GEOHYDROLOGY OF, AND SIMULATION OF GROUND-WATER FLOW IN, THE VALLEY-FILL DEPOSITS IN THE RAMAPO RIVER VALLEY, NEW JERSEY

By M.C. Hill, G.P. Lennon, G.A. Brown, C.S. Hebson, and S.J. Rheaume

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CONVERSION FACTORS AND ABBREVIATIONS

Multiply inch-pound unit	By	To obtain metric unit
	Length	
inch (in.)	2.540	centimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
acre	0.004047	square kilometer
square foot (ft^2)	0.09290	square meter
square mile (mi^2)	2.590	square kilometer
	Flow	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
gallon per minute (gal/min)	0.06308	liter per second
million gallons per day (Mgal/d)	0.04381	cubic meter per second
	Hydraulic conductivity and transmissivity	
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.09290	meter squared per day
	Specific capacity	
gallon per minute per foot [(gal/min)/ft]	0.2070	liter per second per meter
	Ground-water yield	
gallon per day per square mile [$(gal/d)/mi^2$]	0.001461	cubic meter per day per square kilometer

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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 both the United States and Canada, formerly called Sea Level Datum of 1929.

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ABSTRACT

The Ramapo River valley is a narrow valley bordered by bedrock highlands in northeastern New Jersey and southeastern New York. The water resources of the valley are used extensively for public supply.

The valley-fill deposits of the Ramapo River valley form the most productive aquifer in the basin. In 1982, total pumpage from major well fields in the valley exceeded 15 million gallons per day. The valley-fill deposits, which are as much as 200 feet thick along the center of the valley, consist mostly of sand and gravel and generally are under water-table conditions. In the northern part of the study area, however, a silt and clay layer about 2 miles long and 0.5 mile wide confines a basal sand and gravel layer. Near the center of this area, the vertical hydraulic conductivity of the confining unit was 3×10^{-4} feet per day based on the results of permeameter tests, the calculated average transmissivity of the confined aquifer was 15,700 feet squared per day, based on results of two aquifer tests, and the average storage coefficient was 1.3×10^{-4}. The aquifer-test data indicate that recharge to the confined aquifer through the confining unit is less than recharge around its edges.

Three comprehensive base-flow seepage runs were made in the study area to investigate induced seepage from the Ramapo River. Results of these seepage runs indicate that the Ramapo River is hydraulically connected to the underlying aquifer, and that gaining and losing reaches are present under natural conditions. Results of two local seepage runs confirmed the results of the comprehensive seepage runs. Streambed hydraulic conductivity measurements, based on data from the local seepage runs, ranged from 25 to 35 feet per day.

A calibrated three-dimensional numerical model was constructed to quantify the hydrogeologic characteristics of the ground-water system and to evaluate the hydrologic relations between ground-water withdrawals and streamflow in the northern part of the study area. Results of simulations indicate that measured streamflow gains and losses caused by ground-water withdrawals from the valley-fill deposits are affected by variations in the hydrogeologic characteristics of the ground-water system. For instance, differences in hydraulic conductivity, aquifer thickness, and valley width can affect the water-transmitting properties of the valley-fill deposits. Also, the presence of a shallow confining unit in the northern part of the study area reduces the hydraulic connection between the Ramapo River and the Mahwah Township wells screened in the underlying confined aquifer. As a result, streamflow losses caused by ground-water withdrawals are shifted upstream or downstream to areas of increased hydraulic connection where the confining unit is absent.

INTRODUCTION

The valley-fill deposits of northern New Jersey and their associated rivers and streams are productive sources of both ground-water and surface-water supply. Generally, ground water is the source of water for users within the valleys and surface water is the source for reservoirs that supply water to the densely populated areas of northeastern New Jersey, including Jersey City and Newark. In recent years, drought conditions have reduced streamflows and storage in the reservoirs to the extent that the reservoirs have been unable to provide sufficient water to meet the demand. In the mid 1960's, mid 1970's, and twice in the early 1980's, drought conditions resulted in extremely low water levels in the reservoirs, and water-use restrictions were imposed in northern New Jersey.

An investigation of the geohydrology of the valleys that contribute water to the reservoirs and of the stream-aquifer interactions in the valleys was begun by the State of New Jersey. This study was done by the U.S. Geological Survey in cooperation with the New Jersey Department of Environmental Protection and Energy.

Purpose and Scope

This report describes the results of an investigation of the geohydrology of the Ramapo River valley conducted during 1981-84. The report includes information on the hydrogeologic setting, hydraulic properties, rates of pumping, water levels, and stream-aquifer interactions of the valley-fill deposits along the Ramapo and Mahwah Rivers and Masonicus Brook in the New Jersey part of the Ramapo River basin. Results of a simulation of the valley-fill aquifer near Mahwah, New Jersey, by means of a numerical model constructed to quantify the characteristics of the ground-water system and to evaluate the hydrologic relations between ground-water withdrawals and streamflow also are presented.

Methods of Study

The hydrology of the valley-fill deposits was defined by a variety of methods. The hydrogeology was defined on the basis of lithologic and hydrologic data from wells, borings, and potentiometers, and from surface-geophysical studies. Hydraulic properties of the ground-water system were determined by using well yields reported by drillers, and by analyzing data from three aquifer tests and two permeameter tests. Direction of ground-water flow was determined by contouring water levels measured in water-table observation wells, and by determining differences in heads measured in wells screened at different depths. Surface-water hydrology was defined on the basis of stream-discharge data from continuous-recording stations. Ground-water/surface-water interactions were described on the basis of data from both comprehensive and local seepage runs. Ground-water pumpage was determined from data obtained from local water departments.

The elevations of points from which any ground- or surface-water levels were measured were determined to within 0.01 ft (foot) by standard surveying methods. The elevations of other points (for example, borings used only to define the hydrogeology) were determined to within 0.2 ft by altimeter, or to within 5 ft using 1:24,000-scale U.S. Geological Survey quadrangle maps having 10-ft topographic-contour intervals.

These data were used to develop a three-layer finite-difference model to simulate ground-water flow in the valley-fill aquifer system near Mahwah, New Jersey. The model was calibrated by the trial-and-error method, and sensitivity runs were used to determine the accuracy of model-input values.

Previous Studies

The geology of the Ramapo River basin was described by Kummel (1898, 1899) in his study of the rocks that compose the Newark Supergroup and by Salisbury (1902) in his study of the glacial deposits. Kummel (1940) described the geology of the entire State of New Jersey, and Johnson (1950) revised the geologic map of the State. Vecchioli and Miller (1973) summarized existing geologic studies of the area. The rock units in the New York part of the basin also have been mapped (New York State Museum and Science Service, 1961).

The water resources of the Ramapo River basin were described briefly by Vermeule (1894) in his study of the water supply of the State of New Jersey. Weston and Sampson (1924) considered the basin as a potential source of water for the City of Bayonne. Tippetts-Abbett-McCarthy-Stratton, Engineers (1955) described the water-supply potential of the Ramapo River and the general ground-water potential of the basin. The water resources of the basin also were described by the Bergen County Water Study Committee (1957), Widmer and others (1966) and Neglia and others (1967). Carswell and Rooney (1976) discussed the depth to bedrock and relative permeability of the bedrock in the southern end of the basin near Pompton Plains. A detailed evaluation of the water resources of the basin was presented by Vecchioli and Miller (1973). Their study includes a description of the geology of the basin and the available information on the hydraulic properties of the geologic units. They also considered the water quality and discharge characteristics of streamflow originating in areas underlain by Precambrian gneiss and the Newark Supergroup, and the relation between nearby ground-water pumpage and induced recharge to the aquifers from the Ramapo River. A detailed summary of daily streamflow data for 1922-66 near Mahwah (station 01387500) and at Pompton Lakes (station 01388000) is included.

A number of studies focused on the water resources of the Rockland County, New York, part of the Ramapo River basin: Perlmutter (1959) studied primarily the geology and ground-water resources of the Newark Supergroup, Moore and others (1982) studied the geology and ground-water resources of the valley-fill aquifer along the Ramapo and Mahwah Rivers, and Ayer and Pauszek (1963) evaluated the surface-water resources of Rockland County and their potential for water supply. An areal, two-dimensional ground-water flow model of the valley-fill deposits along a 2-mi (mile) reach of the Ramapo River just north of the New York-New Jersey State line was constructed to simulate the influence on the river of pumping from Spring Valley Water Company wells (Leggette, Brashears and Graham, Inc., 1981 and 1982).

None of the existing studies of the Ramapo River valley describes the ground-water system and ground-water/surface-water interactions in sufficient detail for the present study. Specifically, in most of the study area, previous studies have not identified the aquifers and confining units within the valley-fill deposits, determined the hydraulic properties of the ground-water system, described ground-water flow, or identified losing and gaining reaches of the stream in detail.

Stream-discharge measurements made in the basin and surface-water-quality data from the Ramapo River near Mahwah (station 01387500) are reported in the annual water-data reports of the U.S. Geological Survey (U.S. Geological Survey, 1981, 1982, and 1983). Surface- and ground-water-quality data collected in the basin in 1964 and 1965 are presented, analyzed, and discussed in Vecchioli and Miller (1973).

Well-Numbering and Location System

Wells, test holes, and potentiometers discussed in this report are shown on plate 1. The municipality and the latitude and longitude of the location of each well were determined by locating the wells on U.S. Geological Survey 1:24,000-scale quadrangle maps. Construction features and yield characteristics of the wells, test holes, and deep potentiometers are presented in table 1. Construction features of the shallow potentiometers are discussed later in this report.

Each well, test hole, or potentiometer has two identifying numbers. The first is the well number used by the U.S. Geological Survey in New Jersey or New York. The New Jersey well numbers listed in the first column of table 1 consist of a two-digit number, which identifies the county in which the well is located, and a four-digit sequence number. For example, the New Jersey well number 03-196 identifies the 196th well inventoried in county 03, which is Bergen County. The New York well numbers consist of the six-digit latitude of the well, a zero, the six-digit longitude of the well, and a two-digit sequence number. The sequence number is shown in the first column of table 1; the latitude and longitude are in columns 5 and 6. For example, columns 5, 6, and 1 can be combined to form the New York well number 410655074085701, which identifies the first well inventoried at latitude 41° 06′ 55″, longitude 74° 08′ 57″.

The second identifying number is a six-digit location code composed of three two-digit numbers separated by hyphens (column 2 of table 1). The first two-digit number of the location code is the minutes of latitude of the well's location; the second two-digit number is the minutes of longitude. These two numbers define rectangular areas in the grid shown on plate 1. The third number of the location code is the sequence number of the wells within each rectangular area. In this report, where data from these wells are shown on maps, the sequence number is listed beneath the data. The location code can be derived by using the grid shown on each map and the sequence number.

Wells in table 1 are grouped by municipality. Within these groups the wells are ordered by increasing values of the location code.

Acknowledgments

The authors acknowledge the assistance of officials and private individuals in the Ramapo River valley area who furnished information on their wells and permitted access to their wells and property. Mary B.

Patrick, Stuart Ostrow (Mahwah Township Health Department), and officials at the Bergen County Health Department, the Mahwah Township Water Department, the Oakland Borough Water Department, and McGee Brothers Gravel Pit were among those who provided such assistance. Thanks are extended to the officials of the Mahwah Township Water Department and the drillers of Rinbrand Well Drilling Company for their cooperation with our efforts to monitor water levels during the aquifer test they conducted.

DESCRIPTION OF STUDY AREA

Location and Physiographic Setting

The study area, shown on plate 1, is the Ramapo River valley, which consists of the valley-fill aquifer along the Ramapo and Mahwah Rivers and Masonicus Brook, and includes parts of Oakland Borough and Mahwah Township in Bergen County, New Jersey, and Rockland County, New York. Towns in the study area include Oakland, Darlington, Mahwah, and West Mahwah, New Jersey, and Suffern and Hillburn, New York.

The Ramapo River valley is in the Ramapo River basin, which is part of the Passaic River basin (fig. 1). The Ramapo River drains an area of 163 mi^2 (square miles), 113 mi^2 of which is in New York State (R.D. Schopp, U.S. Geological Survey, oral commun., 1985). The headwaters of the Ramapo River are near Monroe, New York. For most of its course the Ramapo River follows a low valley through highlands that rise from 300 to 1,000 ft above the river surface. The Mahwah River, the major tributary of the Ramapo River, flows through similar terrain and joins the Ramapo River just south of the New York-New Jersey State line. Masonicus Brook (pl. 1), a tributary of the Mahwah River, joins the Mahwah River just before the confluence of the Mahwah River and Ramapo River. For most of its length, Masonicus Brook follows a low valley bordered by highlands that rise 150 to 300 ft above the brook. The low valleys of the Ramapo and Mahwah Rivers and Masonicus Brook are underlain by up to about 170 ft of valley-fill deposits in the study area.

Other tributaries to the Ramapo River flow from the surrounding highlands. In New Jersey, these include Stag Brook, which joins the Ramapo River downstream of West Mahwah; Darlington Brook, which joins the Ramapo River near Darlington; Bear Swamp Brook, which joins the Ramapo River 1.2 mi north-northeast of the Oakland Borough-Mahwah Township boundary; Fox Brook, which joins the Ramapo River 0.5 mi north-northeast of the Oakland Borough-Mahwah Township boundary; and Pond Brook, which joins the Ramapo River near Oakland (pl. 1).

The Ramapo River basin lies within the New England and Piedmont provinces of the Appalachian Highlands (Parker and others, 1964, fig. 1). The boundary between the New England and Piedmont physiographic provinces is called the border fault, and its location roughly coincides with that of the Ramapo River in New Jersey and the Mahwah River in New York. The New England province is northwest of this boundary, and in New Jersey consists of the rugged Ramapo Mountains whose peaks commonly are higher than 900 ft above sea level. The Piedmont province to the southeast is marked by less

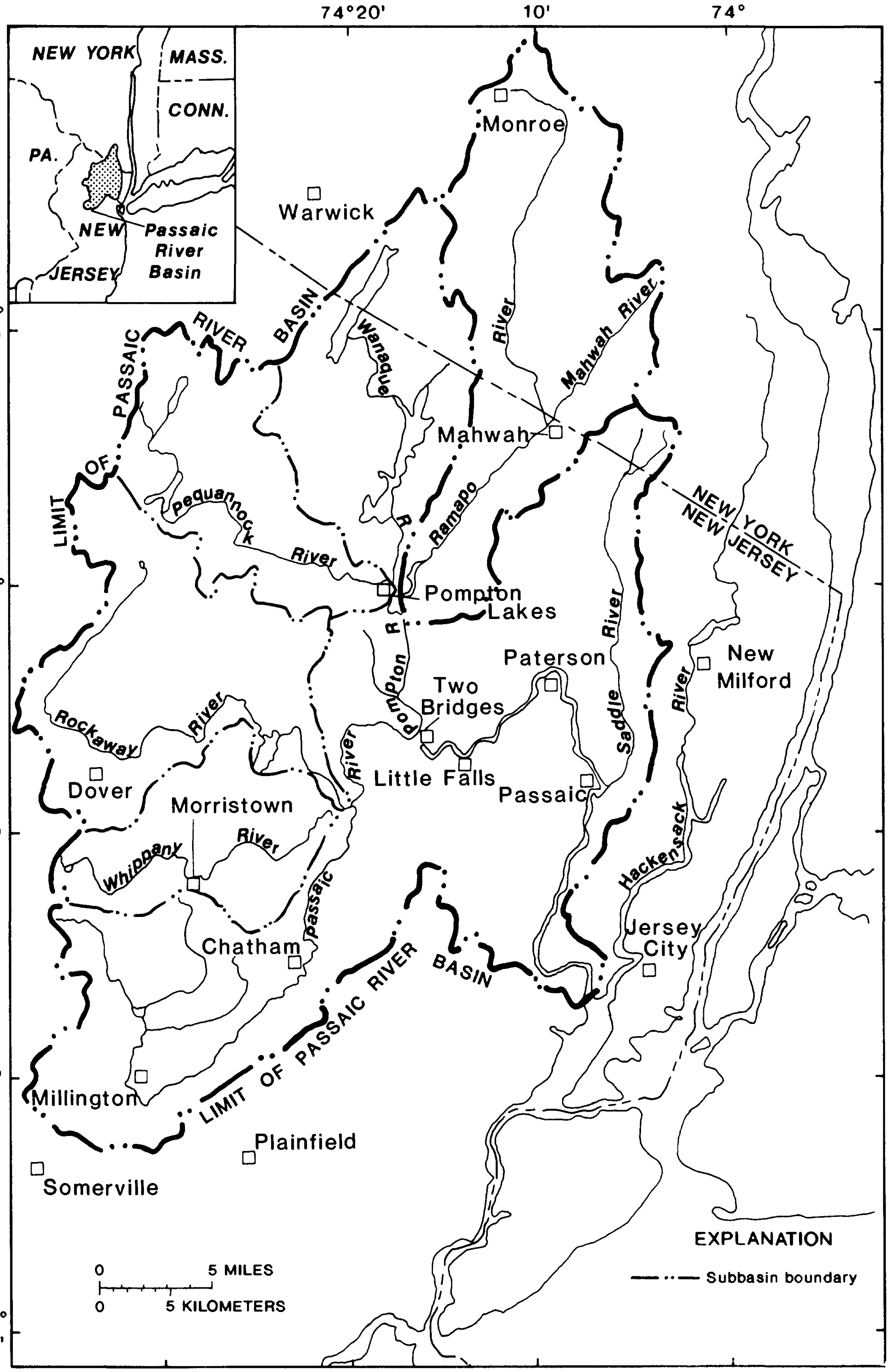


Figure 1.--Location of the Ramapo River basin in the Passaic River drainage system.

rugged hills whose altitudes do not exceed 750 ft. Extensive swampy areas are found at the headwaters of tributary streams in the Piedmont province (Vecchioli and Miller, 1973, p. 6).

Climate

Virtually all water within the Ramapo River valley both above and below the land surface originates as precipitation on the Ramapo River basin. Precipitation is measured at a rain gage near Raymond Dam, which is located 3 mi north of Pompton Lakes Borough on the southeast side of the Wanaque Reservoir (pl. 1). The average annual precipitation for the period 1951-80 was 47.14 in. (inches) (National Oceanic and Atmospheric Administration, 1982). Average monthly precipitation for the period 1951-80 is shown in figure 2. Precipitation occurs relatively uniformly throughout the year; smallest amounts occur in January, February, May, and October, and largest amounts occur in March and August.

Temperature is measured at a station located about 9 miles southeast of Pompton Lakes on the Boonton Reservoir. The average annual temperature for the period 1951-80 was 50.3 °F (degrees Farhrenheit). The lowest mean monthly temperature (18.6 °F) occurred in January; the highest mean monthly temperature (83.5 °F) occurred in August (National Oceanic and Atmospheric Administration, 1982).

On average, about 20 of the 47 in. of average annual precipitation leaves the basin through evapotranspiration (Vecchioli and Miller, 1973, p. 11). Evapotranspiration in the study area varies seasonally. Low temperatures and dormancy of vegetation result in low rates of evapotranspiration from November through April. High temperatures and active growth of vegetation increase rates of evapotranspiration from May through October. Evapotranspiration usually is greatest during the early and middle parts of the growing season.

Land Use

About half of the Ramapo River valley has been developed for industrial, commercial, or medium-density residential use; the other half is sparsely populated residential or public land (Bergen County Planning Board, 1978; Bergen County Planning Board, 1985, p. 96). In the northern part of the valley, development is centered around Mahwah and West Mahwah, New Jersey, and Suffern, New York; the land use is predominantly industrial. In the southern part, development is centered near the Ramapo River in Oakland Borough; the land use is predominantly industrial and medium-density residential. The highlands surrounding the valley are sparsely populated. Most of the highlands west of the Ramapo River in New Jersey and west of the Mahwah River in New York are mountainous and wooded, and are poorly suited for urbanization or major development. The highlands east of the Ramapo and Mahwah Rivers are hilly and mostly wooded; development has been largely residential. Rapid development of the northern Ramapo River valley is anticipated (Bergen County Planning Board, 1985, p. 63), encouraged by the upcoming (as of 1992) completion of Interstate 287, a north-south highway that will traverse Mahwah, New Jersey.

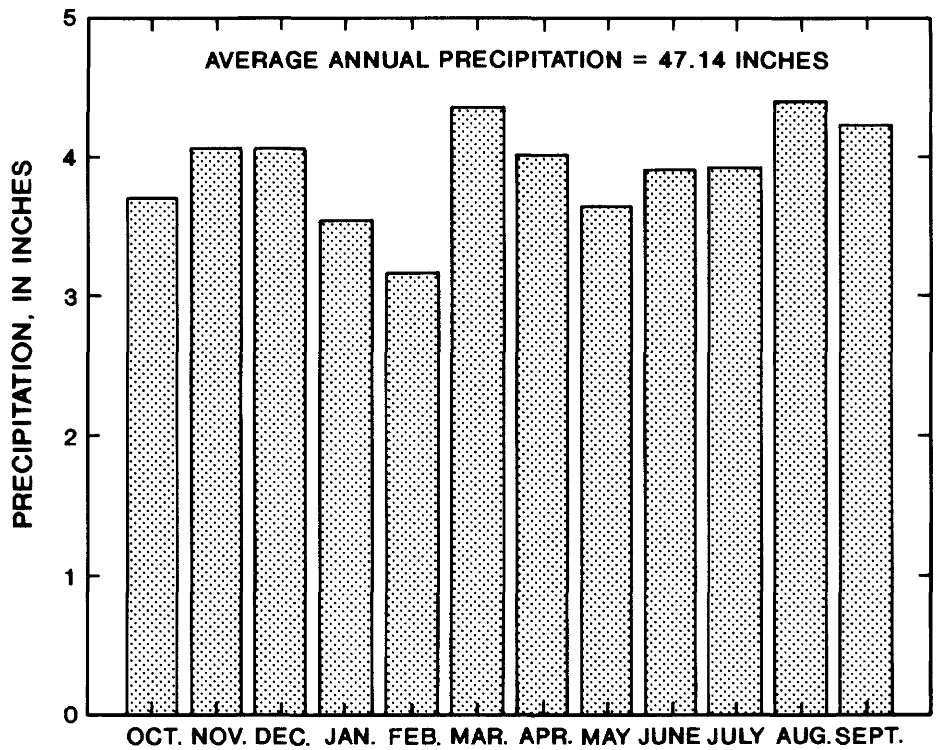


Figure 2.--Average monthly precipitation at Raymond Dam on Wanaque Reservoir, New Jersey, 1951-80.

Geologic Setting

Geologic units, ranging in age from Precambrian to Holocene, crop out in and near the Ramapo River valley. The areal extent of the rock units is shown in figure 3. The older units are rock and include Precambrian gneiss, and Triassic and Jurassic basalt and sedimentary formations of the Newark Supergroup. The younger units, which are unconsolidated deposits, include glacial till and stratified drift. Glacial till generally forms a thin mantle above the consolidated rocks; stratified drift fills some of the large depressions on the bedrock surface.

This study focuses on the stratified drift along the New Jersey part of the Ramapo River valley (fig. 3). This material is called valley-fill deposits throughout the report.

Bedrock

Precambrian Rocks

The Precambrian crystalline rocks, which are composed mostly of gneiss, underlie most of the area west of the Ramapo and Mahwah Rivers (fig. 3). These rocks form part of the Ramapo Mountains.

Newark Supergroup

The Newark Supergroup of Late Triassic and Early Jurassic age (Olsen, 1980, p. 6) underlies the area east of the Ramapo and Mahwah Rivers (fig. 3). These rocks are truncated on the west by a major fault zone where rocks of the Newark Supergroup abut Precambrian crystalline rocks (fig. 3). The Ramapo River follows the trend of the fault zone in New Jersey, and the Mahwah River follows the trend of the fault zone in New York (Ratcliffe, 1980, fig. 1).

The Newark Supergroup is composed of interbedded basalts and sedimentary formations (Olsen, 1980, p. 6). The basalts are much harder and more resistant to erosion than are the sedimentary formations. Near the Ramapo River valley the basalts form the northern extreme of the Watchung Mountains. The sedimentary formations bordering the Ramapo River valley are composed of cemented sandstone and conglomerate with interbedded shale. They form the highlands that border the town of Mahwah on the east and the west, and the low-lying areas between the basalt ridges in the southern part of the basin (pl. 1). The sedimentary formations form higher topographic features in the Ramapo River basin than they do in most other areas of New Jersey, and less weathered bedrock is found at the surface because the upper, weathered layer has been removed by glacial scour.

Unconsolidated Quaternary Deposits

Glacial Till

Glacial till, or unstratified drift, commonly is a mixture of sediments ranging in size from clay to boulders. Glacial till covers the higher altitude areas of the basin, and thin deposits occasionally are present between the consolidated rocks and overlying stratified drift. The till

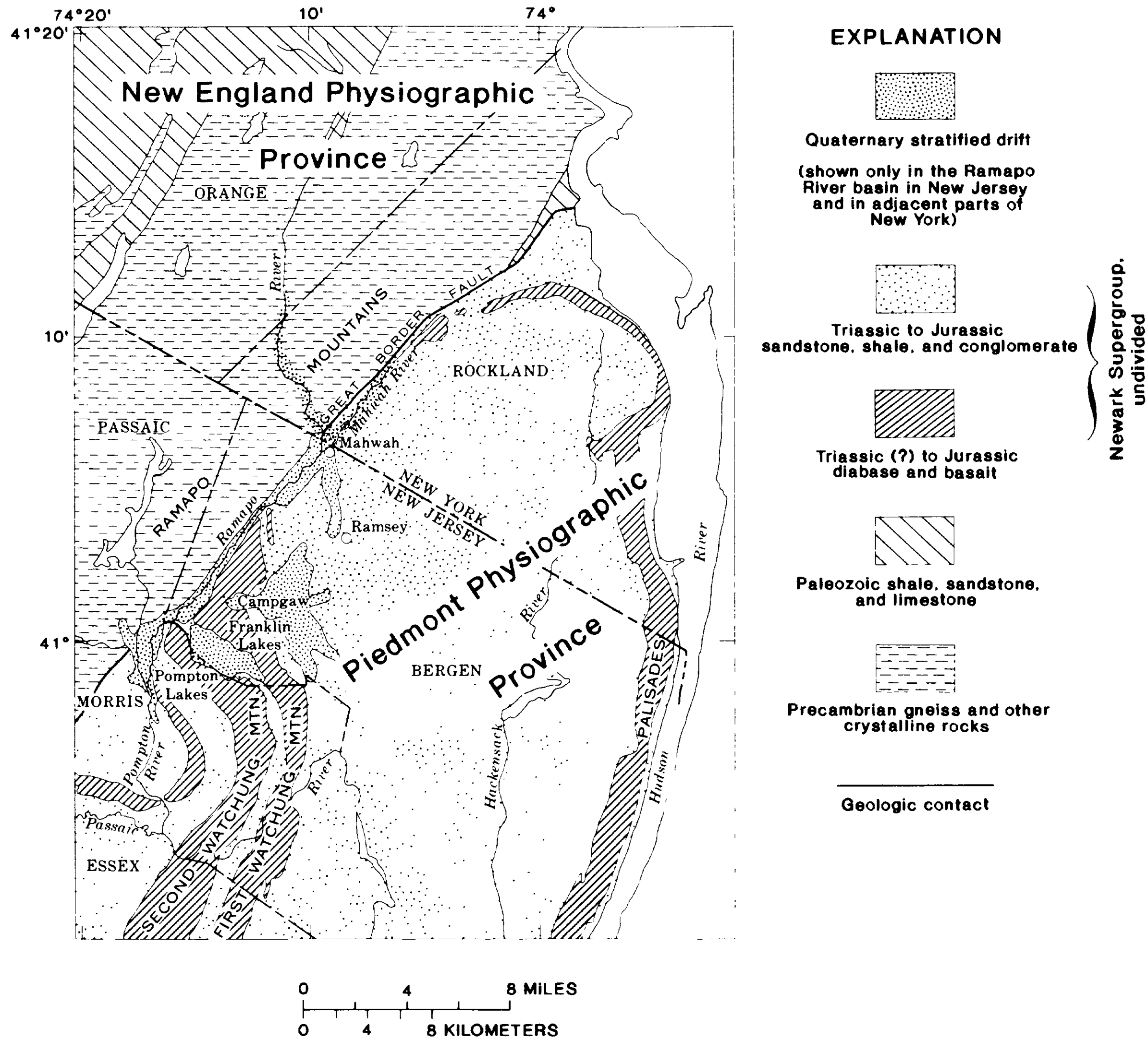


Figure 3.--Generalized geology of northeastern New Jersey and the adjacent part of New York.

cover on the Precambrian rocks typically is thin--a few feet to a few tens of feet thick--and bedrock exposures are numerous, particularly on steep slopes and summits. The till cover on the rocks of the Newark Supergroup is more than 100 feet thick in some places, but more commonly is a few feet to many tens of feet thick, and commonly is thinnest on the crests of the basalt ridges (Vecchioli and Miller, 1973, p. 10).

Stratified Drift

Stratified drift primarily is composed of unconsolidated sand and gravel (outwash or other high-energy deposits) with some silt and clay (glacial-lake deposits), and is found mainly in the valleys of the basin.

The most extensive stratified-drift deposit of the Ramapo River basin is found along the Ramapo and Mahwah Rivers and Masonicus Brook (pl. 1). This deposit is referred to as valley fill in this report. The valley-fill deposits generally are 0.4 to 1.4 mi wide, and are bordered by glacial deltas of stratified drift that range in width from less than 0.25 to nearly 1 mi; the average width is about 0.5 mi (Salisbury, 1902, p. 575). The tops of the deltas are as much as 60 ft above the local altitude of the Ramapo River. The valley-fill deposits along the Ramapo and Mahwah Rivers and Masonicus Brook are up to 200 ft thick, and comprise the most productive aquifer system in the basin.

Other extensive stratified-drift deposits are found in upland areas near Franklin Lakes and Campgaw (fig. 3). Near Franklin Lakes, most of the stratified drift forms a plain between the First and Second Watchung Mountains that extends from the Ramapo River valley on the northwest to about 1 mi east of Franklin Lake. It commonly is 100 ft thick, but locally is as thick as 145 ft. Glacial deltas are found near the eastern border of the plain and north of Franklin Lakes. Near Campgaw, the stratified drift forms glacial deltas, eskers, and irregular ice-contact deposits that are up to 100 ft thick (Vecchioli and Miller, 1973, p. 31).

GEOHYDROLOGY OF VALLEY-FILL DEPOSITS

The depth to bedrock from land surface in the study area, which is equivalent to the thickness of the valley-fill deposits, is shown on plate 2. The depth to bedrock consistently exceeds 100 ft along the Ramapo River. It exceeds 150 ft near Suffern, New York, southwest of Mahwah along line D-D', and at the confluence of Fyke Brook and the Ramapo River; southwest of Crystal Lake it exceeds 200 ft. Along the Mahwah River and Masonicus Brook valleys, depths to bedrock generally are less than 100 ft. The upland stratified drift in the vicinity of Franklin Lakes and Campgaw comprises aquifers that are locally important where the saturated thickness is several tens of feet, but, as a whole, they are of minor regional importance as a source of water (Vecchioli and Miller, 1973, p. 31).

The depth-to-bedrock map shown on plate 2 was developed with data from previous studies, drillers' records from previously installed wells, additional test drilling, and surface-geophysical methods. Three surface-geophysical methods were used. The New Jersey Geological Survey used single-channel seismic-reflection and seismic-refraction methods to obtain point values of depth to bedrock at 165 stations, which were arranged in 14 lines across the valley. At each station a recorder was set up, and seismic impulses were produced at evenly spaced locations along a traverse extending away from the recorder. A sledge hammer and strike plate were used to produce the seismic impulses (Robert Canace and Wayne Hutchinson, New Jersey Department of Environmental Protection and Energy, written commun., 1984). Eighty seismic-reflection stations were positioned in the deeper (greater than 50 ft) parts of the valley. The seismic traverses at these stations were about 50 ft long. Eighty-five seismic-refraction stations were positioned where the depth to bedrock was less than about 60 ft. The seismic traverses at these stations were up to 250 ft long. The data were analyzed by Canace and Hutchinson (New Jersey Department of Environmental Protection and Energy, written commun., 1984). The refraction data probably are accurate to within 10 percent (Haeni, 1984, p. 32); the reflection data probably are less accurate. Geologic sections B-B′, D-D′, and G-G′ were drawn with the data from these seismic traverses (plate 2).

The U.S. Geological Survey used a 12-channel seismic-refraction method along traverses which were either 600 or 1,200 ft long. The seismic sources were explosives positioned at the ends of the seismic lines. Sections drawn from data collected along these traverses are shown in Appendix A. Locations of these traverses are shown on plate 2. The results of these tests probably are accurate to within 10 percent according to Haeni (1984, p. 32).

The relative permeabilities of the valley-fill deposits and bedrock can be characterized by means of the well-yield and specific-capacity measurements shown in table 1. The well yield is the rate at which water was pumped from a well during the well-acceptance test; the specific capacity is the well yield divided by the drawdown observed during the well-acceptance test. Because both well-yield and specific-capacity measurements are affected by the construction of the well, its development, the character of the screen or casing perforation, and the velocity and length of flow up the casing (Lohman and others, 1972, p. 11), a few wells may not represent realistically the hydraulic properties of the material. Meaningful comparisons can be made, however, if many measurements are available from each type of geologic material considered.

The relation between properties of the geologic materials and well yields and specific capacities (reported in table 1) is shown in figure 4. Many of the wells in each of the consolidated units yield 10 gal/min (gallons per minute) or less. Even the sedimentary formations of the Newark Supergroup rocks, which also are aquifers elsewhere in New Jersey, have a median specific capacity of only 0.82 (gal/min)/ft (gallons per minute per foot of drawdown). This value is consistent with those reported by Barksdale and others (1943, p. 149), who found that the estimated sustainable ground-water yield of the sedimentary formations southwest of the Ramapo River basin in Middlesex County is twice as large as the range of 200,000 to 300,000 gallons per day per square mile reported for the Ramapo River basin by Vecchioli and Miller (1973, p. 34). Most wells in the valley-fill deposits yield more than 30 gal/min, and many yield more than 100 gal/min. These data indicate that the valley-fill deposits in the Ramapo River basin are significantly more permeable than the consolidated rocks, and that the consolidated rocks effectively bound ground-water flow in the valley-fill deposits.

