GEOHYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE CARBONATE ROCKS OF THE VALLEY CREEK BASIN, EASTERN CHESTER COUNTY, PENNSYLVANIA

By Ronald A. Sloto

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CONVERSION FACTORS AND ABBREVIATIONS

For the convenience of readers who prefer to use metric (International System) units rather than the inch-pound units in this report, the following conversion factors may be used:

Multiply inch-pound unit	By	To obtain metric unit
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.59	square kilometer (km^2)
gallon (gal)	3.785	liter (L)
million gallons (Mgal)	3,785	cubic meter (m^3)
gallon per minute (gal/min)	0.06308	liter per second (L/s)
gallon per minute per foot [(gal/min)/ft]	0.207	liter per second per meter [(L/s)/m]
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
inch per square mile (in/mi^2)	9.807	millimeter per square kilometer (mm/km^2)

GEOHYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE CARBONATE ROCKS OF THE VALLEY CREEK BASIN, EASTERN CHESTER COUNTY, PENNSYLVANIA

By Ronald A. Sloto

ABSTRACT

Sixty-eight percent of the 22.6-square-mile Valley Creek basin is underlain by Cambrian and Ordovician limestone and dolomite. Ground water flows through a network of interconnected secondary openings; primary porosity is virtually nonexistent. Some of these openings have been enlarged by solution. Secondary porosity and permeability exhibit great spatial variability, and the yield and specific capacity of wells are highly variable. The number of water-bearing zones decreases with depth. Fifty percent of water-bearing zones are encountered within 100 feet of the land surface, and 81 percent are within 200 feet.

Most ground-water flow in the Valley Creek basin is local and discharges to nearby streams. Ground-water discharge comprised an average of 76 percent of the flow of Valley Creek during 1983-87, including both natural ground-water discharge and quarry pumpage discharged to Valley Creek. Discharge from the Cedar Hollow quarry comprised 21 to 26 percent of the base flow of Valley Creek; the average was 23 percent. The average natural base flow of Valley Creek would be 8 percent lower if the quarry were not operating.

Regional ground-water flow is to the northeast to the Schuylkill River. On the western side of the Valley Creek basin, the ground-water divide is 1/2 mile west of the surface-water divide. An estimated 0.75 million gallons per day of ground water flows from the adjacent West Valley Creek basin eastward into the Valley Creek basin. A ground-water divide is not present on the eastern side of the basin; the water table slopes gently eastward toward the Schuylkill River. On the northeastern side, an estimated 1.76 million gallons per day of ground water flows northeastward out of the basin to the Schuylkill River beneath the surface-water divide. On the southeastern side, an estimated 0.85 million gallons per day of ground water flows beneath the surface-water divide into the basin.

Annual water budgets and an average water budget were calculated for 1983-87 for the 20.8-square-mile area above the streamflow-gaging station. Annual precipitation for 1983-87 ranged from 40.61 to 56.55 inches and averaged 47.25 inches; annual streamflow ranged from 15.55 to 28.57 inches and averaged 22.31 inches; annual evapotranspiration ranged from 18.21 to 24.83 inches and averaged 22.90 inches; and annual recharge ranged from 15.89 to 26.84 inches and averaged 21.04 inches.

The Valley Creek basin was modeled as a two-dimensional water-table aquifer. Recharge to, ground-water flow through, and discharge from the rocks of Chester Valley were simulated. In order to include the natural hydrologic boundaries of the ground-water-flow system, the 66.4-square-mile area between the Brandywine Creek and the Schuylkill River was modeled. The model was calibrated under steady-state conditions using average recharge and evapotranspiration rates. Aquifer hydraulic conductivity was estimated from specific-capacity and aquifer-test data. The average (1983-87) annual water budget for the Valley Creek basin was simulated.

The effect of increased ground-water development on base flow and underflow was simulated by locating a hypothetical well field producing 4 million gallons per day in different parts of the basin. Pumpage from a well field near surface-water divides would induce as much as an additional 1.41 inches per year of underflow from an adjacent surface-water basin. Pumpage from a well field near the center of the basin would affect base flow more than underflow.

Increased seepage of ground water into quarries as a result of their expansion was simulated as increased withdrawal by pumping. A 100-percent increase in the pumping rate of the Cedar Hollow quarry, from 3.93 to 7.86 million gallons per day, would reduce the natural base flow of Valley Creek by 18 percent. However, the quarry pumpage would be discharged to Valley Creek, thereby increasing the base flow at the gaging station by about 10 percent.

INTRODUCTION

The Valley Creek basin is underlain by some of the most productive aquifers in Chester County, Pennsylvania. The basin, a major source of water for public water supplies, is undergoing rapid residential and commercial development. The demand for water is increasing, but chemical contamination has reduced the availability of potable ground water. A study to assess the effects of urbanization (Sloto, 1987) pointed out the vulnerability of this area to ground-water contamination.

A thorough understanding of the ground-water-flow system is necessary for prudent management of the ground-water resource. Ground-water planning and management in Chester County uses a surface-water-basin water-budget approach (Reith and others, 1979; Chester County Planning Commission, 1982). The ground-water-flow system in carbonate rock differs from the flow system in the crystalline rocks that underlie most of Chester County. Changes in the water budget of a basin caused by ground-water development in carbonate rock are difficult to estimate without a method that takes into account the entire hydrologic system. A digital model of the flow system is such a method. This study, done by the U.S. Geological Survey in cooperation with the Chester County Water Resources Authority, uses a digital model of regional ground-water flow to estimate the effects of ground-water development on the water budget of the Valley Creek basin.

Purpose and Scope

This report describes ground-water flow in the carbonate rocks of the Valley Creek basin. A digital model of regional ground-water flow was developed and used to simulate the average water budget in the basin and to estimate the effects of increased well and quarry pumping on base flow and underflow. Although the report primarily discusses ground-water flow in the carbonate rocks of the Valley Creek basin, it was necessary to consider noncarbonate rocks north and south of the carbonate rocks as well as the eastern and western boundaries of the hydrologic system to simulate ground-water flow. Therefore, the carbonate rocks between the Schuylkill River and the Brandywine Creek in Chester and Montgomery Counties are included in the modeled area. Data for noncarbonate rocks in these areas also are given in this report.

The report also presents a water budget for the Valley Creek basin upstream from the stream-gaging station at the Pennsylvania Turnpike bridge near Valley Forge (station 01473169), describes the development and calibration of a two-dimensional digital model of ground-water flow, and briefly discusses ground-water quality.

Location and Physiography

The Valley Creek basin is in eastern Chester County in southeastern Pennsylvania (fig. 1). Valley Creek drains 22.6 mi^2 (square miles) and is a tributary to the Schuylkill River. Streamflow from 20.8 mi^2 is measured at streamflow-gaging station 01473169, Valley Creek at Pennsylvania Turnpike Bridge near Valley Forge (fig. 2). Little Valley Creek, the major tributary to Valley Creek, drains 6.92 mi^2 and is confluent with Valley Creek above the gaging station.

The Valley Creek basin is in the Piedmont physiographic province. Sixty-eight percent of the basin is underlain by carbonate rocks; 32 percent of the basin is underlain by noncarbonate rocks. The center of the basin is underlain mostly by easily eroded limestone and dolomite, which form Chester Valley (fig. 2). Chester Valley cuts across the center of Chester County. The northern part of the Valley Creek basin is underlain by resistant quartzites that form the North Valley Hills. The southern part of the Valley Creek basin is underlain by resistant phyllite that forms the South Valley Hills.

Chester County has a modified humid continental climate characterized by warm summers and moderately cold winters. The normal annual temperature (1951-80) recorded at Phoenixville, 3 mi (miles) north of the Valley Creek basin, is 51.3 °F (degrees Fahrenheit). The normal temperature for January, the coldest month, is 30.1 °F. The normal temperature for July, the warmest month, is 74.5 °F. The average annual precipitation at Phoenixville for 81 years of record (1890-95, 1913-87) is 43.85 in. (inches). The minimum annual precipitation, 31.10 in., occurred in 1963. The maximum annual precipitation, 59.55 in., occurred in 1979. The 1951-80 normal precipitation is 43.55 in. (National Oceanic and Atmospheric Administration, 1982). Precipitation is mostly evenly distributed throughout the year, but slightly more falls in July and August than in the other months.

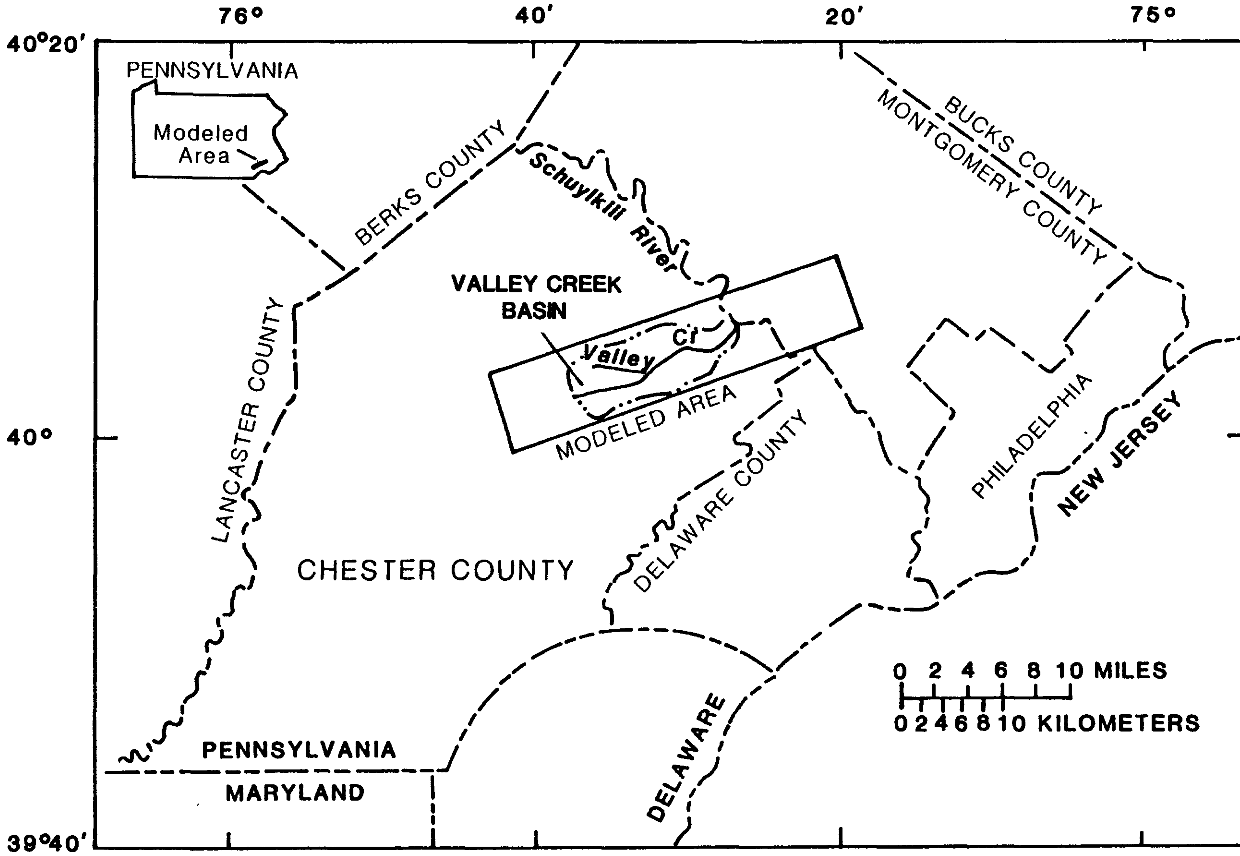


Figure 1.--Location of Valley Creek basin.

Well-Numbering System

The well-numbering system used in this report consists of a county-abbreviation prefix followed by a sequentially-assigned number. The prefix CH denotes a well in Chester County; MG denotes a well in Montgomery County. Data for wells in Chester County used for analysis are given in Sloto (1989). Data for wells in Montgomery County are given in Newport (1971).

Previous Investigations

The geology of the Valley Creek basin was mapped and described by Bascom and others (1909) and Bascom and Stose (1938). Geologic-quadrangle maps for the study area were published by Berg and Dodge (1981). Part of the basin is included in the geologic map of Lyttle and Epstein (1987).

The hydrology of the igneous and metamorphic rocks of central Chester County was described by Poth (1968). McGreevy and Sloto (1977) described the ground-water resources of Chester County. Sloto (1987) described the effect of urbanization on the water resources of eastern Chester County. Sloto (1989) presented ground-water data for Chester County. The ground-water resources of Montgomery County were described by Newport (1971).

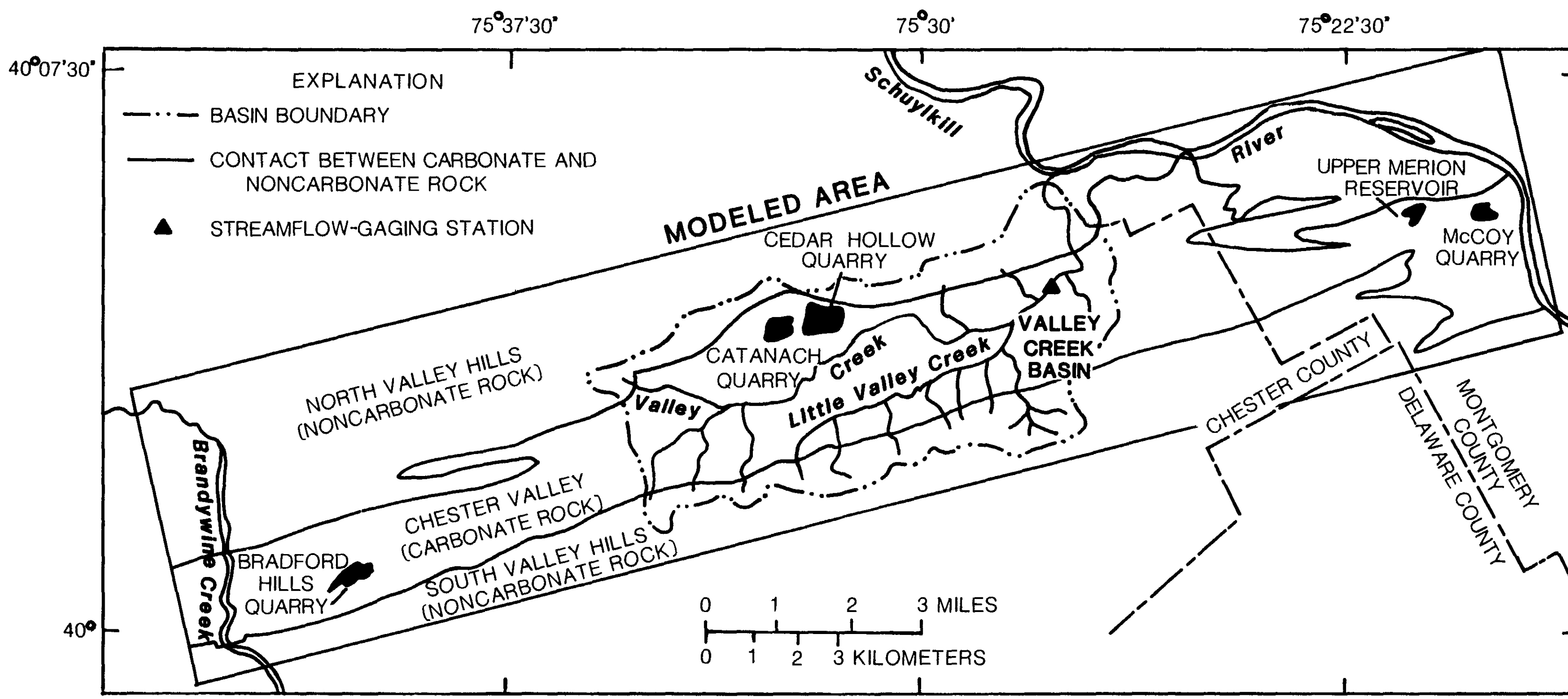


Figure 2.--Location of modeled area.

Acknowledgments

The cooperation of well owners for access to wells for water samples, water-level measurements, and geophysical logging is gratefully acknowledged. The author especially thanks Glasgow, Incorporated; Leggette, Brashears, and Graham, Incorporated; the Pennsylvania Department of Environmental Resources; Philadelphia Suburban Water Company; Thomas G. Keyes, Incorporated; and the Warner Company for providing data and access to their wells and property. Special thanks is given to the Philadelphia Suburban Water Company for providing aquifer-test data for their wells and to Sidney Fox of Leggette, Brashears, and Graham; Preston Leutweiller of the Philadelphia Suburban Water Company; and Joseph Pinto of the Warner Company.

GEOHYDROLOGY

Stratigraphy

The carbonate rocks of Chester Valley form the center of the Valley Creek basin. Noncarbonate rocks underlie the North Valley Hills to the north and the South Valley Hills to the south (pl. 1). The stratigraphic relationships of these units are given in table 1. Geologic descriptions were taken from Bascom and Stose (1938) and Lyttle and Epstein (1987). The nomenclature used in this report is that of Lyttle and Epstein (1987). The nomenclature of Berg and others (1986) given in table 1 is used by the Pennsylvania Geological Survey and has been used for previous reports on the ground-water resources of Chester County.

Carbonate Rocks

A sequence of carbonate rocks of Cambrian and Ordovician age underlies Chester Valley. The principal formations are the Cambrian Ledger Dolomite and Elbrook Formation and Cambrian and Ordovician Conestoga Limestone. The Cambrian Vintage Dolomite and Kinzers Formation, which crop out in a narrow band, are thin and are not important water-bearing units.

Vintage Dolomite

The Vintage Dolomite is of small areal extent in eastern Chester Valley. In the Valley Creek basin, it is present in a fault block north of Mill Lane and in a very narrow band between two parallel faults in the northeastern part of the basin. The lower part of the Vintage Dolomite is a fine-grained, thin- to medium-bedded, argillaceous to sandy dolomite with abundant mica on the bedding planes. The upper part is a fine- to medium-grained, mottled, blue limestone, grading downward into medium-grained, knotty dolomite with blebs of coarse-grained dolomite, grading downward into medium-grained, thick-bedded dolomite. It is less than 200 ft (feet) thick.

Kinzers Formation

The Kinzers Formation crops out adjacent to the Vintage Dolomite in eastern Chester Valley. The upper part is a fine- to medium-grained, irregularly bedded, argillaceous, nodular limestone containing marble lenses.

The lower part is a thin-bedded, impure limestone. The weathered limestone has the appearance of a shaley mica schist. The Kinzers Formation is less than 30 ft thick.

Ledger Dolomite

The Ledger Dolomite has been quarried in many places in Chester Valley. It is white to gray, massive to thick bedded, finely laminated, and has a high magnesium content. The dolomite is interbedded with some siliceous beds and laminated limestone, which is finely speckled in places. The lower part of the unit is characterized by alternating light and dark, porous, cherty layers. The lower contact is gradational with the Kinzers Formation. The Ledger is about 1,000 ft thick.

Table 1.--Stratigraphic section

SYSTEM AND ERA	SERIES	GEOLOGIC UNIT: Lyttle and Epstein (1987)		GEOLOGIC UNIT: Berg and others (1986)	
Triassic	Upper Triassic	Stockton Formation		Stockton Formation	
Ordovician	Lower Ordovician	Conestoga Limestone		Conestoga Formation	
Cambrian	Upper Cambrian	Elbrook Formation	Octoraro Phyllite		
	Middle Cambrian			Elbrook Formation	
		Ledger Dolomite		Ledger Formation	
	Lower Cambrian	Kinzers Formation		Kinzers Formation	Wissahickon ("Octoraro") Schist
		Vintage Dolomite		Vintage Formation	
		Antietam Quartzite and Harpers Phyllite, undivided		Antietam Formation and Harpers Formation, undivided	
		Chickies Quartzite			
Late and Middle Proterozoic		Leucocratic and Intermediate Felsic Gneiss		Chickies Formation	
				Gneiss	

Elbrook Formation

The Elbrook Formation forms low hills in Chester Valley and underlies the ridge south of the Catanach and Cedar Hollow quarries. The Elbrook is a light blue, thin-bedded, fine-grained limestone interbedded with white, thick-bedded, fine-grained, laminated dolomitic marble. Concentrations of coarse-grained mica are left as a pressure-solution residue parallel to regional cleavage. The lower contact is gradational with the Ledger Dolomite. The Elbrook is about 800 ft thick.

Conestoga Limestone

The Conestoga Limestone crops out along the southern edge of Chester Valley. Mica coats most of the bedding and cleavage planes. The lower part of the formation is a coarsely crystalline, light gray to white, medium- to thick-bedded dolomite interbedded with thin-bedded, medium-grained limestone and thin-bedded, fine-grained dolomite. Some of the basal beds are a coarse limestone conglomerate containing large pebbles and irregular masses of coarse white marble in a dark, argillaceous matrix. The middle part of the formation is a thin, dark, graphitic phyllite with thin, sandy schist layers that thicken eastward. It generally forms a line of discontinuous hills. The upper part is a bluish gray, thin-bedded, fine- to medium-grained, highly micaceous limestone with argillaceous, shaley partings that give it a finely laminated appearance. It unconformably overlies the Elbrook Formation. The Conestoga Limestone is 500 to 800 ft thick.

