STREAM-AQUIFER RELATIONS AND YIELD OF STRATIFIED-DRIFT AQUIFERS IN THE NASHUA RIVER BASIN, MASSACHUSETTS

by Virginia de Lima

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CONVERSION FACTORS AND ABBREVIATIONS

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
cubic foot (ft^3)	0.02832	cubic meter (m^3)
	Flow	
gallon per minute (gal/min)	3.785	liter per minute (L/min)
million gallons (Mgal)	3,785	cubic meter (m^3)
million gallons per day (Mgal/d)	3,785	cubic meter per day (m^3/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per square mile [(ft^3/s)/mi^2]	0.01093	cubic meter per second per square kilometer [(m^3/s)/km^2]
inch per year (in./yr)	25.4	millimeter per year (mm/yr)
	Hydraulic Conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity	
foot squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)

Temperature

Temperature in degrees Celsius (°C) can be converted to degrees Fahrenheit (°F) as follows: °F = 9/5 °C + 32

Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Mean Sea Level of 1929."

Stream-Aquifer Relations and Yield of Stratified-Drift Aquifers in the Nashua River Basin, Massachusetts

By Virginia de Lima

ABSTRACT

Aquifer yields were estimated for the Pearl Hill-Willard Brooks, Stillwater River, Wekepeke Brook, Still River, Witch Brook, and Catacoonamug Brook aquifers in the central and eastern parts of the Nashua River basin. The stratified-drift aquifers are currently used for water supply and have been identified as possible sources of additional supply for communities in the basin.

Long-term yields from intercepted ground-water discharge and induced infiltration of surface water were estimated for the six aquifers. Estimates of potential aquifer-yield were reduced to maintain streams at low streamflow (95-percent flow duration) and at very low streamflow (99.5-percent flow duration). If low streamflows were maintained, only two aquifers could sustain well withdrawals 99 percent of the time; most of the aquifers, however, could sustain withdrawals ranging from 0.20 to 0.97 million gallons per day 95 percent of the time. If very low streamflows were maintained, all the aquifers could sustain withdrawals ranging from 0.07 to 1.09 million gallons per day 99 percent of the time and withdrawals ranging from 0.49 to 3.62 million gallons per day 95 percent of the time.

Water stored in each of the aquifers would be available for short-term use during periods of no recharge. After 180 days of no recharge, yields from storage would range from 5.8 million gallons per day in the Pearl Hill-Willard Brooks aquifer to 13.1 million gallons per day in the Still River aquifer. Withdrawing these large quantities of water from storage could adversely affect the environment by lowering the water table and depleting aquifer storage.

During 1984-85, the Witch Brook, Catacoonamug Brook, and Still River aquifers were pumped at the highest rates relative to their potential long-term yields--that is 55, 57, and 51 percent of their potential yields, respectively. The regional aquifer along the Squannacook River, adjacent to Witch Brook, was pumped at the lowest rate relative to its potential yield--12 percent.

Infiltration tests at two sites were made to determine the vertical hydraulic conductivity of streambeds. Vertical hydraulic conductivities ranged from 2.0 to 5.0 feet per day. These values probably are typical of small sandy-bottom streams in the Northeast. Numerical models of ground-water flow were used to estimate the well spacing needed to induce infiltration of stream flow at the maximum rate. Well spacing ranged from 1,000 feet in aquifers that yield 1 million gallons per day to each well to 8,000 feet in aquifers that yield 7 million gallons per day to each well.

INTRODUCTION

Most of the Nashua River basin is a hilly, till and bedrock upland in north-central Massachusetts that includes several narrow valleys filled with stratified drift. In 1982, Interstate highway 190 was completed, bisecting the area and connecting the industrial cities of Fitchburg and Leominster (fig. 1) with Worcester, which is south of the basin. The improved transportation has spurred industrial growth and increased local water demand in the eastern and central parts of the basin. The sources of water to meet this increasing demand are the sand and gravel deposits in the valleys.

To gain additional information on the yield of the major sand and gravel aquifers in the eastern and central parts of the Nashua River basin, the U.S. Geological Survey (USGS), in cooperation with the Massachusetts Department of Environmental Management, Division of Water Resources, identified six aquifers for study. During the 3-year study, aquifer yields were calculated for sustainable long-term withdrawal and for short-term withdrawal from storage during prolonged periods of no recharge.

Induced infiltration of water from streams to water-supply wells is a common occurrence and can represent a substantial percentage of aquifer yield in these narrow, thin, and discontinuous aquifers. Not only can the hydraulic connection between stream and aquifer affect the quantity of water available to a nearby well, but also there is growing concern that surface water of impaired quality may affect the quality of ground water pumped by a well. The importance of accurate estimates of vertical hydraulic conductivity of streambed materials is a need recognized throughout the Northeast. The U.S. Geological Survey's Regional Aquifer Systems Analysis studied several sites where induced infiltration occurs (Alan Randall, U.S. Geological Survey, oral commun., 1987), but the study reported here is the first in Massachusetts that used field determinations of the vertical hydraulic conductivity of the streambed to estimate the potential rates of infiltration from streams to aquifers.

Purpose and Scope

This report describes aquifer yields and stream-aquifer relations of six aquifers in the eastern and central parts of the Nashua River basin (fig. 1) that have been identified as the major potential sources of ground-water supply in the Nashua River basin (Brackley and Hansen, 1977). All of the aquifers are in stratified-drift deposits in the valleys of streams tributary to the Nashua River and range in area from 2.6 to 7.0 mi^2 (square miles). The streams that drain the aquifers are Catacoonamug Brook, Pearl Hill and Willard Brooks, Still River, Stillwater River, Wekepeke Brook, and Witch and Bixby Brooks. Two other aquifers of potentially high yield in the eastern part of the basin were not studied: one in downtown Fitchburg, which is unlikely to be developed for drinking water, and one at Fort Devens, which is under the jurisdiction of the U.S. Army.

Because most of the stratified-drift aquifers and streams in the Nashua River basin are hydraulically connected, pumping of the aquifers affects streamflow. The quantity of water contributed from streams to pumped wells at a site in Shirley and at a site in Pepperell (fig. 1) was estimated from infiltration tests and was used to estimate probable stream contributions to pumped wells in the other similar small aquifers in the basin. During the induced infiltration studies, one surface and several ground-water samples were analyzed to determine if water quality could be used to estimate the degree of connection between a stream and the adjacent aquifer.

Previous Investigations

Numerous reports describing the water resources of the Nashua River basin have been prepared for individual towns by private consultants. Regional studies have been done by the Nashua River Watershed Association (1970), the Montachusett Regional Planning Commission (1978), and the U.S. Geological Survey as part of Hydrologic Atlas program (Brackley and Hansen, 1977).

Acknowledgments

The author is grateful for the assistance from the Massachusetts Division of Water Resources and to the employees of municipal water departments for providing information on ground-water exploration in their communities. A special note of appreciation goes to John Lynch, Joan Briggs, and the staff of the water department in Pepperell, and to Vernon Griffith, Dick Hatch, and the staff of the water department in Shirley for their cooperation during the tests to measure induced infiltration.

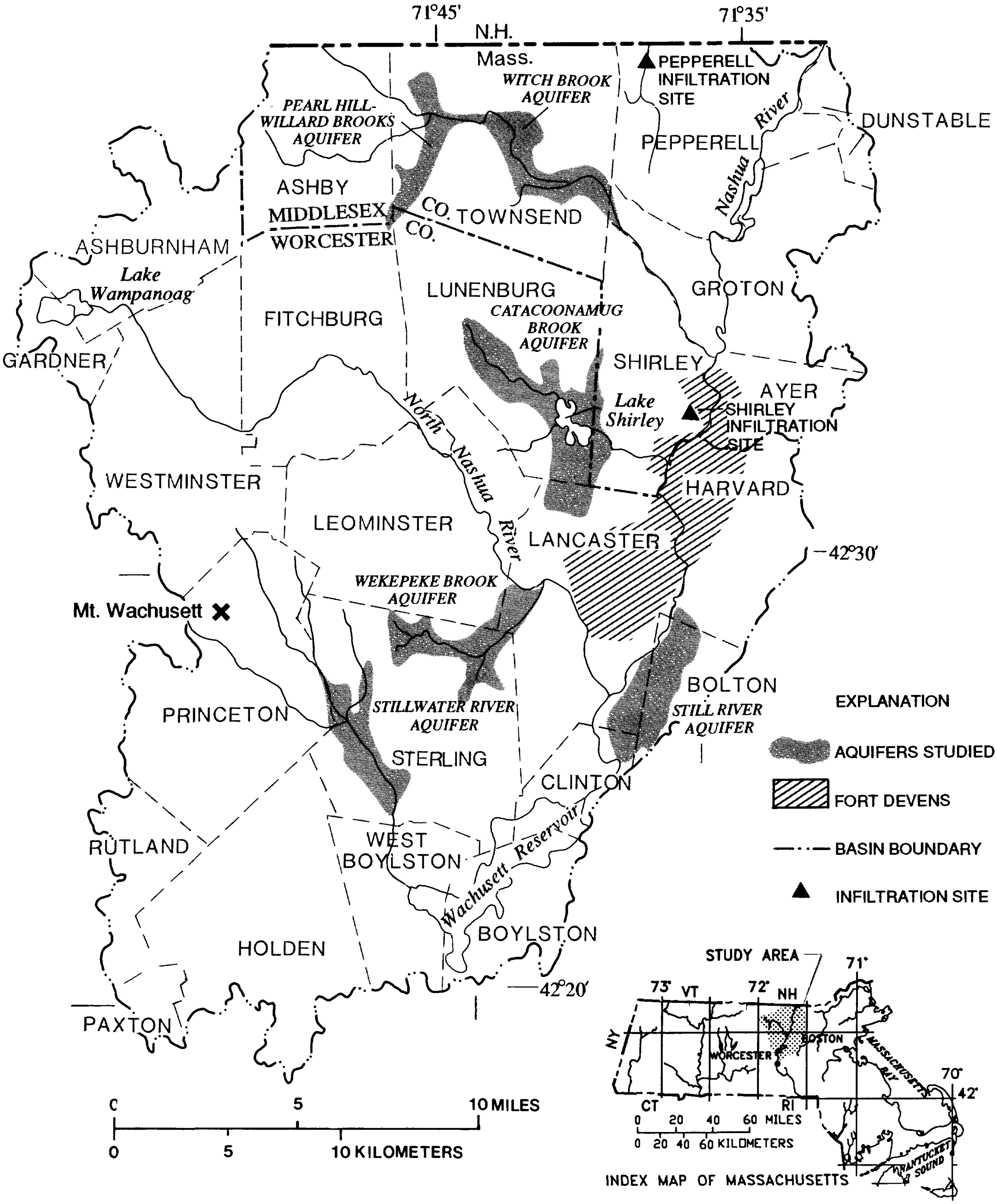


Figure 1.--Location of aquifers studied and infiltration sites in the Nashua River basin.

DESCRIPTION OF THE STUDY AREA

Geographic Setting

The Nashua River basin covers 445 mi^2 of Middlesex and Worcester counties in north-central Massachusetts (fig. 1). The hilly western and northwestern area of the basin is part of the Central Highlands physiographic province and includes Mt. Wachusett, one of the prominent topographic features in eastern Massachusetts. A gently sloping, 300-ft (foot) escarpment forms the western side of the Nashua River valley from Worcester to the New Hampshire border and marks the western edge of the coastal lowlands physiographic province (Denny, 1982). The area east of the escarpment is characterized by low hills and wide flood plains along the northern and southern branches of the Nashua River.

The North Nashua River begins as Whitman River at Lake Wampanoag in Gardner and Ashburnham and flows east and southeast to join the Nashua River (formerly called the South Nashua River) in Lancaster, 5.2 mi (miles) downstream from Wachusett Reservoir. From that point, the Nashua River flows northward into New Hampshire where it joins the Merrimack River at Nashua, New Hampshire. In 1906, the southern branch of the Nashua River was dammed to form Wachusett Reservoir. Since then, almost all the flow of the southern branch has been diverted to become part of the Massachusetts Water Resources Authority's water supply system. The flow released from the reservoir to the river, a minimum of 2.8 ft^3/s (cubic feet per second), is only one fifth of the river's natural flow (Gadoury and others, 1986).

Hydrogeologic Setting

Crystalline bedrock underlies the Nashua River basin. Depths to bedrock range from zero, where rocks are exposed at land surface, to more than 200 ft beneath the Nashua River in Bolton and Lancaster. The topographic relief of the bedrock surface resulted from scour of existing drainage channels by glaciers. Till, a poorly sorted mixture of rock fragments ranging in size from clay to boulders, overlies the bedrock in most of the Nashua River basin. Locally, thick deposits of till form hills in the upland areas of the basin. Stratified drift, a sorted and layered mixture of sediments, is found as ice-contact and deltaic deposits along the sides of valleys and as lake-bottom and outwash deposits in the middle of valleys.

Wells drilled in bedrock are a common source of water to rural homes. Yields of these wells typically are less than 15 gal/min (gallons per minute) and depend on the number, size, and interconnection of the fractures that the well intercepts (Cushman and others, 1953). Dug wells in till have long been used for domestic and agricultural supply. The yield of these wells is usually only a few gallons per minute because of the characteristically low hydraulic conductivity of till. In contrast, wells in stratified drift commonly yield adequate quantities of water for municipal use. In areas of thick, coarse-grained, saturated drift, wells can yield as much as 1,000 gal/min.

Sources of Potential Ground-Water Withdrawals

Ground water withdrawn from stratified-drift aquifers in Massachusetts may be viewed as coming from one or more of the following three sources: intercepted ground-water discharge, induced infiltration of surface water, and storage.

In a humid climate, ground water in an undeveloped aquifer discharges to streams. Pumped wells may intercept this ground-water discharge and, if wells are located near streams or lakes, the lowered water level in the vicinity of the pumped well can cause water to flow from the surface-water body into the aquifer by induced infiltration.

Ground water stored in an aquifer moves towards a pumped well as the water level is drawn down. This withdrawal from storage usually is recharged each year by precipitation and spring snowmelt and is not a major source of water with respect to long-term average aquifer yield. During prolonged periods of no recharge such as during a drought, however, little ground water discharges from an aquifer and streamflow is at a minimum. Consequently, prolonged pumping would withdraw water from aquifer storage near a pumped well. If an aquifer is pumped heavily by many wells, large areas might be dewatered and the available ground-water storage could ultimately be depleted.

STREAM-AQUIFER RELATIONS IN STRATIFIED-DRIFT AQUIFERS

Aquifer yields that can be maintained year after year cannot depend on ground water in storage; it is only because ground water is a renewable resource that aquifers can support large withdrawals. The long-term yield depends on the recharge to the aquifer from precipitation and snowmelt, and on recharge from surface-water infiltration. Ground-water discharge from an aquifer to a stream is approximately equal to recharge to the aquifer from precipitation; therefore, all estimates of long-term aquifer yield were based on the quantity of water available from intercepted ground-water discharge and from induced infiltration of surface water.

The quantity of stream water that can be induced to infiltrate to an aquifer depends on the hydraulic properties of the streambed and the hydraulic conditions created by the pumped well. Detailed analyses at two areas where induced infiltration is occurring were made to estimate typical values of vertical hydraulic conductivity for small sandy-bottomed streams in the Northeast. In this report, the vertical hydraulic conductivity of the streambed is also referred to as the streambed conductivity.

Methods of Investigation

Several different methods were used to determine the vertical hydraulic conductivity of streambed material. Estimates were made from streamflow and head measurements near pumped wells, from permeameters installed in the streambed, from simulation of the hydrologic system with a ground-water-flow model, and from evidence of mixing based on water-quality data.

The quantity of water induced to flow from a stream through a streambed to an aquifer and thence to a pumped well is a function of the vertical hydraulic conductivity of the streambed material, the area of the streambed through which induced infiltration occurs, and the hydraulic gradient across the streambed. Vertical hydraulic conductivity can be calculated by using Darcy's equation (Freeze and Cherry, 1979, eq. 2.4, p. 16):

$$K_v = \frac{Q}{AI} \qquad (1)$$

where $I = (h_1 - h_2)/L$,

and where K_v is the vertical hydraulic conductivity of the streambed, in feet per day;

Q is the flow loss from the stream reach due to induced infiltration, in cubic feet per day;

A is the area of the streambed affected by induced infiltration, in square feet;

I is the hydraulic gradient;

h_1 is the altitude of the head in the stream, in feet;

h_2 is the higher of 1) the altitude of the head in the aquifer beneath the streambed, or 2) the altitude of the bottom of the streambed, in feet; and

L is the streambed thickness, in feet.

If streamflow measurements upstream and downstream from a pumped well indicate a loss in flow, the difference can be assumed to be induced infiltration. Most of this water will probably go to the pumped well although there are some situations where induced infiltration will enter the aquifer and then discharge again farther downstream (Newsom and Wilson, 1988). Head measurements in multilevel piezometers installed in the stream channel can be used to determine the vertical hydraulic gradient across the streambed. The area of the streambed subject to induced infiltration is that portion where the pumped well lowers the aquifer head below the stream stage as determined by comparison of stream and ground-water levels.

Site Selection

The hydrogeology of an area must be simple and well understood if Darcy's equation (1) is to give a reasonable streambed hydraulic conductivity. Ideally, there should be a single stream for which flow measurements can be made and the pumped well should deplete a measurable percentage of the streamflow. Areas with layers of fine-grained material in the stratified drift are not studied unless it is clear that the controlling hydraulic conductivity is that of the streambed, not the fine-grained-aquifer layer.

Induced infiltration sites that could be used as models for the Nashua River basin were identified using the above criteria. Of the sixteen water-supply wells in the Nashua River basin that are within 500 ft of a surface water source, only two, the Bemis Road well adjacent to Gulf Brook in Pepperell and the Patterson Road well adjacent to Morse Brook in Shirley, met the selection criteria. Base flow measurements made in Gulf Brook during spring and summer of 1985 were less than 3.5 ft^3/s, and the average pumping rate of the municipal well was about 1.4 ft^3/s (650 gal/min). Base flow measurements made in Morse Brook were less than 1.5 ft^3/s and the average pumping rate of the well was 0.5 ft^3/s (225 gal/min). Because the pumpage is a large percentage of the base flow in each brook, the potential effect of the pumped wells on streamflow was greater than the expected error of the flow measurements.

Pepperell Infiltration Site

The Bemis Road well is located about 160 ft from Gulf Brook which flows through the center of a stratified-drift-filled valley in Pepperell, Massachusetts (fig. 1). The well is located near the downstream end of the aquifer which is about 2-mi long and about 0.5-mi wide. Several sand pits in the vicinity of the well indicate that the aquifer is composed predominantly of medium sand. Well logs confirm this and indicate that the maximum thickness of the aquifer is about 60 ft. Sediment cores of the streambed indicated that there is about 0.75 ft of fine sand with organic material. Comparison of the streambed material with the well logs indicate that the vertical hydraulic conductivity of the streambed controls the quantity of water flowing from the stream to the well.

In 1983, a 5-day aquifer test was run by SEA Consultants Inc. (1984) as part of a water-resource-protection investigation for the town of Pepperell. Ground-water flow directions, determined from the map of water levels at the end of the aquifer test, were used to locate the reach of the stream through which induced infiltration was expected to occur.

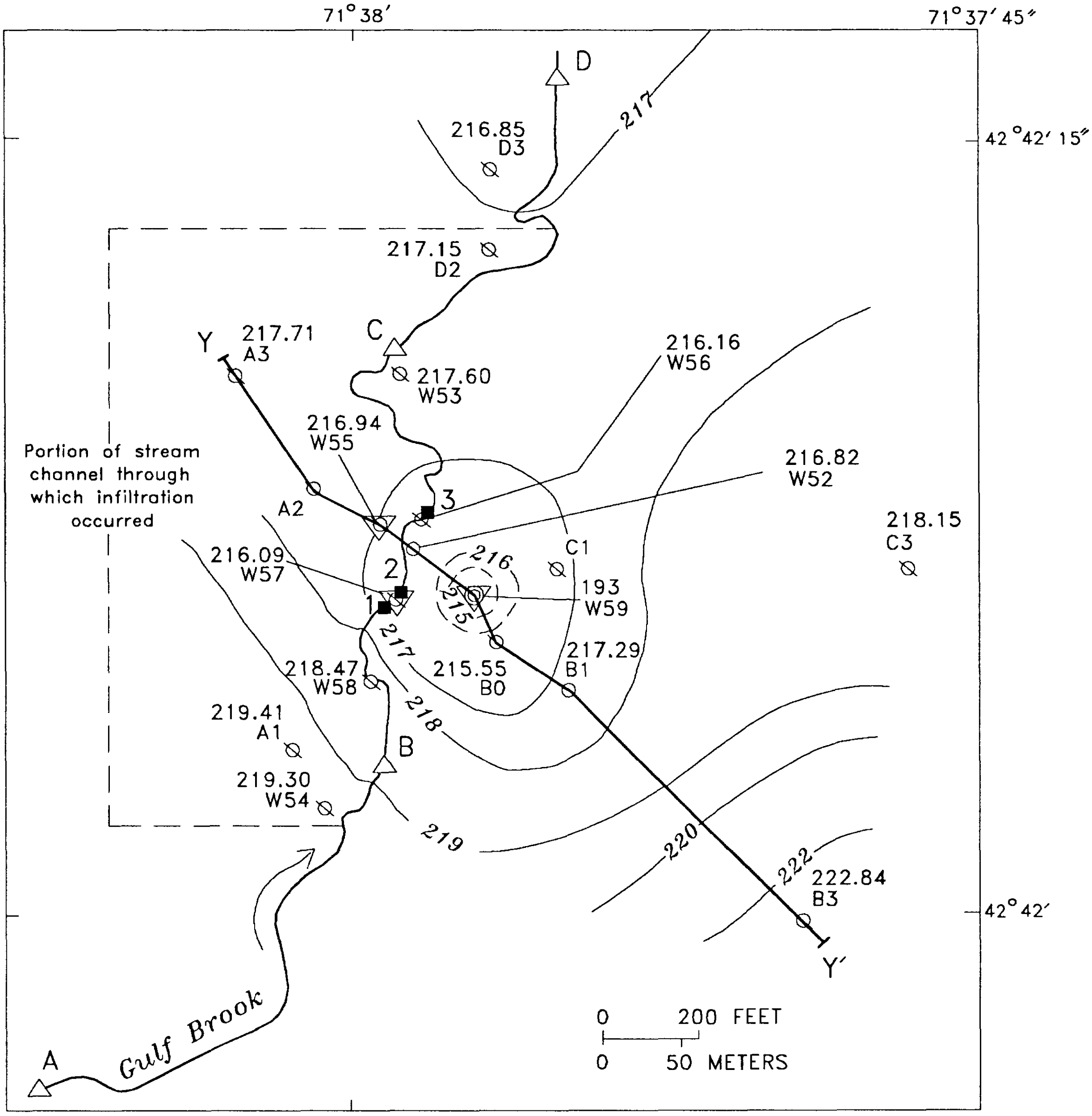
Infiltration Tests and Darcy Calculation

Infiltration tests were made on May 31 and November 27, 1985, in the vicinity of the Bemis Road pumped well. In each case, the pumping of the well was changed from an intermittent schedule to one where the same quantity of water was pumped per day, but at a lower, continuous rate. The well had been pumping at a constant rate for at least 7 days before each test. During each test, water-level measurements were made in 23 observation wells and piezometers in the area to define the vertical and horizontal head distribution in the aquifer (figs. 2 and 3). Three flow measurements were made at each of four measurement sites along the stream. Two of these sites (A and D) were chosen far outside the infiltrating reach. Sites B and C were measured to define more accurately the area through which most of the infiltration occurs.

