Ground-Water Resources of Orange and Ulster Counties, New York

GEOLOGICAL SURVEY WATER-SUPPLY PAPER 1985

Prepared in cooperation with the New York State Conservation Department

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By MICHAEL H. FRIMPTER

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GROUND-WATER RESOURCES OF ORANGE AND ULSTER COUNTIES, NEW YORK

By Michael H. Frimpter

ABSTRACT

The ground-water resources of Orange and Ulster Counties, in southeastern New York, occur in Pleistocene sand and gravel deposits of glacial origin and in Precambrian and Paleozoic consolidated rocks.

The sand and gravel aquifers generally are small and restricted to the valley areas, but some can be recharged rapidly and, therefore, yield large quantities of water. The aquifer in the valleys of the Neversink River and Basher Kill between Port Jervis and Summitville, with an estimated ground-water yield of about 100 mgd (million gallons per day) could be the most productive aquifer in southeastern New York exclusive of Long Island. An additional 20 mgd might be produced from similar deposits between Phillipsport and Wawarsing. Numerous smaller sand and gravel aquifers are scattered throughout the valleys of the two counties. Major parts of the two larger stream valleys, the Wallkill River and Esopus Creek valleys, do not contain sand and gravel aquifers but are filled with relatively impermeable clay and silt.

The consolidated rock has secondary porosity and is a dependable aquifer for small domestic supplies, but not for municipal and industrial supplies. Notable exceptions are the brecciated fault zones in southeastern Orange County and in the carbonate rocks deep in the Rondout Creek and Sandburg Creek valleys.

Iron, manganese, and hydrogen sulfide are the most common and troublesome chemical pollutants of the ground water in Orange and Ulster Counties but the water is generally of good chemical quality for public water supplies.

INTRODUCTION

The basic water-supply problem facing Orange and Ulster Counties is the development of additional water sources capable of supporting the counties' rapidly increasing population and industrial activity. Of prime importance to the solution of this problem is the location and appraisal of the ground-water and surface-water resources of the two counties.

The high yield aquifers of the two-county area are not evenly distributed, a situation which creates many of the problems faced by water-supply planners. A difficult problem that must be solved for the water-supply planner is the determination of the potential yields of the aquifers once they have been located.

Rural homes and farms throughout New York State rely on wells for dependable supplies of good quality water. Adequate domestic water supplies may be obtained from wells drilled almost anywhere in Orange and Ulster Counties; for the owner of a rural home, water-supply cost is a question of well depth.

PURPOSE AND SCOPE

This report presents the results of an investigation of the ground-water resources of Orange and Ulster Counties, N.Y., and is part of a continuing program of hydrologic investigations made in cooperation with the Water Resources Commission of the State of New York. The purpose of these studies is to provide the data needed for planning the development of the natural water resources of the State.

The aquifers discussed in this report were located and their potential yields determined through detailed geologic and hydrologic study. Greatest emphasis was placed on delineating and evaluating those areas where municipal and industrial supplies might be developed. The question of how deep to drill a well was explored by geologic investigation and by statistical analysis of data from existing wells. Water-quality problems, particularly that of hydrogen sulfide content, also were investigated.

An inventory of over 1,700 wells and springs was made during the course of study. Lithologic and hydrologic data were collected from more than 470 test borings, and chemical analyses were made of 128 samples of the ground water. Water-level fluctuations were recorded continuously in four wells, and periodic measurements were made in three additional wells. Rock and soil exposures throughout the two-county area, as well as drill cores and cuttings, were examined for geologic and hydrologic information. Records of the wells and springs, lithologic logs, and chemical analyses of water samples collected for this study are reported by Frimpter (1970) in "Ground-water basic data, Orange and Ulster Counties, New York."

ACKNOWLEDGMENTS

This report was prepared in cooperation with the New York State Conservation Department, Division of Water Resources, F. W. Montanari, Director, and under the supervision of Ralph C. Heath and Garald G. Parker, successive District Chiefs, U.S. Geological Survey, Albany, N.Y.

The many residents of Orange and Ulster Counties who supplied information on their wells and granted permission to measure water levels and the superintendents of municipal water systems and other public officials who supplied test data and records have contributed to the success of this study. The following well drillers gave freely from their valuable records and experience: Roy DeWitt, Sam DeWitt, David Tompkins, Bruce Tompkins, Andrew Wild, Ray Gillespie, William Diegel and James Eckerson. C. W. Lauman and Co., Inc.; Metcalf and Eddy, Engineers; Hazen and Sawyer, Engineers; Eustance and Horowitz, Engineers; and Artersian Well and Equipment Co., Inc., supplied well data. The New York City Board of Water Supply, the Port of New York Authority, and the New York State Department of Transportation supplied lithologic logs from test borings. Officials of the New York State Health Department supplied lists of public water supplies and their superintendents.

LOCATION AND PHYSICAL FEATURES OF THE AREA

Orange and Ulster Counties are located in the southeastern corner of New York (fig. 1) and are bounded by the Hudson River and Rockland County on the east and by the states of Pennsylvania and New Jersey on the south. Ulster County lies immediately north of Orange

FIGURE 1.—Location of the study area.

County and the two are bounded on the west and north by Sullivan, Delaware, and Greene Counties. The area of Orange County is 834 square miles and the area of Ulster County is 1,137 square miles. The two-county area has been divided into three physiographic provinces (fig. 2): the New England, the Valley and Ridge, and the Appalachian Plateaus (Fenneman, 1938). Variations of bedrock lithology are responsible for the development of these distinct physiographic provinces and are shown on plate 1. The southeastern boundary of Orange County lies in the New England province and is mostly a forested area. This part of the New England province is known as the Hudson Highlands and consists mainly of a State-owned park and the Federal military reservation at West Point. Altitudes in the Highlands range

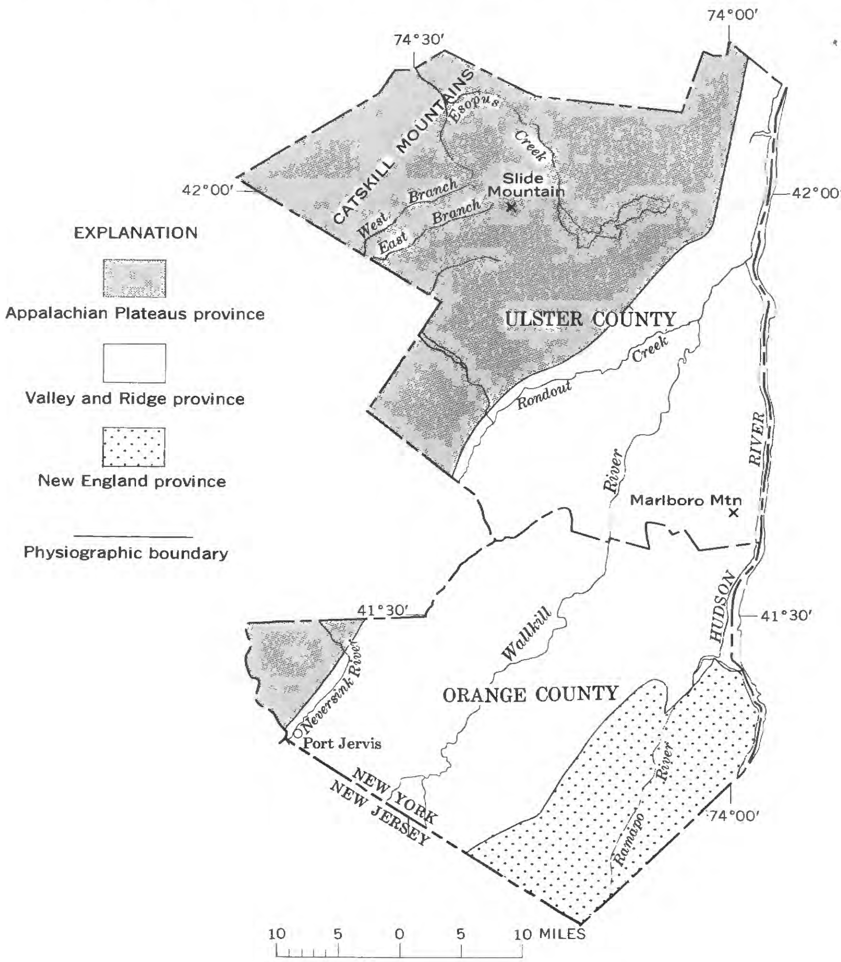

FIGURE 2.—Physiographic provinces.

from near sea level at the Hudson River to about 1,400 feet above mean sea level, and the terrain is composed of erosion-resistant granite and other crystalline rocks. Woodbury Creek and the Ramapo River, as well as numerous mountain brooks, drain this area and add to its scenic beauty.

North of the New England province, the Valley and Ridge province has a low rolling relief. Most of this land has been cleared for farms and orchards; only the more rugged and less tillable areas, such as Shawangunk and Marlboro Mountains, remain forested and sparsely populated. The Valley and Ridge province is underlain by alternating layers of hard sandstone and soft shale that were compressed and crumpled by pressure exerted from the southeast. The long axes of the wrinkle-like folds in the crumpled rock layers consequently trend northeast-southwest. Different rates of erosion of these tilted layers of hard and soft rock give rise to the sequence of narrow ridges and valleys typical of this province. Most of the low lying area of the province is drained northward by the Wallkill River. Along the northwest boundary of the province, the Neversink River, Rondout Creek and Esopus Creek drain narrow valleys developed over soluble limestone and dolostone bedrock. The valleys of these three streams form a trough extending from the southwest corner of Orange County at Port Pervis to the northeast corner of Ulster County.

The Appalachian Plateaus province lies in the western and northern parts of the study area and is a rugged, forested area that is sparsely populated. The plateau in Orange County, northwest of Port Jervis, is generally an area of elevated flatland and is underlain by nearly horizontal beds of sandstone and shale where a few streams have incised steep-walled valleys. In northwestern Ulster County, however, the sandstone and shale are thicker and the land rises higher than the rest of the plateau. In this area streams have eroded the plateau to a greater extent and no flatland lies between the steep-walled stream valleys. The uneroded remnants of the plateau form the high sharp-peaked Catskill Mountains. The peak of Slide Mountain, 4,204 feet above sea level, is the highest in the Catskill Mountains. Although the Catskill Mountain area is generally not suitable for farming or industry, it has enormous potential for recreation.

SURFACE-WATER RESOURCES

The largest single source of water in Orange and Ulster Counties is the Hudson River. The mean annual discharge at Poughkeepsie for the period 1947 to 1965 was 17,700 cfs (cubic feet per second) and the lowest mean monthly discharge was 3,030 cfs occurring in September 1964 (Giese and Barr, 1967, p. 14–15). During periods of low discharge, water slightly contaminated by sea water reaches as far upstream as Poughkeepsie. In 1968 the U.S. Geological Survey studied the discharge and salinity of the Hudson River Estuary in detail.

The headwaters of the major streams that drain the Appalachian Plateaus province and cross the two counties have been tapped for water supplies for the New York City. The drainage basins of Rondout Creek, Esopus Creek, Neversink River, and the Delaware River contain reservoirs that are part of the New York City water supply system. The export of water from these drainage basins represents export of most of the surface-water resources of the two counties, exclusive of the Hudson River. The only remaining large stream is the Wallkill River. The main stem of this river does not contain many sites suitable for reservoirs. The average discharge for the Wallkill River's 711 square mile drainage area above Gardiner is 1,029 cfs, which represents an average yield of about 1.45 cfs per square mile. All the cities of Orange and Ulster Counties obtain their water from reservoirs on streams with rather small drainage areas.

WELL, SPRING, AND BORING LOCATION SYSTEM

The wells, springs, and borings cited in this report are numbered on the basis of a latitude and longitude system. The area has been divided into 1-minute quadrangles with a latitude and longitude grid. Each well, spring, or boring is designated by a composite of two location numbers and a sequential number or letter. The first number indicates the latitude and the second number indicates the longitude of the southeast corner of the grid in which the well is located. The first digits of latitude and longitude remain the same throughout the two-county area and therefore are omitted from the identification numbers. The sequential numbers identify wells and sequential letters identify test borings within a 1-minute quadrangle. Thus, in figure 3, well 125–424–2 is bounded by latitudes 41°25′ and 41°26′ and longitudes 74°24′ and 74°25′ and is the second well inventoried within these boundaries. Springs are identified by the same numbering system used for wells, except that the suffix "Sp" is added, as in 115–410–1Sp.

ORIGIN AND OCCURRENCE OF GROUND WATER

Nearly all the ground water in Orange and Ulster Counties is derived from precipitation that falls on these counties. The part of the rain that falls on the ground and does not evaporate will either seep into the soil or run into streams and lakes.

During the growing season large quantities of soil moisture are transpired by trees, crops, and other vegetation. Although eight of

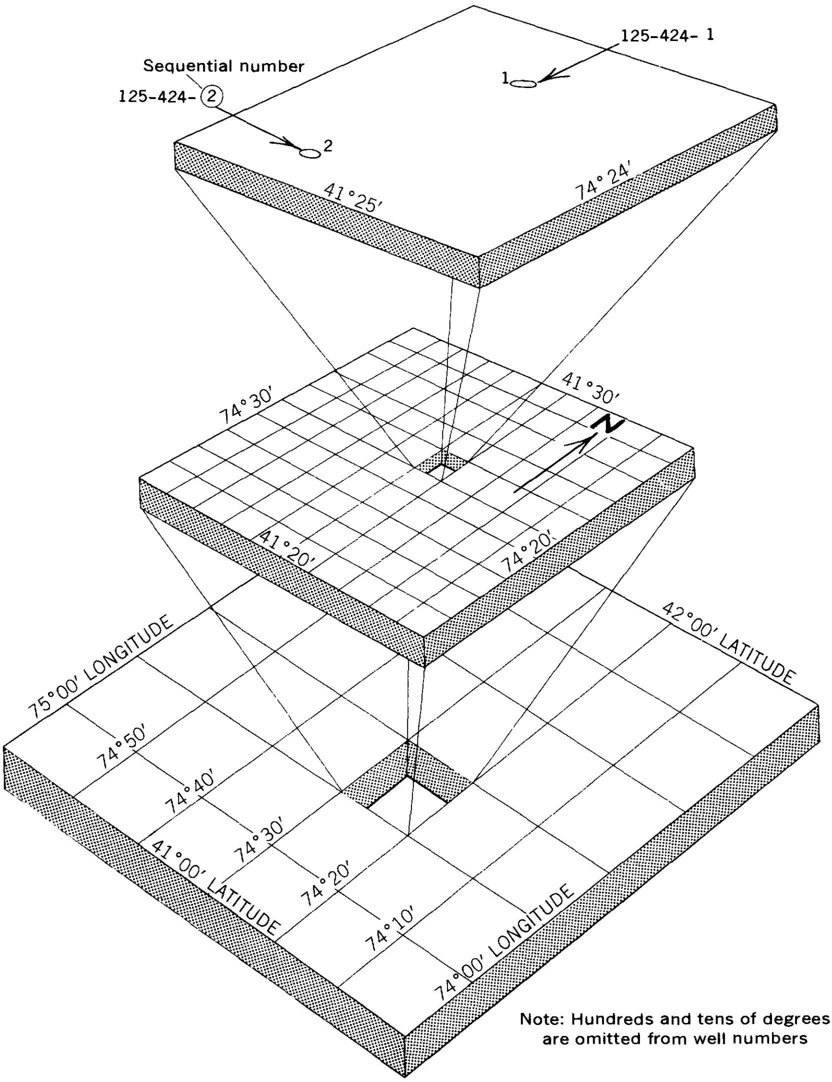

FIGURE 3.—Well-location system.

the 16 U.S. Weather Bureau stations in Orange and Ulster Counties receive their greatest monthly precipitation in July or August (U.S. Weather Bureau, 1964), these are months of low streamflow and ground-water levels. The quantity of water released to the atmosphere by a hardwood forest in eastern United States is 20 to 30 inches of water per year (Lull, 1964, p. 6–22), or more than half of the precipitation that falls in most of the study area. In summary, between one-half and three-quarters of the 40 to 50 inches of precipitation falling on the counties is lost through evaporation and transpiration; most of the rest flows to the sea through the streams and rivers; only a small percentage is diverted through the ground before being discharged to the streams. For more detailed information on precipitation runoff and water lost by evapotranspiration, the reader is referred to Knox and Nordenson (1955) and to Hely, Nordenson, and others (1961).

Ground water is water below the land surface that saturates openings in rock. These openings include the open fractures, cavities, or intergranular spaces that are present in almost all rocks of the earth's surface. Aquifers (water-bearing rock reservoirs) may be divided into two groups, those in which the openings are fractures or solution cavities and those in which they are intergranular spaces. In the study area, the consolidated rocks are aquifers of the first type and the unconsolidated rocks are aquifers of the second type.

UNCONSOLIDATED AQUIFERS

Sand and gravel deposits are the best sources of large quantities of ground water in Orange and Ulster Counties. How a sand deposit might become a source of water may be best explained by a simple example: children playing at the beach dig holes in the sand and marvel at the fact that water appears in the holes. The water drains from spaces between the sand grains which surround the hole. In wells, a strainer or screen is placed in the hole to prevent the sand and gravel from caving in and filling the well, while still allowing the free inflow of water.

The porosity of a material is the percentage of its total volume that is open space. It is an important hydraulic property because it determines how much water a material can hold. For example, 1 cubic foot of coarse sand having a porosity of 40 percent can hold about 3 gallons of water.

The specific yield of any material is the volume of water that will drain from a sample by gravity divided by the gross volume of the sample. When the water is allowed to drain, some of it is retained as a thin film wetting the individual grains or cracks. Specific yield is therefore always somewhat less than total porosity.

The specific yield varies considerably with variation in grain size. Figure 4 shows idealized wetted grains of various sizes. The thickness of the film of water that adheres to or wets the grains and will not drain from the grains remains the same regardless of the size of the grains. As the grain size decreases, the amount of water that will drain from between the grains decreases because a greater percentage of the water adheres to the grains. For this reason silt may have a low spe-

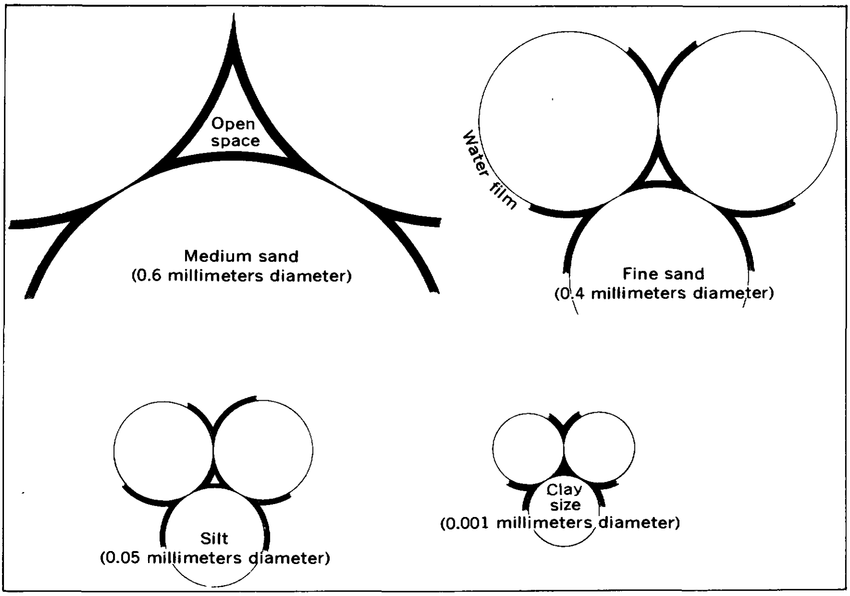

FIGURE 4.—Idealized microscopic view of wet sediments of various grain sizes.

cific yield, although it is just as porous as a high-yield coarse sand.

The degree of sorting in a sand and gravel deposit can also have a great effect on porosity and specific yield. Imagine a gravel aquifer in which all the stones are about 1 inch in diameter. All the space between these grains could be filled with water. However, if sand were mixed with this gravel, the grains of sand would occupy much of the space which water might have filled. As a result, the porosity of the deposit, and therefore the specific yield, would be reduced by the mixing of grains of different sizes. An example of poorly sorted sediment is glacial till (hardpan), which generally contains grains ranging in size from boulders to clay. The porosity of till is lower than the porosity of sand and gravel because the pore space between the larger grains is filled with smaller grains.

CONSOLIDATED AQUIFERS

Water occurs in consolidated deposits (bedrock) somewhat differently than it does in unconsolidated deposits. The bedrock of Orange and Ulster Counties has been subjected to high compression and in some cases to high temperatures. As a result, the grains of the rock have been either squeezed together or cemented together so completely that there are no spaces left between the grains.

The only space for water in the consolidated rocks is in thin tabular fractures where the rock has been broken. Most fractures result indirectly from the pressure and heat which destroyed the original porosity of the consolidated rocks. When the pressure that consolidated the rock was removed through erosion and earth movement, the rock expanded and cracked.

Detailed geologic studies in many areas have revealed that natural rock fratcures generally become scarcer with increasing depth. The porosity of consolidated rock, therefore, also decreases with increasing depth. For this reason, more water is available near the surface than at greater depth. A qualitative representation of this variation is shown in figure 5.

Fractures are also formed where the rock of the earth's crust has broken or faulted into large blocks that have moved in relation to each other. These fracture zones, or fault zones, may contain many more fractures and, therefore, a much larger amount of water than the surrounding rock. Fault zones sometimes contain claylike ground-up rock that will, unfortunately, make well water muddy or turbid. Occasionally, this pulverized rock may, by sealing a fault zone, form a barrier to the movement of water.

MOVEMENT OF GROUND WATER

The storage capacity of a ground-water reservoir, as determined by porosity and specific yield, only measures a part of the reservoir's effectiveness; just as important is the reservoir's ability to transmit

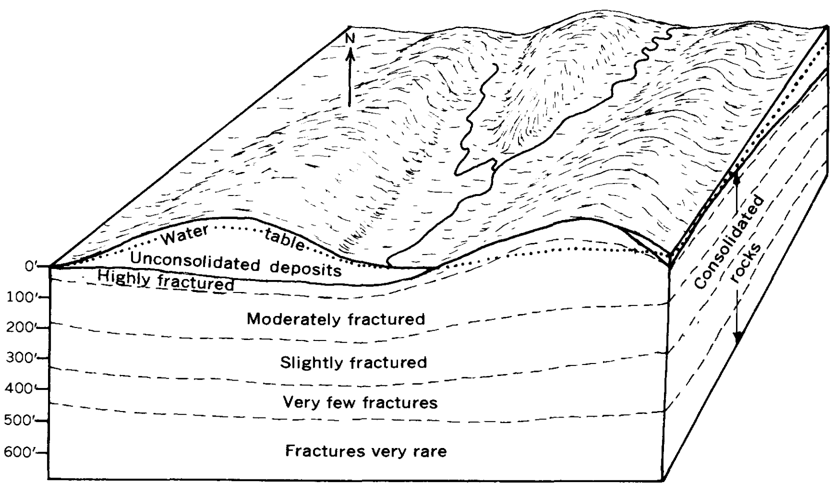

FIGURE 5.—A typical section of the Hudson River lowlands showing the degree of fracturing in bedrock at various depths.

water. Permeability is a measure of the capability of rock to transmit water and is expressed as the gallons of water at 60°F that will flow through a cross-sectional area of 1 square foot under a hydraulic gradient of 100 percent in 1 day.