Noncarbonate Rocks

The area underlain by crystalline rock is divided by Chester Valley. North of Chester Valley, the North Valley Hills are underlain by the Cambrian Chickies Quartzite and the Antietam Quartzite and Harpers Phyllite, undivided, and Middle Proterozoic leucocratic and intermediate felsic gneiss. South of Chester Valley, the South Valley Hills are underlain by the Cambrian to Late Proterozoic Octoraro Phyllite.

Leucocratic and intermediate felsic gneiss

The Middle Proterozoic leucocratic and intermediate felsic gneiss is of small areal extent in the Valley Creek basin. It is a fine- to medium-grained, white to gray, microcline-microperthite-quartz gneiss intimately associated with biotite-oligoclase-microperthite-quartz gneiss and is interlayered with amphibolite.

Octoraro Phyllite

The South Valley Hills are underlain by the Octoraro Phyllite (which was called the albite-chlorite facies of the Wissahickon Formation by Bascom and Stose, 1938). The Octoraro Phyllite is a greenish- to silvery-gray, fine- to medium-grained phyllite and phyllonite. Grain size decreases to the north. Along the northern edge of its outcrop area, the Octoraro Phyllite is bounded by a major thrust fault and is thrust over and onto the Conestoga Limestone.

Chickies Quartzite

The Chickies Quartzite is a medium-grained, cross-bedded, massive to medium-bedded, finely-laminated quartzite and sericitic quartz schist. The basal Hellam Member, which is not mapped as a separate unit, is a coarse-grained, tourmaline-bearing quartzite and arkosic pebble conglomerate with interbeds of black slate and biotite schist. The Chickies Quartzite is about 500 ft thick.

Antietam Quartzite and Harpers Phyllite

The Antietam Quartzite and Harpers Phyllite are undivided north of Chester Valley. The Antietam Quartzite is a fine-grained, laminated quartzite that grades downward into the Harpers Phyllite. The upper beds are coarse grained and calcareous. The Antietam Quartzite is estimated to be less than 200 ft thick north of Chester Valley. The Harpers Phyllite is fine- to medium-grained, sandy, and argillaceous. It is less than 500 ft thick.

Stockton Formation

The Triassic Stockton Formation is part of the sedimentary Newark basin. The Stockton is a thick sequence of interbedded sandstone, arkosic sandstone, arkose, arkosic conglomerate, siltstone, and shale. The beds dip about 10 to 20 degrees to the northeast. The Stockton unconformably overlies older Paleozoic and Precambrian rocks. In the modeled area, the Stockton Formation is as much as 3,000 ft thick.

Hydrology

Ground water flows through a network of fractures slightly enlarged by solution. The ground-water system in Chester Valley more nearly resembles that of fractured rock than that of a classical karst terrane. A karst terrane generally is characterized by sinkholes, dry valleys, and underground drainage; regional ground-water flow is through an arterial network leading to a large main conduit that discharges through large springs. Most flow in the Valley Creek basin is local with discharge to nearby streams. Regional flow in eastern Chester Valley is to the Schuylkill River.

Dissolution is the primary weathering process of carbonate rock. Dissolution generally is most active above and within the zone of water-table fluctuation where water movement is relatively rapid and recharge water is acidic. Below the zone of water-table fluctuation, water movement is comparatively slower, and acidic recharge water becomes neutralized. Near the land surface, dissolution of carbonate rock results in the filling of voids by clay, the collapse of solution openings, and the progressive lowering of the land surface. Clay and unconsolidated material sometimes moves downward through solution openings, plugging water-bearing openings. This plugging can decrease well yields and increase turbidity of ground-water discharged from wells.

The depth of weathering is highly variable. Deeply weathered zones can be found adjacent to outcrops. Carbonate rock commonly exhibits pinnacle weathering (fig. 3). Pinnacle weathering is caused by solution along bedding planes and fractures parallel to bedding planes in steeply-dipping strata. As solution enlarges openings along the bedding planes and fractures, and enlargement moves downward in the formation, the solid rock between the weathered areas is left as pinnacles.

Figure 3.--Pinnacles of the Elbrook Formation exposed in the Cedar Hollow quarry. View looking east.

Water-Bearing Zones

Primary porosity in the carbonate rocks of Chester Valley is virtually nonexistent. Ground water flows through a network of interconnected secondary openings--fractures, joints, faults, parting planes, and bedding planes. Some of these openings have been enlarged by solution. The number and size of the openings determines the secondary porosity of the rock; the degree of interconnection of the openings determines the secondary permeability. The high permeability of carbonate rock is predominantly the result of enlargement of secondary openings by solution. Where solution has been active, permeability may be high; elsewhere, the same unit may be nearly impermeable.

Most openings enlarged by solution are only a fraction of an inch wide, but they are capable of transmitting large quantities of water. A close inspection of five active quarries in Chester Valley during this study revealed very few large solution openings. Most of the solution openings were horizontal enlargements of vertical fractures and were less than 1 ft (foot) wide (fig. 4). Most of the ground water entering the quarries flowed from very narrow fractures. Ground water was observed discharging from only one large solution opening about 15 ft above the third level of the Cedar Hollow quarry (fig. 5); the opening was less than 1 ft across.

Figure 4.--Solution openings along a vertical fracture in the Cedar Hollow quarry. Scale in center of photograph is one foot long.

Figure 5.--Ground water discharging from a solution opening in the Cedar Hollow quarry. Scale on left side of photograph is two foot long.

Well CH-89, drilled south of the Cedar Hollow quarry in the Elbrook Formation, penetrated a 58-ft void from 207 to 265 ft below land surface. From 265 to 295 ft, the opening was filled with mud and rubble. Drilling was halted at 295 ft without penetrating solid rock. The void was probably an enlarged fracture similar to the one in figure 5. The well was drilled near a sinkhole area. The mud and rubble at the bottom of the opening is probably weathered surficial material that moved downward through sinkholes.

The number of water-bearing zones decreases with depth (fig. 6). The distribution of 235 water-bearing zones in 119 wells in the Vintage Dolomite, Kinzers Formation, Ledger Dolomite, Elbrook Formation, and Conestoga Limestone in Chester County was analyzed. These 119 wells represent 16,403 ft of uncased borehole; well depths are as great as 605 ft. Fifty percent of the water-bearing zones are encountered within 100 ft of the land surface, and 81 percent are encountered within 200 ft of land surface (fig. 6). Table 2 shows that more than two water-bearing zones per 100 ft of uncased borehole were encountered in the upper 100 ft of the geologic units, more than one water-bearing zone per 100 ft was encountered in the upper 200 ft, and less than one water-bearing zone per 100 ft was encountered below 200 ft. The large number of water-bearing zones per 100 ft at depths greater than 450 ft is because of small sample size.

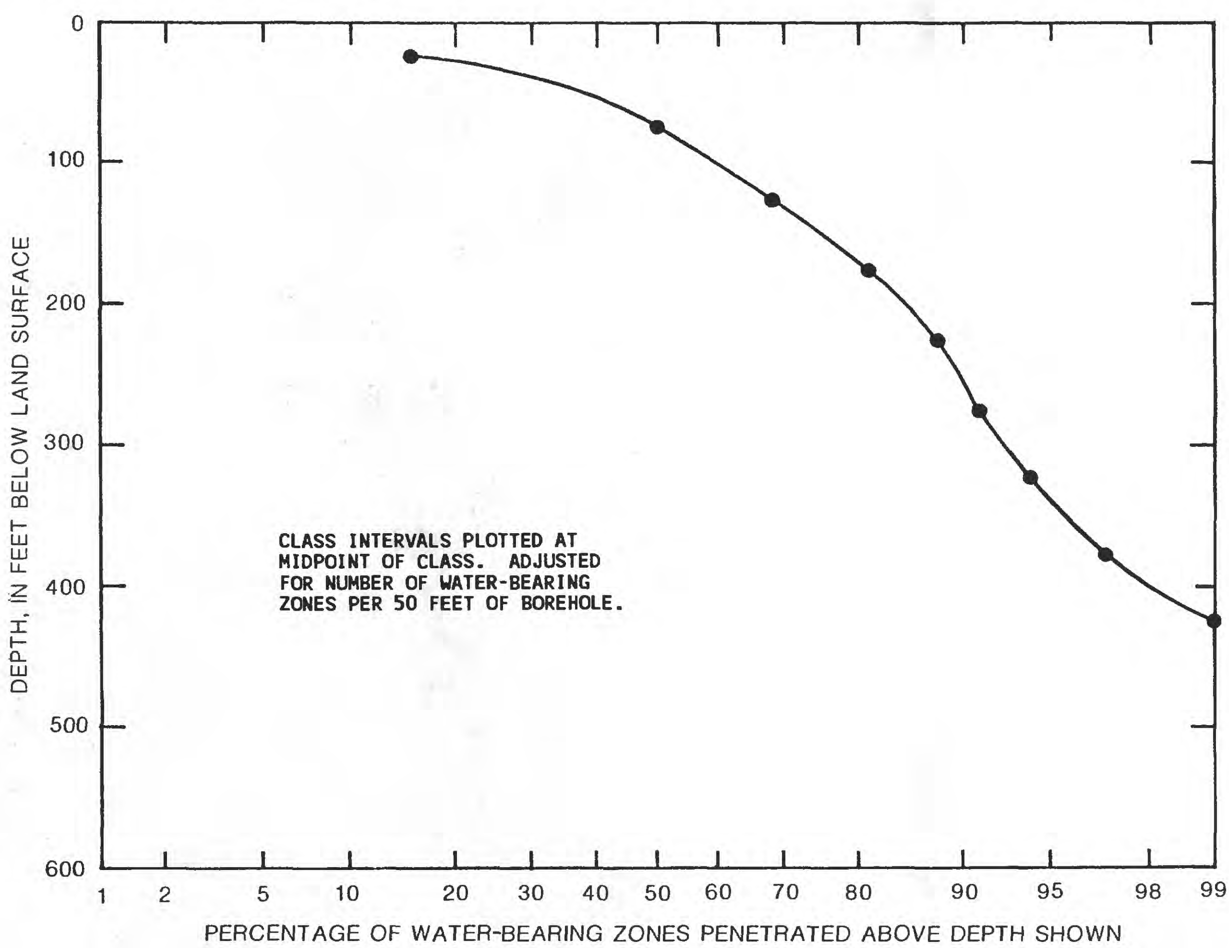


Figure 6.--Distribution of water-bearing zones with depth in carbonate rock.

Table 2.--Number of water-bearing zones per 100 feet of uncased borehole drilled in carbonate rocks

[ft, feet]

Interval (ft)	Number of water-bearing zones penetrated	Uncased footage drilled (ft)	Number of water-bearing zones per 100 ft of uncased borehole
0- 50	35	1,248	2.80
50-100	82	3,774	2.17
100-150	44	3,032	1.45
150-200	30	2,535	1.18
200-250	16	1,781	.90
250-300	6	1,367	.44
300-350	7	914	.77
350-400	4	730	.55
400-450	5	534	.94
450-500	5	389	1.28
500-550	0	73	0
550-600	1	50	2.00
600-650	0	5	0

Hydraulic Characteristics of Stratigraphic Units

Carbonate rocks

Secondary porosity and permeability exhibit great spatial variation in carbonate rocks; therefore, the yield and specific capacity of wells are highly variable. Well yield depends on the number and size of openings penetrated below the water table--the more water-bearing openings intersected and the larger their size, the greater the well yield. Figure 7 shows the caliper logs of two wells. Well CH-79 does not intersect any significant water-bearing zones; the yield is less than 1 gal/min (gallon per minute). Well CH-212 intersects two major and several minor water-bearing zones; the yield is 1,000 gal/min.

The reported yield and specific capacity of wells in Chester County are summarized in tables 3 and 4, respectively. The specific-capacity frequency distributions for wells in the Conestoga Limestone, Elbrook Formation, and Ledger Dolomite in Chester County are shown in figure 8. Wells in the Ledger Dolomite have the highest specific capacities; wells in the Elbrook Formation have the lowest specific capacities.

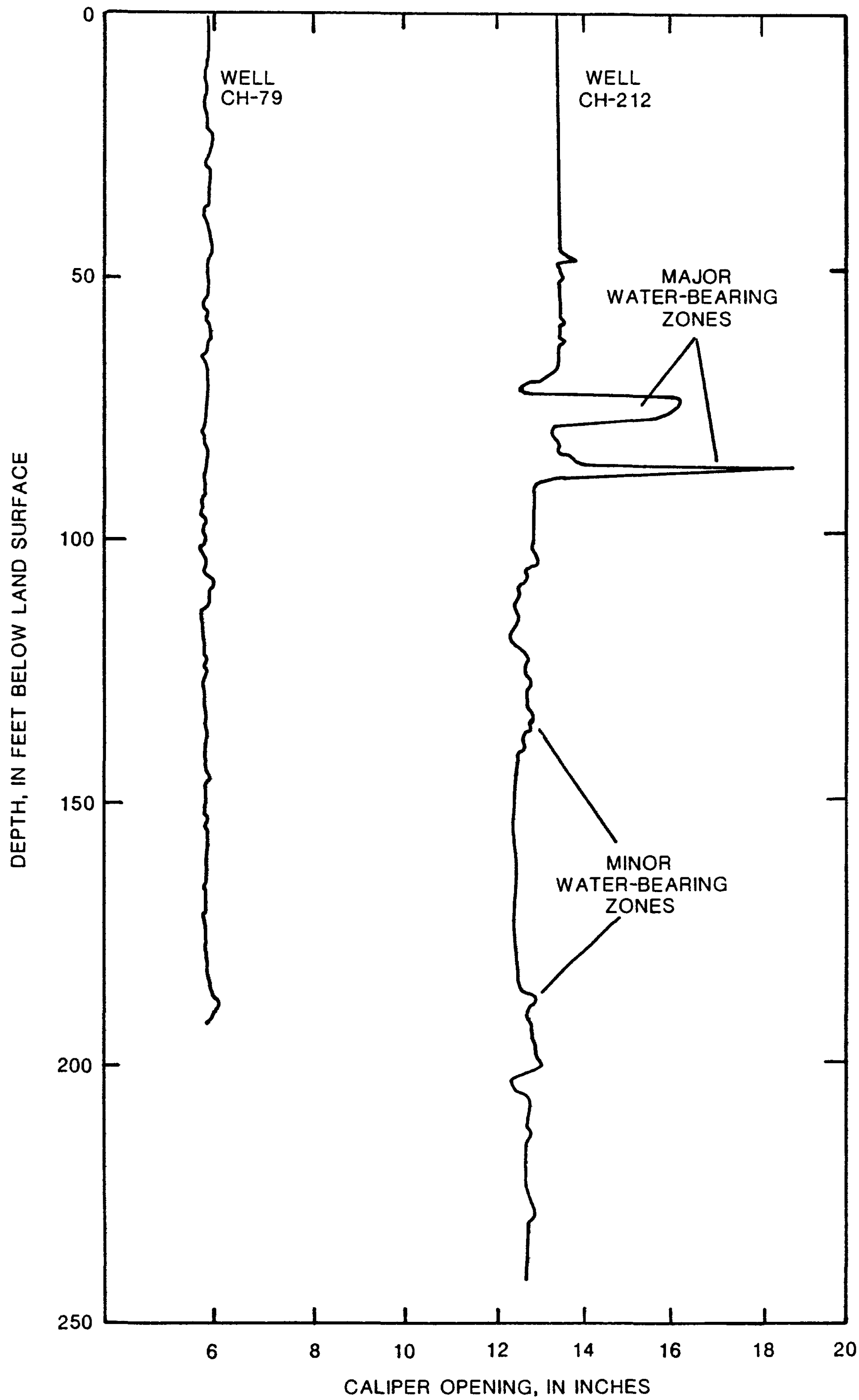


Figure 7.--Caliper logs of wells CH-79 and CH-212.

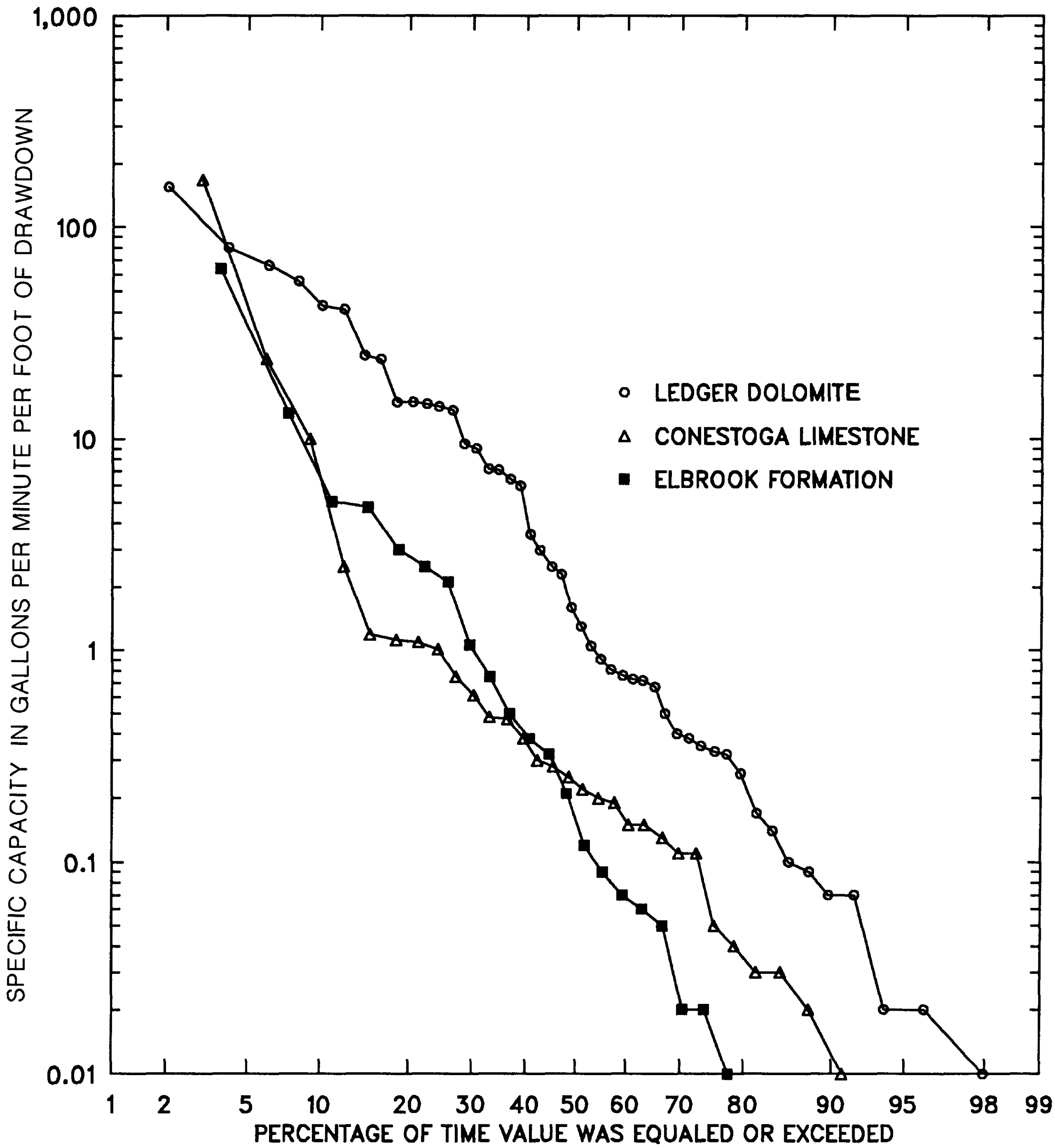


Figure 8.–Distribution of specific capacity of wells in the Conestoga Limestone, Elbrook Formation, and Ledger Dolomite.

Table 3.--Reported yields of wells

[Yields are in gallons per minute; <, less than;
--, too few data to compute range or median]

Geologic unit	All wells: Number of wells	All wells: Range	All wells: Median	Nondomestic wells: Number of wells	Nondomestic wells: Range	Nondomestic wells: Median
Stockton Formation	88	4- 800	20	23	12- 800	70
Conestoga Limestone	76	<1-1,000	18	16	2-1,000	65
Elbrook Formation	59	<1-1,200	15	14	5-1,200	75
Ledger Dolomite	81	<1-1,120	60	28	8-1,120	202
Kinzers Formation	4	10- 100	--	1	17	--
Vintage Dolomite	13	<1- 665	50	4	30- 665	--
Antietam Quartzite and Harpers Phyllite	27	1- 75	12	2	23- 75	--
Chickies Quartzite	100	<1- 100	13	10	<1- 96	25
Octoraro Phyllite	152	<1- 300	15	41	3- 230	40
Leucocratic gneiss	140	<1- 165	18	18	10- 167	55

Table 4.--Reported specific capacity of wells

[Specific capacity is in gallons per minute per foot of drawdown;
<, less than; --, too few data to compute range or median]

Geologic unit	All wells: Number of wells	All wells: Range	All wells: Median	Nondomestic wells: Number of wells	Nondomestic wells: Range	Nondomestic wells: Median
Stockton Formation	32	0.02- 14	0.38	4	2.5 - 14	--
Conestoga Limestone	27	<.01-167	.32	5	.01-167	--
Elbrook Formation	31	<.01- 24	.22	10	.02- 24	.33
Ledger Dolomite	48	.01-155	1.5	14	.5 -155	20
Kinzers Formation	3	.08- 1.4	--	1	.08	--
Vintage Dolomite	8	.27- 29	.81	4	.71- 29	--
Antietam Quartzite and Harpers Phyllite	14	.04- 3	.39	3	.11- .19	--
Chickies Quartzite	40	<.01- 5	.16	7	<.01- 3	.55
Octoraro Phyllite	68	<.01- 38	.32	14	.12- 3.8	1.3
Leucocratic gneiss	74	.01- 25	.28	14	.04- 25	.40

Conestoga Limestone.--The Conestoga Limestone is generally a low-yield carbonate rock. Well yields range from less than 1 to 1,000 gal/min; however, only one yield exceeds 225 gal/min. The median yield of nondomestic wells is 65 gal/min. Specific capacities of four nondomestic wells in Chester County range from 0.01 to 167 (gal/min)/ft (gallons per minute per foot) of drawdown.

