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SIMULATION OF GROUND-WATER FLOW IN AN UNCONFINED SAND AND GRAVEL AQUIFER AT MARATHON, CORTLAND COUNTY, NEW YORK

U.S. GEOLOGICAL SURVEY
Water-Resources Investigation Report 00-4026

Prepared in cooperation with
CORTLAND COUNTY SOIL AND
WATER CONSERVATION DISTRICT

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BY TODD S. MILLER

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Ithaca, New York
2000

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CONTENTS

FIGURES

TABLES

CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	**To Obtain**
	Length	
inch (in.)	2.54	centimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square mile (mi^2)	2.590	square kilometer
	Flow	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
inch per year (in/yr)	25.4	millimeter per year
million gallons per day (Mgal/d)	3.785	cubic meters per day
gallons per minute (gal/min)	0.06309	liter per second
foot per mile (ft/mi)	0.1894	meter per kilometer
	Temperature	
degrees Fahrenheit (˚F)	˚C = 5/9 (˚F-32)	degrees Celsius
	Specific Conductance	
	microsiemens per centimeter at 25˚ Celsius (µS/cm)	

Vertical datum: In this report, “sea level’ refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929) - a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

Simulation of Ground-water Flow in an Unconfined Sand and Gravel Aquifer at Marathon, Cortland County, New York

By Todd S. Miller

Abstract

The Village of Marathon, in Cortland County, N.Y., has three municipal wells that tap a relatively thin (25 to 40 feet thick) and narrow (less than 0.25 mile wide) unconfined sand and gravel aquifer in the Tioughnioga River valley. Only one of the wells is in use because water from one well has been contaminated by petroleum chemicals from a leaking storage tank, and water from the other well contains high concentrations of manganese. The operating well pumps about 0.1 million gallons per day and supplies about 1,000 people.

A three-dimensional, finite-difference ground-water-flow model was used to (1) compute hydraulic heads in the aquifer under steady-state conditions, (2) develop a water budget, and (3) delineate the areas contributing recharge to two simulated wells that represent two of the municipal wells—one 57 feet east of the Tioughnioga River, the other 4,000 feet to the south and 75 feet from a man-made pond. The water budget for simulated long-term average, steady-state conditions with two simulated pumping wells indicates that the principal sources of recharge to the unconfined aquifer are unchanneled runoff and ground-water inflow from the uplands (41 percent of total recharge), precipitation that falls directly on the aquifer (34 percent), and stream leakage (23 percent). Only 2 percent of the recharge to the aquifer is from ground-water underflow into the northern end of the modeled area. Most of the simulated ground-water discharge from the modeled area (78 percent of total discharge) is to the Tioughnioga River; the rest discharges to the two simulated wells (19 percent) and as underflow at the southern end of the modeled area (3 percent).

Results of a particle-tracking analysis indicate that the aquifer contributing area of the northern (simulated) well is 0.10 mile wide and 0.15 mile long and encompasses 0.015 square miles; the contributing area of the southern (simulated) well is 0.20 mile wide and 0.11 mile long and encompasses 0.022 square miles. The average traveltime of ground water from the valley wall to either simulated well is about 1.5 years, calculated on the basis of an assumed aquifer porosity of 0.3. The flowpath analysis indicates that both contributing areas contain surface-water sources of recharge—the Tioughnioga River and Hunts Creek contribute water to the northern well, and a pond and a small tributary contribute water to the southern well.

Ground-water temperature in an observation well between the Tioughnioga River and the municipal well fluctuated several degrees Fahrenheit in response to pumping of the municipal well. This temperature fluctuation, in conjunction with the pumping well causing a ground-water gradient from the Tioughnioga River to the pumping well (ground-water levels in the pumping well were generally greater than 3 ft lower than that of the Tioughnioga River), indicate that there is a hydraulic connection between the river and aquifer at this site.

INTRODUCTION

Aquifers are principal sources of water supply for many communities. Many municipal wells that tap these aquifers have been contaminated or are under threat of contamination because of the lack of knowledge of contributing areas that supply recharge to these wells. An important step in protecting municipal ground-water supplies from contamination is to identify the contributing areas to water-supply wells and the potential sources of chemical contaminants that are within the contributing area. Cortland County promotes implementing protection plans as the most cost-effective means of ensuring a safe drinking-water supply. The U.S. Environmental Protection Agency (1987) has recommended that municipalities adopt "Wellhead Protection" strategies to regulate land use in areas that contribute recharge to public-supply wells. The objectives of these strategies are to delineate the land-surface area from which water enters the ground-water system and flows to municipal wells, and to set regulations that minimize the possibility of contamination within these areas.

The Village of Marathon, in Cortland County, has three wells that tap a relatively thin (25- to 40-ft thick) and narrow (less than 0.25 mi wide) unconfined sand and gravel aquifer in the Tioughnioga River valley (fig. 1). In 1998, only one well (well 452, fig. 1) near the Tioughnioga River in the northern part of the village was used; the second well (well 450, fig. 1), also in that vicinity, has been abandoned because the water has been contaminated by petroleum chemicals. The third well (well 459, fig.1), near a pond in the southern part of the study area, is not used because the water contains high concentrations of manganese. The village is concerned about the quality of water in the Tioughnioga River because ground-water/surface-water interaction is common in shallow water-table aquifers, and induced infiltration of surface water to the one remaining well is likely. Past flooding of the Tioughnioga River has also posed a threat of ground-water contamination of the municipal well. The village plans to have two wells (one for the principal supply and one for backup) and needs information on the aquifer to protect and manage the village supply, which serves about 1,000 people.

During 1997-98, the U.S. Geological Survey (USGS), in cooperation with the Cortland County Soil and Water Conservation District, conducted an investigation of a part of the sand and gravel aquifer at Marathon to collect and compile data needed to delineate contributing areas to two hypothetical wells that represent the municipal supply wells. The study entailed a literature search, compilation of available hydrogeologic data, synoptic ground-water-level and streamflow measurements made during June 11-12, 1997, continuous recording of ground-water levels and water temperature in an observation well between the Tioughnioga River and the municipal pumping well 452, and development of a finite-difference-ground-water-flow model to simulate ground-water flow in the aquifer.

Purpose and Scope

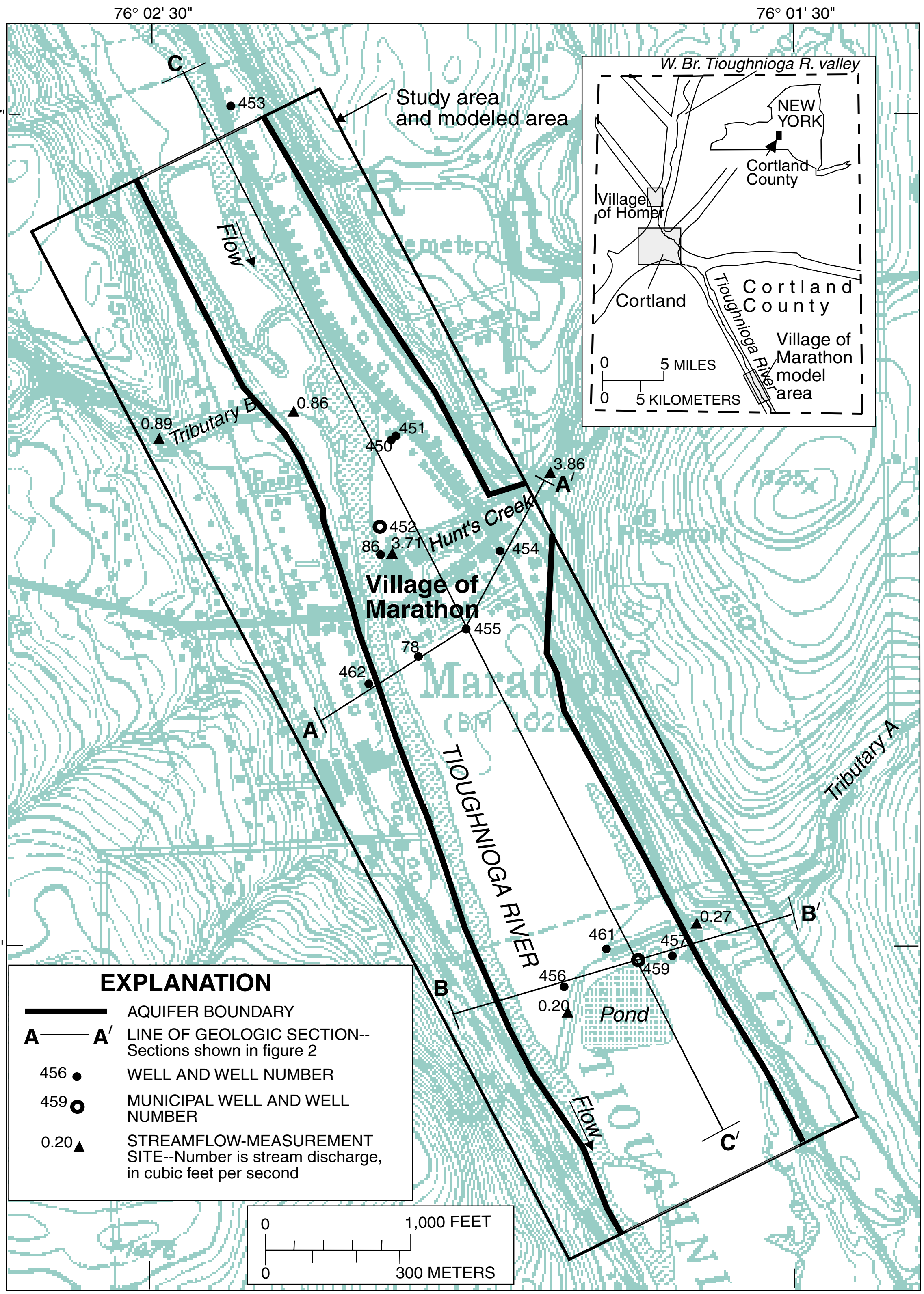
This report describes: (1) the hydrogeologic setting of the Marathon area, (2) the design of the ground-water flow model, and (3) the results of simulations made with the model. The report presents tables of the simulated inflows and outflows to the aquifer as computed for average, steady-state conditions, and includes maps showing the distribution of hydraulic head in the aquifer, and the contributing areas of the two municipal wells.

Methods

The three-dimensional, numerical-ground-water flow model MODFLOW, developed by the USGS (McDonald and Harbaugh, 1988), was used to simulate steady-state conditions in the aquifer and to compute ground-water levels and a water budget for the study area. The particle-tracking program MODPATH, also developed by the USGS (Pollock, 1994), was used to delineate the areas that contribute recharge to two simulated wells that represent the municipal wells. Levels were run to (1) wells used to calibrate the model, (2) to the top and bottom of the pond in the southern part of the study area, and (3) in the channels of Tioughnioga River, Hunts Creek, and several other small streams. Depth soundings were made to obtain bottom elevation of the pond. Water-level and streambed-elevations measurements are accurate to a tenth of a foot where levels were run.

Acknowledgments

Thanks are extended to Walter Ackley and Gary Lawrence (both former water supervisors for the Village of Marathon); and to Michael Root (current



Base from U.S. Geological Survey, Marathon, N.Y., 1:24,000, 1994

Figure 1. *Location of modeled area and lines of geologic sections at Marathon, Cortland County, N.Y.*

water supervisor) for providing data and access to wells. Mr. Ackley knew much more about the aquifer than he gave himself credit for.

