Delineation of Areas Contributing Recharge to Selected Public-Supply Wells in Glacial Valley-Fill and Wetland Settings, Rhode Island

By Paul J. Friesz

In cooperation with the
Rhode Island Department of Health

Scientific Investigations Report 2004-5070

U.S. Department of the Interior
U.S. Geological Survey

U.S. Department of the Interior
Gale A. Norton, Secretary

U.S. Geological Survey
Charles G. Groat, Director

U.S. Geological Survey, Reston, Virginia: 2004
For sale by U.S. Geological Survey, Information Services
Box 25286, Denver Federal Center
Denver, CO 80225

For more information about the USGS and its products:
Telephone: 1-888-ASK-USGS
World Wide Web: http://www.usgs.gov/

Suggested citation:
Friesz, P.J., 2004, Delineation of areas contributing recharge to selected public-supply wells in glacial valley-fill and wetland settings, Rhode Island: U.S. Geological Survey Scientific Investigations Report 2004-5070, 57 p.

Contents

Figures

Tables

Conversion Factors and Datums

Multiply	By	To obtain
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
foot (ft)	0.3048	meters (m)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
gallons per minute (gal/min)	3.78544	liters per minute (L/min)
gram per cubic centimeter (g/cm^3)	62.4220	pounds per cubic foot (lb/ft^3)
inch (in.)	25.4	millimeter (mm)
inches per year (in/yr)	25.4	millimeter per year (mm/yr)
miles (mi)	1.609	kilometers (km)
square miles (mi^2)	2.59	square kilometers (km^2)

Vertical Coordinate Information (altitude): Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD 29), formerly called the Sea Level Datum of 1929.

Horizontal Coordinate Information (location): Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

Delineation of Areas Contributing Recharge to Selected Public-Supply Wells in Glacial Valley-Fill and Wetland Settings, Rhode Island

By Paul J. Friesz

Abstract

Areas contributing recharge and sources of water to one proposed and seven present public-supply wells, screened in sand and gravel deposits and clustered in three study areas, were determined on the basis of calibrated, steady-state ground-water-flow models representing average hydrologic conditions. The area contributing recharge to a well is defined as the surface area where water recharges the ground water and then flows toward and discharges to the well.

In Cumberland and Lincoln, public-supply well fields on opposite sides of the Blackstone River are in a narrow valley bordered by steep hillslopes. Ground-water-level and river-stage measurements indicated that river water was infiltrating the aquifer and flowing toward the wells during pumping conditions. Simulated areas contributing recharge to the Cumberland well field operating alone for both average (324 gallons per minute) and maximum (1,000 gallons per minute) pumping rates extend on both sides of the river to the lateral model boundaries, which is the contact between the valley and uplands. The area contributing recharge at the average pumping rate is about 0.05 square mile and the well field derives 72 percent of pumped water from upland runoff. At the maximum pumping rate, the area contributing recharge extends farther up and down the valley to 0.12 square mile and the primary source of water to the well field was infiltrated river water (53 percent). Upland areas draining toward the areas contributing recharge encompass 0.58 and 0.66 square mile for the average and maximum rates, respectively. By incorporating the backup Lincoln well-field withdrawals (2,083 gallons per minute) into the model, the area contributing recharge to the Cumberland well field operating at its maximum rate is reduced to 0.08 square mile; part of the simulated area which contributes recharge to the Cumberland well field when it is operating alone contributes instead to the Lincoln well field when both well fields are pumped. The Cumberland well field compensates by increasing the percentage of water it withdraws from the river by 11 percent. The upland area draining toward the Cumberland contributing area is 0.55 square mile. The area contributing recharge to the Lincoln well field is 0.08 square mile and infiltrated river water contributes 88 percent of the total water; the upland area draining toward the contributing area is 0.34 square mile.

In North Smithfield, a public-supply well in a valley-fill setting is close to Trout Brook Pond, which is an extension of the Lower Slatersville Reservoir. A comparison of water levels from the pond and underlying sediments indicates that water is not infiltrated from Trout Brook Pond when the supply well is pumped at its maximum rate of 200 gallons per minute. Simulated areas contributing recharge for the maximum pumping rate and for the estimated maximum yield, 500 gallons per minute, of a proposed replacement well extend to the ground-water divides on both sides of Trout Brook Pond. For the 200 gallons-per-minute rate, the area contributing recharge is 0.23 square mile; the well derives almost all of its water from intercepted ground water that normally discharges to surface-water bodies. For the pumping rate of 500 gallons per minute, the area contributing recharge is 0.45 square mile. The increased pumping rate is balanced by additional intercepted ground water and by inducing 25 percent of the total withdrawn water from surface water.

In Westerly, one public-supply well is in a watershed where the primarily hydrologic feature is a wetland. Water levels in piezometers surrounding the well site indicated a downward vertical gradient and the potential for water in the wetland to infiltrate the underlying aquifer. The simulated area contributing recharge for the average pumping rate (240 gallons per minute) and for the maximum pumping rate (700 gallons per minute) extends to the surrounding uplands (surficial materials not covered by the wetlands) and to a ground-water divide separating the watersheds of the Pawcatuck River and Block Island Sound. For the average pumping rate, the upland area contributing recharge is 0.13 square mile and contributes 46 percent of the total water withdrawn from the well; the remaining water withdrawn from the well is derived from the wetlands or indirectly from the uplands through the wetland from an area of 0.54 square mile. For the maximum pumping rate, the area contributing recharge in the uplands is 0.16 square mile and supplies 21 percent of the total water pumped; the

remaining water is from the wetland or indirectly from the uplands through the wetland from an area of 1.27 square miles. Thus, the primary source of water withdrawn from the well originates in the wetland or discharges to the wetland from the surrounding uplands before infiltration.

Hydrologic factors that most affected the simulated areas contributing recharge to wells in valley-fill and wetland settings were recharge rates, the hydraulic connection between surface water and the aquifer, and the location of ground-water divides. In a narrow valley setting, however, with major water sources near a well, the percentage of source water withdrawn from the well was affected by these hydrologic factors, but not the area contributing recharge. A well in the vicinity of a surface-water source may not always induce flow, even if the ground and surface waters are well connected, because the amount of induced flow can also depend on the pumping rate and the quantity of ground water that the well can intercept. In a wetland setting, recharge rates were a major factor in the size of the area contributing recharge in the upland areas, but the area contributing water in the wetland showed only slight change because the connection between the wetland and aquifer was the controlling factor.

Introduction

Accurate delineation of areas contributing recharge to public-supply wells is an essential component of State, local, and Federal strategies for the protection of drinking-water supplies from contamination (U.S. Environmental Protection Agency, 1991). Identification of the area contributing recharge and the source of water to a supply well allows for an assessment of the susceptibility and risk of the water supply to contamination. The Rhode Island Department of Environmental Management (RIDEM), through its Wellhead Protection Program, has determined areas contributing recharge to most public-supply wells in Rhode Island, but the RIDEM and the Rhode Island Department of Health (RIDOH) do not have a high degree of confidence in some of the contributing-area delineations to wells in complex hydrologic settings. Numerical ground-water-flow modeling, coupled with a particle-tracking technique, is a more advanced method for delineating areas contributing recharge than the analytical methods that have previously been used for this purpose. Numerical models can represent a variety of geologic and hydrologic features that may affect the size and location of the area contributing recharge to a well but are difficult to include in analytical methods, such as ground-water systems with irregular geometry and complex lithology, or the interaction between individual pumping wells and hydrologic features such as surface-water bodies. Information provided by a numerical model on the source of water to a well can also aid in the protection of public health in Rhode Island.

Each of the public-supply wells included in this study is screened in sand and gravel deposits in valley-fill and wetland settings near surface-water sources. The degree of hydraulic connection between the surface water and the underlying aquifer can affect the tendency for pumping to induce infiltration of surface water into the aquifer. The quantity of water derived from surface water, if any, can affect the size and shape of the area of the aquifer contributing water to the well. The U.S. Geological Survey (USGS), in cooperation with the RIDOH, began a 3-year study in 2000 to increase understanding of the geohydrology and of the important hydrologic factors required to properly delineate areas contributing recharge and the sources of water to seven present supply wells and one proposed supply well clustered in three study areas in four Rhode Island towns (fig. 1).

Purpose and Scope

This report describes the geohydrology and the areas contributing recharge and sources of water to eight public-supply wells in three study areas: the Cumberland and Lincoln study area includes two Cumberland wells and three Lincoln wells, the North Smithfield study area includes one present and one proposed well, and the Westerly study area includes one well. Numerical models were developed and calibrated for each study area on the basis of geologic and hydrologic data collected during this and previous investigations. Maps depict the simulated area contributing recharge to supply wells in glacial valley-fill and wetland settings for selected pumping rates and steady-state, average hydrologic conditions. Maps also show the effects of selected hydrologic properties on the delineated area contributing recharge to the supply wells. The area contributing recharge to a public-supply well is defined as the surface area where water recharges the ground water and then flows toward and discharges to the well (Reilly and Pollock, 1993).

Description of the Study Areas and Previous Investigations

The three study areas are in northeast, north-central, and southwest Rhode Island in four towns (fig. 1). The climate is humid and temperate with an average annual temperature of about 50°F and average annual precipitation of about 48 in. over the northern and 50 in. over the southern study areas (National Oceanic and Atmospheric Administration, 2002).

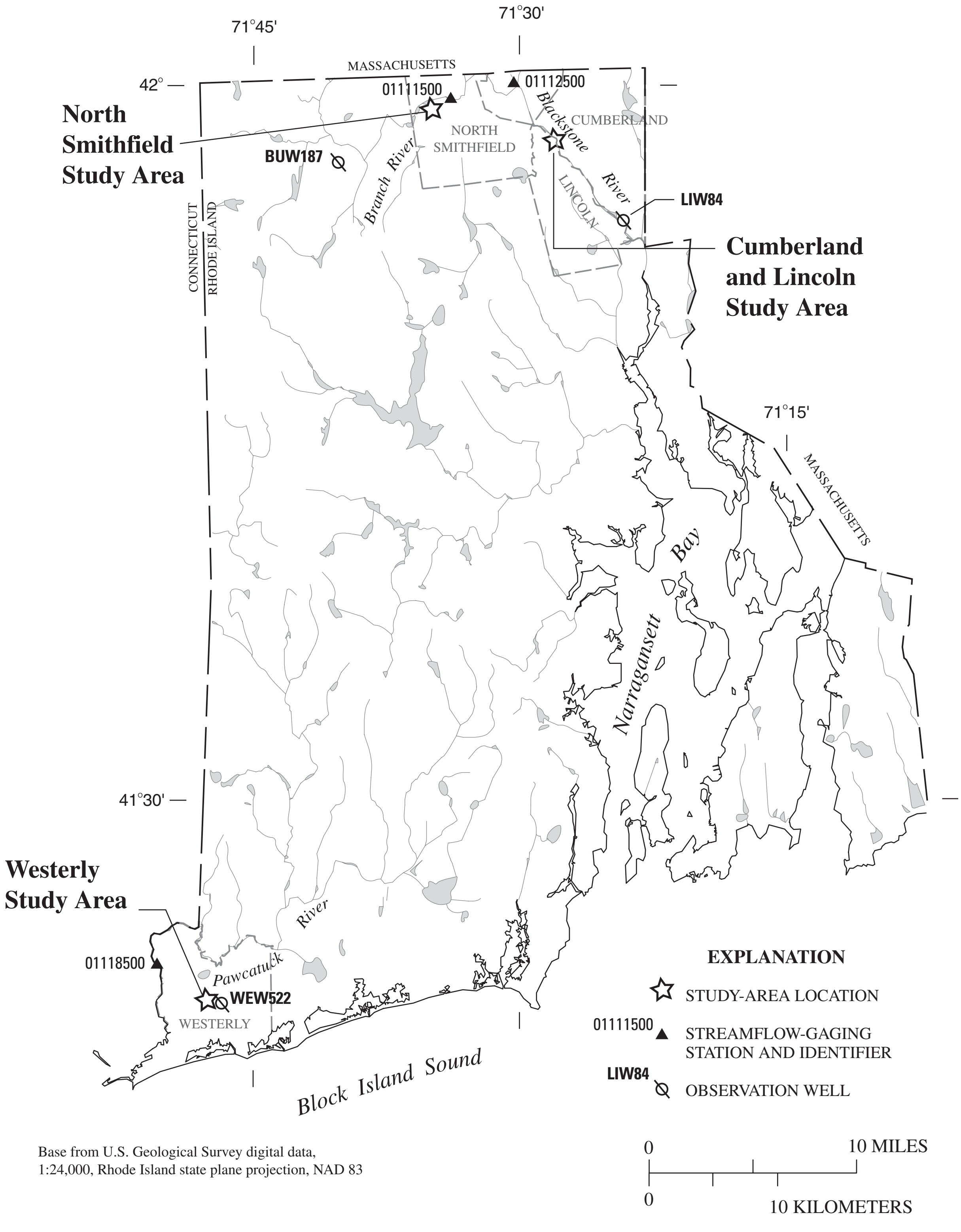


Figure 1. Location of study areas in Rhode Island and selected U.S. Geological Survey long-term network streamflow-gaging stations and observation wells.

Ground water in the study areas is stored and transmitted in surficial sediments of glacial origin—till and stratified deposits—and in the underlying bedrock. A thin, discontinuous layer of till deposited directly on the bedrock by glaciers is composed of a poorly sorted mixture of sediments ranging in size from clay to boulders. Stratified deposits consist of well-sorted, layered sediments ranging in size from clay to gravel deposited by glacial meltwater. Stratified deposits overlie the till in the valleys and also the lowlands in coastal areas. In the Westerly study area, mixed sediments of till and stratified deposits also are present in a moraine, which formed at the ice margin when the retreating glacier paused for a period of time.

The direction of ground-water flow is generally from the till uplands toward discharge areas in the valleys and the coast. The public-supply wells are screened in coarse-grained stratified deposits composed of sand and gravel. Sand and gravel deposits are the primary aquifers in Rhode Island because of the higher storage and transmissive properties of these deposits compared to the other geologic units.

The Cumberland and Lincoln study area and the North Smithfield study area are characterized by narrow, valley-fill settings bordered by steep hillslopes. The two supply wells in Cumberland and the three supply wells in Lincoln are on opposite sides of the Blackstone River. The North Smithfield well site is near the Slatersville Reservoirs, which were created by dams built across the Branch River. Previous investigations by the USGS in the Rhode Island areas of the Blackstone River Watershed (Johnston and Dickerman, 1974a) and the Branch River Watershed (Johnston and Dickerman, 1974b) analyzed the ground-water resources and water quality. These investigations resulted in published and unpublished maps at a regional scale of the bedrock-surface altitude and transmissivity of the surficial sediments, and information helpful in determining the hydraulic connection between the rivers and aquifer. Site-specific information from these investigations was available for well sites in both study areas.

The Westerly study area is in the broad lowlands of the lower Pawcatuck River Watershed and the coast and is characterized by a wetland of large areal extent. A ground-water and water-quality investigation of the lower Pawcatuck River Watershed by the USGS (Gonthier and others, 1974) resulted in regional maps of the bedrock-surface altitude, transmissivity and water-table contours of the surficial deposits, and information concerning the riverbed properties. Site-specific information from this investigation was not available for the well site.

Numerical Modeling Strategy

Finite-difference numerical ground-water-flow models based on the computer code MODFLOW-96 (Harbaugh and McDonald, 1996), each capable of simulating the response of the ground-water system to public-supply-well withdrawals, were developed for each of the three study areas. Model-input parameters were assigned from the literature or calibrated to water levels measured during steady-state, nonpumping conditions and to drawdowns from aquifer tests. Models were calibrated by adjusting model-input parameters within reasonable ranges until the simulated water levels approximated measured water levels at observation wells. Improvements in model-simulation results were achieved by minimizing the differences, or residuals, between simulated and measured water levels. Model-input values were changed for a group of cells, or zones, rather than on a cell-by-cell basis.

Areas contributing recharge and the sources of water to the public-supply wells were determined by use of the particle-tracking program MODPATH (Pollock, 1994). The particle-tracking program calculates ground-water-flow paths and traveltimes based on the head distribution computed with the ground-water-flow-model simulation. Areas contributing recharge were delineated by forward tracking of particles from the top faces of model cells in recharging areas to the discharging wells. Particles were allowed to pass through model cells with weak sinks; weak sinks remove only a part of the water that flows into the cell. Upland areas that drain toward the area contributing recharge but were not simulated directly in the model were delineated on the basis of maps of land-surface contours and the watershed boundary.

The development of a ground-water-flow model for each of the three study areas required that the geometry and hydraulic properties of the ground-water system and fluxes into and out of the model be quantified. Hydraulic properties of stratified glacial sediments and bedrock are discussed below; the thin veneer of till covering the bedrock was not simulated as a separate unit from the bedrock. Hydraulic properties of wetlands and a moraine in the Westerly study area are discussed in that section of the report.

Bedrock surface: The altitude of the bedrock surface in each of the study areas was revised from published and unpublished maps prepared in previous USGS watershed investigations. Lithologic logs from wells and borings completed for municipalities and private landowners since the previous watershed investigations and collection of sediment samples from wells and borings during this study were used to revise bedrock-surface contours.

Hydraulic conductivity: Horizontal hydraulic-conductivity values were assigned to stratified sediments in the models on the basis of lithologic logs and aquifer-test data. Values determined from lithologic logs were based on the relation between horizontal hydraulic conductivity and grain size determined by Rosenshein and others (1968) and modified by Dickerman (1984) for numerous aquifer-test results in Rhode Island. The authors developed this relation by correlating transmissivity values from the aquifer tests to lithology.

The general distribution of horizontal hydraulic conductivity in the models was based on geologic sections that were constructed for each study area from lithologic logs, surficial geology, and an understanding of the geologic processes that formed the stratified glacial deposits. The stratified sediments were deposited by glacial meltwater in sequences during retreat of the last glacial ice sheet (Stone and others, 1992). Each depositional sequence consists of sorted, layered sediments laid down by meltwater in ice-contact, deltaic, and lacustrine environments; sediment size generally decreases and sorting increases with distance from the ice margin. Sediments were grouped into three units on the basis of the dominant grain sizes—sand and gravel; sand; and fine-grained sediments composed of very fine sand, silt, and clay—to construct the generalized geologic sections.

An anisotropic ratio of 10:1 for horizontal to vertical hydraulic conductivity was used with horizontal-conductivity values of 50 ft/d or greater based on average values from aquifer-test analyses in Rhode Island (Dickerman, 1984). An anisotropic ratio of 50:1 was used with horizontal hydraulic-conductivity values less than 50 ft/d because of the high percentage of silt and clay in fine-grained sediments.

Bedrock was assigned a horizontal hydraulic-conductivity value of 0.5 ft/d, representative of crystalline bedrock (Randall and others, 1966), and an anisotropic ratio of horizontal to vertical hydraulic conductivity of 1:1 because of the unstratified nature of bedrock.

Surface water-ground water interaction: Surface-water bodies were simulated in the models as head-dependent flux boundaries. This boundary type simulates flow between the surface-water body and the underlying aquifer as a function of the head gradient and conductance. The conductance term incorporates the geometry and the vertical hydraulic conductivity of the bed sediments of the surface-water body. Values of vertical hydraulic conductivity initially were determined according to reported values in the literature and then changed within the range of reasonable values during model calibration, if necessary, to improve the fit between simulated and measured water levels in observation wells adjacent to the surface-water body. Reported vertical hydraulic-conductivity values ranged from 0.1 to 17 ft/d for bed sediments in Rhode Island (Rosenshein and others, 1968; Gonthier and others, 1974; Johnston and Dickerman, 1974a) and central Massachusetts (Lapham, 1989; de Lima, 1991; Friesz and Church, 2001). The vertical hydraulic conductivity of coarse-grained sediments in these studies typically ranged from 1 to 3 ft/d and from 0.1 to 0.7 ft/d for fine-grained sediments. The altitude of the surface-water stage and the geometry of the bed sediments were determined from water-level measurements, water-depth and sediment-thickness profiles, topographic maps, field observations, and values reported in the literature.

Recharge rates: Recharge rates were estimated based on mean annual runoff data. Randall (1996) constructed lines of equal mean annual runoff for the glaciated Northeast United States on the basis of records from streamflow-gaging stations during the 30-year period 1951–80. Annual runoff, equivalent to precipitation minus evapotranspiration over a watershed, provides an estimate of maximum water available for recharge. Actual recharge for direct infiltration of precipitation into the stratified sediments was assumed to equal the maximum water available for recharge because overland runoff for these permeable sediments is considered minimal. For most cases, recharge to stratified sediments from upland till and bedrock runoff was assumed to be equivalent to water available for recharge. If topographical maps indicated that tributary streams drained the uplands and discharged to valley streams, or if field observations indicated that upland runoff did not fully recharge the stratified sediments, then a fraction of the water available for recharge was used.

Storage: Storage values were specified for transient simulations, which were used only to calibrate simulated drawdown values to drawdowns from aquifer tests. A specific-yield value of 0.25 for the highest active layer of the models was used for the stratified sediments; this value approximates the mean value determined by Allen and others (1963) in the upper Pawcatuck River Basin. A specific-yield value of 0.005 for crystalline bedrock was based on Randall and others (1966). A specific-storage value of 1.0×10^{-5} ft^{-1} determined by Moench and others (2000) for Cape Cod sediments was specified for model layers beneath the highest active layer for both stratified sediments and crystalline bedrock.

Delineation of Areas Contributing Recharge to Public-Supply Wells

The determination of areas contributing recharge and sources of water to the public-supply wells in Rhode Island was based on model simulations of steady-state, average hydrologic conditions. The size of the area contributing recharge for a particular pumping rate is related to the recharge rate from direct precipitation and, if applicable, the quantity of water derived from other sources, such as infiltration of surface water. Each of the three study areas includes a potential surface-water source in close proximity to at least one public-supply well: a river in the Cumberland and Lincoln study area, a reservoir in the North Smithfield study area, and a wetland in the Westerly study area. The accurate delineation of the areas contributing recharge to the wells required an understanding of the hydraulic connections between the surface-water sources and the aquifer for each study area.

Cumberland and Lincoln Study Area

The Cumberland and Lincoln study area includes two well fields adjacent to the Blackstone River in northeastern Rhode Island (fig. 1 and 2). The study area is characterized by a narrow, valley-fill setting bordered by steep hillslopes. Dams built across the river, originally to supply water to mills, define the upstream and downstream extent of the study area and the adjacent uplands form its lateral boundary.

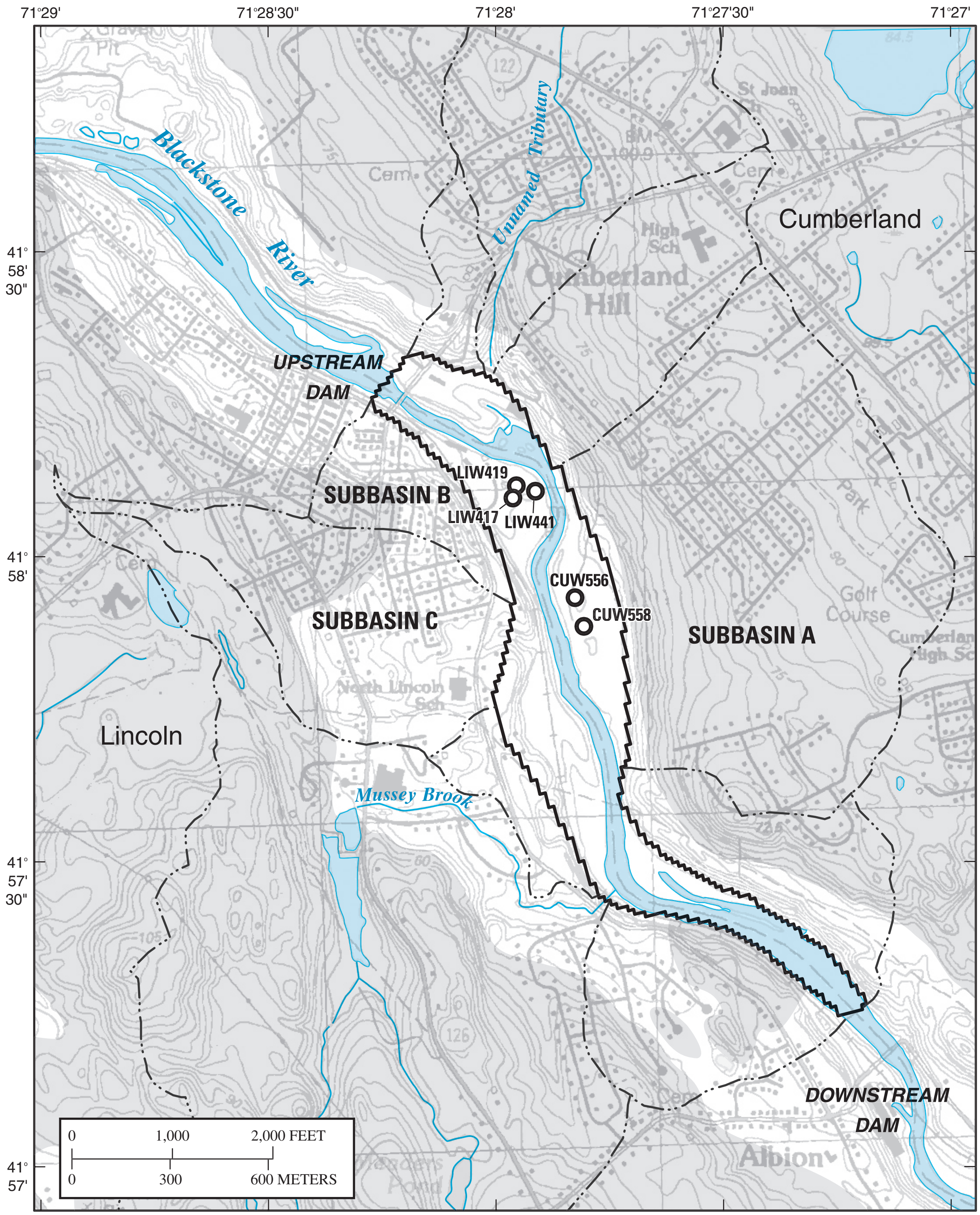
The well field in the Town of Cumberland consists of two active wells, whereas the well field in the Town of Lincoln consists of three backup wells that have not operated since 1985 because of the detection of contaminated ground water at the well field (Roy F. Weston, Inc., 1996). Characteristics of the five wells, whose distances from the Blackstone River range from 170 to 380 ft, are listed in table 1. Well-screen depths range from 42–52 ft to 80–95 ft below land surface; maximum saturated-sediment thickness at the well fields is about 80 ft. The areas contributing recharge were determined for the Cumberland well field operating alone at average and maximum pumping rates. In addition, because areas that contribute water to a well field can be affected by a change in stress on the ground-water system, such as withdrawals by nearby supply wells, areas contributing recharge were determined for the Cumberland and Lincoln well fields pumping simultaneously at maximum pumping rates. Average and maximum pumping rates are listed in table 1.

Geohydrology

The valley-fill deposits consist of stratified glacial deposits and till; the areal extent of these surficial sediments mapped by Chute (1949) is shown in figure 2 and the topography of the underlying crystalline bedrock (Quinn and others, 1949) is shown in figure 3. The bedrock valley trends northwest-southeast at the two well fields. Stratified sediments that cover the western hillslope are mostly unsaturated.

A north-south section through the well fields (*A–A′*) shows the lithology and thickness of the stratified glacial deposits and the vertical placement of the public-supply well screens in relation to the Blackstone River (fig. 4). The generalized geologic section indicates one depositional sequence as the glacier retreated in a northward direction. Fine sediments, consisting of fine sand, silt, and clay in the south part of the Cumberland well field, are overlain by well-sorted sand, which in turn is overlain by sand and gravel. Sand and gravel sediments in the south part of the section, which is near the river, are probably post-glacial alluvial deposits. Sand and gravel sediments between CUW556 and the river, in some cases, consist of poorly sorted material. In addition, the fine sediments in the southern part of the section extend northeastward between the two Cumberland supply wells and the valley-hillslope contact. All five public-supply wells are screened in the sand and gravel sediments, three at medium depth and two near the bottom of this unit.

The study area is drained by the Blackstone River, which is wide relative to the valley. The downstream dam provides a relatively flat river gradient from the dam to the Lincoln well field. In this setting, river stage controls ground-water levels in the valley near the river. Mean annual streamflow, based on records from a USGS continuous streamflow-gaging station in the regulated Blackstone River at Woonsocket, Rhode Island (01112500) (fig. 1), 3 mi upstream from the study area, is 778 ft^3/s for the 72 years of record (1929–2001) (Socolow and others, 2002).



Base from U.S. Geological Survey topographic quadrangle, Pawtucket, Rhode Island, 1975, 1:24,000

Surficial geology from Chute, 1949

EXPLANATION

TILL AND BEDROCK

STRATIFIED DEPOSITS

MODEL BOUNDARY

SUBBASIN BOUNDARY—Only part of two perennial-stream subbasins, for Mussey Brook and an unnamed tributary, are shown. Subbasins A,B, and C are referred to in report text.

CUW558 PUBLIC-SUPPLY WELL AND IDENTIFIER

Figure 2. Public-supply wells, surficial geology, and upland subbasins draining toward model boundary, Cumberland and Lincoln study area, Rhode Island.

Table 1. Characteristics of the public water-supply wells in the Cumberland and Lincoln study area, the North Smithfield study area, and the Westerly study area, Rhode Island.

[Altitudes in feet relative to NGVD 29. No., number; USGS, U.S. Geological Survey; ft, foot; gal/min, gallons per minute; --, individual pumping rates unknown]

Name		Altitude of land surface (ft)	Depth of screen top and bottom below land surface (ft)	Screen altitude top and bottom (ft)	Average pumping rate (gal/min)	Maximum pumping rate (gal/min)	Year installed
USGS	Local						
Cumberland							
CUW556	No. 1	104	69–79	35–25	[1]172	[2]500	[3]1995
CUW558	No. 2	104	42–52	62–52	[1]152	[2]500	[3]1995
Total					324	1,000	
Lincoln							
LIW417	No. 3	105	45–65	60–45	--	--	1963
LIW419	No. 5	105	52–62	53–43	--	--	1969
LIW441	No. 10	105	80–95	25–10	--	--	1979
Total					[4]1,111	[5]2,083	
North Smithfield							
NSW310	Tifft Road	238	49–64	189–174	[6]200	[7]500	1963
Westerly							
WEW584	Crandall	41	47–62	-6 to -21	[8]240	[9]700	1986

[1]Average of years 1997–99 (2000 and 2001 not included because wells were developed).

[2]Neal Fiorio, Cumberland Water Department, oral commun., 2002.

[3]Replacement.

[4]Roy F. Weston Inc. (1996).

[5]Maurice Trudeau, Lincoln Water Commission, oral commun., 2002.

[6]Approximate maximum pumping rate of current well (Michael Romano, formerly of North Smithfield Water Department, oral commun., 2001).

