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Appendix 5. Steady-State Model

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Appendix 5. Steady-State Model

The steady-state model was developed using a variety of observation data and was evaluated primarily with streamflow-discharge and ground-water head data collected in October 2004. The model represents ground-water flow during a seasonal low-flow period. The steady-state model parameters were developed with a combination of parameter estimation, literature values, and "transient model" results discussed later in the report (appendix 7).

Boundary Conditions

Boundary conditions include the movement of water (fluxes) into and out of the model cells. Fluxes were simulated by MODFLOW-2000 (Harbaugh and others, 2000) packages to represent: constant head surface, streams and streamflow routing, recharge, well withdrawals, and spatially distributed withdrawals and returns. Simulated withdrawals and returns are discussed in the steady-state model, "Water Use and Stresses" section.

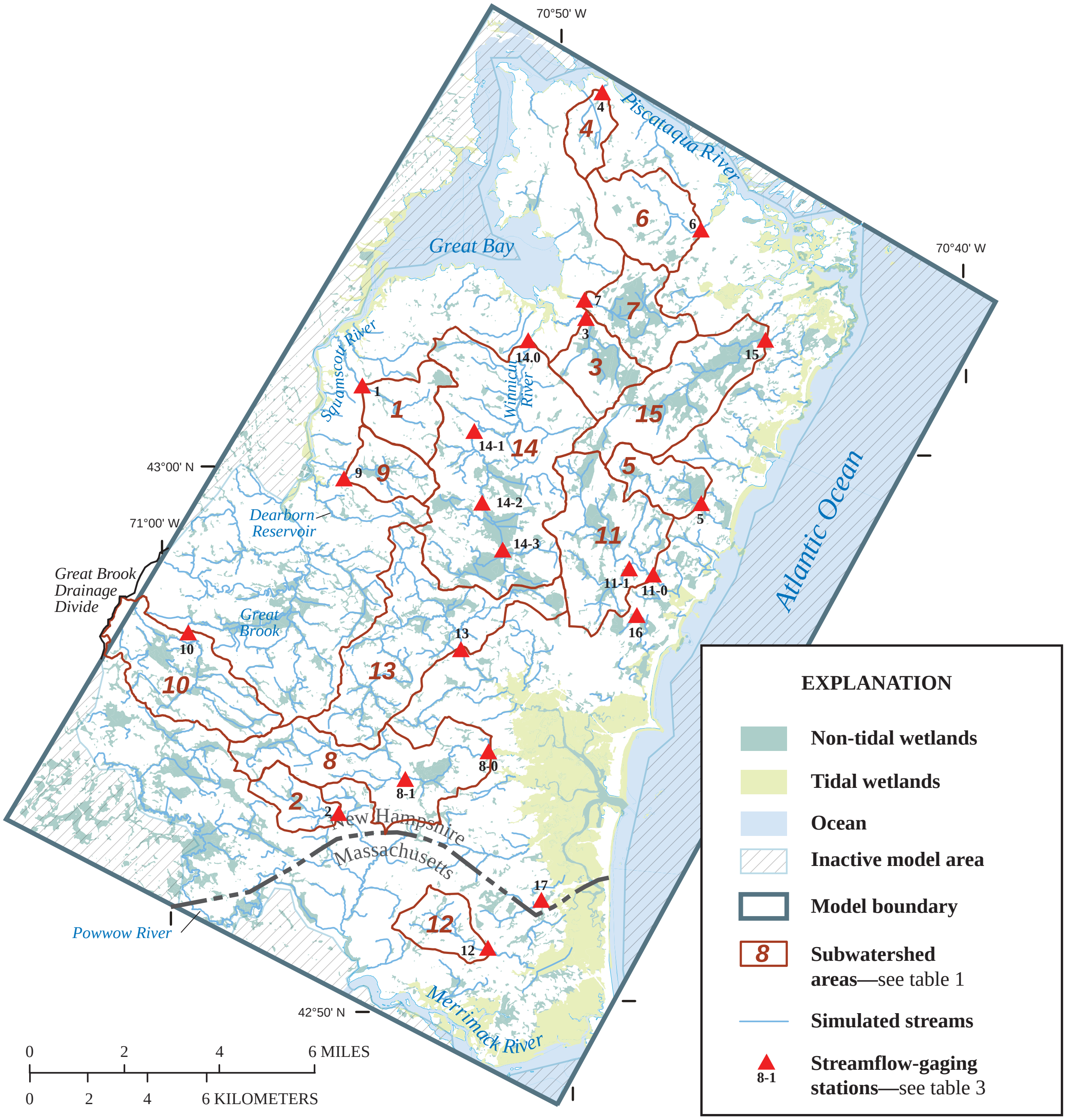
Lateral Extent

The lateral model extent was selected to coincide with major hydrologic features, or boundaries, which consisted primarily of surface water bodies and, in one location, part of a watershed divide (fig. 5–1). A 5-mi segment of the Great Brook drainage divide at the southwestern boundary of the model was defined as a no-flow boundary in model layer 1 (fig. 5–2). Below layer 1, all external lateral boundary cells were simulated as no-flow boundaries. All other lateral boundaries in model layer 1 were coincident with surface-water bodies and were simulated as constant-head boundaries. The lateral 6-mi long boundary south of the Great Brook watershed coincided with the Powwow River in New Hampshire and Massachusetts and was the only freshwater constant-head boundary in the model. More than three quarters of the model's lateral boundaries, a total length of about 50 mi, were defined by saltwater bodies including the Atlantic Ocean to the southeast, the Piscataqua River to the northeast, Great Bay and Squamscott River to the northwest, and the Merrimack River to the southwest.

Constant Head

The location of the fresh ground-water and saltwater interface in the bedrock aquifer of the Seacoast study area is not well known. The ground-water-flow model developed does not explicitly simulate the interface but it is assumed to be coincident with the shoreline on the basis of simulated heads and data from wells near the shoreline. In some low-lying areas immediately adjacent to Great Bay and the estuaries in Hampton Falls and Hampton, completion reports for domestic wells indicate seawater contamination (Frederick Chormann, New Hampshire Geological Survey, written commun., 2005). Seawater contamination can be readily apparent during drilling, and can be distinguished from road salt because it causes the circulated drilling water to foam during air-rotary drilling. In some cases such wells were grouted and another well was completed nearby that yields freshwater. Because such events have been reported only within the past decade, the sustainability of such withdrawals is not known. Locally, the occurrence and movement of water to an individual well depends greatly on the fracture network intersected by the well and the fracture network's connection to the regional fracture network. At the regional scale, the saltwater interface is believed to intersect the land surface at the shoreline (Mack, 2004). However, wells drilled farther inland may intersect the interface in low-lying areas with low hydraulic heads and withdrawal stresses as indicated by well-completion reports.

The saltwater bodies were simulated as constant-head boundaries with an equivalent freshwater head (EFH) (fig. 5–2) determined by the depth of the water body and the density of saltwater. Following this methodology (Langevin, 2003b), the EFH is equal to the depth at the midpoint of the saltwater body (or one-half the water depth) times a factor of 1.025 to account for the greater density of saltwater. Tidal water bodies in the study area generally are deep—in many places more than 50 ft deep—and, therefore, water bodies such as Great Bay form important hydrologic boundaries. Because saltwater depths are shallowest at the shoreline, the cells with the lowest EFH are in the saltwater body adjacent to the shoreline. The effect of the seaward extent of the EFH boundary area was assessed in preliminary model testing (Mack, 2004). It was found that increasing the areal extent of the EFH boundary in the seaward direction increased the flux of water from the sea toward the cells with lowest EFH simulated head; however, increasing the EFH boundary extent did not affect ground-water flow in the interior of the model representing the freshwater system. In a different hydrogeologic setting, with greater withdrawal stresses and hydraulic conductivity,



Streams and water bodies, including tidal and estuaries from 1:24,000
National Hydrography Dataset, 1999

Figure 5–1. Watershed drainage divides, tidal areas, and streamflow-gaging stations in the Seacoast model area, southeastern New Hampshire. (This figure is the same as figure 2 on page 5 in the report.)

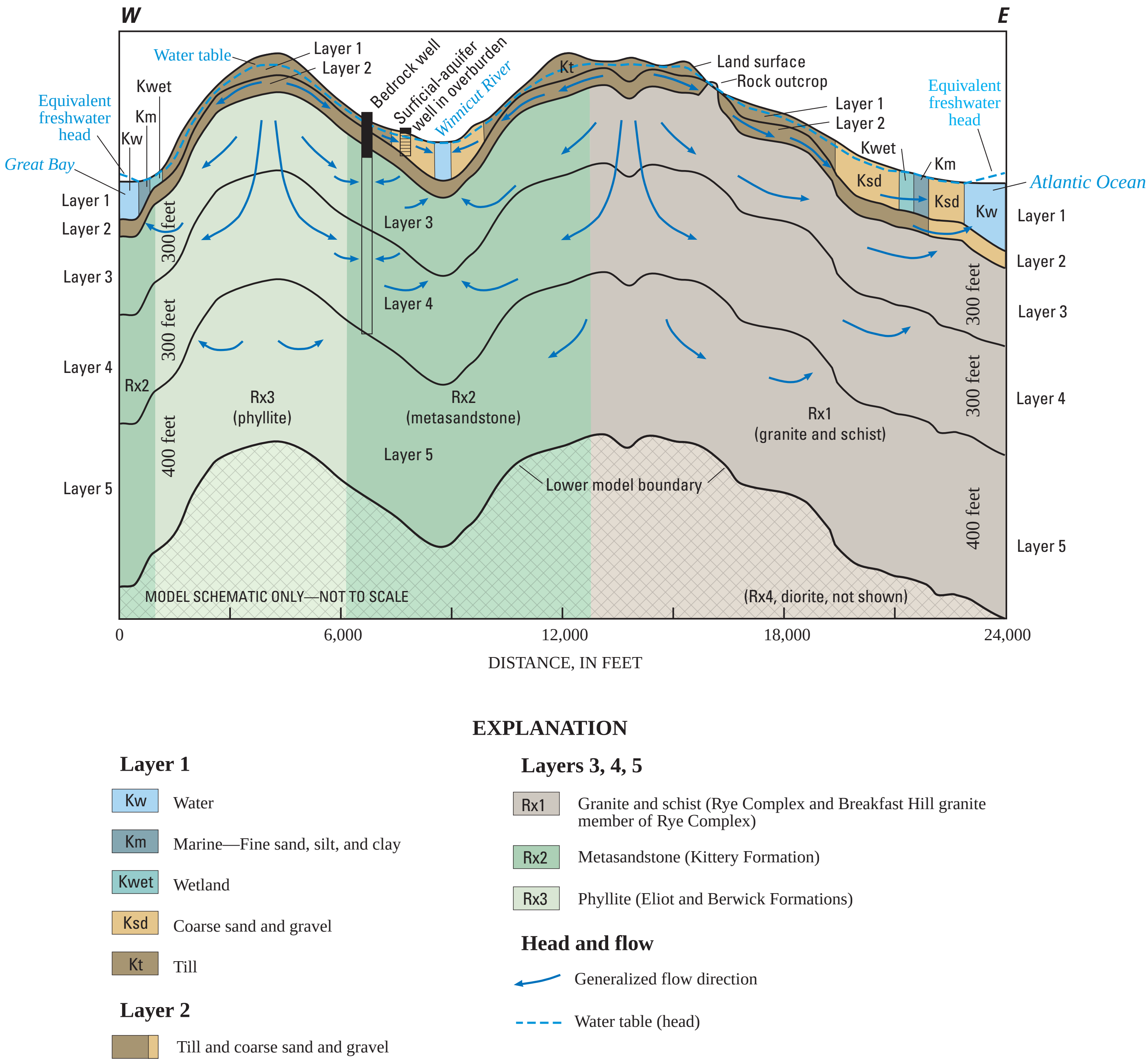


Figure 5–2. Schematic cross section of the ground-water-flow model for the Seacoast model area, southeastern New Hampshire. (This figure is the same as figure 13 on page 28 in the report.)

the lateral extent of the EFH boundary would likely be more important. In regional ground-water-flow simulations of the coastal-plain sediments of Georgia, Payne and others (2006) and Provost and others (2006) found that heads in the freshwater aquifer were not very sensitive to the offshore extent of the EFH boundary unless the offshore extent of that boundary was placed very close (immediately adjacent) to the shoreline. In the Seacoast model, ground-water flow in the freshwater aquifer was insensitive to the lateral extent of the EFH boundary because the withdrawal stresses were low, or not present, near the shoreline and the bedrock aquifer hydraulic conductivity was also low at the shoreline. Although the EFH boundary could have been placed at the point of the lowest EFH head (the shoreline) to represent the Seacoast hydrologic system (accurately), the boundary was placed a few hundred feet to a few thousand feet offshore to ensure that the lateral extent of this boundary was not a concern.

Streams

Stream boundaries were placed at intermittent and perennial streams and rivers in the study area (fig. 5–1), with the exception of tidal rivers. The Streamflow-Routing Package (STR1) (Prudic and others, 2004) for use with MODFLOW-2000 (Harbaugh and others, 2000; Hill and others, 2000) was used because it provides streamflow-routing capability in addition to the use of parameters for estimating streambed conductance, and observations of streamflow for the parameter estimation and observation processes. Stream characteristics required as model input include stream stage, streambed conductance, top and bottom elevation of the streambed, and width.

