GROUND-WATER FLOW IN THE NEW JERSEY COASTAL PLAIN

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Ground-Water Flow in the New Jersey Coastal Plain

By Mary Martin

REGIONAL AQUIFER-SYSTEM ANALYSIS—
NORTHERN ATLANTIC COASTAL PLAIN

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1404–H

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

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 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 beginning with Professional Paper 1400.

Charles G. Groat
Director

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply English unit	*by*	*to obtain metric unit*
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
foot per day per foot ((ft/d)/ft)	1.00	meter per day per meter
gallon per minute (gal/min)	0.06309	liter per second
inch (in.)	2.54	centimeter
inch per year (in/yr)	2.54	centimeter per year
million gallons (Mgal)	3785	cubic meter
million gallons per day (Mgal/d)	3785	cubic meter per day
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

The National Geodetic Vertical Datum of 1929 (NGVD of 1929) is 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."

GROUND-WATER FLOW IN THE NEW JERSEY COASTAL PLAIN

By Mary Martin

ABSTRACT

Ground-water flow was simulated in 10 aquifers and 9 intervening confining units of the New Jersey Coastal Plain, which consists of unconsolidated gravel, sand, silt, and clay of early Cretaceous to Holocene age. A multilayer finite-difference model was used to simulate both prepumping steady-state conditions and transient conditions from 1896 through 1980. Model calibration indicates that the highest transmissivity, greater than 10,000 feet squared per day, is in Camden and Gloucester Counties in the three Potomac-Raritan-Magothy aquifers; Monmouth and Ocean Counties in the middle Potomac-Raritan-Magothy aquifer; in Atlantic and Cape May Counties in the confined Kirkwood aquifer; and in Ocean, Burlington, Atlantic, and Cape May Counties in the lower and upper Kirkwood-Cohansey aquifers. Confining-unit vertical leakance is highest, greater than 1×10^{-3} feet per day, per foot, in updip areas and lowest, less than 1×10^{-5} feet per day, per foot, in downdip areas.

Sensitivity analyses show that the model was useful in refining initial estimates of transmissivity and confining-unit vertical leakance near the major cones of depression and that the assumptions associated with the lateral and downdip boundary conditions do not seriously limit the usefulness of the model results. However, simulated water levels near the major cones of depression in several aquifers are fairly sensitive to parts of the model framework, including confining-unit characteristics along the outcrop of the Potomac-Raritan-Magothy aquifers, the updip limit of the confined Kirkwood aquifer, and the downdip limit of the upper and lower sand units of the Englishtown aquifer system.

Calibration and sensitivity analyses also show that aquifer storage coefficients are about 1×10^{-4}, except in downdip areas of the Potomac-Raritan-Magothy aquifers, where they may be as much as eight times higher. Confining-unit specific storage is about 6×10^{-6} per foot. Areas near the center of the major cones of depression in the Potomac-Raritan-Magothy, Englishtown, Wenonah-Mount Laurel, and Kirkwood-Cohansey aquifers approximated steady-state conditions in 1981. However, downdip and offshore areas are under transient-flow conditions. Simulated changes in water levels along the saltwater-freshwater interface boundary (the occurrence of ground water with greater than 10,000 milligrams per liter chloride concentrations) indicate that the lower Potomac-Raritan-Magothy aquifer and the confined Kirkwood aquifer have the greatest potential for movement of the interface. However, simulated hydraulic gradients within the aquifers near the interface boundary cannot be used to quantify movement of the interface. The simulated sources of water to wells in 1978 included 3 percent from aquifer storage, 3 percent from boundary flows, 4 percent from the ocean and bays, and 90 percent from decreased discharge to or increased recharge from streams; that is, from a reduction in streamflow.

The prepumping regional flow system recharged in upland areas in Mercer, Middlesex, and western Monmouth Counties; in western Ocean and central Burlington Counties; and in central Gloucester and Camden Counties and discharged to the Atlantic Ocean, Delaware River, Delaware Bay, Raritan Bay, and to large rivers in the Coastal Plain. Under pumping conditions, regional cones of depression formed in the three Potomac-Raritan-Magothy aquifers and in the Englishtown aquifer, Wenonah-Mount Laurel aquifer, and the confined Kirkwood aquifer.

INTRODUCTION

The New Jersey Coastal Plain is a major source of ground water for the southern half of New Jersey and is part of the northern Atlantic Coastal Plain aquifer system, which extends from Long Island, N.Y., to the southeast boundary of North Carolina. The New Jersey Coastal Plain covers an area of about 4,200 square miles (mi^2). Coastal Plain sediments are a southeastward-thickening wedge of unconsolidated gravel, sand, silt, and clay of Early Cretaceous to Holocene age. These sediments are more than 6,500 feet thick in southern Cape May County and are underlain primarily by Precambrian bedrock (Zapecza, 1984, p. 6).

Withdrawals from the New Jersey Coastal Plain aquifers were more than 350 million gallons per day (Mgal/d) in 1980 (Vowinkel, 1984, p. 7). Regional cones of depression are present in six major aquifers, and water levels have declined in one aquifer, to an altitude of more than 240 feet below sea level (Walker, 1983). Declines in water levels have increased the potential for movement of saltwater into freshwater aquifers; for movement of contaminants into, within, and between aquifers; and for permanent decreases in stream baseflow. These problems result from local and regional changes in the ground-water flow system. Effective resource management, which minimizes these problems, requires definition of the regional hydrogeologic framework and flow system.

PURPOSE AND SCOPE

The study that generated this report is one of a series that comprises the U.S. Geological Survey's Regional Aquifer Systems Analysis (RASA) program (described in the Foreword). This study is part of the northern Atlantic Coastal Plain RASA whose objectives include a complete description of the hydrogeologic framework, geochemistry of the ground water, and simulation of predevelopment and present ground-water flow conditions by digital models. The purpose of this report is to describe the prepumping and transient flow systems in the New Jersey Coastal Plain aquifer system as part of the New Jersey RASA program. Specifically, this report describes (1) the conceptual hydrogeologic model of the stressed and unstressed flow systems, (2) the methods and approach used in simulating flow, and (3) the results and conclusions of the flow simulations. The simulation of ground-water flow in the New Jersey Coastal Plain was coordinated with RASA modeling studies in the other States within the northern Atlantic Coastal Plain and with a regional RASA modeling study. The regional RASA model simulated ground-water flow in the entire northern Atlantic Coastal Plain by incorporating data from modeling studies in Long Island, N.Y., New Jersey, Delaware, Maryland, Virginia, and North Carolina. As part of the regional RASA, the New Jersey study area is adjacent to the Long Island, N.Y., study area to the northeast and the Delaware-Maryland RASA study area to the southwest. This report describes the methodology and results of the digital simulation of ground-water flow in the New Jersey Coastal Plain.

As part of the New Jersey RASA project, data on aquifer and confining-unit characteristics and on pumpage and water levels from 1918 through 1980 were compiled for the entire New Jersey Coastal Plain. The hydrogeologic framework of the New Jersey Coastal Plain has been described by Zapecza (1984). The framework report describes the sequence of aquifers and confining units and includes thickness maps of the aquifers and confining units and structure contour maps of the tops of the aquifers. Aquifer and confining-unit characteristics are summarized in this report. Pumpage and water-level data are presented by Zapecza, Voronin, and Martin (1987).

These data were incorporated into a digital model of the ground-water flow system, and flow was simulated for prepumping steady-state conditions (pre-1896) and transient conditions from 1896 through 1980. The ground-water flow system in sediments of the New Jersey Coastal Plain and in offshore sediments that contain water with less than 10,000 milligrams per liter (mg/L) chloride concentration was analyzed.

LOCATION AND EXTENT

The New Jersey Coastal Plain is about one-fifth of the northern Atlantic Coastal Plain and is shown in figure 1. The emerged part of the New Jersey Coastal Plain extends from the Fall Line, the northwestern limit of the Coastal Plain sediments, to the Atlantic Ocean in the southeast, and from the Raritan Bay in the northeast, to the Delaware Bay in the southwest. The Delaware River overlies Coastal Plain sediments near the Fall Line between northern Burlington County and western Salem County.

The study area shown in figure 1 covers about 9,000 mi^2 and extends from the Fall Line to about 20 miles (mi) offshore in the southeast. The area extends from Raritan Bay to the center of Delaware Bay. In New Jersey, the study area includes all of Monmouth, Burlington, Ocean, Camden, Gloucester, Salem, Atlantic, Cumberland, and Cape May Counties and parts of Middlesex and Mercer Counties. The model area, which is larger than the study area, also includes parts of New Castle County, Del., and Philadelphia and Buck Counties, Pa.

PREVIOUS INVESTIGATIONS

Numerous reports have been published that describe the hydrogeology of the Coastal Plain sediments. Table 1 lists the major sources of data and information on well identification, water levels, aquifer and confining-unit characteristics, withdrawals, and flow-system characteristics used in this study. The publications are generally of four types. Data reports are a compilation of data for a particular area. Areal studies sometimes include compilations of data but also describe the geology and hydrology of an area. Interpretive studies quantify certain aspects of the ground-water flow system, such as water levels or aquifer thickness, as continuous characteristics over an area using both observed data and principles of hydrology. Simulation studies are similar to interpretive studies; however, digital modeling techniques are used to incorporate large amounts of data. Simulation studies quantify aspects of the ground-water flow system that are not easily quantified from observed data, such as the hydraulic properties of aquifers and confining units and the response of the aquifer system to withdrawals.

Investigations of New Jersey Coastal Plain aquifers consisted primarily of areal studies from the late 1920's to the early 1970's. Simulation studies began in the mid-1970's and generally were analyses of limited parts

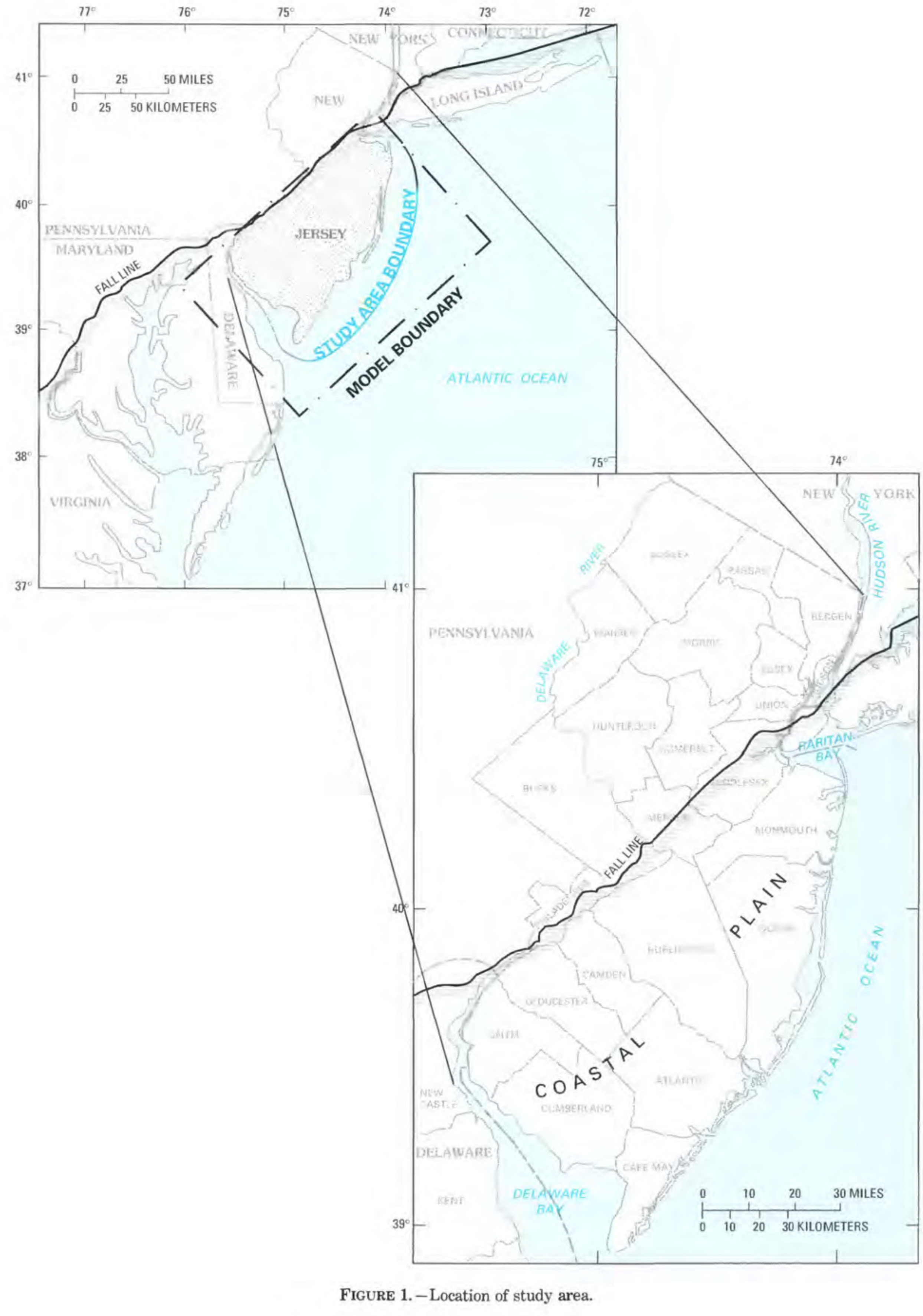


FIGURE 1.—Location of study area.

TABLE 1.—*Previous hydrogeologic investigations*

Reference	Area or subject
	AREAL STUDIES
Anderson and Appel, 1969	Ocean County.
Barksdale and others, 1943	Middlesex County.
Barksdale and others, 1958	Lower Delaware River basin.
Clark and others, 1968	Atlantic County.
Farlekas, Nemickas, and Gill 1976	Camden County.
Gill, 1962a	Cape May County.
Gill, 1959	Cape May Peninsula.
Greenman and others, 1961	Southeastern Pennsylvania.
Hardt and Hilton, 1969	Gloucester County.
Jablonski, 1968	Monmouth County.
Nichols, 1977b	Northern Coastal Plain (Englishtown aquifer).
Rhodehamel, 1973	Mullica River basin, central Coastal Plain.
Rooney, 1971	Cumberland County.
Rosenau and others, 1969	Salem County.
Rush, 1968	Burlington County.
Vecchioli and Palmer, 1962	Mercer County.
	DATA REPORTS AND INTERPRETIVE STUDIES
Eckel and Walker, 1986	1983 Coastal Plain water levels.
Gill and Farlekas, 1976	Potomac-Raritan-Magothy aquifer system geohydrologic maps.
Gill, 1962b	Well records and logs, Cape May County.
Hardt, 1963	Public water supplies, Gloucester County.
Jablonski, 1959 and 1960	Well records, Monmouth County.
Meisler, 1980	Occurrence of salty ground water in the northern Atlantic Coastal Plain.
Rush, 1962	Well records and water quality in Burlington County.
Thompson, 1928, 1930, 1932	Ground-water supplies of the Atlantic City region, Asbury Park area, and Camden area.
Vowinkel, 1984	Generalized withdrawal data-management system.
Vowinkel and Foster, 1981	Generalized geohydrology of the Coastal Plain.
Walker, 1983	1978 Coastal Plain water levels.
Zapecza, 1984	Hydrogeology of the Coastal Plain.
Zapecza, Voronin, and Martin, 1987	Coastal Plain water levels and withdrawals.
	SIMULATION STUDIES
Farlekas, 1979	Farrington aquifer, northern Coastal Plain.
Harbaugh and Tilley, 1984	Water-table aquifer, Mullica River basin, central Coastal Plain.
Luzier, 1980, and Harbaugh, Luzier, and Stellerine, 1980	Potomac-Raritan-Magothy aquifer system, New Jersey Coastal Plain.
Nemikas, 1976	Wenonah Mount Laurel aquifer, northern Coastal Plain.
Nichols, 1977a	Englishtown aquifer, northern Coastal Plain.

of the ground-water flow system using a finite-difference simulation of two-dimensional flow. Previous simulation studies have included analyses of the Farrington, Englishtown, and Wenonah-Mount Laurel aquifers and the Potomac-Raritan-Magothy aquifer system (table 1).

ACKNOWLEDGMENTS

Thanks are given to the New Jersey Department of Environmental Protection for its assistance in supplying withdrawal and well data. Thanks are also given to the many municipalities, industries, and individuals that provided withdrawal and water-level information to the New Jersey Department of Environmental Protection and the U.S. Geological Survey.

CONCEPTUAL HYDROGEOLOGIC MODEL

The New Jersey Coastal Plain sediments are a seaward-dipping wedge of alternating layers of sand, silt, and clay overlying crystalline basement. Cretaceous and Tertiary sediments generally strike northeast-southwest and dip 10 to 60 feet per mile (ft/mi) to the southeast, whereas overlying Quaternary sediments, where present, are flat lying (Zapecza, 1984). The Coastal Plain deposits thicken downdip from a feather edge at the Fall Line to more than 6,500 feet in southern Cape May County (Gill and Farlekas, 1976).

The Coastal Plain sediments have been subdivided into a sequence of aquifers and confining units based on the general hydraulic properties of the sediments. The geologic and hydrogeologic units are listed in table 2. The

TABLE 2.—*Geologic and hydrogeologic units of the New Jersey Coastal Plain and model units used in this study*

SYSTEM	SERIES	GEOLOGIC UNIT	HYDROGEOLOGIC UNIT	MODEL UNITS UPDIP	MODEL UNITS DOWNDIP
Quaternary	Holocene	Alluvial deposits	Undifferentiated	Upper Kirkwood-Cohansey aquifer (A9)	
Quaternary	Holocene	Beach sand and gravel	Undifferentiated	Upper Kirkwood-Cohansey aquifer (A9)	Holly Beach aquifer (A10)
Quaternary	Pleistocene	Cape May Formation	K-C aq. sys.[1]		Cape May confining unit (C9) / Upper Kirkwood-Cohansey aquifer (A9)
Tertiary	Miocene	Pensauken Formation		Upper Kirkwood-Cohansey aquifer (A9)	
Tertiary	Miocene	Bridgeton Formation	Kirkwood-Cohansey aquifer system	Upper Kirkwood-Cohansey aquifer (A9)	Upper Kirkwood-Cohansey aquifer (A9)
Tertiary	Miocene	Beacon Hill Gravel	Kirkwood-Cohansey aquifer system	Upper Kirkwood-Cohansey aquifer (A9)	Upper Kirkwood-Cohansey aquifer (A9)
Tertiary	Miocene	Cohansey Sand	Kirkwood-Cohansey aquifer system	Upper Kirkwood-Cohansey aquifer (A9) / Lower Kirkwood-Cohansey aquifer (A8)	Upper Kirkwood-Cohansey aquifer (A9)
Tertiary	Miocene	Kirkwood Formation	Kirkwood-Cohansey aquifer system	Lower Kirkwood-Cohansey aquifer (A8)	Upper Kirkwood-Cohansey aquifer (A9)
Tertiary	Miocene	Kirkwood Formation	Confining unit		Confining unit overlying the Rio Grande water-bearing zone (C8)
Tertiary	Miocene	Kirkwood Formation	Rio Grande w-bz[2]		Confining unit overlying the Rio Grande water-bearing zone (C8)
Tertiary	Miocene	Kirkwood Formation	Confining unit		Confined Kirkwood aquifer (A8)
Tertiary	Miocene	Kirkwood Formation	Atlantic City 800-foot sand		Confined Kirkwood aquifer (A8)
Tertiary	Miocene	Kirkwood Formation	Composite confining bed	Basal Kirkwood confining unit (C7)	Basal Kirkwood confining unit (C7)
Tertiary	Eocene	Piney Point Formation	Composite confining bed: Piney Point aquifer	Piney Point aquifer (A7)	Piney Point aquifer (A7)
Tertiary	Eocene	Shark River Formation	Composite confining bed	Vincentown-Manasquan confining unit (C6)	Vincentown-Manasquan confining unit (C6)
Tertiary	Eocene	Manasquan Formation	Composite confining bed	Vincentown-Manasquan confining unit (C6)	Vincentown-Manasquan confining unit (C6)
Tertiary	Paleocene	Vincentown Formation	Composite confining bed: Vincentown aquifer	Vincentown aquifer (A6)	Vincentown-Manasquan confining unit (C6)
Tertiary	Paleocene	Hornerstown Sand	Composite confining bed	Navesink-Hornerstown confining unit (C5)	Navesink-Hornerstown confining unit (C5)
Cretaceous	Upper Cretaceous	Tinton Sand	Composite confining bed	Navesink-Hornerstown confining unit (C5)	Navesink-Hornerstown confining unit (C5)
Cretaceous	Upper Cretaceous	Red Bank Sand	Composite confining bed: Red Bank sand	Navesink-Hornerstown confining unit (C5)	Navesink-Hornerstown confining unit (C5)
Cretaceous	Upper Cretaceous	Navesink Formation	Composite confining bed	Navesink-Hornerstown confining unit (C5)	Navesink-Hornerstown confining unit (C5)
Cretaceous	Upper Cretaceous	Mount Laurel Sand	Wenonah-Mount Laurel aquifer	Wenonah-Mount Laurel aquifer (A5)	Wenonah-Mount Laurel aquifer (A5)
Cretaceous	Upper Cretaceous	Wenonah Formation	Wenonah-Mount Laurel aquifer / Marshalltown-Wenonah confining unit	Wenonah-Mount Laurel aquifer (A5) / Marshalltown-Wenonah confining unit (C4)	Wenonah-Mount Laurel aquifer (A5) / Marshalltown-Wenonah confining unit (C4)
Cretaceous	Upper Cretaceous	Marshalltown Formation	Marshalltown-Wenonah confining unit	Marshalltown-Wenonah confining unit (C4)	Marshalltown-Wenonah confining unit (C4)
Cretaceous	Upper Cretaceous	Englishtown Formation	Englishtown aquifer system	Englishtown aquifer (A4)	
Cretaceous	Upper Cretaceous	Woodbury Clay	Merchantville-Woodbury confining unit	Merchantville-Woodbury confining unit (C3)	Merchantville-Woodbury confining unit (C3)
Cretaceous	Upper Cretaceous	Merchantville Formation	Merchantville-Woodbury confining unit	Merchantville-Woodbury confining unit (C3)	Merchantville-Woodbury confining unit (C3)
Cretaceous	Upper Cretaceous	Magothy Formation	Potomac-Raritan-Magothy aquifer system: Upper aquifer	Upper Potomac-Raritan-Magothy aquifer (A3)	Upper Potomac-Raritan-Magothy aquifer (A3)
Cretaceous	Upper Cretaceous	Magothy Formation / Raritan Formation	Potomac-Raritan-Magothy aquifer system: Conf un	Confining unit between the middle and upper Potomac-Raritan-Magothy aquifers (C2)	Confining unit between the middle and upper Potomac-Raritan-Magothy aquifers (C2)
Cretaceous	Upper Cretaceous	Raritan Formation	Potomac-Raritan-Magothy aquifer system: Middle aquifer	Middle Potomac-Raritan-Magothy aquifer (A2)	Middle Potomac-Raritan-Magothy aquifer (A2)
Cretaceous	Upper Cretaceous / Lower Cretaceous	Potomac Group	Potomac-Raritan-Magothy aquifer system: Conf un	Confining unit between the lower and middle Potomac-Raritan-Magothy aquifers (C1)	Confining unit between the lower and middle Potomac-Raritan-Magothy aquifers (C1)
Cretaceous	Lower Cretaceous	Potomac Group	Potomac-Raritan-Magothy aquifer system: Lower aquifer	Lower Potomac-Raritan-Magothy aquifer (A1)	Lower Potomac-Raritan-Magothy aquifer (A1)
Pre-Cretaceous		Bedrock	Bedrock confining unit		

[1] Kirkwood-Cohansey aquifer system
[2] Rio Grande water-bearing zone

☐ Indicates adjacent geologic or hydrogeologic unit not present

Modified from Zapecza (1984, table 2) and Seaber (1965, table 3).

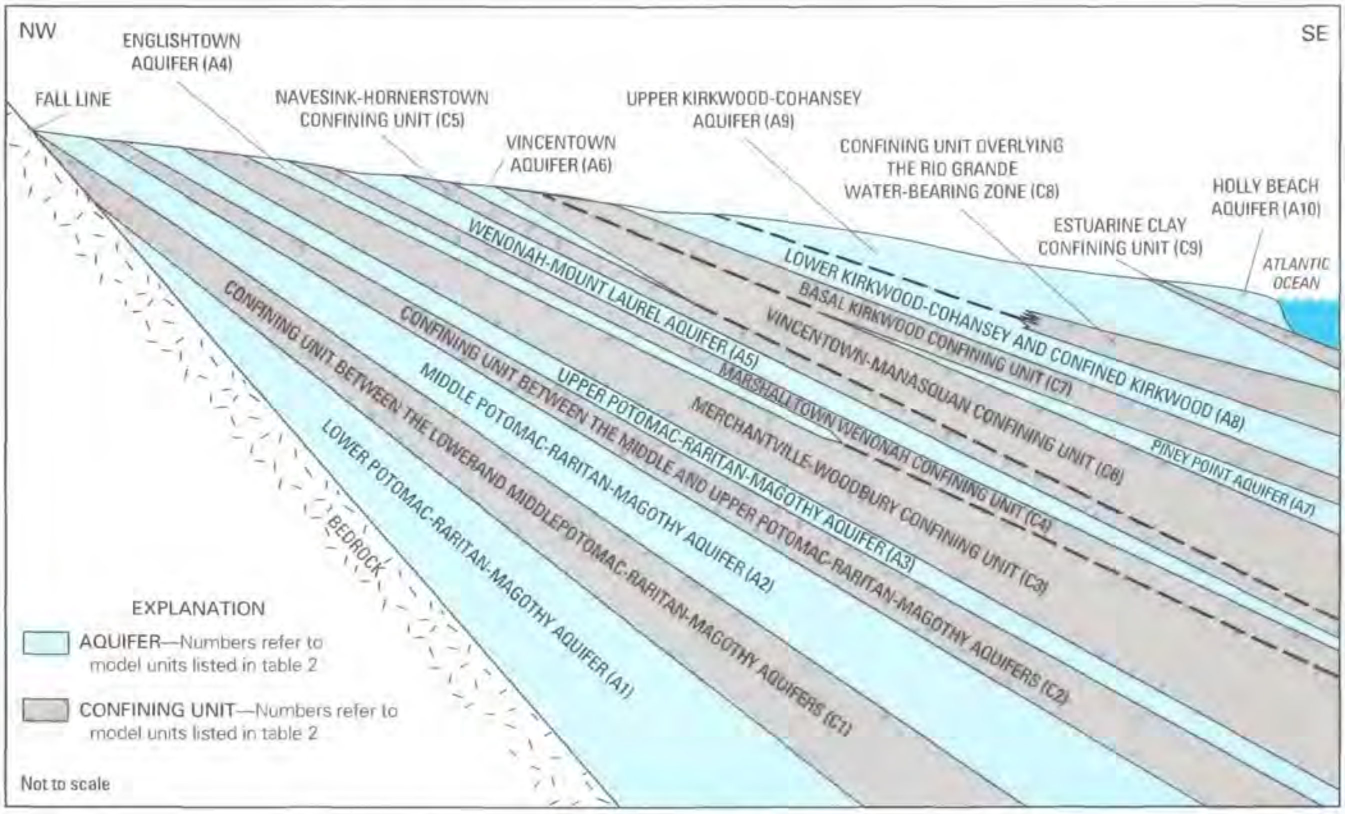


FIGURE 2.—Generalized hydrogeologic section of the New Jersey Coastal Plain.

aquifer and confining-unit designations do not necessarily correspond to time- or rock-stratigraphic designations. The New Jersey Coastal Plain aquifers are composed predominantly of sand but also include interbedded silts and clays. The confining units are composed predominantly of clay and silt with minor amounts of sand. Most of these units crop out near the Fall Line in irregular bands parallel to the strike. The Coastal Plain sediments were modeled as 10 major aquifers and 9 intervening confining units. These model units are identified in table 2 and in the generalized hydrogeologic section of the Coastal Plain in figure 2.

Recharge to the New Jersey Coastal Plain is by infiltration of precipitation, leakage from surface-water bodies, and lateral flow from adjacent States. Water is discharged from the aquifers as flow to surface-water bodies, evapotranspiration, withdrawals from wells, and lateral flow to adjacent States. Previous investigations have reached the general conclusion that the major sources of water to wells are induced infiltration from surface-water bodies, decreased flow to surface-water bodies, increased flow from adjacent States, decreased flow to adjacent States, and water released from storage. Minor sources of water to wells, sources that are important only locally, are increased infiltration of precipitation and decreased evapotranspiration.

AQUIFER AND CONFINING-UNIT CHARACTERISTICS

The model framework is based on the hydrogeologic framework of the New Jersey Coastal Plain presented by Zapecza (1984). Names of geologic and hydrogeologic units used in the following sections describing the model units are from Zapecza, unless otherwise noted. Although the major aquifers and confining units in New Jersey were mapped by Zapecza, several modifications were necessary to adapt the hydrogeologic framework to a model framework. Estimates of unit thickness for areas not mapped by Zapecza were made for most units, and several units were subdivided into separate model layers. Also, some framework modifications were needed to ensure a compatible framework with adjacent States. Lithologic characteristics of the geologic units, aquifers and confining units, and stratigraphic nomenclature have been discussed by Zapecza and are not discussed in this report.

An informal nomenclature is used for the model units in this report. The names used in this report were selected partly for convenience and are not always descriptive or exact. In many cases, an aquifer name represents a regional aquifer and additional local minor aquifers. The sediments represented by each model unit are described in the following sections. The aquifers and confining units are described in order of oldest (deepest) to youngest (shallowest), following the order of presentation by Zapecza (1984).

The hydrogeologic characteristics of each aquifer and confining unit are also described in the following sections. Data on aquifer transmissivity, hydraulic conductivity, and storage coefficient are shown in table 3. Data on vertical hydraulic conductivity of confining units are given in table 4. Structure contour maps of the tops of the confined aquifers and thickness maps of the confining units are shown in figures 3 through 20.

Data presented in tables 3 and 4 are compiled from the literature of previous estimates of the hydraulic characteristics of aquifers and confining units within the New Jersey Coastal Plain. However, these data are not internally consistent because they are from many different methods of collection and analysis and may not represent actual hydraulic characteristics because of the method of collection or analysis. Many early estimates of transmissivity may be as much as 25 percent too high if aquifer-test data from a leaky confined aquifer were analyzed as though the data were from a nonleaky aquifer (Harold Meisler, oral commun., 1985). This is probably true of transmissivity data from Barksdale and others (1958), Rush (1968), Jablonski (1968), and Gill (1962a). Many laboratory analyses of confining-unit hydraulic conductivity were made on disturbed and repacked cores (G.M. Farlekas, oral commun., 1985). Results of such analyses are given in Farlekas and others (1976). Some of the data presented in tables 3 and 4 are shown for aquifers or confining units different from those given in the original reference. The unit names given in the tables are based on the model framework.

Only freshwater flow has been simulated, and, for the purpose of this report, freshwater is water with chloride concentrations less than 10,000 mg/L. Therefore, only sediments containing water with less than 10,000 mg/L chloride concentrations are represented in figures 3 through 20 and discussed in the following sections. The estimated locations of 10,000 mg/L chloride concentrations within the aquifers are based on data presented by Meisler (1981) and are referred to in this report as the downdip limit of freshwater within the aquifers. The assumptions associated with choosing 10,000 mg/L chloride concentrations as the limit of fresh ground water are discussed further in the section on boundary conditions.

The maps shown in figures 3 through 20 are approximate representations of the Coastal Plain hydrogeologic framework. In several areas, altitudes and thicknesses were estimated from sparse geophysical logs. For most units, altitudes and thicknesses in offshore areas are estimated from approximated contour lines in onshore areas.

LOWER POTOMAC-RARITAN-MAGOTHY AQUIFER

The lowermost Coastal Plain aquifer modeled is the lower aquifer of the Potomac-Raritan-Magothy aquifer system. This aquifer is referred to as the lower Potomac-Raritan-Magothy aquifer and is designated A1 in the model. It is present everywhere in Camden County and in parts of Monmouth, Ocean, Burlington, Atlantic, Gloucester, and Salem Counties. This aquifer is underlain by crystalline basement or weathered basement rock. The configuration of the bedrock surface is shown by Zapecza (1984, pl. 1).

Structure contours of the top of the lower Potomac-Raritan-Magothy aquifer are shown in figure 3. Near the Delaware River in northwestern Salem, Gloucester, and Camden Counties, the altitude of the top of the aquifer is the same as that shown by Zapecza (1984, pl. 6). In downdip areas and in the eastern Coastal Plain, the top of the aquifer is estimated from available geophysical logs. Generally, in these areas, model unit A1 represents the sand beds of the lower third of the undifferentiated Potomac-Raritan-Magothy aquifer system.

The altitude of the top of the lower Potomac-Raritan-Magothy aquifer ranges between 100 feet below sea level near the Delaware River to 3,200 feet below sea level near Barnegat Light in Ocean County and to the east beneath the Atlantic Ocean. The aquifer sediments may extend further downdip than shown on figure 3, but the area modeled is limited to the part of the aquifer containing freshwater as shown by the estimated occurrence of 10,000 mg/L chloride concentrations.

The aquifer is 75 feet thick near the Delaware River and is thickest along the 10,000 mg/L chloride boundary. Along the southern boundary, the aquifer is 450 feet thick in eastern Ocean County, 600 feet thick in southern Gloucester County, and 500 feet thick in eastern Salem County.

The lower Potomac-Raritan-Magothy aquifer is continuous into Delaware and is in the lower part of the Potomac Formation in that State (D.A. Vroblesky, written commun., 1985). The aquifer extends south of Long Island, N.Y., but is not part of the sediments found in New York (Henry Trapp, written commun., 1985). The lower aquifer does not crop out in the New Jersey Coastal Plain and is overlain entirely by the confining unit between the lower and middle Potomac-Raritan-Magothy aquifers.

Aquifer-test analyses indicate that the transmissivity of the lower Potomac-Raritan-Magothy aquifer ranges generally from 2,300 to 9,100 feet squared per day (ft^2/d), and possibly as high as 16,600 ft^2/d (table 3). Storage coefficients from these tests range from 3.3×10^{-5} to 1.5×10^{-3}. Aquifer test data are available only for Camden and Gloucester Counties.

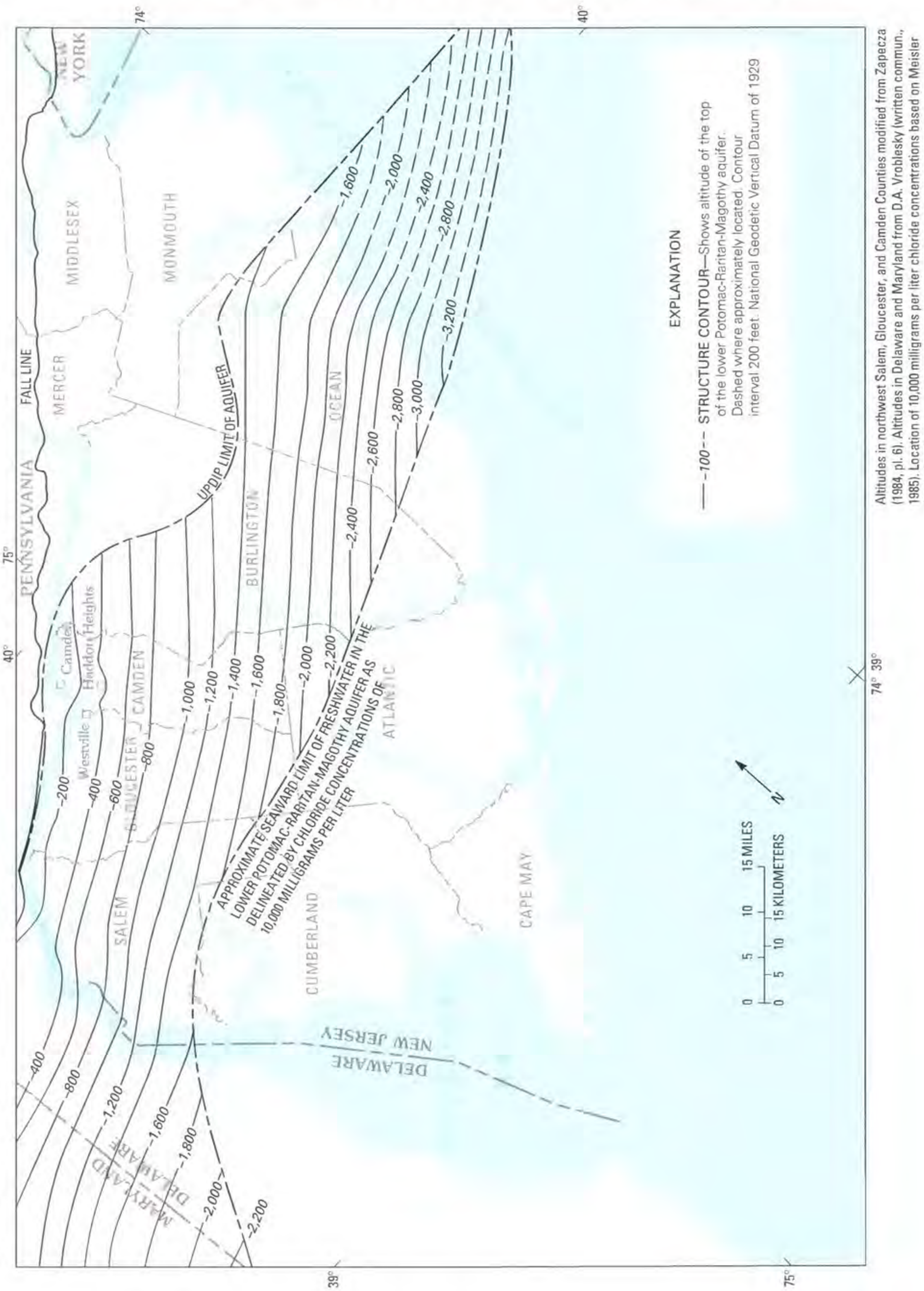
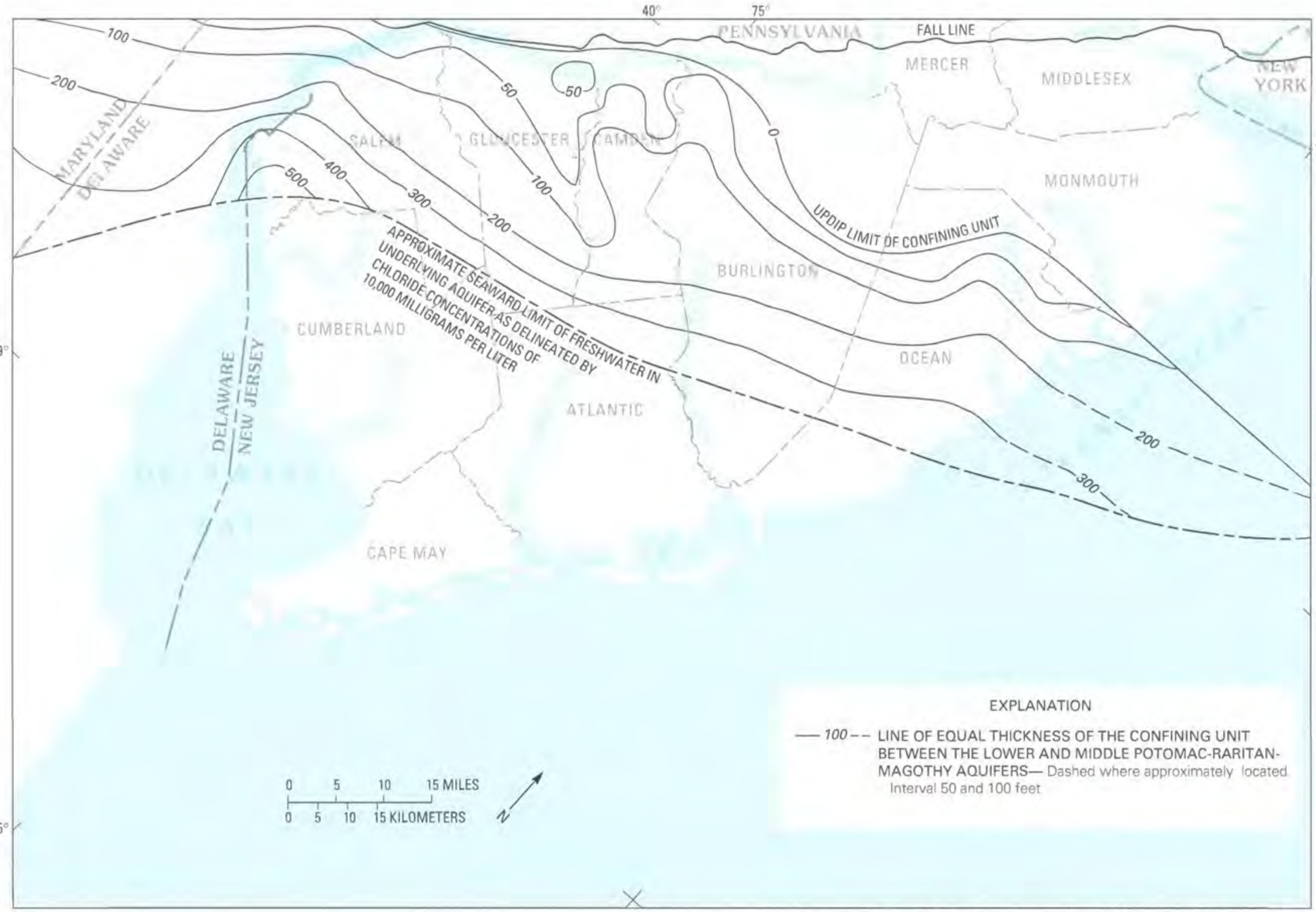
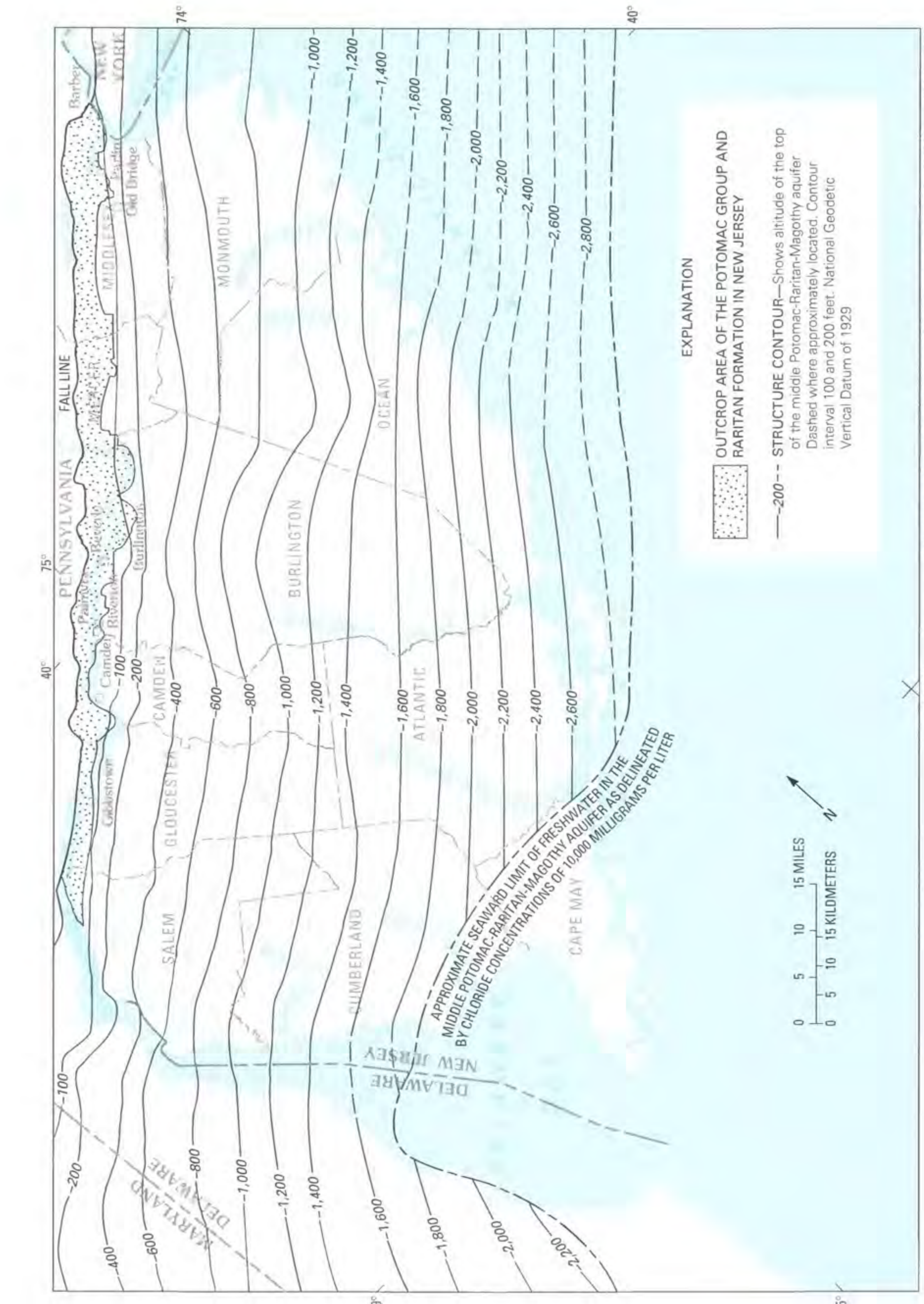


FIGURE 3.—Altitude of the top of the lower Potomac-Raritan-Magothy aquifer (model unit A1).



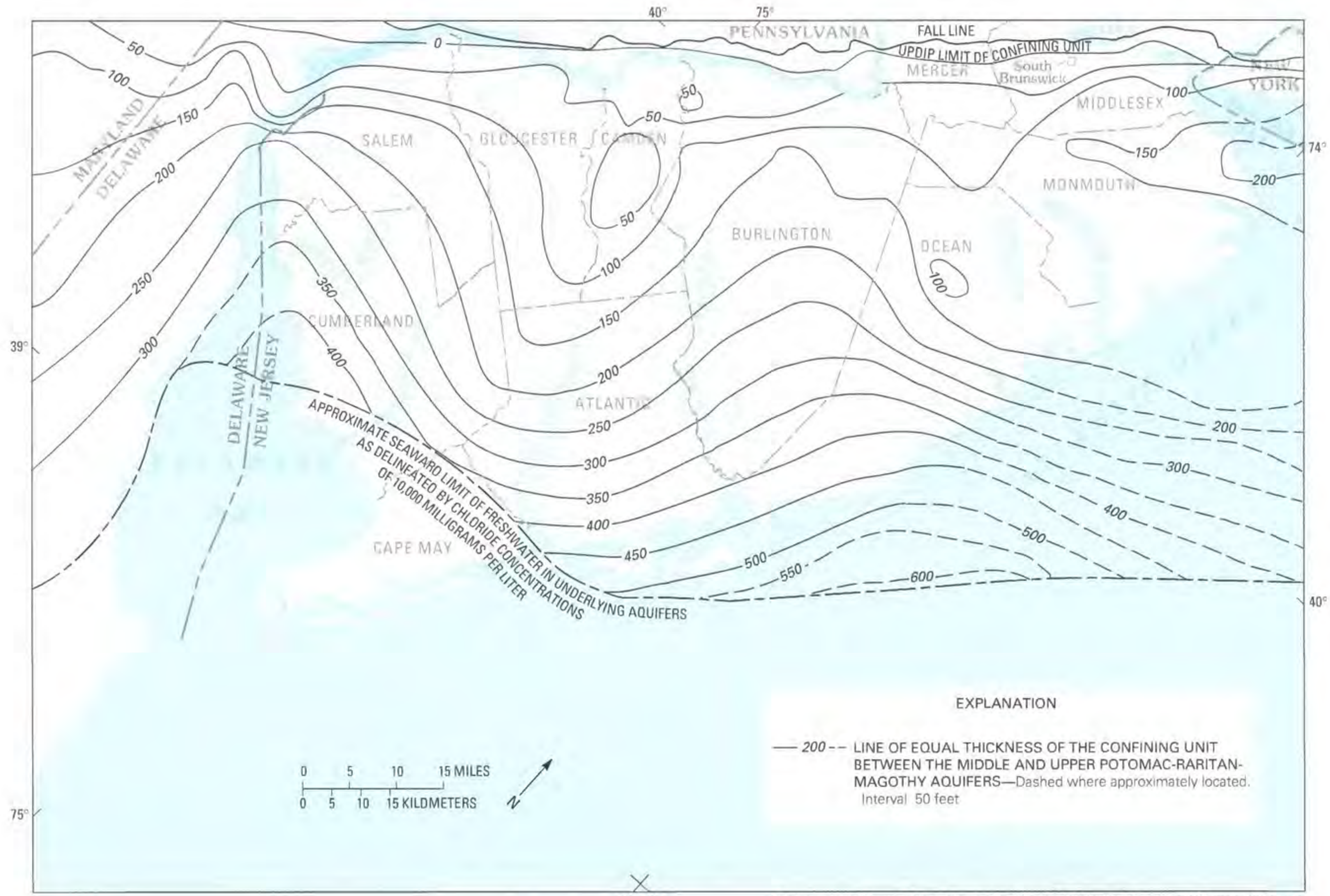
Thickness in northwest Salem, Gloucester, and Camden Counties modified from Zapecza (1984, pl. 6). Thickness in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 4.—Thickness of the confining unit between the lower and middle Potomac-Raritan-Magothy aquifers (model unit C1).



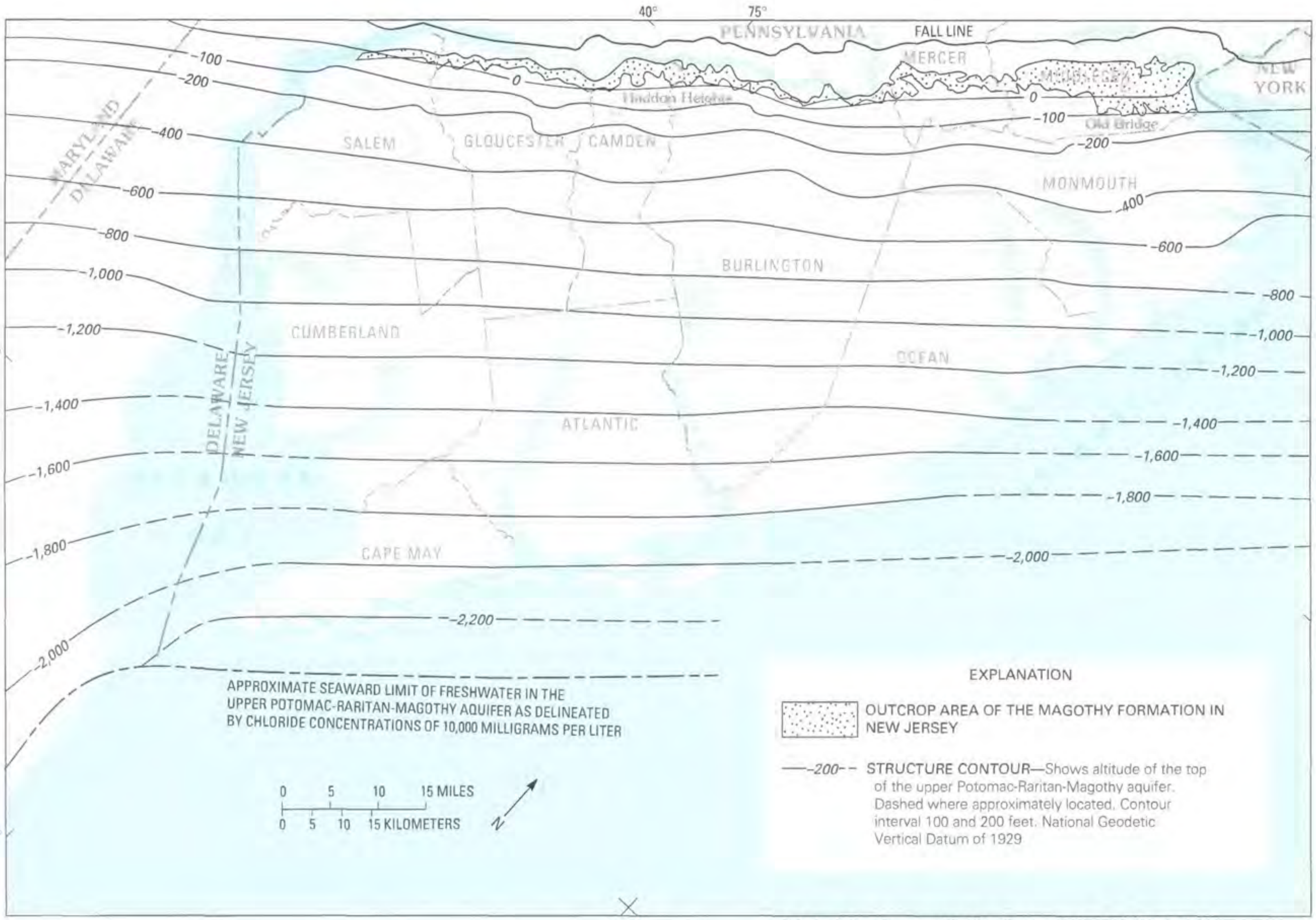
Altitudes in northwest Salem, Gloucester, Camden, Burlington, Ocean, and Monmouth Counties modified from Zapecza (1984, pl. 7). Altitudes in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 5.—Altitude of the top of the middle Potomac-Raritan-Magothy aquifer (model unit A2).



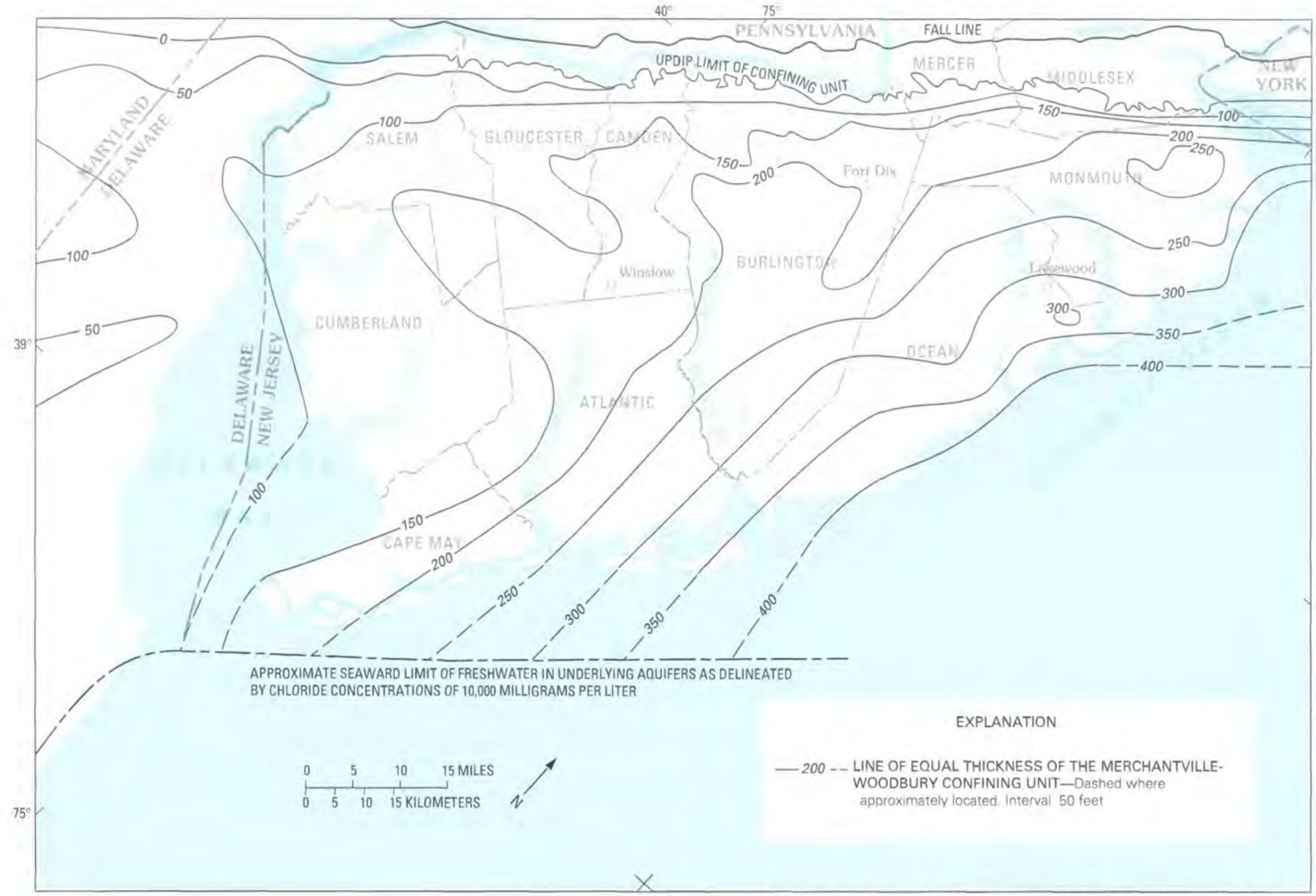
Thickness in northwest Salem, Gloucester, Camden, Burlington, Ocean, and Monmouth Counties modified from Zapecza (1984, pl. 9). Thickness in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 6.—Thickness of the confining unit between the middle and upper Potomac-Raritan-Magothy aquifers (model unit C2).



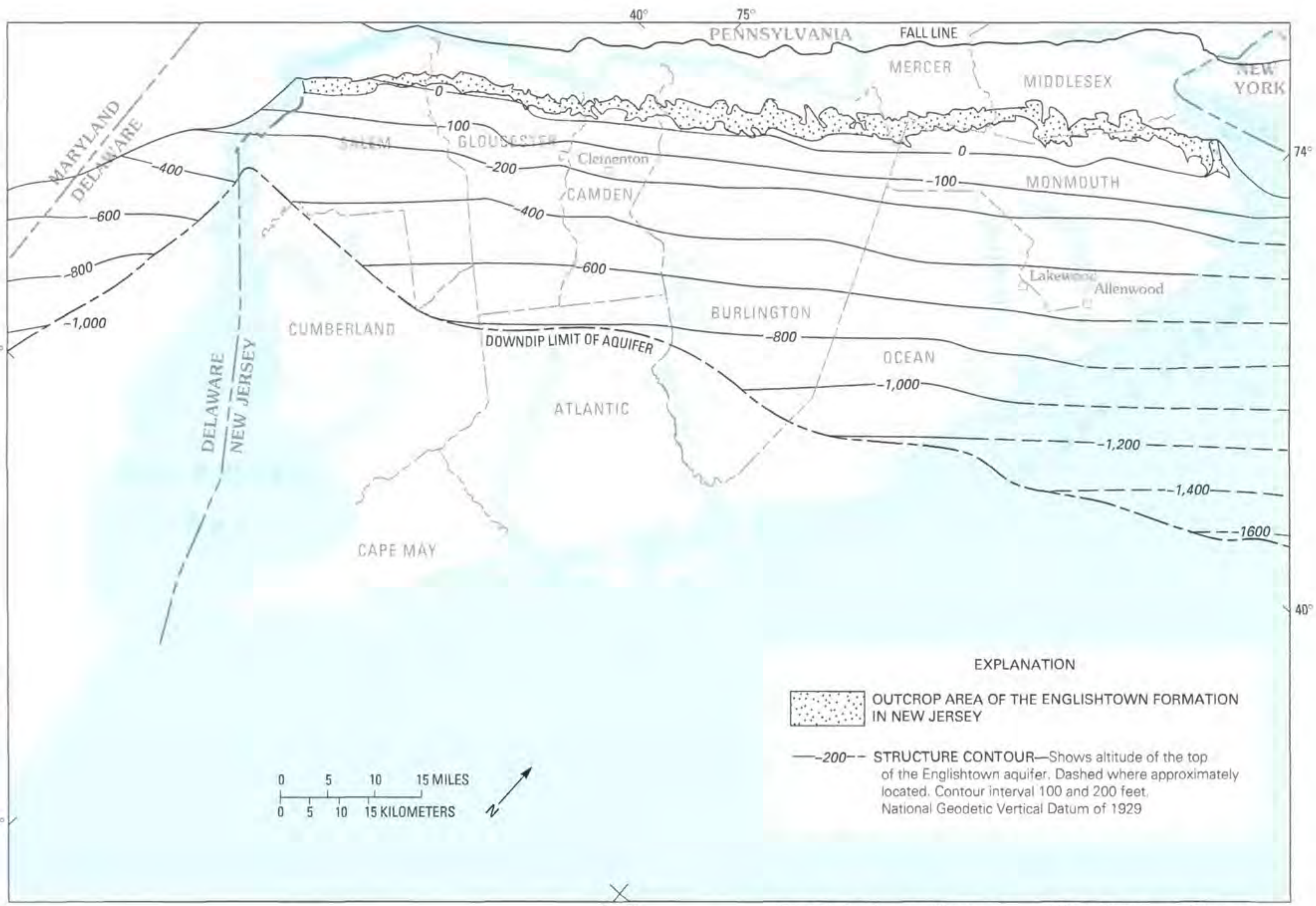
Altitudes in New Jersey modified from Zapecza (1984, pl. 10). Altitudes in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 7.—Altitude of the top of the upper Potomac-Raritan-Magothy aquifer (model unit A3).



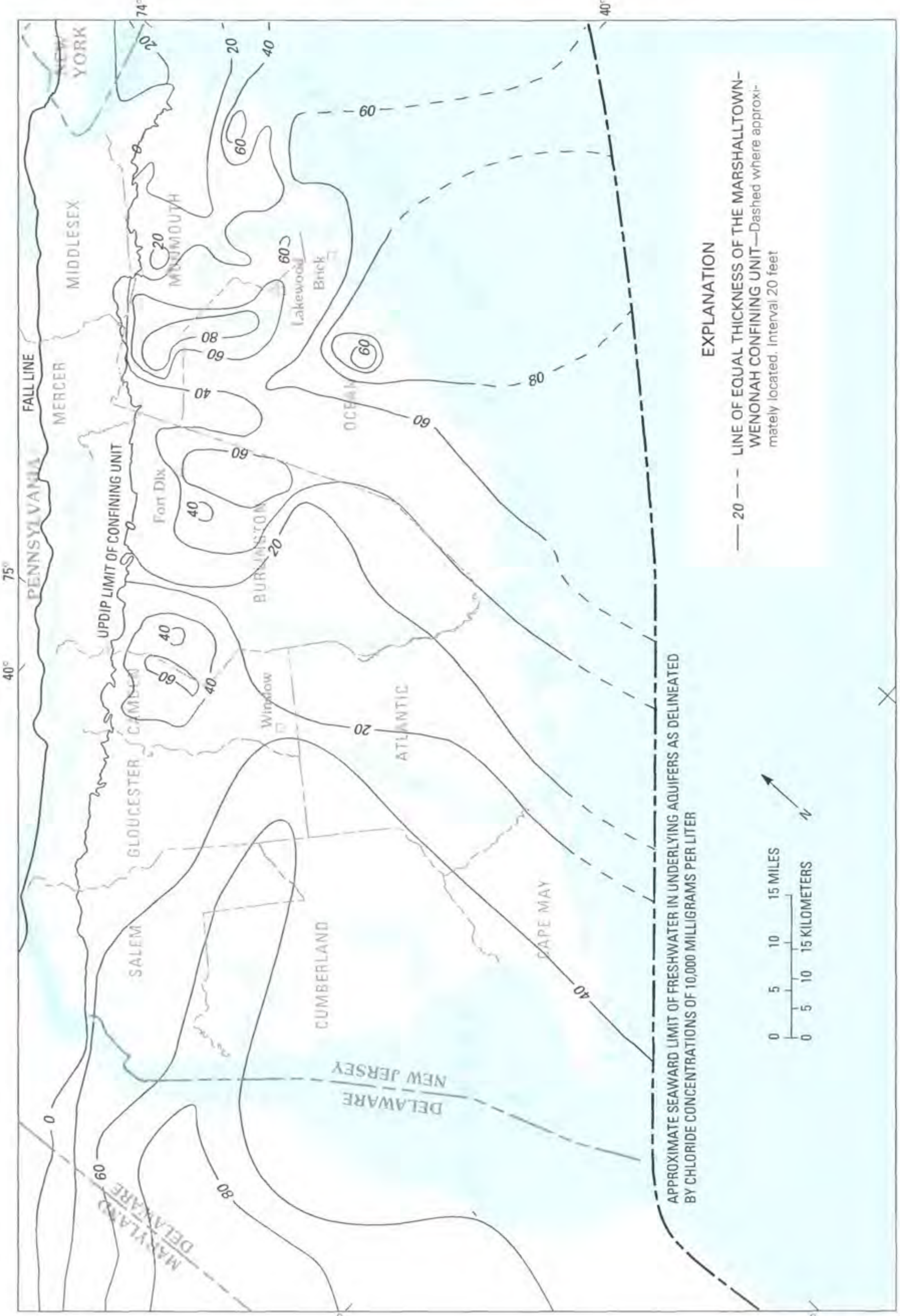
Thickness in northwest Salem, Gloucester, Camden, Burlington, and Monmouth Counties modified from Zapecza (1984, pl.12). Thickness in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 8.—Thickness of the Merchantville-Woodbury confining unit (model unit C3).



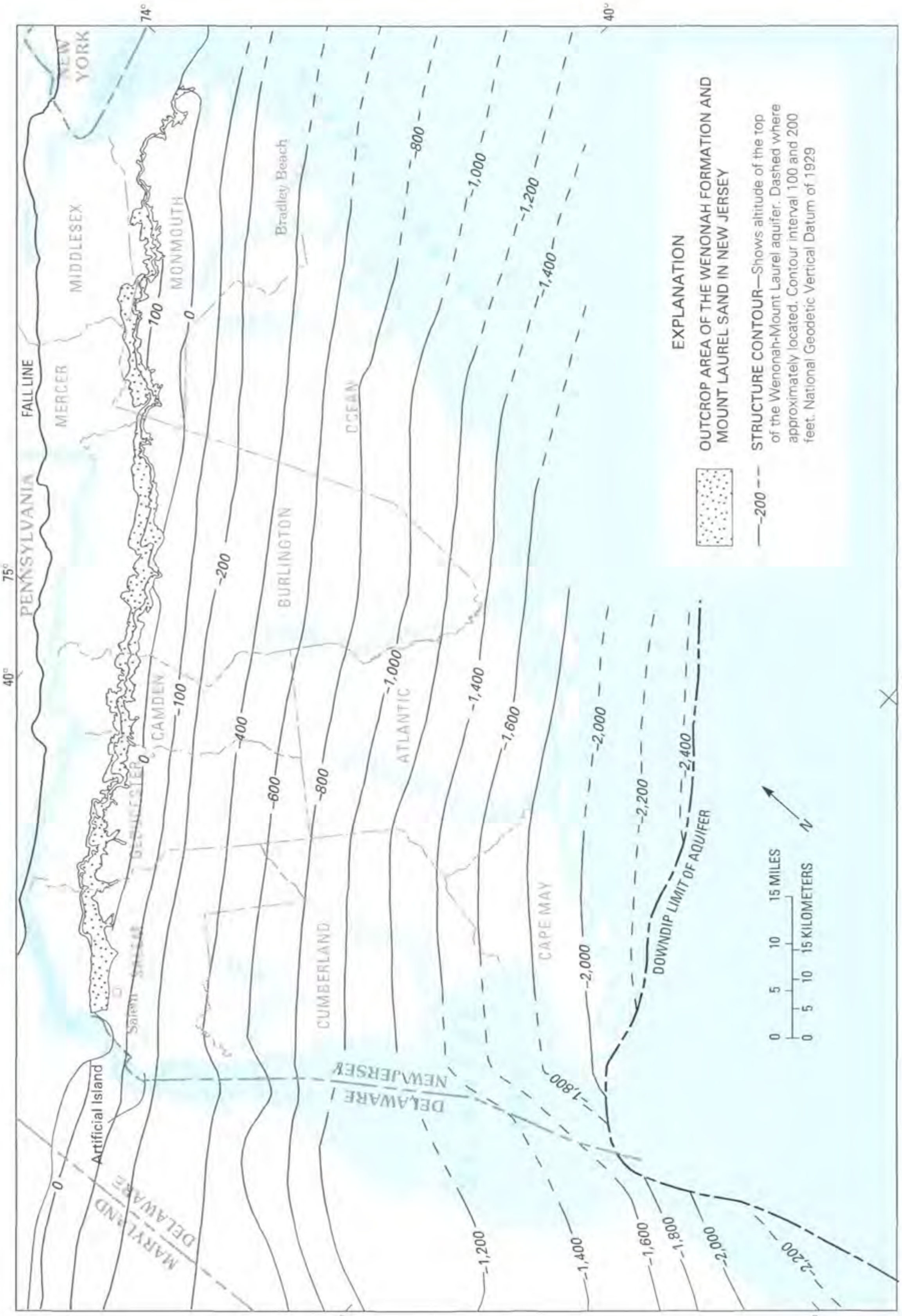
Altitudes in New Jersey modified from Zapecza (1984, pl. 13). Altitudes in Delaware and Maryland from D.A. Vroblesky (written commun., 1985).

FIGURE 9.—Altitude of the top of the Englishtown aquifer (model unit A4).



Thickness in areas overlying the Englishtown aquifer in New Jersey modified from Zapecza (1984, pl. 15). Thickness in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 10.—Thickness of the Marshalltown-Wenonah confining unit (model unit C4).



Altitudes in New Jersey modified from Zapecza (1984, pl. 16). Altitudes in Delaware and Maryland from D.A. Vroblesky (written commun., 1985).

FIGURE 11.—Altitude of the top of the Wenonah-Mount Laurel aquifer (model unit A5).

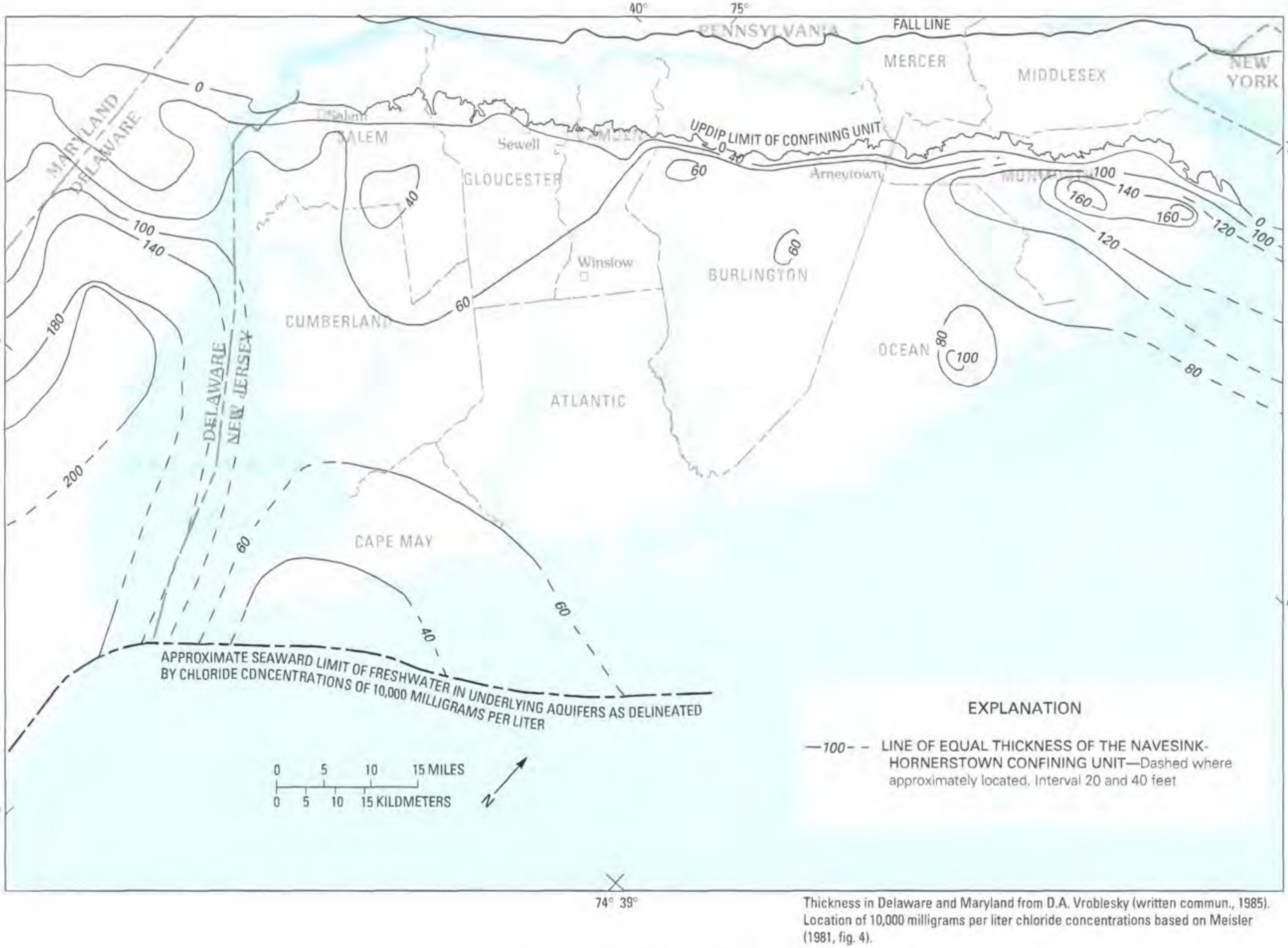
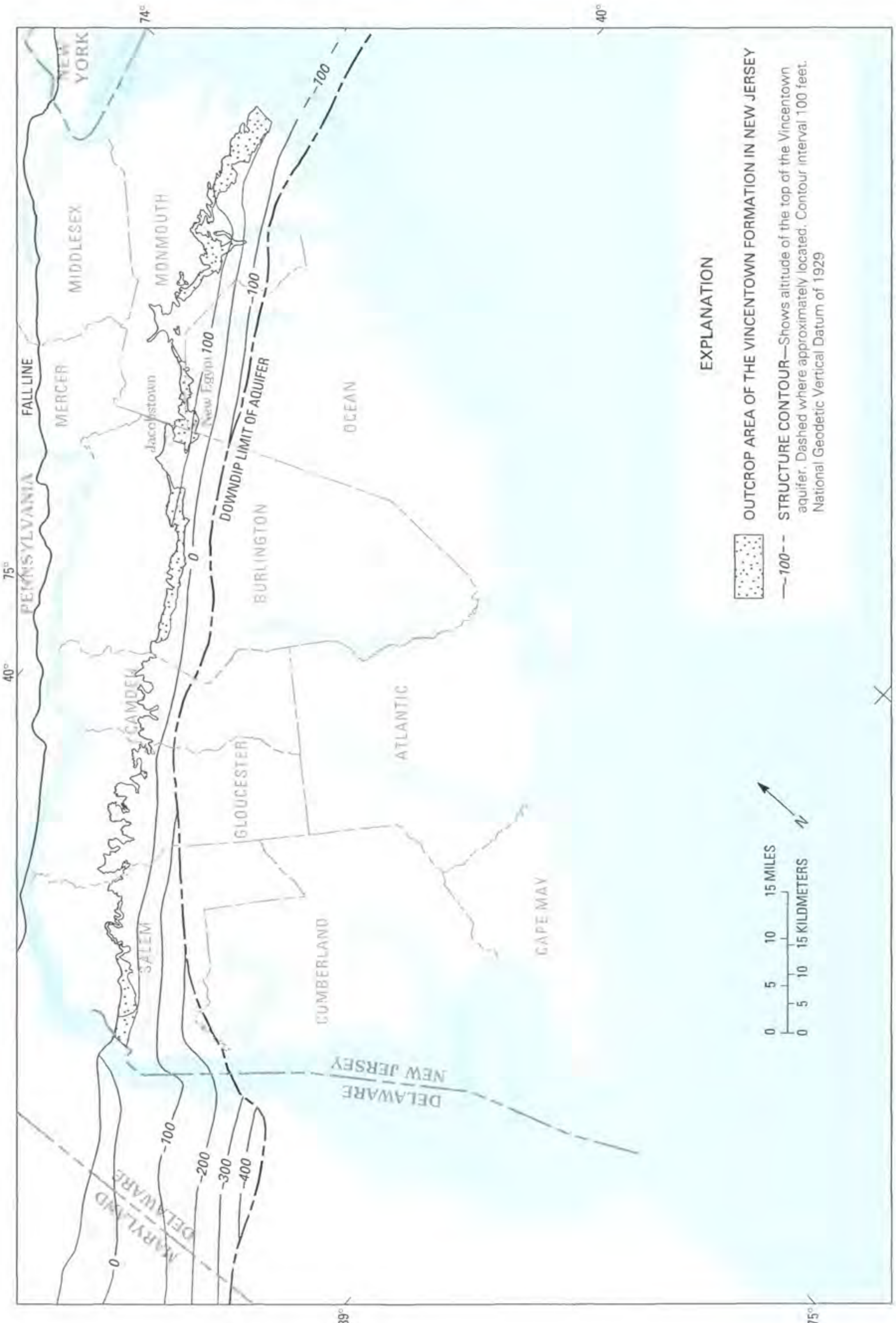


FIGURE 12.—Thickness of the Navesink-Hornerstown confining unit (model unit C5).



