GEOHYDROLOGY OF, AND SIMULATION OF GROUND-WATER FLOW IN, THE MILFORD-SOUHEGAN GLACIAL-DRIFT AQUIFER, MILFORD, NEW HAMPSHIRE

By Philip T. Harte and Thomas J. Mack

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 91-4177

Prepared in cooperation with the

U.S. ENVIRONMENTAL PROTECTION AGENCY
REGION I

Bow, New Hampshire
1992

U.S. DEPARTMENT OF THE INTERIOR

MANUEL LUJAN, Jr., *Secretary*

U.S. GEOLOGICAL SURVEY

Dallas L. Peck, *Director*

For additional information, write to:

Chief, NH-VT Office
U.S. Geological Survey
Water Resources Division
525 Clinton Street
Bow, NH 03304

Copies of this report can be purchased from:

U.S. Geological Survey
Books and Open-File Reports Section
Federal Center, Bldg. 810
Box 25425
Denver, CO 80225

CONTENTS

	Page
Abstract	1
Introduction	2
Purpose and scope	2
Approach and methods	2
Description of the study area	4
Ground-water withdrawals	4
Previous investigations	6
Acknowledgments	8
Geohydrology	8
Milford-Souhegan aquifer	8
Areal extent and saturated thickness	8
Lithologic characteristics and depositional history	10
Hydraulic characteristics	10
Ground-water levels and flow directions	13
Stream-aquifer interaction	17
Recharge	24
Simulation of ground-water flow	26
Model construction and initial data input	29
Grid design	30
Horizontal discretization	30
Vertical discretization	30
Hydraulic conductivity	30
Boundary conditions	38
Recharge	40
Stream-aquifer interaction	40
Model calibration	40
Sensitivity analysis	52
Evaluation of model results	58
Hydraulic characteristics	58
Ground-water flow	59
Estimates of contributing areas to supply wells	60
Summary and conclusions	73
Selected references	74
Appendix A: Data for selected wells and boreholes in the Milford-Souhegan aquifer	A-1
Appendix B: Water levels in selected wells during the October 17, 1988, Keyes aquifer test	B-1
Appendix C: Ground-water-level data for wells in the Milford-Souhegan aquifer	C-1

ILLUSTRATIONS

Page

Figures 1-4. Maps showing:
1. Location of study area ... 3
2. Areal extent of Milford-Souhegan aquifer and locations of streams, Keyes and Savage wells, and major cultural features ... 5
3. Locations of wells in the Milford-Souhegan aquifer ... 7
4. Areal extent and saturated thickness of the Milford-Souhegan aquifer ... 9

5. Graph showing drawdowns at nested observation wells during the Keyes well aquifer test of October 17, 1988 ... 14
6. Hydrograph showing long-term water levels at an observation well in the Milford-Souhegan aquifer ... 15
7. Map showing altitude of the water table in the Milford-Souhegan aquifer, October 1988 ... 18
8. Block diagram showing hydrologic cycle and generalized ground-water-flow patterns in a typical river-valley aquifer ... 19
9. Diagram showing water-level differences in nested wells before and during the Keyes aquifer test ... 20
10. Map showing altitude of potentiometric surface and horizontal ground water-flow directions before and during the Keyes aquifer test ... 21
11. Map showing locations of streamflow-measurement sites and gaining and losing reaches of streams, June and October 1988 ... 23
12. Diagram of a hypothetical pumped well showing a cross-sectional view of the zone of diversion and a plan view of the contributing area ... 27
13. Map showing model grid, inactive and active cells, and simulated streams ... 31
14. Diagram showing vertical discretization along model columns 41 and 100 ... 32

15-19. Maps showing zones of horizontal hydraulic conductivity of the Milford-Souhegan aquifer as used in the model:
15. Layer 1 ... 33
16. Layer 2 ... 34
17. Layer 3 ... 35
18. Layer 4 ... 36
19. Layer 5 ... 37

20-24. Maps showing simulated steady-state heads in the Milford-Souhegan aquifer, October 1988:
20. Layer 1 ... 42
21. Layer 2 ... 43
22. Layer 3 ... 44
23. Layer 4 ... 45
24. Layer 5 ... 46

25. Map showing drainage basins in the Milford-Souhegan aquifer ... 53
26. Graph showing mean differences between measured heads and heads computed by the calibrated model and during sensitivity analysis by hydraulic-conductivity zone for layers 1 and 3 ... 56

27-28. Maps showing simulated steady-state heads in the Milford-Souhegan aquifer with simulated pumpage at the Savage and Keyes wells:
27. Layer 1 ... 61
28. Layer 3 ... 62

29. Generalized hydrogeologic section showing simulated ground-water flow paths through model column 100 at the Keyes well ... 63
30. Generalized hydrogeologic section showing simulated ground-water flow paths through model column 41 at the Savage well ... 64

ILLUSTRATIONS (continued)

Page

31-35. Maps showing possible ranges in contributing areas of the Savage and Keyes wells in relation to variation in:
- 31. Pumping rate 66
- 32. Recharge 67
- 33. Horizontal hydraulic conductivity 68
- 34. Streambed conductance 69
- 35. Vertical hydraulic conductivity 70

TABLES

Page

Table 1. Principal ground-water withdrawals from the Milford-Souhegan aquifer 6
2. Values of transmissivities, hydraulic conductivities, and storage coefficients from tests of wells in the Milford-Souhegan aquifer 11
3. Summary of ground-water levels measured at selected wells in the Milford-Souhegan aquifer 16
4. Base-flow data from streamflow-measurement sites 22
5. Stream inflows and outflows in the study area, October 1988 25
6. Distribution of recharge to the Milford-Souhegan aquifer 26
7. Average hydraulic conductivity estimated from predominant grain size of stratified-drift sediment 38
8. Comparison of horizontal hydraulic-conductivity estimates from well-log descriptions with values from aquifer tests for selected wells in the Milford-Souhegan aquifer 39
9. Differences between simulated and measured heads for selected wells in the Milford-Souhegan aquifer 47
10. Differences between simulated and measured heads in the Milford-Souhegan aquifer, by hydraulic-conductivity zone 51
11. Differences between simulated and measured heads in the Milford-Souhegan aquifer, by model layer 52
12. Comparison of simulated and measured heads in select shallow and deep well nests in the Milford-Souhegan aquifer 52
13. Differences between simulated and measured streamflow gains and losses for each drainage basin in the Milford-Souhegan aquifer 54
14. Mean differences between heads from the calibrated model and heads computed during sensitivity analysis, by hydraulic-conductivity zone 55
15. Mean differences between net seepage from the calibrated model and net seepage computed during sensitivity analysis, by drainage basin 57
16. Range of contributing areas of the Savage and Keyes wells produced by varying pumping rate and other model parameters 71

CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square foot (ft^2)	0.09294	square meter
square mile (mi^2)	2.590	square kilometer
	Volume	
gallon (gal)	3.785	liter
	Flow	
inch per year (in/yr)	25.4	millimeter per year
foot per second (ft/s)	0.3048	meter per second
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per second per square mile [$(ft^3/s)/mi^2$]	0.01093	cubic meter per second per square kilometer
gallon per minute (gal/min)	0.06309	liter per second
million gallons per day (Mgal/d)	0.04381	cubic meter per second
	Hydraulic Conductivity	
foot per day (ft/d)	0.3048	meter per day
	Transmissivity	
cubic foot per day per square foot times foot of aquifer thickness [$(ft^3/d)/ft^2$]ft	0.09290	cubic meter per day per square meter times meter of aquifer thickness

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

Geohydrology of, and Simulation of Ground-Water Flow in, the Milford-Souhegan Glacial-Drift Aquifer, Milford, New Hampshire

By Philip T. Harte and Thomas J. Mack

ABSTRACT

A study was done by the U.S. Geological Survey, in cooperation with the U.S. Environmental Protection Agency, to describe the geohydrology of the Milford-Souhegan glacial-drift aquifer in southern New Hampshire, to understand regional flow in the aquifer, and to estimate the areas in the aquifer contributing water to two discontinued public-supply wells. Water in several wells completed in the aquifer, which underlies an area of 3.3 mi² (square miles), has been affected by contaminants that may have originated at one or more potential sources. Because of the potential for losing the use of a valuable resource, such as the Milford-Souhegan aquifer, the U.S. Environmental Protection Agency concluded that a comprehensive analysis of regional flow in the aquifer was necessary to help in planning possible remediation.

The Milford-Souhegan aquifer consists of as much as 114 feet thick of unconsolidated glacial sediments in a buried pre-Pleistocene valley, and has a maximum saturated thickness of approximately 100 feet. The aquifer is composed predominantly of sand and gravel interbedded with silt; deposits generally are finer in the eastern part than in the western part. Horizontal hydraulic conductivity of stratified-drift deposits ranges from approximately 1 to 1,000 feet per day.

Ground-water flow is governed by the good hydraulic connection between the Souhegan River and its tributaries and the aquifer. In the western reaches of the Souhegan River, the river recharges the aquifer and ground-water flow is away from the river. In the eastern reaches of the Souhegan River, ground water discharges to the river and ground-water flow is towards the river.

Total recharge to the aquifer, based on October 1988 streamflow data, was estimated to be 5.31 ft³/s (cubic feet per second). Estimates of the major rates of recharge to the aquifer from various sources are as follows: 3.19 ft³/s from infiltration of precipitation, 1.44 ft³/s from surface-water infiltration, and 0.64 ft³/s from lateral inflow from till-bedrock upland areas. Ground-water withdrawals were approximately 5 ft³/s in 1988.

A three-dimensional ground-water-flow model was used to simulate ground-water heads, stream-aquifer fluxes, and ground-water-flow directions and rates in the Milford-Souhegan aquifer. A numerical, semianalytical particle-tracking procedure was used to delineate contributing areas to two discontinued public-supply wells.

Ground-water-flow simulations of hydrologic conditions for October 1988 indicate that ground-water flow is largely controlled by stream-aquifer interactions, ground-water recharge, and ground-water withdrawals. Ground-water flow is primarily horizontal except near major production wells. Surface-water infiltration is a major source of water to production wells in the western part of the aquifer.

Simulated pumping of the discontinued public-supply wells, referred to as the Savage and Keyes wells, indicate that the effects of pumping on the ground-water-flow system are highly dependent on aquifer geometry, proximity to hydrologic boundaries, and nature of flow systems--whether in a ground-water discharge or recharge zone. Simulated pumping of the public-supply wells indicates that (1) for average daily withdrawals (0.323 ft³/s (145.0 gal/min (gallons per minute)) at Savage and 0.223 ft³/s (100.1 gal/min) at Keyes), the contributing area of the Savage well was larger (0.148 mi²) than the contributing area of the Keyes well (0.103 mi²); (2) the Savage well induces flow from an industrial discharge ditch; and (3) approximately 53 and 70 percent of the ground water pumped from the Savage well and Keyes well, comes from infiltration of precipitation and lateral flow from aquifer boundaries, the remainder comes from infiltration of surface water.

INTRODUCTION

The Milford-Souhegan aquifer underlies 3.3 mi^2 of the town of Milford, Hillsborough County, southwestern New Hampshire (fig. 1). In this report, the Milford-Souhegan aquifer is defined as the entire sequence of saturated glacial drift and other unconsolidated deposits above the bedrock surface in the Souhegan River valley in Milford. The aquifer consists primarily of stratified sand and gravel with some till and is overlain in places by Holocene alluvium. Laterally, the aquifer is bounded by till-covered bedrock uplands.

The Milford-Souhegan aquifer has been the source of water for nine production wells yielding greater than 300 gal/min (0.67 ft^3/s): two discontinued public-supply wells, referred to hereafter as the Savage well (in the western part of the aquifer) and the Keyes well (in the eastern part of the aquifer); several wells at the Milford Fish Hatchery; and two industrial wells. A small number of houses and condominiums are still supplied by domestic wells completed in the aquifer. The Savage and Keyes wells were the primary source of public water supply for Milford during 1960-84.

In 1983, elevated concentrations of five volatile organic compounds were detected in samples of water from the Savage well: tetrachloroethylene (PCE); 1,1,1-trichloroethane; 1,2 trans-dichloroethylene; trichloroethylene (TCE), and 1,1-dichloroethane (New Hampshire Water Supply and Pollution Control Division, 1985). In that same year, high concentrations of volatile organic compounds were also detected in water from the supply well in a nearby mobile home park. As a result, both wells were removed from service in February of that year. Shortly thereafter, the Savage well was added to the U.S. Environmental Protection Agency's (USEPA) National Priority List (1986). In October 1984, elevated concentrations of 1,2-dichloroethane, 1,1,1-trichloroethane, and PCE were detected in water from the Keyes well, and the Keyes well also was removed from service.

Further sampling of surface and ground waters by the New Hampshire Water Supply and Pollution Control Division (1985) has indicated that ground-water contamination is widespread in the Souhegan River valley in and near Milford. The USEPA, the New Hampshire Department of Environmental Services, and the New Hampshire Division of Public Health Services are conducting investigations into the potential sources of contamination in the valley. These investigations have identified a number of water wells that are affected by contaminants that may have originated from a variety of sources within the aquifer. Because of (1) the importance of the sand and gravel aquifer as a regional water-supply source; (2) the large number of wells at risk; and (3) the number, distribution, and variety of potential sources of contamination; the USEPA determined that an analysis of regional ground-water flow, was needed to evaluate courses of action.

The U.S. Geological Survey (USGS), in cooperation with the USEPA, began a study in 1987 to evaluate the regional ground-water-flow system of the Milford-Souhegan aquifer and to estimate the aquifer recharge areas contributing water to the Keyes and Savage public-supply wells. Field observations and collection of geohydrologic data began in May 1987 and continued through October 1988. Hydrologic analyses and simulations were done from November 1988 through September 1989.

Purpose and Scope

This report describes the regional ground-water-flow system of the Milford-Souhegan aquifer and provides estimates of the aquifer areas contributing water to the Keyes and Savage public-supply wells. Specifically, the scope of this report includes discussions of (1) geohydrologic framework of the Milford-Souhegan aquifer, (2) regional ground-water flow, (3) approach and methods used in characterizing and simulating ground-water flow, (4) results of the flow simulation, (5) estimates of contributing areas of the Keyes and Savage wells based on simulation results, and (6) the sensitivity of model results to adjustments in model parameters. This report deals exclusively with advective flow. The study of solute transport of contaminants is beyond the scope of this work.

Approach and Methods

The geohydrology of the Milford-Souhegan aquifer was investigated by compiling available subsurface data from domestic well records in the files of the New Hampshire Department of Environmental Services, consultants reports for the Town of Milford and private concerns, and files of the USGS (Toppin, 1987). The vertical extent of the aquifer was delineated from logs of test borings and wells and from seismic-refraction data. The areal extent

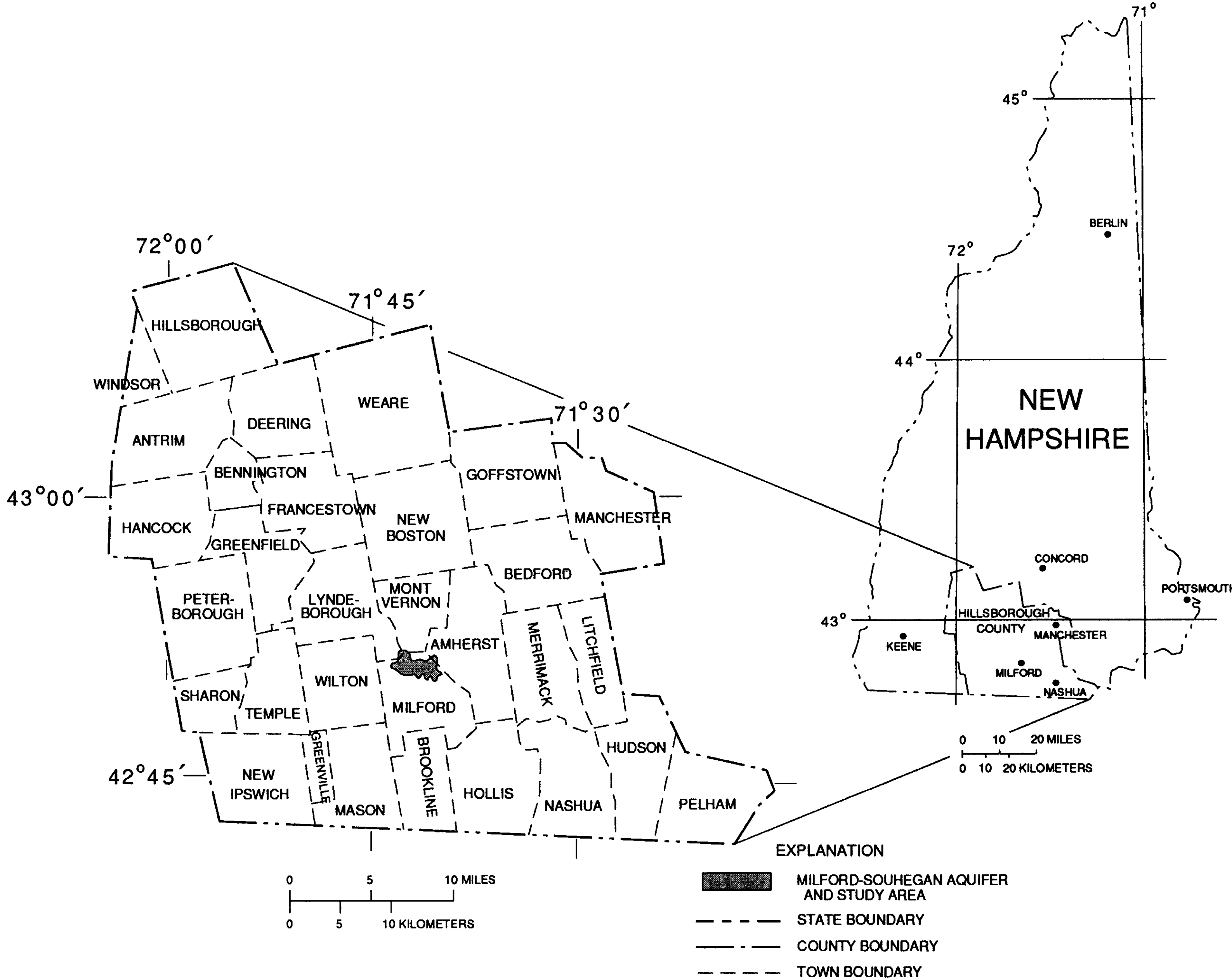


Figure 1.--Location of study area.

of the aquifer was determined by mapping the exposed stratified-drift contact with surrounding till or bedrock.

Hydrologic and lithologic characteristics of the aquifer were determined by (1) examination of more than 100 lithologic logs of test borings and developing relations between grain-size characteristics and hydraulic conductivity, and (2) analysis of results of aquifer tests. Aquifer-test results, published for six test and production wells in the aquifer, were reviewed and incorporated into this study; one comprehensive aquifer test was also conducted during this investigation. Data on hydraulic characteristics of the aquifer were collected for more than 100 wells, and pumpage data from the major users were collected for 7 wells.

Low-flow measurements of streams were made at 31 sites in August and October 1988 to determine stream-aquifer interactions, streamflow gains and losses, and estimate aquifer recharge. Concurrently with low-flow measurements, ground-water levels were measured at more than 70 wells to determine the altitude of the water table and directions of ground-water flow.

A three-dimensional, numerical ground-water-flow model (McDonald and Harbaugh, 1988) was used to simulate October 1988 ground-water flow in the Milford-Souhegan aquifer; a period that water levels were close to long-term averages. The steady-state model of this aquifer was designed with an emphasis on evaluating ground-water flow near the Savage and Keyes wells.

Contributing areas of the Savage and Keyes wells were delineated from the results of the steady-state ground-water-flow simulation and by use of a numerical particle-tracking technique (Pollock, 1989). Recharge, streambed hydraulic properties, and aquifer properties were varied in the ground-water-flow simulations to determine the effects of these parameters on the estimates of contributing areas of these wells.

Description of the study area

The sediments that constitute the Milford-Souhegan aquifer underlie an area approximately 3 mi long, in the direction of the Souhegan River, and range in width from 1,200 ft (at its western and eastern boundaries) to 1.5 mi (fig. 2). The aquifer's western and eastern boundaries are designated at constrictions in the Souhegan River valley. The western boundary is adjacent to the town of Wilton, and its eastern boundary is in downtown Milford. The aquifer's northern boundaries are the contact between saturated stratified drift and till-covered uplands. The aquifer's southern boundaries generally coincide with the same contact but also include a small area of unsaturated stratified drift.

Relief in the area underlain by the aquifer is slight; land-surface elevations range from 230 ft to 290 ft above sea level. The surrounding area is characterized by broad, rounded hills to the south (maximum elevation, 442 ft) and more rugged hills to the north (maximum elevation, 750 ft).

Surface water is drained by the eastward-flowing Souhegan River and its 12 tributaries (fig. 2). The Souhegan River has a gentle slope of about 0.002 ft/ft (foot per foot) throughout much of the valley except at the western and eastern boundaries of the Milford-Souhegan aquifer, where the slope through valley constrictions is steep. All tributaries derive their water from the surrounding uplands, except for Great Brook and the discharge ditch (fig. 2). Great Brook drains an adjacent valley south of the Souhegan River valley. Flow in the discharge ditch is primarily discharged ground water that has been used for industrial cooling and processing. Tributaries are mostly to the north of the Souhegan River and coincide with extensive adjacent upland areas.

Ground-Water Withdrawals

Ground water from the Milford-Souhegan aquifer is the primary source of water for large industrial and commercial users in the valley. The principal ground-water withdrawals from this aquifer are listed in table 1 and their locations are shown in figure 3. The Milford-Souhegan aquifer was a major source of water for the town of Milford before the discovery of contaminants from pumped water at the Savage and Keyes wells. Since 1984, the town of Milford has been dependent on imported water from the Penachuck Water Company and from a public-supply well in an adjacent river valley.

The Milford Fish Hatchery, operated by the New Hampshire Department of Fish and Game, pumps the largest amount of water from the aquifer. The combined continuous discharge from two wells at the hatchery is approximately 3.56 ft^3/s. This water is used nonconsumptively in raising various fish species and is returned to Purgatory Brook.

A manufacturing company and a wire and cable company are also major ground-water users in this valley (table 1). The manufacturing company uses ground water for noncontact cooling and process

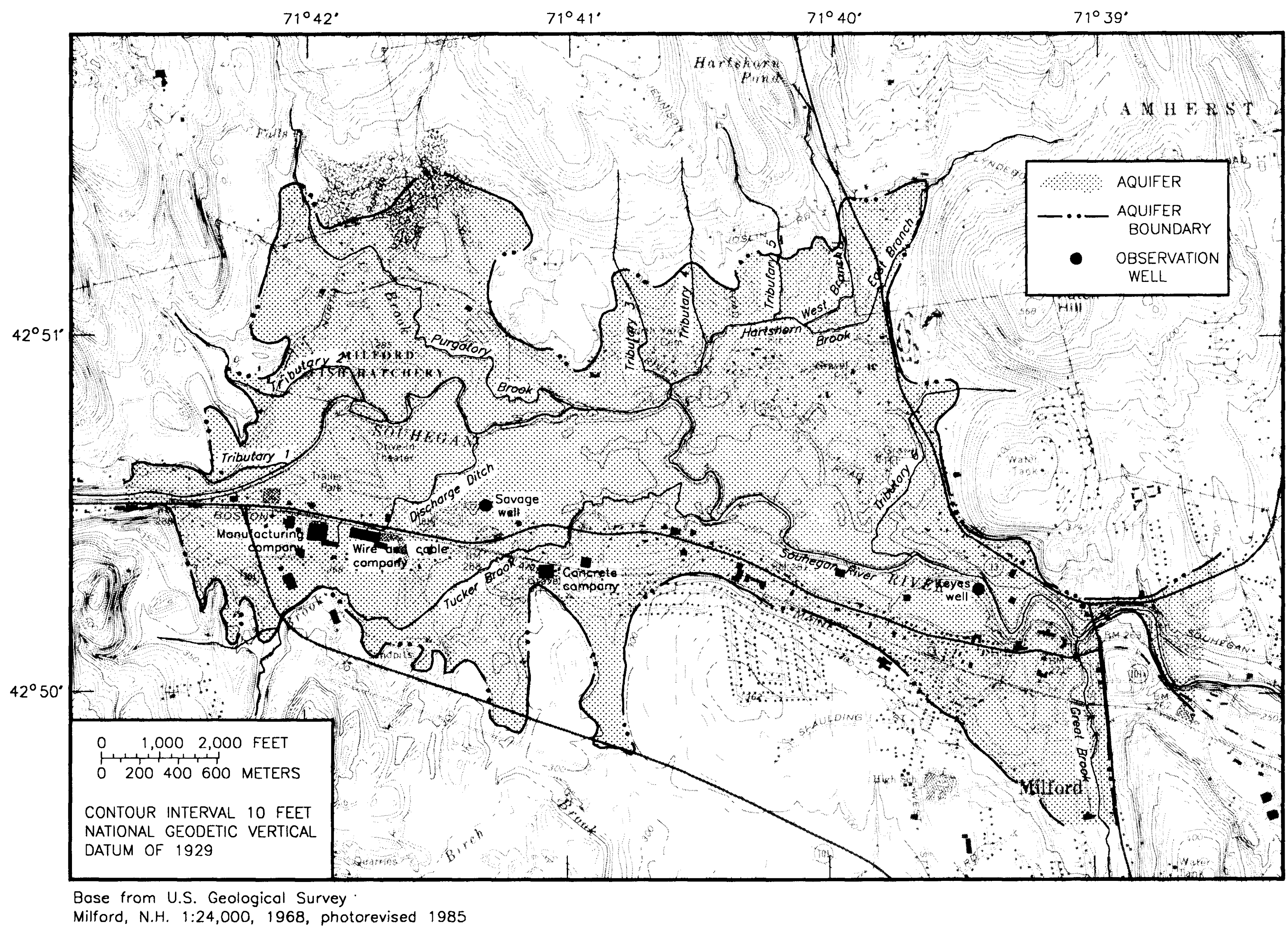


Figure 2.--Areal extent of Milford-Souhegan aquifer and locations of streams, Keyes and Savage wells, and major cultural features.

Table 1.--*Principal ground-water withdrawals from the Milford-Souhegan aquifer*

[ft^3/s, cubic feet per second; Mgal/d, million gallons per day; --, no data; mfg., manufacturing; co., company]

Well number	Site name	Maximum pumping rate (ft^3/s)	Average daily with-drawals 1988 (Mgal/d)	Average daily with-drawals 1983 (Mgal/d)	Used-water discharge	Remarks
87	Fish Hatchery 4	1.78	1.15	1.15	Purgatory Brook	Continuously pumped
208	Fish Hatchery 5	1.78	1.15	1.15	Purgatory Brook	Continuously pumped
47	A mfg. co.	.62	.36	.36	Discharge ditch	Varies with production, reduced 50 percent Sundays
49	A wire and cable co.	.36	.17	.17	Discharge ditch	Varies with production, reduced 67 percent on Sundays
73	A concrete co.	.89	--	--	Discharge onsite	Seasonal use, varies
128	Savage well	1.08	0	.21	Public supply	Pumped 6-8 hours system daily, discontinued in 1983
126	Keyes well	.67	0	.14	Public supply	Pumped 6-8 hours system daily, discontinued in 1984

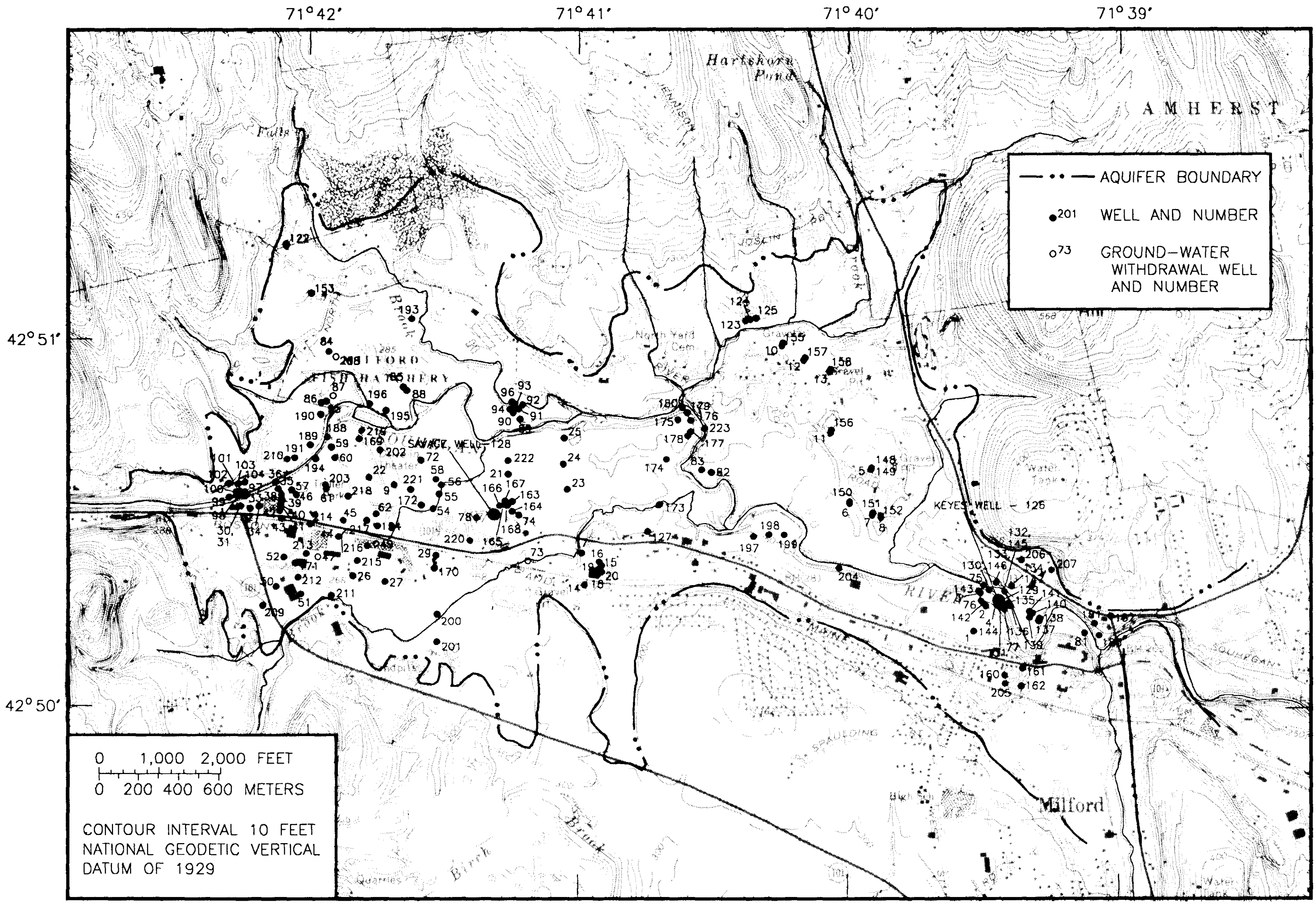
water in the casting of ferrous-alloy metal parts. The wire and cable company uses ground water for contact cooling processes in the production of polyethylene-coated wire and cable. Both companies withdraw water for 6 to 7 days a week, from the southwestern part of the aquifer, and discharge to an artificial stream. Flow in this stream, referred to as the discharge ditch, consists entirely of discharged process water and surface runoff from the two companies (New Hampshire Water Supply and Pollution Control Division, 1985).