The flux (Q_L) from the stream to and from the aquifer in the model is controlled by the variables in equation 1:

$$Q_L = (K_{sb} wL/m)*(h_s - h_a), \quad (1)$$

where

(K_{sb}) is the streambed hydraulic conductivity,
w is stream width,
(L) is stream segment length,
m is streambed thickness, and
h_s and h_a are heads in the stream (stage) and aquifer, respectively.

The streambed hydraulic conductivity (K_{sb}) was assigned a parameter *(Ksb1)* and estimated by parameter-estimation techniques. The streambed hydraulic conductivity (K_{sb}) and bed thickness (m) however, were not known, as is commonly true in ground-water and simulations. Values of 1 to 5 ft/d have been calculated for streambed conductivity in New England (Mack and Harte, 1996; Lyford and others, 2003; DeSimone, 2004). It was not necessary to precisely determine the streambed hydraulic conductivity and thickness to simulate the effects of streams in this regional system; therefore, streambed conductance is effectively a lumped parameter.

Conceptually the aquifer system is well connected to the regional river system, which functions as the primary drain for the hydrologic system of the Seacoast. Parameter-estimation tests indicated that ideally the streambeds should not restrict discharge from the regional aquifer system to the streams. The optimal streambed conductivity was likely greater than 10 ft/d; however, lower values were required to prevent numerical oscillations caused by high fluxes at stream cells. Locally, such as at flow scales of tens or hundreds of feet, variations in streambed conductance limit or enhance the flux of water between the stream and the aquifer. Local effects, however, are not apparent at larger scales and were not apparent in the regional ground-water-flow system of the Seacoast area. For example, an aquifer test of Well 16 in Stratham adjacent to the Winnicut River indicated that discontinuous clays limited the drawdown in surficial sediments at the river in the vicinity of the well (Maher, 1997a). Over a larger area, however, withdrawals are supplied by the river-aquifer system. The flux of water into and out of the stream may depend more on the wider ranging hydraulic conductivity of the underlying aquifer sediments than on the local streambed hydraulic conductance. A spatially variable parameter for streambed conductance was assessed in preliminary models with the River Package (Harbaugh and others, 2000) but could not be applied in the current model using the Stream Package and MODFLOW-2000.

The stream-segment length (L) and elevations can be readily determined from DEM data. Stage was assigned the DEM elevation closest to the stream channel; the streambed top and bottom were assumed to be 1 and 2 ft, respectively, below the stage. Streams were generally small and were assigned a width of 5 ft. Actual stream widths may differ considerably at the cell scale (200 ft). In some wetland areas, due to the 40,000-ft^2 cell area, the cell elevation did not accurately reflect the true stream-surface elevation; as a result, the stages for a few cells did not decrease in downstream order. This problem arose for some wetlands with a poorly defined channel or high points in the wetland. Simulation of a stream channel in the wetland was not necessary because the wetland itself had a greater effect on the system than the stream. For the few areas where this occurred, preventing numerical convergence, streams in the selected cells were deleted, and streamflows were effectively routed through the wetland.

Recharge

Initial estimates of seasonal and average annual recharge in New Hampshire were from Flynn and Tasker (2004). Long-term average annual recharge in the Oyster River watershed, about 5 mi north of the Seacoast model area, was estimated to be approximately 19 in/yr. Recharge was applied to the numerical model by the MODFLOW-2000 Recharge package.

Recharge was represented with a parameter (Rech1), at an initial rate of 4.75×10^{-3} ft/d (19 in/yr), and applied to the topmost active model cell. The two exceptions were constant-head cells representing saltwater bodies for which it is not necessary to simulate a recharge, and exposed bedrock. About 0.2 percent of the model area consisted of exposed bedrock (table 5–1), which was generally represented by isolated cells or small clusters of cells. The application of an areal recharge rate, to a low-hydraulic-conductivity model cell representing bedrock would result in an anomalously high simulated head for that cell. In nature, recharge would not enter the bedrock outcrop at the areal rate, but would flow off the outcrop to an adjacent location; in this model, recharge was not applied to cells representing bedrock.

The areal recharge rate estimated for the model was reduced in two areas; wetlands and areas covered by marine sediments. Wetlands were considered to be hydraulically connected to streams, to have a high horizontal hydraulic conductivity, and therefore, to drain rapidly to surface-water bodies. In areas covered by low-hydraulic conductivity marine sediments, a full recharge rate results in water levels that are excessively high relative to the land surface. For these reasons recharge rates were arbitrarily reduced by half in wetland and marine areas. Similar approaches for recharge on wetlands and marine sediments were used by DeSimone (2004). Because marine sediments covered about 18 percent of the total model area (table 5–1) and do not completely cover any one watershed, their areal extent was too small to allow for the estimation of recharge rates by parameter estimation. Additionally, the effect of reducing recharge rates on wetland and marine areas was slight, a change of approximately a few percent, on the regional water balance.

Hydraulic Properties

The hydraulic properties of the surficial and bedrock aquifers differ with sediment and bedrock type. The surficial aquifers consist of glacial sediments that cover more than 80 percent of the model area (table 5–1), and the remainder consists of wetlands and water bodies (fig. 5–3). These aquifers are generally thin and their properties discontinuous. The bedrock aquifers exhibit regional variations in hydraulic properties that are reflected by well yields. Hydraulic properties of surficial and bedrock aquifers were simulated with MODFLOW-2000 using the Layer Flow Parameter (LFP) package. Required model inputs include aquifer thickness and hydraulic conductivity.

Regionally, high-conductivity stratified-drift aquifers act as a local short circuit in the regional ground-water-flow system (Harte and Winter, 1995); for this reason, vertical head differences within these aquifers are negligible. In addition, accurate calculation of vertical ground-water-flow gradients within the surficial aquifer was not critical in the regional model. Therefore, because the surficial aquifer is relatively thin regionally, and detailed vertical flow is not needed in the surficial unit, subdivision of the surficial aquifer beyond the 2 layers described above was not necessary.

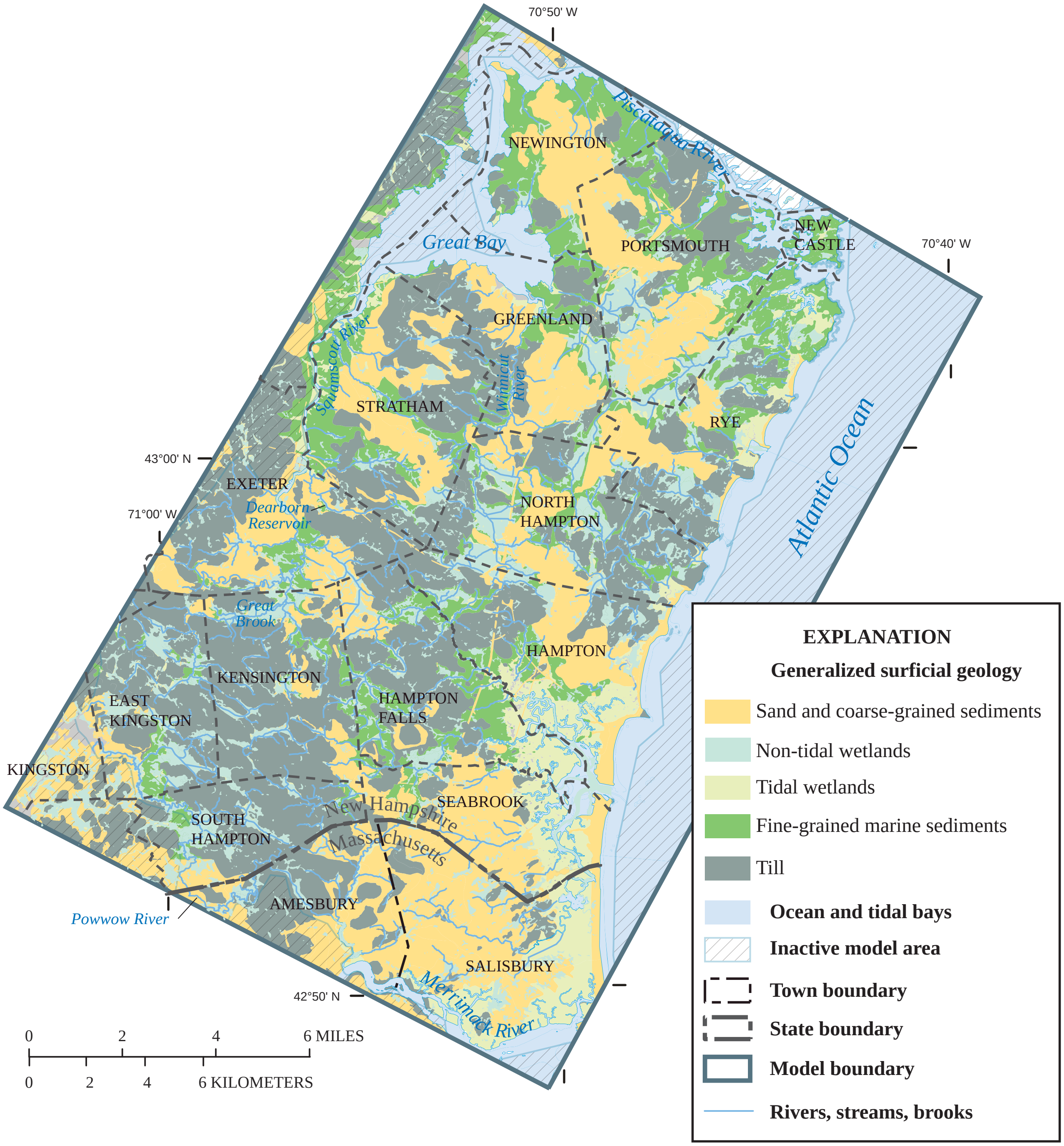
Surficial Aquifers

Most surficial aquifers, which cover approximately 80 percent of the simulated land area, consist of fine-grained till and marine silts and clays (fig. 5–3, table 5–1). Areas of coarse-grained stratified drift, primarily sands and gravels, are discontinuous and cover a smaller percentage of the study area (fig. 5–3, table 5–1). Surficial sediments generally are between 10 and 50 ft thick. The more extensive, high-transmissivity (on the order of thousands of ft^2/d), stratified-drift aquifers in the study area include the Pease aquifer and the stratified-drift aquifer at Great Brook in Kensington (fig. 5–3). Because the specific yields of the overburden aquifer deposits (till, fine and course grained stratified-drift) are similar—about 20 percent—the total amount of water stored in these deposits is controlled by thickness of the deposit.

Table 5–1. Subwatersheds and percentages of surficial geologic sediments, wetlands, surface-water bodies, and bedrock within each subwatershed, Seacoast model area, southeastern New Hampshire. (This table is the same as table 1 on page 10 in the report.)

[All streams are in New Hampshire unless otherwise noted. Subwatershed areas shown on figure 2; mi², square miles; —, not available]

Sub-watershed number (fig. 2)	River or stream name	Area (mi^2)	Stratified drift and other coarse-grained sediments	Marine sediments	Till	Wetland	Surface-water bodies	Bedrock
1	Mill Brook, Stratham	1.98	64.1	24.8	5.3	5.8	0.0	0.0
2	Back River, South Hampton	1.53	1.6	.0	94.1	4.0	.3	.0
3	Packer Brook, Greenland	2.31	49.2	27.0	7.3	14.8	1.7	.0
4	Pickering Brook, Newington	1.29	73.2	22.8	3.9	.0	.0	.0
5	Bailey Brook, Rye	1.95	39.5	10.1	43.0	7.1	.3	.0
6	Hodgson Brook, Portsmouth	3.52	49.4	26.4	23.0	.9	.3	.0
7	Pickering Brook, Greenland	2.97	43.5	11.8	15.8	28.1	.6	.1
8.0	Hampton Falls River, Route 1, Hampton Falls	6.66	22.3	21.7	52.0	3.2	.8	.0
8.1	Hampton Falls River, Mill Lane, Hampton Falls	3.61	10.4	.1	45.0	44.1	.4	.0
9	Parkman Brook, Stratham	1.89	23.8	33.8	42.1	.0	.3	.0
10	Great Brook, Kensington	5.45	1.1	6.8	81.3	10.6	.3	.0
11.0	Little River, North Hampton	6.12	22.9	23.2	46.1	7.4	.4	.0
12	Smallpox Brook, Salisbury, Mass.	1.83	95.0	5.0	—	—	—	—
13	Taylor River, Hampton	8.41	7.1	24.2	59.6	8.5	.6	.0
14.0	Winnicut River, Greenland	14.19	32.5	13.7	43.1	10.0	.6	.0
15	Berrys Brook, Rye	5.38	36.1	35.8	13.7	14.2	.1	.0
	Approximate total area and percentages of materials in the total area	230	24.3	17.6	39.1	7.7	11.1	.2



New Hampshire surficial geology base from 1:24,000 N.H. State Geologist, 2005
Massachusetts base from 1:250,000 MassGIS, 1999

Figure 5–3. Distribution of surficial sediments, wetlands, and water bodies in the Seacoast model area, southeastern New Hampshire. (This figure is the same as figure 3 on page 6 in the report.)

