U.S. Department of the Interior
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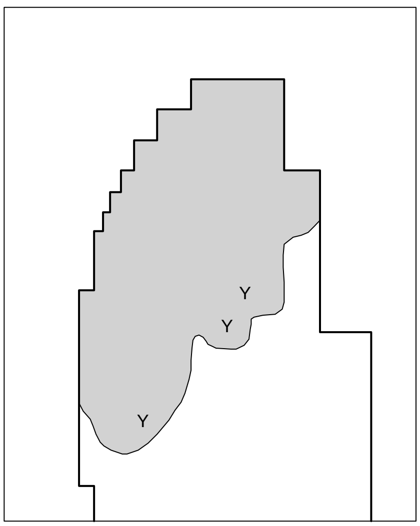
Delineation and Analysis of Uncertainty of Contributing Areas to Wells at the Southbury Training School, Southbury, Connecticut

Water-Resources Investigations Report 00-4158



Prepared in cooperation with the
Connecticut Department of Environmental Protection and the
Connecticut Department of Mental Retardation

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By J. Jeffrey Starn, Janet Radway Stone, and John R. Mullaney

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East Hartford, Connecticut
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Illustrations by: Stewart Crone
Manuscript preparation and editing by: B.A. Korzendorfer

U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, Secretary

U.S. GEOLOGICAL SURVEY
Charles G. Groat, Director

For additional information write to:

District Chief
U.S. Geological Survey
101 Pitkin Street
East Hartford, CT 06108

Copies of this report can be purchased from:

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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATIONS

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
acres	4,047	square meter
square foot (ft^2)	0.09290	square meter
square mile (mi^2)	2.590	square kilometer
	Volume	
million gallons (Mgal)	3,785	cubic meter
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per second per square mile ($ft^3/s/mi^2$)	0.01093	cubic meter per second per square kilometer
gallon per minute (gal/min)	0.06309	liter per second
gallon per day (gal/d)	0.003785	cubic meter per day
inch per year (in/yr)	25.4	millimeter per year
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day
	Transmissivity	
foot squared per day (ft^2/d)	0.09290	meter squared per day

Vertical datum: 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 Sea Level Datum of 1929.

Altitude, as used in this report, refers to distance above or below sea level.

Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness $[(ft^3/d)/ft^2]ft$. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Delineation and Analysis of Uncertainty of Contributing Areas to Wells at the Southbury Training School, Southbury, Connecticut

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ABSTRACT

Contributing areas to public-supply wells at the Southbury Training School in Southbury, Connecticut, were mapped by simulating ground-water flow in stratified glacial deposits in the lower Transylvania Brook watershed. The simulation used nonlinear regression methods and informational statistics to estimate parameters of a ground-water flow model using drawdown data from an aquifer test. The goodness of fit of the model and the uncertainty associated with model predictions were statistically measured.

A watershed-scale model, depicting large-scale ground-water flow in the Transylvania Brook watershed, was used to estimate the distribution of ground-water recharge. Estimates of recharge from 10 small basins in the watershed differed on the basis of the drainage characteristics of each basin. Small basins having well-defined stream channels contributed less ground-water recharge than basins having no defined channels because potential ground-water recharge was carried away in the stream channel.

Estimates of ground-water recharge were used in an aquifer-scale parameter-estimation model. Seven variations of the ground-water-flow system were posed, each representing the ground-water-flow system in slightly different but realistic ways. The model that most closely reproduced measured hydraulic heads and flows with realistic parameter values was selected as the most representative of the ground-water-flow system and was used to delineate boundaries of the contributing areas. The model fit revealed no systematic model error, which indicates that the model is likely to represent the major characteristics of the actual system.

Parameter values estimated during the simulation are as follows: horizontal hydraulic conductivity of coarse-grained deposits, 154 feet per day; vertical hydraulic conductivity of coarse-grained deposits, 0.83 feet per day; horizontal hydraulic conductivity of fine-grained deposits, 29 feet per day; specific yield, 0.007; specific storage, 1.6E-05. Average annual recharge was estimated using the watershed-scale model with no parameter estimation and was determined to be 24 inches per year in the valley areas and 9 inches per year in the upland areas.

The parameter estimates produced in the model are similar to expected values, with two exceptions. The estimated specific yield of the stratified glacial deposits is lower than expected, which could be caused by the layered nature of the deposits. The recharge estimate produced by the model was also lower—about 32 percent of the average annual rate. This could be caused by the timing of the aquifer test with respect to the annual cycle of ground-water recharge, and by some of the expected recharge going to parts of the flow system that were not simulated. The data used in the calibration were collected during an aquifer test from October 30 to November 4, 1996. The model fit was very good, as indicated by the correlation coefficient (0.999) between the weighted simulated values and weighted observed values. The model also reproduced the general rise in ground-water levels caused by ground-water recharge and the cyclic fluctuations caused by pumping prior to the aquifer test.

Contributing areas were delineated using a particle-tracking procedure. Hypothetical particles of water were introduced at each model cell in the top layer and were tracked to determine whether or not they reached the pumped well. A deterministic contributing area was calculated using the calibrated model, and a probabilistic contributing area was calculated using a Monte Carlo approach along with the calibrated model.

The Monte Carlo simulation was done, using the parameter variance/covariance matrix generated by the regression model, to estimate probabilities associated with the contributing area to the wells. The probabilities arise from uncertainty in the estimated parameter values, which in turn arise from the adequacy of the data available to comprehensively describe the ground-water-flow system and the validity of the parameter definitions. The Monte Carlo data sets were conditioned to remove unrealistic parameter sets. Probabilities in the contributing area range from 1 to 100 percent. The highest probabilities (greater than 50 percent) are in the coarse-grained deposits that ring the head of the valley; this area is consistent with the deterministic contributing area defined using the estimated parameter values. The probabilities do not reflect subsurface variabilities within the defined parameter structure, but in this problem, the large-scale variations are expected to dominate contributing area uncertainty.

The contributing area shows that most water enters the stratified glacial deposits through the coarse-grained deposits that ring the head of the lower Transylvania Brook watershed. Some of these deposits are not saturated throughout the year and could not be simulated in the model; however, because the primary public-supply well receives most of its water from this area, the unsaturated deposits should be considered to be within the contributing area. The travel times for ground water in this area are less than 1 year, based on an assumed porosity of 0.20. Travel times for most of the rest of the contributing area are less than 2 years. Contributing areas for alternative models are similar, indicating that nonuniqueness in the design of the model does not seem to be a problem.

INTRODUCTION

The Connecticut Aquifer Protection Program (section 22a-354 of the Connecticut General Statutes) requires water suppliers to delineate aquifer-protection areas for all wells that obtain water from stratified glacial deposits and that provide water to more than 1,000 people. State regulations specify the use of ground-water-flow models to delineate aquifer-protection areas. Ground-water-flow models commonly are calibrated using a trial-and-error method that is subjective, and the models do not provide a measure of the uncertainty of model predictions that is inherent in any type of model.

The Southbury Training School (STS), owned and operated by the Connecticut Department of Mental Retardation (DMR), maintains its own water-supply system from three wells that tap stratified glacial deposits in the Transylvania Brook watershed in central Connecticut (fig. 1). The STS supplies water to approximately 1,862 people; therefore, aquifer-protection areas must be mapped. In 1996, the U.S. Geological Survey (USGS), in cooperation with the Connecticut Department of Environmental Protection (DEP) and the Connecticut DMR, began a study to consider alternatives to traditional ground-water-flow modeling. The study, using the STS as a test site, was undertaken to develop and demonstrate methods that can be used to quantify the uncertainty in model predictions of contributing areas to wells, as well as meet the statutory requirements for delineating contributing areas. Modeling methods were enhanced in this study (1) to provide an objective model calibration that considers the amount of available data and (2) to estimate the uncertainty of model predictions of the contributing areas.

Purpose and scope

This report describes the hydrologic analyses and simulation modeling that were used to delineate contributing areas to public-water supply wells at the STS. The report presents an analysis of the uncertainty in the contributing areas, as determined by simulation models. It also presents geologic information for Transylvania Brook watershed and summarizes previous investigations and data collection in the study area.

Previous investigations

A previous investigation of the aquifer at the STS was done by Mazzaferro (D.L. Mazzaferro, Ground Water Inc., written commun., 1991). This study includes an estimate of the contributing area to the wells, details of water-supply and test-well construction, some geologic data, and estimates of the transmissivity of the aquifer. Details about the operation of the STS well field also are available from the Water-Supply Plan for the Southbury Training School (Corinne Fitting, Connecticut Department of Environmental Protection, oral commun., 1996). Meinzer and Stearns (1929) and Mazzaferro (1986) studied the hydrology of the Pomperaug River watershed, which is

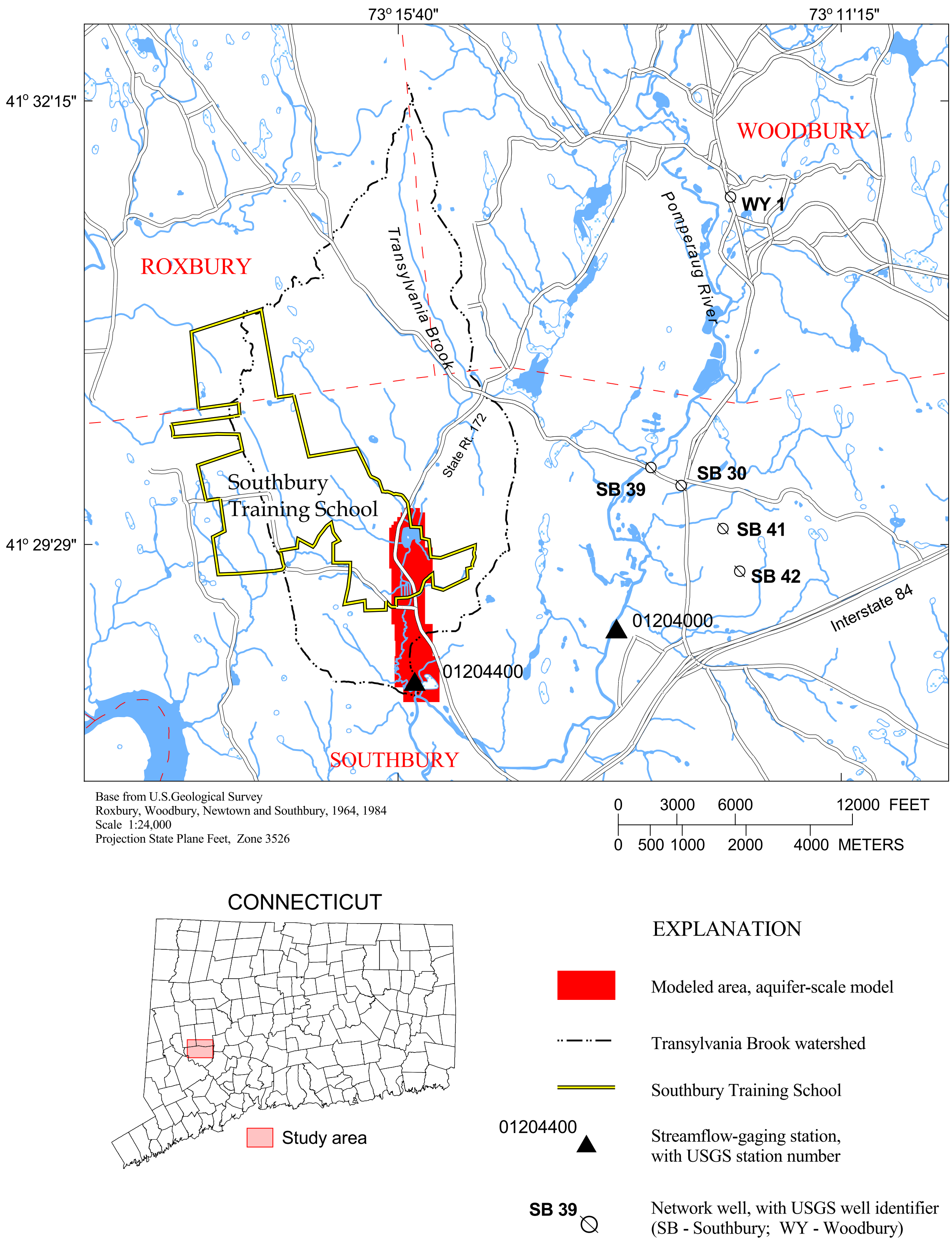


Figure 1. Location of the study area and long-term U.S. Geological Survey data-collection sites, Transylvania Brook watershed, Connecticut.

adjacent to the Transylvania Brook watershed (fig. 1). The climate, topography, and geology of the Pomperaug River watershed are similar to those of the Transylvania Brook watershed, and much of the hydrologic information is transferable, particularly with regard to recharge rates and water budget. Wilson and others (1974) described the hydrology of the lower Housatonic River watershed and presented information on the mean runoff of the Transylvania Brook watershed, the flow frequency duration for the Pomperaug River at Southbury, and the general yield of bedrock wells in the watershed.

Acknowledgments

The authors wish to thank the USGS personnel that contributed substantially to this report: George Casey provided the GIS analysis used in many aspects of the project, Timothy Frick provided the estimates of water use for the study area, Remo Mondazzi logged the boreholes and collected water-level and streamflow measurements throughout the project, and Steven Kiesman surveyed the well locations and instrumented the wells with pressure transducers for the aquifer test. The authors also thank Al Van Geersdaele, STS, for the high level of cooperation received in conducting the aquifer test using the STS water-supply infrastructure.

DATA COLLECTION AND ANALYSIS

Sixteen boreholes were drilled at 11 locations for this study (table 1; figs. 2 and 3) to determine the depth to bedrock, altitude of the water level, and distribution of geologic units. Observation wells were completed in each borehole so that water levels could be monitored during an aquifer test. The observation wells are labelled N (north), E (east), or W (west), for direction from the main pumped well at STS, with a number (1 being closest to the pumped well and increasing in distance from the pumped well). A letter at the end of the well label indicates a deep (D) or shallow (S) well. To supplement geologic information from the boreholes, 189 well-completion reports were obtained from files at the Connecticut DEP, inspected, and plotted on maps. Of these reports, 29 domestic wells were selected for inclusion in the USGS Ground-Water Site

Table 1. Data on observation wells drilled for the aquifer test, Southbury Training School (STS), Connecticut

[—, rock not encountered; all wells shown on fig. 3 except W3, which is shown on fig. 2]

Well number	Distance from well PW-3, in feet	Altitude of land surface, in feet above sea level	Altitude of rock (refusal), in feet above sea level	Altitude of top of screen, in feet above sea level	Altitude of bottom of screen, in feet above sea level	Altitude of water level, in feet above sea level	U.S. Geological Survey local well number[1]
N1S	31.89	173.81	—	160.81	150.81	162.32	SB 101
N1D	34.57	173.77	96.77	114.77	112.77	162.44	SB 100
N2S	110.58	177.42	—	164.42	154.42	164.52	SB 103
N2D	114.13	177.59	90.59	118.59	116.59	162.83	SB 102
N3	260.81	175.23	99.23	130.23	128.23	163.38	SB 104
N5	1,500	175.55	111.55	127.55	125.55	178.22	SB 105
E1S	64.13	169.30	—	163.30	153.30	161.04	SB 107
E1D	60.94	169.33	93.33	110.33	108.33	162.39	SB 106
E2S	135.80	165.48	—	162.48	152.48	159.34	SB 109
E2D	136.50	165.38	85.38	105.38	103.38	162.33	SB 108
E3	470.00	152.10	72.10	87.10	85.10	159.84	SB 110
E4	1,050	215.30	—	121.30	119.30	161.52	SB 111
W1S	61.27	174.50	—	153.50	151.50	162.23	SB 113
W1D	62.52	174.56	112.56	128.56	126.56	162.22	SB 112
W2	227.32	181.52	131.52	151.52	149.52	164.74	SB 114
W3	1,750	185.69	160.69	162.69	160.69	170.02	SB 115

[1]Well number used in the U.S. Geological Survey Ground-Water Site Inventory Database

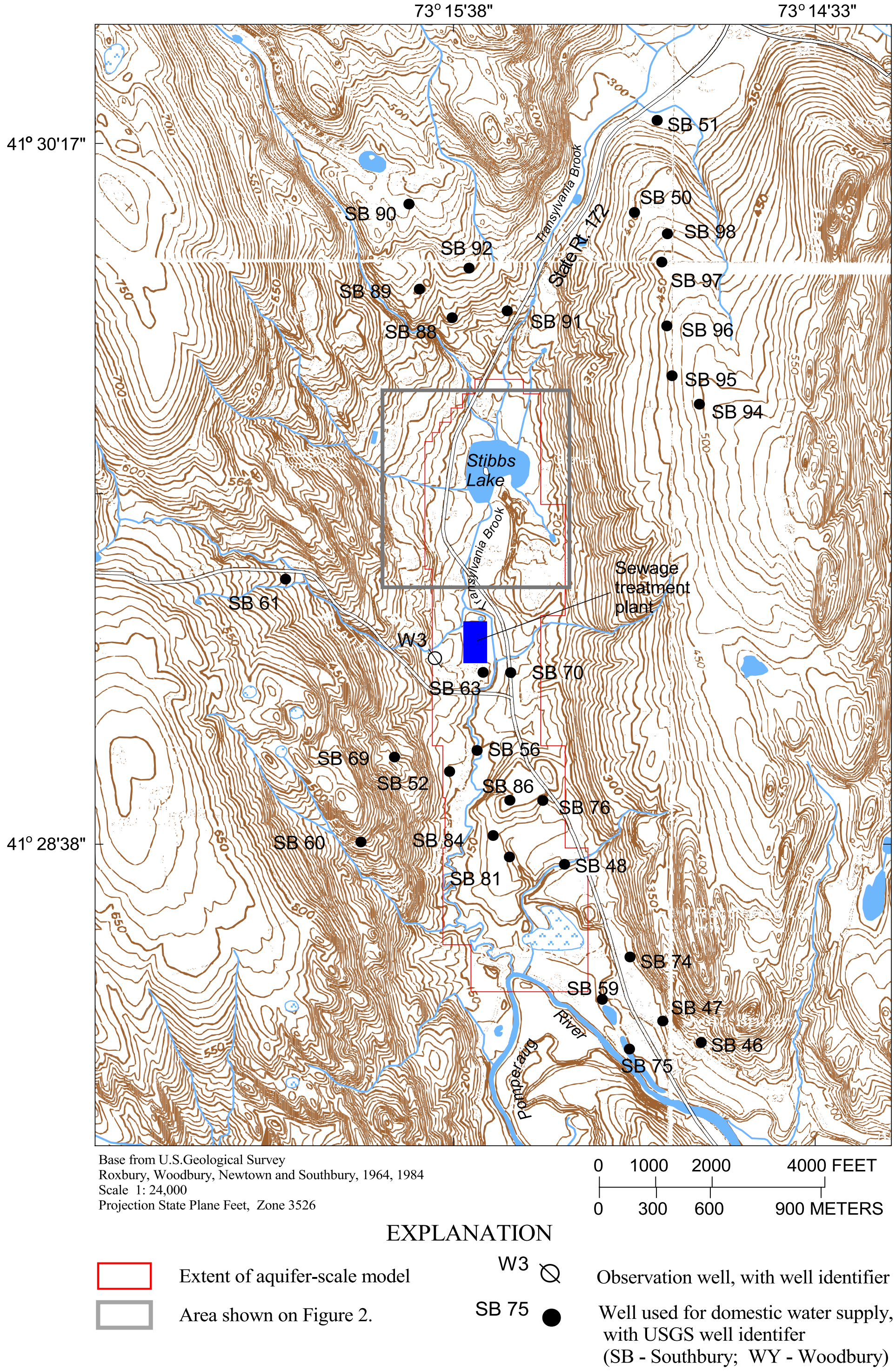
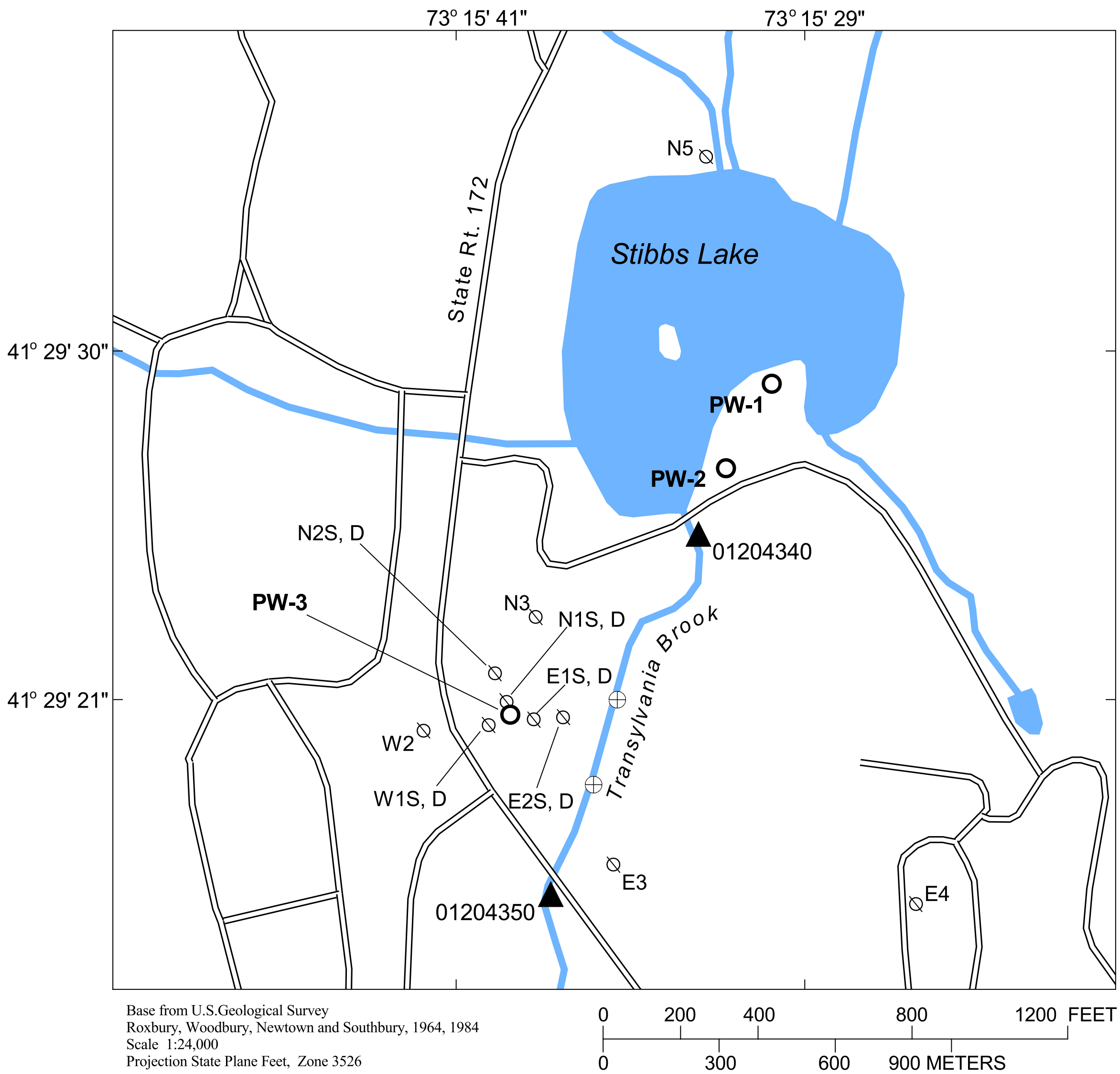


Figure 2. Selected wells, Transylvania Brook watershed, Connecticut.