Because water must travel through the rock's pore space, the permeability is largely dependent on the rock porosity and the interconnection between the pores. Water will move more slowly through rock with smaller pore spaces because more friction is developed. A coarse sand is more permeable than a fine sand, and clay is practically impermeable.

The permeability of unconsolidated rocks varies through a wide range. The following table shows the approximate upper limit of permeabilities found for laboratory samples of different size sedimentary materials (Todd, 1959, p. 53). These permeabilities are presented

Sediment	*Permeability (gallons per day per square foot)*
Gravel	100,000
Sand	1,000
Very fine sand	10
Clay	.01

here to show the variations and orders of magnitude of permeability and are not intended to be a guide to classifying the transmissive properties of aquifers.

The permeability of consolidated rock depends on the degree of interconnection of fractures as well as on the size and number of fractures. For example, during the drilling of a well, openings may be penetrated that initially yield a few gallons per minute. After only a few minutes of discharge, however, the yield may drop considerably or even stop altogether. Such instances indicate the presence of cavities or fractures which are not interconnected or are only poorly interconnected. Drillers commonly refer to these as "pockets" or "blind veins." The lack of interconnection of fractures in rock is demonstrated when the pumping of a well fails to cause a lowering of the water level in a nearby well tapping the same aquifer.

It is important to realize that rock permeability due to fracturing is not a homogeneous feature. The degree of fracturing is extremely variable, even within very small areas. Wells that are only a few feet apart and that tap the same rock aquifer can have vastly different yields and can sometimes exhibit different static water levels.

The variation of permeability of bedrock in the study area can be seen by comparing three neighboring wells (131–419–1, –4, and –5). These wells are no more than 700 feet apart and are located in a similar topographic and geologic setting. All three wells were drilled for

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domestic supplies by the same methods, by the same driller, and for the same builder. Well 131–419–1 penetrated 14 feet of rock for each gallon per minute of yield, while the same ratios for wells 131–419–4 and –5 were 38 and 0.8, respectively.

INTERRELATION BETWEEN SURFACE WATER AND GROUND WATER

The water table may be defined simply as a surface below which all open rock spaces are filled with water. Springs, lakes, ponds, or streams exist where the water table is higher than the the land surface. Streams traversing such areas continue to flow when there hasn't been any rain, because water flows into the streams from the saturated rocks and soil adjacent to the streambeds.

Exceptions to this definition of water table exist when aquifers or other water-saturated rocks are perched above unsaturated rock because of intervening materials of low permeability. Perched water table aquifers, however, are not significant sources of water in the study area.

The interrelation of ground and surface water can be shown in another way. Pumping large quantities of water from the ground lowers the water table and can cause the water levels of nearby lakes and streams to decline. For example, when test well 120–17–7 was pumped at 350 gpm (gallons per minute) for 23 hours, a water-level decline of half a foot was noted in a pond about 110 feet from the well. Also, a 30-day pumping test of three wells in the Woodbury well field was largely responsible for a water-level decline of 16 inches in a pond about 0.4 mile north of the wells.

The interrelationship between ground and surface water also works the other way: draining a lake may lower water levels in nearby wells, as shown in figure 6*A* and *B*. Wells are sometimes drilled near a lake or stream in order to take advantage of the interconnection between water above and below land surface. By withdrawing water from wells located in an aquifer near a stream or lake, water may be induced to infiltrate and recharge the aquifer. Figure 6 shows the path of water circulation in an aquifer which is in contact with a lake when wells adjacent to the lake are pumped.

Pumping wells to add water to adjacent surface reservoirs may be wasted energy (fig. 6*C*). Ground water and surface water in the project area are closely related, and development or control of one requires consideration of the effect on the other.

In many instances aquifers may become contaminated by recharge from polluted surface water. Commonly a village is built over its own aquifer (ground-water reservoir), and water from street gutters,

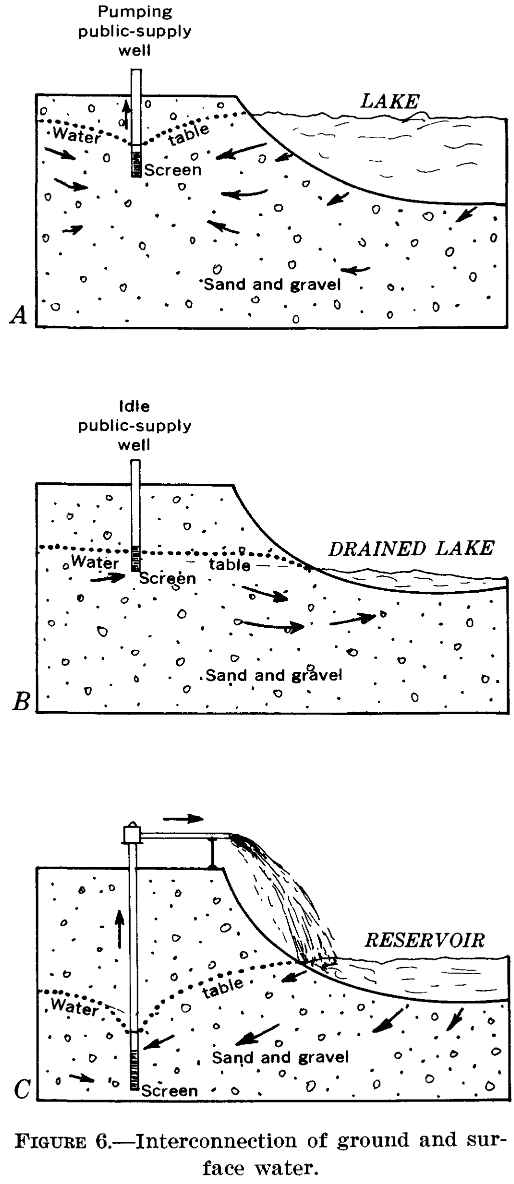


FIGURE 6.—Interconnection of ground and surface water.

disposal pits, septic tanks, leaky sewer mains, or accidental spillage may recharge and pollute the aquifer (Parker and others, 1967, p. 202–204).

Alteration of the courses of streams can have a great effect on the rate of recharge to an aquifer. Enclosing streams in culvert pipe or lining their banks with concrete would stop recharge from these sources. Drainage ditches and storm sewers rapidly transport runoff from recharge areas, thereby reducing the time available for downward percolation of water. Pavement, such as for parking lots, makes an effective impermeable seal over the earth surface, speeds surface runoff and, therefore, reduces ground-water recharge.

As the development of Orange and Ulster Counties proceeds, recharge rates to aquifers will be altered. To maintain the yields of aquifers, control of construction on recharge areas may be necessary. If, for instance, housing, storm sewers, and pavement are added to a recharge area, some compensatory measure must be taken to maintain the preconstruction recharge rate to the aquifer. Perhaps the storm sewers could terminate in recharge basins on the recharge area. Streams could be rerouted over a greater part of the recharge area by including recharge ponds along their courses; in this manner the recharge to an aquifer could be increased. In some instances the removal of vegetative cover would reduce transpiration losses and thereby increase recharge. On steep recharge slopes, however, the removal of vegetation might increase the rate of runoff, decrease the recharge rate, and cause erosion.

The effect of each new construction in the recharge areas of aquifers should be considered when the construction is planned and appropriate measures taken to insure equal or increased recharge after the construction is completed. In the discussion of major aquifers in this report, the recharge and hydrology of each aquifer is presented and can be used as an aid in the solution of specific recharge problems.

WATER LEVELS

Water wells can draw their supply only from the saturated zone in the ground. The top of this zone, the water table, represents the minimum depth to any underground water source. This minimum depth is a very important factor to be considered in the development of a water supply. Fortunately, the relatively humid climate in New York State keeps the water table at a shallow depth. The depth to the water table in Orange and Ulster Counties rarely exceeds 100 feet and generally is only a few tens of feet or less below the land surface.

RELATION OF WATER TABLE TO TOPOGRAPHY

The water table generally is a subdued expression of the land surface (fig. 7). The elevation of the water table tends to parallel the land elevation because precipitation, the source of ground-water recharge, falls fairly uniformly over the land surface. Because the flow paths of water through the ground are restricted and intricate, the flattening of the water table surface is somewhat retarded. Figure 7 shows that

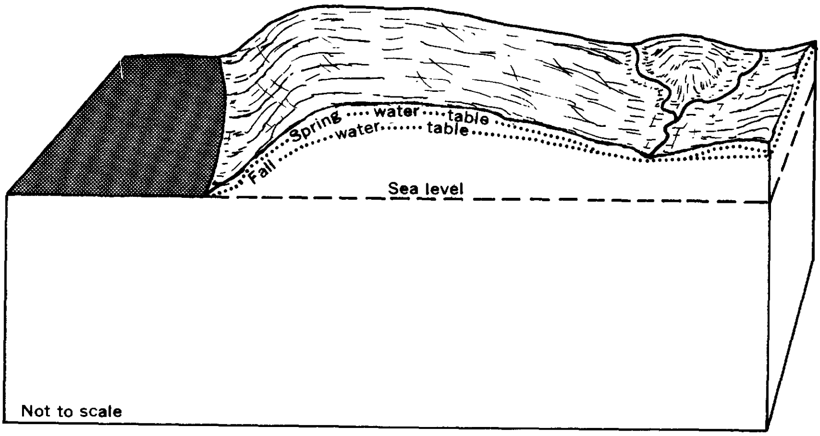


FIGURE 7.—Seasonal relationship between the water table and land surface.

the water table is flatter than the land surface; note the greater depth to the water table near the edge of the cliff. Depths to the water table greater than 100 feet are rare in the two-county area, and most of those recorded are found near the edge of a steep hill or cliff.

Because the shape of the water table does conform closely to the shape of the land surface and is rarely more than 100 feet below it, water-table maps are not included in this report. Water levels for specific locations can be estimated from the well tables and location maps in "Ground-water basic data, Orange and Ulster Counties, New York."

ARTESIAN CONDITIONS

Artesian aquifers occur where ground water is confined by overlying impermeable or semipermeable materials. The static water level in a well penetrating an artesian aquifer is higher than the top of the aquifer. An imaginary surface coinciding with the level to which water would rise in wells tapping an artesian aquifer is called the potentiometric surface. When the potentiometric surface is higher than the land surface, the water pressure is great enough to cause a well to flow. In figure 8 the water-level fluctuations of an artesian well, well 120–413–1, are shown. The artesian aquifer in Woodbury Creek valley is recharged on the western side of the valley at an elevation greater than the tops of the flowing wells (fig. 26). The Warwick Village supply wells are completed in an artesian aquifer. Artesian wells would also be expected in the southern part of the Wallkill valley where organic silt, silt, and clay overlie sand and gravel.

SEASONAL WATER-LEVEL FLUCTUATIONS

The level of the water table depends on the balance between, recharge to, and discharge from aquifers. During prolonged periods of below-normal precipitation, recharge to aquifers decreases, and the water table becomes flatter and lower. Shallow wells and wells on hilltops are most severely affected by drought. Figure 7 shows hydrographs of water levels for two seasons; the higher level occurs in the spring after a period of recharge from snowmelt and the lower level occurs in the fall after a period of slight recharge and high evapotranspiration losses.

Hydrographs of water levels observed in five wells located along a 2-mile reach of Seeley Brook are shown in figure 8. A location map for these wells is shown in figure 9. Generally, the water-level fluctua-

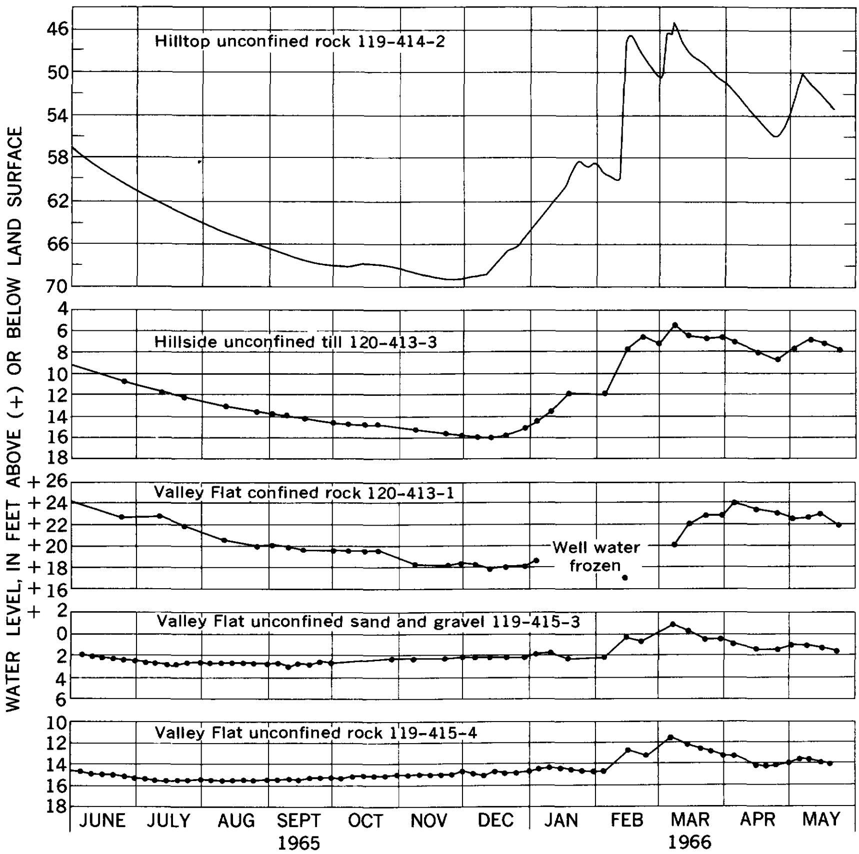


FIGURE 8.—Hydrographs of five wells in the Seeley Brook basin.

tions in these wells are similar, but they vary in amplitude. Well 119–414–2 is located on a hilltop and exhibits a seasonal water-level fluctuation of about 24 feet. During the same period of record, well 119–415–4, located on the valley floor, exhibited a water-level fluctuation of only about 4 feet.The topographic situation is an important factor determining ground-water levels and their range of fluctuation. Wells 119–415–3 and –4 tap different aquifers, but are in similar topographic situations, and therefore, exhibit almost identical water-level fluctuations.

For a comparison between streamflow and fluctuations in ground-water levels, a hydrograph of the discharge of Seeley Brook, which drains the basin in which these wells are located, is given in figure 10. Streamflow is erratic and highly dependent on both precipitation and the melt water from ice and snow. However, the streamflow hydrograph does show the same general trend as the ground-water hydrograph. Low streamflow is sustained by water seeping from the ground.

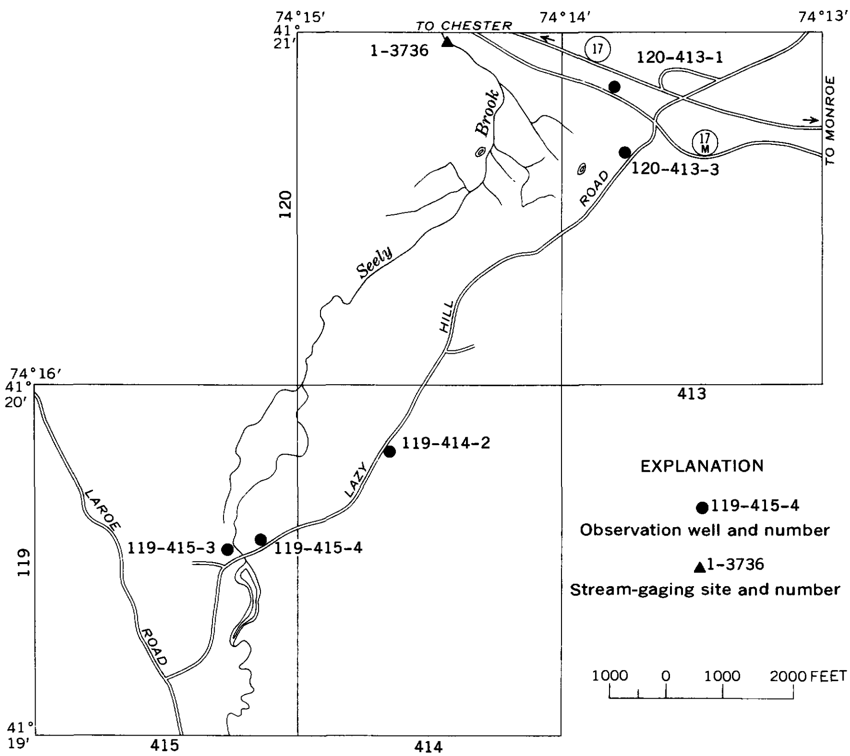


FIGURE 9.—Locations of observation wells.

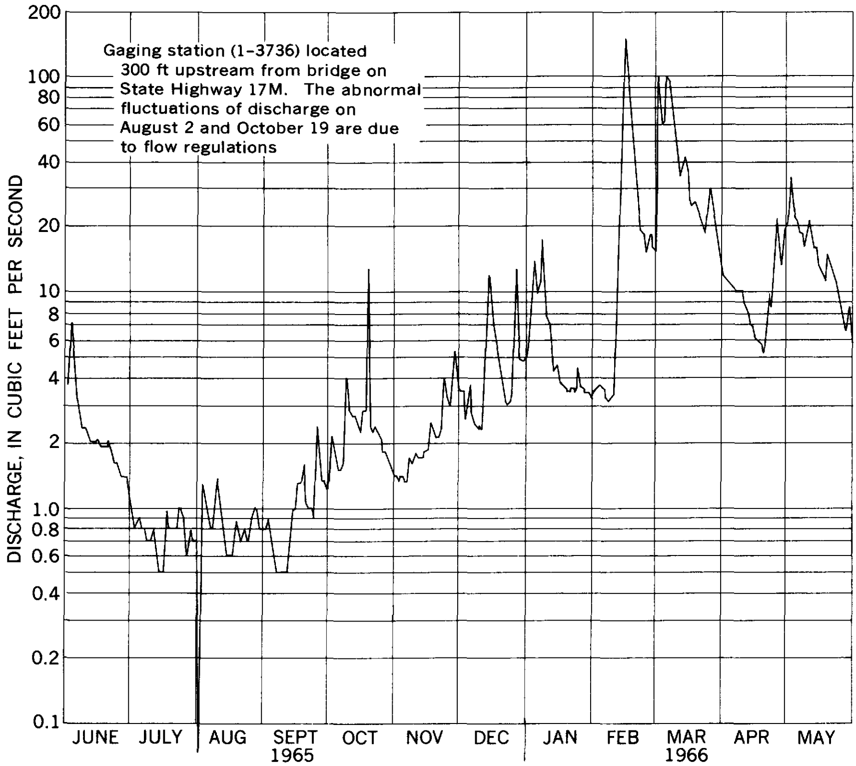


FIGURE 10.—Discharge of Seeley Brook.

The high flows registered in the middle of February and the beginning of March are due to the melting of snow and ice that had accumulated on the ground during the winter. The greatest recharge to the ground-water reservoirs occurs during these snow-melt periods.

LONG-TERM WATER-LEVEL TRENDS

Records of the water level have been obtained from well 144–421–3 at Napanoch in the valley of Rondout Creek by the U.S. Geological Survey for 13 years. These records and the record of precipitation at Ellenville (U.S. Weather Bureau, 1952–1966), 2 miles away, are presented in figures 11 and 12. The correlation between lower water levels and precipitation shown in figure 12 is quite obvious. The water level reached a low in 1964 when precipitation was low; the highest levels occurred in 1955 when two hurricanes were responsible for a total annual precipitation of about 10 inches above normal. The downward trend of the water level during the drought of the early sixties

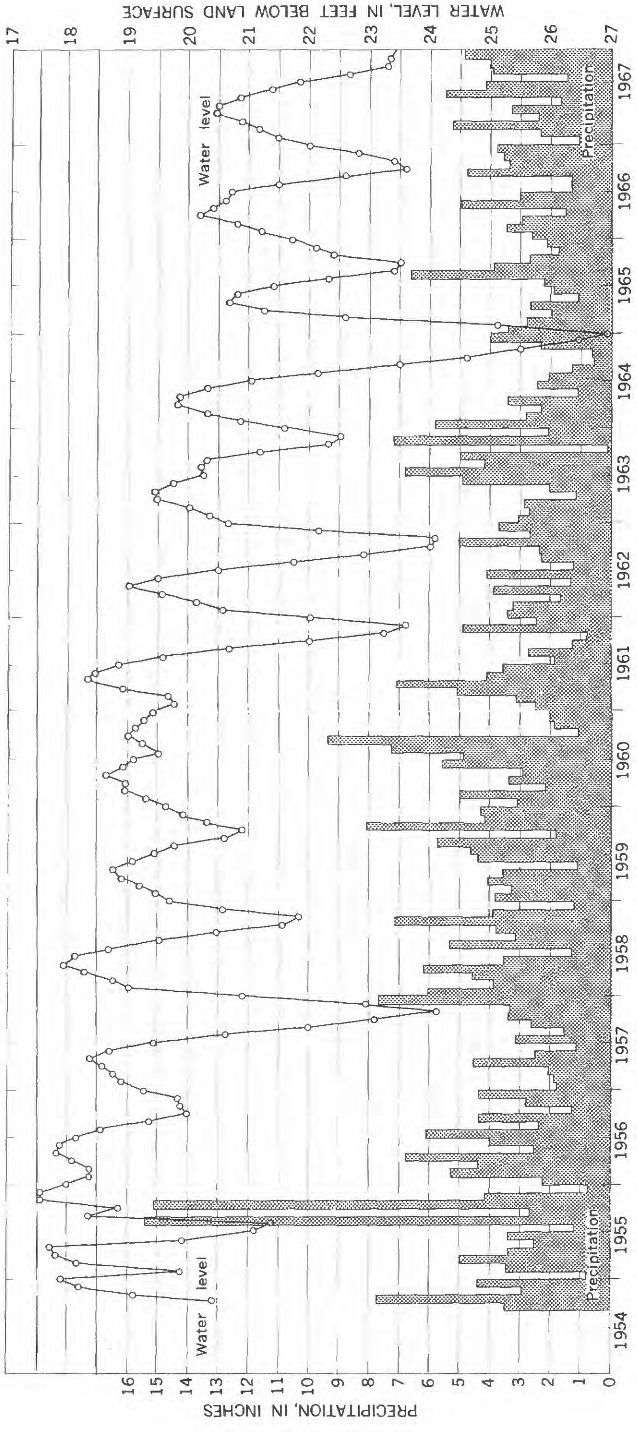


FIGURE 11.—Monthly ground-water levels at Napanoch and precipitation at Ellenville.