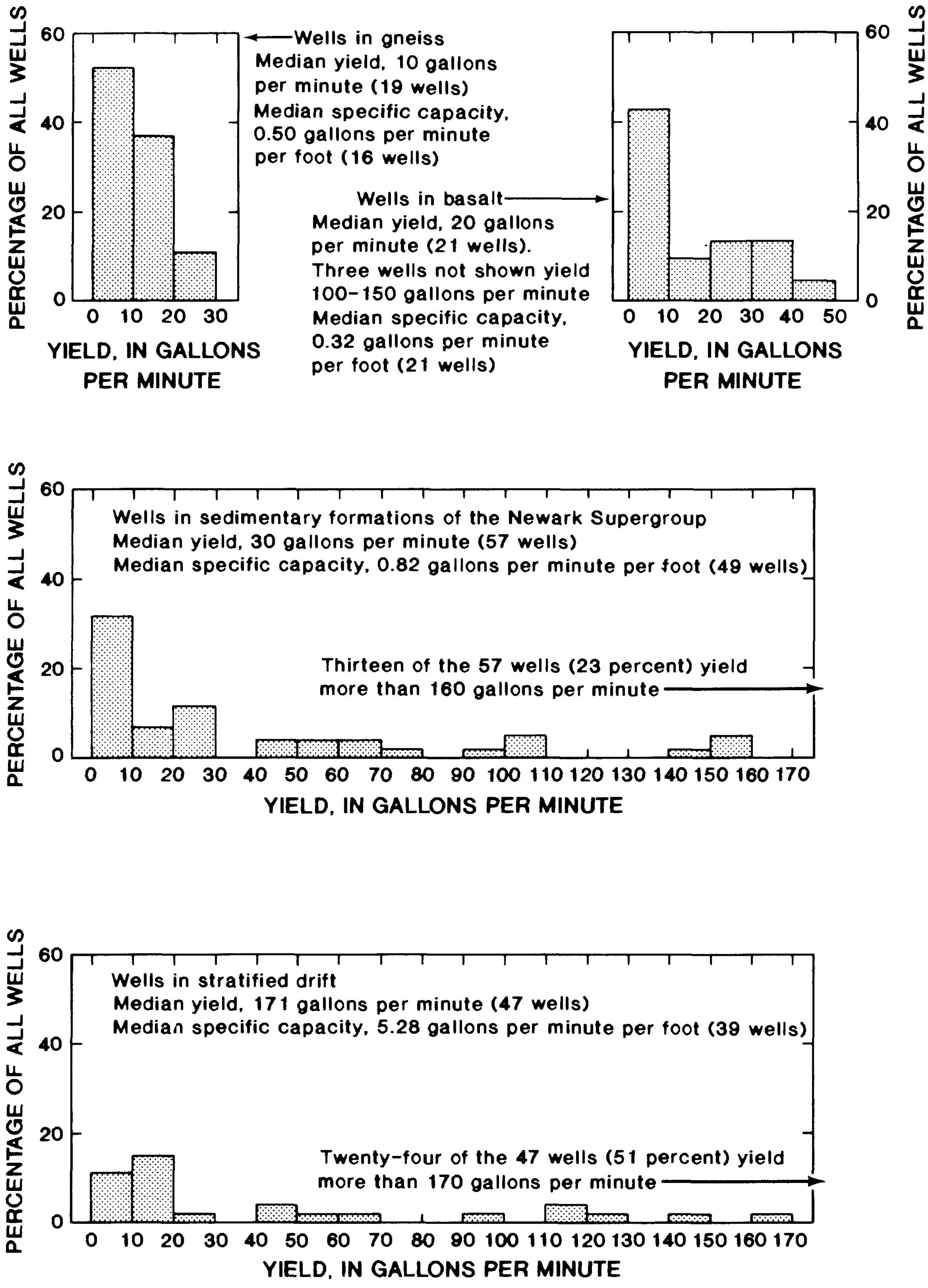


Figure 4.--Yields and specific capacities of wells in the Ramapo River valley, by geologic formation.

Aquifers and Confining Units

Ground water in the valley-fill deposits of the Ramapo River valley is found in aquifers that are composed mostly of sand and gravel layers. Where present, thick, areally extensive silty layers form confining units. Relatively impermeable bedrock bounds the sides and bottom of the valley-fill deposits. Locations of wells, test holes, and potentiometers used to study the valley-fill ground-water system are shown on plate 1; construction features and yield characteristics of the wells, test holes, and deep potentiometers are given in table 1. Construction features of shallow potentiometers used in this study are discussed in the text, and are not included in table 1.

Aquifers and confining units in the valley-fill deposits were defined by using drillers' records from the following sources: New Jersey Department of Environmental Protection and Energy records of materials encountered while drilling public and domestic water-supply wells (generally submitted by local well drillers); U.S. Geological Survey records of materials encountered while drilling test wells and test holes; and New Jersey Department of Transportation records of materials encountered while drilling test holes. The data submitted by local well drillers vary in accuracy and reliability; the other sources are considered to be reliable. Results of an attempt to use borehole-geophysical methods to differentiate between the sand and silt components of the valley-fill deposits were inconclusive.

The valley-fill deposits generally are composed of sand with discontinuous silt layers up to 4 ft thick. A concentration of such layers found near the Suffern well field (fig. 5) probably forms a local confining unit. A more extensive silt layer in the northern part of the study area ranges up to 100 ft thick and is 2 mi long and 0.5 mi wide. This silty layer is both overlain and underlain by saturated sand. Together, the layers form an aquifer system composed of an unconfined aquifer, a confining unit, and a confined aquifer.

Thickness

A thickness map of the silty confining unit is shown in figure 5; a thickness map of the confined aquifer, which rests on bedrock, is shown in figure 6. Thicknesses were approximated on the basis of sparse data and knowledge of the depositional environment.

A geologic section through the valley-fill deposits along line K-K' (figs. 5 and 6) is shown in figure 7. This section shows the relative extents and thicknesses of the layers. The major confining unit thickens to the southwest, culminating in a bulbous shape above the sharp depression in the bedrock surface. It pinches out southwest of Darlington Brook, near where the valley narrows (pl. 2).

Hydraulic Properties

Typical horizontal hydraulic conductivities in valley-fill deposits of glacial origin range from 1 to 13,000 ft/d (feet per day) for aquifer materials, and from 1×10^{-4} to 1 ft/d for confining units. The specific

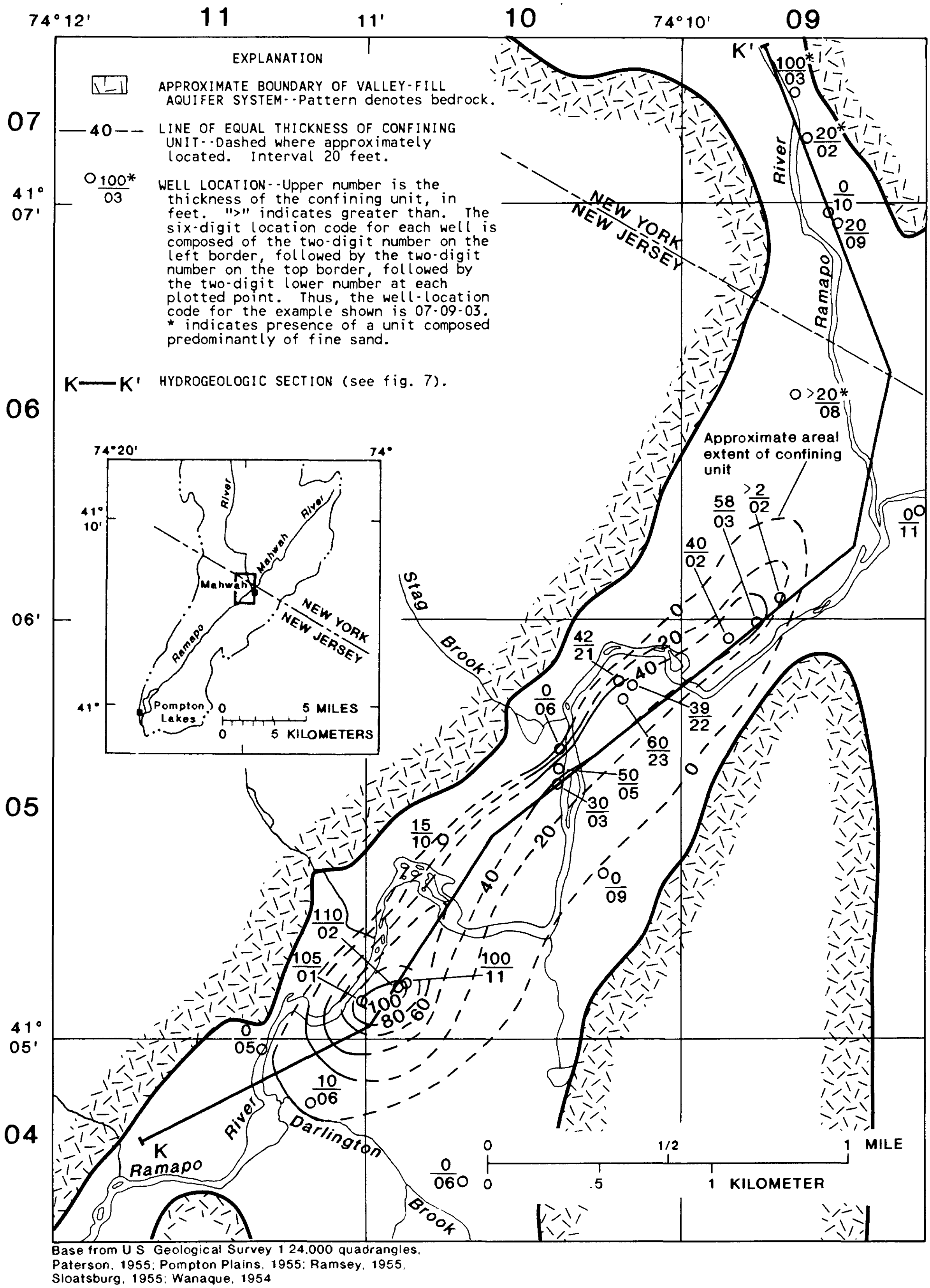


Figure 5.--Thickness of the confining unit in the valley-fill deposits near Mahwah, New Jersey.

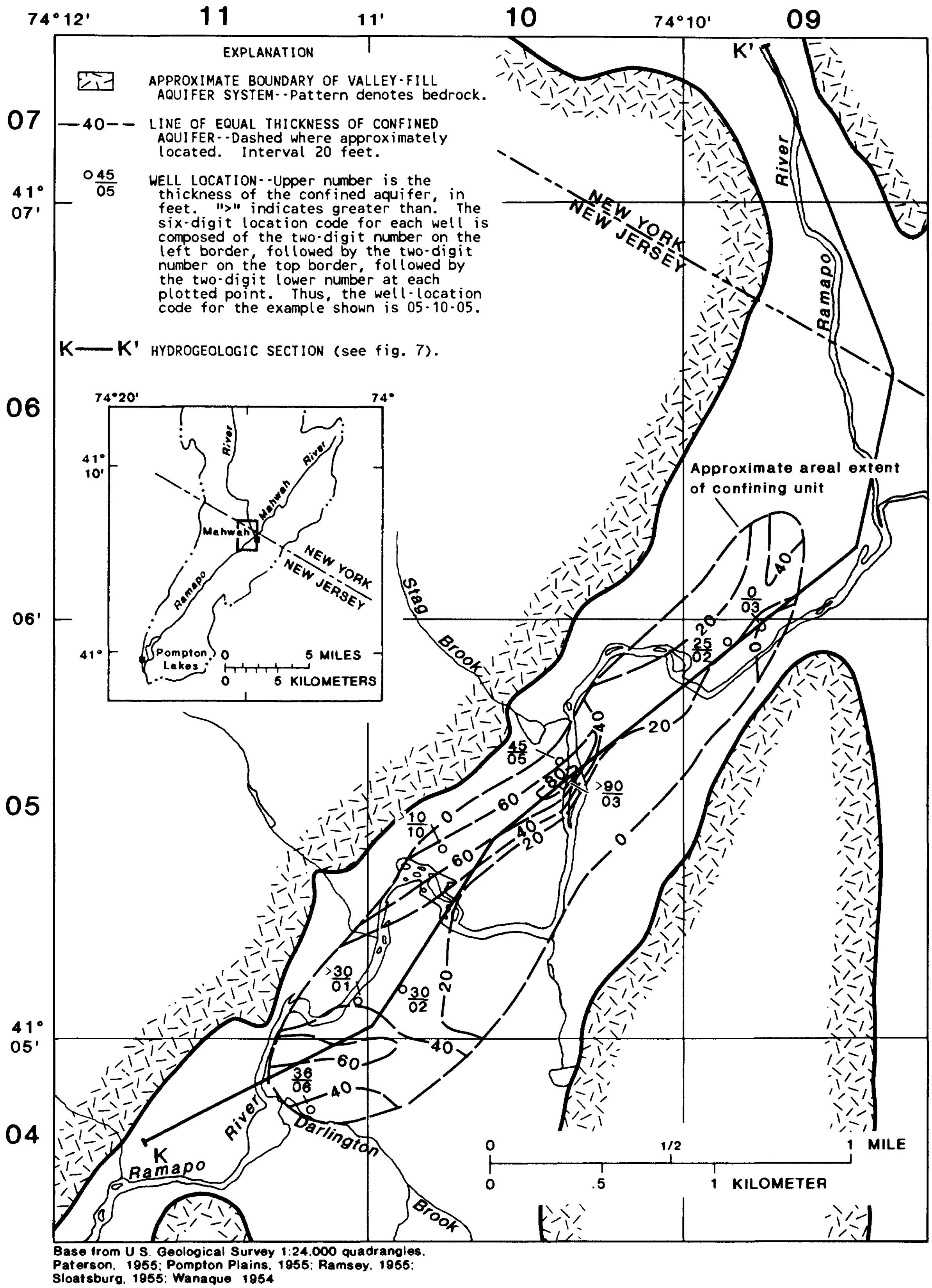
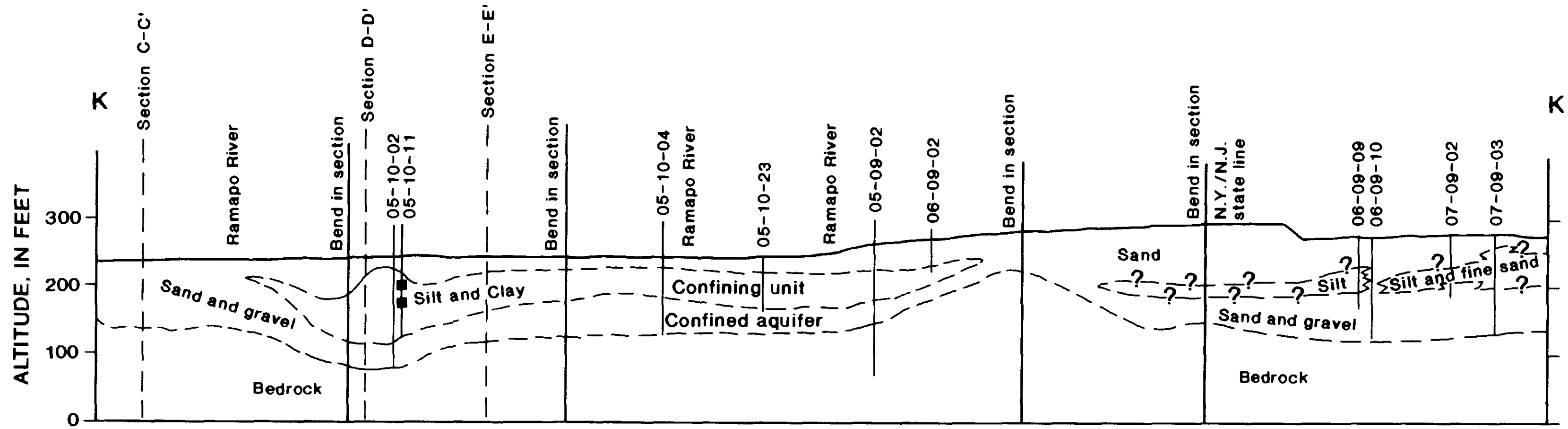


Figure 6.--Thickness of the confined aquifer in the valley-fill deposits near Mahwah, New Jersey.



EXPLANATION

05-10-02 Location code of well listed in table 1. Wells have been projected 150 feet or less onto line of section, except well 05-10-23 which is 500 feet northwest of the line of section

Location of sample analyzed by permeameter test

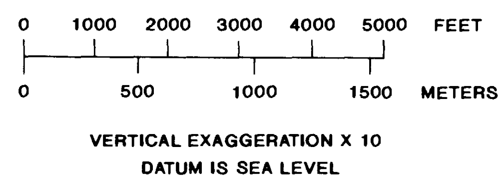


Figure 7.--Hydrogeologic section through the Ramapo River valley near Suffern, New York, and West Mahwah, New Jersey.

yield of aquifer materials normally ranges from 0.15 to 0.30; the storage coefficient for confined aquifers normally ranges from 10^{-4} to 10^{-2} (Lyford and others, 1984, table 1, p. 12).

Two additional reports include estimates of the hydraulic conductivity of valley-fill material in the Ramapo River valley. Leggette, Brashears and Graham (1981, figs. 5 and 6) reported horizontal hydraulic conductivities that range from 13 ft/d to 660 ft/d and storage coefficients that range from 1×10^{-4} to 3×10^{-1} along a 2-mi length of the Ramapo River valley upstream from the Suffern Water Department well field. These values are from a calibrated model of the valley-fill ground-water system and provide a broad range of estimated aquifer characteristics in the area. Vecchioli and Miller (1973, p. 52) reported the hydraulic conductivities of three samples of loose, repacked aquifer material from Oakland, New Jersey, determined with a permeameter in the laboratory by Weston and Sampson (1924, p. 70). The hydraulic conductivities of two samples, one composed of well-sorted medium sand and the other composed of well-sorted coarse sand, were 350 ft/d (2,600 gallons per day per square foot) and 3,300 ft/d (25,000 gallons per day per square foot), respectively. The hydraulic conductivity of a sample composed of poorly sorted gravel was between these two extremes.

As part of the current study, hydraulic properties of the confining unit and confined aquifer (figs. 5, 6, and 7) were estimated by means of permeameter and aquifer tests. Results of two permeameter tests indicate that the vertical hydraulic conductivity of the confining unit is 3×10^{-4} ft/d. Analysis of aquifer tests in the confined aquifer produced estimates of horizontal hydraulic conductivity of 600 and 650 ft/d, estimates of transmissivity of 15,100 and 16,300 ft^2/d (feet squared per day), and estimates of storage coefficient of 1.1×10^{-4} and 1.4×10^{-4}. The analysis indicated that, although recharge through the confining unit is negligible, a recharge boundary is present. This recharge probably is derived from the overlying water-table aquifer. Because the areal extent of the confining unit is limited, water probably flows into the confined aquifer around the edges of the confining unit. The permeameter and aquifer tests are described in detail below.

<u>Permeameter tests</u>

On the basis of permeameter tests conducted on two samples of confining-unit material from the Mahwah, New Jersey, area, the vertical hydraulic conductivity of each is 3×10^{-4} ft/d. The samples were collected during the drilling of well 05-10-11 (table 1, figs. 6 and 7) with 2-1/2-ft-long, 3-in.-inside diameter Shelby[1] tubes. The tested samples, which were about 5 in. long and 3 in. in diameter, were taken from the larger samples; the tested samples were from depths of 43 and 68 ft below land surface, respectively (fig. 7). The approximate layering of the two samples, which were composed of very fine sand to silt, is shown in figure 8. Triaxial pressure

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[1] Use of trade, product, or firm names in this report is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey.

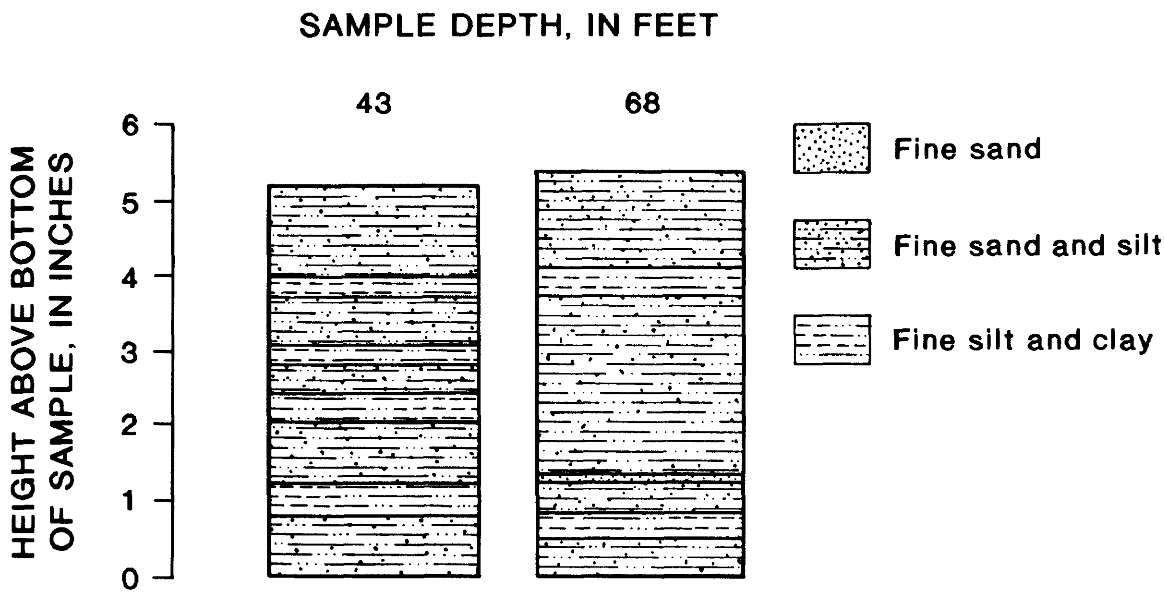


Figure 8.--Generalized distribution of particle sizes in samples of confining-unit material at well 05-10-11.

equivalent to the estimated field overburden pressure was exerted on each sample while the permeameter tests were conducted (R.L. Ladd, Woodward-Clyde Consultants, written commun., 1981). The tests were performed by Woodward-Clyde Consultants of Clifton, New Jersey, in October 1983.

The measurement of identical vertical hydraulic conductivities for both samples indicates that the measured value may be higher than the actual value, and suggests that some other controlling factor, such as leakage between the core and the membrane used to enclose the core during the test, may have occurred. However, the measured value is not unreasonably high and, therefore, is useful as an upper limit of the actual value.

Aquifer tests

Three aquifer tests were conducted in the confined aquifer southwest of Mahwah (figs. 5 and 7). Drawdown data from two of three tests were used to estimate the transmissivity, horizontal hydraulic conductivity, and storage coefficient of the aquifer. Data from the third aquifer test were not used for this purpose because of variations in pumpage. Drawdown data from all three aquifer tests were used to identify the aquifer's hydrologic boundaries. The wells used in the tests--Mahwah Township Water Department wells 16 and 17 and test well 17 (location codes 05-11-01, 05-10-20, and 05-10-02, respectively)--are screened in the confined aquifer. Hereafter, these wells are referred to as production wells 16 and 17 and test well 17. Construction features and yield characteristics of these wells are given in table 1; their locations are shown on plate 1.

Table 2 lists the wells that were included in the three aquifer tests, the function of each well in each test, and the date, duration, and pumping rate of each test. Figure 9 shows the locations of the wells in the well field.

Selected data from the three aquifer tests are listed in Appendix B and are plotted in figure 10. The data from aquifer test 2 have been corrected by as much as 0.2 ft to account for continued recovery from the first aquifer test; this correction was necessary because aquifer test 2 began about 3 hours after the end of aquifer test 1.

The Theis type curves that were matched to the aquifer-test data also are shown in figure 10. Early-time data from aquifer tests 2 and 3 closely matched the type curves. The poor fit between the type curves and the data from aquifer test 1 probably is the result of variations in the rate of pumping. The pumping rate initially was about 503 gal/min, then increased irregularly through the first 12 minutes of the test, and remained at 560 gal/min for the remainder of the test. Although the data from aquifer test 1 were not used to estimate hydraulic properties for this reason, late data from this 72-hour test can be used to identify the boundaries of the confined aquifer.

Table 2.--Wells used in three aquifer tests at one of the Mahwah Township Water Department well fields

[Pumping rates in gallons per minute indicated in parentheses; distances between wells are shown in figure 9]

Mahwah Township Water Department well name and location code	Aquifer-test number and date		
	1 November 8-11, 1982	2 November 11, 1982	3 November 15, 1982
Production well 16 (05-11-01)	measured	measured	pumped (620)
Production well 17 (05-10-20)	pumped (503/560)	pumped (703)	measured
Test well 17 (05-10-02)	measured	measured	measured
Length of test, in hours	72	0.9	5.8

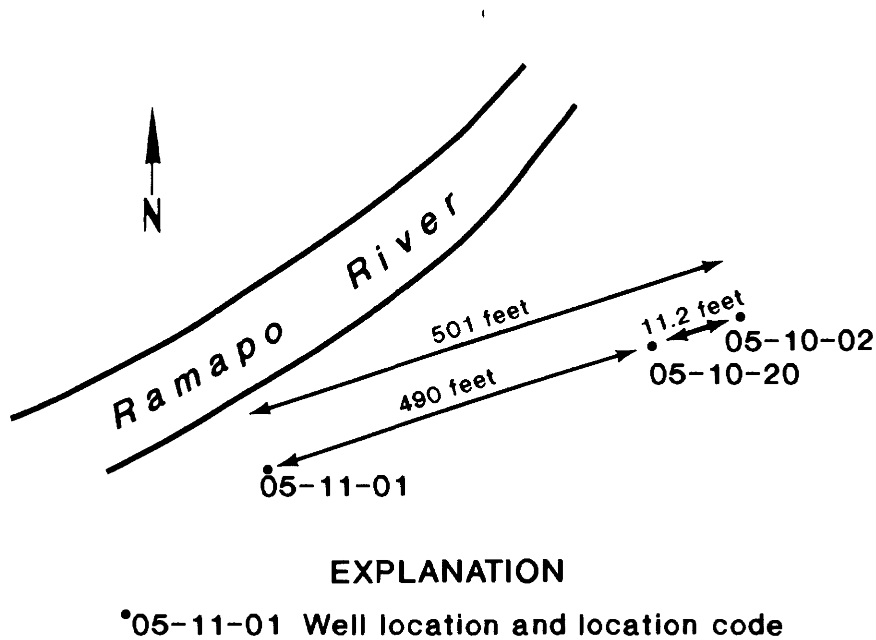
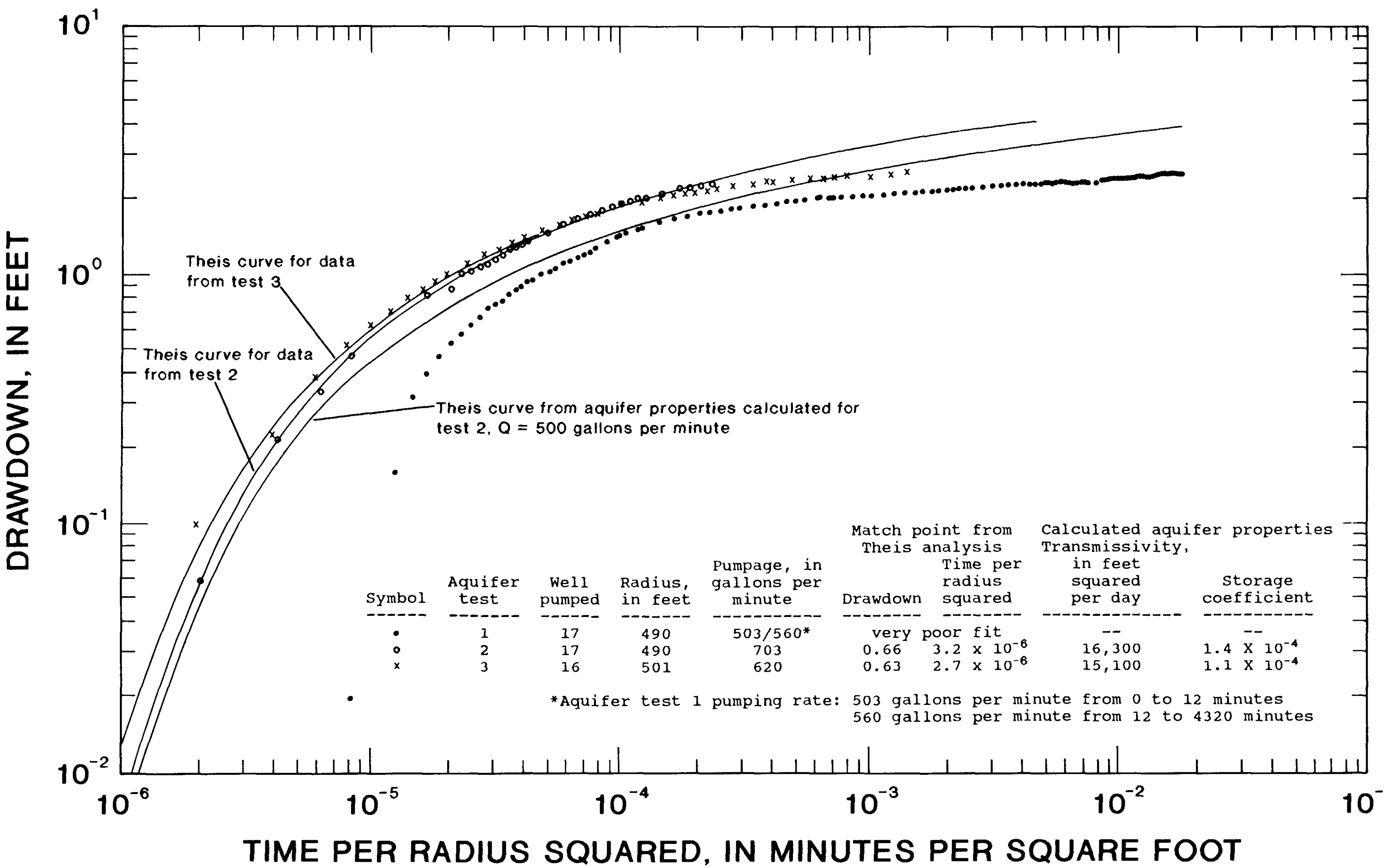


Figure 9.--Distances between wells used in three aquifer tests at Mahwah Township Water Department wells 16 and 17.



					Match point from Theis analysis		Calculated aquifer properties	
Symbol	Aquifer test	Well pumped	Radius, in feet	Pumpage, in gallons per minute	Drawdown	Time per radius squared	Transmissivity, in feet squared per day	Storage coefficient
•	1	17	490	503/560*	very poor fit		--	--
o	2	17	490	703	0.66	3.2×10^{-6}	16,300	1.4×10^{-4}
x	3	16	501	620	0.63	2.7×10^{-6}	15,100	1.1×10^{-4}

*Aquifer test 1 pumping rate: 503 gallons per minute from 0 to 12 minutes
560 gallons per minute from 12 to 4320 minutes

Figure 10.--Theis analysis of three aquifer tests at Mahwah Township Water Department wells 16 and 17.

Drawdowns measured in test well 17 during aquifer tests 1 and 2 differ significantly from the Theis type curve. The finite diameter and partial penetration of the pumped well, 11.2 ft from well 17, probably affected the measured drawdowns at test well 17. Attempts to correct the drawdowns according to the methods of Reed (1980, p. 8-13 and 37-44) produced improved, but still poor, matches with the Theis curve; therefore, these drawdown data are not included in this report.

Theis analysis of the recovery measured in production well 16 after aquifer test 1 and in test well 17 after aquifer test 3 yielded estimates of transmissivity that were as much as 20 percent greater than estimates calculated from the drawdown data. Although the recovery data did not otherwise contradict the drawdown data, they are not presented here.

Transmissivity and storage-coefficient values calculated using the Theis curve-matching technique (Theis, 1935; Reed, 1980, p. 5) were 16,300 ft^2/d and 1.4×10^{-4}, respectively, for aquifer test 2; and 15,100 ft^2/d and 1.1×10^{-4}, respectively, for aquifer test 3 (fig. 10). Average values of aquifer transmissivity and storage coefficient are, therefore, 15,700 ft^2/d and 1.3×10^{-4}, respectively. Aquifer thickness reported in drillers' logs at the three wells used in the test were between 22 and 32 ft. Assuming an average thickness of 25 ft, the horizontal hydraulic conductivities calculated from aquifer tests 2 and 3 are 650 ft/d and 600 ft/d, respectively; the specific-storage values calculated from aquifer tests 2 and 3 are 5.6×10^{-6}/ft and 4.4×10^{-6}/ft, respectively.

The fit between the Theis type curve and observed drawdowns was good for almost all of aquifer test 2 and for the first 25 min (minutes) of aquifer test 3. At later times during both tests, measured drawdowns were less than those predicted by the Theis curve, and the difference increased steadily with time. Late in aquifer test 3, the measured drawdown was nearly constant; the late data from aquifer test 1 suggest that drawdown probably would have remained constant during a much longer test. The observed system differs from the ideal system considered by Theis (1935) in several ways: (1) The confining unit is not impermeable and contains stored water, so that leakage into the confined aquifer from above may occur; (2) the confined aquifer is hydraulically connected to the water-table aquifer and, indirectly, to surface-water bodies at the edges of the confining unit (figs. 5, 6, and 7); and (3) the confined aquifer does not have uniform thickness. The first and second factors would provide the confined aquifer with additional sources of water that could account for the departure of the measured drawdown data from the ideal response; these factors are examined below. Because it is unlikely that the third factor could account for the departure of the measured drawdown data from the ideal response, it is not considered further.

An attempt was made to fit the drawdown data from aquifer tests 2 and 3 to the modified leaky-aquifer curves of Hantush (1960). Although it is likely that some of the deviation between the measured drawdown data and the Theis type curve is caused by recharge from storage in the confining unit, the modified leaky-aquifer curves do not match the data well. The modified leaky-aquifer curves continue to show an increase in drawdown with time; late in the tests, however, the measured drawdowns were nearly constant.

The most likely explanation for the flattening of the drawdown curve late in the tests (fig. 10) is the hydraulic connection between the confined and water-table aquifers at the edges of the confining unit (fig. 6). Downward flow could occur anywhere along the edges of the confining unit, but would tend to occur mainly at the edges closest to the pumped well.

