) because the results for the other layers are similar. The changes in the simulated 1980 heads are discussed below for each sensitivity simulation.

Transmissivity

Simulated heads were tested for sensitivity to transmissivity values that were half as large and twice as large as the values used in the calibration. With reduced transmissivity, simulated heads in the center of the regional cones in the upper Potomac and Magothy aquifers and the middle and lower Potomac aquifers (layers 3, 2, and 1) in Franklin, Va. (figs. *A*, *B*, and *C*, pl. 17), and in the Englishtown aquifer, part of the Black Creek-Matawan aquifer (layer 4) in New Jersey (fig. *C*, pl. 16), were approximately 100 and 65 ft deeper, respectively, than calibrated heads (figs. 25, 24). Changes in head were similar in other aquifers in the areas selected for the sensitivity analysis. The least significant change was for the Aquia aquifer, part of the Beaufort Aquia aquifer (layer 6) in southern Maryland (fig. 23), and the middle Potomac aquifer in New Jersey, where heads were about 40 ft deeper.

For the simulation in which transmissivity values were twice as large as the calibrated values, heads in the deepest parts of the cones increased the most. An increase in head, and thus a decrease in drawdown, caused the head gradient to become less steep and the cone to become shallower. The maximum increase in head was about 70 ft, in the Englishtown aquifer in New Jersey. Least affected by the increase in transmissivity values, with head increases of about 50 ft or less, were the Castle Hayne aquifer (layer 7) in North Carolina, the upper Potomac and Magothy aquifers (layer 3) in Virginia, and the lower Potomac (layer 1) and the middle Potomac (layer 2) aquifers in New Jersey.

Heads were very sensitive to transmissivity values in areas of large withdrawals and areas having relatively low transmissivity values (generally less than 5,000 ft^2/d). In the lower Potomac aquifer (layer 1) in the Franklin area of Virginia, withdrawals of about 40 Mgal/d were concentrated in a small area and changes in transmissivity values had a major effect on both the shape and the depth of the regional cone. The Piney Point aquifer, part of the Castle Hayne-Piney Point aquifer (layer 7) in Delaware, and the Englishtown aquifer, part of the Black Creek-Matawan aquifer (layer 4) in New Jersey, have relatively low transmissivity values and also have pumping centers located near aquifer-pinchout areas. This external geometry further increases the sensitivity to transmissivity values.

Low sensitivity to transmissivity relates to high transmissivity values and proximity of pumping centers to the unconfined system or to surface-water bodies. The Castle Hayne aquifer, part of the Castle Hayne-Piney Point aquifer (layer 7) in North Carolina, typically has transmissivity values ranging from 10,000 ft^2/d to about 100,000 ft^2/d in the area selected for sensitivity analysis. Because of these high values of transmissivity, a small change in drawdown will cause a significant increase or decrease in lateral flow in the aquifer. Therefore, doubling or halving high transmissivity values has relatively little effect on simulated heads compared with changes in low transmissivity values.

In southern New Jersey, transmissivity values of the Magothy aquifer and the middle and lower Potomac aquifers (layers 3, 2, and 1) were typically less than 10,000 ft^2/d in the sensitivity-test areas. However, pumping centers were located updip near the aquifer outcrops, which are adjacent to or underlie the Delaware River. The river was a major source of recharge to the pumping centers and acted as a specified-head boundary on the system. Because of the proximity of this specified-head boundary to the withdrawal centers, simulated heads near the Delaware River were not as sensitive to changes in transmissivity values as were other selected areas having similar transmissivity values.

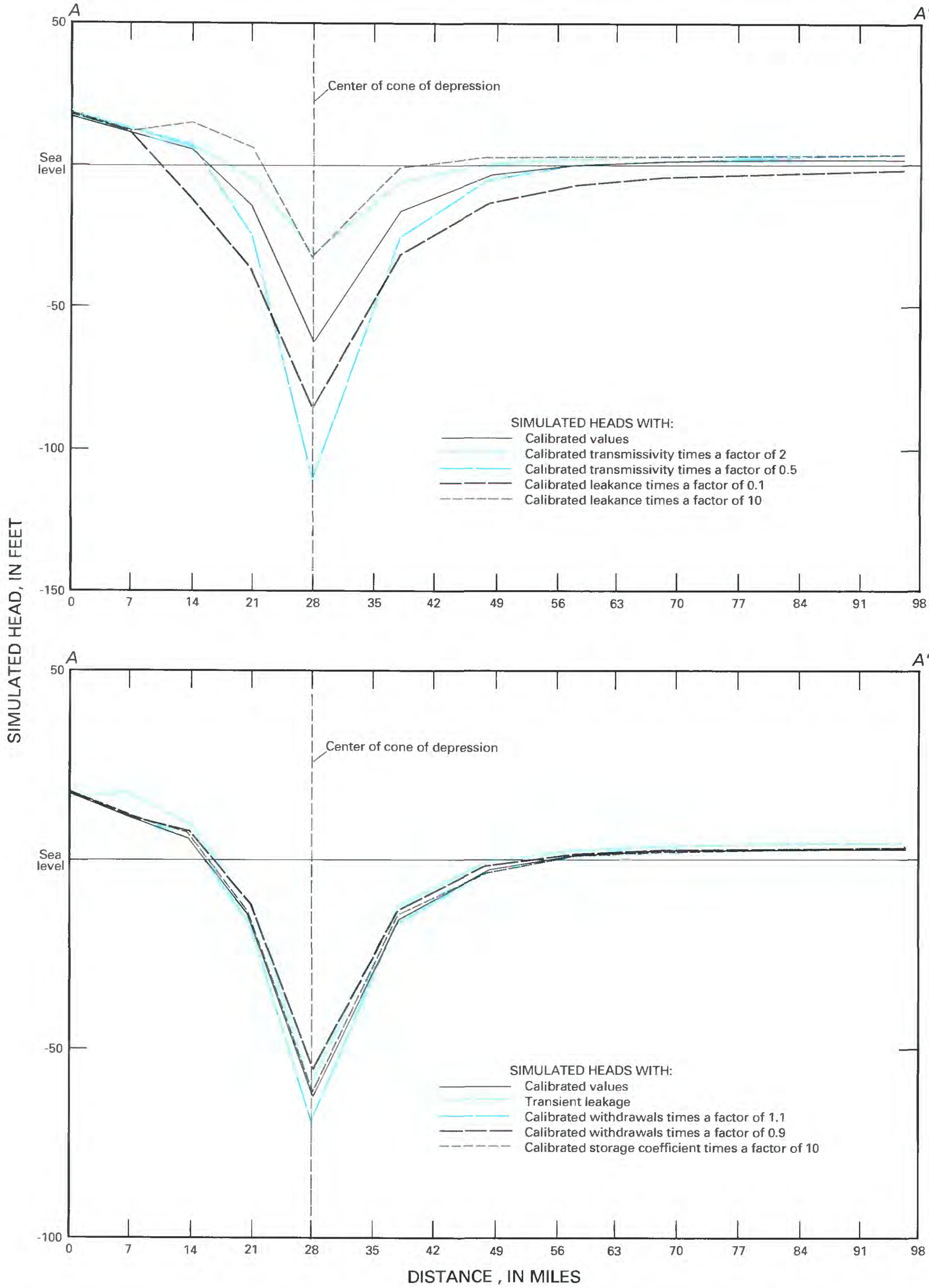


FIGURE 22.—Simulated heads for calibration and sensitivity simulations along line A–A' in the Castle Hayne-Piney Point aquifer (layer 7), 1980 flow conditions. Location of section shown in figure 21.

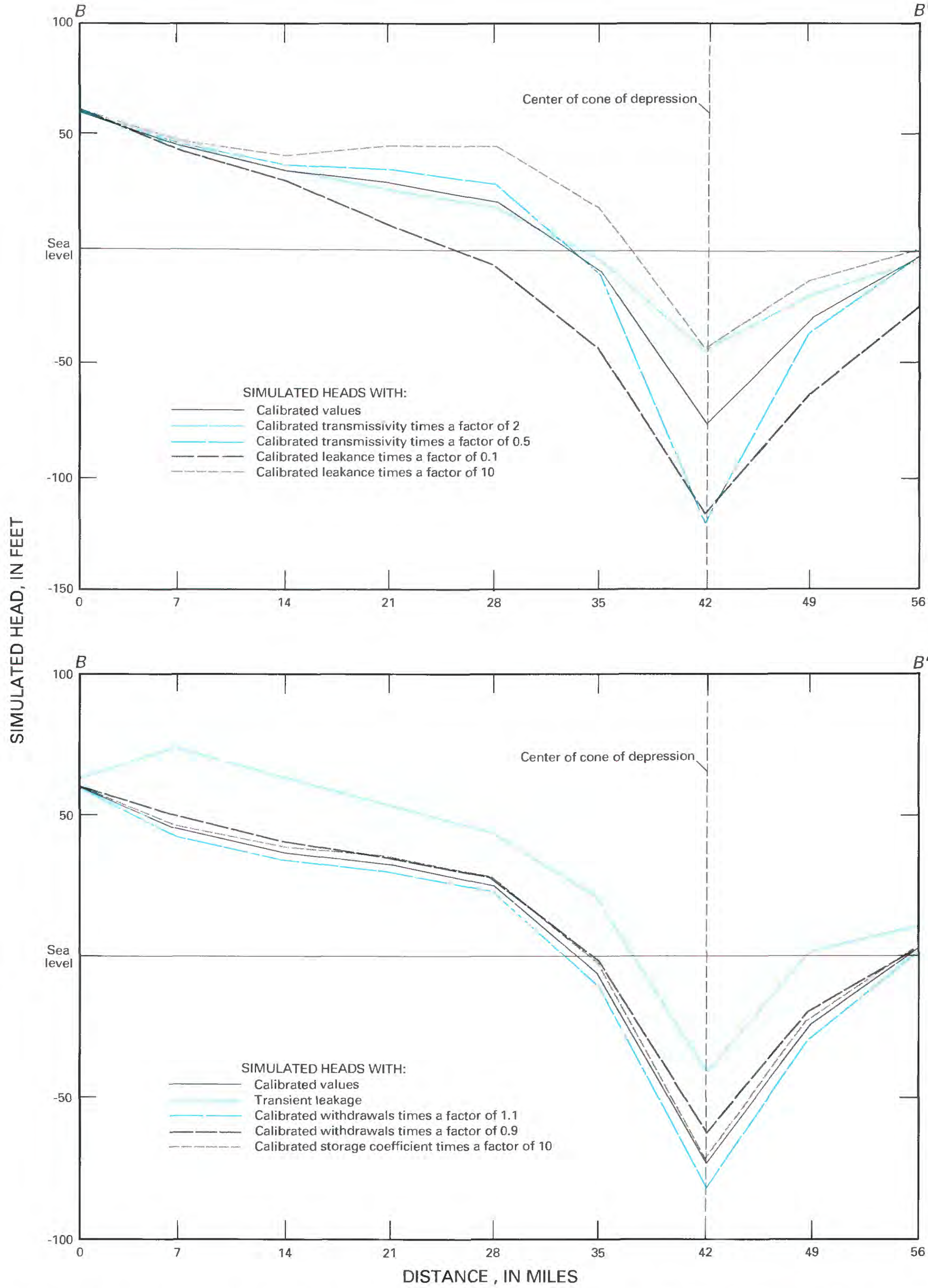


FIGURE 23.—Simulated heads for calibration and sensitivity simulations along line *B–B′* in the Beaufort-Aquia aquifer (layer 6), 1980 flow conditions. Location of section shown in figure 21.