Altitudes in New Jersey modified from Zapecza (1984, pl. 19). Altitudes in Delaware and Maryland from D.A. Vroblesky (written commun., 1985).

FIGURE 13.—Altitude of the top of the Vincentown aquifer (model unit A6).

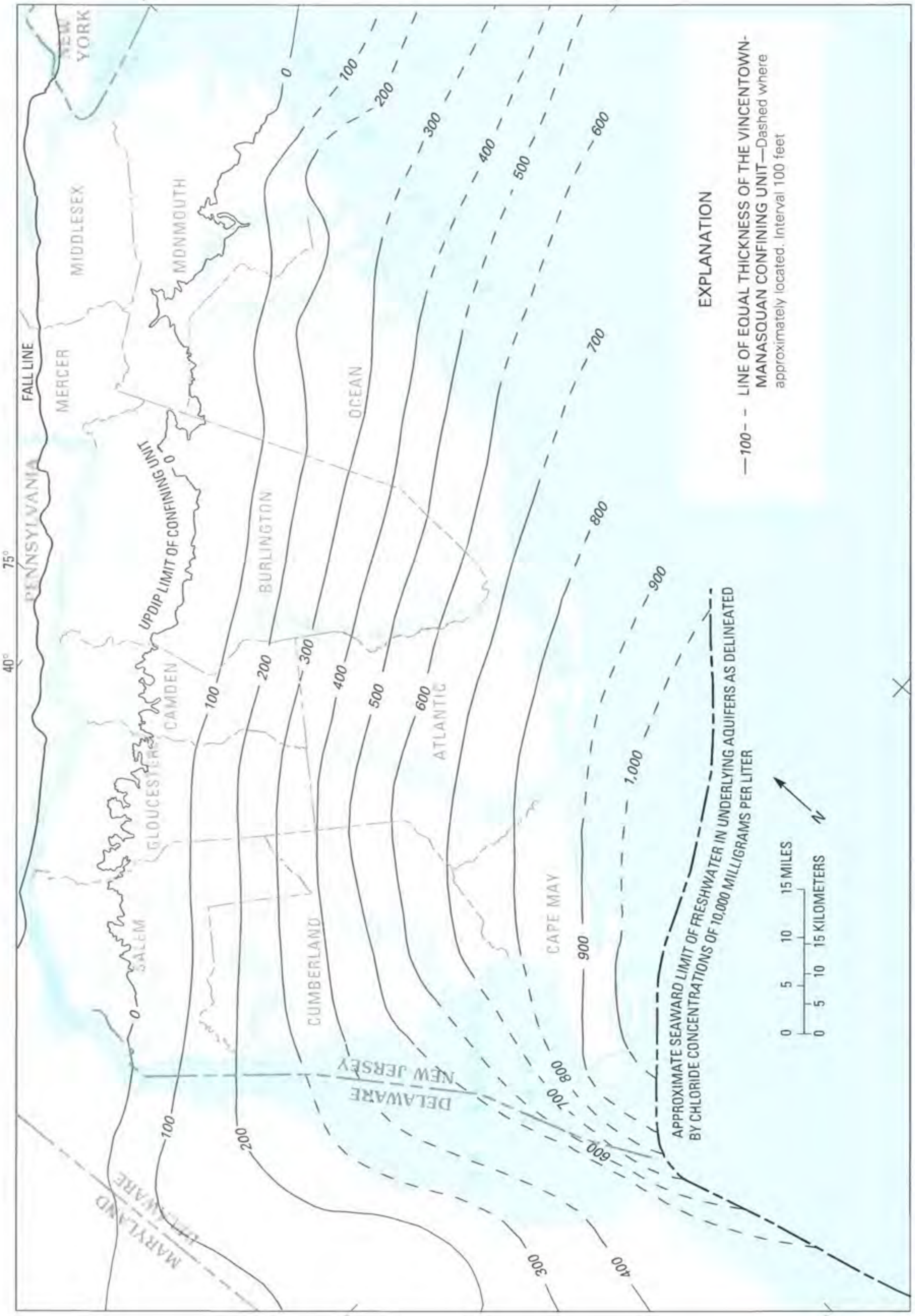
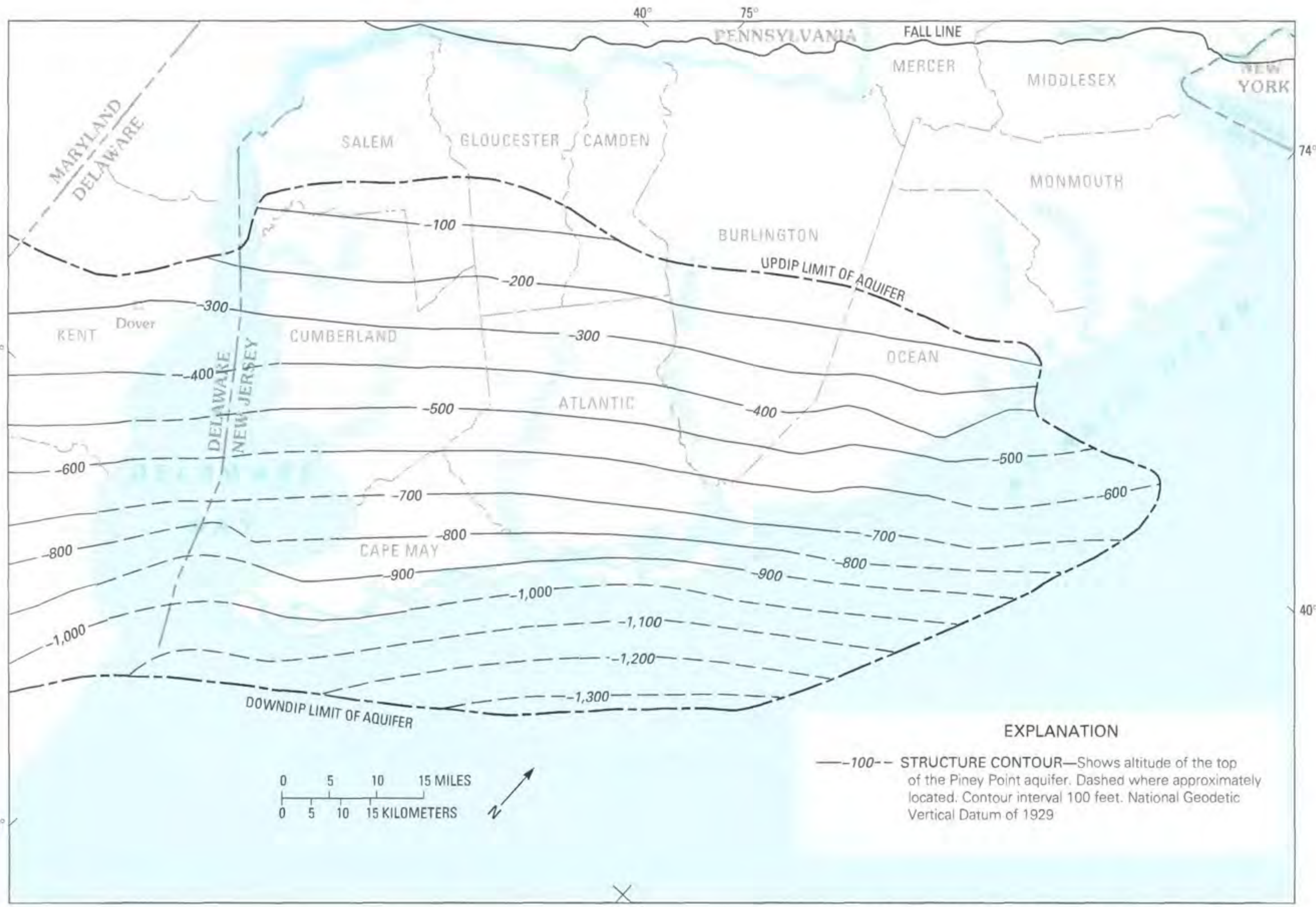
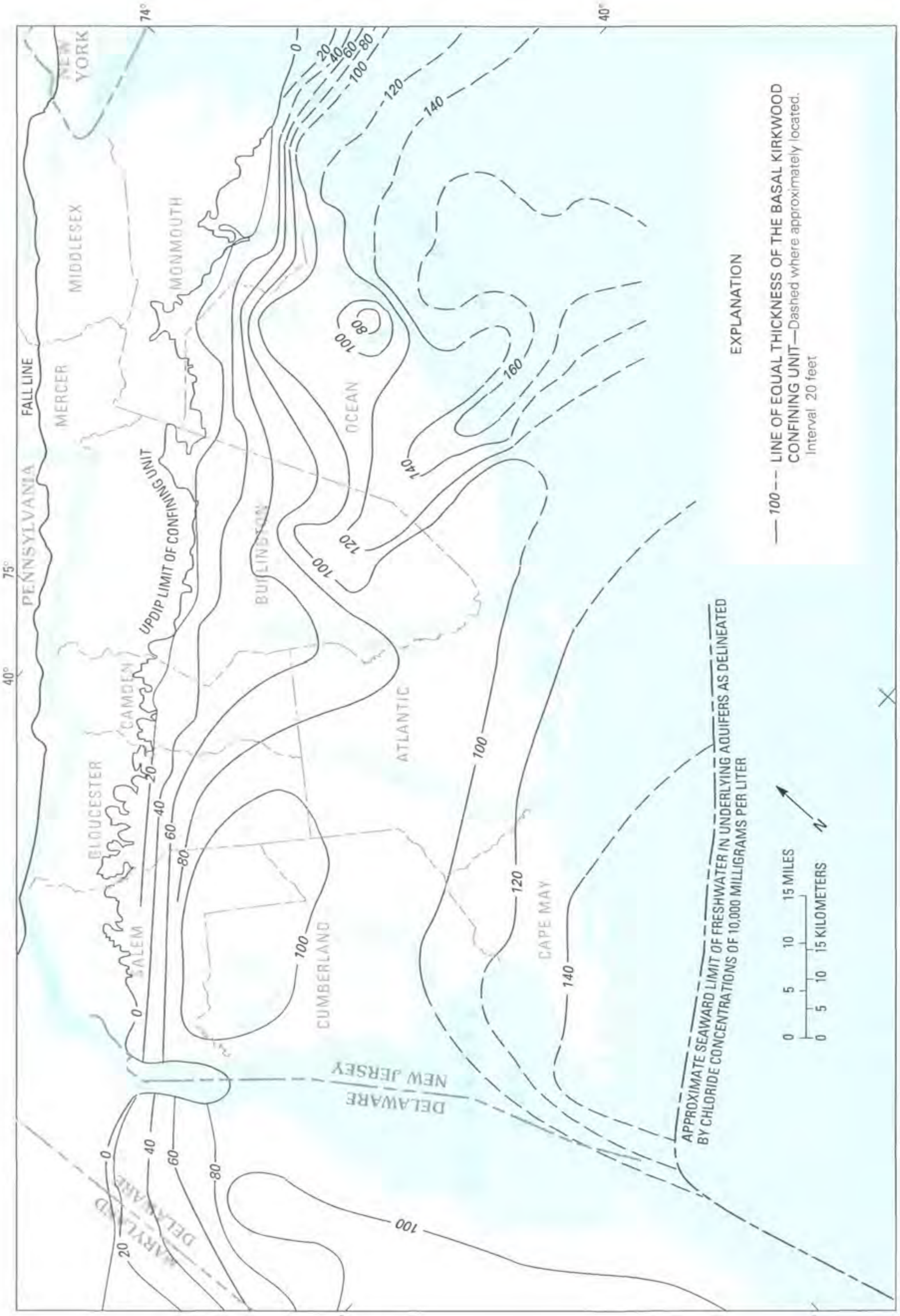


FIGURE 14.—Thickness of the Vincentown-Manasquan confining unit (model unit C6).



Altitudes in New Jersey modified from Zapecza (1984, pl. 19). Altitudes in Delaware and Maryland from D.A. Vroblesky (written commun., 1985).

FIGURE 15.—Altitude of the top of the Piney Point aquifer (model unit A7).



Thickness in Delaware and Maryland from D.A. Vroblesky (written commun., 1985). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 16.—Thickness of the basal Kirkwood confining unit (model unit C7).

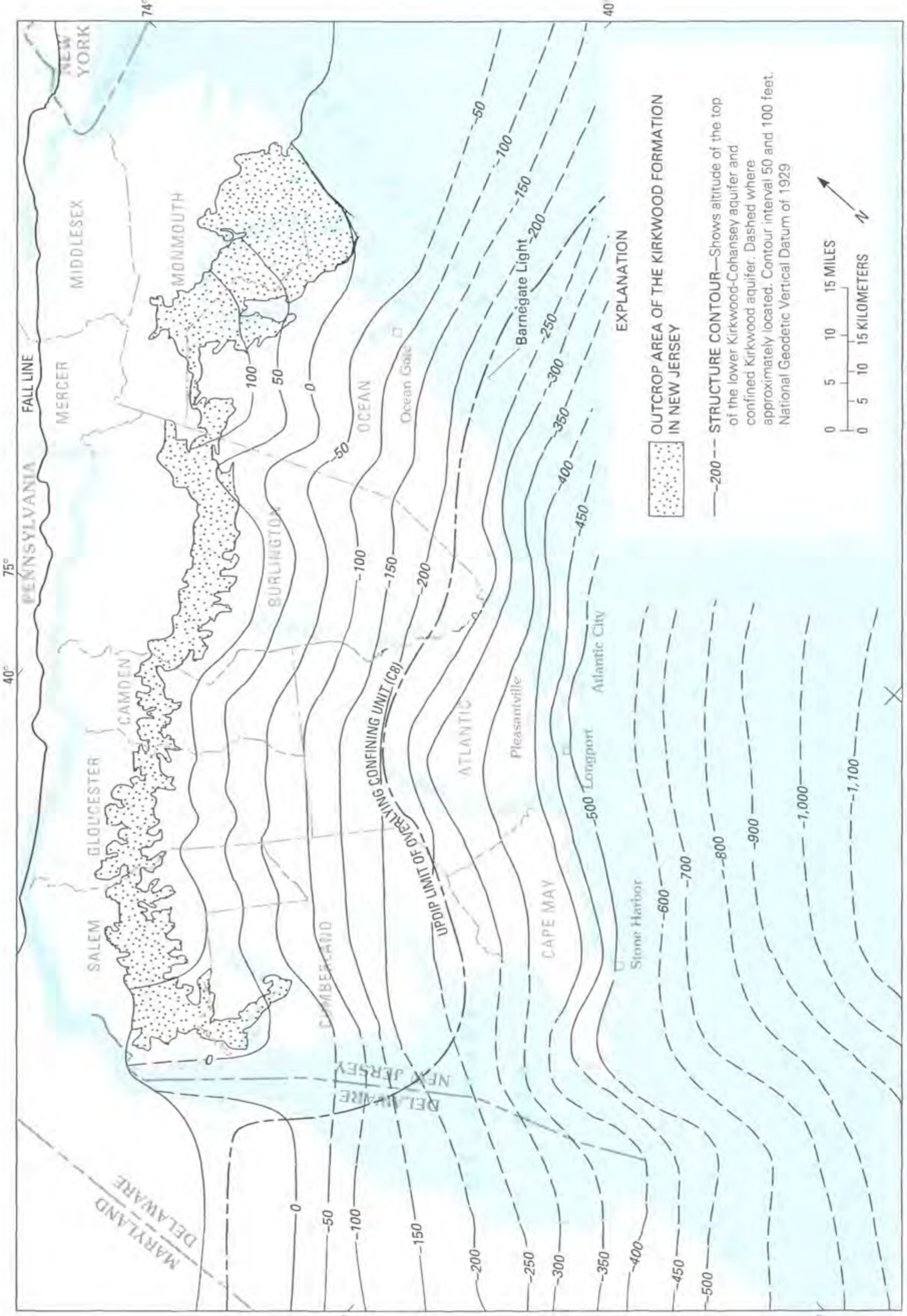


FIGURE 17.—Altitude of the top of the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (model unit A8).

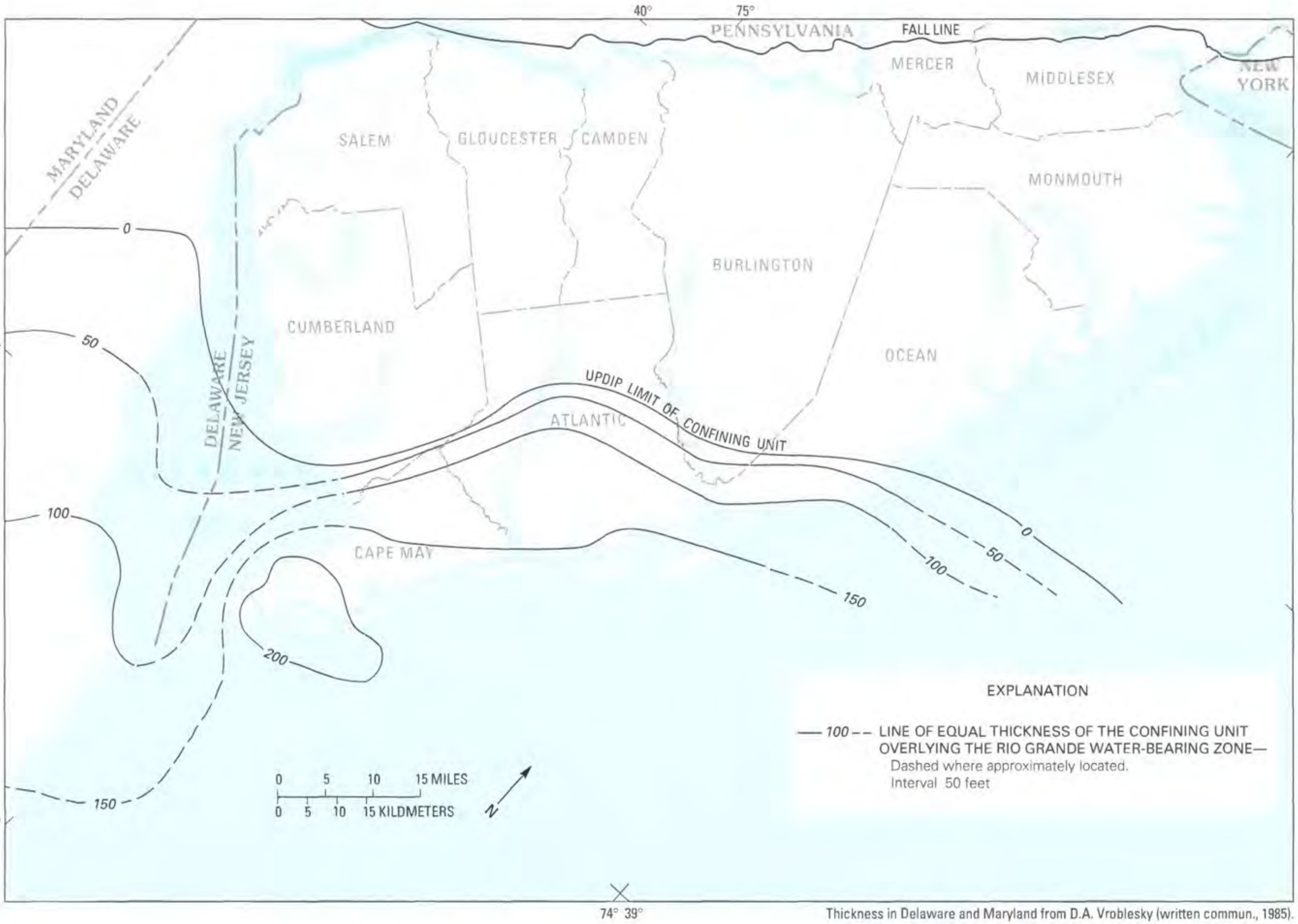


FIGURE 18.—Thickness of the confining unit overlying the Rio Grande water-bearing zone (model unit C8).

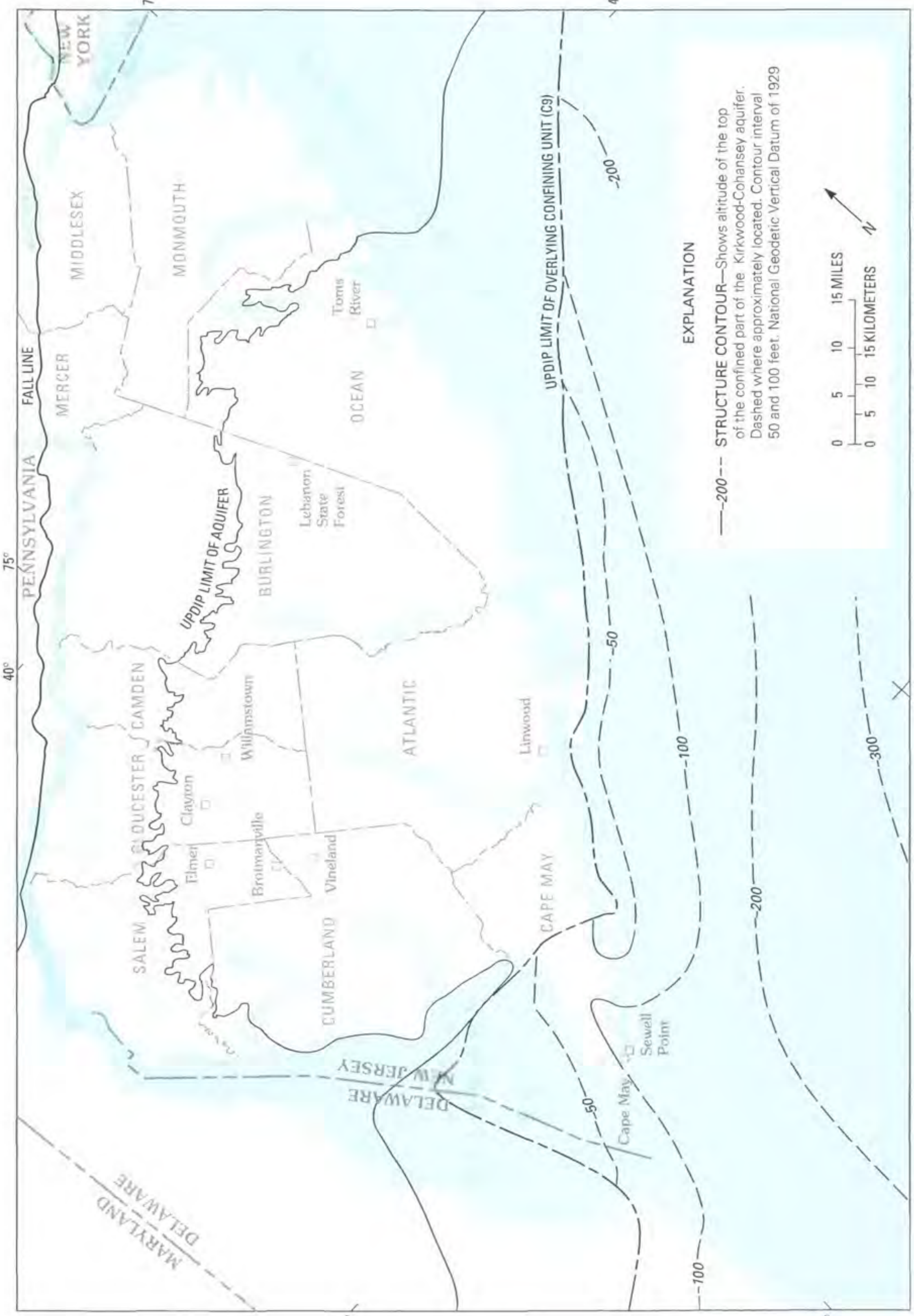


FIGURE 19.—Altitude of the top of the confined part of the upper Kirkwood-Cohansey aquifer (model unit A9).

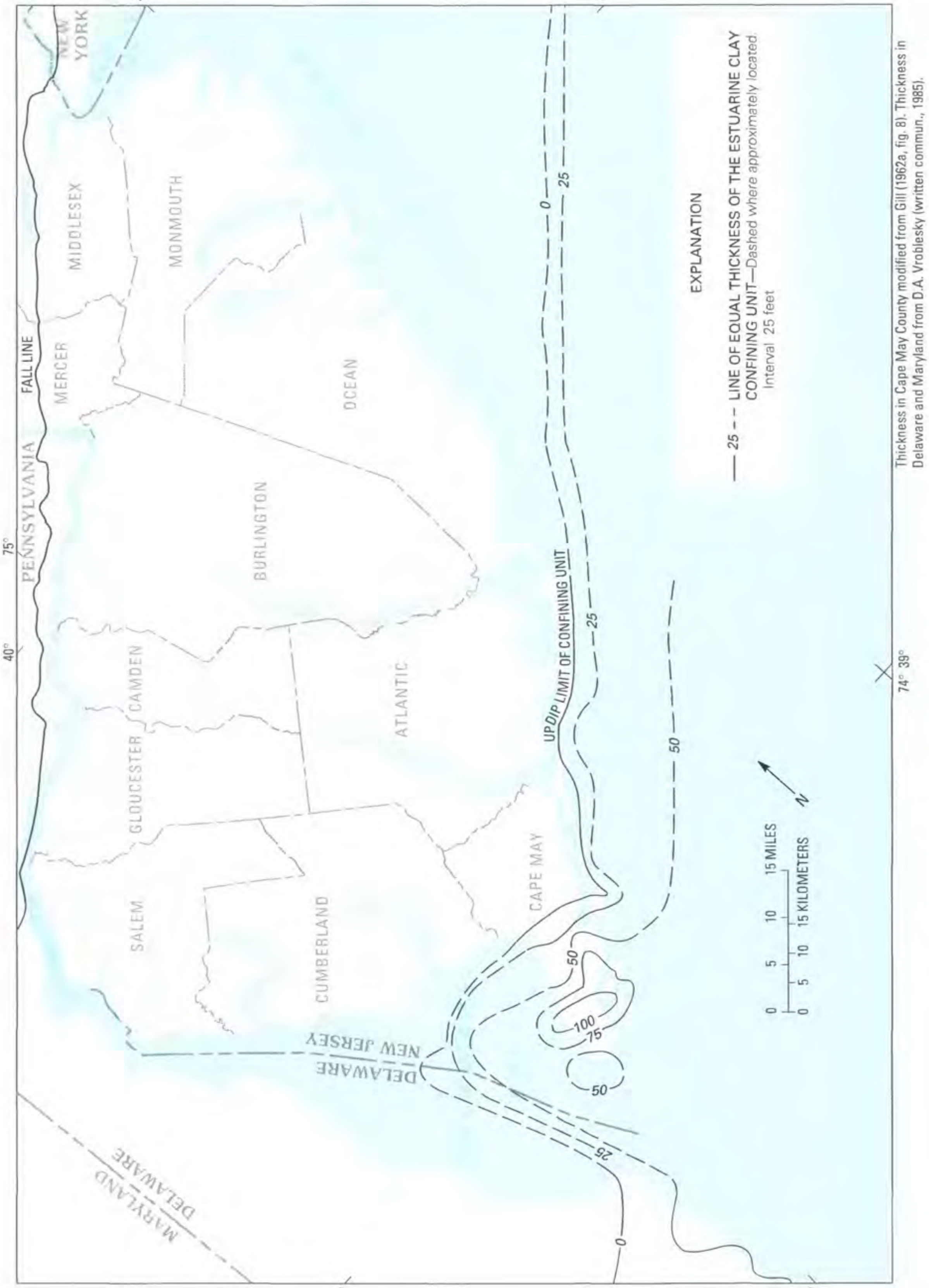


FIGURE 20.—Thickness of the estuarine clay confining unit (model unit C9).

TABLE 3.—*Summary of data on transmissivity, hydraulic conductivity, and storage coefficient for aquifers in the New Jersey Coastal Plain*
Location: Shown on maps of altitude of the top of the aquifers (figs. 3, 5, 7, 9, 11, 13, 15, 17, and 19).

Transmissivity (ft^2/d)	Hydraulic conductivity (ft/d)	Storage coefficient (dimensionless)	Type of data	Location	Reference
			LOWER POTOMAC-RARITAN-MAGOTHY AQUIFER (A1)		
2,300– 6,700	—	1.0×10^{-4} to 3.5×10^{-4}	Aquifer test	Camden, Camden County	Farlekas and others (1976, p. 38).
3,200– 3,700	—	3.3×10^{-5} to 1.5×10^{-3}	Aquifer test	Camden, Camden County	Farlekas and others (1976, p. 38).
[1]8,300	350	1.2×10^{-3}	Aquifer test	Camden, Camden County	Barksdale and others (1958, p. 97).
[2]16,600	240	1.0×10^{-3}	Aquifer test	Haddon Heights, Camden County.	Barksdale and others (1958, p. 97).
6,800– 9,100	140–190	9.0×10^{-5} to 1.7×10^{-4}	Aquifer test	Westville, Gloucester County.	Barksdale and others (1958, p. 97).
[3]6,000–35,000	—	8×10^{-5} to 8×10^{-3}	Model results	New Jersey Coastal Plain	Luzier (1980, p. 44).
			MIDDLE POTOMAC-RARITAN-MAGOTHY AQUIFER (A2)		
6,200–12,000	130–270	2.1×10^{-3}	Aquifer test	Burlington Township, Burlington County.	Rush (1968, p. 33).
22,000	200	6.0×10^{-2}	Aquifer test	Burlington, Burlington County.	Rush (1968, p. 33).
28,200–68,600	—	1.1×10^{-4} to 5.8×10^{-4}	Aquifer test	Palmyra, Burlington County	Rush (1968, p. 33).
13,100–17,400	217–290	1.0×10^{-4} to 2.4×10^{-4}	Aquifer test	Beverly, Burlington County	Rush (1968, p. 33).
20,000	200	1.5×10^{-4}	Aquifer test	Riverton, Burlington County	Barksdale and others (1958, p. 97).
6,300	200	1.5×10^{-4}	Aquifer test	Gibbstown, Gloucester County.	Barksdale and others (1958, p. 97).
[1]8,300	350	1.2×10^{-3}	Aquifer test	Camden, Camden County	Barksdale and others (1958, p. 97).
13,400	2,000	1.6×10^{-3}	Aquifer test	Parlin, Middlesex County	Barksdale and others (1958, p. 97).
6,700–10,200	79–119	3.7×10^{-5} and 8.6×10^{-5}	Aquifer test	Parlin, Middlesex County	Barksdale and others (1958, p. 97).
550– 1,900	32–88	4.0×10^{-5} to 8.1×10^{-2}	Aquifer test	Barber, Middlesex County	Barksdale and others (1958, p. 97).
[4]2,300– 9,000	—	5.8×10^{-4} to 2.4×10^{-3}	Aquifer test	Old Bridge, Middlesex County.	Barksdale and others (1958, p. 97).
42–16,800	105	1.6×10^{-4}	Model results	Southeastern Mercer, Middlesex, Monmouth, and northern Ocean Counties.	Farlekas (1979, p. 32 and 51).
[3]6,000–35,000	—	8×10^{-5} to 8×10^{-3}	Model results	New Jersey Coastal Plain	Luzier (1980, p. 44).
			UPPER POTOMAC-RARITAN-MAGOTHY AQUIFER (A3)		
500– 3,000	—	1.0×10^{-4}	Aquifer test	Delmarva Peninsula	Cushing and others (1973, p. 41).
[2]16,600	240	1.0×10^{-3}	Aquifer test	Haddon Heights, Camden County.	Barksdale and others (1958, p. 97).
[4]2,300– 9,000	—	5.8×10^{-4} to 2.4×10^{-3}	Aquifer test	Old Bridge, Middlesex County.	Barksdale and others (1958, p. 97).
[3]6,000–35,000	—	8×10^{-5} to 8×10^{-3}	Model results	New Jersey Coastal Plain	Luzier (1980, p. 44).
			ENGLISHTOWN AQUIFER (A4)		
2,100	—	2.7×10^{-4}	Aquifer tests	Clementon, Camden County	Farlekas and others (1976, p. 61).
1,300	27	1×10^{-4}	Aquifer tests	Monmouth County	Jablonski (1968, p. 53).
—	45–67	—	Laboratory tests.	Monmouth County	Jablonski (1968, p. 53).
1,100	12	7.6×10^{-5}	Aquifer test	Allenwood, Monmouth County.	Nichols (1977a, p. 25).
1,100	15	2.2×10^{-4}	Aquifer test	Lakewood, Ocean County	Nichols (1977a, p. 25).
400– 2,400	—	—	Model results	Monmouth, Ocean, and northern Burlington Counties.	Nichols (1977a, p. 26).

TABLE 3.—*Summary of data on transmissivity, hydraulic conductivity, and storage coefficient for aquifers in the New Jersey Coastal Plain*—Continued

Transmissivity (ft^2/d)	Hydraulic conductivity (ft/d)	Storage coefficient (dimensionless)	Type of data	Location	Reference
			WENONAH-MOUNT LAUREL AQUIFER (A5)		
360–1,430	17	7.0×10^{-5} to 2.1×10^{-4}	Aquifer test	Bradley Beach, Monmouth County.	Jablonski (1968, p. 62).
360–1,400	13–19	1.5×10^{-5} to 3.5×10^{-4}	Model results	Monmouth, Ocean, and northeastern Burlington Counties.	Nemickas (1976, p. 39).
1,200	13	3.5×10^{-4}	Aquifer test	Salem, Salem County	Rosenau and others (1969, p. 40).
940	19	—	Aquifer test	Artificial Island, Salem County.	Farlekas and others (1976, p. 70).
			VINCENTOWN AQUIFER (A6)		
530	21	—	Laboratory test.	Outcrop area between Jacobstown and New Egypt, Burlington County.	Rush (1968, p. 53).
			PINEY POINT AQUIFER (A7)		
1,200–6,000	—	2×10^{-4} to 4×10^{-4}	Aquifer tests	Delaware	Cushing, Kantrowitz, and Taylor (1973, p. 43).
4,100	—	3.0×10^{-4}	Aquifer test	Dover, Kent County, Delaware.	Leahy (1976, p. 10).
1,000–7,000	—	3.0×10^{-4}	Model results	Kent County, Delaware	Leahy (1979, p. 35 and 40).
1,400	23	3×10^{-4}	Aquifer test	Ancora, Camden County	Rush (1968, p. 56).
			LOWER KIRKWOOD-COHANSEY AQUIFER AND CONFINED KIRKWOOD AQUIFER (A8)		
5,200–5,900	42–48	1.2×10^{-4} to 2.3×10^{-4}	Aquifer test	Atlantic City, Atlantic County.	USGS unpublished data.
9,900–12,500	120–150	2.3×10^{-4} to 2.8×10^{-4}	Aquifer test	Pleasantville, Atlantic County.	Gill (1962a, p. 47).
8,800–9,600	108–120	2.6×10^{-4} to 2.7×10^{-4}	Aquifer test	Pleasantville, Atlantic County.	Gill (1962a, p. 47).
3,400–3,600	38–41	8.5×10^{-5} to 9.0×10^{-5}	Aquifer test	Stone Harbor, Cape May County.	Gill (1962a, p. 47).
6,700	—	3.0×10^{-4}	Statistical	Longport, Cape May County	Gill (1962a, p. 47).
1,500	—	6×10^{-4}	Aquifer test	Ocean Gate, Ocean County	Anderson and Appel (1969, p. 48).
			UPPER KIRKWOOD-COHANSEY AQUIFER (A9)		
16,000	130	4.2×10^{-4}	Aquifer test	Linwood, Atlantic County	Rhodehamel (1973, p. 55).
20,000	130	—	Aquifer test	Batsto, Burlington County	Rhodehamel (1973, p. 55).
12,000	120	—	Aquifer test	Lebanon State Forest, Burlington County.	Rhodehamel (1973, p. 55).
5,500–8,400	86–130	2.2×10^{-4} to 4.9×10^{-4}	Aquifer tests	Cape May City, Cape May County.	Gill (1962a, p. 49).
3,600–4,500	53–67	1.2×10^{-4} to 2.4×10^{-4}	Aquifer tests	Sewell's Point, Cape May County.	Gill (1962a, p. 49).
10,000	170	—	Aquifer test	Vineland, Cumberland County.	Rhodehamel (1973, p. 55).
4,000	130	1.0×10^{-3}	Aquifer test	Clayton, Gloucester County	Rhodehamel (1973, p. 55).
7,500	250	—	Aquifer test	Clayton, Gloucester County	Rhodehamel (1973, p. 55).
8,300	90	—	Aquifer test	Williamstown, Gloucester County.	Rhodehamel (1973, p. 55).
3,800	140	—	Aquifer test	Toms River, Ocean County	Rhodehamel (1973, p. 55).
4,300	150	3×10^{-4}	Aquifer test	Elmer, Salem County	Rhodehamel (1973, p. 55).
20,000	150	4.4×10^{-2}	Aquifer test	Brotmanville, Salem County	Rhodehamel (1973, p. 55).

[1] Model unit A1 or A2.
[2] Model unit A1 or A3.
[3] Model unit A1, A2, and A3 combined.
[4] Model unit A2 or A3.

TABLE 4.—*Summary of vertical hydraulic-conductivity data for confining units in the New Jersey Coastal Plain*

Location: Shown on confining unit thickness maps (figs. 6, 8, 10, and 12).

Geologic unit	Vertical hydraulic conductivity (ft/d)	Type of data	Location	Reference
CONFINING UNIT BETWEEN THE MIDDLE AND UPPER POTOMAC-RARITAN-MAGOTHY AQUIFERS (C2)				
Woodbridge Clay Member of the Raritan Foramtion.	3.6×10^{-2} to 8.6×10^{-6}	Model results	Southeastern Middlesex County to northern Monmouth County.	Farlekas (1979, p. 36).
Woodbridge Clay Member of the Raritan Formation.	8.4×10^{-2}	Aquifer test and model results.	South Brunswick Township, Middlesex County.	Farlekas (1979, p. 12).
MERCHANTVILLE-WOODBURY CONFINING UNIT (C3)				
Merchantville Formation	1.0×10^{-4} to 4.0×10^{-4}	Laboratory test.	Winslow Township, Camden County.	Farlekas and others (1976, p. 133–134).
Merchantville Formation and Woodbury Clay.	3.7×10^{-6} to 6.0×10^{-5}	Laboratory test.	Fort Dix, Burlington County.	Nichols (1977b, p. 58).
Merchantville Formation and Woodbury Clay.	3.6×10^{-6} to 1.4×10^{-5}	Laboratory test.	Lakewood, Ocean County	Nichols (1977a, p. 58).
Merchantville Formation and Woodbury Clay.	4.3×10^{-6}	Model results	Northern Coastal Plain New Jersey.	Nichols (1977a, p. 76).
Merchantville Formation and Woodbury Clay.	8.6×10^{-7} to 1.7×10^{-3}	Model results	New Jersey Coastal Plain	Luzier (1980, p. 29).
Woodbury Clay	1.0×10^{-4} to 3.0×10^{-2}	Laboratory test.	Winslow Township, Camden County.	Farlekas and others (1976, p. 133–134.
Englishtown Formation clayey silt lithofacies.	1.9×10^{-6}	Laboratory test.	Lakewood, Ocean County	Nichols (1977a, p. 58).
MARSHALLTOWN-WENONAH CONFINING UNIT (C4)				
Marshalltown Formation	2.6×10^{-4}	Laboratory test.	Fort Dix, Burlington County.	Nichols (1977a, p. 58).
Marshalltown Formation	1.3×10^{-1}	Laboratory test.	Winslow Township, Camden County.	Farlekas and others (1976, p. 133).
Marshalltown Formation	4.9×10^{-4}	Laboratory test.	Lakewood, Ocean County	Nichols (1977a, p. 58).
Marshalltown Formation and Wenonah Formation.	5.7×10^{-6} to 2.4×10^{-5}	Laboratory test.	Brick Township, Ocean County.	Nichols (1977b, p. 48).
Marshalltown Formation and Wenonah Formation.	1.5×10^{-5}	Model results	Northern Coastal Plain New Jersey.	Nichols (1977a, p. 76).
NAVESINK-HORNERSTOWN CONFINING UNIT (C5)				
Navesink Formation	2	Laboratory test.	Arneytown, Burlington County.	Rush (1968, p. 51).
Navesink Formation	5.0×10^{-4} to 1.3×10^{-1}	Laboratory test.	Winslow Township, Camden County.	Farlekes and others (1976, p. 133).
Navesink Formation	9	Laboratory test.	Sewell, Gloucester County	Rosenau and others (1969, p. 46).
Red Bank Sand	1.3×10^{-1}	Laboratory test.	Arneytown, Burlington County.	Rush (1968, p. 51).
Hornerstown Sand	3.0×10^{-3} and 2.0×10^{-2}	Laboratory test.	Arneytown, Burlington County.	Rush (1968, p. 52).
Hornerstown Sand	8.0×10^{-2} to 6.7×10^{-1}	Laboratory test.	Winslow Township, Camden County.	Farlekas and others (1976, p. 133).
Hornerstown Sand	4	Laboratory test.	Sewell, Gloucester County	Rosenau and others (1969, p. 46).
Navesink Formation and Hornerstown Sand.	5.6×10^{-2}	Aquifer test	Salem, Salem County	Rosenau and others (1969, p. 46).
BASAL KIRKWOOD CONFINING UNIT (C7)				
Alloway Clay Member of the Kirkwood Formation.	2.0×10^{-5} to 5.2×10^{-5}	Laboratory tests.	Cumberland County	Nemickas and Carswell (1976, p. 4).

CONFINING UNIT BETWEEN THE LOWER AND MIDDLE POTOMAC-RARITAN-MAGOTHY AQUIFERS

The confining unit above the lower aquifer of the Potomac-Raritan-Magothy aquifer system is a section of predominantly silt and clay sediments, with some sand, in the lower third of the undifferentiated Potomac-Raritan-Magothy aquifer system. In the model, this confining unit is referred to as the confining unit between the lower and middle Potomac-Raritan-Magothy aquifers and is designated as model unit C1. The thickness of this confining unit is shown in figure 4. Confining-unit thickness data near the Delaware River in northwestern Salem, Gloucester, and Camden Counties, are from Zapecza (1984, pl. 6). In downdip areas in the eastern Coastal Plain, the confining-unit thickness is estimated from available borehole geophysical logs. The thickness of this confining unit ranges from less than 50 feet near the Delaware River to 500 feet in downdip areas near the Delaware River in Salem County. The modeled confining unit has the same areal extent as the lower Potomac-Raritan-Magothy aquifer (A1) and is overlain everywhere by the middle Potomac-Raritan-Magothy aquifer (A2).

No previous estimates of vertical hydraulic conductivity have been made for the confining-unit sediments between the lower and middle aquifers of the Potomac-Raritan-Magothy aquifer system in the New Jersey Coastal Plain. Many previous studies (Barksdale and others, 1958; Farlekas and others, 1976; Luzier, 1980; and Walker, 1983) have discussed various aspects of the hydrogeology of the lower aquifer of Potomac-Raritan-Magothy aquifer system in combination with the middle or middle and upper aquifers of the Potomac-Raritan-Magothy aquifer system. These studies suggest that the vertical hydraulic connection within the Potomac-Raritan-Magothy aquifer system is relatively high and that the hydraulic conductivity of the confining unit between the lower and middle aquifers of the Potomac-Raritan-Magothy aquifer system is somewhat higher than the hydraulic conductivity of the Merchantville-Woodbury confining unit that overlies the upper aquifer of the Potomac-Raritan-Magothy aquifer system. However, similarity in water levels in the lower and middle parts of the aquifer system compared to the upper part of the aquifer system may result from the distribution of ground-water withdrawals rather than from the degree of hydraulic connection.

The hydraulic conductivity of the confining unit between the lower and middle Potomac-Raritan-Magothy aquifers is assumed to be similar to (and possibly higher than) hydraulic conductivity estimates for the confining unit between the middle and upper Potomac-Raritan-Magothy aquifers and as much as two orders of magnitude more than the hydraulic conductivity of the Merchantville-Woodbury confining unit.

MIDDLE POTOMAC-RARITAN-MAGOTHY AQUIFER

The middle Potomac-Raritan-Magothy aquifer completely overlies the confining unit between the lower and middle Potomac-Raritan-Magothy aquifers. In areas where the underlying aquifer and confining unit, model units A1 and C1, do not exist, the middle aquifer overlies bedrock, weathered bedrock, or clays. The modeled aquifer, model unit A2, represents the middle aquifer of the Potomac-Raritan-Magothy aquifer system, which is equivalent to the Farrington aquifer (the Farrington Sand Member of the Raritan Formation) in the northeastern New Jersey Coastal Plain.

Structure contours of the top of the middle Potomac-Raritan-Magothy aquifer are shown in figure 5. The altitude on the top of the middle aquifer ranges from land surface in the outcrop area near the Fall Line to more than an estimated 2,800 feet below sea level in downdip seaward areas east of Atlantic County. The altitude data of the top of the aquifer near the Delaware River and in northwestern Monmouth and Ocean Counties are from Zapecza (1984, pl. 7). The top of the middle aquifer in downdip undifferentiated areas in the southern and southwestern Coastal Plain is estimated from available borehole geophysical logs. Generally, this aquifer is the sandiest part of the middle third of the undifferentiated Potomac-Raritan-Magothy aquifer system. The middle aquifer extends from the Fall Line to some unknown distance downdip beneath the Atlantic Ocean. However, the modeled freshwater part of the middle aquifer is shown to extend from the Fall Line to southern Cumberland and southern Atlantic and Ocean Counties but not beneath most of Cape May County. The thickness of the middle aquifer near the Delaware River ranges from 50 feet to 150 feet and is more than 600 feet in downdip areas offshore of Ocean County.

The middle Potomac-Raritan-Magothy aquifer is continuous into Delaware where it is the sandy part of the upper half of the Potomac Formation (D.A. Vroblesky, written commun., 1985). The middle aquifer is equivalent to the Lloyd aquifer (the Lloyd Sand Member of the Raritan Formation) on Long Island, N.Y. (Henry Trapp, written commun., 1985).

Aquifer-test analyses (table 3) indicate that the transmissivity of the middle aquifer ranges from 550 to 22,000 ft^2/d; however, one aquifer test in Burlington County (Rush, 1968, p. 33) shows unrealistic transmissivity values over 68,000 ft^2/d. Most of the aquifer-test analyses are from the northern Coastal Plain and give transmissivity values that are higher than the few analyses for the lower aquifer. Storage coefficient estimates from these tests range from 3.7×10^{-5} to 8.1×10^{-3} and are similar to storage coefficient estimates for the lower aquifer.

CONFINING UNIT BETWEEN THE MIDDLE AND UPPER POTOMAC-RARITAN-MAGOTHY AQUIFERS

The confining unit between the middle and upper Potomac-Raritan-Magothy aquifers is designated C2 in the model. This confining unit generally represents the clayey part within the middle third of the undifferentiated Potomac-Raritan-Magothy aquifer system and overlies the middle aquifer everywhere except where the middle aquifer crops out near the Fall Line. The thickness of the middle confining unit is shown in figure 6. Near the Fall Line, the thickness is generally less than 50 feet (Zapecza, 1984, pl. 9). In downdip areas, the thickness was estimated on the basis of available geophysical logs. The confining unit generally thickens downdip and is more than 600 feet near the 10,000 mg/L chloride line.

Estimates of vertical hydraulic conductivity for the confining unit range from 8.6×10^{-6} to 8.4×10^{-2} feet per day (ft/d) (table 4). These estimates are derived from a flow modeling study of the underlying aquifer in the northeastern Coastal Plain by Farlekas (1979) and suggest that the vertical hydraulic conductivity varies several orders of magnitude locally.

UPPER POTOMAC-RARITAN-MAGOTHY AQUIFER

The upper Potomac-Raritan-Magothy aquifer overlies the confining unit between the middle and upper aquifers everywhere except where the confining unit crops out near the Fall Line. This aquifer, designated A3 in the model, is the upper aquifer of the Potomac-Raritan-Magothy aquifer as described by Zapecza (1984, p. 18–19) and is primarily the Magothy Formation in New Jersey but is the Old Bridge aquifer (the Old Bridge Sand Member of the Magothy Formation) in the northeastern New Jersey Coastal Plain. The upper aquifer, like the middle aquifer, extends from its outcrop area near the Fall Line to some unknown distance downdip beneath the Atlantic Ocean. The downdip area of the upper aquifer simulated by the model, however, is limited to the area with freshwater. The modeled extent of the upper aquifer is greater than that of the lower and middle aquifers, because freshwater occurs offshore in the upper aquifer. The upper aquifer is present in each county of the New Jersey Coastal Plain.

The altitude of the top of the aquifer (fig. 7) ranges from land surface in the outcrop area to more than 2,200 feet below sea level in downdip areas south of Cape May County. The altitude of the top of the upper aquifer is that shown by Zapecza (1984, pl. 10). The altitude in offshore areas where no borehole geophysical logs are available was estimated on the basis of the onshore structural trends. The thickness of the upper aquifer of the Potomac-Raritan-Magothy aquifer system is shown by Zapecza (1984, p. 11). The aquifer is thinnest near the outcrop area, where it is less than 50 feet thick and near the Delaware Bay where it is about 75 feet thick. The aquifer thickens to more than 200 feet in eastern Monmouth County.

The upper aquifer is continuous with the Magothy Formation of Delaware (D.A. Vroblesky, written commun., 1985) and is modeled in the northern Atlantic Coastal Plain RASA study as the approximate lower third of the Magothy Formation in New York (P.P. Leahy, written commun., 1985). Except in the outcrop area, the upper aquifer is overlain everywhere by the Merchantville-Woodbury confining unit.

Two aquifer tests in New Jersey are summarized in table 3 for the upper aquifer. Although these tests are listed for the upper aquifer, this cannot be confirmed because of incomplete well records. There is no evidence to suggest that the hydraulic conductivity or lithology of the upper aquifer of the Potomac-Raritan-Magothy aquifer system differs greatly from the hydraulic conductivity and lithology of the lower or middle aquifers of the Potomac-Raritan-Magothy aquifer system. Differences in transmissivity among the three aquifers result primarily from differences in thickness.

MERCHANTVILLE-WOODBURY CONFINING UNIT

The Merchantville-Woodbury confining unit generally represents the sediments of the Merchantville-Woodbury confining unit described by Zapecza (1984, p. 19–20). However, several modifications were made to incorporate the unit into the model framework as model unit C3. In areas where the overlying Englishtown Formation includes an upper and lower sand unit within the aquifer (Nichols, 1977b, p. 15), the confining unit between these sands is specified for this report to be part of the Merchantville-Woodbury confining unit. The thickness of model unit C3 (fig. 8) is, therefore, slightly greater than the thickness of the Merchantville-Woodbury confining unit shown by Zapecza (1984, pl. 12) in Ocean, southeastern Monmouth, and southeastern Burlington Counties. Also, in areas southeast of the downdip limit of the Englishtown aquifer system, Zapecza (1984, p. 19) includes all sediments between the overlying Wenonah-Mount Laurel aquifer and the underlying Potomac-Raritan-Magothy aquifer system in the Merchantville-Woodbury confining unit. Model unit C3, however, includes only the lower part of these sediments in these areas. The upper part is assigned to model unit C4. The thickness of the confining unit in Cumberland, Cape May, Atlantic, and southern Burlington and Ocean Counties is, therefore, less than that shown by Zapecza (1984, pl. 12). The sediments represented by model unit C3 are illustrated in figure 2 and table 2. The modeled confining unit increases in thickness downdip, from less than 50 feet at the outcrop to more than 400 feet along the coast of Ocean County.

Estimates of vertical hydraulic conductivity for sediments of the Merchantville-Woodbury confining unit have been made from model simulation studies and laboratory analyses (table 4). Estimates range from 8.6×10^{-7} to 3.0×10^{-2} ft/d. Luzier (1980, p. 29) shows hydraulic conductivity decreasing in the downdip direction. The sediments of the Merchantville-Woodbury confining unit have long been recognized (Barksdale and others, 1958, p. 135–136) as one of the most effective confining units in the New Jersey Coastal Plain.

ENGLISHTOWN AQUIFER

The Englishtown aquifer, designated A4 in the model, overlies the Merchantville-Woodbury confining unit but has a more limited areal extent. Model unit A4 represents both sand units of the Englishtown aquifer system described by Zapecza (1984, p. 20–22). The altitude of the top of the Englishtown aquifer is shown on figure 9. The aquifer extends from its outcrop area several miles southeast of the Fall Line to northern Cumberland, northern Atlantic, southern Burlington, and southern Ocean Counties. The downdip limit of this model unit is at a facies change to silt and clay and is the downdip limit of the permeable sand. The Englishtown aquifer contains freshwater everywhere. The altitude of the top of the aquifer ranges from land surface in the outcrop area to more than 1,600 feet below sea level offshore of Ocean County. The altitudes shown in figure 9 are those shown by Zapecza (1984, pl. 13), with projected altitudes in offshore areas.

The thickness of the modeled aquifer is the same as the thickness of the Englishtown aquifer system shown by Zapecza (1984, pl. 14). The aquifer is generally less than 100 feet thick. However, the Englishtown aquifer has its greatest thickness of 200 feet in southern Monmouth and northeastern Ocean Counties, where two distinct permeable sand units are present. The thickness of the upper sand unit, which has the greatest withdrawals, is less than 120 feet in this area.

The Englishtown aquifer in New Jersey is continuous with the Englishtown Formation in Delaware (D.A. Vroblesky, written commun., 1985) and is modeled as the approximate middle third of the Magothy Formation in New York in the hydrogeologic framework of the northern Atlantic Coastal Plain RASA model (P.P. Leahy, written commun., 1985). The Englishtown aquifer is overlain everywhere except in its outcrop area by the Marshalltown-Wenonah confining unit.

Transmissivity estimates for the Englishtown aquifer derived from model simulations range from 400 to 2,400 ft^2/d (table 3). Transmissivity estimates from aquifer-test analyses are comparable and range from 1,100 to 2,100 ft^2/d. These estimates are considerably lower than transmissivity estimates for the Potomac-Raritan-Magothy aquifers. These lower transmissivities are related to the low hydraulic conductivity and thickness of the Englishtown aquifer. Estimates of the storage coefficient range from 7.6×10^{-5} to 2.7×10^{-4} (table 3).

MARSHALLTOWN-WENONAH CONFINING UNIT

In areas overlying the Englishtown aquifer, the Marshalltown-Wenonah confining unit represents the sediments of the Marshalltown-Wenonah confining unit described by Zapecza (1984, p. 22–23). In areas southeast of the downdip extent of the Englishtown aquifer, this unit is part of the Merchantville-Woodbury confining unit described by Zapecza (1984, p. 19–20). The sediments represented by model unit C4 are shown in table 2 and figure 2. This confining unit is relatively thin, ranging from less than 20 feet to less than 100 feet, and has the same downdip extent as the Merchantville-Woodbury confining unit (C3) (fig. 10). Therefore, southeast of the downdip limit of the Englishtown aquifer (A4), this confining unit directly overlies the Merchantville-Woodbury confining unit (C3).

Estimates of vertical hydraulic conductivity for sediments of the Marshalltown-Wenonah confining unit range from about 5.7×10^{-6} to 4.9×10^{-4} ft/d (table 4). However, one laboratory analysis of cores from Camden County indicated an unusually high vertical hydraulic conductivity of 1.3×10^{-1} ft/d.

WENONAH-MOUNT LAUREL AQUIFER

The Wenonah-Mount Laurel aquifer, designated A5 in the model, overlies the Marshalltown-Wenonah confining unit (C4). The altitude data of the top of the Wenonah-Mount Laurel aquifer (fig. 11) is from Zapecza (1984, pl. 16). The aquifer extends from its outcrop area, several miles southeast of the Fall Line, about 50 to 60 miles downdip, where there is a facies change from sand to silt and clay. The altitude of the top of the aquifer ranges from land surface in its outcrop area and is estimated to be more than 2,400 feet below sea level beneath the Atlantic Ocean. The aquifer is thickest in southeastern Gloucester and Salem Counties and northwestern Cumberland County where it is over 120 feet thick (Zapecza, 1984, pl. 17). The aquifer generally thins downdip and is less than 40 feet thick in southern Cape May County.

The Wenonah-Mount Laurel aquifer is equivalent to permeable sands in the Monmouth Formation in Delaware (D.A. Vroblesky, written commun., 1985) and is modeled as the upper part of the Magothy Formation in New York by the regional RASA study (P.P. Leahy, written commun., 1985).

Estimates of transmissivity from aquifer tests and model simulations range from 360 to 1,430 ft^2/d (table 3). Storage coefficients were estimated to be from 1.5×10^{-5} to 3.5×10^{-4} (table 3). The hydraulic characteristics of the Wenonah-Mount Laurel aquifer are similar to those of the Englishtown aquifer.

NAVESINK-HORNERSTOWN CONFINING UNIT

The Navesink-Hornerstown confining unit, designated C5 in the model, overlies the Wenonah-Mount Laurel aquifer (A5) everywhere except in the aquifer's outcrop area. Model unit C5 represents the lower part of the composite confining unit described by Zapecza (1984, p. 24–25). The unit includes the Navesink Formation, Red Bank Sand, Tinton Sand, and the Hornerstown Sand. These minor aquifers are relatively thin and limited in their areal extent.

Thickness of the Navesink-Hornerstown confining unit is shown in figure 12. Throughout most of the New Jersey Coastal Plain, the confining unit is less than 80 feet thick. However, in central Monmouth County, the unit is 100 to 180 feet thick because of the presence of the Red Bank and Tinton Sands.

Estimates of vertical hydraulic conductivity for sediments within the Navesink-Hornerstown confining unit range from 5.0×10^{-4} to 9 ft/d (table 4). Many of the estimates are several orders of magnitude higher than for sediments in other confining units. Generally, these relatively high vertical hydraulic conductivities are for minor sand layers within the confining unit and are not representative of the confining unit's overall vertical hydraulic conductivity.

VINCENTOWN AQUIFER

The Vincentown Formation overlies the Navesink-Hornerstown confining unit. However, throughout most of its extent the formation is a confining unit (Zapecza, 1984, p. 25). The Vincentown Formation is an aquifer in its outcrop area and for about 8 to 10 miles downdip. The aquifer part of the Vincentown Formation is designated A6 in the model and appears in Salem, Gloucester, Camden, Burlington, Ocean, and Monmouth Counties (fig. 13). The aquifer is not present in New York. In Delaware, the Vincentown aquifer consists of permeable sands in the Rancocas Group (D.A. Vroblesky, written commun., 1985).

The altitude of the top of the aquifer is shown in figure 13 and is the same as that shown by Zapecza (1984, pl. 19). The altitude of the top of the aquifer ranges from altitudes of more than 100 feet above sea level in the outcrop to more than 200 feet below sea level downdip in Salem County. The thickness of the Vincentown aquifer is shown by Zapecza (1984, pl. 19). Generally, the aquifer is less than 100 feet thick. However, thicknesses of 140 feet occur in central Monmouth County.

Only one estimate of transmissivity is available for the Vincentown aquifer. A laboratory analysis of sediments from Burlington County resulted in an estimated transmissivity of 530 ft^2/d (table 3).

VINCENTOWN-MANASQUAN CONFINING UNIT

The Vincentown aquifer is overlain everywhere, except in the outcrop area, by the Manasquan Formation. The Manasquan-Vincentown confining unit, designated C6 in the model, includes the downdip silt and clay portions of the Vincentown Formation, the Manasquan Formation, and, locally, the overlying Shark River Formation. These sediments are part of the composite confining unit described by Zapecza (1984, p. 24–25).

The thickness of the Vincentown-Manasquan confining unit, shown in figure 14, ranges from about 50 feet at the outcrop area to about 1,000 feet in southern Cape May County. This unit has the same downdip extent as the Navesink-Hornerstown confining unit (C5). Therefore, in areas southeast of the downdip extent of the underlying Vincentown aquifer (A6), the Vincentown-Manasquan confining unit (C6) directly overlies the Navesink-Hornerstown confining unit (C5). This relation is shown in figure 2 and table 2. No estimates of vertical hydraulic conductivity are available for the Vincentown-Manasquan confining unit.

PINEY POINT AQUIFER

The Piney Point aquifer described by Zapecza (1984, p. 26–29) is designated as model unit A7. The altitude of the top of the Piney Point aquifer is shown in figure 15. The aquifer does not crop out because it is overlain entirely by the basal clay of the Kirkwood Formation. The updip limit of the Piney Point aquifer is generally at the downdip limit of the Vincentown aquifer (A6) (figs. 2, 13, and 15). The Piney Point aquifer probably thins and is not present several miles downdip of the Atlantic Coast. The altitude of the top of the aquifer is the same as that shown by Zapecza (1984, pl. 20), with altitudes beneath the ocean estimated from onshore contour trends and available offshore data. The altitude of the top of the aquifer ranges from about 50 feet below sea level in southeastern Salem and Gloucester Counties to about

1,300 feet below sea level beneath the Atlantic Ocean southeast of Atlantic County.

The thickness of the Piney Point aquifer is shown by Zapecza (1984, pl. 21). The aquifer is thickest in western Cumberland County where it is over 200 feet thick. The aquifer thins to the northeast where it is less than 40 feet thick in Atlantic County but thickens again to over 120 feet in Burlington and Ocean Counties. The Piney Point aquifer is not present in New York but is continuous into Delaware (D.A. Vroblesky, written commun., 1985). The only estimate of transmissivity of the Piney Point aquifer in New Jersey based on aquifer-test data is 1,400 ft^2/d. Estimates of transmissivity and storage coefficient for the aquifer in Delaware are shown in table 3.

BASAL KIRKWOOD CONFINING UNIT

The basal Kirkwood confining unit, designated C7 in the model, represents the basal clay of the Kirkwood Formation and, locally, silty parts of the Piney Point Formation. These sediments are the uppermost part of the composite confining unit described by Zapecza (1984, p. 24–25). The confining unit completely overlies the Piney Point aquifer (A7) and extends several miles northwest of the aquifer's updip limit. Northwest of the Piney Point aquifer the basal Kirkwood confining unit overlies the Vincentown-Manasquan confining unit (C6).

Thickness of the basal Kirkwood confining unit is shown in figure 16. Generally, the unit thickens downdip. The unit is over 140 feet thick in southern Cape May County, but its greatest thickness exceeds 160 feet in eastern Ocean County and just offshore.

Only one estimate of vertical hydraulic conductivity is available for sediments of the basal Kirkwood confining unit. A laboratory analysis of the Alloway Clay Member of the Kirkwood Formation, part of the confining unit, indicated a vertical hydraulic conductivity ranging from 2.0×10^{-5} to 5.2×10^{-5} ft/d (table 4).

LOWER KIRKWOOD-COHANSEY AQUIFER AND CONFINED KIRKWOOD AQUIFER

The lower Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer are designated as A8 in the model. In downdip areas, the modeled aquifer represents the Atlantic City 800-foot sand and the overlying, relatively minor, Rio Grande water-bearing zone. In this report, the Atlantic City 800-foot sand and the Rio Grande water-bearing zone are together referred to as the confined Kirkwood aquifer. In updip areas, the modeled aquifer represents the lower part of the unconfined Kirkwood-Cohansey aquifer system. The Atlantic City 800-foot sand and the Kirkwood-Cohansey aquifer system are described and mapped by Zapecza (1984, p. 29–34, pls. 22, 23, and 24), but Zapecza does not subdivide the Kirkwood-Cohansey aquifer system into an upper and lower part. He also describes, but does not map, the Rio Grande water-bearing zone (Zapecza, 1984, p. 31–32). The relation between the hydrogeologic units and the modeled aquifer are shown in figure 2 and table 2.

The Kirkwood-Cohansey aquifer system was subdivided in this study into an upper and lower aquifer in updip areas to better represent the vertical head distribution in the unconfined aquifer system and to provide a lateral connection between the confined Kirkwood aquifer and the lower Kirkwood-Cohansey aquifer. Zapecza (1984, p. 29) states that the Atlantic City 800-foot sand has not been mapped beyond its overlying confining unit, and its connection with the unconfined aquifer is not known.

The altitude of the top of the modeled aquifer and the approximate extent of the confined Kirkwood aquifer are shown in figure 17. The updip limit of the confined Kirkwood aquifer extends through southeastern Cumberland, central Atlantic, southeastern Burlington, and southern Ocean Counties and is the same as the limit of the overlying confining unit. This line divides the modeled aquifer into the lower part of the unconfined aquifer in updip areas to the northwest and a confined aquifer in downdip areas to the southeast. Southeast of this line the altitude of the top of the modeled aquifer is the top of the Rio Grande water-bearing zone. In this area, the altitude of the top is 50 to 250 feet above the top of the Atlantic City 800-foot sand shown by Zapecza (1984, pl. 22). Northeast of this line the top of the modeled aquifer continues to rise until it reaches the water table. The altitude of the top of the modeled aquifer (fig. 17) ranges from the water table in its outcrop area, several miles southeast of the Fall Line, to 500 feet below sea level in southern Cape May County and is estimated to be more than 1,100 feet below sea level at the southeast model boundary.

The thickness of the modeled aquifer ranges from about 50 feet in its outcrop area to about 200 feet in Cape May County. The lower Kirkwood-Cohansey aquifer is the updip part of the aquifer and is approximately the lower third of the Kirkwood-Cohansey aquifer system described by Zapecza (1984, p. 32–34). The thickness of the confined Kirkwood aquifer, which is the downdip part of the modeled aquifer, is the combined thickness of the Atlantic City 800-foot sand, the Rio Grande water-bearing zone, and the intervening confining unit.

The confined Kirkwood aquifer extends to the southeast model boundary. However, the aquifer probably contains freshwater for at least 25 miles offshore beyond the model boundary. The unit thins toward the northeast and is not present in Long Island, N.Y. (Henry Trapp, written commun., 1985). To the southwest, the confining units within the modeled aquifer become thicker, and the aquifer includes the Frederica and Cheswold aquifers in Delaware (D.A. Vroblesky, written commun., 1985).

Estimates of transmissivity of the modeled aquifer from Atlantic, Cape May, and Ocean Counties range from 1,500 to 12,500 ft^2/d (table 3). Storage coefficient estimates range from 8.5×10^{-5} to 6.0×10^{-4}. Transmissivity estimates for the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (A8) are lower than transmissivity estimates for the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) but are slightly higher than estimates for the other underlying aquifers (A4–A7).

CONFINING UNIT OVERLYING THE RIO GRANDE WATER-BEARING ZONE

Model unit C8 represents the confining unit overlying the Rio Grande water-bearing zone and is designated as C8 in the model. This confining unit is the upper part of the confining unit overlying the Atlantic City 800-foot sand described by Zapecza (1984, p. 31). The confining unit completely overlies the confined Kirkwood aquifer, does not crop out, and is present mainly in southern Ocean, southeastern Atlantic, and Cape May Counties (fig. 18).

The thickness of the confining unit is shown in figure 18. The unit thickens downdip, ranging from 50 feet near its updip limit to 200 feet in southern Cape May County. The unit is 100 to 200 feet thinner than the confining unit mapped by Zapecza (1984, pl. 22) because he includes the thickness of the Rio Grande water-bearing zone and the confining unit underlying this zone. No estimates of vertical hydraulic conductivity are available.

UPPER KIRKWOOD-COHANSEY AQUIFER

The sediments in the upper part of the unconfined Kirkwood-Cohansey aquifer system are designated as model unit A9 and are referred to in the model as the upper Kirkwood-Cohansey aquifer. This modeled aquifer directly overlies the lower Kirkwood-Cohansey aquifer (updip part of A8) in areas northwest of the updip limit of the confined Kirkwood aquifer. Southeast of the limit, the upper Kirkwood-Cohansey aquifer represents unconfined and confined sediments of the Kirkwood Formation above the confining unit overlying the Rio Grande water-bearing zone (C8). Throughout most of its onshore extent, the modeled aquifer is unconfined; however, in the peninsular part of Cape May County and offshore the aquifer is confined by the overlying estuarine clay facies of the Cape May Formation (Gill, 1962a, fig. 2).

The altitude of the top of the confined part of the upper Kirkwood-Cohansey aquifer is shown in figure 19. Although the top of the sediments is the altitude of land surface, the altitude of the top of the saturated sediments (water table) is estimated to be at or near land surface near streams and surface-water bodies and up to several tens of feet below land surface near local ground-water divides. The altitude of the top of the confined part of the upper Kirkwood-Cohansey aquifer in Cape May County is similar to that shown by Gill (1962a, fig. 6) for the top of the estuarine sand facies. The altitude of the top of the aquifer in offshore areas is the ocean and bay bottoms.

The thickness of the unconfined Kirkwood-Cohansey aquifer system, which includes the unconfined parts of the lower Kirkwood-Cohansey aquifer (updip part of A8) and the upper Kirkwood-Cohansey aquifer (A9), is described and shown by Zapecza (1984, pl. 24 and p. 32–34). The confined part of the upper Kirkwood-Cohansey aquifer is from 150 to 250 feet thick in Cape May County. Estimates of transmissivity for the upper Kirkwood-Cohansey aquifer from aquifer-test analyses range from 3,600 to 20,000 ft^2/d (table 3). The higher estimates are generally for the thicker parts of the unconfined part of the aquifer. Estimates of storage coefficient from these tests range from 1.2×10^{-4} to 4.4×10^{-2}.

ESTUARINE CLAY CONFINING UNIT

The estuarine clay confining unit, designated C9 in the model, represents the estuarine clay facies described by Gill (1962a, p. 25–26). Onshore, this confining unit is present only in the peninsular part of Cape May County where the upper Kirkwood-Cohansey aquifer (A9) is confined. Offshore, the estuarine clay confining unit extends beneath Delaware Bay and the Atlantic Ocean.

The thickness of the estuarine clay confining unit is shown in figure 20. The unit has a fairly uniform thickness of about 50 feet, although locally in southern Cape May County the unit is over 100 feet thick. However, contours in offshore areas are highly approximate. There are no estimates of vertical hydraulic conductivity for this confining unit.

HOLLY BEACH AQUIFER

The Holly Beach aquifer, designated A10 in the model, represents part of the unconfined Holly Beach water-bearing zone described by Gill (1962a, p. 41). The modeled aquifer is the Holly Beach water-bearing zone only on the peninsular part of Cape May County where it is underlain by the estuarine clay facies of the Cape May Formation mapped by Gill (1962a, fig. 8). In the northern part of the county where the clay facies is absent, sediments of the Holly Beach water-bearing zone are combined with the upper Kirkwood-Cohansey aquifer (A9). The lithology and stratigraphy of the sediments comprising this aquifer are discussed in detail by Gill

(1962a, p. 21–32). The aquifer does not extend offshore and is not continuous into Delaware.

The altitude of the top of the sediments comprising the Holly Beach aquifer is at land surface, which ranges from about 25 feet above sea level in the central Cape May Peninsula to sea level along the shore. These sediments are about 30 to 80 feet thick. However, the thickness and altitude of the top of the aquifer are generally taken to refer only to the saturated sediments. Therefore, based on subtracting the altitude of the top of the estuarine clay facies of the Cape May Formation (Gill, 1962a, fig. 7) from the water table (Gill, 1962a, fig. 46), the altitude of the saturated sediments ranges from about 15 feet above sea level to near sea level and the thickness ranges from 30 to 70 feet. No estimates of transmissivity or of the storage coefficient of this unit have been made.

GROUND-WATER FLOW SYSTEM

The ground-water flow system is defined by the flow patterns and rates within the aquifers. The Coastal Plain flow system is dynamic and has changed greatly from prepumping conditions. Movement of ground water in the modeled area under prepumping conditions was controlled by several factors: (1) hydraulic properties of the saturated sediments, (2) topography, and (3) recharge and discharge to and from the aquifers at the model boundaries. After pumping began, ground-water flow was affected by the location and amount of withdrawals. Other factors affecting ground-water flow, such as seasonal or yearly changes in areal recharge or changes in recharge caused by changes in land use, are considered short-term effects; therefore, they are not discussed in this report.

Coastal Plain aquifers generally transmit water easily compared to confining units, which transmit water only slightly. Flow within the aquifers is predominantly horizontal, although some vertical flow exists. Confining units generally have very large vertical hydraulic gradients, and flow is mostly vertical.

Topography has a major effect on the prepumping flow system within the Coastal Plain aquifers because it influences regional hydraulic gradients and the location of recharge and discharge areas. The prepumping regional flow system had ground-water recharge in areas of high water levels corresponding to high land altitudes, primarily in northwestern Monmouth County near the Fall Line and the aquifers' outcrop areas, in western Ocean County, and in central Gloucester and Camden Counties. Discharge areas in the prepumping regional flow system included the Atlantic Ocean, the Delaware River, the Delaware and Raritan Bays, and large rivers in the Coastal Plain.

Recharge to the confined aquifers during prepumping conditions generally flowed vertically downward through confining units and laterally downdip within aquifers and then upward through confining units to be discharged at large surface-water bodies. Figure 21 shows generalized flow patterns in four aquifers. These flow patterns are based on the prepumping potentiometric surface maps by Zapecza, Voronin, and Martin (1987, figs. 4, 5, 8, and 10), which were derived from the earliest recorded water-level measurements. Flow in downdip areas in the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) was not only vertically upward but also lateral and updip. The outcrop area of the middle and upper Potomac-Raritan-Magothy aquifers along the Delaware River was a discharge area for this updip flow. The fairly tight Merchantville-Woodbury confining unit (C3) separates the regional flow system in the underlying Potomac-Raritan-Magothy aquifers from that in the overlying Englishtown aquifer (A4). In many aquifers, updip flow toward outcrop areas occurs locally from adjacent ground-water highs directly downdip of the outcrop areas. Prepumping flow patterns are more local, and flow is faster in the unconfined outcrop areas of the aquifers than in the deeper confined parts. The Piney Point aquifer (A7) is totally confined and has a regional flow system in contrast to numerous local flow systems in the unconfined lower and upper Kirkwood-Cohansey aquifers (A8 and A9) (fig. 21).