GEOLOGY

The Marathon area is within the glaciated and dissected Allegheny Plateau, which consists of gently folded layers of shale with some siltstone, sandstone, and limestone that dip southward from 10 to 40 ft/mi. These rocks are part of the “Catskill Delta” complex (Woodrow and Sevon, 1985), which was deposited in marine seas during late Devonian time. These rocks were then uplifted above sea level millions of years later, then eroded to a nearly flat plain by the middle of the Cenozoic (Isachsen and others, 1991). They were again uplifted during Late Cenozoic time to form the Allegheny Plateau, which was subsequently dissected by streams and eroded by glaciers to form a smooth, hilly terrain.

The entire region was glaciated several times during the Quaternary Period, which began 1.6 million years ago and ended about 10,000 years ago (Isachsen and others, 1991). Some of the sediments that were eroded by the ice either became entrained in the glacier or were dragged along its bottom, where some of the material was ground up and later deposited as till on top of bedrock in most places. The till in the study area consists of poorly sorted clay, silt, sand, and stones that were compacted by the ice. Till and bedrock, which are much less permeable than sand and gravel, form the bottom of the overlying sand and gravel aquifer in the Tioughnioga River valley (fig. 2).

Alluvial and glaciofluvial deposits in the study area form a 25-to-40-ft thick sand and gravel aquifer (fig. 2, geologic sections *A-A', B-B', and C-C'*). The upper part of the aquifer consists of recent alluvial deposits (flood-plain, fan, and channel sediments), and the lower parts may consist of glaciofluvial sediments such as kame and outwash deposits. The low areas adjacent to the Tioughnioga River have flood-plain deposits that consist of silt and fine sand and are only few feet thick and are unsaturated, except during extremely wet periods and floods. The fan and channel deposits consist of sand and gravel with various amounts of silt; the alluvial deposits also contain decaying organic matter.

Kames form hummocky terraces along the lower flanks of the valley walls and probably extend below the alluvial deposits in the valley bottom; they consist of fair- to well-sorted silt, sand, and gravel that were deposited atop and next to glaciers and collapsed when the ice melted. Outwash is fluvial channel deposits deposited by meltwaters that flowed from the glacier southward through the Tioughnioga River valley when the glacier was north of Marathon. Outwash probably underlies the alluvial deposits in the valley and consists of well-sorted sand and gravel (fig. 2A). The outwash and kame deposits in the Marathon area typically contain many erratic clasts (rocks that are foreign to the Marathon area) that were transported by the glacier and meltwater from several tens of miles to the north, whereas the alluvial deposits tend to be mostly local shale and siltstone clasts that were eroded by upland streams and deposited in the valley, where they became reworked by the Tioughnioga River.

The Marathon area undergoes frequent flooding because the Tioughnioga River (about 115 ft wide) has a low stream gradient and occupies a relatively large part (8 to 13 percent) of a fairly narrow valley (900 to 1,400 ft wide). Large streams that occupy relatively small valleys, or small streams that occupy relatively large valleys are known as misfit streams. The Tioughnioga river became a misfit stream because of glacial diversion of drainage. The valley at Marathon during preglacial time was on the south side of a nearby drainage divide (fig. 3B), and only a small, southward-flowing stream occupied the valley in the Marathon area. During the last glaciation, however, several streams near Cortland (about 15 mi north of Marathon) were diverted by the glacier from their preglacial course through the Fall Creek valley to the west (fig. 3B) and were rerouted to the south into its present course through the narrow valley now occupied by the Tioughnioga River. Most land development in the Tioughnioga River valley, from Cortland to Marathon, is on high ground, such as on alluvial fans and hillsides, because flooding is common in the valley.

HYDROLOGY

The upper part of the sand and gravel aquifer at Marathon consists of alluvial-fan and channel deposits, and the lower part probably consists of outwash and kame deposits (fig. 2). The overbank flood deposits are not part of the aquifer because they are only a few feet thick and are probably unsaturated most of the time. The top of the aquifer is the water

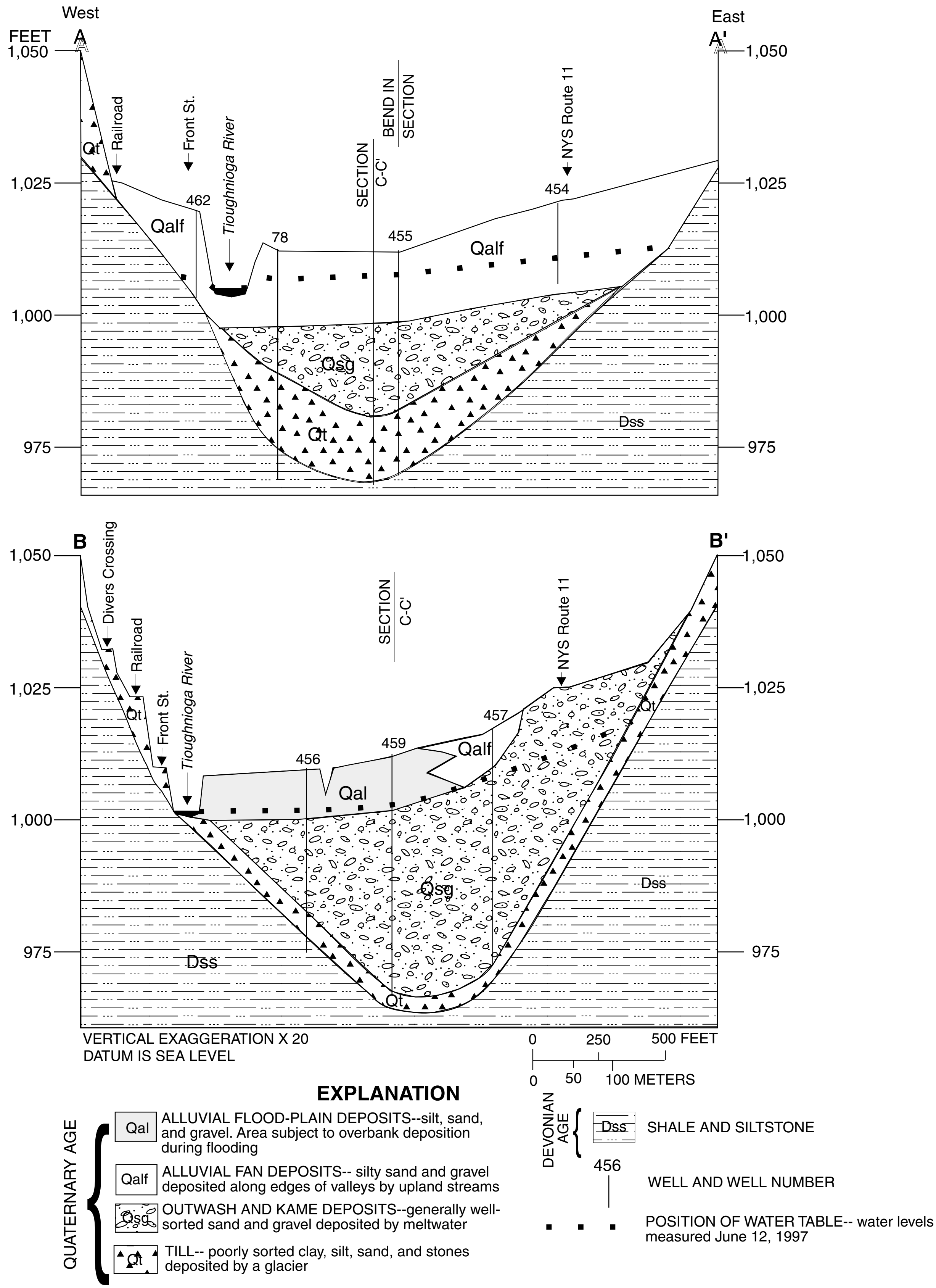
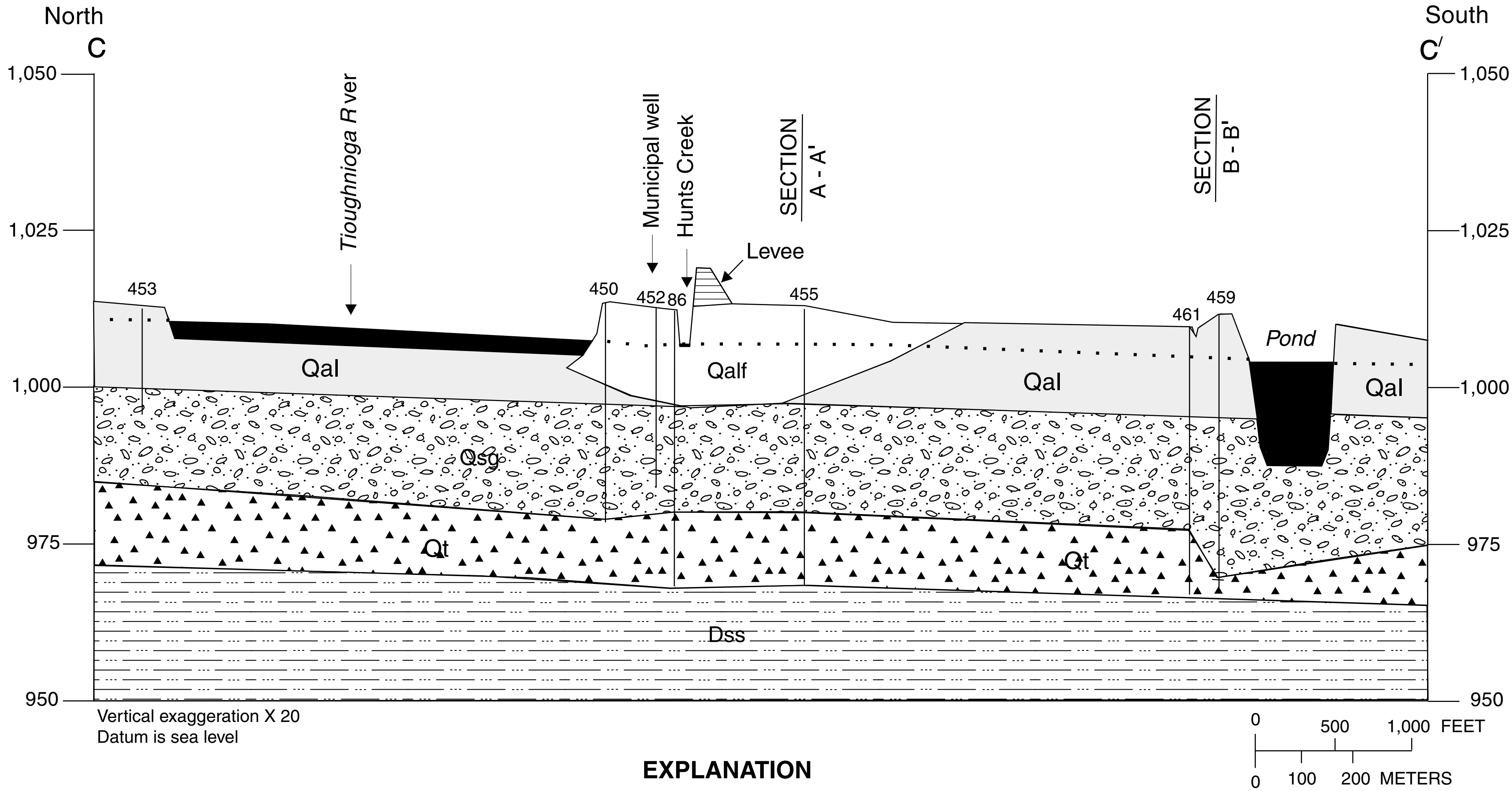


Figure 2. *Hydrogeologic sections A-A′, B-B′ and C-C′ in modeled area, Marathon, N.Y. (Lines of section are shown in fig. 1)*