Water levels in piezometers located between the pumped well and the stream indicate that head in the aquifer near the stream is below the level of the streambed when the well is pumped, causing water to flow from the stream towards the well (fig. 3). Twenty-four hours after pumping ceased, however, the heads at different depths were equal, indicating induced infiltration had stopped.

In May, flow measurements on Gulf Brook showed a loss of 0.49 ft^3/s between measuring sites A and D. The public-supply well had been pumping at 1.39 ft^3/s (625 gal/min) for 7 days; therefore, 35 percent of the water pumped from the well may have originated in Gulf Brook. Only in three places was it possible to measure the head gradient across the streambed layer itself. However, using head measurements made in the stream, in shallow piezometers driven into the streambed, and in shallow wells on the stream bank, and using a streambed thickness of 0.75 ft, estimates of hydraulic gradients along the stream averaged 1.35 ft/ft (feet per foot). The area of the streambed through which induced infiltration occurred was approximately 11,600 ft^2. The vertical hydraulic conductivity of the streambed, estimated using the Darcy equation (1), was 2.7 ft/d (feet per day).

In November, after the well pumped at 1.1 ft^3/s (500 gal/min) for 8 days, the flow loss was 0.70 ft^3/s; therefore, 64 percent of the well water may have derived from the stream. The flow loss was greater in November than in May because the stream stage was higher and some low areas were flooded increasing the



EXPLANATION

219 — WATER TABLE CONTOUR--shows altitude of water table, May 31, 1985. Dashed where approximately located. Contour interval is variable. Datum is sea level.

Y Y′ — SECTION LINE

W59 — MUNICIPAL-SUPPLY WELL (PUMPED WELL) AND IDENTIFIER

219.41 A1 — OBSERVATION WELL OR GROUP OF PIEZOMETERS AND IDENTIFIER. NUMBER IS MEASURED ALTITUDE OF WATER TABLE OR HEAD, MAY 31, 1985. (data not available for well A2).

A — STREAMFLOW MEASUREMENT SITE AND IDENTIFYING LETTER

1 — INFILTROMETER AND IDENTIFYING NUMBER

WATER-QUALITY SAMPLING SITE

DIRECTION OF WATER FLOW

Figure 2.--Water table on May 31, 1985, after pumping 7 days, and location of observation wells and streamflow measurements at the Pepperell infiltration site.

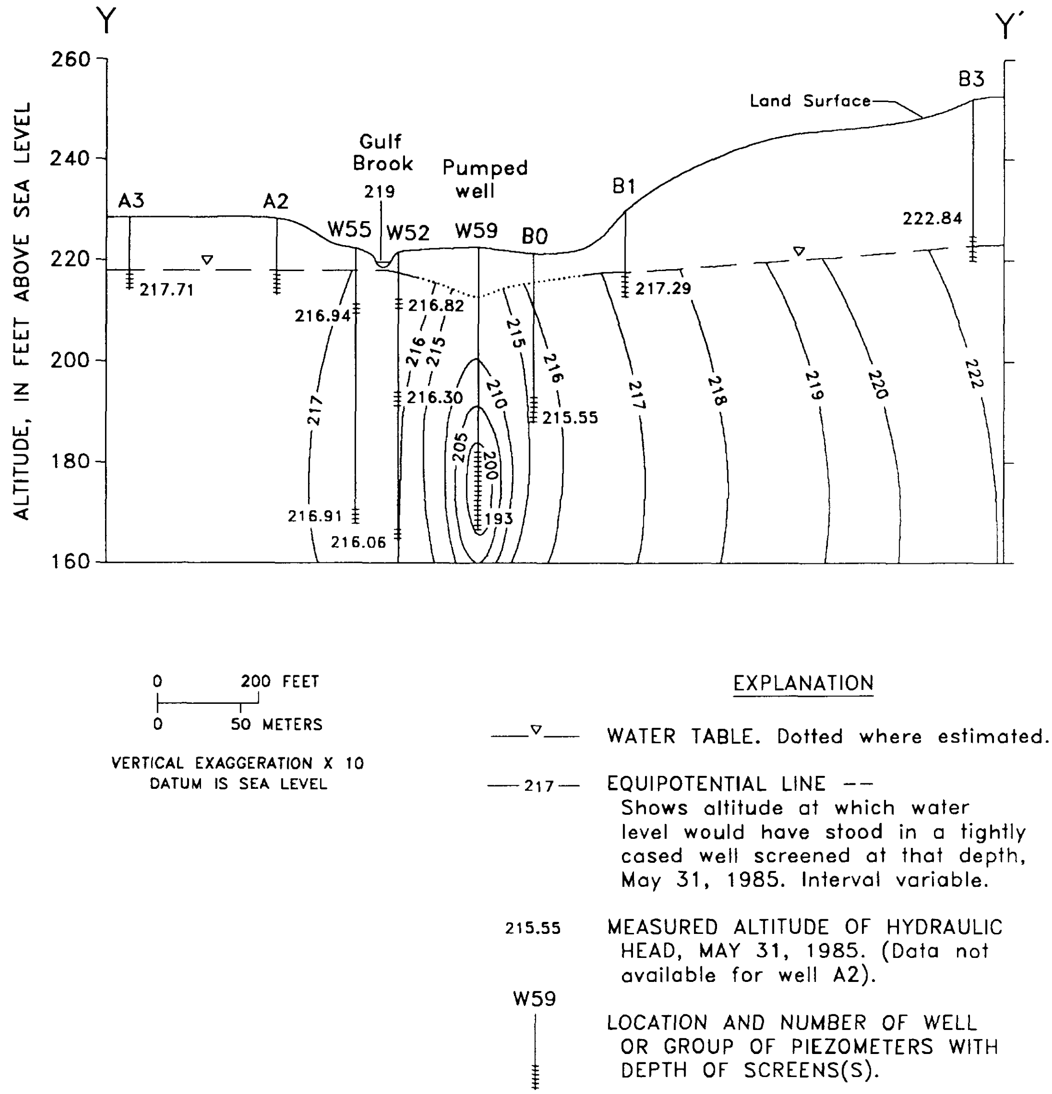


Figure 3.--Hydraulic head on May 31, 1985, after pumping 7 days, and location of well and multilevel piezometer screens at the Pepperell infiltration site (see trace Y-Y′ on figure 2).

streambed area to approximately 48,500 ft^2. The average hydraulic gradient was estimated at 0.59 ft/ft and vertical hydraulic conductivity of the streambed was estimated at 2.1 ft/d. This slight difference in values (2.1 versus 2.7 ft/d) may be attributed to an increase in viscosity of the water in November compared to May, the errors inherent in flow measurements and assumptions, or a combination of both.

Infiltrometer tests

Determination of streambed conductivity from flow and head measurements is difficult and time consuming, therefore an easier method was tried during the November test. Three permeameters, called "infiltrometers", were driven through the Gulf Brook streambed to make direct measurements of the vertical hydraulic conductivity. The infiltrometer equipment, designed for this study, had a 1.25-in. (inch)-diameter piezometer welded inside a 5-ft length of 8-in.-diameter pipe. This equipment was driven into the center of the streambed filling the 8-in.-diameter pipe with sediment. The piezometer had a 2-in.-long screen which measures the aquifer head at the bottom of the infiltrometer (fig. 4). Data were collected using the equipment for both constant-head and falling-head permeameter tests. A constant head, equal to that in the stream, was maintained in the infiltrometer using a siphon attached to a collapsible reservoir filled with a measured volume of water (fig. 4A). When pumping of the aquifer began, the volume entering the system during a given time was equal to the volume lost from the reservoir. When the siphon was disconnected, the equipment became a falling-head permeameter (fig. 4B) and the head change was noted over time. Hydraulic conductivity was calculated using Darcy's equation (1) for the constant-head data. Hydraulic conductivity for the falling head data was calculated using Todd's laboratory permeameter formula (1980, eq 3.20, p. 74):

$$K_v = \frac{r_t^2}{r_c^2} \frac{L}{t} \ln \frac{h_1}{h_2} \qquad (2)$$

where
- K_v is the vertical hydraulic conductivity, in feet per day;
- r_t is the radius of the measuring tube, in feet;
- r_c is the radius of the sediment cylinder, in feet;[1]
- L is the length of the sediment column, in feet;
- t is time of test, in days;
- h_1 is the head at the beginning of the test in feet;[2] and
- h_2 is the head at the end of the test, in feet.[2]

[1] $r_t = r_c$ in this apparatus.
[2] both h_1 and h_2 are referenced to the bottom of the infiltrometer.

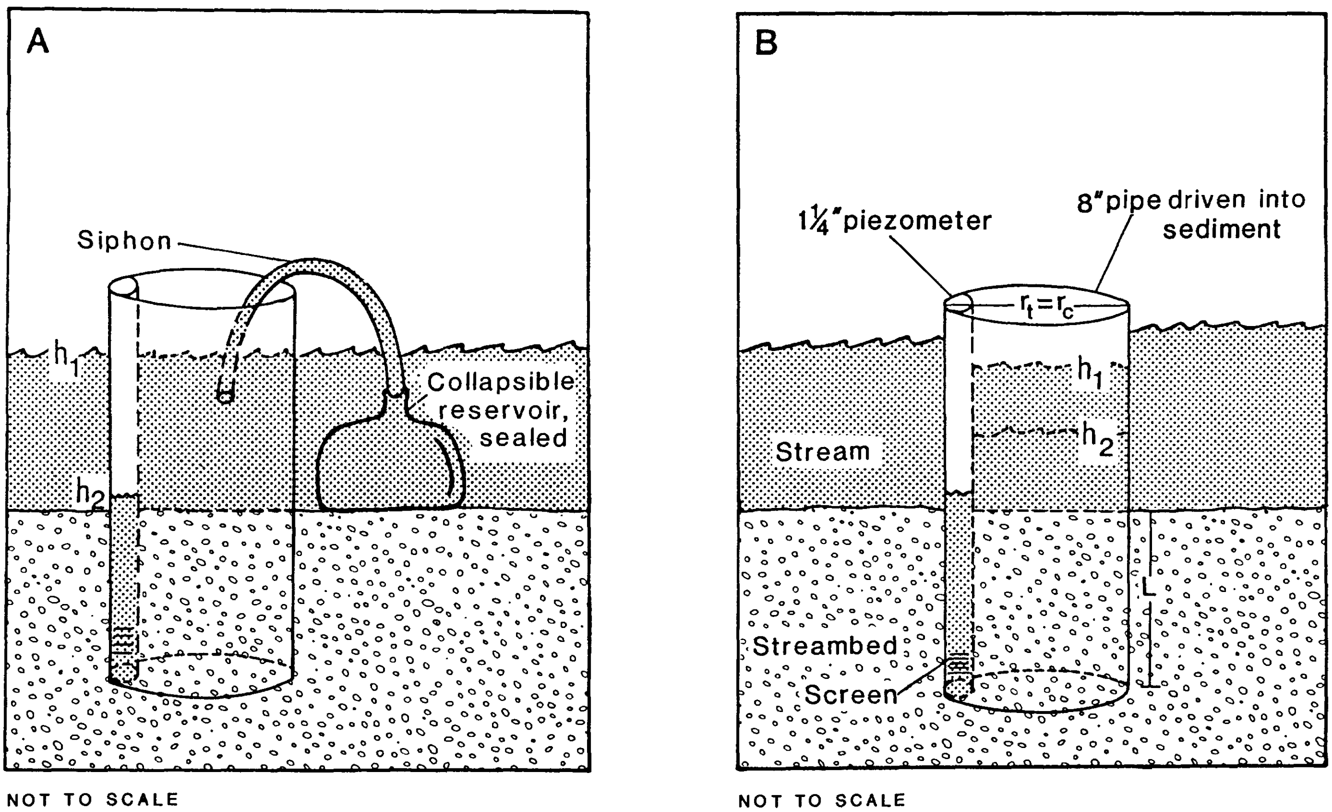


Figure 4.--Infiltrometer equipment; A: constant-head set-up, B: falling-head set-up.

Infiltrometer data were used to determine the following values of vertical hydraulic conductivity.

Vertical hydraulic conductivity, in feet/day		
	Constant head test	Falling head test
Infiltrometer #1	0.26	0.17
Infiltrometer #2	.06	.09
Infiltrometer #3	.90	.19

The vertical hydraulic conductivities determined using the infiltrometers are one-tenth of those calculated from streamflow and head measurements of the entire stream reach. Because values of streambed hydraulic conductivity determined in other studies in the Northeast range from about 2 to 10 ft/d (Joel Dysart, U.S. Geological Survey, oral commun., 1986) and confirm the results of the flow measurement approach, it was assumed that the values of vertical hydraulic conductivity determined from the infiltrometer study are too low. Two factors known to contribute to these low values of conductivity are the accidental introduction of silty wash water into the infiltrometers, which may have created a layer of fine-grained sediments, and the compaction of the streambed material during installation of the equipment. Low values of streambed conductivity would also result if most of the induced infiltration occurs through the coarse material at the sides of the stream rather than through the finer-grained stream bottom, or if currents scour out the fine bed material in places and create "windows" through which most of the infiltration occurs.

Vertical hydraulic conductivity of streambed material would be much easier to measure with infiltrometers than with streamflow and head measurements; therefore, the technique deserves further study. However, until the infiltrometer technique is refined and the results are better understood, the hydraulic conductivities derived from flow measurements and head data will be used.

Shirley Infiltration Site

The Patterson Road well is about 50 ft from Morse Brook in Shirley, Massachusetts (fig. 1). Morse Brook is one of two streams that cross a large sand plain before discharging to the Nashua River. The Town of Shirley plans to develop another municipal well about 1,000 ft north of the Patterson Road well near Walker Brook. Near both the existing and planned wells, the aquifer material is predominantly fine to coarse sand. The saturated thickness of the aquifer is 45 ft near the Patterson Road well and about 30 ft near Walker Brook.

Infiltration Tests and Darcy Calculation

Infiltration tests were made on January 16, and July 12, 1985, in the vicinity of the Patterson Road well. The well had been pumped at a constant rate for at least 5 days before each test. Pumping rates were 0.58 ft^3/s (260 gal/min) in January and 0.47 ft^3/s (210 gal/min) in July.

Water-levels were measured in a three-dimensional array of 17 observation wells and piezometers throughout the area, and flow measurements were made at 4 sites along Morse Brook (figs. 5 and 6). Sites A and D were at the edge of the area affected by the pumped well; sites B and C, located about 150 ft from the well, were within the area. The heads in the piezometers between the well and the brook indicated that water was moving toward the pumped well (fig. 6A). Within an hour after pumping ceased, however, the gradient had reversed and ground water was discharging from the aquifer to the brook. Figure 6B is the cross section showing heads 24 hours after pumping ceased.

The field data from the January and July tests indicated similar streamflow losses (0.03 ft^3/s) and similar head gradients across the streambed (0.1 ft/ft). In January the area of the streambed was approximately 7,800 ft^2 and in July it was 5,200 ft^2. Thus, the vertical hydraulic conductivity of the streambed calculated using Darcy's equation (1) was 3.3 ft/d in January and 5.0 ft/d in July. The losses measured in Morse Brook ranged from 5 to 9 percent of the total flow, close to the expected error of measurement, and must be considered estimates. Therefore, the streambed conductivities based only on these flow measurements and gradients along the stream need to be confirmed using another approach.

Digital Model Simulation

An evaluation of the hydrologic system in the vicinity of the Patterson Road well was made using a ground-

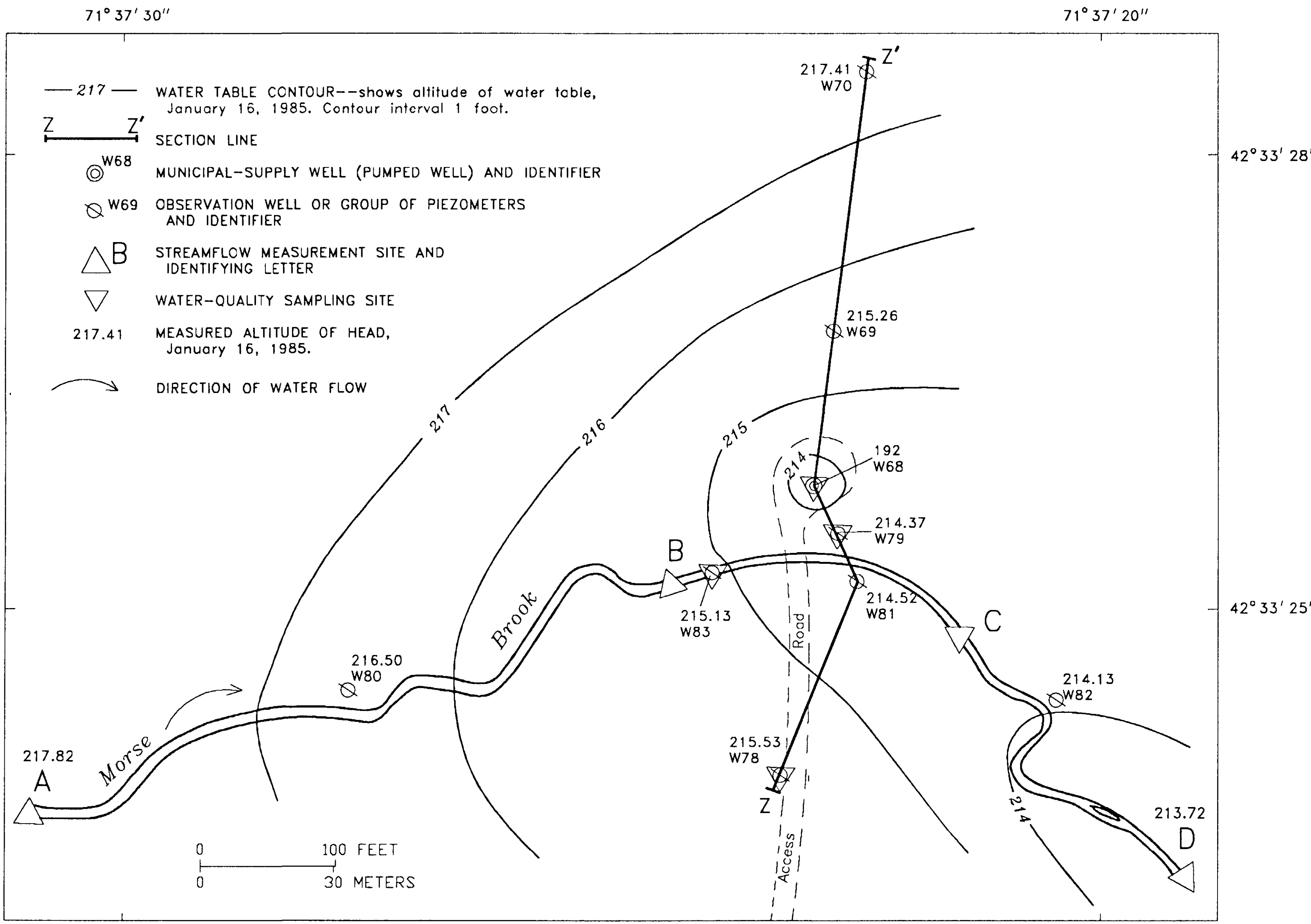


Figure 5.--Water table on January 16, 1985, after pumping 5 days, and location of observation wells and streamflow measurements at the Shirley infiltration site.

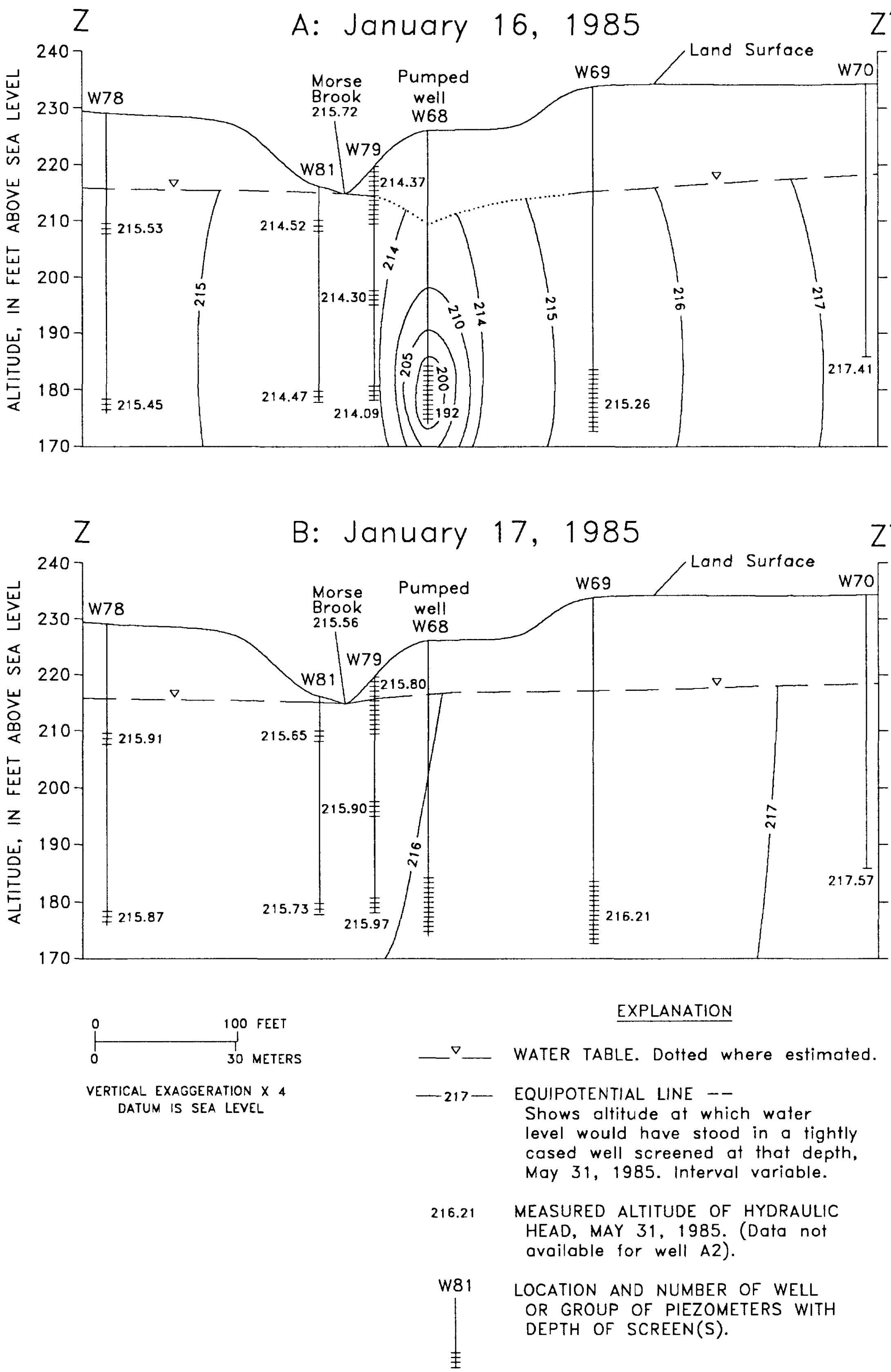


Figure 6.--Hydraulic head and location of well and multilevel piezometer screens at the Shirley infiltration site; A: on January 16, 1985, after pumping 5 days, B: on January 17, 1985, 1 day after pumping ceased (see trace Z-Z′ on figure 5).

water-flow model (McDonald and Harbaugh, 1988). A modeling approach allowed the inclusion of many other hydrogeologic data from the site (Coffin & Richardson, 1976) and was used to give greater confidence in the streambed conductivity estimates.

