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Hydrogeologic Study of Water-Well Failures in Argillite Bedrock of Sourland Mountain, Somerset County, New Jersey, in 1982

Department of Environmental Protection
Division of Water Resources

1988

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HYDROGEOLOGIC STUDY OF WATER-WELL FAILURES IN ARGILLITE BEDROCK OF SOURLAND MOUNTAIN, SOMERSET COUNTY, NEW JERSEY, IN 1982

by

Hugh F. Houghton

New Jersey Department of Environmental Protection
Division of Water Resources
Geological Survey
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CONVERSION FACTORS

Area

1 acre = 43,560 square feet (ft^2)
1 square mile (mi^2) = 640 acres

Volume

1 cubic foot (ft^3) = 7.481 gallons

Flow (precipitation, stream discharge, or recharge)

1 inch/year = 47,580 gallons/day/square mile (gpd/mi^2)

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> Quoted excerpt (in quotation marks), New Jersey Geological Survey Technical Memorandum 88-2, Houghton, 1988.

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ABSTRACT

The Sourland Mountain area is underlain by hard, relatively impermeable sedimentary rock (argillite) of the Lockatong Formation of late Triassic age, and hard intrusive igneous rock (diabase) of early Jurassic age. Average well yield from the diabase is among the lowest of any bedrock type in New Jersey. The median yield from wells tapping argillite in the study area is between 2 and 3 gallons per minute. Some of the lowest reported yields from wells tapping argillite occur where the argillite has been baked by adjacent diabase.

Severe water-level declines were reported in several wells near the top of Sourland Mountain in Hillsborough Township, Somerset County, New Jersey, late in 1982. This study was initiated to determine the cause of the well failures.

The principal cause of ground-water-level decline was overpumping in an area of closely spaced wells. A three-year period of below-normal precipitation (1980-1982) contributed to the problem, but was probably not the main cause of well failures in the study area.

Water budget calculations for the years 1969 though 1986 suggest that a maximum deficit of ground-water and soil-moisture storage occurred in 1980, during the first year of a three-year period of low precipitation. Water levels in wells were probably recovering to nearly normal levels during 1982.

The rate of ground-water pumping on a particular 1.0-acre lot in the study area was estimated to exceed 500 gallons per day. This withdrawal rate was probably the main cause of depression of water levels centered on two wells on this lot. Observed water levels in these wells and two adjacent wells were more than 100 feet below regional ground-water levels in December, 1982. The areal extent of water-level depression was small, indicating an aquifer with low permeability.

INTRODUCTION

Purpose of Study

In the winter of 1982-83 the New Jersey Geological Survey investigated the geology and hydrology of the northeast part of Sourland Mountain in Somerset County. The study area is delineated in figure 1. The purpose of the study was to determine the causes of critically low ground-water levels and to improve understanding of the ground-water system in this area.

Methods of Investigation

The plan of investigation included the following tasks:

1. Compile records of wells in the area affected by lowered ground-water levels.
2. Measure ground-water levels and compare them with older levels, to determine whether long-term changes had occurred.

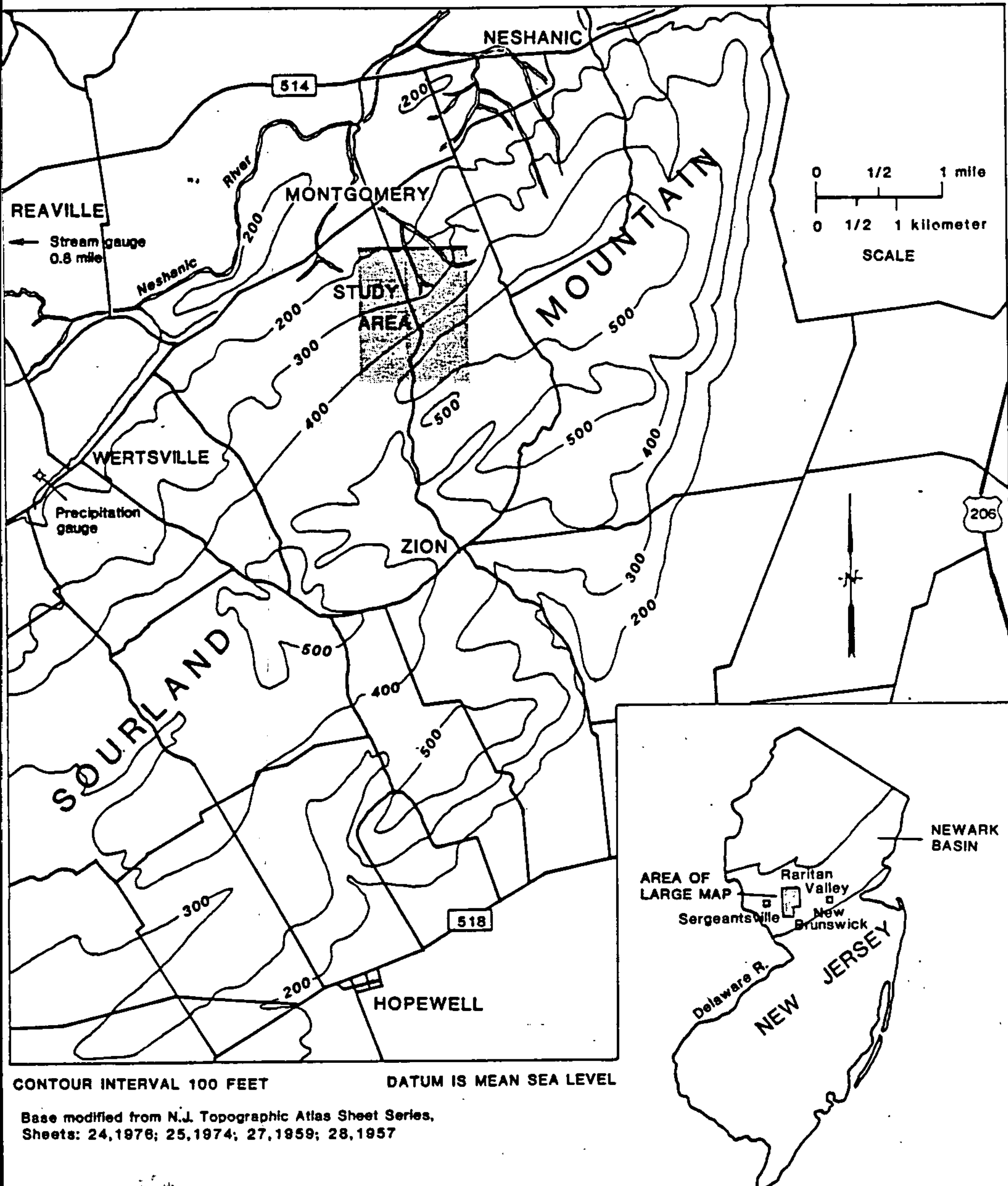


Figure 1. Map showing Sourland Mountain and area of study.

3. Perform pumping tests and observe drawdown and recovery rates in order to determine aquifer characteristics.
4. Obtain continuous water-level data from an observation well in the study area, to determine the magnitude of short-term water-level fluctuations.
5. Inspect domestic water-supply systems to determine pumping capacity and other relevant data.
6. Provide data and interpretations to aid in planning ground- water use and delineate possible preventive or remedial measures.

Data pertaining to ground-water occurrence, including well records, soil logs, soil maps, and geologic maps and reports were compiled for the study area. Precipitation records for Wertsville, stream discharge records for the Neshanic River at Reaville, and pan-evaporation records for New Brunswick were compiled for the period 1969-1985. A hydrologic budget for the Sourland area was prepared from the above data.

This report contains more background information than usual for a technically-oriented readership. It is hoped that this additional material will make the results of the study accessible to a wider audience, particularly among residents and officials of the Sourland Mountain region of New Jersey.

Acknowledgments

This study was initiated in 1982 under the supervision of Frank Markewicz, then Acting State Geologist of New Jersey. I thank several of my colleagues at the New Jersey Geological Survey (NJGS), including Daniel Dombroski who assisted with water-level measurements under adverse conditions, I. G. Grossman who suggested using a water budget and made many helpful editorial comments, and James Boyle who helped develop water-budget methods presented in the report. Richard Walker of the U. S. Geological Survey (USGS) provided water-level recording instrumentation. The manuscript benefited from outside reviews by Otto Zapecza and Charles Wood of the USGS. The methodology used and conclusions made in this report do not necessarily concur with the views expressed by the reviewers.

GEOLOGY

Regional Geology and Topography

Sourland Mountain is formed by the leading edge of a tilted slab of bedrock, or fault block, which extends in New Jersey from the Delaware River northeastward into the Raritan Valley. The fault block is tilted northwestward, causing the rock layers to dip down toward the northwest (fig. 2). Slopes facing northwest are moderate, roughly paralleling the layering of the rocks. The eroded surface of the fault uplift faces to the southeast, where slopes are irregular and generally steeper than northwest- facing slopes. The summit of Sourland Mountain is elongated in a northeast-southwest direction and rises to a maximum elevation of about 560 feet. The highest ridges of the plateau-like summit are underlain by diabase, while secondary summits on the north flank of the mountain are underlain by argillite.