EXPLANATION

PW-1	Water-supply wells, with identifier
E2D	Observation wells, with identifier (Well W3 is shown on figure 3) First letter is direction from PW-1 (E, east; W, west; N, north). Number is relative distance from PW-1. Last letter is depth (S, shallow; D, deep).
	Streambed piezometers
01204350	Streamflow-gaging station, with USGS station number

Figure 3. Observation wells, water-supply wells, and streamflow-gaging stations, Southbury Training School, Connecticut.

Inventory (GWSI) database because of the quality and uniqueness of the information on the report (appendix 1; fig. 2). Domestic wells are labelled with the GWSI site identification numbers (SB, for the town of Southbury, and a sequential well number for that town).

Streamflow was measured at two sites (fig. 3; table 2). The upstream site represents water flowing from Stibbs Lake into Transylvania Brook and into the area of the well field. The downstream site is just south of the well field and represents water that leaves the well field area. Historical measurements were available from the downstream site, as well as a third site at the mouth of Transylvania Brook (table 2; fig. 1).

Table 2. Instantaneous streamflow at Transylvania Brook, Connecticut

[—, not measured; NA, not applicable]

Date	Streamflow at USGS gaging station (locations shown on figs. 1 and 3)			Loss (-) or gain (+), in cubic feet per second between stations 01204340 and 01204350	Standard deviation[1] of loss or gain, in cubic feet per second
	01204340 (upstream from well field)	01204350 (downstream from well field)	01204400 (at mouth of Transylvania Brook)		
June 10, 1965	—	1.27	3.73	—	NA
July 15, 1965	—	0.29	0.98	—	NA
September 27, 1965	—	.15	.89	—	NA
March 2, 1966	—	10.2	30.	—	NA
August 17, 1966	—	.44	1.09	—	NA
July 28, 1967	—	—	1.38	—	NA
October 14, 1966	—	.97	—	—	NA
September 24, 1968	—	—	1.54	—	NA
June 27, 1969	—	—	3.64	—	NA
June 18, 1970	—	—	4.07	—	NA
July 20, 1971	—	—	2.56	—	NA
September 25, 1972	—	—	1.14	—	NA
August 28, 1973	—	—	1.84	—	NA
August 29, 1996	0.77	0.82	—	+0.05	0.06
September 19, 1996	6.77	6.92	—	+.15	.49
October 17, 1996	2.80	2.84	—	+.04	.20
October 21, 1996	55.5	49	—	-6.5	3.71
October 29, 1996	13.7	14	—	+.03	.98
November 4, 1996	8.73	8.58	—	-.15	.62
April 10, 1997	14.0	12.8	—	-1.2	.95

[1]Standard deviation is calculated as the square root of the sum of the variances of each measurement assuming each measurement to be accurate within 5 percent.

An aquifer test was conducted at well PW-3 from October 30 to November 4, 1996 to determine the hydraulic properties of the aquifer (transmissivity, storativity, and boundary conditions). During the aquifer test, water levels were monitored in all observation wells, and inside and outside two streambed piezometers installed in Transylvania Brook (fig. 3); stage was measured at staff plates at each streamflow-gaging station. Streamflow was calculated from stage measurements based on stage/discharge relations determined at each streamflow-gaging station.

GEOHYDROLOGY OF TRANSYLVANIA BROOK VALLEY

The STS covers 1,600 acres, mostly in the watershed of Transylvania Brook, a tributary of the Pomperaug River. The Transylvania Brook watershed, like the rest of western Connecticut, is underlain by three principal hydrogeologic units—bedrock, glacial till, and stratified glacial deposits (often called “stratified drift”) (figs. 4 and 5). The watershed lies along the western edge of a bedrock structural basin (Gates, 1954, 1959; Scott, 1974; Stanley and Caldwell, 1976; Rodgers, 1985). Stratified glacial deposits at the STS partially fill a bedrock valley in the southern half of the Transylvania Brook watershed.

Geology

The bedrock of the structural basin (fig. 4) is similar in character and structure to that of the Hartford Basin in central Connecticut. Highlands to the east of Transylvania Brook are underlain by Mesozoic-age sedimentary bedrock, which is predominantly red-brown and includes arkosic sandstone, conglomerate, and shale. The highest ridges to the east reach 500 to 650 ft in altitude and are composed of extrusive igneous basalt (also of Mesozoic age). Highlands to the west of Transylvania Brook are underlain by Paleozoic-age crystalline bedrock that is predominantly schist; these highlands reach altitudes greater than 650 ft. The contact between sedimentary bedrock to the east and crystalline (metamorphic) bedrock to the west lies beneath glacial sediments in the Transylvania Brook watershed (fig. 4). Red-brown sandstone was encountered in the bottom of several boreholes (N2D, N3, N5, E1D, and E2D), indicating that the contact between sedimentary and crystalline bedrock lies farther west in the valley than shown on published geologic maps (Stanley and Caldwell, 1976; Rodgers, 1985). This contact may be a fault zone or unconformity, although it is not shown as such on published geologic maps.

Unconsolidated glacial deposits cover bedrock in most places in the Transylvania Brook watershed. These materials were deposited during the advance and retreat of Pleistocene continental glaciers, particularly the last (late-Wisconsinan) glaciation. Surficial materials have been mapped at a regional scale (Stone and others, 1992) and include (1) glacial till, which was laid down directly by ice on top of bedrock and is the surficial material on the valley sides and in the uplands; and (2) stratified glacial deposits, which partially fill the bedrock valley to an altitude of 200 to 250 ft (figs. 5, 6, and 7).

Till consists of a nonsorted, generally nonstratified mixture of grain sizes ranging from clay to large boulders. The till matrix is composed dominantly of sand and silt, although boulders in and on the surface commonly are abundant. Till is typically a dense and compact material due its mode of deposition beneath the great weight of the ice sheet and to the presence of silt and clay in the matrix. This type of till commonly is identified as “hardpan” in well logs recorded by well drillers. A sandier and stonier, less dense and less compact facies of till (ablation till) may be present in some places. The color and lithology of till generally reflect the composition of the underlying bedrock to the north from which the till was derived. In the Transylvania Brook watershed, till is described in well logs as gray, having been derived predominantly from the crystalline bedrock underlying the highlands to the northwest. Red-brown till derived from the red-brown sedimentary rocks underlying the eastern side of Transylvania Brook and the highland to the east is likely to be present locally in the eastern part of the study area; thin red-brown till overlying red sandstone was encountered in boreholes E3 and N5 drilled at the STS during this study. Numerous well logs from the residential subdivision on the hillside northeast of the STS indicate the presence of thick gray till overlying red-brown sedimentary rock. Well logs for domestic wells inventoried indicate that till ranges from less than 3 ft to as much 140 ft in thickness in the study area. Several boreholes drilled at the site showed locations where till is absent and stratified glacial deposits directly overlie bedrock; other boreholes showed the presence of till.

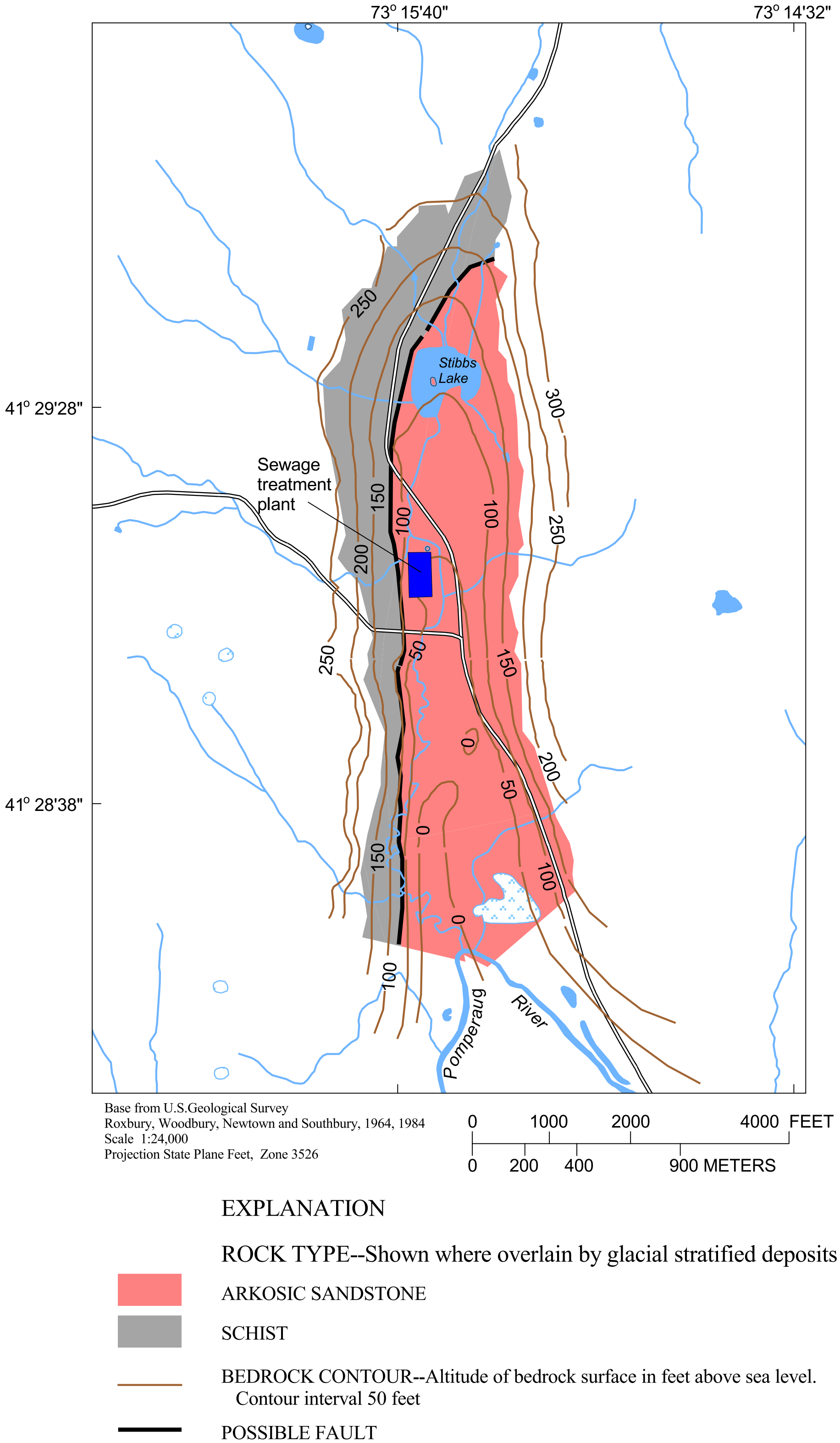


Figure 4. Bedrock geology of the Transylvania Brook valley, Connecticut.

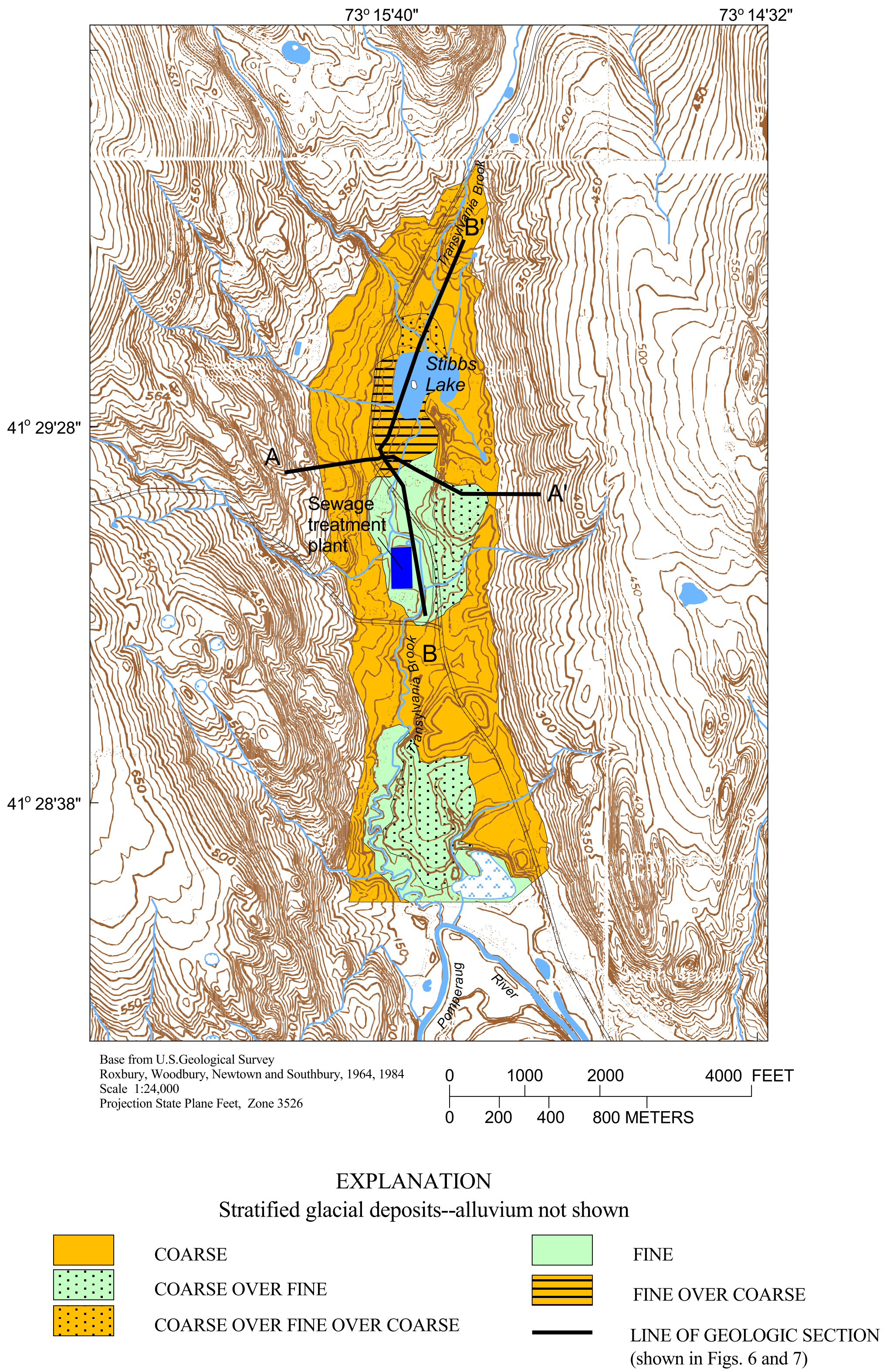


Figure 5. Stratified glacial deposits of the Transylvania Brook valley, Connecticut.

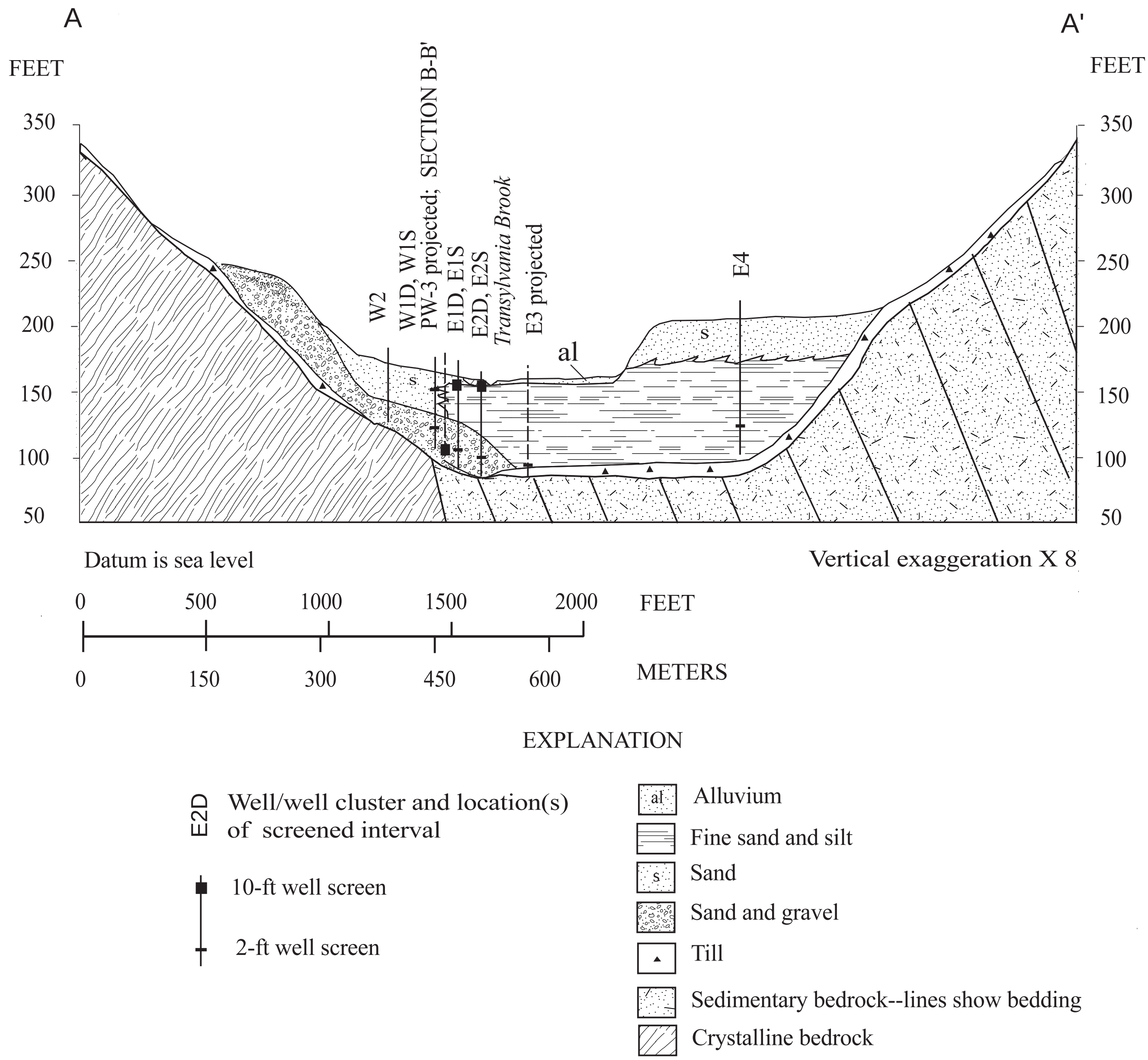


Figure 6. East-to-west geologic section through Transylvania Brook valley, Connecticut.

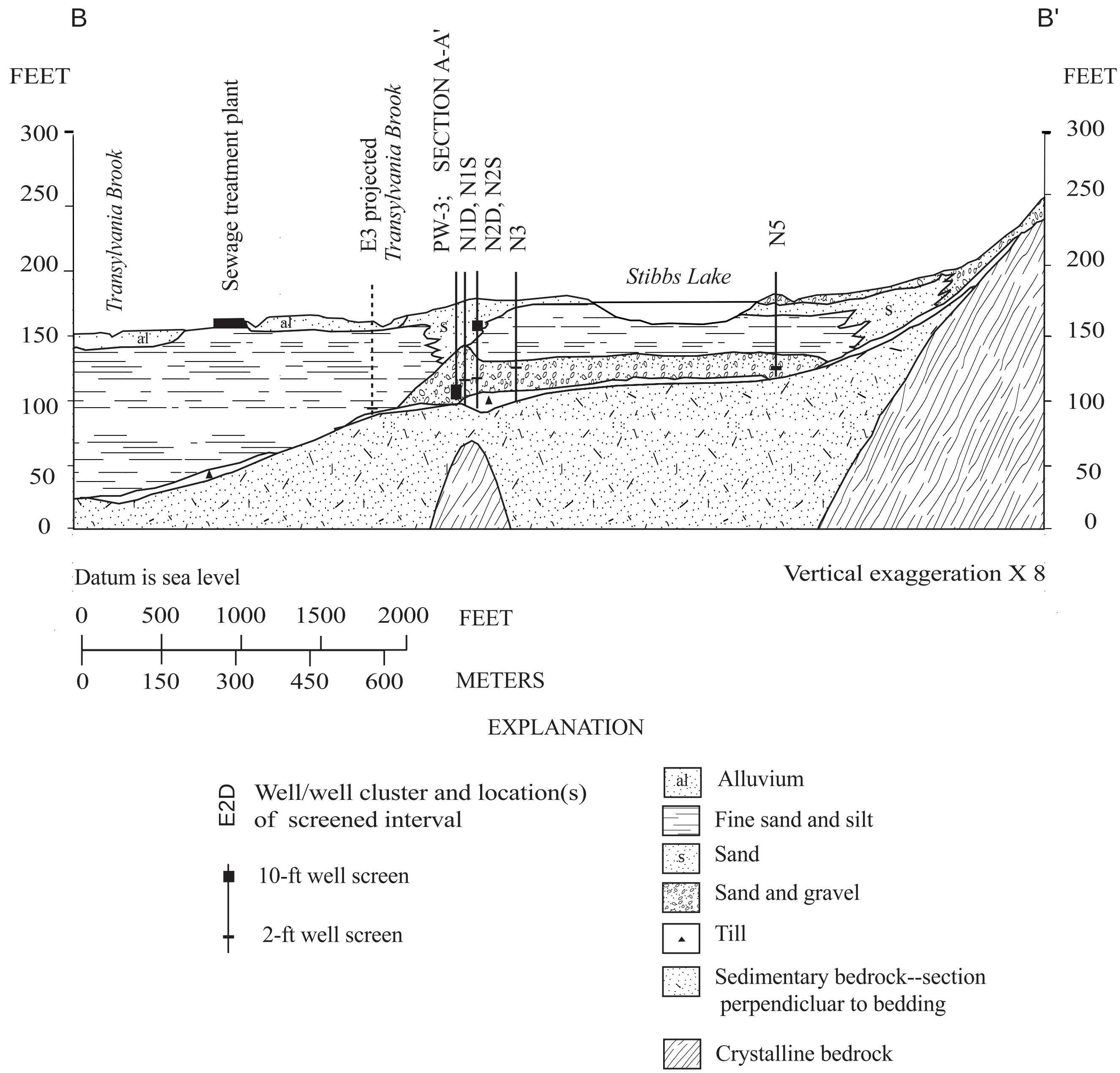


Figure 7. North-to-south geologic section through Transylvania Brook valley, Connecticut.

Stratified glacial deposits consist of layers of well to poorly sorted gravel, sand, silt, and clay laid down by meltwater in glacial lakes and streams that occupied valleys during retreat of the ice sheet. The stratified deposits in the Transylvania Brook watershed near the STS consist of both coarse-grained and fine-grained sediments graded to or deposited in glacial lakes that occupied the Pomperaug and Transylvania Brook valleys in progressively northward positions in front of the retreating ice sheet. Coarse-grained deposits (gravel, sand and gravel, and sand) were laid down as ice-marginal deltas and fluviodeltaic deposits in close proximity to the retreating ice sheet. Fine-grained deposits (very fine sand, silt, and clay) accumulated farther from the ice margin in the quiet water conditions of the glacial lake bottom. Near the STS, coarse-grained deposits, including sand and gravel and sand units, locally make up the entire thickness of the stratified section, particularly west of Rt. 172 and at the north end of the valley. In much of the area east of Rt. 172, coarse-grained deposits underlie, overlie, or interfinger laterally with fine-grained deposits. The main production well (PW-3) at the STS and installed observation wells (W2, W1D, E1D, E2D, N1D, N2D, N3, and N5), are screened in a subsurface sand and gravel unit, which ranges from a few to about 40 ft in thickness. To the east, north, and south of PW-3, the sand and gravel unit is overlain by fine-grained deposits ranging from a few to 60 ft in thickness; to the west of PW-3, the sand and gravel unit is overlain by 30 to 40 ft of sand (fig. 6). The subsurface sand and gravel unit likely extends west of Rt. 172 to connect with coarse-grained deposits on the valley side and north beneath Stibbs Lake at least as far as observation well N5. The unit pinches out to the east and south (figs. 6 and 7). Split-spoon samples of the subsurface coarse-grained deposits indicate that the unit consists of interbedded layers of pebbly, medium to coarse sand, granule to pebble gravel, and coarse to very coarse sand and granules. Large pebbles and cobbles also may be present, although these were not sampled by the split-spoon sampler. The fine-grained deposits consist of thinly laminated very fine sand, silt, and clay. Individual layers are typically 0.1 to 1.0 in. in thickness and alternate between coarser (very fine sand to fine sand) and finer layers (silt and/or clay). Coarser layers are typically thicker than finer layers.

