New Jersey Geological and Water Survey
Open-File Report 16-1

Hydrogeologic Framework and Computer Simulation of Groundwater Flow in the Valley-Fill and Fractured-Rock Aquifers of the Germany Flats Area of Sussex County, New Jersey

New Jersey Department of Environmental Protection

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Cover photo: Water gage at Howells Pond, Andover Township, Sussex County. *Photo by Z. Allen-Lafayette*

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by

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FORWARD

This report is based on an unpublished 1995 master's thesis, "Ground-Water Flow in a Buried Valley and Effects of Quarry Dewatering on Howells Pond, Sussex County, New Jersey" by Laura J. Nicholson and is a published version of the master's thesis. It includes enhanced graphics, minor editorial changes to the report text, and additional stratigraphic information from wells drilled subsequent to 1995. The groundwater flow models and interpretation of modelling results documented in this report are identical to those in the thesis.

Throughout the report, the name "Limecrest Quarry" is used to denote the marble and granite quarry operated by Limecrest Products Corporation from 1919 to 1995. Subsequent to 1995, ownership of the quarry has changed several times. Additionally, groundwater withdrawal rates at the quarry have varied substantially, including times of little or no pumpage resulting in accompanying changes to groundwater and surface water levels in the quarry vicinity. Because this report presents groundwater flow conditions that existed in the Germany Flats area in the mid-1990s, changes in the flow regime that have occurred subsequent to that time are beyond the scope of this report.

HYDROGEOLOGIC FRAMEWORK AND COMPUTER SIMULATION OF GROUND-WATER FLOW IN THE VALLEY-FILL AND FRACTURED-ROCK AQUIFERS OF THE GERMANY FLATS AREA OF SUSSEX COUNTY, NEW JERSEY

ABSTRACT

The Germany Flats area encompasses parts of Sparta, Lafayette and Andover Townships in southern Sussex County, New Jersey. It is part of a buried valley that contains productive glacial-valley-fill and carbonate-fractured-rock aquifers. The headwaters of the Paulins Kill and Pequest River and several small lakes and ponds are located here. A large quarry operates on the eastern wall of the valley, withdrawing approximately 6.5 million gallons of water per day from the Franklin Marble and granite as part of the dewatering process.

In response to a projected increase in groundwater demand in this largely-rural area, this study was conducted to assess the hydrogeology and groundwater resource potential of the valley. It includes 1) development of the hydrogeologic framework of the surficial and bedrock aquifers, 2) assessment of the flow system and quantification of aquifer properties, and 3) development of a numerical groundwater-flow model of the surficial and fractured-rock aquifers to be used as a tool in evaluating effects of existing and potential groundwater withdrawals.

Results of the modeling indicate that dewatering activities at the rock quarry dominate the flow system, lowering the water level more than 100 feet in the quarry vicinity and shifting groundwater divides in both the valley-fill and carbonate rock aquifers. The pattern of drawdown in the valley-fill aquifer is greatly dependent upon the glacial stratigraphy. Where low-permeability glacial confining units are absent, pumping stress in the bedrock is more readily transmitted to the water-table aquifer.

Quarry dewatering also results in changes to the surface-water flow system. With the quarry wells pumping 6.5 mgd, modeling results indicate that 1) average annual streamflow in the East Branch Paulins Kill, which receives discharged water from the quarry, has increased by 2.2 ft^3/s in comparison to pre-development flow, 2) base flow in the modeled part of the Pequest River is reduced as groundwater that formerly discharged to the river is diverted by the quarry withdrawal, and 3) more stream water leaks to underlying aquifers and more losing stream reaches exist than in the pre-development valley due to induced stream-water infiltration in the large cone of depression surrounding the quarry pumping wells. Simulated drawdown in Howells Pond, a glacial kettle pond located 1.3 miles southwest of the quarry, is approximately 6 feet.

INTRODUCTION

Groundwater is the sole source of water supply for residents and industries in the part of southern Sussex County informally known as “Germany Flats” and encompassing parts of Sparta, Andover and Lafayette Townships (fig. 1). It is used for industrial, agricultural and domestic supply, including two golf courses, a nursing home, sand and gravel quarries, domestic residences and a rock quarry. The largest permitted groundwater withdrawal in this region, approximately 6.5 mgd, is for dewatering at the former Limecrest Quarry in Sparta Township. Water that accumulates on the quarry floor from groundwater seepage and precipitation is removed and discharged to an adjacent stream. The quarry has been in operation since 1919 and is mined primarily for the Franklin Marble, a calcium-carbonate-rich rock used for agricultural fertilizer and decorative stone.

The Statewide Water Supply Master Plan identified the Sparta area of Sussex County as a growth center requiring increased water supply from groundwater. In anticipation of increased water demand as a result of projected population growth, the New Jersey Geological and Water Survey conducted a study of the groundwater resource in the Germany Flats area. Funding for the study was allocated through the 1981 New Jersey Water Bond.

Dewatering at Limecrest Quarry exerts a large stress on the groundwater-flow system. An understanding of the flow system in the quarry vicinity is therefore key to assessing the regional availability of groundwater for public water supply. A 12.3 square mile area surrounding the quarry was chosen for detailed study and a numerical groundwater-flow model simulating flow in the valley-fill and fractured-rock aquifers was developed. The model was used to simulate groundwater flow both in the mid-1990's and in a "pre-development" scenario when no pumpage was present. By comparing modeling results from the two scenarios, an assessment was made of the effects of existing pumpage on groundwater levels and flow patterns, stream flow and the water level in Howells Pond, a small kettle pond in Andover Township. Particular attention is given to the pond as data indicate that the water level here dropped in the decades preceding the study as a result of groundwater withdrawals, an indication of the potential hydrologic impacts to surface water from groundwater development.

Purpose and Scope

This report presents 1) the hydrostratigraphic framework of the valley-fill and fractured-rock aquifers, 2) quantification of aquifer properties, 3) a numerical, three-dimensional, steady-state groundwater-flow model developed for the area surrounding Limecrest Quarry, 4) a simulation of groundwater flow under steady-state "average" flow conditions for the mid-1990's (the conditions that existed between 1990 and 1995 when field data for this study were obtained), 5) a simulation of groundwater flow under steady-state pre-development conditions, 6) a comparison of simulated mid-1990's and pre-development groundwater flow conditions, 7) an assessment of the impacts of quarry dewatering on the valley-fill and fractured-rock aquifers, surface water and stream flow, and 8) an assessment of the potential causes of the lowering of the water level in Howells Pond.

Previous investigations

Hydrogeologic investigations of the Germany Flats area were limited prior to this study. A regional investigation of groundwater resources of parts of Sussex and Warren County was conducted by Miller (1974). Harold E. Pellow and Associates, Inc. (1975, unpublished) conducted a water resource study of the Germany Flats area.

Drake and Volkert (1993) mapped the bedrock geology of the Newton East quadrangle in Sussex County. Surficial geology of the Kittatinny Valley and vicinity including southern Sussex County was mapped by Witte (1992). Subsequent to completion of this study, the surficial geology and earth materials of the Newton East quadrangle of Sussex County were mapped by Witte and Monteverde (2006).

Acknowledgements

Numerous people at the New Jersey Geological and Water Survey contributed to this study. These include Ron Witte, Jeff Hoffman, Jim Boyle, Ted Pallis, Bill Graff, and Walt Marzulli; and Dave Hall, Stewart Sandberg, Joe Rich, Evelyn Hall and Bob Canace (formerly of the NJGWS). Zehdreh Allen-Lafayette assisted greatly with preparation of figures for the report. The author wishes to thank all residents, business owners, and municipal officials within the project area that made this study possible by allowing access to their properties for data collection.

Location and description of study area

The study area is located in Sussex County, New Jersey in the New Jersey Highlands and Appalachian Valley and Ridge physiographic provinces and includes parts of Sparta, Andover and Lafayette Townships (fig. 1). It is a northeast-southwest oriented valley floored by carbonate bedrock overlain by glacial valley-fill deposits. The uplands are composed of slate and crystalline rocks. The modeled area encompasses 12.3 square miles within the larger

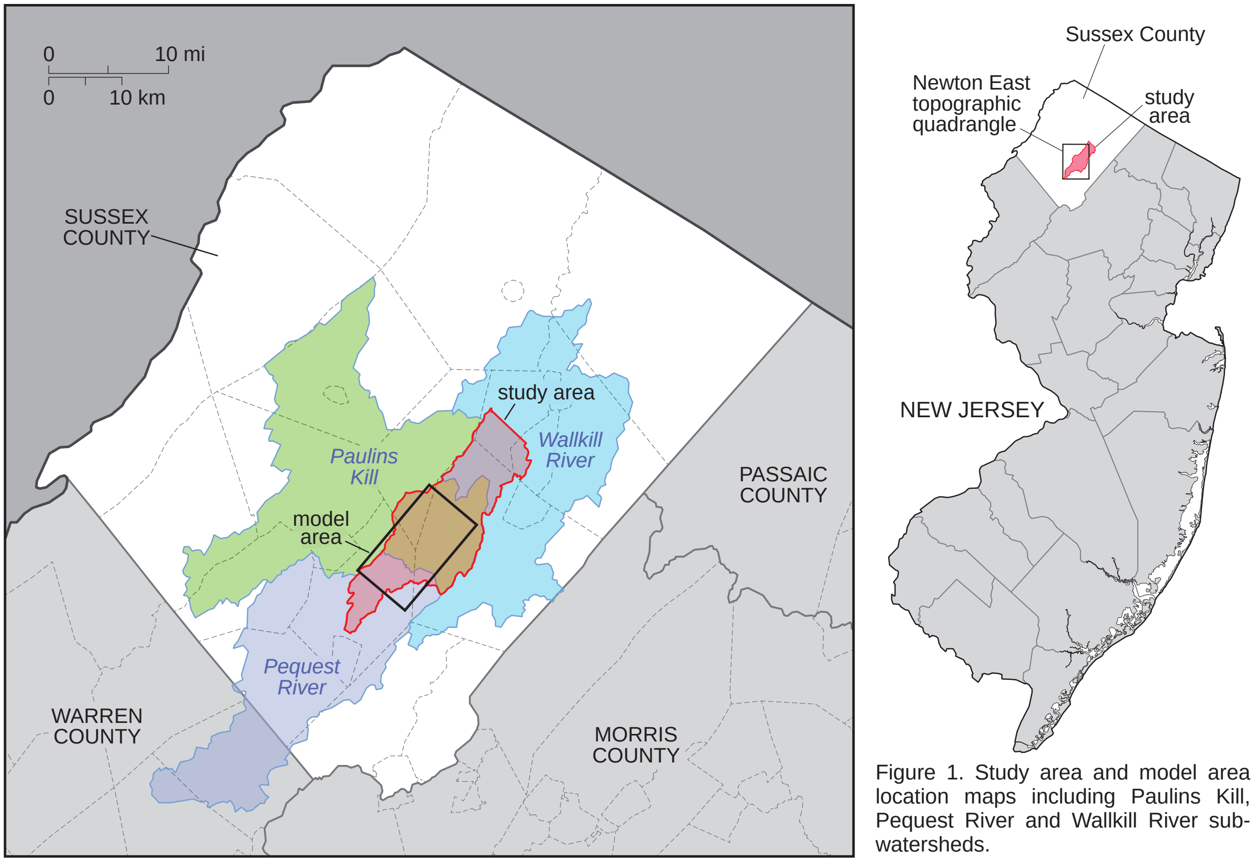


Figure 1. Study area and model area location maps including Paulins Kill, Pequest River and Wallkill River sub-watersheds.

study area.

Headwater tributaries of the Pequest and Wallkill Rivers and the Paulins Kill originate within the study area. The headwaters of the Pequest River are a series of small lakes and ponds connected by streams. In the Pequest River basin, the drainage is southwestward, following the trend of the valley. A surface-water divide between the Pequest River and Paulins Kill basins is mapped just north of Howells Pond in Andover Township (Ellis and Price, 1995). A tributary of the Paulins Kill, the East Branch, flows generally northwestward and exits the study area through a gap in the western bedrock ridge. The surface-water divide between the Paulins Kill and Wallkill River basins is just south of White Lake in Sparta Township. Drainage in the Wallkill River basin is northeastward.

Wells and numbering

Plates 1 and 2 show location of wells, geophysical surveys, stream flow measurements, and geologic sections pertinent to this study. Wells are numbered with a letter indicating the municipality where it is located and an arbitrarily assigned number; Andover, Lafayette and Sparta Township wells are designated with the letters A, L and S respectively. Where possible, the N.J. Department of Environmental Protection well permit number is also included in the report.

GEOLOGY

The oldest rocks are Precambrian-age gneiss, marble, and granitic intrusives (fig. 2). They form a high ridge bordering the southeastern edge of the study area, and underlie younger rocks in the valley. The Franklin Marble and Wildcat Marble occur in two narrow bands within the granite and gneiss (Drake and Volkert, 1993).

Overlying the Precambrian rocks along an

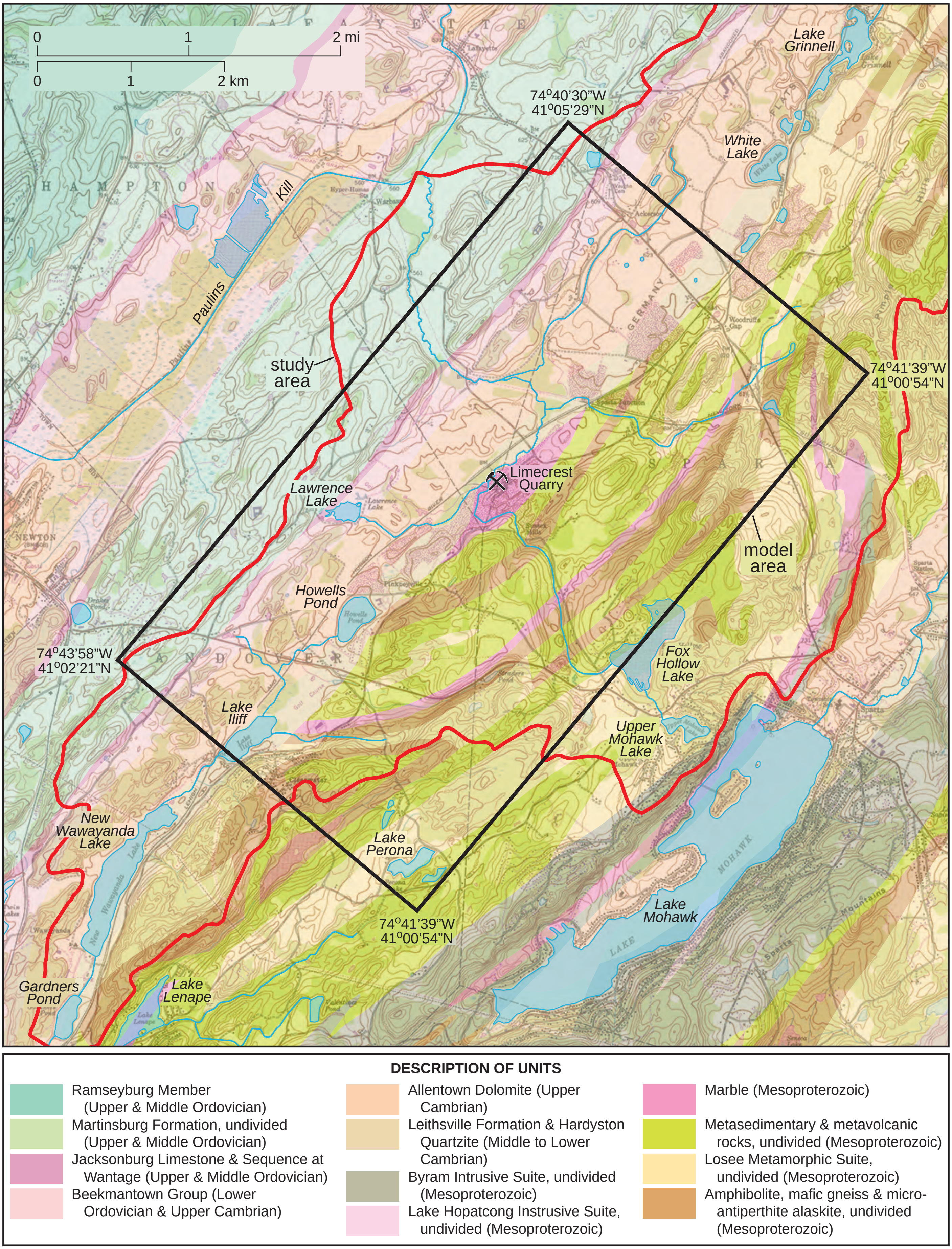


Figure 2. Bedrock map of study and model areas in Sussex County, New Jersey. Modified from Drake and others (1995).

erosional unconformity is the Cambrian-age Hardyston Quartzite. This basal unit grades into carbonate rock of the overlying Kittatinny Supergroup. The stratigraphic thickness of the Hardyston Quartzite is variable in New Jersey, ranging from about 2 feet to 200 feet (Drake and others, 1995). The formation is less than 50 feet thick in the study area.

The Kittatinny Supergroup consists of Cambrian to Ordovician-age carbonate rock of the Leithsville Formation, Allentown Dolomite, Beekmantown Group and Jacksonburg Limestone. The rock is primarily dolomite, with some limestone, dolomitic sandstone, siltstone and shale. The total stratigraphic thickness of the Kittatinny Supergroup in New Jersey is approximately 4000 feet (Herman and Monteverde, 1989). Within the study area, the sedimentary rocks generally strike northeast-southwest, following the trend of the valley, and dip at an angle of approximately 56^{o} to 65^{o} to the northwest.

The Ordovician Martinsburg Formation is the youngest rock unit, and consists of slate, graywacke and siltstone (Drake and others, 1995). The stratigraphic thickness is at least 3000 feet (Kummel, 1940). The resistant slate of the Martinsburg Formation forms a bedrock ridge on the northwestern side of the valley.

Structural deformation of rocks in the area resulted from at least two orogenic events and includes folding and faulting of the Paloezoic rocks and the development of slaty cleavage in the Martinsburg Formation. Alleghanian thrust faulting and folding is superimposed on earlier deformation leading to a structurally complex sequence (Herman and Monteverde, 1989). The Paleozoic strata overlie previously altered Precambrian rock.

Quaternary deposits in the study area are primarily late-Wisconsinan glacial sediments, and recent stream and swamp deposits (fig. 3). The glacial deposits consist of till, and sediments associated with the proglacial lakes that occupied the valley as the ice front retreated northward. The glacial sediments vary in thickness, but can be several hundred feet where valley-fill overlies carbonate bedrock at lower elevations. Till is thickest on northwest-facing bedrock slopes (Witte, 1992).

HYDROGEOLOGY

Bedrock aquifers

For the purposes of this report, the bedrock units are grouped into four aquifers 1) the igneous and metamorphic fractured-rock aquifer, 2) the Franklin Marble, 3) the carbonate fractured-rock aquifer, and 4) slate of the Martinsburg Formation. The igneous and metamorphic fractured-rock aquifer includes undifferentiated Precambrian gneiss and granitoid rock, a narrow band of the Wildcat Marble and Franklin Marble, and the Hardyston Quartzite. The stratigraphically-thin layer of Hardyston Quartzite was grouped with the underlying crystalline rock rather than the carbonate rocks because the hydraulic properties of the quartzite were thought to more closely match that of the granite and gneiss. The Franklin Marble is distinguished as a separate unit only in the area of Limecrest Quarry. The carbonate-rock aquifer consists of rocks of the Kittatinny Supergroup and Jacksonburg Limestone. The Martinsburg aquifer includes all rocks of the Martinsburg Formation. Hydrologic properties of the bedrock aquifers are summarized in table 1.

Igneous and metamorphic fractured-rock aquifer

Groundwater flow in the Precambrian crystalline rocks and the Hardyston Quartzite is attributable to the secondary porosity features of the rock. Groundwater flows primarily in joints and fractures in the rock due to the comparatively low permeability of the rock matrix.

Most water-yielding zones within fractured crystalline rock aquifers occur within a few hundred feet of the ground surface. Fractures

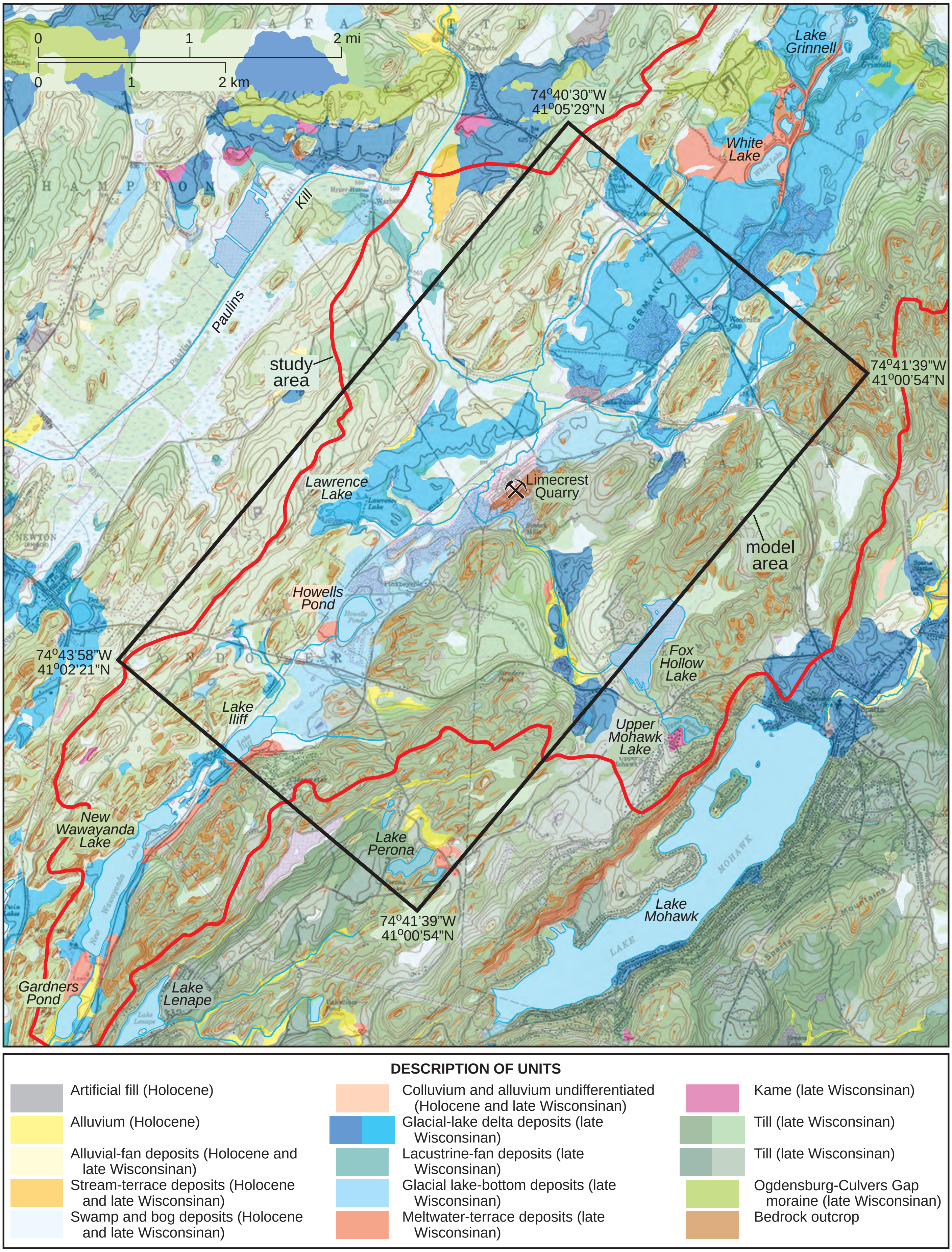


Figure 3. Surficial map of study and model areas in Sussex County, New Jersey. Modified from Witte, 1992.

Table 1. Summary of hydraulic conductivity values, well yield and specific capacities for bedrock and valley-fill aquifers. * Directionality: H = horizontal, V = vertical.

HYDROGEOLOGIC UNIT	HYDRAULIC CONDUCTIVITY (ft/day)		DIREC-TION-ALITY*	WELL YIELD (gpm)		SPECIFIC CAPACITY (gpm/ft)		LOCATION (All locations in NJ unless otherwise specified)	METHOD OF DETERMINATION	REFERENCE	COMMENTS
	Range	Average or single value		Range	Average	Range	Average				
FRACTURED-ROCK AQUIFERS Igneous and metamorphic rocks and Hardyston Quartzite	0.64 - 0.92	0.78	H	-------	-------	-------	-------	Sparta Township	aquifer test	NJGWS	Values are for pyroxene granite
	-------	3.4	H	-------	-------	-------	-------	Lehigh County, PA	estimated from specific capacity	Sloto and others, 1991	Value is for Hardyston Quartzite
	-------	-------		1 - 30	11	-------	-------	Study area vicinity	NJDEP well records	this report	Based on 46 reported values
	-------	-------		-------	-------	0.003 - 3	0.35	Study area vicinity	NJDEP well records	this report	Based on 15 reported values
	-------	-------		0.17 -100	10	-------	-------	Sussex & Warren Counties	reported values	Miller, 1974	Based on data for 1018 domestic wells
	-------	-------		-------	-------	-------	0.30	Sussex & Warren Counties	reported values	Miller, 1974	Based on a sample of 794 wells
Franklin Marble	-------	-------		-------	36	-------	-------	Sparta Township	NJDEP well records	this report	Reported value for 1 well
	-------	-------		0.25 - 100	14	-------	-------	Sussex & Warren Counties	reported values	Miller, 1974	Based on data for 162 domestic wells
	-------	-------		-------	-------	-------	0.57	Sussex & Warren Counties	reported values	Miller, 1974	Based on a sample of 129 domestic wells
Carbonate rock (Jacksonburg Limestone, Kittatinny Supergroup)	-------	327	H	-------	-------	-------	-------	Andover Township	72-hour aquifer test	this report	Transmissivity = 5555 ft²/day
	-------	10	H	-------	-------	-------	-------	Morris County	ground-water-flow model	Voronin and Rice, 1996	
	22 - 432	-------	H	-------	-------	-------	-------	Warren County	ground-water-flow model	Hutchinson, 1981	
	0.1 - 864	-------	H	-------	-------	-------	-------	Morris County	ground-water-flow model	Nicholson and others, 1996	
	-------	-------		<1 - 450	-------	-------	-------	Study area vicinity	NJDEP well records	this report	Based on 85 reported values
	-------	-------		-------	-------	0.01 - 151	5.2	Study area vicinity	NJDEP well records	this report	Based on 43 reported values
Ordovician Martinsburg Slate	-------	-------		0.25 - 120	10.5	-------	-------	Sussex and Warren Counties	reported values	Miller 1974	Based on a sample of 919 domestic wells
	-------	-------		-------	-------	-------	0.39	Sussex and Warren Counties	reported values	Miller 1974	Based on a sample of 495 domestic wells
	-------	1.7	H	-------	-------	-------	-------	Lehigh County, PA	estimated from specific capacity	Sloto and others, 1991	Based on specific capacity data
	-------	-------		<1 - 50	11	-------	-------	Study area vicinity	NJDEP well records	this report	Based on 13 reported values
	-------	-------		-------	-------	0.01-0.80	0.30	Study area vicinity	NJDEP well records	this report	Based on 13 repored values
QUATERNARY DEPOSITS Quaternary stratified-drift aquifer	-------	149	H	-------	-------	-------	-------	Andover Township	aquifer test	this report	Based on transmissivit = 14,878 ft²/day
	10 - 500	-------	H	-------	-------	-------	-------	Morris County	calibrated model value	Voronin, 1991	
	188 - 951	-------	H	-------	-------	-------	-------	Passaic County	aquifer test	J.L. Hoffman, written comm.	
	17 - 130	-------	H	-------	-------	-------	-------	Morris County	calibrated model values	Nicholson and others, 1996	
	-------	-------		3 - 733	60	-------	-------	Study area vicinity	NJDEP well records	this report	Based on 13 reported values
	-------	-------		-------	-------	0.09 - 9.2	2.3	Study area vicinity	NJDEP well records	this report	Based on 10 reported values
Glacial low-permeability layer	-------	0.0059	V	-------	-------	-------	-------	Andover Township	lab permeameter	this report	Lab test by Woodward-Clyde, 1993
	-------	0.93	V	-------	-------	-------	-------	Andover Township	lab permeameter	this report	Lab test by Woodward-Clyde, 1993
	-------	0.04	V	-------	-------	-------	-------	Andover Township	aquifer test	this report	
	-------	0.3	V	-------	-------	-------	-------	Passaic County	slug test	NJGWS unpublished data	
	-------	0.004	V	-------	-------	-------	-------	Morris County	lab permeability	Nicholson and others, 1996	

are more abundant near the land surface due in part to physical weathering of the rock by near-surface hydrologic processes. The overall thickness of the aquifer can be estimated based on depths of water-supply wells in this unit. Miller (1974) reported an average well depth of 141 feet based on 140 wells in the Precambrian rock of Andover and Sparta Townships, and noted that “no fractures occurred below 300 feet”. Based on these data, the thickness of the crystalline bedrock aquifer is assumed to be approximately 200 feet.

