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Appendix 7. Transient Model

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Appendix 7. Transient Model

The transient numerical model was developed to estimate storage characteristics in the aquifer system and evaluate transient ground-water-flow processes. Transient processes include seasonal storage depletion and recovery and assessment of ground-water movement from recharge to discharge. The transient model evaluated the seasonal effects of water use, recharge, and ET on ground water stored in the aquifer system; ET was assessed with the transient model only. This model also provided an analysis of mean monthly and 2003–04 specific monthly recharge and ET conditions. The use of different stress conditions, mean and specific monthly, provided confidence in the estimated model parameters. Parameter descriptions and abbreviations are given in table 7–1.

The transient model was similar to the steady-state model in that the grid and most boundary conditions, such as for streams and tidal water bodies (constant heads), were the same. The parameter zones were the same as in the previous model. Boundary conditions which differed from those used in the steady-state model were recharge, ET, and rates of specified fluxes (withdrawals and returns). Storage terms, including specific yield and specific storage, were assigned to the transient model in the same overburden and bedrock-parameter zones developed for the steady-state model. Transient parameter estimation was used to evaluate bedrock aquifer storage (Ss). Specific yield and specific storage in the surficial aquifers were not estimated, however, because these properties are relatively well known in comparison to those for the bedrock aquifer. Specific yield in the bedrock aquifer also was not estimated because of the probable limited extent and influence of this parameter in the bedrock aquifer.

The transient model consisted of a 4-year simulation developed to assess mean monthly conditions in the first 2 years of the simulation, and specific monthly conditions in the second 2 years of the simulation. Assessing mean monthly conditions is useful in assessing parameters for general or long-term simulations, whereas assessing specific monthly conditions provides more information on the potential range and variability of monthly model parameters. The first 2-year period was calibrated to mean monthly base flows and water levels calculated for the period from 2001 through 2004. The second 2-year simulation represents a calibration to actual monthly conditions observed in 2003 and 2004. The 2003–04 period was selected for detailed simulation because streamflow-gaging stations were operated during this period and detailed water-use data were available or estimated for this period.

Conditions used to initiate the model, particularly starting heads, affect subsequent simulated conditions; care in the selection of initial conditions will prevent the creation of artifacts later in the simulation. The simulation was initiated by using average monthly recharge estimates (fig. 7–1) and an initial head surface equal to the land surface. Model parameters, including monthly recharge, were estimated for the first 2-year simulation period on the basis of the mean of observations for 2001–04. The heads from the end of the first 2-year period were then used to reinitiate the simulation. This process—reinitiation of the model on the basis of the previous results of the 2-year simulation—was repeated several times until the effects of the initial conditions were not apparent and the simulation stabilized over the 2-year period. This stable simulation was then used to initiate the simulation for the second 2-year period; this simulation was calibrated to actual monthly observations for 2003 and 2004. The initial months of the second simulation, the first few months of 2003, were not well simulated because of the transition from mean monthly recharge parameters to actual monthly recharge parameters. Without observation data for a period longer than the 2 years of monthly observations, difficulties with this transition were unavoidable but were minimized by following this two-part simulation process.

This simulation could have also been conducted, but less efficiently, with two separate models which estimated parameters for the recharge conditions described above. Many of the model parameters, such as hydraulic conductivities, however, would have been the same in both simulations. For this reason, the mean monthly and the 2003–04 monthly periods were used in one simulation for more efficient parameter estimation. The time discretization was based on a monthly stress period with mean monthly stresses and observations. Major changes in boundary conditions, such as recharge, ET, and water use, generally occur by season, and monthly stress periods allow for gradations within seasons. The model consisted of 48 monthly stress periods each with two time steps of approximately 15 days in length. Shorter (more) time steps were investigated to provide smoother changes in boundary conditions; however, model improvement was slight with the expense of prohibitively long computation times during parameter-estimation simulations.

Table 7–1. Descriptions of transient parameters for the Seacoast model, southeastern New Hampshire.

[ft/d, feet per day; d.f., dimensionless factor or value]

Parameter name	Unit	Parameter description
Ktill	ft/d	Horizontal hydraulic conductivity, till group.
Ktillv	ft/d	Vertical hydraulic conductivity, till group.
Ksd	d.f.	Multiplier of horizontal hydraulic conductivity, coarse-grained sediment group.
Ksdv	d.f.	Multiplier of vertical hydraulic conductivity, coarse-grained sediment group.
Rxk1	ft/d	Horizontal hydraulic conductivity, bedrock group Rx1.
Rxk2	ft/d	Horizontal hydraulic conductivity, bedrock group Rx2.
Rxk3	ft/d	Horizontal hydraulic conductivity, bedrock group Rx3.
Rxk4	ft/d	Horizontal hydraulic conductivity, bedrock group Rx4.
Rxk1v	ft/d	Vertical hydraulic conductivity, bedrock group Rx1.
Rxk2v	ft/d	Vertical hydraulic conductivity, bedrock group Rx2.
Rxk3v	ft/d	Vertical hydraulic conductivity, bedrock group Rx3.
Rxk4v	ft/d	Vertical hydraulic conductivity, bedrock group Rx4.
Ksw	ft/d	Horizontal hydraulic conductivity, open water.
Kswv	ft/d	Vertical hydraulic conductivity, open water.
Ksb1	ft/d	Streambed hydraulic conductivity.
Km	ft/d	Horizontal hydraulic conductivity, fine-grained sediment group.
Kmv	ft/d	Vertical hydraulic conductivity, fine-grained sediment group.
Kwet	ft/d	Horizontal hydraulic conductivity, wetlands.
Kwetv	ft/d	Vertical hydraulic conductivity, wetlands.
Hani1	ft/d	Horizontal anisotropy, bedrock units Rx1, Rx2, Rx3.
Hani2	ft/d	Horizontal anisotropy, bedrock unit Rx4.
SSR1	d.f.	Specific storage, bedrock unit Rx1.
SSR2	d.f.	Specific storage, bedrock unit Rx2.
SSR3	d.f.	Specific storage, bedrock unit Rx3.
SSR4	d.f.	Specific storage, bedrock unit Rx4.
SSsd	d.f.	Specific storage, surficial sediment groups coarse: grained, till, marine.
SSwet	d.f.	Specific storage, wetlands.
R1	ft/d	Areal recharge, average January.
R2	ft/d	Areal recharge, average February.
R3	ft/d	Areal recharge, average March.
R4	ft/d	Areal recharge, average April.
R5	ft/d	Areal recharge, average May.

Table 7–1. Descriptions of transient parameters for the Seacoast model, southeastern New Hampshire.—Continued

[ft/d, feet per day; d.f., dimensionless factor or value]

Parameter name	Unit	Parameter description
R6	ft/d	Areal recharge, average June.
R7	ft/d	Areal recharge, average July.
R8	ft/d	Areal recharge, average August.
R9	ft/d	Areal recharge, average September.
R10	ft/d	Areal recharge, average October.
R11	ft/d	Areal recharge, average November.
R12	ft/d	Areal recharge, average December.
RCH25	ft/d	Areal recharge, January 2003.
RCH26	ft/d	Areal recharge, February 2003.
RCH27	ft/d	Areal recharge, March 2003.
RCH28	ft/d	Areal recharge, April 2003.
RCH29	ft/d	Areal recharge, May 2003.
RCH30	ft/d	Areal recharge, June 2003.
RCH31	ft/d	Areal recharge, July 2003.
RCH32	ft/d	Areal recharge, August 2003.
RCH33	ft/d	Areal recharge, September 2003.
RCH34	ft/d	Areal recharge, October 2003.
RCH35	ft/d	Areal recharge, November 2003.
RCH36	ft/d	Areal recharge, December 2003.
RCH37	ft/d	Areal recharge, January 2004.
RCH38	ft/d	Areal recharge, February 2004.
RCH39	ft/d	Areal recharge, March 2004.
RCH40	ft/d	Areal recharge, April 2004.
RCH41	ft/d	Areal recharge, May 2004.
RCH42	ft/d	Areal recharge, June 2004.
RCH43	ft/d	Areal recharge, July 2004.
RCH44	ft/d	Areal recharge, August 2004.
RCH45	ft/d	Areal recharge, September 2004.
RCH46	ft/d	Areal recharge, October 2004.
RCH47	ft/d	Areal recharge, November 2004.
RCH48	ft/d	Areal recharge, December 2004.

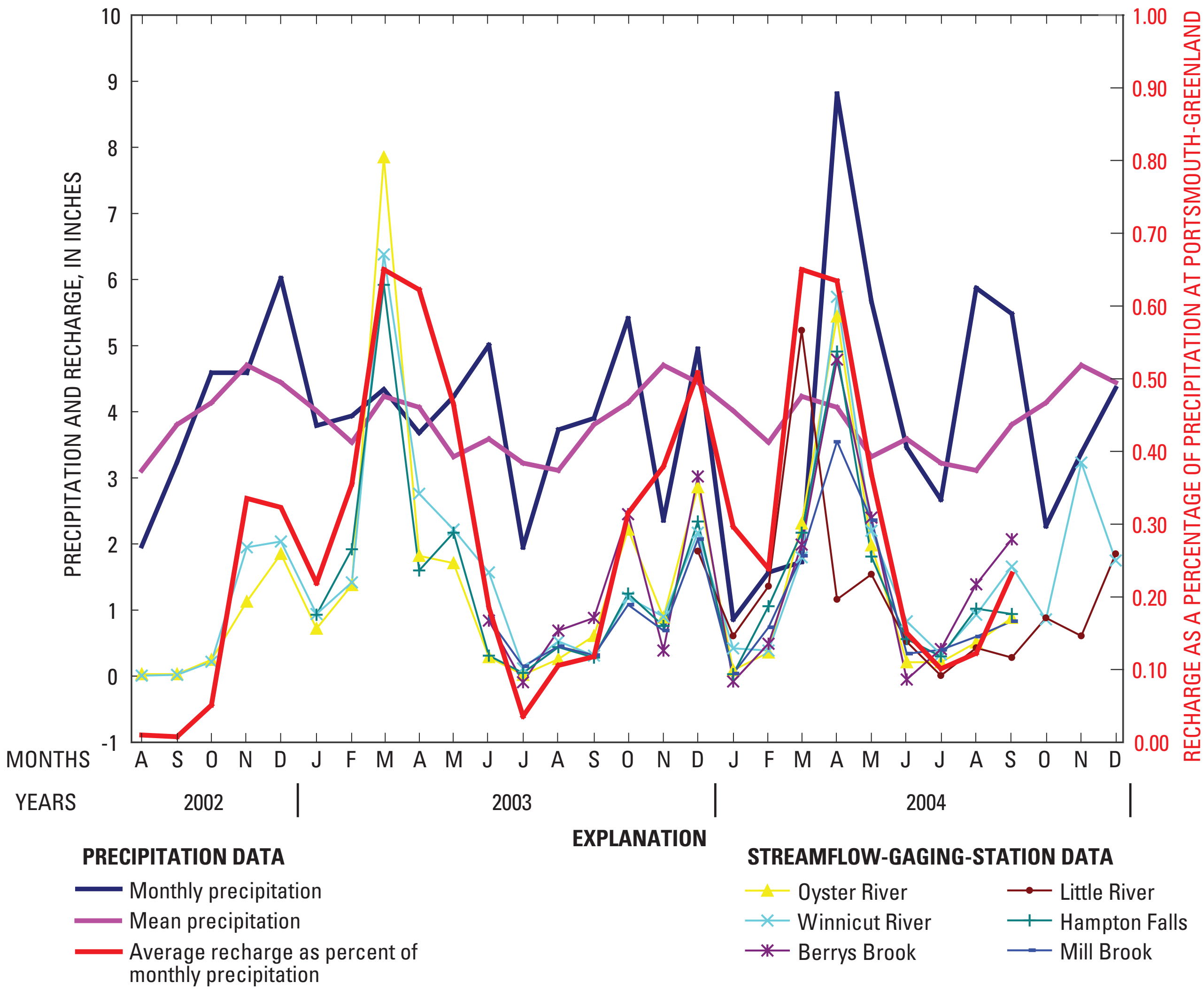


Figure 7–1. August 2002 to December 2004 monthly recharge estimated for Seacoast and Oyster River streamflow-gaging stations and precipitation in New Hampshire. (This figure is the same as figure 6 on page 15 in the report.)

Boundary Conditions

Withdrawals for the first 2 years of the 4-year simulation period were simulated on the basis of average monthly pumpages for 2003–04 registered withdrawals. For the calibration period 2003–04, actual monthly pumpages were used. As in the steady-state model, registered withdrawals were simulated as wells in the layer corresponding to the simulated withdrawal by the WELL package (Harbaugh and others, 2000). For nonregistered water uses, including domestic, community, industrial, and commercial, the rate of estimated water use or return was estimated by census block (Horn and others, 2007). Seasonally estimated returns and withdrawals were applied in the appropriate census block. These uses were distributed in the model as specified fluxes, by the Flow and Head Boundary (FHB) package (Leake and Lilly, 1997) and applied to cells within a 600-ft zone within the perimeter of each census block. Nearly all nonregistered withdrawals were from the bedrock aquifer and were simulated in model layers 3 and 4. Returns were simulated in model layer 2 to prevent returns from being inadvertently eliminated in areas where cells in model layer 1 were seasonally dry. Estimated withdrawal and return rates during winter were applied in December through January, while summer rates were applied in June through August. Average rates were applied in all other months.

Monthly average recharge (fig. 7–1) was used as an initial recharge in the model. Recharge was then refined for the two calibration periods by use of parameter estimation and observations of head and base flows. An observation data set was created for each calibration period, one of mean monthly observations and one of specific monthly observations. For the first calibration period, 12 recharge parameters (RCH1 through RCH12) representing a mean recharge for each month of the year were estimated on the basis of an observation data set consisting of the monthly means of estimated or observed head and base-flow observations for a recent 5-year period (2000–04). For the second calibration period, 24 recharge parameters (RCH25 through RCH48) were estimated to represent recharge in each individual month for the 2003–04 period. Actual monthly head and base-flow observations were used for the second calibration period. Estimation of physical parameters, such as hydraulic conductivities, is affected by all observations over the period represented by the simulation, whereas the parameter for a specific monthly recharge depends on antecedent conditions and observations for that month. In the parameter-estimation process, greater emphasis was given to calibration of the actual monthly recharge parameters than the mean monthly recharge parameters. The two sets of head and base-flow observations used in estimating recharge and other model parameters are described in the observation data section below.