Hydrology

Water in the Transylvania Brook watershed flows from till-covered uplands in stream channels and as subsurface flow over and into the stratified glacial deposits in the valley bottom. Some water from upland stream channels also seeps into the stratified glacial deposits. Most runoff (both ground-water and surface-water) discharges to Transylvania Brook. Some ground-water runoff is intercepted by water-supply wells, both domestic and public; this water is returned after treatment to the aquifer or to Transylvania Brook.

Hydraulic properties of the aquifer system

Three principal types of subsurface materials are found in the watershed—coarse-grained stratified glacial deposits; fine-grained deposits, which consist of till and fine-grained stratified deposits; and bedrock, which consists of sedimentary and crystalline rocks (figs. 4 and 5). The materials are grouped in this way for hydrogeologic purposes, which is somewhat different than the geologic grouping of units, because of the general similarity in their hydraulic properties. Melvin and others (1992) summarized the hydraulic properties of these materials in Connecticut on the basis of published aquifer test and laboratory test results (table 3), and these results are applicable to the materials at the STS site.

Coarse-grained stratified deposits have the highest hydraulic conductivity because of the openness of their interconnected pore spaces, and thus form the most productive part of the ground-water system. Till and fine-grained stratified deposits have a lower hydraulic conductivity than the coarse-grained deposits because fine sand, silt, and clay occupy more of the open pore space within the deposits. Theoretically, the hydraulic conductivity of till, which is nonsorted and nonlayered, should be the same in all directions, but the hydraulic conductivity of the fine-grained deposits, composed of layered material, should be much higher in the horizontal direction (along the relatively coarser layers) than in the vertical direction (through both finer and coarser layers) (table 3).

Table 3. Median values of hydraulic properties of hydrogeologic units in Connecticut

[Modified from Melvin and others, 1992, table 1; —, no data or insufficient data]

Unit	Hydraulic conductivity, in feet per day	Storativity, dimensionless	Porosity, in percent	Orientation
Coarse-grained stratified drift	170	.36	—	—
Fine-grained stratified drift, silt	.14	.29	—	vertical
Fine-grained stratified drift, clay	0.0001	—	—	vertical
	.82	—	—	horizontal
Loose surface till, crystalline-rock provenance	2.7	0.28	—	—
Loose surface till, sedimentary-rock provenance	.71	—	32	horizontal and vertical
Compact surface till	.007	—	—	horizontal
Crystalline bedrock	.6			
Sedimentary bedrock	4.7			

In bedrock, ground water flows primarily through fractures in the rock rather than through the pore spaces in the rock. Water-bearing pathways in fractured bedrock include nearly horizontal stress-relief fractures, layer-parallel fractures in layered crystalline rocks and sedimentary rocks, and high-angle to vertical fractures caused by movements of the earth's crust. Bedrock wells typically penetrate many fractures, but only a few of these fractures may produce water. The passage of water from unconsolidated till and stratified deposits into or out of bedrock can be impeded by a compact till that overlies bedrock in most of Connecticut; this till, shown between stratified glacial deposits and bedrock in figs. 6 and 7, was compacted beneath the glaciers and thus has a lower hydraulic conductivity than loose surface till. The fractured bedrock is the source of water to all domestic wells inventoried in the area. The average yield of inventoried bedrock wells is less than 5 gal/min; however, the possibility of a highly fractured zone in the valley bottom is indicated by reported yields of 50 to 100 gal/min from several domestic wells in the area.

D.L. Mazzaferro (Ground Water, Inc. written commun., 1991) estimated the average transmissivity of the coarse-grained deposits at the STS well field to be 8,200 ft^2/d, using the specific capacity of the three production wells and grain-size analysis of two test borings. Mazzaferro estimated transmissivity at the production wells from 6,400 ft^2/d to 12,100 ft^2/d and at the test borings from 6,400 to 9,500 ft^2/d. These estimates, however, which were based on specific capacity and grain-size analysis, are considered to be less accurate than estimates based on aquifer-test analysis.

An aquifer test was conducted by the USGS from October 30 to November 4, 1996 at STS well PW-3. During the test, well PW-1 was not pumping. Transmissivity and storativity were calculated using the semi-logarithmic method described by Cooper and Jacob (1946) (table 4). The analyses of the aquifer-test data are on file at the Connecticut District office of the USGS. There were slight changes in the slope of the semi-logarithmic plot of drawdown with time that indicated an additional source of water to the aquifer, such as leakage from a surface-water body, recharge from precipitation, or leakage from fine-grained deposits. The drawdown curves were analyzed both at early and late times in the test.

Table 4. Summary of transmissivity and storage from an aquifer test at well PW-3, Southbury Training School, Connecticut

[Well locations shown on fig. 3. S, shallow water-table well; D, deep well]

Well	Distance from pumped well, in feet	Transmissivity, in feet squared per day early/late time	Storage coefficient early/late time	Drawdown at 1,000 minutes into test, in feet
N1S (early)	32	3,100	.007	5.3
N1S (late)		3,600	.004	
N2D (early)	114	3,400	.0004	4.99
N2D (late)		5,000	.00015	
N3 (early)	261	3,400	.0003	3.69
N3 (late)		6,500	.00008	
E1D (early)	61	3,300	.0006	6.01
E1D (late)		4,660	.00006	
E2D (early)	136.5	3,100	.0005	4.98
E2D (late)		5,440	.00004	
W1D (early)	63	3,400	.0006	5.88
W1D (late)		4,660	.0003	
W1S (early)	61.3	2,883	.005	5.03
W1S (late)		3,436	.003	

Average ground-water recharge and relation to model calibration period

To perform a transient ground-water simulation, as required by the aquifer protection mapping process, it is necessary to understand (1) the long-term average steady-state ground-water recharge in the basin of interest and (2) the relation between the simulated time period and average ground-water recharge. The relation of the study period to long-term average conditions will be discussed using long-term streamflow and ground-water level data from the adjacent Pomperaug River watershed. Limited data from the STS site also will be used to assess the hydrologic conditions at the time of the aquifer test.

Effective ground-water recharge is defined in this study as the amount of water that infiltrates from the land surface into the aquifer minus evapotranspiration from the aquifer (fig. 8). Average annual recharge was estimated using a regression-derived linear relation between ground-water outflow and the percentage of the drainage area underlain by coarse-grained deposits for Connecticut (Mazzaferro and others, 1979). In Mazzaferro's study, ground-water outflow was determined by hydrograph separation. Ground-water outflow is assumed to be a conservative estimate of recharge if changes in ground-water storage are small. The relation is

$$Y = 35 + 0.6X , \tag{1}$$

where:

Y is ground-water outflow as a percentage of total runoff, and

X is the percentage of the watershed underlain by coarse-grained stratified glacial deposits.

Stratified glacial deposits occupy 17 percent of the land surface in the Transylvania Brook watershed, so that solution of equation 1 indicates that basin-wide recharge is 45.2 percent of total annual runoff. Mean annual runoff for the Transylvania Brook area is 1.71 $ft^3/s/mi^2$ (Wilson and others, 1974); therefore, ground-water outflow is calculated to be 3.89 ft^3/s, or 10.5 in/yr over the 5.03-mi^2 area of the drainage basin.

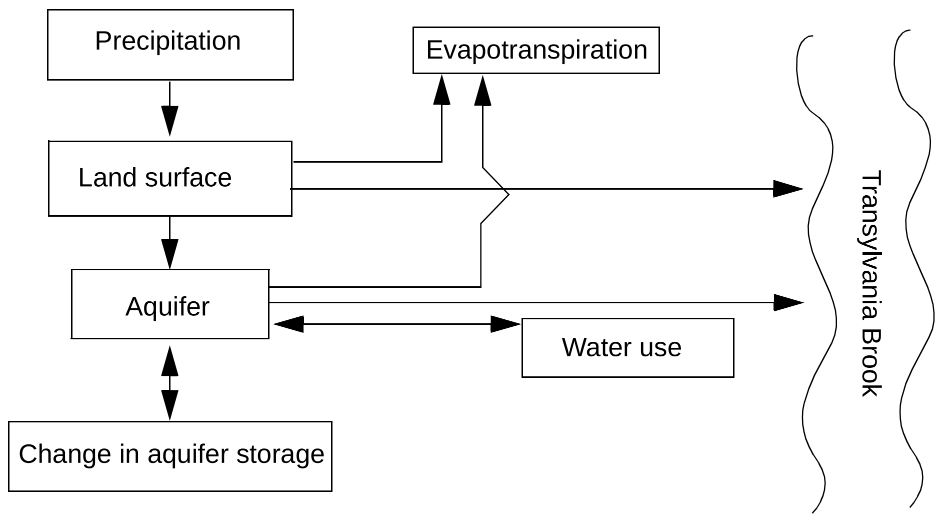


Figure 8. Generalized water budget.

Mazzaferro and others (1979) observed that areas underlain by stratified glacial deposits have a recharge rate that is 2.7 times higher, on average, than areas underlain by till. Using this relation and the percentages of the basin underlain by till and stratified deposits, recharge to stratified glacial deposits is calculated to be about 22 in. (1.41 ft^3/s) and recharge to till is about 8 in. (2.48 ft^3/s). Although the rate of recharge to till is lower, the volume of water recharged is greater because till covers a larger area of the basin than do stratified deposits.

Ground-water recharge occurs in annual cycles, and in order to simulate values of recharge in the model, it is necessary to understand how the time period of the aquifer test relates to the recharge cycle. Also, because steady-state hydrologic conditions are required as starting conditions for the transient model, it is necessary to assess whether the conditions at the time of the aquifer test approximate steady-state conditions, and if not, to determine what type of adjustment could be made to the data to make them comparable to steady-state values. Long-term data (more than 1 year) are needed to analyze the yearly cycle of recharge. No long-term data-collection sites are present in the Transylvania Brook watershed, but five USGS network wells and a streamflow-gaging station are in the adjacent Pomperaug River watershed (fig. 1).

The Pomperaug River watershed has virtually the same surficial geology, climate, and topography as the Transylvania Brook watershed, so the hydrologic response of each basin to precipitation is expected to be similar. The main difference between the watersheds is size—the Pomperaug River watershed above the streamflow-gaging station is about 75 mi^2 and the simulated part of the Transylvania Brook watershed is about 5 mi^2. Long-term records are available for several streams of similar size to Transylvania Brook, but at more distant locations in Connecticut. Peaks on the streamflow hydrographs of these streams correspond closely with peaks on the hydrograph of the Pomperaug River; therefore the Pomperaug River is deemed to be an adequate surrogate for Transylvania Brook.

Ground-water flow sustains streamflow between rains, and ground-water recharge can be estimated by analyzing streamflow records. Streamflow records can be used to calculate ground-water recharge using three computer programs—PART, RORA, and PULSE (Rutledge, 1997; 1998). These programs separate streamflow into ground-water and surface-water components and can be used to estimate ground-water discharge from a basin. PART is based on an empirical analysis of the streamflow record. RORA and PULSE are based on the drainage characteristics of a basin and require an estimate of the master recession curve (Rutledge, 1997; 1998). The master recession curve is fit to the recession segments of peaks; the difference in streamflow between adjacent recession curves is ground-water recharge. In this study, the results of all three programs are similar (fig. 9).

Streamflow and recharge data for the Pomperaug River from 1995-96 generally display the typical hydrologic cycle (fig. 9). Streamflow and recharge are high at the beginning of winter, because precipitation is abundant and plants are not using large amounts of water. As temperatures rise and plants begin to use water, streamflow and recharge decline and reach lowest levels in summer. After the first killing frost in fall, streamflow and recharge increase (Melvin, 1986). The decline in streamflow and recharge in spring 1995 began earlier than normal because, as of June 1995, rainfall was 4.47 in. below normal for the year. As expected, less ground-water flowed to the stream in June to September during both years than in other months of the year, although streamflow and recharge were lower in summer 1995 than in summer 1996. Beginning in September 1995, a general rise in streamflow indicated increased ground-water discharge during this time period. Streamflow and recharge generally remained at this level until the summer of 1996, when the yearly decline began. The aquifer test (October 30 to November 4, 1996) was conducted when the calculated ground-water component of streamflow was midway between the summer low and the rate at the end of the year.

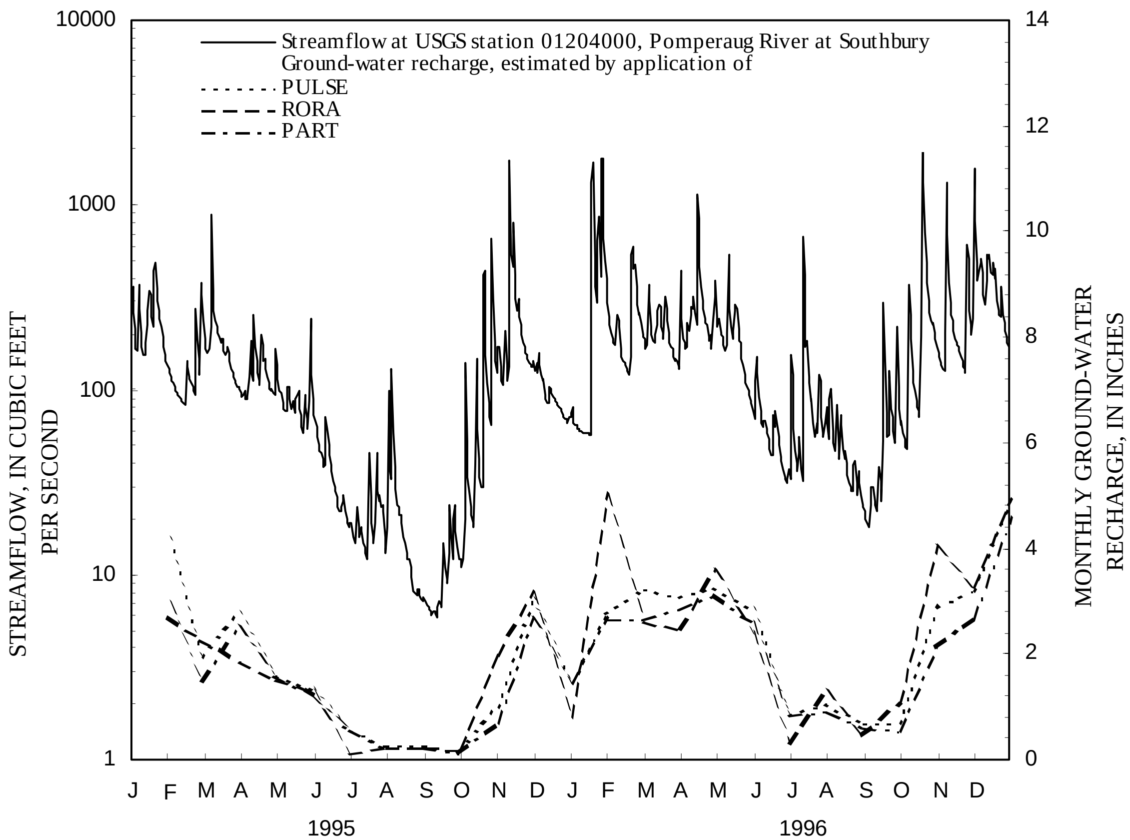


Figure 9. Streamflow and ground-water recharge for the Pomperaug River, 1995-96.

On October 20, 1996 (10 days prior to the aquifer test), 4.71 in. of precipitation fell in the Pomperaug River Basin (fig. 10). Stream stage in the Pomperaug River peaked on October 20 and subsequently declined throughout the aquifer test. An analysis of streamflow using the program PULSE shows that the 2 months preceding the aquifer test was a period of generally increasing ground-water contribution to streamflow (fig. 10). From the lowest streamflow in September 1996 until the aquifer test, there were 5.1 in. of recharge, according to the program PULSE. Most of this recharge (3.1 in.) took place during the large rainfall 10 days before the aquifer test.

Ground-water recharge calculations presented in the previous paragraphs are corroborated by ground-water levels in USGS network wells (fig. 11). Water levels are measured by the USGS at four wells in the Pomperaug River watershed (biweekly October 1991 to October 1996; monthly after October 1996) (figs. 1 and 11; table 5). The wells are at various positions across the river valley—the valley bottom (SB 39), a stream terrace in stratified glacial deposits (SB 30), the side of a thick till deposit (SB 41), and on top of a thick till deposit (SB 42). The valley bottom well has a generally constant water level that is controlled by the stage of the Pomperaug River. Water levels in the other wells show a dampened fluctuation similar to the streamflow hydrograph. Water levels generally declined throughout 1995 until October when recharge increased. Rises in ground-water levels in the fall of 1995 appear to lag behind calculated recharge events (fig. 11). By November, recharge was sufficient to maintain water levels (that is, recharge equaled ground-water outflow), so only small changes in water level in the aquifer took place until the summer of 1996, when ground-water levels began to decline. A period of rising water levels, indicating that recharge exceeded ground-water outflow, began in September 1996.

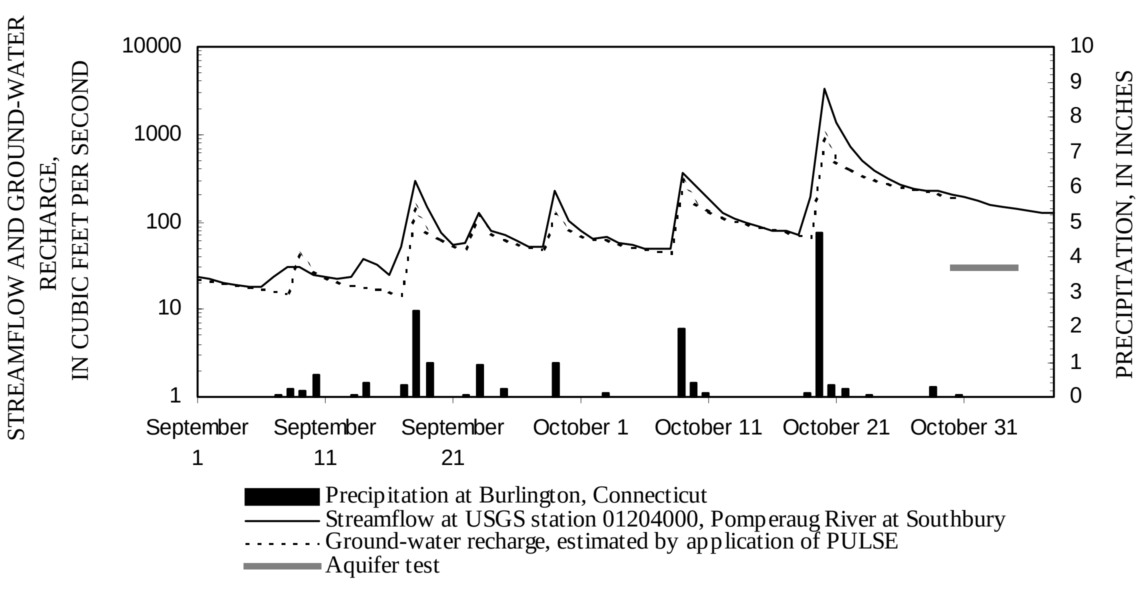


Figure 10. Streamflow, precipitation, and ground-water recharge for the Pomperaug River, September to October 1996.

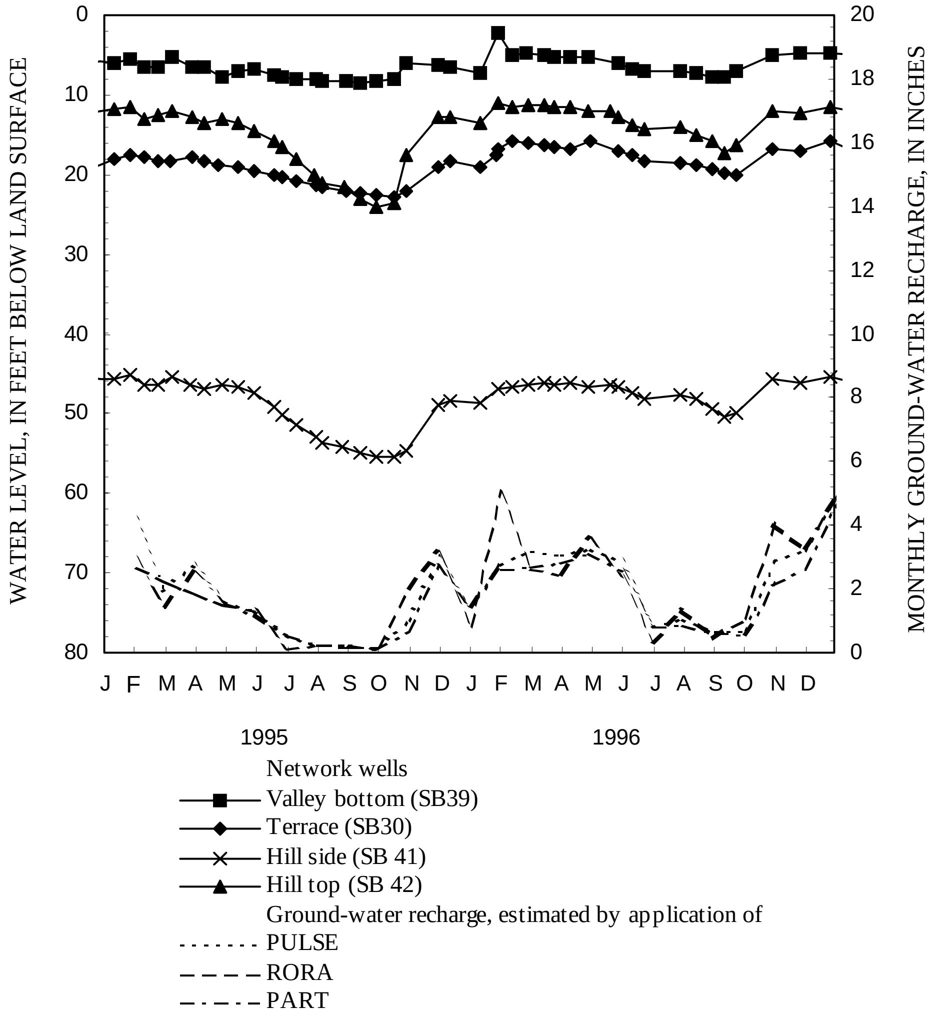


Figure 11. Water levels in four network wells and ground-water recharge in the Pomperaug River watershed, Connecticut, 1995-96.

Table 5. Ground-water levels in network wells in the Pomperaug River watershed, Connecticut

Well	Topographic position	Period of record	Depth to water, in feet below land surface			
			Median	September 23, 1996	October 30, 1996	November 27, 1996
SB 39	Valley bottom	10/24/91-10/29/98	6.62	7.08	4.96	4.83
SB 30	Terrace	1/2/79-10/29/98	18.97	20.05	16.9	16.93
SB 41	Hill side	10/24/91-10/29/98	47.29	49.98	45.65	46.05
SB 42	Hill top	8/19/93-10/29/98	13.60	16.30	12.1	12.3

Ground-water levels in the four USGS network wells generally were close to median annual levels during the fall of 1996 (table 5). Water levels were below the median annual levels prior to the aquifer test on September 23, 1996 and rose from September 23 to October 30, probably in response to the precipitation on October 20. Water levels on November 27 were similar to the previous measurement, indicating that recharge was about equal to or greater than discharge. The frequency of water-level measurements (biweekly) was not sufficient to show when water levels reached their annual high level.

