Appendix Part 3. Development of Groundwater Flow Models

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Appendix Part 3. Development of Groundwater Flow Models

By John P. Masterson

Development of Groundwater-Flow Models

Numerical flow models are useful tools for testing and improving conceptual models of groundwater-flow systems by providing a means to synthesize existing hydrogeologic information into an internally consistent mathematical representation of a real system or process (Konikow and Reilly, 1999). Groundwater-flow models were developed for two areas within the Pawcatuck River Basin, the lower Wood River and eastern Pawcatuck River areas (figs. A3–1, A3–2). These numerical flow models were based on the USGS computer program MODFLOW (Harbaugh and others, 2000) and were used for the purposes of assessing the potential effects of groundwater pumping from irrigation and water-supply withdrawal sites proposed within the study area on streamflows and water levels. Results calculated by MODFLOW were compared to results calculated by the analytical streamflow depletion algorithm in the HSPF model and used to evaluate alternatives for the conjunctive management of the groundwater and surface-water resources in selected areas of the basin.

Model Discretization and Boundaries

The finite-difference model grid consists of a series of orthogonal model cells used to vary user-specified hydraulic parameters, model stresses, and boundary conditions spatially. The conceptualization of how and where water enters, moves through, and leaves the aquifer is critical to the development of an accurate flow model (Reilly, 2001). A detailed discussion of grid discretization, boundary conditions, and the use of finite-difference equations to simulate groundwater flow is presented in McDonald and Harbaugh (1988), Reilly (2001), and Reilly and Harbaugh (2004).

Spatial Discretization

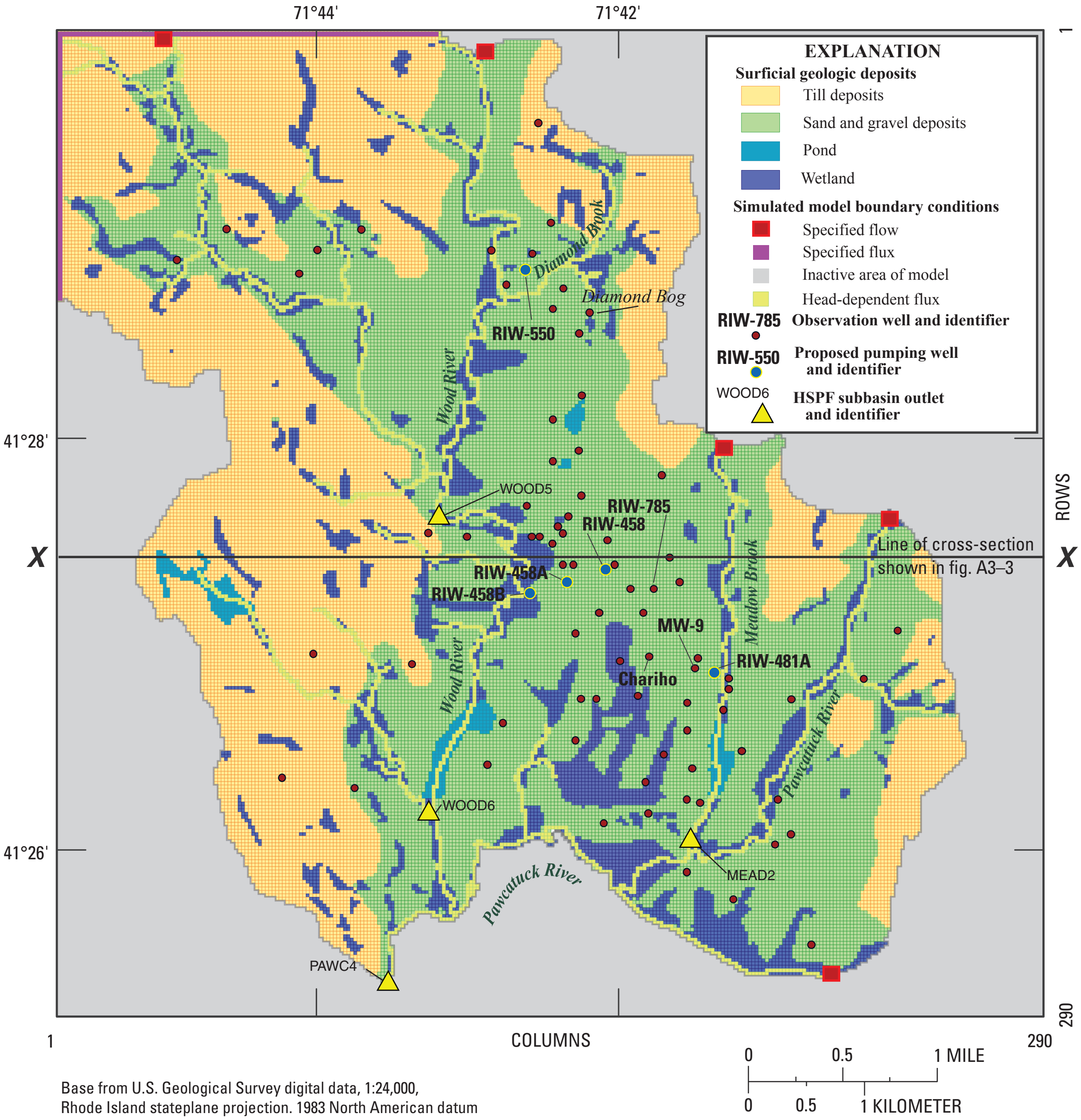
The total active modeled area in the lower Wood River subbasin is about 18 mi^2, and the active modeled area of the eastern Pawcatuck River subbasin is about 114 mi^2 (figs. A3–1, A3–2). The finite-difference grids representing these two areas consist of 290 rows and 290 columns of uniformly spaced model cells that are 100 ft on a side for the lower Wood River model and 400 rows and 210 columns of uniformly spaced model cells that are 250 ft on a side for the eastern Pawcatuck River model. Both modeled areas were subdivided vertically into four layers of different thicknesses that extend from the water table downward to encompass the upper 50 ft of bedrock. The eastern Pawcatuck River model extends from the Pawcatuck River Basin to the southern coastline of Rhode Island to represent groundwater flow better in the southern part of the basin.

Most previous groundwater modeling investigations of areas that include upland till and bedrock surrounding glacial stratified deposits did not explicitly simulate the upland areas, but rather specified the recharge that would have entered the aquifer through these areas as enhanced recharge at the contact between till and glacial stratified deposits (Dickerman and others, 1997; Barlow and Dickerman, 2001; DeSimone and others, 2002; Granato and others, 2003; Friesz, 2004). This approach was used to prevent numerical instabilities that often result from simulating steeply sloping, thinly saturated deposits, such as the upland till areas throughout the northern part of the Pawcatuck River Basin and discussed previously in Part 1.

A more recent investigation of the South Coastal Basin of southern Rhode Island (Masterson and others, 2007) included a fixed-transmissivity approach that allowed for the simulation of both the glacial stratified deposits and the upland till and bedrock areas. This approach is equivalent to simulating a confined aquifer, which, for steady-state conditions that do not represent seasonal fluctuations of the water table, is a reasonable representation of both areas within the aquifer. The benefit of this approach is that it allows for a more realistic representation of flow from these upland areas to the surrounding glacial stratified deposits.

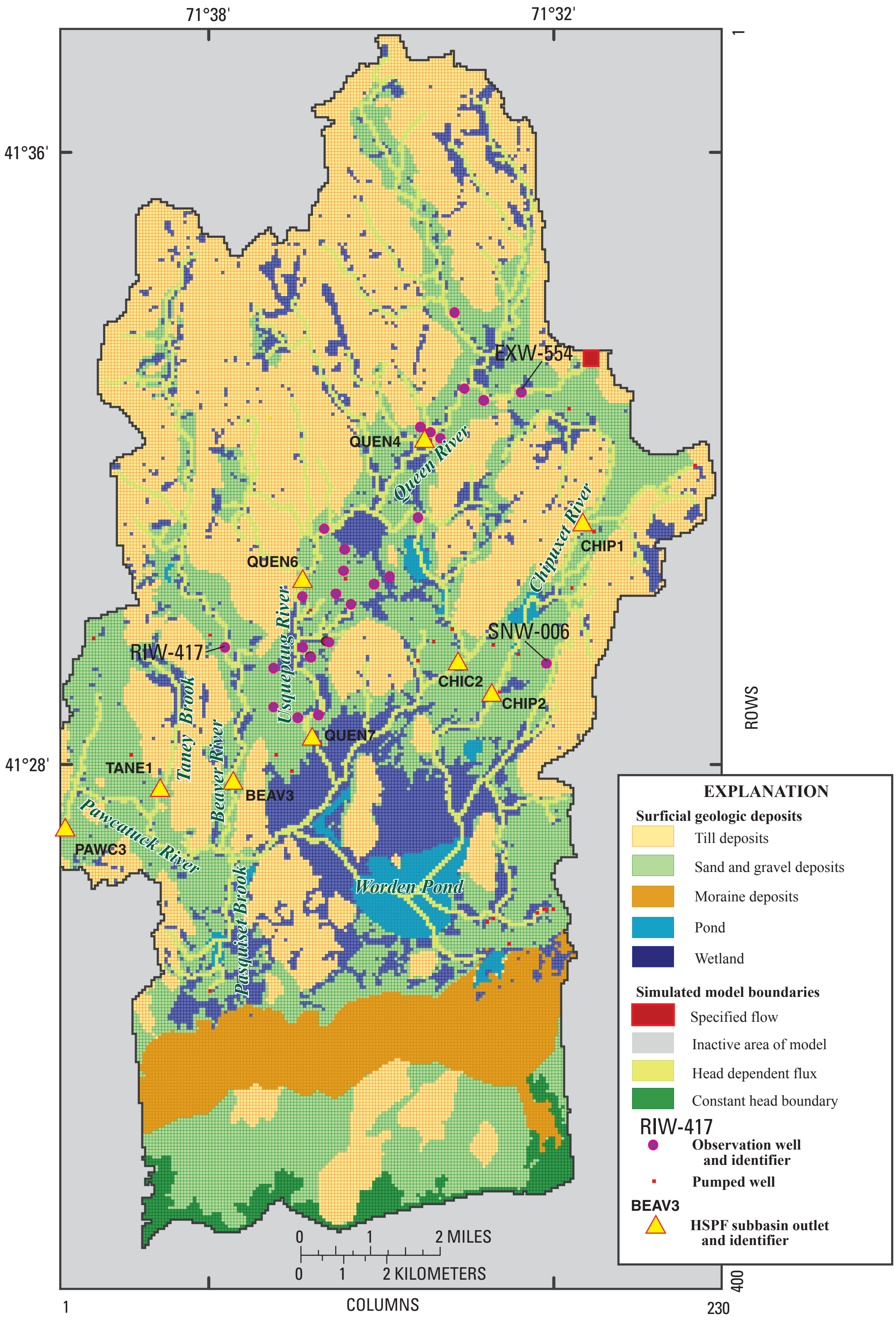
The model layer most affected by this fixed-transmissivity approach is the top layer. In the variable-transmissivity approach, the top of layer 1 is the water table calculated by the model, whereas in the fixed-transmissivity approach, the top of the layer must be specified. The top of the uppermost layer (layer 1) was initially set to the land-surface altitude, which was interpolated from 30-m digital-elevation-model data (P.A. Steeves, U.S. Geological Survey, written commun., 2004) and then reset to the altitude of the simulated water table during the model-calibration process. After several iterations of resetting the top of layer 1 to the altitude of the model-calculated water table, a reasonably close match to the observed water-level data was achieved, and that water-table altitude was then specified as the top of layer 1.