The Sourland fault block consists of a thick section of sedimentary rocks intruded by a sheet-like sill of diabase, an intrusive igneous rock commonly called "trap rock". The sedimentary rocks are late Triassic in age (215 to 200 million years old), while the diabase is somewhat younger, dating from earliest Jurassic time (about 195 million years ago) (Cornet, 1977; Dallmeyer, 1975). The sedimentary rocks at the base of the fault block are tan and red sandstone of the Stockton Formation. This sandstone is overlain by the Lockatong Formation, composed mostly of argillite, a gray, hard siltstone with abundant fine- grained cementing minerals. Recent geologic mapping by the NJGS indicates that the argillite sequence is about 2200 feet thick and the diabase is about 1400 feet thick. The uppermost beds of the Sourland fault block are part of the Brunswick Group, consisting mostly of reddish-brown siltstone and

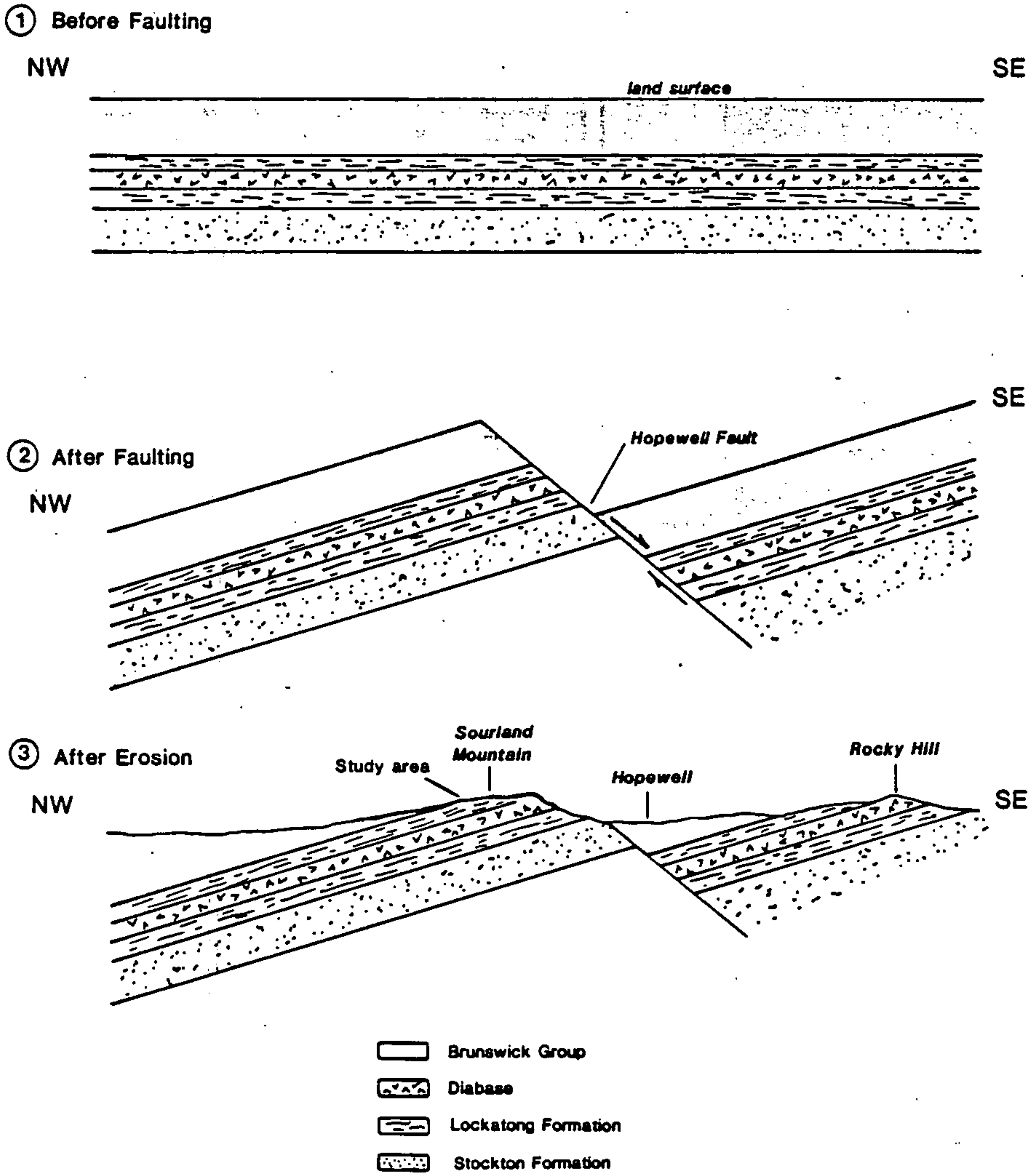


Figure 2. Generalized geologic sections showing evolution of geology and topography of Sourland Mountain region.

mudstone without hard cement. The direction of maximum dip of the sedimentary rocks is about 35 degrees west of north (N35W); the dip angle averages about 18 degrees near the study area.

The argillite below and above the intrusive sill has been baked and hardened. The baked or thermally metamorphosed zone extends more than one hundred feet into the underlying rock, and several hundred feet into the overlying sedimentary rock. At the land surface, the lower and upper baked zones have an outcrop width of about 500 and 1400 feet respectively (H.B. Kummel, written communication, 1896).

Physical properties of the diabase are fairly uniform from the bottom to the top of the sill except for minor changes in crystal size or mineral composition. These subtle changes in the diabase do not greatly affect its hydrologic character. However, vertical variations in sedimentary rocks can significantly affect their hydrologic properties.

The sedimentary rocks consist of both thick beds (about 6 feet to more than 30 feet thick) of hard, sparsely fractured siltstone or argillite, and thinner-bedded interlayers (commonly totaling 1 to 15 feet in thickness) of soft, highly fractured mudstone. Mudstone beds are not abundant or thick in the upper Lockatong Formation. In the lower Brunswick Group mudstone beds increase upwards in number and thickness. Mudstone beds store more water per unit volume than the hard sandstone and argillite, due to their greater porosity. The scarcity of mudstone beds in the upper Lockatong Formation diminishes the available ground-water storage capacity of the bedrock aquifer. This is a factor in low yields and low specific capacities of wells in the study area.

Most of the study area is underlain by argillite of the Lockatong Formation. All domestic wells with sharp water-level declines reported in 1982 tap this formation. Three deep wells also penetrate the underlying diabase. The argillite is thickest in the north, at its contact with the overlying siltstone of the Brunswick Group. There, the argillite extends to a depth of about 800 feet (see fig. 9). The argillite overlies the diabase sill which intersects the ground surface a few hundred feet southeast of the study area.

Joints and Faults

In the rocks of the Sourland region, most of the ground water is stored in and moves through fractures. These fractures are of two types: (1) high-angle joints and faults, and (2) bedding- plane joints.

High-angle joints are fracture planes which developed nearly perpendicular to the bedding of the rocks while they were still nearly horizontal. These joints are nearly perpendicular to bedding everywhere in the Newark basin, even where the bedding dips steeply. These joints occur in regular sets, mostly trending 15 to 50 degrees east of north (N15E to N50E). Spacing between high-angle joints ranges from less than 6 inches in typical siltstone beds to more than 36 inches in massive argillite and diabase. Joints are always more closely spaced in fault zones than they are in the same rocks elsewhere.

The low fracture permeability of some of the argillite beds of Sourland Mountain may have been further reduced by thermal alteration by the intruding diabase. The diabase was probably emplaced before the formation of the prevalent high-angle joints. Thermal hardening by recrystallization of argillite beds adjacent to the igneous intrusion may have increased the resistance of these beds to fracturing. Argillite in the "baked" zones above and below the diabase sill may have fewer joints per unit volume of rock and therefore, lower fracture porosity.

Faults are fractures along which differential movement has occurred. Most faults in the Newark basin occur in subparallel groups referred to as fault zones. Major fault zones, such as the Hopewell fault located southeast of the study area, can consist of hundreds of individual fault planes in a zone thousands of feet wide. Many faults in the Newark basin are oriented in a NE- SW direction, roughly parallel to the most common direction of strike of bedding and the strike of principal joints.

Individual faults commonly have a zone of broken rock ranging in width from less than one inch to more than 36 inches. Some faults, especially those in the major fault zones, are continuous over long distances. Such faults may connect water-bearing joints and beds and cause adjacent wells to interact strongly.

Faults may be discernible during the drilling of a well and may be reported by the driller as soft "seams" (a driller's seam can also be a soft or highly fractured bed). Faults often contain minerals such as calcite, epidote, or pyrite, which may be reported by the driller. Large pieces of rock may appear in well cuttings from a fault zone since the already broken rock can be flushed out of the hole before it is pulverized by the rotary drilling bit.

Porosity and permeability can be either greater or less than usual in fault planes. Many faults produce ample supplies of water, especially from hard rocks such as sandstone, argillite, and diabase. In other instances, and particularly in soft rocks such as mudstone or shale, fault planes may contain large amounts of clay and pulverized rock. This pulverized rock, known as "fault gouge" can be quite impermeable, causing the fault plane to act as a flow boundary to ground water.

Joints are usually closely spaced in fault zones. As a result, fracture porosity of the bedrock may be enhanced within a wide zone (thousands of feet or more) on either side of a major fault. With the exception of the extreme northeastern end, the north flank of Sourland Mountain is distant from major fault zones. This factor may contribute to the low yields of wells in the study area, compared to areas along the southern and eastern flanks of Sourland Mountain near the Hopewell fault zone.

The second general class of fractures is bedding plane joints. The spacing of these joints is dependent upon rock type. Diabase is not bedded and therefore does not have bedding plane joints. Some joints in the diabase are subparallel to the bedding of the enclosing sedimentary rocks; these are typically widely spaced (several feet to tens of feet). In sedimentary rocks bedding-plane spacing can range from tens of feet in massive argillite, to tenths of an inch in fissile siltstone and shale. Bedding-plane fracture porosity can thus range over several orders of magnitude in different rock types.

Soils

Soil logs based on percolation test pits in soil overlying the argillite bedrock of Sourland Mountain typically show a thin (6 to 10 inches) organic-rich layer, and a much thicker (48 to 72 inches) inorganic subsoil made up mostly of dense clay with a putty-like consistency when wet (fig. 3). Soil logs with percolation test results are prepared by engineering firms and filed with other site permits for each lot at the local government offices. The Soil Conservation Service (SCS) soil report for Somerset County describes the subsoil of the Lehigh series, which underlies much of the study area, as a silty clay loam (Kirkham, 1976). Its clay content may be as much as 40 percent (Foth, 1983, p.26). The subsoil contains progressively more broken rock with depth. Fractured argillite with some clay is usually encountered at 5-6 feet below the surface (fig.3).

Soils of the study area are shown in figure 4. Most of the soils are mapped as Lehigh silt loam (LhB and LhC), and Penn shaly silt loam (PnC) (Kirkham, 1976). The "B" indicates slopes of 2-6 percent; the "C" means 6-12 percent slopes. Some soil unit boundaries have been modified for this study to reflect trends of the bedrock formations.

The area where well problems occurred in 1982 is covered by Lehigh series soils. The water-bearing characteristics of this series have been described by Kirkham (1976, p.29) as follows:

> "Permeability is slow in the subsoil, and the available water capacity is moderate. These soils have a water table perched at a depth of 1/2 foot to 4 feet during winter and early in spring."

Thick, clay-rich subsoil which develops on the argillite hinders infiltration of water into the bedrock aquifer. As a result, perched water tables develop within the subsoil zone during periods of high precipitation. This perched water slowly percolates into the underlying bedrock, or returns to the atmosphere by transpiration and evaporation during drier parts of the year.