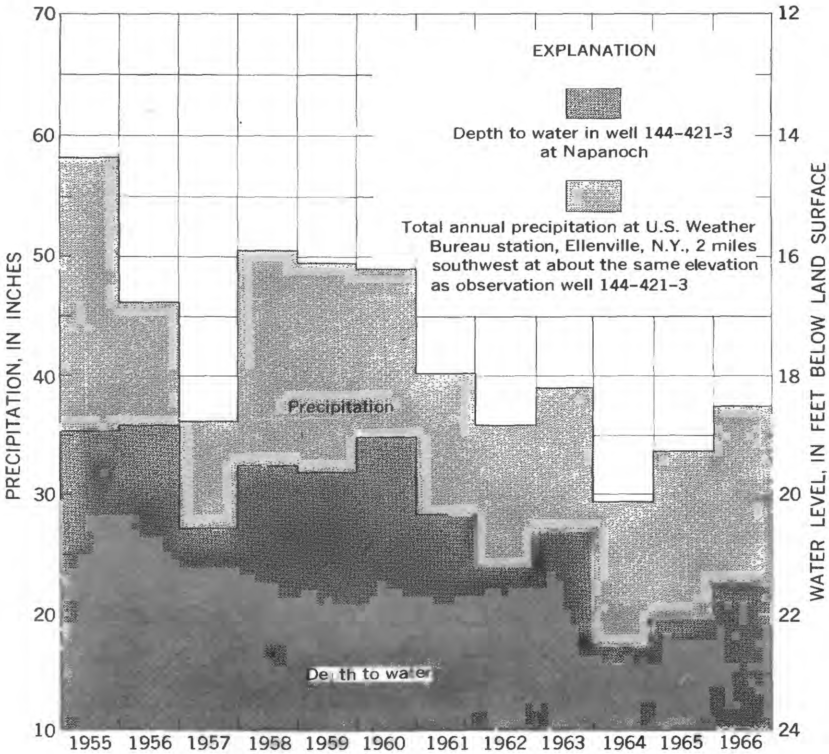


FIGURE 12.—Mean annual water level at Napanoch and mean annual precipitation at Ellenville.

is also a very obvious feature. Notice in figure 11 that the water levels regularly decline during the growing season. They generally reach their highest point at the end of the winter or early in the spring, even though there may be large amounts of precipitation during July and August. The summer precipitation is lost mostly through evaporation and transpiration.

EFFECTS OF WATER WITHDRAWAL

Discharge of water from an aquifer through pumping a well will cause a lowering of the water table or piezometric surface near that well. The depressed water table near a discharging well is cone-shaped and is appropriately called a cone of depression. Figure 13 is a sketch of a cone of depression caused by pumping the well at the house indicated while the other wells were idle. Note that pumping water from this well lowers the water level in the neighboring wells. The effect on

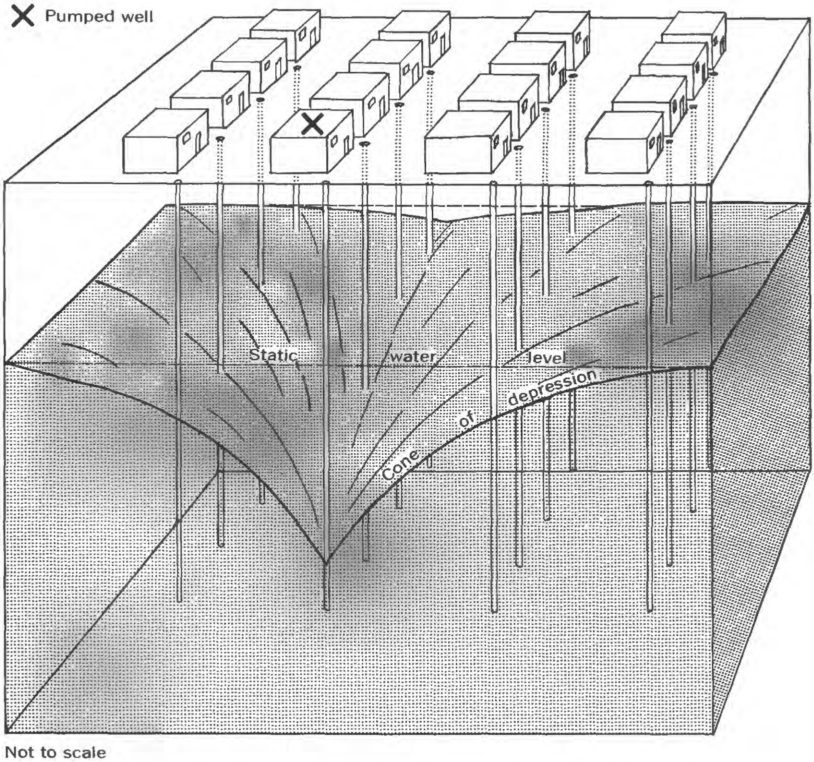


FIGURE 13.—Cone of depression produced by pumping a single well.

a well that is near the pumped well is greater than that on wells located farther away.

Figure 14 is a chart of the water levels observed in a well located about 230 and 330 feet from two production wells in the Woodbury well field. The wells are pumped intermittently at about 200 gallons per minute. When neither well is being pumped, the water level rises to its highest position; when both wells are pumping, the water level declines to its lowest position. The duration of pumping of these wells was recorded by automatic devices connected to the pump controls. These times of pumping are shown on the chart for correlation with the water-level fluctuations.

Occasionally, pumping wells at high rates can lower the water below the producing zones of neighboring wells, thus rendering them dry. Not only does this cut off the underground supply to the neighboring wells, but it may cause damage to their pumping equipment.

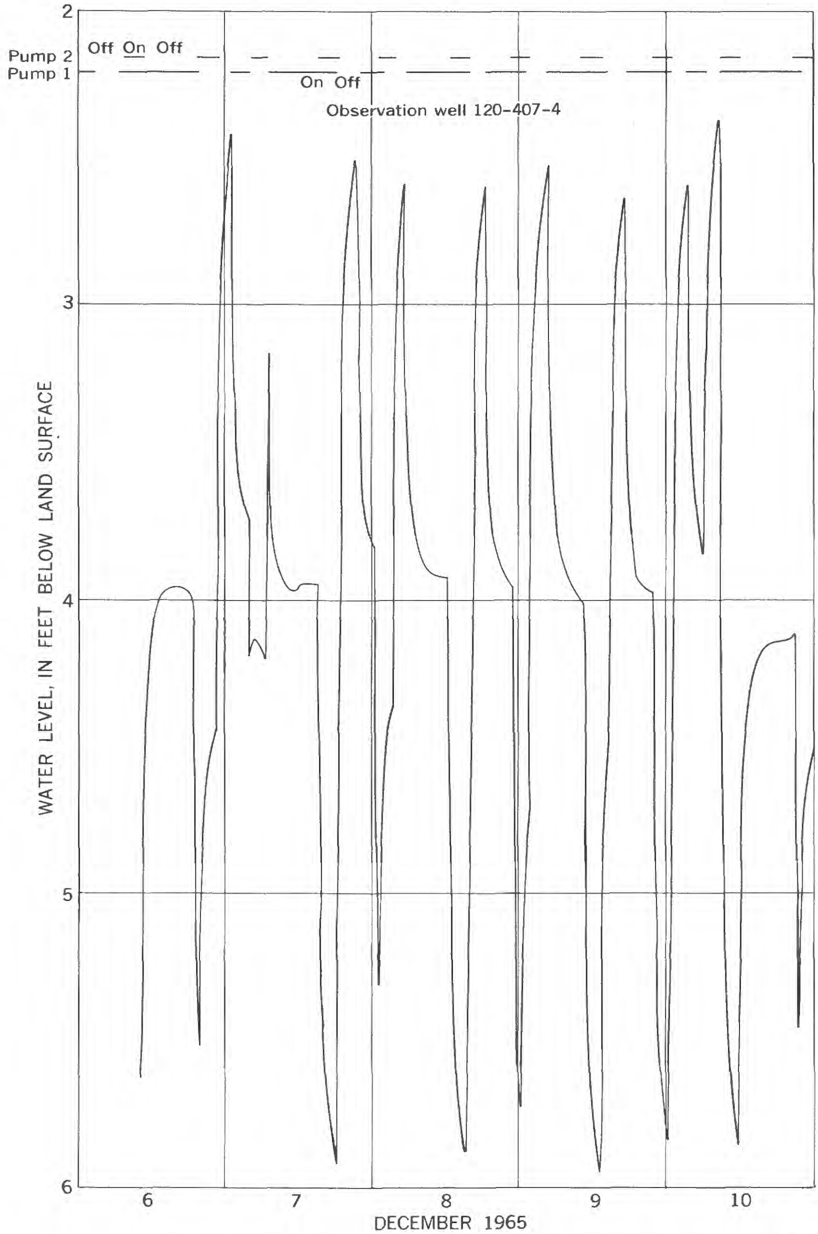


FIGURE 14.—Effects of withdrawal on water levels.

WELL CONSTRUCTION

The construction of a well tapping bedrock is simple to understand. In Orange and Ulster Counties, a layer of soil and unconsolidated rock of varying thickness overlies the consolidated rocks. A hole is drilled through the soil and into consolidated rock. Steel pipe called casing is placed in that part of the hole which penetrates soil, unconsolidated rock, or other earth which might crumble into the hole and block it. The water enters the well through fractures in the consolidated rock penetrated by the uncased well bore. A cut away view of a typical bedrock well is shown in figure 15.

Several alternative methods of construction are possible for wells tapping unconsolidated aquifers. Where a sand or gravel aquifer and the water table are near the land surface, well points may be driven or jetted into the aquifer for small supplies. A well point is a perforated pipe with a steel point on the end and a fine mesh screen over the perforations. Water enters the pipe through the screened perforations, while the screen stops the sand and gravel particles. Many satisfactory wells have been constructed in this inexpensive manner where conditions permit.

In some places open-end wells yield adequate home supplies (fig. 15), but this type of well can only be constructed in coarse-gravel that does not contain much fine material. Coarse sediments will not move up through the open end into the casing when the well is pumped at a low rate. High pumping rates, however, cause sand, or even gravel, to move up into the casing. For the small quantities of water necessary to supply a home, open-end wells finished in gravel can be quite satisfactory.

The highest well yields from sand and gravel aquifers are obtained with screened wells. A simplified diagram of screened well construction is shown in figure 15. The screened part of the well is placed in the aquifer where it allows water to enter the well while excluding sand and gravel. In the normal procedure of constructing this type of well, a cased hole is drilled into the aquifer and a screen is placed in the bottom of the hole. When the casing is pulled back, the screen is left in contact with the aquifer. The water in the well and aquifer adjacent to the well is then agitated in order to remove fine sand from the aquifer and well screen. The removal of this fine material is called development and it greatly increases the rate at which water can flow into the well. Because the removal of fine material may cause the aquifer to compact, the driller must take care to leave some aquifer material above and below the screen. If he does not, compaction of the aquifer might allow overlying confining clay strata to collapse sufficiently to come in contact with the screen.

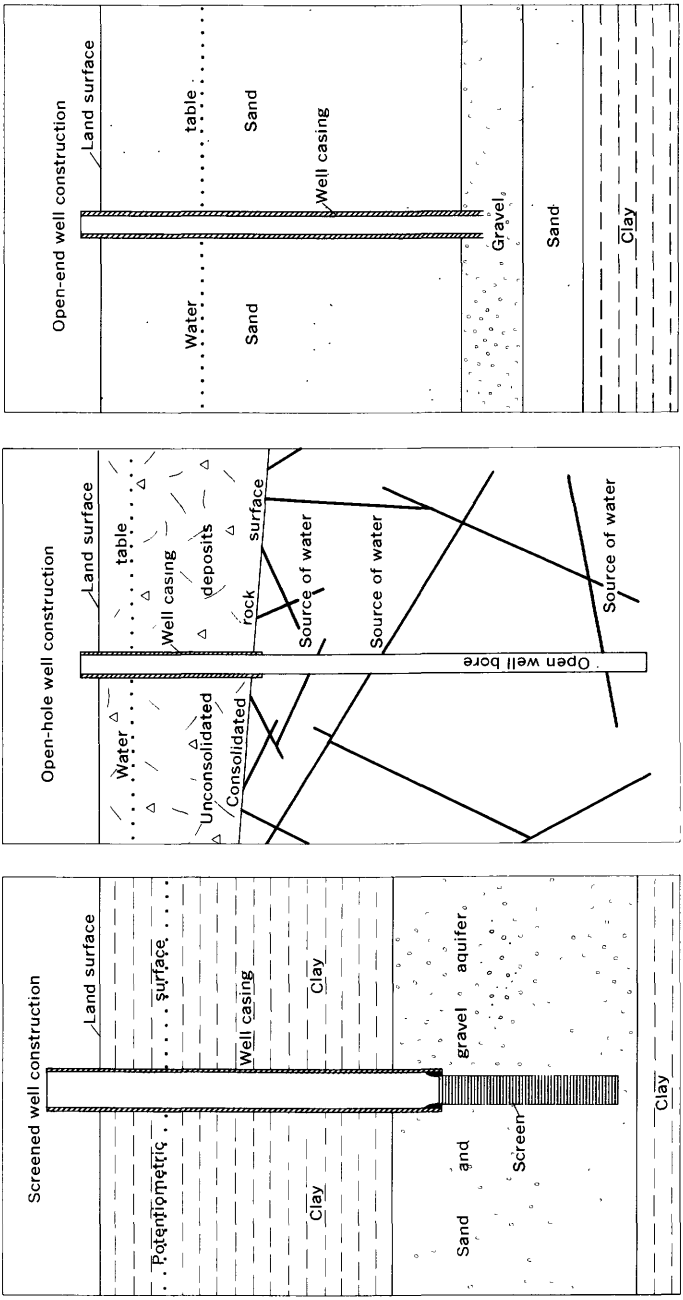


FIGURE 15.—Common types of well construction.

CLASSIFICATION OF UNCONSOLIDATED AQUIFERS

The highest yield aquifers of Orange and Ulster Counties are found in unconsolidated, stratified, glacial deposits. The extent of these deposits is shown on plate 2. During the Pleistocene Epoch a south-moving ice sheet covered the two counties. Erosion by the ice removed the mantle of weathered rock that had formed over the bedrock during previous ages, and bedrock that would not move was abraded by rock fragments carried in the ice. The rock material picked up by the ice was eventually redeposited on bedrock when the ice melted. These glacial deposits are divided into two types: (1) unstratified ground-moraine deposits (till and lodgement till), deposited directly from the ice, and (2) stratified deposits (outwash, ice-contact, and lake deposits), laid down in glacial lakes and streams.

GROUND MORAINE

A thin layer of ground moraine mantles most of the two-county area. The moraine is composed of till—an unsorted mixture of boulders, gravel, sand, silt, and clay. The till almost invariably overlies bedrock and averages about 20 feet in thickness, although the thickness varies greatly from place to place and is generally dependent on irregularities in the underlying bedrock surface. Another type of ground moraine, lodgement till, was deposited in greater thicknesses beneath the moving glacial ice. Drumlins are elongate streamlined hills composed of lodgement till, and they are oriented parallel to the direction of ice movement. A number of drumlins were formed in the Ridge and Valley province of the two counties. The lodgement till in these drumlins is commonly 75 and 200 feet thick. The till also has been plastered in thick deposits on the slopes of some steep bedrock surfaces.

Well drillers in the two-county area refer to till as "hardpan" because it is characteristically difficult to drill. The large boulders in till seriously impede the progress of drilling and sometimes have to be broken with explosives. Because till has a low permeability, it is considered to be a poor aquifer. None of the drilled wells in the inventory sample obtained water from till deposits. Large-diameter (36–inch average) wells dug in till, however, are capable of yielding sufficient supplies for homes and small farms.

OUTWASH

As melt water drained from the glacier, it carried with it rock debris that had been incorporated in the ice. Gravel, sand, silt, and clay were washed down the stream valleys away from the glacier. The swift-moving streams left layers of relatively clean gravel and sand along their valleys and carried the finer silt and clay particles further downstream to quieter waters. Deposits formed in this manner are appropriately called glacial outwash, and such deposits are the best aquifers in Orange and Ulster Counties. They are highly permeable and are generally located in hydrologic situations favorable for high rates of recharge.

During the general retreat of the glacier, the retreating ice front remained for a long time at the present location of Summitville in Sullivan County (Rich, 1934, p. 66). At Summitville the southerly flow of ice was just matched by the rate of melting, and a large amount of rock debris called recessional moraine was deposited. A river of melt water flowed from this ice front past the present site of Port Jervis and on to the Atlantic Ocean and filled the valley of Neversink River between Summitville and Port Jervis and south in the Delaware River valley to Milford, Pa., with 100 to 150 feet of silt, sand, and gravel. A few small streams in Orange County deposited sand and gravel in their valleys in a similar manner. These melt-water channels can be identified on plate 2 as sinuous deposits of sand and gravel.

ICE-CONTACT DEPOSITS

Stratified sediments of sand and gravel that were deposited at the margins of the melting ice are called ice-contact deposits. They overlie the ground-moraine deposits in valleys and along hillsides and may exceed 200 feet in thickness at places but are generally thinner. Ice-contact deposits are generally only slightly less permeable than outwash deposits, but are frequently of very small volume and are generally not located in hydrologic situations favorable for high rates of recharge.

The morphology of the ice-contact deposits was determined principally by the form of the stagnant ice against which they were laid down. Two prevalent morphologic types are kames and kame deltas. A kame is a hillock or mound of sand and gravel having ice-contact slopes. A kame delta is a delta formed in contact with stagnant ice where a glacial stream flowed into a glacial lake. Examples of kames and kame deltas may be seen at Chester, Pellets Island, and in the area about Phillipsburg, Crystal Run, and Stony Ford. They appear as small irregular areas of sand and gravel on plate 2.

LAKE DEPOSITS

Ice- and moraine-dammed lakes were abundant although short lived during postglacial time in the two counties. Because the quiet water of the lakes allowed silt and clay to settle out, deposits as much as 150 feet thick consisting of thin alternating layers of clay and silt were deposited on the lake bottoms. The silt was deposited during the summer months when the lake surfaces were free of ice and the lake waters could be agitated by wind. During the winter months, when the lakes were covered with a layer of ice, the lake waters were quiet and allowed even very fine clay to settle on the bottom. Clay deposits were mined at New Paltz and are still mined at Roseton for the manufacture of bricks.

A large lake in the Wallkill River valley near Durlandville and Pine Island in Orange County has filled with silt and clay and has become dry land since the last glacier melted. When the lake was nearly filled with sediment, a swamp developed on the shallow lake bottom and vegetation thrived. Now, a 10-foot layer of organic soil forms fertile farmland where there was once a lake. All lakes filling with sediment will eventually become land areas in a similar manner.

In nearby Rutgers Creek valley a postglacial lake formed near Westtown. The lake drained when the ice dam at the north end melted. The tops of old deltas built into the lake are still present at about 580 feet above sea level, a fact which indicates that the lake was more than 120 feet deep. The deltas are shown as sand and gravel and the bottom, as clay on plate 2.

The recessional moraine at Summitville formed a dam for a finger-like lake that was over 300 feet deep and extended northward to the village of Accord in Ulster County. Large sandy deltas were deposited in this lake by tributary streams entering at the present locations of Napanoch, Ellenville, and Spring Glen. Eventually, water overflowed from the lake and cut through the dam at Summitville which allowed the lake to partially drain in a southward direction. Later, the ice receded far enough for complete drainage to the north.

Because of damming by glacial ice in the Hudson Valley, another lake formed in the Wallkill River and Rondout Creek valleys between the Hudson valley and a point about 2½ miles south of New Paltz. Sandy deltas were deposited in this lake by tributary streams at New Paltz and Tillson.

As glacial ice melted from the Catskill Mountain area of Ulster County, the first melting took place in the upstream parts of the Esopus Creek valley. The drainage was blocked by ice lying in the downstream channel, and as a result, small lakes formed in the mountain valleys.

The lower Esopus Creek valley contained the largest postglacial lake in the two-county area. This lake was dammed at the downstream end by a tongue of glacial ice in the Hudson Valley. The largest deltas built into this lake were deposited between Katrine Park and Veteran, northwest of Kingston.

Glacial ice lingered longest in the Hudson Valley, but before it left, lakes formed between the stagnant melting ice and the rock walls of the valley. Today, the remnants of deltas (Ries, H., 1895, p. 114) and lake bottoms that were built in these lakes appear as terraces along both sides of the river at altitudes ranging from 120 to 180 feet. With the melting of the glacial ice in the Hudson Valley, the last period of glaciation ended in Orange and Ulster Counties.

The silt and clay deposits laid down in the glacial lakes are not aquifers because they have very low permeability. The sand and gravel deltas built into the lakes, however, are frequently still crossed by streams. Although the deltas are generally small, they may be rapidly recharged and are therefore aquifers that are sometimes capable of yielding moderate supplies of water. Geologic sections through unconsolidated lake sediments in many of the area's major stream valleys are shown on plate 4.

POTENTIAL YIELDS OF THE MAJOR UNCONSOLIDATED AQUIFERS

Potential yields of those aquifiers capable of sustaining industrial and municipal ground-water supplies can be estimated from geologic and hydrologic information through the application of hydrologic principles. Water withdrawn from an aquifer comes from two sources: recharge and water in storage. Because of their small volume, the high-yield aquifers in Orange and Ulster Counties have small storage capacities, and the annual yields of these aquifers usually can be estimated almost entirely from their annual recharge. In fact, the aquifers would be of little importance if abundant precipitation and favorable surface conditions did not allow rapid and frequent recharge. Estimates of the potential yields of aquifers presented in this report are based primarily on estimates of potential recharge from all possible sources, under the recharge conditions that would exist if maximum withdrawal were taking place.