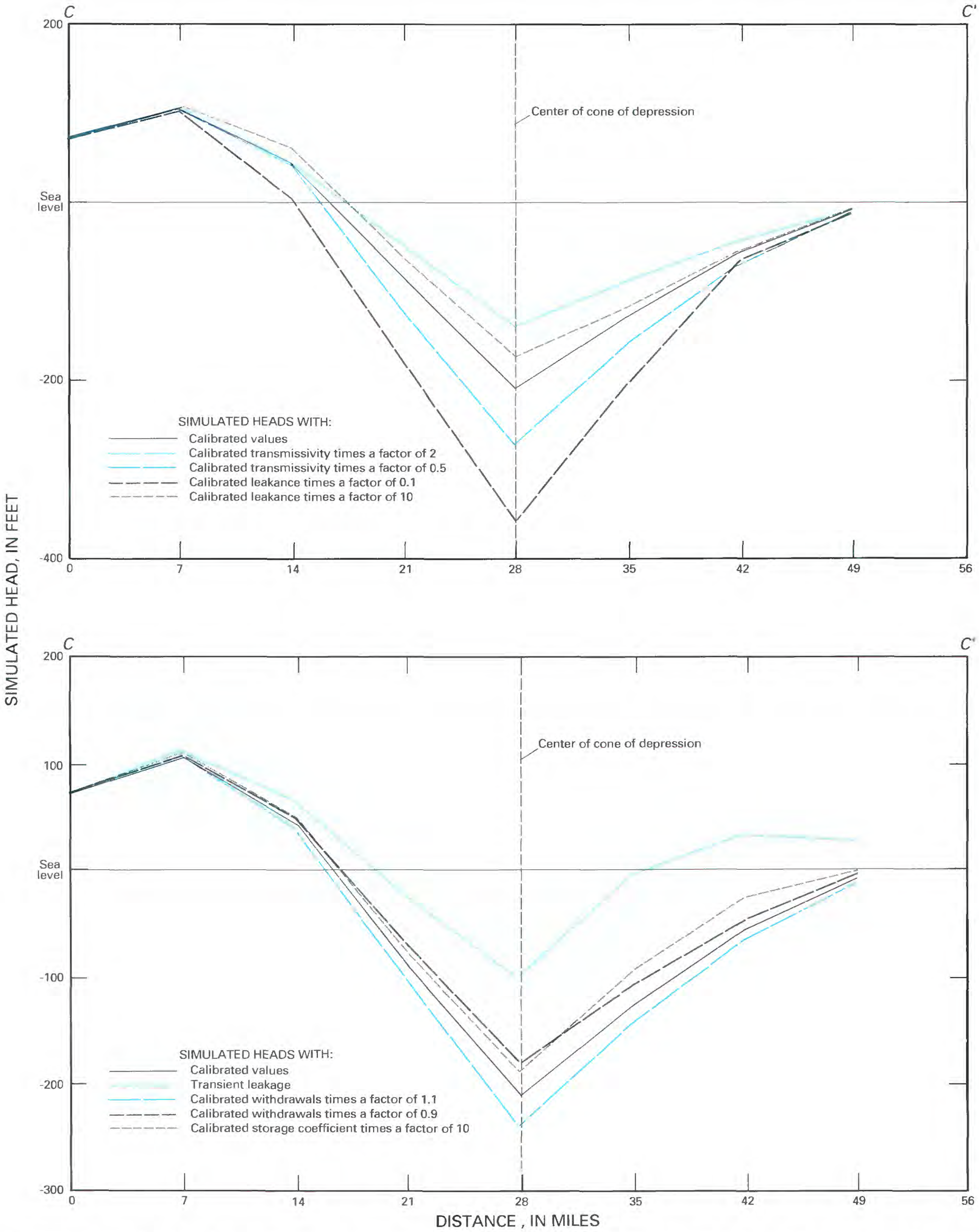


FIGURE 24.—Simulated heads for calibration and sensitivity simulations along line *C–C'* in the Black Creek-Matawan aquifer (layer 4), 1980 flow conditions. Location of section shown in figure 21.

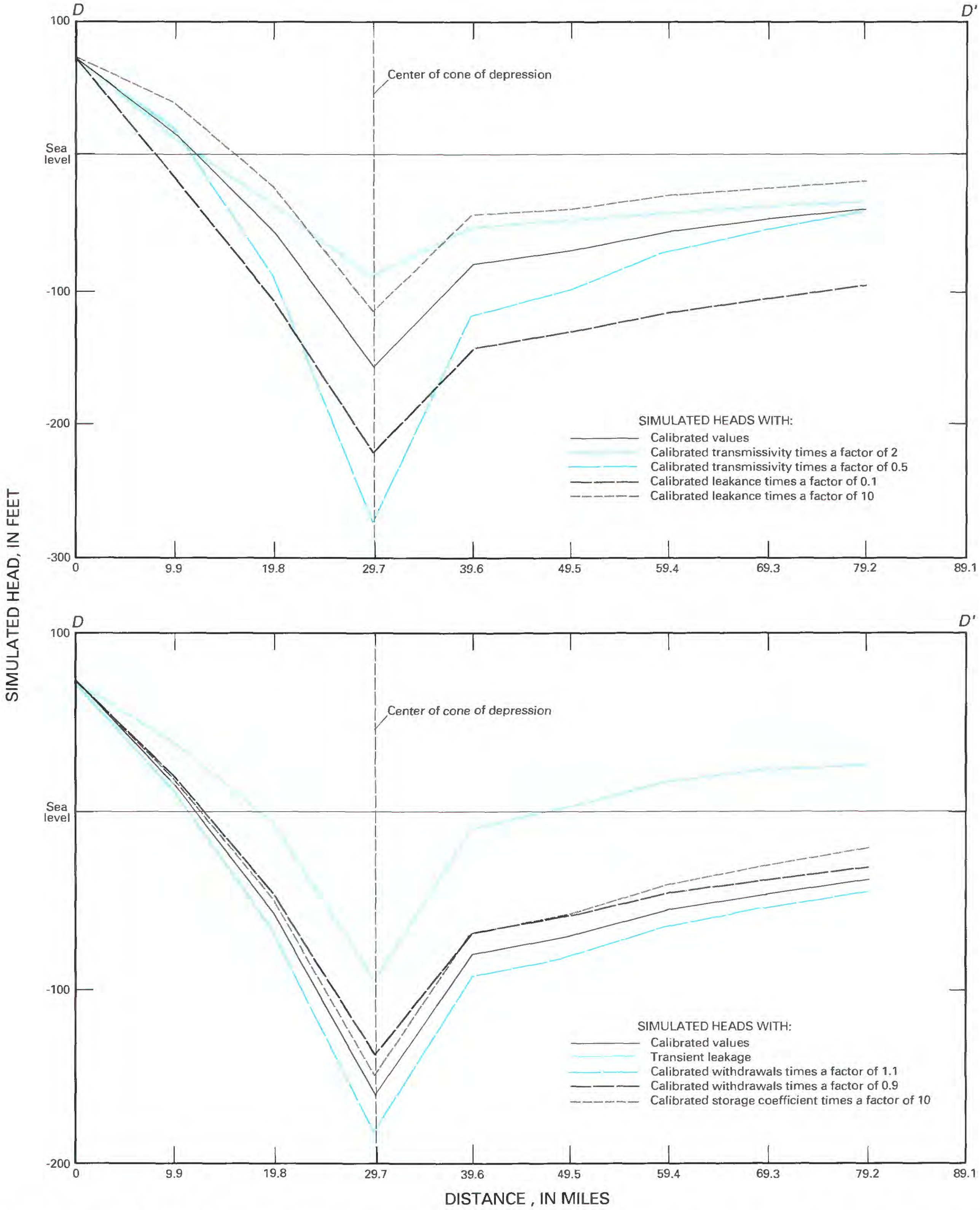


FIGURE 25.—Simulated heads for calibration and sensitivity simulations along line *D–D'* in the middle Potomac aquifer (layer 2), 1980 flow conditions. Location of section shown in figure 21.

Storage Coefficient

The storage coefficient was increased an order of magnitude, to 1.0×10^{-3}, to determine the sensitivity of the simulated heads in selected areas compared with heads simulated with the storage coefficient (1.0×10^{-4}) used in the calibration model for confined aquifers. No simulation was performed in which storage coefficient was decreased, because substantially reduced values were not realistic for the Coastal Plain aquifer system and because minor changes were known, from the calibration process, to have little effect.

Areas showing high sensitivity to the increase in storage coefficient were the Englishtown aquifer (layer 4) in New Jersey (fig. 24), the Piney Point aquifer (layer 7) in Delaware, and the upper Potomac and Magothy aquifers (layer 3) in Virginia, where simulated heads for 1980 were about 25 ft higher than the calibrated values. Areas that showed less than 5 ft of head change were the confined Kirkwood aquifer (layer 8) in New Jersey, the Castle Hayne aquifer (layer 7) in North Carolina, and the Aquia aquifer (layer 6) in Maryland.

The areas that showed the most sensitivity to the increase in aquifer storage coefficient typically had withdrawals that were continuing to increase near the end of the simulation. For example, withdrawals in the Piney Point aquifer (layer 7) in Delaware increased more than 45 percent, from 2.45 to 3.69 Mgal/d, during the period 1965 to 1980. In contrast, areas where withdrawals remained relatively constant during the latter part of the transient simulation period showed little sensitivity to the increase in storage coefficient. For example, withdrawals in North Carolina have remained relatively constant (increasing only 12 percent), at about 100 Mgal/d, since 1966. In these areas, the system has reached a steady-state condition for both values of storage coefficient (calibration value and increased value). In figures 18 and 20, hydrographs for two observation wells in New Jersey and Virginia are shown for the calibration and the increase-in-storage simulation. The hydrograph for the calibration storage coefficient (lower value) reached the steady-state head sooner than the hydrograph for the increased value (higher). However, the duration of the simulation was sufficient for both hydrographs to reach approximately the same head by the end of the simulation (1980).