In addition to the long-term data available for the Pomperaug River, some limited data are available from the STS site that can be used to assess hydrologic conditions at the time of the aquifer test. Stream stage in Transylvania Brook declined throughout the aquifer test. In some situations, the amount of ground-water flow to or from the stream (referred to as streamflow gains and losses, respectively) can be calculated by subtracting the downstream flow from the upstream flow; in this situation, however, the difference between the streamflows was less than the standard deviation of the gain or loss (table 2). Water levels in two piezometers driven about 2 ft into the streambed near the pumped well were compared to water levels in the stream to see if water was leaking into or out of the stream. Throughout the aquifer test, water levels in the aquifer below the streambed were higher than water levels in the stream, indicating that the stream was gaining water throughout the test.This relation persisted as the stream stage declined throughout the test.

Water levels were recorded for several days prior to the aquifer test in observation wells used for the test; one water-level hydrograph is shown in figure 12. Large fluctuations caused by the cycling on and off of the pump in well PW-3 may obscure, to some extent, the natural fluctuations that would take place because of recharge; however, the high and low extremes in each pumping cycle increased before and after the aquifer test, indicating that ground-water recharge may have been taking place, as in the USGS network wells (table 5).

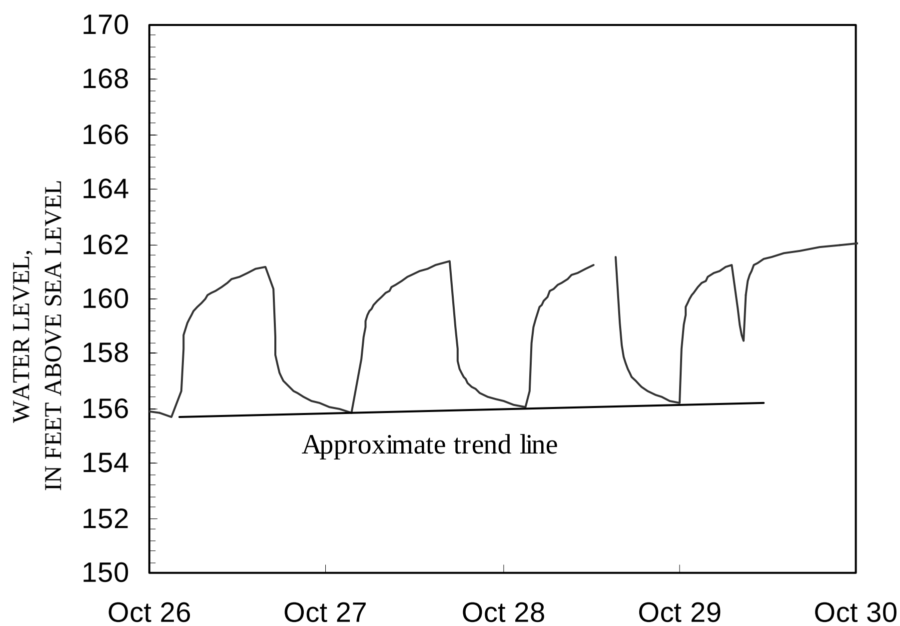


Figure 12. Water levels in observation well E2D, Southbury Training School, Connecticut.

The hydrologic conditions during the aquifer test were complex. Ground water and streamflow relations during the aquifer test were typical of fall, but this relation was complicated by the 4.71 in. of precipitation on October 20. On the basis of existing data, however, it is reasonable to state that ground-water recharge was taking place, but at a rate below normal for fall and winter. To use water levels and streamflow as observations in model calibration, both sets of observations must reflect the same set of hydrologic conditions; in this case, however, ground-water levels were probably rising because of precipitation 10 days before the aquifer test whereas stream stage and streamflow were declining. Model calibration using these data should include the recharge on October 20, or the data should be adjusted before simulation to account for the recharge. Although the stream stage was declining throughout the aquifer test, ground-water levels may still have been rising because recharge is delayed by the time of travel through the unsaturated upper part of the aquifer system and from upland areas.

Water use

The use of water in the watershed falls into two categories—domestic water use, consisting of water withdrawn from small, privately owned wells completed in bedrock, and institutional water use, consisting of water withdrawn from wells completed in stratified glacial deposits. Residents of the STS are served by three wells identified as PW-1, PW-2, and PW-3 (table 6; fig. 2). Well PW-3 is the main production well and is in daily use at a rate of 0.338 Mgal/d (Al Van Geersdaele, Southbury Training School, oral commun., 1996). Well PW-1 is a standby well that automatically turns on and off during the day at a rate of 0.180 Mgal/d as demand dictates. Well PW-2 is classified as an emergency source of supply and is unused because of high concentrations of iron and manganese. Wells PW-1 and PW-2 were drilled in 1938 with reported yields at that time of 0.338 Mgal/d and 0.72 Mgal/d, respectively. By 1967, the yields of these wells had declined to 0.144 Mgal/d for well PW-1 and 0.41 Mgal/d for well PW-2. Well PW-3 was drilled in 1970 and had a reported yield of 0.36 Mgal/d. Water-use records were obtained from STS for January 1989 to June 1996. The pumping rate during this period averaged 0.234 Mgal/d and ranged from 0.231 to 0.263 Mgal/d, with slightly higher rates common during summer months. There was no discernible trend in water use (T.W. Frick, U.S.Geological Survey, written commun., 1996). Wastewater from the STS is collected at the sewage-treatment plant (fig. 2), treated, and discharged to Transylvania Brook downstream from USGS streamflow-gaging station 01204350.

A water-use survey was conducted by the USGS to determine the magnitude of domestic ground-water withdrawals in the Transylvania Brook watershed and what effect, if any, the withdrawals might have on ground-water flow. Annual domestic water use was estimated to be 1.1 Mgal during June to August 1995 and 3.0 Mgal during September 1995 to May 1996 (T.W. Frick, U.S.Geological Survey, written commun., 1996). Domestic water is withdrawn from bedrock and returned through septic systems to the stratified deposits. About 15 percent of the domestic water is estimated to be lost through evapotranspiration. The net ground-water withdrawal from the aquifer system is about 0.0017 Mgal/d. This rate is small relative to pumping at the STS (0.338 Mgal/d) and is distributed over a large area; therefore, domestic ground-water withdrawals were not simulated in models in this study.

Table 6. Construction details of water-supply wells, Southbury Training School, Connecticut

[Data from D.L. Mazzaferro, Ground Water Inc., written commun., 1991; —, information not available]

Well	Date drilled	Depth, in feet	Diameter, in inches	Water level, in feet below land surface (reported by driller)	Screen length, in feet	Screened interval, in feet	Diversion permit, in million gallons per day
PW-1	1938	53	12	—	13	—	0.144
PW-2	1938	43	16	5	10	—	.288
PW-3	1970	79	12	10.4	10	66-76	.36

DELINEATION OF CONTRIBUTING AREAS

Ground-water flow was simulated at two scales, a watershed scale and an aquifer scale. The watershed-scale model, which encompasses upland and valley areas in the Transylvania Brook watershed, was used to understand how the topography, precipitation, and geology of the watershed affect the distribution of recharge to the stratified glacial deposits in the valley. The “distribution of recharge” in this report refers to the contribution of recharge from each upland basin, as a percentage of the total rate of recharge. Contributing areas were not estimated using the watershed-scale model. The aquifer-scale model encompassed the stratified glacial deposits in the valley and used the distribution of recharge estimated by the watershed-scale model. Contributing areas and the magnitude of recharge were estimated in the aquifer-scale model. The three-dimensional, finite-difference ground-water flow computer code known as MODFLOW-96 was used for both the simulations (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996).

Watershed-scale simulation

The watershed-scale model was used to simulate processes in upland areas that affect the rates and distribution of ground-water recharge. The distribution of recharge indicated by this model was used in the aquifer-scale model. The simulation at the watershed scale is very coarse and should not be used to predict water levels in specific wells.

The Variable-Recharge package (Kontis, in press) in MODFLOW was used for the simulation. In this package, upland drainage basins (fig. 13) are simulated by specifying the land-surface altitude and the amount of water available for recharge (mean annual runoff). The model then calculates the head in the aquifer. If the head is above land surface, water is rerouted as channeled or unchanneled flow into the valley aquifer according to a user-specified ratio. Streamflow processes that might be simulated with the River or Stream packages in MODFLOW are not used in upland areas. The stratified glacial deposits in the valley receive water from upland areas and from direct recharge. As in upland areas, if the head is above land surface, water is rerouted. Streamflow losses from tributary streams in the valley bottom are calculated as in the Stream package (Prudic, 1989). The Variable-Recharge Package can be used to estimate the amount of mean annual runoff that becomes ground-water recharge and the distribution, by drainage basin, of that recharge.

Model grid and layers

Ground-water processes were simulated at the watershed scale using a one-layer model (fig. 13). The bottom elevation of the aquifer was generated by subtracting 80 ft from land-surface altitude in the valley areas and 160 ft in the upland areas. The model grid has cells with a uniform spacing of 500 ft on each side. The grid is 65 rows by 27 columns and encompasses an active grid area of 5.03 mi^2. The upland area is represented by 463 cells covering an area of 4.15 mi^2, and the valley is represented by 98 cells covering 0.88 mi^2.

Hydraulic properties of the aquifer

Simulated values for aquifer properties were based on previous studies (tables 3 and 4). Hydraulic conductivity of till and (or) bedrock, fine-grained stratified deposits, and coarse-grained stratified deposits was 0.5, 5, and 80 ft/d, respectively. The vertical hydraulic conductivity divided by the thickness of the riverbed deposits was 1 d^{-1}.

Boundary conditions and stresses

Water levels in shallow water-table aquifers often are a subdued replica of the land surface. No evidence to the contrary is present at the STS; therefore, the lateral boundaries of the aquifer system are assumed to be the watershed of Transylvania Brook (fig. 13) and are treated in the model as no-flow boundaries. In general, some differences could be present between the surface-water and ground-water divides, especially where there are large changes in hydraulic properties of the aquifer or large ground-water withdrawals near the boundary of the aquifer. The only known place in the study area where either of these conditions apply is along the possible fault between the arkose and crystalline bedrock. The fault could be a zone of hydraulic conductivity that is higher than the unfaulted bedrock that surrounds it, or it could be sealed shut by mineral deposits in the fault, by the smearing of rock debris into the fault openings, or by dense till forced into the top of the fault during glaciation. The fault was not simulated in the watershed-scale simulation, but the effects of the fault were tested in the aquifer-scale simulation.

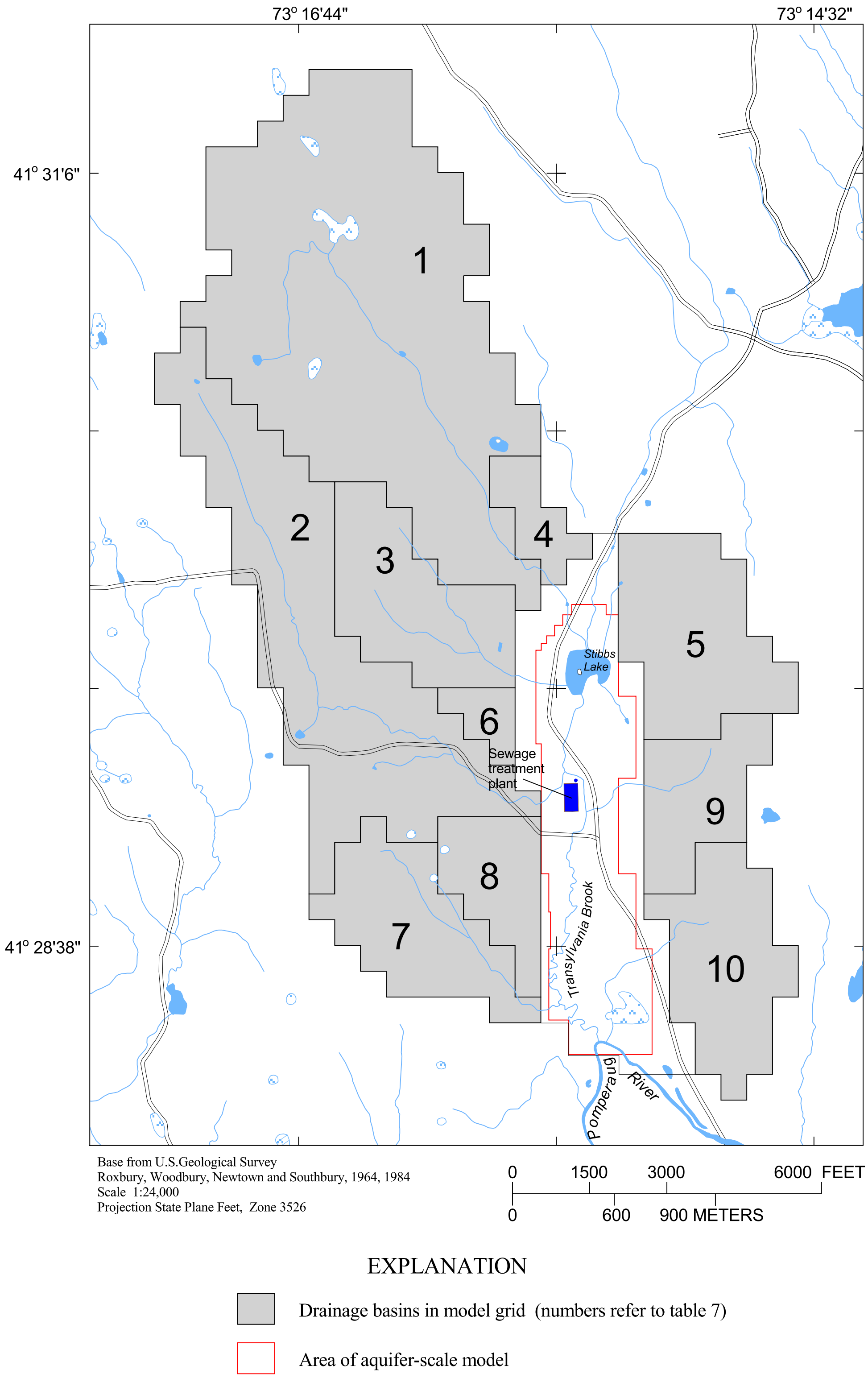


Figure 13. Extent of watershed-scale model grid, Transylvania Brook watershed, Connecticut.

The lower boundary of the ground-water-flow system is formed by the imaginary surface that separates the water that discharges to Transylvania Brook from the water that flows beneath the watershed and discharges elsewhere or is relatively stagnant. Under this definition, the ground-water-flow system includes some water in the bedrock and all water in the glacial deposits. Some water in the bedrock may discharge to surface water in the watershed, but some water could flow out of the watershed through deep fractures or regionally extensive faults in the bedrock. No evidence indicates any significant quantities of water leaving the watershed through subsurface fractures or faults.

The upper boundary of the ground-water-flow system is the water table, which is in coarse-grained deposits, fine-grained deposits, till, or bedrock. The water table fluctuates up and down as the amount of ground water in storage changes.

The water available for recharge was applied over the entire modeled area. The watershed-scale model was intended to simulate natural conditions, and pumping was not simulated.

Model calibration

The watershed-scale model was checked for plausibility by comparing it to historical water levels recorded on well-completion reports and the amounts of recharge estimated using equation 1 for upland and valley areas. No changes were made to assigned values in the model except that mean annual runoff was lowered slightly from 8.60 ft^3/s to 7.85 ft^3/s to achieve a better match between simulated recharge rates and those predicted using the regression relation (equation 1). This reduction was reasonable because the watershed-scale simulation considered only steady-state conditions; some of the mean annual runoff would be stormflow, a non-steady-state process not included in the model.

The watershed-scale model produced reasonable hydraulic heads within the limits of the data. Model heads were evaluated against measurements that were collected over a long period of time and a wide range of hydrologic conditions. Overall, different hydrologic conditions represented by the data probably average out, and the model and measured values are reasonably close. The mean error (observed head minus simulated head) was -1.16 ft. A value close to zero indicates that the model is unbiased and that the errors are random. In this case, the model is judged to be unbiased. The mean absolute error was 16.81 ft; this is the amount of deviation from the mean in the data. The measured heads in the watershed have a range of 349 ft from the lowest to highest measurement; therefore, the mean absolute error is about 5 percent of the total change in head. The root mean square error is 22.5 ft, or about 6 percent of the total change in head. This amount of error, relative to the total head loss in the system, was considered acceptable.

Spatial bias also is important is assessing the plausibility of a model. A model should not consistently over- or under-predict water levels in any region of the modeled area. In the watershed-scale model, spatial bias was difficult to assess on the basis of measured water levels because the measurement locations are not evenly spread across the watershed (they are clustered around the valley), and the water levels were measured at different times of different years. Another way to assess spatial bias is to look at the predicted position of streams in the upland areas. The Variable-Recharge package uses the relation of land-surface and water-table altitudes to determine where a stream should be. In the watershed-scale model, streams were predicted in approximately the correct positions in basins 1, 2, 3, and 7 (fig. 13). Ponds near the headwater of the stream in basin 1 also were predicted by the model. These observations support the conclusion that predicted water levels were not consistently above or below the expected level near these streams.

The watershed-scale model yielded a net recharge rate of 8.8 in/yr to upland areas, slightly more than the 8 in/yr predicted by equation 1. Recharge from upland areas to stratified glacial deposits is equivalent to infiltration minus seepage losses to upland streams plus streamflow loss where upland tributaries cross over the coarse-grained deposits and unchanneled flow from upland basins. Net recharge from individual upland basins is given in table 7.

The model yielded a net recharge rate of 23.6 in/yr to the valley aquifer directly, which is higher than the 22 in/yr predicted by equation 1. Direct recharge from valley areas to stratified glacial deposits is equivalent to infiltration minus seepage losses to valley streams plus streamflow losses from the main channel of Transylvania Brook. Net direct recharge to individual basins in the valley is given in table 7.

Table 7. Recharge to stratified glacial deposits, Transylvania Brook drainage basin, Connecticut

Basin	Recharge from upland area		Direct recharge to valley area		Total recharge to valley, in cubic feet per second
	in cubic feet per second	in inches per year	in cubic feet per second	in inches per year	
1	0.18	1.7	0.05	23.6	0.23
2	.16	2.5	.03	23.6	.19
3	.50	20.9	.10	23.6	.60
4	.14	21.2	.04	23.6	.18
5	.31	11.7	.14	23.6	.45
6	.08	20.2	.10	23.6	.18
7	.27	12.2	.02	23.6	.29
8	.27	21.5	.15	23.6	.42
9	.32	21.1	.25	23.6	.57
10	.47	20.2	.02	23.6	.49

Results of the simulation

The main conclusion to be drawn from the watershed-scale simulation is that upland recharge is not evenly distributed (table 7). In the small, undrained basins, primarily those on the eastern side of Transylvania Brook (fig. 13), almost the entire amount of water available for recharge enters the aquifer and passes through the subsurface into the coarse-grained stratified deposits. In three basins on the western side of Transylvania Brook (fig. 13; basins 1, 2, and 7) that are drained by well-defined streams, most of the upland recharge enters the streams, which then flow out onto the valley aquifer. Where these streams flow over coarse-grained deposits, water flows from the stream into the aquifer and is counted as recharge. Tributary streams that do not have a large upland catchment area (basin 5 in fig. 13, for example) do not recharge the aquifer, rather they drain water from the aquifer.

Aquifer-scale simulation

The aquifer-scale model was used to calculate contributing areas to the two water-supply wells at the STS. Parameters of the model were estimated using nonlinear regression and drawdowns measured during an aquifer test. The distribution of recharge was obtained from the watershed-scale simulation, but the magnitude of recharge was estimated by the aquifer-scale model to reflect hydrologic conditions at the time of the aquifer test. Contributing area simulations were done using hypothetical drought recharge conditions, which were based on information collected in the Pomperaug River Basin during the drought of record in the mid-1960s.

Ground-water-flow models typically are calibrated by a trial-and-error method, in which the model is adjusted by the modeler until a reasonable match between calculated and observed heads and flows is produced. Nonlinear regression makes calibration more efficient and objective because parameter values are adjusted to obtain automatically the best possible fit between simulated and observed values. The model fit is measured by the sum of squared weighted errors (SSE), where error equals the simulated minus observed values. The statistical framework of this process can be used to test the validity of the regression, the reliability of the parameter estimates, and the likelihood that a given model represents the system more accurately than an alternative model. The method is described by Cooley and Naff (1990), Hill (1992, 1994, and 1998), and Poeter and Hill (1997). The computer program used is called UCODE, which stands for **U**niversal Inverse **Code** (Poeter and Hill, 1998). UCODE works with any model code by modifying model input files from user-defined templates, running the model, reading the model output, running the nonlinear regression, and modifying model parameters until the optimal parameter values are reached.

In this study, UCODE was used with MODFLOW-96. In UCODE/MODFLOW-96, the user defines parameters that represent the boundary conditions, stresses, and hydraulic properties of the aquifer system. Inputs to the transient model that apply to areas of the model grid, like aquifer properties and recharge, can be defined using zones (part of the model grid corresponding to a particular geologic unit). Model parameters also can be created using user-defined functions. For example, the vertical conductance between model layers, which is used as input by MODFLOW-96, is calculated by dividing the parameter estimate by the distance between the centers of the overlying and underlying layers.

Model grid, layers, stresses, and time steps

The aquifer-scale model grid was designed to simulate ground-water flow in the stratified glacial deposits and underlying bedrock in the Transylvania Brook watershed (fig. 14). The grid has 44 columns and 56 rows, variably spaced from about 10 ft near the pumped well (PW-3) to about 200 ft near the edges of the model grid. The active cells in the model cover about 1.459×10^7 ft^2, or about 0.5 mi^2.

The ground-water-flow system was simulated using three layers, although in some alternative models, described later in this report, bedrock was simulated as a fourth layer. All layers were simulated as convertible, meaning that MODFLOW-96 would determine if the layer was a water-table layer or a confined layer and then apply the appropriate hydraulic property values. In this study, all geologic units were simulated, so the bottom altitude of a given layer is the same as the top altitude of the underlying layer. Model layers are defined according to a percentage of the total thickness and to the geology, where the total thickness of the stratified deposits is the distance between land surface and the bedrock surface. Where the entire thickness of the deposits is the same geologic unit, layers 1 and 2 each represent one-quarter of the total thickness, and layer 3 is one-half of the total thickness. Where the deposits are in different geologic units, layers 1 and 2 are each one-half of the upper unit, and layer 3 is the lower unit. The top of layer 3 is shown on figure 15.

The stratified deposits are draped along the side of the bedrock valley wall (figs. 6 and 7), particularly on the western side of the valley. The water table probably is flatter than the slope of the bedrock surface along the lateral margins of the valley, and the saturated part of the aquifer gradually thins toward the edges of the model grid. This gradual thinning was incorporated into the model grid design by designating the minimum saturated combined thickness of layers 1 and 2 to be 20 ft. The additional volume of aquifer material simulated in this way was offset by making model grid cells in layer 3 inactive where the total estimated saturated thickness was less than 40 ft.

The pumped well (PW-3) was simulated as an aquifer stress using the Well package in MODFLOW-96. The rate of pumping as measured during the test was used as the stress rate (185 gal/min), and the well was simulated in layer 3. Other wells at the STS were not in use during the test and therefore were not simulated. Domestic water wells also were not simulated, because the total volume of water pumped was small and spread over a large area. In addition, these wells were far from the pumped well at STS, thus minimizing any effect on the aquifer test.

The model consisted of 50 time steps that increase in size from 0.0001 to 0.8334 d by a factor of 1.2. The total length of the stress period was 5 d. The initial conditions for the transient simulation were generated by a steady-state model. The parameter values in the model were estimated simultaneously for the steady-state and transient models.