Some of the largest aquifers were analyzed for the amount of water that could be withdrawn from storage during 1 year. The period of 1 year was chosen because every spring, when the winter snowpack melts, the aquifers generally are completely recharged. These analyses were made by assuming reasonable water transmission and storage properties for the aquifer material and by determining pumping rates that would cause the maximum allowable drawdown in 1 year. Aquifer volumes and maximum allowable drawdown were determined from geologic data; the effects of well interference and impermeable boundaries were taken into account. In two aquifers, storage was found to be a significant factor and was integrated with the yield estimations.

In this report, natural ground-water discharge is considered a water source termed "salvable discharge." This discharge is an indirect measure of aquifer recharge that can be determined relatively easily by stream gaging and analysis of streamflow records.

After periods of little or no precipitation, all streamflow is derived from ground-water discharge through springs and general seepage, and the streams are said to be in a base-flow condition. To measure the ground-water discharge from aquifers, stream discharge was measured under base-flow conditions, at about 90-percent flow duration (the discharge that is exceeded 90 percent of the time).

Infiltration from lakes, ponds, and streams is an additional and important source of recharge for aquifers. To appraise the potential recharge through infiltration, the permeability of the intervening materials, the area of contact, and the maximum gradient between the source and the aquifer are considered. The amount of water available for infiltration is another limit on the amount of water which could be induced to infiltrate and is also considered in the following descriptions.

Unconsolidated aquifers with the greatest potential yield are discussed in the following pages and the estimated yields of these aquifers are shown on plate 3. The locations of relatively low-yield unconsolidated aquifers are shown on the plate.

Yields of individual wells tapping unconsolidated aquifers are highly dependent on well construction. Therefore, considerable caution should be exercised in attempting to estimate well yields from the data in the basic data report by Frimpter (1970). A nominal well construction must be assumed for comparing well yields in different aquifers or parts of aquifers. In this comparison, a fully penetrating, screened, properly developed, 12-inch diameter well is assumed.

The most variable geohydrologic property affecting the yields of individual wells is transmissibility (aquifer permeability multiplied by aquifer thickness). While aquifer permeability varies considerably over short distances within the unconsolidated aquifers of Orange and Ulster Counties, it is not expected to vary much from aquifer to aquifer. This is expected in view of the relatively similar origin (Pleistocene glacial outwash) of all the unconsolidated aquifers. However, aquifer thickness does vary from aquifer to aquifer and controls the transmissibility, and therefore, well yields. On a basis of aquifer thickness, therefore, estimated individual well yields for the unconsolidated aquifers of the two-county area are shown on plate 3. They are classified as having moderate yield (up to 300 gpm) or as having high yield (300 to 1,000 gpm).

LOWER NEVERSINK RIVER VALLEY AND BASHER KILL VALLEY

A sand and gravel aquifer in the valleys of the Neversink River and Basher Kill extends from Summitville in Sullivan County (Soren, 1961, p. 22) to Port Jervis in the southwestern corner of Orange County (fig. 16; aquifer L, pl. 3). Parker and others (1964, p. 92) describe this aquifer as being 28 miles long extending from Summitville to Milford, Pa., averaging 1 mile wide and storing about 11.3 billion cubic feet of water or about 84.4 billion gallons. The sand and gravel generally grades downvalley from coarser material at Summitville to finer material at Port Jervis and from coarser material on the bottom to fine material on the top. A thin layer of fine sandy and silty soil overlies some of the surface of the aquifer, mostly in the low, flat areas as shown on plate 2. The fine sand is more than 50 feet thick at Port Jervis, north of the confluence of the Neversink and Delaware Rivers. The thickness of the aquifer ranges from less than 10 feet to a little more than 150 feet but is variable because of the irregular surface of the underlying bedrock. An example of this irregularity is an islandlike hill of shale and siltstone protruding from the glacial-outwash sediments 1 to 2 miles northeast of Port Jervis. Section *A–A′*, plate 4, is a geologic cross section of the unconsolidated deposits showing the depth to bedrock in the Neversink Valley at Godeffroy.

The aquifer in New York State is 21 miles long, has an average width of about half a mile, and has an average saturated thickness of 100 feet, so the volume is roughly 29 billion cubic feet. "The storage, coefficient S, of an aquifer is defined as the volume of water it releases from or takes into storage per unit surface area of the aquifer per

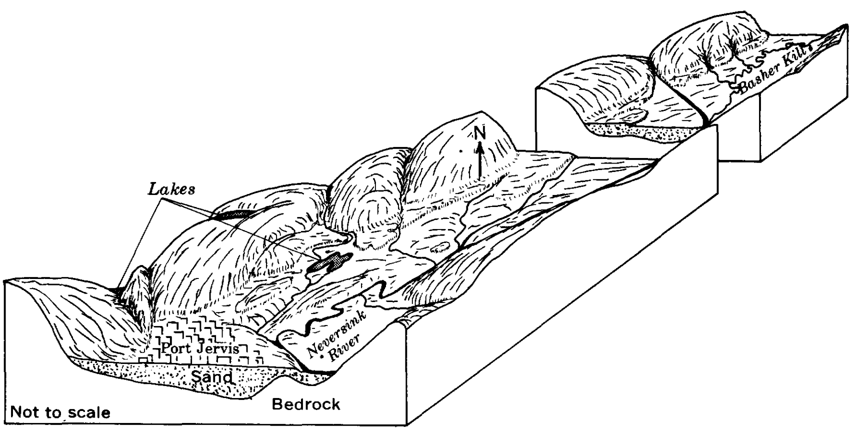


FIGURE 16.—Surface and subsurface structure of Neversink valley near Port Jervis.

unit change in the component of head normal to that surface." (Ferris and others 1962, p. 74). The storage coefficient for sand and gravel deposits generally is about 0.2. If this aquifer could be completely drained, it would yield about 0.2×29 billion cubic feet or about 44 billion gallons of water. Probably less than half of this amount could be withdrawn under the present economic and technological conditions.

The largest source of recharge in the Neversink valley is the Neversink River. The following paragraphs discuss the possible rate of recharge along a 5,000-foot reach where the Neversink River is 30 feet wide.

The flow of water through porous material is described by Darcy's Law, which may be expressed by the formula

$$Q=PIA,$$

where

Q = the quantity of water, in gallons per day,
P = the permeability of the aquifer, in gallons per day per square foot,
I = the hydraulic gradient, in feet per foot, and
A = the cross-sectional area through which the water is moving, in square feet.

Along this stream then

$A = 30 \text{ ft} \times 5{,}000 \text{ ft} = 150{,}000$ square feet, and
$I = 1$ (max).

Because there are no deposits of very low permeability between this stream and the aquifer and because the streambed area through which the water must infiltrate is relatively small, the permeability of the streambed rather than the permeability of the aquifer determines the rate of infiltration. Assuming that the streambed has a permeability similar to that of very fine sand—about 40 gallons per day per square foot and substituting in $Q=PIA$, then

$Q = 40 \times 1 \times 150{,}000$
$Q = 6{,}000{,}000$ gpd (gallons per day) per 5,000 feet of stream which is more than 4,000 gpm (gallons per minute).

Except for small reaches of the river where the streambed is separated from the sand and gravel of the aquifer by silt, 4,000 gpm should be a fair estimate of the infiltration rate. Therefore, an estimated 6 mgd (million gallons per day) per mile or a total of 48 mgd could be induced to infiltrate into the aquifer from the Neversink River through the 8 miles of streambed between Godeffroy and Port Jervis. This estimated 48 mgd which could be induced to infiltrate from the Neversink River is also dependent on the flow available for infiltration. The flow of the Neversink River at Godeffroy occasionally drops below 48 mgd, but only for very short periods of time. As an example, in 1964 when precipitation in the drainage area was about 20 percent below normal, the flow at Godeffroy was below 48 mgd for only 10 days in the period October 1, 1963, to October 1, 1964 (U.S. Geological Survey, 1970, p. 619). From October 1, 1964, to September 30, 1965, the flow at Godeffroy was less than 48 mgd for about 40 days but was never lower than 20 mgd (U.S. Geological Survey, 1970, p. 619). The low flows of this stream are largely controlled by the Neversink Reservoir. Releases from this reservoir are at the discretion of New York City except for conservation releases. Most likely, the export of ground water from this basin would not be allowable because it would reduce the natural ground-water discharge to the Delaware River system by the amount of water exported.

Base flow of Basher Kill near Cuddebackville indicates a ground-water discharge of over 13 mgd for the valley northeast of Cuddebackville. This represents a salvable discharge of about 1.4 mgd per linear mile of valley for that part of the aquifer drained by Basher Kill.

Because withdrawal of water at the 13 mgd rate would dry up Basher Kill at Cuddebackville about 10 percent of the time, the stream cannot be considered as an additional source of recharge through infiltration to the aquifer. Withdrawing 13 mgd from the aquifer would also reduce Basher Kill's contribution to the Neversink River by about 13 mgd. This reduction of streamflow would reduce the quantity of water available in the Neversink River to recharge the aquifer between Godeffroy and Port Jervis. However, if the water withdrawn from the aquifer in the Basher Kill valley were returned to Basher Kill as waste or used water, there would be no appreciable loss from the system and the quantity of water available for infiltration in the Neversink River would not be significantly reduced. The interrelations between surface water and ground water must be fully understood if the water resources of these two valleys are to be developed to their full potential.

An alternative and more accurate method of estimating potential yield of an aquifer is to estimate the quantity of water available in storage between periods of complete recharge. A large part of the spring runoff in the Basher Kill valley is water which might have become ground water if the aquifer had not been completely saturated. If, however, ground water is withdrawn for use during periods of little recharge, additional storage space for spring recharge would be created. Water that is normally rejected as recharge would then be accepted as recharge and be available as usable ground water. No permanent loss in storage would result because the aquifer would be recharged each spring, just as the storage of a surface-water reservoir is replenished during spring runoff.

To analyze this aquifer for potential yield from storage, several simple and logical assumptions were made: the transmissibility of the aquifer averages about 50,000 gallons per day per foot, the storage coefficient is 0.2, the valley walls are impermeable, and the aquifer is completed recharged during the spring. Because the aquifer averages about 100 feet thick, an available drawdown of 50 feet was assumed. By constructing distance drawdown graphs for the stated conditions, by summing the drawdown produced by hypothetical pumping wells, and by summing the drawdown caused by the impermeable valley walls, a maximum possible pumping rate can be found. In this analysis, a pumping rate of about 400 gpm for wells 1,000 feet apart was found to produce a drawdown of 50 feet at the pumping wells after 360 days. This amounts to about 2.8 mgd per 5,000 foot length of valley—a total of 39 mgd for the aquifer between Godeffroy and Summitville, and 22 mgd for the aquifer between Godeffroy and Port Jervis.

The estimates of recharge that would occur if water were drawn from storage are larger and believed to be more accurate than the estimates of recharge determined from measurements of natural ground-water discharge. Therefore, the estimates of potential yield of the aquifer are based on these larger recharge estimates. The aquifer between Godeffroy and Port Jervis has a potential yield of 70 mgd—22 mgd from storage and 48 mgd from stream infiltration. The aquifer between Godeffroy and Summitville has a potential yield of 39 mgd from storage. As mentioned previously, the streamflow may not be considered an infiltration source in this section of the aquifer because streamflow here is derived from ground-water discharge.

In (1968) only a few wells tapped this aquifer and it was virtually undeveloped. As more industry and housing enters the area, the aquifers undoubtedly will be developed to a much greater extent. Planned recharge using infiltration pits or even recharge ditches that could be scarified between recharge seasons could be made to easily double the recharge to this aquifer. A discussion of artificial recharge and numerous references ae included in Parker, Cohen, and Foxworthy (1967).

SANDBURG CREEK VALLEY

Between Phillipsport in Sullivan County and Wawarsing in Ulster County, the valleys of Homowack Kill, Sandburg Creek, and a small part of Rondout Creek contain an important sand and gravel aquifer (aquifer G, pl. 3). The Summitville moraine forms the southern end of this aquifer and is part of the drainage divide between the Delaware River drainage basin and the Hudson River drainage basin.

Deltas of sand and gravel as Spring Glen, Ellenville, and Napanoch are part of the aquifer.

Ground-water discharge from this aquifer sustains the low flow of Sandburg Creek. A flow of about 10 mgd in Sandburg Creek at Ellenville is equaled or exceeded 90 percent of the time. This 10 mgd is salvable ground-water discharge and represents about 1.7 mgd per mile of valley aquifer. The remaining 3-mile stretch of valley aquifer between Ellenville and Napanoch is very similar to the part south of Ellenville and is therefore estimated to discharge ground water at the same rate, giving a total salvable ground-water discharge for the whole aquifer of about 15 mgd. Because we have considered the low streamflow as salvable ground-water discharge, the streams which carry this discharge cannot be considered sources of infiltration except during high flow.

More than enough water generally is available to recharge the aquifer during each spring thaw. Therefore, this aquifer was analyzed for potential yield from storage in a manner similar to the analysis of the aquifer between Summitville and Port Jervis. Assuming complete recharge each spring, about 2 mgd per mile of valley length, or a total of 18 mgd, would be available in the aquifer between Phillipsport and Wawarsing. Because the quantity of recharge determined by the storage analysis is slightly larger than the estimated salvable discharge, it will be used as the best estimate of potential yield from this aquifer.

The yields of a few screened wells in this aquifer are indicative of its potential. A 39-foot-deep well ~~(142–423–9)~~, which was test pumped at 1,000 gpm, is part of the Ellenville water supply. During exploratory tests for the village of Ellenville, some large well yields were reportedly obtained, but at many places the water contained iron. Two industrial wells located north of Ellenville (143–423–1 and –2) were test pumped at 325 and 125 gpm, respectively. Farther north at the Napanoch Correctional Institution, wells 144–421–1 and –2 were test pumped at 222 and 349 gpm, respectively.

RONDOUT CREEK VALLEY

Rondout Creek valley, north of the hamlet of Wawarsing in Ulster County (aquifer F, pl. 3) contains a sand and gravel aquifer buried beneath silt and clay lake sediments as indicated on plate 2. Moraines were deposited near Wawarsing and Kerhonkson in this valley, and the aquifer appears to be in good hydraulic contact with Rondout Creek through the moraine 4,000 feet southwest of Kerhonkson. A test boring (146–418–a) was made in the moraine, and water flowed out of the boring at an elevation slightly higher than Rondout Creek.

At this boring, ground water is discharging from the aquifer through an overlying partially confining sand and silt layer. Water levels in wells tapping this aquifer show a gentle hydraulic gradient toward Rondout Creek (*E–E′*, pl. 4).

Because exposures of sand and gravel along the northwest side of the valley probably form the recharge area for this aquifer, recharge is greatest where this area is crossed by streams. Presumably withdrawal of water from the aquifer would reverse the hydraulic gradient and cause water to flow from Rondout Creek into the aquifer at the moraine near Kerhonkson. Actual recharge and potential recharge to this buried aquifer appear to be relatively small (probably less than 5 mgd). Numerous borings and detailed pumping tests would be required to determine its potential yield. Many open-end drilled wells that tap this aquifer yield water with objectionable quantities of iron.

AQUIFERS BETWEEN LAKE KATRINE PARK AND VETERAN

Deltas of sand and gravel that were deposited where tributary streams entered a glacial lake north of Kingston form aquifers at Veteran (aquifer C, pl. 3), Mount Marion Park and Ruby (aquifer D, pl. 3), and the confluence of Saw Kill and Esopus Creek (aquifer E, pl. 3), in Ulster County. Figure 17 is a block diagram showing part of this area. Except for the aquifer at the confluence of Saw Kill and Esopus Creek, these aquifers are recharged primarily in the spring by streams that cross their recharge areas near the valley wall.

The aquifers at Ruby, Mount Marion Park, and Veteran have to rely on annual spring recharge and their storage capacity to yield

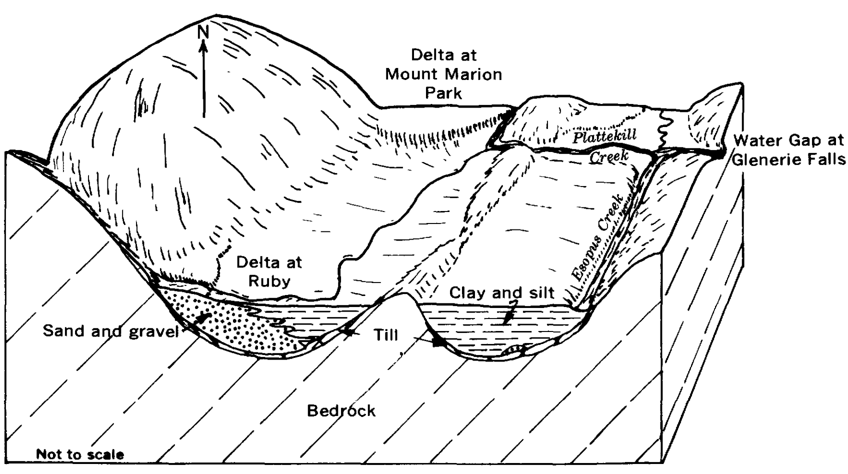


FIGURE 17.—Valley aquifers near Glenerie Falls.

their full potential. Each of these aquifers might yield between 0.5 and 1 mgd. Wells 200–400–4 and –5, drilled in the aquifer near Ruby, yield 200 gpm each.

Storage capacity is not an important factor in the prediction of potential yield from the aquifer at the confluence of Saw Kill and Esopus Creek. This aquifer receives its recharge as infiltration from nearby surface water. The permeability of the streambeds and pond bottoms which lie over this aquifer is the most important factor in determining potential recharge and, therefore, potential yield of the aquifer. Wells yielding a total of 2 mgd might be developed in this aquifer.

Wells 158–400–4 and –5 are within 60 feet of the east bank of Esopus Creek and induce infiltration from the creek. Well 158–400–4 is pumped at 600 gpm but yields water which must be treated for iron and manganese. Well spacing is especially important in this aquifer and should be given careful consideration in selecting additional well sites because of the likelihood of well interference.

RUTGERS CREEK VALLEY

Sand and gravel deltas were built into a postglacial lake between Unionville and Johnson in the Rutgers Creek valley in Orange County (aquifer s, pl. 3). The flat terrace one-half of a mile due west of Westtown is an example of one of these deltas. The terraces along Pine Hill Road, the gravel bank on Ford Lea Road, and the hill on which Westtown is located are of similar origin and composition. The areal extent of these sand and gravel deposits are shown on plate 3 and in figure 19. Most of the adjacent valley bottom is immediately underlain by clay and silt. Figure 18 is a schematic diagram illus-

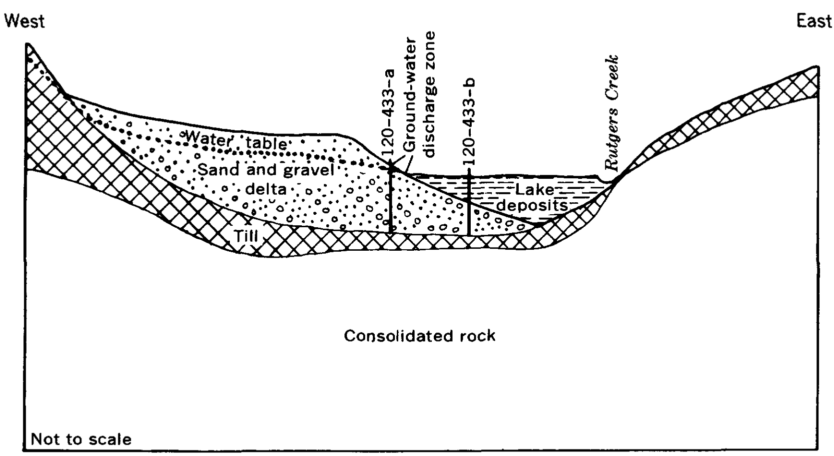


FIGURE 18.—Schematic geologic section of Rutgers Creek valley near Westtown.

trating the geologic structure of the deposits one-half of a mile west of Westtown. The beds of gravel and sand in this delta are wedge shaped and tilt upward in the direction of their source. U.S. Geological Survey test borings, wells 120–433–a and –b, confirm that the aquifer extends below the lake sediments.

During periods of little or no rainfall, the flow of Rutgers Creek and its tributaries is sustained by ground-water discharge from the sand and gravel aquifers. Therefore, estimates of the yield of these aquifers should not include recharge from stream infiltration. Recharge to these deposits occur mainly as infiltration from precipitation. Estimates of this rate of recharge were made from 13 measurements of salvable discharge at four sites along Rutgers Creek and its tributaries. The locations of the measurement sites are shown in figure 19, and the results of the measurements and their relation to flow-duration data for Rutgers Creek at Gardnerville are given in figure 20.

The discharge expected at the four sites when Rutgers Creek at Gardnerville is at 90-percent duration discharge were estimated from figure 20. This discharge was selected because streamflow at the 90-percent duration discharge is derived entirely from ground-water discharge. Table 1 gives these estimates and the areas of drainage and areas of aquifer drained. The last column of table 1 gives the ground-water discharge per square mile of aquifer drained, which is approximately equal to recharge by direct infiltration from precipitation. The variation of discharge per square mile seen in the last column is primarily due to the variation of ground-water storage available in the aquifers at elevations higher than Rutgers Creek. Between sites 3 and 4 there is much more sand and gravel above the level of Rutgers Creek than there is between sites 2 and 3, and therefore, the increase in flow between 3 and 4 is greater than between 2 and 3. The best estimate of possible recharge from direct infiltration of precipitation to these aquifers in table 1 is 0.58 mgd per square mile of aquifer recharge area. Because there are 2.8 square miles of aquifer recharge area in this basin, the estimated recharge is about 1.6 mgd.