The major source of water to the aquifers is precipitation that percolates through the unsaturated zone to the saturated zone of the unconfined parts of the Coastal Plain aquifers. A large part of this ground-water recharge is discharged to nearby surface-water bodies. Only a small amount infiltrates through confining units to recharge the underlining confined aquifers. In the central and southeastern Coastal Plain, precipitation is about 45 inches per year (in/yr), and evapotranspiration is about 22.5 in/yr according to Rhodehamel (1970, p. 6 and 7). He estimated total runoff for the same area to be about 22.5 in/yr with about 2.5 in/yr being surface or overland runoff and about 20 in/yr being ground-water runoff or ground-water discharge to streams. In Rhodehamel's budget, ground-water recharge is equivalent to ground-water runoff. Of the 20 in/yr of ground-water recharge, Rhodehamel estimated 3 in/yr to be deep flow to the confined aquifers. Therefore, most ground-water recharge, about 17 in/yr for this area, remains in the shallow unconfined aquifers.

Under prepumping conditions, deep flow to the confined aquifers was mainly discharged vertically upward through confining units to large surface-water bodies. Flow along the freshwater-saltwater boundary was generally upward discharge to the ocean. Most flow beneath the Delaware and Raritan Bays was upward discharge.

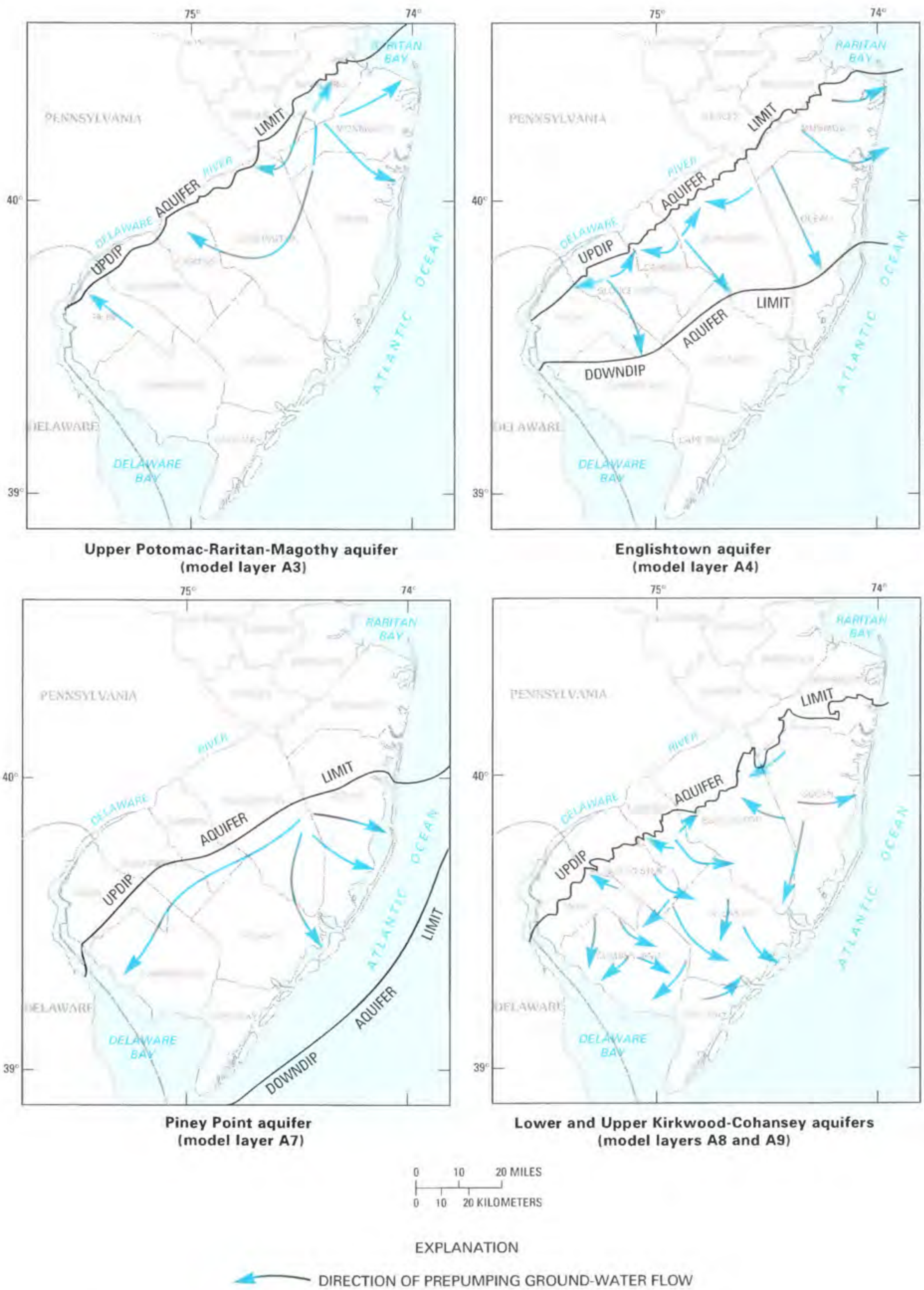


FIGURE 21.—Generalized patterns of prepumping ground-water flow in four Coastal Plain aquifers.

However, some ground-water flow in the deep confined aquifers beneath the bays was lateral as shown by Back (1966, p. A10).

Under pumping conditions, flow patterns changed in response to ground-water withdrawals. Water levels have declined in areas of ground-water withdrawals to create local and regional cones of depression in both confined and unconfined parts of the ground-water system. Several major cones of depression are shown in figure 22.

In the Potomac-Raritan-Magothy aquifer system there is a steep cone of depression in the middle aquifer (A2) near Raritan Bay in Middlesex and Monmouth Counties (fig. 22). On the basis of simulation, Farlekas (1979, p. 51) concludes that vertical leakage from the Old Bridge (upper Potomac-Raritan-Magothy aquifer, A3) is the major source of water for the withdrawals in the Farrington (middle Potomac-Raritan-Magothy aquifer, A2) and that other sources included water from aquifer storage and recharge from the outcrop area of the Farrington aquifer. The most areally extensive cone of depression in New Jersey appears in the upper, middle, and lower Potomac-Raritan-Magothy aquifers (A1, A2, and A3) centered in northwest Camden County (Walker, 1983, pls. 1 and 2). Luzier (1980, p. 68) states that near the city of Camden, the three stratigraphic units of the Potomac-Raritan-Magothy aquifer system function as one hydrologic system and that withdrawals had initially decreased discharge to the Delaware River and later (after 1960) induced recharge from the river.

In southern Monmouth and northeastern Ocean Counties, the Englishtown (A4) and Wenonah-Mount Laurel (A5) aquifers have very large and deep cones of depression (fig. 22) (Walker, 1983, pls. 3 and 4). Nichols (1977a, p. 96) on the basis of model simulation attributes both of these cones to withdrawals from the Englishtown aquifer and attributes the source of the water to vertical leakage from the aquifers above the Wenonah-Mount Laurel aquifer and lateral flow from the outcrop area of the Wenonah-Mount Laurel aquifer. In the confined Kirkwood aquifer (A8), a regional cone of depression centered in Atlantic County has developed that extends from the aquifer's updip limit to downdip areas beneath the Atlantic Ocean (Walker, 1983, pl. 5).

After prepumping conditions, lateral flow in the aquifers between Delaware and New Jersey and between New Jersey and Long Island changed in response to water- level declines. In most aquifers, flow toward New Jersey has increased or flow from New Jersey has reversed direction. However, in the Piney Point aquifer (A7), declining water levels in Delaware (Leahy, 1979, p. 18) have increased flow from New Jersey toward Delaware. In the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3), the confined Kirkwood aquifer (A8), and the upper Kirkwood-Cohansey aquifer (A9), prepumping movement of water downdip toward the saltwater areas within the aquifers has decreased in response to ground-water withdrawals, and a corresponding increase in chloride concentrations in downdip areas is expected.

SIMULATION OF GROUND-WATER FLOW

DIGITAL MODEL

Heads in the 10 aquifers of the New Jersey Coastal Plain were simulated by using a multilayer finite-difference model. A modified version of the Trescott (1975) computer codes was used and is described by Leahy (1982). A schematic representation of the model units used to represent the hydrogeologic units (fig. 2 and table 2) is shown in figure 23. A quasi-three-dimensional representation of the aquifers and confining units was used, in which it is assumed that flow is entirely horizontal within the aquifers and vertical through the confining units. Water levels within the confining units were not simulated, but vertical flow through the confining units is calculated by using vertical leakance (hydraulic conductivity divided by thickness).

APPROACH

Water levels were simulated for transient pumping conditions from 1896 to 1981 and prepumping steady-state conditions. A series of three models was used to simulate the conceptualized steady-state and transient flow systems. A 10-layer steady-state model was used to obtain roughly calibrated aquifer and confining-unit characteristics and to quantify the amount of flow from the unconfined parts of the aquifers into underlying confined aquifers. An 11-layer steady-state model was used to calculate initial water level conditions for an 11-layer transient model. The transient model was used to further calibrate aquifer and confining-unit properties. A flow chart showing the various stages of the modeling process discussed below is shown in figure 24.

The initial estimates of aquifer and confining-unit characteristics were used in a 10-layer steady-state model to simulate prepumping conditions. The 10-layer model has lateral no-flow boundaries in all of the aquifers. The upper boundary is a constant-head boundary representing the long-term average altitude of the water table in the outcrop areas of the aquifers. The 10-layer model was used to simulate only prepumping steady-state conditions, because simulating the water table as a constant-head boundary provides an infinite amount of recharge to the confined aquifers. A detailed discussion of the 10-layer model boundaries is necessary and is discussed in a later section on boundary conditions.

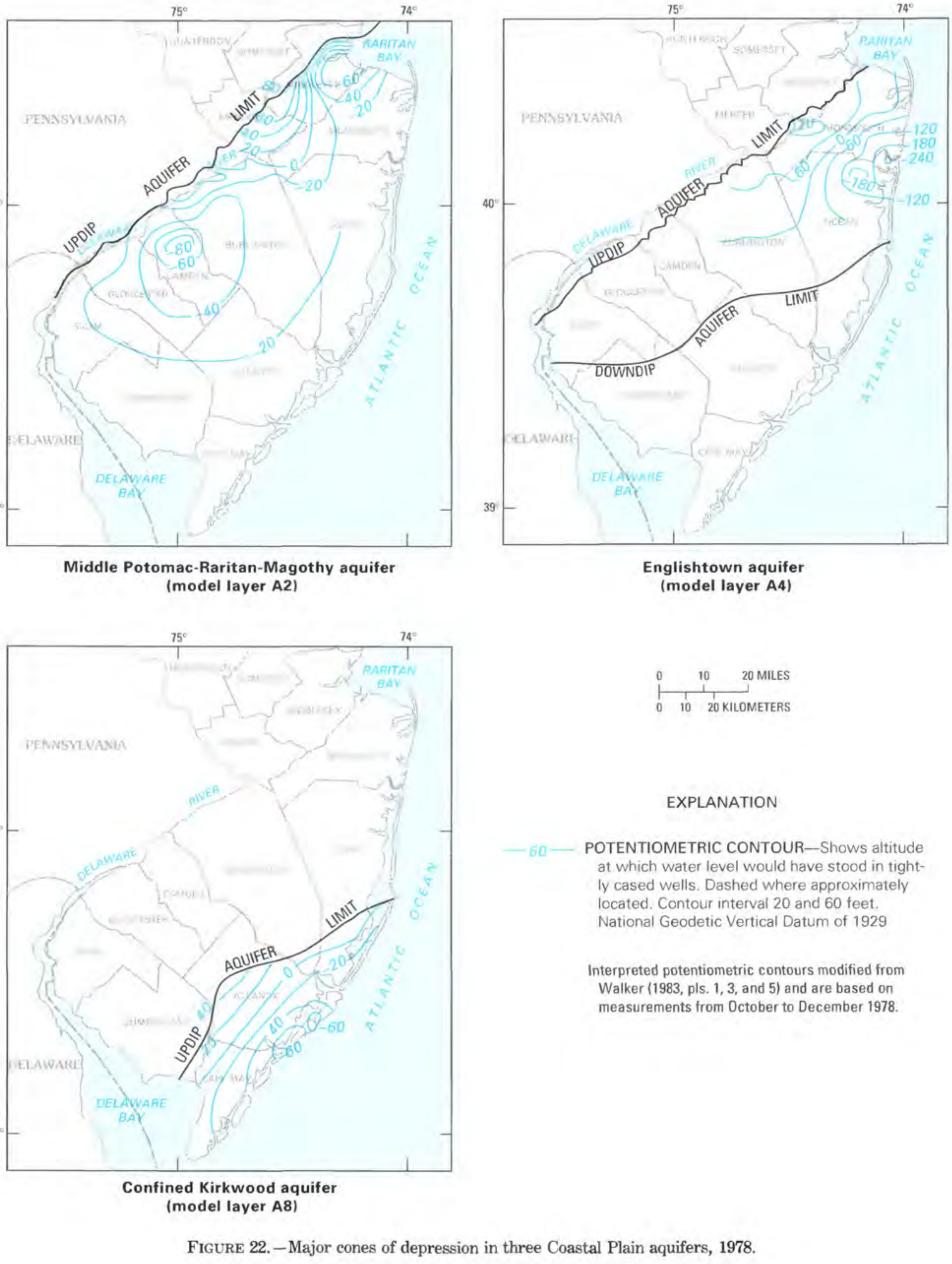


FIGURE 22.—Major cones of depression in three Coastal Plain aquifers, 1978.

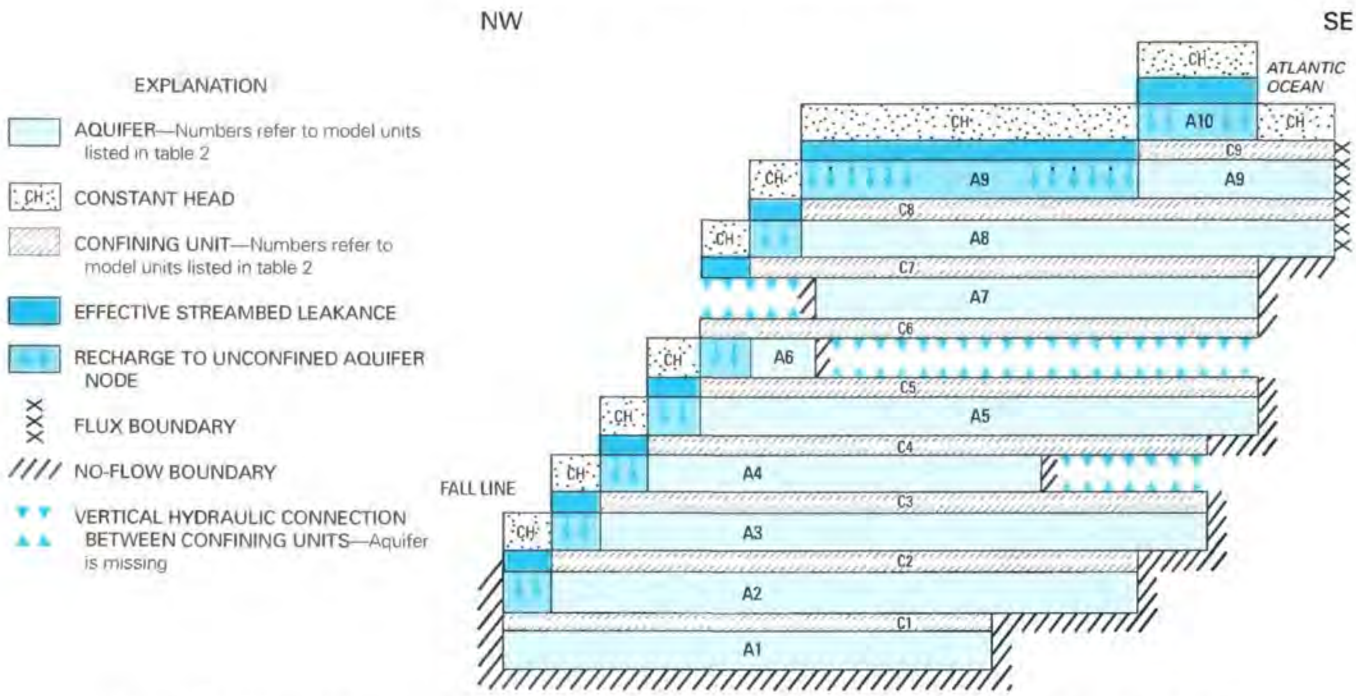


FIGURE 23.—Schematic representation of aquifers, confining units, and boundary conditions in the digital flow model.

To simulate changes in the water table under pumping conditions, an 11-layer model was used. The constant-head boundary representing the water table was removed from the outcrop areas, and recharge was added to the unconfined outcrop areas. In all onshore nodes an overlying constant-head boundary representing streams was added. The recharge boundary at the water table allows the model to simulate declines in water-table altitudes under pumping conditions. The 11-layer model was used to simulate both prepumping steady-state conditions and transient conditions from 1896 to 1981. The simulated prepumping water levels from the 11-layer model were used as initial conditions for transient simulations. Model input data that were changed during calibration were updated in the 10-layer and 11-layer steady-state simulations and in the 11-layer transient simulations to ensure compatible results among the three models.

Both the 11-layer steady-state and transient models have lateral flow boundaries. The boundary flows were calculated by the regional RASA model (Leahy and Martin, 1986). The regional RASA model simulates flow in the northern Atlantic Coastal Plain by using the approach and design similar to that used in the New Jersey subregional model. Model input data changes during calibration of the New Jersey flow model were updated in the New Jersey Coastal Plain area of the regional RASA model, and boundary flows were recalculated as needed to ensure compatible results between the regional and New Jersey subregional models. The lateral boundaries and estimates of lateral flows are discussed in the later sections on boundary conditions and model input data.

GRID DESIGN

The modeled aquifers were discretized by using a finite-difference grid shown in figure 25. The model area was divided by using 29 rows and 51 columns. Nodes in columns 1 and 51 and rows 1, 28, and 29 were not active in the calibration simulation; however, nodes in row 28 were used during some of the sensitivity simulations described later. The block-centered nodes are the center of the cell areas between two adjacent lines in the row direction (southwest to northeast) and two adjacent lines in the column direction (northwest to southeast). Nodes are designated in this report by row number followed by column number. For example, node 23, 18 is in row 23 and column 18. About 57 percent of the 14,790 nodes used to simulate the 10 aquifers are active.

A variable grid spacing is used to minimize the number of nodes in areas where fine-scale definition is not necessary. Grid spacing is largest in Delaware and in areas with little data on water levels or aquifer characteristics. Cell areas are 6.25 mi^2 in the northern and southwestern New Jersey Coastal Plain, 9.375 mi^2 in the southeastern Coastal Plain, and as large as 47.5 mi^2 in offshore areas.

The grid is aligned approximately parallel to the Fall Line and the strike of the Coastal Plain hydrogeologic units. The column direction of the New Jersey

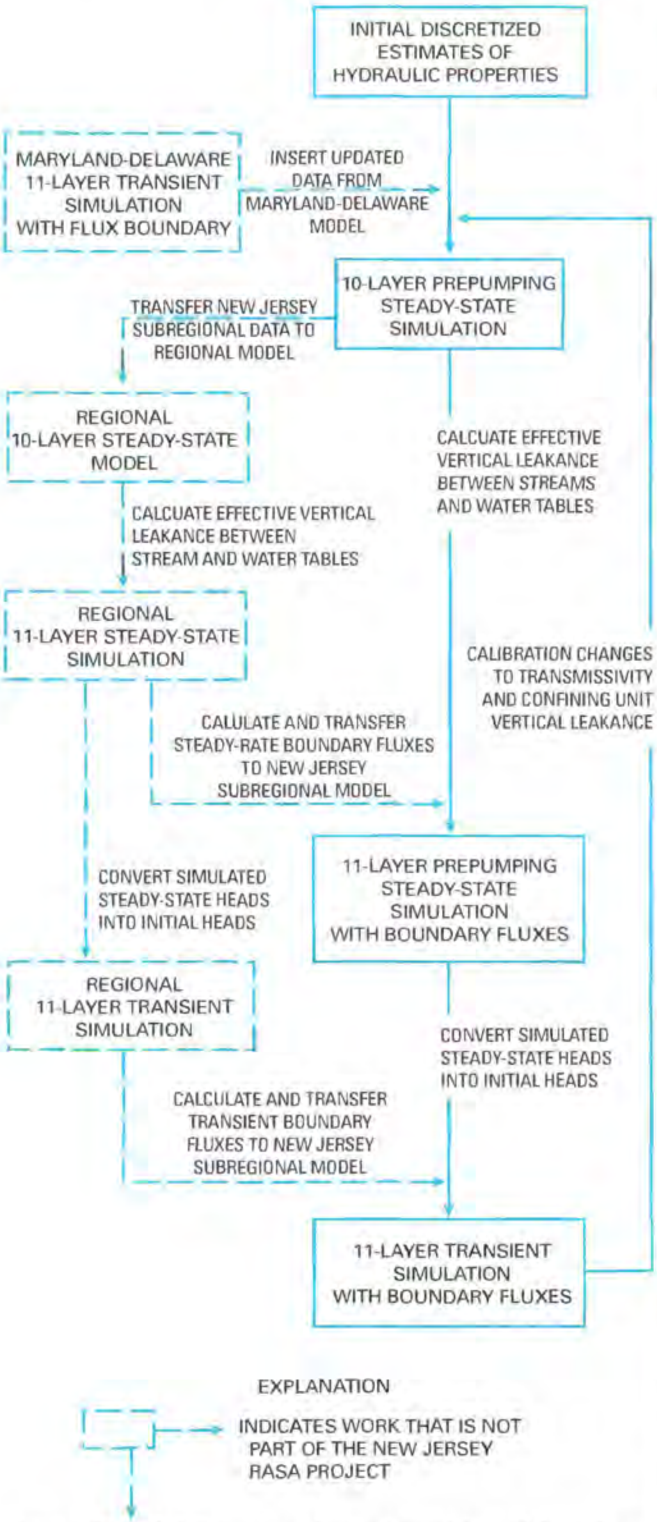


FIGURE 24.—Modeling approach showing relation of New Jersey subregional flow model to the northern Atlantic Coastal Plain regional and Maryland-Delaware subregional flow models.

subregional flow model grid is rotated about 8.5 degrees clockwise from the row direction of the grid used in the regional flow model. The cell size of the regional RASA grid is 49 mi^2 within the New Jersey subregional area. The orientation and cell size of the regional grid are shown in a representative node (31, 8) in figure 25.

BOUNDARY CONDITIONS

Lateral model boundaries are shown in figure 25. The boundaries shown are generalized because they are not the same in every aquifer. The northwestern updip limits of all the aquifers, except the Holly Beach aquifer (A10), are at the Fall Line or are within 15 miles of the Fall Line and generally are parallel to it; therefore, the northwest model boundary approximates the extent of the Coastal Plain sediments at the Fall Line and is represented by a no-flow boundary.

The northeast boundary in the 10-layer model roughly approximates a flow line in a ground-water discharge area in Raritan Bay. The southwest model boundary roughly approximates a flow line along a ground-water divide between the Delaware Bay and the Chesapeake Bay. The northeast and southwest boundaries are simulated as no-flow boundaries in the 10-layer model but as flow boundaries in the 11-layer model. To minimize the modeled area, the flow boundaries in the 11-layer model were located slightly closer to New Jersey than no-flow boundaries in the 10-layer model. Boundary flows are simulated only for those parts of the aquifers intersecting the flow boundaries. Those aquifers with limited extent are bounded on the northeast and southwest by no-flow boundaries.

The southeast boundaries of the modeled Potomac-Raritan-Magothy aquifers (A1, A2, and A3) approximate the downdip limit of freshwater. The positions of these boundaries are based on the depths to 10,000 mg/L chloride concentrations shown by Meisler (1980, fig. 4) and on the altitudes of the aquifer tops (figs. 3, 5, and 7). These boundaries were represented as no-flow boundaries and are referred to as the interface boundaries in this report.

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 with less than 10,000 mg/L chloride concentration is considered freshwater of equal density and nonmixing with higher density saltwater. Also, the interface is assumed to be stationary. Although it is known that the real interface is a transition zone of varying chloride concentrations between freshwater and saltwater, and that this transition zone will move in response to head changes in the freshwater flow system, the static freshwater-saltwater interface is assumed to be a reasonable

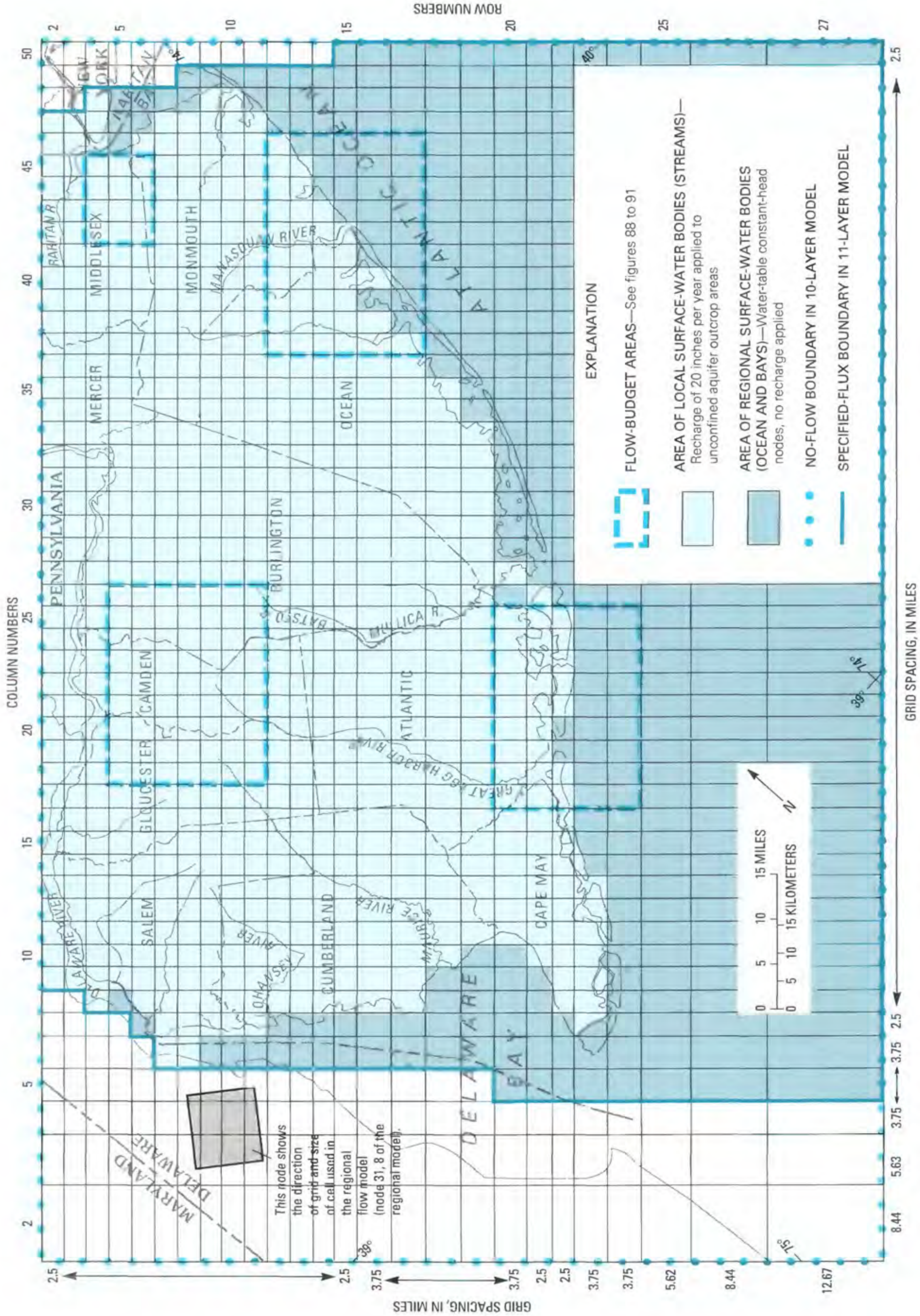


FIGURE 25.—Finite-difference grid, flow-budget areas, and generalized lateral model boundaries.

approximation of this boundary for both prepumping and transient conditions. Sensitivity of simulated water levels to the position of this interface boundary was tested and is described in a later section on sensitivity analysis. The estimated location of the 10,000 mg/L chloride concentration line in the Potomac-Raritan-Magothy aquifers (A1, A2, A3) is shown in figures showing data and results for these aquifers. The same boundary locations are used for simulation of both prepumping and transient conditions. However, it is not suggested that the actual 10,000 mg/L chloride concentration surface has not moved.

The southeastern boundary in the confined Kirkwood aquifer and the upper Kirkwood-Cohansey aquifer (A8 and A9) is the downdip limit of the modeled area. Both aquifers extend beyond the southeast model boundary. Within these units, freshwater is estimated to be about 25 to 35 miles further downdip than the model boundary based on the depth to chloride concentrations greater than 10,000 mg/L as shown by Meisler (1981, fig. 4) and on the estimated altitude of these aquifers offshore. This southeastern model boundary is a lateral flow boundary in the 11-layer model and is assumed to be far enough offshore to have little effect on simulated water levels onshore. However, simulated water levels were tested for sensitivity to the location of this boundary and the amount of flow at this boundary. These results are described in a later section on sensitivity analysis.

The Englishtown, Wenonah-Mount Laurel, Vincentown, and Piney Point aquifers (A4, A5, A6, and A7) have no-flow boundaries (fig. 23) to the southeast that represent their actual downdip limits. The Piney Point aquifer also has a no-flow boundary to the northwest, and the Holly Beach aquifer (A10) is present only in Cape May County and has lateral constant-head boundaries along the shore.

The lower boundary of the model represents the top of the underlying crystalline basement. The boundary is no-flow and underlies the lower Potomac-Raritan-Magothy aquifer (A1) and the extreme downdip parts of the middle Potomac-Raritan-Magothy aquifer (A2) where the lower Potomac-Raritan-Magothy aquifer is absent. The lower no-flow boundary in very limited downdip areas of the middle and upper Potomac-Raritan-Magothy aquifers, the Wenonah-Mount Laurel aquifer, the Piney Point aquifer, and the lower Kirkwood-Cohansey and confined Kirkwood aquifer (A2, A3, A5, A7, and A8) represents the top of sediments saturated with water containing chloride concentrations of 10,000 mg/L or more.

The upper boundary in the 10-layer model is a constant-head boundary representing the altitude of the water table in the unconfined parts of the aquifers. This boundary includes all of the Holly Beach aquifer (A10), most of the upper Kirkwood-Cohansey aquifer (A9), and the outcrop areas of the Potomac-Raritan-Magothy aquifers, the Englishtown aquifer, the Wenonah-Mount Laurel aquifer, the Vincentown aquifer, and the lower Kirkwood-Cohansey aquifer (A1–A6, A8). The Piney Point aquifer (A7) has no outcrop or constant-head boundary in the modeled area. The lower Potomac-Raritan-Magothy aquifer (A1) has no outcrop area or constant-head nodes in New Jersey, although there are several constant-head nodes representing a small outcrop area in Maryland. Surface-water levels in offshore areas are represented as constant-head nodes in the estimated outcrop areas of the hydrogeologic units. The majority of the offshore constant-head nodes are in the Holly Beach aquifer (A10).

The 10-layer model was used to simulate only prepumping steady-state conditions, because the water-table constant-head boundary has the potential to supply an infinite amount of recharge to the confined aquifers. However, the constant-head boundary of the 10-layer steady-state simulations provides the rate of deep percolation (Leahy and Martin, 1986, p. 169). Simulated flow to and from the water table is calculated by the model for each constant-head node. This flow represents the rate of water recharging to or discharging from the underlying confined aquifers and will be referred to as deep percolation in this report. Figure 26 shows the schematic representation of the upper boundary condition for the 10-layer model showing deep percolation.

The onshore upper boundary in the 11-layer model is constant head, representing long-term stream stage. Because of the large cell size in onshore areas (6.25 to 9.375 mi^2) and because there are numerous streams in New Jersey, each cell contains at least one stream and generally many. Therefore, the stream stage represented by the constant-head node is an average for the cell area. In onshore areas of the 11-layer model, the unconfined aquifer outcrop areas are active with simulated recharge to the water table. Changes in water-table altitudes caused by pumping can then be simulated in the 11-layer model. Figure 26 shows the upper boundary for the 11-layer model. Ground-water recharge is applied to nodes that were onshore water-table constant-head nodes in the 10-layer model. The long-term stream-stage constant-head nodes in the 11-layer model are in all nodes directly overlying previous onshore water-table constant-head nodes of the 10-layer model. However, constant-head nodes representing stream stage are not added above the areas of large surface-water bodies such as the ocean and bays (fig. 25). In these offshore areas, constant-head nodes in the 10-layer model are unchanged in the 11-layer model. No recharge is applied in this area.

To represent the flow between the aquifer outcrop areas and the streams, deep percolation calculated at

EXPLANATION

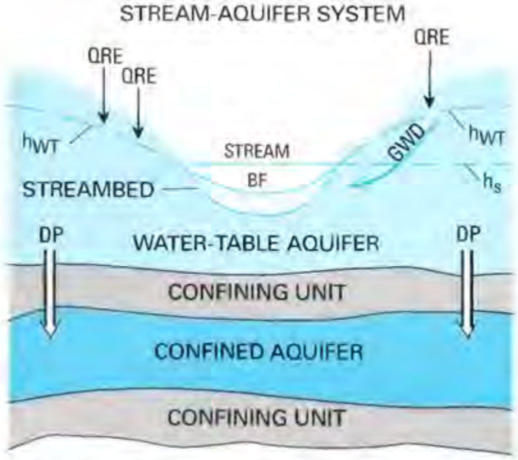
BF	BASEFLOW	h_{WT}	WATER-TABLE HEAD
DP	DEEP PERCOLATION	h_s	STREAM STAGE
QRE	GROUND-WATER RECHARGE	TK_{CB}	CONFINING-UNIT VERTICAL LEAKANCE
GWD	GROUND-WATER DISCHARGE	TK_{SB}	STREAMBED VERTICAL LEAKANCE
		T	TRANSMISSIVITY
		S	AQUIFER STORAGE COEFFICIENT

GWD = BF

BF = QRE − DP

$TK_{SB} = BF/(h_{WT} - h_s)$

STREAM-AQUIFER SYSTEM



10-LAYER MODEL

CONFIGURATION	DATA
CONSTANT-HEAD WATER-TABLE AQUIFER	h_{WT}
CONFINING UNIT	TK_{CB}
AQUIFER	T, S
CONFINING UNIT	TK_{CB}

11-LAYER MODEL

CONFIGURATION	DATA
CONSTANT HEAD AT STREAM STAGE	h_s
STREAMBED	TK_{SB}
WATER-TABLE AQUIFER	QRE, T, S
CONFINING UNIT	TK_{CB}
AQUIFER	T, S
CONFINING UNIT	TK_{CB}

FIGURE 26.—Schematic representation of stream-aquifer flow, deep percolation, and the upper boundary in the 10-layer and 11-layer models.

each onshore node in the 10-layer model was used to calculate the amount of base flow at each node (base flow equals ground-water recharge minus deep percolation). An effective streambed leakance is calculated, by means of Darcy's Law, at each node from the modeled recharge rate, simulated deep percolation, and hydraulic gradient between the water table and streams. Each cell contains many small streams, but one average stream stage was estimated for each node. Regional lateral flow in the aquifer outcrop areas is assumed to be negligible. The calculated streambed leakance does not represent actual streambed properties or characteristics for a single stream but is a lumped parameter of all the factors controlling flow between the water table and streams on a regional scale for each node.

The upper model boundary allows the simulated water-table level to vary in response to ground-water withdrawals but still has the potential to supply an unlimited source of water from the simulated streams. The upper model boundary does not simulate short-term changes in water-table or stream stages caused by seasonal changes in recharge or tides. However, the upper model boundary is a reasonable representation of field conditions as neither long-term or large-scale declines in water-table levels nor elimination of streamflow on a regional scale due to ground-water withdrawals has been observed. The method of determining the altitude of the water table and stream stage for these boundaries is discussed in the following section on model input data.

MODEL INPUT DATA

Initial estimates of transmissivity were based on results from aquifer tests, specific-capacity tests, and model simulations of previous studies. A summary of hydraulic conductivity determined for each aquifer from specific-capacity tests is given in table 5. The hydraulic conductivity values were estimated from specific-capacity tests by using equation 6 of McClymonds and Franke (1972, p. 11) and multiplied by aquifer thickness to obtain the initial transmissivity values. Transmissivity and hydraulic conductivity determined from aquifer-test data and model simulations are listed in table 3. Some transmissivity values are shown in table 3 with no corresponding hydraulic conductivity. Hydraulic conductivity was calculated from these transmissivity values by dividing the values by the estimated saturated thickness of the aquifer at the location. These hydraulic conductivity values from aquifer and specific-capacity tests were used to estimate a range or average hydraulic conductivity for each aquifer. These estimates were then slightly modified near State lines to be consistent with hydraulic conductivity estimates used in adjacent States.

An average hydraulic conductivity value was used for the Englishtown (A4), Wenonah-Mount Laurel (A5), Vincentown (A6), Piney Point (A7), and Holly Beach (A10) aquifers. These aquifers had very little areal variability in hydraulic conductivity. Areally variable hydraulic conductivity values were used for each of the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) and the two Kirkwood-Cohansey aquifers (A8 and

TABLE 5.—*Summary of hydraulic-conductivity values based on specific-capacity tests and initial estimates of transmissivity*

Model unit	Aquifer	Number of tests	Hydraulic conductivity (ft/d)[1]			Initial estimated range of transmissivity[2] (ft²/d)[3]
			Minimum	Maximum	Average	
A1	Lower Potomac-Raritan-Magothy aquifer	35	32	527	199	860–17,300
A2	Middle Potomac-Raritan-Magothy aquifer	104	22	397	155	860–21,600
A3	Upper Potomac-Raritan-Magothy aquifer	131	29	432	135	860–19,900
A4	Englishtown aquifer	30	6	130	30	90– 5,400
A5	Wenonah-Mount Laurel aquifer	22	7	34	16	90– 2,300
A6	Vincentown aquifer	18	10	110	42	860– 3,500
A7	Piney Point aquifer	8	9	95	33	260– 5,200
A8	Lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer.	33	50	320	120	860–19,900
A9	Upper Kirkwood-Cohansey aquifer	45	50	300	178	860–25,900
A10	Holly Beach aquifer	9	120	180	150	5,200– 7,800

[1] ft/d: feet per day.
[2] Transmissivity: Estimated prior to model calibration from hydraulic conductivity values and thickness.
[3] ft²/d: feet squared per day.

A9). Generally, the aquifers with the lowest variability also had the least amount of available data.

Initial estimates of transmissivity are also given in table 5. The Potomac-Raritan-Magothy aquifers, the Kirkwood-Cohansey aquifers, and the Holly Beach aquifer (A1, A2, A3, A8, A9, and A10) have the highest average hydraulic conductivity (greater than 100 ft/d) and the highest initial estimates of transmissivity, except for the Holly Beach aquifer, which has a high hydraulic conductivity (150 ft/d) but a low initial estimate of transmissivity because the unit is relatively thin.

Initial estimates of vertical confining-unit leakance were made by dividing an estimate of the vertical hydraulic conductivity for each unit by the thickness of the unit at each node. Estimates of confining-unit vertical hydraulic conductivity were based on data presented in table 4 and on estimated vertical hydraulic conductivity in adjacent states. Only one vertical hydraulic conductivity was used for each confining unit and generally approximated to the lowest estimate of hydraulic conductivity for the unit. If no vertical hydraulic conductivity was given for a confining unit in table 4, the estimate was based on the vertical hydraulic conductivity of other units with similar lithologic properties.

In preliminary steady-state simulations, the initial estimates of leakance were found to be generally too low, particularly in updip areas. Estimates of vertical hydraulic conductivity were then reevaluated and increased about two orders of magnitude in updip areas. This distribution of a decrease in vertical hydraulic conductivity toward the downdip directions is logical because the sediments tend to be finer grained in the downdip direction and, therefore, less permeable. Luzier (1980, fig. 9) used a similar distribution for the Merchantville-Woodbury confining unit. Estimates of leakance based on the reevaluated vertical hydraulic conductivity values varied from three to four orders of magnitude within each confining unit because confining units thicken and grain size is finer in the downdip direction.

Storage coefficients determined from aquifer-test data (table 3) were used to estimate one representative storage coefficient for all the aquifers. The storage coefficient data range from 1.5×10^{-5} to 8.1×10^{-2}. The highest values are generally near the aquifer outcrop areas and indicate unconfined or partially confined conditions. A single confined storage coefficient of 1.0×10^{-4} was used for each of the aquifers. In unconfined areas, a specific yield of 0.15 was used. These coefficients are approximate means for all of the aquifers and were the same as the values used in other States for the northern Atlantic Coastal Plain RASA study. Generally, these values were not changed during model calibration.

Prepumping water-table altitudes for the upper boundary of the 10-layer model were estimated by using measured water levels and long-term stream stages, which were estimated by using topographic maps. Water-table altitudes between surface-water bodies were based on the altitude and configuration of land surface and measured water levels. Water-table altitudes are shown in figure 27. The altitudes are generalized because they were estimated for periods of average precipitation or ground-water recharge under prepumping conditions.

The effective long-term stream altitudes for the upper boundary of the 11-layer model were estimated to be the average altitude of the largest stream segment from topographic maps. All nodal stream altitudes are less than the water-table altitudes at that node. Therefore, under prepumping conditions, the streams are simulated as discharge points from the aquifers. Although the simulated streams are discrete points and not a continuous surface, stream altitudes have been contoured and

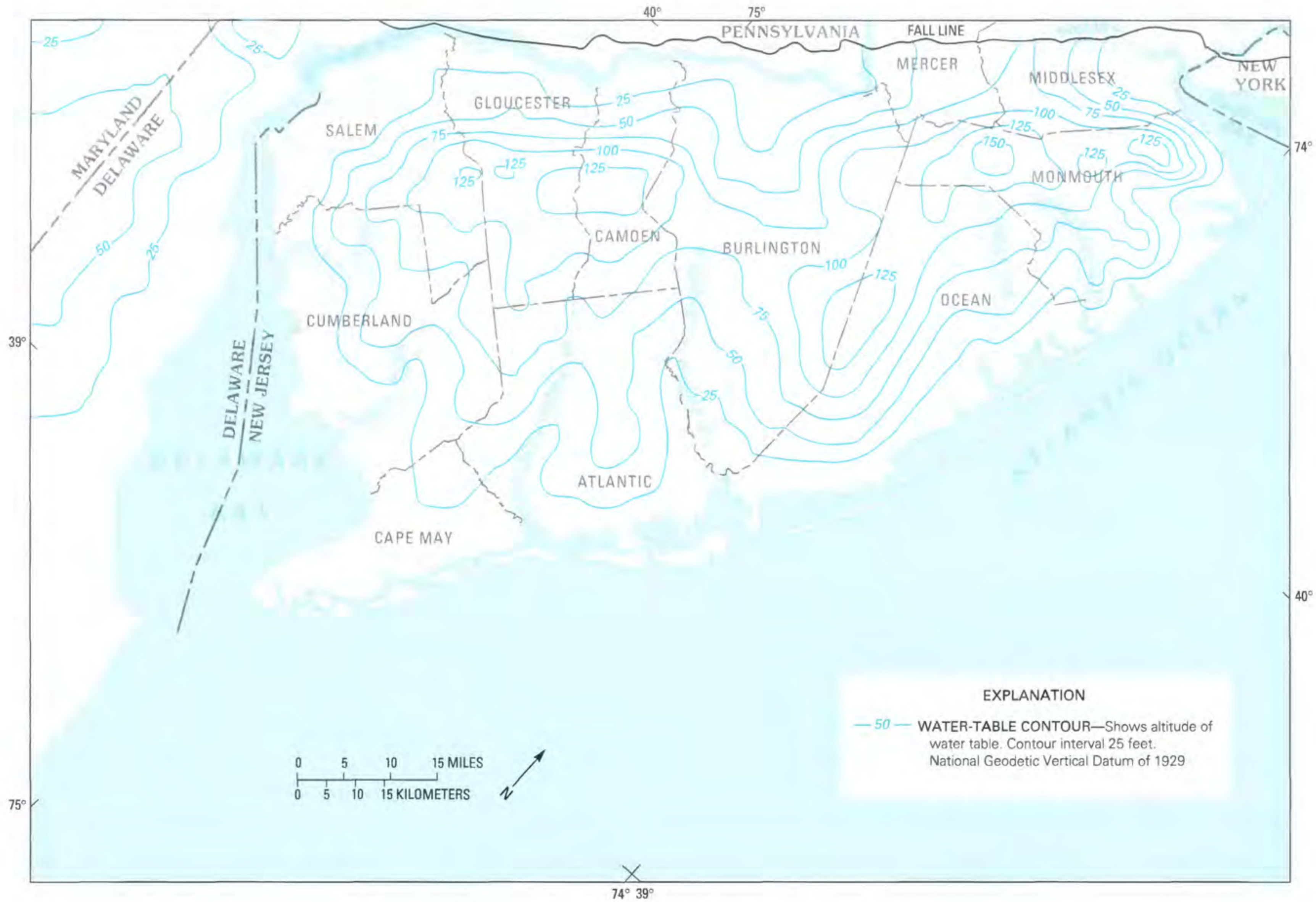


FIGURE 27.—Generalized altitude of the water table, prepumping conditions.

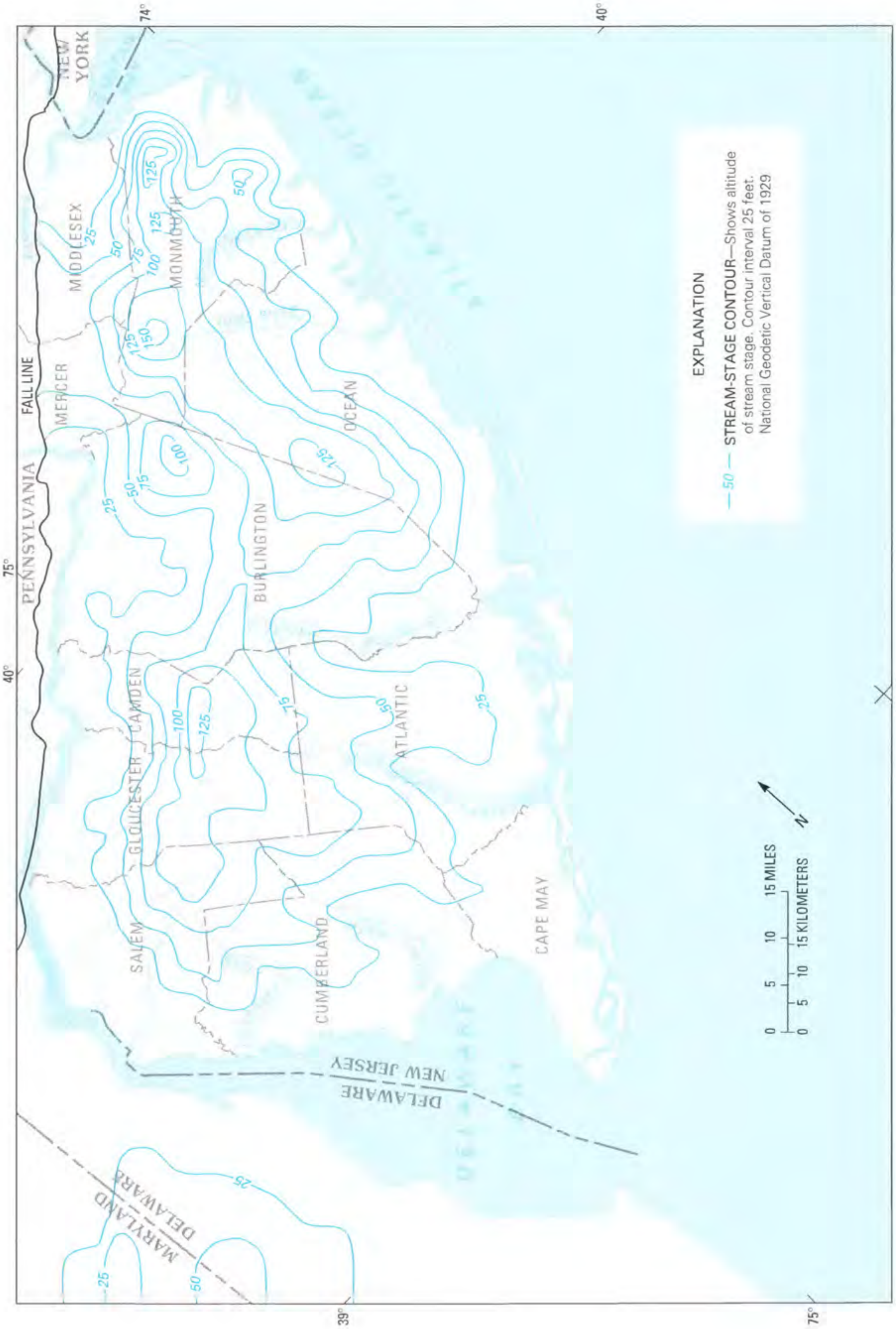


FIGURE 28.—Generalized altitude of streams.

TABLE 6.—*Summary of hydrologic budget analyses used to estimate ground-water recharge*

Precipitation (in/yr)	Evapotranspiration (in/yr)	Runoff (in/yr)[1]			Other (in/yr)	Location	Reference
		Surface	Ground-water	Total			
44	23	—	—	21	—	Delaware River Basin	Parker and others (1964, p. 14).
45	22.5	2.5	20	22.5	up to 2.8[2]	Pine Barrens Region central and southeast Coastal Plain.	Rhodehamel (1970, p. 5, 6, 16, and 18).
44	22	—	—	22	—	Gloucester County	Hardt and Hilton (1969, p. 54).
43–46	23–26	—	—	17–20	—	Burlington County	Rush (1968, p. 16–18).
44.2	24.0	—	—	20.2	—	New Jersey Coastal Plain	Vowinkel and Foster (1981, p.18).
41	25	4	12	16	—	Nanicoke River Basin, Kent and Sussex Co., Del.	Johnston (1976, p. 39).
—	—	—	—	20–21	—	Potomac-Raritan-Magothy outcrop area.	Barksdale (1958, p. 102).
—	—	—	—		4–11[3]	Potomac-Raritan-Magothy outcrop area.	Luzier (1980, p. 32).
—	—	—	—		4–8[3]	Farrington aquifer system outcrop area.	Farlekas (1979, p. 32).

[1] in/yr: inches per year.
[2] Deep regional ground-water recharge, a part of ground-water runoff.
[3] Ground-water recharge, a part of ground-water runoff.

are shown in figure 28. The generalized altitudes are estimated for periods of average precipitation or streamflow. These contours provide a means to compare stream altitudes to water-table altitudes and to analyze the areal variation in stream altitudes. However, there is no intention to suggest that these contours represent a continuous stream surface.

Recharge to the outcrop areas in the 11-layer model is equal to long-term precipitation minus long-term evapotranspiration and can be directly measured as long-term streamflow. Streamflow consists of surface runoff and ground-water discharge. Ground-water discharge is water that initially recharged aquifers and later discharged to streams. Ground-water discharge may be (1) local flow to nearby streams within shallow aquifers, (2) intermediate flow moving along longer flow paths beneath nearby streams and discharging to neighboring larger streams, or (3) regional flow moving through the deep confined aquifers and eventually discharging to large rivers or surface-water bodies. The upper boundary in the 11-layer model, with recharge to the water table and constant-head nodes representing the stream stage, simulates surface runoff and local ground-water discharge as flow through one or two cells (less than 5 mi) of the outcrop areas, whereas intermediate and regional ground-water discharge is simulated as deep percolation that has moved longer distances through the confined aquifers.

Water-budget estimates were used to determine an estimate of ground-water recharge to the water-table aquifer for the 11-layer model (table 6). A value of 20 in/yr was used to represent the long-term and areal average ground-water recharge in the New Jersey Coastal Plain. This value is consistent with values shown in table 6 and those used in the northern Atlantic Coastal Plain RASA studies in adjacent States (15 in/yr in Maryland and 22.5 in/yr in Long Island, N.Y.). This value is also consistent with average stream discharge. The average stream discharge for 16 streams with surface-water basins within the New Jersey Coastal Plain was 22.2 in/yr in 1981.

Ground-water withdrawal data presented by Zapecza, Voronin, and Martin (1987, tables 2 and 3) for 1918 through 1980 are used in the New Jersey Coastal Plain area of the model. The latitude and longitude of wells and well fields are used to identify the nodes to which withdrawals are applied. Figure 29 shows total annual withdrawals in the New Jersey Coastal Plain and the averaged withdrawal rates for each of nine pumping periods from 1896 through 1980.

The pumping periods are 3 to 25 years long. Withdrawals for each pumping period are also given in table 7. Time discretization was based not only on changing withdrawal rates in New Jersey but also on historical withdrawal rates from other States within the northern Atlantic Coastal Plain. The regional RASA model and subregional models for each of the other States used pumping periods equivalent to those in the New Jersey subregional model.

The average withdrawals in New Jersey for pumping period 1 (1896 to 1921) are estimated to be one-half of the 1920 withdrawal rate. Withdrawals for Pennsylvania (R.A. Sloto, written commun., 1983) and Delaware (W.B. Fleck, written commun., 1983) were estimated

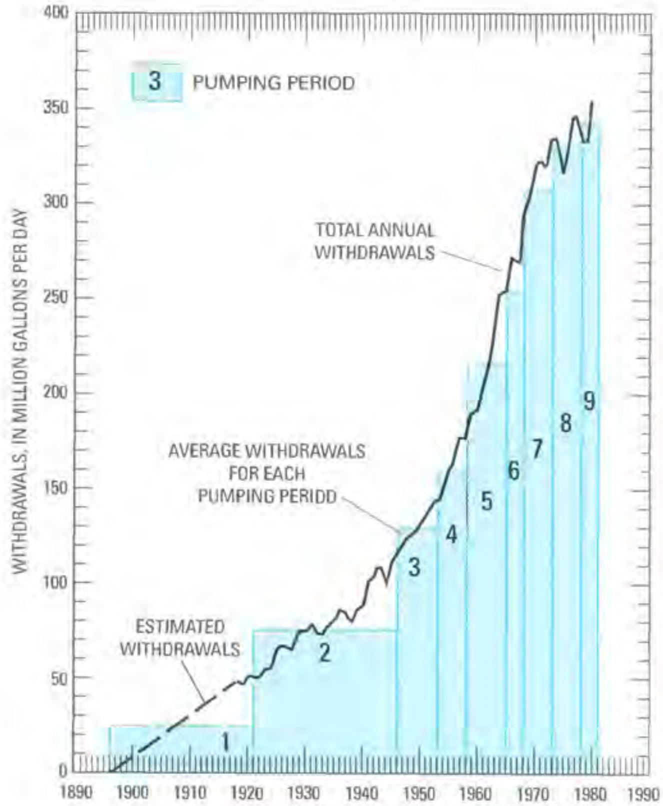


FIGURE 29.—Total annual ground-water withdrawals and average withdrawals for each pumping period for the New Jersey Coastal Plain.

from data compiled for modeling studies in those States by the U.S. Geological Survey. Withdrawals for the relatively small areas of these States that are included in the model are given in table 7. Withdrawals for Pennsylvania and Delaware areas are from 5 to 16 percent of total withdrawals for each pumping period.

Lateral boundary flows for the 11-layer steady-state and transient model were calculated by using simulated water levels and transmissivity values from the regional RASA model. Flows were calculated by using water levels simulated by the prepumping steady-state model and water levels simulated at the end of each pumping period by the transient model. Calculated flows were assumed to represent flow across the boundaries for the following pumping period and to be constant throughout the pumping period. Flows calculated from the regional prepumping simulations were used in the New Jersey subregional flow model for the prepumping simulation and pumping period 1. Total boundary flows for each pumping period are shown in table 7. Flows along each boundary for prepumping conditions and pumping period 8 are discussed further for each aquifer and boundary in the later section on the sensitivity analysis of boundary flows.

TABLE 7.—*Ground-water withdrawals and lateral boundary flows for each pumping period*

[Boundary flows: Positive boundary flows are out of model area, negative boundary flows are into model area. Difference between flows for any pumping period and prepumping flows is the source of water to wells in table 13. Total flow: Sum of withdrawals and boundary flows. Mgal/d: Million gallons per day.]

Pumping period	End date	Withdrawals (Mgal/d)		Boundary flows (Mgal/d)	Total flow (Mgal/d)
		New Jersey	Pennsylvania and Delaware		
prepumping	1–1–1896	0	0	4.7	4.7
1	1–1–1921	24.6	1.8	4.7	31.1
2	1–1–1946	75.2	10.8	4.4	90.4
3	1–1–1953	127.3	23.5	3.8	154.5
4	1–1–1958	158.1	16.7	2.2	177.0
5	1–1–1965	215.3	19.6	0.7	235.6
6	1–1–1968	253.7	20.3	−1.3	272.7
7	1–1–1973	303.7	21.9	−3.2	322.4
8	1–1–1978	331.1	19.8	−5.2	345.7
9	1–1–1981	341.7	16.5	−5.6	352.6

MODEL CALIBRATION

APPROACH

Calibration was achieved primarily by a trial-and-error adjustment of aquifer transmissivity and confining-unit vertical hydraulic conductivity. Generally, simulated water levels are affected most by changes in these hydraulic characteristics. However, the altitude of the water table, stream stages, and aquifer storage coefficients were also adjusted slightly during calibration. Simulated water levels were tested for sensitivity to other model input data and to several boundary conditions during and after the calibration process.

The 10-layer steady-state model was used to roughly calibrate aquifer and confining-unit properties by comparing simulated prepumping potentiometric surfaces to measured prepumping water levels. The measured prepumping water levels and potentiometric surfaces for the Coastal Plain aquifers are shown and discussed by Zapecza, Voronin, and Martin (1987). Prepumping potentiometric surfaces are available for all aquifers except the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2), which had only sparse prepumping water-level data. The steady-state model provided an efficient method of obtaining rough estimates of the hydraulic characteristics. Simulated prepumping water levels changed slightly during calibration of the transient flow system as more measured water levels provided a basis for more rigorous calibration. Therefore, the calibration of the transient flow system and the final calibration of the prepumping flow system were done simultaneously with similar changes in model input data made in the 10-layer and 11-layer steady-state models and in the 11-layer transient model.

The 11-layer model was calibrated to transient water levels by attempting to match interpreted 1978 potentiometric surfaces and long-term well hydrographs. Interpreted 1978 potentiometric surface maps by Walker (1983, fig. 9 and pls. 1–5) are available for the following aquifers: the upper Potomac-Raritan-Magothy aquifer system (A3), the Englishtown aquifer (A4), the Wenonah-Mount Laurel aquifer (A5), the confined Kirkwood aquifer (A8), and the confined part of the upper Kirkwood-Cohansey aquifer (A9). Walker (1983) refers to the lower and middle Potomac-Raritan-Magothy aquifers of this study as the lower aquifer of the Potomac-Raritan-Magothy aquifer system. Data for the 1978 potentiometric surface for the undifferentiated lower and middle aquifers (Walker, 1983, pl. 1) were reanalyzed to make separate maps for each aquifer. Potentiometric surface maps for 1978 are not available for the Vincentown aquifer (A6), Piney Point aquifer (A7), the unconfined lower and upper Kirkwood-Cohansey aquifers (A8 and A9), and the Holly Beach aquifer (A10). Simulated potentiometric surfaces for these aquifers are compared to measured water levels.

Data for the interpreted 1978 potentiometric surfaces were collected over a period of several months but primarily in October or November of 1978. However, pumping period 8 ends closest to this date on January 1, 1978. Although the dates of the simulated and interpreted transient water-level surfaces differ in time by at least 10 months, their comparison is adequate for simulating the regional flow system. In areas with observation wells, water-level declines from 1973 to 1978 averaged 1 to 4 ft/yr (Walker, 1983). At this rate water-level declines from January 1, 1978, to October or November 1978 would be much less than the calibration criteria of 15 feet (described below). The simulated January 1, 1978, and interpreted October/November 1978 potentiometric surfaces are referred to as the simulated and interpreted 1978 potentiometric surfaces, respectively, in this report.

Well hydrographs also were used for calibration. Simulated water levels at the end of each pumping period were compared to the measured water levels closest to the end of the pumping period. Comparison of well hydrographs to simulated water levels provided more accuracy than using only potentiometric surface maps. The use of well hydrographs also minimized the error in comparing the simulated and interpreted 1978 surfaces with dates that differed by 10 months. Simulated hydrographs are compared to measured water levels presented by Zapecza, Voronin, and Martin (1987, pls. 1–10) for the period 1924 to 1984 for each aquifer. However, most wells do not have water-level data for the entire period. Hydrographs for wells in the undifferentiated Potomac-Raritan-Magothy aquifer system are assigned to the lower or middle aquifer based on the altitude of the well screen intervals and the altitude of the lower (fig. 3) or middle (fig. 5) aquifer. Likewise, hydrographs for the undifferentiated Kirkwood-Cohansey aquifer system are assigned to the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (A8) and the upper Kirkwood-Cohansey aquifer (A9) based on the screen intervals and the altitudes of the tops of the aquifers shown in figures 17 and 19.

A total of 89 well hydrographs were selected for model calibration from the hundreds of wells in the New Jersey Coastal Plain measured by the U.S. Geological Survey. These well hydrographs were selected on the basis of their location and period of record. Most wells are in the updip parts of the aquifers, although records for some deeper wells in downdip areas are available. Most wells have periods of recorded water-level measurements of at least 15 years, and several wells have records of more than 50 years. Seven to 13 well hydrographs were used for each aquifer. Only the Vincentown aquifer (A6), with one well, has an insufficient number of wells with recorded water levels for calibration.

Simulated water levels were used to generate simulated hydrographs for observation-well sites at points other than the nodes for comparison with well hydrographs. A linear interpolation based on the location and simulated water levels at the three nodes nearest to the observation well was used to calculate the simulated hydrograph of that well. The simulated water levels at the three nodes define a plane, which is assumed to approximate water levels between these points. The altitude of the plane at the site of the measured water level is determined by the location of the observation-well site with respect to the three nodes. The model code was modified to calculate the simulated hydrographs using this three-node method.

Determining the acceptability of the match between simulated and interpreted potentiometric surfaces and hydrographs is subjective. The attempt was made to match interpreted 1978 potentiometric surfaces as closely as possible and hydrographs to within 10 feet. Calibration was considered acceptable when the following criteria were satisfied:

1. The interpreted 1978 potentiometric surfaces (including the depths of the cones of depression) were reproduced to within 15 feet, and the locations and configurations of the simulated cones were reasonable.
2. The interpreted prepumping steady-state potentiometric surfaces were reproduced to within 15 feet.
3. Simulated hydrographs matched the measured hydrographs to within 15 feet at the end of each pumping period.
4. Simulated water levels in areas with little or no measured data were compatible to water levels in

areas with data and were compatible with flow directions postulated in the conceptual model.

5. Hydraulic characteristics representing the flow system (including aquifer transmissivity, confining-unit vertical hydraulic conductivity, flow rates between the aquifers, and flow between the unconfined aquifer and the streams) were compatible to measured aquifer and confining-unit characteristics and to flow rates postulated in the conceptual model.

The model is more rigorously calibrated in some areas than others. The degree of calibration depends on the distribution of measured data, the character of the hydrogeologic framework and potentiometric surfaces in relation to the cell size, the sensitivity of simulated water levels to data changes, and the ease with which calibration was achieved. Generally, the Vincentown aquifer (A6) and offshore areas of the aquifer system were calibrated only according to the fourth and fifth calibration criteria and are not considered to be well calibrated. Also, calibration within the outcrop areas of the confined aquifers is severely limited by the cell size. The area of the outcrops in any cell is generally much less than the cell size. The model outcrop areas provide a means for simulated regional recharge to the confined aquifers, but local unconfined water levels or flow rates are not simulated accurately in these areas.

SIMULATION OF PREPUMPING STEADY-STATE CONDITIONS

Generally, the match between simulated and interpreted prepumping potentiometric surfaces is considered acceptable, although in some areas of some aquifers there is not a good match. Figures 30 to 39 show the simulated surfaces and, when available, the interpreted prepumping potentiometric surfaces. The interpreted surfaces show contours based on measured data. Discussion of the comparison of the simulated and interpreted surfaces follows.

The simulated prepumping potentiometric surfaces for the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2) are shown in figures 30 and 31. The interpreted and simulated prepumping surfaces for the upper aquifer (A3) are shown in figure 32. The high ground-water altitudes near the outcrop in Middlesex County and the low altitudes near the outcrop area along the Delaware River, near Raritan Bay, and along the Atlantic Coast in Ocean and Monmouth Counties are simulated in all three aquifers. However, the lateral gradient from the outcrop area in Mercer and Middlesex Counties to the discharge area beneath the Atlantic Ocean in the upper aquifer is not closely simulated, and simulated water levels along the Atlantic Coast may be 10 to 15 feet higher than actual water levels.

The simulated and interpreted prepumping potentiometric surfaces for the Englishtown aquifer (A4) are shown in figure 33. The high ground-water altitudes near the outcrop in northwestern Monmouth County, the enclosed ground-water low in the outcrop in western Burlington County, and the low ground-water altitudes beneath Raritan Bay, the Atlantic Coast, and the outcrop area in Gloucester and Camden Counties are all simulated. Two small enclosed ground-water highs near the outcrop in Burlington and Gloucester Counties and the lateral gradient between the outcrop in Monmouth County and the Atlantic Coast were not simulated. Also, the simulated surface shows a ground-water high in Camden County that is lower, smaller, and further north in the interpreted potentiometric surface. The small enclosed highs near the outcrop may be too small in comparison to the cell size to be accurately simulated. Nevertheless, the general ground-water altitudes are approximately simulated in these areas. The larger simulated ground-water high in Camden County differs from the interpreted potentiometric surface because the interpreted surface probably does not represent true prepumping conditions and reflects water levels lowered by small amounts of unrecorded local pumpage. As in the Potomac-Raritan-Magothy aquifers, ground-water gradients in Monmouth County are not closely simulated, and simulated water levels along the coast may be 15 feet higher than actual prepumping water levels in the Englishtown aquifer.

The closely matched simulated and interpreted potentiometric surfaces for the Wenonah-Mount Laurel aquifer (A5) are shown in figure 34. Only minor differences are seen. The local ground-water highs or lows based on a single data point may be incorrect on the interpreted potentiometric surface map or may not be large enough in relation to the cell size to be simulated. Generally, the simulated prepumping flow patterns in the Wenonah-Mount Laurel aquifer are similar to those in the Englishtown aquifer (A4) (fig. 33). Ground-water altitudes are high near the outcrop in Monmouth and Camden Counties and low near the outcrop in Burlington and Salem Counties and in most of the southeastern Coastal Plain and along the Atlantic Coast.

The simulated and interpreted prepumping potentiometric surfaces for the Vincentown aquifer (A6) are shown in figure 35. Generally, the simulated potentiometric surface for the Vincentown aquifer (A6) is similar to the potentiometric surface for the underlying areas of Wenonah-Mount Laurel aquifer (A5) (fig. 34), as water-level differences between these aquifers are less than 5 feet, except at the Atlantic Coast in Monmouth County. The simulated water levels for the Vincentown aquifer approximate the measured water levels only generally. Most of the localized areas of high or low ground-water

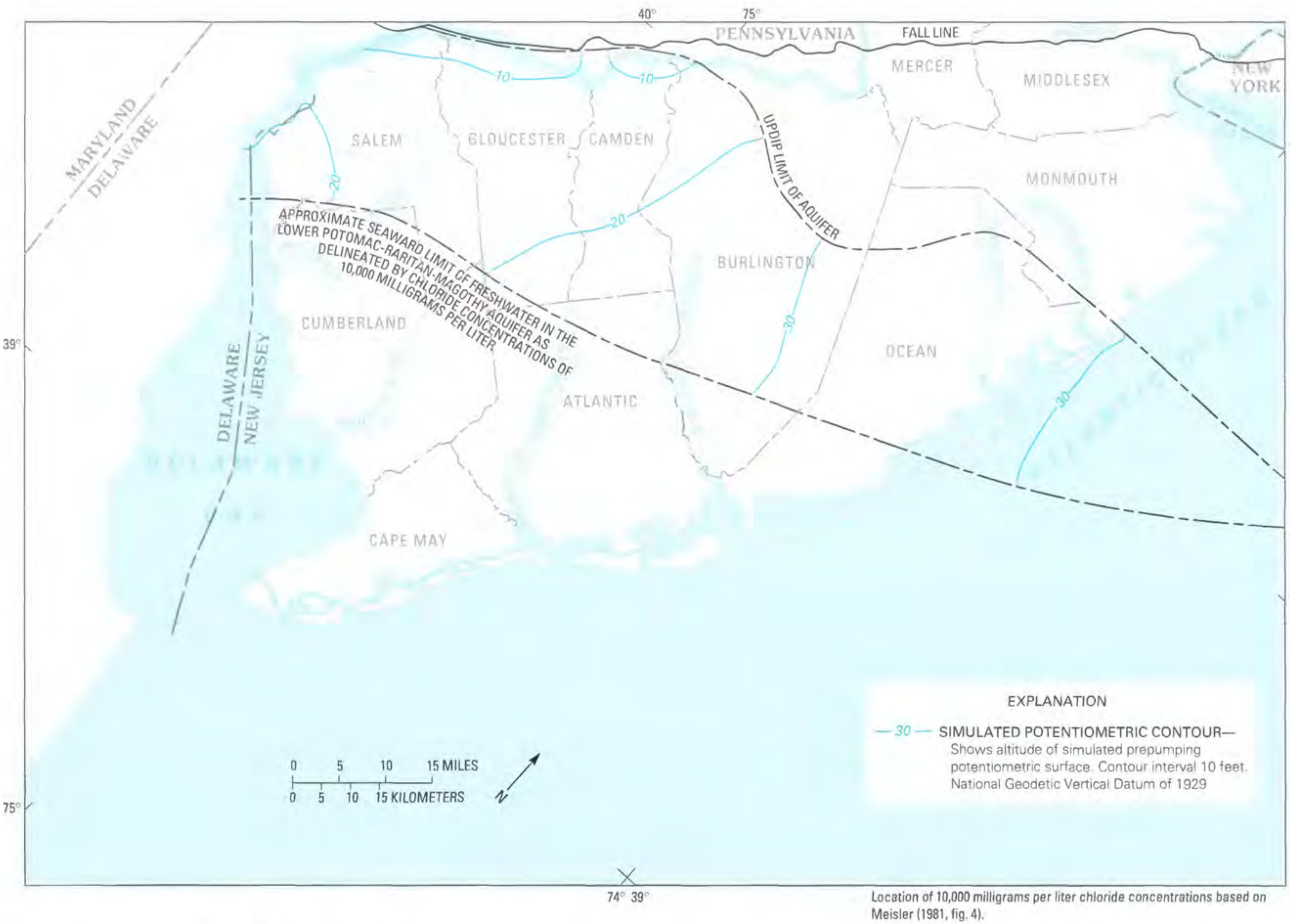
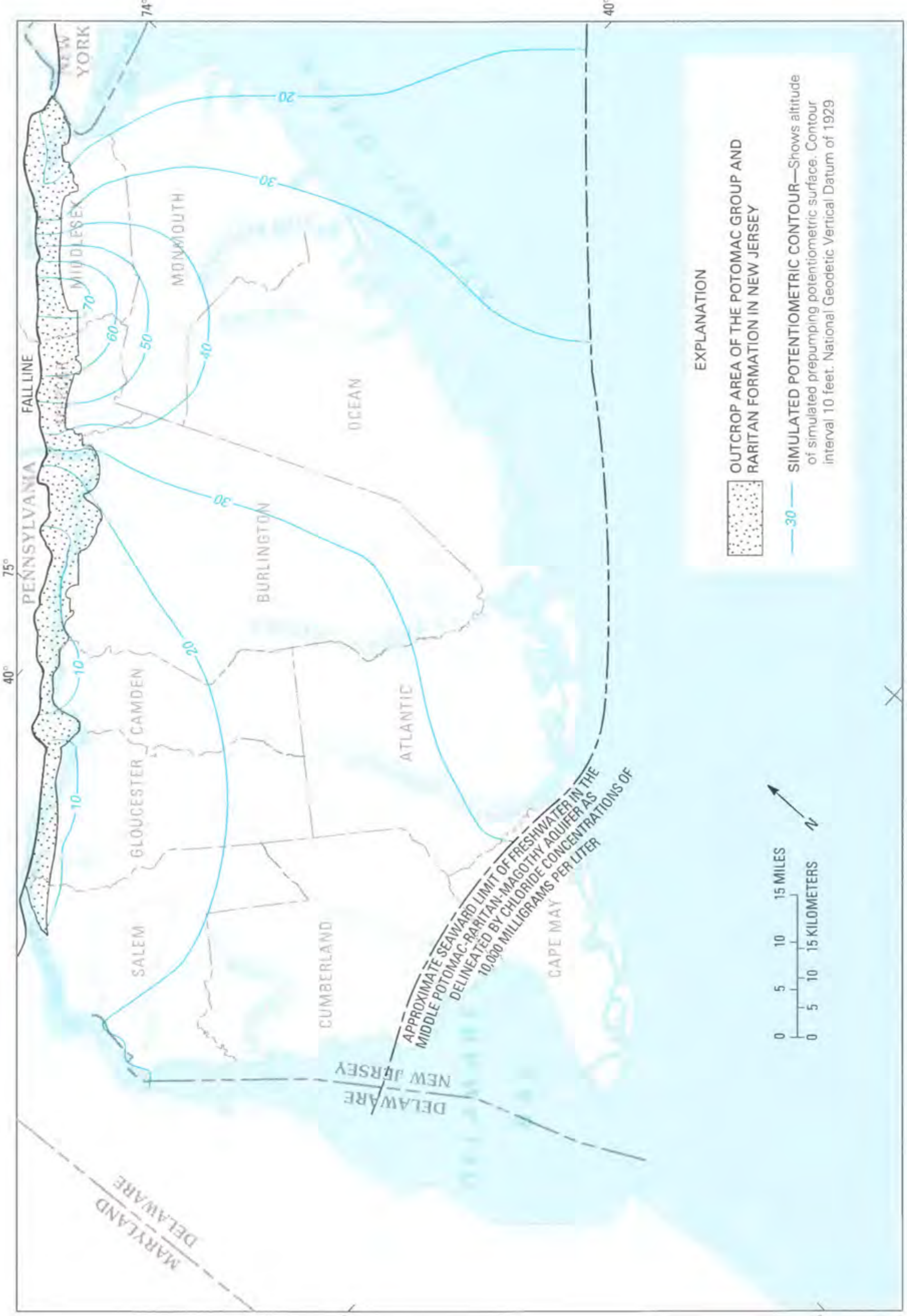
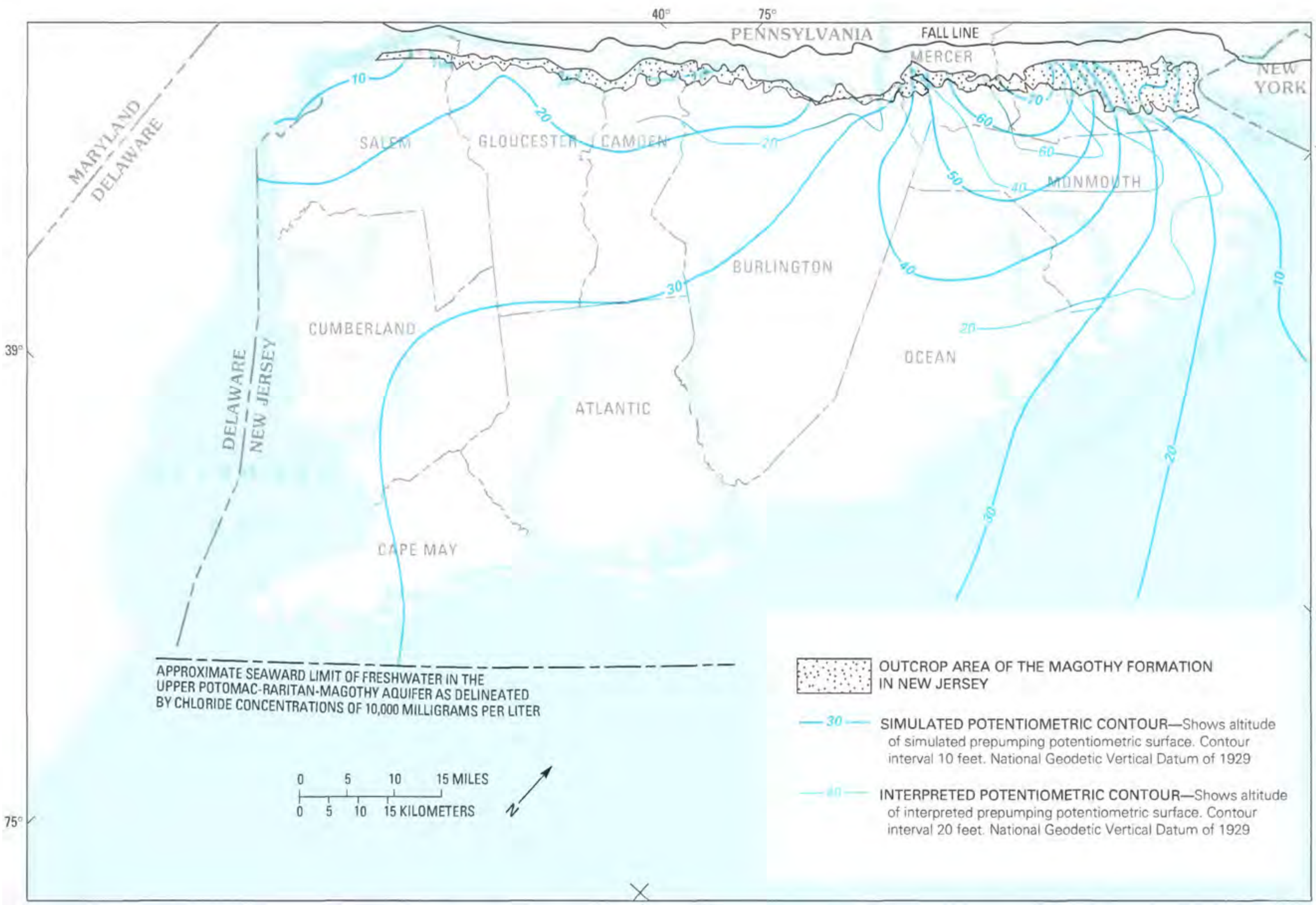


FIGURE 30.—Simulated prepumping potentiometric surface of the lower Potomac-Raritan-Magothy aquifer (model unit A1).