Reported hydraulic conductivity values for pyroxene granite in Sparta Township, New Jersey ranged from 0.64 to 0.92 ft/d and averaged 0.78 ft/d, based on unpublished data of the NJGWS compiled in 1995. The data is now included in Mennel and Canace (2002). A median hydraulic conductivity of 3.4 ft/d was determined for the Hardyston Quartzite in Lehigh County, Pennsylvania (Sloto and others, 1991).

Data on well yields and specific capacities of domestic wells in Sussex and Warren Counties, New Jersey by Miller (1974) provides a measure of the relative productivity of the bedrock aquifers. In a survey of 1,018 domestic wells in the “Precambrian crystallines”, well yields ranged from 0.17 to 100 gpm, with an average of 10 gpm, and median of 8 gpm. Based on a sample of 794 wells, the average specific capacity is 0.30 gpm/ft. In the study area, well yields ranged from 1 to 30 gpm with an average of 11 gpm based on 46 reported values from NJDEP well records. Specific capacity ranged from 0.003 to 3 gpm/ft and averaged 0.35 gpm/ft based on 15 reported values. Values for both well yield and specific capacity are less than those reported for the “Kittatinny Formation” and “Franklin limestone”, and similar to those of the Martinsburg Formation (table 1). Despite the lower yields, the crystalline fractured-rock aquifer has proved adequate for domestic and municipal water.

Franklin Marble

The Franklin Marble is a metasedimentary calcium carbonate, and is therefore susceptible to chemical dissolution and the formation of solution-enhanced fractures and cavernous zones. A number of caves have formed in the Franklin Marble within the vicinity of the study area (Dalton, 1976). Groundwater flow occurs primarily in fractures or openings in the rock matrix.

A survey of depths of 129 water wells in the Franklin Marble in Sussex and Warren Counties reported that most wells were drilled to less than 150 ft, but some were deeper than 300 ft (Miller, 1974).

Because of its relatively limited extent, well data for the Franklin Marble are sparse. In a survey of domestic wells in the study area, only Well S-30 (NJDEP Well Permit Number 22-08556) was reportedly installed in the “white limestone” or Franklin Marble. The yield for this well was 36 gpm, higher than that of wells in the surrounding granite and gneiss, which averaged 11 gpm based on a sample of 46 wells. Based on data for 162 domestic wells, Miller (1974) reports well yields ranging from 0.25 to 100 gpm, with an average yield of 14 gpm, and median of 10 gpm. A specific yield of 0.57 gpm/ft is reported based on a sample of 129 domestic wells. Again, these values are higher than those reported for the surrounding crystalline rock, though not as large as those indicated for the carbonate rocks. Additionally, several million gallons of water per day are pumped from the Franklin Marble by Limecrest Quarry, indicating that the marble is capable of very high yields. However, due to its limited extent, regionally this aquifer is not as important for water supply as the more regionally-extensive carbonate-rock aquifer.

Carbonate-rock aquifer

The carbonate-rock aquifer includes all rocks of the Kittatinny Supergroup and Jacksonburg Limestone. Chemical dissolution of the dolomite and limestone has enlarged fractures and bedding plane partings and led to the development of solution cavities and caverns, thereby enhancing groundwater flow. The carbonate rocks vary greatly in lithology and therefore in their resistance to physical and chemical weathering and their water-bearing potential.

Test drilling in the study area provided some insight into the thickness of the carbonate rock aquifer. Well L-1 (NJDEP 22-33686) is a 650-feet-deep well in the Allentown Dolomite in Lafayette Township. Seams in the rock were observed as deep as 611 feet, but were more abundant above 300 feet. Wells A-2 (NJDEP 22-32508) and A-3 (NJDEP 22-32509) at the Rolling Greens Golf Course in Andover Township were drilled to depths of 300 feet and 378 feet respectively, and completed in weathered carbonate rock below approximately 200 feet of glacial sediments. The highly weathered zone of carbonate rock yielded several hundred gallons per minute. The high degree of weathering observed in the cuttings suggests that this permeable zone extends to depths greater than those drilled.

Other researchers provide a range of values for determining the carbonate rock aquifer thickness. For a groundwater model in western Morris County, New Jersey, Nicholson and others (1996) assumed the entire thickness of the Kittatinny carbonates, as much as 900 feet, constituted the aquifer thickness due to the presence of deep wells with high yields. Sloto and others (1991) examined the frequency of water bearing zones in 27,228 feet of uncased borehole in the Kittatinny and Jacksonburg Limestone of Lehigh County, Pennsylvania and found that 82 percent of water bearing zones are within 250 feet of land surface, whereas only 4 percent occurred below a depth of 350 feet. Miller (1974) noted that most cavernous zones in the Kittatinny occur between 50 to 300 feet below land surface and that the chance of obtaining a good water supply below 600 feet is generally slight.

The data suggest that the depth of open fractures in the Kittatinny carbonate rock is variable and probably depends on topographic position, the stratigraphic unit encountered, and other geologic and hydrologic factors. For this investigation, the uppermost 300 feet of carbonate rock was assumed to constitute the carbonate rock aquifer.

The susceptibility to chemical dissolution is a major factor in determining rock permeability and potential well yields. Markewicz and others (1981) noted that sinkholes and cavernous zones are most common in the Califon and Wallkill members of the Leithsville Formation, the Allentown Dolomite, and the Hope Member of the Rickenbach Formation, and that the highest-yielding wells tap the Leithsville and Allentown formations. During field investigations for this study, several depressions, sinkholes, and a small disappearing stream were observed in carbonate rocks in these units. Sloto and others (1991) also determined that the highest-yielding domestic and industrial wells in the Kittatinny Supergroup in Pennsylvania draw water from the Leithsville Formation and the Allentown Dolomite.

The values of hydraulic conductivity and transmissivity for the carbonate rock are larger than those obtained for the other bedrock aquifers. As part of this study, an aquifer test was conducted in the carbonate bedrock at the Rolling Greens Golf Course in Andover Township (pl. 1). Analysis of the test yielded a hydraulic conductivity value of 327 ft/d, an average transmissivity value of 5555 ft^2/d, and a storage coefficient of 1.37×10^{-3} (fig. A-1). Hydraulic conductivity values for carbonate rock from groundwater models in New Jersey include 10 ft/d (Voronin, 1991) and 22 to 432 ft/d (Hutchinson, 1981). Nicholson and others (1996) obtained values ranging from 0.1 to 864 ft/d, but most of the modeled area was simulated with values of 60 ft/d or less.

Well yields and specific capacity data from

Miller (1974) also indicate that the carbonate-rock aquifer is capable of high yields. Based on a survey of 422 domestic wells, yields ranged from 0.25 to 120 gpm, with an average yield of 14 gpm and a median of 10 gpm. Average specific capacity based on a sample of 298 domestic wells was 1.05 gpm/ft.

The carbonate rock is the most productive bedrock aquifer. It is regionally extensive and capable of large well yields. In the study area, it is second only to the glacial sand and gravel aquifers in water-bearing potential.

Martinsburg Formation slate

The Ordovician Martinsburg Formation is the youngest rock unit in the study area. It is an interbedded, laminated to medium-bedded graywacke and siltstone, and slate (Drake and others, 1995). It is at least 1000 feet thick in Sussex County. The resistant slate of the Martinsburg Formation forms a bedrock ridge along the northwestern side of the valley.

The Martinsburg is the least permeable of the fractured-rock aquifers. Well records for the study area report several dry holes were drilled while attempting to install domestic wells. The relative resistance of the rock is evident in that it forms a bedrock ridge adjacent to the more permeable carbonate rocks. The occurrence of small ponds and a spring near the top of the slate ridge suggest that the rock's low permeability impedes recharging water from penetrating deeper into the aquifer. Miller (1974) notes that most of the successful wells in the Martinburg Formation in Sussex and Warren Counties are completed in the weathered zone within 200 feet of the surface. Based on available data, the aquifer is assumed to be 200 feet thick in the study area.

The low permeability of the Martinsburg slate is documented by previous researchers. An average hydraulic conductivity of 1.7 ft/d and median of 0.8 ft/d for the aquifer in Lehigh County, Pennsylvania was estimated by Sloto and others (1991) based on specific capacity data.

Yields of 13 domestic wells in the Martinsburg Formation within the study area range from less than 1 to 50 gpm, and average 11 gpm. The specific capacity of these wells ranges from .01 to .80 gpm/ft, and averages 0.30 gpm/ft. These results are similar to those reported by Miller (1974). Based on a sample of 919 domestic wells, yields ranged from 0.25 to 120 gpm, with an average of 10.5 gpm and a median of 6 gpm. An average specific capacity of 0.39 gpm/ft is reported based on a sampling of 495 domestic wells.

Glacial valley fill

The thickness of the valley-fill deposits is variable. It is absent where bedrock crops out at the surface but can exceed 200 feet where sediment from glacial meltwater filled in glacial lakes and formed deltas (Witte, 1992). The most permeable sediments are the well-sorted glacial-deltaic and lacustrine-fan sands and gravels. These deposits form the glacial aquifers. The fine-grained lake-bottom deposits of silt, fine sand, and clay, in places overlain by Quaternary swamp deposits, are confining units which impede groundwater flow.

Groundwater flow in the unconsolidated sediments is due to the primary porosity of these deposits. Water flows in pore spaces between the sediment grains. Unlike the fractured-rock aquifers, the valley-fill sediments are capable of storing large volumes of water.

Hydrostratigraphy

The hydrostratigraphy of the study area was developed based on stratigraphic mapping of the valley-fill deposits as part of this study, and mapping of the surficial geology by Witte (1992). Geologic and geophysical logging of test wells, aquifer permeability testing, and water-level monitoring were evaluated within the context of the glacial depositional history and surficial mapping to delineate the distribution of aquifer and confining units, and ascertain the hydraulic connection between the glacial deposits and the underlying bedrock aquifers.

Sections A-A', B-B' and C-C' of figures 4, 5 and 6, respectively, depict the hydrogeologic framework of the valley-fill deposits. Lines of section and locations of wells used to generate the sections are shown on Plates 1 and 2. Well logs for geologic sections in figures 4, 5 and 6 are included in table A-1. South of Howells Pond, the permeable glacial sediments are divided into an upper and lower aquifer separated by a semiconfining unit of silt and minor clay (fig. 4). The maximum thickness of the glacial deposits including the low-permeability layer is about 200 feet. The semiconfining unit is as much as 40 feet thick, and pinches out to the northeast. In the upper permeable layer, which is generally less than 50 feet thick, sediment grain size decreases with depth from the surface until the semiconfining unit is encountered. In the permeable sediments beneath the semiconfining unit, grain size increases downward from a silty sand at the base of the semiconfining unit to gravel and sand at the bedrock contact. The thickness of the lower permeable layer is extremely variable, ranging from 10 feet to over 100 feet thick.

Sediments of the upper aquifer are primarily glacial-deltaic sediments that were deposited as glacial ice stood at a stable ice-margin position about 2000 feet north of Howells Pond (Witte, 1992). Sediments of the lower aquifer are glacial-deltaic and glacial-lacustrine-fan in origin. The fine-grained gray silt and clay of the semiconfining unit are lake-bottom deposits from a small glacial lake that formed south of the ice front.

The drilling of four test wells and aquifer testing at the Rolling Greens Golf Course allowed for a better interpretation of the hydrogeologic framework south of Howells Pond. The results of aquifer testing indicated that the fine-grained layer acts as a semiconfining unit that impedes vertical flow from the upper to the lower glacial aquifer. During the 72-hour test, Well A-3 (NJDEP 22-32509), in the weathered carbonate bedrock, was pumped and drawdown occurred in adjacent bedrock wells A-2 (NJDEP 22-32508) and A-4 (NJDEP 22-32507) screened in the lower glacial aquifer. No response was observed in an adjacent shallow well A-5 (NJDEP 22-32510) screened above the semiconfining unit, indicating that the pumping stress in the bedrock aquifer was not easily transmitted to the water-table aquifer. The time-drawdown data for observation wells in the bedrock and lower glacial aquifer best conform to the type-curve for leaky aquifers (Hantush and Jacob, 1955), which suggests some degree of confinement for the aquifer (fig. A-1).

Drillers' logs from several wells installed north of Howells Pond indicate that the low permeability layer at the Rolling Greens Golf Course is probably absent beneath the pond (figs. 4 and 5). Geologic logs of several wells at the Lifecare Mews facility, approximately 1000 feet north of the pond, report only sand and gravel directly overlying the carbonate bedrock. Here, the semiconfining unit is absent and the entire thickness of the glacial sediments, as much as 100 feet, constitutes a single aquifer. The sediments generally coarsen towards the "head of outwash", a location where the glacial ice stagnated while depositing sediment, north of Howells Pond.

An aquifer test conducted at the Lifecare Mews facility on August 17 to August 21, 1992 (Doncar, Inc., unpublished data available on file at NJGWS), indicates that the undifferentiated sediments of the glacial aquifer are in good hydraulic connection with the dolomite bedrock. Test pumping of Well A-16 (NJDEP 22-32016) in the glacial aquifer at a depth of about 70 feet resulted in drawdown in Well A-17 and Well A-18 (NJDEP 22-32137 and NJDEP 22-32138, respectively) screened in the glacial sediments and in Well A-19 (NJDEP 22-20866) open to the bedrock aquifer.

The thickness of the glacial sediments

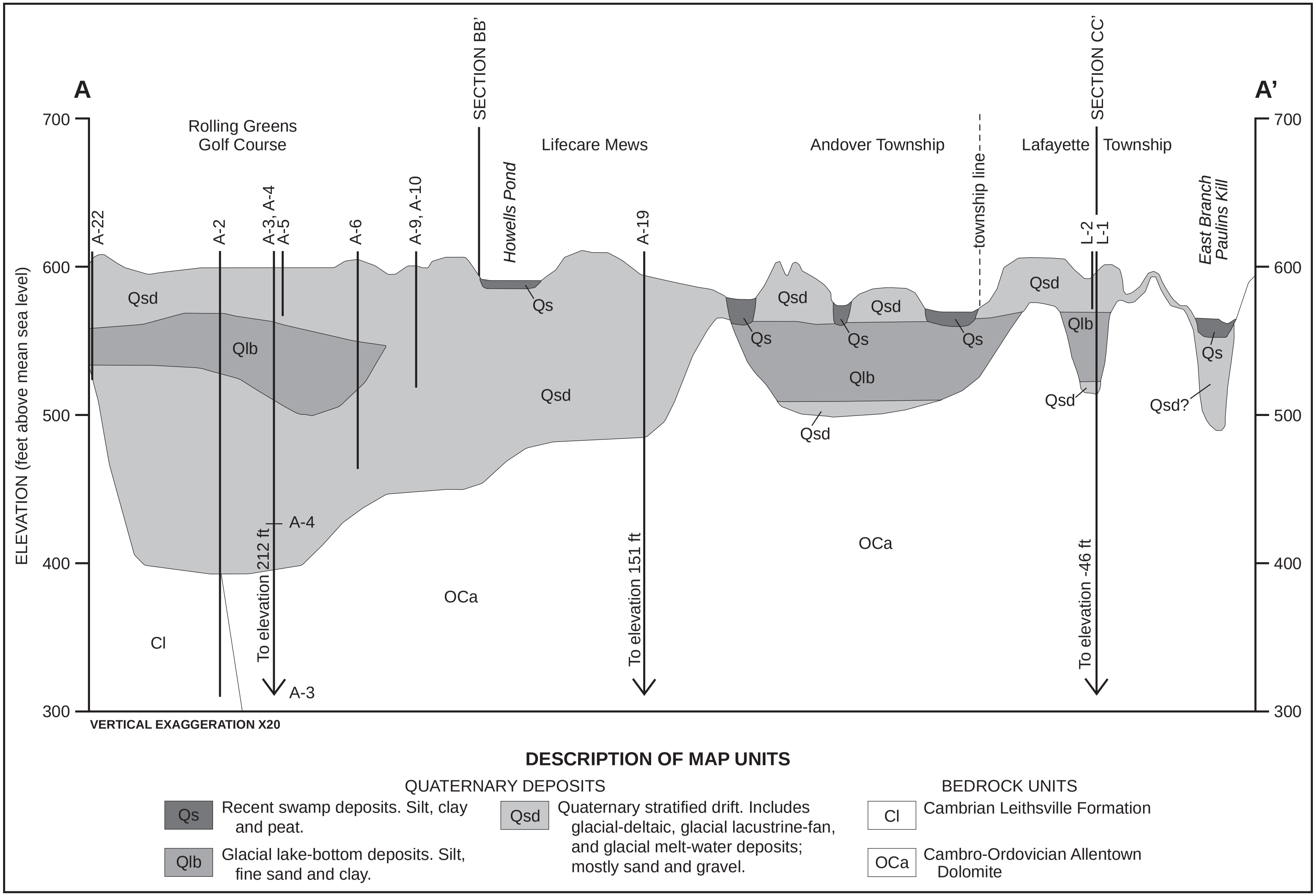


Figure 4. Down-valley geologic section **A-A’** showing valley-fill and bedrock units.

beneath Howells Pond was determined based on a geophysical traverse made in winter when the pond was ice-covered. Transient electromagnetic (TEM) data were collected at four stations (pl. 2, TEM-1 through TEM-4). The data indicate that the depth to bedrock ranges from about 65 feet along the pond's southeastern edge, shallows slightly to 50 feet near its center, and drops steeply from 65 feet to 140 feet along its western edge (S. Sandberg, New Jersey Geological Survey, written commun., 1993). Although the geophysical data provide an estimate of the depth to bedrock beneath the pond, the lithology of the sediment can not be distinguished. Field observations indicate that the shallow subsurface contains peat and muck of recent origin.

North of Howells Pond, and extending into Lafayette and Sparta Townships, a low ridge of carbonate bedrock divides the valley (fig. 6). Here the thickness of the glacial deposits is variable. Surficial deposits are thickest on the western side where glacial lake-bottom sediments occur. Geophysical seismic refraction data collected along the rise (pl. 2, lines GS-1, GS-2) show an undulating bedrock surface; depth to bedrock ranges from 35 to 70 feet (D. Hall, New Jersey Geological Survey, written commun., 1996). Field observations at an abandoned sand and gravel quarry near the Lafayette and Andover Township border also reveal that glacial sediments overlie a hummocky bedrock surface.

Because well data for the central part of the study are sparse, drilling of test wells L-1 (NJDEP 22-33686) and L-2 (NJDEP 22-33635) provided key information about the hydrogeologic framework. Drilling encountered (from the surface) approximately 20 feet of fine to medium sand, silt and some gravel; 60 feet of gray silt and clay; 5 feet of gravel capable of yielding 150 gpm; dolomite bedrock. The groundwater level in shallow well L-2 (NJDEP 22-33635) screened in glacial sands overlying the silt and clay layer was about 30 feet higher than the water level in the lower gravel and bedrock, indicating that the fine-grained layer acts as a confining unit in this area.

Well data for the eastern side of the valley are also sparse. The area directly west of Limecrest Quarry is predominantly wetlands drained by the East Branch Paulins Kill. Although the shallow subsurface is mapped as low-permeability recent swamp deposits, possibly overlying peat and organic-rich silt and clay (fig. 4), hydrologic evidence suggests that more permeable glacial sediments may also be present. Data collected as part of this study from streambed piezometers in the East Branch Paulins Kill indicate that, at various times, stream water is lost to the aquifer along this reach, suggesting a good connection between the groundwater and surface-water systems through relatively permeable deposits.

The area between the East Branch Paulins Kill and the northern boundary of the study area consists predominantly of glacial-deltaic deposits, although glacial lake-bottom deposits and till may occur in the subsurface (Witte and Monteverde, 2006). The glacial deposits thicken northeastward to approximately 200 feet just north of the study area boundary. Several sand and gravel quarries operate in this region. Analysis of aquifer-test results for a test conducted by the Sparta Township Municipal Utilities Authority at the Tanis Sand and Gravel Quarry on March 20, 1991 indicate that the aquifer is unconfined at this location. Time-drawdown data observed in Well S-18 (NJDEP 22-30665) caused by the pumping of Well S-17 (NJDEP 22-20370) shows delayed yield aquifer response; accordingly this test was analyzed using the Neuman (1975) unconfined aquifer type curves (fig. A-2).

Hydrologic properties of valley-fill deposits

The valley-fill aquifers and confining units consist primarily of late-Wisconsinan glacial deposits. The sediments that make up these aquifers were transported by glacial meltwaters,

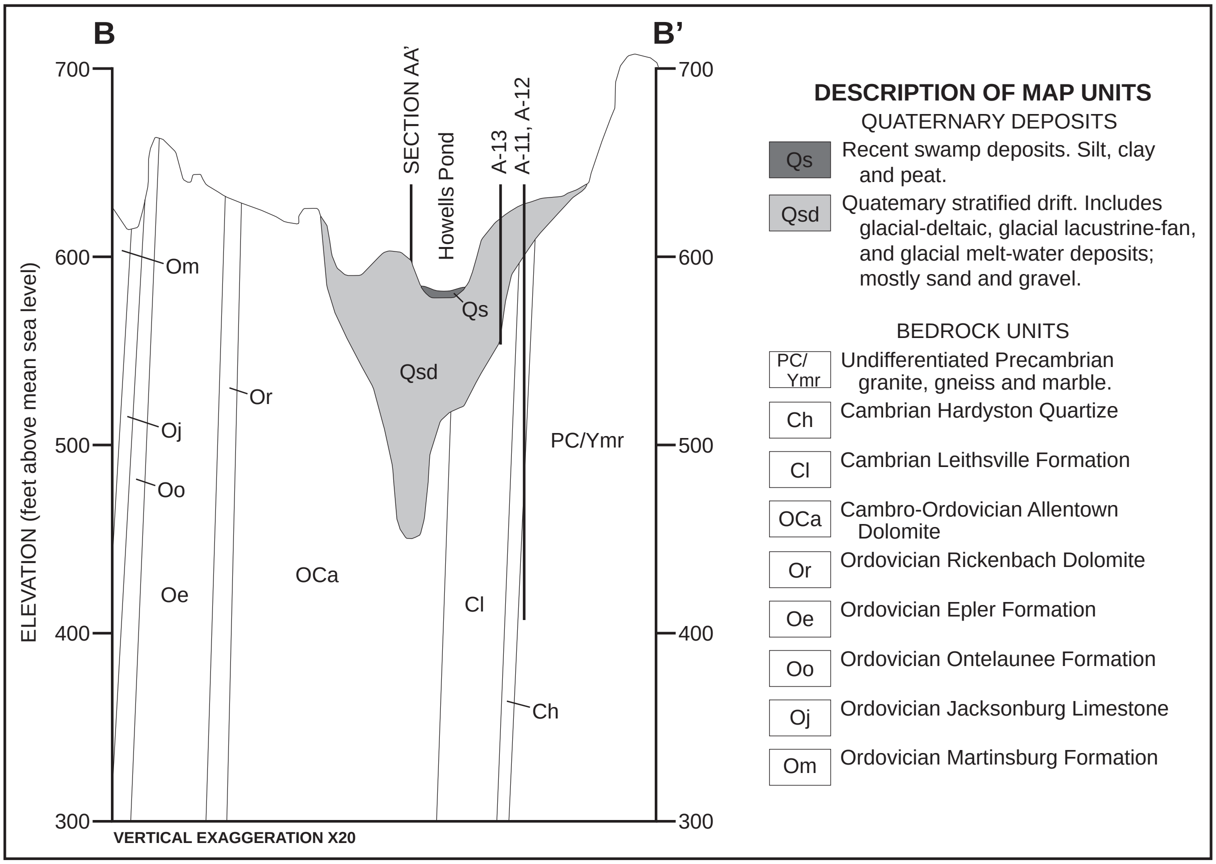


Figure 5. Geologic section **B-B'** through Howells Pond, Andover Township, Sussex County.