The glacial stratified deposits were then represented in three model layers of varying thicknesses (layers 1–3) from the water table to bedrock to allow for the detail necessary to represent the streambed thickness, vertical changes in the lithology, and the screened intervals of pumping wells (fig. A3–3). The top layer (layer 1) is relatively thin (generally less than 5 ft thick) to provide as realistic a representation as



Base from U.S. Geological Survey digital data, 1:24,000,
Rhode Island stateplane projection. 1983 North American datum

Figure A3–1. Geographic extent, surficial geology, observation wells, proposed withdrawal sites, simulated boundary conditions, and outflow points from HSPF subbasins for the lower Wood River model area in the Pawcatuck River Basin, southwestern Rhode Island.



Base from U.S. Geological Survey digital data, 1:24,000,
Rhode Island stateplane projection. 1983 North American datum

Figure A3–2. Model extent and boundary conditions used in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island.

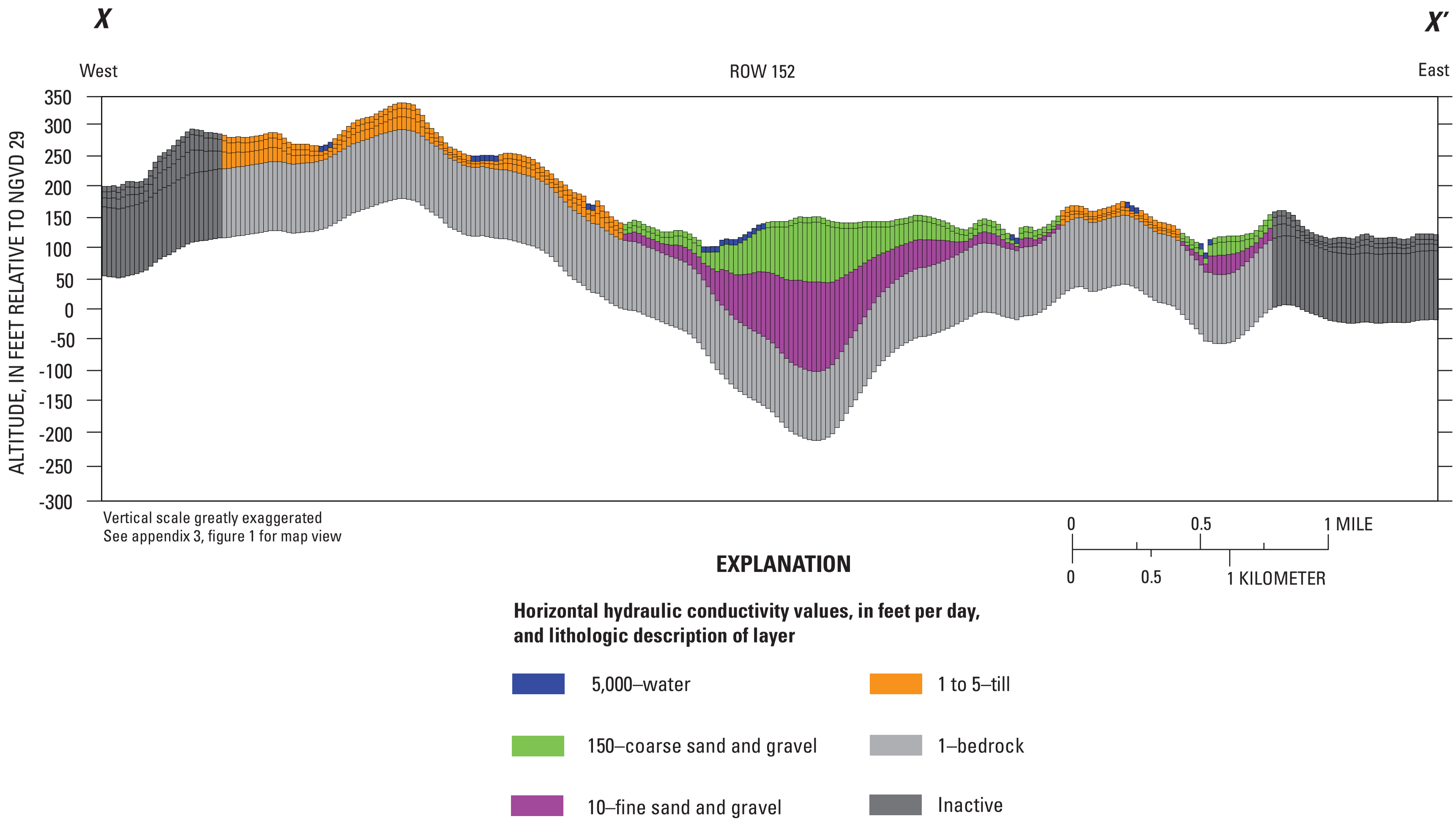


Figure A3–3. Hydraulic-conductivity distribution and aquifer thickness in the lower Wood River model area in the Pawcatuck River Basin, southwestern Rhode Island. (Location of cross section shown in figure A3–1.)

possible of the rivers in the modeled areas (blue cells within layer 1 in figure A3–3). Layer 2 generally represents the coarser glacial stratified deposits of sand and gravel (green cells), and layer 3 represents the finer glacial stratified deposits of silt and clay (purple cells). The bottom layer (layer 4); shown as gray cells (fig. A3–3), extends from the top of the bedrock surface to 50 ft below the bedrock surface to allow for flow in fractured bedrock in areas where the unconsolidated deposits are thin, such as beneath the upland tills (shown as orange cells). An example of the model layering used in this investigation is shown for section X-X′ (fig. A3–3) for the lower Wood River model (fig. A3–1).

Hydrologic Boundaries

The hydrologic boundaries in a flow model are the areas in which and the method by which all of the water entering and leaving the model is specified, and generally coincide with the physical boundaries of the flow system.

In developing the model for the Pawcatuck River Basin, it was assumed that the groundwater and surface-water divides were coincident in the upland till areas and therefore defined the extents of the active areas for both models. One area in which the groundwater and surface-water divides are not coincident is the Chipuxet River subbasin in the eastern Pawcatuck River model (fig. A3–2). In this area, groundwater flows to the east across the surface-water divide into the Annaquatucket River Basin. To account for this, the eastern no-flow boundary in the eastern Pawcatuck River model was shifted to the west from the surface-water divide to be coincident with the groundwater divide determined from the results of groundwater-flow-model simulations of the Hunt-Annaquatucket-Pettaquamscutt stream-aquifer system (Barlow and Dickerman, 2001).

In areas where the models did not extend to an upland-till surface-water divide—namely, across river valleys—specified flows were assigned to the model cells closest to the points where rivers entered the active modeled areas. These specified flows were derived from the hydrograph-separation analysis (Rutledge, 1998) described in Part 1. The lower Wood River model included five of these cells, and the eastern Pawcatuck River model included one (fig. A3–1).

The upper boundary of the model was the water table, which was a free-surface boundary that received spatially variable recharge from precipitation. The lower boundary of the model was the crystalline bedrock that underlies the entire study area. The bottom of the model was set at 50 ft below the bedrock surface to allow for the simulation of groundwater flow in the upper, more weathered, part of the bedrock.

The lateral boundaries of the model, through which all groundwater discharged to streams and the coast, were represented in the model as either head-dependent flux boundaries or constant-head boundaries. The head-dependent flux boundaries were used to simulate the interaction between groundwater and surface water in the streams throughout the basin. The constant-head boundaries were used in the eastern Pawcatuck River model to simulate groundwater discharge to the coastal waters along the southern boundary of the model.

The stream boundaries were simulated by the Stream-Routing Package (STR) (Prudic, 1989), which allows for the hydrologic interaction between the aquifer and the adjoining streams and routes water between adjacent stream reaches. Discharge fluxes to the streams are calculated by the model on the basis of the hydraulic gradient between the model-calculated head in the model cell representing the stream and the specified-boundary head and hydraulic conductance at the boundary face. The boundary heads specified for the streams simulated by the STR package were set equal to the stream-stage altitudes estimated from the 1:24,000 USGS topographic-map hypsography for the study area (P.A. Steeves, U.S. Geological Survey, written commun., 2004). The hydraulic conductance was calculated for each model cell representing a stream (McDonald and Harbaugh,1988) as

$$C = \frac{(K)(W)(L)}{(M)}\ , \tag{1}$$

where

C is the hydraulic conductance of the streambed (ft^2/d),
K is the vertical hydraulic conductivity of streambed deposits (ft/d),
W is the width of the stream (ft),
L is the length of the stream within the model cell (ft), and
M is the thickness of the streambed (ft).

The parameters used in equation 1 to calculate the hydraulic conductance (*C*) were assumed to be the same for all streams and rivers. The simulated vertical hydraulic conductivity of the streambed deposits (*K*) was set equal to 30 ft/d, a value similar to that of the surrounding aquifer. The width (*W*) of the streams was assumed to be 5 ft, the length (*L*) was set equal to the length of the model cell representing the stream (100 ft for the lower Wood River model and 250 ft for the eastern Pawcatuck River model), and the streambed thickness was assumed to be 5 ft.

The resulting conductance value of 7,500 ft^2/d in the eastern Pawcatuck River model was adjusted downward to 2,500 ft^2/d, similar to the value calculated for the lower Wood River model, to minimize the mass-balance errors from simulations that utilized the higher conductance value. Justification for this adjustment can be found in the MODFLOW documentation (McDonald and Harbaugh, 1988), which states that in the absence of detailed field measurements of streambed characteristics, a conductance value must be chosen more or less arbitrarily and may need to be adjusted during model calibration. Adjustments to these parameters may affect the locations and magnitudes of model-calculated fluxes to and from these simulated surface-water bodies; local-scale analyses of specific surface-water bodies may require the collection of more detailed hydrologic data than was possible for this regional analysis.

Constant-head boundaries are used to represent large surface-water bodies, such as the coastal-water boundary of the eastern Pawcatuck River model. It is assumed that flow to or from these water bodies will not affect their water levels. The amount of water flowing between the surface-water body and the aquifer in the model depends upon the groundwater heads in the model cells that surround the specified-head boundary (Reilly, 2001).