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 31.—Simulated prepumping potentiometric surface of the middle Potomac-Raritan-Magothy aquifer (model unit A2).



Interpreted prepumping potentiometric contours from Zapecza, Voronin, and Martin (1987, fig. 4). Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 32.—Simulated and interpreted prepumping potentiometric surfaces for the upper Potomac-Raritan-Magothy aquifer (model unit A3).

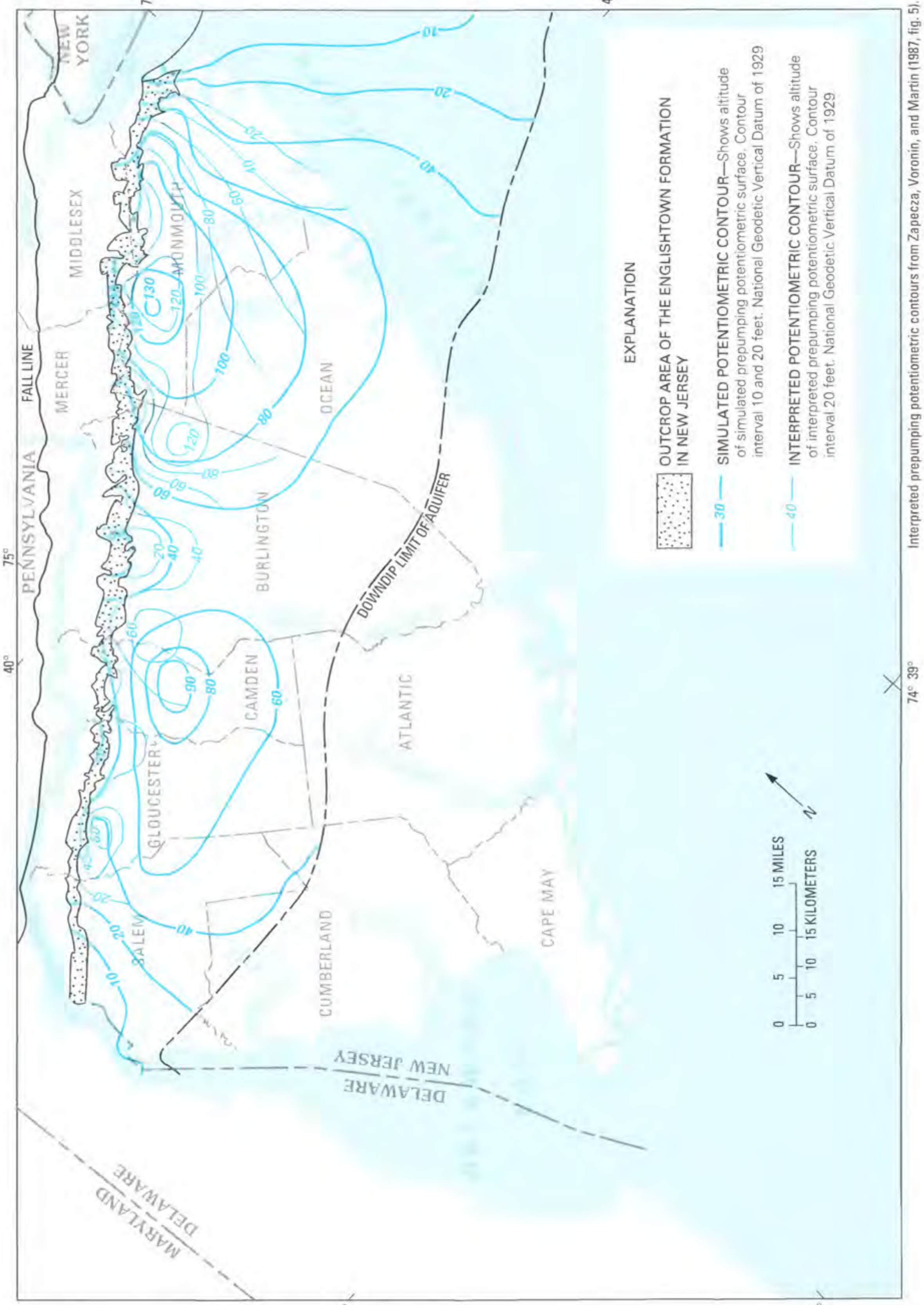


FIGURE 33.—Simulated and interpreted prepumping potentiometric surfaces for the Englishtown aquifer (model unit A4).

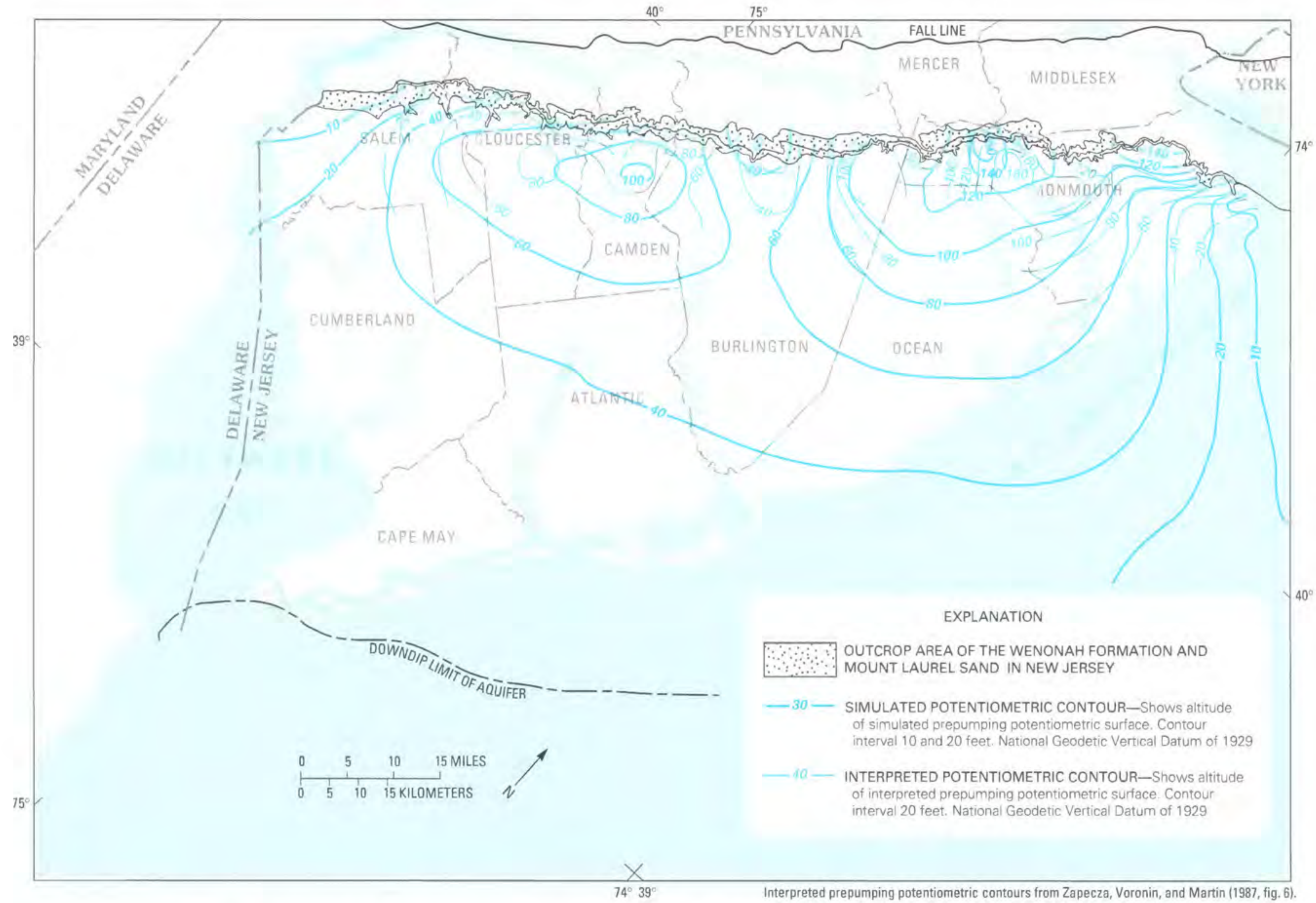
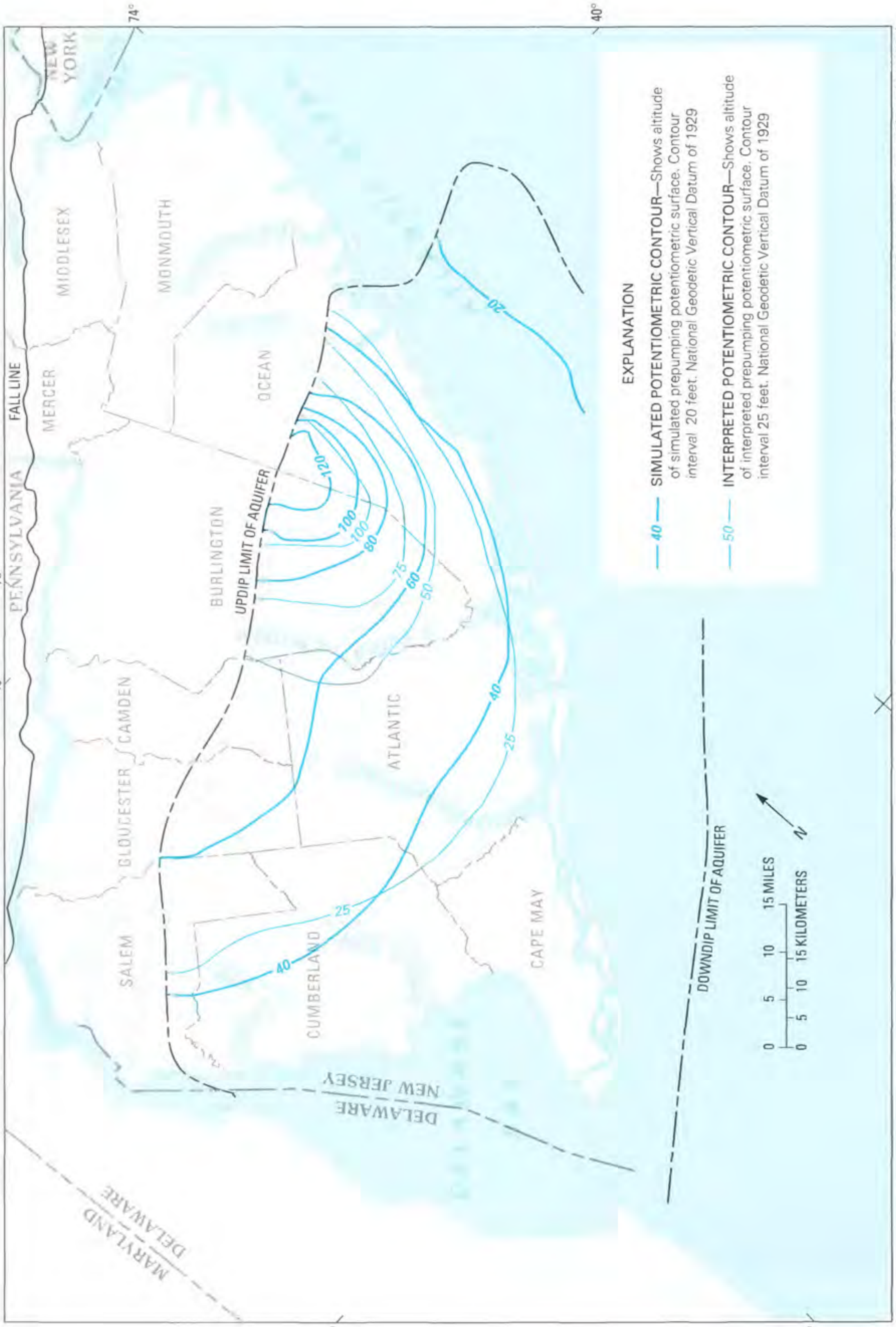


FIGURE 34.—Simulated and interpreted prepumping potentiometric surfaces for the Wenonah-Mount Laurel aquifer (model unit A5).

Interpreted prepumping potentiometric contours from Zapecza, Voronin, and Martin (1987, fig.7).

FIGURE 35.—Simulated and interpreted prepumping potentiometric surfaces for the Vincentown aquifer (model unit A6).



Interpreted prepumping potentiometric contours from Zapecza, Voronin, and Martin (1987, fig.8).

FIGURE 36.—Simulated and interpreted prepumping potentiometric surfaces for the Piney Point aquifer (model unit A7).

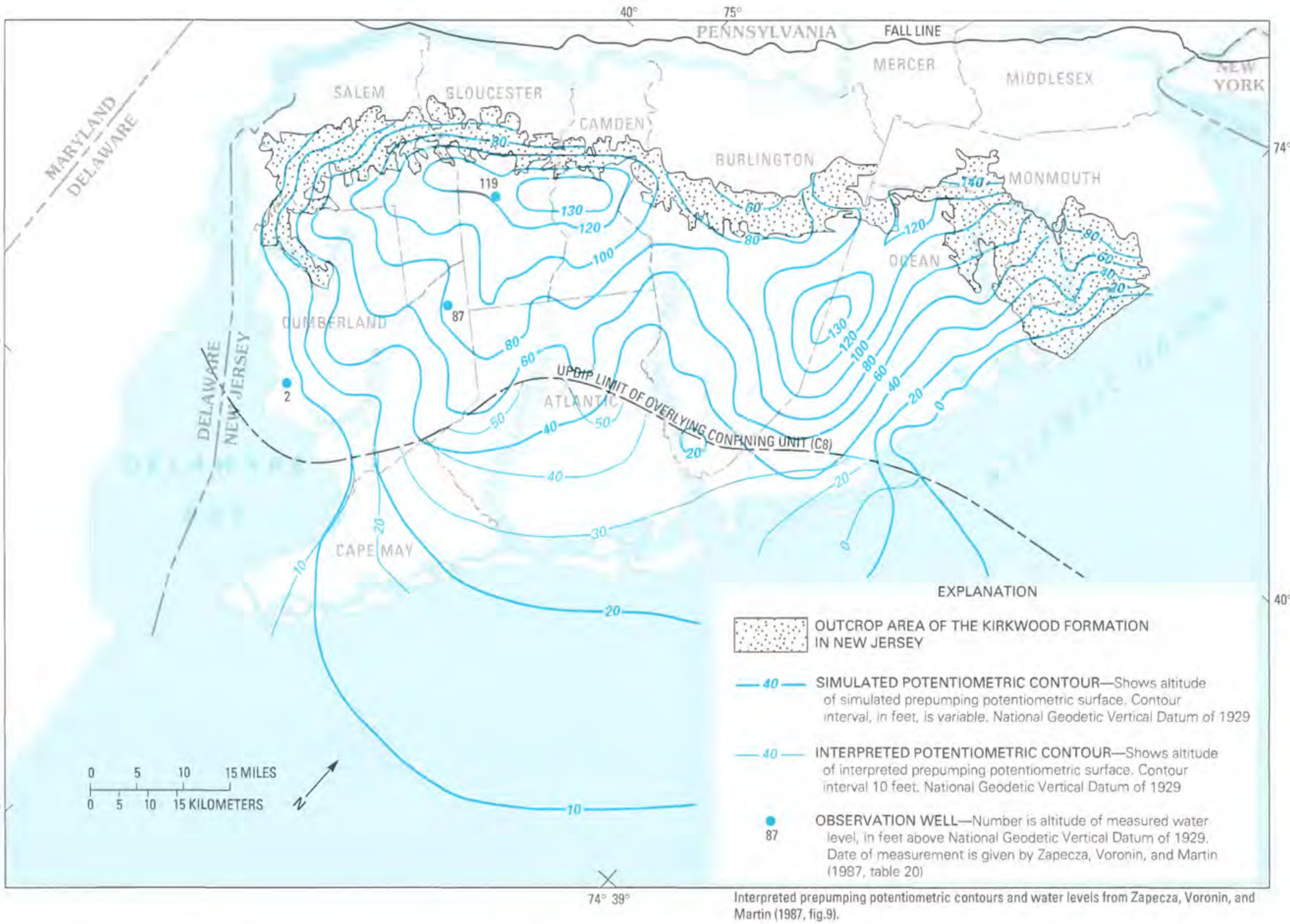


FIGURE 37.—Simulated and interpreted prepumping potentiometric surfaces and measured prepumping water levels for the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (model unit A8).

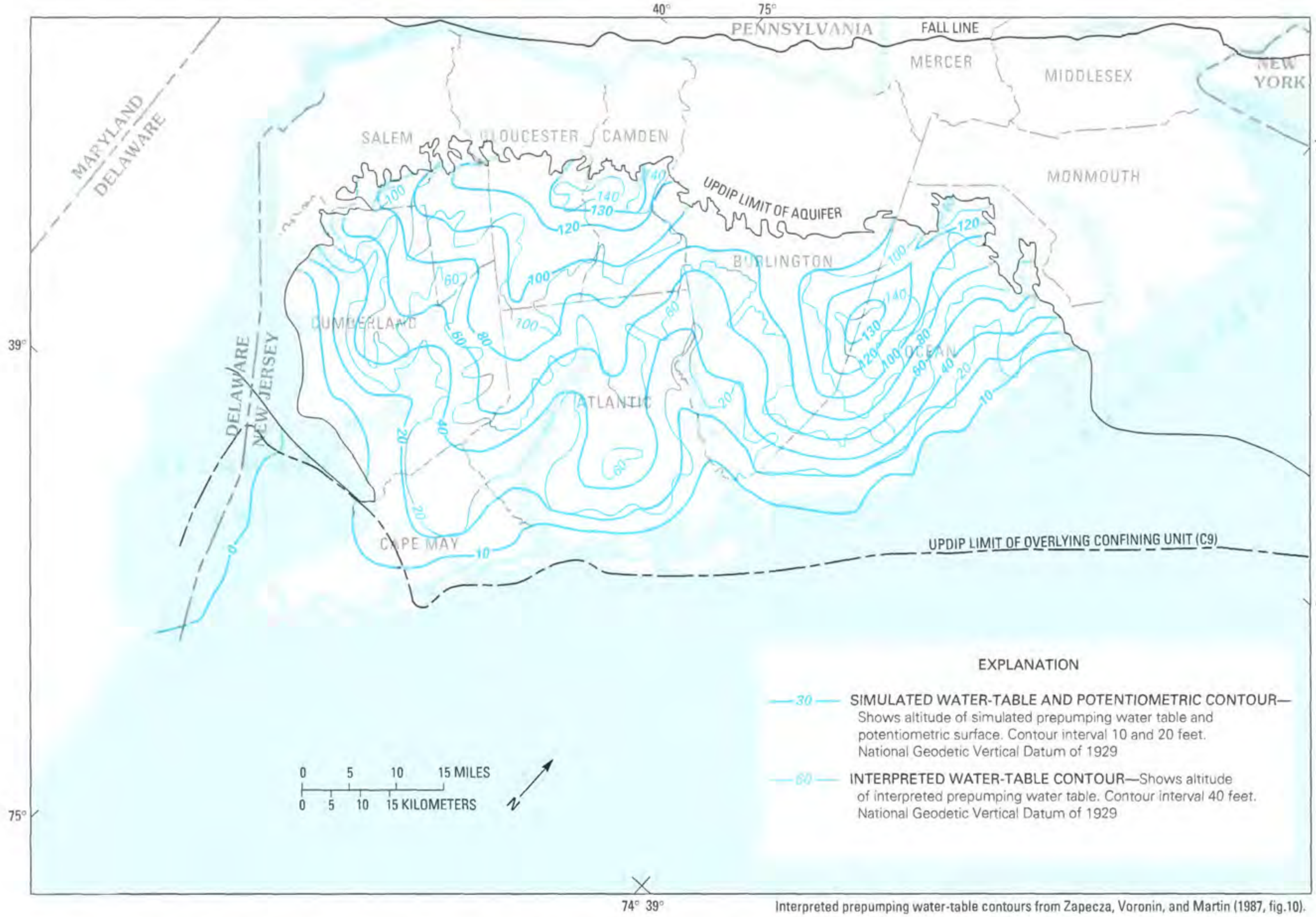


FIGURE 38.—Simulated and interpreted prepumping water tables and potentiometric surfaces for the upper Kirkwood-Cohansey aquifer (model unit A9).

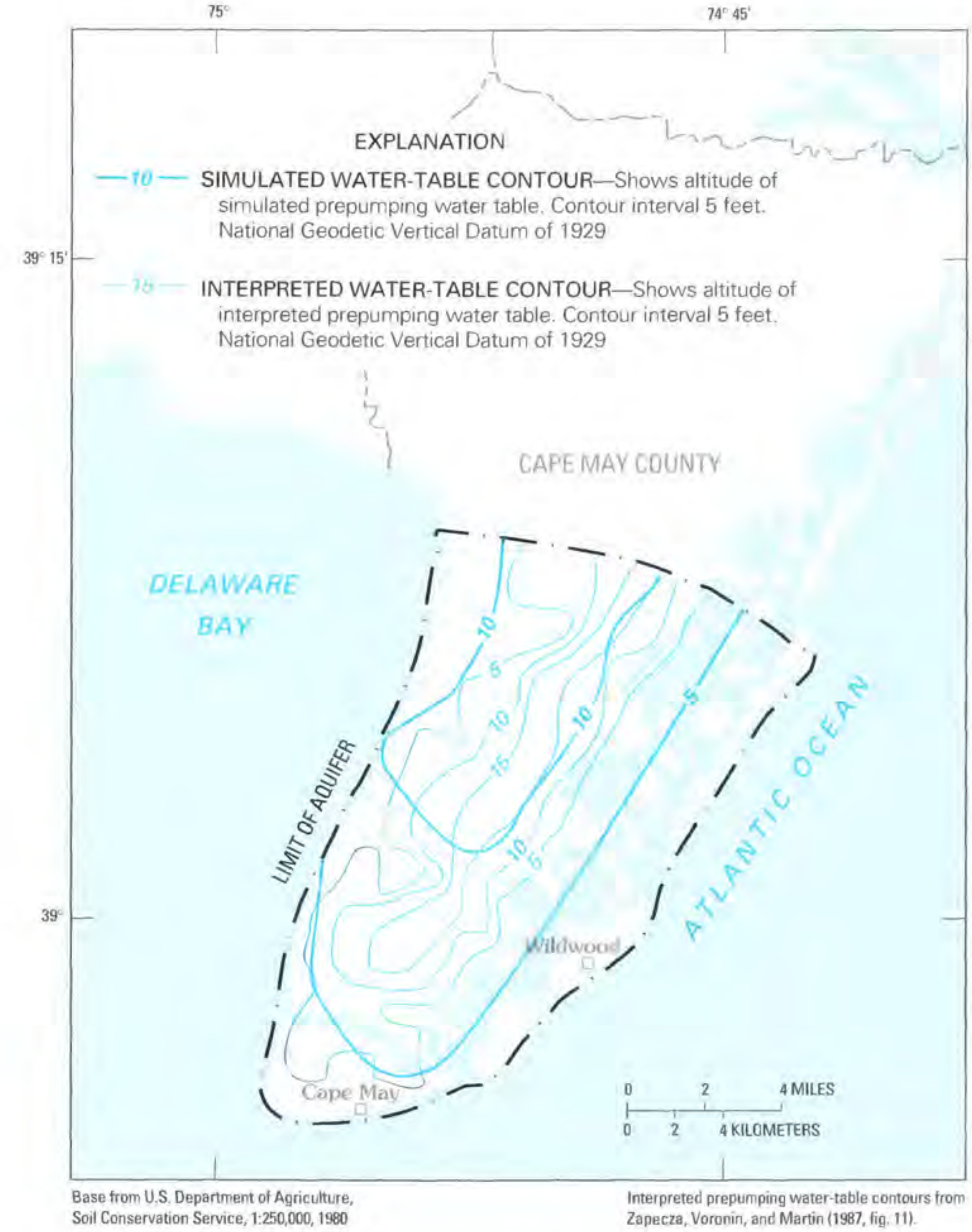


FIGURE 39.—Simulated and interpreted prepumping water tables for the Holly Beach aquifer (model unit A10).

altitudes cannot be simulated because of the large cell size. Also, the outcrop area for the Vincentown aquifer is poorly represented by the large cell size; therefore, local recharge and discharge areas cannot be simulated accurately.

The simulated and interpreted prepumping potentiometric surfaces for the Piney Point aquifer (A7) are shown in figure 36. Potentiometric contours are closely simulated except in the southwestern Coastal Plain and along the Atlantic Coast where simulated contours are about 20 feet higher than the interpreted contours. Normally, a 20-ft difference between the simulated and interpreted potentiometric surfaces over a large area is not acceptable. However, the general trend of the potentiometric surface has been simulated, and relatively few water-level measurements are available; therefore, further calibration was not considered beneficial.

The simulated and interpreted prepumping potentiometric surfaces for the lower Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer (A8) are shown in figure 37. The simulated potentiometric surface matches the interpreted potentiometric surface closely in the confined parts of the aquifer. In the unconfined areas, water-level altitudes in the lower Kirkwood-Cohansey aquifer are very similar to the water-table altitudes in the overlying upper Kirkwood-Cohansey aquifer (A9) and strongly reflect the local flow patterns determined by stream drainage patterns. In the confined Kirkwood aquifer (A8), local flow patterns are less apparent, and simulated water levels decrease downdip from the western extent of the confining unit to beneath the Atlantic Ocean.

The interpreted prepumping potentiometric surfaces for the upper Kirkwood-Cohansey aquifer (A9) and the Holly Beach aquifer (A10) are almost entirely water-table surfaces, and potentiometric contours are closely controlled by local streams. These potentiometric surfaces have been simulated as closely as possible with the cell size used in the model. The simulated and interpreted surfaces for the upper Kirkwood-Cohansey aquifer and the Holly Beach aquifer are shown in figures 38 and 39, respectively. Generally, the altitude of the simulated water table is more than 100 feet above sea level in central Ocean, central Burlington, central Camden, and central Gloucester Counties. Low water-table altitudes are simulated near the Atlantic Ocean and Delaware Bay coastlines and along large rivers.

Flow to and from the confined aquifers and flow to and from streams were analyzed for prepumping steady-state conditions to determine whether the simulated flows are comparable to rates and areas of recharge and discharge postulated in the conceptual model. In the 10-layer steady-state simulations, flow between the unconfined outcrop areas and the underlying confined aquifers is represented by flow to and from the water-table constant-head nodes in the unconfined parts of all of the model units, except the lower Potomac-Raritan-Magothy aquifer (A1) and the Piney Point aquifer (A7), which have no water-table constant-head nodes, as they are completely confined in New Jersey.

Simulated flow to and from the confined aquifers for prepumping conditions is shown in figure 40. Flow to and from the confined aquifers is flow to and from the water-table constant-head nodes in the 10-layer steady-state model. Generally, simulated recharge and discharge rates for the confined aquifers are less than 5 in/yr. However, the largest discharge to the water table is 8 in/yr, and the largest recharge to the confined aquifers is 9 in/yr. These high values are not shown in figure 40 because of the 5 in/yr interval but are in the area where the upper Kirkwood-Cohansey aquifer (A9) directly overlies the lower Kirkwood-Cohansey aquifer (A8) with no intervening confining unit. In this area, 80 percent of the recharge and discharge values are from 0.0 to 4.5 in/yr. Steady-state calibration of the lower and upper Kirkwood-Cohansey aquifers (A8 and A9) included decreasing transmissivity of these units to decrease the amount of flow between the two aquifers in this area to less than 15 in/yr.

In most of the other onshore areas in New Jersey, simulated flow represents flow to or from the unconfined aquifer outcrops through an underlying confining unit to a confined aquifer. In these areas with an underlying confining unit, 90 percent of the recharge and discharge values are less than 1.5 in/yr. Simulated discharge from the confined aquifers in offshore areas is generally less than 0.01 in/yr.

Simulated recharge areas coincide with land-surface highs, and discharge areas coincide with low areas along streams and large surface-water bodies. Highest recharge and discharge rates are simulated in areas where there is no underlying confining unit to impede flow, and flow in areas with an underlying confining unit is only a small part, less than 10 percent, of the assumed 20 in/yr of recharge.

In the 11-layer steady-state model, flow from constant-head nodes representing stream stages simulates prepumping flow between the water-table and the streams. Simulated flow to streams is shown in figure 41. Generally, flow to streams is between 10 and 30 in/yr, although the highest flow is 38 in/yr and the lowest is 7 in/yr. The extreme values of simulated flow to streams are not shown in figure 41 because of the 5 in/yr interval but are in the area where the lower and upper of the Kirkwood-Cohansey aquifers (A8 and A9) have no intervening confining unit. In this area, about 70 percent of the simulated base-flow values are from 13.5 to 26.5 in/yr. In other areas, where the water-table aquifer is underlain by a confining unit, about 90 percent of the flow to streams is from 16.5 to 23.5 in/yr. The difference between the 20 in/yr of simulated recharge to the water table and the flow to streams is vertical and lateral flow from or to the unconfined outcrop areas. In areas where the flow to streams is less than 20 in/yr, ground-water flow is laterally and vertically out of the unconfined outcrop areas into confined aquifers, and in areas where flow to streams is more than 20 in/yr, ground-water flow is laterally and vertically from confined aquifers into the unconfined outcrop areas and then upward to streams.

Areas with more than 20 in/yr of flow to streams coincide with onshore discharge areas along major streams, and areas with less than 20 in/yr of flow to streams coincide with onshore recharge areas. In areas where the unconfined outcrop areas are underlain by a confining unit, most of the 20 in/yr of recharge flows to the streams, and simulated flow to streams is generally within 3.5 in/yr of this recharge value. In the area where there is no confining unit underlying the outcrop areas, flow to the streams is generally within 10 in/yr of the 20 in/yr of the applied recharge.

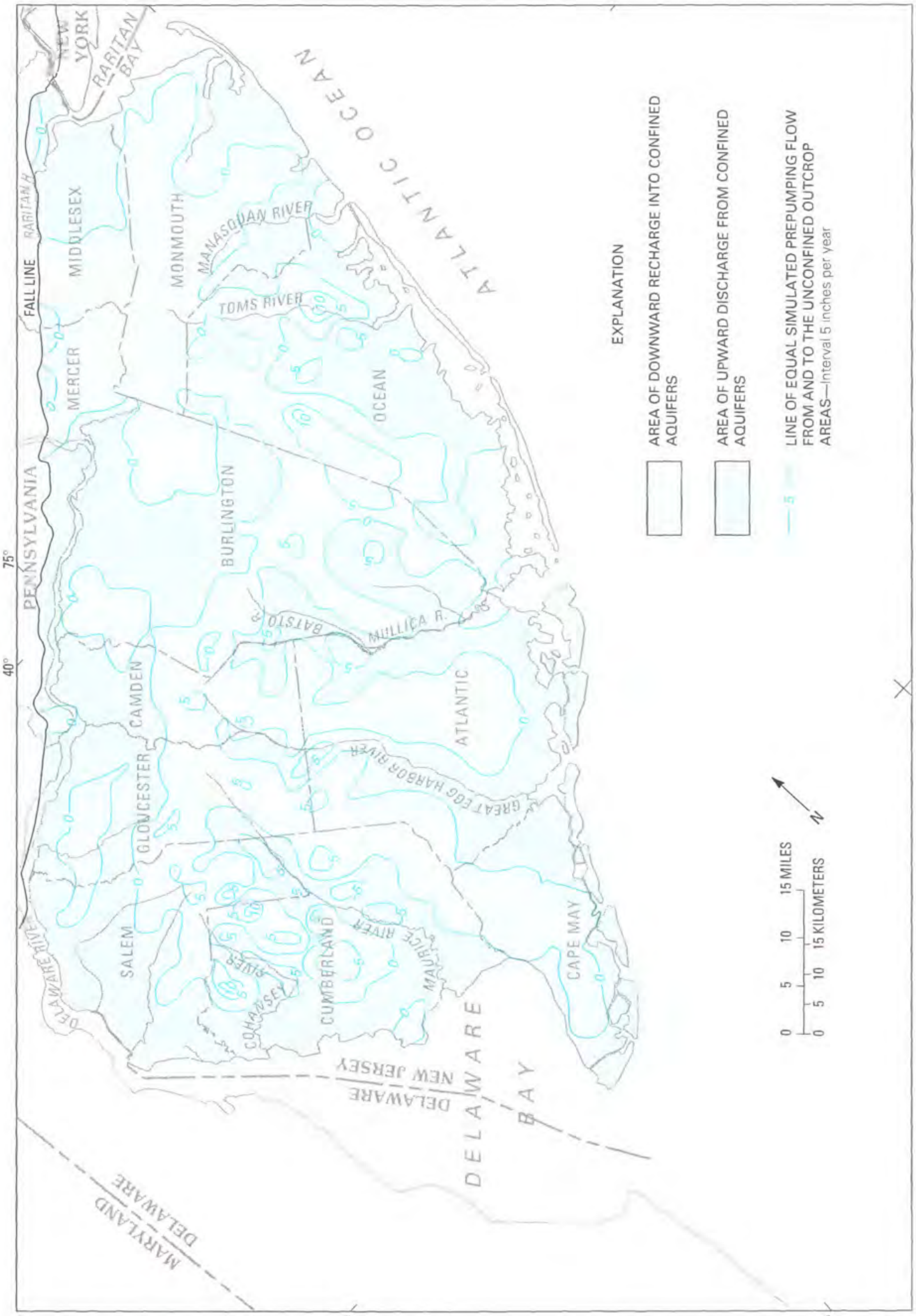


FIGURE 40.—Simulated prepumping ground-water flow from and to the unconfined outcrop areas in the 10-layer model.

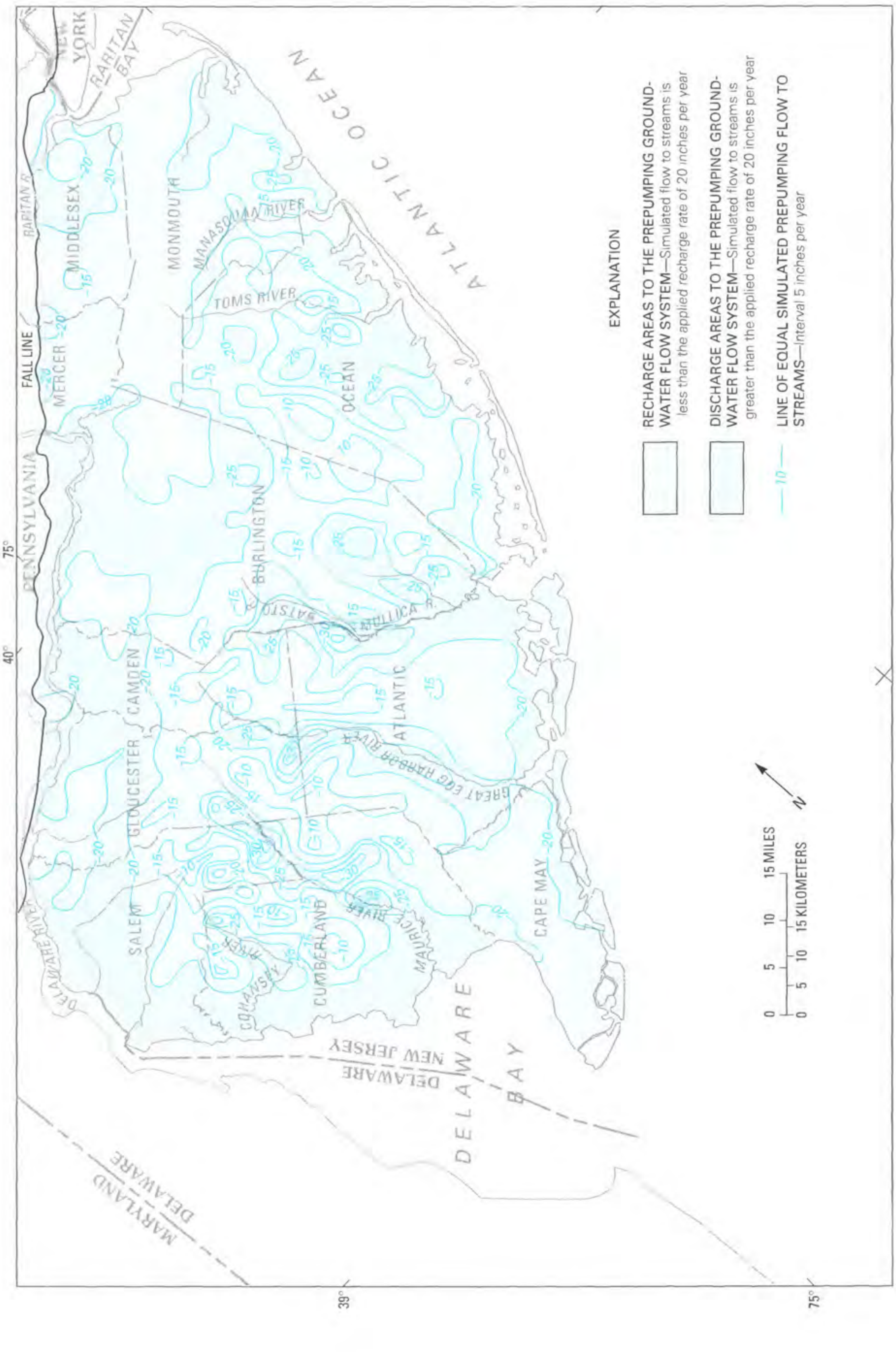


FIGURE 41.—Simulated prepumping ground-water flow to streams in the 11-layer model.

SIMULATION OF TRANSIENT CONDITIONS

Generally, the interpreted 1978 potentiometric surface is simulated closely by the transient model results. Figures 42 to 51 show both the simulated and interpreted 1978 potentiometric surfaces. A comparison of simulated and measured water levels at the end of each pumping period is shown in table 15, which follows the references cited section. Simulated and measured hydrographs are shown in figures 52 to 54 for the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3), the Englishtown aquifer (A4), the Wenonah-Mount Laurel aquifer (A5), and the confined Kirkwood aquifer (A8), and the upper Kirkwood-Cohansey aquifer (A9). The configuration of the potentiometric surfaces has been simulated closely in most of the aquifers, although in some areas, and in some aquifers, the match between simulated and interpreted potentiometric surfaces is poor. However, the match between simulated and measured hydrographs is considered to be good. The comparison of simulated and measured transient water levels is discussed for each aquifer below.

In the lower Potomac-Raritan-Magothy aquifer (A1), the center of the large cone of depression centered in Camden County is simulated (fig. 42) about 10 feet shallower than the interpreted cone. Simulated water levels near the Delaware River do not show local variations in the potentiometric surface. Simulated hydrographs for wells 7–354 and 15–296 (table 15) average about 11 feet lower than the measured water levels and the simulated hydrograph for well 15–323 (table 15) averages about 18 feet above measured water levels. These differences are mostly the result of discretization scale, because the simulated hydrographs were calculated from water levels averaged over large cell areas.

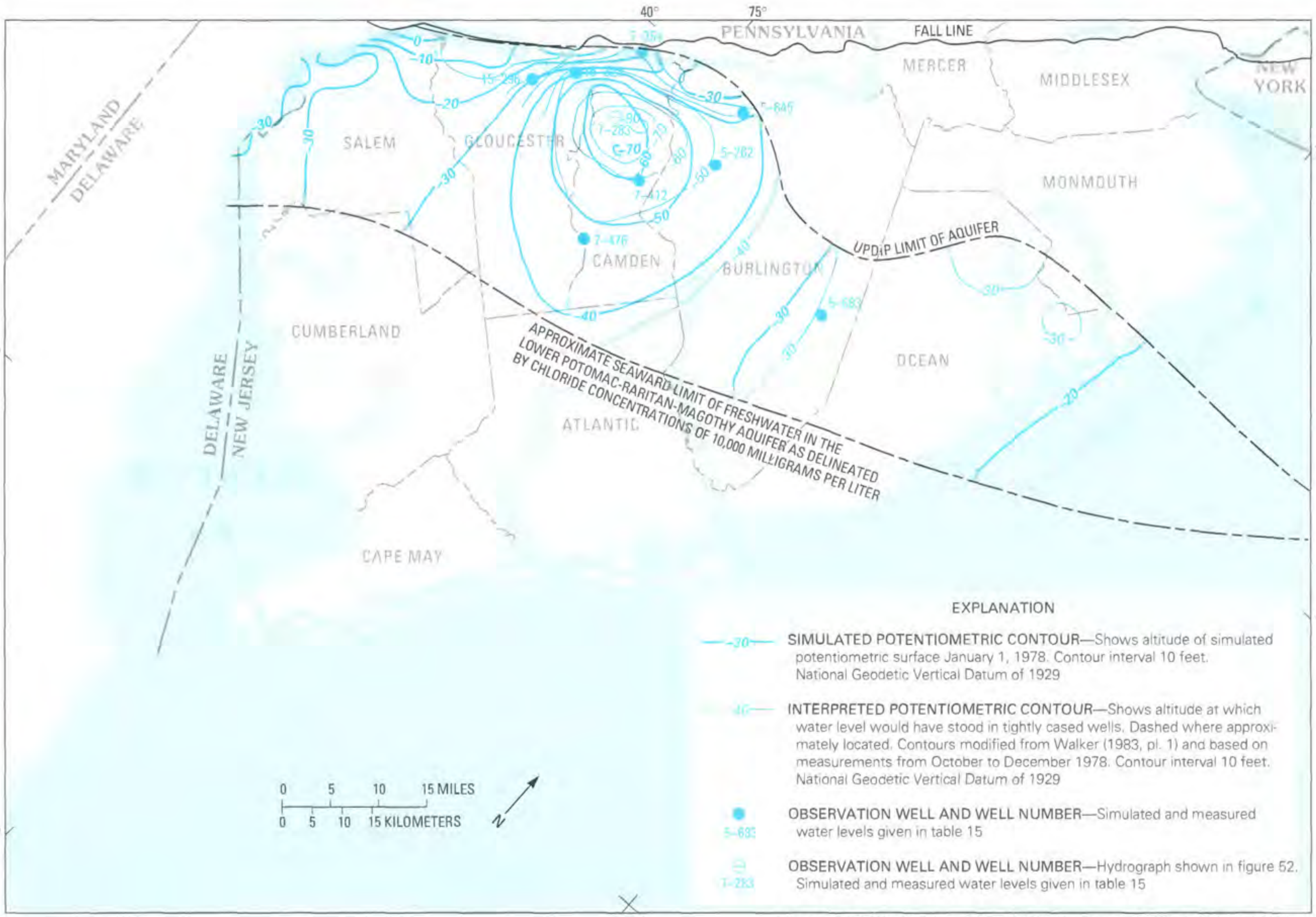
The interpreted potentiometric surface of the middle Potomac-Raritan-Magothy aquifer (A2) has been closely simulated, generally, to within 10 feet (fig. 43). However, the centers of the simulated cones of depression are as much as 20 feet shallower than the interpreted cones because of the large cell size. A moderate-sized cone at Artificial Island in Salem County has not been simulated, possibly because of incorrect or insufficient simulated withdrawal data. The only large area considered to be poorly simulated is downdip in Burlington and Ocean Counties. This area has simulated water levels 15 to 25 feet lower than measured water levels, as shown in table 15 for well 29–19.

The interpreted and simulated 1978 potentiometric surfaces of the upper Potomac-Raritan-Magothy aquifer (A3) flow system are shown in figure 44. Water levels in the large cone centered in Camden County and the high ground-water altitudes near the outcrop areas in Mercer and Middlesex Counties are simulated to within 10 feet of measured water levels. However, the downdip area along the coast in Monmouth and Ocean Counties is poorly simulated. Although no water levels were measured in this area, based on observed and simulated water levels in this area for the middle Potomac-Raritan-Magothy aquifer (A2), simulated water levels probably are 15 to 20 feet lower than actual water levels. Another poorly simulated area is in Salem County near Delaware Bay, where simulated water levels are as much as 20 feet higher than measured water levels. Simulated lateral hydraulic gradients are from Delaware toward Gloucester County, whereas measured water levels indicate gradients are downdip toward Cumberland County. The reason for this difference is unknown.

Simulated 1978 water levels for the Englishtown aquifer (A4) closely match the interpreted 1978 potentiometric surface in the northern part of the aquifer (fig. 45). Although some simulated water levels (wells 25–429 and 29–534, table 15) near the very steep cone of depression in Ocean and Monmouth Counties are about 40 feet higher than measured water levels, the area is simulated as closely as possible with the cell size used in that area. Measured 1978 water levels for the Englishtown aquifer in central Camden County as shown by Walker (1983, pl. 3) were incorrect, and 1983 measured water levels are about 40 feet above sea level (Eckel and Walker, 1986, pl. 4). Therefore, the simulated water levels of 60 feet (fig. 45) are probably 20 feet too high. A local groundwater low near the Burlington-Monmouth County line is not simulated, although the simulated water levels are acceptable in this area.

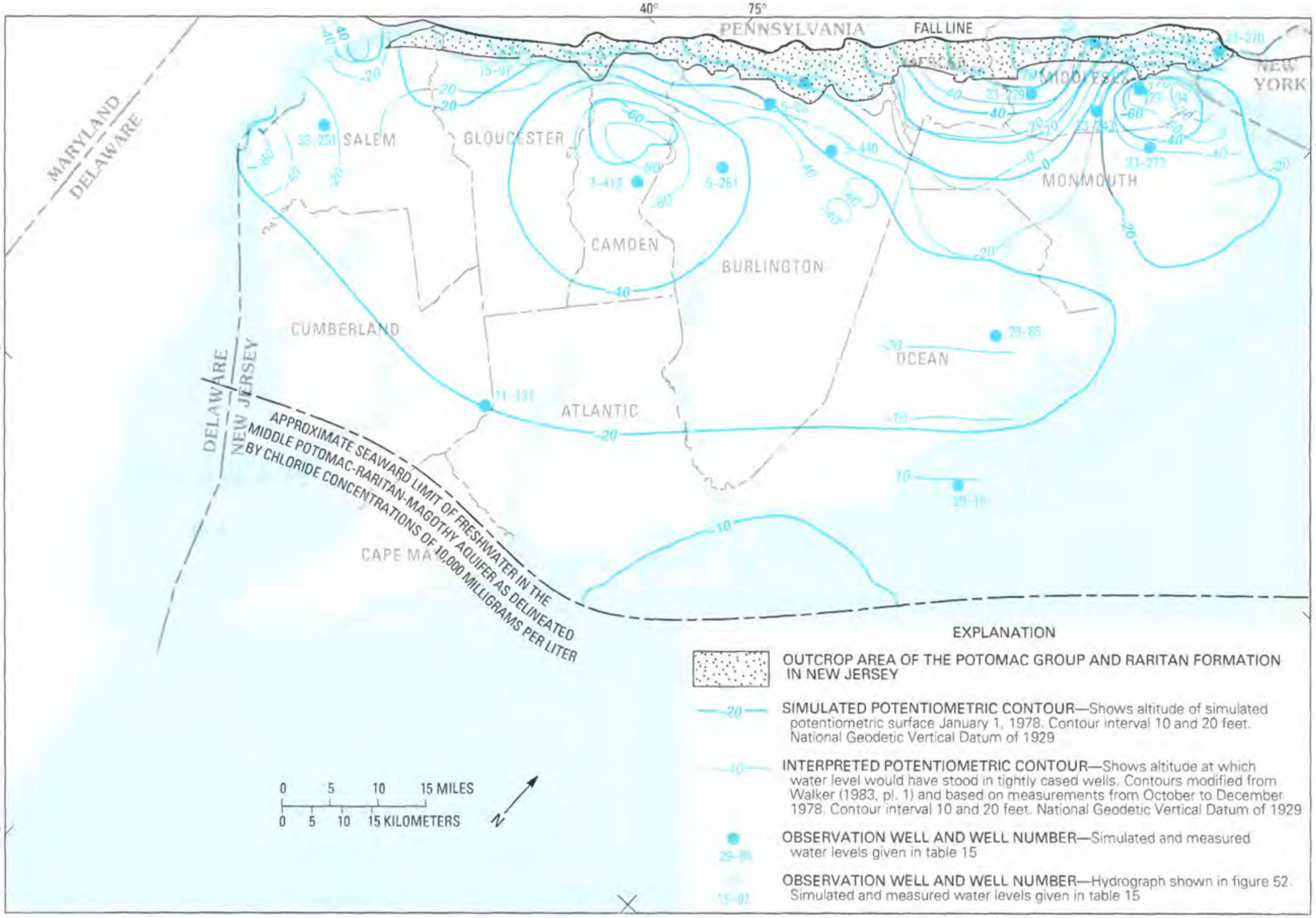
Major features in the 1978 flow system of the Wenonah-Mount Laurel aquifer (A5) are similar to those in the Englishtown aquifer (A4). The high ground-water levels along most of the outcrop, the ground-water low near the outcrop in central Burlington County, and the large cone of depression centered along the coast of Monmouth and Ocean Counties are simulated (fig. 46). A local ground-water low in western Monmouth and Ocean Counties is not simulated, probably because of the cell size in the area. With no water-level data available in downdip areas, the accuracy of the simulated water levels in this area could not be evaluated.

Simulated water levels in the Vincentown aquifer (A6) were not calibrated because there is only one observation well at which measured and simulated water levels can be compared. Simulated water levels are about 12 feet lower than measured water levels at the observation-well site in western Ocean County (well 29–139, table 15). The simulated 1978 potentiometric surface is consistent with the conceptual model and the simulated water levels in adjacent aquifers. Simulated water levels for the Vincentown aquifer (A6) are shown in figure 47. The configuration of simulated potentiometric surface for the Vincentown aquifer is similar to the simulated



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 42.—Simulated and interpreted 1978 potentiometric surfaces for the lower Potomac-Raritan-Magothy aquifer (model unit A1).



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 43.—Simulated and interpreted 1978 potentiometric surfaces for the middle Potomac-Raritan-Magothy aquifer (model unit A2).

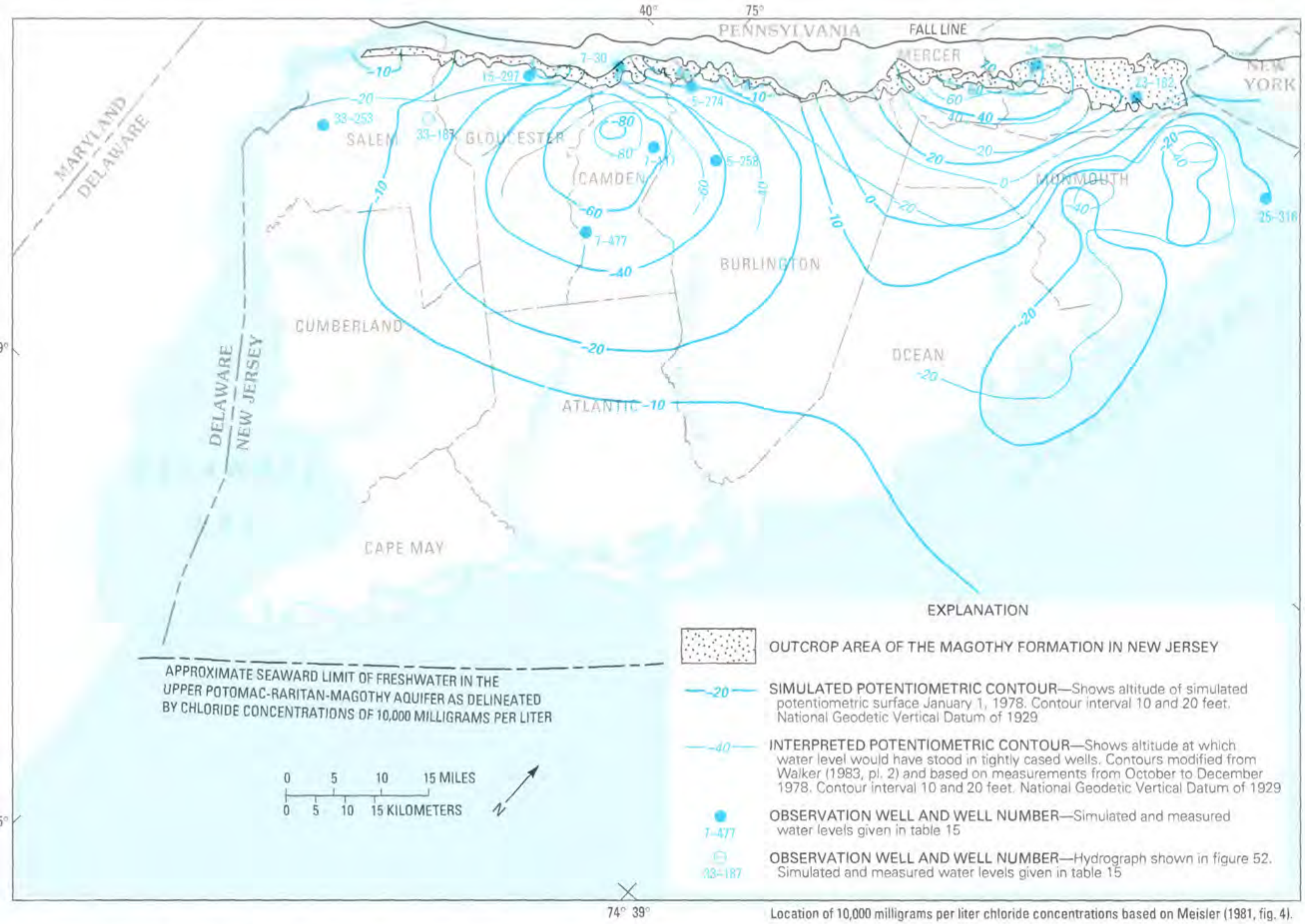


FIGURE 44.—Simulated and interpreted 1978 potentiometric surfaces for the upper Potomac-Raritan-Magothy aquifer (model unit A3).

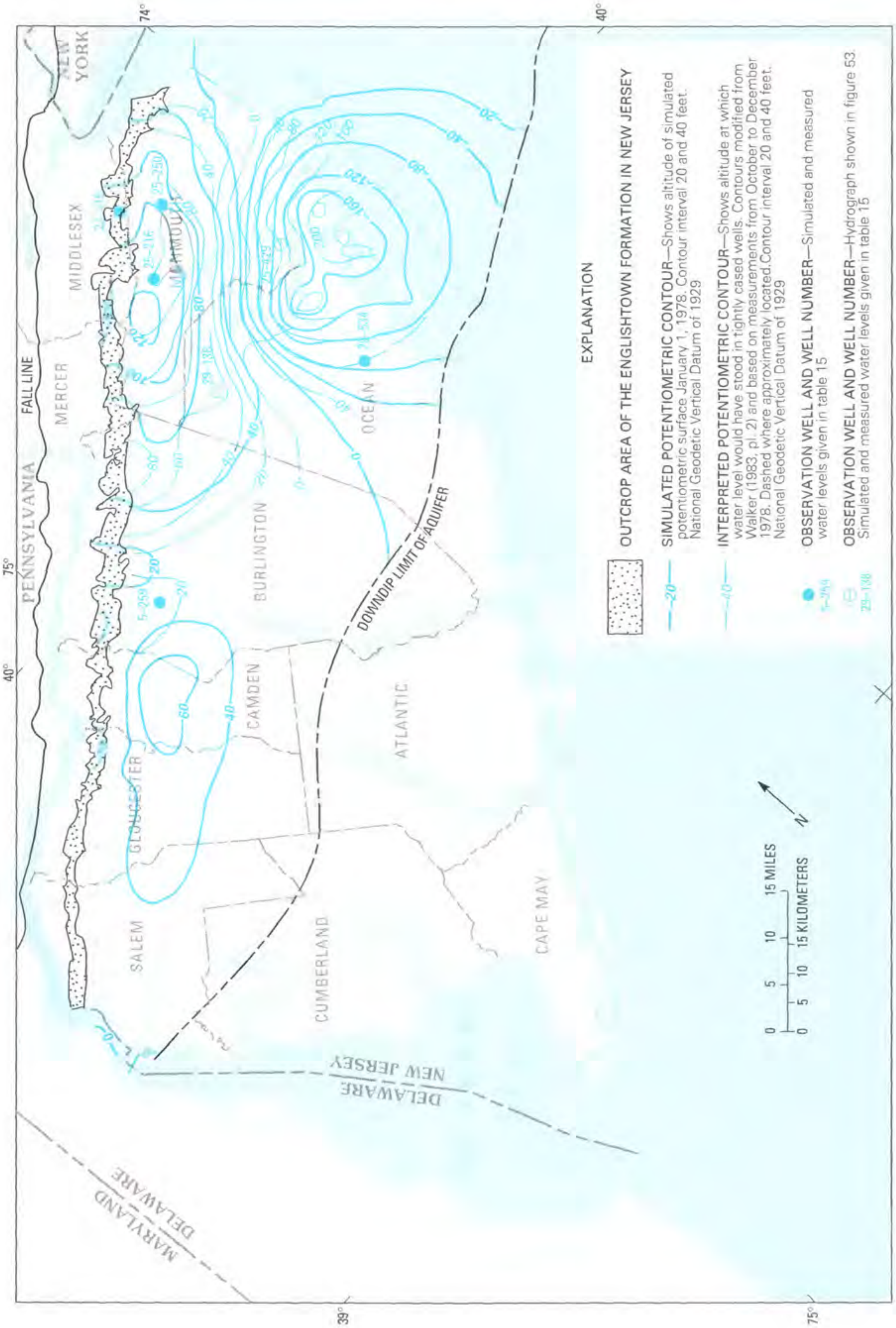


FIGURE 45.—Simulated and interpreted 1978 potentiometric surfaces for the Englishtown aquifer (model unit A4).

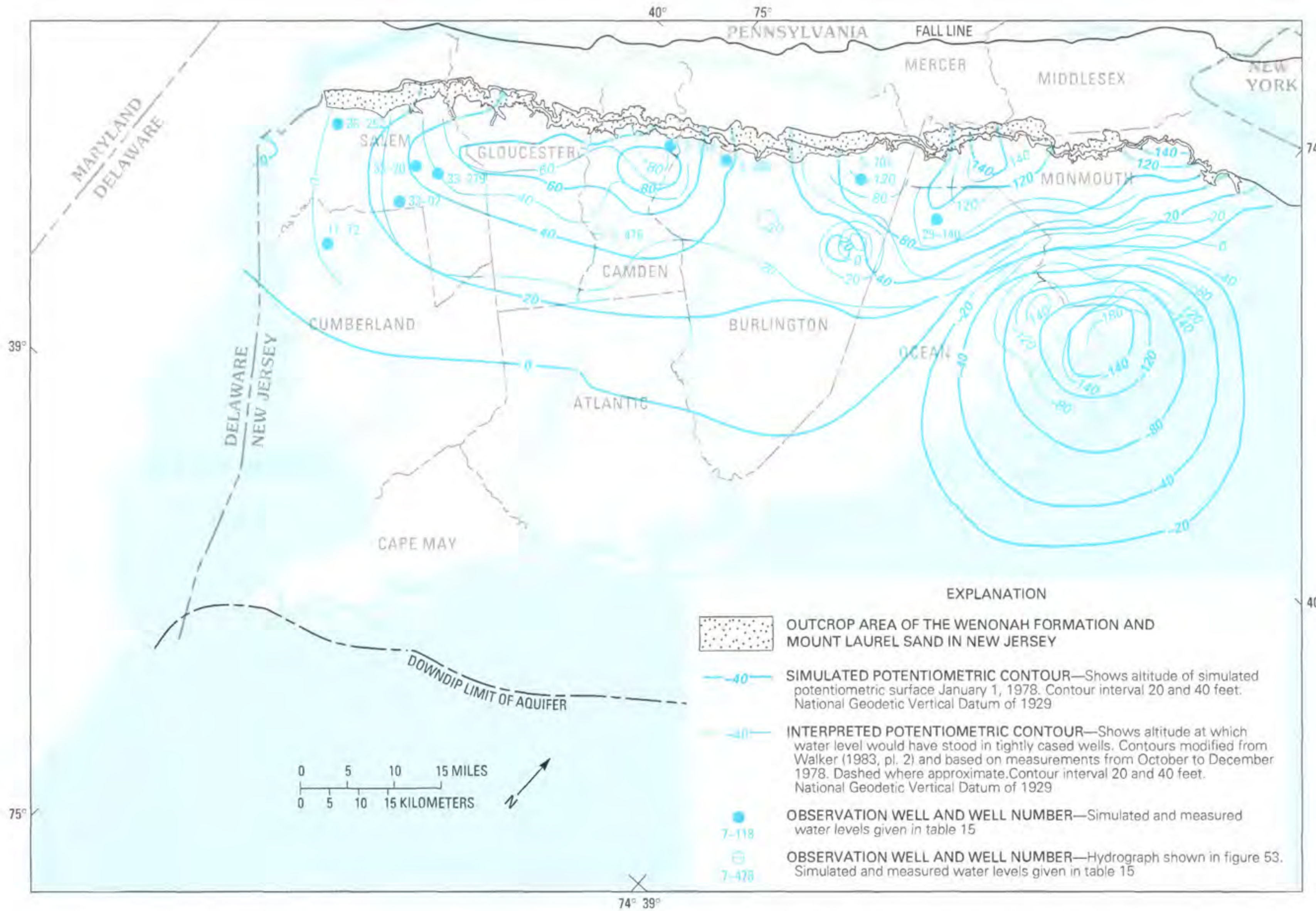


FIGURE 46.—Simulated and interpreted 1978 potentiometric surfaces for the Wenonah-Mount Laurel aquifer (model unit A5).

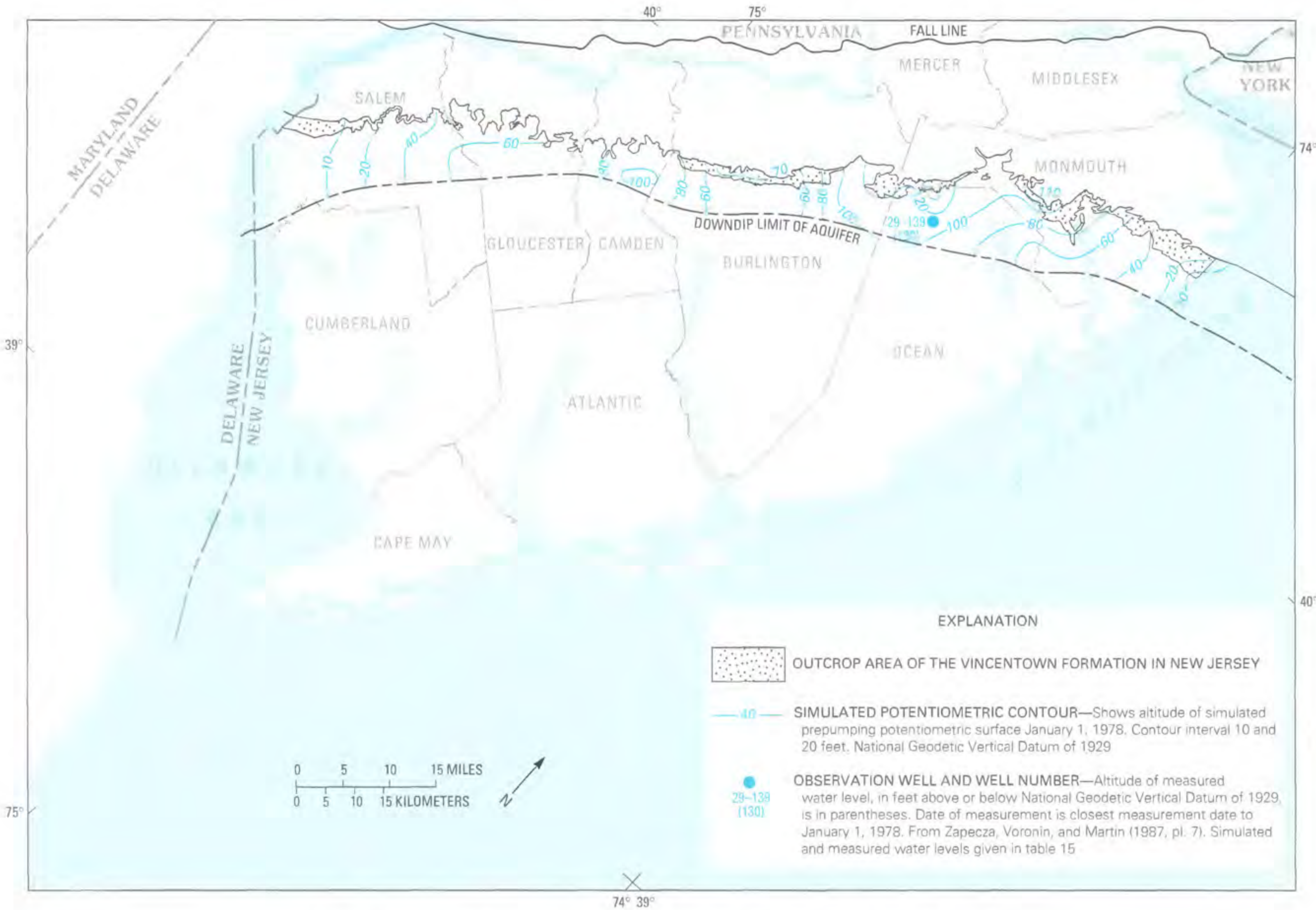


FIGURE 47.—Simulated 1978 potentiometric surface for the Vincentown aquifer (model unit A6).

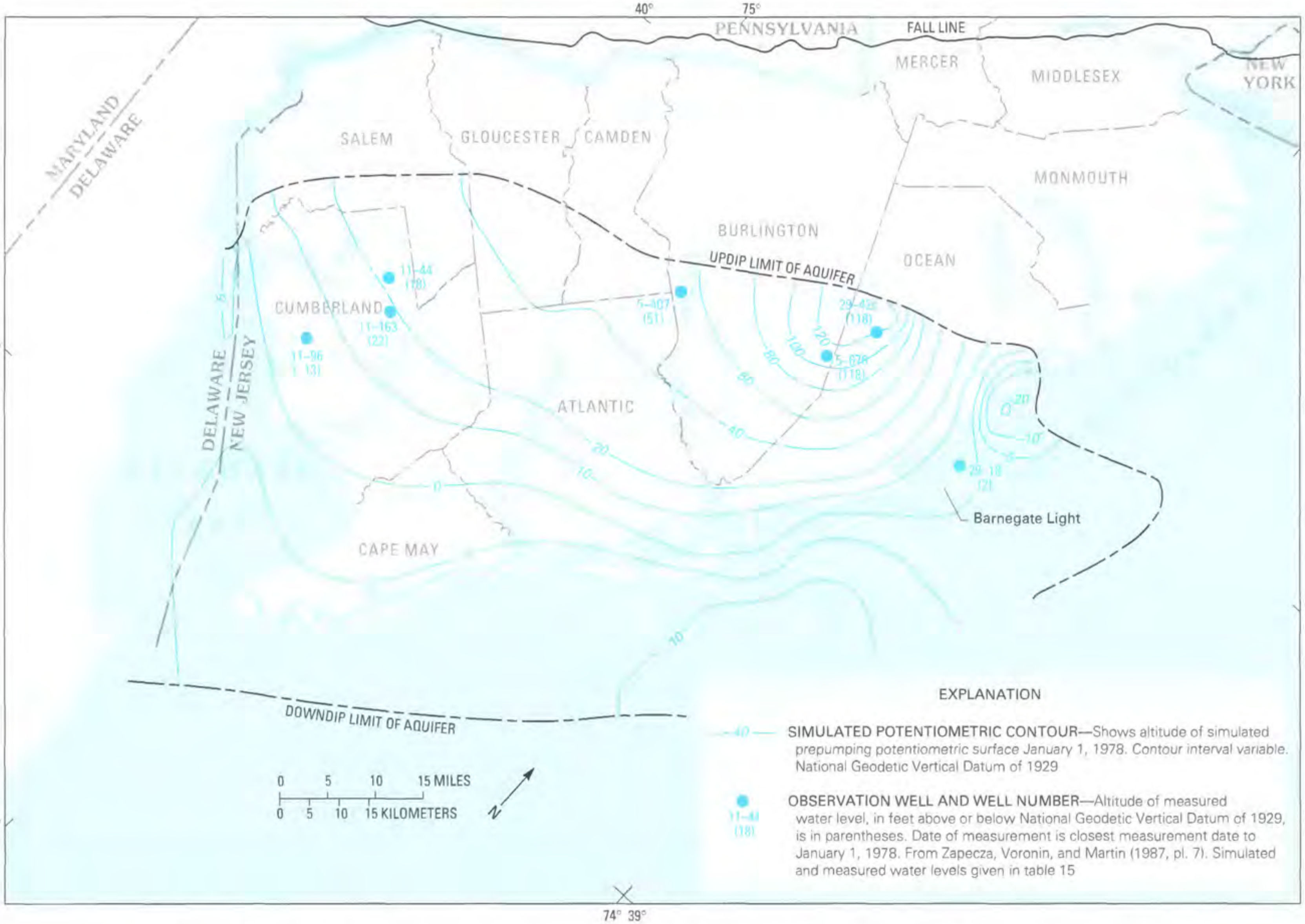


FIGURE 48.—Simulated 1978 potentiometric surface and measured 1978 water levels for the Piney Point aquifer (model unit A7).

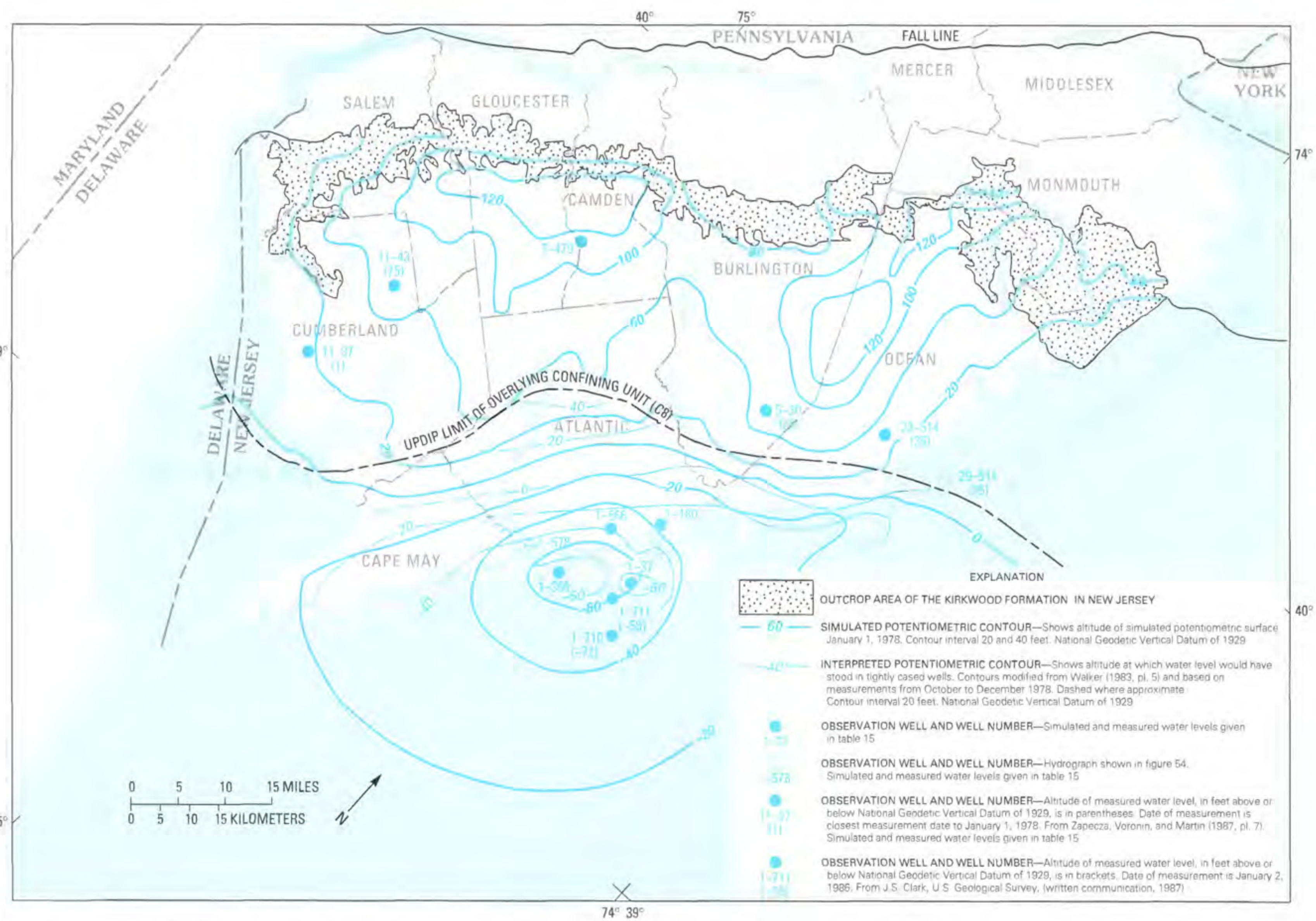


FIGURE 49.—Simulated and interpreted 1978 potentiometric surfaces and measured 1978 water levels for the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (model unit A8).

