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Evaluating Prediction Uncertainty of Areas Contributing Recharge to Well Fields of Multiple Water Suppliers in the Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island

Scientific Investigations Report 2012–5114

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KEN SALAZAR, Secretary

U.S. Geological Survey
Marcia K. McNutt, Director

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Conversion Factors, Datum, and Abbreviations

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
foot (ft)	0.3048	meter (m)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

°C=(°F–32)/1.8

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD 29).

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

Altitude, as used in this report, refers to distance above the vertical datum.

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

Mean annual streamflow and mean annual base flow: Mean annual is defined as the arithmetic mean of yearly flows.

Abbreviations

KCWA	Kent County Water Authority
NKWD	North Kingstown Water Department
RIDEM	Rhode Island Department of Environmental Management
RIDOH	Rhode Island Department of Health
RIEDC	Rhode Island Economic Development Corporation
USGS	U.S. Geological Survey

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Evaluating Prediction Uncertainty of Areas Contributing Recharge to Well Fields of Multiple Water Suppliers in the Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island

By Paul J. Friesz

Abstract

Three river basins in central Rhode Island—the Hunt River, the Annaquatucket River, and the Pettaquamscutt River—contain 15 production wells clustered in 4 pumping centers from which drinking water is withdrawn. These high-capacity production wells, operated by three water suppliers, are screened in coarse-grained deposits of glacial origin. The risk of contaminating water withdrawn by these well centers may be reduced if the areas contributing recharge to the well centers are delineated and these areas protected from land uses that may affect the water quality. The U.S. Geological Survey, in cooperation with the Rhode Island Department of Health, began an investigation in 2009 to improve the understanding of groundwater flow and delineate areas contributing recharge to the well centers as part of an effort to protect the source of water to these well centers. A groundwater-flow model was calibrated by inverse modeling using nonlinear regression to obtain the optimal set of parameter values, which provide a single, best representation of the area contributing recharge to a well center. Summary statistics from the calibrated model were used to evaluate the uncertainty associated with the predicted areas contributing recharge to the well centers. This uncertainty analysis was done so that the contributing areas to the well centers would not be underestimated, thereby leaving the well centers inadequately protected. The analysis led to contributing areas expressed as a probability distribution (probabilistic contributing areas) that differ from a single or deterministic contributing area.

Groundwater flow was simulated in the surficial deposits and the underlying bedrock in the 47-square-mile study area. Observations (165 groundwater levels and 7 base flows) provided sufficient information to estimate parameters representing recharge and horizontal hydraulic conductivity of the glacial deposits and hydraulic conductance of streambeds. The calibrated value for recharge to valley-fill deposits was 27.3 inches per year (in/yr) and to upland till deposits was 18.7 in/yr. Calibrated values for horizontal hydraulic conductivity of the valley-fill deposits ranged from 20 to 480 feet per day (ft/d) and of the upland till deposits was 16.2 ft/d. Calibrated values of streambed hydraulic conductance ranged from 10,000 to 52,000 feet squared per day. Values of recharge and horizontal hydraulic conductivity of the valley-fill deposits were the most precisely estimated, whereas the horizontal hydraulic conductivity of till deposits was the least precisely estimated.

Simulated areas contributing recharge to the well centers on the basis of the calibrated model ranged from 0.19 to 1.12 square miles (mi^2) and covered a total area of 2.79 mi^2 for average well center withdrawal rates during 2004–08 (235 to 1,858 gallons per minute (gal/min)). Simulated areas contributing recharge for the maximum well center pumping capacities (800 to 8,500 gal/min) ranged from 0.37 to 3.53 mi^2 and covered a total area of 7.99 mi^2 in the modeled area. Simulated areas contributing recharge extend upgradient of the well centers to upland till and to groundwater divides. Some areas contributing recharge include small, isolated areas remote from the well centers. Relatively short groundwater traveltimes from recharging locations to discharging wells indicated the wells are vulnerable to contamination from land-surface activities: median traveltimes ranged from 2.9 to 5.0 years for the well centers, and 78 to 93 percent of the traveltimes were 10 years or less for the well centers. Land cover in the areas contributing recharge includes a substantial amount of urban land use for the two well centers in the Hunt River Basin, agriculture and sand and gravel mining uses for the well center in the Annaquatucket River Basin, and, for the well center in the Pettaquamscutt River Basin, land use is primarily undeveloped.

Model-prediction uncertainty was evaluated using a Monte Carlo analysis. The parameter variance–covariance matrix from nonlinear regression was used to create parameter sets that reflect the uncertainty of the parameter estimates and the correlation among parameters. The remaining parameters representing the glacial deposits (vertical anisotropy of valley-fill deposits and of till deposits, maximum groundwater evapotranspiration, and hydraulic conductance for head-dependent cells representing a groundwater divide) that could

not be estimated with nonlinear regression were incorporated into the variance–covariance matrix using prior information on parameters. Thus the uncertainty analysis was an outcome of calibrating the parameters to available observations and to information that the modeler provided. A water budget and model-fit statistical criteria were used to assess parameter sets so that prediction uncertainty was not overestimated. Because of the effects of parameter uncertainty, the size of the probabilistic contributing areas for each well center for both average and maximum pumping rates was larger than the size of the deterministic contributing areas for the well center. Thus, some areas not in the deterministic contributing area may actually be in the contributing area, including additional areas of urban and agricultural land use. Generally, areas closest to the well centers with short groundwater traveltimes are associated with higher probabilities, whereas areas distant from the well centers with long groundwater traveltimes are associated with lower probabilities. The deterministic contributing areas generally corresponded to areas associated with high probabilities (greater than 50 percent). Areas associated with low probabilities extended long distances along groundwater divides in the uplands remote from the well centers.

Introduction

Accurate delineation of areas contributing recharge to production wells is an essential component of Federal, State, and local strategies for the protection of drinking-water supplies from contamination (U.S. Environmental Protection Agency, 1991). The area contributing recharge to a well is defined as the surface area where water recharges the groundwater and then flows toward and discharges to the well (Reilly and Pollock, 1993). At the State level, the Source Water Assessment Program of the Rhode Island Department of Health (RIDOH), Office of Drinking Water Quality, was established as the result of the 1996 Amendments to the Federal Safe Drinking Water Act. Since that time, RIDOH has assessed the susceptibility and risk of public-water supplies to contamination, and the agency has encouraged land-use planning within the areas contributing recharge to a production well.

The basins for three rivers in central Rhode Island—the Hunt (H) River, the Annaquatucket (A) River, and the Pettaquamscutt (P) River—contain 15 production wells from which drinking water is being withdrawn (figs. 1 and 2). The town of North Kingstown Water Department (NKWD), Rhode Island Economic Development Corporation (RIEDC), and Kent County Water Authority (KCWA) withdraw drinking water from these large-capacity production wells, which supply a total average daily rate of 5.0 million gallons a day (Mgal/d) from unconfined sand and gravel aquifers of glacial origin. Groundwater withdrawals within this area, referred to as “the HAP” in this report, meet water-quality standards without need for treatment. Wells in the Hunt River Basin, however, are threatened by commercial, industrial, highway, and dense residential development (DeSimone, 1999; Hickey and Joubert, 2003). Nitrate concentrations are slightly elevated in some of these wells, indicating the presence of fertilizers or wastewater. In addition, the gasoline additive methyl *tert*-butyl ether (MTBE) has been detected in some of these wells (Richard Amirault, Rhode Island Department of Health, written commun., 2008). Wells in the Annaquatucket and Pettaquamscutt River Basins are in mostly rural areas, but 30 to 40 percent of the currently protected areas could qualify to be converted to residential use. Land use upgradient of the wells in the Annaquatucket River Basin includes extensive agriculture and sand and gravel mining.

A U.S. Geological Survey (USGS) study in the HAP by Barlow and Dickerman (2001) designed a numerical model to simulate groundwater flow in the glacial valley-fill deposits; this model was updated by Barlow and Ostiguy (2007). One of the purposes of these studies was to determine areas contributing recharge to the production wells. Uplands bordering the valley-fill deposits were incorporated indirectly into the model by adding streamflow and groundwater discharge from these uplands at the edge of the model. Simulated areas contributing recharge to most production wells in the existing HAP models extended to this model boundary, indicating that uplands outside the model may be contributing water to the wells. Land cover in the uplands near some of the wells includes urban and agricultural land uses. If groundwater flow in the uplands was also to be simulated in a new model, the addition would help ensure that the entire area that contributes water to the production wells is included.

The original HAP models were calibrated to hydrologic data manually to provide a reasonable match between field (observed) and simulated groundwater levels and streamflows. Manual calibration, however, may not provide the optimal set of parameter values that give the best model fit to the observations. Results from a manually calibrated model also do not provide a means for quantitatively assessing model-prediction uncertainty concerning, in this case, the size, shape, and location of the area contributing recharge to a well. Because the simulated contributing areas to the HAP wells are generally narrow and long and, in some cases, do not cover the area above the well, the RIDOH and the Rhode Island Department of Environmental Management (RIDEM) are concerned about the practicality of applying these model results. Without an evaluation of the uncertainty associated with the model predictions, the contributing area to a well may be underestimated, thereby leaving the well inadequately protected.

The USGS, in cooperation with the RIDOH, began a 2-year study in 2009 to increase understanding of groundwater flow in the HAP as part of an effort to protect the source of water to these 15 large-capacity production wells. The new study modified the original HAP models to incorporate the surrounding uplands directly into the model and calibrated the model to hydrologic data by inverse modeling using nonlinear regression. Nonlinear regression estimates the optimal set of model-parameter values. In contrast to parameter values that

Figure 1. Location of study area, selected U.S. Geological Survey (USGS) long-term network streamgages and observation wells, and a National Oceanic and Atmospheric Administration (NOAA) climatological station and tidal gage.

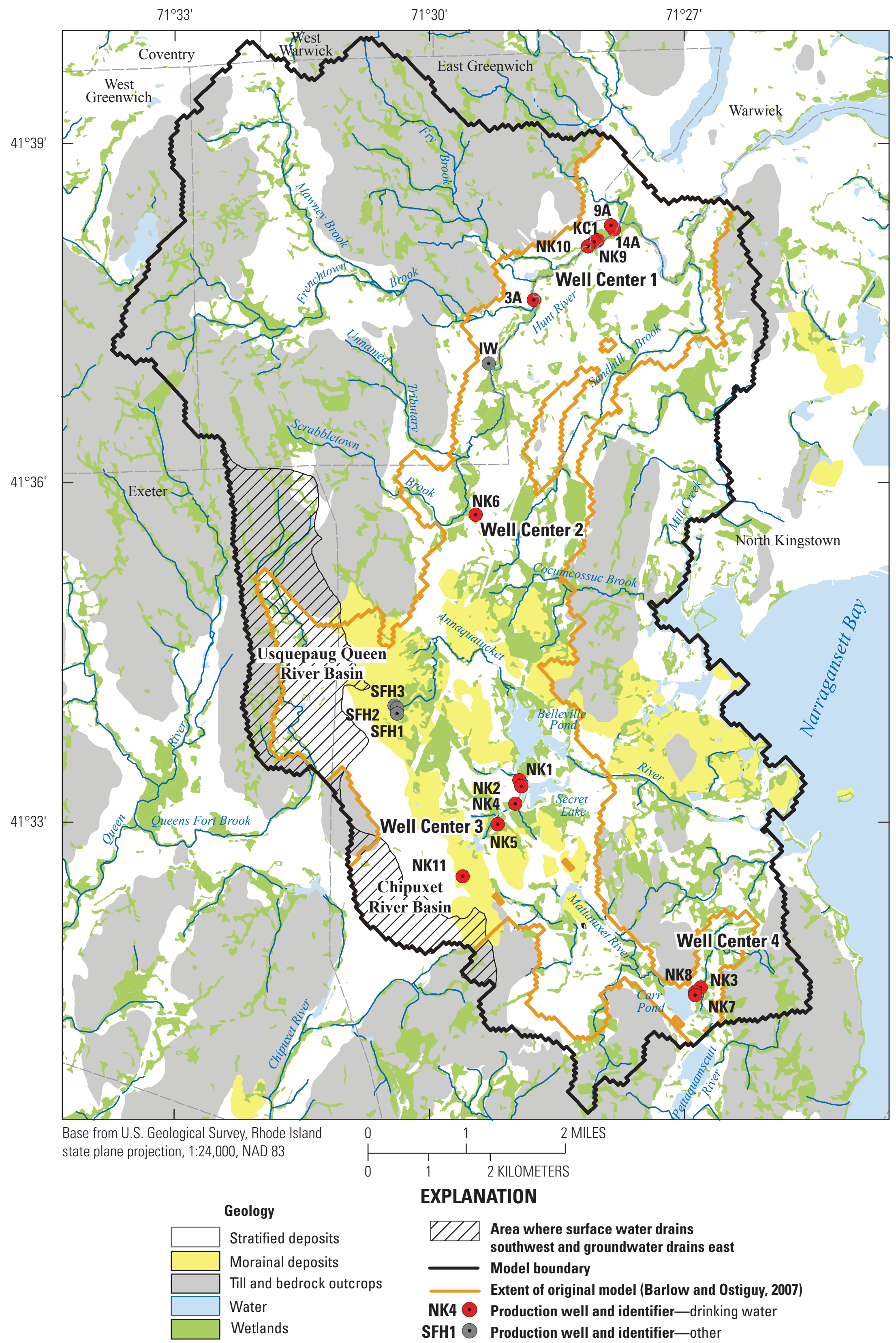


Figure 2. Production wells, model extents, surficial geology, and selected drainage boundaries, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

are calibrated manually, summary statistics from nonlinear regression can be used to provide a quantitative measure of the uncertainty of the predicted contributing area (Starn and others, 2010). The uncertainty that this study considers arises from the observation dataset used for this calibration and not from the model design. Results of the calibrated model (used for a deterministic contributing area) and the uncertainty analysis (used for a probabilistic contributing area) could be useful to water-resource managers when they assess the risk of contamination and implement land-use plans. Quantitative uncertainty analyses of the predicted contributing areas have been applied in only a few studies of small modeled areas. This study applies the uncertainty analysis in a large model with multiple water suppliers and multiple wells.

Purpose and Scope

This report documents the design and calibration of a numerical groundwater-flow model for the purpose of delineating the areas contributing recharge to 15 production wells supplying drinking water in the HAP River Basins. The NKWD operates a total of 11 production wells in the 3 river basins. In the Hunt River Basin, the RIEDC operates three production wells, and the KCWA operates one production well. The original HAP models were modified to simulate groundwater flow in the uplands and for use with inverse modeling. Changes to the lateral extent increased the modeled area from 18.0 to 47.0 square miles (mi^2). The groundwater-flow model was calibrated by inverse modeling using nonlinear regression to estimate the optimal set of parameter values. Simulated areas contributing recharge to the wells based on the optimal set of parameter values (the deterministic contributing area) are shown on maps for selected pumping rates and for average, steady-state hydrologic conditions. Summary statistics from nonlinear regression were used to evaluate the uncertainty associated with the predicted areas contributing recharge to the well fields. Maps depict the results of the uncertainty analysis of the simulated area contributing recharge expressed as a probability distribution (the probabilistic contributing area).

Description of Study Area and Previous Investigations

The HAP study area is in central Rhode Island primarily in the towns of North Kingstown, East Greenwich, and Exeter (figs. 1 and 2). The study area is referred to as the HAP by this and most previous investigations because the major river basins are the Hunt, Annaquatucket, and Pettaquamscutt River Basins; the Pettaquamscutt is called the Mattatuxet in its headwaters. These three rivers, along with Cocumcossuc Brook and other small streams in the eastern part of the study area, discharge to Narragansett Bay. In the southwest, parts of the Usquepaug–Queen and the Chipuxet River Basins are also in the study area (fig. 2). Previous studies, most recently Dickerman and Barlow (1997), have shown that groundwater and surface-water divides do not coincide between the Annaquatucket River Basin and the Usquepaug–Queen and the Chipuxet River Basins. Surface water in these two basins flows southwest of the study area, eventually to the Pawcatuck River and Block Island Sound, but some of the groundwater in these basins flows eastward to the Annaquatucket River Basin. The total size of the HAP study area is 47.0 mi^2.

Previous investigations by the USGS have analyzed the geology and hydrology of part or all of the study area. Surficial and bedrock geology have been mapped by Power (1957, 1959), Quinn (1952, 1963), Schafer (1961), Smith (1955, 1956), and Williams (1964). Reconnaissance studies of groundwater conditions, including measurements of groundwater levels in valley-fill deposits and in upland till, were done by Allen (1956), Allen and others (1959), Hahn (1959), and Johnson and Marks (1959). The first comprehensive investigation of the groundwater and surface-water resources was done by Rosenshein and others (1968). As part of this study they completed contour maps of the bedrock surface and of the water table, saturated thickness, and transmissivity of the valley-fill deposits. Information on streamflows and streambed characteristics was also collected. A water-table map of the valley-fill deposits, based on water-level measurements made during October 1996 at near-average water-level conditions, was prepared by Dickerman and Barlow (1997). A numerical model was developed by Barlow and Dickerman (2001) to simulate groundwater flow in the valley-fill deposits. This same study used optimization techniques to evaluate alternative water-management strategies to reduce streamflow depletion and simulate the areas contributing recharge to the production wells for 1996 average withdrawals. Barlow and Ostiguy (2007) modified this model to account for new hydrogeologic data and used the modified model to assess the effects of existing and proposed well withdrawals at a State fish hatchery in the Annaquatucket River Basin on groundwater and streamflow conditions, especially in nearby wetlands. In addition, the study simulated areas contributing recharge to the production wells for average withdrawals during 2003, and it included a NKWD proposed production well in the Annaquatucket River Basin. An overview of these 2001 and 2007 numerical models is described in the following section of this report.

In the HAP study area, glacial and post-glacial deposits overlie crystalline and metamorphosed sedimentary bedrock (fig. 2). These glacial deposits consist of till, stratified, and morainal deposits; most of these deposits were laid down during advance and retreat of the last (late Wisconsinan) ice sheet, which was retreating northward through southern Rhode Island between 17,000 to 18,000 radiocarbon years ago (Stone and others, 2005). A thin, discontinuous layer of till deposited directly on the bedrock by glacial ice is composed of a poorly sorted mixture of sediments ranging in size from clay to boulders. Stratified deposits consisting of well-sorted, layered sediments ranging in size from clay to gravel that were deposited by glacial meltwater overlie the till in the valleys, including an upland area of Frenchtown Brook Basin. Mixed sediments of till and stratified deposits are present in moraines, which formed at the ice margins (Schafer, 1961). The stratified deposits and the morainal deposits are referred to as valley-fill deposits in this report; they cover about 61 percent of the study area (fig. 2). Deposits in the valley range from a surface altitude of a few feet along the shoreline of Narragansett Bay to 250 feet (ft) along the upland-valley contact in the Usquepaug–Queen River Basin. Till deposits in the uplands rise to 480 ft in the headwaters of Frenchtown Brook. The production wells are screened in coarse-grained valley-fill deposits composed of sand and gravel. Transmissivity of the valley-fill deposits ranges from zero at the upland-valley contacts to about 40,000 feet squared per day (ft^2/d) (Rosenshein and others, 1968); areas with the highest transmissivities (20,000 to 40,000 ft^2/d) are in areas along most of the northeast-flowing Hunt River, including the production wells in the lower Hunt River Basin, and in the Annaquatucket River Basin in an area near Secret Lake and associated NKWD production wells. Post-glacial deposits of alluvium and peat locally overlie glacial deposits in flood-plain and wetland areas (fig. 2).

The climate is humid and temperate with an average annual temperature of about 50°F and average annual precipitation of 48.7 inches (in.), according to records from 1941 through 2008 for climatological station 374266 at Kingston, Rhode Island (National Oceanic and Atmospheric Administration, 2009) (fig. 1). The source of all water in the study area is ultimately from precipitation. Groundwater generally flows from topographical highs in the uplands toward streams and the valley-fill deposits. The groundwater system is recharged by direct infiltration of precipitation, stream leakage, and a small amount of wastewater from septic systems (Barlow and Dickerman, 2001). Groundwater is decreased by discharges to streams and surface-water bodies, by evapotranspiration from the water table, and by pumping at the production wells. The aquifer is generally in close hydraulic connection with the surface-water system (Barlow and Dickerman, 2001). A detailed description of the hydrogeology of the study area is available in Barlow and Dickerman (2001) and Rosenshein and others (1968).

Groundwater is withdrawn by 19 large-capacity production wells (fig. 2, table 1), 15 of which supply drinking water. The focus of this study is these 15 wells: 11 production wells of the NKWD (NK1–NK11), 3 wells of the RIEDC (3A, 9A, and 14A), and 1 well of the KCWA (KC1). The other four wells do not supply drinking water: three (wells SFH1–SFH3) provide water to a State fish hatchery in the Annaquatucket River Basin, and well IW is for an industrial site in the Hunt River Basin.

RIDOH and RIDEM delineate areas contributing recharge and determine susceptibility to contamination by well center. The 15 wells cluster in 4 well centers (fig. 2): well centers 1 and 2 (7 wells) are in the Hunt River Basin, well center 3 (5 wells) is in the Annaquatucket River Basin, and well center 4 (3 wells) is in the Pettaquamscutt River Basin.

Well center 1 consists of six wells (NK9, NK10, KC1, 3A, 9A, and 14A) adjacent to the Hunt River in the lower part of the basin. No withdrawals were made from NK10 during 1996 or 2003, the years that the original models used to simulate areas contributing recharge. Well center 2 has a single well (NK6) near Scrabbletown Brook, a tributary to the Hunt River. Well center 3 consists of five wells (NK1, NK2, NK4, NK5, and NK11) in the Annaquatucket River Basin. Four of these wells are near Secret Lake or a tributary of the lake. The fifth well (NK11), upgradient of the others, began withdrawals in December 2008. Well center 4 consists of three wells (NK3, NK7, and NK8) in the Pettaquamscutt River Basin near Carr Pond and a tributary to the pond.