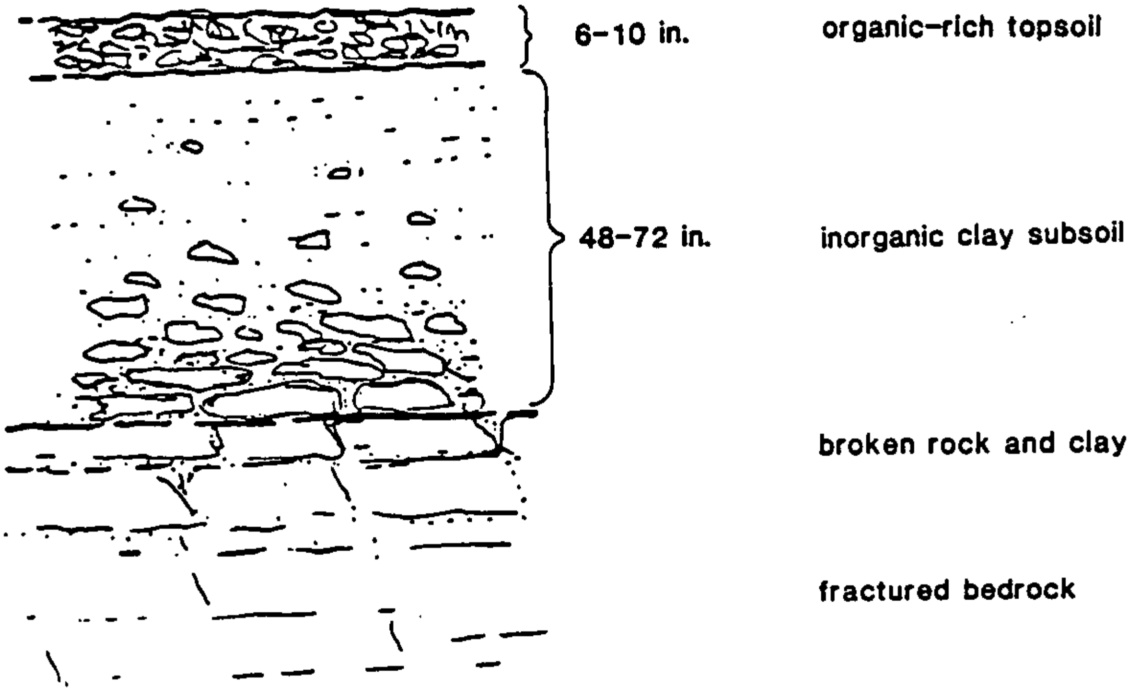
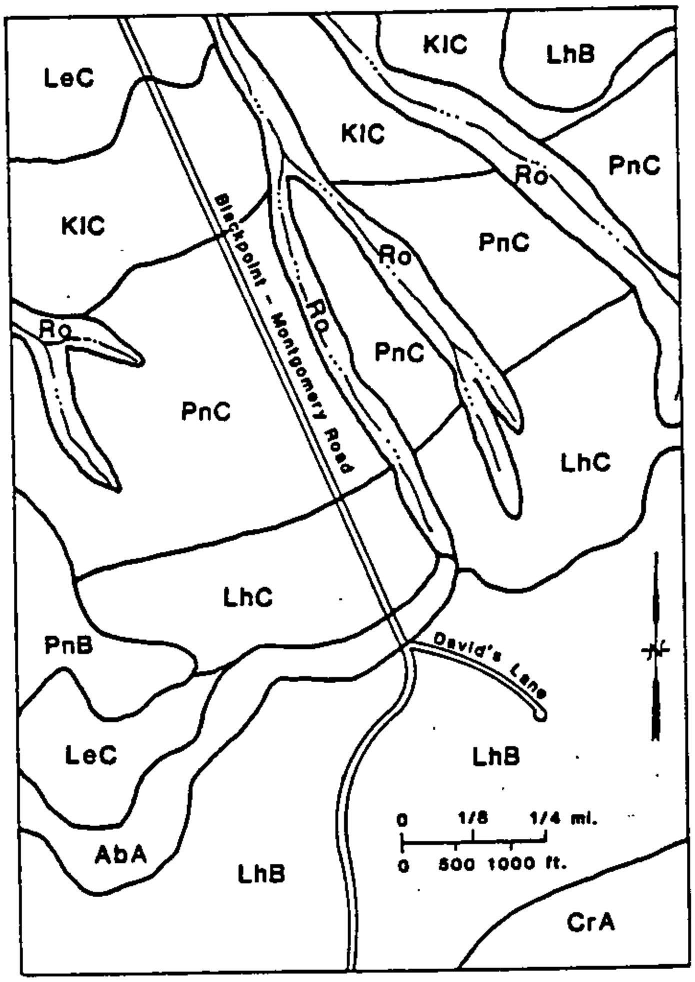


Figure 3. Typical soil profile on Lockatong argillite, Sourland Mountain.



Explanation

AbA - Abbottstown silt loam, 0-2 percent slopes

CrA - Croton silt loam, 0-2 percent slopes

KlC - Klinesville shaly loam, 4-12 percent slopes

LeC - Lawrenceville silt loam, 6-12 percent slopes

LhB - Lehigh silt loam, 2-6 percent slopes

LhC - Lehigh silt loam, 6-12 percent slopes

PnC - Penn shaly loam, 6-12 percent slopes

Ro - Rowland silt loam

Note: CrA-LhB and LhC-PnC contacts coincide approximately with contacts of diabase-Lockatong and Lockatong-Brunswick, respectively.

Figure 4. Soil map of study area, modified from Kirkham (1976).

GROUND WATER

Storage and Movement

Ground water is stored either in pores between grains or in fractures in bedrock. Moderately cemented sedimentary rocks have pores between their grains. Tightly cemented sedimentary rocks such as argillite, and most igneous rocks such as diabase, generally lack pores between grains or minerals. In these rocks, ground water occurs almost entirely in fractures. Storage and flow of ground water within fractured bedrock is largely controlled by the orientation and density of fractures. Ground water flow in systematically fractured bedrock is generally anisotropic, that is, it flows more readily in some directions than others. Ground water flows more easily parallel to continuous fractures such as principal high-angle joints and bedding plane joints. It also flows more easily along beds with higher fracture permeability, such as fissile siltstone beds.

Anisotropic ground-water movement causes wells which are closely spaced along a preferred flow direction to interact more strongly than wells oriented in other directions. Anisotropic flow parallel to the strike of bedding has been reported in the Newark basin and the adjoining Gettysburg basin in Pennsylvania by Herpers and Barksdale (1951), Longwill and Wood (1965), Vecchioli (1967), Vecchioli and others (1969), Wood (1980), and Spayd (1985). Strike-parallel preferred flow under pumping conditions may result from any of several factors, including:

1. Some beds are more permeable than others. Wells drilled along strike of these beds are likely to tap the same permeable horizons.
2. Principal high-angle joint strike is close to strike of bedding throughout much of the Newark basin.
3. Bedding-plane joints have the same strike as bedding.
4. High-angle fractures commonly terminate along bedding surfaces. As a result, the large-scale geometry of fracture systems can be that of tabular zones oriented nearly parallel to bedding planes.

In the study area wells are more likely to interfere with each other if they:

- are aligned parallel to the strike of major high-angle joints, averaging about N35E, or
- tap the same bedding-plane joints or permeable beds, which depends on both depth and direction between wells.

Water occurs in two zones in the subsurface: the shallow unsaturated zone and the underlying saturated zone. The interface between these two zones is called the **water table**. All free water which occurs in the saturated zone is considered ground water. In the saturated zone, all pores in the subsurface material are filled with water. Ground water below the water table moves in response to differences in head from recharge to discharge areas. Recharge areas are mostly uplands, where elevations and potentiometric levels are higher than elsewhere. Discharge areas are found at low elevations, at springs, swamps, ponds, and streams. Ground water also flows toward manmade depressions in the water table, such as the cone of depression around a pumped well.

A semiconfined aquifer is a stratum or bed with relatively high permeability, bounded above and below by less permeable strata, of which one or both are leaky. Semiconfined conditions can occur where interconnected fractures form discrete networks bounded vertically by less fractured rock. The pressure surface for such an aquifer connects points at the highest saturated elevation in the aquifer. Ground water at any point within a semiconfined bed or fracture system will rise in tightly-cased wells to the same height as the top of the pressure surface. Thus, water in a well will rise to a height which is called the **potentiometric surface** or head. The **static head** is the head or potentiometric surface in a well which is not being pumped. The **pumping level** is the level to which the head drops under pumping conditions.

Water Wells

Wells drilled into fractured rock first penetrate an upper water- saturated zone, which is commonly a shallow water-table aquifer. Shallow ground water may be subject to contamination from on-site septic disposal and other sources, so a steel casing long enough to reach below the shallow water table is cemented into the wellbore. This casing inhibits water from flowing directly to the well from this zone. The well is then deepened, usually until one or more semiconfined aquifers are penetrated.

Most modern wells are equipped with a submersible pump which is lowered into the wellbore at the end of a steel or plastic pipe. The pipe serves as a conduit for the well water. The pump is set far enough above the bottom of the well so that it does not stir up sediment and draw it into the pump and piping system.

When the pump is operating, water is withdrawn from the wellbore, which lowers the water level in the well. Lowering of head by pumping is called **drawdown**. Drawdown in the well depresses the water table or the potentiometric surface surrounding the well, forming a **cone of depression**. As drawdown increases, the area of the water-table depression expands. If drawdown is sufficiently deep and the cone of depression expands out to nearby wells, their water levels will drop. This effect is **well interference**. Figure 5 illustrates drawdown in a well drilled in fractured bedrock.

At the time of completion of a well, the driller is required to perform a pumping test to determine the yield of the well and observe the drawdown after a period of time, generally at least one hour. Ideally, the test should be run for a period long enough for the water level to stabilize. For low yield fractured- rock wells, however, the pumping test usually ends when the water level drops below the pump intake.

The **yield** of a well is determined either by blowing air into it or pumping it and measuring the discharge rate. The **specific capacity** of a well is the number of gallons pumped per minute divided by the number of feet of drawdown. Specific capacity is expressed in gallons per minute per foot. Wells with high specific capacity produce more water and recharge more quickly than those with low specific capacity.