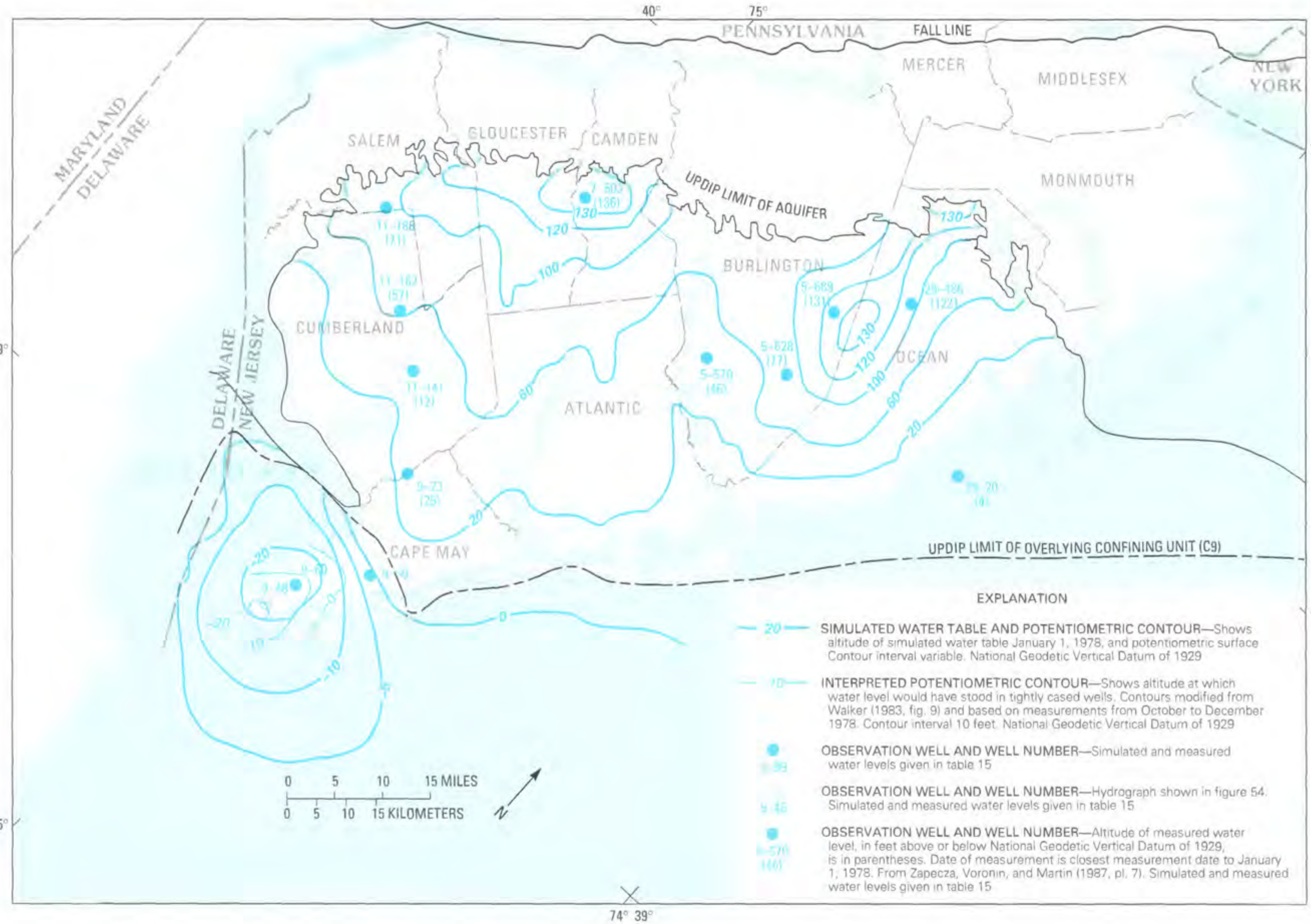
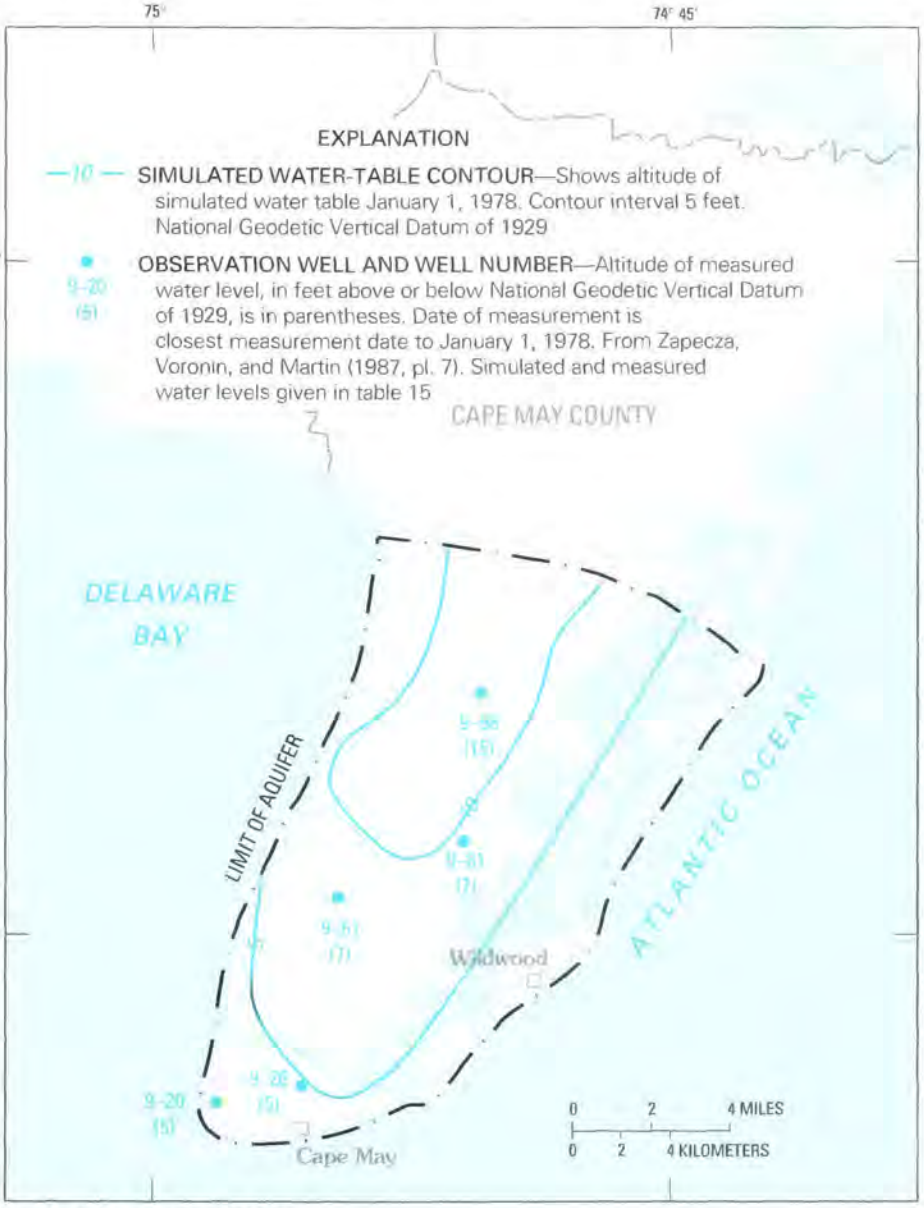


FIGURE 50.—Simulated 1978 water table, simulated and interpreted 1978 potentiometric surfaces, and measured 1978 water levels for the upper Kirkwood-Cohansey aquifer (model unit A9).



Base from U.S. Department of Agriculture, Soil Conservation Service, 1:250,000, 1980

FIGURE 51.—Simulated 1978 water table and measured 1978 water levels for the Holly Beach aquifer (model unit A10).

potentiometric surface for the underlying Wenonah-Mount Laurel aquifer (A5), and the potentiometric surface reflects land-surface altitudes near the Vincentown outcrop area.

The simulated 1978 potentiometric surface and measured 1978 water levels for the Piney Point aquifer (A7) are shown in figure 48. As with the Vincentown aquifer (A6), there are not enough observation wells to draw an interpreted potentiometric surface for the Piney Point aquifer. However, there are enough observation wells to determine that the simulated water levels are within reasonable ranges. Simulated water levels for six of the seven observation wells are within about 10 feet of the measured water levels (table 15). The flow system within the Piney Point aquifer consists of high ground-water levels along the county line between central Burlington and Ocean Counties, low ground-water altitudes beneath the Atlantic Ocean and Delaware Bay, and a small cone of depression along the coast in Ocean County. Ground-water levels in western Cumberland County are simu-

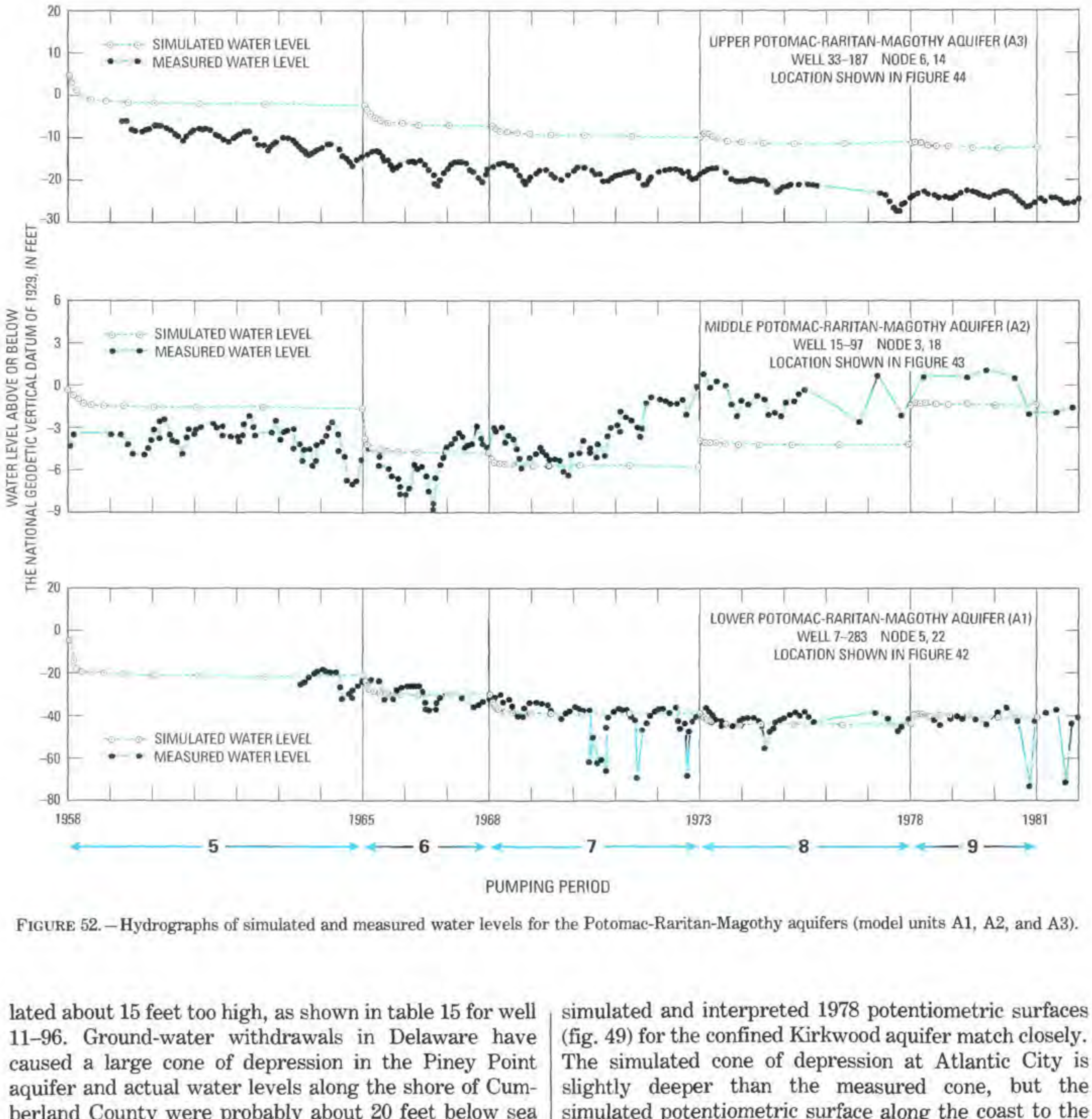


FIGURE 52.—Hydrographs of simulated and measured water levels for the Potomac-Raritan-Magothy aquifers (model units A1, A2, and A3).

lated about 15 feet too high, as shown in table 15 for well 11–96. Ground-water withdrawals in Delaware have caused a large cone of depression in the Piney Point aquifer and actual water levels along the shore of Cumberland County were probably about 20 feet below sea level in 1981. Measured 1983 water levels (Eckel and Walker, 1986, pl. 6) indicate that another small cone of depression near Barnegat Light in Ocean County, which was not indicated in the 1978 measured water levels, was not simulated.

Simulated and interpreted potentiometric surfaces and measured water levels for the lower Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer (A8) are shown in figure 49. Simulated and measured hydrographs for this model unit are shown in figure 54. The simulated and interpreted 1978 potentiometric surfaces (fig. 49) for the confined Kirkwood aquifer match closely. The simulated cone of depression at Atlantic City is slightly deeper than the measured cone, but the simulated potentiometric surface along the coast to the northeast is slightly higher than the interpreted potentiometric surface. Comparison of simulated and measured hydrographs for several wells near Atlantic City (wells 1–37, 1–180, and 1–366 in table 15) shows that, although simulated water levels near the end of the simulation match measured water levels fairly well, the trend of water-level decline is not simulated closely. The poorly simulated trends of water-level decline near Atlantic City are probably the result of poor estimates of pumpage prior to 1970.

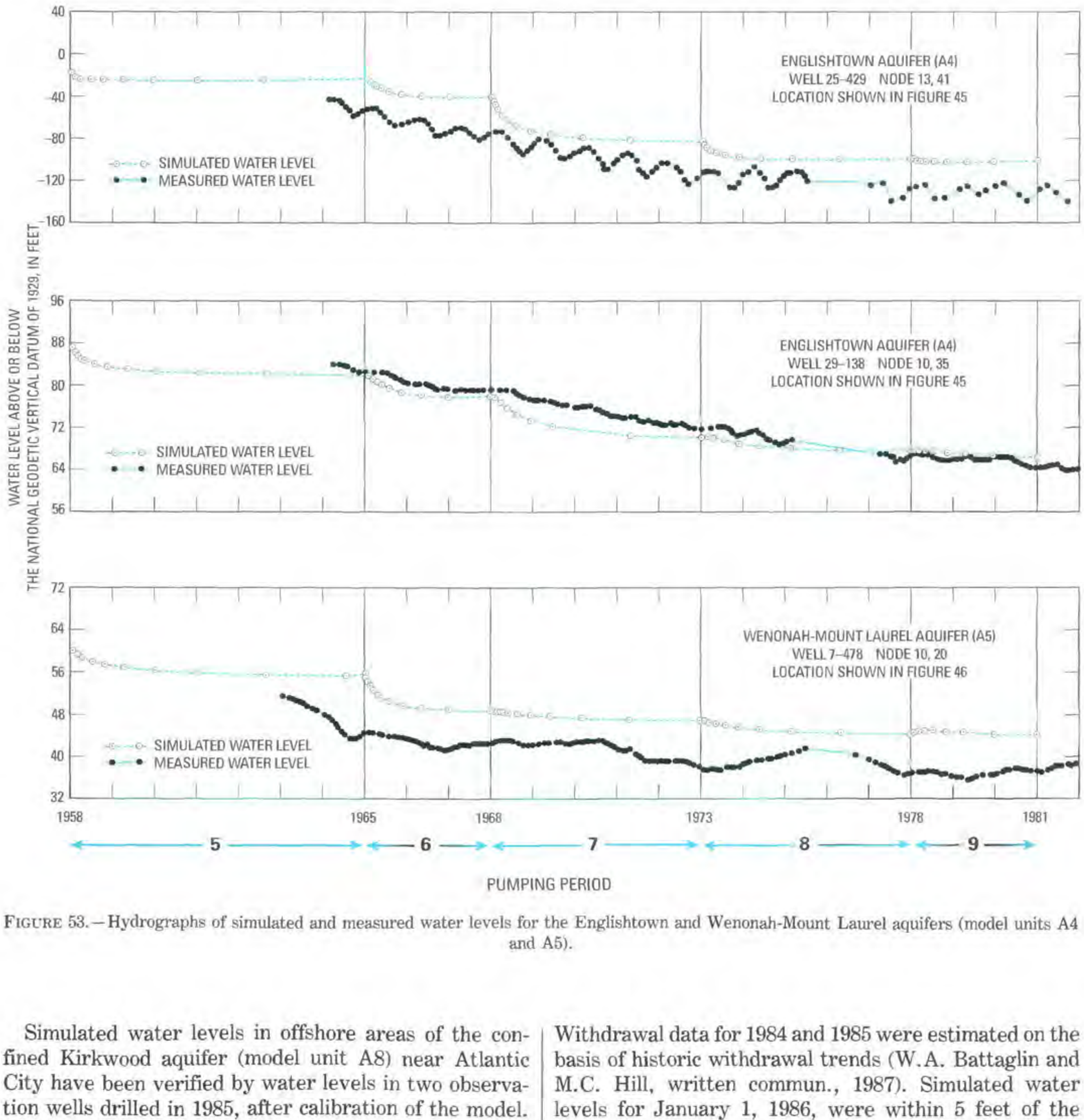


FIGURE 53.—Hydrographs of simulated and measured water levels for the Englishtown and Wenonah-Mount Laurel aquifers (model units A4 and A5).

Simulated water levels in offshore areas of the confined Kirkwood aquifer (model unit A8) near Atlantic City have been verified by water levels in two observation wells drilled in 1985, after calibration of the model. Water levels measured on January 2, 1986, were 72 feet below sea level in well 1–710 approximately 2 miles offshore and 59 feet below sea level in well 1–711 approximately 5 miles offshore, (J.S. Clark, written commun., 1987). Simulated water levels for January 1, 1986, were obtained by extending the period of simulation using the calibrated model. Withdrawals for 1981 through 1983 were estimated on the basis of data in the Ground-Water Withdrawal Inventory data base of the New Jersey District of the U.S. Geological Survey. Withdrawal data for 1984 and 1985 were estimated on the basis of historic withdrawal trends (W.A. Battaglin and M.C. Hill, written commun., 1987). Simulated water levels for January 1, 1986, were within 5 feet of the measured water level for each well. The similarity between the simulated and measured water levels demonstrates the ability of the regional model to aid in the definition and evaluation of the flow system.

Simulated and measured water levels in the lower part of the unconfined Kirkwood-Cohansey aquifer (A8) match closely. Measured 1978 water levels in the unconfined areas are essentially the same as the prepumping water levels for this area and are within a few feet of the water table in the unconfined upper Kirkwood-Cohansey

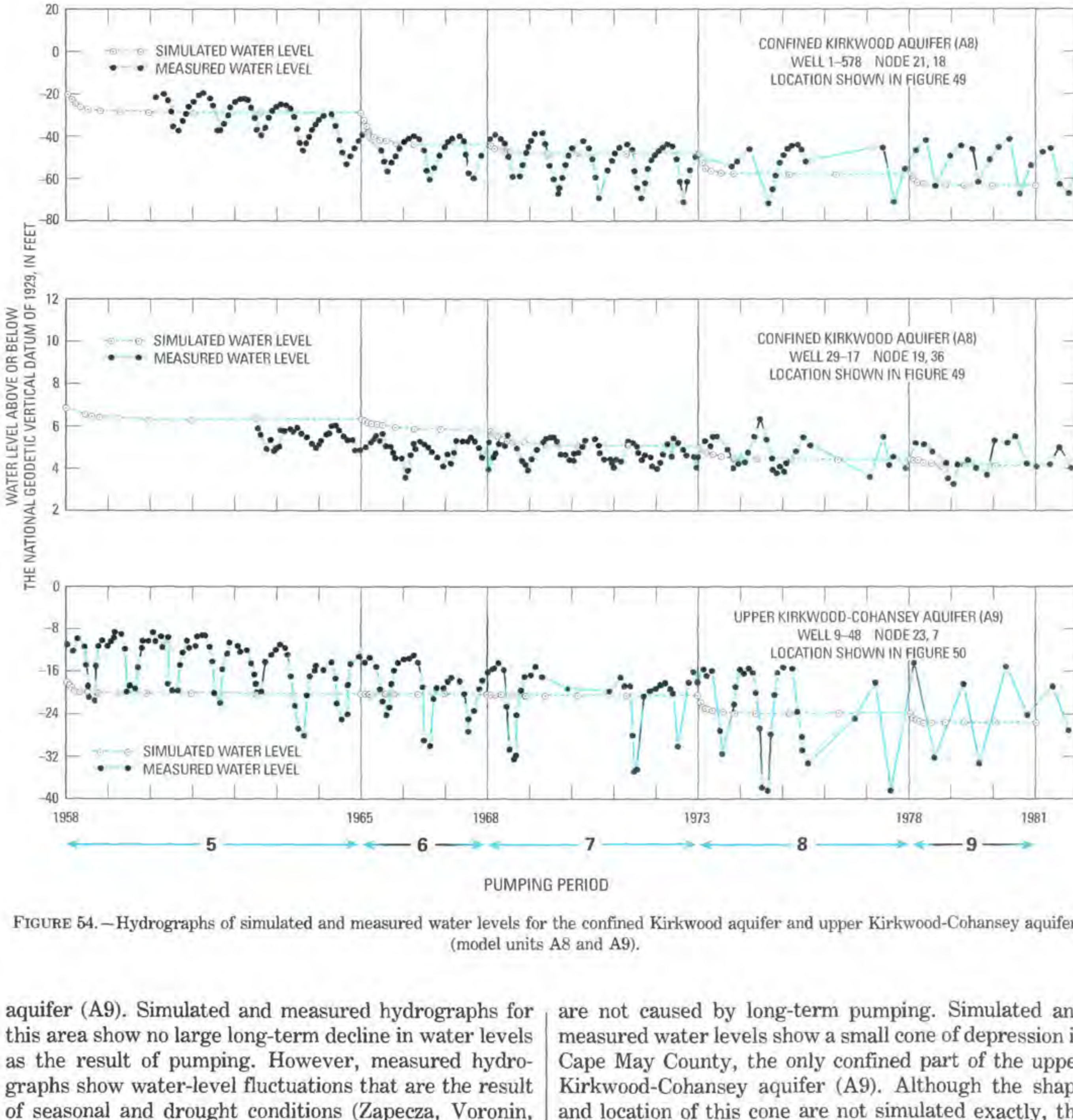


FIGURE 54.—Hydrographs of simulated and measured water levels for the confined Kirkwood aquifer and upper Kirkwood-Cohansey aquifer (model units A8 and A9).

aquifer (A9). Simulated and measured hydrographs for this area show no large long-term decline in water levels as the result of pumping. However, measured hydrographs show water-level fluctuations that are the result of seasonal and drought conditions (Zapecza, Voronin, and Martin, 1987, pl. 9 and p. 10). These short-term fluctuations are not simulated because recharge was not varied over the simulated time period.

Water levels in the upper Kirkwood-Cohansey aquifer (A9) are also simulated closely (fig. 50). Most of this aquifer system is unconfined, and simulated 1978 water levels are nearly the same as simulated prepumping water levels. Measured water levels (Zapecza, Voronin, and Martin, 1987, pl. 9) show short-term fluctuations that are not simulated (well 9–48, fig. 54) because they are not caused by long-term pumping. Simulated and measured water levels show a small cone of depression in Cape May County, the only confined part of the upper Kirkwood-Cohansey aquifer (A9). Although the shape and location of this cone are not simulated exactly, the general trend of the ground-water levels is simulated.

Generally, the 1978 water levels in the Holly Beach aquifer (A10) are assumed to be similar to prepumping water levels because the entire aquifer is unconfined and withdrawals are relatively small. Several measured water levels are shown with the simulated potentiometric surface for the Holly Beach aquifer in figure 51. Neither the measured (Zapecza, Voronin, and Martin, 1987, pl. 10) nor the simulated water levels (table 15) show any long-term decline from pumping.

Simulated flow between the water table and streams in 1978 is considered comparable to the recharge and discharge areas postulated in the conceptual model and to the simulated prepumping flow system. Generally, 1978 flow to streams is similar to flow during prepumping conditions (fig. 41), except in the outcrop area of the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3). In most areas, change in flow to streams decreased less than 5 in/yr. The average flow to streams decreased between 2.5 and 3.0 in/yr. However, along the Potomac-Raritan-Magothy outcrop area locally in Middlesex County and near the Delaware River in Gloucester, Camden, and Burlington Counties, flow to streams decreased 10 to 50 in/yr. These extreme changes in flow are in unconfined parts of the Potomac-Raritan-Magothy aquifers where withdrawals exceed 13 Mgal/d per node. Simulated reductions in streamflow that are greater than the sum of simulated prepumping discharge and the applied recharge rate of 20 in/yr are possible in the model because the constant-head nodes representing stream stages are able to supply an infinite source of water. In areas where simulated transient flow is out of these stream nodes, the model poorly represents recharge and the flow between streams and the water table. Simulated flow from three constant-head stream nodes (3, 23; 3, 24; and 4, 42) for January 1, 1978, is greater than 10 in/yr. This may indicate a critical reduction in streamflow or that the actual increase in recharge to the water table and the corresponding decrease in streamflow occur as smaller rates over larger areas than those simulated by the model.

EVALUATION OF HYDRAULIC CHARACTERISTICS AND FLOW SYSTEM BASED ON SIMULATION

Data values of aquifer and confining-unit hydraulic characteristics derived from the calibrated model are estimates of these properties on a regional scale. The water-transmitting character of the aquifers and confining units is shown by maps of aquifer transmissivity and confining-unit vertical leakance. These transmitting characteristics are lumped coefficients that are dependent on the hydraulic conductivity and thickness of the units. Aquifer hydraulic conductivity can be estimated by dividing transmissivity by saturated aquifer thickness. Vertical hydraulic conductivity of the confining units can be estimated by multiplying the vertical leakance by the unit's thickness.

Flow rates between aquifers have been computed by using simulated water levels and confining-unit vertical leakance values from the calibrated model. Flow rates between aquifers are shown by maps of flow to and from an aquifer through the overlying confining unit. Flow rates into an aquifer are positive; flow rates out of an aquifer are negative. Therefore, because flow rates are taken through the overlying confining unit, positive flow rates represent downward flow, and negative rates represent upward flow. These flow rates are regional estimates of recharge to and discharge from the aquifers.

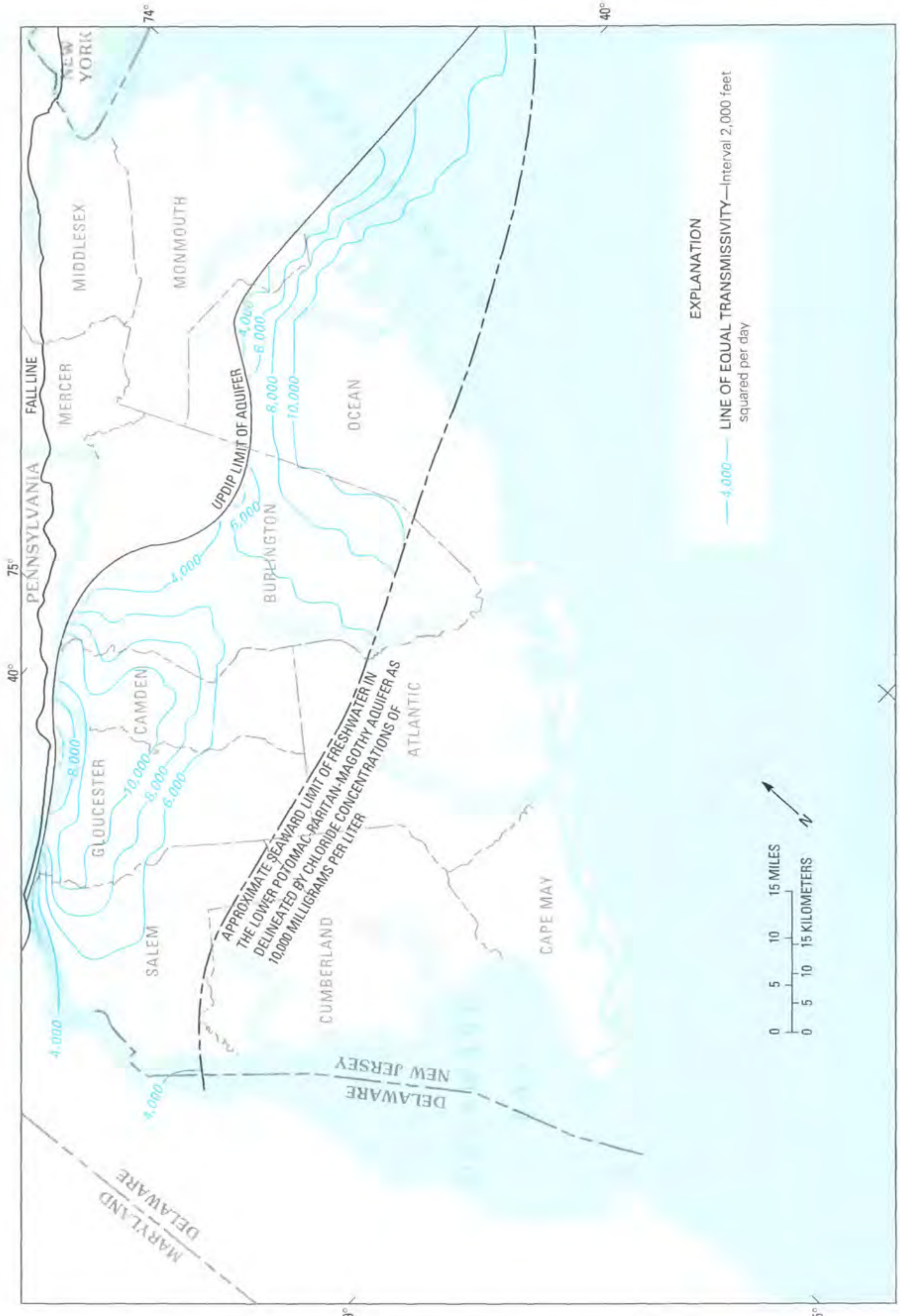
The model estimates of transmissivity, vertical leakance, and flow rates may not be representative for local areas that are smaller than the model cell areas. The accuracy of calibrated model data and data results is dependent on (1) the accuracy of the conceptual model; (2) the cell size; (3) the assumed boundary conditions; (4) the amount and distribution of data on water levels, withdrawals, and aquifer and confining-unit properties; and (5) the accuracy of the model calibration. Hydraulic characteristics used in the calibrated model are generally similar, but not identical, to estimates derived from previous simulation studies or field measurements.

TRANSMISSIVITY

Transmissivity used in the calibrated model for each aquifer is shown in figures 55 through 64. The range of estimated transmissivity from the calibrated model is given in table 8 for each aquifer. Comparison of the model-calibrated transmissivity (table 8) to the initial estimates of transmissivity, based on specific-capacity and aquifer-test data shown in table 5, shows that the maximum transmissivity for each unit in the model is about 20 to 40 percent lower in five aquifers, 55 percent lower in one aquifer, 6 percent higher in one aquifer, and unchanged in three aquifers.

The transmissivity shown in figures 55 through 64 is a regional estimate for the entire modeled extent of each aquifer. In outcrop areas, the transmissivity does not represent actual unconfined transmissivity. Actual transmissivity in the unconfined outcrop areas decreases with declining water levels. Most nodes simulating the aquifer outcrop areas also represent confined parts of the aquifers because of the large grid spacing. The unconfined parts of the upper Kirkwood-Cohansey aquifer (A9) and the Holly Beach aquifer (A10), like the outcrop areas, are simulated as confined aquifers beneath the confining streambeds and modeled transmissivity does not change with declining water levels. In these unconfined areas where withdrawals have caused water-level declines, hydraulic conductivity estimates, based on transmissivity and saturated thickness used in this model, are probably too low.

The transmissivity of the lower Potomac-Raritan-Magothy aquifer (A1) is shown in figure 55 and ranges from 860 to 10,400 ft^2/d (table 8). The low values are in areas where the aquifer thins beneath the Delaware River. The highest transmissivity is updip in northern Camden and Gloucester Counties and downdip in Ocean County. For most of the aquifer's extent, transmissivity



Location of 10,000 milligrams per liter chloride concentrations based of Meisler (1981, fig. 4).

FIGURE 55.—Transmissivity used in the model for the lower Potomac-Raritan-Magothy aquifer (model unit A1).

FIGURE 56.—Transmissivity used in the model for the middle Potomac-Raritan-Magothy aquifer (model unit A2).

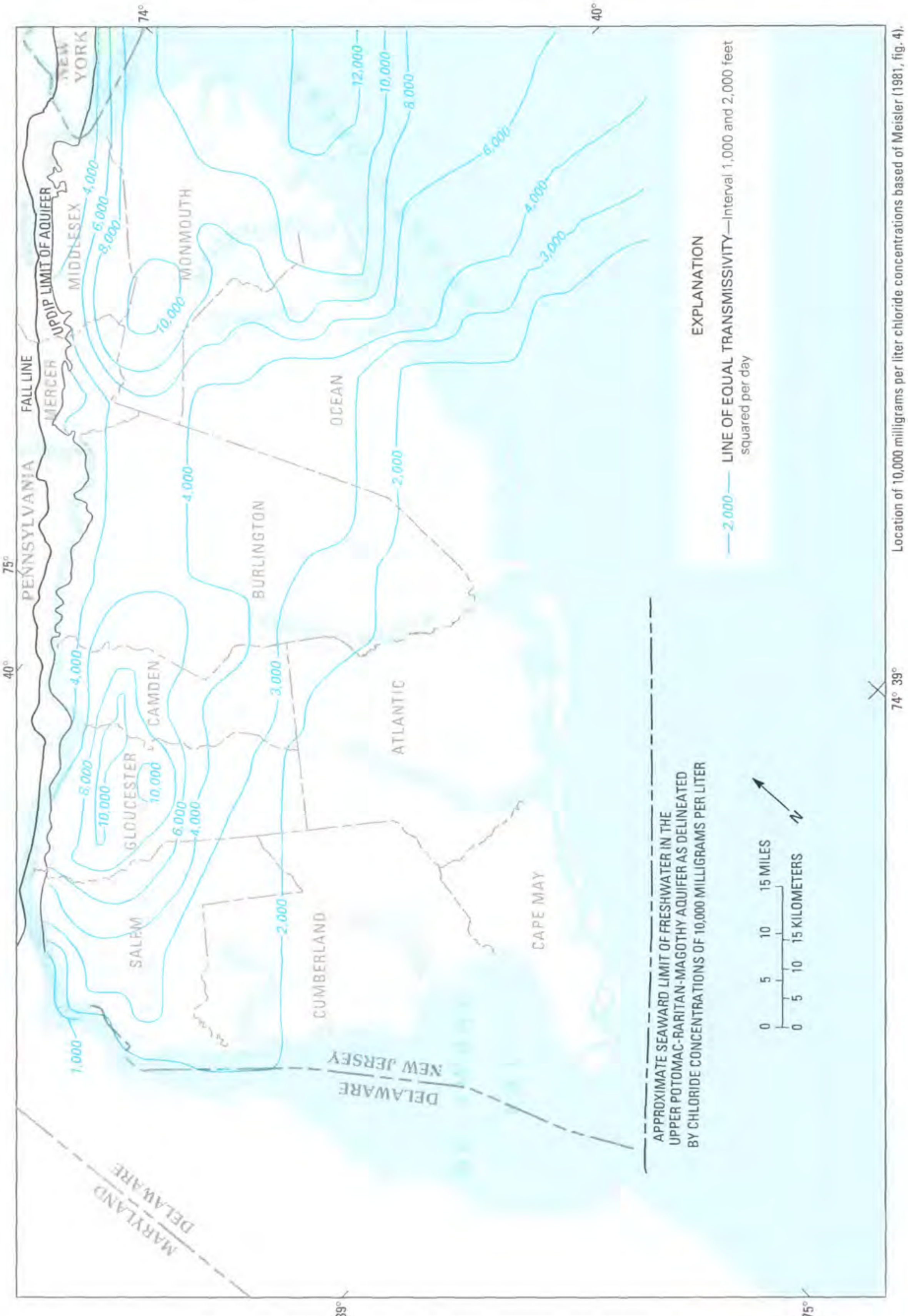


FIGURE 57.—Transmissivity used in the model for the upper Potomac-Raritan-Magothy aquifer (model unit A3).

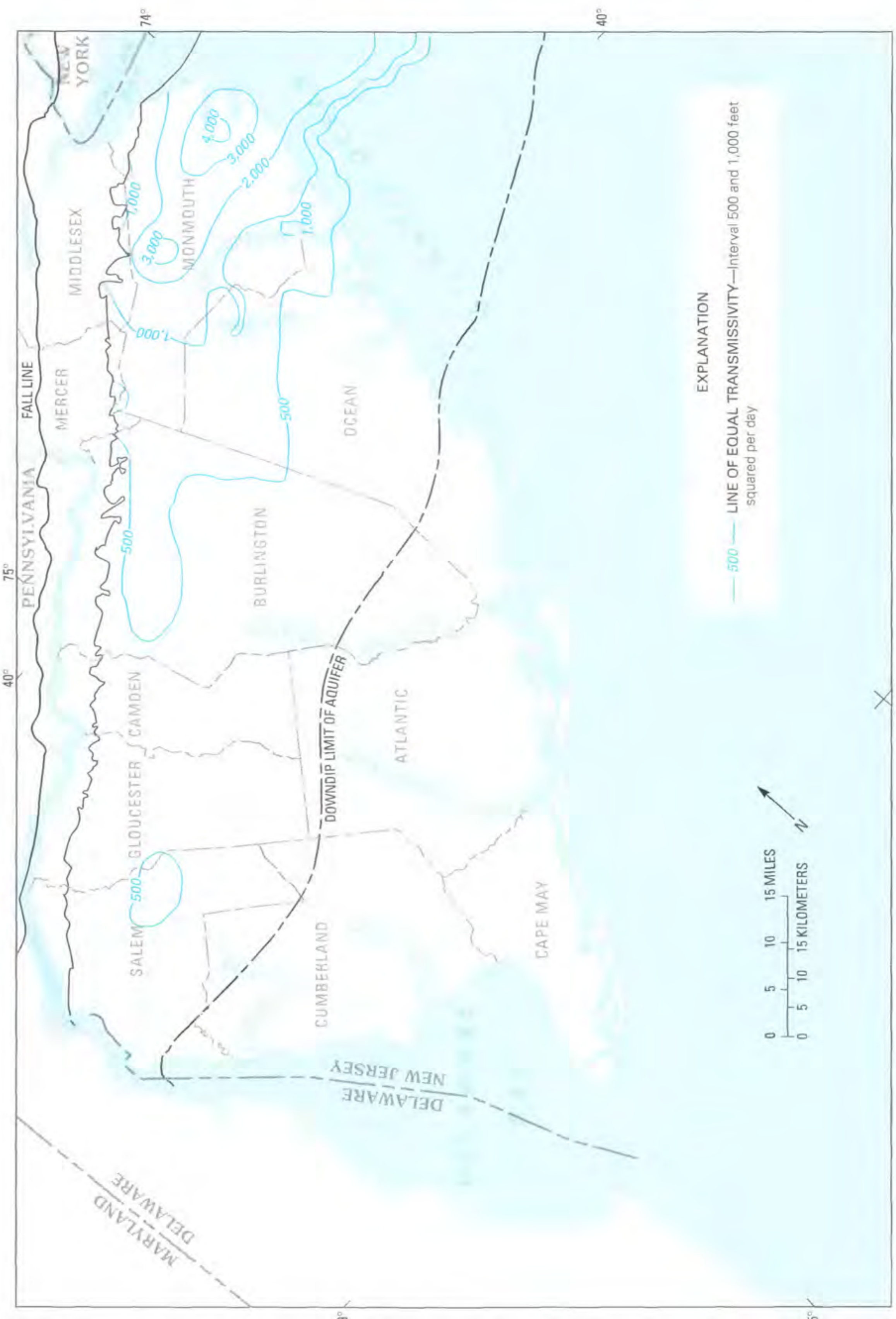


FIGURE 58.—Transmissivity used in the model for the Englishtown aquifer (model unit A4).

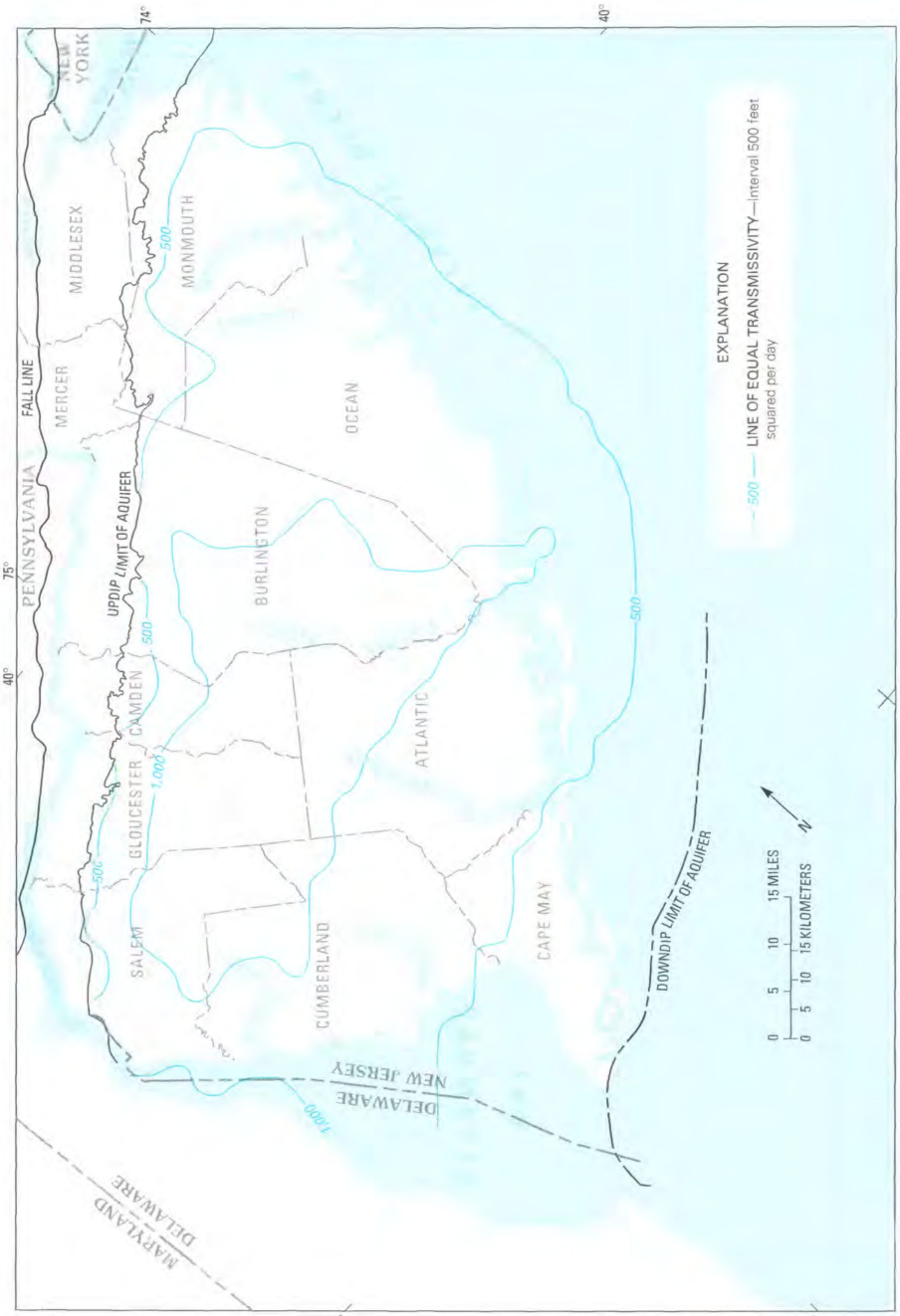


FIGURE 59.—Transmissivity used in the model for the Wenonah-Mount Laurel aquifer (model unit A5).

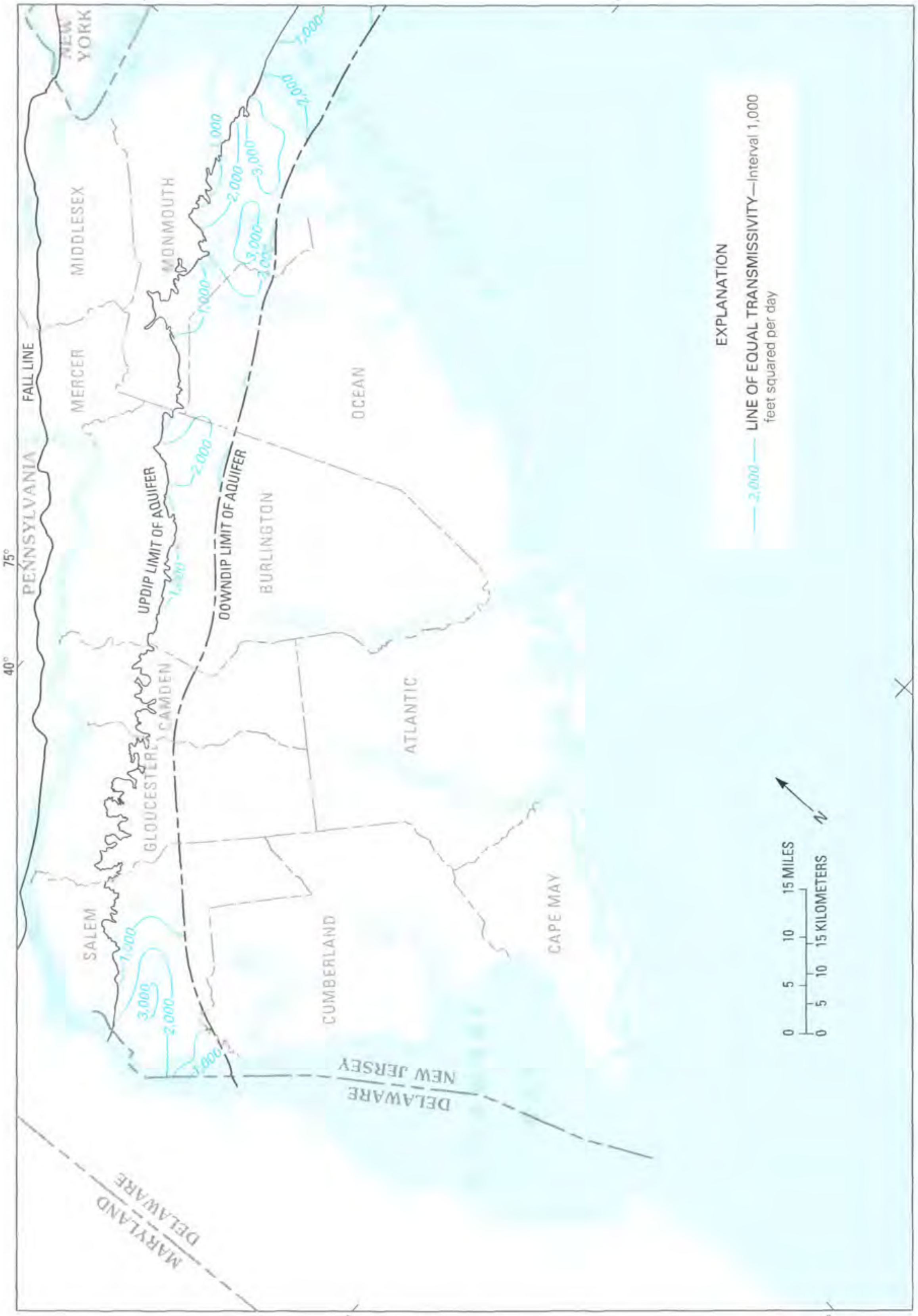


FIGURE 60.—Transmissivity used in the model for the Vincentown aquifer (model unit A6).

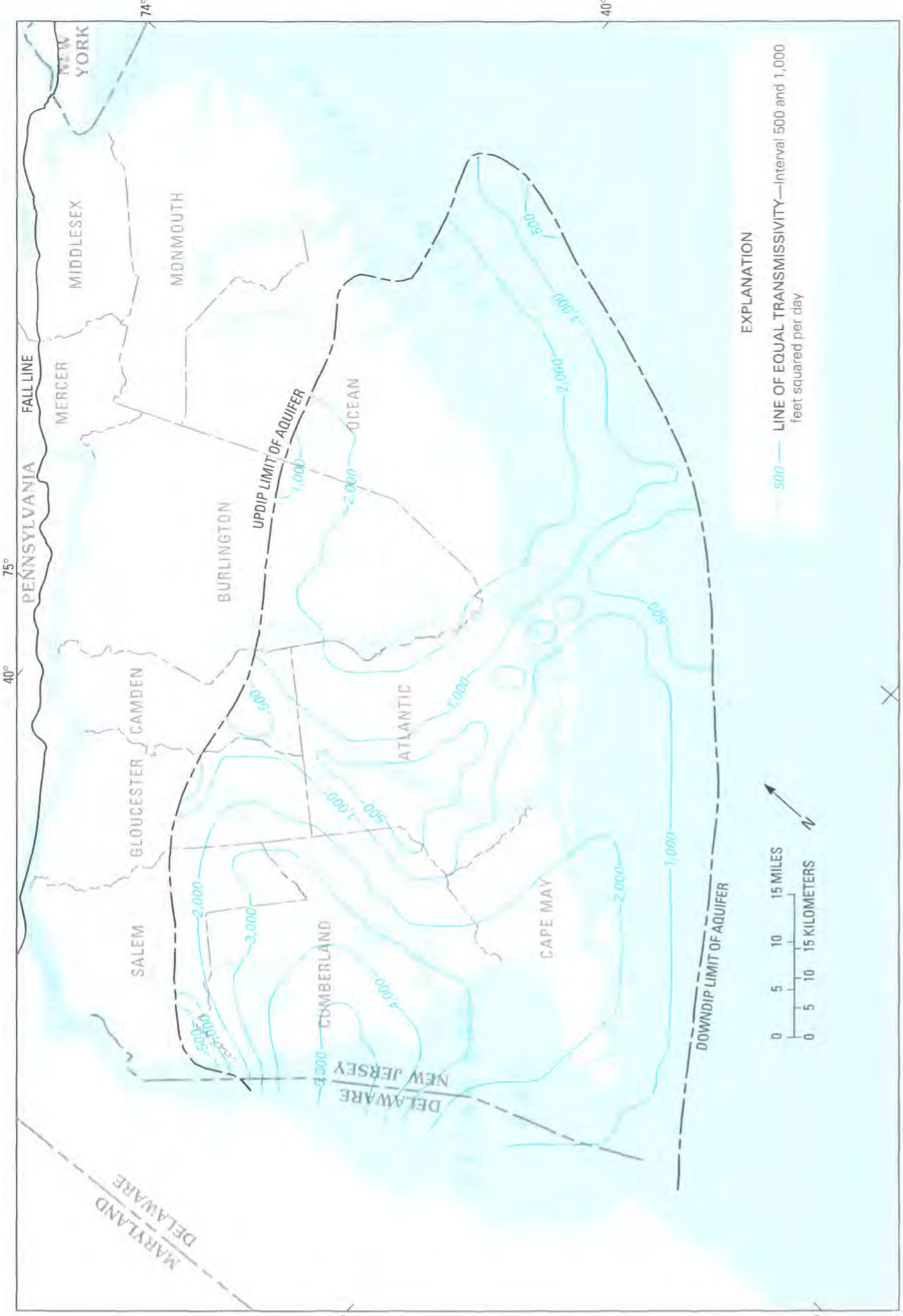


FIGURE 61.—Transmissivity used in the model for the Piney Point aquifer (model unit A7).

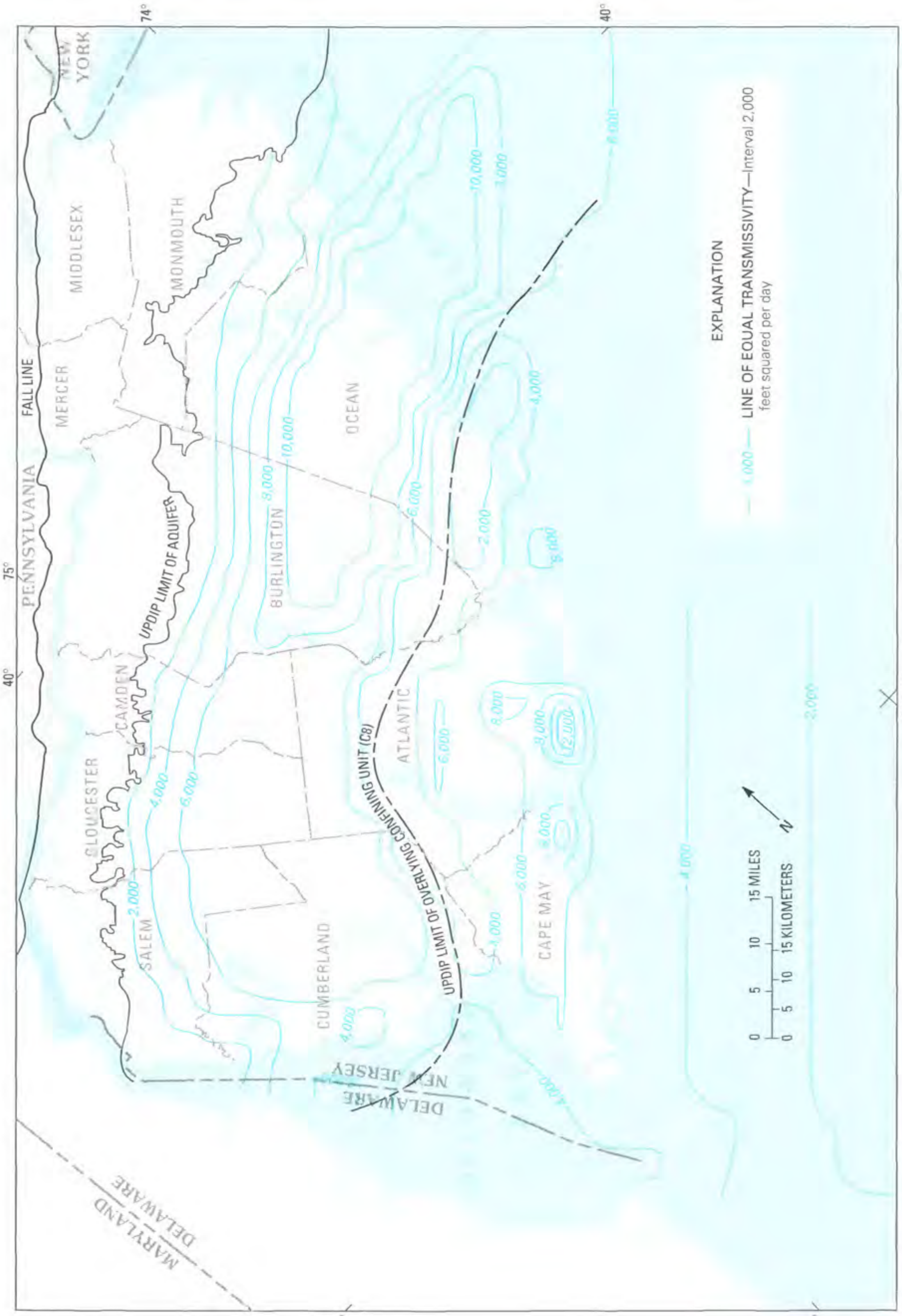


FIGURE 62.—Transmissivity used in the model for the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (model unit A8).

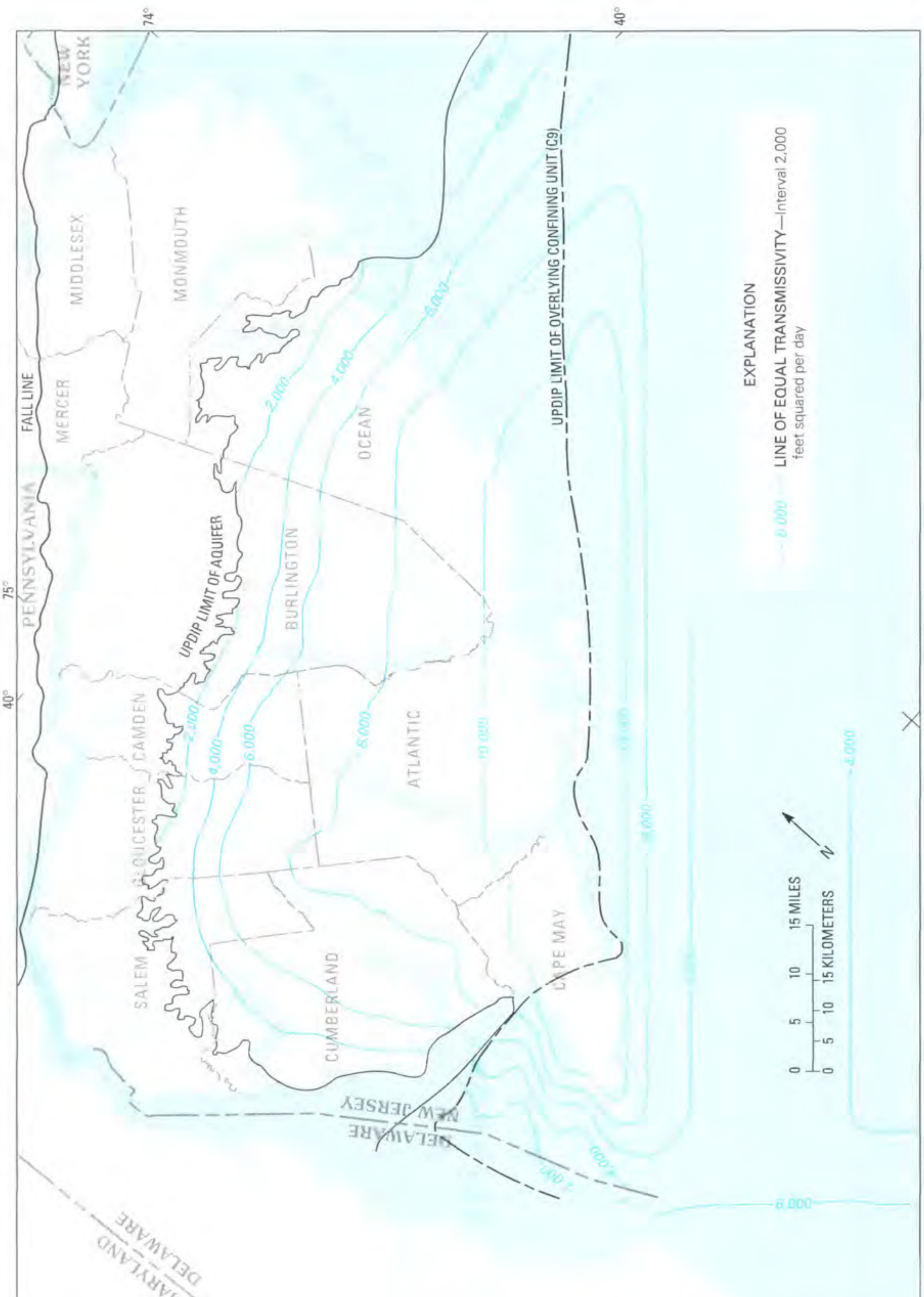


FIGURE 63.—Transmissivity used in the model for the upper Kirkwood-Cohansey aquifer (model unit A9).

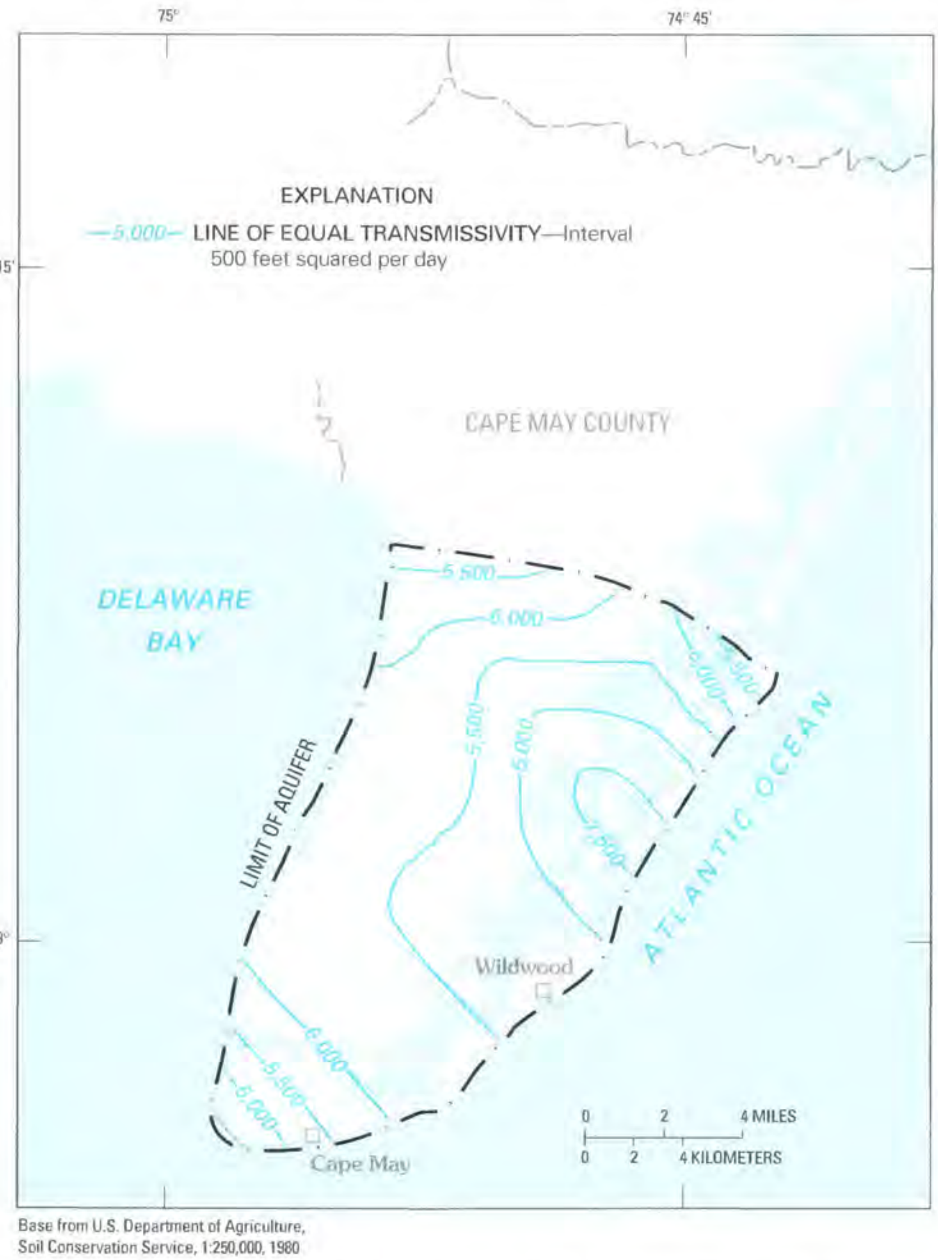


FIGURE 64.—Transmissivity used in the model for the Holly Beach Aquifer (model unit A10).

is between 5,200 ft^2/d and 10,400 ft^2/d. Lateral variability of transmissivity in the lower aquifer reflects changes in aquifer hydraulic conductivity more than changes in aquifer thickness.

Transmissivity of the middle Potomac-Raritan-Magothy aquifer (A2) is shown in figure 56 and ranges from 600 to 23,000 ft^2/d (table 8). The highest transmissivity is in northern Ocean and southern Monmouth Counties; the lowest transmissivity is near the outcrop where the aquifer thins and also in downdip areas. Transmissivity estimates are relatively high, 11,000 ft^2/d, in northern Camden County. Lateral changes in transmissivity in the middle aquifer reflect changes in hydraulic conductivity and do not reflect changes in aquifer thickness. These transmissivity values are generally 30 percent and locally 50 percent, less than those used by Farlekas (1979, p. 31) to simulate flow in the Farrington aquifer in Middlesex and Monmouth Counties. The differences in local transmissivity are probably the result of the smaller grid spacing used by Farlekas

TABLE 8.—*Range of transmissivity and confining-unit vertical leakance used in the calibrated model*

Model unit	Aquifer	Transmissivity (feet squared per day)
A1	Lower Potomac-Raritan-Magothy	860–10,400
A2	Middle Potomac-Raritan-Magothy	600–23,000
A3	Upper Potomac-Raritan-Magothy	300–12,100
A4	Englishtown	70– 4,400
A5	Wenonah-Mount Laurel	60– 1,400
A6	Vincentown	860– 3,500
A7	Piney Point	170– 5,200
A8	Lower Kirkwood-Cohansey	860–10,000
	and confined Kirkwood.	860–13,000
A9	Upper Kirkwood-Cohansey	270–11,700
A10	Holly Beach	5,200– 7,800
Model unit	**Confining unit**	**Vertical leakance (foot per day per foot)**
C1	Between the lower and middle Potomac-Raritan-Magothy aquifers.	2×10^{-6}–4×10^{-4}
C2	Between the middle and upper Potomac-Raritan-Magothy aquifers.	3×10^{-8}–5×10^{-3}
C3	Merchantville-Woodbury	2×10^{-8}–3×10^{-1}
C4	Marshalltown-Wenonah	1×10^{-7}–6×10^{-1}
C5	Navesink-Hornerstown	2×10^{-8}–5×10^{-1}
C6	Vincentown-Manasquan	5×10^{-9}–5×10^{-1}
C7	Basal Kirkwood	7×10^{-8}–4×10^{-2}
C8	Overlying the Rio Grande water-bearing zone.	3×10^{-8}–1×10^{-6}
C9	Estuarine clay	2×10^{-7}–5×10^{-1}

(1979), and regional differences are probably the result of differences in definition of the middle aquifer and different boundary conditions.

Transmissivity of the upper Potomac-Raritan-Magothy aquifer (A3) is shown in figure 57 and ranges from 300 to 12,100 ft^2/d (table 8). The lowest transmissivity is beneath the Delaware River in northern Salem County. Relatively low transmissivity, less than 3,000 ft^2/d, is in areas near the outcrop where the aquifer thins and in downdip areas in Cumberland, Atlantic, Cape May, southern Burlington, and southern Ocean Counties. The highest transmissivity is offshore of Monmouth County. Relatively high transmissivity, greater than 10,000 ft^2/d, is along the shore of Monmouth County, in northern Camden, northern Gloucester, and eastern Monmouth Counties. Unlike the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2), lateral changes in transmissivity of the upper aquifer are partly related to changes in aquifer thickness. The thickest part of the aquifer, near the shore of Monmouth County, also has the highest transmissivity.

Transmissivity of the Englishtown aquifer (A4) is shown in figure 58 and ranges from 70 to 4,400 ft^2/d (table 8). The highest transmissivity is in Monmouth County, where the aquifer is thickest. The lowest transmissivity is in updip and downdip areas where the aquifer thins near its limit. In general, lateral changes in transmissivity reflect changes in aquifer thickness. Transmissivity shown in figure 58 for Ocean and Burlington Counties is generally similar to that used by Nichols (1977a, p. 26) to simulate flow in the Englishtown aquifer. However, in northeastern Monmouth County, transmissivity is about two times higher than that used by Nichols (1977a, p. 26).

Transmissivity of the Wenonah-Mount Laurel aquifer (A5), shown in figure 59, ranges from 60 to 1,400 ft^2/d (table 8) and is generally lower than all the other Coastal Plain aquifers. There is little lateral variability, and 1,000 ft^2/d is a representative average for the aquifer. Calibrated transmissivity estimates fall approximately within the range of transmissivity estimated by Nemickas (1976, p. 39). Lateral changes in transmissivity generally result from changes in aquifer thickness, and the highest transmissivity is in Salem, central Camden, and central Burlington Counties, where the aquifer is thickest.

Transmissivity of the Vincentown aquifer (A6) is shown in figure 60 and ranges from 860 to 3,500 ft^2/d (table 8). Lateral changes in transmissivity result from changes in aquifer thickness. Highest transmissivity values are very localized in Salem and Monmouth Counties. Lowest transmissivity is in Gloucester and northern Salem Counties. Final transmissivity estimates used in the model are the same as the initial estimates based on an average hydraulic conductivity (table 5).

Transmissivity of the Piney Point aquifer (A7) is shown in figure 61 and ranges from 170 to 5,200 ft^2/d (table 8). The areas of transmissivity greater than 2,000 ft^2/d are mostly in Burlington, Ocean, Cumberland, and Cape May Counties where the aquifer is thickest. Transmissivity is relatively low, less than 1,000 ft^2/d, in Atlantic, southern Camden, and southern Gloucester Counties.

Transmissivity of the confined Kirkwood aquifer (the downdip part of A8) is shown in figure 62 and ranges from about 860 to 13,000 ft^2/d (table 8). Generally, transmissivity is about 5,000 to 6,000 ft^2/d with local areas of higher transmissivity. Areas of highest transmissivity, greater than 10,000 ft^2/d, are along the coast southwest of Atlantic City. Low transmissivities, less than 4,000 ft^2/d, are along the updip limit of the confined Kirkwood aquifer. Although there are no data to suggest either low hydraulic conductivity or that the unit is relatively thin in this area, low transmissivity was needed to simulate the cone of depression at Atlantic City. Water levels along the coast were simulated by decreasing the amount of flow from the updip unconfined part of model unit A8 by decreasing transmissivity. Decreasing flow through the overlying confining unit C8 by decreasing vertical leakance did not improve the calibration. The lower parts of the unconfined Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer are assumed to be part of the same model unit (A8) and have a direct lateral hydraulic connection. However, more geohydrologic and geophysical data are needed to refine the aquifer characteristics in this area.

Transmissivity of the lower Kirkwood-Cohansey aquifer, the updip part of model unit A8, is shown in figure 62 and ranges from 860 to 10,000 ft^2/d (table 8). Transmissivity is lowest (less than 4,000 ft^2/d) where the aquifer thins near its outcrop area. Downdip, transmissivity is generally greater than 7,000 ft^2/d. The highest transmissivity (greater than 10,000 ft^2/d) is in central Burlington and central Ocean Counties.

Transmissivity of the upper Kirkwood-Cohansey aquifer (A9) is shown in figure 63 and ranges from 270 to 11,700 ft^2/d (table 8). Transmissivity is lowest (less than 6,000 ft^2/d) in updip areas where the aquifer thins. Highest transmissivity (greater than 10,000 ft^2/d) is along the coast in Cape May, Atlantic, and southern Ocean Counties. The confined part of the upper Kirkwood-Cohansey aquifer in southern Cape May County has transmissivity ranging from 8,000 to 11,700 ft^2/d. Transmissivity for the unconfined Kirkwood-Cohansey aquifer system, represented by the combined lower and upper Kirkwood-Cohansey aquifers (A8 and A9) in the model, is highest (about 14,000 to 18,000 ft^2/d) where the aquifer system is thickest in central Cumberland, northern Atlantic, central Burlington, and central Ocean Counties.

Transmissivity of the Holly Beach aquifer (A10) is shown in figure 64 and ranges from 5,200 ft^2/d to 7,800 ft^2/d (table 8). Final estimates of transmissivity are the same as the initial estimates based on hydraulic conductivities determined from specific-capacity tests (table 5).

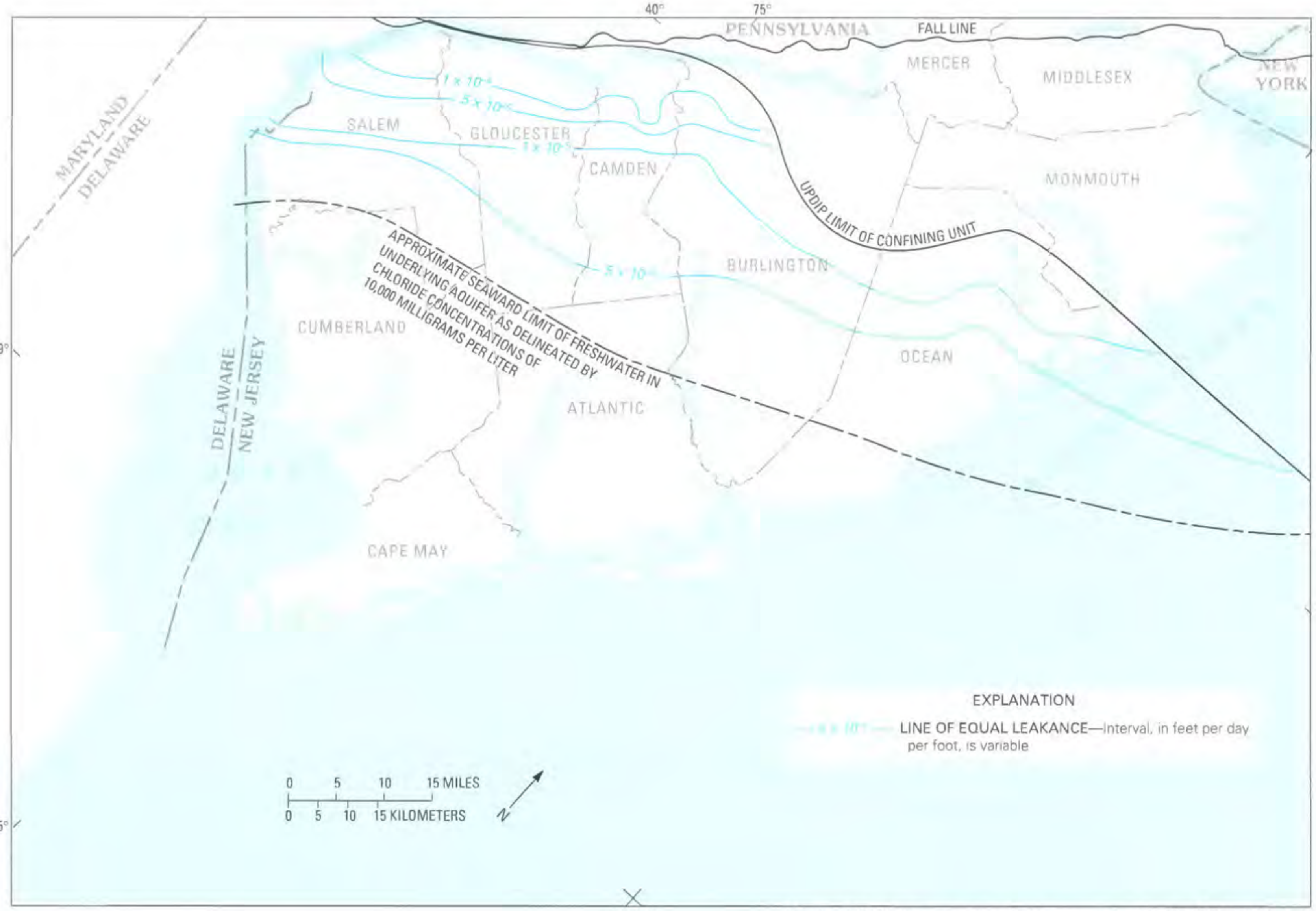
VERTICAL LEAKANCE OF CONFINING UNITS

Vertical leakance used in the calibrated model for each of the confining units is shown in figures 65 to 73 and summarized in table 8. Leakance, when multiplied by confining-unit thicknesses, gives the vertical hydraulic conductivity of the unit. The vertical hydraulic conductivity of the confining units is of the same order of magnitude reported by Luzier (1980, p. 29), Nemickas (1976, p. 37), Nichols (1977a, p. 76), and Farlekas (1979, p. 36); however, the lateral distribution is somewhat different from that reported by those investigators.