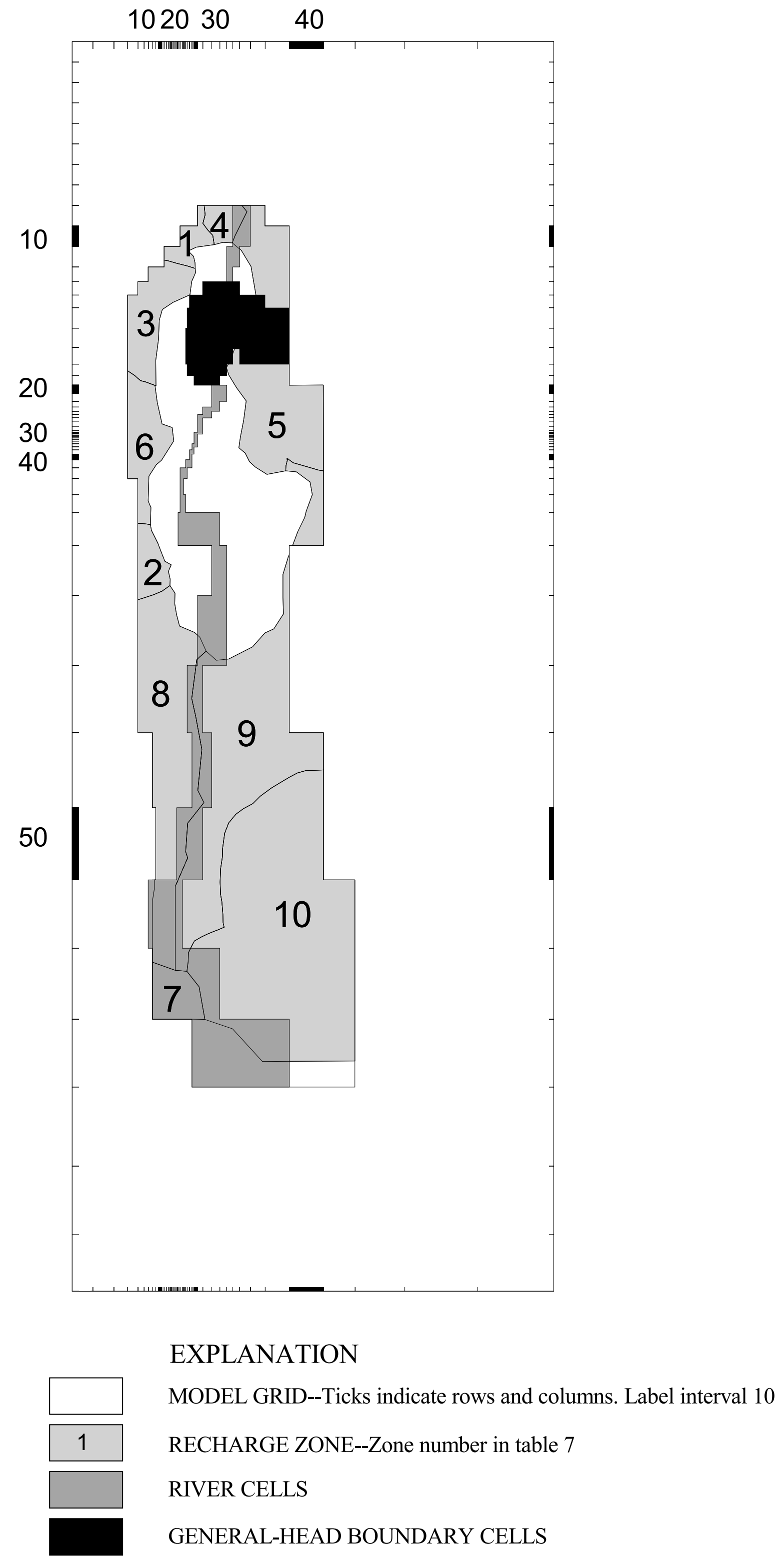


Figure 14. Extent of aquifer-scale model grid and boundary conditions, Transylvania Brook valley, Connecticut.

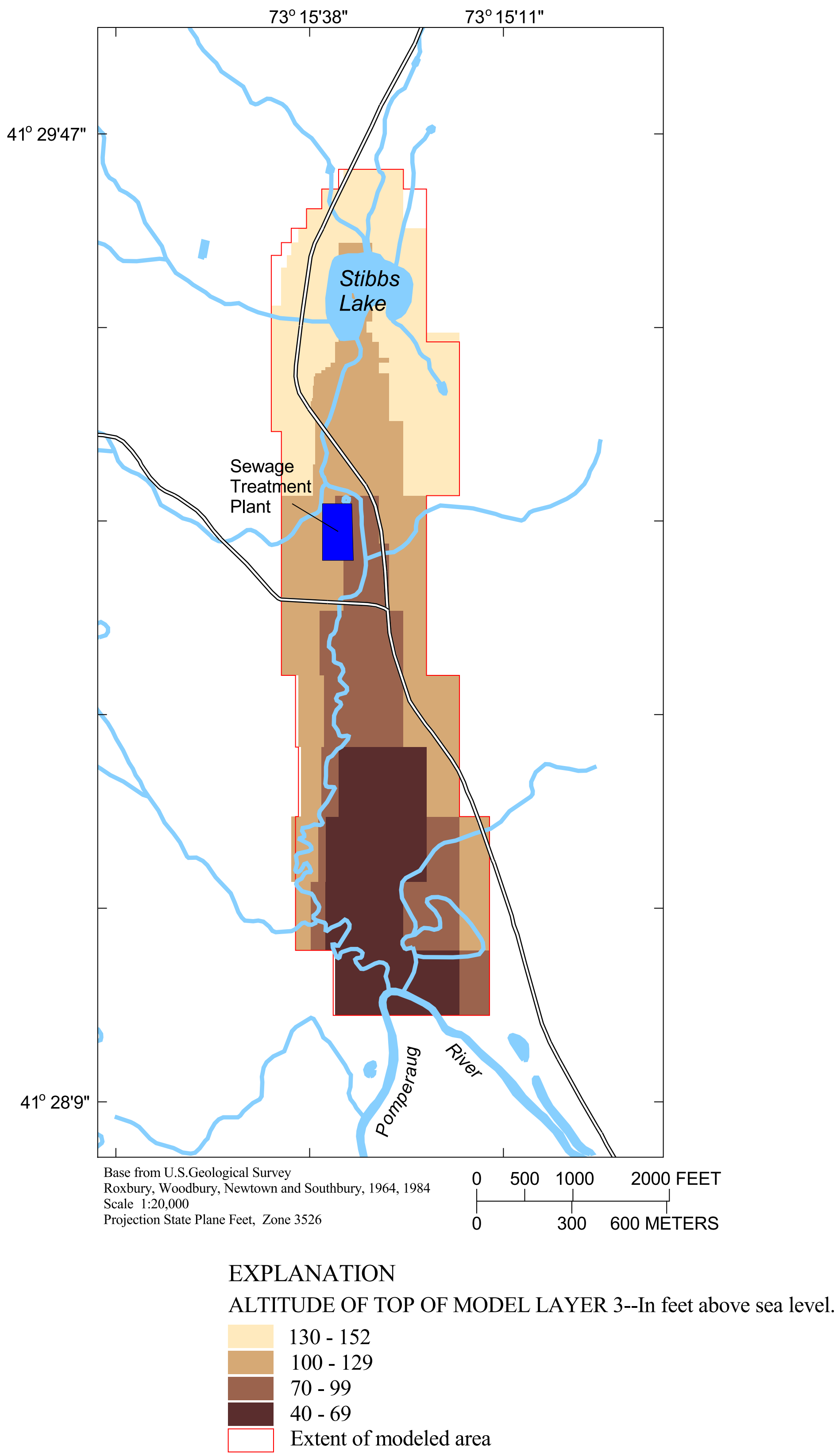


Figure 15. Altitude of top of model layer 3, aquifer-scale model, Transylvania Brook valley, Connecticut.

Boundary conditions and model parameter definition

Nine model parameters were defined, although not all parameters were estimated in the final simulation. Six parameters define aquifer properties, and three define boundary conditions. The simultaneous estimation of all defined model parameters was not possible because the model was not sensitive to all parameters. If calibration data do not contain sufficient information about the parameter, the model is not sensitive to that parameter and the parameter cannot be reliably estimated. Parameters that cannot be estimated may nonetheless be important for model predictions. In this report, independent information on unestimated parameters is used to include their effect in model predictions.

The stratified deposits at the STS were divided into two zones related to the surficial geologic units (fig. 16). Coarse-grained deposits were designated zone 1, and fine-grained deposits were designated zone 2. One hydraulic conductivity parameter is defined for zone 1 in layers 1, 2, and 3, and another for zone 2 in layers 1, 2, and 3. Layer 4 (bedrock), used only in an alternative model, also was divided into two zones, one representing bedrock (schist and arkose, zone 3), and one representing the possible fault contact between the two rock types (zone 4).

The uppermost model layer contains the water table (the “water-table layer”), and the amount of water released from storage in the aquifer is proportional to the specific yield and the decline in water level during pumping. Although the water-table layer contains coarse and fine deposits, both were assigned the same specific yield, which was treated as a parameter. It is expected that the specific yield of the coarse and fine deposits represented by zones 1 and 2, respectively, are different, and the specific yield estimated by the model is more representative of zone 1 than zone 2. Treating them as the same is not expected to affect the model appreciably because drawdown was much less in zone 2 than in zone 1. In layers below the water-table layer, water is not produced from de-watering of the aquifer as in the water-table layer. Instead, water is released from storage in the aquifer in proportion to the storage coefficient and the decline in pressure in the aquifer. One parameter is used for the specific storage in zones 1 and 2. Specific storage is the storage coefficient divided by the thickness of the model layer, which UCODE multiplies by the layer thicknesses.

The stratification of the glacial deposits produces a difference in hydraulic conductivity in the vertical direction relative to the horizontal direction. A function is defined in UCODE to divide vertical hydraulic conductivity by the vertical distance between the center of layers to produce the vertical conductance term required by MODFLOW-96. The vertical hydraulic conductivity was represented by two parameters, one for each of zones 1 and 2. If vertically adjacent model cells contain different zones, the vertical hydraulic conductivity of zone 2 is used in the calibration. The effective hydraulic conductivity is the harmonic mean of the two, which is weighted heavily toward the lower of the two values (zone 2).

Boundary conditions govern the flow of water into and out of the model grid during the simulation (fig. 14). The amount of this flow is governed by boundary condition terms that can be defined as model parameters. Two types of boundary conditions were used in the model, specified flow and head-dependent flow. Specified-flow boundaries were used to simulate the base and lateral boundaries of the model (flux = 0) and recharge. Recharge was applied using the Recharge package in MODFLOW-96. Simulated recharge included the combined ground-water recharge from upland basins and valley bottom (table 7), applied to the corresponding area in the valley bottom (fig. 14). A parameter was defined that proportionally changed this recharge. A parameter value of 1.0 produced the same amount of recharge as the watershed-scale simulation. Recharge was applied only to zone 1 areas (coarse-grained deposits) in the uppermost active model layer. Recharge was not applied to zone 2 because of the expected low vertical hydraulic conductivity of this unit, and because most of the recharge would be expected to occur near the valley wall, which corresponds to the location of zone 1 near the STS. This approach differs from that commonly used in modeling of valley aquifers in New England in which upland recharge is added through imaginary wells at the edge of the valley (Mazzaferro, 1986).

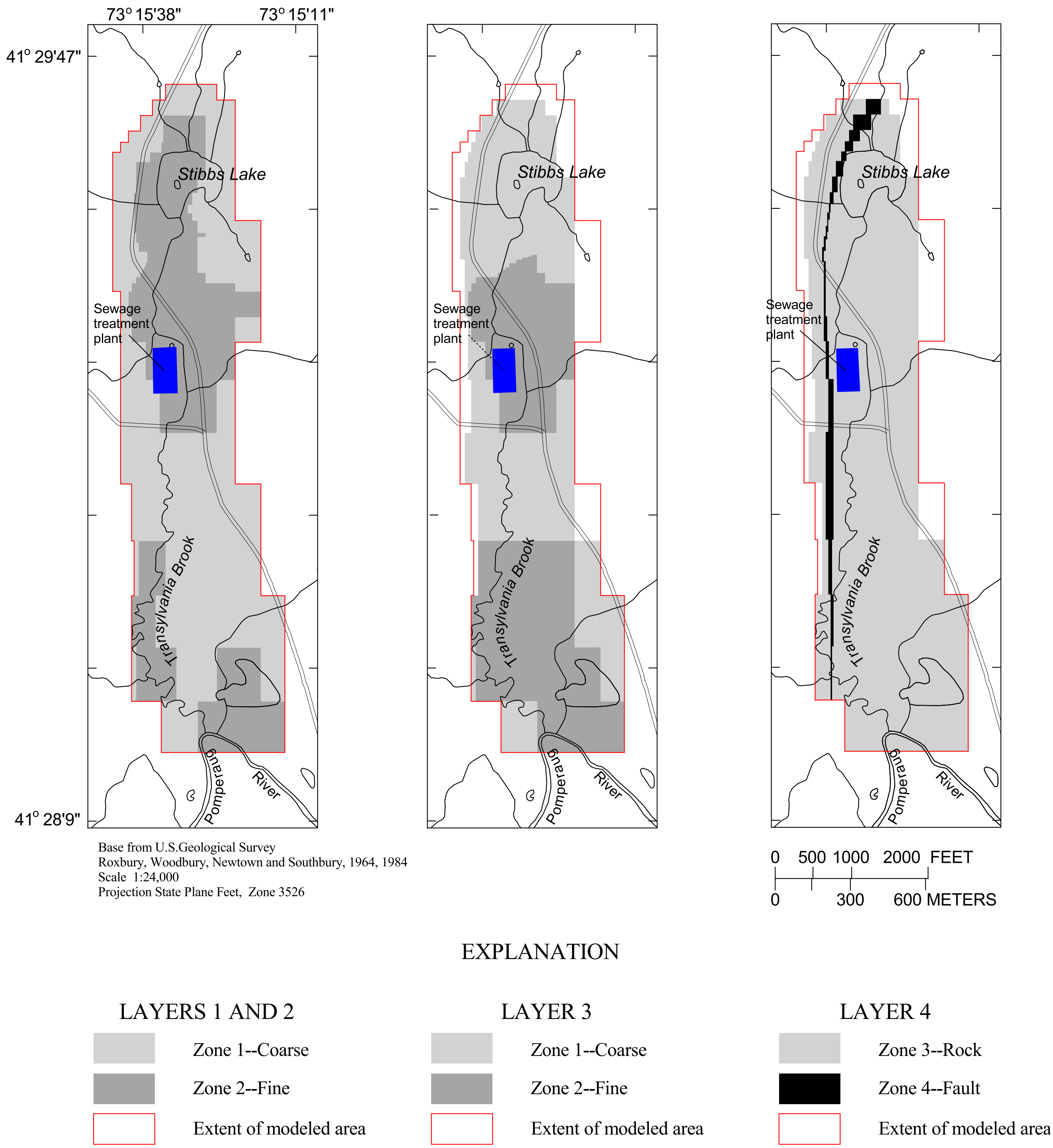


Figure 16. Aquifer zones, aquifer-scale model, Transylvania Brook valley, Connecticut.

In this study, the parameter estimation procedure could change the simulated location of the edge of the valley, which made the use of imaginary wells problematic. Zone 1 constitutes a narrow band of cells (areas 1, 2, 3, 4, and 6 on fig. 14) in the area near the well field that is similar to the one-cell-width band that would receive recharge if imaginary wells were used. The approach used in this study spreads the source of recharge over a larger area than would be the case if wells were used, and thus slightly overestimates the contributing area.

Head-dependent flow boundaries are used to simulate flux to and from Transylvania Brook and Stibbs Lake. Transylvania Brook was simulated using the Stream Package (Prudic, 1989), and Stibbs Lake was simulated using the General-Head Package in MODFLOW-96. A head-dependent flow boundary requires the specification of a conductance, which controls the flow of water to (or from) an external source of water; flow equals the conductance times the difference in head between the boundary and the calculated value. In the case of Transylvania Brook and Stibbs Lake, the external hydraulic head is the average water level in the water body. The Stream Package also requires the altitudes of the top and bottom of the streambed, which were estimated as 1 and 2 ft below the average water level in the stream, respectively. The Stream Package accounts for water as it moves from one model cell to the next in the downstream direction and allows the stream to go dry if the water flowing out of the stream exceeds the amount of water flowing into the stream.

In this implementation of UCODE and MODFLOW-96, parameters are defined for hydraulic conductivity of the stream and lake bed. For the stream boundary, a function is defined in UCODE that multiplies the hydraulic conductivity of the streambed by the area of the stream in each cell, divided by the thickness of the streambed. The stream areas were estimated using streamflow-measurement notes. For the lake boundary, the UCODE function multiplies the hydraulic conductivity of the lake bed by the area of the lake in each cell, determined using the 1:24,000-scale topographic map of the area. The thickness of the lake bed was assumed to be 1.0 ft.

Model calibration

The model was calibrated to (1) steady-state water-level measurements made on October 30, 1996; (2) drawdown measurements made during the aquifer test from October 30-November 4, 1996; and (3) water-level rises after the rainfall on October 20, 1996. The calibration data consist of a set of measurements in each of 5 wells—one steady-state, pre-test water-level altitude in each well and 11 transient drawdown measurements in each well, for a total of 60 measurements. The 11 transient measurements that best represented drawdown in the aquifer were selected from many measurements made during the test. Streamflow measurements were not used in the calibration of the model because stream gains and (or) losses were within the margin of error for streamflow measurements.

Water-level data collected during the aquifer test were affected by factors external to the test that had to be removed from the data to calculate drawdown caused by pumping. External effects include (1) long-term water-level changes caused by years of pumping at the STS, (2) rise in water levels caused by the annual period of ground-water recharge that often takes place in Connecticut in the fall, and (3) rise in water levels caused by the pump being turned off for about 2 days prior to the start of the test.

The effects of long-term water-level changes were assumed to be independent of time (in other words, average annual water levels are lower than they would be naturally, but have stabilized over decades of pumping at the STS); therefore, the long-term water-level correction depends only on distance from the pumped well. The radius of the area affected by pumping was calculated by assuming a circular recharge area around the well and a recharge rate of 22 in/yr. The steady-state ground-water flow equation (Fetter, 1994, p. 218) was used to correct the steady-state water-level measurements made prior to the aquifer test by assuming no drawdown at the radius of the area affected by pumping. A hydraulic conductivity of 80 ft/d was used in this analysis, as determined by preliminary analysis of the aquifer test. This correction resulted in the addition of 0 to 3.50 ft to each of the steady-state water-level altitudes.

In Connecticut, water levels tend to rise from October to May (fig. 11); however, water levels in 1996 probably rose faster than normal because of the extreme precipitation event on October 20 (fig. 10). This rise in water levels was treated as a linear trend defined by the increase in peaks in water levels measured several days before the aquifer test (fig. 12). A linear trend was fit to the rise of the lowest water level each day for each of the wells in zone 1 and was assumed to be steady during the aquifer test. The slope of each trend line was multiplied by the time, in days, since the beginning of the aquifer test and was added to drawdown. The trend correction resulted in a larger drawdown in all wells in zone 1 by 0.3 to 0.8 ft over the length of the test.

The effects of water-level recovery after pumping stopped, prior to the aquifer test, is shown following the last pumping cycle on figure 12. Ideally, the water levels should have reached a stable value by the time the aquifer test started, and corrections to drawdown would not be necessary; however, water levels were still rising at the start of the test. The non-steady ground-water flow equation (the Theis equation; Fetter, 1994, p. 201) was used to correct the drawdowns for water-level recovery by assuming a transmissivity of 5,600 ft^2/d (from the preliminary aquifer-test analysis) and a storage coefficient of 0.0004, beginning when the pump was turned off. A constant pumping rate of 235 gal/min was used. The simulated water-level recovery at each time during the test, subtracted from the water level at the beginning of the test, was added to each measured drawdown. The result of this correction was an increase in drawdowns of 0.09 ft over the duration of the test.

In UCODE, calibration data are assigned weights, thus allowing data that are known with a higher degree of accuracy to have a greater effect on the regression than data that are less well known. In this case, drawdown measurements made during the aquifer test (that help define the hydraulic properties of the aquifer and are accurately known) affect the regression more than the steady-state water-level altitudes (that help define the role of recharge and discharge in the aquifer and are less accurately known). According to regression theory (Hill, 1992, 1994, and 1998), the weights need to reflect possible error in the measurement of the data and are proportional to 1 divided by the variance of the measurement. Weights for measurements of water-level altitude are based on the assumption that 95 percent of the steady-state water level measurements are within 2.0 ft of their true value. This assumption is based on a qualitative assessment of how accurately the measured water level represents the horizontal and vertical heterogeneity of the aquifer within the entire model cell and on the estimated accuracy of the steady-state water-level correction. The measurement errors are assumed to be normally distributed, so that 2.0 equals the standard deviation (the square root of the variance) times the critical value at the 95-percent confidence interval (1.96). The variance for water-level altitudes is about 1.00.

Water levels measured during the aquifer test were converted to drawdowns (the initial water level minus the water level at some time during the test). This subtraction cancels out many sources of error in the water-level altitude measurements. The weights for the drawdown measurements are based on the assumption that 95 percent of the measured drawdowns are within 0.1 ft of their true value. The actual measurements were made with a pressure transducer having an accuracy stated by the manufacturer of 0.03 ft; however, this accuracy may not have been realized under actual field conditions. The variance was calculated as above and equals 0.0026 for drawdown measurements.

Description of alternative models

Ground-water models are nonunique because hydrologic features in the flow system can be simulated in many ways. To evaluate the effect of this nonuniqueness, seven alternative models were posed in this study by simulating possible hydrologic features in various combinations. The alternative models are compared using various statistical measures of model fit and regression performance and, all else being equal, simpler models are given preference. In this study, one model was identified as best representing the flow system. The choice of alternative models is affected by the amount and type of data used for model evaluation. Some hydrologic features that might be important to simulating contributing areas may not be well represented in the model because the data used for model development did not contain enough information about those particular features. Composite-scaled sensitivities, which are computed by UCODE, can be used to evaluate if a model parameter can be estimated with the model. Composite-scaled sensitivities less than about 0.01 times the largest composite-scaled sensitivity indicate that UCODE may not be able to estimate the parameter (Hill, 1998, p. 38). The alternative models are described below and are designated by the letters CAL (for “calibration”) followed by the numerical identifier of the simulation.

CAL0—The first alternative model was designed to include only the minimum number of hydrologic features needed to simulate the aquifer-test data. This model excludes Stibbs Lake (the hydraulic conductivity of the lake bed was set to $1x10^{-7}$ ft/d). This may be reasonable because the lake overlies fine-grained deposits, which limit the lake as a source of water to the aquifer. The hydraulic conductivity of the streambed was set to 1.0 ft/d. Based on the composite-scaled sensitivities computed at the optimal parameter values, the vertical hydraulic conductivity of zone 2 could not be estimated and was fixed at $1x10^{-4}$ ft/d. Six parameters were estimated in the model—recharge, horizontal and vertical hydraulic conductivities of zone 1, horizontal hydraulic conductivity of zone 2, specific yield, and specific storage.

CAL1—The second alternative model was identical to CAL0 except that the streambed hydraulic conductivity was fixed at 5.0 ft/d. A typical range for this parameter is 0.13 to 14 ft/d (Wilson and others, 1974, p. 30); however, the model predicted significant negative recharge rates for streambed conductivities greater than 5.0 ft/d.

CAL2—This model was identical to CAL0 except that Stibbs Lake was included by setting the hydraulic conductivity of the lake bed to 0.1 ft/d, based on values reported by Wilson and others (1974, p. 31). The sensitivities of this model at the optimal parameter values indicate that it would not be possible to estimate recharge while allowing leakage from the lake; therefore, recharge was fixed at the value estimated in CAL0.