ET from April through October in wetland areas was assessed by the EVT package (Harbaugh and others, 2000). In the steady-state model, ET is effectively accounted for in that the simulated annual recharge is a net recharge to the aquifer, that is, the infiltration of precipitation minus ET. ET was initially simulated in the model only in the wetland areas where it is likely to be larger than ET in other model areas, and the head is more likely to be simulated close to the land surface. Because some areas of the model represented considerable land-surface relief and simulated heads were less accurate in these areas, heads were not simulated well enough with respect to the land surface for the required ET extinction depth (generally about 6 ft) to be accurately represented. The sensitivity analyses, however, showed that ET rates could not be estimated by applying estimation to the available data because there was no unique information (observation data) to distinguish ET from net recharge. The base-flow observation data reflects the net recharge, or effective recharge accounting for ET. Therefore, ET was incorporated in the monthly net-recharge terms in the model. For this reason some estimated net recharge rates for summer months were negative; this result indicated that ET was greater than the infiltration of precipitation during those months. In an investigation about 50 mi south of the study area, DeSimone (2004) calculated a wetland ET rate of 29.4 in/yr (-6.7×10^{-3} ft/d) and simulated wetlands connected to streams as areas of no recharge. In the transient simulation reported here, net recharge on wetlands was simulated as zero which, on an annual basis over the entire model area, reduced the total simulated recharge flux by about 5 percent.

The use of 36 recharge parameters, although physically reasonable, resulted in the individual recharge parameters having a low sensitivity in the parameter-estimation process. Parameter insensitivity resulting from a highly parameterized model has been noted by Randall Hunt (U.S. Geological Survey, written commun., 2007). The recharge parameterization was retained, however, to allow for greater model utility. It also is important to note that for some months with similar ET conditions and total precipitation, the calculated base flows differed. For example, August and September 2003 were months with similar monthly precipitation totals (3.7 and 3.9 in.; fig. 7–1); however, hydrograph separations and calibration results indicate that monthly net recharge was much greater in August than September. Although the total precipitation for the two months was similar, daily precipitation data show that September had fewer and more concentrated periods of precipitation. The largest total daily precipitation for September 2003 was 1.7 in. compared to 1.04 in. for August 2003. Concentrated precipitation events results in less infiltration and less base flow.

Base Flows

Monthly base flows were needed for initial estimation of recharge, and observations of base flows and heads were needed for the transient parameter-estimation process. Monthly base flows were estimated from analysis of a 5-year synthetic streamflow record and 2003–04 streamflows. Continuous streamflows were measured at stations in the model area beginning at different times in 2002 and 2003 (table 7–2). Synthetic streamflow records for the model area continuous stations were extended back to January 1, 2000, by the Maintenance of Variance Extension Type 1 (MOVE.1) method (Hirsch, 1982) by developing correlations with records from the Oyster River streamflow-gaging station. The synthetic discharge records created for the period from January 2000 through December 2004 were used to estimate average monthly base flows for this period with the base-flow portioning program PART (Rutledge, 1993, 1998). The program RORA (discussed previously in the Recharge section) was applied to the extended record for the Winnicut River to estimate an initial average monthly recharge for the study area for the 2001–04 period.

Continuous streamflows were measured at five streamflow-gaging stations in the model area beginning in July 2002 at the Winnicut River station and in 2002 and 2003 at other stations until September 30, 2004 (table 7–2). Record extension was used to generate mean daily streamflows for the five streams from 2000 through 2004. The MOVE.1 method (Hirsch, 1982) was used if the logarithm of streamflows at a short-term streamflow-gaging station is linearly correlated to streamflows at a long-term station. Ries (1994) indicated that MOVE.1 can be used for record extension if the linear correlation coefficient between the two sets of concurrent streamflows is at least 0.80. The correlation coefficients between the logarithms of concurrent streamflows in the study area and of the Oyster River flows were Winnicut River, 0.93; Little River, 0.89; Mill Brook, 0.88; Berry's Brook, 0.83; and Hampton Falls, 0.92. These coefficients for the log-linear relations indicated that record extension was suitable for use in the study area. The MOVE.1 relations reflect current water uses, however, and are only suitable for use with streamflow records collected under similar water-use conditions. Changing water-use conditions affect streamflows, and record extension may not be suitable for other periods. The favorable comparison of streamflows generated by MOVE.1 relations to the measured streamflows for the Winnicut River are likely a result of similar water-use conditions throughout the period 2002–04 (fig. 7–2).

Average monthly base flow for the period 2000–04 were calculated by partition of streamflow records at the five stations. Calculated base flows converted to a linear rate per watershed area (in/mo) were compared to monthly average (2000–04) and monthly (2003–04) total precipitation for Portsmouth-Greenland (fig. 7–3). Base flows calculated by this technique for the months of March and April in 2003 and for the 5-year average for March and April were equivalent to or slightly greater than the corresponding precipitation rates. In other months, base flows were not more than 65 percent of the corresponding monthly precipitation. The base-flow partition technique was obviously not reasonable for these periods. Cautions on the use of streamflow-partition techniques for a monthly time scale were given by Rutledge (1993, 2000). A comparison of techniques for a similar setting in Pennsylvania, however, indicated that monthly base-flow estimates by streamflow partition were reasonable (Risser and others, 2005). For the monthly periods in this study when the calculated base-flow rate was more than 65 percent of the monthly precipitation, the base-flow rate was reduced to 65 percent of the precipitation rate. This condition occurred during one other month in the 5-year calculated base-flow record (April 2001), but in this case, the high base flow was influenced by high flows during the preceding month (effects of monthly discretization). Individual monthly results for this period were not used in the observation data for the model; averages were used for this period. Monthly annual base flows for 2000–04 are provided in table 7–3 to provide additional insight on apparent hydrologic conditions in the five watersheds during the investigation.

The calculated annual total base flows portray individual watershed characteristics and regional climatic conditions in the study area or climatic conditions as reflected by streamflows at the Oyster River watershed. The drought of 2002 is apparent in the calculated base flows for 2002, which were about 75 percent of the 5-year average for 2000–04 conditions (table 7–3). Base flows in 2003 (table 7–3) were about 10 percent greater than average, and 2004 base flows were average. Base flows in the study area generally were a little more than 13 in/yr; however, base flows in Little River were lower than in other watersheds. The lower flows probably reflect a greater water use per unit area than in the other watersheds. Flows in the Berry's Brook and Winnicut River watersheds were slightly higher than in other watersheds. In general, the calculated base flows were about 30 percent of precipitation with the exception of 2002 when base flow was about 22 percent of precipitation. More precipitation (46 in.) fell in 2002 than in the preceding year 2001 (40 in.), yet the base flows for 2002 were less than 2001. This observation indicates that, during periods of intense precipitation, a considerable volume of water may bypass the Seacoast aquifer system without becoming recharge. For example, removing the anomalous peaks from the March and April 2003 base-flow calculations resulted in a 15-percent reduction in total annual recharge for that year. This effect would be more pronounced for short periods of high precipitation (storms).

Table 7–2. Streamflow characteristics for streams in the Seacoast model area and for Oyster River, southeastern New Hampshire. (This table is the same as table 3 on page 25 in the report.)

[Site numbers shown on figure 2 unless otherwise indicated; C, continuous-record streamflow-gaging station; P, partial-record streamflow-gaging station; mi^2, square miles; ft^3/s, cubic feet per second; —, not available]

Site number	Station number	Stream	Location	Area (mi^2)	Station code	Mean flow 2004 (ft^3/s)	Calculated annual base flow 2004 (ft^3/s)	Streamflow measured October 7–8, 2004 (ft^3/s)	Streamflow measured January 4, 2004 (ft^3/s)	Base flow estimated January 4, 2004 (ft^3/s)	Base flow estimated January 27, 2004 (ft^3/s)
	01073000	Oyster River	Durham (not in model area)	12.1	C	19.4	16	4.1	24	10.6	3.5
9		Parkman Brook	Portsmouth	1.91	P	—	—	[1] 1	—	—	—
1	01073750	Mill Brook	Route 108, Stratham	2.48	C	—	2.5	1.00	4.9	2.3	.37
4		Pickering Brook	Shattuck Way, Newington	—	P	—	—	.095	—	—	—
6		Hodgson Brook	Cate Street, Portsmouth	3.52	P	—	—	1.18	—	—	—
3		Packer Brook	Ports Avenue, Greenland	2.25	P	—	—	.56	—	—	—
7		Pickering Brook	Ports Avenue, Greenland	2.97	P	—	—	.61	—	—	—
14-1	01073785	Winnicut River	Route 33, Greenland	14.19	C	22.3	19.3	7.3	27	10.3	2.4
14-2		Winnicut River	Winnicut Road, Stratham	—	P	—	—	4.94	—	—	—
14-3		Winnicut River	Walnut Road, North Hampton	—	P	—	—	3.47	—	—	—
14-4		Winnicut River	Route 111, North Hampton	—	P	—	—	1.71	—	—	—
15	01073810	Berrys Brook	Sagamore Road, Rye	5.38	C	9.8	7.6	3.1	8.1	6	.75
5		Bailey Brook	Love Lane, Rye	1.73	P	—	—	1.09	—	—	—
11-0	01073822	Little River	Woodbury Road, North Hampton	6.12	C	8.5	5.9	2.5	8.5	4.4	.4
11-1		Little River	Unnamed tribuary, North Hampton	—	P	—	—	.30	—	—	—
16		Nilus Brook	North Shore Road, Hampton	1.5	P	—	—	.61	—	—	—
13		Taylor River	Old Stage Road, Hampton	8.41	P	—	—	4.83	—	—	—
10		Great Brook	Giles Road, Brentwood	5.5	P	—	—	1.82	—	—	—
8-0		Hampton Falls	Route 1, Hampton Falls	6.66	P	—	—	3.68	—	—	—
8-1	01073848	Hampton Falls	Mill Lane, Hampton Falls	3.61	C	6	4.6	1.72	8	3.3	.72
2		Back River	Amesbury Road, South Hampton	1.53	P	—	—	.535	—	—	—
17		Cains Brook	Route 1, Seabrook	—	P	—	—	[1] 1	—	—	—
12		Smallpox Brook	True Road, Salisbury, Mass.	1.83	P	—	—	1.00	—	—	—

[1] Flow estimated on the basis of previous measurements or qualitative information.

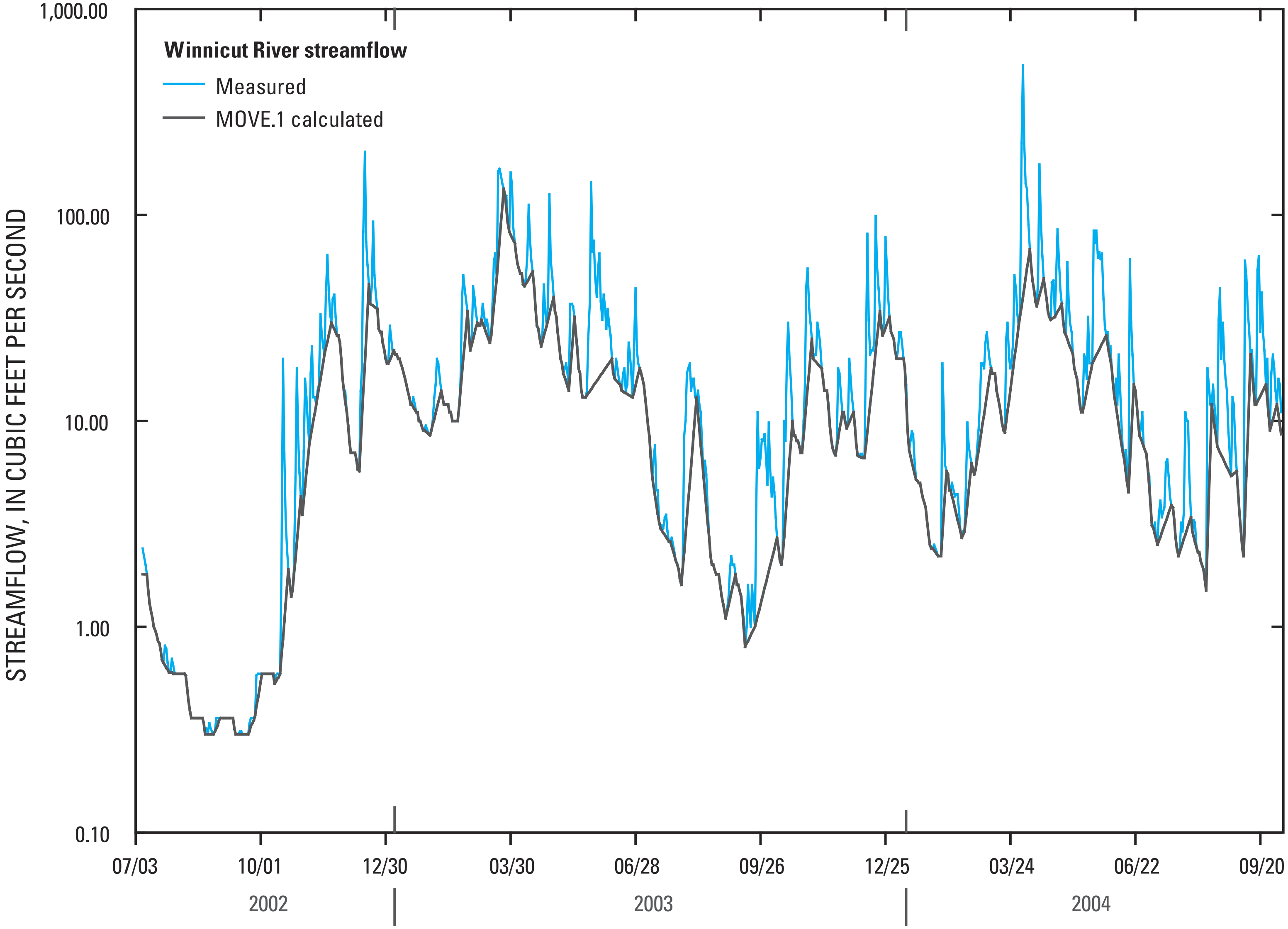


Figure 7–2. Measured and calculated MOVE.1 daily mean streamflows for 2002–04 for the Winnicut River in the Seacoast model area, southeastern New Hampshire.