Lateral variability of vertical leakance reflects variability in vertical hydraulic conductivity and confining-unit thickness. Highest leakance is in updip areas, at and near the outcrops, where confining units are thinnest. In other areas, however, lateral changes in vertical hydraulic conductivity, which varies several orders of magnitude within a confining unit, tend to minimize the effects on leakance by changes in thickness, which varies about one order of magnitude.

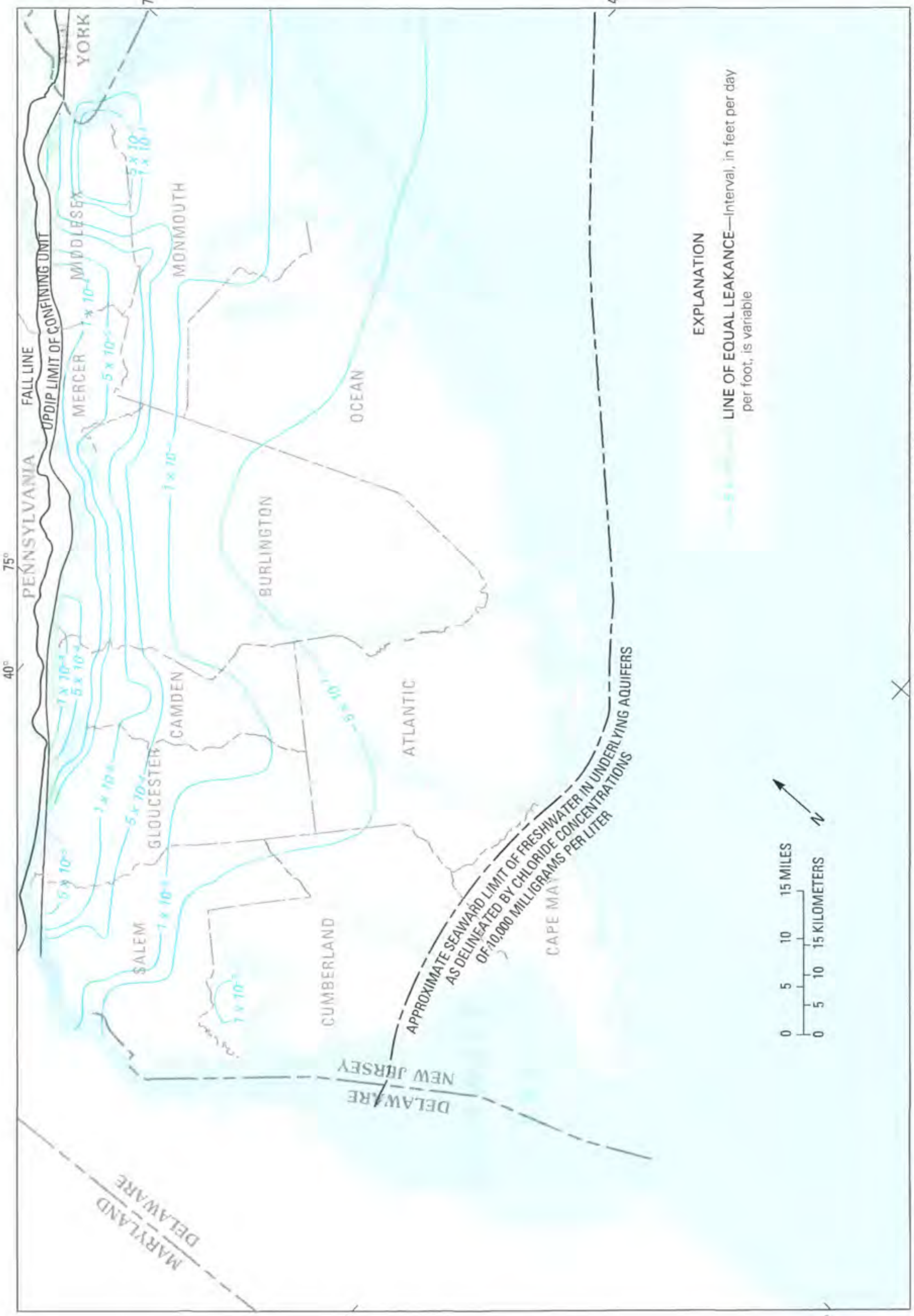
Generally, leakance varies laterally in the updip areas of the confining units. Lateral changes in leakance in updip areas are local, whereas lateral changes in downdip areas are over large areas. Although the leakance values shown in figures 65 to 73 represent the regional transmitting properties of the confining units, the simulated lateral variability of leakance within the units is affected by the distribution of measured water levels used in calibration and by model sensitivity to leakance. Measured water levels are more abundant in shallow updip areas, and calibration on a more local scale is possible. In downdip areas, measured water levels are fewer and only a more regional calibration is possible. Also, simulated water levels in downdip areas are much less sensitive to local changes in leakance. Therefore, leakance distributions with high variability in downdip areas were not considered during the calibration process. Although the regional vertical leakance of the confining units has been adequately simulated, the leakance values may not be locally representative in downdip areas.

The Merchantville-Woodbury confining unit (C3) and the Vincentown-Manasquan confining unit (C6) have relatively low leakance values, less than 1×10^{-8} feet per day per foot ((ft/d)/ft), nearshore and offshore of southern Monmouth and northern Ocean Counties (figs. 67 and 70). Downdip areas of most of the confining units have leakances of about 1×10^{-8} to 1×10^{-6} (ft/d)/ft. The Merchantville-Woodbury, Marshalltown-Wenonah, Navesink-Hornerstown, and Vincentown-Manasquan confining units (C3, C4, C5, and C6) have leakances in downdip areas to the northeast that are at least an order of magnitude lower than leakances in downdip areas to the southwest (figs. 67–70). However, the basal Kirkwood confining unit (C7) has leakances in downdip areas to the northeast that are at least an order of magnitude higher than leakances in downdip areas to the southwest (fig. 71). The confining unit between the lower and middle Potomac-Raritan-Magothy aquifers (C1) and the Marshalltown-Wenonah confining unit (C4) have relatively high leakances in some downdip areas compared to other confining units. The confining unit between the lower and middle Potomac-Raritan-Magothy aquifers (C1) has leakance between 1×10^{-6} and 1×10^{-4} (ft/d)/ft throughout most of its downdip extent (fig. 65). The high leakance is consistent with the hydrogeologic framework described by Zapecza (1984, p. 12). He states that the lower and middle aquifers of the Potomac-Raritan-Magothy aquifer system are largely interbedded sands, silts, and clays in downdip areas and cannot be distinguished from each other. High leakance



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 65.—Vertical leakance used in the model for the confining unit between the lower and middle Potomac-Raritan-Magothy aquifers (model unit C1).



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 66.—Vertical leakance used in the model for the confining unit between the middle and upper Potomac-Raritan-Magothy aquifers (model unit C2).

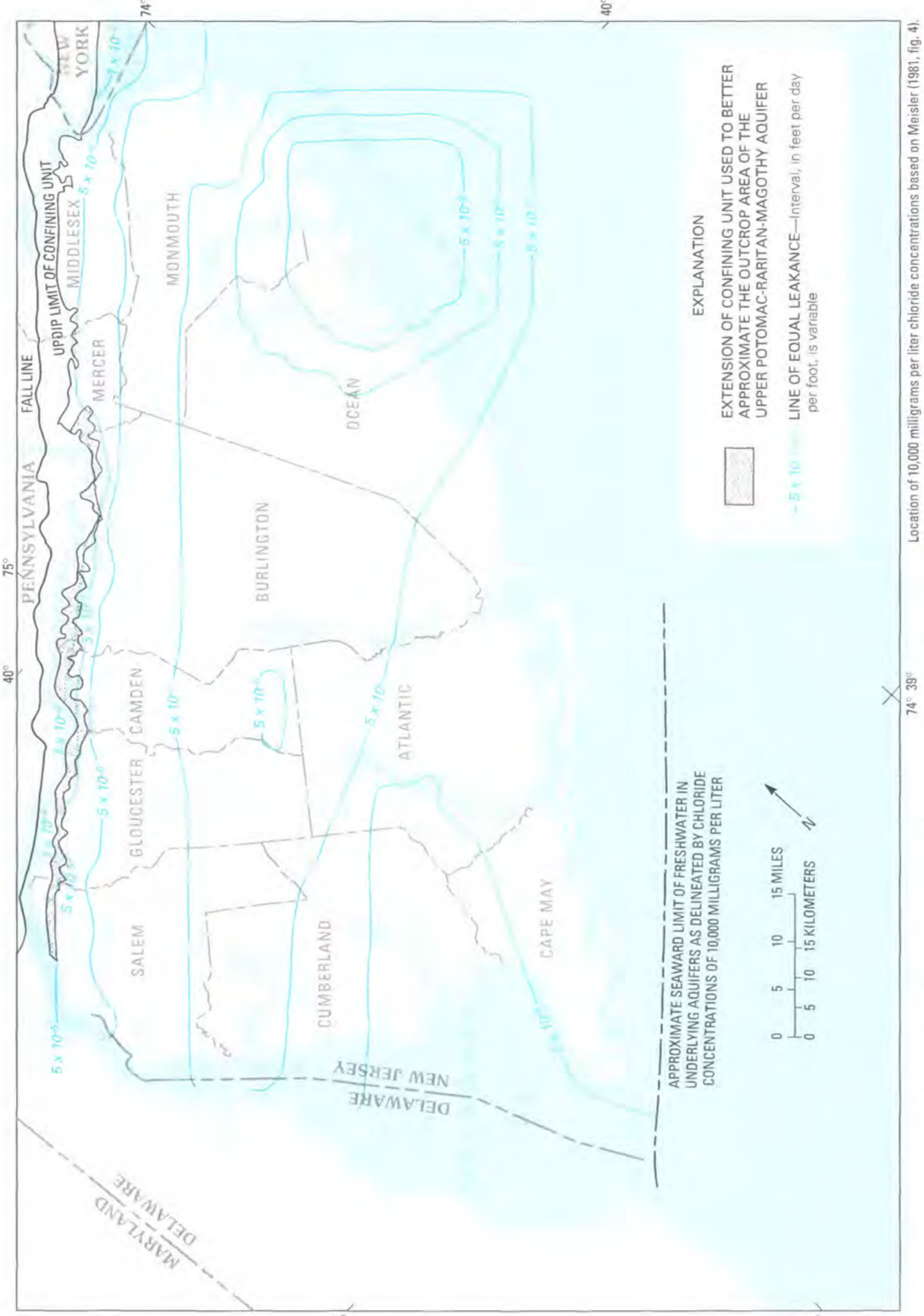


FIGURE 67.—Vertical leakance used in the model for the Merchantville-Woodbury confining unit (model unit C3).

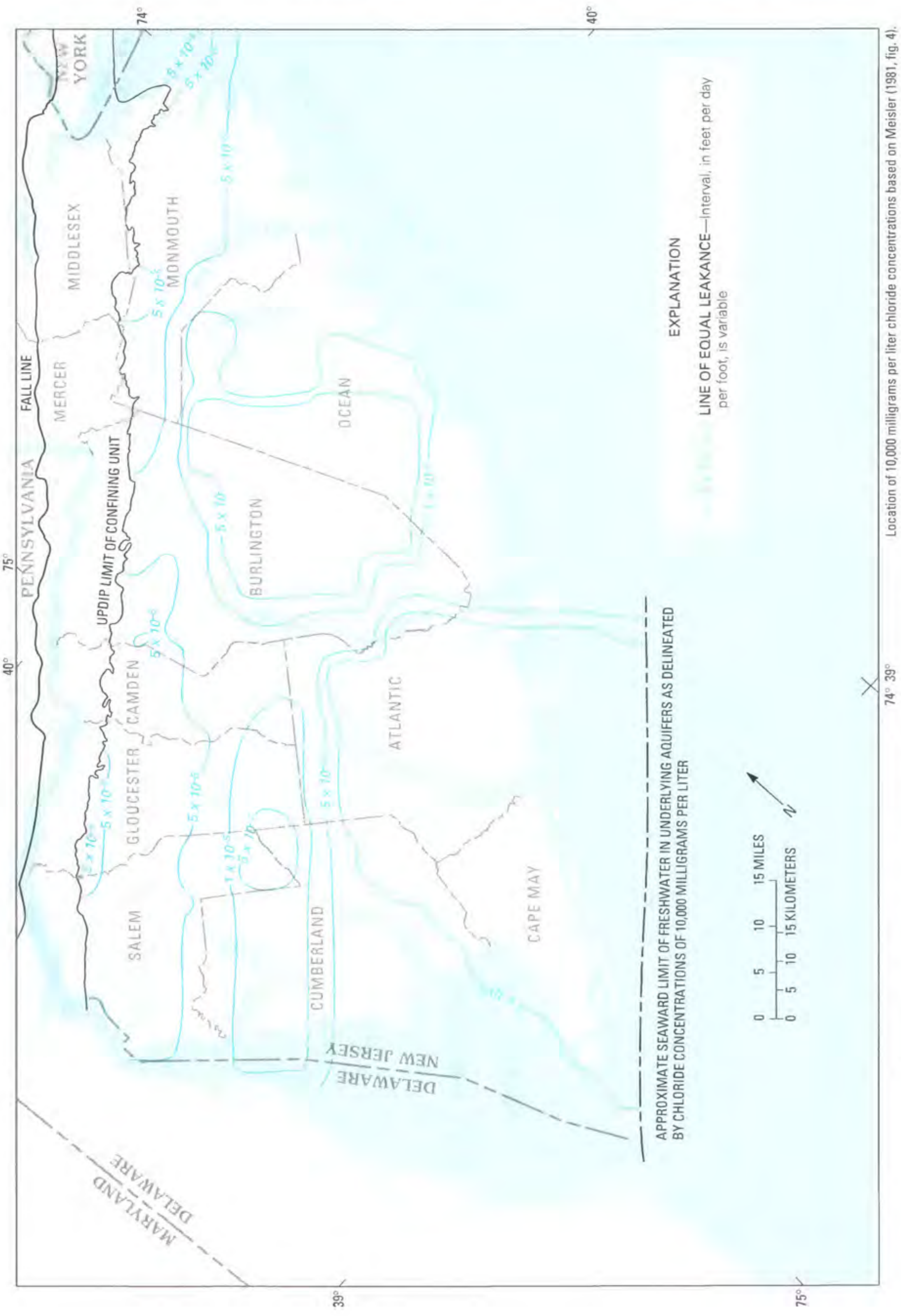
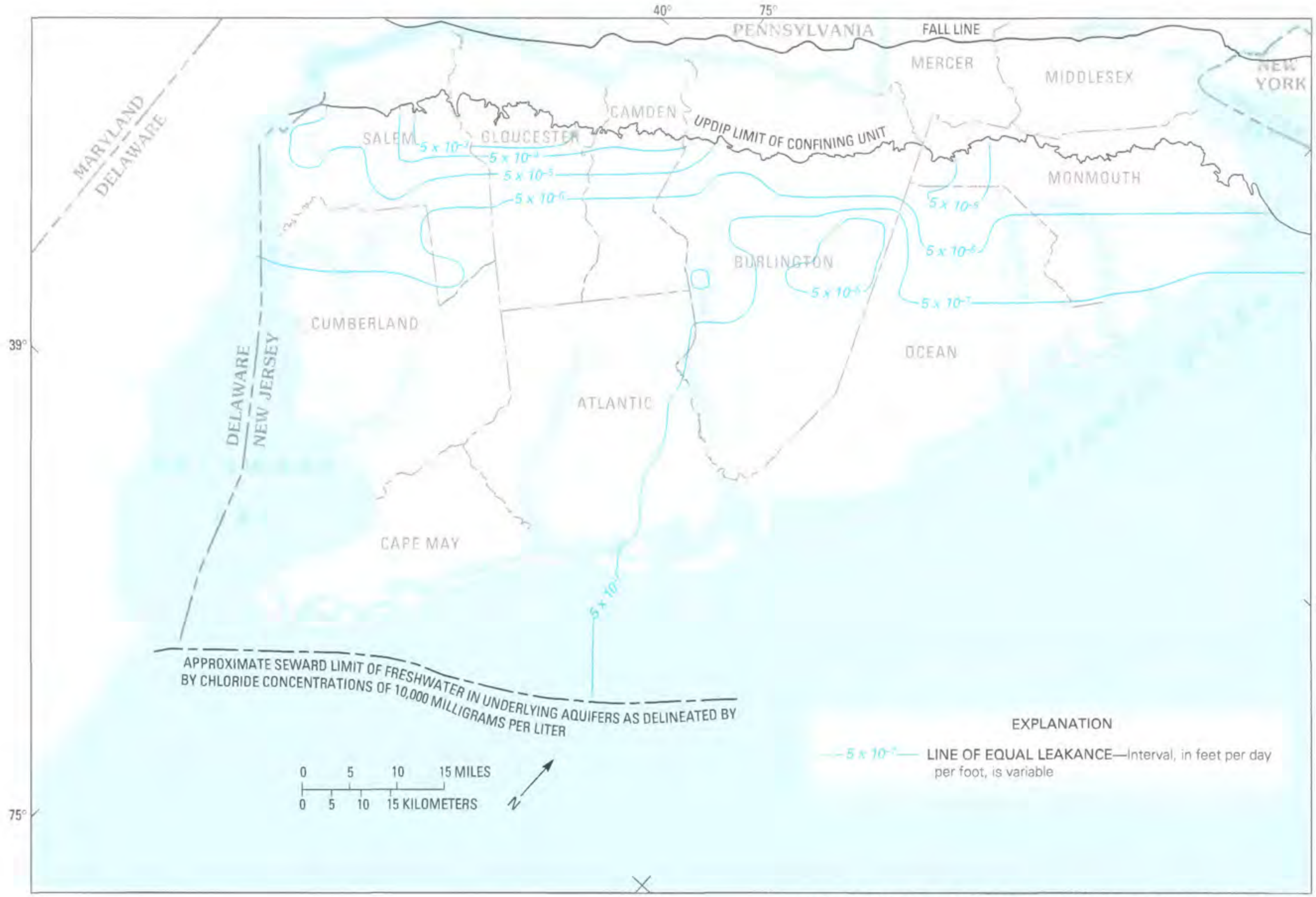
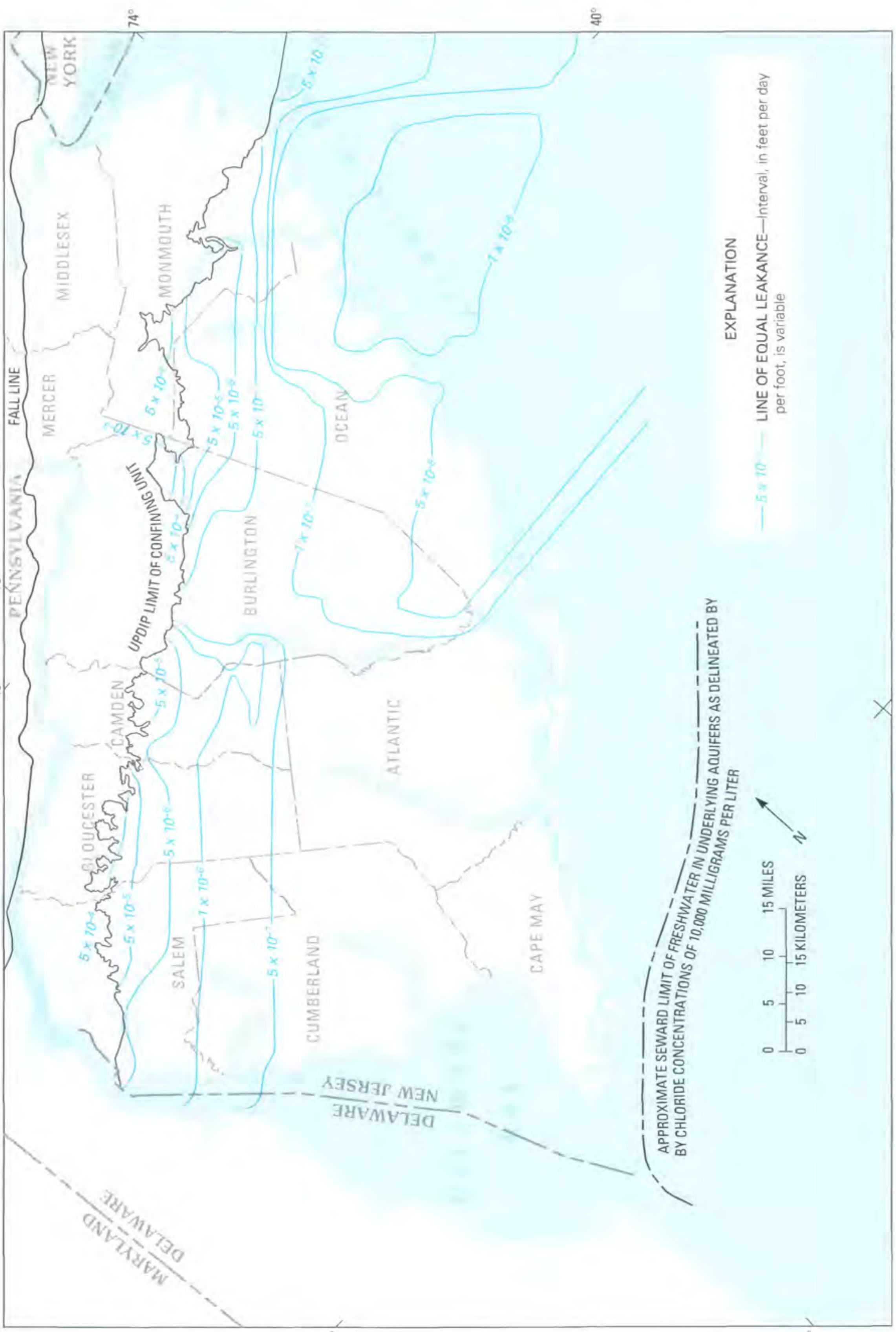


FIGURE 68.—Vertical leakance used in the model for the Marshalltown-Wenonah confining unit (model unit C4).



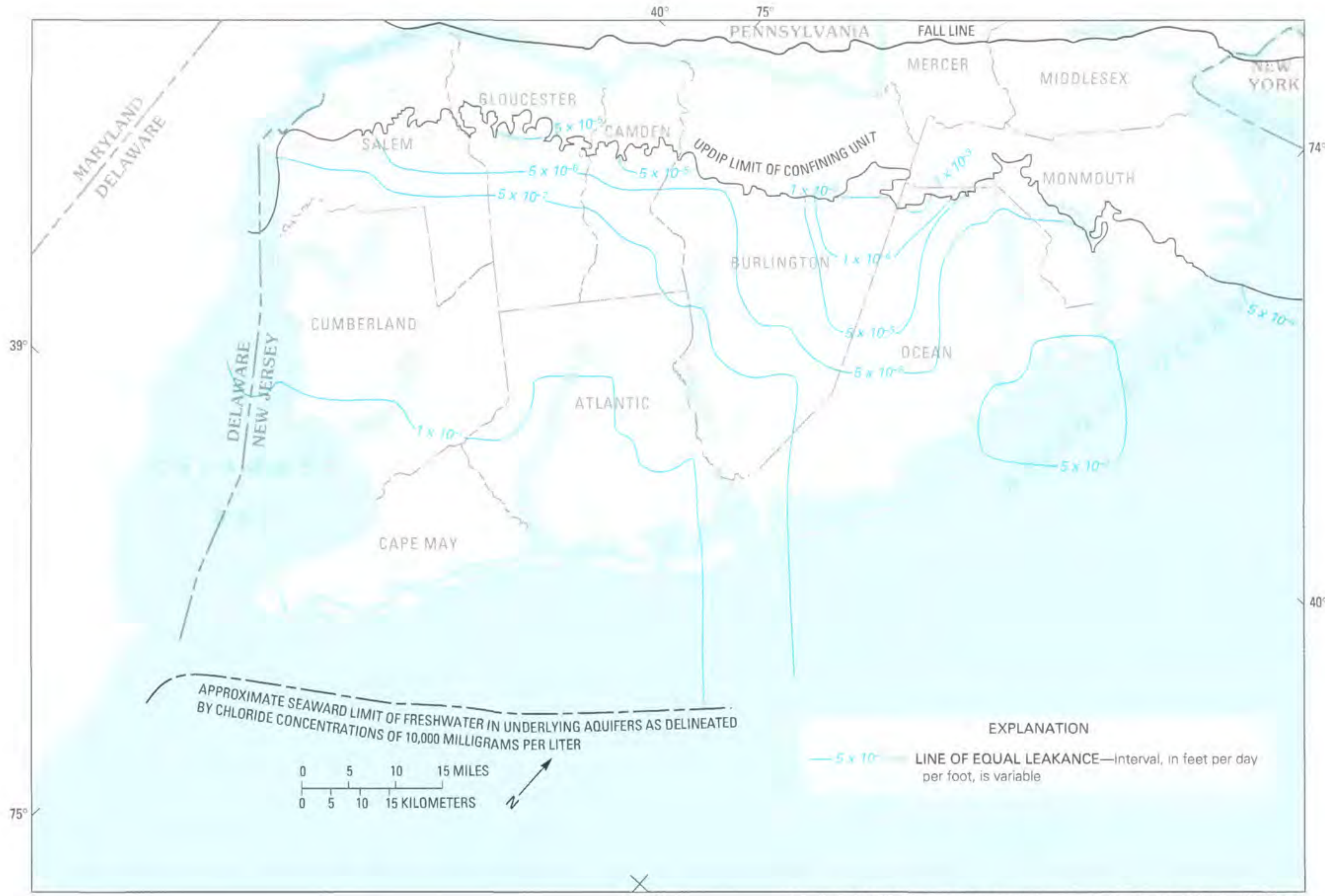
Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 69.—Vertical leakance used in the model for the Navesink-Hornerstown confining unit (model unit C5).



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 70.—Vertical leakance used in the model for the Vincentown-Manasquan confining unit (model unit C6).



Location of 10,000 milligrams per liter chloride concentrations based on Meisler (1981, fig. 4).

FIGURE 71.—Vertical leakance used in the model for the basal Kirkwood confining unit (model unit C7).

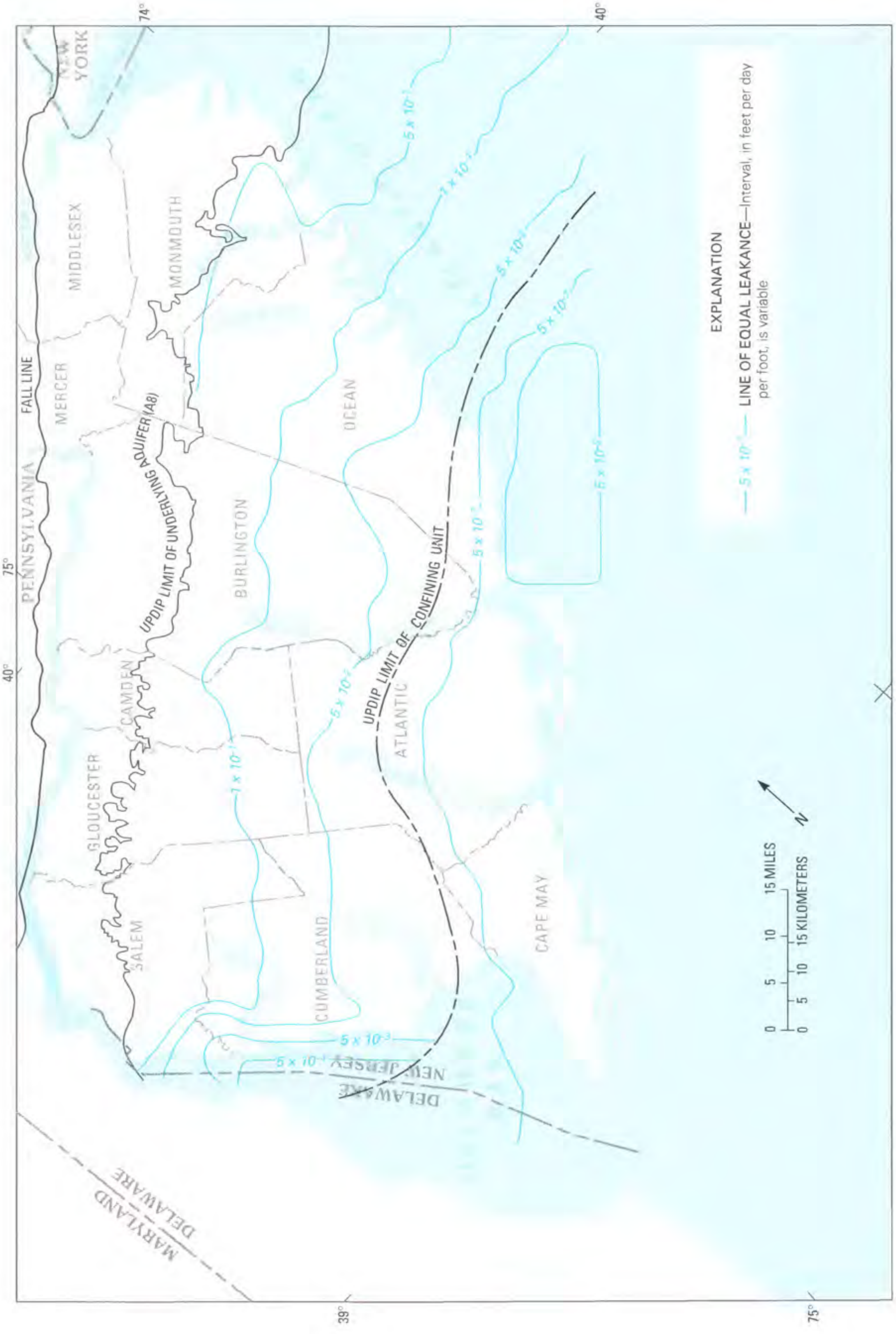


FIGURE 72.—Vertical leakance used in the model for the confining unit overlying the Rio Grande water-bearing zone (model unit C8).

FIGURE 73.—Vertical leakance used in the model for the estuarine clay confining unit (model unit C9).

between the lower and middle aquifers causes them to respond to stresses as one hydrologic unit. The Marshalltown-Wenonah confining unit (C4) has leakance between 1×10^{-6} and 1×10^{-4} (ft/d)/ft in downdip areas in Cape May, southeastern Atlantic, and southeastern Cumberland Counties (fig. 68).

In updip areas where the confining units are relatively thin, leakance ranges between 1×10^{-5} and 1×10^{-3} (ft/d)/ft. The highest leakance, greater than 1×10^{-4} (ft/d)/ft, is along the updip limit of the confining units among the Potomac-Raritan-Magothy aquifers (C1 and C2) and the Navesink-Hornerstown confining unit (C5). Another area of high leakance is where the model framework was adjusted to include an extension of the Merchantville-Woodbury confining unit (C3) over the modeled outcrop area of the upper Potomac-Raritan-Magothy aquifer (A3) (fig. 67). Leakance in this area is as high as 5×10^{-3} (ft/d)/ft. This extension of the confining unit was necessary to simulate water-level declines near the outcrop area of the upper Potomac-Raritan-Magothy aquifer near the Delaware River. This approach was probably necessary because of the large block size used to represent the relatively narrow outcrop area.

The onshore unconfined parts of the aquifers are overlain by an artificial confining unit representing an effective streambed leakance in the 11-layer steady-state and transient models. Flow between the constant-head boundary (representing the stream stages) and the water table is dependent on the leakance of this streambed confining unit. Streambed leakance values are calculated for each node from the long-term areally averaged stream stages, the simulated water-table altitude, the assumed recharge rate, and the simulated deep percolation (fig. 25). These leakances do not represent the hydraulic properties of actual streambed sediments and are not shown in figures 65 through 73. However, they do represent the regional connection between the water-table aquifer and streams. Streambed leakances range from 6×10^{-6} to 1.5×10^{-1} (ft/d)/ft and are generally the highest leakances (table 8) for each confining unit.

Areas where confining units are absent between aquifers are relatively small except between the lower and upper Kirkwood-Cohansey aquifers (A8 and A9 in updip areas). In areas where confining units are absent between aquifers over several nodes, artificial confining-unit thicknesses of about 0.1 foot were used to give very high leakance values to simulate the direct hydrologic connection between overlying and underlying aquifers. These leakance values (not shown in figures 65 to 73) are several orders of magnitude greater than the confining-unit leakance values. Vertical leakance between the upper and lower Kirkwood-Cohansey aquifers was calculated as the harmonic mean of the leakance of the adjacent aquifers. The vertical hydraulic conductivity of the aquifers was estimated to be 0.01 times the aquifers' horizontal hydraulic conductivity. The resulting vertical leakance between the lower and upper Kirkwood-Cohansey aquifers is about 5×10^{-2} to 1×10^{-1} (ft/d)/ft.

AQUIFER STORAGE

The model was calibrated by using a storage coefficient of 0.15 in unconfined areas. A storage coefficient of 1×10^{-4} was used for all confined aquifers except in downdip areas of the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) in Monmouth and Ocean Counties, where storage coefficients of 5×10^{-4} to 8×10^{-4} were used. The higher storage coefficient in the Potomac-Raritan-Magothy aquifers represents storage properties in areas where the aquifers are relatively thick. These higher values were used because simulated water levels in this area did not change significantly during calibration as a result of changing estimates of aquifer transmissivity and confining unit leakance within a reasonable range. Water levels were calibrated in this area by increasing aquifer storage of the three Potomac-Raritan-Magothy aquifers to five to eight times the original estimate of 1×10^{-4}, which was unchanged everywhere else in the modeled area.

PREPUMPING STEADY-STATE FLOW SYSTEM

In the simulated prepumping flow system, ground-water recharge occurs in areas where the land surface is higher than adjacent areas. These areas can be seen as unshaded areas in figure 41 in the southeastern parts of Salem, Gloucester, Camden, Burlington, Mercer, and Middlesex Counties; the western parts of Ocean and Monmouth Counties; and in central Cumberland, Cape May, and Atlantic Counties. Regional discharge areas in the prepumping flow system are shaded in figure 41 and include the Atlantic Ocean, Delaware River, Delaware Bay, Raritan Bay, and relatively large rivers in the Coastal Plain. These rivers include the Cohansey and Maurice Rivers in Cumberland County, the Great Egg Harbor River in Atlantic County, the Batsto and Mullica Rivers in Atlantic and Burlington Counties, the Toms River in Ocean County, and the Manasquan and Navesink Rivers in Monmouth County.

Simulated vertical flow in two cross sections of the Coastal Plain aquifers is shown in figure 74 for prepumping conditions. Simulated vertical hydraulic gradients between aquifers are highly variable but tend to be greater in updip than in downdip areas. Similarly, simulated vertical ground-water flow rates are also higher in updip areas (0 to 4 in/yr) than rates in downdip areas (0 to 0.2 in/yr). The prepumping flow system is described below for each aquifer.

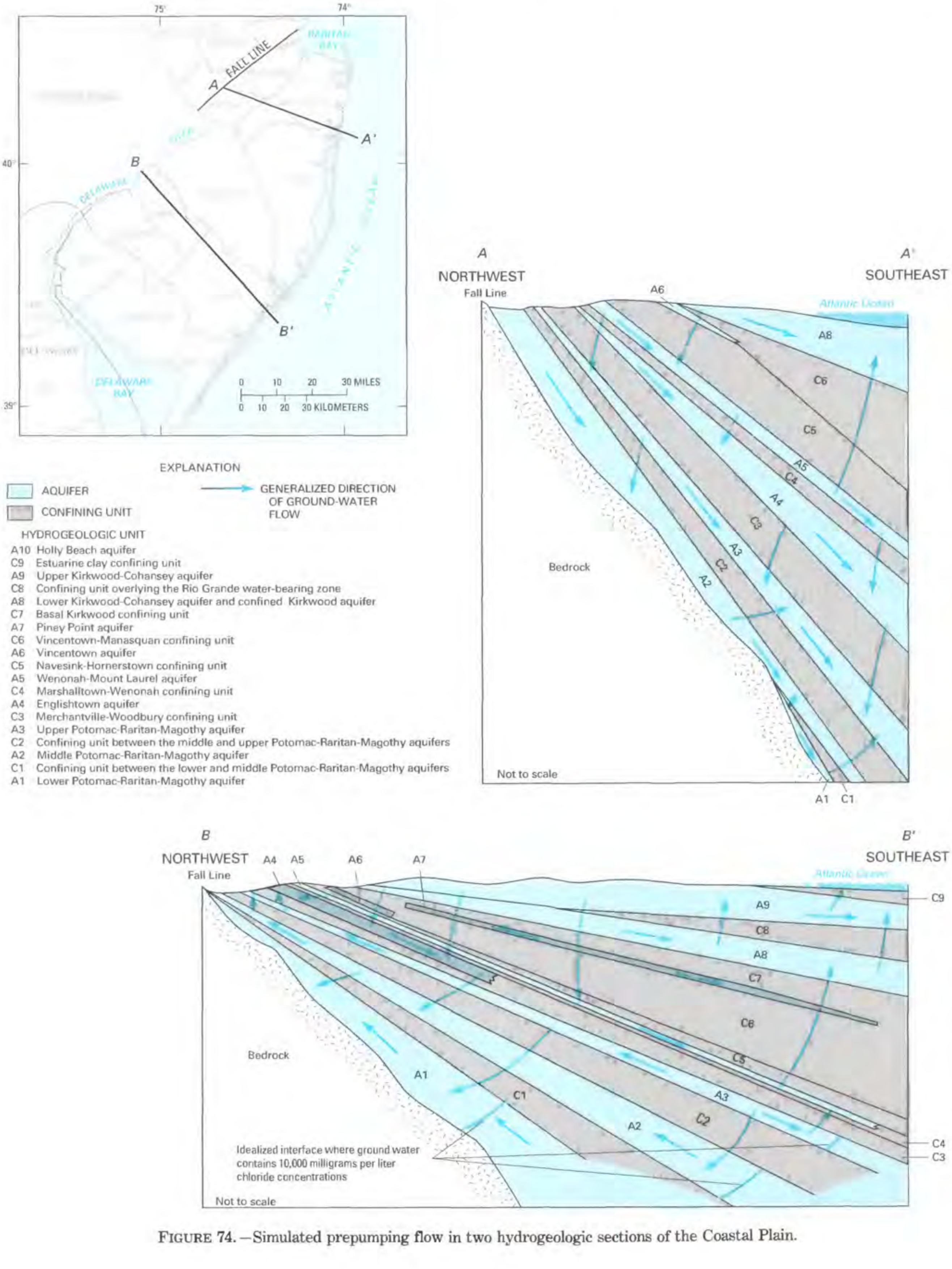


FIGURE 74. —Simulated prepumping flow in two hydrogeologic sections of the Coastal Plain.

Simulated prepumping water levels for the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) are within 10 feet of each other at all locations. Under prepumping conditions, the flow systems in the three aquifers are similar, and the Potomac-Raritan-Magothy aquifer system can be considered as a single hydrologic unit. Recharge to the aquifers is from the overlying Englishtown aquifer (A4) and the outcrop area in Mercer, Middlesex, and Monmouth Counties. Vertical flow through the tops of the Potomac-Raritan-Magothy aquifers is shown in figures 75, 76, and 77. Simulated prepumping flow is mostly downward from the Englishtown aquifer (A4) to the upper Potomac-Raritan-Magothy aquifer (A3) and generally downward into the other two Potomac-Raritan-Magothy aquifers. Vertical water-level differences between the upper aquifer and the Englishtown aquifer are as much as 70 feet in updip areas in western Monmouth County and central Camden Counties but generally 20 to 40 feet in downdip areas (figs. 32 and 33). Areas of upward discharge under prepumping conditions for the three Potomac-Raritan-Magothy aquifers are along the outcrop areas in Salem, Gloucester, Camden, and Burlington Counties; near the Delaware River Estuary in western Salem County; at Raritan Bay in the middle and upper Potomac-Raritan-Magothy aquifers; in downdip offshore areas; and in eastern Monmouth and Ocean Counties in the middle Potomac-Raritan-Magothy aquifer.

The Englishtown aquifer (A4) receives recharge in the simulated prepumping flow system from parts of its outcrop area in western Salem, western Gloucester, Mercer, Middlesex, and western Monmouth Counties and from the overlying Wenonah-Mount Laurel aquifer (A5) through the Marshalltown-Wenonah confining unit (C4) (fig. 78). A very small amount of upward leakage recharges the Englishtown aquifer from the upper Potomac-Raritan-Magothy aquifer (A3) offshore of Monmouth and Ocean Counties (fig. 77). Downward leakage from the Wenonah-Mount Laurel aquifer occurs over most of the Englishtown aquifer, except locally in western Salem County near the Delaware River, northeastern Monmouth County near the outcrop area, and in northeastern Ocean County. Upward discharges from the Englishtown aquifer to the Wenonah-Mount Laurel aquifer are less than 0.5 in/yr, except in northeastern Monmouth County where there are moderate leakage rates of 0.5 to 1.0 in/yr. Simulated water-level differences between the Englishtown and Wenonah-Mount Laurel aquifers for prepumping conditions are as much as 20 feet in updip areas of Monmouth County but less than 10 feet in downdip areas and updip areas in the western Coastal Plain (figs. 33 and 34). These vertical water-level differences are much less than those between the Englishtown aquifer and underlying upper Potomac-Raritan-Magothy aquifer (A3) because the leakance of the Merchantville-Woodbury confining unit (C3) below the Englishtown aquifer is 1 to 1.5 orders of magnitude less than the leakance of the Marshalltown-Wenonah confining unit (C4) above the Englishtown aquifer (table 8).

In the simulated prepumping flow system, the Englishtown aquifer (A4) receives moderate amounts of recharge, 0.5 in/yr or more, in the areas of the ground-water highs directly downdip of the outcrop area in Monmouth, central Camden, central Gloucester, and northeastern Salem Counties (fig. 33). Although some of this recharge flows downdip, some recharge flows updip toward the outcrop areas. Water flowing updip in Camden County discharges at the outcrop, but water flowing updip in Gloucester and Monmouth Counties may become downward leakage to the upper Potomac-Raritan-Magothy aquifer because the Englishtown aquifer outcrop areas in Gloucester and Monmouth Counties are not discharge areas.

The simulated prepumping flow in the Wenonah-Mount Laurel aquifer (A5) (fig. 79) is similar to flow in the underlying Englishtown aquifer (A4). Both aquifers are principally recharged by downward leakage from overlying aquifers. Ground-water highs exist downdip of the outcrop areas in central Camden, central Gloucester, northeastern Salem, and western Monmouth Counties (fig. 34), and local recharge and discharge areas are present along the outcrop. Areas with the highest recharge rates, greater than 0.5 in/yr, are in areas of ground-water highs. The Wenonah-Mount Laurel aquifer has more upward discharge areas than the Englishtown aquifer. Upward leakage from the Wenonah-Mount Laurel aquifer is in offshore areas and along the coast. Prepumping vertical water-level differences are less than 10 feet in updip areas between the Wenonah Mount-Laurel aquifer (A5) and the overlying Vincentown aquifer (A6) and the lower Kirkwood-Cohansey aquifer (A8) (figs. 34, 35, and 37). In downdip areas, vertical water-level differences are up to 60 feet between the Wenonah-Mount Laurel aquifer (A5) and the overlying Piney Point aquifer (A7) (figs. 34 and 36).

The prepumping vertical water-level differences between the Vincentown aquifer (A6) and the overlying lower Kirkwood-Cohansey aquifer (A8) are generally less than 20 feet, but as much as 60 feet in Camden County (figs. 35 and 37). Rates of ground-water recharge from the overlying aquifer are also greatest in Camden County, 0.2 to 3.0 in/yr. Ground water in the Vincentown aquifer is discharged to low-altitude areas along the outcrop or downward to the Wenonah-Mount Laurel aquifer. Hydraulic gradients within the Vincentown aquifer are similar to those in the underlying parts of the Wenonah-Mount Laurel aquifer.

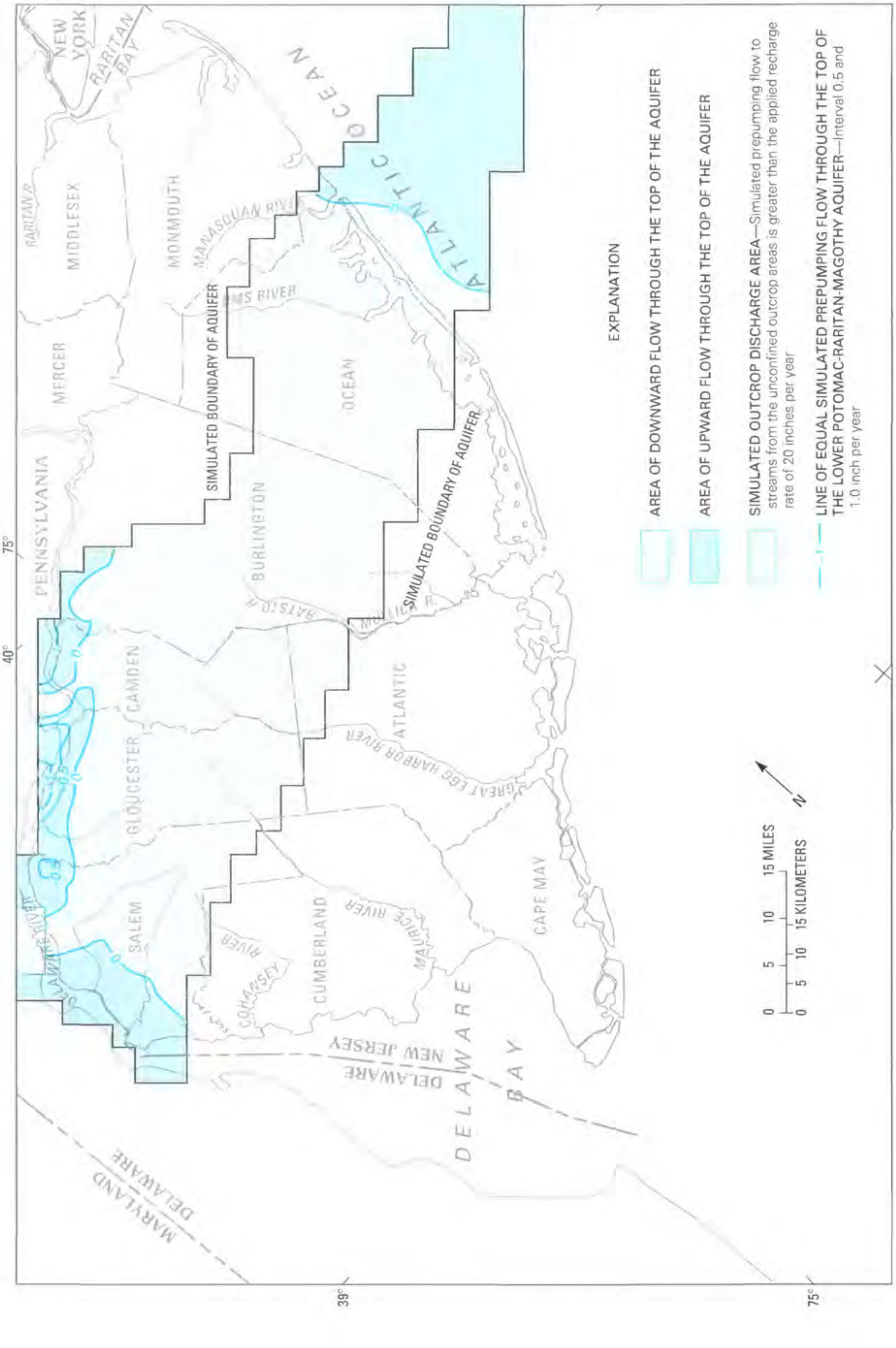


FIGURE 75.—Simulated prepumping flow through the top of the lower Potomac-Raritan-Magothy aquifer (model unit A1).

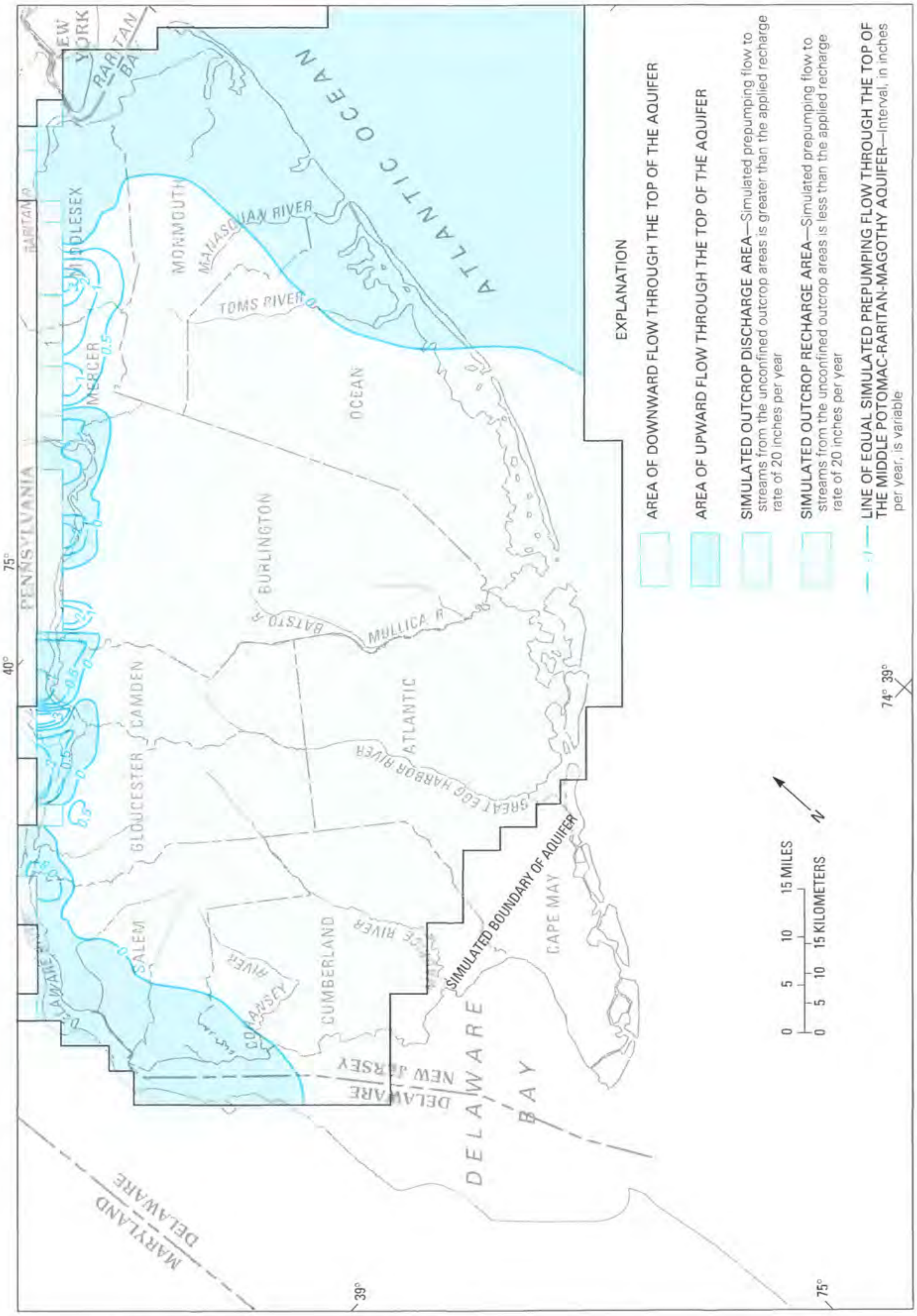


FIGURE 76.—Simulated prepumping flow through the top of the middle Potomac-Raritan-Magothy aquifer (model unit A2).

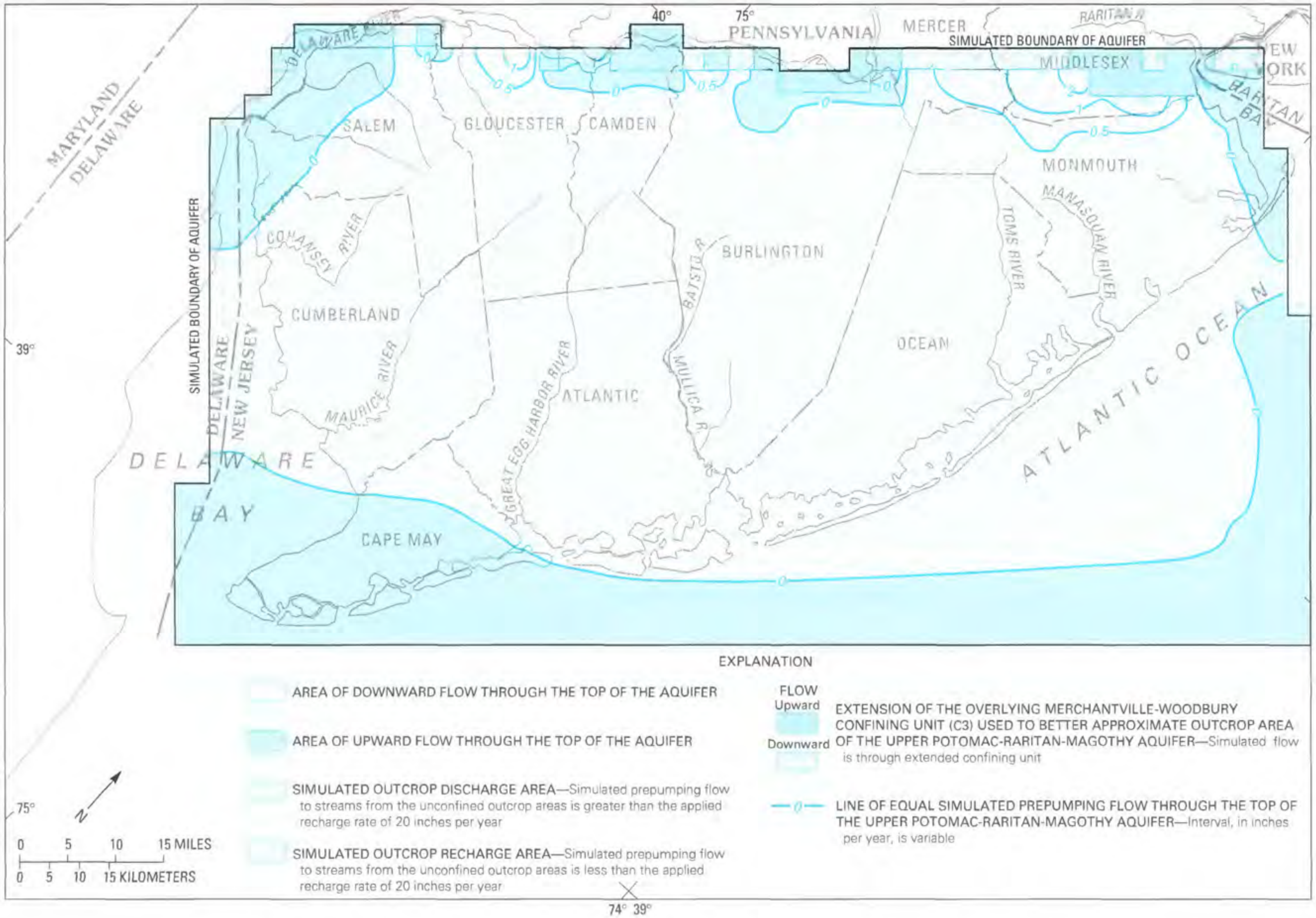


FIGURE 77.—Simulated prepumping flow through the top of the upper Potomac-Raritan-Magothy aquifer (model unit A3).

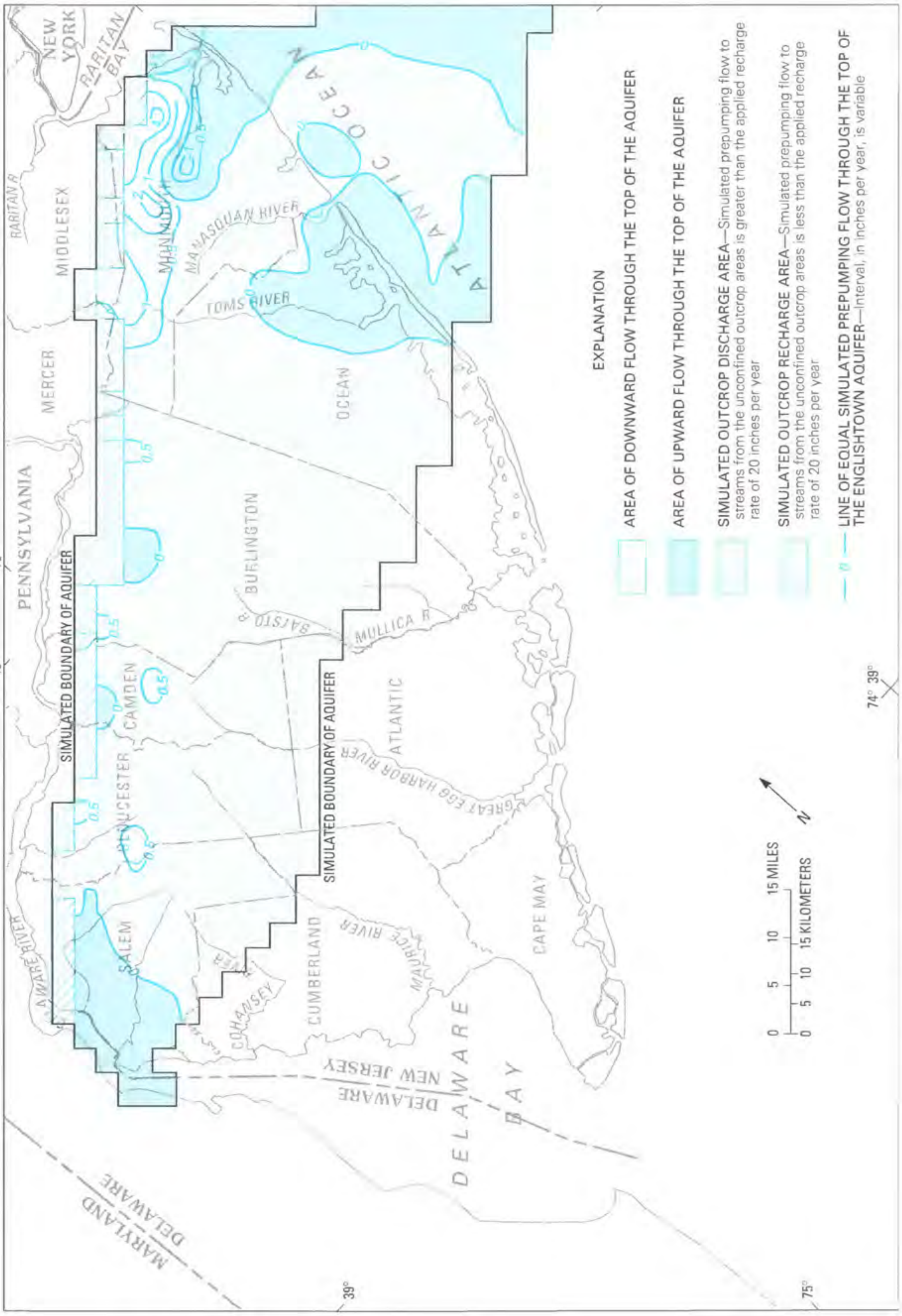


FIGURE 78.—Simulated prepumping flow through the top of the Englishtown aquifer (model unit A4).

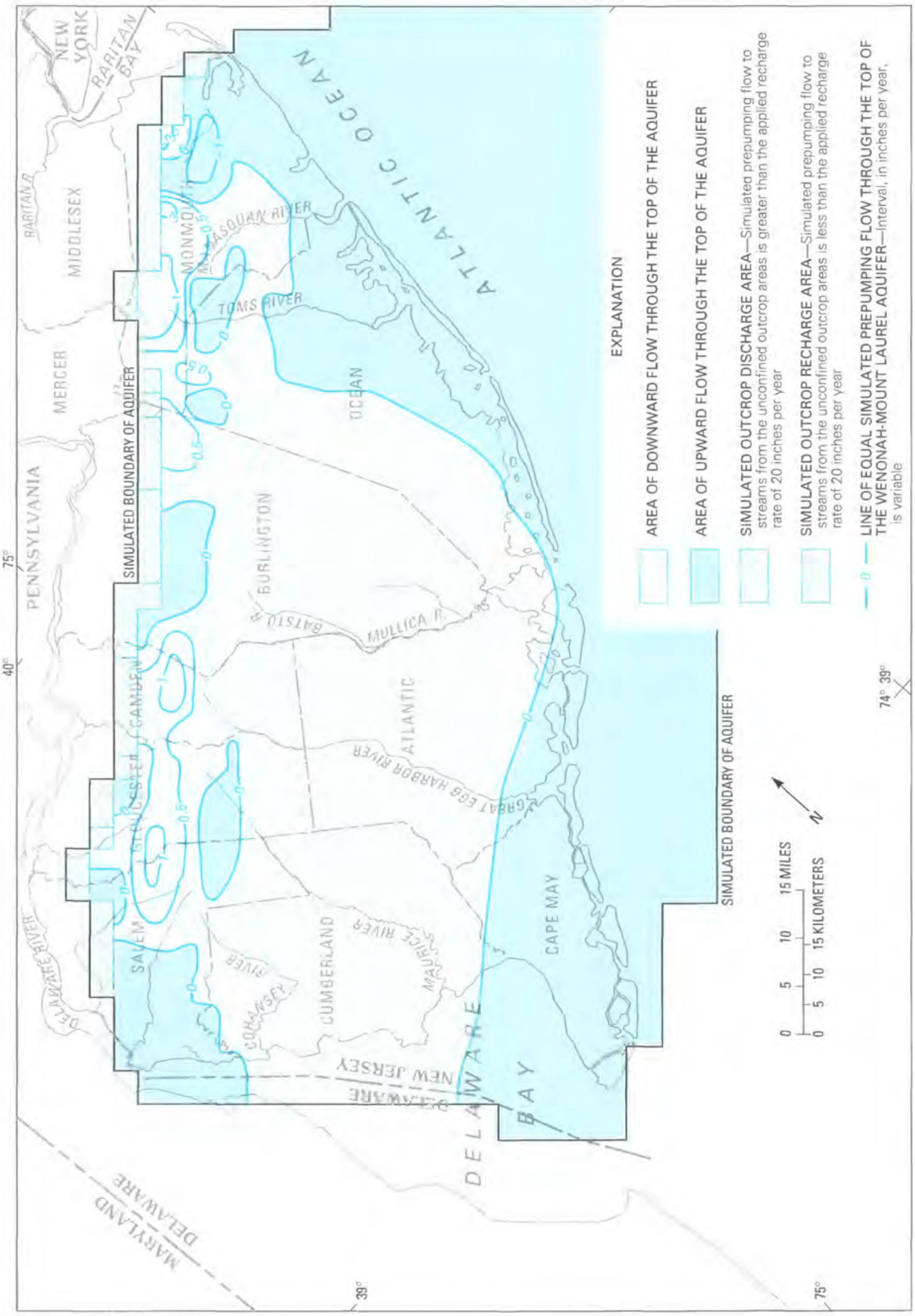


FIGURE 79.—Simulated prepumping flow through the top of the Wenonah-Mount Laurel aquifer (model unit A5).

The simulated prepumping flow in the Piney Point aquifer (A7) is dissimilar to flow in the underlying aquifers. Ground water within the Piney Point aquifer flows from the ground-water high in Burlington and Ocean Counties both downdip and to the southwest. The Piney Point aquifer is confined everywhere and the flow system is characterized as a more regional flow pattern that does not have local upward discharge areas beneath large rivers. This regional flow pattern is similar to flow patterns in downdip areas of the underlying aquifers.

Vertical water-level differences between the Piney Point aquifer (A7) and the overlying lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (A8) are 0 to 20 feet in the eastern part of the aquifer but as much as 60 feet in the western part (figs. 36 and 37). Recharge from the overlying aquifer is greatest, 0 to 1.5 in/yr, near the ground-water high. To the southwest, recharge is less than 0.2 in/yr. Areas of upward discharge, generally less than 0.2 in/yr, are in areas along the coast and in Burlington and Atlantic Counties beneath the Mullica River basin.

Simulated prepumping flow in the confined Kirkwood aquifer (A8) under prepumping conditions is from the updip limit of the aquifer to downdip areas. Vertical water-level differences between the confined Kirkwood aquifer (A8) and the overlying upper Kirkwood-Cohansey aquifer (A9) are from 10 to 30 feet (figs. 37 and 38). Recharge and discharge through the top of the confined Kirkwood aquifer are generally less than 0.2 in/yr (fig. 80). Upward leakage through the top of the confined Kirkwood aquifer is beneath the coast and the Mullica and Great Egg Harbor Rivers.

Recharge to the confined part of the upper Kirkwood-Cohansey aquifer (A9) in southern Cape May County is less than 0.2 in/yr from the overlying Holly Beach aquifer (A10). Prepumping vertical water-level differences between these two aquifers are less than 10 feet (figs. 38 and 39).

1978 TRANSIENT-STATE FLOW SYSTEM

Compared to the analysis of the prepumping flow system, the analysis of the 1978 flow system is more detailed because more measured data (Walker, 1983) are available. The simulated 1978 potentiometric surfaces (figs. 42–51) provide some additional information on the flow system in areas with few or no measured water levels, particularly in offshore areas. Simulated offshore water levels in the aquifers may not be the same as actual offshore water levels, because the model cannot be accurately calibrated without measured water levels in these areas. However, the simulated water levels provide estimates of hydraulic gradients within and between aquifers in offshore areas. Lateral hydraulic gradients within and vertical hydraulic gradients between aquifers are discussed. Simulated flow rates through the tops of selected aquifers are given, and regional recharge and discharge areas and hydraulic gradients are discussed. The reader is referred to Walker (1983) for a detailed discussion of 1978 water levels and water-level fluctuations in the major aquifers.

The steepest simulated lateral hydraulic gradients within the aquifers are near the major cones of depression. Large cones of depression have steeper simulated gradients on the updip side than on the downdip side. In the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3), simulated lateral hydraulic gradients between the Delaware River and the center of the large cone of depression in central Camden County are about 10 ft/mi (figs. 42–44). Downdip of this large cone, hydraulic gradients within these aquifers are generally less than 5 ft/mi. The steepest gradients in these aquifers are in the middle Potomac-Raritan-Magothy aquifer (A2) between the outcrop area and the steep cone of depression in Middlesex and northwestern Monmouth Counties (fig. 43). Hydraulic gradients in this area are 15 to 20 ft/mi. Northeast of this cone of depression beneath Raritan Bay, gradients within the middle Potomac-Raritan-Magothy aquifer are about 10 ft/mi. Hydraulic gradients in the Englishtown aquifer (A4) and the Wenonah-Mount Laurel aquifer (A5) are 25 to 30 ft/mi near the large cone of depression in southeastern Monmouth and northwestern Ocean Counties (figs. 45 and 46). Offshore of this cone, simulated hydraulic gradients are about 10 ft/mi within each of these aquifers. In the confined Kirkwood aquifer (A8), hydraulic gradients updip of the cone of depression at Atlantic City are 6 to 8 ft/mi, but simulated gradients in offshore areas are 2 to 4 ft/mi (fig. 49).

Simulated vertical gradients between aquifers and flow patterns within aquifers for 1978 differ from those in the prepumping flow system in both updip outcrop areas and downdip areas distant from the major cones of depression. Simulated vertical flow for 1978 in two cross sections of the Coastal Plain aquifers is shown in figure 81. Simulated vertical flow rates in the 1978 flow system are between 0.0 and 0.2 in/yr in most areas, but rates of 1 to 5 in/yr are common locally, particularly in updip areas and near the major cones of depression. Lateral flow within the aquifers is dominated by large regional cones of depression.

As in the simulated prepumping flow system, the vertical water-level differences between the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2) are generally less than 10 feet (figs. 42 and 43). However, flow through the top of the lower Potomac-Raritan-Magothy aquifer for January 1, 1978, (fig. 82) is greatly changed from prepumping conditions (fig. 75). Large

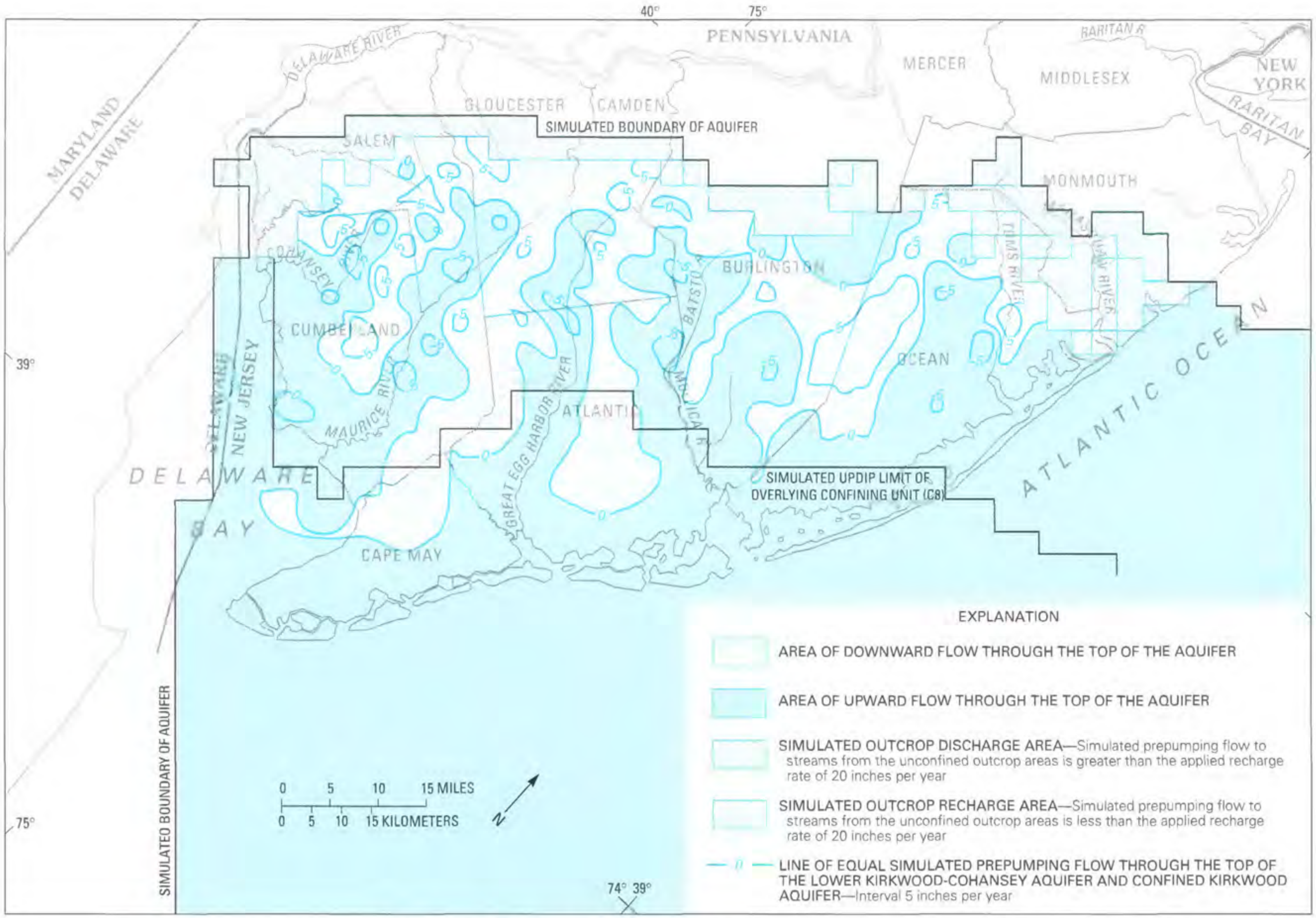


FIGURE 80.—Simulated prepumping flow through the top of the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (model unit A8).

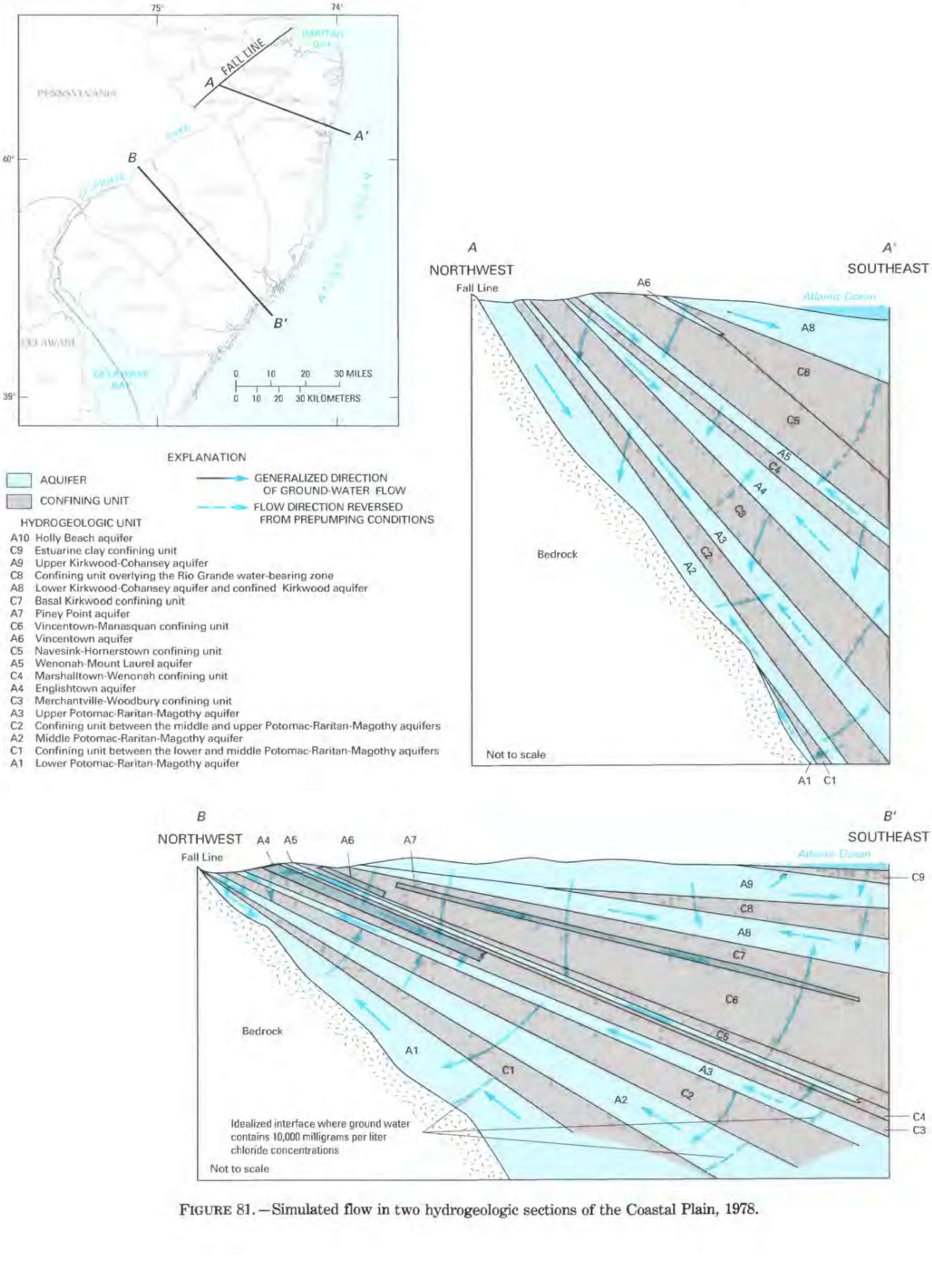


FIGURE 81.—Simulated flow in two hydrogeologic sections of the Coastal Plain, 1978.

