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SIMULATION OF GROUND-WATER FLOW AND MOVEMENT OF THE FRESHWATER-SALTWATER INTERFACE IN THE NEW JERSEY COASTAL PLAIN

Water-Resources Investigations Report 98-4216

Prepared in cooperation with the

NEW JERSEY DEPARTMENT OF ENVIRONMENTAL PROTECTION

USGS
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By Daryll A. Pope and Alison D. Gordon

U.S. GEOLOGICAL SURVEY

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West Trenton, New Jersey

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U.S. DEPARTMENT OF THE INTERIOR

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Volume	
gallon (gal)	3.785	liter
gallon (gal)	0.003785	cubic meter
million gallons (Mgal)	3,785	cubic meters
	Flow	
million gallons per day (Mgal/d)	0.04381	cubic meters per second
gallons per minute (gal/min)	0.06308	liters per second
inch per year (in/yr)	25.4	millimeter per year
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day
	Transmissivity	
foot squared per day (ft^2/d)[1]	0.09290	meter squared per day
	Density	
grams per cubic centimeter (g/cm^3)	62.43	pounds per cubic foot

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

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[1] This unit is used to express transmissivity, the capacity of an aquifer to transmit water. Conceptually, transmissivity is cubic feet (of water) per day per square foot (of aquifer area) times feet (of aquifer thickness), or $(ft^3/d)/ft^2$ x ft. In this report, this expression is reduced to its simplest form, ft^2/d.

SIMULATION OF GROUND-WATER FLOW AND MOVEMENT OF THE FRESHWATER-SALTWATER INTERFACE IN THE NEW JERSEY COASTAL PLAIN

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ABSTRACT

The confined aquifers of the New Jersey Coastal Plain are sands that range in thickness from 50 to 600 feet and are separated by confining units. The confining units are composed of silts and clays that range in thickness from 50 to 1,000 feet. The aquifers are recharged by precipitation on their outcrop areas. This water then flows laterally downdip and vertically to the deeper confined aquifers. The confined aquifers ultimately discharge to the Raritan and Delaware Bays and to the Atlantic Ocean.

In 1988, ground-water withdrawals from confined and unconfined New Jersey Coastal Plain aquifers were approximately 345 million gallons per day, more than 75 percent of which was pumped from the confined aquifers. These withdrawals have created large cones of depression in several Coastal Plain aquifers near populated areas, particularly in Camden and Monmouth Counties. The continued decline of water levels in confined aquifers can cause saltwater intrusion, reduce stream discharge near the outcrop areas, and threaten the quality of the ground-water supply.

SHARP, a quasi-three-dimensional finite-difference computer model that can simulate freshwater and saltwater flow, was used to simulate the ground-water flow system in the New Jersey Coastal Plain, including the location and movement of the freshwater-saltwater interface in nine aquifers and eight intervening confining units. The freshwater-saltwater interface is defined as the hypothetical line seaward of which the chloride concentration is equal to or greater than 10,000 milligrams per liter. Model simulations were used to estimate the location and movement of the freshwater-saltwater interface resulting from (1) eustatic sea-level changes over the past 84,000 years, (2) ground-water withdrawals from 1896 through 1988, (3) and future ground-water withdrawals from 1988 to 2040 from Coastal Plain aquifers. Simulation results showed that the location and movement of the freshwater-saltwater interface are more dependent on the historical sea level than on the stresses imposed on the flow system by ground-water withdrawals from the Coastal Plain aquifers from 1896 to 1988.

Results of a predictive simulation in which pumpage from existing wells was increased by 30 percent indicate that additional withdrawals from each of the eight confined aquifers in the Coastal Plain would broaden and deepen the existing cones of depression and result in significant drawdowns from the 1988 potentiometric surfaces. Drawdowns of 30 feet were simulated at the center of the cone of depression in the Upper, Middle, and Lower Potomac-Raritan-Magothy aquifers in Camden and Ocean Counties. Simulated drawdowns exceeded 80 feet at the center of the cone of depression in the Wenonah-Mount Laurel and Englishtown aquifers in Monmouth County. Drawdowns of 30 feet were simulated in the lower Kirkwood-Cohansey and confined

Kirkwood aquifers in Cape May County. Simulation results showed that the increase in ground-water withdrawals would result in only minimal movement of the freshwater-saltwater interface by 2040, despite large drawdowns.

INTRODUCTION

In 1988, withdrawals from confined and unconfined aquifers in the New Jersey Coastal Plain were approximately 345 Mgal/d, of which more than 75 percent was from the confined aquifers. The development of ground-water resources has occurred primarily near large population centers, creating large cones of depression in several of the Coastal Plain aquifers. Continued decline of water levels in confined aquifers poses the threat of serious adverse effects, including saltwater intrusion and reduction of stream discharge near the aquifer outcrop areas.

Effective management of the ground-water resources in the Coastal Plain requires identification of specific areas where ground-water supplies may be threatened. Threatened areas are currently defined as areas where water levels are more than 30 ft below sea level (Battaglin and Hill, 1989, p. 8). Mandated reductions in withdrawals of 35 to 50 percent have been considered or imposed in these areas (CH2M Hill and others, 1992, p. 4-32). This water-level criterion, however, may not always identify the most severely threatened areas. Local hydrogeologic conditions and proximity of saltwater may enhance the potential for saltwater intrusion in areas where water levels are higher than 30 ft below sea level. Alternatively, areas in which water levels are lower than 30 ft below sea level may not be threatened by saltwater intrusion if no nearby source of saltwater exists or if the hydrogeologic properties of the aquifer limit the movement of saltwater from source areas. Factors such as water-transmitting properties of the aquifer, sources of recharge to and discharge from the aquifer, and the location and movement of the freshwater-saltwater interface in the aquifer also may need to be considered to identify areas where ground-water supplies may be threatened and those where ground water may be safely withdrawn without adverse consequences.

In order to better identify areas in which ground-water supplies may be threatened by saltwater intrusion, the U.S. Geological Survey (USGS), in cooperation with the New Jersey Department of Environmental Protection, conducted a study of ground-water flow and the movement of the freshwater-saltwater interface in the New Jersey Coastal Plain. The objectives of the study were to (1) describe the hydrogeology of each aquifer and confining unit in the New Jersey Coastal Plain, (2) develop a ground-water flow model of the Coastal Plain that includes the parts of the aquifers that contain saltwater, and (3) use the results of the simulations and available hydrogeologic data to describe the flow system in the New Jersey Coastal Plain.

Purpose and Scope

This report describes the hydrogeology of, ground-water flow system in, and freshwater-saltwater interface movement in the New Jersey Coastal Plain sediments. The SHARP computer model (Essaid, 1990) is used to simulate the ground-water flow system, including the location and movement of the freshwater-saltwater interface in nine aquifers and eight intervening confining units in the Coastal Plain. SHARP is a quasi-three-dimensional finite-difference ground-water flow model that simulates both freshwater and saltwater flow separated by a sharp interface. The flow model is based on the USGS New Jersey Regional Aquifer-System Analysis (RASA) model

(Martin, 1998) but has been modified to simulate flow to the seaward limit of the Continental Shelf and to include the saltwater part of the flow system. Model design, boundary conditions, input data, calibration, and assumptions used in the simulations are described. The results of the simulations are used to estimate the location and movement of the freshwater-saltwater interface resulting from (1) eustatic sea-level changes over the past 84,000 years, (2) ground-water withdrawals from Coastal Plain aquifers from 1896 through 1988, and (3) hypothetical ground-water withdrawals from 1988 to 2040.

Approach

The freshwater-saltwater interface in the New Jersey Coastal Plain is still moving landward in response to past (lower) sea levels. Because the current location of the interface is not in equilibrium with the current sea level (Meisler and others, 1985), the past 84,000 years of sea-level fluctuations were simulated to obtain initial conditions for the simulation of the flow system during development (1896-1988). The resulting saltwater-freshwater interface and simulated heads represent the ground-water flow system resulting from eustatic sea-level fluctuations. Water budgets representing the flow system in 1896 were developed.

The stressed ground-water flow system from 1896 through 1988 was simulated by using available ground-water withdrawal data. The model was calibrated by using synoptic water-level data collected during fall 1988 and continuous water-level data recorded at wells during 1984-88. The flow system in 1988, including observed and simulated water levels, the simulated position of the freshwater-saltwater interface, and available chloride data, is described by means of water budgets representing the flow system in each confined aquifer.

Pumpage was increased by 30 percent at existing withdrawal sites to simulate changes in ground-water flow conditions through 2040. The flow model was used to simulate water levels and movement of the saltwater-freshwater interface in response to the increase in withdrawals. Flow budgets are used to describe changes in the flow system and the source of water to wells in the predictive simulation.

Location and Extent of Study Area

The New Jersey Coastal Plain is located in the northern part of the Coastal Plain physiographic province. The Coastal Plain sediments in the study area extend from the Fall Line in the northwest to the edge of the Continental Shelf in the southeast, and from the Delaware Bay in the southwest to Raritan Bay in the northeast. The study area (fig. 1) extends from the Fall Line to the Continental Shelf, about 80 mi offshore from Atlantic City. It includes all of Atlantic, Burlington, Camden, Cape May, Cumberland, Gloucester, Monmouth, and Ocean Counties, and parts of Middlesex and Mercer Counties in New Jersey, parts of New Castle County in Delaware, and Philadelphia and Bucks Counties in Pennsylvania. The extent of the model area is the same as that of the study area.

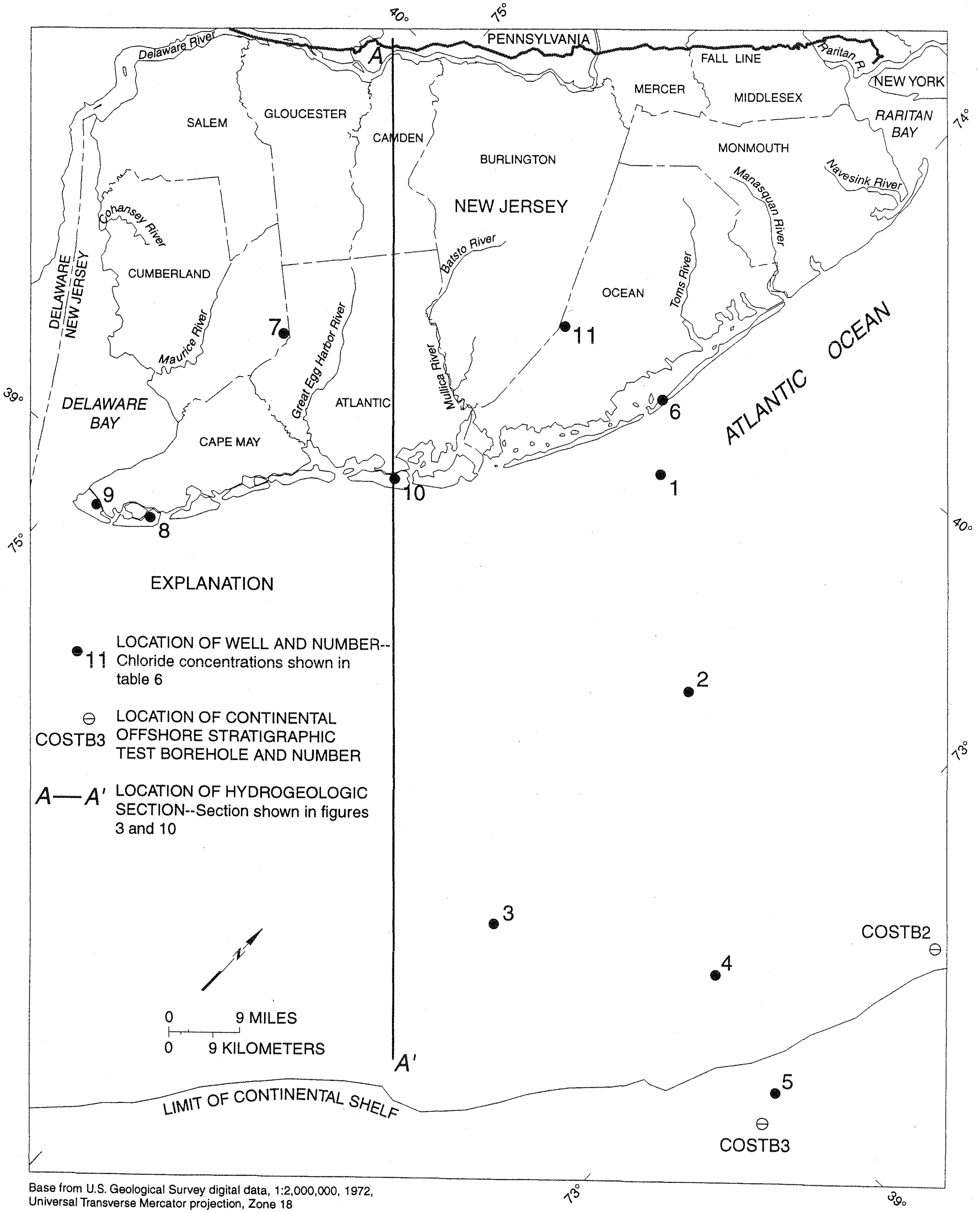


Figure 1. Map showing location of study area, New Jersey Coastal Plain.

Previous Investigations

Many reports have described the hydrogeology of the Coastal Plain sediments in New Jersey. Martin (1998) briefly describes these reports, several of which have been used extensively in this report. These reports are listed below along with a brief description of how each report has been used in this study.

Zapecza (1989) describes the hydrogeologic framework of the New Jersey Coastal Plain aquifers in onshore areas. The freshwater flow system in the New Jersey Coastal Plain was described by Martin (1998) by using a numerical ground-water flow model as part of the USGS RASA program. Martin describes the numerical ground-water flow model and uses the results of the simulation to describe the predevelopment and 1978 flow systems. Available data on aquifer and confining-unit properties in the New Jersey Coastal Plain are presented and discussed. The hydrogeologic framework used to develop the model is based on that in Zapecza (1989). The conceptual model of the flow system and the modeling approach used in this report are based strongly on Martin's work. The conceptual model in this study was modified from Martin (1998) to include the saltwater flow system in downdip areas.

Hathaway and others (1976, 1979), Scholle (1977, 1980), and Mattick and Hennessy (1980) describe the geology offshore. Results of these studies were used to extend the hydrogeologic framework in several aquifers to the limit of the Continental Shelf in order to simulate the saltwater part of the flow system.

Freshwater-saltwater relations in the New Jersey Coastal Plain are described in two reports. Meisler (1989) used chloride-concentration data from wells and boreholes to describe the location of salty ground water in both onshore and offshore areas in the Northern Atlantic Coastal Plain. A map showing the depth to the 10,000-mg/L chloride concentration in Meisler's (1989) report was used in this study to approximate the location of the freshwater-saltwater interface in each aquifer.

The effects of eustatic sea-level changes on freshwater-saltwater relations are discussed in Meisler and others (1985). A cross-sectional flow model through Atlantic City is used to simulate the location of the freshwater-saltwater interface. Meisler (1989) concluded that the observed freshwater-saltwater interface is not in equilibrium with present sea level but is still influenced by lower sea levels that have occurred in the past. He attributes the observed broad freshwater-saltwater transition zone to saltwater circulation caused by eustatic sea-level fluctuations.

Walker (1983), Eckel and Walker (1986), and Rosman and others (1995) present potentiometric-surface maps of the confined parts of most of the Coastal Plain aquifers for fall 1978, 1983, and 1988, respectively. Synoptic water levels in these reports were used to describe the ground-water flow system and calibrate the flow model.

Results of several more recent modeling studies (Pucci and others, 1994; Spitz and Barringer, 1992; and Navoy, 1994) were examined and used to modify the current flow model as needed.

DESCRIPTION OF HYDROGEOLOGIC UNITS USED IN THE MODEL

The Coastal Plain aquifer system in New Jersey is composed of seaward-dipping layers of sand, silt, and clay overlying crystalline bedrock. The sediments generally strike northeast-southwest and dip 10 to 60 ft/mi to the southeast. Detailed descriptions of the hydrogeology of the New Jersey Coastal Plain aquifers and confining units are given in Zapecza (1989) and Martin (1998).

The conceptual model used to represent the aquifer system is based on that used by Martin (1998) to simulate the freshwater flow system in the Coastal Plain. Aquifers in the New Jersey Coastal Plain are composed predominantly of sand, but also may include interbedded silts and clays. The confining units are clay and silt with minor amounts of sand (Martin, 1998, p. 6). In the current study, Martin's (1998) conceptual model was modified and extended downdip to include the flow system in the saltwater parts of the aquifers and confining units. The aquifer and confining-unit designations are the same as those used in the RASA model. The Coastal Plain sediments were simulated as ten major aquifers and nine intervening confining units. The Holly Beach water-bearing zone (model layer A10) and the underlying Cape May confining unit (model layer C9) were included in the model, but they are minor units that are present only in the Cape May County area. Aquifers in the New Jersey Coastal Plain and their corresponding geologic units are shown in table 1. A generalized hydrogeologic section through the Coastal Plain (fig. 2) shows the conceptual model of the aquifers and confining units in onshore areas.

The saltwater part of the flow system was modeled by extending the hydrogeologic framework to the edge of the Continental Shelf. Because little is known of the stratigraphy of the Continental Shelf sediments and even less is known of the hydrogeology offshore, the offshore aquifer system was assumed to form a fairly uniform wedge of units correlating to the onshore system and striking northeast-southwest. The altitudes of the top and bottom of each aquifer and confining unit were estimated by extending the trends observed in the RASA model framework, and by using available data from well logs and lithologic descriptions of two deep boreholes near the edge of the Continental Shelf. These boreholes, shown in figure 1, were drilled during the 1970's as part of the Continental Offshore Stratigraphic Test (COST) program (Scholle, 1977; Scholle, 1980). It was not possible to completely correlate the aquifers onshore to the data from these boreholes. Ages of sediments and sequences of major sand and clay units were used to estimate the altitude and thickness of each aquifer and confining unit at the Continental Shelf, enabling the interpolation of the hydrogeologic framework between the known structure onshore and the inferred structure at the edge of the shelf. Estimates of the framework characteristics of the Potomac-Raritan-Magothy aquifer system onshore in Cumberland, Cape May, and Atlantic Counties and the eastern parts of Burlington and Ocean Counties are based on those in O.S. Zapecza (U.S. Geological Survey, written commun., 1990). A generalized section showing the model representation of the aquifers and confining units to the limit of the Continental Shelf is shown in figure 3.

In the RASA model, the downdip limits of the Lower, Middle, and Upper Potomac-Raritan-Magothy aquifers, the lower Kirkwood-Cohansey and confined Kirkwood aquifers, and the upper Kirkwood-Cohansey aquifer were based on the estimated limit of freshwater in each aquifer. In this study, the ground-water flow in the saltwater part of each aquifer also is included. These aquifers are assumed to be present and to consist of permeable sands to the limit of the Conti-

Table 1. Geologic and hydrogeologic units of the New Jersey Coastal Plain and model units used in this study

[Modified from Zapecza (1989, table 2) and Seaber (1965, table 3); shading indicates adjacent geologic or hydrogeologic unit is not present]

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

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

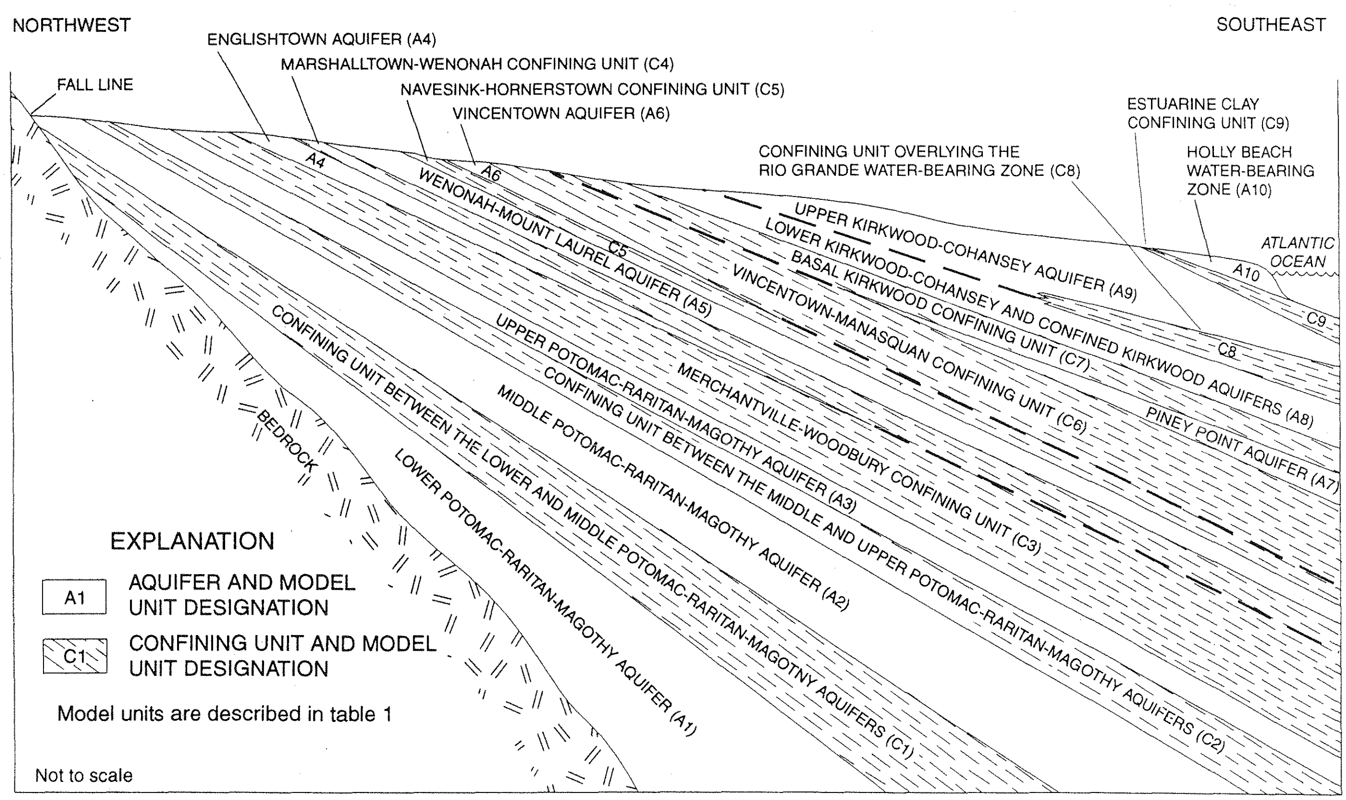


Figure 2. Generalized hydrogeologic section through the onshore part of the New Jersey Coastal Plain. (From Martin, 1998, fig. 2)

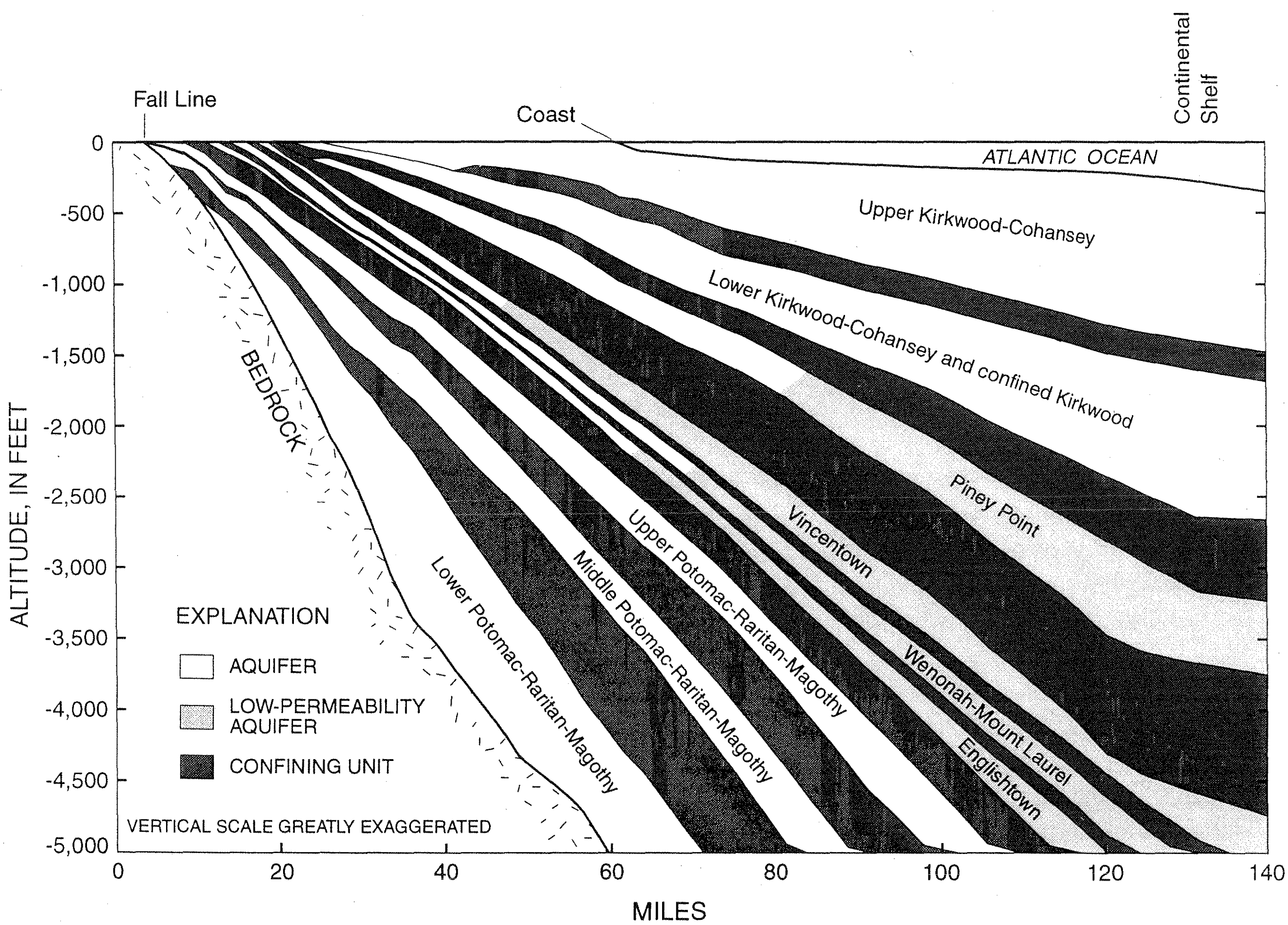


Figure 3. Generalized hydrogeologic section showing model representation of aquifers and confining units from the Fall Line to the Continental Shelf, New Jersey Coastal Plain.

nental Shelf. Hydraulic conductivities in downdip areas of each aquifer are estimated from values for the sediments near the downdip boundary of the RASA model and are assumed to decrease offshore.

Each aquifer and confining unit is described below. The discussion includes a description of the hydrogeologic framework, hydrogeologic properties, and areas where the representation of the unit is different from that used in the RASA model. Maps of the transmissivity of each aquifer and the leakance of each confining unit are presented. The confining-unit leakance maps are limited to areas where leakance in the model actually represents the presence of a confining unit and do not include areas where leakance in the model is used to represent conductances to overlying streams. Additional details on the hydrogeologic framework in the study area can be found in Zapecza (1989). Available data on the hydraulic properties of the aquifers and confining units (including transmissivities, horizontal hydraulic conductivities, and storage coefficients of the aquifers and vertical hydraulic conductivities of the confining units) are available in Martin (1998).

Lower Potomac-Raritan-Magothy Aquifer

The lowermost aquifer simulated is the Lower Potomac-Raritan-Magothy aquifer (model unit A1). The aquifer consists of sand and gravel and is underlain everywhere by crystalline bedrock. In downdip and offshore areas the model unit generally represents the lower one-third of the undifferentiated Potomac-Raritan-Magothy aquifer system. Crystalline bedrock is not penetrated by either of the COST wells (more than 16,000 ft below sea level), and geophysical investigations indicate that the depth to bedrock is between 13,000 and 20,000 ft below sea level. A thick sequence of Jurassic sediments overlies the bedrock in the COST wells. These sediments include large percentages of limestone, dolomite, clay, or claystone, all of which are less permeable than the overlying Lower Cretaceous sediments (Scholle, 1977, p. 22) and in the model are considered to be impermeable. The altitude of the bottom of the Lower Potomac-Raritan-Magothy aquifer near the edge of the Continental Shelf is assumed to be the base of the Lower Cretaceous sediments in the COST wells (Scholle, 1980, p. 6). The underlying Jurassic sediments are considered to be impermeable and serve as the model's lower boundary.

The altitude of the top of the Lower Potomac-Raritan-Magothy aquifer ranges from sea level near the Delaware River to more than 9,600 ft below sea level near the edge of the Continental Shelf. The aquifer is 75 ft thick near the Delaware River and thickens downdip to more than 2,000 ft at the limit of the Continental Shelf.

The transmissivity of the Lower Potomac-Raritan-Magothy aquifer ranges from 480 to 18,900 ft^2/d, with an average of 6,290 ft^2/d (app. 1, fig. 1a). Transmissivities are lowest updip where the aquifer thins near and under the Delaware River; they are highest downdip, offshore from northern Ocean County. Results of aquifer tests indicate that transmissivities range from 2,300 to 16,600 ft^2/d (Martin, 1998). Transmissivity in the Lower Potomac-Raritan-Magothy aquifer is generally controlled more by variations in hydraulic conductivity than by thickness. Hydraulic conductivities from 2 to 10 ft/d were estimated for the downdip areas where the aquifer was extended past the boundaries of the RASA model. The aquifer is assumed to be present and permeable downdip to the limit of the Continental Shelf.

Confining Unit between the Lower and Middle Potomac-Raritan-Magothy Aquifers

The confining unit between the Lower and Middle Potomac-Raritan-Magothy aquifers (model unit C1) overlies the entire extent of the Lower Potomac-Raritan-Magothy aquifer. The confining unit has the same areal extent as model unit A1 and is overlain everywhere by model unit A2. Thickness ranges from less than 50 ft near the Delaware River to more than 1,100 ft southeast of Cape May. Along the coast, thickness increases to the south, from less than 50 ft at the updip limit of the confining unit near the Manasquan River to the maximum thickness near Cape May. At the COST wells, the confining unit is estimated to be about 700 ft thick.

In downdip areas, the Lower and Middle Potomac-Raritan-Magothy aquifers are largely interbedded sands, silts, and clays and cannot be distinguished from one another (Zapecza, 1989, p. 12). The high leakance of the confining unit (app. 1, fig. 1b) causes the Lower and Middle aquifers to respond to stresses as one hydrologic unit. Therefore, this confining unit is more permeable than most of the other confining units in the New Jersey Coastal Plain. Leakance varies from a maximum of 2.5×10^{-4} ft/d/ft in updip areas to less than 5×10^{-7} ft/d/ft in downdip areas. Leakance values updip in Salem, Gloucester, Camden, and southern Burlington Counties are about 50 percent lower than those reported in Martin (1998). Estimates of leakance downdip are based on vertical hydraulic conductivities calculated from RASA model input data. Variations in leakance of the unit are primarily the result of differences in the thickness of the confining unit.

Middle Potomac-Raritan-Magothy Aquifer

The Middle Potomac-Raritan-Magothy aquifer (model unit A2) is equivalent to the Farrington aquifer (the Farrington Sand Member of the Magothy Formation) in the northeastern part of the New Jersey Coastal Plain. Where the Lower Potomac-Raritan-Magothy aquifer and the confining unit between the Lower and Middle Potomac-Raritan-Magothy aquifers are absent, the aquifer overlies bedrock, weathered bedrock, or clays (Martin, 1998). In downdip areas the aquifer generally represents the sandiest part of the middle one-third of the undifferentiated Potomac-Raritan-Magothy aquifer system. In downdip areas onshore the aquifer framework was based on data from O.S. Zapecza (written commun., 1992). The altitude of the top of the aquifer ranges from 100 ft above sea level to 8,000 ft below sea level offshore at the limit of the Continental Shelf. The aquifer thickens from 50 ft thick near the Delaware River to more than 800 ft thick at the COST wells. Along the coast the thickness varies from 100 ft near Monmouth County to more than 600 ft east of Ocean County, and decreases to 250 ft in northern Cape May County.

Transmissivity of the Middle Potomac-Raritan-Magothy aquifer (app. 2, fig. 2a) ranges from less than 1,000 ft^2/d in updip areas to more than 16,000 ft^2/d in Ocean County, with an average value of 2,400 ft^2/d. Estimates of transmissivity from available aquifer tests range from 42 ft^2/d to 68,600 ft^2/d (Martin, 1998). Lateral variations in transmissivity are the result of variations in hydraulic conductivity rather than differences in aquifer thickness. Hydraulic conductivities of 2 to 6 ft/d were estimated downdip from the RASA model boundary. The aquifer is assumed to extend to the limit of the Continental Shelf.

Confining Unit Between the Middle and Upper Potomac-Raritan-Magothy Aquifers

The confining unit between the Middle and Upper Potomac-Raritan-Magothy aquifers (model unit C2) overlies the entire extent of the Middle Potomac-Raritan-Magothy aquifer. The confining unit thickens downdip from less than 50 ft thick near the outcrop to 400 ft thick at the coast and to more than 750 ft thick at the limit of the Continental Shelf. The confining unit is estimated to be between 700 and 800 ft thick at the COST wells.

The leakance of the confining unit between the Middle and Upper Potomac-Raritan-Magothy aquifers (app. 2, fig. 2b) varies from about 2×10^{-4} ft/d/ft in updip areas to less than 5×10^{-7} ft/d/ft in downdip areas and in parts of Middlesex and Monmouth Counties. Variations in leakance are primarily the result of differences in the thickness of the confining unit. Leakance in downdip areas is estimated from vertical-hydraulic-conductivity values calculated from RASA model input data.

Upper Potomac-Raritan-Magothy Aquifer

The Upper Potomac-Raritan-Magothy aquifer (model unit A3) consists primarily of the Magothy Formation in New Jersey, but is the Old Bridge aquifer (the Old Bridge Sand Member of the Magothy Formation) in the northeastern part of the New Jersey Coastal Plain. The altitude of the top of the aquifer ranges from 100 ft in the outcrop area to 6,400 ft below sea level at the limit of the Continental Shelf. Onshore, the aquifer thickens downdip from a featheredge at the outcrop to 1,500 ft thick near the coast in Monmouth County and to about 250 ft thick at Cape May. Offshore, the aquifer is approximately 850 ft thick at the limit of the Continental Shelf and about 900 ft thick at the COST wells.

Transmissivity of the Upper Potomac-Raritan-Magothy aquifer (app. 3, fig. 3a) ranges from less than 500 ft^2/d in updip areas in Salem County to more than 10,000 ft^2/d in Gloucester, Camden, and Monmouth Counties. The average transmissivity of this unit in the model area is 2,800 ft^2/d. Transmissivity determined from aquifer tests ranges from 500 ft^2/d to 16,600 ft^2/d (Martin, 1998). Estimates of hydraulic conductivity downdip range from 2 to 10 ft/d. The aquifer is assumed to extend to the limit of the Continental Shelf.

Merchantville-Woodbury Confining Unit

The Merchantville-Woodbury confining unit (model unit C3) overlies the entire extent of the Upper Potomac-Raritan-Magothy aquifer. The confining unit thickens from a featheredge to 200 ft thick downdip at the coast and to more than 450 ft thick at the downdip boundary of the model. At the COST wells the unit is estimated to be between 500 and 600 ft thick.

Leakance of the Merchantville-Woodbury confining unit (app. 3, fig. 3b) varies from 1×10^{-2} ft/d/ft updip near the Delaware River in Gloucester, Camden, and Burlington Counties to 2×10^{-9} ft/d/ft along the coast in Ocean and Monmouth Counties. The Merchantville-Woodbury confining unit is one of the least permeable confining units in the New Jersey Coastal Plain. The high leakance shown in updip areas in appendix 3 (fig. 3b) represents an extension of the confining unit used to simulate the outcrop of the Upper Potomac-Raritan-Magothy aquifer near the Delaware River (Martin, 1998).

Englishtown Aquifer

The Englishtown aquifer (model unit A4) in this report includes both the upper and lower sand units of the Englishtown aquifer system as described by Zapecza (1989). A facies change from sand to silt and clay occurs in northwestern Cumberland, northwestern Atlantic, southwestern Burlington, and southwestern Ocean Counties and represents the downdip limit of the simulated aquifer. The limit of permeable sediments in the Englishtown aquifer is shown in appendix 4 (fig. 4a). The permeable part of the aquifer contains only freshwater.

The altitude of the top of the Englishtown aquifer ranges from 150 ft in the outcrop area to more than 1,600 ft below sea level downdip in Ocean County. The aquifer is generally less than 100 ft thick except in southern Monmouth County and northeastern Ocean County, where it is as much as 200 ft thick.

The transmissivity of the Englishtown aquifer (app. 4, fig. 4a) is generally less than 500 ft^2/d in the southern, permeable part of the aquifer. Transmissivity is highest in parts of Burlington and Ocean Counties and in Monmouth County, where it reaches a maximum of 1,200 ft^2/d. Estimates of transmissivity from available aquifer tests range from 1,000 to 2,100 ft^2/d (Martin, 1998). Hydraulic conductivity of the aquifer ranges from 4 to 10 ft/d. Variations in transmissivity generally are the result of differences in aquifer thickness.

Marshalltown-Wenonah Confining Unit

The Marshalltown-Wenonah confining unit (model unit C4) overlies the Englishtown aquifer in updip areas (where the Englishtown aquifer is present). In downdip areas, the confining unit directly overlies the Merchantville-Woodbury confining unit (model unit C3). The confining unit is relatively thin; its thickness onshore ranges from 20 to 80 ft. Offshore the unit thickens gradually to 180 ft at the limit of the Continental Shelf in the southeastern corner of the study area.

The leakance of the Marshalltown-Wenonah confining unit (app. 4, fig. 4b) ranges from a maximum of 3×10^{-3} ft/d/ft where the confining unit is thinnest to a minimum of 1×10^{-6} to 6×10^{-8} ft/d/ft where it is thickest (40-80+ ft thick) in northern Burlington County, parts of Ocean County, and southern Monmouth County. The leakance of the Marshalltown-Wenonah confining unit is higher than that of the other confining units in the New Jersey Coastal Plain.

Wenonah-Mount Laurel Aquifer

The Wenonah-Mount Laurel aquifer (model unit A5), where present, overlies the Marshalltown-Wenonah confining unit. The aquifer extends from its outcrop downdip approximately 50 to 60 mi where a facies change from sand to silt and clay occurs. This facies change defines the downdip, permeable part of the aquifer. The downdip limit of the simulated aquifer is shown in appendix 5 (fig. 5a). The altitude of the top of the Wenonah-Mount Laurel aquifer ranges from 200 ft in the outcrop to 2,600 ft below sea level offshore. The aquifer is generally less than 100 ft thick.

The transmissivity of the Wenonah-Mount Laurel aquifer (app. 5, fig. 5a) ranges from 3 to 1,600 ft^2/d, with an average value of 620 ft^2/d. Transmissivity greater than 1,000 ft^2/d is limited to parts of northern Salem and Cumberland Counties, eastern Gloucester and Camden Counties, southern Burlington County, and northwestern Atlantic County. Transmissivity values determined from aquifer tests range from 360 to 1,430 ft^2/d (Martin, 1998). Lateral variations in transmissivity generally are the result of differences in aquifer thickness.

Navesink-Hornerstown Confining Unit

The Navesink-Hornerstown confining unit (model unit C5) overlies the entire extent of the Wenonah-Mount Laurel aquifer. The confining unit is generally less than 100 ft thick onshore except in southern Atlantic County and in Monmouth and northern Ocean Counties. The thickness of the confining unit is approximately 60 ft along the coast and increases to 140 to 180 ft at the limit of the Continental Shelf.

The leakance of the Navesink-Hornerstown confining unit (app. 5, fig. 5b) ranges from 1×10^{-3} ft/d/ft in updip areas to 2×10^{-8} ft/d/ft in north-central Burlington County. Leakance in updip areas is among the highest in the Coastal Plain and represents areas of strong vertical connection between the Wenonah-Mount Laurel aquifer and the Vincentown aquifer.

Vincentown Aquifer

The Vincentown aquifer (model unit A6) extends from its outcrop area approximately 8 to 10 mi downdip, where a facies change from sand to silt and clay occurs. The permeable part of the aquifer is shown in appendix 6 (fig. 6a). The altitude of the top of the aquifer ranges from 200 ft in the outcrop area to 400 ft below sea level downdip. Thickness increases from about 25 ft in the outcrop area to a maximum of 140 ft in Monmouth County, although the aquifer is generally less than 100 ft thick.

The transmissivity of the Vincentown aquifer (app. 6, fig. 6a) ranges from 500 ft^2/d to 3,500 ft^2/d, with an average value of 1,500 ft^2/d. Transmissivity determined from a single laboratory test in the outcrop area in Burlington County was reported as 530 ft^2/d (Martin, 1998). Transmissivity is greatest in local areas in Salem and Monmouth Counties where the aquifer is thickest. Variations in transmissivity are the result of differences in aquifer thickness.

Vincentown-Manasquan Confining Unit

The Vincentown-Manasquan confining unit (model unit C6) overlies the Vincentown aquifer in updip areas, where the Vincentown aquifer is present. Downdip from the limit of the Vincentown aquifer, the confining unit directly overlies the Navesink-Hornerstown confining unit.

The confining unit thickens from a featheredge in the outcrop area of the overlying lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) to more than 900 ft thick at the limit of the Continental Shelf.

The leakance of the Vincentown-Manasquan confining unit (app. 6, fig 6b) ranges from 3.5×10^{-2} ft/d/ft in updip areas to less than 1×10^{-8} ft/d/ft downdip.

Piney Point Aquifer

The Piney Point aquifer (model unit A7) is completely overlain by the clay of the basal Kirkwood confining unit and does not crop out. The updip limit of the Piney Point aquifer is generally at the downdip limit of the Vincentown aquifer. The Piney Point aquifer thins and is probably absent several miles downdip from the coast. The altitude of the top of the aquifer ranges from less than 100 ft below sea level to 1,300 ft below sea level at its downdip limit. The thickness of the aquifer ranges from about 40 ft in Atlantic County to more than 500 ft in western Cumberland County. The Piney Point aquifer is continuous south into Delaware.

The transmissivity of the Piney Point aquifer (app. 7, fig. 7a) ranges from 150 to 2,700 ft^2/d. Transmissivity is highest where the aquifer is thickest in Burlington, Ocean, Cumberland, and Cape May Counties. Areas where transmissivity is less than 1,000 ft^2/d are Atlantic, southern Camden, and southern Gloucester Counties. Transmissivity reported from the only available aquifer test in this aquifer is 1,400 ft^2/d (Martin, 1998).

Basal Kirkwood Confining Unit

The basal Kirkwood confining unit (model unit C7) represents the basal clay of the Kirkwood Formation and silty parts of the Piney Point Formation. The basal Kirkwood confining unit overlies the Piney Point aquifer where the Piney Point is present, and overlies the Vincentown-Manasquan confining unit both updip and downdip from the limit of the Piney Point aquifer. The confining unit thickens gradually downdip to about 140 ft thick at the coast, and to more than 500 ft thick at the limit of the Continental Shelf.

Leakance of the basal Kirkwood confining unit (app. 7, fig. 7b) is low and ranges from 1×10^{-4} to 5×10^{-8} ft/d/ft. Leakance is highest updip in Burlington and Ocean Counties.

Lower Kirkwood-Cohansey and Confined Kirkwood Aquifers

The lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) represent the confined Atlantic City 800-foot sand and the overlying, relatively minor Rio Grande water-bearing zone in downdip areas. In this report the Atlantic City 800-foot sand and Rio Grande water-bearing zone are together referred to as the confined Kirkwood aquifer. This aquifer is assumed to be present and permeable downdip to the limit of the Continental Shelf. In updip areas this unit represents the lower part (approximately the lower one-third) of the unconfined Kirkwood-Cohansey aquifer system and is referred to as the lower Kirkwood Cohansey aquifer. The transmissivity of the lower Kirkwood-Cohansey and confined Kirkwood aquifers and the approximate updip limit of the confined Kirkwood aquifer (which is also the updip limit of the overlying confining unit) are shown in appendix 8 (fig. 8a).

The Kirkwood-Cohansey aquifer system was subdivided into an upper and lower aquifer in updip areas to better represent the vertical head distributions in the unconfined aquifer system and to provide a lateral connection between the confined Kirkwood aquifer and the lower Kirkwood-Cohansey aquifer (Martin, 1998). Where the overlying confining unit is present, the altitude of the top of the modeled aquifer represents the top of the Rio Grande water-bearing zone and ranges from 200 ft below sea level to 1,600 ft below sea level at the limit of the Continental Shelf. The thickness of the confined Kirkwood aquifer ranges from about 200 ft to about 1,000 ft at the limit of the Continental Shelf and represents the combined thickness of the Atlantic City 800-foot sand, the Rio Grande water-bearing zone, and the intervening confining unit.

The altitude of the top of the lower Kirkwood-Cohansey aquifer ranges from 170 ft in the outcrop area to 200 ft below sea level in Cumberland County near the updip limit of the overlying confining unit. The lower Kirkwood-Cohansey aquifer is generally less than 200 ft thick.

Transmissivity of the lower Kirkwood-Cohansey and confined Kirkwood aquifers (app. 8, fig. 8a) ranges from less than 1,000 ft^2/d near the outcrop area to 12,800 ft^2/d in Atlantic County, with an average value of 4,000 ft^2/d. Transmissivity determined from available aquifer tests ranges from 1,500 to 12,500 ft^2/d (Martin, 1998).

Confining Unit Overlying the Rio Grande Water-Bearing Zone

The confining unit overlying the Rio Grande water-bearing zone (model unit C8) completely overlies the confined Kirkwood aquifer. The unit thickens downdip from less than 150 ft over most of its onshore extent to slightly more than 200 ft thick at the limit of the Continental Shelf. The confining unit offshore has a relatively uniform thickness in the model.

Southwest of Cape May County in the Delaware Bay, the location of the updip limit of the confining unit has been modified from that used in the RASA model. Recent investigations of the Cape May County area including the Delaware Bay indicate that the confining unit may be thin or absent there (L.M. Voronin, U.S. Geological Survey, oral commun., 1992). The confining unit in the Delaware Bay may have been eroded by the Delaware River; in this case, the lower Kirkwood-Cohansey and confined Kirkwood aquifers would be in direct connection with the saltwater in the bay or with overlying recent sediments, which also contain saltwater, in this area. Therefore, saltwater is more likely to be moving toward Cape May from the direction of the Delaware Bay than from the offshore areas downdip.

The leakance of the confining unit overlying the Rio Grande water-bearing zone (app. 8, fig. 8b) is relatively low and ranges from 5×10^{-5} to 2.4×10^{-8} ft/d/ft.

Upper Kirkwood-Cohansey Aquifer

The upper Kirkwood-Cohansey aquifer (model unit A9) overlies the confining unit overlying the Rio Grande water-bearing zone, as well as the lower Kirkwood-Cohansey aquifer where the confining unit is absent. The upper Kirkwood-Cohansey aquifer is simulated as unconfined, overlain only by the estuarine clay confining unit (model unit C9) in offshore areas, except in Cape May County where the Holly Beach aquifer exists. The altitude of the top of the aquifer ranges from 180 ft in the outcrop area to about 300 ft below sea level at the limit of the Continental Shelf, and the thickness increases gradually from a featheredge in the outcrop to more than 1,100 ft at the limit of the Continental Shelf. Transmissivity of the upper Kirkwood-Cohansey aquifer ranges from 200 to more than 20,000 ft^2/d in offshore areas.

Estuarine Clay Confining Unit

The estuarine clay confining unit (model unit C9) overlies the upper Kirkwood-Cohansey aquifer, but is present only in peninsular Cape May County and offshore. Its thickness is identical to that in the RASA model, but in this report the top of the confining unit offshore corresponds to the bathymetry. The confining unit was extended downdip to the limit of the Continental Shelf by estimating a uniform thickness of 50 ft. Leakance offshore was determined during model calibration.

DESCRIPTION OF GROUND-WATER-FLOW MODEL

Freshwater and saltwater heads and freshwater-saltwater interface locations in 10 aquifers in the New Jersey Coastal Plain were simulated by using the SHARP model (Essaid, 1990). A schematic representation of the model units used to represent the hydrogeologic units is shown in figure 4.

Freshwater and saltwater heads and the location of the freshwater-saltwater interface were simulated (1) for an assumed steady-state condition in which sea level was relatively constant for 20,000 to 30,000 years (from 110,000 to 84,000 years ago); (2) for the past 84,000 years, by including the effects of changes in sea level (based on the initial conditions in simulation 1); and (3) for 1896 through 1988, by including ground-water withdrawals from the New Jersey Coastal Plain (based on the results of simulation 2).

Sharp-Interface Approach

The SHARP model is a quasi-three-dimensional, finite-difference model that simulates coupled freshwater and saltwater flow separated by a sharp interface (Essaid, 1990). The SHARP model can be used to simulate regional ground-water flow more efficiently than a more numerically demanding solute-transport model, which typically is practical only for simulation of flow in a two-dimensional section.

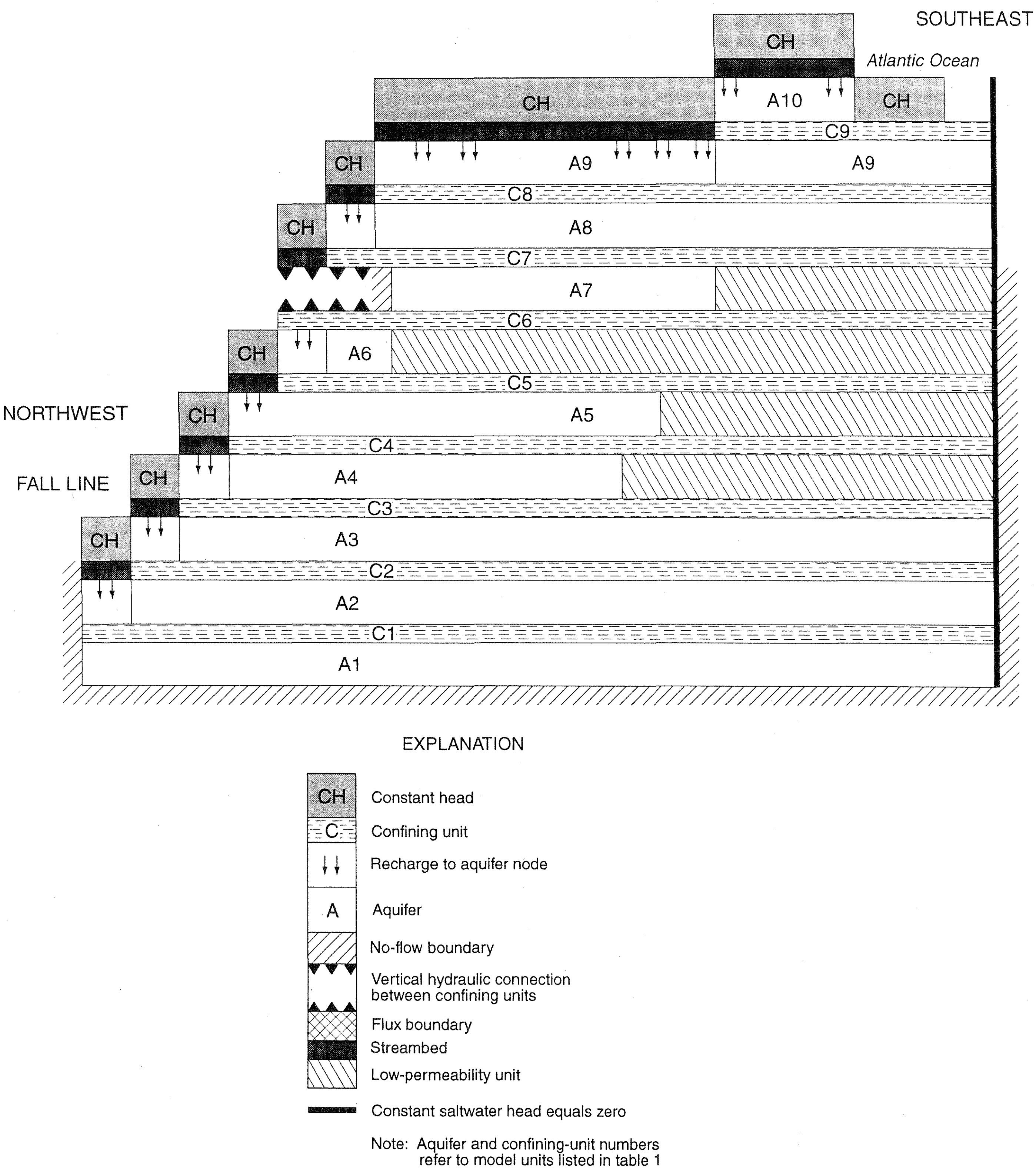


Figure 4. Schematic representation of aquifers, confining units, and boundary conditions used in the New Jersey Coastal Plain flow model.

Compared to the RASA model (Martin, 1998), the SHARP model refines the simulation of ground-water flow in the New Jersey Coastal Plain in several ways. First, it allows the location of the freshwater-saltwater interface in each of the 10 aquifers to be simulated rather than estimated on the basis of the sparse data on chloride concentrations that are available for offshore areas. The SHARP model determines the location of the interface from the simulated freshwater and saltwater heads in the aquifers. Also, heads in the vicinity of the interface represent the actual gradient of the freshwater in the aquifer rather than being strongly influenced by an adjacent no-flow boundary.

The sharp-interface approach is assumed to be valid when the width of the transition zone is small relative to the thickness of the aquifer. Although in some areas in the New Jersey Coastal Plain the zone of diffusion is wide and this assumption is false, the sharp-interface approach still represents a significant improvement over a simulation of the freshwater system only. Errors associated with violating this assumption occur in the downdip, freshwater parts of the aquifers, near the freshwater-saltwater interface. Even though the simulation of the interface in these areas may not be ideal, the ability to simulate the movement of the interface and more accurately simulate the gradients near it make the sharp interface approach more accurate than representing the interface as a stationary no-flow boundary.

In the SHARP model, freshwater and saltwater heads are simulated simultaneously. These heads are then used to solve for the elevation of the saltwater interface at each cell. The vertically integrated freshwater and saltwater flow equations are described in Essaid (1990). In cells that contain the interface, the interface elevation is given by:

$$\zeta_1 = (1 + \delta)\,\Phi_s - \delta\Phi_f ,$$

where ζ_1 is the elevation of the interface,

Φ_s is the saltwater head,

Φ_f is the freshwater head, and

$\delta = \gamma_f / (\gamma_s - \gamma_f)$,

where γ_f and γ_s are the freshwater and saltwater specific weights, respectively.

From this equation, the value of δ is 40 when the freshwater and saltwater densities are 1.0 g/cm^3 and 1.025 g/cm^3, respectively. Heads in model cells in which only freshwater or saltwater are present are simulated by using the appropriate equation for either freshwater or saltwater flow.

The SHARP model code was changed to allow recharge and streams to be simulated in aquifers other than the uppermost aquifer. This allows the simulation of recharge and overlying freshwater constant heads (representing streams) in the outcrop areas of dipping model units. An additional array is used to designate the layer to which recharge is to be applied for each model cell.

Limitations

The SHARP model simulates horizontal movement of saltwater but does not directly simulate vertical movement of saltwater. Because vertical flow between aquifers in some areas can be very large, vertical leakage in the SHARP model is simulated by using the restricted-mixing option (Essaid, 1990, p. 27). Use of this option prevents large volumes of water from being "converted" from one type to another without the user's knowledge. Because the vertical movement of saltwater into either underlying or overlying freshwater aquifers is not simulated, the location of the freshwater-saltwater interface tip and toe simulated by the model represents only horizontal movement of saltwater. Areas in which saltwater may be moving vertically into freshwater areas are identified by analyzing vertical flow rates from the model output and are considered to be possible sources of saltwater.

Horizontal movement of the interface in confining units or low-permeability sediments is much smaller than horizontal movement in aquifers. In these low-permeability areas, vertical movement of saltwater probably is much more important than horizontal movement from updip aquifers or sources of saltwater. Use of the restricted-mixing option prevents saltwater from flowing into either overlying or underlying freshwater aquifers. Because of this limitation, the movement of the freshwater-saltwater interface movement in confining units and low-permeability sediments cannot be reliably simulated.

In downdip parts of the model units representing the Englishtown (A4), Wenonah-Mount Laurel (A5), Vincentown (A6), and Piney Point (A7) aquifers (where the downdip aquifer-limit lines shown in appendixes 4 (fig. 4a), 5 (fig. 5a), 6 (fig. 6a), and 7 (fig. 7a) show a facies change to silt or clay or an aquifer pinchout), ground-water flow was simulated by using a very low hydraulic conductivity (0.001 ft/d). This allowed the position of the interface and freshwater and saltwater heads to be simulated in these areas, even though horizontal movement of water (including the freshwater-saltwater interface) is very small. These areas are not considered to be part of the aquifers as described previously, and are more representative of low-permeability confining-unit material. In general, however, movement of saltwater in confining units cannot be simulated by using the SHARP model.

Model Grid

The block-centered finite-difference model grid used is shown in figure 5. The model dimensions are 50 rows by 49 columns. The outside row and column on each side are not active cells, but are used to establish the lateral model boundaries. Onshore and in the updip part of the model (to 25 mi offshore from the coast at Cape May), the grid spacing is constant at 13,200 ft (2.5 mi) on each side. Farther offshore, the row spacing increases to a maximum of 19,800 ft (3.75 mi), whereas the column spacing remains constant at 13,200 ft.

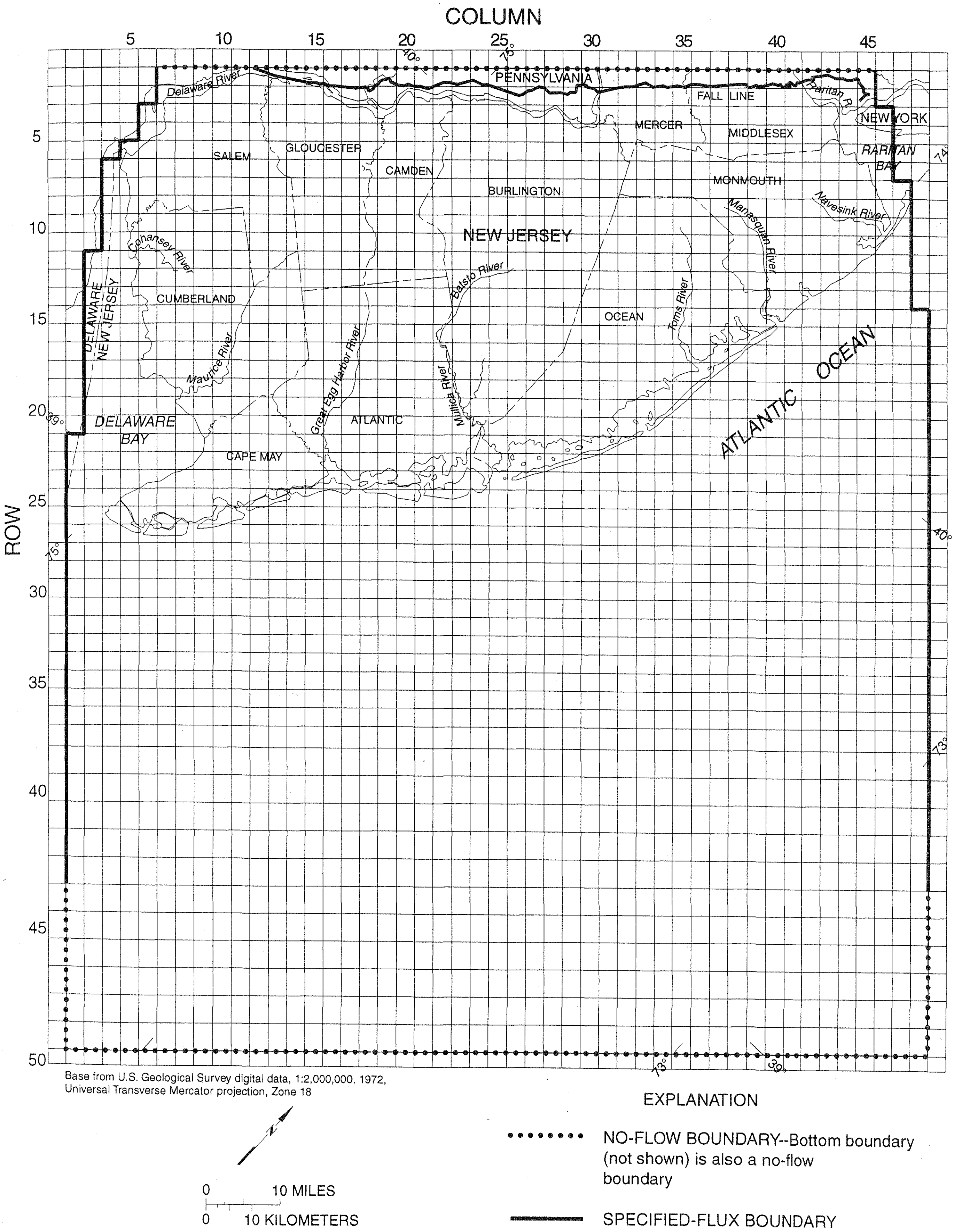


Figure 5. Model grid and generalized lateral boundaries of the New Jersey Coastal Plain flow model.

The grid is aligned approximately parallel to the Fall Line and the strike of the Coastal Plain hydrogeologic units. The grid is aligned with the RASA model grid and, in updip areas, the model cells are the same as in the RASA model. In downdip onshore and offshore areas, the grid spacing is finer than that used in the RASA model. Movement of the interface is simulated more accurately if grid spacing in the vicinity of the interface is small. The location of the freshwater-saltwater interface ranges from the Delaware Bay to more than 50 mi offshore in the upper Kirkwood-Cohansey aquifer (model unit A9), and from Salem and Cumberland Counties in the south to offshore from Ocean County in the north in the Lower Potomac-Raritan Magothy aquifer (model unit A1). Because of the wide area in which the interface is found in different aquifers, the grid was constructed to be finer than the RASA model grid in most downdip onshore and offshore areas.

Boundary Conditions

The modeling approach and boundary conditions in aquifer and confining-unit outcrops onshore and offshore are the same as those used in the New Jersey RASA model (Martin, 1998).

Generalized lateral model boundaries are shown in figure 5. The boundary conditions are slightly different in some aquifers than in others. The northwestern updip limit of all model layers is the limit of Coastal Plain sediments at the Fall Line. This represents the northwestern boundary in model unit A1. In all other model units the northwestern boundary is the updip limit of the aquifer. These updip limits are represented as no-flow boundaries in the model.