The constant-head boundary used for the salt-pond coastal-water bodies along the southern Rhode Island coast in the eastern Pawcatuck River model was set at a uniform value of 2.0 ft above the National Geodetic Vertical Datum of 1929 (NGVD 29), this value was similar to the value used in the previous model developed for coastal Rhode Island (Masterson and others, 2007). The southern part of the eastern Pawcatuck River model was extended southward to the Rhode Island coastline because of the uncertainty in the position of the groundwater divide that separates the Pawcatuck River Basin from the South Coastal Basin. Simulations by the model developed for coastal Rhode Island (Masterson and others, 2007) showed that adjusting the constant head up or down by 1.0 ft had little effect on model-calculated fluxes to these coastal waters.

Temporal Discretization and Initial Conditions

The simulation period for this investigation from January 2000 through September 2004 was subdivided, or discretized, into 75 stress periods. The first stress period simulated the average pumping and recharge conditions for the study period by using the steady-state option in MODFLOW–2000 (Harbaugh and others, 2000). The remaining 74 stress periods were subdivided into monthly periods of 28 to 31 days with the exception of the months May through September 2002, for which 22 weekly stress periods were used. The 2002 summer period was more finely discretized because it included the lowest streamflows observed during the simulation period, and the finer time discretization allowed for a more accurate model representation of the measured changes in streamflow under low-flow conditions. Daily time steps were used for each of the monthly stress periods and half-day time steps were used for the weekly stress periods to improve model stability and to minimize mass-balance errors.

Hydrologic Stresses

The hydrologic stresses simulated in the model included recharge from precipitation and pumping for water supply and turf-farm irrigation. These stresses were varied between successive weekly and monthly stress periods. The recharge estimates were derived from the HSPF model (Part 2), and the pumping rates for production and irrigation wells are presented in Part 1.

Recharge

Recharge rates were specified in the two models on the basis of the active groundwater component of the HSPF model for the Pawcatuck River Basin. Recharge rates were specified for three distinct zones defined according to the hydrogeologic characteristics assigned to these zones in each model. These zones were the glacial stratified deposits, upland till areas, and wetlands (shown in figs. A3–1 and A3–2). The areal average recharge rate used in the two models for the study period was about 22.7 in/yr. In the lower Wood River model, this rate was derived from recharge rates of 32.9 in/yr for the stratified glacial deposits, 21.5 in/yr for the upland till, and -4.2 in/yr for the areas covered by wetlands. In the eastern Pawcatuck River model, this rate was derived from recharge rates of 33.4 in/yr for the stratified glacial deposits, 23.3 in/yr for the upland till, and -4.2 in/yr for the areas covered by wetlands. Because upland till areas are characterized by lower permeability, greater topographic relief, and the potential for more overland runoff than valley-fill deposits, it is generally assumed that recharge in upland till is less than in glacial stratified deposits, although no field data were available to quantify the difference in this study.

The aggregate recharge rate of about 22.7 in/yr is consistent with values that were initially derived from hydrograph-separation techniques and adjusted during model calibration in previous studies in Rhode Island (Dickerman and others, 1997; Barlow and Dickerman, 2001; Granato and others, 2003; Friesz, 2004; Zarriello and Bent, 2004). This rate is nonetheless lower than what was reported previously for long-term average conditions for this area because it represents the average of the study period only, January 2000 through September 2004, rather than the longer time period associated with the previous studies.

Recharge in wetlands is also poorly understood and subject to a large degree of uncertainty, but it is believed to vary greatly depending on the extent of open water, type of vegetation, the location of the wetland within the flow system, and whether the wetlands have surface-water outflows. For this reason, recharge rates for wetlands simulated in numerical models are often determined through model calibration rather than by field measurements.

The development of the models for the Pawcatuck River Basin was based on the assumption that nearly all of the ponds and wetlands are hydrologically connected to the Pawcatuck River and its associated tributaries. Because precipitation on ponds and wetlands was assumed to flow out of the basin through the stream network, no simulated recharge reached the underlying aquifer in these areas. For the period from late spring to early fall, it was assumed that the evapotranspiration rate exceeded the precipitation rate; this excess resulted in a net water change of -4.2 in/yr in wetland areas.

It was assumed for the purpose of this investigation that no water was returned to the aquifer from irrigation or from onsite domestic septic systems. Irrigation of turf farms is the predominant agricultural use of pumped water in the basin, and it was assumed that during the growing season, April to October, all of the water pumped for irrigation was lost to evapotranspiration. The water pumped for municipal supply was exported out of the basin (Wild and Nimiroski, 2004) and thus removed from the groundwater-flow system. Water use from small-capacity domestic-supply wells was not considered in this analysis; it was assumed that, because most of this water was returned to the aquifer through nearby onsite septic systems, there was no net change in flow in the aquifer.

Pumping

Pumping for municipal supply was simulated in the eastern Pawcatuck River model from wells in the town of Kingstown from the Kingstown Water Department, United Water of Rhode Island, and the University of Rhode Island. The average total pumping rate simulated for the study period was about 3.4 Mgal/d. A detailed discussion of the simulated changes in pumping rate for municipal supply and turf irrigation is included in Part 1.

Hydraulic Properties

The hydraulic properties required as input data for the simulation of groundwater flow are the horizontal and vertical hydraulic conductivity of the aquifer and aquifer storage. The hydraulic conductivity values initially specified in the flow model were assigned on the basis of aquifer-test, lithologic, and geologic information (Gonthier and others, 1974; Melvin and others, 1992) as well as information published in previous modeling investigations in southern Rhode Island (Dickerman and others, 1990, 1997; Barlow and Dickerman, 2001; Granato and others, 2003; Friesz, 2004). Hydraulic conductivity values for stratified glacial deposits were assigned on the basis of grain-size categories and geologic processes of deposition. A detailed discussion of the geologic setting of the Pawcatuck River Basin can be found in Part 1. The relation between geologic framework and hydraulic conductivity values from a similar model of groundwater flow in southern Rhode Island (Masterson and others, 2007) was used in this investigation.

The range in hydraulic conductivity values specified in the flow models is listed in table A3–1 for the deposits shown on figures A3–1 and A3–2. An example of the distribution of hydraulic conductivity values used in the calibrated model is shown for cross-section X–X′ (fig. A3–3).

Surface-water bodies, such as ponds and wetlands, were simulated as part of the aquifer system and assigned hydraulic properties. It was assumed that the ponds were in direct hydraulic connection to the aquifer and that they provided no effective resistance to flow. As a result, groundwater-flow lines converged toward ponds in upgradient areas where groundwater discharges to ponds, and diverged in downgradient areas where ponds recharge the aquifer. Ponds in the model were simulated as areas of high hydraulic conductivity (50,000 ft/d), more than two orders of magnitude higher than the values for hydraulic conductivity specified in surrounding aquifer (table A3–1).

Table A3–1. Hydraulic conductivity values of the lithologic units simulated in the groundwater-flow models in the Pawcatuck River Basin, southwestern Rhode Island.

Lithologic deposit	Horizontal hydraulic conductivity (feet per day)	Anisotropy
Stratified glacial deposits		
Sand and gravel	150–350	5:1 to 3:1
Fine/medium sand	70–125	30:1 to 10:1
Fine sand and silt	10–30	100:1 to 10:1
Silt and clay	0.01–10	1:1 to 100:1
Moraines	10–70	100:1 to 30:1
Peat	1.0	100:1
Till	1–5	10:1
Bedrock	1.0	10:1

It was assumed that the surfaces of wetlands had a greater resistance to horizontal flow than the kettle-hole ponds, but less resistance to flow than the surrounding aquifer. As a result, wetland areas were assigned a horizontal hydraulic conductivity value of 5,000 ft/d and a vertical hydraulic conductivity value of 5 ft/d (anisotropy of 1,000:1).

The peat deposits that underlie the wetlands were assigned a hydraulic conductivity value of 1.0 ft/d to account for the highly decomposed peat and organic-rich silts beneath the wetlands (Friesz, 2004). It was assumed that the peat deposits are highly anisotropic; the vertical hydraulic conductivity of these deposits was specified as 0.01 ft/d (anisotropy of 100:1) (table A3–1).

The streambed deposits were assumed to be much more permeable than the wetland peats and more similar to the surrounding aquifer. As a result, the vertical hydraulic conductivity value used to determine the conductance discussed previously in the section “Hydrologic Boundaries” was 30 ft/d, the vertical hydraulic conductivity associated with a horizontal hydraulic conductivity of 150 ft/d and an anisotropy of 5:1 for sand and gravel deposits (table A3–1).

The remaining hydraulic property represented in the models is the storage term, which is required to quantify the volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head (Heath, 1989). The storage capacity of the aquifer controls, in part, the degree to which water levels fluctuate in the aquifer in response to seasonal changes in recharge, changes in pumping rates, and the timing of the release of groundwater to streams.

Storage properties were assigned in the models on the basis of the same hydrogeologic units used to assign the hydraulic conductivity values. Storage was specified in the model as a dimensionless storage coefficient as a feature of the fixed-transmissivity approach, which is based on the assumption of confined conditions in all of the model layers. The storage coefficient terms specified for layer 1 were assumed to represent specific yield and were 0.23 for glacial stratified deposits, 1.0×10^{-7} for upland till, and 1.0 for the surface-water bodies.

The upland till areas were assigned low storage values—much lower than published values to compensate for the inability of the model to simulate properly the seasonally varying dewatering of these upland areas in layer 1. This limitation is a consequence of the fixed-transmissivity approach, which would otherwise result in unrealistically large water releases of water from storage during the dry summer months when presumably the upland till areas would dry out and no longer transmit water to the glacial stratified deposits (DeSimone, 2004).

Surface-water bodies, including ponds, streams, and wetlands, were all assigned a storage value of 1.0 to represent the maximum porosity and storage capacity for the water column which was explicitly represented as aquifer material in the flow models. These high values used in layer 1 resulted in the dampened fluctuations in water levels in response to seasonal changes in recharge.

In the lower model layers, the storage coefficient for the confined glacial stratified deposits and bedrock was set to 1.0×10^{-5} and the storage coefficient for upland till areas remained 1.0×10^{-7}. The aquifer sediments beneath the surface-water bodies consist of glacial stratified deposits and were represented as such with respect to simulated storage properties.

Model Calibration and Limitations

Model calibration is the process by which initial model-input parameters are modified to make simulated results more closely match observed water levels and streamflows (Reilly and Harbaugh, 2004). The adequacy of the calibration is based to a large degree on the intended use of the results and thus can differ among investigations. For this investigation, the primary use of the groundwater-flow models for the lower Wood River and eastern Pawcatuck River subbasins was to provide an analysis of the response of the groundwater-flow system to changes in groundwater withdrawals. This analysis included a comparison of the changes in model-calculated water levels and streamflows for different pumping scenarios.

As a result, the calibration of these particular models focused on matching the magnitudes of measured change in water levels and streamflows during the study period rather than matching the absolute values of measured water levels and streamflows. Water-level data from previous investigations, however, were used to show that the water levels simulated for steady-state conditions in this investigation provided a reasonable match to previously measured water levels (tables A3–2 and A3–3) given that the recharge conditions at the time those measurements were made differ from those observed during the study period. Data from previous investigations included the water-level data published for the lower Wood River model for October 1976 (Dickerman and others, 1990) and for the Usquepaug-Queen area of the eastern Pawcatuck River model for September 1989 (Dickerman and others, 1997).