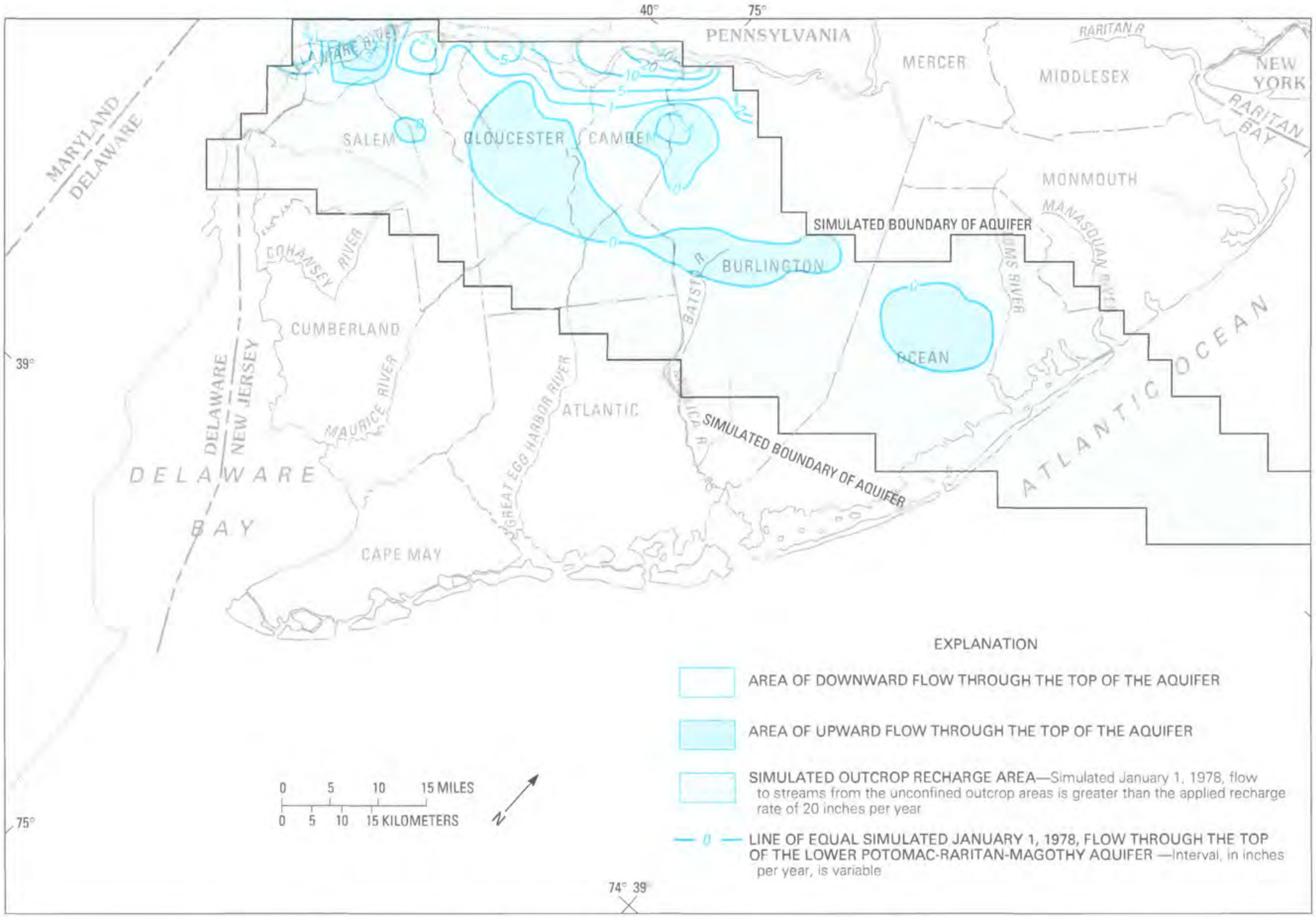


FIGURE 82.—Simulated January 1, 1978, flow through the top of the lower Potomac-Raritan-Magothy aquifer (model unit A1).

areas of upward discharge are shown for the lower Potomac-Raritan-Magothy aquifer in 1978. Although flow rates between the aquifers are generally less than 1.0 in/yr, simulated upward discharge near the cone of depression at Camden is as high as 2.0 in/yr and downward recharge into the aquifer is generally between 1.0 and 20.0 in/yr in northwestern Camden County. Although vertical water-level differences are relatively low, flow rates between the lower and middle Potomac-Raritan-Magothy aquifers are high because of the high leakance of the intervening confining unit. Downward leakage greater than 20 in/yr for two nodes in northwestern Camden County is unrealistically high and reflects similar high flow through overlying confining units. This high flow results from the availability of an unlimited source of water from the overlying constant-head stream boundary and probably indicates that the actual induced recharge into the aquifer is at a smaller rate over a larger area than is simulated by the model.

Simulated vertical water-level differences between the middle and upper Potomac-Raritan-Magothy aquifers (A2 and A3) in 1978 (figs. 43 and 44) are greater than the water-level differences in the prepumping flow system but are relatively small, generally less than 20 feet. The greatest simulated vertical water-level differences are in Salem County, where they are as much as 30 feet, and in Monmouth and Middlesex Counties, where they are as much as 70 feet. Simulated flow through the top of the middle Potomac-Raritan-Magothy aquifer for January 1, 1978, is shown in figure 83. The major changes in flow from prepumping to 1978 conditions (fig. 76) are (1) high upward discharge, more than 1.0 in/yr from the middle Potomac-Raritan-Magothy aquifer near the area of the cone of depression in Camden County; (2) disappearance of upward discharge from the middle Potomac-Raritan-Magothy aquifer under Raritan Bay, in Middlesex and northern Monmouth Counties, under the Delaware River, and under Delaware Bay adjacent to western Salem County; (3) increase in recharge along the outcrop area of the middle Potomac-Raritan-Magothy aquifer in Mercer and Middlesex County from about 3 to 5 in/yr; and (4) relatively high recharge, generally between 1.0 and 20.0 in/yr, to the middle aquifer in northwestern Camden County. The downdip recharge greater than 20 in/yr to the middle Potomac-Raritan-Magothy aquifer for five nodes near the Delaware River in Camden and Burlington Counties is unrealistically high for the same reasons indicated above for similar high flow into the lower Potomac-Raritan-Magothy aquifer (A1). Another important feature of the 1978 flow system is the relatively low recharge, less than 0.5 in/yr, over the steep cone in the middle aquifer near Raritan Bay.

The simulated vertical water-level differences between the upper Potomac-Raritan-Magothy aquifer (A3) and the Englishtown aquifer (A4) in 1978 (figs. 44 and 45) are much greater than those in the prepumping flow system. Head differences in 1978 are as much as 140 feet in Camden County, 100 feet in western Monmouth County, and more than 180 feet in southern Monmouth and northeastern Ocean Counties. Simulated flow through the top of the upper Potomac-Raritan-Magothy aquifer for January 1, 1978, is shown in figure 84. Although most of the upper Potomac-Raritan-Magothy aquifer receives 0.0 to 0.5 in/yr of downward recharge, similar to prepumping conditions (fig. 77), some areas had significant changes in flow. These changes include (1) disappearance of discharge to the Delaware River and Bay near western Salem County; (2) changing of areas from high discharge to areas of high discharge between 1.0 and 10.0 in/yr near the outcrop area in northern Camden, Gloucester, and Burlington Counties; (3) increase in recharge from over 1.0 to over 3.0 in/yr locally near the outcrop area in Middlesex County; (4) change in flow of about 3 in/yr beneath Raritan Bay, from over 1.0 in/yr of upward discharge to more than 2.0 in/yr of recharge; and (5) very low upward discharge (less than 0.2 in/yr) in Ocean and Monmouth Counties (beneath the large cone of depression in the Englishtown aquifer). Flow rates are low beneath the cone in the Englishtown aquifer despite the extreme hydraulic gradients, because of the relatively low hydraulic conductivity of the intervening Merchantville-Woodbury confining unit (C3).

The 1978 transient-flow system within the three Potomac-Raritan-Magothy aquifers includes relatively high rates of recharge in and near the outcrop areas in Mercer and Middlesex Counties, along the Delaware River, and beneath Raritan Bay. Recharge from the Delaware River flows toward the large cone of depression in central Camden County and smaller cones in Salem County (figs. 42–44). Recharge from Mercer and Middlesex Counties flows toward the cones of depression near Raritan Bay or downdip. This downdip flow is discharged upward offshore or flows southwestward, then updip to the Camden cone.

Simulated vertical water-level differences in 1978 between the Englishtown aquifer (A4) and the Wenonah-Mount Laurel aquifer (A5) are much smaller than those between the underlying upper Potomac-Raritan-Magothy aquifer (A3) and the Englishtown aquifer. Vertical water-level differences are as much as 50 feet near the center of the large cone of depression near the coast in Monmouth and Ocean Counties, 0 to 10 feet in downdip areas away from the cone, and about 20 feet in updip areas (figs. 45 and 46). Simulated flow through the top of the Englishtown aquifer is shown in figure 85. Simulated flow rates between the Englishtown and Wenonah-Mount Laurel aquifers are much higher than those between upper Potomac-Raritan-Magothy and

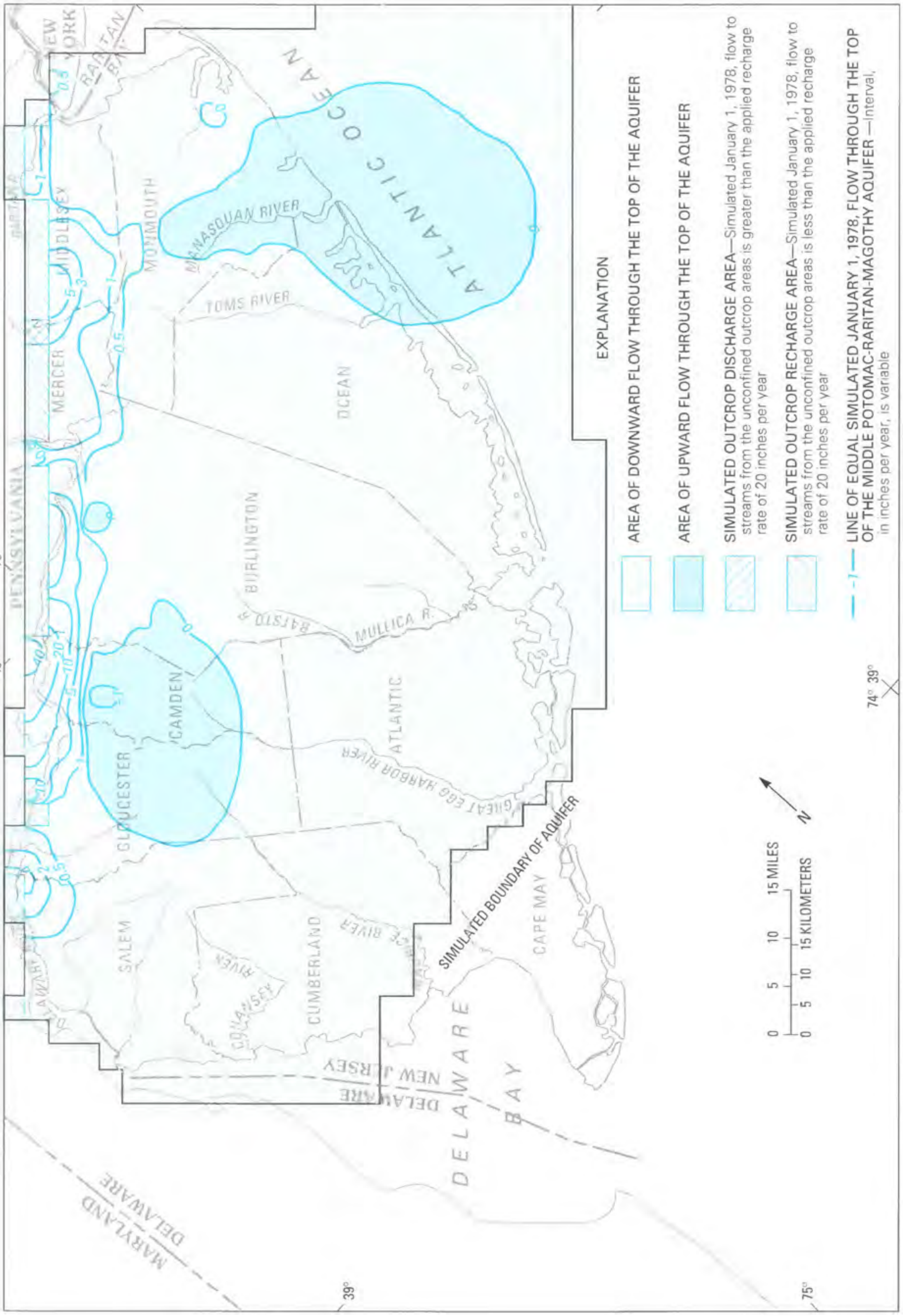


FIGURE 83.—Simulated January 1, 1978, flow through the top of the middle Potomac-Raritan-Magothy aquifer (model unit A2).

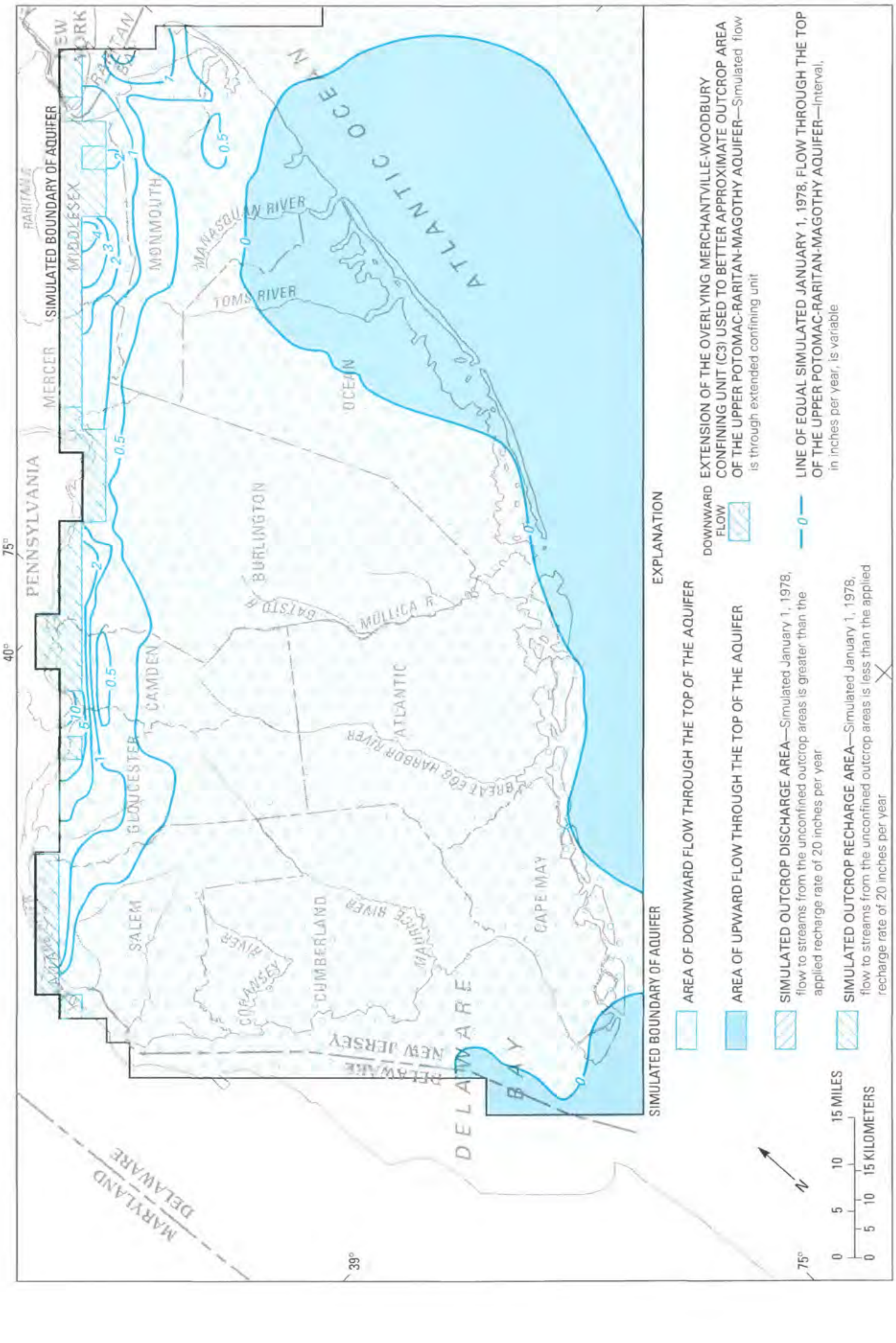


FIGURE 84.—Simulated January 1, 1978, flow through the top of the upper Potomac-Raritan-Magothy aquifer (model unit A3).

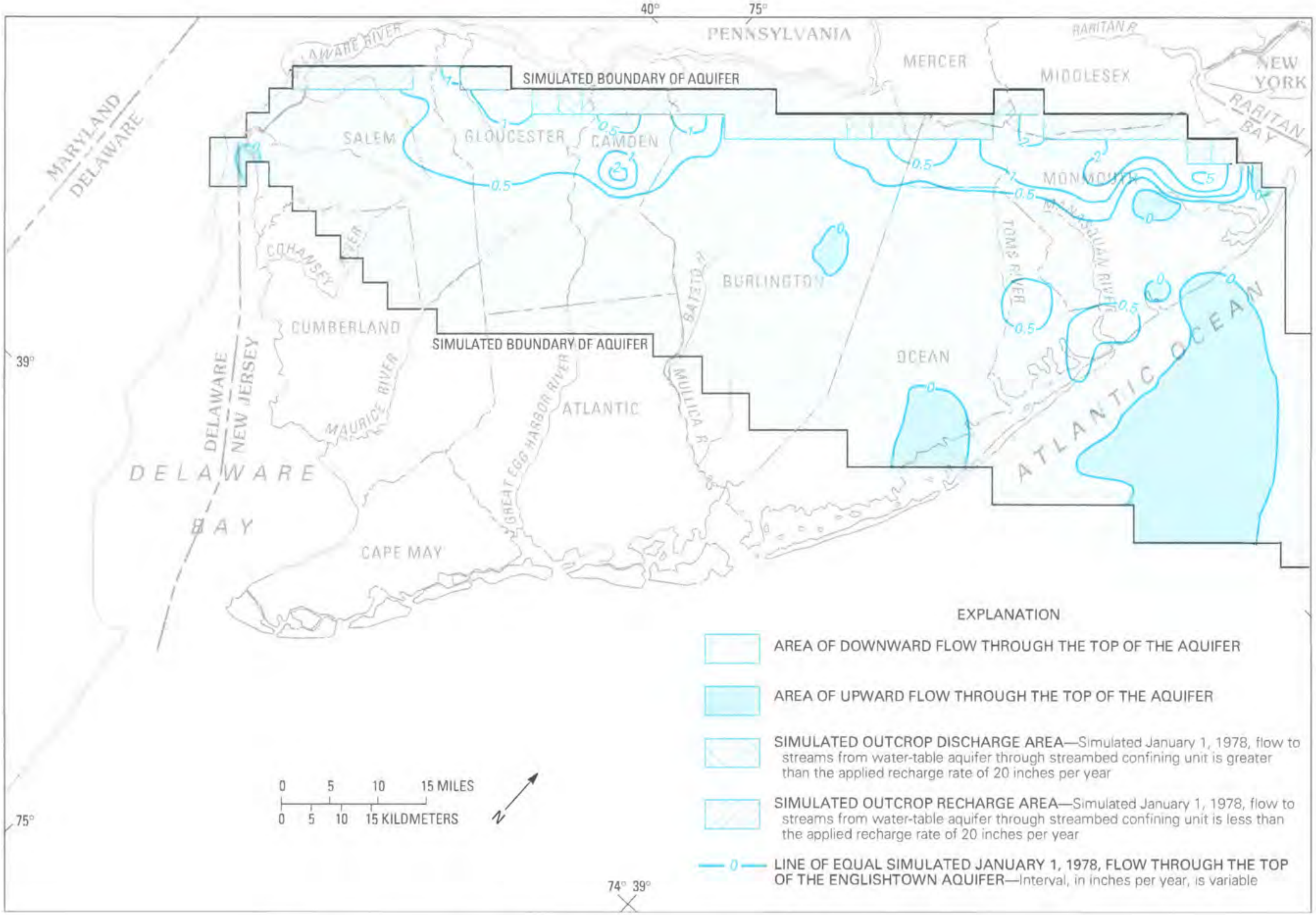


FIGURE 85.—Simulated January 1, 1978, flow through the top of the Englishtown aquifer (model unit A4).

Englishtown aquifers because of the relatively high leakance of the Marshalltown-Wenonah confining unit (C4) (table 8). Downward recharge from the Wenonah-Mount Laurel aquifer over the large area of the cone of depression in the Englishtown aquifer is more than 0.5 in/yr in 1978. Recharge into the Englishtown aquifer in the updip areas has increased since prepumping conditions (fig. 78) by as much as 1.5 in/yr in Camden County and about 1 in/yr in Monmouth County. The prepumping areas of upward discharge from the Englishtown aquifer are either eliminated or greatly reduced in the 1978 flow system, including areas within the outcrop.

Simulated vertical water-level differences between the Wenonah-Mount Laurel aquifer (A5) and the overlying aquifers in 1978 range from about 20 feet in updip and downdip areas in the southwest Coastal Plain to more than 180 feet in the area of the cone of depression in Monmouth and Ocean Counties (figs. 46–48). Flow through the top of the Wenonah-Mount Laurel aquifer is shown in figure 86. Downward recharge near the outcrop in 1978 is as much as 1.0 to 3.0 in/yr. Upward discharge near the outcrop in 1978 is less than that under prepumping conditions (fig. 79). Offshore from Monmouth and Ocean Counties, where there was upward discharge under prepumping conditions, are areas of downward recharge in 1978. Flow beneath Delaware Bay changed directions from prepumping conditions in areas southwest of Salem and Cumberland Counties.

The flow systems in the Englishtown (A4) and Wenonah-Mount Laurel (A5) aquifers are recharged from ground-water highs near the outcrops in Camden, Gloucester, and Monmouth Counties (figs. 45 and 46). Some flow from these areas is to local wells, to the ground-water low in Burlington County, or it is discharged upward offshore and into Delaware Bay. However, most of the recharge flows downdip and eastward to the area of the large cones of depression in Monmouth and Ocean Counties. A unique aspect of the flow system occurs beneath Delaware Bay offshore of Cumberland and Cape May Counties in the Wenonah-Mount Laurel aquifer, where water is discharged both downward to the upper Potomac-Raritan-Magothy aquifer (A3) and upward to overlying Piney Point aquifer (A7). This flow from the Wenonah-Mount Laurel aquifer is caused by withdrawals in the adjacent aquifers in Delaware and New Jersey.

The simulated 1978 flow systems in the Vincentown (A6) and Piney Point (A7) aquifers (figs. 47 and 48) are very similar to the prepumping flow systems. Recharge into the Vincentown aquifer from the lower Kirkwood-Cohansey aquifer (A8) is 0.5 to 5.0 in/yr higher than under prepumping conditions because of increased downward discharge to the Wenonah-Mount Laurel aquifer (A5). Simulated 1978 water levels in the Vincentown aquifer are about 20 feet higher than in most areas of the Wenonah-Mount Laurel aquifer (figs. 46 and 47). In eastern Monmouth County, they are up to 100 feet higher than water levels in the Wenonah-Mount Laurel aquifer. Recharge and discharge rates between the Piney Point and lower Kirkwood-Cohansey aquifers are changed only slightly from prepumping conditions. Prepumping recharge and discharge areas along the coast of Ocean County have changed to an area of downward recharge over the small cone of depression in the Piney Point (fig. 48) and to an area of increased upward discharge, 0.4 in/yr, under the limb of the cone of depression in the confined Kirkwood aquifer (A8) (fig. 49).

The simulated 1978 flow system in the unconfined lower and upper Kirkwood-Cohansey aquifers (A8 and A9) is generally unchanged from prepumping conditions. Simulated changes in water levels within the unconfined aquifers and changes in downward flow rates from the unconfined aquifers are insignificant. However, the flow in the confined Kirkwood aquifer (A8) and in the confined part of the upper Kirkwood-Cohansey aquifer (A9) has changed from prepumping conditions. Essentially all simulated flow in these confined aquifers is to the areas of the two cones of depression (figs. 49 and 50). One cone is centered near Atlantic City in the confined Kirkwood aquifer (A8) (fig. 49), and the other cone is in the confined part of the upper Kirkwood-Cohansey aquifer (A9) in Cape May County (fig. 50). Water-level differences between the confined Kirkwood aquifer and the upper Kirkwood-Cohansey aquifer are more than 70 feet near Atlantic City. Vertical water-level differences between the confined part of the upper Kirkwood-Cohansey aquifer and the Holly Beach aquifer (A10) are more than 20 feet. Simulated upward discharge under prepumping conditions in onshore areas and offshore in the Atlantic Ocean is not present in 1978 (fig. 87). However, flow rates into the confined parts of the lower and upper Kirkwood-Cohansey aquifers through overlying confining units are generally less than 0.2 in/yr in 1978.

SENSITIVITY ANALYSIS

Sensitivity simulations provide information on which the impact of assumptions and model input data on the simulation results can be evaluated. The sensitivity simulations are made with the model data used for calibration but with changes within a reasonable range to one of the hydraulic characteristics, such as aquifer storage, or to one of the model assumptions, such as location of the interface boundary. Data are changed for several model units, a single model unit, or for a limited area depending on the characteristic being tested. Water

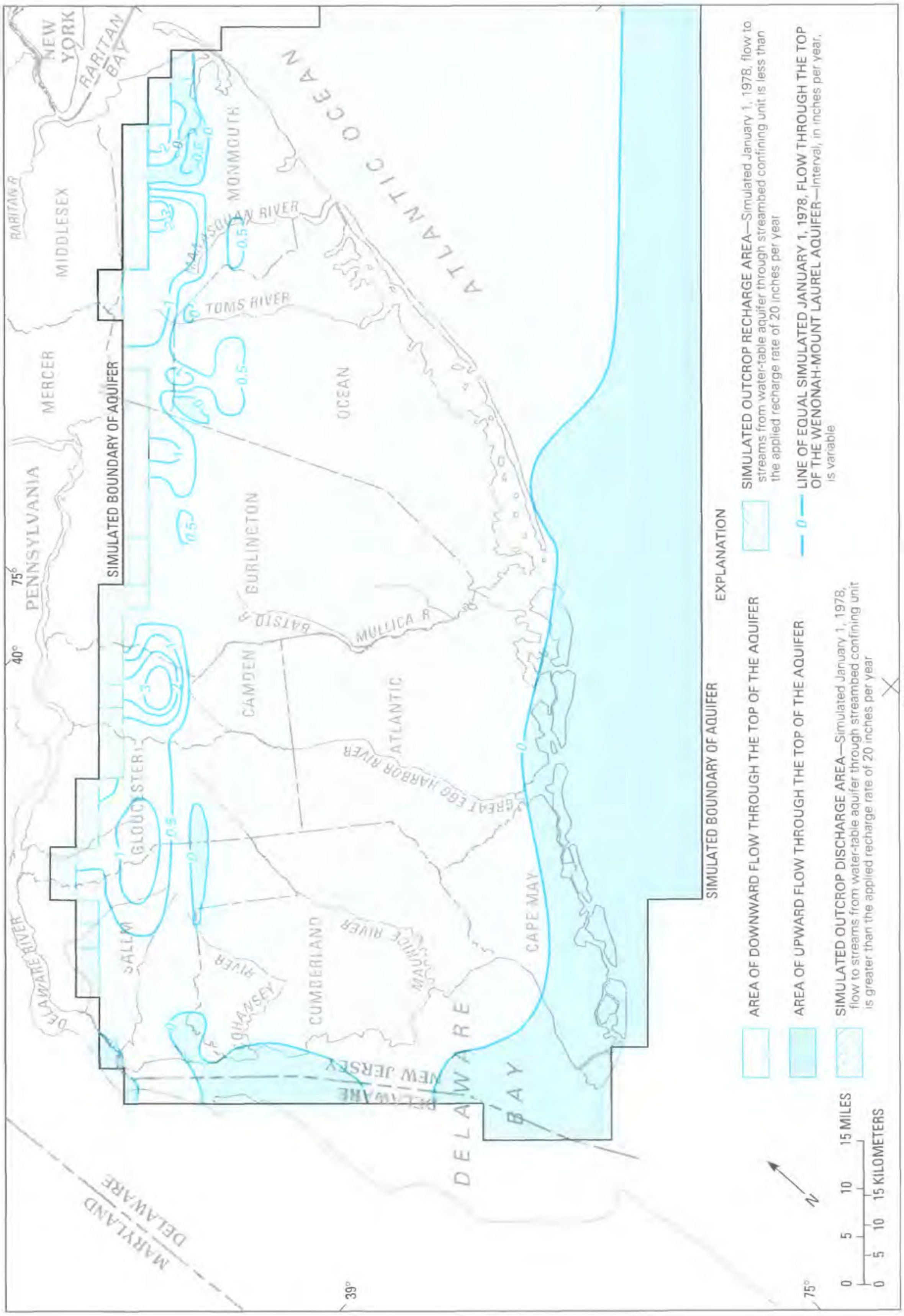


FIGURE 86.—Simulated January 1, 1978, flow through the top of the Wenonah-Mount Laurel aquifer (model unit A5).

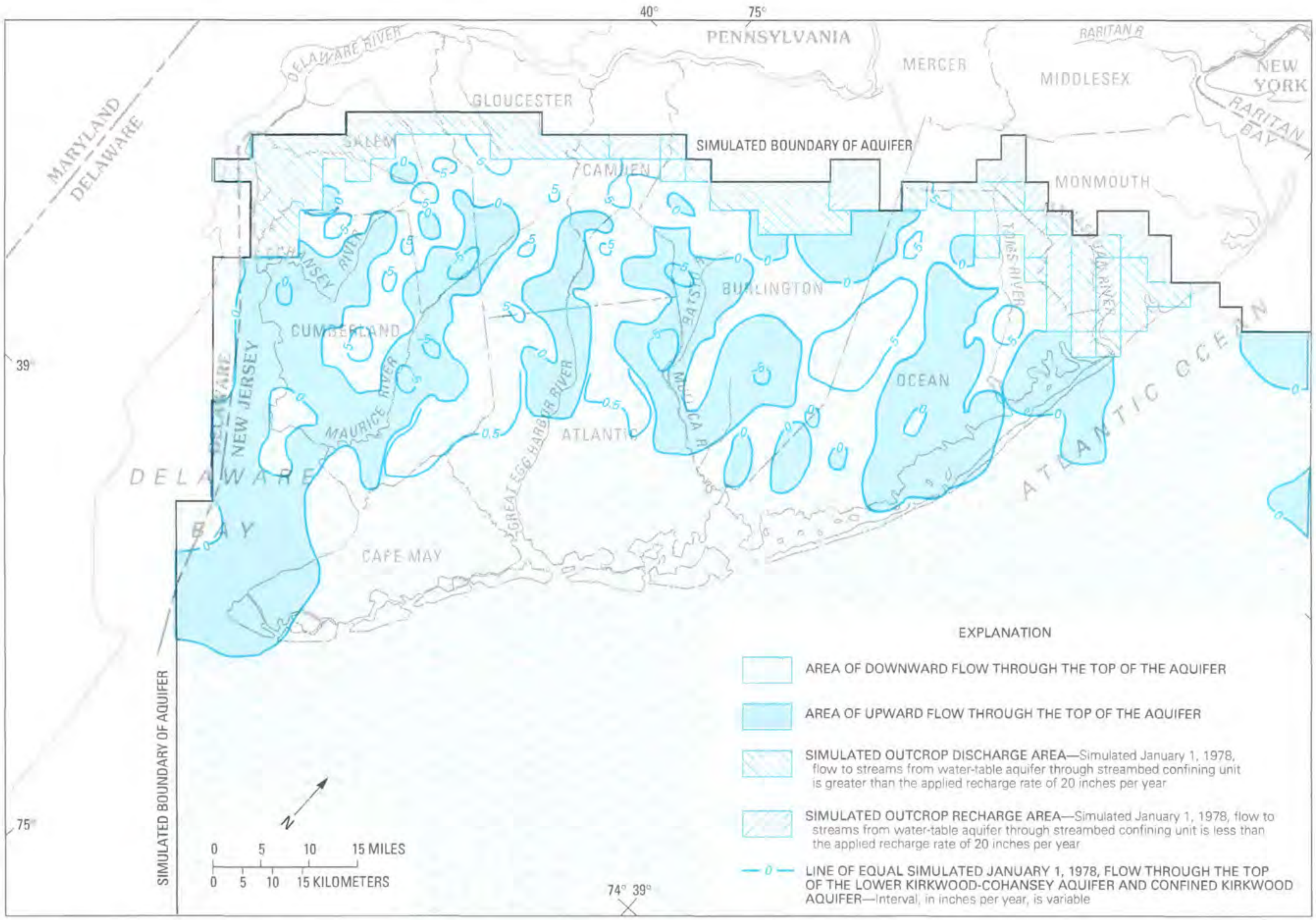


FIGURE 87.—Simulated January 1, 1978, flow through top of the lower Kirkwood-Cohansey aquifer and confined Kirkwood aquifer (model unit A8).

levels from the sensitivity simulation are compared to water levels from the calibration simulation. If a small change in a specified hydraulic characteristic produces large changes in water levels, the model is useful in refining initial estimates of that characteristic. Conversely, if a large change in a specified hydraulic characteristic produces very little water-level difference, the flow model is not effective for estimating that characteristic. Consequently, sensitivity simulations are also useful for identifying types of data needed to improve the simulation of flow within the aquifer system.

Evaluating model assumptions by sensitivity analysis is limited in several ways. Sensitivity analyses can be made only on a limited number of characteristics and areas, and therefore only a small percentage of the factors affecting flow in an area can be tested. Also, the results of model calibration are not unique, and a different combination of aquifer and confining-unit characteristics could have resulted in model calibration. Sensitivity analyses on different calibrated model data may result in different conclusions. However, within these limitations, the following sensitivity analyses provide a means for evaluating the simulation results and assumptions.

TRANSMISSIVITY AND VERTICAL LEAKANCE OF THE CONFINING UNITS

Sensitivity analyses of transmissivity and vertical leakance of the confining units were made for the area along the Delaware River and areas near the center of the large cones of depression in the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) and the confined Kirkwood aquifer (A8). The areas near the center of the large cones were chosen because they are areas with high withdrawal rates and have a large amount of information on water levels; therefore, these areas of the model were better calibrated than other areas. The area along the Delaware River was chosen because it is an outcrop area near an area with high withdrawal rates.

Four simulations were made with transmissivity and leakance changes as listed in table 9 for the area of the cone of depression centered in Camden County. The model area and results of these sensitivity simulations are also summarized in table 9 (simulations 1 to 4). These simulations show that large changes in transmissivity and large increases in the leakance of the overlying and intervening confining units near the center of the cone of depression in Camden County have a significant effect on simulated water levels. These large effects on water levels indicate that additional data on these hydraulic characteristics in the areas of the cones of depression will increase the accuracy of the model results. Decreases in vertical leakance of the Merchantville-Woodbury confining unit (C3) produced very little water-level change; therefore, lower estimates of leakance based on additional data in this area would not improve the accuracy of the model results.

Four sensitivity simulations were made by varying the transmissivity and vertical leakance of the sediments along the Delaware River. The data changes, model area, and results of these sensitivity simulations are given in table 9 (simulations 5 to 8). These simulations show that the rate of flow induced from the Delaware River is controlled more by vertical leakance of the confining units than by the transmissivity of the aquifers along the river. Although the simulated outcrop areas provide regional recharge to the ground-water flow system, the large cell size prevents accurate simulation of water levels in aquifer outcrop areas. Results and conclusions relating specifically to aquifer outcrop areas must be used cautiously and are useful as guidelines for more detailed studies in these areas.

Four sensitivity analyses were also made for the area near the center of the large cone of depression in Atlantic County, varying transmissivity and leakance. The data changes, model area, and results of these sensitivity simulations are given in table 9 (simulations 9 to 12). These simulations show that, like the water levels in the Potomac-Raritan-Magothy aquifers (A1, A2, and A3), simulated water levels in the confined Kirkwood aquifer (A8) are greatly affected by large changes in transmissivity or in leakance of the overlying confining unit. Therefore, the accuracy of the model results would be improved by additional data on these hydraulic characteristics in this area. Some change in transmissivity or leakance could be made in the calibrated model to produce more acceptable simulated water levels, but these changes would have to be much smaller than the range of values tested in these sensitivity analyses.

Two additional sensitivity simulations were made of the transmissivity of the confined Kirkwood aquifer (A8) (simulations 13 and 14, table 9). The water-level changes in these sensitivity simulations are not large; however, the simulations show that horizontal flow through only a one- or two-cell width along the updip limit of the confined Kirkwood aquifer significantly affects water levels near Atlantic City. The hydraulic connection between the unconfined lower Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer is not well defined (Zapecza, 1984, p. 29). Therefore, additional data on the hydraulic characteristics between the unconfined lower Kirkwood-Cohansey aquifer and the confined Kirkwood aquifer will improve model results near the center of the cone.

TABLE 9.—*Results of sensitivity analyses on transmissivity and confining-unit leakance*

Model unit: A1, lower Potomac-Raritan-Magothy aquifer
A2, middle Potomac-Raritan-Magothy aquifer
A3, upper Potomac-Raritan-Magothy aquifer
A4, Englishtown aquifer
A5, Wenonah-Mount Laurel aquifer
A6, Vincentown aquifer
A7, Piney Point aquifer
A8, lower Kirkwood-Cohansey and confined Kirkwood aquifers
A9, upper Kirkwood-Cohansey aquifer
A10, Holly Branch
Nodes: Location shown on figure 25

Simulation	Model unit	Nodes		General description of change made to calibrated data	Resulting water-level changes for January 1, 1978
		Rows	Columns		
1	A1, A2, and A3	5–10	18–25	100 percent increase in transmissivity in the confined Potomac-Raritan-Magothy aquifers and in Camden, Gloucester, and Burlington Counties.	Cone of depression is about 15 ft shallower in A1 and A2 and 30 ft shallower in A3.
2				Same as above, but a 50-percent decrease.	Cone of depression is about 20 ft deeper in A1 and A2 and 30 ft deeper in A3. Less than 5-ft water-level decline in downdip areas of A1 and A2.
3	C3	5–10	18–25	An order-of-magnitude increase in leakance of the Merchantville-Woodbury confining unit in Camden, Gloucester, and Burlington Counties.	Cone of depression is about 15 ft shallower in A1 and A2 and 30 ft shallower in A3. Ground-water high above cone in A4 is about 20 ft lower, 40 ft lower locally. Similar ground-water highs in A5 and A6 are about 10 ft lower.
4				Same as above, but an order-of-magnitude decrease.	Cone of depression is about 5 ft deeper in A1 and A2 and 10 ft deeper in A3. Ground-water high in A4 is about 15 ft higher. Less than 5-ft water-level decline in downdip areas of A1, A2, and A3.
5	A1, A2, and A3	2–4	18–25	100-percent increase in transmissivity in the Potomac-Raritan-Magothy aquifers near the Delaware River in Camden and Burlington Counties.	Cone of depression is about 10 ft shallower in A1, A2, and A3. Water levels near the Delaware River are about 10 ft higher in A1.
6				Same as above except a 50-percent decrease.	Cone of depression is about 10 ft deeper and heads near the Delaware River are up to 20 ft lower in A1, A2, and A3. Water levels in Salem County are about 10 ft lower.
7	C1, C2, and C3	2–4	18–25	An order-of-magnitude increase in leakance in the confining units above the Potomac-Raritan-Magothy aquifers along their outcrops in Camden, Gloucester, and Burlington Counties.	Cone of depression is about 30 ft shallower in A1, A2, and A3. Water levels are 5 ft higher in downdip areas of A1, A2, and A3.
8				Same as above, but an order-of-magnitude decrease.	Cone of depression is about 80 ft deeper in A1, 70 ft deeper in A2, and 40 ft deeper in A3. Water levels near the Delaware River are up to 100 ft lower in A1 and 70 ft lower in A2 and A3. Water levels declined 15–20 ft in downdip areas, 10 ft in western Salem County, and about 5 ft or less near Raritan Bay. Local 20-ft water-level declines in A4.
9	A8	16–27[1]	2–50[1]	100-percent increase in transmissivity in the confined Kirkwood aquifer in Atlantic, Cape May, and Ocean Counties.	Cone of depression is about 30 ft shallower near the center in A8. Water levels near the offshore boundary are 5 ft higher in A8 and 10 ft higher in A7.
10				Same as above, but a 50-percent decrease.	Cone of depression is about 50 ft deeper near the center in A8. Water levels near the offshore boundary are about 10 ft lower in A7.
11	C8	16–27[1]	2–50[1]	An order-of-magnitude increase in leakance in the confining unit overlying the Rio-Grande water-bearing zone in Atlantic, Cape May, and Ocean Counties.	Cone of depression is about 35 ft shallower near the center in A8. Water levels near the offshore boundary are 10 ft higher in A7 and A8.
12				Same as above, but an order-of-magnitude decrease.	Cone of depression is about 20 ft deeper near the center in A8. Water levels near the offshore boundary are 20 ft lower in A8 and 10 ft lower in A7.
13	A8	16–19[2]	11–36[2]	100-percent increase in transmissivity in A8 along the updip limit of the confined Kirkwood aquifer.	Cone of depression is about 10 ft shallower near the center in A8 and water levels near the offshore boundary in A7 and A8 are 5 ft higher.
14				Same as above, but a 67-percent decrease.	Cone of depression is about 15 ft deeper near the center in A8 and water levels near the offshore boundary in A7 and A8 are 5 ft lower.

[1] Only those areas where A8 is confined.
[2] Only those areas near the updip limit of unit A8.

AQUIFER STORAGE

Simulated water levels were tested for sensitivity to changes in the storage coefficient of the confined aquifers. Decreasing the storage coefficient two orders of magnitude everywhere in all confined aquifers changed the 1978 simulated water levels generally less than 5 feet. Increasing the storage coefficient two orders of magnitude for all confined aquifers significantly changed simulated water levels in the large cones of depression in the Englishtown and Wenonah-Mount Laurel aquifers (A4 and A5) in Ocean and Monmouth Counties. Simulated water levels in 1978 were about 50 feet higher in this area for each order of magnitude increase in the storage coefficient. Simulated water levels increased about 10 feet for every order of magnitude increase in storage in other major cones of depression, including cones in the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) near Camden, in the confined Kirkwood aquifer (A8) near Atlantic City, and in the middle aquifer of Potomac-Raritan-Magothy aquifer system (A2) in Middlesex County.

Most of the aquifer-test results shown in table 3 are within an order of magnitude of the storage coefficient used in the calibrated model. The sensitivity simulations indicate that the storage coefficients used in the calibrated model are reasonable, because simulated water levels are not sensitive to decreases in the storage coefficient and increases of one and two orders of magnitude give storage coefficients that are close to storage coefficients of unconfined aquifers.

DOWNDIP BOUNDARIES

Simulated water levels were tested for sensitivity to the position of the southeast downdip boundaries of several aquifers. In the Potomac-Raritan-Magothy aquifers (A1, A2, And A3), the downdip boundary is the idealized interface between the freshwater and saltwater flow systems at the estimated occurrence of 10,000 mg/L chloride concentrations. Several simulations were made with the interface moved to various distances further offshore. The interface was not moved to a closer onshore position because there is no evidence to indicate that 10,000 mg/L chloride concentrations exist further northwest than initially estimated. Simulated water levels were not sensitive to moving the interface boundary seaward. In the sensitivity simulation with the boundary moved as far offshore as possible in the model, the interface was about 50 miles seaward from the boundary used in the calibrated model for the middle and upper Potomac-Raritan-Magothy aquifer system (A2 and A3) and 75 to 100 miles seaward from the boundary in the calibrated model in the lower Potomac-Raritan-Magothy aquifer (A1). In this simulation, simulated 1978 water levels changed less than 5 feet in all aquifers.

Simulated water levels were also tested for sensitivity to the estimated downdip extent of several aquifers that thin or become silty. The downdip extent of the Piney Point (A7) and the Wenonah-Mount Laurel (A5) aquifers was extended about 50 miles farther downdip, and simulated water levels changed less than 5 feet. Water levels in the Englishtown aquifer (A4) are most sensitive to the position of the aquifer's downdip extent. Extending the aquifer's downdip boundary 40 to 75 miles from that used in the calibrated model changed simulated water levels 30 to 50 feet near the center of the large cone of depression in Monmouth and Ocean Counties. Simulated water levels northwest of the cone changed less than 5 feet, offshore water levels and water levels to the south changed about 10 feet. Extending the aquifer's downdip extent 10 to 50 miles changed simulated water levels 10 to 20 feet near the center of the cone of depression. During calibration, water levels near the center of the cone of depression were also sensitive to the estimated limit of the upper sand unit of the Englishtown aquifer system. Although model unit A4 represents both the upper and lower sand units, the upper sand unit is not present downdip and offshore of Ocean and southern Monmouth Counties, and low transmissivity (less than 500 ft^2/d) in this area represents only the lower aquifer (fig. 58). In the calibrated model, the limit of the upper aquifer and the low transmissivities are estimated to be about 5 miles offshore. Moving the limit 3 to 8 miles farther offshore changed simulated water levels in the cone of depression several tens of feet.

These simulations suggest that the estimated position of the freshwater-saltwater interface boundary does not significantly affect simulated water levels and does not seriously limit model results. The estimated downdip extent of the Piney Point (A7) and Wenonah-Mount Laurel (A5) aquifers also does not significantly affect simulated water levels. However, water levels in the Englishtown aquifer (A4) are sensitive to the estimated downdip limit of the aquifer and the downdip limit of upper sand unit offshore of Ocean and Monmouth Counties. Therefore, the accuracy of model results near the cone of depression in Monmouth and Ocean Counties depends on the accuracy of the location of the downdip limit of the Englishtown aquifer.

BOUNDARY FLOWS

Simulated water levels were tested for sensitivity to the amount of flow used to simulate lateral boundary conditions for the aquifers along the southwest, southeast, and northeast boundaries. Total boundary flows used in the calibrated model for each pumping period are

TABLE 10.—*Boundary flows for prepumping conditions and for January 1, 1978, for each aquifer*

Model unit: A1, lower Potomac-Raritan-Magothy aquifer
A2, middle Potomac-Raritan-Magothy aquifer
A3, upper Potomac-Raritan-Magothy aquifer
A4, Englishtown aquifer
A5, Wenonah-Mount Laurel aquifer
A6, Vincentown aquifer
A7, Piney Point aquifer
A8, lower Kirkwood-Cohansey and confined Kirkwood aquifer
A9, upper Kirkwood-Cohansey aquifer
A10, Holly Beach

Boundary flows: Positive flows are into model area, negative flows are out of model area.

Model unit	Boundary flows (million gallons per day)					
	Prepumping			January 1, 1978		
	South-western boundary	South-eastern boundary	North-eastern boundary	South-western boundary	South-eastern boundary	North-eastern boundary
A1	0.850	—	−0.106	0.351	—	0.060
A2	.426	—	−1.431	1.477	—	.947
A3	.106	—	−3.379	−.040	—	1.613
A4	.048	—	−.409	.013	—	.276
A5	.118	—	−.067	.035	—	.039
A6	.153	—	−.051	.072	—	.016
A7	.035	—	—	−1.202	—	—
A8	.093	−0.065	−.246	.065	0.371	−.134
A9	−.811	.101	−.093	1.112	.244	−.095
A10	—	—	—	—	—	—
Total	1.018	.036	−5.782	1.883	.615	2.722

given in table 7. Boundary flows for each boundary in the prepumping simulation and in pumping period 8 (1973–78) of the transient simulation are shown in table 10. Three sensitivity simulations were made with (1) no boundary flows, (2) 2 times the boundary flows used in the calibrated model, and (3) 10 times the boundary flows used in the calibrated model. The results of the sensitivity simulations are shown in table 11.

Interpretation of these sensitivity simulations is limited by the relation of the boundary flows to the aquifer and confining-unit hydraulic characteristics near the boundaries. The amount of lateral flow at a boundary is calculated from the transmissivity and hydraulic gradients. The hydraulic gradients, however, are dependent on the confining-unit properties, which control vertical leakage, and the transmissivity. Changing only the flow at the boundaries, as was done in the sensitivity simulations, makes the boundary flows incompatible with transmissivity and vertical leakance at the boundaries. That is, lateral boundary flows that differ from those used in the calibrated model are not probable without also having different aquifer and confining-unit characteristics near the boundaries. The simulations described in this section do not show the sensitivity of simulated water levels to conditions at the flow boundaries. However, the simulations show the general significance of boundary flows to water levels within the New Jersey Coastal Plain aquifers, assuming the transmissivity and vertical-leakance values are those of the calibrated model.

The results of the three sensitivity simulations indicate that simulated water levels for most of the modeled aquifers are not significantly affected by large changes in boundary flows. Also, the amount of lateral flow at the boundaries has a minimal effect on estimates of transmissivity and leakance derived from model calibration. Large water-levels changes are generally near the boundaries and are not significant 5 to 10 miles away from the boundaries. Simulated water levels are most sensitive to boundary flows along the southwest model boundary between Delaware and New Jersey, in the three Potomac-Raritan-Magothy aquifers (A1, A2, A3), the Piney Point aquifer (A7), and the two Kirkwood-Cohansey aquifers (A8 and A9). However, simulated water levels are affected very little, generally less than 10 feet, for changes in flow up to 100 percent along the northeast boundary.

As an estimate of actual flows at the boundaries, the flows used in the model may be in error by as much as 100 percent locally. However, boundary flows are probably not in error by this much everywhere along the boundaries. Boundary flows are relatively small compared to withdrawals (less than 3 percent) and generally have only a small (less than 10 feet) effect on simulated water levels. Therefore, the lateral boundary flows used in the calibrated model are reasonable regional estimates of the flow.

CONFINING-UNIT STORAGE

Three sensitivity simulations were made to determine the effects of confining-unit storage on simulated water levels. No confining-unit storage had been used in the calibrated model. Specific storages of 6×10^{-6}/ft, 6×10^{-5}/ft, and 6×10^{-4}/ft for the confining units cause extreme changes in 1978 simulated water levels. In the simulation with a confining-unit specific storage of 6×10^{-4}/ft, water levels are 150 to 200 feet higher at the large cone of depression in Monmouth and Ocean Counties in the Englishtown and Wenonah-Mount Laurel aquifers (A4 and A5). Water levels for this simulation were 20 to 80 feet higher near the large cone of depression in Camden County in the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3). In the simulation with a confining-unit specific storage of 6×10^{-5}/ft, water levels in the Englishtown and Wenonah-Mount Laurel aquifers are 75 to 130 feet higher in Monmouth and Ocean Counties. Water levels in the Potomac-Raritan-Magothy aquifers are 10 to 40 feet higher in Camden County. Model calibration would not be possible using these values of confining-unit specific storage, as model transmissivity and confining-unit leakance would have to be

TABLE 11.—*Results of boundary-flow sensitivity analysis*

Simulation 1: No boundary flows in any unit.
Simulation 2: Two times the boundary flows used in calibration for each unit.
Simulation 3: Ten times the boundary flows used in calibration for each unit.
Model unit: A1, lower Potomac-Raritan-Magothy aquifer
A2, middle Potomac-Raritan-Magothy aquifer
A3, upper Potomac-Raritan-Magothy aquifer
A4, Englishtown aquifer
A5, Wenonah-Mount Laurel aquifer
A6, Vincentown aquifer
A7, Piney Point aquifer
A8, lower Kirkwood-Cohansey and confined Kirkwood aquifer
A9, upper Kirkwood-Cohansey aquifer
A10, Holly Beach

Model unit	Simulation number	Resulting water-level changes for January 1, 1978: Southwest boundary	Southeast boundary	Northeast boundary	Other areas
A1	1	10–25 ft lower, 200 ft lower locally.	5 ft lower to 15 ft higher.	—	—
	2	5–10 ft higher	0–5 ft higher	—	—
	3	20–90 ft higher, 180 ft higher locally.	5 ft lower to 90 ft higher.	5 ft lower	Cone of depression in Camden County, 10 ft shallower.
A2	1	10–15 ft lower	0–15 ft higher	—	Cone of depression in Middlesex and Monmouth Counties, 5 ft shallower.
	2	5–15 ft higher	0–15 ft higher	—	Same area as above, 5 ft shallower.
	3	20–140 ft higher	15 ft lower to 140 ft higher.	5–15 ft higher	Same area as above, 5–10 ft shallower. Cone of depression in Camden County, 5 ft shallower.
A3	1	0–5 ft lower	—	0–5 ft higher	—
	2	10 ft higher	5 ft lower to 10 ft higher.	0–15 ft higher	—
	3	35–75 ft higher	15 ft lower to 35 ft higher.	15–35 ft higher	Cone of depression in Camden County, 5–10 ft shallower. Cone of depression in Monmouth County, 5–10 ft shallower.
A4	1	—	—	0–10 ft higher	—
	2	—	—	5–10 ft higher	—
	3	10 ft higher	10–35 ft higher	10–45 ft higher	—
A5	1	—	0–15 ft higher	10 ft higher	—
	2	5 ft higher	5–10 higher	—	—
	3	10–35 ft higher	5–25 ft higher	25–40 ft higher	—
A6	1	—	—	—	—
	2	—	—	—	—
	3	10 ft higher	—	—	—
A7	1	—	—	—	—
	2	—	—	—	—
	3	10–70 ft lower	5–10 ft higher	—	—
A8	1	10 ft lower	0–10 ft lower	—	—
	2	10 ft higher	5–10 ft higher	5 ft higher	—
	3	40–110 higher	30–40 ft higher	5–40 ft higher	Cone of depression in Atlantic County, 10 ft shallower.
A9	1	5–15 ft lower	5 ft lower	—	Cone of depression in Cape May County, 5 ft deeper.
	2	5–15 ft higher	5 ft higher	5 ft higher	Same area as above, 5 ft shallower.
	3	55–130 ft higher	10–55 ft higher	0–40 ft higher	Same area as above, 30 ft shallower.

unreasonably low. These sensitivity simulations show that confining-unit specific storages of 6.0×10^{-4}/ft and 6.0×10^{-5}/ft are probably higher than the specific storage of the New Jersey Coastal Plain confining units.

In the simulation with a confining-unit storage of 6×10^{-6}/ft, water-level changes are much less than in the other two simulations. The greatest increase in water levels occurs in the Englishtown and Wenonah-Mount Laurel aquifers (A4 and A5). In these aquifers, 1978 water levels near the center of the cone of depression in Monmouth and Ocean Counties are 25 to 50 feet higher. Water levels in downdip areas of these aquifers are 10 to 60 feet higher than those of the calibration simulation. In all other aquifers, water levels near the center of the cones of depression are less than 10 feet higher, and water levels in downdip areas are 5 to 25 feet higher.

The results of this simulation suggest that calibration of the New Jersey RASA model would be possible with a confining-unit storage value of 6×10^{-6}/ft or less and somewhat lower values of aquifer storage, transmissivity, or confining-unit leakance. The calibration simulation with no confining-unit storage simulates only steady leakage through the confining units. Although actual transient leakage from the confining units is not directly simulated in the calibrated model, the release of water from confining units is probably indirectly simulated in the release of water from aquifer storage. To better simulate the release of water from storage from aquifers and confining units, aquifer storage should be based on actual aquifer thickness and an estimate of aquifer specific storage rather than a constant storage coefficient as was used for most of the aquifers in the calibrated model.

Present data on confining-unit specific storage are very limited. An estimate of specific storage of 3×10^{-6} to 6×10^{-6}/ft for the confining sediments overlying the Piney Point aquifer in Delaware was derived from an aquifer-test analysis by Leahy (1976, p. 22). Estimates of land subsidence near Atlantic City in Atlantic County are about 0.06 in/yr based on subsidence measurements at a monitoring station by the U.S. Geological Survey (P.P. Leahy, oral commun., 1985). This subsidence is attributed primarily to inelastic compaction of the confining units in the upper part of the Coastal Plain resulting from declining water levels. Elastic specific storage estimated from subsidence data is about 1.3×10^{-6}/ft for the basal Kirkwood confining unit (C7) (P.P. Leahy, oral commun., 1985), and the relatively small amount of subsidence suggests that inelastic confining-unit specific storage within the Coastal Plain is also relatively low. However, subsidence data have been collected for a relatively short period of time, since 1980, and are only from sediments in the upper part of the Coastal Plain. Not enough data are available to justify including confining-unit specific storage in the calibration simulation of the RASA model.

ADDITIONAL EVALUATION OF FLOW AND BOUNDARY CONDITIONS

TRANSIENT-FLOW CONDITIONS

The rate at which water levels change in response to changes in withdrawals and the time it takes for steady-state conditions to be reached after a change in withdrawals are important aspects of the transient-flow conditions within the New Jersey Coastal Plain aquifers. Two simulations were made to analyze these aspects by simulating the effects of (1) continued withdrawals from 1981 to 2011 at the 1978–80 withdrawal rate and (2) no withdrawals after 1980. The second simulation is based on the additional assumption (to those discussed previously) that the aquifers are totally elastic. Both simulations had simulated January 1, 1981, water levels (pumping period 9) as initial conditions.

In the first simulation, with continued withdrawals, water levels declined very little, less than 5 feet, in the upper aquifers (the Vincentown to the upper Kirkwood-Cohansey aquifers, A6 to A9) and near the center of the major cones of depressions. However, water levels farther from the major cones declined as much as 25 feet. In the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3), the center of the large cone of depression in central Camden County declined less than 2 feet after 1980 with continued withdrawals at the 1978–80 rate. Similarly, water levels in the cone of depression in Middlesex and Monmouth Counties in the middle Potomac-Raritan-Magothy aquifer also declined less than 2 feet. Water levels in downdip areas of the three Potomac-Raritan-Magothy aquifers near the northeast model boundary declined between 18 and 25 feet by the end of the simulation. Water levels in downdip areas near the southwest boundary declined less than 5 feet. In the Englishtown and Wenonah-Mount Laurel aquifers (A4 and A5), water levels near the major cone of depression in Monmouth and Ocean Counties declined 4 feet by the end of the simulation, but water levels in downdip areas near the northeast model boundary declined 17 feet. Water levels in downdip areas near the southwest model boundary declined only 6 feet in these aquifers. Water levels in the two Kirkwood-Cohansey aquifers (A8 and A9) declined less than 2 feet everywhere.

In general, the simulation of continued withdrawal at the 1978–80 rate to 2011 showed that near the large cones of depression, the flow system is very close to steady-state conditions, and water levels in these areas would change less than 5 feet if pumpage rates did not change. However, in downdip areas farther from the major cones of depression, water levels are under transient conditions and water levels would decline as much 25 feet with no change in pumpage rates. Also, in areas near the major cones, if pumpage rates did not change, 50 percent of the water-level decline would occur in 6 years and 75 percent would occur in 12 years. In areas away from the major cones, 50 percent of the water-level decline would occur in 8 years and 75 percent of the cones would occur in 16 years. This simulation also showed that after 10 years, changes in water levels would be less than 1 ft/yr everywhere.

The second simulation simulated the effects of no withdrawals after 1980. Generally, water levels near the center of the major cones of depression react quickly to changes in withdrawals compared to areas farther from the cones. The Potomac-Raritan-Magothy aquifers (A1, A2, and A3) in central Camden County and the confined Kirkwood aquifer (A8) in Atlantic County had simulated drawdowns from prepumping conditions to 1981 of as much as 100 feet. With no withdrawals after 1980, simulated water levels recovered about 95 percent, or to within about 5 feet of prepumping water levels, in 2 years. In the Englishtown and Wenonah-Mount Laurel aquifers (A4 and A5) near the center of the large cone of depression in Monmouth and Ocean Counties, drawdowns were as much as 285 feet from prepumping conditions to 1981. With no pumpage after 1980, simulated water levels recovered about 95 percent, or to within 14 feet of prepumping water levels, in 6 years. In downdip areas near the southeast model boundary in the lower Coastal Plain aquifers (the lower Potomac-Raritan-Magothy to the Wenonah-Mount Laurel aquifers, A1 to A5), drawdowns from prepumping conditions to 1981 were from 40 to 55 feet. With no pumpage after 1980, water levels recovered 50 percent in 12 years and 75 percent in 16 years. Water levels in the upper Potomac-Raritan-Magothy, Englishtown, and Wenonah-Mount Laurel aquifers (A3, A4, and A5) continued to decline for 2 years after 1980. In downdip areas of these aquifers near the southwest boundary, drawdowns were about 25 feet from prepumping conditions to 1981. With no pumpage after 1980, water levels in the downdip areas of these aquifers recovered about 75 percent in 12 years.

In general, the simulation with no pumpage after 1980 showed that, assuming the aquifers are totally elastic, water levels near the major cones would recover quickly in the absence of pumpage—95 percent in 2 to 6 years. However, water levels farther from the cones would recover more slowly—75 percent in 12 to 16 years. Also, after about 26 years, water-level recovery would be less than 1 ft/yr everywhere.

FLOW NEAR THE DOWNDIP BOUNDARY

In the steady-state prepumping flow system, areas of upward discharge along the southeast boundary are simulated in the upper Potomac-Raritan-Magothy, Wenonah-Mount Laurel, Piney Point, confined Kirkwood, and upper Kirkwood-Cohansey aquifers (A3, A5, A7, A8, and A9), and in small areas of the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2) and the Englishtown aquifer (A4) (figs. 75–80). The lower and middle Potomac-Raritan-Magothy aquifers and the Englishtown aquifer have downward flow through overlying confining units along most of the southeastern interface boundary. In all of the above aquifers, flows along this boundary are less than 0.01 in/yr and are smaller than the error in estimating these flows using the calibrated model.

The downward direction of simulated flow to the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2) may indicate that the conceptual model of generally upward flow along the interface is too simplified. Prepumping leakage from the overlying upper Potomac-Raritan-Magothy aquifer (A3) flows updip within the middle and lower Potomac-Raritan-Magothy aquifers and is discharged along the outcrop area near the Delaware River. The amount of leakage, transmissivity of the lower and middle Potomac-Raritan-Magothy aquifers, and the vertical leakance of the overlying confining units caused the direction of flow in downdip areas to be lateral along the interface boundary and updip rather than upward through overlying confining units.

After pumping began, the areas of upward flow along the southeastern model boundary were affected by withdrawals from the aquifers. Changes in the direction of vertical flow along the southeastern boundary between prepumping and 1978 flow conditions occur in the three Potomac-Raritan-Magothy aquifers, the Wenonah-Mount Laurel aquifer, the confined Kirkwood aquifer, and the upper Kirkwood-Cohansey aquifer (A1, A2, A3, A5, A8, and A9). Flow along the southeastern boundary is downward in 1978 through the confining unit overlying the Rio Grande water-bearing zone and the estuarine clay confining unit (C8 and C9) overlying the Kirkwood-Cohansey aquifers, although it is upward through the basal Kirkwood confining unit (C7) overlying the Piney Point aquifer. Flow through the Navesink-Hornerstown confining unit (C5) overlying the Wenonah-Mount Laurel aquifer is downward along the southeastern boundary offshore of Ocean County (fig. 86). Flow through the three confining units overlying the Potomac-Raritan-Magothy aquifers (C1, C2, and C3) is downward along the southeastern boundary offshore of Ocean County (figs. 82–84).

The southeast boundaries of the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) represent an idealized freshwater-saltwater interface (figs. 3, 5, and 7). The southeast boundaries of the confined Kirkwood and upper Kirkwood-Cohansey aquifer (A8 and A9) are flow boundaries at the downdip limit of the model, which is several tens of miles updip from the assumed position of the freshwater-saltwater interface in these aquifers (figs. 17 and 19). Although the interface is assumed to be an immovable no-flow surface, it is actually a transition zone of chloride concentrations, the position of which is affected by flow in both the saltwater and freshwater flow systems. This transition zone is expected to move landward as withdrawals change the flow system. Simulated drawdowns along the interface boundary from prepumping conditions to 1978 are as follows: as much as 60 feet in the lower Potomac-Raritan-Magothy aquifer, 45 feet in the middle Potomac-Raritan-Magothy aquifer, and 35 feet in the upper Potomac-Raritan-Magothy aquifer. Simulated drawdowns from prepumping conditions to 1978 along the southeast flow boundary are 20 feet in the confined Kirkwood aquifer and 10 feet in the upper Kirkwood-Cohansey aquifer. Lateral hydraulic gradients in these downdip areas are greatest, 1 ft/mi, in the confined Kirkwood aquifer (A8).

Simulated hydraulic gradients within the aquifers near the interface boundary cannot be used to estimate movement of the interface, although simulated changes in hydraulic gradients indicate a potential for interface movement. Any change in water level at the boundary or change in gradients near the boundary creates the potential for interface movement. Even if the simulated flow direction within an aquifer near the interface is toward the interface, if water levels or lateral gradients near the interface have decreased from prepumping conditions, the interface is under the potential to move updip. By comparing transmissivity values, the amounts of drawdowns, and changes in hydraulic gradients in each aquifer, the lower Potomac-Raritan-Magothy aquifer (A1) and the confined Kirkwood aquifer (A8) have the greatest potential for landward movement of the interface.

SOURCE OF WATER TO WELLS

The source of water to wells is analyzed below by using ground-water budgets. The rates of horizontal and vertical flow are determined for four major cones of depression in seven aquifers and the relative amounts of water derived from aquifer storage, boundary flows, and flow from constant-head nodes are summarized for each pumping period and aquifer. The source of water to wells in the major cones is the difference between prepumping and January 1, 1978, flow rates into and out of the areas. The source of water to wells in these areas is summarized as decreases and increases in horizontal and vertical outflow and inflow. As in other analyses, the budget figures presented in this section are regional estimates and may be useful as initial estimates of the source of water to wells locally. However, the source of water to wells for a small area in the Coastal Plain may be significantly different than these regional estimates.

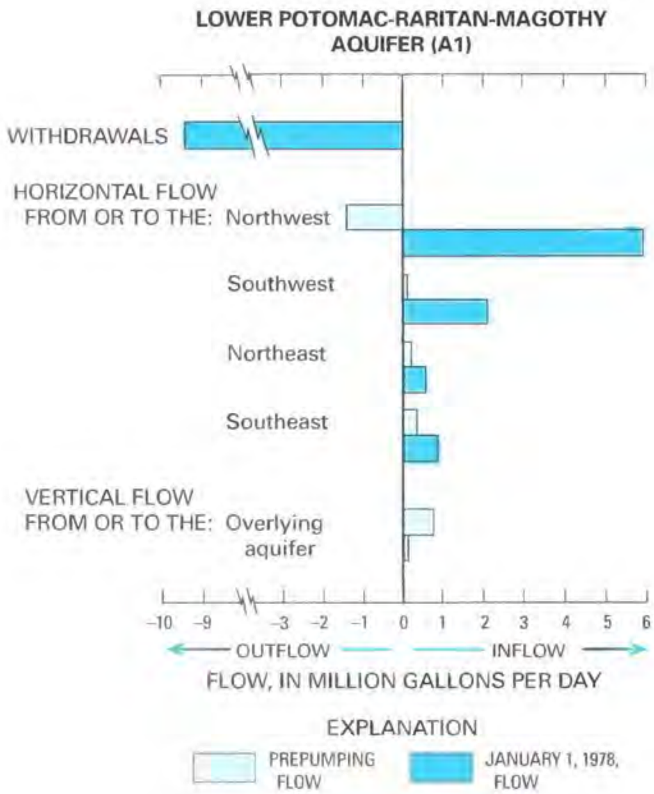
Flow near the centers of four major cones of depression in aquifers was quantified for prepumping and January 1, 1978, (pumping period 8) conditions. The cones of depression are in the Potomac-Raritan-Magothy aquifers (A1, A2, and A3) in central Camden County, the middle Potomac-Raritan-Magothy aquifer (A2) in Monmouth and Middlesex Counties, the Englishtown and Wenonah-Mount Laurel aquifers (A4 and A5) in Monmouth and Ocean Counties, and the confined Kirkwood aquifer (A8) in Atlantic County. The areas for which ground-water flow budgets were compiled are shown in figure 25. These areas include a large percentage of local withdrawals within relatively small areas as compared to the areas of the cones of depression. The areas were selected because the previously discussed analysis of transient conditions indicated that simulated water levels within these areas are near steady state.

The rate of water released from storage at a particular time is not easily estimated from simulation results. The rate of water being released from aquifer storage during the last time step of any pumping period is much smaller than the actual rate as illustrated by the very small simulated water-level changes during the last time step of many pumping periods (figs. 52–54). The average rate of water released from storage during an entire pumping period is a better estimate of the actual rate of water being released from storage at the end of a pumping period. However, the average rate for a pumping period increases the budget error between discharges and sources of water estimated at the end of a pumping period.

Ground-water flow budgets were calculated for areas near the center of the major cones of depression, which are close to steady-state conditions and where water being released from aquifer storage is negligible. The simulated release of water from aquifer storage at the end of pumping period 8, January 1, 1978, is considered insignificant and is not included in the following analysis of source of water to wells for these selected areas. Estimates of the actual rate of water released from storage at January 1, 1978, based on the rate of water released from storage during pumping period 8 (January 1, 1973, to January 1, 1978), are less than 2 percent of the discharges from these areas. The following analysis compares the relative significance of changes in the rates of horizontal and vertical flows as a source of water to wells in these areas. The importance of aquifer storage as a source of water to wells is discussed for each aquifer and pumping period later in this section.

The prepumping and January 1, 1978, horizontal and vertical flow rates are shown in figures 88 to 91 for each area. Under prepumping conditions, the major direction of flow into the budget areas for the Potomac-Raritan-Magothy aquifers (A1, A2, and A3) in central Camden County was from above (fig. 88). A major direction of outflow for these aquifers under prepumping conditions was updip toward the outcrop areas to the northwest. For the middle and upper Potomac-Raritan-Magothy aquifers in the central Camden County area, another major direction of outflow was downward. For the middle Potomac-Raritan-Magothy aquifer in Middlesex and Monmouth Counties (fig. 89), the major direction of inflow under prepumping conditions is from the southwest, and the major direction of outflow is updip (northwest) toward the outcrop. Unlike the Potomac-Raritan-Magothy aquifers, the major direction of prepumping flow into the Englishtown aquifer (A4), Wenonah-Mount Laurel aquifer (A5), and the confined Kirkwood aquifer (A8) is downdip from the northwest (figs. 90 and 91).

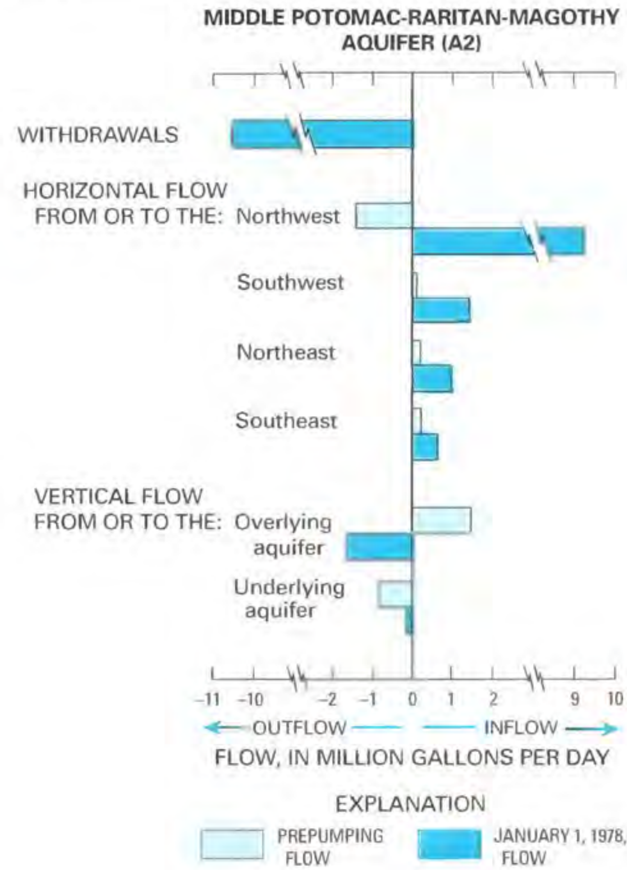
Flow rates for January 1, 1978, are much different from prepumping flow rates for each of the areas. The major outflows from the areas are withdrawals from wells, except in the Wenonah-Mount Laurel aquifer (A5) (fig. 90), in which vertical flow downward into the Englishtown aquifer (A4) is the major outflow from the budget area. The central Camden County area for the middle Potomac-Raritan-Magothy aquifer (A2) is the only other budget area that has flow out of the area in addition to withdrawals. In this area, there is also vertical flow to the overlying and underlying aquifers. Flow into each of the budget areas, except in the



MODEL UNIT: A1	*FLOW BUDGET ROWS: 5–11	COLUMNS: 18–26
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	8.80	86.6
Decreased horizontal outflow	1.35	13.3
TOTAL	10.15	99.9
DISCHARGES		
Withdrawals	9.44	
Increased vertical outflow	0.72	
TOTAL	10.16	

*Flow budget area is shown in figure 25.

FIGURE 88.—Simulated prepumping and January 1, 1978, flow rates and source of water to wells for the Potomac-Raritan-Magothy aquifers (model units A1, A2, and A3), central Camden County area.

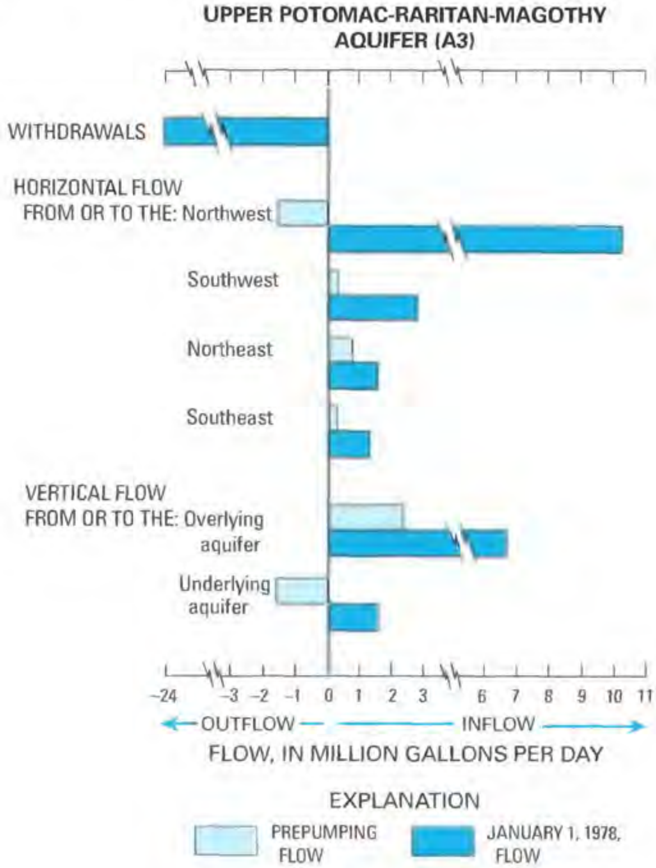


MODEL UNIT: A2	*FLOW BUDGET ROWS: 5–11	COLUMNS: 18–26
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	11.70	89.6
Decreased horizontal outflow	1.35	10.3
TOTAL	13.05	99.9
DISCHARGES		
Withdrawals	10.59	
Increased vertical outflow	0.88	
Decreased vertical inflow	1.60	
TOTAL	13.07	

*Flow budget area is shown in figure 25.