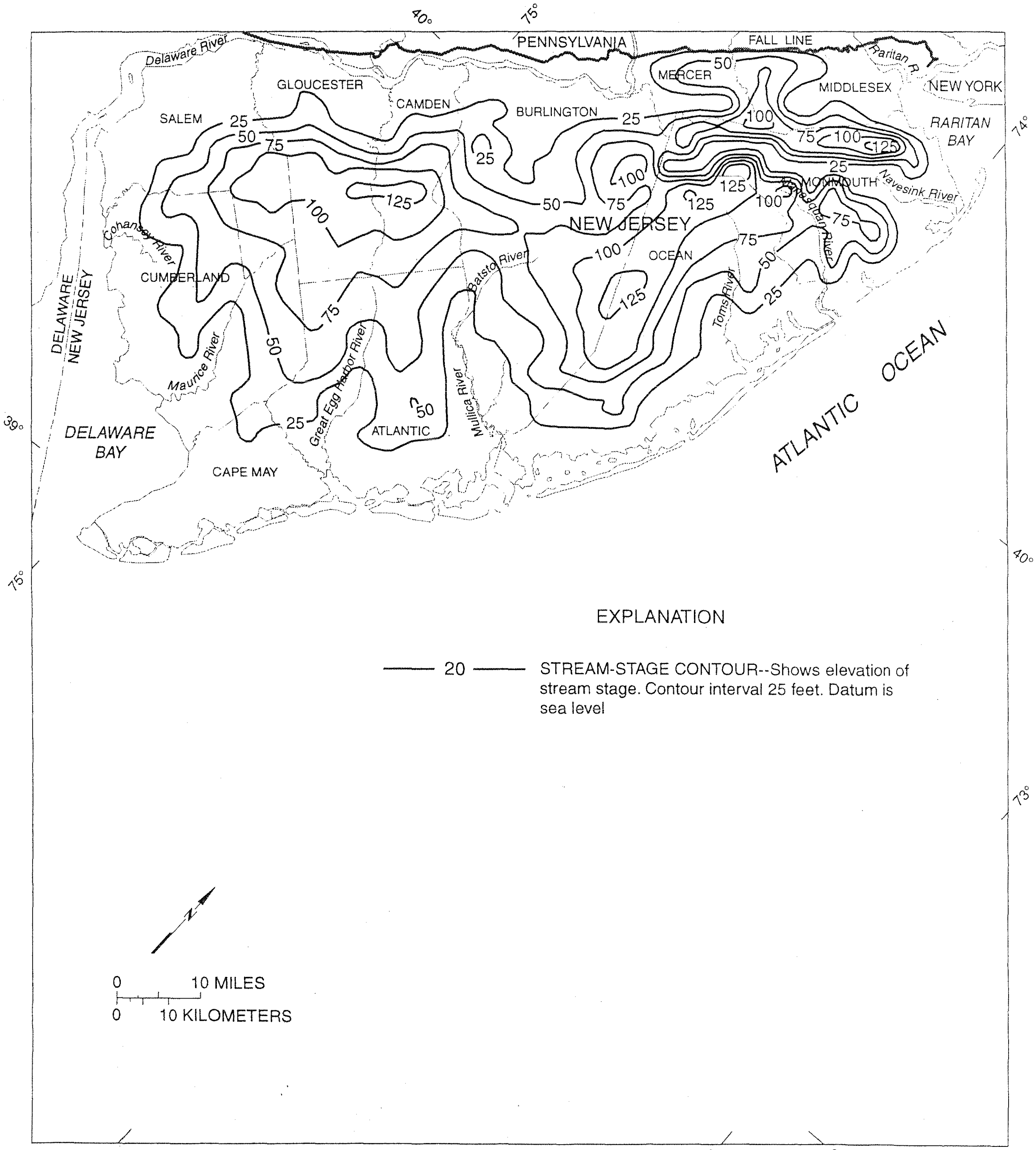
The northeastern boundary of the model approximates a flow line in a ground-water discharge area in Raritan Bay. The southwestern model boundary approximates a flow line along a ground-water divide near the Delaware Bay. These boundaries are simulated as specified-flux boundaries in the transient simulation for 1896-1988 by using flows from the Northern Atlantic Coastal Plain (NACP) model (Leahy and Martin, 1993), as they were in the New Jersey RASA model. The fluxes from the NACP model are the same as those used in the RASA model. Because the fluxes generally were small along these boundaries and the hydrogeologic properties of most of the aquifers were not changed substantially during calibration, these fluxes are assumed to be reasonable. The fluxes were modified for use in the seven stress periods used in this model by linear interpolation. In downdip areas, where the aquifers have been extended past the boundary of the RASA model, the boundaries are represented as no-flow boundaries. This approximation is valid because the gradients along these areas are small, and several of the aquifers are represented as low-permeability units with little ground-water flow.

Boundary flows were simulated by using injection or recharge wells along the edge of the model. In the SHARP model, the flow to or from a well is assumed to be an appropriate percentage of the type of water (fresh or salty) that is present in that cell over the screened interval. The ratio of freshwater flow to saltwater flow is determined by the percentage of the aquifer that contains fresh or salty water. No boundary flows were used in the initial steady-state flow model (sea level = -27 ft) or in the simulation of changes in sea level because these fluxes are unknown and are assumed to be small because the boundaries are near flow divides.

The downdip boundaries in this model are considerably different from those in the RASA model because the current model was extended to the limit of the Continental Shelf. In the RASA model, the downdip boundaries were simulated as a combination of no-flow boundaries (model units A1-A7) and specified-flux boundaries (A8, A9). In the Potomac-Raritan-Magothy aquifer system (model units A1-A3), the downdip limit of the aquifers in the RASA model was the presumed location of the freshwater-saltwater interface. In this model, these units were extended downdip to the limit of the Continental Shelf, where they are represented by a no-flow boundary. The freshwater-saltwater interface was simulated directly in the current model and the freshwater flow system near the interface is simulated more realistically than if the interface were a no-flow boundary. In model units A4 to A7, the limit of the permeable part of the aquifer is the same as in the RASA model; however, the facies change in these units is now represented by low-permeability sediments downdip. The downdip boundaries of model units A8 and A9, which were represented as specified-flux boundaries in the RASA model, are now simulated as constant saltwater heads where these units subcrop at the Continental Slope. The saltwater head used is equal to the depth of saltwater above the midpoint of the aquifer (bathymetry + 1/2 aquifer thickness) and the depth of the saltwater interface is assumed to be 25 ft above the top of the aquifer. This approach allows the freshwater-saltwater interface location to be simulated and improves the accuracy of the simulation of the freshwater flow system in these aquifers.

The lower boundary of the model represents the top of the underlying crystalline bedrock in the onshore areas that were simulated as part of the RASA model. The crystalline rocks are simulated as a no-flow boundary and underlie the Lower Potomac-Raritan-Magothy aquifer (model unit A1), except in the northern part of the New Jersey Coastal Plain onshore where the Lower aquifer is absent and the crystalline rock underlies the Middle Potomac-Raritan-Magothy aquifer (A2). Because the crystalline bedrock was not penetrated by either of the offshore COST wells, the lower boundary of the model in offshore areas is a no-flow boundary that represents the contact with the thick sequence of low-permeability Jurassic sediments that contain high percentages of limestone, dolomite, and clay or claystone observed at the COST wells.

The upper boundary of the model in onshore areas represents the water table and streams in the outcrop areas of the Coastal Plain aquifers and is the same as that used in the New Jersey RASA model (Martin, 1998). Cells in the outcrop areas of the unconfined aquifers receive recharge, and water-table altitudes respond to ground-water withdrawals. Streams in these cells are represented by a constant head in the overlying layer. The head represents the average long-term stream stage for the outcrop cell. (Each outcrop cell is assumed to have at least one stream in it.) The average elevations of these streams are shown in figure 6. The effective streambed leakances and average stream stages documented in the RASA model were used. This approach is a reasonable way to include the effects of the water table in a regional model, but it has limitations. The large cells used limit the resolution of the water table and it is not possible to include many of the local features that can be important in the aquifer recharge areas. Because the aquifer outcrops as modeled can supply an unlimited amount of water to the deeper confined aquifers, analysis of the flow budgets of the aquifers is necessary to confirm that the amount of water supplied is reasonable.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 6. Generalized stream elevations in the New Jersey Coastal Plain. (From Martin, 1998)

The outcrop areas in offshore areas are represented as constant-head nodes. Recharge is not applied in these areas. Most of these cells are in the Holly Beach water-bearing zone. The constant-head values used are equivalent freshwater heads determined from the depth of saltwater above the node, which was calculated by using the bathymetry values described in the next section.

The upper boundary of the model is affected during the simulation of sea-level changes. The SHARP model was modified to update the boundary conditions when sea level rises or falls. In each time step, the elevation of the top of every cell that is designated as an offshore outcrop cell is compared to sea level calculated for that time step to determine whether the cell should be treated as an onshore cell or an offshore cell. Cells that are converted to onshore outcrop cells as a result of changes in sea level are simulated in a similar way to other onshore outcrop cells. Recharge is applied to the cells and overlying leakance values are changed to an average leakance value for stream cells in that layer. Stream stage was approximated by using the values of bathymetry for these cells, to which recharge is applied at a rate of 10 in/yr. When sea level falls below the bathymetry at a cell, it reverts to an offshore cell and model parameters revert to their original values. Saltwater heads are updated at each time step as sea level changes at constant heads overlying offshore outcrop cells and at cells in model units A8 and A9, which represent the constant saltwater heads in the subcrop areas of these units.

Because the model was designed primarily to study the confined aquifers in the New Jersey Coastal Plain, unconfined aquifers, where they were modeled (updip parts of the confined aquifers and the upper Kirkwood-Cohansey aquifer), were included to serve primarily as boundary conditions so that flow to and from these areas to the underlying or adjacent confined aquifers could be simulated. Results of simulations for these areas are discussed only in terms of their effect on the confined parts of other aquifers. Because the transition zone of the freshwater-saltwater interface in the upper Kirkwood-Cohansey aquifer is very wide, the simulated interface position and movement are not given with confidence.

Input Data

Initial estimates of aquifer and confining-unit properties were taken directly from the RASA model. In this section, modifications made to the RASA model input data are described. Changes were required to extend model units past the boundary of the RASA model to the limit of the Continental Shelf. New data arrays that were necessary for the SHARP model include porosity, initial location of the saltwater interface, and the freshwater heads in equilibrium with this initial interface.

Horizontal hydraulic conductivities in the updip areas for each aquifer were calculated by using the calibrated transmissivity from the RASA model and the thickness of the aquifer in the hydrogeologic framework described earlier. Two methods were used to estimate horizontal conductivity downdip from the extent of the RASA model. In the downdip part of model units A1, A2, A3, A8, and A9, conductivity was estimated on the basis of trends in the RASA model conductivity values in onshore areas and by assuming a gradual decrease in conductivity with depth to a minimum value of 2 ft/d at the edge of the Continental Shelf. This minimum value is consistent with the hydraulic-conductivity value of 2.5 ft/d used in regions farthest downdip in the model of Meisler and others (1985). For the remaining model layers (A4, A5, A6, and A7), in the area in which the model was extended downdip from the RASA model boundary the layers are simulated as low-permeability sediments. Because the downdip parts of these aquifers function more as confining units than as aquifers, the horizontal conductivity was assumed to be 30,000 times the vertical conductivity of the overlying and underlying confining units on the basis of the anisotropy described in Meisler and others (1985). The mean vertical hydraulic conductivity of model units C4, C5, C6, and C7, calculated as described below, was multiplied by 30,000 to provide a single horizontal-conductivity value for the low-permeability parts of the aquifers.

Leakance values for this model were taken from the RASA model input data. In downdip areas, the RASA leakances were first converted to vertical-hydraulic-conductivity values, and trends in conductivity were used to extend the model downdip from the RASA model boundary. Vertical hydraulic conductivity is more easily estimated from observed trends than leakance because the variable thickness of the unit does not need to be considered. The RASA model calibrated leakance values were transformed into vertical conductivities by using the thickness from the current framework. Downdip values for model units C1, C2, C7, C8, and C9 were estimated from updip trends. In most cases the lowest value at the downdip limit of available data was used for the entire downdip area; however, because unit C7 showed considerable variability across strike, two zones of vertical conductivity were used. The vertical-hydraulic-conductivity values do not vary with distance downdip because the values used in updip areas range from 1×10^{-8} to 1×10^{-10} ft/d and are reasonable values.

The vertical conductivities of model units C3, C4, C5, and C6 in areas where the overlying aquifer was not simulated (as a result of a facies change) in the RASA model were calculated by using the RASA calibrated leakance values for those units. For each confining unit, a mean vertical hydraulic conductivity was calculated from the leakance values for the five rows farthest "downdip" in the RASA model. The four values were then averaged to yield a single mean vertical conductivity, which was assigned to the downdip parts of the confining units. In model units C3, C4, and C5, this value of 3.3×10^{-8} ft/d is slightly to moderately higher than the values at the RASA model layer limits. In unit C6, however, the averaged value was more than two orders of magnitude greater than the value at the downdip limit of the RASA model. The sharp boundary between the two zones caused stability problems in the simulations. Therefore, the average vertical conductivity at the downdip limit of the RASA model, 2.5×10^{-10} ft/d, was used for the entire downdip part of model unit C6.

Storage terms used in the current model were modified from those used in the RASA model. In the RASA model, a constant storage coefficient was used throughout the modeled area (except for the downdip parts of model units A1-A3). In the current mode, a constant value for specific storage was used and a storage coefficient was calculated for each cell on the basis of the specific storage and the aquifer thickness. This allows more reasonable estimates of the storage term in the thicker, downdip part of the aquifers than those used in the RASA model. Specific-storage values of 5×10^{-7} and 1×10^{-6} were used for model units A1 through A5, as determined during model calibration, and a value of 1×10^{-6} was used for units A6 through A10. Storage values used in the unconfined parts of each aquifer are equal to 0.15, a typical value for specific yield.

Recharge rates used in the model are the same as those used in the calibrated RASA model. A uniform rate of 20 in/yr was applied to the outcrop area of each aquifer. This rate was adjusted during model calibration and had little effect on model results; streamflow changed but net flow to the confined aquifers did not.

Estimates of porosity in the Coastal Plain aquifers are required for the SHARP model. A review of the literature and available data from the area were used to estimate the porosity. Freeze and Cherry (1979) list representative porosity values for unconsolidated sediments as follows: sand, 25 to 50 percent; silt, 35 to 50 percent; and clay, 40 to 70 percent. Gill (1962) reports porosities for the Cohansey Sand in Cape May County that range from 27.2 percent to 40.8 percent and average about 30 percent.

Scholle (1977, 1980) reports offshore porosities from the COST wells. Porosities were determined from density and sonic logs in the COST-B2 well and decrease with depth. Sandstone porosities from 0 to 4,000 ft below sea level range from 40 to 60 percent, from 4,000 to 9,000 ft below sea level range from 20 to 30 percent, and from 9,000 to 16,000 ft below sea level range from 5 to 20 percent. At depths below 7,000 ft, virtually all porosity values are less than 25 percent. Porosities reported for the COST-B3 well were determined from conventional and sidewall core samples beginning at depths of 7,000 ft. Porosities at the COST-B3 well are less variable with depth; most values are between 15 and 25 percent. On the basis of these values, a reasonable estimate of porosity for the New Jersey Coastal Plain aquifers was considered to be between 25 and 40 percent. For this model, a porosity of 30 percent was used for all aquifers. Porosity was varied during model calibration and affects the rate and movement of the freshwater-saltwater interface.

Because the simulation of sea-level rise and fall was a critical aspect of this model, bathymetric data were needed to delineate the onshore areas (areas that receive recharge) during simulation of sea-level changes. Contours at 10- to 20-meter (32.8- to 65.6-ft) intervals from the 1:750,000 bathymetric map of Uchupi (1970) were digitized. The bathymetry at the edge of the Continental Shelf also was modified slightly so that at least one row of cells was always present at the offshore limit of the model, even at the lowest sea level simulated. The bathymetry ranges from sea level at the present coastline to about 250 ft below sea level at the edge of the Continental Shelf. The northeastern corner of the model also includes a small area of the Continental Slope, where the bathymetry drops steeply to more than 450 ft below sea level at the edge of the model.

Initial estimates of freshwater heads and freshwater-saltwater interface locations were used as initial conditions for the steady-state run, in which the flow system resulting from a sea level set 27 ft below present-day sea level was simulated. The results of this simulation were used as a starting point for simulation of sea-level rise and fall.

Initial elevations of the freshwater-saltwater interface were estimated by using the depth to the 10,000-mg/L chloride concentration in Meisler (1989, pl. 5). After the interface elevations were fixed, the freshwater heads were calculated so that the freshwater heads and interface elevations were consistent. The freshwater heads were calculated from the interface elevation and the saltwater heads (which were set to 27 ft below present sea level for the initial simulation) in order to fulfill the condition required by the SHARP model that initial saltwater heads be near sea level (Essaid, 1990). The freshwater heads were calculated by using the following equation from Essaid (1990, p. 52):

$$\Phi_s = \frac{\zeta_1 + \delta\Phi_f}{(1+\delta)},$$

where Φ_s and Φ_f are the saltwater and freshwater heads, respectively; ζ_1 is the interface elevation; and $\delta = \gamma_f/(\gamma_s - \gamma_f)$, where γ_f and γ_s are freshwater and saltwater specific weight, respectively. The model was initialized and then stopped so that the interface could be examined for "pockets," cells of one type (fresh, salty, or mixed) completely surrounded by cells of a different type. These pockets were removed by adjusting the interface elevation and recalculating the freshwater head before continuing the model simulation.

Calibration

Heads in the model were calibrated primarily to potentiometric-surface maps for 1988 and long-term water-level hydrographs, but water-level data for 1978 also were used. Interpreted potentiometric-surface maps of water-level data collected in fall 1988 (Rosman and others, 1995, figs. 1-8) were used for all aquifers. The simulated water levels for 1988 were expected to reproduce regional flow patterns, hydraulic gradients, and locations of cones of depression. Centers of cones of depression may not be closely simulated because the water level at each cell is averaged over the cell area, which is 2-1/2 mi by 2-1/2 mi. Also, the synoptic water-level data (from which the 1988 potentiometric-surface maps were produced) may reflect local cones of depression that cannot be reproduced at the scale of this model. Water levels measured in fall 1983 and fall 1988 were averaged and compared to simulated water levels for the last stress period in the model (November 1983 through October 1988). Model calibration was considered acceptable when the difference between the average measured water level and the simulated water level (for the last stress period) at most observation wells was within 15 ft. Hydrographs of water levels in 141 observation wells (unpublished data on file at the U.S. Geological Survey office in West Trenton, N.J.) also were used to calibrate long-term trends in water levels. Simulated hydrographs also were required to match these measured water levels within 15 ft at the end of the pumping period 1983-88.

Although comparing the average water level over the last stress period to the simulated head is a good calibration tool, sources of error need to be considered. Average measured water levels in 1983 and 1998 could differ from simulated water levels if ground-water withdrawals changed sharply over the last stress period. Ground-water withdrawals were relatively steady over this period, however, so this concern can be disregarded. In addition, the measured water levels represent a specific point in space (the well), whereas the simulated water level at a grid cell is a representative value for the entire cell. The two values can differ in areas of steep gradients or where cones of depression around pumped wells are smaller than can be represented by the model grid. Additionally, simulated and observed water-level-altitude maps may differ because the simulated water level represents the average water level for 1983-88, whereas the observed water level represents conditions in fall 1988.

The calibration of the location of the freshwater-saltwater interface was accomplished by using available chloride-concentration data from Meisler (1989), other chloride-concentration data collected by the USGS (computerized data base available at the U.S. Geological Survey office in West Trenton, N.J.), and interface locations based on the depth to the 10,000-mg/L chloride concentration in Meisler (1989). Meisler's map was used to approximate the location of the freshwater-saltwater interface. High concentrations of chloride were used to verify the location of the interface; however, because most measured chloride concentrations were less than 10,000 mg/L, the chloride-concentration data were used mainly to delineate the extreme updip location of the interface.

SIMULATION OF GROUND-WATER FLOW AND FRESHWATER-SALTWATER INTERFACE MOVEMENT THROUGH 1896

The position of the freshwater-saltwater interface in the New Jersey Coastal Plain aquifers in modern times is more closely related to the predevelopment flow system than to current conditions. The current flow system is not in equilibrium with the present-day sea level; if it were, the interface would be much farther onshore than its current location. The current location of the freshwater-saltwater interface in the New Jersey Coastal Plain and the width of the freshwater-saltwater interface are the result of past sea-level conditions (Meisler and others, 1985). Sea level during the past 900,000 years has generally been lower than current sea level, has fluctuated several times, and has been as much as 300 ft below current sea level (Meisler and others, 1985, p. 6-7). The sea-level curve for the past 300,000 years (Meisler and others, 1985) is shown in figure 7. This curve was developed on the basis of work by Zelmer (1979).

The initial conditions for the simulation of flow from 1896 through 1988 were derived from the simulation of the ground-water flow system and freshwater-saltwater interface movement over the last 110,000 years. Simulated water levels and interface locations in 1896 were needed as initial conditions for a simulation of current conditions (1896-1988). If the current sea level were used, the freshwater-saltwater interface would be too far onshore and would be static; the effects of the latest rise in sea level, to which the flow system is still responding, would be unaccounted for. Use of a sea level that is 75 ft below present sea level would result in a more realistic location of the freshwater-saltwater interface, but the interface would again be static and its landward movement as a result of the recent sea-level rise would be omitted. Therefore, a simulation of the flow system that incorporates the effects of past sea-level conditions was needed.

Simulation of the flow system and freshwater-saltwater interface movement during the time of significant sea-level changes and interface movement requires two major assumptions: (1) The geologic and hydrogeologic characteristics of the aquifers and confining units is assumed to have been the same during the past 100,000 years. Any recent deposits or reworking of these sediments probably had little overall effect on the aquifers at the regional scale of this study. (2) Precipitation is assumed to have been constant over the past 100,000 years and equal to the precipitation observed today.

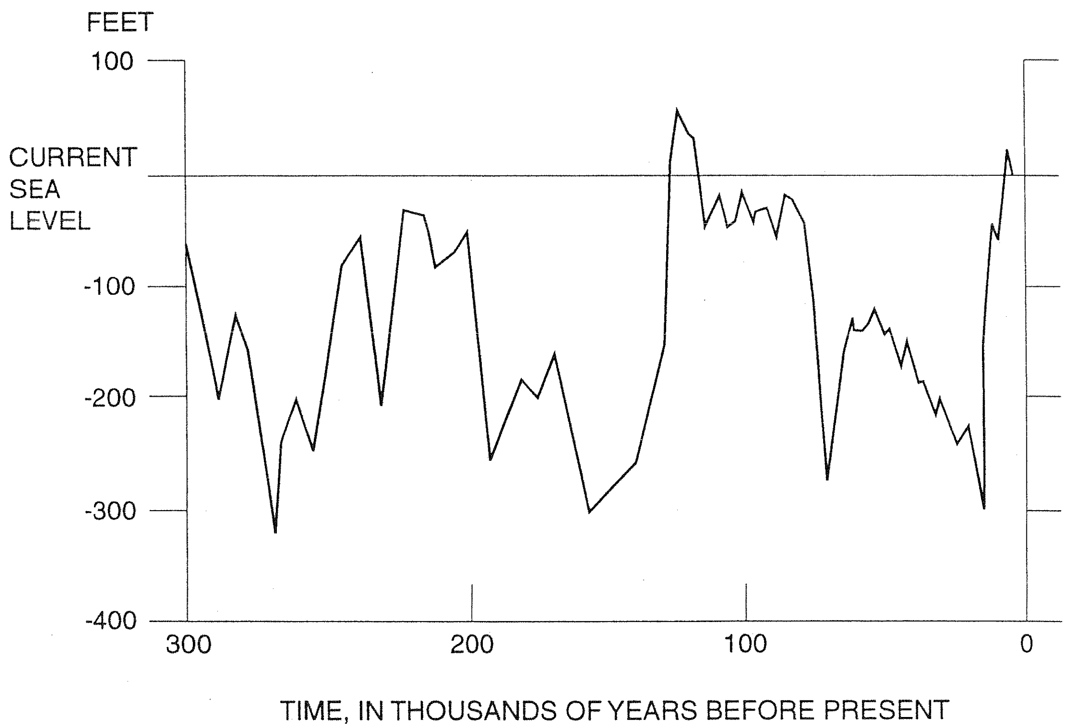


Figure 7. Sea-level curve. (Modified from Meisler and others, 1985, fig. 4B)

Simulation of Steady-State Conditions from 110,000 to 84,000 Years Ago

A steady-state simulation in which sea level was -27 ft (27 ft below current sea level) provided the initial conditions for a transient simulation that included the changing sea level as part of the boundary conditions for the period from 84,000 years ago to 1896. The sea-level curve in Zelmer (1979, fig. 7) shows several cyclic rises and falls of sea level with amplitudes of more than 200 ft over the past 900,000 years. The steady-state simulation represents a period of relatively constant sea level from about 110,000 to 84,000 years ago. This is the period during which the flow system was most likely to have reached equilibrium. Sea level during this period was about 27 ft below current sea level and was fairly stable for about 26,000 years.

The steady-state model with sea level at -27 ft was initialized by using simulated heads from the New Jersey RASA flow model (Martin, 1998) as initial heads and initial interface locations derived from the current observed location of the 10,000-mg/L chloride-concentration line in Meisler (1989, pl. 5). Initial estimates of the locations of the interface tip and toe in each aquifer were determined by using the depth to the 10,000-mg/L chloride concentration in Meisler and others (1985) and the altitude of the top and bottom of each aquifer used in the current model. In several aquifers the initial location of the freshwater-saltwater interface was moved toward the updip limit of the aquifer because the interface moved more easily downdip than updip as a result of onshore recharge, which "drives" the flow. Although the resulting location of the freshwater-saltwater interface would have been the same, the solution was reached more quickly and with fewer numerical problems.

Because the SHARP model does not directly simulate a steady-state condition, a transient simulation with a time step of 500 years was used. The porosity was set to an extremely low value (0.005) so that little water is displaced as the interface moves. This maximizes the simulated rate of movement of the interface. Although the model-output simulation times are no longer accurate, this does not affect the steady-state results. The storage coefficient was set equal to 0.0 for the steady-state simulation. Any freshwater-storage terms in the model budget are the result of water entering or leaving the freshwater system as a result of the movement of the interface. The steady-state simulation was continued until the volume of water contributed from storage was very small, indicating that the freshwater-saltwater interface was no longer moving. The flow system is not presented and discussed in this section; however, the initial location of the interface in aquifers in which the permeability decreases downdip (the Englishtown (A4), Wenonah-Mount Laurel (A5), Vincentown (A6), and Piney Point (A7) aquifers) is discussed.

In the Englishtown (A4) and Vincentown (A6) aquifers, the interfaces were initialized downdip from the permeable part of the aquifer (within the low-permeability areas) and did not move significantly during the steady-state simulation. The low-permeability sediments limit the movement of the interface. When the interface is initially located within the low-permeability zone, or moves downdip to the boundary between the permeable aquifer and the low-permeability zone during the steady-state simulation, the exact location of the interface cannot be simulated. The interface is somewhere downdip from the change to a low-permeability zone but downdip movement has been controlled by the low horizontal hydraulic conductivities. Movement of the interface in low-permeability materials (clays, silts, and silty sands) is controlled more by vertical movement of saltwater through overlying and underlying sediments than by horizontal movement of saltwater.

In the Wenonah Mount-Laurel aquifer (A5), the interface was initialized near the downdip limit of the permeable part of the aquifer, and the tip and toe of the interface moved offshore during the steady-state simulation (sea level = -27 ft). In northern areas (north of a line extending due east from the Great Egg Harbor River at the coast), the interface moved offshore during the steady-state simulation until it reached the low-permeability zone and did not move significantly within the low-permeability area. This result implies that north of this line the interface is located within the low-permeability sediments farther offshore than is simulated. South of this line the steady-state interface moved as much as 3.5 mi farther offshore than the initial estimates within the permeable limits of the aquifer during the steady-state simulation; however, the movement of the interface in this area also may have been affected by proximity to the low-permeability boundary, but to a smaller degree. Therefore, this is probably a good approximation of the location of the interface.

In the Piney Point aquifer (A7), the interface was initialized just downdip from the boundary between the permeable part of the aquifer and the low-permeability sediments in areas south of Absecon Island (in Atlantic County north of the mouth of the Great Egg Harbor River). Therefore, the location of the freshwater-saltwater interface in this part of the aquifer cannot be simulated with confidence. In areas north of Absecon Island, where the interface was initialized within the permeable aquifer sediments, both the tip and toe of the interface moved updip from the estimated initial position during the steady-state simulation. These estimated positions were adjusted during previous steady-state runs and were from 15 to 35 mi farther onshore than the locations estimated from data in Meisler and others (1985).

The lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8) and the upper Kirkwood-Cohansey aquifer (A9) are in closer contact with overlying saltwater than the deeper aquifers and are assumed to be in connection with saltwater at the downdip subcrop; therefore, the freshwater-saltwater interface moves farther in these aquifers than in the deeper aquifers during simulation of sea-level changes. The simulated steady-state interface (representing a sea level of 27 ft below current sea level) is similar to the static condition that the interfaces in the two aquifers would eventually reach under current conditions if no stresses were imposed. The simulated heads and interfaces in these aquifers are shown in figures 8 and 9, respectively.

The confining unit overlying the Rio Grande water-bearing zone (C8) affects the steady-state interface in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8). North of Island Beach State Park, the confining unit is absent and the simulated steady-state tip and toe of the freshwater-saltwater interface in the lower Kirkwood-Cohansey and confined Kirkwood aquifers are located just offshore from the barrier islands. Where the confining unit is present, however, heads offshore are higher than they are where the confining unit is absent. The interface is farther offshore in these areas because the heads are higher.

Southwest of Cape May County in the Delaware Bay, where the overlying confining unit (C8) is absent, the lower Kirkwood-Cohansey and confined Kirkwood aquifers are in direct contact with the overlying, saltwater-containing sediments of the upper Kirkwood-Cohansey aquifer (A9).

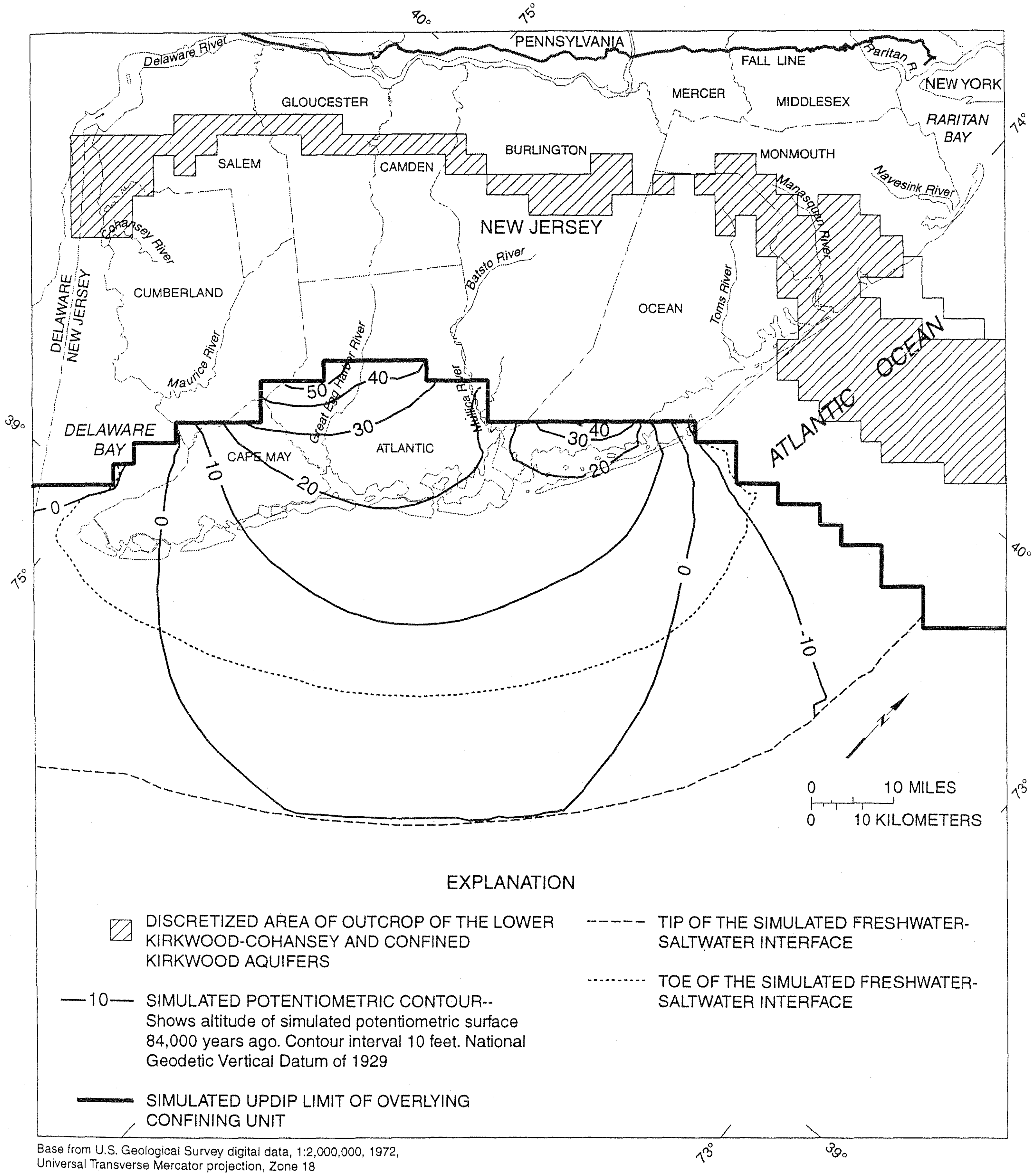


Figure 8. Simulated potentiometric surface in the confined Kirkwood aquifer, and locations of the simulated freshwater-saltwater interface tip and toe in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) in the New Jersey Coastal Plain 84,000 years ago.

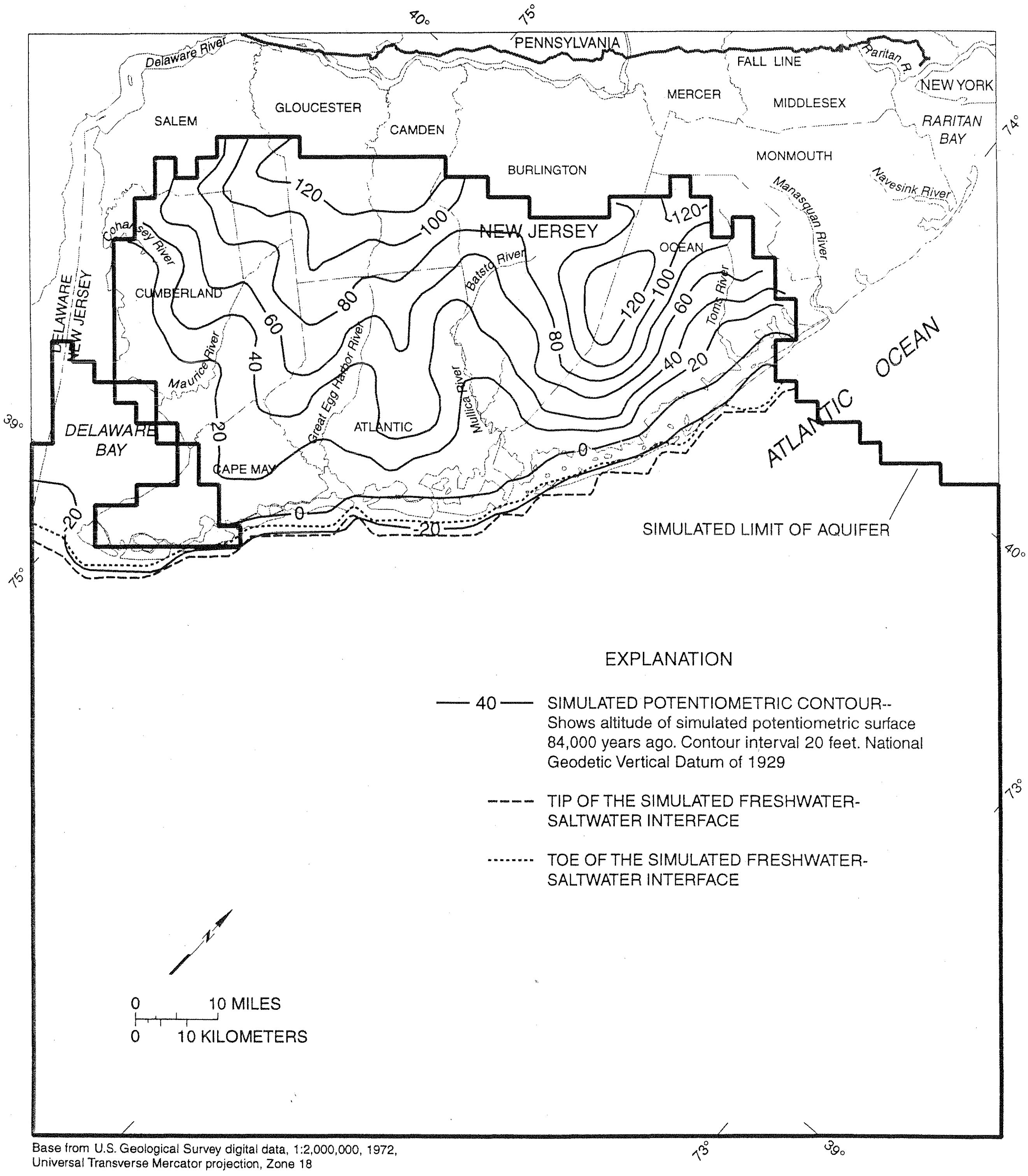


Figure 9. Simulated potentiometric surface and locations of the simulated freshwater-saltwater interface tip and toe in the upper Kirkwood-Cohansey aquifer (model unit A9) in the New Jersey Coastal Plain 84,000 years ago.

The simulated steady-state toe of the freshwater-saltwater interface in model unit A8 gradually curves toward the west around the southern end of Cape May County. The interface is farther onshore near Cape May than farther north because heads near the updip limit of the confining unit overlying the Rio Grande water-bearing zone (C8) are low.

In the upper Kirkwood-Cohansey aquifer, the simulated steady-state interface is narrow, is just offshore, and curves toward the tip of Cape May County. The simulated steady-state interface positions in the lower and upper Kirkwood-Cohansey aquifers (model units A8 and A9) represent the equilibrium condition toward which the flow system is moving.

Simulation of Sea-Level Changes

Changes in sea level affect the flow system and movement of the freshwater-saltwater interface in the New Jersey Coastal Plain in several ways. The depth of overlying saltwater in offshore areas decreases when sea level drops, resulting in lower saltwater heads both in the constant-head cells overlying the uppermost aquifer (representing overlying saltwater in the ocean) and in the constant heads at the subcrop of model units A8 and A9 at the Continental Shelf. Saltwater heads offshore have decreased while the freshwater heads remained constant, causing the freshwater-saltwater interface to move seaward. Decreasing sea level also exposes sediments that had been covered by saltwater; they become part of the onshore freshwater flow system, with freshwater streams and freshwater recharge. The size of the area of freshwater recharge increases, and this additional freshwater in the uppermost aquifer (A9) moves the saltwater farther offshore. Higher heads resulting from increased freshwater recharge cause the interface to move seaward quickly. When sea level rises again, however, no strong force exists to reverse the flow. Therefore, the freshwater-saltwater interface moves seaward in response to a sea-level drop much faster than it moves landward after a sea-level rise. For these reasons, the freshwater-saltwater interface has not yet reached equilibrium with current sea level.

Sea-level changes during the past 84,000 years were simulated by using the heads and interfaces resulting from the steady-state simulation as initial conditions. Storage coefficients and porosity values used were set to reasonable values for a transient simulation. Rates and magnitudes of sea-level changes were estimated from the sea-level curve shown in figure 7 and are as follows:

Time (years before present)	Rate (feet per year)	Sea level at end of period (feet below land surface)
110,000 - 84,000	0	27
84,000 - 71,000	-0.015615	230
71,000 - 61,000	.0092	138
61,000 - 50,000	0	138
50,000 - 18,000	-.0031875	240
18,000 - 100	.0133	0

Calibration of the location of the freshwater-saltwater interface during sea-level changes was based primarily on the positions of the interface in the upper Kirkwood-Cohansey aquifer (A9) and the lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8) because the interface moved farthest and was most sensitive to changes in hydraulic parameters in these aquifers. Most of the changes made during calibration were made in model unit A9. For this reason, the simulated heads and interface in this aquifer are less reliable than those in the deeper aquifers, and this unit was included in the simulation only as a boundary condition so that the heads and interfaces in the deeper units (especially A8) could be more accurately simulated. Changes made during calibration include (1) reducing the leakance of confining unit C9 overlying the upper Kirkwood-Cohansey aquifer (A9), which represents sediments on the ocean floor; (2) adjusting the streambed conductances used when converting offshore cells to onshore cells; and (3) reducing the recharge applied to offshore areas to 10 in/yr when sea level falls. These changes were made to simulate the interface in model unit A9 so that the general shape (slope and width) and position of the interface were reasonable. The model was sensitive to changes in the leakance of model unit C9 and the streambed conductances offshore and was not sensitive to changes in recharge.

The hydrogeologic section in figure 10 shows the simulated interface locations in the steady-state simulation (in which sea level is 27 ft below current sea level) and the simulation of sea-level changes through 1896. (The location of this section, shown in figure 1, is the same as that used for the sections in figures 2 and 3.)

Simulation of changes in sea level generally resulted in downdip movement of the freshwater-saltwater interface from the simulated steady-state interface position. Most of the downdip movement was the result of the drop in sea level from 27 ft below present sea level 84,000 years ago to 230 ft below sea level 71,000 years ago. The freshwater-saltwater interface did not reach equilibrium with the lowest sea level, but was still moving downdip when sea level began to rise again. At the end of the simulation (1896), the freshwater-saltwater interface was moving updip in response to the rise in sea level from 240 ft below sea level 18,000 years ago to present sea level in 1896.

Movement of the Freshwater-Saltwater Interface

The movement of the freshwater-saltwater interface from 84,000 years ago to 1896 was greatest in the upper Kirkwood-Cohansey aquifer and the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model units A8 and A9) and in the three Potomac-Raritan-Magothy aquifers (model units A1, A2, and A3). The interface moved downdip in the Potomac-Raritan Magothy aquifers as much as 3.5 mi as a result of the fluctuations in sea level. The tip and toe of the interface in each aquifer moved seaward about 1.75 to 2 mi. Because these aquifers are permeable downdip to the Continental Shelf, the movement of the interface is not restricted. The smallest amount of movement was observed in the toe of the freshwater-saltwater interface in the Lower Potomac-Raritan-Magothy aquifer in Ocean County. In some areas, the tip and toe of the interface in both the Middle and Upper Potomac-Raritan-Magothy aquifers moved as much as 2.5 mi downdip. At the end of the simulation (1896), the interface was moving landward in all three aquifers in response to the most recent sea-level rise.

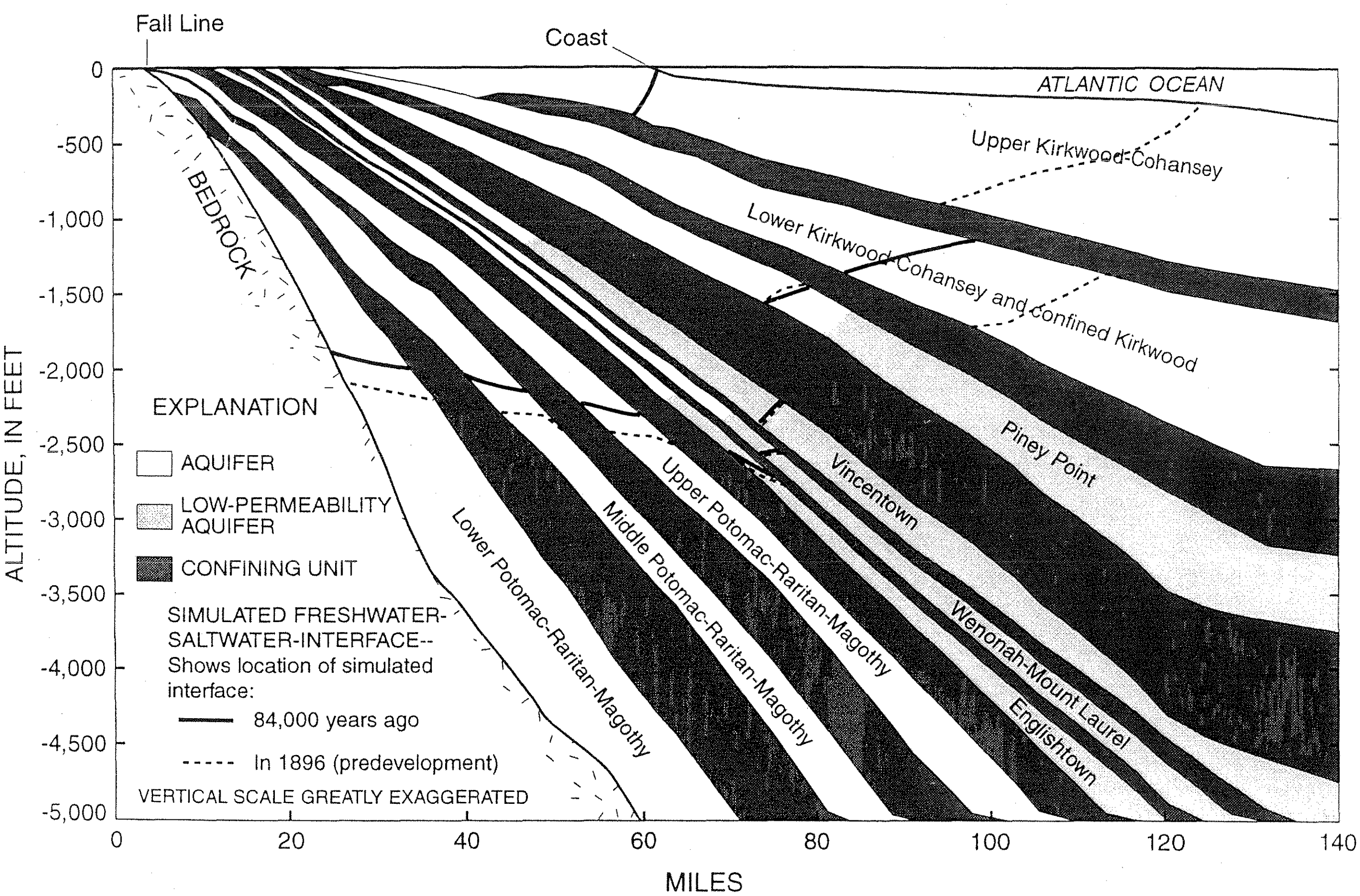


Figure 10. Hydrogeologic section showing simulated locations of the freshwater-saltwater interface in the New Jersey Coastal Plain 84,000 years ago and in 1896 (predevelopment).

During this simulation, the interface in the Wenonah-Mount Laurel aquifer (model unit A5) was relatively static. The freshwater-saltwater interface movement can be simulated only in the south, where the interface is in the permeable part of the aquifer. During the simulation the toe of the interface moved downdip as much as 0.5 mi while the tip of the interface was in the low-permeability sediments; therefore, this result is questionable.

Movement of the interface in the Piney Point aquifer (model unit A7) can be simulated only where the interface is the permeable part of the aquifer, offshore from northern Atlantic County and most of Ocean County. Interface movement in this aquifer is complex and probably is influenced by the proximity of the simulated interface to the low-permeability part of the aquifer. The interface tends to move downdip in the northern part of the aquifer as a result of the sea-level decline. In the area downdip from Long Beach Island, south of the mouth of the Mullica River in Ocean County, the toe of the interface is stationary while the tip moves inland, narrowing the interface in this area.

The tip of the interface in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8) moved 10 to 15 mi downdip during this simulation. Movement was greatest in the southern part of the study area, near the model boundary. The interface generally moved less than 12 mi seaward in areas north of a line extending due east from Atlantic City.

The position of the freshwater-saltwater interface in the upper Kirkwood-Cohansey aquifer (A9) was most affected by changes in sea level. The interface changed from a narrow (generally less than 2-mi-wide), steep interface just off the coast in the steady-state simulation to a much wider (about 12-30 mi wide) interface in 1896. Movement was greatest directly offshore from northern Cape May County, Atlantic County, and southern Ocean County, where the heads in the aquifer were highest and the toe of the interface moved downdip about 15 mi. In areas to the north and south, where simulated heads in 1896 were less than 10 ft above sea level, downdip movement of the toe was much smaller because heads were lower. Heads were lower offshore because the overlying confining unit is absent and the aquifer is unconfined and in connection with the overlying aquifer.

<u>Ground-Water Flow System in 1896</u>

The results of the simulation of the flow system in the New Jersey Coastal Plain ending in 1896 represent predevelopment conditions in the system. The simulated location of the freshwater-saltwater interface, simulated heads in 1896, and overall flow budgets for each aquifer derived from simulation results are described below.

Freshwater flow budgets for the confined part of each aquifer are shown in table 2. A schematic representation of an aquifer showing each of the terms used in the flow budgets in this report is shown in figure 11. Storage in the predevelopment flow system represents movement of the freshwater-saltwater interface; as the interface moves landward, seawater displaces freshwater. The saltwater flow term includes flow from the aquifer outcrop or overlying-confining-unit outcrop in the Delaware or Raritan Bay. This flow term represents freshwater flow from areas that are known to contain saltwater. The model does not simulate inflow of saltwater, but these areas

Table 2. Simulated predevelopment flow budgets of confined aquifers in the New Jersey Coastal Plain, 1896

[Values are in million gallons per day]

	Inflow						
Model unit	Storage	Flow downdip from unconfined aquifer or outcrop	Saltwater	Leakage from overlying unconfined aquifer	Leakage from overlying confined aquifer	Leakage from underlying confined aquifer	Total
Lower Potomac-Raritan-Magothy aquifer (A1)	0	0	0	0	3.4	0	3.4
Middle Potomac-Raritan-Magothy aquifer (A2)	0	3.1	0	5	5.3	2.3	15.7
Upper Potomac-Raritan-Magothy aquifer (A3)	0	1.4	0	6	7.4	.4	15.2
Englishtown aquifer (A4)	0	0	0	5.2	6.6	.6	12.4
Wenonah-Mount Laurel aquifer (A5)	0	.5	0	3.8	7.3	1.6	13.2
Vincentown aquifer (A6)	0	2.1	0	9.2	0	.9	12.2
Piney Point aquifer (A7)	0	0	0	5.3	0	.3	5.6
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8)	1.2	3.2	.1	.1	0	1.4	6.0

	Outflow						
Model unit	Storage	Flow downdip to unconfined aquifer	Saltwater	Leakage to overlying unconfined aquifer	Leakage to overlying confined aquifer	Leakage to underlying confined aquifer	Total
Lower Potomac-Raritan-Magothy aquifer (A1)	0	0.2	0	0.9	2.3	0	3.4
Middle Potomac-Raritan-Magothy aquifer (A2)	0	3	.1	8.8	.4	3.4	15.7
Upper Potomac-Raritan-Magothy aquifer (A3)	0	8.2	.5	.6	.6	5.3	15.2
Englishtown aquifer (A4)	0	2.8	.5	.4	1.6	7.1	12.4
Wenonah-Mount Laurel aquifer (A5)	0	1.9	.1	3.0	1.3	6.9	13.2
Vincentown aquifer (A6)	0	2.5	0	3.4	0	6.3	12.2
Piney Point aquifer (A7)	0	0	.4	3.6	1.4	.2	5.6
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8)	.1	2.6	2.9	.4	0	0	6.0

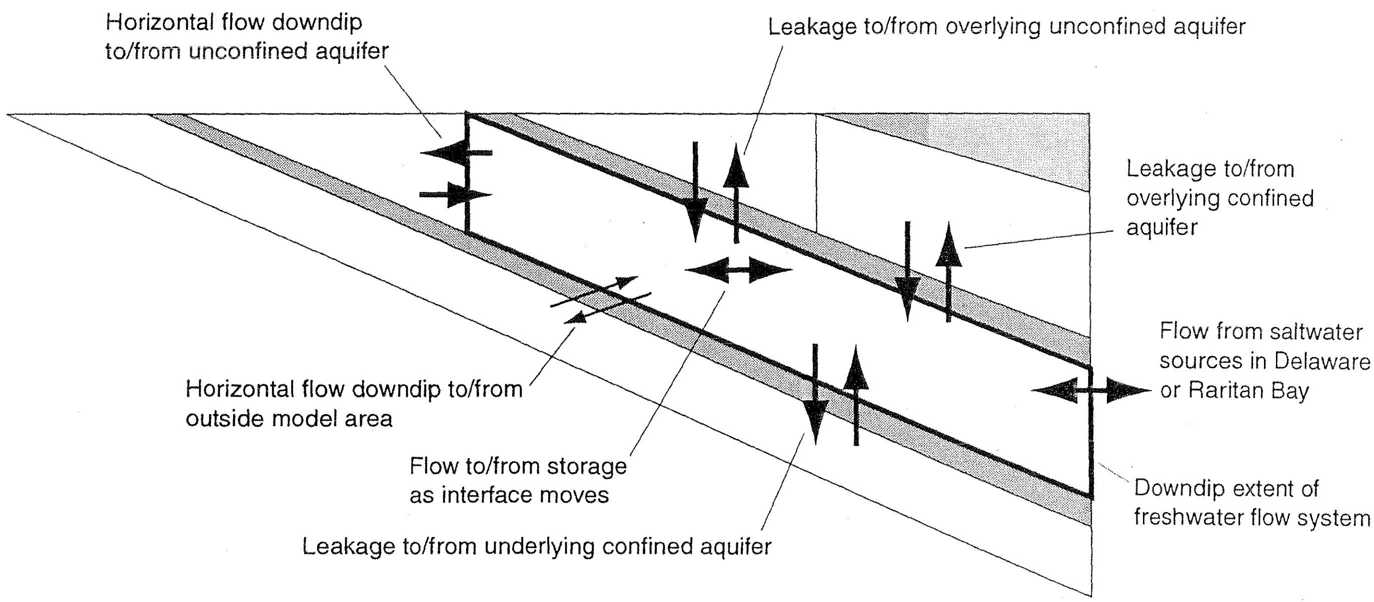


Figure 11. Schematic representation of budget terms used to describe flow budgets in each aquifer in the New Jersey Coastal Plain flow model.

represent potential or real sources of saltwater to the system--for example, the Delaware Bay. The overlying-unconfined-aquifer term includes flow from the outcrop of the overlying unconfined aquifer as well as from the outcrop of an overlying confining unit.

Simulated heads in 1896 in the three Potomac-Raritan-Magothy aquifers are within 10 ft of each other in all areas, indicating that the aquifers behaved as a single hydrologic unit before development. Flow in the Potomac-Raritan-Magothy aquifers is driven by recharge from the outcrop areas of the Middle and Upper aquifers in Middlesex and Mercer Counties. Discharge is to Raritan Bay and the Raritan River to the north and to the Delaware River in the south.

In all three Potomac-Raritan-Magothy aquifers the interface is farther downdip in northern areas and closer to the updip limits of the aquifers in the south. This is especially true in the Lower (A1) and Middle (A2) Potomac-Raritan-Magothy aquifers. High heads near the recharge areas in the north in the Middle and Upper aquifers (A2 and A3) cause the interface in these areas to be located farther downdip than in the southern part of the study area. Because the downdip part of the Potomac-Raritan-Magothy aquifer system functions as a single unit, the interface locations in the three aquifers are related. In the section in figure 8, the interface positions in the Lower, Middle, and Upper aquifers are nearly horizontal, with altitudes between -2,100 and -2,500 ft. In downdip parts of the Middle and Upper Potomac-Raritan-Magothy aquifers, saltwater probably is present in either the underlying aquifer or the underlying confining unit.

The simulated heads and positions of the tip and toe of the freshwater-saltwater interface in the Lower Potomac-Raritan-Magothy aquifer (A1) in 1896 are shown in appendix 1 (fig. 1c). Simulated heads in this aquifer generally are within a few feet of the simulated heads in the Middle Potomac-Raritan-Magothy aquifer. Flow in the Lower aquifer is from the ground-water high in and offshore from Ocean County southeast toward the Delaware River. The flow budget for the Lower Potomac-Raritan-Magothy aquifer in 1896 is shown in table 2. In the predevelopment flow system, this aquifer receives all of its inflow from the overlying Middle Potomac-Raritan-Magothy aquifer in downdip areas. Discharge is to the Middle aquifer offshore from Ocean County, in parts of Burlington County, and updip near the Delaware River and the outcrop.

The freshwater-saltwater interface in the Lower Potomac-Raritan-Magothy aquifer is farthest updip in the southern part of the aquifer, downdip from the discharge area to the Delaware River, where heads in the aquifer are lowest, and is farthest downdip in Ocean County, where heads are highest. Irregularities in the shape of the toe of the interface in the Lower Potomac-Raritan-Magothy aquifer in eastern Burlington and Ocean Counties result from the shape of the bedrock surface (which represents the bottom of the Lower Potomac-Raritan-Magothy aquifer) in this area (Zapecza, 1989, pl 1). The freshwater-saltwater interface narrows offshore from Ocean County, where the aquifer thins.

The simulated heads and the position of the freshwater-saltwater interface in the Middle Potomac-Raritan-Magothy aquifer (A2) in 1896 are shown in appendix 2 (fig. 2c). Predevelopment heads in the Middle aquifer are 40 to 70 ft above sea level in the recharge area in Middlesex and Mercer Counties. Heads downdip from the recharge area are higher than those simulated in the RASA model. Most of the recharge to the Middle Potomac-Raritan-Magothy aquifer (table 2) is vertical flow from the confined and unconfined Upper Potomac-Raritan-Magothy aquifer and horizontal flow from the unconfined part of the Middle aquifer in Middlesex County. Recharge from the underlying Lower Potomac-Raritan-Magothy aquifer occurs in the northeastern part of the study area, near the freshwater-saltwater interface in the Lower aquifer, and updip near the discharge area in the southwestern part. Discharge is to the unconfined part of the Middle Potomac-Raritan-Magothy aquifer, the overlying Upper aquifer near the Delaware River in the southwest, and the overlying Upper aquifer near Raritan Bay.

The interface in the Middle Potomac-Raritan-Magothy aquifer is widest in northern Atlantic, southeastern Gloucester, and eastern Ocean Counties where the aquifer is more than 350 ft thick, and is narrower to the north and south where the aquifer thins. The interface is farthest offshore downdip from the recharge area in the northwestern part of the study area because heads are high in this area.

Simulated heads and the position of the freshwater-saltwater interface in the Upper Potomac-Raritan-Magothy aquifer (A3) in 1896 are shown in appendix 3 (fig. 3c). Heads in the Upper aquifer are 60 to more than 70 ft above sea level in the recharge area in Mercer and Middlesex Counties, and are less than 20 ft below sea level near Raritan Bay and updip near the Delaware River. Most of the recharge to the Upper Potomac-Raritan-Magothy aquifer (table 2) is from overlying areas in the northwestern part of the study area. Slightly more than half (55 percent) of this recharge is from the confined Englishtown aquifer; the remainder is from the unconfined Englishtown aquifer and the simulated outcrop of the Merchantville-Woodbury confining unit. More than half (54 percent) the discharge from the Upper Potomac-Raritan-Magothy aquifer is horizontal flow to the unconfined part of the aquifer near the Delaware River. About one-third of the discharge is flow to the underlying Middle aquifer.

As in the Lower and Middle Potomac-Raritan-Magothy aquifers, the freshwater-saltwater interface in the Upper Potomac-Raritan-Magothy aquifer is farthest offshore downdip from the area of high heads in the recharge area in Middlesex and Monmouth Counties. The interface is farthest offshore in the southern part of the simulated area. The width of the interface ranges from 3 to 5.5 mi in most areas.

The simulated predevelopment flow systems in the Englishtown (A4) and Wenonah-Mount Laurel (A5) aquifers are similar as a result of the high vertical conductivity of the intervening Marshalltown-Wenonah confining unit (the conductance of confining unit is more than an order of magnitude larger than in the Merchantville-Woodbury confining unit (C3)). Heads in the Englishtown and Wenonah-Mount Laurel aquifers are within 10 ft of one another except in updip areas in southern Monmouth, western Ocean, and northern Burlington Counties, where heads in both aquifers are relatively high. Both aquifers are recharged primarily from overlying units. Recharge areas for the aquifers are in Camden and Gloucester Counties in the southern part of the study area and in southern Monmouth County in the northern part, where the heads are more than 120 ft above sea level in both aquifers.

The simulated heads in the Englishtown aquifer (A4) in 1896 are shown in appendix 4 (fig. 4c). Discharge from the aquifer (table 2) is primarily to the underlying Upper Potomac-Raritan-Magothy aquifer (A3) with smaller amounts of lateral flow to the unconfined part of the aquifer and vertical flow to the overlying aquifer. Upward discharge to the Wenonah-Mount Laurel aquifer occurs in updip areas and offshore from western Monmouth County. The Englishtown aquifer contains freshwater throughout its extent.

Simulated heads and the position of the freshwater-saltwater interface in the Wenonah-Mount Laurel aquifer (A5) in 1896 are shown in appendix 5 (fig. 5c). Recharge from the overlying unconfined Vincentown aquifer represents about 29 percent of the total recharge to the Wenonah-Mount Laurel aquifer (table 2). Flow from the overlying confined aquifers includes flow from the overlying confined parts of the Vincentown (A6), Piney Point (A7), and lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8). About half of the discharge from the aquifer is flow to the underlying Englishtown aquifer. Twenty-five percent of the discharge is upward flow to the unconfined Vincentown aquifer. The remainder of the discharge is horizontal flow to the outcrop area and vertical flow to the overlying confined aquifers. The Wenonah-Mount Laurel aquifer contains freshwater throughout most of its extent. The freshwater-saltwater interface is simulated within the Wenonah-Mount Laurel aquifer only in extreme downdip areas in the south-eastern part of the study area, near Cape May County.

The simulated heads in the Vincentown aquifer (A6) in 1896, shown in appendix 6 (fig. 6c), are similar to the simulated heads in the underlying Wenonah-Mount Laurel aquifer. Heads in the two aquifers are within 5 ft of each other except near the coast in Monmouth County. Simulated heads in the Vincentown aquifer are highest in Camden and western Ocean Counties, and are lowest in the areas of discharge to the Delaware River in the south and the outcrop of the Vincentown aquifer offshore from the coast of Monmouth County. About 75 percent of the recharge to the Vincentown aquifer is vertical flow from the overlying unconfined Kirkwood-Cohansey aquifer (table 2). Horizontal flow from the outcrop and vertical flow from the underlying Wenonah-Mount Laurel aquifer provides the rest of the recharge. Discharge is primarily downdip flow to the Wenonah-Mount Laurel aquifer (52 percent of discharge), but also includes flow to the overlying Kirkwood-Cohansey aquifer and horizontal discharge to the outcrop. The Vincentown aquifer contains freshwater throughout its extent.

Simulated heads and the position of the freshwater-saltwater interface in the Piney Point aquifer in 1896 are shown in appendix 7 (fig. 7c). Heads are highest, more than 120 ft above sea level, along the border between Burlington and Ocean Counties where the overlying Kirkwood-Cohansey aquifer recharges the aquifer. Flow is toward the Delaware Bay and the coast in Ocean County, where the water discharges to overlying aquifers. Almost all of the recharge (95 percent) to the Piney Point aquifer is from the overlying lower Kirkwood-Cohansey and confined Kirkwood aquifers. Discharge is primarily (64 percent) to the overlying lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8). Also, 25 percent of the discharge is flow to the overlying confined Kirkwood aquifer (A8) in downdip areas.

The freshwater-saltwater interface in the Piney Point aquifer is from 2 to 4.25 mi wide. The interface is simulated within the permeable part of the Piney Point aquifer only in areas offshore from Ocean County and northern Atlantic County.

Simulated heads and the position of the freshwater-saltwater interface in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8) in 1896 are shown in appendix 8 (fig. 8c). Heads are highest near the updip limit of the confining unit overlying the Rio Grande water-bearing zone (C8), where the confined part of the aquifer receives recharge from the unconfined lower Kirkwood-Cohansey aquifer (53 percent of inflow). Flow from the underlying Piney Point aquifer accounts for about 23 percent of inflow to the confined part of the aquifer (table 2). The remaining source of freshwater recharge to the aquifer is water released from storage. In the sharp-interface approach, as the freshwater-saltwater interface moves inland, freshwater is displaced by encroaching saltwater (the total volume of freshwater has not changed but the area occupied by freshwater in the aquifer is smaller). The storage term in the budget in table 2 represents the freshwater displaced from pores by the encroaching saltwater. Ground-water flow is from the high areas in Atlantic County toward the Delaware Bay where the overlying confining unit is absent offshore. In the predevelopment simulation, about half the outflow from the aquifer is freshwater discharge to lateral or overlying salty parts of the aquifer offshore. Most of the remaining discharge from the confined part of the aquifer is lateral flow to the unconfined part of the aquifer.

Where the overlying confining unit (C9) is present, the freshwater-saltwater interface roughly follows the shape of the coast 30 to 40 mi offshore. The interface is 10 to 15 mi wide in this area.

Near the updip limit of the overlying confining unit (C8) and in areas where the overlying confining unit is not present, south and west of Cape May in the Delaware Bay, and offshore from Ocean County, the toe of the interface is farther onshore than in the rest of the aquifer. As described previously, heads updip from these areas were too low to completely flush the saltwater out of the aquifer as the interface moved downdip. The aquifer in these areas generally contains freshwater with only a small layer of saltwater underlying the freshwater. The thickness of the saltwater layer in these areas is less than 25 ft and represents less than 10 percent of the thickness of the aquifer.

SIMULATION OF GROUND-WATER FLOW AND FRESHWATER-SALTWATER INTERFACE MOVEMENT, 1896-1988

The ground-water flow model was used to simulate the ground-water flow system in the New Jersey Coastal Plain for the period 1896-1988. Initial conditions for the simulation were the results of the 84,000-year simulation of sea-level change described in the previous section. Starting freshwater and saltwater heads and location and altitude of the saltwater interface in each active cell were taken from the simulation of the flow-system response to sea-level changes ending in 1896.

The period from November 1, 1896, through October 31, 1988, was simulated by using seven stress periods ranging in length from 5 to 25 years. Stress periods were selected on the basis of withdrawal data and to coincide with the synoptic water-level data collected every 5 years. The ending dates of the stress periods were changed slightly from those used in the RASA model to coincide more closely with the synoptic measurements collected in the fall of 1978, 1983, and 1988. The lengths of the stress periods and the total boundary flows for each stress period are shown in table 3.