Transmissivities based on analysis of four aquifer tests conducted for 24 to 72 hours on wells in the modeled area in Chester and Montgomery Counties ranged from 1,250 to 1,680 ft^2/d (feet squared per day); the median was 1,540 ft^2/d. The specific capacities of the four wells ranged from 15 to 27 (gal/min)/ft.

Elbrook Formation.--The Elbrook Formation is the least productive carbonate rock in Chester Valley. Yields range from less than 1 to 1,200 gal/min; however, only two yields exceed 500 gal/min. The median yield of nondomestic wells is 75 gal/min. The specific capacities of 10 nondomestic wells in Chester County range from 0.02 to 24 (gal/min)/ft; the median specific capacity is 0.83 (gal/min)/ft.

Transmissivities based on analysis of five aquifer tests conducted for 6.5 to 96 hours on wells in the modeled area in Chester County ranged from 13 to 2,740 ft^2/d; only one transmissivity exceeded 350 ft^2/d. The median was 211 ft^2/d. The specific capacities of the five wells ranged from 0.16 to 15 (gal/min)/ft.

Ledger Dolomite.--The Ledger Dolomite is the most productive aquifer in Chester County. Yields range from less than 1 to 1,120 gal/min. The median yield of nondomestic wells is 100 gal/min. Specific capacities of 14 nondomestic wells in Chester County range from 0.5 to 155 (gal/min)/ft; the median specific capacity is 20 (gal/min)/ft.

Transmissivities based on analysis of six aquifer tests conducted for 63 to 315 hours on wells in the modeled area in Chester and Montgomery Counties ranged from 600 to 15,430 ft^2/d; the median transmissivity was 1,680 ft^2/d. The specific capacities of the six wells ranged from 8 to 74 (gal/min)/ft.

Kinzers Formation and Vintage Dolomite.--Few data are available for wells in the Kinzers Formation and Vintage Dolomite in Chester County. These geologic units crop out together in a narrow band and are thin and of small areal extent. Because they are so thin, wells drilled into the Kinzers and Vintage probably derive water from underlying units, making the Kinzers and Vintage difficult to evaluate. Data on well yield and specific capacity for the few wells identified with these geologic units are summarized in tables 3 and 4.

Yields of four wells in the Kinzers Formation range from 10 to 100 gal/min. The specific capacity of three wells range from 0.08 to 1.4 (gal/min)/ft.

Yields of 13 wells in the Vintage Dolomite range from less than 1 to 665 gal/min; however, only one yield exceeds 100 gal/min. The median yield is 50 gal/min. Specific capacities of four nondomestic wells range from 0.71 to 29 (gal/min)/ft.

Noncarbonate rocks

Well yields in the Stockton Formation range from 4 to 800 gal/min. The median yield of nondomestic wells is 70 gal/min. Specific capacities of four nondomestic wells range from 2.5 to 14 (gal/min)/ft.

Well yields in the Antietam Quartzite and Harpers Phyllite (undivided) range from 1 to 75 gal/min; the median yield is 12 gal/min. Specific capacities of three wells range from 0.04 to 0.19 (gal/min)/ft.

Well yields in the Chickies Quartzite range from less than 1 to 100 gal/min. The median yield of nondomestic wells is 25 gal/min. Specific capacities of seven nondomestic wells range from less than 0.01 to 3 (gal/min)/ft; the median specific yield is 0.55 (gal/min)/ft.

Well yields in the leucocratic and intermediate felsic gneiss range from less than 1 to 165 gal/min. The median yield of nondomestic wells is 55 gal/min. Specific capacities of 14 nondomestic wells in Chester County range from 0.04 to 25 (gal/min)/ft; the median specific yield is 0.4 (gal/min)/ft.

Well yields in the Octoraro Phyllite range from less than 1 to 300 gal/min. The median yield of nondomestic wells is 40 gal/min. Specific capacities of 14 nondomestic wells range from 0.12 to 3.8 (gal/min)/ft; the median specific capacity is 1.3 (gal/min)/ft.

Water-Level Fluctuations

The carbonate rocks of Chester Valley form a complex, heterogeneous water-table aquifer. The water table fluctuates in response to recharge from precipitation, discharge from pumped wells and quarries, evapotranspiration, and discharge to streams. The water table generally rises during the fall and winter when evapotranspiration is at a minimum and recharge is at a maximum; it declines during the spring and summer when evapotranspiration is at a maximum and recharge is at a minimum. Wells in different parts of Chester Valley have similar hydrographs (fig. 9).

Shallow and deep wells have similar water levels and similar responses to precipitation. Two wells in the Conestoga Limestone 200 ft apart--one shallow and one deep--were equipped with continuous water-level recorders. Well CH-80 is 57 ft deep, and well CH-2448 is 300 ft deep. The water-level altitude was about the same in both wells (fig. 10). The water levels in both wells had a similar response to precipitation and similar fluctuations (fig. 11).

Although the ground-water system is generally under water-table conditions, confined ground water is present locally. In carbonate rocks, ground water can be confined by the relatively impermeable sides of a fracture or solution channel. Confined conditions were observed in three wells in the Elbrook Formation in the center of the Valley Creek basin. The water level in well CH-2313 rose above land surface in the spring of 1978, 1983, and 1984; it was 1.50 ft above land surface on April 21, 1983. Well CH-2148 is a flowing well; the head was reported to be 56 ft above land surface on February 13, 1969 (Leggette, Brashears, and Graham, undated written commun.). The continuous water level recorded in observation well CH-323 shows the effect of earth tides, which indicates confined conditions.

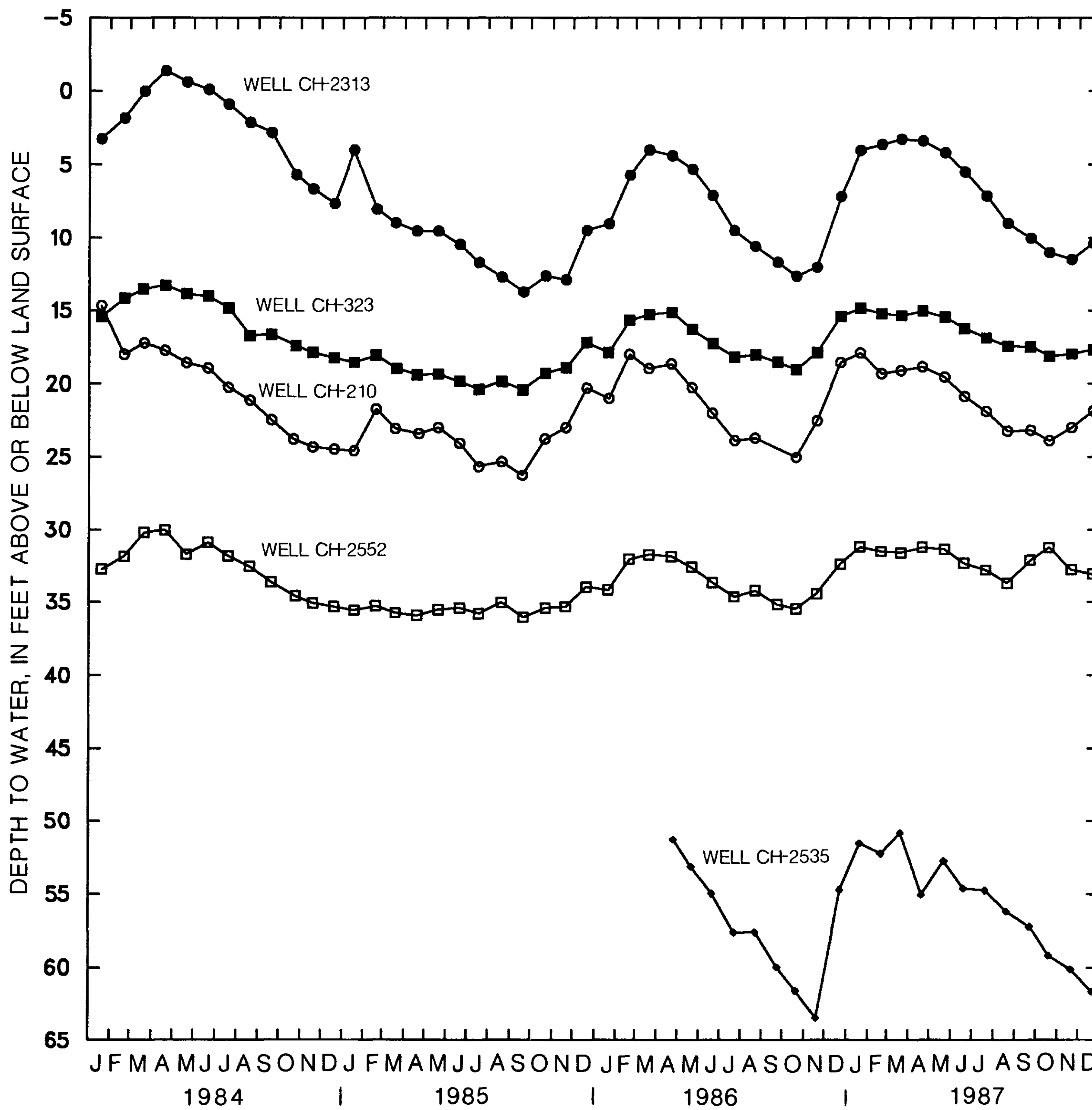


Figure 9.--Hydrographs of five wells in Chester Valley, 1984-87.

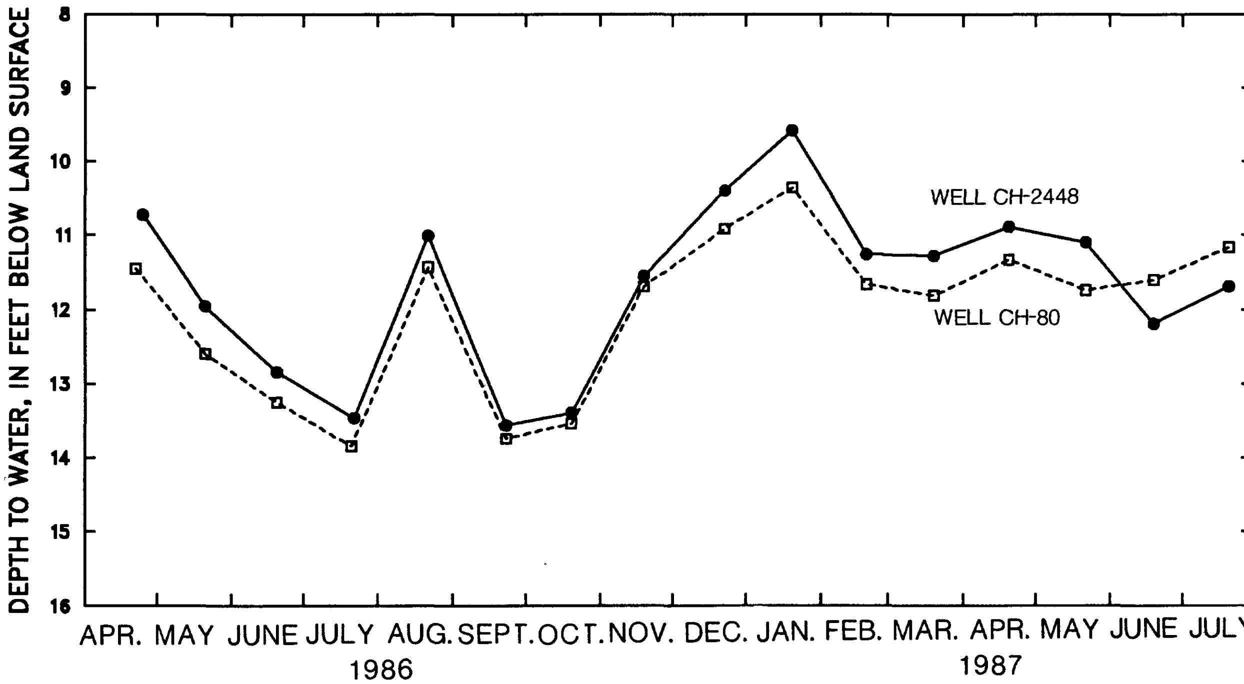


Figure 10.--Hydrographs of water levels measured monthly in wells CH-80 and CH-2448, April 1986 to July 1987.

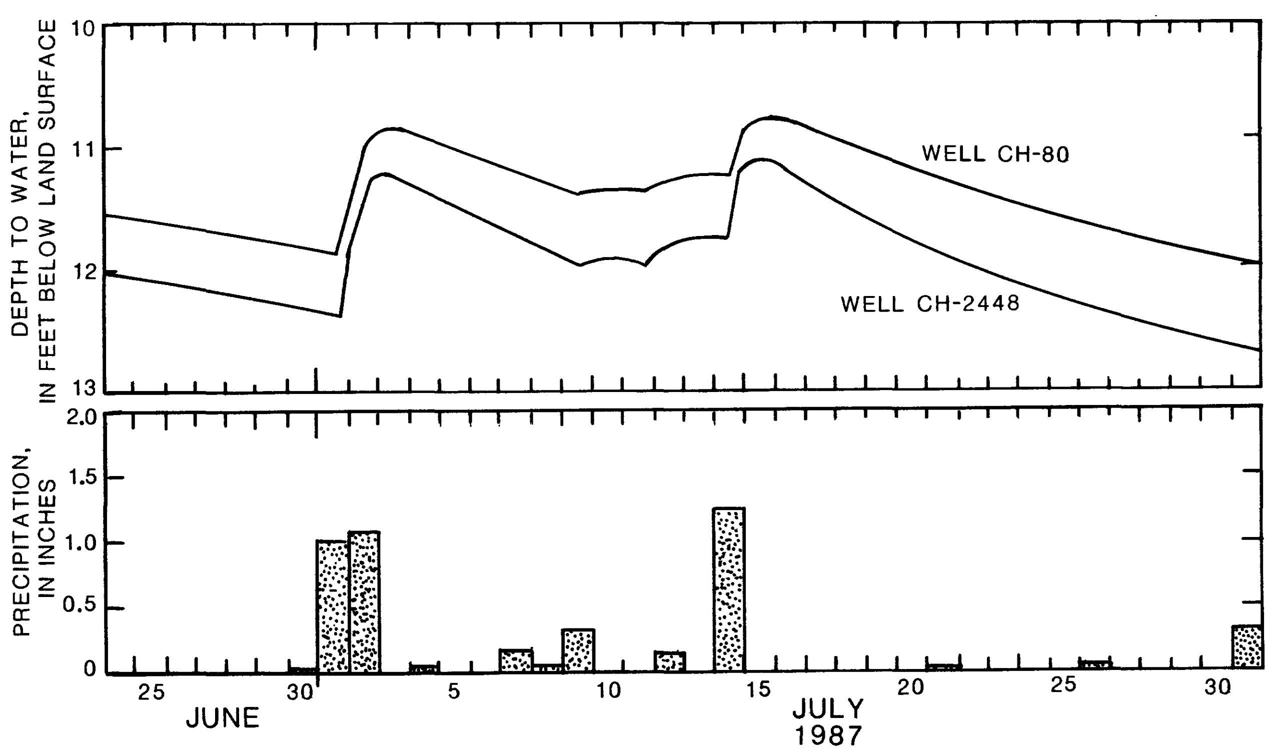


Figure 11.--Precipitation histogram and hydrographs of wells CH-80 and CH-2448 showing similar response to precipitation, June 24 to July 31, 1987.

Ground-Water/Surface-Water Relations

The ground-water and surface-water systems are well connected in the Valley Creek basin. Where Valley Creek flows over fractured bedrock, the aquifer and the surface-water system are in direct contact. In most areas, ground water discharges to streams and makes up the base-flow component of streamflow. In some areas, stream reaches lose water to the ground-water system.

Base-flow separations were made on hydrographs of Valley Creek using a computer program developed by Sloto (in press). The local minimum method was used. Ground-water discharge (base flow) made up 73 to 84 percent of the annual flow of Valley Creek measured at the gaging station during 1983-87 (table 5). The average ground-water discharge was 76 percent of streamflow. This includes both natural ground-water discharge and quarry pumpage discharged to Valley Creek. The base flow of Valley Creek ranged from 11.56 in. in 1985 to 21.87 in. in 1984; the average was 17.00 in. Figure 12 shows streamflow and base flow of Valley Creek for 1984, a year of high base flow, and 1985, a year of low base flow.

Table 5.--Streamflow and base flow of Valley Creek and discharge from Cedar Hollow quarry, 1983-87

[in., inches]

Year	Streamflow (in.)	Base flow (in.)	Base flow Percent of streamflow	Cedar Hollow quarry discharge (in.)	Cedar Hollow quarry discharge Percent of base flow	Cedar Hollow quarry discharge Days of record
1983	26.98	19.69	73	[1]5.19	26	0
1984	28.57	21.87	77	[1]4.83	22	127
1985	15.55	11.56	74	2.94	25	245
1986	20.47	15.16	74	3.30	22	345
1987	19.98	16.72	84	3.54	21	344
Average	22.31	17.00	76	3.96	23	

[1] Estimated

Discharge from the Cedar Hollow quarry made up 21 to 26 percent of the annual base flow of Valley Creek measured at the streamflow-gaging station for 1984-87; the average was 23 percent (table 5). Quarry discharge for 1983-85 was estimated because measurements did not begin until June 1984 and data for 1985 were incomplete. Quarry discharge was subtracted from base flow after hydrograph separation.

Because the water table is lowered by quarry dewatering, ground water does not discharge to streams, and no perennial streams flow in the subbasin where the quarries are located. However, the discharge from the Cedar Hollow quarry sustains a greater than natural base flow in Valley Creek. The

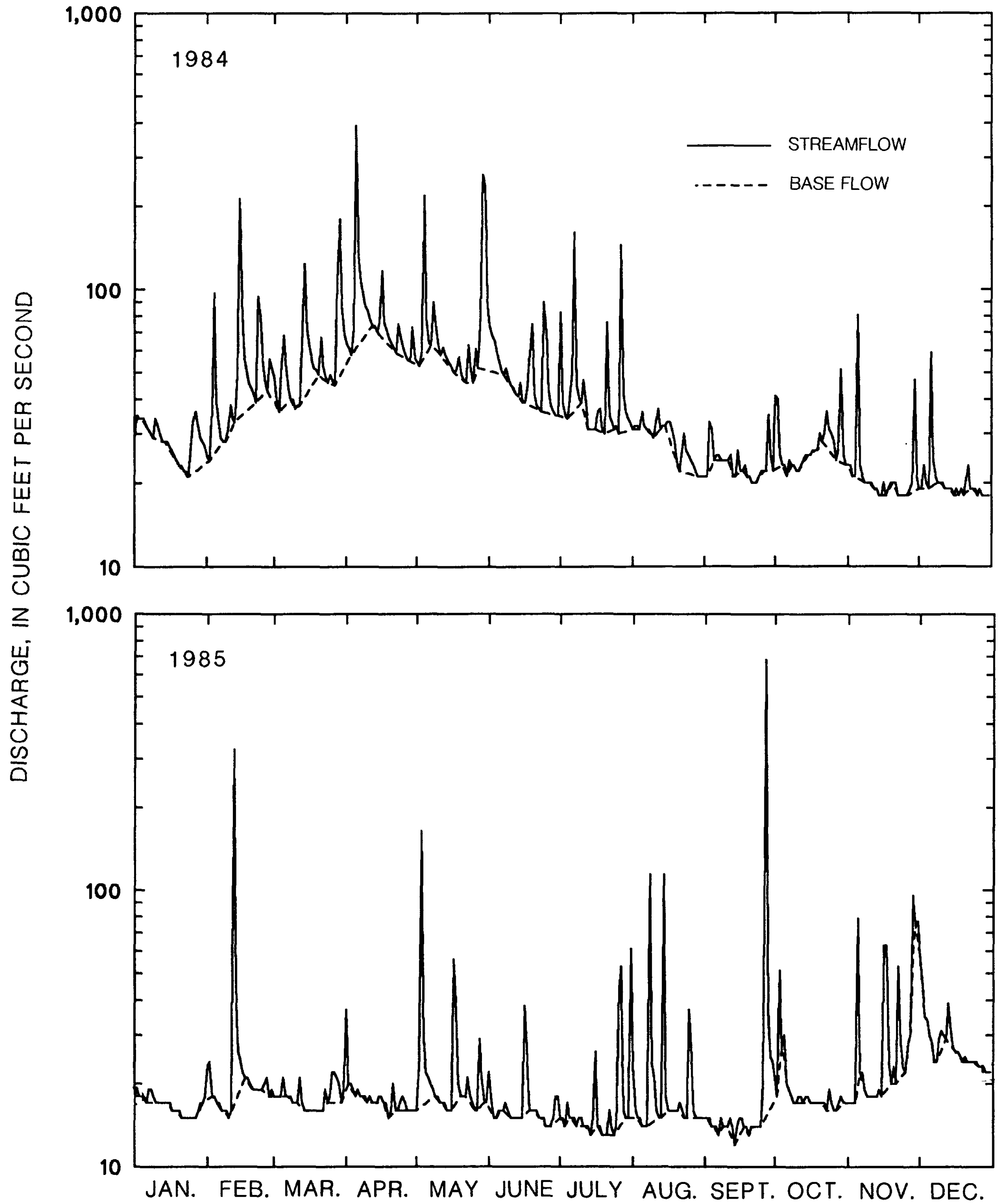


Figure 12.--Streamflow and base-flow hydrographs of Valley Creek at Pennsylvania Turnpike bridge near Valley Forge (station 01473169), 1984-85.

surface-water drainage area around the quarries is 2.62 mi^2. The ground-water basin that provides ground-water inflow to the quarries is 3.22 mi^2. Average annual base flow for 1984-87 for the Valley Creek basin above the gaging station (16.33 in.), excluding the quarry ground-water basin (3.65 in.), is 12.68 in. for 17.58 mi^2 or 0.72 in/mi^2 (inches per square mile). Average quarry discharge for 1984-87 is 3.65 in. for 3.22 mi^2 or 1.13 in/mi^2, which is 57 percent greater than the rest of the Valley Creek basin. Assuming that the natural ground-water discharge from the quarry basin would also be 0.72 in/mi^2, the average annual base flow of Valley Creek would be 14.98 in. or 8 percent lower than if the quarry were not being dewatered.