Another factor affecting sensitivity to the increase in storage coefficient was the location of surface-water bodies and the unconfined system. In the Magothy aquifer (layer 3) in southern New Jersey, increasing the storage coefficient resulted in only a 5-ft increase in the simulated head for 1980 in the updip parts of the aquifer, which are closer to the Delaware River than are the downdip parts. The Delaware River was simulated as a constant-head boundary, and was a major source of water for withdrawals in the updip parts of the aquifer. However, the downdip areas showed a 15-ft increase in simulated 1980 head in response to the increase in storage coefficient.

Confining-Unit Leakance

Two simulations were made to evaluate the sensitivity of the aquifer system to confining-unit leakance. In the simulations, the leakance of the overlying confining unit was changed only in areas selected for sensitivity analysis. In one simulation, an order of magnitude reduction in confining-unit leakance generally lowered simulated heads in the underlying aquifer relative to the simulated heads from the calibrated model. Similarly, in the second simulation, increasing confining-unit leakance an order of magnitude increased heads in the underlying aquifers.

The head declines caused by a reduction in confining-unit leakance ranged from about 10 ft for the confined Kirkwood aquifer (layer 8) in New Jersey to about 150 ft for the Englishtown aquifer (layer 4) in New Jersey (fig. 24). Typically, the shallower aquifers represented by model layers 6 through 8 had simulated declines of about 20 ft, whereas the deeper aquifers (layers 1, 2, and 3) had declines of 50 to 100 ft. The effect of reducing the leakance of the confining units overlying the Englishtown aquifer (layer 4) in New Jersey is noteworthy: Simulated heads in that aquifer were as much as 200 ft lower than heads simulated during calibration in the center of the large regional cone, but were as much as 25 ft higher toward the Delaware River in southern New Jersey. Heads rose near the Delaware River because the leakance of the confining unit underlying the Englishtown aquifer in southern New Jersey was also reduced as part of this sensitivity analysis and because the underlying Magothy aquifer (layer 3) is heavily pumped in the vicinity of the Delaware River. By reducing the leakance and the vertical hydraulic connection between the aquifers, a greater vertical head gradient was needed to provide the same volume of ground-water flow. Therefore, heads in the heavily pumped areas of the aquifers declined, whereas, as the result of many factors, including the location of system boundaries, aquifer hydraulic characteristics, and the properties of underlying confining units, the heads in adjacent aquifers rose. Similarly, the Englishtown aquifer (layer 4) has good hydraulic connection with the overlying Wenonah-Mount Laurel aquifer (layer 5) in New Jersey (Nichols, 1977, p. 2) and was heavily pumped within the area of the regional cone of depression. Reducing the leakance of the confining unit between these two aquifers caused heads in the overlying aquifer to rise more than 50 ft.

Increasing confining-unit leakance by an order of magnitude caused the 1980 simulated heads to be higher than in the calibration simulation in the selected sensitivity areas and aquifers. The simulated 1980 heads in aquifers with the higher leakance were typically 25 to 40 ft higher than the calibrated heads. Head increases in the middle Potomac aquifer (layer 2) in Virginia (fig. 25) were about 40 ft. The lower head increase occurred in the center of the regional cones in most of the sensitivity-test areas. As with the reduction in confining-unit leakance, heads in aquifers overlying the confining unit changed in response to increased leakance. Because the areas selected for sensitivity tests encompass major pumping cones, heads in overlying aquifers generally declined in response to an increase in vertical hydraulic connection between the heavily pumped aquifer and the overlying aquifer.

The model was particularly sensitive to confining-unit leakance for several major reasons. First, the range of values tested was two orders of magnitude, which was considerably larger than the uncertainty associated with transmissivity, aquifer storage coefficient, or withdrawals. Another reason is that leakance affects heads in two aquifers—those underlying and overlying the confining unit. Although changes in the transmissivity values in an aquifer have some effect on adjacent aquifers, the effect was significantly smaller than the effect of leakance changes for the range of values tested in this sensitivity analysis. Also, within the sensitivity-test areas where leakance and transmissivity were changed, vertical flows generally were greater than horizontal flows, and changes in confining-unit leakance (which controls vertical flow) had a greater effect on simulated heads than did changes in transmissivity (which controls horizontal flow).

Ground-water withdrawals

All simulated withdrawals for all 10 pumping periods were reduced and increased by 10 percent. This range of variation was assumed to represent the uncertainty associated with the estimates of ground-water withdrawal rates. Changes in head relative to the 1980 calibrated values were about the same for both simulations (increase or decrease in withdrawals); however, increasing withdrawals decreased heads, while decreasing withdrawals increased heads. Simulated head changes greater than 5 ft occurred only in the areas having large regional cones. In general, simulated pumping cones having 1980 heads more than 100 ft below sea level had head changes in the range of 15 to 25 ft. These deep cones were located in the Peedee-Severn (layer 5) and Black Creek-Matawan aquifers (layer 4) in New Jersey, and in the middle and lower Potomac aquifers (layers 2 and 1) in Virginia and Delaware. In contrast, shallower cones having 1980 heads less than 100 ft below sea level had head changes in the range of 5 to 10 ft.

Transient leakage

The second part of the sensitivity analysis involved hydrologic characteristics not included in the framework of the calibrated model. One aspect was the effect of transient leakage on simulated heads. Transient leakage refers to the physical process by which confining units in the aquifer system, which have finite storage, are able to provide additional water in response to declining heads in the aquifers. The calibration and sensitivity-test simulation discussed earlier had no confining-unit storage and, therefore, no transient leakage.

Transient leakage was simulated by means of modifications to the Trescott (1975) finite-difference flow model code described in Posson and others (1980) and Leahy (1982b). Additional data requirements for the simulation are specific storage, vertical hydraulic conductivity, and confining-unit thickness. An average thickness for each confining unit was used. Although confining-unit thickness is areally variable, a uniform value was assumed to be adequate for this analysis. The ratio of vertical hydraulic conductivity to thickness used in this analysis was the same as the confining-unit leakance used in the calibration simulation. Thus, after confining-unit storage was depleted, the calibration simulation and the transient leakage simulation would reach the same equilibrium conditions and simulate the same potentiometric surfaces. However, by including confining-unit transient leakage, the time required to reach equilibrium would be greater than that of the calibration simulation.

The transient leakage sensitivity simulations were performed using a confining-unit specific storage of 6.0×10^{-4}, 6.0×10^{-5}, and 6.0×10^{-6} per foot of confining unit. The specific storages were the same areally and for all confining units. Because few data were available to define specific storage, a uniform value was assumed to be appropriate. The range selected for the sensitivity analysis represented extremes of the specific storage. The smallest value, 6.0×10^{-6} per foot, was typical of the elastic specific storage of the aquifers. The largest value, 6.0×10^{-4} per foot, was typical of the inelastic specific storage of the deformable confining units. Results of the sensitivity simulation using a specific storage of 6.0×10^{-5} per foot are shown in figures 22–25 because the amount of head change using this value is most similar to the head changes in the other sensitivity simulations.

In the simulation using the smaller specific storage, 6.0×10^{-6} per foot, the 1980 computed heads generally were 10 to 20 ft higher than the calibrated potentiometric surfaces in the areas selected for sensitivity analysis. The area of the regional cone in the Englishtown aquifer (layer 4) in New Jersey was most sensitive to the transient leakage, with 1980 heads 25 to 50 ft higher. The least sensitive aquifers included the lower Chesapeake aquifer (layer 8) and the Castle Hayne-Piney Point

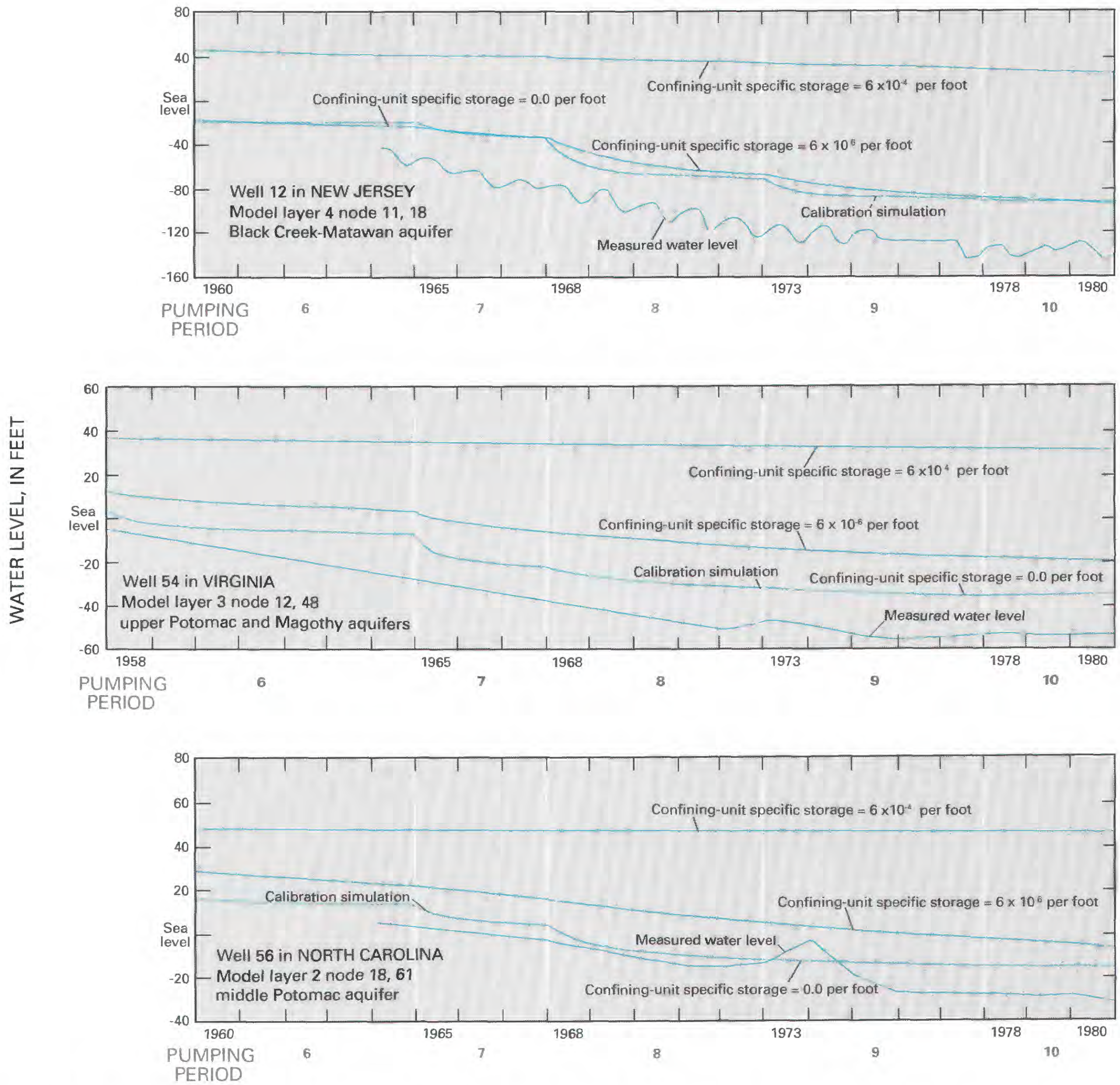


FIGURE 26.—Hydrographs showing sensitivity to confining-unit storage for observation wells in the Black Creek-Matawan aquifer (layer 4) in New Jersey, the upper Potomac and Magothy aquifers (layer 3) in Virginia, and the middle Potomac aquifer (layer 2) in North Carolina. Well locations shown on plate 18*A*.

aquifer (layer 7), where heads were not significantly higher. Hydrographs (fig. 26) for three observation wells in New Jersey, Virginia, and North Carolina show the computed heads for the sensitivity simulations using the highest and lowest specific storage.