Table 3. Ground-water withdrawals from each aquifer and lateral boundary flows by pumping period, New Jersey Coastal Plain, 1896-1988

[Pumpage is in million gallons per day]

Model unit	Aquifer	Pumping period						
		1 11/01/1896- 10/31/1921	2 11/01/1921- 10/31/1946	3 11/01/1946- 10/31/1958	4 11/01/1958- 10/31/1968	5 11/01/1968- 10/31/1978	6 11/01/1978- 10/31/1983	7 11/01/1983- 10/31/1988
	Potomac Raritan-Magothy aquifer system:							
A1	Lower Potomac-Raritan-Magothy aquifer	8.9	27.1	40.7	54.4	64.2	64.7	63.7
A2	Middle Potomac-Raritan-Magothy aquifer	3.1	15.7	27.4	45.9	69.8	70.3	64.5
A3	Upper Potomac-Raritan-Magothy aquifer	1.6	8.6	16.9	37	51.7	51.2	52.3
A4	Englishtown aquifer	1.7	3.4	5.7	7.7	10.9	9.7	9.1
A5	Wenonah-Mount Laurel aquifer	.5	1.0	1.8	3.6	4.5	4.5	4.9
A6	Vincentown aquifer	0	0	0	.1	.6	1.0	.6
A7	Piney Point aquifer	.1	.3	.6	.9	1.8	1.8	1.8
A8	Lower Kirkwood-Cohansey and confined Kirkwood aquifers	1.5	8.7	13.3	21.1	34	37.6	39
A9	Upper Kirkwood Cohansey aquifer	2.3	10.8	12.2	27	42.1	47	50.4
A10	Holly Beach water-bearing zone	0	0	.1	.1	.1	.2	0
Boundary flow		4.6	3.9	3.9	4.5	4.3	2.5	4.3

Total annual ground-water withdrawals from the New Jersey Coastal Plain aquifers are shown in figure 12 for the seven stress periods. Withdrawal data for 1918-80 (from Zapecza and others, 1987) were used as input for the New Jersey RASA model (Martin, 1998). In this model, these data were supplemented with additional data for 1981-88 from the USGS, New Jersey District, water-use data base (unpublished data available at the U.S. Geological Survey office in West Trenton, N.J.). Average ground-water withdrawals for each stress period by aquifer are shown in table 3.

The withdrawal data generally are limited to wells from which withdrawals were greater than 10,000 gal/d, but also include some lower capacity wells. Although many domestic and irrigation wells were not included, they account for only a small percentage of the total withdrawals and tend to be concentrated in and near the aquifer outcrop areas, where the effects of withdrawals are smallest.

In this section, the ground-water flow system in each aquifer in the New Jersey Coastal Plain and locations of ground-water withdrawals during 1983-88 are described. Observed-potentiometric-surface maps for fall 1988 are used to describe cones of depression, ground-water flow directions, and ground-water gradients. Simulated heads in observation wells are compared with measured heads in 1988 and with average water levels for 1983-88 (table 4, at end of report). Simulated and observed water levels were used together to describe the ground-water flow system. Changes in hydrogeologic properties made during model calibration also are discussed.

Simulated heads at the end of the last stress period represent the period from October 1983 through September 1988. Average pumpage for each well during this period was used during the simulation. Water levels in areas near wells from which ground-water withdrawals changed significantly during this period did not closely match the observed 1988 water levels (Eckel and Walker, 1986). In areas where pumpage was relatively constant during the stress period the simulated heads are compared to heads measured in 1988. In order to quantify average flow conditions, monthly water levels at 91 observation wells were used to determine an average measured water level for 1983-88. Most of these observation wells are not near large withdrawal wells; therefore, water levels in these wells are likely to reflect regional water levels more closely than water levels measured near withdrawal wells. These average measured water levels were compared to simulated water levels. Maps showing the locations of the observation wells for which residuals (the difference between the simulated head at the well and the average measured head at the well during 1983-88 (table 4)) were calculated and locations of selected wells in which chloride concentration was determined are shown by aquifer.

Simulated heads near large cones of depression may not match observed water levels for several reasons. Simulation of cones of depression with very steep gradients is difficult because the grid cells used in the model are large. The withdrawals in the model are distributed over the entire area of the model cell rather than at a single point. Water flows into this cell across all six faces; this combined surface is much larger than the flow surface that exists around a well field. In addition, some of the 1988 synoptic water-level measurements were made at pumped wells (in which water levels may not have recovered fully before the measurements were made) and may reflect local drawdown due to withdrawals rather than regional water levels. Measurements made in observation wells a small distance from the pumping centers provide the best estimates of regional water levels.

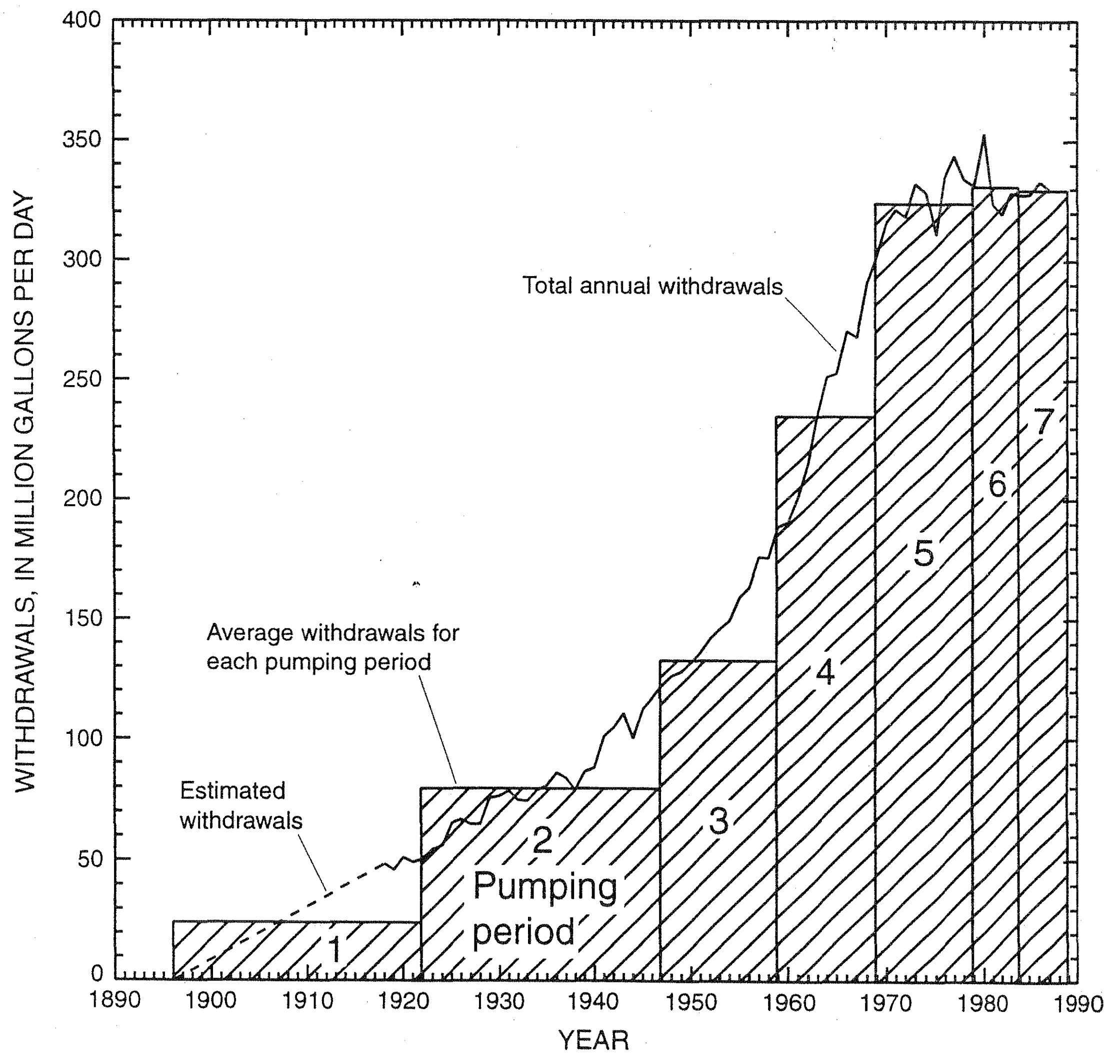


Figure 12. Annual ground-water withdrawals in the New Jersey Coastal Plain from 1917 to 1988 and during model pumping periods.

Description of Ground-Water Flow System, 1983-88

Simulated and measured hydrographs for selected wells completed in the Lower and Middle Potomac-Raritan-Magothy aquifers (A1 and A2), the Upper Potomac-Raritan-Magothy aquifer and the Englishtown aquifer (A3 and A4), and the Wenonah-Mount Laurel aquifer and the confined Kirkwood aquifer (A5 and A8) are shown in figures 13 through 15, respectively. Because of the length of the stress periods and the use of average annual withdrawal data, seasonal variations observed in the measured hydrographs for some wells are not represented in the simulated hydrographs. Both simulated and measured water levels in well 7-412 in the Lower Potomac-Raritan-Magothy aquifer (fig. 13) and in well 7-477 in the Upper Potomac-Raritan-Magothy aquifer (fig. 14) declined approximately 50 ft from 1965 to 1988. Water levels in well 23-194 in the Middle Potomac-Raritan-Magothy aquifer (fig. 13) declined approximately 25 ft from 1965 to 1988. Simulated water levels in well 29-138 in the Englishtown aquifer (fig. 14) are about 10 ft lower than measured water levels. Water levels in this well declined approximately 35 ft from about 1965 to 1988. Simulated water levels in well 7-478 in the Wenonah-Mount Laurel aquifer (fig. 15) are about 5 ft higher than measured water levels. The simulated hydrograph for well 1-180 in the confined Kirkwood aquifer (fig. 15) matched measured water levels from 1975 to 1988, but simulated water levels were higher than measured water levels from 1960 to 1975.

Flow budgets for the confined part of each aquifer were calculated for the period 1983-88 (table 5). Storage is important during the simulation of the stressed ground-water flow system from 1896 to 1988 because, as the freshwater-saltwater interface begins to move inland in response to steepening gradients caused by withdrawals, the encroaching saltwater displaces the freshwater in downdip parts of the aquifer near the interface. This freshwater released from storage is particularly important in the flow budgets for the Potomac-Raritan-Magothy aquifer system.

Potomac-Raritan-Magothy Aquifer System

Heads in the three Potomac-Raritan-Magothy aquifers were similar in 1988, especially near the large regional cone of depression in Camden, Gloucester, and southwestern Burlington Counties. Simulated heads in the Lower and Middle aquifers were within 5 ft of each other except in updip areas and at the centers of cones of depression, where heads generally were within 10 ft of each other. Vertical gradients between the Middle and Upper aquifers were smallest in the southern part of the study area and largest in Ocean, Monmouth, and Middlesex Counties to the north. The flow system in each aquifer is described first by means of simulated and observed water levels. Changes made to hydrogeologic properties in all of the Potomac-Raritan-Magothy aquifers during model calibration are discussed together because similar changes were made to all three aquifers. Withdrawals from the three Potomac-Raritan-Magothy aquifers represent 67 percent of all withdrawals simulated in the model.

Of the three aquifers, the change in the representation of the downdip boundary from a no-flow boundary to a sharp freshwater-saltwater interface most greatly affected the flow system in the Lower Potomac-Raritan-Magothy aquifer. This is because the large cones of depression in the Lower aquifer are near the freshwater-saltwater interface.

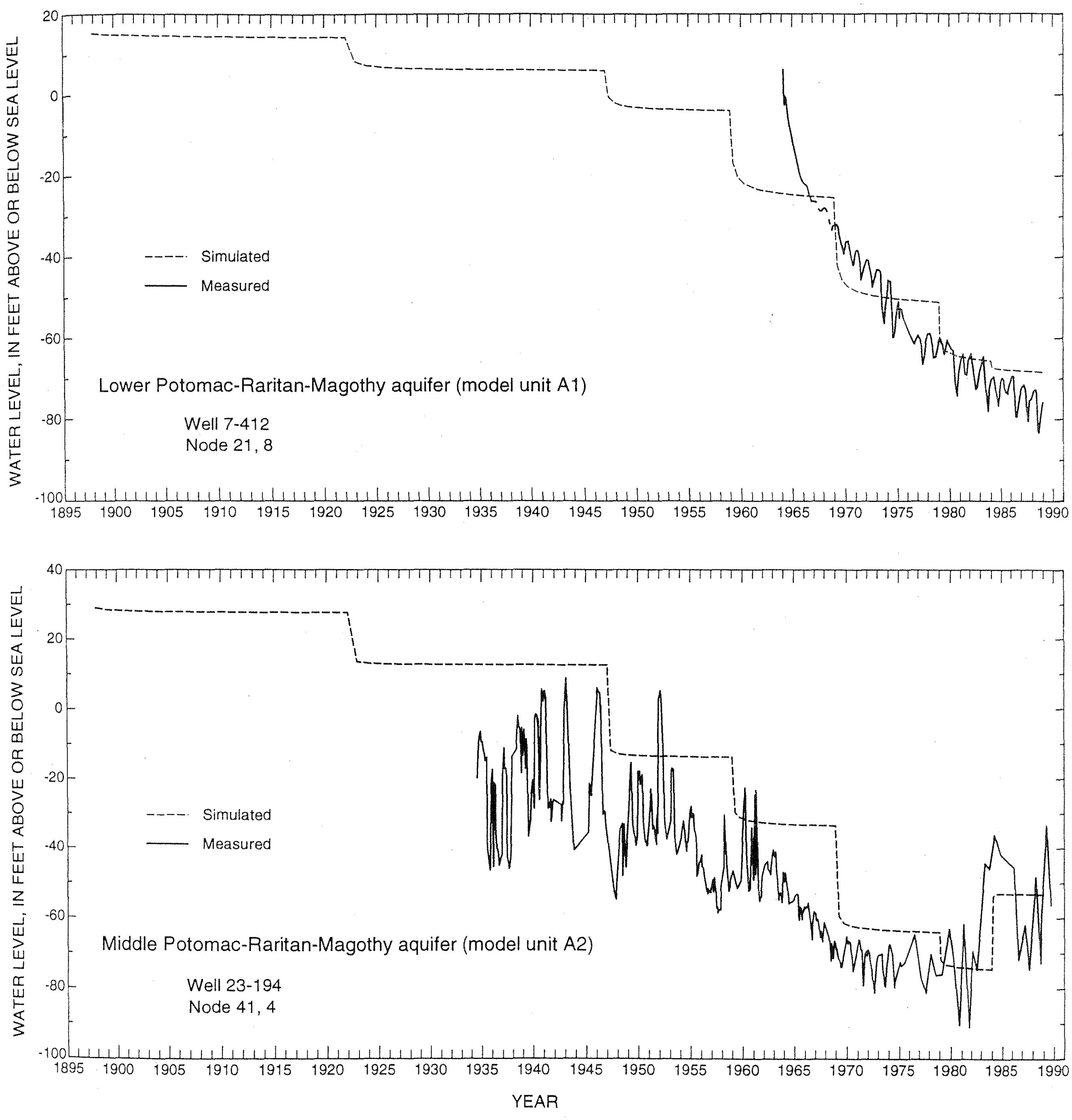


Figure 13. Simulated and measured water levels in the Lower and Middle Potomac-Raritan-Magothy aquifers, New Jersey Coastal Plain, 1896-1989. (Well locations shown in figs.16 and 18, respectively)

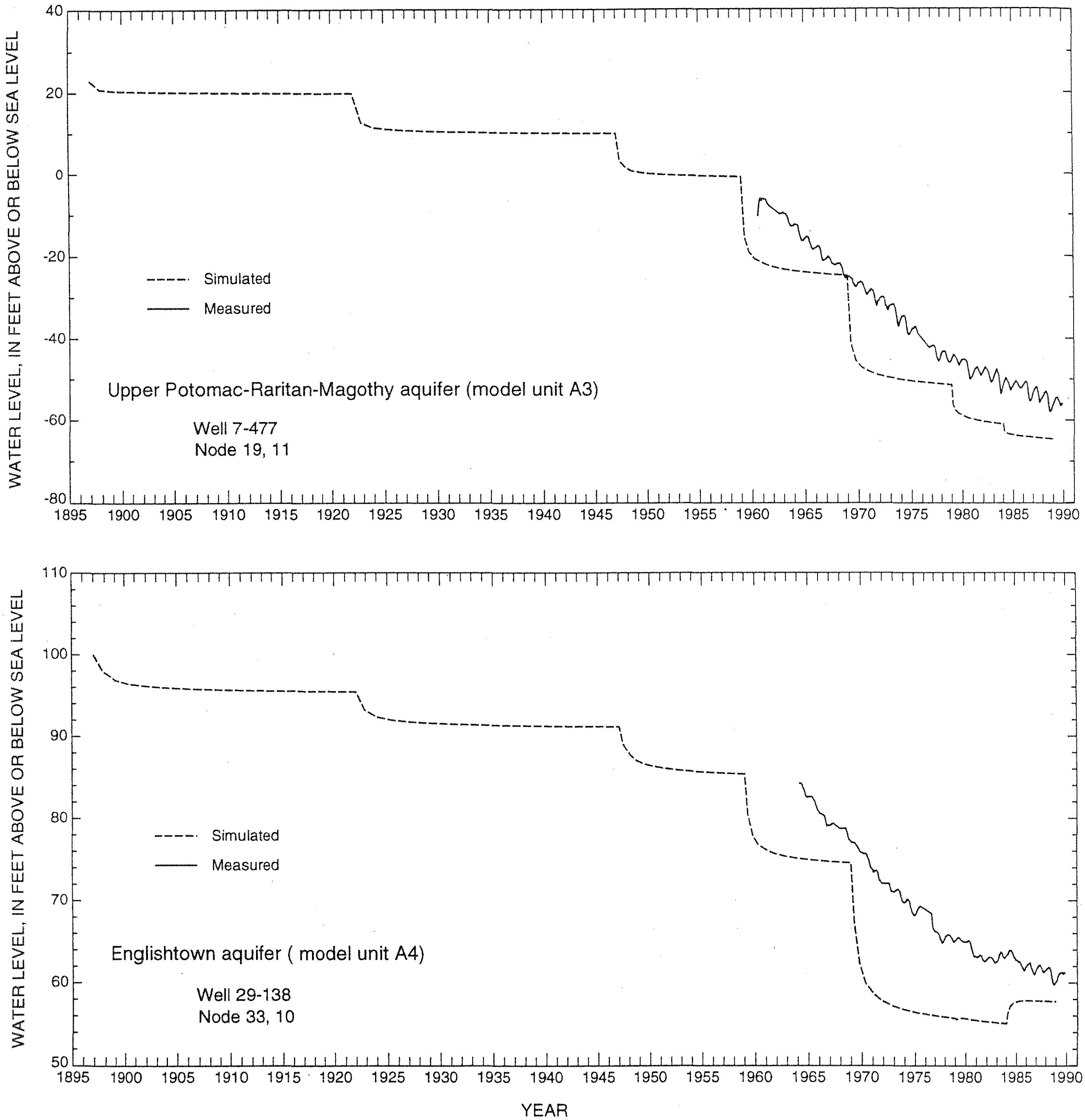
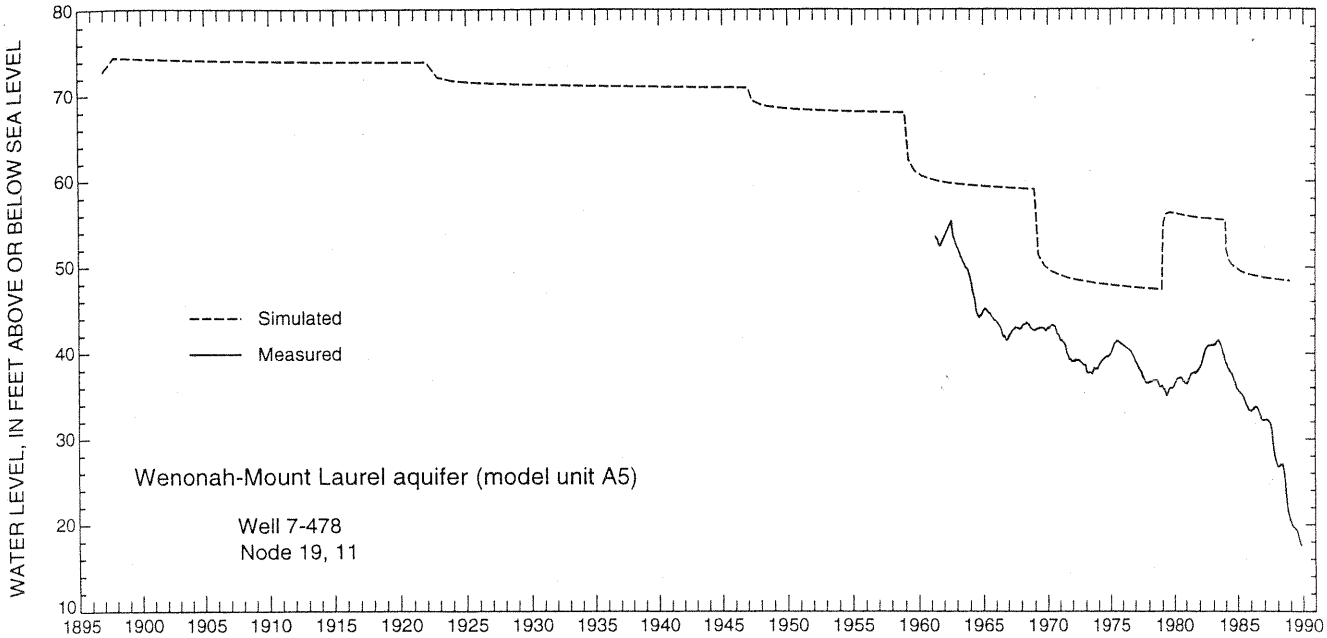


Figure 14. Simulated and measured water levels in the Upper Potomac-Raritan-Magothy and Englishtown aquifers, New Jersey Coastal Plain, 1896-1989. (Well locations shown in figs. 20 and 22, respectively)



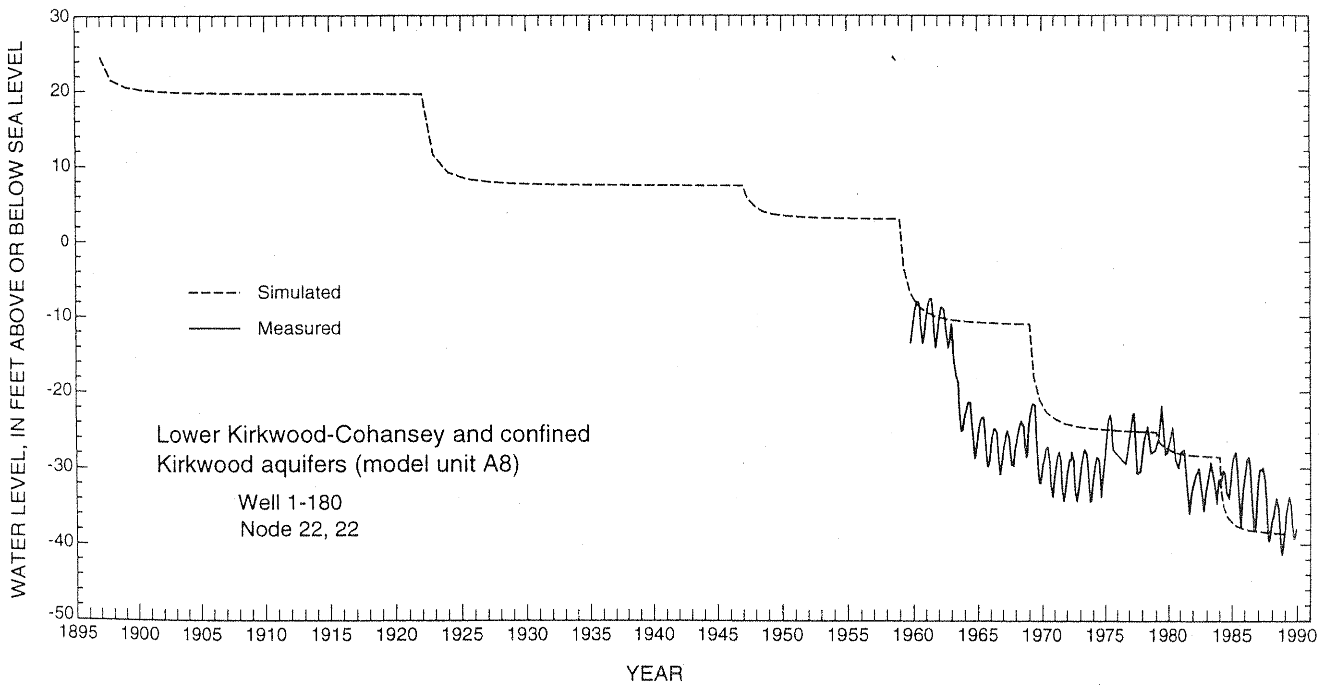


Figure 15. Simulated and measured water levels in the Wenonah-Mount Laurel aquifer and the lower Kirkwood-Cohansey and confined Kirkwood aquifers, New Jersey Coastal Plain, 1896-1989. (Well locations shown in figs. 24 and 30, respectively)

Table 5. Simulated flow budgets of confined aquifers in the New Jersey Coastal Plain, 1983-88

[Values are in million gallons per day]

	Inflow							
Model unit	Storage	Flow downdip from unconfined aquifer	Saltwater	Leakage from overlying unconfined aquifer	Leakage from overlying confined aquifer	Leakage from underlying confined aquifer	Flow across model boundary	Total
Lower Potomac-Raritan-Magothy aquifer (A1)	5.2	2	0	2.6	62	0	0.4	72.2
Middle Potomac-Raritan-Magothy aquifer (A2)	2.0	15.9	.1	93	16.9	3.4	2.7	134
Upper Potomac-Raritan-Magothy aquifer (A3)	1.4	25.3	.5	18.6	18.1	3.6	2.7	70.2
Englishtown aquifer (A4)	0	.1	.2	9.1	18.9	.3	.4	29
Wenonah-Mount Laurel aquifer (A5)	.1	.6	.2	7.4	18.2	.2	.1	26.8
Vincentown aquifer (A6)	0	3.0	.1	16.3	0	.1	0	19.5
Piney Point aquifer (A7)	.4	0	1.0	7.9	0	.8	0	10.1
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8)	2.2	13.8	2.3	1.6	.1	2.7	.8	23.5

	Outflow								
Model unit	Storage	Flow downdip to unconfined aquifer	Saltwater	Leakage to overlying unconfined aquifer	Leakage to overlying confined aquifer	Leakage to underlying confined aquifer	Flow across model boundary	Pumpage	Total
Lower Potomac-Raritan-Magothy aquifer (A1)	0	0	0	0	3.4	0	5.1	63.7	72.2
Middle Potomac-Raritan-Magothy aquifer (A2)	0	.1	0	.3	3.6	62	3.5	64.5	134
Upper Potomac-Raritan-Magothy aquifer (A3)	0	.3	0	0	.3	16.9	.4	52.3	70.2
Englishtown aquifer (A4)	0	1.9	.2	0	.1	17.7	0	9.1	29
Wenonah-Mount Laurel aquifer (A5)	0	1.4	0	.6	.7	19.1	.1	4.9	26.8
Vincentown aquifer (A6)	0	1.3	.1	1.5	0	16.1	0	.6	19.5
Piney Point aquifer (A7)	.1	0	0	1.3	2.7	.6	3.6	1.8	10.1
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8)	.1	.2	.4	0	0	0	.5	22.3	23.5

Inflow from storage near the freshwater-saltwater interface accounted for 7 percent of the water coming into the Lower Potomac-Raritan-Magothy aquifer. Most of this water was released from storage downdip from the two cones of depression. Rates of flow from storage were largest downdip from the cone of depression in Ocean County. This additional source of water initially resulted in simulated heads in the Lower Potomac-Raritan-Magothy aquifer that were 5 to 10 ft above the simulated heads in the RASA model. In the RASA model, this water was supplied to the withdrawal wells as flow from the overlying Middle Potomac-Raritan-Magothy aquifer. During model calibration, leakance values for the confining units overlying the Lower and Middle Potomac-Raritan-Magothy aquifers (model units C1 and C2) near the large cone of depression in Camden County were reduced by as much as 50 percent from leakances used in the RASA model. Because of the interconnection of the three Potomac-Raritan-Magothy aquifers, this additional water supplied from storage as a result of the movement of the interface in the Lower Potomac-Raritan-Magothy aquifer affected heads over large areas in the overlying Middle and Upper Potomac-Raritan-Magothy aquifers (model units A2 and A3) as well.

Inflows from storage to the Middle and Upper Potomac-Raritan-Magothy aquifers were similar, but less significant. Because changes in storage due to the movement of the freshwater-saltwater interface in the Potomac-Raritan-Magothy aquifer system (primarily from the Lower Potomac-Raritan-Magothy aquifer) were important, flow from overlying aquifers was reduced. In general, leakances over large areas of the confining units overlying the aquifers of the Potomac-Raritan-Magothy aquifer system were reduced (about 75 percent) during model calibration.

Leakances of the Merchantville-Woodbury confining unit (model unit C3) in downdip areas were reduced significantly during model calibration to supply less water to the Potomac-Raritan-Magothy aquifer system in general. The isolation between the Potomac-Raritan-Magothy aquifer system and the overlying Englishtown and Wenonah-Mount Laurel aquifers is consistent with Zapecza's (1989, pl. 12) description of the Merchantville-Woodbury confining unit.

Transmissivities in the Lower Potomac-Raritan-Magothy aquifer were reduced near the cone of depression in Camden County during model calibration so that hydraulic conductivities were consistent with values in adjacent areas. Similar reductions in transmissivity were made in the same area in the Middle Potomac-Raritan-Magothy aquifer.

Lower Potomac-Raritan-Magothy aquifer

Locations of withdrawal wells in the Lower Potomac-Raritan-Magothy aquifer used in the model for the period 1983-88 are shown in appendix 1 (fig. 1d). Wells with average annual pumpage greater than 5 Mgal/d (347 gal/min) represent the largest purveyors. Withdrawals are primarily from updip areas near the Delaware River in Camden (73 percent), northwestern Gloucester (12 percent), and southwestern Burlington Counties (9 percent).

The observed potentiometric surface and the ground-water flow direction in the Lower Potomac-Raritan-Magothy aquifer in fall 1988 are shown in figure 16. Ground water in this aquifer flows toward the large cone of depression centered in Camden County and toward a less areally extensive cone along the Delaware River in Salem County. Water levels in large areas of Camden County are 40 to more than 90 ft below sea level. Flow directions have reversed from the prede-

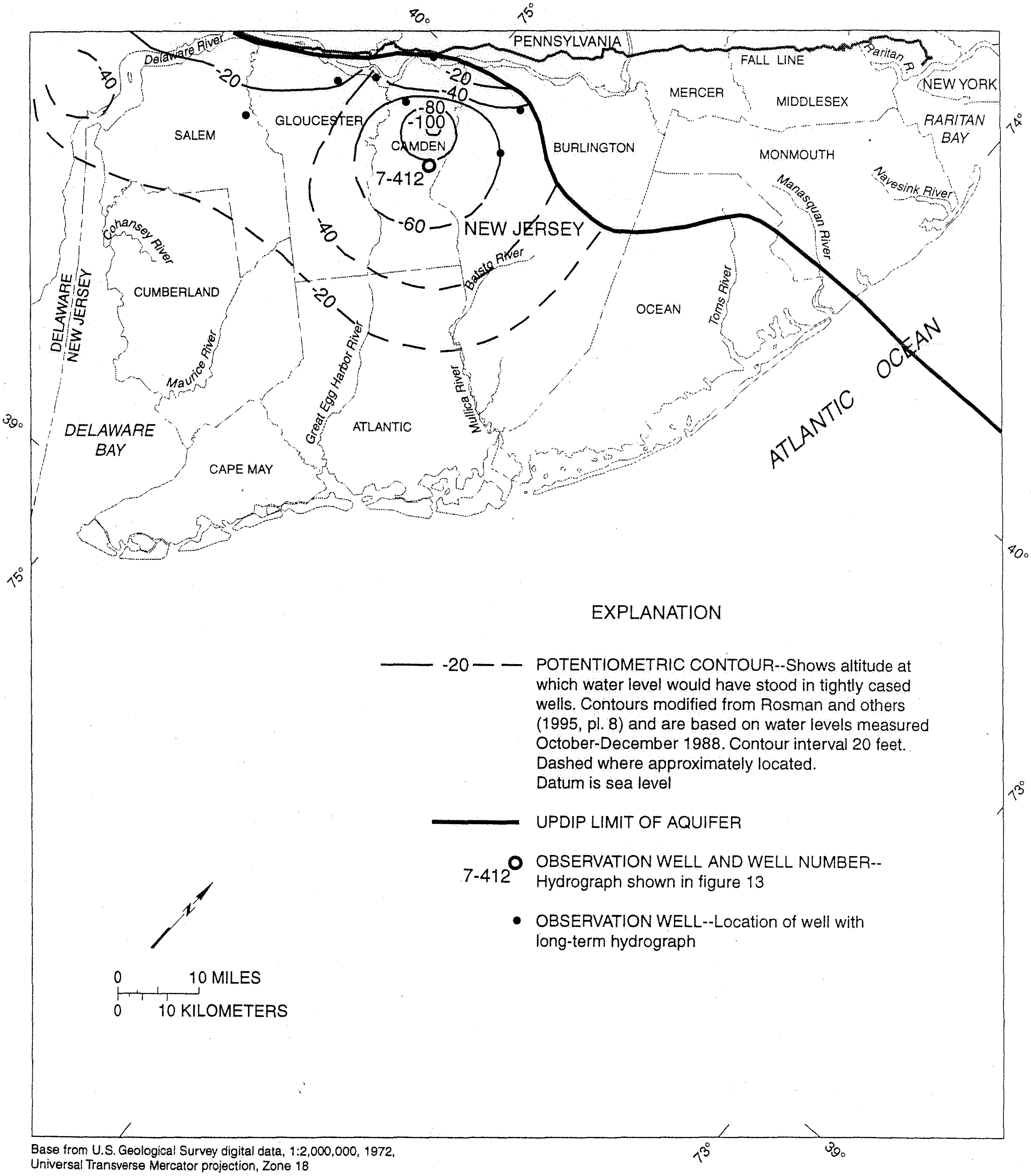


Figure 16. Observed potentiometric surface and locations of selected wells in the Lower Potomac-Raritan-Magothy aquifer (model unit A1) with long-term hydrographs, New Jersey Coastal Plain, 1988.

velopment flow system. In this simulation, water flows from the Delaware River to withdrawal wells instead of discharging to the Delaware River. The location of the tip and toe of the freshwater-saltwater interface in the Lower Potomac-Raritan-Magothy aquifer and residuals at eight observation wells are shown in figure 17. Simulated heads are within 10 ft of the measured heads except at the center of the cone of depression in Camden County, where the simulated heads are more than 10 ft higher than measured heads, and in Salem County near the Delaware River, where the small cone of depression observed in the potentiometric-surface map is not simulated. Residuals at the observation wells are less than 12 ft except at two wells updip near the Delaware River. Gradients near the river are not well simulated because the grid cells in this area are large.

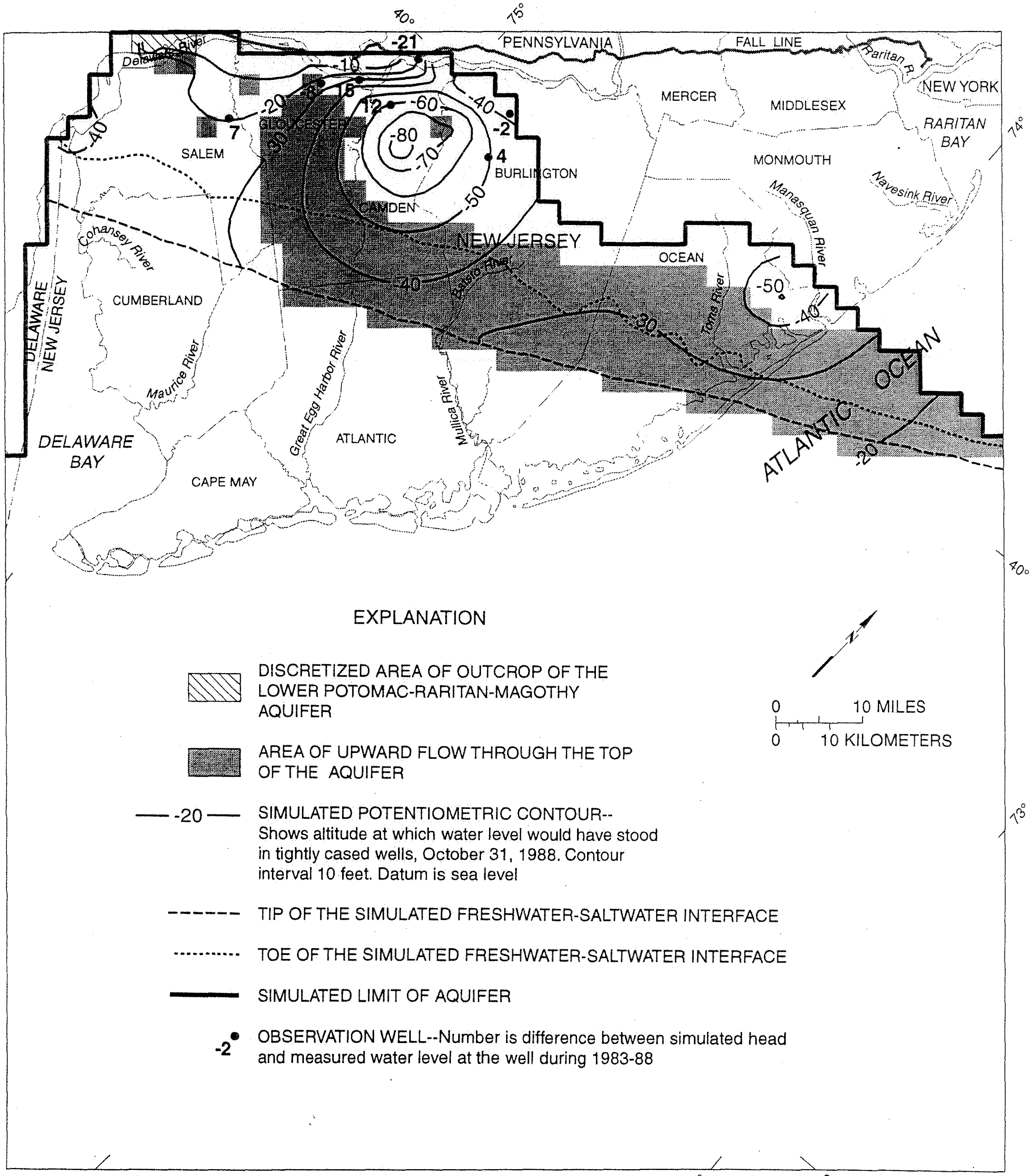
A cone of depression was simulated downdip in northeastern Ocean County, where head measurements for 1988 are not available. Simulated freshwater heads between the tip and toe of the freshwater-saltwater interface are more than 30 ft below sea level except in most of Cumberland and Salem Counties in the south. Low freshwater heads near the interface allow the freshwater-saltwater interface to move landward.

In the flow budget for the confined part of the Lower Potomac-Raritan-Magothy aquifer (table 5), most of the recharge to the updip part of the aquifer occurs as flow from the overlying Middle Potomac-Raritan-Magothy aquifer in Camden County and, to a lesser degree, in Burlington and Gloucester Counties. This vertical flow represents 86 percent of the recharge to the Lower Potomac-Raritan-Magothy aquifer. Inflow from storage occurs downdip as a result of movement of the freshwater-saltwater interface. Discharge from the aquifer is primarily through withdrawal wells. Vertical flow to the overlying aquifer occurs in Gloucester County near the cone of depression and downdip near the freshwater-saltwater interface.

Middle Potomac-Raritan-Magothy aquifer

Ground-water flow in the Middle Potomac-Raritan-Magothy aquifer (model unit A2) also has been altered significantly as a result of ground-water withdrawals. Locations of withdrawal wells for 1983-88 are shown in appendix 2 (fig. 2d). Withdrawals are primarily from updip areas along the Delaware River in Burlington (30 percent), Gloucester (9 percent), and Camden (9 percent) Counties.

The observed-potentiometric-surface map of the Middle Potomac-Raritan-Magothy aquifer (fig. 18) shows that areas along the Delaware River in the southern part of the study area and Raritan Bay in the northern part that were discharge areas in the predevelopment flow system are now recharge areas. Several cones of depression have developed; some are deep regional cones that affect the entire Potomac-Raritan-Magothy aquifer system and others are smaller cones due to local ground-water withdrawals. The large cone of depression centered in Camden County is present in all three Potomac-Raritan-Magothy aquifers. Gradients are much steeper updip from the cone, toward the Delaware River, than downdip. The cone of depression in eastern Middlesex and western Monmouth Counties is also present in the Upper Potomac-Raritan-Magothy aquifer, although water levels in the Middle Potomac-Raritan-Magothy aquifer are much lower. Water levels in the local cone in northeastern Ocean County declined from 30 to 40 ft below sea level in 1983 to more than 70 ft below sea level in 1988.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 17. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, difference between simulated and measured water levels at observation wells, and area of upward flow in the Lower Potomac-Raritan-Magothy aquifer (model unit A1), New Jersey Coastal Plain, 1988.

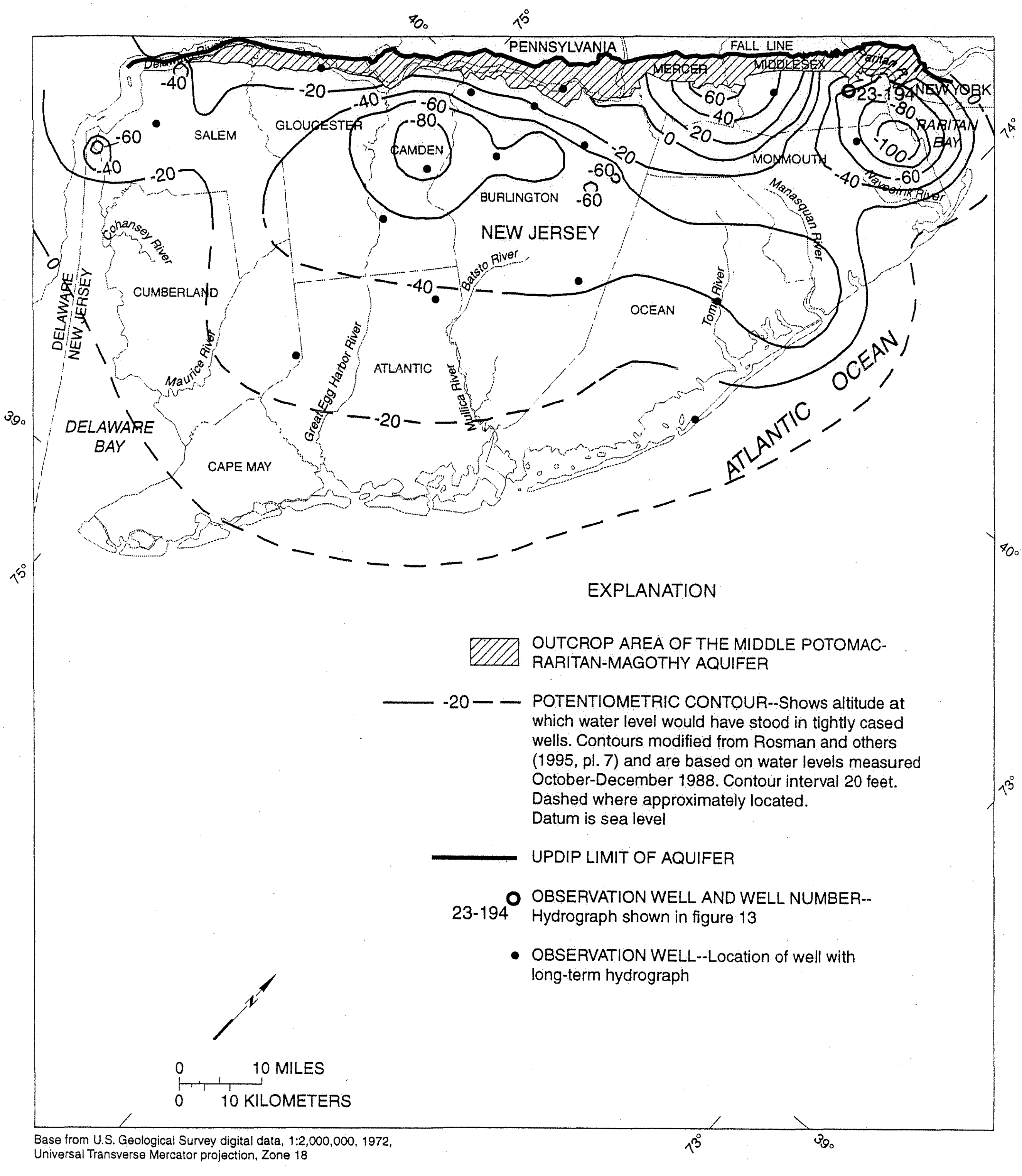
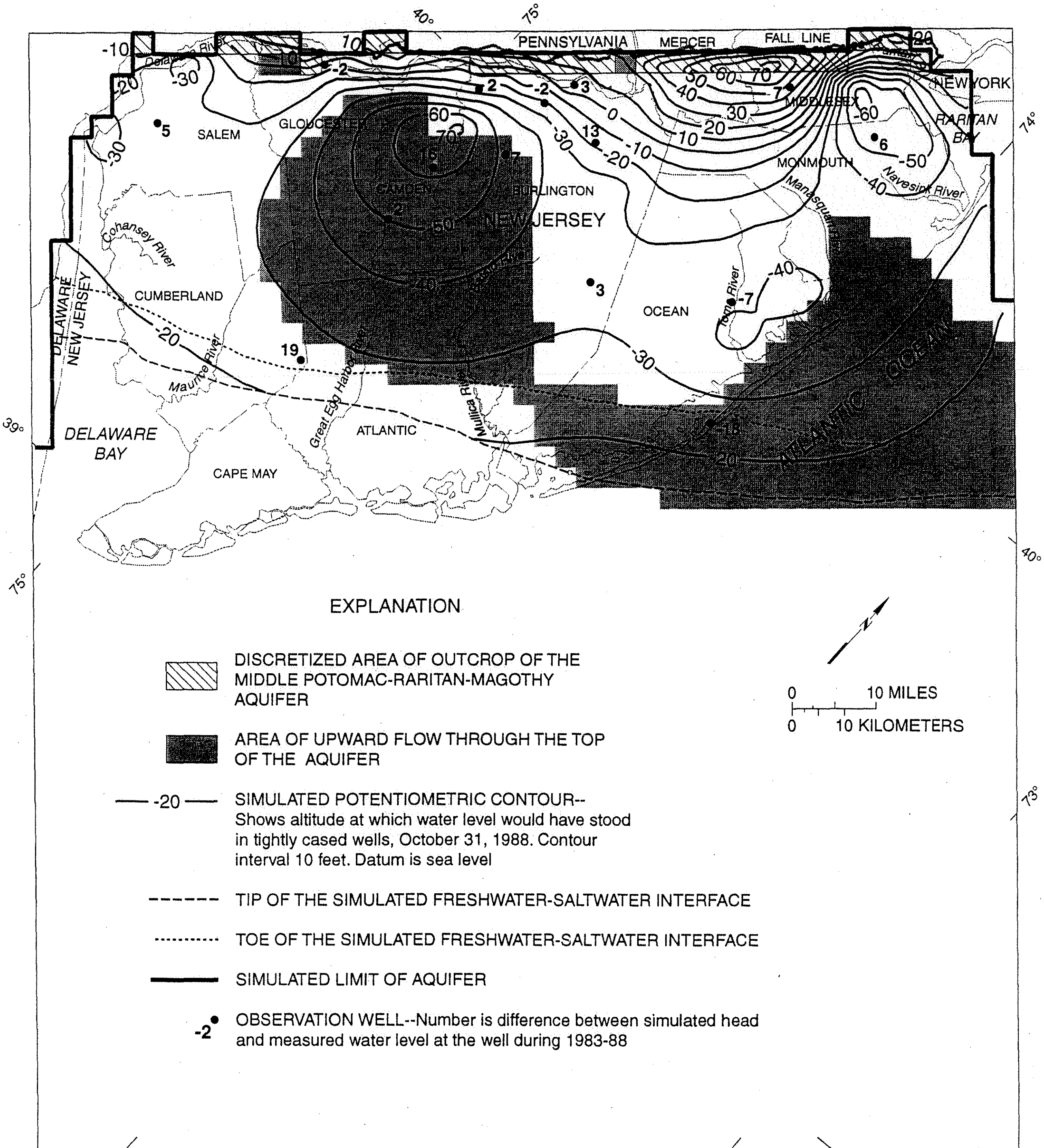


Figure 18. Observed potentiometric surface and locations of selected wells in the Middle Potomac-Raritan-Magothy aquifer (model unit A2) with long-term hydrographs, New Jersey Coastal Plain, 1988.

Simulated heads and the freshwater-saltwater interface tip and toe in the Middle Potomac-Raritan-Magothy aquifer in 1988 are shown in figure 19. Residuals between simulated heads in 1988 and the average measured heads for 1983-88 at 17 selected wells also are shown. Simulated heads at observation wells are within 10 ft of average measured heads except in some wells near the cone of depression in Camden County and in downdip areas. Simulated water levels are similar to observed water levels in 1988 in most areas. The simulated centers of the large regional cones of depression are 10 to 15 ft above the observed water levels. Water levels in observation wells downdip near the interface are not well simulated. Also, local cones of depression in the observed potentiometric surface were not simulated as a result of the scale of the model; however, the general flow directions and gradients in the Middle Potomac-Raritan-Magothy aquifer are well simulated. The deepest part of the cone of depression in northwestern Monmouth County near the coast could not be simulated by using reasonable values of hydrogeologic properties. Because water levels at the center of the cone were measured in a withdrawal well and may not have recovered completely, the measured water level may represent local conditions rather than the regional flow system. The simulated cone of depression along the coast in northern Ocean County is 10 to 20 ft higher than the observed water levels in 1988, but observed water levels in 1983 were only 30 to 40 ft below sea level in this area. The simulated heads are high in this area because ground-water withdrawals in the early part of the stress period were less than withdrawals at the end of the stress period. The simulated heads at observation well 33-251 west of the cone are within about 5 ft of the average water levels at the well. Simulated heads in the recharge area in Mercer and Middlesex Counties are about 10 ft higher than measured water levels.

The major recharge areas in the Middle Potomac-Raritan-Magothy aquifer are the ground-water high in Mercer County and southern Middlesex County and parts of the aquifer along the Delaware River in the southern part of the study area. Ground-water flow directions near the Delaware River in this area have reversed from those in the predevelopment flow system.

The flow budget for the confined part of the Middle Potomac-Raritan-Magothy aquifer (table 5) shows that recharge occurs primarily by vertical flow from the outcrop of the Upper Potomac-Raritan-Magothy aquifer and the confining unit overlying the Middle Potomac-Raritan-Magothy aquifer (69 percent). Most of this flow is near the Delaware River in the southern part of the study area, and two-thirds of this water flows downward to the Lower Potomac-Raritan-Magothy aquifer. This inflow represents water contributed from the Delaware River. Horizontal flow from the outcrop represents about 12 percent of inflow to the Middle Potomac-Raritan-Magothy aquifer and occurs in updip areas. Flow from the overlying Upper Potomac-Raritan-Magothy aquifer represents 13 percent of inflow and is greatest in updip areas in Gloucester, Camden, and Burlington Counties. Ground-water withdrawals account for 48 percent of the outflow from the aquifer and vertical flow to the underlying Lower Potomac-Raritan-Magothy aquifer accounts for 46 percent of the outflow. Discharge to the overlying confined aquifer occurs near the cone of depression in the Upper Potomac-Raritan-Magothy aquifer in Camden County and downdip offshore from Ocean and Monmouth Counties.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 19. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, difference between simulated and measured water levels at observation wells, and area of upward flow in the Middle Potomac-Raritan-Magothy aquifer (model unit A2), New Jersey Coastal Plain, 1988.

Upper Potomac-Raritan-Magothy aquifer

Locations of withdrawal wells in the Upper Potomac-Raritan-Magothy aquifer during 1983-88 are shown in appendix 3 (fig. 3d). Withdrawals from this aquifer represent 25 percent of the withdrawals from Coastal Plain aquifers. More than half (53 percent) of the ground-water withdrawals for the period are from wells in Middlesex and Monmouth Counties.

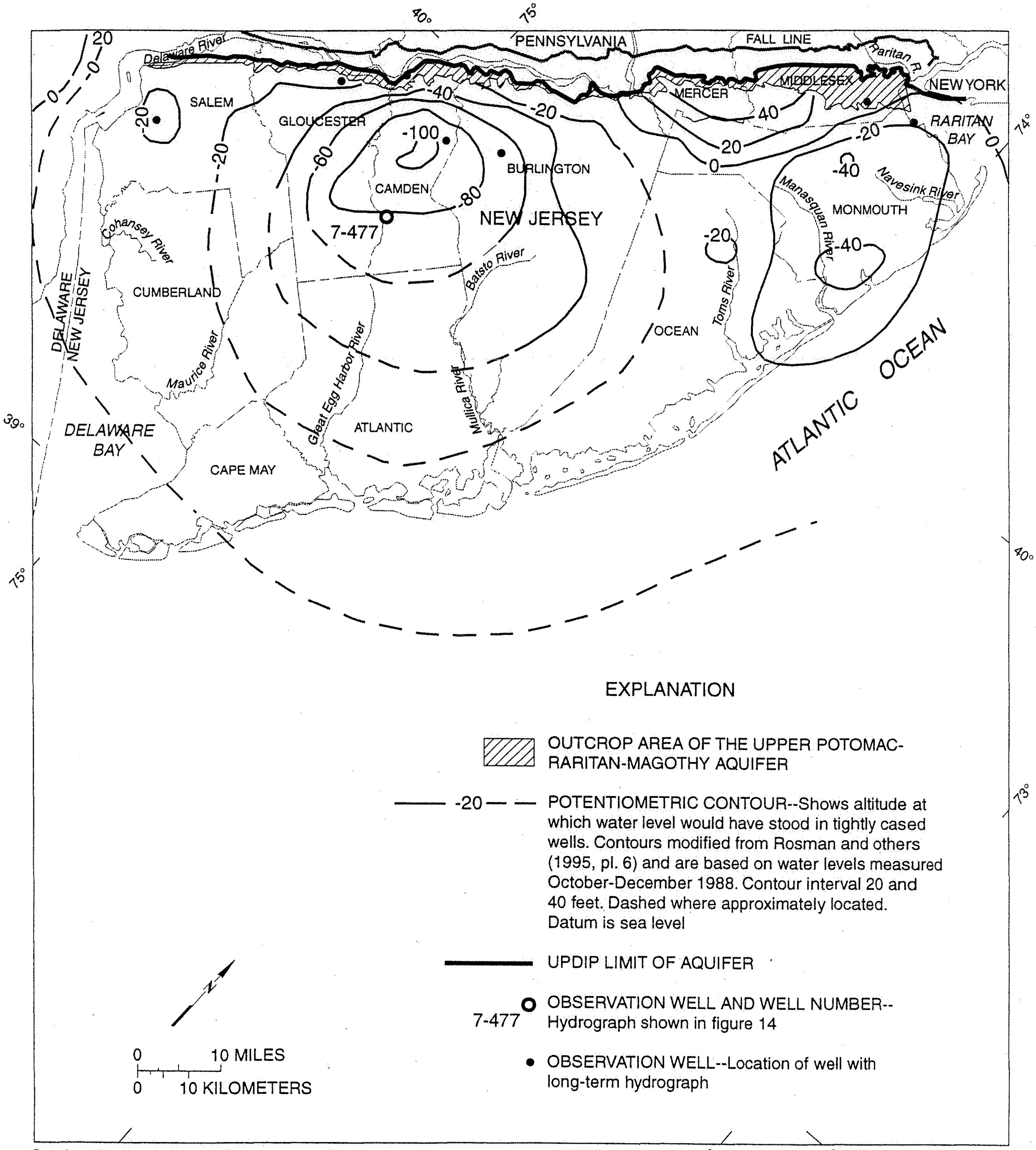
Observed water levels in the Upper Potomac-Raritan-Magothy aquifer in fall 1988 (fig. 20) generally are lower than those in fall 1983 (Eckel and Walker, 1986) at the beginning of the stress period. Water levels are more than 100 ft below sea level at the center of the cone of depression in Camden and Gloucester Counties, and in southwestern Burlington County. The second regional cone of depression occupies most of Monmouth County and parts of northern Ocean County. Water levels generally are 20 to 40 ft below sea level in this area.

Simulated 1983-88 heads in the Upper Potomac-Raritan-Magothy aquifer are shown in figure 21. The simulated location of the tip and toe of the freshwater-saltwater interface and residuals at eight observation wells also are shown. The major observed cones of depression are well simulated and flow directions and gradients generally are consistent with heads in fall 1988. Differences between simulated and observed heads are largest near the centers of the cones of depression in Camden, Burlington, and Gloucester Counties, and near the pumping center in northeastern Monmouth County. Simulated gradients in the recharge area in Mercer and Middlesex Counties are similar to measured gradients in 1988; however, simulated heads are more than 10 ft above the 1988 measured water levels. Observed water levels in this area were higher in 1983 than in 1988; therefore, average conditions for the period are higher than shown in the 1988 water-level map. Heads in this area generally are within 10 ft of the observed water levels in 1983.

Average observed water levels for 1983-88 were within 13 ft of the simulated heads at all eight observation wells. The best matches were in updip areas and in northern Middlesex and Monmouth Counties. Differences were largest near the cone of depression in Camden County and in southern Salem County.

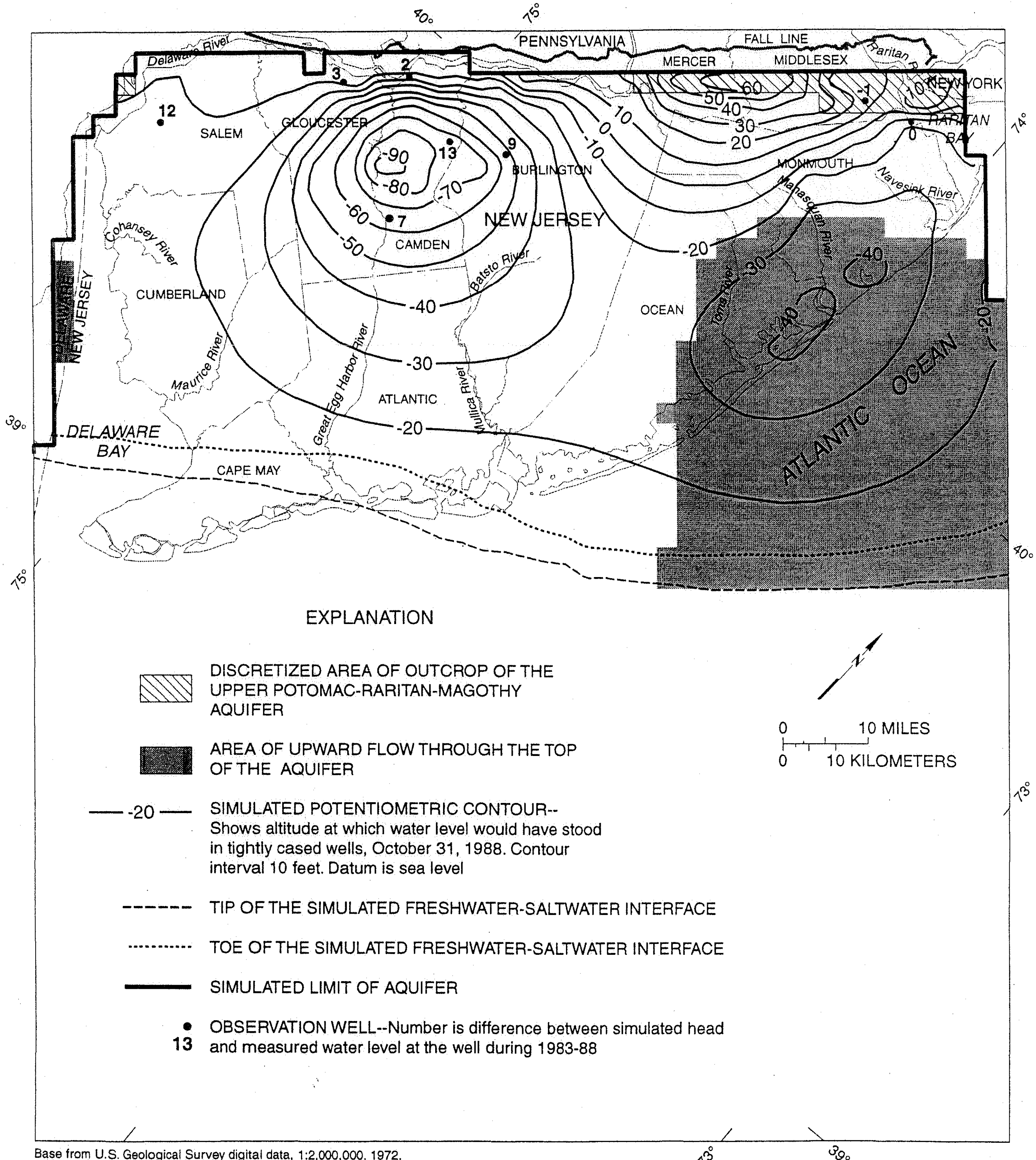
Simulated heads in the Upper Potomac-Raritan-Magothy aquifer generally are lower than heads in the overlying Englishtown and Wenonah-Mount Laurel aquifers except downdip in Monmouth and Ocean Counties, where water discharges from the Upper Potomac-Raritan-Magothy aquifer to the Englishtown aquifer. This area coincides with the location of a large cone of depression in the Englishtown aquifer.

In the flow budget for the confined part of the Upper Potomac-Raritan-Magothy aquifer for 1983-88 (table 5), about 36 percent of the recharge to the aquifer is horizontal flow from the unconfined part of the aquifer. Vertical flow from the overlying unconfined Englishtown aquifer and the outcrop of the Merchantville-Woodbury confining unit and vertical flow from the confined Englishtown aquifer (primarily in updip areas) each provide 26 percent of the inflow to the Upper Potomac-Raritan-Magothy aquifer. Vertical flow from the underlying Middle Potomac-Raritan-Magothy aquifer is greatest near the center of the cone of depression in Camden County.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 20. Observed potentiometric surface and locations of selected wells in the Upper Potomac-Raritan-Magothy aquifer (model unit A3) with long-term hydrographs, New Jersey Coastal Plain, 1988.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 21. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, difference between simulated and measured water levels at observation wells, and area of upward flow in the Upper Potomac-Raritan-Magothy aquifer (model unit A3), New Jersey Coastal Plain, 1988.

Discharge from the aquifer is primarily ground-water withdrawals (75 percent). The remaining 25 percent of discharge is downward flow to the Middle Potomac-Raritan-Magothy aquifer, which is greatest in updip areas in northern Gloucester, Camden, Burlington, Mercer, and southern Middlesex Counties.

Englishtown Aquifer

Locations of withdrawal wells in the Englishtown aquifer are shown in appendix 4 (fig. 4d). Almost all of the pumpage from the Englishtown aquifer is in Ocean and Monmouth Counties. The largest withdrawals are near the coast at the border between Ocean and Monmouth Counties.

Observed water levels in the Englishtown aquifer are shown in figure 22. The major feature of the flow system is the large, deep cone of depression in southern Monmouth and northeastern Ocean Counties where measured water levels are more than 220 ft below sea level. The cone of depression is composed of several distinct pumping centers; the deepest part is near the coast. Observed water levels in the Englishtown aquifer in 1988 are similar to water levels in fall 1983. The shape of the cone of depression is the same but water levels inland, west of the Manasquan Inlet, were lower in 1983 (more than 200 ft below sea level) than in 1988.

Simulated water levels in the Englishtown aquifer for 1983-88 and residuals at six observation wells are shown in figure 23. Simulated heads match the observed water levels well, especially updip in Monmouth County and northern Burlington County. Measured water levels are within 10 ft of simulated values updip and near the center of the cone (downdip). Simulated heads are 1 ft above average measured water levels near the center of the cone. Simulated heads along the edges of the steep cone of depression are 20 to 30 ft above the average observed heads. The steep gradients and large grid cells make simulation of the cone of depression difficult. Flow directions and gradients near the cone, however, are reasonable.

Simulated heads in Camden and Gloucester Counties are higher than observed water levels. The gradient in observed heads is much steeper than the simulated gradient in updip areas, but observed water levels in this area are sparse. Simulated gradients between the local ground-water high in Camden and Gloucester Counties and the large cone of depression were steeper than the observed water levels indicate.

Water-level differences between the Englishtown aquifer and the overlying Wenonah-Mount Laurel aquifer are less than 10 ft throughout most of the aquifer except updip in the recharge areas and near the cone of depression in the northern half of Ocean County and southern Monmouth County. Vertical head differences between the aquifers near the cone exceed 100 ft in some areas.