EXPLANATION

QUATERNARY AGE

- Qal — ALUVIAL FLOOD PLAIN AND CHANNEL DEPOSTS-- silt, sand, and gravel. Area subject to overbank deposition during flooding
- Qalf — ALLUVIAL FAN DEPOSITS-- silty sand and gravel deposited along edges of valleys by upland streams
- Qsg — GLACIOFLUVIAL DEPOSITS-- generally well-sorted sand and gravel deposited by meltwater
- Qt — TILL- poorly sorted clay, silt, sand, and stones deposited by a glacier

DEVONIAN AGE

- Dss — SHALE AND SILTSTONE

453 — WELL AND WELL NUMBER -number assigned by USGS

. . . . — WATER TABLE - water levels measured during June 12, 1997

Figure 2. (continued) *Hydrogeologic sections A-A′, B-B′, and C-C′ in modeled area, Marathon, N.Y. (Lines of section are shown in fig. 1)*

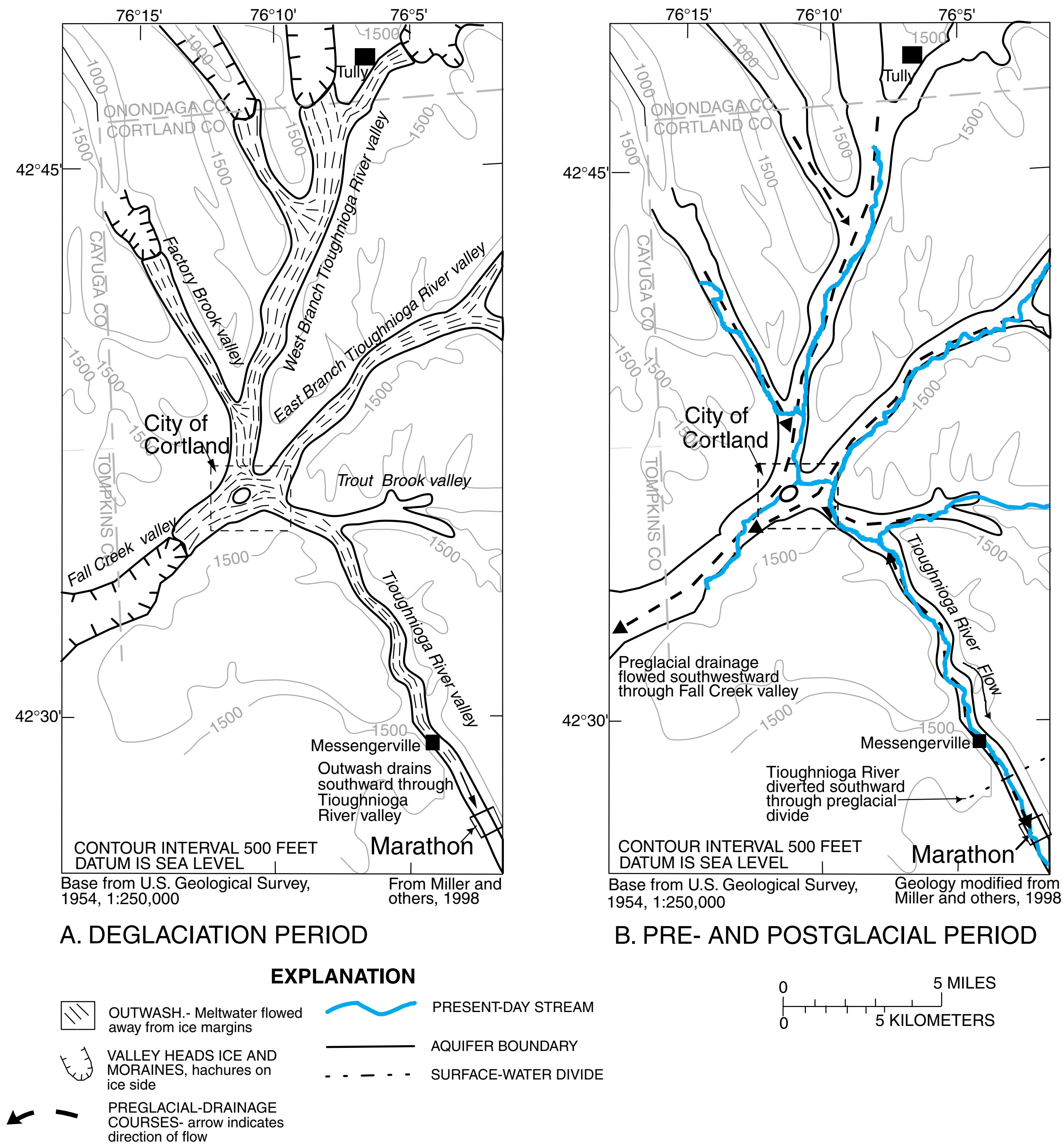


Figure *3. Drainage in the Tioughnioga River valley in Cortland County, N.Y.: A. During deglaciation. B. During pre-and postglaciation period. (Location is shown in fig. 1, inset map)* table, and the bottom is either the top of till or top of bedrock. The saturated thickness (the depth from the water table to the bottom of the aquifer) is typically 20 to 30 ft. The aquifer is bounded laterally by the till-covered bedrock hillsides.

Sources of Recharge

The aquifer in the modeled area receives recharge from five sources: (1) precipitation that falls directly on the aquifer, (2) runoff (sheet flow over land surface) from unchanneled, till and bedrock hillsides that are adjacent to the sand and gravel deposits in the valley, (3) ground-water seepage from adjacent till deposits and bedrock, (4) stream leakage (natural leakage, and leakage induced by nearby pumping wells), and (5) southward inflow from the aquifer at the northern end of the modeled area. Part of the precipitation that falls on the aquifer is returned to the atmosphere by evapotranspiration; the remainder infiltrates to the water table.

Rates of recharge from precipitation, as well as from other sources, vary seasonally. Most of the precipitation that falls during the nongrowing season (typically from mid-October through the end of April) infiltrates into the ground and recharges the aquifer, whereas most of the precipitation that falls during the growing season (May through September) is lost through evapotranspiration. Annual precipitation in the Marathon area is estimated to average about 40 in., slightly less than the 41.2 in. measured at the Cortland weather station during 1973-92 (National Oceanic and Atmospheric Administration, 1992), because the Marathon area receives slightly less lake-effect snow than Cortland.

The average annual recharge to surficial sand and gravel aquifers in the northeastern United States is about equal to the long-term average annual stream runoff, which, for 30 years (1951-80) in southern Cortland County averaged about 22 in. (Lyford and Cohen, 1987). Therefore, the average volume of recharge that the aquifer receives from precipitation that falls directly on the aquifer is 47,700 ft^3/d, as calculated from an average annual recharge rate of 22 in. (0.00502 ft/d) multiplied by the surface area of the aquifer (9,500,000 ft^2) in the modeled area. The average annual evapotranspiration is estimated to be 18 in. (1.5 ft) and equals average annual precipitation (40 in.) minus average annual runoff (22 in).

Recharge to valley-fill aquifers from adjacent unchanneled hillsides includes surface runoff and lateral flow of ground water from the uplands; this water flows toward the valley and seeps into the aquifer along its edges. All precipitation that is not lost through evapotranspiration in the uplands adjacent to the aquifer is assumed to become either runoff or ground-water recharge that eventually flows into the valley and seeps into the aquifer.

The daily amount of recharge from runoff from adjacent unchanneled hillsides can be calculated by the following equation:

$$R = \frac{(P - ET) \times DA}{t}; \qquad (1)$$

where R = average annual rate of recharge from runoff from unchanneled hillsides, in ft^3/d;
P = average annual precipitation of 40 in. (3.4 ft);
ET = average annual evapotranspiration (1.5ft), as calculated from above;
DA = drainage area of hillside (12,425,000 ft^2); and
t = 365 days.

Solution of the above equation indicates that the rate of recharge from runoff from unchanneled hillsides along the edges of the aquifer is 57,900 ft^3/d.

Tributaries to the Tioughnioga River that drain till-and-bedrock basins in the uplands are sources of aquifer recharge where they enter the main valley and flow over permeable sand and gravel deposits. The rate of recharge from these tributaries is controlled largely by the vertical hydraulic conductivity and thickness of the streambed, and by the hydraulic heads in the stream and aquifer. Rates of recharge from tributaries were estimated from the losses of water calculated from streamflow measurements made on most of the large tributaries in the study area on June 11, 1997, during a period judged to represent median annual flow conditions. Values are given in table 1. The median annual condition was defined as that when the flow duration at the stream gage in the Tioughnioga River at Cortland (station 01509000) was near 50 percent (285 ft^3/s) (Hornlein and others, 1997).

Table1. *Losses of flows from tributary streams to the unconfined aquifer in the Marathon, N.Y. modeled area, June 11, 1997.*
*[*Values are in cubic feet per second; location of tributaries and streamflow-measurement sites are shown in fig. 1.*]*

Tributary	*Streamflow loss (-) between measurement site near valley wall and near mouth, during June 11, 1997*
Hunts Creek	-0.15
Tributary A	-0.07
Tributary B	-0.03
Estimated loss from other tributaries not measured	-0.05
Sum of tributary losses	-0.30

Ground-Water Discharge

Water in the aquifer discharges into the Tioughnioga River, to the Village of Marathon's municipal well, and as underflow out of the southern end of the study area. Ground water in most glaciated river valleys in the Northeast typically discharges to the main trunk stream. Water levels in most wells are higher than the surface of the Tioughnioga River, indicating that most ground water is discharging to the river. This is generally the condition except in the vicinity of the municipal pumping well, where ground-water levels are lower than the river during pumping periods.

Withdrawal of water from a shallow aquifer that is directly connected to a surface-water body affects the exchange of water between these two water bodies. Pumping water from the Marathon municipal well not only captures some of the ground water that would have discharged into the Tioughnioga River, it also induces some water from the river to flow into the aquifer.

Surface-Water and Ground-Water Interaction

Ground-water temperatures do not fluctuate as frequently, nor as widely or as rapidly, as do surface-water temperatures. Therefore, the temperature of ground water and surface water at a given site can be used in conjunction with ground-water levels to indicate the degree of hydraulic connection between a stream and an adjacent aquifer. Water temperature at depths greater than 50 ft in coarse-grained aquifers in New England does not fluctuate but, rather, is near the mean annual air temperature of a region (47 ^{o}F in the Marathon area) if the vertical ground-water velocity is negligible (Lapham, 1989). Significant downward flow, such as near pumping wells, increases the depth to which ground-water temperature fluctuations are found.