The concrete company withdraws ground water at up to 0.89 ft^3/s as wash water for sand and gravel operations. This use is seasonal (to augment surface-water supplies). Used water is discharged on site, where it returns to the aquifer.

Domestic wells withdraw minor amounts of ground water from the aquifer. Most domestic wells are to the north of the Souhegan River.

Previous Investigations

The geohydrology of the Milford area has been investigated in regional studies by the USGS (Cotton, 1977; Toppin, 1987), as well as in site-specific



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 3.--Locations of wells in the Milford-Souhegan aquifer.

studies by the New Hampshire Water Supply and Pollution Control Division (1985) and private consultants (for example, Roy F. Weston, Inc., 1984 and 1987; D.L. Maher Company, 1985; and NUS Corporation, 1987). The New Hampshire Water Supply and Pollution Control Division Report (1985) describes ground-water flow and volatile-organic-compound movement near the Savage well; a simple numerical model was used to investigate advective transport. Lithologic, surface-geophysical, and hydrologic data were collected by several consulting firms. The surficial geology of the study area was mapped by Koteff (1970). The bedrock geology of the study area was investigated by Lyons and Bothner (1989).

Acknowledgments

Information for this study was provided by the New Hampshire Water Supply and Pollution Control Division, the Department of Environmental Services, the Water Resources Division, and the Department of Fish and Game. Unpublished borehole data were provided by HMM Associates. The authors wish to acknowledge officials and employees of the Town of Milford for their cooperation and assistance. Thanks are also extended to individuals who granted USGS personnel access to their property.

GEOHYDROLOGY

The study area is situated in the western region of the Massabesic-Merrimack-Rye terrane (Lyons and Bothner, 1989)--a northeast-southwest trending geologic structure. The northeastern boundary of the Massabesic-Merrimack-Rye terrane is defined by the Campbell fault, located approximately 2 mi west of the study area. The southeastern boundary of the Massabesic-Merrimack-Rye terrane is defined by the Atlantic Ocean. The Massabesic-Merrimack-Rye terrane consists of Precambrian to Ordovician plutonic rocks, and Silurian to Devonian metasedimentary rocks (Lyons and Bothner, 1989). Plutonic rocks, granite and gneiss, underlie the Milford-Souhegan aquifer and adjacent upland areas.

The Milford-Souhegan aquifer consists of unconsolidated Pleistocene glacial sediments that fill a buried pre-Pleistocene valley. Bedrock-surface topography, which defines the valley, probably is a result of preglacial drainage patterns that were deepened by glacial erosion. The lithology of the unconsolidated glacial sediments ranges from well-sorted stratified drift consisting of coarse sands and gravels to poorly sorted, dense glacial till.

Water in the Milford-Souhegan aquifer is mostly unconfined, but is semiconfined in places. The aquifer is recharged by infiltrating precipitation, streamflow seepage, and infiltrating runoff from adjacent uplands. Ground water discharges to the Souhegan River and some of its tributary streams. Ground water in the aquifer interacts with water in the bedrock; however, the degree of interaction is not known. Volatile organic compounds, originating from land uses on top of surficial deposits in the river valley, have been detected in bedrock wells. Volatile-organic-contaminant migration suggests that a hydraulic connection exists between the Milford-Souhegan aquifer and the underlying bedrock.

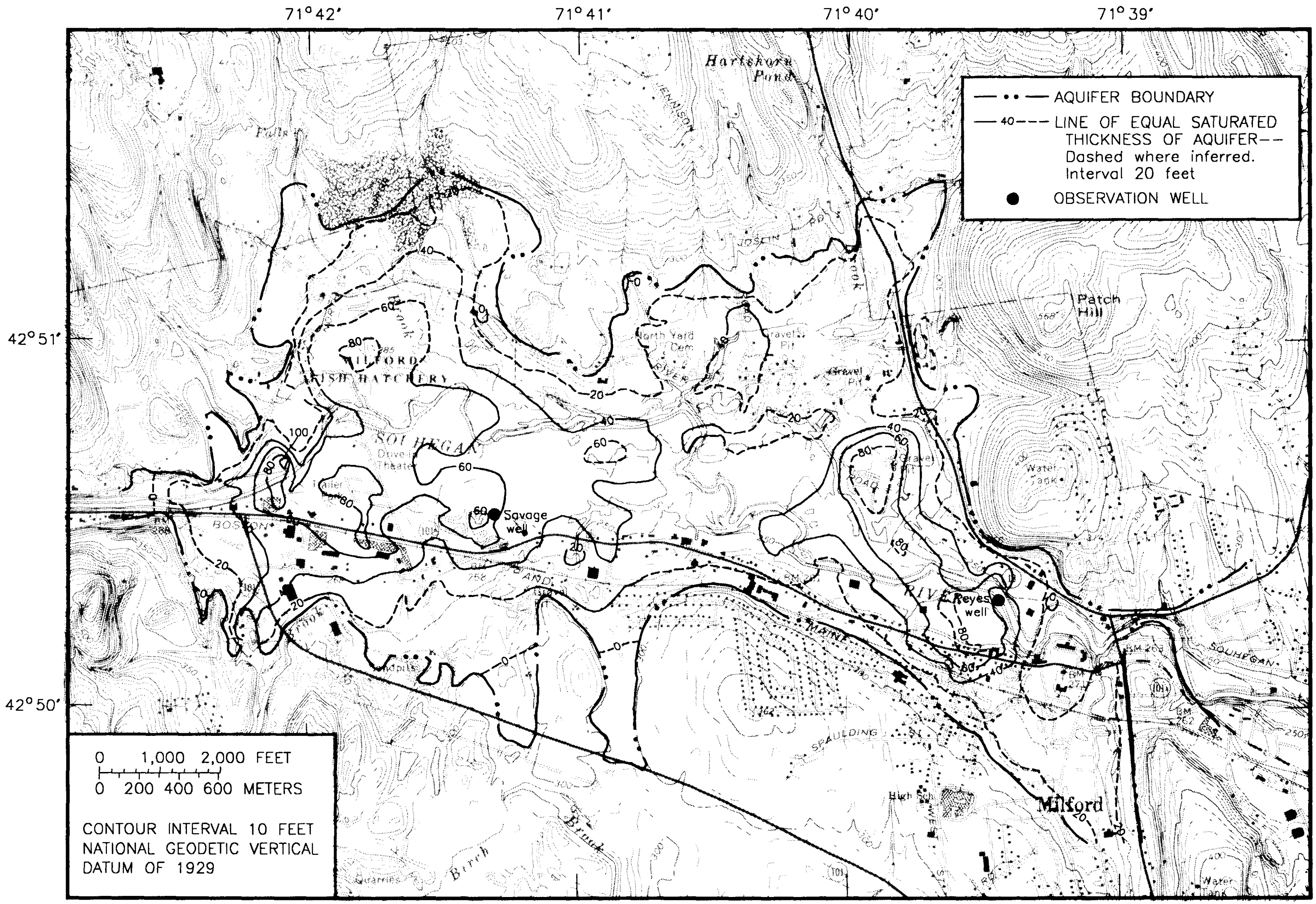
Milford-Souhegan Aquifer

The Milford-Souhegan aquifer is described with respect to geometric configuration, lithology, modes of deposition, and water-bearing properties. Information on aquifer properties was obtained primarily from wells and test-holes.

Areal Extent and Saturated Thickness

The areal extent of the Milford-Souhegan aquifer is shown in figure 4. The aquifer boundary corresponds to the areal limit of the exposed stratified-drift and river-valley deposits. The stratified-drift and river-valley deposits are bounded by exposed till along 90 percent of its border. In general, the aquifer boundary coincides with the zero-saturated thickness of the stratified-drift and river-valley deposits. Unsaturated stratified drift and till, which are important sources of recharge to the aquifer, are present along the southern aquifer boundary outside the zero saturated-thickness contour.

Saturated-thickness contours in figure 4 represent the vertical distance between the average level of the water table and the bedrock surface. Bedrock-altitude and water-table-altitude maps were used to produce the saturated-thickness map. Bedrock surface and the average level of the water table were determined from available borehole data and from measured water levels in wells and estimates of surface-water stage elevations. Locations of wells and borings are shown in figure 3. Borehole



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 4.--Areal extent and saturated thickness of the Milford-Souhegan aquifer.

data on the thickness of till and stratified drift at selected wells in the study area are summarized in Appendix A.

Saturated thickness is greatest in the west-central, east-central, and southeastern parts of the aquifer. Saturated thickness ranges from zero along the aquifer boundary to 112 ft in the west-central part of the aquifer (fig 4). Saturated thickness is zero adjacent to bedrock outcrops in the north-central part of the aquifer, where the Souhegan River bends to the south, and along the Souhegan River approximately 1,000 ft east of the Keyes well.

Stratified-drift deposits comprise more than 80 percent of the total saturated thickness at most locations in the aquifer, and till comprises the remaining percentage. The maximum saturated thickness of stratified drift is 75 ft at well 5[1] in the east-central part of the aquifer (fig. 4). The maximum saturated thickness of till is 47 ft, or 50 percent of the total saturated thickness, at well 40 in the west-central part of the aquifer (figs. 3 and 4).

Lithologic Characteristics and Depositional History

In the study area, till generally is a poorly sorted mixture of sand, gravel, silt and clay. The till forms a discontinuous mantle overlying bedrock. Two distinct tills occur in the study area--a dense, brown lower till overlying bedrock and a sandy, gray upper till--correspond to two glacial advances (Koteff, 1970). In many locations, till is absent and stratified drift rests directly on bedrock.

Stratified-drift deposits are glacial sediments that have been transported, reworked, and sorted by fluvial action. Stratified drift in the study area was deposited during four stages of Pleistocene glacial depositional activity (Koteff, 1970). During the first two stages, the Souhegan valley was occupied by ice. Ice-contact materials were deposited by meltwater streams flowing southwest, and then southeast, over bedrock spillways a few miles south of the study area. During the third stage, the ice retreated northward to what is now downtown Milford. Ice-contact deposits, controlled by a bedrock spillway at an elevation of about 300 ft, formed from streams originating at the southern margin of the study area. The streams flowed east in the area now occupied by Great Brook.

The stratified drift that comprises the Milford-Souhegan aquifer was deposited during the fourth stage. This material represents outwash plain rather than ice-contact sedimentation. The Souhegan valley by this time was relatively free of ice, and a west-to-east drainage pattern caused coarser material to be deposited in the western part of the aquifer than in the eastern part. The drainage pattern was complicated by channels near Purgatory and Hartshorn Brooks (fig. 2), which flowed south from the melting glacier. Water ponded behind bedrock along what is now the railroad bed in downtown Milford, south of the present river outlet. The fine-grained sediments found at depth near the Keyes well may have been deposited in the glacial lake. Water leaving the study area flowed into Glacial Lake Merrimack, an arm of which stretched up the Souhegan valley. During the last stages of deglaciation, Glacial Lake Merrimack was drained and terraces were cut into the stratified drift in the Souhegan valley (Koteff, 1970).

Stratified drift in the study area is composed chiefly of sand and gravel interbedded with some silt. Stratified-drift deposits logged at test holes in the study area form a complex pattern laterally and vertically. Individual lithologies cannot be traced across the region, and correlations can rarely be made except between wells a few feet apart. This kind of heterogeneity is typical of an outwash-plain environment, where braided stream channels migrate across a valley floor (Koteff, 1970). In such an environment, sedimentation varied with changes in the melting glacier.

Although lithology is highly variable in the study area, sediments are coarser in the west than in the east. At the Savage well, the drift consists of coarse-grained sands and some gravel. At the Keyes well, drift grades with depth from medium sand to fine sand and silt.

Hydraulic Characteristics

Estimates of horizontal hydraulic conductivity, transmissivity, and storage coefficient are given in table 2 for stratified drift in the Milford-Souhegan

---

1 The well-numbering system used in this report is unique to this report and is designed to identify wells in a simple numerical fashion.

Table 2.--*Values of transmissivities, hydraulic conductivities, and storage coefficients from tests of wells in the Milford-Souhegan aquifer*

[no., number; ft, feet; ft^2/d, feet squared per day; ft/d, feet per day; --, no data]

Local well name	Pumped well no.	Observation well no.	Screened interval depth below land surface (ft)	Date of test	Transmissivity (ft^2/d)	Horizontal hydraulic conductivity (ft/d)	Storage coefficient	Method of test analysis[1]
Keyes	126	--	50 - 60	10/88	1,500	17	0.00900	7
Keyes 2D	126	2	54 - 56	10/88	3,280	50	.002	1
Keyes 3D	126	3	49 - 51	10/88	4,600	74	.0016	1
Keyes 4D	126	4	50 - 52	10/88	1,210	20	.0005	1
Potter 1D	126	132	55 - 57	10/88	780	12	.0013	1
Potter 2D	126	133	56 - 58	10/88	180	3	.00002	7
Potter 3D	126	134	56 - 58	10/88	1,390	23	.0013	1
Keyes	126	--	50 - 60	6/72	--	--	--	--
Keyes 2	126	130	52 - 60	6/72	6,400	98	.12000	2
Keyes 3	126	129	41 - 51	6/72	6,000	12	--	--
Ford 34	135	--	40 - 50	9/68	--	--	--	--
Ford 1	135	139	35 - 50	9/68	4,000	110	--	2
Ford 4	135	141	--	9/68	1,400	83	--	2
Ford 5	135	140	--	9/68	1,400	60	--	2
FH-3	86	--	33 - 43	3/85	--	540	--	3
FH-1	84	--	51 - 66	3/85	--	340	--	3
FH 85-6	93	--	22 - 25	3/85	18,000	--	.00400	2
FH-5	208	--	50 - 65	3/85	86,700	1,240	.03800	2
FH-5	208	--	50 - 65	3/85	67,900	970	--	3
MI-28	--	43	35 - 55	8/83	1,100	39	--	4
MI-29	--	171	31 - 51	8/83	540	13	--	4
MI-30	--	44	27 - 72	8/83	1,200	35	--	4
MI-31	--	45	36 - 54	8/83	220	6	--	4
MI-32	--	46	30 - 75	8/83	100	3	--	4

Table 2.--*Values of transmissivities, hydraulic conductivities, and storage coefficients from tests of wells in the Milford-Souhegan aquifer*--Continued

Local well name	Pumped well no.	Observation well no.	Screened interval depth below land surface (ft)	Date of test	Transmissivity (ft^2/d)	Horizontal hydraulic conductivity (ft/d)	Storage coefficient	Method of test analysis[1]
Savage	128	--	42 - 52	6/60	29,400	490	--	3
Savage	128	--	42 - 52	3/57	7,600	130	.10500	2
Savage	128	--	42 - 52	4/81	9,700	150	--	5
MI-2	128	163	37 - 47	4/81	7,200	--	--	5
MI-3	128	164	44 - 49	4/81	9,200	--	.03000	5
MI-4	128	165	39 - 49	4/81	8,500	--	--	5
MI-5	128	166	39 - 49	4/81	7,300	--	.03000	5
MI-6	128	167	--	4/81	9,100	--	--	5
MI-6A	128	168	--	4/81	9,300	--	--	5
MI-7	128	21	--	4/81	8,300	--	.06000	5
RFW-1	14	--	8 - 28	11/86	20	1	--	6
RFW-2	15	--	10 - 35	11/86	500	12	--	6
RFW-3	16	--	13 - 43	11/86	660	9	--	6

[1] Method of test analysis and source of data:
1. Aquifer test--Nueman (1974).
2. Aquifer test--Jacob (1946).
3. Single-well pumping test--Meyer (Meyer, 1963, p. 83).
4. Slug tests (New Hampshire Water Supply and Pollution Division, 1985).
5. Walton (New Hampshire Water Supply and Pollution Division, 1985).
6. Single-well recovery test (Weston, 1986).
7. Aquifer test--Walton (1962).

aquifer. Stratified-drift deposits are the most permeable deposits in the Milford-Souhegan aquifer. Hydraulic conductivities, transmissivities, and storage coefficients were calculated from aquifer tests. Hydraulic conductivities are also listed for slug tests done by the New Hampshire Water Supply and Pollution Control Division (1985) and for single-well recovery tests done by Roy F. Weston, Inc. (1987).

Aquifer tests have been conducted at wells 86, 84, 93, and 208 by the New Hampshire Fish and Game Department (D.L. Maher and Company, 1985 and 1988) and near a manufacturing company, at wells 43, 171, 44, 45, and 46 by New Hampshire Water Supply and Pollution Control Division (1985). The Savage well, previously a public-supply well, was tested on several occasions (R.E. Chapman Co., 1957a, 1957b, 1960, and 1981; New Hampshire Water Supply and Pollution Control Division, 1985). A limited aquifer test was conducted at the Keyes well in 1972 (R.E. Chapman Co., 1972) and at an adjacent test well site across the Souhegan River in 1957 and 1968 (R.E. Chapman Co., 1957a and 1968). During October 1988, USGS personnel conducted a 7-day aquifer test at the Keyes well. The Keyes well was pumped continuously for 7 days at a rate of 0.67 ft^3/s and recovery was monitored for 24 hours. Thirteen observation wells were installed for this test to supplement available wells. Data collected during the 1988 Keyes aquifer test are presented in Appendix B.

Estimates of hydraulic conductivity from aquifer tests (table 2) were determined by calculating transmissivity from aquifer tests and dividing by saturated thickness. Transmissivity was calculated according to the methods of Theis (1935), Jacob (1946) (corrected for dewatering), and Neuman (1974) for unconfined conditions. Transmissivity was also calculated according to the methods described by Walton (1962) for confined flow with leaky conditions (Kruseman and de Ridder, 1983, p. 81-84). The application of the Neuman methods on analyzing aquifer-test results is probably the most appropriate method to evaluate aquifer tests in the Milford-Souhegan aquifer. The Neuman method considers the effects of partial penetration and delayed gravity response to the pumped aquifer. The Neuman method was used along with the Walton method to evaluate drawdown data collected from the Keyes well aquifer tests. Values of transmissivity, horizontal hydraulic conductivity, and storage calculated by the Neuman method were approximately the same as those calculated by the Walton method (table 2, footnote 1).

Horizontal hydraulic conductivity was also estimated on the basis of the specific capacity of wells by methods described by McClymonds and Franke (1972) and Meyer (1963). These methods are used to analyze single-well aquifer-test data in which drawdown data are available only for the pumped well.

Horizontal hydraulic conductivities are highest in the western part of the aquifer and lowest in the eastern part of the aquifer. These data correspond to regional glacial stratigraphic trends (Koteff, 1970). Hydraulic conductivity near the Savage well is generally an order of magnitude greater than that at the Keyes well.

Most estimates of hydraulic conductivity are less than 500 ft/d, except values reported for FH-5 (table 2). It is probable that some of the observed variation in hydraulic conductivity is caused by variations in pump well-design and efficiency and methods of analysis.

The Keyes aquifer test of October 1988 provides information on hydraulic characteristics of the aquifer, including horizontal hydraulic conductivity, vertical hydraulic conductivity, and storage coefficient. Horizontal hydraulic conductivities and storage coefficients are given in table 2 for the Keyes well and for six deep observation wells. Six well couples, or nested wells, were installed for this test; a deep well was screened at the same altitude as the Keyes well, approximately 30 ft below the water table, and a shallow well at the same location was screened at the water table. Information from nested-well sites were used to determine vertical hydraulic conductivity.

Drawdowns in the deep observation wells during the Keyes aquifer test responded in a fashion characteristic of a semiconfined aquifer. Figure 5 shows the drawdown at a nested-well site resulting from constant pumping of the Keyes well. Delayed-yield effects, commonly observed in an unconfined aquifer, were not observed. The probability exists that the aquifer test was terminated before the effects of delayed yield could impact drawdowns in the deeper wells. Drawdowns in the deep wells stabilized after one-half day of pumping. Drawdowns in the shallow, water-table wells did not stabilize during the 7-day test and can be attributed to low-gravity drainage. Drawdowns at the shallow wells were generally one-half the drawdowns at the deep wells (Appendix B).

Vertical hydraulic conductivity ranges from 0.15 to 2.5 ft/d in the Keyes well field area. These rates are approximately one-tenth the horizontal hydraulic conductivity (table 2). Vertical hydraulic conductivity (Kv) was calculated from vertical-leakage rates calculated for the Keyes aquifer test following methods described by Neuman (1974).

Storage coefficients determined from aquifer tests at the Keyes well and deep observation wells (wells 2, 3, 4, 132, 133 and 134) are indicative of a semiconfined or confined aquifer (table 2). The low estimates of storage coefficient are partly attributed to effects of the partial-penetration of the wells. The screened intervals of the observation wells are short and are positioned approximately 30 ft below the water table. At the screened interval, the aquifer remained saturated for the duration of the October 1988 aquifer test. Therefore, virtually all water released from storage at this interval is from water expansion and aquifer compression.

Ground-Water Levels and Flow Directions

Ground-water levels and flow in the Milford-Souhegan stratified-drift aquifer were determined by reviewing data collected during previous studies and by observations made during this study. Observations made during this study include measurements of ground-water levels to determine directions of ground-water flow and measurement of streamflow along the

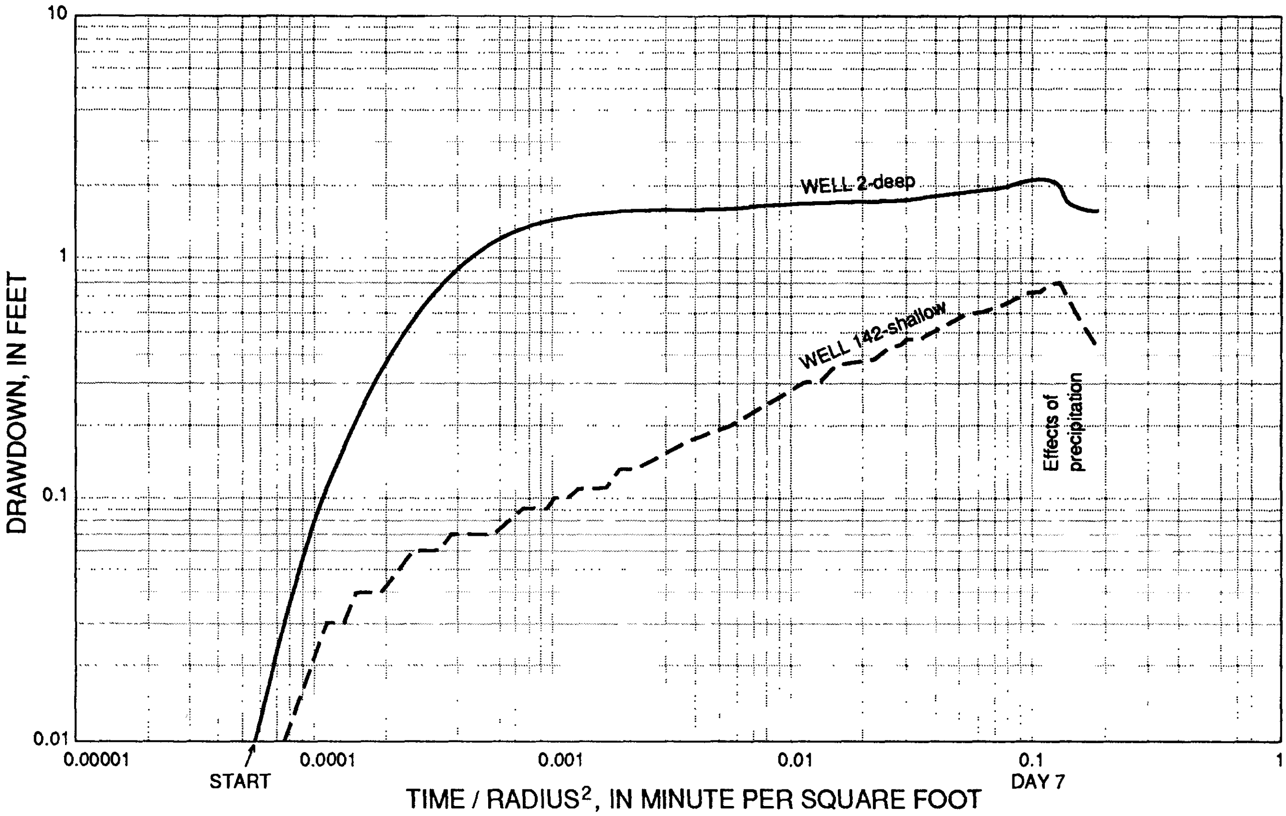


Figure 5.--Drawdowns at nested observation wells during the Keyes well aquifer test of October 17, 1988.

Souhegan River and its tributaries to identify gaining and losing reaches of the streams.

Fluctuations in the elevation of the water table are caused principally by seasonal variation in recharge to the aquifer and by variations in ground-water withdrawals. A hydrograph for well 29 (USGS observation well MOW-36), which is southwest of the Savage well, is shown in figure 6. Data from this well for the period 1962-88 show that the water level rises in response to recharge in March and April, whereas the water level declines during the summer and early fall. The maximum observed annual fluctuation in water level from 27 years of record is 6 ft in 1978. During most years, annual water level fluctuations are 3.5 to 4 ft. Since 1981, annual water-level fluctuations have ranged from approximately 1.5 to 3 ft. It is not clear whether the decrease in the magnitude of water-level fluctuations since 1981 is due to the 1983 shutdown of the Savage well, approximately 1,200 ft away from well 29, or to increased uniformity in annual distribution of rainfall in Milford during the past few years (National Oceanic and Atmospheric Administration, 1989).

Water-level-elevation data for 154 wells in the study area are listed in Appendix C; a statistical summary for wells with more than 4 measurements is given in table 3. Measured water-level fluctuations generally range from 2 to 4 ft. Water-level fluctuations exceeding 6 ft are usually caused by ground-water withdrawals. Water levels affected by pumpage include (1) production wells 84 and 208 at the Milford Fish Hatchery and observation well 87 nearby (table 3) and (2) production well 49 at the wire and cable company and observation well 164 nearby (Appendix C). Water-level fluctuations greater than 6 ft also occur at well 46 and are caused by recharge from seepage losses of the Souhegan River.