A comparison of the model-calculated water levels to those that were measured throughout the study period provides a better measure of how well the models in this investigation simulated the changes in water levels with time. The water levels used in this comparison were measured in one observation well in the lower Wood River model area (RIW–785) (fig. A3–1) and in three observation wells in the eastern Pawcatuck River model area (EXW–554, RIW–417, and SNW–006) (fig. A3–2). The comparisons between the measured and model-calculated water levels for these wells are shown in figures A3–4A–D. In general, the changes in model-calculated water levels throughout the study period agreed well with the measured water levels.

The differences in the absolute values between measured and modeled water levels in these observation wells can be partially attributed to the effects of the representation of the stream network in the modeled area. In narrow valley-fill sand and gravel aquifers, such as those in the northeastern part of the Pawcatuck River Basin, the simulated streambed altitudes assigned to the model cells can greatly affect the water levels in the surrounding aquifer. Discrepancies between streambed altitudes specified in the model according to the digital-elevation data and the actual streambed altitudes can result in several feet of deviation of the model-calculated water levels from those measured in the field. Also, the effect of grid discretization on the geometry of the simulated stream network may result in an overrepresentation of the effect of the stream on the simulated water levels in the surrounding aquifer. This effect is more apparent when the model cells representing streams are much wider than the actual stream channel.

The calibration of the groundwater-flow model with respect to streamflow was based on a comparison of model-calculated streamflows to those determined by the hydrograph-separation techniques described in Part 1. A comparison of model-calculated to measured streamflows is presented in figures A3–5A–I for the outlet points of the HSPF subbasins described in Part 2.

Given the limitations of the calibration data coupled with the objectives of the model-calibration process, only minor adjustments were made to the initially specified hydraulic conductivity values. The storage coefficient term used to represent the upland till areas was adjusted downward to a value much lower than values previously reported for similar hydrogeologic settings so that the responses of the model-calculated water levels and streamflows to seasonal changes in recharge better matched the observed transient data. The recharge rates simulated in the models were determined by the HSPF model and were not modified during the calibration process in order to maintain a consistent water balance between the HSPF and MODFLOW models.

Adjustments also were made to the hydraulic properties to resolve large-scale, regional inconsistencies between the field data and the simulation results rather than to match individual calibration targets of observed water levels and streamflows. The fixed-transmissivity approach used to develop the flow models limits the maximum pumping rates that can be simulated in the flow models. If the simulated pumping rate of a given well site would result in a drawdown of the water table that is greater than 10 percent of the saturated thickness in that area then the fixed-transmissivity approach may result in an overrepresentation of the aquifer transmissivity, and therefore, an underrepresentation of the actual drawdown from that pumping rate (Reilly and Harbaugh, 2004).

The flow models developed for this investigation at best represent idealized conceptualizations of actual hydrogeologic conditions. The accuracy of the model predictions presented in this analysis are a function of the quality of the input calibration data, and how realistic were the assumptions of future pumping and recharge rates. Although, individual pumping rates and well locations will invariably change as water-management plans evolve over time; however, the use of numerical models developed according to a well-reasoned conceptual model of the aquifer system, with an understanding of the limitations of the simulation results, can provide valuable tools for improving the understanding of groundwater and surface-water interactions in the Pawcatuck River Basin.

Table A3–2. Model-calculated and measured groundwater levels in the lower Wood River model area in the Pawcatuck River Basin, southwestern Rhode Island.

[ID, identifier; Residual, residual equals measured minus model-calculated water levels. Measured water-level data derived from Dickerman and others (1990). Water-level values are in feet relative to NGVD 29]

Well ID	Stateplane-X coordinate	Stateplane-Y coordinate	Model layer	Water levels		
				Measured	Model-calcluated	Residual
CHW–226	274897.94	127118.23	1	42.1	41.9	0.2
CHW–426	277568.58	129237.94	1	43.0	42.3	0.7
CHW–429	277947.3	128224.99	1	46.0	43.8	2.2
RIW–231	273926.49	135420.07	1	59.2	59.6	-0.4
CHW–226	274897.94	127118.23	1	42.1	41.9	0.2
CHW–265	278549.97	124984.86	1	50.7	46.2	4.5
CHW–299	280089.27	132775.27	1	53.2	45.4	7.8
CHW–305	281082.1	134190.31	1	51.1	45.7	5.3
CHW–428	277489.63	127922.31	1	42.0	42.8	-0.8
HOW–291	268447.57	136951.38	1	54.1	52.6	1.5
HOW–300	264053.79	145363.72	1	80.4	75.8	4.6
HOW–425	269153.64	145350.54	1	57.0	55.0	2.0
RIW–253	275283.3	129141.67	1	47.1	48.0	-0.9
RIW–254	270201.09	137858	1	47.7	53.9	-6.2
RIW–257	271418.63	137551.42	1	57.2	56.1	1.1
RIW–284	274902.68	129243.74	1	48.2	48.9	-0.7
RIW–288	274907.18	131268.03	1	52.7	52.4	0.3
RIW–289	273759.22	128841.45	1	47.4	48.5	-1.1
RIW–291	272463.63	128540.79	1	52.0	47.7	4.3
RIW–294	269500.16	131483.13	1	49.0	48.8	0.2
RIW–418	270371.26	145246.35	1	60.0	57.8	2.2
RIW–419	269607.85	144337.26	1	58.0	55.3	2.7
RIW–420	270976.32	143625.43	1	66.0	59.0	7.0
RIW–421	271282.26	144231.99	1	66.0	59.7	6.3
RIW–422	272041.8	143521.68	1	66.0	60.9	5.1
RIW–428	271735.87	142915.11	1	65.7	60.3	5.4
RIW–430	271807.7	141093.05	1	64.3	59.6	4.7
RIW–432	270968.56	140386.54	1	63.3	58.3	5.0
RIW–438	271727.74	139473.79	1	62.6	58.8	3.8
RIW–440	271800.77	138157.81	1	60.9	57.5	3.4
RIW–443	271113.33	137248.51	1	56.6	55.3	1.3
RIW–444	271113.33	137248.51	1	56.6	55.3	1.3
RIW–447	271265.14	137045.71	1	56.5	55.2	1.3
RIW–448	270959.84	136742.79	1	56.3	53.8	2.5
RIW–449	270959.84	136742.79	1	56.1	53.8	2.3

Table A3–2. Model-calculated and measured groundwater levels in the lower Wood River model area in the Pawcatuck River Basin, southwestern Rhode Island.—Continued

[ID, identifier; Residual, residual equals measured minus model-calculated water levels. Measured water-level data derived from Dickerman and others (1990). Water-level values are in feet relative to NGVD 29]

Well ID	Stateplane-X coordinate	Stateplane-Y coordinate	Model layer	Water levels		
				Measured	Model-calcluated	Residual
RIW–450	271265.14	137045.71	1	56.5	55.2	1.3
RIW–452	270351.17	136946.69	1	55.8	52.3	3.5
RIW–453	272559.11	136840.23	1	55.7	57.9	-2.2
RIW–454	270579.6	136946.14	1	55.9	53.1	2.8
RIW–456	274385.44	136329.97	1	64.0	61.2	2.8
RIW–457	274688.43	135620.78	1	62.0	60.6	1.4
RIW–462	272785.9	136131.19	1	58.6	58.1	0.5
RIW–463	272785.9	136131.19	1	58.6	58.1	0.5
RIW–465	273241.15	135421.63	1	58.5	58.7	-0.2
RIW–467	271567.55	136134.05	1	57.5	55.3	2.2
RIW–468	271262.97	136134.78	1	56.5	54.2	2.3
RIW–470	272325.71	134715.25	1	58.2	57.3	0.9
RIW–471	272325.71	134715.25	1	58.1	57.3	0.8
RIW–472	273620.28	134712.26	1	58.2	58.9	-0.7
RIW–475	271638.91	134109.57	1	61.0	55.7	5.3
RIW–476	275216.54	133392.86	1	56.0	56.9	-0.9
RIW–477	272931.64	133296.82	1	65.0	56.1	8.9
RIW–479	273462.42	132283.45	1	63.0	57.1	5.9
RIW–485	276129.08	132783.57	1	52.3	54.0	-1.7
RIW–486	276129.08	132783.57	1	52.8	54.0	-1.2
RIW–488	276128.42	132479.92	1	52.4	52.9	-0.5
RIW–489	276128.42	132479.92	1	54.0	52.9	1.1
RIW–490	275974.78	131872.97	1	51.3	51.2	0.1
RIW–491	275974.78	131872.97	1	51.6	51.2	0.4
RIW–492	276505.28	130657.23	1	47.2	47.9	-0.7
RIW–493	276505.28	130657.23	1	48.8	47.9	0.8
RIW–495	277955.59	132172.37	1	53.0	44.8	8.2
RIW–496	271786.68	132186.13	1	59.0	53.0	6.0
RIW–497	272243.64	132185.05	1	64.0	54.8	9.2
RIW–498	274908.99	132077.75	1	57.0	54.4	2.6
RIW–524	271631.49	130971.91	1	57.0	51.6	5.4
RIW–525	269040.19	130269.68	1	49.0	47.9	1.1
RIW–526	274220.13	130561.06	1	56.0	52.1	3.9
RIW–528	273685.13	129752.56	1	51.0	52.8	-1.8
RIW–631	275057.04	130154.33	1	46.3	50.0	-3.7

Table A3–3. Model-calculated and measured groundwater levels in the Eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island.