In the ground-water flow budget for the confined part of the Englishtown aquifer (table 5), most of the recharge to the Englishtown aquifer is from the overlying Wenonah-Mount Laurel aquifer near ground-water highs in Camden and Gloucester Counties and western Monmouth County. Almost one-third of the inflow occurs where the Wenonah-Mount Laurel aquifer crops out. Flow from the confined part of the Wenonah-Mount Laurel aquifer is more than 0.5 in/yr in Gloucester and Camden Counties, near the ground-water high. Downdip flow from the unconfined

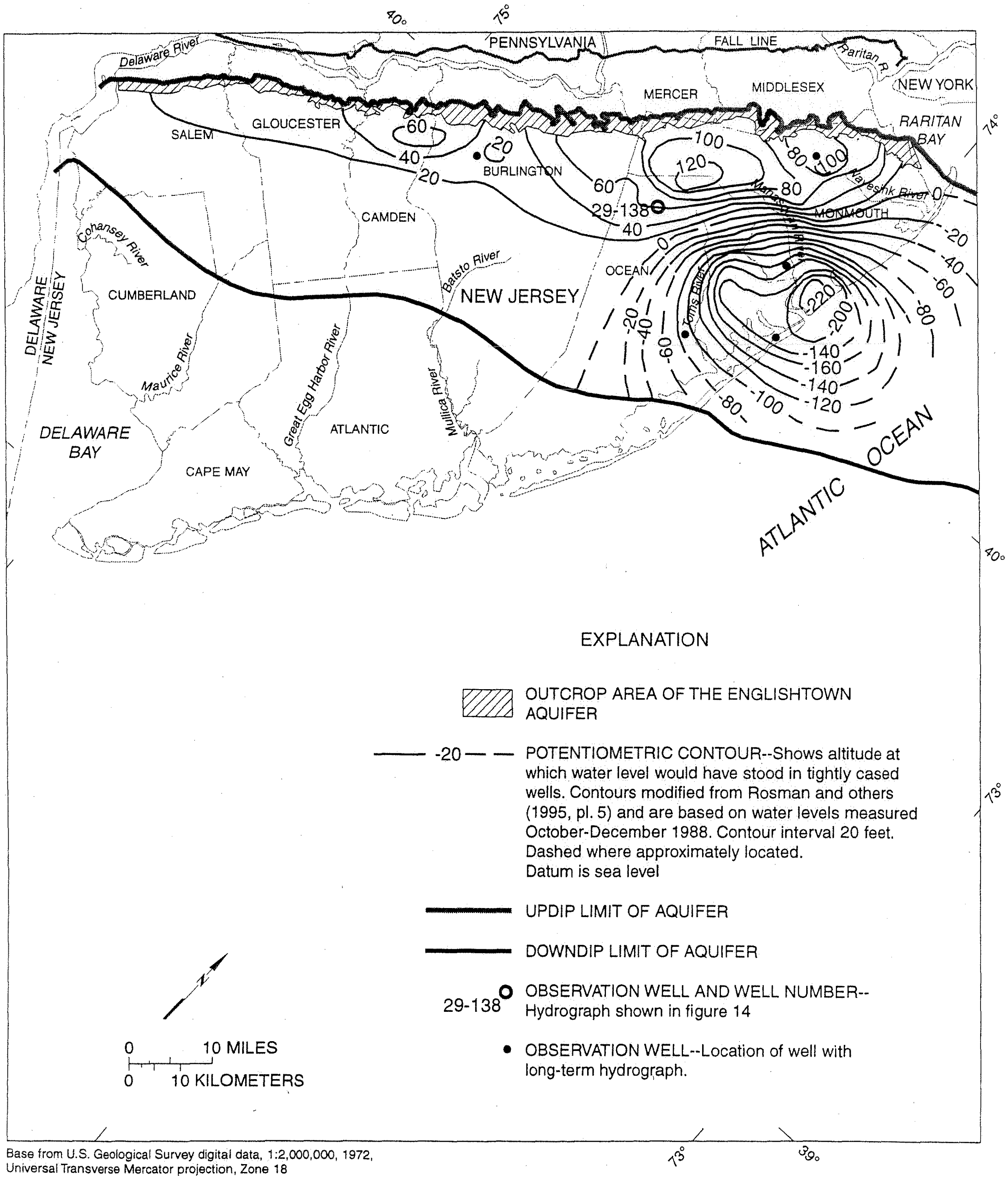
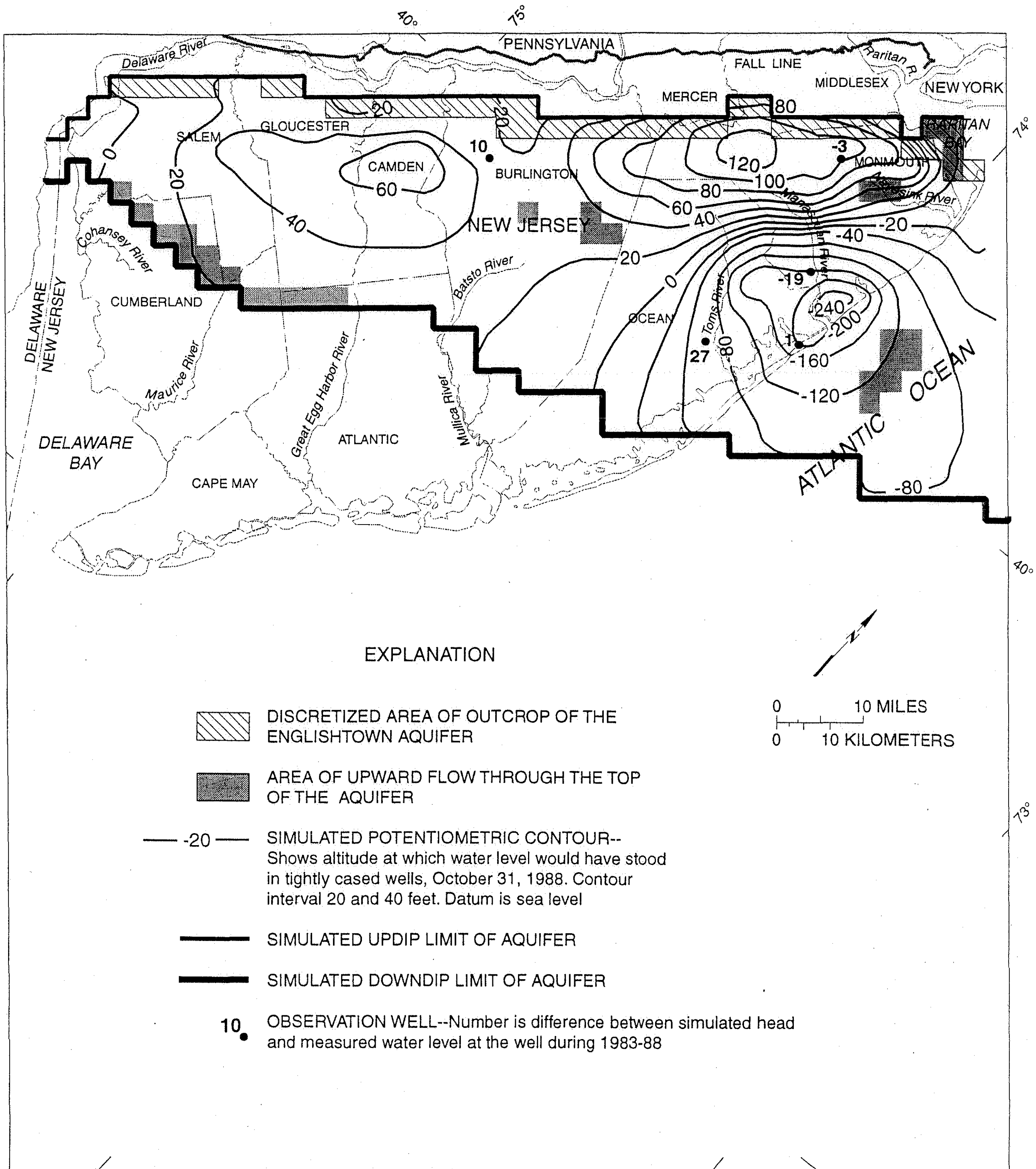


Figure 22. Observed potentiometric surface and locations of selected wells in the Englishtown aquifer (model unit A4) with long-term hydrographs, New Jersey Coastal Plain, 1988.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

Figure 23. Simulated potentiometric surface, difference between simulated and measured water levels at observation wells, and area of upward flow in the Englishtown aquifer (model unit A4), New Jersey Coastal Plain, 1988.

part of the Englishtown aquifer accounts for very little of the recharge. Much of the recharge to the outcrop of the Englishtown aquifer flows downward to recharge the underlying Upper Potomac-Raritan-Magothy aquifer. Discharge from the confined part of the Englishtown aquifer is to ground-water withdrawals, the Upper Potomac-Raritan-Magothy aquifer, and the outcrop of the Englishtown aquifer. Vertical flow to the underlying Upper Potomac-Raritan-Magothy aquifer represents 61 percent of outflow. Most of this flow occurs in updip areas, where the Merchantville-Woodbury confining unit is thinnest.

Several changes were made to hydrogeologic properties of the Englishtown aquifer and the overlying Marshalltown-Wenonah confining unit during model calibration. The hydraulic conductivity used for the Englishtown aquifer (6 to 8 ft/d) was more uniform than that used in the RASA model. This resulted in reduced transmissivity in updip areas and increased the gradient toward the major cone of depression. Transmissivities downdip are larger than those used in the RASA model. As discussed earlier, the leakance of the underlying Merchantville-Woodbury confining unit was decreased to reduce flow to the Potomac-Raritan-Magothy aquifers. The leakance of the overlying Marshalltown-Wenonah confining unit also was decreased (about 75 percent) during calibration. Reductions were largest along the coast in the northern part of the study area near the cone of depression.

Wenonah Mount-Laurel Aquifer

Locations of ground-water withdrawals from the Wenonah-Mount Laurel aquifer during 1983-88 are shown in appendix 5 (fig. 5d). Withdrawals from updip areas were made in Salem, Gloucester, Camden, and Burlington Counties. Withdrawals from the deeper parts of the aquifer were made downdip in Monmouth and Ocean Counties. Many of the simulated wells are small; only a few of those shown have average withdrawals for the period of more than 0.5 Mgal/d. Withdrawals were largest in Burlington (41 percent), Camden (25 percent), and Monmouth (21 percent) Counties.

The observed water levels in the Wenonah Mount-Laurel aquifer in 1988 are shown in figure 24. The flow system in the Wenonah-Mount Laurel aquifer is similar to that in the underlying Englishtown aquifer. Water levels are more than 140 ft above sea level in southwestern Monmouth and western Ocean Counties, where the aquifer receives recharge. Water levels are lowest in southern Monmouth and northern Ocean Counties near the coast, where water levels in the center of the large cone of depression are more than 200 ft below sea level. Large drawdowns in this area are primarily the result of withdrawals from the underlying Englishtown aquifer. Ground-water flow in the Wenonah-Mount Laurel aquifer is from recharge areas in southern Monmouth, western Ocean, and northwestern Burlington Counties toward the cone of depression near the coast. Water from the recharge area at the ground-water high in Camden and Gloucester Counties flows both downdip toward the cone and toward the Delaware Bay, and updip toward the aquifer outcrop in Gloucester County.

Simulated heads, the location of the tip and toe of the freshwater-saltwater interface, and residuals at eight observation wells in the Wenonah-Mount Laurel aquifer (model unit A5) for 1983-88 are shown in figure 25. Simulated heads are within 13 ft of the average measured water level at all eight wells. Heads updip from the cone of depression in Monmouth and Ocean Counties

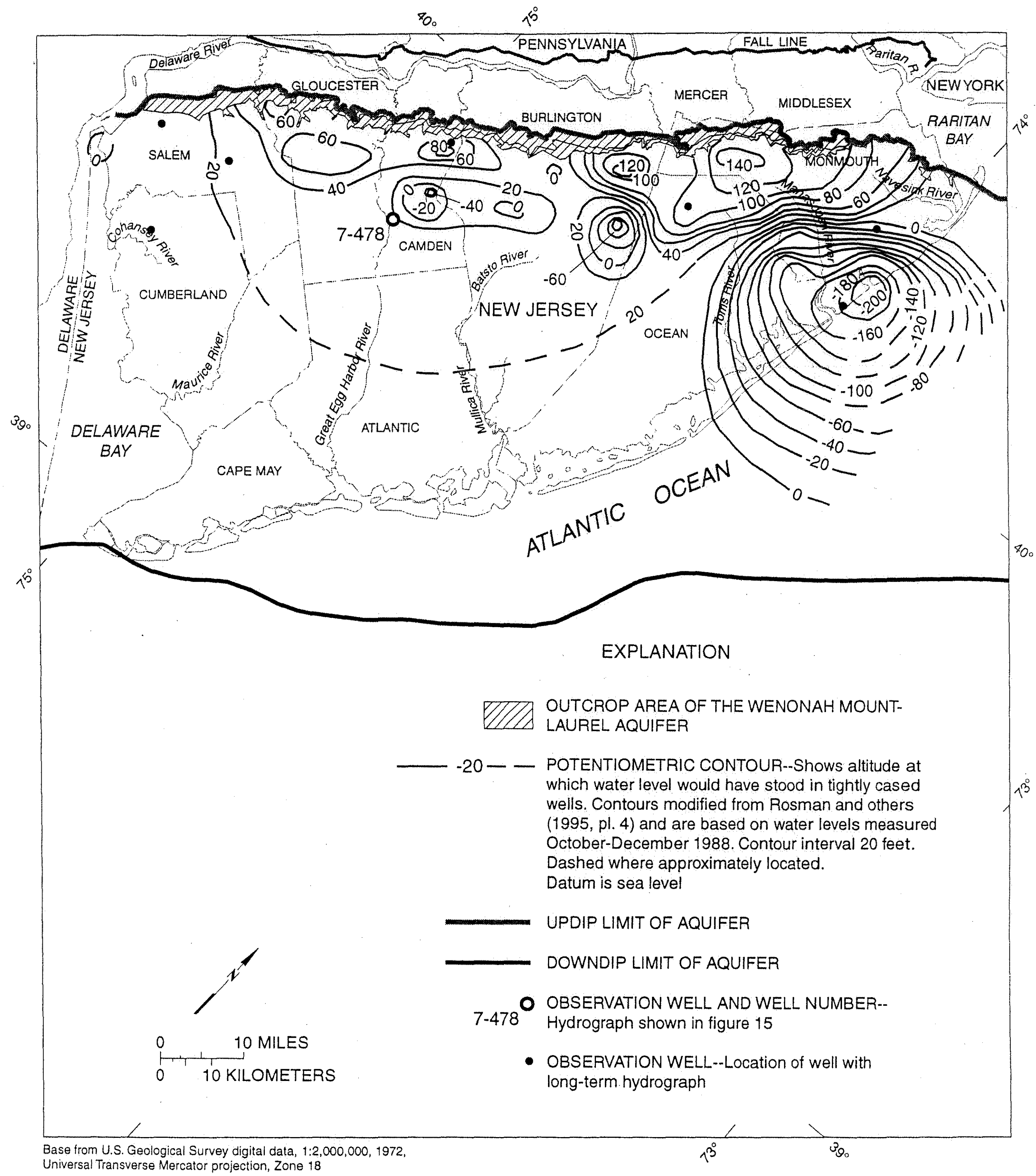


Figure 24. Observed potentiometric surface and locations of selected wells in the Wenonah-Mount Laurel aquifer (model unit A5) with long-term hydrographs, New Jersey Coastal Plain, 1988.

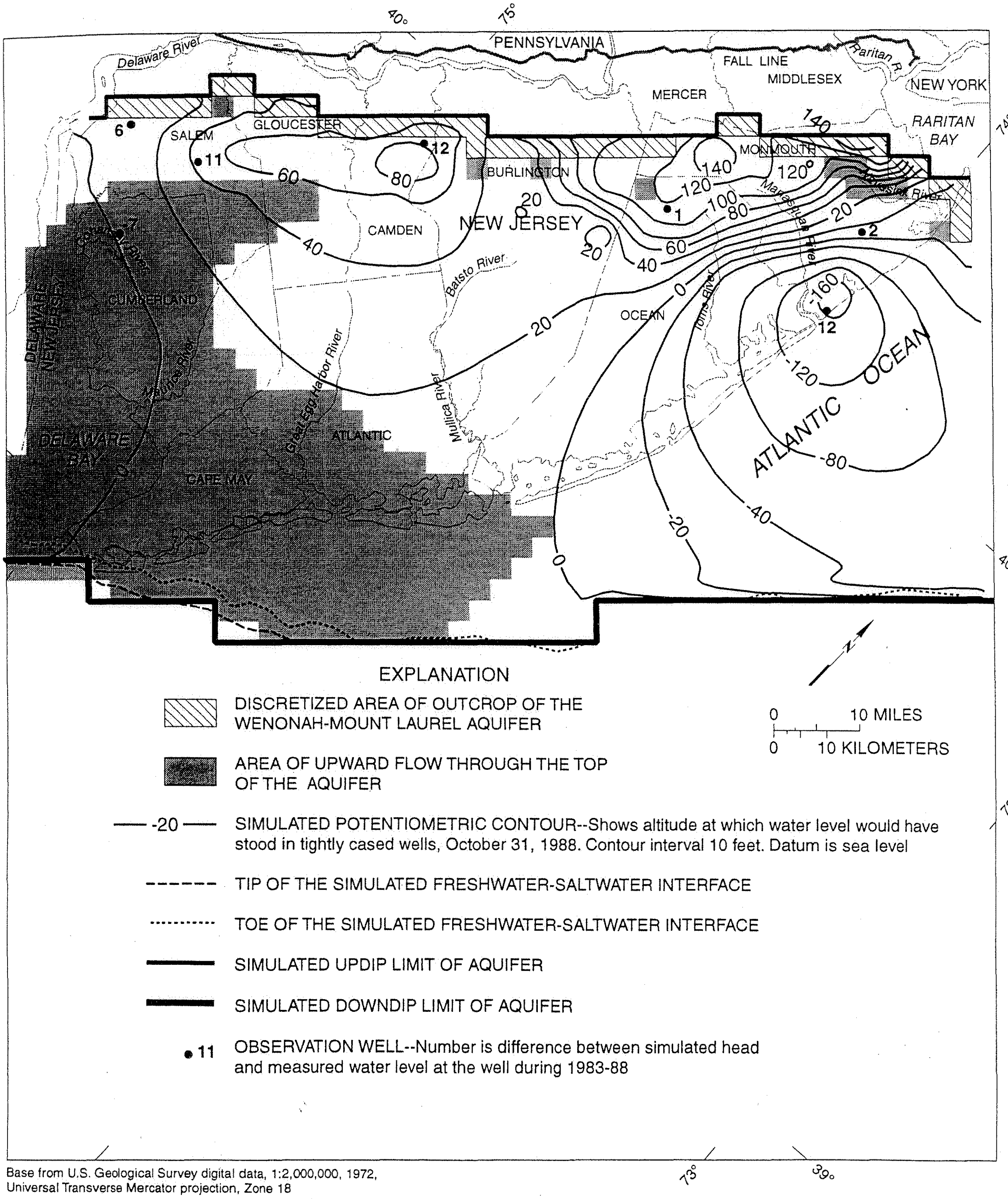


Figure 25. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, difference between simulated and measured water levels at observation wells, and area of upward flow in the Wenonah-Mount Laurel aquifer (model unit A5), New Jersey Coastal Plain, 1988.

and near the Delaware Bay in Salem and Cumberland Counties match the observed water levels well. The simulated water level at well 25-486 near the center of the cone is only 12 ft higher than the average observed water level. Simulated heads in Camden and Gloucester Counties are from 10 to 15 ft above the observed water levels near the recharge area. The simulated water levels are more similar to the 1983 water levels than the 1988 water levels. Downdip from the recharge area, simulated heads are higher and gradients are less steep than measured water levels and observed gradients. In general, simulated flow directions and gradients in the Wenonah-Mount Laurel aquifer are similar to those in the observed flow system.

In the flow budget for the confined part of the Wenonah-Mount Laurel aquifer (table 5), 95 percent of inflow comes from the overlying aquifers and discharges downward to the underlying Englishtown and Upper Potomac-Raritan-Magothy aquifers. Horizontal flow from the outcrop of the Wenonah-Mount Laurel aquifer supplies only 2 percent of inflow. Vertical flow from the overlying unconfined aquifer and confining-unit outcrop supplies almost 28 percent of inflow, and vertical flow from the overlying confined Vincentown, Piney Point, and lower Kirkwood-Cohansey and confined Kirkwood aquifers supply 68 percent of inflow. Flow from the overlying confined aquifers occurs primarily in updip areas. Areas in which vertical flow is 0.5 to more than 7 in/yr coincide with the ground-water highs in western Monmouth County and parts of Gloucester, Burlington, and Salem Counties.

Flow from the Wenonah-Mount Laurel aquifer to the underlying Englishtown aquifer represents 71 percent of aquifer discharge. Most of the vertical flow occurs in updip areas and near the cone of depression in the Englishtown aquifer. Water also discharges from the Wenonah-Mount Laurel aquifer to the overlying aquifer in southern and downdip areas in Cumberland and Cape May Counties and parts of Salem and Atlantic Counties. Ground-water withdrawals from the confined Wenonah-Mount Laurel aquifer account for 18 percent of outflow. As in the Englishtown aquifer, ground-water discharges horizontally to the aquifer outcrop in Gloucester, Camden, and northern Salem Counties.

Both the transmissivity of the Wenonah-Mount Laurel aquifer and the leakance of the overlying Navesink-Hornerstown confining unit (model unit C5) were adjusted during model calibration. Transmissivity was decreased in northern Ocean and southern Monmouth Counties to better simulate gradients updip from the cone of depression. Hydraulic conductivities in this area are 6 to 85 ft/d and are similar to conductivities in the Englishtown aquifer. Leakance of the overlying confining unit was decreased during calibration to reduce heads in the Wenonah-Mount Laurel aquifer.

Vincentown Aquifer

Ground-water withdrawals from the Vincentown aquifer (model unit A6) generally are less than 1 Mgal/d. Locations of withdrawal wells in this aquifer during 1983-88 are shown in appendix 6 (fig. 6d). Withdrawal wells are present only in the northern part of the Coastal Plain in Monmouth and Ocean Counties.

The observed water levels in the Vincentown aquifer in 1988 are shown in figure 26. Water-level data are sufficient to draw an observed-potentiometric-surface map only in Monmouth and Ocean Counties. Ground-water flow is from the topographic high in northwestern Ocean County and southern Monmouth County toward the Atlantic Coast.

The simulated potentiometric surface of the Vincentown aquifer during October 1983-October 1988 is shown in figure 27. Measured water levels are available for this period at only one observation well. The simulated head at this well is 8 ft below the average measured water level for the period. In general, the simulated heads match the measured water levels in 1988 closely in the northern part of the aquifer where measured water levels are available.

Simulated flow is from the ground-water highs in Camden, Burlington, Ocean, and Monmouth Counties toward the Delaware Bay to the south and the Atlantic Ocean to the north. Between the two ground-water highs, water flows downward to the Wenonah-Mount Laurel aquifer. In the southern part of the aquifer in Salem, Gloucester, and Camden Counties, water levels in the Vincentown aquifer and underlying Wenonah-Mount Laurel aquifer are very similar. In the northern part of the aquifer, water levels in the Vincentown aquifer generally are higher than in the southern part, and water near the topographic highs flows downward to recharge the Wenonah-Mount Laurel aquifer.

In the flow budget for the confined part of the Vincentown aquifer (table 5), recharge is by vertical flow from the overlying lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) and by horizontal flow from its outcrop areas. Recharge from the outcrop occurs in Burlington, Ocean, and southern Monmouth Counties, near the topographic highs. Vertical flow from the overlying aquifer occurs everywhere except in northern Burlington County and parts of Monmouth County. In Monmouth County, near the Manasquan River, water flows downdip toward the downdip limit of the aquifer and discharges to both the underlying Wenonah-Mount Laurel aquifer and the overlying lower Kirkwood-Cohansey and confined Kirkwood aquifers. Discharge from the aquifer is primarily vertical flow to the underlying aquifer. Water also discharges horizontally to the outcrop near Raritan Bay in the south and near the Atlantic Coast in Monmouth County.

The transmissivity of the Vincentown aquifer was not changed during model calibration. The leakance of the overlying Vincentown-Manasquan confining unit (model unit C6) was decreased about 70 percent.

Piney Point Aquifer

Ground-water withdrawals from the Piney Point aquifer (model unit A7) are relatively minor; locations of withdrawal wells in this aquifer are shown in appendix 7 (fig. 7d). Wells in Ocean County are near the coast and account for 84 percent of the withdrawals.

The observed water levels in the Piney Point aquifer in fall 1988 are shown in figure 28. Ground water flows from updip areas in eastern Burlington and southern Ocean Counties southeast toward the Atlantic Coast and south toward the Delaware Bay. Water levels in Cumberland County

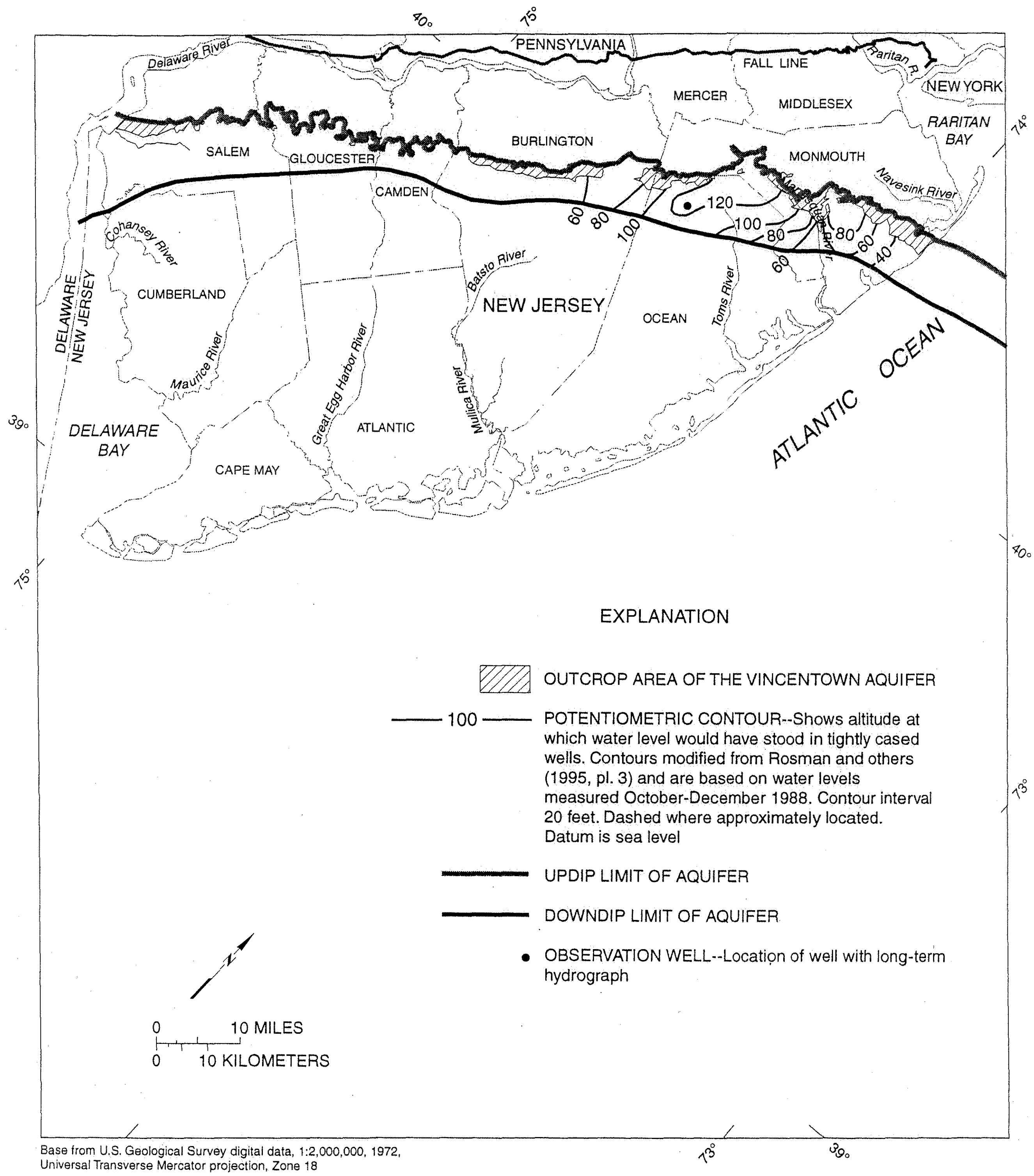


Figure 26. Observed potentiometric surface and locations of selected wells in the Vincentown aquifer (model unit A6) with long-term hydrographs, New Jersey Coatal Plain, 1988.

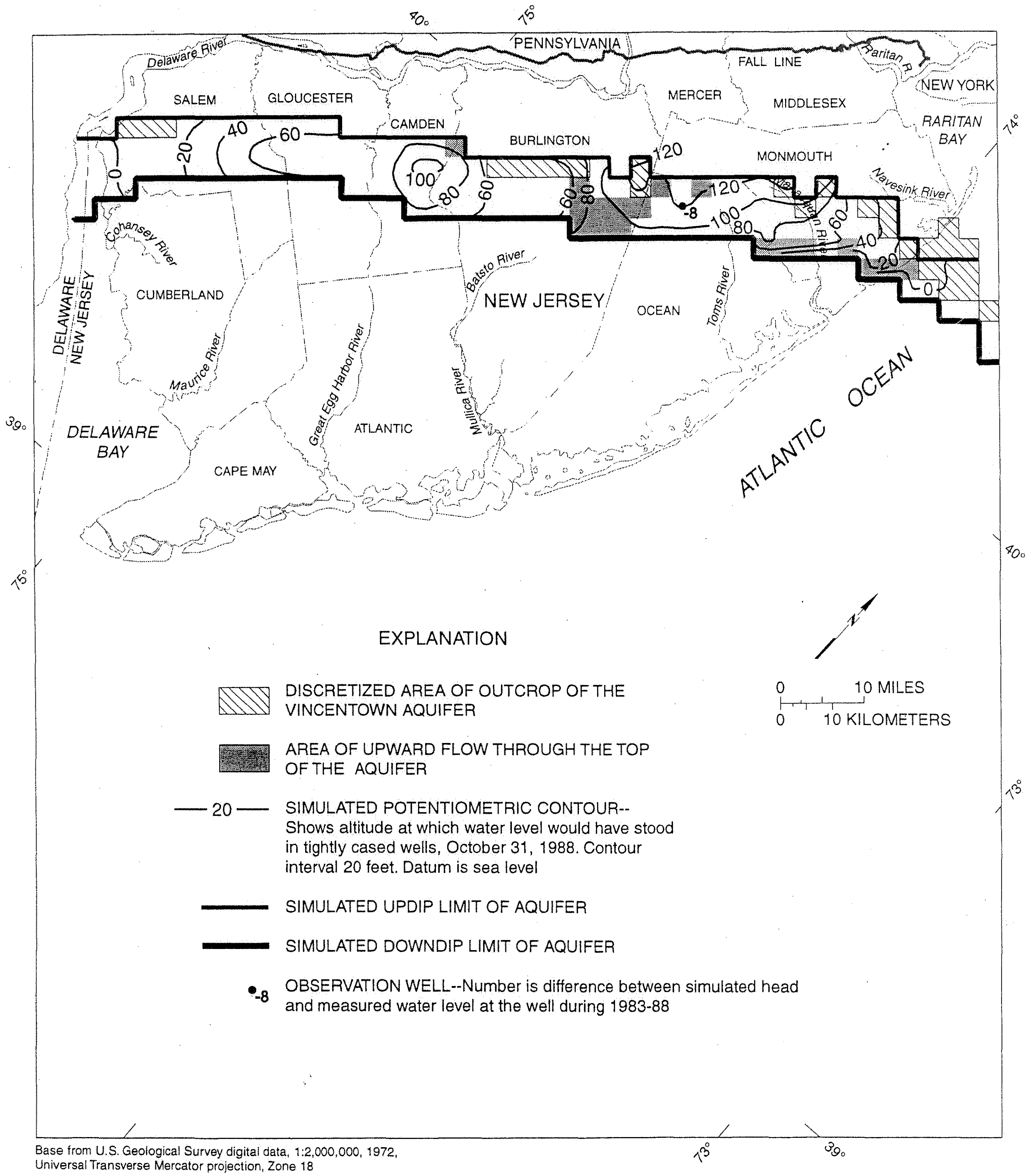


Figure 27. Simulated potentiometric surface, difference between simulated and measured water levels at observation wells, and area of upward flow in the Vincentown aquifer (model unit A6), New Jersey Coastal Plain, 1988.

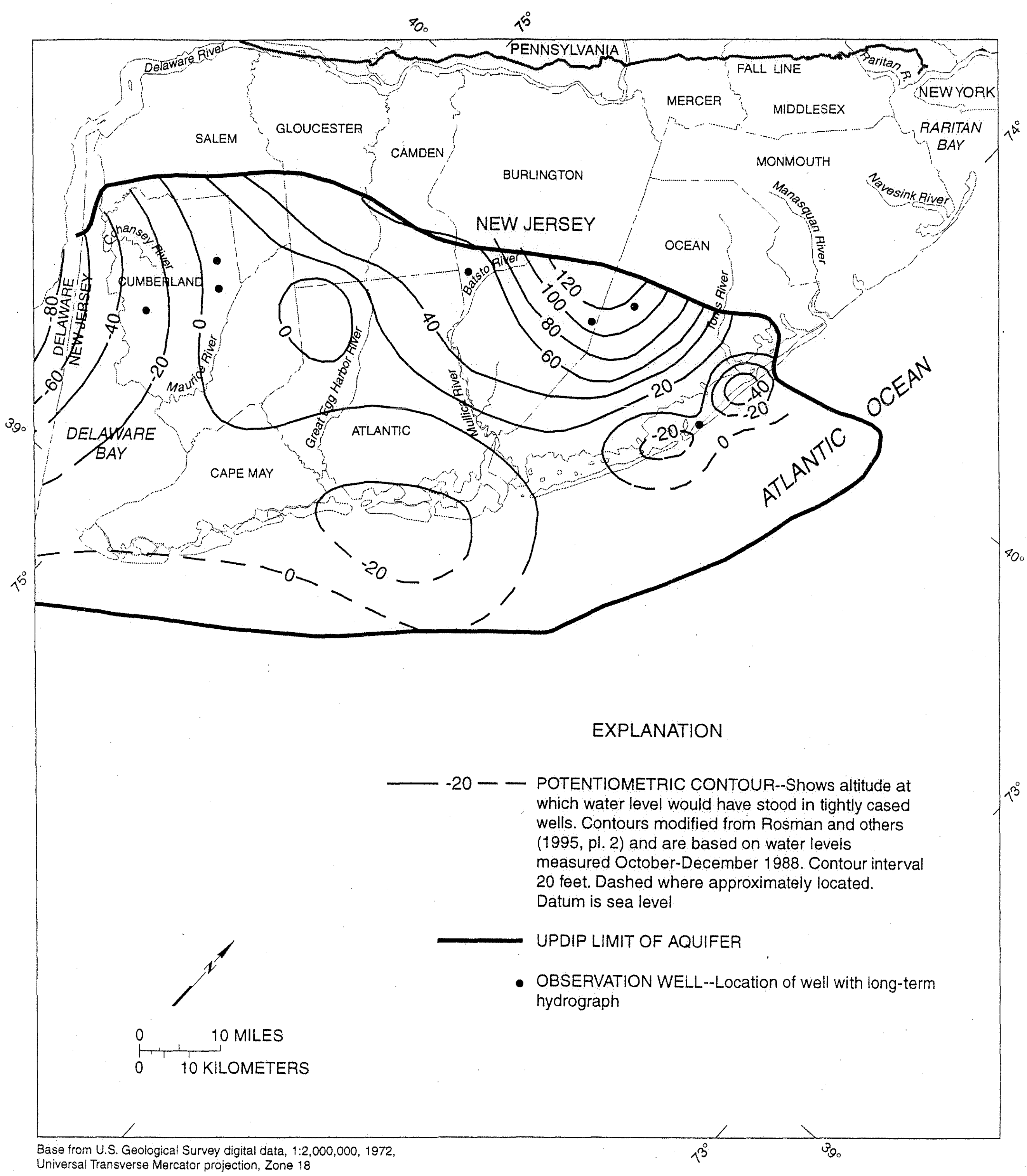


Figure 28. Observed potentiometric surface and locations of selected wells in the Piney Point aquifer (model unit A7) with long-term hydrographs, New Jersey Coastal Plain, 1988.

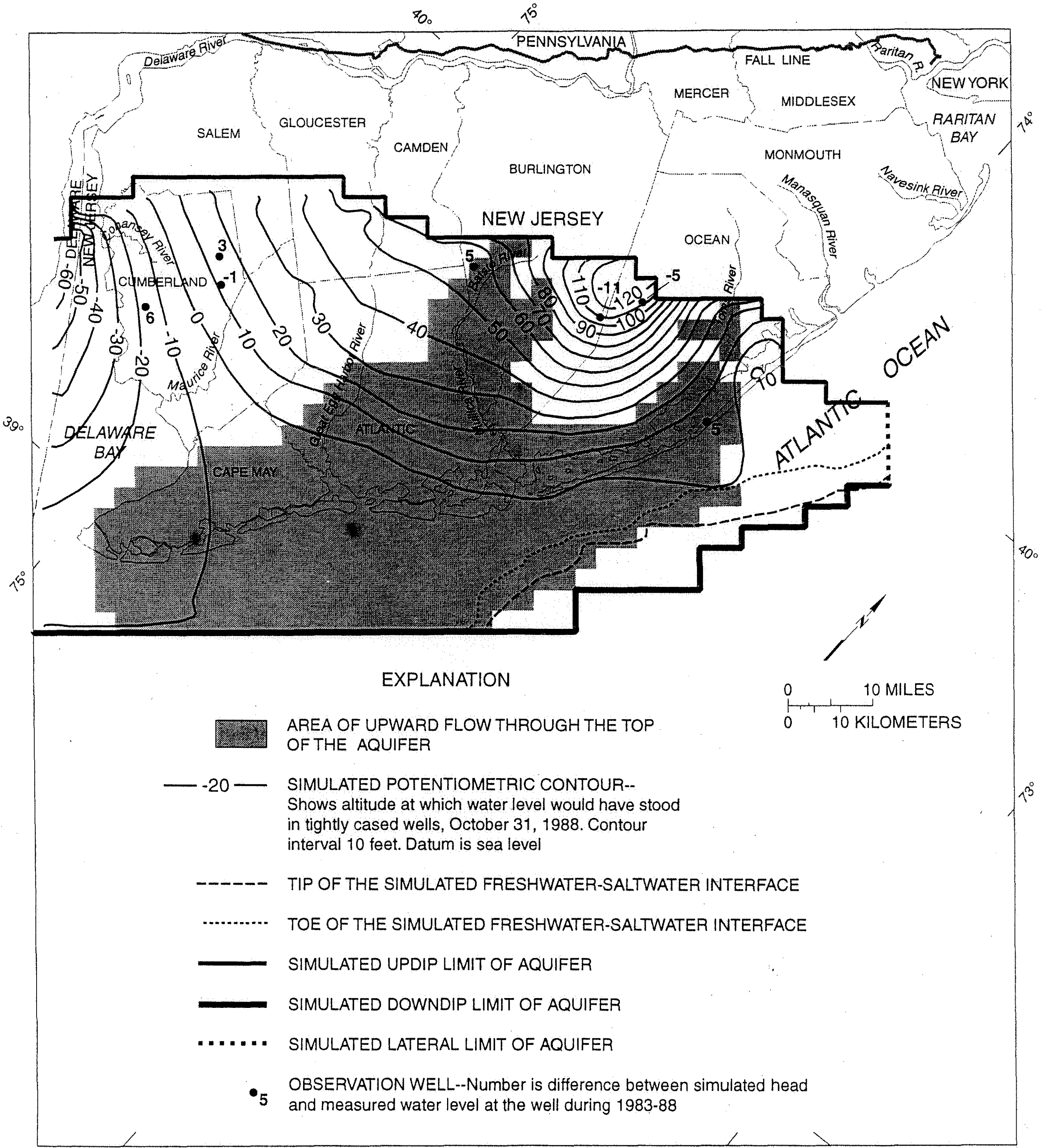
are affected by withdrawals in Delaware (where water levels are more than 120 ft below sea level). Ground-water withdrawals from the barrier islands off the coast of Ocean County have resulted in two small cones of depression where water levels are more than 20 ft below sea level.

Simulated heads and the location of the tip and toe of the freshwater-saltwater interface in the Piney Point aquifer for 1983-88 are shown in figure 29. Water levels at the seven observation wells are all within 11 ft of the average measured water levels and only one well had a residual greater than ±6 ft. The configuration of the simulated potentiometric surface and flow directions match those in the observed flow system. Water levels in Burlington and Ocean Counties are within 10 ft of the measured water levels. One of the small cones of depression on the barrier islands (north of the mouth of the Mullica River) is not well simulated because of the large grid cells.

Simulated water levels in the southern part of the study area in Cumberland and Cape May Counties and parts of Gloucester, Camden, and Atlantic Counties are as much as 20 ft higher than measured water levels in 1988. Simulated water levels, however, are very similar to measured water levels in 1983 (Eckel and Walker, 1986, pl. 6), and simulated water levels at three observation wells in the area are within 6 ft of the average water level during 1983-88. Differences between simulated and measured heads are the result of changes in the flow system that occurred during the last model stress period (1983-88).

In the flow budget for the confined part of the Piney Point aquifer (table 5), recharge is primarily flow from the overlying unconfined lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8). Most of this recharge occurs updip along the border between Burlington and Ocean Counties, where the confining unit is thin and water levels in the overlying aquifer are greater than 120 ft above sea level (as a result of a topographic high). Other sources of water to the aquifer are upward flow from the Wenonah-Mount Laurel aquifer and downward flow from the overlying lower Kirkwood-Cohansey and confined Kirkwood aquifers in the Delaware Bay (included in the saltwater-flow term). Flow in the Piney Point aquifer differs from that in the other aquifers because less of the water that recharges the aquifer updip flows downward to underlying aquifers. The largest discharge from the aquifer is flow across the model boundary near the Delaware Bay. Because these flows were such an important part of the budget of the aquifer, they were adjusted by using fluxes estimated from 1983 and 1988 heads near the Delaware Bay and estimates of transmissivity in order to calibrate the model. Water discharges to the overlying aquifer only in small areas where the overlying aquifer is unconfined, but discharges over most of the area where the aquifer is confined.

The leakance of the overlying basal Kirkwood confining unit (model unit C7) was decreased 75 percent over much of the extent of the confining unit during model calibration. Transmissivity was reduced in a small area in Cumberland County to obtain a more uniform transmissivity distribution in the area.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

Figure 29. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, difference between simulated and measured water levels at observation wells, and area of upward flow in the Piney Point aquifer (model unit A7), New Jersey Coastal Plain, 1988.

Lower Kirkwood-Cohansey and confined Kirkwood Aquifers

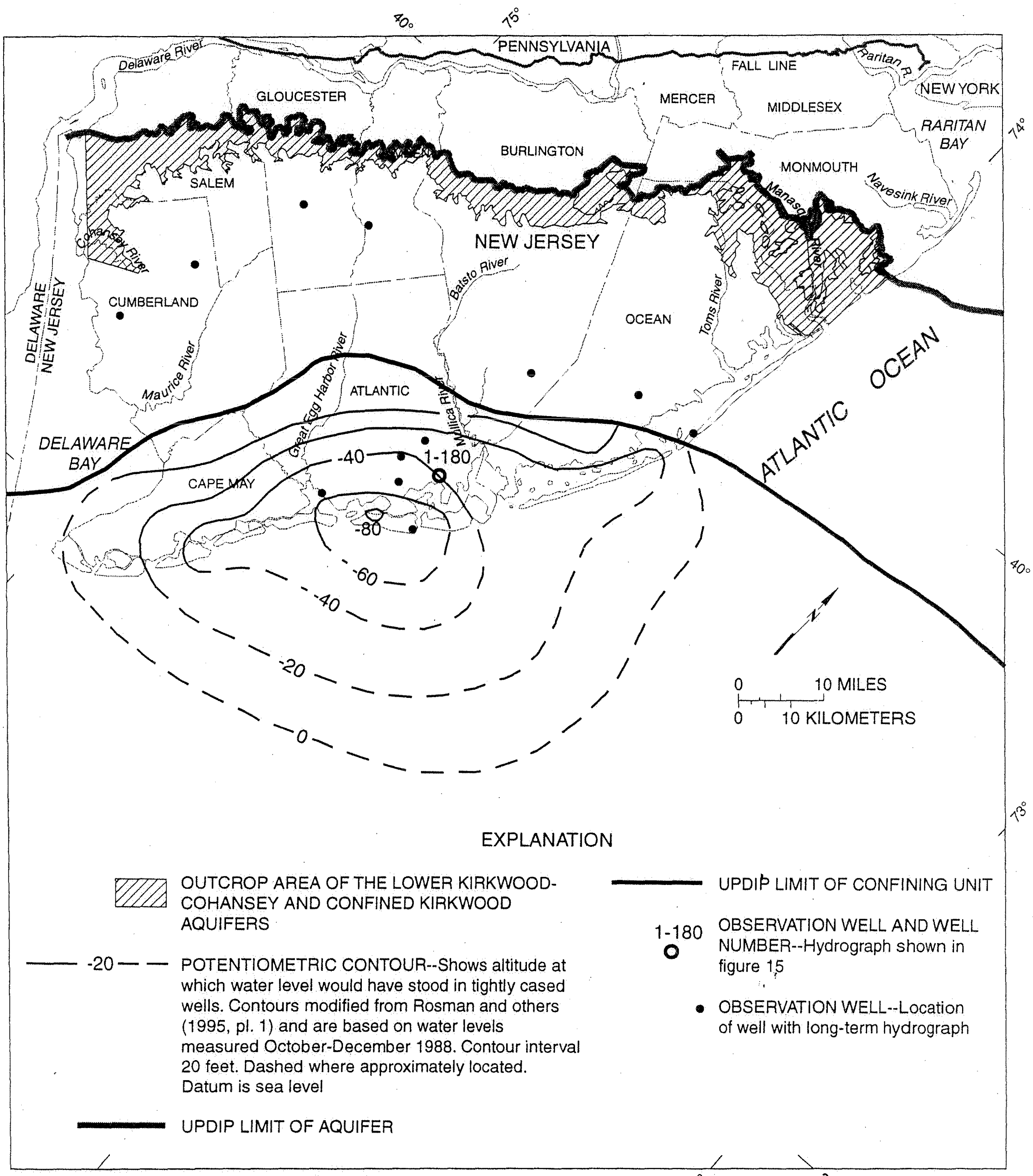
Withdrawals from the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) total about 22 Mgal/d and are primarily from wells along the Atlantic Coast. This aquifer is a major source of water supply in the New Jersey Coastal Plain. Locations of withdrawal wells in the lower Kirkwood-Cohansey and confined Kirkwood aquifers for 1983-88 are shown in appendix 8 (fig. 8d). Withdrawals in Atlantic, Cape May, and southern Ocean Counties account for 48 percent, 30 percent, and 22 percent, respectively, of the total withdrawals from the confined aquifer.

The observed water levels in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) in fall 1988 are shown in figure 30. Only water levels for the confined part of layer A8 (the confined Kirkwood aquifer) are shown. A large cone of depression is centered around Atlantic City and water levels have fallen 30 to 50 ft from predevelopment levels over most of the aquifer. Water levels at the center of the cone are more than 80 ft below sea level. The ground-water flow direction in Cape May County, near the Delaware Bay, has reversed since predevelopment; water now flows from the Delaware Bay (where the aquifer is unconfined) toward Cape May and Atlantic Counties rather than discharging to the Delaware Bay.

Simulated heads and the location of the tip and toe of the freshwater-saltwater interface in the confined Kirkwood aquifer during 1983-88 are shown in figure 31. Simulated heads closely match the heads observed in 1988 except near the center of the cone of depression, where simulated heads are higher than observed heads as a result of the large model cells. Differences between simulated heads and average measured water levels are less than 11 ft at all six wells and are less than ±7 ft at five of the six wells. Where the aquifer is unconfined (and model unit A8 represents the lower Kirkwood-Cohansey aquifer), residuals at observation wells are reasonable and the simulated water levels closely match the water levels observed in 1988.

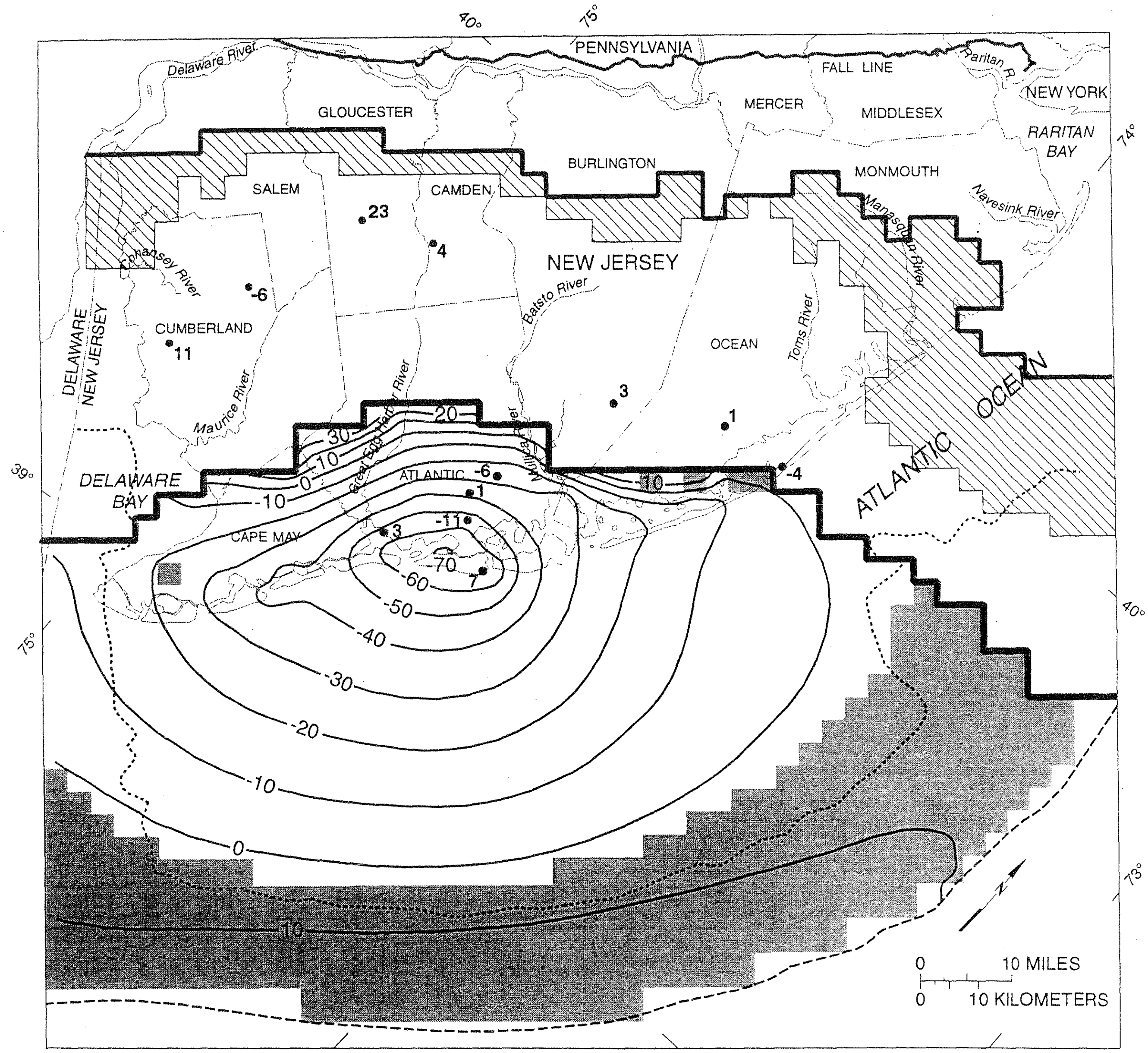
Simulated water levels west of Cape May in the Delaware Bay are 10 to 20 ft below sea level near the updip limit of the overlying confining unit. Where the confining unit is absent, the aquifer is in contact with overlying saltwater in the upper Kirkwood-Cohansey aquifer (model unit A9); however, saltwater in unit A9 was not simulated in this area. Water flows from the overlying upper Kirkwood-Cohansey aquifer into the confined Kirkwood aquifer over most of the aquifer. Freshwater discharges to the overlying aquifer downdip, near the freshwater-saltwater interface.

In the flow budget for the confined Kirkwood aquifer (model unit A8) (table 5), recharge to the aquifer is primarily by horizontal flow from the unconfined part of the lower Kirkwood-Cohansey aquifer (59 percent of inflow). The underlying Piney Point aquifer (model unit A7) contributes more inflow (11 percent) than do the overlying unconfined parts of the upper Kirkwood-Cohansey aquifer (model unit A9). Flow from saltwater areas, unconfined areas in the lower Kirkwood-Cohansey aquifer (model unit A8) in the Delaware Bay, and offshore areas in the upper Kirkwood-Cohansey aquifer (model unit A9) is significant and accounts for 10 percent of the inflow to the aquifer. Water displaced by updip movement of the freshwater-saltwater interface is represented as inflow from storage and accounts for about 9 percent of inflow. This occurs



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 30. Observed potentiometric surface in the confined Kirkwood aquifer and locations of selected wells in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) with long-term hydrographs, New Jersey Coastal Plain, 1988.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

39°
73°

EXPLANATION

DISCRETIZED AREA OF OUTCROP OF THE LOWER KIRKWOOD-COHANSEY AND CONFINED KIRKWOOD AQUIFERS

AREA OF UPWARD FLOW THROUGH THE TOP OF THE AQUIFER

-20 SIMULATED POTENTIOMETRIC CONTOUR--Shows altitude at which water level would have stood in tightly cased wells, October 31, 1988. Contour interval 10 feet. Datum is sea level

TIP OF THE SIMULATED FRESHWATER-SALTWATER INTERFACE

TOE OF THE SIMULATED FRESHWATER-SALTWATER INTERFACE

SIMULATED UPDIP LIMIT OF AQUIFER

SIMULATED UPDIP LIMIT OF OVERLYING CONFINING UNIT

7 OBSERVATION WELL--Number is difference between simulated head and measured water level at the well during 1983-88

Figure 31. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, difference between simulated and measured water levels at observation wells, and area of upward flow in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8), New Jersey Coastal Plain, 1988.

offshore between the tip and toe of the interface, east of Atlantic and Ocean Counties. Ground-water withdrawals from the aquifer represent 95 percent of the outflow from the confined Kirkwood aquifer. Other outflows are relatively minor.

The hydrogeologic properties of the lower Kirkwood-Cohansey and confined Kirkwood aquifers and the confining unit overlying the Rio Grande water-bearing zone (model unit C8) were adjusted slightly during model calibration. Leakance of the confining unit was decreased 75 percent, a reduction similar to those made in other confining units. Transmissivity was increased in a small area updip from the cone of depression in order to match observed gradients near the cone of depression.

Freshwater-Saltwater Interface Location and Movement, 1896-1988

Available chloride-concentration data in downdip and offshore parts of the aquifers were used to simulate the location of the freshwater-saltwater interface and to describe the zone of diffusion in each aquifer. Because chloride-concentration data near the interface are limited, it was not always possible to determine whether the interface was well or poorly simulated or whether the zone of diffusion was narrower or wider than expected.

Meisler (1989) provides chloride measurements made at 11 onshore and offshore locations (fig. 1). In most cases chloride concentrations were measured at various depths in each well. These measurements are shown in table 6 (at end of report). For this study, the sample depths were compared to the hydrogeologic framework and each sample was assigned to the appropriate model unit. Where the aquifer contains only freshwater, chloride values at both the top and bottom of the aquifer are likely to be low (much less than 10,000 mg/L). Where the aquifer contains both fresh-water and saltwater (contains the interface), chloride values at the top of the aquifer are likely to be less than 10,000 mg/L, and values from near the bottom of the aquifer are likely to be greater than 10,000 mg/L. Where the aquifer contains only saltwater, chloride values at all depths are expected to be significantly greater than 10,000 mg/L (that is, similar to concentrations in pure seawater). These guidelines were used to determine, where possible, where the location of the interface may not be well simulated and where the zone of diffusion near the interface may be wider than the sharp-interface assumptions in the model indicates. The chloride concentrations of some brines in the study area are greater than those of seawater (18,000 mg/L) (table 6). Meisler and others (1985) conclude that the transition zones probably were produced by the mixing of fresh and salty ground water of either brine or salty ground-water origin. The chloride concentration of brine is several times that of seawater (Trapp and Meisler, 1992).

All chloride measurements of samples from the 106 wells were made by the U.S. Geological Survey National Water Quality Laboratory (table 7, at end of report). Most of these data were collected during 1983-88. Where data were not available for this period, a sample collected on another date was substituted. Seventeen samples, therefore, were collected during 1989-90, generally in Ocean and Salem Counties. These samples were collected as part of the New Jersey Coastal Plain saltwater monitoring network (Bauersfeld and others, 1989, 1990a, 1990b). These wells are 9-067, 9-302, 11-061, 11-691, 15-194, 25-001, 25-320, 29-004, 29-454, 29-626, 29-807, 29-815, 33-035, 33-108, 33-346, 33-364, and 33-459. Eighteen samples collected from wells during 1971-82 were used when more recent data were not available. These wells are 9-153, 11-063, 11-066, 11-133, 11-137, 15-231,15-324, 15-349, 25-009, 25-142, 25-243, 25-321, 33-074, 33-106, 33-251, 33-253, 33-368, and 33-426.

Most of the chloride concentrations were determined in samples collected from production wells. Some data from observation and test wells also were available. The wells were selected on the basis of proximity to an area of documented saltwater intrusion or to an area in which simulation results indicated that ground water may be moving vertically from an area of saltwater to an area of freshwater. Chloride concentrations were compared to the simulated position of the tip and toe of the freshwater-saltwater interface.

Potomac-Raritan-Magothy Aquifer System

The simulated freshwater-saltwater interface in the Potomac-Raritan-Magothy aquifer system is fairly consistent among the aquifers because the sediments downdip are well connected. The depth to the interface is between 2,000 and 2,500 ft below sea level in all three aquifers. The interface is deepest in the Upper Potomac-Raritan-Magothy aquifer and shallowest in the Lower Potomac-Raritan-Magothy aquifer. This means that freshwater in downdip parts of the Middle and Upper Potomac-Raritan-Magothy aquifers overlies saltwater in the underlying Lower and Middle Potomac-Raritan-Magothy aquifers, respectively, creating areas where saltwater could potentially flow upward into freshwater in overlying aquifers.

Lower Potomac-Raritan-Magothy aquifer

The position and shape of the freshwater-saltwater interface in the Lower Potomac-Raritan-Magothy aquifer (app. 1, fig. 1f) are basically unchanged from those in 1896. The interface moved less than 0.1 mi in all areas; however, large amounts of water (5.2 Mgal/d, or 7 percent of inflow) have been released from storage near the interface as a result of the landward movement of saltwater and the displacement of freshwater. This release from storage occurs primarily in Burlington and Ocean Counties between the tip and toe of the freshwater-saltwater interface.

The simulated location of the interface in the southern part of the New Jersey Coastal Plain matches the observed data fairly well. Minimum and maximum chloride concentrations in a borehole in Cape May County and two wells on the northern boundary of Cumberland County (table 6) are 11,600 and 38,100 mg/L. Chloride values from the Anchor Gas Dickinson 1 well (number 9, fig. 1) in Cape May County show that the aquifer contains saltwater. At the Anchor Gas Ragovin well (number 7, fig. 1) in Cumberland County, concentrations range from 11,600 mg/L near the top of the aquifer to 27,400 mg/L near the bottom. This indicates that the water in the aquifer is primarily salty but becomes less salty near the top, an interpretation that matches the simulated location of the interface closely. Chloride concentrations in the Island Beach Test well (number 6, fig. 1) near the coast in Ocean County, about 6 mi downdip from the tip of the interface, range from 5,400 to 11,300 mg/L. The presence of fresher water near the top of the aquifer than near the bottom may indicate that the interface is farther offshore than is simulated in the model or that the width of the zone of transition may be greater than that of the sharp interface between freshwater and saltwater used in the model. Because the highest concentrations at this well are only slightly greater than 10,000 mg/L, the interface location in this area probably is farther offshore than the simulated interface location.

High chloride concentrations measured in updip areas in Salem and Gloucester Counties (170 to more than 800 mg/L) are the result of the movement of salty water from the southeast (Schaefer, 1983, p. 46-47). The only wells that withdraw water from the Lower Potomac-Raritan-Magothy aquifer in this area tap a pocket of relatively fresh water near the Delaware River in Salem County. This indicates that the zone of transition in this aquifer in the southwestern part of the study area probably is wide. Water in the updip part of the aquifer near the Delaware River is relatively fresh; therefore, movement of water from the Delaware River into the aquifer has not increased chloride concentrations in these areas.

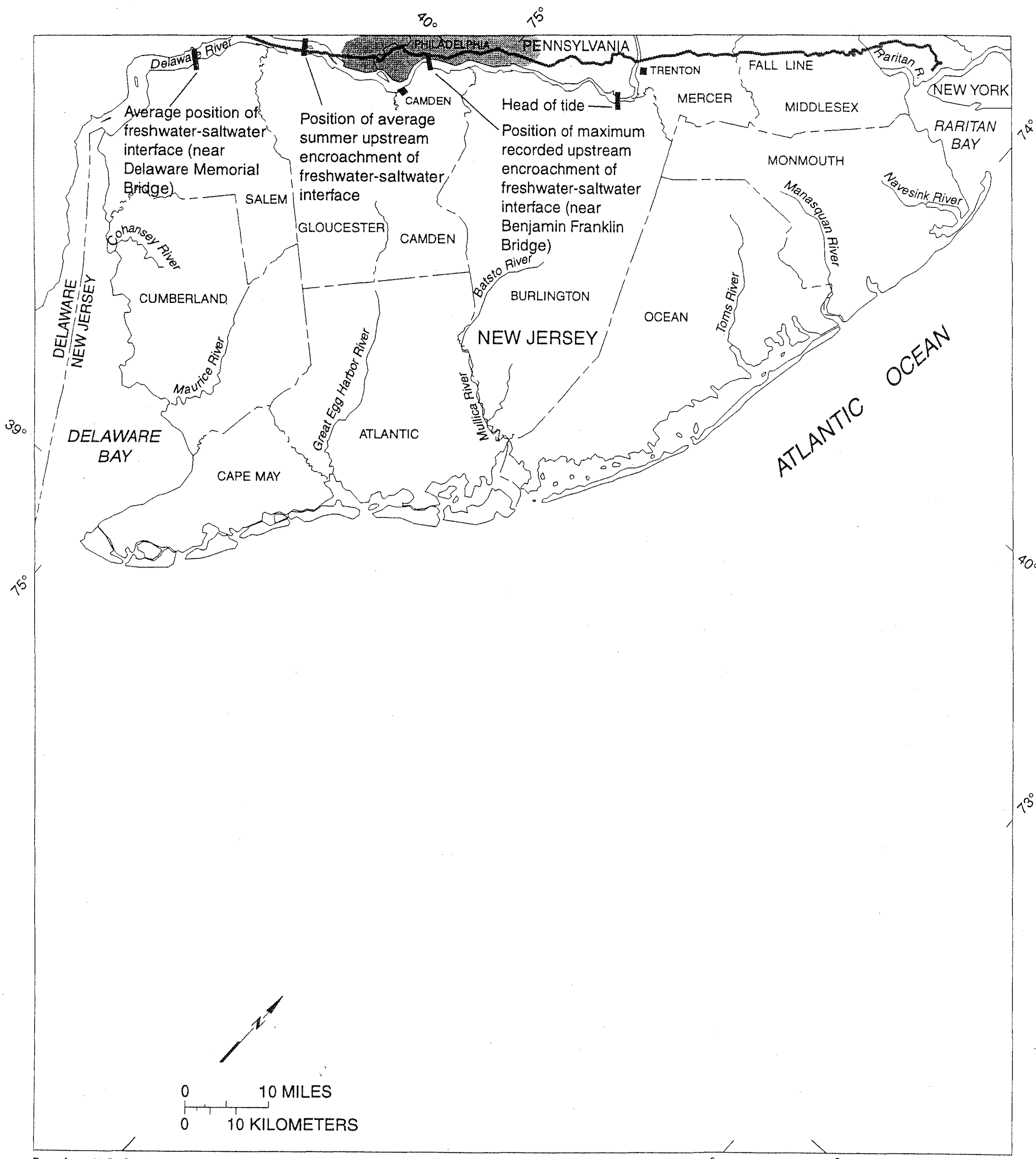
The extent of saltwater in the Delaware River under various conditions is shown in figure 32. The average position of the freshwater-saltwater interface in the estuary is near the Delaware Memorial Bridge in Salem County. The average position of the interface under summer (low-flow) conditions is south of Philadelphia and the farthest upstream encroachment of saltwater is near the Benjamin Franklin Bridge.