Head and bedrock elevations from 17 observation wells and piezometers and from a seismic-refraction survey in the vicinity of the pumped well were used as input to the model. In addition, previous aquifer test results and sediment samples were used to estimate aquifer conductivity and specific yield. The hydraulic conductivity of the aquifer was set at 100 ft/d in the medium sands at the northern edge of the model area and at 200 ft/d in the coarser sands at the center. These values were based on aquifer tests run for the town which indicate a transmissivity of 11,000 ft^2/d near the Patterson Road well (Randy Fouch, Coffin & Richardson, oral commun., 1984). The aquifer was modeled as a single-layer sand plain drained by a stream. A constant flux-boundary simulated leakage into and out of the model area from the surrounding sand plain. The average rate of flux into the model from areas upgradient was 0.33 ft^3/s per 1000 ft of model boundary for a total inflow of 0.85 ft^3/s. The flux out of the model to areas downgradient was 0.24 ft^3/s per 1000 ft of model boundary for a total outflow across the boundaries of 0.35 ft^3/s. These rates were determined from earlier model runs with constant head cells around the steady-state model. Recharge and evapotranspiration were assumed to be zero in January; in July, no recharge was simulated but ground-water evapotranspiration was estimated to be 5 in./yr (inches per year). The model was run first as a steady-state simulation adjusting the conductivity of the 1-ft-thick streambed until the heads and streamflow matched the prepumping field conditions. The model was then run as a transient simulation, using a storage coefficient of 0.3, for the 5-day period of the January test and verified against the 7-day July test. When the streambed conductivity was 2 ft/d, simulated drawdowns best matched measured drawdowns in both January and July (fig. 7). The simulated streamflow loss in July was 0.03 ft^3/s which matched the measured streamflow loss exactly.

For comparison, the model was rerun using a streambed conductivity of 0.2 ft/d, the value determined from the infiltrometer tests. The simulated drawdowns (fig. 8) did not match the field data as well as when the conductivity was 2 ft/d. More importantly, the simulated flow loss of 0.002 ft^3/s in July was lower than the measured value by an order of magnitude.

The values of vertical hydraulic conductivity calculated from field data at the two infiltration sites ranged from 2.1 to 5.0 ft/d. The ground-water-flow model of the Patterson Road site indicated a vertical hydraulic conductivity of 2.0 ft/d, supporting the direct calculations. This value of streambed hydraulic conductivity is probably typical for small sandy streams in the Northeast. Therefore the aquifer yields, presented later in this report, include estimates of water available from induced infiltration based on the conservative value of 2 ft/d for vertical hydraulic conductivity when the streambed is 1-ft thick.

Evidence of Induced Infiltration Based on Water Quality

During the infiltration tests, the water quality of Gulf and Morse Brooks and of the adjacent aquifers was sampled to determine if water quality could be used to estimate the degree of connection between the stream and the aquifer and to assess the impact of stream-water quality on the well-water quality. Selected sampling sites included:

1. the stream, at the well point closest to the pumped well,
2. a well point beneath the streambed,
3. multilevel piezometers between the stream and the pumped well,
4. the pumped well, and
5. the ambient ground-water quality as represented by a deep observation well unaffected by induced infiltration.

The water samples were analyzed for common constituents, physical properties such as temperature and dissolved oxygen, and nutrients. The chemical and physical constituents were used to trace mixing of the surface and ground waters under infiltrating conditions.

In areas of induced infiltration, the water pumped by the well might be a mixture of water from the stream and water from the aquifer. The infiltration tests were scheduled during low-flow periods so that water loss in the stream due to induced infiltration should have been a significant and measurable percentage of total streamflow. However, because the water in

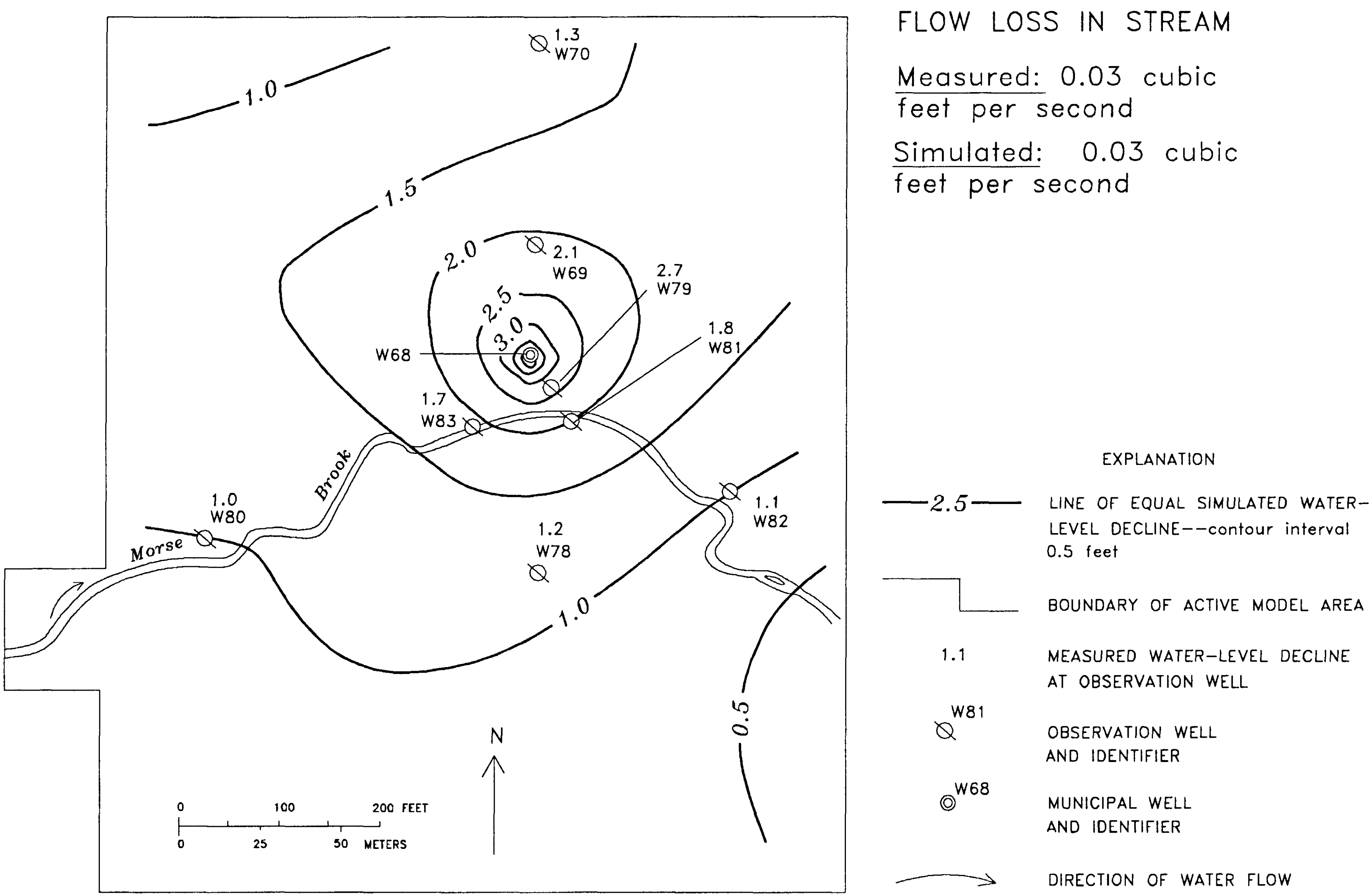


Figure 7.--Comparison of measured drawdowns and simulated drawdowns, based on a streambed conductivity of 2 feet per day, during July infiltration test at the Shirley infiltration site.

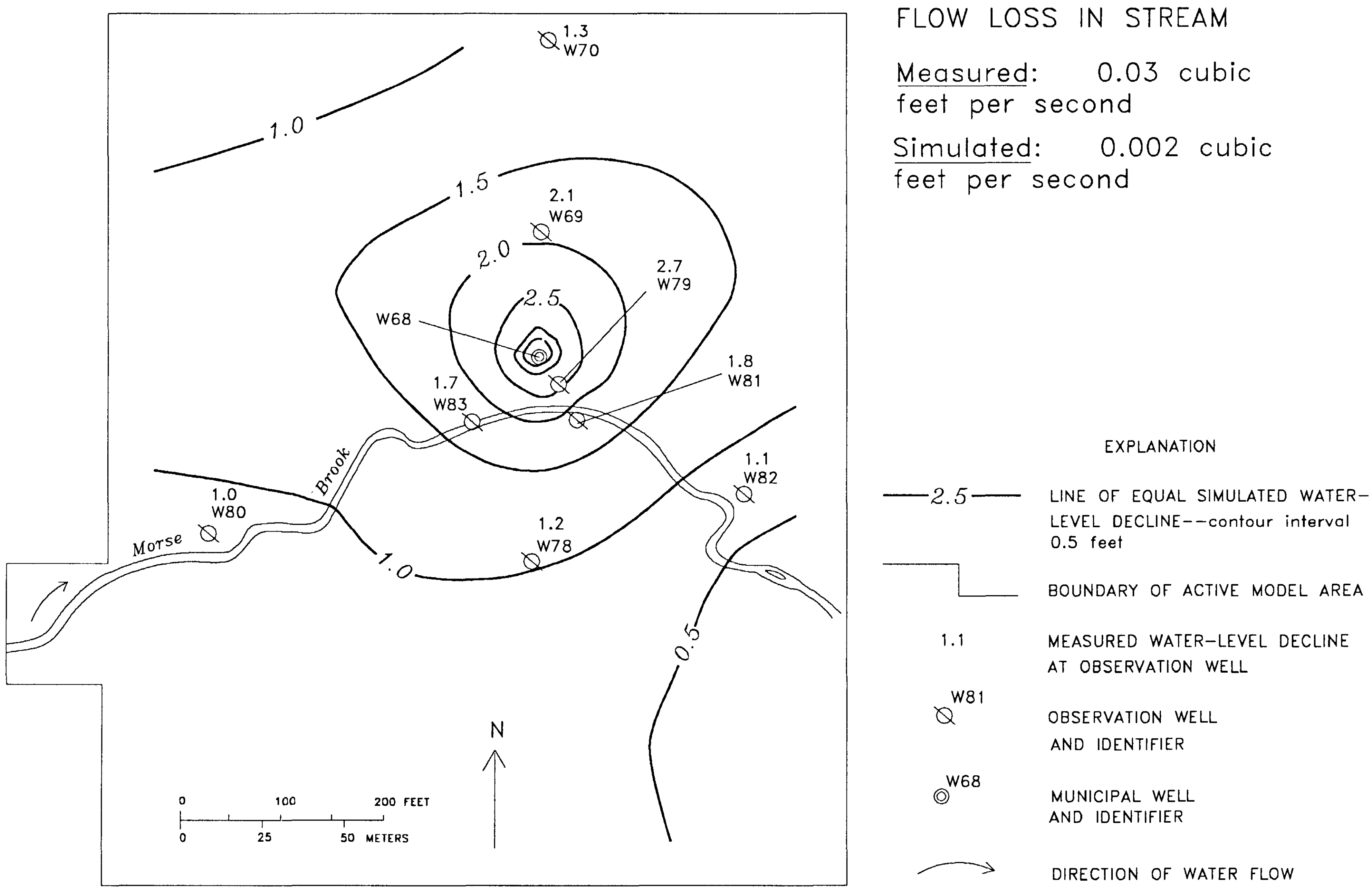


Figure 8.--Comparison of measured drawdowns and simulated drawdowns, based on streambed conductivity of 0.2 feet per day, during July infiltration test at the Shirley infiltration site.

streams during low flow is predominantly ground-water discharge, chemical evidence of mixing is difficult to demonstrate.

Pepperell Infiltration Site

In November, at the Pepperell site, the quality of the stream and aquifer water differed significantly in nine of the 19 water-quality constituents sampled on November 27, 1985 (table 1). In seven of these nine, the pumped-well value fell between the stream and the aquifer values, indicating mixing from induced infiltration. Of these seven constituents, chloride is a conservative element and might show mixing most accurately. The percentage of water that may have originated in the stream was calculated using the following mass balance equation.

$$C_W V_W = C_S V_S + C_A V_A \qquad (3)$$

where C is the concentration, in milligrams per liter;
V is the volume, in liters;
$_W$ indicates the pumped well;
$_S$ indicates the stream; and
$_A$ indicates the aquifer.

Using chloride as an example, $C_W = 4.8$ mg/L, $C_S = 5.9$ mg/L, and $C_A = 2.7$ mg/L (table 1). If the total volume is 100 L of well water and x represents the volume coming from the stream, substitution into eq. (3) gives x = 66 L. Therefore, 66 percent of the water pumped from the well may have originated in the stream. Similar calculations for silica (60 percent) and calcium (71 percent), which are both relatively conservative in the geochemical environment of the aquifer, verify that approximately two thirds of the water pumped by the well may have come from the stream. This supports the results of the Darcy calculation which indicated that 64 percent of the water pumped by the well may have originated in the stream.

Piper diagrams classifying the water quality were drawn for three samples (fig. 9). The values for the pumped well (numbers 2 and 5) fall between the stream and aquifer values. This illustrates the mixing of stream and aquifer waters demonstrated by the mass balance equations of the conservative elements.

Analysis of mixing based on instantaneous sampling of both stream and pumped well water is useful but not completely representative of the mixing process. This is because the water pumped from the well is made up of ground water that has been in the aquifer for different lengths of time and that had entered the aquifer with varying physical and chemical properties. If the water quality of the stream is highly variable, as it would be during quickly changing weather conditions in the spring, instantaneous water-quality data might not give accurate evidence of mixing. In May, of the fourteen constituents which differed significantly between the surface and ground water, only five showed mixing. However, the dissolved oxygen and temperature data confirm that induced infiltration and mixing were occurring.

Although the results of the water-quality study in Pepperell can not be used as the primary method of estimating the surface-water contribution to the pumped well, the data confirm that the water quality in the stream has an effect on the quality of the water pumped from the well. Stream-water quality is a factor to consider when locating or assessing water-supply systems in areas where some of the water pumped by the well comes from the stream.

Shirley Infiltration Site

As in Pepperell, the stream and the ground water in Shirley were similar in composition and, because the flow loss in Morse Brook was less than 10 percent of the quantity pumped by the well, it was unlikely that the water-quality analysis would indicate any mixing caused by induced infiltration. The physical properties of the water from the well, however, showed possible mixing from induced infiltration. In the July test, the aquifer temperature was 8°C, the stream temperature 13°C; the pumped well temperature was between these at 10°C. The dissolved-oxygen values decreased between Morse Brook and below the streambed. A rise in dissolved iron (from 9 to 300 µg/L (micrograms per liter) and manganese (from 5 to 9 µg/L) at these sampling sites was due to the increased solubility of iron and manganese in a reducing environment; this condition may occur in an area of induced infiltration.

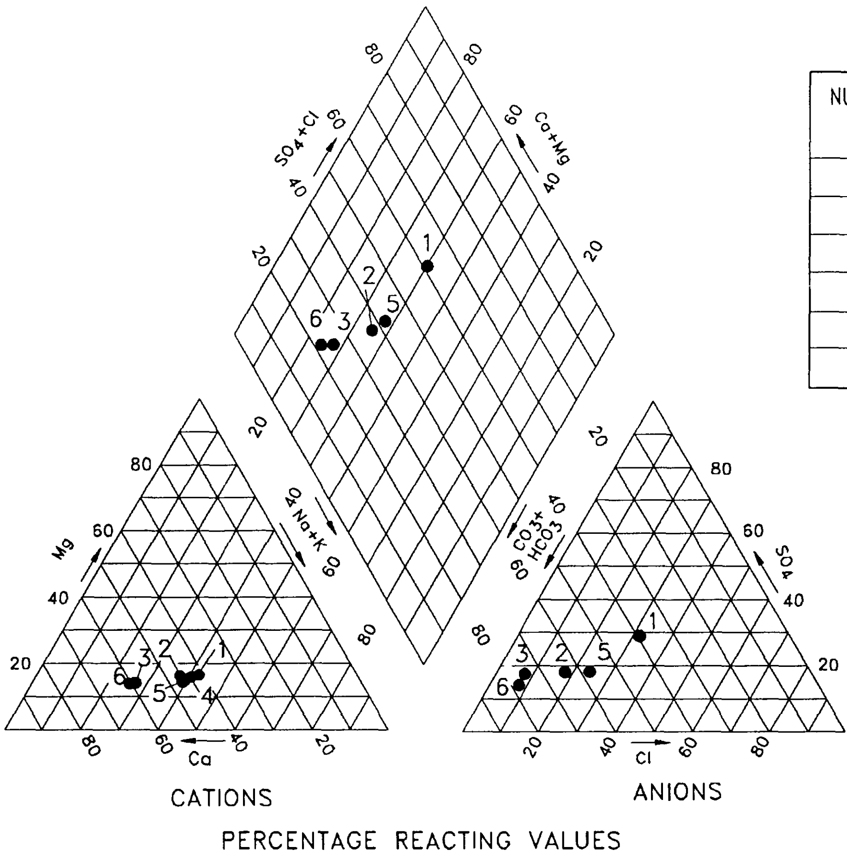
Maximizing Induced Infiltration

Surface water would enter an aquifer at the maximum rate if the head in the aquifer were lowered beneath the streambed along the entire stream channel. Under this hydraulic condition, the quantity of in-

Table 1.--*Water quality of the stream, pumped well, and aquifer in Pepperell, November 27, 1985.*

	Gulf Brook	Pumped well[1] (W59)	Aquifer[1] (W55)
Water-quality constituents in which stream sample and aquifer sample differ significantly (more than the precision of the analytical method)			
Alkalinity, field (milligrams per liter as $CaCO_3$)	5	11	18
Carbon, total organic (milligrams per liter as C)	6.4	1.5	.4
Aluminum, dissolved (micrograms per liter as Al)	80	30	20
Calcium, dissolved (milligrams per liter as Ca)	4.1	5.0	7.2
Silica, dissolved (milligrams per liter as SiO_2)	8.0	10	13
Iron, dissolved (micrograms per liter as Fe)	150	78	6
Chloride, dissolved (milligrams per liter as Cl)	5.9	4.8	2.7
Manganese, dissolved (micrograms per liter as Mn)	7	17	3
Oxygen, dissolved (milligrams per liter as O_2)	11.7	3.9	9.4
Water-quality constituents in which the stream and aquifer samples do not differ significantly			
Temperature (degrees celsius)	7.5	9.0	7.0
pH (standard units)	6.7	6.2	6.7
Specific conductance (microsiemens per centimeter)	41	47	51
Solids, dissolved (residue at 180 degrees celsius)	39	39	41
Magnesium, dissolved (milligrams per liter as Mg)	1.0	1.0	1.0
Sodium, dissolved (millgrams per liter as Na)	4.1	4.0	3.0
Potassium, dissolved (milligrams per liter as K)	.9	.9	1.4
Sulfate, dissolved (milligrams per liter as SO_4)	7.1	6.4	7.3
Fluoride, dissolved (milligrams per liter as F)	.06	.06	.08
Bromide, dissolved (milligrams per liter as Br)	.02	.02	.01

[1]See figure 2.



NUMBER	DATE	AQUIFER/STREAM
1	11/20/85	GULF BROOK
2	11/20/85	PUMPED WELL
3	11/20/85	AQUIFER
4*	5/31/85	GULF BROOK
5	5/31/85	PUMPED WELL
6	5/31/85	AQUIFER
*anion analysis not available		

Figure 9.--Piper diagram showing mixture of aquifer and stream waters in the Bemis Road pumped well November 27, 1985, at the Pepperell infiltration site.

duced infiltration would be a function of the vertical hydraulic conductivity and thickness of the streambed, the head in the stream, and the area of the streambed. Theoretically, this hydraulic condition could be achieved with many pumped wells located in a line parallel to the stream. Ground-water-flow models (McDonald and Harbaugh, 1988) of hypothetical stream-aquifer systems representing a variety of conditions typical of the region were used to estimate the approximate spacing of pumped wells necessary to create the hydraulic conditions required to induce infiltration at the maximum rate.

Ground-water-flow models representing a 1-mi wide and 4-mi long section in the middle of thirty hypothetical aquifers were run using all combinations of aquifer hydraulic conductivity (100, 200, or 500 ft/d), initial saturated thickness (50 or 80 ft), and distance between the pumped well and the stream (50, 100, 200, 500, or 1000 ft). In each simulation, all sides of the aquifer were simulated with no flow boundaries (the ends of the model area were far enough from the pumped well not to affect the drawdown). The stream was simulated as 20-ft wide and 1-ft deep, and the streambed was simulated as 1-ft thick with a vertical hydraulic conductivity of 2 ft/d. Ground-water levels everywhere were initially set equal to stream level. Thus, when the aquifer drawdown beneath the stream was 2 ft (creating a hydraulic gradient of 2 ft/ft), the induced infiltration rate would be at a maximum, 4 $ft^3/d/ft^2$ (cubic feet per day per square foot) of stream channel. To create a conservative analysis in the 180-day transient simulations, the storage coefficient was set to 0.3, the upper end of the range expected for most unconfined aquifers (Freeze and Cherry, 1979, p. 61). If the storage coefficient was 0.2, wells could be spaced further apart.

Because most of the small aquifers in Massachusetts are thin, withdrawal from public-supply wells is limited by aquifer saturated thickness and available drawdown in the well. Thus, simulated drawdown at the pumped well in each hypothetical aquifer was also limited--to 30 ft when the saturated thickness was 50 ft, and to 50 ft when the saturated thickness was 80 ft. Drawdown was controlled by representing the pumped well as a constant head node 3 ft on a side. The average rate of discharge from the constant head node is equivalent to the discharge from a well that would create the specified drawdown after 180 days. Only one pumped well was included in each simulation, and the results used to estimate the spacing between wells needed to cause induced infiltration at the maximum rate along the entire channel of an aquifer extending beyond the limits of the model. The well spacing was estimated by noting the distance to where the drawdown beneath the stream, after 180 days, was one foot. A similar well, equidistant from this point also would cause a 1-ft drawdown. Thus the two wells would create the hydraulic conditions necessary for maximum induced infiltration along the channel between them. The estimate of the well spacing would be double the distance to where the drawdown was one foot. This analysis assumes that the water-level declines resulting from different pumping wells are independent and that the effects of each well operating alone can be added together to give the net effect of all wells operating simultaneously. Because the aquifer is unconfined, the sum of the effects of pumping individual wells may not equal the effect of pumping multiple wells. Nonetheless, it is believed that the errors introduced by these assumptions will not alter the qualitative character of the results.