[ID, identifier; Residual, residual equals measured minus model-calculated water levels. Measured water-level data derived from Dickerman and others (1997). Water-level values are in feet relative to NGVD 29]

Well ID	Stateplane-X coordinate	Stateplane-Y coordinate	Model layer	Water levels		
				Measured	Model-calcluated	Residual
EXW–553	311039	175548	1	145.3	149.3	-4.0
EXW–554	316357	169269	1	144.9	143.2	1.7
EXW–555	313315	168562	1	132.9	132.9	0.0
EXW–556	309891	165528	1	125.2	122.7	2.5
EXW–558	309131	166035	1	120.6	120.5	0.1
EXW–559	308294	166440	1	122.4	123.8	-1.4
EXW–560	308136	159254	1	122.6	122.3	0.3
EXW–561	311795	169474	1	129.1	128.2	0.9
RIW–188	296634	147323	1	99.5	99.4	0.1
RIW–780	296629	144185	1	100.3	98.0	2.3
RIW–782	298531	143373	1	98.4	96.2	2.2
SNW–311	300680	158350	1	113.2	114.2	-1.0
SNW–314	302198	155009	1	117.1	114.3	2.8
SNW–515	299603	148129	1	101.4	99.5	1.9
SNW–1192	305850	154600	1	119.4	117.0	2.4
SNW–1193	304632	153994	1	118.6	115.7	2.9
SNW–1194	302276	156729	1	117.4	114.7	2.7
SNW–1195	302804	152376	1	111.9	111.3	0.6
SNW–1196	301587	153187	1	113.1	111.6	1.5
SNW–1197	298924	152988	1	108.2	107.1	1.1
SNW–1198	301050	149342	1	104.8	102.8	2.0
SNW–1199	300206	143573	1	95.6	94.1	1.5
SNW–1200	298995	148939	1	100.8	99.4	1.4

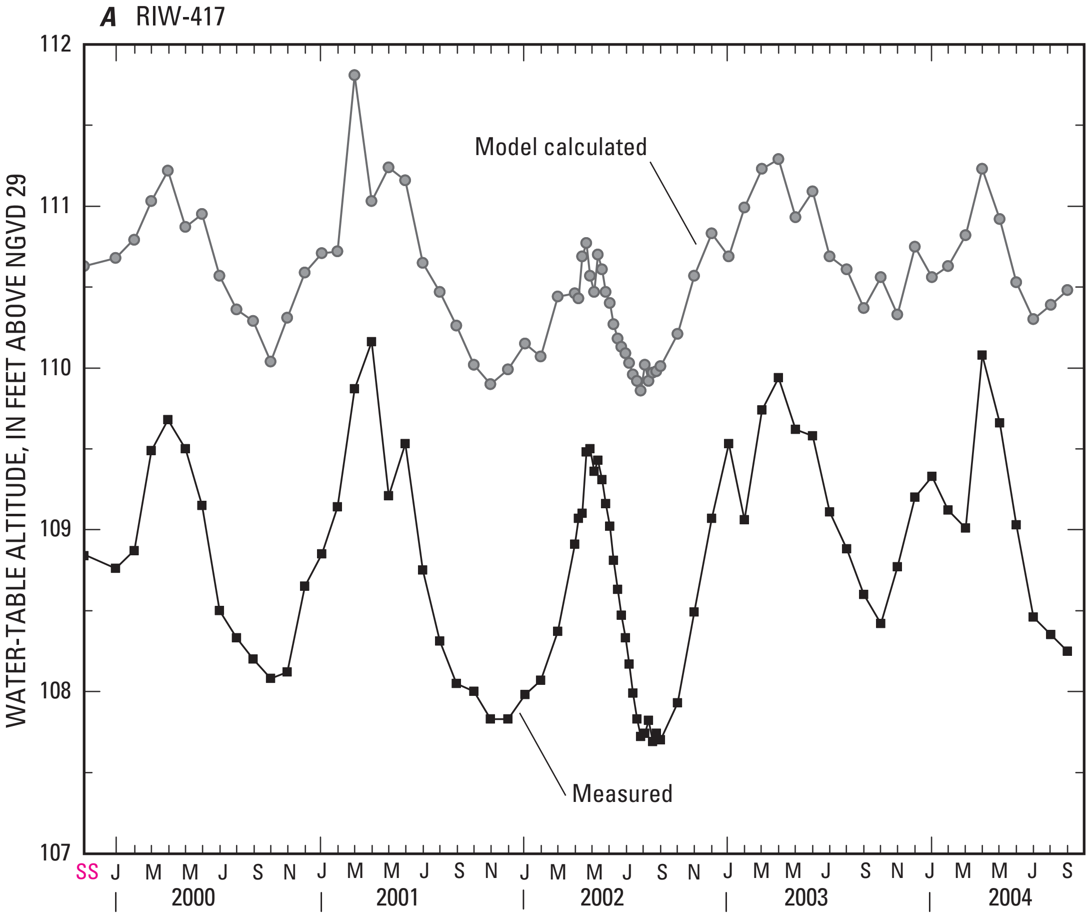


Figure A3–4A. Comparison of measured and model-calculated groundwater levels for well RIW–417 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

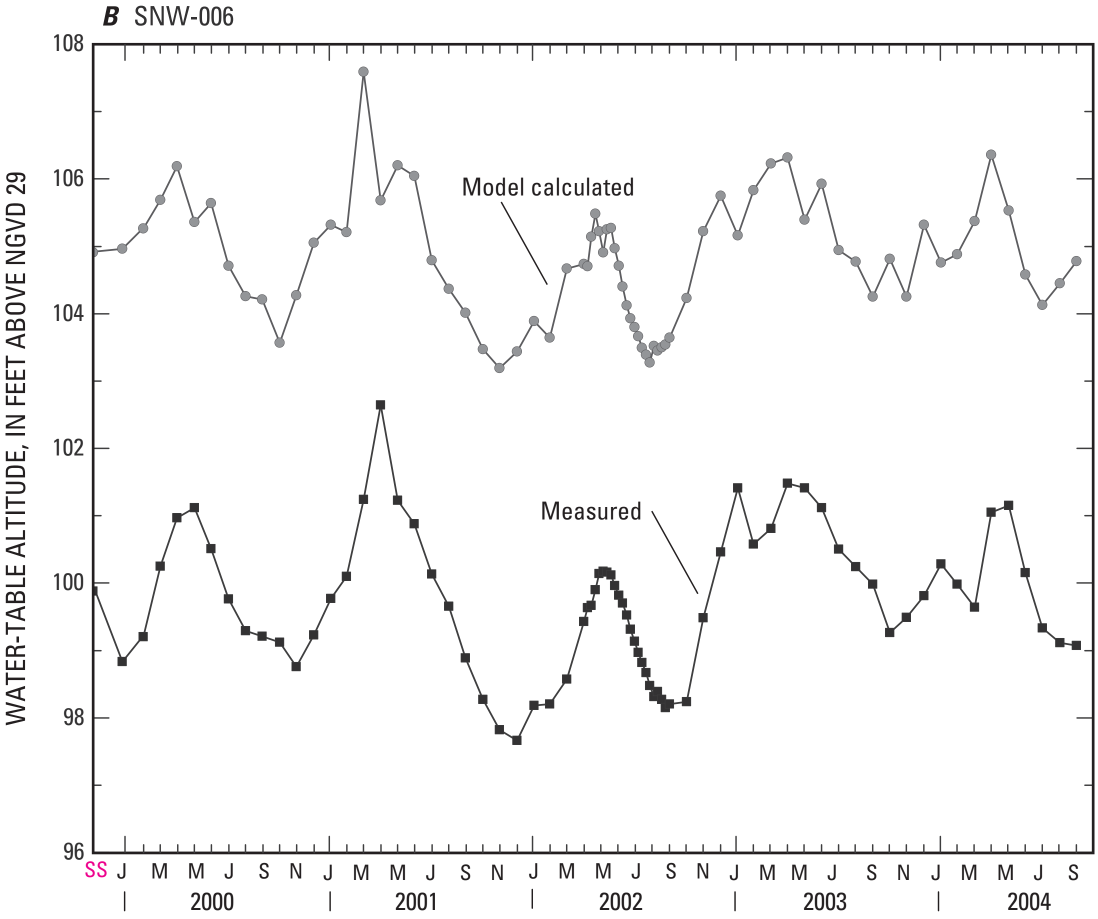


Figure A3–4B. Comparison of measured and model-calculated water levels for well SNW–006 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

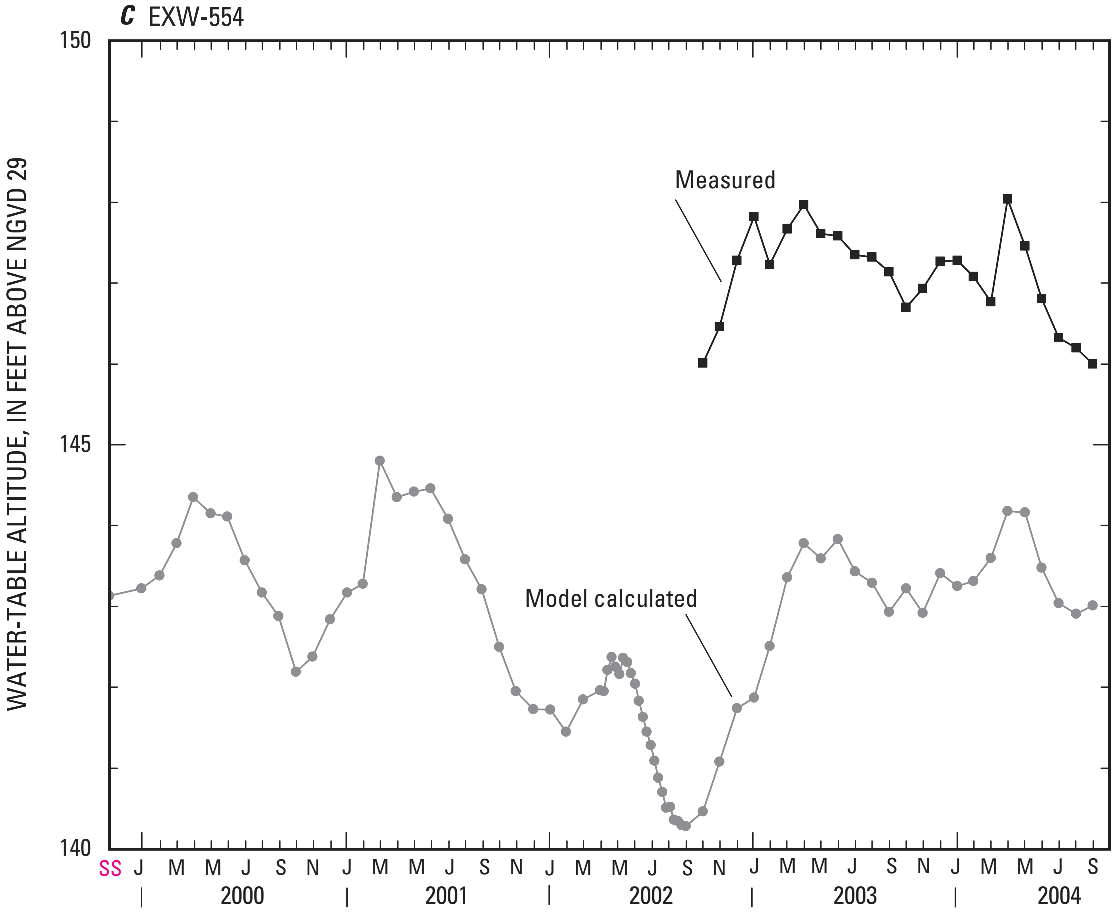


Figure A3–4C. Comparison of measured and model-calculated water levels for well EXW–554 in the lower Wood River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