The simulations using the larger value of specific storage, 6.0×10^{-4} per foot, showed that deeper aquifers (layers 1, 2, and 3) generally had significantly greater increases in head relative to the calibration results. In the areas most sensitive to transient leakage, heads were 50 to 100 ft higher. In the Black Creek-Matawan aquifer (layer 4), heads in the deep cone of depression in New Jersey were as much as 200 ft higher than the calibrated values. However, in most areas that were less sensitive to transient leakage, heads increased 10 to 25 ft.

In areas where heads were sensitive to the large value (6.0×10^{-4} per foot) of confining-unit specific storage, it was difficult to match the simulated heads with the interpreted or measured heads, unless the values of transmissivity and vertical hydraulic conductivity of the confining units were changed greatly. Many of those changes were out of the range of reasonable estimates.

For these reasons, it can be concluded that the large confining-unit specific storage value (6.0×10^{-4} per foot) is not appropriate for the Coastal Plain aquifer system. Because of the uncertainties in coefficient of storage of aquifers and confining units, and the relative insensitivity of simulation results to the smaller values of specific storage of the confining units, as discussed above, the effects of transient leakage probably are not important on a regional scale. However, they may be significant locally in some coastal areas.

Seaward boundary

The final sensitivity analysis consisted of determining the sensitivity of the model calibration to the seaward no-flow boundary condition. Sensitivity analysis was performed to determine the effect of this no-flow boundary, which represents the limit of ground water having a chloride concentration of less than 10,000 mg/L, on the simulated prepumping and 1980 potentiometric surfaces. It was important to define the sensitivity of the model results to this boundary because the simulated no-flow boundary did not move in response to drawdowns of approximately 50 ft that were simulated under the 1980 pumping conditions along this seaward no-flow boundary in the deeper Potomac aquifers (layers 1, 2, and 3) in Virginia. In some other aquifers, particularly the Black Creek-Matawan and Peedee-Severn aquifers (layers 4 and 5) in New Jersey, Delaware, and Maryland, this seaward no-flow boundary corresponds to the physical limit of the aquifer; therefore, the assumption of a no-flow boundary is justified.

The sensitivity test was made using a variable-density flow model to simulate the aquifer system on the basis of a theoretical development discussed by Weiss (1982) and Kontis and Mandle (1988). Modeling techniques that incorporated the effects of a variable-density fluid were particularly appropriate to define the sensitivity of the model results to the location of a seaward no-flow boundary because the ground water in the northern Atlantic Coastal Plain aquifer system increases in dissolved solids seaward and regionally the dissolved solids do not appear to be moving inland.

The variable-density modeling techniques described in this report have been successfully used to simulate ground-water flow in the Madison and its associated aquifers in Montana, Nebraska, North Dakota, South Dakota, and Wyoming (Downey, 1984, p. G19) and to analyze regional ground-water flow in the northern Great Plains aquifer system (Weiss, 1982, p.2) and in the northern Midwest aquifer system (Kontis and Mandle, 1988).

The variable-density formulation of the flow equation assumes that the density distribution within the system varies spatially but not temporally. Transport of solute within the system was not simulated. However, ground-water flow in a system in which fluid density varies in space was simulated. The assumption of the variable-density formulation being in a steady-state (equilibrium) condition is approximately true if the ground-water velocities in the saline water parts of the system are low. Analysis of the results of the simulation of the regional flow system containing water having variable density in the nonfreshwater part of the model area showed that the computed velocities were very small, typically less than 1 ft/yr. Comparison of the velocities, the grid spacing, and the time simulated showed that the density in an individual grid block does not change with time, thereby justifying the use of the variable-density model for simulation of transient conditions in this study.

The approach used in this sensitivity analysis was to include the saltwater parts of the aquifer system in the model and move the no-flow boundary farther seaward. The seaward no-flow boundary used in this sensitivity simulation corresponded to the farthest downdip extent of any individual aquifer. The simulated aquifer system with this boundary configuration was much larger than the area simulated in the calibration simulation. In the calibration simulation, the farthest seaward extent of any of the aquifers was generally the upper Chesapeake aquifer (layer 9). Therefore, the deeper aquifers had the greatest increase in areal extent because they generally did not extend as far seaward as the upper Chesapeake aquifer.

The technique used to simulate flow in an aquifer system containing water of variable density requires the use of terms describing the gravitational effect of saltwater on the flow system and the conversion of freshwater density and hydraulic properties, such as transmissivity, storage coefficient, and confining-unit leakance, to corresponding variable-density hydraulic properties. The terms describing the effects of gravity on the flow equation were represented in the model as additional sources or sinks on the system. These source and sink terms and the variable-density hydraulic parameters were used directly in constant-density finite-difference flow simulations.

Data requirements for this simulation included the dissolved-solids distribution of water in the aquifers, a depth-temperature profile, depth to top of aquifers, and thickness of confining units. The dissolved-solids distributions of water were developed from chloride-concentration estimates (Meisler, 1981), the aquifer tops and confining-unit thicknesses were adapted from Trapp (in press), and the depth-temperature profile was adapted from data from deep boreholes located in the Continental Shelf (Scholle, 1980). The simulated ground-water flow system was bounded offshore by a no-flow boundary located farther offshore in the Continental Shelf area. Ground water having varying dissolved-solids concentration was simulated in all aquifers to the extended downdip boundary. The dissolved-solids concentrations ranged from 300 to 85,000 mg/L. The higher value was about 2.5 times the concentration generally associated with seawater. The spatial and temporal discretization and the hydraulic characteristics used in the variable-density simulation were identical to or derived directly from those used in the calibration simulation.

Initial estimates were made of the aquifer and confining-unit hydraulic characteristics for the expanded part of the system. In general, the confining units thicken offshore; this causes an accompanying decrease in confining-unit leakance in a seaward direction. Transmissivity in the saltwater parts of the system was initially assumed to be equal to transmissivity at the 10,000-mg/L chloride-concentration line in each aquifer.

As a first approximation, freshwater was assumed to be present throughout the enlarged system, and therefore the whole modeled area was assumed to contain freshwater. This simulation resulted in significant changes from the calibrated flow system. With the additional area included, the large regional cones that were located near the offshore no-flow boundary in the calibration simulation expanded into the enlarged part of the aquifer system. For example, in the middle Potomac aquifer (layer 2), the 1980 simulated heads in Virginia were 10 to 15 ft higher in the enlarged variable-density system in the area of the seaward boundary of the calibrated model. Similarly, in New Jersey, the 1980 simulated variable-density heads were as much as 20 to 25 ft higher than those in the area of the same boundary.

Preliminary results of prepumping simulations with areally varying dissolved-solids concentrations were unsatisfactory using initial estimates of the hydraulic characteristics. Freshwater heads computed for prepumping, steady-state conditions were significantly below sea level in the deeper, saltwater parts of the system. This is not physically possible under prepumping conditions. The results suggested that the prepumping freshwater flow system was not in balance with the dissolved-solids distribution specified in the model. In other words, the results suggested that if a transport simulation had been made using the same hydraulic characteristics, the freshwater flow system would have flushed or transported the water containing higher dissolved solids seaward. Initial attempts to improve the model results focused on adjusting the dissolved-solids distribution. However, these adjustments resulted in a very minor increase in the heads, which were still below sea level. Adjustments in dissolved-solids distribution were small because the data used to estimate the dissolved-solids distributions were assumed to be reliable (H. Meisler, U.S. Geological Survey, written commun., 1981).

Confining-unit leakance and transmissivity were expected to decrease in a seaward direction as a result of an increase in confining-unit thickness, a decrease in aquifer thickness, and a decrease in hydraulic conductivities of both aquifers and confining units because of increases in overburden pressure. Although data for offshore areas are sparse, they suggest a decrease in hydraulic conductivity of the aquifers in a seaward direction (Scholle, 1980, p. 81, fig. 60). Initial estimates of hydraulic characteristics included a seaward decrease in confining-unit leakance and a nondecreasing value of transmissivity in the saltwater part of the system. In later simulations, the initial estimates of transmissivity were also reduced in the extended downdip areas. By reducing the transmissivity, the downdip extent of the freshwater flow would be reduced and the flow system would be more in balance with the specified dissolved-solids distribution. Values of transmissivity in the saltwater areas were reduced to 50 and 25 percent of their initial estimates in two separate simulations.

The simulation that used transmissivity values in the saltwater areas reduced to 25 percent of the initial estimates best matched the observed behavior of the system. Although further reduction in the confining-unit leakance in the saltwater part of the system would also probably have resulted in a similar head distribution, the reduction of transmissivity values was assumed to be the most appropriate approach for this sensitivity analysis.