One measure for assessing well vulnerability to contamination that the RIDOH uses is the amount of high-intensity land use in the contributing area to a well or well center. These urban and agricultural land uses store, apply, or generate pollutants that have the potential to contaminate nearby water resources (Hickey and Joubert, 2003). Such uses include commercial, industrial, high- and medium-high density residential (one-fourth acre or less), medium-density residential (one-fourth to one acre), waste disposal, institutional, transportation, sand and gravel mining, and selected agriculture (cropland, orchards, and nurseries). The distribution of these urban and agricultural land uses (fig. 3) is important in assessing the potential sources of contaminants near the well centers. The remaining land cover not shown in figure 3 is not considered high-intensity land use, including forest, brushland, wetland, low density residential, and other types of agriculture (for example, pasture). Of the four well centers, well center 1, which has the highest pumping rate, is in an area with the most urban land cover, whereas well center 4, with the lowest pumping rate, is in a mostly undeveloped area.

Overview of Original Models

Numerical models of the HAP aquifer were developed by Barlow and Dickerman (2001) and Barlow and Ostiguy (2007) to simulate groundwater flow in the surficial deposits in the valleys (fig. 2). A brief overview of the original steady-state models provides background information for the present study's modifications that more accurately simulate groundwater flow in the uplands bordering the valleys and for the subsequent recalibration of the model with nonlinear regression. A detailed description of the original models is in Barlow and Dickerman (2001) and Barlow and Ostiguy (2007).

The original models simulated groundwater flow in the valley-fill deposits by a four-layered model with a uniformly spaced grid; each model cell was 200 by 200 ft. The model grid was oriented parallel to the northeast-trending valleys of the Hunt River and of Sandhill Brook and to the southeast-trending valleys of the Annaquatucket and Pettaquamscutt Rivers. In the vertical dimension, the model extended from the water table, defined by Dickerman and Barlow (1997), to the bedrock surface, defined by Rosenshein and others (1968)

Table 1. Withdrawal rates for production wells in the groundwater-flow model, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

[gal/min, gallons per minute; ft^3/s, cubic feet per second; KCWA, Kent County Water Authority; RIEDC, Rhode Island Economic Development Corporation; NKWD, North Kingstown Water Department; RIDEM, Rhode Island Department of Environmental Management; --, not applicable]

Water supplier well name	Water supplier	Well center	1996 withdrawal rate		2004–08 withdrawal rate		[1]Maximum withdrawal rate	
			(gal/min)	(ft^3/s)	(gal/min)	(ft^3/s)	(gal/min)	(ft^3/s)
Hunt River Basin								
KC1	KCWA	1	226	0.50	563	1.25	1,200	2.67
3A	RIEDC	1	155	0.35	135	0.30	1,100	2.45
9A	RIEDC	1	113	0.25	142	0.32	1,100	2.45
14A	RIEDC	1	272	0.61	171	0.38	1,100	2.45
NK6	NKWD	2	172	0.38	319	0.71	900	2.01
NK9	NKWD	1	819	1.82	470	1.05	2,000	4.46
NK10	NKWD	1	0	0	377	0.84	2,000	4.46
IW	Industrial	--	174	0.39	[2]174	[2]0.39	[2]174	[2]0.39
	Total for basin		1,931	4.30	2,351	5.24	9,574	21.34
Annaquatucket River Basin								
NK1	NKWD	3	127	0.28	313	0.70	1,000	2.23
NK2	NKWD	3	95	0.21	95	0.21	700	1.56
NK4	NKWD	3	122	0.27	267	0.59	1,000	2.23
NK5	NKWD	3	343	0.76	237	0.53	750	1.67
NK11	NKWD	3	--	--	[3]113	[3]0.25	750	1.67
SFH1	RIDEM	--	438	0.98	[4]290	[4]0.65	[5]290	[5]0.65
SFH2	RIDEM	--	438	0.98	[4]400	[4]0.89	[5]400	[5]0.89
SFH3	RIDEM	--	0	0	[4]600	[4]1.34	[5]600	[5]1.34
	Total for basin		1,563	3.48	2,315	5.16	5,490	12.24
Pettaquamscutt River Basin								
NK3	NKWD	4	124	0.28	60	0.13	250	0.56
NK7	NKWD	4	13	0.03	95	0.21	325	0.72
NK8	NKWD	4	6	0.01	80	0.18	225	0.50
	Total for basin		143	0.32	235	0.52	800	1.78
	Total for all basins		3,637	8.10	4,901	10.92	15,864	35.36

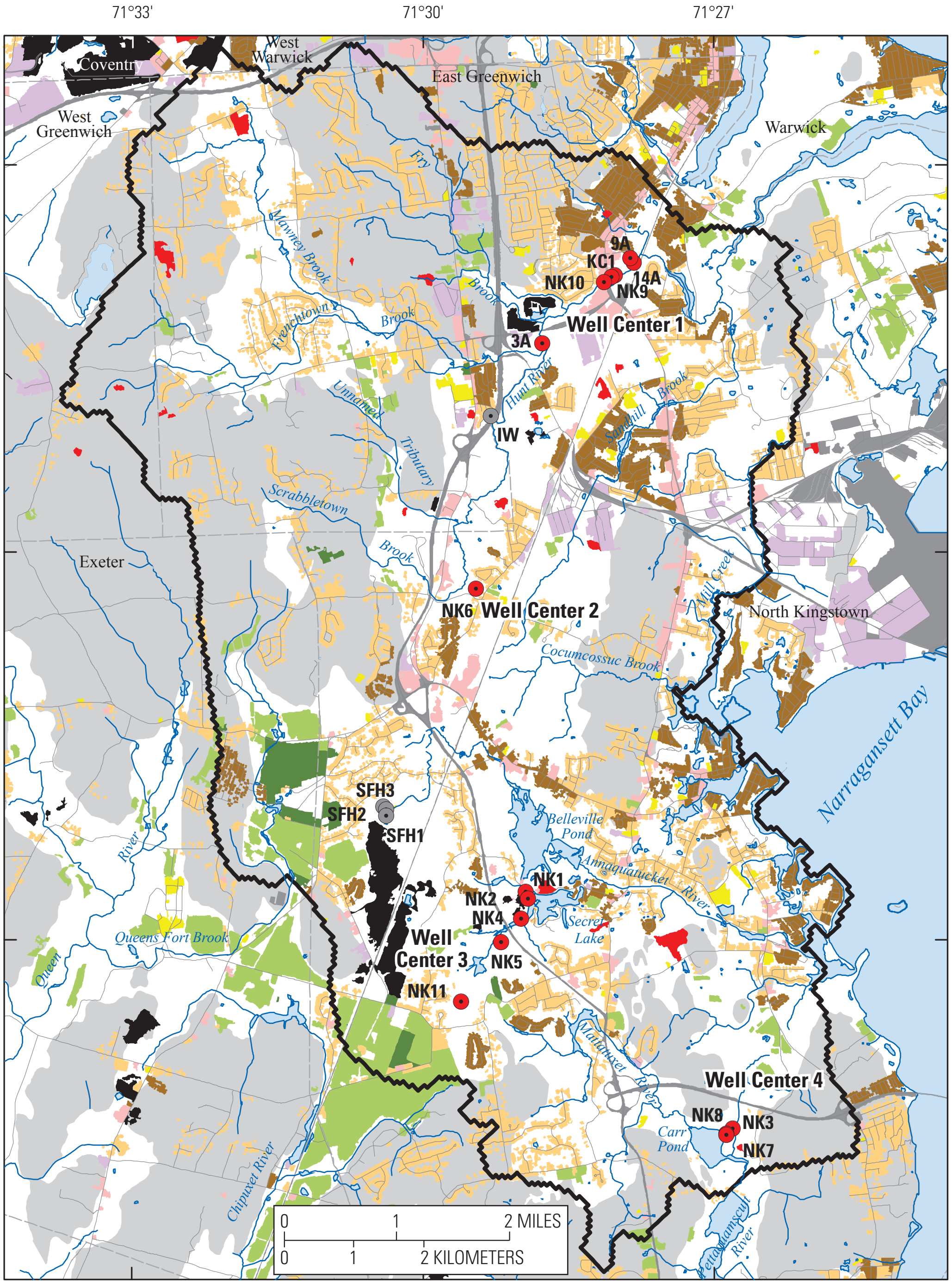
[1]Maximum-rated capacity of pump.

[2]The 1996 average rate was used for Well IW.

[3]The 2009 average rate was used for Well NK11.

[4]Theodore Peters, Rhode Island Department of Environmental Management, written commun., 2010.

[5]The 2004–08 average rate was used for Wells SFH1, SFH2, and SFH3.



EXPLANATION

Geology

- Valley-fill deposits
- Till and bedrock outcrops
- Water

Selected land-use types

- High- to medium-high density residential (less than 1/4 acre)
- Medium-density residential (1/4 to 1 acre)
- Commercial
- Industrial
- Institutional (for example, hospital, schools)
- Transportation
- Waste disposal
- Sand and gravel mining and quarries
- Orchards and nurseries
- Cropland
- **Model boundary**
- NK4 **Production well and identifier**—drinking water
- SFH1 **Production well and identifier**—other

Figure 3. Selected land uses, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

and modified by Barlow and Dickerman (2001). Thin surficial deposits, mostly along the valley edges (fig. 2), were excluded from the active area of the model because simulating thin deposits created numerical instabilities. Barlow and Ostiguy (2007) increased the lateral extent of the active area of the 2001 model to include thin valley-fill deposits west and south of the State fish hatchery wells, and they resolved numerical instabilities by simulating all layers, including the top layer, with aquifer transmissivities that did not change with changing groundwater levels. This change to the lateral extent increased the size of the Barlow and Ostiguy (2007) model from 16.9 to 18.0 mi^2.

Hydrologic processes of recharge, evapotranspiration from the water table, well withdrawals, and the interaction between the aquifer and surface water (streams and ponds) were simulated in the original models. Recharge was from direct infiltration of precipitation and from wastewater in areas that received water supplies but were not sewered. Inflow from the uplands bordering the valley, either by streamflow or by groundwater, was applied to cells along the edge of the model. The southwest extent of the model was along a groundwater divide between the Annaquatucket River Basin and the Chipuxet and Usquepaug–Queen River Basins. Because NK11, a proposed well that would not begin withdrawals until late 2008, would be closer to this model boundary than were existing production wells (fig. 2), Barlow and Ostiguy (2007) modified that part of the groundwater divide in transmissive materials (valley-fill deposits) from a no-flow boundary to a general-head boundary. This type of boundary condition allows groundwater to flow across the simulated divide in response to changing stress conditions in the model.

The original steady-state models simulated long-term average hydrologic conditions represented by the 1941–96 period. The models were manually calibrated to 22 groundwater levels and 5 streamflows. The groundwater levels had been measured on October 8, 1996, at near-average water-level conditions. Long-term average streamflow was calculated at one continuous-record streamgage and estimated at four partial-record streamgages.

Simulation of Groundwater Flow

For this study, groundwater levels and flows were simulated in the surficial deposits and the underlying bedrock in the HAP study area using a three-dimensional finite-difference numerical model code MODFLOW-2000 (McDonald and Harbaugh, 1988; Harbaugh and others, 2000). The original models were modified to (1) simulate groundwater levels and flows in the uplands (fig. 2) and (2) represent boundary conditions and hydraulic properties as parameters (table 2) for calibration by nonlinear regression and for evaluating model-prediction uncertainty. The groundwater-flow model was calibrated to long-term average, steady-state hydrologic conditions based on 172 groundwater-level and streamflow (base flow) observations.

Model Extent and Spatial Discretization

The geographic extent of the active area of the original models was expanded to include all of the thin surficial deposits in the valley and the predominately till uplands draining toward the valley (fig. 2). Topographical divides in the relatively low-permeability till where groundwater and surface-water divides are most likely to coincide were used for most of the lateral extent. In addition, the model was extended eastward to Narragansett Bay and one of its tributaries, Mill Creek—hydrologic features that serve as model boundaries (fig. 4). Thickness of surficial deposits near Narragansett Bay range from 0 to about 60 ft (Rosenshein and others (1968)). Modifying the southwest extent of the model along the groundwater divide in transmissive deposits separating the Annaquatucket River Basin from the Chipuxet and Usquepaug–Queen River Basins was beyond the scope of this study. These changes to the lateral extent of the model were made to (1) improve representation of the groundwater system, (2) move the edge of the model farther from pumping centers in most places, and (3) allow for additional streamflow observations in model calibration, or to more accurately simulate the drainage area of streamflow observations used in the original models. Information for constructing the model in these new areas was available from Rosenshein and others (1968) and from USGS surficial geological and topographical quadrangles. These changes to the 2007 model extent increased the active area from 18.0 to 47.0 mi^2. The model grid was extended using the same cell size (200 by 200 ft) and orientation (43 degrees north of east) as the original models. The model grid consisted of 205 rows and 293 columns, and it included a total of 130,928 active cells.

Model layers were simulated by using a fixed transmissivity, including the top layer, to linearize and thereby simplify the numerical calculations and increase numerical stability. Barlow and Ostiguy (2007) changed the top layer of the model by Barlow and Dickerman (2001) from a variable transmissivity to a fixed transmissivity because of numerical instability caused by incorporating into the model thin surficial deposits near the State fish hatchery wells. Similar numerical instability would have occurred when simulating thin layers on sides of steeply sloping hills, such as the uplands. Calibration by inverse modeling when using nonlinear regression can also present convergence difficulties for the model. Another advantage of using a fixed transmissivity is that it increases the number of model simulations that converge for the analysis of the probabilistic contributing area. Simplifying the numerical calculations by using a fixed transmissivity is described by Hill (1998) and Hill and Tiedeman (2007). Barlow and Ostiguy (2007) reported that changes made to the model, including the fixed transmissivity for layer 1, did not result in substantial changes to simulated groundwater levels, streamflow, or hydrologic budgets.

In the original models, the valley-fill deposits were subdivided vertically into four model layers that extend from the water table to the bedrock surface. The lateral extent of

Table 2. Definition of model parameters and statistics on parameter values, whether estimated or specified, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

[ft/d, feet per day; in/yr, inches per year; --, dimensionless or not applicable]

Parameter name	Parameter description	Units	Optimal or specified value	95-percent confidence interval (Lower value–Upper value)	Coefficient of variation (dimensionless)
	Estimated by nonlinear regression				
K_MULT	Multiplier of horizontal hydraulic conductivity of valley-fill deposits	--	0.81	0.67–1.00	0.10
K_TILL	Horizontal hydraulic conductivity of till deposits	ft/d	16.2	10.4–25.1	0.23
R_TILL	Effective recharge rate on till deposits	in/yr	18.7	13.1–24.3	0.15
R_VF	Recharge rate on valley-fill deposits	in/yr	27.3	22.2–32.4	0.09
SB_MULT	Multiplier of streambed hydraulic conductance	--	2.6	2.0–3.5	0.15
	Specified by prior information from the literature				
ETM	Maximum evapotranspiration from the water table in valley-fill deposits	in/yr	21	13.8–28.3	--
GHB_MULT	Multiplier of hydraulic conductance	--	1	0.25–4.0	--
K_ROCK	Horizontal hydraulic conductivity of bedrock	ft/d	0.1	--	--
K_SW	Horizontal hydraulic conductivity of surface water	ft/d	50,000	--	--
KV_ROCK	Ratio of horizontal to vertical hydraulic conductivity of bedrock	--	1	--	--
KV_TILL	Ratio of horizontal to vertical hydraulic conductivity of till deposits	--	5	1–25	--
KV_VF	Ratio of horizontal to vertical hydraulic conductivity of valley-fill deposits	--	5	1–25	--
R_SW	Effective recharge rate on surface water	in/yr	19.5	--	--

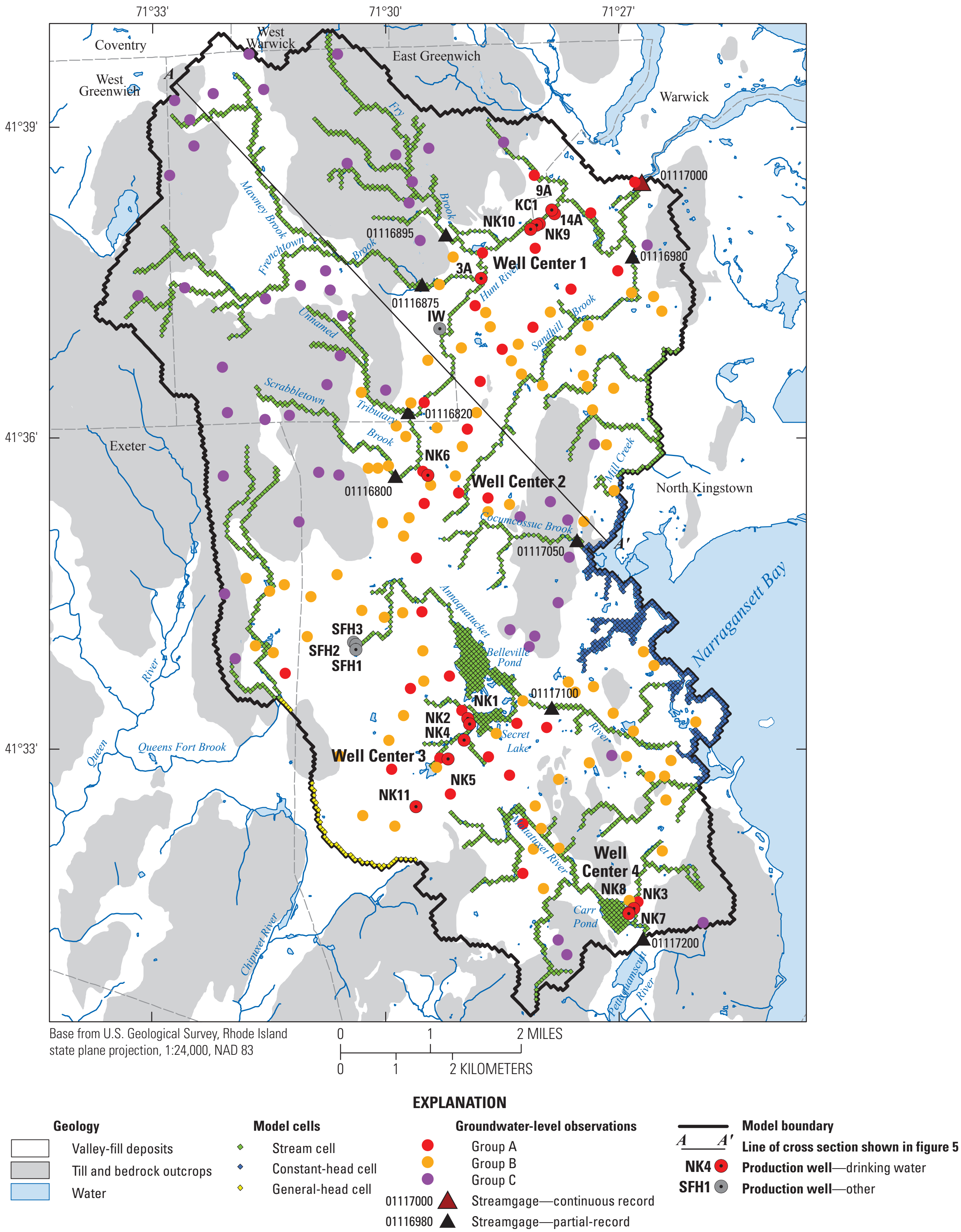


Figure 4. Section line, model-boundary types, groundwater-level observations, and continuous-record and partial-record streamgages, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

the active area decreased from the top layer to the bottom layer in conformity with the shape of the bedrock surface. For this study the model was relayered to provide a realistic representation of groundwater flow from the thin deposits in the uplands and near the valley-upland contact to the thicker valley deposits. A layer to represent bedrock was also added to allow for groundwater flow in bedrock areas, especially where surficial deposits are thin, such as beneath the uplands. Although the model layering was changed, the total thickness of the saturated valley-fill deposits was approximately the same in areas coincident to both models.

The model was relayered by first representing surficial deposits in one layer (the top layer), and bedrock in a second layer (the bottom layer). The top of the model grid, which is used to determine transmissivity of the top layer, was initially set at land-surface altitude, but then it was reset by running the two-layered model, temporarily adjusting hydraulic properties until the simulated water table compared favorably to available data (water-table map by Dickerman and Barlow (1997), and water-level and land-surface altitudes). This simulated water-table altitude was then used as the top of the model in subsequent simulations. The top layer was next subdivided into three layers (layers 1 to 3) based on surface-water features, lithology, and placements of the production well screens. Layer 1 is 5 ft thick to represent surface water in the ponds and to simulate shallow groundwater flow near surface water accurately. Layers 2 and 3 each represented 50 percent of the remaining deposits; the production wells are screened in these layers. Stratified, morainal, and till deposits are represented in all three layers. Shallow bedrock areas less than 7 ft from the top of the model were incorporated into surrounding surficial materials. The bottom layer (now layer 4) represented bedrock with a constant thickness of 200 ft throughout the model beneath the surficial deposits. An example of the model layering from the uplands of Hunt River Basin to Narragansett Bay is shown in figure 5.

Boundary Conditions

The model specified the same types of boundary conditions that the original models specified to represent sources of recharge and areas of discharge, except for a new boundary type to represent Narragansett Bay (fig. 4). Well withdrawals and wastewater return flow are discussed in a subsequent section. Recharge was defined on the basis of surficial geology and surface water (table 2). A single recharge parameter (R_VF) represented both the stratified deposits and morainal deposits; conceptually, because morainal deposits consist of permeable materials, recharge rates should be similar for both surficial materials. A recharge rate to ponds and lakes, defined by parameter R_SW, was not considered for parameter estimation but was instead specified for all model simulations as 19.5 inches per year (in/yr) by subtracting the average annual evaporation rate of a free-water surface from the average annual precipitation rate (Barlow and Dickerman, 2001). In new areas of the model, a third recharge parameter was defined for upland till (R_TILL).