Results of the hypothetical models suggest that the quantity of water pumped from a well with a specified drawdown has a linear relation with well spacing required to cause maximum induced infiltration. Figure 10, based on data from all 30 hypothetical models, is a plot of the well spacing required to cause maximum induced infiltration at different withdrawal rates. If an existing well yields 2 Mgal/d (million gallons per day), the maximum rate of induced infiltration would be maintained if other wells were located at 2,250-ft intervals along the stream. The well spacing ranged from 1,000 ft in an aquifer with each well yielding 1 Mgal/d to 8,000 ft in an aquifer with each well yielding 7 Mgal/d. Because of well

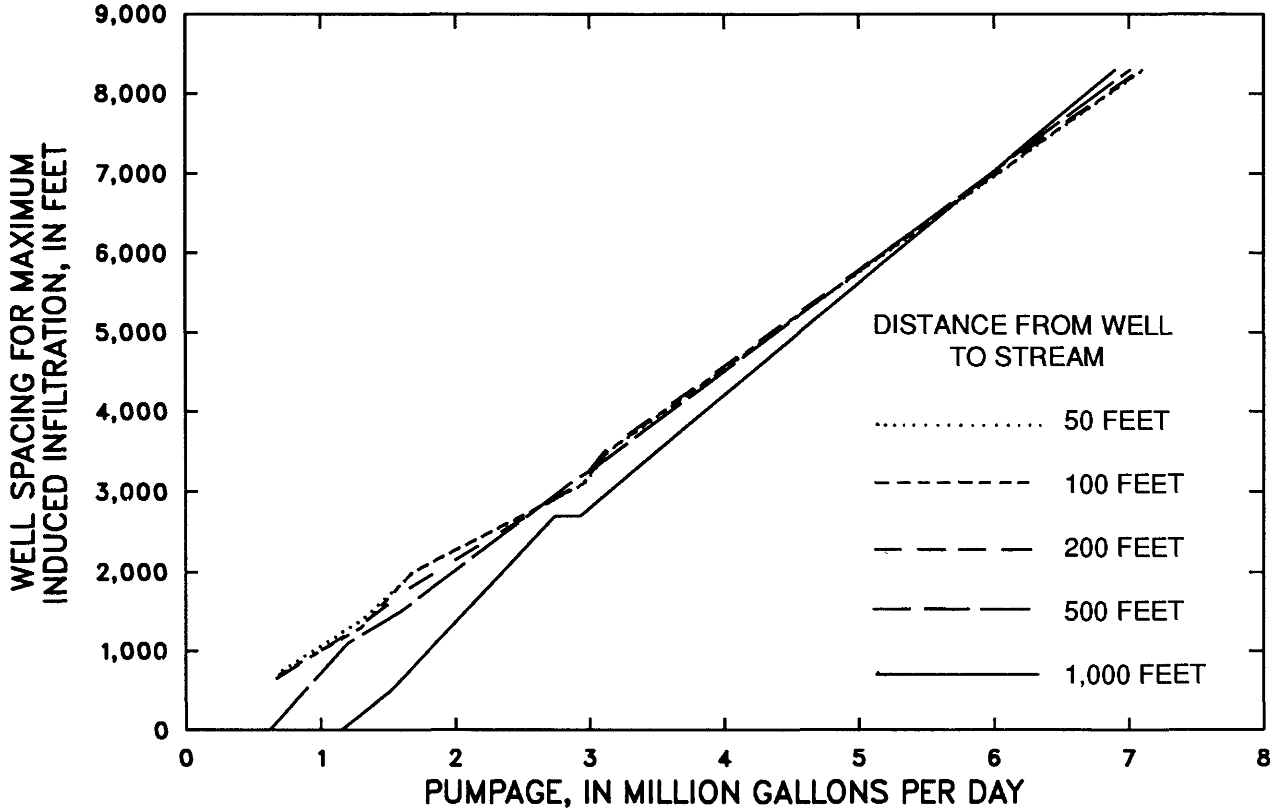


Figure 10.--Well spacing needed for maximum induced infiltration in hypothetical aquifers typical of the study area.

interference, these wells would no longer pump their original quantity. The total contribution to the well from induced infiltration is limited to the 4 $ft^3/d/ft^2$ of stream channel. Also, yield would be less at the ends of the aquifer. Also evident from the graph is that distances from the pumped well to the stream of 500 ft or less have similar effects on the drawdown under the stream. Commonly the quantity of water derived from infiltration is controlled by available streamflow rather than by the properties of the streambed; therefore yield of a pumped well would be less than those given in figure 10; the graph should only be used if there is ample streamflow.

YIELD OF STRATIFIED-DRIFT AQUIFERS

Long-term aquifer yields based on intercepted ground-water discharge and induced infiltration of surface water were estimated for six aquifers in the Nashua River basin. These aquifers, each named for the stream it underlies, are the upper and lower Catacoonamug Brook, Pearl Hill-Willard Brooks, Still River, Stillwater River, Wekepeke Brook, and Witch Brook aquifers (see fig. 1).

Theoretically, a pattern of pumped wells or well fields could be designed to intercept all ground-water discharge. Wells or well fields located along streams could be pumped to induce infiltration of streams that flow across an aquifer. Theoretically, if all ground-water discharge could be intercepted and if all the surface-water inflows could be induced to infiltrate, the aquifer's potential yield would equal the natural streamflow. Therefore, in this report, maximum long-term yield is calculated as the sum of the current withdrawals and the streamflow, which is a measure of water potentially available from interception of ground-water discharge and induced infiltration.

Streamflow and, therefore, aquifer yield varies over time. In hydrology, streamflow is often presented as a flow duration, which describes a given flow in the stream as the percentage of time that flow is equaled or exceeded. For example, the 90-percent flow duration is the streamflow that is equaled or exceeded 90 percent of the time. A flow of 99-percent duration is a smaller flow than a 90-percent duration flow. In this report the term "yield duration" is analogous. A 90-percent yield duration is the quantity of water that can be withdrawn from an aquifer 90 percent of the time; 10 percent of the time this yield would not be available.

During prolonged periods of no recharge such as during a drought, little ground water discharges from an aquifer and streamflow becomes minimal. Therefore, most water pumped during these conditions is withdrawn from aquifer storage. To compare the quantities of ground water in storage in different aquifers, estimates were made of the average rate at which water can be withdrawn during 180 days of no recharge while dewatering the aquifer by no more than 50 percent. If withdrawal from storage is not replenished, water levels will decline and aquifer storage will be depleted. Therefore, these estimates of ground water available from storage in aquifers are a measure of the source's short-term capacity to deliver water in excess of the long-term yield or during isolated periods of no recharge. Other factors limit the short-term yield that can actually be withdrawn from aquifer storage. The cost of the many wells or well fields needed to achieve 50 percent dewatering would be prohibitive. Also, the drastically lowered water table would cause environmental impacts such as streamflow depletion and dewatering of wetlands.

The estimates of water available from storage provide a means for comparing the relative potential short-term yield of aquifers under extreme climatic and pumping conditions. However, because these large withdrawals from aquifer storage cannot be maintained over time and have large economical and environmental costs, estimates of long-term aquifer yield are a better assessment of the quantity of water available from the aquifer. All the yield estimates represent long-term averages and cannot be used to predict how much ground water will be available in a given year.

Methods of Estimating Long-Term Yield

Long-term yield of the six aquifers was estimated using one of two methods. In areas where a major stream drains the most transmissive part of the aquifer, ground-water discharge and surface inflow available for infiltration were measured. In these areas, yield estimates were based on the flow measurements and the flow-duration curves developed for the stream. In areas where the most transmissive part of the aquifer is not drained by the main stream, ground-water discharge and surface inflow could not be measured directly. Therefore, long-term yield was estimated using curves that relate baseflow to the percentage of the basin covered by stratified drift.

Determining Long-Term Aquifer Yield from Stream-Discharge Measurements

Flow duration at an ungaged stream site can be estimated by comparing the ungaged flows with concurrent daily flows at a gaged site, which is referred to as an index station. During this study the outflow from each aquifer was measured at least three times. Estimates of the flow duration were improved by incorporating flow data from past studies. All measurements at each outflow site were plotted against concurrent daily flows at each of three long-term, continuous-record gaging stations at which flow durations are known. The long-term stations used were Sevenmile River at Spencer, Quaboag River at West Brimfield, and Squannacook River near West Groton. The established station having the best relation with the outflow site was used as the index station for the flow-duration analysis of that site. In every case, the relationship between the outflow site and the index station was better at higher flow durations (lower flows).

The flows at an ungaged site are usually compared to the flows at an index station either graphically or by a mathematical equation. The Stedinger-Thomas mathematical method (1985) was used in this study. Flow-duration values at the ungaged outflow site of each aquifer were estimated from the flow duration of the index site using the Stedinger-Thomas line of relation drawn for that stream. An example of a generic line of relation is shown in figure 11A. Using the figure, if the 95-percent flow duration at the index station is 28 ft^3/s, a flow of 2.5 ft^3/s is an estimate of the 95-percent flow duration at the ungaged outflow site.

Streamflow at the downstream end of each aquifer comprises both ground-water discharge from the aquifer and surface-water inflows to the aquifer. The surface-water component of the measured outflow of each aquifer was determined by measuring the inflows of the main stream and of the tributaries where they entered the aquifer. The ground-water discharge was the difference between the measured inflows and outflow.

Estimates of long-term yield of each of the six aquifers include the maximum potential induced infiltration

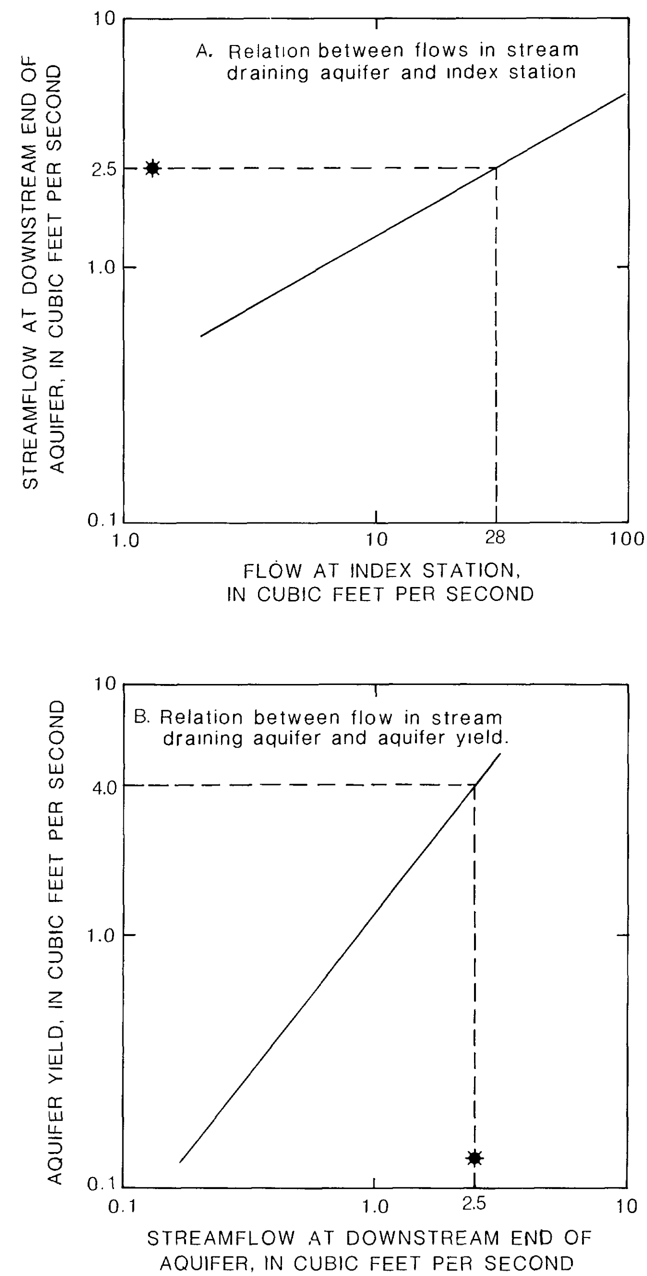


Figure 11.--Example of method for determining aquifer-yield durations.

from the stream. The potential induced infiltration values were calculated using a streambed conductivity of 2 ft/d and a thickness of 1 ft, as determined from the infiltration tests. Stream area and depth were determined from field reconnaissance. Potential induced infiltration, computed as above, was compared to measured stream inflows, and the lesser value was adopted for use in estimating long-term aquifer yield.

Long-term aquifer yield was calculated as the sum of the increase in streamflow through the reach, the adopted induced infiltration value, and the current ground-water withdrawal. Because large supply wells withdraw water from most of the aquifers, and because available streamflow does not always infiltrate, the potential yield of the aquifer usually does not equal the streamflow measured at the downstream end of the aquifer.

To compare streamflow to potential aquifer yield, the measured streamflow must be adjusted to account for development and for variations in the quantity of induced infiltration. For each aquifer, streamflows measured at the downstream end of the aquifer were each plotted against potential aquifer yield (the sum of ground-water discharge, induced infiltration, and withdrawals), at the time the flow measurement was made, and a line of relation drawn between the two. This line was used to graphically convert flow durations to equivalent yield durations for that aquifer. Figure 11B shows an example of the method. If the 95-percent flow duration was 2.5 ft^3/s, the 95-percent yield duration would be 4.0 ft^3/s. The yield is larger than the streamflow because of current withdrawals. Typically during summer when the streamflows are lower (at a higher flow duration), water demand is greater and more water is pumped from wells than during winter. Therefore, the difference between streamflow and aquifer yield is not constant.

If an aquifer were developed to yield the estimates given, all ground water that would naturally discharge to streams would be intercepted and all surface water would be infiltrated; the stream would cease to flow during at least part of the year. Aquifer yield values were therefore adjusted to maintain streamflow at a level presently equaled or exceeded 95 or 99.5 percent of the time. These flow durations encompass the range that water resources planners generally consider to indicate low and very low flows. Where no flow duration information is available, some water planners choose to set streamflow requirements based on the size of the drainage area. To compare our results to this method, estimates of aquifer yield were also adjusted to maintain streamflow equal to 0.2 $ft^3/s/mi^2$ (cubic feet per second per square mile) of drainage area.

Determining Long-Term Aquifer Yield from Percent of Basin Covered by Stratified Drift

Flow-duration curves, which relate a given flow to the percentage of time that flow is equaled or exceeded, range from gentle to steep depending on the hydrogeologic characteristics of the basins drained by the streams. In areas of stratified drift where the ground is highly permeable and the storage capacity is relatively high, precipitation enters the ground and is released to the stream gradually during both wet and dry periods. Thus, the flow-duration curve for this area would have a gentle slope. In areas of till where the ground is less permeable, little precipitation is taken into storage, thus surface-water runoff is high during wet periods and little ground water is discharged to streams during dry periods. The flow-duration curve for a stream draining this area would be steep (fig. 12).

Several independent studies in New England (Thomas, 1966, Cervione and others, 1972, and Lapham, 1988) derived families of flow-duration curves comparing the percentage of a basin covered by stratified drift to the stream discharge per square mile at different flow durations. Streamflow for Stillwater River and Wekepeke Brook was estimated using each of the sets of curves, and compared to actual flow measurements. Estimates of streamflow using

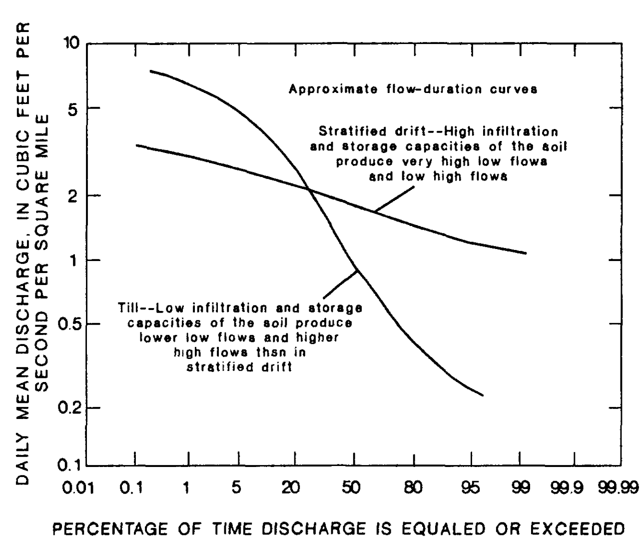


Figure 12.--Typical flow-duration curves for streams draining till and stratified drift.

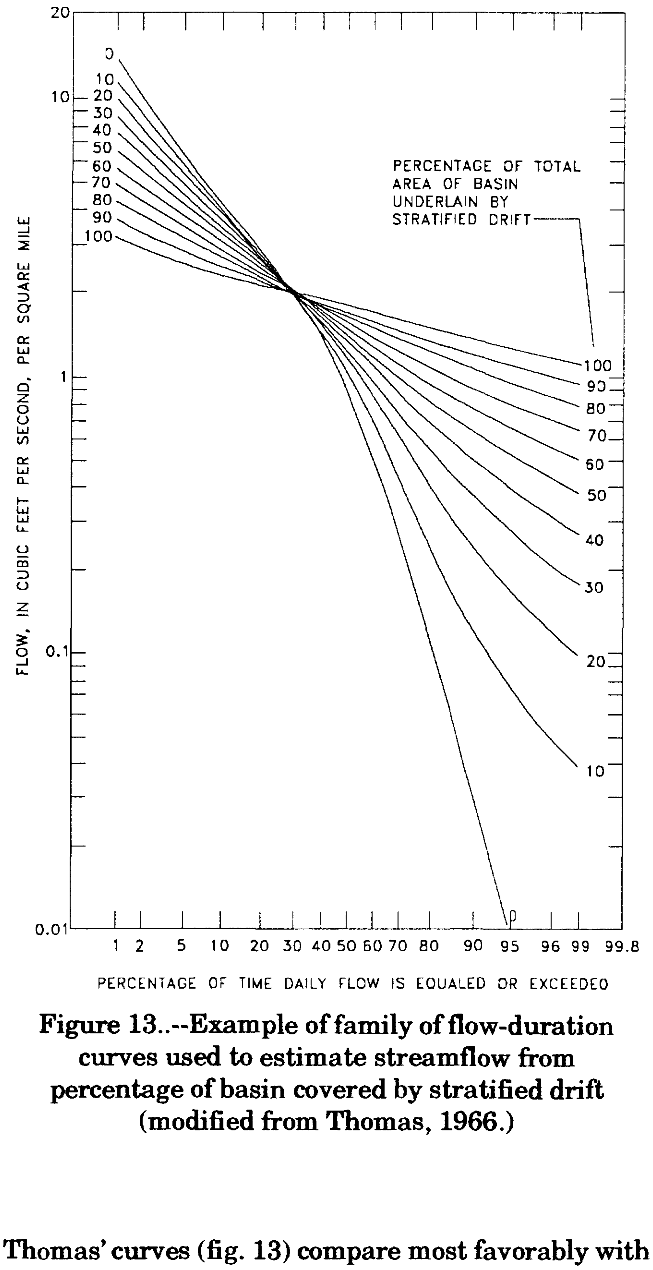


Figure 13..--Example of family of flow-duration curves used to estimate streamflow from percentage of basin covered by stratified drift (modified from Thomas, 1966.)

Thomas' curves (fig. 13) compare most favorably with measured streamflow particularly at the important lower flows (fig. 14) and were used to estimate streamflow and yield in the other aquifers.

Estimates of flow in the Upper Catacoonamug Brook, Still River, and Witch Brook aquifers based on Thomas' curves are for total natural streamflow. Comparison of potential induced infiltration to the inflow to these aquifers, measured during 1984 and 1985, indicates that all surface water in Still River and Witch Brook could be infiltrated and all surface water in the Upper Catacoonamug Brook could be infiltrated at flow durations greater than about 70 percent. Therefore long-term aquifer yield is equivalent to the streamflow estimated using Thomas' method. As with the estimates calculated from streamflow measurements, these yields were adjusted to maintain specified streamflows.

Methods of Estimating Short-Term Yield from Storage

Several methods are available for estimating aquifer yield available from storage. In this study, ground-water-flow models of McDonald and Harbaugh (1988) were used instead of an analytical (image well) approach because of the ease of distributing well fields in the modeled area and the possibility of simulating

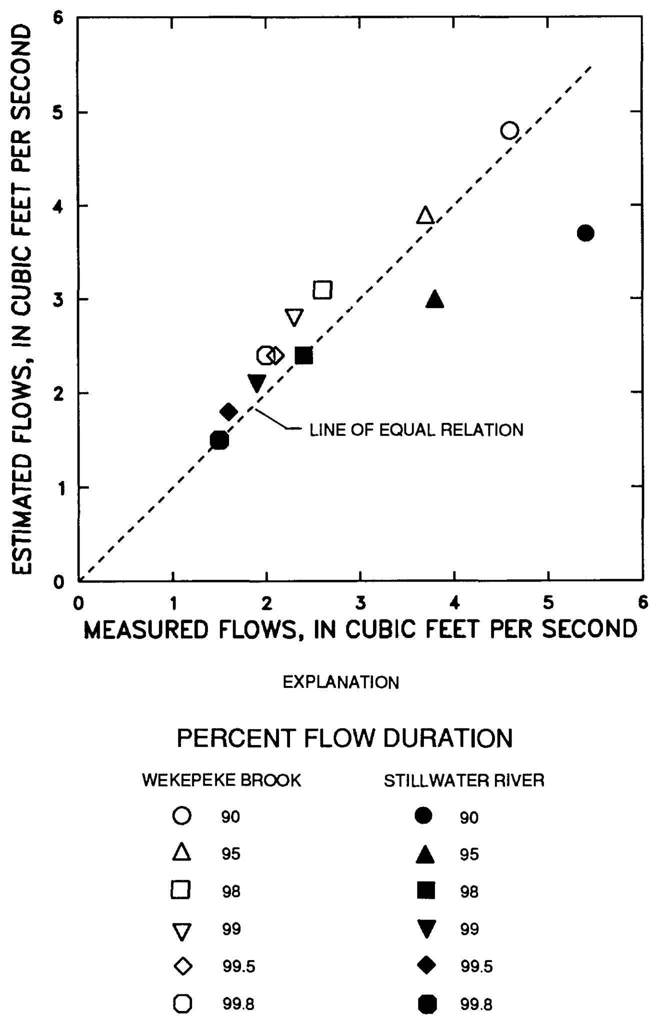


Figure 14.--Comparison of measured streamflow and streamflow estimated from percentage of basin covered by stratified drift.

leakage from a surface-water body which does not fully penetrate the aquifer.