Ground-water temperature [sprobably] fluctuates near the municipal well in the northern part of the study area (well 452, fig. 1) because the well is screened at a shallow depth (22-28 ft below land surface), is only 57 ft east of the Tioughnioga River (fig. 4), and pumping is likely to induce the infiltration of surface-water. A temperature probe was installed 20 ft below land surface in an observation well (screened from 20-25 ft below land surface) between the municipal well and the river—19 ft west of the pumping well and 38 ft east of the Tioughnioga River (fig. 4)—to verify this hydraulic connection. Small

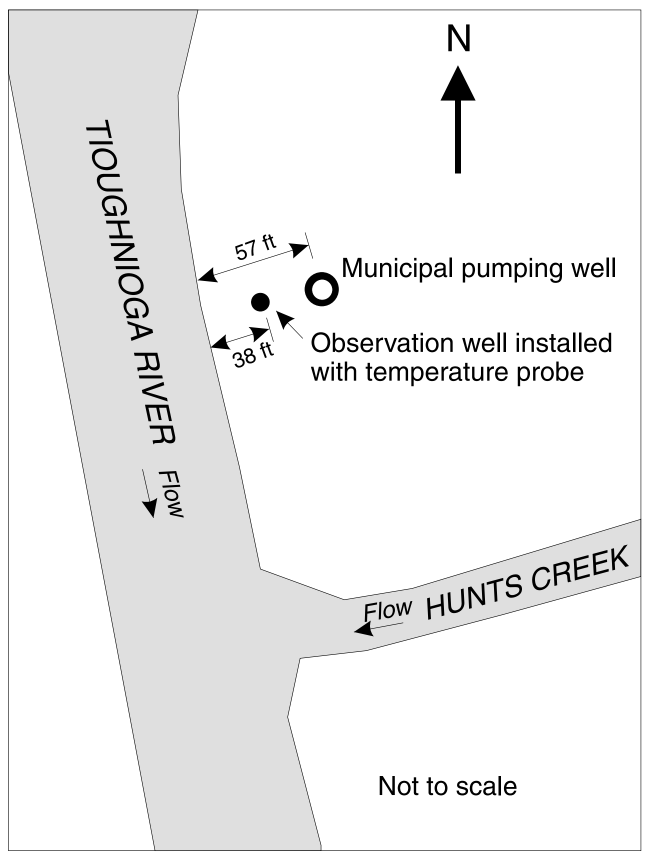


Figure 4. *Location of northernmost municipal pumping well (location is shown in fig. 1) and observation well with temperature probe in relation to Tioughnioga River and Hunts Creek at Marathon, N.Y.*

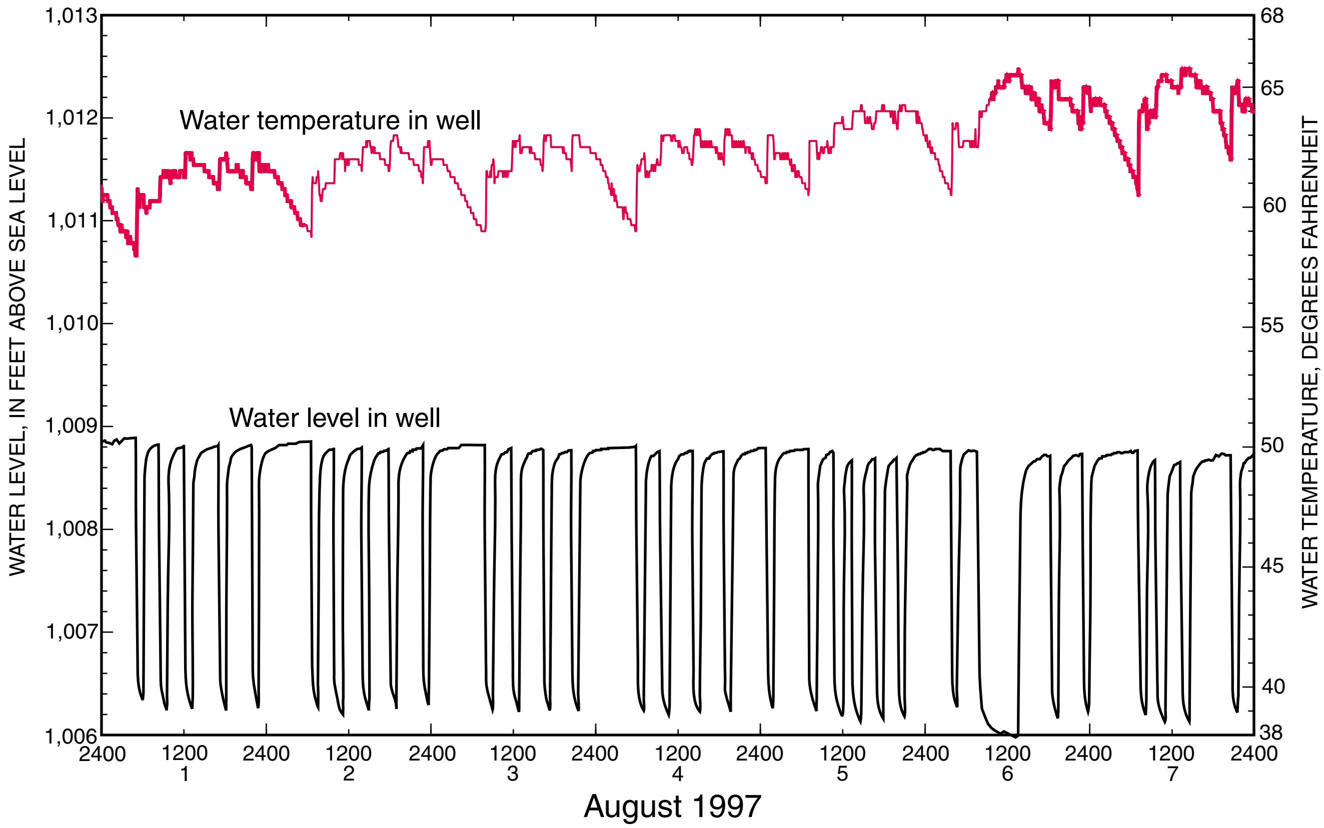


Figure 5. *Ground-water temperature and water level in observation well between municipal pumping well 452 and Tioughnioga River, Marathon, N.Y., August 1-7, 1997. (Location shown in fig. 4.)*

fluctuations in temperature would be expected even during nonpumping periods because the aquifer is thin in this area, whereas large fluctuations in temperature, which were found in the observation well and corresponded to the river-water temperature, indicate induced recharge from the river to the pumping well.

Plots of ground-water temperature and water levels at this observation well during August 1-7, 1997 are shown in figure 5; those from July 1997 through June 1998 are shown in figure 6. The ground-water temperature fluctuations correspond to the daily water-level drawdowns and recoveries caused by the pumping well, which is turned on and off two or three times a day. The temperature fluctuation in response to pumping indicates that the pumping well is inducing water from the Tioughnioga River in addition to capturing water from the aquifer's shallow zone (fig. 7), which undergoes warming and cooling in response to air temperature. Pumping of the municipal well typically causes a drawdown of about 3 ft below river stage in the observation well; this indicates that the hydraulic gradient slopes from the river to the well during pumping and water is induced from the river.

In summer, when surface water is warmer than ground water, the water temperature in the observation well rises when the municipal well is being pumped and falls when the well is turned off (fig. 5). The reverse is true in winter, when surface water is colder than ground water-the water temperature in the well falls when the well is being pumped and rises when the well is turned off. These relations do not necessarily hold during seasonal transitions when ground-water temperature lags about 1 month behind the seasonal trend of surface-water temperature. For example, surface water is warmest during July and August; whereas ground water is warmest during August and September (fig. 6A). Surface water is coldest during January and February; whereas ground water is coldest during February and March (fig. 6C).

Ground-water temperatures in the observation well were found to "spike" according to the season and water temperature of the Tioughnioga River. The spikes occurred during (1) moderately high

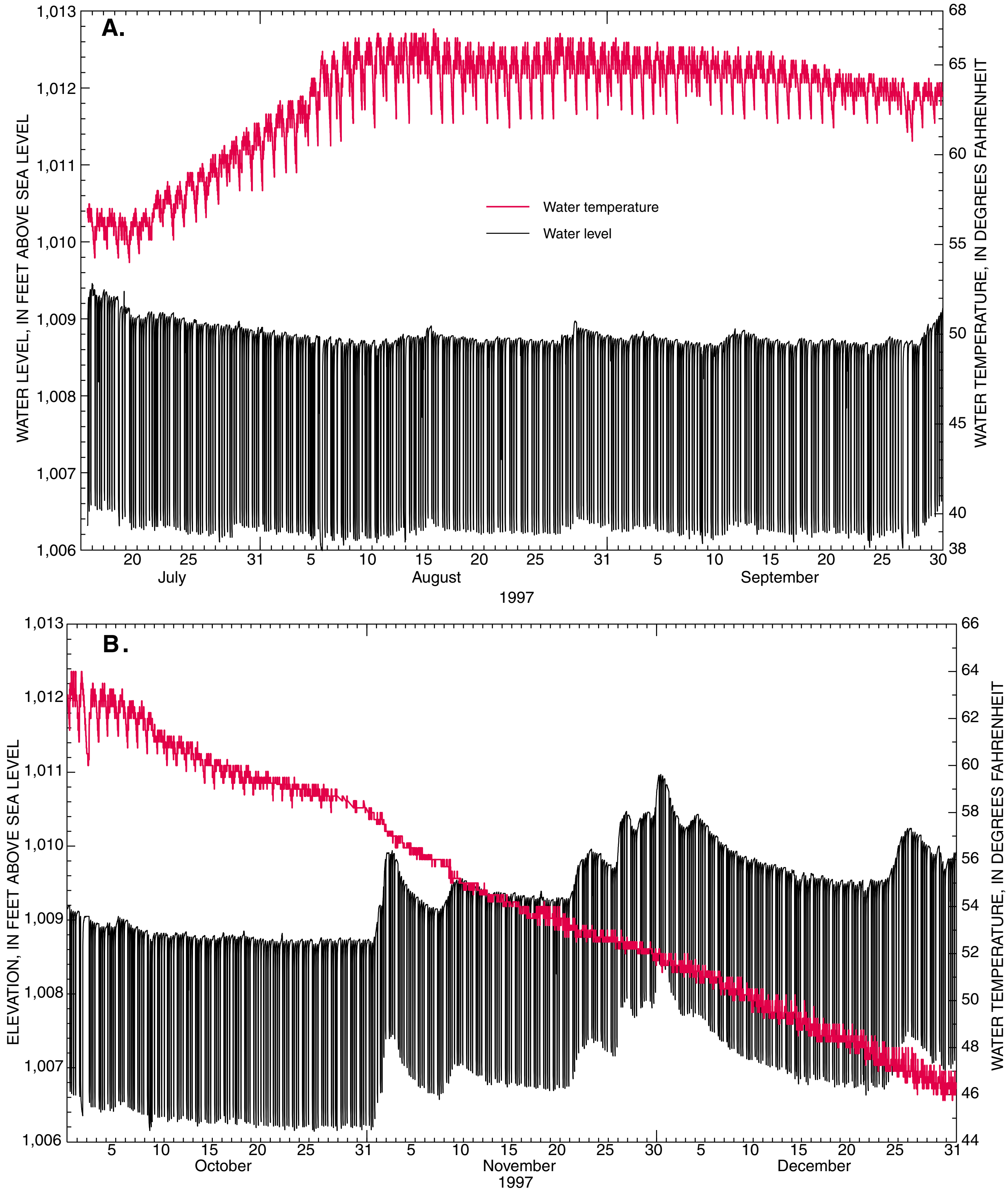


Figure 6. *Ground-water temperature and water level in observation well between municipal pumping well 452 and Tioughnioga River, Marathon, N.Y.: A. July through September 1997. B. October through December 1997.*