FIGURE 88.—Continued.

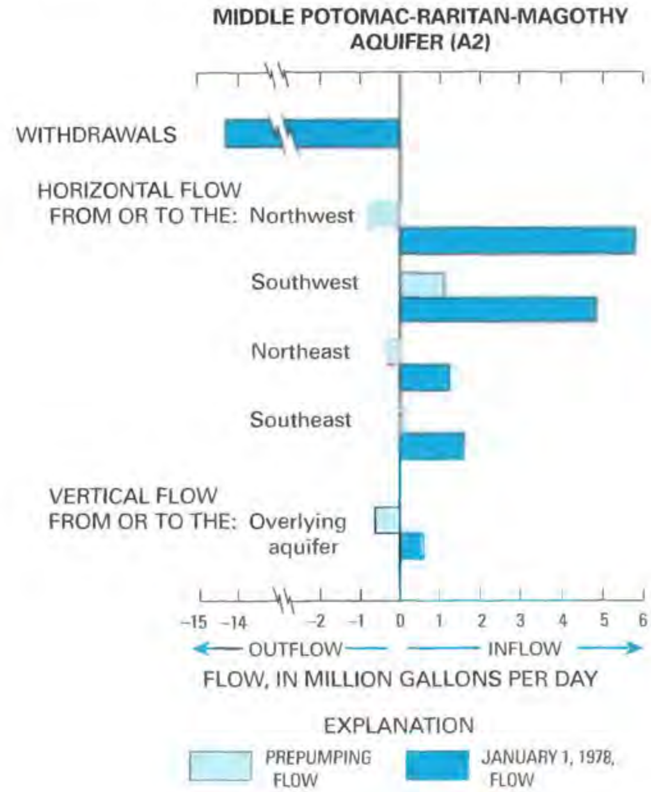
Englishtown aquifer, is greatest, 40 to 75 percent of the total flow, from the northwest (downdip from the outcrop areas). In the Englishtown aquifer in Monmouth and Ocean Counties, about 40 percent of the flow into the area is from the northwest, but about 47 percent of the flow enters the area from the overlying Wenonah-Mount Laurel aquifer. Other significant flows into the budget areas occur in the central Camden County area for the upper Potomac-Raritan-Magothy aquifer, where 27 per cent of the flow into the area is from the overlying



MODEL UNIT: A3	*FLOW BUDGET ROWS: 5–11	COLUMNS: 18–26
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	14.97	62.4
Decreased horizontal outflow	1.49	6.2
Increased vertical inflow	5.81	24.2
Decreased vertical outflow	1.60	6.7
TOTAL	23.87	99.5
DISCHARGES		
Withdrawals	24.00	
TOTAL	24.00	

*Flow budget area is shown in figure 25.

FIGURE 88.—Continued.



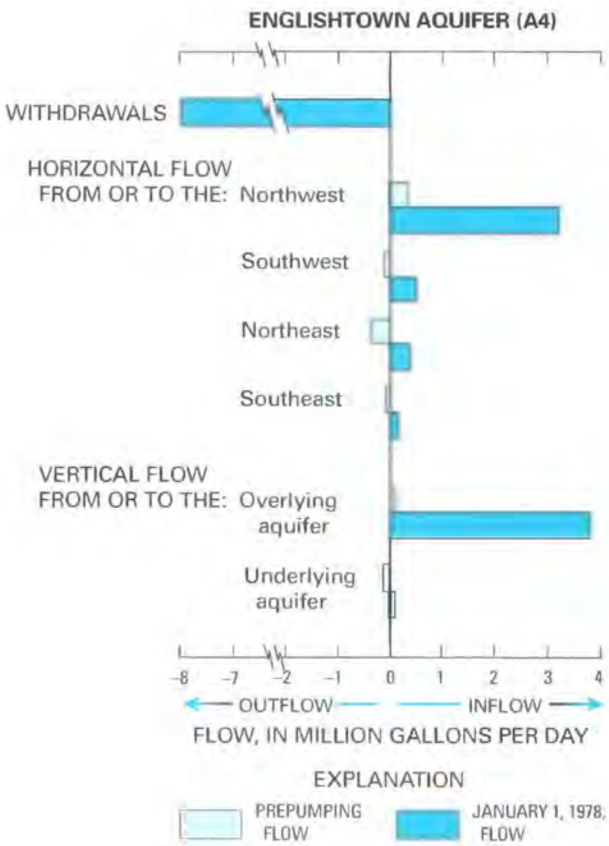
MODEL UNIT: A2	*FLOW BUDGET ROWS: 4–6	COLUMNS: 42–45
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	12.57	88.0
Decreased horizontal outflow	1.05	7.4
Increased vertical inflow	0.50	3.5
Decreased vertical outflow	0.06	0.4
TOTAL	14.18	99.3
DISCHARGES		
Withdrawals	14.28	
TOTAL	14.28	

*Flow budget area is shown in figure 25.

FIGURE 89.—Simulated prepumping and January 1, 1978, flow rates and source of water to wells for the middle Potomac-Raritan-Magothy aquifer (model unit A2), Middlesex and Monmouth Counties.

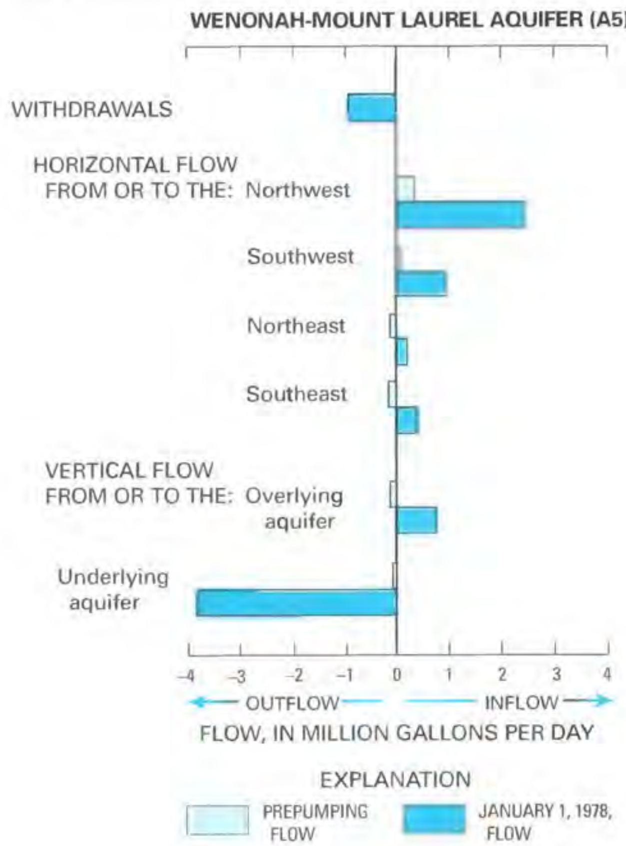
Englishtown aquifer, and in the Middlesex and Monmouth County area for the middle Potomac-Raritan-Magothy aquifer, where 35 percent of the flow into the area is from the southwest (southwestern Middlesex County).

The major source of water to wells, the difference between prepumping and January 1, 1978, inflows and outflows, is increased horizontal inflow into the budget areas (figs. 88–91). This flow is 79 to 90 percent of the total flow to wells in five of the seven areas. In the



MODEL UNIT: **A4**	*FLOW BUDGET ROWS: 12–17	COLUMNS: 37–46
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	3.80	47.7
Decreased horizontal outflow	0.34	4.2
Increased vertical inflow	3.80	47.7
Decreased vertical outflow	0.02	0.3
TOTAL	7.96	99.9
DISCHARGES		
Withdrawals	7.97	
TOTAL	7.97	

*Flow budget area is shown in figure 25.

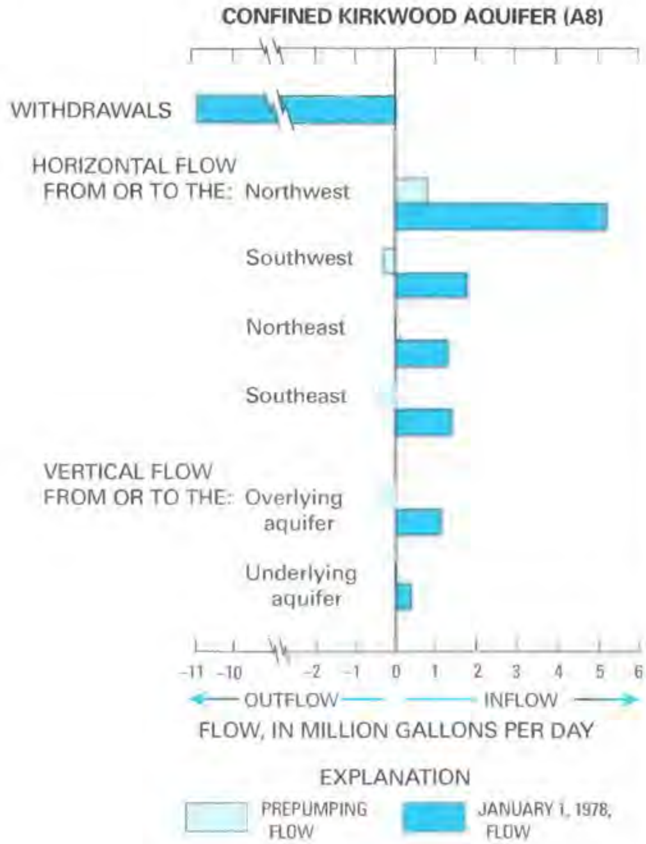


MODEL UNIT: **A5**	*FLOW BUDGET ROWS: 12–17	COLUMNS: 37–46
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	3.61	78.7
Decreased horizontal outflow	0.22	4.8
Increased vertical inflow	0.73	15.8
TOTAL	4.56	99.3
DISCHARGES		
Withdrawals	0.90	
Increased vertical outflow	3.69	
TOTAL	4.59	

*Flow budget area is shown in figure 25.

FIGURE 90.—Simulated prepumping and January 1, 1978, flow rates and source of water to wells for the Englishtown and Wenonah-Mount Laurel aquifers (model units A4 and A5), Ocean and Monmouth Counties.

central Camden County area for the upper Potomac-Raritan-Magothy aquifer (A3), increased horizontal inflow is 62 percent of the total flow to wells, and in the Monmouth-Ocean County area for the Englishtown aquifer (A4), this source is 48 percent of the total. The other major source of water to wells in these two areas is increased vertical inflow. Most of the increased horizontal inflow into each of the areas is from the northwest, water moving downdip from the outcrop areas. Decreases in horizontal and vertical outflows from the budget areas are about 0 to 13 percent of the total flow of water to wells. For the Potomac-Raritan-Magothy



MODEL UNIT: A8	*FLOW BUDGET ROWS: 20–24	COLUMNS: 17–25
SOURCES	MILLION GALLONS PER DAY	PERCENT OF DISCHARGES
Increased horizontal inflow	8.63	79.3
Decreased horizontal outflow	0.61	5.6
Increased vertical inflow	1.29	11.8
Decreased vertical outflow	0.35	3.2
TOTAL	10.88	99.9
DISCHARGES		
Withdrawals	10.89	
TOTAL	10.89	

*Flow budget area is shown in figure 25.

FIGURE 91.—Simulated prepumping and January 1, 1978, flow rates and source of water to wells for the confined Kirkwood aquifer (model unit A8), Atlantic County.

aquifers (A1, A2, and A3) in the central Camden County area, the decrease in horizontal outflow is the elimination of prepumping flow toward the outcrop areas. In the Middlesex and Monmouth County area for the middle Potomac-Raritan-Magothy aquifer (A2), the decrease in horizontal outflow is the elimination of prepumping flow toward the outcrop area and Raritan Bay. In the Englishtown (A4), Wenonah-Mount Laurel (A5), and the confined Kirkwood (A8) aquifers, the decrease in horizontal outflow from the budget areas constitutes the elimination of prepumping flow toward offshore discharge areas to the northeast, southeast, and southwest.

The total amount of water released from storage within all the aquifers was analyzed for each pumping period, and the amount of water released from storage within each aquifer was analyzed for pumping period 8 (1973 to 1978) and for prepumping conditions (1896) through pumping period 8 (1978). Table 12 shows the simulated source of water to wells for each pumping period and includes the rate of water released from storage and the percentage of withdrawals that comes from water released from storage for each pumping period. The rate of water released from storage near the end of the simulation (pumping periods 7 to 9, 1968 to 1980) is about 10 Mgal/d or 3 percent of the withdrawals during that time. The highest rates of water released from storage, 13 Mgal/d, were during the late 1960's (pumping period 6, 1965 to 1968) or about 5 percent of the withdrawals during that period. The simulated rates of water released from aquifer storage for each aquifer are shown in table 13. The rates for 1896 to 1978 (pumping periods 1 through 7) and 1973 to 1978 (pumping period 8) show about the same relative rates among aquifers. The highest rates of water released from storage ranged from about 1.4 to 3.0 Mgal/d and are from the Potomac-Raritan-Magothy aquifers (A1, A2, and A3) and the upper Kirkwood-Cohansey aquifer (A9). About 80 percent of the water being released from storage at any time is from these aquifers.

The amount of withdrawals derived from changes in boundary flows is similar to the amount derived from storage and is also shown on table 12. The amount of water flowing to wells is the difference between the boundary flows for any pumping period and prepumping conditions (table 10). Near the end of the simulation (during pumping periods 8 and 9), changes in boundary flows provided about 10 Mgal/d to wells or about 3 percent of the withdrawals. Of this 10 Mgal/d, about 8.5 Mgal/d is from change in flow rates along the northeast boundary, about 0.9 Mgal/d is from the southwest boundary, and about 0.6 Mgal/d is from the southeast boundary (table 10). The major source of water from boundary flows is from increased flow into the model area along the northeast boundary in the middle and upper Potomac-Raritan-Magothy aquifers (A2 and A3) and the southwest boundary in the middle Potomac-Raritan-Magothy aquifer (A2) and the upper Kirkwood-Cohansey aquifer (A9) (table 10). Not all changes in boundary flows were increased inflow over the period

TABLE 12.—*Simulated source of water to wells for each pumping period*

Pumping period	End date	Withdrawals	Amount of withdrawals from: Storage		Boundary flows		Ocean and bays		Streams		Error (Mgal/d)[1]
			Mgal/d	Percent	Mgal/d	Percent	Mgal/d	Percent	Mgal/d	Percent	
1	1–1–1921	26.4	0.8	3.0	0.0	0.0	1.1	4.2	24.6	93.2	0.1
2	1–1–1946	86.0	1.6	1.9	.3	.3	3.1	3.6	81.1	94.2	.1
3	1–1–1953	150.8	5.8	3.8	.9	.6	5.0	3.3	139.1	92.2	.0
4	1–1–1958	174.8	5.2	3.0	2.5	1.4	6.4	3.7	160.7	91.9	.0
5	1–1–1965	234.9	9.0	3.8	4.0	1.7	9.2	3.9	212.8	90.6	.1
6	1–1–1968	274.0	13.3	4.9	6.0	2.2	10.3	3.8	244.4	89.2	.0
7	1–1–1973	325.6	11.3	3.5	7.9	2.4	13.0	4.0	293.5	90.1	.1
8	1–1–1978	350.9	9.7	2.8	9.9	2.8	14.7	4.2	316.6	90.2	.0
9	1–1–1981	358.2	10.1	2.8	10.3	2.9	15.0	4.2	322.9	90.1	.1

[1] Mgal/d: million gallons per day.

TABLE 13.—*Simulated rate of water released from aquifer storage for each aquifer*

[Pumping periods 1 through 8 are from January 1, 1896, to January 1, 1978. Pumping period 8 is from January 1, 1973, to January 1, 1978. Mgal/d: million gallons per day.]

Model unit	Aquifer	Water from storage (Mgal/d): Pumping periods 1 through 8	Water from storage (Mgal/d): Pumping period 8	Withdrawals pumping period 8 (Mgal/d)	Percent of withdrawals from storage pumping period 8
A1	Lower Potomac-Raritan-Magothy	0.36	1.39	74.40	1.87
A2	Middle Potomac-Raritan-Magothy	.89	2.93	92.31	3.17
A3	Upper Potomac-Raritan-Magothy	1.08	1.73	84.11	2.06
A4	Englishtown	.23	.49	11.86	4.13
A5	Wenonah-Mount Laurel	.30	.69	4.47	15.44
A6	Vincentown	.07	.23	0.95	24.21
A7	Piney Point	.08	.18	1.97	9.14
A8	Lower Kirkwood Cohansey and confined Kirkwood	.20	.49	36.91	1.33
A9	Upper Kirkwood-Cohansey	.81	1.50	43.82	3.42
A10	Holly Beach	.00	.00	.01	.00
	Totals	4.02	[1]9.63	[1]350.81	[2]2.75

[1] Totals differ slightly from values in table 13 because of rounding errors in different methods of calculation.
[2] Percent of total withdrawals from storage in all aquifers for pumping period 8 (not a column total but calculated on the basis of total water from storage and total withdrawals for pumping period 8).

of simulation. Seven aquifers (the lower and upper Potomac-Raritan-Magothy, Englishtown, Wenonah-Mount Laurel, Vincentown, and the lower Kirkwood-Cohansey and confined Kirkwood aquifers) have decreased inflow or increased outflow along the southwest boundary. The major flow out of the model area in 1978 (pumping period 8) is 1.2 Mgal/d from the Piney Point aquifer (A7) toward Delaware.

The source of water to wells was also analyzed to determine the relative amounts of flow from local and regional surface-water bodies. The distinction between local and regional was based on the size of the surface-water body relative to the cell size. Streams including the Delaware River, which are less than a cell in width, are considered to be local surface-water bodies. The Atlantic Ocean, Delaware Bay, and Raritan Bay are regional surface-water bodies. The areas of local and regional surface-water bodies are shown in figure 25.

The amount of flow to streams and rivers relative to the ocean and bays was determined by totaling the rate of flow for constant-head nodes representing streams and totaling those representing the ocean and bays for each pumping period (table 14). The simulated amount of flow to streams at any time is dependent on the amount of recharge used in the model and is not directly comparable to the amount of flow to constant-head nodes in the bays and ocean or to withdrawals. However, the change in flow to rivers and streams from prepumping conditions to any pumping period is not dependent on recharge but is determined by withdrawals, as is the change in flow to the ocean and bays. Therefore, the amount of withdrawals from changes in flow to streams relative to changes in flow to the ocean and bays was determined by comparing (1) the amount of withdrawals to (2) changes from prepumping conditions in the flow rates to constant-head nodes representing each source (table 12).

TABLE 14.—*Simulated flow to and from constant-head nodes for each pumping period*

[Flow to and from constant-head nodes: Positive flows are into constant-head nodes, negative flows are out of constant-head nodes. Difference between flows for any pumping period and prepumping flows is the source of water to wells in table 13.
Streams: Flow to stream constant-head nodes is directly dependent on simulated recharge to the water-table aquifer, 20 inches per year (4507 million gallons per day).]

Pumping period	End date	Flow to and from constant-head nodes (million gallons per day)	
		Ocean and bays	Streams
prepumping	1-1-1896	12.7	4489.8
1	1-1-1921	11.6	4465.2
2	1-1-1946	9.6	4408.7
3	1-1-1953	7.7	4350.7
4	1-1-1958	6.3	4329.1
5	1-1-1965	3.5	4277.0
6	1-1-1968	2.4	4245.4
7	1-1-1973	−.3	4196.3
8	1-1-1978	−2.0	4173.2
9	1-1-1981	−2.3	4166.9

In general, throughout the simulation, the rates of flow to wells from the ocean and bays are slightly more than flow to wells from storage or lateral boundary flows. Flow from the ocean and bays is about 14 Mgal/d near the end of the simulation, or about 4 percent of withdrawals. The major source of water to wells is from decreased streamflow. Streamflow provides about 320 Mgal/d or 90 percent of withdrawals. Decreased streamflow is equivalent to increased deep percolation to the confined aquifers. Based on data in table 14 and on the area of local surface-water bodies on figure 25, deep percolation increased from 0.1 in/yr during prepumping conditions to about 1.5 in/yr near the end of the simulation (pumping periods 8 and 9, 1978 to 1981).

SUMMARY AND CONCLUSIONS

Flow in the interbedded unconsolidated sand, silt, and clay of the New Jersey Coastal Plain was simulated with a digital model representing an alternating sequence of 10 aquifers and 9 intervening confining units. These units generally thicken downdip from the Fall Line toward the Atlantic Ocean. Near the Fall Line, aquifers and confining units are less than 50 feet thick. In downdip freshwater areas, maximum aquifer thicknesses range from about 100 to 600 feet, and maximum confining-unit thicknesses range from about 50 to 1,000 feet.

The prepumping ground-water flow system was recharged in upland areas of high altitudes in Mercer, Middlesex, and western Monmouth Counties; in western Ocean and central Burlington Counties; and in central Gloucester and Camden Counties and was discharged to the Atlantic Ocean, Delaware River, Delaware Bay, Raritan Bay, and large rivers in the Coastal Plain. A large part of ground-water recharge was discharged to nearby surface-water bodies, and only a small amount recharged the confined aquifers. Recharge to the confined aquifers during prepumping conditions flowed vertically downward through confining units and laterally downdip in aquifers, then upward through confining units or sometimes laterally updip in aquifers to be discharged to the regional surface-water bodies.

Under pumping conditions, regional cones of depression formed in the three Potomac-Raritan-Magothy aquifers, Englishtown aquifer, Wenonah-Mount Laurel aquifer, and the confined Kirkwood aquifer. In 1980, withdrawals were greater than 350 Mgal/d. The sources of water to wells include water released from aquifer storage and decreased discharge to large surface-water bodies but mostly water from reduction in streamflow.

Water levels in the 10 aquifers were simulated by using a multilayer finite-difference model for prepumping steady-state conditions and transient conditions from 1896 to 1980. The model included 10 aquifer layers, a grid with 29 rows and 51 columns, with cell sizes ranging from 6.25 to 47.5 mi^2. Calibration was achieved primarily by trial-and-error adjustment of initial estimates of transmissivities and vertical leakance of the confining units. Minor adjustments were made to water-table and stream altitudes and the confined aquifer storage coefficient. After calibration, the model was tested for sensitivity to several aquifer and confining-unit properties and several boundary conditions.

The lower model boundary and the northwestern aquifer limits are no-flow boundaries. The other lateral boundaries are flow or no-flow boundaries depending on whether an aquifer extends beyond the modeled area. Downdip no-flow boundaries represent either the limits of the aquifers or a freshwater-saltwater interface at the estimated location of 10,000 mg/L chloride concentrations within the aquifers. Flows for the lateral flow boundaries were calculated from simulated water-level and transmissivity data from the regional RASA model of the northern Atlantic Coastal Plain.

Flow was simulated in the 10 aquifers by using a 10-layer model with a constant-head boundary representing water-table altitudes and an 11-layer model with recharge of 20 in/yr to the water table and with a constant-head boundary representing altitudes of long-term stream stages. Simulations with the 10-layer model provided estimates of deep percolation to the confined aquifers, which were used to calculate streambed vertical leakance between the water table and stream nodes. The upper 11-layer model boundary allows simulated water levels in the water-table aquifer to change in response to withdrawals but provides an infinite source of water from overlying streams.

Model calibration is partly based on a comparison of simulated to interpreted potentiometric surfaces and of simulated to actual well hydrographs. Differences between simulated and measured water levels are usually less than 15 feet. However, in some areas aquifers are not considered calibrated. Generally, these are offshore areas, where there are no measured water levels for calibration, and in outcrop areas where the cell size is large compared to outcrop widths. Downdip areas of the three Potomac-Raritan-Magothy aquifers (A1, A2, and A3) in Monmouth and Ocean Counties and the Piney Point aquifer (A7) in Cumberland County are not well calibrated, as simulated water levels are more than 15 feet higher than measured water levels. Although areas near the major cones of depression are considered calibrated, simulated water levels near the center of the cones are 10 to 40 feet higher than measured water levels because of the large cell size.

Model calibration is also based on whether simulated flow to and from constant-head nodes is reasonable in relation to recharge and discharge areas postulated in the conceptual model. Areas of simulated recharge to the confined aquifers coincide with areas of high land-surface altitudes, and discharge areas coincide with low altitude areas along large surface-water bodies. Highest simulated flow rates to and from the confined aquifers and between the water table and streams are in areas where there is no confining unit underlying the aquifers' outcrops. In areas where confining units underlie the unconfined outcrops, simulated recharge and discharge rates to the confined aquifers are generally less than 1.5 in/yr, and recharge and discharge to streams is generally within 3.5 in/yr of the assumed recharge rate of 20 in/yr. Under pumping conditions, simulated flow to streams decreased less than 5 in/yr in most areas. Simulated deep percolation to the confined aquifers from the unconfined outcrop areas is 0.1 in/yr under prepumping conditions and 1.5 in/yr under pumping conditions.

The calibrated model provided regional estimates of aquifer transmissivity and confining-unit vertical leakance as shown in table 8. The highest transmissivity, greater than 10,000 ft^2/d, is in Camden and Gloucester Counties in the Potomac-Raritan-Magothy aquifers (A1, A2, and A3); in Monmouth and Ocean Counties in the middle Potomac-Raritan-Magothy aquifer (A2); in Atlantic and Cape May Counties in the confined Kirkwood aquifer (A8); and in Ocean, Burlington, Atlantic, and Cape May Counties in the lower and upper Kirkwood-Cohansey aquifers (A8 and A9). Leakance is highest, greater than 1×10^{-3} (ft/d)/ft, in updip areas near the outcrops and lowest, less than 1×10^{-5} (ft/d)/ft, in downdip areas. The model was calibrated by using a specific yield of 0.15 in unconfined areas. A storage coefficient of 1×10^{-4} was used for all the confined aquifers except in downdip areas of Monmouth and Ocean Counties for the Potomac-Raritan-Magothy aquifers (A1, A2, and A3) where storage coefficients of 5×10^{-4} to 8×10^{-4} were used.

In the simulated prepumping flow system, vertical hydraulic gradients between aquifers are highly variable but tend to be greatest in updip areas. Simulated prepumping vertical flow rates are also highest, up to about 4 in/yr, in updip areas compared to rates of less than 0.2 in/yr in downdip areas. In the simulated 1978 transient flow system, steepest vertical and lateral hydraulic gradients are near the major cones of depression. Lateral gradients are steepest on the updip and onshore sides of the cones. Flow rates between aquifers in 1978 are less than 0.2 in/yr in many areas, but rates from about 1 to 5 in/yr are common locally in updip areas and near large cones of depression.

Sensitivity simulations show that in the major cones of depression tested, simulated water levels are fairly sensitive to changes in transmissivity and confining-unit leakance. Sensitivity simulations also show that the simulated water levels near the center of the cone in Camden County are most sensitive to confining-unit vertical leakance near the outcrop areas of the Potomac-Raritan-Magothy aquifers (A1, A2, and A3). Simulated water levels in the cone of depression near Atlantic City are sensitive to changes in transmissivity in the confined Kirkwood aquifer (A8) along the updip extent of the overlying confining unit, where the aquifer becomes unconfined. Sensitivity analyses also indicate that the aquifer storage coefficients used in the calibrated model were reasonable.

Sensitivity analysis also indicated that simulated water levels in most areas with measured water-level data were not very sensitive to the position of the aquifers' no-flow boundaries or to the amount of flow at the flow boundaries. The location of the downdip aquifer limits, the location of the freshwater-saltwater interface, and the amount of flow at the lateral flow boundaries are reasonable approximations of the physical system and have minimal effects on estimates of transmissivity and vertical leakance. However, in Monmouth and Ocean Counties in the Englishtown aquifer (A4), water levels are sensitive to the position of the downdip limit of the Englishtown aquifer and to the position of the offshore limit of the upper sand unit within the aquifer. The location of these limits should be better defined for more accurate simulation in these areas.

Sensitivity analyses of confining-unit storage show that a preliminary estimate of 6×10^{-6} per foot for specific storage is probably reasonable. However, aquifer storage should be based on aquifer-specific storage and thickness for better simulation of the flow system.

Analysis of flow near the downdip aquifer boundaries indicates some vertically downward flow under prepumping conditions in the lower and middle Potomac-Raritan-Magothy aquifers (A1 and A2) and in the Englishtown aquifer (A4). Although simulated water levels within the aquifers near the freshwater-saltwater interface boundary cannot be used to calculate movement of the interface, the lower Potomac-Raritan-Magothy aquifer (A1) and the confined Kirkwood aquifer (A8) have the greatest potential for inland migration of saltwater.

Analysis of transient conditions showed that the flow system is very near steady-state conditions near the center of the major cones of depression. Water levels in these areas would change less than 5 feet if withdrawal rates remained at the 1978–80 rates. However, in downdip areas farther from the major cones of depression, the flow system is more transient, and water levels would decline as much as 25 feet if there were no change in withdrawal rates after 1980. If there were no withdrawals after 1980, water levels near the major cones would recover quickly—95 percent in 2 to 6 years—but water levels farther from the cones would recover more slowly—75 percent in 12 to 16 years.

The major source of water to wells near the center of the major cones of depression in 1978 is increased lateral flow from the outcrop areas. In the upper Potomac-Raritan-Magothy aquifer (A3) and the Englishtown aquifer (A4), another major source of water is increased vertical flow through the overlying confining units. About 80 percent of the water released from storage is from the Potomac-Raritan-Magothy aquifers (A1, A2, and A3) and the upper Kirkwood-Cohansey aquifer (A9). On a regional scale, the simulated sources of water to wells in the late 1970's include (1) 3 percent from aquifer storage, (2) 3 percent from boundary flows, (3) 4 percent from the ocean and bays, and (4) 90 percent from decreased discharge to or increased recharge from streams.

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TABLE 15

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*

[Simulated water level: In feet above or below sea level. Estimated from simulated water levels for the three nodes nearest to actual well location. Measured water level: In feet above or below sea level. Date of measured water level is for measurement nearest to end date in Zapecza, Voronin, and Martin (1987, pls. 1–10). Water-level difference: Simulated water level minus measured water level, in feet. Well locations shown in figures 42 to 51. —, data unavailable.]

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
LOWER POTOMAC-RARITAN-MAGOTHY AQUIFER (A1)				
WELL 5–262	NODE 7, 26			
1	1/1/1921	17.89	—	—
2	1/1/1946	12.14	—	—
3	1/1/1953	5.20	—	—
4	1/1/1958	1.04	—	—
5	1/1/1965	−12.47	—	—
6	1/1/1968	−21.40	—	—
7	1/1/1973	−33.02	−34.93	1.91
8	1/1/1978	−45.51	−45.48	−0.03
9	1/1/1981	−48.49	−48.92	−0.43
WELL 5–645	NODE 5, 27			
1	1/1/1921	16.45	—	—
2	1/1/1946	12.94	—	—
3	1/1/1953	8.64	—	—
4	1/1/1958	5.69	—	—
5	1/1/1965	−7.51	—	—
6	1/1/1968	−18.38	−15.35	−3.03
7	1/1/1973	−30.28	−24.04	−6.24
8	1/1/1978	−40.26	−29.39	−10.87
9	1/1/1981	−37.14	−33.36	−3.78
WELL 5–683	NODE 14, 30			
1	1/1/1921	28.76	—	—
2	1/1/1946	23.94	—	—
3	1/1/1953	18.46	—	—
4	1/1/1958	15.66	—	—
5	1/1/1965	2.16	−3.59	5.75
6	1/1/1968	−4.58	−8.06	3.47
7	1/1/1973	−14.87	−16.36	1.49
8	1/1/1978	−28.14	−29.44	1.29
9	1/1/1981	−36.60	−31.94	−4.66
WELL 7–283	NODE 5, 22			
1	1/1/1921	9.35	—	—
2	1/1/1946	−1.71	—	—
3	1/1/1953	−16.39	—	—
4	1/1/1958	−24.94	—	—
5	1/1/1965	−41.82	−43.82	2.00
6	1/1/1968	−49.90	−51.49	1.59
7	1/1/1973	−59.58	−58.57	−1.01
8	1/1/1978	−65.19	−62.09	−3.10
9	1/1/1981	−62.11	−62.37	0.26
WELL 7–354	NODE 3, 23			
1	1/1/1921	1.14	—	—
2	1/1/1946	−6.61	—	—
3	1/1/1953	−12.24	0.42	−12.67
4	1/1/1958	−12.27	−2.14	−10.13
5	1/1/1965	−12.49	−1.41	−11.08
6	1/1/1968	−11.79	−2.25	−9.54
7	1/1/1973	−14.59	−2.68	−11.92
8	1/1/1978	−15.42	0.25	−15.67
9	1/1/1981	−14.10	−0.41	−13.69
LOWER POTOMAC-RARITAN-MAGOTHY AQUIFER (A1)—Continued				
WELL 7–412	NODE 8, 23			
1	1/1/1921	15.36	—	—
2	1/1/1946	7.50	—	—
3	1/1/1953	−2.16	—	—
4	1/1/1958	−7.63	—	—
5	1/1/1965	−21.81	—	—
6	1/1/1968	−30.73	−28.30	−2.43
7	1/1/1973	−41.98	−43.67	1.69
8	1/1/1978	−58.61	−59.25	0.65
9	1/1/1981	−62.75	−67.14	4.38
WELL 7–476	NODE 10, 20			
1	1/1/1921	17.75	—	—
2	1/1/1946	10.10	—	—
3	1/1/1953	1.03	—	—
4	1/1/1958	−3.86	—	—
5	1/1/1965	−16.72	−15.87	−0.85
6	1/1/1968	−25.05	−21.97	−3.08
7	1/1/1973	−34.65	−31.94	−2.71
8	1/1/1978	−46.34	−43.61	−2.73
9	1/1/1981	−50.03	−48.96	−1.07
WELL 15–296	NODE 4, 18			
1	1/1/1921	8.62	—	—
2	1/1/1946	3.14	—	—
3	1/1/1953	−4.36	—	—
4	1/1/1958	−8.25	—	—
5	1/1/1965	−17.06	−13.12	−3.93
6	1/1/1968	−24.68	−13.24	−11.44
7	1/1/1973	−28.74	−13.69	−15.05
8	1/1/1978	−29.40	−15.32	−14.07
9	1/1/1981	−26.54	−18.84	−7.70
WELL 15–323	NODE 4, 20			
1	1/1/1921	7.73	—	—
2	1/1/1946	−0.05	—	—
3	1/1/1953	−15.07	−34.47	19.40
4	1/1/1958	−20.06	−47.99	27.93
5	1/1/1965	−29.19	−60.50	31.31
6	1/1/1968	−34.91	−49.57	14.66
7	1/1/1973	−39.52	−46.14	6.61
8	1/1/1978	−40.86	−47.89	7.03
9	1/1/1981	−36.31	−47.32	11.01
MIDDLE POTOMAC-RARITAN-MAGOTHY AQUIFER (A2)				
WELL 5–63	NODE 5, 28			
1	1/1/1921	15.53	—	—
2	1/1/1946	13.34	—	—
3	1/1/1953	10.53	—	—
4	1/1/1958	8.63	—	—
5	1/1/1965	0.53	—	—
6	1/1/1968	−7.48	−4.49	−2.99
7	1/1/1973	−15.20	−10.03	−5.17
8	1/1/1978	−21.80	−14.86	−6.94
9	1/1/1981	−21.88	−18.18	−3.71

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

MIDDLE POTOMAC-RARITAN-MAGOTHY AQUIFER (A2)—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
WELL 5–101 NODE 4, 29				
1	1/1/1921	12.28	—	—
2	1/1/1946	11.80	—	—
3	1/1/1953	10.28	−1.29	11.57
4	1/1/1958	9.26	−0.92	10.17
5	1/1/1965	6.82	−4.37	10.20
6	1/1/1968	4.98	−5.02	10.00
7	1/1/1973	4.07	−4.12	8.20
8	1/1/1978	3.32	1.52	1.80
9	1/1/1981	3.22	1.24	1.99
WELL 5–261 NODE 7, 26				
1	1/1/1921	18.23	—	—
2	1/1/1946	12.62	—	—
3	1/1/1953	5.84	—	—
4	1/1/1958	1.77	—	—
5	1/1/1965	−11.79	—	—
6	1/1/1968	−20.78	—	—
7	1/1/1973	−32.53	−35.04	2.51
8	1/1/1978	−44.89	−45.80	0.91
9	1/1/1981	−47.54	−49.13	1.58
WELL 5–440 NODE 7, 30				
1	1/1/1921	25.41	—	—
2	1/1/1946	21.44	—	—
3	1/1/1953	16.30	—	—
4	1/1/1958	14.65	—	—
5	1/1/1965	2.25	—	—
6	1/1/1968	−4.92	—	—
7	1/1/1973	−12.08	−21.03	8.95
8	1/1/1978	−19.03	−28.76	9.73
9	1/1/1981	−21.49	−27.19	5.70
WELL 7–413 NODE 8, 23				
1	1/1/1921	15.82	—	—
2	1/1/1946	7.65	—	—
3	1/1/1953	−2.29	—	—
4	1/1/1958	−7.84	—	—
5	1/1/1965	−22.55	−30.43	7.88
6	1/1/1968	−31.87	−38.50	6.63
7	1/1/1973	−44.07	−52.88	8.81
8	1/1/1978	−57.07	−66.34	9.26
9	1/1/1981	−60.77	−72.14	11.36
WELL 11–137 NODE 17, 16				
1	1/1/1921	25.75	—	—
2	1/1/1946	18.91	—	—
3	1/1/1953	12.92	—	—
4	1/1/1958	9.46	—	—
5	1/1/1965	0.27	—	—
6	1/1/1968	−5.52	—	—
7	1/1/1973	−12.25	—	—
8	1/1/1978	−20.08	−36.10	16.03
9	1/1/1981	−24.48	−40.27	15.80
WELL 15–97 NODE 3, 18				
1	1/1/1921	4.88	—	—
2	1/1/1946	2.52	—	—
3	1/1/1953	0.35	—	—
4	1/1/1958	−0.61	−4.71	4.10
5	1/1/1965	−1.71	−5.50	3.79
6	1/1/1968	−4.81	−4.08	−0.73
7	1/1/1973	−5.71	0.16	−5.88
8	1/1/1978	−4.44	−0.83	−3.61
9	1/1/1981	−1.72	−2.16	0.44
WELL 23–70 NODE 2, 41				
1	1/1/1921	61.29	—	—
2	1/1/1946	61.14	59.95	1.19
3	1/1/1953	61.00	57.87	3.13
4	1/1/1958	61.00	56.75	4.25
5	1/1/1965	60.94	56.08	4.86
6	1/1/1968	60.88	58.33	2.56
7	1/1/1973	60.86	59.23	1.63
8	1/1/1978	60.84	57.93	2.91
9	1/1/1981	60.80	55.31	5.49
WELL 23–194 NODE 4, 43				
1	1/1/1921	27.89	—	—
2	1/1/1946	11.66	1.46	10.20
3	1/1/1953	−19.84	−33.76	13.92
4	1/1/1958	−23.46	−49.43	25.97
5	1/1/1965	−32.92	−54.04	21.11
6	1/1/1968	−38.04	−62.02	23.98
7	1/1/1973	−56.04	−70.70	14.66
8	1/1/1978	−70.80	−74.51	3.70
9	1/1/1981	−71.56	−77.34	5.78
WELL 23–229 NODE 4, 39				
1	1/1/1921	66.47	—	—
2	1/1/1946	64.56	—	—
3	1/1/1953	62.54	—	—
4	1/1/1958	62.31	—	—
5	1/1/1965	59.83	—	—
6	1/1/1968	57.12	61.89	−4.77
7	1/1/1973	53.94	64.27	−10.33
8	1/1/1978	52.65	57.18	−4.52
9	1/1/1981	51.83	57.11	−5.28
WELL 23–243 NODE 5, 41				
1	1/1/1921	43.97	—	—
2	1/1/1946	33.17	—	—
3	1/1/1953	21.41	—	—
4	1/1/1958	21.80	—	—
5	1/1/1965	16.47	—	—
6	1/1/1968	11.36	—	—
7	1/1/1973	0.34	—	—
8	1/1/1978	−8.13	—	—
9	1/1/1981	−11.93	−9.50	−2.43

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
MIDDLE POTOMAC-RARITAN-MAGOTHY AQUIFER (A2)—Continued				
WELL 23–270	NODE 3, 46			
1	1/1/1921	6.98	—	—
2	1/1/1946	6.95	—	—
3	1/1/1953	6.91	3.49	3.42
4	1/1/1958	6.81	5.85	0.96
5	1/1/1965	6.63	3.89	2.74
6	1/1/1968	6.57	3.49	3.08
7	1/1/1973	6.53	0.08	6.45
8	1/1/1978	6.48	−2.02	8.51
9	1/1/1981	6.44	−4.57	11.01
WELL 25–272	NODE 7, 43			
1	1/1/1921	30.93	—	—
2	1/1/1946	24.23	—	—
3	1/1/1953	14.86	—	—
4	1/1/1958	12.57	—	—
5	1/1/1965	2.90	—	—
6	1/1/1968	−2.97	—	—
7	1/1/1973	−20.53	—	—
8	1/1/1978	−34.79	−41.06	6.27
9	1/1/1981	−40.03	−50.41	10.37
WELL 29–19	NODE 19, 36			
1	1/1/1921	29.66	—	—
2	1/1/1946	26.13	—	—
3	1/1/1953	22.84	—	—
4	1/1/1958	20.21	—	—
5	1/1/1965	9.67	—	—
6	1/1/1968	4.61	—	—
7	1/1/1973	−4.81	9.14	−13.95
8	1/1/1978	−17.00	0.65	−17.65
9	1/1/1981	−26.86	−2.78	−24.09
WELL 29–85	NODE 14, 37			
1	1/1/1921	32.47	—	—
2	1/1/1946	28.81	—	—
3	1/1/1953	24.58	—	—
4	1/1/1958	22.00	—	—
5	1/1/1965	8.54	—	—
6	1/1/1968	1.54	—	—
7	1/1/1973	−8.62	−1.49	−7.13
8	1/1/1978	−23.57	−21.08	−2.49
9	1/1/1981	−37.96	−27.13	−10.83
WELL 33–251	NODE 6, 10			
1	1/1/1921	17.97	—	—
2	1/1/1946	9.12	—	—
3	1/1/1953	2.68	—	—
4	1/1/1958	−0.43	—	—
5	1/1/1965	−11.74	—	—
6	1/1/1968	−21.69	−14.60	−7.09
7	1/1/1973	−20.66	−17.71	−2.95
8	1/1/1978	−27.78	−26.05	−1.73
9	1/1/1981	−34.46	−26.94	−7.52

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
UPPER POTOMAC-RARITAN-MAGOTHY AQUIFER (A3)				
WELL 5–258	NODE 7, 26			
1	1/1/1921	22.21	—	—
2	1/1/1946	15.96	—	—
3	1/1/1953	8.92	—	—
4	1/1/1958	3.84	—	—
5	1/1/1965	−11.66	−18.90	7.24
6	1/1/1968	−19.82	−24.67	4.85
7	1/1/1973	−31.37	−38.56	7.19
8	1/1/1978	−43.32	−50.18	6.86
9	1/1/1981	−48.93	−55.20	6.26
WELL 5–274	NODE 4, 25			
1	1/1/1921	14.03	—	—
2	1/1/1946	11.01	—	—
3	1/1/1953	7.41	—	—
4	1/1/1958	4.58	—	—
5	1/1/1965	−1.83	—	—
6	1/1/1968	−6.20	—	—
7	1/1/1973	−11.72	−15.12	3.40
8	1/1/1978	−16.07	−20.65	4.58
9	1/1/1981	−16.19	−22.23	6.04
WELL 7–30	NODE 4, 21			
1	1/1/1921	7.03	—	—
2	1/1/1946	−4.35	—	—
3	1/1/1953	−20.32	−17.80	−2.52
4	1/1/1958	−25.07	−20.00	−5.07
5	1/1/1965	−35.23	−24.50	−10.73
6	1/1/1968	−41.13	−26.54	−14.59
7	1/1/1973	−48.37	−30.92	−17.45
8	1/1/1978	−46.88	−23.04	−23.84
9	1/1/1981	−39.20	−23.65	−15.55
WELL 7–117	NODE 7, 23			
1	1/1/1921	18.25	—	—
2	1/1/1946	9.42	—	—
3	1/1/1953	−0.92	—	—
4	1/1/1958	−7.44	—	—
5	1/1/1965	−27.90	—	—
6	1/1/1968	−40.34	−44.35	4.01
7	1/1/1973	−55.09	−62.10	7.01
8	1/1/1978	−64.81	−72.74	7.93
9	1/1/1981	−70.09	−76.57	6.49
WELL 7–477	NODE 10, 20			
1	1/1/1921	23.60	—	—
2	1/1/1946	14.22	—	—
3	1/1/1953	3.65	—	—
4	1/1/1958	−1.64	—	—
5	1/1/1965	−15.36	−28.65	13.29
6	1/1/1968	−25.42	−35.89	10.46
7	1/1/1973	−38.05	−48.06	10.01
8	1/1/1978	−49.00	−61.31	12.31
9	1/1/1981	−55.52	−68.62	13.09

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
UPPER POTOMAC-RARITAN-MAGOTHY AQUIFER (A3)—Continued				
WELL 15–297	NODE 4, 18			
1	1/1/1921	14.92	—	—
2	1/1/1946	11.11	—	—
3	1/1/1953	6.37	—	—
4	1/1/1958	3.57	—	—
5	1/1/1965	−2.76	−8.92	6.16
6	1/1/1968	−10.23	−8.48	−1.75
7	1/1/1973	−12.27	−8.28	−3.99
8	1/1/1978	−12.07	−10.59	−1.47
9	1/1/1981	−13.47	−12.57	−0.89
WELL 23–182	NODE 5, 43			
1	1/1/1921	18.25	—	—
2	1/1/1946	17.97	26.41	−8.43
3	1/1/1953	17.79	24.37	−6.58
4	1/1/1958	17.65	23.91	−6.26
5	1/1/1965	17.24	19.41	−2.17
6	1/1/1968	17.02	20.44	−3.43
7	1/1/1973	16.82	19.52	−2.71
8	1/1/1978	17.02	18.31	−1.29
9	1/1/1981	16.82	16.02	0.79
WELL 23–292	NODE 3, 39			
1	1/1/1921	76.58	—	—
2	1/1/1946	76.49	—	—
3	1/1/1953	76.40	—	—
4	1/1/1958	76.33	—	—
5	1/1/1965	76.10	78.82	−2.72
6	1/1/1968	75.92	80.20	−4.28
7	1/1/1973	75.68	81.48	−5.80
8	1/1/1978	75.48	76.42	−0.94
9	1/1/1981	75.40	75.17	0.23
WELL 25–316	NODE 9, 49			
1	1/1/1921	8.03	—	—
2	1/1/1946	5.72	—	—
3	1/1/1953	1.97	—	—
4	1/1/1958	−0.34	—	—
5	1/1/1965	−5.35	—	—
6	1/1/1968	−5.67	−3.14	−2.53
7	1/1/1973	−10.81	−6.46	−4.35
8	1/1/1978	−12.98	−5.85	−7.13
9	1/1/1981	−12.37	−7.47	−4.91
WELL 33–187	NODE 6, 14			
1	1/1/1921	19.35	—	—
2	1/1/1946	13.92	—	—
3	1/1/1953	8.31	—	—
4	1/1/1958	4.69	—	—
5	1/1/1965	−2.34	−14.05	11.71
6	1/1/1968	−7.32	−17.02	9.70
7	1/1/1973	−9.72	−18.48	8.76
8	1/1/1978	−11.88	−24.44	12.56
9	1/1/1981	−12.83	−25.64	12.81

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
UPPER POTOMAC-RARITAN-MAGOTHY AQUIFER (A3)—Continued				
WELL 33–253	NODE 6, 10			
1	1/1/1921	14.10	—	
2	1/1/1946	9.31	—	—
3	1/1/1953	5.94	—	—
4	1/1/1958	5.01	—	—
5	1/1/1965	2.13	—	—
6	1/1/1968	−0.87	−12.15	11.28
7	1/1/1973	−1.74	−14.45	12.71
8	1/1/1978	−4.12	−20.79	16.67
9	1/1/1981	−6.92	−22.60	15.68
ENGLISHTOWN AQUIFER (A4)				
WELL 5–259	NODE 7, 26			
1	1/1/1921	44.60	—	—
2	1/1/1946	43.37	—	—
3	1/1/1953	42.02	—	—
4	1/1/1958	40.83	—	—
5	1/1/1965	37.81	20.44	17.37
6	1/1/1968	36.25	25.33	10.92
7	1/1/1973	33.99	25.95	8.05
8	1/1/1978	31.70	25.05	6.66
9	1/1/1981	30.03	23.53	6.50
WELL 23–516	NODE 6, 42			
1	1/1/1921	90.02	—	—
2	1/1/1946	89.99	98.85	−8.86
3	1/1/1953	89.95	96.95	−7.00
4	1/1/1958	89.90	95.87	−5.97
5	1/1/1965	89.78	96.24	−6.45
6	1/1/1968	89.74	97.64	−7.90
7	1/1/1973	89.67	99.36	−9.70
8	1/1/1978	89.56	97.97	−8.41
9	1/1/1981	89.55	94.73	−5.18
WELL 25–216	NODE 7, 39			
1	1/1/1921	122.55	—	—
2	1/1/1946	122.22	—	—
3	1/1/1953	121.78	—	—
4	1/1/1958	121.46	—	—
5	1/1/1965	120.60	—	—
6	1/1/1968	119.98	—	—
7	1/1/1973	118.77	121.48	−2.71
8	1/1/1978	118.05	120.71	−2.67
9	1/1/1981	117.94	118.60	−0.66
WELL 25–250	NODE 7, 42			
1	1/1/1921	108.61	—	—
2	1/1/1946	108.18	—	—
3	1/1/1953	107.53	—	—
4	1/1/1958	106.20	—	—
5	1/1/1965	103.46	—	—
6	1/1/1968	103.56	—	—
7	1/1/1973	102.81	101.90	0.90
8	1/1/1978	101.82	101.26	0.57
9	1/1/1981	101.63	100.29	1.34

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
ENGLISHTOWN AQUIFER (A4)—Continued				
WELL 25–429	NODE 13, 41			
1	1/1/1921	48.08	—	—
2	1/1/1946	31.14	—	—
3	1/1/1953	2.60	—	—
4	1/1/1958	−15.09	—	—
5	1/1/1965	−22.35	−54.33	31.98
6	1/1/1968	−40.86	−77.46	36.60
7	1/1/1973	−84.20	−119.43	35.23
8	1/1/1978	−104.26	−135.06	30.80
9	1/1/1981	−106.67	−139.38	32.71
WELL 29–138	NODE 10, 35			
1	1/1/1921	95.26	—	—
2	1/1/1946	92.59	—	—
3	1/1/1953	88.91	—	—
4	1/1/1958	86.95	—	—
5	1/1/1965	82.20	82.61	−0.41
6	1/1/1968	77.77	78.73	−0.96
7	1/1/1973	69.67	71.00	−1.33
8	1/1/1978	66.48	65.31	1.17
9	1/1/1981	65.13	63.15	1.97
WELL 29–534	NODE 16, 36			
1	1/1/1921	53.74	—	—
2	1/1/1946	41.86	—	—
3	1/1/1953	24.76	—	—
4	1/1/1958	15.31	—	—
5	1/1/1965	9.39	—	—
6	1/1/1968	0.74	−36.64	37.37
7	1/1/1973	−30.59	−58.01	27.42
8	1/1/1978	−44.24	−73.99	29.75
9	1/1/1981	−47.01	−83.12	36.10
WENONAH-MOUNT LAUREL AQUIFER (A5)				
WELL 5–260	NODE 7, 26			
1	1/1/1921	48.94	—	—
2	1/1/1946	48.52	—	—
3	1/1/1953	48.10	—	—
4	1/1/1958	47.66	—	—
5	1/1/1965	46.72	50.64	−3.92
6	1/1/1968	46.22	52.45	−6.23
7	1/1/1973	45.57	52.12	−6.56
8	1/1/1978	44.76	51.72	−6.96
9	1/1/1981	44.41	49.44	−5.03
WELL 5–701	NODE 8, 32			
1	1/1/1921	101.24	—	—
2	1/1/1946	100.79	—	—
3	1/1/1953	99.91	—	—
4	1/1/1958	99.44	—	—
5	1/1/1965	98.01	122.96	−24.95
6	1/1/1968	97.75	123.80	−26.05
7	1/1/1973	97.12	123.04	−25.91
8	1/1/1978	96.84	123.80	−26.96
9	1/1/1981	97.08	125.40	−28.31
WENONAH-MOUNT LAUREL AQUIFER (A5)—Continued				
WELL 7–118	NODE 7, 23			
1	1/1/1921	77.81	—	—
2	1/1/1946	77.58	—	—
3	1/1/1953	77.40	—	—
4	1/1/1958	77.25	—	—
5	1/1/1965	77.09	—	—
6	1/1/1968	76.94	—	—
7	1/1/1973	76.75	71.13	5.62
8	1/1/1978	76.43	69.71	6.72
9	1/1/1981	75.80	68.81	7.00
WELL 7–478	NODE 10, 20			
1	1/1/1921	71.09	—	—
2	1/1/1946	67.17	—	—
3	1/1/1953	63.63	—	—
4	1/1/1958	60.93	—	—
5	1/1/1965	55.95	44.69	11.26
6	1/1/1968	49.15	42.78	6.37
7	1/1/1973	46.80	38.49	8.30
8	1/1/1978	43.87	36.45	7.42
9	1/1/1981	43.40	36.38	7.02
WELL 11–72	NODE 11, 9			
1	1/1/1921	28.67	—	—
2	1/1/1946	24.69	—	—
3	1/1/1953	22.52	—	—
4	1/1/1958	21.06	—	—
5	1/1/1965	17.95	—	—
6	1/1/1968	15.23	—	—
7	1/1/1973	11.72	—	—
8	1/1/1978	7.85	9.84	−1.99
9	1/1/1981	3.24	9.70	−6.46
WELL 29–140	NODE 10, 35			
1	1/1/1921	117.93	—	—
2	1/1/1946	117.25	—	—
3	1/1/1953	116.32	—	—
4	1/1/1958	115.69	—	—
5	1/1/1965	114.61	118.15	−3.54
6	1/1/1968	113.54	118.00	−4.46
7	1/1/1973	111.29	117.20	−5.91
8	1/1/1978	110.01	114.82	−4.80
9	1/1/1981	109.55	113.18	−3.63
WELL 33–02	NODE 9, 12			
1	1/1/1921	42.19	—	—
2	1/1/1946	39.70	—	—
3	1/1/1953	37.87	—	—
4	1/1/1958	36.59	—	—
5	1/1/1965	33.92	—	—
6	1/1/1968	31.61	—	—
7	1/1/1973	29.38	—	—
8	1/1/1978	26.76	23.08	3.68
9	1/1/1981	24.16	23.40	0.77

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
WENONAH-MOUNT LAUREL AQUIFER (A5)—Continued				
WELL 33–20 NODE 8, 13				
1	1/1/1921	50.32	—	—
2	1/1/1946	48.96	—	—
3	1/1/1953	47.82	—	—
4	1/1/1958	47.02	—	—
5	1/1/1965	45.40	33.99	11.41
6	1/1/1968	44.05	35.85	8.21
7	1/1/1973	42.98	35.57	7.42
8	1/1/1978	41.78	33.52	8.27
9	1/1/1981	40.67	32.95	7.72
WELL 33–252 NODE 6, 10				
1	1/1/1921	7.35	—	—
2	1/1/1946	7.57	—	—
3	1/1/1953	7.67	—	—
4	1/1/1958	6.45	—	—
5	1/1/1965	4.90	—	—
6	1/1/1968	4.81	−1.10	5.92
7	1/1/1973	6.41	1.78	4.63
8	1/1/1978	7.10	0.96	6.14
9	1/1/1981	6.72	0.12	6.60
WELL 33–279 NODE 8, 14				
1	1/1/1921	60.11	—	—
2	1/1/1946	58.53	—	—
3	1/1/1953	57.14	—	—
4	1/1/1958	56.20	—	—
5	1/1/1965	54.27	46.64	7.63
6	1/1/1968	52.54	47.86	4.68
7	1/1/1973	51.14	47.57	3.57
8	1/1/1978	49.61	44.96	4.65
9	1/1/1981	48.28	44.58	3.70
VINCENTOWN AQUIFER (A6)				
WELL 29–139 NODE 10, 35				
11	1/1/1921	121.57	—	—
22	1/1/1946	121.22	—	—
33	1/1/1953	120.73	—	—
44	1/1/1958	120.43	—	—
55	1/1/1965	119.91	129.95	−10.04
66	1/1/1968	119.34	130.40	−11.06
77	1/1/1973	118.14	131.13	−12.99
88	1/1/1978	117.44	131.14	−13.70
99	1/1/1981	117.27	129.23	−11.97
PINEY POINT AQUIFER (A7)				
WELL 5–407 NODE 13, 25				
1	1/1/1921	64.43	—	—
2	1/1/1946	63.02	—	—
3	1/1/1953	61.92	—	—
4	1/1/1958	60.90	—	—
5	1/1/1965	58.36	52.17	6.18
6	1/1/1968	56.52	52.02	4.50
7	1/1/1973	55.23	52.09	3.14
8	1/1/1978	53.80	50.93	2.87
9	1/1/1981	52.92	50.35	2.57
PINEY POINT AQUIFER (A7)—Continued				
WELL 5–676 NODE 15, 31				
1	1/1/1921	113.53	—	—
2	1/1/1946	113.29	—	—
3	1/1/1953	113.06	—	—
4	1/1/1958	112.91	—	—
5	1/1/1965	112.60	118.97	−6.37
6	1/1/1968	112.38	118.60	−6.22
7	1/1/1973	112.03	122.84	−10.81
8	1/1/1978	111.74	119.10	−7.37
9	1/1/1981	111.61	119.38	−7.77
WELL 11–44 NODE 12, 12				
1	1/1/1921	43.31	—	—
2	1/1/1946	40.03	—	—
3	1/1/1953	38.05	—	—
4	1/1/1958	36.49	—	—
5	1/1/1965	33.49	—	—
6	1/1/1968	30.25	—	—
7	1/1/1973	25.79	22.89	2.90
8	1/1/1978	21.92	18.00	3.92
9	1/1/1981	17.39	14.63	2.75
WELL 11–96 NODE 15, 9				
1	1/1/1921	31.83	—	—
2	1/1/1946	27.89	—	—
3	1/1/1953	25.78	—	—
4	1/1/1958	23.98	—	—
5	1/1/1965	20.98	—	—
6	1/1/1968	16.86	—	—
7	1/1/1973	10.22	−3.44	13.66
8	1/1/1978	4.90	−12.75	17.65
9	1/1/1981	−2.21	−18.25	16.04
WELL 11–163 NODE 14, 13				
1	1/1/1921	41.40	—	—
2	1/1/1946	37.97	—	—
3	1/1/1953	35.93	—	—
4	1/1/1958	34.31	—	—
5	1/1/1965	31.22	—	—
6	1/1/1968	27.88	—	—
7	1/1/1973	23.10	—	—
8	1/1/1978	19.03	21.85	−2.82
9	1/1/1981	14.30	18.53	−4.24
WELL 29–18 NODE 19, 36				
1	1/1/1921	26.23	—	—
2	1/1/1946	23.77	—	—
3	1/1/1953	20.00	—	—
4	1/1/1958	18.31	—	—
5	1/1/1965	15.87	6.54	9.34
6	1/1/1968	13.93	5.38	8.56
7	1/1/1973	9.55	3.88	5.67
8	1/1/1978	4.68	2.25	2.43
9	1/1/1981	3.49	0.48	3.02

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
PINEY POINT AQUIFER (A7)—Continued				
WELL 29–425	NODE 15, 33			
1	1/1/1921	121.17	—	—
2	1/1/1946	121.10	—	—
3	1/1/1953	121.00	—	—
4	1/1/1958	120.95	—	—
5	1/1/1965	120.87	118.60	2.27
6	1/1/1968	120.81	118.72	2.09
7	1/1/1973	120.66	120.62	0.04
8	1/1/1978	120.47	118.83	1.65
9	1/1/1981	120.42	118.33	2.09
LOWER KIRKWOOD-COHANSEY AQUIFER AND CONFINED KIRKWOOD AQUIFER (A8)				
WELL 1–37	NODE 22, 22			
1	1/1/1921	14.87	—	—
2	1/1/1946	−5.06	—	—
3	1/1/1953	−14.03	−59.20	45.18
4	1/1/1958	−23.56	−54.61	31.05
5	1/1/1965	−38.81	−62.44	23.64
6	1/1/1968	−54.58	−66.58	11.99
7	1/1/1973	−62.40	−64.47	2.07
8	1/1/1978	−63.71	−58.19	−5.52
9	1/1/1981	−67.54	−65.17	−2.38
WELL 1–180	NODE 20, 24			
1	1/1/1921	18.72	—	—
2	1/1/1946	3.85	—	—
3	1/1/1953	−1.76	—	—
4	1/1/1958	−6.25	—	—
5	1/1/1965	−15.37	−25.47	10.11
6	1/1/1968	−31.15	−26.99	−4.16
7	1/1/1973	−37.35	−30.83	−6.52
8	1/1/1978	−37.49	−28.29	−9.21
9	1/1/1981	−40.93	−28.75	−12.17
WELL 1–366	NODE 22, 19			
1	1/1/1921	14.10	—	—
2	1/1/1946	−10.57	−36.61	26.04
3	1/1/1953	−22.41	−51.51	29.10
4	1/1/1958	−33.47	−46.97	13.50
5	1/1/1965	−46.13	−79.65	33.53
6	1/1/1968	−59.11	−77.98	18.87
7	1/1/1973	−60.25	−68.70	8.45
8	1/1/1978	−68.34	−59.54	−8.79
9	1/1/1981	−73.01	−53.21	−19.80
WELL 1–566	NODE 20, 22			
1	1/1/1921	16.74	—	—
2	1/1/1946	−4.80	−18.93	14.13
3	1/1/1953	−11.20	−27.65	16.45
4	1/1/1958	−15.52	−19.15	3.63
5	1/1/1965	−28.47	−41.15	12.68
6	1/1/1968	−51.44	−41.43	−10.00
7	1/1/1973	−57.25	−45.00	−12.25
8	1/1/1978	−55.29	−39.74	−15.55
9	1/1/1981	−58.50	−39.95	−18.54

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
LOWER KIRKWOOD-COHANSEY AQUIFER AND CONFINED KIRKWOOD AQUIFER (A8)—Continued				
WELL 1–578	NODE 21, 18			
1	1/1/1921	18.71	—	—
2	1/1/1946	1.84	—	—
3	1/1/1953	−5.94	—	—
4	1/1/1958	−12.83	—	—
5	1/1/1965	−22.08	−30.06	7.97
6	1/1/1968	−33.80	−33.28	−0.51
7	1/1/1973	−37.16	−39.22	2.05
8	1/1/1978	−45.01	−39.18	−5.82
9	1/1/1981	−49.23	−39.77	−9.46
WELL 5–30	NODE 17, 28			
1	1/1/1921	71.98	—	—
2	1/1/1946	71.97	—	—
3	1/1/1953	71.97	—	—
4	1/1/1958	71.97	—	—
5	1/1/1965	71.97	66.50	5.47
6	1/1/1968	71.97	66.35	5.62
7	1/1/1973	71.97	69.42	2.55
8	1/1/1978	71.96	66.61	5.36
9	1/1/1981	71.96	65.83	6.13
WELL 7–479	NODE 10, 20			
1	1/1/1921	113.28	—	—
2	1/1/1946	113.28	—	—
3	1/1/1953	113.27	—	—
4	1/1/1958	113.27	—	—
5	1/1/1965	113.27	110.04	3.23
6	1/1/1968	113.22	109.47	3.76
7	1/1/1973	113.18	110.52	2.66
8	1/1/1978	113.15	110.20	2.95
9	1/1/1981	113.14	110.06	3.08
WELL 11–43	NODE 12, 12			
1	1/1/1921	74.18	—	—
2	1/1/1946	74.13	—	—
3	1/1/1953	74.06	—	—
4	1/1/1958	74.05	—	—
5	1/1/1965	74.05	—	—
6	1/1/1968	73.97	—	—
7	1/1/1973	73.93	78.96	−5.03
8	1/1/1978	73.95	76.07	−2.12
9	1/1/1981	73.74	76.04	−2.30
WELL 11–97	NODE 15, 9			
1	1/1/1921	10.66	—	—
2	1/1/1946	10.66	—	—
3	1/1/1953	10.66	—	—
4	1/1/1958	10.65	—	—
5	1/1/1965	10.64	—	—
6	1/1/1968	10.64	—	—
7	1/1/1973	10.64	1.87	8.77
8	1/1/1978	10.64	1.65	8.99
9	1/1/1981	10.63	0.92	9.70

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
LOWER KIRKWOOD-COHANSEY AQUIFER AND CONFINED KIRKWOOD AQUIFER (A8)—Continued				
WELL 29–17 NODE 19, 36				
1	1/1/1921	8.56	—	—
2	1/1/1946	7.91	—	—
3	1/1/1953	7.28	—	—
4	1/1/1958	6.90	—	—
5	1/1/1965	6.26	4.89	1.37
6	1/1/1968	5.78	4.39	1.38
7	1/1/1973	4.93	4.36	0.57
8	1/1/1978	4.24	4.46	−0.22
9	1/1/1981	3.92	3.85	0.07
WELL 29–514 NODE 18, 33				
1	1/1/1921	45.04	—	—
2	1/1/1946	45.02	—	—
3	1/1/1953	45.00	—	—
4	1/1/1958	44.99	—	—
5	1/1/1965	44.97	35.90	9.07
6	1/1/1968	44.96	36.44	8.52
7	1/1/1973	44.93	37.83	7.10
8	1/1/1978	44.84	34.48	10.37
9	1/1/1981	44.81	35.91	8.89
UPPER KIRKWOOD-COHANSEY AQUIFER (A9)				
WELL 5–570 NODE 16, 25				
1	1/1/1921	44.12	—	—
2	1/1/1946	44.11	—	—
3	1/1/1953	44.11	—	—
4	1/1/1958	44.11	48.17	−4.06
5	1/1/1965	44.11	49.72	−5.61
6	1/1/1968	44.10	49.82	−5.72
7	1/1/1973	44.10	56.30	−12.20
8	1/1/1978	44.09	48.37	−4.27
9	1/1/1981	44.09	49.05	−4.96
WELL 5–628 NODE 16, 29				
1	1/1/1921	75.71	—	—
2	1/1/1946	75.71	77.06	−1.35
3	1/1/1953	75.71	77.23	−1.52
4	1/1/1958	75.71	76.56	−0.85
5	1/1/1965	75.71	76.55	−0.85
6	1/1/1968	75.70	76.97	−1.26
7	1/1/1973	75.70	77.85	−2.15
8	1/1/1978	75.70	77.60	−1.90
9	1/1/1981	75.70	75.74	−0.05
WELL 5–689 NODE 14, 31				
1	1/1/1921	123.52	—	—
2	1/1/1946	123.52	—	—
3	1/1/1953	123.51	—	—
4	1/1/1958	123.51	127.70	−4.19
5	1/1/1965	123.50	129.15	−5.64
6	1/1/1968	123.50	130.10	−6.60
7	1/1/1973	123.49	135.32	−11.83
8	1/1/1978	123.49	131.75	−8.27
9	1/1/1981	123.48	128.44	−4.96
UPPER KIRKWOOD-COHANSEY AQUIFER (A9)—Continued				
WELL 7–503 NODE 9, 20				
1	1/1/1921	137.97	—	—
2	1/1/1946	137.96	—	—
3	1/1/1953	137.94	—	—
4	1/1/1958	137.93	—	—
5	1/1/1965	137.91	—	—
6	1/1/1968	137.90	—	—
7	1/1/1973	137.87	143.61	−5.75
8	1/1/1978	137.84	137.40	0.44
9	1/1/1981	137.82	136.19	1.63
WELL 9–23 NODE 19, 13				
1	1/1/1921	24.76	—	—
2	1/1/1946	24.74	—	—
3	1/1/1953	24.74	—	—
4	1/1/1958	24.73	24.48	0.25
5	1/1/1965	24.72	24.81	−0.09
6	1/1/1968	24.71	25.81	−1.10
7	1/1/1973	24.71	26.38	−1.68
8	1/1/1978	24.70	25.77	−1.08
9	1/1/1981	24.69	25.11	−0.42
WELL 9–48 NODE 23, 7				
1	1/1/1921	−2.21	—	—
2	1/1/1946	−7.28	—	—
3	1/1/1953	−10.53	—	—
4	1/1/1958	−17.39	−10.98	−6.41
5	1/1/1965	−20.23	−13.76	−6.47
6	1/1/1968	−20.35	−17.11	−3.24
7	1/1/1973	−20.90	−17.97	−2.93
8	1/1/1978	−24.31	−21.37	−2.94
9	1/1/1981	−26.18	−22.91	−3.27
WELL 9–60 NODE 22, 8				
1	1/1/1921	1.73	—	—
2	1/1/1946	−2.66	—	—
3	1/1/1953	−5.71	—	—
4	1/1/1958	−11.96	—	—
5	1/1/1965	−17.37	−3.67	−13.69
6	1/1/1968	−18.49	−5.53	−12.96
7	1/1/1973	−19.63	−8.71	−10.92
8	1/1/1978	−24.41	−14.49	−9.92
9	1/1/1981	−26.79	−15.35	−11.45
WELL 9–99 NODE 22, 11				
1	1/1/1921	8.43	—	—
2	1/1/1946	6.81	—	—
3	1/1/1953	5.67	—	—
4	1/1/1958	3.41	5.06	−1.65
5	1/1/1965	1.23	4.70	−3.47
6	1/1/1968	0.72	5.58	−4.86
7	1/1/1973	0.29	6.11	−5.81
8	1/1/1978	−1.40	4.84	−6.24
9	1/1/1981	−2.19	4.04	−6.23

TABLE 15.—*Comparison between simulated and measured water levels at the end of each pumping period*—Continued

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
UPPER KIRKWOOD-COHANSEY AQUIFER (A9)—Continued				
WELL 11–141	NODE 16, 13			
1	1/1/1921	22.07	—	—
2	1/1/1946	22.00	—	—
3	1/1/1953	22.00	—	—
4	1/1/1958	21.94	—	—
5	1/1/1965	21.65	12.78	8.87
6	1/1/1968	21.20	13.00	8.20
7	1/1/1973	21.07	15.31	5.76
8	1/1/1978	21.01	13.25	7.76
9	1/1/1981	20.86	12.62	8.23
WELL 11–162	NODE 14, 13			
1	1/1/1921	61.88	—	—
2	1/1/1946	61.79	—	—
3	1/1/1953	61.78	—	—
4	1/1/1958	61.79	—	—
5	1/1/1965	61.64	—	—
6	1/1/1968	61.46	—	—
7	1/1/1973	61.31	62.11	−0.80
8	1/1/1978	61.24	57.68	3.56
9	1/1/1981	61.19	58.36	2.83
WELL 11–188	NODE 9, 12			
1	1/1/1921	83.23	—	—
2	1/1/1946	83.22	—	—
3	1/1/1953	83.22	—	—
4	1/1/1958	83.21	—	—
5	1/1/1965	83.21	—	—
6	1/1/1968	83.03	—	—
7	1/1/1973	82.96	71.20	11.76
8	1/1/1978	82.93	70.80	12.14
9	1/1/1981	83.00	70.52	12.48
WELL 29–20	NODE 19, 36			
1	1/1/1921	6.34	—	—
2	1/1/1946	5.97	—	—
3	1/1/1953	5.68	—	—
4	1/1/1958	5.52	—	—
5	1/1/1965	5.25	4.29	0.96
6	1/1/1968	5.03	4.40	0.63
7	1/1/1973	4.67	4.77	−0.10
8	1/1/1978	4.40	4.57	−0.17
9	1/1/1981	4.28	4.12	0.16
WELL 29–486	NODE 13, 34			
1	1/1/1921	114.42	—	—
2	1/1/1946	114.42	—	—
3	1/1/1953	114.41	126.65	−12.23
4	1/1/1958	114.41	123.65	−9.24
5	1/1/1965	114.40	123.85	−9.45
6	1/1/1968	114.39	124.55	−10.16
7	1/1/1973	114.22	128.60	−14.37
8	1/1/1978	113.64	123.95	−10.31
9	1/1/1981	113.39	123.93	−10.54

Pumping period	End date	Simulated water level	Measured water level	Water-level difference
HOLLY BEACH AQUIFER (A10)				
WELL 9–20	NODE 23, 6			
1	1/1/1921	1.44	—	—
2	1/1/1946	1.44	—	—
3	1/1/1953	1.44	—	—
4	1/1/1958	1.44	—	—
5	1/1/1965	1.44	3.66	−2.22
6	1/1/1968	1.44	4.04	−2.60
7	1/1/1973	1.44	5.19	−3.75
8	1/1/1978	1.44	4.93	−3.49
9	1/1/1981	1.44	3.95	−2.51
WELL 9–26	NODE 23, 7			
1	1/1/1921	5.03	—	—
2	1/1/1946	5.03	—	—
3	1/1/1953	5.03	—	—
4	1/1/1958	5.03	1.80	3.23
5	1/1/1965	5.03	1.61	3.42
6	1/1/1968	5.03	1.79	3.23
7	1/1/1973	5.03	2.24	2.78
8	1/1/1978	5.03	1.67	3.35
9	1/1/1981	5.02	1.50	3.53
WELL 9–61	NODE 22, 8			
1	1/1/1921	7.83	—	—
2	1/1/1946	7.83	—	—
3	1/1/1953	7.83	—	—
4	1/1/1958	7.83	—	—
5	1/1/1965	7.83	7.57	0.26
6	1/1/1968	7.83	8.38	−0.55
7	1/1/1973	7.82	8.42	−0.61
8	1/1/1978	7.81	7.50	0.31
9	1/1/1981	7.80	7.30	0.51
WELL 9–81	NODE 23, 10			
1	1/1/1921	10.43	—	—
2	1/1/1946	10.43	—	—
3	1/1/1953	10.43	—	—
4	1/1/1958	10.42	6.95	3.47
5	1/1/1965	10.42	5.48	4.94
6	1/1/1968	10.42	7.55	2.87
7	1/1/1973	10.42	7.96	2.46
8	1/1/1978	10.42	7.50	2.92
9	1/1/1981	10.42	5.46	4.96
WELL 9–98	NODE 22, 11			
1	1/1/1921	13.18	—	—
2	1/1/1946	13.15	—	—
3	1/1/1953	13.12	—	—
4	1/1/1958	13.10	—	—
5	1/1/1965	13.07	—	—
6	1/1/1968	13.06	—	—
7	1/1/1973	13.17	17.14	−3.98
8	1/1/1978	13.16	16.02	−2.86
9	1/1/1981	13.16	14.37	−1.21

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