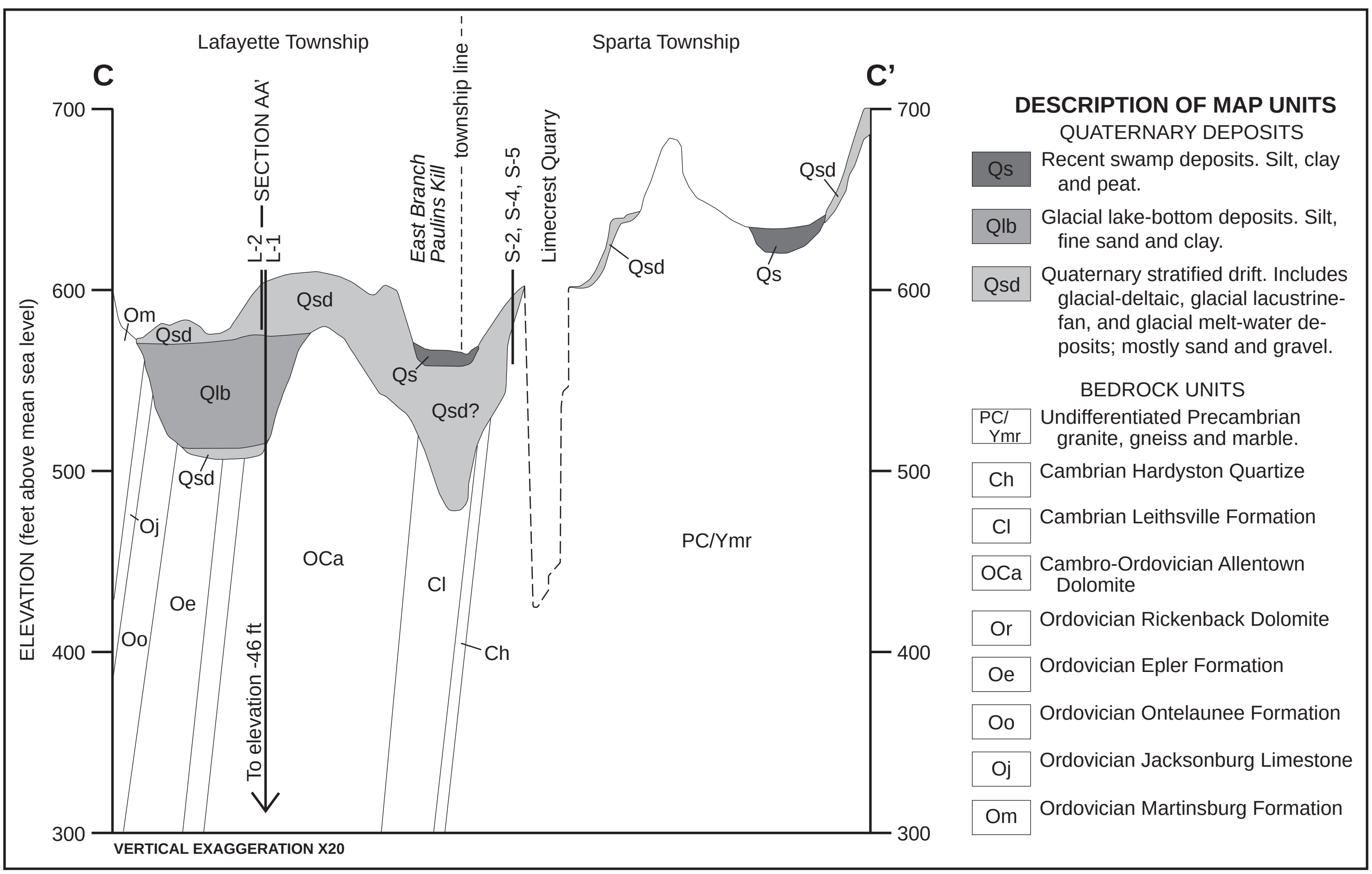


Figure 6. Geologic section **C-C'** showing valley-fill and bedrock units near Limecrest Quarry.

carried in the glacial ice, or settled in quiet-water environments. The deposits are therefore highly variable in lithology, and grain size can range from boulder to clay-size particles. For these reasons, large variations in aquifer properties characterize glaciated terrains. Hydrologic characteristics of the valley-fill aquifers and confining units are summarized in table 1.

Valley-fill aquifers

Analysis of recovery data from a 72-hour aquifer test conducted August 17 through August 21, 1992 by Doncar, Incorporated in the glacial sediments at the Lifecare Mews facility yielded a transmissivity value of 14,878 ft^2/d, and a hydraulic conductivity value of 149 ft/d.

Hydraulic conductivity values for glacial sands and gravels obtained by previous researchers in New Jersey include 10 to 500 ft/d (Voronin, 1991), and 188 to 951 ft/d (J.L. Hoffman, New Jersey Geological Survey, written commun., 1995). Values used in a groundwater model for a buried valley in Morris County ranged from 17 to 130 ft/d for the upper glacial aquifer, and averaged 81 ft/d for the lower glacial aquifer (Nicholson and others, 1996).

Well yields for 13 wells installed in glacial sediments in the study area and vicinity range from 3 to 733 gpm, and average 60 gpm. Specific capacities of 10 wells in the valley-fill aquifer range from .09 to 9.2 gpm/ft, and average 2.3 gpm/ft.

Valley-fill confining units

The vertical hydraulic conductivity of the low-permeability sediments is based on undisturbed samples of the glacial lake-bottom semiconfining unit. The samples were collected using a Shelby tube sampler at test wells A-2 (NJDEP 22-32508) and A-3 (NJDEP 22-32509) at the Rolling Greens Golf Course in Andover Township. They were analyzed by the Woodward-Clyde Laboratory Facility in Clifton, New Jersey for grain-size distribution using the combined sieve and hydrometer method (American Society for Testing of Materials (ASTM) D422-63), hydraulic conductivity using the tube permeameter method (ASTM D5084-90), and supplemented by a visual description of the sediments.

Sample no. 1 was collected at Well A-2 (NJDEP 22-32508) at an interval of 65 to 67 feet below land surface. The material is a brown, silty, fine sand composed of 40 percent silt. Its vertical hydraulic conductivity is 0.93 ft/day.

Sample no. 2 was obtained from Well A-3 (NJDEP 22-32509) at a depth of 60 to 62 feet below land surface. It is a gray, non-plastic silt with a trace of fine sand and a clay layer. Silt and clay made up 93 percent of the sample. Its vertical hydraulic conductivity is 5.9×10^{-3} ft/day.

An estimate of vertical hydraulic conductivity of the semiconfining unit is based on aquifer test data at the Rolling Greens Golf Course and the following equation:

(1) $K' = L' * b'$

where L' is leakance, or 9.9×10^{-4} /d from the aquifer test solution; b' is the average saturated thickness of the semi-confining unit, or 40 ft; K' is vertical hydraulic conductivity of the semi-confining unit, or 4.0×10^{-2} ft/d from the previous values.

Slug test data for tests in glacial varved silts and clay in Passaic County, New Jersey yielded an average vertical hydraulic conductivity value of 0.3 ft/d (J. L. Hoffman, New Jersey Geological Survey, written commun., 1994). A value of 4.0×10^{-3} ft/d was obtained from laboratory permeability testing of glacial lake-bottom sediments from Morris County, New Jersey (Nicholson and others, 1996).

Seasonal water-level trends

The seasonal trend in groundwater and surface-water levels is evident in the hydrographs of figures 7 and 8. The graphs are based on data from observation wells and a staff gage in Lake Iliff. The seasonal fluctuations are typical of water-level trends in temperate climates where precipitation is distributed fairly evenly throughout the year. The water table is highest in early spring and declines during the summer and early fall due to evapotranspiration. It rises again in winter after the growing season has ended.

Seasonal water-level fluctuations were more pronounced in groundwater recharge areas (fig. 7) than in groundwater discharge areas near valley streams and lakes (fig. 8). The maximum seasonal fluctuations recorded in the carbonate rock aquifer and valley-fill aquifer in groundwater recharge areas were 15 feet and 14 feet, respectively. The corresponding fluctuations near groundwater discharge areas were 7 feet and 6 feet, respectively. Surface-water fluctuations in Lake Iliff were smaller, reaching a maximum of only 3 feet. This is due to the larger storage capacity of the lake and an artificial control at its outlet that helps to maintain the lake level during periods of low precipitation.

Stream-aquifer interaction

Surface-water bodies in the study area are generally in good connection with the underlying aquifers. Exceptions occur where streams overlie fine-grained glacial-lake-bottom sediments. Some small streams in the study area are intermittent and go dry when the water table declines in summer and early autumn.

The interaction between ground and surface water must be analyzed within the context of groundwater withdrawal in the valley. As discussed in the subsequent Groundwater

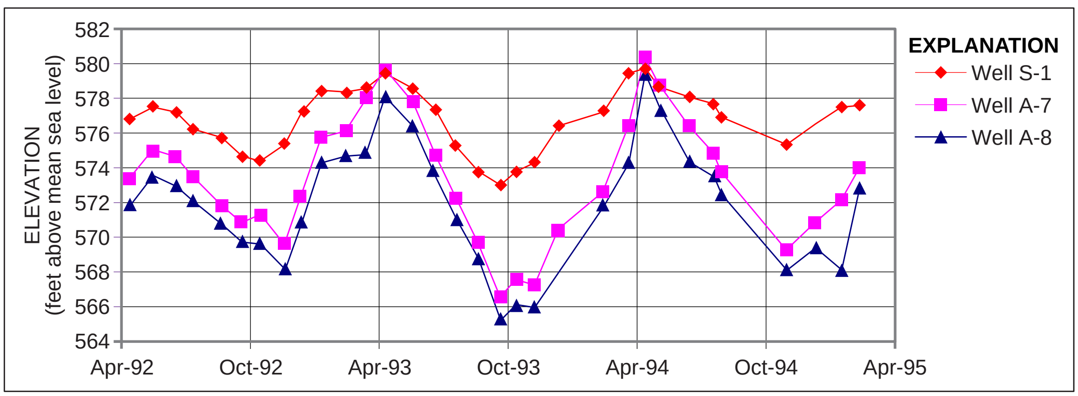


Figure 7. Seasonal water-level fluctuations in groundwater recharge areas.

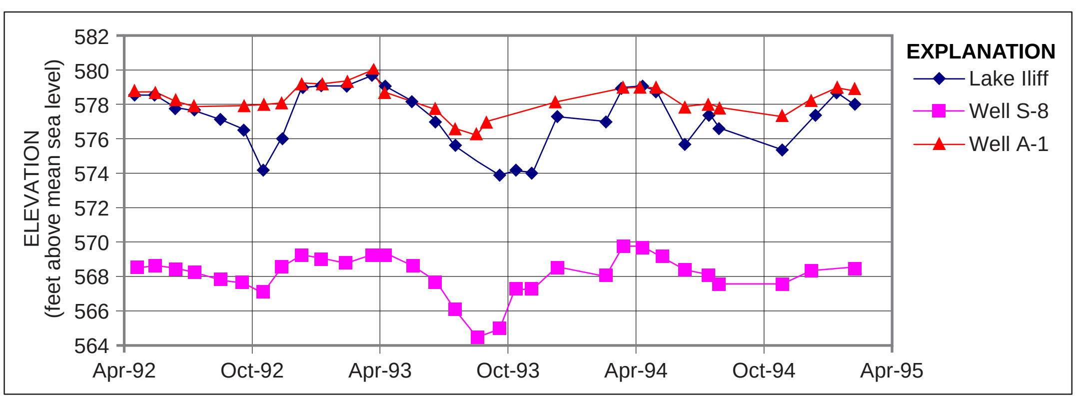


Figure 8. Seasonal water-level fluctuations in groundwater discharge areas.

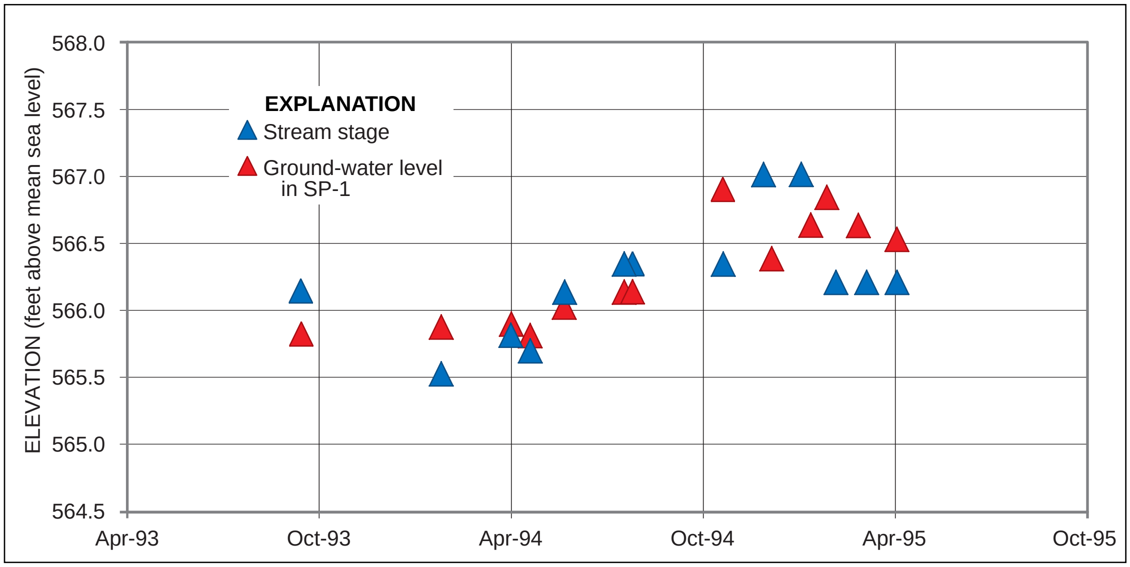


Figure 9. Comparison of water level in streambed piezometer SP-1 and stream stage in East Branch Paulins Kill.

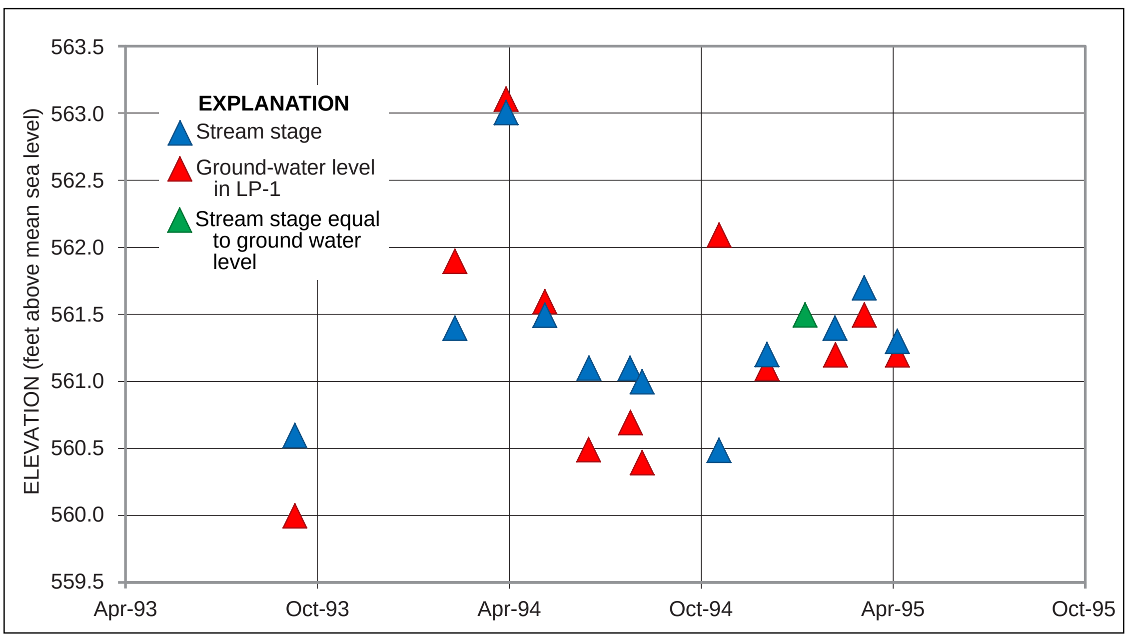


Figure 10. Comparison of water level in streambed piezometer LP-1 and stream stage in the East Branch Paulins Kill.

Table 2. Comparison of groundwater level in piezometer SP-1 and stream stage in the East Branch Paulins Kill. (ft above msl = feet above mean sea level)

Date of measurement	**Ground-water elevation** (ft above msl)	**Stream stage** (ft above msl)	**Elevation of piezo-metric surface above stream stage** (ft)	**Stream status**
30-Sep-93	565.9	566.1	-0.2	losing
15-Feb-94	565.9	565.5	0.4	gaining
21-Apr-94	565.9	565.8	0.1	gaining
10-May-94	565.8	565.7	0.1	gaining
20-Jun-94	566.0	566.1	-0.1	losing
29-Jul-94	566.1	566.3	-0.2	losing
9-Aug-94	566.1	566.3	-0.2	losing
8-Nov-94	566.9	566.3	0.6	gaining
20-Dec-94	566.4	567.0	-0.6	losing
25-Jan-95	566.6	567.0	-0.4	losing
22-Feb-95	566.8	566.2	0.6	gaining
21-Mar-95	566.6	566.2	0.4	gaining
21-Apr-95	566.5	566.2	0.3	gaining

Budget section of this report, a large-scale groundwater diversion and surface water discharge at Limecrest Quarry affects the flow of water in the valley. Water-level data and modeling results indicate that dewatering at Limecrest Quarry has created a downward vertical gradient throughout much of the study area, inducing leakage of streamwater to the underlying aquifers.

To investigate stream-aquifer interactions, piezometers SP-1 and LP-1 were installed several feet below the stream bottom in the East Branch Paulins Kill downstream from the Limecrest Quarry surface-water discharge (pl. 2). Data collected at both piezometers record periods of upward and downward vertical flow. Figures 9 and 10 and accompanying tables 2 and 3 show water-level trends in SP-1 and LP-1, respectively, compared with stream stage in the East Branch Paulins Kill based on 13 measurements obtained between September 1993 and April 1995. From September 1993 to September 1994, the stream reach gained water from the aquifer during autumn and winter,

Table 3. Comparison of groundwater level in piezometer LP-1 and stream stage in the East Branch Paulins Kill. (ft above msl = feet above mean sea level)

Date of measurement	**Ground-water elevation** (ft above msl)	**Stream stage** (ft above msl)	**Elevation of piezo-metric surface above stream stage** (ft)	**Stream status**
30-Sep-93	560.0	560.6	-0.6	losing
15-Feb-94	561.9	561.4	0.5	gaining
21-Apr-94	563.1	563.0	0.1	gaining
10-May-94	561.6	561.5	0.1	gaining
20-Jun-94	560.5	561.1	-0.6	losing
29-Jul-94	560.7	561.1	-0.4	losing
9-Aug-94	560.4	561.0	-0.6	losing
8-Nov-94	562.1	560.5	1.6	gaining
20-Dec-94	561.1	561.2	-0.1	losing
25-Jan-95	561.5	561.5	0.0	gaining
22-Feb-95	561.2	561.4	-0.2	losing
21-Mar-95	561.5	561.7	-0.1	losing
21-Apr-95	561.2	561.3	-0.1	losing

and lost water to the aquifer in the spring and summer months. However, these trends are not observed in the subsequent data, suggesting that the amount and timing of precipitation also affect stream-aquifer interaction.

SIMULATION OF GROUNDWATER FLOW IN THE MID-1990's

A major goal of this report is to assess how groundwater withdrawals affect groundwater flow in the study area. To do this, average groundwater flow conditions in the mid-1990's were modeled using a computer groundwater-flow model. The calibrated flow model was then used to simulate predevelopment conditions in the valley, by removing the pumping stresses and observing the response.

Groundwater flow was modeled using the U.S. Geological Survey modular three-dimensional finite-difference groundwater-flow model, or MODFLOW (McDonald and Harbaugh, 1988). The computer program uses a finite-difference method to solve a set of algebraic equations based on the partial differential equation for groundwater flow in three dimensions and user-defined boundary conditions and input parameters to generate values for head and flow at each model cell. ZONEBUDGET (Harbaugh, 1990) a program which provides mass-balance information, was used to quantify flow volumes between aquifer zones and to streams and lakes.

Conceptual model

In the conceptual model of groundwater flow, water is supplied by precipitation. Water that is not lost to evapotranspiration or evaporation from the shallow subsurface recharges the aquifers. The ridge-top crystalline and slate fractured-rock aquifers are recharged directly by precipitation. The shallow valley-fill and outcropping carbonate-rock aquifers are also recharged directly by precipitation, in addition to subsurface flow from the adjacent uplands. The carbonate-rock and deeper valley-fill aquifers receive water primarily from downward flow through permeable glacial sediments and subsurface flow from the adjacent ridges. Streams in the study area may gain or lose water along a given reach under natural conditions or as a result of induced infiltration, the diversion of stream water into the aquifer due to pumping stress. Locally, the carbonate rock aquifer may be semiconfined by overlying low-permeability glacial lake-bottom sediments.

Groundwater leaves the flow system in several ways. It may discharge to streams or lakes, or exit the modeled area as subsurface aquifer flow or underflow beneath surface-water bodies. Withdrawals for residential, industrial, or agricultural use also remove water from the flow system, although some of this water returns to the subsurface either directly through a subsurface discharge (septic system), or, as near Limecrest Quarry, indirectly via surface-water discharge to a losing stream reach.

Simplifying assumptions

The hydrogeologic framework of the study area is complex. The aquifers consist of fractured rock and unconsolidated sedimentary deposits. Groundwater flows under unconfined (water-table) or semiconfined conditions. The sedimentary deposits vary greatly in their lithologic and hydrologic properties, and in their spatial distribution throughout the study area. The fractured-rock aquifers include dolomite, limestone, slate, granite, gneiss, quartzite and marble, with widely varying hydrologic properties. The water-table elevation varies by several hundred feet between hilltops and the valley floor. A realistic simulation of the groundwater flow system requires an accurate representation of the hydrogeologic framework. Nevertheless, some simplifying assumptions were necessary to model the complex system.

<u>Assumptions</u>

1) The carbonate fractured-rock aquifer is 300

feet thick.

2) The igneous and metamorphic fractured-rock aquifer that includes the Franklin Marble is 200 feet thick.

3) The aquifer thickness of the Martinsburg Formation is 200 feet.

4) Groundwater flow in the fractured-rock aquifers approximates flow in a porous medium on a regional scale.

5) Drawdown due to pumping stresses at Limecrest Quarry has stabilized. Although this may not be strictly true, water-level trends provide no evidence to the contrary.

Discretization and model framework

The model area was discretized using a 2-layer grid of 62 rows and 29 columns, encompassing a geographic area of 12.3 square miles (figs. 11 and 12). The areal dimensions of most of the model cells are 400 by 400 feet. Larger model cells of 400 by 600 feet and 400 by 800 feet are assigned to the southeastern edge of the model to simulate part of the crystalline-rock aquifer. A coarser discretization is feasible here owing to the relative homogeneity of the crystalline rocks compared to the valley-fill sediments. Layer elevations and aquifer parameters represent averaged values throughout the respective cell areas. The model grid is oriented along the principal direction of groundwater flow that follows the northeast-southwest trend of the valley and the strike of the carbonate rock.

The hydrogeologic framework for the modeled area was developed based on available stratigraphic and hydrologic information. Model layer tops and bottoms were chosen largely based on stratigraphic information obtained from driller's well logs and wells installed as part of this study. Stratigraphic information at well locations is summarized in table A-1 and locations of wells are indicated on plate 1. The schematic drawing in figure 13 shows model layers 1 and 2. Layer 1 is the uppermost model layer and simulates the unconfined (water-table) aquifer. Its top is the ground surface and its base coincides with the top of layer 2. Layer 1 includes the upper glacial sand and gravel aquifer and, where present, the underlying semiconfining unit. Both high and low-permeability sediments are included in layer 1 because initial attempts to separate the two presented problems for model calibration; a thin upper layer consisting of permeable glacial deposits "went dry" and impeded model calibration. In areas where layer 1 contains the semiconfining unit, a low leakance is assigned between layer 1 and layer 2, representing the slower downward flow of groundwater through the less permeable sediments. Where permeable sediments make up the bulk of the composite aquifer, the simulated horizontal hydraulic conductivities in layer 1 reflect the higher values associated with these deposits. In places where only permeable sand and gravel overlies carbonate bedrock, layer 1 includes the unconsolidated sediments and the base of the layer corresponds to the top of the carbonate rock aquifer. To simplify model discretization of the crystalline and slate fractured-rock aquifers that flank the valley, the top 100 feet of both is arbitrarily assigned to layer 1, the bottom 100 feet to layer 2. Assignment of appropriate vertical conductivities between model layers 1 and 2 allow flow to occur between the two layers as if each were a single, intact aquifer.

The top of layer 2 everywhere corresponds to the bottom of layer 1. Layer 2 may alternate between unconfined and confined conditions depending upon the relationship between the simulated head values and specified top of the aquifer. It consists of permeable valley-fill deposits underlying glacial lake-bottom semiconfining units, the carbonate rock aquifer (except for small areas where it crops out and is included in layer 1), and the lower 100 feet

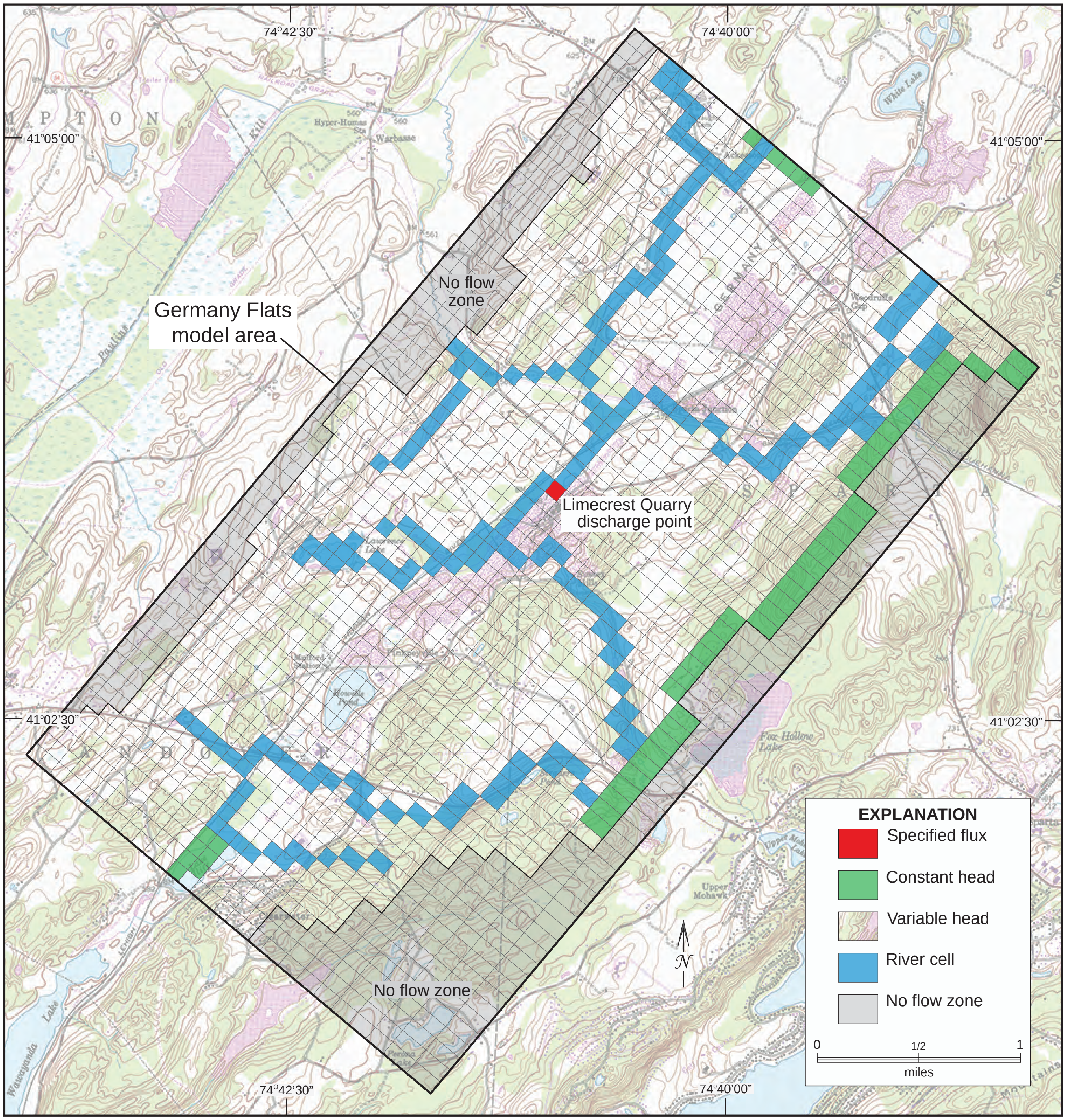


Figure 11. Boundary conditions in model layer 1.