For each of the six aquifer areas studied, a ground-water-flow model was used to estimate the average rates at which water could be pumped from storage with 180 days of no recharge and with a drawdown at pumped well fields of 50 percent of the original saturated thickness. The models were highly simplified representations of the aquifers and were not calibrated to produce accurate water-table simulations using different pumping conditions. Transmissivity data from Brackley and Hansen (1977), updated with new information, were used to construct models of each of the six aquifers. Hydraulic conductivity was constant throughout each model area, and the variation in transmissivity was simulated by varying the saturated thickness. The storage coefficient was assigned a value of 0.20 for each aquifer. No streams were simulated in the models because during severe drought most would be dry; in areas where streams might continue to flow, the estimates of yield from storage would be conservative. Lake Shirley was simulated because it covers a large percentage of the Catacoonamug aquifer. It was assumed that the vertical hydraulic conductivity of the silty lake-bottom deposits was 0.4 ft/d.

A single, high-yielding well lowers the water level in a "cone of depression" around the well; a well field of many small wells pumping simultaneously, however, produces many cones of depression that together lower the water table to a relatively uniform level beneath the well field. Thus, to approximate a 50 percent dewatering of each aquifer, individual model nodes, each measuring 1,000 ft by 500 ft, distributed throughout the most highly transmissive sections of the aquifer, were designated as well fields. The head in these nodes was set constant at 50 percent of the total saturated thickness (Lapham, 1988); the rate of discharge from each of these constant-head nodes equals the pumping rate of a well field that would cause a drawdown of 50 percent of the saturated thickness. In each aquifer simulation, the sum of the average pumping rates from the individual well fields was determined at the end of an 180-day pumping period.

The short-term yields calculated for each aquifer represent large withdrawals from many wells that produce water through substantial dewatering of the aquifer. Users cannot expect to withdraw water at these rates for long periods of time.

Aquifer Yields

Potential long-term yields of the Pearl Hill and Willard Brooks, Stillwater River, and Wekepeke Brook aquifers were estimated from streamflow measurements; long-term yields of the Still River, Catacoonamug Brook, and Witch Brook aquifers were estimated from the percentage of the basin underlain by stratified drift or by a combination of the two methods.

Data collected for this study were used to update and modify the transmissivity map of Hydrologic Atlas 276 (Brackley and Hansen, 1977). Therefore, the aquifer maps in the following sections are based on all available data. High-yielding wells, streamflow-measurement sites, and seismic-refraction-survey lines are also on the maps.

The graphs of potential aquifer yield included in the following sections of this report depict the percentage of time that a given yield can be equaled or exceeded; they are <u>yield-duration</u> curves. Because withdrawal of the total aquifer yield (solid line on the yield duration curves) might cause the streams to cease flowing, an adjusted yield is also shown for which streamflow can be maintained at or above the present 95- and 99.5-percent <u>flow duration</u> on the main streams. These flow duration values were chosen to represent low and very low flow in the streams. In addition, an adjusted yield is shown for which streamflow can be maintained at 0.2 ft^3/s per square mile of the drainage area. The graphs show a yield-duration line for each of the selected streamflow criteria. Thus, for a given minimum required streamflow, water managers can use the appropriate line to determine the potential yield of the aquifer and determine the potential effect a given withdrawal will have on the stream. In all cases, the potential yields given would require extensive development of the aquifer. To compare the current withdrawal with the potential yield, the higher of the average 1984 and average 1985 withdrawal rates is also plotted on the graphs.

The streamflow duration values are based on streamflow measurements made after the aquifers had been developed. Therefore, even if streamflow must be maintained at 95-percent flow duration, the graphs show there is yield available more than 95 percent of the time because the aquifer yield is equal to the measured streamflow plus the current ground-water withdrawal. The actual withdrawal varies with the seasons; the withdrawal value plotted on the graphs is the average for the year and does not always match the field conditions on which the analysis is based. Changes in withdrawal rates or minimum streamflow requirements will affect the percentage of time that a given yield can be equaled or exceeded.

Pearl Hill-Willard Brooks aquifer

Pearl Hill Brook rises in the northeastern corner of Fitchburg and flows north to join Willard Brook near Ash Swamp in West Townsend where the two streams become the Squannacook River (fig. 23, located with the other site maps, beginning on page 36). There is very little stratified drift underlying Willard Brook. Stratified drift, composed mainly of medium sand, fills the north-south trending valley of Pearl Hill Brook. The aquifer is about 50-ft thick, 0.75-mi wide, and 4.5-mi long. The most transmissive part of the aquifer is along Pearl Hill Brook and along Mason and Walker Brooks, which drain into Ash Swamp.

Townsend's main water supply is a well field in a gravelly zone in this aquifer near the confluence of Pearl Hill and Willard Brooks. This field supplied 143.9 Mgal of water to the town in 1984 (an average of 0.39 Mgal/d) and 135.4 Mgal in 1985 (0.37 Mgal/d).

The estimate of short-term yield from aquifer storage for the Pearl Hill-Willard Brooks aquifer was based on pumpage from 5 hypothetical well fields located in the most transmissive part of the aquifer. If the water level under each well field were drawn down 30 ft (50 percent of the original saturated thickness), water could be pumped from storage at an average rate of 5.8 Mgal/d for short periods of time. However, because large withdrawals from aquifer storage cannot be maintained over time, the estimate of long-term aquifer yield is a better assessment of the quantity of water available from the aquifer.

Long-term aquifer yields estimated for the Pearl Hill-Willard Brooks aquifer at various yield durations are given in figure 15. For example, if streamflow at the Mason Road outflow site is at least 0.7 ft^3/s (99.5-percent flow duration) the 1984 withdrawal of 0.39 Mgal/d can be maintained about 99.2 percent of the time. If streamflow is at least 2.6 ft^3/s (95-percent flow duration) the 1984 withdrawal would be available only 95 percent of the time. Withdrawal of 1 Mgal/d would be possible 97 percent of the time for a minimum streamflow of 99.5-percent flow duration.

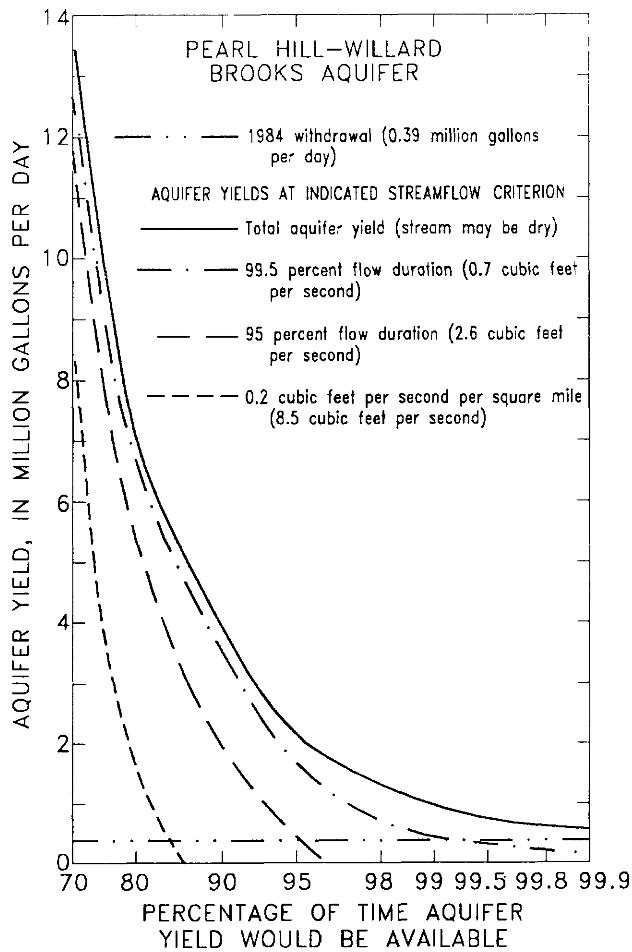


Figure 15.--Yield of the Pearl Hill-Willard Brooks aquifer.

Stillwater River aquifer

The Stillwater River begins at the confluence of Keyes Brook and Justice Brook, on the Princeton-Sterling town boundary at the northwest corner of figure 24, and flows south through the western part of Sterling to Wachusett Reservoir, (fig. 1) where the water becomes part of the metropolitan Boston supply. The aquifer follows the river and is about 1-mi wide and 4-mi long. Data from wells and a seismic-refraction survey indicate that the aquifer fills a bedrock channel about 100-ft deep (fig. 29) and consists of fine to medium sand with some lenses of gravel. Many tributaries flow into the Stillwater River, but because the aquifer is narrow, most of the tributaries flow only a short distance over the stratified drift. The aquifer materials extend up the tributary valleys only along Houghton and Wilder Brooks.

Sterling has a gravel-packed well in a wetland area near the upper end of the aquifer between Moores Corners and West Sterling. This well pumped 87.8 Mgal in 1984 (0.24 Mgal/d) and 65.9 Mgal in 1985 (0.18 Mgal/d). The town has another well field (Sterling Well #1) in sand and gravel to the east of the main aquifer, but this area does not drain to the Stillwater River and was not included in the study.

The estimate of short-term yield from aquifer storage for the Stillwater River aquifer was based on pumpage from 8 hypothetical well fields distributed throughout the aquifer. If the water level under each well field were lowered 40 ft, water could be pumped from storage at an average rate of 10.3 Mgal/d for short periods of time.

Long-term yield of the Stillwater River aquifer was estimated from the measured outflow below Stones Bridge on Muddy Pond Road and from the inflows at the upstream end of the aquifer and on several tributary streams. Figure 16 shows the yield estimates for the aquifer. The average 1984 withdrawal of 0.24 Mgal/d can be met about 99.3 percent of the time when streamflow is at least 1.6 ft^3/s (99.5-percent flow duration) and 94 percent of the time when streamflow is at least 2.7 ft^3/s (95-percent flow duration). Withdrawal of 1 Mgal/d would be possible 94 percent of the time when a streamflow of at least 99.5-percent flow duration is maintained.

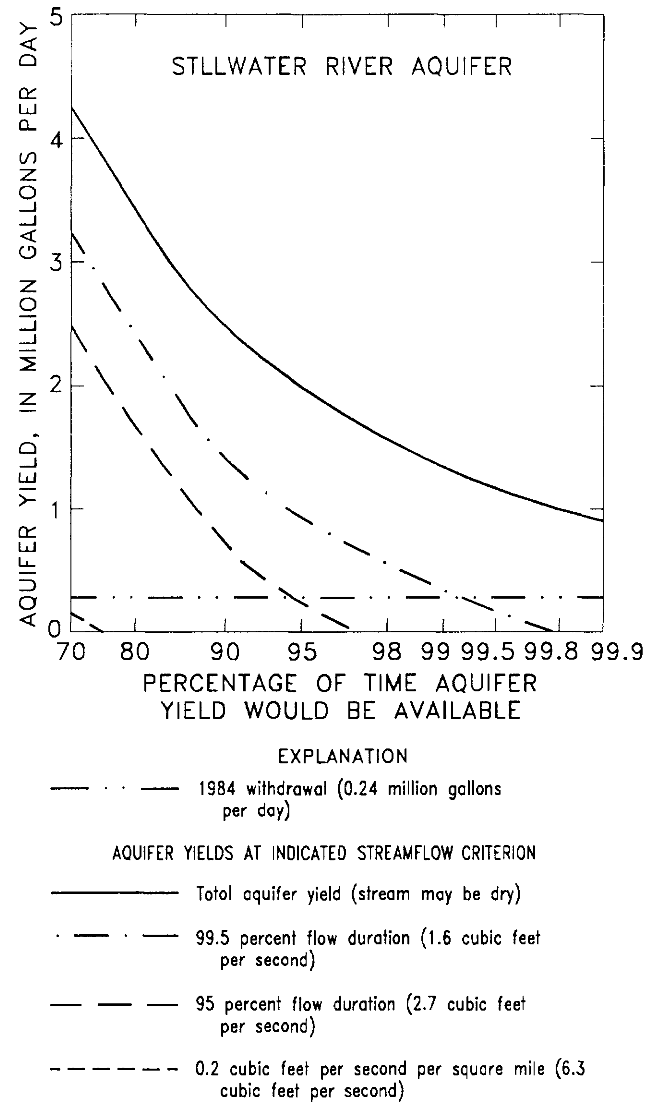


Figure 16.--Yield of the Stillwater River aquifer.

Wekepeke Brook aquifer

Wekepeke Brook rises above Heywood Reservoir in northern Sterling and flows east to Pratt Junction. The area upstream of State Highway 12 (fig. 25), is fed by several springs which were once used by Clinton for water supply. Clinton (south of the basin) now holds the area in reserve. The area has also been considered for inclusion in the Metropolitan District Commission's User Sources Alternative program as a surface-water source (Wallace Floyd Assoc, Inc, 1985). Another branch of Wekepeke Brook flows north from the center of Sterling to join the main brook just south of Pratt Junction. From there, the brook flows north to the North Nashua River.

The most transmissive part of the Wekepeke Brook aquifer is a deposit of fine to medium sand and gravel up to 90-ft thick located northeast of Pratt Junction, along a tributary of Wekepeke Brook. This area is about 2-mi long and less than 0.5-mi wide (fig. 29). Leominster has three wells in this area which produced 198.7 Mgal in 1984 (0.54 Mgal/d), and 53.0 Mgal in 1985 (0.15 Mgal/d). The wells produce less than their capacity during the dry summer months and are used as a reserve to the surface-water reservoirs that are the city's main source of supply. In 1985, Leominster completed a connection to Wachusett Reservoir and began purchasing water from the Massachusetts Water Resources Authority.

There is much public interest in the Wekepeke Brook aquifer because the towns of Leominster, Sterling, and Lancaster have developed, or plan to develop, supply wells in the area. In addition, Interstate 190 (not shown on fig. 25) crosses the aquifer and a new industrial park is planned on land adjacent to the aquifer.

Sterling has several test holes near Wekepeke Brook in the northeastern corner of town and Lancaster has test holes along the North Nashua River 0.5 mile upstream from its confluence with Wekepeke Brook. The Lancaster test site is in a stratified-drift deposit that does not drain to Wekepeke Brook and therefore it was not included in this study. However, there is the potential that further development in the Wekepeke Brook aquifer could affect this site because the stratified drift is continuous between the two areas.

Average short-term yield from storage in the Wekepeke Brook aquifer was 6.6 Mgal/d. This estimate was based on pumpage from 6 hypothetical well fields and a water level decline of 25 ft under each well field.

The long-term yield estimates (fig. 17) indicate that the 0.54 Mgal/d of water withdrawn by Leominster in 1984 would be available about 98.5 percent of the time if streamflow is at least 2.1 ft^3/s (99.5-percent flow duration) and about 94 percent of the time when streamflow is at least 3.3 ft^3/s (95-percent flow duration). One Mgal/d could be maintained 96 percent of the time with a minimum streamflow of 99.5-percent flow duration.

Surface water in the Wekepeke Brook was also measured at State Highway 12 because of the interest in the area upstream of this point as a surface-water supply. The potential induced infiltration from Wekepeke Brook below Leominster Road is greater than the baseflows measured at that site. Therefore, any surface water removed from the upstream area would lower the potential long-term yield of the downstream area. In addition, if the upstream area were developed for ground water, the yields downstream would be even lower because ground-water underflow to the lower part would also be intercepted.

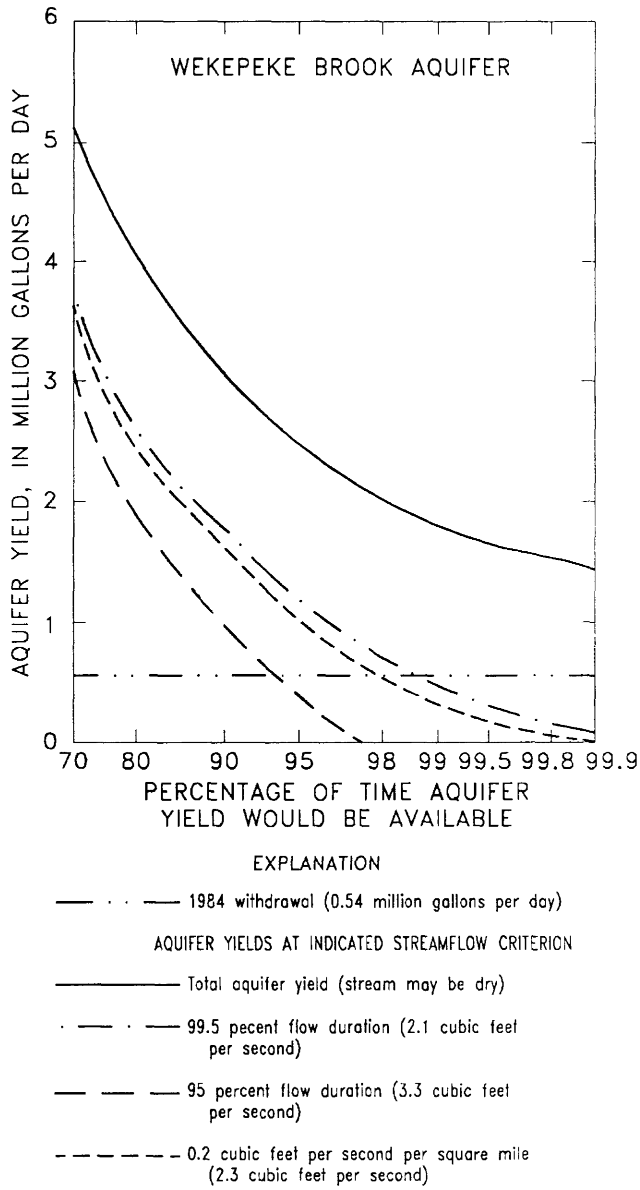


Figure 17.--Yield of the Wekepeke aquifer.

Still River aquifer

The Still River flows west from the center of Bolton and then north along the flood plain on the eastern side of the Nashua River (fig. 26). Several unnamed tributaries to the Still River drain the aquifer; one flows directly to the Nashua River at the southern end of the area studied.

Data from lithologic logs of wells suggest that the coarse sand and gravel exposed in sand pits at the eastern side of the aquifer extends under thick, stratified deposits of fine silt and clay that cover the main river valley. The aquifer, which is about 4-mi long and 1.5-mi wide, is nearly 200-ft thick in some areas (figs. 29 and 30), with 25 to 50 ft of coarse-grained material at the bottom. The town of Lancaster has two gravel-packed wells in the coarse-grained material in the southern part of the aquifer that together yielded 181.2 Mgal of water in 1984 (0.50 Mgal/d) and 178.6 Mgal in 1985 (0.49 Mgal/d).

Short-term yield from storage in the Still River aquifer was estimated with pumpage from 8 hypothetical well fields distributed throughout the aquifer. A water-level decline of 45 ft would produce an average of 13.1 Mgal/d from storage for short periods of time.

Estimates of long-term yield from the Still River aquifer (fig. 18) were based on the percentage of the drainage area underlain by stratified drift because not all of the ground water from the Still River aquifer discharges to streams in which flow measurements were made. The sum of the baseflow measurements was one-tenth of the flows estimated from the percentage of stratified drift for similar durations because aquifer material extends beneath the fine silt and clay which impede ground-water flow to the stream. Data collected for this study do not indicate where the ground water is discharging; it may leave the area as underflow to the north. Wells developed in the coarse material at the bottom of the aquifer might capture this underflow.

The aquifer and streams on the flood plain of the Nashua River are poorly connected and well withdrawals have little effect on flow in the streams. For this reason, aquifer yield estimates in figure 18 are adjusted to maintain prescribed quantities of flow only in the reaches of the streams that cross the exposed stratified drift. The average 1984 withdrawal by Lancaster of 0.50 Mgal/d can be maintained 99.3 percent of the time if streamflow is at least 1.6 ft^3/s (99.5-percent flow duration) and 95 percent of the time if streamflow is at least 2.4 ft^3/s (95-percent flow duration). One Mgal/d could be maintained 95 percent of the time with a minimum streamflow of 99.5-percent flow duration.

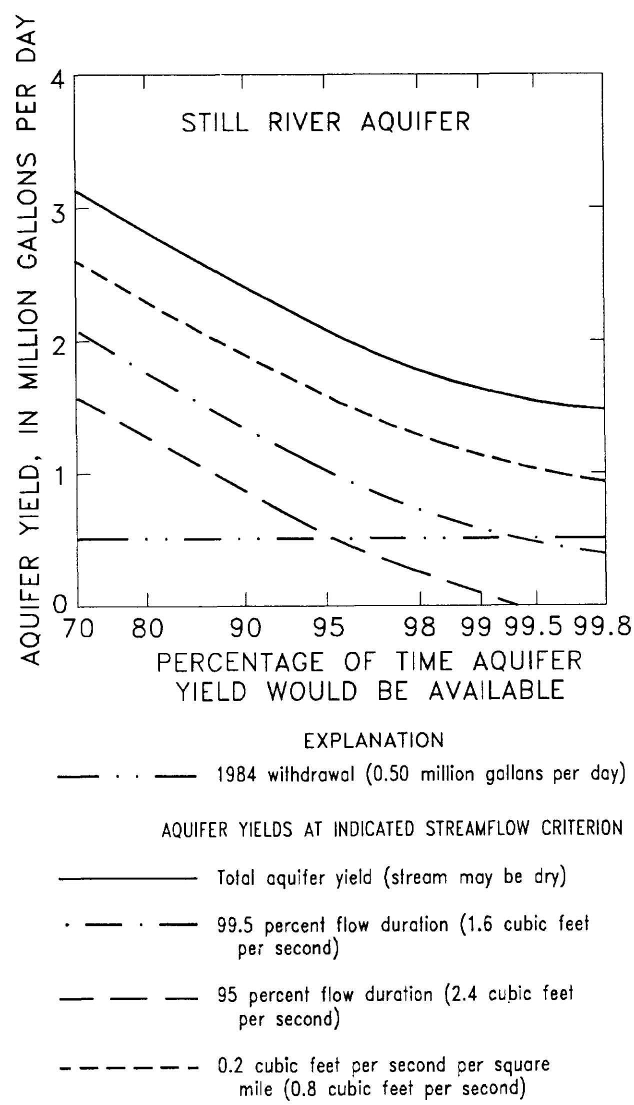


Figure 18.--Yield of the Still River aquifer.

Catacoonamug Brook aquifers

Catacoonamug Brook rises in the center of Lunenburg and flows southwest to the Nashua River on the eastern edge of Shirley (fig. 27). In the 1850s, Shakers repaired and enlarged a small dam on the brook at the Lunenburg-Shirley town line, creating Lake Shirley (William Kelly, town of Shirley, oral commun., 1986). Catacoonamug Brook flows into the northern end of this lake, exits at the dam on the eastern side, and flows through Shirley Village. The 360-acre, spring-fed lake averages 6-ft deep and is surrounded by about 300 houses.

Several streams drain the aquifer to the west and upstream from Lake Shirley. These include a major unnamed tributary of Catacoonamug Brook that flows through White Rabbit Swamp, an unnamed stream that flows from Lake Whalom (not shown on map, located west of Cross Road) near Leominster through Massapoag Pond, and Easter Brook which flows from North Leominster into the western shore of Lake Shirley. Spruce Swamp Brook and Bow Brook are major tributaries to Catacoonamug Brook downstream from the lake.