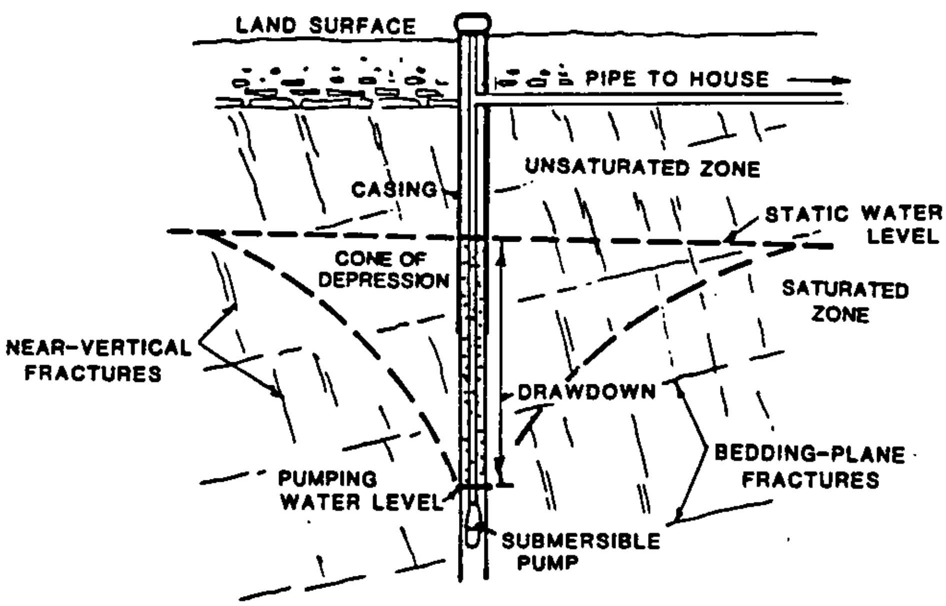


Figure 5. Diagrammatic cross section of a well in a fractured bedrock aquifer, showing nonpumping (static) and pumping levels.

The Hydrologic Cycle

The hydrologic cycle is the closed system through which water moves from the atmosphere to the earth as precipitation, enters surface water and ground water bodies, flows toward oceans, and returns to the atmosphere by evaporation and transpiration. The source of all fresh ground water is precipitation. The amount of water which circulates through the earth-atmosphere system varies from place to place, depending on the climate. The total amount of ground water available in any particular area depends mainly on two factors: 1) annual excess of precipitation over evaporation and, 2) volume of ground-water storage and permeability of the local aquifer(s).

The hydrologic cycle can be expressed by the following equation:

PRECIPITATION = RUNOFF + EVAPOTRANSPIRATION + CHANGES IN STORAGE

Precipitation is moisture in any form which falls from the atmosphere. Runoff includes all forms of surface-water flow and some shallow subsurface flow, as explained below. Evapotranspiration includes moisture which evaporates directly from the land surface or from free-water surfaces to the atmosphere, and water lost to the air from plants (transpiration). Storage changes include changes in surface-water volume, soil moisture content, and ground-water volume.

One of the most easily understood depictions of the hydrologic cycle is the system chart of Freeze and Cherry (1979, p.4). The chart is reproduced here as figure 6. The three main components of the cycle (precipitation, runoff, and evapotranspiration) are in tablet-shaped boxes. Terms involving storage are in rectangular boxes, and those involving rates of movement are in hexagonal boxes. The part of the cycle involving ground water is at the bottom of the chart. In this chart throughfall is the component of precipitation which falls directly on land or water surfaces, while interception includes any precipitation which is intercepted by foliage. The system chart assumes no underflow of ground water into or out of drainage basin.

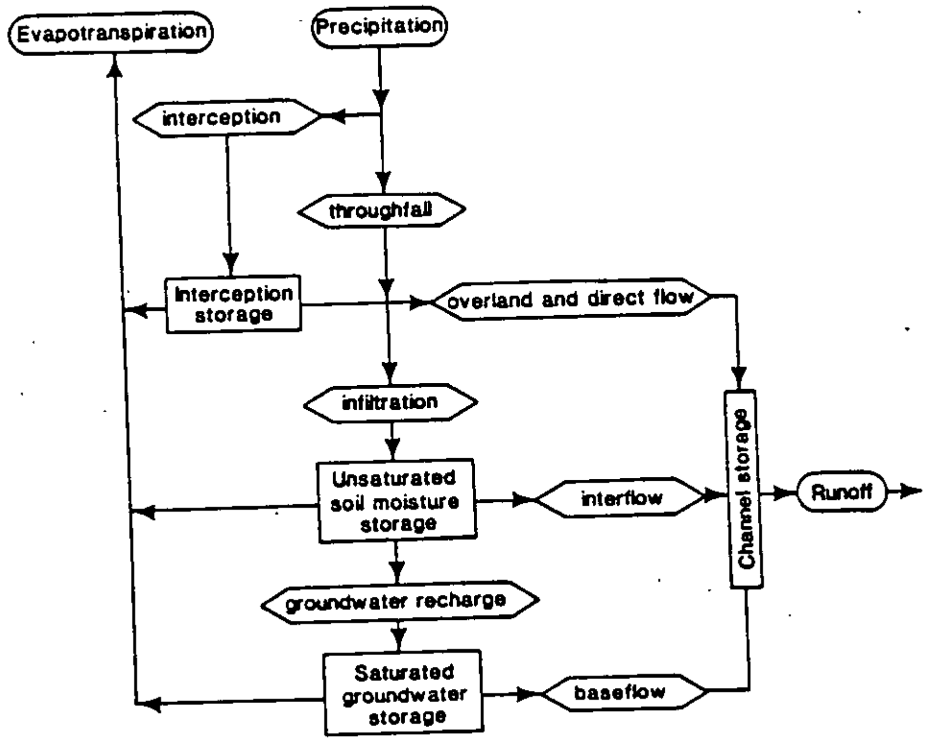


Figure 6. System chart of the hydrologic cycle. Modified from Freeze and Cherry (1979). Reproduced by permission of Prentice- Hall, Inc.

From the relationships shown in figure 6, an equation can be written for ground-water recharge:

GROUND-WATER RECHARGE = GROUND-WATER DISCHARGE + EVAPOTRANSPIRATION (from ground water) + STORAGE CHANGE (in ground-water storage)

The equation above shows that ground-water recharge is greater than ground-water discharge, except when ground-water storage change is both negative and greater in absolute value than evapotranspiration from ground water.

HYDROLOGIC DATA

Precipitation

Precipitation data from the U.S. Weather Service record station at Wertsville for the years 1969-1986 are shown in figure 7 (National Oceanic and Atmospheric Administration, 1969-1986). The location of the precipitation gauge is shown in figure 1. The average annual precipitation for the years shown was 47.33 inches. The cumulative deficit for the dry years of 1980-82 was over 28 inches, based on the 18-year average. Reports of well failures in the fall of 1982 correspond with the later stages of this three-year period.

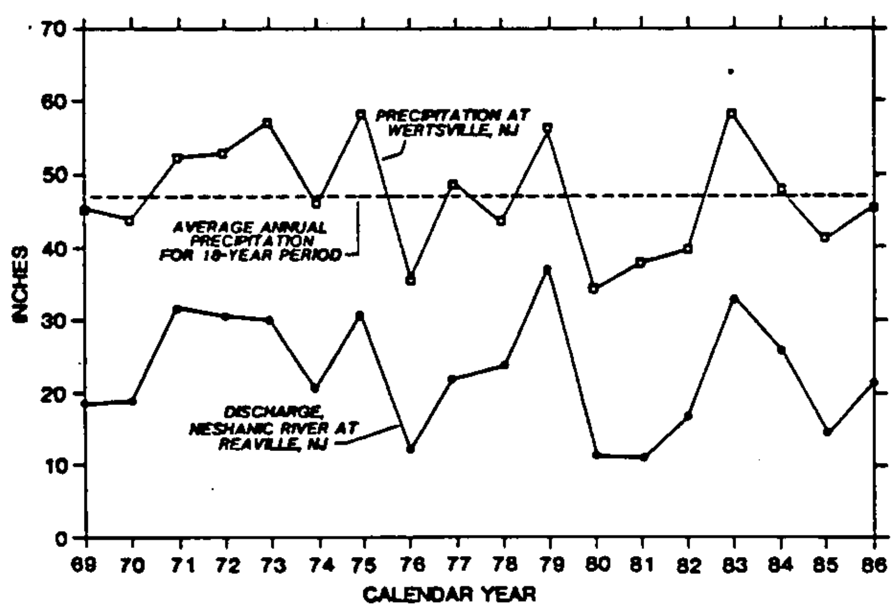


Figure 7. Annual precipitation and annual stream discharge near Sourland Mountain. Discharge given in inches of water over the area of the drainage basin. Sources of data given in text. Location of precipitation and stream gaging stations shown in figure 1.

Runoff

Discharge data are available for two streams draining watersheds partly in the Sourland Mountain area for the period 1969-1985. The records are for the Neshanic River at Reaville, and Stony Brook at Princeton. Back Brook tributary near Ringoes, which drains a small watershed on the north slope of Sourland, has records from 1977. The quality of records from Back Brook tributary is fair to poor (U.S.G.S., 1982, for example). Stream flow data are published annually by the U.S. Geological Survey (1970-87).

The patterns of discharge of the Neshanic River and Stony Brook are similar, therefore only Neshanic River data are shown in figure 7. The location of the stream gauge is indicated in figure 1. Stream discharge is given in inches of water (over the area of the drainage basin) so it may be compared directly with precipitation at Wertsville. Precipitation and stream-discharge data are also listed in table 4.

Evaporation

Evaporation measurements in central New Jersey are made by the meteorology department at Cook College in New Brunswick (Keith Arnesen, written communication, 1987). Measurements are of pan evaporation, that is the amount of water in inches that evaporates daily from an open pan of water. The data reflect daily changes in relative humidity and air temperature; they do not, however, accurately reflect transpiration or evaporation from foliage. Pan evaporation data are listed and discussed in a later section, as part of a water-budget calculation.

WELL AND AQUIFER CHARACTERISTICS

Well Yields and Specific Capacities

Water-well statistics for the bedrock of the Sourland Mountain area are given in table 1. Median values rather than means of yield and specific capacity were reported by Hordon (1984), because the median is less affected by extreme values. This is especially important for small data sets, such as those in this report. The median is the middle value in a set, that is, 50 percent of the values are larger and 50 percent are smaller than the median.

Table 1 summarizes water-well data for Brunswick and Lockatong Formations and diabase in three reporting areas: Hunterdon County (Kasabach, 1966), the Sourland Mountain region (Hordon, 1984), and the Newark basin in New Jersey (Carswell, 1976; Carswell and Rooney, 1976; Kasabach, 1966; Miller, 1974; Nichols, 1968; Vecchioli and Miller, 1973; Widmer, 1965). Median yields could not be computed from statewide data.

Records are available for 24 wells in the study area; 18 include drawdown and specific capacity data (table 2). Well locations are shown in figure 8.