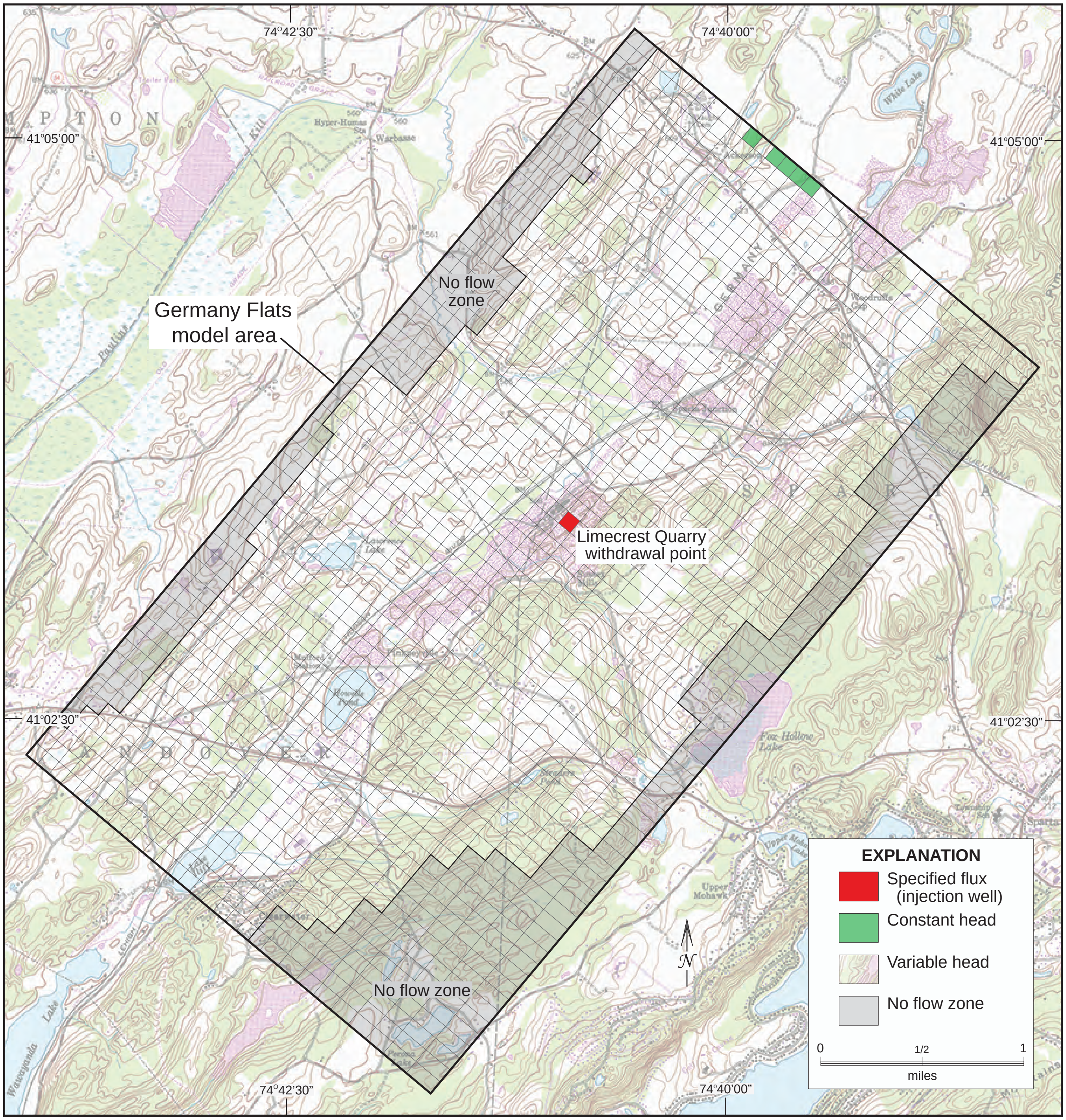


Figure 12. Boundary conditions in model layer 2.

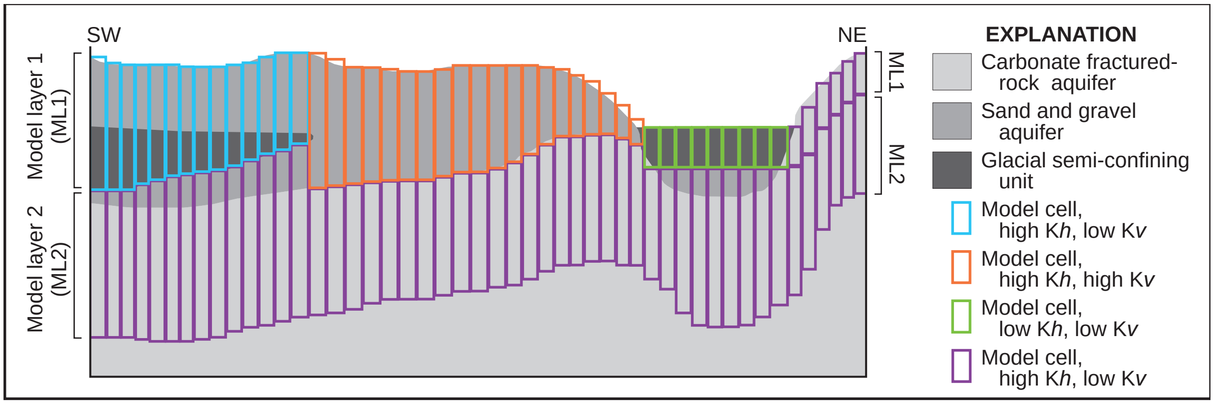


Figure 13. Schematic diagram depicting the assignment of hydrogeologic units to model layers 1 and 2.

of both the crystalline and slate fractured-rock aquifers. As described previously, in many places the carbonate rock forms a composite aquifer with the overlying glacial sand and gravel. Where the composite aquifer is present, the hydraulic conductivity of layer 2 is modified to reflect an average permeability of both units.

Boundary conditions

Boundary conditions in MODFLOW consist of three types 1) constant head, 2) specified flux, and 3) head-dependent flux. A constant head is assigned to a water-level elevation that remains unchanged during the simulation. The specified-flux boundary is an assigned inflow or outflow; a discharging well, for example. A special case of the specified-flux boundary is the "no-flow" boundary in which the flux across a cell is specified as zero. Head-dependent-flux boundaries allow the volume of water entering or leaving a cell to be determined by heads in adjacent cells and user-assigned cell conductances. The MODFLOW River Package and General Head Boundary (GHB) Package (McDonald and Harbaugh, 1988) were used to assign head-dependent flux boundaries in this model. Cells within the model boundary to which no boundary condition is assigned are termed "variable-head" cells, as the head is determined by flows into and out of the cell from adjacent cells.

Where possible, model boundaries were chosen to correspond with natural hydrologic boundaries such as groundwater divides. The northwestern edge of the model generally corresponds to a surface-water divide at the top of the slate bedrock ridge. It is assumed that the surface-water and groundwater divides coincide and that no groundwater flows across this boundary. The divide is simulated as a no-flow boundary in layers 1 and 2 (figs. 11 and 12).

The southeastern limit of the modeled area is specified by either a constant-head or no-flow boundary. Groundwater flowing into the model area along the crystalline bedrock ridge is represented by constant-head cells in layer 1. The constant-head cells are set to the elevation of the water table based on field measurements and preliminary water-table contour maps. Model cells in layer 2 that underlie this boundary are recharged from the downward flow of water from layer 1. Although caution must be used when assigning a constant-head boundary because it can provide unlimited, and possibly unrealistic flows, Anderson and Woessner (1992) note that in some instances constant-head boundaries may be preferable to specified-flux because it is easier to measure head than flow, and specifying head values may simplify model calibration. Based on modeling results, the use of a constant-head boundary to represent inflow along the bedrock ridge is valid because flow is not affected by pumping well stress; simulated flow in this area is nearly identical whether or not wells are pumping.

The resulting model-predicted flow rate from these constant-head boundaries, $2.47x10^5$ ft^3/d (1.9 mgd), also seems reasonable. Where the southeastern model boundary coincides with a surface-water divide, and presumed groundwater divide, it is represented by the no-flow boundary surrounding the model in layers 1 and 2, in a manner similar to that described for the slate fractured-rock aquifer.

The northeastern edge of the model is represented by either constant head, no-flow or head-dependent flux boundaries in layers 1 and 2. To simulate down-valley flow into the modeled area in the valley-fill deposits (layer 1) and carbonate-rock aquifer (layer 2), model cells at the boundary were assigned constant-head values based on measured groundwater levels in wells in the vicinity. The amount of flow entering the modeled area along this constant-head boundary, $2.73x10^5$ ft^3/d (2.0 mgd) was determined and deemed reasonable based on aquifer transmissivities and the hydraulic gradient. A stream entering the modeled area in layer 1 is designated as a "river cell" (head-dependent flux boundary) and no flow was assigned to the cell in layer 2 underlying the stream. In the bedrock ridges that flank the valley, groundwater flow is generally downslope, paralleling the northeastern model boundary. Because little or no flow enters or exits the modeled area here, flow is simply constrained to the variable-head cells within the model area.

The southwestern edge of the model crosses 30-acre Lake Iliff. The lake is represented by a constant-head boundary in layer 1. Groundwater flow is predominantly towards the lake, and all water that exits the southern part of the model discharges from it. In reality, there is probably some underflow in the aquifers beneath the lake. The underflow is assumed to be minimal in comparison to streamflow and for simplification it is not simulated.

Recharge

Estimates of long-term average groundwater recharge to water-table aquifers in New Jersey were prepared as part of an update to the New Jersey Statewide Water Supply Master Plan (N. J. Department of Environmental Protection, 1996). In the analysis, it was assumed that, over long time periods, recharge to a watershed equals base flow as determined by hydrograph separation techniques. A long-term average recharge value of 12.1 in/yr was determined for the Paulins Kill River Basin at Blairstown, New Jersey (USGS station 1443500) located many miles downstream of the study area. This value was determined using the Posten hydrograph separation method (Posten, 1984). A value of 15.4 in/yr was calculated from the less conservative sliding interval method (Pettyjohn and Henning, 1979). However, the area being considered in this report is located at the headwaters of the Paulins Kill. Runoff in the study area may be greater than in the rest of the basin due to uplands of steeply-sloping, low-permeability rock aquifers. During calibration, the lower recharge value, 12.1 in/yr, determined by the Posten method provided the best results and was therefore used in the simulation. Applying an areally distributed recharge rate of 12.1 in/yr as a specified flux resulted in a total recharge of 5.2 mgd over the model area.

Surface-water boundaries

Streams and Lake Lawrence, a small man-made lake, were modeled using the MODFLOW RIVER package that simulates flow between surface-water features and the groundwater system. Flow into or out of the stream reach is calculated as:

$$(2) \qquad QRIV = CRIV\ (HRIV - h_{i,j,k})$$

where QRIV is the flow between the stream and the aquifer, CRIV is the hydraulic conductance of the stream-aquifer interconnection, and HRIV is the stream stage, and $h_{i,j,k}$ is the head in the cell node underlying the stream reach, which is the head in the underlying aquifer as determined by the MODFLOW program (McDonald and Harbaugh, 1988).

Streambed conductance is defined as:

(3) CRIV= (K L W) / (HRIV-RBOT)

where K is the vertical hydraulic conductivity of the saturated material beneath the stream, L is the length of the stream as it crosses model cell i,j,k, W is the stream width, and RBOT is the bottom of the streambed.

When the stream stage is below the head in the aquifer, water flows from the aquifer to the stream and is removed from the cell, simulating a gaining stream. When the head in the underlying aquifer is lower than the stream stage, water in the stream seeps into the aquifer, simulating a losing stream reach. Seepage to the aquifer is limited by the specified elevation of the streambed bottom (RBOT). When the head in the aquifer underlying the stream is lower than RBOT, water flows to the aquifer at a constant rate. Flow will not increase even if the head should drop far below RBOT.

For this study, HRIV was estimated from topographic maps and surveyed stream and lake elevations. RBOT was assumed to be 1 foot lower than the stream elevation; many streams in the study are less than 1 foot deep, and a few are more. Stream length, L, was assumed to equal the cell length and ranged from 400 to 800 feet. Stream width was approximated as 10 feet. As noted in the MODFLOW documentation, vertical hydraulic conductivity of the streambed is difficult to determine because field data are usually not available. Therefore streambed vertical conductivity, K, was adjusted during model calibration to provide reasonable agreement between measured and simulated groundwater levels. A value of 6 ft/d provided adequate head matches in most of the modeled area. Higher estimates of K were used for a reach of the East Branch Paulins Kill between Limecrest Quarry discharge point and USGS streamgaging station 0144328 based on October, 1994 seepage run data which indicated that a large volume of water leaks from the stream to the aquifer along this reach. Streambed conductance (CRIV), calculated on the basis of vertical hydraulic conductivity values, ranged from 24,000 to 112,000 ft^3/d per ft.

As previously described, Lake Iliff is represented by a constant-head boundary at the southwestern model boundary. Howells Pond is simulated as part of the water-table aquifer (variable-head cells) so that the water level in the pond is free to fluctuate during the simulation.

Groundwater withdrawals and discharges

Groundwater-withdrawal and water-use data were obtained from the DEP Bureau of Water Allocation which records groundwater withdrawals exceeding 100,000 gpd. Additionally, groundwater is withdrawn by numerous domestic wells. Because the study area is not sewered, much of the water is returned to the aquifer via septic systems. Any consumptive water use is assumed to have negligible impacts on regional groundwater levels. These small-scale domestic withdrawals were not simulated in the model.

Limecrest Quarry imposes the major pumping stress in the study area. The quarry withdraws several million gallons of water per day to dewater the quarry pit. Ponded water on the quarry floor is removed by two wells, piped to two lined settling ponds, and discharged to the East Branch Paulins Kill. The withdrawal point is modeled as a specified-flux boundary, or discharging well, in layer 2, row 28, column 18 (fig. 12). The point where withdrawn water is discharged to the East Branch Paulins Kill is modeled as a specified-flux boundary, or injection well, in layer 1, row 27, column 16 (fig. 11). The volume of water lost between the withdrawal point and the stream discharge point is assumed to be negligible.

Figure 14 shows the rate of dewatering at the quarry from 1945 to 1993. Prior to approximately 1945, dewatering was not necessary (G.A. Brandon, Jr., Limecrest Corp., oral commun., 1995). Estimates by Limestone Products Corporation indicate that pumpage peaked in the 1970's at about 7.4 mgd, before declining in the early 1980's (L. Carroll, Limestone Products Corp., written commun., 1992). Pumpage rates before 1990 are estimated based on pump capacity.

Pumpage rates from 1990 to 1993, which range from 5.6 to 6.7 mgd, are average annual daily flows based on flow meter readings (New Jersey Department of Environmental Protection Private Water Diversion Reports 1990-1993, unpublished). The average daily discharge from December 1992 to May 1993, 6.5 mgd, was selected to simulate mid-1990's average quarry pumpage. This value is the average pumpage for the 6 months prior to the synoptic survey to which the model was calibrated.

The Limecrest Quarry discharge includes sources other than groundwater. Ponded water on the quarry floor includes precipitation that falls directly in the quarry area and direct runoff intercepted by the quarry pit. Conversely, evaporation decreases the volume of water in the quarry. Because there is no reliable way to separate these various components, all of the quarry discharge is assumed to be derived from groundwater sources. Field observations indicate that water enters the quarry primarily as seepage through fractures in the marble and granite.

Water from the Limecrest Quarry dewatering operation is discharged to the East Branch Paulins Kill. To represent this, the total volume of water removed by the quarry wells in model layer 2, is put back into the model in layer 1. Surface-water discharge is simulated by assigning a specified flux ("injection well") to a model River cell.

Simulated aquifer properties

Horizontal and vertical hydraulic conductivity values were assigned to the various hydrogeologic units based on available data and were adjusted during model calibration within

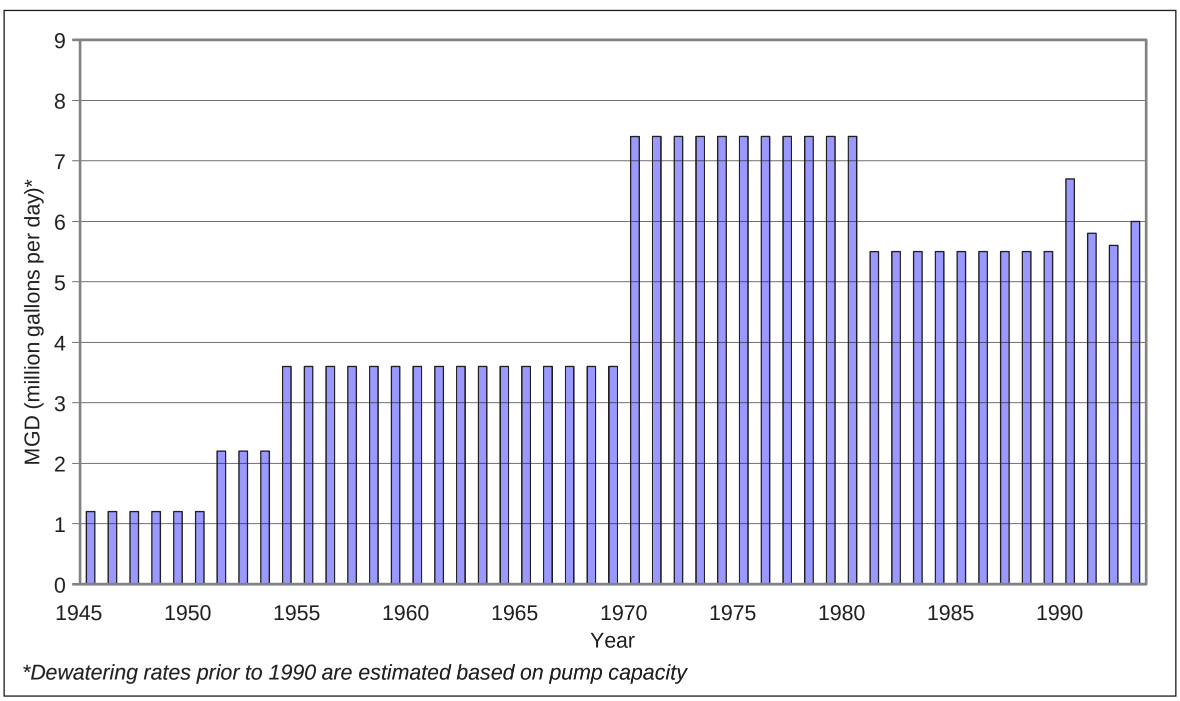


Figure 14. Groundwater withdrawal at Limecrest Quarry.

acceptable ranges of published values for carbonate rock aquifers. The calibrated values for model layers 1 and 2 are presented in figures 15 and 16, respectively.

Leakance values (V_{cont}), which control the rate of leakage between the two model layers, were calculated using the ModelCad386 preprocessing modeling software (Rumbaugh, 1993) based on assigned vertical hydraulic conductivities and model layer thickness. Leakance values vary on a cell-by-cell basis.

The equation for leakance used in this model is:

For cell i,j,k to underlying cell i,j,k+1…

$$(3)\quad Vcont_{i,j,k+1/2} = \frac{1}{((\triangle v_k)/2) / K_{z\,i,j,k}) + ((\triangle v_{k+1})/2) / K_{z\,i,j,k+1})}$$

where: $\triangle v_k$ is the thickness of model layer k

$\triangle v_{k+1}$ is the thickness of model layer k+1

$K_{z\ i,j,k}$ is the vertical hydraulic conductivity of the upper layer in cell i,j,k

$K_{z\ i,j,k}$ is the vertical hydraulic conductivity of the lower layer in cell i,j,k

(McDonald and Harbaugh, 1988)

In the MODFLOW program, leakance values are then multiplied by the horizontal cell dimensions to calculate vertical conductance, the rate of flow between model layers.

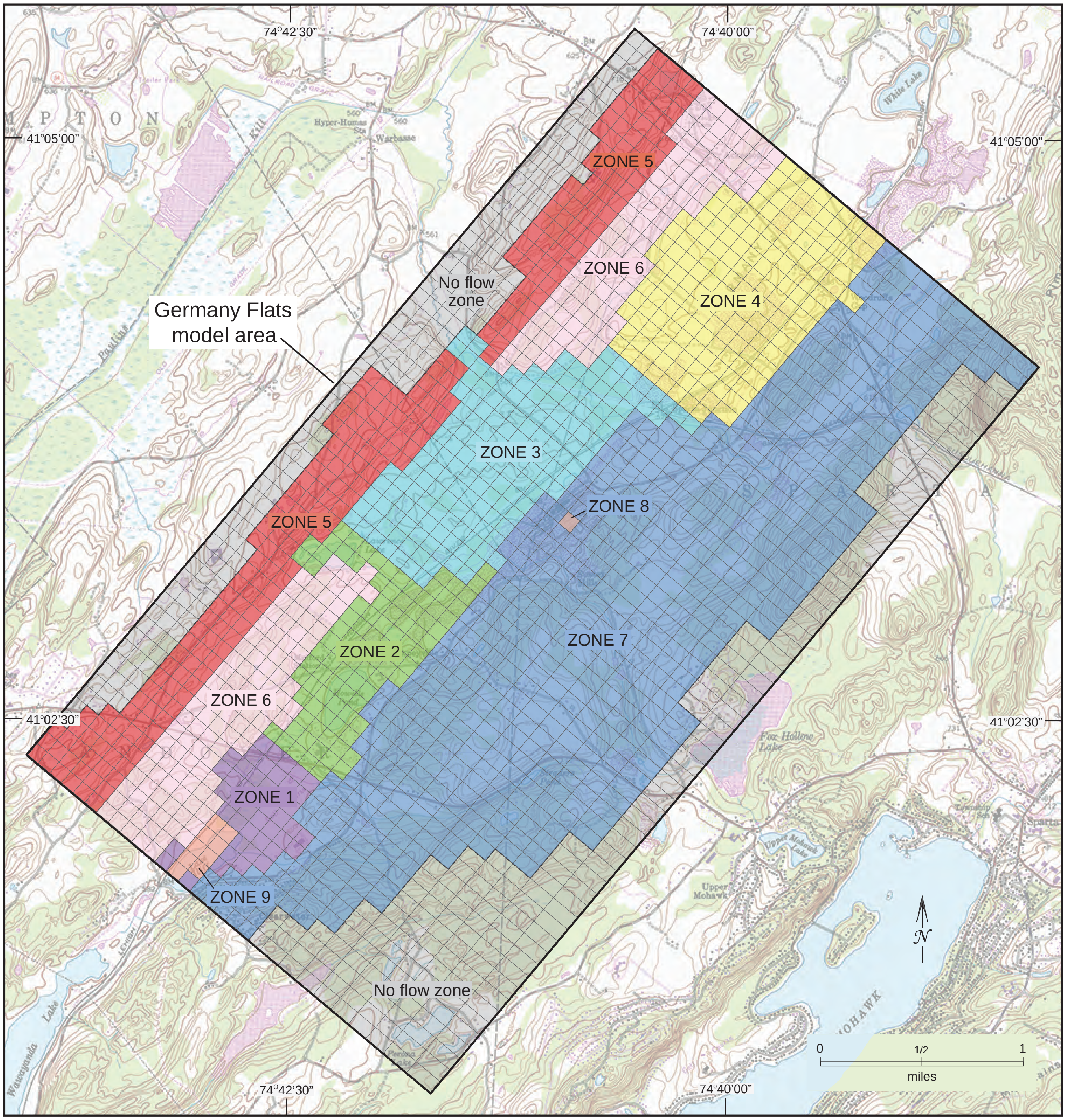
Horizontal and vertical hydraulic conductivity values for the glacial aquifers reflect the widely varying hydraulic properties of the sediments (fig. 15). The open surface water of Lake Iliff is represented with a larger vertical hydraulic conductivity than that of the surrounding aquifer.

Hydraulic conductivity values for the bedrock aquifers and composite carbonate rock/valley-fill aquifer also range over several orders of magnitude (figs. 15 and 16). An anisotropy ratio of 2:1, with higher hydraulic conductivity along bedrock strike, was assigned to the carbonate bedrock aquifer and the composite carbonate rock/valley-fill aquifer in layer 2. The anisotropic nature of fractured-rock aquifers is well documented in the literature, and was evident from aquifer-test data that showed an elongation of the cone of depression in the direction of bedrock strike. A ratio of 2:1 was determined during model calibration. The increased hydraulic conductivity in the direction of bedrock strike may be largely due to enhanced permeability along the intersections of bedding plane partings and fractures.

Because MODFLOW requires that the anisotropy factor be applied as a constant to all cells in a model layer, the Martinsburg slate and Precambrian crystalline rocks in layer 2 are also, by default, assigned a 2:1 anisotropy ratio. Similarly, no anisotropy factor is assigned to the small area of carbonate bedrock in layer 1. Although this may not reflect the true nature of the rock, it is thought to have little effect on the model output.