Hydrology

Surface Water

Daily streamflow records are available from three U.S. Geological Survey gaging stations on the Ramapo River in or near the study area. Streamflow at the station at Suffern, New York (01387420), has been measured since 1979; streamflow near Mahwah, New Jersey (01387500), has been measured since 1922; and streamflow at Pompton Lakes, New Jersey (01388000), has been measured since 1921. For water year 1983 (October 1, 1982 through September 30, 1983), mean annual flows at these stations were 220, 297, and 386 ft^3/s (cubic feet per second), respectively; minimum daily flows were 7.3, 13, and 18 ft^3/s, respectively. Station locations are shown on plate 4.

Streamflow at the station at Pompton Lakes (01388000) is affected by nearby water users. During 1921-39, streamflow was routed past the gage through a nearby power plant. Since 1953, water has been diverted from the Ramapo River just north of the station at Pompton Lakes into the Wanaque Reservoir. In calendar year 1982, when seepage runs were conducted as part of the present study, water was diverted from January through March at an average rate of 123 ft^3/s. No diversions occurred during the remainder of 1982. Therefore, the seepage runs conducted on May 18 and October 13, 1982, were not affected by diversions to Wanaque Reservoir.

Seasonal Variations in Streamflow

Mean monthly streamflows in the Ramapo River near Mahwah during water years 1923-83 are compared with the maximum and minimum monthly mean streamflows in figure 11. Water year 1923, for example, began on October 1, 1922, and ended on September 30, 1923. Although a seasonal trend is clear in all three curves, the minimum monthly mean streamflows vary most. The seasonal streamflow patterns shown in figure 11 do not directly reflect the precipitation record (fig. 2) because the rainfall-runoff relation is affected by evapotranspiration and infiltration.

Low-Flow Frequency

Low-flow-frequency data for the Ramapo River near Mahwah (station 01387500) are shown in figure 12. The smooth curves in figure 12 are Pearson type III distributions fitted to the logarithms of the low-flow discharges (Riggs, 1968; Riggs, 1972). These curves represent the magnitude and frequency of the lowest flow each year for the indicated number of consecutive days for climatic years 1924-83. Climatic year 1924, for example, began on April 1, 1923, and ended on March 31, 1924.

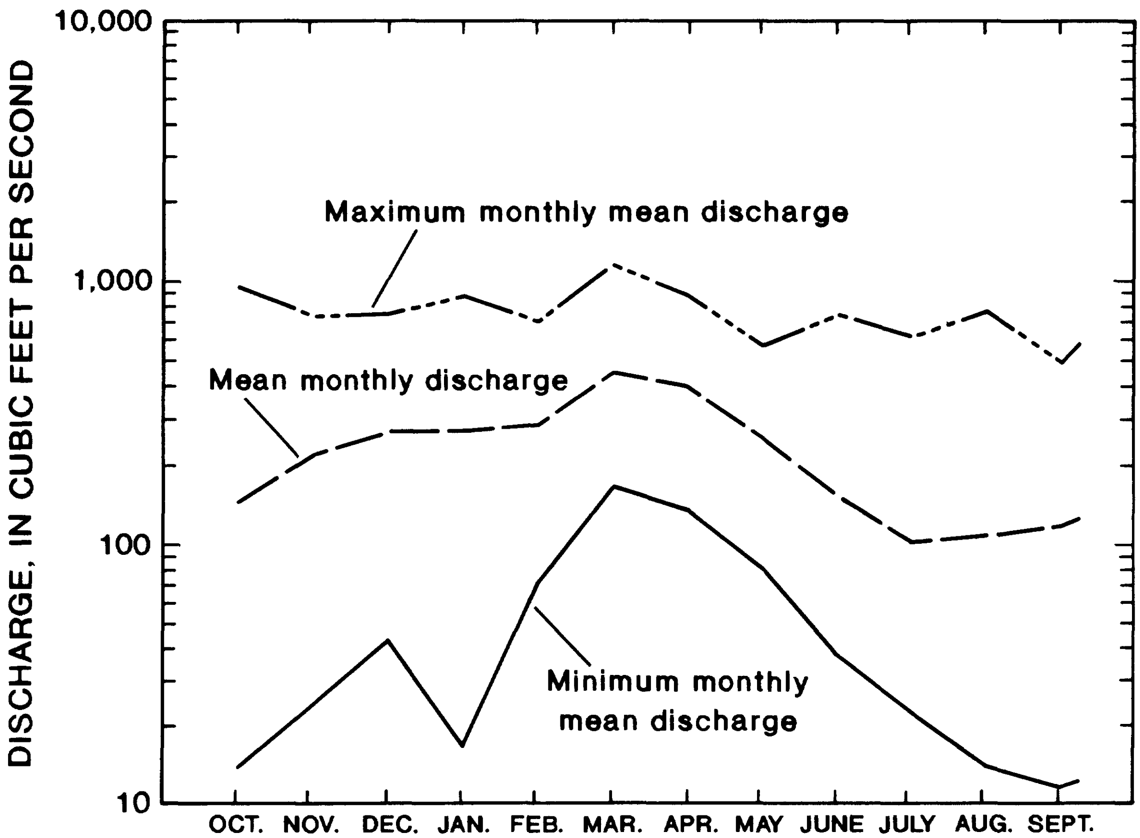


Figure 11.--Mean monthly discharge and maximum and minimum monthly mean discharge for Ramapo River near Mahwah, New Jersey (station 01387500), water years (October 1 to September 30) 1923-83.

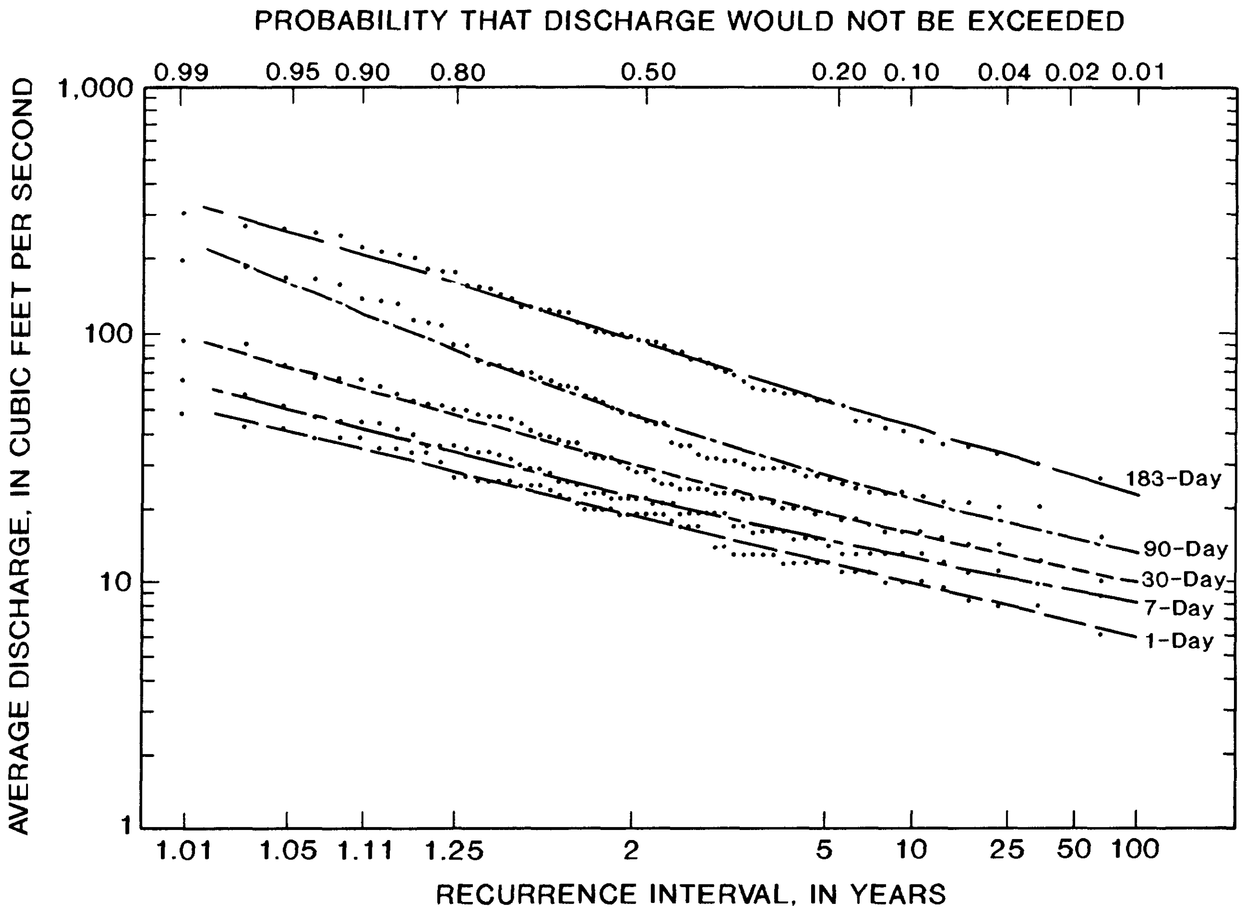


Figure 12.--Low-flow frequency curves for Ramapo River near Mahwah, New Jersey (station 01387500), climatic years 1924-83. (A climatic year extends from April 1 through March 31 and is designated by the calendar year in which it ends.)

The importance of individual low-flow frequency curves depends on the type of water use considered. For example, minimum average discharges for periods of 1 or 7 consecutive days are most important if continuous supplies are required. In the context of the ground-water/surface-water interactions considered in this study, a duration that corresponds to the length of the dry season (typically several months) probably is most applicable.

Minimum Streamflow Requirements

Minimum flow requirements were established at two locations along the Ramapo River to protect downstream water users. Since 1976, the New York State Department of Environmental Conservation has permitted the Spring Valley Water Company to pump water from its Ramapo Valley well field only when flow in the Ramapo River at Suffern, New York (station 01387420, pl. 4), is at least 12 ft^3/s (New York State Department of Environmental Conservation, 1982, p. II-3). The well field is adjacent to the Ramapo River, less than 1 mi from the New York-New Jersey State line. The New Jersey Department of Environmental Protection and Energy allows[2] the Hackensack Water Company and the North Jersey District Water Supply Commission together to divert as much as 4,650 million gallons of water per month (an average of 239 ft^3/s) from the Ramapo River to the Wanaque Reservoir, except when flow in the Ramapo River at Pompton Lakes, New Jersey (station 01388000), is less than 62 ft^3/s (Robert Canace, New Jersey Department of Environmental Protection and Energy, oral commun., 1986).

Streamflow at the Suffern and Pompton Lakes stations frequently falls below the indicated values. During water years 1978-83, streamflow in the Ramapo River at Pompton Lakes (station 01388000) was less than 62 ft^3/s during 77 to 270 days per year. During water years 1980-83, streamflow in the Ramapo River at Suffern, New York (station 01387420), was less than 12 ft^3/s during 7 to 74 days a year. Despite the regulations, restrictions at both locations have been modified to allow pumpage or diversions at lower river flows during drought (Vlado Michna, New Jersey Department of Environmental Protection and Energy, oral commun., 1986; New York State Department of Environmental Conservation, 1982, p. II-3).

Ground Water

Ground water in the saturated valley-fill sediments of the Ramapo River valley generally is under water-table conditions. The major exception is ground water in the confined aquifer shown in figures 6 and 7. Major sources of recharge to the ground-water system are infiltration of precipitation, infiltration of runoff from adjacent highlands, and seepage from streams (Vecchioli and Miller, 1973, p. 56). After flowing through the ground-water system, water eventually discharges to the Ramapo River or its tributaries, or is intercepted by wells.

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[2] Diversions are allowed under New Jersey Department of Environmental Protection and Energy Water Supply permits W.S.-1685 and W.S.-1651, issued in 1982.

Levels

Water levels were measured on October 13, 1982, at 44 sites in the valley-fill deposits in the study area, and were measured at 6 additional sites at other times. Water levels also were measured monthly from July 1982 through December 1983 at 24 of these sites. Locations of the wells and potentiometers are shown on plates 1 and 3; water-level measurements are shown on plate 3.

The 50 measured sites include 27 shallow potentiometers and 5 deeper observation wells installed as part of this project, and 18 private and municipal wells. The shallow potentiometers were installed by the drive-point method and were constructed of 1.25-in. (inch) nominal-inside-diameter steel pipe with drive-point screens 1 to 3 ft long. All of these shallow potentiometers were set with the top of the screen 3 to 8 ft below the water table; total depths ranged from 6 to 32 ft. All potentiometers were developed on completion to ensure their connection to the aquifer. Well numbers used by the U.S. Geological Survey in New Jersey (NJWRD) and New York (NYWRD) and location codes for the shallow potentiometers are as follows:

NJWRD well number	Location code	NJWRD well number	Location code	NYWRD well number	Location code
03- 95	01-15-06	03-113	04-11-10	410649074092001	06-09-13
03- 96	01-15-07	03-114	04-10-07	410714074081601	07-08-01
03- 97	01-15-08	03-115	05-11-03		
03-101	03-13-10	03-117	05-11-02		
03- 99	03-13-11	03-124	05-10-15		
03-105	03-13-12	03-125	05-10-16		
03-103	03-12-13	03-127	05-10-17		
03-102	03-12-14	03-126	05-10-18		
03-107	03-12-15	03-129	05-09-04		
03-108	04-12-04	03-131	05-09-05		
03-110	04-11-07	03-130	05-08-04		
03-109	04-11-08	03-128	06-09-12		
03-112	04-11-09				

The five deeper observation wells, which are called deep potentiometers in this report, were constructed of 1-in. polyvinyl chloride (PVC) pipe with the bottom 1 ft slotted and the bottom end capped. They were installed by augering 3-in.-diameter holes, placing the PVC pipe in the hole, and then back-filling the annulus between the pipe and hole with material removed from the hole. The tops of the screened zones ranged from 24 to 65 ft below the land surface. Construction features of the deep potentiometers are presented in table 1, in which these potentiometers are called "USGS test wells."

Seventeen of the 18 private and municipal wells used for water-level measurements were constructed with steel casing ranging from 6 to 12 in. nominal inside diameter, and were screened in sand and gravel at depths ranging between 40 and 190 ft. The exception was a dug well. Detailed construction features and yield characteristics of the wells are included in table 1.

Altitudes of the measuring points of potentiometers and wells were determined by using standard surveying methods. Water-level measurements shown on plate 3 are accurate to within 0.1 ft.

Flow

The predominant direction of ground-water flow in the valley-fill deposits is parallel to the long axes of the valleys in the study area, except where tributaries from the highlands recharge the valley-fill aquifer from the sides (pl. 3). The downvalley hydraulic gradient in the water-table aquifer of the Ramapo River valley is 0.0014 ft/ft (foot per foot) on average, and is fairly constant. From the New York State line to 0.3 mi south of Fox Brook, a head loss of 41 ft was measured over a distance of 5.3 mi, resulting in an average hydraulic gradient along this length of the valley in the northern part of the study area of 7.7 ft/mi (feet per mile), or 0.0015 ft/ft. From 0.3 mi south of Fox Brook to just south of Crystal Lake, a head loss of 13 ft was measured over a distance of about 2 mi, resulting in an overall hydraulic gradient in the southern part of the study area of 6.5 ft/mi, or 0.0012 ft/ft. Slightly greater downvalley gradients were measured near Stag and Fox Brooks (0.0022 ft/ft and 0.0026 ft/ft, respectively); much steeper downvalley gradients were measured in the valley-fill deposits along Masonicus Brook in New Jersey and along the Mahwah and Ramapo Rivers in New York.

Steep lateral gradients were measured near the edges of the valley-fill deposits, where tributaries enter the valley from the surrounding highlands. These steep lateral gradients, shown on plate 3, are associated with steep vertical gradients and substantial downward flow into the aquifer. In July 1984, for example, a creek flowing southwestward from the small lake at the base of Stag Brook (pl. 3) lost all its flow to the aquifer before reaching the Ramapo River.

Significant downward gradients were measured in the center of the valley at a few locations, as shown in the inset on plate 3. In the area of the confining unit, the observed downward gradients resulted from pumpage from the confined aquifer. The reason for the observed downward gradient in well 03-12-08 (near the southwest end of the inset, pl. 3) is unknown, because no confining unit is present in the area. Residual drawdown from pumpage is unlikely at this well because it is not located near known pumping wells; variations in hydraulic conductivity that could produce the low head at depth also seem unlikely.

Withdrawals

Ground-water withdrawals from the valley-fill deposits in the New Jersey part of the study area increased from less than 1 Mgal/d (million gallons per day) in 1950 to more than 7 Mgal/d in the early 1970's, and were fairly constant from the early 1970's to the early 1980's (E.F. Vowinkel, U.S. Geological Survey, written commun., 1982). Withdrawals from the New York part of the study area were about 1 Mgal/d in 1969, and increased to more than 11 Mgal/d by the early 1980's. Most of the ground water withdrawn from the valley-fill deposits is pumped from well fields in Spring Valley and Suffern, New York; at one well field and two individual wells in Mahwah Township, New Jersey; and at two well fields (Soons and Bush) and one individual well in Oakland Borough, New Jersey. Some of these wells are listed in table 1 and located on plate 1. The well-location codes of the most productive of these wells are listed below:

Municipality	Well-location code shown in table 1
Spring Valley, New York	07-09-01 through 07-09-06
Suffern, New York	06-09-09; 06-09-10; 06-09-14
Mahwah Township, New Jersey	
Ford field	05-10-03; 05-10-06 through 05-10-08
Well 16	05-11-01
Well 17	05-10-20
Oakland Borough, New Jersey	
Soons field	03-13-01 through 03-13-03
Bush field	01-15-04; 01-15-05
Well 9 (Route 208)	01-13-01

Mean annual withdrawals during 1969-83 at the New York and New Jersey well fields are shown in figure 13. Mahwah Township well 16 started production in 1982. Yearly withdrawals from well 16 during 1982 and 1983 averaged 0.31 and 0.47 Mgal/d, respectively. Withdrawals at Mahwah Township well 17 began in November 1986 (Patrick Malone, Township of Mahwah, Department of Public Works, oral commun., 1990). Withdrawals at Oakland Borough well 9 on Route 208, which has been active since the late 1960's, have increased to 0.5 Mgal/d.

Water levels in some wells may have been affected by pumping from nearby production wells. The average withdrawals from production wells in the study area for the period October 9-13, 1982, are shown on plate 3. Withdrawal values were nearly constant over the 5-day period in all cases. Drawdowns caused by pumping are evident in the water levels measured at the Suffern Water Department well field and the Mahwah Township Ford well field.

Ground-Water/Surface-Water Interactions

Sustained high yields from wells in the sand and gravel deposits of the Ramapo River valley are possible because the permeable valley-fill aquifer is recharged by induced seepage from the Ramapo River. The relation between ground-water withdrawals and induced seepage is important because heavy withdrawals can reduce streamflow below minimum flow requirements during drought. Vecchioli and Miller (1973, p. 57-58) showed that the valley-fill aquifer is hydraulically connected to the Ramapo River and its tributaries.

Leggette, Brashears and Graham, Inc. (1982, appendix B, p. 11), measured large changes in streamflow caused by ground-water withdrawals during an aquifer test at the Spring Valley well field, New York, in the valley fill just north of the study area. A total of 8.5 Mgal/d was pumped from six

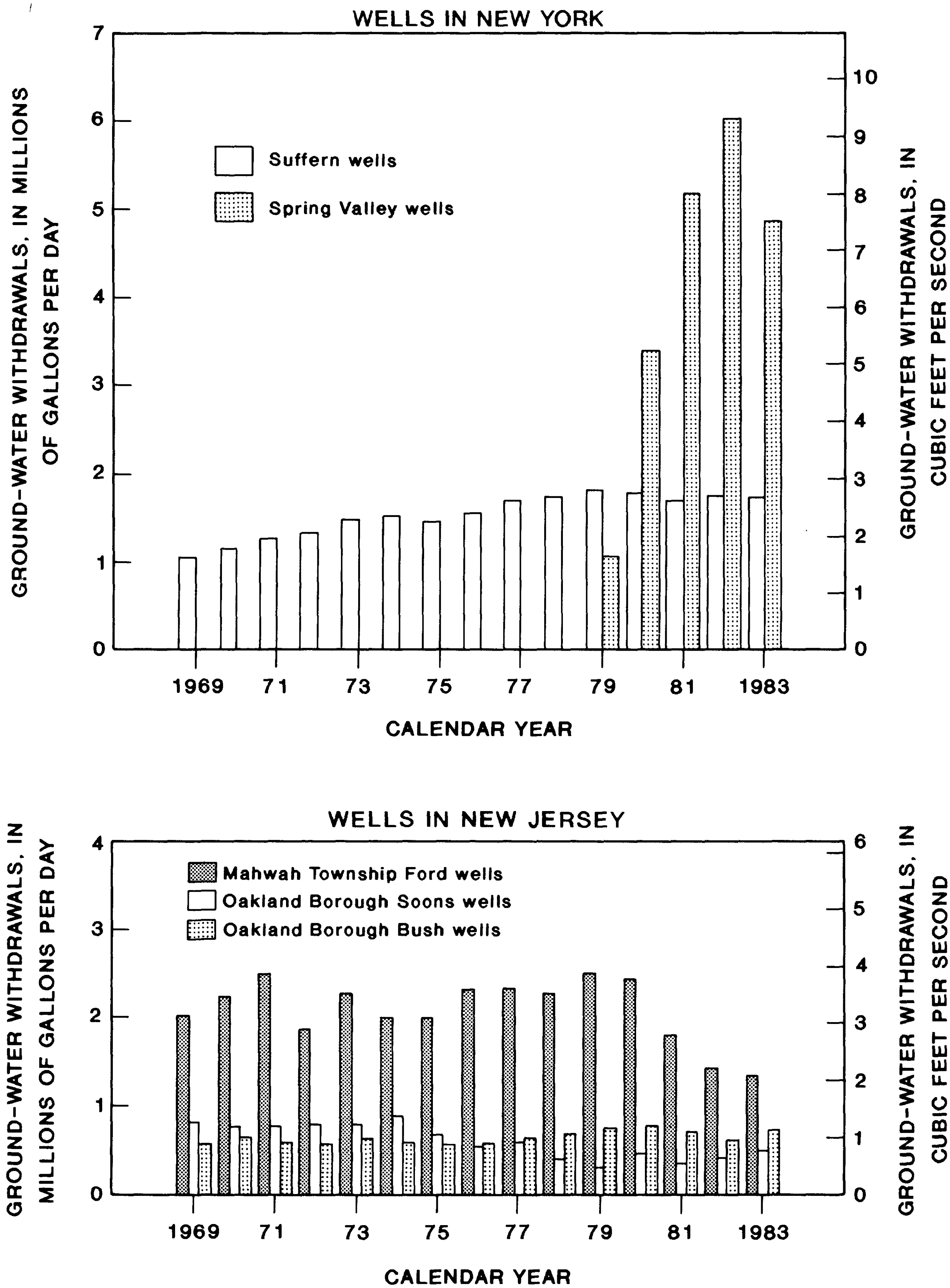


Figure 13.--Average yearly ground-water withdrawals from the valley-fill deposits at major well fields in the New York and New Jersey parts of the Ramapo River valley, 1969-83.

wells along a 1.5-mi reach of the Ramapo River. After only 30 hrs, 60 percent of the water withdrawn was accounted for by measured changes in the flow of the Ramapo River.

In this study, seepage runs were conducted to measure the flow between the river and the aquifer system. A seepage run consists of a set of discharge measurements taken along the river under base-flow conditions over a short period of time, generally less than a day. Base-flow conditions occur when precipitation directly on the river and overland flow are insignificant, so that changes in river flows result only from ground-water/surface-water interactions. Discharge measurements made during seepage runs can be used to identify river reaches that are either "losing" (water flows from the stream to the aquifer) or "gaining" (water flows from the aquifer to the stream), depending on whether stream discharge decreases or increases in the downstream direction. Even under ideal conditions, however, the error in discharge measurements made by using the 0.6-depth rule and 25 sections generally exceeds 3.8 percent in about one-third of the measurements (Carter and Anderson, 1963, p. 113). Errors can be greater where velocities are less than 0.2 ft/s (foot per second) or depths are less than 0.5 ft, as they are at some sites measured for this study (R.D. Schopp, U.S. Geological Survey, oral commun., 1987). For this study, measurement error was assumed to be 5 percent. This will be called the potential measurement error in this report. If a calculated gain or loss is of the same magnitude as the potential measurement error, it is difficult to conclude whether a stream reach is gaining or losing water. For this reason, a calculated gain or loss is significant only if it equals or exceeds 5 percent of the larger of the two streamflow measurements used in the calculation.

Three comprehensive seepage runs and two smaller scale seepage runs were conducted. In the comprehensive seepage runs, streamflow was measured at sites distributed throughout the New Jersey part of the Ramapo River valley; the smaller scale seepage runs, conducted along losing reaches that appeared to be affected by induced recharge from nearby production wells, provided additional detail on streamflow gains and losses along these reaches.

Comprehensive seepage runs

Results of the three comprehensive seepage runs indicate that losing and gaining reaches alternate along the main stem of the Ramapo River. The discharge data collected during these runs are tabulated on plate 4. Adjusted main-stem discharge values were calculated by subtracting cumulative tributary inflows from measured total discharge values. Ground-water inflow or outflow (gains or losses) along each reach between measurement sites was evaluated by subtracting successive adjusted main-stem discharge values. Losses and gains in streamflow along each reach of the main stem of the Ramapo River are compared with ground-water withdrawals in table 3.

The streamflow gains and losses measured during this study confirm results reported by Vecchioli and Miller (1973, p. 66), who showed that much of the seepage loss along the main stem of the Ramapo River probably results from ground-water withdrawals along the channel, which induce recharge from the river. The increased number of measurements made along the main stem

Table 3.--Adjusted stream-discharge measurements, gains and losses, and ground-water withdrawals along the main stem of the Ramapo River during base-flow conditions

[Flow values in cubic feet per second; all values rounded to the nearest tenth; see plate 4 for actual measured values; dashes indicate discharge or value not determined]

Station number	Adjusted main-stem discharge[2] 10-14-81	Adjusted main-stem discharge[2] 5-18-82	Adjusted main-stem discharge[2] 10-13-82	Gain or loss (-) of base-flow from previous station 10-14-81	Gain or loss (-) of base-flow from previous station 5-18-82	Gain or loss (-) of base-flow from previous station 10-13-82	Ground-water withdrawals[4] 10-14-81	Ground-water withdrawals[4] 5-18-82	Ground-water withdrawals[4] 10-13-82
[Combined flow at sites 0138742503 and 0138749205]	14.7	72.7	23.3						
[1]01387500	16.0	67.0	24.0	1.3	-5.7	[3]0.7			
[1]01387525	16.0	72.9	26.2	[3]0	5.9	2.2			
								Ford well field	
							2.9	3.7	[5]2.5
01387540	14.8	75.4	24.0	-1.2	[3] 2.5	-2.2			
[1]01387610	13.6	72.9	20.9	-1.2	[3]-2.5	-3.1			
01387660	--	75.2	23.7	--	[3] 2.3	2.8			
[1]01387670	10.0	73.9	23.4	-3.6	[3]-1.3	[3]-.3			
01387765	--	82.7	28.0	--	8.8	4.6			
								Soons well field	
							0.2	1.6	0.4
01387769	--	--	24.6	--	--	-3.4			
01387811	--	88.3	25.8	--	5.6	[3]1.2			
								Bush well field	
							1.3	1.2	1.0
[1]01387910	12.8	82.6	21.9	2.8	-5.7	-3.9			
		Total		-1.9	9.9	-1.4	4.4	6.5	3.9

[1] Stations reported in Vecchioli and Miller (1973, pl. 4).
[2] These values were calculated from the stream-discharge values in the table on plate 4 by deducting cumulative tributary flows and rounding to the nearest tenth.
[3] These values are less than 5 percent of the larger of the two streamflow values used to calculate a gain or loss and are not considered significant.
[4] Average of ground-water withdrawals on the day of seepage run and the 3 days prior to seepage run.
[5] Includes 0.7 cubic feet per second from Mahwah Township Water Department well 16 (location code 05-11-01, table 1 and plate 1).

during this study permits additional analysis. For example, losses between stations 01387540 or 01387525 and 01387610 (table 3) are too large to be the result of ground-water withdrawals alone. These losses also are affected by variations in hydrogeologic characteristics of the ground-water system. Seepage measurements made on October 13, 1982, show that the losing reaches are just downstream from areas where the transmissivity of the water-table aquifer increases abruptly, indicating that a local decline in head resulting from an increase in the water-transmitting capacity of the aquifer system probably causes local variations in inflow to and outflow from the river. Between stations 01387540 and 01387610, the transmissivity of the water-table aquifer increases as the confining unit pinches out 0.4 mi upstream from station 01387610. Under low base-flow conditions, the losing reach extends upstream toward station 01387525 (see seepage runs of October 14, 1981, and October 13, 1982). No data are available for horizontal or vertical hydraulic conductivity just upstream from station 01387765 and the Oakland Soons well field.

Seepage runs conducted on October 14, 1981, and May 18, 1982, generally are consistent with results of the October 13, 1982, seepage run.

Local seepage runs

Two small scale seepage runs were carried out on September 15-16, 1983, to (1) verify the results of the comprehensive seepage runs, (2) provide additional detail on stream-aquifer interactions in two areas where streamflow losses exceeded local pumpage, and (3) estimate the vertical hydraulic conductivity of the streambed. The two reaches considered were (1) the reach between stations 01387525 and 01387540, adjacent to the Mahwah Township Water Department Ford well field and just upstream from Mahwah Township well 16; and (2) the reach between stations 01387670 and 01387769, adjacent to the Oakland Borough Water Department Soons well field (pl. 4).

Ford well field.--Stream discharges were measured on September 15-16, 1983, at four stations along the Ramapo River near the Mahwah Township Water Department Ford well field and well 16. Figure 14 shows the station locations, the discharge values, and the locations of nearby production wells, among other information. The Ford well field consists of four production wells in the confined aquifer aligned nearly parallel to and on the western side of the river. The wells are screened from 65 to 140 ft below land surface (table 1). Mean daily rate of withdrawal from the Ford well field varied from 2.6 to 3.7 ft^3/s and averaged 3.0 ft^3/s during the 30-day period before September 15. The mean rate of withdrawal during the 2 days before, and on the days of, the seepage run was 2.9 ft^3/s. Production well 16 is located adjacent to the river 0.8 mi southwest of the Ford field and is screened in the confined aquifer from 116 to 149 ft below land surface. Mean daily rates of withdrawal from well 16 varied from 0.33 to 1.53 ft^3/s and averaged 0.95 ft^3/s during the 30-day period before September 15. The mean rate of withdrawal during the 2 days before, and on the days of, the seepage run was 0.82 ft^3/s.

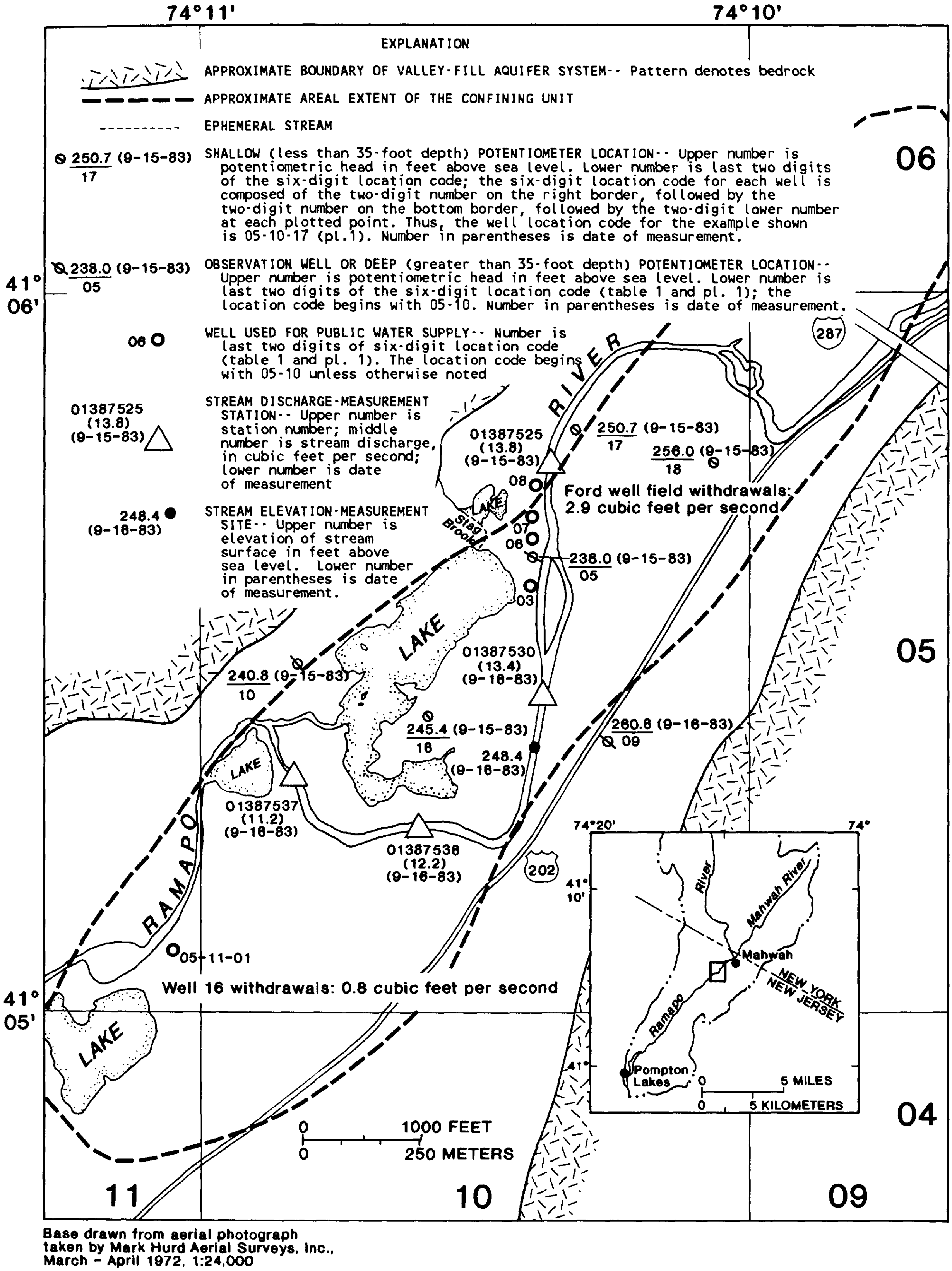


Figure 14.--Streamflow and ground-water levels near the Mahwah Township Water Department Ford well field and well 16, September 15-16, 1983.