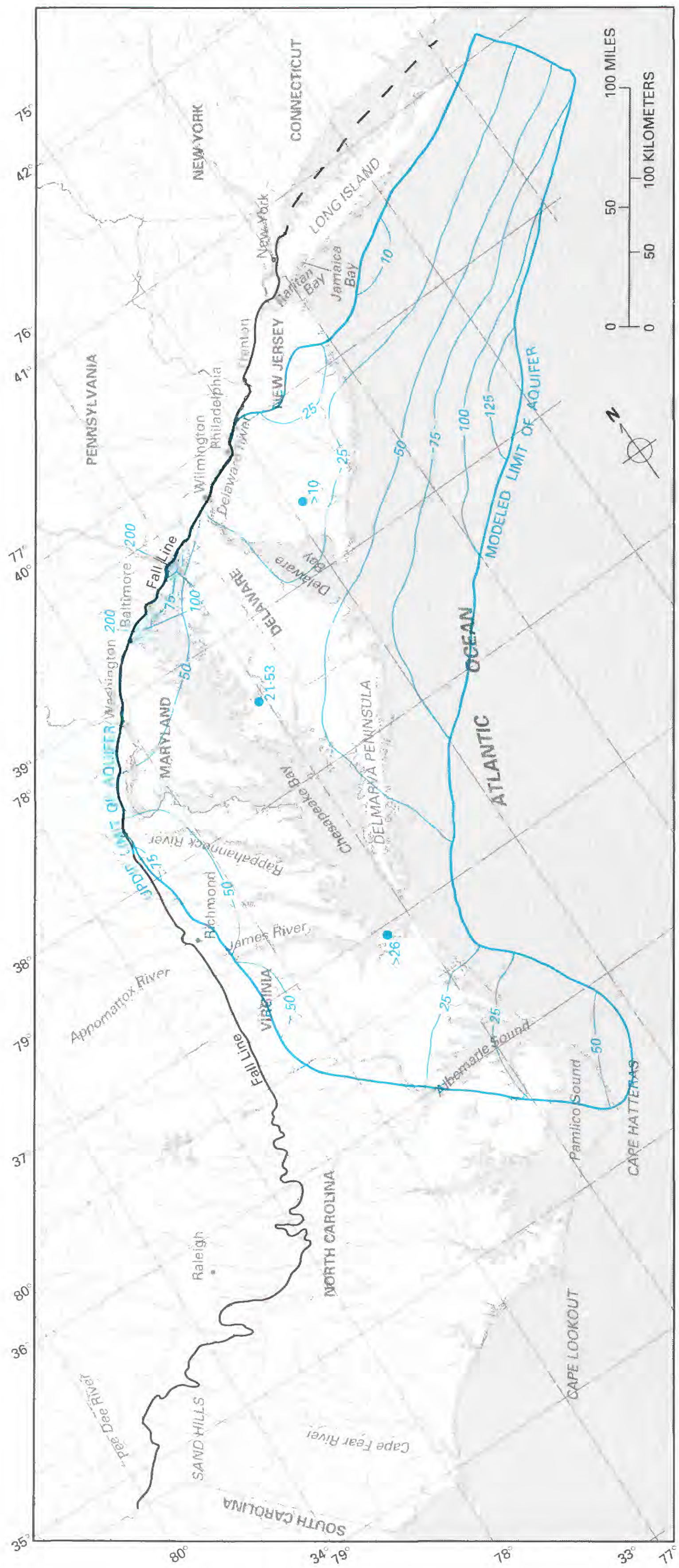
Figure 27 shows the simulated prepumping freshwater potentiometric surface for the lower Potomac aquifer (layer 1) using the variable-density model with transmissivity in the saltwater area reduced to 25 percent of the values used in the calibrated model at the seaward no-flow boundary. Heads in the freshwater areas of the lower Potomac aquifer and the other aquifers in the system agreed closely with the heads in the calibration simulation (fig. *C*, pl. 12), and freshwater-equivalent heads in the saltwater parts of the aquifers were above sea level.

The saltwater sections of the variable-density model were calibrated to three measurements of freshwater-equivalent heads (fig. 27) from wells tapping the lower Potomac aquifer (layer 1) near Norfolk, Va., and Cambridge, Md., and in southern New Jersey. These measurements were made between 1971 and 1981, and although they may be slightly affected by pumping, they represent a minimum estimate of the prepumping freshwater-equivalent head in these areas. The range in head for the Cambridge well in Maryland (Trapp and others, 1984, p. 36) was caused by the rapidly changing density profile in the lower Potomac aquifer. The simu-

EXPLANATION

— 10 — SIMULATED POTENTIOMETRIC CONTOUR—Shows altitude of simulated prepumping potentiometric surface for ground water of freshwater density. Contour interval, in feet, is variable. Datum is sea level

● 21-53 LOCATION OF WELL AND ALTITUDE OF FRESHWATER-EQUIVALENT HEAD—Two numbers indicate range in observed freshwater head at varying depths within aquifer. Number with greater than sign (>) is minimum freshwater head and may be slightly affected by pumping. Altitude of head, in feet above sea level



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

FIGURE 27.—Simulated potentiometric surface for the lower Potomac aquifer (layer 1) using the variable-density model, prepumping conditions.

lated freshwater-equivalent head for this location should, and does, fall within this measured range of values.

Computed Darcy velocities in the saltwater parts of the system were about 0.1 ft/yr for prepumping conditions. The simulated direction of ground-water flow near the seaward boundary may be inaccurate because the variable-density system was terminated at the previously described maximum seaward limit of freshwater. Meisler and others (1985, p. 6, fig. 3) simulated ground-water flow in a cross section through southern New Jersey, and included the natural offshore boundary (the Continental Slope) of the system in their analysis. The variable-density simulation did not incorporate this boundary but did extend the system as much as 60 mi offshore of New Jersey (the Continental Slope is located about 80 mi offshore). The computed direction of ground-water flow may be affected by this simplification. For the purposes of this sensitivity analysis, the limitation caused by the artificial offshore no-flow boundary appears to be insignificant. It is noteworthy, however, that the Darcy velocities computed by the variable-density model were in agreement with those of 0.02 ft/yr computed by the cross-sectional model (Meisler and others, 1985, p.15) in the area having 10,000 mg/L chloride concentrations in the upper Chesapeake aquifer (layer 9).

The variable-density model was also used to simulate the 1980 heads in the aquifer system. The differences between the simulated 1980 potentiometric surfaces resulting from the calibration and the variable-density simulations were generally small (less than 10 ft) along the seaward no-flow boundary used for the calibration model. The maximum head difference between the 1980 calibration and variable-density simulations occurred in the cones of depression located near the seaward no-flow boundary used in the calibration simulation. The maximum difference was about 25 ft in the center of the cones of depression in the Black Creek-Matawan aquifer (layer 4) in New Jersey and the lower Potomac aquifer (layer 1) in Virginia. In general, heads in the variable-density simulation were less than 5 ft higher than those in the 1980 calibration simulation. Major exceptions were in the middle Potomac aquifer (layer 2) in Virginia (15 ft higher), the upper Potomac and Magothy aquifers (layer 3) in Virginia (10 ft higher), and the Peedee-Severn aquifer (layer 5) in New Jersey (20 ft higher).

Figure 28 shows the simulated 1980 head profiles from the model calibration and variable-density simulations through major cones of depression. The location of the profiles is shown of figure 21. Figure 28 shows head profiles in the Castle Hayne-Piney Point aquifer (layer 7) in North Carolina and the Beaufort-Aquia aquifer (layer 6) in Maryland. In both the profiles, heads were less than 10 ft higher in the variable-density simulation than those in the calibration simulation.

Figure 29 shows head profiles in the Black Creek-Matawan aquifer (layer 4) in New Jersey and in the middle Potomac aquifer (layer 2) in Virginia. These profiles show that heads in the center of cones of depression were as much as 25 ft higher in the variable-density simulation than those in the calibration simulation. These differences decrease toward the seaward boundary, where the head difference was less than 10 ft.

Analysis suggests that the sensitivity of model results to the seaward no-flow boundary condition was minimal. However, heads in some regional cones were as much as 25 ft higher when the density was specified in the model. The cones most affected were located in deeper aquifers near the freshwater-saltwater interface. Simulated heads along the interface were generally within 10 ft of the calibrated heads. These small head differences and the assumed freshwater-saltwater interface are appropriate and probably do not significantly affect the model results or the model calibration.

SCALE DEPENDENCE

One of the principal aspects of modeling aquifer systems on two distinctly different scales was the integrating effect of the larger grid spacing. Martin and Leahy (1983) discussed the impact of areal discretization scale on model results. The methodology of interfacing regional and subregional models was discussed by Martin (1987). The larger size mesh uses areally averaged values for hydraulic characteristics, and thus the larger regional model blocks average local variations in potentiometric surface and in recharge or discharge.

Comparison of heads and flows simulated at regional and subregional scales by models having equivalent hydraulic parameters has demonstrated the effect of discretization scale on model results. Figure 30 compares simulations of the 1980 pumping conditions for the Englishtown aquifer (layer 4) from the regional model and the New Jersey subregional model. The potentiometric surface simulated by the regional model showed the same general configuration as the surface simulated by the subregional model. The regionally simulated surface lacks the resolution or detail of the subregionally simulated surface, because heads were averaged over a larger cell area in the regional model. Thus, local flow features tend to be lost in the regional model. An increase in model resolution provided by the finer mesh of the subregional models provided a more accurate calibration than did the regional model. In general, the deepest parts of the large regional cones (heads more than 100 ft below sea level) tended to be simulated

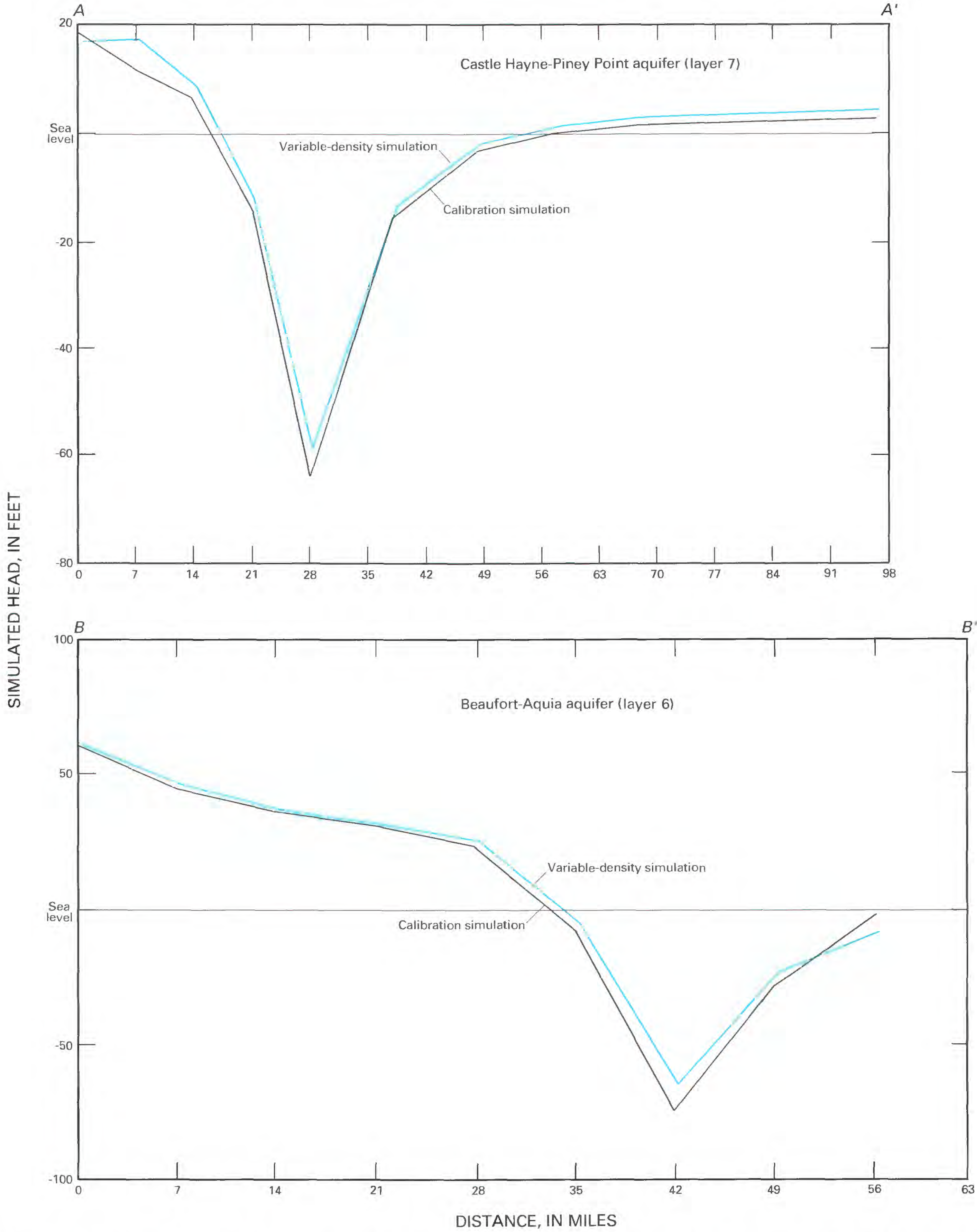


FIGURE 28.—Simulated heads along line of section *A–A'* in the Castle Hayne-Piney Point aquifer (layer 7) and along line of section *B–B'* in the Beaufort-Aquia aquifer (layer 6) using the variable-density and calibration simulations, 1980 flow conditions. Location of sections shown in figure 21.