[7]Estimated yield of replacement well (HydroSource Associates, Inc., 1999).

[8]Average of years 1997–2001.

[9]Paul Corina, Westerly Water Department, oral commun., 2002.

Three sources of water potentially recharge the stratified deposits in the valley and, thus, could provide water to be withdrawn by the supply wells. One source is direct infiltration of precipitation that falls on the valley. A second source is runoff from the adjacent uplands, either by ground-water flow or surface-water runoff. Surface-water runoff from the uplands discharges onto the valley surface in defined channels or fans out over the land surface at the valley and upland contact. Because the valley surface is usually above the water table, water can infiltrate to the saturated zone. In valley-fill settings under nonpumping conditions, upland runoff can be a major source of recharge to the valley aquifer, if the area of the valley floor is much smaller than the upland area draining towards the valley. For example, in a similar valley-fill setting in a subbasin of the Deerfield River Watershed, northwestern Massachusetts, at least 70 percent of the recharge to the valley aquifer was derived from upland sources (Friesz, 1996).

Hillslope areas that drain toward the valley were divided into subbasins on the bases of land-surface contours (fig. 2). Field observations of the upland-and-valley contact east of the two Cumberland supply wells (fig. 2, subbasin A) indicated three major intermittent streams draining this upland subbasin. Water from two of these intermittent streams, which are east and northeast of CUW556, infiltrate the stratified deposits at, or near, the valley-and-upland contact, except during the peak of the spring runoff when small amounts of streamflow discharge to the Blackstone River. The third intermittent stream, southeast of CUW558, discharges water to the Blackstone River during the spring runoff and after intense precipitation events. Thus, the actual recharge to the aquifer from the upland subbasin east of the Cumberland well field is less than the total upland runoff.

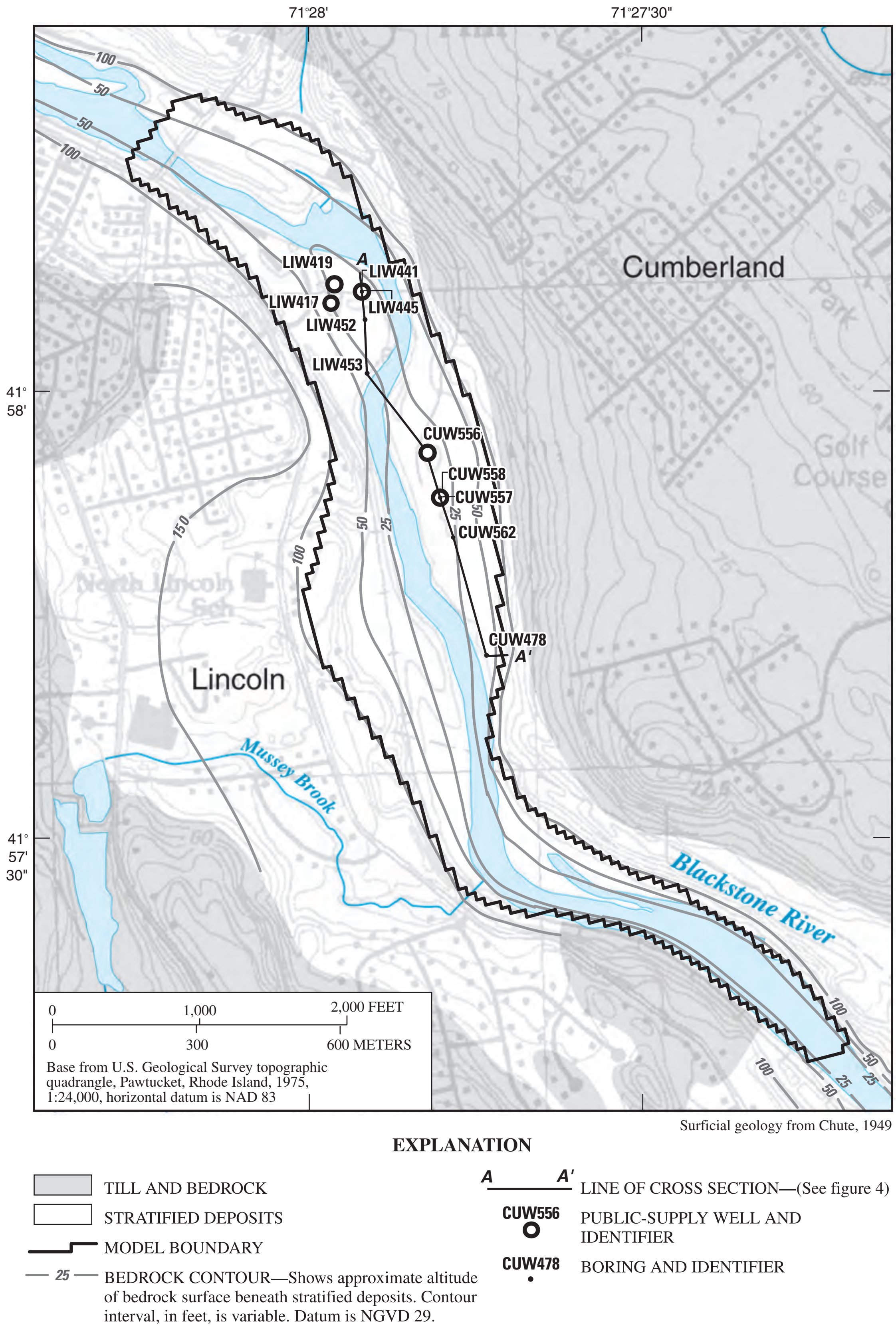


Figure 3. Bedrock-surface contours, section line, public-supply wells and selected borings, Cumberland and Lincoln study area, Rhode Island.

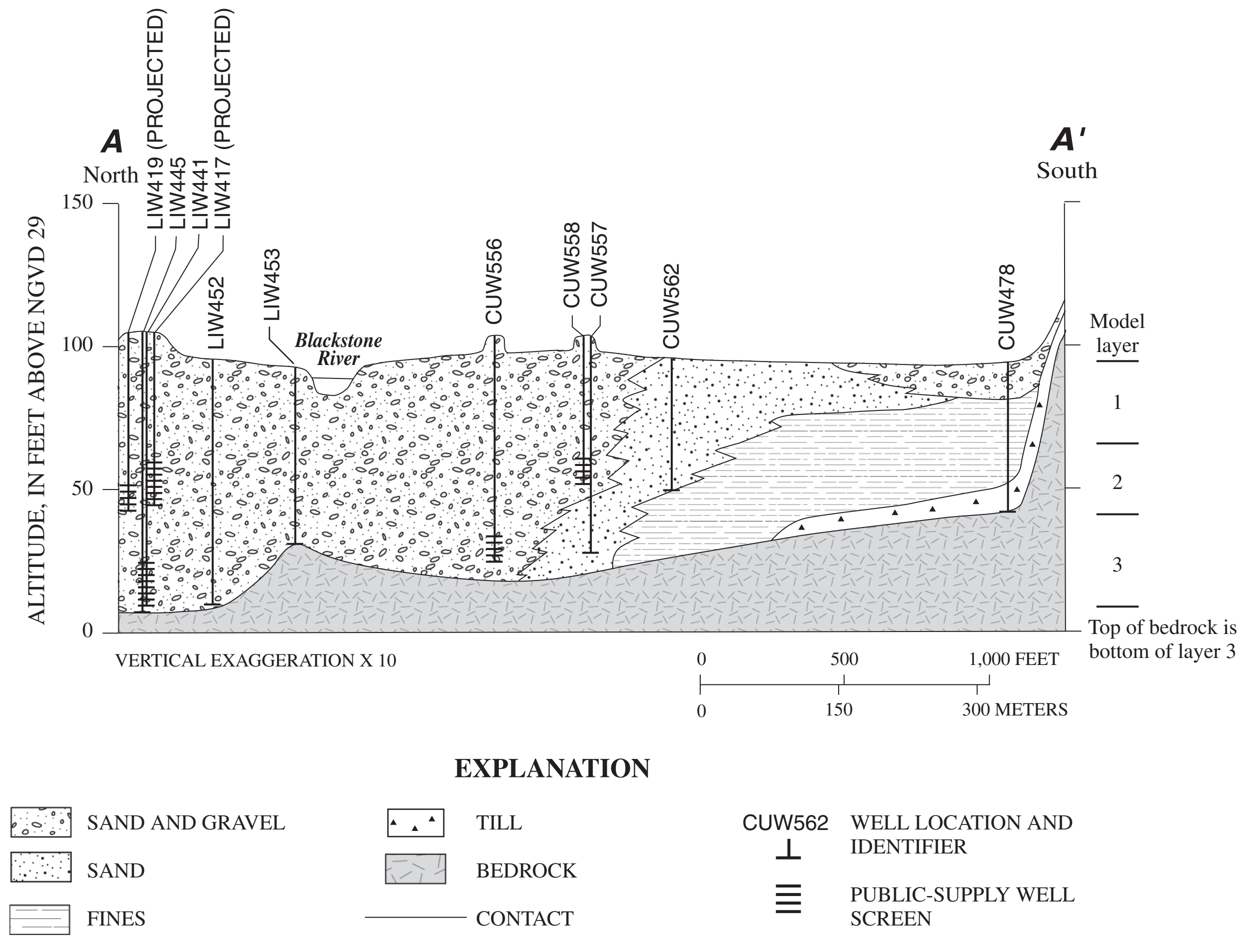


Figure 4. Geologic cross section and model layers (line of section *A*-*A*′ shown in fig. 3), Cumberland and Lincoln study area, Rhode Island.

The hillslope west of the Lincoln well field (fig. 2, subbasin B) is covered by permeable stratified deposits, but land use is urban. Storm sewers draining impervious surfaces channel the hillslope runoff and discharge water at the valley-and-upland contact; field observations indicate that most of this runoff infiltrates the aquifer near the contact and the remaining overland flow discharges into a detention basin, where it infiltrates the aquifer. Thus, all runoff from this hillslope is assumed to recharge the valley deposits.

Another important hillslope area in terms of the supply wells is directly west of the Cumberland well field on the Lincoln side of the Blackstone River (fig. 2, subbasin C). An intermittent stream drains this hillslope, which supports wetlands and developed urban areas in its watershed. This intermittent stream discharges to the Blackstone River, thereby reducing the percentage of runoff from this hillslope that recharges the valley deposits.

Two perennial streams, Mussey Brook and an unnamed tributary, drain the uplands in the study area, according to the USGS Pawtucket quadrangle. Both streams, however, probably contribute little recharge to the valley aquifer because one stream (Mussey Brook) empties onto the valley floor a short distance from the Blackstone River and the other stream (unnamed tributary) travels through a human-made tunnel where it empties on the valley floor before discharging to the river.

A third potential source of water to the stratified deposits is leakage from the Blackstone River, either naturally, induced by pumping of a well, or both. The quantity of river water available to recharge the aquifer under average conditions (the mean annual flow rate of 778 ft^3/s for the Woonsocket, RI, streamflow-gaging station) far exceeds the maximum pumpage withdrawal rate of the Cumberland well field (1,000 gal/min or 2.23 ft^3/s) or the maximum withdrawal rate from both well fields (3,083 gal/min, or 6.87 ft^3/s).

Unpublished profiles of water depth and thickness of fine-grained riverbed sediments from a previous USGS study (Johnston and Dickerman, 1974a) indicated that, within most reaches of the Blackstone River, the riverbed is composed of coarse-grained sediments, except near its banks. The riverbed between the upstream dam and the Lincoln well field, beneath a reach of rapidly moving water, consists of all coarse-grained sediments. Conceptually, ground water flows into the river and river water infiltrates the aquifer through the permeable coarse-grained material. The average river depth is 8 ft at the Cumberland well field and 4 ft near the Lincoln well field.

Water-level and temperature measurements indicate that water infiltrates from the Blackstone River into the aquifer and discharges from the supply wells. A generalized potentiometric map shown in figure 5 was drawn from water-level measurements on June 15, 2001, in the river and in observation wells screened at different elevations when CUW556 and CUW558 were each pumping at a rate of 400 gal/min during an on-off pumping cycle of 24 hours. The direction of ground-water flow, approximately perpendicular to the contours, is from both the valley-hillslope contact and the river. Water levels measured in shallow observation wells adjacent to and on both sides of the Blackstone River were lower than the river stage. These water levels indicate the area through which surface water likely infiltrates into the aquifer under these pumping conditions. A comparison of the seasonal variation in river temperature to the seasonal variation of the temperature of water withdrawn from CUW409, which is near CUW556, indicated that part of the water withdrawn from the supply wells is derived from the Blackstone River (Johnston and Dickerman, 1974a).

Model Design

Ground-water flow in the stratified glacial deposits in the narrow Blackstone River Valley was simulated as a three-layer numerical model with a uniformly spaced grid. The model grid represents an area of 0.24 mi^2 (fig. 2) and consists of 149 rows and 64 columns with each cell 50 ft on a side. The model grid consists of 5,920 active cells and was oriented parallel to the northwest-southeast trending valley at the well fields. The dams built across the Blackstone River represent the upstream and downstream extent of the model. Because all water upstream of these dams was assumed to flow over the dams, no ground water was simulated as flowing into or out of the model beneath the dams. The valley and hillslope contact forms the eastern extent of the model. The 10-ft saturated contour line mapped by Johnston and Dickerman (1974a) on the western hillslope originally represented the western extent of the model, but because of the numerical instability created by trying to simulate thin saturated sediments on the side of a hill, a smaller area of this hillslope was simulated in the final model.

The stratified deposits were subdivided vertically into three model layers that extend from the water table to the bedrock surface (fig. 4). The vertical spacing of the layers was chosen to represent the lithology and the vertical placement of the observation wells and public-supply well screens. Three supply wells are screened in layer 2 and two supply wells are screened in layer 3. The lateral extent of the active area decreases from the top layer (layer 1) to the bottom layer (layer 3) in conformity with the shape of the valley-fill bedrock channel.

Boundary conditions specified in the model to represent sources of recharge and areas of discharge are shown in figure 6. The interaction between the Blackstone River and the underlying aquifer was simulated as a head-dependent flux boundary by using the MODFLOW river package (Harbaugh and McDonald, 1996) in layer 1. River stage was interpolated between stage measurements made during this study adjacent to the Cumberland well field and the downstream dam, and was taken from topographical contours on the USGS Pawtucket quadrangle. Riverbed elevations were obtained from unpublished USGS water-depth profiles from a USGS study (Johnston and Dickerman, 1974a) and the predominately coarse-grained riverbed was simulated as 1 ft thick. An analytical method based on Darcy's Law was used by Johnston and Dickerman (1974a) to determine the vertical hydraulic conductivity of the riverbed at the well fields. Water levels, measured in an array of shallow piezometers near the river during a 7-day aquifer test at the well fields, were used to determine the riverbed area of infiltration and the average head loss over this area. Along with an estimate of the percentage of water withdrawn from the pumping wells that is derived from the river, a vertical hydraulic conductivity equal to 0.2 ft/d was calculated with this method. This same method was used by Johnston and Dickerman (1974a) at a site 4 mi downstream of the modeling area with similar riverbed characteristics and resulted in a vertical hydraulic conductivity of 1.8 ft/d. A vertical hydraulic conductivity of 0.2 ft/d was used as an initial value in the model for most of the riverbed of the Blackstone River, although this value is more typical of fine-grained than coarse-grained sediments, and could indicate fine-grained sediments beneath the coarse-grained riverbed sediments. The river reach between the upstream dam and the Lincoln well field, which consists entirely of coarse-grained material, was assigned a value of 1 ft/d.

The top boundary of the aquifer is the water table, which is simulated by the model. Recharge from precipitation directly on the valley floor was uniformly distributed over the top of the model. Recharge was applied at a rate of 26 in/yr based on mean annual runoff data for 30 years, 1951–80 (Randall, 1996). This recharge rate, representative of the water available for recharge for long-term average annual conditions, is confirmed by 61 years of runoff record (1940–2001) from an unregulated subbasin to the Blackstone River, Branch River at Forestdale (01111500) (fig 1). The geologic contact between the permeable stratified sediments and the underlying, relatively impermeable till and bedrock, is represented as a no-flow boundary.

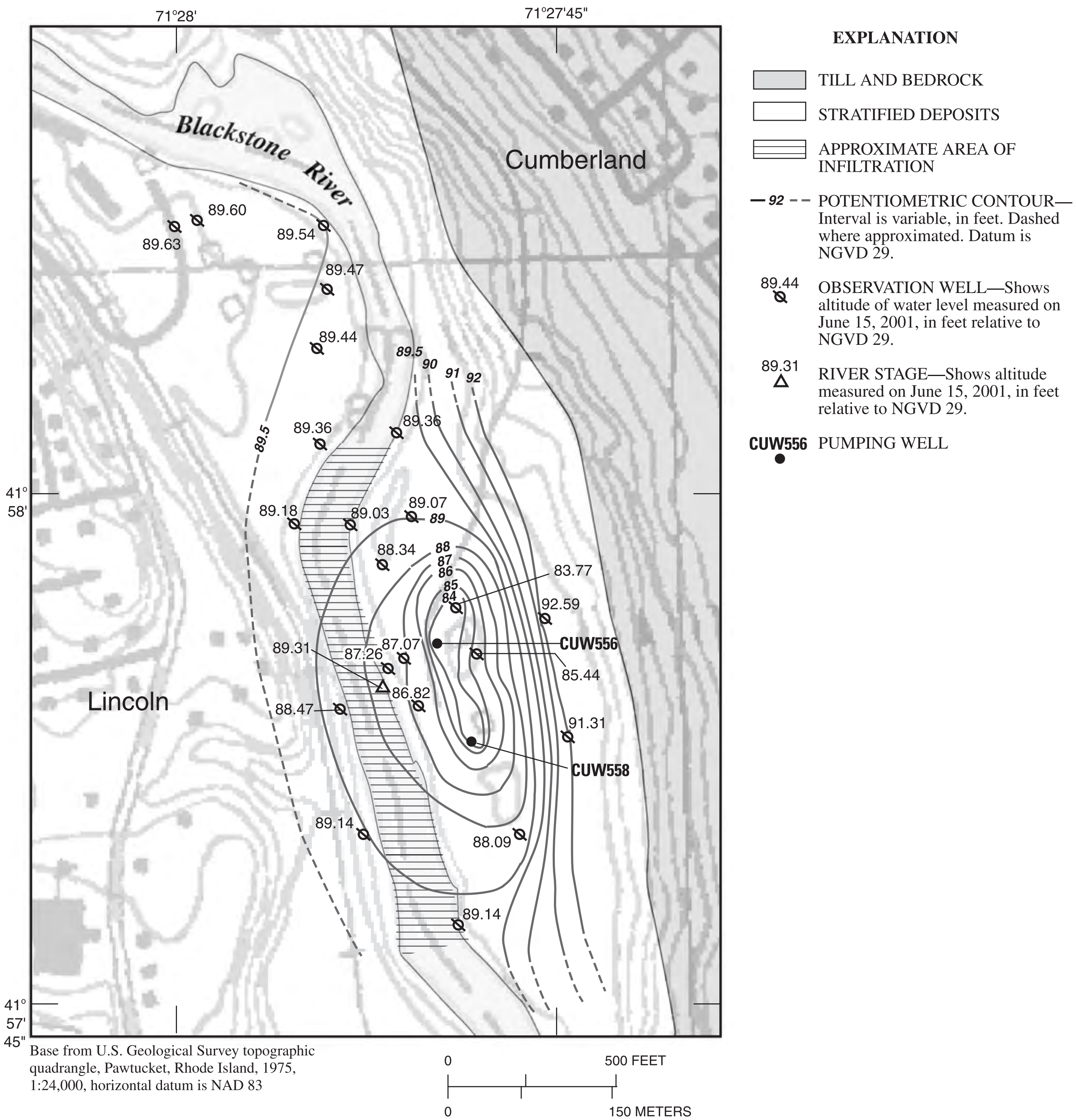


Figure 5. Estimated potentiometric contours and area of river infiltration for a total pumping rate of 800 gallons per minute during a cycle of 24 hours on and then off, June 15, 2001, Cumberland well field, Cumberland and Lincoln study area, Rhode Island.

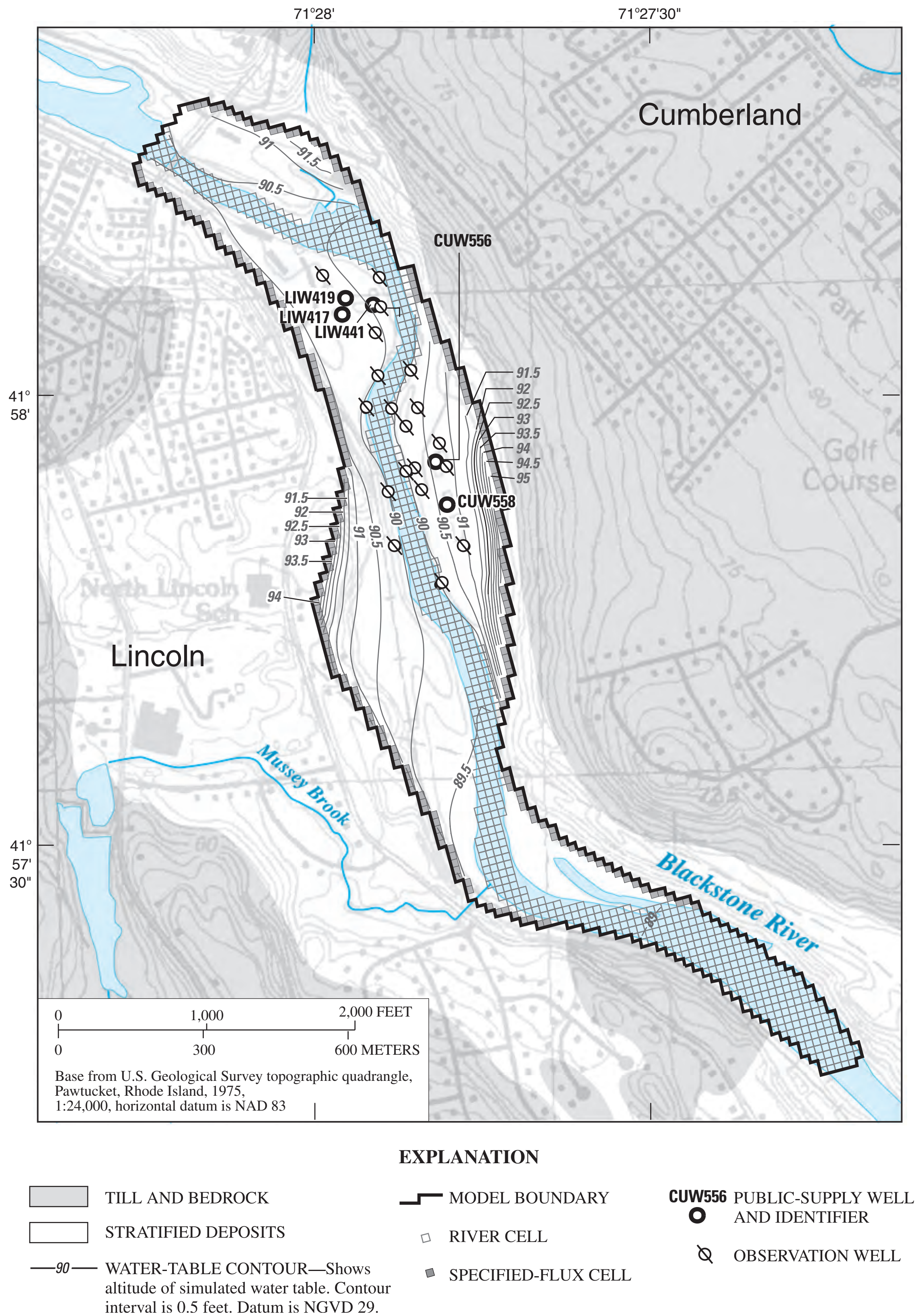


Figure 6. Model-boundary types, simulated water-table contours for steady-state, nonpumping conditions, and location of public-supply wells and observation wells, Cumberland and Lincoln study area, Rhode Island.

Recharge from hillslopes draining toward the valley was simulated by specified fluxes. Inflow from these hillslopes was calculated by multiplying an upland runoff rate by the subbasin hillslope area and then evenly distributing the inflow among the cells in the top layer adjacent to the hillslope. For most hillslope subbasins, 75 percent of the water available for recharge, or 20 in/yr, was chosen on the basis of observations of hillslope runoff at the Cumberland well field. This value was considered to be typical for the study area because most valley areas are similarly narrow in extent compared to the hillslope area draining toward them. Exceptions were the hillslopes directly west of the Lincoln and Cumberland well fields, where recharge rates of 26 and 13 in/yr, respectively, were applied on the basis of hillslope-runoff observations discussed in the previous section. In addition, if the base of a hillslope intersected the Blackstone River, then no recharge from the uplands was applied to the model.

Initial horizontal hydraulic-conductivity values were specified on the basis of lithology and aquifer-test results at both well fields reported by Johnston and Dickerman (1974a). Coarse-grained sediments, representing both sand and gravel and sand, were assigned a single value of 180 ft/d. The fine-grained sediments were assigned a value of 15 ft/d.

Model Calibration

The ground-water-flow model was calibrated through steady-state and transient simulations. The steady-state calibration was based on water-level measurements made on February 21, 2001, after several months of minimal pumping at the supply wells, and the transient calibration was based on unpublished USGS drawdown measurements collected during a 7-day aquifer test from April 25 to May 2, 1971, by Johnston and Dickerman (1974a). Ground-water levels based on the steady-state nonpumping simulation were used as initial water levels for the transient simulation. The input parameters adjusted during model calibration were horizontal hydraulic conductivity and vertical hydraulic conductivity of the riverbed sediments.

Nineteen ground-water levels and two river-stage measurements made on February 21, 2001, are near long-term mean annual conditions determined on the basis of records from USGS long-term well LIW84 and from the streamflow-gaging station Branch River at Forestdale, Rhode Island (01111500) (fig. 1 and 7). LIW84 is 4 mi downstream of the study area in a similar valley-fill setting. The two Cumberland supply wells were operated for only a few hours a week in December 2000 and January 2001, and not at all in February 2001, because of plans to redevelop the supply-well screens; as a result, water levels in observation wells are assumed to represent nonpumping steady-state conditions. The observation network (fig. 6) provided an areal definition of water levels in the valley, including adjacent to the river, for the nonpumping steady-state simulation.

Initial input values of vertical hydraulic conductivity of the riverbed and horizontal hydraulic conductivity of the aquifer caused simulated water levels throughout the ground-water system to be greater than measured water levels, especially near the river. Varying these conductivity values indicated that simulated water levels throughout the narrow aquifer were sensitive to vertical hydraulic conductivity of the riverbed, whereas water levels in the aquifer away from the riverbanks were more sensitive to horizontal hydraulic conductivity than water levels close to the riverbanks. For the steady-state model, the initial value of vertical hydraulic conductivity of the riverbed was raised from 0.2 to 0.8 ft/d, which lowered water levels throughout the aquifer and improved the match between simulated and measured water levels near the river, but also lowered simulated water levels below measured water levels in the coarse sediments away from the river. By lowering horizontal hydraulic conductivity of coarse sediments from 180 to 120 ft/d, a reasonable match between simulated and measured water levels away from the river was achieved. In addition, horizontal hydraulic conductivity of the fine sediments was raised from 15 to 20 ft/d to lower water levels below the land surface in areas near the valley-hillslope contact.

The altitude and configuration of the simulated water table for long-term steady-state, nonpumping conditions are shown in figure 6. The direction of ground-water flow is generally perpendicular to the valley axis from the valley-hillslope contact to the Blackstone River because the uplands provide a major source of recharge to the aquifer and because of the relatively flat river gradient. The calculated total inflow and outflow rate to the modeled area is 141,936 ft^3/d. Of the total inflow, 111,188 ft^3/d or 78 percent is from upland runoff, 28,364 ft^3/d or 20 percent is from direct precipitation recharge, and 2,384 ft^3/d or 2 percent is from natural seepage of river water. Outflow is through the river.

The upland subbasin east of the Cumberland well field potentially provides a substantial quantity of water to the aquifer in the vicinity of the well field. The upland subbasin area, 0.45 mi^2, contributes 57,157 ft^3/d or 297 gal/min to the aquifer. The valley floor at the Cumberland well field, defined as the area bounded by the hillslope and the river, is 0.04 mi^2, and contributes only 6,608 ft^3/d or 34 gal/min by precipitation recharge. Under nonpumping, steady-state conditions, total recharge to the valley at the Cumberland well field from upland runoff and precipitation recharge, 331 gal/min, is nearly equivalent to the average pumping rate and one-third of the maximum pumping rate. Thus, to sustain withdrawals at the maximum pumping rate, other sources of water from either the opposite side of the river or the river itself, or both, must be withdrawn from the Cumberland well field.

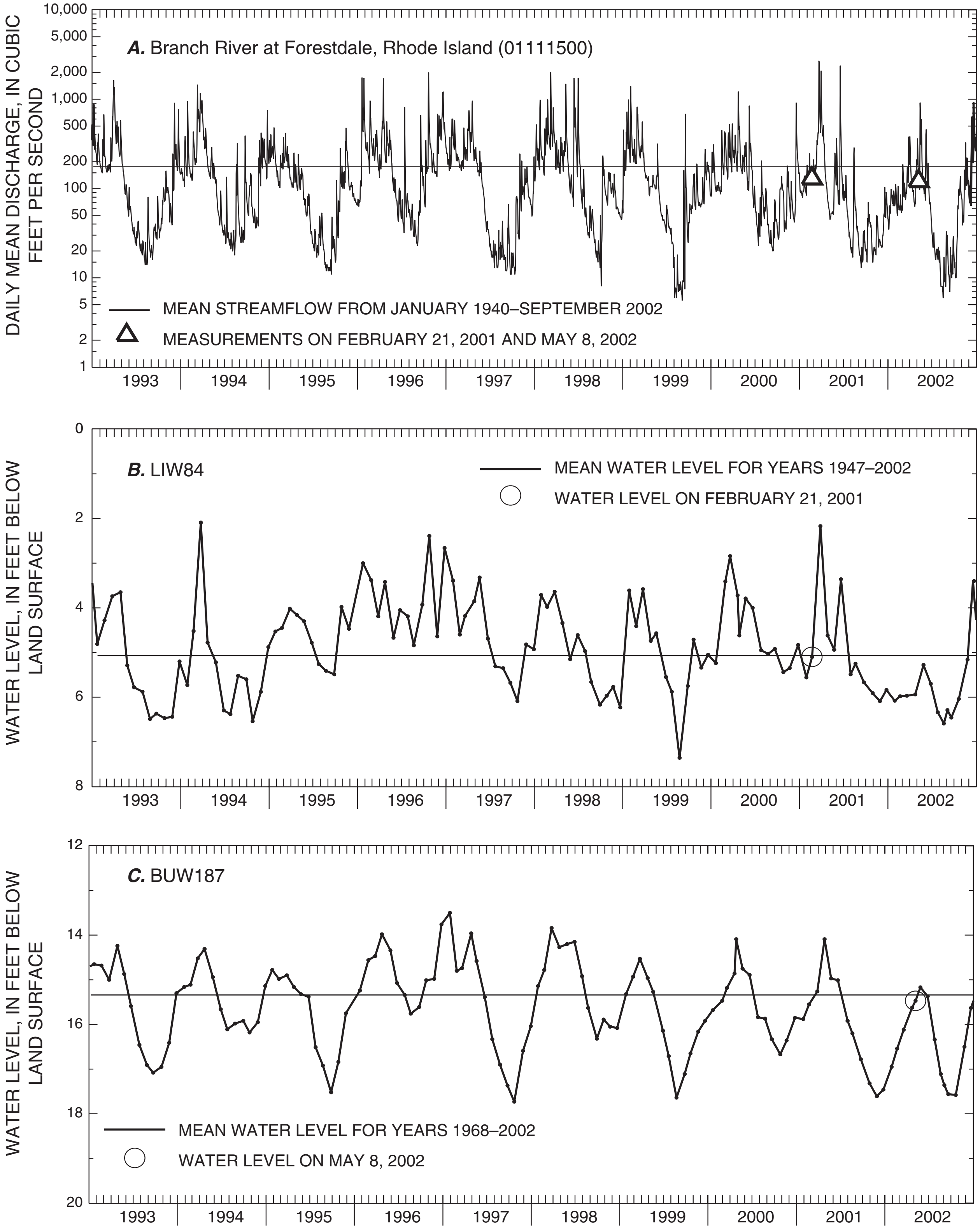


Figure 7. Streamflow for *A*, the Branch River at Forestdale, Rhode Island; and water levels at *B*, well LIW84; and *C*, well BUW187, 1993–2002.