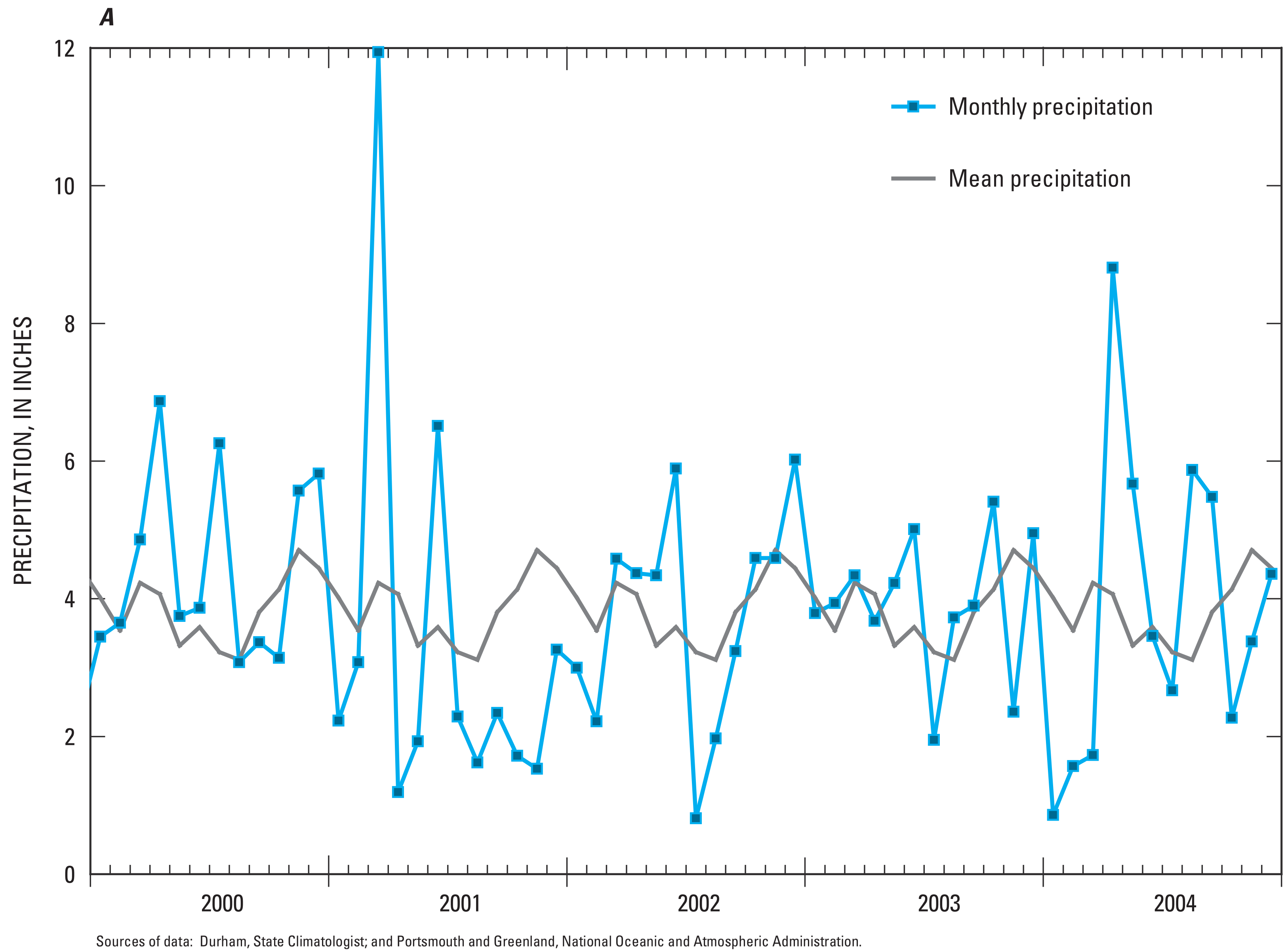
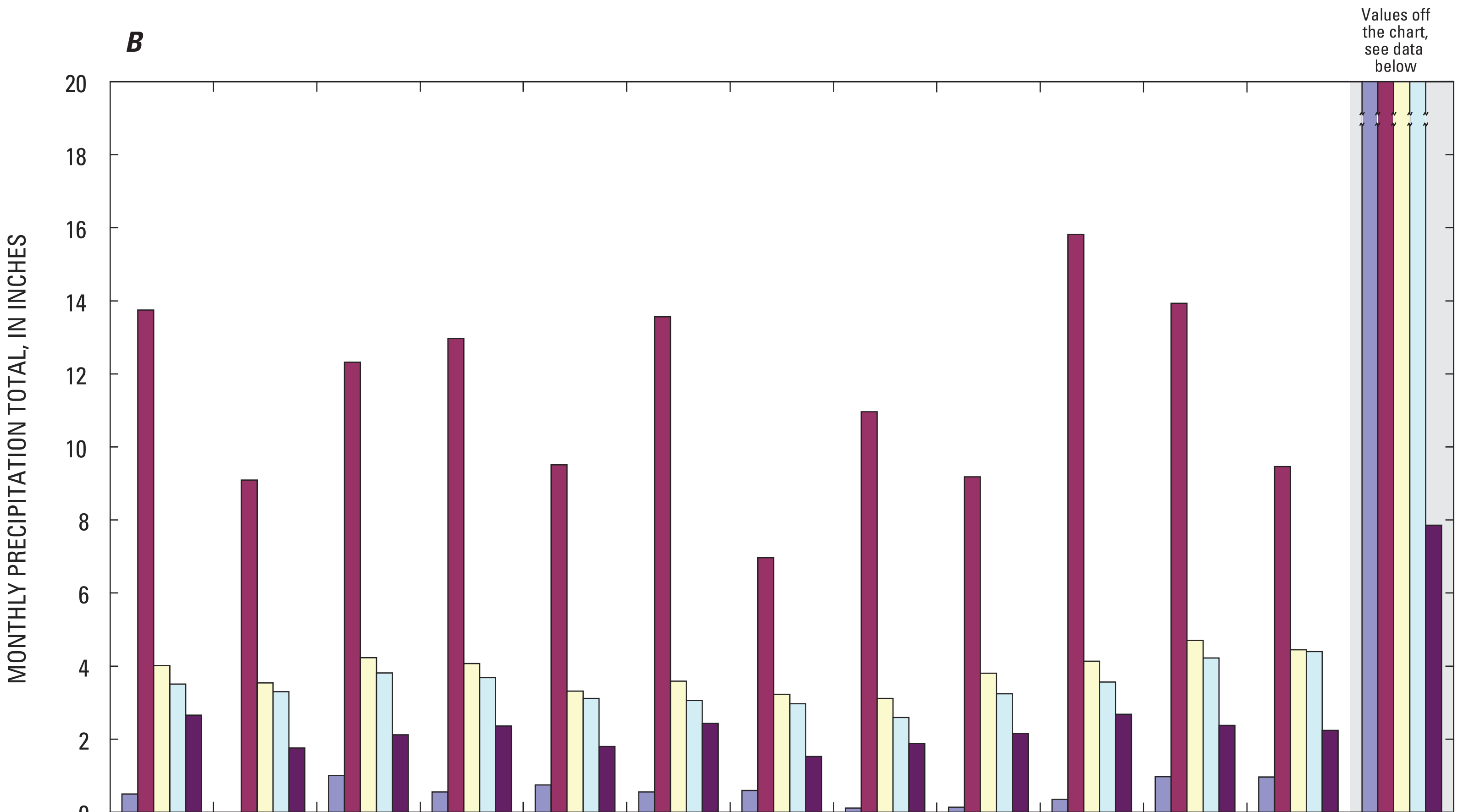


Figure 7–3. *(A)* Monthly and mean precipitation at Portsmouth and Greenland, New Hampshire, from January 2000 through December 2004. (This figure is the same as figure 1–2 on page 56–57 in appendix 1.)



	JAN.	FEB.	MAR.	APR.	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	NOV.	DEC.	Annual
MINIMUM	0.5	0	1	0.55	0.75	0.55	0.59	0.11	0.14	0.35	0.97	0.96	29.19
MAXIMUM	13.75	9.09	12.32	12.97	9.51	13.56	6.97	10.96	9.18	15.82	13.93	9.46	71.26
MEAN	4.02	3.54	4.23	4.07	3.32	3.59	3.22	3.11	3.81	4.14	4.71	4.44	45.37
MEDIAN	3.51	3.3	3.815	3.68	3.11	3.06	2.97	2.59	3.24	3.56	4.22	4.4	44.84
STD. DEVIATION	2.66	1.75	2.12	2.36	1.80	2.43	1.53	1.88	2.16	2.68	2.38	2.24	7.86

Sources of data: Durham, State Climatologist; and Portsmouth and Greenland, National Oceanic and Atmospheric Administration.

Figure 7–3. *(B)* Monthly precipitation statistics at Portsmouth and Greenland, New Hampshire, from 1955 through 2005.—Continued (This figure is the same as figure 1–2 on page 56–57 in appendix 1.)

Table 7–3. Mean monthly and annual average base flows calculated for selected watersheds in the Seacoast model area, southeastern New Hampshire.

[ft^3/s, cubic feet per second; $ft^3/s/mi^2$, cubic feet per second per square mile]

Stream	January (ft^3/s)	February (ft^3/s)	March (ft^3/s)	April (ft^3/s)	May (ft^3/s)	June (ft^3/s)	July (ft^3/s)	August (ft^3/s)	September (ft^3/s)	October (ft^3/s)	November (ft^3/s)	December (ft^3/s)	Annual total (ft^3/s)	Annual total ($ft^3/s/mi^2$)
								2000–04 average						
Winnicut River	10.32	11.92	33.98	37.64	20.45	12.75	3.20	3.40	3.13	4.88	12.17	17.32	171.2	12.1
Little River	3.32	3.93	12.02	11.92	6.32	3.65	.87	1.17	.80	1.51	3.31	4.98	53.8	8.8
Berrys Brook	4.00	5.01	16.48	15.53	8.56	4.95	1.17	1.66	1.01	2.31	4.11	6.90	71.7	13.3
Mill Brook	1.73	2.09	4.93	4.78	3.43	2.15	.67	.59	.50	.74	1.53	2.49	25.6	11.1
Hampton Falls River	2.74	3.43	8.57	8.24	5.09	3.12	.83	.82	.72	1.06	2.43	3.93	41.0	11.4
								2003						
Winnicut River	14.53	15.82	34.61	33.12	23.40	20.49	5.79	5.05	1.91	7.27	14.00	20.20	196.2	13.8
Little River	6.10	4.82	11.71	8.44	4.88	5.05	1.11	2.28	0.93	1.86	3.89	6.42	57.5	9.4
Berrys Brook	5.60	6.04	19.61	16.55	9.89	7.67	1.63	2.33	1.40	6.81	6.56	10.64	94.7	17.6
Mill Brook	2.25	2.41	4.82	4.81	4.33	2.87	.62	.98	.49	.90	1.69	2.75	28.9	12.6
Hampton Falls River	3.79	4.19	7.74	6.65	5.54	4.08	.66	1.22	.58	1.50	2.94	4.95	43.8	12.1
								2004						
Winnicut River	9.24	4.64	17.61	51.04	23.53	11.33	3.45	7.14	10.69	10.84	15.15	30.55	195.2	13.7
Little River	1.86	1.53	6.21	12.78	8.65	3.02	1.22	2.18	2.25	3.82	4.33	8.44	56.3	9.2
Berrys Brook	3.22	1.91	7.05	10.56	9.85	3.91	1.59	4.15	2.60	2.29	4.44	11.90	63.5	11.8
Mill Brook	1.40	1.28	2.43	5.67	3.97	1.90	.96	.80	1.18	1.24	1.71	4.19	26.7	11.6
Hampton Falls River	2.32	2.50	4.85	11.00	5.95	2.65	1.10	1.35	2.01	1.75	2.88	6.73	45.1	12.5

Ground-Water Levels

The model also was calibrated to monthly ground-water levels. There were very few monitoring wells with long-term continuous water-level data in the region. Continuous water levels were collected for varying periods between October 2003 and August 2005 at five monitoring wells in the study area by (HEW-44, HEW-45, SSW-7, GTW-141, GTW-157) (figs. 7–4 and 7–5). The longest water-level record in the study area consists of monthly measured water levels, from July of 1997, to the present (2005) at a bedrock (SSW-248) and overburden (SSW-249) well pair in Stratham (Raymond Talkington, Geopshere, Inc., written commun., 2006). Study-area water-level records and records from other nearby wells were extended to generate monthly synthetic water levels at the continuously monitored wells for additional periods. Water levels at wells SSW-248 and SSW-249 and long-term USGS wells generally were measured near the end of the month, and measurements within a week of each other were considered concurrent. Comparisons of the monthly measurements at SSW-248 with same-day readings at wells monitored for this study showed poor relations (R^2 less than 0.50). Only 7 to 14 concurrent monthly observations were made, however, at the monitoring wells during the study period.