Hydraulic Conductivity

The hydraulic conductivity of surficial sediments was estimated for zones of materials with similar identifiable properties. Initial values of hydraulic properties were assumed on the basis of results from previous investigations described earlier in this report. The hydraulic-conductivity zones used were till (Kt): silt, clay, and undifferentiated marine sediments (Km); wetlands (Kwet); open water bodies (Ksw); and all other glacial sediments not designated as marine fine-grained, which primarily comprised stratified drift but included alluvium and fill (Ksd).

The horizontal hydraulic conductivity of stratified-drift aquifers was calculated as transmissivity divided by saturated thickness both of which were derived from contours determined in previous investigations (Moore, 1990; Stekl and Flanagan, 1992). The calculated horizontal hydraulic conductivity of coarse-grained sediments differed spatially from a few feet per day to more than 100 ft/d. Unlike the other parameter zones, the parameter zone Ksd was defined as a multiplier applied to the spatially differing hydraulic conductivity. The hydraulic conductivity within all other zones was kept constant.

Open water bodies (Ksw), including ponds and tidal water (ocean), were assigned horizontal and vertical hydraulic conductivities of 10,000 ft/d to effectively create flow-through cells and were not evaluated with parameter estimation. Wetland areas were simulated in a similar manner; the assignment of large horizontal hydraulic conductivities (100 ft/d) allowed water to be transmitted horizontally to streams or other features in or adjacent to the wetland, whereas the use of a low vertical hydraulic conductivity (1 ft/d) restricted vertical water movement.

The sensitivities of all parameters (except Ksw) were calculated, and based on the sensitivity analysis, selected values were estimated by using parameter estimation techniques (Hill and others, 2000). Parameter sensitivities and estimated values are discussed under the calibration section later in the report. Because of the regional nature of the Seacoast model, multiple layers of sediment were, in some areas, represented within a single layer (layer 1). Therefore, the effective hydraulic conductivities simulated by the regional parameter zone might have been higher or lower than the hydraulic conductivity that would have been simulated in a finely discretized, or site-specific, representation of a particular sediment. Therefore, the effective hydraulic conductivity calculated for a layered sediment as a whole is greater than that of a single low-hydraulic-conductivity layer within the unit.

Thickness

The thicknesses of layers 1 and 2 (fig. 5–2), which were primarily surficial deposits, were determined by subtracting each layer's thickness from the DEM elevation to produce elevations for layer 1 and layer 2. Layer 1 was meant to account for the varying thickness of stratified drift or other surficial deposits and was assigned a minimum thickness of 3.3 ft. Layer 2 was assigned a uniform thickness of 6.6 ft and was primarily meant to represent a till layer beneath other sediment types. Layer 2 was designed also to provide a thin saturated layer above the bedrock surface to accommodate some fluxes (recharge and water-use-return flows).

Saturated thickness, mapped by stratified-drift aquifer investigations (Stekl and Flanagan, 1992; Moore, 1990), was used as the primary thickness data for layer 1. The depth to bedrock, or secondarily the well-casing length minus 10 ft (the length of casing typically set into bedrock) for wells completed in bedrock, was used where available for the bottom of layer 1. For a group of boreholes within a few hundred feet of each other, such as in the study area for a site-specific investigation, the average depth to bedrock for the borings was used.

For areas where stratified-drift thickness contours or well data were not available, surficial geology (Bennett and others, 2004) was used to approximate the thickness of layer 1 for an area coded as bedrock, 0 ft; for an area designated as "thin," 15 ft; for areas designated as till (not including thin till), 25 ft; and for all other areas (primarily coarse-grained sediments, wetlands, and open-water bodies), 30 ft (a typical thickness). The thicknesses of surficial sediments in Massachusetts (Salisbury and Amesbury) were approximated by using geographic information system (GIS) coverages from MassGIS (http://www.mass.gov/mgis): for areas coded as till or bedrock, 20 ft; areas coded as unclassified sand, 20 ft; areas coded as sand 0–50 ft thick, 30 ft; for areas coded as sand 50–100 ft thick, 50 ft. Although MassGIS information indicated thick surficial sediments in Salisbury, few areas were expected to have thick surficial deposits. For areas with little or no thickness information, generally till areas, layer 1 was assumed to be 15 ft. In the model, till was assumed to be present in all non-stratified drift areas inland of the Atlantic Ocean. Till often is discontinuous beneath stratified-drift aquifers; therefore, for purposes of the regional-scale model, it was not assumed to be present in areas of stratified drift. Layer 1 also includes ocean and tidal-water bodies for which bathymetry data were used to estimate the thickness.

In areas of mapped bedrock outcrops, the thickness of all sediments is zero. Such areas generally were small and discontinuous, usually less than the area of a few model cells or even a single model cell. The incorporation of isolated inactive cells reduced the stability of the numerical model without increasing its capability. For this reason, a minimum thickness of 15 ft for a sediment layer (till) was applied in areas of bedrock outcrop, as described above, and bedrock parameters (discussed below) were assigned to the cell.

Bedrock Aquifers

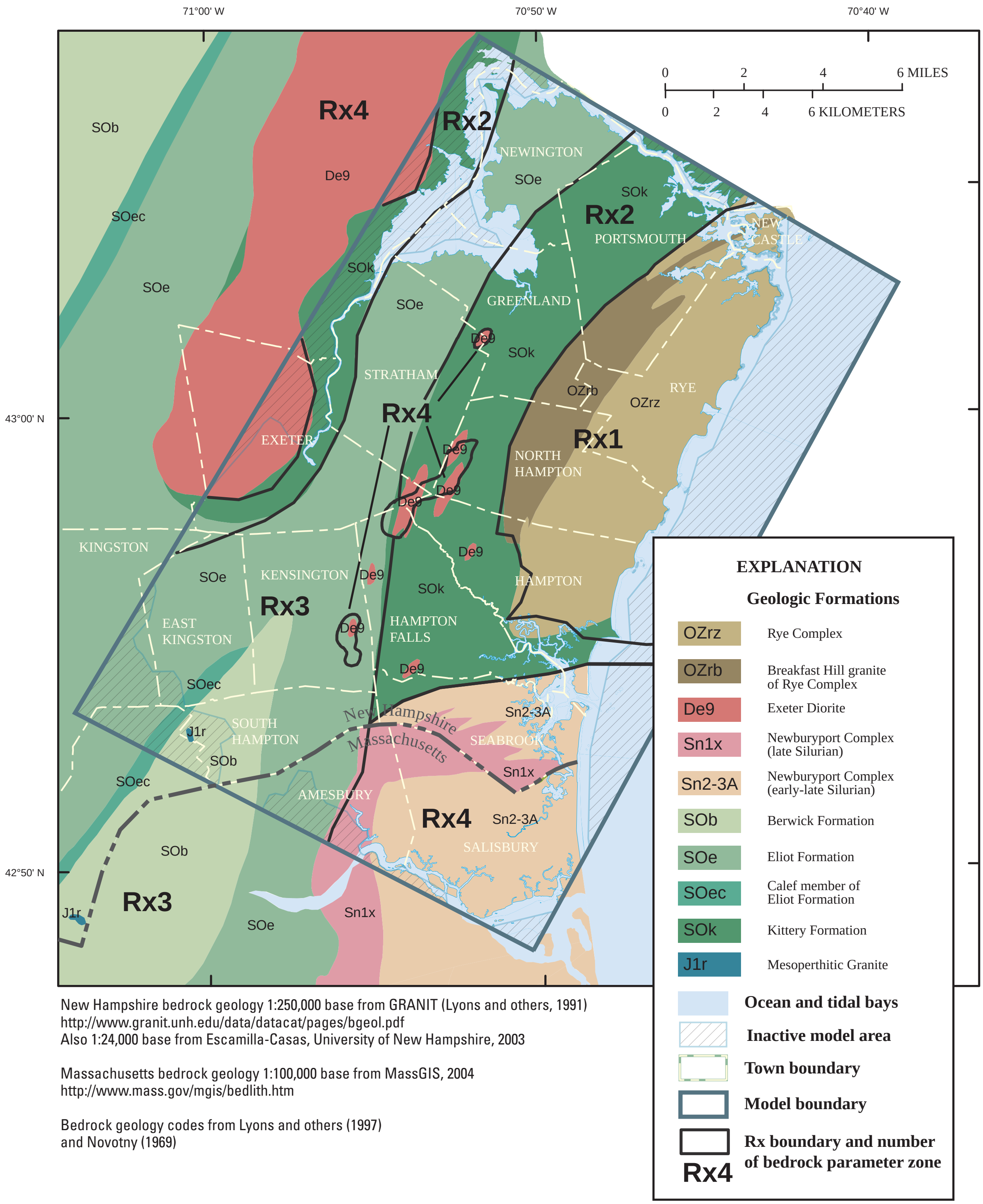
Bedrock in the study area consists of crystalline metasedimentary and igneous rocks (Novotny, 1969). Heterogeneity in the bedrock aquifer was indicated by an analysis of statewide bedrock well-yield probabilities (Moore and others, 2002) that revealed distinct differences from formation to formation in the model area. Some of the highest well-yield probabilities in the State were associated with the Seacoast's Kittery Formation and Rye Complex. Other bedrock units in the study area, such as the Exeter Diorite and Newburyport Formations, were found to have low well-yield probabilities (Moore and others, 2002). Bedrock well-yield probabilities in the Seacoast area (Moore and others, 2002) differed regionally with geologic units (Lyons and others, 1997) and indicated regional variations in bedrock-aquifer hydraulic conductivity. The higher yield probabilities corresponded to areas of coarser grained bedrock such as the well bedded schist and gneiss components of the Rye Complex. Data from high-yielding bedrock wells in the Kittery Formation (Frederick Chormann, New Hampshire Geological Survey, written commun., 2006) that were not available at the time of the investigation by Moore and others (2002) show higher well yields than previously found (appendix 2, table 2–1).

The model grid was oriented parallel to the northeast-southwest regional structural pattern (fig. 5–4). Hydraulic conductivities of the bedrock aquifers in the Seacoast model were examined by zones (fig. 5–4) in the MODFLOW-2000 parameter-estimation process.

The actual hydraulic conductivity of bedrock aquifers in the study area, as in many investigated areas, is not known. The hydraulic conductivity of crystalline bedrock may span several orders of magnitude, as shown by Hsieh and others (1993), and also depends on the scale of investigation. An individual fracture or fracture zone can have a high hydraulic conductivity of hundreds of feet per day. At the regional scale, however, the hydraulic conductivity of the crystalline-bedrock aquifer is determined by the smaller fractures of the pervasive fracture network that forms the connection between high-conductivity zones (Tiedeman and others, 1997, 1998). The regional hydraulic conductivities estimated at the USGS's fractured-rock research site at Mirror Lake, N.H., ranged from about 0.01 to 0.1 ft/d (Tiedeman and others, 1997), whereas the hydraulic conductivities of fractures measured in boreholes at Mirror Lake ranged seven orders of magnitude (Hsieh and others, 1993).

In explorations for bedrock water-well locations, consultants generally assess lineaments as a first step in searching for fracture zones. For example, an assessment of a fractured-bedrock aquifer in coastal Maine found fracture-correlated lineaments, including a buffer zone of about 100 ft, to be positively correlated with greater bedrock well yields (Mabee and others, 1994). A spatial analysis of New Hampshire bedrock well yields (Moore and others, 2002) found lineaments within a 100-ft buffer zone to be positively related to yields. In a similar use of lineament data in finite-difference EPM model of a carbonate-aquifer system, which he terms a "fracture-zone continuum model," Langevin (2003a) simulated fracture traces (lineaments) with a hydraulic conductivity 100 times the matrix conductivity. Langevin (2003a) stochastically assessed bulk (regional) and matrix-block transmissivities and the effects of the matrix-block conductivity of fracture zones on ground-water-flow paths and traveltimes to large ground-water withdrawals. In the Seacoast model, ground-water head and discharge data were not sufficiently detailed to calculate fracture-zone block conductivity or estimate the sensitivities. Bedrock aquifer heterogeneity was incorporated into the model, however, by simulating the hydraulic conductivity of bedrock at cells containing lineaments (Degnan and Clark, 2002; Ferguson and others, 1997a,b) at 10 times the bulk hydraulic conductivity. Within a model-cell area of 200 by 200 ft, this increased value of hydraulic conductivity represents a buffer zone of slightly more than 100 ft around the lineament. A fracture zone may be less than the width of the model cell (200 ft), and the conductivity of fractures in such a zone is likely very high. A preliminary simulation of fracture zones with a hydraulic conductivity two orders of magnitude higher than the bulk conductivity, indicated that fracture zones, although locally very important, do not have a large effect on the regional flow system. The regional aquifer system is more sensitive to the degree of fracturing in the bulk rock matrix (Tiedeman and others, 1997, 1998).