The approximate configuration of the water table in the Milford-Souhegan aquifer on October 13 and 14, 1988, is shown in figure 7. The data were obtained from 49 measurements of water levels made by the USGS and the New Hampshire Department of Environmental Services (Richard Pease, New Hampshire Department of Environmental Services, written commun., 1988). Ground-water-level contours coincide with surface-water elevations

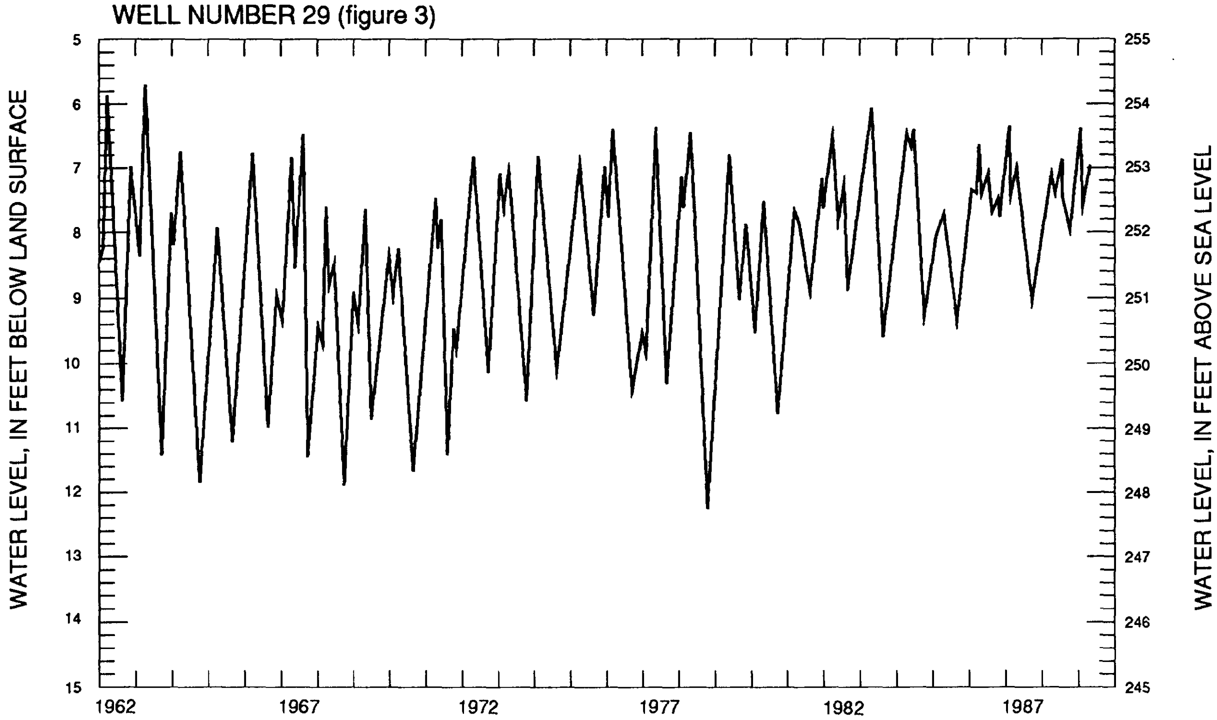


Figure 6.--Hydrograph showing long-term water levels at an observation well in the Milford-Souhegan aquifer.

along the Souhegan River and its major tributaries and are consistent with seepage loses and gains along stream reaches.

Ground-water levels in the Milford-Souhegan aquifer on October 13 and 14, 1988, are believed to be representative of long-term average conditions. The October 1988 measured depth to water at well 29 (fig. 3) of 8.1 ft, was relatively close to the long-term mean depth to water of 8.5 ft considering annual water-level fluctuations are approximately 6 ft. Although the October 1988 water level was slightly above the long-term mean for this well, the level was low for the year, and 1988 levels at this well were higher than average (1961-88). The 27-year water-level record at well 29 may be weighted lower by exceptionally dry periods in the late 1960s and late 1970s.

The generalized horizontal direction of ground-water flow is also illustrated in figure 7. As indicated, the flow is from areas of high water-table altitude to areas of low water-table altitude. The flow pattern has a regional component from the valley sides to the Souhegan River and a secondary component in the downstream direction, eastward along the Souhegan River.

The vertical flow of ground water in an idealized stratified-drift aquifer with till and bedrock is shown in figure 8. Ground water that recharges the stratified-drift aquifer generally moves from an aquifer boundary, or ground-water divide, and discharges as streamflow. Water that enters the aquifer farther from its point of discharge follows a deeper path through the aquifer than water that enters near a discharge point (fig. 8). Ground-water withdrawals will alter directions of natural flow in a valley-fill aquifer.

Vertical-flow directions in the Milford-Souhegan aquifer follow the general pattern shown in figure 8 except where natural flow is significantly affected by ground-water withdrawals. Vertical-flow direction and rates can be inferred from vertical-head gradients at nested wells. Vertical-head gradients are slight throughout the aquifer, generally less than 0.001 ft/ft, except near ground-water withdrawals. Ground-water withdrawals cause an increase in vertical flow. Increases in vertical-head gradients, shown in figure 9, were observed during the Keyes aquifer test of October 1988. Vertical-head gradients increased from near zero to 0.07 ft/ft during the aquifer test. Monitoring of nested wells

Table 3.--*Summary of ground-water levels measured at selected wells in the Milford-Souhegan aquifer*

[no., number; ft, feet]

Well no.	No. of observations	Measured high (ft above sea level)	Date	Measured low (ft above sea level)	Date	Range (ft)
1	7	237.51	09/28/88	235.18	02/02/89	2.33
2	7	236.08	10/05/88	234.12	02/02/89	1.96
3	7	235.74	09/28/88	233.99	02/02/89	1.75
4	7	235.75	10/31/88	233.99	02/02/89	1.76
6	11	238.67	10/25/88	235.25	02/02/89	3.42
7	11	238.02	10/25/88	234.87	02/02/89	3.15
8	11	237.73	10/25/88	234.52	02/02/89	3.21
14	5	255.31	12/08/86	253.62	09/18/87	1.69
15	5	251.46	12/08/86	248.52	09/18/87	2.94
16	6	251.06	12/08/86	248.30	09/18/87	2.76
17	6	251.00	12/08/86	248.45	09/18/87	2.55
29	12	251.92	10/21/88	253.60	11/21/88	1.68
30	10	267.31	05/11/84	263.86	10/06/83	3.45
31	10	267.38	05/11/84	263.60	09/29/83	3.78
32	10	266.54	05/11/84	263.28	09/27/83	3.26
33	9	266.39	02/10/84	264.45	10/06/83	1.94
34	8	265.26	05/11/84	261.51	09/27/83	3.75
35	9	262.97	05/11/84	257.24	10/06/83	5.73
36	9	263.57	05/11/84	261.08	09/27/83	2.49
37	9	263.14	05/11/84	259.89	10/21/83	3.25
38	8	263.02	05/11/84	259.04	10/21/83	3.98
39	7	265.07	05/11/84	260.10	09/12/84	4.97
40	9	262.95	11/05/84	258.78	10/21/83	4.17
41	9	263.06	05/11/84	258.57	10/21/83	4.49
42	8	263.16	05/11/84	258.57	10/21/83	4.59
43	10	263.18	11/28/83	258.86	09/27/83	4.32
44	10	259.40	11/28/83	254.31	09/27/83	5.09
45	10	259.89	11/28/83	255.06	01/26/84	4.83
46	9	265.09	01/26/84	258.00	09/27/83	7.09
47	5	254.67	11/28/83	250.71	11/09/84	3.96
84	5	240.57	02/02/89	231.73	10/21/88	8.84
87	8	250.16	09/01/88	239.91	10/21/88	10.25

Table 3.--*Summary of ground-water levels measured at selected wells in the Milford-Souhegan aquifer--Continued*

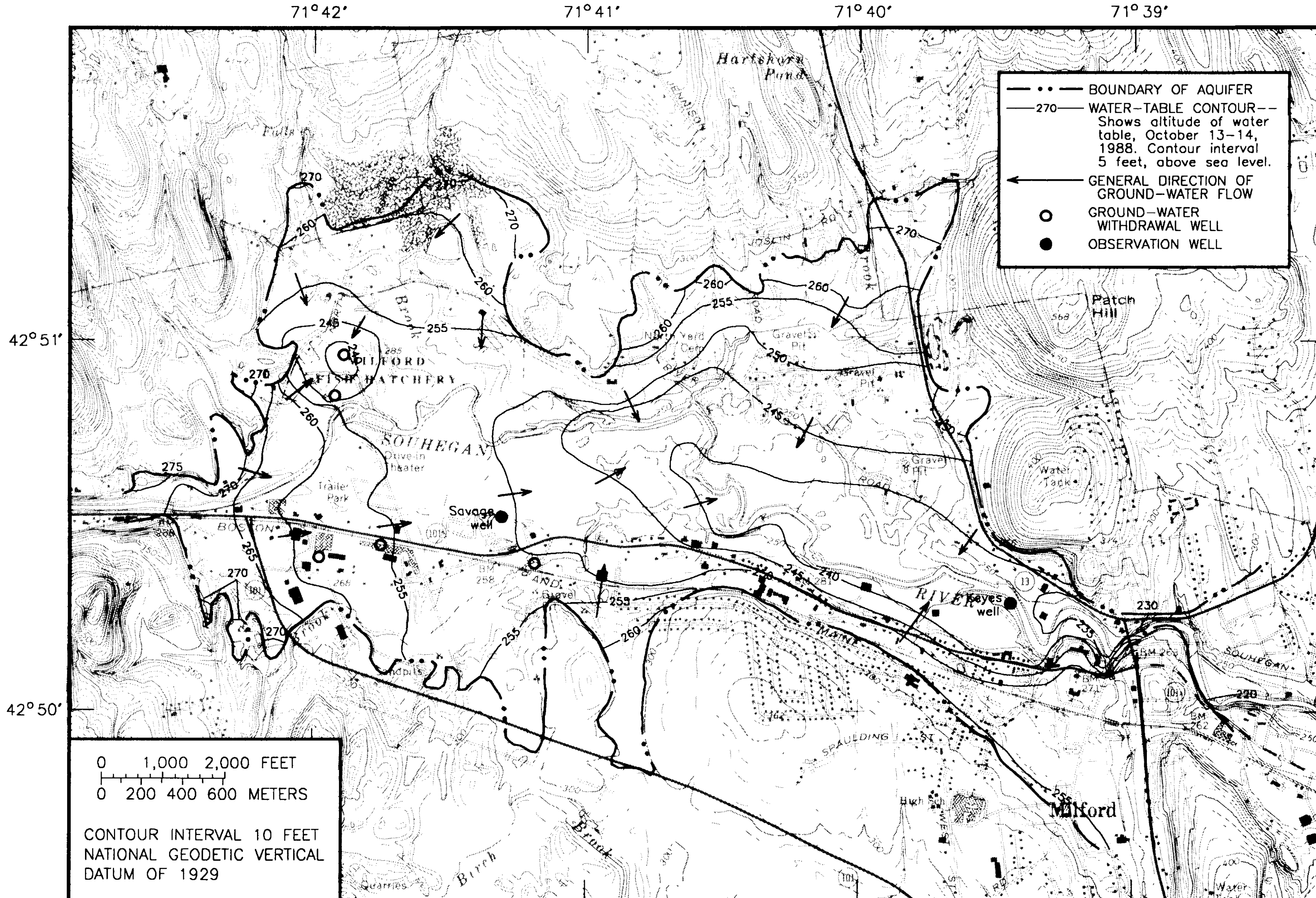
Well no.	No. of observations	Measured high (ft above sea level)	Date	Meassured low (ft above sea level)	Date	Range (ft)
123	11	251.70	10/22/88	248.75	02/02/89	2.95
124	11	252.55	10/24/88	249.33	02/02/89	3.22
125	11	251.95	10/22/88	248.95	02/02/89	3.00
126	5	235.76	10/31/88	231.69	05/10/72	4.07
132	6	235.80	10/31/88	234.10	02/02/89	1.70
133	6	235.80	10/31/88	234.11	02/02/89	1.69
134	6	236.03	10/31/88	234.36	02/02/89	1.67
142	7	236.08	10/31/88	234.07	02/02/89	2.01
143	7	235.67	09/28/88	233.88	02/02/89	1.79
144	7	235.75	10/31/88	233.97	02/02/89	1.78
145	6	235.81	10/31/88	234.08	02/02/89	1.73
146	6	235.82	10/31/88	234.09	02/02/89	1.73
147	6	236.12	10/05/88	234.37	02/02/89	1.75
150	11	238.69	10/25/88	235.53	02/02/89	3.16
151	11	237.96	10/25/88	234.98	02/02/89	2.98
152	11	239.83	10/25/88	236.27	02/02/89	3.56
171	10	262.05	11/28/83	257.01	09/27/83	5.04
208	11	234.62	12/20/88	223.89	10/21/88	14.98

beyond the wells 6,000 ft northwest of the Keyes well revealed no significant variation in vertical-head differences during the Keyes aquifer test.

Ground-water withdrawals at the Keyes well alters natural ground-water flow direction and rates in the vicinity of the Keyes well. Potentiometric surfaces and inferred horizontal ground-water-flow directions before and during the Keyes aquifer test are shown in figure 10. Figure 10 was constructed using ground-water levels from wells screened at the same depth as the pumped Keyes well. Pumping of the Keyes well increased head gradients southwest of the well and reversed natural head gradients southeast of the well. Pumping of this well also induces infiltration of water from the Souhegan River. Ground-water withdrawals at the Keyes well had less of an impact on ground-water levels from water-table wells--wells screened at or near the water table (fig. 9). The least affected water levels were from water-table wells on the northeastern side of the Souhegan River.

Stream-Aquifer Interaction

The Souhegan River and its tributaries provide recharge to the Milford-Souhegan aquifer and receive discharge from the aquifer. Stream-aquifer interactions are important in terms of understanding ground-water flow and quantifying available ground-water supplies. Induced infiltration from surface water sustains ground-water withdrawals during periods of low natural recharge from infiltrating precipitation. Streamflows for the Souhegan River and its tributaries at base flow are listed in table 4. Locations of streamflow-measurement sites, losing



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 7.--Altitude of the water table in the Milford-Souhegan aquifer, October 1988.

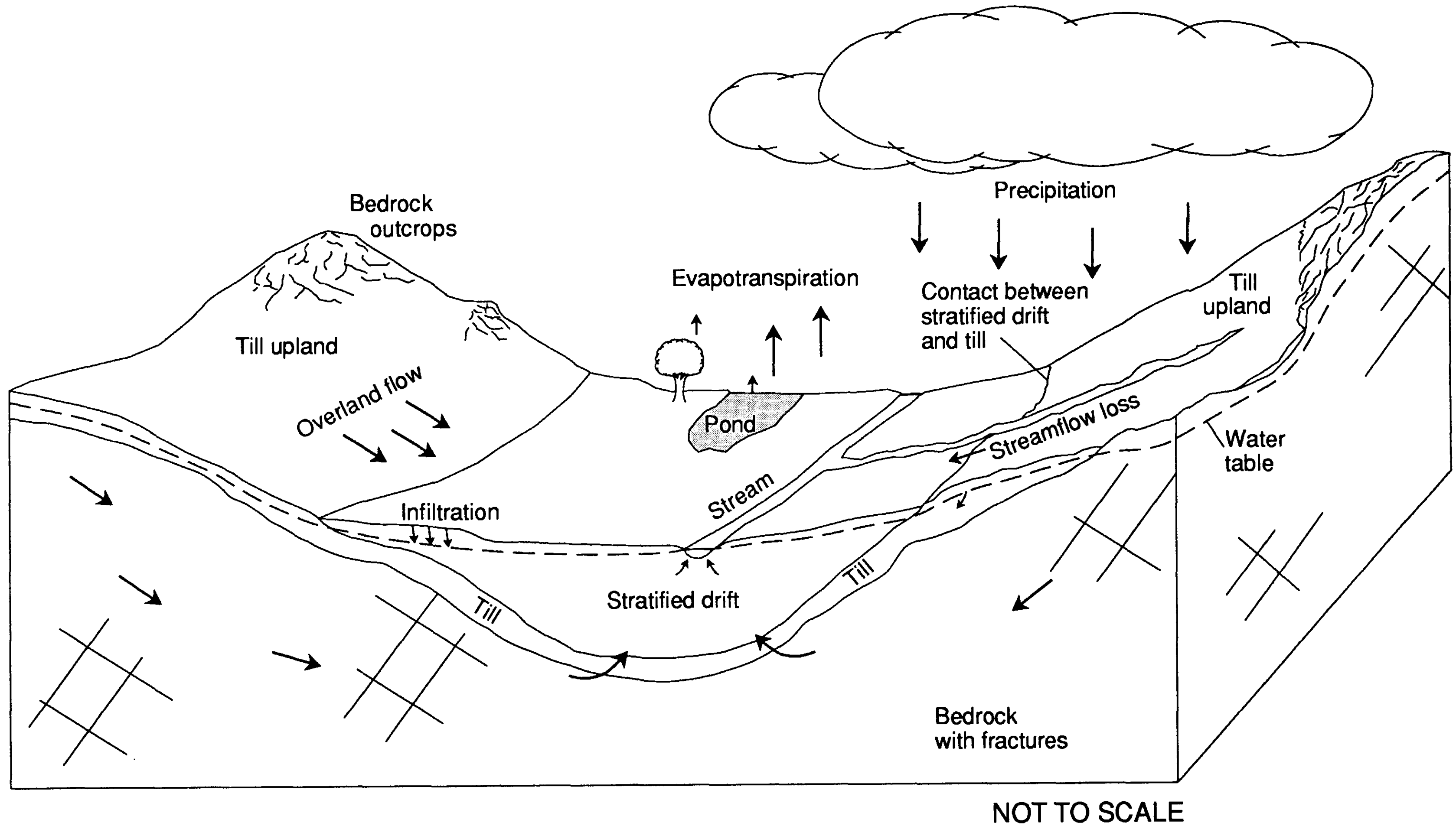


Figure 8.--Hydrologic cycle and generalized ground-water-flow patterns in a typical river-valley aquifer (From Morrissey, 1989, fig. 2).

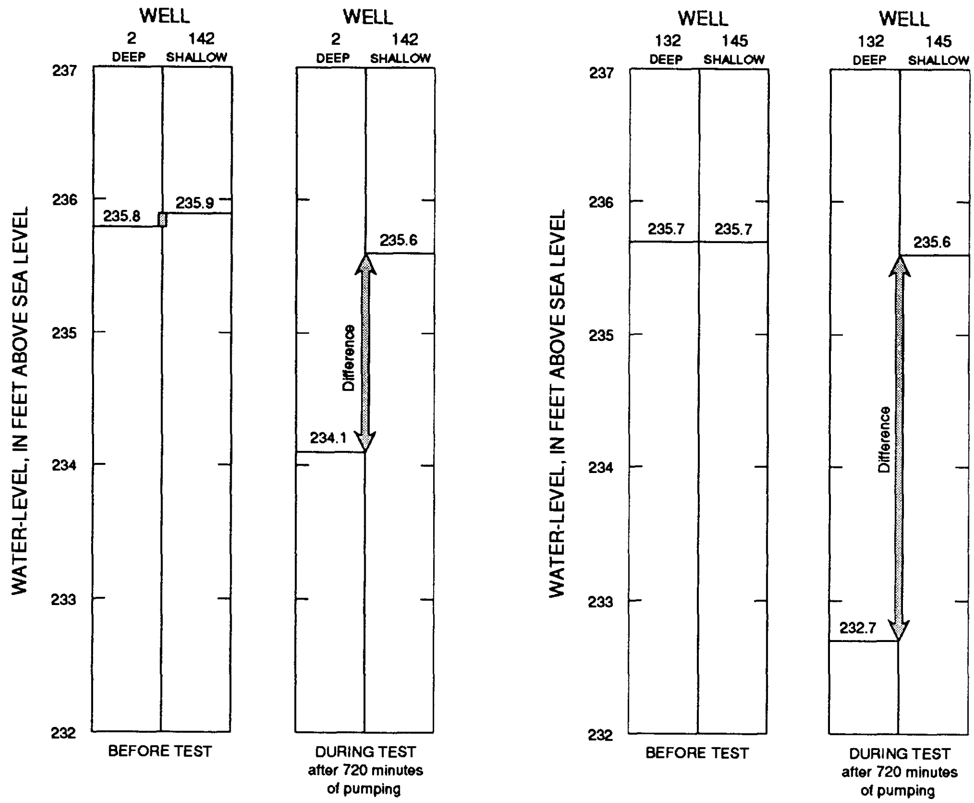


Figure 9.--Water-level differences in nested wells before and during the Keyes aquifer test.

and gaining stream reaches for June and October 1988, (and Milford Fish Hatchery discharge location) are shown in figure 11. The natural pattern of stream-aquifer interaction has been altered in the Milford-Souhegan aquifer by ground-water withdrawals at production wells 87 and 208 at the Fish Hatchery and return of nonconsumptive waters into Purgatory Brook. A similar situation occurs at production wells 47 and 49 and return of nonconsumptive waters into the discharge ditch.

June and October streamflows represent moderate to low base flows. Base flow or ground-water runoff is defined as that part of the runoff which was passed into the ground, has become ground water, and has been discharged into a stream channel as spring flow or seepage water (Langbein and Iseri, 1960). Streamflow was at 75- and 85-percent durations on June 14 and October 3 and 14, 1988, at the USGS's Piscataqua River streamflow-gaging station (0109100) (K.W. Toppin, U.S. Geological Survey, written commun., 1989), which is 14 mi north of the study area.

Ground-water recharge, indicated by losing stream reaches, occurs along most upland-draining tributaries, the western reaches of the Souhegan River, and the discharge ditch. Upland-draining tributaries that recharge the aquifer, in order of decreasing quantity of recharge, are Purgatory Brook, Tucker Brook, and tributaries 1, 2, and 5 (fig. 11). Tucker Brook and tributary 2 and 5 lost all streamflow to the aquifer during observed base-flow periods. MacNish and Randall (1982), in a study of a stratified-drift river valley in New York, similar to the Milford-Souhegan aquifer, observed that streamflow losses are greatest in upland-draining tributaries at the valley wall and decrease downstream. The Souhegan River lost 4 to 10 percent of its total streamflow between stations 31 and 22 on October 13 and June 14, 1988. Streamflow losses are attributed to a coarse-grained and permeable streambed in combination with a high stream-stage elevation relative to aquifer head. Streamflow losses are also partly attributed to large ground-water withdrawals at the Milford Fish Hatchery, which lower aquifer head in this area, and possibly from withdrawals at other nearby production wells. The discharge ditch loses water over most of its course. The elevation of stream stage in the discharge ditch

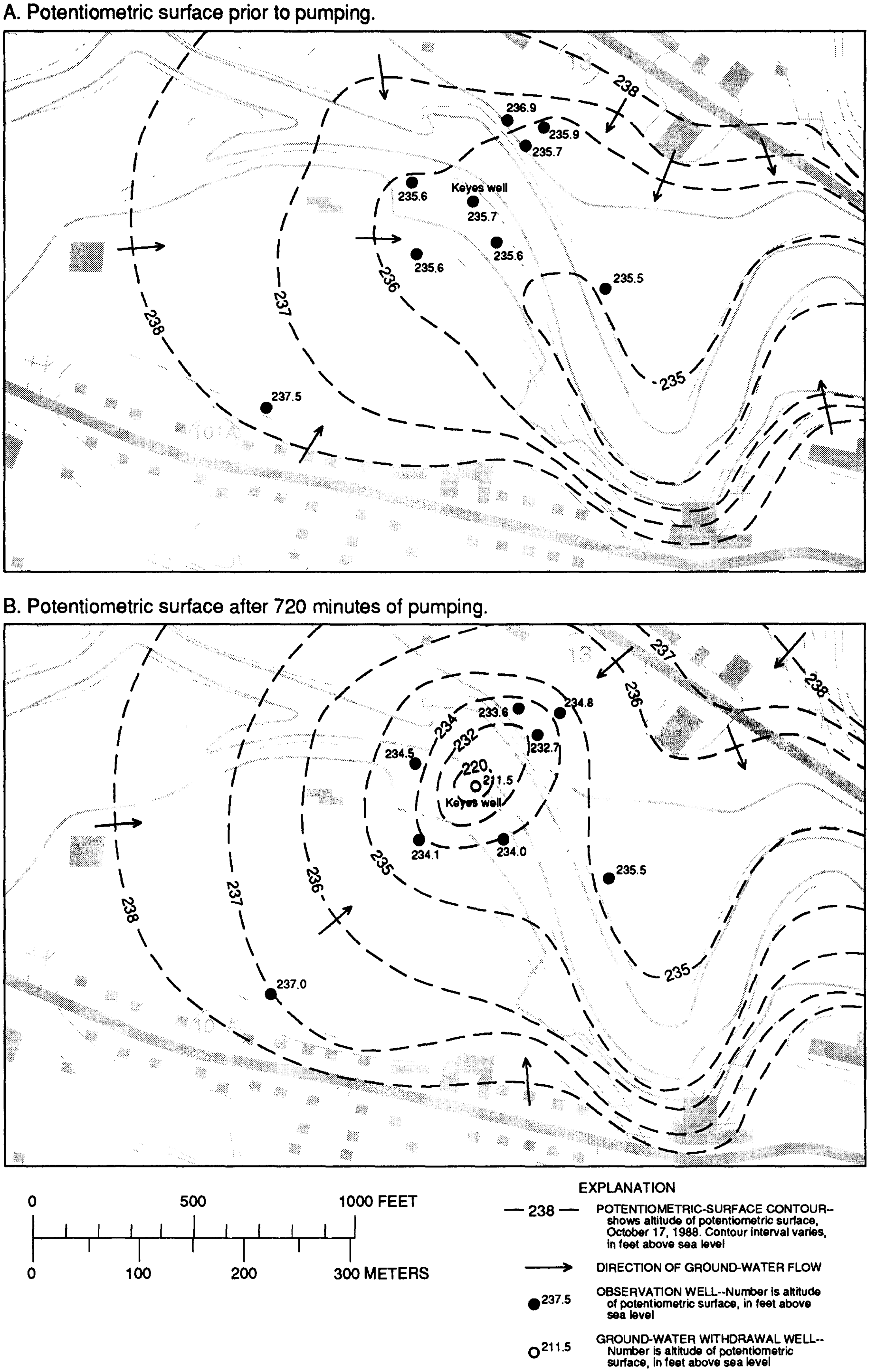


Figure 10.--Altitude of potentiometric surface and horizontal ground-water-flow directions before and during the Keyes aquifer test.