The quality of relations between the records of the Stratham wells and the records of long-term monitoring wells outside the study area were assessed for use in extending water-level records. Good linear relations (R^2 greater than 0.70) were found in a comparison of the full record for SSW-248 (1997 to present) to observations at long-term monitoring wells (HTW-5, LIW-1, DDW-46, and PBW-148; Keirstead and others, 2005) with about 40 to 80 concurrent observations. Records for the well closest to the study area with a long-term water-level record, LIW-1 in Lee (1953 to present), and for the Stratham wells linearly correlated ($R^2 = 0.74$). The Lee well (LIW-1) is screened in well drained highly conductive outwash deposits; consequently, the water-level record fluctuates little from climatic events. A slightly better correlation ($R^2 = 0.77$) was found between the record for DDW-36, a well screened in stratified drift about 18 mi to the west, and the records for the Stratham wells. Coefficients of determination were 0.87 between the records for SSW-248 and PBW-148 and 0.82 between the records for HTW-5 and PBW-148. These analyses indicated that regional climatic patterns and well responses in southeastern New Hampshire permit record extension by using the records of PBW-148. Both HTW-5 and PBW-148 bedrock wells are about 100 ft deep. Well HTW-5, with more than 30 years of monthly record, is about 25 mi west of the Stratham wells, and PBW-148, with 5 years of continuous daily record, is about 30 mi west. On the basis of concurrent water-level measurements between 1997 and 2005, MOVE.1 relations were developed between the records of monitoring wells HTW-5 and SSW-248 to estimate water levels at the Stratham well for 10 months of missing record in 2002 and 2003.

Although the Stratham well record is valuable, it included too few observations to provide correlations with other wells. Because of the limited availability of concurrent monthly water levels for monitoring wells in the study area, relations with continuous records outside the study area were assessed to provide the necessary additional records. Available water-level records from monitoring-well networks were used for the 2003–04 period. Most wells in these networks were associated with specific large ground-water-withdrawal permits, however, and were clustered in a few areas of the model. Few transient water levels were available in records for wells outside of the network areas; as a result, initial calibrations indicated that the model was insensitive to some parameters. Surface-water levels were used to provide about 300 head observations for model layer 1. The observations generally were made in discharge areas with little seasonal surface water-level change where the water table is at or near land surface. Surface-water levels were not used as head observations for the months of June, July, and August because ground-water levels may drop below the topographic altitudes of stream surfaces (representing a dry stream) particularly at the higher altitudes.

Water levels in stratified-drift aquifers generally show very little seasonal variation (less than 5 ft), whereas water levels in till aquifers generally have an annual range of about 10 ft. Twenty-eight synthetic observation points were created on the basis of the depth-to-water ranges (about 10 ft) and patterns of till observation wells HEW-45 and SSW-7. Points were centrally placed in till bodies between streams without water-level observations and far from large registered withdrawals. Synthetic observation points also were created at the same locations using the seasonal (unstressed) bedrock-aquifer water-level patterns observed at HEW-44 and GTW-141.

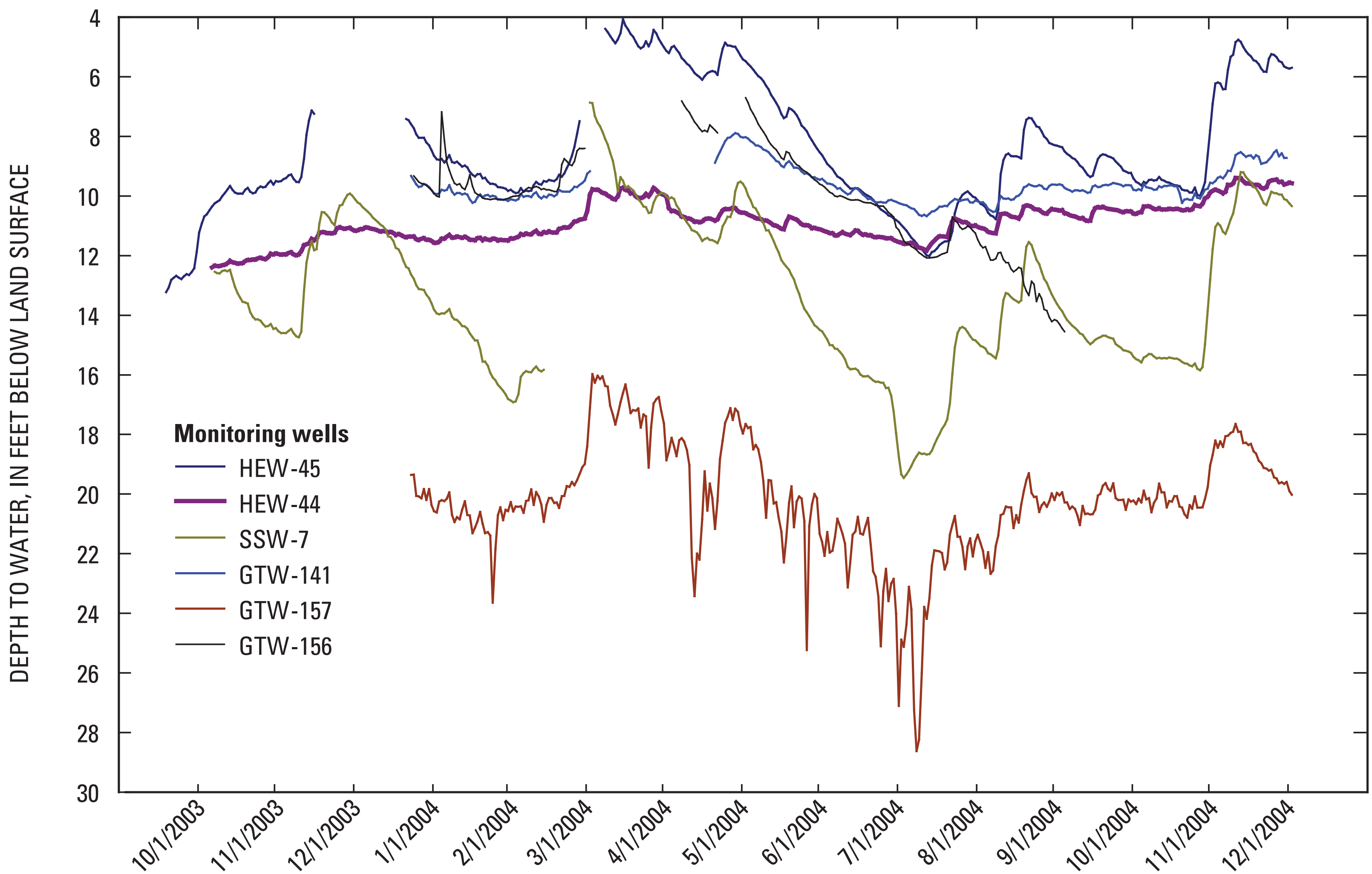


Figure 7–4. Daily mean depth to ground water in six monitoring wells in the Seacoast model area, southeastern New Hampshire, between October 2003 and January 2004. Areas of no data (data gaps) are missing record. (Well locations shown in figure 7. This figure is the same as figure 9 on page 20 in the report.)

Specific Storage

Initial values of specific storage were obtained from literature and other nearby investigations. Values of specific storage are relatively well known for surficial (glacial) aquifers, whereas the storage characteristics of crystalline bedrock aquifers throughout the Northeast, including the Seacoast, are not well known. It was necessary to estimate specific storage parameters with all model layers confined in order to simplify the parameter estimation calculations. Additionally, all surficial-aquifer properties were held constant during calibration simulations because they were better known than bedrock storage. Specific storage values determined from aquifer tests in glacial sediments (Randall, 2001) were typically about 1×10^{-4}. Imposing confined conditions on the surficial aquifers, however, can cause an unrealistic amount of water to be released from the surficial aquifers during transient seasonal (summer) simulations. For this reason DeSimone (2004) used specific storage value of 2.5×10^{-6} for till in seasonal simulations. Use of a low specific storage value for surficial sediment was tested in this investigation but was found not to be necessary; a value of 1×10^{-4} was used for all surficial sediments.

An investigation of the properties of a similar crystalline bedrock in North Carolina (Daniel and others, 1997) found that transmissivity, specific storage, and well yield varied in a consistent manner. In that investigation, it was found that well yield could be used as an index for specific storage. Specific storage in the Seacoast bedrock aquifers differs with the overall fracturing and connectivity as indicated by the well yields of each formation (appendix 2). The differences in bedrock well-yield probability observed in the Seacoast bedrock aquifer (Moore and others, 2002) are likely also to indicate differences in specific storage. This indication is supported by the fact that the Kittery Formation and Rye Complex have sustained several large ground-water withdrawals, whereas the formations to the west have been able to sustain fewer large withdrawals despite exploratory efforts by water suppliers. For this reason, the same four zones (fig. 7–6) used for other bedrock characteristics were used for parametization of specific storage in the bedrock aquifer. The bedrock aquifer was considered confined in all cases, and specific yield of the bedrock aquifer was not examined. Lyford and others (2003) estimated a specific storage of 1×10^{-5} for a high-yield bedrock formation (the Eliot Formation) on the basis of a long-term (30-day) aquifer test in West Newburyport, Mass. From observation of regional bedrock-well yields, the specific storage of the Eliot Formation in the vicinity of the West Newburyport well field is likely to be greater than the regional specific storage of the Eliot Formation in the Seacoast model area.

To limit inverse-model simulation time, specific-storage parameters for the bedrock aquifers were estimated with all other parameters held constant. The model was updated with the estimated value for specific storage, and then other model parameters, such as recharge and hydraulic conductivities, were re-estimated. Final specific storage values used in the model were approximately 8×10^{-7} (SSR1), 3×10^{-6} (SSR2), and 1×10^{-7} (SSR3). Because there were not enough observation data to estimate storage for bedrock-aquifer zone Rx4, and it was believed to be low, a value of 1×10^{-7} (SSR4) was used. It is important to note that the estimated specific storage values represented a bulk characteristic rather than the specific storage of individual fracture zones. Specific storage in the immediate vicinity of a bedrock well with a sustainable high yield is likely to be greater, possibly an order of magnitude greater, than the regional specific storage.

Estimated Model Parameters

Estimated model parameters are listed in table 7–4 along with 95-percent confidence intervals and sensitivities are shown in figure 7–7. Model parameters, confidence intervals and sensitivities reflect not only the observation data and the weights placed on data, but also the conceptualization and parameterization of the ground-water-flow system. The parameter values were fairly unique in that similar values were calculated from different parameter starting values. Nonunique parameters were indicated by correlation coefficients between two parameters greater than 0.95 (appendix 8).

The problem of conceptual model uncertainty, and the fact that conceptual models are often changed because of additional data, is discussed by Bredehoft (2005). A principal of parsimony (Hill, 1998) was followed according to which model parameters were defined and zoned on the basis of physically defensible units such as major bedrock formations (fig. 7–6) or surficial-sediment groups (fig. 7–8). This approach lends confidence that the model was not overparameterized to fit observations rather than simply created to fit the available data. Use of fewer parameters is preferred in terms of model stability and parameter correlation. Some parameter zones may be too simple, however, and may not reflect important subzone hydraulic variations. For example, the bedrock zones used do not reflect variations within each bedrock unit, and the case could be made for increased parameterization based on regularized inversion techniques (Hunt and others, 2007).