Ground-water discharge from the quarry area is greater than ground-water discharge from the rest of the Valley Creek basin for several reasons. Little vegetation grows in the quarry or on stripped areas, active berms, and spoil piles; therefore, evapotranspiration in the quarry area is much less than in the rest of the basin. Ground-water evapotranspiration is probably negligible because the water table has been significantly lowered around the quarries. Reduction of soil-moisture evapotranspiration and ground-water evapotranspiration increase the quantity of ground-water discharge. In addition, the precipitation falling on the quarry and surface runoff from the surrounding contributing area drain to the quarry sumps and are discharged with the ground-water inflow to the quarry.

All tributaries to Valley Creek have their headwaters in the crystalline rocks north and south of Chester Valley. Because some streams cross the contact between the crystalline and carbonate rock, they lose water or become perched above the water table for a short distance. The hydraulic conductivity of carbonate rock is much greater than that of the crystalline rock, and the water table generally is lower on the carbonate-rock side of the contact. The quantity of streamflow lost is small because the discharge of these headwater streams is small.

Regional ground-water flow through the Valley Creek basin is northeast to the Schuylkill River. Figure 13 shows the generalized direction of ground-water flow. The diagram shows three ground-water mounds and two ground-water sinks. The mounds correspond to topographically high areas of low hydraulic conductivity. The sinks are cones of depression around active quarries.

On the western side of the Valley Creek basin, the ground-water divide is about 1/2 mi west of the surface-water divide (fig. 13). Some ground water flows from the adjacent West Valley Creek basin eastward into the Valley Creek basin. Based on Darcy's law and water-table maps of Sloto (1987, pl. 2) and Wood (1984), the estimated inflow from the west is about 0.75 Mgal/d (million gallons per day) or 0.76 in/yr (inches per year). A saturated thickness of 550 ft, the zone of active ground-water circulation, was assumed.

A ground-water divide is not present on the eastern side of the Valley Creek basin. The water table slopes gently eastward toward the Schuylkill River. On the northeastern side of the Valley Creek basin, ground water flows out of the basin beneath the surface-water divide and flows northeast beneath Valley Forge National Historical Park to the Schuylkill River. Based on Darcy's law and the water-table map of Sloto (1987, pl. 2), the estimated underflow out of the Valley Creek basin to the east is about 1.76 Mgal/d or

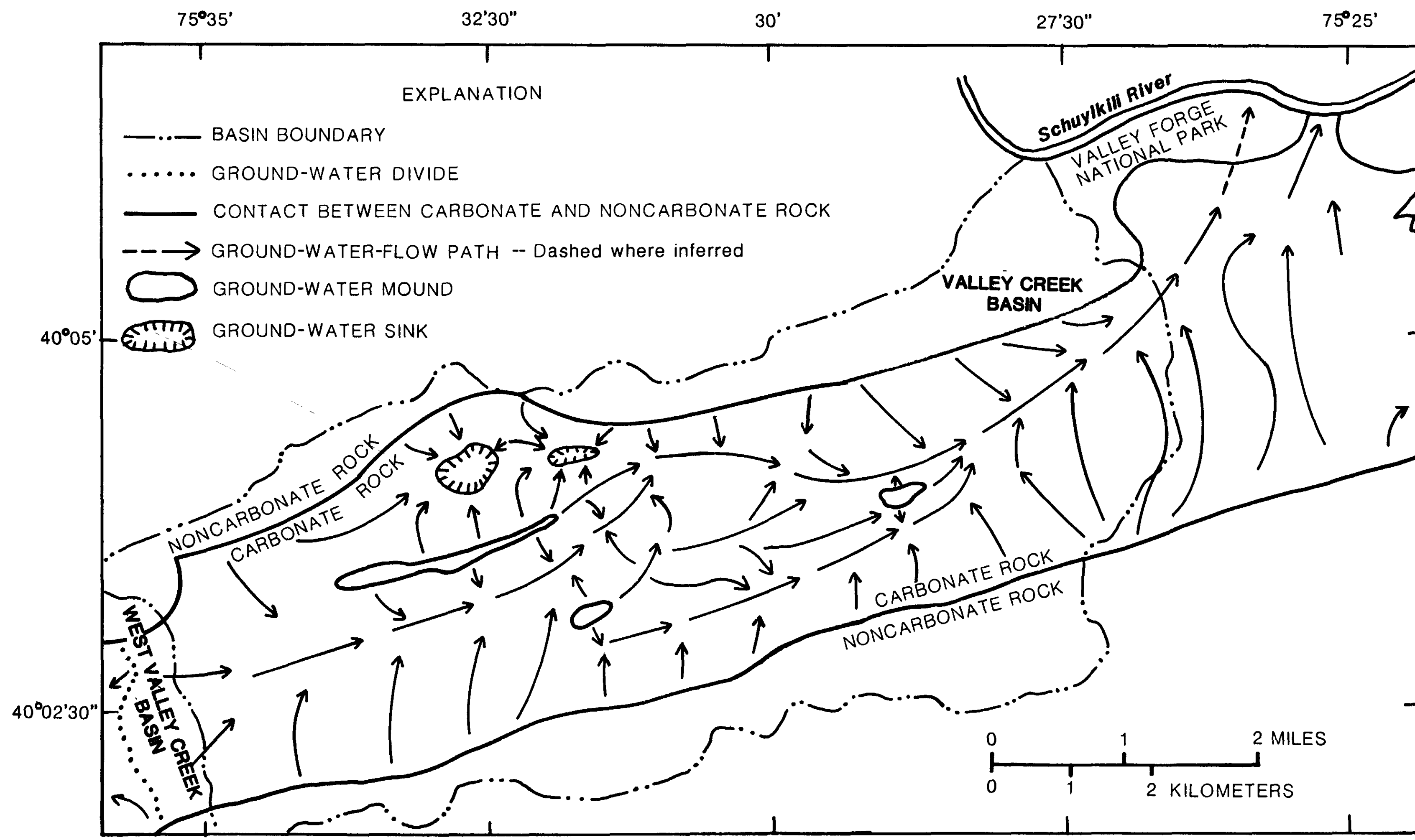


Figure 13.--Generalized direction of ground-water flow in the carbonate rocks of the Valley Creek basin.

1.78 in/yr. On the southeastern side of the Valley Creek basin, some ground water flows beneath the surface-water divide into the basin. Based on Darcy's law and the water-table map of Sloto (1987, pl. 2), estimated ground-water inflow to the basin from the southeast is 0.85 Mgal/d or 0.86 in/yr.

During low streamflow when the water table is low, Valley Creek loses water between the streamflow-gaging station and the contact between the carbonate and crystalline rocks where Valley Creek leaves Chester Valley. The measured loss in this reach was 3.23 Mgal/d on October 10, 1984, and 0.94 Mgal/d on July 1, 1985. The streamflow measured at the bridge on Wilson Road (station number 01473170 on pl. 1) was 20.1 and 17.7 ft^3/s (cubic feet per second), respectively. An increase in streamflow between these points was measured on April 28, August 13, and November 11, 1975, when the streamflow at the bridge on Wilson Road was 50, 38.1, and 34.5 ft^3/s, respectively.

Water Budget

A water budget is an estimate of water entering and leaving a basin plus or minus changes in storage for a given time period. Water enters as precipitation and leaves as ground-water flow, streamflow, evapotranspiration, and diversions, such as ground-water pumpage. Water also is taken into or released from ground-water and soil-moisture storage.

Annual water budgets for 1983-87 and an average water budget for those years were calculated for the 20.8-mi^2 part of the basin above the streamflow-gaging station. The streamflow gage measures the quantity of water leaving the basin above the gage as streamflow. Water-level records from observation well CH-323 were used to estimate the annual change in ground-water storage. Ground-water withdrawals from the basin are calculated from pumpage data supplied by the Philadelphia Suburban Water Company; estimated commercial, industrial, and domestic pumpage (Sloto, 1987, p. 22); and water pumped by the Catanach quarry not returned to the ground-water system. These ground-water withdrawals are consumptive use or are removed from the Valley Creek basin as exported ground water. Records of pumpage from the Cedar Hollow quarry were supplied by the Warner Company. Precipitation was measured at a rain gage maintained in the basin by the U.S. Geological Survey.

The annual change in ground-water storage was estimated using water-level data from observation well CH-323. Specific yield was calculated from the change in water level in well CH-323 for six periods between August 24, 1984, and June 10, 1986. The selection criteria for these periods were:

(1) Streams were at base flow. No rain fell immediately prior to the start of the period.

(2) No recharge was observed in the hydrograph of well CH-323. Little or no rain fell during the period; any precipitation that fell went to satisfy the soil-moisture deficit.

(3) The period was longer than 10 days.

(4) Streamflow, water-level, rainfall, and quarry-pumpage data were available preceding and during the period.

Six periods, from 10 to 26 days each, met these criteria. For each period, daily streamflow was summed to calculate the quantity of water leaving the basin. Quarry pumpage discharged to Valley Creek was subtracted from the streamflow, and the result was divided by 0.85 (1 minus the ratio of the 3.22-mi^2 ground-water basin around the quarries to the 20.8-mi^2 Valley Creek basin above the gaging station) to remove the contribution of the ground-water basin around the quarries. The result was then divided by the change in water level observed in well CH-323 to calculate specific yield. This specific yield is assumed to be representative of the basin. Figure 14 shows the relation between the water level in well CH-323 and specific yield for the six periods. Specific yield decreases with depth because aquifer storage decreases with depth. This is consistent with gravity yield calculations by Olmsted and Hely (1962, p. 17) for the adjacent Brandywine Creek basin.

The change in ground-water storage in the water budgets was calculated by averaging the water level in CH-323 at the begining and end of the year. The specific yield from figure 14 corresponding to the average water level was multiplied by the change in water level during the year to calculate the annual change in ground-water storage.

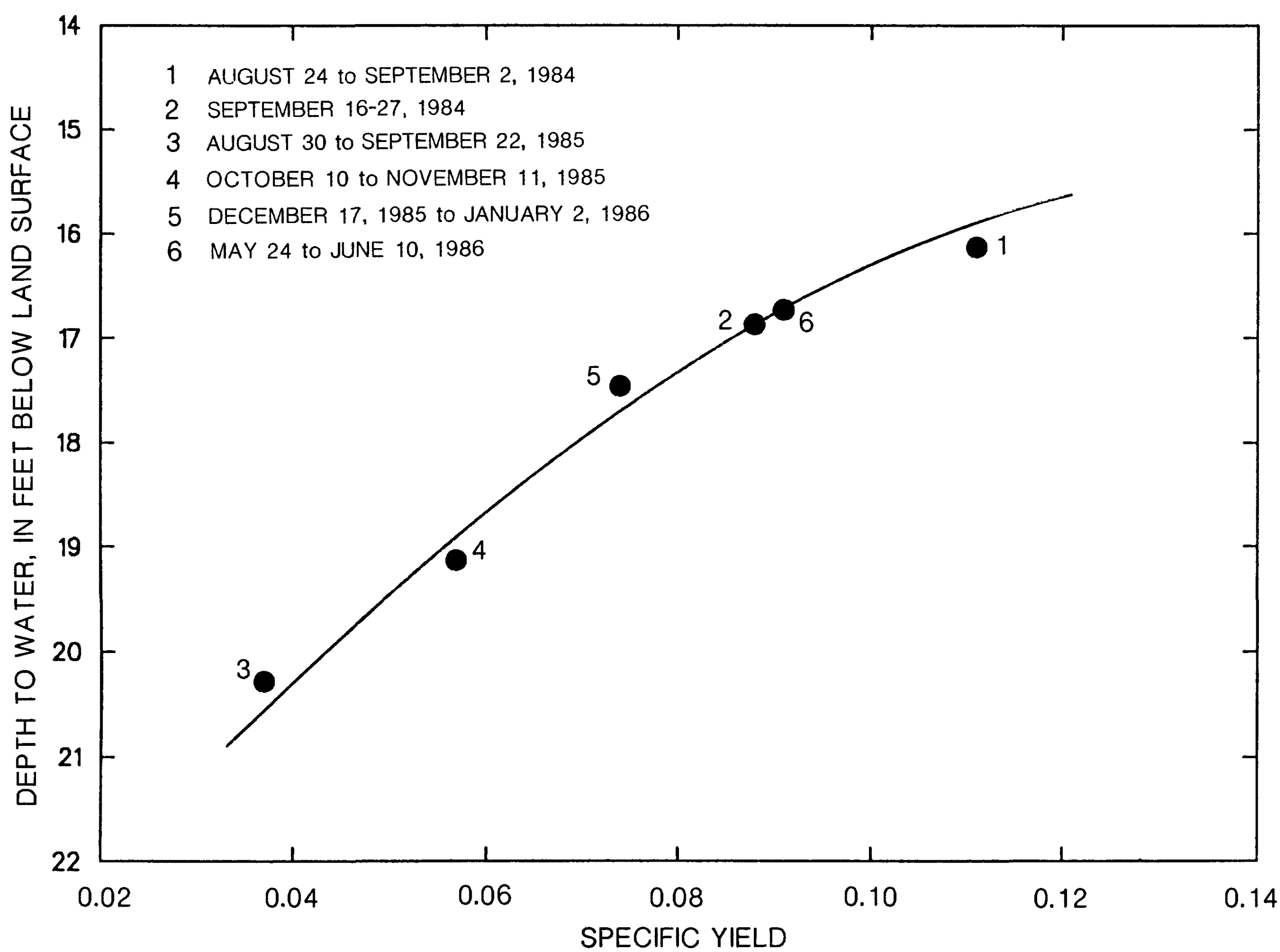


Figure 14.--Relation between specific yield and depth to water in well CH-323.

Because the water budget begins and ends in winter when soil moisture is usually at field capacity, the change in soil moisture is equal to zero, and a soil-moisture term is not included in the water-budget equation. The annual water budget can be expressed as

$$P = SF + GW + DS + ET, \qquad (1)$$

where

P = precipitation,
SF = streamflow,
GW = ground-water withdrawals from wells removed from the basin,
DS = change in ground-water storage, and
ET = evapotranspiration.

All of the terms in the water-budget equation are known except evapotranspiration; the equation is solved for evapotranspiration.

Water budgets for 1983-87 and an average water budget for the 20.8-mi^2 area above the gaging station, expressed in inches of water, are given in table 6. In the water budgets, ground-water underflow into the basin from the west and ground-water loss to sewers (Sloto, 1987, p. 24-30) were not considered. The quantity of these two components are assumed to be equal and to balance each other.

The average water budget for 1983-87 is assumed to approximate long-term or steady-state conditions. The average water budget shows a decrease in ground-water storage of only 0.09 in. during this period. Evapotranspiration ranges from 18.21 to 24.83 in. The average for 1983-87 is 22.90 in.; this is a reasonable approximation for the long-term average.

Recharge

Precipitation that infiltrates and does not replenish soil moisture recharges the ground-water system. Because carbonate rocks have a higher hydraulic conductivity than crystalline rocks, the rate of recharge of carbonate rocks also tends to be higher.

Annual and average recharge for the 20.8-mi^2 area of the Valley Creek basin above the gaging station were estimated for 1983-87 using the following equation:

$$R = BF + GW + DS + GWET, \qquad (2)$$

where

R = recharge,
BF = base flow,
GW = ground-water withdrawals from wells removed from the basin,
DS = change in ground-water storage, and
GWET = ground-water evapotranspiration.

Annual and average recharge estimates for 1983-87, expressed in inches, are given in table 7. Base flow was taken from table 5 and ground-water withdrawals and change in ground-water storage were taken from table 6. Ground-water evapotranspiration was estimated. Ground-water underflow into the basin from the west and ground-water loss to sewers are assumed to be equal and to balance each other.

Recharge for 1983-87 ranged from 15.89 to 26.84 in.; the average was 21.04 in. Base flow was equal to 73-79 percent of recharge except for 1984, when base flow was equal to slightly more than the total recharge because 4.13 in. of water was released from ground-water storage. For 1983-87, average annual base flow was 81 percent of average annual recharge.

Table 6.--Annual and average water budgets for the Valley Creek basin above the gaging station, 1983-87

[Values are in inches per year]

Year	Precipitation	Streamflow[1]	Ground-water withdrawals	Change in ground-water storage	Evapotranspiration
1983	56.55	26.98	1.84	+3.31	24.42
1984	48.74	28.57	2.03	-4.13	22.27
1985	42.71	15.55	1.91	+.42	24.83
1986	47.65	20.47	2.32	+.10	24.76
1987	40.61	19.98	2.56	-.14	18.21
Average	47.25	22.31	2.13	-.09	22.90

[1] Includes quarry pumpage discharged to Valley Creek

Table 7.--Recharge in the Valley Creek basin, 1983-87

[Values are in inches per year]

Year	Recharge	Base flow[1]	Ground-water withdrawals	Change in ground-water storage	Ground-water evapotranspiration[2]
1983	26.84	19.69	1.84	+3.31	2.00
1984	21.77	21.87	2.03	-4.13	2.00
1985	15.89	11.56	1.91	+.42	2.00
1986	19.58	15.16	2.32	+.10	2.00
1987	21.14	16.72	2.56	-.14	2.00
Average	21.04	17.00	2.13	-.09	2.00

[1] Includes quarry pumpage discharged to Valley Creek
[2] Estimated

Ground-Water Quality

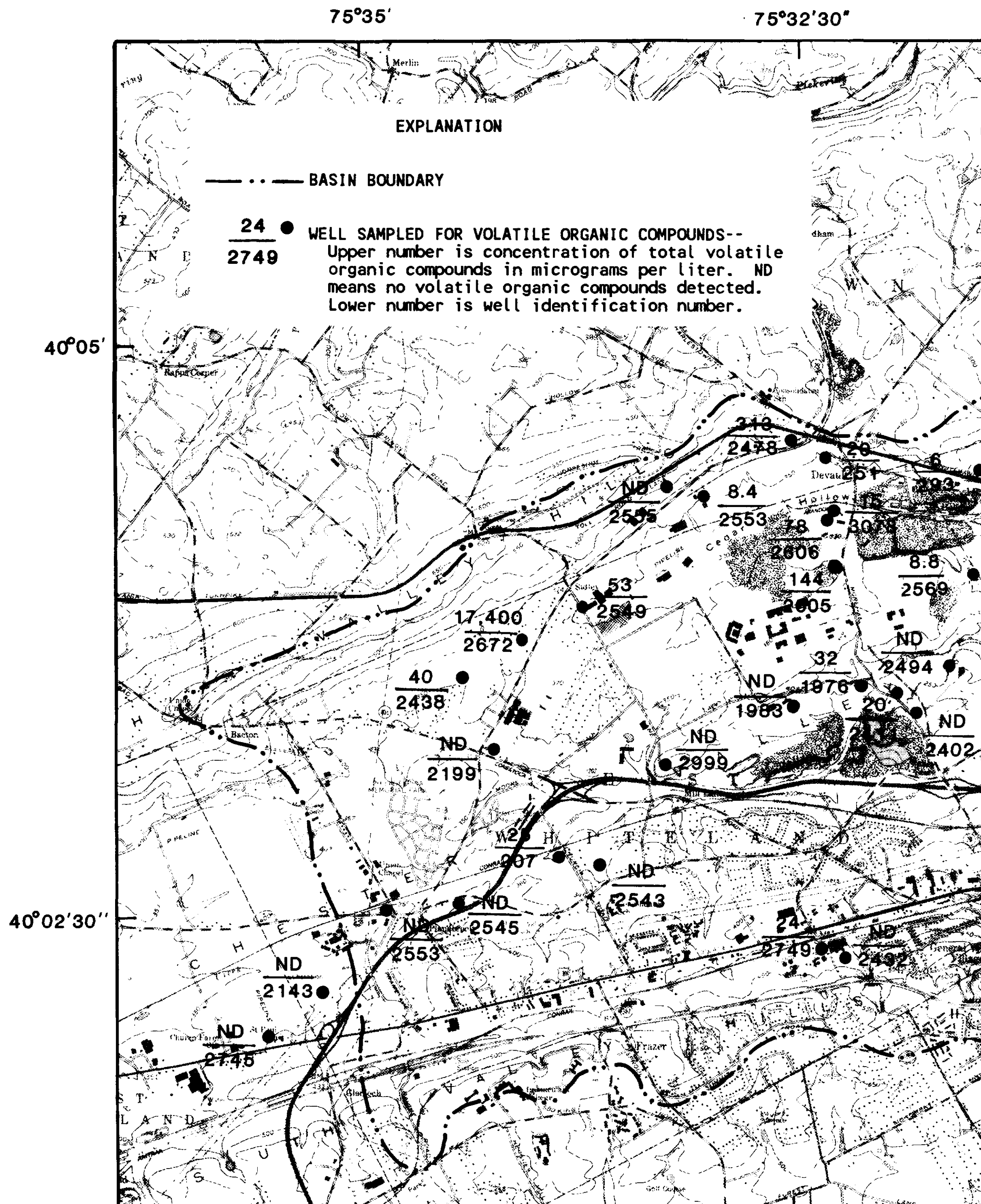
The quality of water is determined by the type and quantity of substances dissolved in it. As water moves through the hydrologic cycle, it dissolves gasses and mineral matter from the atmosphere, soil, and rock. Additional substances may be added by human activities.