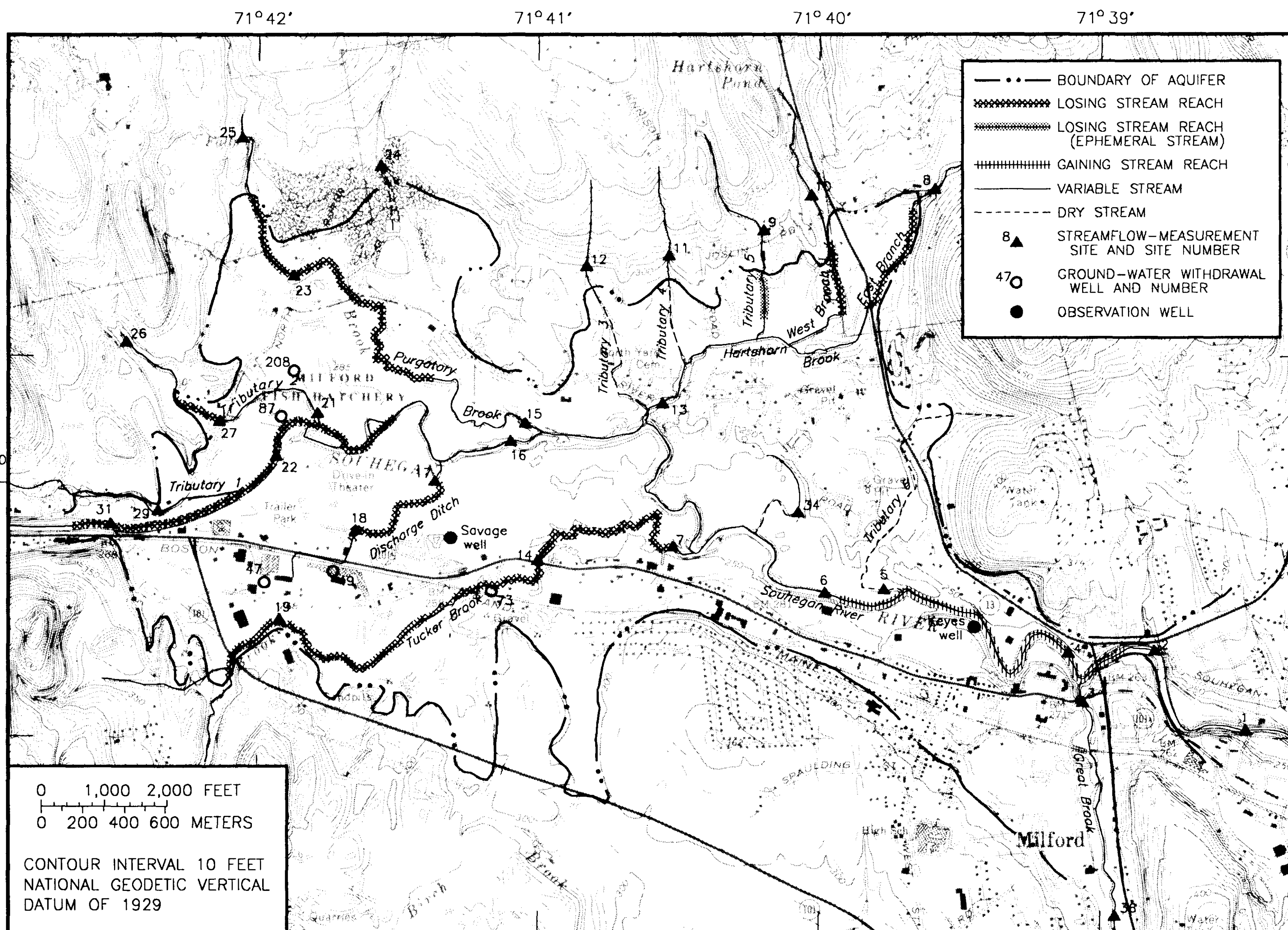
Table 4.--*Base-flow data from streamflow-measurement sites*

[Locations of measurement stations shown in figure 11; --, no data]

Measurement site	Streamflow, in cubic feet per second, on given measurement date					
	6/14/88	9/17/88	9/28/88	10/03/88	10/14/88	11/01/88
1	59.12	50.58	66.86	24.67	24.07	71.62 66.96
2	--	--	--	--	0	--
3	6.23	--	--	--	2.35	--
4	0	--	--	--	--	--
5	0	--	--	--	0	--
6	47.00	--	--	18.99	18.73	--
7	0	--	--	0	0	--
8	.23	--	--	--	.02	--
9	.03	.05	--	--	.004	--
10	.45	--	--	--	.65	--
11	0	--	--	--	0	--
12	0	--	--	--	0	--
13	.56	--	--	--	.72	--
14	.03	--	--	--	.02	--
15	5.18	--	--	--	[a]5.78 5.22	--
16	37.28	--	--	--	[b]14.91	--
17	.37	--	--	--	.52	--
18	.52	--	--	--	.85	--
19	.56	--	--	--	.47	--
21	--	0	--	--	0	--
22	35.30	--	--	13.74	14.10	--
23	2.30	--	--	--	1.24	--
24	0	--	--	--	.05	--
25	--	--	--	--	1.38	--
26	--	--	--	--	.06	--
27	0	0	--	0	0	--
29	.07	--	--	--	.08	--
30	.06	--	--	--	[b].04	--
31	39.10	31.36	46.90	14.39	[b]14.10	54.84
32	32.60	--	--	--	--	--
34	0	0	--	--	0	--

[a] Sixty-one percent of streamflow from discharge of ground-water withdrawals at Fish Hatchery.

[b] Streamflow estimated to be 60 percent of June 1988 streamflows.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 11.--Locations of streamflow-measurement sites and gaining and losing reaches of streams, June and October 1988.

is above the water table and is maintained by water discharged from industrial facilities (New Hampshire Water Supply and Pollution Control Division, 1985).

The Souhegan River is the final discharge point for most water in the Milford-Souhegan aquifer, except for consumptive water losses and evapotranspiration. Ground-water discharges to most of the eastern two-thirds of the river. The highest observed streamflow gains were between sites 6 and 1; 5.92 ft^3/s in June and 3.47 ft^3/s in October.

October and June streamflows (fig. 10) were similar with respect to stream gains and losses. The most significant difference in patterns of gains and losses was between measurement stations 16 and 6. October streamflow measurements indicate a losing reach, whereas June streamflow measurements indicate a gaining reach (fig. 11, table 4). Further low-flow measurements would be needed to quantify stream-aquifer losses and gains in this reach. In general, October base flow was approximately 60 percent less than that of June. As a result, October gains and losses were much lower than those in June.

Recharge

The Milford-Souhegan aquifer receives recharge from three principal sources: (1) direct infiltration of precipitation, (2) natural and induced infiltration from surface water, and (3) lateral inflow at the aquifer boundary from adjacent, predominantly till-covered uplands. The aquifer is also recharged from the underlying bedrock, but probably to a much lesser degree. Although data are not available to quantify the rate of recharge from the underlying bedrock, upward vertical head gradients from the bedrock to the drift at nested wells 124, 125, and 123 (Appendix C) support the hypothesis that some recharge must occur. Ground-water levels at bedrock well 124 are higher than levels at adjacent drift wells 123 and 125 (fig. 3).

Ground-water recharge varies seasonally as well as from year to year. During the growing season, May through mid-October in the northeast, most rainfall is retained in the soil to replenish soil moisture lost to evapotranspiration. Consequently, recharge occurs infrequently and in small amounts between May and October, except during an unusually wet late spring or summer. Ordinarily, most recharge occurs during the remainder of the year, from mid-October through April, except when frost, frozen soil, snow, or ice impedes or prevents infiltration.

Under natural conditions, water entering an isolated, bounded aquifer as recharge is ultimately discharged to streams or evapotranspired. The component of recharge that becomes ground-water discharge is termed effective recharge and is available for recovery and use. Effective recharge is less than total recharge because some recharge is lost to ground-water evapotranspiration. Base flows provide a useful measure of effective recharge. Observed low flows of the Souhegan River and its tributaries (table 4) are assumed to consist entirely of base flow.

Effective recharge to the Milford-Souhegan aquifer is assumed to be equal to the net gain in base flow across the aquifer. This assumption implies that changes in aquifer storage and water consumption have negligible effects on gain in net base flow. The assumption and its implications are believed to be valid because (1) ground-water levels changed only slightly during October 1988; and (2) major ground-water withdrawal centers in the aquifer, the manufacturing and concrete companies, and the Milford Fish Hatchery return virtually all withdrawn ground water back to the Souhegan River and its tributaries. In essence, the return flow consists of captured water that would have eventually discharged directly to the Souhegan River and (or) become induced surface-water infiltration.

Effective recharge to the Milford-Souhegan aquifer in October 1988, is 5.31 ft^3/s. The effective recharge was determined from October 1988 streamflow data, which is summarized in table 5. The effective recharge is calculated by subtracting all stream inflows entering the aquifer from stream outflows leaving the aquifer.

Use of streamflow measurements from October 3 and October 14, 1988, (table 4) was necessary to complete table 4. Streamflow on October 3 is highly correlated to streamflow on October 14; differences in streamflow on the two days are small. This correlation was observed at several stations on the Souhegan River and at the Piscataquog streamflow-gaging station (01091000) nearby (K.W. Toppin, U.S. Geological Survey, written commun., 1989). For example, the ratio of discharges at the Piscataquog River station to those at station 1 were similar, approximately 13 percent on October 3 (3.22 to 24.07 ft^3/s) and on October 14 (3.30 to 24.72 ft^3/s).

Ground-water-recharge estimates are given in table 6 for the principal sources of recharge to the Milford-Souhegan aquifer. The total recharge from all listed sources is based on the net gain in base flow

Table 5.--*Stream inflows and outflows in the study area, October 1988*

[ft^3/s, cubic feet per second]

Inflows			Outflows		
Stream and measurement station	Stream flow (ft^3/s)	Date of measurement	Stream and measurement station	Stream flow (ft^3/s)	Date of measurement
31 Souhegan River	14.39	10/03/88	1 Souhegan River	24.67	10/03/88
25 Purgatory Brook	1.38	10/14/88			
30 Tributary 1	[a].04	--			
26 Tributary 2	.06	10/14/88			
9 Tributary 4	0	10/14/88			
10 Hartshorn Brook 1	.65	10/14/88			
8 Hartshorn Brook 2	.02	10/14/88			
19 Tucker Brook	.47	10/14/88			
3 Great Brook	2.35	10/14/88			
Total inflows	19.36		Total outflows	24.67	

[a] Streamflow estimated to be 60 percent of June 1988 streamflow.

across the aquifer (5.31 ft^3/s), which is equivalent to a rate of 21.8 in/yr based on an aquifer area of 3.3 mi^2.

Recharge from domestic wastewater leach fields in the study area is insignificant because it was estimated based on water and sewer records to be less than 0.01 ft^3/s. Domestic-wastewater recharge includes wastewater from all nonsewered households and businesses supplied with public-supply water from outside the study area. Domestic-wastewater recharge does not include wastewater from dwellings with private wells because the amount of water returned to the aquifer approximately equals the amount of ground-water withdrawn.

Ground-water inflow from Great Brook valley contributes recharge to the aquifer, but probably accounts for less than 1 percent of total recharge (table 6). Inflow from Great Brook valley occurs at the southeastern boundary of the Milford-Souhegan aquifer (fig. 2), where saturated stratified-drift deposits in Great Brook valley are hydraulically connected with the Milford-Souhegan aquifer. Ground-water inflow was calculated by use of the Darcy equation (which assumes one-dimensional flow) and information on the cross-sectional flow area of the saturated aquifer in Great Brook valley (32,000 ft^2; approximately 1,200 ft across with an average depth of 26.7 ft), the mean horizontal hydraulic conductivity in this area (65 ft/d), and the hydraulic gradient in the valley (0.0015 ft/ft). The hydraulic gradient is assumed to be equal to the slope of Great Brook where it enters the aquifer area (fig. 2).

Recharge from till-covered uplands reaches the aquifer by streams draining the upland areas and by lateral inflow at the aquifer boundary. Runoff from till-covered uplands was measured at 14 streams near the aquifer boundary at base flow. During base flow, the upland ground-water discharge was 3.99 ft^3/s, June 14, 1988, and 2.72 ft^3/s, October 14, 1988, from adjacent till-covered uplands during base-flow measurement (table 3). The average upland ground-water discharge was 0.21 $(ft^3/s)/mi^2$ of upland area.

Lateral recharge from till-covered uplands not drained by upland streams, referred to as lateral till seepage, is assumed to be equal to the ground-water discharge factor (0.21 $(ft^3/s)/mi^2$) times the total adjacent upland area not drained by streams. The total adjacent upland area not drained by streams is approximately 3.139 mi^2. Total lateral till seepage is estimated to be 0.64 ft^3/s (table 5).

Table 6.--*Distribution of recharge to the Milford-Souhegan aquifer*

[ft^3/s, cubic feet per second; in/yr, inches per year; aquifer area is 3.3 mi^2]

Source of recharge	Amount of recharge (ft^3/s)	Amount of recharge (in/yr)	Percentage of total recharge
Ground-water inflow from Great Brook Valley	0.04	0.2	0.7
Lateral seepage from uplands	.64	2.6	12.1
Surface water	1.44	5.9	27.1
Direct infiltration of precipitation	3.19	13.1	60.1
Total	5.31	21.8	100.0

Natural and induced surface-water infiltration contributes more than 26 percent of the total estimated recharge to the aquifer (table 5). Recharge from surface water is equal to the combined streamflow losses on October 1988 from stream reaches delineated in figure 11. The western reaches of the Souhegan River account for approximately one-half of streamflow losses.

Direct infiltration of precipitation is the most significant contributor of recharge to the aquifer and amounts to 3.19 ft^3/s or 60.1 percent of the total recharge. Recharge from direct infiltration of precipitation is calculated as a residual of the base-flow gain by subtracting lateral till seepage, lateral ground-water inflow, and streamflow loss from net base-flow gain.

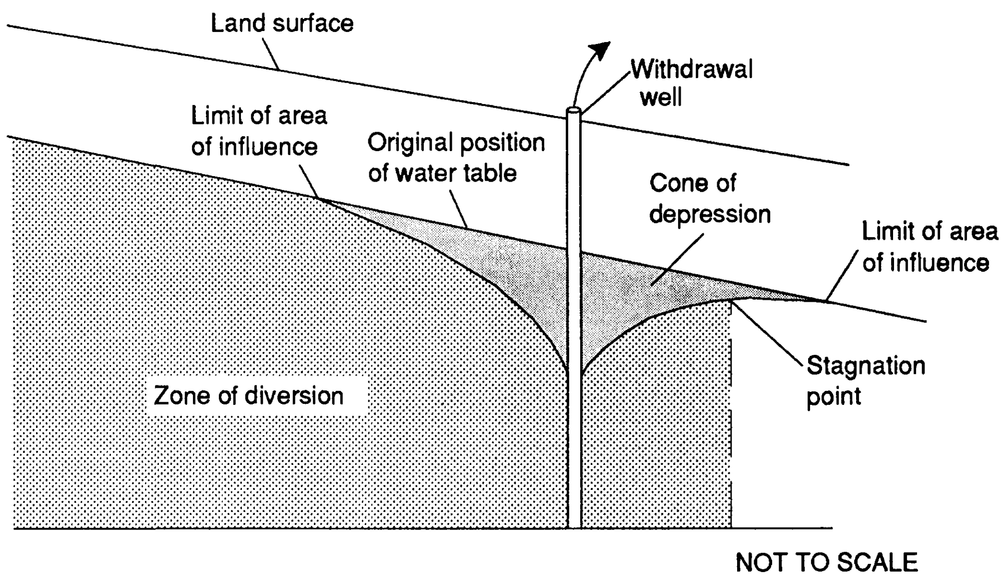
Recharge from direct infiltration of precipitation, calculated by use of the above approach, is 13.1 in/yr; 8.9 in/yr less than a maximum average annual potential rate of 22 in/yr. The maximum average annual potential recharge estimate is based on the assumption that for a typical stratified-drift, river-valley aquifer approximately one-half of the total annual precipitation is consumed by evapotranspiration; the residual component is the maximum potential recharge available from direct infiltration. This relation between precipitation and recharge for river-valley aquifers has been observed by MacNish and Randall (1982) and Lyford and Cohen (1988) at sites with a hydrogeologic setting similar to the Milford-Souhegan aquifer. Any number of environmental factors, such as rejected recharge (precipitation that can not infiltrate the ground) at wetlands may contribute to the calculated infiltration recharge being less than the maximum potential; a further discussion of this is purely speculative in nature. Average annual precipitation, as determined from long-term (1945-60) precipitation records at Milford, New Hampshire (U.S. Weather Bureau, 1964), is 44 in/yr, total precipitation for 1988 was 48.28 in. (National Oceanic and Atmospheric Administration, 1988).

SIMULATION OF GROUND-WATER FLOW

The contributing area of a pumped well is the areal extent of the zone of diversion. The zone of diversion of a pumped well is the volume of an aquifer from which ground-water flow is diverted (Morrissey, 1989). The contributing area and zone of diversion to a hypothetical pumped well is shown in figure 12. A contributing recharge area is stationary under steady-state conditions but is dynamic under transient conditions.

The area of influence of a pumped well is the areal extent of the part of the water table or potentiometric surface that is perceptibly lowered by the withdrawal of water (Meinzer, 1923, p. 61). The contributing area of a pumped well seldom corresponds to its area of influence.

A. Sectional view



B. Plan view

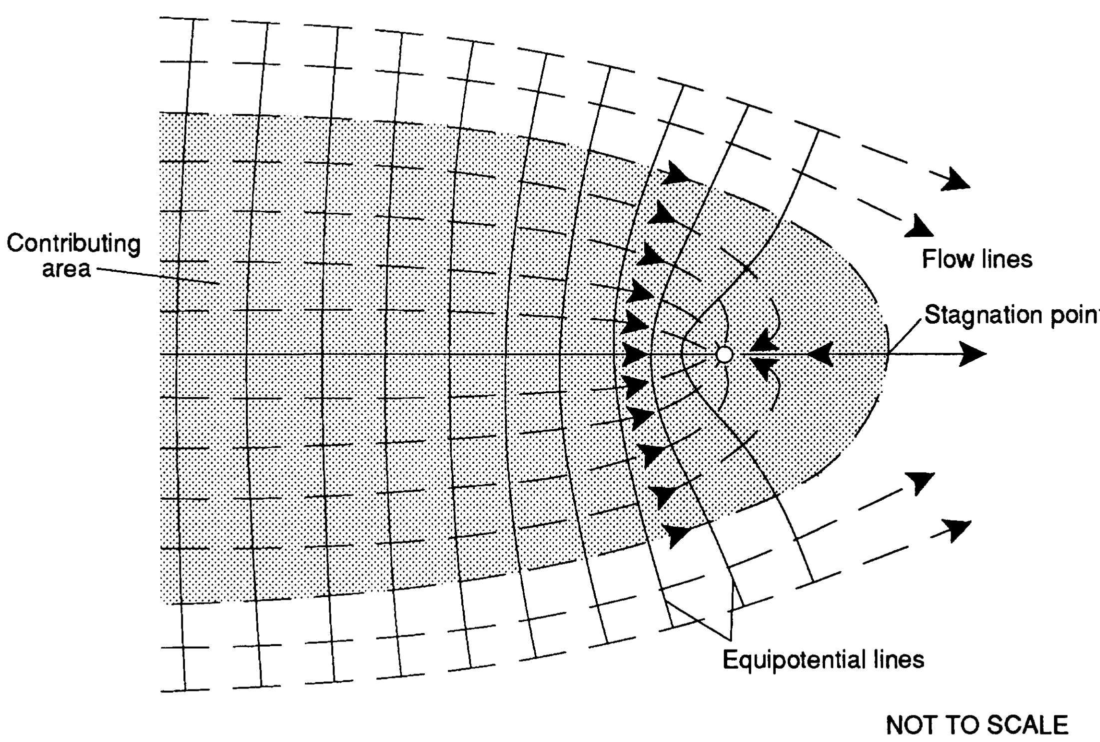


Figure 12.--A hypothetical pumped well showing a cross-sectional view of the zone of diversion and a plan view of the contributing area (From Morrissey, 1987, fig. 7).

Unlike delineating an area of influence, delineating the contributing area of a pumped well requires additional analyses beyond simply measuring drawdowns at observation wells. In certain situations, contributing areas can be effectively delineated by use of flow nets. Flow nets are often used to distinguish between water that is diverted to a pumped well and water that moves past the well to other discharge points. Unfortunately, flow-net analysis requires gross simplifications of aquifer characteristics and ground-water flow, such as the assumption that an aquifer is homogeneous and isotropic. Numerical models are superior to flow nets and analytical models in delineating contributing recharge areas because numerical models can more effectively incorporate spatial variations in aquifer properties and the effects of boundary conditions.

The following sections describe the use of a three-dimensional, numerical ground-water-flow model used in conjunction with a particle-tracking procedure to estimate contributing recharge areas of the Keyes and Savage wells for steady-state conditions. Areas contributing recharge to wells depend on discharge rates of wells, duration of pumping, hydraulic properties of the aquifer and streambeds, proximity of the well to aquifer boundaries, recharge to the aquifer, and well construction (Morrissey, 1989). Because hydraulic properties of the aquifer and streambeds are bulk approximations, contributing areas described in this report are considered estimates.

A steady-state flow model of the Milford-Souhegan aquifer was developed to simulate regional ground-water flow in the Milford-Souhegan aquifer and to simulate local flow around the Savage and Keyes wells. Simulations fall under two distinct categories. First, ground-water flow was simulated for October 1988, at which time the Savage and Keyes wells were not operating. Secondly, flow was simulated with the addition of pre-1983 ground-water withdrawals at the Savage and Keyes wells. Sensitivity of results to variation in model parameters was analyzed for both simulations. Contributing areas were delineated for simulations of pumping at the Savage and Keyes wells. Possible ranges in contributing areas were produced by adjusting, within a reasonable range of error, model parameters of aquifer hydraulic conductivity, streambed hydraulic conductivity, ground-water recharge, and ground-water withdrawals.

The model was calibrated to hydrologic conditions for October 1988. Simulated ground-water levels and simulated stream seepage were compared to ground-water levels and stream seepage measurements for October 1988. Because the number of ground-water level observations from October 1988 was limited, additional ground-water levels from other periods were incorporated into the analysis. The additional data consist primarily of summer and early fall water-level measurements. The mean water levels in wells at which multiple measurements were made are generally within 1 foot of October 1988 levels (Appendix C); thus, these mean water levels were used to help guide calibration.

Observations of the ground-water-flow system during October 1988 indicate that it was in a state of little change--that is, approximately in steady state. Measured ground-water levels at six observation wells (1, 123, 125, 142, 143, and 144; fig. 4) show an average net change of only 0.06 ft from the beginning to the end of October 1988 and a mean observed absolute fluctuation of 0.30 ft during October 1988. Two sets of October 1988 base-flow measurements (table 4) also show few differences. Most low-flow measurements in October are within 3 percent for the same station. Precipitation was uniformly distributed in October 1988. Total monthly precipitation was 2.81 in. and precipitation occurred on 11 days throughout the month. Daily precipitation totals did not exceed 0.20 in. on 9 out of 11 days. The highest daily precipitation total was 2.02 in. on October 23; ground-water levels rose less than 0.20 ft as a result of precipitation on October 23 at wells 123 and 125 (Appendix C).

Simulated recharge rates are based on recharge estimates from October 1988 streamflow data. Simulated ground-water withdrawals are based on October 1988 daily mean ground-water withdrawals. A mean daily rate, required for steady-state simulations, was determined by averaging the typical daily withdrawals over 24 hours. Simulated withdrawal wells include the two Milford Fish Hatchery wells (87 and 208) and the production wells at manufacturing and wire and cable companies (47 and 49). Small ground-water withdrawals at a concrete company (well 73) were not simulated because withdrawn water is discharged onsite and, therefore, returns to the aquifer at approximately the same point. The Savage and Keyes production wells were excluded from the initial simulation because neither well was active during October 1988, except for a brief aquifer test at the Keyes well.

Contributing areas of the Savage and Keyes wells were estimated by simulating October 1988 conditions with hypothetical ground-water withdrawals at these wells. Ground-water withdrawals at the Savage and Keyes wells were simulated with the steady-state model of October 1988 using October 1982 daily mean ground-water withdrawals (table 1). All other model parameters were kept constant. Simulated ground-water levels at the Savage and Keyes wells were compared with ground-water-level data collected during aquifer tests and during periods when the Savage and Keyes wells were being used for water supply.

Although simulation of withdrawals at the Savage and Keyes wells are for a set of hypothetical hydrologic conditions, available hydrologic and climatological data indicate simulations may approximate actual hydrologic conditions of October 1982. Hydrologic stresses including ground-water withdrawals and precipitation (and resultant ground-water recharge) were quite similar for the two periods--the exception being differences in ground-water withdrawals at the Savage and Keyes wells (table 1). Total monthly precipitation (for October 1988 and October 1982) was within 15 percent (National Oceanic and Atmospheric Administration, 1982, 1988). Ground-water levels were similar excluding effects imposed by ground-water withdrawals at the Savage and Keyes wells. Observed monthly ground-water levels at well 29 were slightly higher, 0.6 ft, in October 1988 compared to October 1982. Water-level differences could be a result of ground-water withdrawals at the Savage well in 1982. Streamflow also appears to be quite similar for October 1988 and October 1982. The daily mean discharge was 2.49 ft^3/s and 2.83 ft^3/s at the nearby Stony Brook streamflow-gaging station 01093800 (F.E. Blackey, U.S. Geological Survey, written commun., 1989). These observations indicate that hypothetical simulations, involving simulated withdrawals of the Savage and Keyes wells, approximate actual hydrologic conditions of October 1982, when these wells were active.

The advection model, used to delineate contributing recharge areas to pumped wells, is a semianalytical, particle-tracking procedure (Pollock, 1989). The advection model is a postprocessor for steady-state output from the ground-water-flow model by McDonald and Harbaugh (1988). The particle-tracking method is based on the assumption that each directional-velocity vector varies linearly within a grid cell in its own coordinate direction and that it is constant with respect to other coordinate directions (Pollock, 1989). Given the initial position of a particle, the particle's flow path at any time can be calculated throughout the model grid by computing directional-velocity vectors and multiplying by a time step. Contributing areas of wells were estimated by forward and backward tracking of particles to and from pumped wells to areas of recharge, such as the water table and streams.

Model Construction and Initial Data Input

A program for a block-centered, finite-difference ground-water-flow model (McDonald and Harbaugh, 1988) was used to simulate steady-state ground-water flow in three dimensions in the Milford-Souhegan stratified-drift aquifer. The program consists of independent subroutines that simulate ground-water flow, ground-water/surface-water interaction, recharge, evapotranspiration, several types of boundary conditions, and pumping stresses. Discrete layers within an aquifer can be simulated as unconfined, confined, or convertible from confined to unconfined. Simulated ground-water flow is horizontal within the model layers representing the aquifer; vertical flow occurs between layers. This is an inherent limitation in the computer model, and the effect of this limitation is to permit only two-dimensional flow within each aquifer layer and one-dimensional flow between layers.

Simulated heads were computed using an iterative solver called the strongly implicit procedure (McDonald and Harbaugh, 1988). Heads were computed during successive iterations until they satisfied the head closure criteria of 0.001 ft. A small head closure criteria is needed to ensure that, in addition to head changes, cell fluxes are also small.

Construction of a ground-water-flow model of an aquifer entails compiling certain geologic and hydrologic data into arrays for use in the numerical program. The finite-difference model is discretized--that is, mathematically divided--into horizontal and vertical cells. Aquifer properties are assigned to each cell and represent an integrated value over the cell area. In addition to aquifer properties, streambed conductance, stream stage, and hydrologic stresses including recharge and discharge are also assigned to appropriate cells.

A ground-water-flow model was designed and constructed with consideration for the objectives of the investigation. The Savage and Keyes wells are of concern to this investigation; therefore, the model was designed to provide greater detail near these wells. A relatively fine horizontal grid and sufficient vertical discretization were used in these areas to enable detailed simulation of ground-water flow near the two wells.

Grid Design

The model grid, shown in figure 13, is aligned parallel to the axis of the Souhegan River valley and the general trend of the Souhegan River. The grid is composed of 76 rows and 122 columns, creating 9,272 cells per layer. The active area of the uppermost layer (layer 1) includes 5,636 cells that encompass 2.58 mi^2. The 5-layer model totals 46,360 cells, of which 15,196 cells are active.

Horizontal discretization

The horizontal dimensions of grid cells range from 50 to 200 ft along rows and columns. The grid is fine around the Savage and Keyes wells where cell sizes are 50 by 50 ft. A gradual change in cell size ensures numerical stability (Trescott and others, 1976) and is also necessary to smooth intercell head changes and flow paths between the fine- and coarse-grid areas. Cell sizes are at least 75 percent of the size of the adjacent larger cell. Cell sizes range from coarse to fine--200, 150, 110, 80, 60, and 50 ft.

Vertical discretization

The model is vertically discretized into a maximum of five layers representing the stratified-drift aquifer, each approximately 20 ft thick, to simulate vertical ground-water flow. Layer 1 is simulated with the unconfined option of the model, layer 2 is simulated as convertible, either unconfined or confined, and layers 3, 4, and 5 are simulated as confined. Simulation of vertical flow is important for describing hydrologic conditions near pumped wells. Thicknesses and areal extent of individual layers differ because model layers thin to extinction at aquifer boundaries and because the horizontal extent of the aquifer decreases with depth. For the lower layers, active cells are clustered in bedrock lows, where the saturated thickness is greatest. The total number of model layers and thickness of an individual model layer at a given point depend on the total saturated thickness of the aquifer as determined from well and test-hole data and from estimates of bedrock depth at the valley wall. For cells along the valley edge, where a cell was determined to have a saturated thickness of less than 5 ft, generally the lowest active cell, the saturated thickness was added to the cell in the layer above to avoid creating a very thin layer. Thin cells typically go dry during head convergence of iterative solution techniques (such as those used in numerical models) because of oscillations in values of computed head beyond cell dimensions. Cells dewater when computed heads fall below the altitude of the cell bottom. The numerical model excludes dry cells from subsequent iterations and final simulation results can be inaccurate.

Diagrams representing the vertical discretization for model columns 41 and 101 are shown in figure 14. These columns include the locations of Savage and Keyes wells. The lowermost layers are limited in thickness by the residual thickness remaining from the discretization of the upper layers.

The slope of the model layers is parallel to the general slope of the water table. The average lateral model-layer slope is 0.0100 ft/ft along columns and is 0.0024 ft/ft along rows.