The two most transmissive areas on the western side of Lake Shirley are located along Catacoonamug

Brook near the confluence of the tributary from Massapoag Pond. The area farthest north consists of sand and gravel deposits 30- to 40-ft thick. Lunenburg has developed three gravel-packed wells and one well field in this area, although one well is no longer used because of bacterial problems (James Deming, Town of Lunenburg, oral commun., 1984). The total volume pumped from the Lunenburg municipal wells in 1984 was 99.1 Mgal (0.27 Mgal/d); in 1985 it was 106.1 Mgal (0.29 Mgal/d). Lunenburg has drilled test wells in the more southern area, which has about 45 ft of saturated stratified drift. Lithologic logs of wells suggest that this material has discontinuous lenses of coarse sand and gravel.

The area south of Lake Shirley has deposits of stratified drift 80-ft thick. Wells in this area yield 0.72 Mgal/d for industrial use. The homes around the lake depend on water from shallow sand points in the stratified drift. Some of these wells go dry each fall when the level of the lake is lowered to kill the shoreline vegetation and to allow repair of docks and beaches.

Downstream from the lake, the town of Shirley has a gravel-packed well in 35 ft of gravel close to Catacoonamug Brook. Although this well is capable of pumping 400 gal/min, the water needs chlorination and the well has not been used heavily since construction of the gravel-packed well on Patterson Road in the eastern section of town (Vernon Griffith, town of Shirley, oral commun., 1984). In 1984, the well pumped 11.9 Mgal (0.03 Mgal/d) and, in 1985, 11.7 Mgal (also 0.03 Mgal/d). In addition, there are five dug public-supply wells near Catacoonamug Brook in the center of Shirley Village, 0.4 miles above the brook's confluence with the Nashua River. Although these wells are outside the area studied, their yield could be affected by development upstream because of reduced surface water available for induced infiltration, and reduced ground-water underflow.

The estimate of short-term yield from storage in the Catacoonamug aquifer was based on pumpage from 11 hypothetical well fields around the perimeter of Lake Shirley. If the water level under each well field were drawn down 20 ft, an average of 10.3 Mgal/d would be available from storage for short periods of time.

Because the Catacoonamug Brook aquifer is divided by Lake Shirley, yield estimates were calculated for two areas. The yield of the entire area was estimated from the measured outflow at Lancaster Road in Shirley and inflows on Catacoonamug Brook and on the two major tributaries west of the lake. (The tributaries east of the lake rise within the stratified drift.) Long-term yield estimates for the entire aquifer (fig. 19) assume no change in the volume of Lake Shirley and include the quantity of water currently withdrawn from the public-supply wells in Lunenburg and Shirley and the large industrial well. The 1985 withdrawal of 1.04 Mgal/d can be pumped about 99.1 percent of the time if the streamflow is maintained at 99.5-percent flow duration (2.5 ft^3/s) and 95 percent of the time with streamflow at 95-percent flow duration (3.8 ft^3/s).

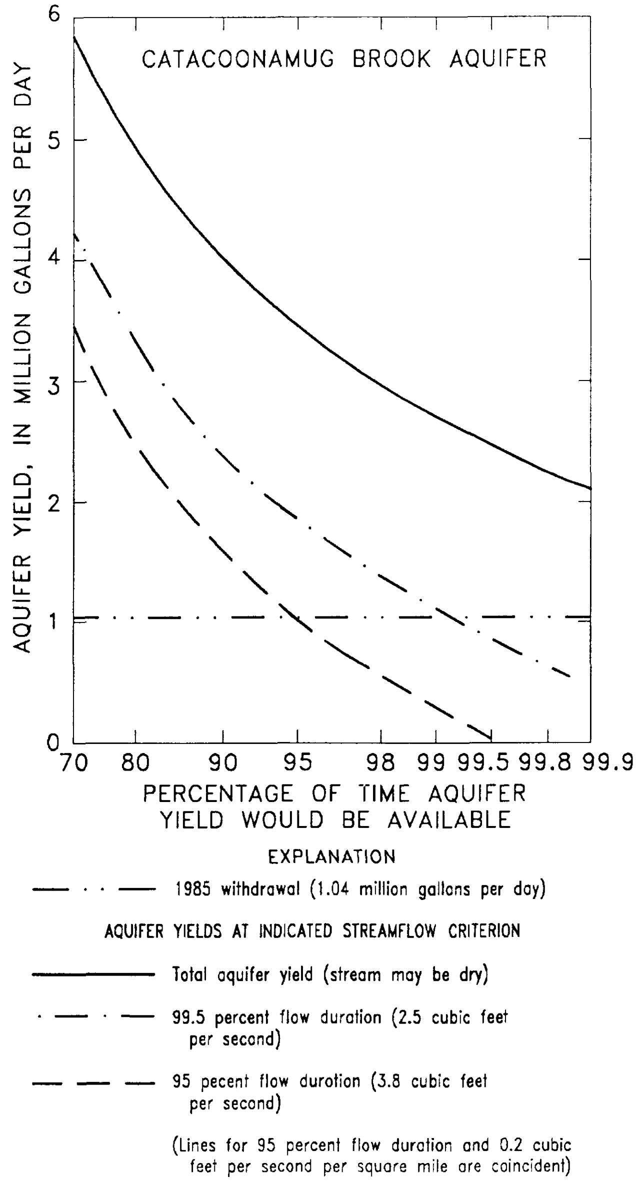


Figure 19.--Yield of the Catacoonamug Brook aquifer.

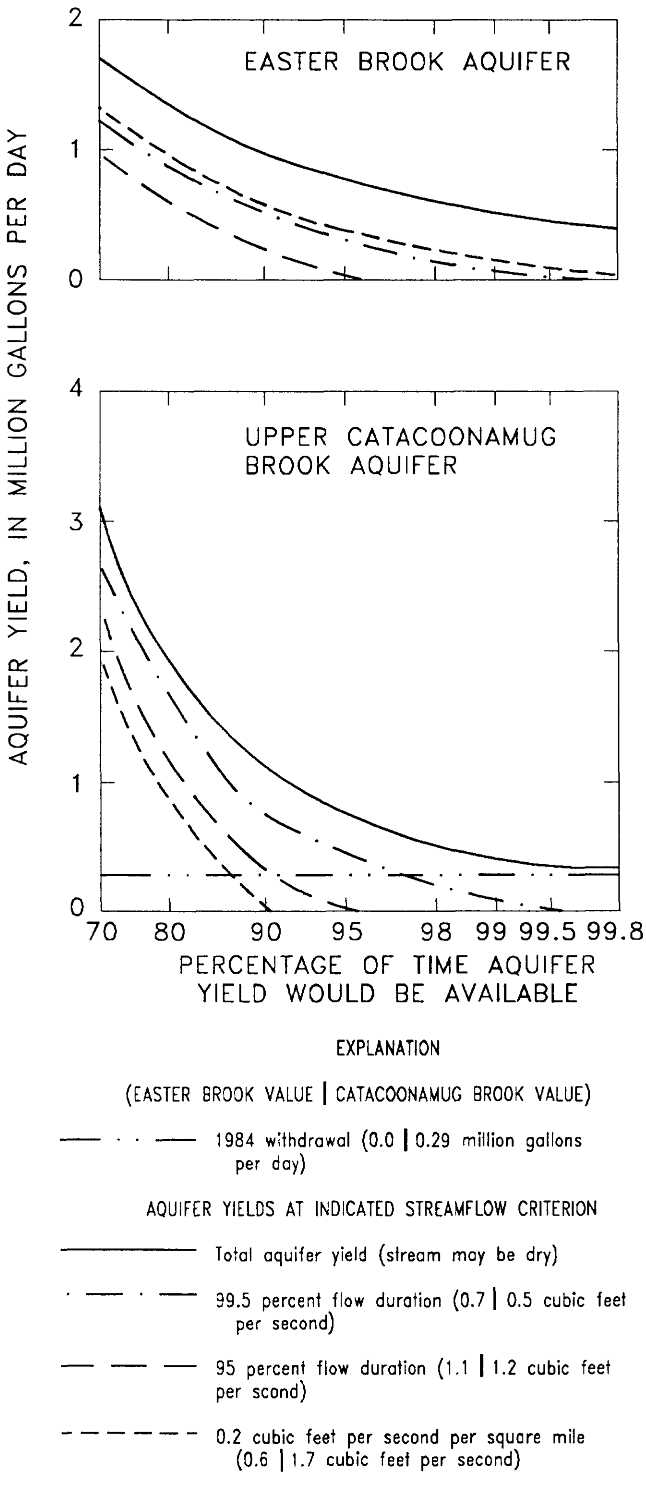


Figure 20.--Yield of the aquifers along tributaries to Lake Shirley.

Long-term aquifer yield estimates given in figure 20 for Catacoonamug and Easter Brooks west of Lake Shirley are based on the percentage of stratified drift in each drainage area because some of the ground water from these aquifers discharges directly to the lake. In the Upper Catacoonamug Brook aquifer, the 1985 withdrawal of 0.29 Mgal/d by Lunenburg can be pumped about 97 percent of the time if streamflow is at least 0.5 ft^3/s (99.5-percent flow duration) and about 90 percent of the time if streamflow is at least 1.2 ft^3/s (95-percent flow duration). One Mgal/d could be maintained 86 percent of the time with a minimum streamflow of 99.5-percent flow duration.

Witch Brook aquifers

The Witch Brook aquifers are an area of stratified drift south of the Squannacook River between Townsend Center and Townsend Harbor (fig. 28). Witch Brook drains a swampy area near South Row Road and flows north and east to the Squannacook River near the Groton town line. The aquifer area is also drained by Bixby Brook, which flows from above Bixby Reservoir in the southern part of Townsend, to Harbor Pond. The most transmissive part of the Witch Brook aquifers is a sand and gravel deposit about 4-mi long, 1-mi wide, and 50-ft thick, which fills the deepest part of a bedrock channel (fig. 30). Stratified-drift extends across the Squannacook River and yields of this regional aquifer were also estimated.

The Witch Brook Water Company, a private water supplier to a housing development, has two gravel-packed wells near a small tributary to Witch Brook. The two wells pumped 99.4 Mgal (0.27 Mgal/d) in 1984, and 94.4 Mgal (0.26 Mgal/d) in 1985.

Short-term yield from aquifer storage was calculated with pumpage of 8 hypothetical well fields in the Witch Brook aquifer. A water-level decline of 35 ft under each well field would produce an average of 10.8 Mgal/d for short periods of time. The long-term yield of the aquifers along Witch and Bixby Brooks (fig. 21) was estimated from the percentage of stratified drift covering each basin. Results show that average 1984 withdrawal of 0.27 Mgal/d from the Witch Brook area can be withdrawn 93 percent of the time if streamflow is at least 0.5 ft^3/s (99.5-percent flow duration) and 86 percent of the time if streamflow is at least 0.8 ft^3/s (95-percent flow duration). One Mgal/d could not be withdrawn from the Witch Brook area with minimum streamflow at any of the given criteria. One Mgal/d

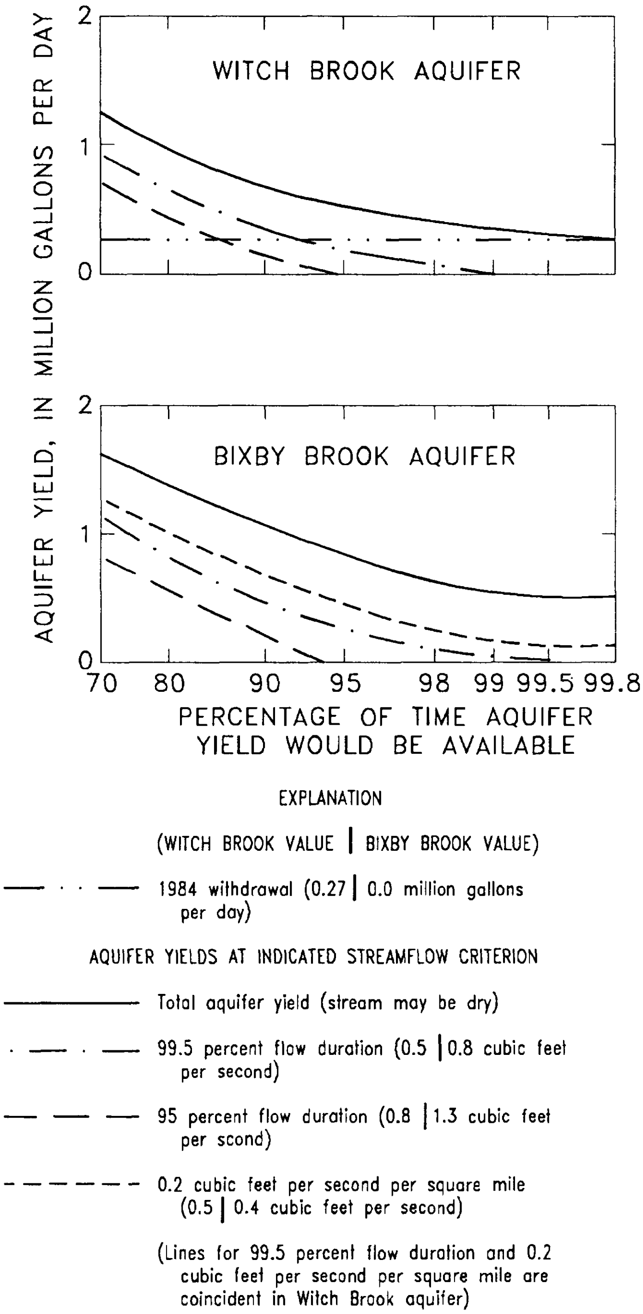


Figure 21.--Yield of the Witch Brook aquifers.

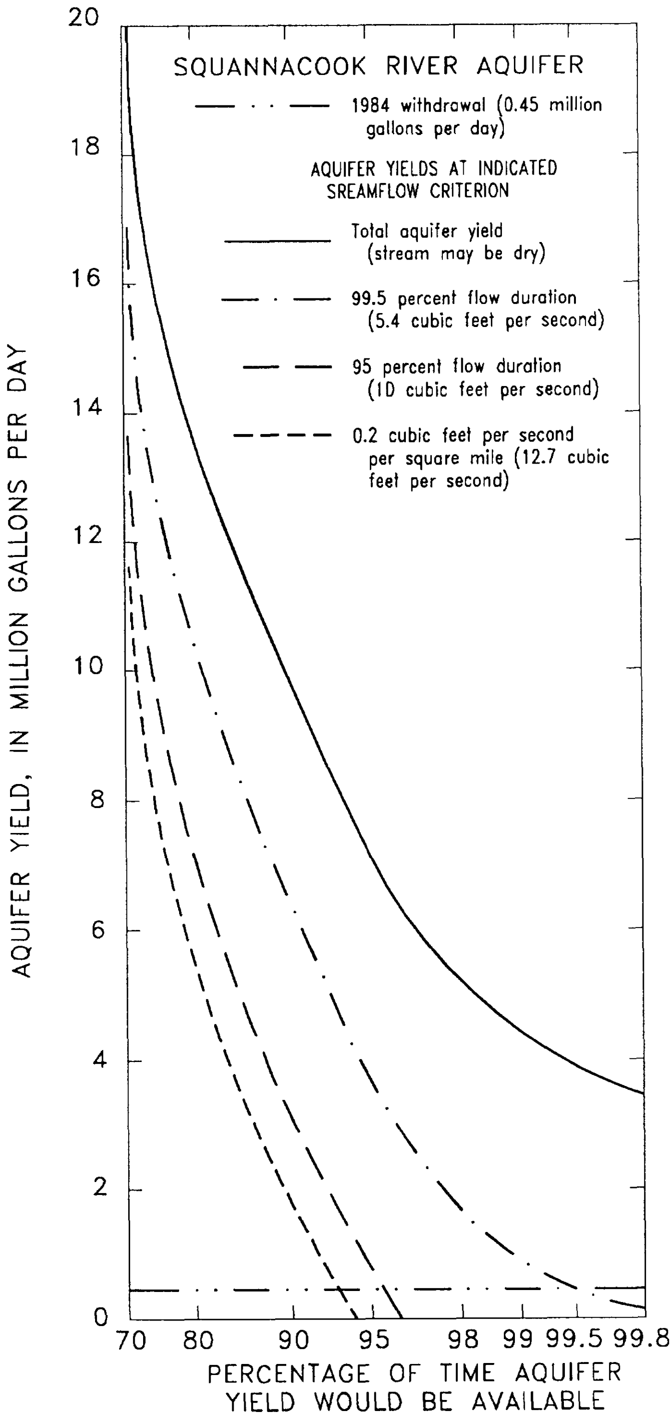


Figure 22.--Yield of the regional aquifer along the Squannacook River adjacent to Witch Brook.

could be maintained in the Bixby Brook area 74 percent of the time with a minumum streamflow of 99.5-percent flow duration.

The long-term yields estimated from percentage of stratified drift are higher than the sum of the baseflows measured because some ground water discharges directly to the Squannacook River rather than to Witch or Bixby Brook. To include this ground-water discharge in aquifer yield estimates, an analysis was done of the regional aquifer along the Squannacook River based on streamflows recorded at the Squannacook River gage (fig. 22).

Townsend has a gravel-packed well north of the town center (Townsend Well #2). This well pumped 63.7 Mgal (0.17 Mgal/d) in 1984 and 66.2 Mgal (0.18 Mgal/d) in 1985. The combined withdrawal from the regional aquifer of 163.1 Mgal (0.45 Mgal/d) in 1984 and 160.6 Mgal (0.44 Mgal/d) in 1985 can be maintained about 99.5 percent of the time with streamflow of at least 5.4 ft^3/s (99.5-percent flow duration) and 95 percent of the time with streamflow of at least 10 ft^3/s (95-percent flow duration). One Mgal/d could be maintained 98.8 percent of the time with a minimum streamflow of 99.5-percent flow duration. Induced infiltration accounts for the difference in these two analyses. Approximately one third of the flow at the Squannacook River gage at the downstream end of the regional aquifer entered the aquifer area as surface-water inflow. Any reduction of this inflow by aquifer development upstream would reduce the potential yields.

The Pearl Hill-Willard Brooks and the Witch Brook aquifers are part of a single system along the Squannacook River. Thus, the yield from the downstream Witch Brook aquifer or the regional Squannacook River aquifer will decrease if the Pearl Hill-Willard Brooks area is developed. This decrease would be the result of less ground-water flow through the stratified drift along the Squannacook River and less surface water available for infiltration. If the Pearl Hill-Willard Brooks aquifer were developed maintaining a flow in the Squannacook River of at least 0.7 ft^3/s (99.5-percent flow duration), only this quantity would be available for induced infiltration to the downstream area. Baseflow measurements made at the outflow site of the Pearl Hill-Willard Brooks aquifer in 1984 and 1985 ranged from 2.6 to 5.3 ft^3/s, all of which could be infiltrated. These data suggest that yields from the downstream area would be lowered by at least 2 ft^3/s (1.3 Mgal/d).

Summary of Aquifer Yields

In the six aquifers studied, current withdrawals can be pumped 93 to 99.5 percent of the time while maintaining very low streamflow at 99.5-percent flow duration, and from 86 to 95 percent of the time while maintaining low streamflow at 95-percent flow duration. Table 2 summarizes the yields. The percentage of stratified drift is given only for those areas where it was used to estimate the yield. The streamflow criteria are listed for each aquifer in cubic feet per second, and the aquifer yield estimates are given in million gallons per day. All yield estimates include the current water withdrawals (the higher of the 1984 and 1985 average rates).

In all of the aquifers, the yield which can be maintained 95 percent of the time with minimum streamflow of 99.5-percent flow duration is greater than or equal to the current average withdrawals. In seven of the eight developed areas, current withdrawals approach or exceed the yields that can be maintained 95 percent of the time with streamflow at 95-percent flow duration; withdrawals exceed yields that can be maintained 99 percent of the time in all the aquifers for a 95-percent flow duration; and in seven of the eight areas, withdrawals approach or exceed yields that can be maintained 99 percent of the time at 99.5-percent flow duration.

Comparing current withdrawals with the yields available 95 percent of the time while maintaining streamflow at 99.5-percent flow duration shows that the Still River, Catacoonamug Brook, and Witch Brook aquifers are pumped at the highest rates relative to their potential yields and that the regional aquifer along the Squannacook River is pumped at the lowest rate relative to its potential yield.

Aquifer	Pumpage as percent of aquifer yield
Pearl Hill-Willard Brooks	24
Stillwater River	26
Wekepeke Brook	49
Still River	51
Catacoonamug, total	57
Upper Catacoonamug	39
Witch and Bixby Brooks	55
Regional Squannacook River	12

No flow measurements or detailed analyses are required to determine streamflow if the 0.2 $ft^3/s/mi^2$ criterion is used because the value is calculated from the size of the drainage area. Table 2 illustrates that this criterion does not have a consistent relation to the flow duration criteria because the percentage of stratified drift in the drainage area is not considered. In five cases, the flow that is equivalent to 0.2 $ft^3/s/mi^2$ is less than the flow at 95-percent flow duration; in one case, it is equal to it; in four cases, it is greater. The flow at 0.2 $ft^3/s/mi^2$ is greater than the 99.5-percent flow duration in all of the aquifers except the Witch Brook, Bixby Brook, Still River, and Easter Brook aquifers.

Table 2.--*Estimated yield from stratified-drift aquifers in the Nashua River basin (streamflow maintained at specified minimum low flow).*

[--, data not applicable; mi², square miles; Mgal/d, million gallons per day; ft³/s, cubic feet per second]

Aquifer	Drainage area (mi²)	Percent stratified drift[1]	Short term yield[2] (Mgal/d)	Streamflow criteria[3], in cubic feet per day			Long-term yield, in Mgal/d at 95-percent duration maintaining streamflow			Long-term yield, in Mgal/d at 99-percent duration maintaining streamflow			Higher of 1984/85 pumpage (Mgal/d)
				percent 95	percent 99.5	(ft³/s)/mi² [4]0.2	percent 95	percent 99.5	(ft³/s)/mi² [4]0.2	percent 95	percent 99.5	(ft³/s)/mi² [4]0.2	
Pearl Hill-													
Willard Bks	42.3	--	5.8	2.6	0.7	8.5	0.39	1.62	0	0	0.45	0	0.39
Stillwater R	31.6	--	10.3	2.7	1.6	6.3	.20	.91	0	0	.33	0	.24
Wekepeke Bk	11.6	--	6.6	3.3	2.1	2.3	.36	1.10	.96	0	.42	.28	.54
Still R[5]	4.2	67	13.1	2.4	1.6	.8	.50	.99	1.53	.09	.58	1.12	.50
Catacoonamug Bk													
Upper and lower	19.1	--	10.3	3.8	2.5	3.8	.97	1.81	.97	.25	1.09	.25	[6]1.04
Upper[7]	11.6	--	--	2.3	1.2	2.3	0	.74	.33	0	.17	.10	.29
Catacoonamug Bk	8.6	16	--	1.2	.5	1.7	0	.45	0	0	.11	0	.29
Easter Bk	3.0	37	--	1.1	.7	.6	0	.29	.33	0	.06	0.10	0
Witch Bks[8]	5.1	--	10.3	2.1	1.3	.9	0	.49	.67	0	.07	.26	.27
Witch Bk	2.3	36	--	.8	.5	.5	0	.21	.23	0	.03	.06	.27
Bixby Bk	2.8	44	--	1.3	.8	.4	0	.28	.44	0	.04	.20	0
Squannacook R[9]	63.7	--	--	10	5.4	12.7	.65	3.62	0	0	.90	0	.45

[1]Percent stratified drift was used to estimate yield only in aquifers where percent is given.
[2]Available from aquifer storage.
[3]These values were used to adjust the aquifer yields presented in the next columns.
[4]Streamflow criterion based on 0.2 ft³/s/mi² of drainage area.
[5]Streamflow criteria based on tributaries upstream of the fine-grained deposits.
[6]Includes withdrawal by the industrial well.
[7]The Upper Catacoonamug Brook aquifer yield is the sum of the two areas below.
[8]The Witch Brook aquifer yield is the sum of the two areas below.
[9]The regional area adjacent to the Witch Brook aquifer is drained by the Squannacook River.