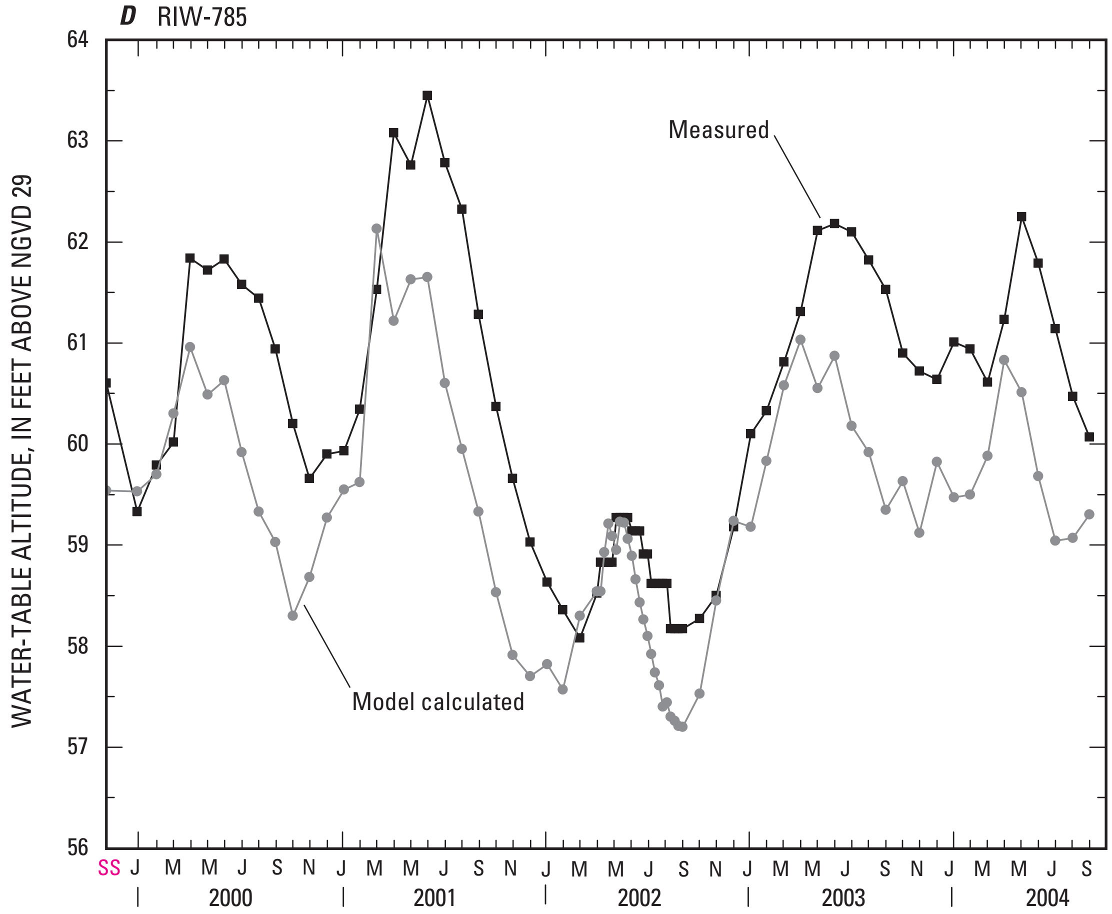


Figure A3–4D. Comparison of measured and model-calculated water levels for well RIW–785 in the lower Wood River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–1. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

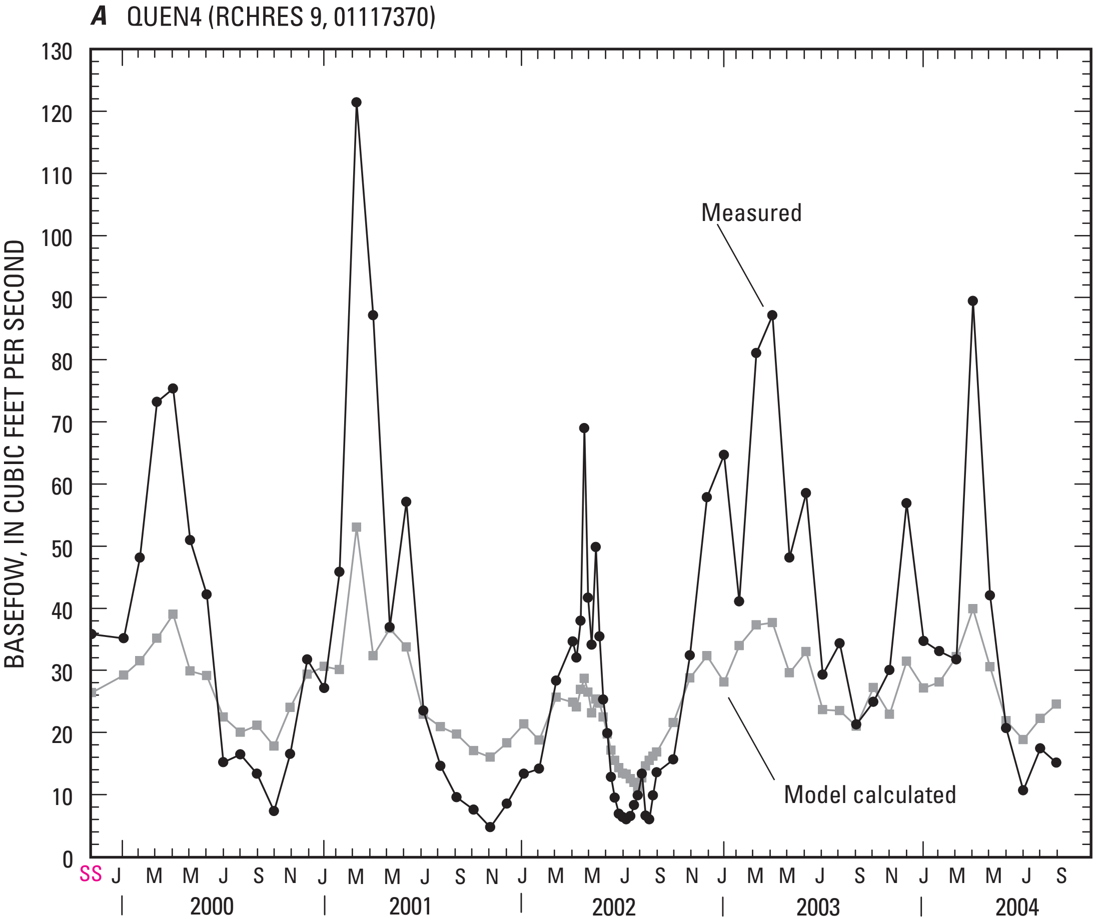


Figure A3–5A. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet QUEN4 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

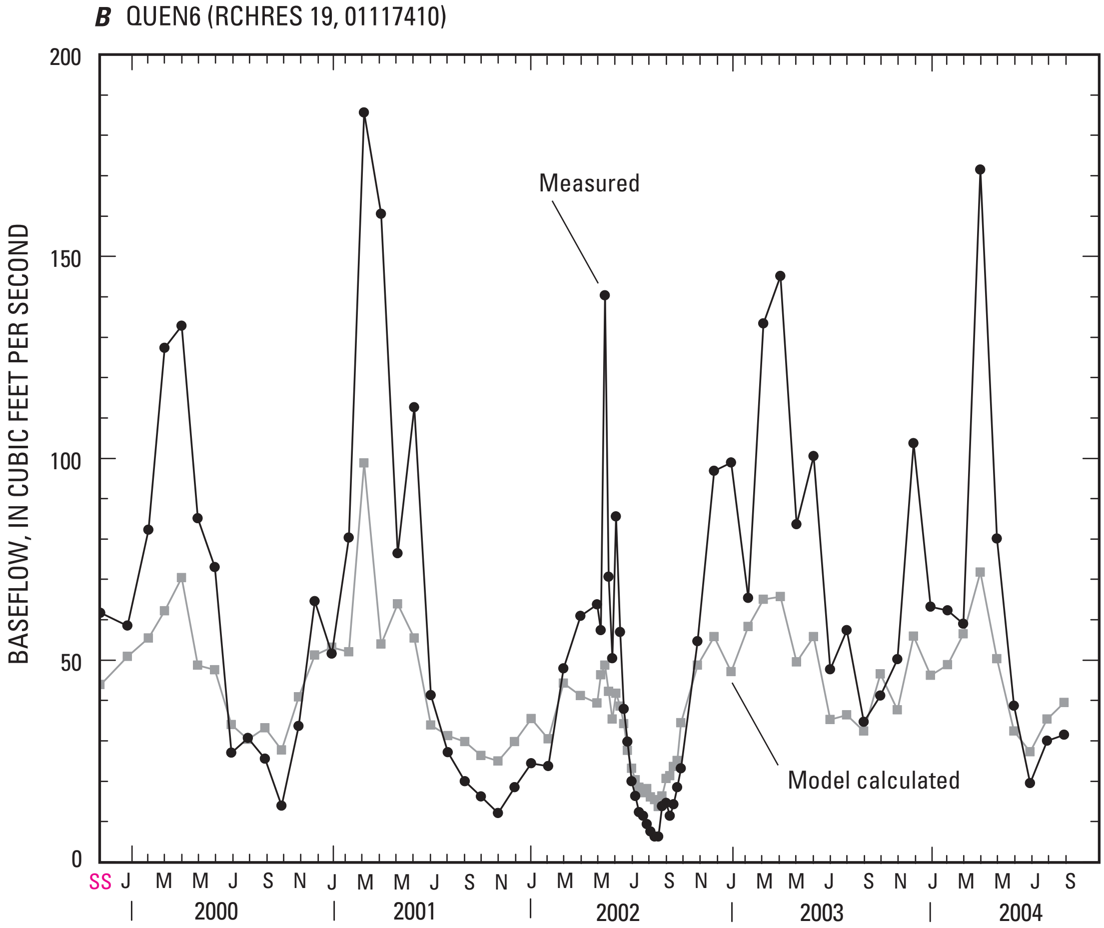


Figure A3–5B. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet QUEN6 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