Evapotranspiration from the water table in the valley-fill deposits was simulated with the evapotranspiration package of MODFLOW. Conceptually most evapotranspiration of groundwater in the study area occurs in wetlands and near surface-water features. A maximum evapotranspiration rate was simulated where the water table was at land surface, and this rate decreases linearly to zero where the water table was 4 ft or more below land surface (Barlow and Dickerman, 2001). Parameter ETM (table 2) was defined to represent this maximum evapotranspiration rate. Evapotranspiration from the water table in till deposits was not simulated directly in the model because simulated water levels may not be accurate enough to accurately simulate evapotranspiration. R_TILL, therefore, represents an effective recharge rate, which accounts for the effects of groundwater evapotranspiration, whereas the recharge parameter (R_VF) for stratified and for morainal deposits represents water that reaches the saturated zone, some of which may be lost through evapotranspiration. This same approach was used by DeSimone (2004) in a similar setting in eastern Massachusetts.

Stream-aquifer interactions were simulated as a head-dependent flux boundary in layer 1 by using the stream-routing package (Prudic, 1989) developed for MODFLOW. The stream-routing package accounts for gains and losses of water in each stream cell, and it routes streamflow from upstream cells to downstream cells. Streams that flow into and out of ponds and lakes were simulated as flowing through these water bodies. Each stream-routing cell requires a conductance term that incorporates the geometry and the vertical hydraulic conductivity of the streambed. Field measurements of the vertical hydraulic conductivity of streambed sediments in the study area by Rosenshein and others (1968) ranged from 0.1 to 15.2 feet per day (ft/d) for materials ranging from organic-rich sediments to coarse sand and gravel. These values were used in the original models for different stream reaches based on field observations or proximity of the stream to known bed sediments. Most streambed conductance values in the original models for small streams generally were 4,000 ft^2/d, for medium streams were 10,000 ft^2/d, and for large streams were 20,000 ft^2/d. These same conductances were used as a guide when assigning bed conductances to stream-routing cells in new areas of the model. Stream altitudes in new areas of the model were determined from, or interpolated between, topographical contours that intersected streams listed on USGS topographical quadrangles. The new model contained 2,277 stream-routing cells. A dimensionless parameter (SB_MULT, table 2) was defined that multiplied the streambed conductance values. Thus a SB_MULT value of 1 would be equivalent to the streambed conductances assigned based on the original models.

A constant-head boundary was used to simulate Narragansett Bay and associated coves. The stage of the Bay and the coves was specified as 0.6 ft NGVD29 based on a long-term tidal gage in Narragansett Bay, Newport, R.I.

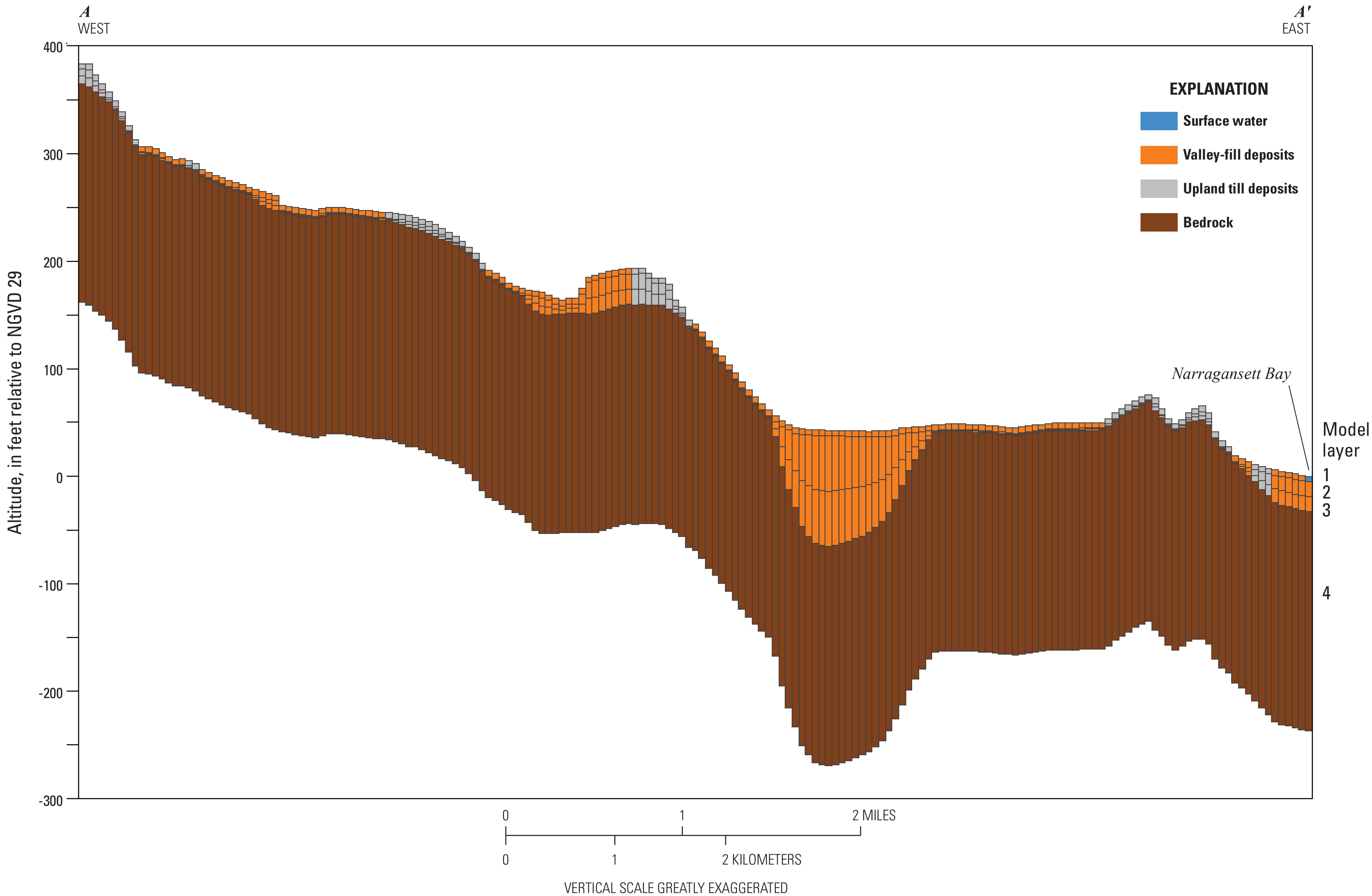


Figure 5. Model cross section, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

(National Oceanic and Atmospheric Administration, 2010) (fig. 1). This altitude represents the stage of the Bay in 1996 when the most accurate groundwater levels used in nonlinear regression were measured. The model contained 306 constant-head cells.

A general-head boundary, which represented the groundwater divide in transmissive materials between the Annaquatucket River Basin and the Chipuxet and Usquepaug–Queen River Basins, required a hydraulic conductance and groundwater head for each boundary cell. The model contained a total of 131 general-head boundary cells in three layers along this divide. The hydraulic conductance value was calculated in the same manner as that in Barlow and Ostiguy (2007). This hydraulic conductance value was determined from aquifer transmissivity by Rosenshein and others (1968) and from model cell dimensions; values ranged from 250 to 1,500 ft^2/d. A dimensionless parameter (GHB_MULT, table 2) was defined that multiplied the hydraulic conductance values. The hydraulic conductivity (transmissivity divided by saturated thickness) in this area of the model was 50 ft/d. A GHB_MULT value of 1 is equivalent to these hydraulic conductances or to this hydraulic conductivity value. For groundwater head at each boundary cell, Barlow and Ostiguy (2007) initially assigned values based on water-table maps by Dickerman and Barlow (1997), Hahn (1959), and Johnson and Marks (1959) and then, during manual calibration, modified the groundwater head specified at each boundary cell so that simulated flow across the boundary was close to zero. After optimal model-parameter values were estimated by nonlinear regression for the present study, this boundary condition was activated and groundwater heads were again adjusted so that simulated flow across the divide was minimal. Although GHB_MULT was not calibrated using nonlinear regression, the uncertainty in the specified value was included in the prediction uncertainty analysis.

Hydraulic Properties

Hydraulic conductivity parameters were assigned on the basis of lithologic units (table 2). Horizontal hydraulic conductivity of valley-fill deposits in the original models was initially calculated as aquifer transmissivity divided by its saturated thickness. Transmissivity contours by Rosenshein and others (1968) and as modified by Barlow and Dickerman (2001) were drawn on the basis of aquifer-test results and lithologic logs. The aquifer-test results provided data on the relation between hydraulic conductivity and grain size, and the values determined from lithologic logs were based on this relation. These horizontal hydraulic values are 20 ft/d for very fine sand, 50 ft/d for fine sand, 100 ft/d for sand, 200 ft/d for sand and gravel, and 500 ft/d for gravel. Along with additional values determined by Dickerman (1984), these values have since been used in most Rhode Island groundwater studies. Saturated thickness was calculated by subtracting bedrock-surface altitude contours (Rosenshein and others, 1968, modified by Barlow and Dickerman, 2001) from water-table contours in Dickerman and Barlow (1997). However, Barlow and Dickerman (2001) adjusted some hydraulic conductivity values during manual calibration because of sparse data in some areas and because the transmissivity contours did not take into account the geologic framework of stagnation-zone retreat of glacial ice and of morphosequence deposition as described by Stone and others (2005) for other valley-fill settings. The final values from the manually calibrated model ranged from 25 to 587 ft/d, and they averaged approximately 180 ft/d (Barlow and Ostiguy, 2007). This range included 1,069 different values for horizontal hydraulic conductivity.

Model cells in new areas of the model between the edge of the 2007 model and the upland-valley contact that included valley-fill deposits were initially assigned horizontal hydraulic conductivity values based on values from the 2007 model for adjacent model cells. Stratified and morainal deposits in new, large areas of the model that are in the Narragansett Bay and Frenchtown Brook drainage areas were initially assigned a horizontal hydraulic conductivity value of 100 ft/d, which is representative of sand. Transmissivity contours by Rosenshein and others (1968) in the Narragansett Bay area were based on few data, and the upland area of Frenchtown Brook was not included in their analysis.

Three different parameterizations of the horizontal hydraulic conductivity of the valley-fill deposits were evaluated during initial parameter estimation. One of these schemes grouped the hydraulic conductivity values into three parameter zones, and a second into four parameter zones. The range of hydraulic conductivities in each parameter zone was guided by the relation between hydraulic conductivity and grain size mentioned earlier in this section. A third parameterization defined a dimensionless parameter, K_MULT, that multiplied the spatially varying hydraulic conductivity values. A K_MULT value of 1 would be equal to the hydraulic conductivity values in the original models and to those values initially assigned for new areas of valley-fill deposits. Model-fit statistics that can be used to compare models with alternative parameterizations, AIC and BIC (Hill and Tiedeman, 2007), indicated that the model using K_MULT had a slightly better model fit than the models using the other two parameterization schemes. Also, a comparison of weighted residuals indicated that the model using K_MULT more randomly distributed weighted residuals than did the other two schemes. Finally, the model that used K_MULT was more stable than the other two models during initial parameter estimation. Because these model-fit statistics and weighted residual comparisons indicated that the model with K_MULT was more accurate than the other two models, and because of its regression stability, the K_MULT model was used in subsequent model simulations.

Two lithologic units were not in the original models: till and bedrock. Parameter K_TILL represented horizontal hydraulic conductivity of till, and parameter K_ROCK represented horizontal hydraulic conductivity of bedrock. A parameter representing surface-water bodies, K_SW, which includes ponds, lakes, and Narragansett Bay and its coves, was specified for all model simulations. Model cells containing surface-water bodies were assigned a horizontal hydraulic conductivity value of 50,000 ft/d to simulate minimal resistance to flow and the corresponding flat gradient across these water bodies (Barlow and Dickerman, 2001).

The ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity for the surficial deposits was represented by two parameters: KV_VF for valley-fill deposits and KV_TILL for upland till deposits. The ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity for bedrock was defined by parameter KV_ROCK.

Groundwater Withdrawals and Return Flow

Average annual withdrawal rates for 1996 at the production wells were specified in the model using the rates listed in table 1. Barlow and Dickerman (2001) determined that the total withdrawal rates from each basin for 1996 were representative of total long-term average withdrawals from each basin during 1941–96 based on available records. Withdrawals by the State fish hatchery wells are discharged to the headwaters of the Annaquatucket River after use at the hatchery.

Wastewater return flow to groundwater, such as by septic systems, was specified in the model for all model simulations. Wastewater return flow in new areas of the model that received water supplies but are not sewered was determined in the same manner as in the original areas of the model. The amount of wastewater return flow was determined from information on the location and rates of water-supply deliveries available for 1996 (Barlow and Dickerman, 2001). The total amount of wastewater return flow applied to the model was 2.1 cubic feet per second (ft^3/s), which compared to 1.2 ft^3/s in the original modeled area.

Observation Data

Parameter values in the groundwater-flow model were estimated using two types of observations—groundwater levels and base flow (groundwater discharge). Observations were weighted on the basis of methods described by Hill and Tiedeman (2007); the weights account for measurement error and for the difference in units between groundwater levels and base flows. More accurate observations are given larger weights than less accurate observations and thus have more influence in the regression and on the estimated parameter values. Different weights also allow for more observations to be included in the regression. The model was calibrated to 172 observations located throughout the modeled area.

Groundwater Levels

Groundwater-level observations used in model calibration were selected from available groundwater levels compiled from previous USGS investigations. A subset (165 groundwater levels) was chosen to provide a generally areal distribution of observations. Near the valley-upland contact, which are areas that have large differences in hydraulic properties, a closer spacing was used if observations were available. Also, groundwater-level observations were selected that were at, or near, average groundwater-level conditions, except in some areas of the model where either none or few groundwater levels were available at average conditions. Groundwater-level conditions were determined by long-term observation well NKW255 in the southern part of the study area, or for measurements that preceded those for this observation well, long-term observation well CHW18 in southern Rhode Island (fig. 1). Water levels have been measured monthly at NKW255 since August 1954 and at CHW18 since October 1946. Both observation wells are screened in valley-fill deposits.

Groundwater-level observations were divided into three groups (A, B, and C), depending on their estimated accuracy (fig. 4). Group A included 33 groundwater levels measured October 8, 1996, near average groundwater-level conditions (Dickerman and Barlow, 1997). Group A observations were in valley-fill deposits screened in model layers 1, 2, and 3. A subset (22 groundwater levels) of this group was used in the manually calibrated models. Group B and Group C groundwater levels were originally measured to construct groundwater maps at a USGS quadrangle scale (Allen, 1956; Allen and others, 1959; Hahn, 1959; Johnson and Marks, 1959). Because most of these water levels were measured in shallow dug wells, they were included in model layer 1. Group B included 83 groundwater levels measured August 1952 and periods of August–November 1954 at, or near, average groundwater conditions. These observations were located in valley-fill deposits. Group C included 49 groundwater levels measured in till or in a thin veneer of stratified deposits in the upland part of Frenchtown Brook Basin. Some of these groundwater levels were measured during the same near-average groundwater-level conditions of August–November 1954 that were used for Group B. Group C also included measurements made August 1949 and August 1951 at below average water-level conditions; most of these observations were in the upland part of Fry Brook and Mawney Brook Basins.

Observation weights are equal to the inverse of the sum of the individual variances for each type of measurement error (Hill and Tiedeman, 2007). The variance is the square of the standard deviation. Three types of measurement error were determined for each group: (1) water-level measurement error, (2) measuring-point altitude error, and (3) uncertainty in average conditions due to seasonal water-level fluctuations.

Water levels were measured with a steel tape or an electrical tape to within 0.01 ft for all three groups. To relate the water-level error to a normal distribution, the study used a 95-percent confidence interval (1.96).

Measuring-point altitudes for Group A were surveyed to within 0.001 ft, and a 95-percent confidence interval (1.96) was also used. Measuring-point altitudes for Group B and Group C were based on land-surface altitude determined from USGS quadrangles. Land-surface contour intervals on these quadrangles are within 5 ft at the 90-percent confidence interval (1.65) (U.S. Geological Survey, 1980). In the uplands, where land-surface topography is steeper than in the valley and where the exact location of the well may not be known (thereby adding more uncertainty to the measuring-point altitude), an accuracy to within 10 ft was used for Group C.

Uncertainty in the groundwater-level measurements to represent average steady-state conditions because of seasonal water-level fluctuations was estimated for each group. One long-term observation well was used to assess average water-level conditions in the study area, a dynamic system with numerous hydrologic settings. Usually one water-level measurement was made at each observation well, and selected water-level observations used in nonlinear regression were made over several decades. The standard deviation of this measurement error was calculated as the range in the seasonal water-level variation divided by 4, which assumes that the observation represents the average and that the range represents the 95-percent confidence interval (Hill and Tiedeman, 2007). Seasonal water-level fluctuations in valley-fill deposits typically vary less than in till deposits because of the larger storage capacity of valley-fill deposits. For 14 observation wells measured monthly from November 1995 to 1996 in the valley-fill deposits, water levels fluctuated about 2 to 4 ft, with the largest fluctuations at the highest water-level altitudes (Barlow and Dickerman, 2001). Hahn (1959) and Johnson and Marks (1959) reported seasonal fluctuations ranging from 3 to 4 ft in the valley-fill deposits. Water levels at long-term well NKW255, screened in valley-fill deposits, typically fluctuated 4 ft annually. For Group A and Group B, screened in valley-fill deposits, a range of 4 ft was assumed for all observations resulting in a standard deviation of 1 ft. For till deposits in the study area, Allen and others (1959) and Hahn (1959) reported seasonal fluctuations of 16 to 17 ft. Four USGS long-term network observation wells (EXW238, EXW278, WCW59, and WGW206), screened in till deposits in central Rhode Island near the study area (fig. 1), typically fluctuated annually from 2 to 15 ft. For Group C, a range of 16 ft was assumed for all observations, or a standard deviation of 4 ft.

The standard deviation of the total error (square root of the total variance for each group) ranged from 1.0 ft for Group A to 7.3 ft for Group C. Corresponding observation weights ranged from 1.0 ft^{-2} for Group A to 0.02 ft^{-2} for Group C (table 3).

Base Flow

Long-term mean annual base flow calculated at eight partial-record sites and at the Hunt River streamgage (01117000) (fig. 4) were used for seven base-flow observations. In the original models, streamflows made at the partial-record sites were related to streamflow at the same continuous streamgage by application of a graphical-correlation technique developed by Searcy (1959) to estimate mean annual streamflow at the partial-record site. For this study, the partial-record sites were related to multiple streamgages by a mathematical method termed Maintenance of Variance Extension, Type 1 (MOVE.1) developed by Hirsch (1982) to calculate both mean annual streamflow and base flow. Base flow, or groundwater discharge, is a measure of effective recharge (recharge minus groundwater evapotranspiration).

Streamflow measurements made at the 8 partial-record sites were related to concurrent mean daily streamflow at several unregulated continuous streamgages in Rhode Island and Connecticut (fig. 1) to estimate the selected statistics. Hunt River streamgage (01117000) was not used in the analysis to determine selected streamflow statistics at the partial-record sites because of groundwater withdrawals upstream of the gage. In addition to instantaneous streamflow measurements at Annaquatucket River at Belleville (01117100), mean daily streamflow from continuous records (September 1961 through 1964) were also available and used in the analysis. The drainage areas for the continuous streamgages have topographical, geologic, and climatic conditions similar to those of the drainage areas of the partial-record sites. Plots of log-transformed data were made to determine the quality and linearity of the relation between the partial-record sites and the continuous

Table 3. Summary information on weighted residuals (observed minus simulated values) for groundwater-level observations, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

[--, not applicable]

Groundwater-level observations, by group (figure 4)	Number of observations	Weight (feet^{-2})	Average (feet)	Minimum (feet)	Maximum (feet)	Standard deviation (feet)	Percentage within two standard errors of the regression
A	33	1	-0.97	-9.26	5.20	3.65	91
B	83	0.1	0.19	-8.28	10.45	3.02	92
C	49	0.02	0.11	-4.15	3.85	1.74	100
All	165	--	-0.07	-9.26	10.45	2.87	94

streamgages. Streamgages with the highest correlation coefficient (two to three gages) were used to estimate the selected streamflow statistics at the partial-record sites. The MOVE.1 was used to provide an equation that related streamflow at the partial-record site to that at the continuous streamgage. Mean annual streamflow and mean annual base flow at the continuous streamgages for complete years of record were entered into the equation to estimate the corresponding mean annual streamflow and base flow at the partial-record sites. Base flow at the continuous streamgages was calculated by use of the automated hydrograph-separation technique PART (Rutledge, 1998). The associated mean square error for each relation was used to combine the multiple estimates for each partial-record site into weighted-average estimates of the selected streamflow statistics to obtain the single best estimate. A detailed description of the MOVE.1 technique for streamflow analysis is described in Ries and Friesz (2000). Computer programs by Granato (2009) facilitated the streamflow analysis.

Information concerning the analysis for the partial-record sites is summarized in table 4 and that for the continuous streamgages in table 5. In addition to mean annual streamflow and mean annual base flow estimated for each partial-record site, table 4 lists the equivalent rate over the drainage area for six of the eight partial-record sites where the surface-water and groundwater divides are most likely to be similar. Mean annual base flow ranged from about 14.8 to 27.7 in/yr over the drainage areas, with a median of 22 in/yr. The lowest value was for Fry Brook (01116895), which drains predominately till deposits (98 percent). Base flow at all eight partial-record sites ranged from 62 to 92 percent of total streamflow with a median of 79 percent. The difference between total streamflow and base flow is overland runoff. The lowest percentage was for the till-dominated drainage area of Fry Brook, and the highest percentage was for Annaquatucket River (01117100). The Annaquatucket River Basin is covered by 90-percent valley-fill deposits, and its surface-water drainage area and its groundwater contributing area are not the same.