The numerical model was then calibrated through simulation of transient conditions during the 7-day aquifer test. The aquifer test consisted of one supply well operating at each well field. Cumberland well CUW409 (fig. 8), which is near CUW556 and screened in layer 3, was pumped at a rate of 780 gal/min from April 25 to May 2, 1971. On the opposite side of the river 24 hours after the start of the aquifer test, Lincoln well LIW417, screened in layer 2, began operating at a rate of 590 gal/min for the remaining 6 days of the test. An initial specific-yield value of 0.25 and specific-storage value of 0.00001 were added to the model for the transient simulation. No precipitation was recorded during the aquifer test, so no recharge from direct infiltration was applied to the model; however, a constant seepage from the hillslopes equal to the steady-state rate was assumed in the simulation.

The transient model was calibrated by comparing simulated drawdowns to measured drawdowns at the end of the aquifer test at 16 observation wells. These observation wells were on both sides of the river (fig. 8) and provided an areal definition of the drawdown distribution; however, most of these wells—13 of 16—were screened in the top part of the aquifer in layer 1.

Model-input parameters from the steady-state simulation did not provide enough drawdown throughout most of the aquifer in the transient simulation compared to measured drawdowns. To improve the match, specific yield was first lowered from 0.25 to 0.15, which is at the low end of the range of values determined by laboratory methods by Allen and others (1963) for sand and for sand and gravel deposits. Specific-yield values at the low end determined from laboratory methods may be more realistic than the average value because the rate at which water drains from pore spaces at the water table is time dependent and gradually approaches its maximum value (Bear, 1979). After 7 days of pumping, water may not have completely drained from the pore spaces during the aquifer test. The vertical hydraulic conductivity of the riverbed was then lowered from 0.8 to 0.2 ft/d to increase drawdowns, and the horizontal hydraulic conductivity of the coarse deposits was raised from 120 to 150 ft/d to extend the drawdown cone farther from the pumping wells. These input values provided a reasonable match between simulated and measured drawdowns except at observation wells near CUW409 and at shallow observation wells adjacent to the river near CUW409 (table 2). The range in differences between simulated and measured drawdowns was -1.00 to 7.45 ft and the mean absolute value was 1.69; the absolute error was 0.54 ft or less for the observation wells not in the vicinity of CUW409. To simulate the drawdown distribution more accurately, a finer grid and a thinner layer 1 would be needed because of the complex flow system near a partially penetrating river and because of model boundaries in close proximity to the observation wells.

Model-input values determined from the transient calibration, 0.2 ft/d for vertical hydraulic conductivity of the riverbed and 150 ft/d for horizontal hydraulic conductivity of the coarse deposits, were used to simulate steady-state ground-water flow under pumping conditions to determine the areas contributing recharge and sources of water to each well field. The model-input values determined by the transient calibration are similar to values determined from analytical methods and lithologic logs reported by Johnston and Dickerman (1974a). The transient calibration also was based on a simulation which includes a stress applied to the aquifer. The steady-state simulation based on nonpumping conditions provided head data for the beginning of the transient simulation and provided insights into the configuration of the water table and water budget for the study area; however, the steady-state model based on nonpumping conditions was not used for any predictions.

Delineation of Areas Contributing Recharge and Sensitivity Analysis

The areas contributing recharge and the sources of water to the public-supply well fields are based on model-input values determined from the calibration process and from average recharge conditions. The focus of this section is on the active Cumberland well field, but the areas contributing recharge determined for both the Cumberland and Lincoln well fields pumping simultaneously at their maximum rates also were determined in case the Lincoln wells, considered backup wells, are used for water supply in the future. Insights into the importance of selected model-input parameters on the delineated extent of the area contributing recharge and sources of water to the Cumberland well field also should be applicable to the Lincoln well field.

The two Cumberland wells are operated together and total 324 gal/min at the average pumping rate and 1,000 gal/min at the maximum pumping rate (table 1). The location and extent of the areas contributing recharge for the average and maximum pumping rates in the modeled area are illustrated in figure 9 and the sizes of the areas and the percentages of the total water withdrawn from each water source are listed in table 3. For both pumping rates, the areas contributing recharge extend west of the Blackstone River and include upland runoff from the west. The areas contributing recharge extend beyond the river despite that the river, a major source of water, is in close proximity to the well field. The valley area contributing recharge for the well field at the average pumping rate covers about 0.05 mi^2; the well field derives most of its water from upland runoff (72 percent) compared to direct precipitation recharge (21 percent) and infiltration from the Blackstone River (7 percent). At the maximum pumping rate, the area contributing recharge (0.12 mi^2) extends farther up and down the valley to capture enough water to balance the pumping rate. At the increased pumping rate, however, the primary source of water pumped at the well field is from the river (53 percent) compared to upland runoff (35 percent) and precipitation recharge (12 percent).

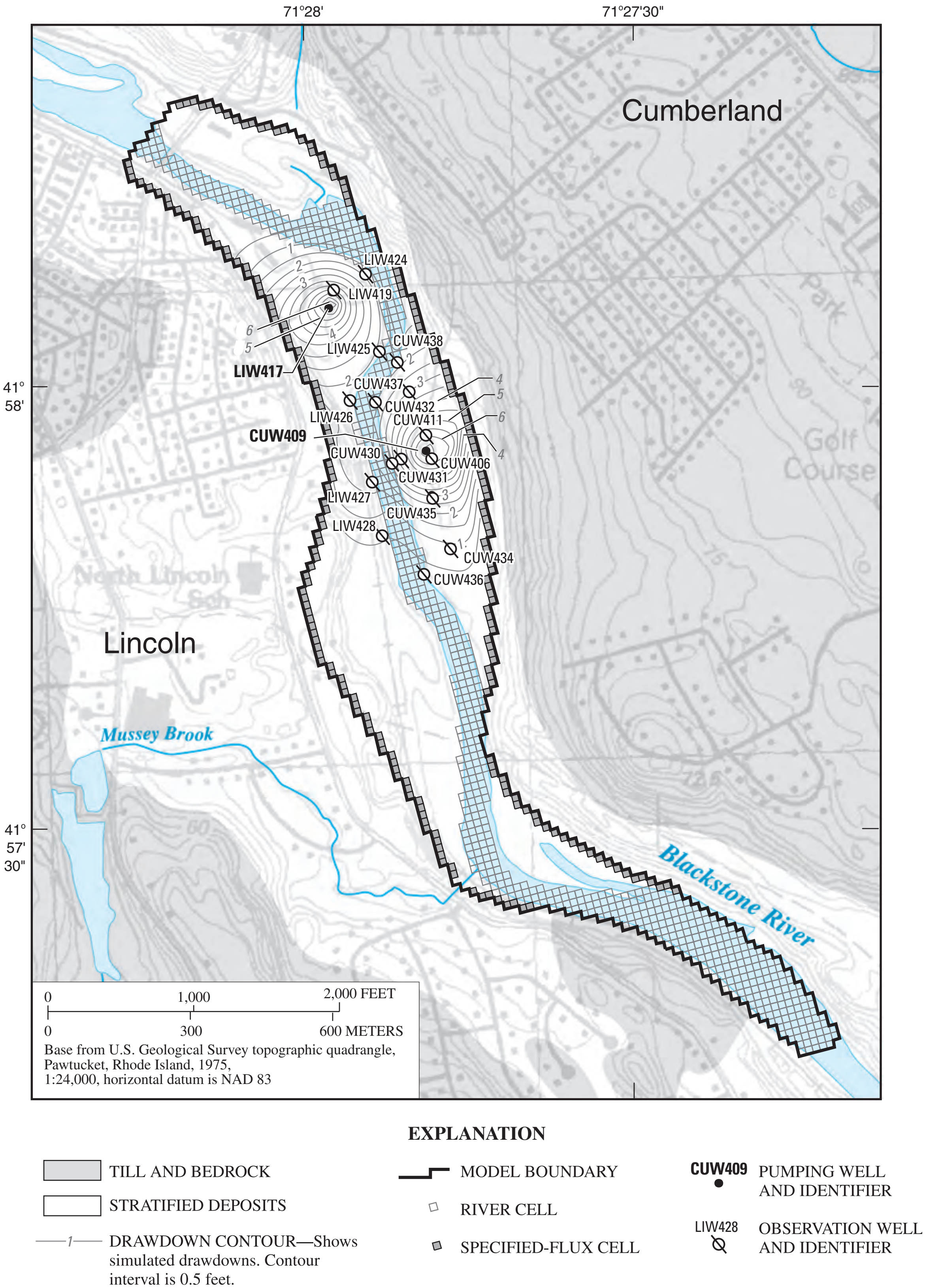


Figure 8. Model-boundary types, simulated drawdowns at end of 7-day aquifer test, and location of pumping wells and observation wells, Cumberland and Lincoln study area, Rhode Island.

Table 2. Measured drawdowns after 7 days of pumping on May 2, 1971, and simulated drawdowns for the transient model, Cumberland and Lincoln study area, Rhode Island.

[ft, foot]

Observation well	Model layer	Measured (ft above NGVD 29)	Simulated (ft above NGVD 29)	Difference (measured minus simulated, in ft)
Cumberland				
CUW406	2	14.39	7.6	6.79
CUW411	2	13.94	6.49	7.45
CUW430	1	5.71	3.57	2.14
CUW431	1	6.51	4.33	2.18
CUW432	1	3.56	3.02	.54
CUW434	1	1.39	0.85	.54
CUW435	1	2.98	2.84	.14
CUW436	1	.15	.06	.09
CUW437	1	3.5	1.96	1.54
CUW438	1	2.05	1.83	.22
Lincoln				
LIW419	2	3.87	4.87	-1.00
LIW424	1	1.48	1.86	-.38
LIW425	1	1.95	1.86	.09
LIW426	1	3.19	1.65	1.54
LIW427	1	3.68	1.69	1.99
LIW428	1	.46	.61	-.15

The effects of operating the backup Lincoln well field for an extended period at its maximum pumping rate of 2,083 gal/min on the area contributing recharge and sources of water to the Cumberland well field are shown in figure 9 and table 3. Simulated drawdowns at the Lincoln well field initially caused model cells near the well field adjacent to the uplands to go dry. The hydraulic conductivity of these model cells was then lowered to raise heads above the model-cell bottom; this adjustment enabled the specified fluxes representing upland runoff to be simulated in the model. By incorporating the Lincoln well-field withdrawals, the area contributing recharge to the Cumberland well field was reduced to 0.08 mi^2; the northern part of the area contributing recharge when the Cumberland well field was operating alone became part of the contributing area for the Lincoln well field. The Cumberland well field compensated by withdrawing an increased percentage of its total water from the river, by 11 percent from 53 to 64 percent, and by extending its contributing area slightly farther downgradient to the south. The area contributing recharge to the Lincoln well field is 0.08 mi^2 and the predominant source of water is the river (88 percent).

Upland areas outside the modeled area that drain toward the areas contributing recharge are shown in figure 9 and the size of the areas are listed in table 3. These upland areas were drawn based on land-surface contours and the boundary of the watershed. The area that drains toward the Lincoln well field's contributing area on the opposite side of the river from the Lincoln well field does not include an upland-stream watershed. As previously mentioned, recharge from this upland stream was not added to the model because the stream moves through a human-made tunnel where it enters the valley floor near the Blackstone River.

A sensitivity analysis of the area contributing recharge and sources of water to the Cumberland well field for alternative model-input values was done for the maximum withdrawal rate from that well field alone (1,000 gal/min). The alternative values provide insights into the importance of selected input parameters on the delineated area contributing recharge in a narrow, valley-fill setting. Results of the alternative model-input values on the area contributing recharge are shown in figure 10 and table 3.

The hydraulic connection between the river and aquifer was reduced by one-half by decreasing the vertical hydraulic conductivity of the riverbed sediments in the vicinity of the well field area from 0.2 to 0.1 ft/d. The reduced connection caused the quantity of river water withdrawn at the well field to decrease by 7 percent (from 53 to 46 percent) and the area contributing recharge to expand slightly (from 0.12 to 0.13 mi^2). The horizontal hydraulic conductivity of coarse-grained sediments was doubled (from 150 to 300 ft/d). This change caused less drawdown in the vicinity of the well field and river, therefore, there was less infiltration of river water (from 53 to 46 percent) and a slightly larger area contributing recharge (from 0.12 to 0.13 mi^2). Recharge rates were reduced by 6 in/yr, a plausible alternative given the results of field observations of upland runoff. The decreased recharge rate caused an increase in the river water withdrawn from the well field by 10 percent (from 53 to 63 percent) and also increased the area contributing recharge slightly (from 0.12 to 0.13 mi^2). The sensitivity analysis indicated that in this hydrologic setting with two major water sources in close proximity to the well field—a wide river in relation to the width of the valley floor and large fluxes from upland runoff applied to the model boundary—the area contributing recharge was not particularly sensitive to selected model-input values for the ranges considered.

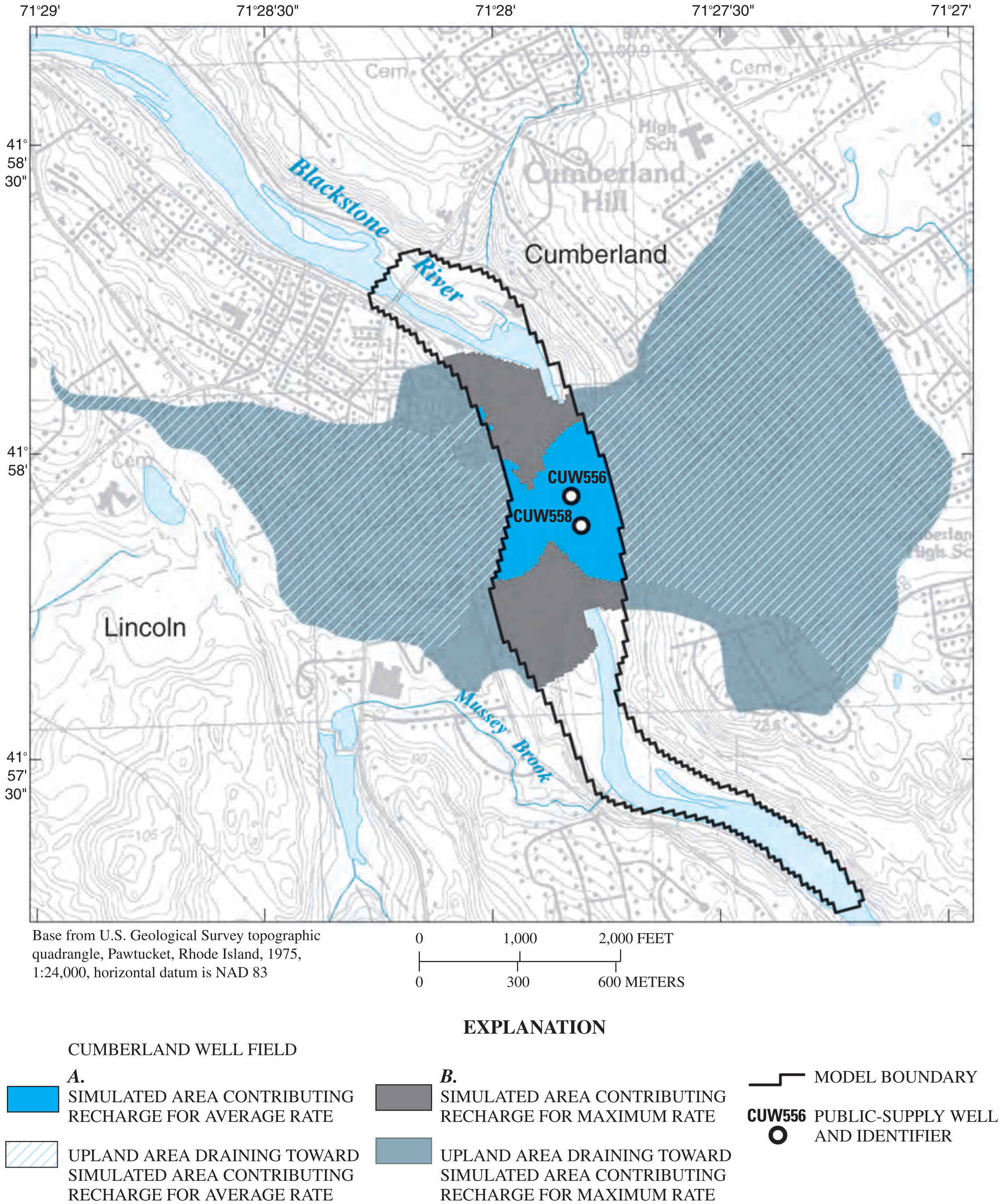


Figure 9. Areas contributing recharge in the modeled area and uplands draining toward the area contributing recharge for *A*, Cumberland well field average pumping rate of 324 gallons per minute; *B*, Cumberland well field maximum rate of 1,000 gallons per minute; and *C*, Cumberland and Lincoln well fields pumping simultaneously at their maximum pumping rates of 1,000 gallons per minute and 2,083 gallons per minute, respectively, Cumberland and Lincoln study area, Rhode Island.

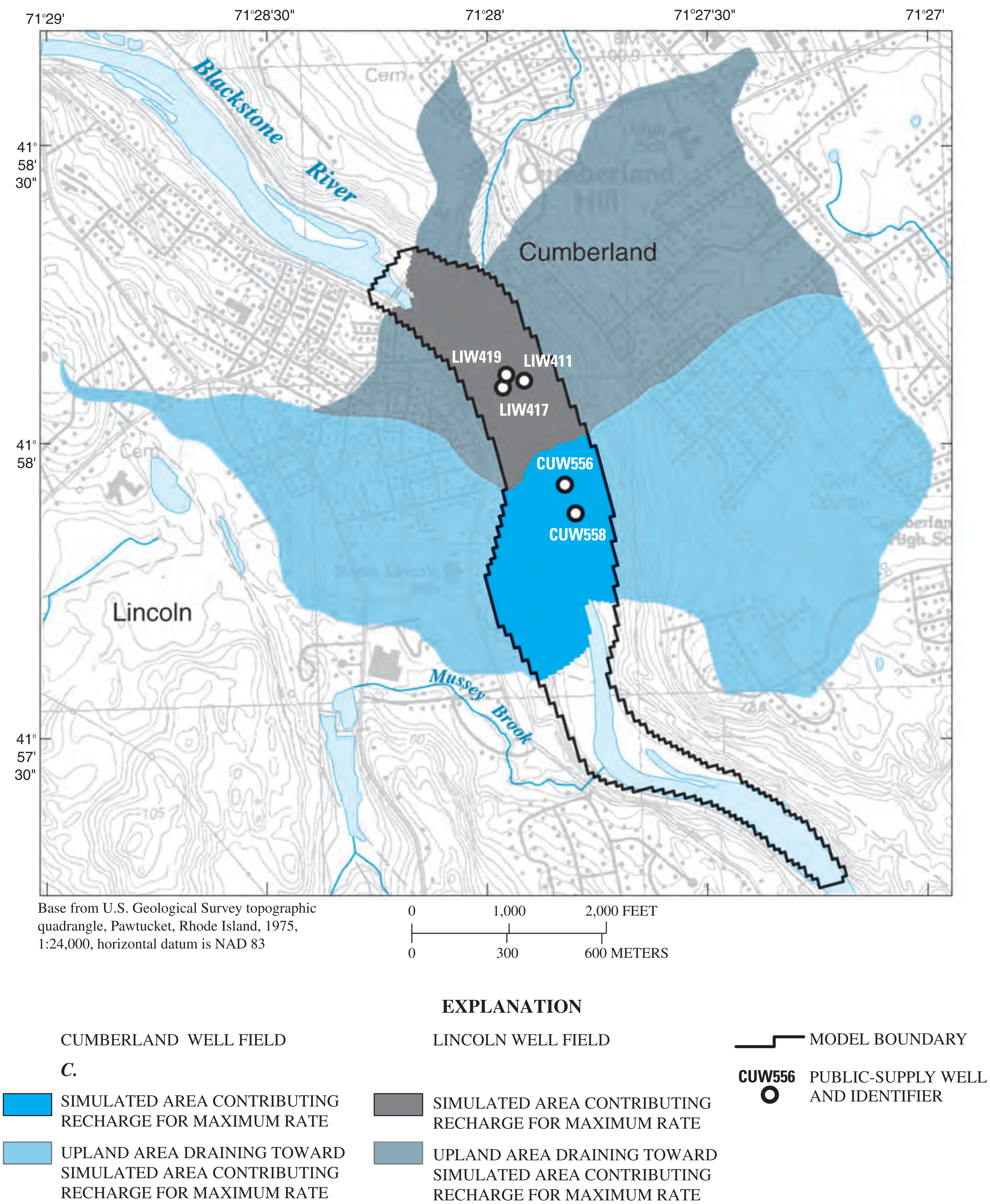


Figure 9—Continued. Areas contributing recharge in the modeled area and uplands draining toward the area contributing recharge for *A*, Cumberland well field average pumping rate of 324 gallons per minute; *B*, Cumberland well field maximum rate of 1,000 gallons per minute; and *C*, Cumberland and Lincoln well fields pumping simultaneously at their maximum pumping rates of 1,000 gallons per minute and 2,083 gallons per minute, respectively, Cumberland and Lincoln study area, Rhode Island.

Table 3. Sizes of areas contributing recharge to public-supply well fields and upland areas that drain toward the areas contributing recharge, and the percentage of the total water withdrawn from different source water, Cumberland and Lincoln study area, Rhode Island.

[gal/min, gallons per minute; in/yr, inches per year; mi^2, square mile; --, not computed]

Model scenario	Size of area contributing recharge (mi^2)	Upland area (mi^2)	Direct precipitation recharge (percent)	Upland runoff (percent)	Infiltration from river (percent)
Cumberland well field					
324 gal/min	0.05	0.58	21	72	7
1,000 gal/min	.12	.66	12	35	53
1,000 gal/min and vertical hydraulic conductivity of riverbed × 0.5	.13	--	17	37	46
1,000 gal/min and horizontal hydraulic conductivity × 2	.13	--	17	37	46
1,000 gal/min and recharge reduced by 6 in/yr	.13	--	12	25	63
Cumberland and Lincoln well fields pumping simultaneously					
Cumberland well field—1,000 gal/min	0.08	0.55	10	26	64
Lincoln well field—2,083 gal/min	.08	.34	2	10	88

North Smithfield Study Area

The North Smithfield Tifft Road public-supply well (NSW310) is in the Trout Brook Watershed, a subbasin of the Branch River in the Town of North Smithfield, north-central Rhode Island (figs. 1 and 11). The supply well is about 100 ft west of Trout Brook Pond, which is a ponded section of Trout Brook because of the Lower Slatersville Reservoir. The study area extends to features that serve as hydrologic boundaries in the numerical model: these features include the watershed boundary of Trout Brook in upland areas, the Slatersville Reservoirs, and the Branch River. A capped landfill designated a Superfund site by the U.S. Environmental Protection Agency is 0.7 mi south of NSW310 (fig. 11). Information about the direction of ground-water flow between the capped landfill and NSW310 would help water managers to determine if landfill contamination might affect the quality of water withdrawn from NSW310.

Well NSW310, constructed in 1963 originally to supply water to an industrial park, consists of a 15-ft screen completed 49 to 64 ft below land surface in 60 ft of saturated sediments (table 1). The maximum pumping rate of the well is about 200 gal/min (Michael Romano, North Smithfield Water Department, oral commun., 2001), about two-thirds less than its original projected yield, which has declined because of iron oxidation of sediments in the vicinity of the well and possible deterioration of the well screen (HydroSource Associates, Inc., 1999). A proposed replacement well has been estimated to yield 400–500 gal/min (HydroSource Associates, Inc., 1999). Areas contributing recharge were thus determined for the maximum pumping rate of both the current well (200 gal/min) and the proposed well (500 gal/min).

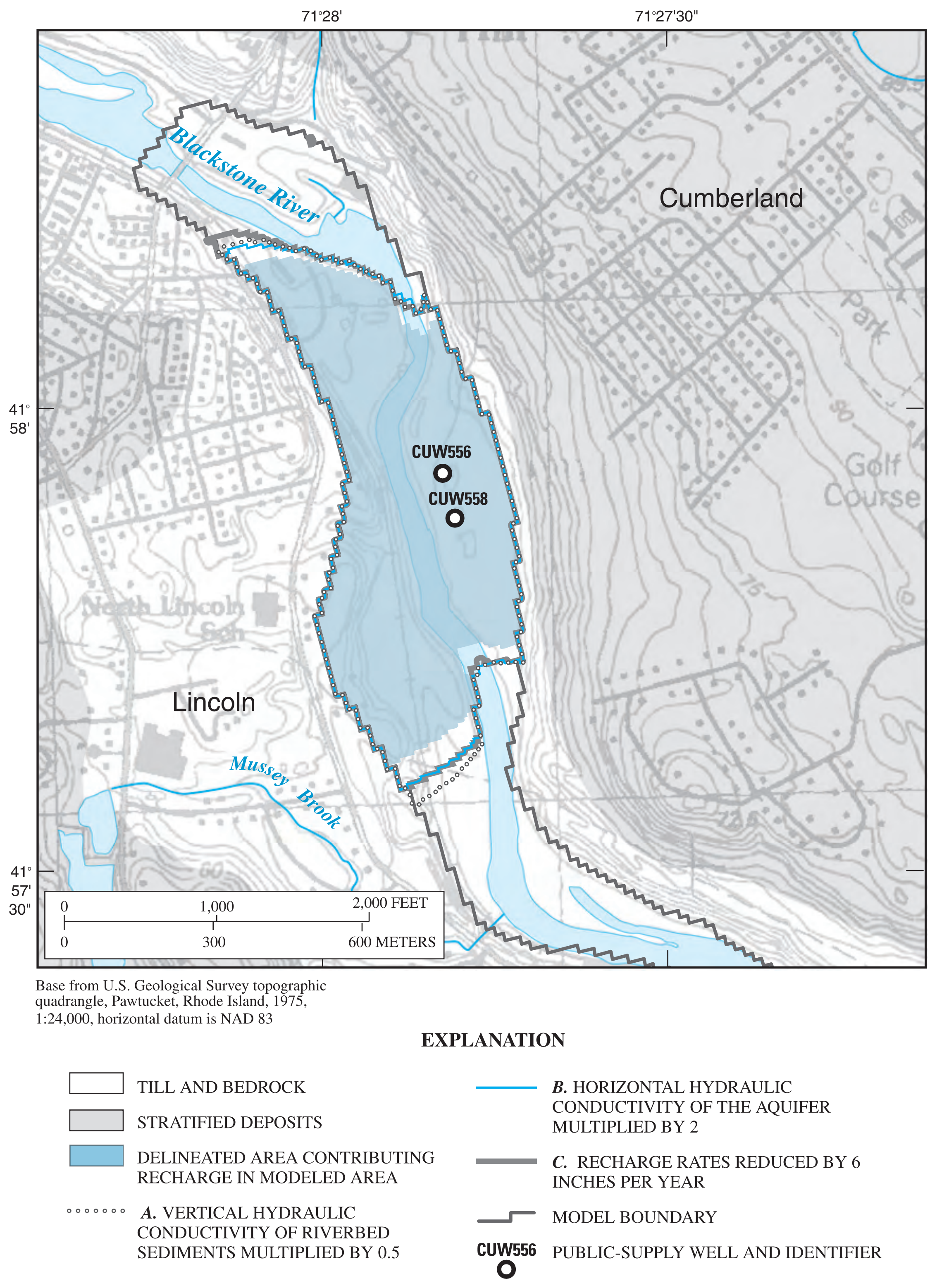


Figure 10. Areas contributing recharge in the modeled area calculated by a sensitivity analysis: *A*, vertical hydraulic conductivity of the riverbed sediments multiplied by 0.5; *B*, horizontal hydraulic conductivity of the aquifer multiplied by 2; and *C*, recharge rates reduced by 6 inches per year, compared to the delineated area contributing recharge to the Cumberland well field at a maximum rate of 1,000 gallons per minute, Cumberland and Lincoln study area, Rhode Island.