Streamflow declined steadily along the measured reach--from 13.8 ft^3/s at station 01387525 to 13.4 ft^3/s and 12.2 ft^3/s at stations 01387530 and 01387536, respectively, and to 11.2 ft^3/s at station 01387537. The total measured change in streamflow was -2.6 ft^3/s, of which 2.2 ft^3/s was lost between stations 01387530 and 01387537.

Streamflows were measured over 2 days, as indicated on figure 14. Streamflow at station 01387530 was measured on both days to assess the change in flow through time. The flow declined from 13.8 ft^3/s on September 15, 1983, to 13.4 ft^3/s the following day. This 2.9-percent decline was less than the potential measurement error of 5 percent of the streamflow measurement and thus was not considered to be significant.

The absence of exact spatial correlation between the ground-water withdrawals and decreasing streamflow shown in figure 14 indicates that streamflow losses probably are not derived from stream reaches nearest the pumped wells. Only 0.4 of the 2.6-ft^3/s loss between stations 01387525 and 01387537 was measured over the reach nearest the source of pumping, and other sources of recharge to the wells are nearby. Two lakes, Stag Brook, and an ephemeral stream, all of which are tributary to the main stem of the Ramapo River, are much closer to the Ford wells than are the main-stem reaches between stations 01387530 and 01387537 (fig. 14). Although the withdrawals are derived from the river, streamflow gains or losses may be measured a considerable distance downstream from the source of the withdrawals, where the tributaries enter the main stem. The results of the local seepage run near the Ford well field support the results of the comprehensive seepage runs along this part of the Ramapo River (table 3). Results of both runs support the theory that observed streamflow losses caused by withdrawals need not necessarily occur in the stream reach nearest the source of ground-water withdrawal, but can result from local differences in hydrogeologic characteristics within the ground-water system. These variations include (1) changes in water-transmitting properties of the valley-fill deposits, which can be caused by changes in hydraulic conductivity, aquifer thickness, or valley width; (2) the discontinuity of confining units, which can shift the effects of pumping wells screened below the confining unit to an area of increased hydraulic connection with the stream (where the confining unit is absent); and (3) natural flow patterns caused by irregular channel topography, which can result in streamflow gains along some reaches but result in losses along other stream reaches.

Streambed vertical hydraulic conductivity was calculated by using the results of the seepage run in conjunction with Darcy's Law in the following form:

$$K_{rb} = \Delta Q \ / \ (L \ b \ \Delta h / \Delta l),$$

where

K_{rb} is the vertical hydraulic conductivity of the streambed,
ΔQ is the loss of streamflow over the reach,
L is the length of the reach,
b is the width of the reach,
Δh is the hydraulic head at the top of the streambed (equal to the elevation of the water in the river) minus the hydraulic head at some distance below the streambed, and

Δl is the distance below the streambed at which hydraulic head is measured.

Streambed vertical hydraulic conductivity was calculated by use of the 2.2 ft^3/s loss measured between stations 01387530 and 01387537. The head in the streambed was measured at station 01387536 at two depths. At Δl equal to 2.10 ft, Δh was equal to 0.05 ft; at Δl equal to 2.90 ft, Δh was equal to 0.09 ft. Given a reach length of 3,300 ft and width of 70 ft, streambed vertical hydraulic conductivities calculated on the basis of head losses measured at Δl, equal to 2.10 and 2.90 ft, were 35 ft/d and 26 ft/d, respectively.

Soons well field.--Stream discharges along the Ramapo River near the Soons well field were measured at four stations on September 15, 1983. Figure 15 shows the location of measured points and discharge values, and the location of local production wells. The Soons well field consists of three production wells aligned perpendicular to the river. Mean daily rate of withdrawal from the Soons well field varied from 0.73 to 4.68 ft^3/s and averaged 3.1 ft^3/s during the 30-day period preceding the seepage run. The mean daily rate of withdrawal during the 3 days before, and on the day of, the seepage run was 2.33 ft^3/s; mean rates of withdrawal on these 4 days were 4.68, 1.80, 2.11, and 0.73 ft^3/s.

In both the small scale seepage run on September 15, 1983, and the comprehensive seepage run on October 13, 1982, gains were measured in the Ramapo River upstream from the Soons well field, and losses that exceeded local withdrawals were measured in the reaches adjacent to the Soons well field. Figure 14 shows that stream-discharge measurements made on September 15, 1983, increased 2.7 ft^3/s from station 01387670 to 01387765 and decreased 2.9 ft^3/s from station 01387765 to 01387769; the loss measured for the downstream reach (01387765 to 01387769) on October 13, 1982, was 3.4 ft^3/s (table 3). Local ground-water withdrawals alone cannot account for the large losses in streamflow measured adjacent to the Soons well field; stream discharge probably also is affected by variations in the hydrogeologic characteristics of the ground-water system.

Streambed vertical hydraulic conductivity was calculated by use of the 1.7 ft^3/s loss measured between stations 01387765 and 01387767 (fig. 15). The head in the streambed was measured halfway between the locations of the two discharge measurements at two depths (Δl) in the streambed. At Δl equal to 1.76 ft, Δh was equal to 0.10 ft; at Δl equal to 2.26 ft, Δh was equal to 0.13 ft. Given a reach length of 1,180 ft and an average width of 88 ft, a streambed vertical hydraulic conductivity calculated by using the head losses measured at Δl equal to 1.76 and 2.26 ft was 25 ft/d.

SIMULATION OF GROUND-WATER FLOW

A numerical model of the northern part of the Ramapo River valley-fill aquifer system was constructed to quantify the hydrogeologic characteristics of the ground-water system and to evaluate the hydrologic relations between ground-water withdrawals and streamflow. The southern part of the study area was not included in the modeled area.

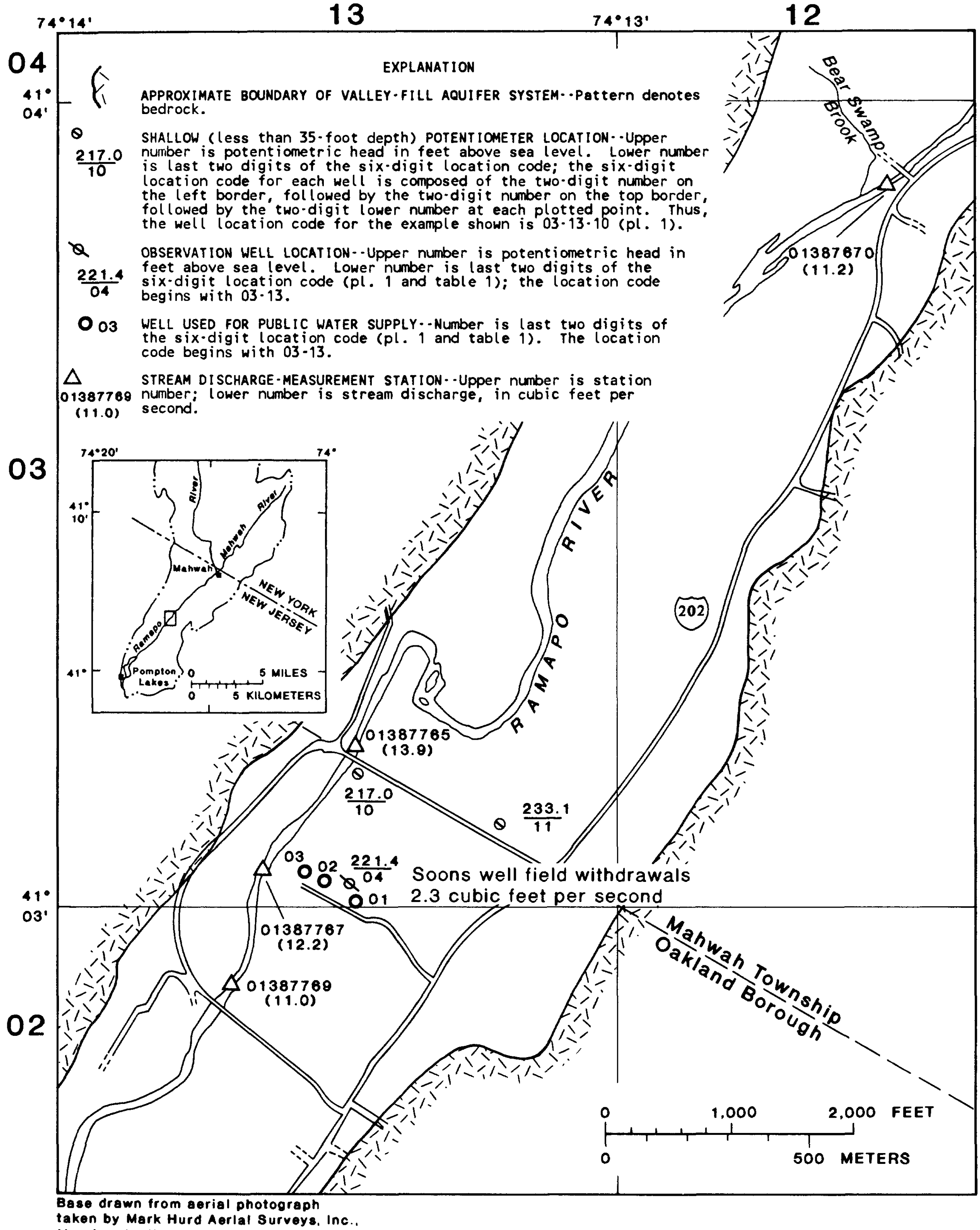


Figure 15.--Streamflow and ground-water levels near the Oakland Borough Water Department Soons well field, September 15, 1983.

The numerical model is based on a differential equation that describes the physics of water flowing through unconsolidated deposits. The differential equation relates water levels to aquifer shape and thickness, to hydraulic properties such as hydraulic conductivity and storage coefficient, and to stresses such as pumpage and areal recharge. For complex system geometry with multiple layers and ground-water/surface-water interactions, numerical methods are used to solve the differential equation. In the resulting numerical model, water levels are calculated only at discrete points, each of which represents a part of the aquifer system. All properties of the system, including aquifer thicknesses, hydraulic properties, and stresses, are defined by values specified at the discrete points. The process of representing a continuous system with a limited number of discrete points is called discretization.

The finite-difference numerical method used in this study has been discussed extensively in the literature (for example, Wang and Anderson, 1982). In the finite-difference method, the discrete points are located along rows and columns, and each point is associated with a rectangular area or cell. The finite-difference grid is formed by the "checkerboard" of rectangular cells. In order to represent an aquifer, the spacing between rows or columns of grid points is adjusted so that the grid is coordinated with aquifer-boundary locations and the smallest cells are in the areas of primary interest. "Active" cells represent part of the aquifer and are assigned aquifer properties such as hydraulic conductivity and storage coefficient; "inactive" cells are outside the simulated system, and are ignored in the simulations. The multilayered aquifer system is represented by using layered finite-difference grids. The finite-difference model used in this study, and the differential equation that describes the physics of water flowing through unconsolidated deposits on which the model is based, is described by McDonald and Harbaugh (1984). (An updated version of the model is found in McDonald and Harbaugh (1988).)

The numerical model of the Ramapo River valley-fill aquifer system was developed and used in the following way. First, a conceptual model of the system was developed using the previously discussed hydrogeology and hydrology of the study area. The available data were used to identify aspects of the system to include in the model, and to develop hypotheses about the flow of water through the system. Second, the numerical model was built based on the conceptual model of the system. Third, the model was calibrated--that is, model inputs (such as hydraulic conductivity and streambed vertical hydraulic conductivity) were adjusted within reasonable limits so that simulated water levels and flows closely matched measured values. Fourth, a sensitivity analysis was conducted to identify which aspects of the system were estimated most accurately, and which most significantly influenced the simulation results.

Model Design and Input Data

Design and Grid

The valley-fill aquifer system is simulated as a three-dimensional ground-water flow system with three layers: layer 1, a water-table aquifer that is hydraulically connected to the Ramapo River and its tributaries; layer 2, a confining unit (or 1 ft of water-table-aquifer material where the confining unit is absent); layer 3, a lower aquifer that is confined in some areas and is the lower part of the water-table aquifer in others. Where the aquifer system is thin and no vertical hydraulic gradients were indicated on the basis of available data, the system was simulated by using only layer 1. The simulated areal extent of the layers is shown in figure 16.

The variable-size finite-difference grid shown in figure 16 was constructed to provide the greatest resolution for the valley near Mahwah; therefore, the valley-fill deposits in the central part of the grid are represented realistically. Because the grid coarsens with increasing distance from the center of the valley, the model does not represent accurately the areal extent of the unconsolidated deposits and the stream locations in the area beyond the center of the valley. The grid can be used, however, to simulate accurately water levels and flow to and from the major streams in the central part of the grid. Beyond the center of the valley, the simulation results are accurate only from a regional perspective, and cannot be considered to be an accurate representation of the local flow system.

In the model, the confining unit was represented by a layer of nodes in the same manner as the aquifer layers. Vertical leakance (vertical hydraulic conductivity divided by the vertical distance between cell nodes in each layer) was calculated as described by McDonald and Harbaugh (1984, p. 142-147); the cell node of the confining unit was located halfway between the top and bottom of the confining unit.

Aquifer System

All active finite-difference cells are assigned values of horizontal and vertical hydraulic conductivity and layer thickness. In areas where no confining unit is present, the same hydraulic conductivity was assigned to all three model layers. In areas where a confining unit is present, each of the model layers may have a different hydraulic-conductivity value. All hydraulic-conductivity values were adjusted during calibration.

The thickness of each layer was estimated from the depth-to-bedrock map (pl. 2), the maps showing lines of equal aquifer and confining-unit thickness (fig. 5 and 6), and the hydrogeologic section (fig. 7). Initially, layer 2 was assigned a thickness of 1 ft in areas where the confining unit, shown in figures 5 and 16, is absent. These data sets were modified during model calibration when it became evident that confined conditions also are present in the vicinity of the Suffern well field, where the thickness of layer 2 was increased, whereas the thicknesses of layers 1 and 3 were adjusted so that the total thickness remained constant.

Horizontal hydraulic conductivity was estimated to be 10 times the vertical hydraulic conductivity of the aquifers throughout the model, so that the anisotropy is 10:1. Vertical hydraulic conductivity tends to be less than horizontal hydraulic conductivity in stratified drift because, at a small scale, the stratified drift is composed of many layers of sediment of differing compositions. The less permeable layers tend to impede the vertical flow of ground water more than they impede horizontal flow.

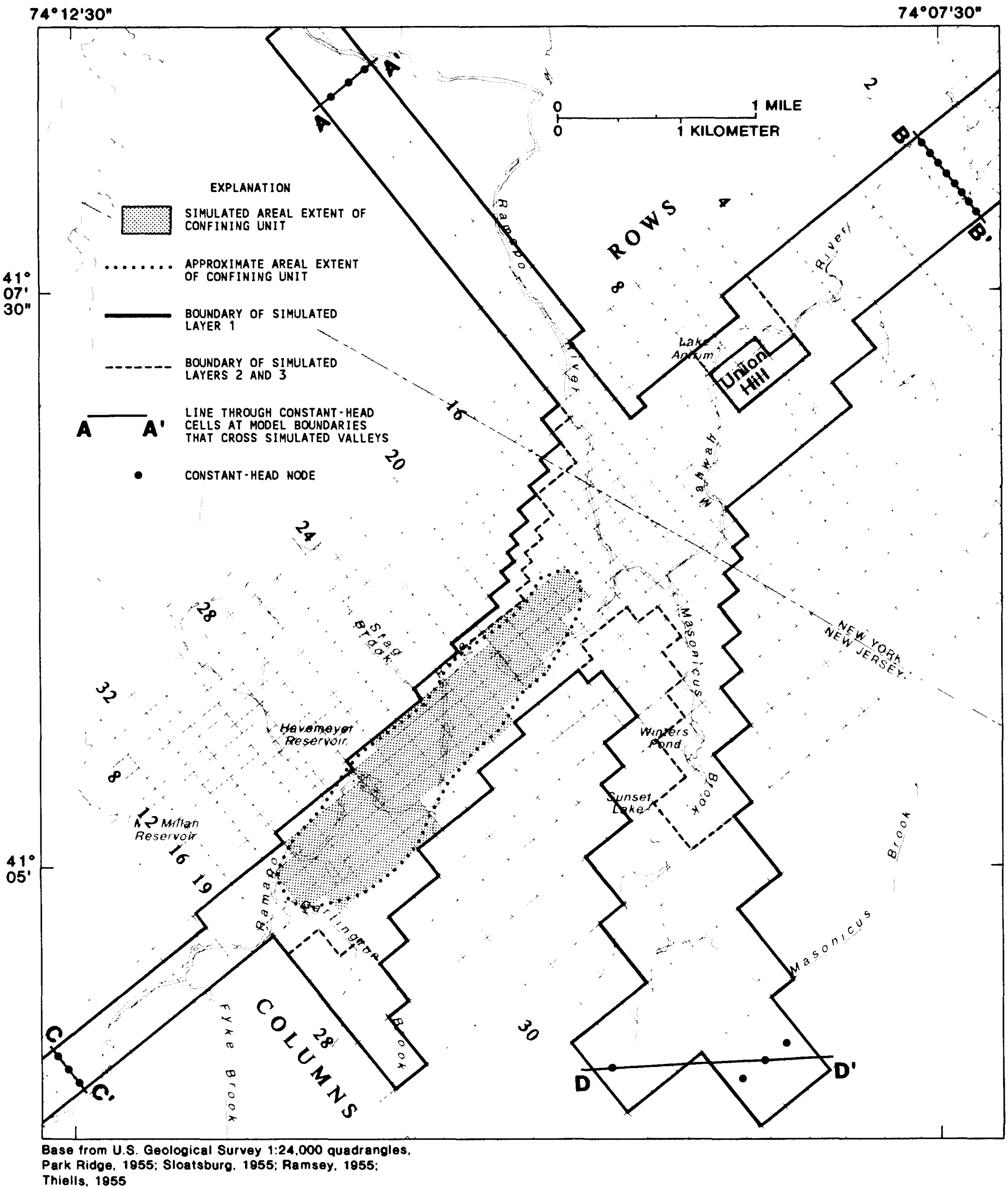


Figure 16.--Finite-difference grid for the model of the valley-fill aquifer system near Mahwah, New Jersey; the simulated areal extent of the three layers; constant-head cells; and the extent of the confining unit southwest of Mahwah.

Little information was available to define the magnitude of the anistropy in the valley-fill deposits. Laboratory measurements of vertical and horizontal hydraulic conductivities of fluvial and lacustrine sediments from the San Joaquin Valley, California, reported by Johnson and Morris (1962), indicate that anisotropy was greater than 1:1 in 46 of the 61 samples tested, and that the horizontal to vertical anisotropy of these 46 samples ranged from 2:1 to 10:1. These results are significant to the current study because fluvial and lacustrine deposits frequently are similar in composition and depositional environment to stratified drift.

The anisotropy used here is the maximum anisotropy given above--that is, 10:1. Freeze and Cherry (1979, p. 148) noted that the horizontal to vertical anisotropy of laboratory samples tends to increase with increasing sample size. Therefore, the actual degree of anisotropy could exceed 10:1. Anisotropy was not modified during calibration, but is included in the analysis of model sensitivity.

Streams

The ground-water-flow model developed by McDonald and Harbaugh (1984) was used to simulate leakage between the water-table aquifer and the streams in the study area. The Ramapo River, Masonicus Brook, Mahwah River, and Darlington Brook were represented in the model at 100 finite-difference cells in model layer 1. These cells are referred to here as river cells. The 100 river cells were combined into 16 reaches, as shown in figure 17. A schematic representation of the simulated streams is shown in figure 18. The boundaries between groups are located near streamflow-measurement stations so that simulated streamflow and changes in streamflow can be compared to measured values. Simulated streamflow at any station is obtained by summing simulated gains or losses for the river cells in the reach upstream from the station and adding that sum to the simulated streamflow at the next upstream station(s). At the farthest upstream stations included in the model on the Ramapo River (station 01387420), the Mahwah River (01387480), and Masonicus Brook (01387485 and 01387486), measured streamflows are used in the calculation. Note that no simulated streamflows for these stations appear in figure 18.

The following five values are required as model input at each river cell: (1) The altitude of the water in the stream (stage), (2) the altitude of the base of the streambed, (3) the thickness of the streambed, (4) the vertical hydraulic conductivity of the streambed, and (5) the area occupied by the stream. In the model, the streambed conductance is calculated as the product of items (4) and (5) divided by item (3). Leakage between the aquifer and the stream generally is calculated by multiplying the streambed conductance by the difference between the stage in the river and the head in the aquifer directly below the river. If the head in the aquifer is below the base of the streambed, however, leakage equals the streambed conductance multiplied by the difference between the water level in the stream and the elevation of the base of the streambed. This procedure is described in McDonald and Harbaugh (1984, p. 214).

The stage of the stream was measured directly at five sites along the Ramapo River, as shown on plate 3. These values are considered to be

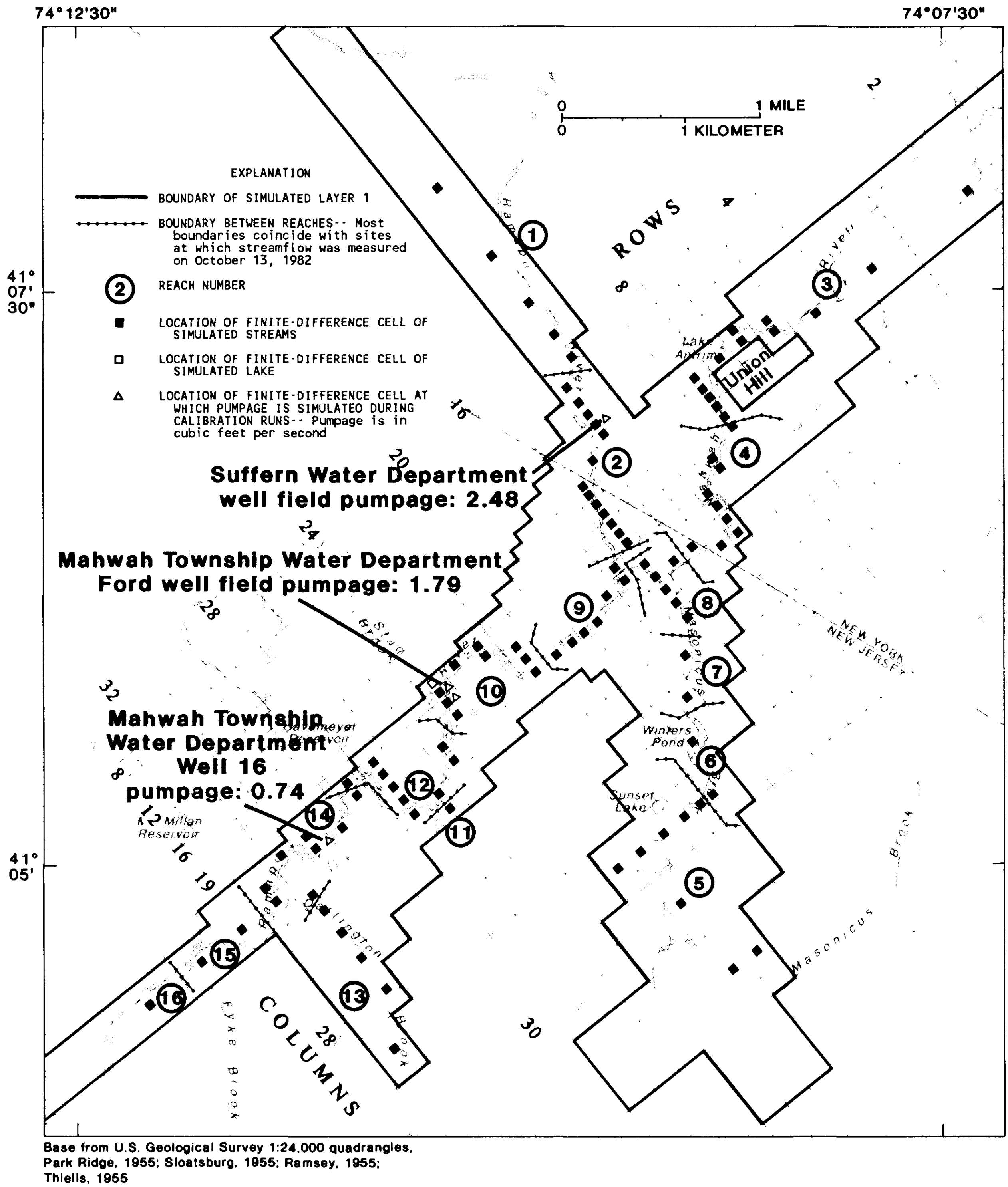


Figure 17.--Finite-difference grid for the model of the valley-fill aquifer system near Mahwah, New Jersey; simulated stream and lake cells; and cells at which pumpage was simulated during calibration.

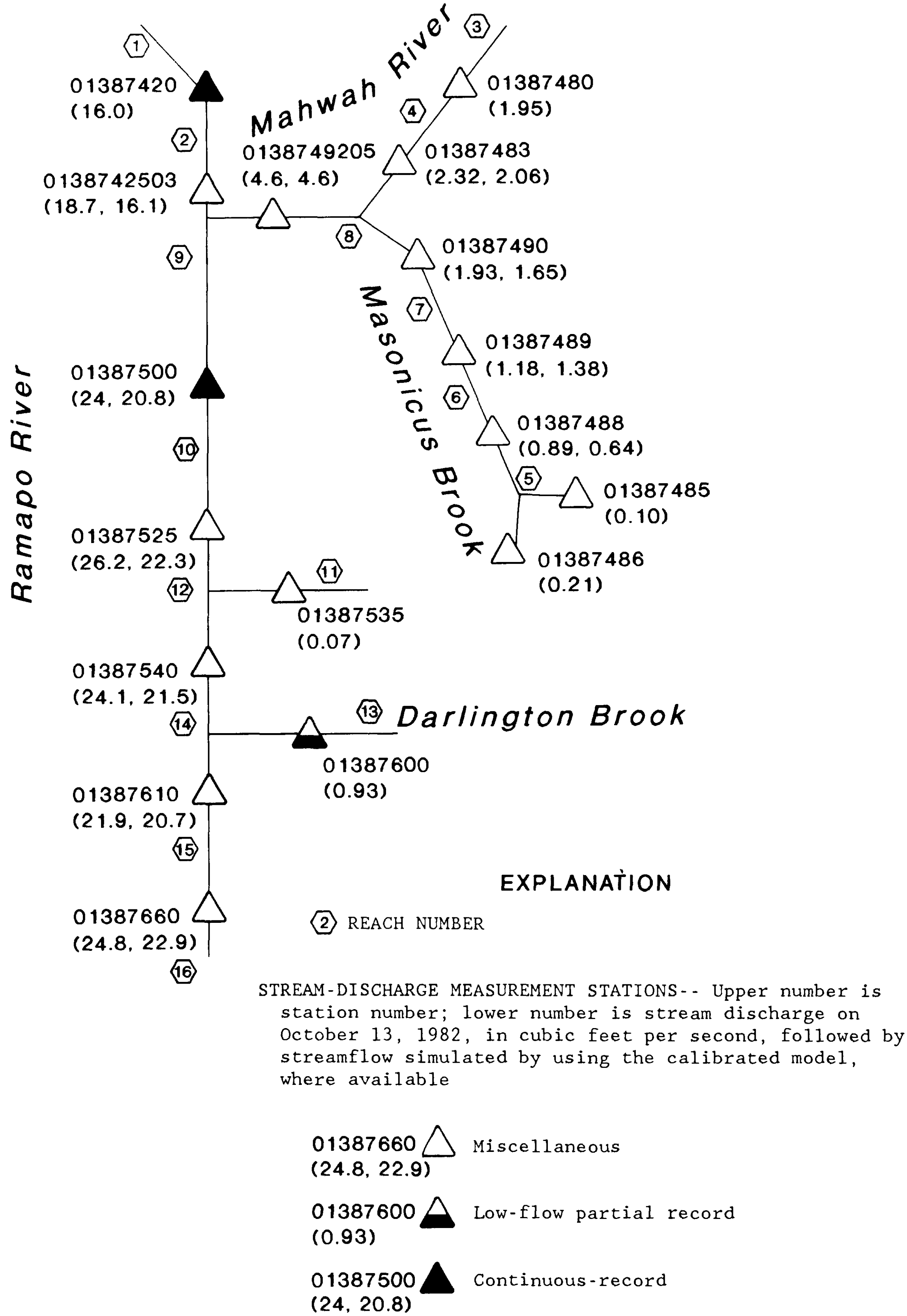


Figure 18.--Streams represented in the numerical model; streamflow measurements made on October 13, 1982; and, except at the farthest upstream stations, streamflows simulated by the calibrated model.

accurate to the nearest tenth of a foot. Elsewhere, stream stages were estimated by using the nearest measured values and 1:24,000-scale U.S. Geological Survey topographic maps with 10-ft contour intervals. These values are considered to be accurate near the five sites where the stage was measured, and less accurate elsewhere. The maximum error probably is less than 5 ft.

The streambed thickness represents the material beneath the stream; the hydraulic conductivity of this material is different from that of the surrounding aquifer material as a result of the past and present influence of the stream. The stream can import fine sediments that cause the hydraulic conductivity of the streambed to be lower than that of the aquifer, and (or) the stream can scour the streambed and cause the hydraulic conductivity of the streambed to be greater than that of the aquifer. Although no data are available on the thickness of the streambed in the Ramapo River valley, it is reasonable to assume that the streambed is more than a few feet thick but less than 25 ft thick. A 10-ft streambed thickness was simulated in most river cells in the model; however, the streambed was simulated as 20 ft thick in the four upstream river cells of Darlington Brook. This change was developed during calibration, and is discussed in more detail later.

The altitude of the base of the streambed was calculated as the simulated stage of the stream minus the streambed thickness. Although this formula implies that the river has no depth, calculating the altitude of the base of the streambed in this way does not produce erroneous results for the following reasons. First, the simulated streams generally are less than 2 ft deep in the modeled area, and often are less than 1 ft deep. Therefore, the error in the altitude of the base of the streambed generally is less than 2 ft. This error probably is similar to the error that results from assuming a streambed thickness of 10 ft. Second, the altitude of the base of the streambed was used in the calculations only if the simulated water-table level was lower than the base of the streambed.

The simulated vertical hydraulic conductivity of the streambed is 30 ft/d, which equals the average of the measurements made during the local seepage run near the Ford well field discussed earlier in this report. This value was not modified during calibration because no data were available to determine reasonable variations in this property. It also was assumed that spatial variations in the vertical hydraulic conductivity of the streambed produce relatively minor variations in streamflow gains and losses. The results of model calibration and sensitivity analysis supported this assumption.

The area occupied by the stream was determined by superimposing the model grid on 1:24,000-scale U.S. Geological Survey topographic maps and estimating the fraction of each river cell covered by water. The area of the river within each cell was calculated as this fraction multiplied by the total area of the cell.

Lake

The method used to calculate leakage between the water-table aquifer and streams also was used to calculate leakage between the water-table aquifer and a lake near Stag Brook (pl. 3). The lake was represented in the model by one river cell at column 18, row 24, as shown in figure 17.

The water level in the lake was determined from a 1:24,000-scale U.S. Geological Survey topographic map with a 10-ft contour interval. The elevation of the water surface in the lake is considered to be accurate to within 5 ft and is about 20 ft above surrounding ground-water levels. The simulated vertical hydraulic conductivity of the lakebed was considered to be less than the vertical hydraulic conductivity of the streambed material (30 ft/d), and was modified during calibration. The area occupied by the lake was estimated from a U.S. Geological Survey topographic map to be 140,000 ft^2 (square feet).

Boundary Conditions

Several types of boundary conditions were specified in the digital model to represent the following boundaries of the simulated aquifer system: (1) The four boundaries that cut across the river valleys at the northwestern, northeastern, southwestern, and southeastern ends of the modeled area (near lines A-A′, B-B′, C-C′ and D-D′ in fig. 16); (2) the boundary along the bottom and sides of the aquifer system where the valley fill borders the adjacent bedrock; and (3) the top boundary of the aquifer system (the water table).

The four boundaries that cut across the river valleys were represented as constant-head boundaries to allow the simulation of flow between the simulated aquifer system and the aquifer system beyond the model boundaries. These boundaries can be represented accurately as constant-head boundaries as long as they are sufficiently far from the simulated points of ground-water withdrawal to ensure that the withdrawals do not affect flow at the boundary. In the model, however, constant-head cells can provide an infinite amount of flow. In order to ensure that such unrealistic results were not produced in the present study, simulations were conducted in which simulated withdrawals were not excessively large. Flows from the constant-head boundaries were not affected by the simulated pumpage in any of the simulations, indicating that the constant-head cells are sufficiently far from the points of withdrawal. The finite-difference cells designated as constant-head are shown in figure 16.

The boundary between the valley fill and the adjacent bedrock is simulated as a no-flow boundary. As discussed earlier, the adjacent bedrock is much less permeable than the valley-fill aquifers. The contrast in permeabilities allows the boundaries to be represented realistically as impermeable. Although some water can seep across the boundary between the valley fill and the sedimentary formations of the Newark Supergroup, which are more permeable than the other bedrock formations, the amount of seepage is unknown and is presumed to be negligible.

The top boundary of the aquifer system is the water table, which is represented in the model as a moving boundary across which areal recharge--a defined flux--flows. Areal recharge equals precipitation that falls directly on the simulated area, plus runoff that flows into the area from the surrounding highlands, minus evapotranspiration. Because the water table is close to the land surface in the modeled area and most of the surficial materials are sand, it is assumed that all precipitation and runoff immediately recharges the water-table aquifer. This assumption is consistent with the rapid response to precipitation measured in water-table observation wells. Because only base-flow conditions are considered for the modeled area, runoff from the surrounding highlands is assumed to be negligible except where major streams are found. This assumption is consistent with field conditions observed during the October 13, 1982, seepage run. The two major streams that flow from the highlands, Darlington and Stag Brooks, are represented in the model as discussed above in the section on streams.