Mean annual base flow at six of the eight partial-record sites was used directly as observations in model calibration (table 6). Mean annual base flow at Scrabbletown Brook (01116800) and at Tributary to Hunt River (01116820) were not used as observations themselves but were used to compute an observation that represented a net gain in base flow along the main stem of the Hunt River. Mean annual base flow at the partial-record sites was estimated from multiple measurements made over several years and related to multiple long-term streamgages. For the purpose of weighting these base-flow observations in parameter estimation, these mean annual base flows were assumed to be accurate within 10 percent at the 95-percent confidence interval.

The observation that represented a net gain in base flow along the main stem of the Hunt River (Hunt River observation, 14.1 ft^3/s, table 6) was calculated by subtracting the total mean annual base flow estimated at the five partial-record sites (gages 01116800–1116980) in the Hunt River Basin (23.9 ft^3/s) (fig. 4 and table 4) from a mean annual base flow calculated from streamflow records at the Hunt River streamgage (01117000; 38.0 ft^3/s) (fig. 4, table 5). Streamflow records at the Hunt River streamgage, although they may have been affected by groundwater withdrawals upstream of the streamgage, were used to determine mean annual base flow with the hydrograph-separation technique PART (Rutledge, 1998). The Hunt River base-flow observation of 14.1 ft^3/s is equivalent to 28.0 in/yr over 6.97 mi^2 of mostly valley-fill deposits. The accuracy of the mean annual base flow at the Hunt River streamgage was assumed to be 5 percent; the total variance of the Hunt River observation was calculated by adding the variances of the five partial-record sites and the streamgage. The coefficient of variation of the Hunt River observation (standard deviation divided by the mean) from which the observation weight was calculated was 0.08 (8 percent), which compares to a coefficient of variation of 0.05 (5 percent) for each of the observations determined at the partial-record sites.

Calibration

The groundwater-flow model was calibrated with the Parameter Estimation (inverse modeling) process of MODFLOW-2000 (Hill and others, 2000; Hill and Tiedeman, 2007) using nonlinear regression that minimizes the differences, or residuals, between field (observed) and simulated water levels and base flows to obtain an optimal set of parameter values. The quality of this calibration was determined by analysis of the residuals and the accuracy of the estimated parameter values. Some parameters, however, may be insensitive to the available observations, or some parameters may be highly correlated with each other and therefore cannot be estimated by nonlinear regression. Values from the original models and the literature (prior information) were used to specify parameter values that could not be estimated by nonlinear regression.

Estimation of Model Parameters

Ten model parameters were evaluated with parameter estimation: two for recharge (R_VF, R_TILL), three for horizontal hydraulic conductivity (K_MULT, K_TILL, K_ROCK), three for vertical anisotropy (KV_VF, KV_TILL, KV_ROCK), one for streambed hydraulic conductance (SB_MULT), and one for the maximum evapotranspiration from the water table in valley-fill deposits (ETM) (table 2). As mentioned previously, parameter GHB_MULT was not considered for parameter estimation, but the uncertainty in the parameter value was included in the uncertainty analysis. Parameter sensitivities, shown by their composite scaled sensitivities in figure 6, indicate whether groundwater-level and base-flow observations provided sufficient information to permit an estimate of a given parameter (Hill and Tiedeman, 2007). Parameters with higher sensitivities generally can be more precisely estimated than can parameters with lower sensitivities. Parameters with composite scaled sensitivities

Table 4. Summary of the long-term mean annual streamflow and base-flow analysis for partial-record streamgages, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

[USGS streamgage number: Locations shown on figure 4. mi^2, square mile; ft^3/s, cubic feet per second; in/yr, inches per year; Base-flow index is mean annual base flow divided by mean annual streamflow; --, not applicable]

USGS streamgage number	Streamgage name	Drainage area (mi^2)	Area of glacial stratified deposits (percent)	Mean annual streamflow				Mean annual base flow				Base-flow index (percent)	Continuous-record streamgage used in relation			
					90-percent confidence limits				90-percent confidence limits							
				Streamflow (ft^3/s)	Lower (ft^3/s)	Upper (ft^3/s)	[1]Streamflow (in/yr)	Base flow (ft^3/s)	Lower (ft^3/s)	Upper (ft^3/s)	[1]Base flow (in/yr)		Number of measurements used in relation	USGS streamgage number	Streamgage name	Correlation coefficient
01116800	Scrabbletown Brook near Davisville, RI	1.32	24	2.72	1.92	3.86	27.9	2.03	1.43	2.88	20.9	0.75	16	01115098	Peeptoad Brook near North Scituate, RI	0.96
													25	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.86
													25	01123000	Little River near Hanover, CT	0.86
01116820	Tributary to Hunt River near Davisville, RI	1.26	20	3.15	2.15	4.61	33.9	2.57	1.76	3.76	27.7	0.82	17	01115098	Peeptoad Brook near North Scituate, RI	0.86
													17	01117800	Wood River near Arcadia, RI	0.90
													26	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.85
01116875	Frenchtown Brook at State Route 2 near Davisville, RI	6.74	45	12.6	9.77	16.2	25.3	9.69	7.51	12.5	19.5	0.77	17	01115098	Peeptoad Brook near North Scituate, RI	0.94
													17	01117800	Wood River near Arcadia, RI	0.96
													17	01123000	Little River near Hanover, CT	0.94
01116895	Fry Brook near East Greenwich, RI	3.11	2	5.47	3.03	9.87	23.8	3.39	1.88	6.12	14.8	0.62	16	01115098	Peeptoad Brook near North Scituate, RI	0.92
													16	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.90
													16	01123000	Little River near Hanover, CT	0.86
01116980	Sandhill Brook near East Greenwich, RI	3.53	[2]96	7.92	4.44	14.1	30.4	6.19	3.47	11.0	23.8	0.78	17	01118000	Wood River at Hope Valley, RI	0.71
													17	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.81

Table 4. Summary of the long-term mean annual streamflow and base-flow analysis for partial-record streamgages, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.—Continued

[USGS streamgage number: Locations shown on figure 4. mi², square mile; ft³/s, cubic feet per second; in/yr, inches per year; Base-flow index is mean annual base flow divided by mean annual streamflow; --, not applicable]

USGS streamgage number	Streamgage name	Drainage area (mi²)	Area of glacial stratified deposits (percent)	Mean annual streamflow: Streamflow (ft³/s)	Mean annual streamflow, 90-percent confidence limits: Lower (ft³/s)	Mean annual streamflow, 90-percent confidence limits: Upper (ft³/s)	[1]Streamflow (in/yr)	Mean annual base flow: Base flow (ft³/s)	Mean annual base flow, 90-percent confidence limits: Lower (ft³/s)	Mean annual base flow, 90-percent confidence limits: Upper (ft³/s)	[1]Base flow (in/yr)	Base-flow index (percent)	Continuous-record streamgage used in relation: Number of measurements used in relation	USGS streamgage number	Streamgage name	Correlation coefficient
01117050	Cocumcussoc Brook near Wickford, RI	1.52	[2]58	3.42	2.33	5.01	30.5	2.74	1.87	4.02	24.4	0.80	17	01115098	Peeptoad Brook near North Scituate, RI	0.88
													17	01117800	Wood River near Arcadia, RI	0.88
													17	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.90
01117100	Annaquatucket River at Belleville, RI	6.44	[2]90	20.5	16.0	26.3	--	18.9	14.7	24.3	--	0.92	1,154	01117500	Pawcatuck River at Wood River Junction, RI	0.71
													1,154	01118000	Wood River at Hope Valley, RI	0.72
													1,154	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.71
01117200	Mattatuxet River near Saunderstown, RI	4.81	[2]62	11.1	7.07	17.4	--	9.33	5.94	14.7	--	0.84	17	01117468	Beaver River near Usquepaug, RI	0.71
													24	01118300	Pendleton Hill Brook near Clarks Falls, CT	0.67
													24	01123000	Little River near Hanover, CT	0.61

[1]Assumes drainage area is approximately the same as groundwater contributing area. Not applicable to Annaquatucket and Mattatuxet River basins.

[2]Includes morainal deposits.

Table 5. Streamflow and drainage-area characteristics of selected continuous-record streamgages in Rhode Island and Connecticut.

[USGS streamgage number: Locations shown on figure 1. mi^2, square mile; ft^3/s, cubic feet per second; in/yr, inches per year. Base-flow index is mean annual base flow divided by mean annual streamflow]

USGS streamgage number	Streamgage name	Drainage area (mi^2)	Area of glacial stratified deposits (percent)	Period of record analyzed	Mean annual streamflow (ft^3/s)	Mean annual streamflow (in/yr)	Mean annual base flow PART (ft^3/s)	Mean annual base flow PART (in/yr)	Base-flow index (percent)
01115098	Peeptoad Brook near North Scituate, RI	4.96	24	1995–07	10.4	28.5	7.79	21.3	75
01117000	Hunt River near East Greenwich, RI	22.9	52	1941–07	47.3	28.1	38	22.5	80
01117468	Beaver River near Usquepaug, RI	9.18	25	1975–2007	21.4	31.6	19.0	28.1	89
01117500	Pawcatuck River at Wood River Junction, RI	99.3	47	1941–07	197.0	27.0	177.0	24.2	90
01117800	Wood River near Arcadia, RI	35.2	23	1965–80; 1983–07	76.9	29.7	67.0	25.9	87
01118000	Wood River at Hope Valley, RI	73.5	26	1942–07	156.7	29.0	133.2	24.6	85
01118300	Pendleton Hill Brook near Clarks Falls, CT	3.99	8.4	1959–06	8.72	29.7	6.46	22.0	74
01123000	Little River near Hanover, CT	30.0	18	1952–06	56.2	25.4	40.0	18.1	71

Table 6. Comparison of observed and simulated base flows, Hunt–Annaquatucket–Pettaquamscutt Rvier Basins, Rhode Island.

[USGS streamgage number: Locations shown on figure 4. ft³/s, cubic feet per second; --, not applicable]

USGS streamgage number	Streamgage or observation name	Observed base flow (ft³/s)	Simulated base flow (ft³/s)	Observed minus simulated base flow (ft³/s)	Weighted residual
01116875	Frenchtown Brook at State Route 2 near Davisville, RI	-9.69	-10.0	0.31	0.65
01116895	Fry Brook near East Greenwich, RI	-3.39	-3.40	0.01	0.11
01116980	Sandhill Brook near East Greenwich, RI	-6.19	-5.36	-0.83	-2.62
--	Hunt River net gain	-14.1	-13.5	-0.60	-0.52
01117050	Cocumcussoc Brook near Wickford, RI	-2.74	-3.47	0.73	5.24
01117100	Annaquatucket River at Belleville, RI	-18.9	-9.39	-9.51	-9.86
01117200	Mattatuxet River near Saunderstown, RI	-9.33	-8.69	-0.64	-1.35

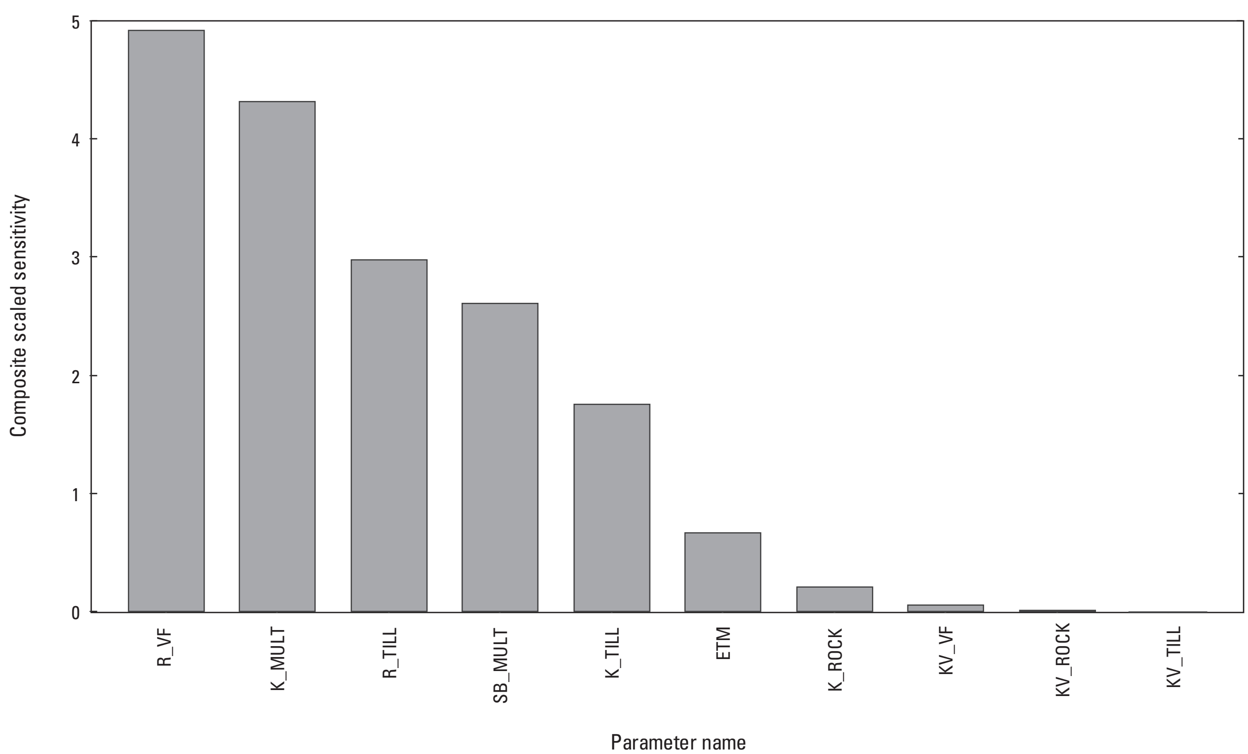


Figure 6. Composite scaled sensitivities for model parameters (parameter information is provided on table 2), Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

that are about two orders of magnitude lower than that of the parameter with the highest value, or those with composite scaled sensitivities less than one, indicate that nonlinear regression may not be capable of estimating the parameter (Hill and Tiedeman, 2007).

Low sensitivities were associated with five parameters: KV_VF, KV_TILL, K_ROCK, KV_ROCK, and ETM (fig. 6). Including any of these five parameters in nonlinear regression resulted in the model not converging with reasonable values. Thus, these five parameters were either assigned values used in the original models or, if a parameter representing a geologic unit or hydrologic process was not in the original models, values from the literature. Both parameters representing the ratios of horizontal to vertical hydraulic conductivity of glacial deposits (KV_VF and KV_TILL) had low sensitivities. The original model value, 5 (table 2), for vertical anisotropy was used for KV_VF. This value was assigned based on stratified deposits of sand and gravel in Rhode Island and Cape Cod, Massachusetts (Dickerman and others, 1990; Masterson and Barlow, 1997; Barlow, 1997). Vertical anisotropy values in these referenced studies ranged from 1:1 to 50:1, with the highest values for fine sand and silt. Melvin and others (1992) summarized the hydraulic properties of till, which can be highly variable, from previous studies in southern New England. For till derived from crystalline bedrock, horizontal hydraulic conductivities ranged from 0.004 to 65 ft/d, and vertical hydraulic conductivities ranged from 0.013 to 96 ft/d. The present study used the same vertical anisotropic value of 5 for KV_TILL that was also used for KV_VF.

Low sensitivities were also associated with bedrock parameters K_ROCK and KV_ROCK. Hydraulic conductivity of crystalline bedrock is generally low. Analysis of specific-capacity data from bedrock wells in eastern Connecticut indicated an average hydraulic conductivity of 0.5 ft/d (Randall and others, 1966). Lower values of 0.02 and 0.09 ft/d for crystalline bedrock in northern New Hampshire were determined through model calibration (Tiedeman and others, 1997). Parameter K_ROCK was specified a value of 0.1 ft/d, and KV_ROCK was specified a ratio of 1 (table 2).

The final low-sensitivity parameter was for maximum evapotranspiration (ETM) from the simulated water table in valley-fill deposits. The original model value of 21.0 in/yr was used (table 2); this value is equal to the estimated mean growing-season rate of free-water-surface evaporation from shallow lakes (Farnsworth and others, 1982).

The five remaining parameters representing recharge (R_VF and R_TILL), horizontal hydraulic conductivity (K_MULT and K_TILL), and streambed hydraulic conductance (SB_MULT) were not highly correlated (parameter correlation coefficients greater than 0.95). Thus these five parameters were estimated by nonlinear regression.

The quality of model calibration can be determined by analysis of the weighted residuals, both numerically and graphically, and by the reasonableness of optimal parameter values and their associated confidence intervals; Hill and Tiedeman (2007) describe this analysis in detail. Weighted residuals should be randomly distributed and close to zero. The average weighted residual was -0.11 ft for all groundwater-level and all base-flow observations; it ranged from a minimum of -9.86 ft to a maximum of 10.45 ft. The sum of squared weighted residuals was 1,488 for the calibrated model. The calculated error variance (sum of squared weighted residuals divided by the difference between the number of observations and the number of parameters estimated by nonlinear regression) was 8.91, and the standard error of the regression (square root of the calculated error variance) was 2.98. Although these measures of the overall magnitude of the weighted residuals should, theoretically, equal 1, that is not commonly the case for groundwater models (Hill and Tiedeman, 2007).

Summary information concerning weighted residuals for groundwater-level observations is listed in table 3 for each head group category and for all heads combined. The average weighted head residual for all groundwater levels was close to zero (-0.07 ft), ranging from a minimum of -9.26 ft to a maximum of 10.45 ft. Ninety-four percent of all weighted head residuals were within two standard errors of the regression. Group A, the highest weighted head group in the regression, had the largest average weighted residual and the largest standard deviation, but 91 percent of these weighted residuals were within two standard errors of the regression.

A comparison of observed and simulated base flow is listed in table 6. Values for six of seven simulated base flows compared well with the observed values: five of seven weighted residuals are within one standard error of the regression, and six of seven are within two standard errors of the regression. For the observation of Annaquatucket River streamgage (01117100) at Belleville, R.I., however, simulated base flow is one-half the observed flow. This partial-record site is downstream of two large ponds, and its watershed does not coincide with its groundwater-contributing area, making it difficult for streamflow data to be well correlated with a similar long-term streamgage.

Weighted residuals for base flows and for groundwater levels are shown graphically in figure 7, which symbolizes the weighted head residuals by the category of their groundwater-level observation group. A comparison of weighted observed values and weighted simulated values (fig. 7A) indicated a good agreement; the correlation between them was 0.99. Figure 7B shows that weighted residuals are generally randomly distributed around zero for all weighted simulated values and that most weighted residuals are within two standard errors of the regression.

The spatial distribution of weighted head residuals is shown in figure 8. The spatial distribution of weighted base-flow residuals is not shown in the figure, but they are distributed randomly throughout the model. The spatial distribution of weighted head residuals is generally random in most areas of the model, except for clusters of weighted residuals that are either all positive or all negative in two areas.