Gradients in the downdip part of the Lower Potomac-Raritan-Magothy aquifer are steepest in eastern Camden County. Fresh ground water near the toe of the freshwater-saltwater interface moves updip at a rate of 0.002 ft/d.

Middle Potomac-Raritan-Magothy aquifer

The simulated freshwater-saltwater interface tip and toe in the Middle Potomac-Raritan-Magothy aquifer in 1988 (app. 2, fig. 2f) continue to move updip in response to past sea-level rise and ground-water withdrawals downdip, which have reversed flow gradients. The interface location is similar to that in predevelopment and has moved updip less than 0.1 mi; however, saltwater has displaced freshwater near the interface and contributes 2 Mgal/d of inflow to the aquifer as water released from storage.

The simulated freshwater-saltwater interface generally matches the available chloride concentrations from about 1988. Chloride concentrations at the Anchor Gas Dickinson well in Cape May County (number 9, fig. 1) are much greater than 10,000 mg/L, and the aquifer contains only saltwater. At the Anchor Gas Ragovin well in Cumberland County (number 7, fig. 1), chloride concentrations range from 8,700 to 11,000 mg/L, which means that the freshwater-saltwater interface is near the well. The simulated interface is near this location but probably is farther offshore than the actual interface. At the Island Beach well at the coast in Ocean County (number 6, fig. 1) chloride concentrations are lower, and water in the well is fresh. This interpretation corresponds well with the simulated location of the toe of the freshwater-saltwater interface. The chloride concentrations slightly less than 10 mi updip from the toe of the interface were between 210 and 300 mg/L at the Warren Grove well in Ocean County (number 11, fig. 1), indicating that the zone of transition in the northern part of the study area is wide. Chloride concentrations as high as 100 mg/L in wells within 1.5 mi of the Delaware River in Salem County are a result of induced recharge from the river (Schaefer, 1983, p. 44). Flow of salty water from the Delaware River is



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

Figure 32. Location of various freshwater-saltwater interface positions in the Delaware Estuary. (From Ayers and Leavesley, 1989, fig. 9)

limited to areas where withdrawal wells are located near the salty parts of the Delaware River. The high chloride concentration at well 33-251 in Salem County (1,900 mg/L) cannot be explained in terms of the ground-water flow model. Contamination probably is the result of either updip movement of saltwater from the east, upward flow from the underlying Lower Potomac-Raritan-Magothy aquifer, or local activities.

Saltwater contamination in downdip parts of the aquifer may be the result of vertical flow from the underlying Lower Potomac-Raritan-Magothy aquifer in areas where it contains salty water (see shaded areas in app. 2, fig. 2f). In most of the model cells in this area water flows from the Lower Potomac-Raritan-Magothy aquifer, which contains saltwater, into the Middle Potomac-Raritan-Magothy aquifer, which contains freshwater. As stated previously, in downdip areas the three Potomac-Raritan-Magothy aquifers behave as a single hydrologic unit and heads in the three aquifers are similar. Vertical movement of saltwater from below is more likely than horizontal movement of the freshwater-saltwater interface from the east because the vertical conductance of the confining unit is high and the distance the saltwater must travel vertically is much smaller than the distance it must travel horizontally.

Upper Potomac-Raritan-Magothy aquifer

The simulated freshwater-saltwater interface and available chloride concentrations for the Upper Potomac-Raritan-Magothy aquifer are shown in appendix 3 (fig. 3f). The interface is moving inland as it was during predevelopment conditions, but has moved less than 0.1 mi in all areas. As in the Lower and Middle Potomac-Raritan-Magothy aquifers, water is released from storage during the simulation as a result of the landward movement of saltwater and the displacement of freshwater. The amount of water contributed from storage is smaller than in the Lower and Middle Potomac-Raritan-Magothy aquifers but greater than in any of the other aquifers except the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8).

The simulated freshwater-saltwater interface matches the limited available chloride data. At the Anchor Gas Dickinson well (number 1, fig. 1) in Cape May County, all chlorides concentrations are greater than 10,000 mg/L, indicating that the aquifer contains saltwater. This well is about 10 mi downdip from the simulated interface, and chloride concentrations near 10,000 mg/L may indicate that the zone of transition downdip from the interface is wide. Updip from the interface, chloride concentrations at the Anchor Gas Ragovin well (number 7, fig. 1) in Cumberland County range from 3,300 to 8,000 mg/L, even though the well is 10 mi updip from the interface, also indicating a wide zone of transition. Chloride concentrations distant from the simulated interface both updip and downdip are about 10,000 mg/L, or half of the concentration in seawater, which indicates that the simulated position of the interface is accurate and the zone of transition is very wide. Chloride concentrations are low (1.0 mg/L) at well 29-577 in Ocean County, which is about 20 mi updip from the interface toe; this may indicate that the maximum width of the zone of transition updip from the interface is 10 to 20 mi.

The presence of high chloride concentrations at the coast in Monmouth County has been discussed in several reports (Schaefer, 1983, p. 20-23; Schaefer and Walker, 1981; Pucci and others, 1994). Pucci and others (1994) concluded that the source of the contamination in the Upper Potomac-Raritan-Magothy aquifer is ground water that flows from the area near the freshwater-saltwater interface beneath Raritan Bay, where it was located in predevelopment, rather than from the submerged outcrop of the Upper Potomac-Raritan-Magothy aquifer or a break in the overlying confining unit. The location of the freshwater-saltwater interface was not simulated in Raritan Bay with the current model. High chloride concentrations are limited to wells near the coast, and encroachment of saltwater stopped when withdrawals from the Upper Potomac-Raritan-Magothy aquifer near the coast were discontinued in the 1970's.

Chloride concentrations in downdip parts of the aquifer in Salem and Gloucester Counties are above 250 mg/L in some areas. The chloride probably is derived from vertical flow from the underlying aquifers rather than lateral movement of saltwater from the interface to the east. Simulated flow is from the Lower (where chloride concentrations exceed 250 mg/L (app. 1, fig. 1f)) to the Middle Potomac-Raritan-Magothy aquifer and then from the Middle to the Upper Potomac-Raritan-Magothy aquifer in most of Gloucester County. Vertical movement of saltwater through the relatively permeable confining units over a distance of only thousands of feet is more likely than horizontal movement of water from the interface miles away. The source of the high chloride concentration measured at well 33-253 in Salem County is unknown, but could be saltwater moving vertically through permeable underlying confining units.

Englishtown Aquifer

Available chloride-concentration data for the Englishtown aquifer are shown in appendix 4 (fig. 4f). The simulated freshwater-saltwater interface is not present in this aquifer. Chloride concentrations within the aquifer boundaries are less than 10 mg/L, although a chloride concentration of 10,300 mg/L was measured in water from the less permeable part of the Englishtown aquifer sediments in Cape May County, about 40 mi downdip from the simulated boundary of the aquifer.

Wenonah-Mount Laurel Aquifer

The simulated freshwater-saltwater interface and the available chloride-concentration data for the Wenonah-Mount Laurel aquifer are shown in appendix 5 (fig. 5f). Chloride concentrations in a borehole in Cape May County range from 1,500 to 6,400 mg/L downdip from the permeable part of the Wenonah-Mount Laurel aquifer. In the southern part of the study area, the freshwater-saltwater interface is relatively near the simulated interface, whereas in the northern part it probably is farther offshore than the simulated interface. In general, it is difficult to draw any conclusions about the location of the interface other than that it is located within the low-permeability sediments downdip from the limit of the Wenonah-Mount Laurel aquifer. The interface probably does not move horizontally over large distances in this aquifer.

Vincentown Aquifer

Available chloride-concentration data for the Vincentown aquifer are shown in appendix 6 (fig. 6f). The simulated freshwater-saltwater interface in the Vincentown aquifer is not present within the aquifer boundaries. All observed chloride concentrations within the aquifer are less than 10 mg/L. Chloride concentrations are greater in the less permeable sediments south of the aquifer's downdip limit and are as high as 14,100 mg/L offshore from Ocean County.

Piney Point Aquifer

The simulated freshwater-saltwater interface and available chloride-concentration data for the Piney Point aquifer are shown in appendix 7 (fig. 7f). The simulated freshwater-saltwater interface in the Piney Point aquifer is about 10 mi offshore from Ocean County, where a chloride concentration of 8,100 mg/L was measured, to more than 10 mi offshore from Atlantic County. Onshore chloride concentrations in this aquifer range from less than 10 mg/L in Burlington County to 330 mg/L in coastal Atlantic County.

Lower Kirkwood-Cohansey and Confined Kirkwood Aquifers

The simulated freshwater-saltwater interface and available chloride-concentration data for the lower Kirkwood-Cohansey and confined Kirkwood aquifers for 1983-88 are shown in appendix 8 (fig. 8f). As described previously, the interface is many miles offshore except in the southern part of the study area in the Delaware Bay and in the northern part near the estimated location of the updip limit of the confining unit and aquifer outcrop offshore. In these areas a thin layer of saltwater, less than 50 ft thick, remains at the bottom of the aquifer.

Chloride concentrations in samples collected at multiple depths in the AMCOR 1 well (number 1, fig. 1) range from 820 to 8,000 mg/L, indicating that water updip from the simulated toe of the interface is fresh. The 8,000-mg/L concentration was measured in one of the shallowest samples from the well. Samples from deeper parts of the aquifer contain less chloride (820-1,800 mg/L); these low values probably are a local phenomenon. Chloride concentrations at this well generally indicate that the zone of transition is wide, probably as a result of the cyclic landward and seaward movement of the interface in response to changes in sea level. The simulated interface in the aquifer moved large distances during simulation of sea-level changes. This wide mixing zone may correspond with the "finger" of saltwater that underlies freshwater in the area near the AMCOR 1 well. Chloride concentrations in samples from Wildwood Pines wells 1 and 2 (number 8, fig. 1) in Cape May County range from 90 to 200 mg/L and other measurements at wells in this area also indicate high chloride concentrations.

PREDICTIVE SIMULATION OF GROUND-WATER FLOW AND FRESHWATER-SALTWATER INTERFACE MOVEMENT, 1988-2040

An increase in ground-water withdrawals of 30 percent by 2040 was simulated to predict water levels and approximate the location of the freshwater-saltwater interface in the modeled aquifers in 2040. Heads and interface locations from the calibrated transient 1988 model were used as initial conditions. Ground-water withdrawals were increased from 216.6 Mgal/d in 1983-88 to 285.9 Mgal/d in 2040 in the confined aquifers. A 30-percent increase in average 1983-88 pumpage was applied uniformly to each well by aquifer. Recharge and boundary fluxes remained constant. The results of this simulation show the response of each aquifer to an increase in withdrawals from existing wells.

Simulated water levels in all eight confined aquifers are shown in figures 33-40. The configuration of water-level contours are similar to those simulated for 1988, but the cones of depression are broader and deeper. By 2040, water levels are, on average, about 30 ft lower near pumping centers than water levels simulated for 1988. A 30-ft decline in water levels was simulated in the Lower and Upper Potomac-Raritan-Magothy aquifers (model units A1 and A3) at a pumping center in northern Camden County; a 20-ft decline was simulated in the Middle Potomac-Raritan-Magothy aquifer (model unit A2) in the same area. A 30-ft decline in water levels in the Middle Potomac-Raritan-Magothy aquifer was simulated at a pumping center in Middlesex County and in the Upper Potomac-Raritan-Magothy aquifer in Monmouth County. Water-level declines of more than 100 ft were simulated in the Englishtown aquifer (model unit A4) along the border between Ocean and Monmouth Counties. In the Wenonah-Mount Laurel aquifer (model unit A5) simulated water levels declined 80 ft around a pumping area in Monmouth County. A 30-ft decline in water levels was simulated for a pumping center in Atlantic County in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8).

The flow budget for the predictive simulation is shown in table 8. The 55.1-Mgal/d increase in withdrawals from the aquifers of the Potomac-Raritan-Magothy aquifer system from those in 1988 resulted in an increase in downward leakage to the Lower and Middle Potomac-Raritan-Magothy aquifers. Vertical flow to the Lower Potomac-Raritan-Magothy aquifer from the overlying Middle Potomac-Raritan-Magothy aquifer increased 30 percent from the 1988 simulation; vertical flow from the overlying unconfined aquifers to the Middle Potomac-Raritan Magothy aquifer increased 32 percent. Lateral flow from the outcrop or downdip from unconfined aquifers to the Upper Potomac-Raritan-Magothy aquifer increased about 41 percent.

Although a comparison of the interface locations between the maps of simulated results for 1988 (figs. 17, 19, 21, 25, 29, and 31) and the maps of simulated results for 2040 (figs. 33, 34, 35, 37, 39, and 40) shows no apparent movement of the freshwater-saltwater interface, water levels declined near the interface. The water-level declines signify changes in the gradient near the interface that may affect the salinity at wells near the interface. A 20-ft decline was simulated near the interface in the Lower Raritan-Potomac-Magothy aquifer. A 30-ft decline was simulated in the Middle and Upper Potomac-Raritan-Magothy aquifers and the Wenonah-Mount Laurel aquifer. A 10-ft decline was simulated near the interface in the Piney Point and lower Kirkwood-Cohansey and confined Kirkwood aquifers. The interface in the latter two model units is located offshore.

In general, the simulated freshwater-saltwater interface did not move significantly in any aquifer during 1988-2040. Simulated movement was greatest (30 ft) inland in the Lower Potomac-Raritan-Magothy aquifer in Ocean County. The interface moved less than 5 ft inland in the other confined aquifers.

SUMMARY AND CONCLUSIONS

The confined aquifers of the New Jersey Coastal Plain are composed of sand, range from 50 to 600 ft in thickness, and are separated by confining units. The confining units are composed of silt and clay and range in thickness from 50 to 1,000 ft. The aquifers are recharged by precipitation on their outcrop areas that flows laterally downdip. Water in the confined aquifers ultimately discharges to the Raritan and the Delaware Bays and to the Atlantic Ocean.

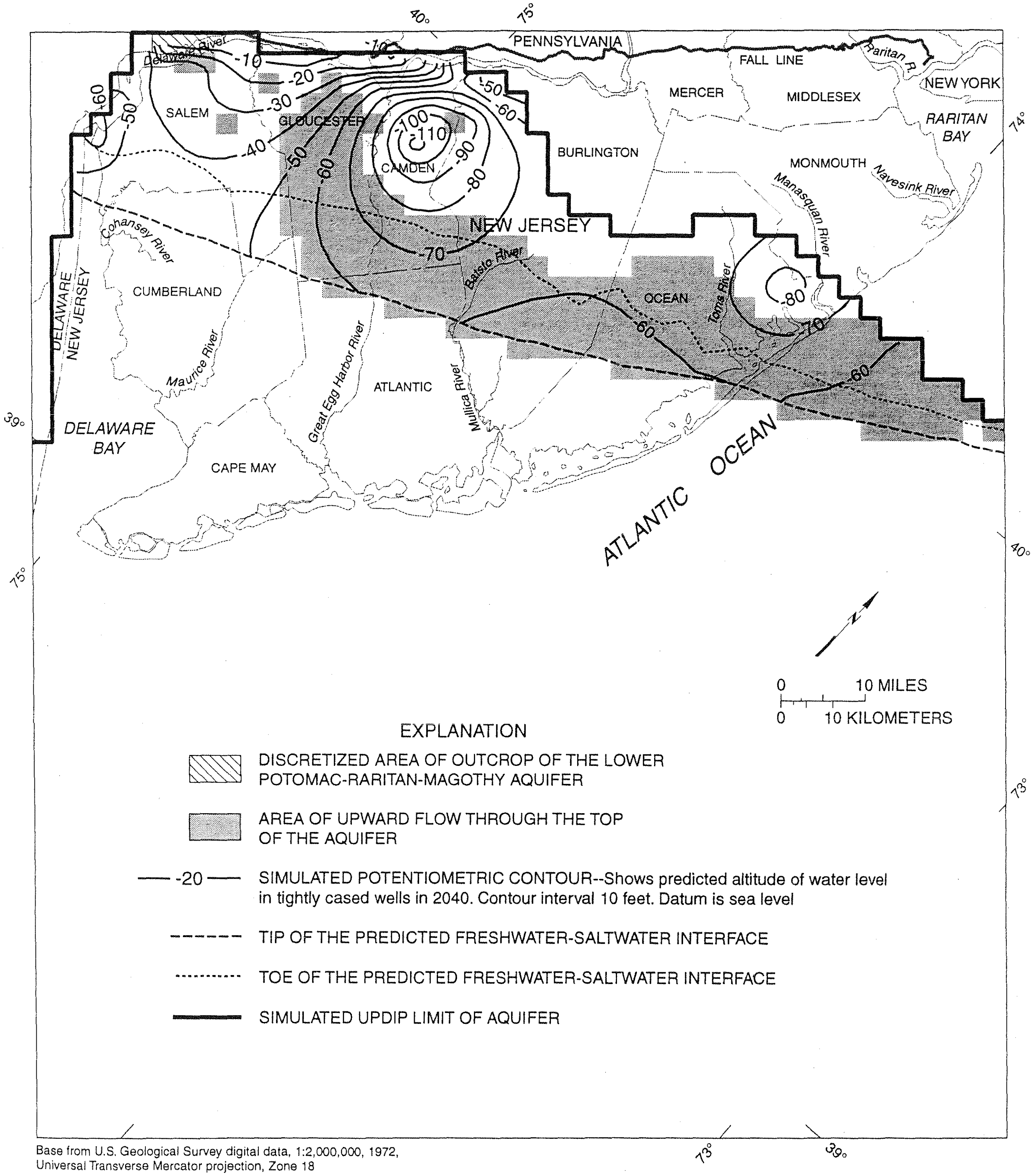
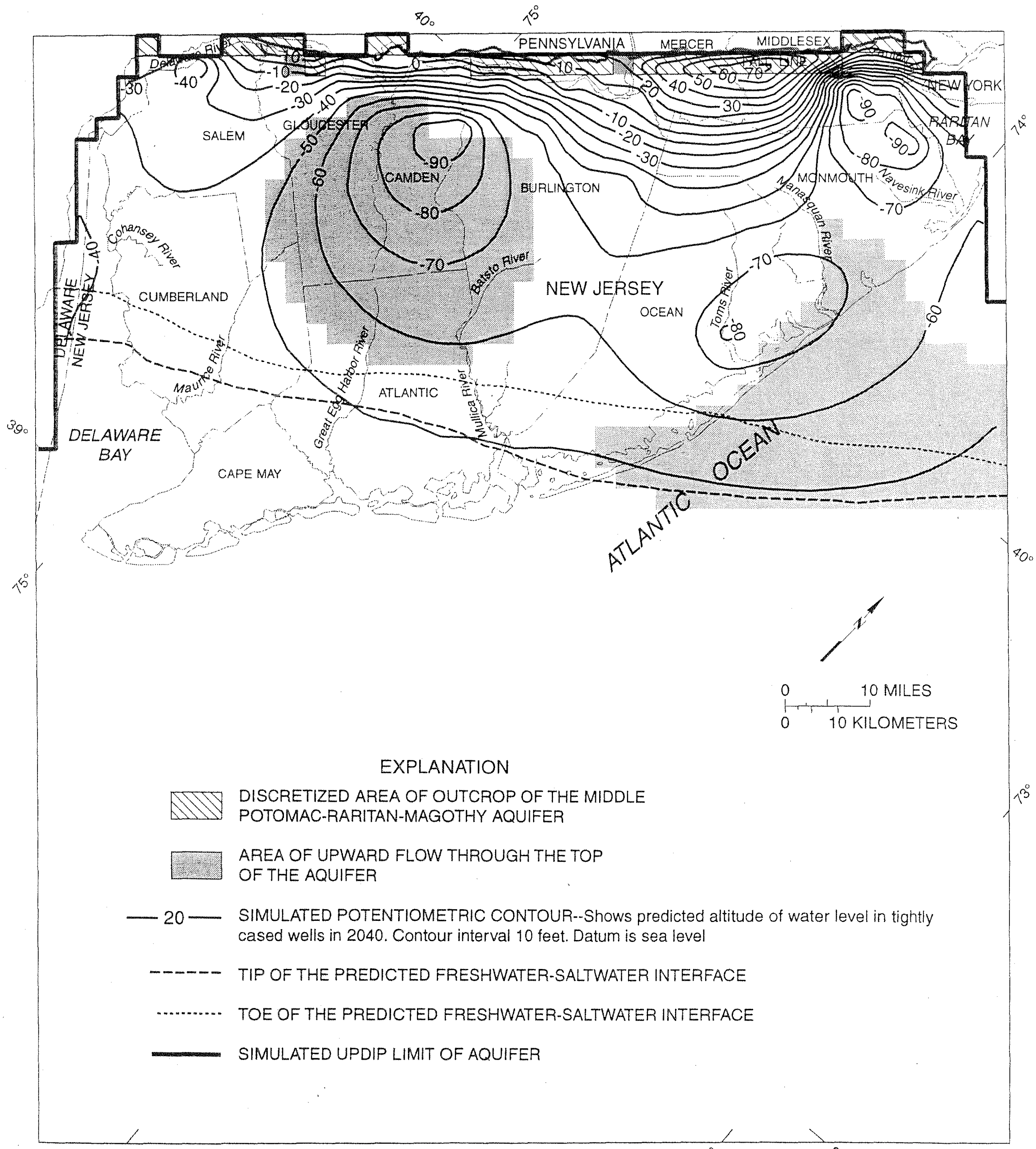


Figure 33. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Lower Potomac-Raritan-Magothy aquifer (model unit A1), New Jersey Coastal Plain, 2040.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 34. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Middle Potomac-Raritan-Magothy aquifer (model unit A2), New Jersey Coastal Plain, 2040.

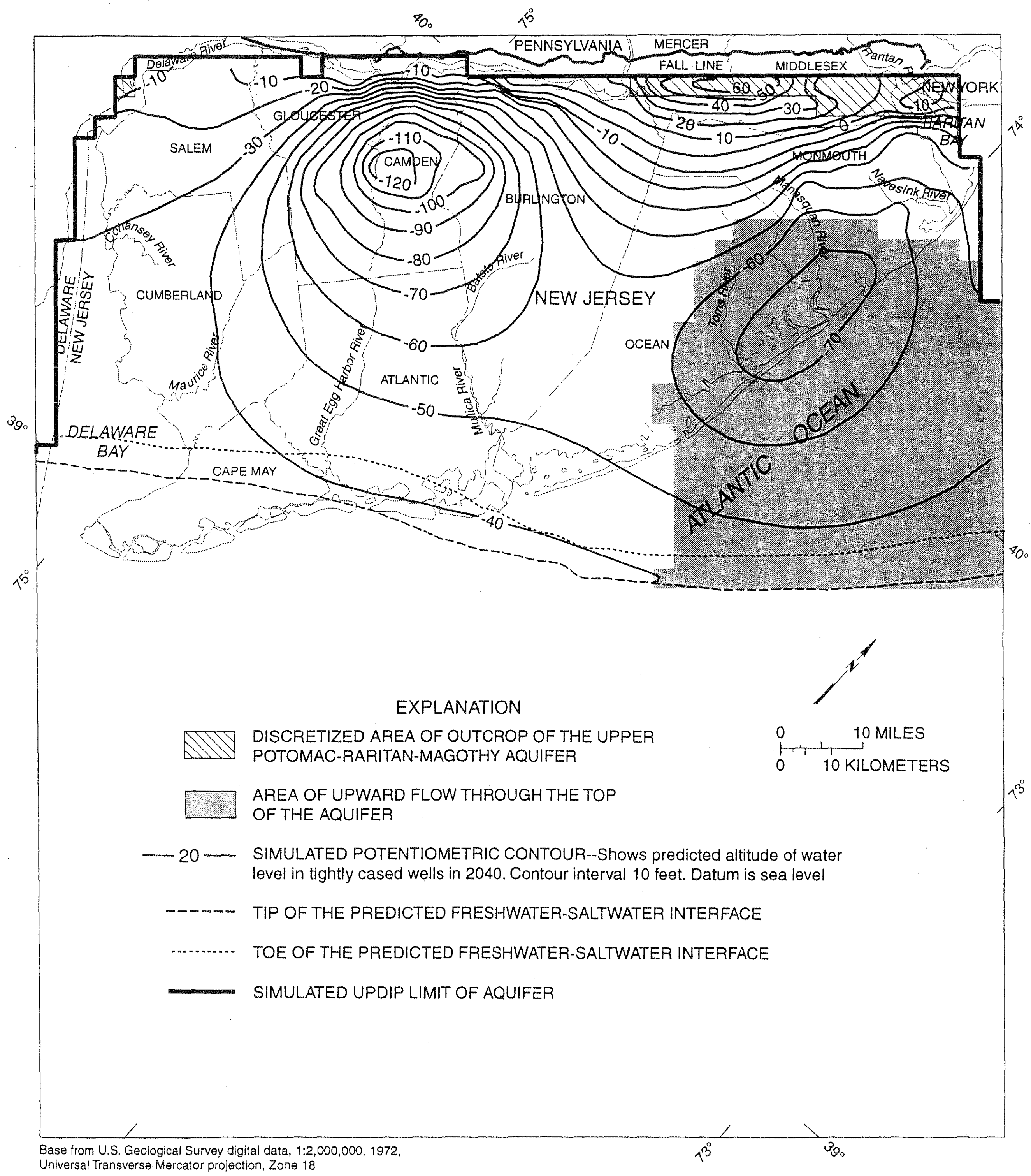
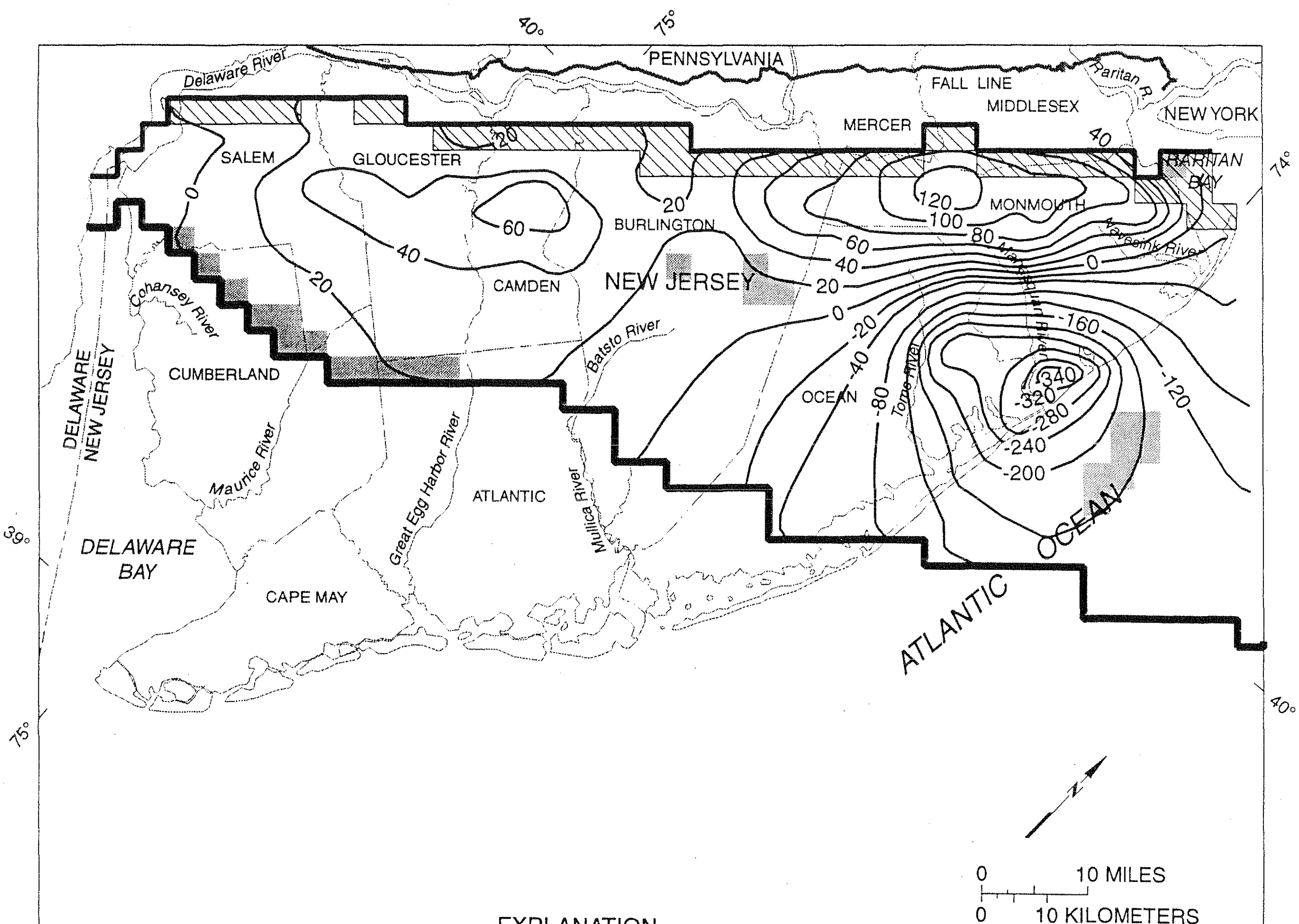


Figure 35. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Upper Potomac-Raritan-Magothy aquifer (model unit A3), New Jersey Coastal Plain, 2040.



EXPLANATION

DISCRETIZED AREA OF OUTCROP OF THE ENGLISHTOWN AQUIFER

AREA OF UPWARD FLOW THROUGH THE TOP OF THE AQUIFER

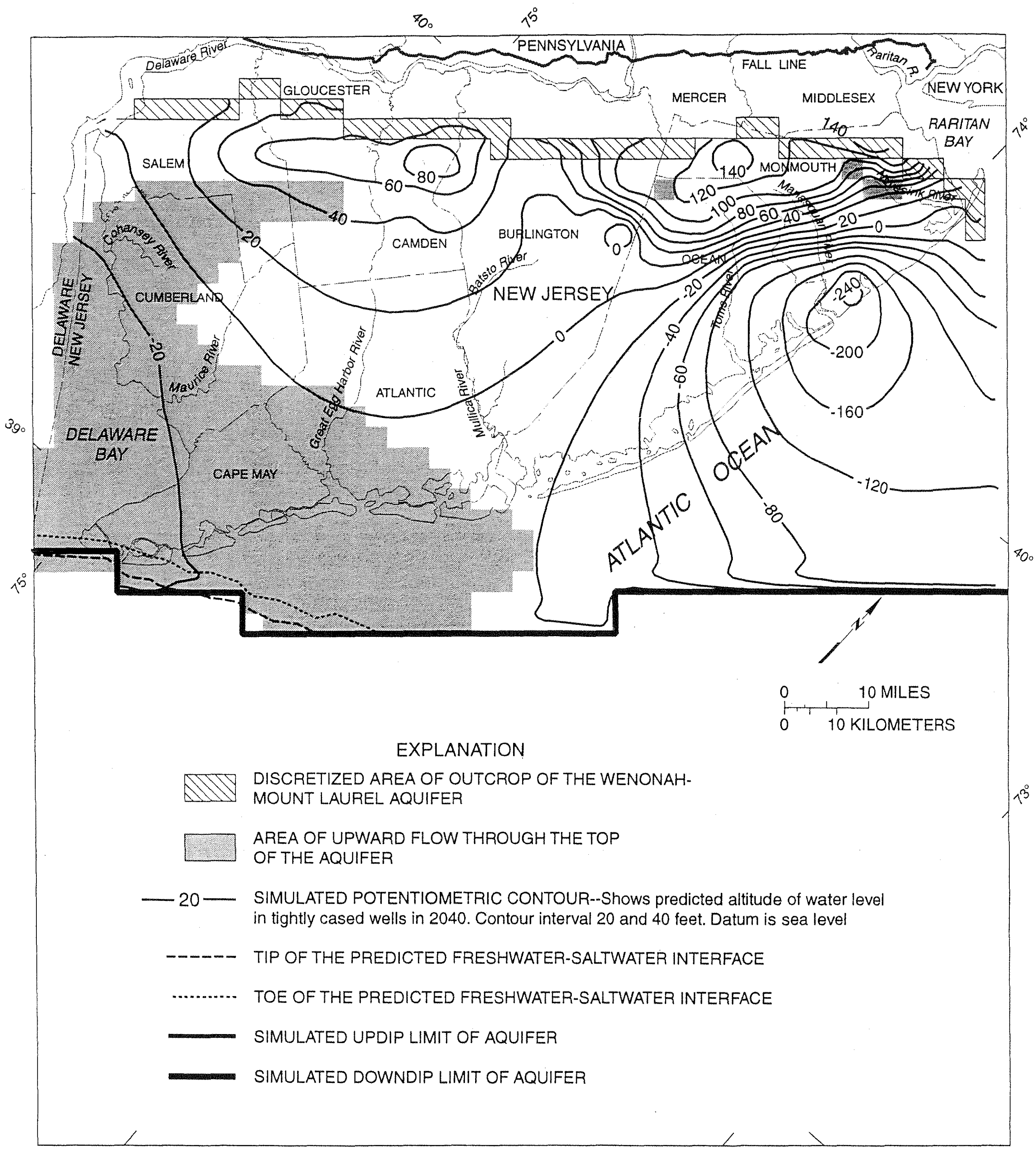
— 20 — SIMULATED POTENTIOMETRIC CONTOUR--Shows predicted altitude of water level in tightly cased wells in 2040. Contour interval 20 and 40 feet. Datum is sea level

SIMULATED UPDIP LIMIT OF AQUIFER

SIMULATED DOWNDIP LIMIT OF AQUIFER

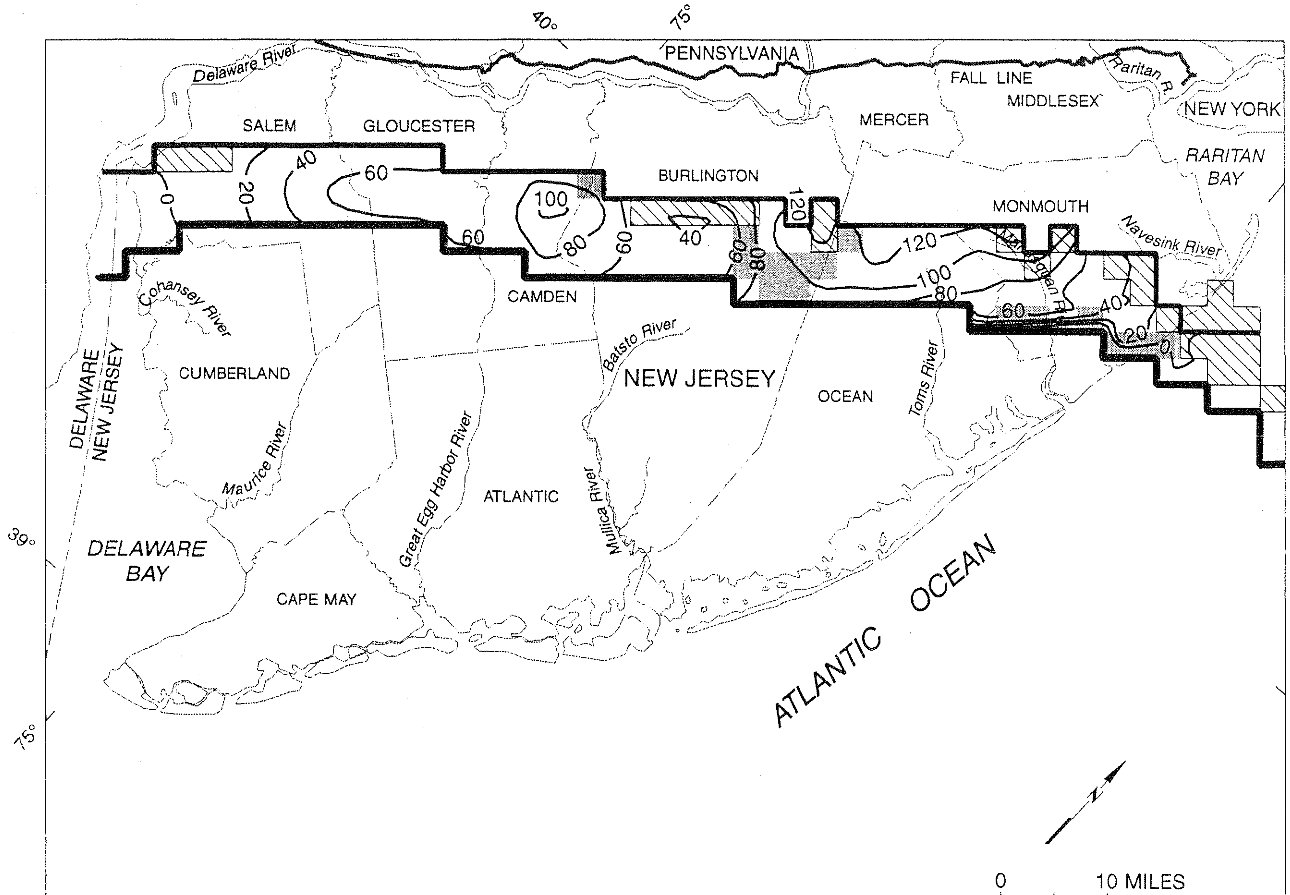
Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 36. Simulated potentiometric surface and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Englishtown aquifer, (model unit A4), New Jersey Coastal Plain, 2040.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 37. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Wenonah-Mount Laurel aquifer (model unit A5), New Jersey Coastal Plain, 2040.



EXPLANATION

DISCRETIZED AREA OF OUTCROP OF THE VINCENTOWN AQUIFER

AREA OF UPWARD FLOW THROUGH THE TOP OF THE AQUIFER

—— -20 —— SIMULATED POTENTIOMETRIC CONTOUR--Shows predicted altitude of water level in tightly cased wells in 2040. Contour interval 20 feet. Datum is sea level

SIMULATED UPDIP LIMIT OF AQUIFER

SIMULATED DOWNDIP LIMIT OF AQUIFER

Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

Figure 38. Simulated potentiometric surface and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Vincentown aquifer (model unit A6), New Jersey Coastal Plain, 2040.

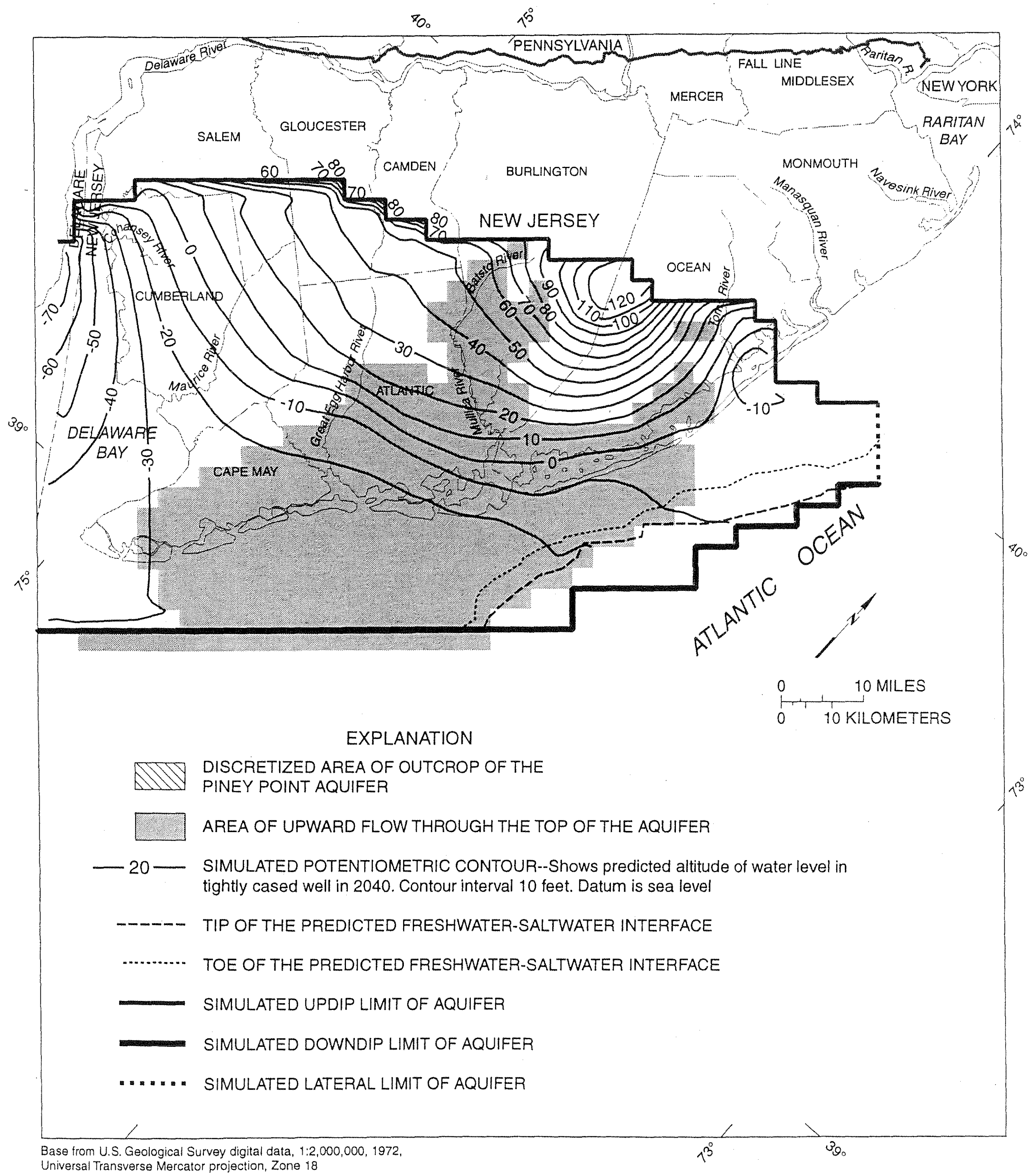


Figure 39. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, Piney Point aquifer (model unit A7), New Jersey Coastal Plain, 2040.

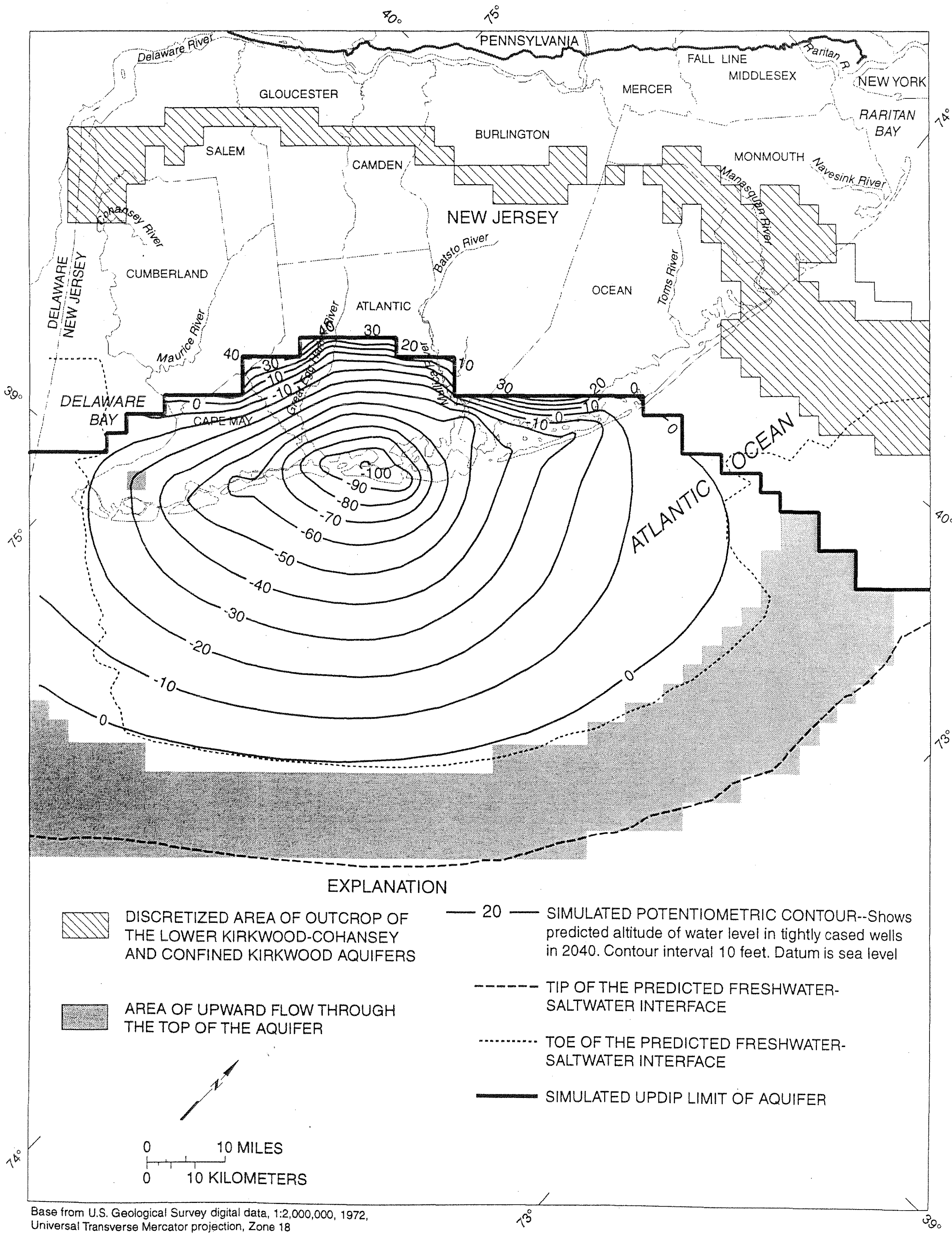


Figure 40. Simulated potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow resulting from a 30-percent increase in withdrawals at all wells, lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8), New Jersey Coastal Plain, 2040.

Table 8. Simulated flow budgets of confined aquifers in the New Jersey Coastal Plain, 2040

[Values are in million gallons per day]

Model unit	Inflow							
	Storage	Flow downdip from unconfined aquifer	Saltwater	Leakage from overlying unconfined aquifer	Leakage from overlying confined aquifer	Leakage from underlying confined aquifer	Flow across model boundary	Total
Lower Potomac-Raritan-Magothy aquifer (A1)	5	2.3	0	3.6	80.9	0	0.4	92.2
Middle Potomac-Raritan-Magothy aquifer (A2)	1.7	20.9	.2	123.1	21.7	4.0	2.7	174.3
Upper Potomac-Raritan-Magothy aquifer (A3)	1.2	35.8	.9	23.1	22.2	4.7	2.7	90.6
Englishtown aquifer (A4)	0	.2	.4	10.6	23.4	.3	.4	35.3
Wenonah-Mount Laurel aquifer (A5)	.1	.7	.4	9	22.1	.1	.1	32.5
Vincentown aquifer (A6)	0	3.4	.2	19	0	.1	0	22.7
Piney Point aquifer (A7)	.2	0	1.1	8.9	.1	.6	0	10.9
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8)	2.2	17.8	3.9	2.1	.1	3.1	.8	30

Model unit	Outflow								
	Storage	Flow downdip to unconfined aquifer	Saltwater	Leakage to overlying unconfined aquifer	Leakage to overlying confined aquifer	Leakage to underlying confined aquifer	Flow across model boundary	Pumpage	Total
Lower Potomac-Raritan-Magothy aquifer (A1)	0.4	0	0	0	4.0	0	5.1	82.7	92.2
Middle Potomac-Raritan-Magothy aquifer (A2)	0	.1	0	.2	4.7	80.9	3.5	84.9	173.3
Upper Potomac-Raritan-Magothy aquifer (A3)	0	.3	0	0	.3	21.6	.4	68	90.6
Englishtown aquifer (A4)	0	1.6	.2	0	.1	21.6	0	11.8	35.3
Wenonah-Mount Laurel aquifer (A5)	0	1.3	0	.2	.6	23.9	.1	6.4	32.5
Vincentown aquifer (A6)	0	1	.2	1.3	0	19.5	0	.7	22.7
Piney Point aquifer (A7)	.1	0	0	.9	3.1	.9	3.6	2.3	10.9
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (A8)	.1	.1	.1	0	0	.1	.5	29.1	30

In 1988, withdrawals from confined and unconfined aquifers in the New Jersey Coastal Plain were approximately 345 Mgal/d, of which more than 75 percent was from the confined aquifers. The development of ground-water resources, primarily near large population centers, has created large regional cones of depression in several of the Coastal Plain aquifers. Continued decline of water levels in confined aquifers poses the threat of serious adverse effects, including saltwater intrusion and reduction of stream discharge near the aquifer outcrop areas.

In order to better identify areas in which ground-water supplies may be threatened by saltwater intrusion, the U.S. Geological Survey (USGS), in cooperation with the New Jersey Department of Environmental Protection, conducted a study of the ground-water flow and freshwater-saltwater interface movement in the New Jersey Coastal Plain.

Ground-water flow and the saltwater-freshwater interface were simulated by using a sharp-interface flow model that is based on a model previously developed as part of the USGS New Jersey Regional Aquifer System Analysis (RASA) program. The lower model boundary is a no-flow boundary representing the top of the underlying crystalline bedrock. The upper boundary is a constant-head boundary representing streams (and recharge) in the aquifer outcrop areas onshore and ocean levels in aquifer outcrop areas offshore. The downdip boundary of the model was changed from a no-flow boundary in the New Jersey RASA model to a direct simulation of the saltwater interface in all aquifers. In the uppermost aquifers, a constant saltwater-head boundary was used to represent presumed outcrops of the aquifers at the downdip limit of the Continental Shelf. A no-flow boundary was applied in the northwestern part of the study area at the updip limit of the Coastal Plain sediments (the Fall Line). Specified-flux boundaries applied at the north-eastern and southwestern boundaries represent flow to the Raritan and the Delaware Bays, respectively.

Water levels and the location of the freshwater-saltwater interface were simulated (1) under steady-state conditions in which sea level was relatively constant from 110,000 to 84,000 years ago; (2) from 84,000 years ago to 1896, including the effects of glaciation on the rise and fall of sea level; (3) under transient conditions from 1896 to 1988 by including ground-water withdrawals from the Coastal Plain aquifers; and (4) in a predictive simulation to the year 2040 with a 30-percent increase in withdrawals at pumping centers existing in 1988. From 110,000 to 84,000 years ago, the freshwater-saltwater interface stabilized in equilibrium with a sea level 27 ft below current sea level. Sea level fell from -27 ft 84,000 years ago to -230 ft 71,000 years ago, which caused the interface to move offshore. As sea level rose beginning 71,000 years ago, the fresh-water-saltwater interface began moving landward. The interface has not yet stabilized and is still moving landward in response to this rise in sea level. The freshwater-saltwater interface is considered to be the boundary seaward of which chloride concentrations equal or exceed 10,000 mg/L.

The model was then used to simulate conditions from 84,000 years ago to 1896 by adjusting areas receiving recharge (onshore areas) and boundary conditions to account for the rise and fall of sea level during the glacial periods. The model was calibrated to present-day (1988) conditions by simulating water levels measured in fall 1988, by matching simulated water levels to hydrographs at 141 observation wells, and by approximating the location of the freshwater-saltwater interface from chloride-concentration measurements. Under pumping conditions, regional cones of depression developed in Camden County in the Lower Potomac-Raritan-

Magothy aquifer; in Camden and Monmouth Counties in the Middle and Upper Potomac-Raritan-Magothy aquifers; and in southern Monmouth County in the Englishtown and Wenonah-Mount Laurel aquifers.

The results of the predevelopment simulation from 84,000 years ago to 1896, the transient simulation from 1896 to 1988, and the predictive simulation from 1988 to 2040 show minimal movement of the interface between freshwater and saltwater in response to increased ground-water withdrawals. The position and movement of the freshwater-saltwater interface are more dependent on historical sea level than on the stressed flow system resulting from ground-water withdrawals from the Coastal Plain aquifers from 1896 to 1988. The freshwater-saltwater interface has not reached equilibrium with current sea level because the interface moves seaward in response to a fall in sea level faster than it moves landward after a sea-level rise.

Results of the predictive simulation to 2040 indicate that additional withdrawals from each of the eight confined aquifers in the Coastal Plain would broaden and deepen the existing cones of depression. Because the withdrawals were increased at current pumping locations (those in use during 1983-88), the configurations of the potentiometric surfaces of the aquifers were not significantly different from those in 1988. Compared with the 1988 potentiometric surfaces, a 30-percent increase in ground-water withdrawals caused drawdowns of 30 ft at the centers of the cones of depression in the Lower, Middle, and Upper Potomac-Raritan-Magothy aquifers in Camden and Ocean Counties; caused drawdowns of more than 80 ft at the center of the cone of depression in the Wenonah-Mount Laurel and Englishtown aquifers in Monmouth County; and caused drawdowns of 30 ft in the lower Kirkwood-Cohansey and confined Kirkwood aquifers in Cape May County. The simulated interface did not move significantly in any of the confined aquifers.

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Table 4. Simulated water-level altitudes for 1988, average measured water-level altitudes for 1983-88, and the difference between simulated and measured water-level altitudes for selected wells, New Jersey Coastal Plain

U.S. Geological Survey well number	Simulated water-level altitude (feet below land surface)	Average measured water-level altitude (feet below land surface)	Difference (simulated-measured) (feet)
Lower Potomac-Raritan-Magothy aquifer (model unit A1)			
5-262	-50.39	-54.35	3.96
5-645	-38.24	-36.35	-1.89
7-283	-60.44	-72.72	12.29
7-354	-18.99	1.84	-20.82
7-412	-68.41	-73.42	5.01
15-296	-25.12	-17.52	-7.60
15-323	-30.93	-46.15	15.22
33-187	-19.42	-26.38	6.96
Middle Potomac-Raritan-Magothy aquifer (model unit A2)			
5- 63	-20.98	-18.96	-2.02
5-101	5.95	2.64	3.31
5-261	-47.53	-54.45	6.93
5-274	-23.92	-25.84	1.92
5-440	-17.71	-30.24	12.53
5-683	-32.11	-34.75	2.64
7-413	-62.07	-78.50	16.44
7-476	-50.21	-52.35	2.14
11-137	-25.68	-44.69	19.01
15- 97	-2.82	-1.18	-1.64
23-194	-53.28	-52.76	-0.52
23-229	45.90	52.80	-6.90
25-272	-53.55	-59.73	6.18
29- 19	-24.89	-6.98	-17.92
29- 85	-38.62	-31.30	-7.32
33-251	-24.18	-29.33	5.15
Upper Potomac-Raritan-Magothy aquifer (model unit A3)			
5-258	-53.15	-61.80	8.65
7- 30	-15.95	-18.00	2.06
7-117	-74.36	-87.21	12.85
7-477	-64.61	-71.94	7.34
15-297	-7.35	-10.72	3.38
23-182	15.02	16.20	-1.18
25-206	-14.54	-14.24	-0.29
33-253	-12.84	-24.67	11.83

Table 4. Simulated water-level altitudes for 1988, average measured water-level altitudes for 1983-88, and the difference between simulated and measured water-level altitudes for selected wells, New Jersey Coastal Plain--Continued

U.S. Geological Survey well number	Simulated water-level altitude (feet below land surface)	Average measured water-level altitude (feet below land surface)	Difference (simulated-measured) (feet)
Englishtown aquifer (model unit A4)			
5-259	30.87	20.95	9.92
25-250	97.18	100.43	-3.25
25-429	-159.75	-140.74	-19.01
29-138	57.72	62.41	-4.69
29-503	-192.51	-193.70	1.20
29-534	-57.86	-85.36	27.49
Wenonah-Mount Laurel aquifer (model unit A5)			
7-118	81.42	69.33	12.09
7-478	48.38	35.42	12.95
11- 72	1.90	9.02	-7.12
25-353	-9.13	-11.14	2.01
25-486	-163.71	-176.18	12.47
29-140	114.85	113.63	1.22
33- 20	43.87	32.76	11.12
33-252	7.42	1.19	6.23
Vincentown aquifer (model unit A6)			
29-139	121.78	130.04	-8.27
Piney Point aquifer (model unit A7)			
5-407	55.91	51.34	4.56
5-676	108.10	118.69	-10.59
11- 44	14.65	11.31	3.34
11- 96	-16.36	-22.23	5.88
11-163	11.31	12.75	-1.44
29- 18	5.54	0.44	5.10
29-425	114.38	118.94	-4.57

Table 4. Simulated water-level altitudes for 1988, average measured water-level altitudes for 1983-88, and the difference between simulated and measured water-level altitudes for selected wells, New Jersey Coastal Plain--Continued

U.S. Geological Survey well number	Simulated water-level altitude (feet below land surface)	Average measured water-level altitude (feet below land surface)	Difference (simulated-measured) (feet)
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8)			
1- 37	-68.45	-75.02	6.57
1-180	-38.63	-32.73	-5.90
1-566	-54.66	-43.94	-10.72
1-578	-52.00	-54.78	2.78
1-703	-37.62	-38.79	1.18
1-706	-26.22	-20.37	-5.85
5- 30	68.60	65.44	3.15
7-479	113.51	109.94	3.56
11- 43	70.58	76.97	-6.39
11- 97	12.31	1.13	11.18
15-792	119.80	97.28	22.52
29- 17	0.65	4.38	-3.73
29-514	37.06	36.43	0.63

Table 6. Concentrations of chloride at various depths in deep wells and boreholes in onshore and offshore areas, New Jersey Coastal Plain, 1888-1974

[Data from Meisler (1989); mg/L as Cl, milligrams per liter as chloride; 00 in sample date indicates unknown day]

Well number (fig. 1)	Local well identifier	Chloride concentration (mg/L as Cl)	Sample depth (feet below sea level)	Sample date	Model unit
1	Amcor 6011	16,200	103	8-17-76	A9
		16,200	241	8-17-76	A9
		8,000	306	8-17-76	A8
		820	305	8-17-76	A8
		1,800	337	8-17-76	A8
		1,600	460	8-17-76	A8
		1,000	523	8-17-76	A8
		1,160	585	8-17-76	A8
		1,350	648	8-17-76	A7
		2,090	710	8-17-76	A7
		8,100	802	8-17-76	A7
		2,720	895	8-17-76	A7
		1,800	926	8-17-76	A7
2	Amcor 6020	14,950	154	9-13-76	A9
		12,800	211	9-13-76	A9
		10,900	240	9-13-76	A9
		4,020	272	9-13-76	A9
3	Amcor 6009, 6009B	17,000	213	8-00-76	A9
		18,200	243	8-00-76	A9
		11,260	273	8-00-76	A9
		6,340	304	8-00-76	A9
		5,140	335	8-00-76	A9
		3,700	335	8-00-76	A9
		3,790	429	8-00-76	A9
		3,520	460	8-00-76	A9
		3,170	460	8-00-76	A9
		3,390	491	8-00-76	A9
		3,530	522	8-00-76	A9
		3,7,70	554	8-00-76	A9
		4,830	616	8-00-76	A9
		4,960	647	8-00-76	A9
		5,640	709	8-00-76	A9
		7,330	771	8-00-76	A9
		6,190	832	8-00-76	A9
		9,070	926	8-00-76	A9
		7,210	1,018	8-00-76	A9
		10,970	1,081	8-00-76	A9
		14,270	1,112	8-00-76	A8

Table 6. Concentrations of chloride at various depths in deep wells and boreholes in onshore and offshore areas, New Jersey Coastal Plain, 1888-1974--Continued

Well number (fig. 1)	Local well identifier	Chloride concentration (mg/L as Cl)	Sample depth (feet below sea level)	Sample date	Model unit
4	Amcor 6010	17,600	276	8-15-76	A9
		15,800	335	8-15-76	A9
		14,000	397	8-15-76	A9
		15,500	522	8-15-76	A9
		14,600	585	8-15-76	A9
		13,900	772	8-15-76	A9
		13,400	832	8-15-76	A9
		13,200	957	8-15-76	A9
		13,500	1,019	8-15-76	A9
		13,400	1,081	8-15-76	A9
		14,100	1,144	8-15-76	A6
		18,000	1,237	8-15-76	A8
5	Amcor 6021	17,600	1,001	9-16-76	A9
		18,000	1,013	9-16-76	A9
		18,400	1,028	9-16-76	A9
		18,100	1,072	9-16-76	A9
		18,000	1,196	9-16-76	A9
		18,000	1,258	9-16-76	A9
		18,200	1,414	9-16-76	A9
		18,000	1,502	9-16-76	A9
		18,000	1,634	9-16-76	A9
		18,000	1,728	9-16-76	A9
		18,200	1,821	9-16-76	A9
6	U.S. Geological Survey Island Beach Test Well	6	390	11-17-77	A7
		780	2,730	5-24-67	A2
		1,100	2,740	5-07-62	A2
		2,700	2,860	5-07-62	A2
		5,400	3,030	5-07-62	A1
		7,000	3,090	5-07-62	A1
		7,000	3,170	5-07-62	A1
		11,300	3,370	5-07-62	A1
		11,300	3,480	5-07-62	A1

Table 6. Concentrations of chloride at various depths in deep wells and boreholes in onshore and offshore areas, New Jersey Coastal Plain, 1888-1974--Continued

Well number (fig. 1)	Local well identifier	Chloride concentration (mg/L as Cl)	Sample depth (feet below sea level)	Sample date	Model unit
7	Anchor Gas, Ragovin	11,000	1,980	10-22-74	A2
		12,000	2,490	10-15-74	A1
		18,000	3,010	10-08-74	A1
		22,000	3,180	10-01-74	A1
		27,000	3,310	9-24-74	A1
		600	860	10-31-64	A6
		3,300	1,230	10-31-64	A3
		8,000	1,480	10-31-64	A3
		8,700	1,840	10-31-64	A2
		11,600	2,410	10-31-64	A1
		12,400	2,660	10-31-64	A1
		16,700	2,940	10-31-64	A1
		25,900	3,030	10-31-64	A1
		27,400	3,110	10-31-64	A1
		27,400	3,240	10-31-64	A1
8	Wildwood Pines 2	90	350	4-21-66	A8
	Wildwood Pines 1	200	880	8-22-61	A8
9	Anchor Gas, Dickinson 1	1,300	1,840	7-14-63	A6
		1,500	2,100	7-14-63	A5
		6,400	2,210	7-14-63	A5
		10,300	2,420	7-14-63	A4
		10,300	2,580	7-14-63	A3
		11,600	2,690	7-14-63	A3
		21,000	3,010	7-14-63	A2
		28,800	3,110	7-14-63	A2
		28,000	3,680	7-14-63	A2
		32,200	3,910	7-14-63	A1
		38,100	4,360	7-14-63	A1
		22,100	5,200	7-14-63	A1
		22,100	5,380	7-14-63	A1
10	Atlantic City	330	1,020	00-00-1888	A7
11	Warren Grove	210	1,515	11-00-82	A2
		300	1,785	11-00-82	A2

Table 7. Concentrations of chloride in selected samples from wells in the New Jersey Coastal Plain, 1971-90

[mg/L, milligrams per liter; --, missing data]

U.S. Geological Survey well number	New Jersey permit number	Latitude	Longitude	Well depth	Sample date	Chloride concentration (mg/L)
Lower Potomac-Raritan-Magothy aquifer (model unit A1)						
11-133	--	392512	745212	3,410	09/24/1974	27,000
15-139	30-01223	394608	752135	345	11/10/1986	820
15-308	--	395044	751242	271	09/26/1985	79
15-312	51-00063	395107	750946	372	12/17/1986	42
15-324	31-00036	395236	750821	224	11/19/1982	21
15-349	--	394650	752316	220	10/01/1980	110
15-398	30-02016	394935	751938	60	11/17/1986	140
15-671	--	394957	750530	670	06/03/1986	10
15-680	30-03602	395038	751605	196	09/22/1986	73
15-712	30-04347	394808	751724	295	03/19/1987	660
33-137	50-00003	394112	753028	361	08/17/1988	76
33-187	--	394037	751914	672	10/07/1985	170
33-346	30-00563	394256	752718	357	08/22/1989	220
Middle Potomac-Raritan-Magothy aquifer (model unit A2)						
5-683	--	395122	743017	2,120	01/28/1988	3.5
11-137	--	392514	745217	2,090	10/22/1974	11,000
15- 72	30-00037	394936	751747	101	11/08/1984	87
15-140	30-01248	394608	752135	184	11/20/1985	33
15-166	30-00410	394755	752108	88	06/16/1986	14
15-374	31-13385	394843	750728	489	07/12/1985	15
15-780	31-26244	395223	751117	90	08/27/1987	34
25-153	29-05942	402444	741010	690	09/19/1985	2.3
25-320	--	402705	735959	878	08/11/1989	5.0
29- 19	--	394829	740535	2,760	12/12/1984	850
29-626	33-10224	395721	741230	1,870	08/18/1989	.9
33-106	--	393514	752917	366	10/10/1980	460
33-108	30-00052	393641	753322	319	08/15/1989	100
33-127	30-00698	394100	753030	188	09/25/1986	69
33-251	--	393348	752755	709	11/22/1982	1,900
33-364	34-01031	392743	753158	840	08/23/1990	26
33-459	30-03336	393928	752147	457	08/15/1989	16