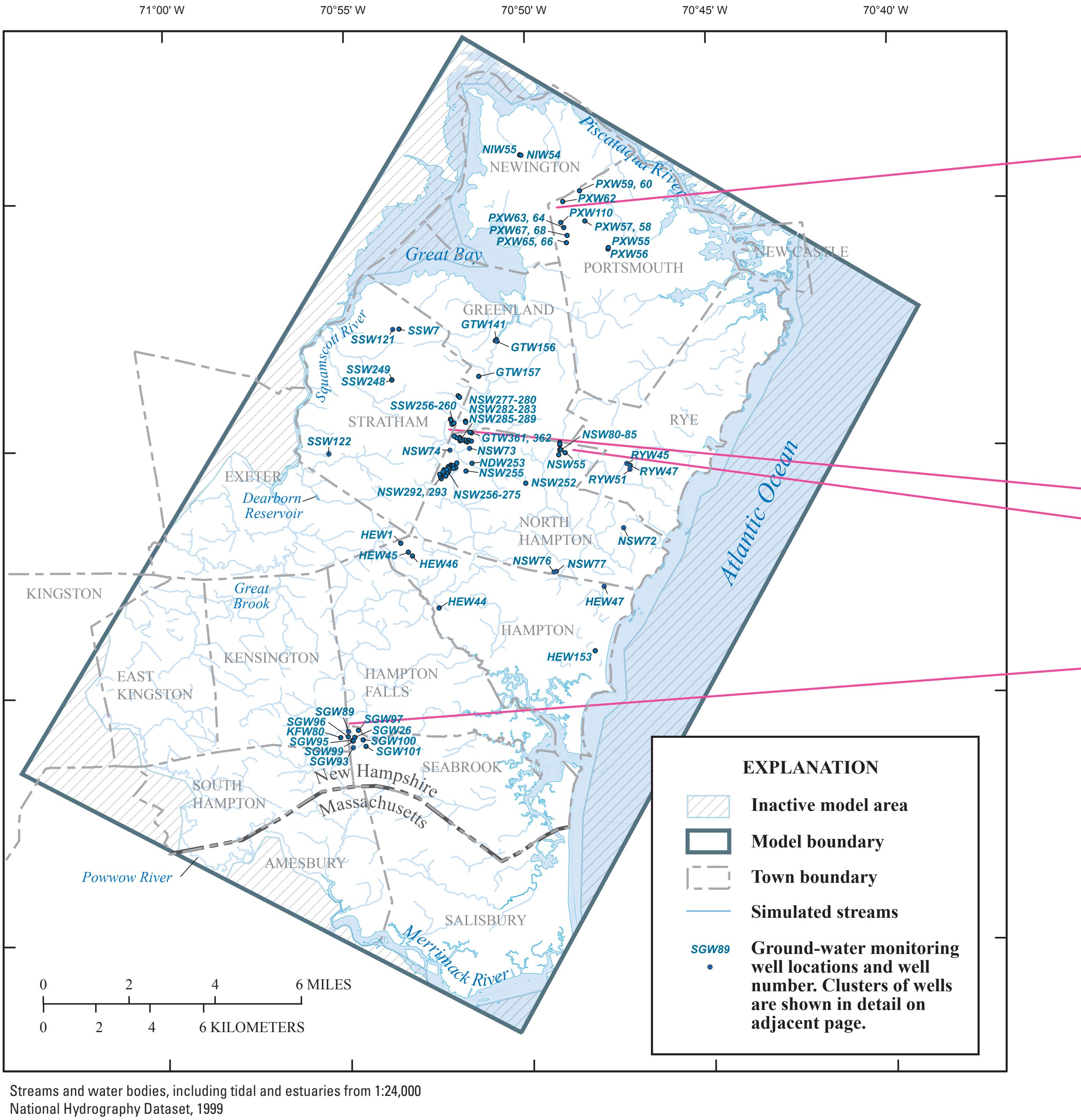
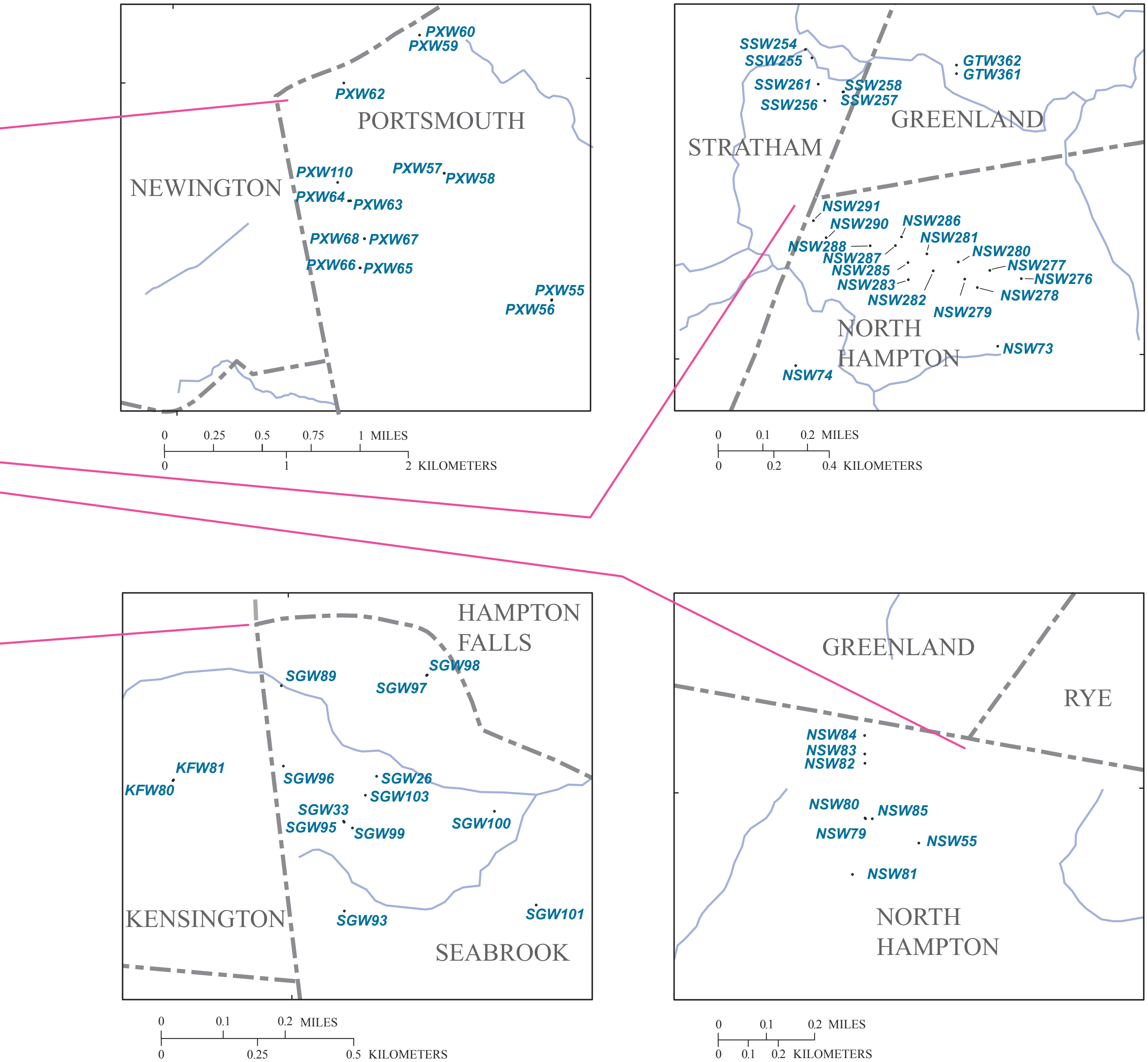
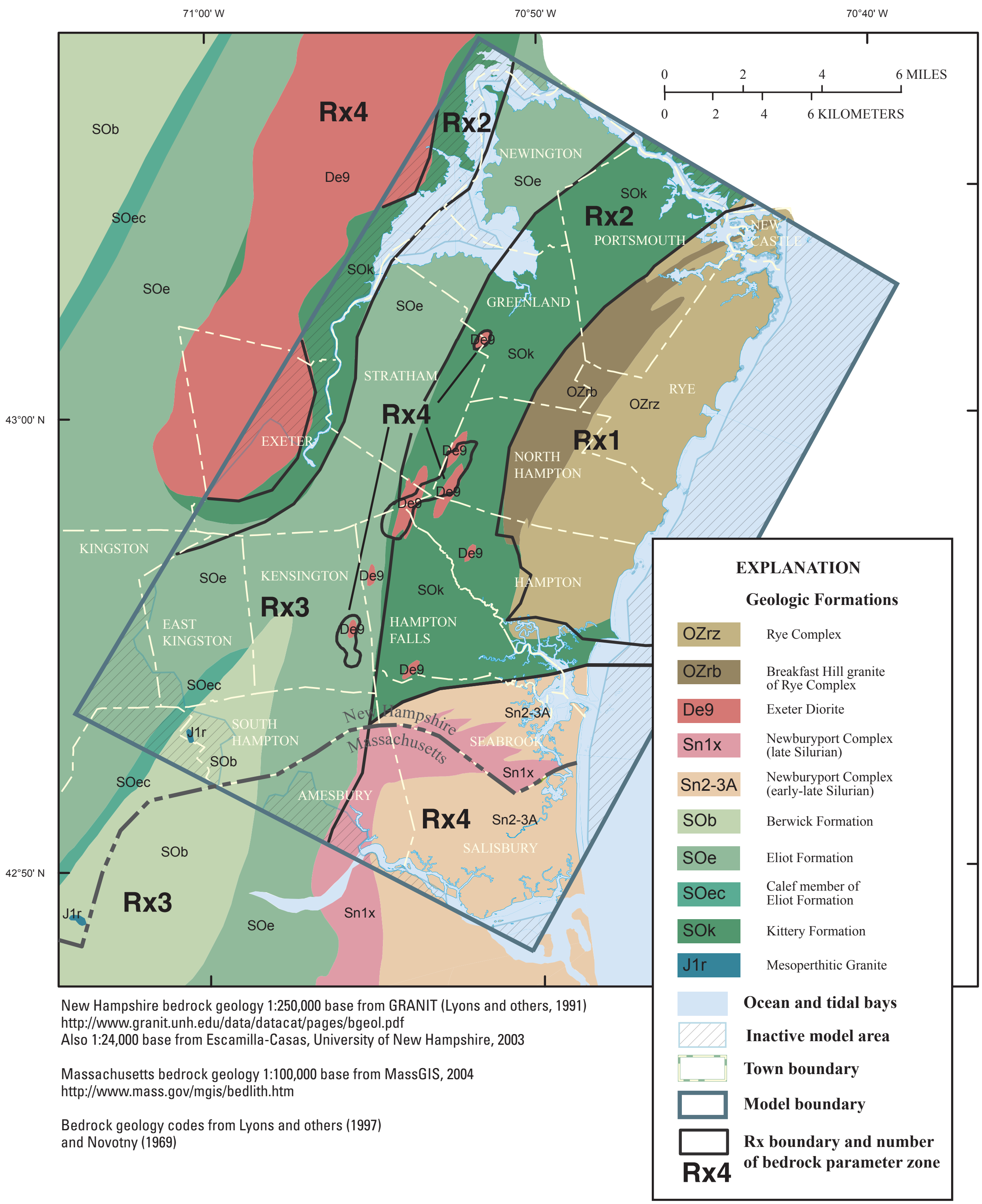


Figure 7–5. *(A)* Ground-water monitoring wells in the study area. (This figure is the same as figure 7 on page 16–17 in the report.)


PXW60
PXW59
PXW62
PORTSMOUTH
NEWINGTON
PXW110
PXW57
PXW58
PXW64
PXW63
PXW68
PXW67
PXW66
PXW65
PXW55
PXW56
0 0.25 0.5 0.75 1 MILES
0 1 2 KILOMETERS
SSW254
SSW255
GTW362
GTW361
SSW261
SSW258
SSW256
SSW257
GREENLAND
STRATHAM
NSW291
NSW290
NSW286
NSW281
NSW288
NSW287
NSW280
NSW285
NSW277
NSW276
NSW283
NSW278
NSW282
NSW279
NORTH HAMPTON
NSW73
NSW74
0 0.1 0.2 MILES
0 0.2 0.4 KILOMETERS
HAMPTON FALLS
SGW98
SGW89
SGW97
KFW81
SGW96
SGW26
KFW80
SGW103
SGW33
SGW100
SGW95
SGW99
KENSINGTON
SGW93
SGW101
SEABROOK
0 0.1 0.2 MILES
0 0.25 0.5 KILOMETERS
GREENLAND
RYE
NSW84
NSW83
NSW82
NSW80
NSW85
NSW79
NSW55
NSW81
NORTH HAMPTON
0 0.1 0.2 MILES
0 0.1 0.2 KILOMETERS



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 7–6. 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.)

Table 7–4. Transient model parameters and confidence intervals for the Seacoast model, southeastern New Hampshire.

[Parameter descriptions listed in appendix H1; ft/d, feet per day; d.f., dimensionless or dimensionless factor; C.I., 95-percent confidence interval]

Parameters for surficial-sediment hydraulic properties														
	Hydraulic conductivity, or conductivity multiplier, and streambed conductance										**Specific storage**			
	Ktill (ft/d)	**Ktillv (ft/d)**	**Ksd (d.f.)**	**Ksdv (d.f.)**	**Km (ft/d)**	**Kmv (ft/d)**	**Kwet (ft/d)**	**Kwetv (ft/d)**	**Ksb1 (ft/d)**		**SSsd (d.f.)**	**SSwet (d.f.)**		
Upper C.I.	1.8E+0	3.6E-1	1.4E+1	4.5E+8	3.4E+0	4.4E-2	2.3E+1	2.9E-1	5.6E+0		1.4E-4	4.2E-4		
Final values	1.0E+0	1.0E-1	1.0E+1	1.0E+0	1.0E-1	1.0E-2	1.0E+1	1.0E-1	5.0E+0		1.0E-4	1.0E-4		
Lower C.I.	5.6E-1	2.8E-2	7.3E+0	2.2E-9	2.9E-3	2.3E-3	4.3E+0	3.5E-2	4.5E+0		7.0E-5	2.4E-5		
Parameters for bedrock hydraulic properties														
	Hydraulic conductivities								**Horizontal anisotropy**		**Specific storage**			
	Rxk1 (ft/d)	**Rxk2 (ft/d)**	**Rxk3 (ft/d)**	**Rxk4 (ft/d)**	**Rxk1v (ft/d)**	**Rxk2v (ft/d)**	**Rxk3v (ft/d)**	**Rxk4v (ft/d)**	**Hani1 (d.f.)**	**Hani2 (d.f.)**	**SSR1 (d.f.)**	**SSR2 (d.f.)**	**SSR3 (d.f.)**	**SSR4 (d.f.)**
Upper C.I.	5.8E-1	1.2E+0	1.2E-1	7.4E-1	1.3E+0	2.8E+0	1.4E-1	8.1E-1	6.2E+0	7.9E+0	5.0E-6	5.3E-6	1.2E-2	1.1E+3
Final values	5.0E-1	1.0E+0	1.0E-1	2.0E-1	5.0E-1	1.0E+0	1.0E-1	2.0E-1	5.0E+0	1.0E+0	7.8E-7	2.8E-6	1.2E-7	1.0E-7
Lower C.I.	4.3E-1	8.2E-1	8.4E-2	5.4E-2	1.9E-1	3.6E-1	7.0E-2	4.9E-2	4.0E+0	1.3E-1	1.2E-7	1.5E-6	1.3E-12	8.8E-18
Parameters for 2000–04 average recharge														
	R1 (ft/d)	**R2 (ft/d)**	**R3 (ft/d)**	**R4 (ft/d)**	**R5 (ft/d)**	**R6 (ft/d)**	**R7 (ft/d)**	**R8 (ft/d)**	**R9 (ft/d)**	**R10 (ft/d)**	**R11 (ft/d)**	**R12 (ft/d)**		
Upper C.I.	4.5E-3	4.3E-3	1.7E-2	1.4E-2	6.9E-3	3.7E-3	1.4E-4	1.3E-3	1.2E-3	2.4E-3	6.9E-3	7.1E-3		
Final values	4.0E-3	3.8E-3	1.5E-2	1.3E-2	6.0E-3	3.0E-3	-5.0E-4	9.6E-4	9.5E-4	2.1E-3	6.2E-3	6.4E-3		
Lower C.I.	3.6E-3	3.3E-3	1.4E-2	1.2E-2	5.1E-3	2.3E-3	-1.1E-3	6.7E-4	7.2E-4	1.8E-3	5.6E-3	5.8E-3		
Parameters for 2003 recharge														
	RCH25 (ft/d)	**RCH26 (ft/d)**	**RCH27 (ft/d)**	**RCH28 (ft/d)**	**RCH29 (ft/d)**	**RCH30 (ft/d)**	**RCH31 (ft/d)**	**RCH32 (ft/d)**	**RCH33 (ft/d)**	**RCH34 (ft/d)**	**RCH35 (ft/d)**	**RCH36 (ft/d)**		
Upper C.I.	6.8E-3	4.9E-3	1.6E-2	1.1E-2	5.3E-3	4.5E-3	-6.5E-5	2.8E-3	5.1E-4	4.2E-3	6.3E-3	8.0E-3		
Final values	6.2E-3	4.3E-3	1.5E-2	1.0E-2	4.5E-3	3.9E-3	-7.3E-4	2.6E-3	1.7E-4	3.8E-3	5.7E-3	7.3E-3		
Lower C.I.	5.7E-3	3.8E-3	1.4E-2	9.1E-3	3.7E-3	3.3E-3	-1.4E-3	2.4E-3	-1.7E-4	3.4E-3	5.2E-3	6.6E-3		
Parameters for 2004 recharge														
	RCH37	**RCH38**	**RCH39**	**RCH40**	**RCH41**	**RCH42**	**RCH43**	**RCH44**	**RCH45**	**RCH46**	**RCH47**	**RCH48**		
Upper C.I.	1.7E-3	1.4E-3	8.7E-3	2.0E-2	7.0E-3	2.5E-3	6.0E-4	3.5E-3	3.2E-3	4.2E-3	6.1E-3	10.0E-3		
Final values	6.7E-4	9.3E-4	7.8E-3	1.8E-2	6.0E-3	1.8E-3	-1.0E-5	3.2E-3	2.9E-3	3.8E-3	5.6E-3	9.0E-3		
Lower C.I.	2.7E-4	6.3E-4	7.0E-3	1.7E-2	5.0E-3	1.1E-3	-6.2E-4	2.9E-3	2.6E-3	3.4E-3	5.1E-3	8.1E-3		