The central part of the model along Sandhill Brook and the headwaters of Cocumcussoc Brook has a cluster of negative head residuals (simulated water level greater than

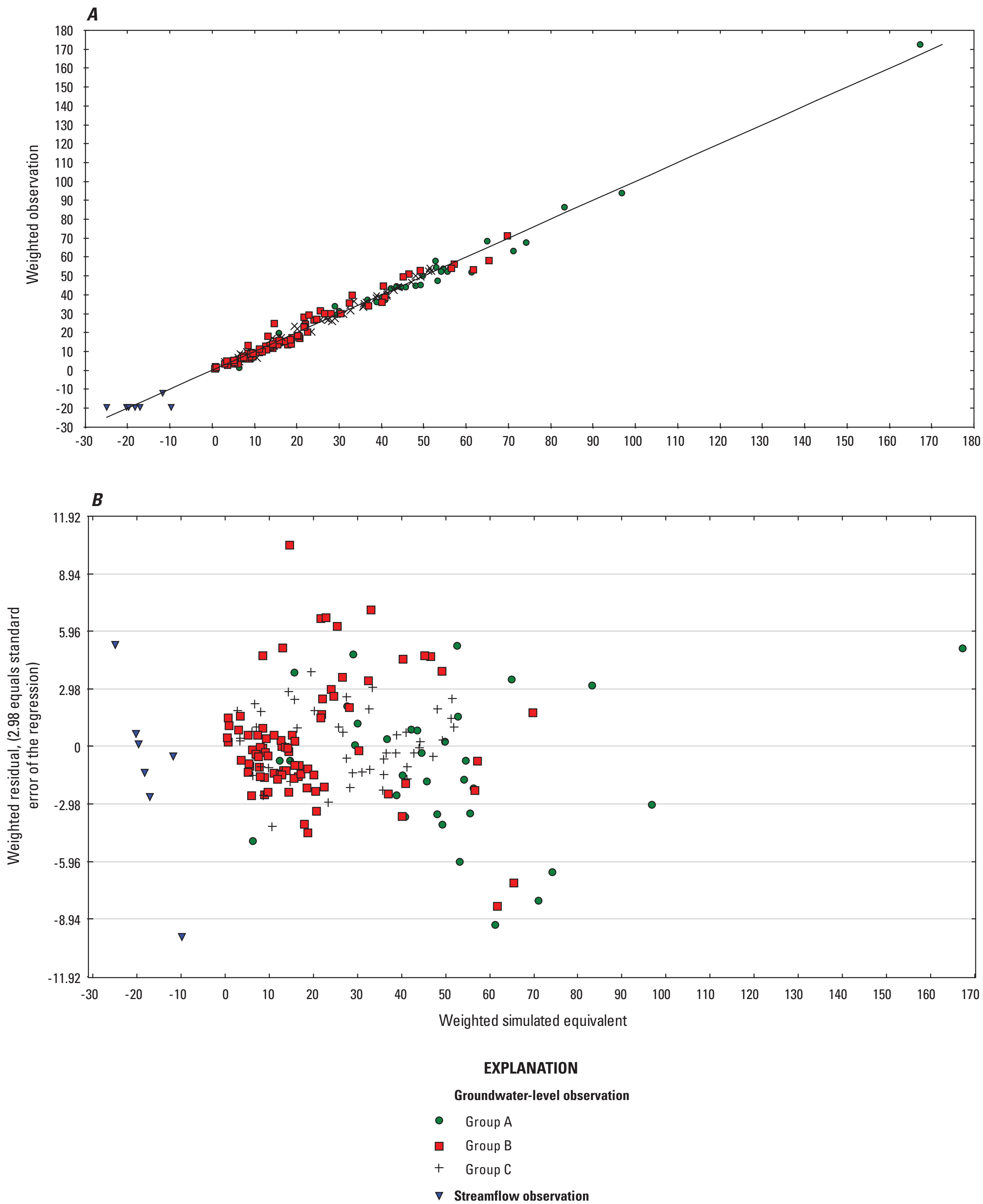


Figure 7. Relation of *A,* weighted observation to weighted simulated equivalent, and of *B,* weighted residual to weighted simulated equivalent, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

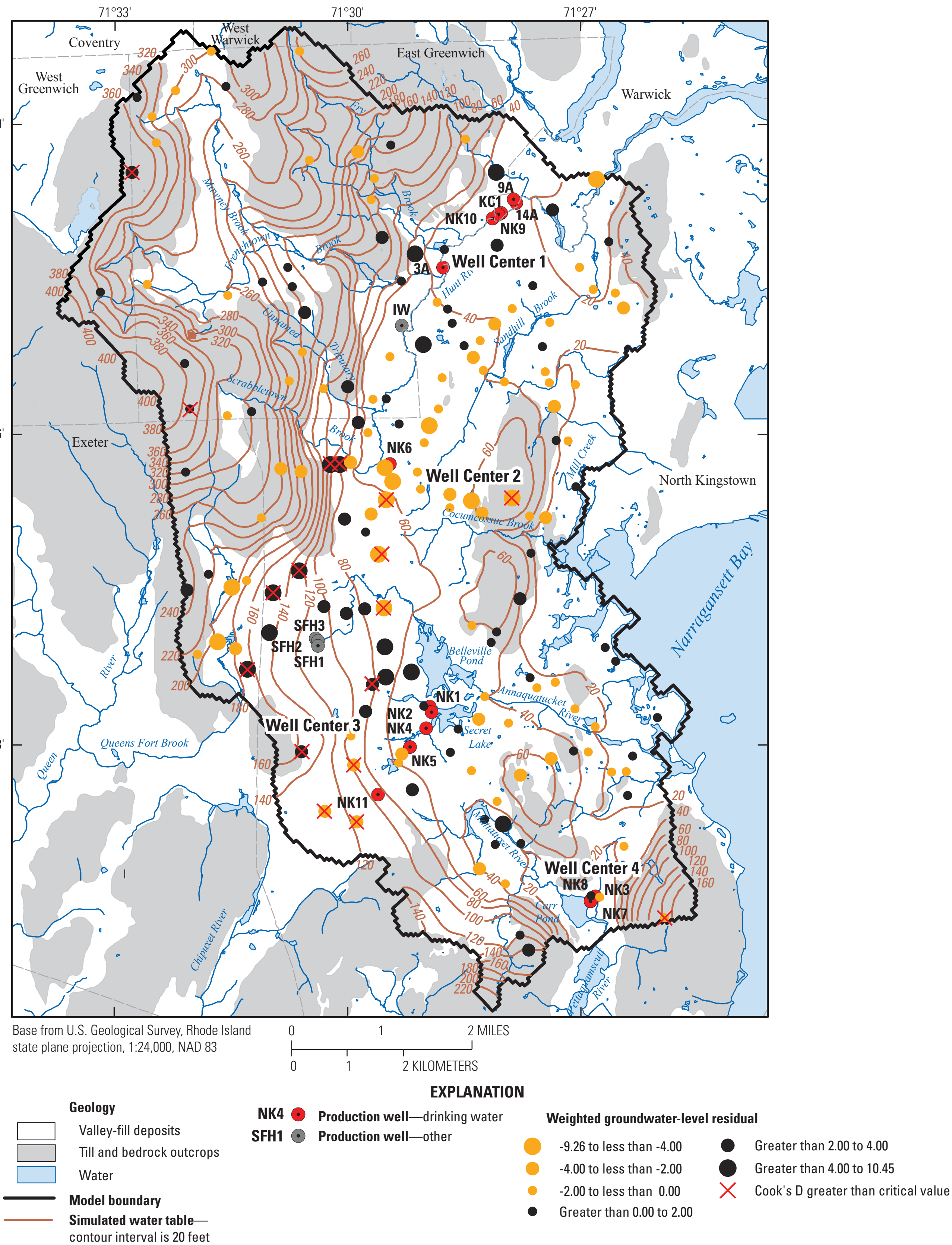


Figure 8. Spatial distribution of weighted residuals and simulated water-table contours for calibrated, steady-state conditions, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

observed water level), some of which are the largest in the model. Areas of the model between Sandhill Brook and Hunt River have thin surficial deposits, including some exposed till deposits and bedrock in the central part of the valley. The lower half of the area drained by Sandhill Brook contains high-density commercial and residential development where impervious surfaces may reduce recharge. The headwater area of Sandhill and Cocumcussoc Brooks consists of extensive wetlands. This complex surface-water feature is difficult to simulate accurately in a groundwater-flow model. In some models, wetlands in Rhode Island have been simulated with a high hydraulic conductivity to flatten the hydraulic gradient across the wetland (Dickerman and others, 1997; Friesz, 2004; Masterson and others, 2007). Finally, stream reaches are more extensive in these wetlands based on a comparison of 1:5,000 stream extent to the 1:24,000 used in the model; some of these stream reaches drain areas near the highest negative weighted residuals.

The southwest part of the model, roughly between Queens Fort Brook and Belleville Pond, has a cluster of positive head residuals (simulated water level less than observed water level). Although these observations are not in the immediate vicinity of the State fish hatchery wells, most were collected when pumping rates at these wells were less than that used in the calibrated model. These observations are also in a complex area of the model where loss from Queens Fort Brook flows toward the headwater reaches of Annaquatucket River.

These two areas indicate that the weighted residuals are not entirely randomly distributed and that there is some bias in model fit. Simplification of the aquifer and processes in these areas of the model may be the source of this model bias.

Insights into observations most important to the parameter estimates were calculated from influence statistics Cook’s D and DFBetas using the RESAN-2000 program (Hill and others, 2000). Observations with the most overall influence in the regression and on the resulting set of estimated parameters were determined from Cook’s D statistics. Twenty-three observations (6 base flows and 17 water levels), or 13 percent of the observations, had a Cook’s D value greater than the critical value of 0.023 (4 divided by the number of observations). Six of seven base-flow observations were considered influential to the set of estimated parameter values, thus indicating the importance of streamflow in this model calibration. In addition, two of the seven base-flow observations (Frenchtown Brook, 01116875, and Cocumcussoc Brook, 01117050) (fig. 4) were the only observations important to all of the individual estimated parameters, according to DFBetas statistics. Although the base-flow observations important to the set of estimated parameters are distributed throughout the model, head observations that dominate the regression are mostly in the southern half of the model (fig. 8). In addition, some of these head observations have relatively large residuals, both negative and positive. These observations with large residuals are in the headwaters of several streams previously mentioned and near the valley-upland contact in areas with steep hydraulic gradients.

Optimal values for the five recharge and hydraulic parameters (R_VF, R_TILL, K_TILL, K_MULT, SB_MULT) that were estimated using nonlinear regression are within a plausible range of values (table 2). Parameter R_VF, which represents precipitation recharge that reaches the water table, was 27.3 in/yr. A mean annual recharge rate of 25.5 in/yr (Barlow and Ostiguy, 2007) was calculated by analysis of streamflow records of the Hunt River streamgage (01117000) during 1941–03 using the computer program RORA (Rutledge, 1993). This recharge rate represents an average over the entire Hunt River Basin, including areas of stratified deposits, till, wetlands, surface-water bodies, and a variety of land uses. Conceptually, recharge rates for stratified deposits is higher than this basin-wide average rate and lower for less permeable till. Effective recharge rate to till deposits, represented by R_TILL, was 18.7 in/yr, close to mean annual base flow of 22.5 in/yr estimated by PART from the streamflow records at the Hunt River streamgage (table 5). This effective recharge rate for till was, as expected, less than this basin-wide average rate. The optimal value for R_TILL was also within the range of effective recharge rates from till-dominated basins in southern New England, as calculated by computerized hydrograph-separation techniques from long-term streamflow records: effective recharge rates ranged from 16 to 24 in/yr when mean annual runoff ranged from 27 to 31 in/yr (Bent, 1995, 1999; Friesz and Stone, 2007). The horizontal hydraulic conductivity of till (K_TILL), 16.2 ft/d, although within the range of plausible values summarized by Melvin and others (1992) and mentioned in the “Hydraulic Properties” section of this report, was higher than values of 1 to 10 ft/d that have been used in manually calibrated models (DeSimone, 2004; Masterson and others, 2007; Friesz and Stone, 2007) and values of 2 to 8 ft/d estimated from parameter estimation (Friesz, 2010) for till in Rhode Island and Massachusetts. Few data were available to define the contact between till and the bedrock surface in the uplands by Rosenshein and others (1968). Bedrock-surface altitude also varied substantially between data points in the uplands when assigning model-cell values. Optimal values for the remaining hydraulic parameters (0.81 for K_MULT and 2.6 for SB_MULT) were considered reasonable based on values reported by Rosenshein and others (1968) and because of the number of terms required to define streambed geometry.

The calibration results of the parameterized model and the original models are not directly comparable because of design changes and the difference in the number of observations and type of observations (base flow versus streamflow). However, the R_VF value of 27.3 in/yr in the parameterized model is relatively close to the manually calibrated value of 28.0 in/yr in the original models, as is the K_MULT of 81 percent of the original model values. SB_MULT, another parameter estimated by nonlinear regression and representing a process common to both models, was 2.6 times the values in the original models.

The uncertainty of the parameter estimate is indicated by the 95-percent linear confidence interval for each optimal value (table 2). For these linear confidence intervals to be valid, weighted residuals should be normally distributed and the model linear near the estimated optimal values (Hill and Tiedeman, 2007). If weighted residuals are independent and normally distributed, they plot on an approximately straight line on a normal probability graph (fig. 9). The correlation between weighted residuals and the normal order statistics for the calibrated model was 0.963. This value is near but less than the critical value for 172 observations, 0.984, at the 5-percent significance level. The degree of model linearity can be quantified using the modified Beale's measure, calculated with the BEALE-2000 program (Hill and others, 2000). The model is considered effectively linear if the modified Beale's measure is less than 0.039 and nonlinear if it is greater than 0.44. The modified Beale's measure for the model was 0.46, indicating that the model is nonlinear. The confidence intervals listed in table 2 are thus approximate values.

The 95-percent confidence intervals for the parameter estimates are all within the ranges of reasonable values reported in the literature. A comparison of the relative precision of different parameter estimates can be made using the coefficient of variation (standard deviation of the estimated value divided by the optimal value) (table 2); a smaller coefficient of variance indicates a more precisely estimated value for the parameter. The coefficient of variations ranged from 0.09 to 0.23. Recharge parameter R_VF and horizontal hydraulic conductivity parameter K_MULT, with coefficient of variation values of 0.1 or less, were the most precisely estimated, whereas the horizontal hydraulic conductivity parameter K_TILL, with a coefficient of variation of 0.23, was the least precisely estimated. The order of the most to least precisely estimated parameter values follows the same order as that of the parameter sensitivities (fig. 6) because of the information provided by the observations in the regression.

The analysis of the weighted residuals and optimal parameter values indicated that the groundwater model is acceptable for the purposes of the study. Optimal parameter values are realistic, and their confidence intervals include reasonable values. Although the spatial distribution of weighted head residuals was not entirely random, model-fit statistics indicated that simulated values for groundwater levels and for base flow are generally close to observed values.

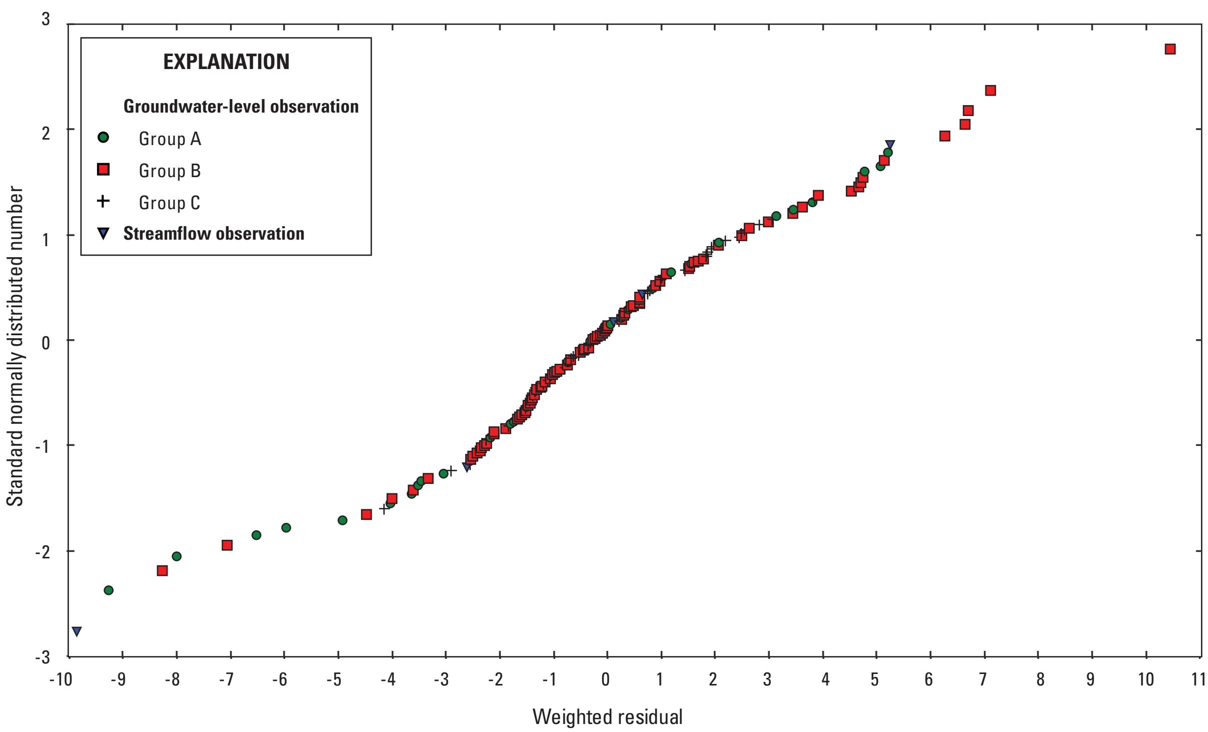


Figure 9. Normal probability of the weighted residuals, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

Simulated Water Table and Water Budget

The altitude and configuration of the simulated water table for the calibrated model, shown in figure 8 at 20-ft contour intervals, are consistent with the conceptual model of groundwater flow in the study area and with regional groundwater maps by Dickerman and Barlow (1997) and Rosenshein and others (1962), as well as with groundwater maps on a quadrangle scale by Allen and others (1959), Hahn (1959), and Johnson and Marks (1959). Groundwater generally flows from topographically high areas and discharges to streams and surface-water bodies. In the uplands, the simulated water table approximately parallels the land surface, and simulated groundwater divides generally coincide with watershed divides. In some upland areas, however, especially in the Mattatuxet River Basin, simulated groundwater levels are above the land surface but the direction of groundwater flow is realistic. The water-table gradient is steepest in the till uplands and in valley-fill deposits near the contact and then it flattens in the more transmissive valley-fill areas.

The simulated groundwater budget for the calibrated modeled area, summarized in table 7, indicated that direct precipitation recharge accounts for most of the total inflow (77 percent). Streamflow loss accounts for most of the remaining inflow (21 percent); most of this streamflow loss occurs from tributaries downstream of the upland-valley contact, which are near areas of abrupt changes in transmissivity, and at the downstream ends of ponds and lakes. Of the total inflow, 80 percent of the groundwater discharges to streams, and another 5 percent discharges directly to Narragansett Bay. Evapotranspiration from the water table and well withdrawals constitute about 8 percent each of the outflow. About one-half of the total well withdrawals are either returned to the modeled area by wastewater return flow or, in the case of the fish hatchery wells, to the headwaters of the Annaquatucket River; the remaining well withdrawals are exported from the modeled area.

In addition to the simulated groundwater budget for the whole model, water budgets for four major units of the modeled area (table 7) provide further insight. To simplify the water-budget calculations, the stratified deposits in the uplands west of Hunt River are included in the upland till category. Recharge is the dominant inflow to the uplands (87 percent). Groundwater discharge in the uplands is mostly to streams (52 percent), but a significant amount is to valley-fill deposits (38 percent). For the valley-fill deposits, recharge is also the main source of inflow (60 percent), although it is less of a percentage of the total inflow compared to that in the uplands. Upland sources, either directly or indirectly, also contribute a significant amount of inflow to valley-fill deposits (about one-third). These upland sources include shallow, lateral groundwater flow from till (15 percent); deep, vertical flow from bedrock (2 percent); and a large percentage of the streamflow loss (17 percent) to the valley-fill deposits.

Simulation of Areas Contributing Recharge and Prediction Uncertainty Analysis

Calibration of the HAP groundwater-flow model by inverse modeling using nonlinear regression provided an optimal set of parameter values. This optimal parameter set was estimated by minimizing the weighted residuals between the observation dataset (165 groundwater levels and 7 base flows) and simulated values. A predicted area contributing recharge to a production well in the HAP based on this optimal parameter set provides a single, most likely contributing area (deterministic contributing area). However, the parameter values were estimated with different levels of uncertainty; this uncertainty in the optimal values was based on the information that the observation dataset provided on the parameters. Parameter uncertainty and its associated effects on model predictions (spatial variability of the simulated contributing area to a well) can be evaluated by a stochastic Monte Carlo analysis. The parameter variance–covariance matrix from nonlinear regression can be used to create plausible parameter sets for the Monte Carlo analysis (Starn and others, 2000). The parameter variance–covariance matrix incorporates the uncertainty of the parameter estimates and the correlation among parameters from the calibrated model. The Monte Carlo analysis was done by replacing the parameter set in the calibrated model by a plausible parameter set multiple times. The probability of a particular location being in the contributing area to a production well was computed from these multiple model simulations.

A Monte Carlo analysis of this type whereby summary statistics from nonlinear regression were used to create parameter sets was described and applied by Starn and others (2000) for a well field screened in valley-fill deposits in a small modeled area. Lindsey (2005) applied the method for contributing areas to bedrock wells for those model parameters that could be estimated by the observations. The uncertainty analysis was also applied to wells in wetland and in coastal settings (Friesz, 2010). Starn and others (2010) compared the uncertainty of contributing areas to a well, depending on whether a model that used observations of only groundwater levels and streamflows or whether the model also included observations of atmospheric tracers. In both these referenced 2010 reports, prior information on model parameters was used to incorporate inestimable parameters into the uncertainty analysis, and a model-fit statistic was used to assess parameter sets so that prediction uncertainty would not be overestimated.

The present study delineated the areas contributing recharge for the 15 production wells that supply drinking water based on their 2004–08 average withdrawal rate and their maximum pumping rate (table 1). The total average annual withdrawal rate for the 15 production wells that supply drinking water increased from 2,587 gallons per minute (gal/min) (5.8 ft^3/s) for 1996 to 3,437 gal/min (7.7 ft^3/s) for

Table 7. Simulated steady-state average annual hydrologic budget for modeled area and for major units in the model, Hunt–Annaqutucket–Pettaquamscutt River Basins, Rhode Island.

[ft^3/s, cubic feet per second; Inflow and outflow may not equal and percent may not sum to 100 percent because of budget error and rounding]

Hydrologic budget component	Flow rate (ft^3/s)	Percent
Modeled area		
Inflow		
Recharge from precipitation	82.4	77
Recharge from wastewater return	2.1	2
Streamflow loss	22.5	21
Groundwater divides	0.0	0
Total inflow	107.0	
Outflow		
Streamflow	85.2	80
Narragansett Bay	5.6	5
Groundwater evapotranspiration	8.2	8
Well withdrawal	8.1	8
Groundwater divides	0.0	0
Total outflow	107.1	
Upland deposits (mostly till)		
Inflow		
All recharge (precipitation and wastewater)	30.8	87
Streamflow loss	1.2	3
From valley-fill deposits	2.0	6
From bedrock	1.4	4
Total inflow	35.4	
Outflow		
Streamflow	18.4	52
Groundwater evapotranspiration	1.5	4
To valley-fill deposits	13.3	38
To bedrock	2.3	6
Total outflow	35.5	
Valley-fill deposits		
Inflow		
All recharge (precipitation and wastewater)	52.9	60
Streamflow loss	15.3	17
Groundwater divides	0.0	0
From till	13.3	15
From ponds	5.3	6
From bedrock	1.3	2
Total inflow	88.1	
Valley-fill deposits—Continued		
Outflow		
Streamflow	54.7	62
Narragansett Bay	5.6	6
Groundwater evapotranspiration	6.7	8
Well withdrawal	8.1	9
Groundwater divides	0.0	0
To till	2.0	2
To ponds	10.6	12
To bedrock	0.5	1
Total outflow	88.2	
Freshwater ponds		
Inflow		
Recharge from precipitation	0.8	5
Streamflow loss	6.0	34
From valley-fill deposits	10.6	61
Total inflow	17.4	
Outflow		
Streamflow	12.0	69
To valley-fill deposits	5.3	31
Total outflow	17.3	
Bedrock		
Inflow		
From valley-fill deposits	0.5	18
From till	2.3	82
Total inflow	2.8	
Outflow		
To valley-fill deposits	1.3	48
To till	1.4	52
Total outflow	2.7	

the 2004–08 period, although individual wells may show an increase or decrease in withdrawal rate. Because there were no withdrawals from NK10 in 1996 or in 2003, areas contributing recharge to this well were not simulated in the original models. For the 2004–08 period, withdrawals from this well averaged 377 gal/min. For NK11, the average pumping rate listed represented the average rate for 2009 because this well did not begin pumping until December 2008.