Analyses of water from wells in the Valley Creek basin are reported by Sloto (1989). Except where affected by human activity, water from wells in the carbonate rocks is of good quality and suitable for most purposes.

Volatile organic compounds

The most serious consequence of urbanization in the Valley Creek basin is the degradation of ground water by man-made organic compounds, particularly volatile organic compounds (VOCs). VOC contamination reduces the quantity of water available for potable supply and is expensive to treat. Sloto (1987, p. 42-47) described the movement of VOCs through the carbonate rocks of the Valley Creek basin under the influence of head gradients caused by quarry pumping. Movement of ground water through the quarries causes the VOCs to volatilize and reduces their concentrations below detection limits, except for trichloroethylene (TCE), which was still detectable in water discharged from the Cedar Hollow quarry.

Other areas in the Valley Creek basin also are affected by VOC contamination (fig. 15). The compounds detected and maximum concentrations are given in table 8. Out of 27 VOCs analyzed, 13 were detected. VOCs were found in 49 percent of the wells sampled. The concentration of total VOCs was as high as 17,400 $\mu g/L$ (micrograms per liter). TCE was the most frequently detected VOC; it was detected in water from 37 percent of the wells sampled. 1,1,1-trichloroethane was detected in water from 20 percent of the wells sampled, 1,2-trans-dichloroethylene and methylene chloride were detected in water from 11 percent of the wells sampled, and tetrachloroethylene (PCE) was detected in water from 9 percent of the wells sampled. These compounds are commonly used solvents and degreasers. 1,2-trans-dichloroethylene is a breakdown product of TCE and PCE.



Base from U.S. Geological Survey Malvern 1:24,000, 1973 and Valley Forge 1:24,000, 1981.

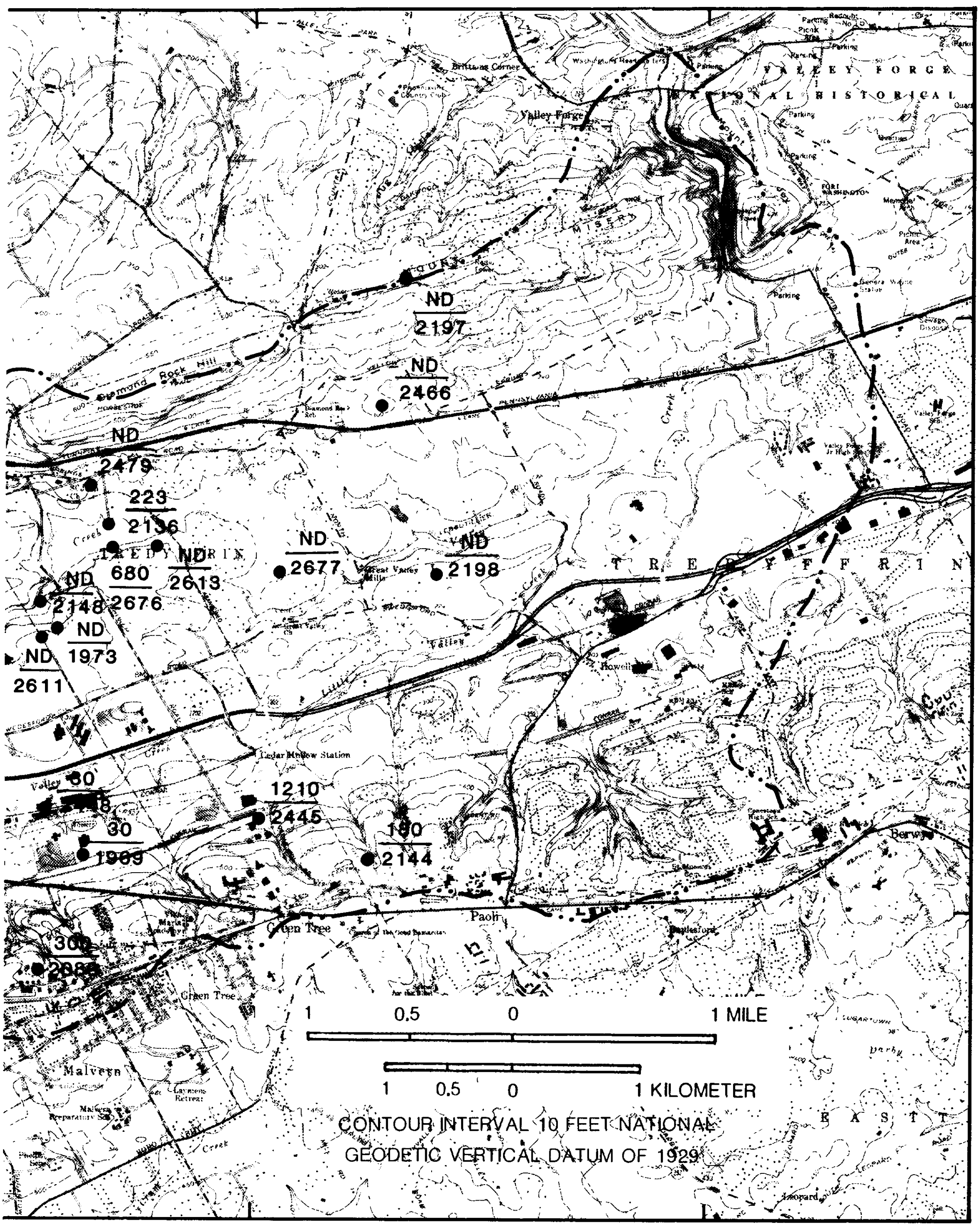


Figure 15.--Concentration of total volatile organic compounds in ground water in the Valley Creek basin.

Table 8.--Frequency of occurrence of volatile organic compounds in water from wells in carbonate rocks

[μg/L, micrograms per liter; --, all concentrations were below detection limit]

Compound	Number of wells sampled	Number of wells with concentration above detection limit	Maximum concentration (μg/L)	Percentage of wells with concentration above detection limit
Benzene	35	2	3.0	6
Bromoform	35	0	--	0
Carbon tetrachloride	35	0	--	0
Chlorobenzene	35	1	2.0	3
Chlorodibromomethane	35	0	--	0
Chloroethane	24	0	--	0
2-Chloroethyl vinyl ether	27	0	--	0
Chloroform	35	2	190	6
Dichlorobromomethane	35	0	--	0
Dichlorodifluoromethane	31	0	--	0
1,1-Dichloroethane	35	1	39	3
1,2-Dichloroethane	35	1	140	3
1,1-Dichloroethylene	35	2	5,400	6
1,2,-trans-Dichloroethylene	35	4	560	11
1,2-Dichloropropane	35	0	--	0
1,3-Dichloropropane	27	0	--	0
Ethylbenzene	35	0	--	0
Methylbromide	24	0	--	0
Methylene chloride	35	4	49	11
1,1,2,2-Tetrachloroethane	35	0	--	0
Tetrachloroethylene	35	3	1,200	9
Toluene	35	1	2.0	3
1,1,1-Trichloroethane	35	7	6,700	20
1,1,2-Trichloroethane	35	0	--	0
Trichloroethylene	35	13	4,400	37
Trichlorofluoromethane	32	1	3.0	3
Vinyl chloride	24	0	--	0
Total volatile organic compounds	35	17	17,400	49

Inorganic constituents

The dominant cation in water from the carbonate rocks is calcium; the dominant anion is bicarbonate. Acidic precipitation readily dissolves calcium, magnesium, and carbonate from limestone ($CaCO_3$) and dolomite [$CaMg(CO_3)_2$]. This causes water from wells in carbonate rock to have greater concentrations of total dissolved solids than water from wells in less soluble noncarbonate rock. Water from 23 wells was analyzed for inorganic constituents. The concentration of total dissolved solids in water from six wells exceeded the U.S. Environmental Protection Agency (USEPA) maximum contaminant level of 500 mg/L (milligrams per liter) for total dissolved solids in drinking water (U.S. Environmental Protection Agency, 1983, p. 293). The dissolved manganese concentration in water from one well exceeded the USEPA maximum contaminant level of 50 µg/L.

In the Planebrook area, a plume of ground water containing elevated concentrations of lithium as high as 13,000 µg/L and boron as high as 20,000 µg/L is moving through the carbonate rocks and discharging to Valley Creek. Boron and lithium in the ground water of this area are described by Sloto (1987, p.54-61).

Water from some wells contains elevated concentrations of chloride caused by the use of highway deicing salt. Sloto (1987, p. 66-70) described elevated chloride concentrations in ground water at a former salt storage site and in water from wells downgradient from the Pennsylvania Turnpike.

SIMULATION OF GROUND-WATER FLOW

Ground-water flow in the Valley Creek basin was simulated using a numerical computer model. The model is a simplified mathematical representation of the complex hydrologic system in the basin. In order to simulate ground-water flow mathematically, certain assumptions regarding the hydrologic system were made and a simplified conceptual model developed. These are described in the following sections. The model approximates the hydrologic system within these imposed constraints and other limitations, which also are discussed below.

Description of Flow Model

Ground-water flow was simulated using the computer program of McDonald and Harbaugh (1984). The model is a finite-difference, two-dimensional model that uses block-centered nodes. The geologic units in the modeled area were simulated as a single water-table aquifer. Recharge to, ground-water flow through, and discharge from the rocks of Chester Valley were simulated.

Sources of water to the modeled hydrologic system are areally varied recharge and quarry pumpage discharged to a sinkhole. Discharge of water from the modeled hydrologic system is by pumpage from wells and quarries, ground-water discharge to streams, and ground-water evapotranspiration.

Simplified Conceptual Model

Continuum methods of ground-water-flow analysis, including modeling, assume laminar flow through a medium with primary porosity and permeability (porous media). The geologic units in Chester Valley have low primary porosity or permeability; ground water flows mainly through secondary openings. However, to permit analysis by continuum methods, the geologic units in Chester Valley are assumed to approximate porous media because of the regional scale of analysis. Secondary-opening density is sufficiently great at a regional scale to permit the use of a porous-media model. A block of aquifer material is assumed to have the equivalent properties of the same-size block of porous media. Water-table maps of Chester Valley (Sloto, 1987, pl. 2; Wood, 1984; Wood, 1985) indicate that ground-water flow is continuous on a regional scale.

In order to analyze ground-water flow with a digital model, a simplified conceptual model of the complex physical system was developed. The conceptual model includes the following assumptions:

(1) The geologic units in Chester Valley act together as a single heterogeneous water-table aquifer.

(2) Hydraulic properties of each geologic unit differ spatially, but are averaged for model simulation. The average is considered representative of the geologic unit. An average hydraulic conductivity is specified for each geologic unit.

(3) The ground-water system is recharged by precipitation. Recharge to noncarbonate rocks is half of recharge to carbonate rocks because the noncarbonate rocks have much steeper slopes, are much less permeable, and accept recharge at a lower rate than carbonate rocks. Recharge is distributed evenly over each rock type.

(4) Streams are in direct hydraulic contact with the aquifer.

(5) The lower limit of ground-water flow is 600 ft below land surface based on analysis of water-bearing zones. Ground-water flow below 600 ft is considered negligible.

(6) Secondary permeability is a function of topography and generally increases down gradient from hilltop to valley bottom (Gerhart and Lazorchick, 1984, p. 11-13). Hydraulic conductivity is a function of secondary permeability; therefore, hydraulic conductivity in the same unit is greater in valley-bottom areas than in hilltop areas. The effect of topography is approximated by adjusting hydraulic conductivity for the topographic position of each cell.

Discretization

Because ground-water and surface-water divides do not coincide in Chester Valley, the Valley Creek basin cannot be modeled separately. The area between the Brandywine Creek and the Schuylkill River was modeled to include the natural hydrologic boundaries of the ground-water-flow system (pl. 2). The modeled area was discretized into a rectangular grid of 16 rows and 52 columns containing 572 active cells. The cell location notation used in this report is (row, column). For example, (6, 25) denotes a cell in the 6th row and 25th column of the model grid. Cell dimensions range from 1,000 to 4,000 ft on a side. The total area covered by active cells is 66.4 mi^2. The Valley Creek basin is represented by 304 active cells and has an area of 23.3 mi^2. The West Valley Creek basin, which drains to the Brandywine Creek west of the Valley Creek basin, has a modeled area of 16.2 mi^2. The basins east of the Valley Creek basin, which drain to the Schuylkill River, have a modeled area of 26.9 mi^2. The grid was oriented parallel to the major direction of ground-water flow, which generally is parallel to geologic contacts. Physical and hydraulic properties are averaged over the area represented by each cell and are assigned to a node in the center of the cell.

Boundary Conditions

The modeled area is defined by natural hydrologic boundaries (pl. 2). Three types of boundary conditions are used: (1) specified flux; (2) head dependent; and (3) specified head. On the northwestern and southern sides of the modeled area, the surface-water divide in the noncarbonate rocks, which coincides with the ground-water divide, is a specified-flux boundary with a specified flux of zero (no-flow boundary). On the western side, the boundary is the Brandywine Creek, which is represented by a head-dependent boundary (stream cells). The Brandywine Creek is a regional sink, and ground-water flow is to the Brandywine from both the east and west (Wood, 1984). On the northeastern and eastern sides of the modeled area, the Schuylkill River is simulated as a specified-head (constant-head) boundary. Here, the Schuylkill is as much as 1,000 ft wide. Ground water discharges to the Schuylkill from most of Chester Valley. The Schuylkill is also a source of induced recharge to the ground-water-flow system in the part of Chester Valley where a cone of depression caused by quarry pumping reaches the Schuylkill River. The model lower boundary is a specified-flux (no-flow) boundary 600 ft below land surface. The model upper boundary is represented by the water-table surface and streams. The water-table is a specified flux boundary; the flux is areal recharge. The model program sets the conductance across the exterior faces of the cells in the first and last rows and columns of the model grid to zero; this produces a no-flow boundary around the exterior cells of the grid (McDonald and Harbaugh, 1984, p. 41).

Streams are simulated as head-dependent boundaries. Leakage to streams (McDonald and Harbaugh, 1984, p. 209) is approximated by

$$Q = k' A (h_r - h_a)/m, \qquad (3)$$

where

Q = leakage,
k' = streambed hydraulic conductivity,
A = cross-sectional area of the streambed,
h_r = stream stage,
h_a = head in the aquifer, and
m = streambed thickness.

The locations of stream cells are shown on plate 2.

Steady-State-Model Calibration

The ground-water-flow model of the Valley Creek basin was calibrated under steady-state conditions using average recharge, evapotranspiration, and pumping rates. The objectives of steady-state calibration, to simulate the average water budget for the Valley Creek basin and regional ground-water flow, were met.

Recharge

The recharge rate was based on the 1983-87 average water budget (table 7). Recharge from precipitation on noncarbonate rocks is assumed to be half that on carbonate rocks because noncarbonate rocks have much steeper slopes, are much less permeable, and accept recharge at a lower rate. Recharge rates of 5.67×10^{-3} ft/d (feet per day) (24.84 in/yr) for carbonate rocks and 2.83×10^{-3} ft/d (12.40 in/yr) for noncarbonate rocks were used; this produced a recharge rate of 4.91×10^{-3} ft/d (21.05 in/yr) for the Valley Creek basin (table 7). Recharge from precipitation is distributed evenly over each rock type.

Evapotranspiration

The evapotranspiration (ET) rate, 5.23×10^{-3} ft/d (22.90 in/yr), was based on the 1983-87 average water budget (table 6). This rate produced a simulated ground-water ET rate of 1.59 in/yr for the Valley Creek basin, which compares favorably with the estimated ground-water ET rate of 2.00 in/yr used in table 8. The ET rate determined by the model depends on the position of the head in the aquifer relative to two given ET reference elevations--ET surface and ET extinction depth (McDonald and Harbaugh, 1984, p. 316). At and above the ET surface, the ET rate is the maximum ET rate. At and below the ET extinction zone, the ET rate is zero. The ET rate varies linearly from the maximum ET rate at the ET surface to zero at the ET extinction elevation (extinction depth). The ET surface was set to the average land-surface elevation for each cell. The ET extinction depth was set to 10 ft.

Pumping Rates

Average pumpage from the modeled area is 22.01 Mgal/d. Pumpage from the Valley Creek basin above the gaging station is 7.14 Mgal/d, which is 32 percent of the pumpage in the modeled area. Pumpage from quarries and the Upper Merion Reservoir, a former quarry, is 16.24 Mgal/d, which is 74 percent of the total pumpage. Pumping rates (table 9) used in model simulations are 1983-87 average rates. The Philadelphia Suburban Water Company supplied pumpage data for their production wells and for the Upper Merion Reservoir (pl. 2). The Upper Merion Reservoir is the abandoned Bridgeport quarry, which is used as a source of supply by the Philadelphia Suburban Water Company. Quarry pumpage for the Cedar Hollow quarry was measured by the Warner Company for 1984-87 and estimated for 1983. Quarry pumpage for the Bradford Hills, Catanach, and McCoy quarries was estimated on the basis of current-meter measurements of inflow to quarry sumps. The Catanach quarry discharges to a sink hole (node 5,24), which recharges the ground-water system; this recharge is simulated as a recharge well. Pumpage by the Foote Mineral Company and the Sheraton Hotel was estimated.

Table 9.--Pumping rates used for simulations

[Mgal/d, million gallons per day]

Node Row	Node Column	Pumping rate (Mgal/d)	Well owner
11	5	0.18	Uwchlan Township Municipal Authority, CH-263, 264
14	6	.42	Bradford Hills quarry
11	7	.66	Uwchlan Township Municipal Authority, CH-2669, 2770
10	8	.45	Uwchlan Township Municipal Authority, CH-1228, 1229
11	10	.12	West Whiteland Municipal Authority, CH-1315, 1316
12	13	.10	Foote Mineral Company, CH-241
6	23	2.64	Catanach quarry
5	24	-1.32	Catanach quarry recharge
7	27	3.92	Cedar Hollow quarry
9	17	.18	Philadelphia Suburban Water Company CH-2199
11	18	.40	Philadelphia Suburban Water Company CH-207
8	40	.96	Philadelphia Suburban Water Company CH-209
4	44	1.23	Philadelphia Suburban Water Company MG-882
5	44	.10	Sheraton Hotel
11	45	.54	Philadelphia Suburban Water Company MG-781
12	46	.85	Philadelphia Suburban Water Company MG-881
10	51	2.91	McCoy quarry
9	49	6.35	Philadelphia Suburban Water Company Upper Merion Reservoir

Aquifer Characteristics

Aquifer characteristics required by the model include altitude of the top and bottom of the aquifer, aquifer thickness, aquifer horizontal hydraulic conductivity, and streambed vertical hydraulic conductance.

The top of the aquifer is the land surface. Average land-surface elevation for each cell was determined from 7.5-minute topographic maps.

Aquifer thickness was assumed to be 600 ft for model simulations on the basis of analysis of water-bearing zones and geophysical logging. Few water-bearing zones are penetrated below 500 ft (table 2), and ground-water circulation below 600 ft is considered negligible. The elevation of the bottom of the aquifer for each cell was set at 600 ft below the average land-surface elevation for that cell.

Each geologic unit was assigned a different hydraulic conductivity. If a cell contained two geologic units of nearly equal area, the mean hydraulic conductivity of the two units was used; otherwise, the hydraulic conductivity of the predominant unit was assigned to the cell. The quartzites and leucocratic gneiss were considered as one geologic unit.