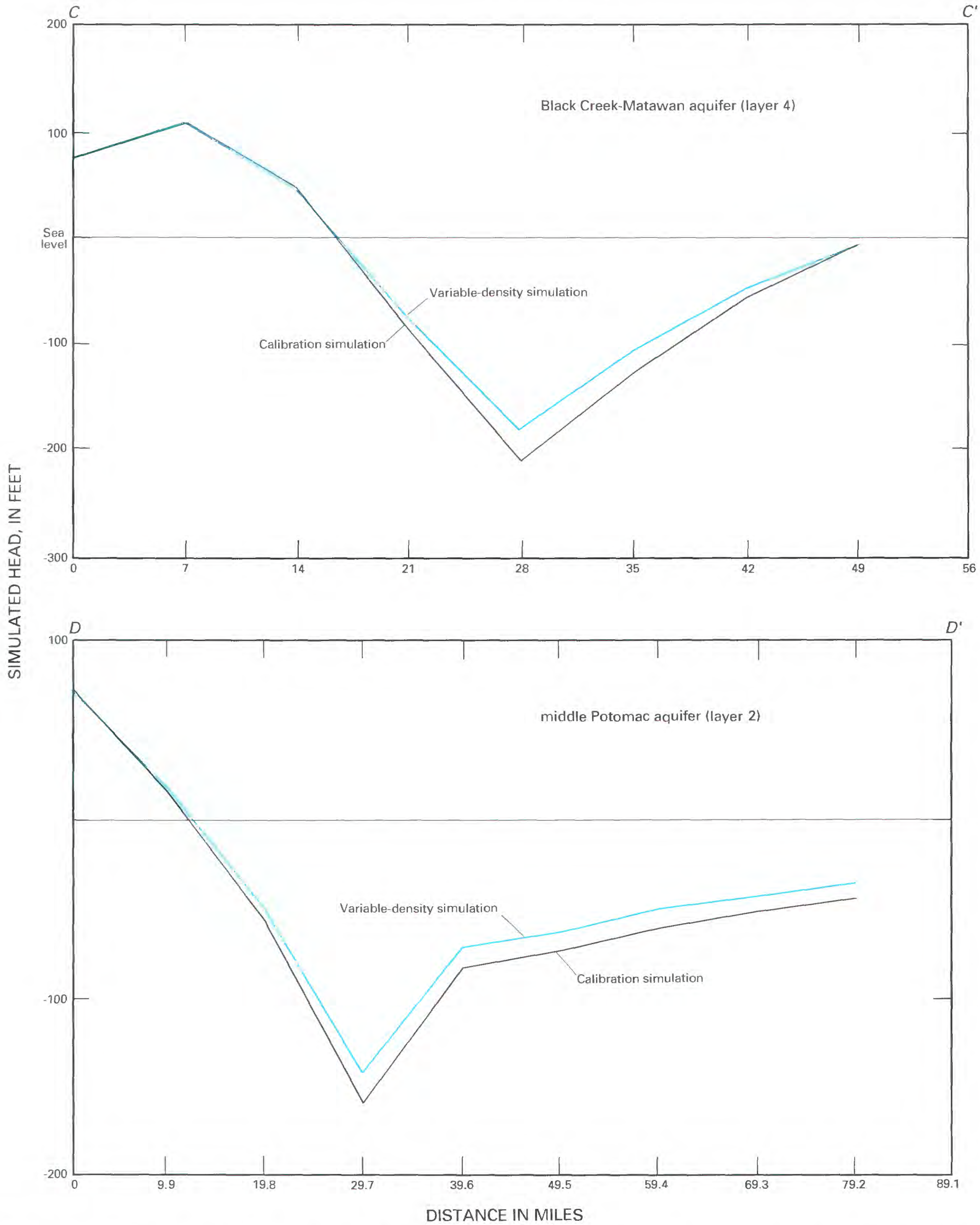


FIGURE 29.—Simulated heads along line of section *C–C′* in the Black Creek-Matawan aquifer (layer 4) and along line of section *D–D′* in the middle Potomac aquifer (layer 2) using variable-density and calibration simulations, 1980 flow conditions. Location of sections shown in figure 21.

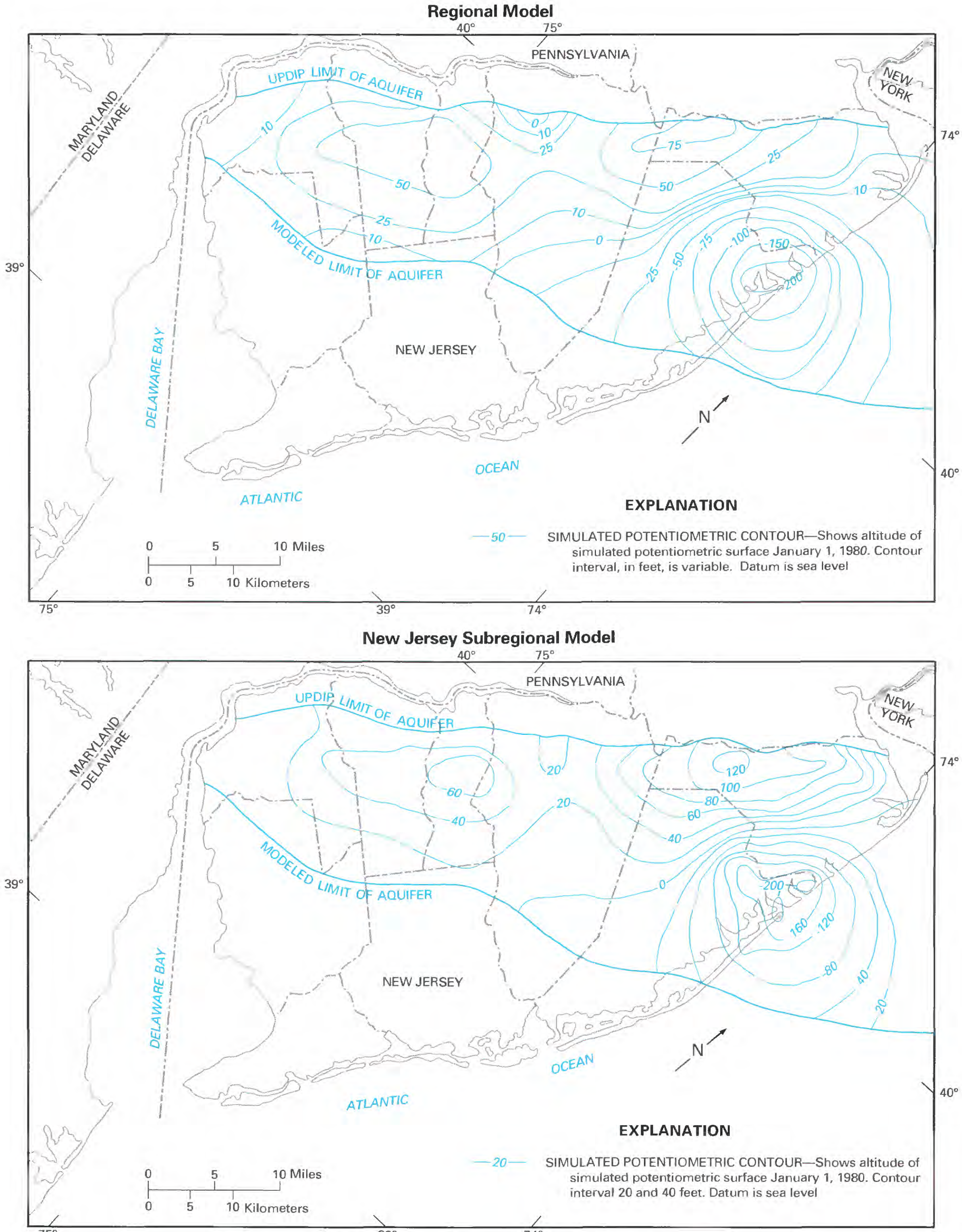


FIGURE 30.—Simulated potentiometric surfaces for the Black Creek-Matawan aquifer (layer 4) using the regional model and the New Jersey subregional model, 1980 flow conditions.

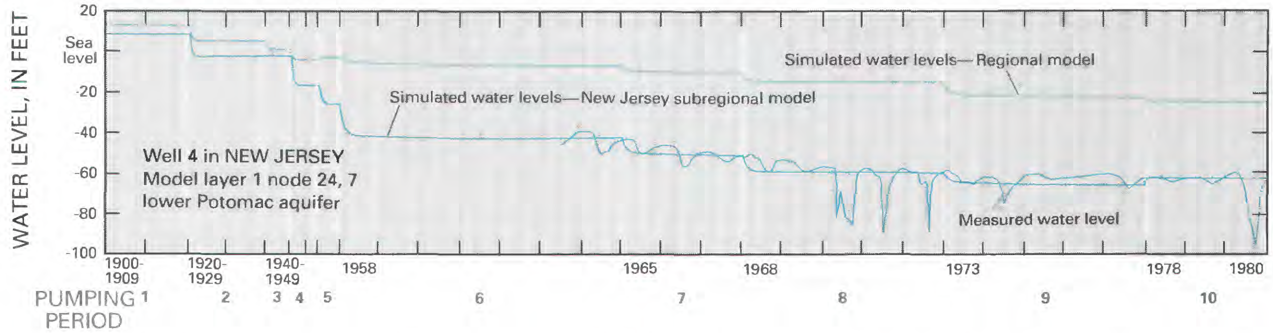


FIGURE 31.—Comparison of hydrographs derived from simulations using the regional model and the New Jersey subregional model. Well location shown on plate 18*A*.

shallower (as much as 50 ft shallower) by the regional model than the subregional models.