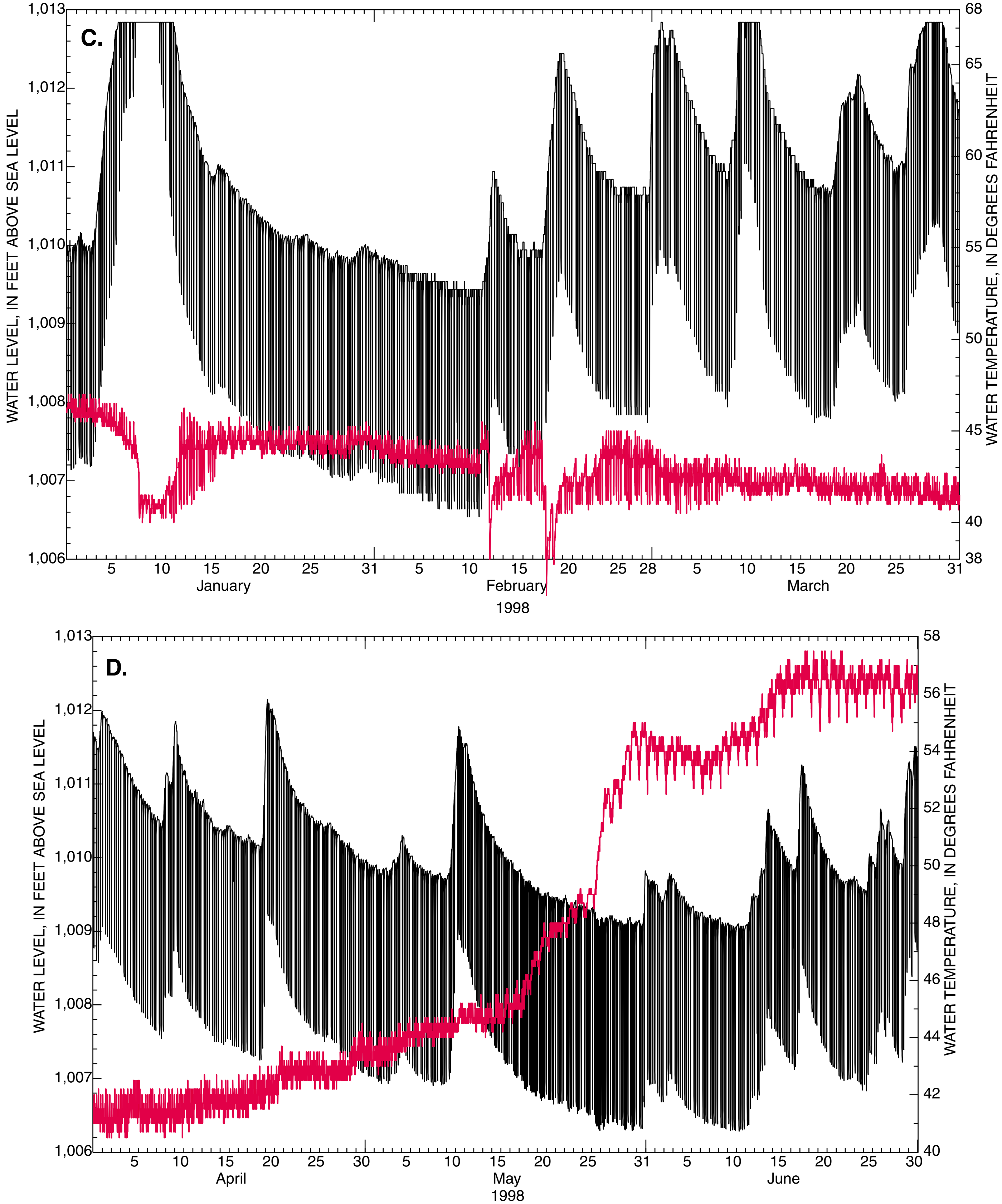


Figure 6. (continued). *Ground-water temperature and water level in observation well between municipal pumping well 452 and Tioughnioga River, Marathon, N.Y. C. January through March 1998. D. April through June 1998.*

stormflows, when the river water rises to near bankfull stage and is higher than the adjacent ground-water levels and seeps into the streambanks; and (2) extremely high stormflows, such as that of early January 1998 (fig. 6C), when a large snowmelt caused the river to overflow its banks, and temporary flooding of the well area resulted in aquifer recharge with relatively cold surface water (fig. 8). This recharge caused rapid lowering of ground-water temperature in the observation well (fig. 6C).

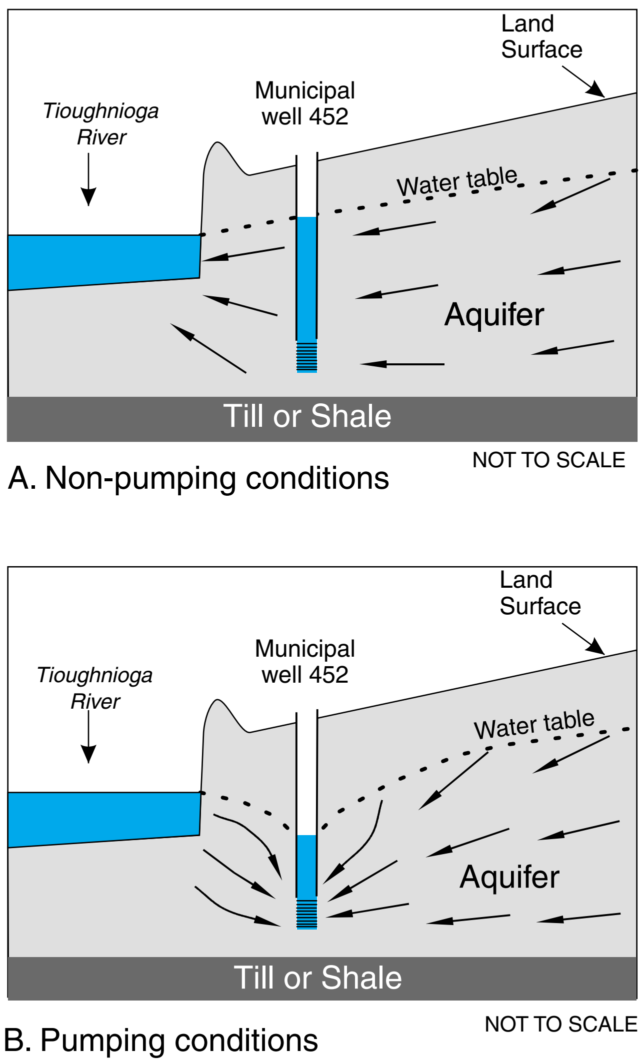


Figure 7. *Hydrologic settings where ground water discharges to the Tioughnioga River: A. Under non-pumping conditions. B. Where pumping by municipal well 452 intercepts ground water that would have flowed to Tioughnioga River and induces water to flow from river to well.*

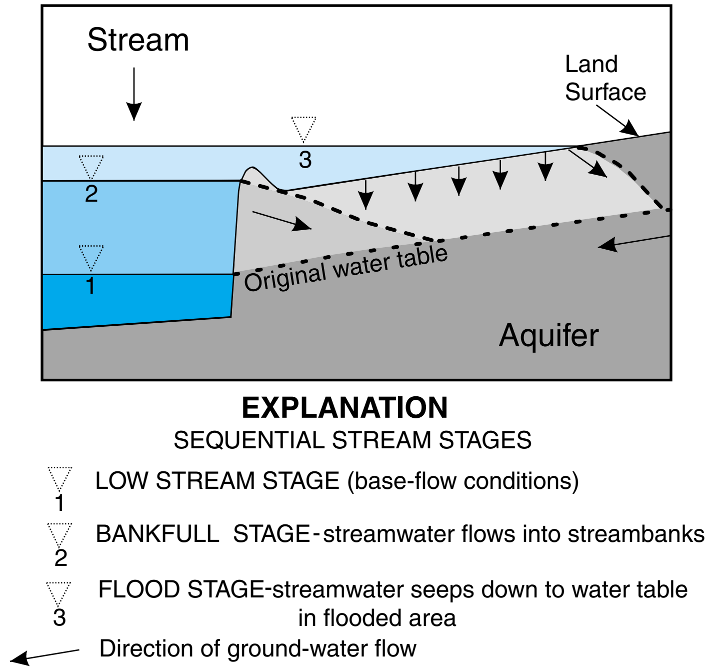


Figure 8. *Processes that affect exchange of water between Tioughnioga River and adjacent aquifer: When river stage rises from low level (1) to bankfull level (2) and is higher than adjacent water table, river water moves laterally into streambank. When the river overflows its banks (3), floodwaters seep downward to the water table.*

SIMULATION OF GROUND-WATER FLOW

A three-dimensional numerical ground-water flow model, MODFLOW (McDonald and Harbaugh, 1988), was used to compute (1) hydraulic heads (hereafter referred to as head) in the aquifer under steady-state conditions, and (2) the water budget for the modeled area. The output from MODFLOW was used in conjunction with a particle-tracking program, MODPATH (Pollock, 1994), to delineate the areas contributing recharge to two simulated wells representing the municipal wells, one near the Tioughnioga River in the northern part of the modeled area and the other near a pond in the southern part.

Model Description and Design

MODFLOW is a widely used code that is based on block-centered, finite-difference equations that simulate ground water flowing through a porous medium. MODFLOW views a three-dimensional system as a sequence of layers that may be designated as always confined, unconfined, or capable of being confined or unconfined.

Simplifying Assumptions

Five simplifying assumptions about the ground-water flow system were used for the simulations:

(1) Ground-water flow is horizontal within the model layers and vertical between layers. (The assumption that ground water moves only horizontally within layers applies reasonably well throughout the modeled area except near pumped wells and directly beneath recharge and discharge areas, where the vertical-flow component within layers may be appreciable.

(2) Recharge to the aquifer is areally uniform.

(3) The modeled aquifer can be divided into a finite number of square blocks or cells, each of which has uniform hydraulic properties. The water level calculated for the center of each block, termed the node, is assumed to be representative of water levels over that entire block.

(4) A simulated pumping well in a model cell is considered to be screened through the full saturated thickness of the cell.

(5) Ground-water levels measured in seven wells on June 12, 1997 represent median annual conditions. The median annual condition is defined as that when the flow-duration is near 50 percent (285 ft^3/s) as measured at the stream gaging station (01509000) on the Tioughnioga River at Cortland (Hornlein and others, 1997), 15 mi north of Marathon.

Model Grid

The model represents the unconfined aquifer as two layers (fig. 9), each 10 to 15 ft thick. The saturated thickness of these layers 1 and 2 was determined from results of test drilling by the Village of Marathon and by several consulting engineering companies for a variety of projects in the village. The two-layer representation provides adequate vertical resolution to simulate ground water that could flow beneath cells that represent streams (such cells are known as “weak sinks” because not all flow into the cell may be captured by the cell). All pumping wells are completed in the lower part of the aquifer and, therefore, are simulated as discharge from model layer 2.

The model grid has 45 rows and 165 columns and was superimposed on a map of the Marathon area (fig. 10). The model contains a total of 14,850 cells (7,425 cells for each layer), 7,606 cells are active (3,803 active cells in each layer). A uniform cell size representing a 50-x 50-ft surface area was used. The model represents a 0.34 mi^2 area.