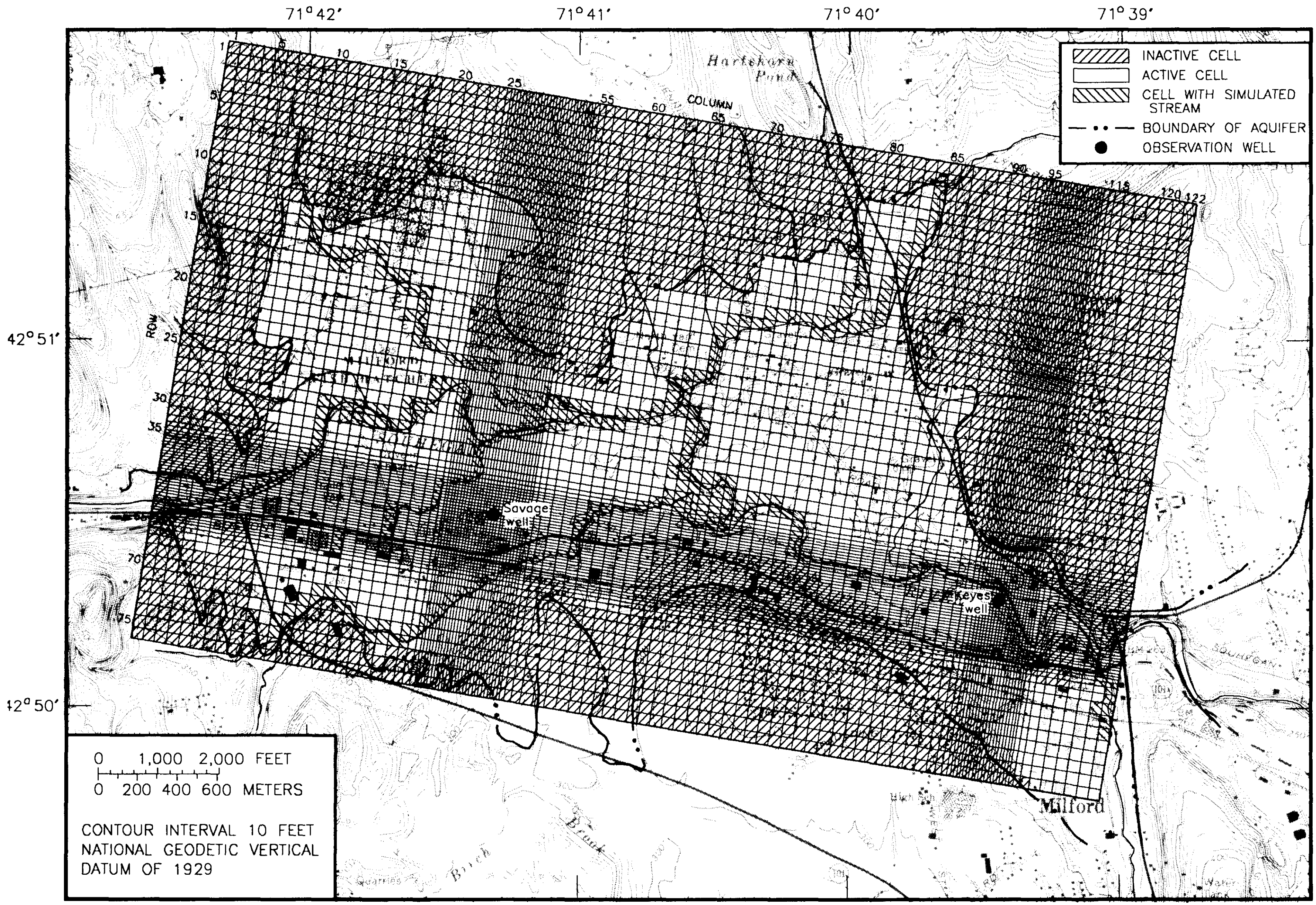
Hydraulic Conductivity

Hydraulic conductivity in the Milford-Souhegan aquifer is isotropic and heterogenous in the horizontal direction. It is anisotropic and heterogenous in the vertical direction with respect to the horizontal. Horizontal hydraulic conductivity is differentiated into three to six zones per model layer; within each zone, horizontal hydraulic conductivity is considered to be homogeneous and isotropic.

The horizontal hydraulic-conductivity zones for each model layer are shown in figures 15 through 19. Zones are numbered according to model layer; for example, there are five horizontal hydraulic-conductivity zones in layer 1 (labeled 1-1 through 1-5) and six zones in layer 2 (labeled 2-1 through 2-6). Some zones in a layer have similar hydraulic conductivities but are delineated as separate zones because they are in different parts of the aquifer.

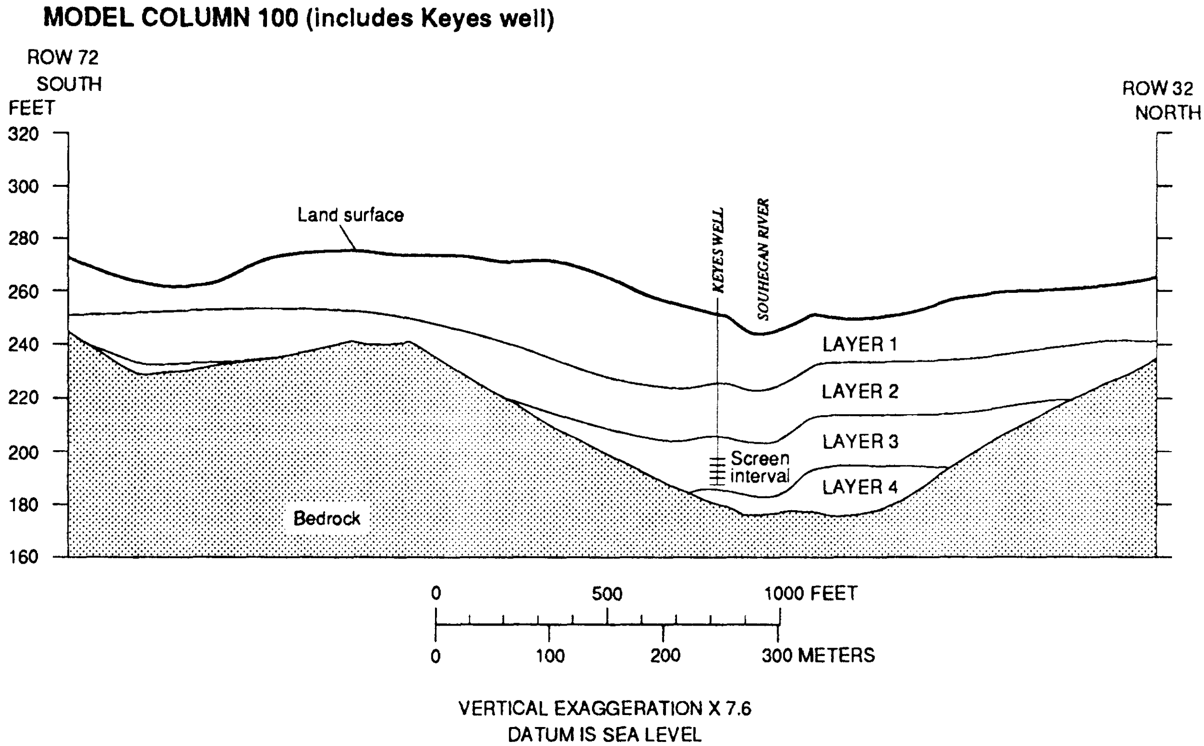
Horizontal hydraulic conductivity was delineated into zones within each layer by examining the distribution of values of hydraulic conductivity and grouping similar values into zones of equal hydraulic conductivity. Zonal hydraulic conductivities for each model layer were determined by averaging hydraulic conductivities computed from stratigraphic logs of test holes that penetrate the layer. Hydraulic conductivities average 45 to 210 ft/d for zones with predominantly stratified-drift deposits, 5 ft/d for zones with sandy till, and 1 ft/d for clayey till.

Hydraulic conductivities representative of stratified-drift material in southern New Hampshire



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 13.--Model grid, inactive and active cells, and simulated streams.



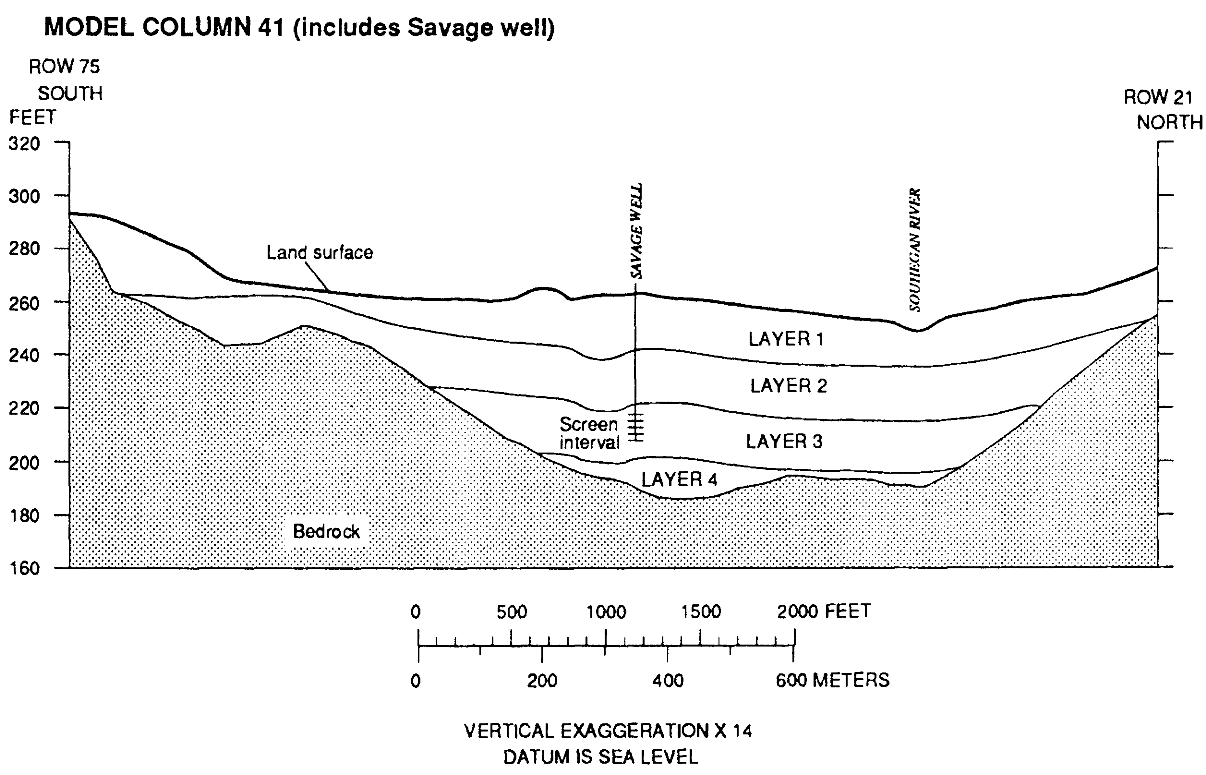
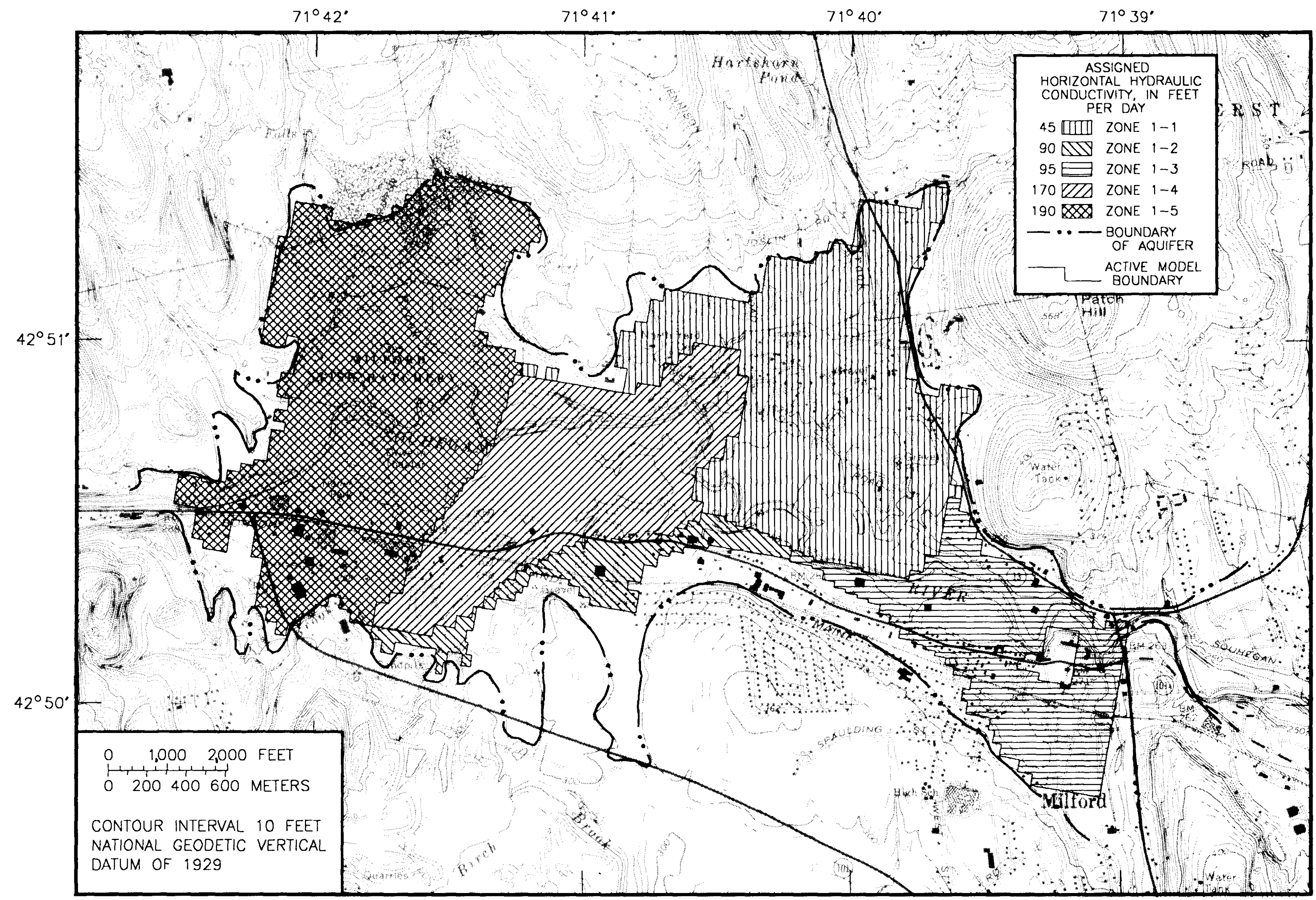
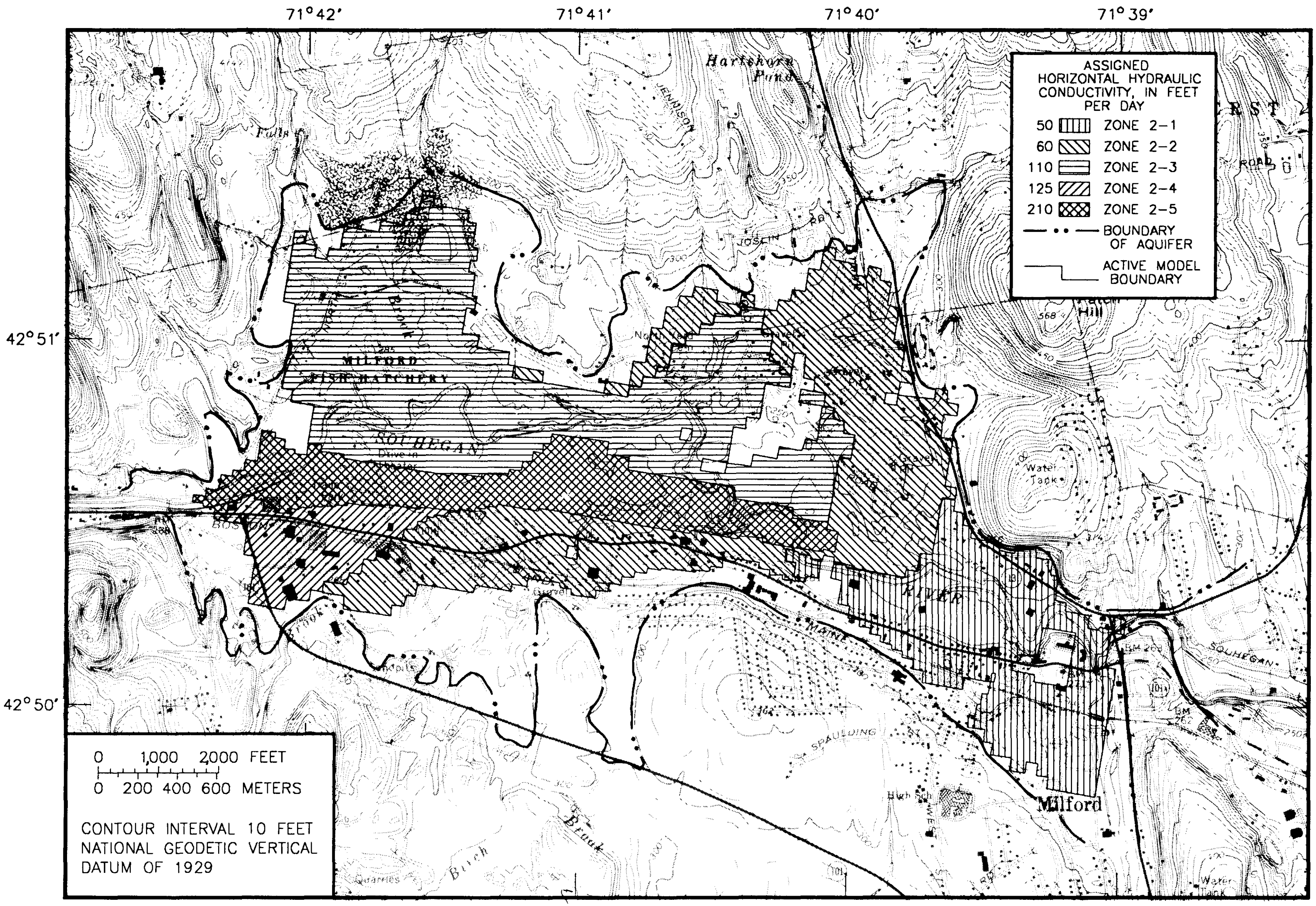


Figure 14.--Vertical discretization along model columns 41 and 100.



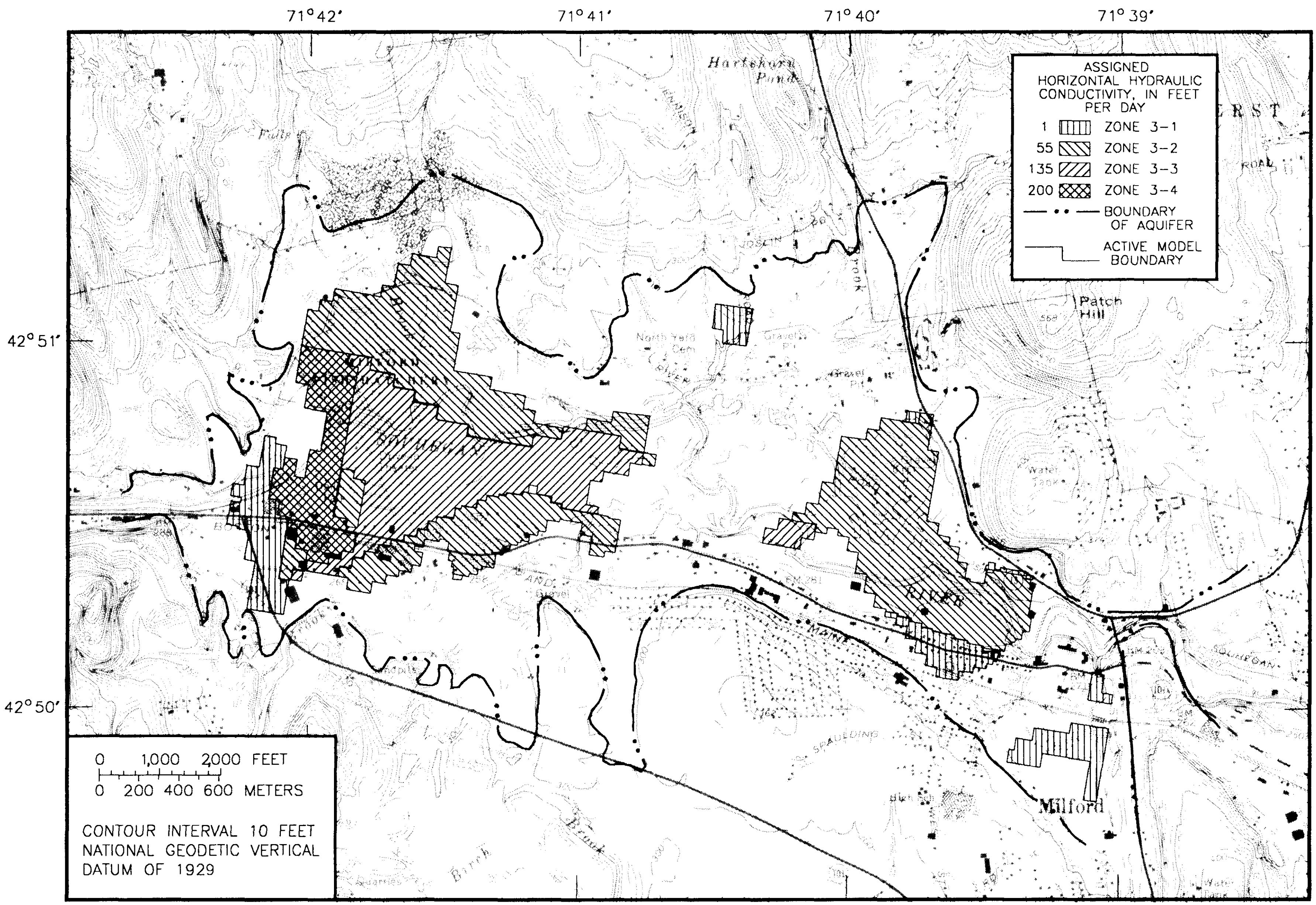
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 15.--Zones of horizontal hydraulic conductivity of the Milford-Souhegan aquifer as used in the model: Layer 1.



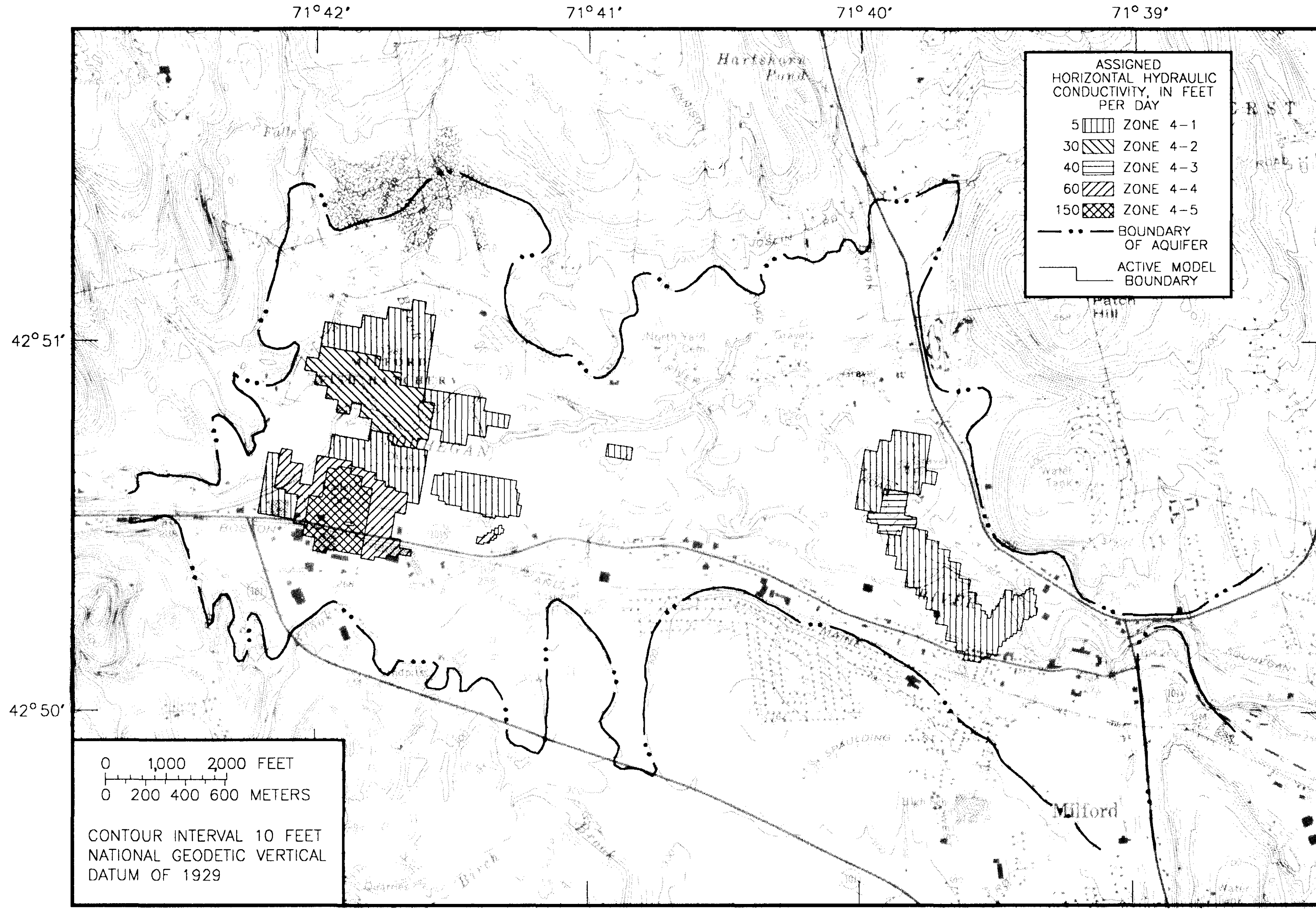
Bose from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 16.--Zones of horizontal hydraulic conductivity of the Milford-Souhegan aquifer as used in the model: Layer 2.



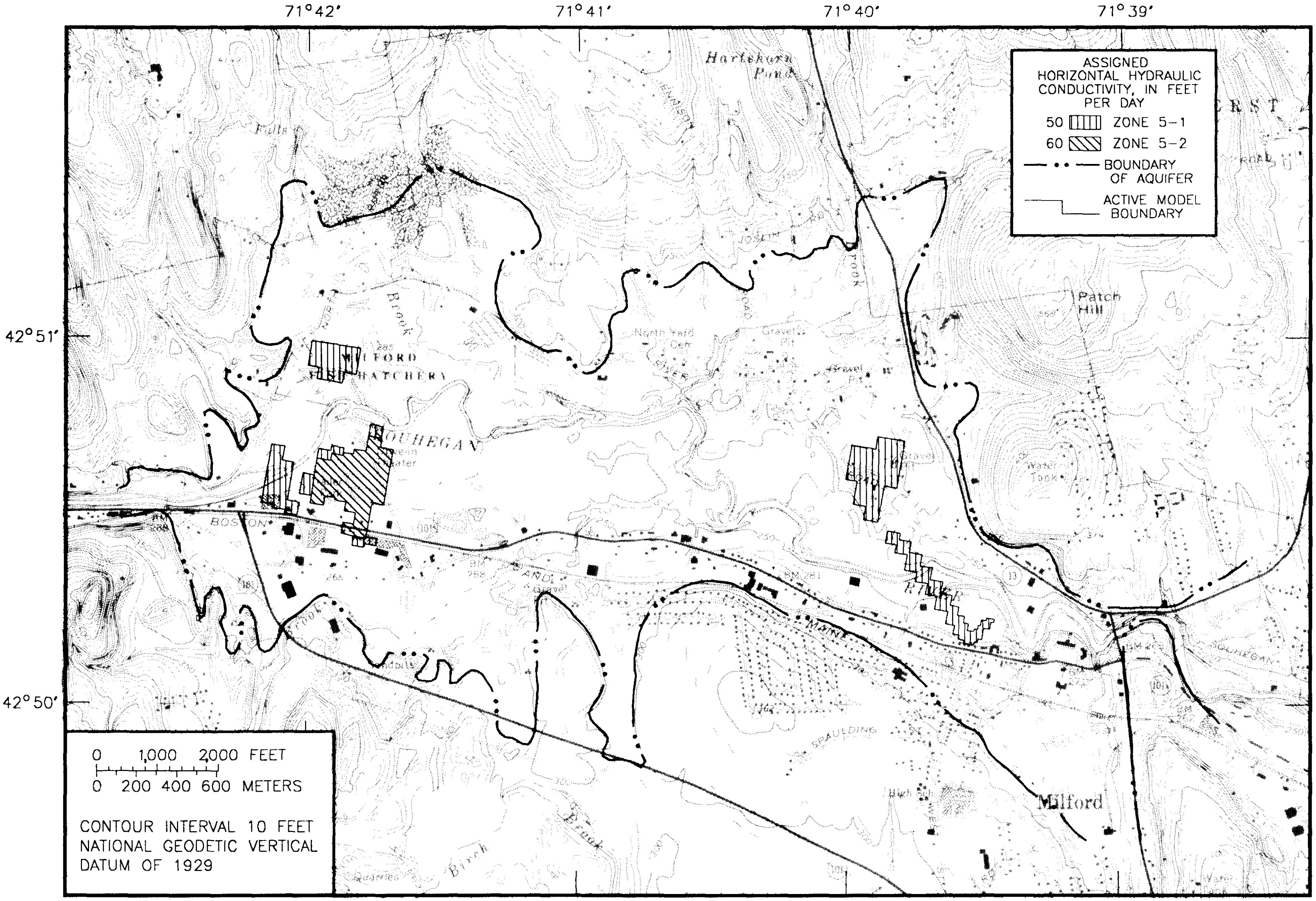
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 17.--Zones of horizontal hydraulic conductivity of the Milford-Souhegan aquifer as used in the model: Layer 3.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 18.--Zones of horizontal hydraulic conductivity of the Milford-Souhegan aquifer as used in the model: Layer 4.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 19.--Zones of horizontal hydraulic conductivity of the Milford-Souhegan aquifer as used in the model: Layer 5.

are listed in table 7; this table was used to estimate hydraulic conductivity for sediments described in well logs. Reported values for a given sediment size were determined from an empirical relation developed by Olney (1983) and grain-size observations on 175 stratified-drift sediment samples from southern New Hampshire (T.J. Mack, R.B. Moore, and P.J. Stekl, U.S. Geological Survey, written commun., 1989). The observations of grain size and hydraulic conductivity were used to develop a summary of average hydraulic conductivity compared to predominant grain size and type of stratified drift (table 7).