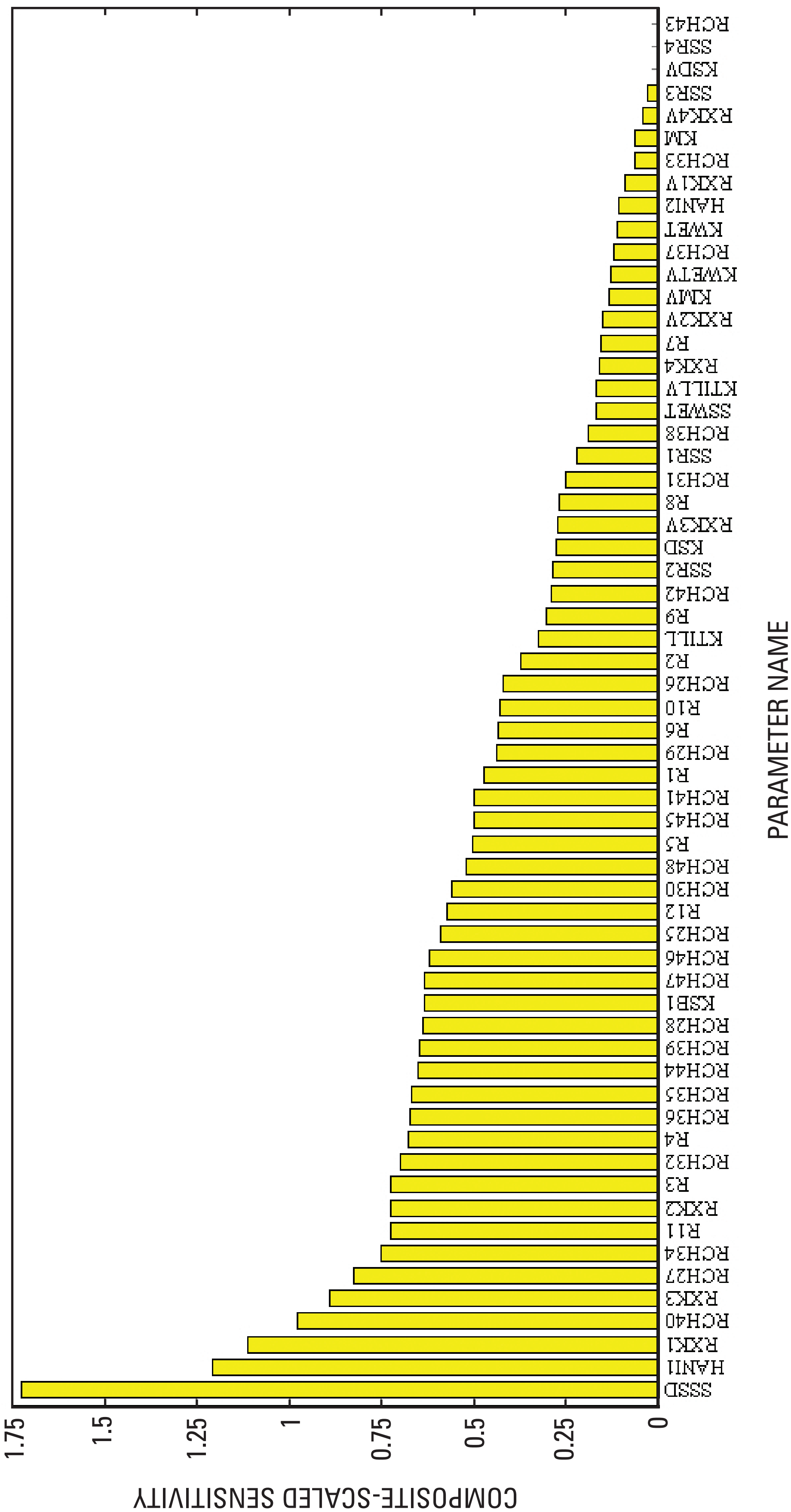
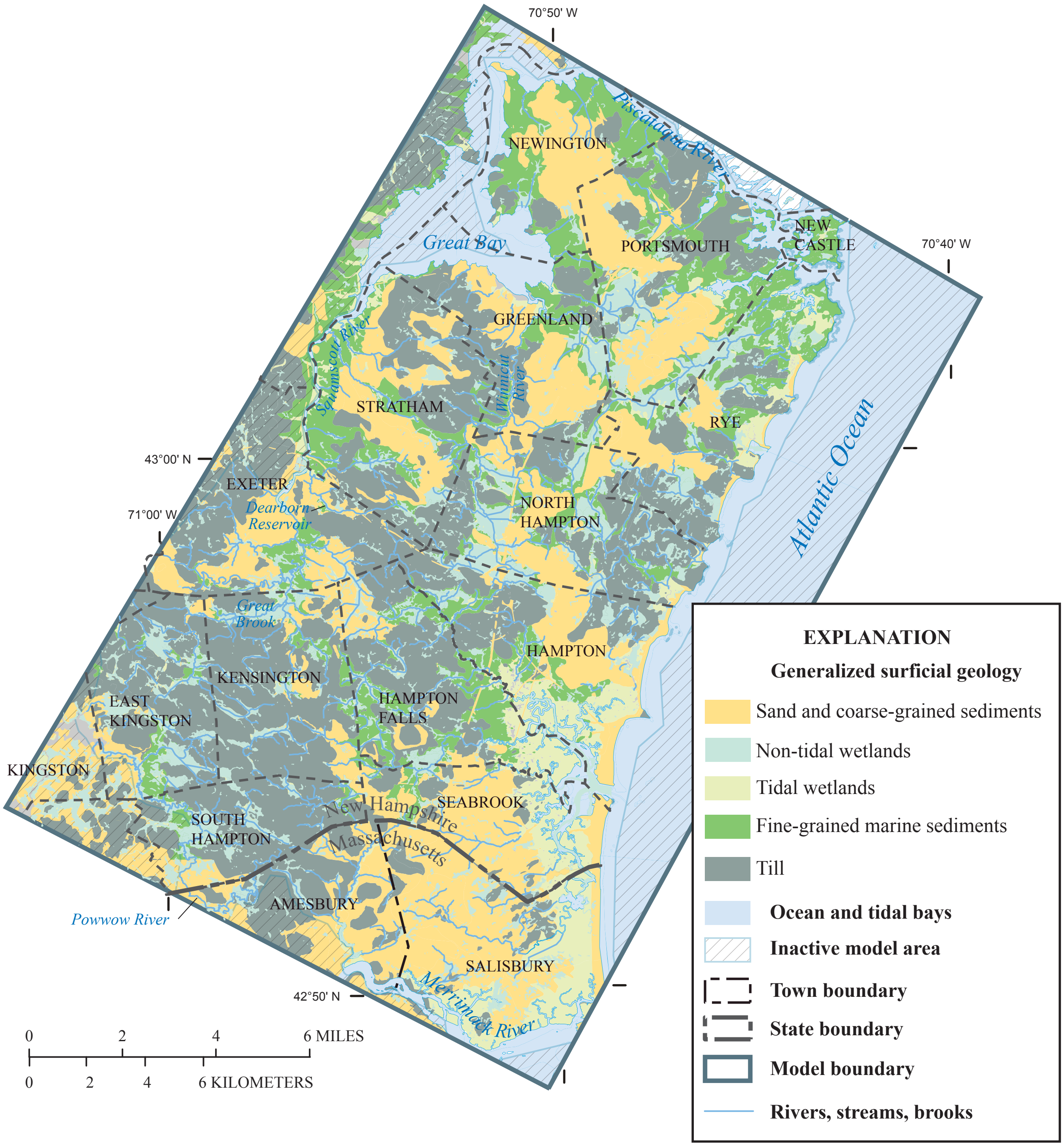


Figure 7–7. Composite-scaled sensitivity plot for transient model parameters for the Seacoast model, southeastern New Hampshire. List of parameters is in table 7–1.



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

Figure 7–8. 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.)

Parameter-sensitivity analysis indicated that ground-water-flow simulation is most sensitive to the overburden sediment storage; bedrock horizontal hydraulic conductivity for units Rx1, Rx2, and Rx3 (fig. 7–6); recharge; and anisotropy (HANI1, fig. 7–7). The simulation was least sensitive to overburden or bedrock vertical hydraulic conductivities, bedrock storage, and wetland and fine-grained sediment properties. The simulation was insensitive to vertical hydraulic conductivity of coarse-grained sediments (Ksdv), specific storage in bedrock zone 4 (SSR4), and recharge in stress period 43 (RCH43) (fig. 7–7). Vertical hydraulic conductivity is well known to be about 1:10 in glacial sediments in New Hampshire. Because of the small area and thickness of stratified-drift sediments relative to the dimensions of the entire regional-flow system, the associated parameters (Ksd, Ksdv) had little effect on the regional-flow system. The sensitivity of the specific storage in bedrock zone 4 was estimated as low as a result of the few large stresses (large ground-water withdrawals) and observations in this zone. This zone consists of the Exeter Diorite and Newburyport Complex where ground-water explorations in general have resulted in the completion of few high-yielding bedrock wells. The recharge in stress period 43, July 2004, was very close to zero (-1E–5 ft/d), and the observation residuals were relatively small compared to other stress periods (fig. 7–7).

The bedrock hydraulic conductivities in table 7–4 are bulk values for the regional scale. Regional bedrock hydraulic conductivities in the Seacoast model area appear to be in the range of 0.1 to 1 ft/d. The calculated values reflect the general density and connectivity of bedrock fracturing at the regional scale for bedrock zones which are based on mapped geologic units. The values do not reflect the hydraulic conductivity of individual fracture zones or variations within zones. The bedrock hydraulic-conductivity parameters calculated with the transient model were about an order of magnitude greater than those calculated with the steady-state model. The reason for this is that the transient observation data, which includes higher base flows than the steady-state observation data, require simulation with greater hydraulic conductivities. This illustrates the effect of different observation data on the estimated parameters. The transient calibrated parameters generally were more robust because of the greater range of observation values. The poor distribution of head-observation points in the transient model compared to the steady-state model, however, limits the ability with which the transient inverse model can solve for parameters important to simulating head variations.

Local bedrock hydraulic conductivities at high-yield bedrock wells will be much higher than the regionally estimated values. As with the steady-state model, calculated heads within a few model cells of large withdrawals will not be realistic. The calculated heads in the bedrock aquifer that were more than a few model cells from the large ground-water withdrawals are regionally realistic because flow in the regional bedrock aquifer is controlled by regional bedrock aquifer properties and not individual fracture zones. For example, the drawdowns reported at large ground-water withdrawals in the model area generally were localized and did not propagate great distances away from the well because the hydraulic conductivity of the bulk rock matrix is lower than the conductivity near a high-yield bedrock well.

Few observation data were obtained for bedrock parameter zone Rx4, particularly base-flow data in either the steady-state or transient model for use with inverse modeling. The lack of stress data is reflected in the large confidence interval, several orders of magnitude, calculated for bedrock storage in this zone. In the transient model, hydraulic conductivity in this zone was simulated with greater values (Rxk4/Rxk4v, 0.02/.02) than the values used in the steady-state model (Rxk4/Rxk4v, 0.001/0.01). Because the model was not very sensitive to this parameter, however, the use of different estimated values was not very meaningful. Parameter values differed between the models, as should be expected, because of the different stresses and observation data used.

Investigation of bedrock well yields in the geologic units corresponding to parameter zone Rx4 (Exeter Diorite Formation and Newburyport Complex) indicated that well yields were higher inside the model area than they were outside the model area and were greater than may have been expected on the basis of examination of the statewide bedrock-yield-probability map. These differences reflect the value of the additional bedrock-well data that were collected for this investigation. It is important to note, however, that the geologic contacts as presently mapped, which form the boundaries of the bedrock parameter zones, are only approximately known and may not reflect the true rock at depth. Moore and others (2002) demonstrated the importance of detailed geologic mapping with respect to analysis of bedrock-well yields. Although bedrock hydraulic conductivity and specific storage cannot be estimated with a high degree of certainty for bedrock zone Rx4, lower well yields and the small number of large water-supply systems in this area indicate that the regional water availability in this zone is low relative to the other bedrock zones.