In this investigation the bedrock aquifer was divided into three layers (fig. 5–2). The upper two layers (layers 3 and 4), each 300 ft thick, represented the layers in which most bedrock wells were completed. The lowest bedrock layer (layer 5) was 400 ft thick and represented the depth of a few deep bedrock wells in the study area. Layer 5 likely represents the greatest depths of



New Hampshire bedrock geology 1:250,000 base from GRANIT (Lyons and others, 1991)
http://www.granit.unh.edu/data/datacat/pages/bgeol.pdf
Also 1:24,000 base from Escamilla-Casas, University of New Hampshire, 2003

Massachusetts bedrock geology 1:100,000 base from MassGIS, 2004
http://www.mass.gov/mgis/bedlith.htm

Bedrock geology codes from Lyons and others (1997)
and Novotny (1969)

Figure 5–4. Dominant bedrock formations in the Seacoast area and bedrock parameter zones used in the Seacoast model, southeastern New Hampshire. (This figure is the same as figure 4 on page 7 in the report.)

the principal bedrock-aquifer flow system. Although fracture density most likely decreases with increasing depth, fracture density was assumed to be uniform for the depths investigated (less than 1,000 ft below land surface) and was held constant throughout all of the bedrock model layers (3–5). Johnson and Dunstan (1998) found no uniform pattern of fracturing with depth in a detailed investigation of bedrock boreholes at Mirror Lake, N.H. Hydraulic properties of the bedrock aquifers were simulated by the MODFLOW-2000 Layer-Property Flow (LFP) package (Harbaugh and others, 2000). Initial bedrock hydraulic conductivities of about 0.2 ft/d were modified by using the parameter-estimation process.

The bedrock aquifer was subdivided into zones representing the major geologic formations. Preliminary assessments evaluated nine bedrock parameter zones based on bedrock-formation subdivisions: Rye Complex, Breakfast Hill Granite, Eliot Formation, Kittery Formation east and west of Great Bay, Newburyport Complex, Exeter Diorite and inclusions of Exeter Diorite in other formations, and the Berwick Formation (Escamilla-Casas, 2003; Lyons and others, 1997). Individual zonation of the nine mapped bedrock formations, however, was not supported by the available data used in the regional ground-water-flow simulation. In consultation with Dr. Wallace Bothner (University of New Hampshire, oral commun., 2006) the bedrock units were grouped into four zones of similar hydraulic conductivity on the basis of bedrock-well yields. The formation groupings and the four associated abbreviations for hydraulic conductivity were (fig. 5–4) Rye Complex, and related Breakfast Hill Granite (Rx1); Kittery Formation (Rx2); Berwick and Eliot Formations (Rx3); and the Exeter Diorite Formation and Newburyport Complex (Rx4).

The anisotropy of the bedrock aquifer may have little influence on the quantity of water in the bedrock aquifer. Anisotropy is of interest in this investigation, however, because it is a variable affecting recharge source areas and residence time of ground-water flow to bedrock water-supply wells. Additionally, anisotropy is particularly important with respect to water-supply protection in the model area. In an investigation of coastal bedrock aquifers in Maine, few bedrock wells along Maine’s coastline were found to be affected by saltwater intrusion (Caswell, 1979a, b). Caswell suggests that the absence of extensive intrusion is probably the result of the water-bearing fractures of the crystalline rock being parallel to the coastline. In another investigation (Richard, 1976), saltwater was to found to have contaminated wells in a coastal area of Maine where the water-bearing fractures were perpendicular to the regional structure (the coastline). The model area in this study was also characterized by bedrock with a prominent southwest-to-northeast structure. Studies of fracture orientations in the Seacoast area (Escamilla-Casas, 2003; Novotny, 1969) indicated that fractures were predominantly oriented along the regional structure parallel to the coastline. An investigation of bedrock-fracture orientations (Mack and Degnan, 2003) in boreholes in the Kittery Formation also found the dominant fracture orientation to be aligned with the regional structure.

Anisotropy was assessed in two parameter zones; one anisotropy zone (Hani1) included the bedrock zones Rx1, Rx2, and Rx3 because the general fabric of these three bedrock zones follows the northeast-trending bedrock structure. The second anisotropy zone (Hani2) is the fourth bedrock zone (Rx4), which consists primarily of intrusive bedrock with less fabric-controlled anisotropy (fig. 5–4).

Water Use and Stresses

The distribution of water uses simulated in the Seacoast model is illustrated in figure 5–5. Registered withdrawal wells, withdrawals greater than 40,000 gal/d and community-supply wells were simulated in layer 1 (overburden) and layers 3 and 4 (bedrock) by the Well (WEL) package (Harbaugh and others, 2000). All other withdrawals for distributed commercial, industrial, and domestic water use were simulated in layers 3 and 4 by the Flow and Head Boundary (FHB) package (Leake and Lilly, 1997). Returns associated with the distributed withdrawals were simulated in layer 2 with the FHB package.

Water uses in the Seacoast area at the time of this study, include surface- and ground-water withdrawals and returns for supply, domestic, commercial, and industrial uses. Water use in the Piscataqua River and coastal watersheds has been described by Horn and others (2007). Water-use data used as model input were based on water-use and return coefficients (Horn and others, 2007) and compiled for the ground-water-flow model (Marilee A. Horn and Laura Hayes, U.S. Geological Survey, written commun., 2006). Commercial, industrial, and domestic uses were estimated from use coefficients determined from metered data, business, and census data. Large withdrawals greater than 20,000 gal/d reported to NHDES were incorporated into the model as wells in layer 1 and layer 3 (fig. 5–5A). Minor water users included community water systems (CWS) and industrial, commercial, and domestic users. Water use at CWSs was calculated on the basis of the number of homes connected to a system and water-use coefficients (Horn and others, 2007). Industrial and commercial water-use rates were calculated on the basis of water-use coefficients and business employee and process information. Large withdrawals and CWSs registered with NHDES were simulated as wells and totaled about 7.8 Mgal/d in the model area.

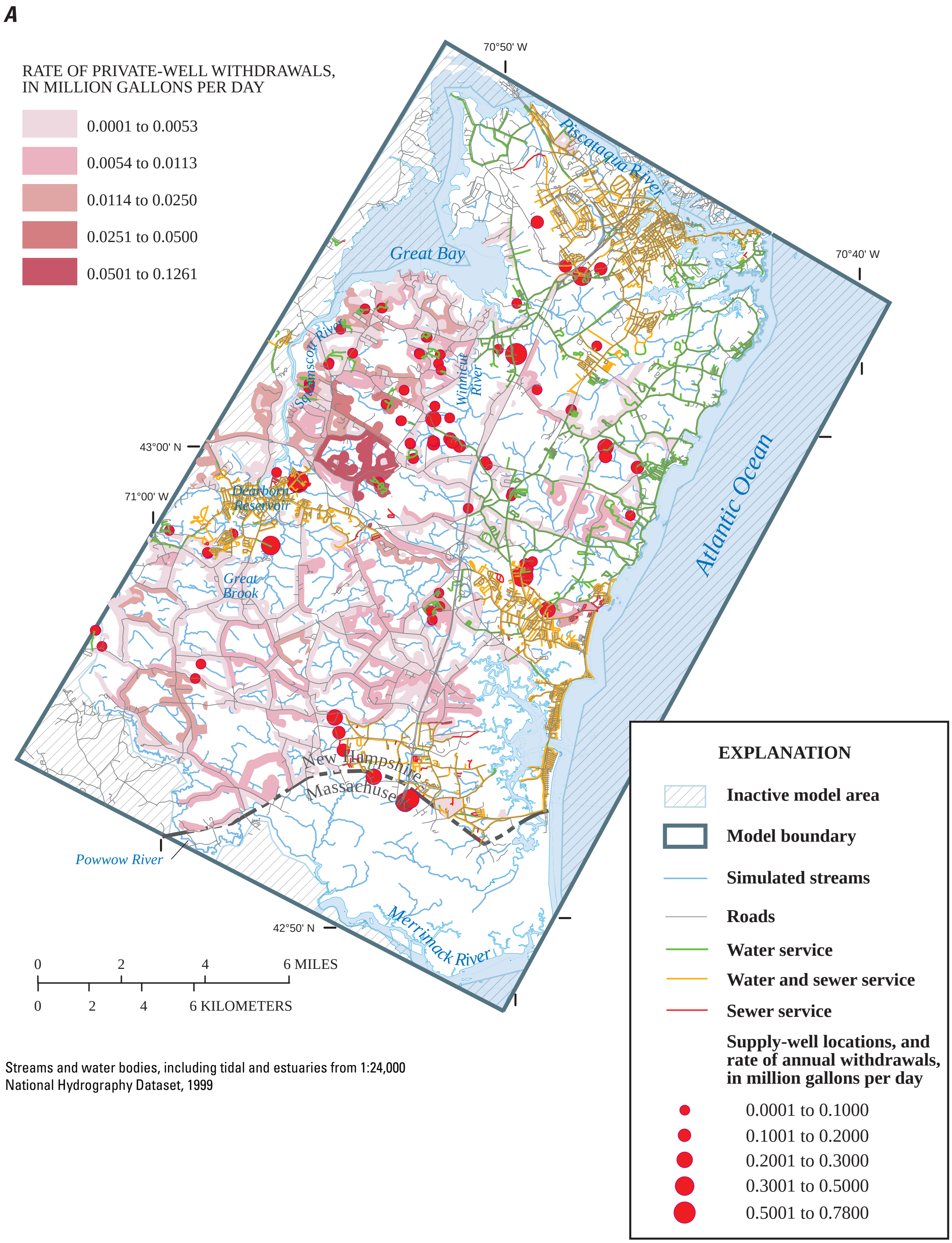


Figure 5–5. *(A)* Private-well and community-supply withdrawal rates, water-distribution, and sewer-service systems in the Seacoast model area in 2003, southeastern New Hampshire.

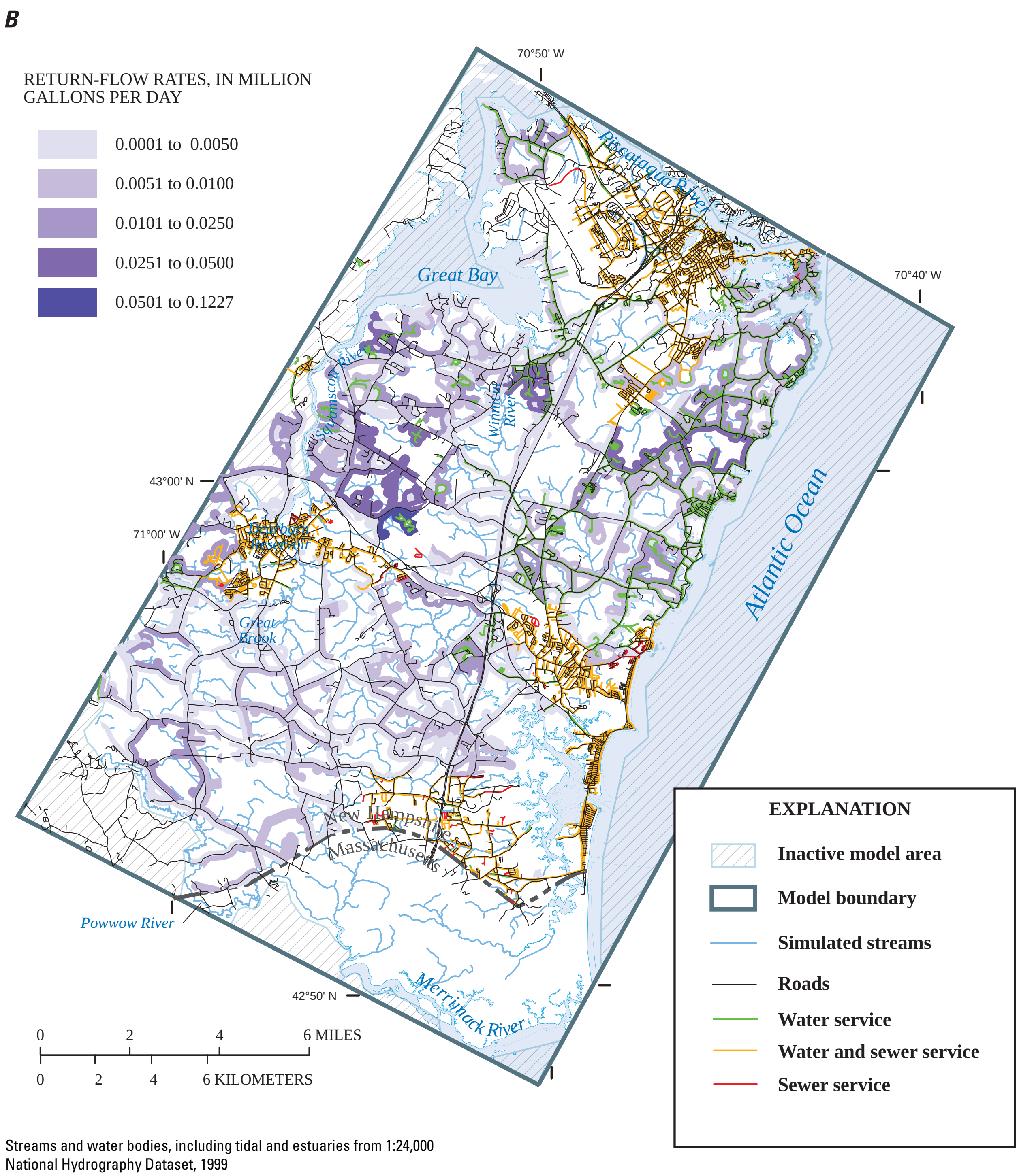


Figure 5–5. *(B)* Water-return rates, and water-distribution and sewer-service systems in the Seacoast model area in 2003, southeastern New Hampshire.—Continued

Smaller withdrawals (not large withdrawals or CWS systems) were almost entirely from bedrock wells and were simulated as distributed uses by the FHB package in model layers 3 and 4. The distributed withdrawals amounted to about 2.7 Mgal/d in the model area. Water returns from the distributed withdrawals were simulated in model layer 2 to approximate infiltration of water from a leach field to the surficial aquifer in all areas that were not sewered (fig. 5–5B). These distributed water uses were estimated by census block for the model area (Horn and others, 2007). The census blocks differed in area from a city block to nearly one-half a square mile, depending on local demographics, and therefore, each block was represented by a few to hundreds of model cells (fig. 5–5B). For most census blocks, the water is actually used near the perimeter of the census block by the homes and businesses predominantly along the streets that form the block boundary. In this case, the water use for the block was applied to model cells on the inside perimeter of the block. At a census-block boundary formed by a stream, the boundary was assumed not to be associated with water use, and the use was distributed to model cells on the inside perimeter of the other three boundaries. The distributed returns amounted to about 2.2 Mgal/d in the model area.