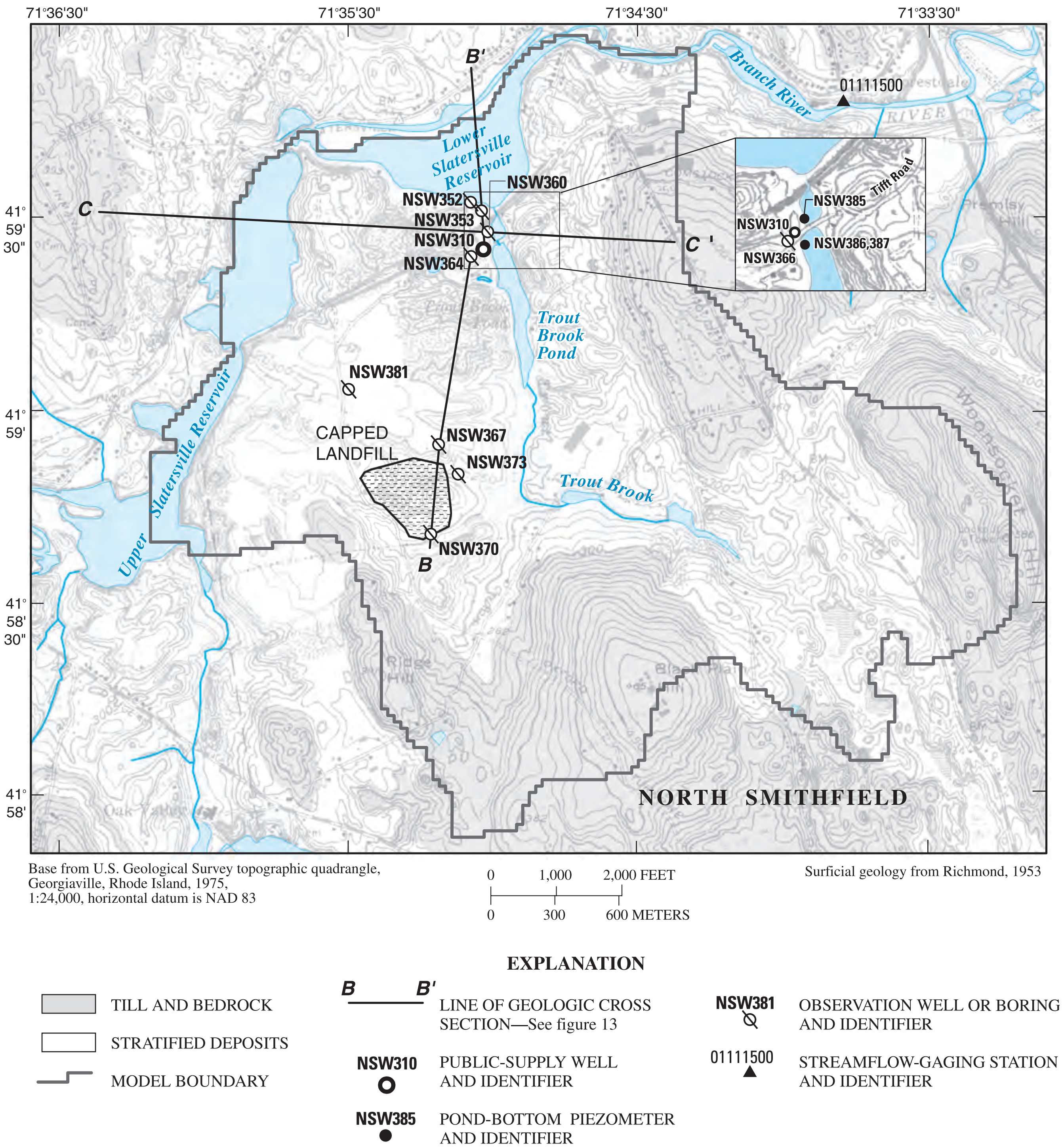


Figure 11. Public-supply well, section lines, selected borings, observation wells, and pond-bottom piezometers, model extent, and surficial geology, North Smithfield study area, Rhode Island.

Geohydrology

Surficial materials consist of stratified glacial deposits and till and are underlain by crystalline bedrock (Richmond, 1952). The areal extent of the surficial sediments mapped by Richmond (1953) is shown in figure 11 and the bedrock topography underlying the stratified deposits is shown in figure 12. The preglacial valley does not coincide with the current channel of the Branch River but trends in a southwest to northeast direction between the Upper Slatersville Reservoir and Trout Brook Pond. Stratified deposits thicker than 150 ft fill this preglacial valley, but its width was found to be narrower than the previously published width.

The vertical distribution and thickness of surficial sediments along a south-north geologic section (*B-B′*) and a west-east section (*C-C′*) through the public-supply-well site are shown in figure 13. Available lithologic logs, surficial geology, and interpretation of the geologic processes that formed the deposits were used in preparing the sections. Land-surface altitudes are based on the USGS Georgiaville topographical map, which was completed before sand and gravel mining altered the surface and thickness of surficial deposits in parts of the cross sections. Section *B-B′* shows three depositional sequences as the glacier retreated in a north-to-northeast direction (J.R. Stone, U.S. Geological Survey, oral commun., 2003). The collapsed end of the first sequence consists of poorly sorted sand and gravel and underlies the capped landfill. The second sequence extends from the capped landfill to the supply well. NSW310 is screened in sand and gravel sediments, which are laterally extensive (section *C-C′*). The third sequence consists primarily of sand that overlies the sand and gravel and also consists of fine-grained sediment beneath the Lower Slatersville Reservoir.

Dams built across the Branch River, originally to supply water to mills, created the Upper and Lower Slatersville Reservoirs. The average depths of the Upper and Lower Slatersville Reservoirs are 8 and 9 ft, respectively (Guthrie and Stolgitis, 1977). Mean annual streamflow of the unregulated Branch River was determined as 175 ft^3/s on the basis of records from the Branch River at Forestdale (01111500) streamflow-gaging station near the downstream end of the study area (figs. 1 and 11) for the period of record from 1940 to 2001.

Recharge from precipitation infiltrates the surficial materials and discharges to the reservoirs or other surface-water bodies in the study area. Water levels measured in observation well NSW381 (fig. 11) and stage measurements of the Upper Slatersville Reservoir and Trout Brook Pond indicate that the reservoir and pond are separated by a ground-water divide. Ground-water altitudes ranged from 0.60 to 2.70 ft higher than the stage of the Upper Slatersville Reservoir during this study. Water levels collected by GZA GeoEnvironmental, Inc. (1999) in July 1999 from three observation wells in a line between NSW381 and the Upper Slatersville Reservoir also indicated a ground-water divide.

In addition to precipitation recharge and upland runoff as potential sources of water withdrawn at NSW310, there are several potential surface-water sources in the vicinity of NSW310: Trout Brook Pond, Lower Slatersville Reservoir, and small kettle ponds. Trout Brook Pond, which is actually an extension of the Lower Slatersville Reservoir and, thus, at the same surface-water altitude, is connected to the reservoir by a narrow waterway about 2 ft deep. The withdrawal rate of 500 gal/min from the proposed replacement well, or 1.11 ft^3/s, is equal to 0.6 percent of the mean annual flow measured at the Branch River streamflow-gaging station.

Three profiles of water depth and the thickness of organic-rich, fine-grained, pond-bottom sediment across Trout Brook Pond, were measured with a hand-held steel rod, and indicated that the average water depth was 5 ft and that the thickness of fine-grained, pond-bottom sediments ranged from 0 to 5 ft. Ground-water levels measured in piezometers installed at two locations at shallow depths beneath the pond (NSW385 and NSW386-387, fig. 11) were consistently higher than the pond stage; these levels indicate that the pond was gaining water from the aquifer even at the 200 gal/min pumping rate. Measurements were taken intermittently during different seasons and during different pumping cycles; one set of measurements was made as long as 24 hours after pumping was started. Ground-water levels also were measured continuously from January 14 to 18, 2002 (fig. 14) in one of the pond-bottom piezometers. Although ground-water levels responded to pumping cycles, these levels never dropped below the two stage measurements taken January 14 and 18, 2003, confirming that the pumping rate of 200 gal/min did not induce pond water from Trout Brook Pond closest to the supply well.

Model Design

Ground-water flow in the study area was simulated by a three-layered variable-grid model. Grid spacing ranged from 10 ft to 100 ft with finer spacing designed to increase detail in areas near NSW310 and Trout Brook Pond. The model subdivides an area of 3.3 mi^2 into 145 rows and 151 columns with 38,223 active cells; the simulated area includes both stratified glacial deposits and till-covered bedrock. The model extent coincides with the study-area boundary (fig. 11); the ground-water divide between the Upper Slatersville Reservoir and Trout Brook Pond was not used as a lateral boundary because the exact location of this divide is unknown and because different pumping rates may shift the location of this divide.

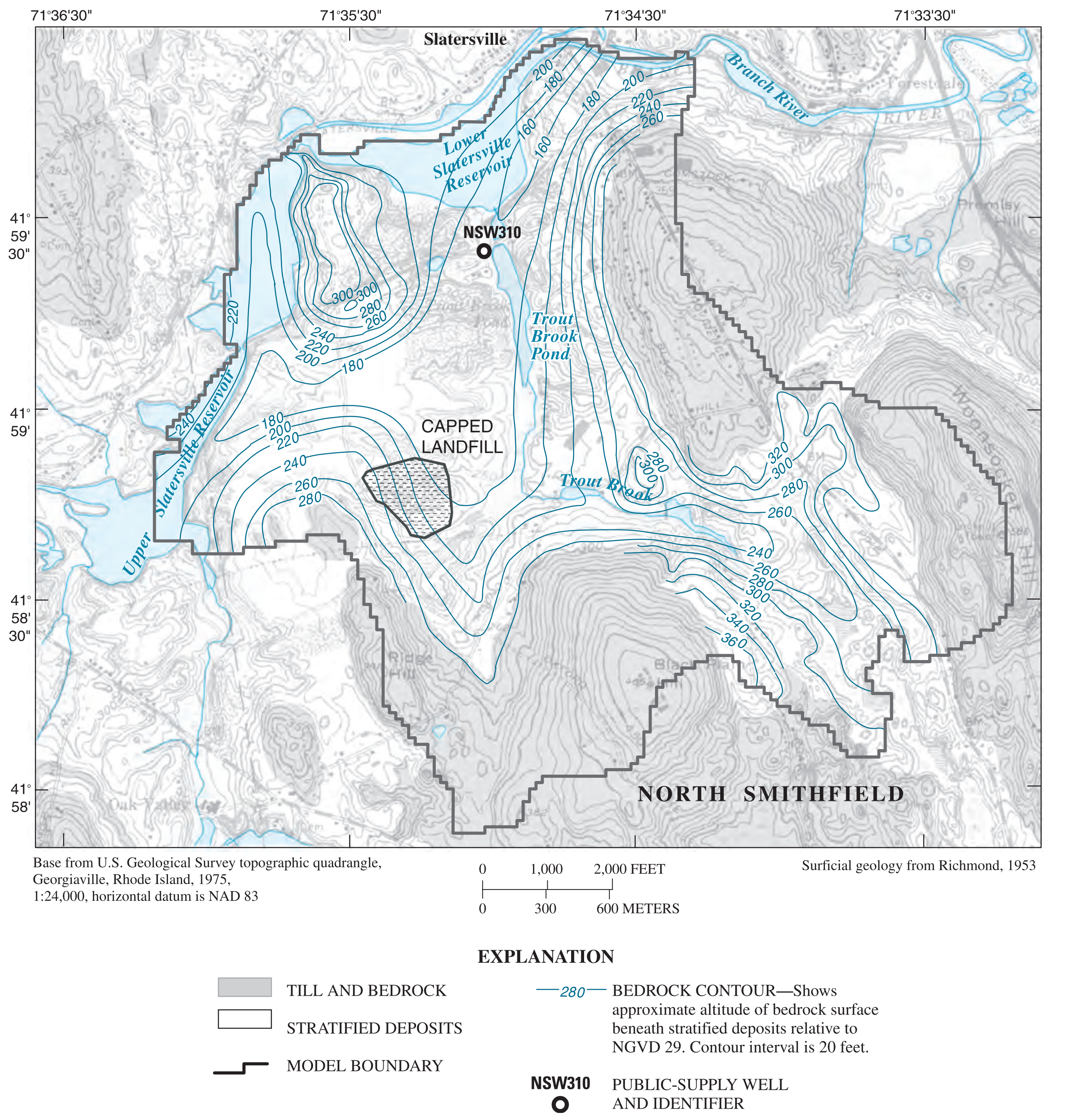


Figure 12. Bedrock-surface contours, North Smithfield study area, Rhode Island.

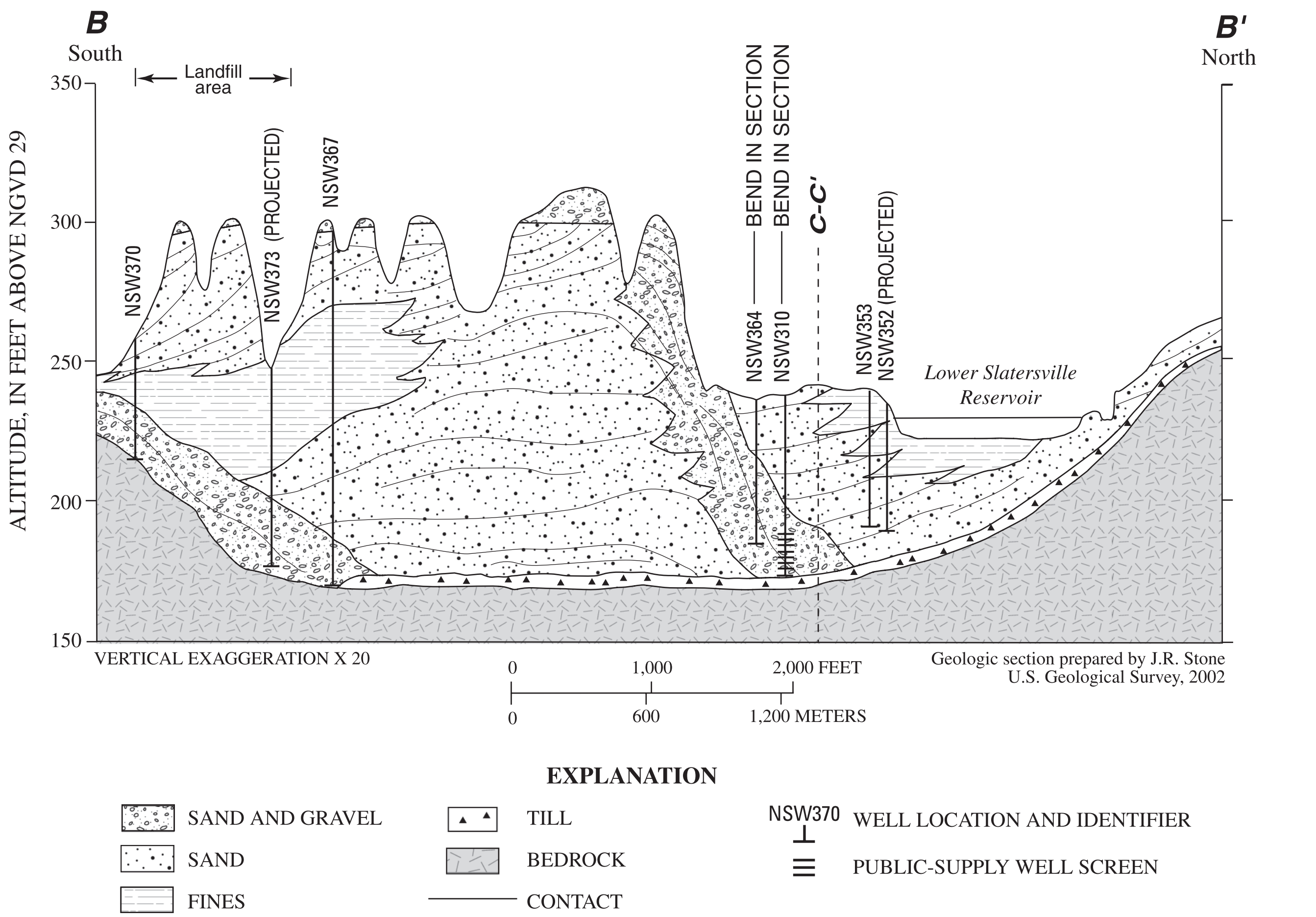


Figure 13. Geologic cross sections (lines of sections *B-B′* and *C-C′* are shown in fig. 11), North Smithfield study area, Rhode Island.

Vertical discretization is based on lithology and screen placements of the supply and observation wells. Stratified glacial deposits are represented in the top two layers (layers 1 and 2) of the model, whereas till and bedrock are represented in all three layers (fig. 15). The bottom altitude of layer 1 is 210 ft or higher where it forms the geologic contact between the stratified deposits and bedrock. Where saturated stratified deposits are less than 20 ft thick, layer 1 represents bedrock. The bottom altitude of the middle layer, layer 2, represents the contact between stratified deposits and bedrock at altitudes less than 210 ft; layer 2 also includes the public-supply-well screen. Layer 3 represents only bedrock. Bedrock has a constant thickness of 300 ft throughout the model, either beneath the stratified deposits or the land surface in the uplands.

Surface-water bodies were simulated in the model as head-dependent flux boundaries by using the MODFLOW river and drain packages (Harbaugh and McDonald, 1996) in layer 1 (fig. 16). Surface-water altitudes were determined from field measurements made for this study, a landfill study (E.C. Jordan, Inc., 1988), and interpolated between USGS Georgiaville topographical-map contours. The altitude of the reservoirs and Trout Brook Pond beds were based on average water depths, whereas the rivers and streams were assumed to be 1 ft deep. Bed thickness is unknown for most of the surface-water bodies, so a bed thickness of 1 ft was assigned in the model. Initial values of vertical hydraulic conductivity for the bed sediments were 0.5 ft/d for water bodies with slow velocities because fine-grained sediments tend to settle on the bottom, and 2 ft/d for the beds of the fast-moving river and streams. The extent of Trout Brook Pond near NSW310 was determined from a large-scale orthophoto (1:5,000); small kettle-hole ponds near NSW310, some of which do not appear on the USGS topographical map, were not simulated in the model because of the minimal amount of water stored in these ponds. Narrow water bodies, such as Trout Brook and intermittent streams near the landfill, were simulated as 10 ft and 5 ft wide, respectively.

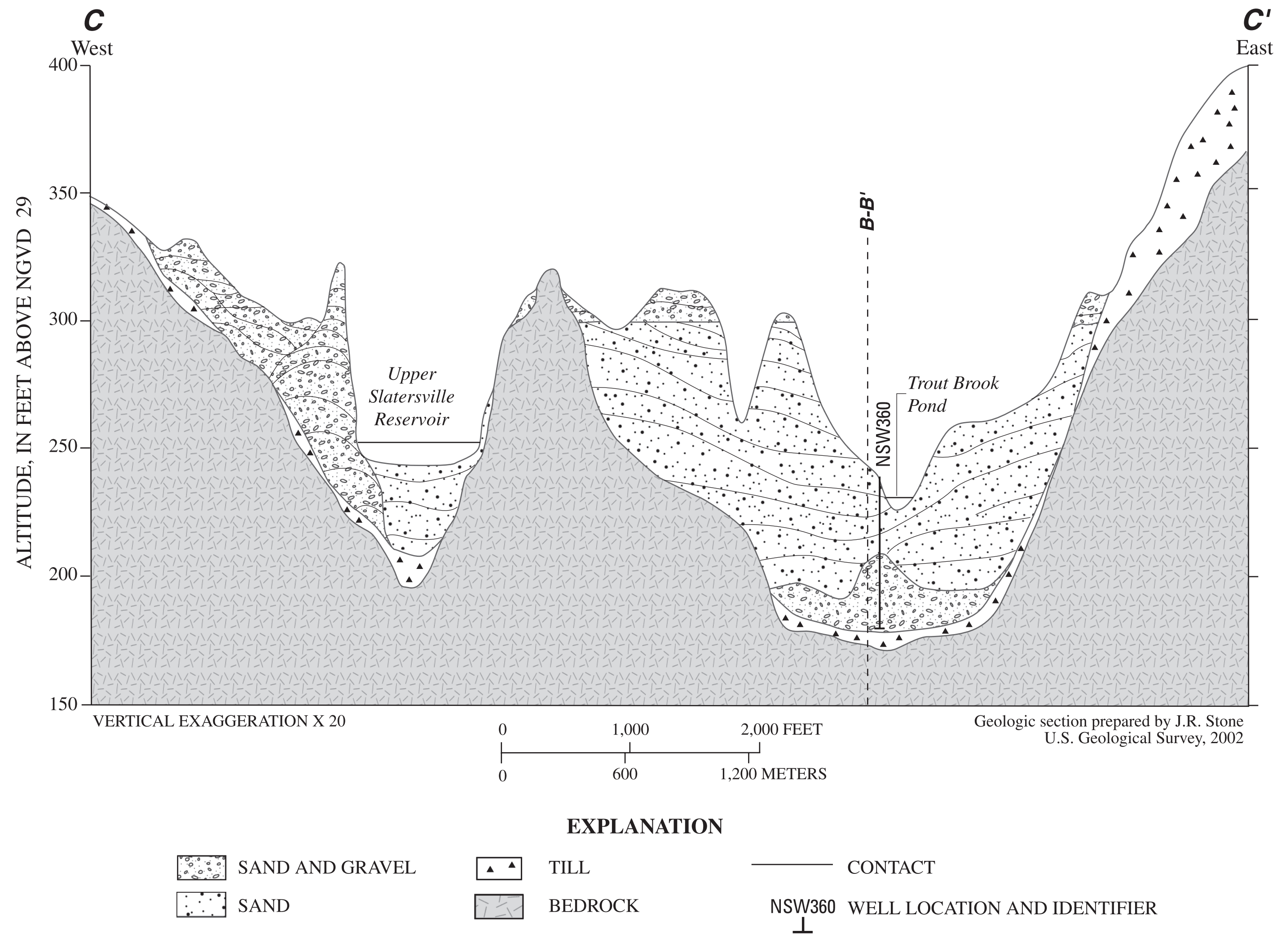


Figure 13—Continued. Geologic cross sections (lines of sections *B-B′* and *C-C′* are shown in fig. 11), North Smithfield study area, Rhode Island.

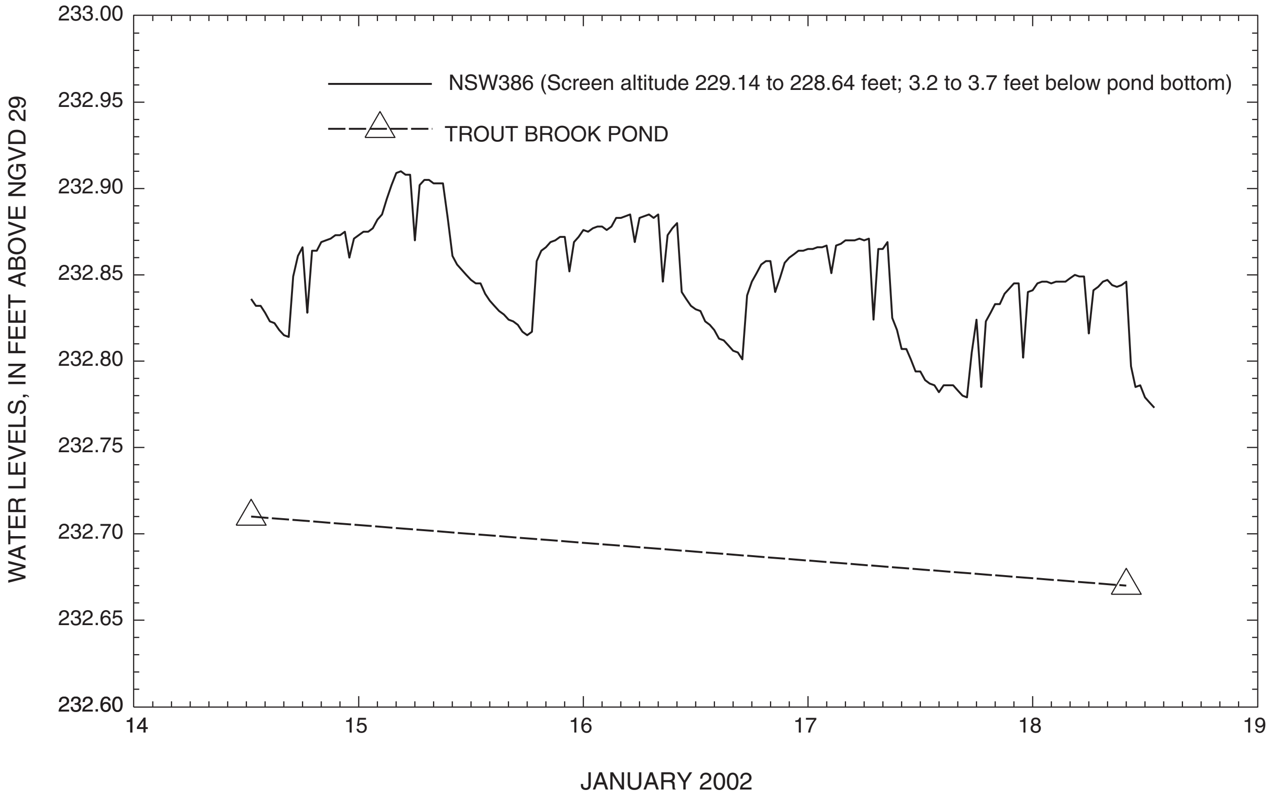


Figure 14. Water levels in pond-bottom piezometer NSW386 and stage of Trout Brook Pond, North Smithfield study area, Rhode Island.

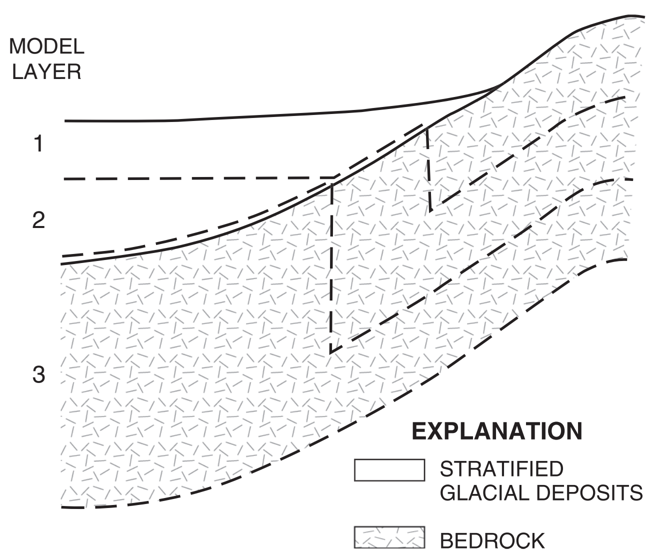


Figure 15. Schematic section showing model layers, North Smithfield study area, Rhode Island.

The top boundary of the model is the water table and is calculated by the model. Twenty-six inches of areal recharge (Randall, 1996), a value which is consistent with long-term mean annual runoff from the Branch River streamflow-gaging station, was applied to the stratified deposits to represent water from direct infiltration of precipitation. This same rate of flux was applied to the uplands but represents both recharge to ground water in the uplands and runoff that recharges the stratified deposits near the contact between the uplands and the stratified deposits. Thus, all water that is added to the ground-water system through precipitation is accounted for in the model. Because the former landfill is capped with an impermeable layer, most of the recharge that would normally infiltrate was applied to three detention basins surrounding the capped landfill. The remaining recharge, 6 in/yr, was applied as infiltrating the sediments at the downgradient edges of the landfill.

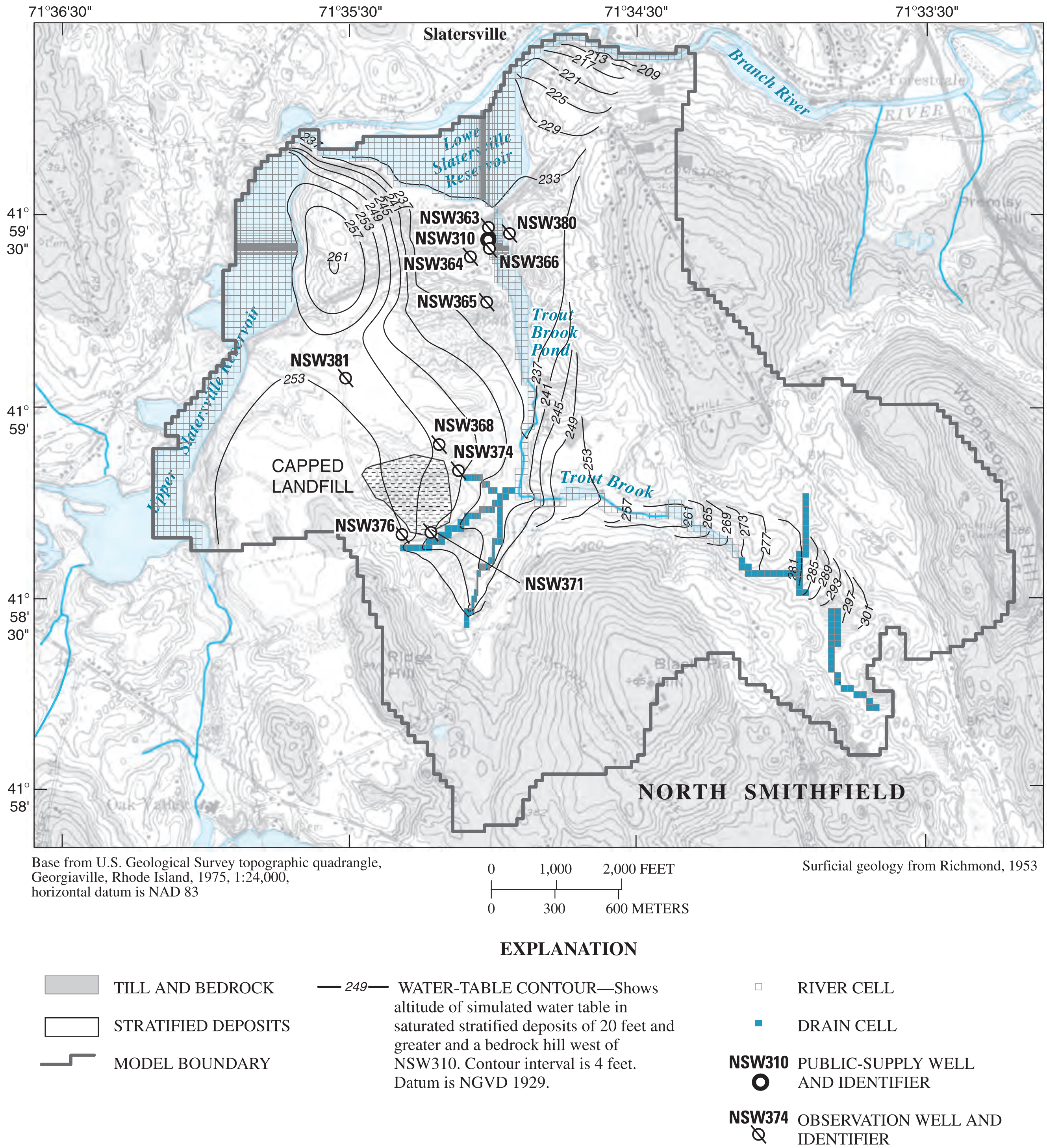


Figure 16. Model-boundary types, simulated water-table contours for steady-state, nonpumping conditions, and location of observation wells, North Smithfield study area, Rhode Island.

Initial horizontal hydraulic-conductivity values specified in the model for stratified deposits were assigned on the basis of lithology and the results of aquifer tests at NSW310 (Johnston and Dickerman, 1974b) and the capped landfill (E.C. Jordan, Inc., 1988). Stratified deposits were distributed among six zones, two each for sand and gravel, sand, and fine-grained deposits. Sand and gravel was assigned values of 200 ft/d for sediments intersecting the supply well-screen and 60 ft/d for poorly sorted sand and gravel at the capped landfill. Sand, consisting mostly of fine to coarse or fine to medium sand, and representative of sand deposits near the supply well and most sand in the modeled area, was assigned a value of 100 ft/d. Average hydraulic conductivity calculated at NSW310 from specific-capacity data (Johnston and Dickerman, 1974b) compares favorably to assigned hydraulic-conductivity values based on lithology. Fine sand was assigned a horizontal hydraulic conductivity of 50 ft/d. Values assigned for fine-grained sediments ranged from 15 ft/d near the Lower Slatersville Reservoir to 1 ft/d at the capped landfill. The till and bedrock unit was assigned a value of 0.5 ft/d, representative of crystalline bedrock.