The areal-recharge rate simulated for the water-table aquifer in the steady-state calibration simulation is deliberately different from the actual rate of areal recharge (precipitation minus evapotranspiration) because, in the steady-state model, the simulated value is designed to balance the dynamic decline in water-table levels and the recession in streamflow that were in process as of October 13, 1982. This concept can be explained by means of a hydrologic budget. Consider the area downstream from stations 01387420, 01387480, 01387535, 01387600, 01387486, and 01387485, and upstream from station 01387660 (pl. 4)--an area of about 2.7 mi^2. The hydrologic budget for this area in terms of rates can be expressed as--

$$P + GW_I = R + ET + \Delta S_G + GW_o + W,$$

where P is precipitation, GW_I is the inflow of ground water, R is the gain (+) or loss (-) in streamflow over the area, ET is evapotranspiration, ΔS_G is the increase (+) or decrease (-) in ground-water storage (ΔS_G is nonzero because of changing water levels), GW_o is the outflow of ground water, and W is the withdrawal of water through wells. For the purpose of the ground-water budget, GW_I, R, ΔS_G, GW_o, and W are each divided by the area to produce average rates over the area. For the steady-state calibration simulation, areal recharge rate equals:

$$\text{Areal recharge rate} = P - ET - \Delta S_G = R + GW_o - GW_I + W.$$

The long-term areal-recharge rate (P - ET) is 0.074 in./d (inches per day), based on the average daily precipitation calculated from the data shown in figure 2 and an estimated evapotranspiration rate of 20 in./yr (inches per year) (0.055 in./d) reported by Vecchioli and Miller (1973, p. 11). The areal-recharge rate required in the steady-state calibration simulation can be calculated from the above equation by estimating P, ET, and ΔS_G, or R, GW_o, GW_I, and W. Unfortunately, only P, R, and W can be determined with precision from available data, so that areal recharge cannot be estimated accurately in this way. It was concluded, however, that the areal-recharge rate probably was less than the long-term areal-recharge rate because precipitation (P) during the 7 days before and including October 13, 1982, averaged 0.025 in./d, or about 20 percent of the average daily precipitation, and 1982 was a drier-than-average year as indicated by the low streamflows shown on plate 4. It was anticipated that neither temporal variations in ET nor the inclusion of ΔS_G would produce an areal-recharge rate larger than the long-term areal-recharge rate of 0.074 in./d. The areal-recharge rate used in the steady-state calibration simulation was

0.034 in./d. Because measured values of areal recharge were not available, the effects of other values were included in the sensitivity analysis.

Ground-Water Withdrawals

Well locations and ground-water withdrawals from production wells represented in the steady-state model are shown on plates 3 and 4. The withdrawal values are average values for the period October 9-13, 1982. In general, withdrawals varied little during this 5-day period, and the error introduced by ignoring these variations probably is small. The locations of the four model cells at which well discharge was simulated are shown in figure 17. Total pumpage was 5.01 ft^3/s; all ground water was withdrawn from cells in model layer 3. At the Mahwah Township Ford well field, withdrawals were divided equally between the two cells indicated in figure 17.

Steady-State Calibration

The purpose of the steady-state calibration was to (1) match the measured and simulated October 13, 1982, hydraulic heads in the water-table aquifer (pl. 3); (2) match the measured and simulated October 13, 1982, hydraulic heads that indicate significant vertical gradients (pl. 3, inset); (3) match measured and simulated October 13, 1982, streamflow, streamflow gains and losses along individual reaches, and the streamflow gain over the entire modeled area; (4) maintain physically reasonable flows at constant-head nodes; (5) maintain physically reasonable values of all model parameters; and, (6) minimize the number of independent parameter values subject to calibration.

The trial-and-error method was used to calibrate the steady-state model. This method begins by running the model with initial estimates of all model inputs. Differences between measured and simulated values of the model performance criteria (discussed below) are used to indicate changes in selected model-input data that can produce simulated values that closely approximate the measured values. After making the appropriate changes to one or more of the model inputs, the model is run again. Simulated and measured values are compared again, and the process is repeated. The calibration is complete when simulated values are close to measured values, as required by the convergence criteria given below. In this study, changes were made to the following model inputs during the steady-state calibration: (1) horizontal hydraulic conductivity of the three model layers, (2) layer thickness, (3) vertical hydraulic conductivity of the lakebed (fig. 17), and (4) rate of areal recharge.

Accuracy of Simulations

Criteria

The calibration was considered to be satisfactory when (1) calibrated water levels were within 5 ft of measured or contoured water levels everywhere in the system, (2) calibrated streamflows displayed the gain-and-loss patterns indicated by available data, (3) calibrated streamflows approximated measured values, and (4) the total calibrated change in streamflow over the modeled area approximated the measured value. Because the streamflow-measurement errors were assumed to be about 5 percent in the study area, some differences between measured and calibrated streamflows and streamflow gains and losses were expected.

Differences between simulated and measured values

Differences between simulated and measured water levels and streamflow gains and losses are presented below to indicate the degree to which simulated values match measured values. These differences were calculated as simulated values minus measured values. The mean squared error, calculated as the sum of the squared differences divided by the number of measured values, is a measure of the overall degree of difference between simulated and measured values. Note that the units of the mean squared error are in units of feet squared for the mean squared error of hydraulic heads. The mean squared error of streamflow gains and losses is not used because the measurements are not equally reliable, as indicated by the different potential measurement errors in table 4. The potential measurement error is assumed to be 5 percent of the larger of the two streamflow values used to calculate a gain or loss (see discussion on p. 33).

Hydraulic heads and streamflow.--Simulated steady-state hydraulic heads for layers 1 and 3 are shown in figures 19 and 20, respectively. The contoured simulated water-table levels in layer 1 (fig. 19) closely approximate the contoured measured water-table levels shown on plate 3 in most of the modeled area. Both contour maps indicate steep hydraulic gradients along Masonicus and Darlington Brooks and the upstream parts of the Ramapo and Mahwah Rivers; flatter gradients are seen elsewhere.

Simulated water-table levels in the wide parts of the valley, where hydraulic gradients are small (fig. 19), were strongly dominated by simulated stream elevations, and generally were insensitive to the changes made during calibration. In these areas, simulated hydraulic heads commonly were within 1 ft of measured values in all calibration runs.

The only area in which the contours in figure 19 and plate 3 are dissimilar is in the southwestern part of the aquifer, northeast of Darlington Brook. The water-table contours on plate 3 show the presence of steep gradients along the southeastern edge of the valley in this area. The simulated water-table altitudes shown in figure 19 have much lower gradients and hydraulic heads in this area. This difference between calibrated and observed ground-water levels indicates that, locally, simulated aquifer transmissivity is too high, simulated recharge is too small, and (or) streams in the area are simulated improperly. The fact that nearby heads and gradients in the center of the valley are close to measured values suggests that the differences are local and do not have a significant effect on simulation results elsewhere in the system.

Differences between measured and simulated heads at 28 measurement locations, shown in figures 19 and 20, range from -8.2 to 4.4 ft; 16 values are between -1.0 and 1.0 ft. The greatest differences, -5.6 and -8.2 ft, are seen at the measurement sites with location codes 07-08-01 and 04-10-07, respectively. The -5.6-ft difference is found in the upstream part of the Mahwah River, where the shape of Union Hill is not represented accurately (fig. 17) because the model grid is coarse. The -8.2-ft difference is found in the southwestern part of the modeled area, northeast of Darlington Brook,

Table 4.--Measured and simulated streamflow gains and losses associated with four areal-recharge rates in the modeled area

[Positive numbers indicate gains; negative numbers indicate losses; streamflow gains and losses are in cubic feet per second; areal-recharge rates are in inches per day; values in parentheses equal simulated minus measured streamflow gain or loss; see plate 4 for actual measured values for each reach; all values are reported to either the nearest tenth (reaches 2, 8, 9, 10, 12, 14, and 15) or the nearest hundredth (reaches 4, 5, 6, and 7); totals are reported to the nearest tenth]

Reach	Measured streamflow gain or loss	Potential measurement error[1]	Simulated streamflow gain or loss associated with an areal-recharge rate of: 0.024	[2]0.034	0.044	0.052
2	2.7	0.9	0.0 (-2.7)	0.2 (-2.5)	0.3 (-2.4)	0.4 (-2.3)
4	.37	.12	.03 (-.34)	.11 (-.26)	.22 (-.15)	.30 (-.07)
5	.58	.04	.09 (-.49)	.33 (-.25)	.57 (-.01)	.73 (.15)
6	.29	.06	.67 (.38)	.74 (.45)	.79 (.50)	.84 (.55)
7	.75	.10	.23 (-.52)	.27 (-.48)	.31 (-.44)	.34 (-.41)
8	.4	.2	.8 (.4)	.9 (.5)	.9 (.5)	1.0 (.6)
9	.6	1.2	.0 (-.6)	.1 (-.5)	.2 (-.4)	.3 (-.3)
10	2.3	1.3	1.4 (-.9)	1.5 (-.8)	1.5 (-.8)	1.6 (-.7)
12	-2.2	1.3	-.8 (1.4)	-.7 (1.5)	-.6 (1.6)	-.6 (1.6)
14	-3.1	1.2	-1.9 (1.2)	-1.8 (1.3)	-1.6 (1.5)	-1.5 (1.6)
15	2.8	1.2	2.2 (-.6)	2.3 (-.5)	2.3 (-.5)	2.4 (-.4)
Total	5.5	[3]1.3	2.7 (-2.8)	4.0 (-1.5)	4.9 (-0.6)	5.8 (0.3)
Total, excluding reach 2	2.8	[3]1.3	2.7 (-0.1)	3.8 (1.0)	4.6 (1.8)	5.4 (2.6)

[1] Determined as 5 percent of the larger of the two streamflow values used to calculate a gain or loss.

[2] Final calibrated value.

[3] Not a total; 5 percent of the largest streamflow (26.2 cubic feet per second) used to calculate gains and losses in all reaches.

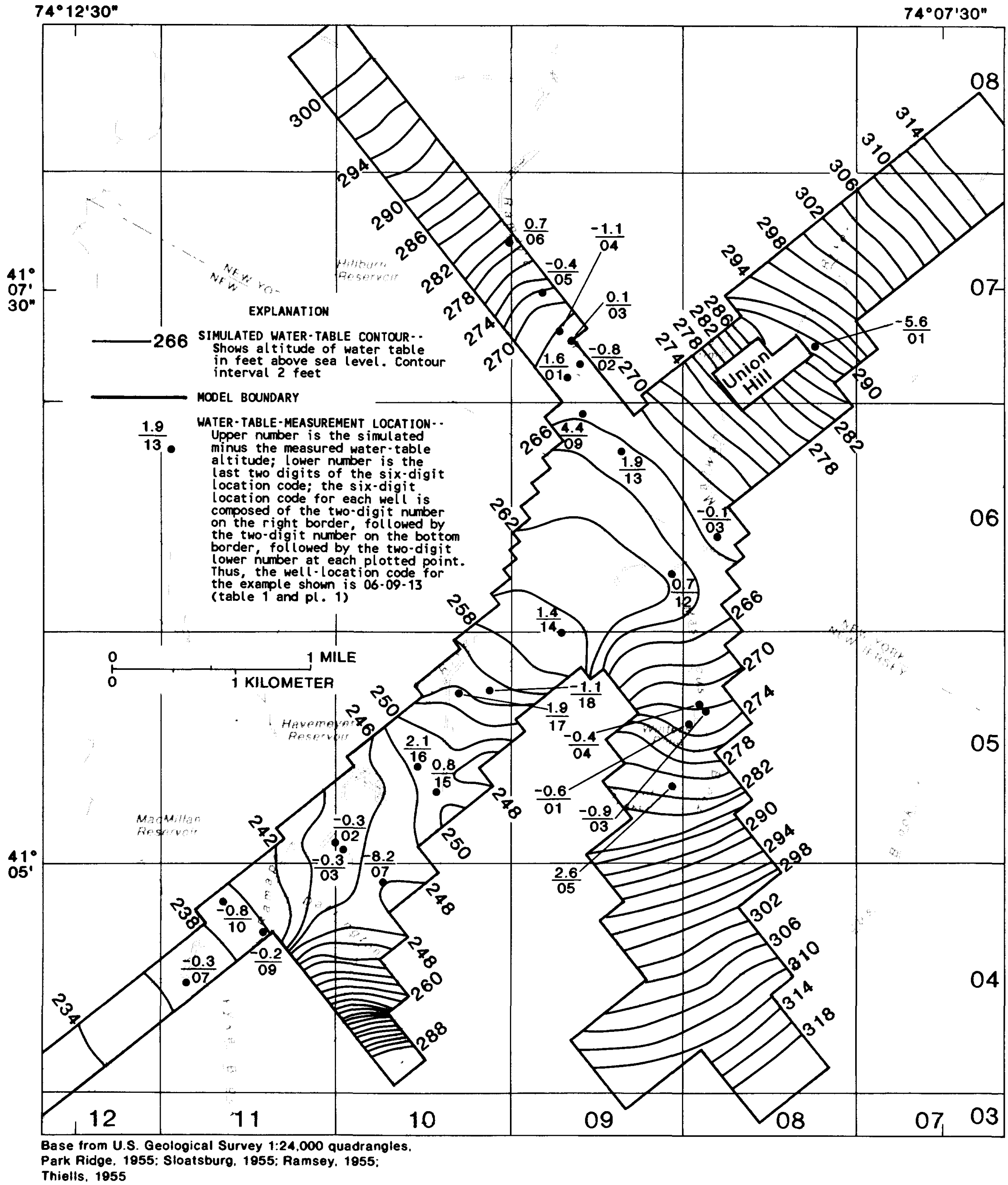
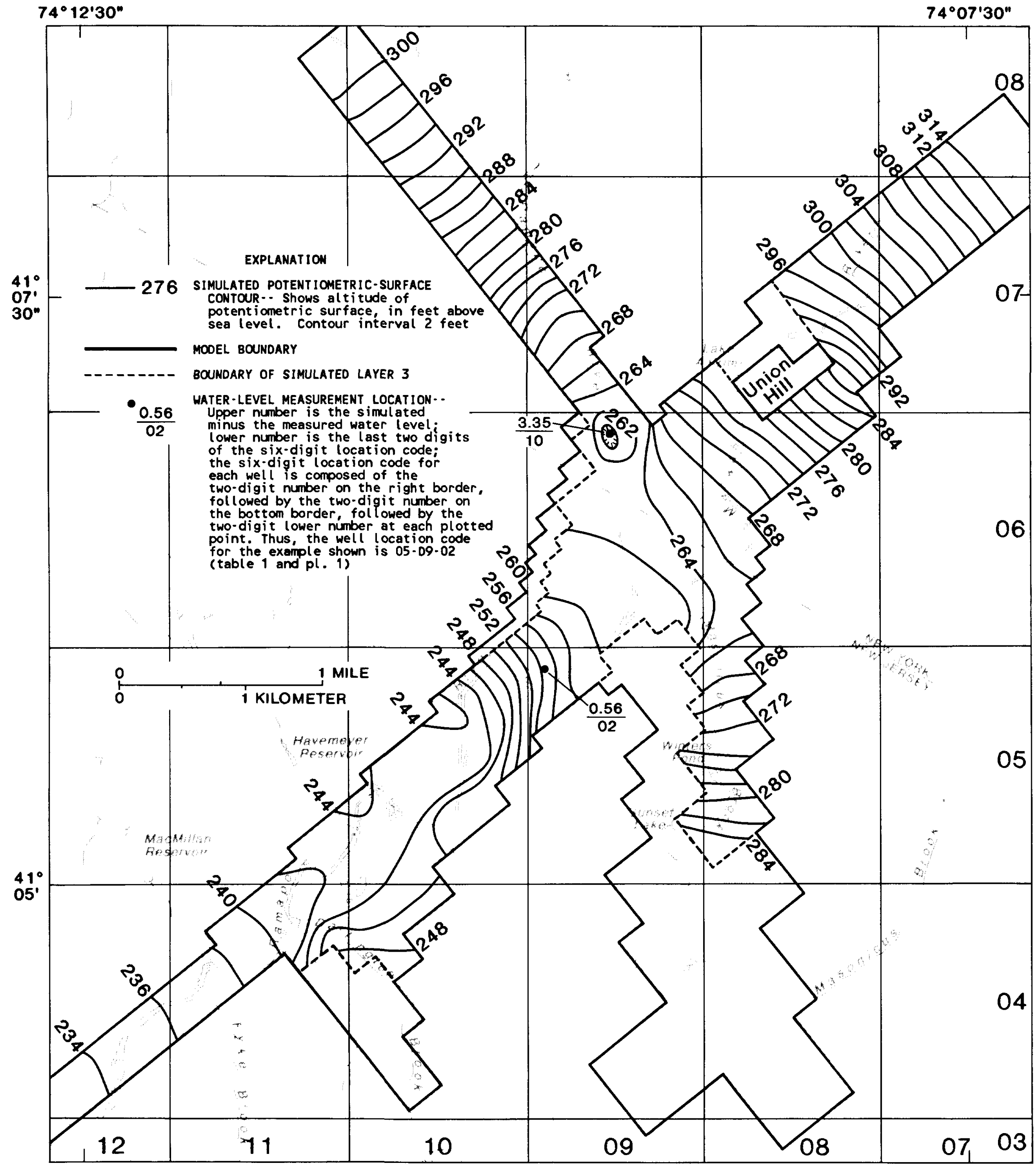


Figure 19.--Simulated water-table altitudes in layer 1 of the Ramapo River valley-fill aquifer-system model, October 13, 1982.



Base from U.S. Geological Survey 1:24,000 quadrangles, Park Ridge, 1955; Sloatsburg, 1955; Ramsey, 1955; Thiells, 1955

Figure 20.--Simulated potentiometric surface of layer 3 of the Ramapo River valley-fill aquifer-system model, October 13, 1982.

where contoured simulated water-table levels also differ from calibrated values, as discussed earlier. Simulated heads are 3 ft higher than measured heads near the Suffern well field (figs. 19 and 20). The high simulated heads in this area are attributable, in part, to the inability of the model grid to represent accurately the locations of the pumped well, the river, and the measurement points. In the final calibrated model, the mean squared error for the 28 measurement locations shown in figures 19 and 20 was 2.1 ft^2.

Streamflows simulated with the calibrated model are shown in figure 18. Simulated gains and losses along individual reaches and the simulated gain in streamflow over the entire modeled area are compared with measured values in table 4. The potential measurement error also is shown in table 4.

The simulated streamflow gains and losses closely approximate the pattern of measured gains and losses. Generally, small gains simulated in the upstream reaches of the modeled area were followed by a large gain, large loss, and large gain in the successive downstream reaches.

Many of the simulated gains and losses differ from the measured values by more than the potential measurement error. The largest differences between simulated and measured gains and losses are found in reaches 2, 12, and 14, although only the difference for reach 2 is substantially larger than the potential measurement error (table 4). The differences are caused at least in part by the simplified model representation of streambed hydraulic conductivity and areal recharge. With a few exceptions, these values are constant for the entire modeled area. Although spatial variations in these values probably could have produced a closer match between simulated and measured streamflow gains and losses, such changes were not simulated because independent estimates of realistic variations were unavailable, and the accuracy of the resulting simulated values would be questionable. The effects on simulated streamflow gains and losses produced by changes in streambed hydraulic conductivity and areal recharge were considered, however, in sensitivity runs.

The differences between simulated and measured streamflow gains and losses for reach 2 also are the result of simulating withdrawals from the Suffern well field in the same node that contains the stream reach, a consequence of the coarseness of the grid in this part of the model (fig. 17). The model accurately represents the vertical separation between the stream and the well field (the river cell is connected to layer 1 and the withdrawals are from layer 3), but the effect of the lateral distance between the well and the river is not simulated.

The differences between simulated and measured gains and losses in reaches 4, 5, 6, 7, and 8 are much larger than their potential measurement errors. However, the total simulated streamflow gain in these reaches is 2.3 ft^3/s, which is close to the measured gain of 2.4 ft^3/s. The large individual differences probably result from the large grid-cell size used to simulate most of these reaches.

The results of the steady-state calibration indicate that the measured streamflow gains and losses probably are affected by variations in the hydrogeologic characteristics of the ground-water system. Ground-water withdrawals in the modeled area under the calibration conditions generally decrease gains in streamflow and increase losses in streamflow in stream reaches near pumped wells. Wells that withdraw water from beneath the confining unit near Mahwah, New Jersey (fig. 5), however, affect stream reaches along the edges of the confining unit.

Hydraulic conductivity.--Calibrated horizontal hydraulic conductivities for the three model layers are shown in figures 21, 22, and 23. The hydraulic conductivity is the same for all three layers except where confining units are present. Estimates of hydraulic conductivity were uniquely determined in 23 zones in the calibrated model. Although this is a large number relative to the number of head and streamflow measurements used in the calibration, it reflects the variability common to valley-fill deposits of glacial origin. Hypotheses concerning the deposition of glacial deposits used to develop these hydraulic conductivities are discussed later.

Simulated hydraulic conductivities are within the ranges suggested by Lyford and others (1984) and are near the values reported by Weston and Sampson (1924, p. 70). Simulated hydraulic conductivities along the Ramapo River in New York State generally are consistent with values used in the model developed by Leggette, Brashears and Graham, Inc. (1981). The permeameter and aquifer tests discussed earlier also were used to help determine hydraulic-conductivity values at row 30, column 21, in layers 2 and 3, respectively. The permeameter-test results suggest that the vertical hydraulic conductivity of layer 2 is 3×10^{-4} ft/d; the simulated value is 2.6×10^{-4} ft/d. (Horizontal hydraulic conductivity is shown in fig. 22; vertical hydraulic conductivity is equal to 0.1 times horizontal hydraulic conductivity.) The aquifer-test results suggest that the horizontal hydraulic conductivity of layer 3 is between 600 and 650 ft/d; the simulated value is 520 ft/d. The low simulated value probably is more accurate because the wells used in the aquifer test were screened in the coarsest sand and gravel in the confined aquifer. Less permeable silty sand and gravel adjacent to the coarse material was represented in model layer 3.

Initial attempts at calibration with hydraulic-conductivity distributions that were relatively homogeneous in the vicinity of the confining unit southwest of Mahwah failed to produce simulated streamflow gains and losses in reaches 10, 12, 14, and 15 that approximated measured gains and losses. In order to improve the match between the simulated and measured streamflow gains and losses for these reaches, simulated hydraulic conductivities were modified to produce a heterogeneous distribution (shown in figs. 21, 22, and 23) on the basis of the following two hypotheses concerning depositional environments:

1. Coarse clastic material that was eroded from the steeply sloping Precambrian rocks on the northwestern side of the valley could have produced coarse, permeable deposits on the northwestern side of the valley fill. Except for the presence of the steeply sloping Precambrian rocks, little evidence is available to contradict or substantiate this hypothesis.

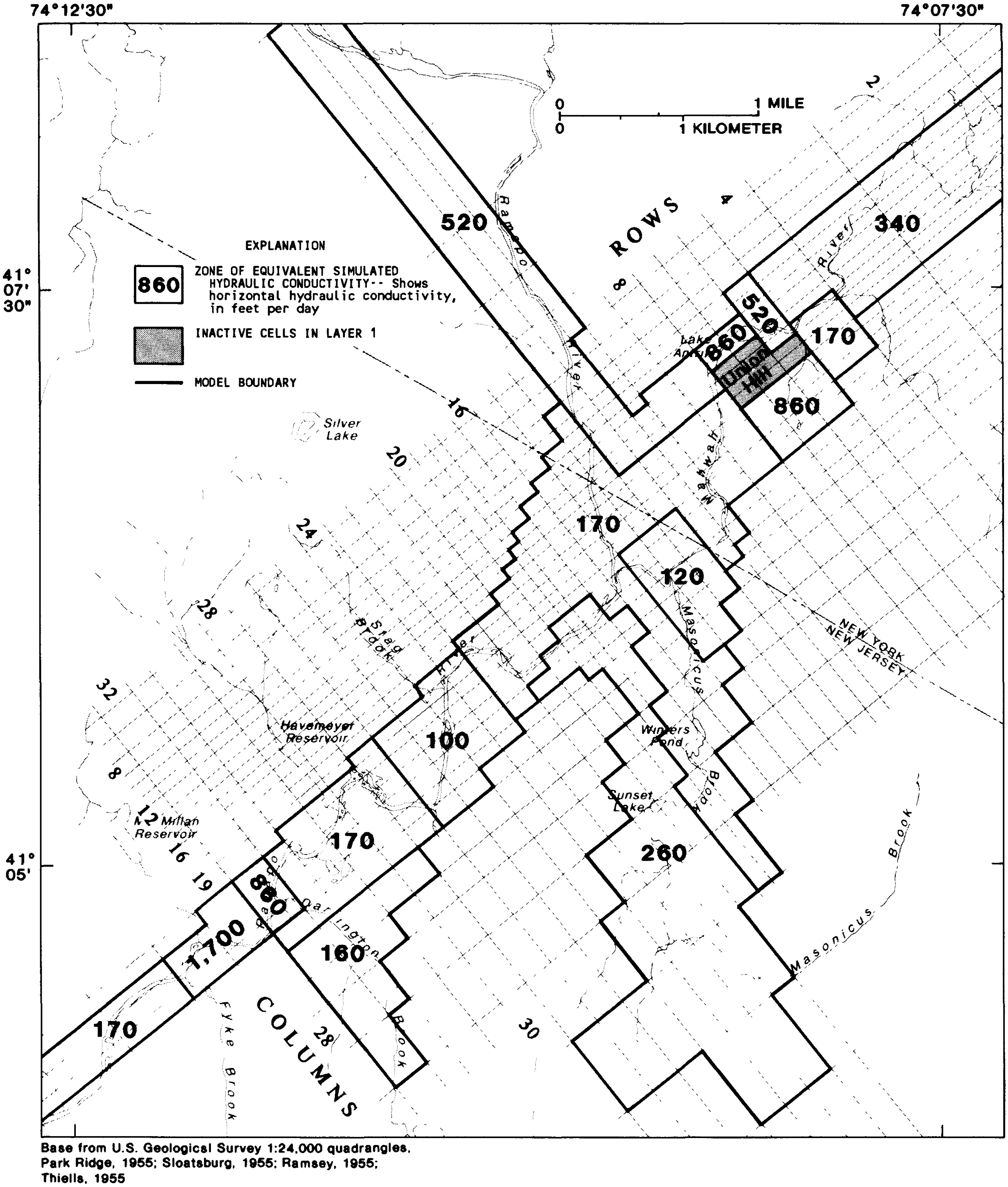


Figure 21.--Calibrated horizontal hydraulic conductivity of layer 1 of the Ramapo River valley-fill aquifer-system model.

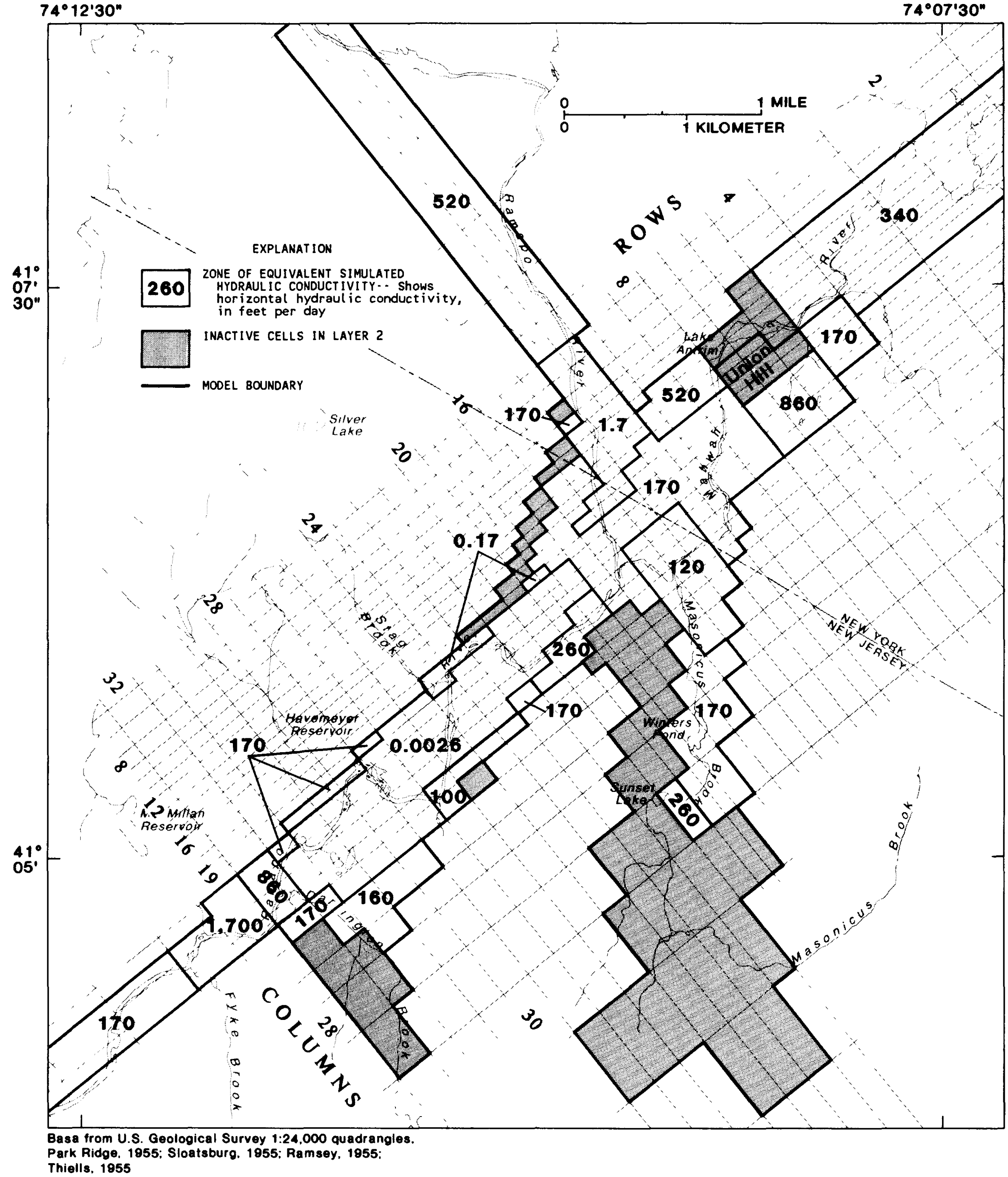


Figure 22.--Calibrated horizontal hydraulic conductivity of layer 2 of the Ramapo River valley-fill aquifer-system model.

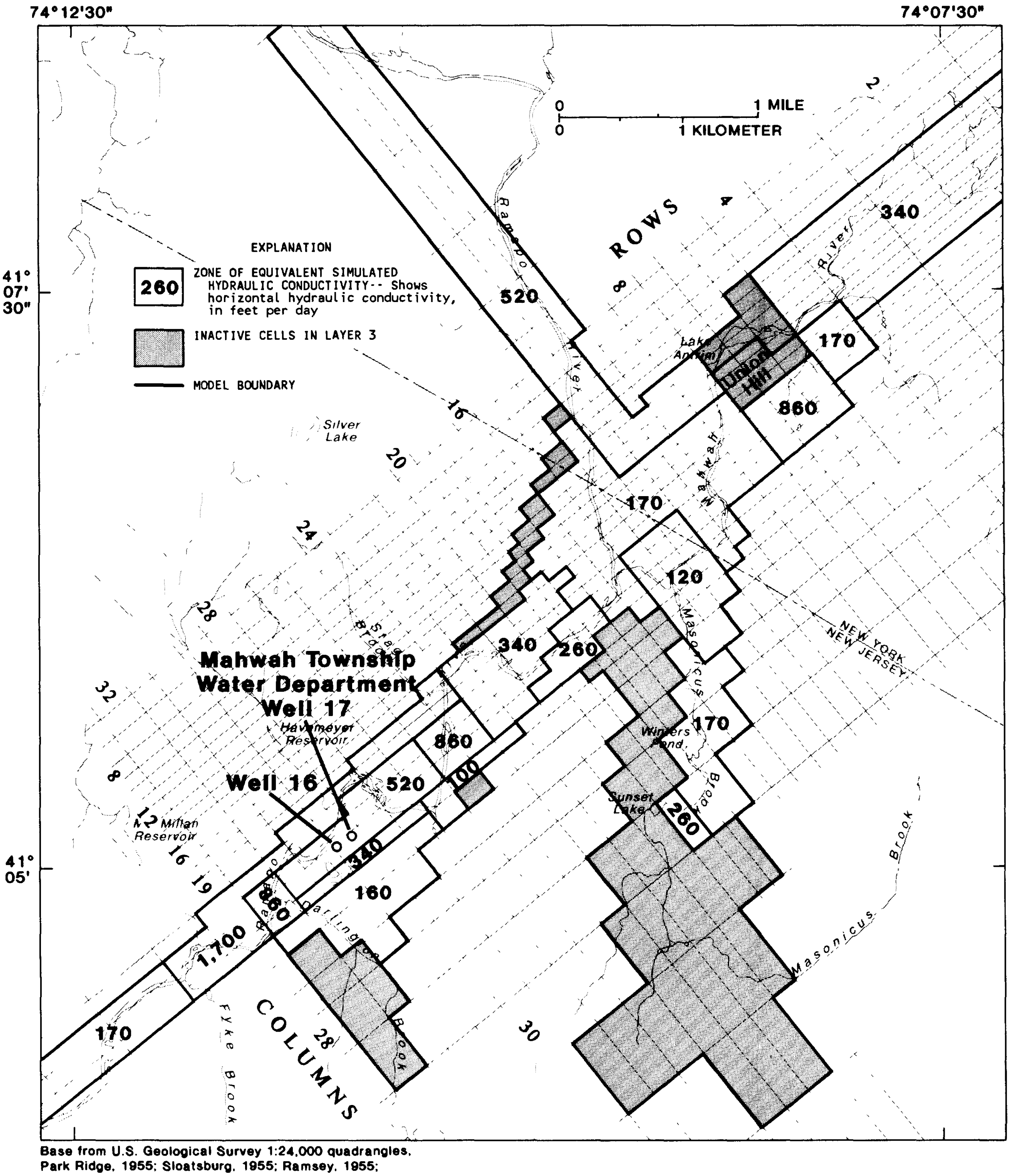


Figure 23.--Calibrated horizontal hydraulic conductivity of layer 3 of the Ramapo River valley-fill aquifer-system model.

2. An end moraine could have existed near the area in which the valley narrows, at the southwestern end of the modeled area. The presence of an end moraine is strongly suggested by the presence of the confining unit southwest of Mahwah that probably is a glacial-lake deposit. A lake could have formed there only if southward drainage was blocked; an end moraine is the most likely mechanism for blocking the drainage. The end moraine would have been coarsest and most permeable adjacent to the ice front to the north; grain size and permeability would decrease to the south. (This depositional sequence was developed on the basis of the discussion of glacial geology in Koteff and Pessl, 1981.)