CAL3—This also was identical to CAL0 except that a fourth layer was added to represent the bedrock. Previous alternative models were based on the assumption that the bedrock was impermeable; however, most domestic water wells in the area get their water from bedrock, so the bedrock is not completely impermeable. In this model, the hydraulic conductivity of the bedrock was set to 5 ft/d, based on table 3, and the thickness was assumed to be 200 ft. The specific storage of this layer was set to $1x10^{-5}$.

CAL4—This model is identical to CAL3 except that the fault (fig. 4) in the bedrock is represented. The hydraulic conductivity of the fault was set to 200 ft/d to reflect the hypothesis that the fault might be a highly permeable zone within the bedrock. The thickness and specific storage of this layer are the same as for CAL3.

CAL5—This model is identical to CAL0, except that the vertical hydraulic conductivities of zone 2 and the lakebed were fixed at 0.1 ft/d. The model was designed to test the hypothesis that the vertical hydraulic conductivity of zone 2 is 0.1 ft/d (table 3).

CAL6—This model was similar to CAL2, but bedrock was modeled as in CAL4. The model was designed to test the hypothesis that there might be interaction between Stibbs Lake and the fault.

Results of the simulation

Alternative models CAL0, CAL1, and CAL5 were very similar, based on their statistical properties (table 8), and all three fit the data more closely than the other alternatives. Model CAL1 was found to be sensitive to initial parameter estimates. In other words, different starting values resulted in different estimated parameters. In CAL1, recharge estimates tended to be lower than in CAL0 to offset the increased amount of water that could leak from the stream because of the higher streambed hydraulic conductivity. This was reflected in the model by a high correlation (0.94) between streambed hydraulic conductivity and recharge. In CAL1 and CAL5, the source of almost all water in the model is a few stream-boundary cells in the northern area of the model, and the optimal recharge rate was negative. This situation is considered unrealistic, and CAL0 was chosen as the most representative model.

Table 8. Statistical measures of model fit of alternative models

Statistical measure	CAL0	CAL1	CAL2	CAL3	CAL4	CAL5	CAL6
Least squares objective function	278	264	343	313	482	278	460
Calculated error variance	5.15	4.89	6.23	5.80	8.93	5.15	8.52
Standard error of the regression	2.27	2.21	2.50	2.41	2.99	2.27	2.92
Correlation coefficient for observations	.999	.999	.999	.999	.998	.999	.999
Correlation coefficient for normal residuals, R^2_N	.955	.951	.983	.956	.974	.959	.974

The range of estimated parameter values among the alternative models was small (table 9). Estimated values generally were within the expected ranges based on typical values (table 3); however, recharge and specific-yield estimates were somewhat different than expected. The multiplication value for recharge was significantly smaller (0.24) than the value estimated by the watershed-scale model (1.0). Taken together with stream losses, this represents about 32 percent of the mean annual recharge estimated in the watershed-scale model. This is probably because the aquifer test was conducted at a time when ground-water recharge was just beginning for the year, and the full amount of annual recharge had not yet taken place. Also, there are some areas where coarse deposits overlie fine deposits (see, for example, the sand unit on the eastern edge of fig. 6). These deposits receive recharge but are not part of the coarse-grained aquifer and so were not simulated. This means that recharge estimated in the aquifer-scale model should be less than in the watershed-scale model.

Specific-yield estimates also were lower than expected. This could be because of the layered nature of the aquifer, which causes a delay in the time between when the head is lowered in the aquifer and when the water can physically drain downward through the various layers to the water table. This phenomenon has been noted in a study of a glacial aquifer with small-scale bedding in which short-term aquifer tests give specific yields that are much lower than values determined in laboratory studies (Nwankwor and others 1984; 1992). In a review of Nwankwor's studies, Moench (1994) found that a drainage delay can affect specific yield estimates, particularly in water-table wells and that unrealistically small specific yields can result from aquifer heterogeneity.

Drawdowns in wells in deep, coarse-grained deposits were accurately simulated by model CAL0 (fig. 17). Of particular significance is the fact that the simulated water levels followed the "s" shape of the data that is typical of water-table aquifers. This shape is caused by a change in the source of water from storage in the aquifer; the early part of the aquifer is dominated by specific storage and the later part is dominated by specific yield. The sensitivity of the model to each drawdown measurement shows the sensitivity to specific storage peaks early in the simulation and sensitivity to specific yield peaks at the end of the simulation period.

Table 9. Optimal parameter values for alternative models

[Kh, horizontal hydraulic conductivity; Kv, vertical hydraulic conductivity; shading indicates parameter not estimated; —, not simulated]

Model parameter	CAL0	CAL1	CAL2	CAL3	CAL4	CAL5	CAL6
Kh, zone 1 (ft/d)	154	155	157	148	132	165	133
Kh, zone 2 (ft/d)	29	26	21	18	12	13	11
Kv, zone 1 (ft/d)	.83	0.87	.94	.66	.60	.73	.62
Kv, zone 2 (ft/d)	1.0E-04	1.0E-04	1.0E-04	1.0E-04	1.0E-04	1.0E-01	1.0E-04
Specific yield, unitless	.0069	.0073	.006	.0082	0.010	.0048	.0095
Specific storage (/ft)	1.6E-05	1.6E-05	1.6E-05	1.6E-05	1.5E-05	1.6E-05	1.5E-05
Lakebed Kv (ft/d)	1.0E-07	1.0E-07	1.0E-01	1.0E-07	1.0E-07	1.0E-07	1.0E-01
Streambed Kv (ft/d)	1.0	5.0	1.0	1.0	1.0	1.0	1.0
Recharge multiplier, unitless	.23	-.002	.24	.23	.20	-.007	-.095
Bedrock Kh (ft/d)	—	—	—	5	5	—	5
Fault Kh (ft/d)	—	—	—	—	200	—	200

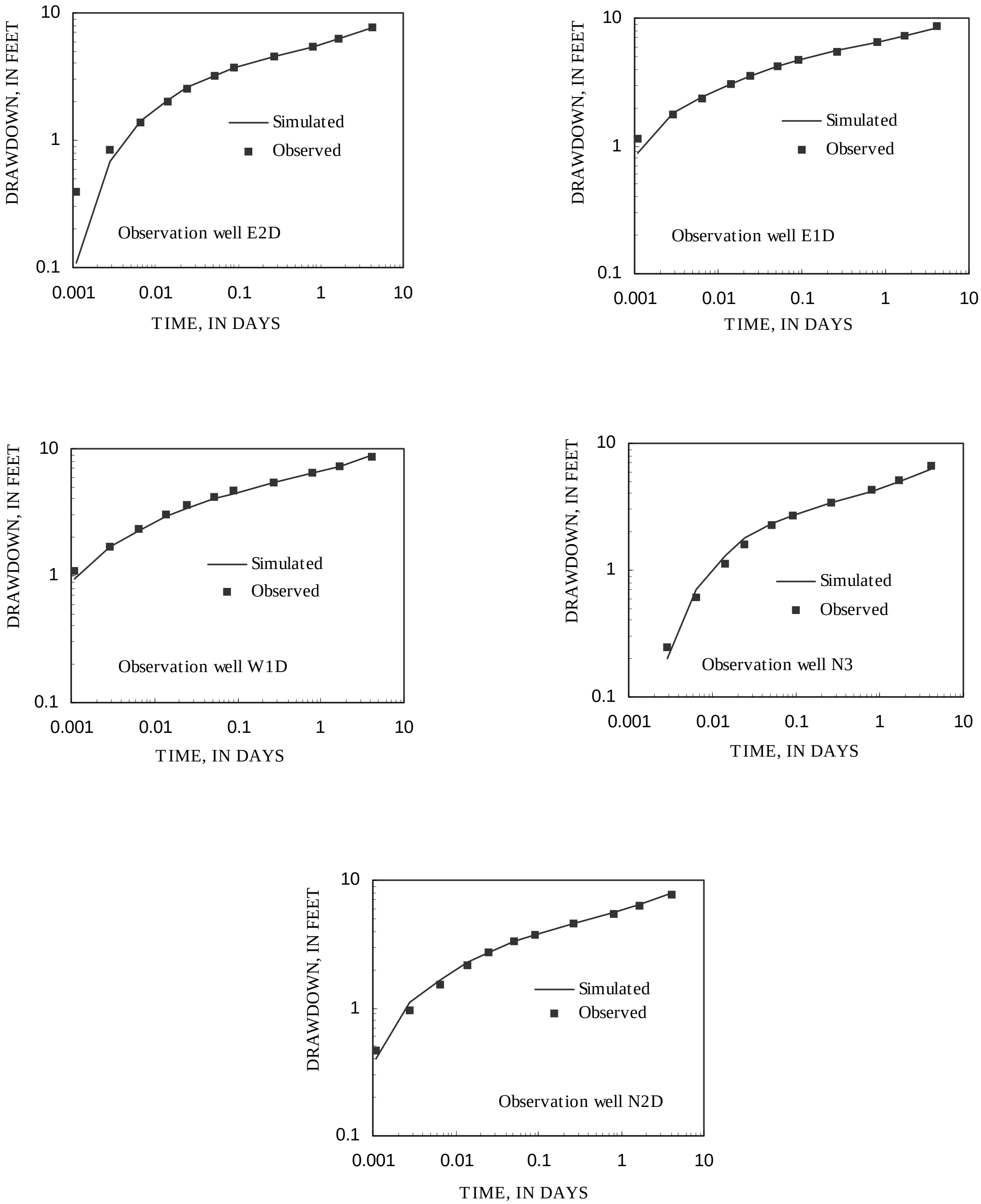


Figure 17. Simulated and observed water-level changes in coarse-grained deposits during the aquifer test, October 30 to November 4, 1996, Southbury Training School, Connecticut.

Steady-state water levels were accurately simulated in most, but not all wells (fig. 18). Observed water levels in coarse-grained deposits in two wells (W3 and N5) were significantly above simulated values. There are plausible explanations why these two wells were not simulated accurately. Well W3 is at the very edge of the model grid. In this area, two important features affect the water level in the well, and either could cause higher water levels than those predicted by the model: (1) the coarse deposits outside the model area may not have drained completely following the rainfall on October 20, 1996, and (2) the abrupt steepening of the bedrock valley in this area causes high vertical gradients over a short horizontal distance. The other well, N5, is a flowing well. The water level measured in that well was above the altitude of the top of the confining layer as it was mapped. The model could only simulate the flowing condition if the extent of the confining layer were increased to higher altitudes. The extent of the confining layer was not changed because no boring or other geologic data exist to support such an extension. The extent of the confining layer is very important in calculating contributing areas; however, an underestimate of the extent, as in this case, exposes more coarse-grained material at the land surface and probably slightly overestimates the contributing area.

Drawdowns in the fine-grained deposits were not as well simulated as levels in coarse-grained deposits, perhaps because of the relation among the layers in those deposits, the length of the well screen, and the thickness of the model layer. The fine-grained deposits are composed of many thin individual layers; this results in high vertical hydraulic gradients. This may cause observed water levels in wells with screens much shorter than the model layer thickness to be different from simulated water levels.

The simulated water-table map shows the expected pattern of ground-water flow. The shape of the 170-ft contour (fig.19) shows that streamflow recharges the aquifer in the northern part of the modeled area where it first comes in contact with the coarse-grained deposits. The 160-ft contour (fig.19) shows that flow is generally down valley toward the lower reaches of Transylvania Brook. At the end of the aquifer test, the 160-ft contour shows that streamflow was contributing flow to the aquifer. The 150-ft contour (fig. 19) shows little change caused by the aquifer test and that the stream is losing water to the aquifer. South of the area shown on figure 19, the water-table gradient is steeper where the brook is underlain by fine-grained deposits than where it is underlain by coarse-grained deposits. Ground water generally flows down valley until an area is reached where the brook is underlain by coarse-grained deposits. In these areas, ground water discharges to the stream at greater rates than where the stream is underlain by fine-grained deposits.

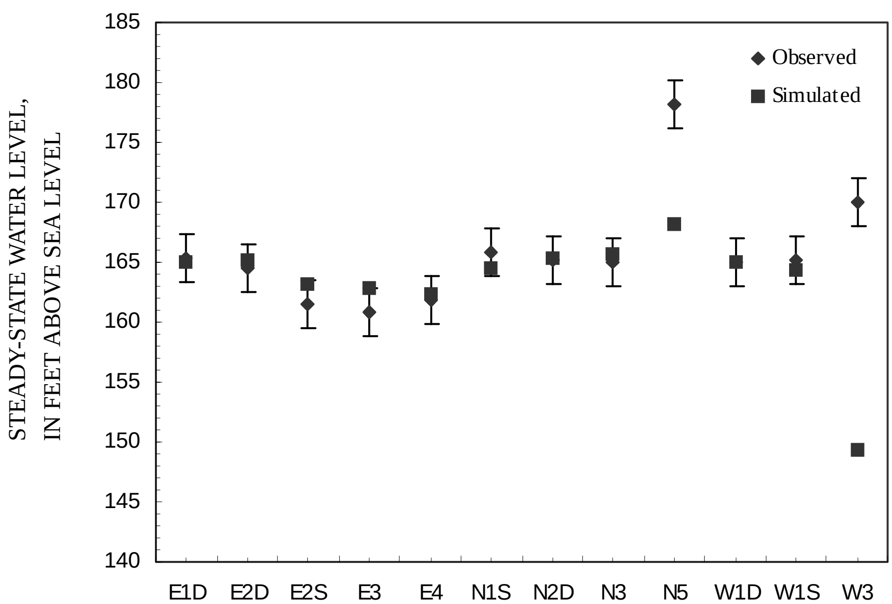


Figure 18. Simulated and observed steady-state water levels, October 30 to November 4, 1996, Southbury Training School, Connecticut. [Error bars show an interval of plus or minus 2 feet.]

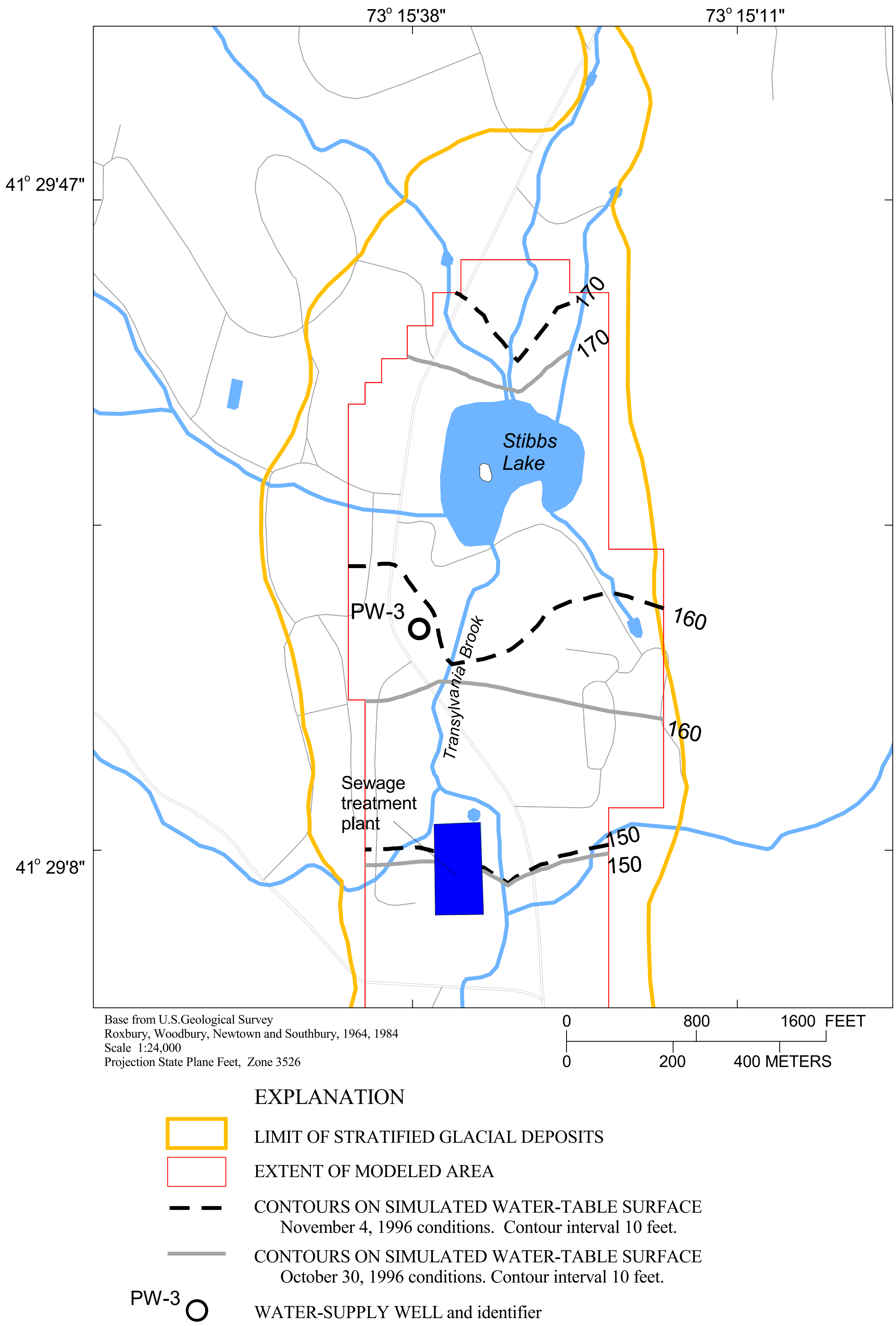


Figure 19. Altitude of simulated water table before and after the aquifer test, October 30 to November 4, 1996, Southbury Training School, Connecticut.

The change from a gaining stream to a losing stream during the aquifer test is consistent with the streamflow measurements made before and after the aquifer test (table 3). The model simulates a gain in streamflow before the aquifer test from the upstream USGS streamflow-gaging station (01204340 on fig. 3) to the downstream USGS station (01204350 on fig. 3) of 0.02 ft^3/s. The measured rate of streamflow gain on this date (October 29, 1996) was 0.03 ft^3/s (table 2). At the end of the aquifer test, the model simulates a streamflow loss of 0.10 ft^3/s. The measured rate of streamflow loss on this date (November 4, 1996) was 0.15 ft^3/s. As mentioned previously, this relation was contradicted by data from the streambed piezometers, which indicated that the stream was gaining water throughout the aquifer test. This may be because (1) the stream really was gaining and the streamflow measurements are too imprecise to determine this small flow, or (2) the piezometers were not in places where the stream was losing water.

Model CAL0 was run and the results were compared to a second set of transient data to verify that the model is reasonable under other hydrologic conditions. Model response was compared to water levels collected in well E2D (fig. 20) for 4 days prior to the start of the aquifer test on October 30, 1996. These data show two hydrologic responses—an overall and gradual rise in response to the large amount of precipitation on October 20, 1996, and a daily cyclic fluctuation in response to the normal pumping at STS. To approximate this situation, a seasonal pattern of recharge was determined from the hydrograph-separation programs (fig. 9). The 140 days prior to October 21, 1996 were a period of low ground-water recharge. Drought conditions (defined in next section) were assumed for this period. The beginning of the recharge period was simulated by assuming recharge at half the average annual rate. These conditions only approximate the true events, because the amount and distribution of recharge to the water table due to the storm are unknown. Although the simulated water levels were about 6 ft higher than the actual water levels, the model reproduces the rising trend and the cyclic fluctuations. Although this analysis is not quantitative, it supports model CAL0 as a reasonable representation of the flow system.

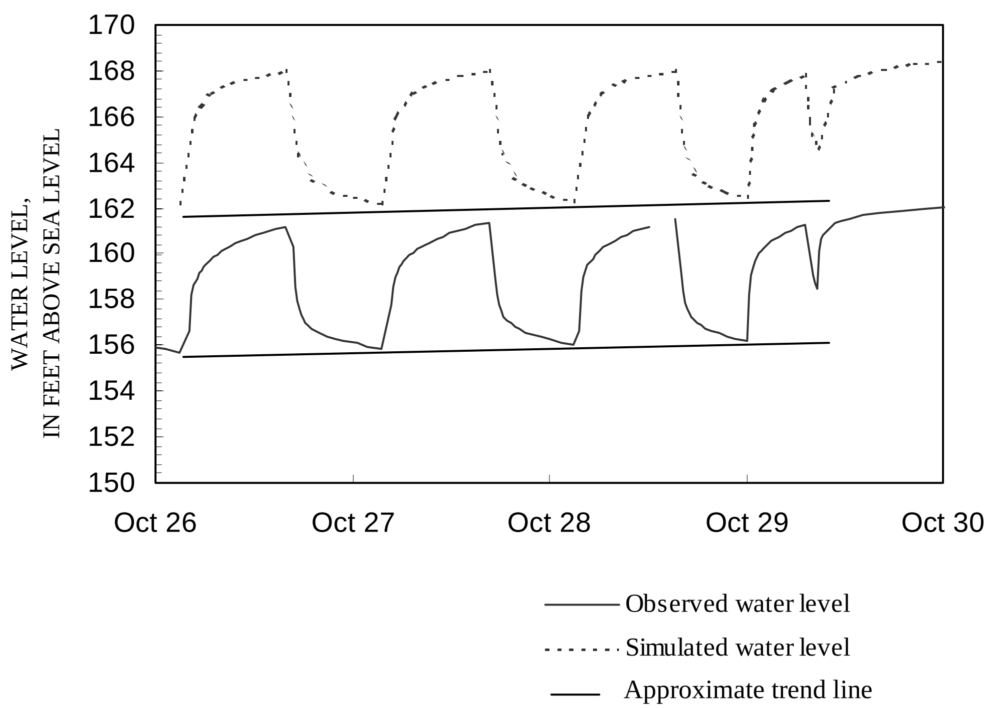


Figure 20. Observed and simulated water levels in observation well E2D, Southbury Training School, Connecticut.

Diagnostic and inferential statistics

The application of statistics to diagnose and analyze ground-water models is well documented (Cooley and Naff, 1990; Hill, 1992, 1994, 1998; Cooley, 1997; Poeter and Hill, 1997; Hill and others, 1998). The discussion of model statistics in this report largely comes from that body of work and follows the methodology outlined by Hill (1998). The use of statistics with UCODE falls into two main categories: (1) the quality of the model calibration and (2) the accuracy of estimated parameters.

For the regression to produce a valid calibrated model, several conditions must be met. Assuming that the model is correct (the true geology and hydrology have been accurately represented by model parameters), the weighted residuals (the difference between the simulated and observed data divided by the variance) must come from a random distribution. Analysis of residuals, both numerical and graphical, can be used to determine how well the model is calibrated.

The calculated error variance (s^2) and the standard error of the regression (the square root of the calculated error variance) (table 8), are quantitative indications of model fit. If the model fit is consistent with the data accuracy in the weighting matrix, these measures equal 1.0. Significant deviations of the calculated error variance or the standard error from 1.0 indicate that the fit is inconsistent with the weighting. Values of the calculated error variance and the standard error are typically greater than 1.0 in practice, reflecting the presence of model error as well as measurement error, which is not represented in the weight matrix.

The model fit also is indicated by the graphical relation of weighted residuals to weighted simulated values. Ideally, weighted residuals are scattered evenly about the line y=0, and their magnitude is not related to the simulated values. Plots were constructed for each alternative model, and weighted simulated residuals seem to be independent of weighted simulated values; therefore, the regression was judged to be valid according to this criterion.

Another measure of model calibration is that the observed values should be reasonably reproduced by the model, as reflected by the correlation coefficient between weighted simulated values and weighted observation that summarizes this relation. This correlation coefficient generally needs to be greater than 0.90 (Hill, 1998). In all alternative models, the correlation coefficient is greater than 0.99 (table 8); therefore, the models are an adequate fit to the data by this criterion.