Model Calibration

The ground-water-flow model was calibrated to water-level measurements made on May 8, 2002, under steady-state, nonpumping conditions, and to drawdown measurements collected during a transient 5-day aquifer test done on April 21 through 26, 1999, by HydroSource Associates, Inc. (1999). Simulated steady-state ground-water levels were used as initial conditions for the transient simulation.

Ten ground-water-level and two surface-water-level measurements made in the study area on May 8, 2002, approximate long-term average annual conditions on the basis of a nearby USGS long-term well BUW187, 4 mi west of the study area, and streamflow-gaging station Branch River at Forestdale (figs. 1 and 7). Water levels were measured 1.5 days after NSW310 stopped pumping, when ground-water levels near the well returned to prepumping levels. Surface-water levels from the Upper Slatersville Reservoir and Trout Brook Pond were used to specify the altitudes of these water bodies in the model; the altitude of the water level in Trout Brook Pond also was used for the altitude of the Lower Slatersville Reservoir.

Initial values of horizontal hydraulic conductivity and the degree of hydraulic connection between surface water and the underlying aquifer were adjusted within reasonable limits during model calibration to improve the match between measured and simulated water levels. Horizontal hydraulic conductivities for the two sand zones, representing predominately fine to coarse sand and fine sand, were lowered from 100 ft/d to 70 ft/d and from 50 ft/d to 20 ft/d, respectively, to increase ground-water altitudes and form a divide in the preglacial valley between the Upper Slatersville Reservoir and Trout Brook Pond. Horizontal hydraulic conductivity of the fine sediments at the capped landfill was raised from 1 ft/d to 10 ft/d. Vertical hydraulic conductivity of the bed sediments in Trout Brook Pond and Lower Slatersville Reservoir was changed from 0.5 to 1.0 ft/d to increase the connection between the surface-water bodies and the underlying aquifer; this change lowered water levels in the adjacent aquifer. A comparison of measured to simulated ground-water altitudes is shown for each observation well in table 4. The range in differences between measured and simulated ground-water altitudes is from -0.76 to 0.77 ft. The mean absolute value is 0.46 ft, which is about 2 percent of the total difference in ground-water altitude, 19.28 ft, measured at the observation wells.

The altitude and configuration of the simulated water table for steady-state nonpumping conditions in saturated stratified deposits of 20 ft and greater and a bedrock hill west of NSW310 are shown in figure 16. Simulated ground-water flow directions, which are perpendicular to the water-table contours, generally agree with results of a numerical model encompassing most of the North Smithfield study area by GZA GeoEnvironmental, Inc. (1999), and reported by the U.S. Environmental Protection Agency (Anna Krasko, U.S. Environmental Protection Agency, written commun., 2000). The calculated total flow rate through the ground-water system is 560,800 ft^3/d. Of the total inflow rate, 498,050 ft^3/d is from recharge, which represents direct precipitation recharge and upland runoff, and the remaining inflow, 62,750 ft^3/d, is seepage from the downgradient ends of the Slatersville Reservoirs, especially the lower reservoir where the valley is relatively wide. The actual seepage rate is unknown, but the downstream ends of the reservoirs are likely outside of the source area for the public-supply well.

Table 4. Ground-water altitudes measured on May 8, 2002, and simulated ground-water altitudes for the steady-state model in the North Smithfield study area, Rhode Island.

[ft, foot]

Observation well	Model layer	Measured (ft above NGVD 29)	Simulated (ft above NGVD 29)	Difference (measured minus simulated, in ft)
NSW363	2	233.71	233.95	-0.24
NSW364	2	234.81	234.57	.24
NSW365	1	235.63	234.86	.77
NSW366	2	233.96	234.11	-.15
NSW368	1	248.78	248.44	.34
NSW371	1	245.70	245.03	.67
NSW374	1	245.29	246.05	-.76
NSW376	1	252.49	253.02	-.53
NSW380	2	233.33	233.72	-.39
NSW381	1	252.61	252.07	.54

The calibrated model based on steady-state conditions was then extended to transient conditions of the 5-day aquifer test, when NSW310 was pumped at 200 gal/min, by incorporating storage properties and changing the recharge rate and pond stage to actual conditions during the aquifer test. During the aquifer test, HydroSource Associates, Inc. (1999), measured 0.60 in. of precipitation. All of this precipitation was added to the model as recharge because evapotranspiration and the soil-moisture deficit were assumed to be negligible in April. In addition to this recharge from precipitation, a constant inflow from the uplands to the stratified deposits equal to the steady-state rate was applied. A visual comparison of measured drawdowns at observation well NSW366, 45 ft south of NSW310, to simulated drawdowns confirms that the input properties from the calibrated steady-state model provide a reasonably close match (fig. 17); however, this analysis represents only a small part of the aquifer near the supply well.

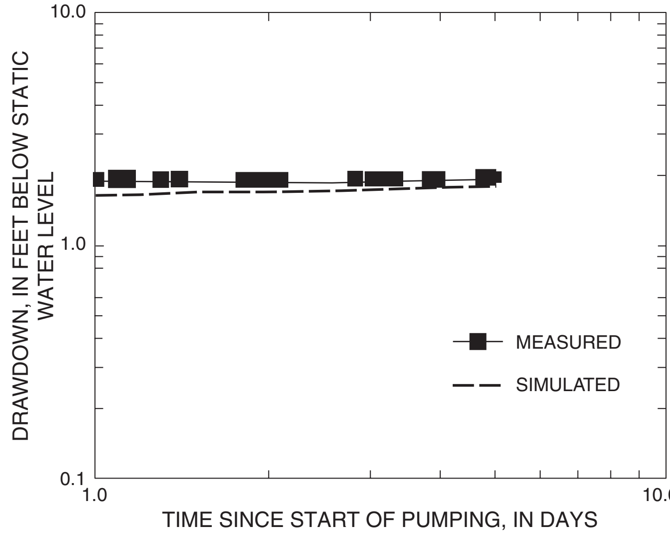


Figure 17. Measured and simulated drawdowns in observation well NSW366 during NSW310 aquifer test, North Smithfield study area, Rhode Island.

Delineation of Areas Contributing Recharge and Sensitivity Analysis

Areas contributing recharge to NSW310 and to the proposed replacement well were determined on the basis of the calibrated model and average recharge rates. The location and extent of the area contributing recharge to NSW310 pumping at its maximum rate of 200 gal/min and to the replacement well at its estimated maximum yield of 500 gal/min are shown in figure 18; the sizes of the areas and sources of water are listed in table 5. The areas contributing recharge for both pumping rates extend to the ground-water divides on both sides of Trout Brook Pond and exclude recharge from the capped landfill south of the well site. Traveltime estimates based on a porosity of 0.35 for stratified sediments (Allen and others, 1963) and 0.02 for bedrock (Randall and others, 1966) indicate that water that recharges the aquifer in the vicinity of the capped landfill and within the area contributing recharge takes about 30 years to reach the well site.

The area contributing recharge for the 200 gal/min rate covers about 0.23 mi^2. At this rate, a small quantity of pumped water (0.04 percent) is induced infiltration from Trout Brook Pond; this quantity is consistent with water levels measured during this study discussed in the Geohydrology section. Even by increasing the vertical hydraulic conductivity between the pond and the aquifer by an order of magnitude from 1.0 ft/d to 10 ft/d, which is still less than the vertical hydraulic conductivity of the aquifer, the well derived only 1 percent of its water from induced infiltration of pond water. Thus, even with the well pumping at a rate of 200 gal/min only 100 ft from a surface-water source, the well derived almost all of its water from intercepted ground water that normally discharges to the pond and the Lower Slatersville Reservoir.

At the proposed replacement well's withdrawal rate of 500 gal/min, the area contributing recharge is 0.45 mi^2. The increased pumping rate is balanced by expansion of the area contributing recharge to intercept additional ground water that would have flowed to surface-water bodies and by inducing 25 percent of the total pumpage from surface water. Trout Brook Pond contributes most of the induced infiltration compared to the Lower Slatersville Reservoir.

Sensitivity analysis of the area contributing recharge and sources of water based on reasonable alternative model-input values was examined for the proposed well-withdrawal rate of 500 gal/min. Alternative values were chosen that may provide larger areas contributing recharge than the delineated contributing area. The model, however, was not recalibrated; the analysis is intended to provide insights into the effects of selected model-input values on the extent of the area contributing recharge. Results of the sensitivity analysis are shown in figure 19 and table 5.

The hydraulic connection between surface-water bodies near the well site was reduced by an order of magnitude by decreasing the vertical hydraulic conductivity of bed sediments from 1 to 0.1 ft/d. Infiltration of surface water decreased from 25 to 14 percent and the area contributing recharge increased from 0.45 mi^2 to 0.52 mi^2. Reducing the annual recharge rate from 26 in/yr to 20 in/yr caused the area contributing recharge to increase from 0.45 mi^2 to 0.51 mi^2 and the proportion of water infiltrating from the river to increase from 25 to 34 percent.

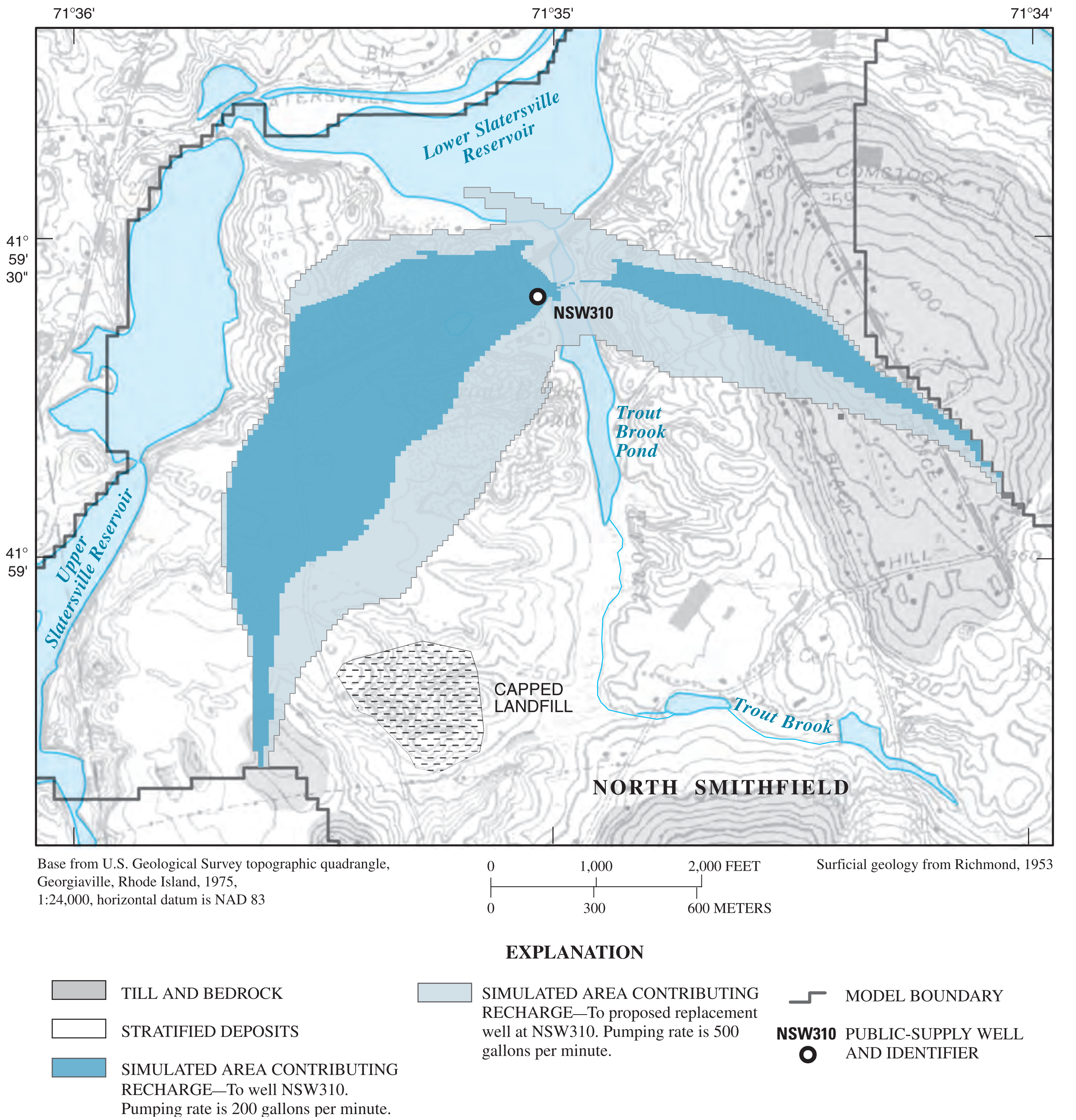


Figure 18. Simulated area contributing recharge to public-supply well NSW310 pumping at a rate of 200 gallons per minute and a proposed replacement well at NSW310 pumping at a rate of 500 gallons per minute, North Smithfield study area, Rhode Island.

Table 5. Sizes of areas contributing recharge to public-supply well NSW310 and a proposed replacement well and the percentage of the total water withdrawn from different source water, North Smithfield study area, Rhode Island.

[gal/min, gallons per minute; in/yr, inches per year; mi^2, square mile]

Model scenario	Size of area contributing recharge (mi^2)	Infiltration of surface water (percent)	Direct precipitation recharge and upland runoff (percent)
NSW310			
200 gal/min	0.23	0.04	99.96
Proposed replacement well			
500 gal/min	0.45	25	75
500 gal/min and vertical hydraulic conductivity of bed sediments × 0.1	.52	14	86
500 gal/min and recharge reduced by 6 in/yr	.51	34	66
500 gal/min and ground-water divide shifted eastward	.46	28	72

The location of the ground-water divide between the Upper Slatersville Reservoir and Trout Brook Pond during average conditions is unknown and could not be accurately defined unless water levels were measured in several observation wells on both sides of the divide. The effect of shifting the simulated ground-water divide eastward on the area contributing recharge, especially in relation to the capped landfill, was made by decreasing horizontal hydraulic-conductivity values for sand and sand and gravel by one-half east of the simulated divide and doubling values for the same variables west of the divide. These changes in hydraulic conductivity shifted the simulated divide 400 to 600 ft eastward, caused the area contributing recharge to extend farther north, and induced additional water to infiltrate from the Lower Slatersville Reservoir. The capped landfill was still excluded from the area contributing recharge.

Westerly Study Area

The Westerly Crandall public-supply well (WEW584) is in the Town of Westerly in southwestern Rhode Island (figs. 1 and 20). Characteristics of the public-supply well are listed in table 1. WEW584 is screened 47 to 62 ft below land surface in about 70 ft of saturated stratified sediments. These permeable sediments underlie a wetland of large areal extent. Conceptually, the degree of hydraulic connection between the wetland and the stratified sediments is the most important factor affecting the size of the area contributing recharge to the well. Areas contributing recharge were determined for both the average pumping rate, 240 gal/min, and for the maximum rate, 700 gal/min.

The geographic extent of the study area includes a subbasin of the Pawcatuck River that contains the well. The study area also extends to Block Island Sound, which serves as the southern boundary of the numerical model (fig. 20).

Geohydrology

Well WEW584 is screened in coarse stratified sediments in a watershed dominated by a densely vegetated wetland (fig. 20). The wetland is considered a palustrine forested and scrub-shrub wetland (Tiner, 1989), according to a classification scheme by Cowardin and others (1979). Wetland deposits consist of peat—organic matter in various stages of decomposition. An investigation of the wetland deposits in the watershed by Hughes (1982) was done to determine the fuel potential of the peat and included an extensive data set based on examination of core samples and probing of the peat to determine its thickness and type. Peat deposits range in thickness from 0 ft at the edges of the wetland to about 8 ft in its center and generally include a thin layer of organic-rich silt and clay up to 8 in. thick near the bottom of the layer.

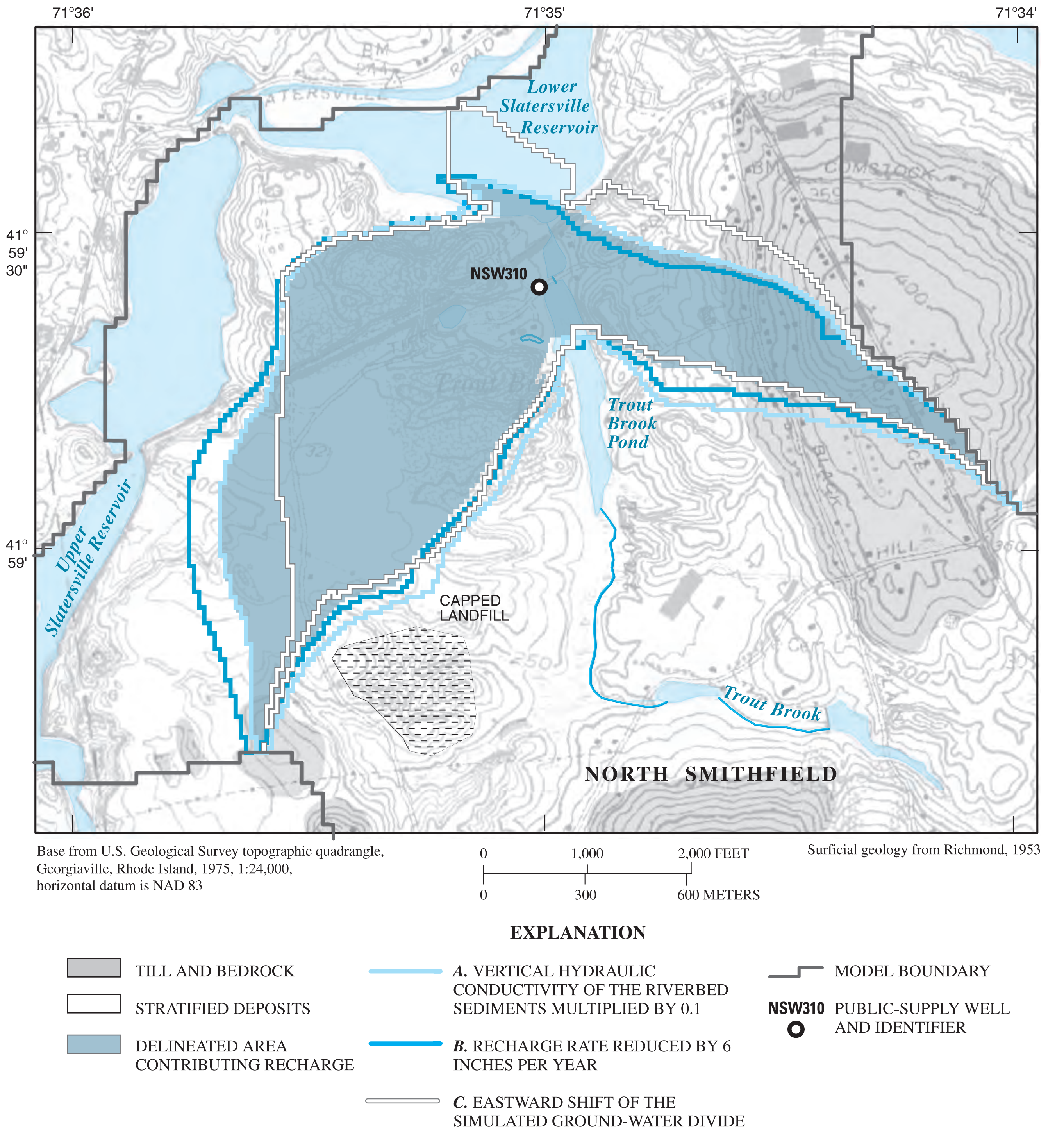
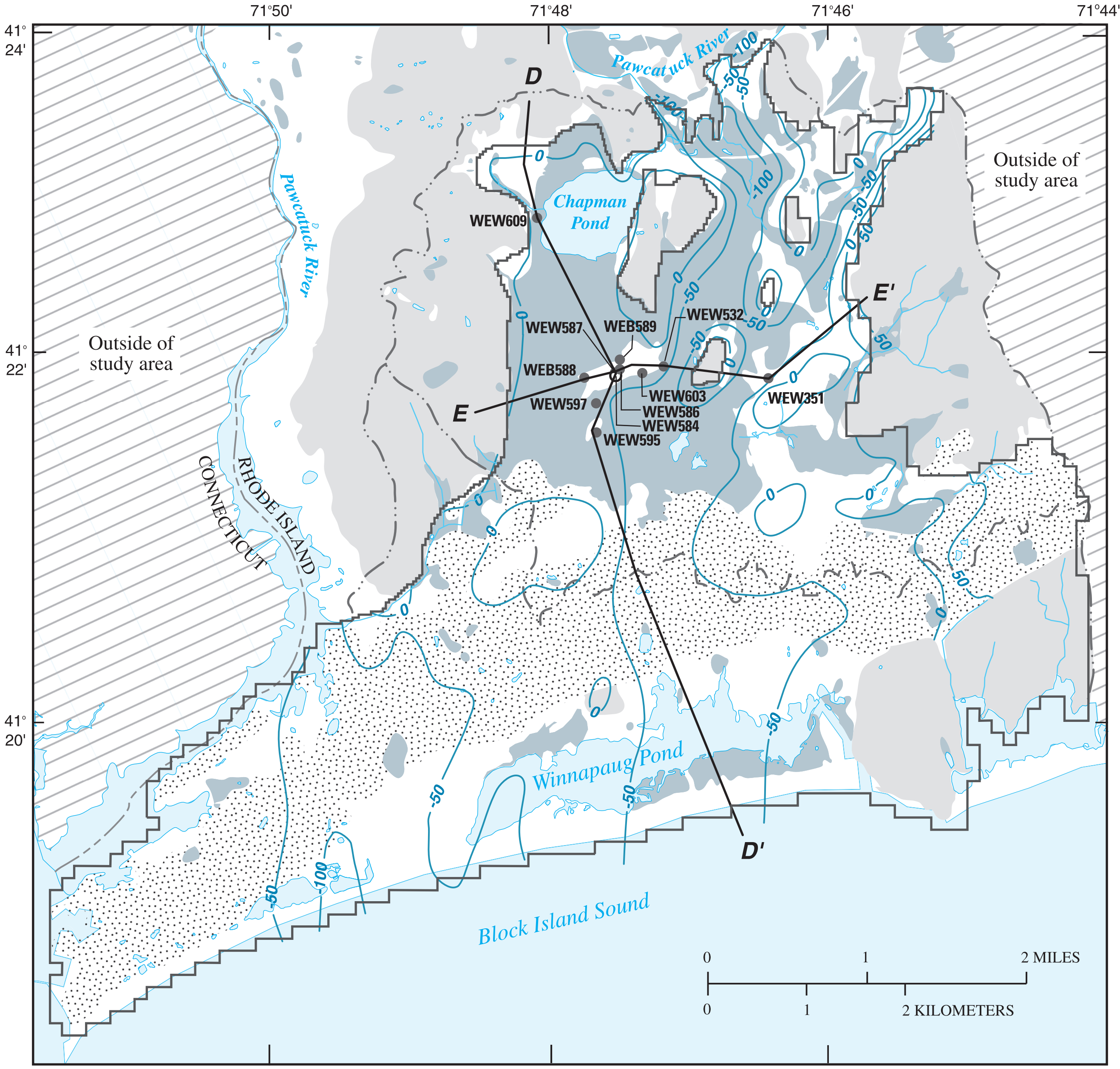


Figure 19. Areas contributing recharge calculated by a sensitivity analysis: *A*, vertical hydraulic conductivity of the riverbed sediments multiplied by 0.1; *B*, recharge rates reduced by 6 inches per year; and *C*, an eastward shift of the simulated ground-water divide between Upper Slatersville Reservoir and Trout Brook Pond, compared to the delineated area contributing recharge for the proposed replacement well at NSW310 pumping at a rate of 500 gallons per minute, North Smithfield study area, Rhode Island.



Base from U.S. Geological Survey topographic quadrangles, Ashaway, 1975, Watch Hill, 1984, Carolina, 1970, Quonochontaug, 1970, Rhode Island, 1:24,000, horizontal datum is NAD 83

Surficial geology from Shafer, 1965, 1968

EXPLANATION

- TILL AND BEDROCK
- STRATIFIED DEPOSITS
- MORAINE
- WETLAND AND PEAT
- 50 BEDROCK CONTOUR—Shows approximate altitude of bedrock surface. Contour interval is 50 feet. Datum is NGVD 29.
- MODEL BOUNDARY
- WATERSHED BOUNDARY
- WEW584 PUBLIC-SUPPLY WELL AND IDENTIFIER
- WEW597 BORING AND IDENTIFIER
- D D' LINE OF GEOLOGIC SECTION—See figure 21.

Figure 20. Public-supply well, section lines and borings, surficial geology, bedrock-surface contours, and model boundary, Westerly study area, Rhode Island.

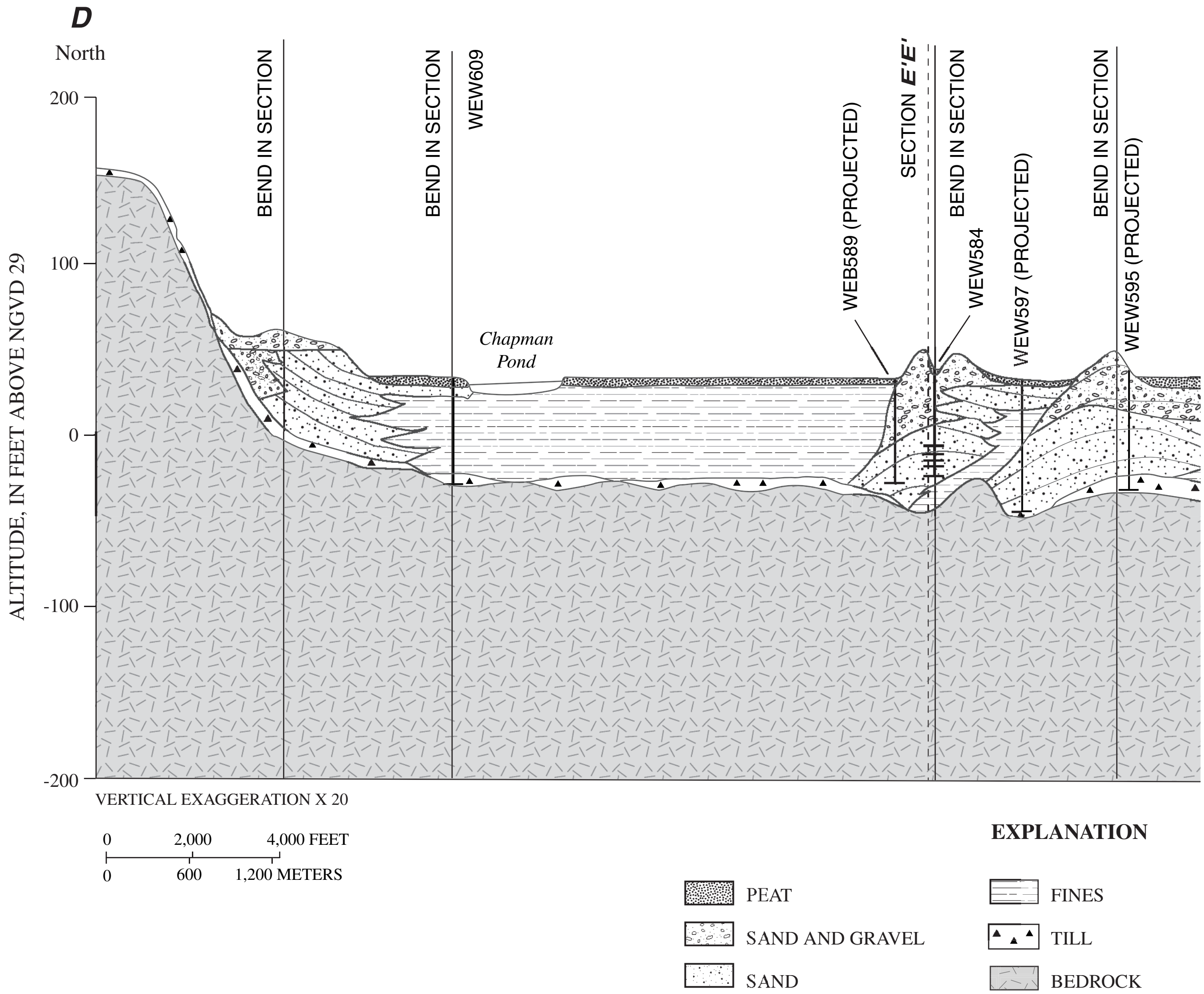
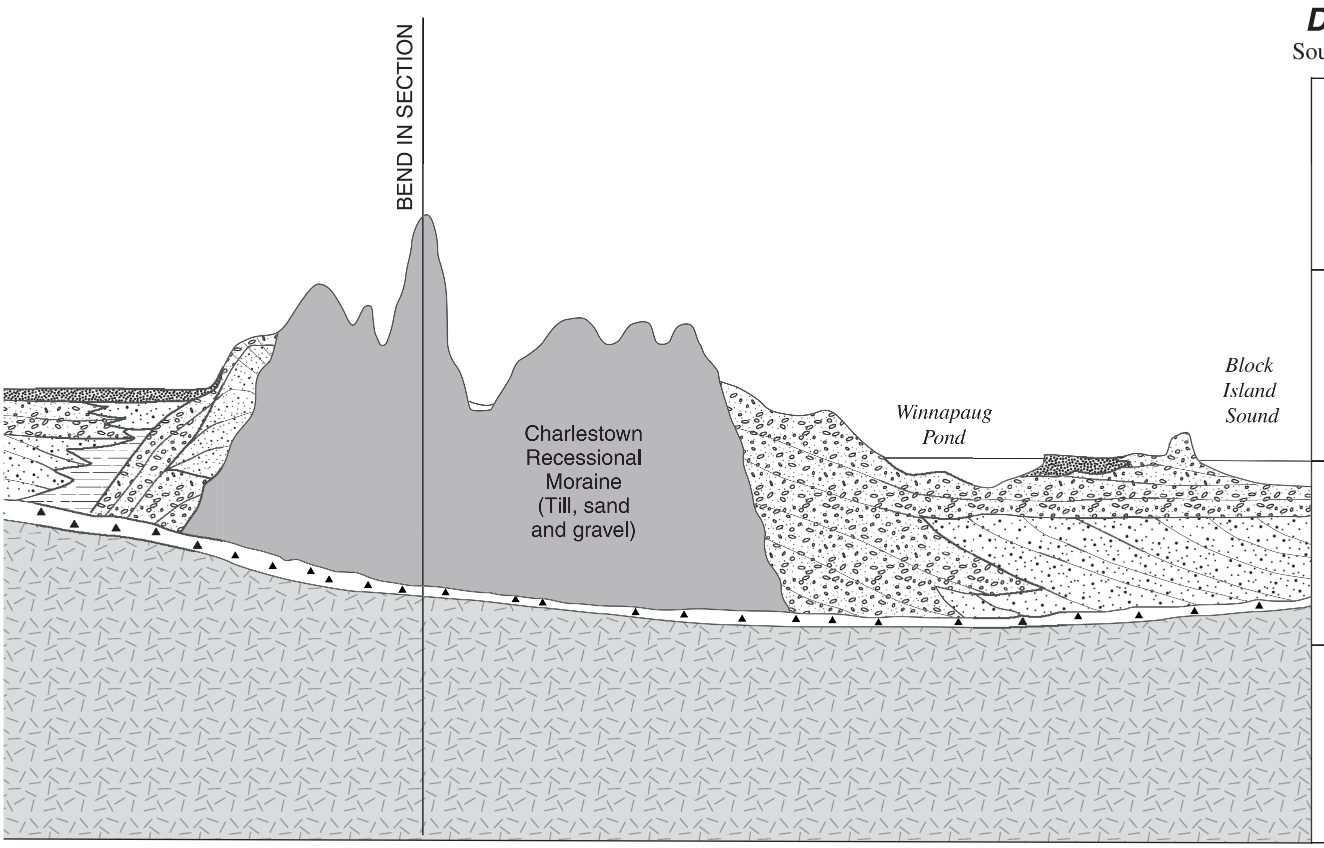


Figure 21. Geologic cross sections (lines of sections *D-D′* and *E-E′* are shown in figure 20), Westerly study area, Rhode Island.