For census blocks with no sewer systems, water returns were simulated by the same method as the distributed withdrawals. For blocks with withdrawals and sewers, withdrawals but no water returns were simulated for that block (fig. 5–5). For blocks with a water-supply system (supplied water) but no sewers, only returns were simulated for that block (fig. 5–5). For blocks with water-distribution systems or sewers, both withdrawals and returns were simulated for that block (fig. 5–5). Consumptive use was estimated by comparison of summer and winter water-use rates (Horn and others, 2007). Returns differed with water-use category and season but were about 85 percent on an annual basis. In some areas with supplied water and sewering, major water uses include the watering of lawns and landscaped areas. It was assumed that most such water use is evaporated or transpired from the surface and does not leach to the underlying aquifers—in other words, that this water is consumed. Such uses were accounted for in the model as a loss.

Surface-water withdrawals, such as agricultural or golf-course withdrawals, were often from small ponds and were simulated as a withdrawal (by the WEL or FHB package) in model layer 1. Surface-water withdrawals from the Dearborn Reservoir in Exeter (fig. 5–1), the only surface-water supply in the model area, was simulated as a withdrawal (well) in a cell of the cluster of cells that represent the reservoir in the model.

Calibration and Observation Data

The steady-state model was calibrated to observations of ground- and surface-water levels and stream base flows from September and October 2004 with additional ground and surface-water-level observations from other periods. Nine broad geohydrologic zones used were used in the model to form consistent geologic or hydrologic areas. The zones consisted of four zones unconsolidated sediment (stratified drift, till, wetlands, and marine sediments), four bedrock zones, and one zone representing open water bodies. Parameter zones were used to represent the hydraulic properties of the units (horizontal and vertical hydraulic conductivity and horizontal anisotropy) and other features such as recharge or streambed conductivity. During calibration, model characteristics were adjusted by parameter zone; for example, the hydraulic conductivity of each zone was consistent within that zone and changes were made to the zone as a whole. Although a better model fit could have been obtained by adjusting the characteristics for individual model cells within parameter zones, such modifications made without a conceptual basis do not improve the quality of information about hydrologic processes in the area and may not result in a realistic model. Instead, the goal of the calibration process was to provide information about regional hydrogeologic processes as opposed to matching discrete observations.

Model calibration depends on observations of the ground-water-flow system. Observations were used to find the best-fit model parameters, and the known or estimated errors of the observations were used to calculate the sensitivity of the ground-water-flow model to the model parameters. Observations included measured or estimated ground-water heads, ground-water discharges (base flows), and the ages of ground-water samples. The accuracies of sets of observations were determined, following the methods of Hill (1998), by weighting observations according to their measurement error. By this process, data sets with greater accuracy were given greater weight in the parameter-estimation process; this permits data sets with different levels of accuracy to be used simultaneously in the parameter-estimation process. Although observational error is rarely known in practice, it can be estimated, and parameter values generally were not sensitive to moderate changes in the weights (Hill, 1998).

The steady-state parameter-estimation process was done for all model layers simulated as nonconvertible (saturated) to linearize the numerical calculations. Although some areas in the natural system were likely to become unsaturated, simulating all layers in the numerical model as saturated greatly simplifies the numerical calculations for a highly heterogeneous model. As was true in this investigation, simulation of low hydraulic-conductivity aquifers can be subject to convergence difficulties and thus unstable; the same problems apply to nonlinear parameter estimation. The simplification (linearization) approach is presented as a guideline for effective model-calibration method described by Hill (1998). Hill (1998) also describes recent publications that indicate that linear confidence intervals were accurate enough for many ground-water flow problems. To ensure that the parameter estimates produced by linear processes were acceptable, the ground-water-flow model was simulated with the same linear parameter estimates while layers 1 and 2 (surficial aquifer) were allowed to convert to unsaturated (nonlinear) conditions. The unconfined (nonlinear) ground-water-flow simulation was found to be less stable, as a result of cells drying; however, the simulation converged with less than 1 percent discrepancy in the volumetric budget.

Base Flows

During periods with little recharge, the streamflow in a watershed is primarily the result of ground-water discharge and is called base flow. Continuous streamflow data can be used to calculate ground-water discharge from a watershed (Rutledge, 2000; Risser and others, 2005). Streamflows were measured throughout the study area on October 7 and 8, 2004, during a period of streamflow recession (table 3). The flow duration at the Oyster River streamflow-gaging station during this period was about 70 percent. Estimated annual average base flows were used as observations of long-term steady-state conditions, and October 2004 base flows were used as observations of a low base flows.

Streamflow measurements in the model area were expressed with an accuracy that represents the estimated percentage of values within 95 percent of their true value (Keirstead and others, 2005). The accuracies of streamflow records were described as excellent, good, fair, or poor indicating that 95 percent of the daily discharges were within 5, 10, 15, or less than 15 percent, respectively, of their true values (Keirstead and others, 2005). Continuous streamflow records in the model area were rated on the basis of site conditions and station operation during the study as: Mill Brook, fair; Winnicut River, fair; Berry's Brook, fair; Little River, good; and Hampton Falls River, good (Keirstead and others, 2005). Spillway operation at the Winnicut River Dam and beaver activity at Berry's Brook prevented the collection of better records at those sites. For comparison, the streamflow record for the Oyster River station was rated good for the same period of record. Individual (miscellaneous) measurements, collected at several streamflow stations in the model area, generally were considered to be good (within 10 percent of their true values). A standard deviation for the measurements within a 95-percent confidence interval can be calculated as (observation × accuracy)/ 1.96 (Hill, 1998). In the parameter-estimation package for MODFLOW-2000, the coefficient of variation can be specified directly with each observation. In the model area, base flows derived from streamflow measurements were observations of the net gain in streamflow over the streamflow in a subwatershed upstream of the measurement point.

Observed and simulated base flows were listed in table 5–2 and shown in figure 5–6. In general simulated base flows agree well with observed base flows. Base flows in most of the larger streams in the model area were simulated well, including Winnicut River, Little River, and Hampton Falls River, as indicated by a weighted residual (dimensionless) within one standard error of regression. The standard error of regression, or standard model error, is defined as the square root of the calculated error variance. If the fit of the model is consistent with the data accuracy, as reflected in the weighting, the expected value of the standard model error is 1.0 (Hill and Tiedeman, 2007). In practice, the standard model error is commonly greater than 1 and reflects model and measurement errors (Hill and Tiedeman, 2007). The standard error of regression was 5.6 for this simulation.

Fourteen streams were simulated within 2 standard model errors (table 5–2, fig. 5–6), and 9 were simulated within one standard model error. Streams that were simulated poorly included some of the small watersheds with the exception of Great Brook, a midsized watershed (5.50 mi^2). The simulated base flow in Great Brook was about twice the measured base flow. A similar anomalous condition was found by Flynn and Tasker (2004) for a station on Dudley Brook in Exeter, about 1 mi west of the model area in a similar surficial and bedrock setting. The measured base flow at Dudley Brook was found to be much lower than the value estimated by hydrograph separation. It is possible that the hydraulic conductivity and storage in the principal bedrock unit underlying the Great Brook drainage were lower than estimated by the model or that more connected wetlands than were simulated drained the Great Brook watershed. The results for Great Brook indicate that this watershed system is not well understood or that it may not be more complex than the modeled watershed. Streamflow measurements were collected only once at Great Brook, and those measurements may not be representative of base-flow conditions.

Nilus Brook is in a small (1.5 mi^2) watershed where the model calculated essentially no base flow. Streamflow measured in Nilus Brook (table 3) in October 2004, however, was drainage from a pond and likely does not include ground-water discharge. For this reason, the Nilus Brook streamflow measurement was removed from the observation data set.

Table 5–2. Observed and simulated base flows for October 2004 in the Seacoast model area, southeastern New Hampshire.

[Site number shown on figure 2; mi^2, square miles; ft^3/s, cubic feet per second; —, not applicable]

Site number	Stream	Area (mi^2)	Observed (ft^3/s)	Simulated (ft^3/s)	Weighted residual (dimensionless)
14-0	Winnicut River, Greenland	14.2	-7.3	-7.6	0.4
13	Taylor River, Hampton Falls	8.41	-4.8	-6.9	4.4
15	Berrys Brook, Rye	5.38	-3.1	-3.9	2.5
11-0	Little River, North Hampton	6.12	-2.5	-2.8	1.1
10	Great Brook, Kensington	5.5	-1.8	-3.7	10.3
8-1	Hampton Falls River, Hampton Falls	3.62	-1.7	-1.4	-1.9
6	Hodgson Brook, Portsmouth	3.52	-1.2	-0.3	-7.1
5	Bailey Brook, Rye	1.73	-1.1	-.5	-5.0
1	Mill Brook, Stratham	2.48	-1.0	-.9	-1.1
12	Smallpox Brook, Salisbury	1.83	-1.0	-1.6	6.4
16	Nilus Brook, Hampton	1.5	-0.61	—	—
7	Pickering Brook, Greenland	2.97	-.6	1.0	5.9
9	Parkman Brook, Stratham	1.91	-.6	-.6	.0
3	Packer Brook, Greenland	2.25	-.6	-.6	1.5
2	Back River, South Hampton	1.53	-.5	-1.1	10.8

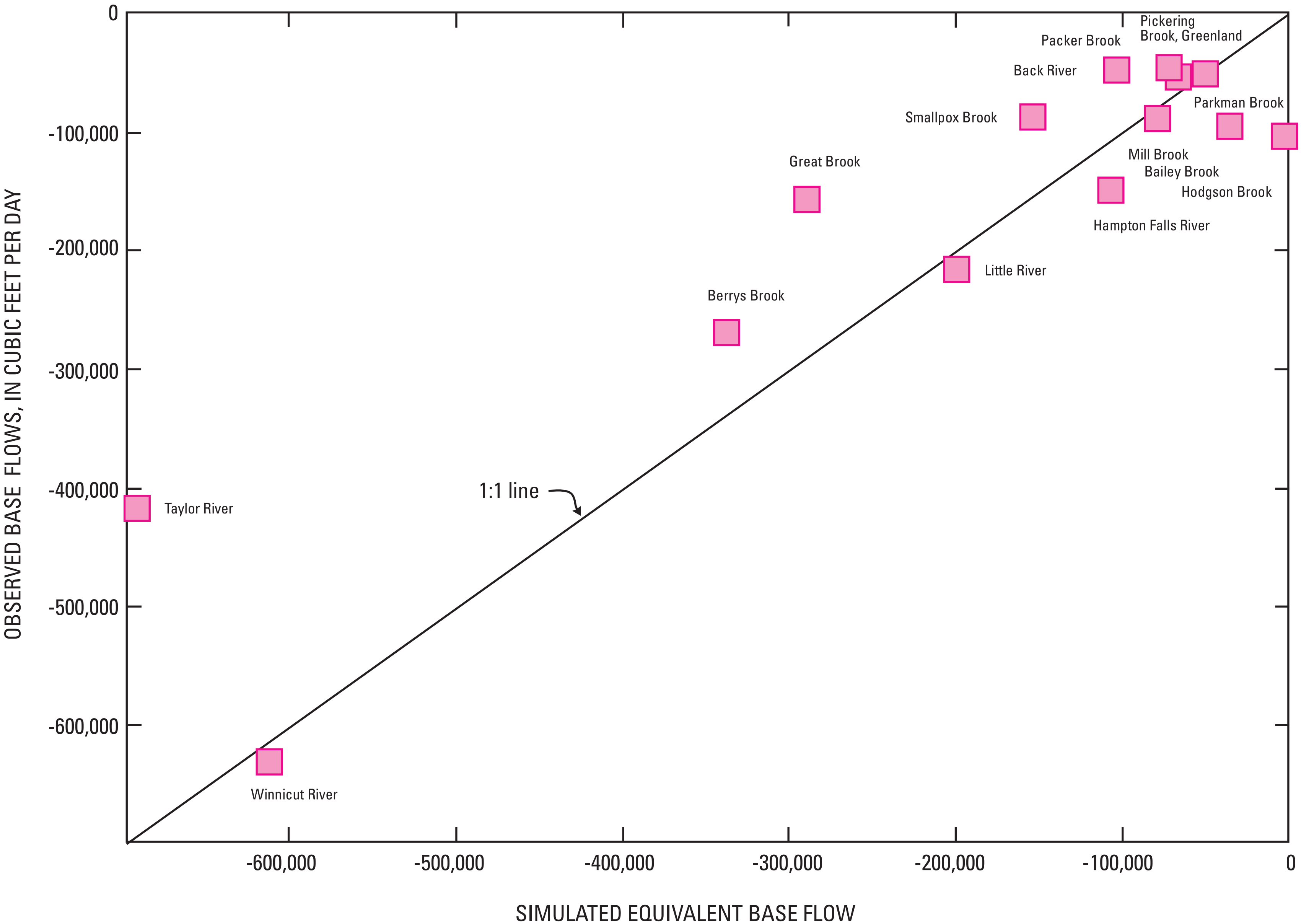


Figure 5–6. Observed and simulated base flows for October 2004 in the Seacoast model area, southeastern New Hampshire. The base flows are negative because they represent flow out of the aquifer system.