For a valid regression, the weighted residuals (simulated minus observed values, times the weight) need to be random and uncorrelated or correlated in a way that can be explained by the fitting imposed by the regression. If the weighted residuals are random, independent, and normally distributed, they fall on an approximately straight line in a normal probability graph. The summary statistic R^2_N (table 8) is the correlation coefficient of this line and can be used to test the weighted residuals (Hill, 1992). For the number of observations in this model (60), the critical value for this statistic is 0.962 at the 0.05-percent significance level. Most alternative models have values close to or above the critical value (table 8), indicating that the residuals are nearly normal. The variations present are not considered to be important.

The estimated variance/covariance and correlation matrices (table 10) produced by UCODE can be used to measure the precision and correlation of the parameter estimates. The variance/covariance matrix is based on the optimal parameter values and includes the effects of parameters that are not estimated, such as lakebed hydraulic conductivity (Hill, 1998). In this model, streambed hydraulic conductivity and recharge are highly correlated (table 10), indicating that change in one can be offset by a change in the other to produce an identical model (correlation coefficient = 0.94). Uniqueness of a regression problem can be tested by varying the initial estimates of the parameters. If the regression converges to the same values, the estimated parameter values are likely to be unique. If the regression converges to different values, the parameters are too correlated and are not unique. It is possible that additional pairs of parameters are correlated, but that the correlation is obscured by inaccuracies in the sensitivities calculated by UCODE (Poeter and Hill, 1998). In this study, numerous initial estimates were tried in CAL0, and the final parameter values were judged to be unique.

Table 10. Correlation and variance/covariance matrices for alternative model CAL0

[Correlations are in shaded part of table. Both matrices are symmetric, so the upper or lower diagonal of either matrix is the transpose of the part of the matrix shown. Correlations of parameters with themselves are equal to 1.0; italic type indicates parameter was fixed and not estimated. Kh, horizontal hydraulic conductivity; Kv, vertical hydraulic conductivity

]

Parameter number	*Lakebed Kv* (1)	*Streambed Kv* (2)	Recharge (3)	Specific yield (4)	Specific storage (5)	Kh, zone 1 (6)	Kh, zone 2 (7)	Kv, zone 1 (8)	*Kv, zone 2* (9)
(1)	5.51E+0 0	0.00E+00	0.00E+00	0.00E+00	0.00E+00	0.00E+00	0.00E+00	0.00E+00	0.00E+00
(2)	0.00	3.42E+00	-7.48E-01	1.69E-01	-1.73E-02	4.13E-02	-4.67E-01	1.03E-01	-4.43E+00
(3)	0.00	-.94	.17	-5.05E-02	2.87E-03	-9.66E-03	1.10E-01	-2.16E-02	1.01E+00
(4)	0.00	.49	-.69	.04	1.75E-03	3.89E-03	-4.84E-02	3.40E-03	-5.14E-01
(5)	0.00	-.22	.14	.15	0.00	-1.03E-03	1.11E-03	-2.07E-04	7.57E-03
(6)	0.00	.55	-.57	.52	-.49	0.00	-8.06E-03	-3.83E-04	5.21E-03
(7)	0.00	-.84	.88	-.83	.11	-.66	.09	-1.42E-02	7.23E-01
(8)	0.00	.56	-.48	.10	-.04	-.18	-.45	.01	-2.74E-01
(9)	0.00	-.17	.19	-.22	.11	-.46	.25	.11	1.21E+02

Ground-water models are characteristically nonlinear; that is, the calculated sensitivities (here, derivatives of simulated hydraulic heads and drawdowns with respect to estimated parameters) of the model are related to the parameter value that is being estimated. Thus, changes in recharge, for example, could produce large changes in model head at one value of recharge but could produce small changes in model head at other values of recharge. The nonlinearity of ground-water models poses potential problems for the analysis of confidence intervals. Some models are approximately linear near the optimal parameter values, and Beale's measure (Hill, 1994) can be used to quantify the degree of nonlinearity in a particular model. For model CAL0, if Beale's measure is greater than 0.27, the model is nonlinear; if Beale's measure is less than 0.024, the model is effectively linear. Beale's measure for CAL0 is 0.28; therefore, the model is nonlinear.

ANALYSIS OF UNCERTAINTY OF CONTRIBUTING AREAS

In this study, deterministic and probabilistic contributing areas were estimated using a particle-tracking procedure. Deterministic contributing areas are estimated using the calibrated model and are correct if the geology has been mapped correctly, model parameters correctly represent the geology, ground-water recharge and discharge are well understood, and the calibration data have no errors. The effect of model construction on contributing areas can be analyzed by comparing the contributing areas produced by each alternative.

Estimation of probabilistic contributing areas uses additional information to better define the uncertainty in the calculated contributing area. Probabilistic contributing areas can be calculated using the calibrated model, the variance/covariance matrix generated by the nonlinear regression parameter estimation, and application of a probabilistic simulation technique, such as the Monte Carlo technique. In the type of probabilistic modeling done in this study, the calibration data are not assumed to be free from errors. Errors in the calibration data include (1) the inaccuracy of field measurements; (2) data that do not accurately represent the modeled feature (for example, water levels in wells with very short screens may not accurately represent water levels in very thick model layers; (3) factors external to the phenomena being modeled (such as steady-state water levels affected by a large rainfall that is not simulated); and (4) subsurface variation in hydraulic properties that is smaller in extent than the modeled feature (for example, water levels in a well screened in a clay layer of small extent may not represent water levels in the coarse-grained aquifer surrounding it). The latter type of error in the calibration data alternatively could be attributed to model error, because the model could be refined to accommodate the error (Hill, 1998).

The main difference between the two types of contributing areas is that the deterministic method generates only one piece of information—either an area is inside the contributing area, or it is not. The probabilistic model generates more information: each area has a probability, ranging from 0 to 100 percent, of being within the true contributing area, based on errors in the model calibration data. In this way, the probabilistic method incorporates the uncertainty that is inherent, but often unstated, in deterministic contributing areas and offers water managers more information on which to base decisions. Probabilistic contributing areas generally are larger than deterministic areas because they include areas that have a low probability of contributing water to a well. This information can be used to make decisions, such as to err on the side of caution and manage land use over a greater area, or to collect additional data that could reduce the uncertainty, and the size, of the contributing area.

Particle tracking commonly is used to delineate areas that contribute water to wells (Franke and others, 1998). In this study, a particle-tracking computer program known as MODPATH was used (Pollack, 1994). Hypothetical "particles" are placed in the simulated system and moved in accordance with the ground-water velocities calculated by the model. In this study, one particle was placed in the center of each model cell at the water table. The particles were tracked forward until they discharged from the aquifer system to a pumped well or to Transylvania Brook. Each cell containing a particle that eventually discharged to a cell in which a pumped well was simulated was considered to be in the contributing area of the well. Contributing areas were calculated using calibrated model CAL0, with modifications to pumped well stresses and recharge rates.

Particle tracking also requires an estimate of aquifer porosity. In this study, a uniform porosity was assumed; the porosity estimate does not affect the size, shape, or location of the contributing area, only the time-of-travel calculation. In this study, a uniform porosity of 0.20 was used, based on the low end of a range of typical values for a sand and gravel aquifer (Walton, 1984, p. 19). Using a low value of porosity produces a shorter travel times than if a higher value of porosity was used.

Conditions used to estimate contributing areas

The contributing area simulations were done using the registered diversion rate of 100 gal/min at well PW-1 and 250 gal/min at well PW-3. These wells were simulated in layer 3, which corresponds to the location of their screened interval. Well PW-2 was not simulated because it is classified as an emergency well only.

Recharge was modified from CAL0 to simulate contributing areas under drought conditions to provide a conservative (larger area) estimate of the contributing area to the well. The drought is defined as (1) the maximum historical period with no direct recharge to the valley (180 days; R.L. Melvin, oral commun., 2000), and (2) recharge from upland sources that would be typical of historical drought conditions, as determined from long-term USGS ground-water and surface-water records.

Long-term USGS network well WY-1 (fig. 1) was used for drought analysis because it (1) has a long period of record (since 1944); (2) is in the same physiographic setting as the STS (Southwest Hills; Melvin, 1986); and (3) is in a topographic setting that could be considered representative of the aquifer (valley terrace; Melvin, 1986). The lowest recorded water level for this well was 30.81 ft below land surface on September 24, 1986. The mean water level for this well during 1944-96 is 25.39 ft below land surface. The drought criterion based on WY-1 was defined to be an average 5.42-ft drop in water level in the aquifer from average annual conditions. Upland recharge was varied, using a multiplier that proportionally changed recharge to produce the target drop in water level; 80 percent of the average annual upland recharge produces an average water level drop of 5.79 ft, and 70 percent of the average annual upland recharge produces an average water level drop of 7.08 ft.

Several streamflow measurements were made on Transylvania Brook during a drought in the mid-1960s (table 2). The lowest recorded streamflow at USGS station 01204350 (fig. 3) was 0.15 ft^3/s on September 27, 1965. At no time during the simulation did streamflow losses exceed this flow; therefore, the amount of available streamflow did not limit recharge to the aquifer. On this date, the gain in streamflow between USGS streamflow-gaging stations 01204350 and 01204400 was 0.74 ft^3/s (table 2). A recharge multiplier of 80 percent produced 1.18 ft^3/s of streamflow; a multiplier of 70 percent produced 1.01 ft^3/s. In simulations of contributing areas, 70 percent was chosen as the multiplier for the rate of recharge from uplands because it best met the combined criteria of ground-water level drop and streamflow gains.

The flow system at the end of the 180-day transient simulation was used to calculate contributing areas. This assumption fixes the velocity of the ground water at the end of the drought period, and particles were tracked through this flow system as if the system were at steady state. In all likelihood, the flow system would change because of recharge at or before the end of 180 days. By assuming that drought conditions persist until all particles reach the pumped well, which could take several years, the contributing areas are overestimated.

Deterministic contributing area

The contributing area calculated using CAL0 includes the northern part of the Transylvania Brook valley within the model boundary (fig. 21). The gravel terrace (fig. 21) was outside the model boundary, but it should be considered to be in the contributing area to the wells. This terrace may not be saturated to any significant depth for most of the year, but rainfall infiltrates this area rapidly and then recharges the aquifer. The model is not detailed enough to simulate this ground-water runoff, but any surface contaminant could be introduced into the aquifer from this area. Travel times within the contributing area ranged from 64 (0.18 year) to 1,808 days (4.95 years). The areas with the shortest travel times (less than 1 year) are at the base of the gravel terrace (fig. 21; PW-3) and the northeastern corner of Stibbs Lake (PW-1). Travel from most of the rest of the area is less than about 2 years, with only a few small areas having longer travel times of up to 5 years.

Selected pathlines of ground-water flow to well PW-3 (fig. 22) show the three-dimensional nature of the ground-water flow system at the STS. Pathline A begins on the gravel terrace, descends to layer 2, and remains in this layer until it comes close to the pumped well in layer 3. Pathline B travels in layer 1 perpendicular to the water-table contours through mostly fine deposits. The low vertical hydraulic conductivity of this material prevents the pathline from entering layer 2. When pathline B passes onto the coarse-grained deposits, it descends over a short distance through layer 2 into layer 3 and changes direction toward the pumped well. Pathline C travels through layer 1, descends over a short distance through layer 2 into layer 3, and then changes direction toward the pumped well. Other pathlines on figure 22 show that there is little vertical travel in the fine-grained deposits. Pathlines in layer 1 are perpendicular to the water-table contours until they pass over coarse-grained deposits, where they enter the underlying layer.

Contributing areas from the alternative models were computed, but are not published here; however, the contributing areas from the alternative models were very similar. The main difference was that contributing areas for two of the models (CAL2 and CAL6) were smaller than the contributing area calculated using CAL0. The contributing areas calculated using CAL2 and CAL6 were more horseshoe-shaped than CAL0 and included only the coarse-grained deposits that ring the head of the valley. The addition of bedrock in CAL3 and CAL4 did not produce significantly different contributing areas than CAL0. None of contributing areas produced by alternative models extend south to the sewage-treatment facility (fig. 21). The general similarity among contributing areas indicates that nonuniqueness in model construction does not greatly affect the predicted contributing areas.

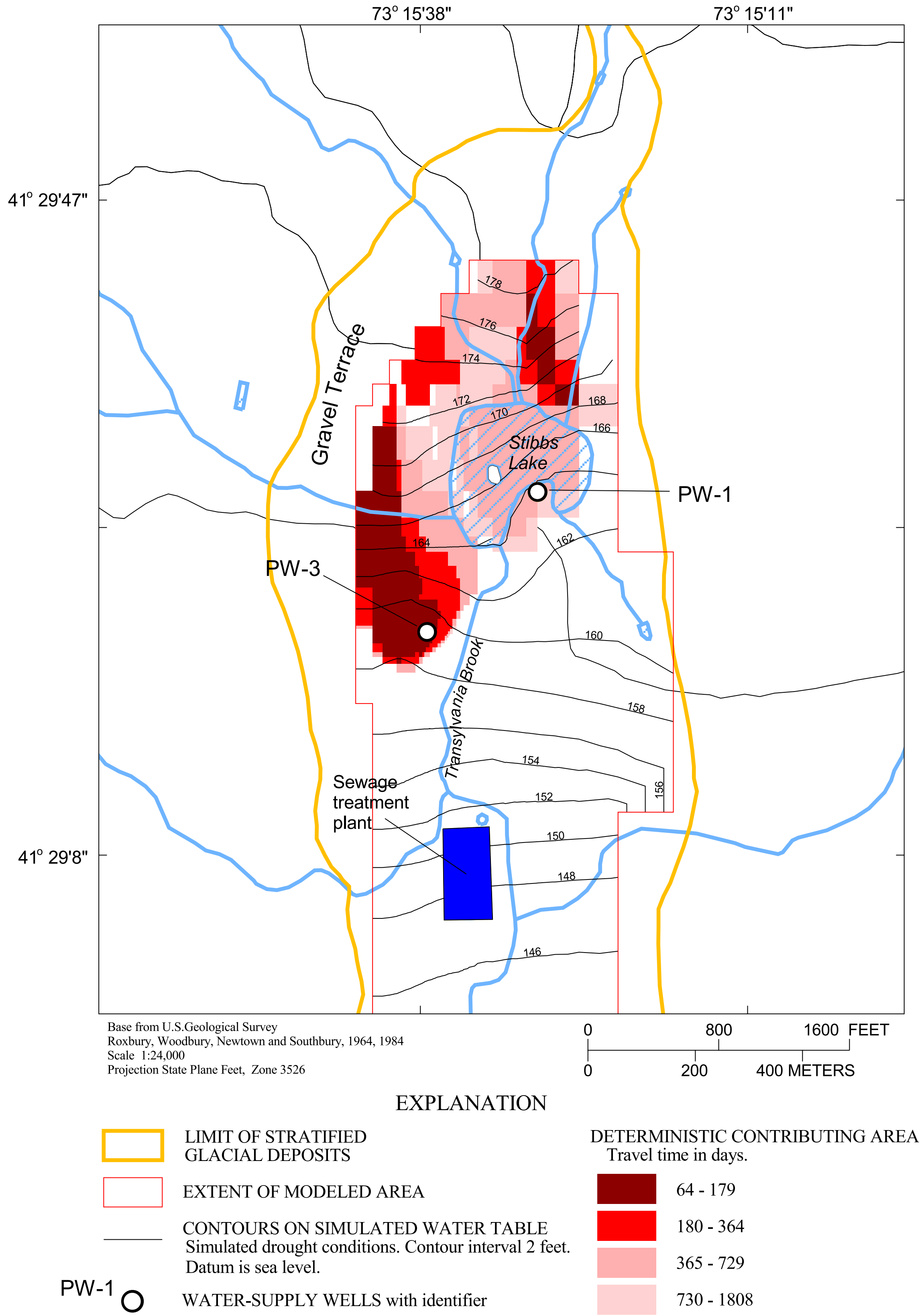


Figure 21. Deterministic contributing areas to wells PW-1 and PW-3, Southbury Training School, Connecticut.

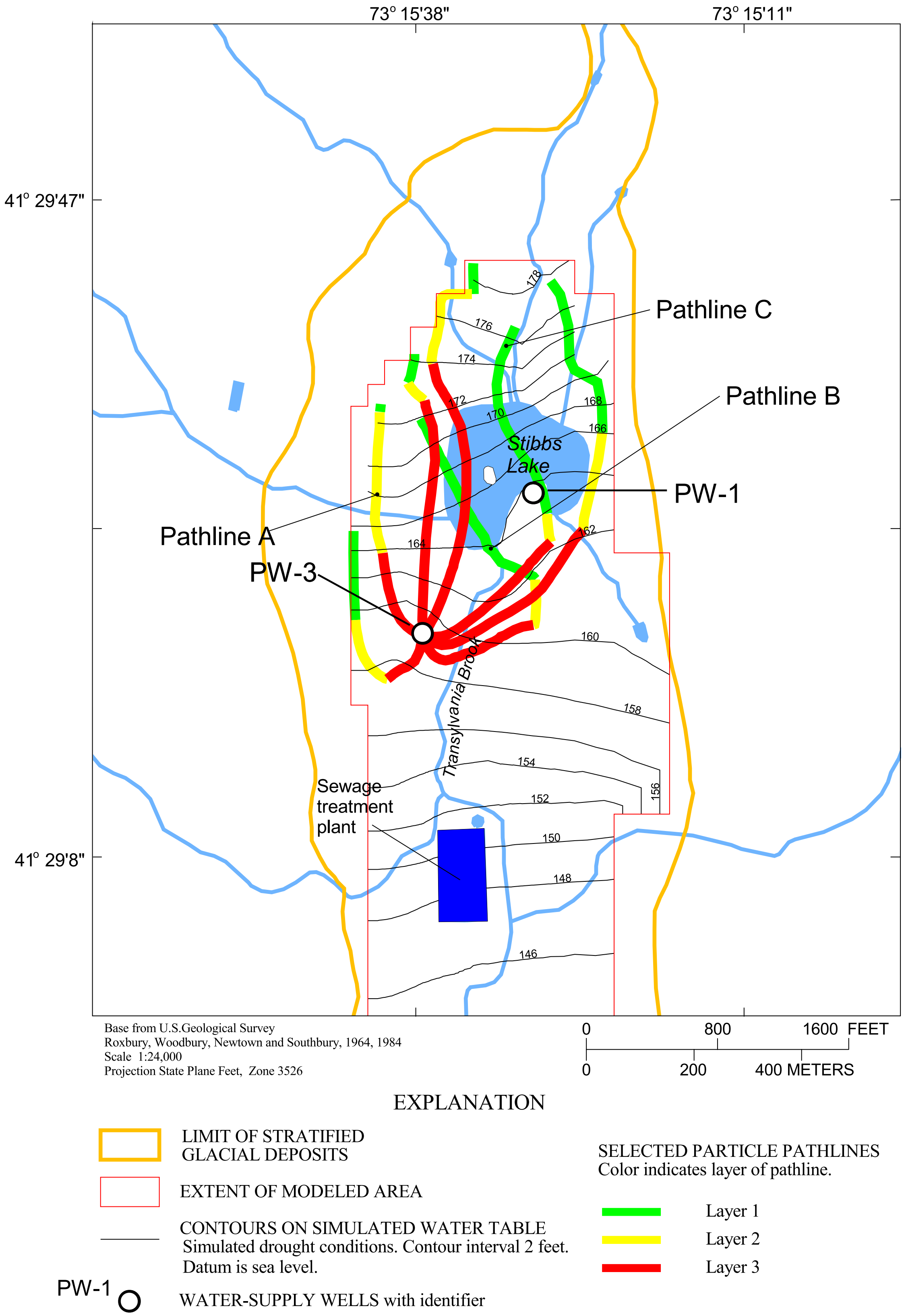


Figure 22. Selected pathlines to well PW-3, Southbury Training School, Connecticut.

Probabilistic contributing areas based on variations in the defined parameters

The uncertainty in the estimated parameters as determined by UCODE is propagated through the analysis of the contributing areas through use of a Monte Carlo analysis. The Monte Carlo simulation involves generating sets of parameter values that, taken as a group, have the same statistical properties as the calibrated model parameters. Using the regression-derived variance/covariance matrix preserves the uncertainty with which parameters are estimated and preserves the correlation among the parameters. The variance/covariance matrix generated by UCODE for all parameters, whether they were estimated or not, is used in this study to quantify the effects of parameter uncertainty on the location and shape of the contributing area.

The Monte Carlo simulation is based on the definition of a random normal variable,

$$\underline{b} = \underline{z}\underline{\underline{\sigma}} + \underline{\mu}, \tag{2}$$

where

$\underline{b}$ is a vector of model parameter values,

$\underline{z}$ is a vector of normally distributed random numbers,

$\underline{\underline{\sigma}}$ is the square root of the variance/covariance matrix, and

$\underline{\mu}$ is a vector of optimal parameter values.

The Monte Carlo analysis was conducted as described by the following steps.

Step 1. Flow model—Estimate optimal parameters of the ground-water-flow model and the variance/covariance matrix using UCODE (already discussed).

Step 2. **Deterministic contributing area—** Determine contributing areas by conducting particle tracking using MODPATH (Pollack, 1994). Place one particle in each model cell. Track the particles forward in time until all the particles have discharged to Transylvania Brook or the pumped wells (PW-3 and PW-1). Flag each cell according to the discharge point of its particle. Combine the flagged cells by discharge locations. Plot contributing areas on a map (fig. 21) (already discussed).

Step 3. Generate parameter sets— Compute the square root of the variance/covariance matrix by taking its Cholesky decomposition matrix (Press and others, 1986). Multiply the resulting matrix by vectors of normally distributed random numbers to generate a large number of sets of model parameter values.

Step 4. Monte Carlo simulation— Run the flow model, substituting one of the generated parameter sets for the optimal parameters. Run the particle-tracking simulation using this flow model. Save the particle discharge locations in a file.

Step 5. Compile results—Sum the number of times, over all simulations, that a particle reaches the pumped well(s) and divide by the number of simulations; thus, each cell will be associated with the number of times, in percent, a particle traveled from the cell to the pumped well.

In this application of the Monte Carlo method, 10,000 parameter sets were generated. The variance of two of the parameters, streambed hydraulic conductivity and hydraulic conductivity of zone 2, were very high. This led to many of those 10,000 parameter sets having unrealistically extreme (both high and low) values for those two parameters. The 95-percent linear confidence intervals for streambed hydraulic conductivity is 0.02 to 51 ft/d and for vertical hydraulic conductivity of zone 2 is 10^{-6} to 1,620 ft/d. These parameter ranges were not considered realistic, so the parameter sets were first conditioned to remove unrealistic sets. Parameter sets were excluded in which the vertical hydraulic conductivity of either the streambed or zone 2 was more than a factor of 10 different from the optimal value for CAL0 (table 9). The value for lakebed hydraulic conductivity was set to 10^{7} ft/d for these simulations. The number of reasonable data sets resulting from the conditioning was 490. No check was made for how the parameter sets affected model fit, and this would contribute to the uncertainty represented by the Monte Carlo analysis; thus, the actual uncertainty is overestimated.

Caution should be used in interpreting probabilistic contributing areas. A low probability at a given location does not mean that the location is likely to be outside the contributing area, rather it indicates that the model calibration data are not adequate to determine if the location is or is not outside the contributing area. The true contributing area is always unknown because one can never have a perfect description of the subsurface. Simulation is the best means of determining contributing areas (Franke and others, 1998) and model calibration data will always be limited; therefore, simulated contributing areas always will have some uncertainty associated with them. A location may be in the true contributing area but not in the simulated contributing area because a hydrologic feature, about which the calibration data do not contain much information, affects the simulation. The reverse is also true—a location may be in the simulated contributing area and not in the true contributing area.