Sediments of glacial origin—stratified sediments, till, and moraine—border and underlie the post-glacial wetland deposits and fill a generally north-south trending preglacial river valley (fig. 20). The axis of this preglacial valley was interpreted by Gonthier and others (1974) to exit to Block Island Sound southeast of the well site. Borings and seismic-refraction surveys completed by Gonthier and others (1974) west and east of the valley axis indicated shallow bedrock depths below the moraine. Bedrock altitude in the area interpreted as the axis of the preglacial valley is unknown but two borings indicated the bedrock surface to be at least 3 and 24 ft below sea level. Lithologic logs collected during this study indicate shallower bedrock altitude in the public-supply-well area and surrounding wetland and thus thinner saturated sediments than those published by Gonthier and others (1974).

Geologic sections drawn approximately parallel and perpendicular to the valley axis through the public-supply-well site show the thickness and lithology of the surficial deposits (fig. 21). Section *D-D′* shows the Charlestown moraine and a series of depositional sequences as the glacier retreated to the north-northwest (J.R. Stone, U.S. Geological Survey, oral commun., 2002). The Charlestown moraine represents a long-term recessional position of the retreating glacier and is described as consisting of loose, sandy till, including lenses of



Geologic section prepared by J.R. Stone
U.S. Geological Survey, 2002

CONTACT

WEW584 WELL LOCATION AND IDENTIFIER

PUBLIC-SUPPLY WELL SCREEN

sorted sand and gravel, which is in contrast to the compacted, unsorted till that forms a thin layer over the bedrock throughout the study area (Schafer, 1965). The few lithologic logs available from the moraine indicate that the peninsula formed by the confluence of the Pawcatuck River and Block Island Sound consists of material that is more stratified and permeable than the material in the eastern half of the moraine.

Well WEW584 is screened in sediments of predominately sand underlying sand and gravel from one of the depositional sequences (fig. 21, section *D-D′*). Section *E-E′* shows the interpreted lateral extent of the coarse sediments in which the public-supply well is screened. The sequence north of the public-supply-well site was interpreted to consist of mostly fine sediments on the basis of the lithologic log from WEW609. Another sequence containing coarse deposits may be present between WEW584 and Chapman Pond or coarse deposits at WEW584 may extend further north; however, the wetlands limited access to collect deep lithologic logs of the stratified sediments.

The Charlestown moraine, deposited across the southern part of the pre-glacial valley, forms a 100-ft ridge above the land surface of the wetland and forms the surface-water divide between the watersheds of the Pawcatuck River and Block Island Sound. According to Tiner (1989), the Charlestown

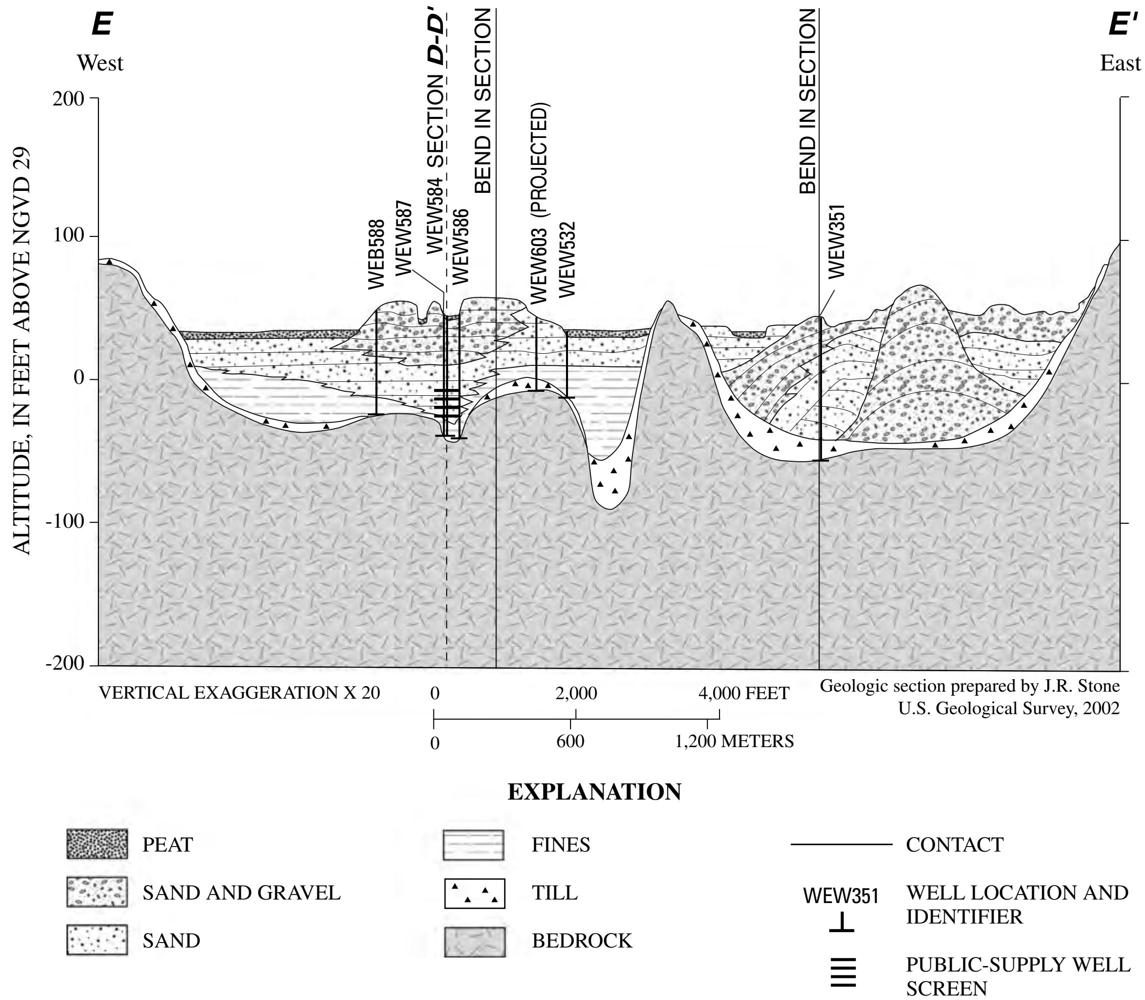


Figure 21—Continued. Geologic cross sections (lines of sections *D-D′* and *E-E′* are shown in figure 20), Westerly study area, Rhode Island.

moraine promoted wetland formation by blocking surface drainage from the north. A water-table map drawn by Gonthier and others (1974) on the basis of water-level measurements made in 1958 and 1966 show that the ground-water divide in the moraine generally coincides with the surface-water divide west and east of the pre-glacial valley axis. Measuring-point altitudes of the wells were estimated from USGS topographical map land-surface contours and are considered accurate within ±5 ft. The location of the ground-water divide near the axis of the pre-glacial valley was uncertain because of the contour interval and the relatively flat gradient shown on the water-table map. Two water-table altitudes in the axis of the pre-glacial valley, however, are above the altitude of the wetland; these altitudes indicate that the divide is in the moraine. Conceptually, the more transmissive the moraine, because of saturated thickness or hydraulic conductivity or both, the more water can move through the moraine to Block Island Sound; thus, the ground-water divide would be farther north in areas of the moraine with higher transmissivity.

In addition to precipitation recharge and runoff from till-covered bedrock, surface water in the wetland is a potential source of water to WEW584. Sources of this surface water in the wetland may be direct precipitation and surface-water and ground-water inputs. Under pumping conditions, water that may have flowed across the surface may be induced into the aquifer and withdrawn at the supply well. Although limited in areas of dense vegetation, observations of the wetland under average hydrologic conditions indicated that most of the eastern half of the main wetland is covered by water. Wetland areas near Chapman Pond, which is shallow (less than 4 ft deep; Guthrie and Stolgitis, 1977) and underlain by peat, are also probably covered with water during average conditions. The southwestern part of the main wetland near the supply well is characterized by separate pools of water and the water table is a few inches below the peat surface between the pockets of surface water under average hydrologic conditions.

An aquifer test at WEW584 was done on December 12–17, 1986, by D.L. Maher (1987) to determine the yield for the new supply well. Drawdowns were measured in WEW584 and in observation wells surrounding the well site but not in sediments below the wetland; three of these observation wells, which were from 10 to 30 ft from WEW584, were screened in the deep part of the aquifer at the same level as WEW584. These drawdown data were analyzed by means of an analytical method to provide quantitative information for numerical model simulation.

A semilogarithmic plot of drawdown against time in WEW584 and the three deep observation wells is shown in figure 22. Drawdowns generally plot as a straight line indicating a confined response during the 5-day aquifer test. According to field notes from the aquifer test, WEW584 was pumped continuously at a rate of 700 gal/min. The change in slope during the third day (3,500 to 4,200 minutes from start of pumping) of the test, however, is interpreted as a change in the pumping rate and not as a response to a boundary of the aquifer because the slope before and after the third day is the same. Drawdowns from a long-term aquifer test or from observation wells screened below the wetland deposits might eventually deviate from a straight line. Such a deviation would indicate vertical leakage from the wetlands.

The drawdown data were analyzed on the basis of a graphical technique by Jacob (1950). This technique includes several simplifying assumptions, such as a homogeneous, isotropic aquifer and horizontal flow to a well screened the full vertical thickness of the aquifer. A transmissivity of 8,200 ft^2/d was determined from the slope of the water-level responses through the late-time drawdown data (fig. 22). This transmissivity value corresponds to a horizontal hydraulic conductivity of about 120 ft/d for the 70 ft of saturated sediments at the well site. The transmissivity value determined from the aquifer test compares favorably to the transmissivity of 8,000 ft^2/d determined from the lithologic logs. The storage coefficient was also determined from the semilogarithmic plot based on Jacob's method (1950) for two of the three observation wells (the distance from the supply well, which is required to determine the storage, was accurately known only for two wells). Storage-coefficient values ranged from 0.0023 to 0.0075 for the two observation wells, which results in a specific storage for 70 ft of saturated sediments similar to that determined for Cape Cod sediments by Moench and others (2000).

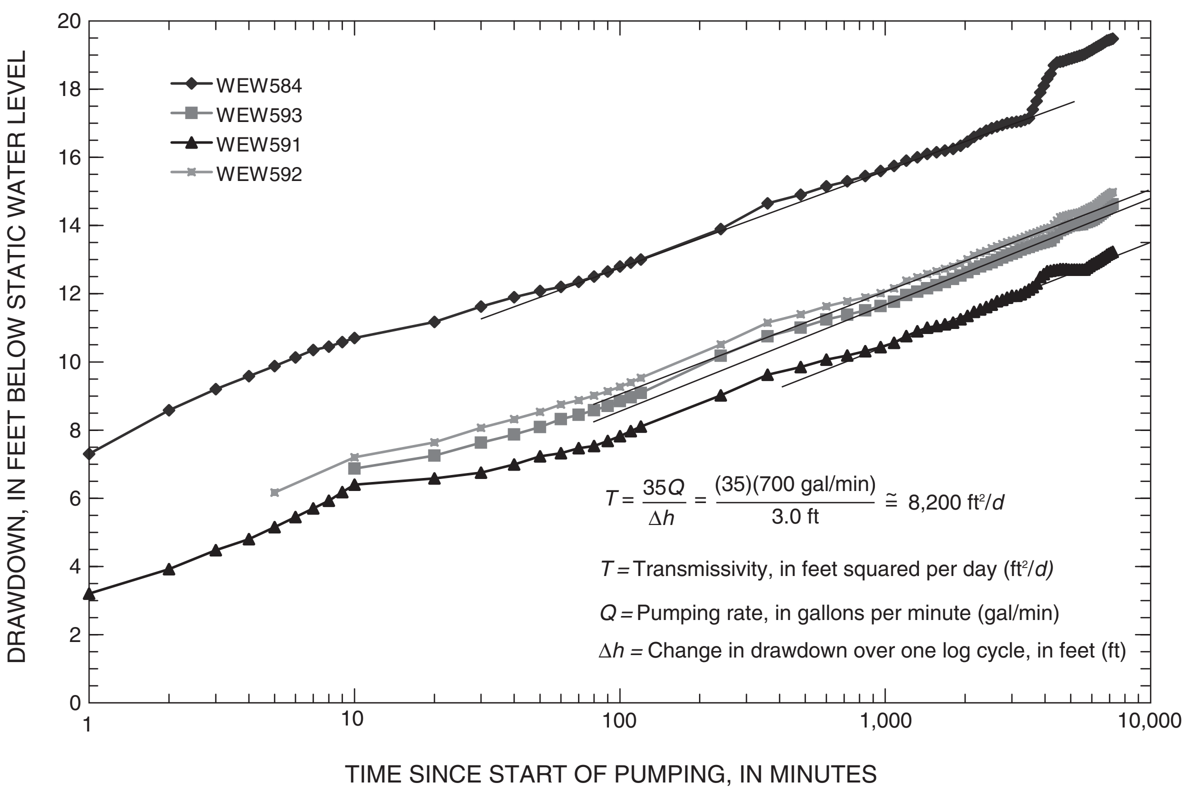


Figure 22. Water-level drawdowns in public-supply well WEW584 and in observation wells during the December 12–17, 1986, aquifer test, Westerly study area, Rhode Island.

During this study, water levels were measured in vertical nests of piezometers installed in the peat and underlying stratified sediments in the vicinity of the well site (fig. 23). At one of these piezometer nests, WEW604-606, water levels were recorded at 30-minute intervals and the surface-water stage was recorded intermittently from July 16 to August 30, 2001 (fig. 24). During this period, the depth of surface water overlying 7 ft of peat at this piezometer nest was about 1 ft and the water covered approximately one acre. Small-diameter piezometers were screened from 2.5 to 3.0 ft and from 5.5 to 6.0 ft deep in the peat layer and a third piezometer was screened from 10 to 13 ft deep in sand, 3 to 6 ft below the bottom of the peat layer. Ground-water levels in the sand responded to the cyclical nature of the supply-well operation, whereas the water levels in the peat did not show a response to these short-term changes in stress. A comparison of the surface-water stage and the ground-water levels, however, indicates a downward vertical gradient, which indicates the potential for surface water in the wetland and ground water in the peat to infiltrate the underlying aquifer.

The head difference between water levels in the peat and the underlying stratified sediments on December 21, 2001, at all the nests of piezometers is shown in figure 23. Peat thickness ranges from 3 to 7 ft at the piezometer nests. Water levels in the wetlands near the public-supply-well site on this date were generally a few inches above or below land surface. Water-level measurements were made when WEW584 was pumping at a rate of 700 gal/min during a pumping cycle of 1 day on and 2 days off. Water levels in the stratified sediments were lower than water levels in the peat because of pumping, or a combination of pumping and possibly, under natural conditions, areas of ground-water recharge. The lowered head in the stratified sediments has the potential to induce leakage from the wetlands.

The estimated value of the hydraulic conductivity of peat, along with other relevant hydraulic properties, was based on the degree to which the peat is decomposed. Boelter (1969) determined the hydraulic conductivity of peat as a function of fiber content and bulk density, two measures of the degree of decomposition. In general, the hydraulic conductivity of peat decreases with decreases in the fiber content and increases in bulk density, because the process of decomposition causes decreases in pore sizes. Boelter (1965, 1969) reported hydraulic-conductivity values for peat in Minnesota that varied over several orders of magnitude, from less than 0.1 to 100 ft/d, and with depth; in general, high values were for shallow, loose peat consisting of recent deposits with large interconnected pore spaces and low values were for highly decomposed compacted peat with small pore spaces at greater depths. Conceptually, most of the water in the wetland travels over the surface and through permeable shallow peat, whereas the highly decomposed peat at depth impedes vertical flow.

The average fiber content of 304 samples from 94 cores and the average bulk density of 20 samples from 10 cores collected by Hughes (1982) in the study area indicate that the peat is, on average, of medium decomposition. Boelter (1969) reported the relation between hydraulic conductivity and fiber content based on a minimum sieve size of 0.10 mm, whereas the fiber content for the study-area deposits was based on a minimum sieve size of 0.25 mm; thus, the less extensive bulk-density data set was used to estimate the average hydraulic conductivity. The average bulk density, 0.15 g/cm^3, corresponds to a hydraulic conductivity of 0.3 ft/d. This average value provides a starting point for assigning horizontal hydraulic conductivity, and most importantly, vertical hydraulic conductivity, which represents the connection between the wetland and underlying aquifer in the model.

Model Design

Ground-water flow in the study area was represented by a five-layered, variable-spaced grid model. In general, flow was simulated in the surficial deposits overlying the relatively impermeable till and bedrock. The lateral extent of the model in the Pawcatuck River and Block Island Sound Watersheds generally coincides with the geologic contact between surficial deposits and till-covered bedrock (fig. 20). Some of the till and bedrock in the Block Island Sound Watershed also is included in the model in order to simulate the ground-water divide between the two watersheds. The modeled area, 16.9 mi^2, consists of 242 rows and 200 columns with 112,963 active cells. Cell sizes range from a minimum of 50 ft on a side in the vicinity of the public-supply well to a maximum of 400 ft on a side near Block Island Sound.

Vertical discretization was based on land-surface altitude in the wetland, lithology, and well-screen placement (fig. 25). Overland flow in the wetland is simulated in layer one; the bottom of this layer coincides with the land surface in wetland areas. A uniform 6-ft-thick layer of peat is represented in layer 2. Stratified deposits, the moraine, and till-covered bedrock in Block Island Sound Watershed are represented in all five layers; WEW584 is screened in layer 5. Three model layers were simulated below the peat in the vicinity of WEW584 originally to calibrate the hydraulic connection between the peat and aquifer from vertical head data. This vertical head data, however, were taken after a short period of pumping, about 18 hours at WEW584, and proved to be too sensitive to storage to determine accurately the vertical hydraulic conductivity of the peat.

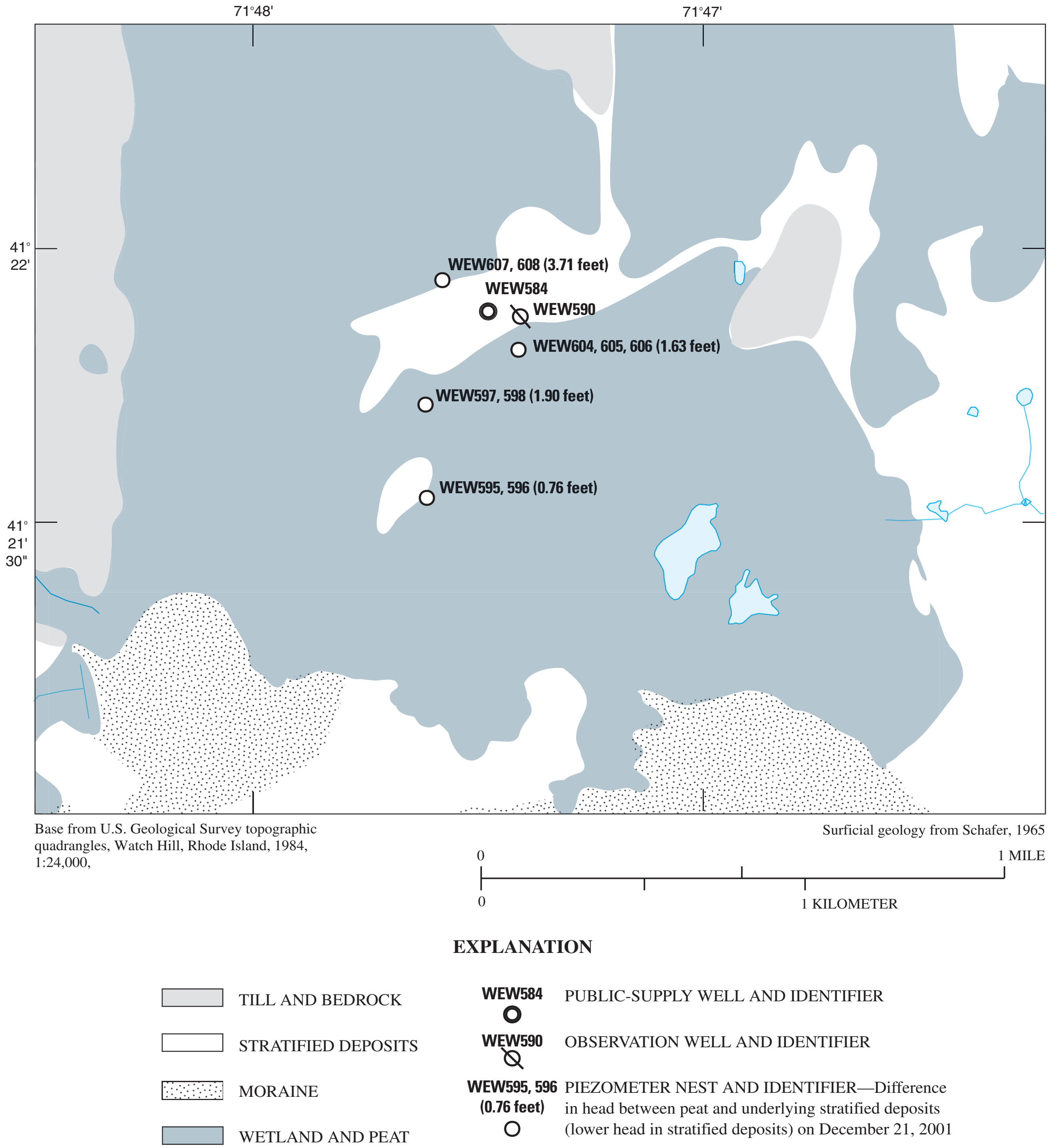
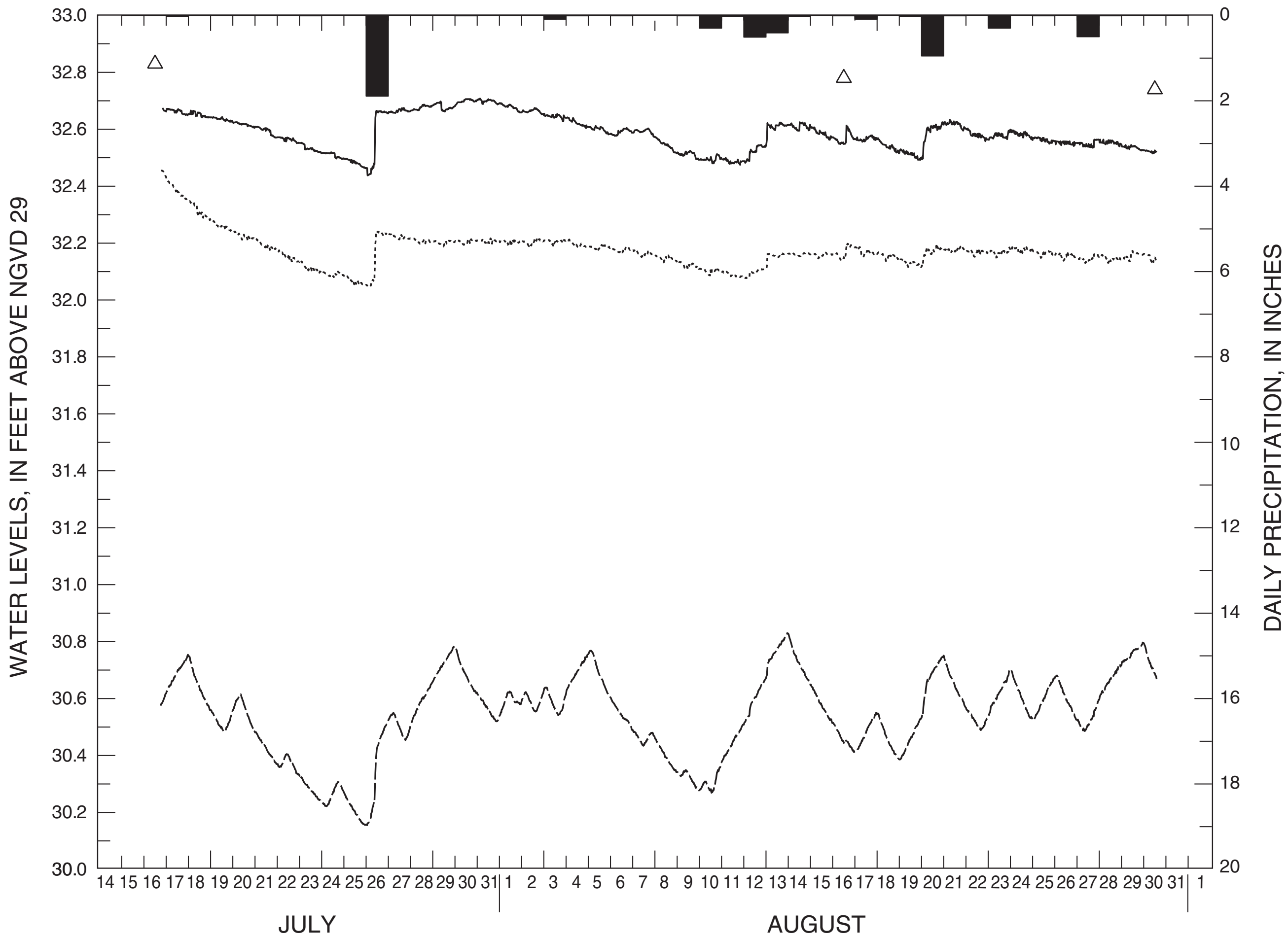


Figure 23. Vertical head difference between peat and stratified glacial deposits at four piezometer nests on December 21, 2001, and observation well WEW590 used in the transient simulation, Westerly study area, Rhode Island.



EXPLANATION

Piezometer	Altitude of screen interval (feet above NGVD 29)	Depth of screen interval below top of peat (feet)
WEW604	29.58–29.08	2.5–3.0, in peat
WEW605	26.58–26.08	5.5–6.0, in peat
WEW606	22.15–19.15	10–13, in sand, or 3–6 below peat layer

DAILY PRECIPITATION

△ WETLAND STAGE

Figure 24. Stage of wetland and water levels in piezometers screened in peat and sand, and daily precipitation amounts, Westerly study area, Rhode Island.

Several types of boundary conditions were specified in the model to represent areas of discharge and sources of recharge (fig. 26). The Pawcatuck River, Chapman Pond, and numerous streams were simulated as head-dependent flux boundaries by using the MODFLOW river, drain (Harbaugh and McDonald, 1996) or stream-routing (Prudic, 1989) packages. Streams draining into and through the wetlands in the eastern part of the study area were simulated with the stream-routing package because some of this surface water may recharge the aquifer, particularly in the vicinity of the supply well. Because the stream-routing package accounts for water in each stream cell, induced infiltration of surface water would cease if simulated surface flow also ceases. Block Island Sound, its confluence with the Pawcatuck River, and two ponds in the moraine were simulated as constant heads.

Surface-water altitudes for the head-dependent flux boundaries were estimated from USGS topographical maps except for Chapman Pond's altitude, which was measured for this study. For most cases, water depths and bed thicknesses of 1 ft were assumed in order to determine top and bottom bed

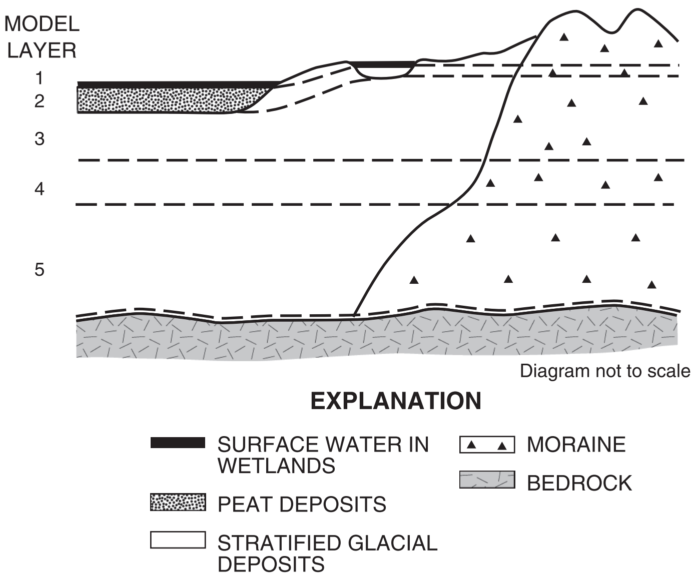


Figure 25. Schematic section showing model layers, Westerly study area, Rhode Island.

elevations from surface-water altitudes; an 8-ft water depth was used for the Pawcatuck River on the basis of a cross-section through the river within the study area by Gonthier and others (1974). The Pawcatuck River was simulated as 100 ft wide and the narrow streams overlying glacial deposits were simulated as 10 ft wide. The vertical hydraulic conductivity of streambeds overlying glacial deposits was assigned a value of 2 ft/d on the basis of the average value for coarse-grained sediments determined by Gonthier and others (1974) from field measurements made with a variable-head permeameter in the Pawcatuck River Watershed. Head-dependent flux boundaries within wetland areas were assigned bed thicknesses, widths, and vertical-hydraulic conductivity values that resulted in conductances about 50 times greater than the conductance value for streambeds overlying glacial deposits. High conductance values allowed surface water to move relatively unimpeded between the head-dependent flux boundary and the wetland simulated in model layer 1; thus, it is the peat layer that controls the degree of connection between the wetland and the aquifer.