Ground-Water Levels

Preliminary analyses indicated that in the regional simulation, a greater distribution of less accurate observations is more important than fewer highly accurate observations (Mack, 2003). For this reason, ground-water levels (heads) used in model calibration were obtained from six sources to provide areally distributed calibration points. The sources included well completion reports, waste-site reports, previous investigations, hydrographic coverages of water suppliers water-level networks, and heads measured in this study. Overburden-well data, or other surficial-head data, were used for model layer 1, and bedrock-well data were used for model layers 3 and 4. Head data generally were expressed as a depth or elevation with an estimated error (length). Ground-water-head observations were weighted according to measurement accuracy according to the guidelines of Hill (1998) and Hill and others (2000).

As of March 2006, the New Hampshire database of well-completion reports (Frederick Chormann and Derek Bennett, New Hampshire Geological Survey, written commun., 2006) included nearly 2,800 georeferenced wells in the model area. Of the ground-water-level measurements in this data set, termed "Historical" heads in the calibration analysis, about 75 were in the overburden aquifer and 820 in the bedrock aquifer. Most of the measurements were in drillers' completion reports for wells constructed during the past 20 years. Ninety-eight percent of those wells were bedrock wells drilled for domestic use. The water-level observations represent measurements recorded by drillers in all seasons with a slight seasonal bias. Seventy-three percent of the bedrock wells for which water levels were recorded were drilled in the 7 months from April through October, while 27 percent of wells were drilled in the remaining 5 months. Most of the water-level measurements were equally distributed in time between April and October, the months during which the range in water levels is generally the greatest. The seasonal variation in overburden or bedrock ground-water levels, which was not accounted for in the model, is about 2 to 4 ft. The drillers' measurements of depth to water may be accurate to less than a foot; however, the ground-water altitudes used contain other sources of error. Of greater magnitude than seasonal bias is resulting from estimation of the measurement-point elevation. Ground-water altitudes derived from this data set were calculated by subtracting depth to water from the measurement point, which in this case was the model-cell surface elevation interpreted from the DEM. The DEM can be considered to be accurate to within one half the contour interval, or 10 ft. If half of one contour interval (10 ft) is assumed to be the 95-percent confidence interval, the standard deviation of the DEM value would be about 5 ft. Because the measurement-point elevations were the primary sources of error in this water-level data set, the water-level data completion in the report were weighted on the basis of the DEM accuracy. Given the uncertainties involved in well location and measurement, a standard deviation of 10 ft was used for water-level observations from this data set.

Overburden and bedrock heads also were obtained from the NHGS GEOLOGS database for waste sites (Gregory Barker, New Hampshire Geological Survey, written commun., 2005). The water levels generally were measured quarterly or biannually at a specific site. Sites with a total of 78 overburden and 10 bedrock-aquifer monitoring wells with multiple water-level measurements were selected, and the measurements for each well were averaged to provide a mean level. Although the data generally were surveyed and collected with an accuracy of 0.01 ft, water-level data were averaged and a standard deviation of 7.5 ft was assumed.

Heads also were collected primarily in September and October 2004 from ground-water-level networks at 133 wells. The head data collected during this period were termed "Synoptic" heads. Ground-water levels also included 122 bedrock heads measured adjacent to Great Bay during June 2000 (Roseen, 2002) and heads measured at the former PAFB in Newington. These data generally were reported to be within 0.01 ft, and for the purposes of the regional ground-water-flow model, the standard deviation was 5 ft. These data and the GEOLOGS data were termed "Accurate" heads in the calibration analysis.

The well data used in calibration provided a widely distributed set of calibration points; however, areas of low relief were not represented by the wells because development was not in low-lying (stream and wetland) areas. Surface-water altitudes of stream and wetland surfaces are commonly used to approximate the water-table surface. The presence of streams and wetlands containing streams in the study area indicate that the water table in those areas is at the land surface. The hydrographic coverage in the National Hydrography Dataset (http://nhd.usgs.gov) was used to identify model cells representing persistent streams. Water-surface altitudes obtained from the DEM were used as head observations (water table) in model layer 1. Stream surfaces, particularly for the gaining streams which represent points of discharge, have less annual range in altitude than ground-water levels. Observations based on stream locations were considered to have the same accuracy as the DEM, one-half the contour interval, and therefore, the same standard deviation of 5 ft. The altitude of the water table is considered to be equal to the altitude of the cell in model layer 1, which represents the stream or water body. For the large areas with little relief at lower altitudes, surface-water heads were selected at 2-ft intervals at altitudes less than 60 ft, and at 5-ft intervals at altitudes over 60 ft.

Preliminary simulations were tested with thousands of surface-water points and with and without surface-water head data. The number of points used in the analysis was reduced to approximately 550 to avoid the potential of too much influence from one data source in the analysis. Because a low weight was assigned to these observations, this dataset did not excessively influence the results of the calibration. These data were termed “Stream” heads in the calibration analysis.

The average weighted-head residual for all head categories combined, including all simulated minus observed head data, was 0.3 ft (table 5–3). The average residual of unweighted simulated minus observed heads was -2.2 ft. In general, the model fit to ground-water levels and surface-water point elevations is good. On average, simulated heads were slightly less than observations. This is understandable because many of the observations were collected during periods of the year with higher flow conditions. Only the synoptic head data were specifically collected during low-flow conditions. Because head observations had varying degrees of accuracy, heads were compared by weight, or accuracy; group and weighted equivalent residuals and actual head residuals are shown in table 5–3. Weighted simulated and observed values were used in parameter estimation and sensitivity calculations. The unweighted, or actual, simulated minus observed values indicate that for the historical head values, although the standard deviation of simulated minus observed heads is nearly 22 ft, over an entire regional aquifer an unweighted average head difference of less than 2 ft is good.

The overall fit of simulated observed heads is shown by a plot of weighted head observations against weighted simulated equivalent heads coded by head category (figs. 5–7A, B). The scatter of observations about the 1:1 line (fig. 5–7A) indicated that the model calculates heads fairly well at a regional scale, and residuals (simulated minus observed weighted heads) were equally distributed above and below the zero residual (fig. 5–7B). The head observations, however, were clustered by location; therefore, the residuals also are clustered. An analysis of residuals indicates that the residuals are not independent and normally distributed likely because of the clustered nature of the data. Figures 5–7B illustrate that the historic data and the water surface, which are much less clustered than data for the other head categories, were simulated very well but have less weight in the calibration.

Most of the weighted residuals, or 91 percent (table 5–3), were within one standard model error of regression (5.6). The data set with the largest residuals was the synoptic data set, for which only 35 percent of the observations were within one standard error (table 5–3). Most of the synoptic head residuals were within 2 standard errors (fig. 5–7B). This analysis, however, indicated that the calibrated ground-water-flow model, although satisfactory for regional analysis, is not sufficiently accurate for detailed assessment of local heads. Table 5–3 indicated that the widely distributed but less accurate data were sufficient for regional-model calibration as suggested in the preliminary model analysis (Mack, 2003). Some large errors that were caused by a poorly determined measurement-point altitude by the model-grid discretization can result in differences between the cell altitude and

Table 5–3. Residuals between steady-state observed and model-calculated water levels by data group in the Seacoast model area, southeastern New Hampshire.

[ft, feet]

Head category[2]	Weight	Number of observations	Percentage within one standard error of regression (5.6)	Weighted simulated minus observed head[1]		Actual simulated minus observed head		Figure symbol (fig. 5–7)
				Average difference (ft)	Standard deviation	Average difference (ft)	Standard deviation	
Historical	0.1	711	98	0.2	2.2	-1.9	21.8	Red circle.
Accurate	.2	204	90	1.6	3.1	-8.3	16.4	Yellow square.
Synoptic	1	134	55	-.5	17.9	.5	17.9	Green square.
Stream	.16	546	99	.2	1.2	-.9	7.4	Blue triangle.
Combined	Mixed	1,595	91	.3	5.6	-2.2	17.2	

[1] Plot of weighted simulated equivalent and observed heads shown in figure 5–7B.

[2] Head categories discussed in text.

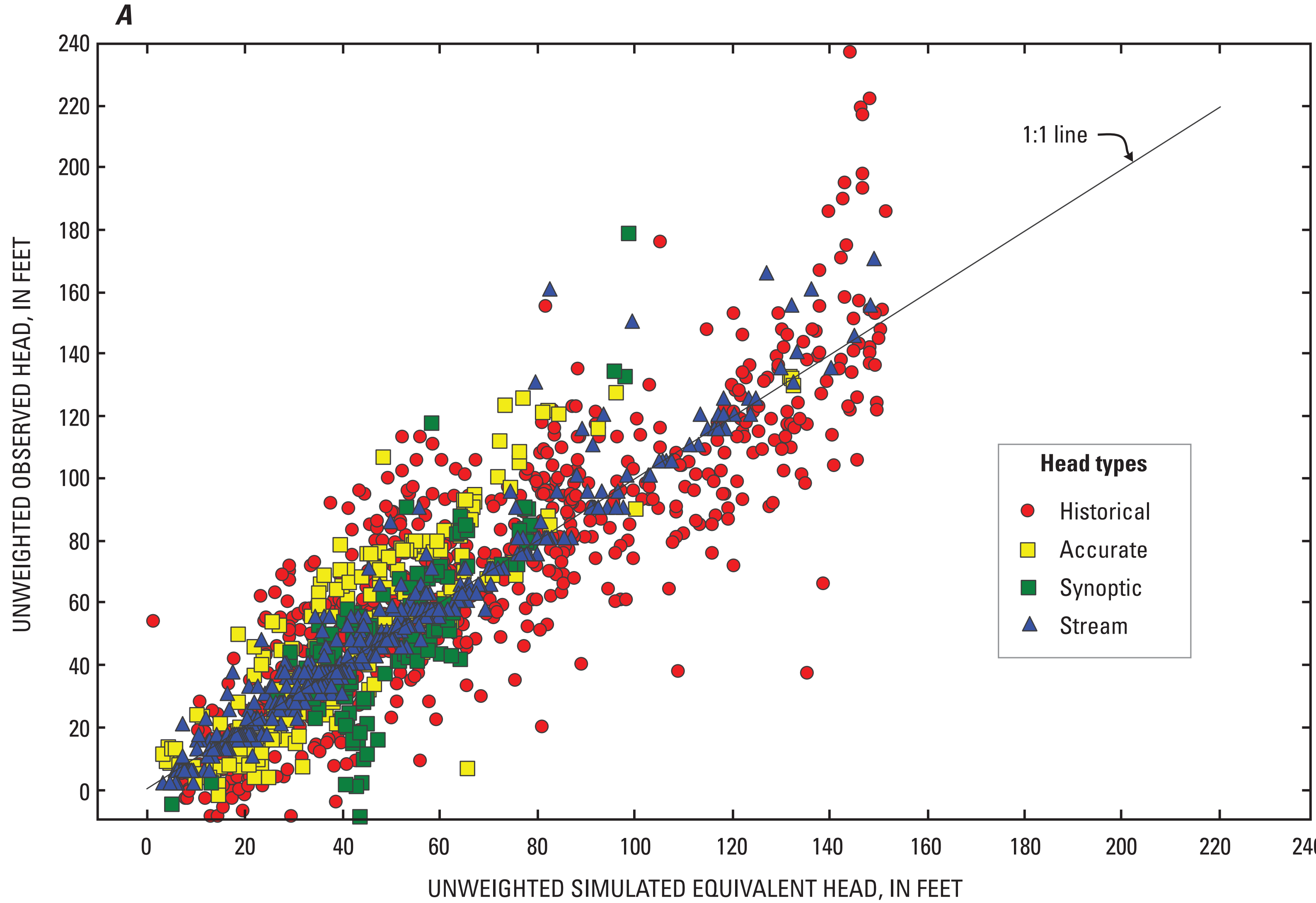


Figure 5–7. Simulated and observed ground-water heads for all categories, except for stream-observation data: *(A)* unweighted simulated and observed heads compared to a 1:1 line.