Model Calibration and Calculated Water Balance

Simulated flows and heads in the transient model were calibrated to monthly observations. The simulated recharge rate was an important variable in the monthly transient simulation. The annual totals of the 2003 and 2004 simulated monthly recharge rates (23 and 22 in.) were comparable to the simulated average recharge rate (22 in.) (fig. 7–9). Peak recharge was earlier in simulation year 2003, and late summer recharge was lower. Recharge rates were similar in the summer and fall for both years with greater differences in the spring recharge rates (fig. 7–9). Base flows in 2003 and 2004 were about 30 percent of the total annual precipitation and simulated recharge was about 48 percent of the total annual precipitation (table 7–5). Differences between the simulated rates of net recharge (fig. 7–9) and the calculated rates of base flow per unit of drainage area (tables 7–3 and 7–4), which were less than recharge rates, were caused by runoff during intense precipitation events, consumptive water use, and water transfers out of the system. Figures 7–10A–B illustrate the calibration in terms of simulated and observed base flows for all observations (fig. 7–10A) and base flows by drainage area by month (fig. 7–10B).

The steady-state simulated recharge rate (approximately 11 in/yr or about 1 in/mo), based on the October 2004 synoptic observations of head and flows, is about the same as the October 2004 transient simulated recharge rate (1.4 in/mo). The average October monthly recharge rate (0.8 in.) for 2000–04, which includes a drought period (fig. 7–3A), is lower than the 2004 rate (fig. 7–9). The October simulated recharge rate used in the steady-state model is about half of the mean annual rate of recharge and represents a seasonal low-flow condition. The transient model results indicated that the ground-water-flow system changes by season and over the course of a year and does not remain in a true steady-state condition. The fall is the most stable period of the year (has the least change) and may be considered to be the season which the ground-water-flow system is in a quasi-steady-state condition. The October synoptic conditions are representative of a fall flow rate but do not represent long-term or average conditions.

Simulated average recharge in January and February is low, about 1.5 and 1.3 in. (fig. 7–9), compared to actual precipitation (3.5 to 3.3 in.) (fig. 7–3B); this difference reflects an accumulation of precipitation in the snowpack. In March and April, simulated net recharge (5.6 and 4.7 in.) was greater than median precipitation (3.8 and 3.7 in.) (fig. 7–3B) because of the addition of recharge from the melting snowpack. Simulated net recharge in the month of May (2.2 in.) is slightly below the median precipitation (3.1) and likely reflects both runoff and a low ET rate. During the month of July, average net recharge is negative, (-0.2 in.) and in August and September it is about 0.3 in. for each month. Net recharge in October is less than 1 in. and increases to about 2.2 in. by December. On average, more than half of the total annual recharge occurs in the spring—about 25 percent in March, 20 percent in April, and 10 percent in May—whereas about 20 percent of the total recharge occurs in the fall (10 percent each in November and December).

Table 7–5. Annual base flows calculated for selected watersheds in the Seacoast model area from 2000 to 2004, southeastern New Hampshire.

[in., inches; in/yr, inches per year; —, not available]

Year	Precipitation (in.)	Winnicut River (in/yr)	Little River (in/yr)	Berrys Brook (in/yr)	Mill Brook (in/yr)	Hampton Falls River (in/yr)	Average (in/yr)	Base flow as percentage of precipitation	Recharge as percentage of precipitation
2000	53.7	17.0	13.4	19.5	15.9	16.5	16.5	31	41
2001	39.6	12.5	10.3	15.0	12.1	13.1	12.6	32	55
2002	45.6	11.0	7.8	10.9	10.3	10.1	10.0	22	48
2003	47.3	18.3	12.6	23.9	16.6	16.0	17.5	37	49
[1]2003	47.3	15.6	10.6	20.0	14.2	13.7	14.8	31	49
2004	46.1	15.6	10.4	13.4	13.2	14.1	13.3	29	48
Averages									
2000–04	46.5	14.3	10.5	15.7	13.2	13.5	13.4	—	—
2003–04	46.7	15.6	10.5	16.7	13.7	13.9	14.1	—	—

[1] With streamflow peaks reduced.

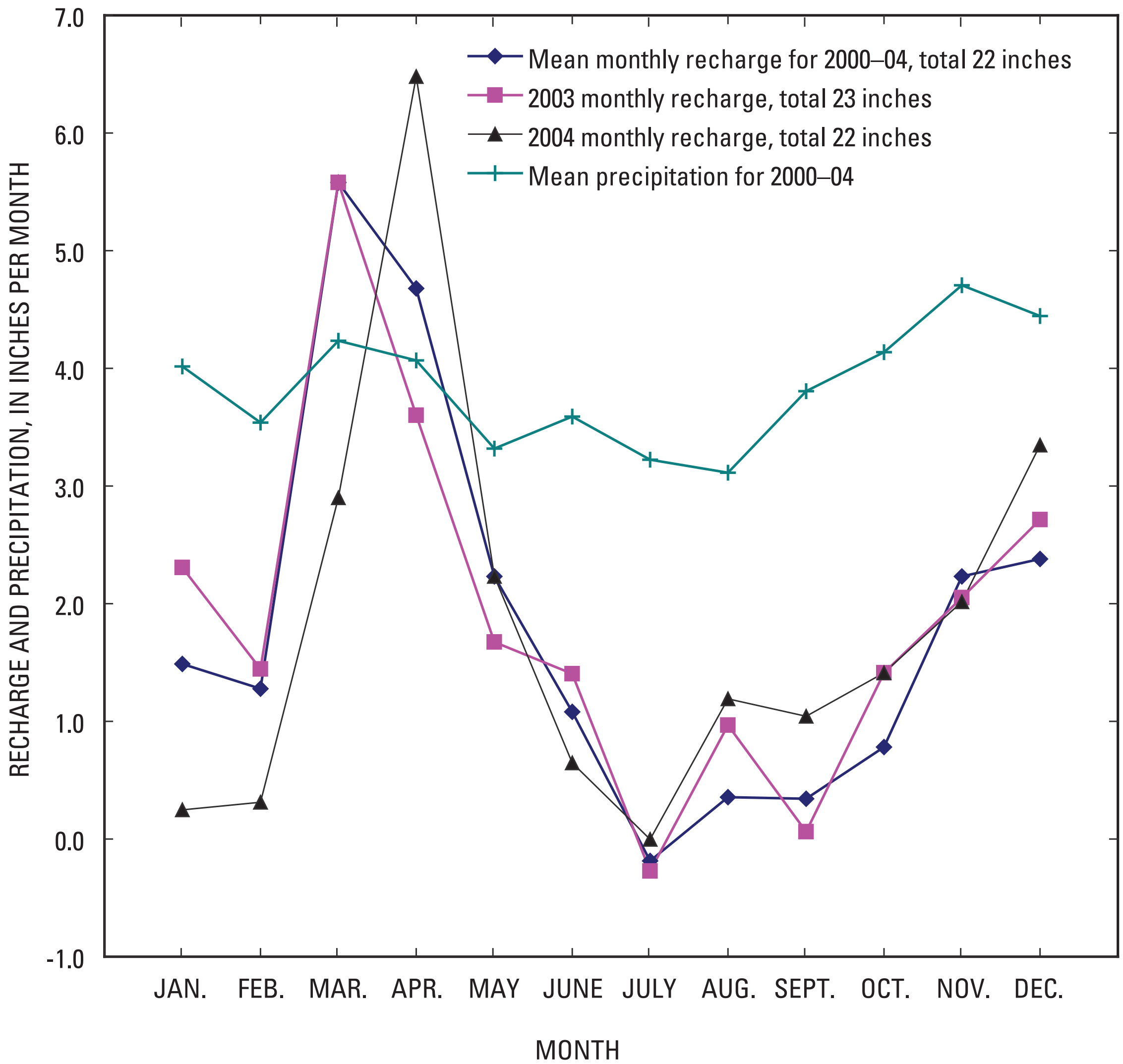


Figure 7–9. Average monthly recharge for 2000–04, monthly recharge for 2003 and 2004, and mean precipitation for 2003–04 for the Seacoast model area, southeastern New Hampshire. (This figure is the same as figure 15 on page 31 in the report.)

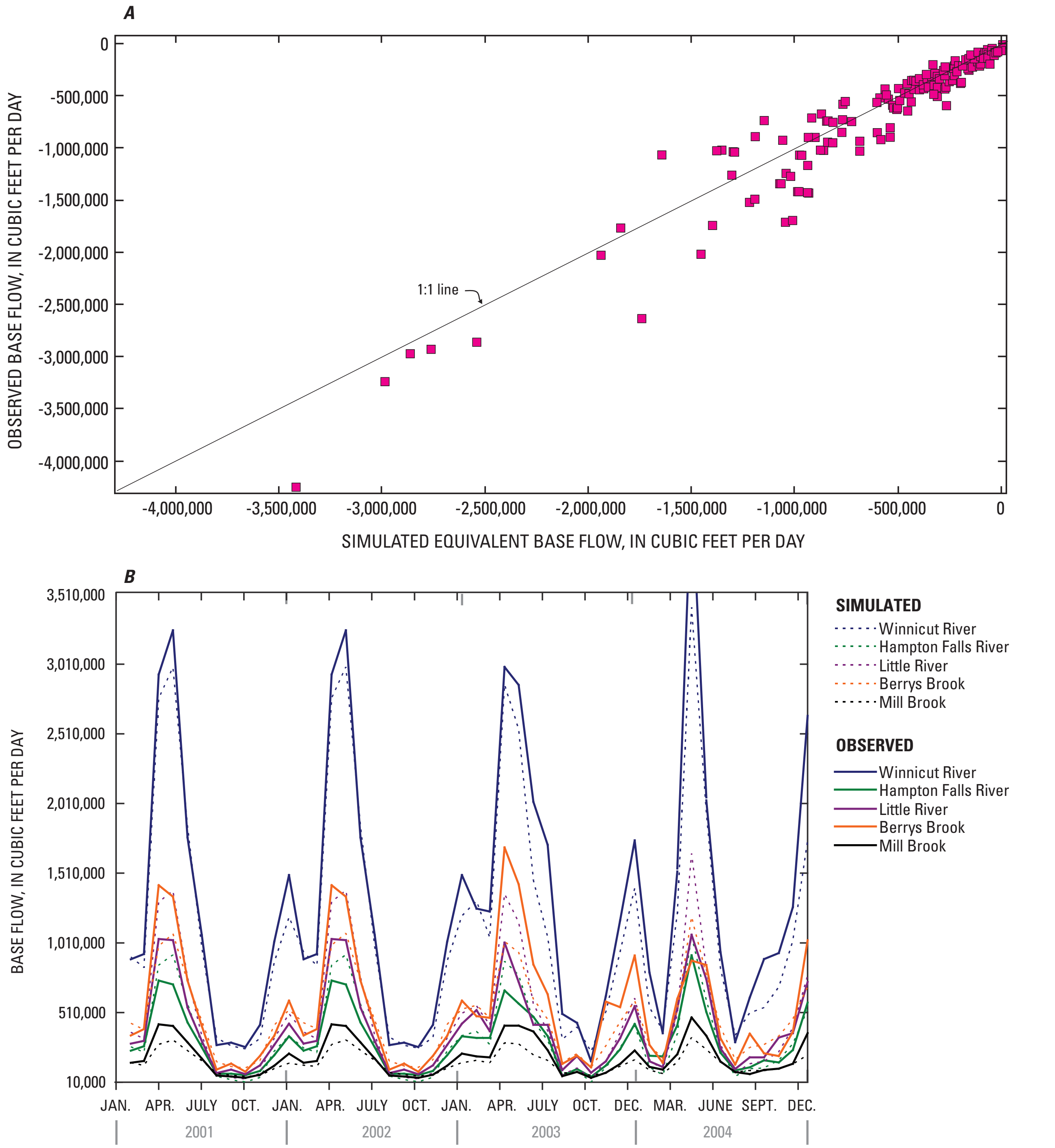


Figure 7–10. *(A)* Simulated and observed monthly base flows compared to a 1:1 line, and *(B)* simulated and observed base flows by month, in the Seacoast model area, southeastern New Hampshire.

A commonly used water-balance method of estimating actual ET for a land area can be obtained by subtracting streamflow and base flow from precipitation (Dingman, 1994). In this investigation, where monthly net recharge is defined as the direct recharge minus ET, a comparison of monthly total and average precipitation with the simulated monthly and average net recharge (fig. 7–9) provided insight into the magnitude and timing of runoff and ET in the Seacoast aquifer system. In the months November through February, the melting of the snowpack runoff was generally about 2 in. per month and recharge was about 50 percent (in the fall) and 40 percent (in the winter) of the monthly precipitation. ET in the study area increased through the growing season, May to September, but depended on climatic factors and vegetation. Precipitation fell as rainfall during this period and was consistent at about 3 to 4 in/mo. However, simulated net recharge ranged from about 2 in. in May to a slightly negative value in July (-0.2 in.), to about 0.3 in. in August and September. Monthly ET in the Seacoast ground-water system was inferred to be about 0.9 in. in May, 2 in. in June, 3.2 in. in July, 2.2 in. in August, and 2.9 in. in September. ET in October was probably about 0.75 to 1 in.; however, it was difficult to assess by this method. The inferred ET rates were slightly less than rates calculated for the Winnicut River watershed (table 8; GeoInsight, Inc., 1999) by the Thornthwaite method (Dingman, 1994), which gave a maximum rate of 4 in/mo for July and August.