Median yield of 24 wells in the study area, as reported by well drillers, is between 2 and 3 gpm (gallons per minute), compared to 7 gpm for the Lockatong argillite in the entire Sourland Mountain region (Hordon, 1984). Median specific capacity for 18 wells in the study area is 0.02 gpm/ft, less than half of the 0.06 gpm/ft for argillite in the entire Sourland region (Hordon, 1984). Reported yield data are based on pumping tests that are usually of short duration and generally do not reflect pumping rates at which water level stabilizes in the well. Therefore, actual long-term yields are probably lower, especially where the record indicates that pumping level was near the bottom of the well at the end of the test.

TABLE 1. Yield and specific capacity of bedrock wells. Yield given in gallons per minute (gpm). Specific capacity units are gpm per foot of drawdown (gpm/ft). Number of wells included in computation denoted by letter n. Sources of data given in text.

Formation	Median yield (gpm)	Mean yield (gpm)	Median specific capacity (gpm/ft)
Brunswick			
Hunterdon County	15 (n=528)	19 (n=528)	1.41 (n=272)
Sourland Mountain	12 (n=323)	--	0.24 (n=112)
Statewide		16.3 (n=1196)	
Lockatong			
Hunterdon County	6 (n=186)	12 (n=186)	0.37 (n=63)
Sourland Mountain	7 (n=949)	--	0.06 (n=426)
Statewide		9.5 (n=393)	
Diabase			
Hunterdon County	5 (n=65)	8 (n=65)	0.35 (n=21)
Sourland Mountain	5 (n=213)	--	0.07 (n=91)
Statewide		7.4 (n=141)	

Water Levels

Static-water levels of wells in the study area following completion of drilling are listed in table 2. Some are shown along with historic water-level data in figure 9. These data show an area of decline in water level in several wells located downslope from David's Lane. The maximum decline in nonpumping level was more than 120 feet, in well 67.

In December 1982 water levels were measured in several wells during a two-day period to observe typical ranges in drawdown and recovery. The results are shown graphically in figure 10. Wells 65 and 67 were not operated on the morning of December 16. The decline of more than 90 feet measured in well 66F at 11:00 am on December 16 followed the use of a washing machine in the residence. Declines of water level in wells 65 and 67 correspond with the time of greatest drawdown in well 66F, suggesting that some interference was occurring.

Continuous Water-Level Measurement

A clock-driven water-level recorder was installed on well 117R and records were kept for a 3-month period in 1982-83. Water levels rose gradually from 25.6 feet to 23.5 feet below top of casing during December. Between early January and mid-February the water level rose from 23.5 feet to 16.7 feet below top of casing. Rhythmic daily fluctuations of water level averaged about 0.1 foot, with a maximum of about 0.2 foot. Daily declines occurred usually between 8 PM and 2 AM, presumably corresponding to the hours of maximum water use in the front well (117F), located approximately 250 feet from the recorder well.

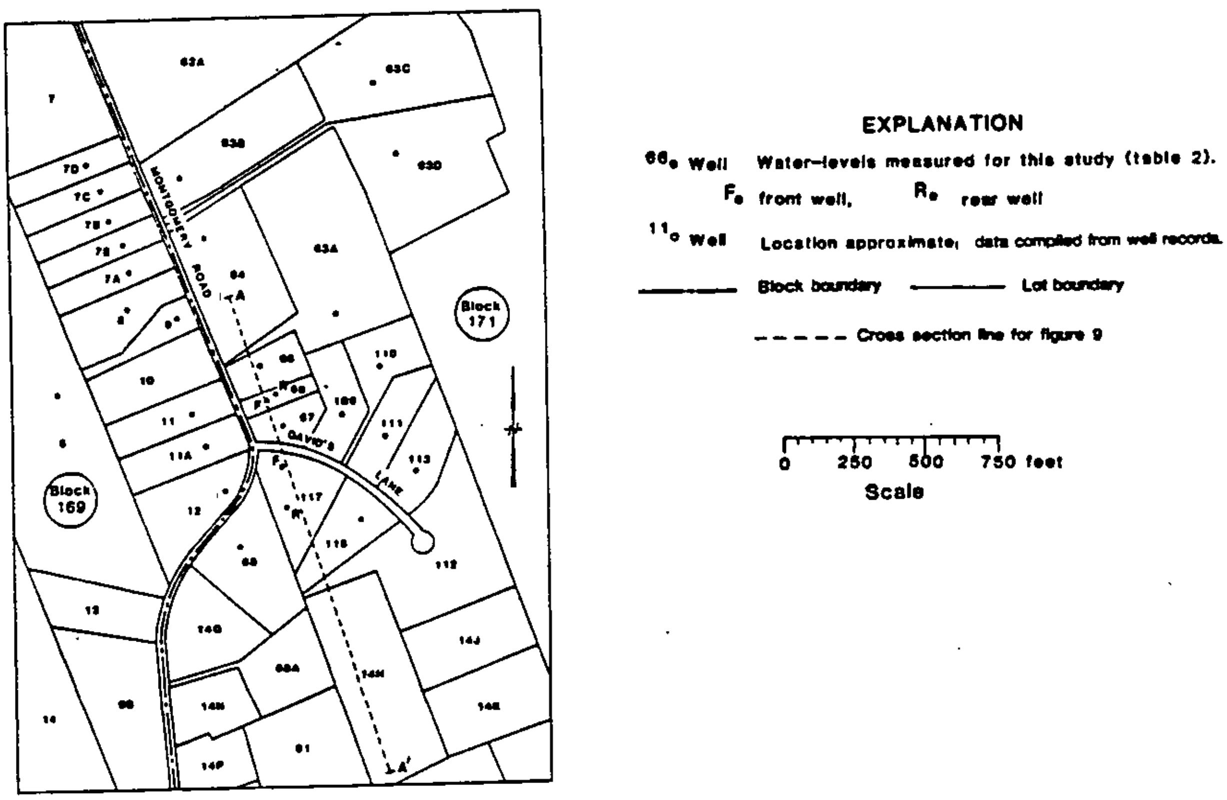


Figure 8. Well-location map. Modified from Hillsborough Township tax map, 1959, Aero Service Corp.

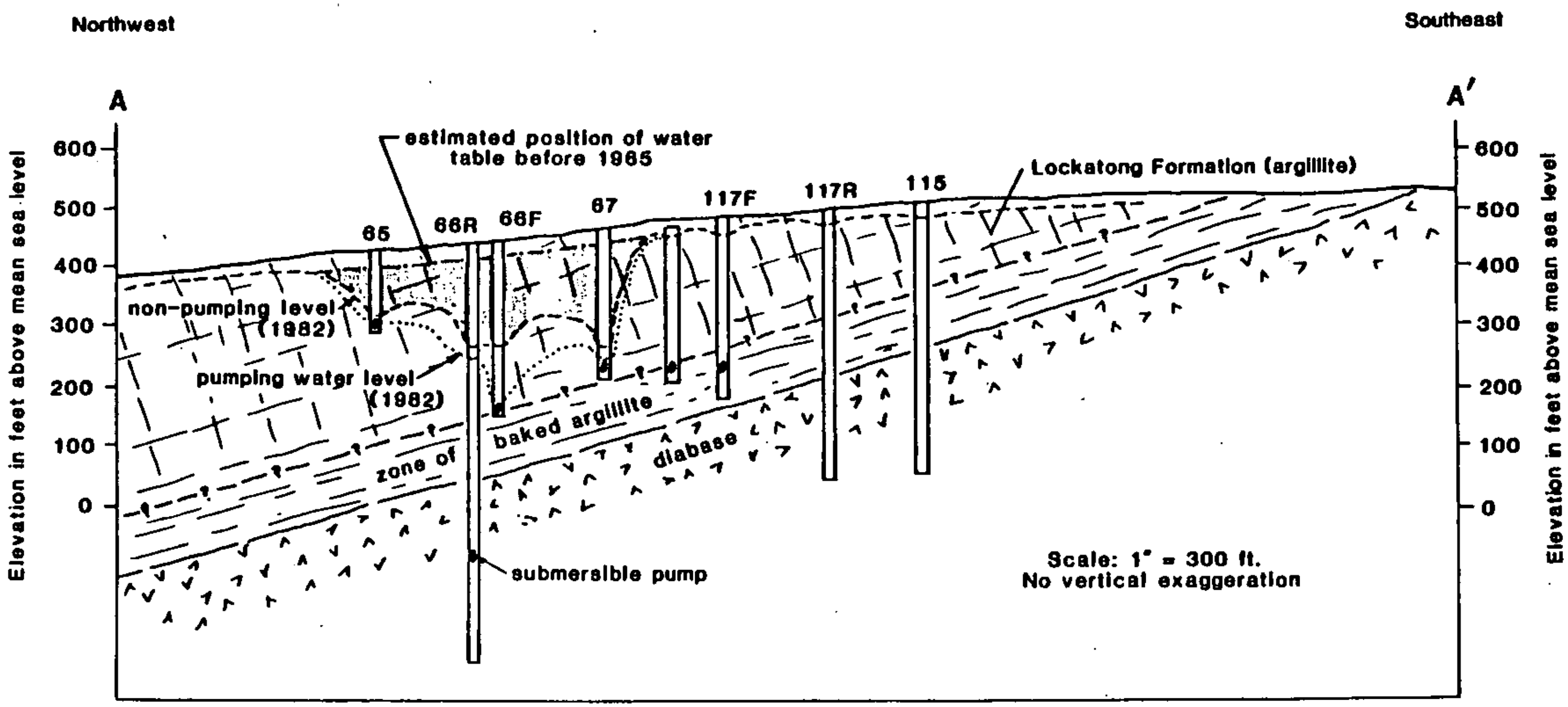


Figure 9. Hydrogeologic section of study area. Water levels based on data in table 2. Line of section shown in figure 8.

TABLE 2. - Records of wells in study area.

Depths are in feet below land surface. All wells are 6-inch nominal diameter. Gpm, gallons per minute. All wells tap the Lockatong Formation. Wells 7D and 63C also penetrate the Brunswick Formation; wells 66R, 115, and 117R also penetrate diabase.