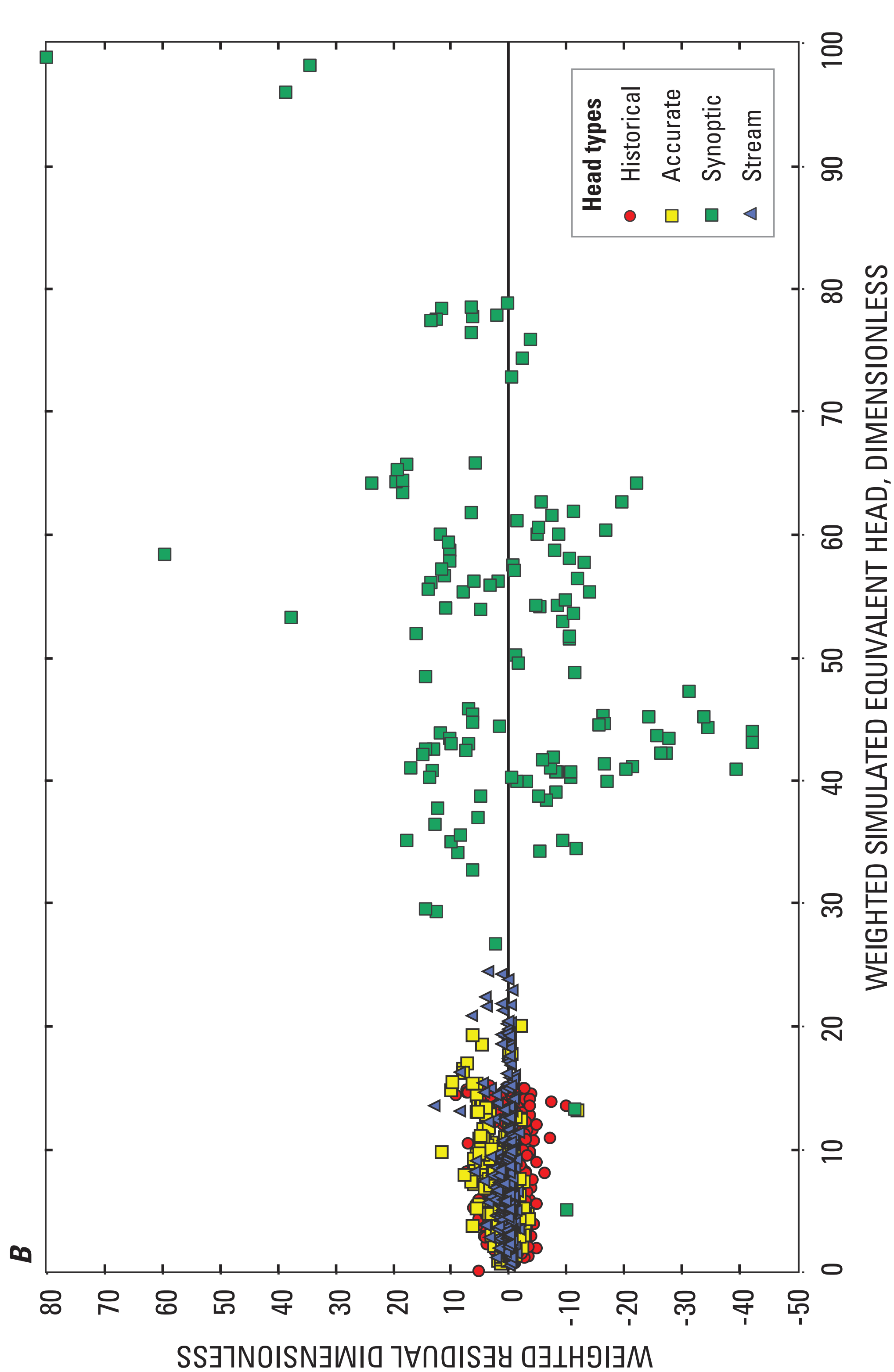


Figure 5–7. Simulated and observed ground-water heads for all categories, except for stream-observation data: *(B)* weighted residuals (simulated minus observed heads) for weighted simulated heads.—Continued

actual altitude. Some of the larger negative residuals could have been caused by local, or transient, pumping effects not explicitly accounted for in the simulation. Larger residuals also were present in geographically clustered data. Areas with large residuals generally correspond to areas with great relief. For example, the calculated head at SSW-7 is more than 50 ft lower than the measured head. This well is located at the base of a till drumlin, however, where the simulated head surface in the surficial aquifer has much less topographic relief than the actual head surface. The model standard error may be improved if local or transient observation errors were identified and some clustered data were excluded. However, it is important to note that ground-water levels can be measured with a far greater degree of accuracy, as indicated by table 5–3, than can be simulated with, or is necessary for, the regional ground-water-flow model.

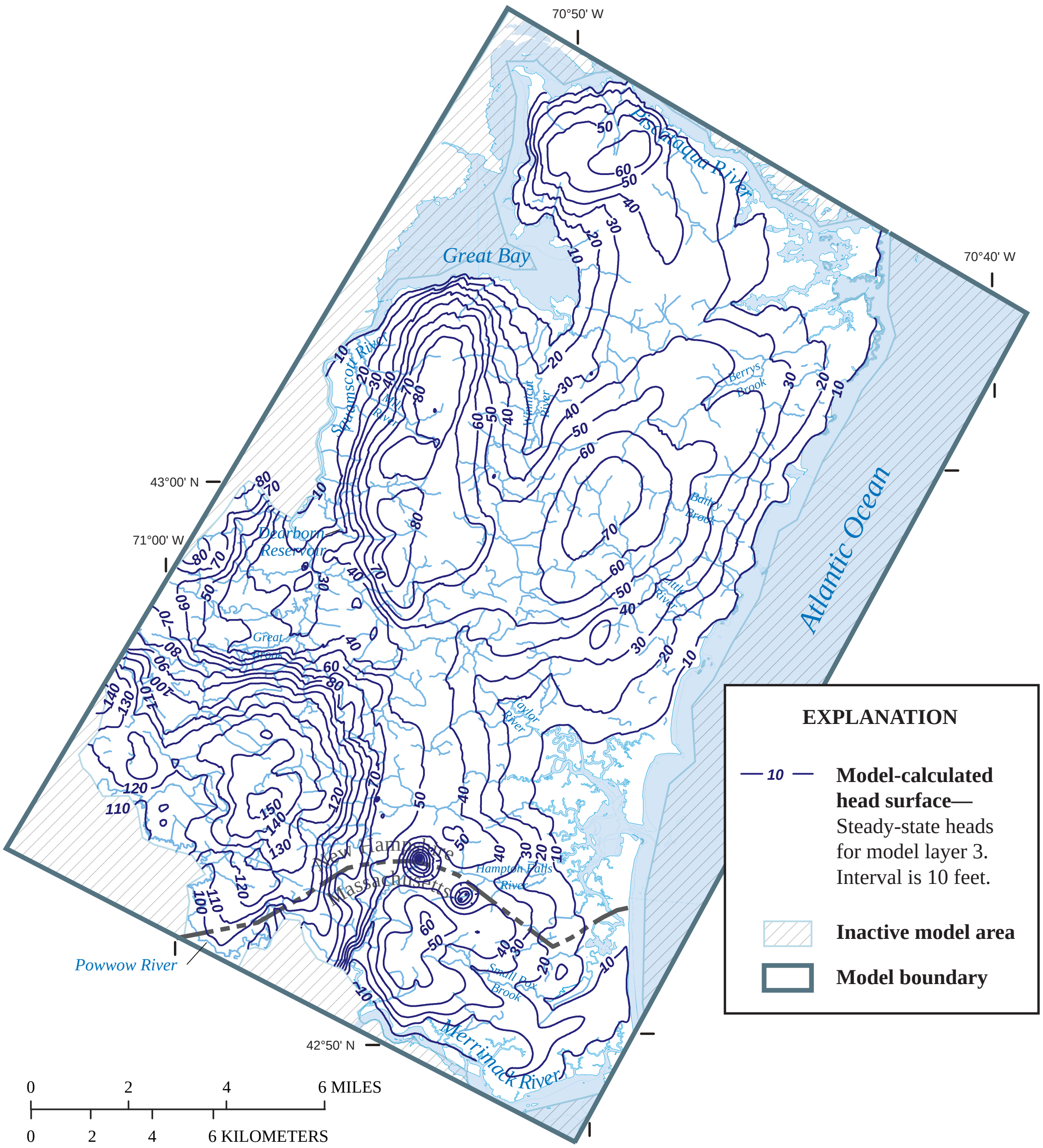
The simulated head surface is shown in figure 5–8. The simulation of heads with a fully linear model, necessary for parameter estimation, shows that in some areas the calculated head surface is above the land surface. In such areas, simulated ground-water flow in the surficial layers may be greater than is realistic. Areas with large positive residuals (heads more than 20 ft greater than DEM values) include parts of Newington, Stratham, and Kensington. Using estimated parameters with the top layer (1) convertible between saturated and unsaturated, allows areas of the model, particularly upland areas and areas with narrow ridges, to dewater. The overall model fit and simulated flows were similar for the fully linear (layer 1 confined) and nonlinear (unconfined) models. Although dewatering resulted in local differences between the linear and nonlinear models, the calculated heads and components of flow generally differed by less than 5 percent between the models. Residuals were analyzed on the basis of parameters estimated with the linear model and heads calculated with the nonlinear model; this approach is suggested by Hill (1998) for use with complex ground-water-flow models.

Estimated Model Parameters

Model parameters were estimated and their sensitivity evaluated using MODFLOW-2000 by the methods described by Hill (1998) and Hill and others (2000). Parameters were used in defining most components of the ground-water-flow equation as described above. In developing a ground-water-flow model, however, it is necessary to make simplifications and assumptions. The model, therefore, is not an exact representation of the ground-water-flow system. Likewise, the statistical analysis of the ground-water-flow system also incorporates simplifications and assumptions. For these reasons, the initial observation weights were adjusted to increase or decrease the relative importance of head and discharge information. Specifically, the calculation of observation weights described above resulted in greater weight placed on head observations as a group than stream observations as a group. This is true even if the surface-water head observations are removed from the analysis. Although the relative differences in head observations are appropriately accounted for, statistical weighting by measurement error does not adequately reflect the importance of the watershed-scale nature of the streamflow observations. Because the focus of this investigation was the regional water balance, observations of base flows on a watershed scale were of greater importance than head observations. For this reason, weights for surface-water observations were increased by a factor of about 5 so that their group weight was comparable to the group weight of the head observations. This method of weighting is currently favored by some researchers over a weighting scheme based entirely on statistical error (Randall Hunt, U.S. Geological Survey, written commun., 2007).

Table 5–4 lists 21 parameters that were assessed for sensitivity along with the final value and lower and upper 95-percent confidence intervals. A correlation matrix for the steady-state model parameters is provided in appendix 6. The final parameter values (table 5–4) are a result of estimation through inverse modeling, literature values, and results of a transient analysis discussed later in the report (appendix 7). A number of parameters were not sensitive enough, or were based on too few observation data, to estimate by inverse techniques. Less sensitive parameters (table 5–4) were removed from the estimation process and were assigned values estimated in other investigations, values published for similar materials, or through trial and error. Therefore, some manual steps were involved in the calibration process where parameters were removed from the parameter estimation process, and the parameter estimation procedures were repeated.

The relative sensitivity of the parameters and the calculated range of the confidence interval provides information about the ground-water-flow system and the relative importance of the components of flow. For example, areal recharge (Rech1) is the most important parameter in this ground-water-flow model. The aerial recharge, calibrated to low-flow conditions, was approximately 11 in/yr. Horizontal anisotropy for bedrock groups Rx1, Rx2, and Rx3 (Hani1) also was relatively significant and indicated that the hydraulic conductivity of the bedrock was greater in the direction of the dominant bedrock structure than transverse to the structure. Using horizontal anisotropy of about 2.5:1 (column to row) for Hani1 improved the model fit. Preliminary variographic analysis of bedrock well yields in the model area indicated similarities in yield in the northeast-southwest (column direction).



Base from USGS and GRANIT, New Hampshire State Plane Coordinate System, North American Datum 1983
Model-calculated head surface from steady-state model layer 3

Figure 5–8. Surface representing calculated steady-state heads for model layer 3, October 2004, Seacoast model area, southeastern New Hampshire. (This figure is the same as figure 14 on page 30 in the report.)

Table 5–4. Steady-state sensitivity analysis, parameter values, and confidence intervals for the Seacoast ground-water-flow model, southeastern New Hampshire.