Aquifer hydraulic conductivity was estimated from specific-capacity and aquifer-test data. Transmissivity was estimated from specific capacity using Theis's method for a water-table aquifer (Theis, 1963, p. 332-336):

$$T' = 0.134 \frac{Q}{s} (K - 264 \log_{10} 5\ Sy + 264 \log_{10} t), \quad (4)$$

and

$$K = -66 - 264 \log_{10} (3.74\ r^2 \times 10^{-6}) \quad (5)$$

where

T' = estimated transmissivity (ft^2/d),
Q = pumping rate (gal/min),
s = drawdown (ft),
Sy = specific yield,
t = duration of pumping (d),
K = a constant, and
r = well radius (ft).

Because the wells used for analysis have small diameters and tap consolidated rock, r was set equal to well radius (Theis, 1963, p. 335). A specific yield of 0.08, the average used to calculate change in ground-water storage in the water budgets (table 6), was assumed. Hydraulic conductivity was calculated by dividing transmissivity by the depth of uncased borehole. The initial estimates of hydraulic conductivity shown in table 10 are averages for each geologic unit.

The initial hydraulic conductivity of each geologic unit was adjusted on the basis of the simulated water budget of the Valley Creek basin and regional ground-water flow. The hydraulic conductivity of individual cells was not adjusted. Final calibrated hydraulic conductivities are given in table 10.

Although the Triassic Stockton Formation is not present in the Valley Creek basin, it is present in Montgomery County in the northeastern part of the modeled area. Specific-capacity data for wells in Montgomery County were used to calculate the hydraulic conductivity in table 10.

The hydraulic conductivity of the Conestoga Limestone was adjusted downward based on model simulations. The hydraulic conductivity calculated from specific-capacity data is probably too high and influenced by two wells with very large specific capacities. Comparison of data for well yield (table 3) and specific capacity (table 4) indicates that the hydraulic properties of the Conestoga Limestone are probably much closer to those of the Elbrook Formation than to those of the Ledger Dolomite.

The Kinzers Formation and Vintage Dolomite are thin in Chester County. Wells drilled into these formations derive water from them and from underlying formations as well. The calculated hydraulic conductivities presented in table 10 are probably not representative of the Kinzers Formation and Vintage Dolomite as they also may reflect the hydraulic properties of underlying units.

Table 10.--Aquifer hydraulic conductivity used for steady-state simulation

[Initial estimates of hydraulic conductivity based on transmissivity calculated from specific-capacity data by the method of Theis (1963, p. 332). Final estimates of hydraulic conductivity are values used in the model after adjusting the initial estimates.]

Geologic unit	Hydraulic conductivity (feet per day) Initial	Final
Stockton Formation	1.9	8.0
Conestoga Limestone	32	10
Elbrook Formation	6.7	7.0
Ledger Dolomite	44	45
Kinzers Formation and Vintage Dolomite	53	6.0
Antietam Quartzite, Harpers Phyllite, Chickies Quartzite, and leucocratic gneiss	2.6	3.0
Octoraro Phyllite	5.2	5.0

Calculation of hydraulic conductivity using specific capacity data from aquifer tests gave values similar to those in table 10. The hydraulic conductivity of the Ledger Dolomite calculated from specific-capacity data was 58 ft/d, the hydraulic conductivity of the Elbrook Formation was 6.2 ft/d, and hydraulic conductivity of the Conestoga Limestone was 20 ft/d.

Hydraulic conductivity of consolidated-rock aquifers is affected by topography (Poth, 1968, p. 21-23). Generally, hydraulic conductivity is less on hilltops and greater in valley-bottom areas. On hilltops, fractures are widely spaced, and the unit is more resistant to weathering. In valley-bottom areas, fractures are closely spaced, and the unit is less resistant to weathering.

Cells were classified as hilltop, hillside, or valley bottom from their topographic position on 7.5-minute topographic maps. The hydraulic conductivity for each cell was multiplied by a topography-adjustment coefficient (table 11) using the method of Gerhart and Lazorchick (1984, p. 27) to adjust hydraulic conductivity for topographic position. The distribution of final hydraulic conductivity is shown on plate 3.

The water-table aquifer system is anisotropic. The preferential direction of ground-water flow in Chester Valley is parallel to bedding along strike. The strike of the geologic units in Chester Valley is northeast. The model grid is oriented parallel to strike. A column-to-row anisotropy multiplication factor of 0.2, based on simulated regional ground-water flow directions, was used, resulting in hydraulic conductivity along strike being five times greater than hydraulic conductivity across strike.

Table 11.--Coefficients for adjusting hydraulic conductivity for topographic effect

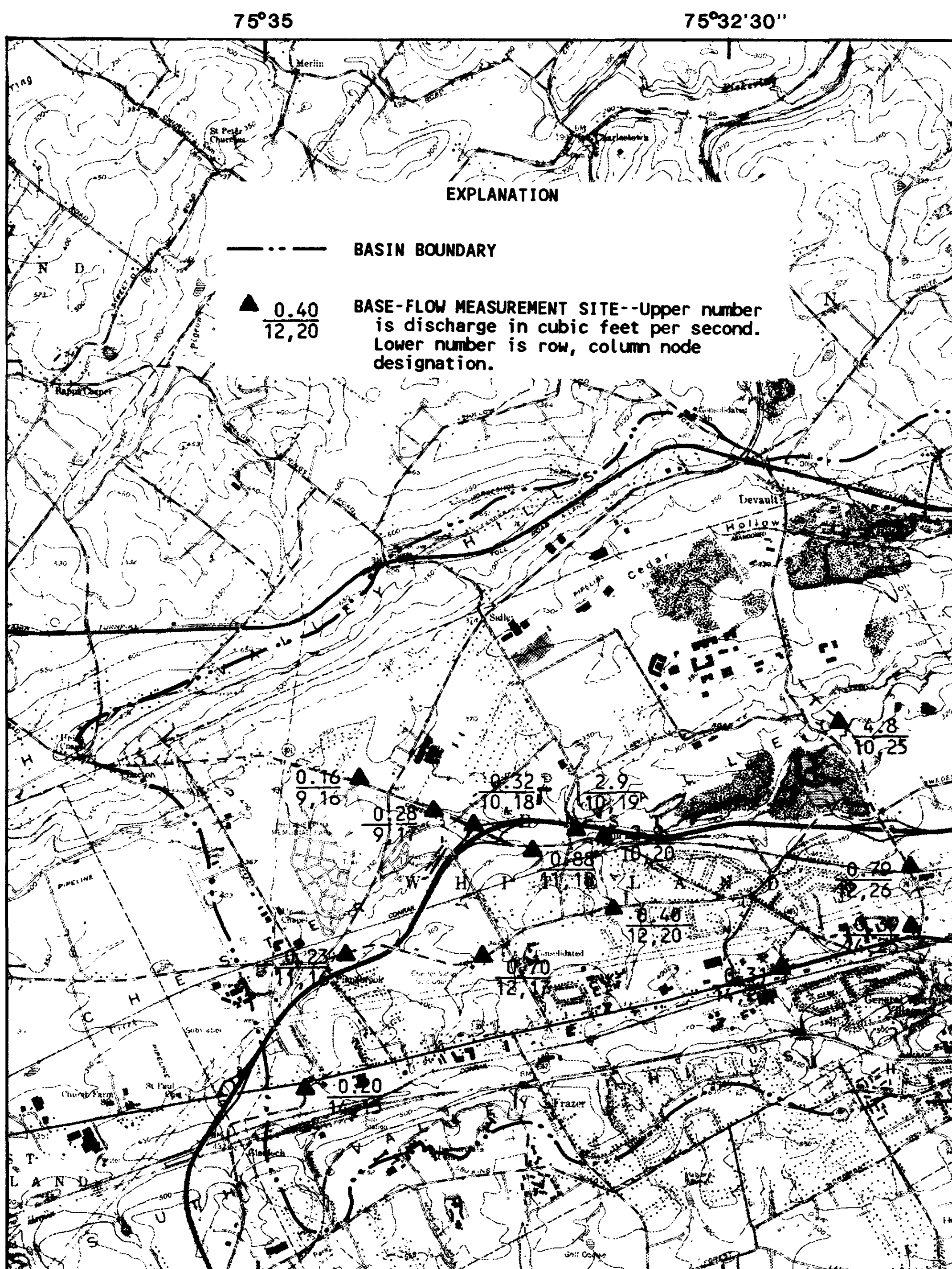
	Topography-adjustment coefficient		
	Hilltop	Hillside	Valley bottom
This study	0.5	1.0	1.5
Gerhart and Lazorchick (1984, p. 29):			
Stockton Formation	0.1	1.0	1.5
Conestoga Limestone	.4	1.0	3.0
Ledger Dolomite	.3	1.0	1.2
Kinzers Formation and Vintage Dolomite	.4	1.0	3.0
Antietam Quartzite, Harpers Phyllite, and Chickies Quartzite	.1	1.0	2.3

The ground-water and surface-water systems in the Valley Creek basin are well connected, and water moves freely between the two systems. The direction and rate of water movement between the two systems are controlled by the hydraulic conductivity of the streambed material, thickness of streambed material, and the difference between head in the aquifer and stream stage. Streambed material differs greatly from place to place and consists of gravel, sand, clay, or fractured bedrock. The hydraulic conductivity of the streambed materials is highly variable and is not known. Therefore, streambed hydraulic conductivity was calibrated by adjusting the hydraulic conductance of each stream node (McDonald and Harbaugh, 1984, p. 214). Streambed thickness was assumed to be 1 ft. The simulated and measured base flow of Valley Creek is given in table 12. Base-flow measurements were made October 11-18, 1984. Locations of base-flow measurements are shown on figure 16. Simulated baseflow at each stream node in table 12 includes the ground-water discharge at that node plus the ground-water discharge at all upstream stream nodes. Simulated and observed base flow are compared in figure 17. Average simulated base flow is 3 percent greater than average observed base flow.

Table 12.--Observed and simulated base flow of Valley Creek

[Measurement sites shown on fig. 16; base flow measured October 11-18, 1984; ft^3/s, cubic feet per second]

Node	Observed base flow (ft^3/s)	Simulated base flow (ft^3/s)
14,15	0.20	0.19
11,16	.23	.25
12,17	.70	.84
11,18	.88	1.2
9,16	.16	.21
9,17	.28	.31
10,18	.32	.32
10,19	2.9	3.2
10,20	2.9	3.2
12,20	.40	.36
10,25	4.8	4.7
9,28	5.2	5.3
7,28	8.4	6.1
8,30	14	12
10,32	14	13
9,33	.04	.05
10,33	14	13
10,35	13	13
14,22	.31	.18
14,25	.39	.35
12,26	.79	.78
13,29	.33	.32
14,29	.09	.10
12,29	1.5	1.5
14,31	.04	.04
12,32	2.7	2.5
14,34	.17	.16
13,34	.38	.35
12,34	3.2	3.0
14,35	.52	.52
13,35	.46	.52
12,35	4.6	3.5
11,35	3.3	3.5
8,36	.05	.05
9,37	26	23
10,39	.05	.03
8,38	20	23
6,39	19	23



Base from U.S. Geological Survey Malvern 1:24,000, 1973 and Valley Forge 1:24,000, 1981.

Figure 16.--Base-flow measurement sites on Valley Creek. Base flow measured October 11-18, 1984.

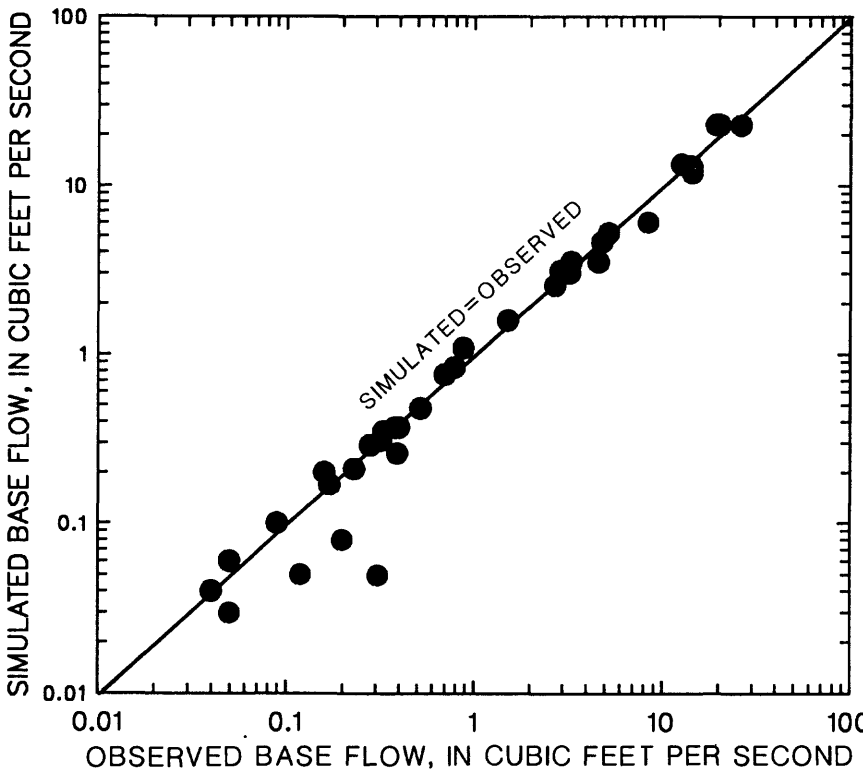


Figure 17.--Relation between observed and simulated base flow of Valley Creek.

Simulated Water-Table Surface

The simulated water-table surface in the Valley Creek basin (fig. 18) was compared to the observed water-table surface (fig. 19) for October 1983. The observed water-table surface is from Sloto (1987, pl. 2) and was mapped using 20-ft contours. Simulated heads in the carbonate rocks of the Valley Creek basin were compared with observed heads using the root mean square error (RMSE). The RMSE is the square root of the sum of the squared difference between the observed and simulated head divided by the number of cells. The RMSE was 19.54 ft.

Simulated Average Water Budget

The average (1983-87) water budget for the Valley Creek basin was approximated by a steady-state simulation. The simulated water budget is compared to the calculated water budget and calculated recharge rate in table 13. Calculated base flow is based on the 1983-87 average base flow (17.00 in.) minus the estimated 1983-87 average quarry pumpage (3.96 in.) discharged to Valley Creek (see table 7).

Table 13 gives the simulated ground-water underflow beneath the surface-water divides on the eastern and western sides of the Valley Creek basin. Simulated underflow is 0.26 in. into the basin from the west and a net outflow of 1.33 in. out of the basin to the east. The simulated underflow into the basin from the west is less than the estimated underflow in table 13, which was based on Darcy's Law, the October 1983 water-table map of Sloto (1987, pl. 2), and the May-June 1984 water-table map of Wood (1984). The estimated underflow is for one point in time; the simulated underflow is an estimate of the long-term average underflow, rather than the underflow for a particular season or year.

Table 13.--Simulated average water budget for the Valley Creek basin

[in., inches]

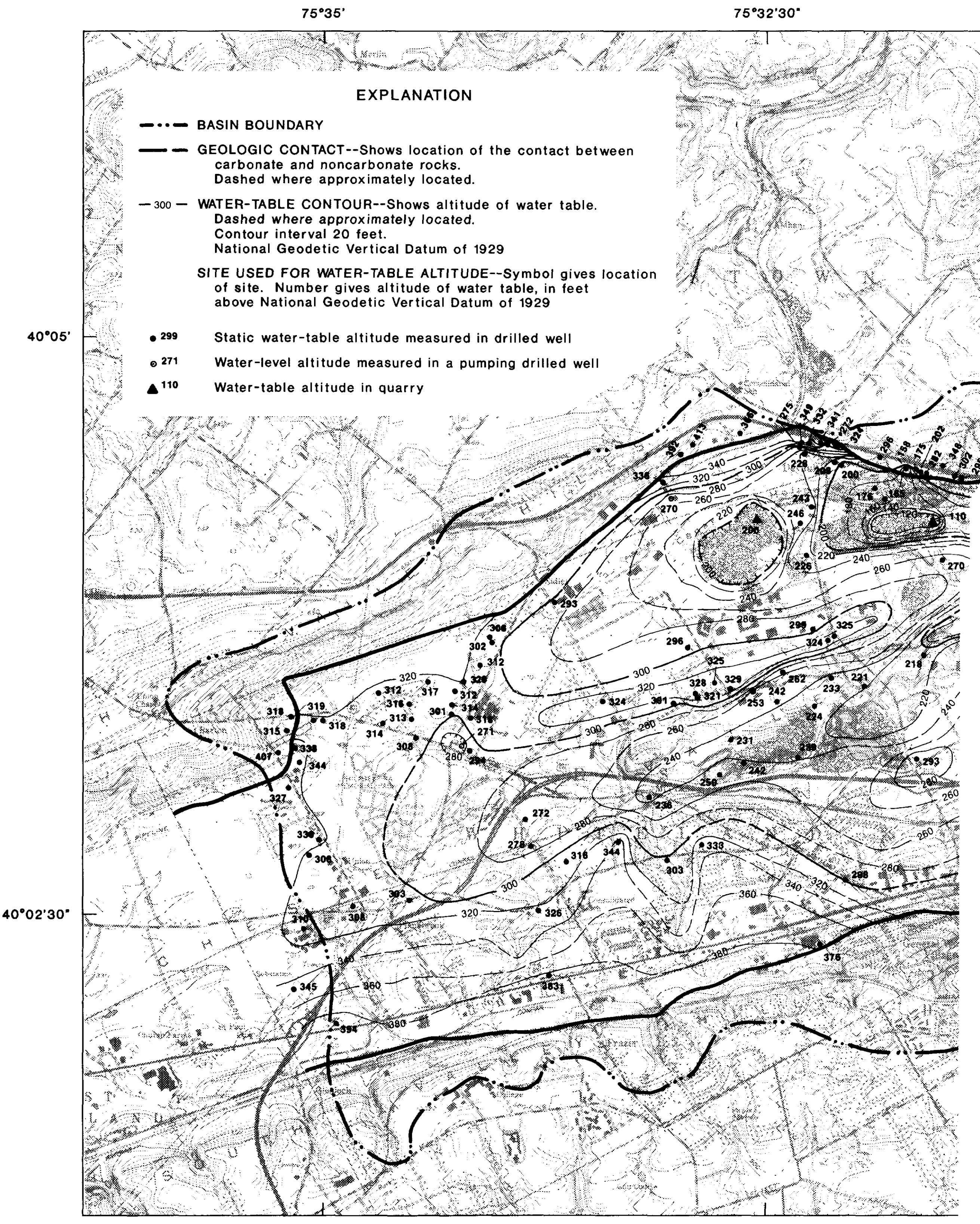
Water-budget component	Water budget: Simulated (in.)	Water budget: Calculated (in.)
Recharge	21.05	21.04
Base flow	13.06	13.04
Evapotranspiration	1.59	2.00
Ground-water withdrawals	[1]1.92	2.13
Cedar Hollow quarry pumpage	3.96	3.96
Underflow into basin:		
from east	.71	[2].86
from west	.26	[3].76
Underflow out of basin:		
to east	2.04	[2]1.78

[1]Does not include small commercial, industrial, and domestic pumpage.
[2]Estimated by Darcy's Law from October 1983 water-table map (Sloto, 1987, pl. 2).
[3]Estimated by Darcy's Law from October 1983 (Sloto, 1987, pl. 2) and May-June 1984 (Wood, 1984) water-table maps.

Sensitivity Analysis

A sensitivity analysis of model variables involves varying the value of a single model variable while holding the others constant. The effect of varying the value of a particular model variable on the simulated water budget was determined by varying the value of the variable being tested over a reasonable range, while the values of the rest of the variables remained fixed. Then, any changes in the simulated water budget are caused only by the change in the value of the variable being tested. If the change in the value of a variable causes a large change in the simulated water budget, the model is said to be sensitive to that variable. Conversely, if changes in the simulated water budget are slight, the model is insensitive to that variable.