TABLE 1.—*Estimated surface- and ground-water discharge, Rutgers Creek at 90-percent duration*

Measuring site		Increase in discharge over next upstream site (cfs)	Area drained less area drained by next upstream site (square miles)	Discharge for drainage areas listed in column 4 (cfs per square mile)	Discharge per square mile of sand and gravel drained between measuring sites	
No.	discharge (cfs)				(cfs)	(mgd)
1	0.43	0.43	14.2	0.031		
2	1.25	.82	4.2	.20	0.63	0.41
3	1.68	.43	3.9	.11	.48	.31
4	2.22	.54	1.7	.32	.90	.58

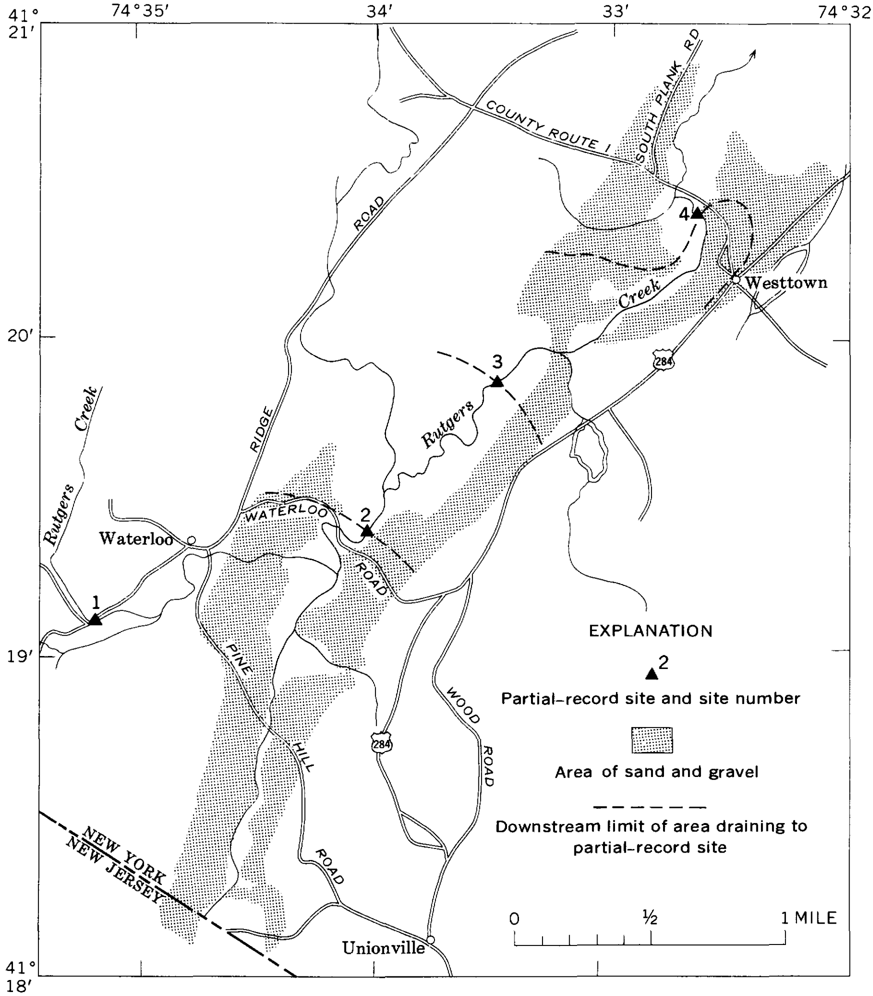


FIGURE 19.—Location of partial record sites and sand and gravel deposits in the Westtown to Unionville part of the Rutgers Creek basin.

This recharge rate of 1.6 mgd is also an estimate of the potential yield of these aquifers.

No screened wells have been constructed in the aquifers, but open-end wells 120–432–1 and –2 are domestic supply wells tapping the gravel at Westtown. The sand and gravel aquifer is preferred in this area because wells tapping the deeper shale bedrock frequently yield water containing hydrogen sulfide.

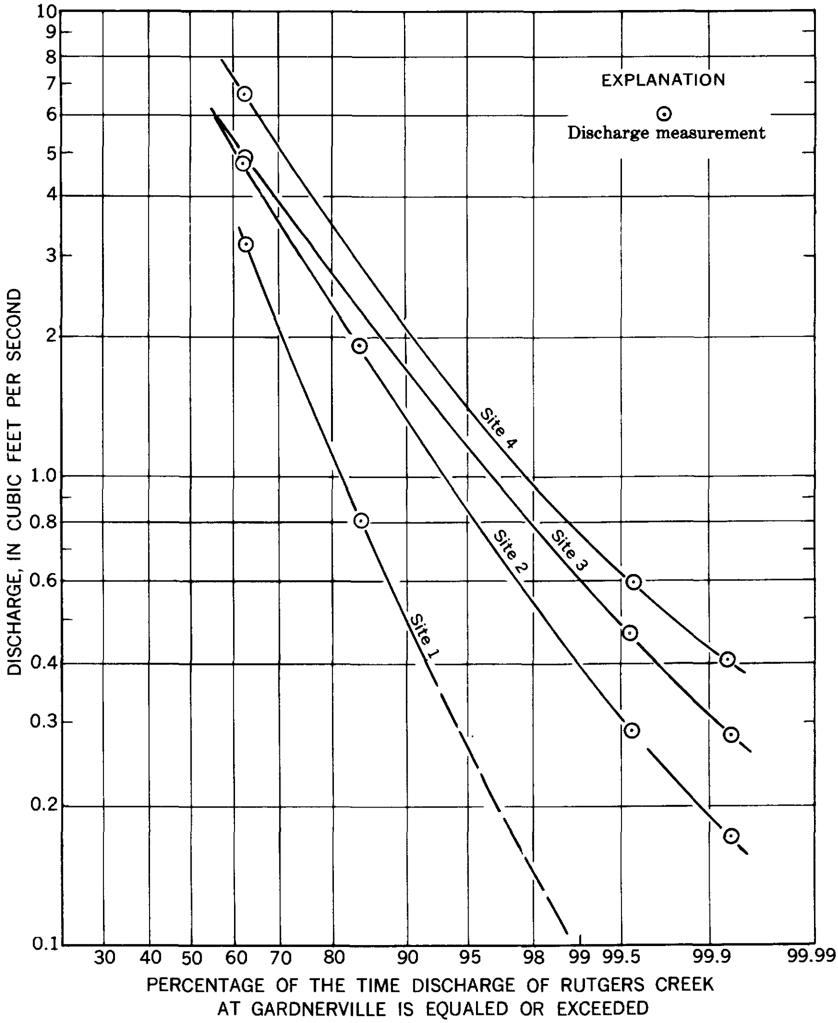


FIGURE 20.—Approximate duration curves for partial record sites on Rutgers Creek. Site numbers are those shown in figure 19 and in table 1.

SOUTHERN WALLKILL RIVER VALLEY

The western side of the Wallkill River valley between Denton in Orange County and the New Jersey State line contains a sand and gravel aquifer under organic soil and clay (aquifer U, pl. 3). The aquifer is exposed at U.S. Route 6 near Denton, near Pellets Island, east of Breeze Hill, and at a few other locations (pl. 2). Test-boring data presented in figure 21 indicate the extent and geologic structure

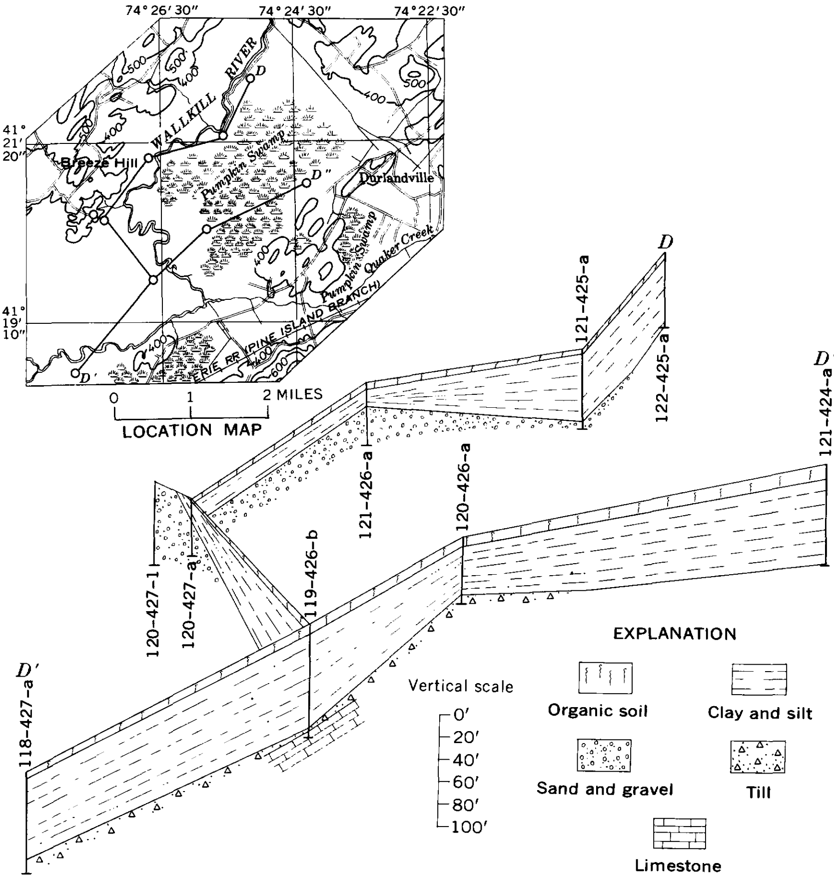


FIGURE 21.—Fence diagram of southern Wallkill River valley.

of the unconsolidated deposits in this area. The aquifer's eastern boundry is unboubtedly more irregular than indicated by the dashed line on plate 2.

The permeability of the aquifer is high, and individual wells may yield as much as 500 gpm. The rate of recharge, however, ultimately limits the safe yield of the aquifer. Recharge occurs primarily from precipitation falling directly on the limited surface exposures of sand and gravel. Based on the area of outcrop, the annual recharge rate from direct infiltration of precipitation to this aquifer is about 1 mgd.

The clay and silt lake sediments that cover most of the aquifer transmit water very slowly. In effect, these lake deposits are a barrier to water which might otherwise enter the aquifer. Very little water could be induced to flow from the streams to the buried aquifer because the streams rarely cross aquifer outcrops.

Under certain conditions the sand and gravel aquifer might receive a small amount of recharge from the organic soil. There is only a small zone where the organic soil and the aquifer are not separated by an impermeable clay layer. Normally, the water table slopes down from the sand and gravel recharge areas toward the areas of organic soil. This slope could be reversed by removing water from storage in the aquifer, and reversal of the gradient would cause water in the porous organic soil to flow into the aquifer (fig. 22). This situation was created by pumping well 121–416–1 in Chester, which is in a similar hydrogeologic situation. Withdrawing water and causing recharge in this manner would partly dewater the organic soil and have an undesirable effect on crops, as well as accelerate the rate of oxidation of the soil. As an example of the availability of water in organic soil, the city of Middletown was able to pump over 0.5 mgd from ditches dug in peat in a small basin near Shawangunk Lake during the drought in 1965.

Insufficient data are available to estimate the quantity of water which might enter the aquifer from the organic soil. Additional induced recharge could be obtained and aquifer yield increased by routing surface streams over areas where the aquifer is exposed.

Very few high-yield wells have been constructed in this aquifer. Screened wells 121–423–3 and –4 supply water for the New Hampton Training School Annex, but attempts to locate a large ground-water supply for the main campus of the New Hampton Training School in Denton have been unsuccessful. Dug wells and drainage ditches in the organic soil have been utilized as sources of irrigation water in the area.

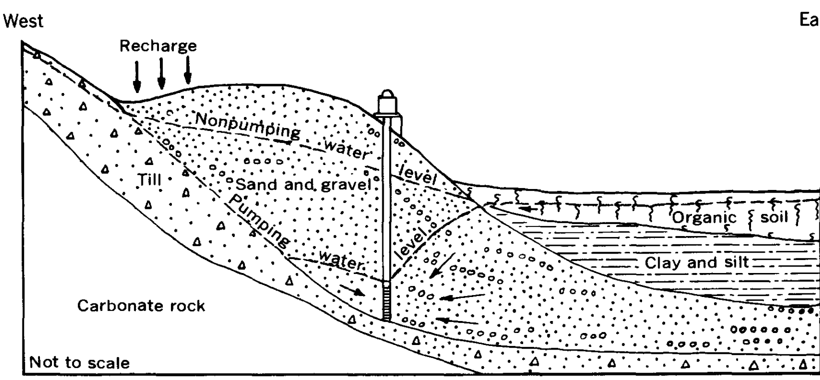


FIGURE 22.—Idealized view of the hydrogeologic conditions in an aquifer in the Wallkill River valley.

BLACK MEADOW CREEK VALLEY

In the vicinity of Chester in Orange County, Black Meadow Creek crosses an old lake bed similar to the one in the Wallkill River valley near Durlandville and Pine Island that was mentioned previously. Irregularly shaped hills of sand and gravel called kames are scattered through the areas as shown on plate 2. Parts of these kames are overlain by younger impermeable lake deposits.

Because these aquifers are not crossed by streams that could be induced to infiltrate, direct infiltration from precipitation is the only large source of recharge. The yields of these aquifers may be estimated by multiplying their recharge area by an estimated recharge rate; the recharge area may be determined from geologic maps.

A sand and gravel deposit crops out as a kame along Baird's Crossroad, extends northward beneath the lake deposits, and is tapped by well 120–417–9 (fig. 23). Although the outcrop of this aquifer is only about 0.04 square mile, the infiltration rate on this area should be very high. Because of sand and gravel mining operations, the surface of the kame has been stripped of vegetation and soil and has been reduced to a saucer shape that traps all the precipitation that falls on it. The sand and gravel at land surface is very permeable and most of the precipitation must percolate down to the water table. Under these special conditions, the annual recharge rate might be 1 mgd per square mile. Recharge to the aquifer would then be $0.04 \times 1{,}000{,}000$ or about 40,000 gpd.

The quantity of available water stored in this aquifer can be estimated from the volume of the saturated aquifer. It apparently is wedge

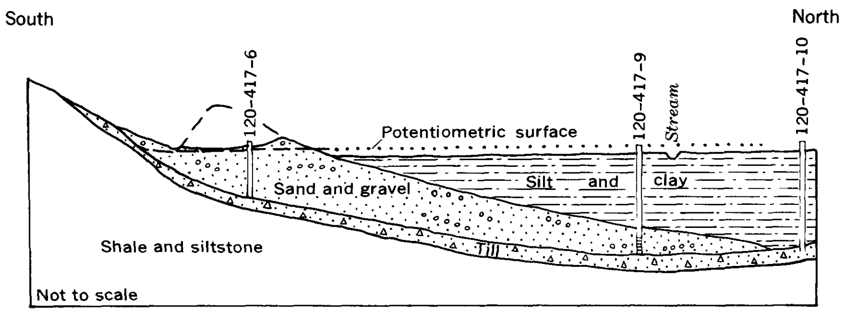


FIGURE 23.—Schematic geologic section of aquifer at Baird's Crossroad.

shaped, and its volume can be determined from its approximate dimensions which are:

h=70 feet (maximum thickness),
l=2,500 feet (length), and
w=2,500 feet (width).

The volume of available water in this wedge is then:

$$\frac{h\,l\,w}{2}\times S,$$

where S is the storage coefficient (0.2). Therefore,

$$\text{volume}=\frac{70\times 2{,}500\times 2{,}500}{2}\times 0.2=43{,}750{,}000 \text{ cubic feet.}$$

This is equivalent to about 327 million gallons. A well discharging 700 gpm continuously for 1 year produces over 367 million gallons. Clearly then, if this aquifer were not frequently recharged, pumping would rapidly exhaust the supply. Therefore, the estimated yield of 40,000 gpd for this aquifer is based on the estimated annual recharge rate from direct infiltration of precipitation.

Well 120–417–9 in this aquifer was pumped at 460 gpm for 149 hours. After 100 hours it had caused a water-level decline of 1.3 feet in observation well 120–417–6 located in the recharge area adjacent to Baird's Crossroad about 1,600 feet southeast of the pumping well (fig. 23).

Well 120–417–7 was drilled in another aquifer about 4,500 feet north of observation well 120–417–6 and test pumped at 350 gpm. Sand and gravel pit operations here had leveled a kame. In addition, gravel was mined from below the water table in some places now marked by ponds. During the pumping test of well 120–417–7, the water level in these ponds declined rapidly, proving the excellent hydraulic contact between the ponds and the well. Unfortunately, the water produced by this well was salty and could not be used for household purposes. A local industry that uses large quantities of salt is probably the source of the contamination.

Wells 120–417–9 and 121–416–1 were drilled into kames in the vicinity of Chester and pumped to supplement the village water supply. Although the short-term yields of these wells may be high, their annual yield is relatively low as shown in the previous paragraphs.

WALLKILL RIVER VALLEY AT PHILLIPSBURG

A sand and gravel deposit is exposed at Phillipsburg, where New York Route 17 crosses the Wallkill River in Orange County. On the northeast side of the crossing, the sand and gravel lies below the water table and is in hydraulic contact with the Wallkill River (aquifer N, pl. 3).

On Mount Joy Road about 200 feet east of Route 17, a U.S. Geological Survey boring penetrated 23 feet of water-saturated sand and gravel, and nearby well 126–421–3 reportedly penetrated about 81 feet of sand and gravel before entering rock. The areal extent of the aquifer material at and below land surface may be seen on plate 2.

Because this aquifer's volume and outcrop area are very small, its potential yield from storage and recharge from precipitation are also small. However, the aquifer is in hydraulic connection with the Wallkill River, and the quantity of water which might be induced to infiltrate is relatively large. During low streamflow a 50-foot-wide and 1,000-foot-long reach of the stream would be in contact with the gravel aquifer. The permeability of the gravel streambed in this reach is estimated to be 50 or more gallons per day per square foot, and the maximum gradient which could be developed across the streambed is 1 foot per foot. The quantity of water which could be induced to recharge the aquifer is, therefore, estimated by use of Darcy's law to be 2.5 mgd. Based on the minimum daily flow of 9 mgd recorded at a U.S. Geological Survey gaging station maintained for 23 years at a point on the Wallkill River about 5,000 feet upstream, 2.5 mgd is always available to recharge the aquifer. The aquifer has a very small storage capacity because of its small volume, and as a result, only a relatively small amount of water could be drawn from storage.

Test wells 126–422–3, –4, and –5 were drilled in the thin sand and gravel deposits on the west side of the river, and although yields indicated by these tests were high, the water contained objectionable quantities of iron.

GREENWOOD LAKE

Greenwood Lake lies in the highlands of southern Orange County, N.Y., and northern Passaic County, N.J. The village of Greenwood Lake is located on a sand and gravel aquifer at the north end of the lake (pl. 2; aquifer Z, pl. 3).

Water enters the sand and gravel near the hillsides and discharges from it into the lake through springs and seeps. A large subsurface spring discharges close to the west shore of the lake, about 1,100 feet north of the intersection of Route 17A and Lake Road. In the winter, when ice forms on the lake, this spring area remains free of ice because of the relatively warm water being discharged. Pumping from the aquifer could salvage this water being discharged into the lake.

High rates of withdrawal could reverse the ground-water gradient and induce recharge from the lake. It is conservatively estimated that this aquifer could produce 4 mgd. Detailed study of the lakebed permeability would be necessary to refine this estimate. Quantities greater than 4 mgd might be obtained by utilizing infiltration galleries, but for higher rates of withdrawal, direct intake from the lake would probably be more economical. More than enough water is available in the lake to support a ground-water supply of 4 mgd. Based on 46 years of record, the average daily overflow from the lake at Awosting, N.J., is about 32 million gallons.

Lowering of the lake level, as was suggested during the drought of 1964, would lower the water table in the aquifer, reduce the available drawdown in nearly all wells adjacent to the lake, and therefore reduce their yield. If the water table dropped below the bottoms of the wells and infiltration galleries, the wells would be useless until the water table once again rose above the bottoms of the wells.

The village of Greenwood Lake was once supplied by two screened wells 87 feet deep along the western side of the valley (113–417–2 and –3). These wells became plugged and their yields dropped off. A new large-diameter shallow screened well was dug for the village about 2,000 feet northeast of the older wells. Numerous shallow test wells have been drilled in the area; a few of them yielded water with objectionable quantities of iron. Because these are little or no barrier between the aquifer and the land surface, and since much of the village is built over the aquifer, the aquifer is highly susceptible to contamination.

WOODBURY CREEK VALLEY

The valley of Woodbury Creek at Highland Mills in Eastern Orange County contains an artesian aquifer (aquifer X, pl. 3). Some of the recharge area for this aquifer lies above the water table on the western side of the valley, but most of the aquifer is overlain by clayey and silty lake deposits in the valley flat, as may be seen on plate 2. A delta of sand and gravel forms a terrace extending across the valley at Cemetery of the Highlands just north of Highland Mills. Figure 24 is a block diagram of the Woodbury Creek valley near Highland Mills.

Woodbury Creek flows on either lake sediments or bedrock and cannot be considered a source of recharge to the aquifer, but a few small unnamed tributaries to Woodbury Creek do cross the exposed aquifer on the western side of the valley and are sources of recharge to the

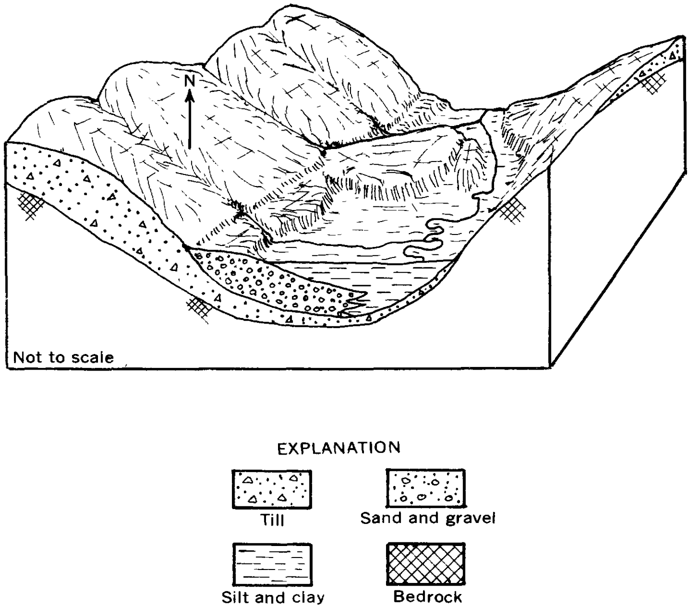


FIGURE 24.—Woodbury Creek valley near Highland Mills.

aquifer. Although the streams are small, their gravel beds are very permeable. An estimated annual average of 500,000 gpd could be induced to infiltrate from these streams. Most of this water would infiltrate during periods of high streamflow in the spring; very little water would infiltrate during periods of low streamflow in the summer.

Direct infiltration of precipitation on the one-half of a square mile of the aquifer that is exposed would occur primarily during the spring thaw. A warm spell in mid-February 1966 caused a high rate of recharge to the aquifer and a subsequent rise of the water level (fig. 25). The recharge area includes the sand and gravel exposed on the west side of the valley and the cemetery area to the north. An estimated annual average of 500,000 gpd could infiltrate the aquifer directly from precipitation.