The streamflow analysis was limited to the major streams; some of the small tributaries might cease flowing even using the adjusted yields. Estimates of aquifer yield that can be maintained more than 90 percent of the time (greater than 90-percent yield duration) are more accurate than estimates for lower durations because the relation between the outflow site and the index station was better at higher flow durations.

Short-term yields from aquifer storage ranged from 5.8 Mgal/d in the Pearl Hill-Willard Brooks aquifer to 13.1 Mgal/d in the Still River aquifer. The latter aquifer has a greater storage capacity because the stratified-drift deposits along the Still River occupy a wider and deeper bedrock channel than in the other areas. The Catacoonamug Brook aquifer covers the largest area, but the yield from storage after 180 days ranks third because Lake Shirley covers a large percentage of the area and limits the number of well fields that can be located in the areas of highest transmissivity.

Appraisal of Aquifer-Yield Estimates

Estimates of aquifer yield are based on all available information about the physical and hydraulic properties of the aquifers. Estimated yield may differ from actual yields because field and pumping conditions may invalidate one or more of the assumptions on which the estimates are based. If, for example, estimated values of transmissivity or specific yield differ from those assumed in the short-term-yield model, the predicted yield might vary. The estimates of long- and short-term aquifer yield are likely to be higher than the quantities that can practically be withdrawn from an aquifer for several reasons:

1) Land availability--Many areas that could support municipal wells have already been developed for other, incompatible uses.

2) Aquifer materials--Some aquifer areas of high transmissivity are difficult to develop for water supply. Some deposits, although thick and containing large quantities of water, consist of fine material that does not transmit water readily. Other areas of coarse material have little saturated thickness, which limits the available drawdown in a well.

3) Well construction--Wells are not completely efficient and do not withdraw all the available water.

4) Water quality--No attempt was made to exclude from the yield estimates those areas where water quality problems exist.

5) Cost--The cost of fully developing the aquifers might be prohibitive.

All aquifer yield estimates represent long-term averages and cannot be used to predict how much ground water will be available in a given year. In a dry year, aquifer yields will be less than those presented here. Conversely, in a wet year, more ground water will be available. If an aquifer has a lot of storage, variations in aquifer recharge may be lessened; the storage has a stabilizing effect.

Estimates of long-term aquifer yield were made assuming that water derived only from intercepted ground-water discharge and induced infiltration. Estimates of short-term aquifer yield were made assuming that water derived only from storage. However there are other sources that may increase yield, such as water captured from reduced evapotranspiration when the water table is lowered and return flow from waste-water discharge.

Estimates of short-term aquifer yield from storage were made based on 180 days of no recharge during a severe drought. The analysis assumed that aquifer storage was filled to capacity at the beginning of the 180-day pumping period. If the water table was already lowered by a previous dry period, yields from storage would be less. Because water derived only from aquifer storage, steady-state conditions are never achieved; continued pumping will cause water levels to continue to decline.

The aquifer-yield estimates in this study were made from a regional perspective; they cannot be used to assess the impact of ground-water development on local water levels or ground-water flow patterns. This type of site-specific analysis must be based on detailed study of the local aquifer.

Operational considerations are also a factor in maintaining aquifer yield. When streamflow must be maintained above a specified flow, ground-water withdrawals from the aquifer must consider the time lag between changing of pumping rates and the effect at the streams. Because it will be difficult to predict and control these effects, withdrawals may have to be less than the yields presented, to assure maintaining the required streamflow.

SUMMARY AND CONCLUSIONS

Vertical hydraulic conductivities typical of small sandy-bottomed streams in Massachusetts were determined from four infiltration tests: two tests on each of two streams. The vertical conductivity of the two streambeds ranged from 2.0 to 5.0 ft/d. The conductivity value of 2 ft/d was used to estimate water available from induced infiltration as part of long-term aquifer yield.

Water quality was sampled near the pumped wells during the induced infiltration tests. Although the water-quality data could not be used as the primary method of estimating the surface-water contribution to the well, the data confirmed that water quality in the stream has an effect on the quality of water pumped from the well.

Ground-water-flow models of hypothetical aquifers were used to determine what well spacing along the streams would produce the hydraulic conditions needed to cause induced infiltration at the maximum rate. The necessary spacing ranged from 1,000 ft in low-yielding aquifers (withdrawal of about 1 Mgal/d to individual wells) to 8,000 ft in extremely high yielding aquifers (about 7 Mgal/d).

During prolonged periods of no recharge, water is available from aquifer storage for short periods of time. Short-term yields of the six aquifers ranged from 5.8 Mgal/d in the Pearl Hill-Willard Brooks aquifer to 13.1 Mgal/d in the Still River aquifer.

Long-term yields (calculated as the sum of intercepted ground-water discharge, induced infiltration from surface water, and current withdrawals) were estimated for the six aquifers by using one of two methods. One was based on flow measurements at the downstream end of the aquifer; the second related streamflow to percentage of the basin covered by stratified drift.

Long-term aquifer yields are presented as the percentage of time a particular yield is available with streamflow of at least 95-percent flow duration (low flow) or 99.5-percent flow duration (very low flow). Yields that can be maintained 95 percent of the time with streamflow of at least 99.5-percent flow duration range from 0.49 Mgal/d in the Witch Brooks aquifer (Witch and Bixby Brooks combined) to 3.62 Mgal/d in the regional aquifer along the Squannacook River. Yields that can be maintained 95 percent of the time with streamflow of at least 95-percent flow duration range from 0 Mgal/d in the Witch Brooks and Upper Catacoonamug Brook aquifers to 0.97 Mgal/d in the total Catacoonamug aquifer (upper and lower sections combined). Yields that can be maintained 99 percent of the time with streamflow of at least 99.5-percent flow duration range from 0.07 Mgal/d in the Witch Brooks aquifer to 1.09 Mgal/d in the total Catacoonamug Brook aquifer. Withdrawals are possible 99 percent of the time with streamflow of at least 95-percent flow duration only in the Still River and total Catacoonamug Brook aquifers; 0.09 and 0.25 Mgal/d respectively.

All of the aquifers are used for water supply. The total Catacoonamug Brook, Witch Brooks, and Still River aquifers were pumped at the highest rates relative to their 95-percent yield durations: 57, 55, and 51 percent respectively, while maintaining streamflow at 99.5-percent flow duration. The regional aquifer along the Squannacook River is pumped at the lowest rate relative to its yield, 12 percent.

The quantity of water available for induced infiltration and the potential aquifer yields in the Nashua River basin probably are typical of small valley-fill aquifers in the Northeast and may be useful for predicting yields of these aquifers.

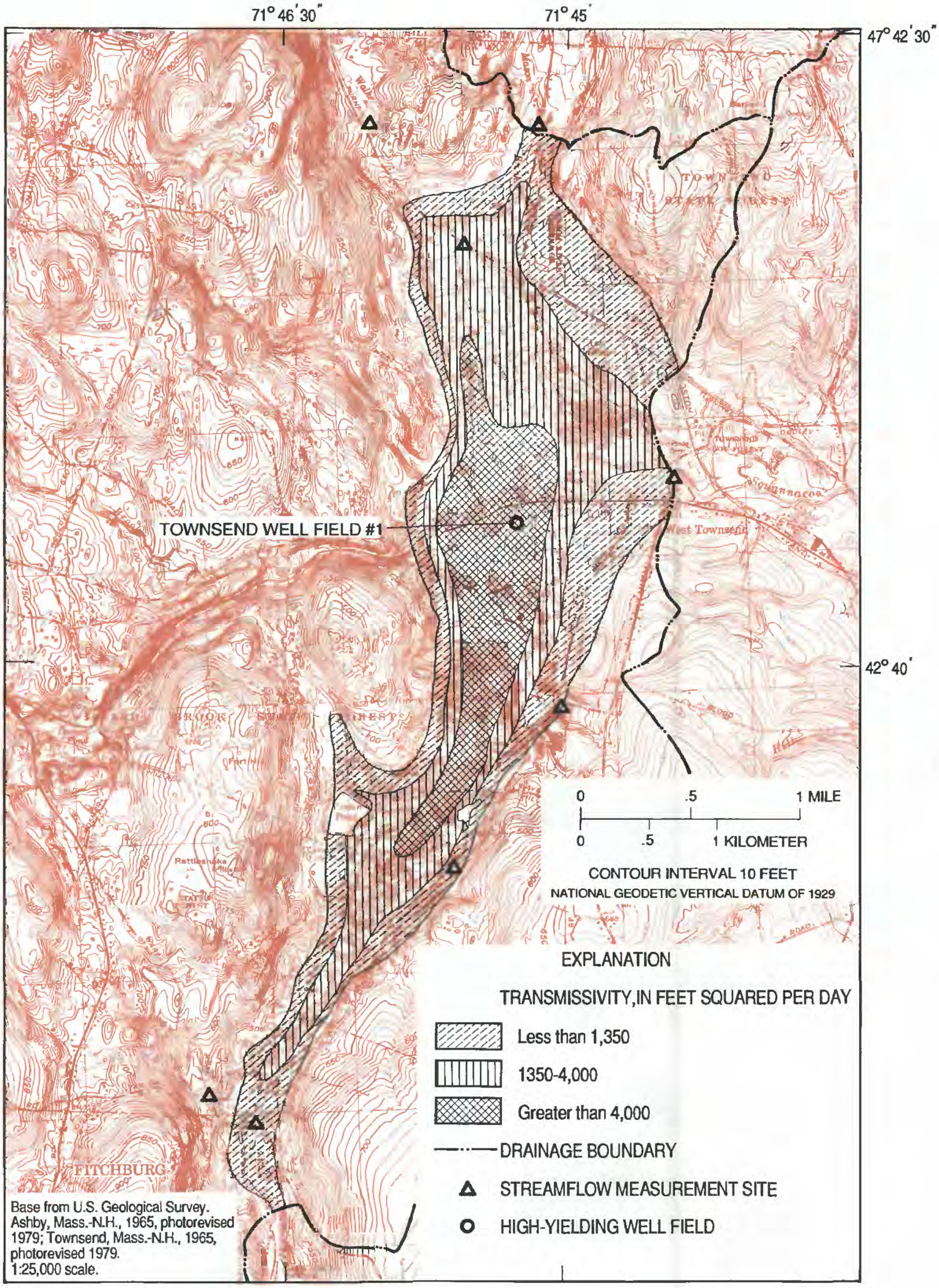


Figure 23.--Areal distribution of transmissivity in the Pearl Hill-Willard Brooks aquifer and location of the water-supply well field and streamflow measurement sites. (Modified from Brackley and Hansen, 1977).

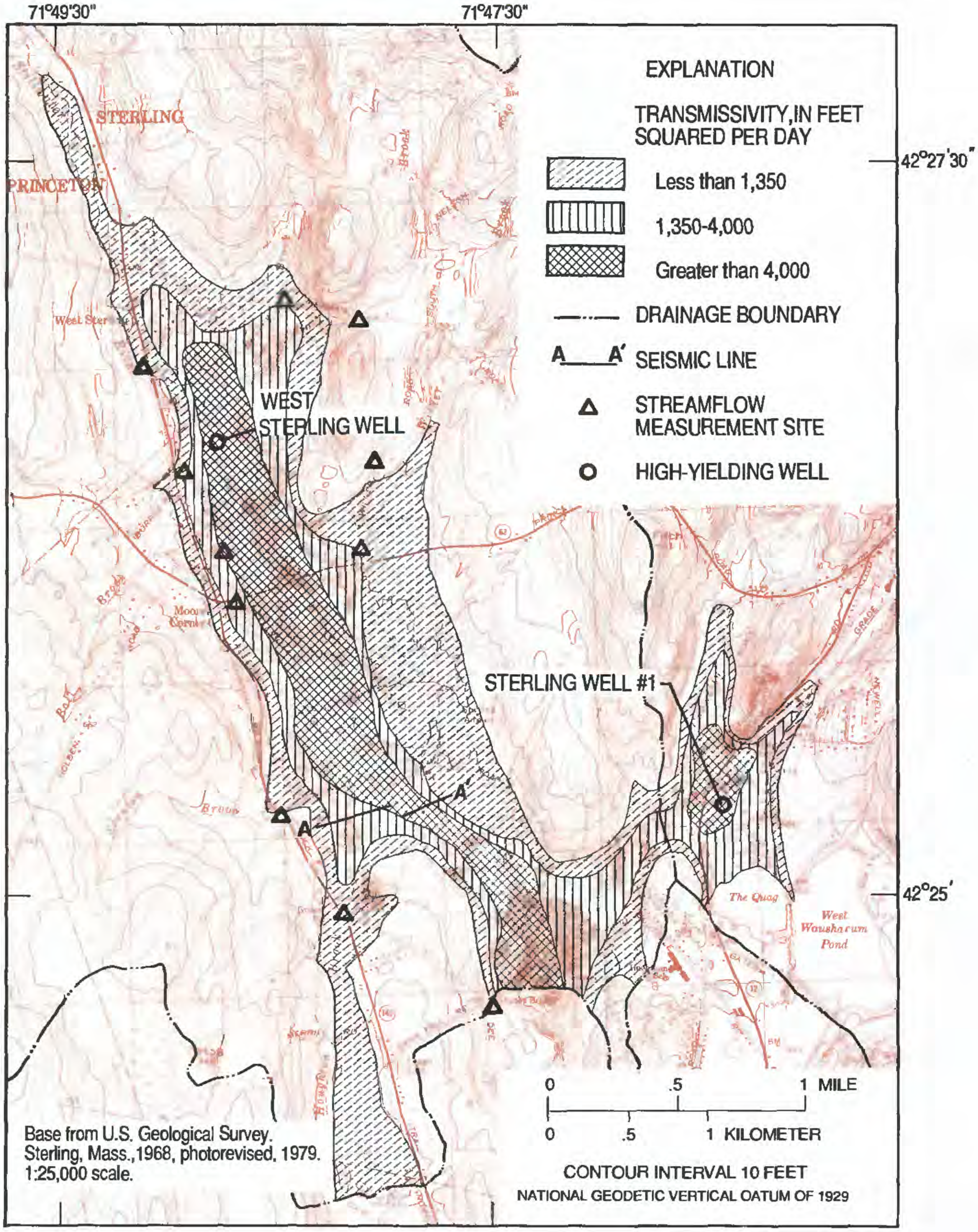


Figure 24.--Areal distribution of transmissivity in the Stillwater River aquifer and location of the water-supply well, streamflow measurement sites, and seismic-refraction survey. (Modified from Brackley and Hansen, 1977).

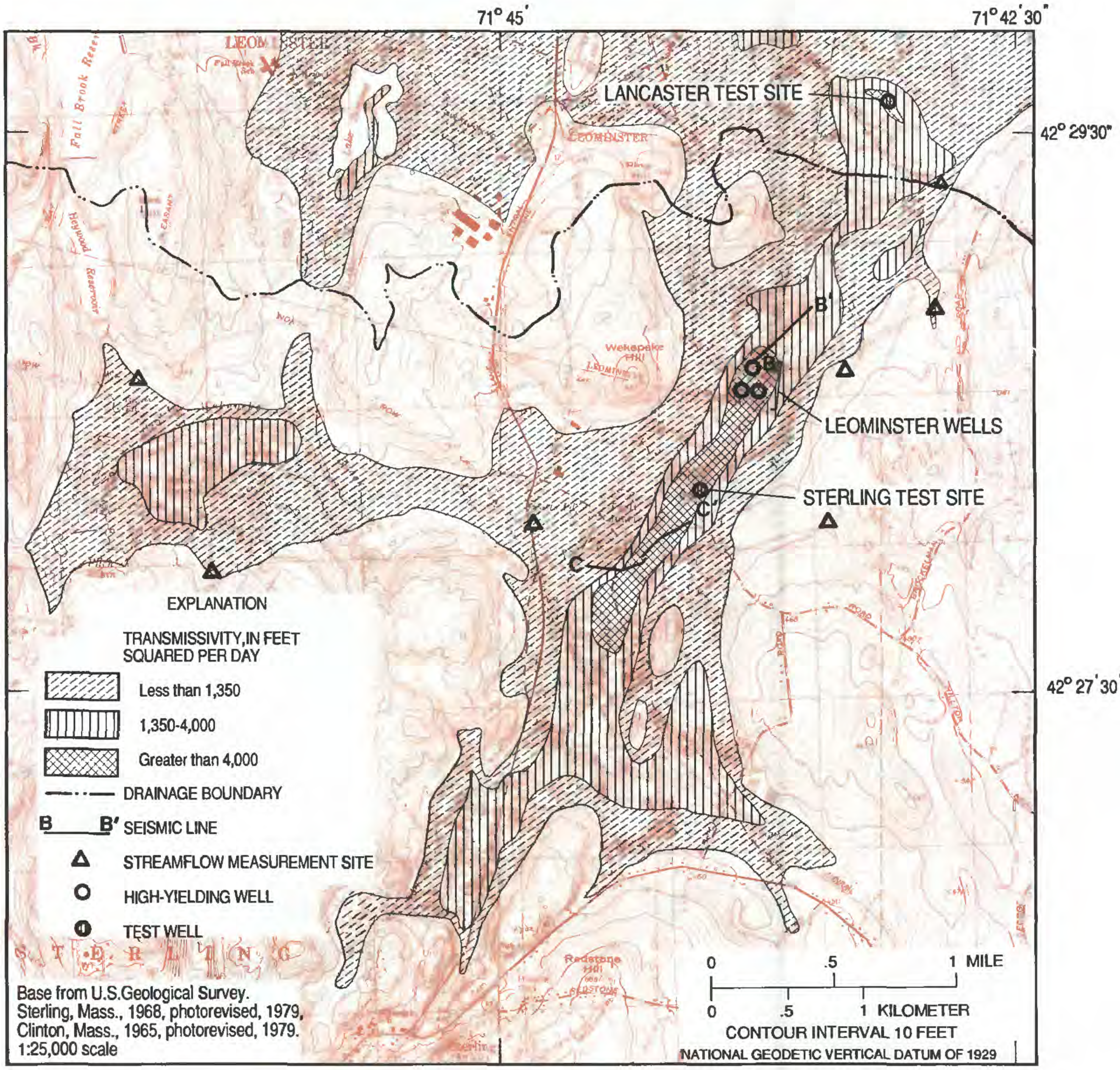


Figure 25.--Areal distribution of transmissivity in the Wekepeke Brook aquifer and location of the water-supply wells, streamflow measurement sites, and seismic-refraction surveys. (Modified from Brackley and Hansen, 1977.)

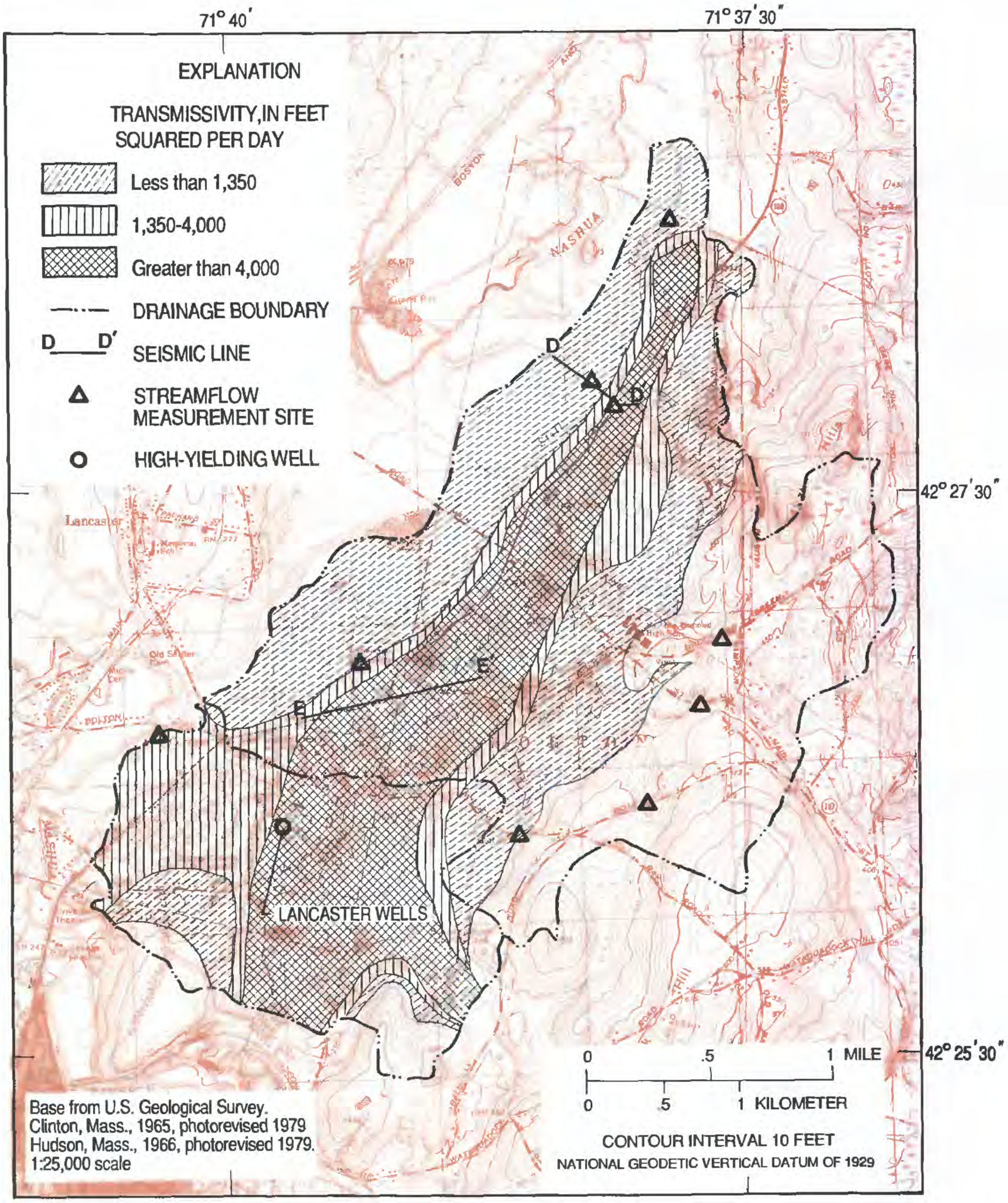


Figure 26.--Areal distribution of transmissivity in the Still River aquifer and location of the water-supply wells, streamflow measurement sites, and seismic-refraction surveys. (Modified from Brackley and Hansen, 1977.)