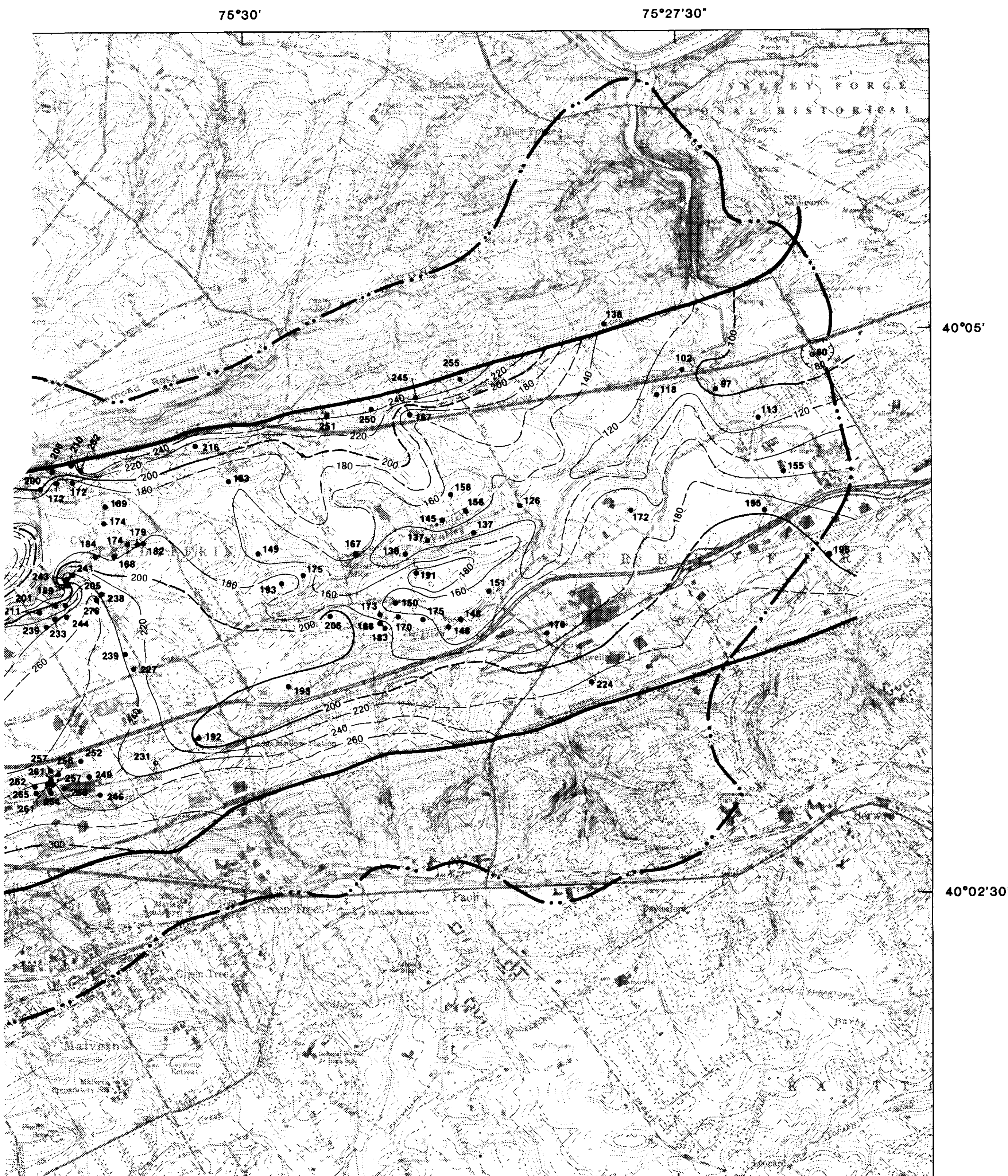
Many sensitivity analyses were made during model calibration. The variables tested for sensitivity are aquifer hydraulic conductivity, streambed hydraulic conductance, recharge rate, ground-water evapotranspiration rate, aquifer thickness, and aquifer anisotropy. The results of a final sensitivity analysis, made when model calibration was completed, is presented below. Model variables were varied over a range from half to double the calibrated value. The effects of changing the value of a model variable on base flow and RMSE are shown on figures 20 and 21, respectively. The effects of changing aquifer anisotropy on base flow and RMSE are shown on figures 22 and 23, respectively. In figures 20 and 21, slope is directly proportional to sensitivity. A low slope indicates low sensitivity (or insensitivity); a high slope indicates high sensitivity.



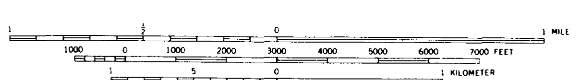
Base from U.S. Geological Survey Malvern 1:24,000, 1973 and Valley Forge 1:24,000, 1981.

MN GN
10°
178 MILS
0°02'
1 MIL
UTM GRID AND 1966 MAGNETIC NORTH DECLINATION AT CENTER OF SHEET

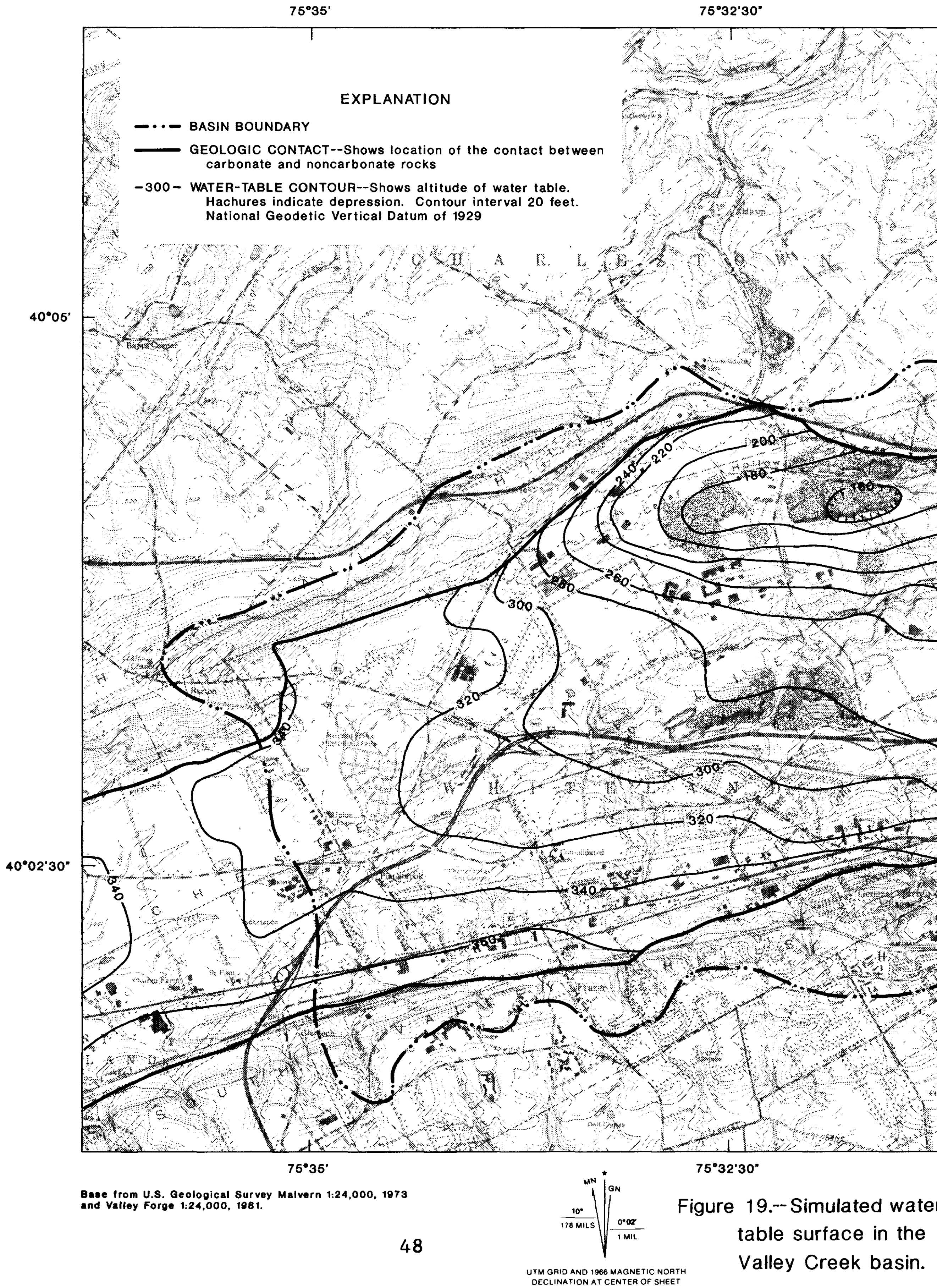
Figure 18.--Observed water-table surface in the Valley Creek basin.



Hydrology by R.A. Sloto, 1983

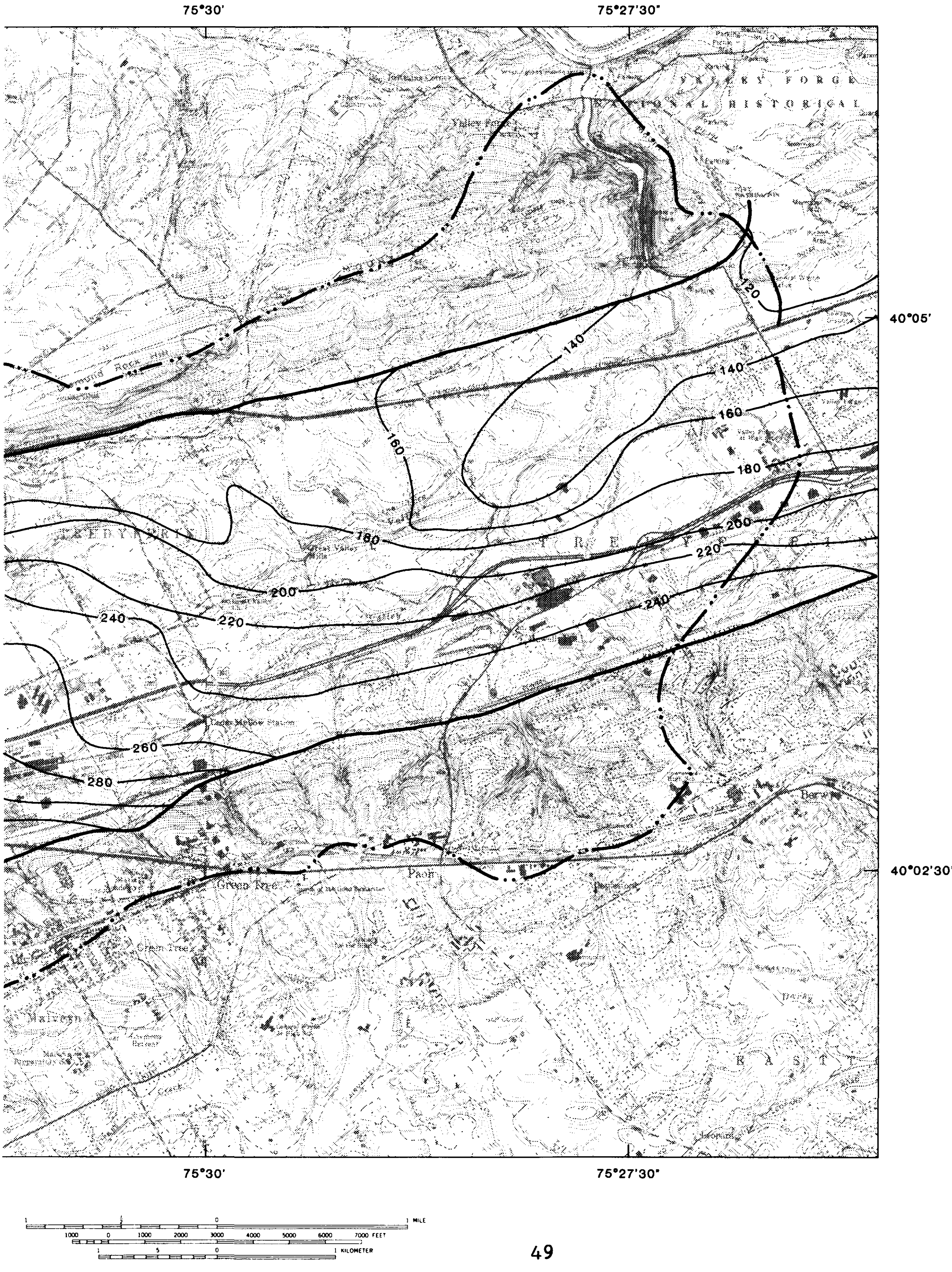


CONTOUR INTERVAL 10 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929



Base from U.S. Geological Survey Malvern 1:24,000, 1973 and Valley Forge 1:24,000, 1981.

Figure 19.--Simulated water-table surface in the Valley Creek basin.



CONTOUR INTERVAL 10 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

The model was found to be most sensitive to the recharge rate (figs. 20 and 21). When the recharge rate was varied from half to double the 1983-87 average rate (10.53 to 42.10 in.), the simulated base flow ranged from 5.58 to 26.79 in. (57 percent less to 105 percent greater than the 1983-87 average base flow). The RMSE ranged from 19.54 to 38.21 ft. The model was less sensitive to streambed hydraulic conductance and aquifer anisotropy. When the streambed hydraulic conductance was varied from half to double the calibrated values, the simulated base flow ranged from 11.43 to 14.45 in. (12.5 percent less to 10.6 percent greater than the 1983-87 average base flow). The RMSE ranged from 18.66 to 22.25 ft. When aquifer anisotropy was varied from 0.1 to 5 (figs. 22 and 23), the simulated base flow ranged from 11.87 to 13.20 in. (9.1 percent less to 1.1 percent greater than the 1983-87 average base flow). The RMSE ranged from 18.46 to 34.69 ft. The model was least sensitive to the ground-water evapotranspiration rate, aquifer thickness, and aquifer hydraulic conductivity. When the ground-water evapotranspiration rate was varied from half to double the 1983-87 average rate (11.45 to 45.80 in.), the simulated base flow ranged from 12.21 to 13.72 in. (5.1 percent less to 6.5 percent greater than the 1983-87 average base flow). The RMSE ranged from 18.71 to 20.14 ft. When aquifer thickness was decreased by half to 300 ft, several nodes around the active quarries became dewatered because of the the low saturated thickness of the aquifer. When aquifer thickness was varied from 330 to 1,200 ft (55 percent less to 100 percent greater than the calibrated value), the simulated base flow ranged from 12.30 to 13.11 in. (5.8 percent less to 0.4 percent greater than the 1983-87 average base flow). The RMSE ranged from 18.44 to 30.49 ft. When aquifer hydraulic conductivity was varied from half to double the calibrated values, the simulated base flow ranged from 12.56 to 13.08 in. (3.8 percent less to 0.2 percent greater than the 1983-87 average base flow). The RMSE ranged from 18.38 to 26.89 ft.

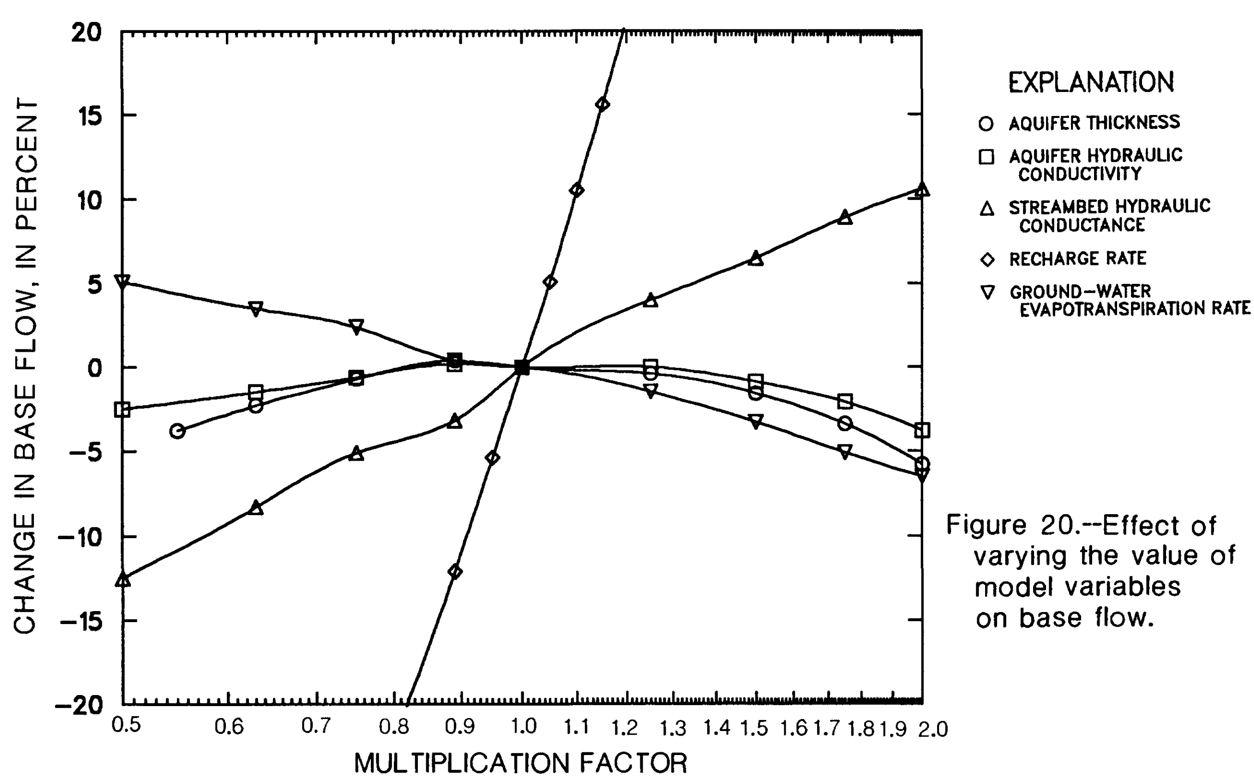


Figure 20.--Effect of varying the value of model variables on base flow.

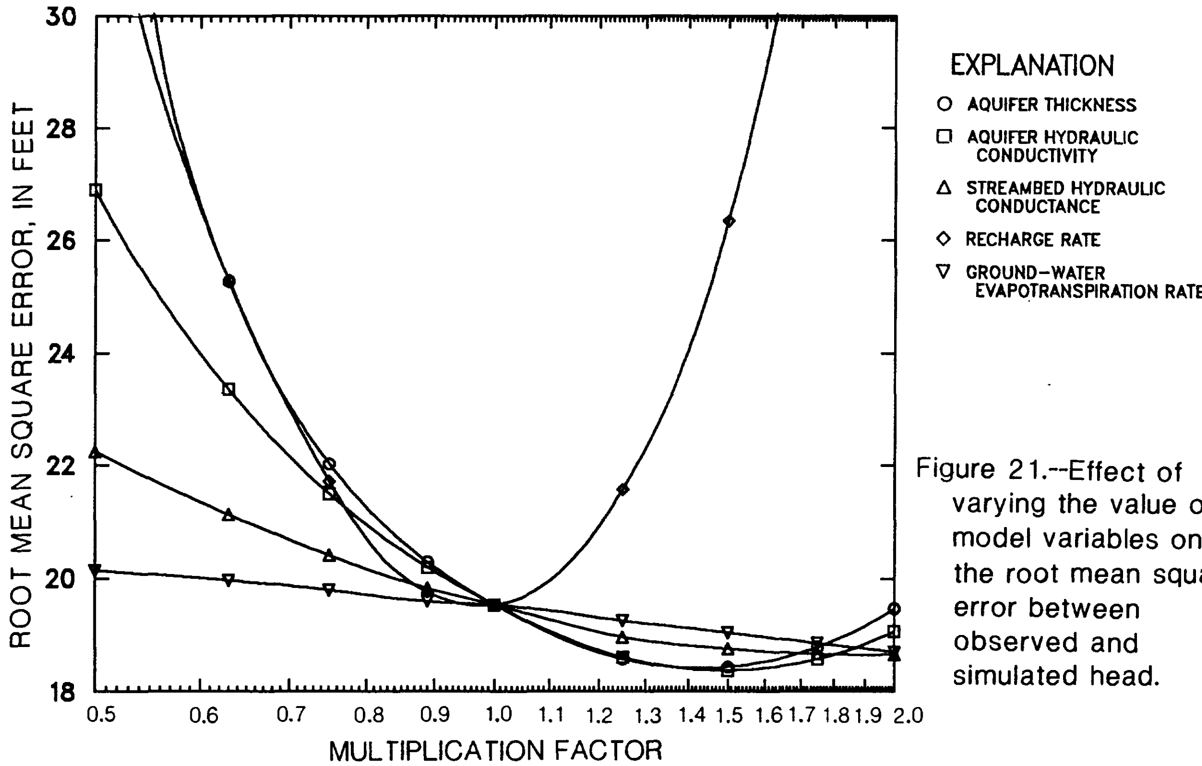


Figure 21.--Effect of varying the value of model variables on the root mean square error between observed and simulated head.

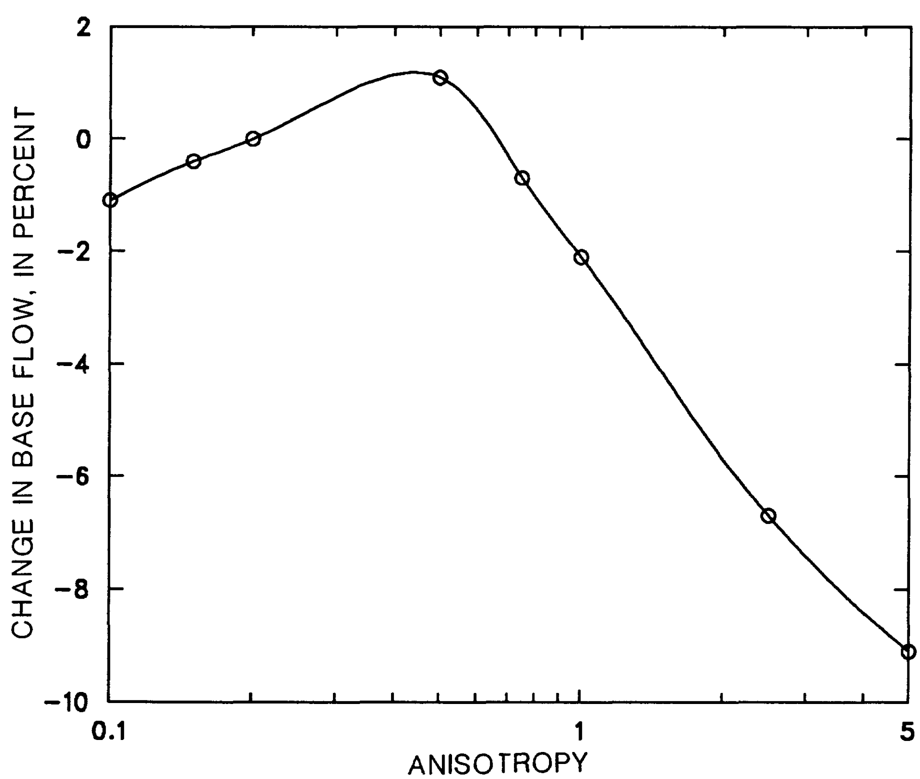


Figure 22.--Effect of varying aquifer anisotropy on base flow.