The aquifer has sufficient volume to yield water from storage to maintain withdrawal rates during periods of little or no recharge. The total recharge that the aquifer could be expected to receive is 1 mgd, and withdrawal at this rate could be maintained without permanently lowering water levels. However, seasonal fluctuations of water level are magnified by withdrawal, as may be seen in figure 25.

Wells 120–407–1 and –2 tap this aquifer for the Woodbury Water District. A record of pumpage and water levels in this well field is given in figure 25. During the winter months of 1966 the two wells produced an average of 225,000 gpd, and during the summer months they produced an average of about 360,000 gpd. The production of these two wells apparently represents the present total production from the aquifer. Well 120–407–3, also in the well field but not used at present, was pumped at 550 gpm for 168 hours while wells 120–407–1 and –2 continued to produce at their normal rates. During this pumping test, the level of a pond about 0.4 mile to the north declined about 16 inches. The pond is located at the site of an abandoned gravel pit and is in hydraulic contact with the aquifer.

When the first test wells were drilled through the confining lake sediments into the permeable gravel, they flowed profusely, indicating that the aquifer is recharged at a higher elevation (fig. 26). Because pumping has developed a cone of depression around the well field, the water levels now rise above land surface only during early spring when recharge rates are highest.

An aquifer test was made at the Woodbury well field; the data obtained from the test were analyzed by the Theis nonequilibrium method (Ferris and others, 1962, p. 92–100). The transmissibility was determined to be about 60,000 gpd per foot. Therefore, the average permeability of the 40-foot-thick aquifer is about 1,500 gpd per square foot. The analysis also indicated a storage coefficient of about 0.0005, which reflects the artesian conditions at the well field. The permeability of this aquifer is probably representative of the sand and gravel aquifers in the two-county area.

SEELEY BROOK VALLEY

A small sand and gravel aquifer in Seeley Brook valley (aquifer W, pl. 3), between Chester and Monroe in Orange County, is in excellent hydraulic contact with Seeley Brook (pl. 2). The water-level changes shown in figure 27 were observed in wells 119–415–3 and –4, at distances of 5 and 500 feet respectively from the streambed. These fluctuations were caused by a change of stage in Seeley Brook and prove the good hydraulic connection between the aquifer and the stream. Although there is no record of the stage of the creek adjacent to the wells, a 2-inch decrease in stage was recorded about 6 hours later at gaging station 1–3736 located a little less than 2 miles downstream.

The flow of Seeley Brook at the gaging station that is equaled or exceeded 90 percent of the time is about 2 cfs or 1.3 mgd. This flow is ground-water discharge that might be salvaged by withdrawal from the aquifer. The potential yield of the aquifer is therefore estimated to be about 1.3 mgd.

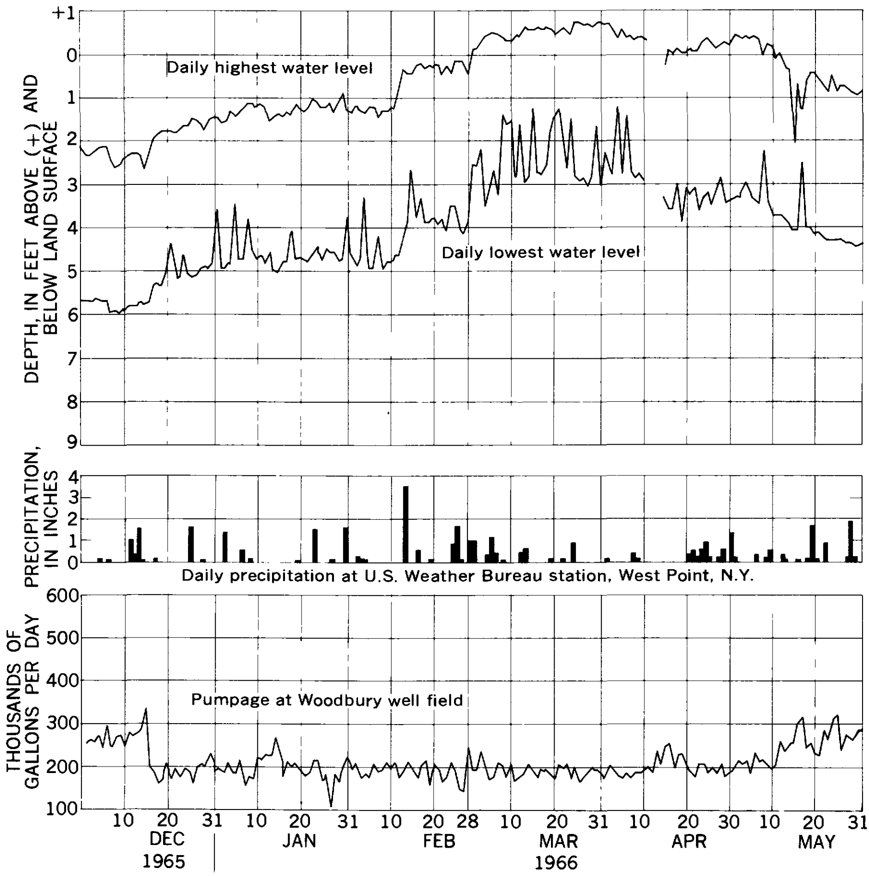
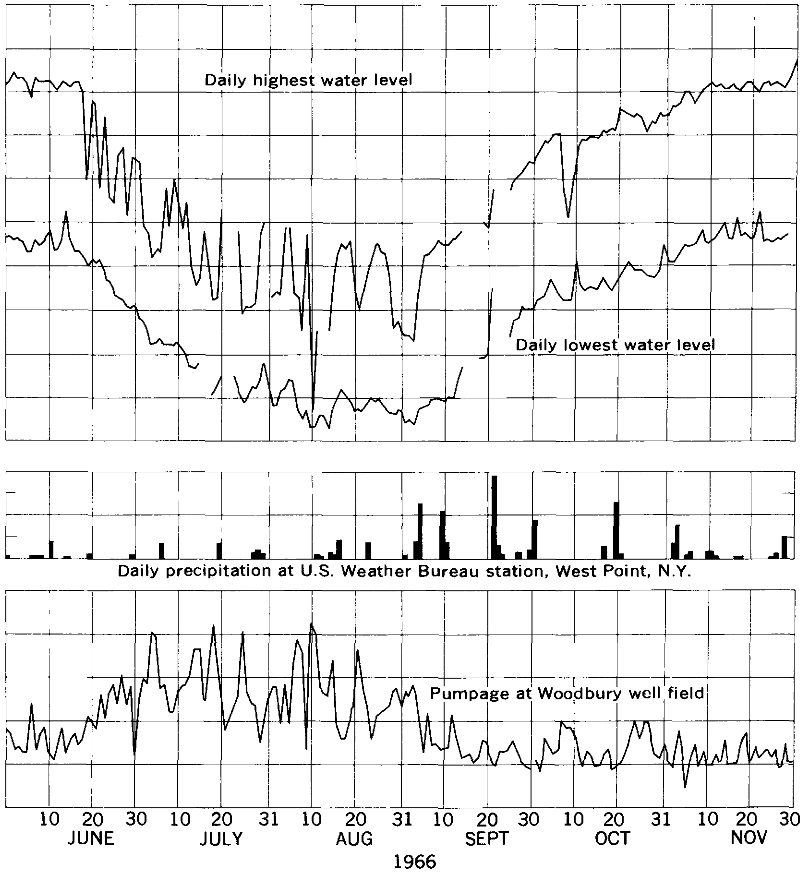


FIGURE 25.—Water levels and pumpage at the Woodbury

Screened well 118–415–2 taps this sand and gravel aquifer and is reported to be capable of producing at least 110 gpm. Screened wells drilled near well 119–415–3 should be capable of yielding over 200 gpm.

TIN BROOK VALLEY

A pitted outwash plain lies in both Orange and Ulster Counties northeast and east of the village of Walden (pl. 2). This sand and gravel deposit was laid down over and around remnant blocks of ice, and depressions or pits formed when the ice melted. Presently many of these pits intersect the water table in the gravel and form lakes such as Lake Osiris.



well field and record of precipitation at West Point.

Tin Brook drains the southern end of this sand and gravel aquifer (aquifers I and J, pl. 3). Low base-flow measurements of Tin Brook indicate a ground-water discharge of about 0.5 mgd from the aquifer. At the same time, wells withdrawing water from the aquifer along this reach of Tin Brook were salvaging an additional 0.5 mgd and diverting it to the Wallkill River via the village of Walden's water-supply and sewer systems. Therefore, the total ground-water discharge from the Tin Brook valley part of the aquifer was about 1 mgd.

The estimated recharge that might enter the aquifer through its 2 square miles of surface exposure during the spring thaw is also about 1 mgd, assuming an estimated recharge rate of 500,000 gpd per square

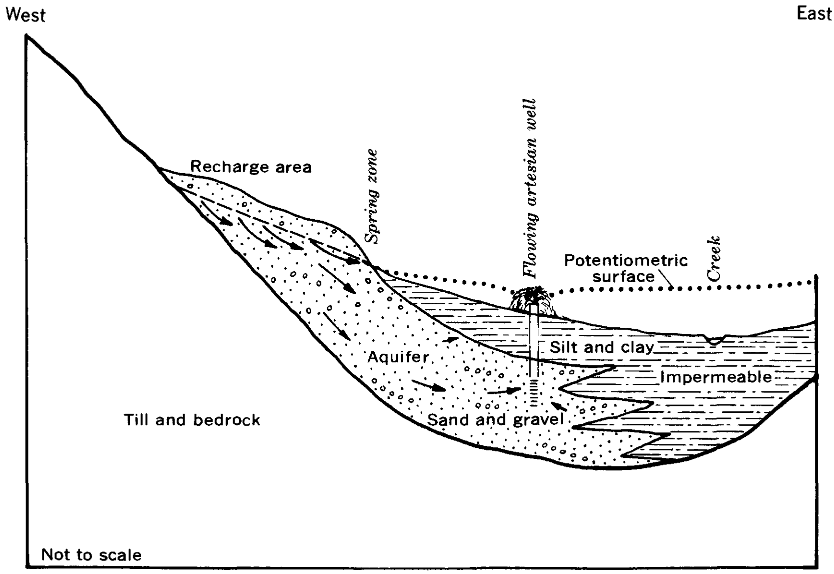


FIGURE 26.—Hydrology of the aquifer in Woodbury Creek valley.

mile (Heath, 1964). This estimated yield agrees with the value obtained from discharge measurements.

The village of Walden is supplied by wells adjacent to Tin Brook near East Walden. Wells 133–409–4, –6, and –7 tap this sand and gravel aquifer below as much as 33 feet of clay. Infiltration from Tin Brook near the well sites is not likely, but upstream reaches of Tin Brook and some of its tributaries appear to be in hydraulic contact with the aquifer.

SHAWANGUNK KILL VALLEY

The narrow V-shaped bedrock valley cut by preglacial Shawangunk Kill has been partly filled with irregularly distributed sand and gravel deposits (pls. 3, 4). From Otisville to Pine Bush, mixed deposits of outwash and moraine occur along the stream in both Sullivan and Orange Counties. While most of the deposits lie above the water table, thick deposits of sand and gravel occur in some of the deep parts of the bedrock valley and form small but highly permeable aquifers. These aquifers might be particularly productive where they are crossed by Shawangunk Kill or its tributaries and water could be induced to infiltrate.

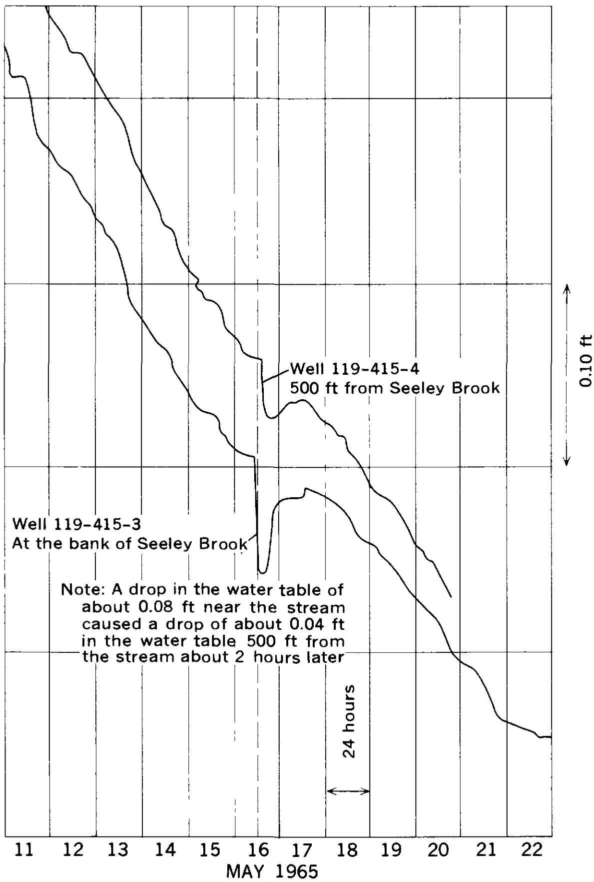


FIGURE 27.—Water level fluctuations in wells 119-415-3 and -4.

The aquifer at Pine Bush (aquifer H, pl. 3) is probably not capable of yielding over 1 mgd. The aquifer tapped by wells for the Pine Bush water supply does not appear to be in good hydraulic contact with any streams and has a rather small recharge area of low permeability. Numerous test wells drilled in the area have not been successful.

The aquifer at Otisville (aquifer M, pl. 3) was tapped in 1966 by test well 128-432-1. This test well yielded at least 200 gpm but is not located in the thickest or most permeable part of the aquifer. This aquifer is virtually unexplored but can probably yield between 1 and 2 mgd. It has a much larger highly permeable recharge area than the aquifer at Pine Bush and, in addition, is crossed by streams.

WAWAYANDA CREEK VALLEY

Irregularly distributed and partly buried outwash and ice-contact deposits lie in the valley of Wawayanda Creek in southern Orange County. These deposits constitute a complex but important group of aquifers (aquifer V, pl. 3) and recharge areas in this valley. Much of the recharge to these aquifers enters through sand and gravel deltas located at the edges of the main valley. Most of the recharge occurs during the spring when the tributary streams that cross the deltas are at their highest stage. Recharge also occurs from direct infiltration of precipitation, primarily in the spring. The fact that well 115–421–1 in the valley flat flows in the spring indicates that the aquifer's recharge area is at a higher elevation.

Well 117–417–5 on the north shore of Wickham Lake taps an artesian sand and gravel aquifer that extends below the lake but is not in good hydraulic contact with the lake. This well is reportedly screened from 72 to 92 feet and was pumped at 200 gpm. Recharge to this aquifer enters through the ice contact sand and gravel deposits exposed to the north between Kings Highway and the Lehigh and Hudson River Railroad tracks. Most of the recharge enters as direct infiltration of precipitation, but some may enter as infiltration from small streams which cross the sand and gravel. Most likely, the potential yield of this aquifer does not exceed 1 mgd.

Sand and gravel ice-contact deposits have been extensively explored at the southeast end of Wickham Lake. Wells at this location supply the State Training School for Boys, but the well yields are reportedly barely adequate. Undoubtedly some water is induced to infiltrate the aquifer from Wickham Lake.

East of the Warwick village boundary the Warwick Water Co. has three supply wells: 115–421–1, 115–420–1, and –2. Wells 115–420–1 and –2, 72 inches in diameter and less than 50 feet deep, were tested at 400 and 325 gpm respectively. This aquifer is thin, covers a small area, and therefore has a small storage capacity. Recharge to the aquifer occurs as direct infiltration of precipitation and from infiltration from two small streams that cross the recharge area. The permeability of the Wawayanda Creek bottom is very low, and recharge from this source is not likely. The total yield that could be sustained by this aquifer is less than 1.5 mgd. The well field is being operated to supplement a surface-water supply (1965).

There may be an aquifer immediately southwest of Warwick along New York Route 94; well 114–423–4 taps sand at a depth of 130 feet and nearby well 114–422–2 reportedly penetrated water-bearing sand and gravel before entering bedrock.

VALLEYS IN THE CATSKILLS

Rondout Creek and Esopus Creek valleys in the Catskill Mountains of Ulster County contain shallow surficial sand and gravel aquifers (pl. 2). The aquifer in Esopus Creek valley is shown as A on plate 3 and an aquifer in Saw Kill valley is shown as B on plate 3. The geologic structure of Esopus Creek valley is shown on plate 4. The sand and gravel aquifers vary in thickness and may yield large quantities of water where they are thickest and adjacent to streams. Recharge might easily be induced to flow through the gravel bottoms of the streams into the gravel aquifers. Potential yields from these aquifers are, therefore, limited only by the flow of the infiltration sources.

ADDITIONAL SMALL SAND AND GRAVEL AQUIFERS

Some additional small areas of sand and gravel which might yield 0.3 mgd or more are located on plate 3. Specifically these are:

- Beaverdam Brook valley, Orange County (aquifer O)
- Moodna Creek valley at Mountainville, Orange County (aquifer Q)
- Moodna Creek valley at Washingtonville, Orange County (aquifer P)
- Moodna Creek valley at New Windsor, Orange County (aquifer R)
- Hudson River valley at Marlboro, Ulster County (aquifer K)
- Monhagen Brook valley south of Middletown, Orange County (aquifer T)
- Ramapo River valley, at Harriman, Orange County (aquifer Y)

The shallow sand and gravel aquifer in Washingtonville induces recharge from Moodna Creek and is tapped by two wells, 125–410–1 and –2, supplying the village. The potential yield of this aquifer is in excess of 1 mgd.

The New Windsor Water Co.'s well 127–401–2, located about 200 feet north of Moodna Creek where U.S. Highway 9W crosses Moodna Creek, will reportedly yield as much as 1,500 gpm. This aquifer (aquifer R, pl. 3) is recharged by water from Moodna Creek, and its potential yield is in excess of 1 mgd. However, the water contains manganese, which is difficult and costly to remove.

The small aquifer in the Hudson Valley at Marlboro (aquifer K, pl. 3) is in an interesting hydrologic situation. When well 136–357–1, which taps this aquifer, was first drilled, the static water level was higher than the level of the Hudson River but fluctuated with the tide cycle in the Hudson. The acquifer is artesian and the fluctuations are caused by loading and unloading of the aquifer by the tides. The natural recharge for this aquifer comes from the small stream which cuts through a sand and gravel terrace about 1,500 to 2,500 feet west of the well and at a higher elevation. Before the well was pumped, discharge from the aquifer was upward through the confining clay and silt at a very low rate. Since pumping began, the artesian pressure in the aquifer has been lowered, and this flow has been reversed.

The first withdrawals from this aquifer were reportedly free of iron and manganese, but with continued withdrawal, iron and manganese contents have risen. Because both iron and manganese are known to be loosely adsorbed to clay, it is speculated that the ground-water flow through the confining beds is flushing these contaminants from the clay in the confining beds.

BEDROCK AQUIFERS

The occurrence of bedrock in Orange and Ulster Counties is shown on plate 1. It is estimated that about one-half of the 300,000 inhabitants of Orange and Ulster Counties depend on bedrock aquifers for their domestic water supply. Although bedrock generally yields sufficient quantities of water for individual home or farm needs, it rarely can be depended upon for larger industrial or municipal supplies. Only about 8 percent of the population is served by public water supplies obtained from bedrock aquifers (U.S. Public Health Service, 1963, p. 42–98).

INDUSTRIAL AND MUNICIPAL SUPPLIES

The consolidated rocks of the two-county area have virtually no primary or intergranular porosity for ground-water storage or transmittal. Ground water obtained from the bedrock aquifers comes from fractures or cracks in the rock. In a few instances, zones of high porosity and permeability have been found in the bedrock of the two-county area. These zones are primarily due to faulting and fracturing or to solution of the otherwise impermeable rock. Solution openings are restricted to the carbonate rocks and are caused by the circulation of ground water through preexisting fractures.

The New York City Board of Water Supply (Freund, 1942) reported that during construction of the Delaware Aqueduct a highly permeable zone of faulted rock was penetrated at a depth of 650 feet beneath Rondout Creek valley at Wawarsing. The zone is in the steeply dipping Manlius Limestone, Binnewater Sandstone, and High Falls Shale and is overlain by about 300 feet of unconsolidated sediments. To reduce the inflow of ground water, 3,640 cubic yards of liquid grout were injected into this 234-foot-wide fracture zone. The Board of

Water Supply estimated that more than 200,000 gpm would have flowed into the tunnel if it had been driven without special treatment. The Board also reported that the water pressure at the borings in the tunnel did not vary appreciably from the originally measured pressure of 290 pounds per square inch, even after long periods of heavy inflow.

This fractured fault zone is potentially a high-yield bedrock aquifer. The fault appears to closely parallel the tabular rock formations that strike in a northeasterly direction and dip steeply toward the northwest. On plate 1 the easternmost narrow band of carbonate rocks extending from Port Jervis to the northeast corner of Ulster County appears to contain this fault zone. There are evidences of this fault zone both north and south of Wawarsing. Water flows at about 50 gpm from test boring 149–409–1, which was drilled into these formations near High Falls. In the vicinity of Cuddebackville in Orange County, a small stream disappears into a sinkhole where there are many limestone caves that are probably related to this fault.

Fault zones between the crystalline rocks and sedimentary rocks in the eastern part of Orange County may contain large supplies of water. A few high-yield wells, which reportedly end in deep gravel, probably end in a brecciated (highly fractured) fault zone. Southeast of Washingtonville, well 124–408–2, capable of producing 250 gpm, is reportedly an open-end well in gravel at a depth of 215 feet. Five thousand feet to the east, wells 124–407–1 and –2 tap bedrock for yields of 180 gpm each. In addition, well 125–406–3 reportedly taps gravel at a depth of 175 feet and well 125–405–15 taps gravel at a depth of 300 feet. All these wells are probably located in the same northeast-trending fault zone.

Some wells on the east side of Woodbury Creek valley at Mountainville reportedly tap deep gravel deposits under till. As an example, well 124–404–2 taps gravel at a depth of 232 feet. The aquifer may be a brecciated fault zone. A fault zone is known to separate limestone from granitic gneiss in this area. Farther south, but along the same fault zone, at Arden Station Road and at Harriman interchange on the Thomas E. Dewey Thruway, borings have penetrated what appears to be a fault zone.

Locally, fault zones may be capable of producing municipal or industrial supplies in southeastern Orange County as well as in Rondout Creek valley near Wawarsing. Because the fault zones are narrow tabular zones of fractured rock that are nearly vertical in attitude, they may be easily missed by drilling. None of the faults are being used as water sources for municipal or industrial supplies, and little is known about their recharge characteristics. However, a high degree of well interference might be expected in widely separated wells tapping the same fault zone. Due to its high permeability, small volume, narrow width, and long length, the fault zone acts hydraulically like a long pipe. When the head is reduced at one end of the zone, the reduction is rapidly transmitted through its length.