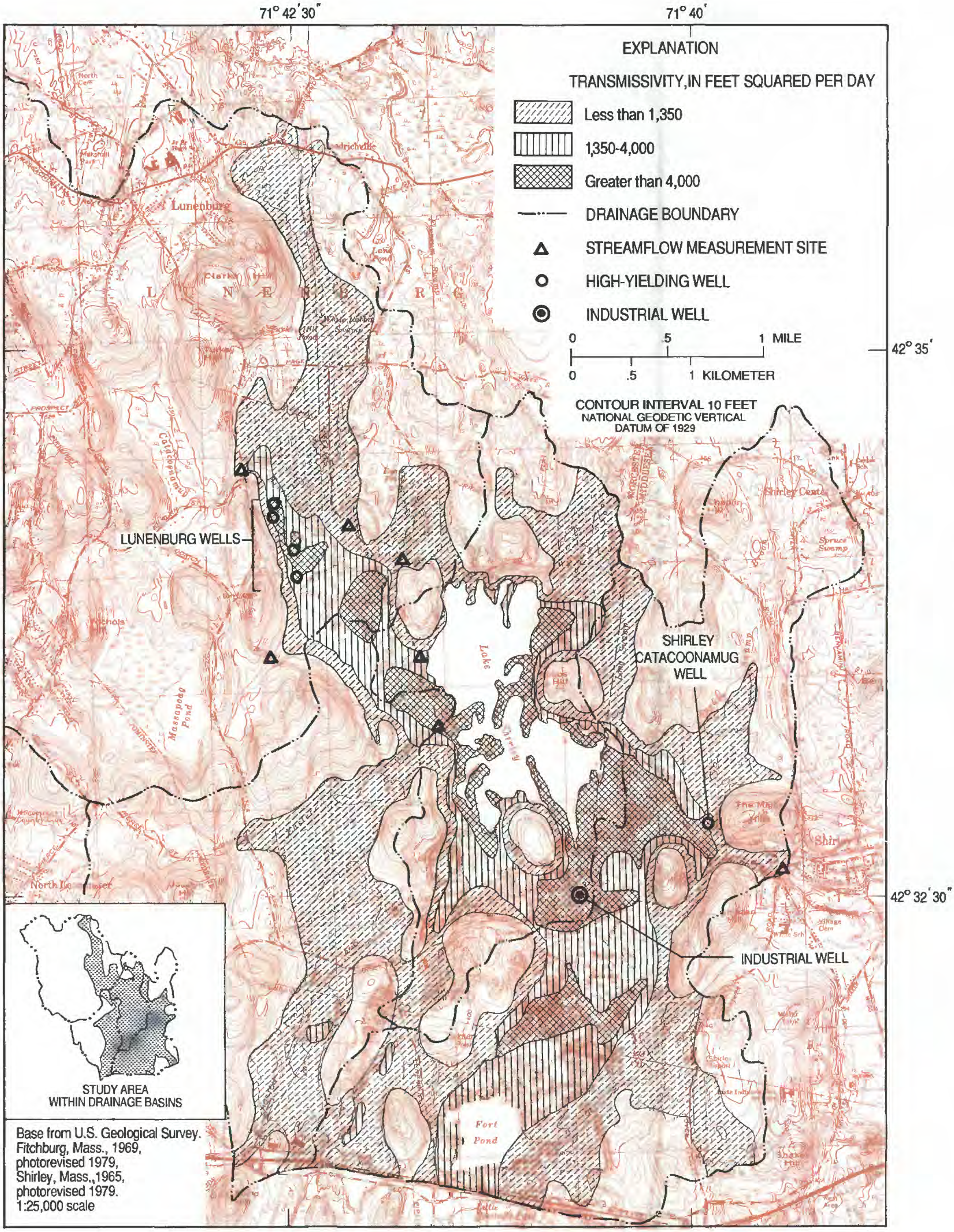


Figure 27.--Areal distribution of transmissivity in the Catacoonamug Brook aquifer and the location of water-supply wells and streamflow measurement sites. (The index map shows the entire drainage area.) (Modified from Brackley and Hansen, 1977.)

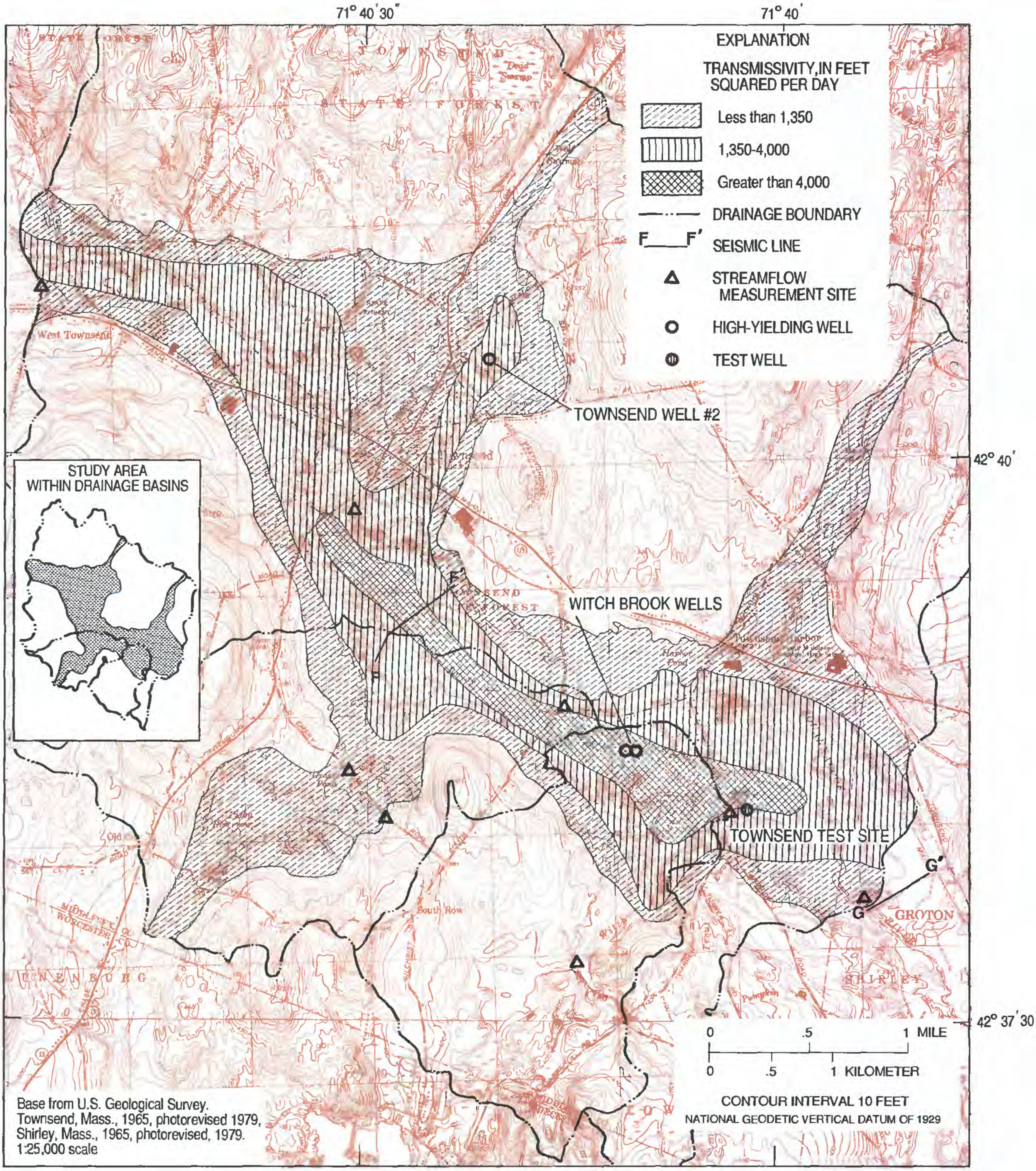


Figure 28.--Areal distribution of transmissivity in the Witch Brook aquifers and location of the water-supply wells, streamflow measurement sites, and seismic-refraction survey. (The index map shows the entire drainage area.) (Modified from Brackley and Hansen, 1977.)

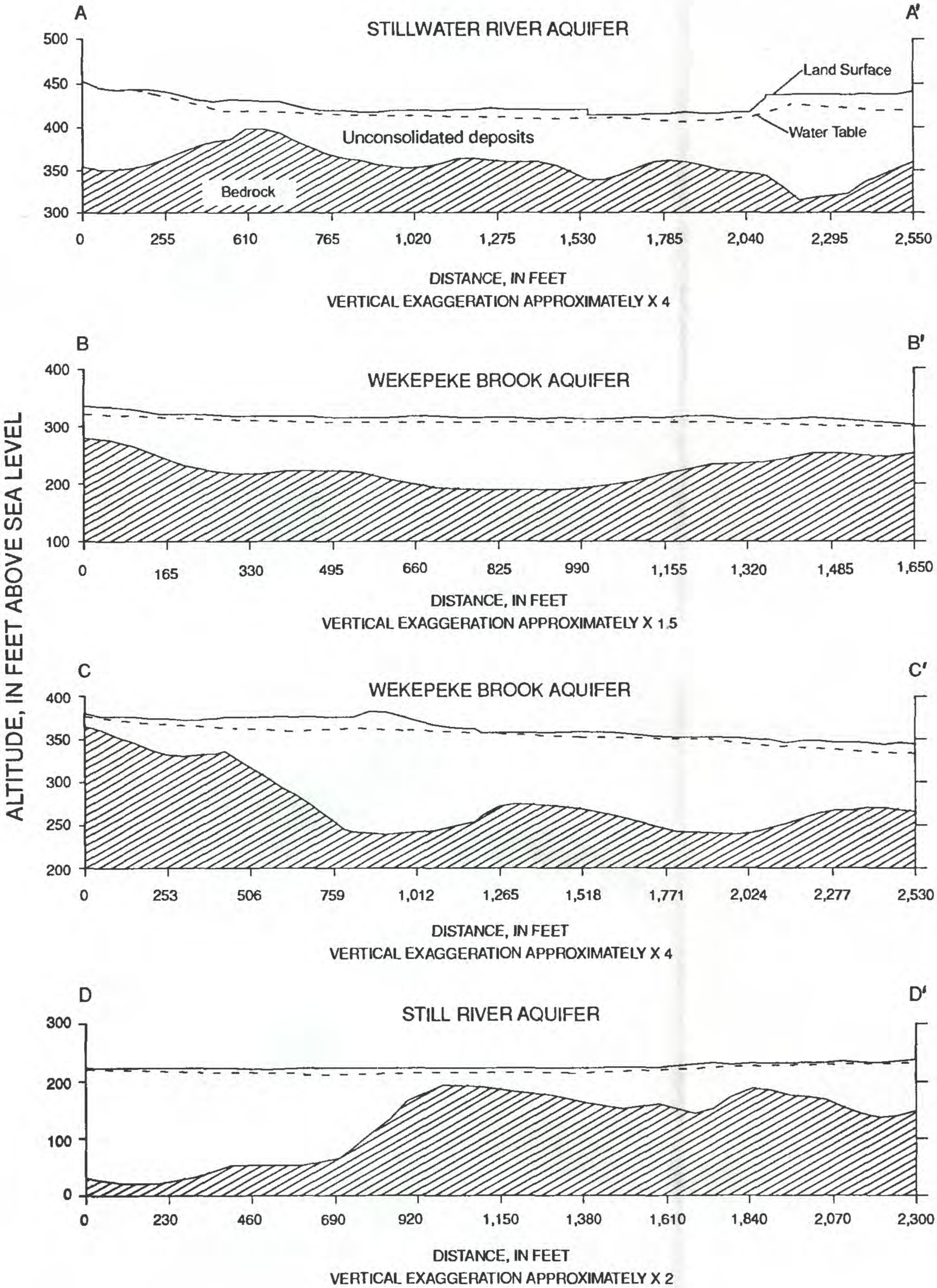


Figure 29.--Geologic sections based on seismic-refraction surveys (traces shown on figures 24, 25, and 26.)

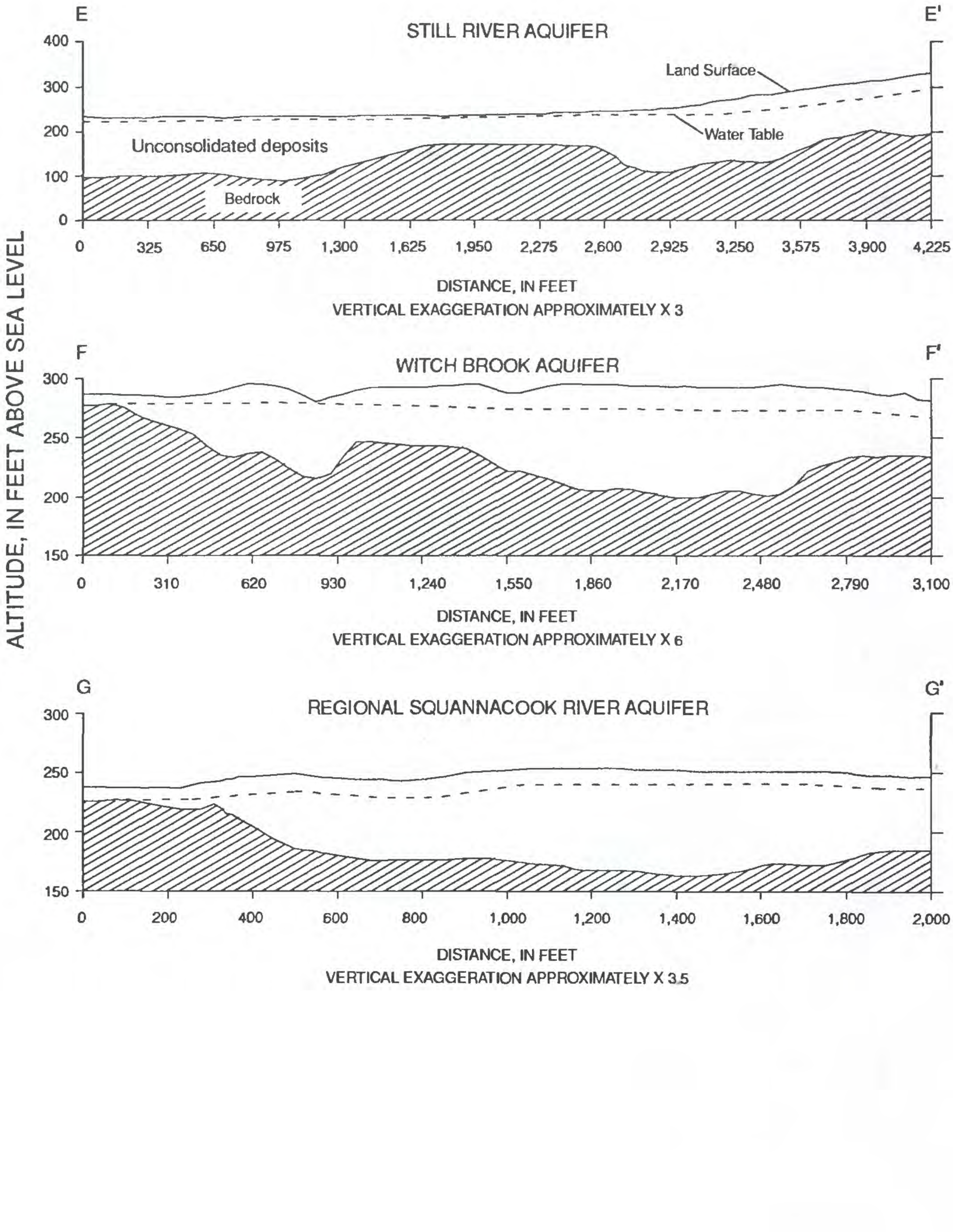


Figure 30.--Geologic sections based on seismic-refraction surveys (traces shown on figures 26 and 28.)

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GLOSSARY

Definitions of terms used in this report were sometimes simplified for clarity; some terms are not universally accepted.

Active model area: That part of the area simulated by a numerical model for which equations describing ground-water flow are solved. In this report, it is the area representing the aquifer.

Aquifer: A permeable geologic material (sand or sandstone, for example) that will yield water in significant quantity to a well or spring.

Aquifer test: A controlled field experiment made to determine the hydraulic properties of water-bearing material. The test involves withdrawing a measured quantity of water from a well and measuring the resulting changes in water level in observation wells surrounding the pumped well.

Base flow: Sustained streamflow that is derived largely from ground-water discharge.

Bedrock: Solid rock, locally called "ledge," that forms the earth's crust. It is locally exposed at the surface as an "outcrop" but more commonly is buried beneath unconsolidated deposits that range in thickness from a few inches to hundreds of feet.

Cone of depression: The area of lowered water level around a pumped well caused by withdrawal of water from the well.

Cubic feet per second (ft^3/s): A unit of flow or discharge. For example, 1 ft^3/s is equal to the flow of a stream 1-ft wide and 1-ft deep flowing at an average velocity of 1 ft/s.

Data: Factual information used as a basis for analysis.

Discharge: The rate of flow of water at a given moment in time. In this report, discharge is expressed in cubic feet per second. See also ground-water discharge and stream discharge.

Drainage basin: The area that gathers water originating as precipitation and contributes it ultimately to a particular stream channel, lake, or ocean.

Drawdown: The lowering of the water level in an aquifer owing to pumping a well.

Drift: Loose rock material transported by a glacier and deposited either directly by ice or by running water emanating from the ice.

Drought yield: Aquifer yield from storage after pumping wells or well fields for 180 days with no recharge while lowering the water level no more than 50 percent of the original aquifer saturated thickness.

Dug well: A shallow, large-diameter well dug in the surficial sediments.

Evapotranspiration: Loss of water to the atmosphere by evaporation from water surfaces and moist soil, and by transpiration from plants.

Flow duration: The percentage of time a given flow in a stream is equaled or exceeded. A flow of 99-percent duration is smaller than a flow of 95-percent duration.

Gage or gaging station: A site on a stream with instruments to measure the changing height of the water surface.

Glacier: A large perennial mass of ice formed by the compaction and recrystallization of snow. A glacier moves slowly due to its own weight. A continental glacier can be as much as 1-mi thick.

Gravel-packed well: A well with gravel surrounding the well screen. Frequently the well screen is 1 to 2 ft in diameter with a 6-in. thick gravel pack. The gravel increases the effective diameter of the well screen and allows water to flow into the well more easily.

Ground water: Water that occurs beneath the land surface. If the water moves to the land surface, it is then called surface water.

Ground-water discharge: Water that is released from the saturated zone of the ground. It includes leakage of water into stream channels, lakes, and oceans; evapotranspiration; and withdrawal from wells.

Ground-water-flow model: As used in this report, a computer program to solve a set of equations which simulate ground-water flow.

Head: Height of water level as measured in a well or stream. The head is a function of location and pressure. Water flows from higher head to lower head.

Head gradient: ratio of difference between heads to the distance between the points at which the head is measured.

Hydraulic conductivity: The capacity of a cube of porous material to transmit water at the prevailing temperature; expressed in a volume per area per day ($ft^3/ft^2/d$ (cubic feet per square foot per day) or ft/d). A material has a hydraulic conductivity of 1 ft/d if, in 1 day, it transmits 1 cubic foot of water through a 1 square foot cross section measured at right angles to the direction of flow, where there is a 1-foot change in water level over a 1-foot flow path.

Hydraulic connection: A stream and aquifer are hydraulically connected if fluctuations in flow or water level in one can affect the flow or water level in the other.

Induced infiltration: Recharge to the ground water from a surface-water body caused by the pumping of a nearby well and the resultant lowering of the ground-water level below the surface-water level.

Infiltrometer: Equipment, designed for this study, which allows permeameter measurements in the stream. It can be used as either a constant-head or a falling head permeameter.

Model: Physical, analytical, or numerical representation of a natural system.

Node: In this report, the center point of a rectangular block of a numerical-simulation model. Often used to refer to the entire block.

Permeable: A material is permeable if it has pores or openings that permit liquids to pass through.

Permeameter: Equipment which measures the head at each end of a sand column in order to determine the hydraulic gradient. Vertical hydraulic conductivity is then determined by timing the flow of a measured volume through the sand column.

Piezometer: An observation well that is open to a small length of aquifer. The water level in the well would be the head at that depth in the aquifer.

Recharge: Water, usually derived from precipitation, that is added to the ground water in the saturated zone.

Saturated thickness: Thickness of the saturated portion of an aquifer. In unconfined aquifers, the difference in altitude between the water table and the bedrock surface or the shallowest underlying zone of relatively low hydraulic conductivity.

Saturated zone: A subsurface zone in which all open spaces are filled with water. The water table is the upper limit of this zone.

Seismic refraction: A geophysical method often useful for determining the depth to the water table or to bedrock. A seismograph is used to determine the time it takes sound energy created by a small explosion to reach a series of sensors. Because sound travels at different velocities in different materials and is refracted (bent) at the boundary between these materials, depths to different types of material can be determined.

Stage: The water-surface elevation of a stream.

Storage: Water filling pore space in aquifer material. Storage is usually depleted during dry seasons and replenished during wet seasons.

Stratified drift: A sorted and layered sediment deposited by meltwater from a glacier; may include separate layers of sand, gravel, silt, and clay.

Stream-aquifer system: An aquifer and a stream that are hydraulically connected.

Streambed conductivity: The vertical hydraulic conductivity of the streambed.

Stream discharge: Flow of water in a stream measured in cubic feet per second.

Surface runoff: Water that moves over the land surface directly to streams or lakes. Surface runoff usually occurs shortly after rainfall or snowmelt.

Surface water: Water when it is on the surface of the land in lakes and rivers. If it seeps into the ground, it is called ground water.

Surficial sediments (deposits): Unconsolidated deposits lying on top of bedrock.

Till: An unsorted, unstratified sediment deposited directly by a glacier. Till may be composed of boulders, gravel, sand, silt, and clay.

Transmissivity: The product of the hydraulic conductivity and the saturated thickness. It is the rate at which water is transmitted through a section of aquifer 1-ft wide where there is a 1-ft change in water level over a 1-ft flow path.

Transpiration: The release of water vapor to the atmosphere by plants.

Unconsolidated: Loose, not firmly cemented or interlocked; for example, sand in contrast to sandstone.

Underflow: Water flowing through an aquifer beneath a stream; it does not discharge to the stream.

Water table: The upper surface of the saturated zone. The altitude of the water table is indicated by the altitude of the water level in an observation well which penetrates the material just far enough to hold standing water.

Well field: A group of small-diameter (usually 2.5 inch) wells connected to a single pump.

Well screen: Slotted section of a well, usually at the bottom, through which water can enter the well.

Withdrawal: Volume of water pumped from a well.

Yield duration: Analogous to flow duration. The percentage of time a given yield is equaled or exceeded. A yield of 99-percent duration is less than a yield of 95-percent duration. A 95-percent yield duration is the quantity of water that can be withdrawn from an aquifer 95-percent of the time; 5 percent of the time this yield would not be available.