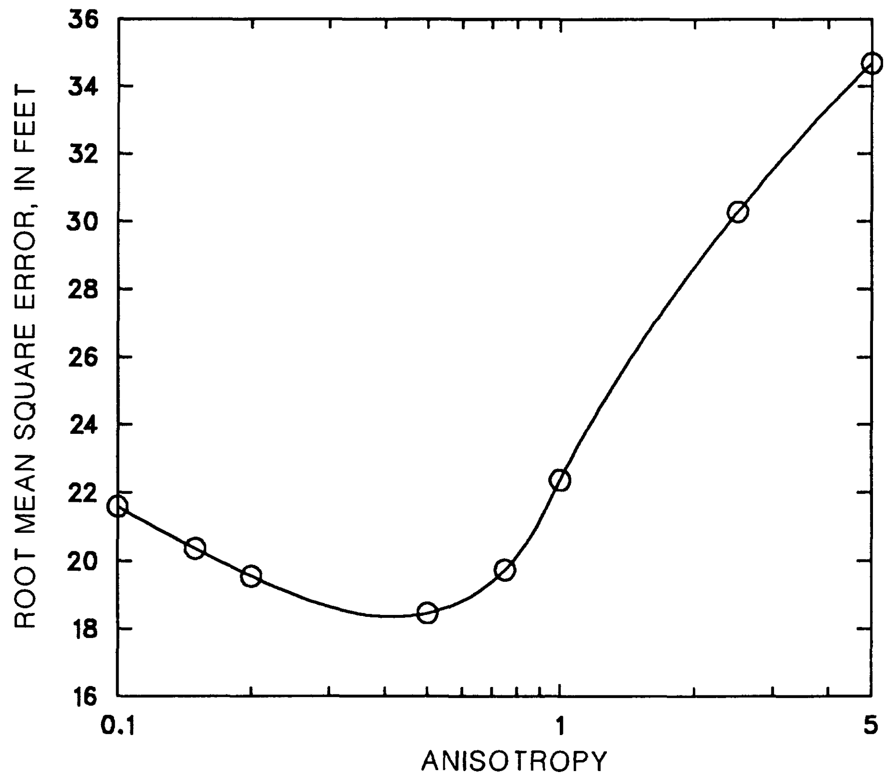


Figure 23.--Effect of varying aquifer anisotropy on the root mean square error between observed and simulated head.

Reliability of Model Simulations

The ground-water-flow model is considered to be calibrated for the Valley Creek basin under average conditions. It is useful for simulating the effects of stresses in the basin on the average water budget. It cannot provide estimates of site-specific effects such as head or drawdown at a particular well site or stream infiltration at a particular stream site.

Simulated Effects of Increased Ground-Water Development

The effects of increased ground-water development on base flow and underflow into and out of the Valley Creek basin were simulated by locating a hypothetical well field in different parts of the basin. Steady-state simulations were used to represent equilibrium conditions, which would be the maximum expected long-term effect.

Starting heads used for the simulations (fig. 18) were based on results of the final steady-state model calibration. In addition to the hypothetical pumping, average ground-water pumping rates used for steady-state calibration (table 9) were used for these simulations.

Increased ground-water development in the Valley Creek basin was simulated as a hypothetical 4 Mgal/d well field; this is equal to an additional 4.04 in. of ground-water withdrawal from the basin annually and is one-third the 1983-87 average base flow of Valley Creek. The well field was simulated as 2 Mgal/d of pumpage from each of two nodes. The well field was simulated at five locations in the Valley Creek basin (fig. 24): well field 1

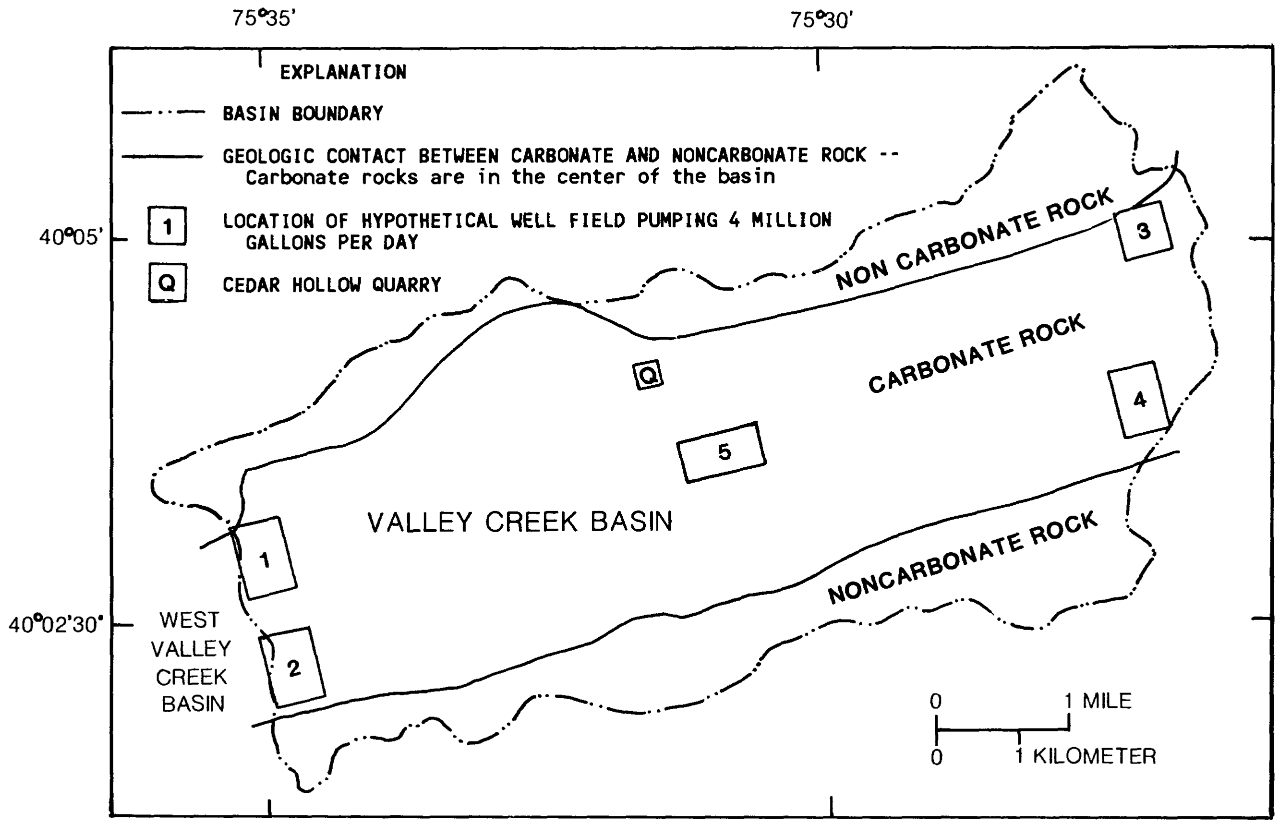


Figure 24.--Locations of hypothetical 4 million gallon per day well fields to simulate increased ground-water development.

is located near the northwestern surface-water divide, well field 2 is located near the southwestern surface-water divide, well field 3 is located near the northeastern surface-water divide, well field 4 is located near the southeastern surface-water divide, and well field 5 is located near the center of the basin. Changes in base flow and underflow caused by locating a hypothetical 4 Mgal/d well field in different parts of the Valley Creek basin are given in table 14. The changes reflect the source of water to wells, such as an increase in underflow into the basin, a decrease in underflow out of the basin, or a reduction of base flow. Additional water to wells is provided by a decrease in ground-water evapotranspiration and dewatering in the vicinity of pumped wells.

Locating the well field near the western surface-water divide would induce additional underflow into the Valley Creek basin from the West Valley Creek basin. Well field 1 (fig. 24), located near the northwestern surface-water divide, would increase underflow into the Valley Creek basin by 1.41 in/yr. The base flow of Valley Creek would be reduced by 1.88 in/yr or 14 percent. Well field 2 (fig. 24), located near the southwestern surface-water divide, would increase underflow into the Valley Creek basin by 0.97 in/yr. Base flow would be reduced by 1.08 in/yr or 8 percent.

Table 14.--Simulated changes in base flow and underflow for the Valley Creek basin caused by increased ground-water development

[Well field locations are shown on figure 24; in., inches]

Well field location	Water-budget component (in.)			
	Base flow	Underflow into basin from the: east	Underflow into basin from the: west	Underflow out of basin to the east
No additional well field	13.06	0.71	0.26	2.04
Northwest surface-water divide	11.18	.71	1.67	2.02
Southwest surface-water divide	11.98	.71	1.23	2.04
Northeast surface-water divide	12.35	1.50	.26	.26
Southeast surface-water divide	11.47	2.11	.26	2.04
Center of the basin	10.09	.71	.38	2.02

Well field 3 (fig. 24), located near the northeastern surface-water divide, would change the underflow on the eastern side of the basin from a net outflow of 1.33 in/yr to a net inflow of 1.24 in/yr. Underflow into the Valley Creek basin from the east would increase by 0.79 in/yr and underflow out of the basin to the east would decrease by 1.78 in/yr. Base flow would be reduced by 0.71 in/yr or 5 percent.

Well field 4 (fig. 24), located near the southeastern surface-water divide, would change the underflow on the eastern side of the basin from a net outflow of 1.33 in/yr to a net inflow of 0.07 in/yr. The underflow into the Valley Creek basin from the east would increase by 1.40 in/yr, and the underflow out of the basin to the east would not change. Base flow would be reduced by 1.59 in/yr or 12 percent.

Well field 5 (fig. 24), located near the center of the Valley Creek basin, would have more of an effect on base flow and less of an effect on underflow than would locating a well field near the surface-water divides. Little additional ground water would be induced to flow into the Valley Creek basin from the adjoining basins. The source of water for additional withdrawal would mainly be intercepted base flow; additional water would be provided by reduced ground-water evapotranspiration and dewatering in the vicinity of pumping wells. The well field would reduce the base flow of Valley Creek by 2.97 in/yr or 23 percent. Underflow into the Valley Creek basin from the west would increase by 0.12 in/yr. Underflow out of the basin to the east would decrease by 0.02 in/yr, and underflow into the basin from the east would remain the same.

The model simulations demonstrate the difficulty of ground-water-resource planning in carbonate-rock terranes. Ground-water-resource planning in Chester County is based on a surface-water-basin approach with ground-water withdrawals causing a one-to-one reduction in base flow (Reith and others, 1979; Chester County Planning Commission, 1982). This approach has major drawbacks in eastern Chester Valley because (1) surface-water and ground-water divides do not coincide and (2) ground-water development, especially near surface-water divides, can cause ground-water divides to shift and induce ground-water underflow from adjacent basins. Large-scale ground-water pumping may not produce expected reductions of base flow in the basin because of shifts in the ground-water divide; however, such shifts may reduce base flow in the adjacent surface-water basin.

Simulated Effects of Increased Quarry Dewatering

As the quarries in Chester Valley expand, ground-water seepage into the quarries probably will increase. Increased seepage was simulated as increased pumping. The effect of increased pumping on the base flow of Valley Creek was simulated by increasing pumping rates (table 9) at the node representing the Cedar Hollow quarry (fig. 24). Steady-state simulations were used to represent equilibrium conditions, which would be the maximum expected long-term effect.

Starting heads used for the simulations (fig. 18) were based on results of the final steady-state-model calibration. Average ground-water pumping rates used for steady-state calibration (table 9) also were used for these simulations.

Increases in ground-water pumping from the Cedar Hollow quarry of 25, 50, 75, and 100 percent were simulated. A 100 percent increase in pumping rate, from 3.93 to 7.86 Mgal/d, would reduce the natural base flow of Valley Creek by 2.33 in. or 18 percent (table 15). Water pumped from the quarry is discharged to Valley Creek, increasing the base flow (natural base flow plus quarry discharge) at the gaging station. Assuming no loss in the quarry discharge to Valley Creek, the base flow at the gaging station would increase by about 10 percent (table 15).

Increasing ground-water pumpage by the same quantities at the Catanach quarry would have nearly the same effect on base flow (table 15) as increased pumpage at the Cedar Hollow quarry.

Table 15.--Simulated changes in the base flow of Valley Creek caused by increased quarry dewatering

[Mgal/d, million gallons per day; in., inches]

Cedar Hollow quarry pumping rate			Base flow of Valley Creek without quarry discharge		Base flow of Valley Creek plus quarry discharge	
(Mgal/d)	(in.)	(percent increase)	(in.)	(percent decrease)	(in.)	(percent increase)
3.93	3.96	0	13.06	0	17.02	0
4.91	4.95	25	12.46	4.6	17.41	2.3
5.90	5.94	50	11.94	8.6	17.88	5.1
6.88	6.93	75	11.32	13	18.25	7.2
7.86	7.92	100	10.73	18	18.65	9.6

SUMMARY

Valley Creek, a tributary to the Schuylkill River, drains 22.6 mi^2 in eastern Chester County. Streamflow from 20.8 mi^2 is measured at a streamflow-gaging station on Valley Creek at the Pennsylvania Turnpike Bridge near Valley Forge. The Valley Creek basin is mostly underlain by easily eroded Cambrian and Ordovician limestone and dolomite, which form Chester Valley. Resistant quartzites form the North Valley Hills to the north, and resistant phyllite forms the South Valley Hills to the south.

Ground water flows through a network of interconnected secondary openings--fractures, joints, faults, parting planes, and bedding planes; primary porosity is virtually nonexistent. Some of these openings are enlarged by solution. Most openings enlarged by solution are only a fraction of an inch wide, but they are capable of transmitting large quantities of water.

Secondary porosity and permeability have large spatial variation in carbonate rock; therefore, the yield and specific capacity of wells are highly variable. Well yield depends on the number and size of openings below the water table. The frequency of water-bearing zones decreases with depth; fifty percent of water-bearing zones are within 100 ft of the land surface, and 81 percent are within 200 ft of the surface.

The carbonate rocks form a complex, heterogeneous water-table aquifer. The water table fluctuates in response to recharge from precipitation, discharge from pumped wells and quarries, evapotranspiration, and discharge to streams. Although most of the ground-water system is under water-table conditions, the water is confined locally.

Most ground-water flow in the Valley Creek basin is local with discharge to nearby streams. Regional ground-water flow is to the northeast to the Schuylkill River. In most areas, the ground-water discharge to streams comprises the base-flow component of streamflow. Ground-water discharge comprised 73 to 84 percent of the flow of Valley Creek measured at the gaging station during 1983-87. The average ground-water discharge was 76 percent of streamflow. Ground-water discharge includes both natural discharge and quarry pumpage for dewatering. For 1983-87, the base flow of Valley Creek ranged from 11.56 to 21.87 in.; the average was 17.00 in.

For 1983-87, discharge from the Cedar Hollow quarry comprised 21 to 26 percent of the base flow of Valley Creek; the average was 23 percent. The discharge from the Cedar Hollow quarry sustains a greater-than-natural base flow in Valley Creek. The average natural base flow of Valley Creek would be 8 percent lower if the quarry were not operating.

On the western side of the Valley Creek basin, the ground-water divide is about 1/2 mi west of the surface-water divide. An estimated 0.75 Mgal/d of ground water flows from the adjacent West Valley Creek basin eastward into the Valley Creek basin. A ground-water divide does not exist on the eastern side of the Valley Creek basin; the water table slopes gently eastward toward the Schuylkill River. On the northeastern side of the basin, an estimated 1.76 Mgal/d of ground water flows northeasterly out of the basin beneath the surface-water divide to the Schuylkill River. On the southeastern side of the basin, an estimated 0.85 Mgal/d of ground water flows beneath the surface-water divide into the basin.

Annual water budgets for 1983-87 and an average water budget for those years were calculated for the 20.8-mi^2 part of the basin above the streamflow gaging station. The average water budget approximates long-term or steady-state conditions. Annual precipitation ranged from 40.61 to 56.55 in. and averaged 47.25 in., annual streamflow ranged from 15.55 to 28.57 in. and averaged 22.31 in., and annual evapotranspiration ranged from 18.21 to 24.83 in. and averaged 22.90 in. Annual recharge ranged from 15.89 to 26.84 in. and averaged 21.04 in. For 1983-87, average base flow was 81 percent of average recharge.

Ground-water flow in the Valley Creek basin was simulated using a digital computer model. The basin was modeled as a two-dimensional water-table aquifer. Recharge to, ground-water flow through, and discharge from the rocks of Chester Valley was simulated. Input to the modeled hydrologic system is areally varied recharge and pumpage from quarry dewatering discharged to a sinkhole. Output of the modeled hydrologic system is withdrawal from wells and quarries, ground-water discharge to streams, and ground-water evapotranspiration.

In order to analyze ground-water flow with a digital model, the following assumptions were made: (1) the geologic units in Chester Valley act as a single heterogeneous water-table aquifer; (2) hydraulic properties differ spatially, but are averaged for model simulation; (3) recharge to noncarbonate rocks is half of recharge to carbonate rocks; (4) streams are in direct hydraulic contact with the aquifer; (5) the lower limit of ground-water flow is 600 ft below land surface; and (6) hydraulic conductivity increases down gradient from hilltop to valley bottom.

The 66.4-mi^2 area between the Brandywine Creek and the Schuylkill River was modeled to include the natural hydrologic boundaries of the ground-water-flow system. On the northwestern and southern sides of the modeled area, the surface-water divide in the noncarbonate rocks, which coincides with the ground-water divide, is a no-flow boundary. The western boundary is the Brandywine Creek, which is a regional ground-water sink. On the northeastern and eastern sides of the modeled area, the Schuylkill River is a specified-head boundary. The lower model boundary is a no-flow boundary 600 ft below land surface. The upper model boundary is the water-table surface and streams. The modeled area was discretized into a variably-sized grid of 16 rows and 52 columns.

The ground-water-flow model of the Valley Creek basin was calibrated under steady-state conditions using average recharge rates of 5.67×10^{-3} ft/d (24.84 in/yr) for carbonate rocks and 2.83×10^{-3} ft/d (12.40 in/yr) for noncarbonate rocks; this produced a recharge rate of 4.91×10^{-3} ft/d (21.05 in/yr) for the Valley Creek basin. The average evapotranspiration rate, 5.23×10^{-3} ft/d (22.90 in/yr), produced a simulated ground-water evapotranspiration rate of 1.59 in/yr. Pumping rates used in model simulations are 1983-87 average rates.

Aquifer hydraulic conductivity was estimated from specific-capacity and aquifer-test data. Hydraulic conductivities were averaged for each geologic unit, and each geologic unit was assigned a different hydraulic conductivity. The hydraulic conductivity for each cell was multiplied by a topography adjustment coefficient: 0.5 for hilltop cells, 1.0 for hillside cells, and 1.5 for valley-bottom cells. A column to row anisotropy multiplication factor of 0.2 was used.

Streambed hydraulic conductivity was calibrated by adjusting the hydraulic conductance of each stream node. Average simulated base flow is 3 percent greater than observed base flow.

The average (1983-87) water budget for the Valley Creek basin was approximated by a steady-state simulation. Simulated underflow is 0.26 in. into the basin from the west and 1.33 in. out of the basin to the east. The calibrated model was most sensitive to recharge rate, less sensitive to streambed hydraulic conductance and aquifer anisotropy, and least sensitive to aquifer thickness, aquifer hydraulic conductivity, and evapotranspiration rate.

The effect of increased ground-water development on base flow and underflow into and out of the Valley Creek basin was simulated by locating a hypothetical 4 Mgal/d (4.04 in/yr) well field in different parts of the basin. Locating the well field near the northwestern surface-water divide would increase underflow into the Valley Creek basin by 1.41 in/yr and reduce the base flow of Valley Creek by 1.88 in/yr or 14 percent. Locating the well field near the southwestern surface-water divide would increase underflow into the Valley Creek basin by 0.97 in/yr and reduce base flow by 1.08 in/yr or 8 percent. Locating the well field near the northeastern surface-water divide would change the underflow on the east side of the basin from a net outflow of 1.33 in/yr to a net inflow of 1.24 in/yr and reduce base flow by 0.71 in/yr or 5 percent. Locating the well field near the southeastern surface-water divide would change the underflow out of the basin to the east from a net outflow of 1.33 in/yr to a net inflow of 0.07 in/yr and reduce base flow by 1.59 in/yr or 12 percent. Locating the well field near the center of the basin would reduce the base flow of Valley Creek by 2.97 in/yr or 23 percent. Underflow into the Valley Creek basin from the west would increase by 0.12 in/yr. Underflow out of the basin to the east would decrease by 0.02 in/yr, and underflow into the basin from the east would remain the same.

As the quarries in Chester Valley expand, ground-water seepage into the quarries probably will increase. Increased inflow was simulated as increased withdrawal by pumping. A 100 percent increase in the pumping rate of the Cedar Hollow quarry, from 3.93 to 7.86 Mgal/d, would reduce the natural base flow of Valley Creek by 18 percent. However, the water pumped from the quarry is discharged to Valley Creek, and the base flow at the gaging station would increase by about 10 percent.

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