Accuracy of simulated water levels

Modeled water levels were calibrated to measured water levels obtained during a May 26-27, 1993 synoptic water-level survey, and supplemented with water levels measured at other various times using a trial and error method. It is assumed that the May 1993 values represent average groundwater levels in the study area in the mid-1990's. Figure 17 compares the May 1993 water level in an observation well in the carbonate rock aquifer in Sussex County, New Jersey with long-term water-level trends. The May 1993 water level is in the middle range of the reported maximum and minimum monthly values for 1993 and the



EXPLANATION

Zone 1: K*x* = 70 ft/d, K*y* = 70 ft/d, K*z* = 0.3 ft/d
Glacial-deltaic sand and gravel, and glacial lake-bottom sediments

Zone 2: K*x* = 70 ft/d, K*y* = 70 ft/d, K*z* = 10 ft/d
Glacial deltaic sand and gravel

Zone 3: K*x* = 10 ft/d, K*y* = 10 ft/d, K*z* = 0.001 ft/d
Glacial lake bottom sediments and recent swamp deposits

Zone 4: K*x* = 70 ft/d, K*y* = 70 ft/d, K*z* = 0.03 ft/d
Glacial sand, gravel, silt and clay

Zone 5: K*x* = 1 ft/d, K*y* = 1 ft/d, K*z* = 0.01 ft/d
Slate

Zone 6: K*x* = 20 ft/d, K*y* = 20 ft/d, K*z* = 0.5 ft/d
Carbonate rock with little or no glacial cover

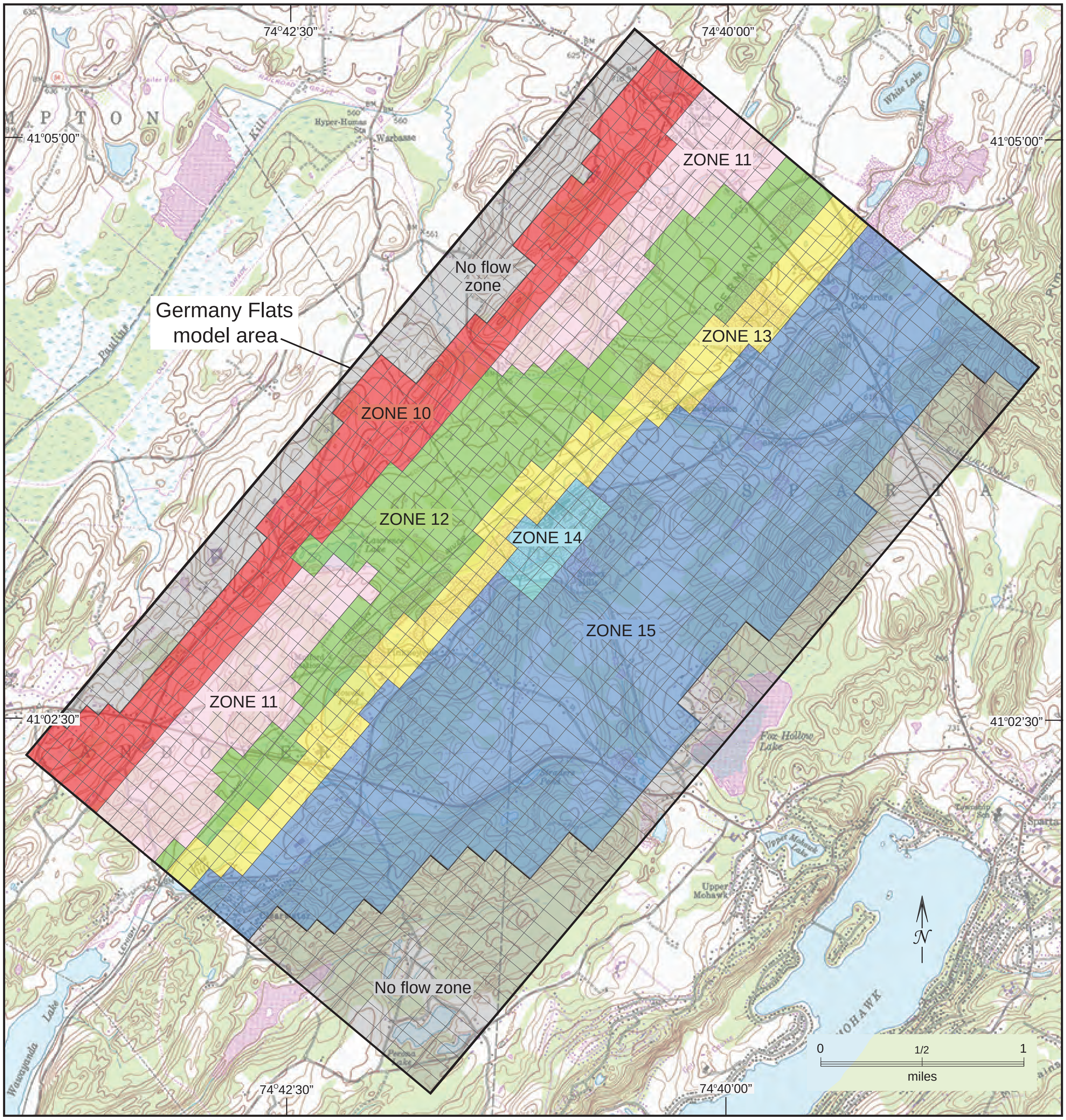
Zone 7: K*x* = 3 ft/d, K*y* = 3 ft/d, K*z* = 0.03 ft/d
Undifferentiated gneiss, granite, marble, and quartzite

Zone 8: K*x* = 74 ft/d, K*y* = 74 ft/d, K*z* = 10 ft/d
Mined out area of Limecrest Quarry

Zone 9: K*x* = 70 ft/d, K*y* = 70 ft/d, K*z* = 3 ft/d
Lake Iliff

No flow zone

Figure 15. Hydraulic conductivity zones in model layer 1. K*x* is value along model rows; K*y* is value along columns; K*z* is vertical hydraulic conductivity.



EXPLANATION

Zone 10: K*x* = 0.5 ft/d, K*y* = 1 ft/d,K*z* = 0.01 ft/d
Slate

Zone 11: K*x* = 10 ft/d, K*y* = 20 ft/d, K*z* = 0.5 ft/d
Carbonate rock with little or no glacial cover

Zone 12: K*x* = 37 ft/d, K*y* = 74 ft/d, K*z* = 0.8 ft/d
Glacial-deltaic sand and gravel and carbonate rock

No flow zone

Zone 13: K*x* = 150 ft/d, K*y* = 300 ft/d, K*z* = 0.8 ft/d
Glacial-deltaic sand and gravel and carbonate rock of the Leithsville Formation

Zone 14: K*x* = 37 ft/d, K*y* = 74 ft/d, K*z* = 0.8 ft/d
Franklin Marble

Zone 15: K*x* = 1.5 ft/d, K*y* = 3 ft/d, K*z* = 0.03 ft/d
Undifferentiated gneiss, granite, marble and quartzite

Figure 16. Hydraulic conductivity zones in model layer 2. K*x* is value along model rows; K*y* is value along columns; K*z* is vertical hydraulic conductivity.

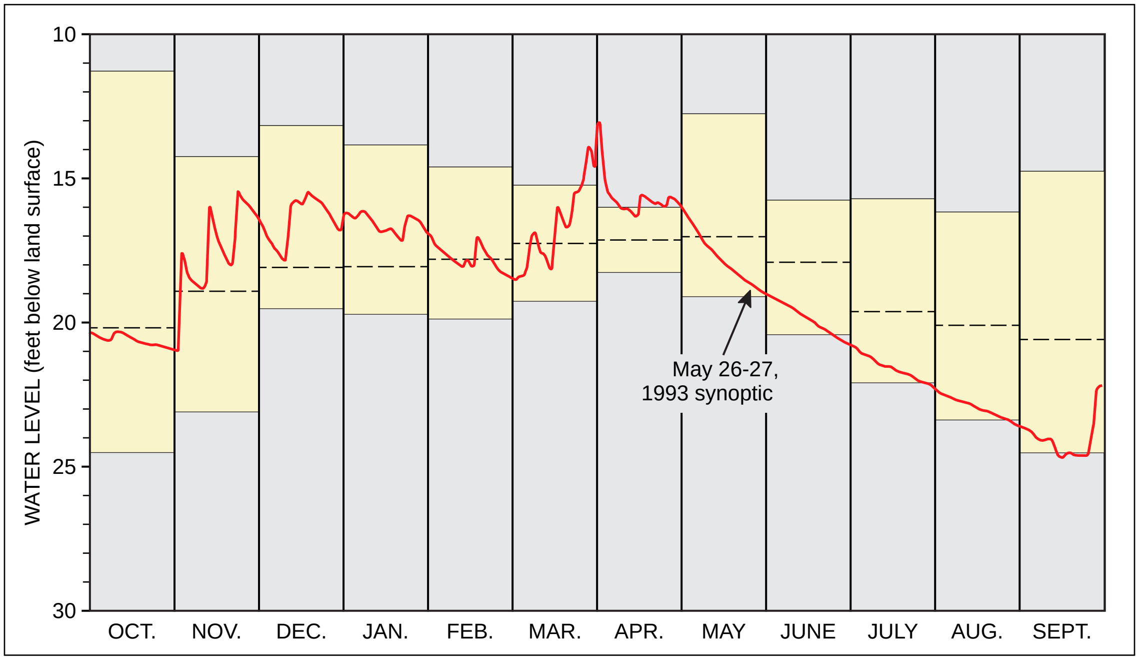


Figure 17. Hydrograph of ground-water level in a well in carbonate rock in Sussex County, New Jersey for water year 1993, and range of monthly maximum and minimum recorded water levels for previous years. Source: U. S. Geological Survey Water-Data Report NJ-93-2.

EXPLANATION

Range between highest and lowest instantaneously recorded monthly water level, prior to the current year.

Average of monthly mean water levels, prior to the current year.

Daily mean water level for the current year.

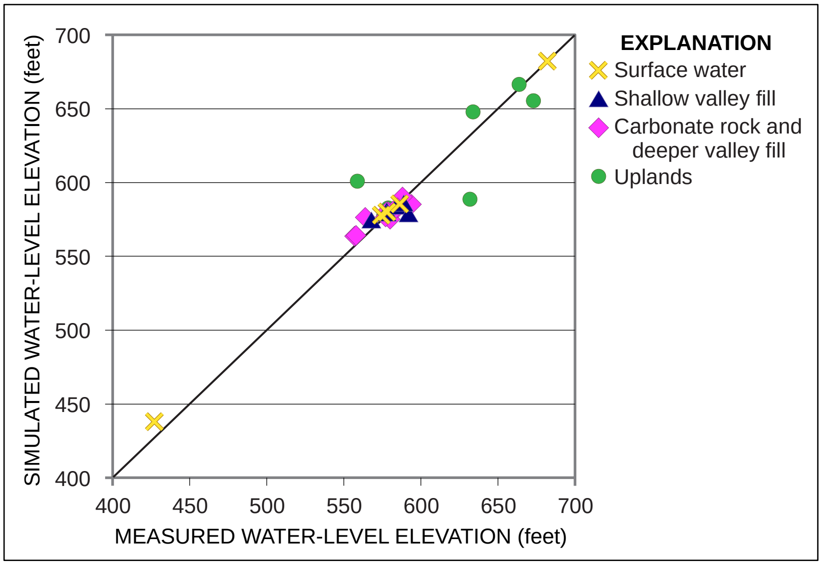


Figure 18. Comparison of measured and simulated water levels.

Table 4. Summary of measured and simulated water levels within model area (ft above msl = feet above mean sea level)

Model cell node			Well or surface-water ID	Elevation (ft above msl)		Residual (ft)	Measurement Date	Measured By
Layer	Row	Column		Measured water level	Simulated water level[1]			
1	17	18	S-11	574.5	573.7	-0.8	4/27/92	Sparta Township
1	17	20	S-8	568.4	574.2	5.8	5/26/93	NJGS
1	19	16	S-9, S-10	565[3]	565.3	0.3	4/27/92	Sparta Township
1	26	10	L-2	592.0	578.6	-13.4	7/15/94	NJGS
1	29	16	S-1, S-4, S-5	552[3]	560.5	8.5	5/26/93	NJGS
1	29	17	S-2, S-3	553[3]	546.4	-6.6	5/26/93	NJGS
1	34	8	A-21	588.8	585.8	-3.0	5/26/93	NJGS
1	39	8	A-25	586.4[4]	585.4	-1.0	6/14/93	NJGS
1	42	12	A-16, A-17, A-18	557.0[2,3]	575.1	18.1[2]	5/27/93	NJGS
1	43	24	S-23	681.8[5]	682.0	0.2	6/8/93	NJGS
1	46	13	A-26	573.7[6]	577.6	3.9	6/1/93	NJGS
1	47	15	A-13	578.0	578.7	0.7	5/26/93	NJGS
1	48	16	A-7	577.8	580.5	2.7	5/26/93	NJGS
1	53	14	A-5	584.0	583.6	-0.4	5/18/93	NJGS
1	61	12	A-27	578.0[7]	578.0	0.0	6/2/93	NJGS
2	1	16	S-16	587.6	590.0	2.4	5/26/93	NJGS
2	4	12	S-15	594.2	585.1	-9.1	5/26/93	NJGS
2	5	16	S-14	585.0	582.6	-2.4	5/26/93	NJGS
2	7	16	S-13	581.8	580.5	-1.3	5/26/93	NJGS
2	8	21	S-21	579.5	582.7	3.2	5/26/93	NJGS
2	13	14	S-12	579.6	575.1	-4.5	5/27/93	NJGS
2	23	12	L-5	578.9	564.0	-14.9	5/27/93	NJGS
2	24	12	L-4	556.9	563.6	6.7	5/26/93	NJGS
2	26	10	L-1	558.0	564.0	6.0	7/15/94	NJGS
2	28	18	S-22	427.1[8]	437.8	10.7	3/31/93	NJGS
2	31	24	S-24	664.4	666.3	1.9	5/27/93	NJGS
2	32	23	S-25	633.9	647.7	13.8	5/20/92	NJGS
2	34	23	S-26	672.6	655.4	-17.2	5/27/93	NJGS
2	39	4	A-30	559.3	600.7	41.4	5/27/93	NJGS
2	41	9	A-20	564.3	576.2	11.9	5/26/93	NJGS
2	42	14	A-15	551.5[2]	573.7	22.2[2]	5/26/93	NJGS
2	47	16	A-12	576.0	577.8	1.8	5/26/93	NJGS
2	48	16	A-8	576.6	576.6	0.0	5/26/93	NJGS
2	53	15	A-3	578.5	579.7	1.2	5/28/93	NJGS
2	56	15	A-2	577.6	580.1	2.5	5/18/93	NJGS
2	61	14	A-1	578.6	580.2	1.6	5/26/93	NJGS
2	61	17	A-24	628.4	588.6	-39.8	5/20/92	NJGS

[1]Calculated head at centroid of cell node
[2]Production well in operation during measurement
[3]Averaged value of measurements from more than one sampling location
[4]Lake Lawrence surface-water elevation
[5]Straders Pond surface-water elevation
[6]Howells Pond surface-water elevation
[7]Lake Iliff surface-water elevation
[8]Limecrest Quarry pit surface-water elevation

previous years.

The measured water levels in wells and surface-water bodies, and the model-simulated water levels, are summarized in table 4. A graph of the relationship between the observed and simulated head values is also shown in figure 18.

Some of the discrepancy between the measured and simulated head values is a result of the way the model head is calculated. The simulated head is an average value determined for the cell node, the centroid of the cell. However, the observation wells fall at various locations and depths within the model cell. No correction has been made for the location or depth of the observation wells. Less agreement between the two values is therefore expected in areas where hydraulic gradients are steep, for example on hill slopes where land surface elevations may vary more than 80 ft within a single model cell, and where the model cell size is large. For these reasons, simulated water levels for upland wells, those on the bedrock ridges, generally show less agreement with measured values than those in the valley aquifers.

Other discrepancies between the simulated and field-measured water levels are due to limitations in the discretization of the geologic framework. For example, the model underestimates the elevation of the water table as measured in Well L-2 (NJDEP 22-32635) screened in the shallow valley-fill aquifer. It is possible that the water-table aquifer is perched in this area as evidenced by low-permeability silt and clay in the subsurface, a large (34 ft) head difference between the measured water level in Well L-2 (NJDEP 22-32635) and adjacent Well L-1 (NJDEP 22-33686) finished in the bedrock aquifer, and the presence of groundwater seeps and wetlands in the vicinity. The localized low-permeability unit is not easily simulated within the more regional scale of the model. Additionally, the water level in the vicinity of the Andover Intermediate Care Center and Andover Convalescent is likely effected by small-scale pumpages of less than 100,000 gpd that were not modeled, leading to higher simulated water levels than were measured in that area. Estimated pumpage for the Andover Nursing Center is between 75,000 to 80,000 gpd and for Andover Convalescent, approximately 20,000 gpd (R. Kolson, Andover Nursing Center, oral commun., 1993.)

Overall, close agreement between the measured and simulated water levels was achieved throughout the modeled area. For the thirty-seven wells used in the model calibration (table 4), the mean error (ME) for the measured head minus simulated head is -1.4, the mean absolute error (MAE) is 7.6, and the root mean-squared error (RMSE) is 12.4. Excluding the six upland wells, in which surface elevations vary considerably, and two near the Andover Nursing Home that may be affected by local pumping, results in a ME of -0.2, a MAE of 4.3 and a RMSE of 6.0. All simulated water levels are considered to be adequate for the intended purposes of the model.

Groundwater levels and flow directions

Figures 19 and 20 are contour maps of the potentiometric surfaces in model layers 1 and 2, respectively. The arrows indicate the general direction of groundwater flow in the valley aquifers. The dashed lines indicate the locations of groundwater divides.

Groundwater flow in the crystalline-rock aquifer and Martinsburg slate is predominantly downslope, normal to the trend of the valley, and locally towards small streams. Away from streams, vertical gradients are downward. A downward flow gradient is expected in these ridge-top aquifers, as the water originates as precipitation that seeps through rock fractures to recharge the aquifers. However, as discussed later in this report, pumping at Limecrest Quarry has enhanced this process though induced infiltration from streams to the aquifer in some upland locations.

Groundwater flow in the valley-fill deposits (fig. 19) is generally towards local discharge points such as streams or lakes. Streams are generally discharge areas, although losing reaches exist. A small steep-sided cone of depression in the immediate area of Limecrest Quarry represents the seepage face at the quarry wall and the lowered water table surrounding the quarry. An east-west trending groundwater divide is located about a half mile northwest of the quarry. North of this divide, groundwater flow in the East Branch Paulins Kill Basin is predominantly down-valley (southwestward) or towards streams which flow westward through a gap in the slate bedrock ridge and into the adjacent valley. Directly south of the divide, groundwater flows towards the quarry, opposite

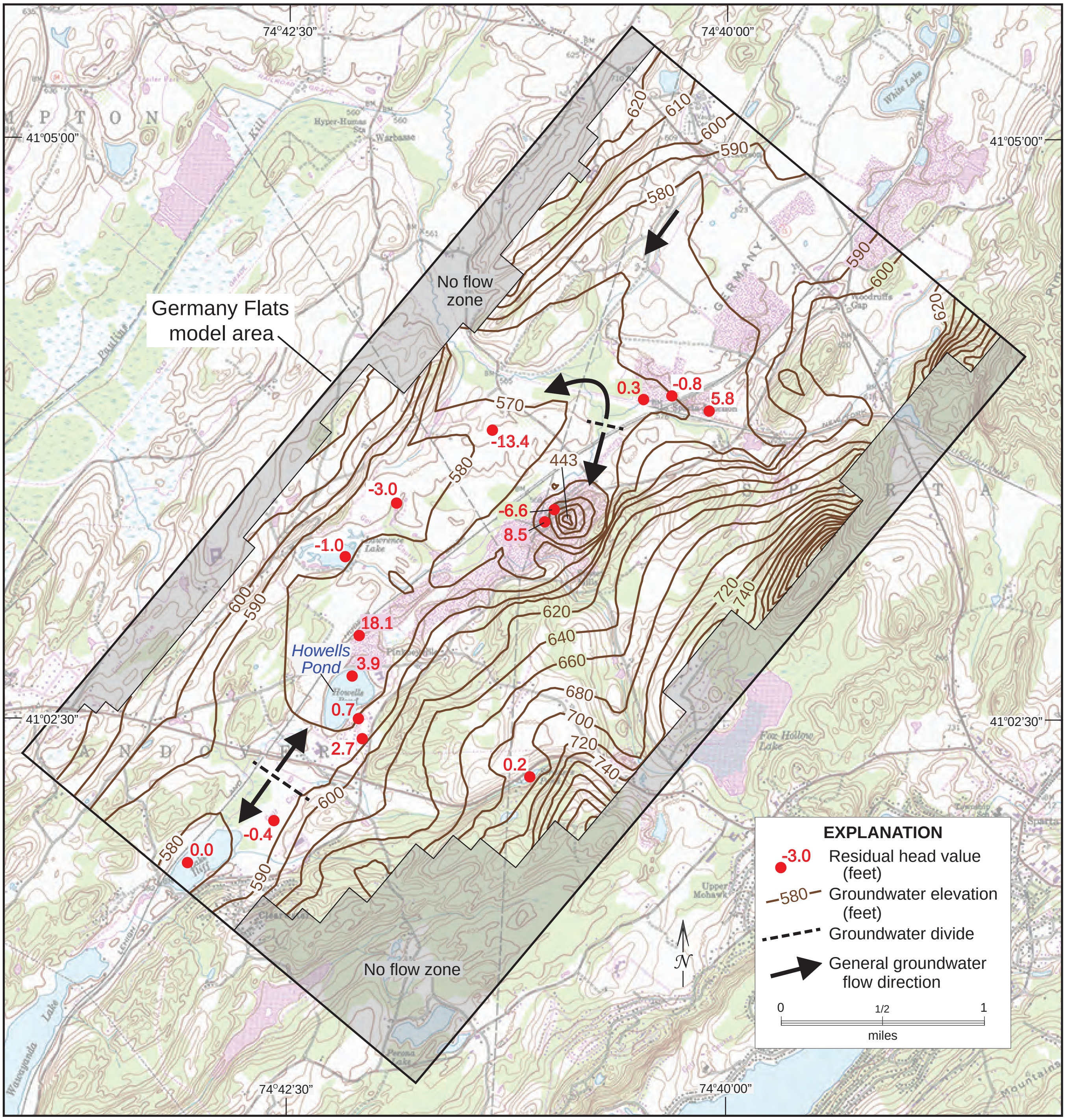


Figure 19. Simulated piezometric surface in model layer 1 with Limecrest Quarry pumping 6.5 mgd. The difference between measured and simulated water levels is shown at well locations. (A minus sign indicates that the simulated water level is lower than the measured water level.)

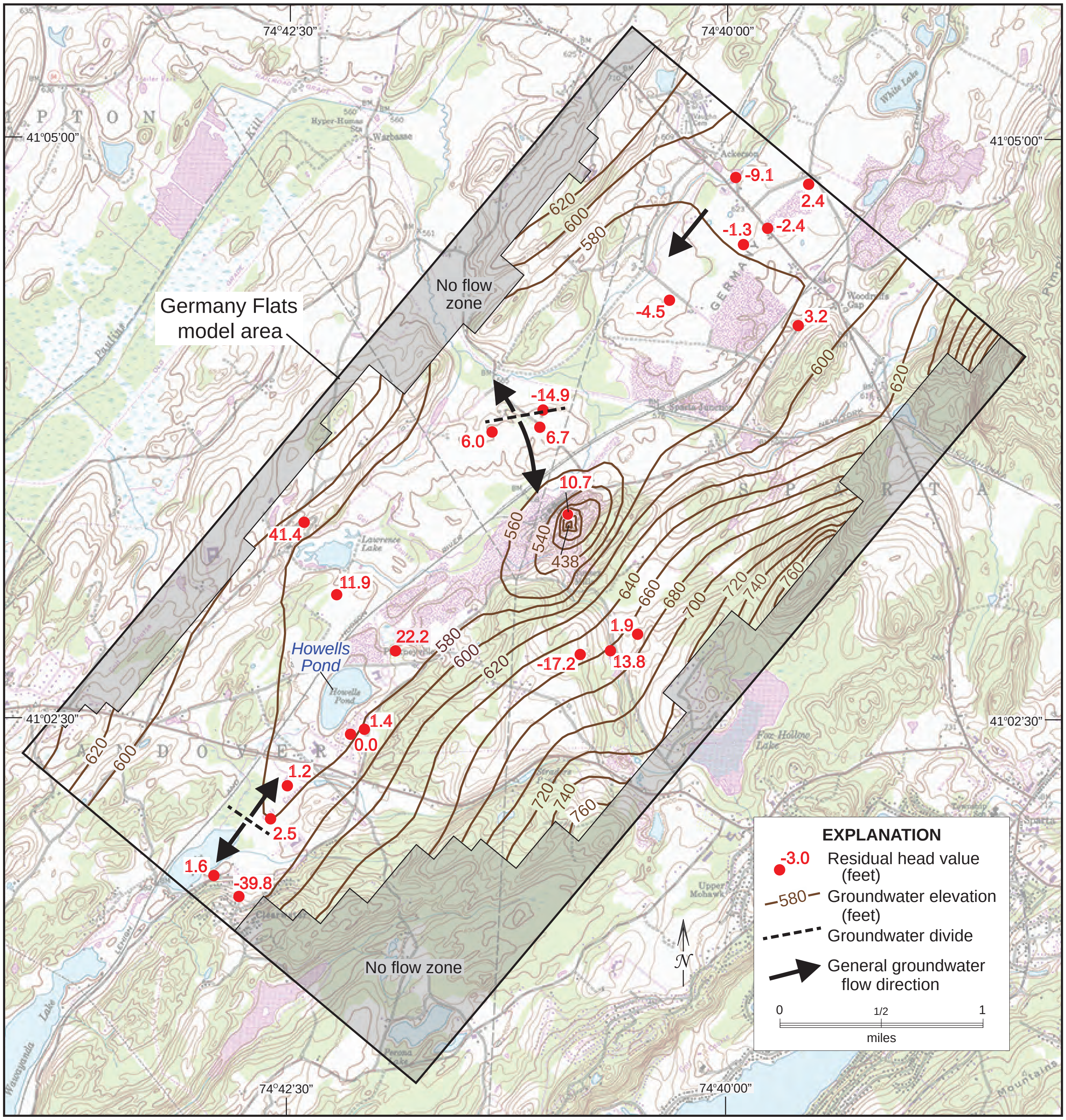


Figure 20. Simulated potentiometric suface in model layer 2 with Limecrest Quarry pumping 6.5 mgd. The difference between measured and simulated water levels is shown at well locations. (A minus sign indicates that the simulated water level is lower than the measured water level.)