Horizontal hydraulic-conductivity estimates from lithologic log data are comparable to hydraulic-conductivity estimates from hydraulic tests in the Milford-Souhegan aquifer (table 8). Estimates from logs include hydraulic conductivities determined for sediments along the entire logged interval and for sediments adjacent to the screened interval. The former represents an average computed hydraulic conductivity for the aquifer at that site. Hydraulic conductivities inferred from sediments are generally within an order of magnitude of estimates from hydraulic tests. Different methods of estimation result in hydraulic conductivities that agree for the Savage and Keyes wells. These results suggest that the use of well logs to assign initial hydraulic conductivities to the model is appropriate. The use of lithologic log descriptions to compute initial hydraulic conductivity for the model allowed for finer discretization of hydraulic conductivity than would have been possible using hydraulic conductivity estimated from hydraulic tests alone.

The aquifer test at the Keyes well in October 1988 provided the only data sufficient for determining vertical hydraulic conductivity of the aquifer. Vertical hydraulic conductivity was found to be approximately one tenth the horizontal hydraulic conductivity. Initial vertical hydraulic conductivity in the model was set to one-tenth the horizontal hydraulic conductivity assigned to each zone in each layer.

Boundary Conditions

The upper model boundary, the water table, is a specified-flux boundary. Specified fluxes are assigned to layer 1 to represent recharge from precipitation and lateral inflow of water from adjacent till and bedrock-covered uplands outside the active model area. Recharge from adjacent upland areas is assigned to the outermost active cells in layer 1.

The lower model boundary represents the top of the underlying crystalline-bedrock surface. This boundary, simulated as a no-flow boundary, underlies all of layer 5 and those parts of layers 4 through 1 not underlain by active cells of another layer. Although ground water flows between bedrock and the

Table 7.--*Average hydraulic conductivity estimated from predominant grain size of stratified-drift sediment*

Stratified-drift sediment	Predominant grain-size range (millimeters)	Estimated average hydraulic conductivity (foot per day)
Sand		
Very fine	Above 3	3 or less
Fine	2 to 3	10
Medium	1 to 2	30
Coarse	0 to 1	130
Very coarse	-1 to 0	190
Gravel		
Fine	-1 to -2	250
Coarse	below -2	300 or greater

Table 8--*Comparison of horizontal-hydraulic-conductivity estimates from well-log descriptions with values from aquifer tests for selected wells in the Milford-Souhegan aquifer*

[ft, feet; ft/d, feet per day]

Local well name	Well number	Screen interval (ft)	Date of test	Hydraulic conductivity from aquifer tests (ft/d)	Method of analysis[1]	Hydraulic conductivity, (from well logs) Entire section (ft/d)	Screened interval (ft/d)
Keyes 2D	2	54 - 56	10/88	20	1	41	1
Potter 1D	132	55 - 57	do.	10	1	37	1
Ford 1	139	35 - 50	9/68	110	2	46	66
FH-5	208	50 - 65	3/85	970	3	230	400
MI-28	43	35 - 55	8/83	39	4	84	14
MI-29	171	31 - 51	do.	13	4	150	43
MI-31	45	36 - 54	do.	6	4	760	38
Savage	128	42 - 52	3/57	120	5	120	70
RFW-1	14	8 - 28	11/86	1	6	86	84
RFW-2	15	10 - 35	do.	12	6	78	72
RFW-3	16	13 - 43	do.	9	6	78	69

[1] Method of test analysis and source of data:

1. Aquifer test--Walton (Kruseman and deRidder, 1983, p. 81).
2. Aquifer test--Jacob (Kruseman and deRidder, 1983, p. 63).
3. Single-well pumping test--Meyer (Meyer, 1963, p. 83).
4. Slug tests reported values from New Hampshire Water Supply and Pollution Division (1985).
5. Walton reported values from New Hampshire Water Supply and Pollution Division (1985).
6. Single-well recovery test (Weston, 1987).

Milford-Souhegan aquifer, little is known about the magnitude of those flows. The net recharge or discharge of any such interaction is implicitly incorporated into ground-water recharge estimates to the aquifer because all recharge to the aquifer ultimately discharges to the Souhegan River.

Outermost active cells at the western and eastern ends of the valley are not adjacent to upland areas; however, they were treated as no-flow boundaries. The model boundary cuts across stratified-drift aquifer material at both ends of the valley, along the course of the Souhegan River; however, the saturated thickness at these locations is 10 ft or less. Ground-water flow through stratified-drift material at these no-flow boundaries is assumed to be negligible because the saturated aquifer material is thin and the cross-sectional area across which flow could occur is small.

The model boundary adjacent to Great Brook valley, at the southeastern model boundary, is a specified-flux boundary. Ground-water flow from Great Brook valley was simulated by assigning a specified flux to the six outermost active cells adjacent to this valley.

Perennial streams were simulated as head-dependent flux boundaries. Ephemeral streams, tributaries 2 and 5 (fig. 11), were simulated as specified-flux boundaries. Tributaries 1, 3, 4, and 6 were not simulated because of little or no streamflow. Streams simulated as head-dependent flux boundaries largely control the position of the water table because stream stage represents the lowest head in the aquifer, with the exception of heads near pumped wells. Simulated perennial streams are the Souhegan River, the discharge ditch, Great Brook, and the upland-draining tributaries;

Purgatory Brook, Tucker Brook, and Hartshorn Brook (fig. 11).

Recharge

Recharge was applied to the upper most active cells (fig. 13) to simulate infiltration of direct precipitation onto the aquifer, lateral inflow of water from adjacent uplands, infiltration of surface waters from ephemeral streams, and lateral inflow from Great Brook valley. The total recharge applied to the model from these sources was 3.92 ft^3/s. Recharge from induced and natural infiltration of perennial streams (Purgatory, Hartshorn, Tucker Brooks, and the discharge ditch) was not applied as recharge but was accounted for in river simulations.

Lateral inflow of water to the aquifer from upland areas was simulated by specifying increased recharge rates to the outermost active cells. Recharge from upland areas, termed lateral inflow, was nonuniformly distributed to outermost active cells on the basis of drainage area of uplands adjacent to each cell. Recharge applied to these cells was determined by multiplying each drainage area by the ground-water-discharge factor of 0.205 $(ft^3/s)/mi^2$ from upland areas not drained by streams. Lateral inflow was not applied to outermost active cells associated with upland areas drained by Purgatory and Hartshorn Brooks and tributaries 2 and 5 (fig. 11).

Recharge was applied to cells in contact with tributaries 2 and 5 to simulate surface-water infiltration to the aquifer. Simulated recharge was equivalent to observed streamflow losses of 6.0×10^{-5} $(ft^3/s)/ft$ along tributary 2 and 2.0×10^{-3} $(ft^3/s)/ft$ along tributary 4.

Lateral ground-water inflow from Great Brook valley, a stratified-drift aquifer outside the model boundary, is simulated in the model by specifying increased recharge rates to the six cells adjacent to this aquifer area. A total recharge rate of 0.04 ft^3/s, calculated from the estimated hydraulic gradient, cross-sectional area, and horizontal hydraulic conductivity of the aquifer, was apportioned uniformly among these cells.

Stream-Aquifer Interaction

Flow between the perennial streams and aquifer is simulated as a function of head gradient and streambed conductance by use of the river package of the ground-water-flow model (McDonald and Harbaugh, 1988). Streambed conductance is calculated as the product of the hydraulic conductivity, width, and length of the streambed within the cell, divided by streambed thickness. For each stream cell, streambed conductance, altitude of the streambed, and stage are entered into the model.

Streambed hydraulic conductivity was assigned an initial estimate of 3 ft/d for all stream cells. For cells simulating the discharge ditch, streambed hydraulic conductivity was set at 1 ft/d because this streambed appeared to contain fine sediment and organic material. During model calibration, streambed conductances were varied to make simulated stream seepage match measured stream seepage as closely as possible.

Stream-stage elevations were measured at surveyed stream-stage measurement points and interpolated from altitudes taken from USGS topographic maps. Stream depths and widths were based on measurements made at streamflow-measurement stations (fig. 11); between stations, values were interpolated. Streambed thicknesses were inferred from observations of channel geometry and typical streambed thicknesses in drift-filled river valleys (D.J. Morrissey, U.S. Geological Survey, written commun., 1989). Streambed thickness of tributaries was assumed to be 1.5 ft. The streambed thickness of the Souhegan River was assumed to be 3 ft for the western part of the river grading to 5 ft for the eastern part of the river where water velocities are slower and fine-grained bottom sediment has accumulated.

Model Calibration

The ground-water-flow model was calibrated to ground-water-flow conditions in the Milford-Souhegan aquifer in October 1988. Although the water levels in 1988 were high, the October water level was near the annual low for 1988. However, the water levels remain nearly constant from September to October 1988 and it is reasonable to assume a steady-state condition comparable to long-term average hydrologic conditions existed during October 1988. Simulated ground-water-flow rates are probably at or near average annual rates.

Model calibration involved adjusting model parameters--such as recharge rates, hydraulic conductivity of the aquifer, and streambed characteristics--from their initial values to reduce the difference between computed and measured heads and stream seepage. The final calibrated model estimates of streambed hydraulic conductivity, stream stage, and horizontal hydraulic conductivity differ from their initial values; recharge was kept the same. Most model parameters are lumped terms and are only approximately known. The parameters for this model were altered, one at a time, within realistic ranges to improve the simulations. The point at which calibration is achieved is somewhat arbitrary; for this model, differences of 3 ft between simulated and measured heads were considered acceptable.

Simulated heads for October 1988 are shown in figures 20 through 24 for the five model layers. In general, the simulated water table for layer 1 (fig. 20) compares well with the interpreted water table based on measurements during October 1988 (figs. 5 and 20). An exception is near the production wells at the Milford Fish Hatchery, where simulated drawdowns desaturate layer 1 and a comparison cannot be made. The measured water levels compare well with the simulated heads for layer 2 in this area (fig. 21). Simulated heads in layer one differ the most from the measured, water levels at the aquifer boundary near Hartshorn Brook (figs. 2 and 20) where the water table is based on stream-stage elevations. Simulation indicates stream stages are probably at a higher altitude than the regional water table. The configuration of the simulated water table also agrees with the measured water table relative to gaining and losing stream reaches (fig. 11). For example, simulated losing reaches, indicated by head contours that bend upstream, correspond to measured losing reaches for the western reaches of the Souhegan River and for tributaries near the valley wall. Simulated gaining reaches, indicated by head contours that bend downstream, correspond to measured gaining reaches.

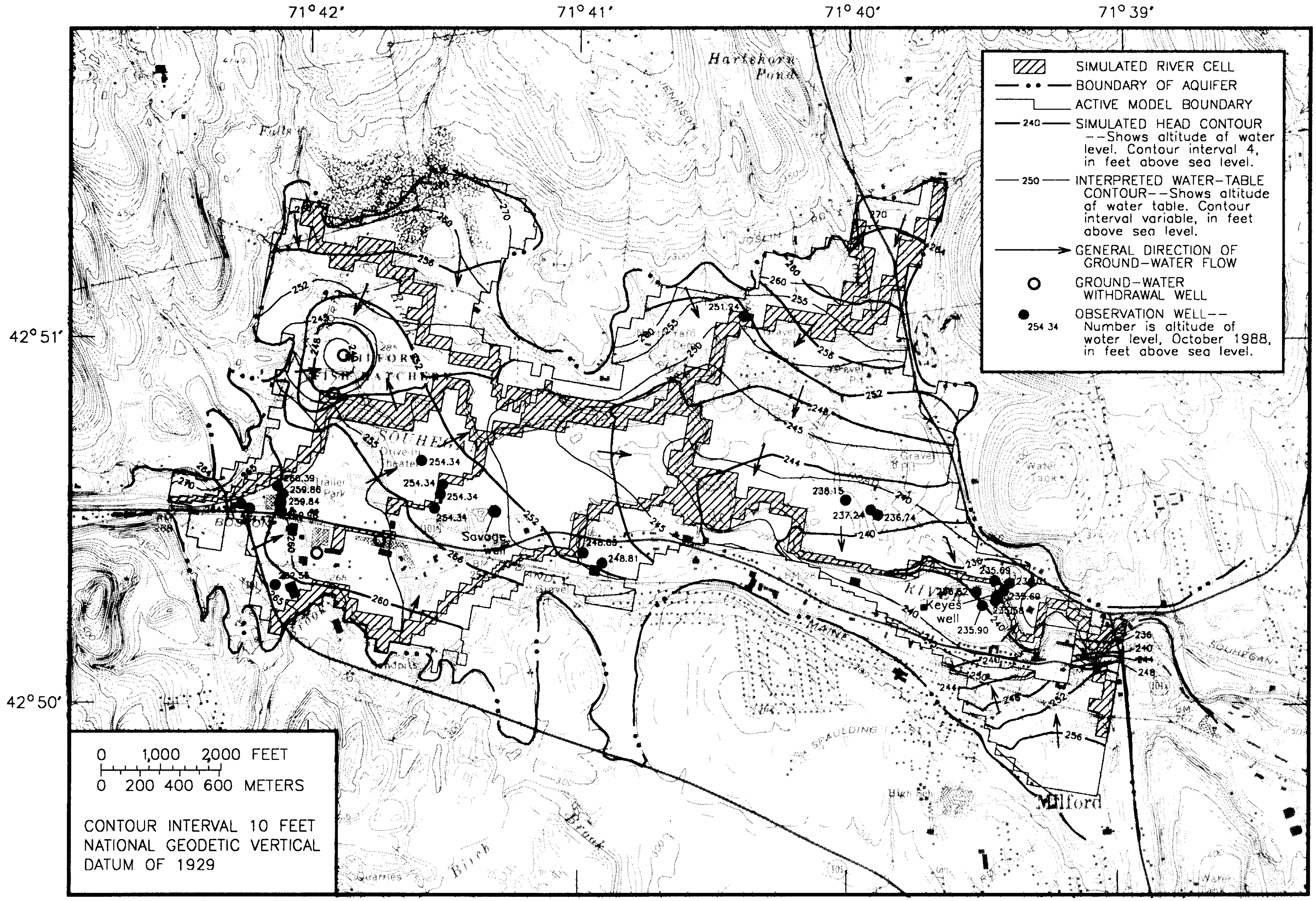
Model calibration was quantified by comparing simulated heads with 41 heads measured in October 1988 (table 9). Observed mean heads from water-level measurements at 116 locations (Appendix C) were also compared to simulated heads to expand the number of locations where head comparisons could be made (table 9). These data, herein called the average heads, were generally within 2 ft of heads measured in October 1988. The term average heads, used in this report, represents the average of the observed heads and is not a long-term mean head value.

Measured heads (October 1988 and average head) were compared individually to simulated heads at the closest model node in the layer representing the screened interval (table 9). Because locations of a well may not coincide closely with the center of the corresponding model cell, some error is introduced into the comparison. Generally, this error is negligible (less than 0.2 ft); however, the error may be considerable at the largest cells in the grid (where the distance between the well and the center of the cell can be as much as 140 ft) and at cells in which pumping is simulated (where the head may change substantially within a cell).

Differences between simulated and measured heads were compared by hydraulic-conductivity zone (figs. 15-19) and model layer. Two statistical means were used to compare the difference between simulated and measured heads within each zone. An absolute mean was used to examine the total error inherent in the head simulation, and a standard mean was used to show the fit of the head simulation in the zone. For example, a standard mean head difference shows whether the simulated heads are, on the average, greater than measured heads (positive difference) or less than measured heads (negative difference). Random error in the head simulation, even if large, would result in a standard mean that is near zero; however, the absolute mean difference would reflect random error.

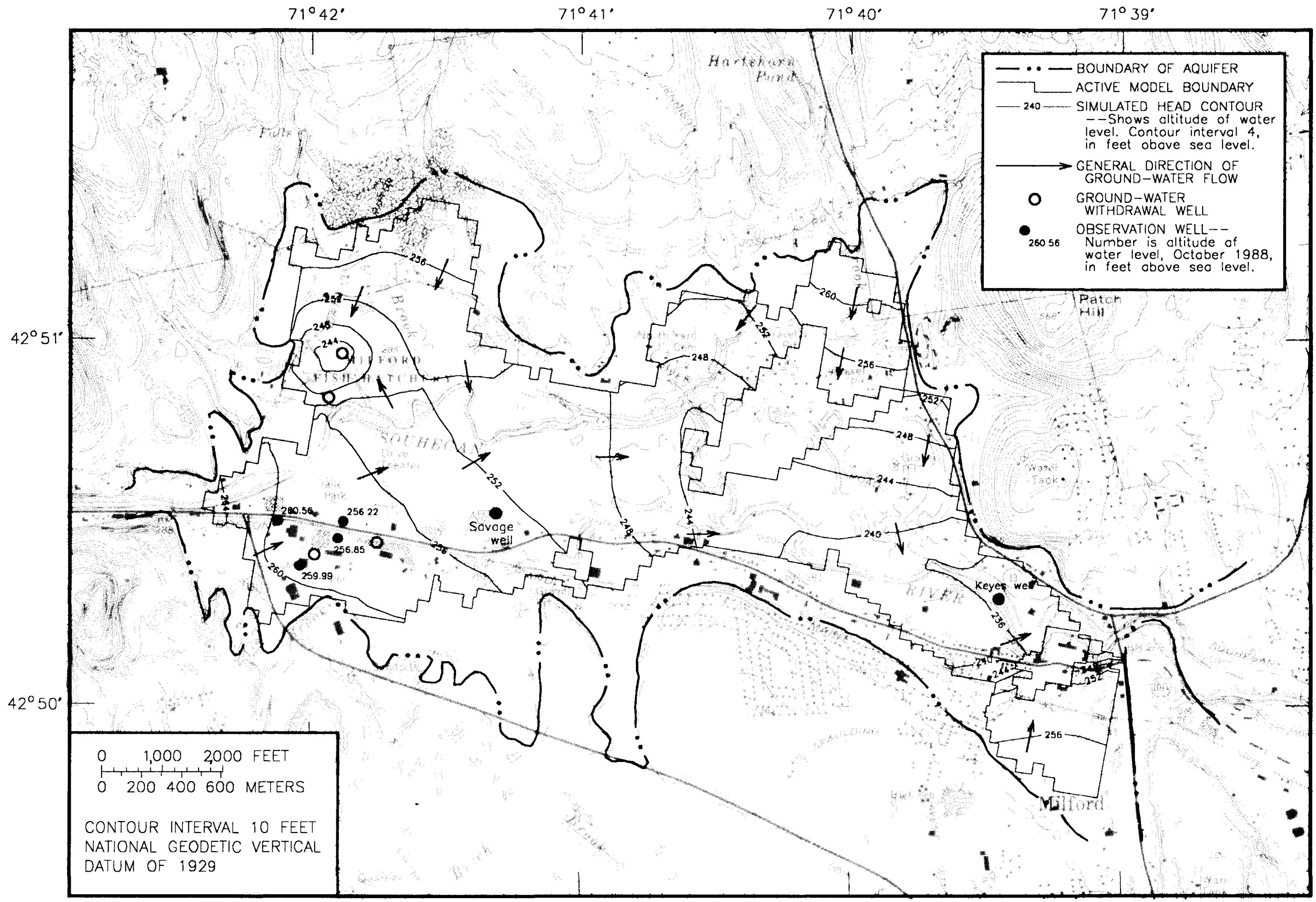
Simulated heads are generally within 3 ft of measured heads for October 1988 except at 8 of 42 model cells (table 9). Most wells where head differences are greater than 4 ft are near the river in hydraulic-conductivity zone 1-1. Measured and computed heads compare favorably for the cell corresponding to the inactive Keyes well (126) (table 9). Measured-head data for October 1988 were not available for the Savage well (128). The largest discrepancies (7.93 and 7.11 ft) are between the simulated heads for layer 3 and observed head at Fish Hatchery wells--pumped well 208 and observation well 84. Simulated pumpage of well 208 was equally divided between model layers 3 and 4; this procedure creates some difficulty in allowing a direct comparison between the simulated head in layer 3 and the measured head. The average simulated head for layers 3 and 4 at well 208 is within 3 ft of the measured head.

Simulated heads are generally within 3 ft of average heads except at 37 of the 116 model cells. Most large differences in head, greater than 5 ft, were (1) at cells at which average heads were calculated from fewer than four water-level measurements, (2) close to the model boundary or in areas where the grid scale is coarse, or (3) at cells where corresponding well data are questionable. At wells with few measurements, average heads may not reflect hydrologic conditions during October 1988. Comparisons of simulated and average heads guided calibration of the model if no other data were available. In general, differences between simulated and



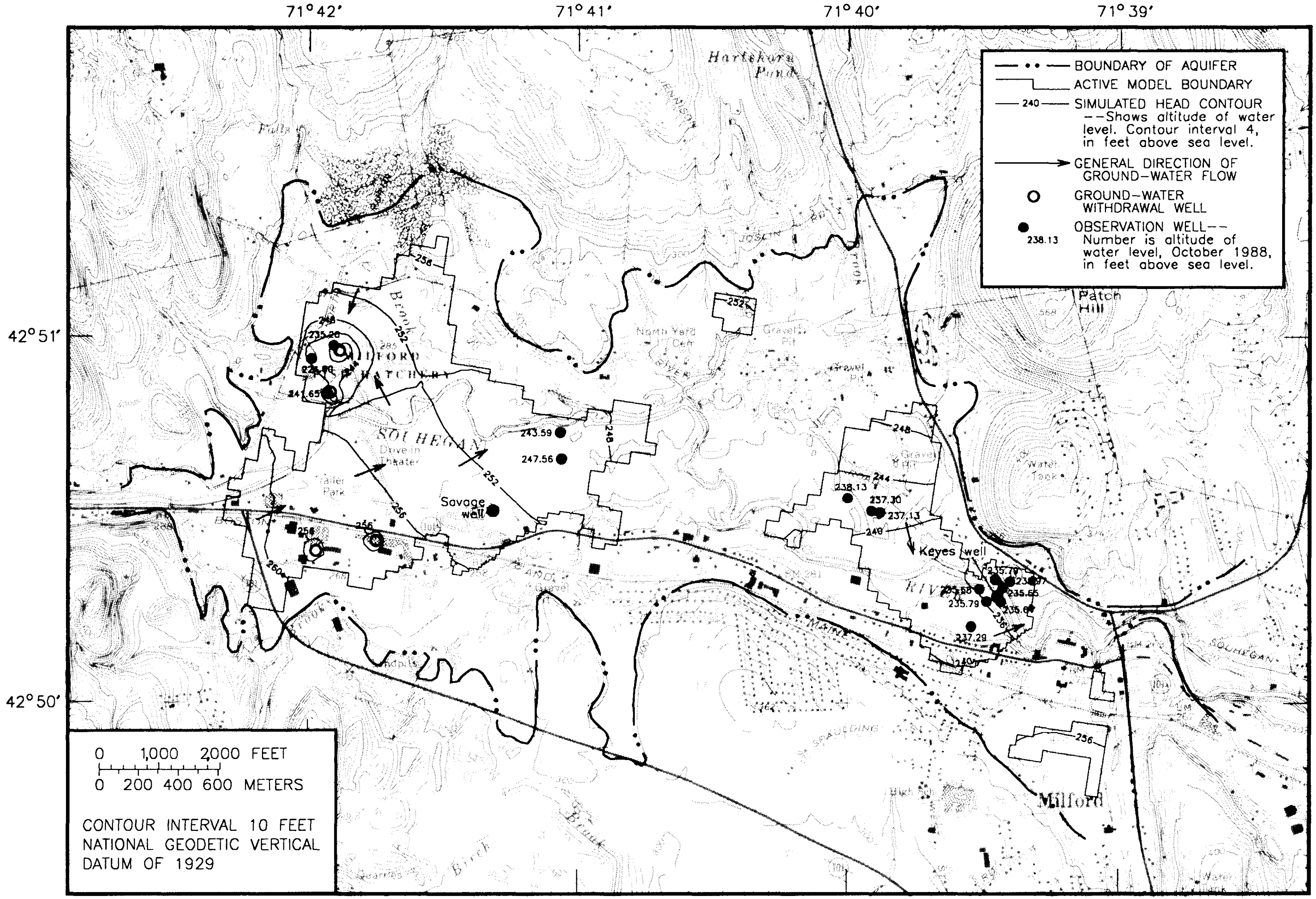
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 20.--Simulated steady-state heads in the Milford-Souhegan aquifer, October 1988: Layer 1.



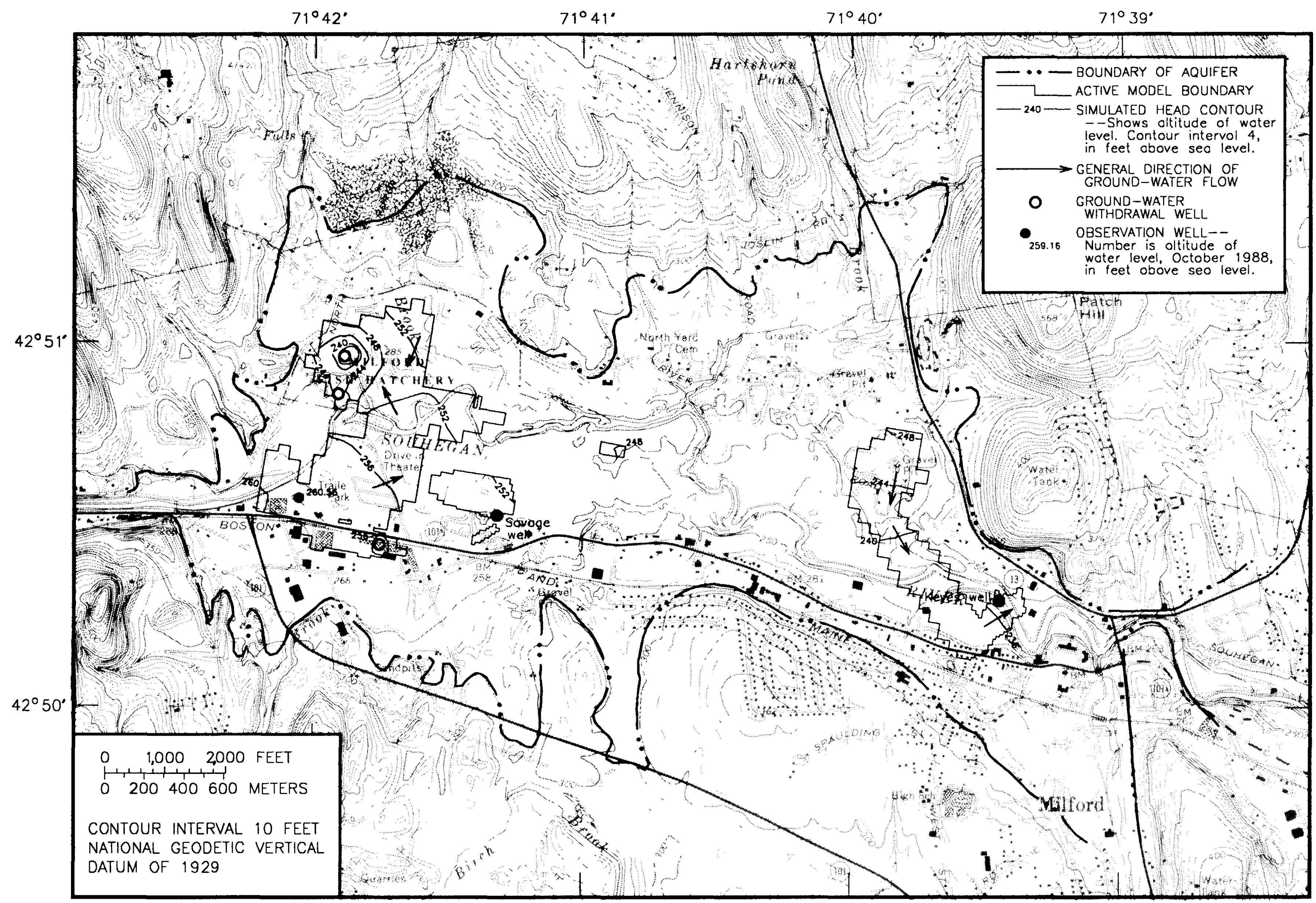
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 21.--Simulated steady-state heads in the Milford-Souhegan aquifer, October 1988: Layer 2.