The uncertainty of the simulated contributing area could be reduced by collecting more data or by reformulating the model. If alternative models have been tested and ruled out, as they have in this study, reformulating the model without adding new data may not lead to great reductions in uncertainty. Parts of the probabilistic contributing areas at the STS are underlain by thinly laminated fine-grained deposits that have low probabilities. For reasons cited in this report, measuring representative hydraulic heads in thinly laminated fine-grained deposits may be problematic, and it may be difficult to improve the model calibration by collecting more data in these areas.

The number of times, expressed as a percentage of the total number of simulations (490), that a particle from a cell reached PW-3 ranged from 1 to 100 (fig. 23). Particles from the coarse deposits along the western edge of the valley and from an area underlain by fine-grained deposits beneath and northwest of Stibbs Lake reached PW-3 more than 50 percent of the time. Particles from an area underlain by fine-grained deposits beneath and southwest of Stibbs Lake reached PW-3 less than 50 percent of the time. The probability is low here because of a combination of two factors that are particular to this model—this area is in the contributing area only if the vertical hydraulic conductivity of the fine-grained deposits is high, and most of the Monte Carlo parameter sets do not have high vertical hydraulic conductivity of the fine-grained deposits. If the calibration data better described the vertical hydraulic conductivity, there would be less variation of vertical hydraulic conductivity in the conditioned Monte Carlo data set, and the uncertainty that this area was either inside or outside the contributing area would be lower.

The number of times, expressed as a percentage, that a particle from a cell reached PW-1 ranged from 1 to 90 (fig. 24). Particles reached the pumped well more than 50 percent of the time from the area at the northernmost end of the modeled area. Other areas north of Stibbs Lake had lower probabilities (1 to 50 percent).

The combined probabilistic contributing areas depict the contributing area for both wells pumping simultaneously. Two lobe-shaped areas of high probability (in red on fig. 25) contribute water to wells PW-1 and PW-3. Pathlines in the western lobe (pathline A on fig. 22) go directly from the contributing area to well PW-3. Pathlines in the eastern lobe area go directly from the contributing area to PW-1 (not shown on fig. 22) and across the valley to PW-3 (pathlines B and C on fig. 22). Both lobes contain areas having short travel times and high probabilities. Short travel times are expected in these areas because of the high hydraulic conductivity of the coarse-grained deposits. High probabilities are expected in these areas because the calibration data were collected mainly in the coarse-grained deposits; therefore, the properties of the coarse-grained deposits are well known, and the probabilities are high.

An area of long travel times and low probability separates the above mentioned lobes of the contributing area. Pathlines in this area go through fine-grained deposits, which have lower hydraulic conductivity and less well known hydraulic properties. The deterministic contributing area (figs. 21) includes part of this area, but the probabilistic contributing area does not (fig. 25) because the probabilities are low and the Monte Carlo data set, being a random selection of likely parameter values, did not include the particular combination of parameter values used in the deterministic simulation. Any area in the deterministic contributing area should be considered also to be in the probabilistic contributing area.

The Monte Carlo simulation shows that some areas outside the deterministic contributing area may actually be in the contributing area. The area having probabilities greater than 1 percent is larger than the deterministic contributing area. The deterministic contributing area for wells PW-1 and PW-3 (fig. 21) is roughly similar in size and shape to the area having probabilities greater than 25 percent (fig. 25). The two contributing areas (figs. 21 and 25) are significantly different in two places—southeast of Stibbs Lake and between well PW-3 and Transylvania Brook. These differences may have implications for management of the aquifer. For example, the area southeast of Stibbs Lake is underlain by coarse deposits and may appear to be appropriate for use as a septic tank/leach field site. In the deterministic simulation, such a decision might seem reasonable; however, the probabilistic simulation shows that water from this area may go to well PW-3. The area between well PW-3 and Transylvania Brook also is not in the deterministic area but is in the probabilistic area; therefore, water from Transylvania Brook may go to well PW-3. It is reasonable that these two areas are in the actual contributing area based on the geology of the site, but the calibration data were not adequate to define the contributing area with a higher probability. A greater probability might be achieved with more observation wells in these areas, a longer aquifer test at a higher pumping rate, and (or) an aquifer test within the area in question.

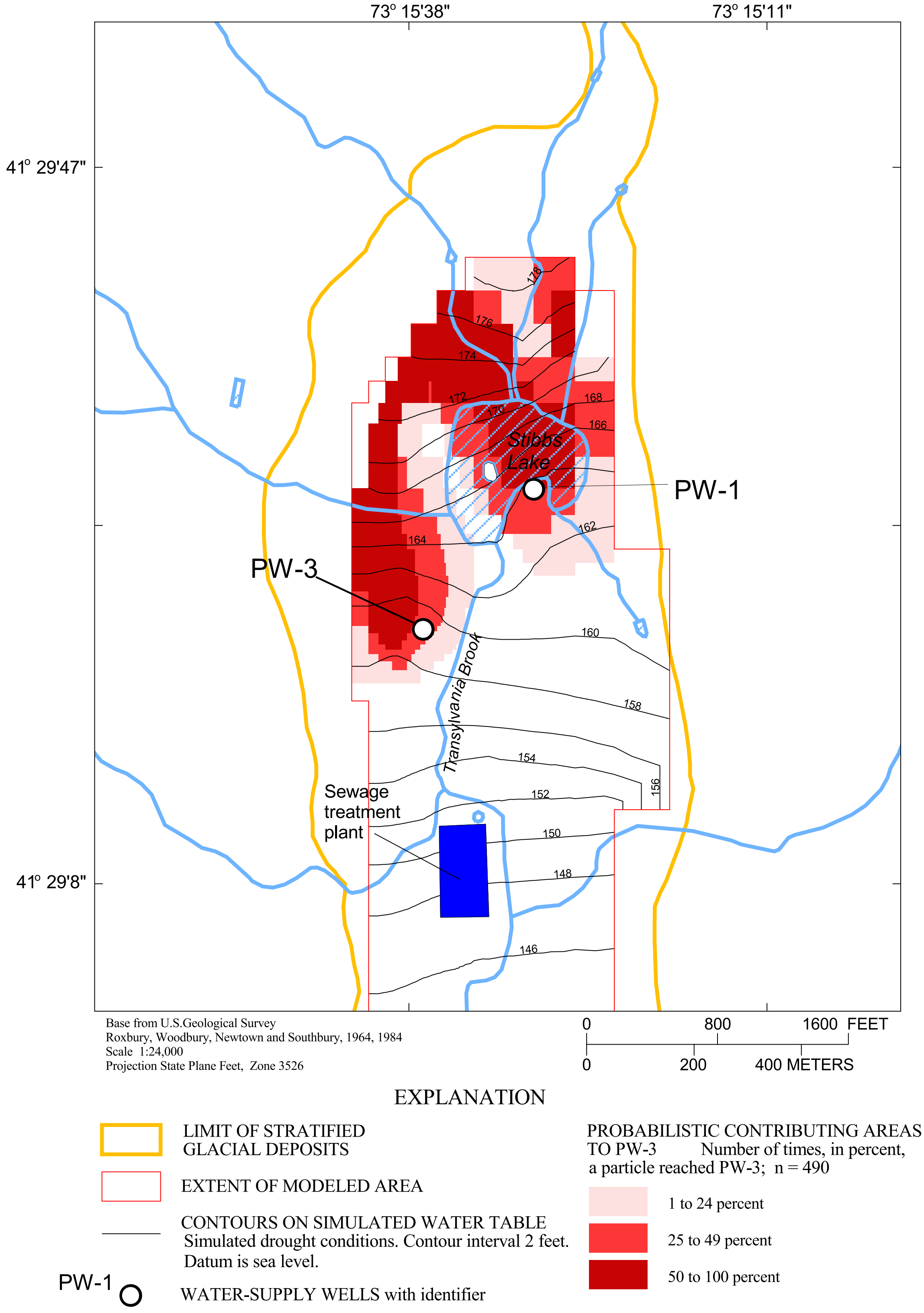


Figure 23. Probabilistic contributing area to well PW-3, Southbury Training School, Connecticut.

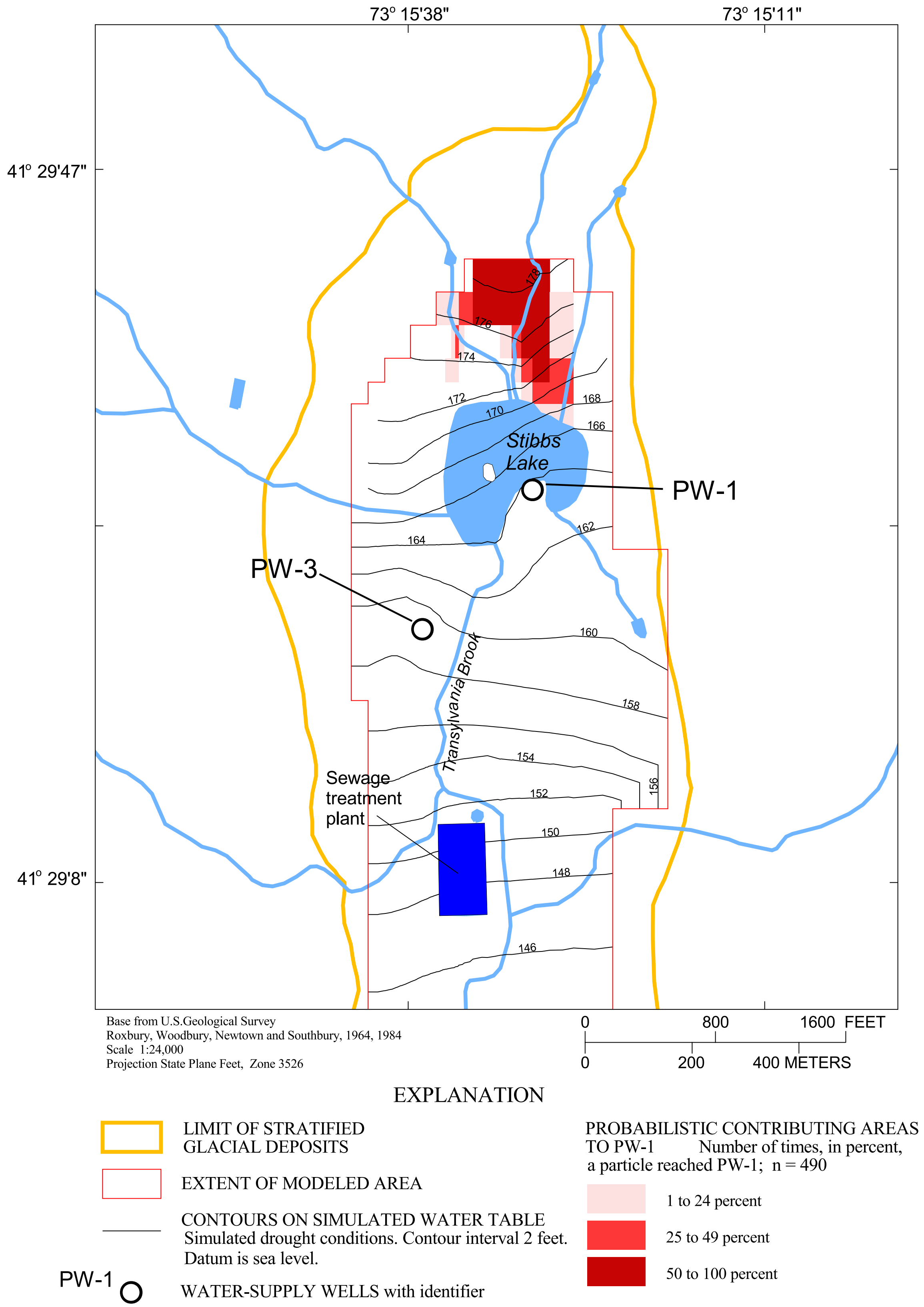


Figure 24. Probabilistic contributing area to well PW-1, Southbury Training School, Connecticut.

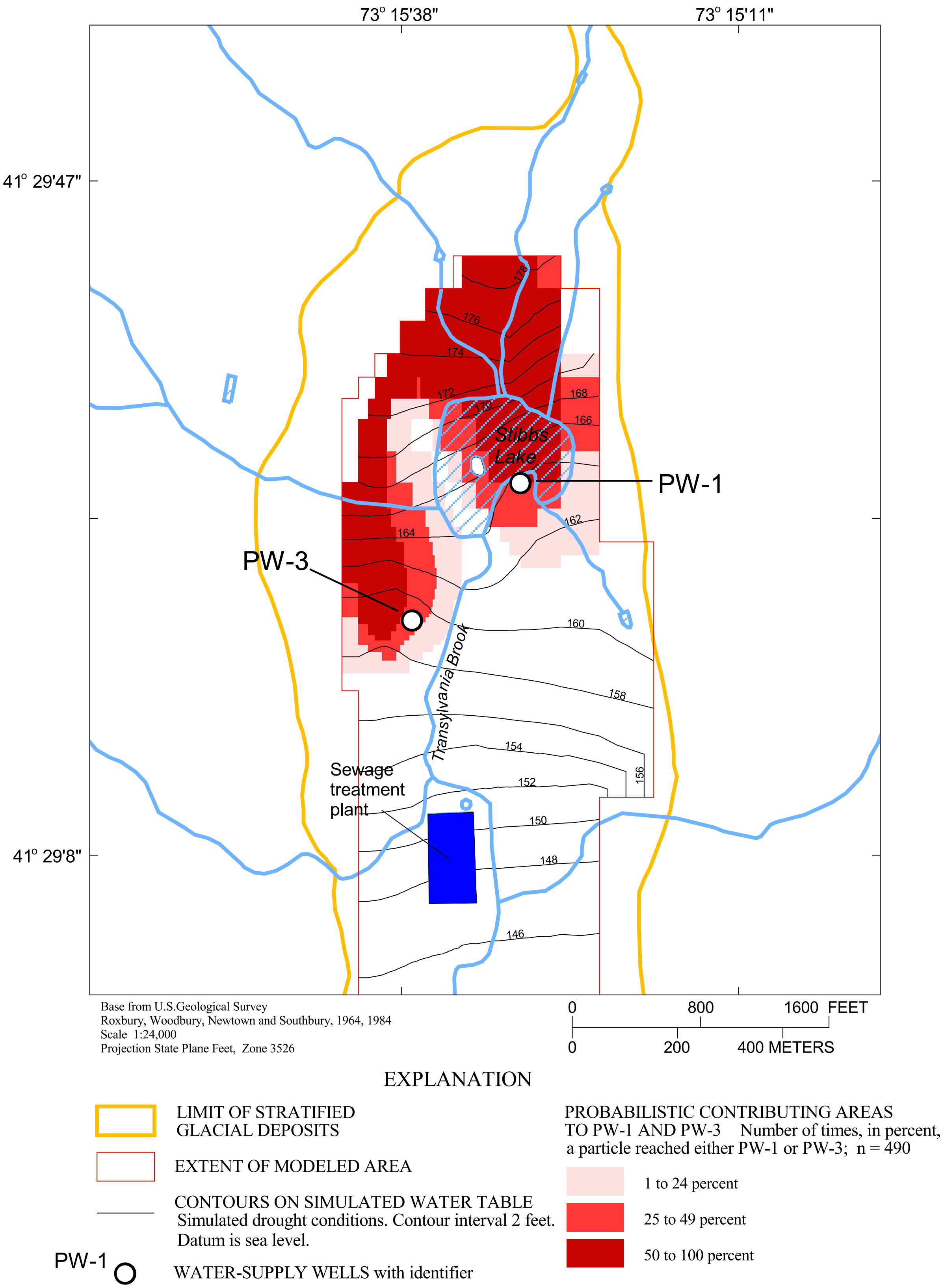


Figure 25. Probabilistic contributing areas to wells PW-1 and PW-3, Southbury Training School, Connecticut.

SUMMARY AND CONCLUSIONS

The Southbury Training School (STS) lies mainly in the Transylvania Brook watershed in western Connecticut and withdraws ground water from stratified glacial deposits in the lower reaches of Transylvania Brook. The stratified deposits are found beneath the relatively (compared to the surrounding uplands) flat surfaces of the valley bottom. The deposits consist of coarse-grained deposits, which were deposited as ice-marginal deltas and fluviodeltaic deposits in close proximity to the retreating Pleistocene ice sheet, and fine-grained deposits, which were deposited in the quiet water conditions of a glacial lake bottom south of the retreating ice. The fine and coarse deposits interfinger and overlie one another in complex ways. The public-supply wells at STS withdraw most of their water from the coarse-grained deposits.

Water flows into the stratified glacial deposits from upland runoff at the valley margins, through infiltration of stream water, and from direct precipitation on the valley bottom. Some water is captured by the water-supply wells at the STS, but most discharges to Transylvania Brook. Analysis of an aquifer test conducted at the primary public-supply well at the STS resulted in preliminary estimates of transmissivity for the stratified glacial deposits that ranged from about 2,900 to 6,500 ft^2/d. The rate of ground-water recharge is estimated to be 24 in/yr over the valley bottom and 9 in/yr over the adjacent upland areas.

Streamflow and ground-water-level data from nearby USGS network sites show the general hydrologic conditions during the aquifer test conducted from October 30 to November 4, 1996. During this period, the annual cycle of ground-water recharge was beginning, which led to generally rising ground-water levels. For the same period, stream levels were generally declining after a large amount of precipitation on October 20, 1996, 10 days before the aquifer test.

Ground-water flow was simulated for the entire watershed of lower Transylvania Brook to understand how topography, precipitation, and geology in the upland parts of the watershed interact to recharge the stratified glacial deposits in the valley. This simulation generated estimates of the distribution of recharge from 10 small drainage basins in the Transylvania Brook watershed that were used in an aquifer-scale simulation. The estimates from each small drainage basin differed on the basis of the drainage characteristics of the basin. Small basins having well-defined stream channels contributed less recharge to the valley than basins having no defined channels because potential ground-water recharge was carried away in the stream channel.

Ground-water flow was simulated in the stratified glacial deposits to define the contributing areas to wells at STS. This simulation was done using a computer code that estimated the parameters of the ground-water-flow model and provided statistical measures of the goodness of fit of the model and the uncertainty associated with model predictions. The ground-water-flow simulation was done using MODFLOW-96, a general purpose, three-dimensional, finite-difference, ground-water-flow model. Contributing areas were computed using MODPATH, a particle-tracking code that works with MODFLOW-96 models. The parameter estimation was done using UCODE, a universal nonlinear regression computer code.

Parameters of the ground-water-flow model are defined in distinct layers and zones. Four zones are used in the simulation—zone 1 is the coarse-grained deposits, zone 2 is the fine-grained deposits, zone 3 is the bedrock, and zone 4 is the fault in the bedrock. Six hydraulic parameters are defined in the model—the horizontal and vertical hydraulic conductivities of zones 1 and 2, the specific yield of the uppermost layer and the specific storage of layers other than the uppermost. Parameters in zones 3 and 4 were not estimated in the simulation. Three additional parameters were used to estimate recharge and the hydraulic conductivities of the lake bed and streambed deposits.

Regression modeling was done using aquifer-test data collected at STS from October 30-November 4, 1996. The data were first corrected for outside influences, which included the effects of long-term pumping at the STS, the effects of generally rising ground-water levels, and the effects of shutting off the pump at the STS prior to the test. Seven alternative models were posed, each representing the ground-water-flow system in slightly different but realistic ways. On the basis of the statistical measures of model fit and using the available data, one model was chosen as being the most representative of the ground-water flow system. The standard error of the regression for the chosen model is 2.27. The correlation coefficient between the weighted residuals from the regression and a normal distribution is 0.955, indicating that the residuals are nearly normal.

Parameter values estimated during the simulation are as follows: horizontal hydraulic conductivity of coarse-grained deposits, 154 feet per day; vertical hydraulic conductivity of coarse-grained deposits, 0.83 feet per day; horizontal hydraulic conductivity of fine-grained deposits, 29 feet per day; specific yield, 0.007; specific storage, 1.6E-05. Average annual recharge was estimated using the watershed-scale model with no parameter estimation and was determined to be 24 inches per year in the valley areas and 9 inches per year in the upland areas.

The parameter estimates produced in the model are similar to expected values, with two exceptions. The estimated specific yield of the stratified glacial deposits (0.007) is lower than expected; this could be caused by the layered nature of the deposits. The recharge estimate produced by the model was also lower—about 32 percent of the annual average rate. This could be caused by the timing of the aquifer test with respect to the annual cycle of ground-water recharge and by some expected recharge going to parts of the flow system that were not simulated. The model reproduced the aquifer-test data accurately, as measured by the correlation coefficient (0.999) between the weighted simulated values and weighted observed values. The model also reproduced the general rise in ground-water levels caused by ground-water recharge and the cyclic fluctuations caused by pumping prior to the aquifer test.

The simulated water table shows the effect of the fine-grained deposits on the shape of the contributing area to the well. The contributing area shows that most water enters the stratified glacial deposits through the coarse-grained deposits that ring the head of the lower Transylvania Brook valley. Some of these deposits are not saturated throughout the year and could not be simulated in the model; however, because the primary public-supply well receives most of its water from this area, the unsaturated deposits should be considered to be in the contributing area. Some travel times for ground water are less than 1 year; travel times for most of the contributing area are less than 2 years.

A Monte Carlo simulation was done, using the variance/covariance matrix generated by the regression model, to estimate the probabilities associated with the contributing area to the wells. These probabilities arise from uncertainty in model, which in turn arises from the adequacy of the data available to comprehensively describe the ground-water-flow system. The Monte Carlo data sets were conditioned to remove unrealistic parameter sets. Probabilities in the contributing area range from 1 to 100 percent, and the highest probabilities (greater than 50 percent) are in the coarse-grained deposits that ring the head of the valley. The deterministic contributing area corresponds to the areas having probabilities of greater than 25 percent.

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Appendix 1. Geohydrologic data for selected wells, Transylvania Brook watershed, Connecticut

[-, not available; 112DFSF, stratified glacial deposits; 112TILL, glacial till; 231 SDMN, sedimentary rock; NCBC, noncarbonate crystalline rock; SB, Southbury]

USGS identifier	USGS aquifer code	Altitude of top of rock (feet above sea level)	Altitude of water level (feet above sea level)	Specific capacity (gallons per minute per foot)
SB 46	231SDMN	185.00	185.00	0.01
SB 47	300NCBC	45.00	135.00	0.05
SB 48	231SDMN	55.00	153.00	0.19
SB 50	300NCBC	295.00	285.00	0.01
SB 51	300NCBC	285.00	285.00	0.01
SB 52	300NCBC	125.00	149.00	0.01
SB 56	231SDMN	42.00	135.00	0.01
SB 59	300NCBC	35.00	126.00	0.23
SB 60	300NCBC	447.00	350.00	0.01
SB 61	300NCBC	400.00	430.00	0.00
SB 63	231SDMN	65.00	165.00	0.83
SB 69	300NCBC	315.00	275.00	0.01
SB 70	231SDMN	25.00	155.00	0.45
SB 74	231SDMN	139.00	137.00	0.28
SB 75	231SDMN	95.00	130.00	20.00
SB 76	300NCBC	37.00	181.00	0.11
SB 81	300NCBC	39.00	153.00	0.00
SB 84	300NCBC	15.00	153.00	0.20
SB 86	300NCBC	5.00	183.00	0.11
SB 88	300NCBC	289.00	279.00	0.01
SB 89	300NCBC	343.00	338.00	0.01
SB 90	231SDMN	335.00	379.00	0.20
SB 91	300NCBC	265.00	267.00	0.02
SB 92	231SDMN	358.00	345.00	0.04
SB 94	231SDMN	357.00	475.00	0.09
SB 95	231SDMN	280.00	437.00	0.01
SB 96	231SDMN	375.00	435.00	0.05
SB 97	300NCBC	250.00	420.00	0.01
SB 98	300NCBC	428.00	423.00	0.04