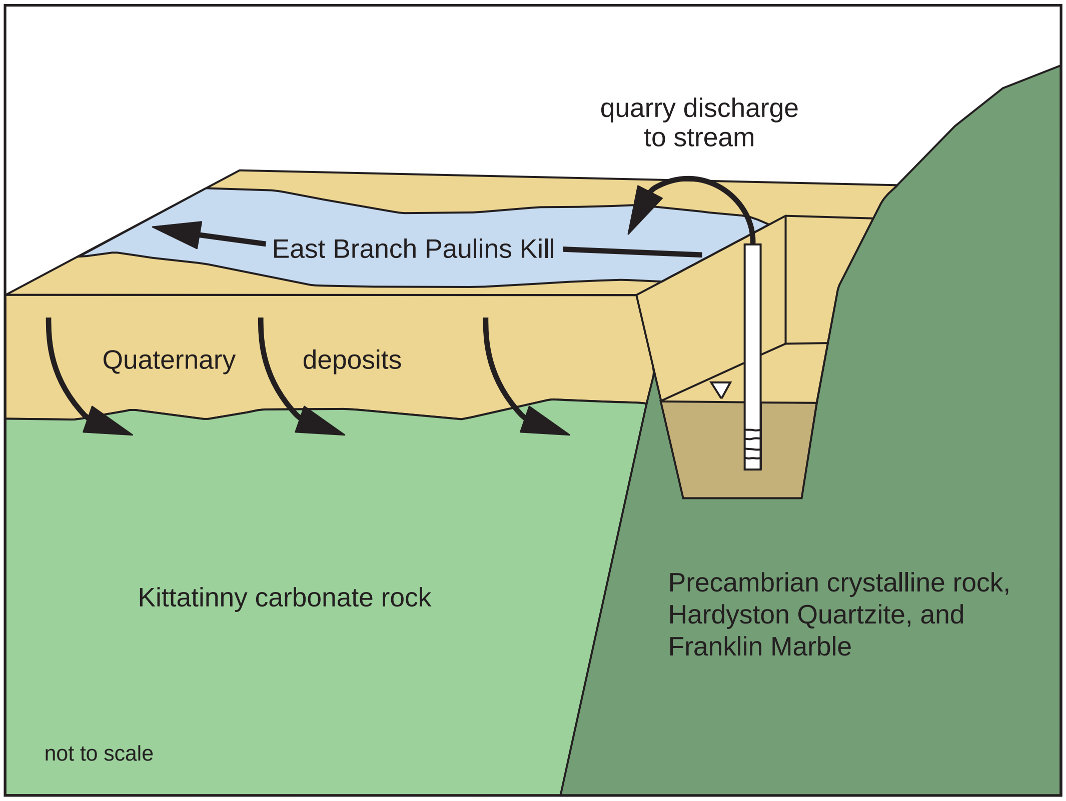


Figure 21. Schematic diagram showing circulating groundwater flow pattern in area of Limecrest Quarry.

the direction of flow in the East Branch Paulins Kill. Another groundwater divide is almost 2 miles southwest of the quarry, between Howells Pond and Lake Iliff. From this location, groundwater flows northeastward following the trend of the valley and southwestward to discharge to Lake Iliff in the Pequest River Basin.

In the carbonate rock aquifer and composite carbonate rock/valley-fill aquifer (fig. 20), groundwater flow is dominated by dewatering at Limecrest Quarry. Most flow within the deeper aquifers is towards the quarry where it is then removed by the quarry pumping wells. Some flow is upward to discharge to streams or Lake Iliff.

As discussed previously, downward vertical gradients exist between the shallow and deeper aquifer throughout much of the modeled area due to lowered heads in the carbonate rock aquifer caused by dewatering at Limecrest Quarry. Northwest of the quarry, where downward vertical gradients are coupled with northwestward flow in the East Branch Paulins Kill and southeastward flow in the shallow and deep aquifers, a circulating groundwater flow pattern has developed (fig. 21). Water is removed from the bedrock aquifer by the quarry pumps and discharged to the East Branch Paulins Kill. Some of the water leaves the study area as streamflow and some seeps downward into the underlying valley-fill deposits and carbonate-rock aquifer. Water in these aquifers may then flow back to the quarry to repeat the cycle. The concept of recirculating water in the area of Limecrest Quarry is supported by measured streamflow losses between the quarry surface-water discharge point and a downstream gage, and downward vertical gradients in streambed piezometers in the East Branch Paulins Kill.

Groundwater levels and flow patterns in both the shallow and deep aquifers have been impacted by pumping at Limecrest Quarry. However, the extent of the influence is not readily seen by observing only the present-day flow system. The changes in water levels due to pumping at the quarry become apparent by comparing present-day and pre-quarry flow conditions. This comparison is made in subsequent sections of the report.

Base flow

Stream baseflows, defined as flow from aquifers to streams minus leakage from streams to aquifers, were calibrated to field-measured values where possible. In October 1994, after a long period without precipitation, measured streamflow at station EBPK-1 on the East Branch Paulins Kill under low base flow conditions was 10.5 ft^3/s (6.8 mgd), which is assumed to represent baseflow conditions. This value was considered to represent baseflow due to the preceding dry climatic conditions occurring in the basin. An average annual base flow for water year 1993, 21.4 ft^3/s (13.8 mgd), was determined for USGS station 0144328, located approximately 0.5 mile downstream of the modeled area, using the sliding interval base-flow-separation technique. Because Station 0144328 drains a larger area than that modeled, 21.4 ft^3/s is considered to be an upper limit for the simulated base flow. The simulated base flow for the East Branch Paulins Kill Basin, 14.9 ft^3/s (9.6 mgd), falls between the upper and lower extremes and is therefore believed to be a reasonable estimate of mid 1990's average base flow.

Throughout the year, stream flow in the East Branch Paulins Kill fluctuates widely, and the Limecrest Quarry surface-water discharge makes up a variable percentage of this flow. Because the steady, continuous release of water from the quarry is indistinguishable from the natural base flow contribution in stream hydrographs, any calculation of base flow on this stream reach inherently includes the contributions from the quarry discharge. Figure 22 shows base flow in the East Branch Paulins Kill at station 0144328 compared with reported quarry discharge from October 1992 to September 1993. During this period, the volume of water discharged by the quarry was 17 percent to more than 100 percent of the total base flow. The quarry discharge exceeded base flow at the downstream gage in October 1992, July 1993 and August 1993 despite contributions from intervening tributaries. This indicates that during the summer and fall, when ground-water levels are typically declining or at their lowest, large volumes of streamwater are lost to the aquifer between the quarry discharge point and the gaging station. Some of this loss occurs as water evaporates from the surface of the wetland area surrounding the quarry discharge point. However, as shown in figures 9 and 10, downward hydraulic gradients in streambed piezometers in the East Branch Paulins Kill,

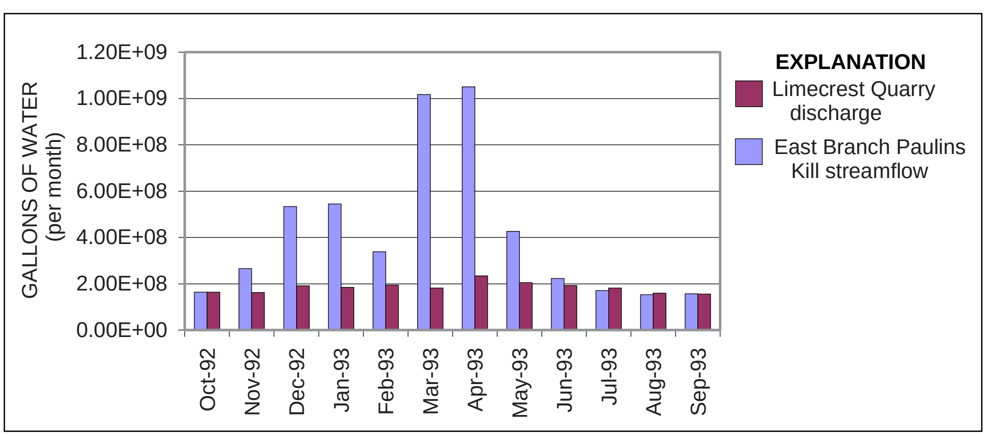


Figure 22. Comparison of Limecrest Quarry discharge with baseflow in the East Branch Paulins Kill at USGS stream gauge station 0144328.

measured at various times, suggest that part of the loss occurs as seepage of water through the streambed along losing stream reaches.

No measured streamflow data are available for the modeled area of the Pequest River basin. However, observations at the outlet of Lake Iliff indicate that, at times, no surface flow exits the lake. This suggests that baseflow in the modeled part of the Pequest River Basin is minimal.

Model-simulated flows from streams to the aquifer exceed the calculated modeled flow from the aquifer to streams; baseflow in the modeled part of the Pequest River Basin shows a net loss of 0.68 ft^3/s (0.4 mgd). This “negative baseflow” is an artifact resulting from the use of head-dependent flux boundaries (the MODFLOW River Package) to simulate streams. The River Package allows a cell containing a stream to contribute water to the underlying aquifer under a unit gradient even when the simulated water level in that cell falls far below the stream bottom; a situation which may cause a real stream to go dry. This condition occurs in the model only when the Limecrest Quarry wells are pumping. When the quarry wells are off, streams gain more water than they lose. (Groundwater discharges to Lake Iliff in both the pre-quarry and mid-1990’s simulations.) Although in reality, baseflow in the modeled part of the Pequest River Basin is not less than zero, the introduction of an additional 0.4 mgd to the model water budget is not believed to significantly affect the overall accuracy of the budget. Further refinement of the baseflow estimate could be gained by limiting the amount of water seeping from losing stream reaches by using the MODFLOW Drain package, which limits flow to the aquifer once the head in the aquifer drops below a designated drain elevation.

In both the Pequest and Paulins Kill Basins, some of the model-simulated losing stream reaches coincide with intermittent streams. Many are near Limecrest Quarry where steep downward hydraulic gradients induce streamwater to leak into the underlying aquifer. The losing reaches occur in both upland streams and streams crossing the valley.

The lack of detailed base flow data, lack of a long-term stream hydrograph, in combination with the uncertainty in the streambed conductance term do not allow for a rigorous calibration of the model discharges to measured baseflow values. Based on available streamflow data, however, the simulated baseflows are reasonable, and provide an independent check on the accuracy of the model.

SIMULATION OF PRE-QUARRY GROUNDWATER FLOW

Flow in the pre-quarry valley was simulated and compared with modeling results for the mid-1990’s simulation to assess how the quarry dewatering has affected groundwater flow in the valley. The steady-state groundwater-flow model for the pre-quarry scenario is identical to the previous model, with the following exceptions:

1) The specified-flux boundary in layer 2, (row 28, column 18) which simulates quarry dewatering was removed; the pumping well was “turned off”.

2) The specified-flux boundary in layer 1 (row 27, column 16) which simulates the Limecrest Quarry discharge to surface water was removed.

3) The higher Kh value of 74 ft/d used to simulate the mined-out area of the quarry (zone 8 in fig. 15), was reduced to 3 ft/d to match the value assigned to the surrounding crystalline rocks.

Groundwater budget

Figures 23 and 24 are a comparison of the pre-quarry groundwater budget with the simulated present-day groundwater budget. Recharge to both models is the same.

Pre-quarry simulated groundwater flow

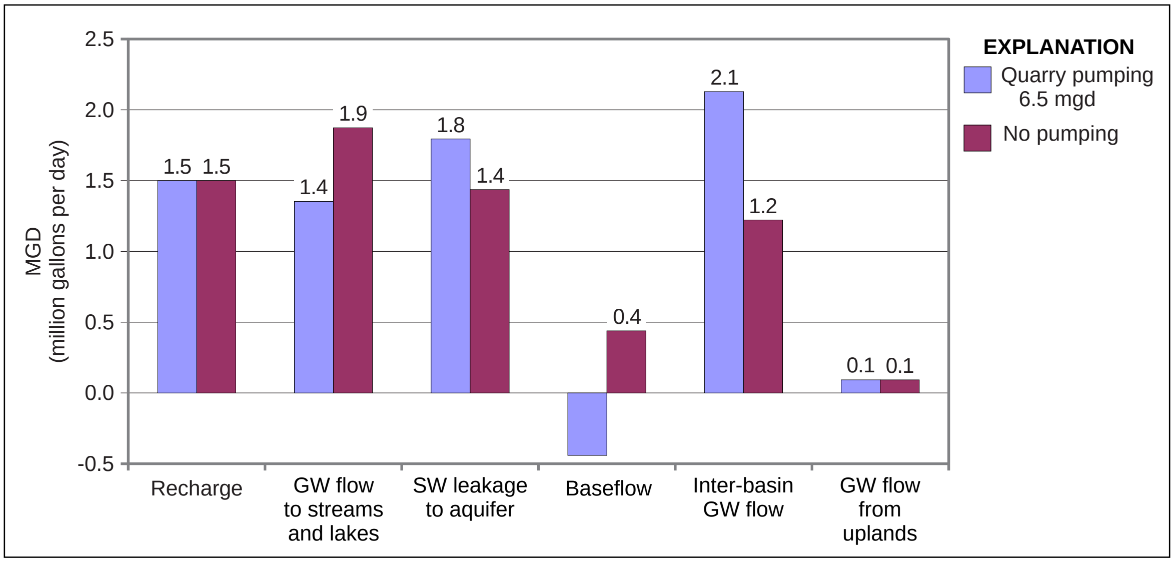


Figure 23. Simulated water budget for the modeled area of the Pequest Basin.

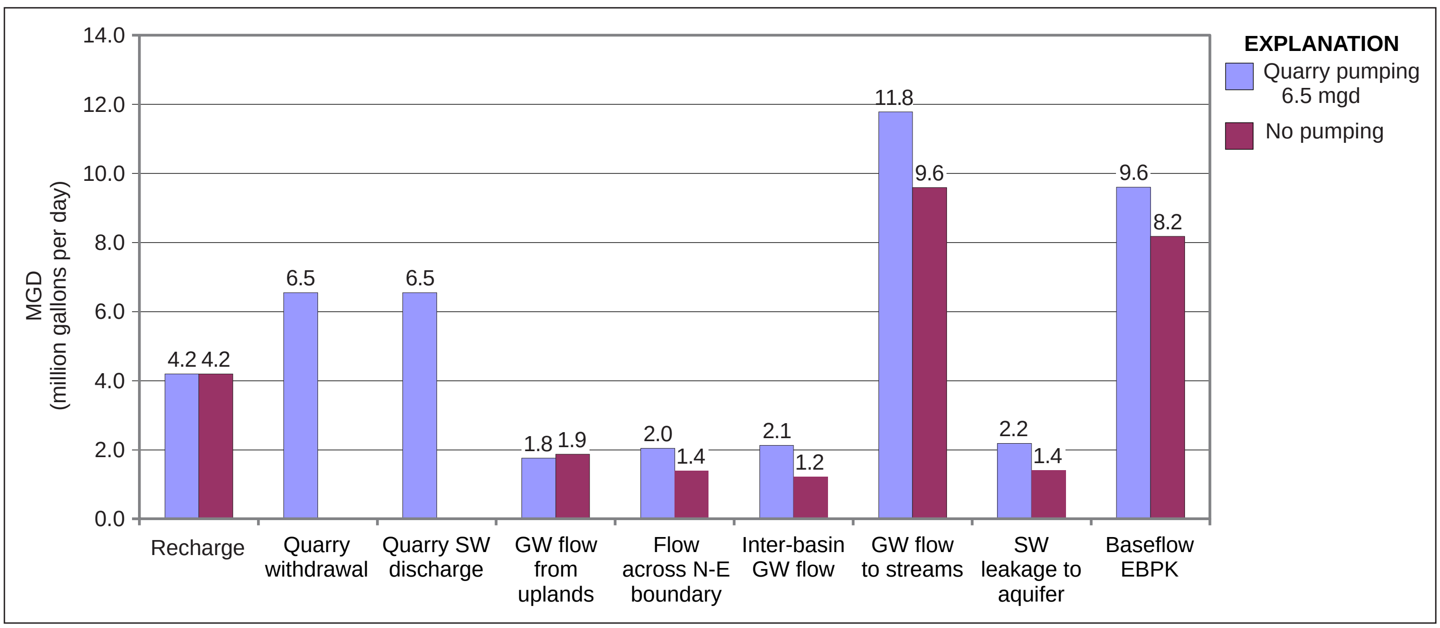


Figure 24. Simulated water budget for the modeled area of the East Branch Paulins Kill Basin.

Table 5. Summary of historic static water level and simulated level at well location (ft above msl = feet above mean sea level).

Model cell node			**Well or surface-water ID**	**Elevation** (ft above msl)		**Residual**[2] (ft)	**Measured date**	**Measured by**
Layer	Row	Column		Measured water level[1]	Simulated water level[1]			
1	51	14	A-6	582	584.8	3	3/1/59	Elvin Hill
1	47	15	A-13	586	583.9	-2	10/1/59	Raymond S. Sipple & Bro.
2	45	44	A-14	587	582.4	-5	2/26/57	Donald Kitchen
2	28	8	L-3	565	573.6	9	7/1/54	Raymond S. Sipple & Bro.

[1] Measured water level as reported on NJDEP well record.
[2] A minus sign indicates that the simulated water level is lower than the measured value.

into the modeled area along the Precambrian crystalline upland boundary, 2.51 x $10^5 ft^3/d$ (1.87 mgd), is very similar to the simulated mid-1990's flow of 2.46 x $10^5 ft^3/d$ (1.84 mgd) indicating that quarry pumping did not result in increased flow from this model boundary.

Groundwater flow into the modeled area in the glacial and carbonate rock aquifers along the northeast model boundary increases by 8.76 x 10^4 ft^3/d (0.7 mgd), or approximately 47%, when the quarry wells are pumping; more water is induced to flow across the boundary when the quarry is dewatering. The impact of quarry dewatering therefore extends beyond the northeast model boundary. It is possible that pumping at Limecrest Quarry has diverted water from the adjacent Wallkill River Basin although this can not be confirmed by available data.

Stream leakage to the aquifer is greater when the quarry pumps are active. In the mid 1990s, 6.2 ft^3/s (4.0 mgd) of stream water seeps into the underlying aquifer, and streams lost water along approximately 32 percent of their total length. In the pre-quarry scenario, only 4.4 ft^3/s (2.8 mgd) leaked from streams to the aquifer, and losing reaches made up approximately 27 percent of the total stream lengths. The difference is due to induced infiltration of stream water within the quarry cone of depression. When the quarry stress is removed, many of the streams within the quarry cone of depression are shown to be gaining rather than losing.

Simulated base flow in the East Branch Paulins Kill Basin, as defined by total discharges from groundwater to surface water minus total leakage from streams to the aquifer, is 12.7 ft^3/s (8.2 mgd), or 2.2 ft^3/s (1.4 mgd) less, when the quarry pumps are inactive. Streamflow in the East Branch Paulins Kill Basin is therefore augmented by the quarry discharge. The additional flow in the East Branch Paulins Kill when the quarry wells are pumping is derived from the quarry surface-water discharge, groundwater diverted from the Pequest River Basin, and increased flow from the northeastern part of the study area near the boundary with the Wallkill River Basin.

Simulated base flow in the pre-quarry Pequest River Basin is 0.68 ft^3/s (0.4 mgd). As discussed previously, when the quarry wells are on, there is a net loss of stream water to the aquifer in this basin. Therefore, in the pre-quarry valley, more groundwater flowed in the modeled area of the Pequest River Basin than at present. Dewatering at Limecrest Quarry has diverted groundwater flow from the Pequest River Basin.

The accuracy of the pre-quarry base flow estimates is difficult to determine because no local stream hydrographs are available for the period prior to the installation of streamflow gaging station 0144328 in 1992. Qualitatively, however, the observed base flow trends are reasonable. Base flow in the East Branch Paulins Kill Basin increases at the expense of base flow in the Pequest River Basin as groundwater is diverted from the Pequest River Basin.

Accuracy of simulated pre-quarry water levels

Because historic water-level data for the pre-quarry valley are sparse, a rigorous calibration to pre-quarry water levels in wells was not possible. Limecrest Quarry has probably affected groundwater levels since at least the 1940's when dewatering began. However, several static water-level measurements from wells drilled prior to 1970 were used as a guide to simulating pre-quarry water levels. Their usefulness is limited because they were collected using unknown methods in different years and under diverse seasonal conditions. Only one of the reported measurements is from a well in which a surveyed water-level elevation was obtained as part of this study, Well A-13 (NJDEP 22-04399). Greater weight was therefore given to this measurement during model calibration due to the better estimate of the actual groundwater elevation and the

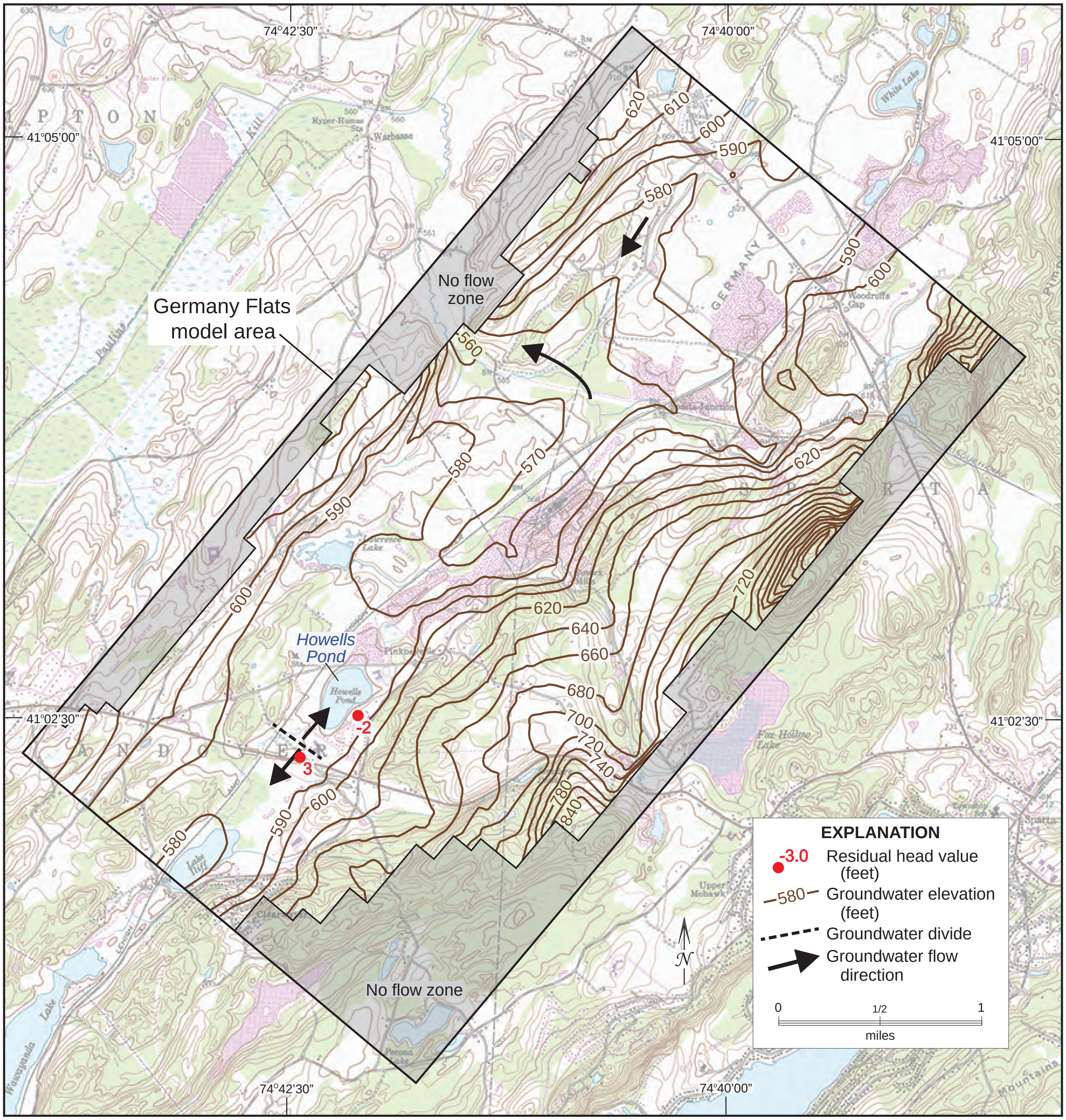


Figure 25. Simulated pre-quarry peizometric surface in Model Layer 1. The difference between simulated and historic static water levels is shown at well locations. (A minus sign indicates that the simulated water level is lower than the historic static water level.)

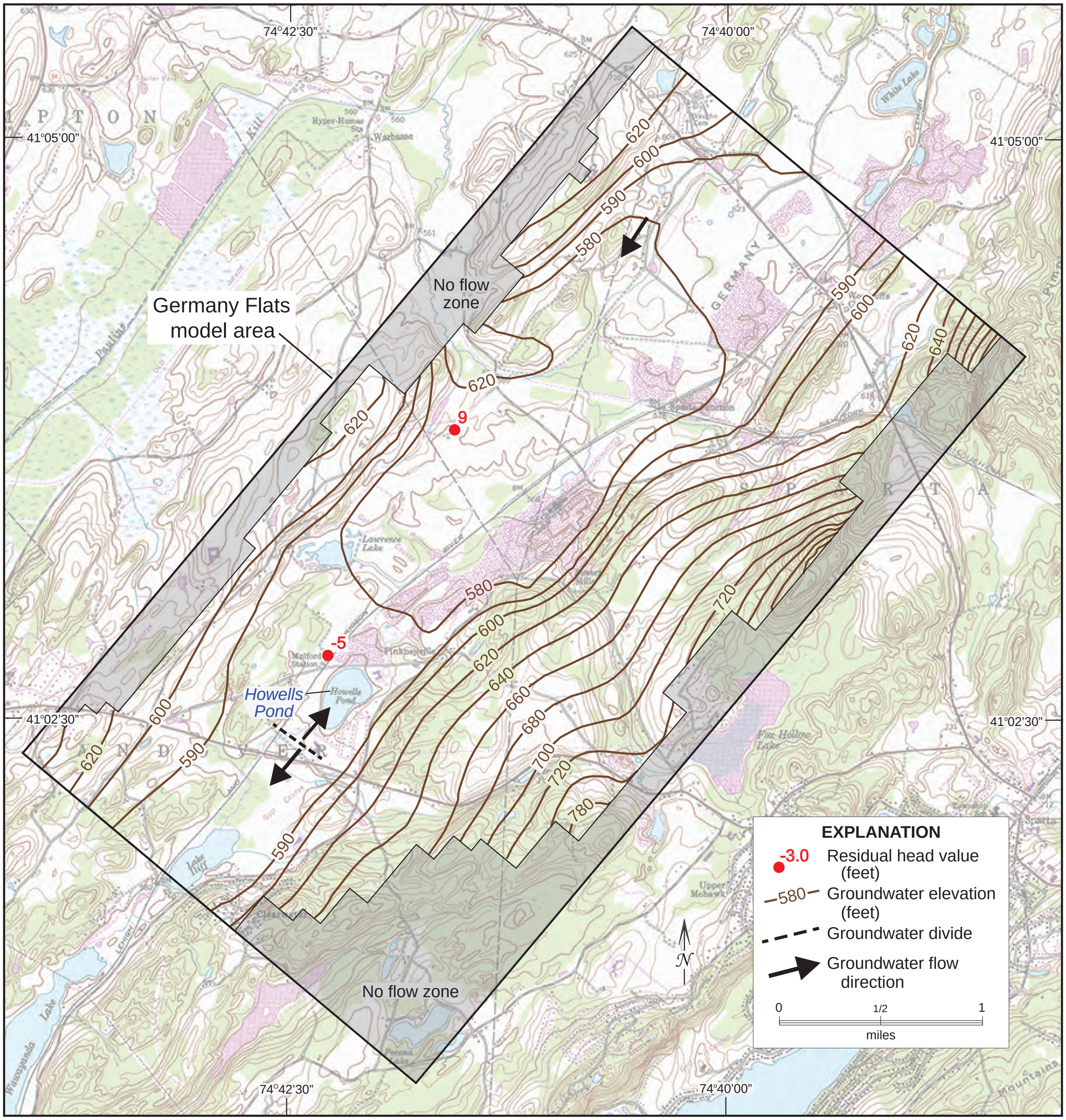


Figure 26. Simulated potentiometric surface in Model Layer 2. The difference between simulated and historic static water levels is shown at well locations. (A minus sign indicates that the simulated water level is lower than the historic static water level.)

proximity of the well to Howells Pond, a focus of this report. The numerous limitations in the pre-1970 data precluded a close match between reported and simulated heads. The residual head value between the historic static level and the simulated head is summarized in table 5. Despite the uncertainty in the historic water-level data, the simulated heads appear to be reasonable. A close match was obtained at Well A-13 (NJDEP 22-04399) near Howells Pond.