RELATIVE YIELDS

A study of occurrence of water in the various types of bedrock in the two-county area indicates that the water-bearing properties of the rocks are almost identical. Results of information collected for 702 wells in consolidated rocks are presented in table 2. As can be seen in this table, there is very little difference in the yield per foot of aquifer penetrated for the different consolidated rock types. The little variation present can be attributed to the topographic location of wells which penetrate the different rock types. Most of the wells that tap crystalline rock are located on hillsides, whereas most of the wells that tap carbonate rock are located in valleys. Most of the wells in sandstone are located in the valleys of the Catskill Mountains, and those in shale occur in a wide variety of topographic situations.

TABLE 2.—*Relative productivity of aquifers*

Rock type	Number of wells	Gpm per foot of well exposed to aquifer
Crystalline rock	32	0.12
Sandstone	79	.18
Carbonate	79	.17
Shale	512	.15
Sand and gravel	62	21.9

The figures in this table are presented only for comparative purposes and should not be used for prediction of well yields.

Also shown in table 2 for comparative purposes are yield and penetration data for screened sand and gravel wells. As can be readily appreciated, there is a vast difference between the yields from bedrock wells and the yields from screened wells in unconsolidated deposits.

DOMESTIC WELL-DEPTH PROBABILITY

To obtain a better estimate of the probable depth required for a domestic well, data from 450 shale, 112 carbonate, and 75 sandstone wells are plotted cumulatively in figure 28. Only data pertaining to wells drilled for approximately the same yield (home and farm supplies) were used. Smooth cumulative frequency curves were easily fitted to the plotted data.

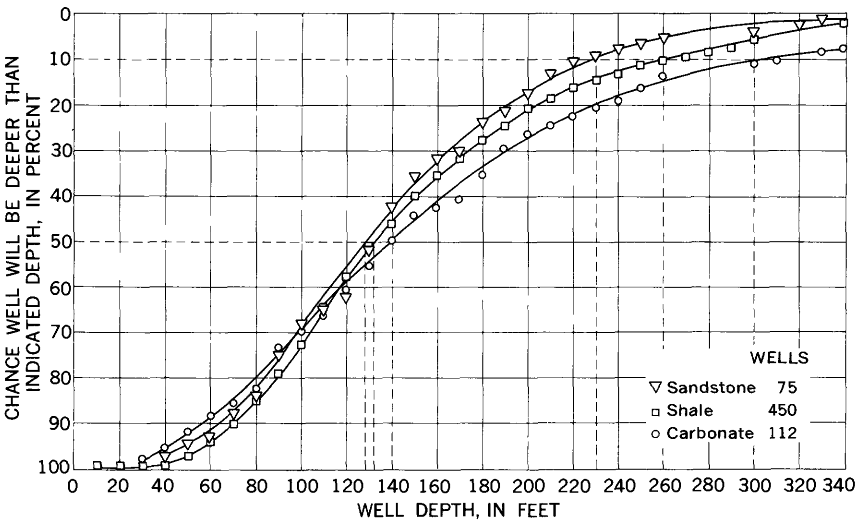


FIGURE 28.—Frequency distribution of well depths with respect to rock type.

The most important observation made from the graphic presentation in figure 28 is that the depths of domestic wells in shale, sandstone, and carbonate rock are nearly identical. About 50 percent of the bedrock wells are deeper than 135 feet. The lack of dependence of well depth on consolidated aquifer rock type is due to the nature of the porosity and permeability in these rocks. As previously stated, the porosity and permeability of consolidated aquifers in Orange and Ulster Counties are dependent upon the degree of fracturing in the rock.

Below about 130 feet, wells drilled for the same yield in carbonate rock are generally deeper than wells drilled in sandstone and shale. Ten percent of the sandstone wells are deeper than 230 feet, 10 percent of the shale wells are deeper than 260 feet, and 10 percent of the carbonate wells are deeper than 300 feet. The curves in figure 28 may be used to determine probable well depths and hence, probable costs for a home-supply well. The figure indicates that there is only a 10-percent chance that a well in shale will be deeper than 260 feet, but there is a 50-percent chance that a well in shale will be deeper than 132 feet.

As mentioned before, topographic location also influences well depth. Figure 29 shows cumulative depth-frequency curves for wells tapping shale located on hilltops, hillsides, and valley flats. Although wells drilled on hilltops are only slightly deeper than wells drilled on hillsides, wells drilled in valley flats are significantly shallower than

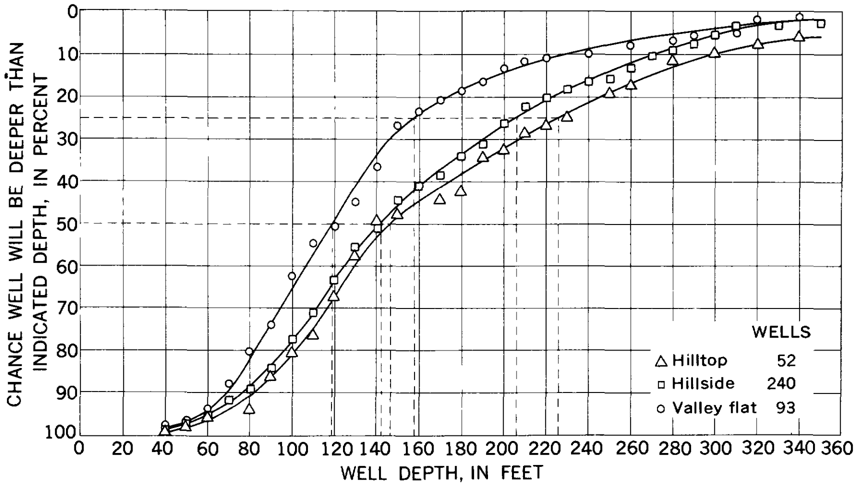


FIGURE 29.— Frequency distribution of well depths with respect to topographic situation.

hilltop and hillside wells. For example, 25 percent of wells in the valley flats are deeper than 158 feet, whereas 25 percent of wells on hilltops are deeper than 226 feet. The curves presented in figures 28 and 29 are a guide to estimating the average cost of privately owned domestic wells and therefore are particularly useful for water-supply planning. Sufficient data were not available to compare wells in various topographic situations in sandstone and carbonate aquifers, but the relationships between topography and depth observed in shale are expected to be similar in sandstone and carbonate rock.

When estimating probable well depth for a specific site, other hydrologic and geologic data can be very useful. That part of a bedrock well which penetrates overburden is cased and does not yield water. Table 3 shows that wells located in areas of thick overburden are deeper because they have greater thicknesses of overburden to penetrate before they reach the bedrock aquifer.

TABLE 3.—*Well depths and thickness of overburden (depth to bedrock aquifer) in feet*

Thickness of overburden	Average depth of wells	Number of wells in sample
<10	123	52
10 to 25	121	96
25 to 50	123	67
50 to 100	166	112
100 to 200	278	51
>200	293	10

Individual well sites may be evaluated with respect to depth of overburden to estimate probable well depths. The well-data table and well-location maps in "Ground-water basic data Orange and Ulster Countries, New York," by Frimpter (1970) may be used to estimate depth to bedrock near existing wells, because the casing length reported in the well table is almost always equal to depth of overburden. The geologic sections (pl. 4) in this report and graphic logs in the basic data report are also sources of depth to bedrock information. Geologic field observation including identification of bedrock outcrops is particularly useful where existing well data are not available.

The porosity and permeability of the unconsolidated sediments overlying the bedrock aquifer have a great effect on the rate of recharge. Thick layers of till with low permeability allow only very slow recharge to the underlying bedrock aquifer, whereas water-saturated sand and gravel overlying the bedrock acts as a source of rapid recharge. Wells tapping bedrock beneath sand and gravel have adequate yields at shallow depths because of the abundant supply of recharge water in the sand and gravel. Many bedrock wells with high yields tap aquifers overlain by saturated sand and gravel.

Geologic and hydrologic conditions must be evaluated for each well location to make reasonably accurate estimates of probable well depths. The records of existing wells near a proposed drilling site are frequently the best indication of local hydrologic and geologic conditions, but because of the variable permeability of bedrock aquifers, many exceptions should be expected.

QUALITY OF WATER

Chemical analyses of 128 samples of well water are given in "Ground-water basic data Orange and Ulster Counties, New York" by Frimpter (1970). The occurrence and effect of various chemical constituents commonly found in ground water are outlined in table 4, and a more detailed discussion of these constituents is presented in the following text. All constituents in this text are reported in mg/l, (milligrams per liter), which is a ratio of the weight of dissolved mineral to 1 liter of the solution.

An outstanding characteristic of ground water is the general constancy of chemical quality with the passage of time. Analyses have been made almost yearly since 1952 on samples from wells 130–405–1, 129–405–3, 129–405–2, 129–406–1, and 130–415–2 in the Hudson River lowlands. There is no large variation in chemical quality of these wells, which all tap the same aquifer; most importantly, the quality of water from these wells has been constant with respect to time. The constancy of ground-water quality as compared with the variability of surface-water quality (in terms of dissolved solids) is shown in figure 30. Although the chemical quality of ground water changes only slightly with time, the chemical quality of stream water fluctuates seasonally.

TABLE 4.—*Chemical constituents in water, their occurrence, and effect*

Chemical constituent	Occurrence	Effect
Silica (SiO_2)	In all natural water in varying concentrations; ground water generally contains more silica than surface water.	Forms boiler scale and deposits on turbine blades.
Iron (Fe) and manganese (Mn).	In practically all natural water; generally, smaller amounts are found in surface water than in ground water.	Iron concentrations of 0.3 mg/l or more stain laundry, porcelain fixtures and other materials; imparts a mineral taste to water and impairs the taste of beverages including coffee and tea.
Calcium (Ca) and magnesium (Mg).	In all natural water; highest concentrations found in water in contact with carbonate rock and gypsum.	Soap consuming, forms an insoluble curd; deposits in pipes and boiler tubes; is the main cause of hardness in water.
Sodium (Na) and potassium (K).	In all natural water; in very low concentrations of alkalies, concentrations of sodium and potassium are about equal. As concentration of alkalies increases, proportion of sodium increases. De-icing road salt is the most common source of abnormal amounts of sodium in ground water in the two counties.	Large amounts may cause foaming in boiler operation; in irrigation water, large amounts are injurious to the soil; low sodium diets are recommended for persons with some circulatory diseases.
Bicarbonate (HCO_3)	In all natural water; larger concentrations present in water in contact with decaying organic matter and carbonate rocks.	Large amounts may affect taste of drinking water. Large quantities in combination with sodium are injurious to the soil.
Sulfate (SO_4)	In most natural water; larger amounts in water in contact with gypsum.	In conjunction with calcium and magnesium forms permanent hardness and hard scale in boiler operation; concentrations of sodium sulfate and magnesium sulfate greater than 200 mg/l may have a laxative effect.
Chloride (Cl)	In most natural water; larger amounts in contaminated waters. De-icing road salt is one of the most common sources of abnormal quantities of chloride in ground water in the two county area.	Taste of drinking water affected when amounts of more than about 250 mg/l are present; corrosiveness is also increased.
Fluoride (F)	In most natural water in small concentrations.	About 1.0 mg/l is helpful in reducing incidence of tooth decay in small children, but higher concentrations may cause mottled enamel on teeth.
Nitrate (NO_3)	In most natural water; contamination by sewage, fertilizer, and organic material increases quantity present.	Small amounts have no effect; 45 mg/l or more has been reported to produce methemoglobinemia in infants, a disease which may be fatal; large amounts of NO_3 indicate pollution.
Hydrogen sulfide (H_2S)	Found dissolved in ground water. In Orange and Ulster Counties, minute quantities occur only in the dark shale and sandstone. About 24 percent of the wells tapping shale in these two counties yield some H_2S.	Disagreeable odor, may tarnish silverware, deleterious effect on photographic developing solutions, usually causes water to be acidic and corrosive.
Natural gas	Occassionally dissolved in ground water in sedimentary rock.	Difficult to detect because it is odorless, tasteless, and colorless; this gas can form a dangerous explosive mixture with air.

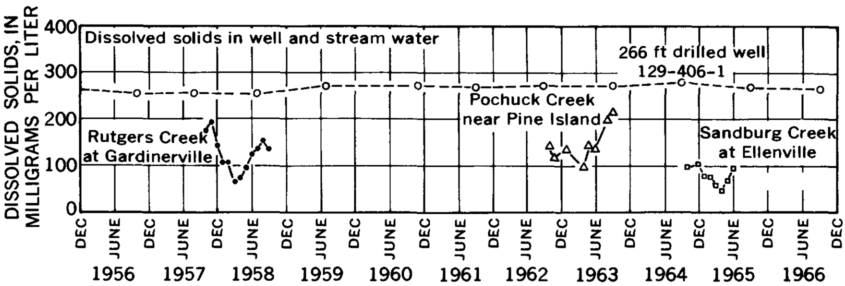


FIGURE 30.—Dissolved solids in ground and surface water.

DISSOLVED SOLIDS

The concentration of dissolved solids in samples of water from wells and springs in Orange and Ulster Counties ranged from 42 to 1,470 mg/l. Only two samples had a concentration of dissolved solids in excess of 500 mg/l, the limit recommended for drinking water by the U.S. Public Health Service (1962, p. 7).

The concentration of dissolved solids in ground water varies with the solubility of the host rock. The sample from well 142–408–2 is not believed to represent natural conditions and therefore was not included in the averages presented in the table below. Carbonate rocks are the most soluble rocks occurring in Orange and Ulster Counties, and the relatively high solubility of the shale and greywacke is due to the carbonate minerals they contain. Greywacke is a dark sandstone with grains of many different minerals.

Average dissolved solids

Rock type	Number of samples	Milligrams per liter
Sandstone	5	109
Crystalline	3	105
Shale and greywacke	26	218
Carbonate	4	266

The concentration of dissolved solids in water from sand and gravel aquifers is variable but is somewhat dependent on the mineral composition of the sand and gravel grains. Gravel containing limestone pebbles generally yields a more highly mineralized water than gravel containing quartz pebbles because limestone is more soluble than quartz. The amount of dissolved solids normally increases with depth, although this does not appear to be a significant trend in Orange and Ulster Counties.

SPECIFIC CONDUCTANCE

Specific conductance is a measurement of the ability of water to conduct electricity and is determined to a large extent by the amount of dissolved mineral matter in the water. Specific conductance is generally proportional to the concentration of dissolved solids, as may be inferred from the straight-line plot in figure 31. As indicated by the slope of the line on this graph, the amount of dissolved solids can be approximated by multiplying the specific conductance by 0.61. The type of mineral matter dissolved in the water also can affect the specific conductance, but because the same constituents are normally dissolved in all the water in this two-county area, this effect is insignificant.

HARDNESS

The alkaline earth metals, calcium and magnesium, are largely responsible for hardness in water. In this report, the hardness produced by all substances in the water is expressed as equivalent concentrations of calcium carbonate. Elsewhere, it is sometimes reported in units of weight called grains, one grain per gallon being equivalent to about 17.1 mg/l.

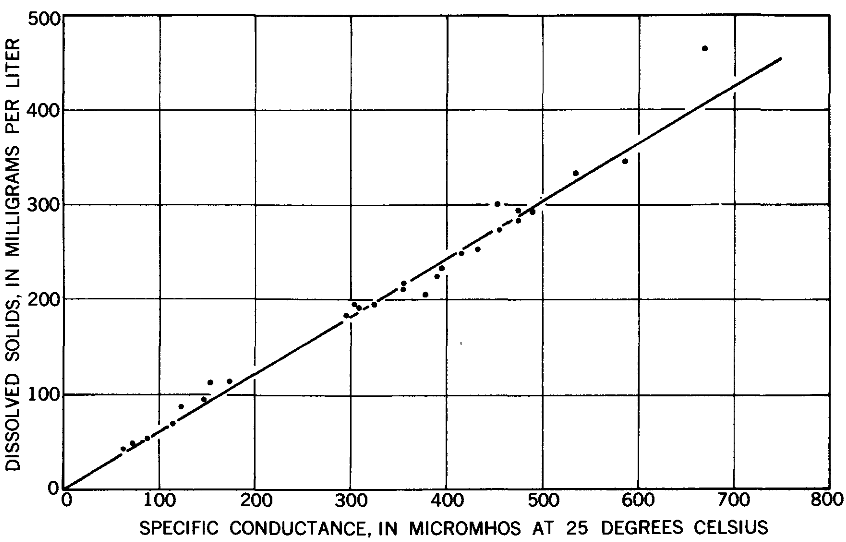


FIGURE 31.—The relation of specific conductance to dissolved solids.

The hardness of 50 samples of well and spring water ranged from 23 to 1,470 mg/l, and averaged 142 mg/l. Hardness of water is commonly described as:

Soft	0–60 mg/l
Moderately hard	61–120 mg/l
Hard	121–180 mg/l
Very hard	over 180 mg/l

Therefore, well and spring water in this area ranges from soft to very hard, but is generally hard. The general statements in the section of this report pertaining to the relation between dissolved solids and type of aquifer are also valid for the relation between hardness and aquifer because those elements which cause hardness in water are also responsible for most of the dissolved solids in the ground water of these two counties.

The average calcium content of 21 samples of well and spring water was 35 mg/l, and the average magnesium content of the samples was 9.5 mg/l. Calcium and magnesium generally occur in water with bicarbonate and sulfate, with bicarbonate being more common than sulfate in the area. The hardness of water which contains mostly bicarbonate is said to be temporary. Heating water with temporary hardness causes precipitation of calcium and magnesium carbonate and, therefore, removal of the hardness. The precipitate thus formed can be a problem because it cakes as a film or crust on surfaces of containers, such as on the bottoms and sides of cooking utensils. Heating coils of electric hot water heaters are particularly susceptible to the formation of such mineral crust. This crust of calcium carbonate, sometimes called a lime deposit, seriously reduces the effectiveness of the heating coil.

Fortunately, water can be softened through the removal of calcium and magnesium by ion-exchange water softeners. In these softeners, water is passed through a tank containing synthetic resins that exchange sodium for the calcium and magnesium in the water. For example, water from well 129–405–2 is naturally hard; a sample taken from this well had a total hardness of 150 mg/l. A later sample, taken after the water passed through a softener, had a total hardness of only 2 mg/l. While the calcium and magnesium are drastically reduced by softening, the sodium content is increased from 13 to 77 mg/l.

Water softening units must be recharged periodically with sodium because sodium is lost to the water during the softening process. This recharging can be done by putting a strong sodium chloride brine through the softener. The exchange of ions is reversed—calcium and magnesium are removed from the resin and replaced with sodium. Disposal of the brine should be controlled so that it will not enter the aquifer. Salt used for regeneration of a water softener caused the increase of sodium and chloride in well 130–405–1 in 1965.

HYDROGEN SULFIDE

Sulfur water is water that contains the dissolved gas, hydrogen sulfide (H_2S). The gas has a disagreeable odor, frequently described as the odor of rotten eggs. The average person can detect as little as 0.05 mg/l of hydrogen sulfide (McKee and Wolf, 1963, p. 200). In larger concentrations in air (0.5 percent), the gas is lethal to man (Frey, 1958, p. 372), but the minute quantities present in the ground water of Orange and Ulster Counties have never been reported to produce ill effects. In a few instances in the two-county area, sulfur water has been reported to tarnish household silverware. The presence of this gas is also undesirable in water used for photographic developing processes.

With rare exceptions, interlayered dark gray to black shale and sandstone is the only aquifer that yields hydrogen sulfide bearing water in the two-county area. Nearly 24 percent of the wells drilled in this aquifer yield, or have yielded, water containing the gas.

Plate 1 indicates those areas underlain by rocks that may yield sulfide-water bearing hydrogen sulfide. Quantities of dissolved hydrogen sulfide vary from well to well and vary seasonally in some wells. Many well owners report that, whereas their well originally produced sulfur water, daily use seems to curtail the appearance of the gas. On the other hand, prolonged heavy pumping may draw sulfur water into a well that does not normally produce such water.

The exact source of hydrogen sulfide has not been determined. The mineral pyrite (fool's gold, FeS_2) is found disseminated throughout these shaly rocks and is a possible source of the sulfur (Hendrickson and Kriefer, 1964, p. 57–58). Although pyrite concentrations vary from place to place, the mineral occurs nearly everywhere in the dark shale. The mineral's apparent absence in other rocks in the two-county area may explain the absence of hydrogen sulfide in the water from those rocks. Hydrogen sulfide may be produced by the action of anaerobic bacteria living in the ground water (Hem, 1959, p. 103, 223). Natural gas might also be involved in the formation of hydrogen sulfide because it also occurs in these same rocks.

Generally, the deeper a well is in the shale aquifer, the more likely it is to produce water bearing hydrogen sulfide. This situation is modified by ground-water circulation within the aquifer. In ground-water discharge areas (valley bottoms) even shallow wells yield sulfur water. The circulation of ground water in discharge areas is upward from zones that possess conditions favorable to the formation of hydrogen sulfide. In recharge areas, such as hilltops and hillsides, only the very deep wells were found to yield water bearing hydrogen sulfide. Water circulation in recharge areas is predominantly downward, and the addition of fresher, oxygenated meteoric water depresses the zone of anaerobic conditions where the gas can form.

Figure 32 is an idealized cross-section view of Rutgers Creek valley near Westtown in Orange County. The dashed line in this cross section represents an imaginary boundary between nonsulfurous and sulfurous water, and the arrows indicate the direction but not magnitude of flow of ground water. Wells producing from below the dashed line will yield sulfurous water and those producing from above it will not produce sulfurous water. Well drillers with experience in this two-county area frequently advise home owners to stop drilling once a sufficient supply has been obtained rather than risk drilling deeper for unneeded extra yield that might contain hydrogen sulfide.