The total maximum pumping rate for the production wells that supply drinking water, 14,400 gal/min (32.1 ft^3/s), is 4.2 times greater than the 2004-08 average pumping rate of 3,437 (7.7 ft^3/s). The maximum pumping rate for the production wells represents the maximum-rated capacity of the well pumps. Maximum pumping rates used in the model simulation are not proposed, long-term (continuous) withdrawal rates. Instead, RIDOH considers the simulated areas contributing recharge at these maximum rates only when implementing land-use planning that protects the quality of the water that the production wells supply. Average withdrawals may change because of changes in water usage or changes in State policies. Areas contributing recharge for the maximum pumping rate therefore represent conservative, or larger, areas for land-use planning than if low pumping rates were to be used.

The rate of wastewater return was not changed from that used in the calibrated model for either of these pumping rate scenarios. Although 2004–08 average withdrawal rates were greater than those of the 1996 withdrawals, wastewater return was a small part of the total inflow to the groundwater system (table 7). Maximum pumping rates are much larger than the 1996 withdrawal rates, but these maximum rates are hypothetical, and the distribution of the wastewater return in the model and the percent exported from the model are unknown. Thus the area contributing recharge to the production wells, especially at the maximum pumping rates, represents a conservative or a larger area than if wastewater return had been modified.

Deterministic Areas Contributing Recharge

Simulated deterministic areas contributing recharge and groundwater traveltimes to the production wells were determined on the basis of the calibrated steady-state model, for simulated pumping conditions, and by use of the particle-tracking program MODPATH (Pollock, 1994). The particle-tracking program calculates groundwater-flow paths and traveltimes on the basis of the head distribution computed by the groundwater-flow simulation. Areas contributing recharge were delineated by forward tracking of particles from recharge locations to the discharging wells. Consistent with the original models, particles were stopped at weak sinks, which remove only a part of the water that flows into the model cell. Barlow and Dickerman (2001) reported that using either option (of being stopped by or passing through a weak sink) minimally affected the simulated contributing areas to the majority of wells. Four particles for each model cell were used, thus each particle represents a surface area of 100 by 100 ft.

Shapes of the individual areas contributing recharge are strongly affected by nearby pumping wells and hydrogeologic features for both pumping rate scenarios (figs. 10 and 11). Most simulated areas contributing recharge for both average and maximum withdrawal rates extend upgradient from the wells to upland till and to groundwater and topographical divides, some of which serve as model boundaries. Some areas contributing recharge include small, isolated areas remote from the well. In addition, some of the contributing areas do not overlie a well. At the average pumping rate, four wells (NK5, NK7, NK8, and 3A) are not overlain by their contributing area, and two wells (NK3 and NK11) are overlain by the contributing area to a downgradient well (fig. 10). At the maximum pumping rate, although all wells are covered by a contributing area, three wells (NK3, NK10, and KC1) have the contributing area to a nearby well overlying it (fig. 11). Wells are screened in either layer 2 or layer 3, and recharge travels along paths above and around the screened interval of the well.

RIDOH and RIDEM assess the vulnerability of drinking water to contamination by well center. The size of the areas contributing recharge for the four well centers for the average pumping rate ranged from 0.19 mi^2 for well center 4 to 1.12 mi^2 for well center 1 and covered a total area of 2.79 mi^2 in the model (table 8). The size of the areas contributing recharge for the maximum pumping rate ranged from 0.37 mi^2 for well center 4 to 3.53 mi^2 for well center 3 and covered a total area of 7.99 mi^2 in the model. Land cover in the areas contributing recharge to the well centers in the Hunt River Basin include substantial amount of urban land use (fig. 3). Land cover in the areas contributing recharge include agriculture and sand and gravel mining for the well center in the Annaquatucket River Basin, and, for the well center in the Pettaquamscutt River Basin, land cover is primarily undeveloped.

The size of the area contributing recharge to a well for a particular pumping rate is related to effective recharge rates from precipitation and, if applicable, from the quantity of water derived from other sources. These sources may include wastewater return flow and surface-water infiltration. Surface-water infiltration may be from natural leakage, induced by pumping of a well, or both. For example, for the average pumping rate, well NK6, with the fourth largest withdrawal rate (319 gal/min) had the largest area contributing recharge (0.43 mi^2), in part, because all of its water was derived from precipitation recharge and wastewater return flow. Three wells in well center 1 on the lower part of the Hunt River (NK9, NK10, and KC1), which have the three largest average pumping rates (377 to 563 gal/min), had smaller contributing areas than NK6 (0.24 to 0.39 mi^2) because part of their water was derived from the Hunt River, and for KC1, part of its water was also derived from an unnamed tributary to the Hunt River that drains the uplands. Well 3A, which has an average pumping rate (135 gal/min or 0.30 ft^3/s) greater than five other wells, had the smallest contributing area in the model. This

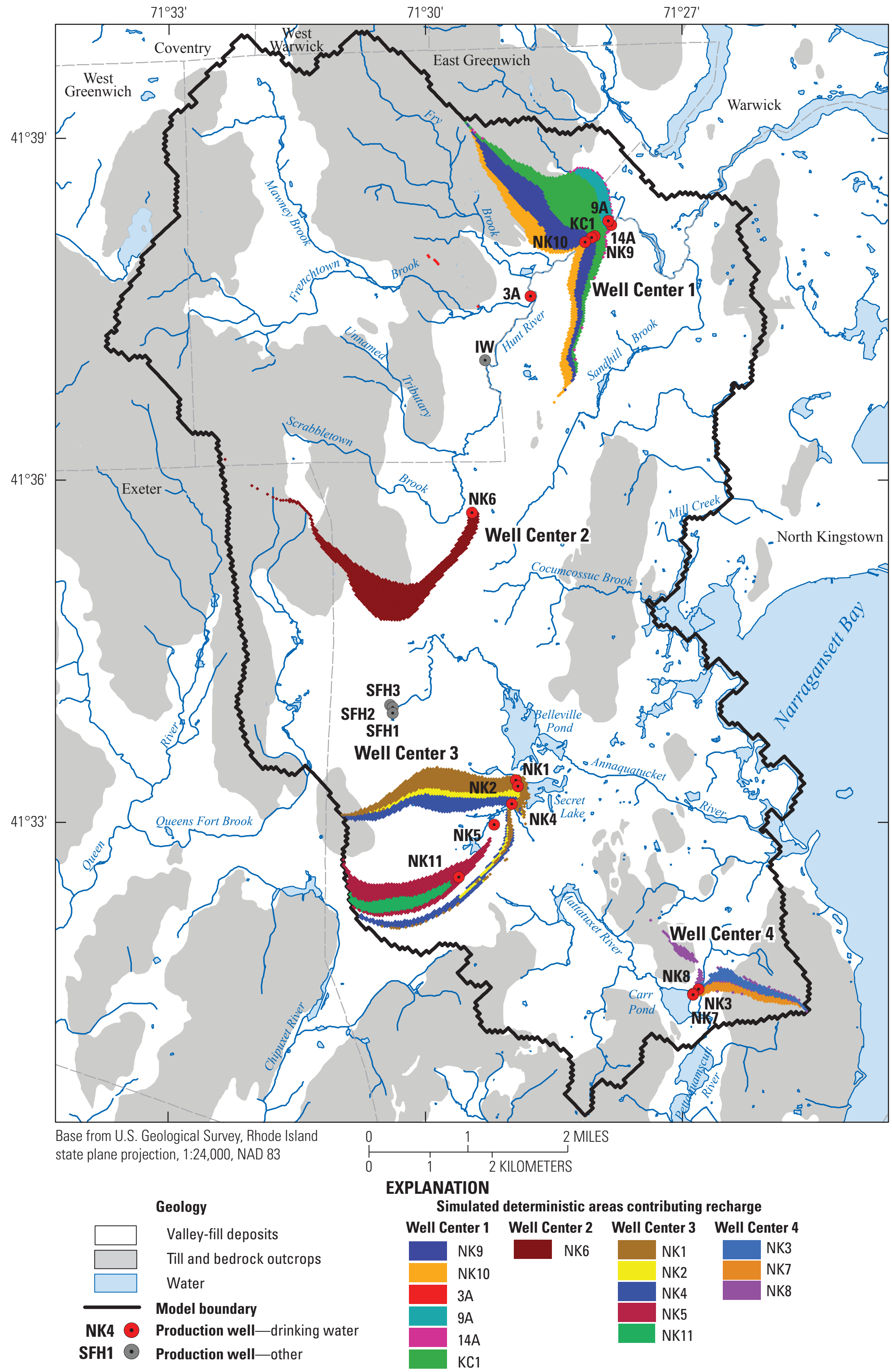


Figure 10. Simulated deterministic areas contributing recharge to the production wells for 2004–08 average pumping rates, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

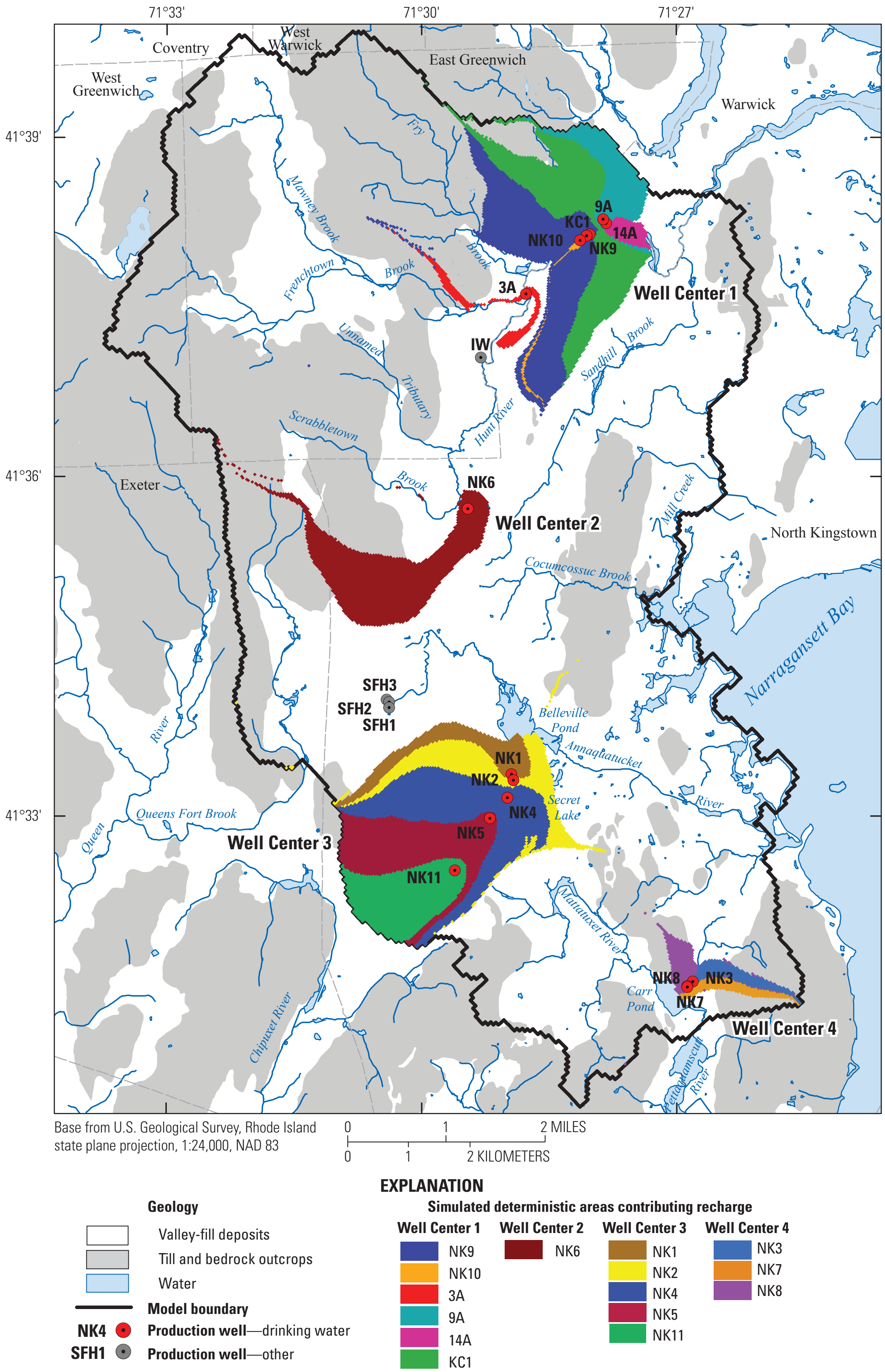


Figure 11. Simulated deterministic areas contributing recharge to the production wells for their maximum pumping rates, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

Table 8. Summary information concerning deterministic areas contributing recharge by well center, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

[gal/min, gallons per minute; mi², square mile; yr, year]

Well center	Total pumping rate (gal/min)		Size of area contributing recharge (mi²)		Groundwater traveltimes			
					For 2004–08 average rate		For maximum rate	
	2004–08 average rate	Maximum rate	For 2004–08 average rate	For maximum rate	Median (yr)	Percent 10 years or less	Median (yr)	Percent 10 years or less
1: (NK9, NK10, KC1, 3A, 9A, 14A)	1,858	8,500	1.12	3.03	4.7	78	5.0	78
2: (NK6)	319	900	0.43	1.06	4.8	90	4.6	87
3: (NK1, NK2, NK4, NK5, NK11)	1,025	4,200	1.05	3.53	5.3	70	4.3	78
4: (NK3, NK7, NK8)	235	800	0.19	0.37	4.1	90	2.9	93
All well centers	3,437	14,400	2.79	7.99	4.9	77	4.5	80

well is near the confluence of Frenchtown Brook and the Hunt River where most of its water was derived from streamflow loss from Frenchtown Brook downstream of the upland-valley contact. The location of an area contributing recharge to a well can be strongly affected by nearby pumping and associated withdrawal rates. For example, NK9 and NK10 have the two largest maximum pumping rates (2,000 gal/min each). NK9 had the largest contributing area in the model for the maximum rates, and its shape and sources of water were similar to its contributing area at the average rate. NK10, however, had the smallest contributing area in the model for maximum rates, even smaller than its contributing area at its lower pumping rate. Most of its pumped water was derived from the Hunt River (about 90 percent), and its remaining water was intercepted groundwater discharge in the valley-fill deposits southeast of the well. In contrast, at the average pumping rate, the well also intercepted precipitation recharge originating in the upland till northwest of the well.

At the maximum pumping rate, the area contributing recharge for the well centers expanded in all directions to capture enough water to balance the increased pumping rate, including downvalley from the wells. For two of the well centers (3 and 4) the contributing areas extended downvalley beneath ponds (Belleville and Carr), and additional areas extended beneath a lake (Secret Lake) and on the opposite side of these surface-water bodies to include small, isolated areas remote from the well centers (fig. 11). For NK2 this included an area in upland till and valley-fill deposits near the upland-valley contact northeast of Belleville Pond. Particle tracks showed that recharging water originating in this till and in valley-fill deposits travels along deep groundwater-flow paths in the valley-fill deposits and, under pumping conditions, passes beneath Belleville Pond to NK2. Recharging water between this contributing area and Belleville Pond travels along shallow and intermediate-depth flow paths before it discharges to the pond.

The areas contributing recharge to the Annaquatucket River Basin wells (well center 3) extend westward from the wells to the edge of the model defined by topographical divides in upland till and a groundwater divide in transmissive materials represented by the general-head boundary (figs. 10 and 11). In addition to recharge, wastewater return flow, and surface-water infiltration from the modeled area, the well center derived part of its water from induced groundwater from the Chipuxet River Basin. As previously mentioned in the "Boundary Conditions" section, simulated groundwater flow across this boundary was close to zero for the pumping rates used in the calibrated model (1996 average pumping rates). Although the average rates for 2004–08 for this well center increased by 338 gal/min from 687 to 1,025 gal/min, including withdrawals from the new well NK11 closest to the groundwater divide, the quantity of groundwater induced from the adjoining basin was minimal (about 1 percent each of NK11 and NK5 withdrawals). At the maximum pumping rate for the well center (4,200 gal/min), the area contributing recharge not only expanded to capture enough water to balance the pumping rate, but also induced enough groundwater to supply 4 percent each of NK4 and NK5 and 16 percent of NK11 withdrawals. This is a total of 190 gal/min or 0.43 ft^3/s. An additional 0.23 ft^3/s of groundwater was induced by pumping along the easternmost general-head boundary cells, but this groundwater discharged to the headwaters of the Mattatuxet River. The induced groundwater withdrawn by the wells would be equivalent to 0.21 mi^2 of land surface if the optimal value for precipitation recharge for valley-fill deposits (27.3 in/yr) was the only source of water and there was no groundwater evapotranspiration. This additional land-surface area in the Chipuxet River Basin, if adjacent to the divide, would include high-intensity land use that is mostly agricultural (cropland).

Simulated traveltime estimates from recharging locations to the production wells for the maximum pumping rate are shown in figure 12. Porosity values were specified in the model for MODPATH, but they affect only groundwater velocity and do not change the contributing areas to the wells. As in the original models, the valley-fill deposits were assigned a porosity of 0.35 based on values determined for similar deposits in southern Rhode Island (Allen and others, 1963) and Cape Cod, Massachusetts (Garabedian and others, 1991). Also, as in the original models, a porosity of 1 was specified for surface water. A porosity of 0.35 was specified for till deposits based on a range of values (0.22–0.50) determined for till deposits in southern Rhode Island (Allen and others, 1963) and southern New England (Melvin and others, 1992). For bedrock, a porosity of 0.02 was assigned based on values for crystalline rock summarized in Meinzer (1923).

Traveltimes generally depend on where recharge enters the aquifer in relation to the production well. Water that recharges the aquifer near the wells has the shortest traveltimes and youngest water, whereas water originating in the till uplands, and for well center 1, also along the groundwater divide between the Hunt River and Sandhill Brook, had the longest traveltimes and the oldest water. Traveltimes ranged from less than 1 year to more than 350 years at each well center. Median traveltimes for the well centers ranged from 2.9 to 5.0 years (table 8). A comparison of median traveltimes between maximum and average pumping rates indicated that increased pumping caused the percentage of young water to increase for three of the four well centers (table 8). Areas contributing water to the wells where traveltimes are 10 years or less ranged from 78 to 93 percent for the maximum pumping rate and from 70 to 90 percent for the average pumping rate for the four well centers. These percentages for traveltimes 10 years or less, and relatively short median traveltimes, indicate that the wells are vulnerable to contamination from activities on the land surface.

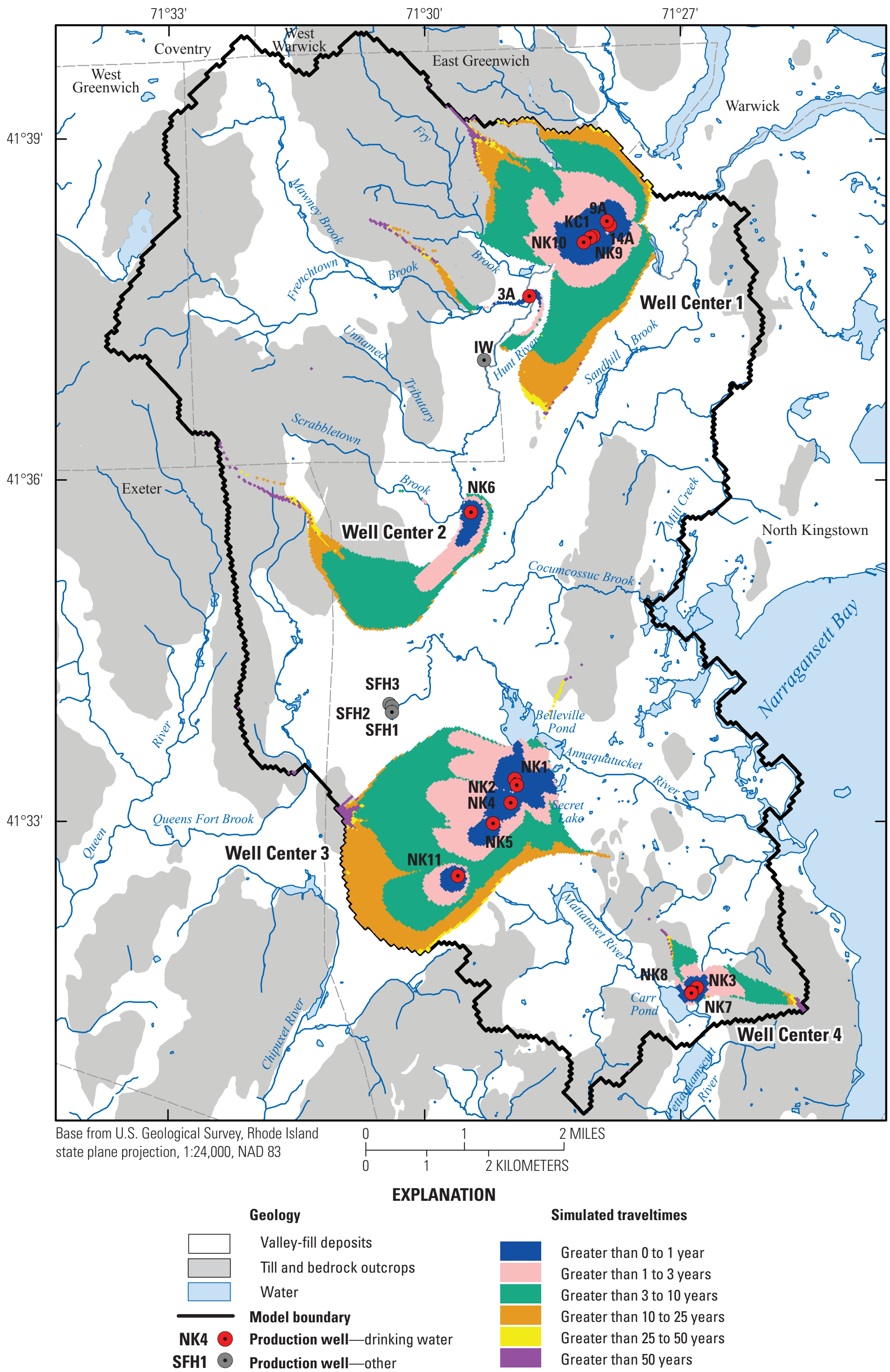


Figure 12. Simulated groundwater traveltimes to the production wells at their maximum pumping rates, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

Probabilistic Areas Contributing Recharge

A quantitative measure of the effects of parameter uncertainty on model predictions (the predicted contributing area) was done by a Monte Carlo analysis. A Monte Carlo analysis was used to obtain the probability of a recharge location being in the contributing area of a well center. The probability distribution is related to the information that the observation dataset provided on the estimated parameters, to prior information on specified parameters, and to the sensitivity of the simulated contributing area to the parameters. Hundreds of parameter sets generated from summary statistics of the calibrated model were used to run hundreds of model simulations in the Monte Carlo analysis. Because combinations of reasonable parameter values may result in unrealistic groundwater levels and streamflows, the parameter sets were evaluated using the pumping rates and associated observation dataset from the calibrated model. Those parameter sets that simulated realistic results were then used in Monte Carlo analyses for the two pumping scenarios.