Discretization scale also affects the distribution and rate of vertical flow. In the subregional model, simulation included recharge and discharge from the ground-water flow system that occurred within small distances. Because the regional model did not include local flow, the rates of recharge and discharge were generally lower than in the subregional model. For example, regionally simulated recharge to the confined aquifers was 0 to 2 in/yr (fig. *A*, pl. 13), whereas in the New Jersey subregional model, simulated recharge to the confined aquifers was 0 to 10 in/yr. The regional model simulated regional recharge and discharge trends but did not simulate local flow rates, as did the subregional models. Flow rates for the regional blocks were very similar to the average rates for the equivalent areas in the subregional models. The simulated areal discharge and recharge to the confined system were similar at both model scales.

The effects of discretization scale through time was also investigated. Figure 31 shows a comparison of simulated hydrographs for an observation well tapping the lower Potomac aquifer (layer 1) in southern New Jersey. Initially, both hydrographs showed very similar trends. However, starting in the mid-1940's, the hydrographs diverge and, ultimately, the hydrograph from the regional model was about 30 ft higher than that from the subregional model. Both hydrographs were derived by linear interpolation of heads at those nodes closest to the well to account for the well not being located at the center of the block. The difference in the hydrographs was related to spatial discretization of the system. This observation well was near many major well fields. The finer discretization of the subregional model allowed for a more accurate representation of the areal distribution of withdrawals. In contrast, in the regional model, the withdrawals were located at nodes that were 7 mi apart, because all withdrawals within a grid block were simulated at the node at the center of the block. For the hydrograph as shown, withdrawals in the regional model had to be located farther from the observation well because of coarse discretization, whereas in the New Jersey subregional model, these same withdrawals were located closer to their actual location. Therefore, the effect of withdrawals on heads in the well where the hydrograph was obtained could be simulated more accurately by the New Jersey subregional model than by the regional model.

This comparison of model results shows the importance of discretization scale. Regional modeling of any ground-water system is subject to limitations imposed by the scale of the discretization. As shown, the scale dependence can sometimes cause significant differences in heads and flows. However, a model designed with a large block size, as used in this study, and carefully calibrated is adequate for understanding the ground-water flow system on a regional scale. For better resolution and analysis of local aspects, finer discretization is needed.

SUMMARY

The sediments of the northern Atlantic Coastal Plain range in age from Jurassic to Holocene. The sediments form a wedge type of deposition that thickens from its western edge along the Fall Line to more than 10,000 ft at Cape Hatteras, N.C., and attain a thickness greater than 45,000 ft beneath the Continental Shelf and Slope. The sediments consist of sand, silt, and clay, with some limestone in North Carolina.

Ground-water flow in the northern Atlantic Coastal Plain aquifer system was simulated using a multilayer finite-difference model of the 10 aquifers and 9 intervening confining units. The grid system used to describe the hydraulic properties of aquifers and confining units had 85 rows and 32 columns. Most of the grid blocks had an area of 49 mi^2. The simulated aquifer system was bounded below and laterally, except for one small area, by no-flow boundaries. The no-flow boundary at the bottom of the model corresponded to the sloping contact between Coastal Plain sediments and underlying low-permeability crystalline rocks or other consolidated rocks. The no-flow boundary to the north and west represented the Fall Line.

An assumed stationary saltwater-freshwater interface at the place where ground water has a 10,000-mg/L chloride concentration corresponded to the no-flow boundary at the seaward limit of the model to the east and south. The southwestern no-flow boundaries, for all but the deepest aquifer, coincided with a ground-water divide beneath the Pee Dee River in South Carolina. Because ground water was flowing northeastward from South Carolina to North Carolina in the deepest aquifer, the southwestern boundary was defined as a specified head to represent this inflow. The northeastern boundary of the aquifer system was the submerged limit of the Coastal Plain aquifer system and the estimated 10,000-mg/L chloride concentration northeast of Long Island, N.Y. The top boundary was represented both as a specified-head boundary at water-table altitudes for prepumping conditions and as a specified-recharge boundary with specified-head streams for prepumping and transient conditions.

The prepumping system was simulated with a steady-state solution and initially used an overlying water-table specified-head boundary. In offshore areas, the flow system was bounded at the ocean bottom by overlying specified-head values of equivalent freshwater heads corresponding to sea water depth. Simulation of the prepumping flow system used the overlying specified-head boundary and yielded heads in the freshwater parts of the aquifers and flow to or from the specified-head boundary which represented the water-table aquifer.

The second conceptualization of simulating the flow system was to treat the top boundary in the onshore area as both a flux and a specified-head boundary. This change in the boundary condition was necessary to simulate transient heads in aquifer outcrop areas that were affected by pumping. This conceptualization included areally distributed ground-water recharge (15 to 22.5 in/yr) to the water-table aquifer, and streams as a specified-head drain for part of the recharge. Streambed leakance (hydraulic conductivity divided by streambed thickness) controls the amount of ground-water recharge that discharges to streams. A water budget for each cell was used to compute an effective streambed leakance in each cell for prepumping conditions. Effective streambed leakance was calculated using an average (long-term) water-table altitude and stream elevation for each cell, and therefore represented the combined leakance properties of all streams within a cell.

Yearly withdrawal estimates from 1900 to 1980 were used to calculate the average withdrawals for 10 pumping periods used to simulate the history of ground-water development. Development of ground water began in the northern Atlantic Coastal Plain in the late 1800's. Withdrawals for 1900 were estimated at 100 Mgal/d—most of this amount for the water supply of Brooklyn and Queens, N.Y. Withdrawals have steadily increased, and total withdrawals in 1980 were estimated to be 1,210 Mgal/d. These estimates do not include domestic and irrigation withdrawals from shallow aquifers, because their effect was considered negligible on a regional scale. Water use of this type is returned to the aquifer with a negligible amount of consumption.

Calibration of the model consisted of trial-and-error adjustment of hydraulic parameters using four smaller subregional models. Lateral boundary fluxes for these subregional models were calculated on the basis of Darcy's Law using heads from regional simulations. Aquifer transmissivity and confining-unit leakance were changed in the subregional models until steady-state and transient-model behavior compared favorably with the measured or interpreted behavior of the aquifer system. Similar parameter adjustments were made in the regional model to ensure compatibility of boundary fluxes.

Steady-state calibration of the model was achieved by adjusting hydraulic parameters and then comparing simulated heads with measured or estimated prepumping heads in the confined aquifers, and flow from overlying constant-head cells with estimated flow to confined aquifers. Calibration of the model for transient conditions was achieved by further adjustment of hydraulic parameters until the computed response of the model for the 1980 pumping conditions approximated the measured heads or the interpreted potentiometric surfaces. In addition, 74 hydrographs were used to evaluate calibration of the regional model through time. Parameter adjustments made for calibration were used to resimulate prepumping conditions to obtain the initial conditions for the transient simulation. This procedure ensured calibration compatibility between prepumping and transient conditions. Heads in the simulated hydrographs generally were within 15 ft of the measured heads. The simulated hydrographs tended to be higher than the measured heads because of the 49-mi^2 block size used in the regional model. The subregional grid blocks, typically a quarter of the size of the regional cells, provided greater model resolution and a better match of the simulated and measured hydrographs.

Areal distributions of transmissivity for the aquifers and leakance for the confining units were refined through model calibration from initial estimates based on hydrologic data. An areally constant value of 1.0×10^{-4} was used for the aquifer storage coefficient except in areas of New Jersey, where storage values of up to 8.0×10^{-4} were used in the Potomac-Raritan-Magothy aquifers (layers 1, 2, and 3). Transmissivity was highest, more than 100,000 ft^2/d, in parts of the Castle Hayne aquifer (layer 7) in North Carolina. Other aquifers that had local transmissivities greater than 20,000 ft^2/d were the upper Chesapeake aquifer (layer 9) on the Delmarva Peninsula, and the upper glacial aquifer (layers 6 and 7) and the middle part of the Magothy aquifer (layer 4) on Long Island. Typically, transmissivities ranged from 500 to about 10,000 ft^2/d for most aquifers. The Beaufort-Aquia (layer 6), Peedee-Severn (layer 5), and Black Creek-Matawan (layer 4) aquifers had transmissivities exceeding 2,000 ft^2/d only in parts of North Carolina.

Simulated leakance of the confining units ranged from about 1.0×10^{-8} to 1 (ft/d)/ft. Lower values were typically found in the deeper units. For example, leakance of the confining unit overlying the lower Potomac aquifer (layer 1), the deepest confining unit in the system, ranged from about 1.0×10^{-8} to about 1.0×10^{-4} (ft/d)/ft, whereas leakance of the confining unit overlying the upper Chesapeake aquifer (layer 9), the uppermost confining unit in the system, ranged from about 1.0×10^{-4} to about 1 (ft/d)/ft. The range of leakance values within a single unit was commonly several orders of magnitude. Updip leakance values tended to increase with decreasing thickness and increasing vertical hydraulic conductivities of the confining units. Extremely high leakance values were present where adjacent aquifers were in direct hydraulic connection. This situation was found throughout the Coastal Plain and was especially common in the upper and lower Kirkwood-Cohansey aquifer (layers 9 and 8) in areas of New Jersey and in the upper, middle, and lower parts of Magothy aquifer (layers 5, 4, and 3) on Long Island. Leakance values tended to decrease in a downdip direction. The tightest (lowest leakance) regional confining units were those overlying the Beaufort-Aquia (layer 6), upper Potomac and Magothy (layer 3), and lower Potomac (layer 1) aquifers.

Sensitivity analysis was performed on hydraulic parameters and model assumptions to evaluate the reliability of the model calibration. The hydraulic parameters tested included transmissivity, storage coefficient, ground-water withdrawals, and confining-unit leakance. Individual parameters were changed on the basis of subjective estimates of uncertainty. In this process, transmissivity was doubled and halved, storage coefficient was increased by an order of magnitude, confining-unit leakance was increased and decreased an order of magnitude, and withdrawals were increased and decreased by 10 percent. Model assumptions that were tested included the importance of (1) transient leakage from confining units and (2) the location and nature of the assumed seaward boundary. Results of the analyses showed variable sensitivities to each of the hydraulic parameters and assumptions.

Sensitivity of the simulated 1980 heads to changes in transmissivity varied by aquifer and location. Simulated heads for 1980 using transmissivity values twice the calibrated values caused heads in the regional pumping cones in confined aquifers to be about 40 ft higher than heads from the calibrated model. The maximum difference was about 100 ft in the Englishtown aquifer (layer 4) in New Jersey. Heads in areas of high transmissivity (greater than 20,000 ft^2/d)—for example, the Castle Hayne aquifer (layer 7) in North Carolina—were least sensitive to changes in transmissivity.