Well number	Date completed	Well depth(ft)	Casing length (ft)	Static water level at completion(ft)	Static level on 12/16/82 (feet)	Yield at completion (gpm)	Pump depth (feet)	Pump capacity (gpm)	Duration of pumping test(hrs)	Water level, end of pump test (ft)	Draw-down (feet)	Specific capacity (gpm/ft)	Approximate elevation of well head(ft)
6	11/82	300	32	10	--	2	290	4	1	290	280	0.01	--
7A	12/65	265	60	30	--	4	265	8	3	265	235	0.02	385
7B	--	135	41	10	--	12	--	--	--	80	70	--	340
7C	6/65	305	60	20	--	12	250	--	--	250	230	0.05	355
7D	3/72	460	60	flows	--	2 to 3	300	--	5	300	300	0.01	345
8	5/78	250	40	42	--	4.5	230	5	8	230	188	0.02	400
11	8/73	475	67	45	--	2.5	--	--	4	400	355	0.01	445
12	1981	285	--	--	22	2 to 3	260(?)	--	--	--	--	--	465
63A	10/73	640	64	20	--	2	--	--	5	378	358	0.01	440
63B	9/66	250	60	10	--	10	--	--	--	90	80	0.13	355
63C	12/79	500	50	--	--	2 to 3	--	--	--	--	--	--	380
63D	12/78	500	50	50	--	3	300(?)	--	1/2	300	250	0.01	410
65	1965(?)	140	--	30(?)	120	2 to 3	135(?)	--	--	--	--	--	430
66R	1972(?)	720	60	--	180	1/2	546(?)	--	--	546	146	0.00	440
66F	11/82	300	30	100	190	1/2	290	4	1	290	190	0.00	445
67	1970(?)	260	60	80	205	4	240	--	--	240	160	0.03	465
68	6/82	285	32	15	--	3	270	5	1	270	255	0.01	490
109	4/73	400	65	40	--	2	350	--	4	350	310	0.01	480
110	--	350	63	5	--	4	250	--	--	250	245	0.02	475
111	--	245	33	10	--	4.5	190	--	--	190	180	0.03	495
113	9/78	400	50	50	--	1	390	1	3	380	330	0.00	510
115	1981	460	--	--	27	1/2	none	--	--	--	--	--	510
117R	5/78	465	42	35	25	1/4	none	--	--	--	--	--	500
117F	5/78	305	40	50	43	3	252	8	--	252	202	0.02	480

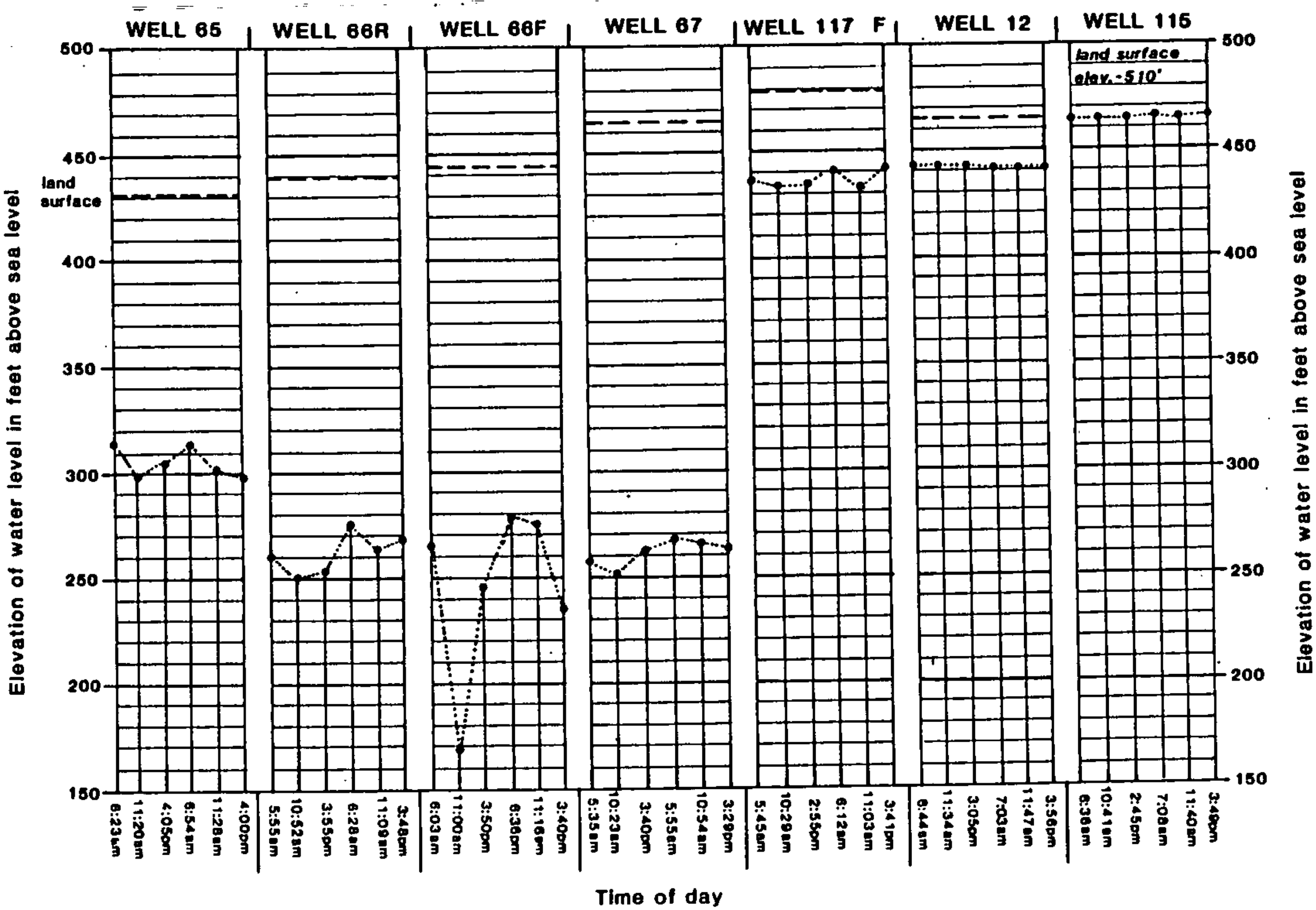


Figure 10. Fluctuations of water levels in seven wells on December 16-17, 1982.

HYDROGEOLOGIC INTERPRETATIONS

Estimates of Recharge and Sustainable Yield

Ground-water studies usually include an estimate of recharge to the aquifer. The average annual recharge may be referred to as perennial or sustained yield, based on the idea that long-term withdrawal of ground water at a rate exceeding natural recharge will reduce storage. The actual relationship between ground- water pumping and aquifer storage change is complex and depends on porosity and permeability of the aquifer, topography, climate, vegetation, impervious cover, sewering, and other variables.

Recharge and sustained ground-water yield have been estimated by several different methods, including hydrograph separation, flow- duration analysis, and water-budget analysis. Buller and Sweigard (1978) used water-budget analysis in a study in Hopewell Township. Posten (1982) used the hydrograph separation method on several fractured-rock drainage basins in New Jersey, including one basin underlain by Lockatong argillite near Flemington. Hordon (1984) conducted hydrograph separation and stream-flow- duration analyses on data from several watersheds in the vicinity of Sourland Mountain. Wright (1982) used the hydrograph separation method in other areas of the central Delaware valley. The results of these analyses and some other published estimates of recharge or sustainable yields are given in table 3.

Water-budget analysis has been used to estimate recharge (Williams, 1981, for example), although some components of recharge cannot be directly assessed from hydrologic and meteorological data. Ground-water discharge is approximately equivalent to base flow of streams and has been used directly as an indicator of "safe" yield (Hordon, 1984). A valid estimation of recharge must include estimates of ground-water discharge or base flow, direct evapotranspiration from ground water, and changes in ground-water storage.

Water Budget for the Sourland Mountain Area

In order to determine whether long-term meteorologic conditions could have had an effect on ground-water levels, a yearly water- budget analysis was conducted. No water-budget method was found in the literature which was suitable for yearly hydrologic data. Most published water-budget methods are designed to track and predict monthly evaporation for use in irrigation planning. The most accurate methods use an energy-balance approach (Penman, 1956; Tanner and Pelton, 1960; Van Bavel, 1966; among others). The data requirements of these methods are beyond the scope of this study.

A water-budget method was developed for this study which is based on a simple water-balance equation and uses annual hydrologic data. Hydrologic data include precipitation, stream discharge (runoff), and pan evaporation. The equation used is the simplest expression of the hydrologic cycle:

PRECIPITATION = RUNOFF + EVAPOTRANSPIRATION + CHANGES IN STORAGE

Subtracting runoff from precipitation yields a remainder which is equal to evapotranspiration (ET) plus changes in storage. When summed over a large number of years, storage changes will approach zero (unless the climate is changing, or the physical environment is altered significantly). The long-term difference between precipitation and runoff, therefore, is equal to average ET. Annual estimates of ET can be normalized to match the calculated average value.

Yearly ET was estimated by using the ratio of pan evaporation to precipitation. Annual ET plus storage change was multiplied by this ratio, to yield a preliminary or unadjusted estimate of ET. This step is equivalent to assuming that the water which does not leave the watershed as stream runoff is evaporated at the same ratio as that between measured rainfall and evaporation from a water-filled pan. The final estimate of ET is made by multiplying the first estimate by a factor which makes the

TABLE 3. Published estimates of recharge or sustained yield. Results given in gallons per day per square mile (gpd/mi2). Recurrence interval is the average time interval within which the predicted volume will be equaled or exceeded once.

Aquifer & rock type	Drainage basin	Variable estimated	Analytical method	Sustained Yield (gpd/mi2)	Recurrence interval	Reference
Brunswick						
shale & siltstone	Middle Delaware River Basin	Baseflow	Hydrograph separation	53,000	1 yr. in 10	Wright(1982)
siltstone & shale	Sourland region	Safe yield	Flow duration	110,000	1 yr. in <5	Hordon(1984)
shale	Royce Brook trib., near Belle Mead	Baseflow	Hydrograph separation	125,000	1 yr. in 10	Hordon(1984)
shale & siltstone	Middle Delaware River basin	Baseflow	Hydrograph separation	220,000	1 yr. in 2	Wright(1982)
siltstone & shale	Flemington area.	Baseflow	Hydrograph separation & specific yield ratio	231,000	long-term average	Hordon(1984)
siltstone & shale	W. Conewago Creek, PA (Gettysburg Fm.)	Recharge	Water budget	288,000	N.A.	Wood(1980)
siltstone & shale	Stony Brook at Princeton	Recharge	Discharge frequency (=flow duration)	300,000	long-term average	Geraghty and Miller(1973)
Lockatong						
argillite	Walnut Brook, NJ	Safe yield	Flow duration	45,000	1 yr. in <5	Hordon(1984)
mixed, mostly Lockatong	Stony Brook at Glenmoore, NJ	Baseflow	Hydrograph separation	88,000	1 yr. in 10	Hordon(1984)
argillite	Walnut Brook, near Flemington, NJ	Baseflow	Hydrograph separation	92,000	99% exceedence	Posten(1982)
argillite	Stony Brook at Princeton	Recharge	Flow duration	100,000	gross average	Geraghty and Miller(1973)
mixed lithologies	Stony Brook at Princeton	Baseflow	Hydrograph separation	119,000	1 yr. in 10	Hordon(1984)
Diabase						
diabase	Hunterdon area	Safe yield	Flow duration	40,000	1 yr. in <5	Hordon(1984)
diabase	Stony Brook at Princeton	Recharge	Flow duration	50,000	gross average	Geraghty and Miller(1973)
diabase	Walnut Brook, NJ	Baseflow	Hydrograph separation & specific yield ratio	83,000	long-term average	Hordon(1984)

average ET equal to that computed from the hydrologic data. Evaporation data were available for the years 1969 through 1986. Results of the water budget for those years are given in table 4. The source of data or method of calculation for each column is discussed in footnotes to the table.