Table 7. Concentrations of chloride in selected samples from wells in the New Jersey Coastal Plain, 1971-90--Continued

U.S. Geological Survey well number	New Jersey permit number	Latitude	Longitude	Well depth	Sample date	Chloride concentration (mg/L)
Upper Potomac-Raritan-Magothy aquifer (model unit A3)						
15- 03	31-06676	394015	750559	740	08/18/1987	110
15- 60	31-02358	394206	750758	612	09/25/1984	69
15-129	50-00049	394409	751330	263	09/26/1984	170
15-194	31-05309	394732	751037	265	08/28/1989	34
15-231	--	394147	751651	358	10/20/1980	22
15-253	31-04741	394437	750249	652	07/24/1985	22
15-332	51-00100	395009	750922	188	10/29/1986	31
15-353	--	394649	752316	17.50	04/18/1985	67
15-519	30-01788	394649	751738	87	11/18/1986	15
15-626	30-33900	394729	752101	19	12/05/1986	14
25- 13	29-07461	401137	740121	1,160	09/21/1983	2.1
25- 34	--	401558	740908	836	05/06/1985	2.5
25-119	29-06480	402403	735923	779	07/15/1988	1.1
25-294	49-00042	402428	741345	252	10/06/1986	2.0
25-321	--	402706	735952	486	09/01/1977	42
25-419	29-03786	402632	741049	285	04/25/1986	1,300
25-462	29-05558	402717	740816	260	08/07/1985	45
25-560	28-11985	401904	742102	306	06/26/1985	2.0
29-577	33-05553	395741	740437	1,500	08/09/1988	1.0
33- 74	30-01151	394241	752201	206	10/03/1980	1.7
33-117	30-00451	393954	753013	102	09/19/1985	19.0
33-253	--	393348	752755	340	11/22/1982	670.0
33-360	28-10466	393750	753131	125	07/25/1986	7.2
33-370	30-01800	394449	752554	52	07/17/1986	8.7
Englishtown aquifer (model unit A4)						
25- 01	29-00116	401401	740025	570	08/25/1989	1.3
25- 09	49-00050	402441	740234	200	08/20/1980	5.6
25- 30	29-00069	400645	740345	750	09/01/1987	.8
25-386	29-04721	400952	740149	670	09/01/1987	.6
25-563	28-07643	401703	742304	49	05/30/1985	6.6
25-638	29-18401-1	401105	741202	499	10/06/1987	2.8
29-138	--	400414	742702	427	05/19/1983	1.9
29-454	53-00002	395808	740421	1,140	08/17/1989	1.9

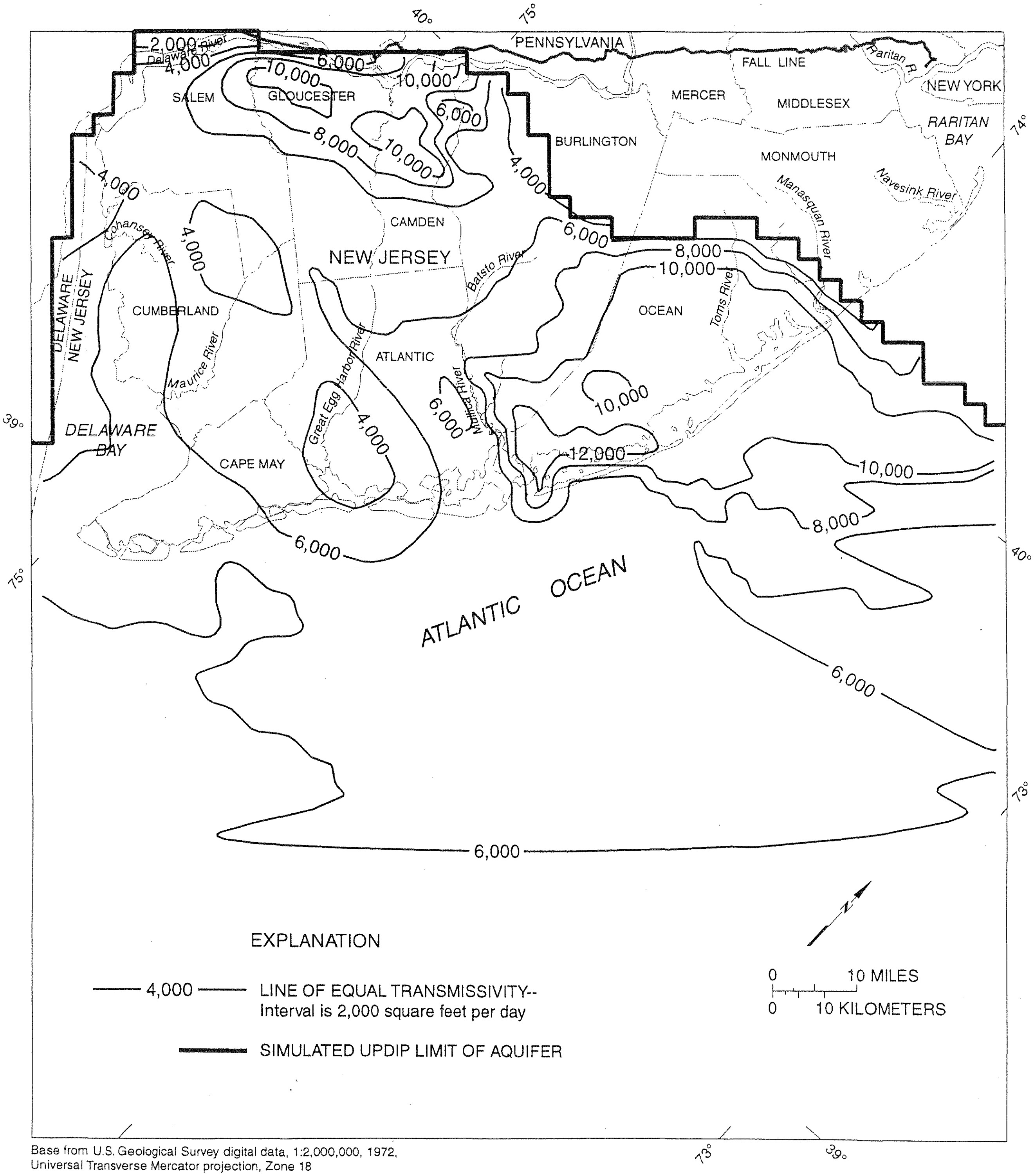
Table 7. Concentrations of chloride in selected samples from wells in the New Jersey Coastal Plain, 1971-90--Continued

U.S. Geological Survey well number	New Jersey permit number	Latitude	Longitude	Well depth	Sample date	Chloride concentration (mg/L)
			Wenonah-Mount Laurel aquifer (model unit A5)			
11- 66	--	392124	751904	310	03/17/1972	227
25- 11	19-00018	401136	740120	501	09/02/1987	2.1
25-142	29-01935	402255	741010	121	08/02/1971	41
25-243	--	401854	741325	80	08/02/1971	3.9
33- 33	--	392751	752441	340	11/07/1984	42
33- 35	34-00757	392744	753206	281	08/22/1989	390
33-279	--	393622	751531	425	03/13/1985	1.9
33-426	--	393451	752718	127	09/29/1981	17
			Vincentown aquifer (model unit A6)			
25-636	29-18404-5	401105	741202	100	10/15/1987	9.0
29-139	28-04784	400414	742702	171	05/23/1983	1.9
33-368	--	393253	752425	133	10/04/1974	6.5
			Piney Point aquifer (model unit A7)			
1-701	35-03992	393148	745617	460	02/18/1986	27
1-713	35-04656	392902	745051	535	10/10/1985	110
5-407	--	394422	744309	260	02/20/1986	1.2
5-676	--	394914	742546	540	09/23/1985	4.8
11- 44	35-01197	392732	750929	376	06/26/1984	45
11- 61	34-01191	391926	751921	354	08/28/1989	66
11- 92	--	391746	751510	417	09/01/1988	82
11- 96	--	391829	751208	375	09/22/1986	4
11-364	34-02333	391617	751355	420	09/05/1986	174
29- 04	33-00364	394524	740632	646	08/22/1989	2.6
29-541	53-00022	395451	740455	525	07/30/1985	2.5
			Lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8)			
1- 42	56-00010	392456	742121	788	08/16/1988	2.4
1-710	--	391726	742221	1,020	09/13/1985	77
1-711	--	391955	742507	871	08/14/1985	15
5-451	--	394536	743542	170	09/11/1987	8.3
9- 67	37-00271	390135	745352	592	09/01/1989	75
9-153	--	385932	744851	931	03/16/1972	114
9-161	--	390704	744750	654	03/20/1985	40

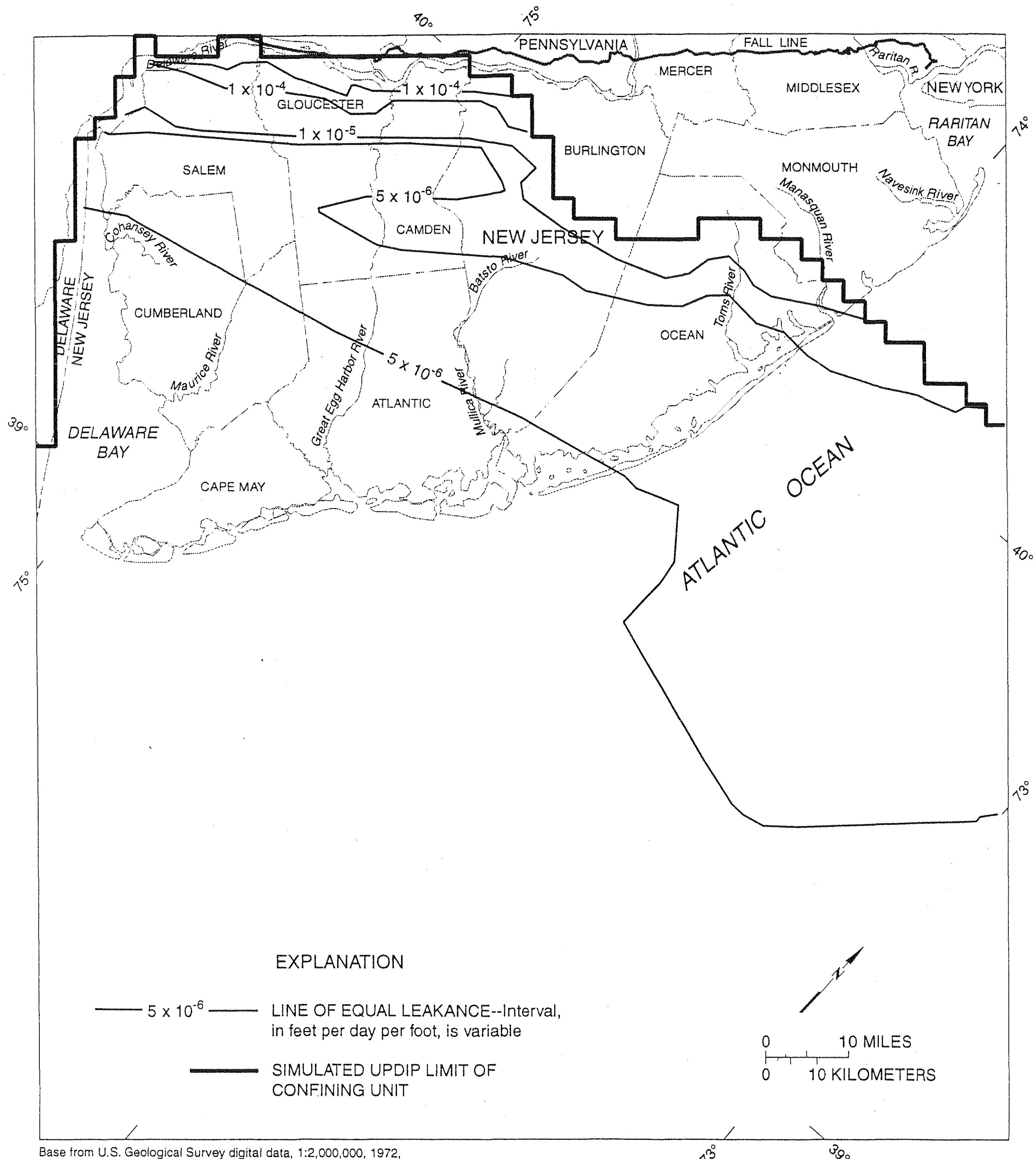
Table 7. Concentrations of chloride in selected samples from wells in the New Jersey Coastal Plain, 1971-90--Continued

U.S. Geological Survey well number	New Jersey permit number	Latitude	Longitude	Well depth	Sample date	Chloride concentration (mg/L)
Lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8)--Continued						
9-173	37-00579	390314	744532	862	08/27/1987	20
9-302	37-03628-9	385709	745128	903	09/06/1990	570
11- 52	35-01299	391420	751023	303	08/31/1988	5.5
11- 63	34-00846	391959	752152	72	06/08/1978	3,600
11-100	34-00460	391840	751336	74	08/11/1987	42
11-309	--	392247	751313	130	12/13/1988	2.6
11-691	35-09802-3	391116	745705	160	09/04/1990	15
15-730	--	393154	745811	98	07/30/1987	33
15-785	31-20754	393917	750149	56	09/19/1988	9.9
25- 29	49-00013	400644	740344	150	09/01/1987	2.5
25-512	29-10880	400802	740230	124	09/01/1987	9.8
29-508	33-00896	395528	740826	153	08/30/1984	6.3
29-544	33-00219	393839	741052	578	08/13/1987	2.9
29-807	29-12178	400536	740251	132	08/17/1989	410
29-815	33-18281	395643	740443	154	08/08/1989	150

APPENDIXES

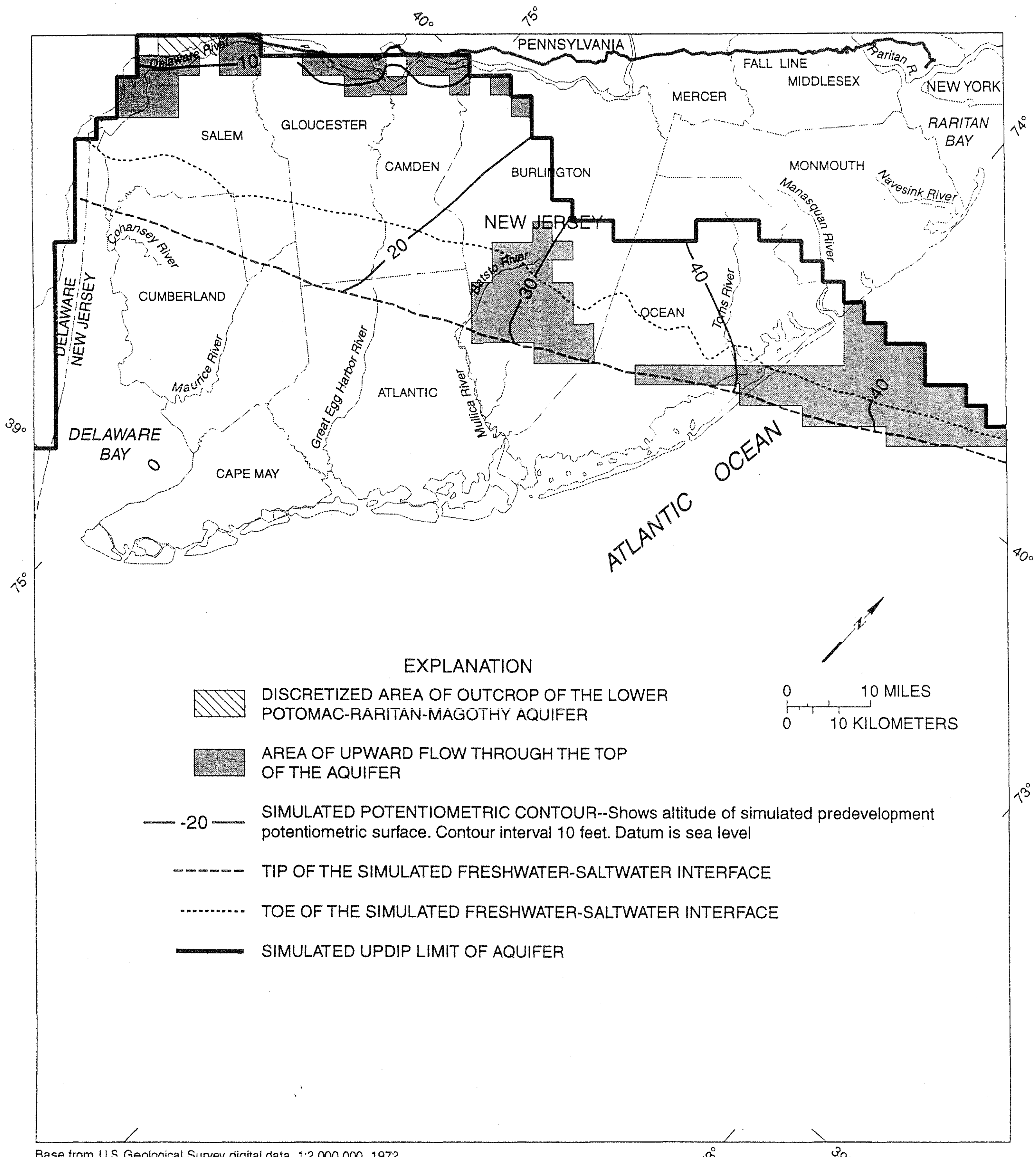


1a. Transmissivity of the Lower Potomac-Raritan-Magothy aquifer (model unit A1) used in the ground-water flow model of the New Jersey Coastal Plain.



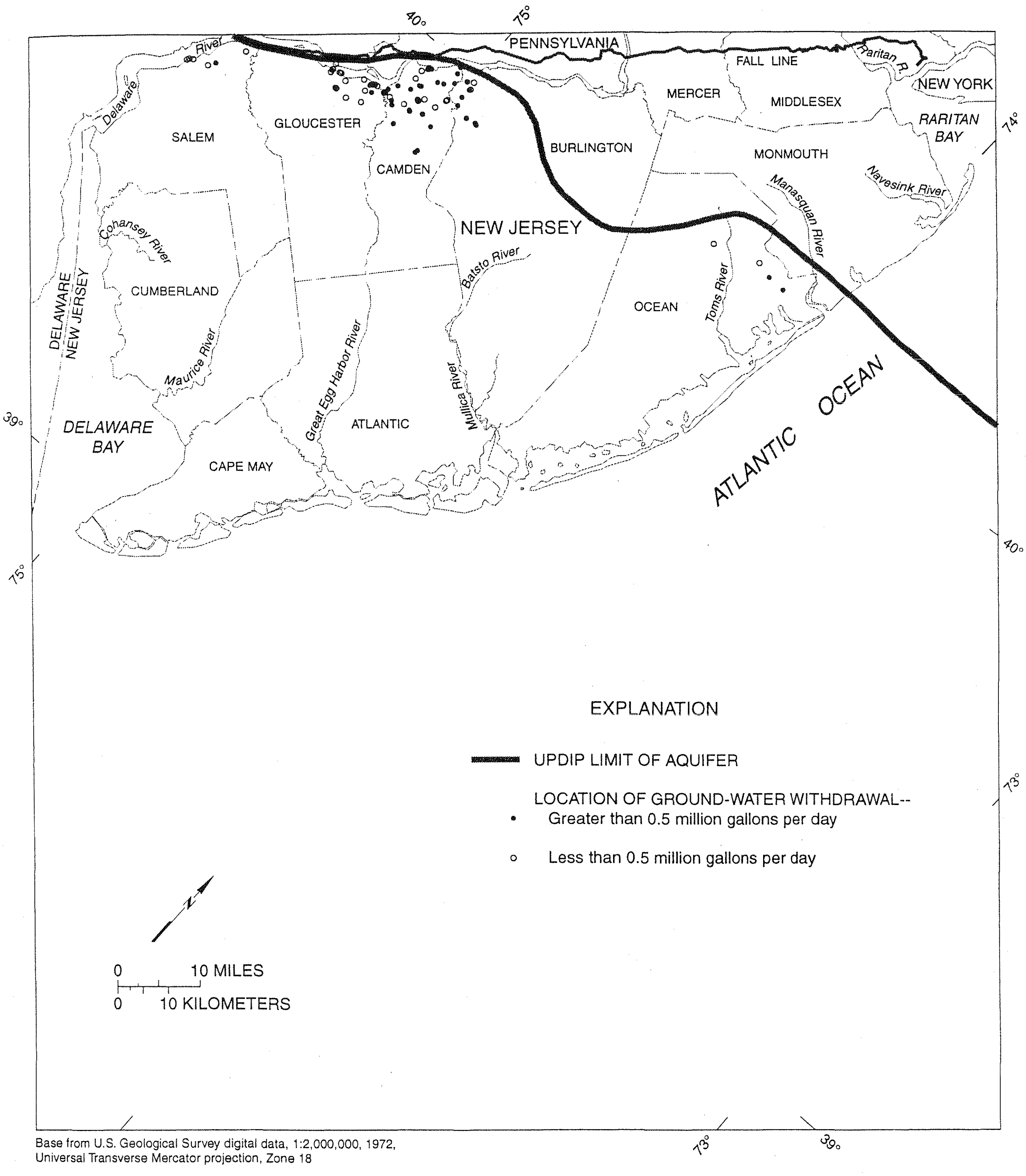
Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

1b. Vertical leakance of the confining unit between the Lower and Middle Potomac-Raritan-Magothy aquifers (model unit C1) used in the ground-water flow model of the New Jersey Coastal Plain.

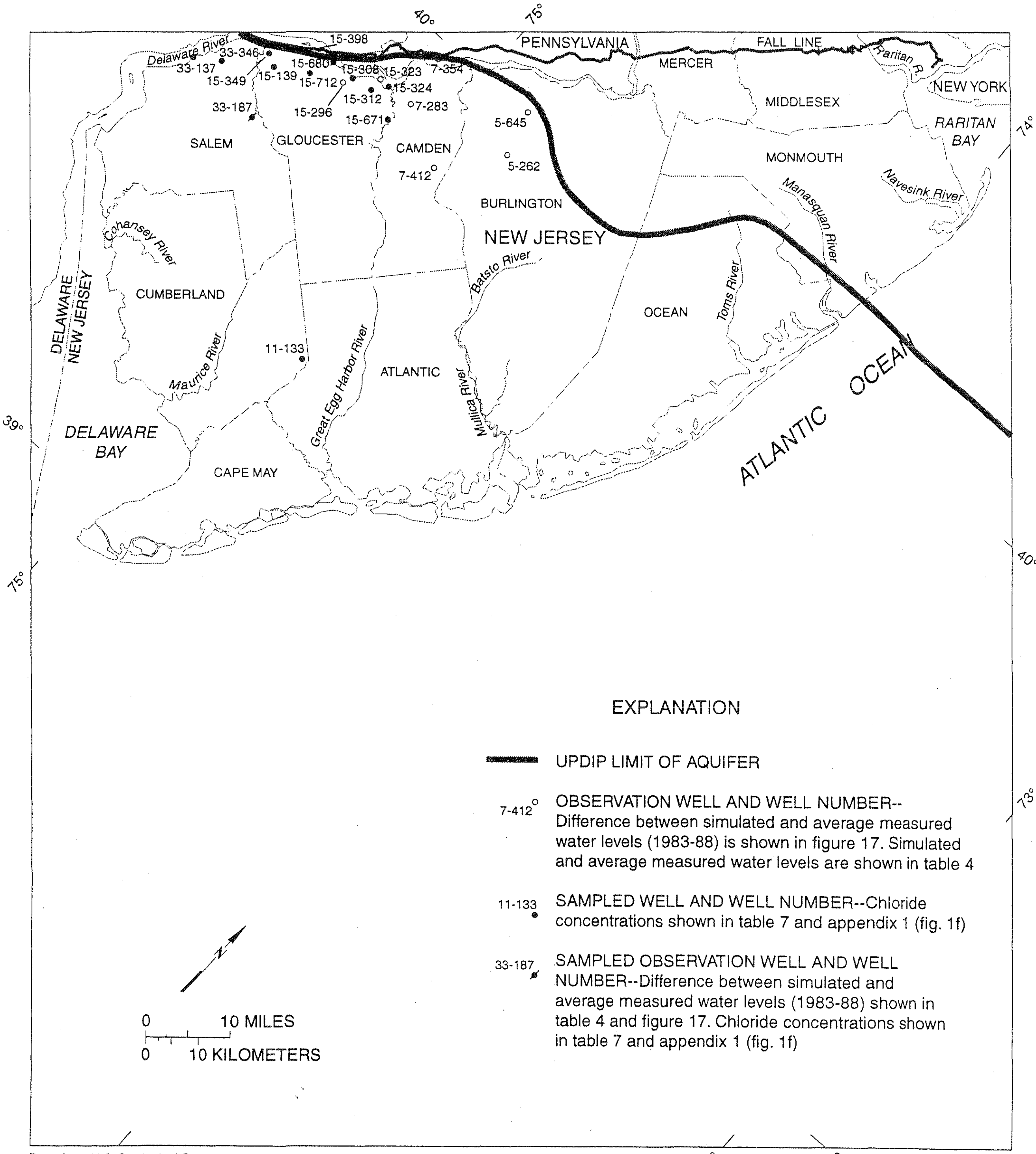


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

1c. Simulated predevelopment potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow, Lower Potomac-Raritan-Magothy aquifer (model unit A1), New Jersey Coastal Plain.

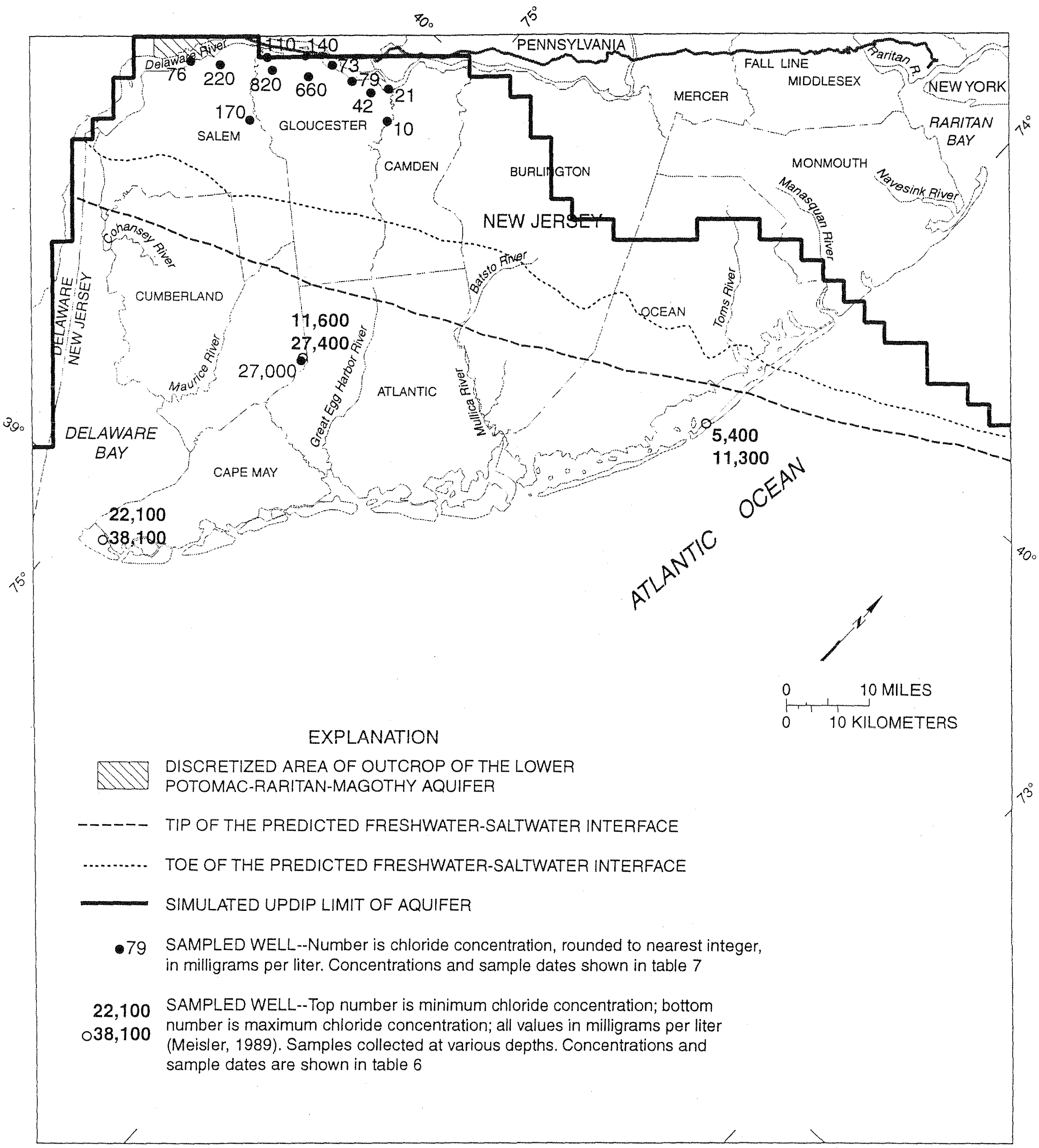


1d. Locations of ground-water withdrawal sites in the Lower Potomac-Raritan-Magothy aquifer (model unit A1), New Jersey Coastal Plain, 1983-88.



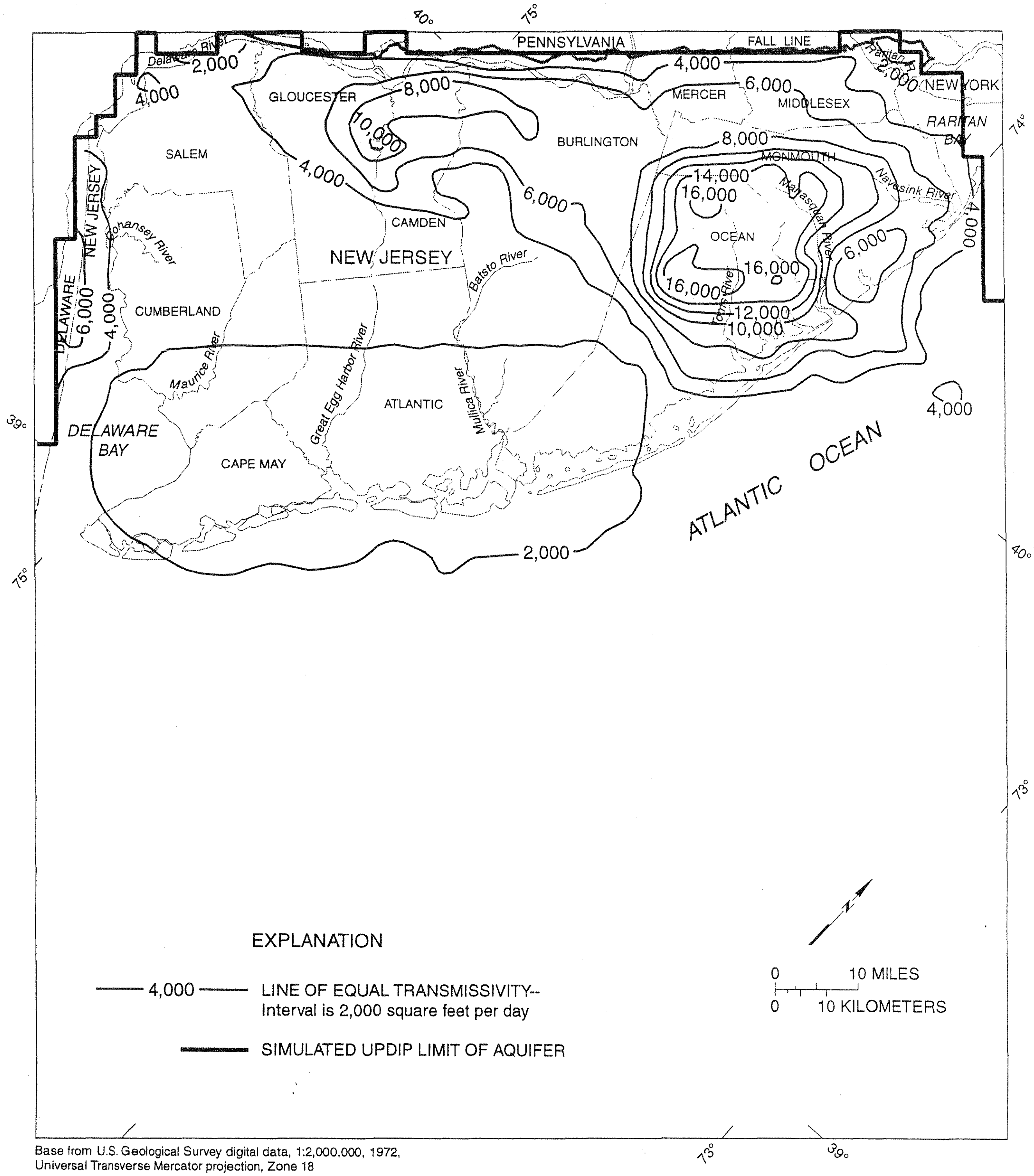
Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

1e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Lower Potomac-Raritan-Magothy aquifer (model unit A1), New Jersey Coastal Plain.

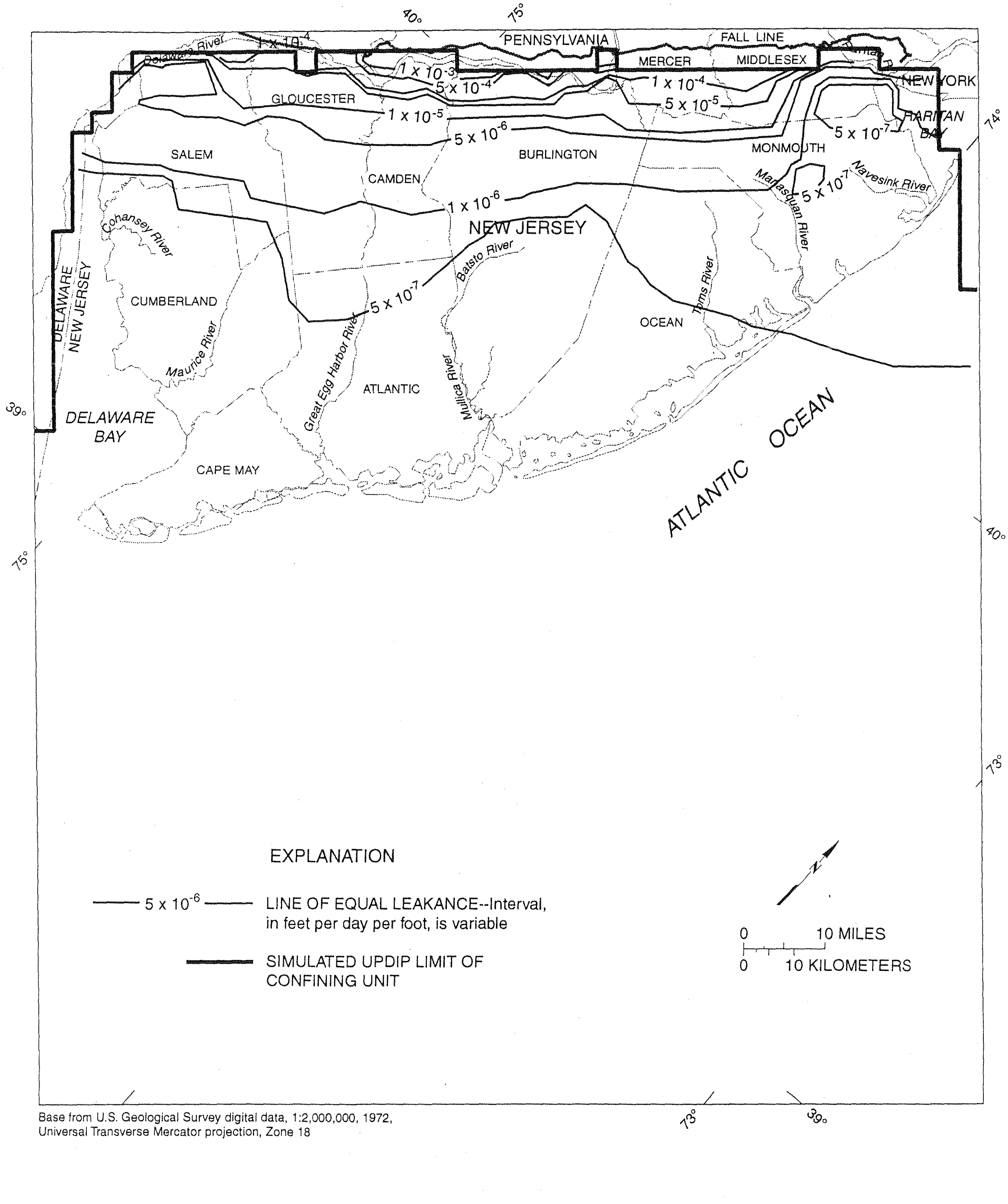


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

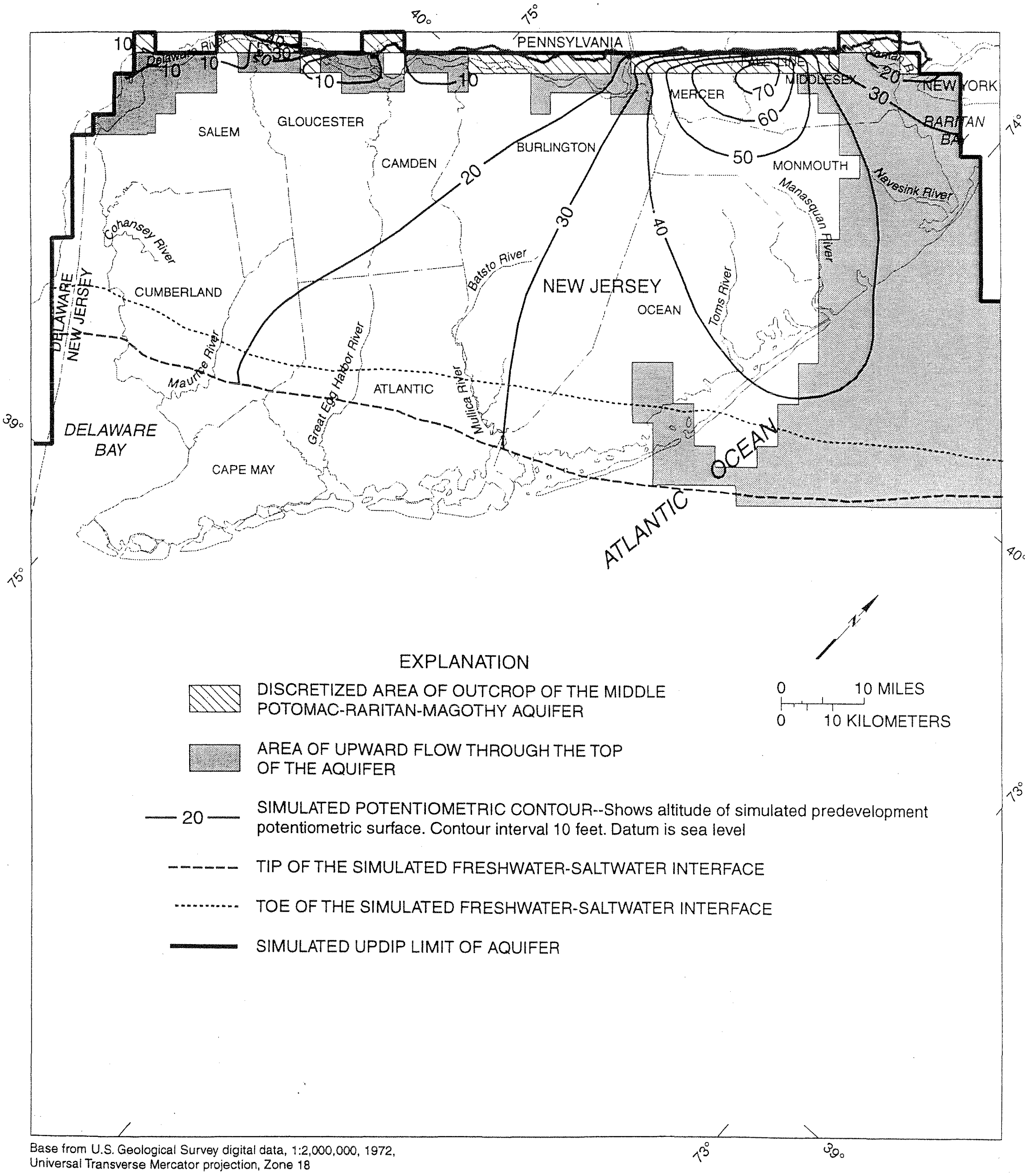
1f. Measured chloride concentrations and location of simulated freshwater-saltwater interface tip and toe, Lower Potomac-Raritan-Magothy aquifer (model unit A1), New Jersey Coastal Plain, 1983-88.



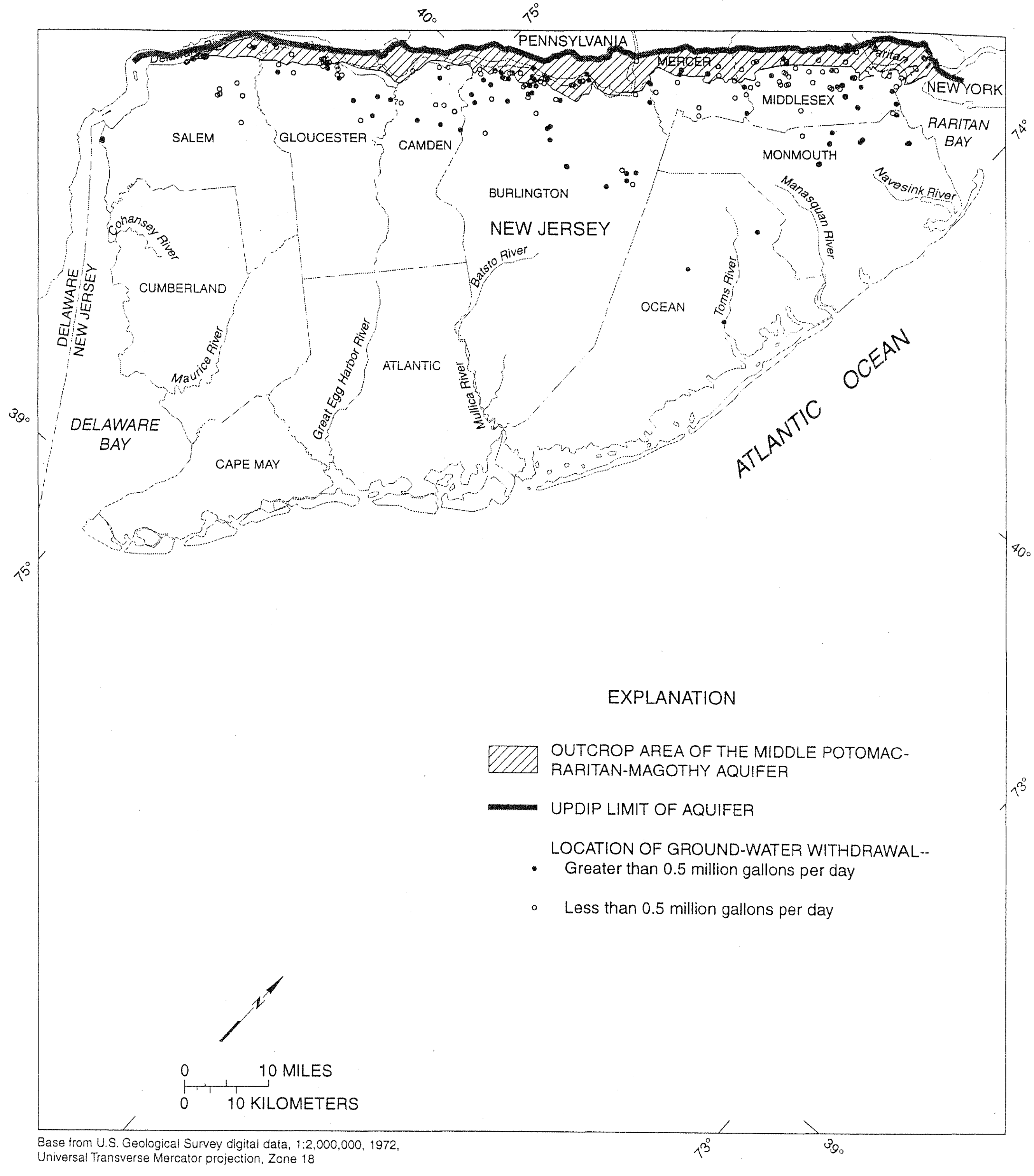
2a. Transmissivity of the Middle Potomac-Raritan-Magothy aquifer (model unit A2) used in the ground-water flow model of the New Jersey Coastal Plain.



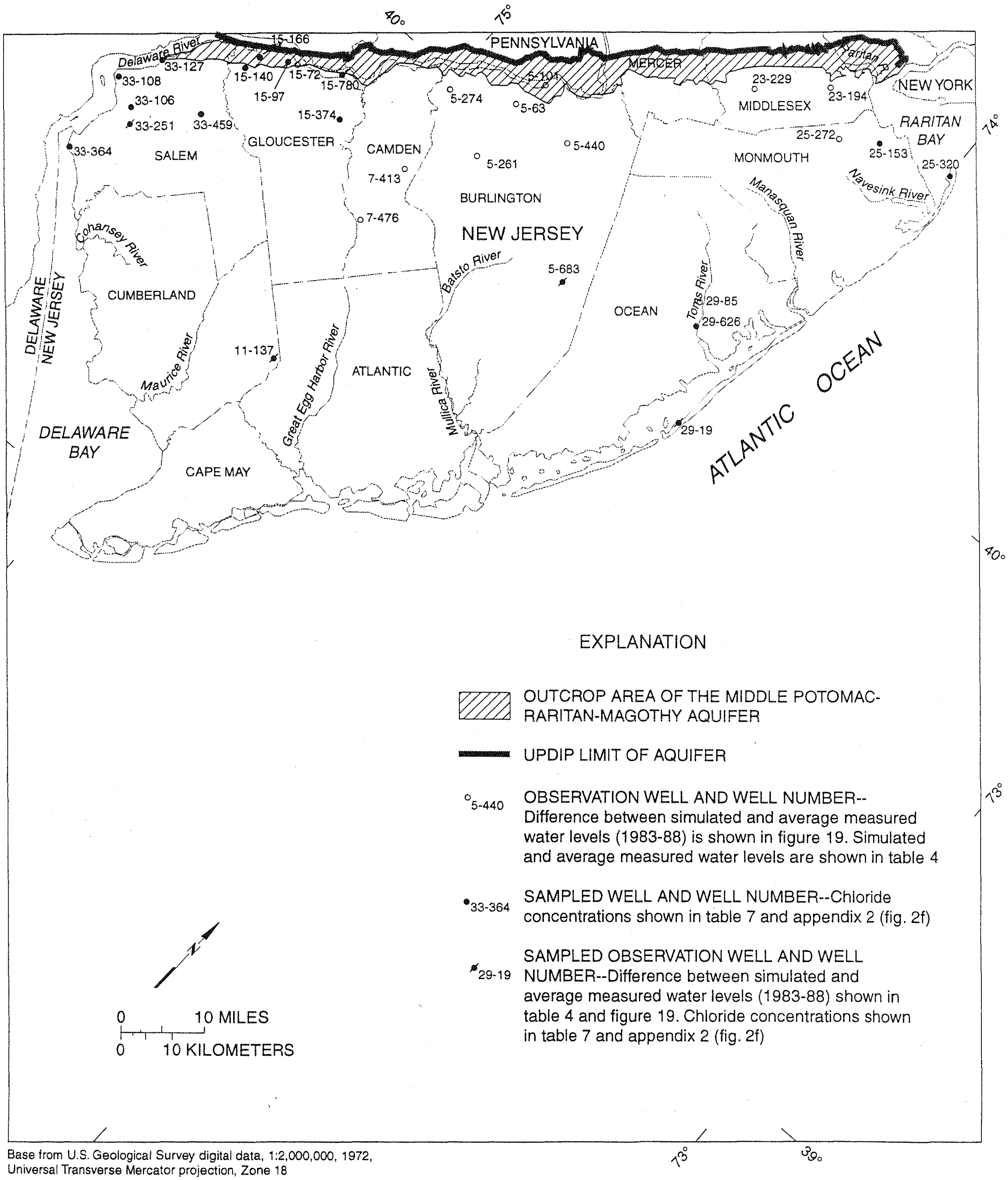
2b. Vertical leakance of the confining unit between the Middle and Upper Potomac-Raritan-Magothy aquifers (model unit C2) used in the ground-water flow model of the New Jersey Coastal Plain.



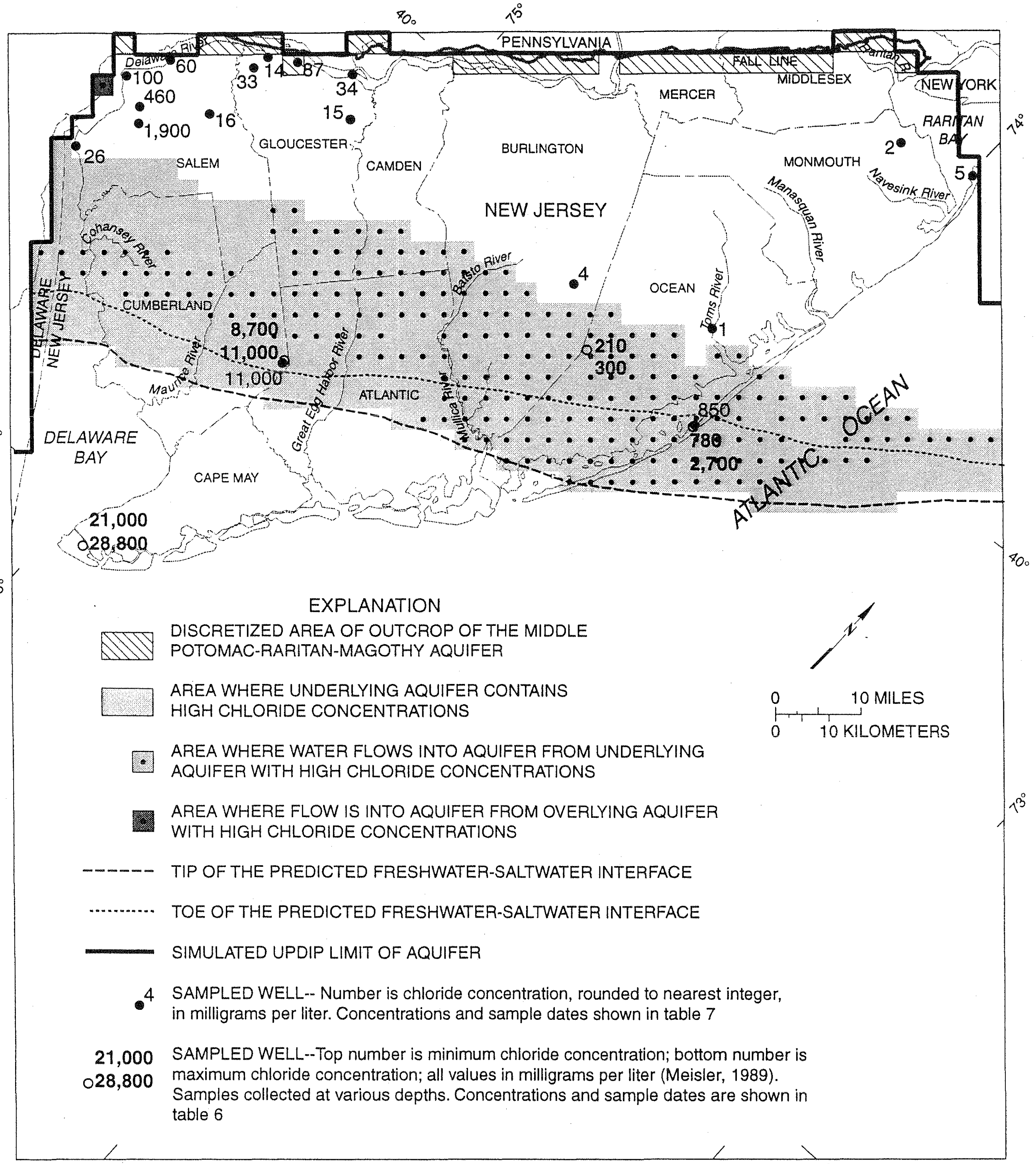
2c. Simulated predevelopment potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow, Middle Potomac-Raritan-Magothy aquifer (model unit A2), New Jersey Coastal Plain.



2d. Locations of ground-water withdrawal sites in the Middle Potomac-Raritan-Magothy aquifer (model unit A2), New Jersey Coastal Plain, 1983-88.

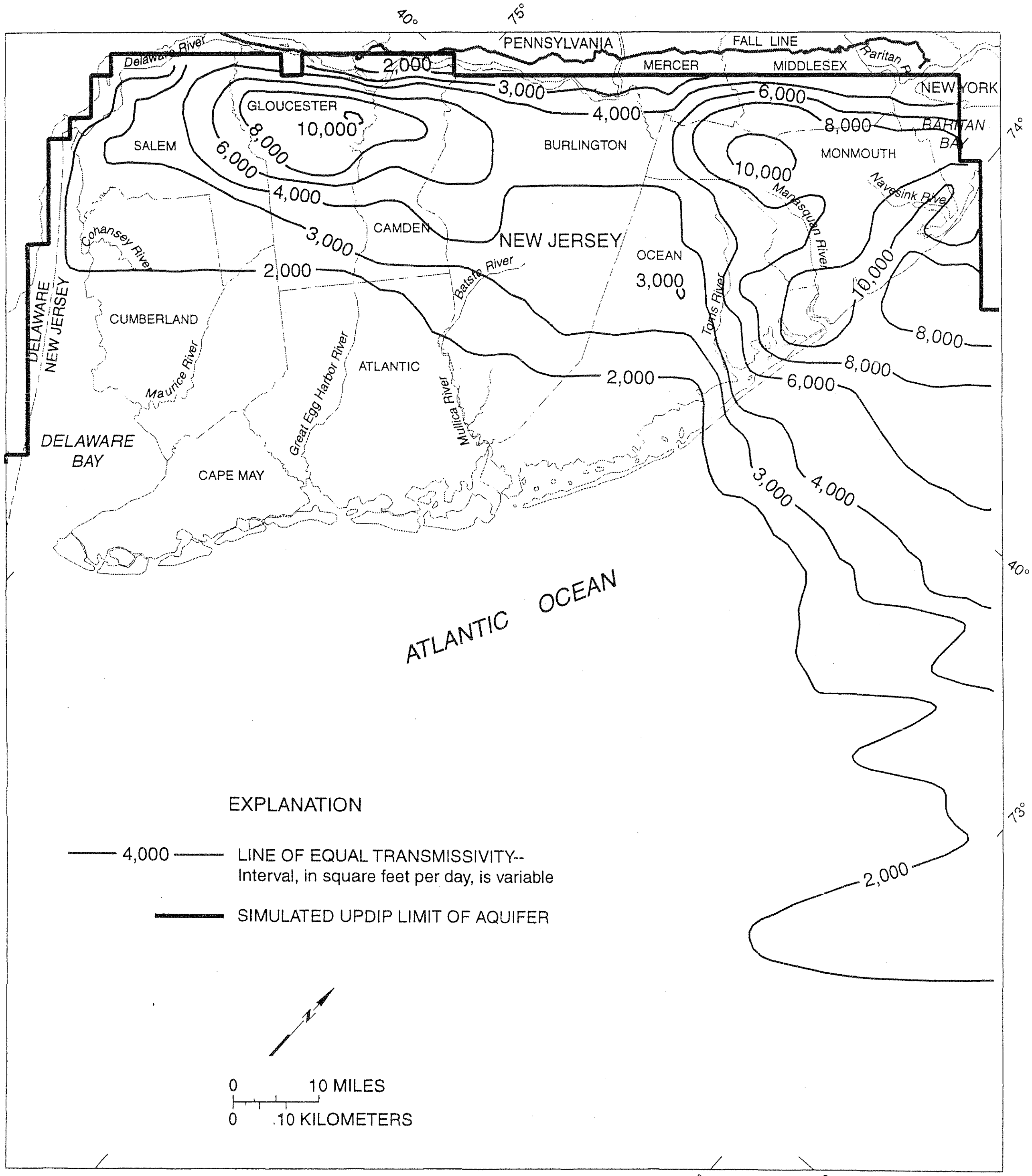


2e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Middle Potomac-Raritan-Magothy aquifer (model unit A2), New Jersey Coastal Plain.



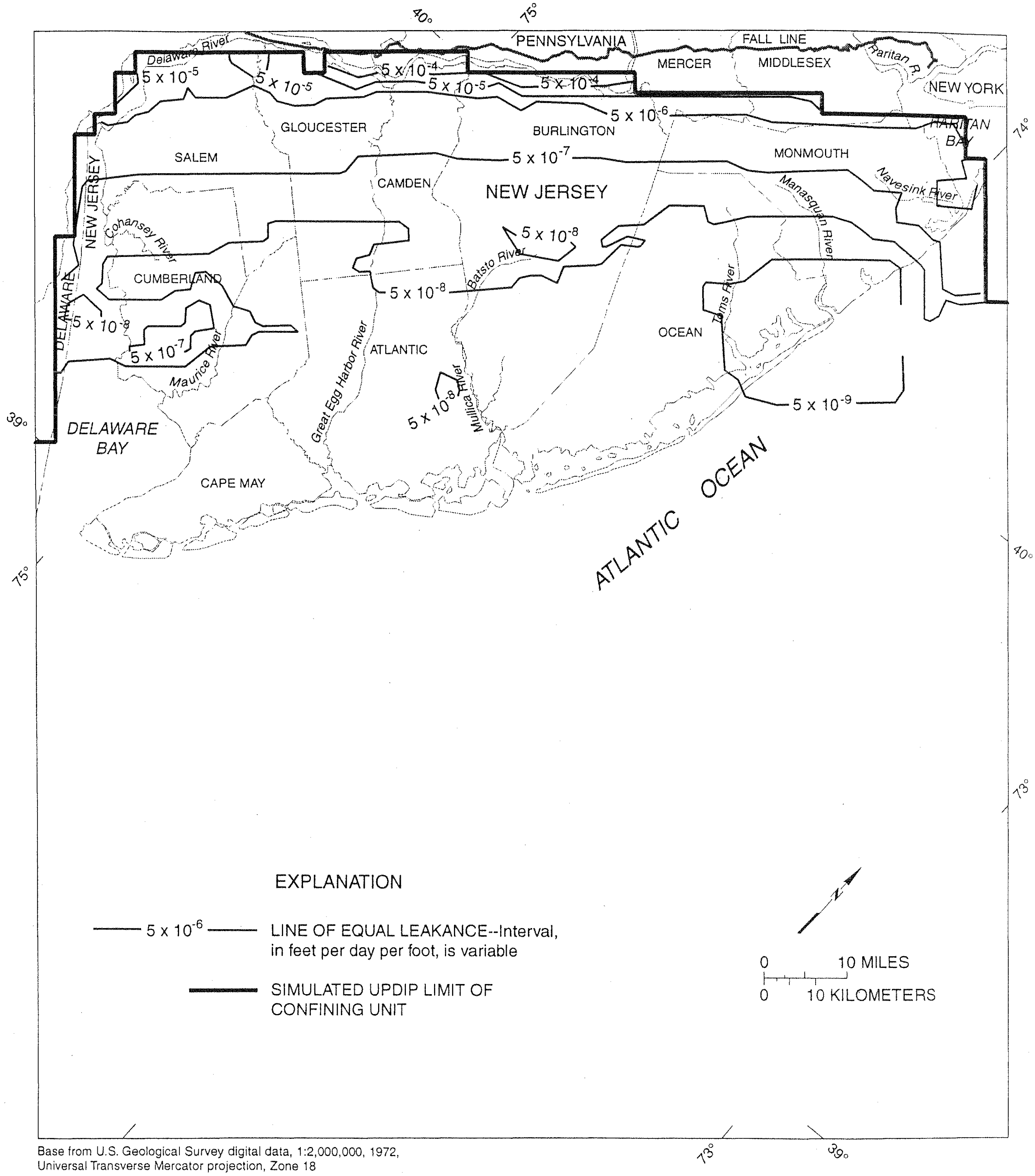
Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

2f. Measured chloride concentrations, location of simulated freshwater-saltwater interface tip and toe, and areas of saltwater flow from overlying and underlying aquifers, Middle Potomac-Raritan-Magothy aquifer (model unit A2), New Jersey Coastal Plain, 1983-88.

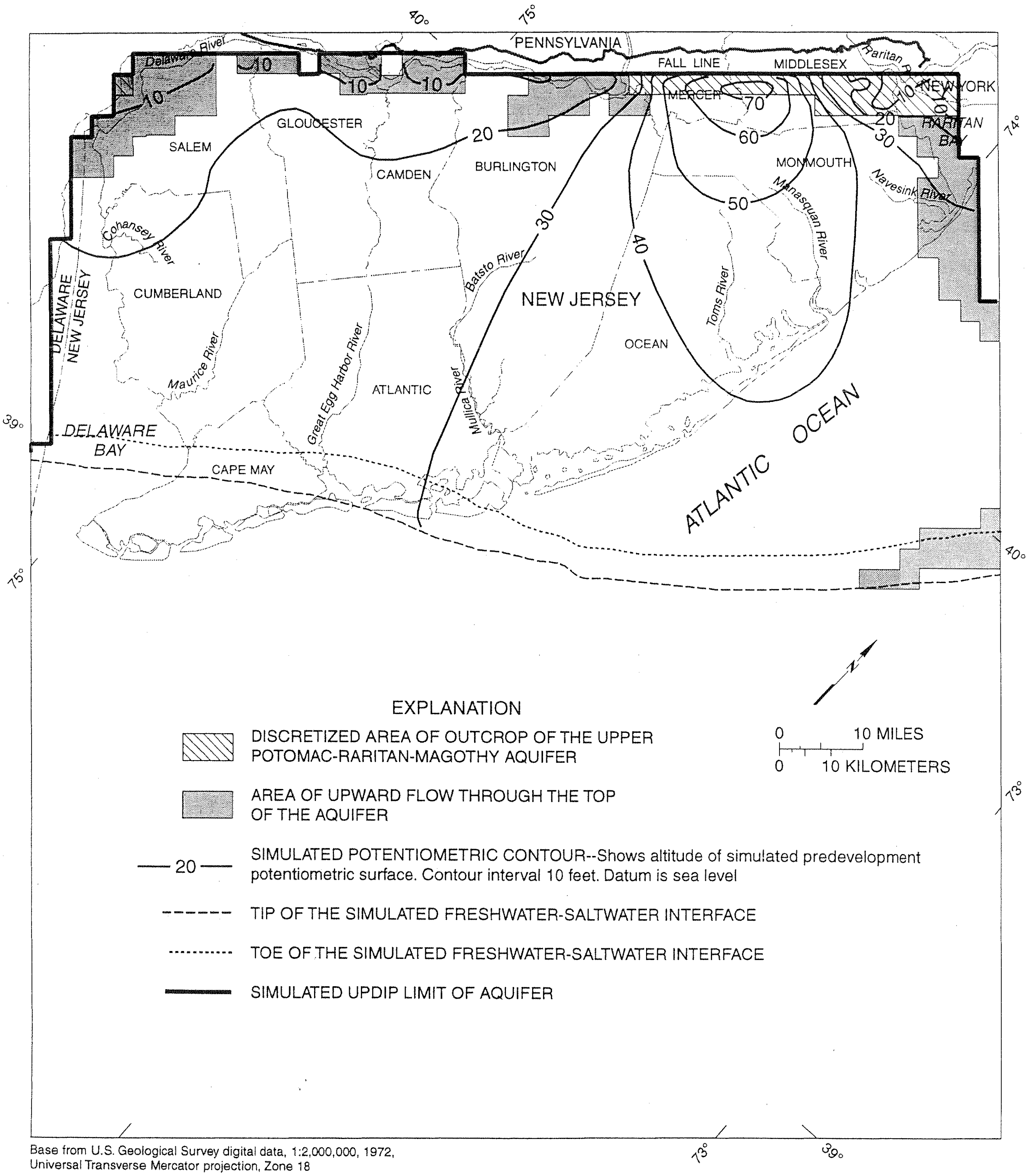


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

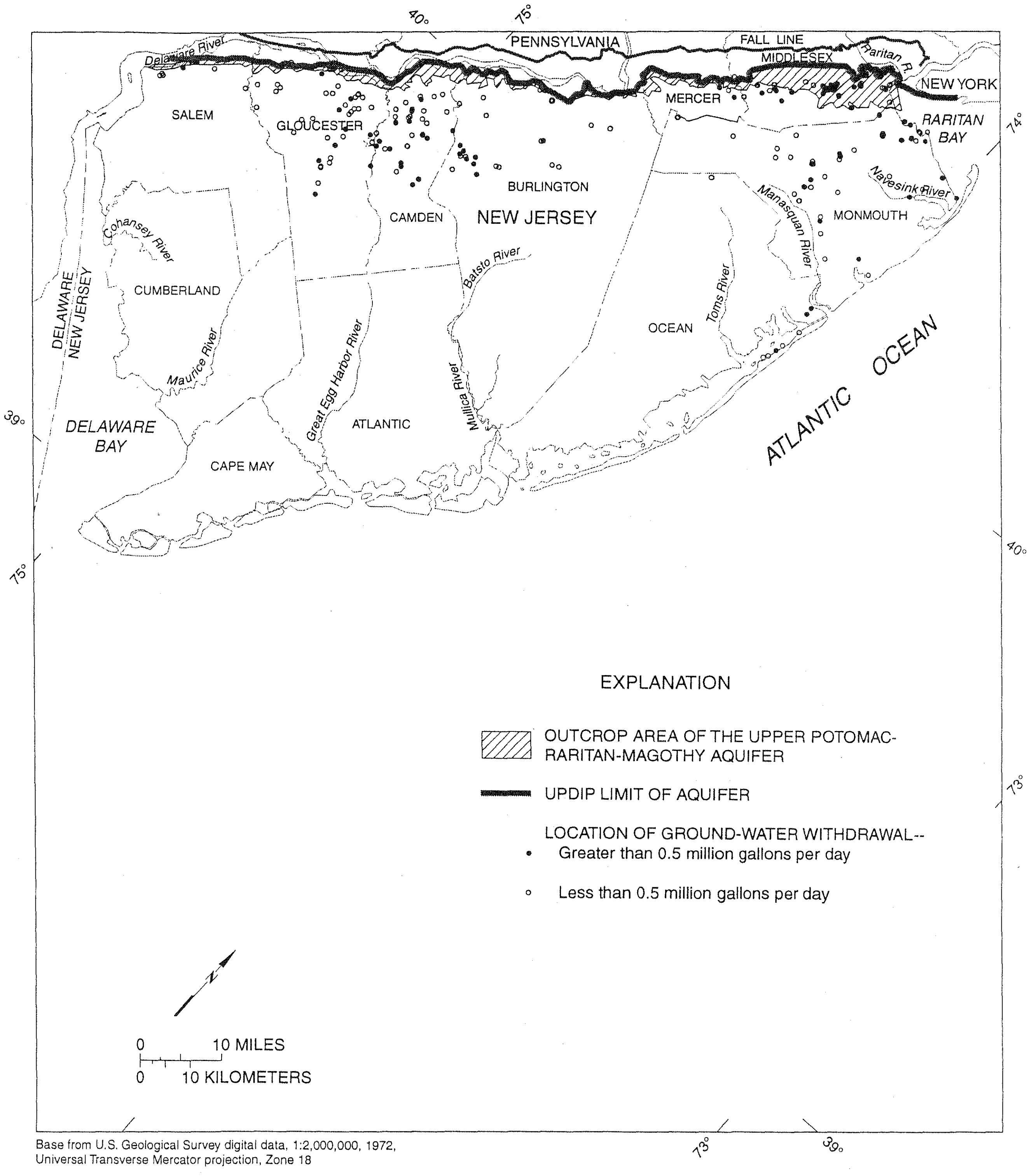
3a. Transmissivity of the Upper Potomac-Raritan-Magothy aquifer (model unit A3) used in the ground-water flow model of the New Jersey Coastal Plain.