The sensitivity of simulated heads for 1980 pumping conditions to changes in the values of storage coefficient varied by aquifer and by location within the aquifer. The aquifer most sensitive to changes in storage coefficient was the Englishtown aquifer (layer 4) in New Jersey, where an order of magnitude increase in storage caused simulated heads to be as much as 50 ft higher than heads in the calibrated model. In contrast, shallower aquifers (layers 6, 7, 8, and 9) generally showed a less than 5-ft change in head for the same order of magnitude increase of the storage coefficient. Areas of high sensitivity to aquifer storage coefficient typically had withdrawals that increased through time.

Increasing and decreasing withdrawals by 10 percent in each pumping period lowered and raised simulated heads equally relative to the calibrated heads for 1980. Generally, deep pumping cones where 1980 simulated heads were more than 100 ft below sea level had head changes of about 15 to 25 ft. For shallower cones where 1980 heads were less than 100 ft below sea level, head changes were typically 5 to 10 ft.

Simulated heads for 1980 were most sensitive to changes in confining-unit leakance, which was varied by an order of magnitude. Heads in the Englishtown aquifer (layer 4) in New Jersey were most sensitive to a reduction of leakance and were about 200 ft lower than calibrated values. The confined Kirkwood aquifer (layer 8) in New Jersey was the least sensitive to changes in leakance. Increasing the leakance by an order of magnitude caused simulated heads to be typically 25 to 60 ft higher than the 1980 calibrated heads in regional pumping cones.

The simulated heads were most sensitive to changes in confining-unit leakance because the range of the values tested was two orders of magnitude. This tested range was considerably larger than the range tested for transmissivity, aquifer storage, or withdrawals.

In the calibrated model, the effects of transient leakage from confining-unit storage were assumed to be negligible. However, the sensitivity of the model was tested in three transient simulations in which values of confining-unit specific storage were assumed to be 6×10^{-4}, 6×10^{-5}, and 6×10^{-6} per foot. These simulations resulted in 1980 heads that were higher than the calibrated values. Results using the lowest value of specific storage of the confining unit indicated that transient leakage could be significant in some parts of the Coastal Plain under particular pumping conditions, but in most areas the effect was small. A maximum head increase of 50 ft was simulated in the Black Creek-Matawan aquifer (layer 4). In contrast, the highest value of the specific storage (6×10^{-4} per foot) produced head increases ranging from about 10 ft in the Castle Hayne aquifer (layer 7) in North Carolina to more than 200 ft in the Englishtown aquifer (layer 4) in New Jersey. The sensitivity analysis showed the potential importance of transient leakage in the aquifer system. The specific storage of the confining units was poorly known, and only limited data on compaction were available from which specific storage could be calculated. However, limited data suggest that only minor compaction has occurred during the period of record, and the specific storage of the confining units was probably low (10^{-6} to 10^{-5} per foot). Additional interpretation and data are needed to define the importance of transient leakage in the aquifer system.

A model capable of simulating flow in a multilayer ground-water system containing a fluid with density that varies areally (Weiss, 1982) was used to evaluate the sensitivity of heads to the location of the assumed seaward no-flow boundary located where ground water had a 10,000-mg/L chloride concentration. The variable-density model simulated flow in the system beyond the area where ground water had chloride concentrations of more than 10,000 mg/L (beyond the seaward boundary in the original model). To simulate a reasonable steady-state condition for the system in terms of heads and flows, transmissivity was varied areally in the area beyond the original seaward limit of the calibrated model. The transmissivity value was reduced to 25 percent of the calibrated values used along the 10,000-mg/L-chloride-concentration line. Vertical leakance of the confining units was also reduced to reflect the increasing thickness of the units in the offshore areas. These decreased values were justified by the known progressive thickening of confining units and decreasing aquifer transmissivity in the seaward direction. These relatively low values for hydraulic characteristics restricted flow in the saltwater part of the system. Sparse data on prepumping heads in the saltwater parts of the system agreed with the simulated prepumping heads using the reduced values of the hydraulic parameters.

The variable-density model was then used to simulate 1980 conditions. At the 10,000-mg/L-chloride-concentration line, the head difference between the calibrated and variable-density models generally was less than 10 ft. This minimal difference suggested that the use of the 10,000-mg/L-chloride-concentration line as the seaward no-flow boundary in the calibrated model produced no serious errors. Furthermore, because hydraulic characteristics were relatively low on the seaward side of the 10,000-mg/L-chloride-concentration line, only a small percentage of water pumped from the system flows from areas containing saltwater. These parts of the system have extremely slow lateral Darcy velocities (less than 0.01 ft/yr) even under simulated pumping conditions.

Definition of the flow system was accomplished through examination of the following results derived from the calibrated model: (1) regional water budget, (2) potentiometric surfaces, (3) vertical leakage between aquifers, and (4) lateral flow directions and velocities in the aquifers. Average areal ground-water recharge to the Coastal Plain was estimated to be about 40,000 Mgal/d, or 15.4 in/yr. The majority of this recharge discharges to the nearest surface-water body. Model results indicate that under prepumping conditions, about 592 Mgal/d, or 0.5 in/yr, of ground water recharged the underlying confined system. This recharge occurred over approximately 25,000 mi^2, or 26 percent of the total area. Discharge from the deeper aquifers occurred over the remaining 74 percent of the study area and accounted for only about 2 percent of the total recharge to the surficial aquifer. Interpretation of the simulated prepumping potentiometric surfaces indicates that (1) recharge to the outcrops of regionally confined aquifers occurred in areas of downward hydraulic gradient, generally along or near the Fall Line, and (2) discharge from the deeper confined aquifers occurred by upward leakage through confining units into the ocean or coastal estuaries and bays.

Simulated prepumping flow to and from the overlying surficial (water-table) aquifer ranged from more than 20 in/yr of water recharging the underlying confined aquifers to more than 16 in/yr of water discharging from the confined aquifers. The higher rates of recharge to and discharge from the confined aquifers were in areas where the surficial aquifer was in direct hydraulic contact with the underlying aquifer (no intervening confining unit). Examples of this relation occur between the upper and lower parts of the upper glacial aquifer (layer 6 and 7) and the upper, middle, and lower parts of the Magothy (layers 5, 4, and 3) aquifer on Long Island, between the upper Kirkwood-Cohansey (layer 9) and lower Kirkwood-Cohansey (layer 8) aquifers in New Jersey, and between the surficial (layer 10) and Castle Hayne (layer 7) aquifers in North Carolina.

The simulated prepumping potentiometric surfaces of the shallower aquifers showed relatively local flow systems. The influence of major rivers, estuaries, and embayments on the flow system was apparent. In contrast, the potentiometric surfaces of the deeper aquifers showed more regional flow patterns. Although the streams and rivers affected the flow pattern in the updip areas near outcrops, the flow pattern generally was not influenced by the presence of overlying surface-water bodies throughout most of the extent of the deeper aquifers. Notable exceptions included Raritan, Delaware, and Chesapeake Bays. These large surface-water bodies affected ground-water flow patterns in all aquifers and had significant influence on the location of the 10,000-mg/L-chloride-concentration line. In the deeper aquifers, ground water may have flowed a distance of several tens of miles. Low lateral hydraulic gradients and estimated values of hydraulic conductivity showed that the Darcy ground-water flow velocities were slow (less than 1 ft/yr) along these regional flow lines.

By 1980, withdrawals had caused heads in several aquifers to decline regionally to more than 100 ft below sea level in areas of North Carolina, Virginia, Delaware, and New Jersey and generally to more than 50 ft below sea level in areas of Maryland. Total withdrawals from the system in 1980 were estimated to be about 1,210 Mgal/d, or about 3 percent of the estimated ground-water recharge to the surficial aquifer. However, 1980 simulated withdrawals were about twice the simulated recharge to the confined system prior to development and caused recharge to the confined aquifers to increase to 1,330 Mgal/d in 1980.

Simulation results suggest that in 1980, the pumped water was provided from the following sources: (1) 61 percent from an increase in downward percolation to the confined system derived from a decrease in discharge to streams; (2) 37.2 percent from a decrease in discharge from the confined system through overlying confining units to large surface-water bodies; (3) 1.7 percent from a reduction in aquifer storage; and (4) 0.1 percent from an increase in inflow from South Carolina in the middle Potomac aquifer (layer 2). In 1980, the area of recharge to the confined system from the unconfined system was approximately 45 percent of the study area. This was about a 19 percent increase from the prepumping conditions.

Results of the simulated history of withdrawals from 1900 to 1980 showed that heads approach steady-state conditions within a 5-year pumping period. Therefore, water was released from aquifer storage with an accompanying head decline for only a short period of time after each increase in withdrawals. Continued increases in withdrawals through time caused additional water to be released from aquifer storage and additional head decline as the system approached a new equilibrium. As a result, in many areas, simulated heads, like actual heads, have steadily declined. However, despite continuing head declines, less than 2 percent of the water pumped in 1980 was supplied by a reduction in aquifer storage.

Results of the regional simulation indicate that most of the pumped water has come from reduction in ground-water discharge to streams in updip areas. In some areas, head declines have propagated updip to outcrop areas and caused some of the recharge that once discharged to streams to be diverted toward pumping centers. For example, on Long Island and in New Jersey, more than 10 in/yr of localized discharge to streams was diverted to pumping centers in shallow aquifers. This trend was evident to a lesser extent along the updip outcrop areas throughout the study area.

Water that formerly discharged upward through overlying confining units in coastal areas was diverted to sustain withdrawals. The amount of this diversion was small compared with the amount that has come from a reduction in discharge to streams from the shallow aquifers. However, locally reduced discharge from deep confined aquifers in coastal areas was shown to be an important source of water to wells in areas having large withdrawals, such as in the upper, middle, and lower Potomac aquifers (layers 3, 2, and 1) in Virginia.

Ground-water withdrawals in the northern Atlantic Coastal Plain aquifer system have caused the development of large regional cones of depression that cover hundreds of square miles in some aquifers. The size and shape of the pumping cones depend on the quantity of water being pumped, the location of the pumping center relative to aquifer outcrop, and the hydraulic characteristics of the aquifers and the confining units. Large cones have developed in downdip areas of the regional Castle Hayne-Piney Point aquifer (layer 7) in Maryland and Delaware, the subregional Wenonah-Mount Laurel (layer 5) and Englishtown (layer 4) aquifers in New Jersey, the Black Creek aquifer (layer 4) in North Carolina, and the Potomac aquifers (layers 3, 2, and 1) in Virginia. Hydraulic gradients, coupled with confining-unit leakance, suggest that vertical leakage from adjacent aquifers was an important source of water to these cones. Large regional cones have developed in updip areas near outcrops, such as those in the Castle Hayne-Piney Point aquifer (layer 7) in North Carolina, the Patapsco and Patuxent aquifers, part of the middle and lower Potomac aquifers (layers 2 and 1) in Delaware, and the upper, middle, and lower Potomac-Raritan-Magothy aquifers (layers 3, 2, and 1) in New Jersey. Hydraulic gradients indicate that in these areas, both vertical leakage from adjacent aquifers and lateral flow from the outcrop areas were important sources of water.

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March 3, 1849