The water-balance method presented here, and two variations of the method, were checked for internal consistency by conducting linear regression analysis on various pairs of values. This technique evaluates how well a straight line fits a Cartesian (X,Y) plot of two variables. The water-balance method shown yields high positive correlation values for critical pairs of variables, including precipitation vs. estimated storage change and stream flow vs. estimated storage change, and high negative correlation between stream flow and estimated evapotranspiration. The high correlations are mostly the result of factoring precipitation into estimated evapotranspiration by the ratio method used. Two other methods tested involved scaling of pan evaporation by (1) the difference between average pan evaporation and average ET plus storage change and (2) the ratio between those variables. These methods had very low correlation between critical variables. The method shown in table 4 was judged most appropriate on the basis of these correlations.

Another way to check the validity of water-budget results is to compare estimated values with observed water-level records from an observation well. The U. S. Geological Survey has maintained an observation well tapping the Stockton Formation at Sergeantsville, New Jersey since 1965 (well number 19-0002). The well is located about 9 miles from the study area, at an elevation of 342 feet. Topography is gentler at Sergeantsville than in the study area. Annual mean water levels were computed from published and unpublished data of the USGS (U. S. Geological Survey, 1978-1987; Frederick L. Schaefer, USGS, written communication, 1988).

Annual mean water levels at Sergeantsville were compared with annual values for several other variables for the period 1969- 1986. Other variables were:

1. estimated storage change (dSe) from water budget,
2. cumulative dSe from water budget,
3. precipitation, and
4. cumulative departure from mean precipitation.

The last two variables were checked to see if they were more closely correlated with water levels than values derived from the water budget.

The mean ground-water level record was closely matched by both the storage change estimated by the water budget (dSe) and by annual precipitation at Wertsville. These variables are plotted in Figure 11. Cumulative storage change does not predict ground water levels as well as annual storage change. The close match of curves indicates that the water budget works at least moderately well. Estimated annual storage change can thus be used directly to estimate ground-water level changes. Figure 11b shows that precipitation could also be used to estimate water- level changes; this would require additional calculation to convert to estimated storage change.

The estimated storage change curve (fig. 11a.) reaches a low point in 1980 and recovers somewhat during 1981-1982. The Sergeantsville observation well follows a similar pattern. The observation-well level was only 0.4 foot below the long-term mean in 1982. The estimated storage change for 1982 in the study area was -1.9 inches.

Estimated storage change can be converted to an estimate of water-level decline if two other parameters are known or can be estimated: 1) specific yield of the aquifer, and 2) ratio of soil-moisture to aquifer-storage change. The storage-change ratio will vary throughout the year, but can be assigned an approximate average yearly value. Using 1.9 inches storage deficit and varying the ratio of soil-moisture to aquifer-storage change from 3:1 to 1:2, in an aquifer with 3 percent specific yield, results in water-level decline between 16 and 42 inches. This suggests that below-average precipitation could not have produced the decline in water levels observed in the most severely impacted wells in the study area in 1982. This conclusion is supported by the fact that in two wells in the study area,water levels measured in December,1982 were higher than those reported at the time of drilling. Although the two

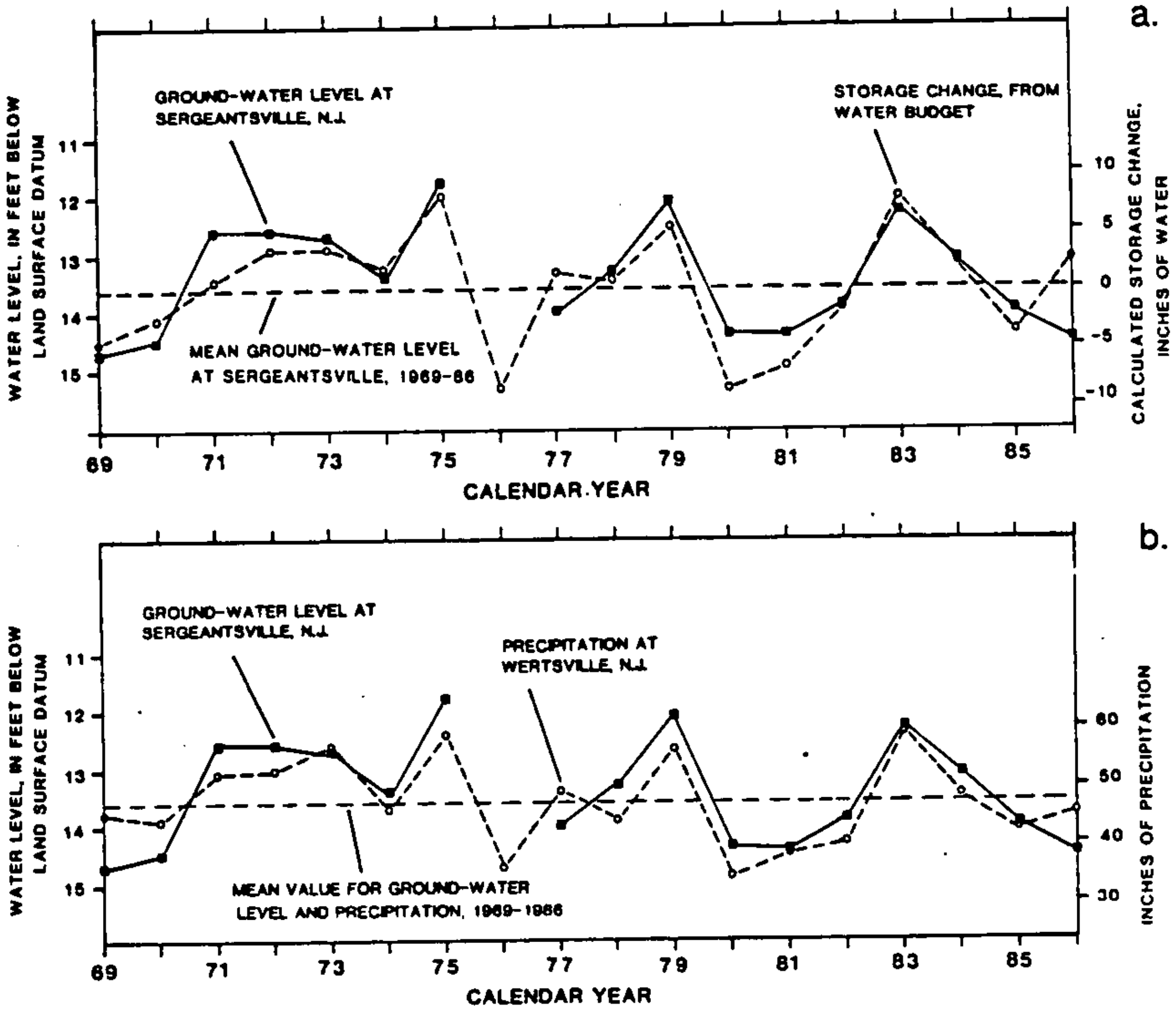


Figure 11. Mean annual ground-water levels Seargentsville, New Jersey observation well, plotted along with a) estimated annual storage changes from water budget (table 4), and b) annual precipitation at Wertsville, New Jersey. Ground-water levels before 1977 are based on periodic manual measurements; levels from 1977-1986 are based on average monthly values from recorder data (U.S. Geological Survey 1978-1987).

TABLE 4. - Annual water budget for the study area, 1969-1986. All data in inches of water except E ratio.

Calendar Year	Precip (A)	Stream Flow (B)	Pan E (C)	ET+dS (A-B) (D)	E ratio (C/A) (E)	Unadj ET (ExD) (F)	ETe (G)	dSe (D-G) (H)	dSe cumulat (I)
1969	45.77	18.41	42.3	27.36	0.92	25.3	32.0	-4.6	-4.6
1970	44.37	19.01	38.7	25.36	0.87	22.1	28.0	-2.6	-7.3
1971	52.59	32.10	40.1	20.49	0.76	15.6	19.8	0.7	-6.5
1972	53.23	30.91	35.7	22.32	0.67	15.0	18.9	3.4	-3.2
1973	57.40	30.48	39.6	26.92	0.69	18.6	23.5	3.4	0.3
1974	46.36	20.33	34.3	26.03	0.74	19.3	24.4	1.7	1.9
1975	59.20	31.28	33.4	27.92	0.56	15.8	19.9	8.0	9.9
1976	35.98	11.90	38.6	24.08	1.07	25.8	32.7	-8.6	1.3
1977	49.32	22.33	37.0	26.99	0.75	20.2	25.6	1.4	2.7
1978	44.00	23.90	33.7	20.10	0.77	15.4	19.5	0.6	3.3
1979	56.48	38.40	31.5	18.08	0.56	10.1	12.8	5.3	8.6
1980	34.39	11.59	37.4	22.80	1.09	24.8	31.4	-8.6	0.0
1981	38.14	11.24	37.6	26.90	0.99	26.5	33.6	-6.7	-6.6
1982	40.20	16.91	34.4	23.29	0.86	19.9	25.2	-1.9	-8.5
1983	59.02	33.78	32.5	25.24	0.55	13.9	17.6	7.7	-0.9
1984	48.44	26.12	34.6	22.32	0.71	15.9	20.2	2.1	1.3
1985	42.05	14.11	37.6	27.94	0.89	25.0	31.6	-3.7	-2.4
1986	45.00	22.86	31.7	22.14	0.70	15.6	19.7	2.4	0
Average	47.33	23.09	36.2	24.24		19.2	24.2		

(A) **Precipitation** - annual measured rainfall at Wertsville, New Jersey. Source: NOAA, National Weather Service, Summary of Climatic Data for New Jersey, 1969-1986.

(B) **Stream flow** - calendar-year total discharge in inches for the watershed of the Neshanic River above Reaville, NJ. Source: USGS, Water Resources Data for New Jersey, 1969-1985;also Robert Schopp, U.S. Geological Survey (oral communication, 1987).