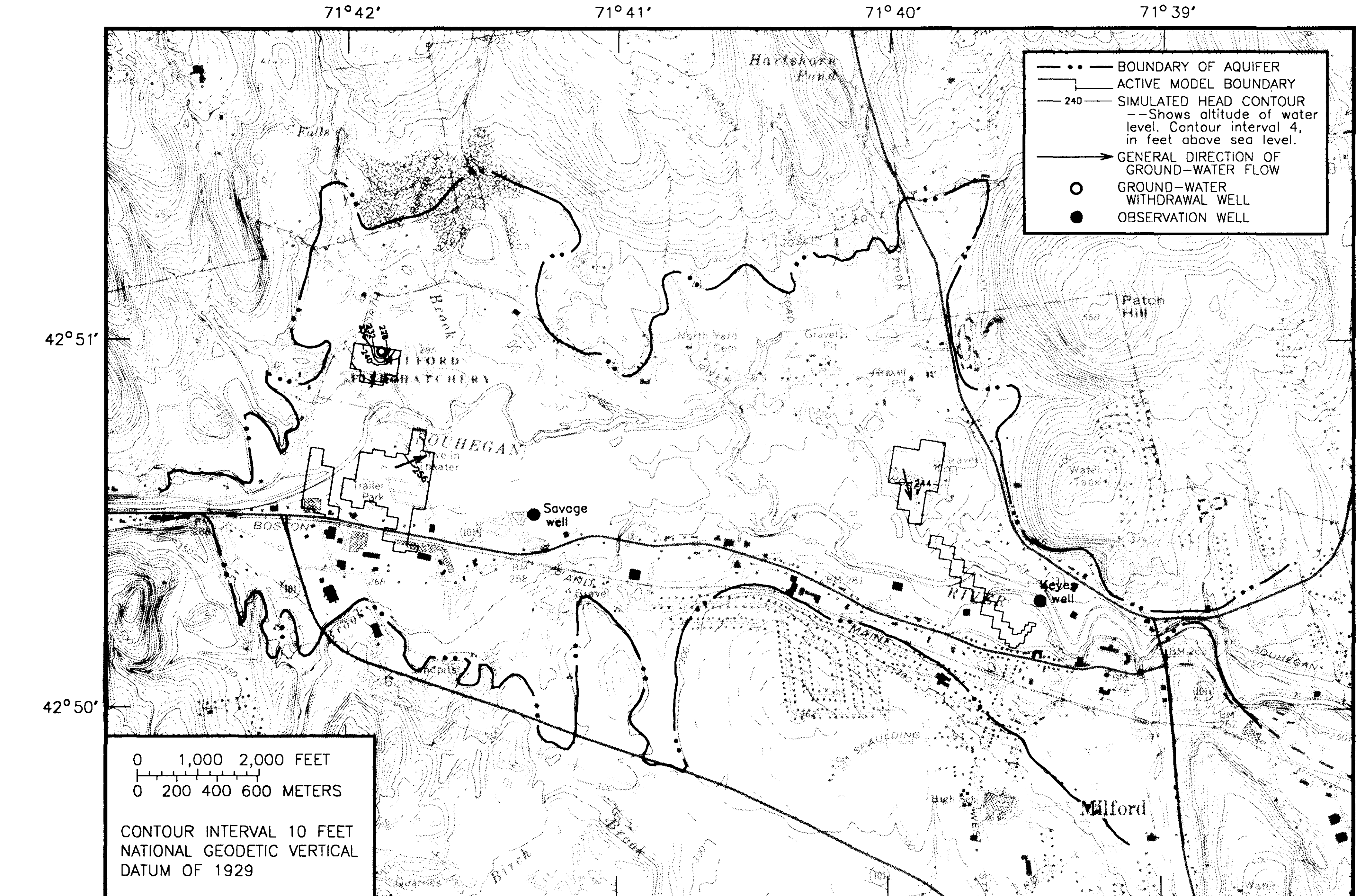
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 22.--Simulated steady-state heads in the Milford-Souhegan aquifer, October 1988: Layer 3.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 23.--Simulated steady-state heads in the Milford-Souhegan aquifer, October 1988: Layer 4.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 24.--Simulated steady-state heads in the Milford-Souhegan aquifer, October 1988: Layer 5.

Table 9.--*Differences between simulated and measured heads for selected wells in the Milford-Souhegan aquifer*

[ft, feet; Meas., measured head; Diff., difference between simulated and measured heads; --, no data]

Well number	Row	Column	Zone[1]	Simulated head (ft)	Average head Meas. (ft)	Average head Diff. (ft)	October 1988 head Meas. (ft)	October 1988 head Diff. (ft)
Wells whose screened interval corresponds to model layer 1								
123	16	70	1-1	251.76	250.8	1.0	251.24	0.52
150	29	81	1-1	242.51	237.8	4.7	238.15	4.36
151	30	83	1-1	241.49	237.0	4.5	237.24	4.25
152	30	83	1-1	241.49	238.7	2.8	238.74	2.75
14	63	60	1-2	250.37	254.3	-3.9	--	--
15	55	61	1-2	249.03	250.1	-1.1	--	--
18	59	61	1-2	249.63	250.4	-.8	--	--
19	57	61	1-2	249.34	249.9	-.6	--	--
20	58	61	1-2	249.48	250.2	-.7	--	--
142	48	97	1-3	236.02	235.4	.6	235.90	.12
143	44	94	1-3	235.82	235.0	.8	235.52	.30
144	46	101	1-3	235.27	235.0	.3	235.58	-.31
145	42	102	1-3	234.87	235.1	-.2	235.69	-.82
146	40	100	1-3	235.01	235.1	-.1	235.69	-.68
147	40	105	1-3	235.12	235.5	-.4	236.01	-.89
160	68	107	1-3	247.79	255.3	-7.5	--	--
204	43	81	1-3	237.23	245.1	-7.9	--	--
179	24	66	1-4	244.39	242.9	1.5	--	--
223	25	68	1-4	242.38	236.5	5.9	--	--
31	56	8	1-5	262.47	265.1	-2.6	264.95	-2.48
32	56	8	1-5	262.47	264.1	-1.6	264.47	-2.00
33	52	8	1-5	262.58	265.1	-2.5	--	--
36	46	10	1-5	260.33	261.9	-1.6	--	--
37	48	10	1-5	260.36	260.9	-.5	260.39	--
38	50	11	1-5	259.51	260.4	-.9	259.86	-.35
39	52	11	1-5	259.49	261.4	-1.9	--	--
41	53	11	1-5	259.48	260.3	-.8	259.84	-.36
42	55	11	1-5	259.48	260.2	-.7	259.98	-.50
50	70	12	1-5	260.54	263.0	-2.5	262.55	-2.01
54	46	24	1-5	254.90	253.1	1.8	253.80	1.10
55	41	24	1-5	254.42	252.9	1.5	253.54	.88
56	38	24	1-5	254.12	252.5	1.6	252.99	1.13
72	32	22	1-5	253.82	254.3	-.5	254.34	-.52
172	46	22	1-5	255.54	253.5	2.0	--	--

Table 9.--*Differences between simulated and measured heads for selected wells in the Milford-Souhegan aquifer*--Continued

Well number	Row	Column	Zone[1]	Simulated head (ft)	Average head Meas. (ft)	Average head Diff. (ft)	October 1988 head Meas. (ft)	October 1988 head Diff. (ft)
Wells whose screened interval corresponds to model layer 2								
137	50	114	2-1	234.52	227.0	7.5	--	--
140	48	113	2-1	234.64	230.0	4.6	--	--
78	47	37	2-2	253.88	242.0	11.9	--	--
220	55	36	2-2	254.92	252.8	2.1	--	--
89	27	45	2-4	249.35	244.4	5.0	--	--
90	26	43	2-4	250.40	247.4	3.0	--	--
91	26	43	2-4	250.40	246.2	4.2	--	--
92	26	44	2-4	250.29	245.6	4.7	--	--
93	26	43	2-4	250.40	246.2	4.2	--	--
94	26	42	2-4	250.48	247.9	2.6	--	--
95	28	13	2-4	249.53	256.3	-6.8	--	--
153	19	10	2-4	254.18	248.1	6.1	--	--
174	28	65	2-4	245.47	236.3	9.2	--	--
175	25	66	2-4	244.21	243.9	+.3	--	--
176	25	67	2-4	242.76	242.7	.1	--	--
180	24	66	2-4	244.44	241.8	2.6	--	--
43	58	11	2-5	259.46	260.5	-1.0	260.56	-1.10
44	61	16	2-5	257.68	256.3	1.4	256.85	.83
45	55	16	2-5	257.45	256.4	1.0	256.22	1.23
171	68	13	2-5	258.44	259.1	-.7	259.99	-1.55
58	36	23	2-6	254.14	251.6	2.5	--	--
98	53	7	2-6	263.51	260.9	2.6	--	--
100	51	7	2-6	263.33	267.7	-4.4	--	--
101	51	7	2-6	263.33	269.3	-6.0	--	--
103	49	7	2-6	263.13	273.2	-10.1	--	--
163	40	47	2-6	252.10	249.8	+2.3	--	--
166	41	44	2-6	252.49	250.2	+2.3	--	--
173	33	65	2-6	246.02	245.1	0.9	--	--
197	38	73	2-6	241.05	250.1	-9.1	--	--
198	36	75	2-6	240.44	247.0	-6.6	--	--
199	35	76	2-6	240.17	241.0	-0.8	--	--

Table 9.--*Differences between simulated and measured heads for selected wells in the Milford-Souhegan aquifer*--Continued

Well number	Row	Column	Zone[1]	Simulated head (ft)	Average head Meas. (ft)	Average head Diff. (ft)	October 1988 head Meas. (ft)	October 1988 head Diff. (ft)
Wells whose screened interval corresponds to model layer 3								
97	51	8	3-1	262.57	264.0	-1.4	--	--
99	54	7	3-1	263.53	261.7	1.8	--	--
104	48	8	3-1	262.40	272.9	-10.5	--	--
209	72	11	3-1	262.06	264.6	-2.5	--	--
1	57	95	3-2	237.22	236.6	0.6	237.26	-.04
2	48	97	3-2	236.02	235.3	.7	235.79	.23
3	44	95	3-2	235.75	235.1	.6	235.59	.16
4	46	102	3-2	235.37	235.0	.4	235.61	-.24
6	29	81	3-2	242.53	237.7	4.8	238.13	4.40
7	30	83	3-2	241.50	237.0	4.5	237.30	4.20
8	30	83	3-2	241.50	236.8	4.7	237.13	4.37
74	44	50	3-2	252.41	257.0	-4.6	--	--
77	48	104	3-2	235.34	235.4	-.1	--	--
126	44	100	3-2	235.29	235.7	- .4	235.66	-.37
128	43	41	3-2	253.03	250.6	2.4	--	--
129	46	103	3-2	235.30	233.2	2.1	--	--
130	43	97	3-2	235.57	231.4	4.1	--	--
131	48	103	3-2	234.44	233.7	1.7	--	--
132	42	102	3-2	235.20	235.1	.1	235.65	-.45
133	40	99	3-2	235.27	235.1	.2	235.72	-.45
134	40	104	3-2	235.17	235.4	-.2	235.97	-.80
135	46	105	3-2	235.19	231.4	3.8	--	--
136	47	105	3-2	235.20	234.1	1.1	--	--
164	43	48	3-2	252.43	249.8	2.6	--	--
165	41	45	3-2	252.42	250.0	2.4	--	--
215	67	18	3-2	257.55	261.3	-3.8	--	--
9	40	20	3-3	255.53	256.0	-0.5	--	--
23	33	58	3-3	249.66	247.4	2.3	--	--
24	30	57	3-3	249.44	247.1	2.3	247.56	1.88
25	28	57	3-3	248.96	245.3	3.7	243.59	5.37
154	55	19	3-3	256.34	258.8	-2.5	--	--
195	28	18	3-3	252.78	254.4	-1.6	--	--
196	28	17	3-3	252.45	254.5	-2.0	--	--
202	32	18	3-3	254.20	253.5	1.7	--	--
216	62	18	3-3	256.54	258.4	-1.9	--	--
219	30	16	3-3	254.83	255.4	-0.6	--	--

Table 9.--*Differences between simulated and measured heads for selected wells in the Milford-Souhegan aquifer*--Continued

Well number	Row	Column	Zone[1]	Simulated head (ft)	Average head Meas. (ft)	Average head Diff. (ft)	October 1988 head Meas. (ft)	October 1988 head Diff. (ft)
Wells whose screened interval corresponds to model layer 3								
221	41	21	3-3	255.30	257.4	-2.1	--	--
222	30	44	3-3	256.21	250.8	-.3	--	--
47	67	15	3-4	256.21	252.6	3.6	--	--
49	61	19	3-4	254.09	252.9	1.2	--	--
57	48	12	3-4	258.70	258.6	-.1	--	--
84	25	14	3-4	243.13	235.3	+7.8	235.20	7.93
87	28	14	3-4	237.07	244.6	-7.5	241.65	-4.58
188	31	14	3-4	256.20	251.9	4.3	--	--
194	36	13	3-4	257.82	258.4	-.6	--	--
208	25	14	3-4	[2]236.71	228.7	3.0	224.60	7.11
213	67	14	3-4	254.41	258.3	-3.9	--	--
214	58	13	3-4	257.65	260.1	-2.5	--	--
Wells whose screened interval corresponds to model layer 4								
75	41	95	4-1	234.84	229.7	5.1	--	--
217	54	18	4-4	256.20	251.8	4.9	--	--
46	49	12	4-5	258.57	259.8	-1.2	259.16	-.59

[1] Zone of horizontal hydraulic conductivity as shown on figures 15-19.

[2] Simulated head was adjusted to represent head at a pumped well using the method described by Trescott (1976, p. 9).

average heads are similar to those between simulated and October 1988 heads.

Head differences are summarized, by hydraulic-conductivity zone, in table 10. Comparison of head differences by zone specifically shows how close simulated heads matched measured heads. Absolute mean differences between simulated and October 1988 heads are less than 3 ft except in zones 3-3 and 3-4. Absolute mean differences between simulated and average heads are greater, and exceed 2 ft in most zones (table 10). The absolute mean difference between simulated and October 1988 heads for all zones is 2.23 ft. The absolute mean difference between simulated and average heads for all zones is 3.3 ft. The standard mean difference between simulated and October 1988 heads are not strongly biased either positively or negatively; thus, distribution of simulated heads is similar to the distribution of measured heads. Mean differences between simulated and average heads are large for zones 2-1 and 2-2 in layer 2, and for layer 4; the significance of these large differences in layer 4 is unknown because of the small number of observations in these zones.

Head differences are summarized by model layer in table 11. Simulated heads in layers 1-4 agree closely with October 1988 heads and in layers 1-3 with average heads (table 11). There were few data for comparisons in layer 4 and no data for comparison in layer 5. The standard mean difference between simulated and October 1988 heads shows that the simulation for layer 1 and 2 are relatively unbiased, whereas simulated heads for layer 3 are somewhat higher than measured heads.

Simulated vertical head gradients are small (less than 0.003 ft/ft) except near production wells. The differences between simulated and observed heads per model layer in table 11 are probably

Table 10.--*Differences between simulated and measured heads in the Milford-Souhegan aquifer, by hydraulic-conductivity zone*

[--, no data]

Zone	Average heads				October 1988 heads			
	Number of obser-vations	Difference from simulated head, in feet			Number of obser-vations	Difference from simulated head, in feet		
		Max-imum	Absolute mean	Standard mean		Max-imum	Absolute mean	Standard mean
1-1	4	4.7	3.3	3.3	4	4.36	2.95	2.95
1-2	7	-3.9	1.2	-.5	0	2.0	--	--
1-3	8	-7.9	2.2	-1.7	6	-.9	.52	-.38
1-4	2	5.9	3.7	3.7	0	--	--	--
1-5	15	-2.7	1.5	-.6	11	-2.48	1.04	-.47
2-1	2	7.2	5.7	5.7	0	--	--	--
2-2	2	8.3	4.5	3.8	0	--	--	--
2-3	0	--	--	--	0	--	--	--
2-4	12	11.9	6.2	5.1	0	--	--	--
2-5	4	1.4	1.1	.2	4	-1.55	1.15	-.15
2-6	11	-10.1	4.4	-3.2	0	--	--	--
3-1	4	-10.3	4.1	-3.2	0	--	--	--
3-2	22	4.8	1.3	1.2	11	4.40	1.42	1.0
3-3	12	3.7	1.8	-.1	2	5.37	3.65	3.65
3-4	10	7.5	3.5	-.5	3	7.93	6.54	3.48
4-1	1	5.1	5.1	5.1	0	--	--	--
4-2	0	--	--	--	0	--	--	--
4-3	0	--	--	--	0	--	--	--
4-4	1	4.9	4.9	4.9	0	--	--	--
4-5	1	-1.2	1.2	-1.2	1	-.59	.59	.59

caused by the geographic bias of the sample population for each layer rather than vertical simulation bias. This means that well location has a larger impact on calibration error than the vertical position of its screened interval in the aquifer. Simulated vertical head gradients for cells representing well nests generally match measured vertical head gradients for October 1988. Comparisons were made between the shallow and deep wells at some of the well nests (table 12). Simulated vertical head differences poorly match observed heads at well nests 152-8 where there is a locally elevated water table; this water table was not simulated in the model. The discrepancy in simulated and observed flow directions at well nests 145-132 and 147-134 suggest the recharge from lateral-till seepage might be overestimated north of the Souhegan River across from the Keyes well field (fig. 2).

Simulated stream seepage compares well with net seepage measured in October 1988 for the drainage basins shown in figure 25. For calibration of the simulated water budget, the drainage area to each tributary to the Souhegan River and four sections of the mainstem of the Souhegan River crossing the Milford-Souhegan aquifer are delineated as shown in figure 25. In general, stream reaches that lost water according to streamflow measurements also lost water in the simulations, and reaches that gained water according to streamflow measurements also gained in the simulations. Simulated stream

Table 11.--*Differences between simulated and measured heads in the Milford-Souhegan aquifer, by model layer*

[--, no data]

Layer	Average heads: Number of observations	Average heads: Difference from simulated head, in feet: Absolute mean	Average heads: Difference from simulated head, in feet: Standard mean	October 1988 heads: Number of observations	October 1988 heads: Difference from simulated head, in feet: Absolute mean	October 1988 heads: Difference from simulated head, in feet: Standard mean
1	34	2.0	-0.3	20	1.32	0.22
2	31	4.1	1.1	4	1.15	-.60
3	48	1.8	.4	16	2.36	1.80
4	3	3.7	2.9	1	.6	-.6
5	0	--	--	0	--	--

seepage, however, fails to reproduce adequately the magnitude of gain along the mainstem of the Souhegan River in drainage basin 10 (table 13 and fig. 25). Results of simulations of tributary streams were satisfactory, considering the relatively small amount of seepage. The differences between simulated and measured seepage for basins 4 and 5 are attributed, to an unknown degree, to unresolved discrepancies associated with streamflow estimates at measurement site 15 (table 4). In contrast, the large differences for basin 10 could not be resolved by reasonable variations in model parameters; additional streamflow gaging and simulation of the river would be necessary to resolve these differences.

The following are possible reasons for the large differences in simulated and measured seepage at basin 10. First, basin 10 comprises only the western part of the drainage area between streamflow sites 1 and 6 (figs. 11 and 25); therefore, streamflow gains beyond the simulated area are unaccounted for in the model. Second, the simulation may not account for all recharge sources to the aquifer. One possible recharge source could be leakage from the Milford public-supply water-distribution system. Water supply in basin 10 is exclusively from a public-supply system that obtains water from outside the simulated area. It is conceivable that as much as 0.50 ft^3/s leaks from that system; such leakage would account for 25 percent of the discrepancy.

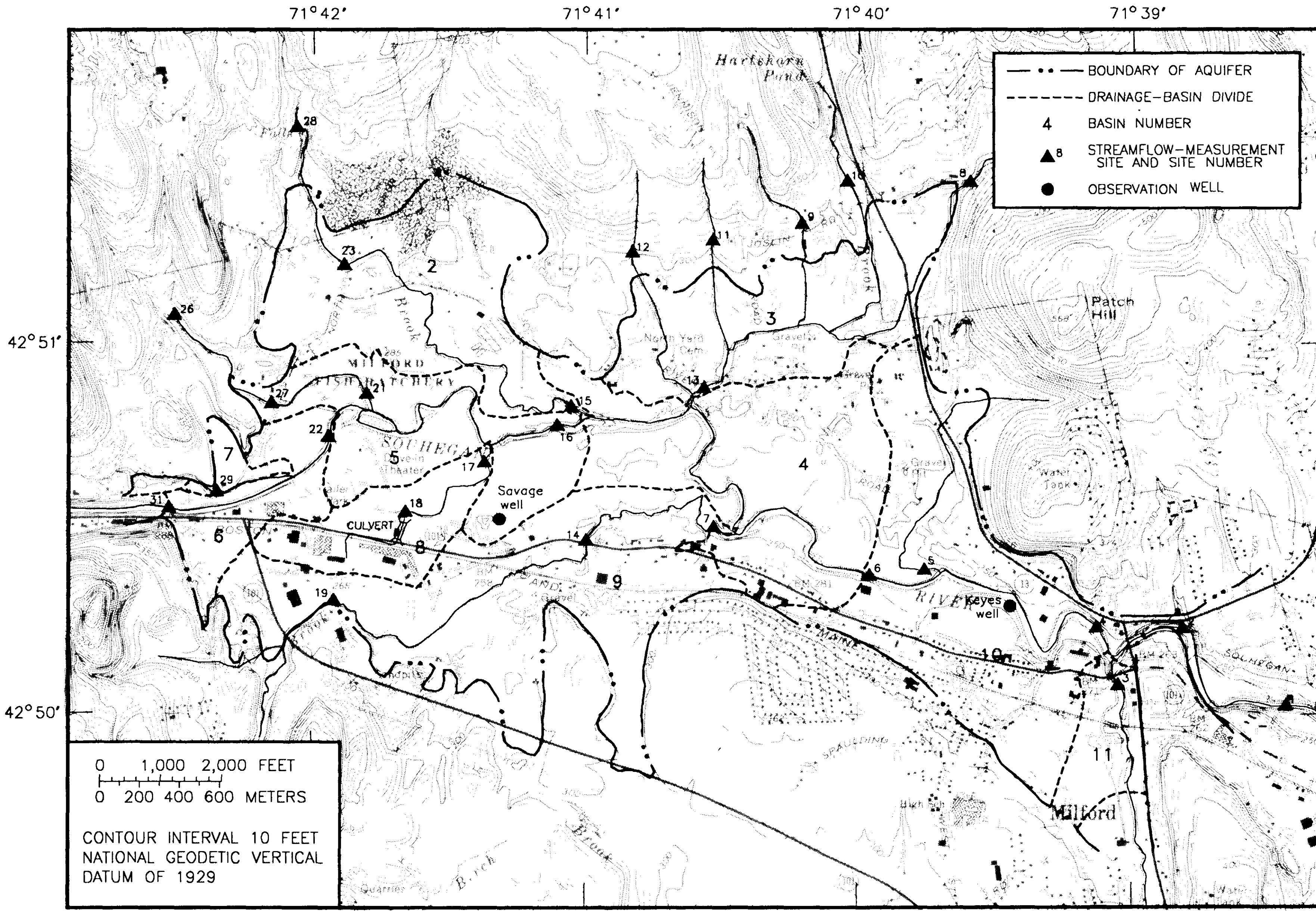
Sensitivity Analysis

A sensitivity analysis of the model was done to determine the response of the calibrated model to changes in model parameters. A secondary reason

Table 12.--*Comparison of simulated and measured heads in select shallow and deep well nests in the Milford-Souhegan aquifer*

[-, denotes an upward flow direction; +, denotes a downward flow direction]

Well nests (well number)	Row	Column	Simulated head, in feet	Measured head, in feet
143-3	44	94	0	-0.07
151-7	30	83	- .01	-.06
144-4	46	101	- .10	- .03
142-2	48	97	0	+ .11
145-132	42	102	- .33	+ .04
147-134	40	105	- .05	+ .04
152-8	30	83	- .01	+1.61



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 25.--Drainage basins in the Milford-Souhegan aquifer.

Table 13.--*Differences between simulated and measured streamflow gains and losses for each drainage basin in the Milford-Souhegan aquifer*

[ft^3/s, cubic feet per second; --, no data]

Drainage basin	Simulated stream seepage calibrated model [1] (ft^3/s)	Measured stream seepage October [1] 1988 (ft^3/s)	Difference between simulated and measured stream seepage [2] (ft^3/s)
2	-.67	-.18	-.49
3	.13	.05	+.08
4	1.28	[3]1.58	-.30
5	-.50	.27	-.77
6	-.87	-.73	-.14
7	0	-.06	+.06
8	-.35	-.31	-.04
9	-.34	-.47	+.13
10	1.05	3.47	-2.42
11	-.24	--	--

[1] Negative values denote streamflow loss to the aquifer; positive values denote streamflow gain.

[2] A negative value means either simulated seepage losses are greater than measured seepage losses or simulated seepage gains are less than measured seepage gains; a positive value means either simulated seepage losses are less than measured seepage losses or simulated seepage gains are greater than measured seepage gains.

[3] Seepage estimated as 60 percent of June 1988, basin 4, seepage (see table 4).

for this analysis was to determine if differences between simulated and measured data could be accounted for by changing parameters from their values in the calibrated model. For a given simulation, the principal input parameters--recharge, horizontal and vertical hydraulic conductivity of the aquifer, and streambed conductance--were independently and uniformly increased and decreased by a factor of 50 percent (by an order of magnitude for vertical hydraulic conductivity) while all other parameters were kept constant.

Each parameter was varied independently, and differences from heads in the calibrated model were examined for each hydraulic conductivity zone. Results of the sensitivity analysis are given in table 14 and graphically displayed in figure 26. Head comparisons were made for the same cells as were compared with measured heads (table 9). The difference between heads in the calibrated model and heads generated during a subsequent simulation will show solely the effect of changes produced by that subsequent simulation. A standard mean was used to examine head differences by zone, for each parameter variation. The direction of change in head difference is important in this analysis. This change can be compared with the difference between calibrated-model head and measured head, also listed in table 14, to determine if varying a parameter improves or worsens the match between simulated and measured heads. The difference between computed heads and measured heads per hydraulic conductivity zone for layers 1 and 3 (the two most critical model layers) are shown in figure 26, which clearly illustrates any improvement of model-fit as a result of further parameter adjustment.

Simulated net seepage is listed in table 15, by drainage basin (fig. 11), for each independent parameter variation. Streamflow data from the October 1988 measurement are not available for basin 11 and could not be included in table 15.

Table 14.--*Mean differences between heads from the calibrated model and heads computed during sensitivity analysis, by hydraulic-conductivity zone*

[HC, hydraulic conductivity; --, no observed head data for comparison. All head values are in feet]

Zone	Number of obser-vations	Calibrated-model head minus meas-ured head	Standard mean differences between calibrated-model heads and heads computed during sensitivity analysis							
			Recharge		Horizontal HC		Vertical HC		Streambed HC	
			× 0.5	× 1.5	× 0.5	× 1.5	× 0.1	× 10	× 0.5	× 1.5
1-1	4	3.35	-1.28	1.15	1.80	-0.75	0.25	-0.10	0.30	-0.10
1-2	5	-.53	-1.21	1.04	1.71	-.84	.10	-.06	-.03	.01
1-3	8	-1.72	-.39	.40	.52	-.17	.15	-.05	.21	-.08
1-4	2	3.70	-.15	.20	-.45	-.35	.00	.00	.95	-.40
1-5	15	-.61	-.39	.35	.26	-.21	-.18	.02	-.88	.42
2-1	2	5.7	-.10	.15	.05	.00	.20	.00	.20	.00
2-2	2	3.8	-.55	.45	.70	-.45	-.20	.00	-.65	.30
2-3	0	--	--	--	--	--	--	--	--	--
2-4	12	5.1	-.21	.20	-1.60	.29	.13	-.06	-.21	.12
2-5	4	.20	-.72	.62	-.20	-.18	-.13	-.05	-1.85	1.12
2-6	11	-3.2	-.22	.24	.25	-.13	-.05	.01	-.09	.05
3-1	4	-3.2	-.42	.35	.83	-.45	-.22	-.02	-1.07	.60
3-2	22	1.27	-.51	.47	.55	-.22	.28	-.14	.04	.01
3-3	12	-.13	-.34	.32	-.01	-.09	-3.29	.39	-.79	.36
3-4	10	-.51	-.62	.57	-.63	1.27	[a]-7.80	1.71	2.06	.91
4-1	1	5.1	-.15	.14	.04	.00	.23	-.04	.23	-.08
4-2	0	--	--	--	--	--	--	--	--	--
4-3	0	--	--	--	--	--	--	--	--	--
4-4	1	4.9	-.60	.60	-.10	-.20	-3.50	.40	1.60	.90
4-5	1	-1.20	-.50	.60	.10	-.30	-4.00	.30	-1.20	.50

[a] Some heads produced within zone 3-4 in this simulation are unrealistic.