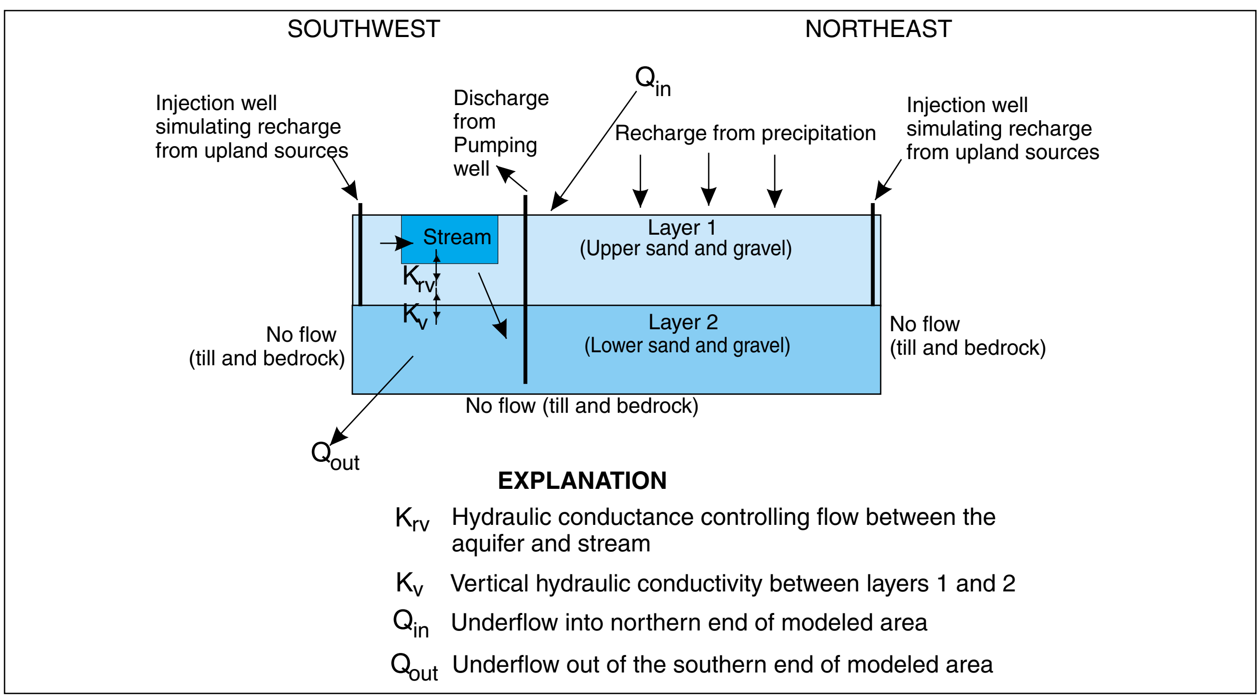
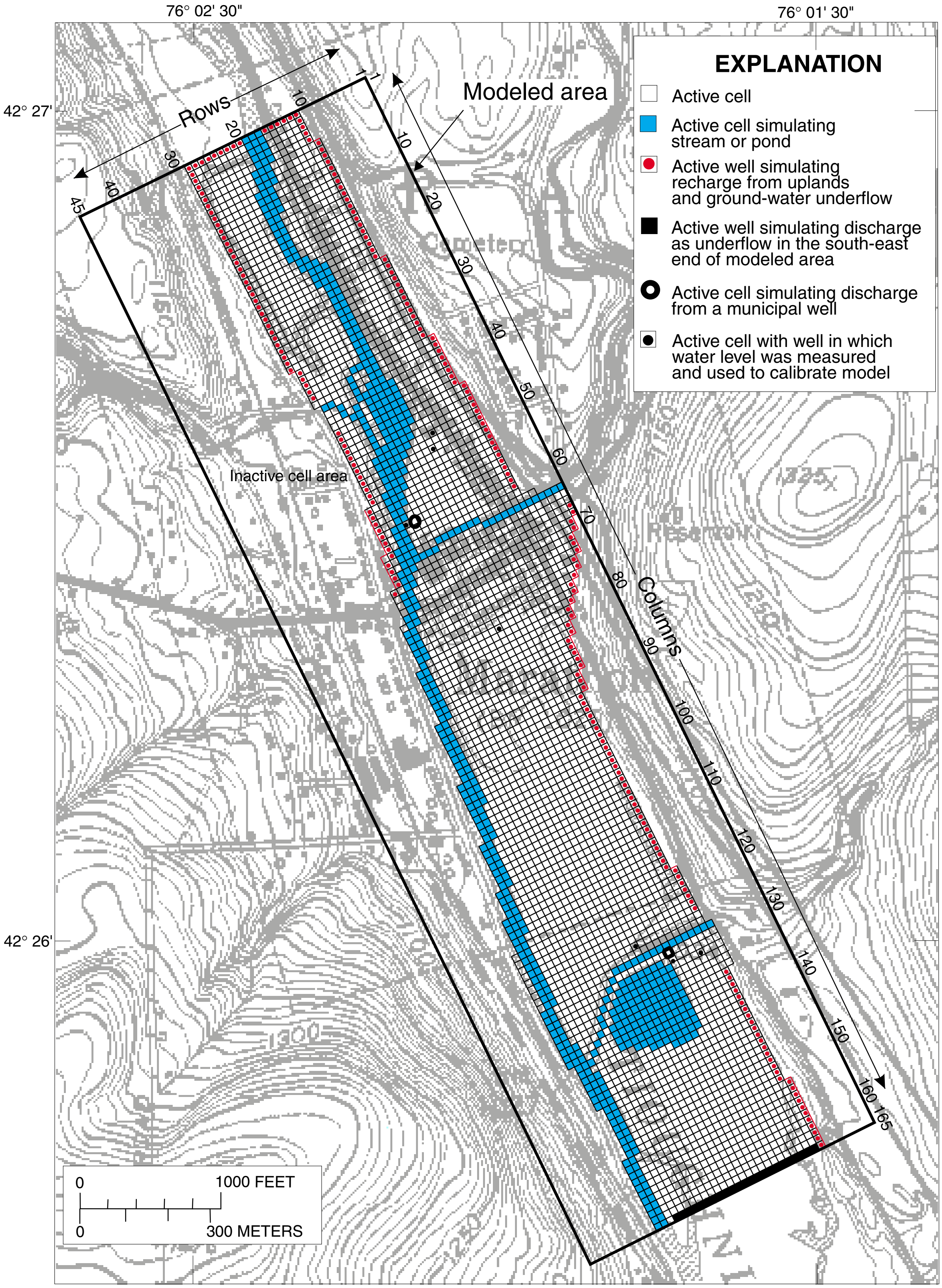


Figure 9. *Model representation of ground-water flow in the sand and gravel aquifer at Marathon, N.Y. (Location is shown in fig. 1.)*



Base from U.S. Geological Survey, Marathon, N.Y., 1994

Figure 10. *Active cells and boundaries in model of aquifer in Tioughnioga River valley at Marathon, N.Y.*

Boundary Conditions

Several types of boundary conditions were specified in the model (figs. 9 and 10). Natural boundaries were used where possible, but arbitrary boundaries were used to limit the northern and southern extent of the model because the aquifer extends many miles beyond the modeled area. A specified-flux boundary was used to represent areal recharge from precipitation that falls directly on the aquifer. A specified-flux boundary, represented by recharge wells (injection wells), also was used along the valley walls to simulate recharge from surface runoff and ground-water seepage from bordering unchanneled uplands into the aquifer (layer 1). The drainage area of the unchanneled hillsides was delineated, and the recharge from these upland area was calculated from equation 1 (p. 8). The calculated recharge from the unchanneled upland area was then divided by the number of bordering active model cells to obtain the recharge rate for each of these cells.

The water withdrawn by the municipal supply well in the northern part of the aquifer was simulated as being pumped from layer 2 of the aquifer. Specified-flux boundaries, represented by wells, also were used to simulate ground-water flow into the northern boundary of the modeled area and out of the southern boundary of the modeled area. The flow into the northern end of the model was represented by recharge wells, whereas the flow out of the southern end was represented by discharge wells. The assigned flux in these boundary cells ranged from 100 to 300 ft^3/d and was calculated from Darcy's equation for one-dimensional flow in a prism of porous material:

$$Q = \frac{KA(h_2 - h_1)}{L} ; \tag{2}$$

where
Q = flow (ft^3/d)
K = hydraulic conductivity of the aquifer (ft/d);
A = cross-sectional area perpendicular to flow (ft^2);
h_2-h_1 = head difference across the prism of flow (ft);
L = length of flow path (ft).

The following hydraulic values were used to calculate underflow (Q) out the Tioughnioga River valley: hydraulic conductivity (K), 100 ft/d, cross-sectional area (A), 4,500 ft^2, head difference (h^2-h^1) of 0.00033 to 0.00067 ft, and length (L) of 1 ft.

The River Package in MODFLOW was used to simulate flow of water between (1) the aquifer and the river and streams, and (2) the aquifer and the pond near the southern municipal well (well 459, fig. 1). The River Package was also used to simulate the recharge from upland tributaries that lose water where they flow over the aquifer. Streambed-conductance values in the River Package were adjusted until simulated losses were close to those losses measured in streams on July 11, 1997 (the period judged to represent median annual conditions and selected for steady-state simulation).

Leakage between streams and aquifer, or stream and pond, depends on the head difference between the surface water and the water table and is adjusted according to a streambed-conductance term. The amount of leakage is computed by Darcy's Law as follows:

$$Q = C_{str}(H_s - H_a), \tag{3}$$

where
Q = leakage to or from the aquifer through the streambed (ft^3/d);
H_s = head in the stream (ft);
H_a = head in the aquifer side of streambed (ft); and
C_{str} = streambed conductance (ft^2/d), defined as the vertical hydraulic conductivity of the streambed, times the width of the stream reach, times its length, divided by the thickness of the streambed.

The pond and streams are in layer 1 of the model. Each stream cell was assigned values for the following: (1) layer, row, and column number, (2) average water-surface elevation, (3) streambed conductance, and (4) elevation of bottom and top of the streambed. A value of 5 ft/d was estimated for the vertical conductivity of the streambed, and a streambed thickness of 1.5 ft was used to compute streambed conductances. A streambed conductance of 7,000 ft^2/d was used for the Tioughnioga River and pond, and values ranging from 800 to 1,900 ft^2/d were used for the small streams.

Hydraulic Values

Transmissivity values were calculated from discharge and drawdown data that were collected by drillers at three aquifer-test sites—one in the northern part of the modeled area, one in the central part, and one in the southern part. Transmissivity was calculated from the following equation, which was derived from (Todd, 1980, eq. 4.70):

$$T = \left[\frac{2.3}{4\pi}\frac{Q}{s}\log 2.25\frac{Tt}{r_w^2 S}\right]^{-1} ; \qquad (4)$$

where Q = well discharge (ft^3/d),
s = drawdown (ft),
T = transmissivity (ft^2/d),
t = time of pumping (d),
r_w = well radius (ft), and
S = storage coefficient (dimensionless).

Transmissivity values calculated for the three aquifer-test sites in the north, central, and south parts of the modeled area were 1,900, 1,640, and 2,400 ft^2/d, respectively. The storage coefficient was assumed to be 0.1. The average horizontal hydraulic conductivity of layers 1 and 2 was estimated to be 88 ft/d, which was calculated by dividing the transmissivity values by the saturated thickness at each pumping test site, then averaging the three values. A value of 100 ft/d was finally used; this was based-on-trial and error calibration of the model.

The vertical hydraulic conductivity for layers 1 and 2 was estimated to be 20 ft/d; this value was calculated by multiplying the horizontal hydraulic conductivity value of 100 ft/d by 0.2 to represent an assumed anisotropy of 1:5. Vertical hydraulic conductivity in stratified drift tends to be less than horizontal hydraulic conductivity because, at a small scale, the stratified drift contains many layers of sediment, some of which consist of plate-shaped particles that tend to settle horizontally and, thus, impede the vertical flow of ground water.

Vertical leakance (Vcont) between most cells representing layers 1 and 2 was 1.1 ft/d. The vertical leakance was computed from the following equation (equation 51 of McDonald and Harbaugh, 1988, p. 5-13) for two vertically adjacent geohydrologic units, each unit with its own value of hydraulic conductivity, and an average thicknesses of 20 ft for layer 1 and 15 ft for layer 2.

$$Vcont_{i,j,k+1/2} = \frac{1}{\dfrac{(v_k)/2}{k_{z\ i,j,k}} + \dfrac{(v_{k+1})/2}{k_{z\ i,j,k+1}}}; \qquad (5)$$

Where v_k = thickness of the model layer k,
v_{k+1} = thickness of model layer k+1,
$k_{z\ i,j,k}$ = vertical hydraulic conductivity of the upper layer in cell i,j,k,
$k_{z\ i,j,k+1}$ = vertical hydraulic conductivity of the lower layer in cell i,j,k+1,
i = row number in model grid,
j = column number in model grid, and
k = layer number in model grid.