Pre-quarry groundwater levels and flow directions

Figures 25 and 26 show simulated groundwater levels in model layers 1 and 2, respectively, prior to pumping at Limecrest Quarry. Groundwater-flow directions in the crystalline fractured-rock aquifer resemble present-day flow paths, except in the immediate area of the quarry (compare figs. 19 and 25 and figs. 20 and 26). Groundwater generally flows downslope following the surface topography or it discharges to upland streams. Away from streams, groundwater moves from the upper to the lower part of the aquifer following downward hydraulic gradients. Figures 27 and 28 show the drawdown in model layers 1 and 2, respectively, due to the quarry pumping stress. Drawdown in the crystalline rocks, which form the southeastern ridge, ranges from less than 1 foot near the hilltop to 139 feet in the Franklin Marble at the quarry pumping wells.

Pre-quarry groundwater-flow directions in the Martinsburg aquifer are nearly identical to present-day flow paths. As in the crystalline bedrock ridge, groundwater flow is downslope and from shallow to deeper parts of the aquifer. The modeling indicates that several feet of drawdown has occurred in this ridge-top aquifer, most likely in response to lowered groundwater levels in the valley aquifers.

In the valley-fill aquifer, under pre-quarry conditions, groundwater flow in the East Branch Paulins Kill Basin was predominantly towards streams and lakes (fig. 25). As previously discussed in the Groundwater Budget section of this report, prior to groundwater withdrawals at the quarry, there were more gaining stream reaches in the East Branch Paulins Kill Basin. The steep cone of depression and groundwater divide near Limecrest Quarry, both results of quarry dewatering, are absent in the pre-quarry valley. Groundwater flow in the Pequest River Basin follows present-day flow directions, southwestward towards Lake Iliff.

Water levels in the valley-fill aquifer have been affected by dewatering at Limecrest Quarry, with 139 feet of drawdown at the quarry withdrawal site. Figure 27 shows the magnitude of drawdown in the water-table aquifer. The impact of the pumpage is minimized at the quarry surface-water discharge site where the large volume of released water recharges the underlying aquifer. Losing stream reaches also help to recharge the underlying aquifers, maintaining water levels that may otherwise be lowered by the pumpage. Additionally, the pattern of drawdown in the water-table aquifer is influenced by the glacial stratigraphy; in particular, the presence or absence of low-permeability glacial lake-bottom sediments. For example, near Howells Pond (zone 2, fig. 15), coarse-grained glacial sediments directly overlie high-permeability carbonate rock of the Leithsville Formation and the Allentown Dolomite. The absence of a confining unit here enables pumping stresses in the bedrock to be more readily transmitted to the shallow aquifer, and drawdown, which exceeds five feet, closely matches that in the underlying bedrock. In contrast, drawdown in the shallow valley-fill aquifer in the vicinity of the Rolling Greens Golf Course (zone 1, fig. 15) is less than one foot, although several feet of drawdown occurs in the bedrock aquifer beneath it. The presence of intervening low-permeability lake-bottom deposits in the subsurface retard movement of groundwater from the shallow aquifer to the pumping wells. This same phenomenon was observed during aquifer testing at the site.

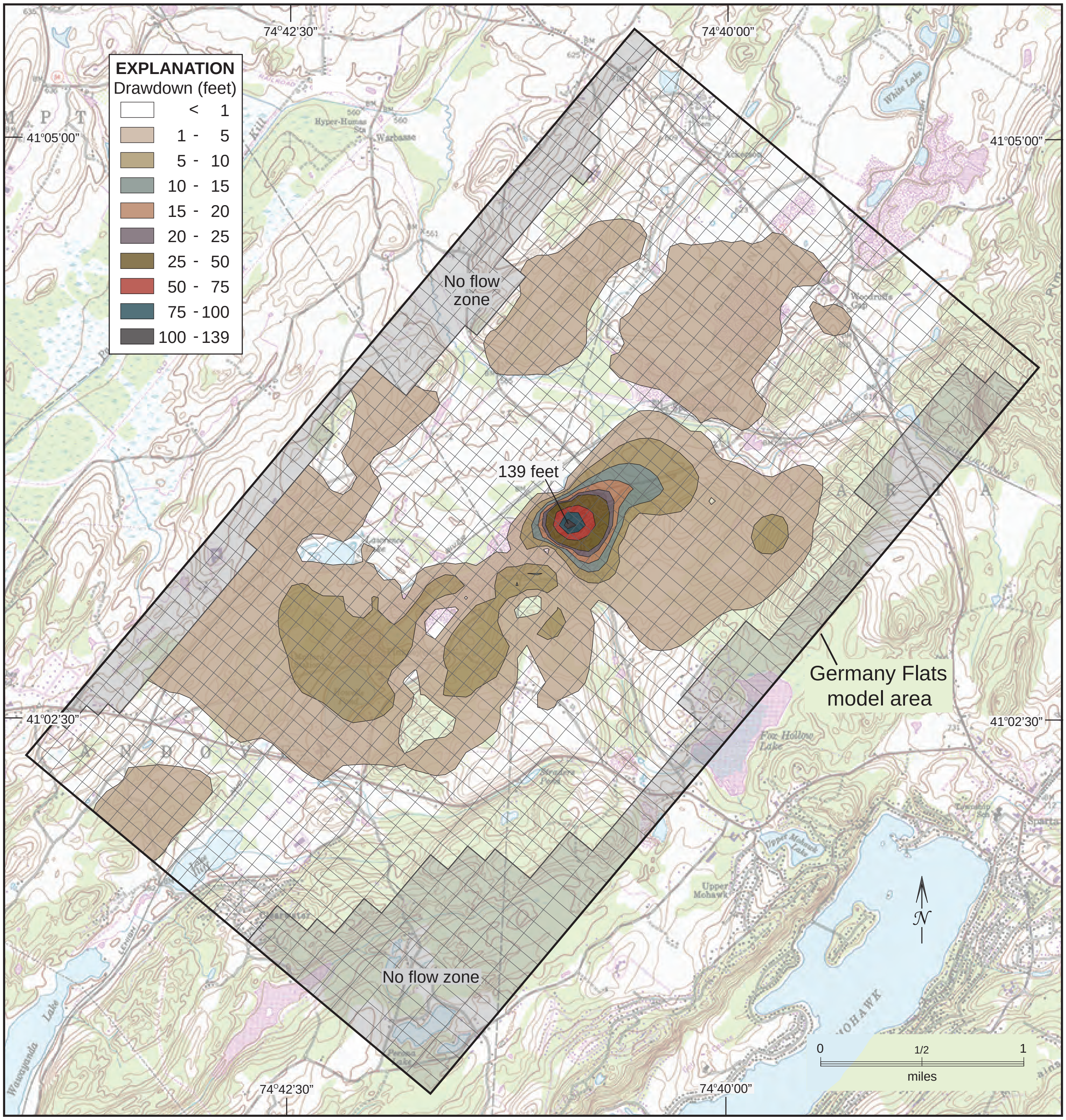


Figure 27. Simulated drawdown in model layer 1 as a result of Limecrest Quarry pumping 6.5 mgd.

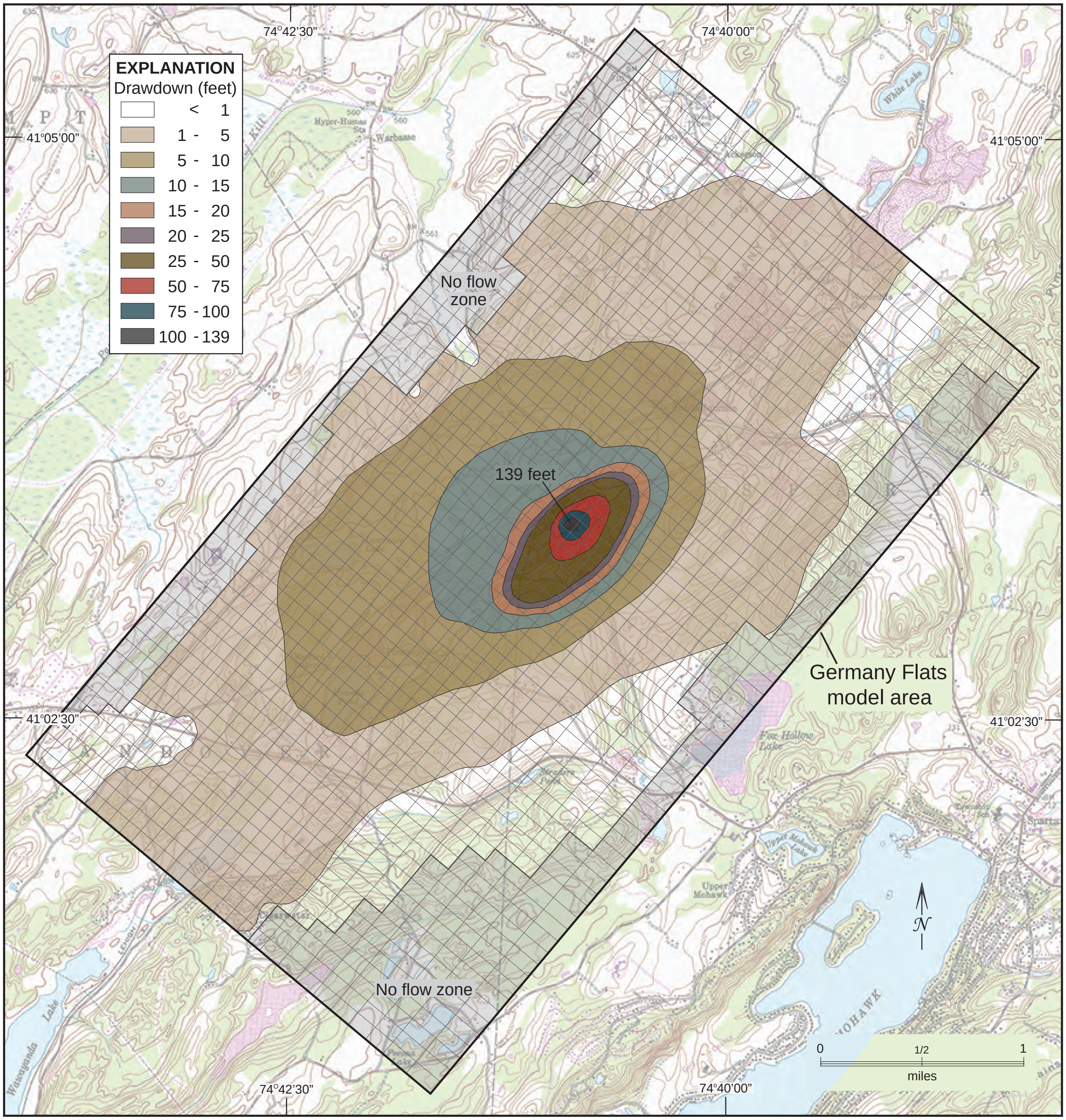


Figure 28. Simulated drawdown in model layer 2 as a result of Limecrest Quarry pumping 6.5 mgd.

The lowering of water levels in the valley-fill aquifer has shifted the location of the groundwater divide near Howells Pond. In the mid-1990's, the divide is approximately 800 feet southwest of the simulated pre-quarry divide (figs. 19 and 25). (The nearly flat hydraulic gradient in the water-table aquifer required use of a contour interval of less than 1 foot to determine the exact location of the simulated divide. The location of the simulated divide also depends on the coarseness of the model grid; a finer grid here may improve delineation of the divide.) Some groundwater that previously flowed southwestward in the Pequest River Basin now flows towards the quarry. The water-budget data confirm that groundwater in the Pequest River Basin has been diverted to the East Branch Paulins Kill Basin.

The impact of quarry dewatering on the groundwater flow pattern is evident in the carbonate rock and composite carbonate rock/valley-fill aquifers. A comparison of figures 20 and 26 indicates that the east-west trending groundwater divide was absent in the pre-quarry valley. The model-simulated piezometric surface indicates that prior to the quarry pumping, the groundwater-flow direction in the East Branch Paulins Kill Basin was predominantly northwesterly or upward towards discharge areas such as streams. As in the water-table aquifer, a southwestward shift in the groundwater divide near Howells Pond may be attributed to Limecrest Quarry dewatering.

Drawdown in the carbonate rock aquifer ranges from 0 to approximately 35 feet, and covers much of the modeled area (fig. 28). The lowering of water levels far from the quarry pumping wells may reflect the high transmissivity of this aquifer, as shallow cones of depression covering large areas characterize high-transmissivity aquifers (Freeze and Cherry, 1979). The overall pattern of drawdown is elliptical, with more drawdown occurring in the direction of bedrock strike, generally northeast-southwest. Transmissivity, as determined during model calibration, is twice as high in this direction. The pumping stress may propagate more readily in more-highly transmissive bedding-plane fractures.

Modeling results indicate that downward flow has increased throughout the valley as a result of the lowered water levels and steeper gradients associated with quarry dewatering. In the simulated pre-quarry scenario, net flow is upward from model layer 2 to model layer 1 at approximately 1.6 x 10^5 ft^3/d (1.2 mgd), representing flow between vertically adjacent aquifers or upward flow within an aquifer. With the quarry dewatering, net downward flow is 5.1 x 10^5 ft^3/d (4.1mgd) from model layer 1 to model layer 2. Flow volumes were determined using ZONEBUDGET (Harbaugh, 1990).

Effects of quarry dewatering on Howells Pond

Howells Pond is a small glacial kettle pond located in the Pequest River Basin near the surface-water divide with the Paulins Kill Basin (pl. 1). The pond is the northernmost in a series of lakes extending southwestward to Gardners Pond. The lakes have been described as kettle lakes, formed by the deposition of glacial sediments around stagnant ice blocks. It has also been theorized that the lakes, which overlie carbonate bedrock, formed over collapsed sinkholes (New Jersey Department of Conservation and Economic Development, 1952). Both glacial and karst processes have probably played a role in the formation of these surface-water bodies. Howells Pond has no inflowing streams. It is fed by groundwater seepage and to a minor extent by stormwater drainage from a nearby road. The pond stage is therefore generally a reflection of the water table.

The water level in Howells Pond in the mid-1990s is lower than that indicated by historic data. Aerial photographs of the pond taken in 1951 and 1961 (Source: Intera/Aero Service, Pittsburgh, PA) show open surface water covering

Figure 29a. Howells Pond, 1951.

Figure 29b. Howells Pond, 1961.

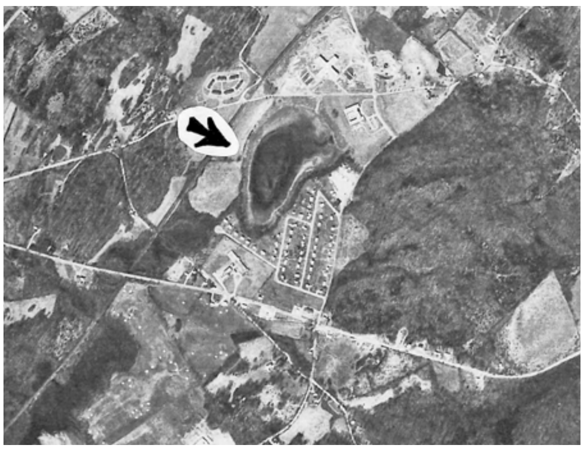


Figure 29c. Howells Pond, 1974.

Figure 29d. Howells Pond, 1995.

approximately 30 acres of land surface (fig. 29 a and b). A U.S. Geological Survey topographic map (revised in 1971) shows Howells Pond at the same extent, with a mapped stream connecting the pond to downstream Lake Iliff. A report by the New Jersey Division of Fish and Game from 1952 notes that the stream exiting Howells Pond is a source of water to Lake Iliff (New Jersey Department of Conservation and Economic Development, 1952). Additionally, in conversations with the author, local residents recalled fishing on Howells Pond as late as the 1960's.

Aerial photographs showing the pond in 1961 and 1974 (Source: Intera/AeroService, Pittsburgh, PA) reveal that its surface area was reduced significantly during this period (fig. 29 b and c). The diminished size and exposed banks are indicative of a drop in the water level in the pond. The pond remains smaller, and therefore shallower, in aerial photographs from 1974 to 1995 (fig. 29 c and d). Field reconnaissance from 1990 to 1995 reveals that the pond occupies a shallow depression with little open surface water, although the pond level varies seasonally. Vegetation, such as cattails and phragmites, cover most of its former area, and its outlet stream is dry.

Several possible causes for the drying-up of the pond were investigated and discounted. The lowered water level is a local phenomenon and therefore would not be attributed to a regional climate change. It also does not appear that the pond has become shallower due to siltation, the accumulation of fine-grained sediments which may impede water flow. The exposed banks and diminished surface area evident in aerial photographs and field observations indicate a drop in the water level rather then a change in the pond bathymetry.

The impact of large groundwater withdrawals was also investigated. Limecrest Quarry is located approximately 1.5 miles northeast of Howells Pond. The water-level decline in the pond, which began sometime between 1961 and 1974, coincides with a large increase in Limecrest Quarry pumpage in 1970. From August 1954 to April 1970, estimated dewatering rates at the quarry ranged from 2.9 mgd to 4.3 mgd, and averaged 3.6 mgd. The May 1970 estimated flow rate, 7.4 mgd, is nearly double the earlier rate.

An assessment of drawdown in Howells Pond due to quarry dewatering was made by comparing simulated water levels in the pond vicinity under pumping and non-pumping conditions. Howells Pond drawdown is simulated using variable-head cells in a manner similar to the surrounding shallow aquifer instead of the constant-head boundary used to designate other surface-water bodies. This allows the water level in Howells Pond to fluctuate in response to pumping stresses during model simulations, so that a determination of drawdown in the pond area can be made. An assumption was made that the pond level would fluctuate in a manner similar to the surrounding shallow aquifer because it is filled principally by groundwater seepage and is essentially a visible expression of the water table.

The simulated head in model cell layer 1, row 46, column 13, located in the center of Howells Pond, is 577.6 feet when the quarry wells are operating at 6.5 mgd, and 583.4 feet when the quarry pumping stress is removed; a difference of 5.8 feet. The modeling therefore indicates that approximately 6 feet of drawdown has occurred in the vicinity of Howells Pond in response to the quarry wells operating at 6.5 mgd. Again, as noted previously, because no attempt was made to specifically model the pond as a surface-water body, 6 feet is simply the model-predicted drawdown in the water-table aquifer in the vicinity of Howells Pond.

As a check on the accuracy of the simulated pre-quarry pond elevation, the elevation was compared with the U.S.G.S. topographic map of the Newton East quadrangle, compiled in 1942 and showing the pond at its full size. On the topographic map, the pond surface is at an elevation greater than 580 feet and less than 600 feet above mean sea level. The simulated elevation of 583.4 feet falls within this range.

Historic water-level data in the area are limited. Only one water-level measurement obtained prior to the large increases from Limecrest Quarry in the early 1970s is available. The reported static elevation of Well A-13 (NJDEP 22-04399) in the shallow glacial aquifer is 586 feet on October 1, 1959 at the time of well installation. The model-simulated pre-quarry water-level elevation is 583.9 feet, or 2 feet lower than the reported value. Although additional measurements would be beneficial in establishing the pre-quarry water-table elevation, the simulated drawdown in Howells Pond appears reasonable based on the available data.

Drawdown in the shallow aquifer is greater in the area of Howells Pond than in some areas adjacent to the quarry due in part to the hydrostratigraphy of the glacial and bedrock units. The pond overlies permeable glacial sand and gravel in good hydraulic connection with a highly transmissive carbonate rock aquifer. These high-permeability units provide a conduit through which pumping stresses at the quarry are transmitted to the shallow aquifer, lowering the water level in the pond. Where low-permeability glacial-lake-bottom deposits are present, drawdown in the shallow aquifer is generally less. The pond is especially vulnerable to the effects of a lowered water table because it receives no surface-water inflow, but is fed by groundwater discharge. When this flow is diverted, as when the quarry wells are pumping, the source of water to the pond decreases and the water level in the pond drops.

SUMMARY AND CONCLUSIONS

The study area is a 12.3 square mile area in Sussex County, New Jersey including parts of Andover, Sparta and Lafayette Townships. The area is a buried valley in which glacial valley-fill deposits overlie a valley floor of Cambrian to Ordovician age carbonate bedrock. The eastern valley wall is Precambrian gneiss, granite and marble and Cambrian quartzite. The western valley wall is slate of the Ordovician Martinsburg Formation. The study area includes the headwaters of the Paulins Kill and Pequest River. Several small lakes and ponds are located in the valley. Limecrest Quarry mines the Franklin Marble along the eastern wall of the valley in the East Branch Paulins Kill River basin. In the mid-1990s, dewatering rates at the quarry were approximately 6.5 mgd.

This study investigates the effects of pumpage at Limecrest Quarry on the valley-fill and bedrock aquifers and surface-water bodies. To do so, the hydrogeologic framework of the study area was developed, hydraulic characteristics of the aquifers were determined through aquifer permeability testing and compiled from previous researchers, and a steady-state numerical groundwater-flow model was developed to simulate groundwater flow in the mid-1990's and pre-quarry valley.

Bedrock aquifers in the study area are 1) an igneous and metamorphic fractured-rock aquifer consisting of Precambrian-age gneiss, granite and marble and the Cambrian Hardyston Quartzite, 2) the Franklin Marble, a meta-sedimentary calcium carbonate, 3) a carbonate rock aquifer consisting of dolomite of the Cambro-Ordovician Kittatinny Supergroup and the Ordovician Jacksonburg Limestone, and 4) slate of the Ordovician Martinsburg Formation. The carbonate rock aquifer and Franklin Marble represent the highest-yielding bedrock aquifers due to their susceptibility to chemical dissolution. The igneous and metamorphic fractured-rock aquifer and the Martinsburg slate represent the least permeable units; groundwater flow is confined to fractures and partings in the rock matrix.

Simulated horizontal hydraulic conductivity (k_x and k_y) values for the igneous and metamorphic fractured-rock aquifer ranged from 1.5 to 3 ft/d; simulated vertical hydraulic conductivity (k_v) was .03 ft/d. Simulated k_x and k_y in the Franklin Marble ranged from 37 to 74 ft/d; k_v was 0.8 ft/d. In the carbonate rock aquifer, simulated k_x ranged from 10 to 150 ft/d, k_y from 20 to 300 ft/d, and k_v from 0.5 to 0.8 ft/d. An anisotropy ratio of 2:1, with higher conductivity in the direction of bedrock strike (northeast-southwest), was assigned to the carbonate rock aquifer. The higher conductivities in this direction reflect enhanced groundwater flow along the intersections of bedding plane partings and sub-parallel fractures. Larger hydraulic conductivities were assigned to the more permeable rocks of the Leithsville Formation and areas where the carbonate rock forms a composite aquifer with permeable glacial sand and gravel. Simulated k_x and k_y of the Martinsburg Formation ranged from 0.5 to 1 ft/d; k_v was .01 ft/d.

Quaternary deposits in the study area are predominantly late-Wisconsinan glacial sediments and recent stream alluvium and swamp deposits. The glacial-deltaic and glacial lacustrine-fan sands and gravels make up the most permeable aquifers, and in some locations are in direct hydraulic connection with the underlying bedrock aquifers. The fine-grained glacial-lake-bottom sediments, in places overlain by recent swamp deposits, are confining units and semi-confining units that impede the vertical flow of groundwater.

Hydraulic conductivity values for the valley-fill deposits varied widely as is characteristic of glaciated terrains. Simulated k_x and k_y values ranged from 10 to 300 ft/d with the highest value representing the composite glacial and carbonate rock aquifer. Simulated k_v ranged from .001 ft/d where a confining unit is present to 10 ft/d in areas of permeable deposits.

Surface-water bodies are generally in good hydraulic connection with the underlying aquifers. Both losing and gaining stream reaches exist in the valley and are simulated in the pre-quarry valley. Streams may change from losing to gaining and vice versa in response to seasonal groundwater fluctuations and precipitation events. Some losing stream reaches are a result of induced streamwater infiltration due to the Limecrest Quarry pumping stress.

Dewatering at Limecrest Quarry has resulted in changes in the water budget for the East Branch Paulins Kill and Pequest River Basins. Under mid-1990's average groundwater-flow conditions, with the quarry wells pumping 1) base flow in the East Branch Paulins Kill is an average of 2.2 ft^3/s greater than in the pre-quarry valley due to the quarry surface-water discharge to this tributary, part of which is derived from water diverted from the Pequest River Basin, 2) base flow in the modeled part of the Pequest River is reduced, and 3) simulated stream leakage to underlying aquifers is greater than in the pre-quarry valley due to induced stream-water infiltration.

Groundwater levels have been lowered and groundwater flow directions in all aquifers have been impacted by Limecrest Quarry dewatering. In the shallow valley-fill aquifer, groundwater flows towards streams and follows surface-water drainage patterns as it did in the pre-quarry valley. Changes in flow directions attributable to the quarry result from a southwestward shift in the groundwater divide between the East Branch Paulins Kill and Pequest River Basins, and the development of a steep-sided cone of depression surrounding the quarry. Just west of Limecrest Quarry, a circulating groundwater flow pattern has developed. Groundwater is pumped from the quarry and discharged to the East Branch Paulins Kill. Some of the water seeps into the underlying glacial and carbonate rock aquifers and flows back towards the quarry to repeat the cycle.

An irregularly-shaped drawdown cone has developed in the shallow valley-fill aquifer due to the quarry dewatering. The pattern of drawdown is controlled, in part, by the glacial stratigraphy, with more drawdown possible where low-permeability glacial confining units are absent. The absence of confining units allows pumping stresses in the bedrock aquifer to be readily transmitted to the water table through the highly-transmissive carbonate rock aquifer and permeable glacial deposits. Drawdown in the valley-fill deposits in the water-table aquifer is less than 10 feet.