Parameter values for the Monte Carlo analysis were created by the following equation (Starn and others, 2000, 2010):

$$b = z\sigma + \mu \qquad (1)$$

where

- b is a set of parameter values,
- z is a vector of normally distributed random numbers with a mean of 0 and standard deviation of 1,
- σ is the square root of the variance–covariance matrix calculated using Cholesky decomposition, and
- μ is a vector of optimal parameter values.

Parameter values that could not be estimated by nonlinear regression and thus were not included in the parameter variance–covariance matrix of the calibrated model may still be important for model predictions (for this study the size, shape, and location of the area contributing recharge to the production wells). The remaining three glacial hydraulic parameters (KV_VF, KV_TILL, and GHB_MULT) and the parameter representing the maximum groundwater evapotranspiration in the valley-fill (ETM) were, therefore, incorporated into the parameter variance–covariance matrix using the specified parameter values. Parameters representing bedrock and surface water were not changed in this analysis. Incorporating these four parameters into the parameter variance–covariance matrix, however, caused large unrealistic uncertainties around the specified parameter values because the information that observations provided was insufficient. Prior information on these parameters from the literature was used to constrain this uncertainty.

A prior weight was used for KV_VF so that the lower limit of the 95-percent confidence was not less than 1. However, because the specified value for KV_VF was 5, which is relatively close to 1, the upper limit of the 95-percent confidence interval was constrained to 25, which may not include most plausible values based on study results in Rhode Island and Cape Cod, Massachusetts, mentioned in the section, “Estimation of Model Parameters.” This 95-percent confidence interval, however, does include most of the values for coarse-grained stratified deposits listed in these studies. A prior weight was applied to KV_TILL in the same manner as that for KV_VF, and it too resulted in a 95-percent confidence interval of 1 to 25. Prior weight for ETM was ±35 percent around the specified value of 21 in/yr, which is the growing-season rate of free-water-surface evaporation. The 95-percent confidence interval (13.8 to 28.3 in/yr) includes the annual rate of free-water-surface evaporation (28 in/yr; Farnsworth and others, 1982), a possible maximum value of groundwater evapotranspiration. A prior weight was used for parameter GHB_MULT so that the 95-percent confidence interval ranged from 0.25 to 4, which is equivalent to a hydraulic conductivity of 12 ft/d (close to a value representative of very fine sand) and 200 ft/d (sand and gravel) around the specified value of 50 ft/d (fine sand). Conceptually, a higher hydraulic conductivity value for the general-head boundary would allow pumping wells to induce more groundwater from adjacent basins than would a lower hydraulic conductivity value. The more groundwater induced from an adjacent basin, the smaller the contributing area to a well in the active model area when compared to a model simulation in which no groundwater or less water was induced across the groundwater divide.

The addition of these four inestimable parameters (KV_VF, KV_TILL, ETM and GHB_MULT) into the parameter variance–covariance matrix incorporated into the Monte Carlo analysis all parameter uncertainty potentially important for model predictions. Parameter uncertainties are from the observation dataset, but also from prior information on parameters that the modeler provided.

Parameter sets created by equation 1 are shown in figure 13A. The hydraulic parameters were log-transformed in the model. The parameter sets have a lognormal or normal distribution around the optimal or specified parameter value; the spread of these data indicates the certainty with which each parameter was estimated, considering the available observations and prior information. Parameters incorporated into the variance–covariance matrix using prior information (KV_VF, KV_TILL, GHB_MULT, and ETM) have the least certainty, which would be expected because a prior weight was used so as to incorporate most plausible values. Parameters sets, however, with glacial anisotropies less than 1 (horizontal hydraulic conductivity less than vertical hydraulic conductivity) or greater than 50 were removed (conditioned) from the parameters sets because they were considered unrealistic for the aquifer as a whole.

For the Monte Carlo analysis, the model was first run with 500 parameter sets, and with the 1996 average pumping rates used in calibrating the model. (A Monte Carlo analysis with 600 parameter sets showed similar results for the probabilistic contributing area, indicating that 500 sets

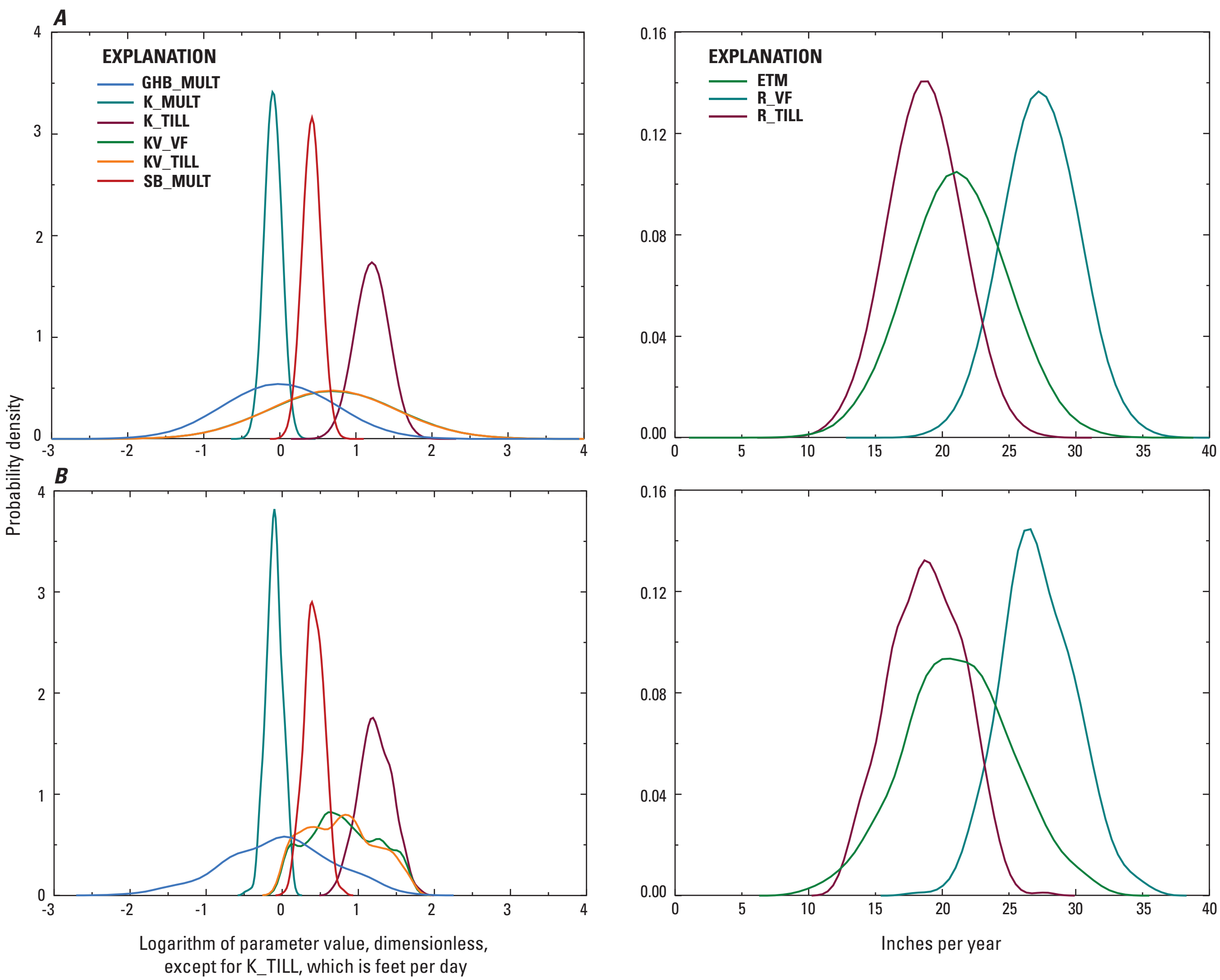


Figure 13. Model-parameter distributions *A,* before acceptance criteria and *B,* after acceptance criteria were applied to determine the probabilistic areas contributing recharge to the well centers, (parameter information is provided on table 2), Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

were sufficient to use in the analysis). The nine hydraulic, recharge, and groundwater evapotranspiration parameter values in each dataset replaced the corresponding parameter values in the calibrated model. The following three criteria for accepting a given parameter set were used: (1) the model converged, (2) the model mass balance was 2 percent or less, and (3) a model-fit statistic (calculated error variance) was less than a specified value. Of the 500 parameter sets run with MODFLOW, 425 sets (85 percent) converged, and of these, 420 sets (84 percent) had a water budget of 2 percent or less, thus meeting two of the three criteria.

The third acceptance criterion, the model-fit statistic, was used so that model-prediction uncertainty would not be overestimated by using a parameter set that produced unrealistic groundwater levels or streamflows compared to that for the calibrated model. The value used for this criterion, however, can be model dependent and subjective. For example, a calculated error variance of 20 or less was used for a Monte Carlo analysis of areas contributing recharge to wells in a wetland setting, for which the calibrated model had a calculated error variance of 5.3 (Friesz, 2010); this criterion did not remove any datasets. In contrast, Starn and others (2010) used a standard error of regression of 20 (calculated error variance of 400) for a comparable Monte Carlo analysis (prior information on selected parameters and the same types of observations) in a valley-fill setting, for which the calibrated model had a standard error of regression of 6.2 (calculated error variance of 38.4); this value for the criterion removed 26 percent of the datasets after the dataset met a water-budget criterion.

For this model application of the Monte Carlo analysis, the third criterion was a calculated error variance of 20. Most parameter sets had a calculated error variance of 15 or less, and 4 sets had greater than 20. Common to these four parameter sets was the fact that the K_MULT and K_TILL both had values greater or less than their 95-percent confidence intervals. All four K_TILL values were less than its confidence interval, and three K_MULT values were greater, and one less, than its confidence interval. Although the individual parameter values were considered reasonable based on model calibration, the combination of these parameter values produced a poor model fit. These four parameter sets were removed from the analysis, and therefore, 416 sets (83 percent) of the 500 parameter sets run with MODFLOW fit the three acceptance criteria. The distribution of parameters after the unrealistic glacial anisotropies were removed, and after the acceptance criteria were applied, are shown in figure 13B. Except for the glacial anisotropies, the distribution of parameters, although slightly altered from the original parameter sets, indicated a generally lognormal or normal distribution. Using a higher, yet plausible, specified value of 10 instead of 5 for the glacial anisotropies in the calibrated model would require less conditioning of these parameter sets and less altering of the distribution.

Monte Carlo analyses were then done by use of the parameter sets that fit the acceptance criteria for the 1996 average withdrawal rates but using the 2004–08 average withdrawal rates and the maximum withdrawal rates. The criteria for the Monte Carlo analyses that used these 416 parameter sets and these pumping rates were that the water budget be 2 percent or less for models that converged. For the 2004–08 average withdrawal rate, 402 parameter sets (97 percent) fit these criteria and thus were run with the particle-tracking program. For the maximum withdrawal rates, 408 parameter sets (98 percent) fit these criteria. The probability that a recharge location would be in the area contributing recharge to the production wells was determined by dividing the number of times a particle at a given location was captured by a well by the total number of accepted particle-tracking simulations; this probability was expressed as a percentage.

The probabilistic areas contributing recharge to the well centers at each center’s 2004–08 average pumping rate are shown in figure 14, and those at the maximum pumping rate are shown in figure 15. For this analysis, the probabilistic distribution is not by well but by well center. Probabilistic contributing areas to individual wells may overlap, even though the deterministic contributing areas do not overlap under a steady-state simulation. The total size of the probabilistic contributing area for each well center for both average and maximum pumping rates was larger than the deterministic contributing areas for the well center because of the effects of parameter uncertainty. This indicated that some areas not in the deterministic contributing area, including additional areas of urban and agricultural land use, may actually be in the contributing area. Generally, areas closest to the well centers with short traveltimes are associated with higher probabilities, whereas areas distant from the well centers with long traveltimes are associated with lower probabilities. In most cases, areas with high probabilities (greater than 50 percent) generally coincided with the deterministic contributing areas for this model.

For well center 1, the deterministic contributing area for the average pumping rate (fig. 10) for the five closely spaced wells adjacent to the lower Hunt River extended southward from the wells in a relatively narrow width to the groundwater divide between the Hunt River and Sandhill Brook, and it extended northwestward from the wells to include most of a till hillslope between two tributary streams that drains toward the valley. The probabilistic contributing area (fig. 14) indicated that additional areas southward toward the groundwater divide and northwestward in the till uplands, mostly east of Fry Brook, may be in the contributing area. Additional recharge originating in upland areas, mostly along groundwater divides between Fry and Frenchtown Brooks and between Frenchtown Brook and small tributaries to Hunt River, including Scrabbletown Brook, may be in the contributing area to the well center. These are areas associated with low probabilities. Two separate Monte Carlo analyses were done for this well center to provide additional insights (results not shown): one to determine the probabilistic contributing area to only the five wells along the lowest part of the lower Hunt River, and one to well 3A near the confluence with Frenchtown Brook and the Hunt River. Areas of low

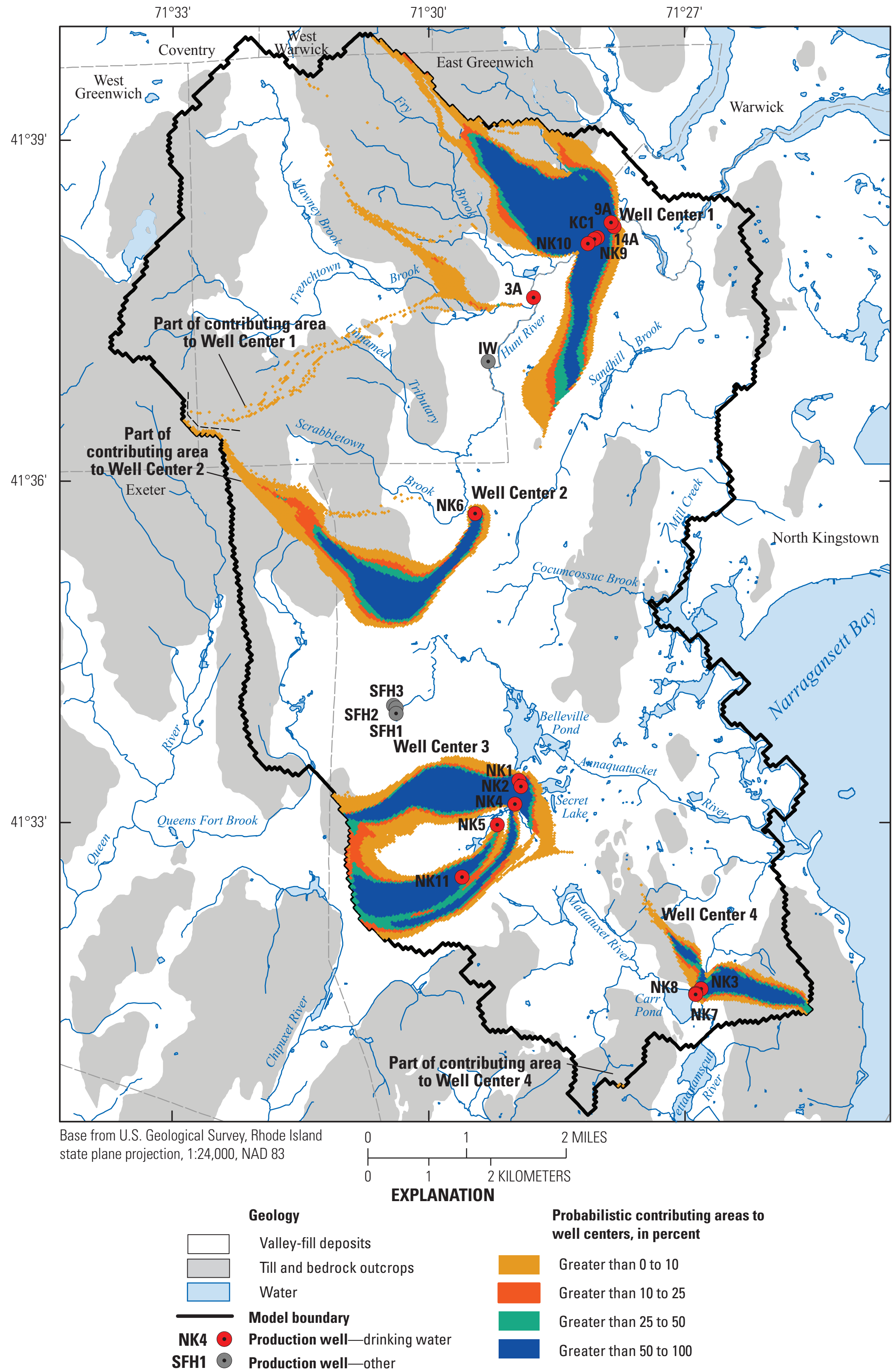


Figure 14. Simulated probabilistic areas contributing recharge to the well centers at their 2004–08 average pumping rates, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

probabilities along both upland groundwater divides can be either to the well field along the lower Hunt River or to well 3A. The probabilistic contributing area to well 3A did not include areas of high probabilities (greater than 50 percent), indicating that for most parameter sets used in the Monte Carlo analysis the well derived most of its water from stream infiltration. Areas of highest probabilities for this well (greater than 10 to 50 percent) are near the upland-valley contact and west of the well in the uplands; the deterministic contributing area for this well corresponded to these areas. In addition to areas west of the Hunt River, the probabilistic contributing area for well 3A also included an area of low probabilities south of the well near the groundwater divide between Hunt River and Sandhill Brook. As with the deterministic contributing area, the probabilistic contributing area did not overlie well 3A.

At the maximum pumping rate, the deterministic contributing area (fig. 11) to the five wells adjacent to the lower Hunt River expanded to include more valley-fill deposits south and east of the wells toward groundwater divides and to include more valley-fill deposits and till north and northwest of the wells toward topographical divides that serve as model boundaries. The deterministic contributing area to one of these five wells also included small areas in the till between Fry and Frenchtown Brooks. The deterministic contributing area for the maximum pumping rate for well 3A also extended upgradient on both sides of the Hunt River. The deterministic contributing areas generally corresponded to high probabilities (fig. 15A), including well 3A. Along margins of the probabilistic contributing area south and east of the well center and north and northwest of the five wells along the lower Hunt River, there is minimal spread in low probabilities because these areas are close to the wells and are constrained by hydrologic and model boundaries. Extensive areas associated with mostly low probabilities to both well 3A and to the wells along the lower Hunt River are along the groundwater divides in upland areas. The deterministic contributing areas in the uplands between Fry and Frenchtown Brooks are associated with mostly high probabilities for well 3A, but with probabilities mostly between 10 to 50 percent for the wells adjacent to the lower Hunt River.

For well center 2, the deterministic contributing area to NK6 for both average and maximum pumping rates extended southwestward from the well to the upland-valley contact and then northwestward in till uplands (figs. 10 and 11). The deterministic contributing area for the maximum pumping rate also included small, isolated areas west of the well adjacent to and beneath Scrabbletown Brook. The probabilistic contributing area to the well for both pumping rates had a large spread in low probabilities in the uplands compared to those for the valley-fill deposits (figs. 14 and 15B). The probabilistic contributing area for the average pumping rate included small, isolated areas west of the well near Scrabbletown Brook and adjacent uplands, areas that are associated with low probabilities. The probabilistic contributing areas for the maximum pumping rate included more areas west of the well adjacent to and beneath Scrabbletown Brook and toward the uplands; these areas are associated with a large range in probabilities. The deterministic contributing area in this area for the maximum pumping rate corresponded to an area of high probabilities (greater than 50 percent) and to probabilities greater than 25 to 50 percent. For both pumping rates, these are additional areas that may be in the contributing area to the well that also have relatively short groundwater-flow paths and traveltimes.