Model Calibration

The model was calibrated to flow measured in three small streams on June 11, 1997 and water levels measured in seven wells on June 12, 1997, a period judged to represent median annual conditions. The model was not calibrated to flows in the Tioughnioga River because the measurement error (typically 5 percent) in large streams would be much greater than the gains or losses that would be expected to occur in each measured reach. For example, a measurement error of ± 13.2ft^3/s would be expected for June 11, 1997, when streamflow in the Tioughnioga River was 265 ft^3/s, but the model-generated gain of water from ground-water discharge into the Tioughnioga River in the modeled area was only 1.3 ft^3/d.

The model was calibrated through a trial-and-error process. Calibration entailed running the model with initial estimates of input values and noting the differences between measured and simulated water levels and streamflows, and then adjusting the input values (within reasonably expected ranges) until the simulated water levels and streamflows were acceptably close to the measured values. Changes were made one at a time to the horizontal hydraulic conductivity, recharge rate, or the hydraulic conductance between aquifer and stream. Final horizontal hydraulic conductivity value for layers 1 and 2 was 100 ft/d (the average hydraulic conductivity calculated from drawdown analyses during aquifer tests, as mentioned earlier, was 88 ft/d). Calibration was considered complete when the simulated streamflow losses were close to those measured (table 2) and the root mean squared error of differences between simulated and measured water levels in the seven wells was 0.5 ft or less (table 3).

The water budget for simulated long-term average, steady-state conditions with one simulated discharging well representing the municipal well pumping is given in table 4. The principal sources of recharge to the unconfined aquifer is from unchanneled runoff and ground-water inflow from the uplands (43 percent of total recharge); precipitation that falls directly on the aquifer (36 percent); stream leakage (19 percent), and ground-water underflow into the northern end of the modeled area (2 percent). Most ground-water discharge in the simulation was into the Tioughnioga River (87 percent of the total); the rest discharged to the simulated pumping well (10 percent) and out the southern end of the modeled area as underflow (3 percent).

Table 2. *Measured and simulated streamflow gains and losses in selected stream reaches near Marathon, N.Y.*

[Values are in cubic feet per second. Negative numbers indicate losses. Locations are shown in fig. 1.]

Tributary stream	Measured June 11, 1997	Simulated
Hunts Creek	-0.15	-0.15
Tributary A (near pond)	-0.07	-0.07
Tributary B (west side of valley	-0.03	-0.03
Sum	-0.25	-0.25

Model Sensitivity

A model-sensitivity analysis, wherein a single aquifer property or flux is varied while all other properties and fluxes are held constant, was conducted to identify the relative effect of adjustments of aquifer properties on simulated heads and stream-aquifer leakage. Future data-collection efforts can be directed to those aquifer properties to which the simulated heads were most sensitive.

Recharge, horizontal hydraulic conductivity, vertical conductance between layers 1 and 2, and the vertical conductances between the aquifer and the streams were increased or decreased, one at a time, by multiplication factors within probable ranges; the effects of these adjustments on calculated heads at seven model cells representing the water-level measurement sites are plotted in figure 11. The smallest root mean square of the difference between

Table 3. *Difference and root mean square of the difference between measured and simulated heads at seven selected wells in the unconfined aquifer at Marathon, N.Y.*

[Values are observed head minus simulated head in layer 1, in feet. Well locations shown in fig. 1].

	Location					
Well no.	Model row	Model column	Measured head June 12, 1997	Head calculated by model	Difference between heads (X)	X^2
457	13	134	1008.2	1008.2	0.0	0.00
451	14	50	1007.2	1007.5	.3	.09
463	16	53	1007.3	1007.5	.2	.04
459	17	132	1002.7	1003.0	.3	.09
458	17	133	1003.1	1002.9	-.2	.04
455	18	79	1007.3	1007.8	.5	.25
461	21	129	1003.2	1003.7	.5	.25
					Total	.76
					Root Mean Square	0.33

Table 4. *Steady-state water budget for the calibrated model with one simulated discharging well*

[Values are in cubic feet per day]

Budget component	Amount (cubic feet per day)	Percent of total
A. Recharge to the aquifer		
Precipitation on aquifer	47,580	36
Unchanneled runoff and ground-water inflow from uplands	57,490	43
Stream leakage	25,240	19
Ground-water inflow into northern end of modeled area	2,900	2
TOTAL	133,210	100
B. Discharge from aquifer		
Municipal wells	13,100	10
Discharge from aquifer to surface water	115,610	87
Ground-water underflow out of southern end of modeled area	4,500	3
TOTAL	133,210	100

calculated and measured heads in the final, calibrated model (multiplication factor equal to 1 in fig. 11) was 0.33 ft (table 3). All other multiplication factors resulted in root mean squared errors equal to or greater than 0.33 feet.

Results of the sensitivity analyses indicate that varying the recharge rate and horizontal hydraulic conductivity of layer 1 had relatively large effects on simulated heads (as much as 5 ft from the observed value), whereas varying the conductance between the aquifer and the streams, and the vertical conductance between layers 1 and 2, had little or no effect on simulated heads (fig. 11). Recharge was varied from 0.5 to 1.5 times the average annual rate of 22 in/yr, horizontal hydraulic conductivity was varied from 0.2 to 5 times the final value that was used in the model (100 ft/d), conductance between the aquifer and the streams was varied from 0.2 to 5 times the value used for the Tioughnioga River (7,000 ft/d) and the small streams (800 to 1,900 ft/d), and the vertical conductance between layers 1 and 2 was varied from 0.1 to 10 times the values used for the final model.

Model Application

The Village of Marathon would like to have available two wells that are capable of meeting the water needs of the village—one main supply well and one backup well, and needs to know the recharge areas that supply water to the pumping wells. As mentioned earlier, only the well in the northern part of the study area is being used at present because water from the well near the pond in the southern part contains high concentrations of manganese. At the time of this study (1997-98), the village was considering either treatment of the water from the southern well to remove the manganese, or installing another well elsewhere. Delineations of the areas that contribute recharge to these wells would indicate where land-use restrictions are needed to protect the ground water from contamination. MODPATH was used to

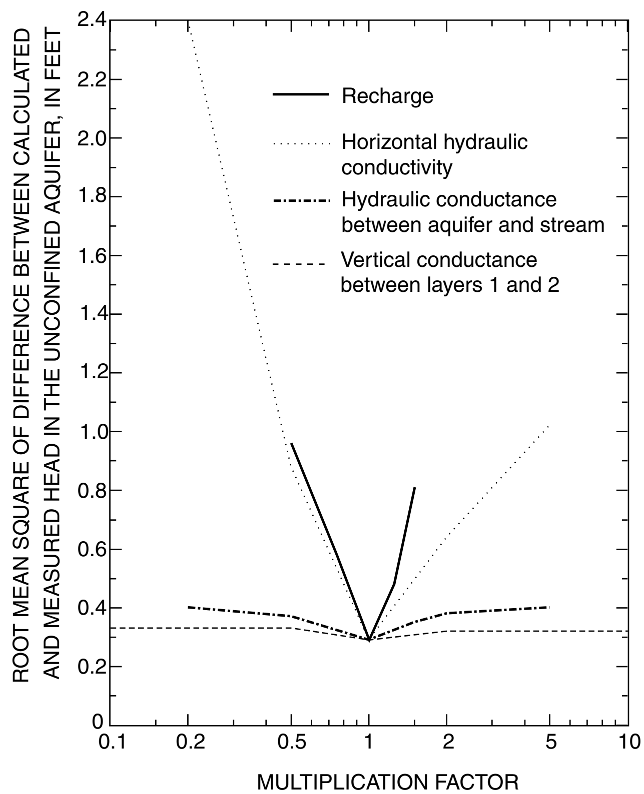


Figure 11. *Results of sensitivity analyses for selected hydraulic properties in layer 1 of the model of the unconfined sand and gravel aquifer at Marathon, N.Y.*

determine the aquifer contributing area to a pumped well (defined as the land-surface area within the valley in which water that infiltrates to the water table eventually flows to the pumping well) but the uplands that contribute surface runoff and ground-water flow to the adjacent aquifer contributing area are also part of the contributing area, however.

In the calibrated model, two simulated wells were used to represent the present municipal wells. Each well was simulated as discharging at a average daily rate of 13,100 ft^3/d (68 gal/min) which, individually, is the average amount of water now pumped daily (98,000 gal/d) by the village. The simulated wells are far enough apart (about 4,000 ft) that their drawdowns do not interfere with each other. The simulated water budget for long-term average, steady-state conditions, and with both municipal wells pumping at an average daily rate of 98,000 gal/d is given in table 5. The principal sources of aquifer recharge is from unchanneled runoff and ground-water inflow from the uplands (41 percent of total recharge); precipitation that falls directly on the aquifer (34 percent); stream leakage (23 percent), and ground-water underflow into the northern end of the modeled area (2 percent). Most ground-water discharge in the simulation was into the Tioughnioga River (78 percent of the total); the rest discharged to the two simulated pumping wells (19 percent) and out the southern end of the modeled area as underflow (3 percent).

Table 5. *Steady-state water budget with two simulated discharging wells*

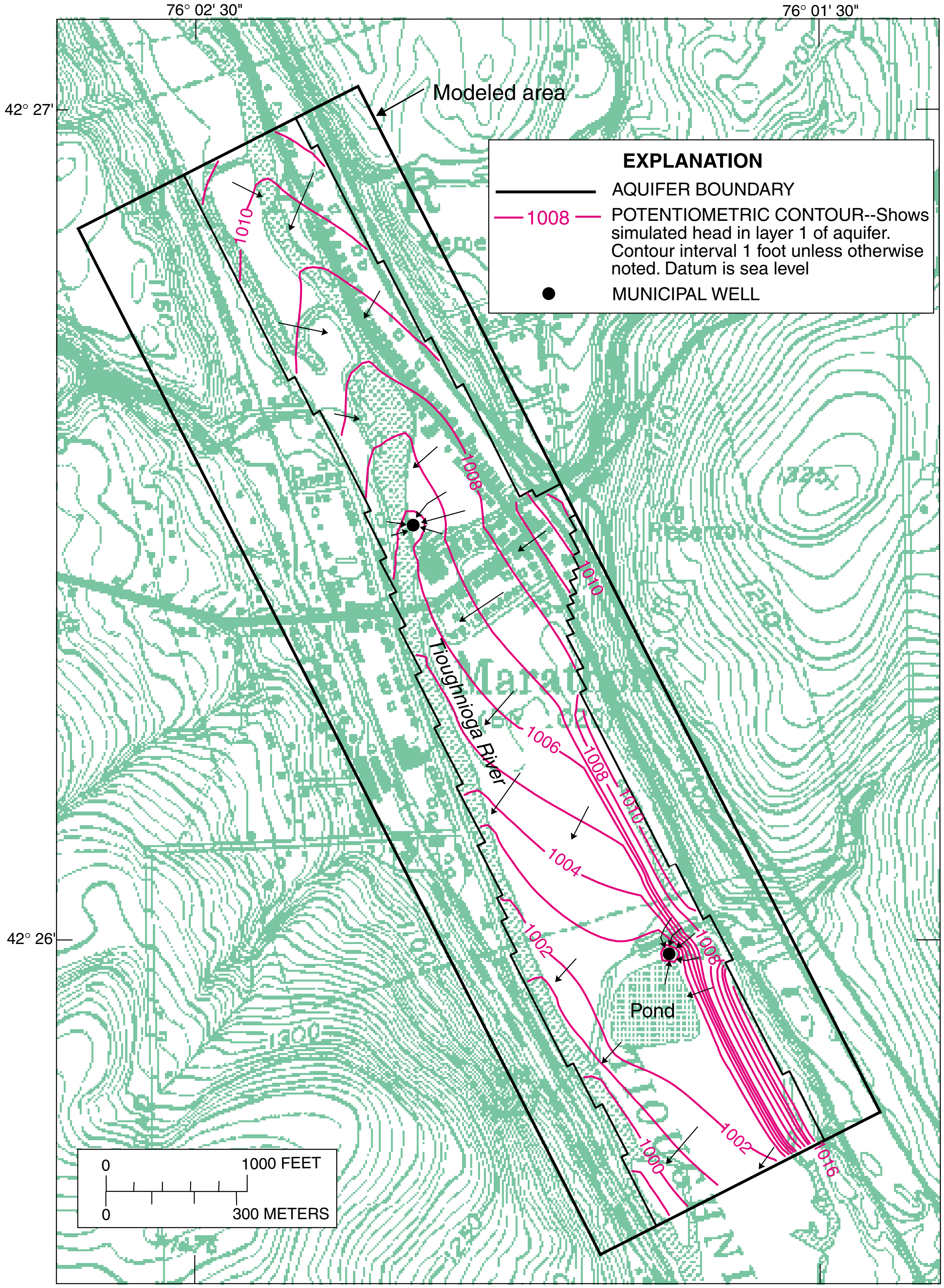
[Values are in cubic feet per day]