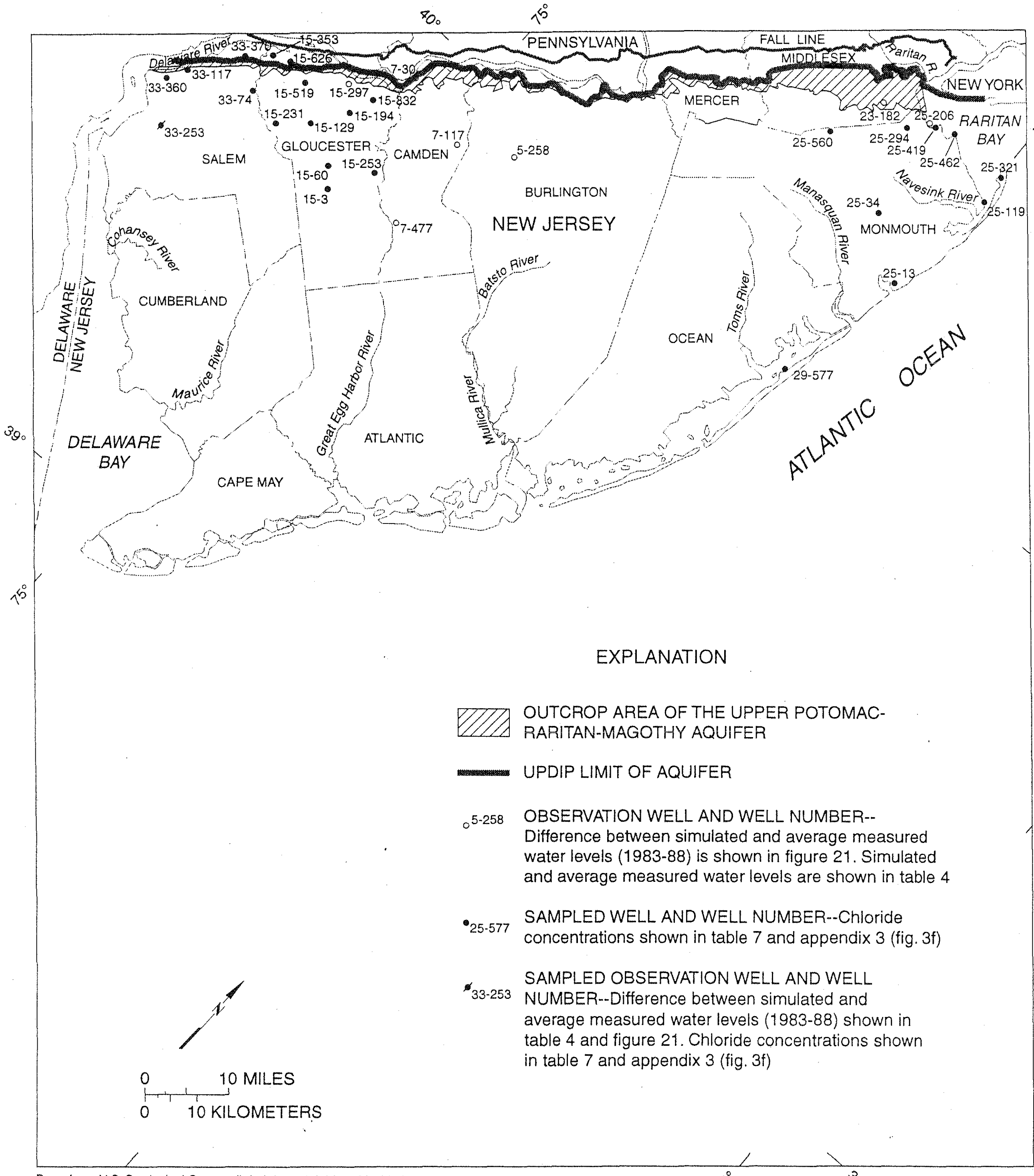
3b. Vertical leakance of the Merchantville-Woodbury confining unit (model unit C3) used in the ground-water flow model of the New Jersey Coastal Plain.



3c. Simulated predevelopment potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow, Upper Potomac-Raritan-Magothy aquifer (model unit A3), New Jersey Coastal Plain.

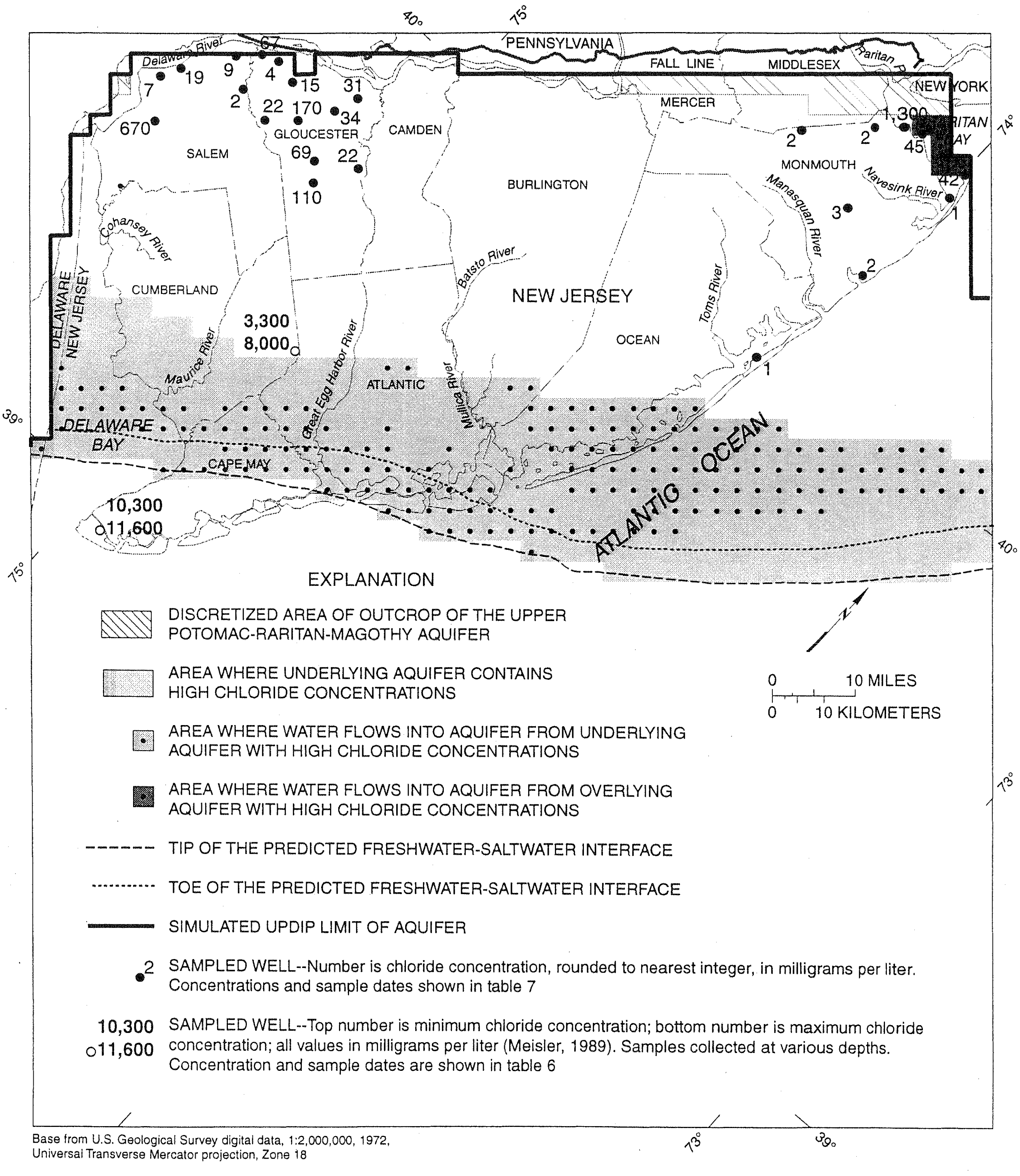


3d. Locations of ground-water withdrawal sites in the Upper Potomac-Raritan-Magothy aquifer (model unit A3), New Jersey Coastal Plain, 1983-88.

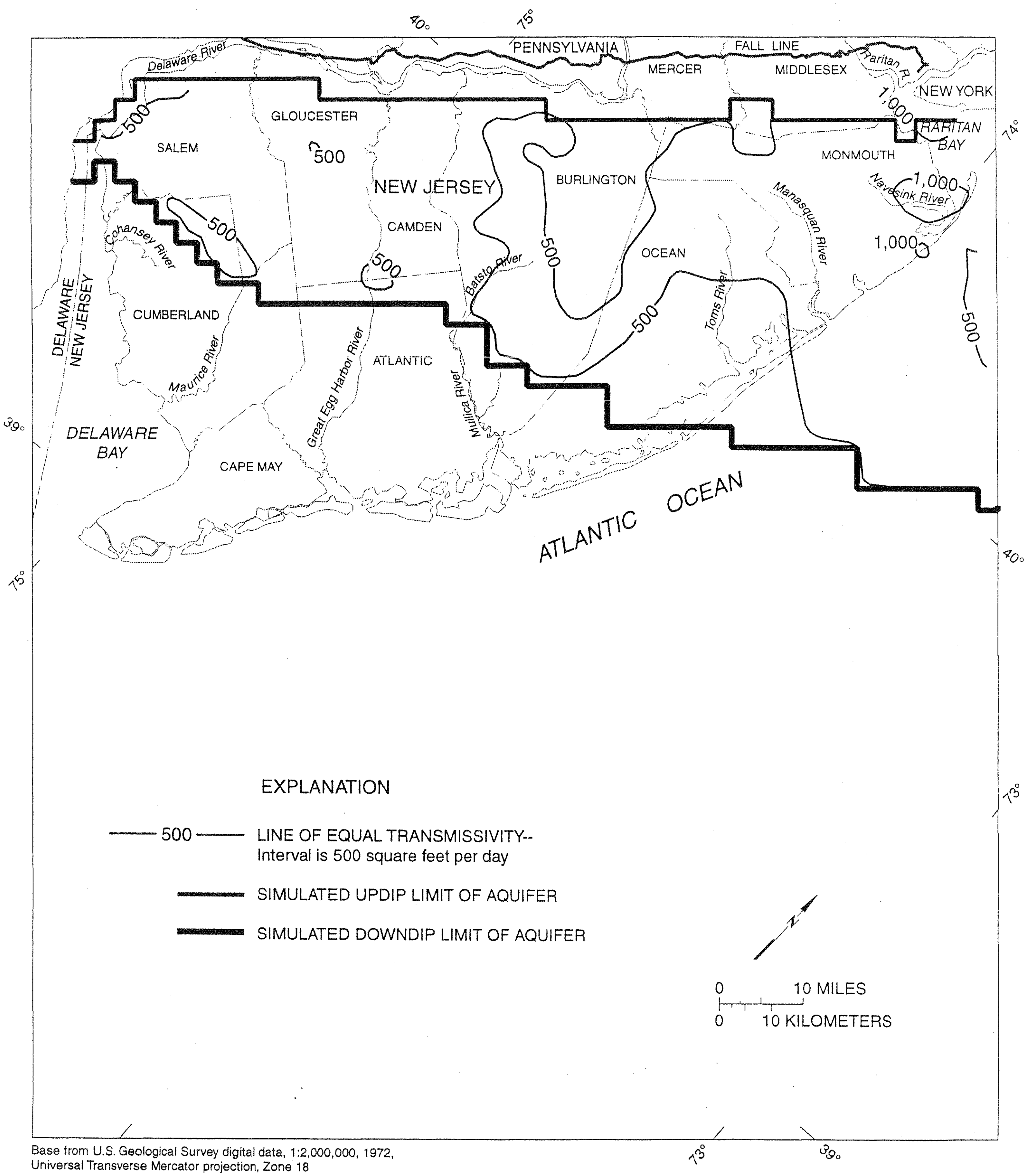


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

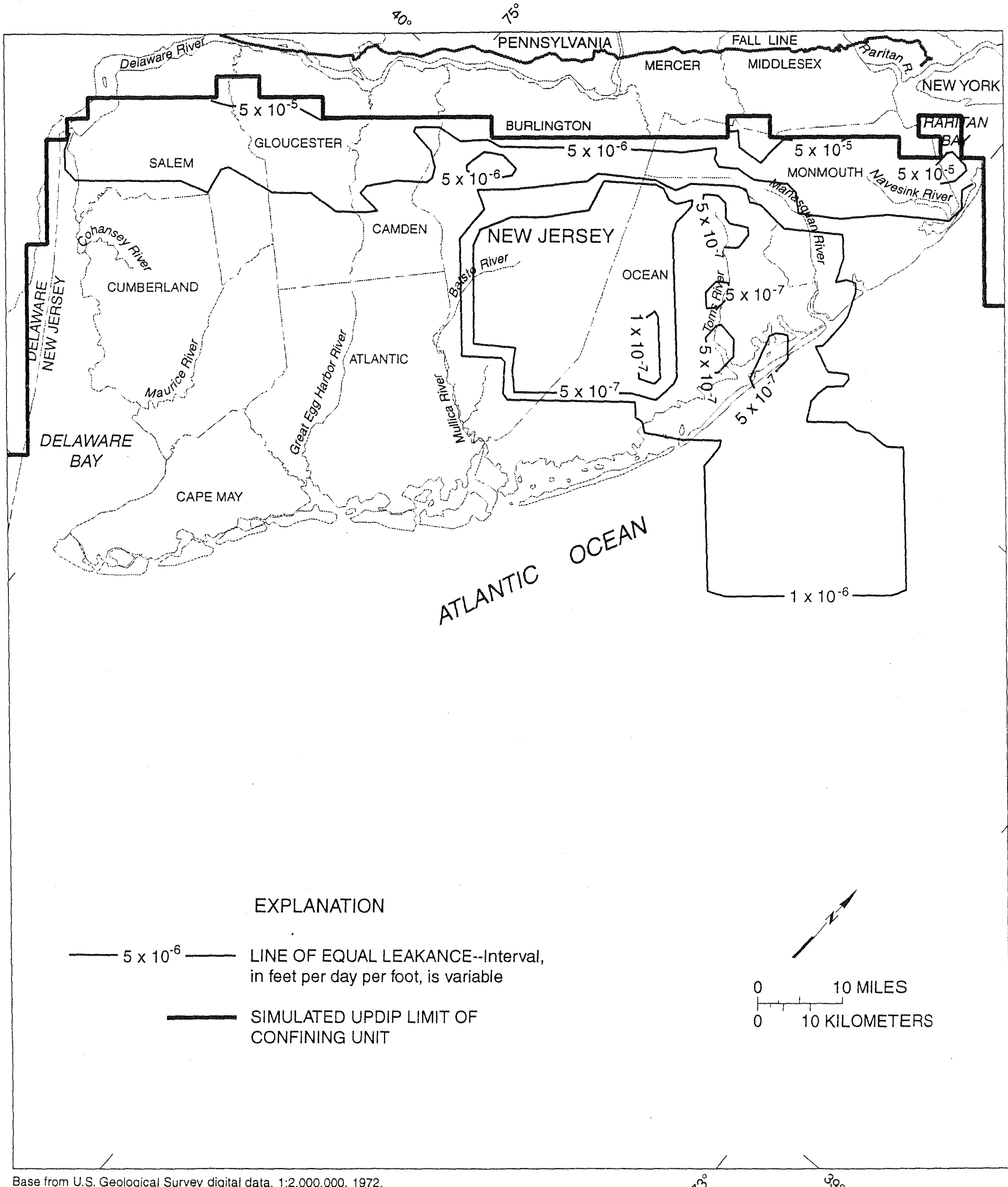
3e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Upper Potomac-Raritan-Magothy aquifer (model unit A3), New Jersey Coastal Plain.



3f. Measured chloride concentrations, location of simulated freshwater-saltwater interface tip and toe, and areas of saltwater flow from overlying and underlying aquifers, Upper Potomac-Raritan-Magothy aquifer (model unit A3), New Jersey Coastal Plain, 1983-88.

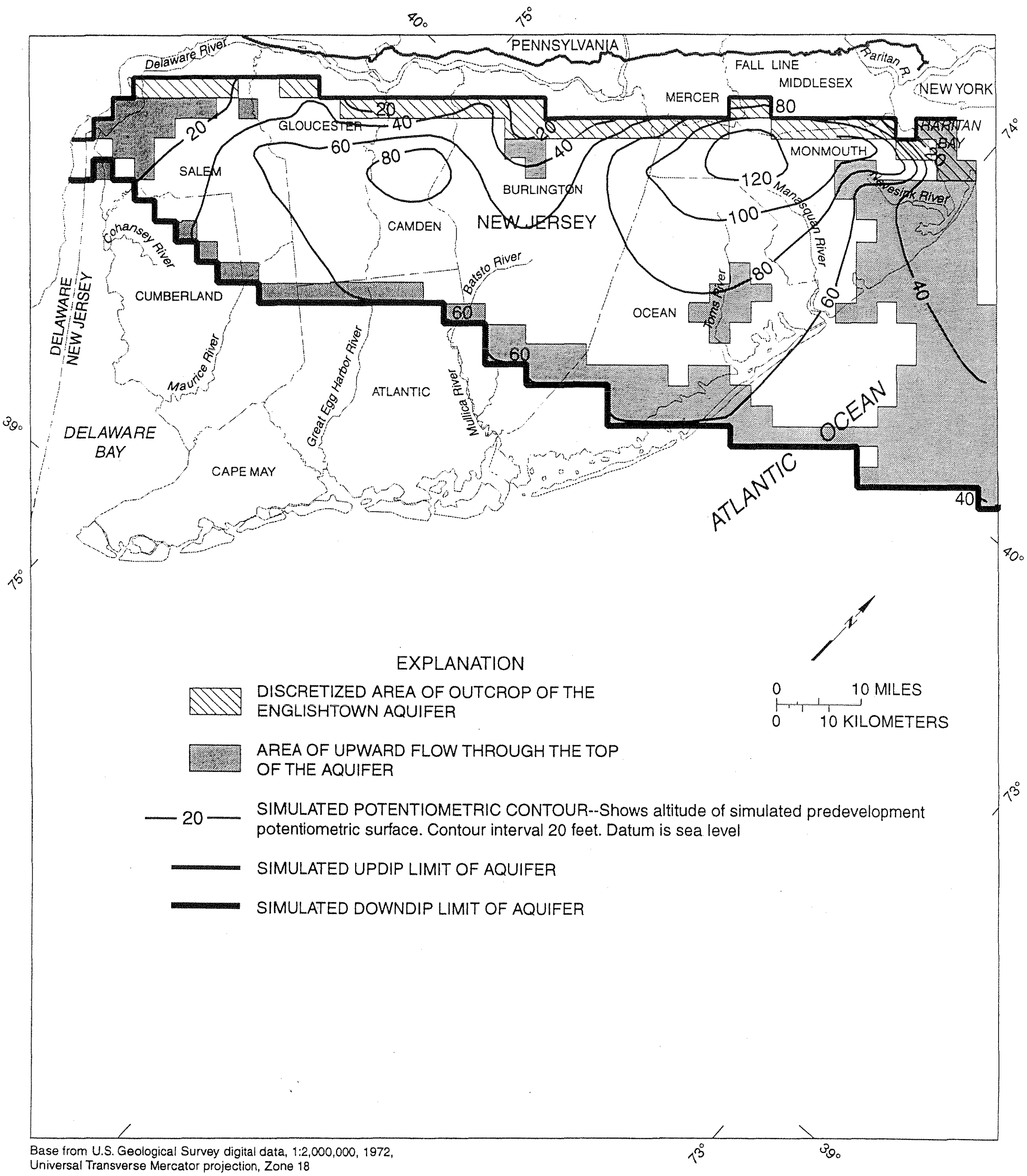


4a. Transmissivity of the Englishtown aquifer (model unit A4) used in the ground-water flow model of the New Jersey Coastal Plain.

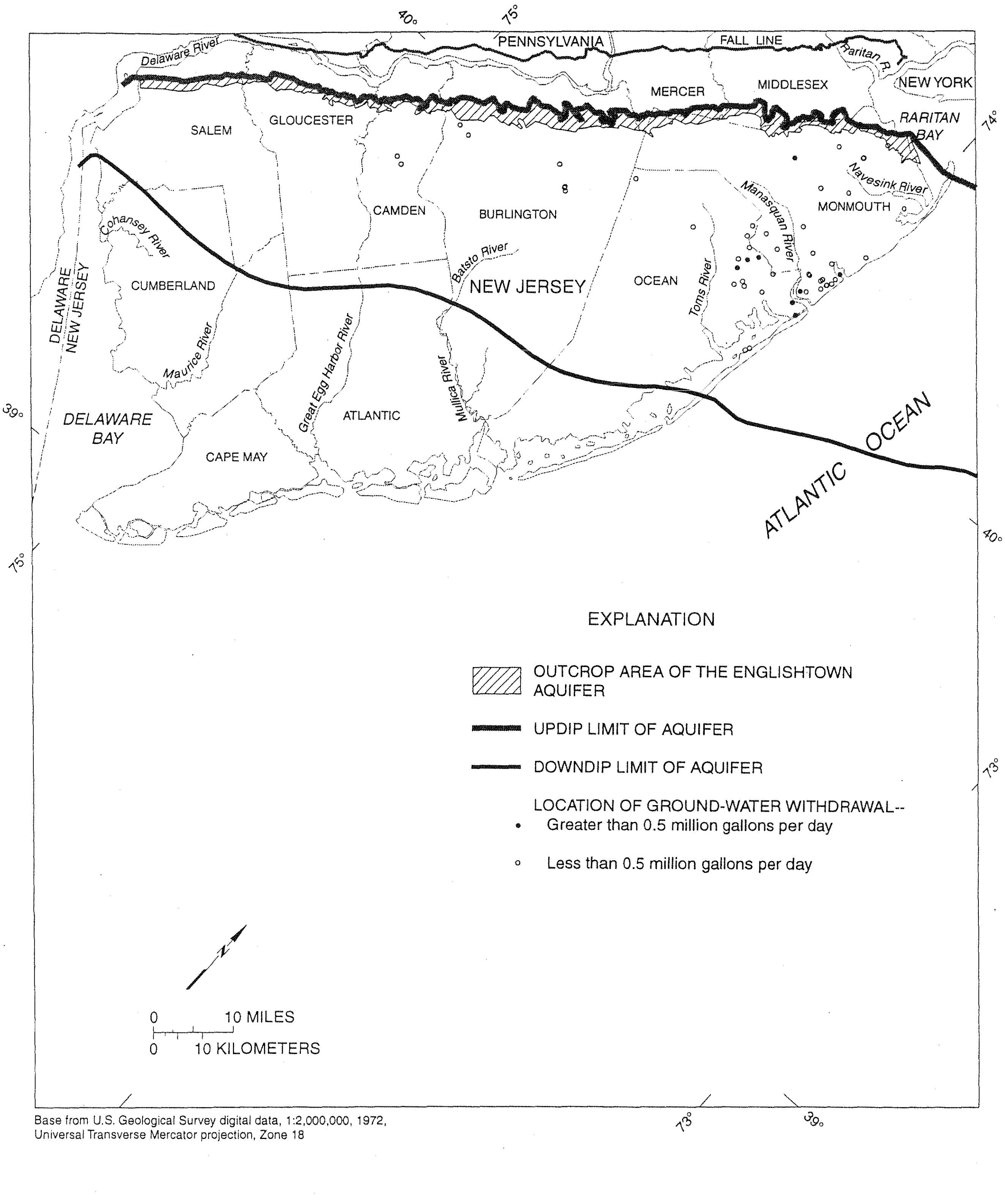


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

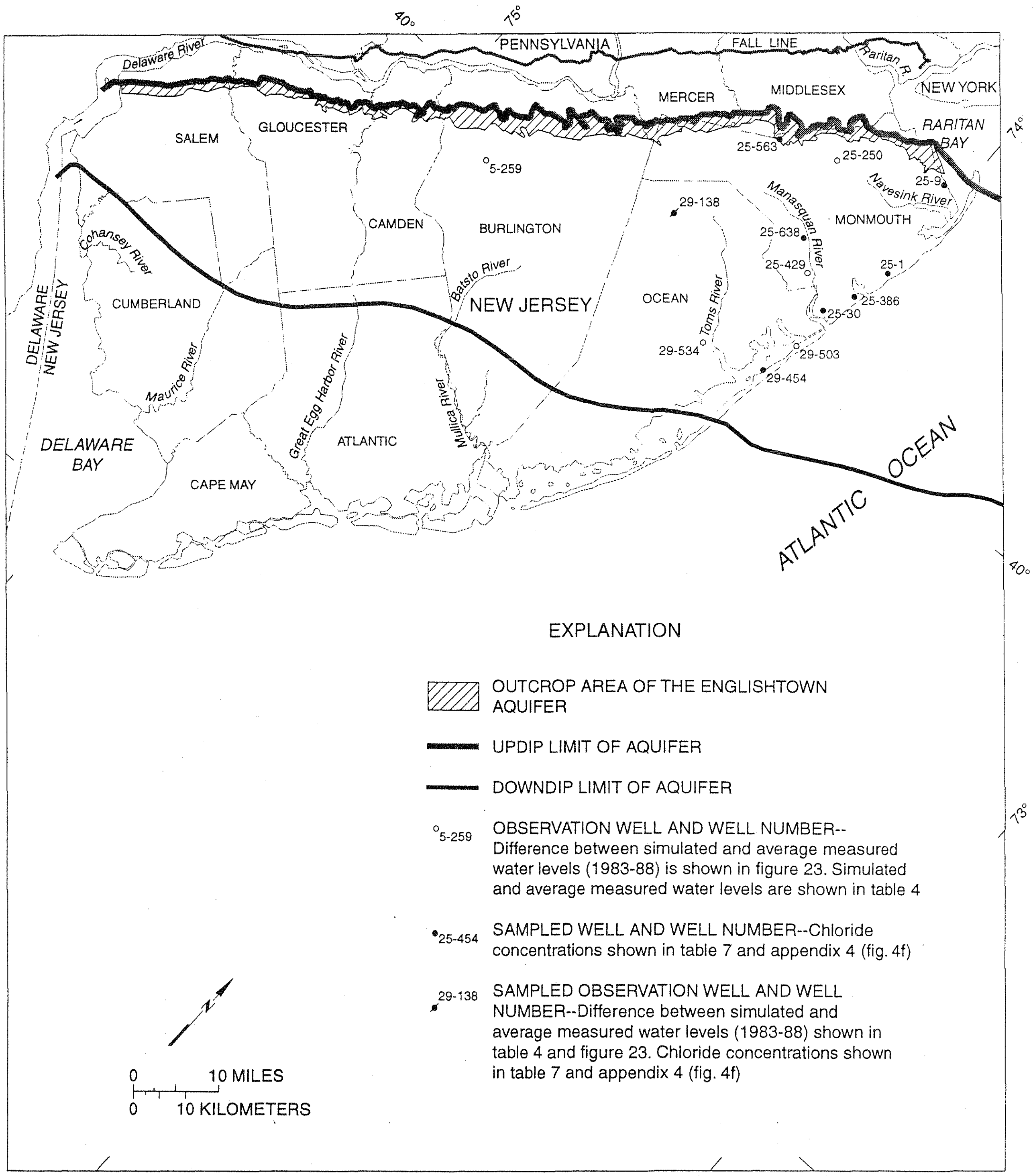
4b. Vertical leakance of the Marshalltown-Wenonah confining unit (model unit C4) used in the ground-water flow model of the New Jersey Coastal Plain.



4c. Simulated predevelopment potentiometric surface and area of upward flow, Englishtown aquifer (model unit A4), New Jersey Coastal Plain.

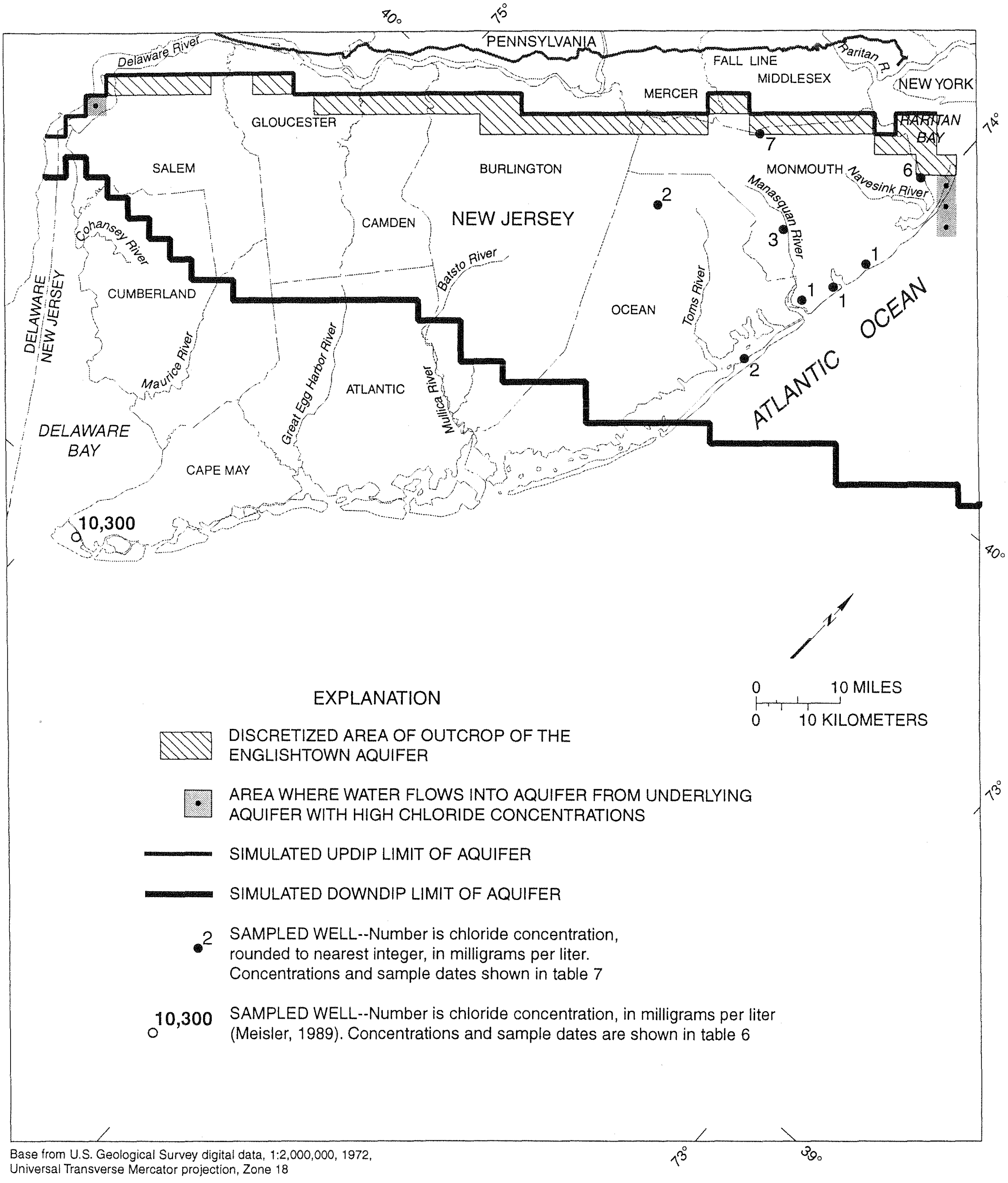


4d. Locations of ground-water withdrawal sites in the Englishtown aquifer (model unit A4), New Jersey Coastal Plain, 1983-88.

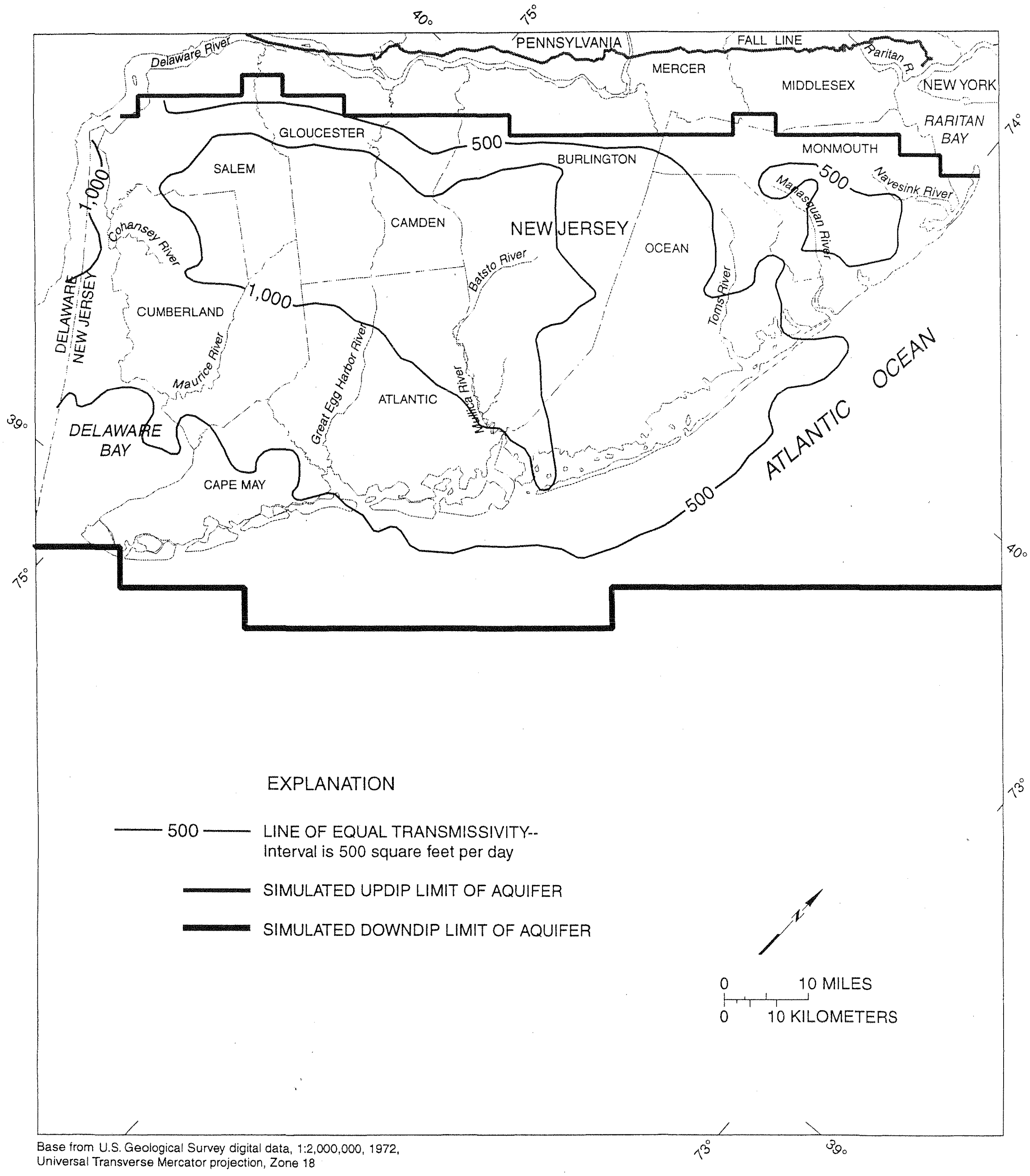


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

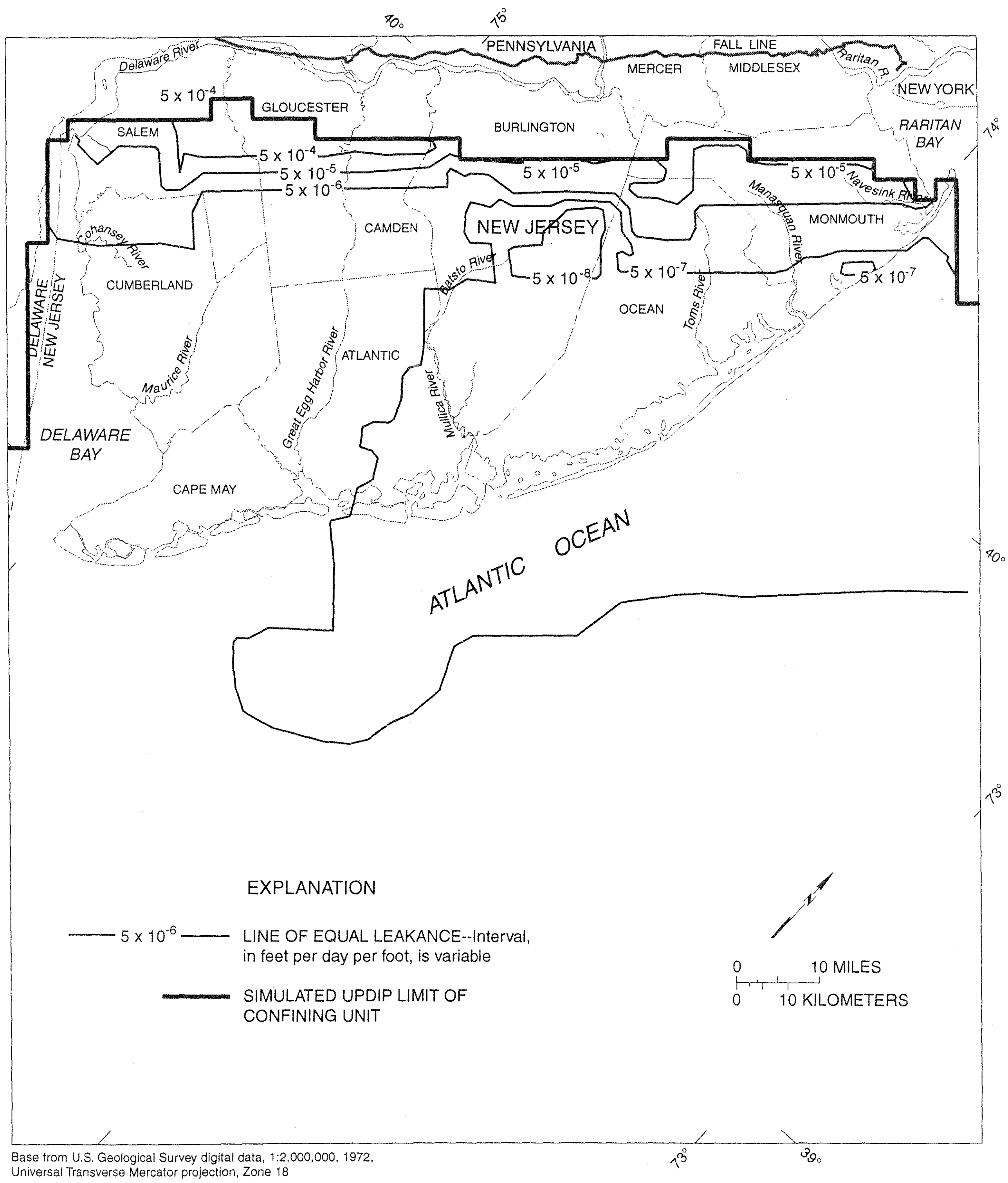
4e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Englishtown aquifer (model unit A4), New Jersey Coastal Plain.



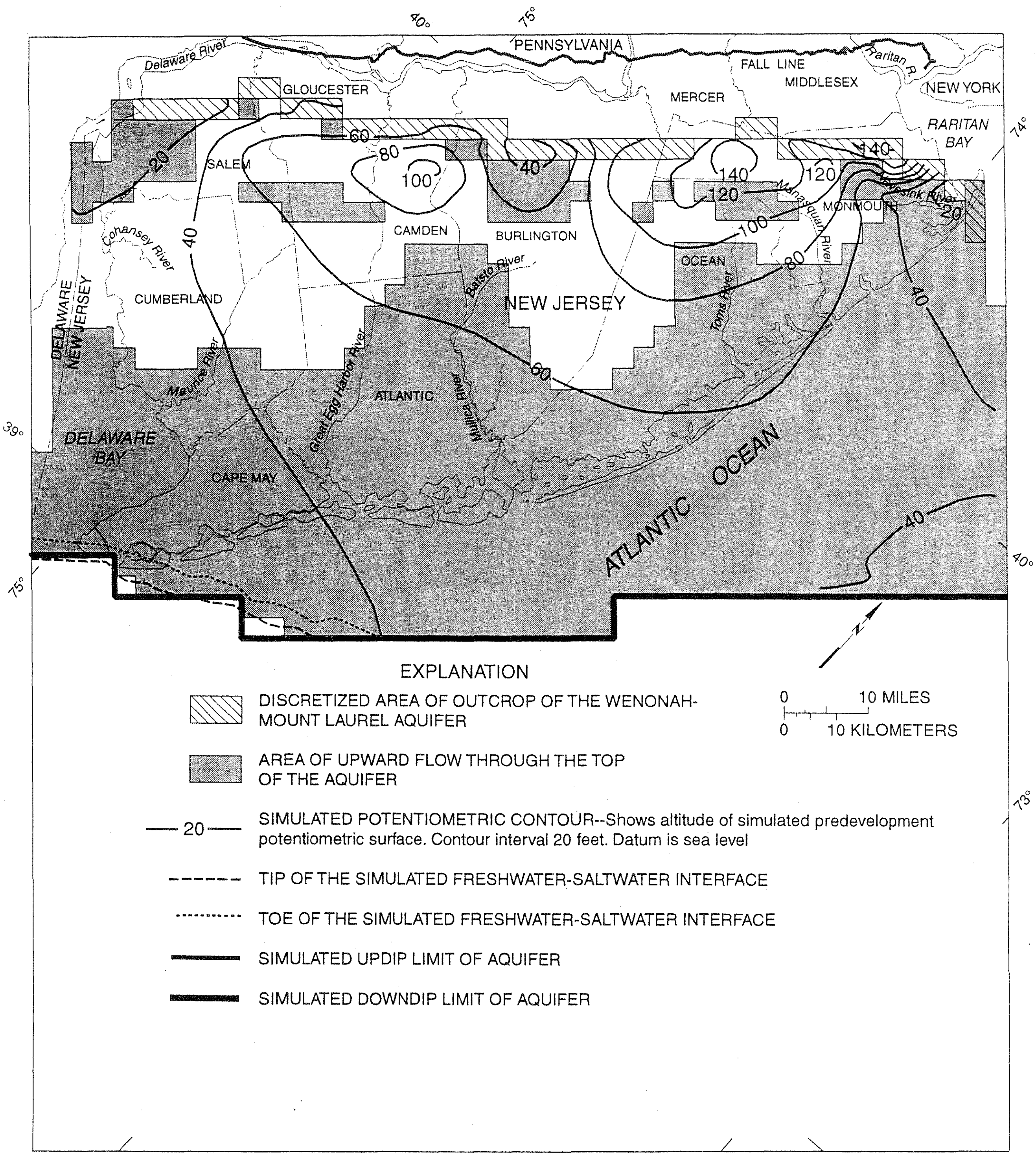
4f. Measured chloride concentrations and areas of saltwater flow from underlying aquifers, Englishtown aquifer (model unit A4), New Jersey Coastal Plain, 1983-88.



5a. Transmissivity of the Wenonah-Mount Laurel aquifer (model unit A5) used in the ground-water flow model of the New Jersey Coastal Plain.

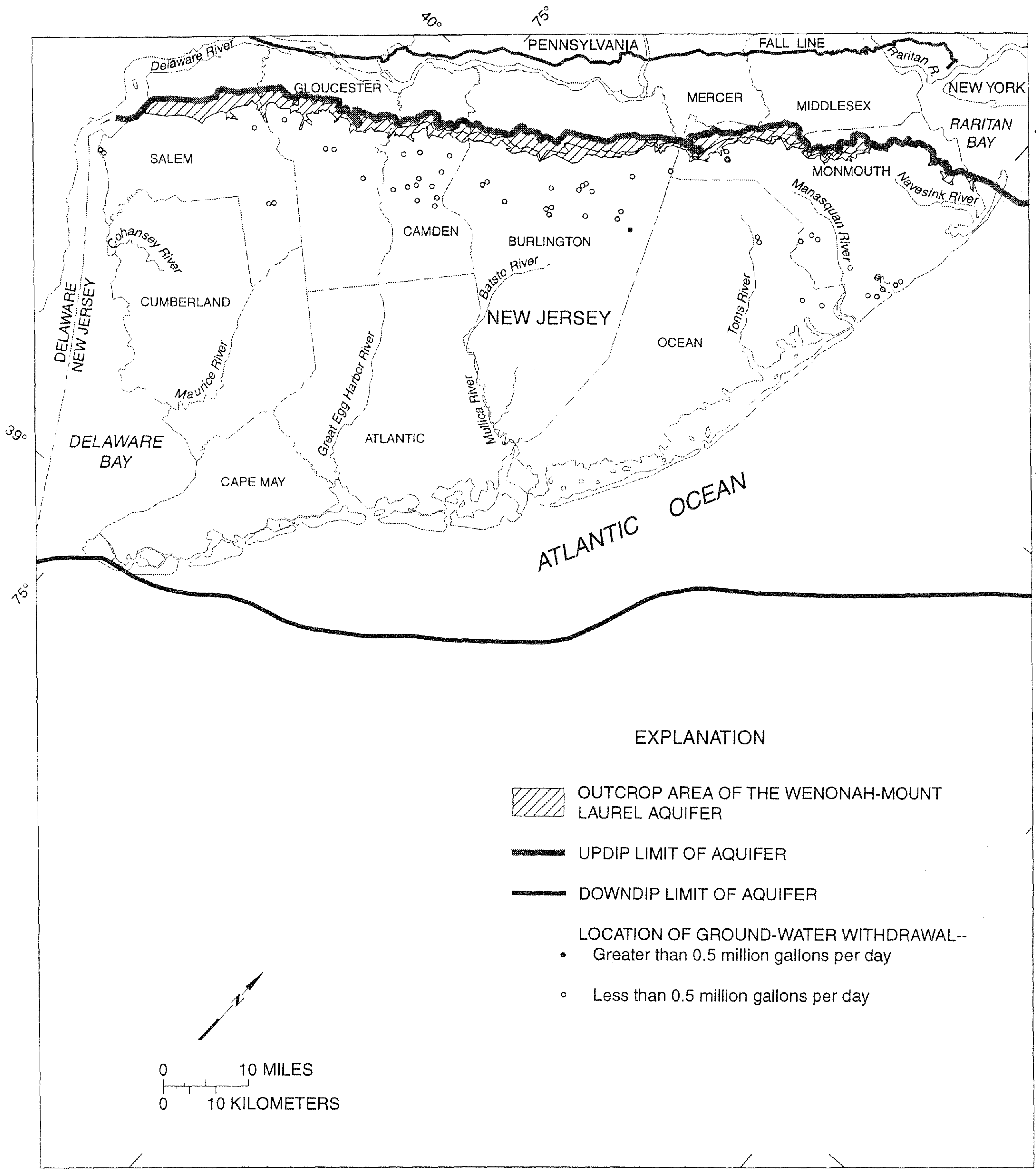


5b. Vertical leakance of the Navesink-Hornerstown confining unit (model unit C5) used in the ground-water flow model of the New Jersey Coastal Plain.



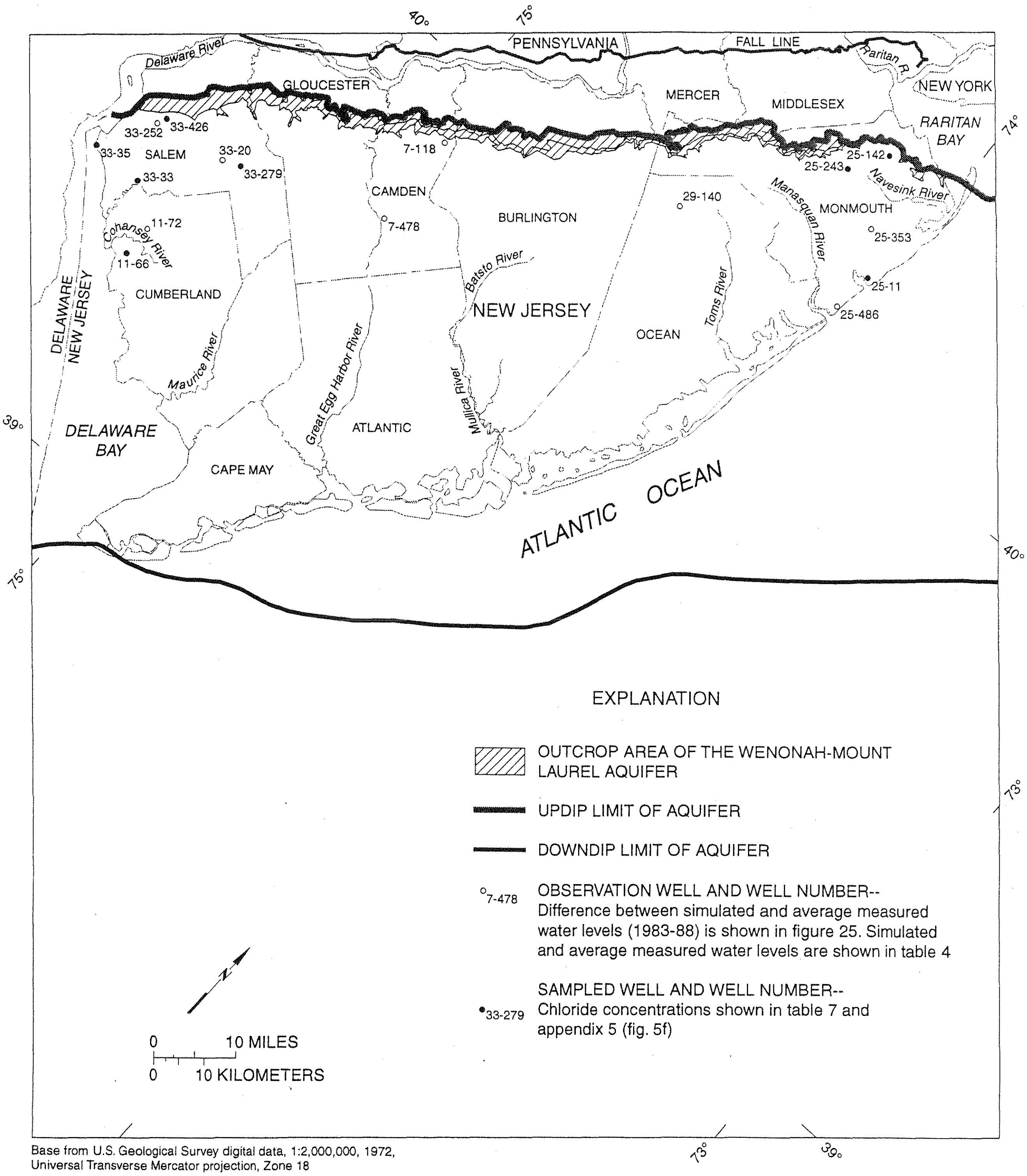
Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

5c. Simulated predevelopment potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow, Wenonah-Mount Laurel aquifer (model unit A5), New Jersey Coastal Plain.

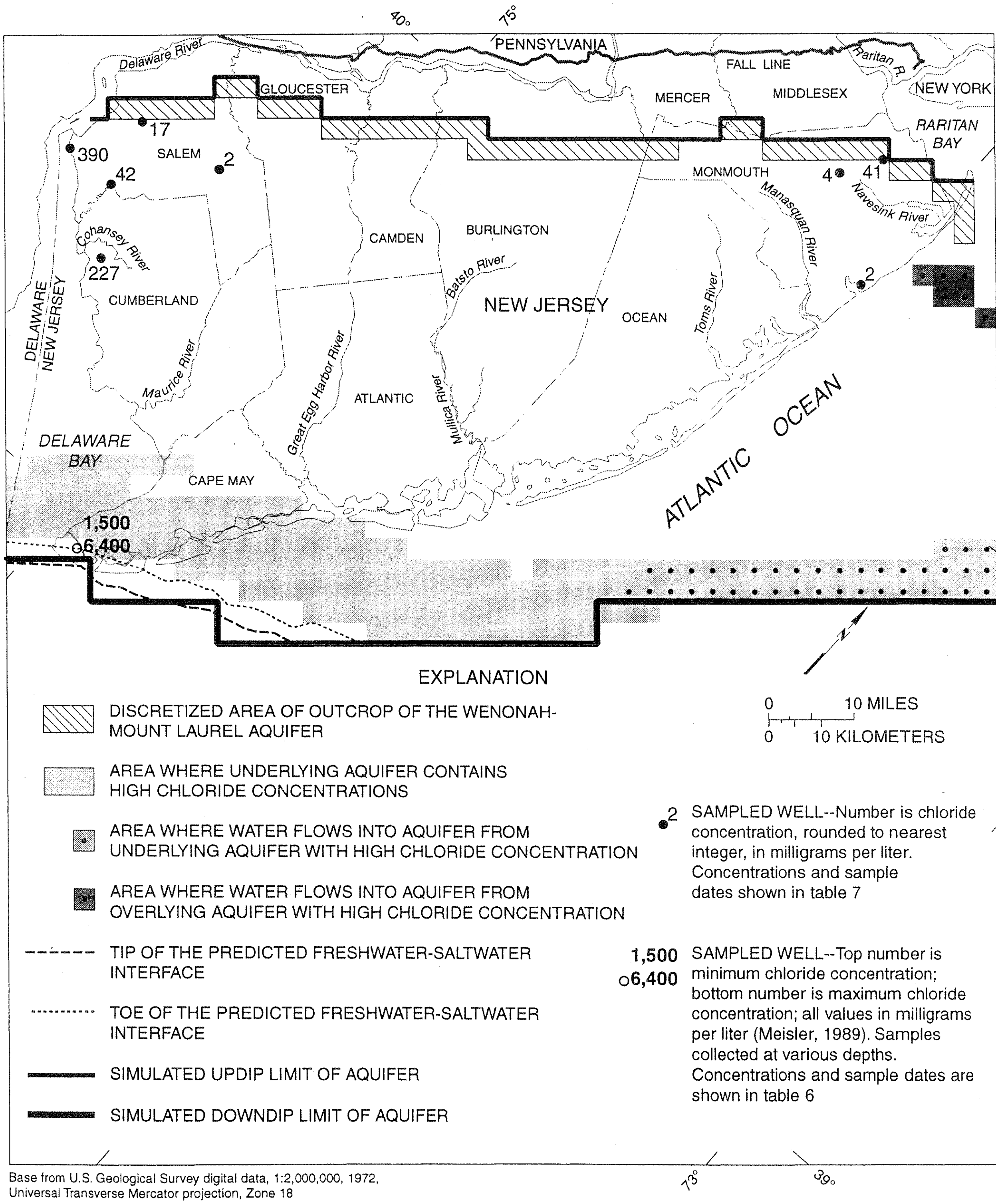


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

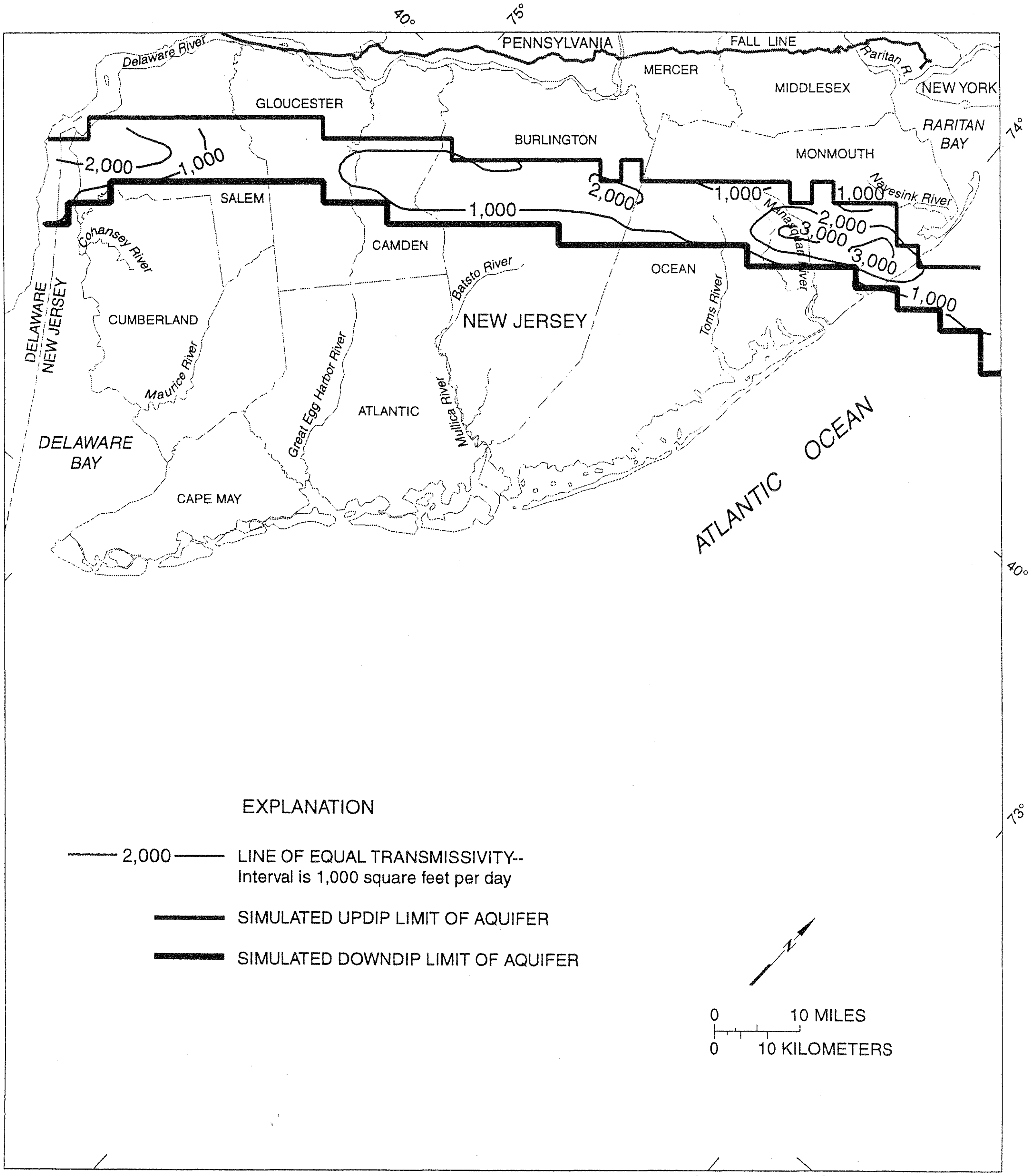
5d. Locations of ground-water withdrawal sites in the Wenonah-Mount Laurel aquifer (model unit A5), New Jersey Coastal Plain, 1983-88.



5e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Wenonah-Mount Laurel aquifer (model unit A5), New Jersey Coastal Plain.

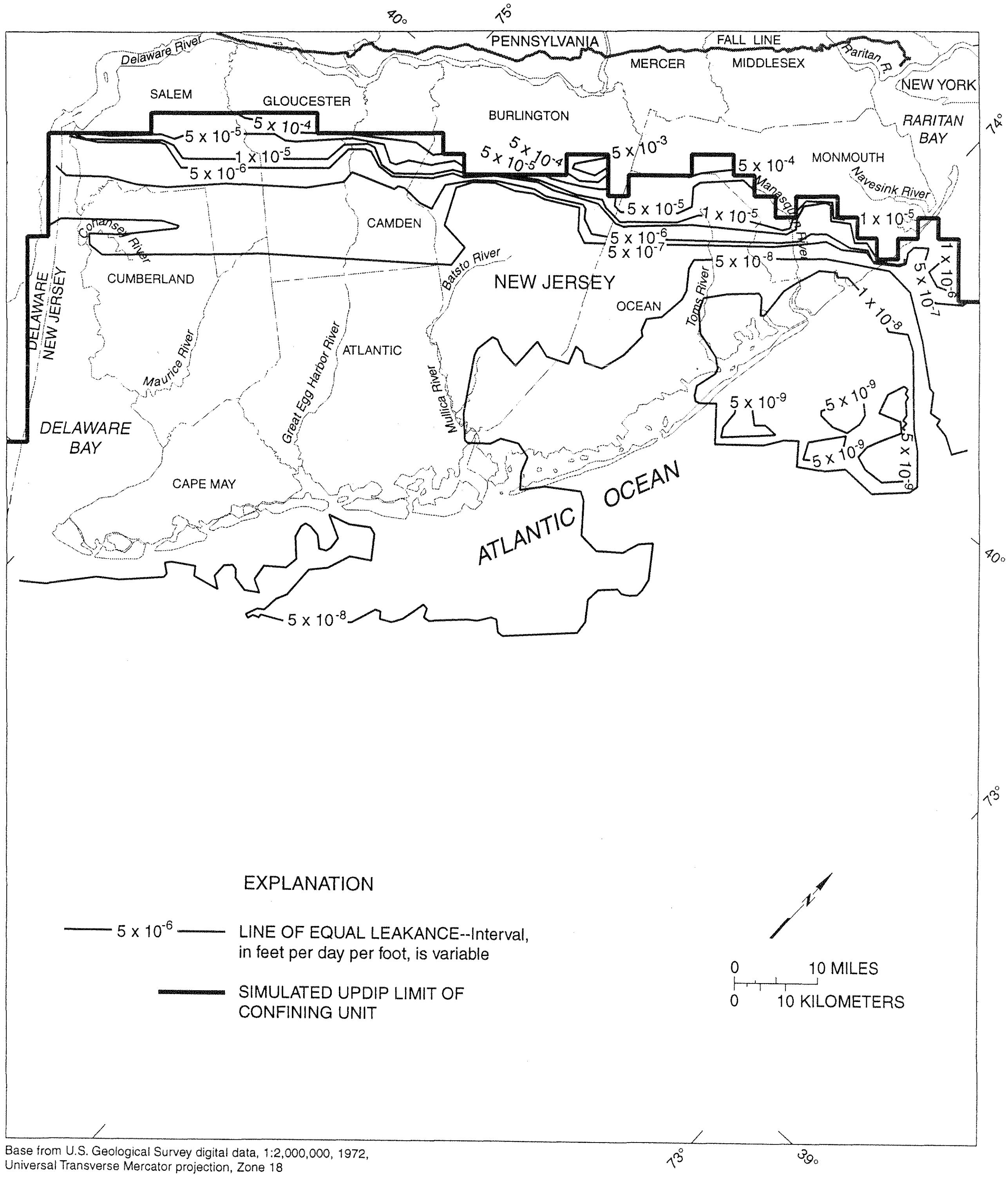


5f. Measured chloride concentrations, location of simulated freshwater-saltwater interface tip and toe, and areas of flow from overlying and underlying aquifers, Wenonah-Mount Laurel aquifer (model unit A5), New Jersey Coastal Plain, 1983-88.