The simulation indicated that the gain in streamflow along reach 10 was a direct result of infiltration from the lake at the end of Stag Brook (fig. 17) and that the hydraulic head near well 10-05-05 (pl. 3, inset) was strongly affected by the vertical hydraulic conductivity of the confining unit in the area. The best match between measured data and simulation results was achieved by using a lakebed vertical hydraulic conductivity of 6.0 ft/d, a vertical hydraulic conductivity for layer 2 of 0.017 ft/d (0.1 times the horizontal hydraulic conductivity shown in fig. 22), and a thickness for layer 2 of 10 to 40 ft. Under these conditions, simulated leakage from the lake was 1.53 ft^3/s.

Recharge rate.--Areal recharge rates ranging from 0.024 to 0.052 in./d were simulated. Table 4 shows the simulated and measured streamflow gains and losses over individual reaches and over the entire modeled area. Gains and losses over the modeled area excluding reach 2 also are included because the difference between the simulated and measured gain along reach 2 is much greater than the other differences.

A comparison of simulated and measured heads and streamflow gains and losses for the simulated recharge rates does not clearly identify a "best" value. The fit between simulated and measured streamflow gains and losses, as measured by the difference between simulated and measured total gain in streamflow, improves as the recharge rate increases. However, if the large loss along reach 2 is omitted, the fit improves as the recharge rate decreases. The fit between simulated and measured hydraulic heads, as measured by the mean squared error, improves as the recharge rate increases, but the improvement is small (less than 0.4 ft^2). An areal recharge rate of 0.034 in./d was used in the calibrated model as a compromise between these contradictory results.

Sensitivity Analysis

Model calibration produces a well-calibrated estimate of a model input, such as hydraulic conductivity or areal recharge rate, when the simulated model outputs, such as hydraulic heads and streamflow gains and losses, are sensitive to the model input. The sensitivity of model outputs to selected model inputs was investigated through sensitivity runs of the model. Each sensitivity run was executed by changing the value of one model-input parameter, calculating new model outputs, and comparing the new model outputs to the calibrated model outputs. Large differences indicate that the model outputs are sensitive to the simulated change in the model input and suggest that the model input is well calibrated. Because the model was calibrated by trial and error, the terms "large" and "well calibrated" can be defined only qualitatively. "Large" means a sufficient difference to have been important in the trial-and-error calibration process; "well calibrated" means that the model input is estimated by calibration within the range of the change made for the sensitivity run. The changes in model inputs in the sensitivity runs were equal to the probable errors in the model inputs, as determined by the modeler. The probable errors were determined based on the degree to which the model inputs were well-measured--that is, the independent field or laboratory data that were available--and on the results of the trial-and-error calibration.

The approach of changing the value of only one model parameter in the sensitivity simulations ignores possible correlations between model inputs. Correlation and its effect on the precision of calibrated model inputs can be explained through the following example.

Consider a confined aquifer of infinite areal extent, which is cut vertically by one straight stream that extends to infinity in either direction. Water in the stream is at elevation h_o. Assuming a homogeneous, isotropic aquifer with one-dimensional flow toward or away from the stream, the head in the aquifer, h, at a distance, x, from the stream, can be calculated as--

$$h = h_o + \frac{q}{Kb} x,$$

where q is the streamflow gain (positive value) or loss (negative value) per unit length of the stream, b is the constant thickness of the aquifer, and K is the hydraulic conductivity of the aquifer. In this case, the calculated hydraulic head is constant at any distance, x, as long as the ratio q/K is constant. Thus, q and K are completely correlated--that is, they could not be estimated independently by using only measured values of hydraulic head. This correlation between streamflow gains and losses or other ground-water-flow rates and hydraulic conductivity is common in ground-water-flow problems. For some problems the highest correlations are those between areal recharge and hydraulic conductivity.

If the type of sensitivity analysis performed for the model were performed on the example problem, and changes in K and q each produced a large change in h, it would be concluded that both K and q were well calibrated. Because of the high correlation between K and q, however, this conclusion is in error. High correlations between K and q are less likely for the model developed for the Ramapo River valley-fill deposits than in many ground-water-flow problems because measured values of both hydraulic heads and streamflow gains and losses are used as calibration criteria. In the above example, this situation is analogous to one in which measurements of both q and h are available, facilitating the accurate estimation of K.

The model-input changes used in the sensitivity analysis are described in table 5. Zones with assigned hydraulic conductivities used in the sensitivity analysis are shown in figure 24. Selected results of the sensitivity runs, which include simulated streamflow gains and losses and changes in simulated hydraulic heads, are shown in figures 25 through 29. The last column of table 5 lists which, if any, model outputs used to calibrate the model were changed sufficiently from their calibrated values

Table 5.--Model inputs changed for each of the sensitivity runs

[K_H, horizontal hydraulic conductivity; K_V, vertical hydraulic conductivity]

Run number	Model input changed	Change from calibrated value	Part of model changed	Model outputs showing sufficient change to affect calibration
1	None	None	None	None
2	Streambed and lakebed conductance	x 1.3	All 101 river cells (including the lake cell)	Do.
3	...do...	x 0.7	...do...	Do.
4	...do...	x 10.0	...do...	Reaches 5, 10, 14, 15, heads
5	River stage	- 0.1 foot	...do...	None
6	...do...	- 0.5 foot	...do...	Heads
7	Anisotropy (K_H/K_V)	x 10.0 (K_V x 0.1)	All active cells	Reaches 2, 12, 15, heads
8	K_H and K_V	x 1.3	[1]Zone 1 (unconfined aquifer, model layer 1)	Reaches 2, 5, 8, 14, 15, heads
9	K_H and K_V	x 0.7	...do...	Do.
10	K_H and K_V	x 1.3	[1]Zone 2 (Darlington Brook, all three model layers)	None
11	K_H and K_V	x 0.7	...do...	Do.
12	K_H and K_V	x 1.3	[1]Zone 3 (northwest of the confining unit, all three model layers)	Reach 10
13	K_H and K_V	x 0.7	...do...	Do.
14	K_H and K_V	x 1.3	[1]Zone 4 (confined aquifer, model layer 3)	Heads
15	K_H and K_V	x 0.7	...do...	Do.
16	K_H and K_V	x 0.5	...do...	Do.
17	K_H and K_V	x 0.2	...do...	Do.
18	K_H and K_V	x 2.0	[1]Zone 5 (confining unit, model layer 2)	Do.
19	K_H and K_V	x 0.7	...do...	None
20	K_H and K_V	x 1.3	All active cells	Reaches 2, 5, 8, 10, 14, 15, heads
21	K_H and K_V	x 0.7	...do...	Do.
22	Areal recharge	x 1.3	All active cells in model layer 1	All reaches, heads
23	...do...	x 0.7	...do...	Do.

[1] Shown in figure 24.

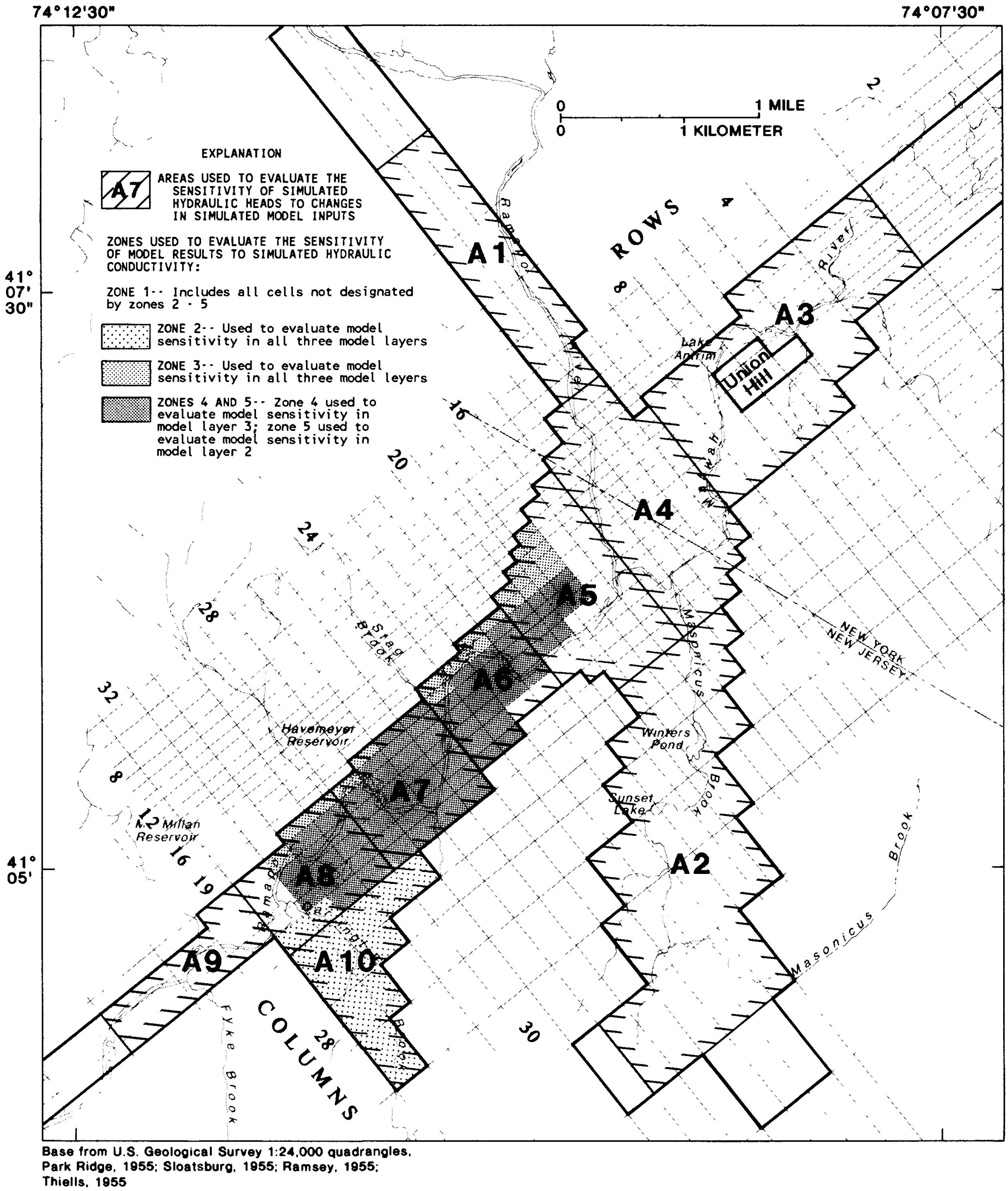


Figure 24.--Areas used to evaluate the sensitivity of simulated hydraulic heads to changes in simulated model inputs, and zones assigned hydraulic conductivities used in the sensitivity analysis.

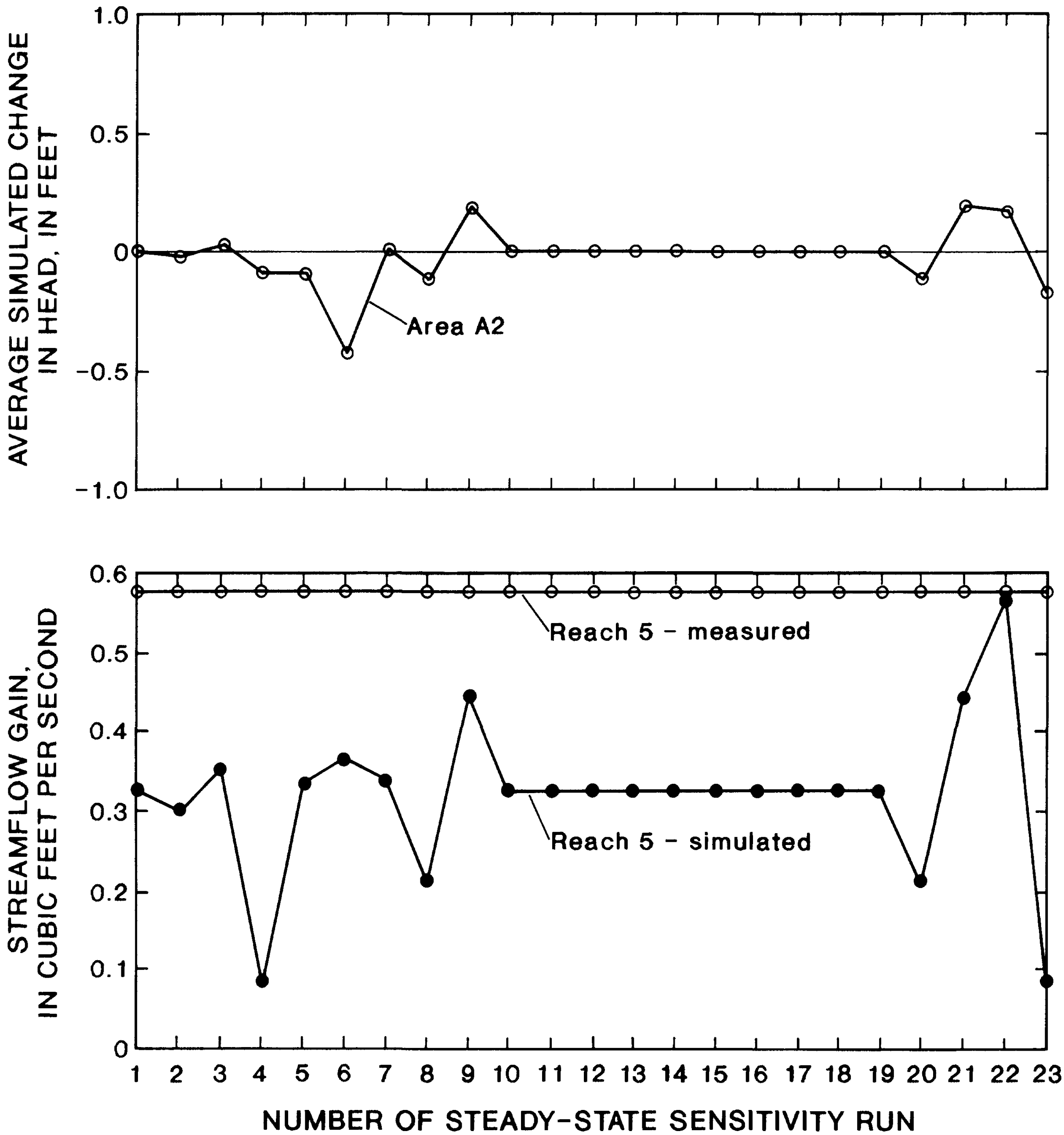


Figure 25.--Streamflow gains and average changes in hydraulic head along the upstream reach of Masonicus Brook (Reach 5 and Area A2) simulated by the calibrated model and in sensitivity runs (table 5); and gains in streamflow measured on October 13, 1982.

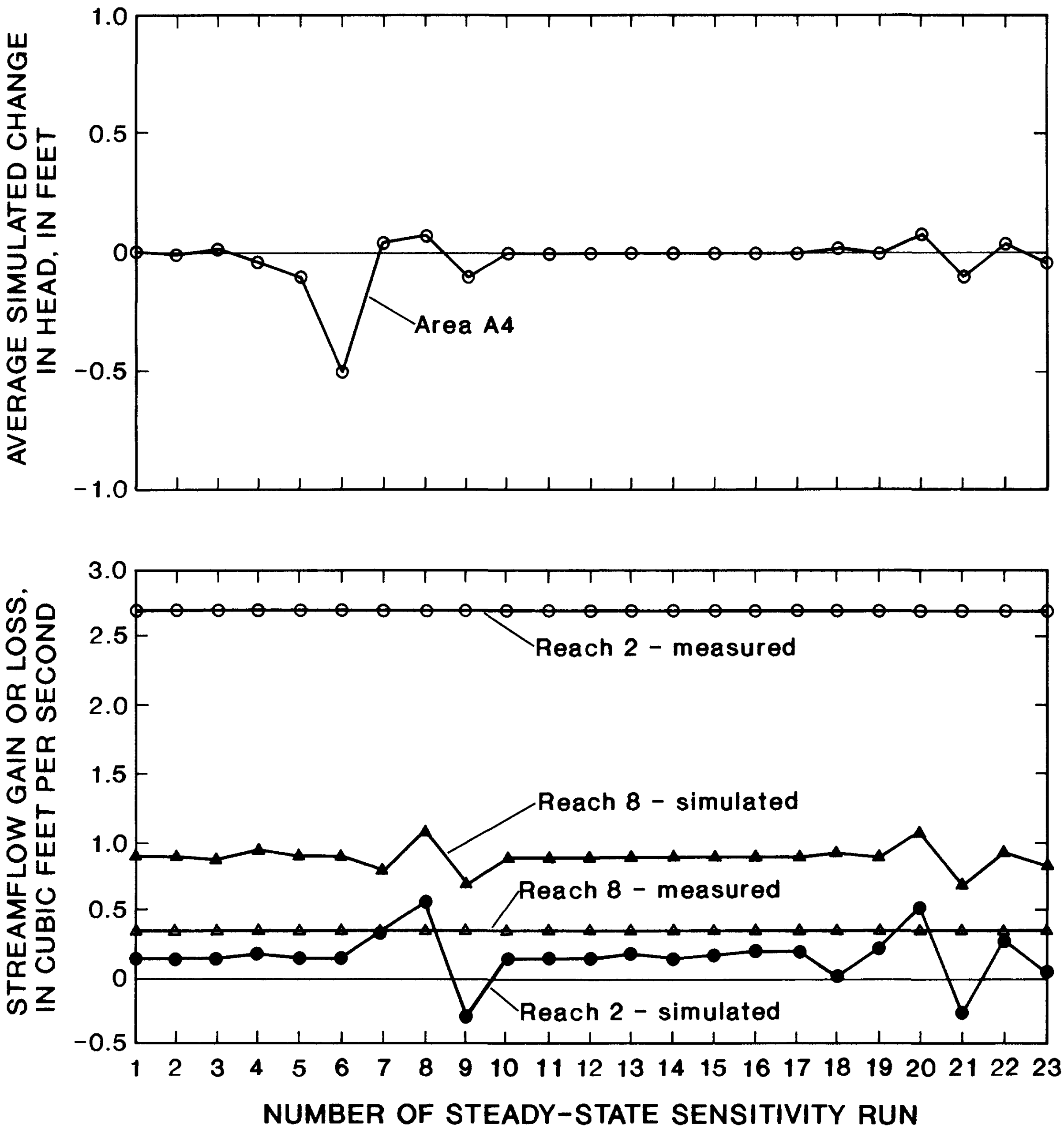


Figure 26.--Streamflow gains and losses and average changes in hydraulic head along the Ramapo River north of Mahwah, New Jersey (Reaches 2 and 8 and Area A4), simulated by the calibrated model and in sensitivity runs (table 5); and gains in streamflow measured on October 13, 1982.

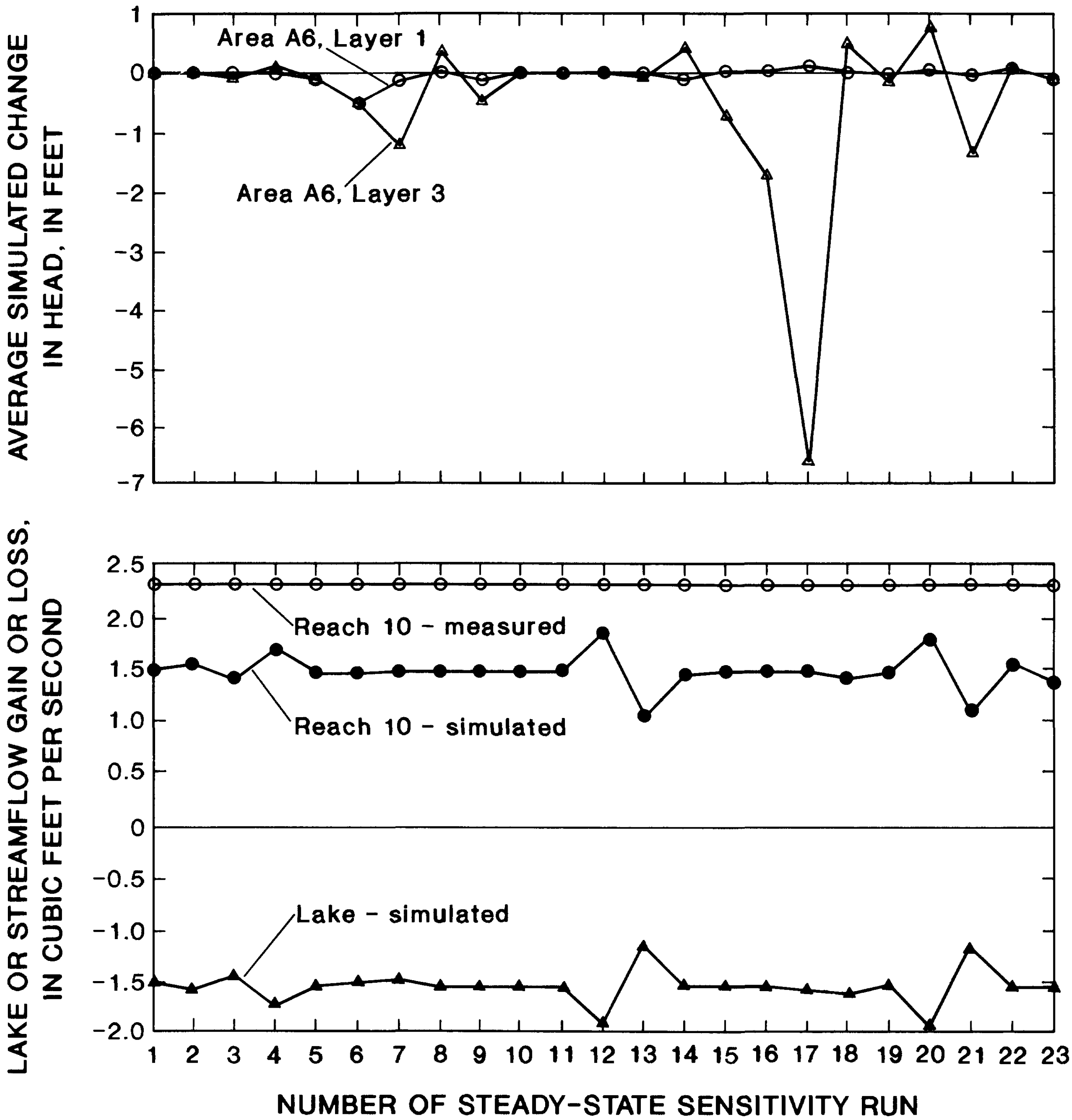


Figure 27.--Ground-water recharge from the lake, streamflow gains, and average changes in hydraulic head along the Ramapo River adjacent to the Mahwah Township Ford wells (Reach 10 and Area A6), simulated by the calibrated model and in sensitivity runs (table 5); and gains in streamflow measured along Reach 10 on October 13, 1982.

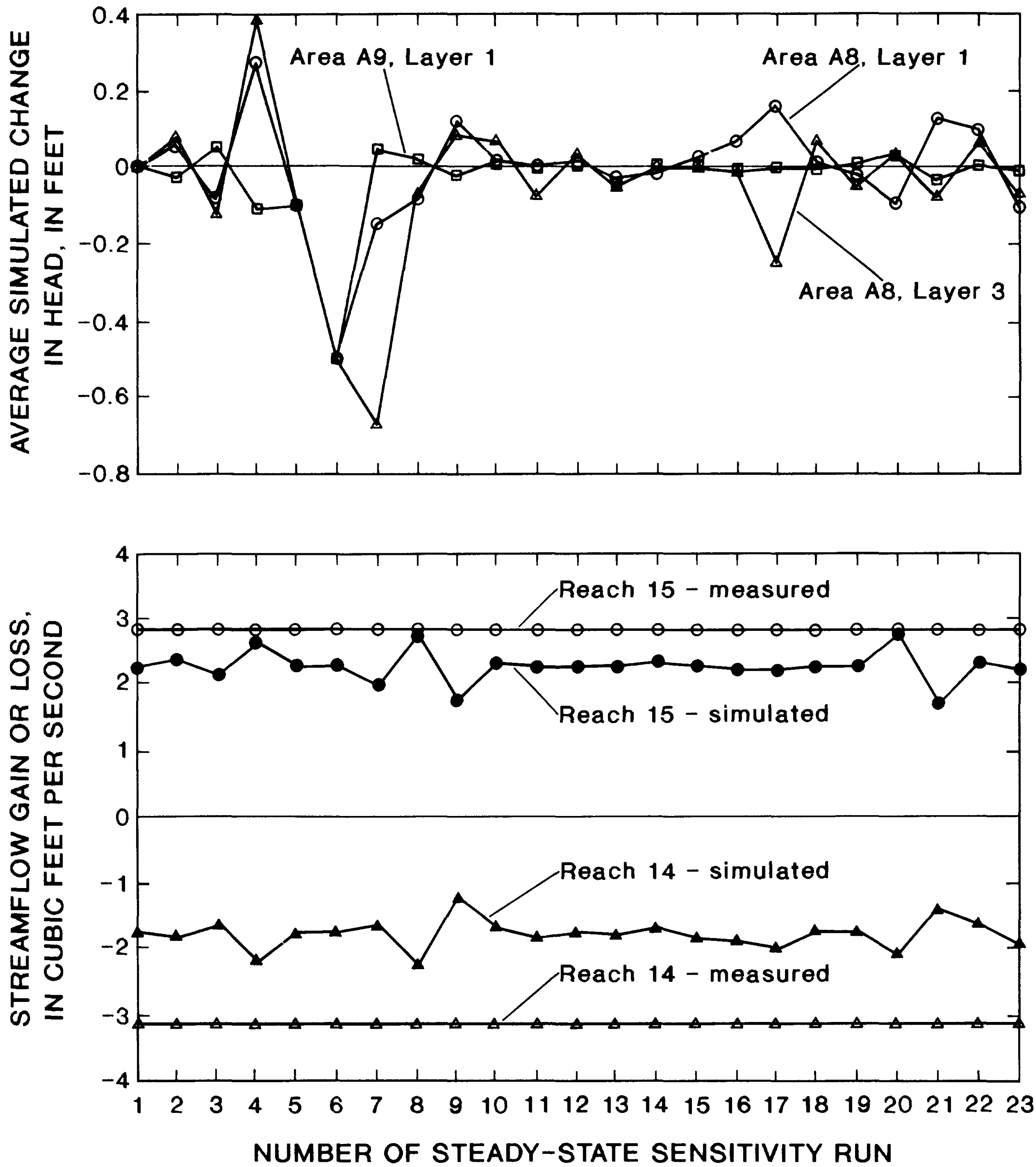


Figure 28.--Streamflow gains and losses and average changes in hydraulic head along the Ramapo River southwest of Mahwah, New Jersey (Reaches 14 and 15 and Areas A8 and A9), simulated by the calibrated model and in sensitivity runs (table 5); and gains in streamflow measured on October 13, 1982.

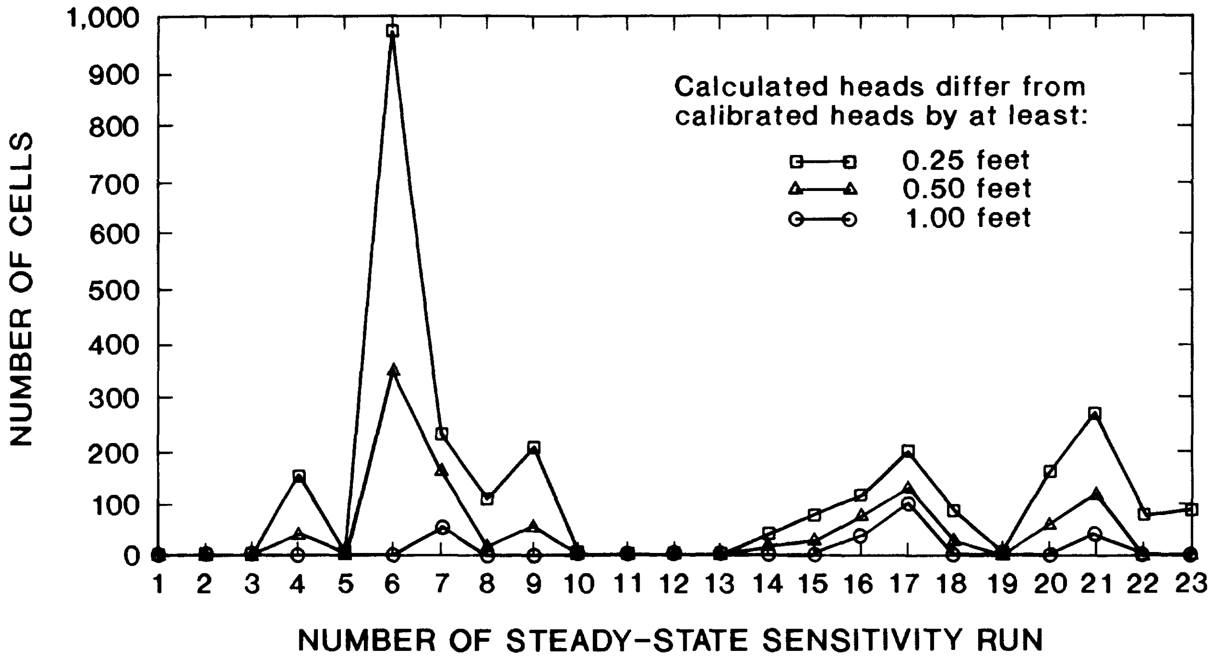


Figure 29.--Number of cells with heads that differ from the calibrated heads by at least 0.25, 0.5, and 1.0 foot for each sensitivity run (table 5).

to influence the calibration. "None" indicates that none of the model outputs was affected significantly by the parameter change, and that the amount of uncertainty in the calibrated value of this parameter is greater than the amount of change from the calibrated value listed in table 5.

Figures 25 to 28 are graphs of simulated streamflow gains and losses along reaches 2, 5, 8, 10, 14, and 15, and at the cell that represents the lake (fig. 17). Measured gains and losses are shown for comparison. Reaches 4, 6, 7, 9, and 12 are not included in the figures because the simulated streamflow gains and losses along these reaches were near calibrated values in all, or nearly all, sensitivity runs. Reaches 1, 3, 11, 13, and 16 are not included in the figures or discussion because measured gains and losses are not available, and these five reaches did not affect model calibration except to verify that their simulated gains or losses did not contradict other information. For example, although no gain or loss was measured for reach 16, plate 4 shows that reach 16 is part of a longer reach that extends from station 01387660 to 01387670. Thus, the simulated gain or loss along reach 16 should not contradict the loss measured along the longer reach.

Figures 25 to 28 also include graphs of the average change in simulated head in areas of the model adjacent to reaches 2, 5, 8, 10, 14, and 15. These areas, referred to as areas A2, A4, A6, A8, and A9, are shown in figure 24. The following steps were used to calculate average head changes: The calibrated head was subtracted from the simulated head at each cell and the difference was multiplied by the cell area; then these values were added and then divided by area A2, A4, A6, A8, or A9. Figure 29 shows the number of model cells at which simulated and calibrated heads differed by at least 0.25, 0.50, and 1.00 ft for each sensitivity run. The total number of active cells in the model was 977.

Streambed hydraulic conductivity

In sensitivity runs 2, 3, and 4, the simulated vertical hydraulic conductivity of the streambed and lakebed sediments of all 101 river cells were multiplied by 1.3, 0.7, and 10.0 (table 5), respectively. These changes are equivalent to either multiplying streambed and lakebed hydraulic conductivity by 1.3, 0.7, and 10.0, or dividing streambed and lakebed thickness by these values. Although the hydraulic conductivity of the streambed generally is not expected to be as much as 10 times the simulated value, because this would be inconsistent with the values of streambed hydraulic conductivity calculated from the results of the local seepage run, sensitivity run 4 was included to test whether a large streambed hydraulic conductivity would produce the large streamflow gains and losses measured in reaches 10, 12, 14, and 15 (table 4).

Compared to the calibration simulation (run 1), the magnitude of simulated gains and losses generally increased in response to increased streambed conductance values (runs 2 and 4) and decreased in response to decreased streambed conductance values (run 3) (figs. 26-28), with the exception of reach 5 (fig. 25), where the resulting change in hydraulic head produced the opposite response. Reaches 5, 10, 14, and 15 were most responsive, but the largest change produced by multiplying stream and lakebed conductance by 1.3 in run 2 was only 0.14 ft^3/s (reach 14). In run

4, the tenfold increase in streambed and lakebed conductance changed simulated gains and losses along these reaches from -0.44 to 0.38 ft^3/s, and it produced a closer match between simulated and measured gains and losses along reaches 10, 14, and 15, but not along reach 5. This tenfold increase in run 4 did not eliminate the difference between simulated and measured values along reaches 10, 12, 14, and 15, in the area of the confining unit.

Simulated hydraulic heads were affected only by the tenfold increase (run 4) in lake and streambed conductance. Generally, heads in the area of the confining unit and Darlington Brook (areas A6, A7, A8, and A10) increased, and heads elsewhere (areas A1, A2, A3, A4, A5, and A9) decreased slightly.

The streambed and lakebed vertical hydraulic conductivities are calibrated within an order of magnitude. However, independent measurements of the streambed hydraulic conductivity performed as part of the local seepage runs support the simulated value. The hydraulic conductivity of the lakebed was not measured directly; it was calibrated on the basis of its substantial effect on the simulated gain in streamflow along reach 10. If the model is conceptually correct in this area, the estimated value of lakebed hydraulic conductivity probably is accurate.

River stage

In sensitivity runs 5 and 6, the simulated river stage in all 101 river cells decreased by 0.1 and 0.5 ft, respectively. In the real system, an increase in river stage accompanies other changes in the system, such as an increase in recharge and an increase in streamflow. The relation between river stage and streamflow is indicated by the rating curve for the Ramapo River near Mahwah (station 01387500; R.D. Schopp, U.S. Geological Survey, written commun., 1983). The rating curve indicates that, near the 24 ft^3/s flow measured October 13, 1982, a 0.1-ft decline in river stage results in a decline in streamflow from 20 to 14 ft^3/s; a 0.5-ft decline results in a decline in streamflow from 55 to 14 ft^3/s.