In the igneous and metamorphic fractured-rock aquifer and Franklin Marble, groundwater flow patterns in the pre-quarry valley are similar to present-day flow paths with the exception of the immediate area of Limecrest quarry where a steep cone of depression directs flow towards the quarry pumping wells. Groundwater flow generally follows topography and is downslope towards the valley. Locally, flow is towards upland streams. Away from streams, downward vertical gradients exist and groundwater moves downward from the upper to the lower parts of the aquifer. Drawdown ranges from less than 1 foot near the ridge top, to 139 feet at the quarry pumping wells.

In the pre-quarry valley, simulated ground-water flow in the carbonate rock aquifer and composite carbonate-rock/valley-fill aquifer was northwestward towards a gap in the slate bedrock ridge, and southwestward towards Lake Iliff. Under mid-1990's average conditions, almost all flow in the carbonate rock aquifer is towards the quarry. Due to lowered water levels in the mid-1990's valley, the groundwater divide between the East Branch Paulins Kill and Pequest River Basins has shifted southwestward, closer to the model boundary, indicating that the quarry is capturing ground-water from the Pequest River Basin. The long axis of the elliptical drawdown cone is oriented northeast-southwest reflecting the increased hydraulic conductivity along bedrock strike and the permeability contrast between the carbonate aquifer and the less permeable crystalline rocks. Drawdown in the carbonate rock aquifer is less than 25 feet.

Groundwater-flow directions in the Ordovician Martinsburg slate have not changed significantly as a result of the quarry dewatering. As in the crystalline bedrock aquifer, groundwater flows downslope towards the valley, locally towards small streams, and from shallow to deeper parts of the aquifer. Several feet of drawdown have occurred in this unit due to the pumpage withdrawals at the quarry.

Simulated drawdown in the vicinity of Howells Pond, a glacial kettle pond located 1.3 miles southwest of Limecrest Quarry, is approximately 6 feet. The water level in the pond is especially vulnerable to groundwater withdrawals at Limecrest Quarry because it overlies permeable glacial sand and gravel and high-permeability carbonate rocks of the Leithsville Formation and Allentown Dolomite. These permeable aquifers provide a conduit through which the quarry pumping stress is readily transmitted to the pond. Because the pond receives no surface-water inflow and is fed solely by shallow groundwater flow, any drop in the water-table elevation lowers the water level in the pond. Results of this study indicate that the drying-up of Howells Pond is a result of dewatering activities at Limecrest Quarry. The study demonstrates the importance of the glacial hydrostratigraphy in controlling the distribution of drawdown in the water-table aquifer.

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APPENDIX A. Figures

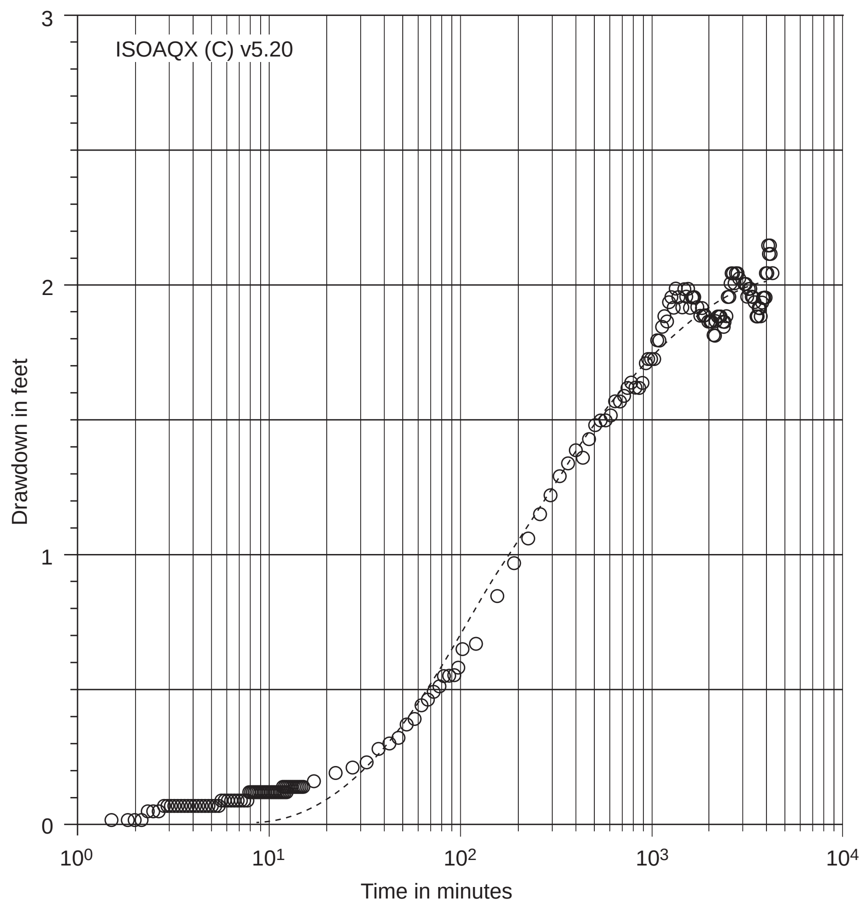

Figure A-1. Time-drawdown data and type curve match for Well A-2 in response to pumping of Well A-3 during a 72-hour aquifer test at the Rolling Greens Golf Course, Andover Township, NJ (July 27-30, 1993). Calculated aquifer properties are: T=5555 ft^2/day, S=1.368E-03, L'=9.907E-04 day-1 using the Hantush-Jacob partial penetration with a RMS error of -7.904E-02 ft. (Small fluctuations in drawdown data are from an unidentified pumping source.)

APPENDIX A. Figures (cont)

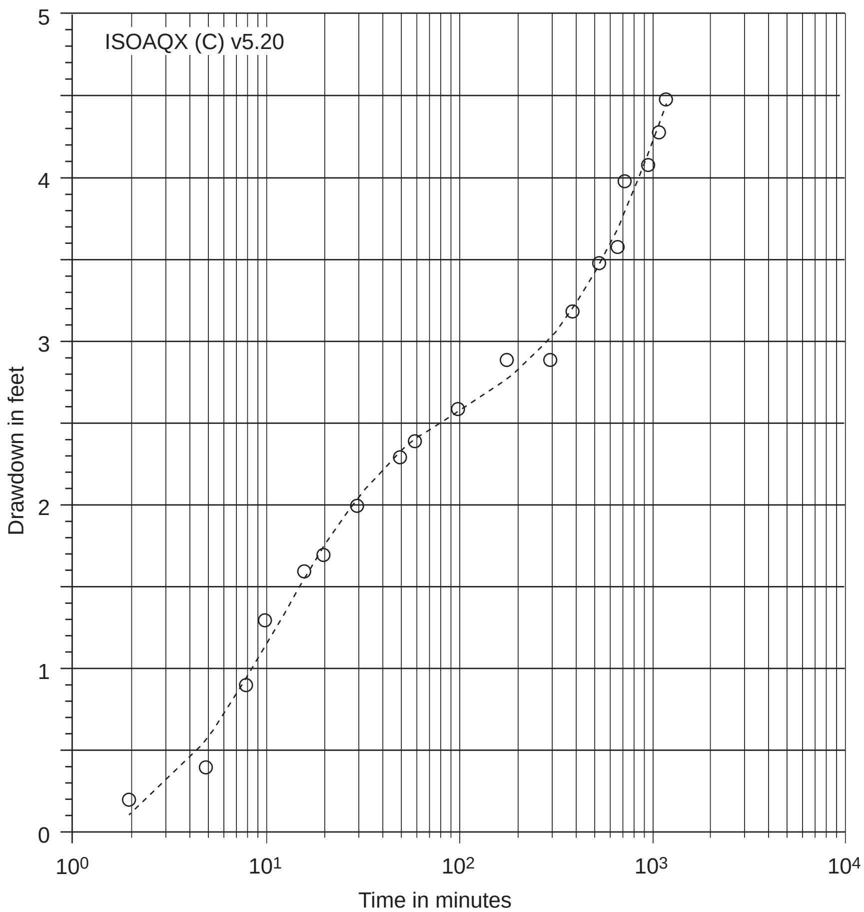

Figure A-2. Time-drawdown data and type curve match for Well S-18 in response to pumping of Well S-17 during a 24-hour aquifer test (March 20, 1991) at the Tanis Sand and Gravel Quarry, Sparta Township, NJ. Calculated aquifer properties are: K=217 ft/day, S=8.739E-04, Sy=2.316E-02, Kz=1.88 ft/day using the Neuman unconfined “A/B” curve model with a RMS error of -9.231E-03 ft.

APPENDIX A. Tables

Table A-1. Well logs used in the study.

Well identifier	NJDEP Well Permit Number	Owner of record	Local identifier	Elevation of land surface or measuring point	Method used to detrmine elevation of land surface or well	Thickness of overburden (ft.)	Bedrock elevation (ft. above mean sea level)	Geologic log (ft) (b - boulder, bl - black, cl - clay, g - gravel, Gn - gneiss, Gr - granite, grn - green, Lms - limestone, occ - occasional, PC - Precambrian, pnk - pink, s - sand, Shl - shale, wh - white)
A-1	22-20818	Lake Iliff Community	domestic well	602.30	surveyed elevation at top of casing	not available	not available	Domestic well; total depth 148 ft.
A-2	22-32508	Kunesch, Ian	Rolling Greens Golf Club OW-2	591.80	surveyed at top of cap	198	392	See detailed log for Geologic Section A-A' in this report
A-3	22-32509	Kunesch, Ian	Rolling Greens Golf Club OW-3	587.90	surveyed at top of casing	210	376	See detailed log for Geologic Section A-A' in this report
A-4	22-32507	Kunesch, Ian	Rolling Greens Golf Club OW-1	590.10	surveyed at recorder deck	>164	<426	See detailed log for Geologic Section A-A' in this report
A-5	22-32510	Kunesch, Ian	Rolling Greens Golf Club OW-4	586.10	surveyed at recorder deck	>40	<543	See detailed log for Geologic Section A-A' in this report
A-6	22-03935	Current, Wilson		600	land surface estimated from USGS topographic quadrangle	>140	<460	0-60 g and s; 60-110 s; 110-130 s; 130-140 g
A-7	22-10810	Janro Realty	John R. Toomey well	608.10	surveyed elevation at top of casing	>65	<543	0-65 s, g
A-8	22-10823	Janro Realty	Hannah Lancaster well	609.80	surveyed elevation at top of casing	50	560	0-50 overburden; 51-98 Lms, soft at 71-72 (10 GPM)
A-9	22-10825	Janro Realty	J.F. Winkelman well	602	land surface estimated from USGS topographic quadrangle	>81	<521	0-70 cl and g; 71-81 g
A-10	22-10862	Janro Realty		602	land surface estimated from USGS topographic quadrangle	>80	<522	0-55 cl and boulders; 56-73 cl and g; 74-80 g
A-11	22-10813	Valliere, Theodore and Barbara		608	land surface estimated from USGS topographic quadrangle	35	573	0-35 overburden; 35-110 Gr; 110-112 soft (1 GPM); 112-198 Gr
A-12	22-10822	Donnely, Michael and Elizabeth		608.60	surveyed elevation at top of casing	38	571	0-38 overburden; 38-122 Gr; 122-123 soft (1GPM); 124-198 Gr
A-13	22-04399	Woehtet, Walter		609.20	surveyed at concrete at top of pit	>74	<538	0-74 s, g
A-15	22-24251	Andover Nursing Center	STP suppliment well	590.10		>113	<470	0-113 s, cl, g; gravel at 113 feet

Well identifier	NJDEP Well Permit Number	Owner of record	Local identifier	Elevation of land surface or measuring point	Method used to detrmine elevation of land surface or well	Thickness of overburden (ft.)	Bedrock elevation (ft. above mean sea level)	Geologic log (ft) (b - boulder, bl - black, cl - clay, g - gravel, Gn - gneiss, Gr - granite, grn - green, Lms - limestone, occ - occasional, PC - Precambrian, pnk - pink, s - sand, Shl - shale, wh - white)
A-16	22-32016	Sussex & Warren Holding Co.	Proposed supply well (TW-2)	589.00	surveyed at top of casing	64	525	0-64 s; 64-74 g (water bearing)
A-17	22-32137	Sussex & Warren Holding Co.	Lifecare MW-1-92	590.10	surveyed elevation at top of casing	>48	<542	0-48 s; 48-58 g
A-18	22-32138	Sussex & Warren Holding Co.	Lifecare MW-2-92	586.70	surveyed at top of cap	>73	<514	0-63 s; 63-73 g
A-19	22-20866	Jeryl, Inc.	Rolling Hills #1	578	land surface estimated from USGS topographic quadrangle	92	503	0-92 s and g overburden; 92-124 Lms
A-20	not available	Farmstead Golf and Country Club	Rock well	602.80	surveyed elevation at top of casing	not available	not available	limestone well (Source: Bob Pheobus, golf course owner)
A-21	not available	Farmstead Golf and Country Club	Overburden well	594.20	surveyed elevation at top of casing	not available	not available	overburden well (Source: Bob Pheobus, golf course owner)
A-22	22-17744	Coolack, Joseph		600	land surface estimated from USGS topographic quadrangle	64	536	0-20 s; 20-50 s and g; 50-64 fine silt; 64-73 brown Lms (water area)
A-24	22-20321	McDaly		676.20	surveyed elevation at top of casing	0	676	0-198 Lms
A-25	22-31703	Sussex & Warren Holding Co.	Lifecare TW-1	587.70	surveyed elevation at top of casing	102	486	0-102 s and g overburden; 102-400 Lms bedrock
A-30	22-26454	Pheobus, Bob		671.10	surveyed elevation at top of casing	12	659	0-12 overburden; 12-659 Shl
L-1	22-33686	Sussex & Warren Holding Co.	nearlime deep	603.96	surveyed elevation at top of casing	85	519	0-5 brown medium to coarse s, some g; 5-20 gray; 20-35 gray slt; 35-80 gray slt and cl; 80-85 g
L-2	22-33635	Sussex & Warren Holding Co.	nearlime shallow	605.81	surveyed elevation at top of casing	>54	<552	0-14 tan s; 14-20 brown fine s and slt; 20-28 brown fine s, g fragment; 28-38 brown fine s and slt; 38-44 gray-brown fine sand and slt; 44-54 gray slt
L-4	22-19313	Raymond International Builders	Dahn well	607.10	surveyed at top of cap	not available	not available	industrial well; total depth 516 ft.
L-5	not available	Seventh Day Adventist	domestic well	606.90	surveyed at top of cap	not available	not available	not available

Well identifier	NJDEP Well Permit Number	Owner of record	Local identifier	Elevation of land surface or measuring point	Method used to detrmine elevation of land surface or well	Thickness of overburden (ft.)	Bedrock elevation (ft. above mean sea level)	**Geologic log (ft)** (b - boulder, bl - black, cl - clay, g - gravel, Gn - gneiss, Gr - granite, grn - green, Lms - limestone, occ - occasional, PC - Precambrian, pnk - pink, s - sand, Shl - shale, wh - white)
S-1	22-31479	Limestone Products Corp.	MW-4s	590	land surface estimated from USGS topographic quadrangle	not available	not available	monitoring well; total depth 45 ft.
S-2	22-30516	Limestone Products Corp.	MW-3s	590	land surface estimated from USGS topographic quadrangle	>32	<558	0-32 brown fine to medium s, some g
S-3	22-31480	Limestone Products Corp.	MW-5s	590	land surface estimated from USGS topographic quadrangle	not available	not available	monitoring well; total depth 33 ft.
S-4	22-30515	Limestone Products Corp.	MW-s2	590	land surface estimated from USGS topographic quadrangle	>31	<559	0-31 brown fine to medium s, some g
S-5	22-30514	Limestone Products Corp.	MW-sl	590	land surface estimated from USGS topographic quadrangle	>35	<555	0-35 brown fine to medium s, some g
S-8	22-27504	Nicholson, Robert B.	Eastern Propane	573.70	surveyed elevation at top of casing	>155	<419	0-132 s, cl, g; 132-155 g
S-9	22-30838	Sparta Twp.	MW-4s	572.98	surveyed elevation at top of casing	>14	<557	0-10 disturbed area - soil, s, refuse; 10-11.5 gray clayey-silt, trace fiber, trace fine s; 11.5-14 s and g
S-10	22-30823	Sparta Twp.	MW-7	573.53	surveyed elevation at top of casing	>88	<483	0-9 coversoils, refuse, fill; 9-15 s and g; 15-88 glaciolacustrine silts and cl
S-11	22-30824	Sparta Twp.	MW-5	580.23	surveyed elevation at top of casing	>16	<562	0-11 coversoils, refuse, fill; 11-14 clayey silt (gray, green, black) with occ. peat; 14-16 s and g
S-12	22-17883	Geldback Refridge Co.		592.60	surveyed at top of cap	24	569	0-24 s and g; 24-280 Lms
S-13	22-27834	Sparta Twp.	Well #2 Commerce Park	617.80	surveyed at top of casing	50	568	0-50 s, cl, g; 50-350 Lms rock (dolomite)
S-14	22-19406	Challenge Industries, Inc.		618.10	surveyed at top of casing	84	534	0-84 cl and g; 84-225 blue lime rock
S-15	22-25441	Spetz, Anna	Unicar Autobody	602.60	surveyed at top of casing	0	603	0-200 Lms
S-16	22-28915	State of New Jersey	soccer well	622.80	surveyed elevation at top of casing	141	482	light to dark grey dolomite

Well identifier	NJDEP Well Permit Number	Owner of record	Local identifier	Elevation of land surface or measuring point	Method used to detrmine elevation of land surface or well	Thickness of overburden (ft.)	Bedrock elevation (ft. above mean sea level)	**Geologic log (ft)** (b - boulder, bl - black, cl - clay, g - gravel, Gn - gneiss, Gr - granite, grn - green, Lms - limestone, occ - occasional, PC - Precambrian, pnk - pink, s - sand, Shl - shale, wh - white)
S-17	22-20370	Tannis Sand and Gravel Co.		590.85	surveyed at marked measuring point on well	>132	<459	0-20 s and g; 20-119 fine s and cl; 119-132 s and g; 128-132 s and g
S-18	22-30665	Sparta Twp.	TW-4	599.17	surveyed at top of hinge	143	456	0-10 no samples; 10-12 s and g; 12-20 no log; 20-123 s, light grey or brown with some silt, cl; 123-143 light brown poorly sorted s and g with clayey silt (till?); 143-157 weathered dolomite
S-21	22-22983	L & S Precision Co.		616.10	surveyed elevation at top of casing	51	565	0-51 cl and g; 51-175 Gr
S-24	not available	Hettrick, William (resident)	Deire Drive	692.10	surveyed elevation at top of casing	not available	not available	Rock well
S-25	22-27602	Carriage Hill Associates	3 Farmbrook Rd. well	641.50	surveyed elevation at top of casing	10	632	0-10 overburden; 10-200 dolomite
S-26	not available	Brookside Homes/ Carol Rickson	20 Farmbrook Road	730.30	surveyed elevation at top of casing	not available	not available	Rock well
S-30	22-08556	Howe, Raymond A.W	ell #1	790	land surface estimated from USGS topographic quadrangle	10	780	0-10 hardpan and boulders; 10-15 “broken up” lime rock; 25-78 white lime rock; 78-113 soft brown Gr

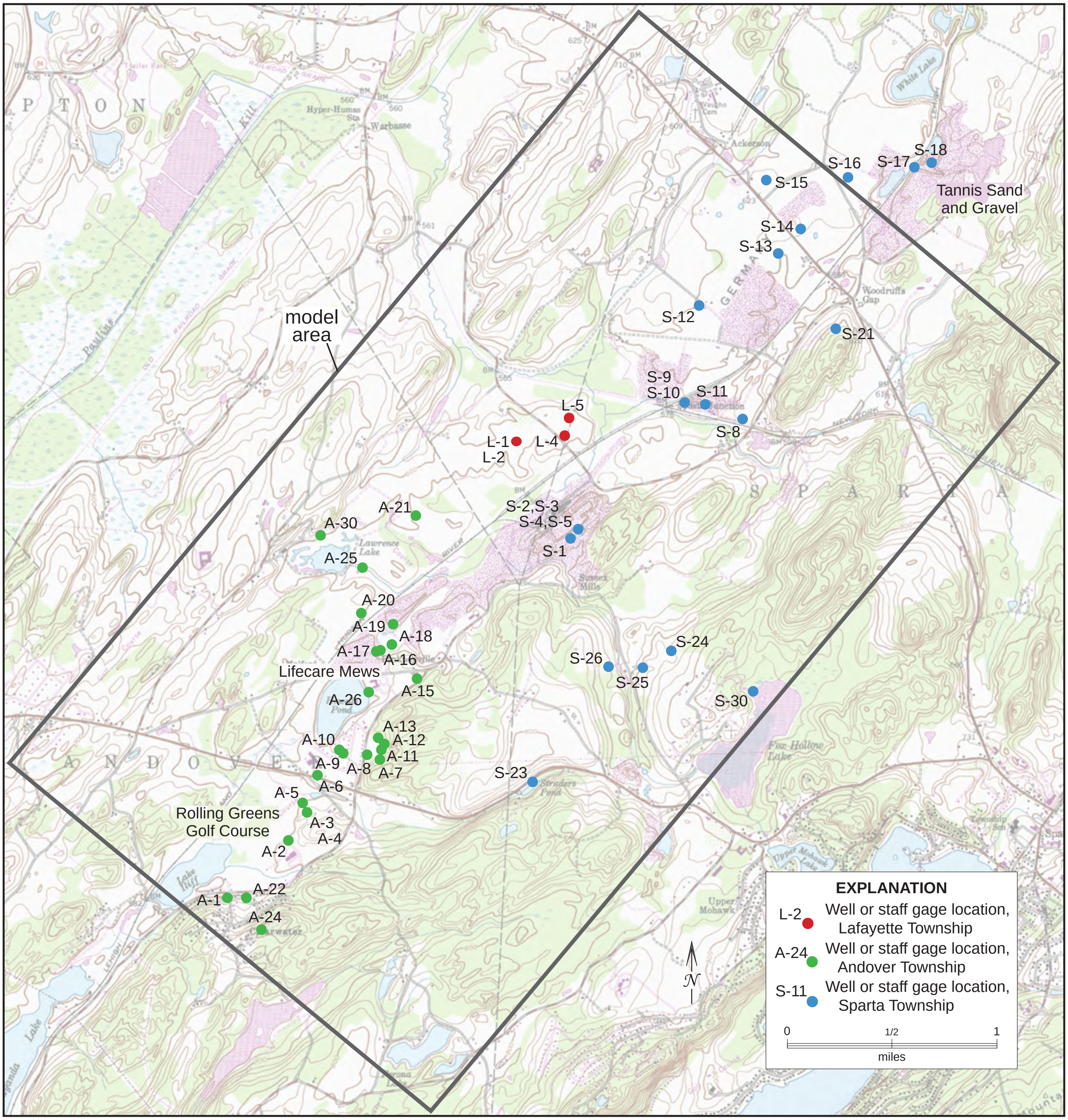

Plate 1. Well locations in Germany Flats model area. Well prefix indicates township location (A = Andover Township, L = Lafayette Township, S = Sparta Township).

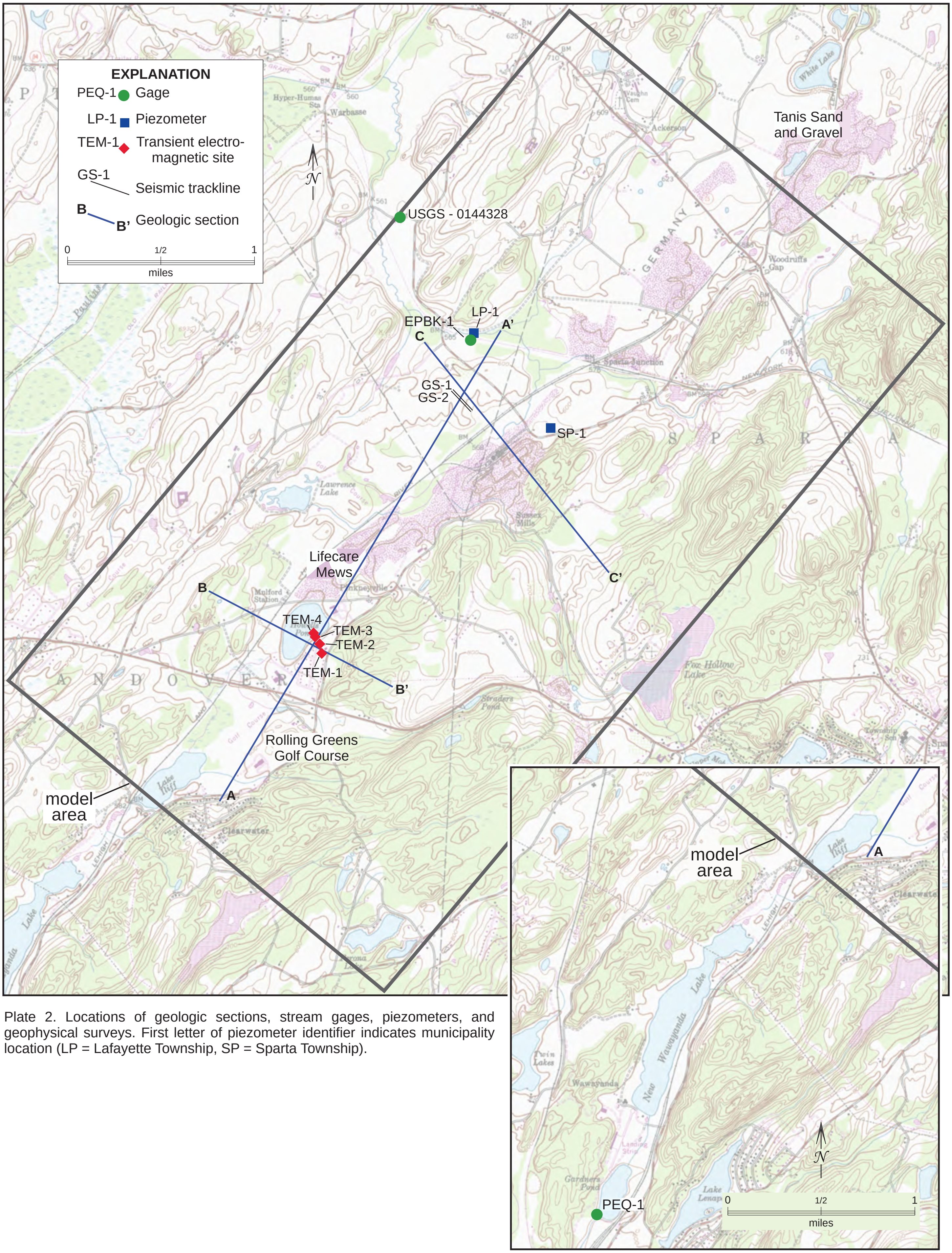

Plate 2. Locations of geologic sections, stream gages, piezometers, and geophysical surveys. First letter of piezometer identifier indicates municipality location (LP = Lafayette Township, SP = Sparta Township).