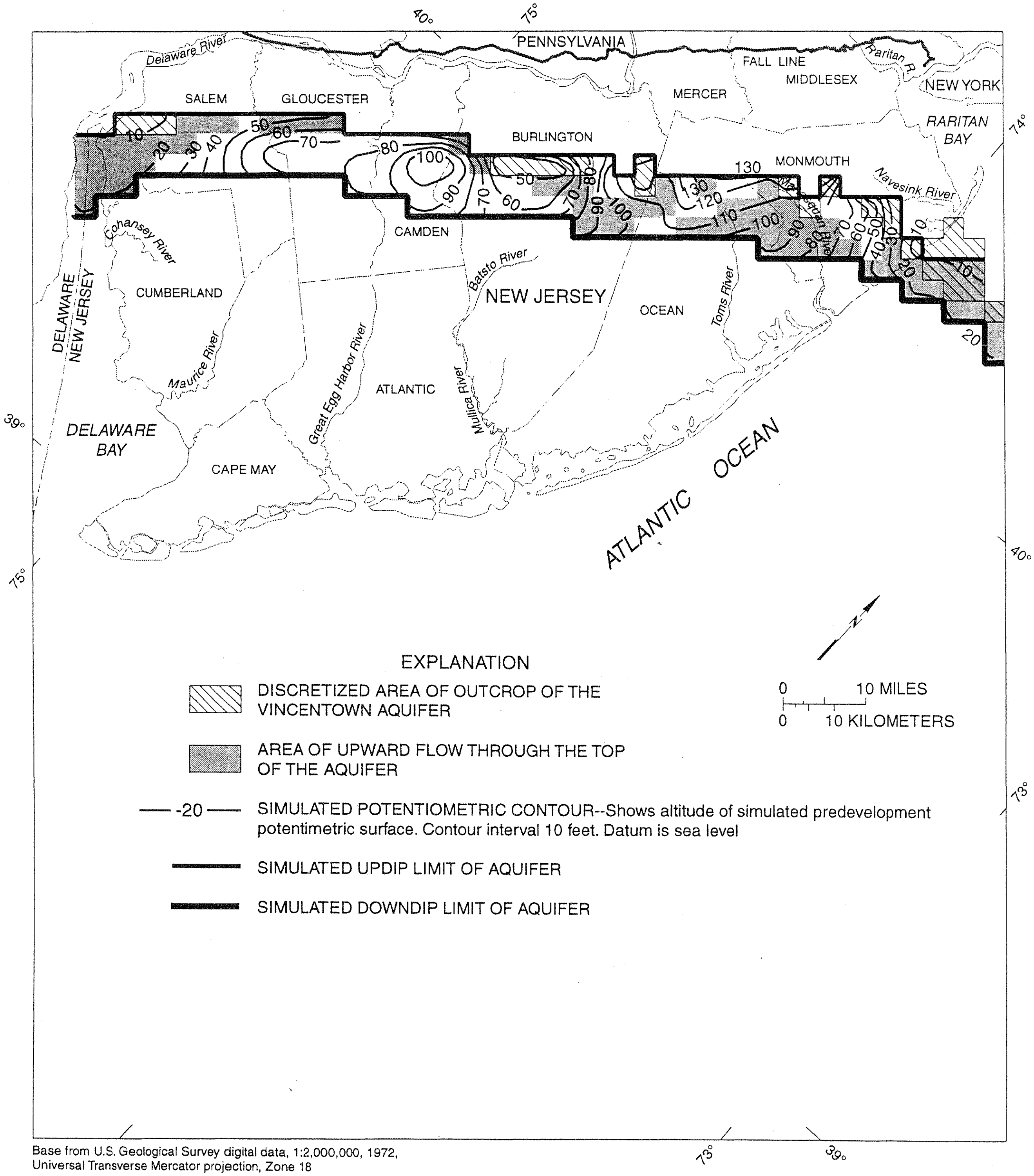


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

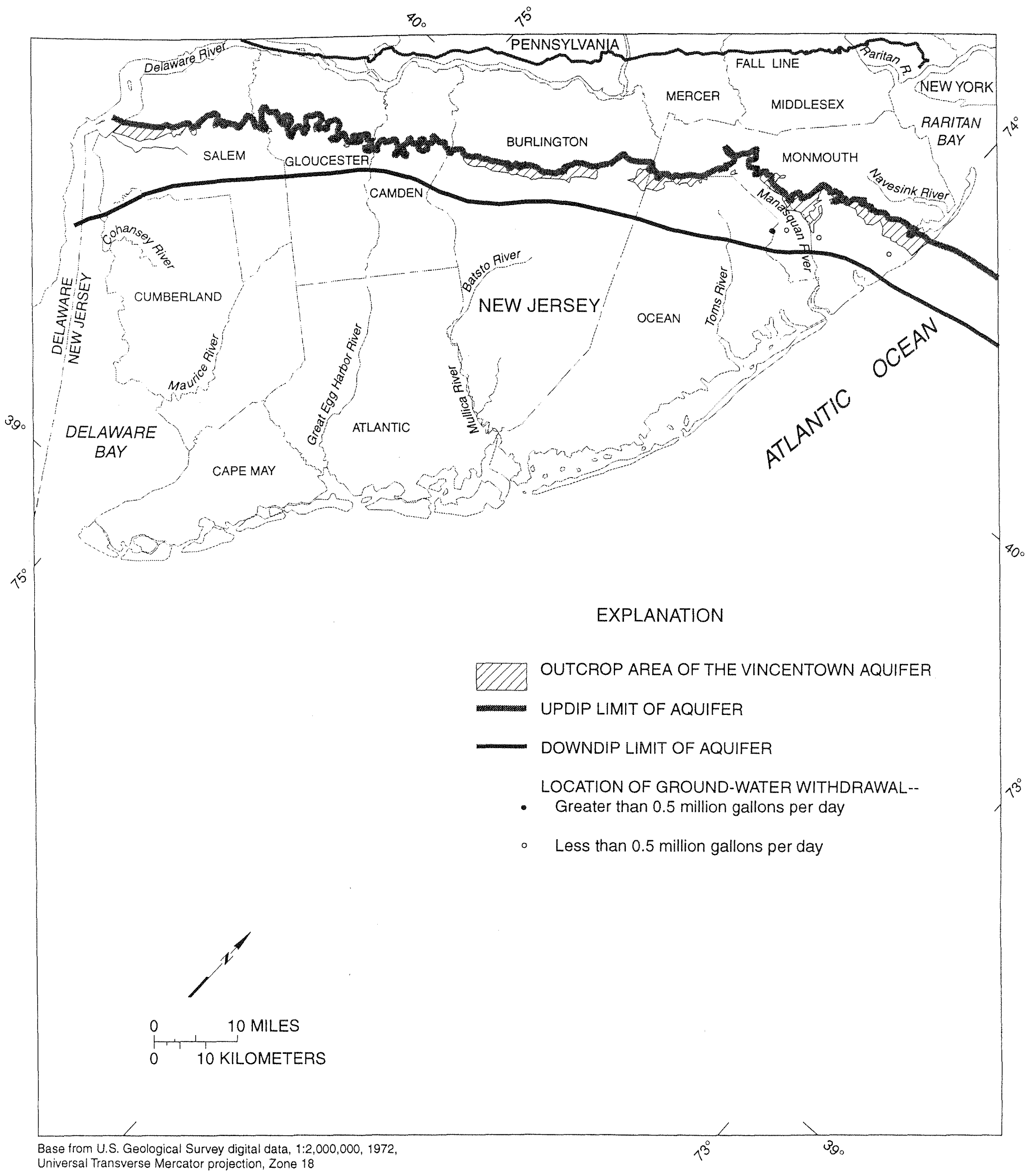
6a. Transmissivity of the Vincentown aquifer (model unit A6) used in the ground-water flow model of the New Jersey Coastal Plain.



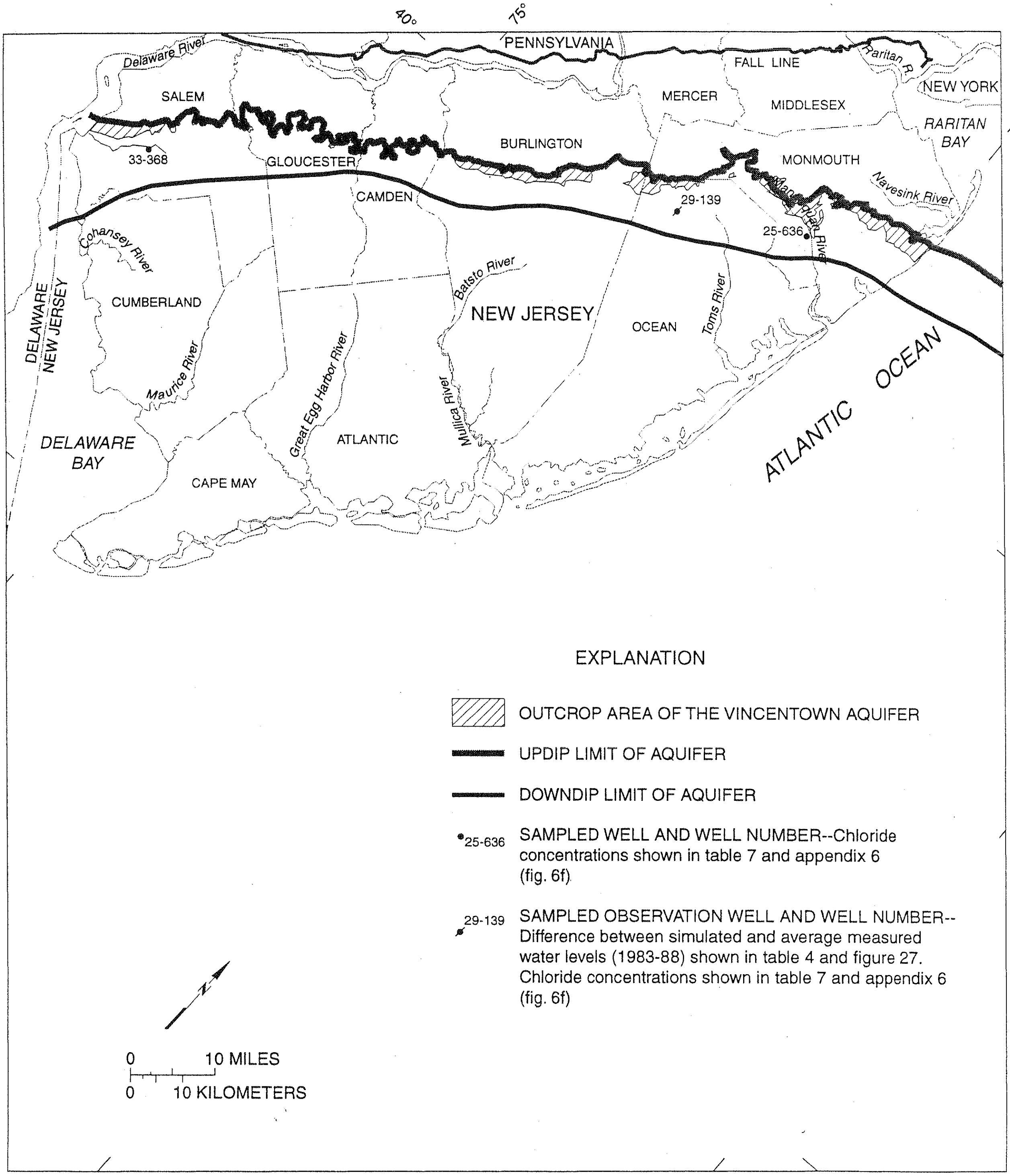
6b. Vertical leakance of the Vincentown-Manasquan confining unit (model unit C6) used in the ground-water flow model of the New Jersey Coastal Plain.



6c. Simulated predevelopment potentiometric surface and area of upward flow, Vincentown aquifer (model unit A6), New Jersey Coastal Plain.

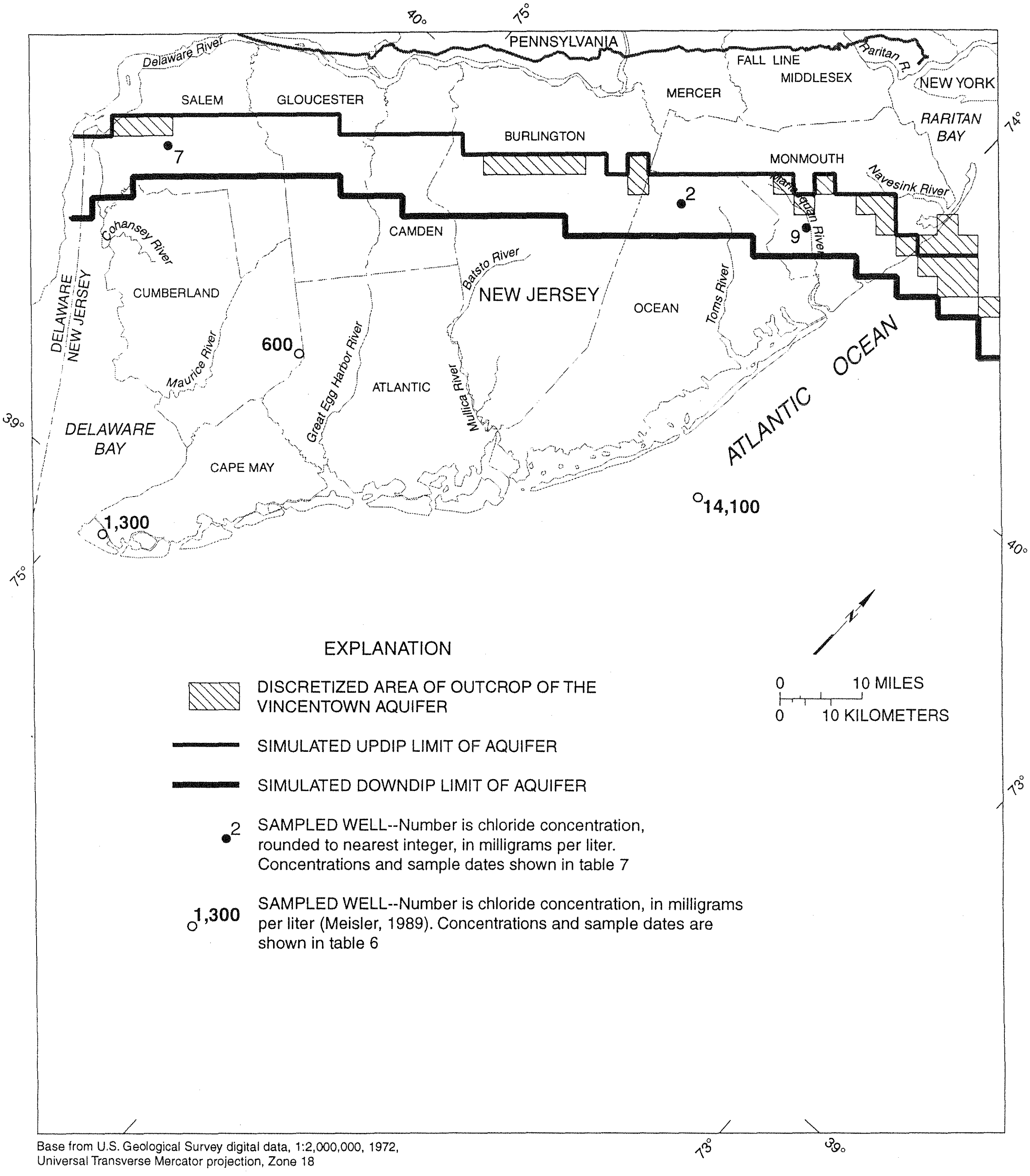


6d. Locations of ground-water withdrawal sites in the Vincentown aquifer (model unit A6), New Jersey Coastal Plain, 1983-88.

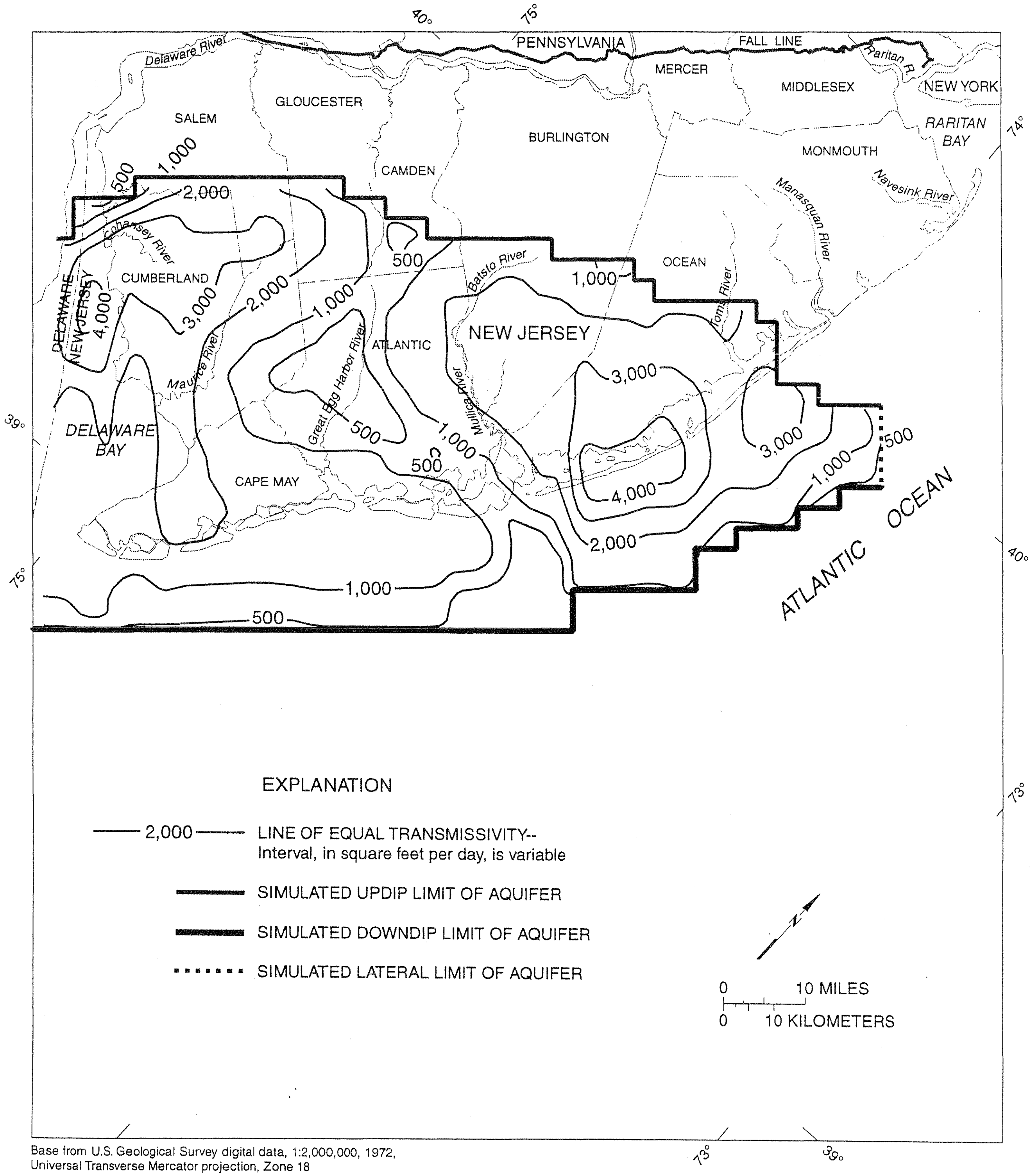


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972,
Universal Transverse Mercator projection, Zone 18

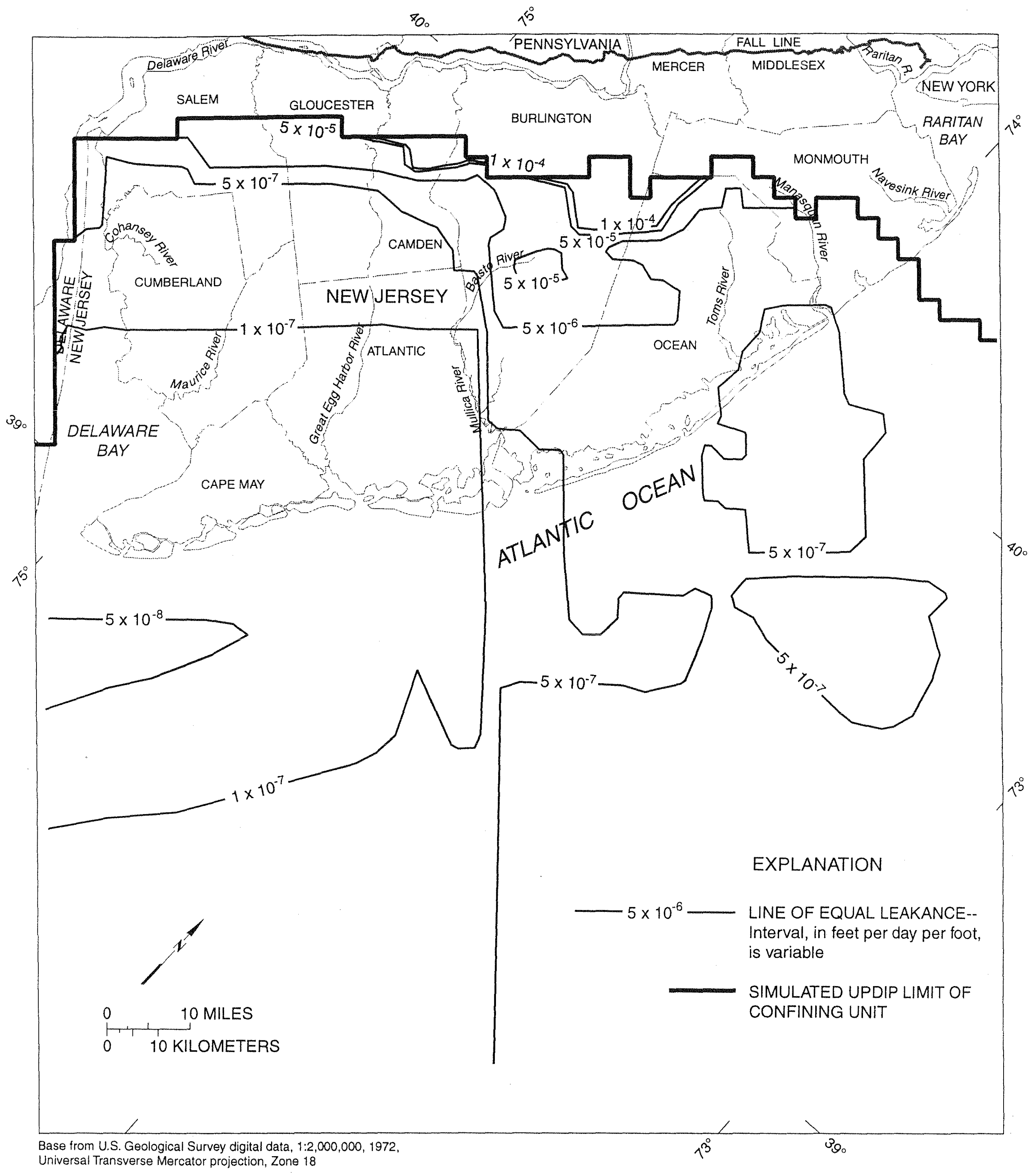
6e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Vincentown aquifer (model unit A6), New Jersey Coastal Plain.



6f. Measured chloride concentrations, Vincentown aquifer (model unit A6), New Jersey Coastal Plain, 1983-88.

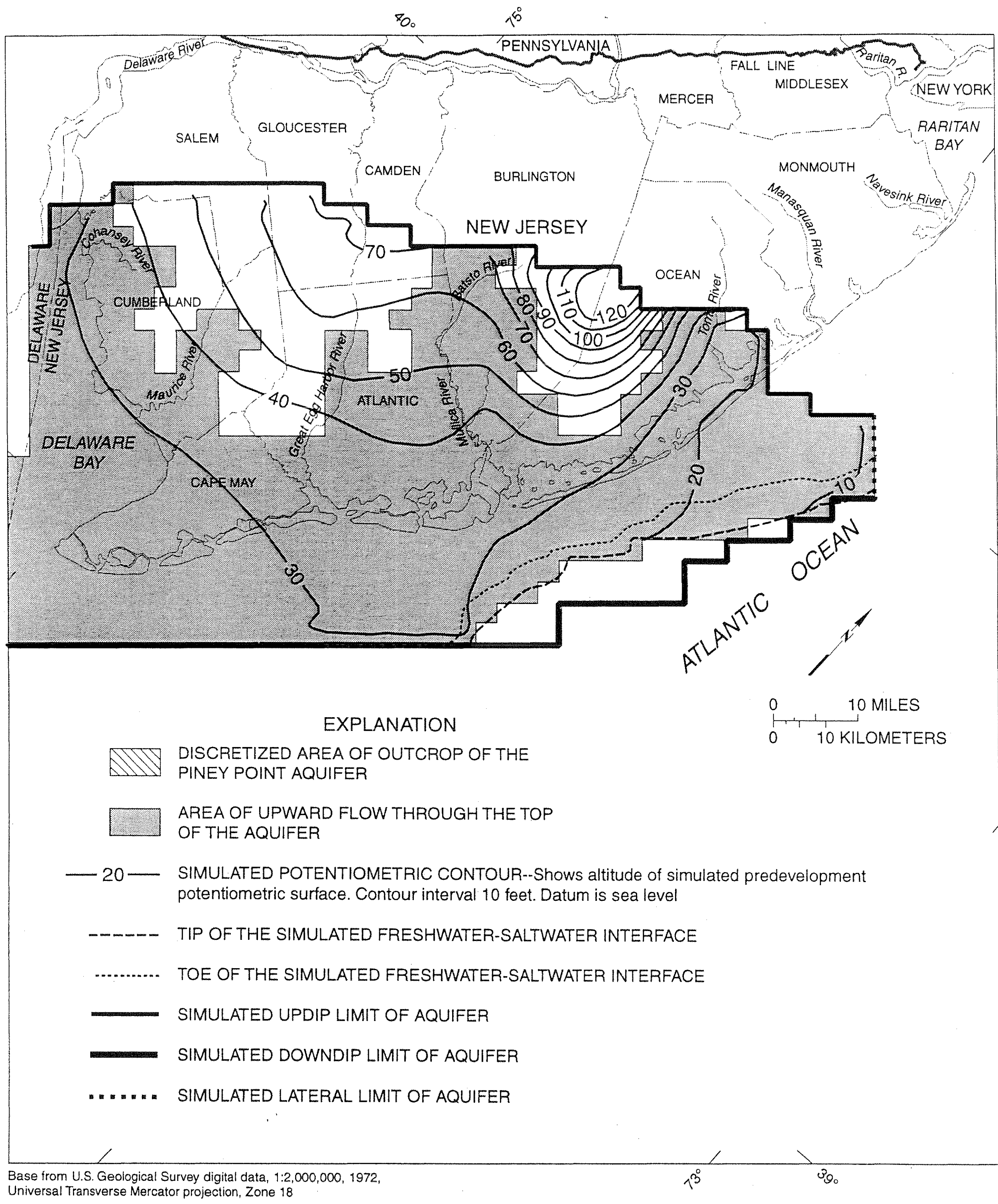


7a. Transmissivity of the Piney Point aquifer (model unit A7) used in the ground-water flow model of the New Jersey Coastal Plain.

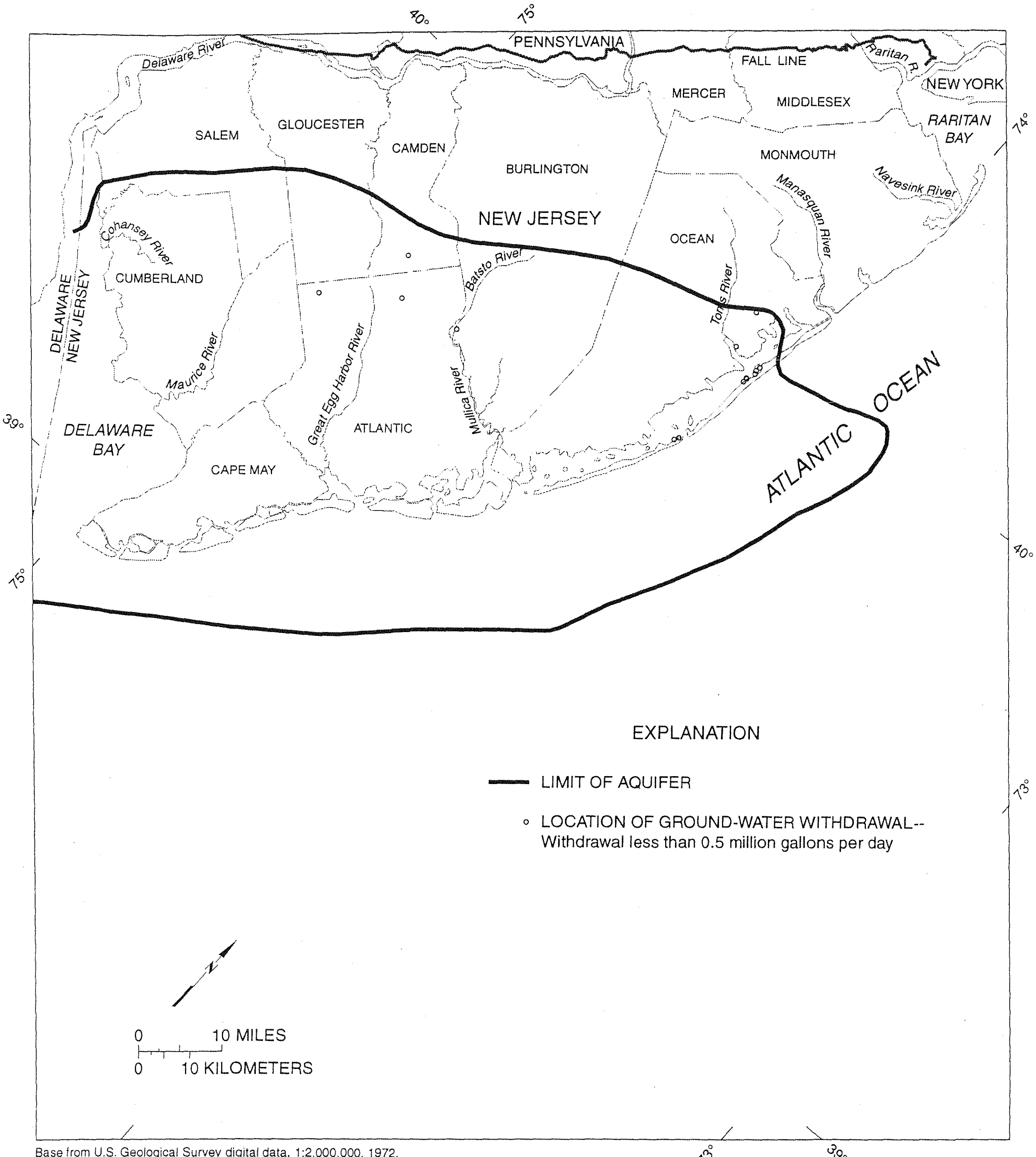


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

7b. Vertical leakance of the basal Kirkwood confining unit (model unit C7) used in the ground-water flow model of the New Jersey Coastal Plain.

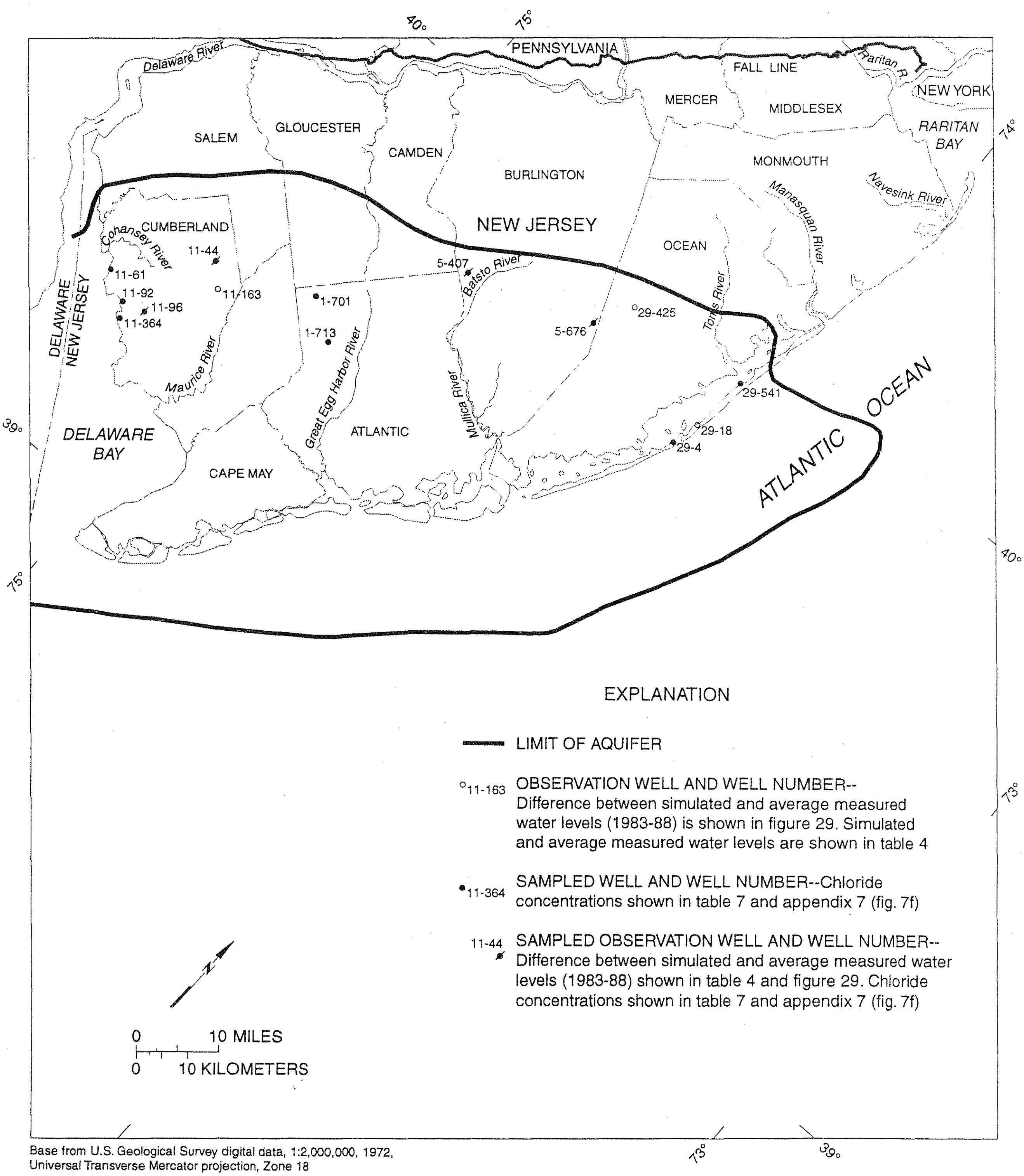


7c. Simulated predevelopment potentiometric surface, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow, Piney Point aquifer (model unit A7), New Jersey Coastal Plain.

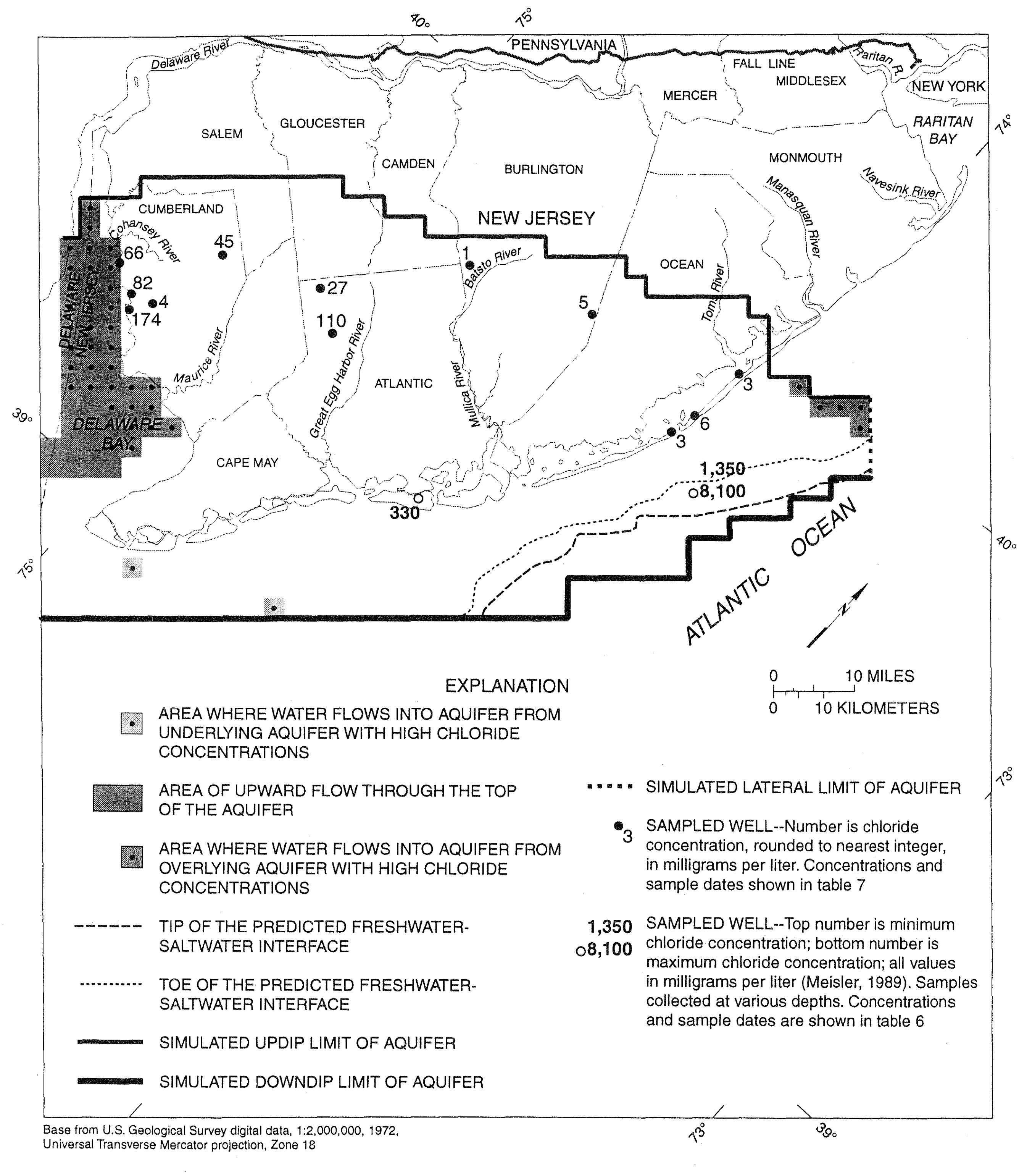


Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

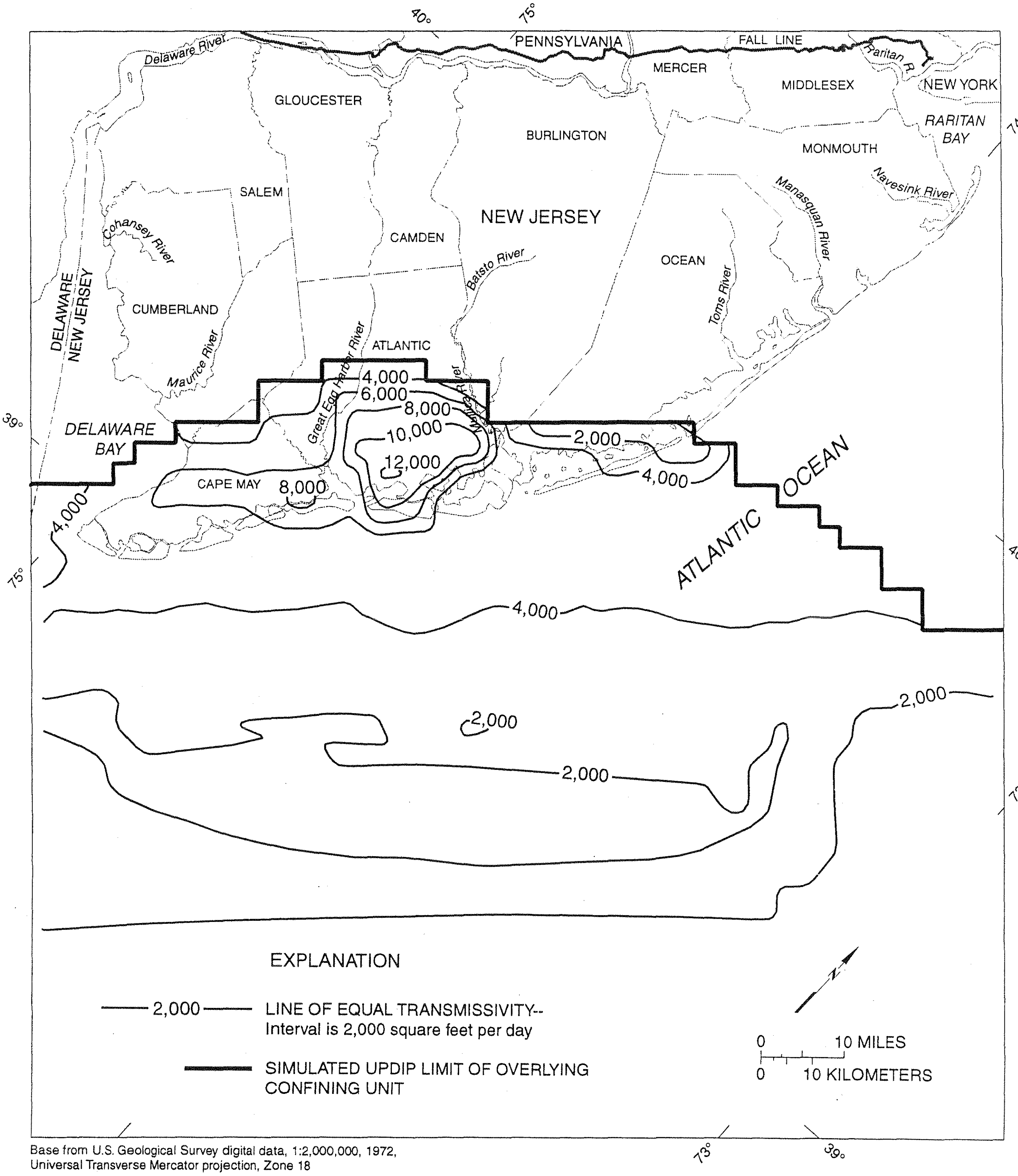
7d. Locations of ground-water withdrawal sites in the Piney Point aquifer (model unit A7), New Jersey Coastal Plain, 1983-88.



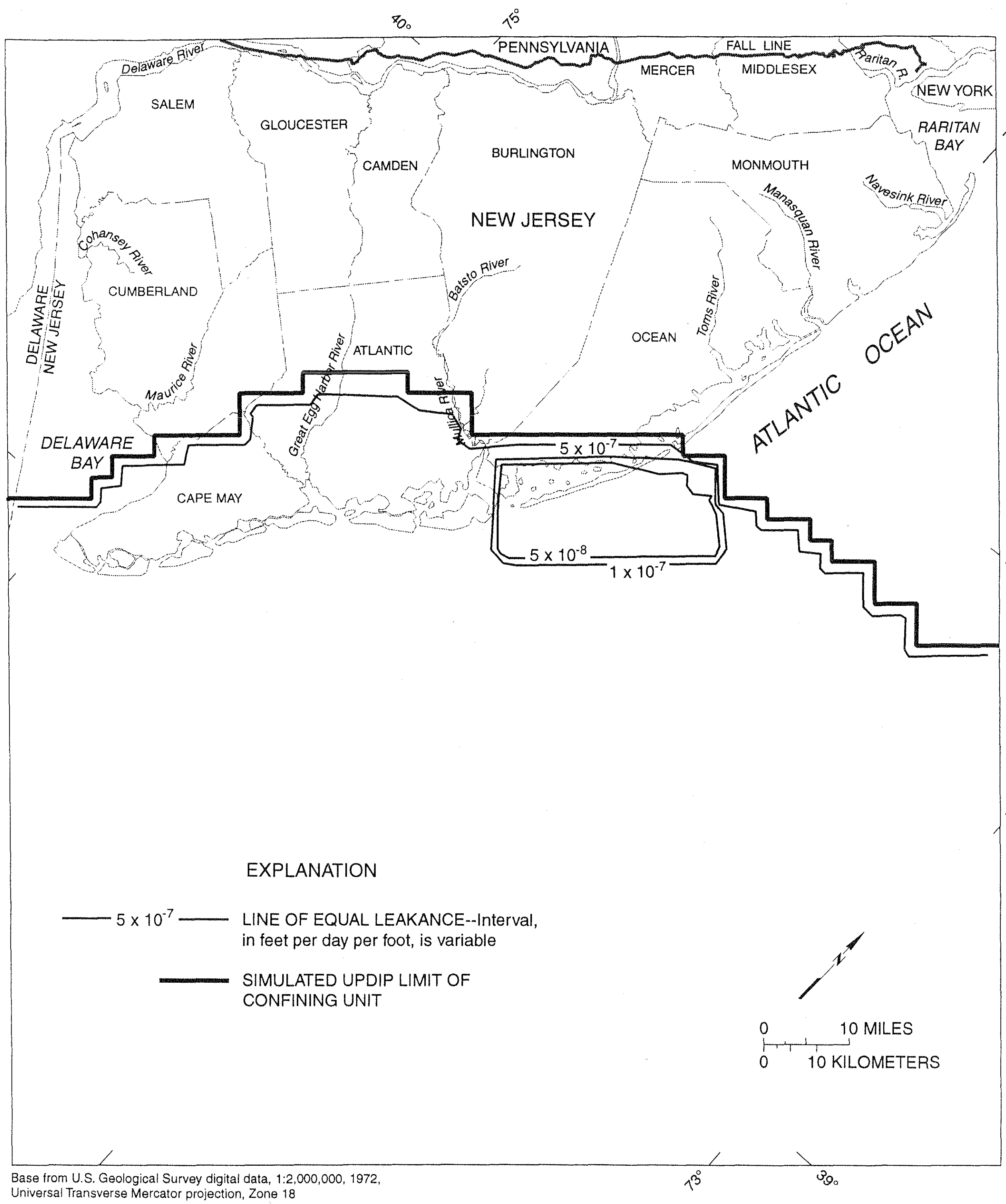
7e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, Piney Point aquifer (model unit A7), New Jersey Coastal Plain.



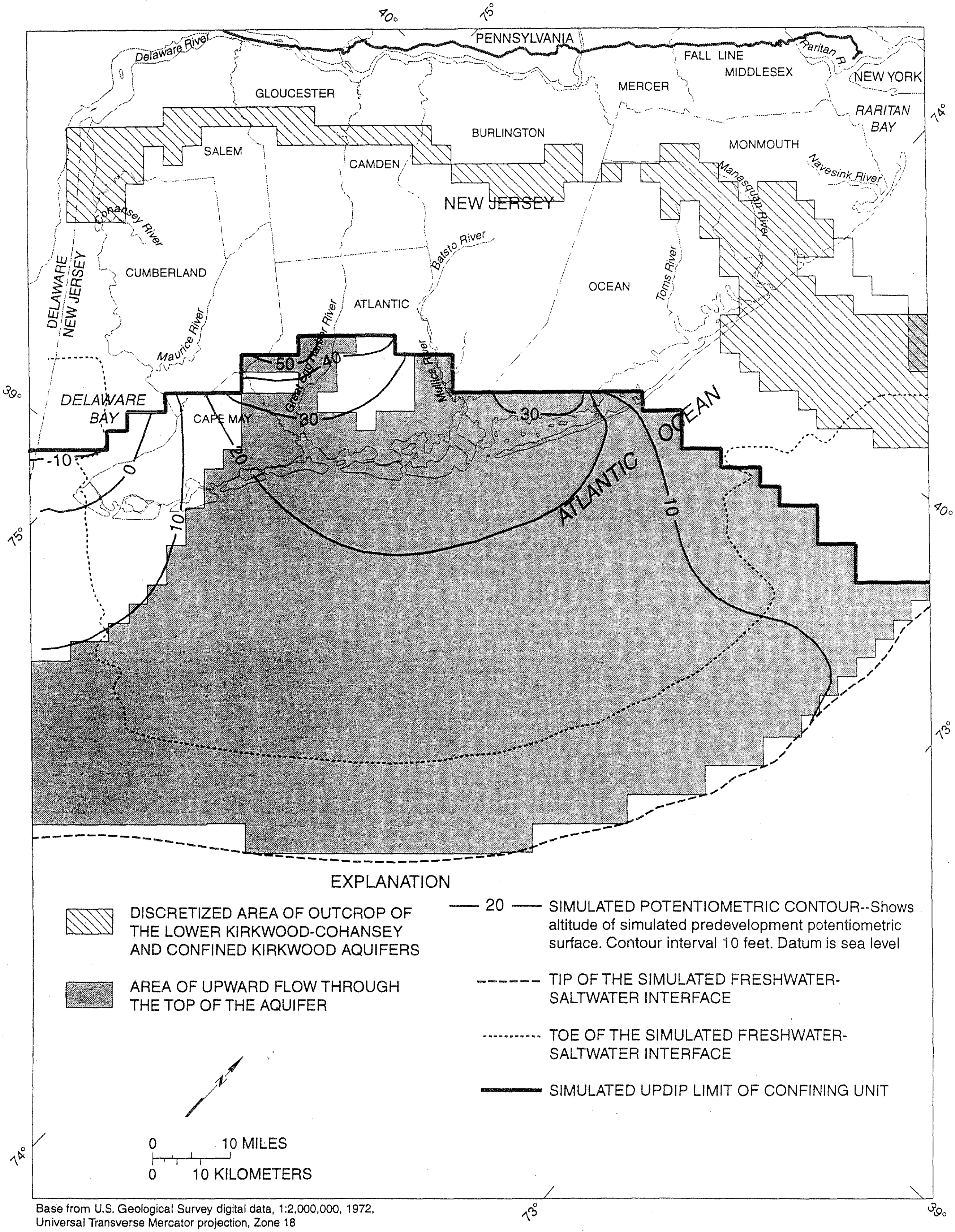
7f. Measured chloride concentrations, location of simulated freshwater-saltwater interface tip and toe, and areas of saltwater flow from overlying and underlying aquifers, Piney Point aquifer (model unit A7), New Jersey Coastal Plain, 1983-88.



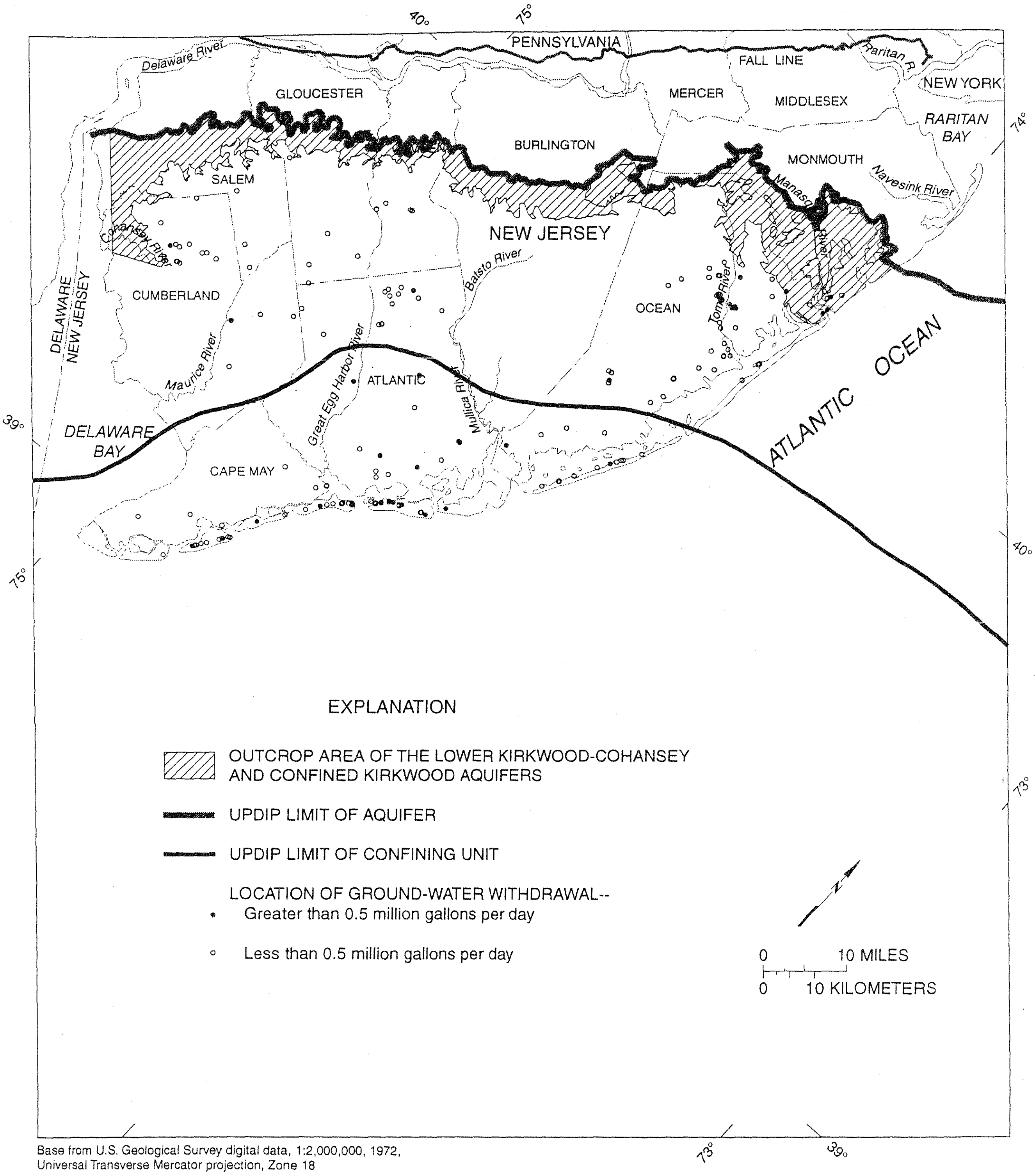
8a. Transmissivity of the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8) used in the ground-water flow model of the New Jersey Coastal Plain.



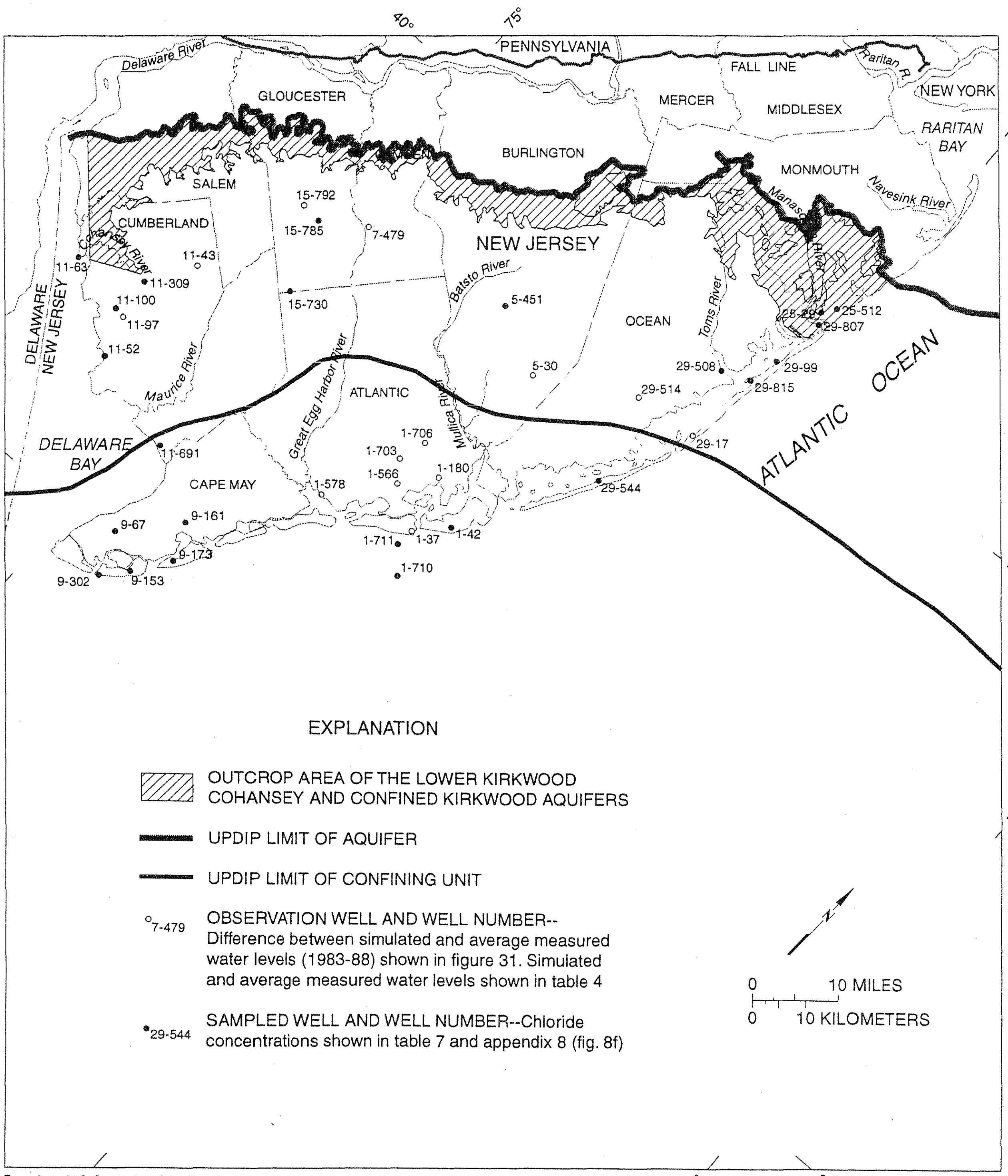
8b. Vertical leakance of the confining unit overlying the Rio Grande water-bearing zone (model unit C8) used in the ground-water flow model of the New Jersey Coastal Plain.



8c. Simulated predevelopment potentiometric surface of the confined Kirkwood aquifer, location of the simulated freshwater-saltwater interface tip and toe, and area of upward flow, lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8), New Jersey Coastal Plain.

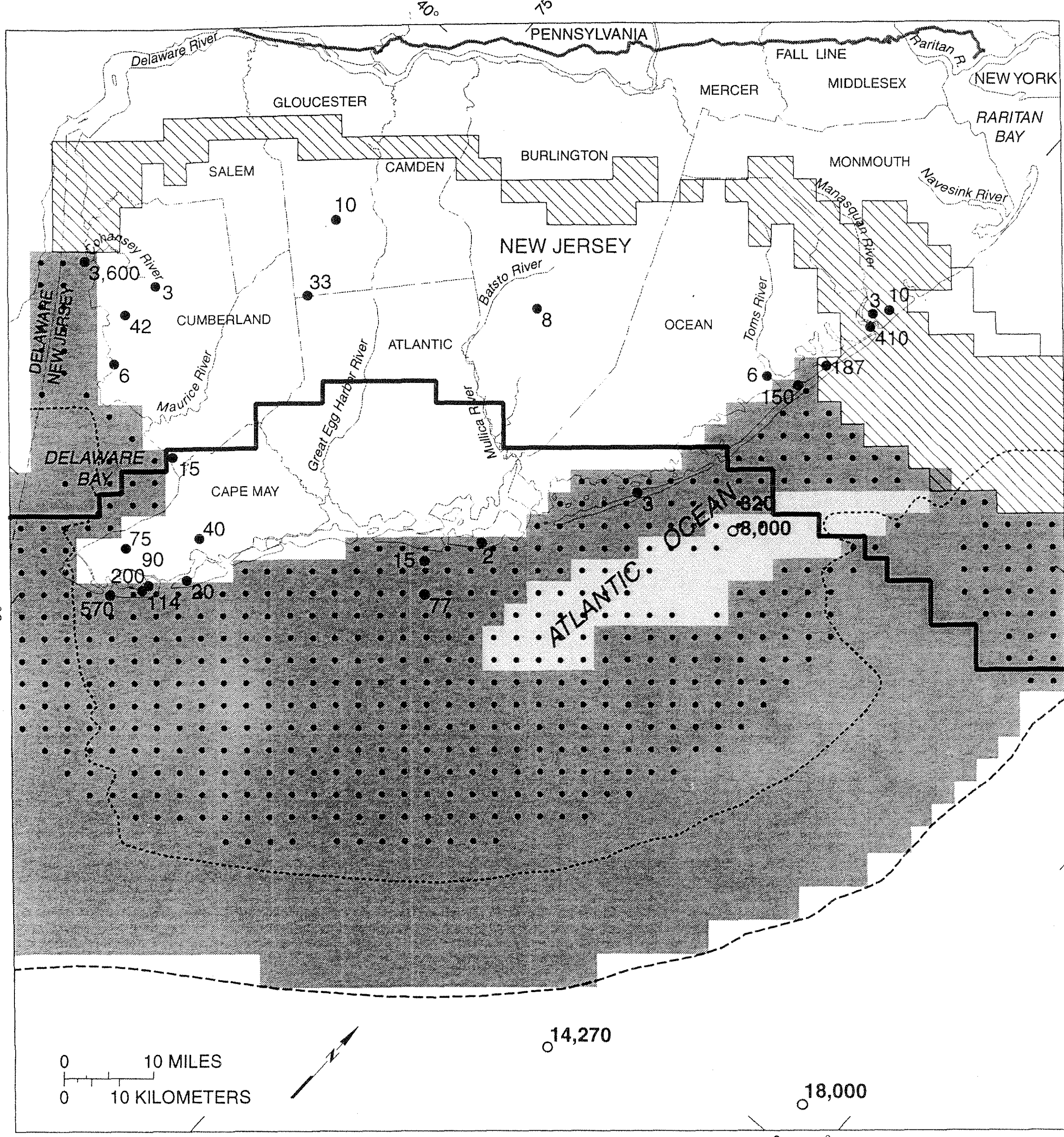


8d. Locations of ground-water withdrawal sites in the lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8), New Jersey Coastal Plain, 1983-88.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

8e. Locations of observation wells for which water-level residuals are available and selected wells for which measured chloride concentrations are available, lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8), New Jersey Coastal Plain.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972, Universal Transverse Mercator projection, Zone 18

EXPLANATION

DISCRETIZED AREA OF OUTCROP OF THE LOWER KIRKWOOD-COHANSEY AND CONFINED KIRKWOOD AQUIFERS

AREA WHERE UNDERLYING AQUIFER CONTAINS HIGH CHLORIDE CONCENTRATIONS

AREA WHERE WATER FLOWS INTO AQUIFER FROM UNDERLYING AQUIFER WITH HIGH CHLORIDE CONCENTRATIONS

AREA OF UPWARD FLOW THROUGH THE TOP OF THE AQUIFER

AREA WHERE WATER FLOWS INTO AQUIFER FROM OVERLYING AQUIFER WITH HIGH CHLORIDE CONCENTRATIONS

TIP OF THE PREDICTED FRESHWATER-SALTWATER INTERFACE

TOE OF THE PREDICTED FRESHWATER-SALTWATER INTERFACE

SIMULATED UPDIP LIMIT OF AQUIFER

3 — SAMPLED WELL--Number is chloride concentration, rounded to nearest integer, in milligrams per liter. Concentrations and sample dates shown in table 7

1,350 / 8,100 — SAMPLED WELL--Top number is minimum chloride concentration; bottom number is maximum chloride concentration; all values in milligrams per liter (Meisler, 1989). Samples collected at various depths. Concentrations and sample dates are shown in table 6

8f. Measured chloride concentrations, location of simulated freshwater-saltwater interface tip and toe, and areas of saltwater flow from overlying and underlying aquifers, lower Kirkwood-Cohansey and confined Kirkwood aquifers (model unit A8), New Jersey Coastal Plain, 1983-88.