A recharge rate of 26 in/yr from direct precipitation was applied to the top of the model except in wetland areas. This recharge rate, which is equivalent to precipitation minus evapotranspiration, was based on mean annual runoff data from 1951 to 1980 (Randall, 1996). This recharge rate is consistent with 60 years of runoff data (1940–2001) from the USGS streamflow-gaging station Pawcatuck River at Westerly, RI (01118500) (figs. 1 and 26). Because evapotranspiration rates in the study area wetlands are unknown, an average evapotranspiration rate equivalent to the evaporation rate from a shallow, open water surface was assumed to represent the wetlands adequately. A specified flux of 21 in/yr was applied to the wetland by subtracting the evaporation rate from a free-water surface (29 in/yr) (Farnsworth and others, 1982) from the rate of precipitation for southern Rhode Island (50 in/yr). Some of this water may infiltrate or move across the wetland as overland flow. Recharge from ground-water and surface-water runoff originating from till-covered bedrock uplands not simulated directly in the model was applied to cells at the edge of the model. The quantity of recharge applied to each model cell was based on the mean annual runoff, 26 in/yr, and the size of the till-covered bedrock upland area draining toward each cell.

The geologic sections provided a general framework for distributing hydraulic conductivity throughout the model. Because the thickness and extent of coarse-grained stratified sediments in the well-site area are well-defined from numerous lithologic logs, two values were used: sand and gravel sediments were assigned a horizontal hydraulic conductivity of 200 ft/d, and sand, consisting primarily of well-sorted fine sand and medium sand layers, was assigned a value of 80 ft/d. Outside of the well-site area, however, sand sediments and sand and gravel, were assigned one value of 150 ft/d, which represents an intermediate value between medium sand and sand and gravel. Because the extent of fine-grained sediments beneath the wetland is not well known, a value of 50 ft/d, representative of fine sand, was used. The moraine was represented by one value, 50 ft/d, also equivalent to the horizontal hydraulic conductivity of fine sand. Till-covered bedrock was represented as one unit and assigned a value of 0.5 ft/d.

Peat was assigned a horizontal hydraulic conductivity of 1 ft/d, slightly greater than the average value determined for the peat based on its degree of decomposition, because most water moving through the peat probably does so through the top portion. Vertical hydraulic conductivity of the peat, the model-input parameter controlling the connection between the wetland and aquifer, was assigned a value slightly less than the average value to represent the highly decomposed peat and organic-rich silts at depth. Field measurements of vertical hydraulic conductivity made with the variable-head permeameter by Gonthier and others (1974) in silty highly decomposed peat underlying slow moving backwater reaches of the Pawcatuck River ranged from 0.1 to 0.7 ft/d. The low end of this range, 0.1 ft/d, was specified in the model to represent a conservative value that would allow less leakage from the wetland and result in a larger area contributing recharge to the well. Surface water in the wetlands was simulated in layer 1 with a high horizontal and vertical hydraulic-conductivity value, initially 100,000 ft/d. The high hydraulic conductivity is intended to simulate the relatively flat gradient that was measured across the wetland. Conceptually, the high hydraulic conductivity value represents minimal resistance to overland flow, and in areas of the wetland where the water level is below land surface, to flow through any large, interconnected pores in poorly decomposed peat in the top few inches of the peat layer.

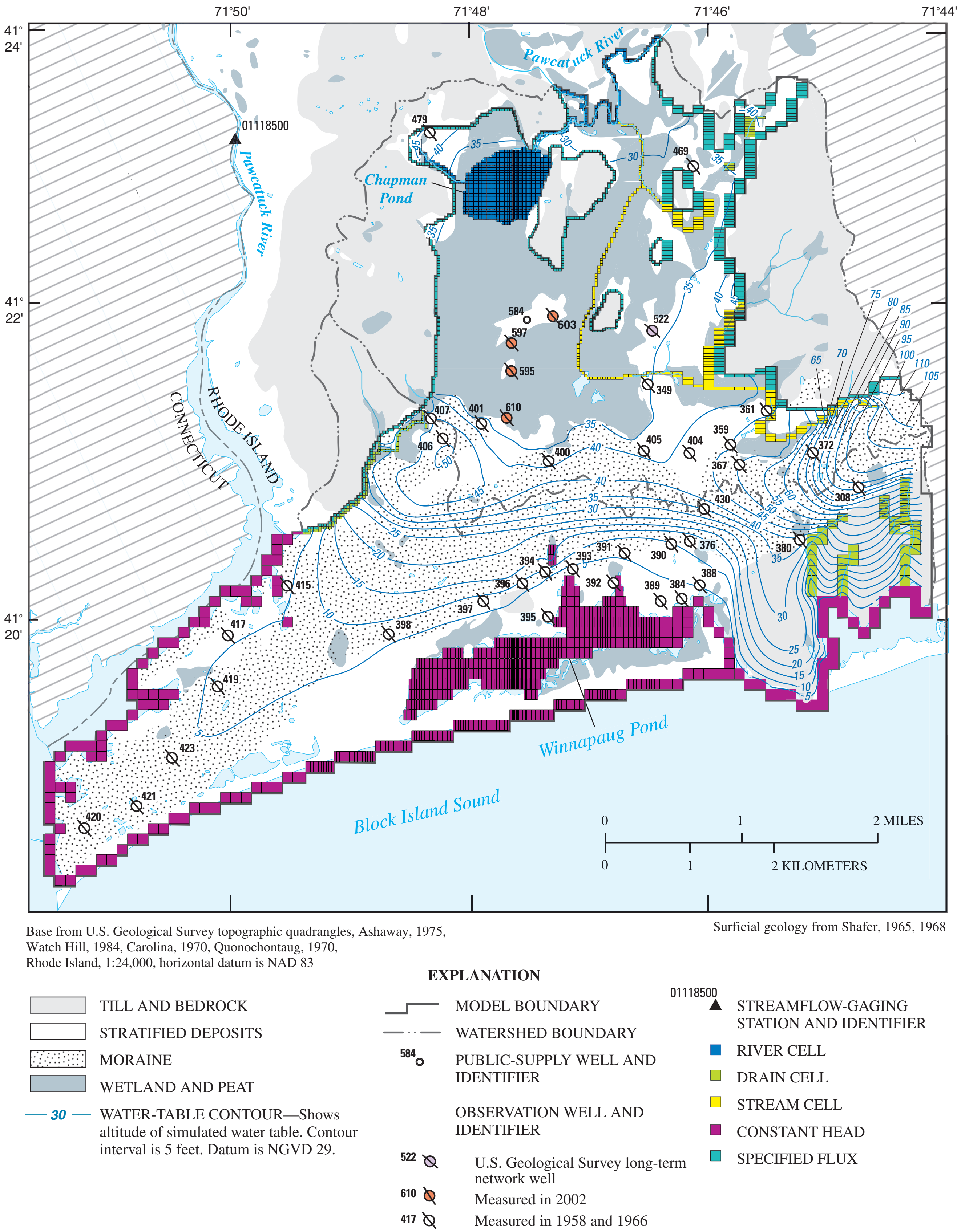


Figure 26. Model-boundary types, streamflow-gaging station, observation wells, and simulated water-table contours for steady-state, nonpumping conditions, Westerly study area, Rhode Island.

Model Calibration

Calibration simulations were done for both steady-state and transient ground-water-flow conditions. Steady-state, nonpumping conditions were based on water-level measurements made for this study in 2002 and water levels measured by the USGS in 1958 and 1966 (Gonthier and others, 1974). The transient calibration was based on the 5-day aquifer test conducted in 1986 (D.L. Maher Co., 1987). The overall purpose of the model calibration was to simulate the relatively flat gradient across the wetland and to simulate water levels and flow directions that were comparable to measured water levels and to the conceptual understanding of the study area hydrology. Horizontal hydraulic conductivity and vertical hydraulic conductivity of bed sediments were adjusted during model calibration. Recharge fluxes and vertical hydraulic conductivity of the peat were held constant; field data were insufficient to calibrate the vertical hydraulic conductivity of the peat in model simulation.

One surface-water-level and four ground-water-level measurements made on June 2, 2002, approximate long-term average annual conditions based on long-term records from WEW522 and from streamflow-gaging station Pawcatuck River at Westerly, Rhode Island (01118500) (fig. 27). WEW522 is in the study area 0.9 mi east of the supply well and the gaging station is just west of the study area at a point which drains most of the Pawcatuck River Watershed (figs. 1 and 26). The ground-water levels, which are from wells in low-lying areas adjacent to the wetland (fig. 26), were measured when WEW584 had been operated for one of the previous 12 days and may have been slightly affected by water withdrawals. The surface-water level used was the altitude of Chapman Pond in the model.

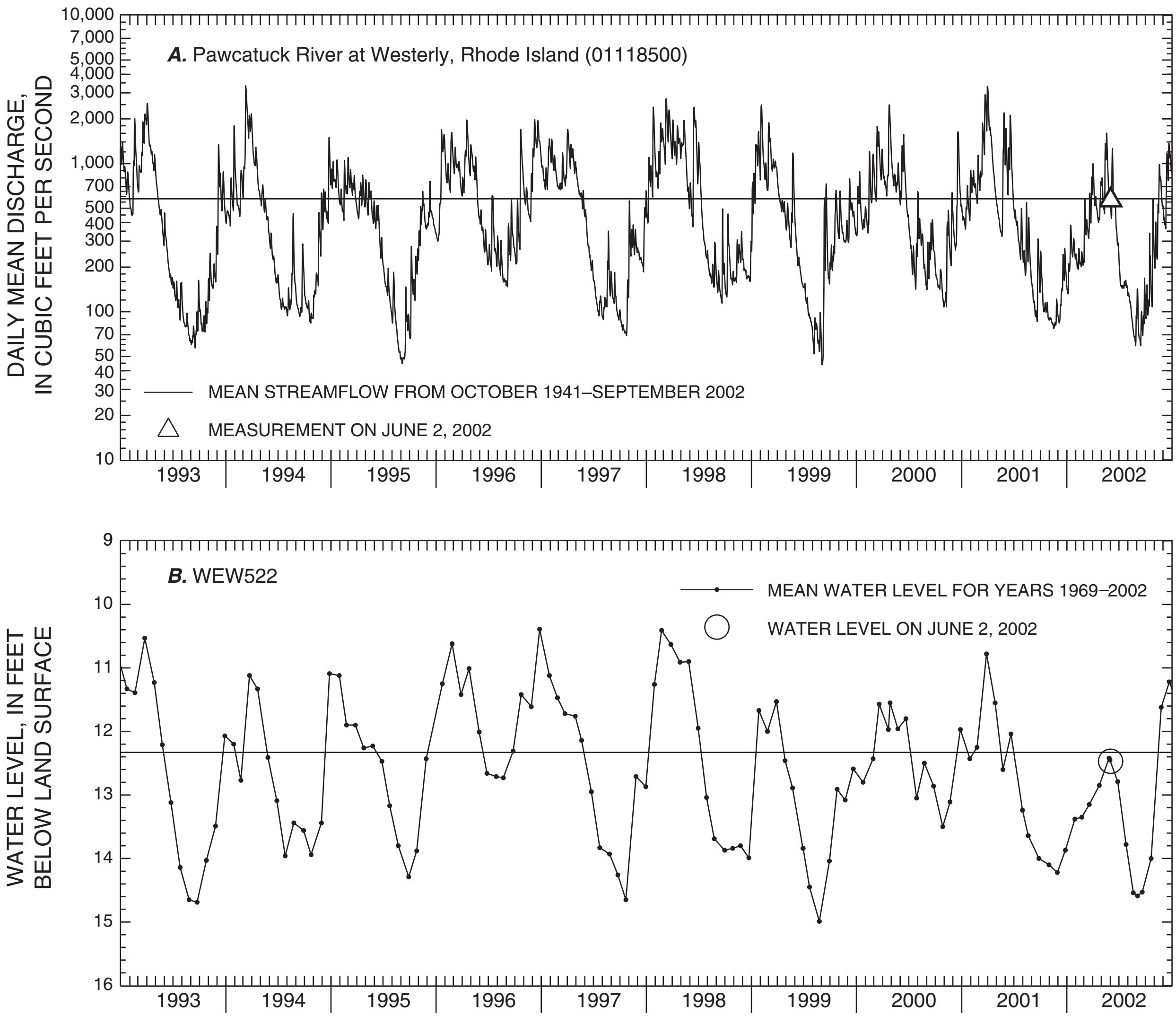


Figure 27. Streamflow for *A*, the Pawcatuck River at Westerly, Rhode Island; and water levels at *B*, well WEW522, Westerly study area, Rhode Island, 1993–2002.

Water levels measured in August 1958 and August 1966 from 35 mostly domestic dug wells were used to aid model calibration. Measuring-point altitudes of these domestic wells, which includes the long-term network well WEW522, are considered accurate within approximately ±5 ft. In contrast to the data set from 2002, water levels from 1958 and 1966 represent below-average natural conditions. The sites for most of these water-level measurements, however, are in the moraine (fig. 26) and provide a means of qualitatively calibrating water levels in the south part of the model.

Initial model-input values resulted in simulated gradients across the large wetland that were slightly too steep and water levels in the eastern half of the moraine that were too low; however, simulated flow was consistent with the conceptual understanding of flow patterns in the study area. The calibration process involved a series of steady-state simulations in which wetland and moraine horizontal hydraulic conductivities were adjusted to improve the match between simulated and measured water levels. Large changes in the wetland hydraulic conductivity resulted in only small changes in water levels; the final value for the wetland, 600,000 ft/d, lowered water levels in the well-site area about 0.4 ft and in the southern extent of the large wetland about 0.6 ft/d. Wetlands of small extent, most of which drain into the large wetland, were simulated with final hydraulic-conductivity values ranging from 5,000 to 50,000 ft/d. Horizontal hydraulic conductivity in the eastern half of the moraine was lowered from 50 ft/d to 10 ft/d. Hydraulic-conductivity values for till-covered bedrock were raised from 0.5 ft/d to 3 ft/d to lower water levels below land surface.

The altitude and configuration of the simulated water table for steady-state, nonpumping conditions are shown in figure 26 and a comparison of simulated to measured ground-water altitudes is listed for the observation wells in table 6. Residuals for water levels measured in June 2002 ranged from -0.45 to -0.10 ft and residuals for the remaining observation (dug) wells ranged from -11 to 9 ft. Six of the seven dug wells with the largest residuals (greater than 5 ft) are 2 mi from the supply well; all seven are located in or south of the moraine. In addition, the ground-water altitudes have an accuracy of ±5 ft. Because the large residuals are not expected to affect the delineation of an area contributing recharge to the supply well substantially, further calibration through adjustment of model parameters was not undertaken.

The total inflow rate for the calibrated, steady-state model, 3,307,000 ft^3/d, includes 2,652,000 ft^3/d or 80 percent from precipitation and 655,000 ft^3/d or 20 percent from runoff from till-covered bedrock. Of this total inflow, 1,527,000 ft^3/d or 46 percent eventually discharges to the Pawcatuck River along the northern boundary, and the remaining outflow, 1,780,000 ft^3/d or 54 percent, eventually discharges to Block Island Sound and the mouth of the Pawcatuck River.

The steady-state calibrated model was then used in a transient simulation of the 5-day aquifer test at WEW584. Drawdowns in observation wells near the supply

Table 6. Ground-water altitudes measured in 1958, 1966, and 2002, and simulated ground-water altitudes for the steady-state model in the Westerly study area, Rhode Island.

[ft, foot]

Observation well	Model layer	Measured (ft above NGVD 29)	Simulated (ft above NGVD 29)	Difference (measured minus simulated, in ft)
		2002		
WEW595	5	32.73	32.87	-0.14
WEW597	5	32.64	32.74	-.10
WEW603	3	32.52	32.68	-.14
WEW610	3	33.17	33.62	-.45
		1958,1966, 2002		
WEW308	1	92	89.71	2
WEW349	1	34	34.78	-1
WEW359	1	48	44.63	3
WEW361	1	48	43.39	5
WEW367	1	54	50.37	4
WEW372	1	78	70.73	7
WEW376	4	4	12.34	-8
WEW380	2	33	41.99	-9
WEW384	4	1	2.33	-1
WEW388	4	3	5.65	-3
WEW389	4	3	1.95	1
WEW390	4	9	10.06	-1
WEW391	4	13	5.91	7
WEW392	4	1	0.56	0
WEW393	4	4	4.29	0
WEW394	4	10	9.03	1
WEW395	4	1	0.75	0
WEW396	4	13	8.58	4
WEW397	4	13	7.87	5
WEW398	4	5	6.87	-2
WEW400	1	48	39.36	9
WEW401	1	37	37.19	0
WEW404	1	41	41.01	0
WEW405	1	35	37.64	-3
WEW406	1	53	48.27	5
WEW407	1	45	44.59	0
WEW415	4	1	3.63	-3
WEW417	4	1	3.97	-3
WEW419	4	1	5.75	-5
WEW420	4	1	1.70	-1
WEW421	4	4	2.66	1
WEW423	4	6	4.26	2
WEW430	3	24	34.94	-11
WEW469	1	35	33.25	2
WEW479	1	43	43.31	0
WEW522	1	34	33.52	0

well and screened in the deep part of the aquifer could not be simulated accurately because of the grid size in the numerical model in the vicinity of the supply well. Drawdowns in observation well WEW590, 400 ft southeast of the supply well (fig. 23) and screened in shallow sand and gravel sediments, were compared to simulated drawdowns in the transient simulation. Boundary conditions were assumed to be the same as in the steady-state calibrated model except for direct recharge from precipitation, because no precipitation occurred during the aquifer test. Storage properties for glacial deposits and bedrock were added tothe transient model; storage properties of the stratified deposits were assumed to represent the moraine and peat in the model adequately. The specific yield of the wetland was set to 1. Head data from the steady-state simulation were used as heads at the start of the aquifer test. The model accurately simulated the drawdowns at the observation well (fig. 28). This agreement indicates that model-input values for the calibrated steady-state model provide reasonable results under stressed conditions caused by the supply well; however, this analysis represents only a small volume of the aquifer near the supply well. Ideally, vertical gradients caused by a long-term aquifer test could be calculated for a nest of piezometers screened at multiple depths in the peat and the underlying aquifer and simulated with the model to calibrate the vertical hydraulic conductivity of the peat.

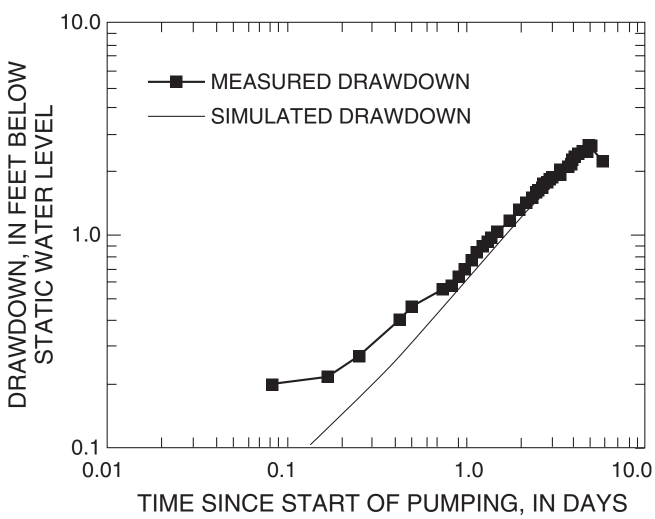


Figure 28. Measured and simulated drawdowns in observation well WEW590 during WEW584 aquifer test, Westerly study area, Rhode Island.

Delineation of Areas Contributing Recharge and Sensitivity Analysis

Average recharge conditions and the average and maximum pumping rates at WEW584 were used as stresses in the calibrated steady-state model to determine areas contributing recharge to the supply well. In the wetlands, MODPATH particles were placed on the bottom faces of cells instead of at the water surface. Particles placed at the water surface may not infiltrate through the bottom face if a portion of water from the wetland cell is withdrawn at the well because of the high horizontal hydraulic conductivity used to simulate the wetland, thereby underestimating the area contributing water to the supply well.

The area representing the starting locations of particles that are withdrawn from WEW584 for the average pumping rate of 240 gal/min and the maximum rate of 700 gal/min was classified into three categories (figs. 29 and 30, and table 7) from which particles flow (1) directly to the well from uplands, (2) indirectly from uplands through the wetlands, and (3) from the wetlands. (The term “uplands” in the Westerly study area refers to surficial material not covered by wetlands.) The first category includes starting points for particles that track from the water table in the uplands to the well without entering the wetland; these particles represent precipitation recharge and runoff from till-covered bedrock. Particles in the second category originate in the uplands but discharge to the wetland first along shallow ground-water-flow paths before they are withdrawn from the well. Particles in the third category infiltrate and discharge to the well from the wetlands.

The total area representing the starting location of particles withdrawn from WEW584 for the average pumping rate is about 0.67 mi^2 (fig. 29). The area extends southeastward to the ground-water divide in the moraine and includes isolated areas in the southeast and a ground-water mound and divide southwest of the supply well. Of this total area, 0.13 mi^2 is directly from uplands and contributes 46 percent of the total pumpage. The remaining water withdrawn from the well, 54 percent, is derived from the wetlands or indirectly from the uplands through the wetland. At the withdrawal rate of 700 gal/min, the total area covers about 1.43 mi^2. The expanded area includes more of the wetland, moraine, and stratified deposits, but also includes runoff from 0.008 mi^2 of till-covered bedrock. Of the total area, 0.16 mi^2 is directly from uplands and represents 21 percent of the total water pumped at the well, whereas the remaining water, 79 percent, is derived from the wetlands or indirectly from the uplands through the wetland. Thus for both the average and maximum pumping rates, the primary source of water withdrawn at WEW584 originates in the wetland or discharges to the wetland from the surrounding uplands before infiltrating.

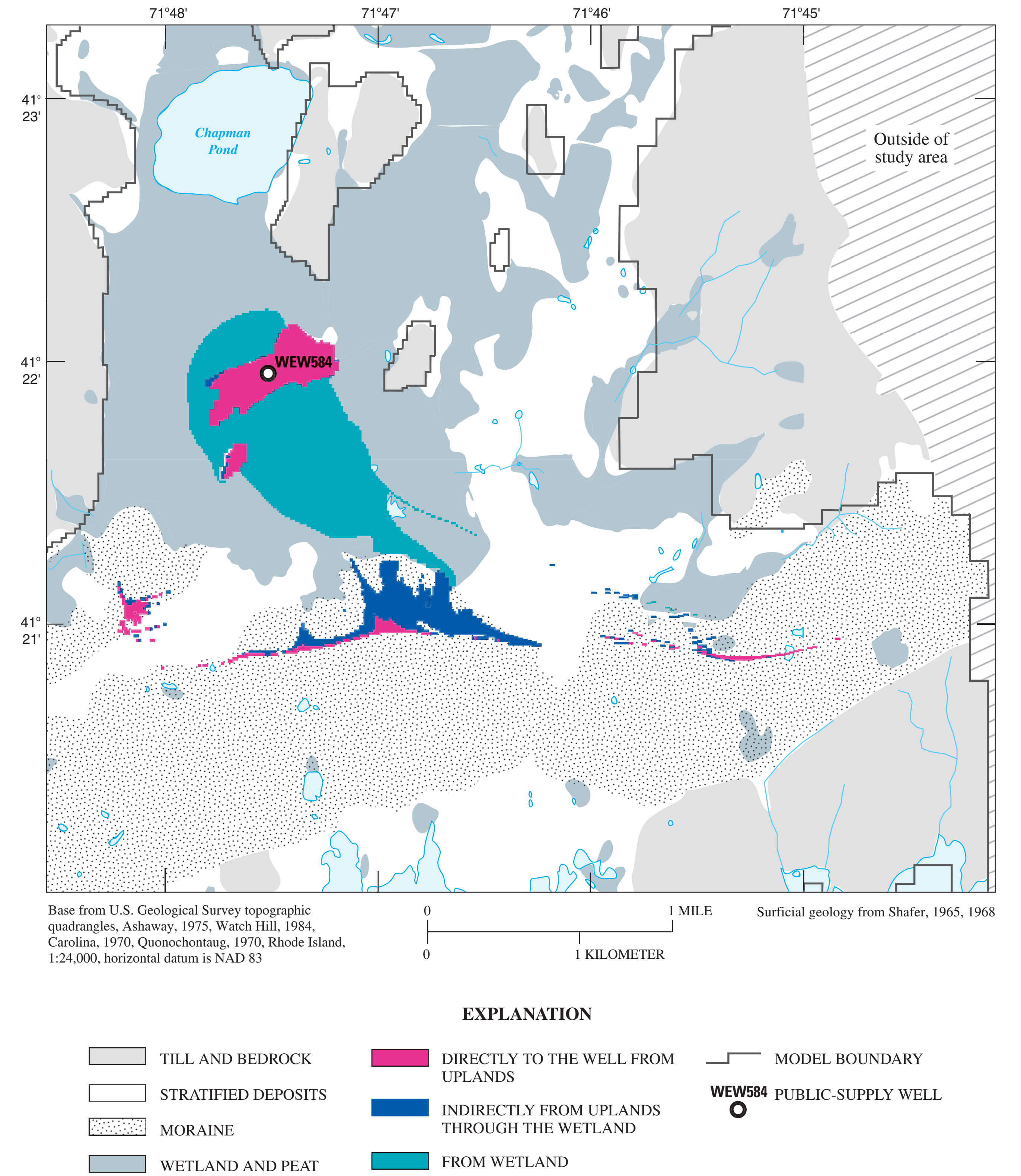


Figure 29. Simulated area representing starting locations of particles withdrawn from WEW584 at its average pumping rate of 240 gallons per minute directly to the well from uplands, indirectly from uplands through the wetland, and from the wetland, Westerly study area, Rhode Island.

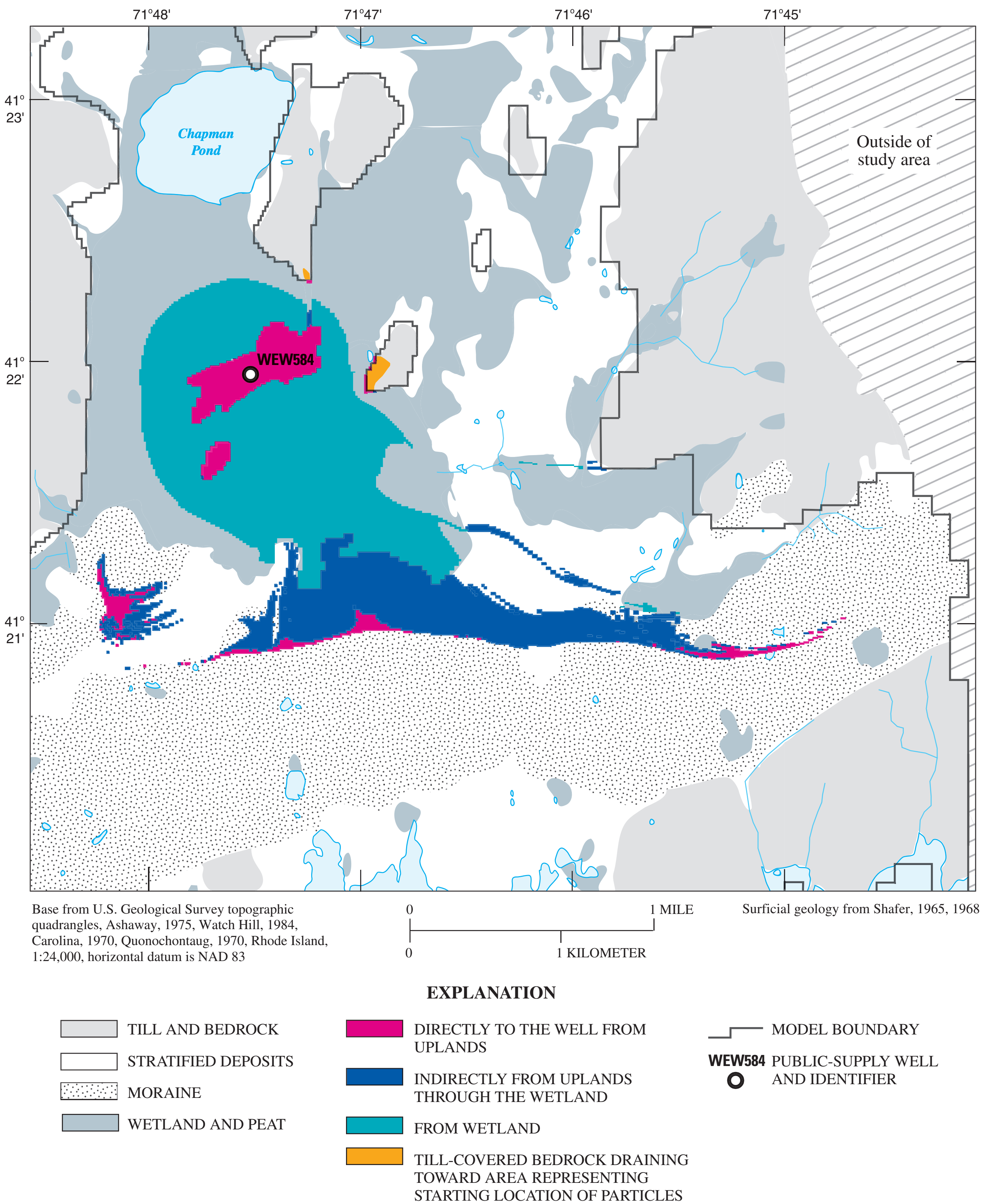


Figure 30. Simulated area representing starting locations of particles withdrawn from WEW584 at its maximum pumping rate of 700 gallons per minute directly to the well from uplands, indirectly from uplands through the wetland, from the wetland, and from till-covered bedrock draining toward the area representing the starting locations of particles, Westerly study area, Rhode Island.

Table 7. Sizes of areas contributing water to public-supply well WEW584 and till-covered bedrock areas draining toward the areas contributing water, and the percentage of the total water withdrawn from different sources of water, Westerly study area, Rhode Island.

[Uplands defined in the Westerly study area as surficial material not covered by wetlands. gal/min, gallons per minute; mi^2, square mile]

Model scenario	Area contributing water to well in modeled area (mi^2)	Area directly to well from uplands (mi^2)	Area indirectly to well from uplands through the wetlands (mi^2)	Wetland area (mi^2)	Till-covered bedrock draining toward area contributing water to well (mi^2)	Upland sources—directly to well from upland (percent)	Wetland sources—wetland and indirectly from upland through wetlands (percent)
240 gal/min	0.67	0.13	0.11	0.43	0	46	54
700 gal/min	1.43	.16	.40	.87	.008	21	79

A sensitivity analysis of the effects of selected hydraulic properties and of recharge on the delineated area representing the starting locations of all particles was done for WEW584 at the maximum pumping rate. The model was not recalibrated and the results are not more plausible than results from the calibrated model; however, the analysis is intended to evaluate the effects of uncertainties in selected model-input parameters on the area contributing water to a well in a wetland setting. The results of the sensitivity analysis are illustrated in figures 31 and 32.