Seasonal variations in net recharge provide a complicated picture of ground-water availability. The ground-water-flow system is continually discharging water (draining) throughout the year, and the net recharge is greater than the discharge in the spring and late fall. During a typical annual cycle, the aquifer system is storing water during the winter and spring when inflows are greater than outflows and losing water during the summer and early fall when outflows are greater than inflows. Figure 7–11 illustrates this process by showing the total amount of water moving into and out of storage in the bedrock aquifer for the model area over a simulated average annual period. Larger inflows are simulated in the spring and outflows in the summer than in the fall and winter when both fluxes are less. As noted above, the fall is the most stable period of the year, a quasi-steady state period, when fluxes are less and inflows and outflows are approximately balanced.

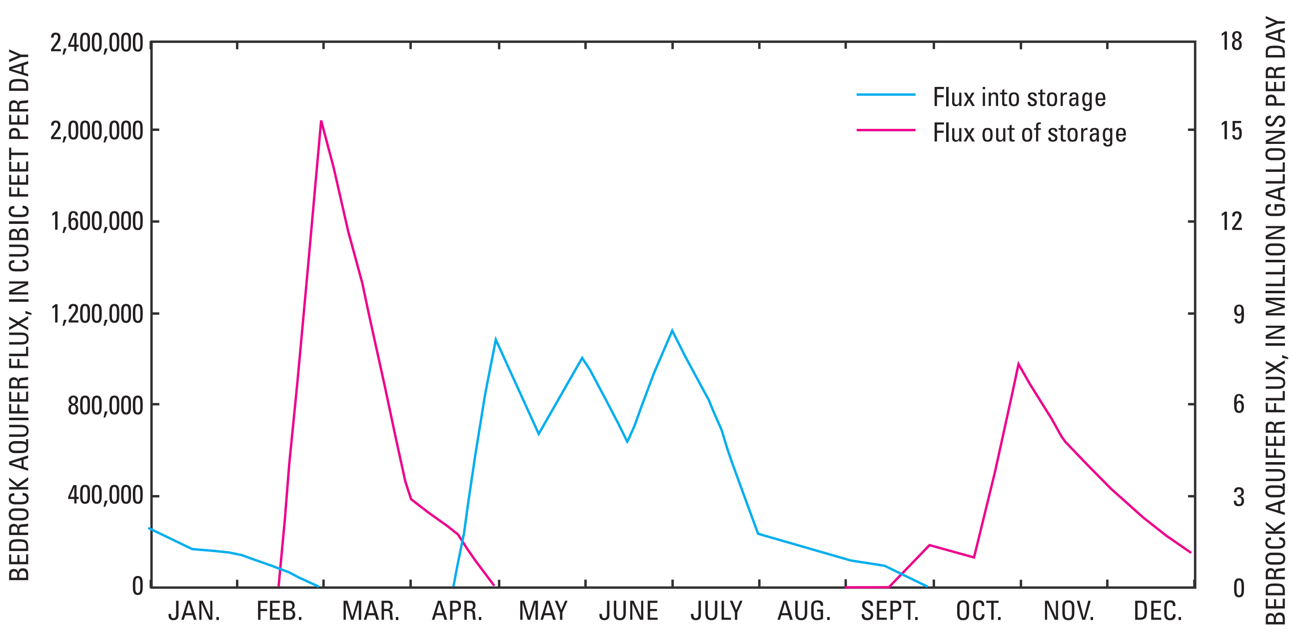


Figure 7–11. The total flux of water into and out of storage in the bedrock aquifer for a simulated average annual period in the Seacoast model area, southeastern New Hampshire.

Unweighted residuals for simulated heads are shown in figure 7–12. The weighted residuals are dimensionless but can be compared to the model-calculated error variance (s^2) or to the standard error of regression (s), which represents the expected residual based on data accuracy and weighting. In this case the model-calculated standard error of regression is ±2.77. Overall, the simulated heads fit the regional trend, and the average weighted residual for all observations was 0.15. Eighty-six percent of all weighted residuals were within one standard error of regression, and 95 percent were within two standard errors of regression. The average weighted residual for the network observations was -1.47, indicating that in general network heads were higher than observations. Regionally, discrepancies calculated by the transient model between the simulated and observed heads were similar to those that were calculated by the steady-state model. The distribution of monthly head observations available for the transient-head simulation, however, was poor and was almost entirely limited to a few areas with head networks. Synthetic heads were used to create observations outside the network areas, but model fit to the synthetic observations was poor. The average weighted residual for synthetic observations was 3.20, indicating that the simulated heads for some till covered areas were poorly fit and less than the expected heads. Some error can be attributed to the fact that the monthly head observations, unlike the base-flow observations, were made at a point in time that may not be representative of a monthly mean value. Other head errors were attributed to the use of approximate (DEM) reference elevations and topographic variations. Differences between actual and monthly approximated water uses also contributed to differences between observed and simulated heads. Water levels in the till aquifer generally rose to the land surface during the spring; these high water levels exceeded the pressure range on the continuous water-level recorders and thus were not recorded (breaks in the records shown in fig. 7–4). It is possible that the potentiometric surface in the till aquifer is above the land surface during the spring melt and that the water-level fluctuations were greater than the ranges indicated by the data.

Comparisons of simulated and observed monthly ground-water heads for four wells are shown in figure 7–13A–D. Simulated seasonal head variations generally were greater than the observed variations. The fit between simulated and observed heads was closer for wells in areas with greater ranges in head (fig. 7–13B–D) caused by local withdrawals and poorer in areas with few withdrawals (fig. 7–13A). This result indicated that the ground-water-flow model simulated the dynamics of ground-water flow better in areas with stresses, but that the simulated hydrologic parameters were not as well suited to other areas. It is likely that a model designed for use solely in an area without complex stresses may provide better results than a model designed for use in both areas; however, it is difficult to estimate hydraulic properties without stresses on the aquifer system. Efforts to reduce the simulated seasonal head variations by manually increasing hydraulic conductivities and decreasing recharge were not successful and reduced the overall model fit. The overall head configurations for the October 2004 transient and steady-state simulations (appendix 5) were comparable; differences resulted from the use of slightly different water-use tabulations for the two simulations.

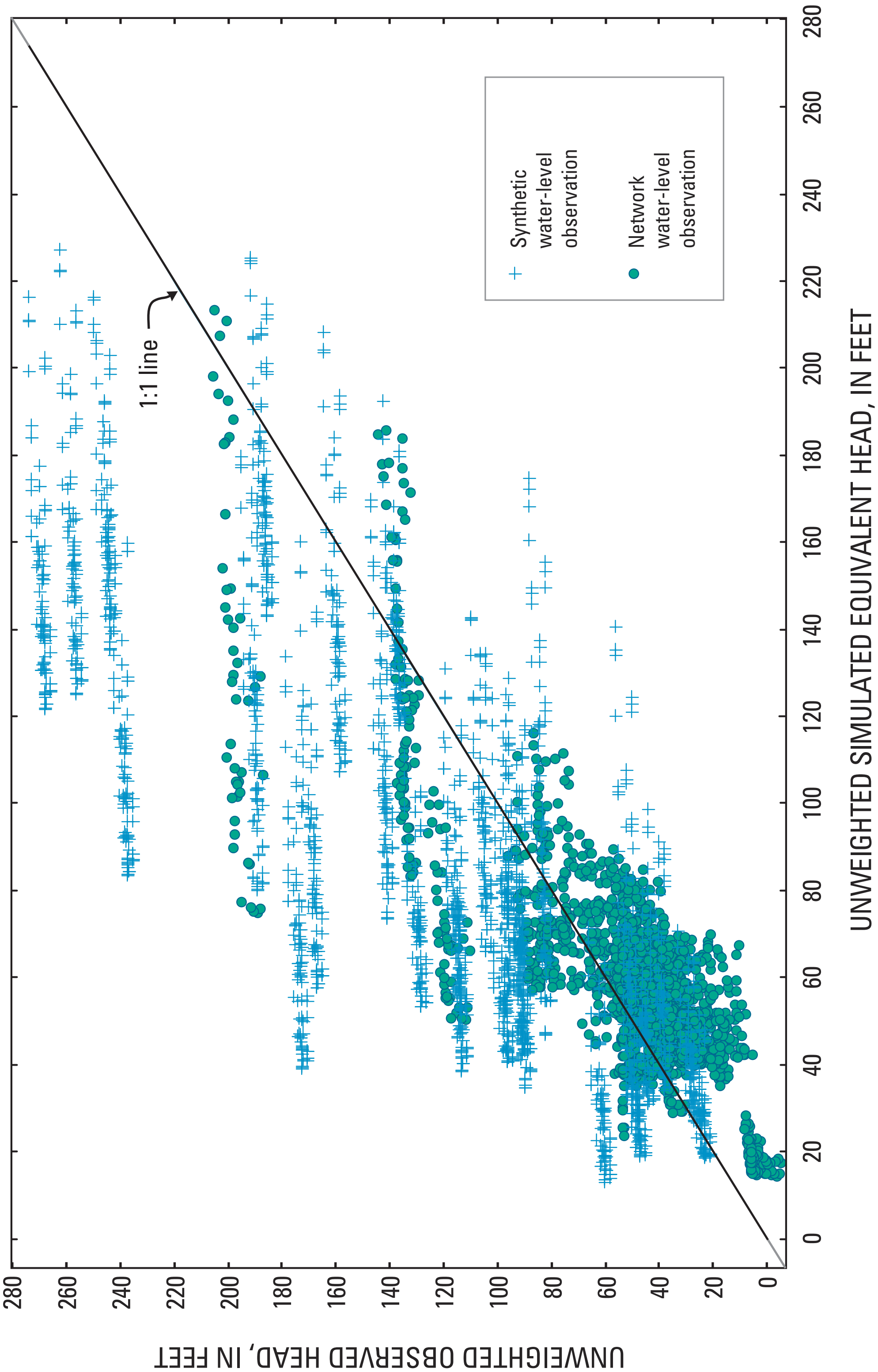


Figure 7–12. Selected simulated and observed monthly ground-water heads for the Seacoast model area, southeastern New Hampshire.

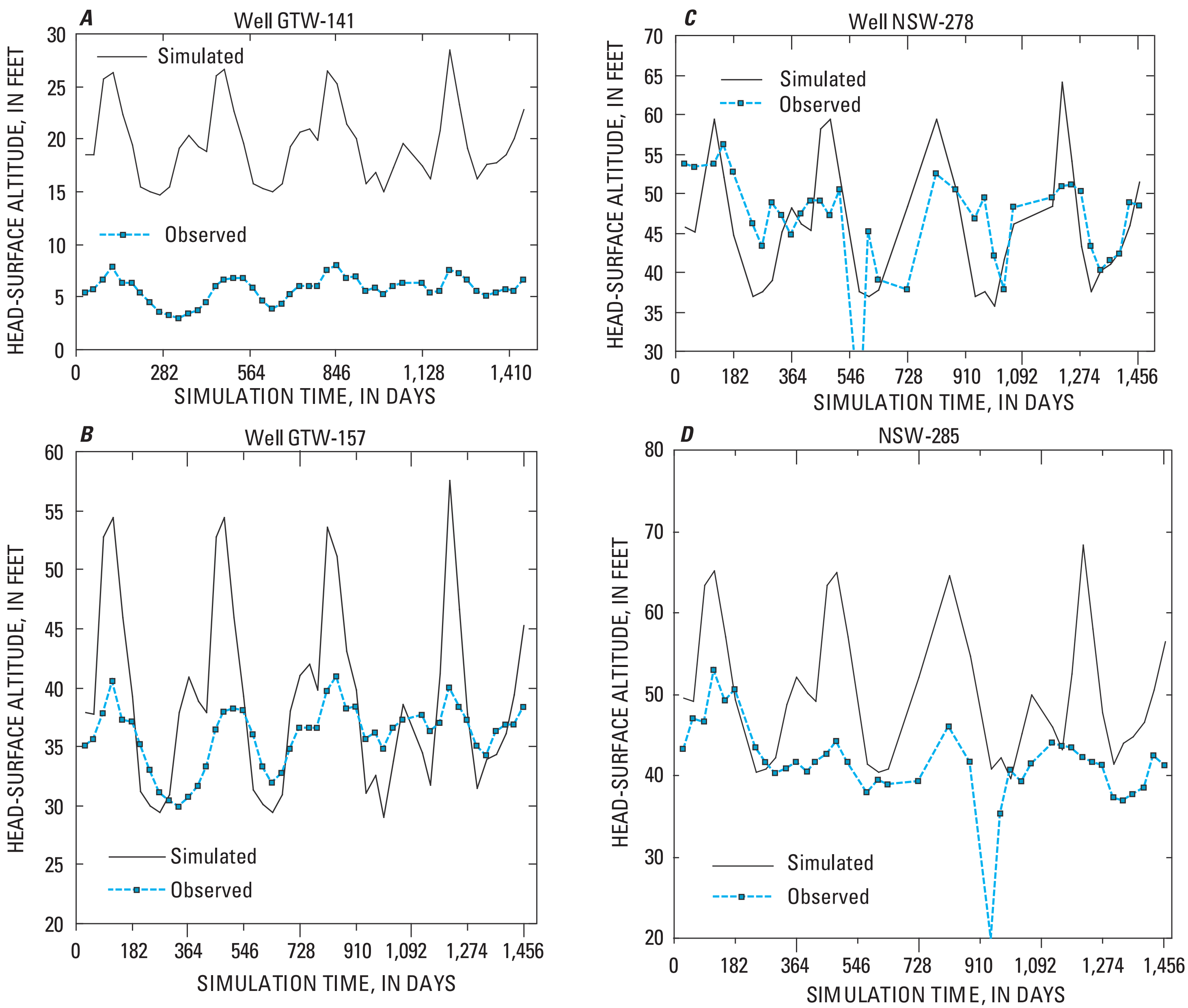


Figure 7–13. Simulated and observed ground-water heads at monitoring wells *(A)* GTW-141, *(B)* GTW-157, *(C)* NSW-278, and *(D)* NSW-285 for the Seacoast model, southeastern New Hampshire. (Location of wells shown on figure 7 on page 16–18 in the report.)

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