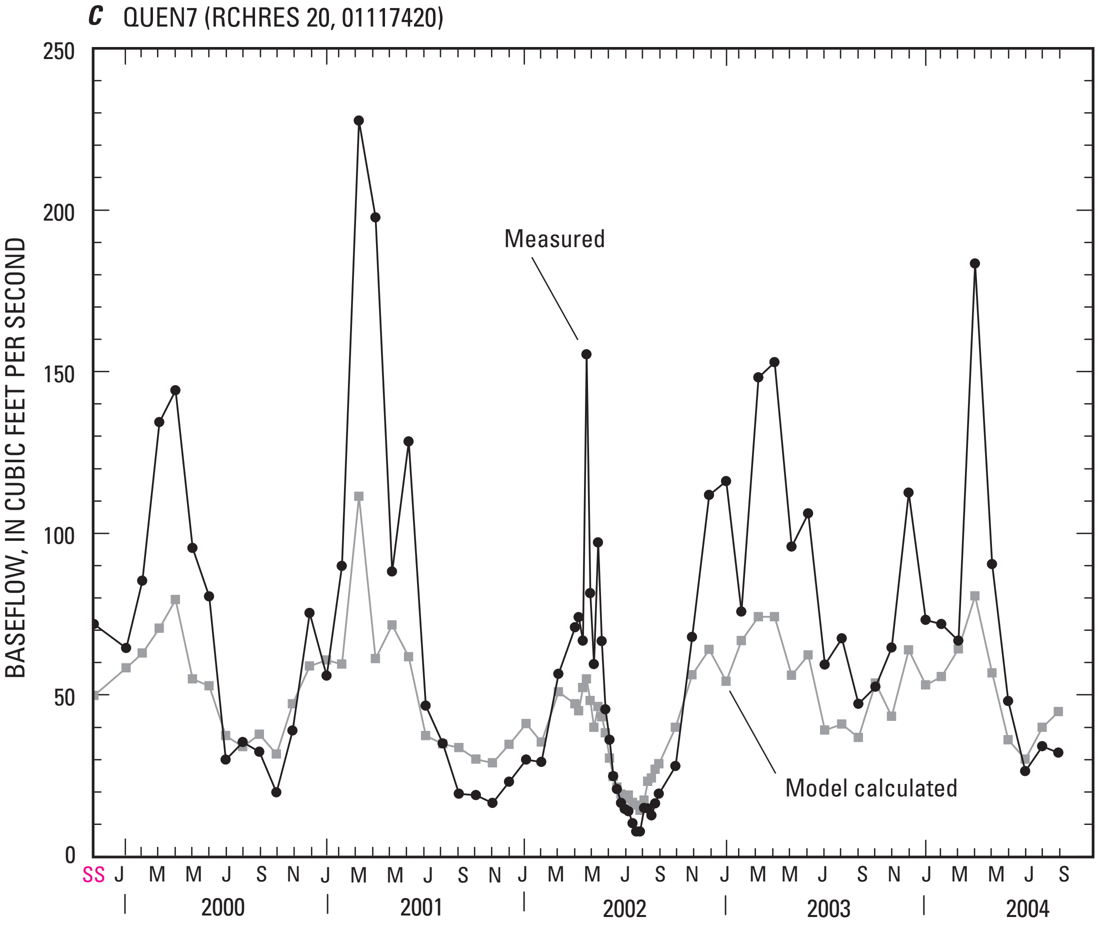


Figure A3–5C. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet QUEN7 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

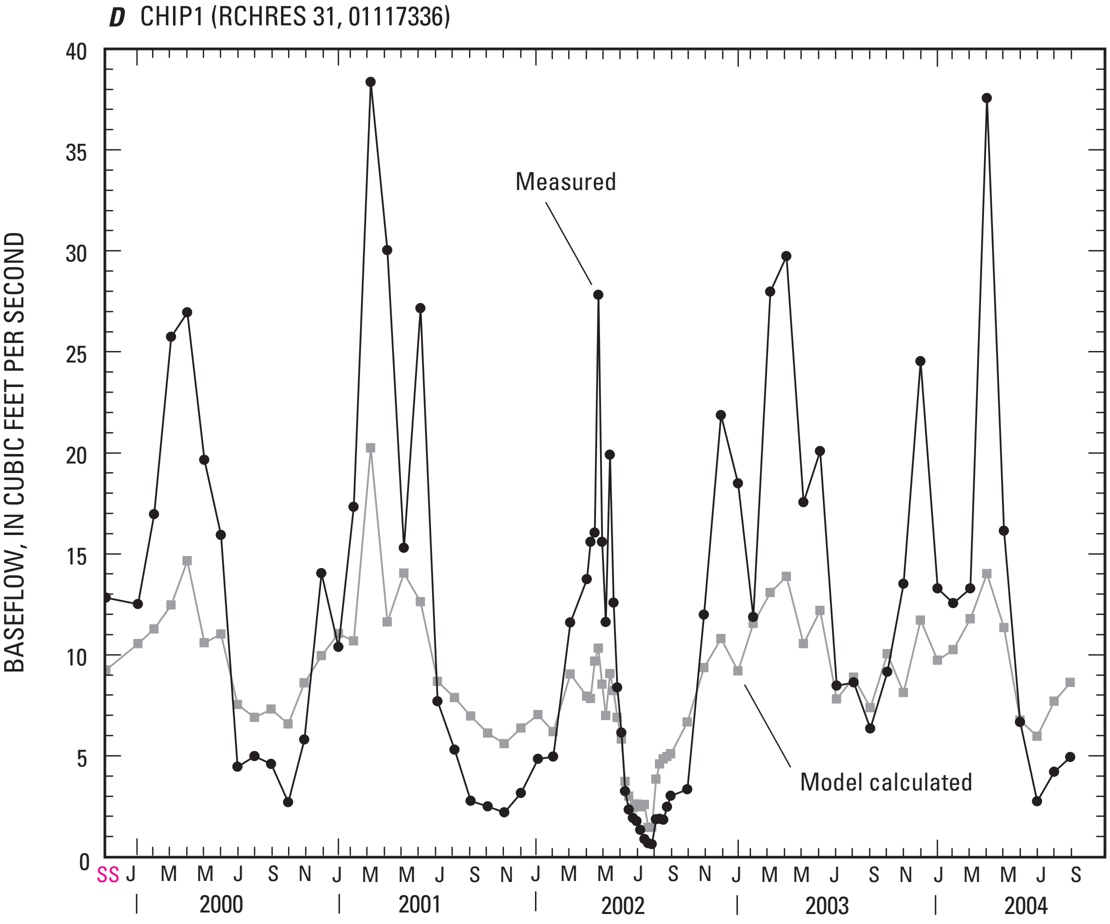


Figure A3–5D. Comparison of measured and model-calculated baseflows for RCHRES outlet CHIP1 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

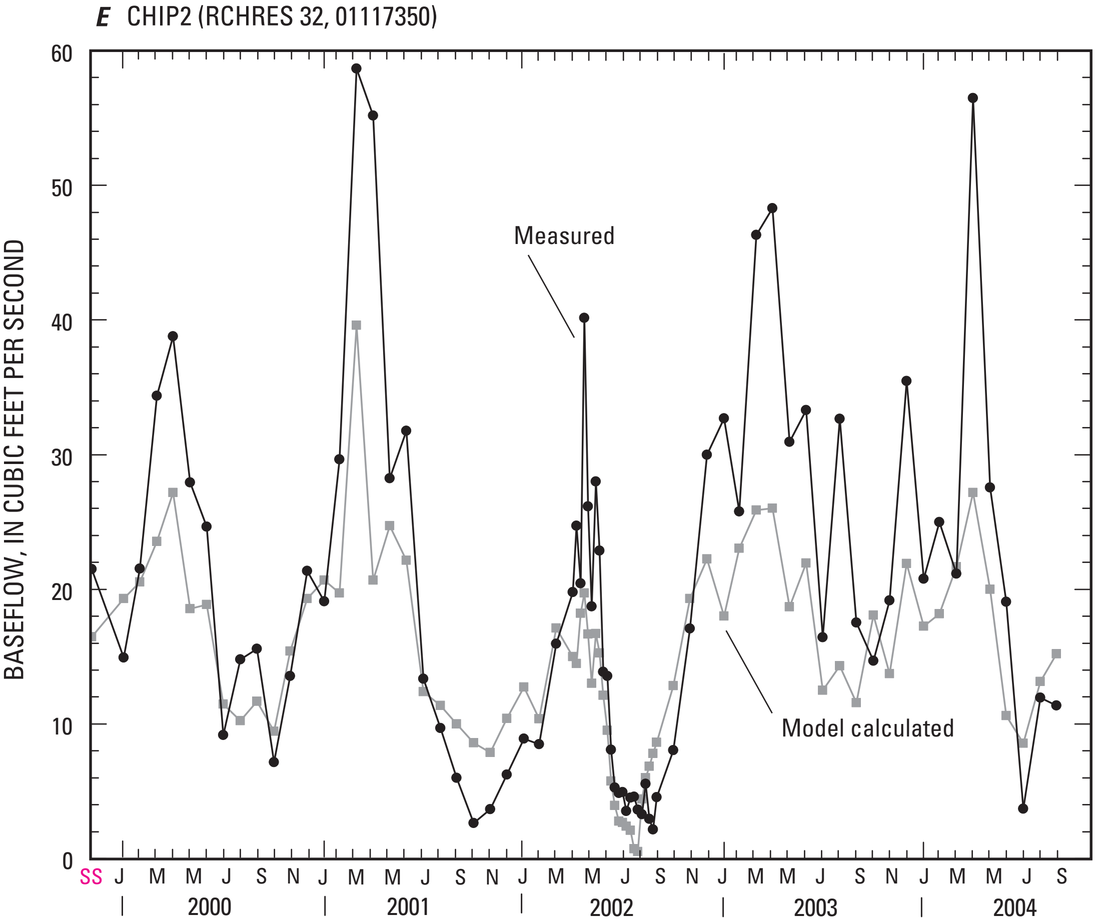


Figure A3–5E. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet CHIP2 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

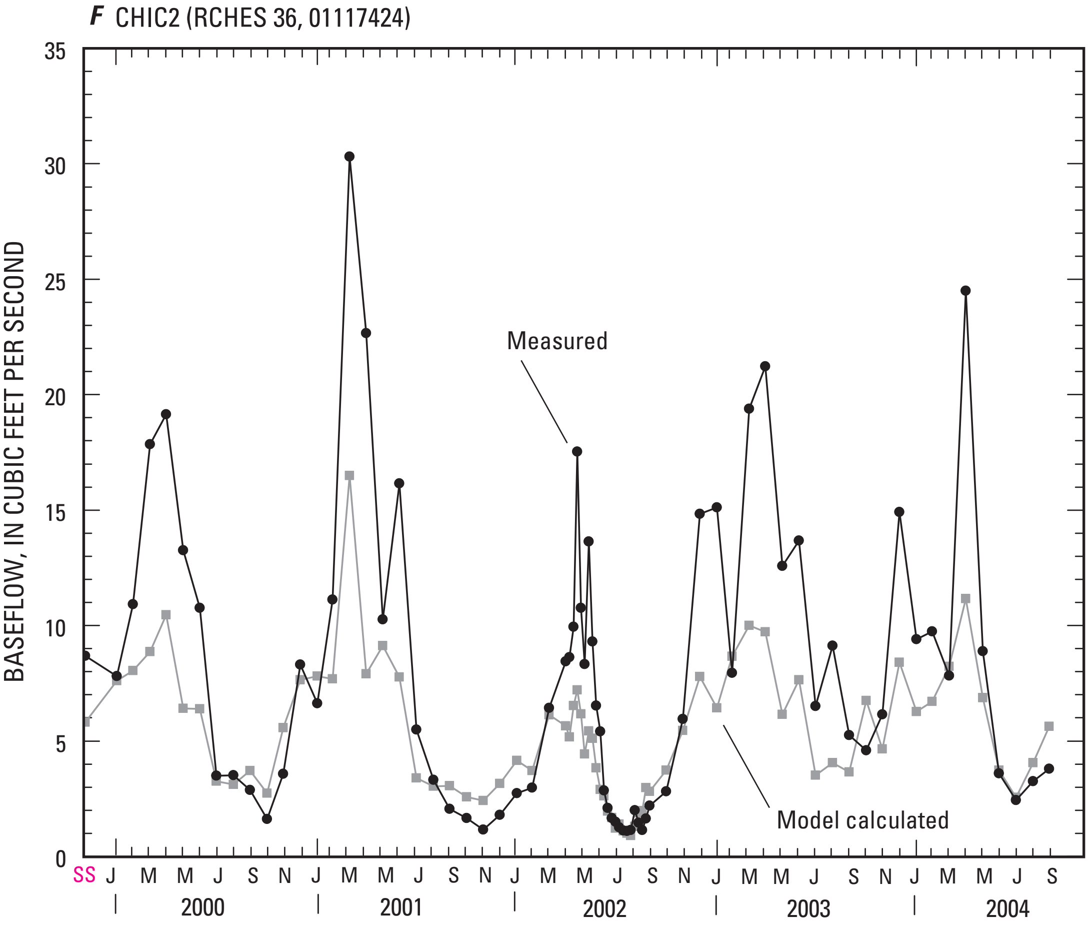


Figure A3–5F. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet CHIC2 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