The degree of hydraulic connection between wetlands and underlying sediments is not well understood. For example, root cavities in the peat in this densely vegetated wetland may provide preferential pathways for water to infiltrate. In addition, an effective thickness representative of low-permeability peat is unknown. The hydraulic connection between the wetland and the underlying aquifer was decreased by reducing the vertical hydraulic conductivity of the peat by half to 0.05 ft/d. The area contributing water to the well expanded minimally to 1.50 mi^2, most likely because the degree of connection used in the calibrated model was already at a sufficiently low value to limit leakage. The vertical hydraulic conductivity of the peat also was increased by an order of magnitude to 1 ft/d; the vertical hydraulic conductivity of 1 ft/d for the simulated 6 ft of peat is equivalent to a vertical hydraulic conductivity of 0.1 ft/d for 0.6 ft of low-permeability peat. The increased hydraulic connection caused the area in the wetland to decrease substantially and the area in the upland to decrease slightly; the total area was reduced to 1.13 mi^2.

The recharge rate and the specified flux applied to the wetland were reduced by 6 in/yr; runoff from till-covered bedrock was not changed. The area contributing water to the well in the wetland showed only slight changes because the vertical hydraulic conductivity of the peat is the limiting factor. The reduced recharge rate over the stratified deposits and the moraine, however, increased the area in these deposits and shifted the southern extent of the area slightly northward because of the new location of the simulated divide. The total area contributing water to the well increased to 1.69 mi^2.

Determination of the exact location of the ground-water divide between the Pawcatuck River Basin and Block Island Sound during average hydrologic conditions would require an extensive water-level network. Because the delineated area contributing recharge extends to the ground-water divide, the effects of the simulated location of the divide on the area contributing water to WEW584 were determined by adjusting horizontal hydraulic-conductivity values within reasonable ranges. The hydraulic conductivity in the eastern half of the moraine was decreased by one-half from 10 ft/d to 5 ft/d and, in the stratified deposits north of the moraine, increased by a factor of two. These changes in hydraulic conductivity shifted the simulated ground-water divide slightly southward, which in turn shifted the southern extent of the area contributing water slightly southward; however, the size of the area contributing water to the well, 1.42 mi^2, was nearly the same as the delineated area. The simulated ground-water divide was then shifted northward by using a horizontal hydraulic conductivity of 50 ft/d for the entire moraine and by decreasing hydraulic conductivity of the stratified deposits north of the moraine by one-half. The southern extent of the area contributing water to the well moved farther north but the area also expanded southeast of the well. The size of the area contributing water to the well, 1.43 mi^2, however, was equivalent to the delineated area.

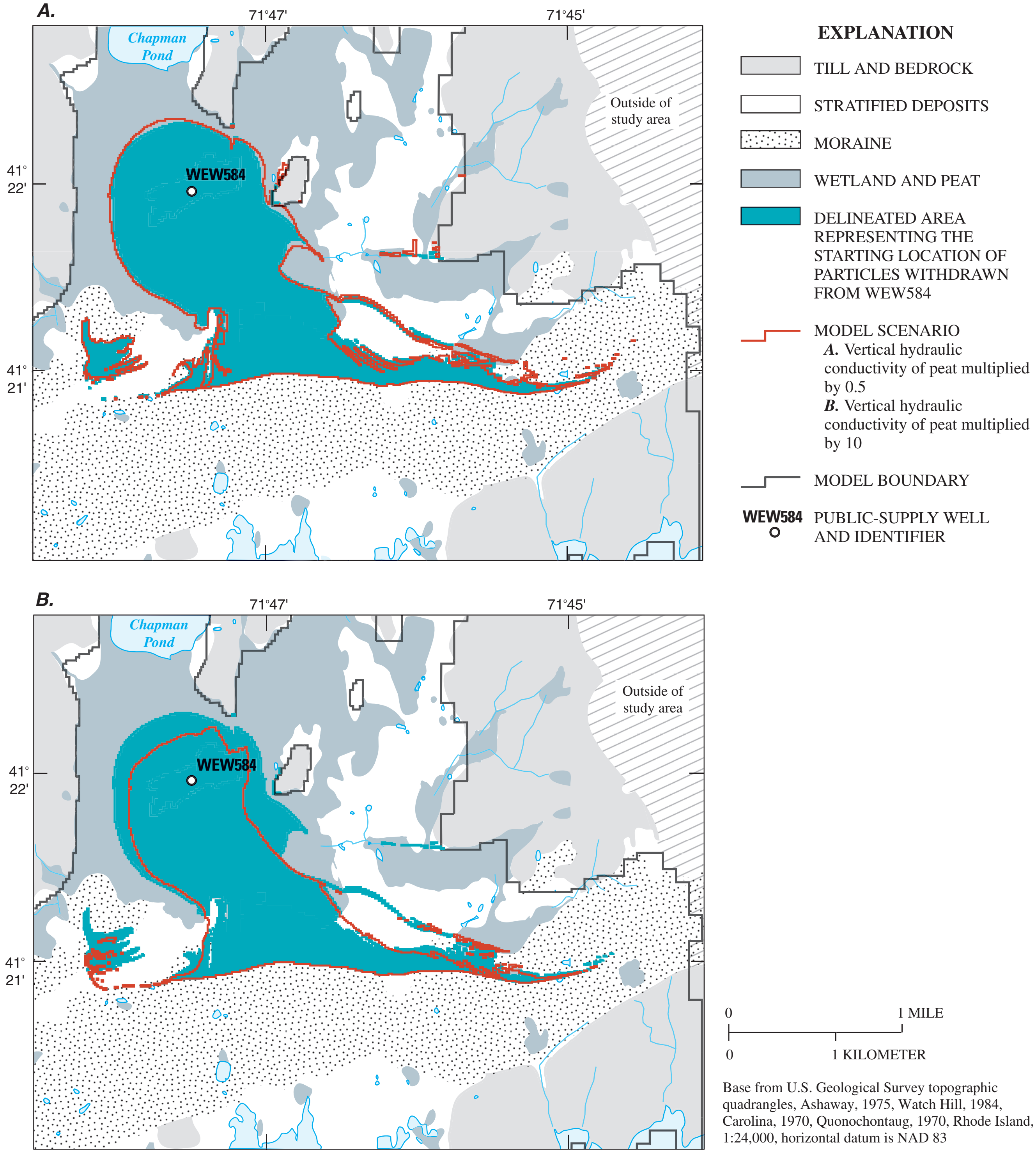
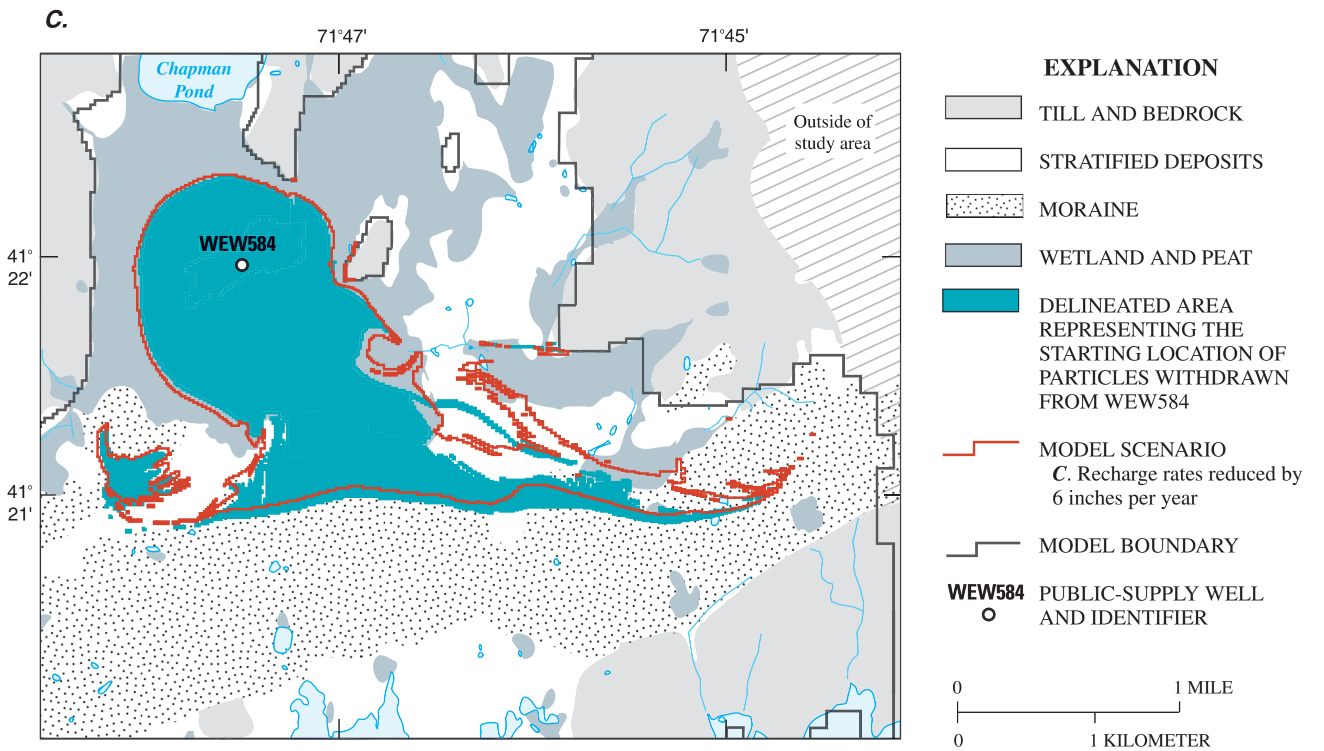


Figure 31. Sensitivity analysis of the effects of vertical hydraulic conductivity of peat *A*, multiplied by 0.5; *B*, multiplied by 10; and *C*, recharge rates reduced by 6 inches per year, on the delineated area representing the starting locations of all particles withdrawn from WEW584 at its maximum pumping rate of 700 gallons per minute, Westerly study area, Rhode Island.



Base from U.S. Geological Survey topographic quadrangles, Ashaway, 1975, Watch Hill, 1984, Carolina, 1970, Quonochontaug, 1970, Rhode Island, 1:24,000, horizontal datum is NAD 83

Figure 31—Continued. Sensitivity analysis of the effects of vertical hydraulic conductivity of peat *A*, multiplied by 0.5; *B*, multiplied by 10; and *C*, recharge rates reduced by 6 inches per year, on the delineated area representing the starting locations of all particles withdrawn from WEW584 at its maximum pumping rate of 700 gallons per minute, Westerly study area, Rhode Island.

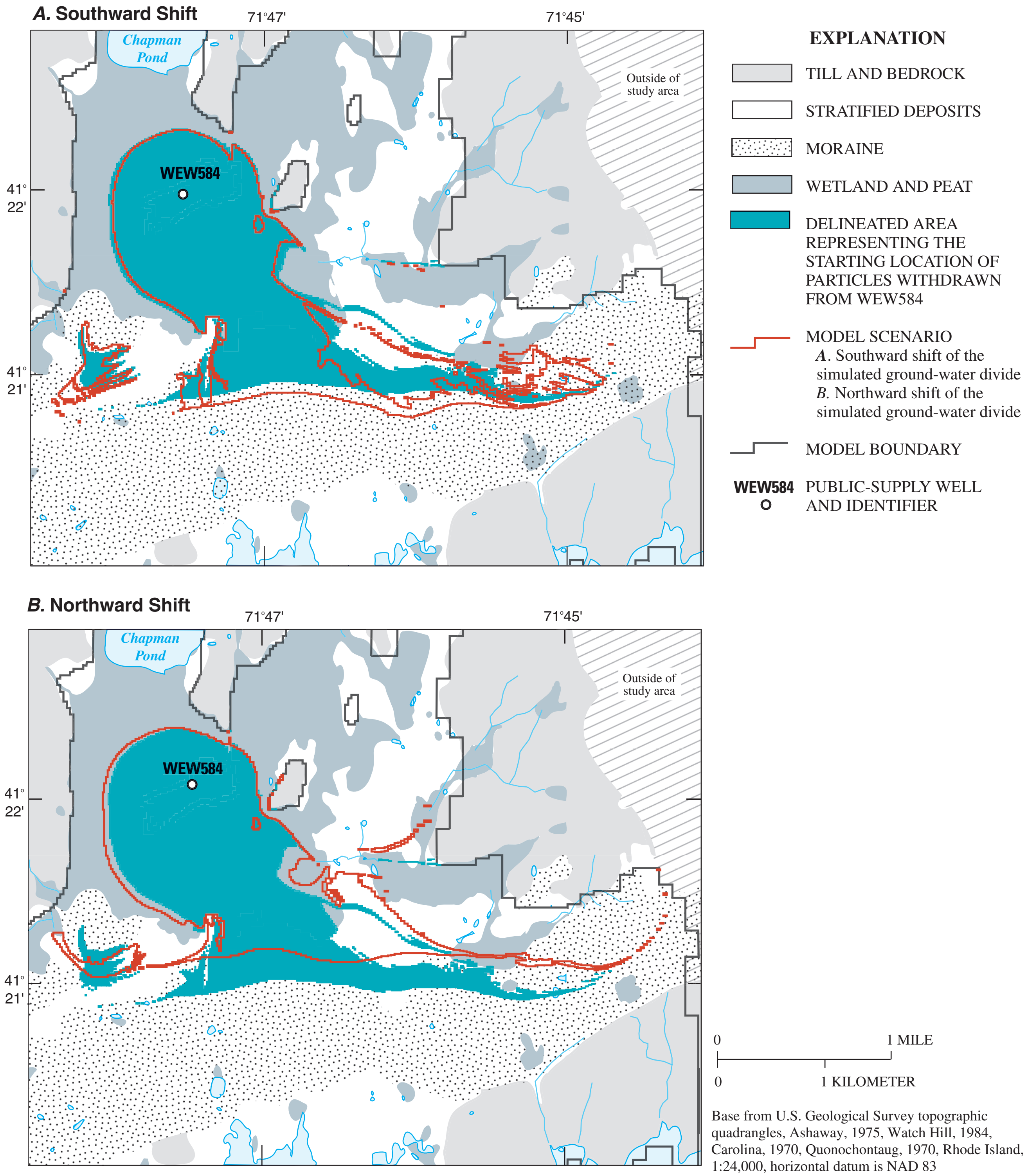


Figure 32. Sensitivity analysis of the effects of *A*, a southward shift; and *B*, a northward shift of the simulated ground-water divide between the watersheds of the Pawcatuck River and Block Island Sound on the delineated area representing the starting locations of all particles withdrawn from WEW584 at its maximum pumping rate of 700 gallons per minute, Westerly study area, Rhode Island.

Summary and Conclusions

Areas contributing recharge and sources of water to seven present public-supply wells and one proposed public-supply well clustered in three study areas in four Rhode Island towns were determined based on calibrated, steady-state numerical ground-water-flow models representing long-term average hydrologic conditions. The U.S. Geological Survey, in cooperation with the Rhode Island Department of Health, began a 3-year investigation in 2000 to increase understanding of the geohydrology and of the important hydrologic factors required to accurately delineate areas contributing recharge and the sources of water to the public-supply wells, which are screened in sand and gravel deposits in valley-fill and wetland settings near potential surface-water sources. The area contributing recharge to a well is defined as the surface area where water recharges the ground water and then flows toward and discharges to the well. Upland areas outside of the modeled area that drain toward the area contributing recharge were drawn on the basis of maps of land-surface contours and the watershed boundary.

In Cumberland, a well field composed of two wells adjacent to the Blackstone River, and on the opposite side of the river in Lincoln, a well field composed of three backup wells, are in the narrow valley of the river between steep hillslopes. Runoff from the uplands is the predominant source of recharge (78 percent) to the valley aquifer during nonpumping conditions. Ground-water-level and river-stage measurements made during this study when the Cumberland well field was operating, and during a previous U.S. Geological Survey study when one well from each well field was operating, indicated that river water infiltrates the aquifer and flows toward the wells during pumping conditions.

Ground-water flow was simulated with a three-layered model representing the stratified glacial deposits in the valley. Areas contributing recharge were determined for the Cumberland well field operating alone at average and maximum pumping rates and for both well fields operating simultaneously at their maximum rates. Simulated areas contributing recharge to the Cumberland well field for both average and maximum pumping rates extend on both sides of the river to the lateral boundary of the model, which is the contact between the valley and uplands. The area contributing recharge for the average pumping rate (total rate of 324 gal/min) is about 0.05 mi^2 and the well field derives most of its water from upland runoff (72 percent) compared to direct infiltration of precipitation (21 percent) and infiltration from the Blackstone River (7 percent). At the maximum pumping rate (total rate of 1000 gal/min), the 0.12-mi^2 area contributing recharge extends farther up and down the valley and the primary source of water to the well field was infiltrated river water (53 percent) compared to upland runoff (35 percent) and precipitation recharge (12 percent). Upland areas draining toward the areas contributing recharge encompass 0.58 and 0.66 mi^2 for the average and maximum rates, respectively. By incorporating the Lincoln well-field withdrawals (total rate of 2,083 gal/min) into the model, the area contributing recharge to the Cumberland well field is reduced to 0.08 mi^2; part of the simulated area that contributes recharge to the Cumberland well field when it is operating alone contributes instead to the Lincoln well field and the Cumberland well field compensates by increasing the percentage of water it withdraws from the river by 11 percent and extends its contributing area slightly down the valley. The upland area draining toward the Cumberland contributing area covers 0.55 mi^2. The area contributing recharge to the Lincoln well field covers 0.08 mi^2 and the main source of water is from the river at 88 percent; upland areas draining toward the area contributing recharge encompass 0.34 mi^2.

In North Smithfield, a public-supply well in a valley-fill setting is adjacent to Trout Brook Pond, which is an extension of the Lower Slatersville Reservoir. Simulated areas contributing recharge were determined for the maximum pumping rate of 200 gal/min and the estimated maximum yield, 500 gal/min, of a proposed replacement well. Water-depth and pond-bottom-sediment profiles indicate that the average water depth of Trout Brook Pond is 5 ft and the thickness of fine-grained bottom sediments ranged from 0 to 5 ft. A comparison of water levels from the pond and underlying sediments indicates that water is not induced from Trout Brook Pond closest to the supply well at the 200-gal/min pumping rate.

Ground-water flow was simulated with a three-layered model representing both stratified glacial deposits and bedrock. The simulated areas contributing recharge for both pumping rates extend to the ground-water divides on both sides of Trout Brook Pond and exclude a capped landfill designated a Superfund site by the U.S. Environmental Protection Agency. Traveltime estimates indicate that water that recharges the aquifer in the vicinity of the capped landfill and within the area contributing recharge takes approximately 30 years to reach the well site. For the 200-gal/min rate, the area contributing recharge is 0.23 mi^2; only a small fraction (0.04 percent) of the total water withdrawn is infiltrated from the pond. When the hydraulic connection between the pond and aquifer was increased by an order of magnitude, the well derived only

1 percent of its water from the pond. Thus, even with the well pumping at a rate of 200 gal/min in close proximity to a surface-water source, the well derived almost all of its water from intercepted ground water that normally discharges to surface-water bodies. For the pumping rate of 500 gal/min, the area contributing recharge covers 0.45 mi^2. The increased pumping rate is balanced by additional intercepted ground water and by induced infiltration (25 percent) from Trout Brook Pond and the Lower Slatersville Reservoir.

In Westerly, one public-supply well is in a Pawcatuck River subbasin, which consists of a densely vegetated freshwater wetland of large areal extent. Wetland deposits, composed of peat, range in thickness from 0 ft at the edges to about 8 ft in the central areas of the wetland. Stage measurements in the wetland and water levels in nests of vertical piezometers installed in the peat and stratified deposits around the well site indicated a downward gradient and, thus, the potential for water in the wetland to infiltrate the underlying aquifer.

Ground-water flow was simulated by a five-layered model representing glacial sediments (stratified deposits and a moraine), peat, and in some areas, directly incorporating the bedrock. Overland flow in the wetlands was simulated by model cells assigned a high hydraulic-conductivity value so as to simulate the lack of resistance to flow and the resulting flat gradient across the wetland. Areas that represent the starting locations of particles that are withdrawn from the well for the average pumping rate (240 gal/min) and the maximum rate (700 gal/min) extend to the surrounding uplands (surficial materials not covered by the wetlands) and to a ground-water divide separating the watersheds of the Pawcatuck River and Block Island Sound. Some particles originating in the uplands discharge to the wetland through shallow ground-water-flow paths, move through the wetland, and infiltrate again into the aquifer before discharging from the well. Some particles originating in the wetland also track as overland flow before infiltrating and discharging from the well.

For the average pumping rate, the upland area contributing recharge directly to the well without passing through the wetland is simulated as 0.13 mi^2 and contributes 46 percent of the total water withdrawn from the well. The remaining water withdrawn from the well is derived from the wetlands or indirectly from the uplands through the wetland from an area of 0.54 mi^2. At the maximum pumping rate, the area contributing recharge from the uplands is simulated as 0.16 mi^2 and supplies 21 percent of the total water pumped; the remaining water is from the wetland or indirectly from the uplands through the wetland from an area of 1.27 mi^2. Thus, for both the average and maximum pumping rates, the primary source of water withdrawn from the well is precipitation recharge to the wetland or discharges to the wetland from the surrounding uplands.

In general, the hydrologic factors that most affect the size and shape of simulated areas contributing recharge and sources of water to wells in valley-fill and wetland settings are recharge rates, the degree of hydraulic connection between surface water and the underlying aquifer, and the location of upgradient ground-water divides. In a narrow valley setting, however, such as the Cumberland and Lincoln study area, with major water sources near a well—a river and large fluxes from upland runoff—the percentage of source water withdrawn from the well is sensitive to these hydrologic factors, but not particularly sensitive to the area contributing recharge. Study results indicate that the area contributing recharge to a well also may extend to the opposite side of a surface-water body, even with this major source of water in close proximity to the well. A well near a surface-water body may not always draw surface water, even when surface and ground waters are well connected; the amount of water that the well can draw from the surface-water body can also depend on the pumping rate and the quantity of ground water that can be intercepted by the well. In a wetland setting, recharge rates are a major factor in the size of the area contributing recharge from uplands, but the area contributing water directly from the wetland shows only slight change because the connection between the wetland and aquifer is the controlling factor and not the flux applied to the wetland.

Acknowledgments

The author expresses appreciation to Clay Commons (Source Water Assessment Coordinator, Rhode Island Department of Health) for initiating and supporting the study and to Neal Fiorio (Superintendent of the Cumberland Water Department), Maurice Trudeau (retired, Superintendent of the Lincoln Water Commission), Michael Romano (formerly in charge of the North Smithfield Water Department), and Paul Corina (Superintendent of Utilities, Westerly Water Department) for providing information on ground-water withdrawals and assisting in the coordination of field activities. The author also thanks Janet Stone (USGS), who constructed geologic sections for the North Smithfield and Westerly study areas and provided insights into glacial geology for all three study areas, and to land owners who granted access to their property to collect sediment samples and to measure water levels.

References Cited

Allen, W.B., Hahn, G.W., and Tuttle, C.R., 1963, Geohydrologic data for the upper Pawcatuck River Basin, Rhode Island: Rhode Island Water Resources Coordinating Board GWM-3, 1 sheet.

Bear, J., 1979, Hydraulics of groundwater: New York, McGraw-Hill, 569 p.

Boelter, D.H., 1965, Hydraulic conductivities of peats: Soil Science, v. 100, no. 4, p. 227–231.

Boelter, D.H., 1969, Physical properties of peats related to degree of decomposition: Soil Science Society of America Proceedings, v. 33, no. 4, p. 606–609.

Chute, N.E., 1949, Surficial geology of the Pawtucket Quadrangle, Rhode Island–Massachusetts: U.S. Geological Survey Geologic Quadrangle Map 2, scale 1:31,680.

Cowardin, L.W., Carter, V., Golet, F.C., and LaRoe, E.T., 1979, Classification of wetlands and deepwater habitats of the United States: Washington, D.C., U.S. Fish and Wildlife Service, FWS/OBS-79/31, 103 p.

de Lima, Virginia, 1991, Stream-aquifer relations and yield of stratified-drift aquifers in the Nashua River Basin, Massachusetts: U.S. Geological Survey Water-Resources Investigations Report 88-4147, 47 p.

Dickerman, D.C., 1984, Aquifer tests in the stratified drift, Chipuxet River Basin, Rhode Island: U.S. Geological Survey Water-Resources Investigations Report 83-4231, 39 p.

D.L. Maher, Co., 1987, Final report on the Crandall property in Westerly, Rhode Island: North Reading, MA, variously paged.

E.C. Jordan, Inc., 1988, Remedial investigation—feasibility study, Landfill and Resource Recovery, Town of North Smithfield, Providence County, Rhode Island, June 1988: E.C. Jordan, Inc., variously paged.

Farnsworth, R.K., Thompson, E.S., and Peck, E.L., 1982, Evaporation atlas for the contiguous 48 United States: U.S. Department of Commerce, National Oceanic and Atmospheric Administration, Technical Report NWS 33, 26 p., 4 pls.

Friesz, P.J., 1996, Geohydrology of stratified drift and streamflow in the Deerfield River Basin, northwestern Massachusetts: U.S. Geological Survey Water-Resources Investigations Report 96-4115, 49 p., 1 pl.

Friesz, P.J., and Church, P.E., 2001, Pond-aquifer interaction at South Pond of Lake Cochituate, Natick, Massachusetts: U.S. Geological Survey Water-Resources Investigations Report 01-4040, 42 p.

Gonthier, J.B., Johnston, H.E., and Malmberg, G.T., 1974, Availability of ground water in the Lower Pawcatuck River Basin, Rhode Island: U.S. Geological Survey Water-Supply Paper 2033, 40 p., 4 pls.

Guthrie, R.C., and Stolgitis, J.A., 1977, Fisheries investigations and management in Rhode Island lakes and ponds, Fisheries Report No. 3: Rhode Island Department of Natural Resources, Division of Fish and Wildlife, 256 p.

GZA GeoEnvironmental, Inc., 1999, Groundwater flow model, proposed Indeck-North Smithfield LLC power plant, North Smithfield, Rhode Island: Providence, RI, GZA GeoEnvironmental, Inc., variously paged.

Harbaugh, A.W., and McDonald, M.G., 1996, User's documentation for MODFLOW-96, an update to the U.S. Geological Survey modular finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 96-485, 56 p.

Hughes, W.D., 1982, Peat resources and glacial geology of Chapman Swamp and adjacent area, Westerly, Rhode Island: Kingston, RI, University of Rhode Island, 135 p.

HydroSource Associates, Inc., 1999, Hydrogeologic study—Tifft Road wellfield, Town of North Smithfield, Rhode Island, October, 1999: Ashland, NH, variously paged.

Jacob, C.E., 1950, Flow of ground water, *in* Rouse, H., Engineering hydraulics: New York, John Wiley & Sons, p. 321–386.

Johnston, H.E., and Dickerman, D.C., 1974a, Availability of ground water in the Blackstone River area, Rhode Island and Massachusetts: U.S. Geological Survey Water-Resources Investigations Report 4-74, 2 sheets.

Johnston, H.E., and Dickerman, D.C., 1974b, Availability of ground water in the Branch River Basin, Providence County, Rhode Island: U.S. Geological Survey Water-Resources Investigations Report 18-74, 39 p., 1 pl.

Lapham, W.W., 1989, Use of temperature profiles beneath streams to determine rates of vertical ground-water flow and vertical hydraulic conductivity: U.S. Geological Survey Water-Supply Paper 2337, 35 p.

Moench, A.F., Garabedian S.P., and LeBlanc, D.R., 2000, Estimation of hydraulic parameters from an unconfined aquifer test conducted in a glacial outwash deposit, Cape Cod, Massachusetts: U.S. Geological Survey Open-File Report 00-485, 139 p.

National Oceanic and Atmospheric Administration, 2002, Climatological data annual summary—New England: Asheville, NC, National Climatic Data Center, v. 114, no. 13, variously paged.

Pollock, D.W., 1994, User's guide for MODPATH/MODPATH_PLOT, version 3—A particle tracking post-processing package for MODFLOW, the U.S. Geological Survey finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 94-464, variously paged.

Prudic, D.E., 1989, Documentation of a computer program to simulate stream-aquifer relations using a modular, finite-difference, ground-water flow model: U.S. Geological Survey Open-File Report 88-729, 113 p.

Quinn, A.W., Ray, R.G., and Seymour, W.L., 1949, Bedrock geology of the Pawtucket quadrangle, Rhode Island–Massachusetts: U.S. Geological Survey Geologic Quadrangle Map 1, scale 1:31,680.

Randall, A.D., 1996, Mean annual runoff, precipitation, and evapotranspiration in the glaciated Northeastern United States, 1951–80: U.S. Geological Survey Open-File Report 96-395, 2 sheets.

Randall, A.D., Thomas, M.P., Thomas, C.E., Jr., Baker, J.A., 1966, Water resources inventory of Connecticut; Part 1, Quinebaug River Basin: Connecticut Water Resources Bulletin no. 8, 102 p.

Reilly, T.E., and Pollock, D.W., 1993, Factors affecting areas contributing recharge to wells in shallow aquifers: U.S. Geological Survey Water-Supply Paper 2412, 21 p.

Richmond, G.M., 1952, Bedrock geology of the Georgiaville Quadrangle, Rhode Island: U.S. Geological Survey Geologic Quadrangle Map 16, scale 1:31,680.

Richmond, G.M., 1953, Surficial geology of the Georgiaville Quadrangle, Rhode Island: U.S. Geological Survey Geologic Quadrangle Map 22, scale 1:31,680.

Rosenshein, J.R., Gonthier, J.B., and Allen, W.B., 1968, Hydrologic characteristics and sustained yield of principal ground-water units, Potowomut–Wickford area, Rhode Island: U.S. Geological Survey Water-Supply Paper 1775, 38 p., 5 pls.

Roy F. Weston, Inc., 1996, Final site inspection prioritization report for Manville wellfield, Lincoln, Rhode Island, October 1996: Wilmington, MA, Roy F. Weston, Inc., variously paged.

Schafer, J.P., 1965, Surficial geologic map of the Watch Hill Quadrangle, Rhode Island–Connecticut: U.S. Geological Survey Geologic Quadrangle Map 410, scale 1:24,000.

Schafer, J.P., 1968, Surficial Geologic Map of the Ashaway Quadrangle, Rhode Island–Connecticut: U.S. Geological Survey Geologic Quadrangle Map 712, scale 1:24,000.

Socolow, R.S., Leighton, C.R., Whitley, J.F., and Ventetuolo, D.J., 2002, Water resources data for Massachusetts and Rhode Island, water year 2001: U.S. Geological Survey Water-Data Report MA-RI-01-1, 307 p.

Stone, J.R., Schafer, J.P., London, E.H., and Thompson, W.B., 1992, Surficial materials map of Connecticut: U.S. Geological Survey Special Map, scale 1:125,000, 2 sheets.

Tiner, R.W., 1989, Wetlands of Rhode Island, U.S. Fish and Wildlife Service, National Wetlands Inventory, Newton Corner, MA, variously paged.

U.S. Environmental Protection Agency, 1991, Protecting the nation’s ground water: EPA’s strategy for the 1990’s: U.S. Environmental Protection Agency, Office of the Administrator, Report 21Z-1020, 84 p.