In the sensitivity run, the change in river stage is executed without simulated increase in recharge, and simulated streamflows were insensitive to changes in river stage. However, simulated hydraulic heads were affected directly. The average change in head in all areas (fig. 24) was approximately equal to the change in river elevation. The change in average simulated heads in the confined aquifer (layer 3 of areas A6, A7, and A8) was slightly greater than the change in river elevation (figs. 27 and 28). In area A6, the average simulated head in the confined aquifer declined 0.54 ft in run 6. The additional 0.04-ft decline probably was caused by the change in the saturated thickness of the water-table aquifer. This decrease in saturated thickness results in a decrease in transmissivity which, in turn, effectively causes the source of recharge to the wells in the confined aquifer to include areas of the water-table aquifer and the river that are outside the areas included for the model calibration.

Although simulated streamflows were not sensitive to changes in river stage applied to all river cells, they probably would be locally sensitive to nonuniform changes in river stage, because such changes could affect local flow directions and magnitudes. The simulated river stage correctly represents the actual river stage within 0.5 ft for the calibrated conditions.

Anisotropy

In sensitivity run 7, the vertical hydraulic conductivity of all model layers was decreased by a factor of 10, so that the ratio of horizontal to vertical anisotropy was increased from 10:1 to 100:1. A large value of anisotropy was considered in this sensitivity run because, as discussed previously, the actual value may exceed 10. Simulated gains and losses changed more than 0.10 ft^3/s only in reaches 2, 15 (figs. 26 and 29), and 12; the changes were 0.26, 0.32, and 0.18 ft^3/s, respectively. The change in reach 2 improved the match between simulated and measured values; the other changes worsened the match.

Simulated hydraulic heads generally were sensitive to the large change in anisotropy. The greatest changes occurred in layer 3 of areas A6 and A8 (figs. 24, 27, and 28), and increased the simulated difference in head between the water-table and confined aquifers. At measurement location 05-09-02 (fig. 20), the difference between measured and simulated heads changed from 0.56 to -0.76 ft. The mean squared error for heads in sensitivity run 7 was 2.0 ft^2, slightly less than 2.1 ft^2 calculated for the calibrated model.

These results indicate that simulated hydraulic heads and streamflow gains and losses are sensitive to an order-of-magnitude change in anisotropy. The increase in anisotropy effectively reduced the flow of ground water between the confined and unconfined aquifers and lowered heads in the confined aquifer. Simulated anisotropy is considered to be calibrated within an order of magnitude.

Hydraulic conductivity

Simulated hydraulic conductivity was varied in sensitivity runs 8 through 21. For runs 8 through 19, the model was divided into five zones so that changes in simulated streamflow gains and losses and hydraulic head caused by changing hydraulic conductivity throughout the model (runs 20 and 21) could be attributed to local changes. The five zones are shown in figure 24. Zone 2 was designed to test the importance of hydraulic conductivity in the Darlington Brook area on model results, particularly those for the confined aquifer. Zone 3 was designed to test the importance of the connection between the confined and unconfined aquifers northwest of the confining unit (pl. 3) on model results. Zones 4 and 5 were designed to test the importance of the hydraulic conductivity of the confined aquifer and confining unit, respectively, on model results. Possible inaccuracy in the simulated areal extent of the confined aquifer was not addressed directly because of the existence of a large number of possible variations. However, the pronounced sensitivity of model results to the hydraulic conductivity in zone 3 implies that the simulation results also are sensitive to some changes in the simulated areal extent of the confining unit. Zone 1, which includes all cells not in zones 2 through 5, was designed primarily to test the effect of the hydraulic conductivity of the unconfined aquifer on model results.

The changes made in sensitivity runs 8 through 21 are listed in table 5. These changes were applied to vertical and horizontal hydraulic conductivity so that the initial ratio of horizontal to vertical anisotropy (10:1) was maintained. Hydraulic conductivities in zones 1 through 4 were increased and then decreased by 30 percent. Additional changes in the hydraulic conductivity of the confined aquifer (zone 4) were made because of the importance of this aquifer to the dynamics of the ground-water system. Hydraulic conductivities in zone 5 were doubled and decreased by 30 percent.

Simulated heads and streamflow gains and losses were least sensitive to changes in the hydraulic conductivity in zone 2, along Darlington Brook (runs 10 and 11), and in zone 5, the confining unit as shown on plate 3 (runs 18 and 19). The insensitivity of model results to the hydraulic conductivity of the confining unit probably is related to the low hydraulic conductivity of the confining unit relative to that in other parts of the valley-fill deposits.

Changes made in the hydraulic conductivity in zone 3, northwest of the confining unit (runs 12 and 13), did not affect the flow between the confined and unconfined aquifers. Simulated flows along reaches 14 and 15, simulated heads in areas A8 and A9, and heads throughout the model were near their calibrated values (figs. 28 and 29). (Reaches are shown in figure 17; areas are shown in figure 24.) The effect of these changes on the simulated gain in reach 10 and the simulated loss from the lake (fig. 27) are caused, in part, by local variations in hydraulic conductivity.

Changes in hydraulic conductivity in zone 4 (runs 14 through 17), the confined aquifer, changed the simulated hydraulic heads in areas A6 (fig. 27), A7, and A8 (fig. 28), and slightly changed the streamflow loss along reach 14 (fig. 28), but had little effect on other simulation results. Increases in the hydraulic conductivity in zone 4 resulted in heads in the confined aquifer that were nearer to those in the overlying part of layer 1, whereas decreases had the opposite effect. Simulated heads in area A6, at the northeastern limit of the confined aquifer, were most sensitive; heads in area A8, at the southwestern limit of the confined aquifer, were least sensitive.

Simulation results generally were most sensitive to changes in the hydraulic conductivity in zone 1. Changes in zone 1 caused changes in simulated streamflow gains and losses along reaches 2, 5, 8, 14, and 15, and changes in simulated heads in areas A2, A4, A6, and A8 (figs. 26, 27, and 28).

In sensitivity runs 20 and 21, the hydraulic conductivity of the entire model was increased and then decreased by 30 percent. The graphs in figures 25 through 28 can be used to associate the resulting changes in model results with the changes that were caused by changing the hydraulic conductivity in zones 1 through 5.

To summarize, the horizontal hydraulic conductivity of the aquifers is calibrated to within 30 percent of the simulated values in zones 1 (unconfined aquifer) and 4 (confined aquifer). The hydraulic conductivity of the aquifers is calibrated less accurately in zones 2 (Darlington Brook), and 3 (northwest of the confining unit). The hydraulic conductivity of the confining unit in zone 5 also is calibrated less accurately. Independent permeameter-test measurements of the vertical hydraulic conductivity of the confining unit (zone 5) show that this unit is much less permeable than the surrounding aquifer material, and the sensitivity runs can be interpreted to indicate that the vertical hydraulic conductivity of the confining unit is so low that a more accurate estimate is unnecessary.

These results indicate that predictive simulations made with this model would be inaccurate if the simulated predicted flow system is sensitive to the hydraulic conductivity of the aquifers in zone 2 (Darlington Brook) or 3 (northwest of the confining unit). The lack of sensitivity of the calibrated system to the hydraulic conductivity in zone 3 also indicates that the areal extent of the confining unit is not well calibrated in this area. Accuracy of these values could be improved by collection of additional data, including additional measurements of hydraulic head in zones 2 and 3.

Areal recharge

In sensitivity runs 22 and 23, the simulated areal recharge rate of 0.034 in./d was increased 30 percent to 0.044 in./d and decreased 30 percent to 0.024 in./d, respectively. The streamflow gains and losses for these simulations and a simulation in which the areal recharge rate equals 0.052 in./d are shown in table 4. This last simulation is included in table 4 to display the effect of an increased areal recharge rate but is excluded from table 5 and figures 25 through 29 because its effects are similar to, although more exaggerated than, those of run 22.

In these three simulations, increasing the simulated areal recharge rate improved the match between measured and simulated streamflow gains and losses along reaches 2 (fig. 26), 4, 5, (fig. 25), 7, 9, 10 (fig. 27), 15 (fig. 28), and the total of all 11 reaches, and worsened the match along reaches 6 and 8 (fig. 26) and the total of all reaches except reach 2. Decreasing the areal recharge rate had the opposite effect on the match between measured and simulated streamflow gains and losses. The changes in the areal recharge rate produced changes in average simulated hydraulic heads in areas A2 and A5 (fig. 24) of about 0.17 ft and 0.15 ft, respectively. Heads in these areas rose with increased areal recharge rate and declined with decreased areal recharge rate; in other areas, the effect was smaller.

The results of sensitivity runs 22 and 23 and the additional run in table 4 indicate that simulation results are sensitive to the simulated changes in the rate of areal recharge. However, as noted previously, the simulated changes did not consistently improve or worsen the match between measured and simulated streamflow gains and losses. In summary, the areal recharge rate is calibrated within 30 percent of its simulated value of 0.034 in./d.

SUMMARY AND CONCLUSIONS

The most productive aquifer system in the Ramapo River basin is the valley-fill deposits along the Ramapo River. These deposits are as much as 200 ft thick and most consist of gravel, sand, and thin local silt layers.

Most of the valley-fill deposits are saturated to within 10 ft of land surface, and water-table conditions prevail. In the northern part of the study area, confined conditions exist beneath a confining unit about 2 mi long and 0.5 mi wide. The measured vertical hydraulic conductivity of this confining unit in two permeameter tests was 3×10^{-4} ft/d, although the actual value might be smaller. The average transmissivity and storage coefficient of the confined aquifer--15,700 ft^2/d and 1.3×10^{-4}, respectively--were calculated on the basis of data from two aquifer tests. Results of these tests and one additional aquifer test indicate that recharge to the confined aquifer through the overlying confining unit is negligible but that recharge around the edges of the confining unit is substantial.

Contoured water-table levels in the study area indicate that the dominant flow direction is parallel to the Ramapo River valley and major connecting valleys, except where tributaries from the highlands recharge the valley fill from the sides. Along the length of the valley, the hydraulic gradient averages 0.0014 ft/ft but is twice as large in some locations. Much steeper gradients are found in the valley-fill deposits along Masonicus Brook in New Jersey, along the Mahwah and Ramapo Rivers in New York, and where tributaries enter these streams from the surrounding highlands. Cones of depression around pumped wells were apparent at the Suffern Water Department well field and the Mahwah Township Ford well field.

Base-flow stream discharges were measured at sites throughout the valley in October 1981 and May and October 1982, and at sites along two reaches near the Mahwah Township Water Department Ford well field and Oakland Borough Water Department Soons well field in September 1983 to quantify stream-aquifer interactions. The 1981 and 1982 seepage runs showed that, along the Ramapo River, reaches alternately gain and lose water. Many of the measured losses exceeded local ground-water withdrawals and occurred just downstream from an abrupt increase in transmissivity of the valley-fill deposits. The results of the 1983 seepage runs substantiated the results of previous seepage runs and produced estimates of the vertical hydraulic conductivity of the streambed that range from 25 to 35 ft/d.

A three-dimensional, finite-difference model was constructed to quantify the hydrogeologic characteristics of the ground-water system and to evaluate the hydrologic relations between ground-water withdrawals and streamflow in the northern part of the study area. Simulation results indicate that measured streamflow gains and losses caused by ground-water withdrawals from the valley-fill deposits also are affected by variations in the hydrogeologic characteristics of the ground-water system. Changes in hydraulic conductivity, aquifer thickness, and valley width can affect the water-transmitting properties of the valley-fill deposits. The presence of shallow confining units can reduce the hydraulic connection between the river and wells screened in the underlying confined aquifer; streamflow losses from ground-water withdrawals then would move upstream or downstream to an area of increased hydraulic connection where the confining unit is absent. The lack of sensitivity of the calibrated model to the hydraulic conductivity of the aquifers in the area northwest of the confining unit indicates that the hydraulic conductivity of the aquifers and the areal extent of the confining unit are not well calibrated in this area. Additional data would be required to determine these values more accurately.

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Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York

[Well locations shown on plate 1; construction features of shallow potentiometers at locations shown on plate 1 are described on p. 29, and are not included in this table; Aquifer codes: SFDF, stratified drift; PSSC, sedimentary formations of the Passaic Formation of Olsen (1980) of the Brunswick Group of the Newark Supergroup; ORGM, basalt flows of the Orange Mountain Basalt of the Brunswick Group of the Newark Supergroup; FLVL, sedimentary formations of the Feltville Formation of the Brunswick Group of the Newark Supergroup; PCMB, Precambrian gneiss; gal/min, gallons per minute; (gal/min)/ft, gallons per minute per foot of drawdown; --, no data available; Ap. loc., approximate location (see footnote 1); additional abbreviations listed in footnote 11]

NJWRD well number	Location code\1	Owner	Local name or number\2	Latitude	Longitude	Primary use of water\3	Aquifer code	Driller	Date completed	Altitude of land surface\4 (feet)
				BOROUGH OF FRANKLIN LAKES						
03-196	00-13-01	M A V Construction	1	410001	741354	H	SFDF	Wittwer, W.E.	03-10-67	453
03- 74	00-13-02	Ensing, O Richard	2	410020	741339	H	ORGM	Burrows, Inc	06-24-56	470
03- 41	01-10-02	Hackensack WC	Tw Frnkln Lks	410145	741024	U	PSSC	Artesian	11-10-69	325
03- 39	01-11-01	Hackensack WC	Tw Mable Ann	410114	741155	U	ORGM	Artesian	01-26-65	390
03- 36	01-11-02	Franklin Lks Boro WD	Frnkln Lks '57	410154	741123	P	SFDF	Artesian	05-02-57	350
03- 35	01-11-03	Franklin Lks Boro WD	Pulis Ave 1	410126	741139	I	ORGM	Rinbrand	09-12-67	370
03-200	01-12-01	Knoeller, Elmer	1	410144	741222	H	PSSC	Burrows, Inc	08-16-55	410
03- 42	01-12-02	Hackensack WC	Tw Frnkln Lks	410130	741201	U	PSSC	Artesian	06-00-65	193
03- 40	01-13-03	Hackensack WC	Rockland	410156	741337	P	PSSC	Rinbrand	01-27-73	440
03-201	02-11-01	Hall, Raymond E	1	410213	741129	H	ORGM	Wittwer, W.E.	01-14-55	385
03-202	02-11-02	Wylie, John	1	410216	741128	H	ORGM	Wittwer, W.E.	01-02-54	387
03-199	02-12-01	Brokaw, Percy	1	410224	741207	H	ORGM	Wittwer, W.E.	08-14-56	637
03- 38	59-12-01	Hackensack WC	Frnkln Lks 1	405930	741249	P	SFDF	Rinbrand	01-11-80	420
03-195	59-13-01	Urban Farms Inc	1	405929	741301	I	SFDF	Burrows, Inc	07-29-58	420
03-198	59-13-02	Ackerman, John E	1	405936	741314	H	ORGM	Rinbrand	08-02-62	455
03-194	59-13-03	Woodmere Inc	1	405936	741324	H	SFDF	Wittwer, W.E.	03-06-67	453
03-197	59-13-04	Rotelle Construction	1	405937	741325	H	SFDF	Wittwer, W.E.	04-17-64	455
03- 37	59-14-01	Hackensack WC	Hilltop Ter Tw	405947	741414	U	FLVL	Artesian	03-03-66	430
				TOWNSHIP OF MAHWAH						
03-159	01-10-01	Herlihy, Timothy F	1	410159	741022	H	PSSC	Burrows, Inc	12-13-57	380
03-165	02-10-01	Knichel, Russell W	1	410240	741015	H	PSSC	Burrows, Inc	02-01-55	410
03-164	02-10-02	Knichel, Raymond	1	410240	741016	H	PSSC	Burrows, Inc	08-07-55	410
03-154	02-10-03	Blokker, John	1	410240	741035	H	PSSC	Sikkema Co	03-04-65	375
03-162	02-10-04	Boucher, R E	1	410240	741044	H	PSSC	Wittwer, W.E.	03-07-56	350
03-155	02-10-05	Colonial Realty	1	410245	741051	H	PSSC	Sikkema Co	06-06-65	373
03-153	03-10-01	Darlington C C	1	410344	741037	H	PSSC	Sikkema Co	02-22-63	325
03- 64	03-10-02	Wehran, F L	1	410349	741033	H	PSSC	Nann, William	05-17-52	310
03- 63	03-10-03	Okonski, Constant	1	410359	741057	H	PSSC	Nann, William	12-22-52	390
03-143	03-11-01	Bergen Co Park Comm	Fyke Rd 2	410313	741128	R	ORGM	Rinbrand	05-15-62	445
03- 28	03-11-02	Mahwah Twp WD	4096	410318	741101	P	PSSC	Rinbrand	08-01-65	450
03-144	03-11-03	Bergen Cn Park Comm	Fyke Rd 7	410321	741116	R	ORGM	Rinbrand	12-06-61	463
03-142	03-11-04	Bergen Cn Park Comm	Fyke Rd 1	410334	741137	I	ORGM	Rinbrand	01-10-69	420
03- 77	03-12-01	Moramarco, Dr	1	410313	741259	H	ORGM	Sikkema Co	03-21-64	290
03-157	03-12-02	Fagen, Eloise	1	410317	741256	H	SFDF	Roger Inc	08-05-63	292.5
03-158	03-12-04	Lord, Donald C	1	410324	741254	H	ORGM	Sikkema Co	09-10-64	295
03-149	03-12-05	Kelley, Joe	1	410330	741255	H	FLVL	Ackerman	12-06-80	280
03- 78	03-12-07	Stadler, Walter L	1	410340	741234	H	FLVL	Nann, William	06-07-50	280
03-150	03-12-08	Smith, Virgil	1	410340	741252	H	SFDF	Modern Inc	04-26-79	285.4L
03-151	03-12-09	J & H Construction	2	410342	741252	H	SFDF	Modern Inc	05-01-79	273
03- 53	03-12-10	Harrington, John	1	410352	741250	H	PCMB	Slater Bros	07-21-75	260
03- 56	03-12-11	Klecha, Albert G	1	410354	741230	H	SFDF	Rinbrand	05-24-57	240
03- 55	03-12-12	Ranslow, A G	1	410358	741250	H	PCMB	Rinbrand	02-26-57	250
03- 50	03-12-16	Little, Ronald	1	410354	741250	H	PCMB	Slater Bros	08-18-75	270
03- 47	03-13-06	Davis, Donald J	1	410322	741326	H	PCMB	Kieffer, H	12-26-61	290
03- 57	03-13-07	Patterson, W W	1	410335	741301	H	SFDF	Wittwer, W.E.	09-24-57	240
03- 51	03-13-08	Lawson, William	1	410340	741308	H	PCMB	Nann, William	08-16-51	310
03-160	03-13-09	Chase, Harry H	1	410348	741311	H	PCMB	Kiser, A.W.	05-25-51	520
03- 48	03-13-13	Marbar Realty Co	1	410325	741326	H	PCMB	Rinbrand	04-30-57	320
03- 49	03-13-14	Ackley, Raymond	1	410333	741313	H	PCMB	Slater Bros	08-10-75	300
03- 52	03-13-15	Owens, Fred	1	410341	741308	H	PCMB	Algeier Bros	08-29-55	300
03- 7	03-14-01	Boy Scouts of Amer	Camp Glen 3	410350	741430	T	PCMB	--	00-00-59	640
03-163	04-09-02	Templin, AF	1	410429	740939	H	PSSC	Wittwer, W.E.	09-30-54	368
03- 65	04-09-03	Malacrida, S	1	410430	740945	H	PSSC	Peerless	12-06-57	390

See footnotes at end of table.

Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

Location code	Depth to bedrock from land surface (feet)	Minimum casing diameter (inches)	Depth of interval below land surface\5 Top (feet)	Bottom (feet)	Well acceptance test\6 Depth of water level below land surface Static (feet)	Pumping (feet)	Yield\7 (gal/min)	Specific capacity ((gal/min)/ft)	Length of test (hours)	Remarks\1
BOROUGH OF FRANKLIN LAKES										
00-13-01	--	6	--	--	28	33	10	2.00	1.5	Screened (?)
00-13-02	8	8	21	90	19	46	6	.22	2.0	Ap. loc.
01-10-02	60	8	65	302	3	30	10	.37	0.5	Hole filled; Ap. loc.
01-11-01	40.5	12	40	47	--	--	--	--	--	Hole filled; Ap. loc.
01-11-02	--	22	40	56	7	31	200	8.33	24.0	Screened; Ap. loc.
01-11-03	69	6	73	300	28	54	112	4.35	8.0	Second aquifer PSSC; Ap.loc.
01-12-01	45	6	46	100	16	59	9	.21	1.0	
01-12-02	169	8	169	193	--	--	--	--	--	Hole filled; Ap. loc.
01-13-03	67	8	67	400	29	134	65	.62	5.0	Ap. loc.
02-11-01	44	6	45	85	15	37	10	.48	1.5	
02-11-02	42	6	43	115	15	57	5	.12	1.5	
02-12-01	15	6	15	305	27	200	3	.02	1.5	
59-12-01	--	14	108	138	34	46	700	57.9	72.0	Screened; Ap. loc.
59-13-01	108	10	96	106	10	107	50	.52	--	Screened
59-13-02	152	6	152	200	45	100	35	.64	4.0	
59-13-03	--	6	105	107	35	56	12	.57	1.5	
59-13-04	--	6	81	85	32	48	10	.62	3.0	
59-14-01	115	8	112	330	48	76	222	8.20	11.0	Second aquifer ORGM; Ap.loc.
TOWNSHIP OF MAHWAH										
01-10-01	85	6	87	130	--	--	4	--	--	
02-10-01	25	6	26	102	19	40	20	.95	2.0	
02-10-02	28	6	33	100	15	20	25	5.00	2.0	
02-10-03	53	6	53	180	8	88	30	.38	2.0	
02-10-04	61	6	62	150	14	94	8	.10	1.5	
02-10-05	44	6	44	120	20	60	100	2.50	2.0	
03-10-01	23	8	23	308	20	220	225	1.12	2.0	
03-10-02	10	6	20	50	18	38	8	.40	2.0	
03-10-03	44	6	88	95	38	45	40	5.71	2.5	
03-11-01	11	6	21	125	5	25	15	.75	3.0	Bedrock is ORGM over PSSC
03-11-02	60	8	85	320	50	114	155	2.42	24.0	
03-11-03	11	8	20	150	30	140	35	.32	4.0	Bedrock is ORGM over PSSC
03-11-04	50	8	54	513	2	200	25	.13	8.0	Bedrock is ORGM over PSSC
03-12-01	103	6	103	265	60	250	7	.04	1.0	
03-12-02	--	6	--	97	--	--	15	--	--	Screened (?)
03-12-04	60	6	60	235	20	90	7	.10	1.0	
03-12-05	110	6	110	450	50	280	7	.03	8.0	
03-12-07	12	6	17	152	31	64	15	.45	2.5	
03-12-08	--	6	--	--	72	--	60	--	--	
03-12-09	--	6	--	--	65	--	8	--	--	Screened (?)
03-12-10	91	6	91	130	15	31	20	1.25	4.0	
03-12-11	--	6	92	93	20	30	10	1.00	4.0	
03-12-12	30	6	30	114	30	50	10	.50	3.0	
03-12-16	21	6	21	270	60	250	8	.04	5.0	Ap. loc.
03-13-06	7	6	21	175	42	125	10	.12	5.0	
03-13-07	--	6	--	--	5	15	25	2.50	1.5	Screened (?)
03-13-08	34	6	34	134	66	72	10	1.67	2.5	
03-13-09	16	6	16	254	46	90	7	.16	6.0	
03-13-13	98	6	98	100	10	15	20	4.00	3.0	Ap. loc.
03-13-14	53	6	53	135	50	80	15	.50	4.0	Ap. loc.
03-13-15	21	6	21	90	2	25	15	.65	6.0	Ap. loc.
03-14-01	3	6	21	125	Flowing	50+	14	<.28	5.0	Ap. loc. (9)
04-09-02	34	6	35	90	8	32	9	.38	1.5	
04-09-03	54	7	57	170	37	100	10	.16	0.5	Ap. loc.

Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

NJWRD well number	Location code\1	Owner	Local name or number\2	Latitude	Longitude	Primary use of water\3	Aquifer code	Driller	Date completed	Altitude of land surface\4 (feet)
				TOWNSHIP OF MAHWAH (CONT.)						
03- 27	04-09-04	Ramsey Boro WD	Test 4	410440	740903	U	PSSC	Rinbrand	05-04-53	298.5
03- 12	04-09-07	Ramsey Boro WD	Cent Av 1	410447	740905	P	PSSC	--	--	300
03- 62	04-10-01	Dixon, Allen	1	410420	741045	H	PSSC	Rinbrand	12-27-57	290
03- 66	04-10-02	Helms, Thomas	1	410441	741058	H	PSSC	Wittwer, W.E.	10-04-58	260
03- 67	04-10-03	Degenaars, Isaac	1	410446	741057	H	PSSC	Sikkema Co	12-29-68	250
03- 68	04-10-04	Gesner, Robert	1	410452	741046	H	PSSC	Wittwer, W.E.	10-05-66	265
03- 69	04-10-05	Havermeyer, H.O.	1	410456	741048	H	PSSC	Rinbrand	10-22-71	270
03-205	04-10-06	Ramapo Vall. College	USGS TB2	410440	741043	U	SFDF	U.S.G.S.	06-09-83	262.0L
03- 60	04-11-01	Immac Conception	Seminary 3	410412	741141	H	PRGM	Rinbrand	10-12-57	290
03- 04	04-11-02	Immac Conception	Seminary 1	410414	741117		PSSC	- -	10-19-49	340
03-152	04-11-03	Immac Conception	Seminary 2	410417	741114	H	PSSC	Rinbrand	11-01-48	360.2
03-156	04-11-04	Taub, Arnold	1	410428	741141	H	PSSC	Sikkema Co	01-09-78	286
03-206	04-11-05	Bergen Cn Park	USGS Test 3	410459	741121	U	SFDF	U.S.G.S.	06-09-83	253.6L
03-210	04-11-06	Charles Elmes	USGS Test 7	410450	741110	U	SFDF	U.S.G.S.	06-10-83	254.7L
03- 54	04-12-01	Jordon, Robert C	1	410401	741245	H	PCMB	Burrows, Inc	04-16-63	290
03- 76	04-12-02	Richter, Arthur	3122	410400	741243	H	PCMB	DF Well	04-05-66	270
03- 61	04-12-03	Wehran, Fred	1	410419	741212	H	ORGM	Sikkema Co	10-10-65	250
03-136	05-08-01	Mahwah Twp WD	Old Sta TW 2	410537	740857	U	SFDF	Burrows, Inc	09-28-70	281.9L
03-134	05-08-02	Mahwah Twp WD	E Ramapo Test	410538	740853	U	SFDF	Burrows, Inc	08-26-70	285.5L
03- 22	05-08-03	Mahwah Twp WD	Old Station	410539	740853	U	SFDF	Burrows, Inc	02-05-71	280
03- 71	05-09-01	Morgan, Carrie	1	410520	740937	H	PSSC	Ives & Sons	07-29-58	400
03-148	05-09-02	Twin Bar	1	410557	740950	C	SFDF	Wittwer, W.E.	05-05-60	265.2L
03-207	05-09-03	Suburban Propane	USGS Test 4	410559	740946	U	SFDF	U.S.G.S.	06-09-83	270.3L
03- 70	05-10-01	Sioux Lane Builders	1	410500	741053	H	PSSC	Wittwer, W.E.	08-12-68	260
03-138	05-10-02	Mahwah Twp WD	Test 17	410506	741055	U	SFDF	De Nure, W	07-25-81	250.9L
03-139	05-10-03	Mahwah Twp WD	Ford 2	410536	741023	P	SFDF	Rinbrand	00-00-53	264.4
03-135	05-10-04	Mahwah Twp WD	Ford Test 1	410537	741022	U	SFDF	Rinbrand	11-15-59	255
03-137	05-10-05	Mahwah Twp WD	Ford Test 2	410537	741023	U	SFDF	Rinbrand	10-06-60	260.2
03- 19	05-10-07	Mahwah Twp WD	Ford 4	410541	741023	P	SFDF	Rinbrand	00-00-53	254
03-140	05-10-08	Mahwah Twp WD	Ford 3	410543	741023	P	SFDF	Rinbrand	00-00-53	262.6
03-204	05-10-09	Ramapo Vall. College	USGS Test 1	410523	741014	U	SFDF	U.S.G.S.	06-08-83	270.6L
03-209	05-10-10	Mary B. Patrick	USGS Test 6	410528	741045	U	SFDF	U.S.G.S.	06-10-83	249.0L
03-211	05-10-11	Mahwah Twp WD	USGS TB8	410506	741055	U	SFDF	U.S.G.S.	09-03-83	250
03- 13	05-10-06	Mahwah Twp WD	Ford 1	410539	741023	P	SFDF	Rinbrand	00-00-53	254
03-120	05-10-20	Mahwah Twp WD	17	410506	741055	P	SFDF	Rinbrand	10-29-82	250.9L
03-212	05-10-21	Jersey Boring Co	J37-26	410551	741012	U	SFDF	Jersey Boring	09-15-83	256.2
03-214	05-10-22	Jersey Boring Co	J37-19	410550	741009	U	SFDF	Jersey Boring	09-13-83	254.9
03-213	05-10-23	Jersey Boring Co	J37-10	410549	741011	U	SFDF	Jersey Boring	09-09-83	256.1
03- 23	05-11-01	Mahwah Twp WD	MTWD 16	410505	741101	P	SFDF	Rinbrand	03-26-80	251.2L
03- 24	05-11-04	Mahwah Twp WD	MTWD T-R1	410505	741101	U	PSSC	Rinbrand	11-22-78	251 L
03- 5	05-13-01	Boy Scouts of Amer	Camp Yah Paw	410514	741346	T	PCMB	- -	00-00-36	800
03-147	06-08-01	American Brake Shoe	3	410604	740855	N	SFDF	Stothoff, W	09-09-53	276.9
03-146	06-08-02	American Brake Shoe	2	410610	740859	N	PSSC	Stothoff, W	08-01-53	271.3
03-145	06-09-01	American Brake Shoe	1	410607	740904	N	PSSC	Stothoff, W	06-24-53	282.4
03-170	06-09-02	Ford Motor Co	7013	410604	740942	U	SFDF	George Inc	08-18-81	267.6L
03-169	06-09-03	Ford Motor Co	7012	410608	740937	U	SFDF	George Inc	08-14-81	270.2L
03-168	06-09-04	Ford Motor Co	7011	410613	740932	U	SFDF	George Inc	08-13-81	268.2L
03-167	06-09-05	Ford Motor Co	7010	410615	740933	U	SFDF	George Inc	08-13-81	268.8L
03-166	06-09-06	Ford Motor Co	7009	410617	740938	U	SFDF	George Inc	08-13-81	284.6L
03-171	06-09-07	Ford Motor Co	7014	410622	740953	U	SFDF	George Inc	08-21-81	287.57
03-172	06-09-08	Ford Motor Co	7042	410630	740937	U	SFDF	George Inc	08-25-81	288.07
03-208	06-09-11	Mahwah Twp WD	USGS TB5	400616	740915	U	SFDF	U.S.G.S.	06-10-83	267.0L
03- 72	06-10-01	Sachs, Arthur	1	410605	741050	H	PCMB	Richardson,E	11-30-51	535
03- 73	06-11-01	Mills, Julius U	1	410659	741104	H	PCMB	Algeier Bros	08-29-52	820

See footnotes at end of table.

Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

Location code	Depth to bedrock from land surface (feet)	Minimum casing diameter (inches)	Depth of interval below land surface\5 Top (feet)	Depth of interval below land surface\5 Bottom (feet)	Well acceptance test\6 Depth of water level below land surface Static (feet)	Well acceptance test\6 Depth of water level below land surface Pumping (feet)	Well acceptance test\6 Yield\7 (gal/min)	Well acceptance test\6 Specific capacity ((gal/min)/ft)	Well acceptance test\6 Length of test (hours)	Remarks\1
TOWNSHIP OF MAHWAH (CONT.)										
04-09-04	105	6	105	107	96	--	--	--	--	
04-09-07	--	--	--	450	--	--	--	--	--	
04-10-01	25	6	25	115	3	10	10	1.43	4.0	
04-10-02	36	6	37	71	4	9	50	10.0	1.5	
04-10-03	75	6	75	400	10	60	10	.20	3.0	
04-10-04	83	6	85	148	16	90	7	.09	2.0	
04-10-05	90	6	90	210	20	60	25	.62	4.0	
04-10-06	40	--	--	35	--	--	--	--	--	U.S.G.S. test boring
04-11-01	55	6	55	110	20	50	4	.13	2.0	
04-11-02	--	8	38	415	21	140	150	1.2	8.0	(9)
04-11-03	61	8	61	435	30	150	110	.92	8.0	(9)
04-11-04	5	6	30	580	60	250	2	.01	1.0	Driller's log--PCMB over PSSC
04-11-05	35	1	29	30	11	--	--	--	--	U.S.G.S. test well (4,8)
04-11-06	--	1	24	25	9	--	--	--	--	U.S.G.S. test well (4,8)
04-12-01	42	6	42	133	--	--	2	--	--	
04-12-02	83	6	90	122	43	80	22	.59	1.5	
04-12-03	43	6	43	100	10	50	100	2.50	4.0	
05-08-01	90	8	75	90	14	57	697	16.53	72.0	Screened (8)
05-08-02	88	8	78	88	15	78	115	1.82	1.0	Screened
05-08-03	--	12	71	86	16	65	1125	22.96	72.0	Screened (8)
05-09-01	20	6	24	60	27	--	30	--	--	
05-09-02	116	6	65	70	5	40	20	.57	32.0	Screened (8)
05-09-03	86	1	65	66	10	--	--	--	--	U.S.G.S. test well (4)
05-10-01	105	6	106	125	17	20	30	10.00	3.0	
05-10-02	168	8	142	164	10	98	465	5.28	72.0	Screened
05-10-03	--	12	110	140	16	96	688	8.60	72.0	Screened (9)
05-10-04	115	--	--	--	16	72	310	5.54	8.0	Screen pulled after test (8)
05-10-05	120	8	110	120	18	54	516	14.33	--	Screened (8)
05-10-07	--	18	70	95	--	--	1230	--	--	Screened (9)
05-10-08	--	18	65	95	8	58	420	8.40	72.0	Screened (9)
05-10-09	--	1	36	37	18	--	--	--	--	U.S.G.S. test well (4,8)
05-10-10	--	1	36	37	3	--	--	--	--	U.S.G.S. test well (4)
05-10-11	--	--	--	125	--	--	--	--	--	U.S.G.S. test boring
05-10-06	--	18	83	103	--	--	700	--	--	Screened (9)
05-10-20	--	14	146	169	9	96	560	6.44	72.0	Screened
05-10-21	--	--	--	--	--	--	--	--	--	Test boring for I-287
05-10-22	--	4	--	--	--	--	--	--	--	Test boring for I-287
05-10-23	--	--	--	--	--	--	--	--	--	Test boring for I-287
05-11-01	--	14	116	149	25	54	900	31.01	80.0	Screened (8)
05-11-04	95	8	66	145	30	143	5	.04	8.0	TW for 16. Hole filled
05-13-01	<5	6	60	152	--	--	8	--	--	Ap. loc. (9)
06-08-01	39	10	29	39	11	30	170	8.95	8.0	Screened (9)
06-08-02	71	10	84	301	11	--	5	--	--	(4,9)
06-09-01	83	10	91	203	--	--	18	--	--	(9)
06-09-02	--	3	8	27	12	--	--	--	--	
06-09-03	--	3	7	27	11	--	--	--	--	
06-09-04	--	3	5	25	10	--	--	--	--	
06-09-05	--	3	6	25	10	--	--	--	--	
06-09-06	--	3	8	27	24	--	--	--	--	Screened
06-09-07	18.5	3	4	16	--	--	--	--	--	
06-09-08	--	3	28	48	--	--	--	--	--	Screened (8)
06-09-11	27	--	--	27	--	--	--	--	--	U.S.G.S. test boring
06-10-01	11	6	11	100	25	50	25	1.00	1.0	Ap. loc.
06-11-01	0	6	12	90	10	50	10	.25	6.0	Ap. loc.

Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

NJWRD well number	Location code\1	Owner	Local name or number\2	Latitude	Longitude	Primary use of water\3	Aquifer code	Driller	Date completed	Altitude of land surface\4 (feet)
				BOROUGH OF OAKLAND						
03-193	00-14-01	Oakland Industrial	Steiner 1	410009	741443	I	FLVL	Ackerman Co	01-10-69	437
03-192	00-14-02	Oakland Industrial	Park 1	410009	741440	I	SFDF	Ackerman Co	04-01-65	420
03-189	00-14-03	Witco Chemical Co	1	410013	741433	I	SFDF	Ackerman Co	09-15-66	450
03-190	00-15-01	Molly's Fish Market	1	410052	741517	A	FLVL	Burrows, Inc	03-01-55	235
03-180	00-15-02	Muller Park	1	410057	741525	R	SFDF	Rinbrand	05-16-73	218
03-184	01-13-01	Oakland Boro WD	OBWD 9 (Rt 208)	410134	741349	P	SFDF	Rinbrand	08-01-67	310
03- 43	01-13-02	Carafa, Dolcino	1	410146	741300	H	PSSC	Rinbrand	07-01-51	430
03-185	01-14-01	Oakland Boro WD	TW2 Oak St	410127	741441	U	SFDF	Rinbrand	08-20-68	263.4
03- 44	01-14-02	Oakland Boro WD	TW Oak St	410127	741449	U	SFDF	Rinbrand	10-04-68	270
03-188	01-14-03	Raritan Plastics	1	410130	741418	N	ORGM	Rinbrand	03-06-62	290
03-181	01-14-04	Oakland Boro WD	Pine St 1	410139	741453	P	SFDF	Rinbrand	08-26-31	266.6L
03-191	01-15-01	Pleasure Swim Club	1	410100	741536	R	SFDF	Ives & Son	05-04-72	210
03-186	01-15-02	Oakland Boro WD	TW Spruce St	410108	741520	U	SFDF	Rinbrand	08-22-67	212.9L
03-187	01-15-03	Long Hill Plaza Inc	1	410113	741528	R	SFDF	Ackerman Co	01-10-72	212
03- 14	01-15-04	Oakland Boro WD	Bush 5	410125	741504	P	SFDF	- -	00-00-54	220
03- 94	01-15-05	Oakland Boro WD	Bush 4	410126	741508	P	SFDF	Rinbrand	09-01-50	220
03- 45	02-13-01	Steidten, George	1	410227	741338	H	ORGM	Wittwer, W.E.	03-07-57	300
03- 58	02-15-01	Smyrychyn, Steven	1	410203	741502	H	PCMB	Rinbrand	05-29-58	375
03- 6	03-13-01	Oakland Boro WD	Soons 7	410301	741327	P	SFDF	- -	08-06-57	250
03-182	03-13-02	Oakland Boro WD	Soons 6	410301	741330	P	SFDF	Rinbrand	04-14-56	238.2
03- 9	03-13-03	Oakland Boro WD	Soons 8	410302	741330	P	SFDF	- -	04-11-59	240
03-183	03-13-04	Oakland Boro WD	Soons 9	410303	741328	U	SFDF	Rinbrand	09-18-58	230.4L
03- 46	03-13-05	Koedam, Cornelis	1	410303	741313	H	FLVL	Sikkema Co	02-10-69	235
				BOROUGH OF RAMSEY						
03- 16	02-09-01	Ramsey Boro WD	Martis Av Well	410244	740927	P	PSSC	- -	01-01-56	350
03-175	03-07-01	Ramsey Golf & CC	1	410300	740756	R	PSSC	Peerless Co	03-24-66	358.
03-173	03-07-02	Ramsey Boro WD	Dixon St	410328	740744	P	PSSC	Burrows, Inc	03-26-65	379 L
03-178	03-08-01	Ramsey Boro WD	Test 1	410351	740841	U	SFDF	Rinbrand	03-24-53	349.5
03-176	03-09-01	Berthold, Charles A	1	410342	740942	H	PSSC	Nann, William	05-10-54	366.5
03-179	03-09-02	Fecanin, John	1	410347	740937	H	PSSC	Rinbrand	06-29-52	395.5
03- 15	03-09-03	Ramsey Boro WD	Woodland Well	410317	740953	--	PSSC	- -	00-00-53	350
03- 17	03-09-04	Ramsey Boro WD	Dar Well	410334	740935	P	PSSC	- -	00-00-56	355
03-177	04-07-01	Kuncik, John	1	410404	740736	H	SFDF	Wittwer, W.E.	08-06-53	425
03- 33	04-07-02	Ramsey Boro WD	TW Airmont Ave	410415	740735	P	PSSC	Rinbrand	12-12-63	450
03- 34	04-08-01	Ramsey Boro WD	Spring St	410426	740835	P	PSSC	Rinbrand	02-03-67	345
03-174	04-09-01	Ramsey Boro WD	Elbert St	410403	740900	U	PSSC	Burrows, Inc	05-20-65	340.3

See footnotes at end of table.

Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

Location code	Depth to bedrock from land surface (feet)	Minimum casing diameter (inches)	Depth of interval below land surface\5		Well acceptance test\6					Remarks\1
					Depth of water level below land surface		Yield\7 (gal/min)	Specific capacity ((gal/min)/ft)	Length of test (hours)	
			Top (feet)	Bottom (feet)	Static (feet)	Pumping (feet)				
					BOROUGH OF OAKLAND					
00-14-01	145	6	156	252	75	200	55	.44	6.0	
00-14-02	--	8	59	69	6	60	119	2.20	8.0	Screened
00-14-03	--	6	80	90	35	45	150	15	8.0	Screened
00-15-01	51	6	51	115	27	100	60	.82	3.0	(9)
00-15-02	--	8	42	52	12	41	100	3.45	--	Screened; Ap. loc.
01-13-01	93	12	85	95	-25	125	582	3.88	72.0	Screened; Second aquifer PSSC
01-13-02	70	6	70	135	38	40	5	2.50	--	
01-14-01	150	6	90	100	58	88	49	1.63	8.0	Screened
01-14-02	149	6	133	143	43	58	310	20.67	24.0	Screened
01-14-03	24	8	31	333	8	250	42	.17	24.0	Driller's log--ORGM over PCMB
01-14-04	--	8	175	190	60	130	125	1.78	24.0	Screened; Not in service (8)
01-15-01	--	8	40	50	10	18	65	8.12	0.5	(8)
01-15-02	--	12	65	91	6	16	1016	101.60	24.0	Screened (8)(9)
01-15-03	--	8	81	98	35	46	405	36.82	8.0	Screened
01-15-04	--	12	108	128	--	--	1160	--	--	Screened
01-15-05	--	6	112	126	5	50	200	4.44	16.0	Screened (9)
02-13-01	49	6	50	280	25	150	3	.02	1.5	Ap. loc.
02-15-01	20	6	20	300	50	300	2	.01	5.0	
03-13-01	76	12	63	83	10	59	419	8.55	12.0	Screened (9)
03-13-02	93	12	85	93	14	78	308	4.81	--	Screened (9)
03-13-03	--	12	87	112	7	14	970	138.57	16.0	Screened (9)
03-13-04	84	12	72	84	14	60	171	3.72	14.5	Screened; observation well(8,9)
03-13-05	60	6	60	210	5	30	10	.40	5.0	
					BOROUGH OF RAMSEY					
02-09-01	--	10	--	--	--	--	200	--	--	Ap. loc. (9)
03-07-01	120	7	132	348	30	285	80	.31	8.0	
03-07-02	105	12	103	400	3	89	151	1.76	3.5	(9)
03-08-01	--	6	99	115	20	90	15	.21	2.0	Casing pulled, Hole filled.
03-09-01	32	6	39	160	28	46	23	1.28	4.0	
03-09-02	46	6	47	140	40	60	10	.50	--	
03-09-03	<47	10	47	300	2	157	200	1.5	24.0	Ap. loc. (9)
03-09-04	21	10	66	400	2	170	220	1.3	34.0	Ap. loc. (9)
04-07-01	--	6	55	60	20	53	11	.33	72.0	Screened
04-07-02	100	10	120	500	50	245	105	.54	72.0	Ap. loc. (9)
04-08-01	84	10	104	600	-5	200	250	1.22	72.0	Flows naturally. Ap.loc.
04-09-01	82	10	104	400	2	271	151	.56	72.0	Test well (9)

Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

NJWRD well number	Location code\1	Owner	Local name or number\2	Lati-tude	Longi-tude	Primary use of water\3	Aqui-fer code	Driller	Date completed	Altitude of land surface\4 (feet)
NY WRD Sequential Number/10			ROCKLAND COUNTY-NEW YORK							
01	06-08-03	Imperial Laundry	1	410655	740857	C	SFDF	- -	- -	270.0L
01	06-08-04	Avon Allied Products	1	410656	740856	U	PSSC	- -	00-00-39	310
02	06-09-09	Suffern WD	2	410657	740932	U	SFDF	Artesian	09-02-36	274
01	06-09-10	Suffern WD	4	410659	740937	U	SFDF	Artesian	00-00-74	277.2L
01	06-09-14	Suffern WD	3	410654	740928	P	SFDF	Artesian	01-15-37	272
01	07-09-01	Spring Valley WC	TW	410710	740937	U	SFDF	Layne	00-00-73	278.0L
01	07-09-02	Spring Valley WC	TW 20	410712	740936	U	SFDF	Layne	07-18-73	273 L
01	07-09-03	Spring Valley WC	TW 19	410716	740939	U	SFDF	Layne	07-16-73	272.7L
01	07-09-04	Spring Valley WC	TW 17	410722	740942	U	SFDF	Layne	07-02-73	276 L
01	07-09-05	Spring Valley WC	TW 15	410730	740948	U	SFDF	Layne	06-26-73	284 L
01	07-09-06	Spring Valley WC	TW 13	410750	740959	U	SFDF	Layne	06-14-73	287.9L

(1) The first two digits of the location code are the minutes of latitude of the well location. The middle two digits of the location code are the minutes of longitude of the well location. These two numbers define a square in the grid on plate 1. Wells in each square are numbered sequentially; this sequential number is the last two digits of the location code. Well locations were determined by field reconnaissance except when "Ap. loc." (approximate location) appears as a remark; in this case, the well data were used only in the preparation of figure 4.

(2) Local name or number appears as in the U.S. Geological Survey Ground-Water Site Inventory (GWSI) data base, except that some words are abbreviated because of space limitations.

(3) A, air conditioning; C, commercial; H, domestic; I, irrigation; P, public supply; R, recreation; T, institutional; U, unused.

(4) Generally determined from 10-foot-contour topographic maps with an accuracy of $\pm$ 5 feet. Altitudes reported to the nearest tenth of a foot were either measured by altimeter (accurate to the nearest foot), or they were calculated by subtracting the height of the well above land surface from the altitude of the top of the well as determined by standard surveying techniques (accurate to the nearest 0.1 foot). The latter wells are indicated by an L.

(5) The interval is an open hole unless otherwise noted in the remarks column. The total drilled depth generally is within 2 feet of the lower end of the opening. Values are rounded to the nearest foot.

(6) Generally measured within 2 weeks of well-completion date. Water levels are rounded to the nearest foot.

(7) Maximum short-term yield commonly exceeds reported value. Reported values are rounded to the nearest gallon per minute.

(8) Used for measurement of potentiometric-surface levels (see plate 3).

(9) Reported in Vecchioli and Miller, 1973, table 1.

(10) The New York USGS identifier for these wells is a 15-digit station number that consists of the latitude and longitude of the well location separated by a zero, followed by a sequential number. The sequential number is used to distinguish among multiple wells located at the same latitude and longitude.

(11) Abbreviations used: Amer, America; Ave, Avenue; Boro, Borough; Bros, Brothers; CC, Country Club; Cn, County; Co, Company; Comm, Commission; Dar, Darlington; Frnkln, Franklin; Immac, Immaculate; Inc, Incorporated; Lks, Lakes; Rac, Racquet; St, Street; Ter, Terrace; TW, Test Well; Twp, Township; USGS (or U.S.G.S.) U.S. Geological Survey; Vall, Valley; WC, Water Company; WD, Water Department.

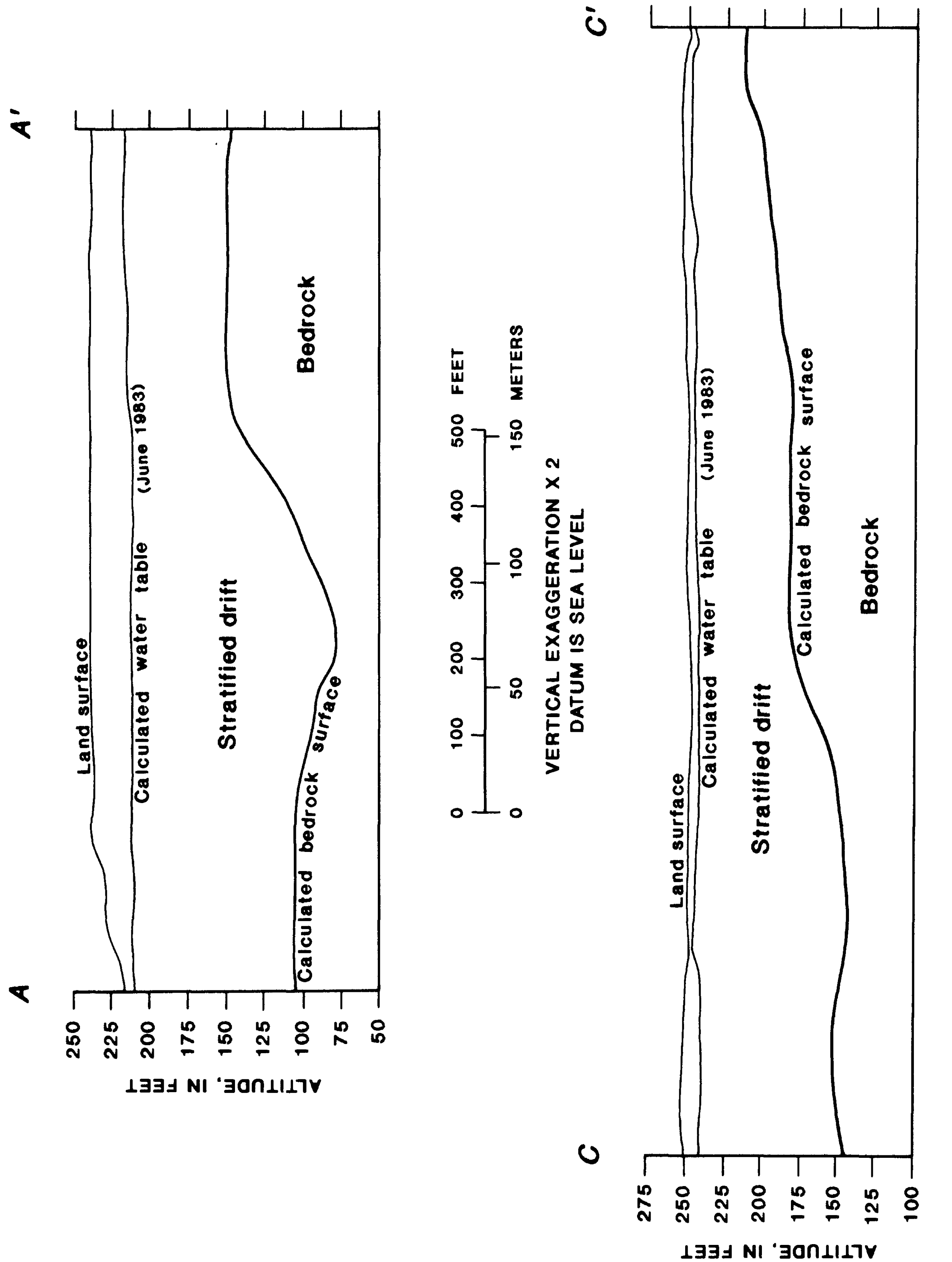
Table 1.--Construction features and yield characteristics of selected wells, test holes, and deep potentiometers in and near the Ramapo River valley, New Jersey and New York--Continued

Location code	Depth to bedrock from land surface (feet)	Minimum casing diameter (inches)	Depth of interval below land surface\5 Top (feet)	Bottom (feet)	Well acceptance test\6 Depth of water level below land surface Static (feet)	Pumping (feet)	Yield\7 (gal/min)	Specific capacity ((gal/min)/ft)	Length of test (hours)	Remarks\1
					ROCKLAND COUNTY NEW YORK					
06-08-03	--	--	--	--	--	--	--	--	--	Dug well (8)
06-08-04	108	8	123	718	14	--	--	--	--	
06-09-09	--	16	50	97	4	17	1350	103.8	7.0	Screened
06-09-10	155	--	110	120	--	--	--	--	--	Screened
06-09-14	--	8	66	95	3	--	400	--	--	Screened
07-09-01	--	--	--	--	--	--	--	--	--	Screened (8)
07-09-02	74.4	2.5	49	68	8	--	56	--	--	Screened (8)
07-09-03	124.9	2.5	115	118	8	--	13	--	--	Screened (8)
07-09-04	131.5	2.5	116	120	9	--	56	--	--	Screened (8)
07-09-05	97.8	2.5	64	68	17	--	51	--	--	Screened (8)
07-09-06	58.9	2.5	49	53	8	--	85	--	--	Screened (8)

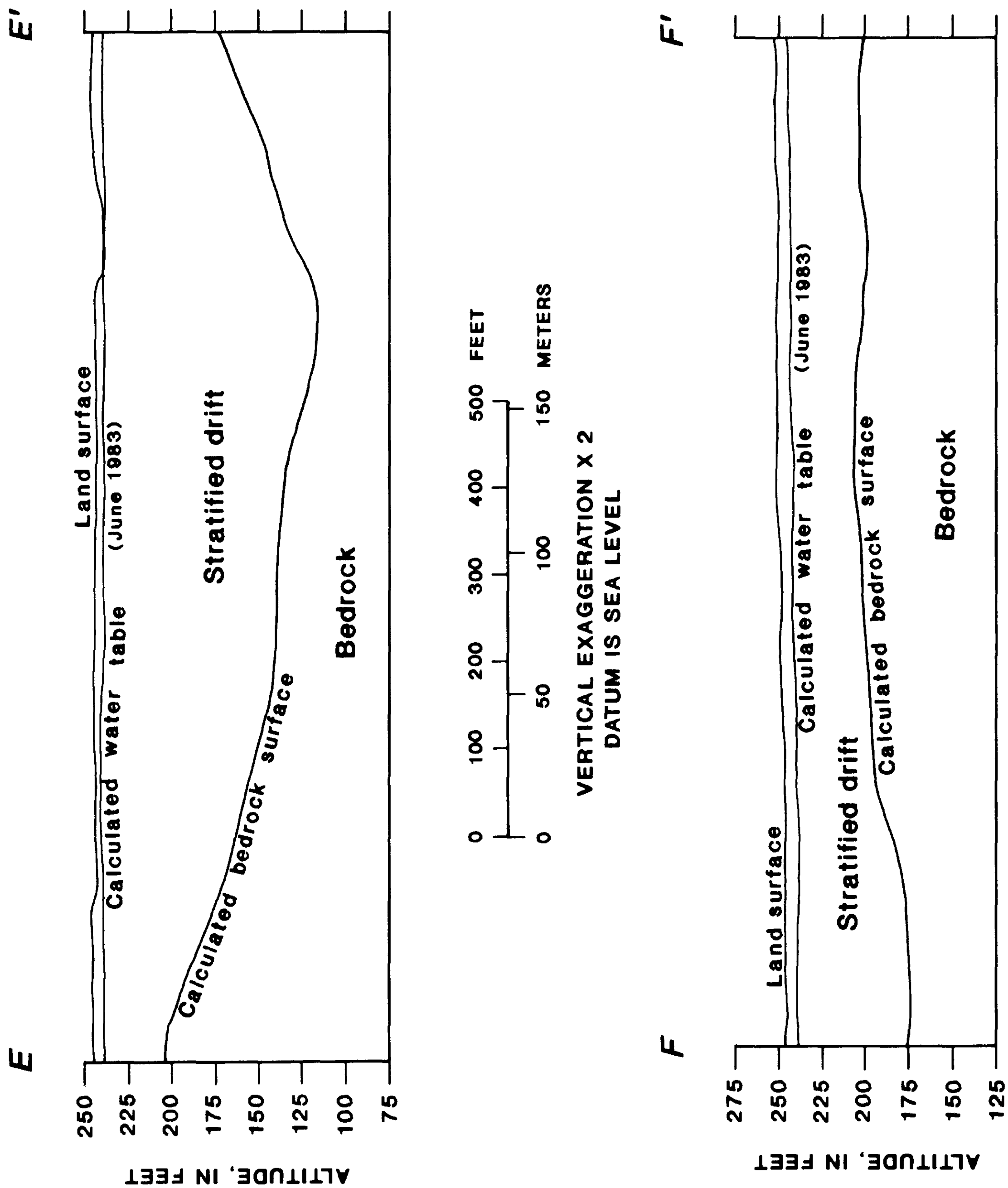
APPENDIXES

Appendix A.--Seismic-Refraction Profiles

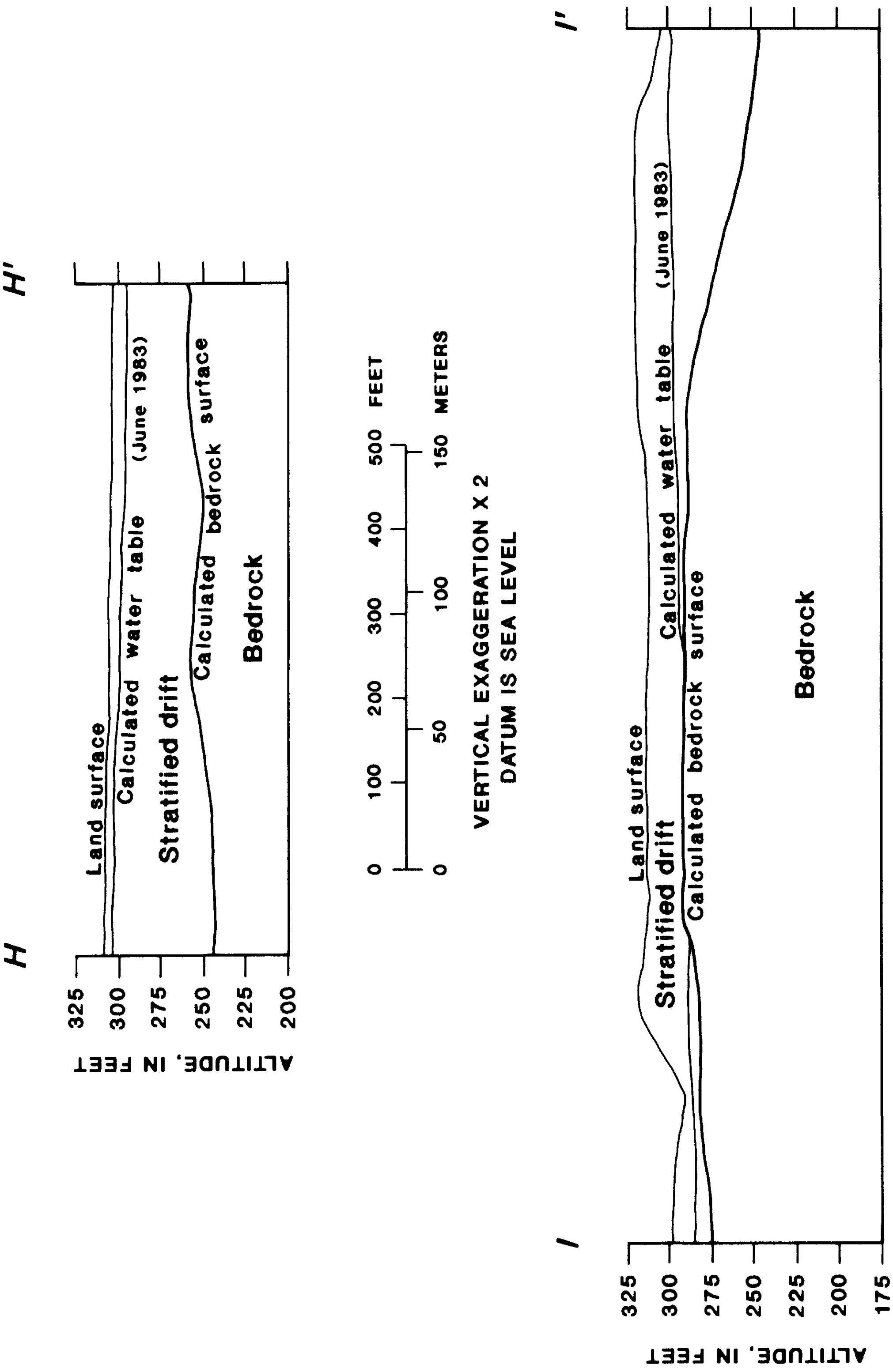
These hydrogeologic sections were constructed from seismic-refraction surveys conducted by the U.S. Geological Survey in June 1983. Locations of individual profiles are shown on plate 2. Interpretation of field geophysical data is based on a computer program described by Scott and others (1972).



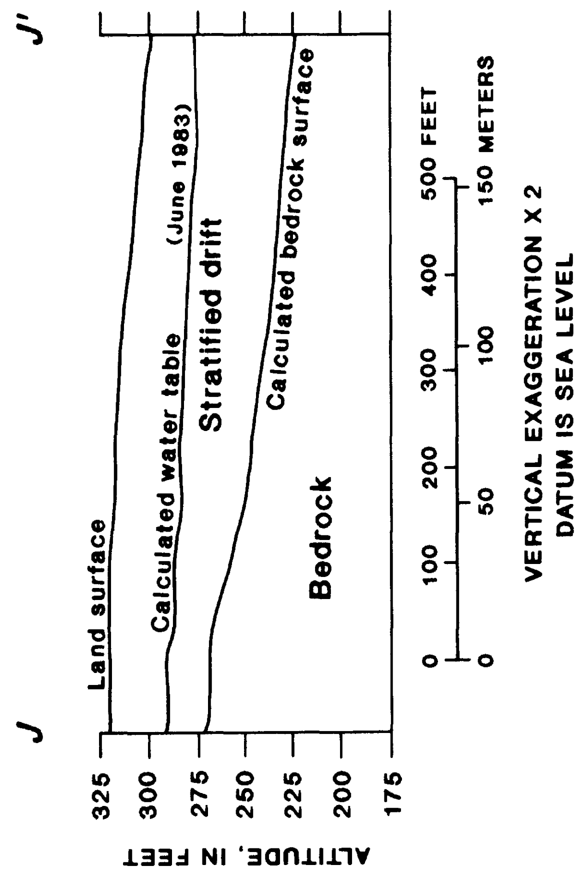
Appendix A.--Seismic-Refraction Profiles--Continued.
Locations of individual profiles are shown
on plate 2.



Appendix A.--Seismic-Refraction Profiles--Continued.
Locations of individual profiles are shown
on plate 2.



Appendix A.--Seismic-Refraction Profiles--Continued.
Locations of individual profiles are shown
on plate 2.



Appendix B.--Aquifer-test drawdown data: Mahwah Township Water Department production wells 16 and 17 and test well 17

Drawdown data from three aquifer tests conducted in the confined aquifer southwest of Mahwah, New Jersey, are listed below. These aquifer tests are described on p. 20. Three wells screened in the confined aquifer were used in these tests:

	NJ WRD well number (see table 1)	Well-location code (see table 1 and plate 1)
Mahwah Township Water Department production well 16	03- 23	05-11-01
Mahwah Township Water Department production well 17	03-120	05-10-20
Mahwah Township Water Department test well 17	03-138	05-10-02

A schematic diagram of the relative locations of these wells is shown in figure 9.

Drawdown data from aquifer test 1

[Production well 17 was pumped at about 503 gallons per minute for the first 12 minutes, then at 560 gallons per minute thereafter; production well 16, at a distance of 490 feet, was observed]

Time since start of pumping (minutes)	Corrected drawdown (feet)	Time since start of pumping (minutes)	Corrected drawdown (feet)	Time since start of pumping (minutes)	Corrected drawdown (feet)
0.1	0.00	90.0	1.93	1686.0	2.40
1.0	0.00	100.0	1.94	1752.0	2.40
2.0	0.02	110.0	1.98	1797.0	2.40
3.0	0.16	120.0	1.99	1932.0	2.41
3.5	0.32	130.0	2.02	2049.0	2.44
4.0	0.40	147.0	2.07	2118.0	2.47
4.5	0.47	150.0	2.08	2174.0	2.47
5.0	0.53	162.0	2.05	2233.0	2.47
5.5	0.58	165.0	2.06	2290.0	2.48
6.0	0.63	170.0	2.06	2352.0	2.48
6.5	0.68	182.0	2.07	2415.0	2.49
7.0	0.74	202.0	2.08	2473.0	2.50
7.5	0.77	222.0	2.11	2531.0	2.49
8.0	0.79	240.0	2.12	2591.0	2.51
8.5	0.84	270.0	2.13	2655.0	2.51
9.0	0.87	298.0	2.15	2715.0	2.51
9.5	0.90	328.0	2.17	2775.0	2.52
10.0	0.94	362.0	2.18	2836.0	2.55
10.5	0.96	391.0	2.19	2890.0	2.56
11.33	1.01	423.0	2.20	2960.0	2.54
12.38	1.04	451.0	2.21	3010.0	2.54
13.0	1.07	486.0	2.22	3077.0	2.54
14.0	1.12	513.0	2.23	3130.0	2.54
15.0	1.14	544.0	2.24	3176.0	2.54
16.0	1.18	575.0	2.26	3253.0	2.55
17.0	1.21	608.0	2.26	3309.0	2.55
18.0	1.24	674.0	2.29	3360.0	2.56
19.0	1.28	740.0	2.31	3431.0	2.58
21.0	1.37	797.0	2.32	3491.0	2.58
23.0	1.42	852.0	2.33	3555.0	2.59
23.5	1.43	915.0	2.34	3608.0	2.59
25.0	1.47	980.0	2.37	3669.0	2.61
28.0	1.53	1043.0	2.37	3729.0	2.59
29.0	1.55	1095.0	2.36	3793.0	2.60
34.0	1.64	1157.0	2.37	3850.0	2.62
39.0	1.68	1216.0	2.38	3911.0	2.60
44.0	1.72	1277.0	2.36	3972.0	2.60
49.0	1.78	1334.0	2.38	4034.0	2.60
54.0	1.80	1395.0	2.41	4092.0	2.60
60.0	1.82	1457.0	2.39	4157.0	2.60
66.0	1.85	1511.0	2.38	4212.0	2.60
71.0	1.86	1557.0	2.38	4276.0	2.61
80.0	1.90	1625.0	2.39	4320.0	Stopped pumping

Drawdown data from aquifer test 2

[Production well 17 was pumped at 703 gallons per minute; production well 16, at a distance of 490 feet, was observed]

Time since start of pumping (minutes)	Corrected drawdown (feet)	Time since start of pumping (minutes)	Corrected drawdown (feet)	Time since start of pumping (minutes)	Corrected drawdown (feet)
0.1	0.02	7.5	1.16	22.0	1.87
.5	.06	8.0	1.20	24.0	1.91
1.0	.22	8.5	1.26	26.0	1.95
1.5	.34	9.0	1.29	28.0	1.99
2.0	.47	9.5	1.33	30.0	2.02
4.0	.84	10.0	1.38	35.0	2.10
5.0	.87	12.0	1.49	41.0	2.18
5.5	1.00	14.0	1.59	45.0	2.19
6.0	1.03	16.0	1.67	50.0	2.24
6.5	1.08	18.0	1.74	55.0	2.28
7.0	1.11	20.0	1.81		

Drawdown data from aquifer test 3

[Production well 16 was pumped at 620 gallons per minute; test well 17, at a distance of 501 feet, was observed; data provided by R. J. Canace (New Jersey Department of Environmental Protection and Energy, written commun., 1982)]

Time since start of pumping (minutes)	Corrected drawdown (feet)	Time since start of pumping (minutes)	Corrected drawdown (feet)	Time since start of pumping (minutes)	Corrected drawdown (feet)
0.0	0.00	9.0	1.34	60.0	2.25
.5	.10	10.0	1.42	70.0	2.29
1.0	.23	12.0	1.53	84.0	2.32
1.5	.39	14.0	1.59	95.0	2.38
2.0	.52	16.0	1.67	100.0	2.39
2.5	.63	18.0	1.73	120.0	2.42
3.0	.72	20.0	1.78	142.0	2.45
3.5	.81	25.0	1.90	160.0	2.48
4.0	.88	30.0	1.97	180.0	2.49
4.5	.95	36.0	2.06	200.0	2.53
5.0	1.01	40.0	2.09	250.0	2.55
6.0	1.12	45.0	2.14	300.0	2.58
7.0	1.23	50.0	2.17	350.0	2.61
8.0	1.28	55.0	2.21		

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