Budget component	*Amount (cubic feet per day)*	*Percent of total*
A. Recharge to the aquifer		
Precipitation on aquifer	47,600	34
Unchanneled runoff and ground-water inflow from uplands	57,500	41
Stream leakage	32,800	23
Ground-water inflow into northern end of modeled area	2,900	2
TOTAL	140,800	100
B. Discharge from aquifer		
Municipal wells	26,200	19
Discharge from aquifer to surface water	110,100	78
Ground-water underflow out of southern end of modeled area	4,500	3
TOTAL	140,800	100

The calibrated model was used to compute heads in the aquifer (fig. 12) that would result during average-annual steady-state conditions with two simulated municipal wells pumping at their expected daily average rate. The lateral direction of ground-water flow is perpendicular to the potentiometric contours and is shown by several arrows in figure 12. Most ground water flows from the edges of the valley toward the Tioughnioga River, and some flows to the two wells; the rest leaves the southern end of the modeled area as underflow (fig. 12).

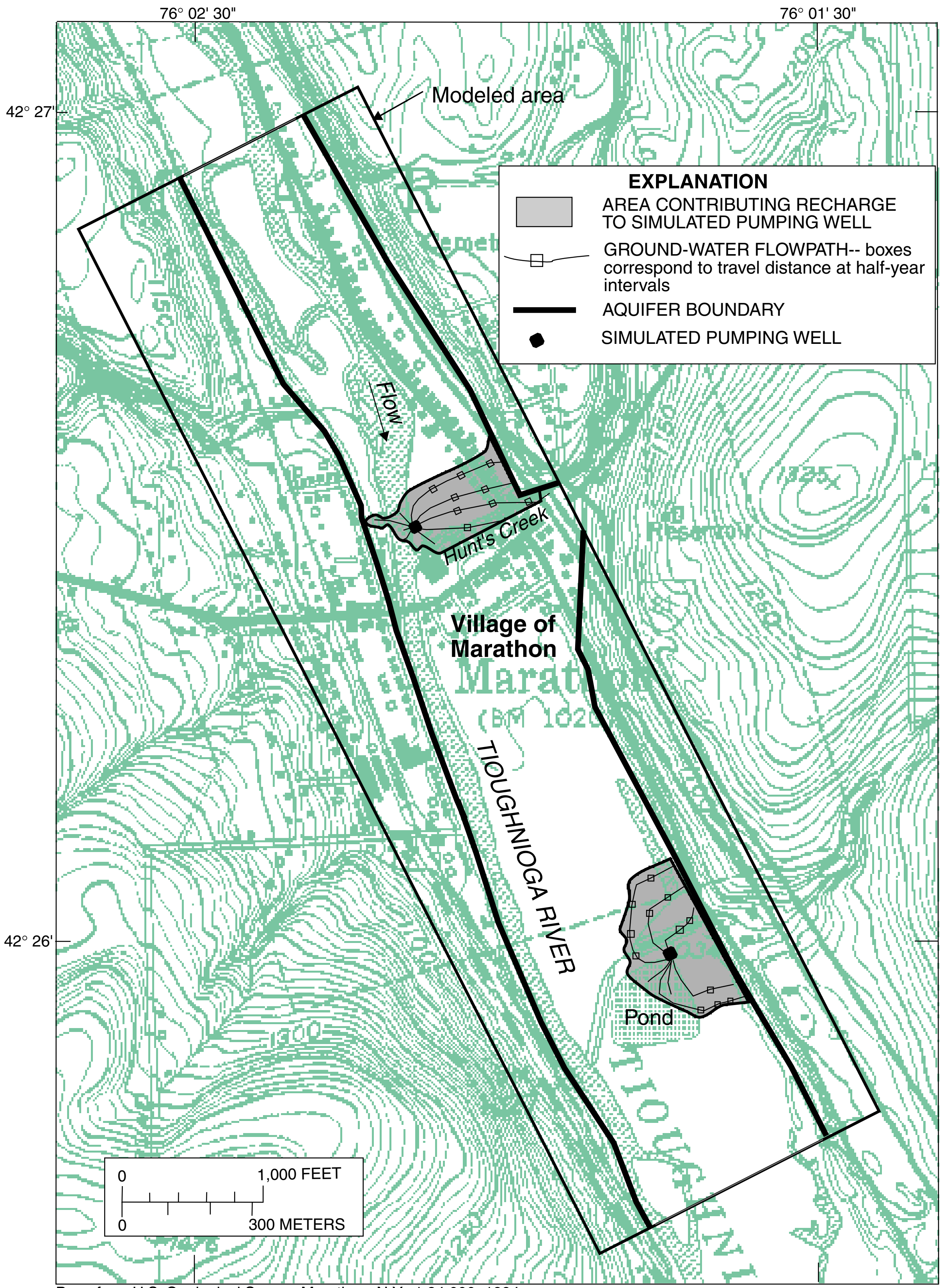
The particle-tracking program, MODPATH, was applied to the calibrated model to delineate the aquifer areas that contribute recharge to the two simulated pumping wells (fig. 13). The flowpath analysis indicates that, (1) the northern well's contributing area is 0.10 mi wide and 0.15 mi long and covers an area of 0.015 mi^2 within the valley, and (2) the southern well's contributing area is 0.20 mi wide and 0.11mi long and covers an area of 0.022 mi^2 within the valley (fig. 13). The average traveltime for ground water to flow from the valley wall to either simulated pumping well is about 1.5 years, as estimated from an assumed aquifer porosity of 0.3. The flowpath analysis also indicates that the contributing areas of both wells contain surface-water bodies—reaches of the Tioughnioga River and Hunts Creek are within the contributing area of the northern well, and a reach of a small unnamed tributary and the northern part of the pond are within the contributing area of the southern well (fig. 13). Both wells are relatively close to these surface-water bodies—the northern well is 57 ft east of the Tioughnioga River, and the southern well is 75 ft northeast of the pond and 90 ft south of the tributary stream; therefore, induced infiltration is likely.

The uplands that contribute surface runoff and ground-water flow to the adjacent valley are also part of the contributing area to the pumped wells. The contributing areas beyond the valleys were not delineated by the model, but the upland contributing



Base from U.S. Geological Survey, Marathon, N.Y., 1:24,000, 1994

Figure 12. *Simulated heads in the unconfined aquifer at Marathon, N.Y., for average annual steady-state conditions with two simulated discharging wells. (Location is shown in fig. 1.)*



Base from U.S. Geological Survey, Marathon, N.Y., 1:24,000, 1994

Figure 13. *Areas that contribute recharge to two simulated municipal pumping wells tapping the unconfined aquifer at Marathon, N.Y. (Location is shown in fig. 1.)*

areas warrant consideration when the potential for aquifer contamination is evaluated. The upland contributing areas (not shown) would include: Hunts Creek basin and the unchanneled hillside adjacent to the valley-contributing area for the northern well, and the unnamed tributary's upland basin and the unchanneled hillside adjacent to the valley-contributing area for the southern well.

About 31 percent of the water pumped by the simulated northern well (13,100 ft^3/d) was derived from the Tioughnioga River, 41 percent was from Hunts Creek, 16 percent from unchanneled runoff from adjacent hillsides, and 12 percent from precipitation that fell on the aquifer in the contributing area to the well. Increasing the simulated pumping rate resulted in an increase the amount of induced infiltration and in the percentage of pumped water that is derived from the Tioughnioga River; for example, at pumping rates of 38,400 ft^3/d (200 gal/min) and 48,000 ft^3/d (250 gal/min), the percentage of pumped water derived from the river was 51 percent and 60 percent, respectively.

SUMMARY

The Village of Marathon has three municipal wells that tap a relatively thin (typically 25-to 40-ft thick), narrow (less than 0.25 mi wide) unconfined sand and gravel aquifer in the Tioughnioga River valley. Only one of the wells was used by the village at the time of the investigation discussed in this report. The second well was abandoned as a result of contamination from a petroleum spill, and a third well, which is near a pond in the southern part of the study area, is not used because the water contains high concentrations of manganese. The village pumps about 0.1 Mgal/d and supplies about 1,000 people.

The ground-water-temperature fluctuations in an observation well between the Tioughnioga River and a municipal pumping well correspond to the daily water-level drawdowns and recoveries caused by the pumping well, which is turned on and off two or three times a day. The temperature fluctuation in response to pumping indicates that the pumping well is inducing water from the Tioughnioga River, in addition to capturing water from the aquifer's shallow zone, which undergoes warming and cooling in response to air-temperature changes. Pumping of the municipal well typically causes a drawdown of about 3 ft below river stage in the observation well; this indicates that the hydraulic gradient slopes from the river to the well during pumping and water is probably induced from the river.

A three-dimensional, finite-difference-ground-water-flow model was used to (1) compute hydraulic heads in the aquifer under steady-state conditions, (2) develop a water budget, and (3) delineate the areas within the valley that contribute recharge to two simulated wells that represent two of the municipal wells. The contributing area of the contaminated well was not delineated because the well was abandoned. The simulated water budget for long-term average, steady-state conditions, and with both municipal wells pumping at an average daily rate of 68 gal/min (98,000 gal/d), indicated that the principal sources of aquifer recharge is from unchanneled runoff and ground-water inflow from the uplands (41 percent of total recharge); precipitation that falls directly on the aquifer (34 percent); stream leakage (23 percent), and ground-water underflow into the northern end of the modeled area (2 percent). Most ground-water discharge in the simulation was into the Tioughnioga River (78 percent of the total); the rest discharged to the two simulated pumping wells (19 percent) and out the southern end of the modeled area as underflow (3 percent).

The flowpath analysis indicates that the northern well's contributing area is 0.10 mi wide and 0.15 mi long and covers an area of 0.015 mi^2 within the valley, and the southern well's contributing area is 0.20 mi wide and 0.11mi long and covers an area of 0.022 mi^2 within the valley. The average traveltime of ground water from the valley wall to the pumping wells is about 1.5 years for both wells, as estimated from an assumed aquifer porosity of 0.3. The flowpath analysis indicates that surface water enters the contributing areas of both wells; the contributing area for the northern well contains a reach of the Tioughnioga River, and contributing area for the southernmost well contains part of a pond. Both wells are relatively close to these surface-water bodies; the northern well is 57 ft east of the river and the southern well is 75 ft northeast of the pond; therefore induced infiltration can be expected.

Surface-water runoff and ground-water flow from uplands that drain to the adjacent contributing areas within the valley also enter the contributing area to the discharging wells, but the contributing areas beyond the valley were not delineated.

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