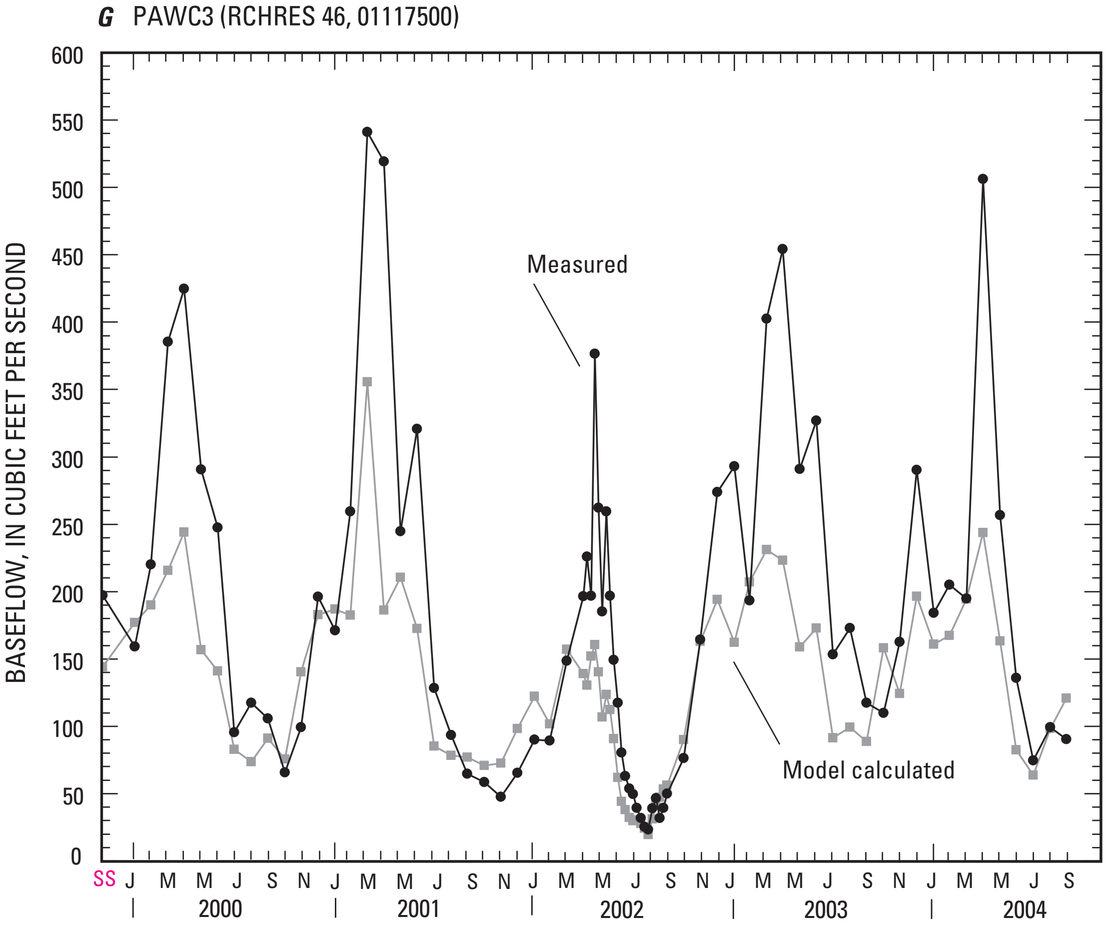


Figure A3–5G. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet PAWC3 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

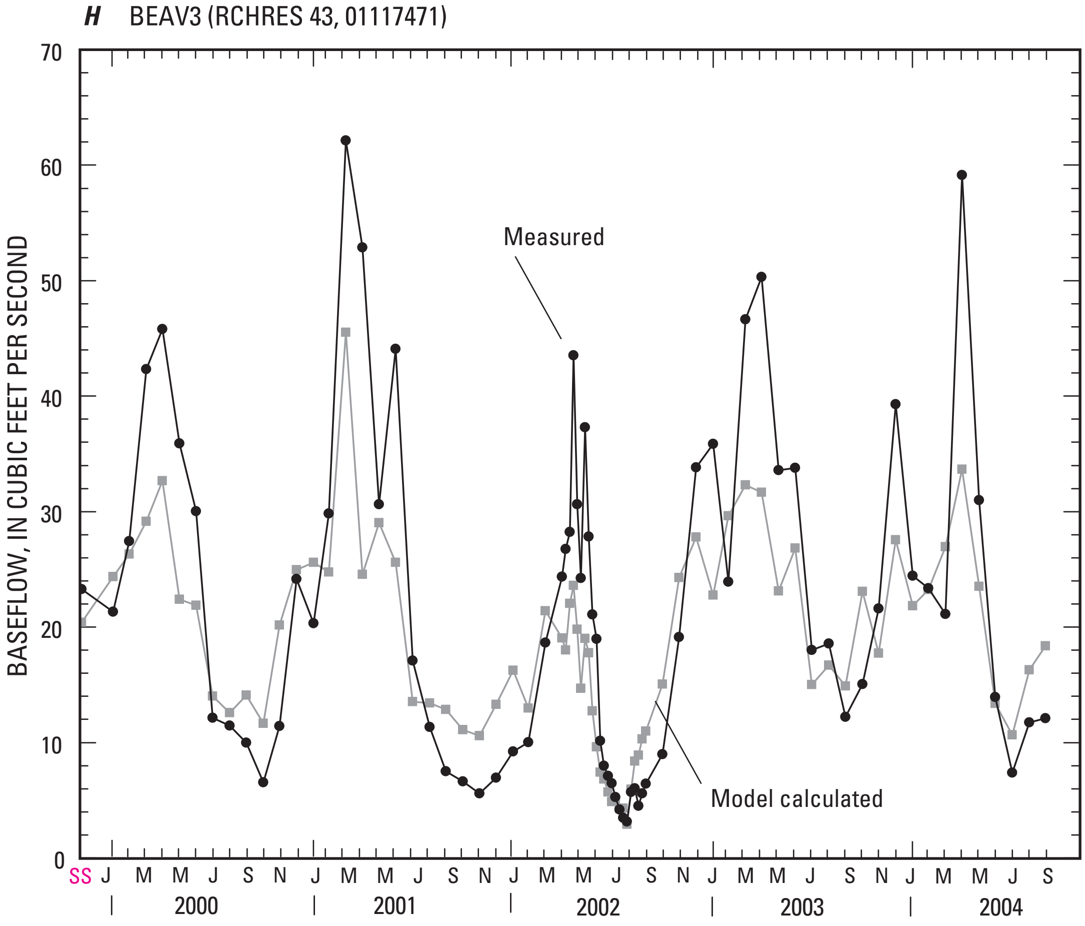


Figure A3–5H. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet BEAV3 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

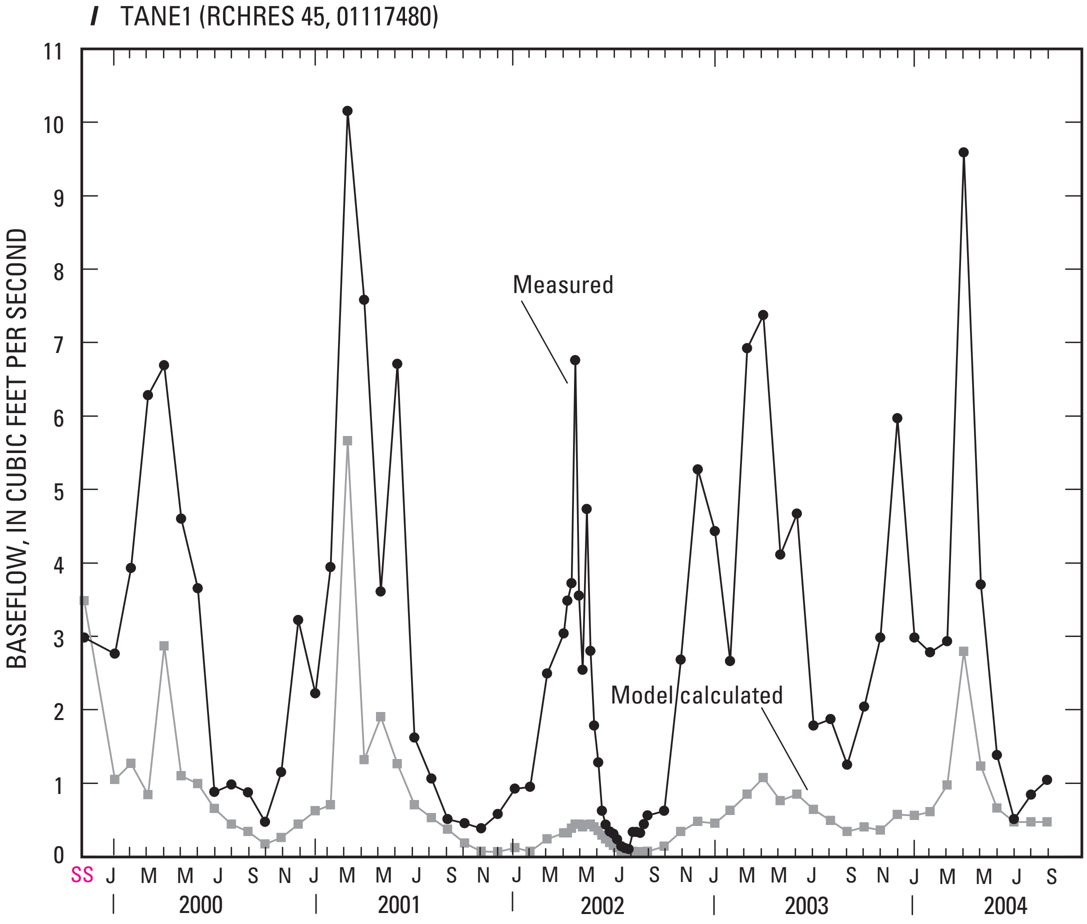


Figure A3–5I. Comparison of measured and model-calculated baseflows for HSPF subbasin outlet TANE1 in the eastern Pawcatuck River model area in the Pawcatuck River Basin, southwestern Rhode Island. Site location shown in figure A3–2. Months shown on x-axis are January, March, May, July, September, and November for years shown. Steady state (SS) represents average steady-state conditions for simulation period.

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