Frequently, wells may be cleared of sulfurous water by prolonged pumping. By pumping a well continuously for a few days, recharge may be induced to replace the sulfurous water. This replacement water would be drawn primarily from the overlying, more permeable rock

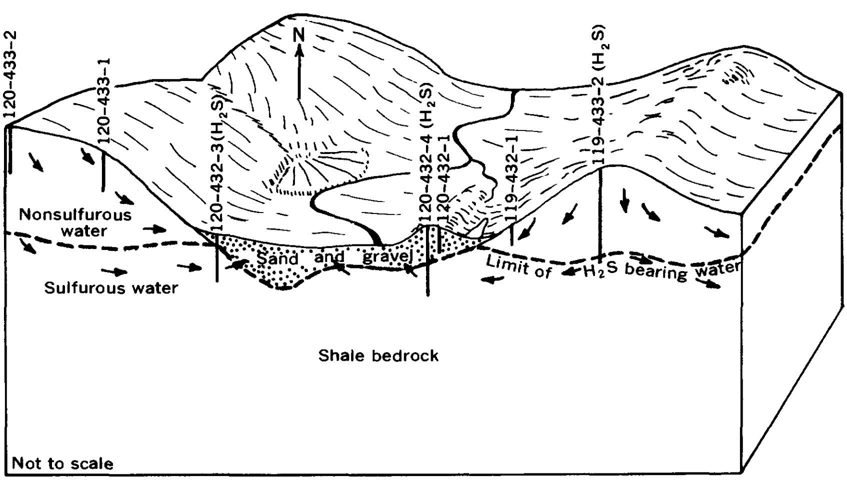


FIGURE 32.—Schematic block diagram of Rutgers Creek Valley.

which contains nonsulfurous water. In this manner, the quality of the well water may be improved. This technique has been found to be successful in many places and is recommended by well drillers in the two-county area. If a well's supply is obtained from fractures more than a few tens of feet below the lower limit of sulfur free water, however, it is unlikely that pumping will reduce the amount of sulfurous water.

On the other hand, abnormally high pumping rates or extended pumping have been known to draw sulfurous water into wells which normally do not produce sulfurous water. This occurs because the additional volume of water must come from deeper parts of the rock aquifer that normally do not yield water to the well.

In a few instances, sulfurous water occurs above nonsulfurous water, and this statement would at first glance appear to contradict statements in the foregoing paragraphs. Such occurrences can be expected in view of the nature of the bedrock aquifer. A well may penetrate a fracture supplied by deeper fractures bearing hydrogen sulfide, and below this fracture the well may penetrate a fracture supplied only by higher fractures which do not bear the gas (fig. 33).

Hydrogen sulfide gas may be removed from water by aeration, chlorination, or potassium permanganate treatment. In the last two of these methods, the hydrogen sulfide is oxidized to form water and elemental sulfur which is odorless and tasteless but which is also insoluble and may cause the treated water to be cloudy or turbid. Aeration releases the gas but does not oxidize it to a measurable extent.

CHLORIDE

There are no rock strata that contain natural rock salt in Orange and Ulster Counties. Abnormally high chlorides in this area can almost always be attributed to manmade sources, such as highway deicing salt. Contamination of wells near stockpiles of road salt is becoming increasingly common. Rainwater leaches salt from the stockpiles and carries it into the ground and thus to the wells. To prevent this sort of local contamination, the salt or sand and salt mixtures must be protected from the leaching action of water. If water is not permitted to dissolve the salt, it cannot contaminate well supplies.

Well 123–420–2, drilled at a site where calcium chloride and sodium chloride were previously stockpiled, yielded nonpotable water. The water reportedly contained more than 1,900 mg/l chloride soon after the well was drilled in 1953. Five years later the chloride concentration was reported to be 220 mg/l, which is below the 250 mg/l limit recommended by the U.S. Public Health Service (1962, p. 7) for drinking water.

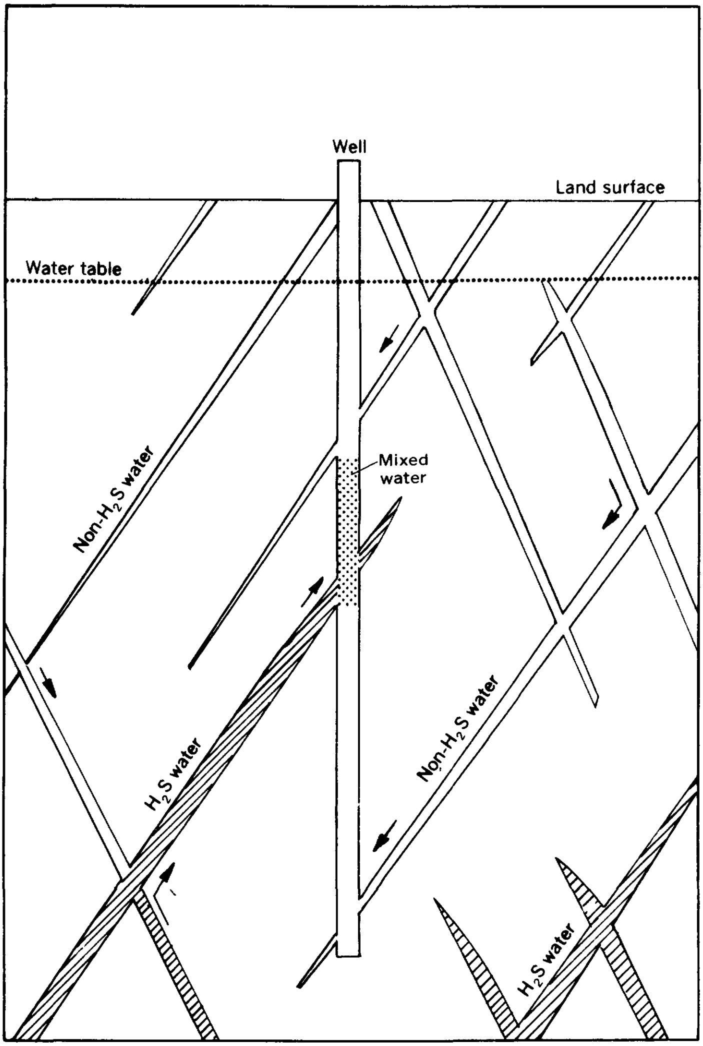


FIGURE 33.—Occurrence of sulfurous water above nonsulfurous water.

Calcium chloride and sodium chloride, well known for their corrosive effect on metal, aided in the almost complete corrosion of the lower 10 feet of the 8-inch-diameter steel casing in well 123–420–2. The salts probably aided the development of leaks in a fuel-oil storage tank buried nearby which caused the subsequent appearance of fuel oil in the well.

The Hudson River may be another source of chloride in ground water. Salt water from the Atlantic Ocean mixes with fresher Hudson River water to form brackish water. The position of the diffuse boundary between fresh water of the upper Hudson and brackish water of the lower Hudson is largely determined by the quantity of fresh water flowing toward the ocean. In the spring, when more water is flowing, the boundary has been as far downstream as Haverstraw Bay. In a dry season it has been as far upstream as Hyde Park. Therefore, water with a high chloride concentration might possibly be pumped from wells in aquifers that are recharged by the Hudson River south of Hyde Park. Normally, the ground water is flowing toward the river and is not saline. However, withdrawal of ground water from wells adjacent to the river could reverse the hydraulic gradient and induce recharge from the river. This induced recharge could have a high chloride content if the brackish water boundary were upstream of the infiltration site. The high concentration of chloride in well 127–401–2 in August 1961 may have been caused by the infiltration of brackish water from Moodna Creek, a tidal tributary to the Hudson River at New Windsor.

In 1965 the chloride content of water from well 130–405–1 increased sharply from a 12-year average of 3.1 to 198 mg/l. This abnormal change in chloride content was traced to spillage of salt used for the regeneration of a water softener.

Other cases of high chloride concentrations can be traced to industrial waste water because some industrial wastes are very high in chloride. Septic tanks, cesspools, and fertilizers are also common contributors of chloride to ground water. High chloride content is, therefore, frequently an indication of contamination resulting from man's activities.

IRON AND MANGANESE

One of the most troublesome chemical constituents in ground water in Orange and Ulster Counties is iron. Concentrations of iron greater than 0.3 mg/l impart an undesirable metallic mineral taste to water and stain laundry and plumbing fixtures brown. In addition, water containing iron may have an undesirable brown color. Manganese is much less common than iron in ground water, but generally appears with iron and has very similar properties. The U.S. Public Health Service (1962, p. 7) recommends that water containing more than 0.3 mg/l of iron of 0.05 mg/l of manganese should not be used where more suitable supplies are available.

Sand and gravel aquifers in Orange and Ulster Counties may yield water with high iron and manganese concentrations. This situation is sometimes troublesome because the greatest ground-water reserves of the two-county area are in sand and gravel aquifers. Exploration of sand and gravel aquifers frequently discloses variable iron concentrations within short horizontal and vertical distances, but the exact cause of this phenomenon is not known.

A reducing and acid environment is favorable for the solution of iron. Consumption of oxygen during the decay of organic material can produce a reducing environment containing carbonic acid. Production of carbon dioxide by oxidation of carbonaceous matter and by reduction of sulfate in the presence of carbonaceous matter makes water acidic and reducing and, therefore, capable of dissolving iron. Wells might logically obtain iron-bearing water if they tap aquifers which are recharged by acidic oxygen-deficient water, and many streams in the study area have an oxygen deficiency due to an abundance of organic material from industrial and municipal wastes.

Bacteria that use iron in their life processes can cause serious problems in screened wells and distribution systems. These bacteria, which precipitate iron from solution, clog well screens and reduce well yields. Clots of bacteria may also grow in a distribution system only to break loose and appear at a tap.

Iron is not a common problem in the water from the dark shales and graywacke sandstones, but it does appear in the shallower parts of the shale aquifer northeast of New Paltz. Extending casing to a greater depth in the shale can shut off this water, but the remedy may also reduce well yield. In Orange and Ulster Counties limestone seems to yield iron-bearing water more frequently than the dark shale and sandstone. Iron-bearing water from carbonate aquifers sometimes occurs in weathered fractures and fault zones containing brown clay. Red shale and sandstone that contain iron-bearing water extend from the New Jersey border at Greenwood Lake northward to Mountainville. An unfiltered sample of water from well 114–417–1 contained 9.2 mg/l iron. Red shale layers in the quartzite conglomerate of the Shawangunk Mountain ridge sometimes also yield water of high iron concentration.

NATURAL GAS

Although occurrences are rare, a few wells in the two counties are reported to have yielded natural gas. The 605-foot well, 147–419–3, at Pautaukunk in Ulster County is one of these. In most wells the quantity of gas produced is small and generally is rapidly exhausted.

The odorless, tasteless, colorless natural gas is methane (CH_4), which burns with a blue flame and may form an explosive mixture with air. It is an explosion and fire hazard anywhere it is allowed to collect—in the well, in the plumbing, or in poorly ventilated areas such as basements. Any gas may separate from water in the plumbing, more noticeably in the hot water pipes, and cause a bumping sound and spurting at the spigot when the water is turned on.

Natural gas occurs dissolved in ground water very much like carbon dioxide (CO_2) occurs dissolved in soda water, but generally it occurs in much lower concentrations. Heating, agitating and depressurizing allow the most rapid and complete degassing of water. The removal of methane from water may best be accomplished by aeration.

NITRATE

Seventeen samples of well or spring water were analyzed for nitrate (NO_3) content, and concentrations ranging from 0.0 to 5.6 mg/l were found. High nitrate concentrations in water have been shown to be toxic to infants, but the tolerance of adults to nitrate is much greater. High nitrate concentrations can usually be traced to manmade sources such as industrial wastes, cesspools, chemical fertilizers, and farmyard manure piles.

FLUORIDE

Fluoride concentrations were measured in water from 17 groundwater sources in the two-county area. These concentrations ranged from 0.1 to 1.0 mg/l. The U.S. Public Health Service (1962) recommends lower, optimum, and upper control limits for fluoride based on the annual average of maximum daily air temperatures. For the two counties exclusive of the Catskill Mountain area these limits are 0.8 mg/l (lower), 0.9 to 1.0 mg/l (optimum), and 1.2 to 1.3 mg/l (upper).

SILICA

No limits of silica (SiO_2) concentration have been recommended for domestic water supplies. Low silica cencentration is required in water used for ice manufacture and high-pressure steam generation. Of the 17 samples analyzed for silica content, the highest concentration was 16 mg/l and the lowest was 5 mg/l.

SODIUM

No limits have been recommended for sodium in domestic water supplies. Low sodium intake, however, is prescribed for certain heart diseases. Thirteen samples of ground water were analyzed for sodium, and concentrations ranging from 1.1 to 134 mg/l were found. Eighty-two samples were analyzed for sodium plus potassium, and concentrations averaged 32.7 mg/l and ranged from 7.6 mg/l to 159 mg/l. High sodium contents may exist in softened water. The exchange of sodium for calcium and magnesium in water softeners was explained more completely in the section pertaining to water hardness.

TEMPERATURE

Ground-water temperatures are more constant than surface-water temperatures. Air and land surface temperatures fluctuate through a wide range during the year and are higher than the temperature of ground water in the summer and lower than the temperature of ground water in the winter. For this reason spring water seems cool in the summer but rarely freezes in the winter. A study of ground-water temperatures was made to gain a more complete knowledge of their variations.

Water temperature profiles were measured in three wells near Monroe at different times of the year by using a remote sensing thermometer. The results of these measurements are shown in figure 34. The seasonal ground-water temperature fluctuations are greatest just below land surface and decrease rapidly with increasing depth. At wells 119–415–3 and –4, the seasonal temperature fluctuations became insignificant about 23 feet below the water table. At well 119–414–2 the water table is over 50 feet below land surface, and seasonal water temperature fluctuations are small. Aquifers in which ground-water temperatures fluctuate greatly are probably near land surface or are recharged by nearby surface-water sources. When water-temperature changes lag far behind seasonal air temperature changes, more remote recharge areas are probable. Ground-water temperatures from depths greater than 50 feet nearly always averaged between 9.5° and 12°C (49° and 54°F).

SUMMARY

Among the aquifers of ~~Orange and~~ Ulster Counties, the glacially derived sand and gravel deposits in the valleys possess the greatest potential as ground-water reservoirs. The largest ground-water reservoir is contained in sand and gravel deposits in the Neversink River and Basher Kill valleys between Port Jervis and Summitville. An estimated 70 mgd could be developed in the Neversink River valley between Port Jervis and Cuddebackville, and an estimated 40 mgd could be developed between Cuddebackville in Orange County and Summitville in Sullivan County. This yield of 110 mgd could be greatly increased, perhaps doubled, by planned recharge through pits and ditches. About 18 mgd is available from sand and gravel in the valley between Phillipsport and Wawarsing. The village of Ellenville taps this sand and gravel aquifer with a single shallow well that can yield over 1 mgd.

Streams flow across, and are in hydraulic contact with, the sand and gravel aquifers between Port Jervis and Wawarsing. These streams are the main sources of recharge to the aquifers but during base-flow

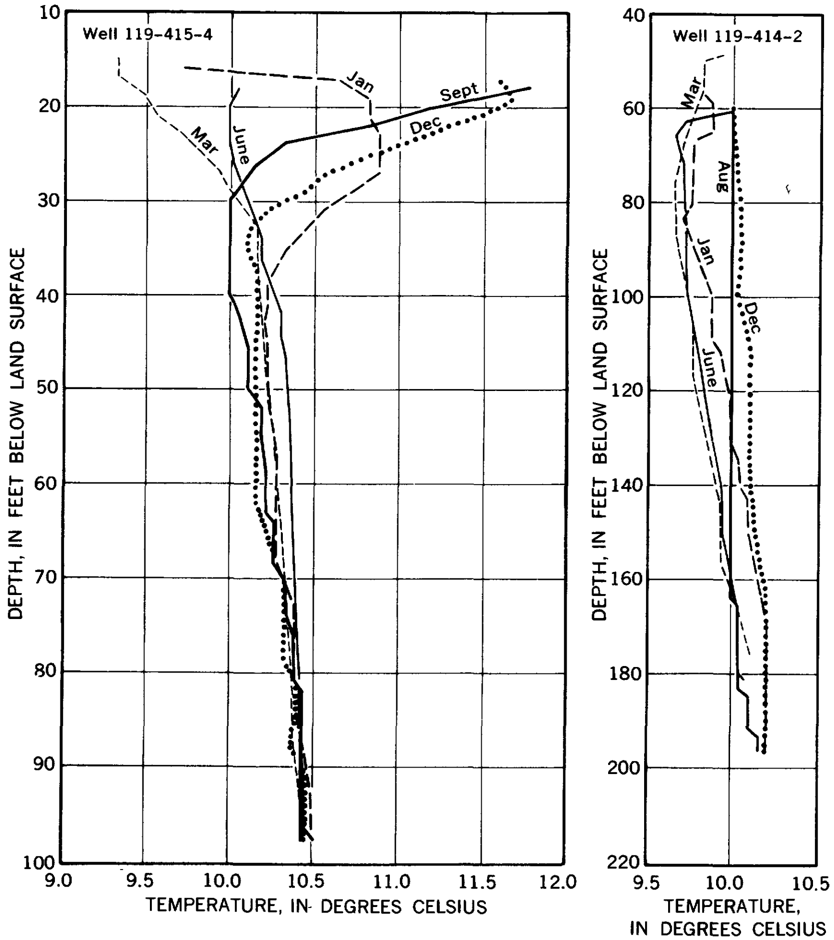
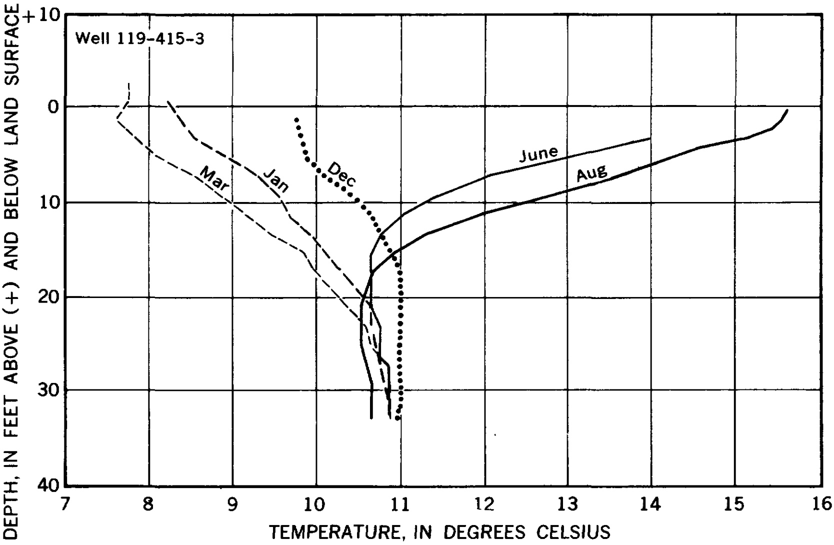


FIGURE 34.—Seasonal ground-water temperature variations.

conditions the aquifers became the main sources of water to the streams. The water in the streams and aquifers forms a single aquifer-stream drainage system, and within this system ground water and surface water cannot be considered independent entities. Efficient management of these water resources requires an understanding of the interchange within this system.

In Ulster County some of the flat-bottomed valleys that look most promising as ground-water sources are filled with glacial-lake clay and generally do not contain suitable aquifer material. Both the northern part of the Wallkill River valley and the lower Esopus Creek valley contain very little coarse granular sediment. Although no buried sand



and gravel aquifers have been found in the northern part of the Wallkill River valley, a few gravel lenses have been found in the lower Esopus Creek valley. Numerous test holes in the Esopus Creek valley near Kingston reveal no thick or extensive layers of either sand or gravel.

Small deltas of sand and gravel form aquifers at the confluence of Saw Kill and Esopus Creek and at Ruby, Mount Marion, and Veteran where other tributaries enter the valley. The aquifer at the confluence of Saw Kill and Esopus Creek is probably capable of producing more than 2 mgd because the two nearby streams are sources of recharge to the aquifer. The deltas farther to the north could probably yield a total of almost 3 mgd.

A number of smaller glacial outwash and kame deposits in Orange County may be capable of yielding around 1 mgd each. Those aquifers that are in direct hydraulic contact with surface streams generally are capable of producing more water than the aquifers that can be recharged only by direct infiltration of precipitation. A sand and gravel aquifer at the north end of Greenwood Lake is in hydraulic contact with the lake and might yield over 4 mgd.

The stream valleys in the Catskill Mountains contain only thin shallow deposits of sand and gravel. The downwasting of glacial ice in this region precluded the formation of outwash sand and gravel, but later stream action has produced a layer of coarse sand and gravel on the surface of these valley bottoms. Infiltration galleries or wells near the streams and in the shallow gravel may yield larger quantities of ground water, but their yields could not exceed the adjacent stream discharge.

Water produced from sand and gravel aquifers in Orange and Ulster Counties is generally hard, and in a few areas dissolved iron and manganese are present in quantities greater than the limits recommended by the U.S. Public Health Service. Water treatment can generally reduce the iron and manganese to acceptable concentrations.

Ground water of good quality may be obtained in sufficient quantities for individual home use almost anywhere in Orange and Ulster Counties. Fractured consolidated rock is the source of most of this supply, and there seems to be little variation of well yield in the consolidated rock types present in the two counties. Moderate quantities of water, such as those capable of supplying housing developments of about 100 homes, are sometimes obtained from wells in the consolidated rock aquifers. Most of these supplies are found in valleys, and frequently more than one well must be drilled to obtain an adequate supply. Because of the rather small storage capacity of consolidated rock, these wells are often quite susceptible to drought conditions.

Large quantities of water suitable for municipal and industrial use are sometimes found in the consolidated rock, but such occurrences are uncommon. The major fault zones associated with the crystalline rocks of southeastern Orange County and with the carbonate rocks in the valley system extending from Port Jervis to Rosendale may be sources of large quantities of ground water.

Water obtained from the bedrock aquifers is generally of good chemical quality but is often moderately hard to very hard. Deeper wells in shale frequently yield water containing hydrogen sulfide. Wells tapping the red shale and sandstone formations in southeastern Orange County commonly yield water containing undersirable quantities of iron, and other consolidated rock aquifers occasionally yield water with high iron concentrations. At depths greater than 50 feet, below the zone affected by seasonal temperature fluctuations, sampled ground-water temperatures average about 11°C (52°F).

RECOMMENDATIONS FOR FUTURE STUDY

Detailed mapping of the geology and the hydraulic properties of the aquifer northeast of Port Jervis should be given first consideration for future ground-water studies in the two-county area. The aim of this work should be to refine the ground-water yield estimate and to provide the basis for a master plan for optimum development of the aquifer. Items such as optimum well spacing and positioning of recharge basins should be determined or planned. On the basis of the geologic and hydrologic data obtained, an analog model could be built and used to manage the development of the aquifer. A similar study of the aquifer in the Sandburg Creek valley near Ellenville might also be undertaken. The smaller unconsolidated aquifers will probably be explored by municipalities and private enterprise as the demand for water increases.

Drilling and testing of the fault-zone aquifers in the Rondout Creek valley and in the Hudson Highlands might reveal additional large supplies. Study of these aquifers would also add to the understanding of the local geology and hydrology of fault-zone aquifers.

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