(C) **Pan evaporation** - annual evaporation totaled from monthly data measured at New Brunswick, NJ. Source: Department of Meteorology, Cook College, Rutgers, the State University.

(D) **Total evapotranspiration plus storage changes** - calculated by subtracting stream discharge from precipitation. The average of this column is equal to average evapotranspiration only, since average storage change (dS) will be zero, unless the climate is changing.

(E) **Evaporation ratio** - pan evaporation divided by precipitation.

(F) **Unadjusted evapotranspiration** - evaporation ratio multiplied by total ET plus dS (column D).

(G) **Estimated evapotranspiration** - Values in column (F) are multiplied by a factor which makes the average of this column equal to the average of column (D). This causes estimated average storage change to equal zero. The factor is 24.2 divided by 19.2, or 1.27.

(H) **Estimated storage change** - Evapotranspiration plus storage change (column D) minus estimated evapotranspiration (column G) equals net estimated storage change.

(I) **Cumulative storage change** - Running total of annual storage changes sets of measurements were not taken under the same conditions, the data do suggest that water- level decline outside the severely impacted area was minimal. The conclusion is also supported by water levels in the Sergeantsville observation well, which were only slightly below average in 1982. Even if the specific yield of the Lockatong is ten times less than the Stockton Formation, water levels would be expected to be only 4 feet below normal in the study area in 1982, based on the observation well record.

The water-budget calculations can be used in at least one other way- to estimate partitioning of storage changes between soil moisture and aquifer storage. If a reasonable value for specific yield of the Stockton Formation is used, say 10 percent, the storage changes estimated from the water budget for the Neshanic watershed will produce close correlations with water level changes in the Sergeantsville well if a storage change ratio of 3:1 is used (soil-moisture to aquifer storage). This estimate is very rough, due to uncertainty in specific yield and because of differences in bedrock, topography, and soils between the Neshanic watershed and the observation well.

The water budget approach was an expedient method of estimating the magnitude of ground-water fluctuations produced by near- drought conditions of 1980-1982. The method is unproven, and was not verified experimentally for this study. The water budget appears to work reasonably well for the area investigated. If the method can be validated by further study it may have wider applicability.

Rate of Ground-Water Withdrawal

Withdrawal rate of ground water in the study area was estimated by assuming 100 gallons per day (gpd) use per person in each household, and dividing the total usage by the number of square miles in the area of interest. The use estimate is somewhat high, but has been used as a planning criterion in New Jersey (Hoffman and Canace, 1986). Withdrawal rates for each of four lots are given in units of gpd/mi^2 in table 5, to facilitate direct comparison with safe yield and recharge estimates in table 3. High withdrawal rates occurred in an area much smaller than one square mile; the high rates should not be interpreted as large pumping volumes.

TABLE 5. - Estimated ground-water-withdrawal rates in 1982 for part of study area where water level declines were measured.

Lot # (Block 171)	Lot size (acres)	Number of occupants	Estimated daily water usage(gpd)	Withdrawal rate (gpd/mi^2)
65	2.30	4	400	111,000
66	1.00	6	600*	384,000
67	1.84	4	400	140,000
117	3.12	2	200	41,000
4-lot total	8.26	16	1600	124,000

*Census at lot 66 indicated 9 occupants, therefore water use estimate is conservative.

Estimated water use on lot 66, with two deep wells in an area of 1.0 acre is 600 gallons per day. Extended to an area of one square mile the estimated withdrawal rate on lot 66 is equivalent to 380,000 gpd/mi^2. Lot 117 has the lowest estimated withdrawal, about 41,000 gpd/mi^2. Pumping and recovery levels in well 117F were not particularly low. The average for the 8.26- acre area in the table is approximately 124,000 gpd/mi^2, equivalent to about 2.6 inches of water per year. This exceeds published estimates of sustainable ground-water yield for the Lockatong Formation (table 3). During drought periods the ground-water deficit is increased by continued high rates of pumping. Water pumped from wells is removed from ground-water storage and is subject to evapotranspiration from septic fields or from lawns and other surfaces. Only a small proportion of water pumped from wells returns to the aquifer as recharge, particularly during hot and dry weather.

DOMESTIC WATER SYSTEMS

In the study area, the typical domestic water-supply system consists of a submersible pump of 1/2 to 3/4 horsepower and a 42- gallon compressed-air pressure tank. The above-ground storage capacity of these systems is approximately 20-25 gallons. No auxiliary storage tanks or separate pressure pumps were observed in any systems. One household had two independent supply systems of standard design, each hooked up to a separate well.

In wells of exceptionally low yield, the quantity of water available is little more than that stored in the borehole, because the rate of recovery is insignificant compared to pumping rates. Many wells in argillite and diabase bedrock are in this category. Measured recovery rates are as low as 4 gallons per hour, equivalent to less than 100 gallons per day (see well 67, figure 10). Even this quantity may be unavailable unless adequate storage capacity is present in a well bore of sufficient depth or in an auxiliary tank.

Wells 65 and 67 illustrate some of the problems resulting from low yields, diminished storage, and heavy pumping. The well at lot 65 is 140 feet deep, with the pump set at about 135 feet below land surface. The water level was observed to rise to about 112 feet below land surface overnight, during December, 1982. At that time the well contained only about 35 gallons of recoverable water. At lot 67 the pump setting is about 240 feet below land surface. The highest observed overnight recovery was to a level of about 190 feet. This is equivalent to about 75 gallons of water in the well. Because of localized decline in water levels during the period preceding this study, storage in wells was inadequate to supplement the low yields and provide a dependable supply.

CONCLUSIONS

Large water-level declines in the study area in 1982 resulted from two principal factors. These are:

1. Heavy pumping centered on wells 66R and 66F created a local cone of depression affecting these wells and nearby wells 65 and 67.
2. A 3-year period of below-normal precipitation from 1980- 1982 reduced recharge to the bedrock and probably lowered the regional water table. Regional decline of water levels by this cause was probably not more than a few feet in 1982.

The effect of local overpumping alone was probably sufficient to cause the water level declines observed. In addition to the two immediate causes, there were additional underlying causes which made wells in the study area susceptible to failure; these are:

1. Wells in the study area all tap argillite, a rock which has low capacity to store and transmit ground water. Soils that form on this bedrock have a high clay content and impede infiltration of water into the subsurface.
2. The study area is on high ground with moderately steep slopes that accelerate runoff and inhibit infiltration.

3. The affected area is distant from major fault zones. Bedrock in the area is probably not as highly fractured as it would be if it were closer to major faults.
4. Much of the recharge area for the aquifer is forested. Evapotranspiration rates are higher in areas of mature forest than in areas with crops, grasses, or bare ground (Johnston, 1970).

DISCUSSION AND RECOMMENDATIONS

Spacing of Wells and Lot Size

The cone of depression of the potentiometric surface around wells on lot 66 had a diameter of at least 360 feet, which is the distance between wells 65 and 67. The water-level decline in these 2 wells was at least 100 feet, indicating the cone extended beyond these wells. The area in which ground-water level declines were observed to be greater than 100 feet was about 3 acres. The total area of the first 3 lots in table 5, with severe water- level declines in 1982 is 5.14 acres.

Hillsborough Township had a minimum lot size requirement of 5 acres for the Sourland Mountain zone in 1982. Evenly spaced wells in square lots of 5 acres would be about 470 feet apart; spacing could be less for rectangular lots. This is more than double the spacing between wells with severe water level declines. Serious well interference is less likely to occur in areas of 5-acre zoning with similar geology, even under extreme circumstances as observed in the study area. Serious interference can still occur in areas of older development with more closely spaced wells.

Widely spaced wells are less likely to fail than closely spaced ones, other factors being equal, particularly if they have a substantial depth of uncased hole and pump at low or moderate rates. Some wells will continue to be subject to failure, especially during protracted droughts, if they are shallow, closely spaced, or are heavily pumped.

Depth of Wells

Well failure on lot 65 would have been less likely to occur if the well had been at least 100 feet deeper, or more than 235 feet deep, with a corresponding increase in depth of pump setting. Deepening of well 67 by the same amount, to more than 340 feet, probably would have enabled it to continue yielding some water, even if it were only from well storage, under the prevailing conditions.

Recent drilling practice in the Sourland Mountain area is to construct wells about 300 feet deep, or deeper where yields are very low in order to increase storage capacity in the well. Shallower wells generally date from the early 1960s or earlier; this trend is partly due a change in drilling methods from cable- tool to air-rotary.

Roadway Drainage

An extensive roadway drainage system was installed along the south side of David's Lane in 1975, uphill from the impacted wells. Open-joint cement pipe drains were buried near the top of bedrock, in the most permeable shallow subsurface zone. The system directs storm runoff down Montgomery Road. The alignment of David's Lane roughly perpendicular to the slope of the mountain may increase the tendency of the drainage system to collect runoff water and intercept water in the shallow zone above unweathered bedrock which is periodically saturated. It is possible that the drains could either reduce or increase recharge under varying flow conditions. Available data are insufficient to assess the role of the drain, if any, in well failures.

Because of the possibility that storm drains could have an adverse impact on recharge, it is suggested that roadway drainage design be carefully assessed, in areas of extremely low-yield wells such as the summit and steeply-sloped parts of Sourland Mountain.

Possible Remedies for Well Failures in the Study Area

Deepening of Wells

The most expedient remedy for the situation in Hillsborough Township is to deepen the affected wells. This solution does not address the question of long-term pumping in excess of recharge, however. This study suggests that ground-water pumpage may locally exceed the average rate of recharge. Unless recharge is increased or pumpage rate decreased, the ground water deficit may continue to accumulate. In that case deepening of wells would only be a temporary remedy.

Recharge Augmentation

Recharge basins have been used in some areas, particularly Long Island , New York, to dispose of storm runoff (Seaburn, 1970). These basins are mostly located in areas underlain by highly permeable sand and gravel deposits. It is unlikely that open recharge basins would be practical in the study area, because of low bedrock permeability, steep slopes, and thin soil. Slow infiltration would require extremely large basins to handle peak flows. Under these conditions artificial recharge basins modeled after infiltration beds for septic systems might be designed to accept part of storm runoff. More study would be required to evaluate the feasibility of such a system.

Alternative Water Supply

A public-water supply for the Sourland Mountain area would be a reliable but expensive solution to well failures. A public- supply well could be drilled in the Brunswick siltstone north of the study area and water could be pumped uphill to a standpipe on the highest part of Sourland Mountain. Such a system would also serve fire-fighting purposes. The high cost of a community system is the main obstacle to its implementation. Public funding sources would have to be sought, because housing density in the area is probably too low to attract a commercial water purveyor.

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