[Most parameter groups shown on figures 5–2 and 5–4; C.I., confidence interval; ft/d, feet per day]

Sensitivity rank	Parameter name	Sensitivity (dimensionless)	Lower 95-percent C.I.	Final value	Upper 95-percent C.I.	Units	Parameter description, recharge zone or hydrogeologic material group
1	Rech1	6.02	2.0E-03	2.5E-03	3.0E-03	ft/d	Areal recharge.
2	Hani1	2.15	1.4E+00	2.5E+00	4.6E+00	ft/ft	Horizontal anisotropy, bedrock units Rx1, Rx2, Rx3.
3	Rxk2	2.12	6.0E-01	1.0E+00	1.7E+00	ft/d	Horizontal hydraulic conductivity, bedrock unit Rx2.
4	Rxk3	1.92	5.0E-02	1.0E-01	2.0E-01	ft/d	Horizontal hydraulic conductivity, bedrock unit Rx3.
5	Ksb1	1.07	1.8E+00	2.5E+00	3.5E+00	ft/d	Streambed hydraulic conductivity.
6	Rxk1	0.86	2.3E-01	5.0E-01	1.1E+00	ft/d	Horizontal hydraulic conductivity, bedrock unit Rx1.
7	Ksd	.64	3.7E+00	1.0E+01	2.7E+01	ft/d	Multiplier of horizontal hydraulic conductivity, coarse-grained sediment group.
8	Rxk2v	.37	5.4E-03	1.0E+00	1.9E+02	ft/d	Vertical hydraulic conductivity, bedrock unit Rx2.
9	Rxk3v	.35	1.5E-02	1.0E-01	6.9E-01	ft/d	Vertical hydraulic conductivity, bedrock unit Rx3.
10	Ktill	.31	4.2E-02	1.0E+00	2.4E+01	ft/d	Horizontal hydraulic conductivity, till group.
11	Kwet	.17	1.8E-01	1.0E+01	5.6E+02	ft/d	Horizontal hydraulic conductivity wetlands.
12	Rxk4	.17	1.2E-02	2.0E-01	3.4E+00	ft/d	Horizontal hydraulic conductivity, bedrock unit Rx4.
13	Kmv	.13	3.1E-04	1.0E-02	3.2E-01	ft/d	Vertical hydraulic conductivity, fine-grained sediments.
14	Kwetv	.13	2.1E-04	1.0E+01	4.9E+01	ft/d	Vertical hydraulic conductivity, wetlands.
15	Rxk1v	.11	1.4E-03	5.0E-01	1.8E+02	ft/d	Vertical hydraulic conductivity, bedrock unit Rx1.
17	Km	.08	1.9E-06	1.0E-01	5.4E+03	ft/d	Horizontal hydraulic conductivity, fine-grained sediment group.
16	Hani2	.07	3.1E-03	1.0E+00	3.3E+02	ft/ft	Horizontal anisotropy, bedrock unit Rx4.
18	Ksdv	.06	6.8E-12	1.0E+00	1.5E+11	ft/d	Multiplier of vertical hydraulic conductivity, coarse-grained sediment group.
19	Rxk4v	.03	2.2E-06	2.0E-01	1.8E+04	ft/d	Vertical hydraulic conductivity, bedrock unit Rx4.
20	Ktillv	.02	9.3E-15	1.0E+00	1.1E+14	ft/d	Vertical hydraulic conductivity, till.
21	Rech2	.02	-3.8E-03	1.3E-04	4.0E-03	ft/d	Areal recharge in wetland areas.

The simulations were found to be insensitive to horizontal anisotropy in the anisotropy zone for bedrock parameter zone Rx4 (Hani2), and therefore, Hani2 was kept at 1:1. The hydraulic conductivity estimated for Rxk2 was larger (1.0 ft/d) than for other bedrock groups; the hydraulic conductivity of Rx2 may be more strongly anisotropic than that of the other bedrock groups. There was not enough observation data, however, to assess the anisotropies of each bedrock group (Rx1, Rx2, and Rx3) independently in the inverse process. The ground-water-flow system also was very sensitive to streambed conductivity (Ksb1). This parameter is a controlling factor in limiting ground-water discharge to streams; ground-water discharge is the primary component of discharge from the aquifer system. The multiplier of hydraulic conductivity of coarse-grained sediments (Ksd), and the horizontal conductivities of till (Ktill), and of bedrock groups Rx3 and Rx1 (Rxk3 and Rxk1) were equally important in the ground-water-flow process, and their values were estimated with narrow (less than an order of magnitude) confidence intervals. The parameter Ksd is a multiplier of the hydraulic conductivity determined from mapped transmissivity and saturated thickness (Moore, 1990; Stekl and Flanagan, 1992). The multiplier (Ksd) initially estimated in the steady-state inverse process was near 1, indicating that the mapped values used provided a reasonable approximation of the hydraulic conductivity of coarse-grained sediments in the stratified-drift aquifer systems of the regional model during low flow conditions. However, a final value of 10 was used (table 5–4) based on results of the transient model analysis discussed later.

The regional ground-water-flow system was not sensitive to vertical hydraulic conductivities, recharge on wetland areas (Rech2), and anisotropy in bedrock group Rx4 (Hani2). These parameters have orders-of-magnitude ranges between their lower and upper confidence intervals. This indicated that these parameters were not important in the regional ground-water-flow system and that the use of published values would be sufficient. Sensitivities were low also because there was not enough hydrologic information (observations) to characterize the parameters. Some parameters may have a low sensitivity because of the nature of the regional aquifer system. For example, the sensitivity of parameter Rech2, recharge on wetlands, is low because recharge on wetlands in the model as in the natural system is generally removed from the system by hydraulically connected stream networks. The low level of sensitivity of Rech2 indicated that recharge on wetlands can be assigned a range of values with little effect on the flow system.

Bedrock hydraulic conductivities at high-yield bedrock wells are obviously much higher than the regionally estimated values. Calculated heads for areas within a few model cells of large withdrawals will, therefore, not be realistic. The calculated heads for areas more than a few model cells away from the large ground-water withdrawals in the bedrock aquifer should be regionally realistic because flow in the regional bedrock aquifer is determined by regional bedrock aquifer properties and not individual fracture zones. Similarly, drawdowns measured at large ground-water withdrawals in the model area generally were localized and did not extend great distances from the wells because the hydraulic conductivity of the bulk-rock matrix is lower than that at a high-yield bedrock well. Although the flow to a high-yield bedrock well may originate from a few high-yielding fractures or fracture zones such as those observed in a high-yield well in Seabrook (Mack and others, 1998), the extents of individual fractures in crystalline rock, even around high-yield well fields (Johnson and others, 1999; Mack and others, 1998), generally are local (up to hundreds of feet). Fracture zones that extend thousands of feet generally consist of a collection of fractures, and the fractures may even be oriented en echelon or off the axis of the lineament. The flow to a fracture zone depends on connections to other fractures or fracture zones and the regional or bulk-rock hydraulic conductivity (Shapiro, 2004).

The parameter values estimated in the inverse process were fairly robust in that similar final values were estimated with different starting values. Although a high degree of confidence is associated with most of the parameter values, the estimated values and the confidence intervals depends on the conceptualization of the ground-water-flow system and the observation data used in the inverse modeling. The conceptualization should provide a realistic representation of the regional ground-water-flow system. With any model, however, different conceptualizations are possible which can result in different confidence intervals or estimated values. Most parameters were not highly correlated to other parameters (appendix 8) with the exception of Hani1 and Rxk2 (-0.77) and Ksdv and Rxk2v (-0.95).

Calculated Water Balance and Flows

The simulated water balance, a steady-state representation of fall 2004 flow conditions, is provided in table 5–5. With model layer 1 unconfined, some areas of the model dewater; these areas include topographic high areas, particularly those in till, or areas near withdrawal wells. Such dewatering occurs in the natural system particularly during seasonal low flows. Unconfined conditions were difficult to simulate because model cells representing withdrawals or returns may dewater, and a flux in a dewatered cell would incorrectly be eliminated from the simulated water balance. Although flow in the natural system includes unconfined conditions, particularly in the overburden represented in model layer 1, the calculations were done with all model

Table 5–5. Model-calculated October 2004 steady-state water balance and components of flow, in the Seacoast model, southeastern New Hampshire.

[Mgal/d, million gallons per day; flux, flow into or out of a model layer or component of the aquifer system; —, not available or not calculated; positive values are fluxes into a model layer, negative values are fluxes out of a model layer]

Model layer	Total flux into layer (Mgal/d)	Flux into layer (land area) (Mgal/d)	Recharge (Mgal/d)	Stream leakage into layer (Mgal/d)	Stream leakage out of layer (Mgal/d)	Net stream-flow (Mgal/d)	Constant head into layer (Mgal/d)	Constant head out of layer (Mgal/d)	Net tidal discharge (Mgal/d)	Wells (Mgal/d)	Distribu-ted water uses (Mgal/d)	Flux into layer compared to total (percent)	Flux into layer com-pared to recharge (percent)
Total	167.0	—	—	—	—	—	72.0	98.7	-26.7	-7.8	—	—	—
Layer 1	—	92.4	82.9	9.9	-58.8	-49.0	—	—	-12.1	-6.2	—	100	—
Layer 2	115.7	74.4	—	—	—	—	—	—	-13.4	—	2.2	69	90
Layers 3 and 4	73.6	—	—	—	—	—	—	—	[1]-12.3	-1.5	-2.7	44	89
Layer 5	14.5	—	—	—	—	—	—	—	—	—	—	9	17

[1] Net discharge from layer 3 (horizontal and vertical flow) to layer 2 at ocean bottom.

layers confined. However, the confined model was believed to represent natural conditions because flows calculated with model layer 1 unconfined differed from those calculated with all layers confined by only about one percent or less.

The net recharge applied to the steady-state ground-water-flow model was 77.7 Mgal/d, or equivalent to an areal rate of about 1.4 in/mo. The ground-water-flow system is rarely likely to be completely in a steady-state condition, but during seasonal low-flow conditions, the flow system may be relatively stable and in a quasi-steady state. The estimated recharge for the quasi-steady state includes some release of water from storage and, therefore, may be slightly higher than a transient recharge for the same period.

In the natural system, there were both streamflow gains and losses along streams in the study area; and the fluxes calculated in the model include flow into, out of, and between cells simulating constant-head and stream boundaries. Simulated flow at the stream boundaries is the sum of all streamflow gains and losses at multiple locations along the stream. For example, the total streamflow into the model was 9.7 Mgal/d while 53.2 Mgal/d flowed out at stream boundaries for a net streamflow of -43.5 Mgal/d (table 5–5). For this reason, the calculated total flux at a boundary, or in a layer, may not portray the regional ground-water-flow pattern very well. Net boundary flows are presented in table 5–4 and were evaluated when discussing the simulated water balance.

Simulated tidal water bodies were represented by a constant-head boundary. In the model, this boundary, where the thickness of saltwater was represented by an EFW head, is greater than zero. The thickness of saltwater in some areas of the model was large, more than 50 ft in some areas of the Piscataqua River and Great Bay; this thickness created fluxes that are not part of the ground-water-flow system within the constant-head cells in layer 1, representing ocean water. The lowest elevation constant-head cells are immediately adjacent to the shoreline in the ocean. This area forms the ultimate point of discharge in the simulation for both freshwater fluxes from the land area and open-water fluxes within the constant-head boundary. A net constant-head flux of 26.1 Mgal/d represented the regional freshwater discharge to tidal water bodies (table 5–5).

Table 5–5 presents the calculated flux into model layers which is balanced by flow out of the layer and fluxes at specified boundaries (recharge, constant heads, streams, wells, and specified fluxes). In general, the total flux into a layer indicates the amount of water moving through that layer. A large part of the total flux in model layer 1, however, is attributed to flow at the constant-head boundary. Of the 83 Mgal/d water recharging the aquifer system, about 90 percent of the recharge (74 Mgal/d) flows into the bedrock aquifer (layers 3 and 4) and 17 percent (14 Mgal/d) of the recharge flows into the lowest model layer (layer 5) representing deep bedrock areas. Surficial-aquifer wells withdraw about 7 percent (6.2 Mgal/d) of the total flow in the system, and bedrock wells remove about 5 percent (4.2 Mgal/d) of the total flow in the system. Simulated return flows, representing supplied-water returns (leach fields), were about 2.2 Mgal/d.

Because the Seacoast hydrologic system is one in which the surficial aquifers generally are thin and the underlying bedrock aquifer is highly fractured, most recharge at the surface would be expected to continue into the bedrock aquifer. In the model, only a small percentage of the recharge (17 percent), however, flows into the lower bedrock aquifer. The amount and rate of recharge to the bedrock aquifer depends on the recharge available at the surface, the degree of bedrock fracturing and its connectivity, and stresses (withdrawals) in the bedrock which would cause water to flow into the lower bedrock aquifer. Under natural conditions, one with no deep stresses, the rate of ground-water flow into the lower part of the aquifer system will be low compared to shallower parts of the aquifer (table 5–5). Flow into the bedrock aquifer under current conditions is likely to be greater than it had been in the past with fewer withdrawals.

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