For well center 3, the deterministic contributing area for the average pumping rate extended westward and southwestward from the wells to the edge of the model in relatively narrow bands, and it also underlies part of Secret Lake (fig. 10). The probabilistic contributing area indicated that additional areas west and southwest of the wells along the margins of the deterministic contributing area may actually be in the contributing area to the well (fig. 15B). This included some recharge near NK4 and NK5 that discharged to small tributaries to Secret Lake in the deterministic model. In contrast to the deterministic model, the Monte Carlo analysis indicated that NK5 may be overlain by the contributing area to the well center (probabilities of 2 percent and less). Additional areas beneath Secret Lake may also be in the contributing area to the well center.

The deterministic contributing area for the maximum pumping rate included all of the area between the well center and the west edge of the model, beneath most of Secret Lake and part of Belleville Pond, and small areas on the opposite side of Belleville Pond and upland till in Queens Fort Brook Basin (fig. 11). Along margins of the probabilistic contributing area (fig. 15B), there is minimal spread in low probabilities because of hydrologic and model boundaries; Belleville Pond and induced groundwater from Chipuxet River Basin can be major sources of water to the well center, depending on the values of a parameter set used in the Monte Carlo simulation. Withdrawals by the fish hatchery wells may also constrain the extent of the contributing area northwest of the well center. On the side of Belleville Pond opposite the well center, probabilities decrease in a radially outward pattern; most of the deterministic contributing area on this side of the pond coincided with probabilities greater than 10 to 50 percent. Additional areas in Queens Fort Brook that are not in the deterministic contributing area are in the probabilistic contributing area. These are areas of mostly low probabilities; the deterministic contributing area in the till uplands corresponded with probabilities of 25 percent and less.

For well center 4, the deterministic contributing area for both pumping rates extended upgradient of the well center to upland till on both sides of a small valley (figs. 10 and 11). The deterministic contributing area for the maximum pumping rate also included a small area downvalley beneath Carr Pond and on the opposite side of the pond in the uplands. The probabilistic contributing area for both pumping rates (figs. 14 and 15B) indicated that additional areas upgradient of the well center in the valley and uplands may be in the contributing area, but the deterministic and probabilistic contributing areas also indicated that recharge discharges to headwater streams of

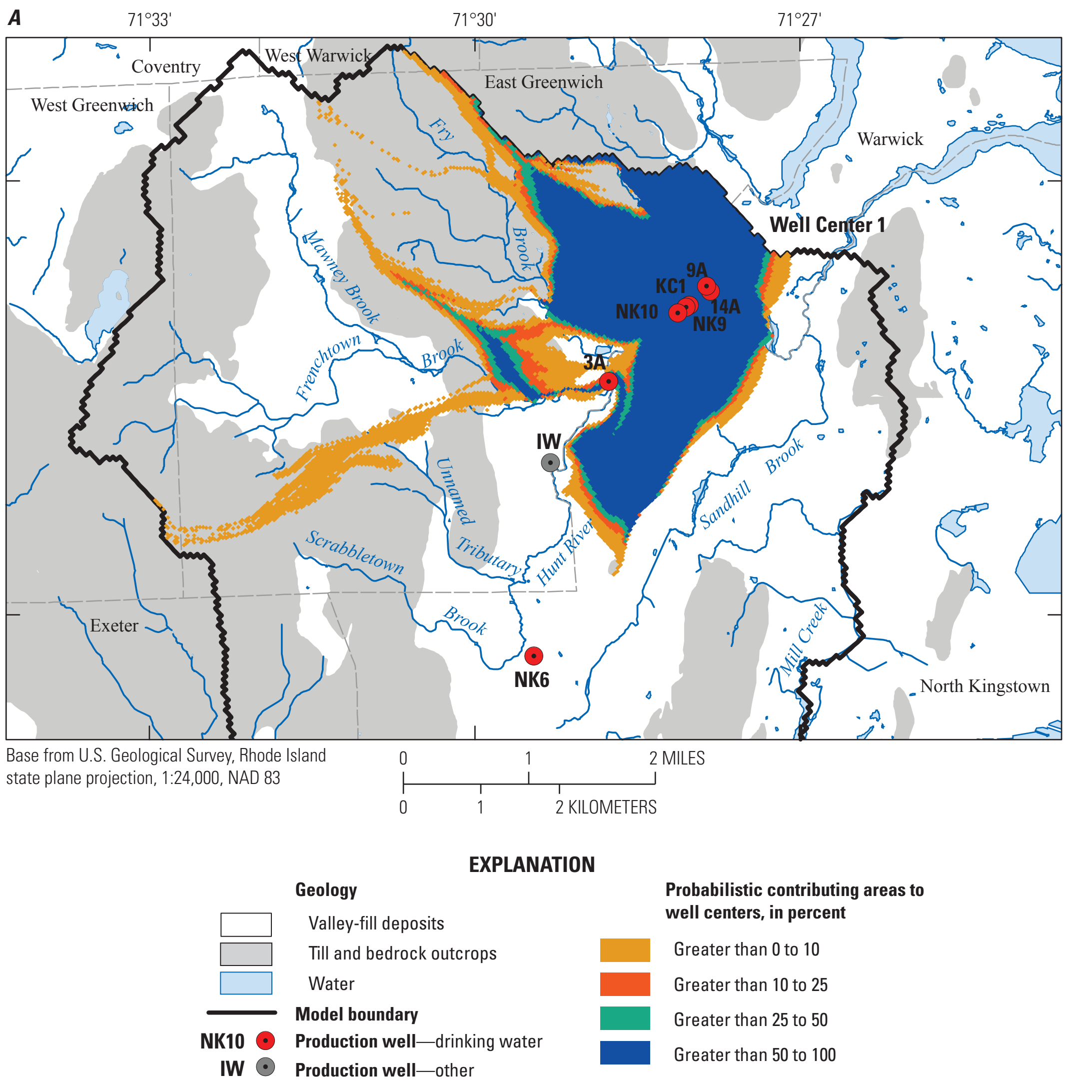


Figure 15. Simulated probabilistic areas contributing recharge to *A,* well center 1 and *B,* well centers 2, 3, and 4, at their maximum pumping rates, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.

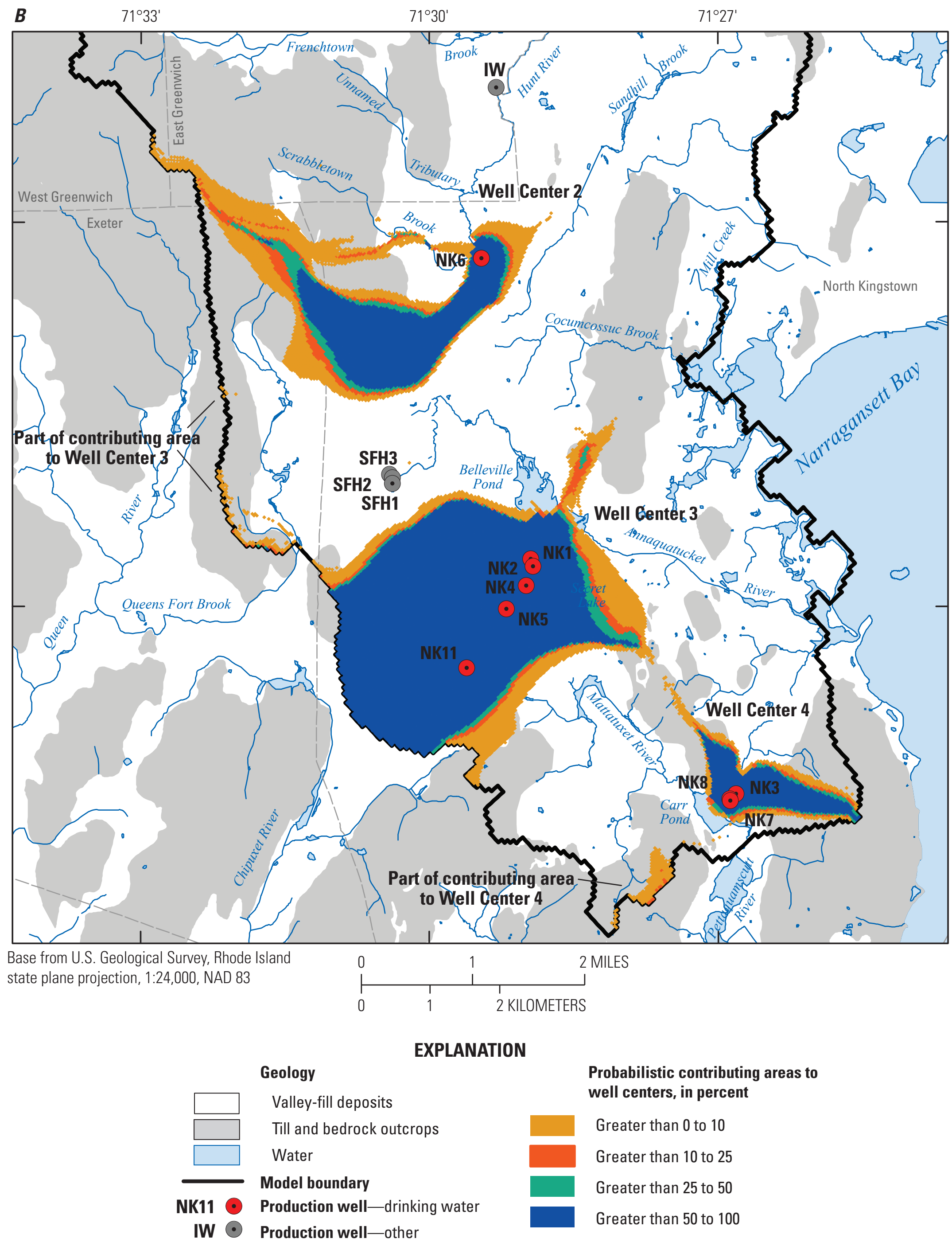


Figure 15. Simulated probabilistic areas contributing recharge to *A,* well center 1 and *B,* well centers 2, 3, and 4, at their maximum pumping rates, Hunt–Annaquatucket–Pettaquamscutt River Basins, Rhode Island.—Continued

this valley and uplands before it is available for infiltration as the stream flows near the well center. In addition to additional areas beneath Carr Pond for the maximum pumping rate, the probabilistic contributing area for both pumping rates included areas on the opposite side of the pond from the wells; these are areas with low probabilities. In the deterministic model for the average pumping rate, NK7 and NK8 were not covered by the contributing area for the well center but the probabilistic contributing area indicated, with probabilities greater than 25 to 50 percent, that this area may be in the contributing area.

Limitations of Model

The finite-difference numerical model of the HAP study area is a regional-scale simulation of groundwater flow, water levels, and the interaction between groundwater and surface water. Simplification included consolidating parameters that represented hydraulic properties and boundary conditions into homogenous units and assigning these parameters to groups of model grid cells sized 200 by 200 ft. Further, recharge rates and their distribution do not take into account impermeable surfaces in urban areas, which are likely to reduce or redistribute recharge in these areas. For these reasons, the model may not be appropriate for simulating local-scale results, but, for evaluating aquifer vulnerability to contamination at a regional scale, the model is useful.

Groundwater flow in upland till and bedrock was greatly simplified in the model. Although hydraulic conductivity values and recharge rates can be highly variable in till deposits, only three parameters represented hydraulic properties and recharge rates in these unsorted materials. For example, recharge rates may range from near zero in low-permeability tills in areas of steep topography to values approaching those for water available for recharge in sandy tills on moderate slopes. Thus the parameter values represented an average value for these deposits. Groundwater flow in bedrock represented the bulk flow in the regional system instead of the flow through bedrock fractures. Thus the model accounts for the overall movement of groundwater through the upland tills and bedrock before it discharges to the valley-fill deposits where the production wells are located.

The areas contributing recharge to the well center in the Annaquatucket River Basin extended to the edge of the model in valley-fill deposits that a groundwater divide defined and that a general-head boundary represented. Although the quantity of groundwater induced across this boundary from the adjacent basin was simulated for the average and maximum pumping rates, it would be necessary to extend the model into the adjacent basin in order to delineate the actual area that contributes water to the well center. Additional geologic and hydrologic data would be needed in order to expand the model and to improve the understanding of the geohydrology in this area.

Uncertainty in the simulated areas contributing recharge to the wells was based on the observation dataset and not from model design. Additional groundwater-level and base-flow observations and other types of field observations may help reduce the uncertainty about the extent of the simulated contributing area by increasing the precision of the parameter value estimates. Additional observations may also help to increase the number of parameters that could be estimated using nonlinear regression, thereby decreasing the need for using prior information from the literature. The resulting uncertainty analysis would then be based solely on objective model-calibration data. Also, the seven base-flow observations used in the calibration represented the net gain in base flow over relatively large segments of stream reaches upstream of the observation. To gain a better understanding of groundwater and surface-water interactions and of streambed conductance near the pumping wells requires observations that represent a net gain or loss in base flow over shorter stream reaches near the wells.

The simulated groundwater traveltimes were based on the calibrated model and uniform porosity for each of the lithologic units. An uncertainty analysis of groundwater traveltimes based on the spatial variability and the plausible range in porosity was beyond the scope of this study. However, in a steady-state model, porosity does not affect the location or size of the simulated area that contributes recharge to a well. Finally, traveltimes do not take into account traveltime in the unsaturated zone between the land surface and the water table.

Summary and Conclusions

The U.S. Geological Survey (USGS), in cooperation with the Rhode Island Department of Health, Office of Drinking Water Quality, began a 2-year investigation in 2009 to increase understanding of groundwater flow and of the areas contributing recharge to 15 production wells in the Hunt, Annaquatucket, and Pettaquamscutt (HAP) Rivers Basins in central Rhode Island. These large capacity production wells are operated by the town of North Kingstown, the Rhode Island Economic Development Corporation, and the Kent County Water Authority. A total average daily rate of 5.0 million gallons per day is withdrawn from these production wells, which are screened in coarse-grained valley-fill deposits of glacial origin.

The area contributing recharge to a well is defined as the surface area where water recharges the groundwater and then flows toward and discharges to the well. Areas contributing recharge to the production wells were determined on the basis of a numerical steady-state groundwater-flow model representing long-term average hydrologic conditions. The study modified an existing model that simulated groundwater flow in the valley-fill deposits in order to (1) simulate flow in the adjoining uplands and (2) represent boundary conditions and hydraulic properties as parameters for calibration by inverse modeling using nonlinear regression and for evaluating model-prediction uncertainty. These changes increased the extent of the active modeled area from 18 to 47 square miles (mi^2). The optimal parameter set from the calibrated model provided a single, best representation of the areas contributing recharge (deterministic contributing areas). The uncertainty analysis led to contributing areas expressed as a probability distribution (probabilistic contributing areas).

Groundwater flow in the HAP study area was simulated by a four-layer model representing surficial deposits and the underlying bedrock. The model was calibrated to 165 groundwater-level observations and 7 streamflow (base flow) observations. Ten parameters representing boundary conditions and hydraulic properties were evaluated for nonlinear regression: two recharge parameters, three horizontal hydraulic-conductivity parameters, three vertical anisotropy parameters, a streambed hydraulic conductance parameter, and a maximum groundwater evapotranspiration parameter. Five of these parameters were estimated by nonlinear regression. The remaining five parameters, which represented vertical anisotropy of the surficial deposits, hydraulic properties of bedrock, and the maximum groundwater evapotranspiration from the valley-fill deposits, were specified on the basis of values previously reported in the literature because observations alone did not provide sufficient information on them. A hydraulic conductance parameter for head-dependent cells representing a groundwater divide was not considered for estimation, but it was included in model-prediction uncertainty. A model-fit statistic, the calculated error variance, which is a measure of the overall magnitude of the weighted residuals, was 8.91. Influence statistics indicated that 13 groundwater-level and 6 base-flow observations had the most overall influence in the regression and in the optimal set of estimated parameters. Six of seven base-flow observations were influential to the set of estimated parameters, and in addition, two of the base-flow observations were the only observations important to all of the individually estimated parameters, thus indicating the importance of this type of observation in model calibration.

The estimated optimal parameter value for recharge to the valley-fill deposits was 27.3 inches per year (in/yr), and the estimated optimal parameter value for effective recharge to upland till deposits was 18.7 in/yr. The estimated optimal parameter value for horizontal hydraulic conductivity of upland till deposits was 16.2 feet per day (ft/d). Two estimated parameters defined multipliers, one of which applied to spatially varying horizontal hydraulic conductivity of the valley-fill deposits and the other to spatially varying streambed conductance. Calibrated horizontal hydraulic conductivity values of the valley-fill deposits ranged from 20 to 480 ft/d and calibrated streambed conductance values ranged from 10,000 to 52,000 feet squared per day (ft^2/d). Optimal parameter values representing processes in the valley-fill deposits (recharge and horizontal hydraulic conductivity) were the most precisely estimated, whereas the horizontal hydraulic conductivity of till was the least precisely estimated parameter value.

The simulated hydrologic budget indicated that recharge is the dominant inflow to the upland deposits. Groundwater in the uplands discharged mostly to streams (52 percent) and to the valley-fill deposits (38 percent). Most inflow to the valley-fill deposits was from recharge (60 percent), but upland sources contributed a significant amount (about one-third), either directly or indirectly. These upland sources of inflow to the valley-fill deposits include lateral flow from till (15 percent), upward flow from bedrock (2 percent), and a large percentage of the streamflow loss (17 percent).

The 15 production wells are clustered in four pumping centers: two centers in the Hunt River Basin contain seven of the wells, one center in the Annaquatucket River Basin contains five wells, and the fourth center in the Pettaquamscutt River Basin contains three wells. Areas contributing recharge to the well centers were simulated for each center's 2004–08 average withdrawal rates (ranging from 235 to 1,858 gallons per minute (gal/min)) and for each center's maximum pumping capacities (ranging from 800 to 8,500 gal/min). Simulated areas contributing recharge extend upgradient of the well centers to upland till and to groundwater and topographical divides. Some areas contributing recharge include small, isolated areas remote from the well center including, for the maximum pumping rate scenario, on the opposite side of surface-water bodies from the well center. Particle tracks indicated that recharge originating in the upland till and valley-fill deposits travels along deep groundwater-flow paths in the valley-fill deposits and, under pumping conditions, passes beneath the surface-water body to the well center. At the average pumping rate, four of the wells are not overlain by the simulated contributing area because recharge travels along paths above and around the screen interval to the well. Simulated surface-water loss within some of the contributing areas affected their size. For the average pumping rates, the size of the areas contributing recharge to the four well centers ranged from 0.19 to 1.12 mi^2 and covered a total area of 2.79 mi^2, and for the maximum pumping rate, the size ranged from 0.37 to 3.53 mi^2 and covered a total area of 7.99 mi^2. The well center in the Annaquatucket River Basin also derived part of its water from induced groundwater across a simulated groundwater divide with an adjoining basin; this induced groundwater was equivalent to recharge over an area of 0.21 mi^2. Land cover in the areas contributing recharge includes a substantial amount of urban land use for the well centers in the Hunt River Basin, agriculture and sand and gravel mining for the well center in the Annaquatucket River Basin, and primarily undeveloped land for the well center in the Pettaquamscutt River Basin. Simulated groundwater traveltimes from recharge locations to production wells for the maximum pumping rate ranged from less than 1 year to greater than 350 years for each well center. Median traveltimes ranged from 2.9 to 5.0 years, and traveltimes of 10 years or less ranged from 78 to 93 percent for the well centers. These relatively short traveltimes indicated that the wells are vulnerable to contamination from activities on the land surface.

Parameter uncertainty and its associated effects on the simulated areas contributing recharge to the well centers were evaluated using a stochastic Monte Carlo analysis. Optimal parameter values and the parameter variance–covariance matrix from nonlinear regression were used to create parameter sets for the analysis. The parameter variance–covariance matrix preserves the uncertainty of the parameter estimates and the correlation among parameters from the calibrated model. The four inestimable parameters representing glacial hydraulic properties and maximum groundwater evapotranspiration were also incorporated into the parameter variance–covariance matrix. Because observations did not provide enough information to constrain the uncertainty of these four parameters within realistic ranges around the specified values, prior information was required. Thus all parameters that may be important for model predictions were incorporated into the analysis. The uncertainty analysis was an outcome of calibrating the model to available observations, but it also depended on information provided by the modeler. Three acceptance criteria were used to assess parameter sets so that prediction uncertainty was not overestimated: the model converged, model mass balance was 2 percent or less, and the calculated error variance was 20 or less. Of 500 parameter sets using pumping values from the calibrated model, 416 fit the acceptance criteria. Four parameter sets that fit the first two criteria had a calculated error variance greater than 20. Common to these four sets was the fact that the two parameters representing horizontal hydraulic conductivity of the glacial deposits had values outside their 95-percent confidence interval. The 2004–08 average pumping rate and the maximum pumping rate scenarios used these 416 parameter sets with the first two acceptance criteria.

The size of the probabilistic contributing areas for each well center for both average and maximum pumping rates was larger than the size of the deterministic contributing areas for the well center because of the effects of parameter uncertainty. Thus, some areas not in the deterministic contributing area, including additional areas with urban and agricultural land cover that has the potential to contaminate groundwater resources, may actually be in the contributing area. Generally, areas closest to the well centers with short groundwater traveltimes are associated with higher probabilities, whereas areas distant from the well centers with long groundwater traveltimes are associated with lower probabilities. In most cases, areas with high probabilities (greater than 50 percent) generally coincided with the deterministic contributing areas. In some cases, areas near the well centers where simulated streams and ponds intercepted recharge in the calibrated model were in the probabilistic contributing area, indicating that this recharge may instead go directly to a well. Three of the four wells not overlain by the deterministic contributing area for the average pumping rate were in the probabilistic contributing area that was associated with probabilities of 50 percent or less. Areas associated with low probabilities extended long distances along groundwater divides in the uplands remote from the well centers.

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