Increasing or decreasing recharge has a substantial effect on the simulated ground-water-flow system (fig. 26). Decreasing recharge by 50 percent of the amount in the calibrated model caused simulated heads to be lower, as shown by negative standard-mean head differences (table 14). For example, decreasing recharge by 50 percent in zone 1-1 resulted in a mean difference of -1.28 ft from the head in the calibrated model. In zone 1-1, increasing recharge by 50 percent resulted in a difference of 1.15 ft from the head in the calibrated model. Varying recharge had the greatest effects on zones 1-1 and 1-2, in which head changes were greater than 1 ft. This pronounced effect in zones 1-1 and 1-2 is probably due to the low horizontal hydraulic conductivity of these zones. In zones for which differen-

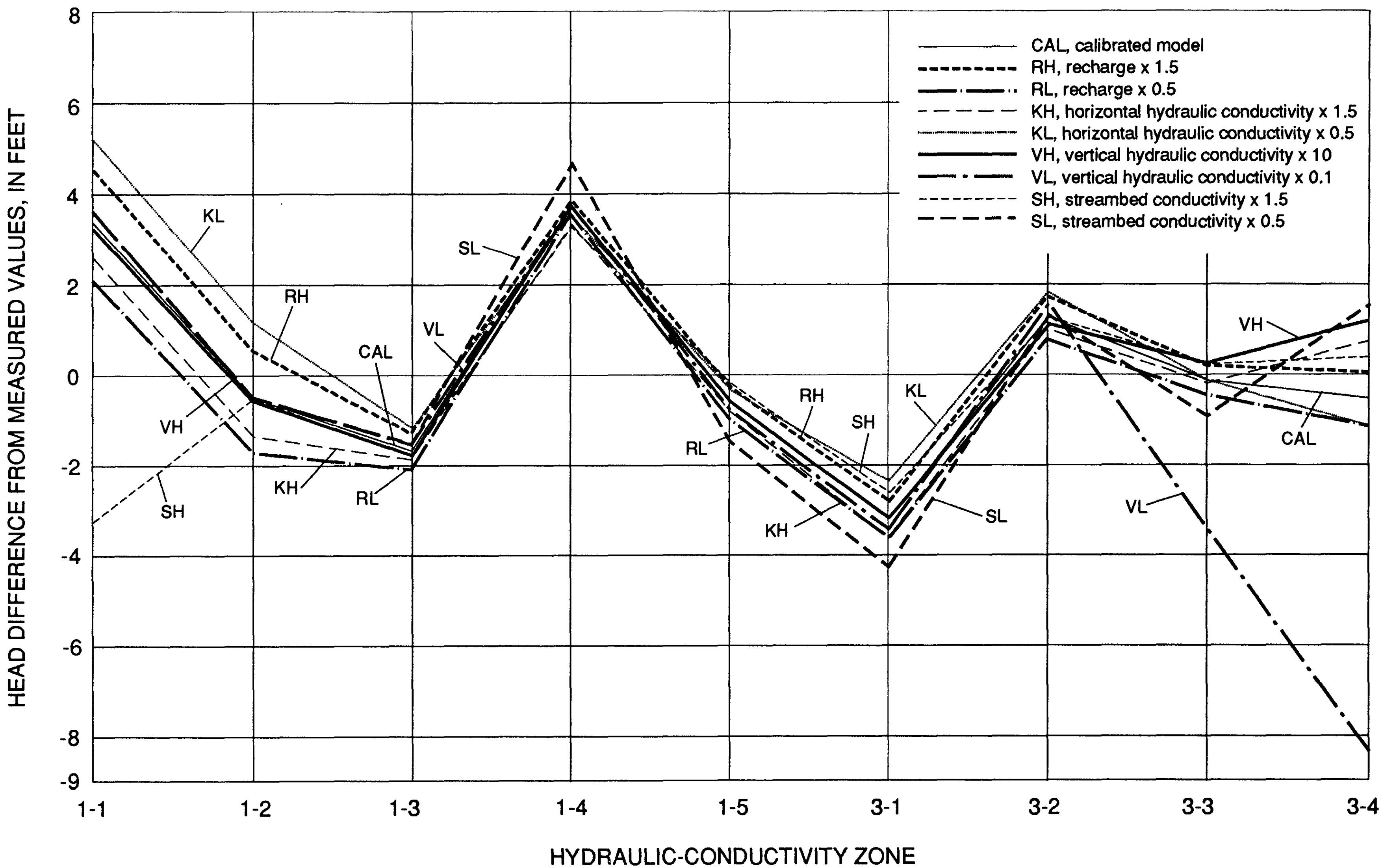


Figure 26.--Mean differences between measured heads and heads computed by the calibrated model and during sensitivity analysis by hydraulic-conductivity zone for layers 1 and 3.

Table 15.--*Mean differences between net seepage from the calibrated model and net seepage computed during sensitivity analysis, by drainage basin*

[HC, hydraulic conductivity. All seepage values are in cubic feet per second]

Basin	Calibrated-model seepage minus measured seepage[1]	Mean differences between calibrated-model seepage and seepage computed during sensitivity analysis[2]							
		Recharge		Horizontal HC		Vertical HC		Streambed HC	
		× 0.5	× 1.5	× 0.5	× 1.5	× 0.1	× 10	× 0.5	× 1.5
2	- 0.49	-0.32	0.32	0.21	-0.13	-0.12	0.04	0.07	-0.06
3	.08	-.24	.25	.21	-.19	.01	.00	-.01	.01
4	-[3].30	-.36	.35	-.34	.29	.00	-.01	-.49	-.02
5	- .77	-.27	.26	-.76	.48	.14	-.03	-.27	.13
6	- .14	-.17	.14	.43	-.27	.03	-.03	-.08	.13
8	- .04	-.07	.07	.04	-.04	-.07	.00	.13	-.13
9	.13	-.16	.17	.24	-.13	.00	-.01	.13	-.06
10	- 2.42	-.25	.26	.44	.25	-.03	.01	-.01	.02
11	([4])	-.08	.09	.20	-.19	.01	.00	.02	-.01

[1] A negative value means either simulated seepage losses are greater than measured seepage losses or simulated seepage gains are less than measured seepage gains; a positive value means either simulated seepage losses are less than measured seepage losses or simulated seepage gains are greater than measured seepage gains.

[2] A negative value means that sensitivity seepage losses are greater than calibrated seepage losses or sensitivity seepage gains are less than calibrated seepage gains. A positive value means sensitivity seepage losses less than calibrated seepage losses or the sensitivity seepage gains greater than calibrated seepage gains.

[3] Values estimated as 40 percent of June, 1989 data.

[4] Seepage measurements not made in this basin.

ces between measured head and calibrated-model head are small (less than 1 ft), that difference could be accounted for by variations in recharge. For most zones, however, the calibration difference cannot be accounted for by varying recharge.

Recharge also has a substantial effect on simulated net seepage (table 15). The greatest effect is in drainage basin 4, where net seepage increased by 0.35 ft^3/s when recharge increased by 50 percent. Increasing or decreasing recharge by 50 percent affected simulated seepage in the drainage basins in different ways. Recharge could account for most of calibration difference (table 15) in some basins; for example, increasing recharge by 50 percent would reduce the difference between measured seepage and calibrated-model seepage to zero in basin 6. In other basins, however, differences between the computed and measured seepage could not be accounted for by varying recharge by 50 percent, and uniformly increasing or decreasing recharge would not uniformly improve the match between simulated and measured seepage. The maximum change in net seepage during sensitivity analysis was in basin 2, which varied by as much as 0.32 ft^3/s from the calibrated model seepage. The smallest range in seepage, ± 0.07 ft^3/s, was in basin 8 and is because of its small drainage area. Increased recharge improved the fit of calibrated-model seepage to measured seepage in six of eight basins.

Varying horizontal hydraulic conductivity by 50 percent resulted in similar to slightly larger head changes than varying recharge by 50 percent (fig. 26). The mean head differences generally were less than 1 ft, the major exceptions were in zones 1-1, 1-2, 2-4, and 3-4 (table 14). Simulated heads are more sensitive to decreases in horizontal hydraulic conductivity than to increases. The effect of varying horizontal hydraulic conductivity was most noticeable near production wells, especially in layers tapped by those wells. In zone 3-4, which contains the simulated wells for the Fish Hatchery, the manufacturing company, and the wire and cable company, decreasing horizontal hydraulic conductivity by 50 percent lowered simulated heads dramatically (table 14).

Increasing or decreasing horizontal hydraulic conductivity caused significant changes in net seepage, which were, in some basins, greater than the effects of varying recharge. The effects of varying horizontal hydraulic conductivity does not produce a consistent pattern on net seepage (table 15). Decreasing horizontal hydraulic conductivity by 50 percent caused the greatest change in net seepage in basin 5, -0.76 ft^3/s (table 15). Increasing horizontal hydraulic conductivity caused the greatest change in seepage, 0.48 ft^3/s, also in basin 5. Seepage in basin 8 was relatively insensitive to changes in horizontal hydraulic conductivity. Increasing horizontal hydraulic conductivity significantly improves the fit of the model to measured seepage, except in basins 4, 5, and 9.

The sensitivity of the model to vertical hydraulic conductivity was examined by varying this parameter by a factor of 10 from the calibrated-model values. These variations produced little change from the calibrated model in either head differences (table 14, fig. 26) or net seepage (table 15). The exception was in the immediate vicinity of pumped wells, where decreasing this parameter produced strong vertical head gradients. For example, head is reduced significantly in zone 3-4, which contains four major production wells, when vertical hydraulic conductivity is reduced by an order of magnitude. Similarly, the greatest change in net seepage (0.14 ft^3/s, table 15) is in basin 5, which contains two major production wells.

Increasing or decreasing streambed conductance also produced substantial changes in simulated head in some zones (fig. 26) and less significant changes in net seepage in most basins. Reducing streambed conductance by 50 percent produced large head decreases (more than 1 ft) in zones 2-5, 3-1, 3-4 (table 14). These zones all are near the main stem of the Souhegan River, and zones 2-5 and 3-4 contain simulated pumped wells. Basins 4 and 5 (table 15) show large net seepage losses (-0.49 and -0.27 ft^3/s) because streamflow gains in these basins were significantly decreased by reducing streambed conductance by 50 percent. Increasing streambed conductance caused some stream reaches to lose additional water but also caused parts of the same reach to gain additional water; thus, changes in net seepage were relatively small for the most part. The effects of variations in streambed conductance had differing effects on the ground-water-flow system. Variations in streambed conductance could account for some difference in calibrated-model heads, but not uniformly, and could not account for the magnitude of difference in calibrated-model seepage in most basins. The large difference in net seepage (basin 10) could not be accounted for by variations in streambed conductance (table 15). (Larger variations in streambed conductance than shown in table 15 were, in fact, experimented with, but even the larger variations could not account for the discrepancy.)

In summary, the calibrated model is most sensitive to recharge and horizontal hydraulic conductivity among the model parameters and more sensitive to decreases than to increases in the parameters examined. The model is generally insensitive to vertical hydraulic conductivity except near pumped wells where vertical hydraulic conductivity is important in controlling drawdowns. The analysis revealed that some changes in parameters could be made in places that would improve the match between computed and measured heads and net seepages. As illustrated in figure 26, most parameter variations, however, do not universally produce a better fitting model. The model, therefore, is believed to be relatively well calibrated, and further changes and refinements are not warranted for the intended use of this model.

Evaluation of Model Results

The degree of confidence placed in the calibrated model depends on several factors, including the original conceptual model and the interpreted boundary conditions, the validity of assuming steady-state conditions, grid discretization, the amount and distribution of water-level and streamflow data for calibration, and the accuracy of all flux estimates. The model reproduces observed results more accurately in some areas than in others depending on distribution of data. In general, simulated data closely matched observed data where there were more data to refine the model. Parameters used in the calibrated model, and in the ground-water-flow simulation itself, are evaluated below.

Hydraulic Characteristics

Final parameters used in the calibrated model are generally similar to initial estimates of hydraulic characteristics derived from previous studies or field measurements. These parameters are regionally averaged and may not be representative of areas smaller than the hydraulic-conductivity zones or drainage basins in the model.

Modifications were made to streambed conductance and stream stage during calibration. Streambed conductance was adjusted by varying the streambed vertical hydraulic conductivity component of the conductance term. An initial estimate of 3 ft/d for streambed hydraulic conductivity was used in the model; final values differ, by as much as 1.5 ft/d, from initial estimates at many locations. Streambed hydraulic conductivity strongly influences head values near the stream. Final streambed hydraulic conductivities used for tributaries are 4 ft/d near the valley wall and lower, 3 to 1.5 ft/d, downstream toward the main stem of the river. Simulated seepage from the Souhegan River is more influenced by streambed hydraulic conductivity in the western reaches of the river, such as in drainage basins 5 and 6, than downstream in the eastern reaches. Streambed hydraulic conductivity is highest (4 ft/d) in basin 6, lowest (2 ft/d) in basins 4 and 5, and was kept at 3 ft/d in basin 10. Stream stage strongly affects simulated head in a broad region around streams and, thus, in the flow system overall. Where stage was measured, stream stages were kept at the measured values. Where stage was interpolated between measured locations, it was varied slightly to improve the fit of simulated heads or fluxes.

Simulations were relatively sensitive to horizontal hydraulic conductivity of the aquifer. Final hydraulic conductivities differ from initial values in zones 1-1, 1-2, 1-5, 2-5, 2-6, and 3-2. Horizontal hydraulic conductivity was increased by 25 percent of the initial value in zone 1-1 and was decreased by 25 percent in zones 1-2, 1-3, and 1-5. Horizontal hydraulic conductivity in layer 2 was decreased by 25 percent in zones 2-5 and 2-6. In the northern part of zone 3-2, horizontal hydraulic conductivity was increased by 100 percent and, in the eastern part of this zone, was decreased by 25 percent. Horizontal hydraulic conductivities in layers 4 and 5 were unchanged from initial estimates.

Variations in vertical hydraulic conductivity had little effect on simulated heads or river seepage except near simulated production wells. Decreasing vertical hydraulic conductivity to the ratio 0.001 caused some cells to dry in layer 1 near major ground-water withdrawals. Final vertical-hydraulic-conductivity ratios are 0.1 between layers 1 and 2 and 0.01 between all other layers. It is difficult to determine how reasonable these estimates are because the model is relatively insensitive to this parameter.

As a qualitative calibration analysis, drawdowns resulting from simulation of the Keyes well at the maximum daily withdrawal rate were compared to observed drawdowns from the aquifer test at the Keyes site in October 1988. Simulated drawdowns in layer 1 were twice the observed drawdown. This overprediction is acceptable because the simulation is for steady state, whereas the water table was still declining at the end of the aquifer test (fig. 5). Simulated drawdown was greater than observed drawdown (in layer 3), which may also be attributed to the heads at the end of the aquifer tests not reaching steady state. The simulation indicated that the initial estimate of horizontal hydraulic conductivity for zone 3-2 (fig. 17) originally was high.

Additional work to verify hydraulic conductivity estimates for the aquifer and streambed could prove useful to future investigations. A more thorough calibration than was possible during this investigation would also be beneficial and could include transient simulations for comparison with the October 1988 aquifer test at the Keyes site. Further calibration could produce improved estimates of model parameters and greater confidence in model simulations.

Ground-Water Flow

The simulation of October 1988 ground-water flow indicates that (1) ground-water flow is primarily horizontal; (2) one major component of flow is downvalley, from west to east, and a second major component is perpendicular to the Souhegan River; (3) ground-water flow is influenced, to a large degree, by stream stage because of the close hydraulic connection between the aquifer and streams; (4) the Souhegan River gains water from the aquifer along its eastern two-thirds, in the simulated area, and loses water to the aquifer along its western third; (5) regional flow throughout the aquifer is least affected by variations in vertical hydraulic conductivity and most affected by variations in recharge and horizontal hydraulic conductivity, in contrast, local flow near pumped wells is affected by variations in vertical hydraulic conductivity; and (6) simulated ground-water flow to pumped wells is strongly affected by aquifer geometry and nearby boundary conditions.

Horizontal head gradients are typically 0.003 to 0.005 ft/ft. The horizontal head gradient is locally steep, 0.032 ft/ft, downstream of the Great Brook dam near the center of Milford. Here, ground-water heads are controlled downgradient by stage in the Souhegan River and upgradient by Great Brook stage upstream from a dam (fig. 14).

Vertical head gradients are generally less than 0.003 ft/ft, except near major pumped wells. Downward head gradients are induced by withdrawals at wells 208 and 87 (Milford Fish Hatchery), 47 (the manufacturing company), and 49 (the wire and cable company). The largest downward gradient, 0.36 ft/ft, is at well 208. Vertical head gradients are smaller at other pumped wells. Throughout most of the aquifer, slight downward gradients are produced by recharge.

Simulated head gradients are upward in cells below gaining river reaches. Most upward head gradients are small, several orders of magnitude less than horizontal head gradients. The exception is at the eastern model-boundary where cells containing the simulated Souhegan River are characterized by forced upward flow to the river because of a reduction in aquifer thickness.

Ground-water withdrawals at well 208 (at the Milford Fish Hatchery) cause significant lowering of the water table and affect horizontal and vertical flow over a much larger area than at other pumped wells. The magnitude of the well's effect can be attributed to the large amount of water withdrawn (table 7), and to the proximity of the well to the valley wall. Similar amounts are withdrawn from well 87, but, because of its proximity to the Souhegan River, heads at well 87 are affected much less than those at well 208. The aquifer near well 87 receives significant recharge from induced infiltration, unlike at well 208.

The simulated water budget shows that simulated seepage losses exceed simulated gains by 0.55 ft^3/s. Simulated seepage losses are 11 percent lower than measured seepage losses; the difference being the inability of the model to simulate seepage gains in basin 10 (fig. 25).

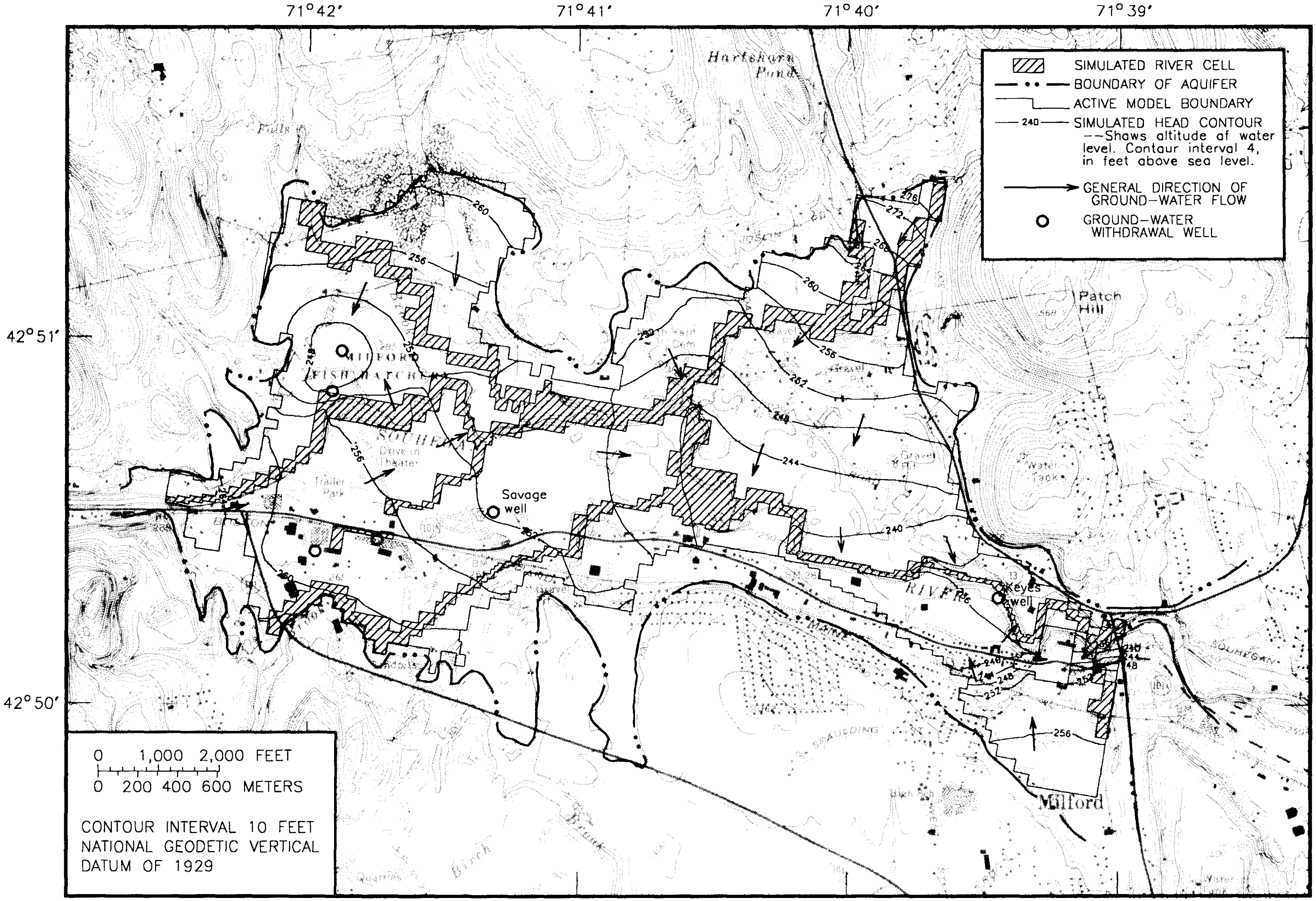
Steady-state head distributions for model layers 1 (representing the upper part of the aquifer including the water table) and 3 (representing the screen zone of the pumped well) are shown in figures 27 and 28 for pre-1983 pumpage simulated at the Savage and Keyes wells with the calibrated model. Analysis of simulated heads and the ground-water budget show that (1) withdrawals at the Savage well lower heads over a much larger area than at the Keyes well (fig. 27); (2) withdrawals at the Savage well induce infiltration from the discharge ditch; (3) the Savage well receives a larger component of its pumped water from stream losses (47 percent) than the Keyes well (30 percent) at average pumping rates; (4) the source of water pumped at the Keyes well is to a large extent (70 percent) recharge by infiltrating precipitation and boundary flow (water from lateral till seepage at the model boundary); and (5) withdrawals at the Keyes well affect ground-water heads on both side of the river, as indicated by the configuration of simulated heads around the well (figs. 27 and 28).

Withdrawals affect the local-flow systems differently at the Savage and Keyes wells because of the differences in well proximity to the Souhegan River and the differences in horizontal hydraulic conductivity of the aquifer at the two well fields. The combined effect of higher horizontal hydraulic conductivity and greater distance to the Souhegan River at the Savage well causes ground-water heads to decline over a larger area than at the Keyes well.

Sectional head profiles through model columns containing the Savage and Keyes wells, shown in figures 29 and 30, illustrate the effect of withdrawals on heads. Ground-water heads are affected over most of the cross-section through the Savage well but not at the Keyes well. Ground-water withdrawals create a ground-water divide between the Savage well and the Souhegan River (fig. 30). Ground-water-head profiles also indicate that flow is essentially horizontal for simulations of nonpumping and pumping at both wells. The simulations show that steep vertical head gradients are present only within 200 ft of the wells.

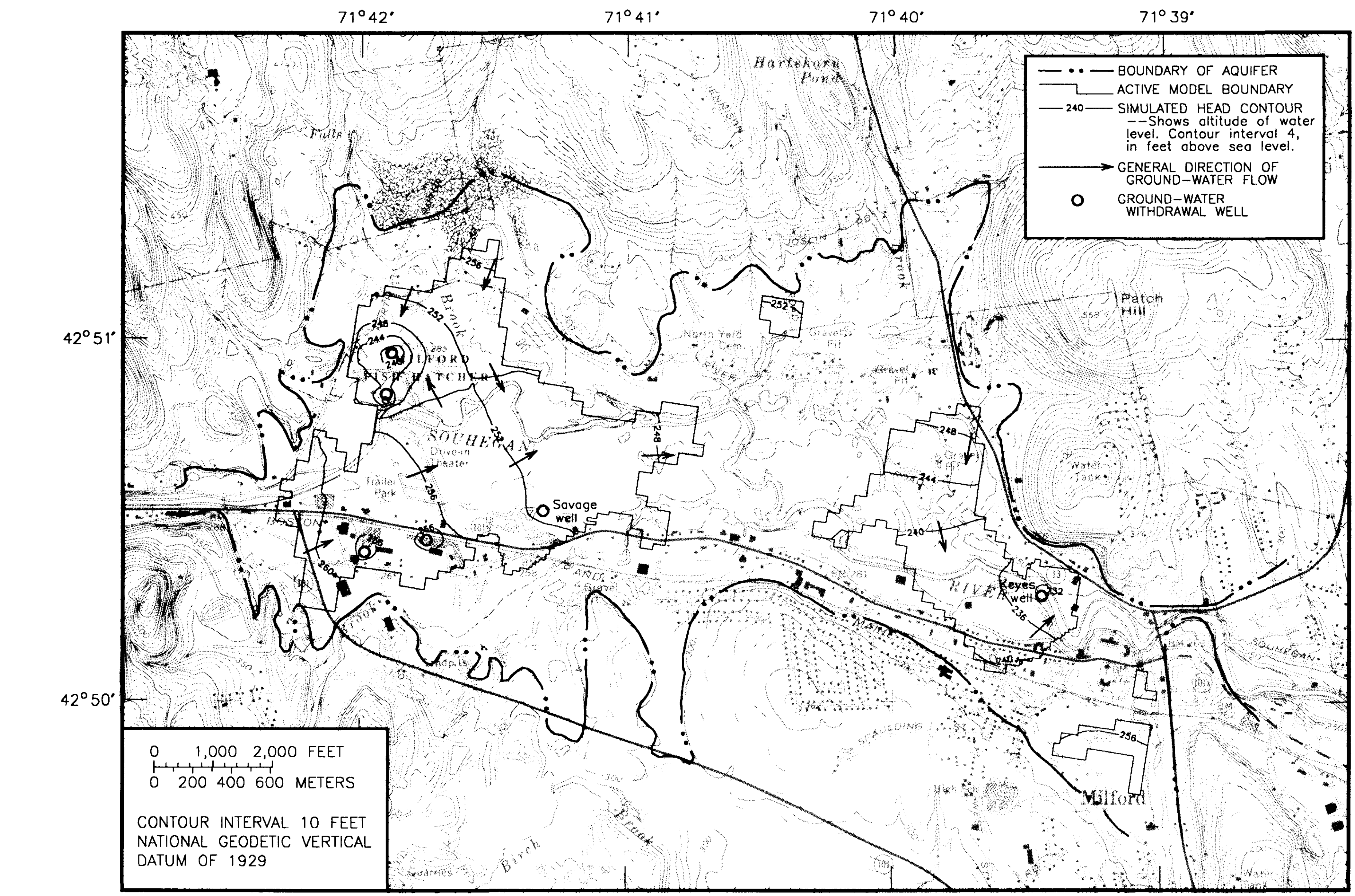
Estimates of Contributing Areas to Supply Wells

The contributing recharge area of a pumped well determines the source of water to that well. The sources of water to a pumped well in a natural ground-water-flow system are water stored in the aquifer, induced infiltration from streams, and captured discharge (ground water that would have discharged to streams had the wells not been pumped). In the steady-state representation of the Milford-Souhegan ground-water-flow system, the source of water, or recharge, to a pumped well can be derived from only three sources--induced stream infiltration, infiltrating precipitation, and boundary flow. Captured discharge in a steady-state model consists of infiltrating precipitation and boundary flow. A decrease in one source must be balanced by an increase in the other to maintain the same rate of withdrawal.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 27.--Simulated steady-state heads in the Milford-Souhegan aquifer with simulated pumpage at the Savage and Keyes wells: Layer 1.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 28.--Simulated steady-state heads in the Milford-Souhegan aquifer with simulated pumpage at the Savage and Keyes wells: Layer 3.

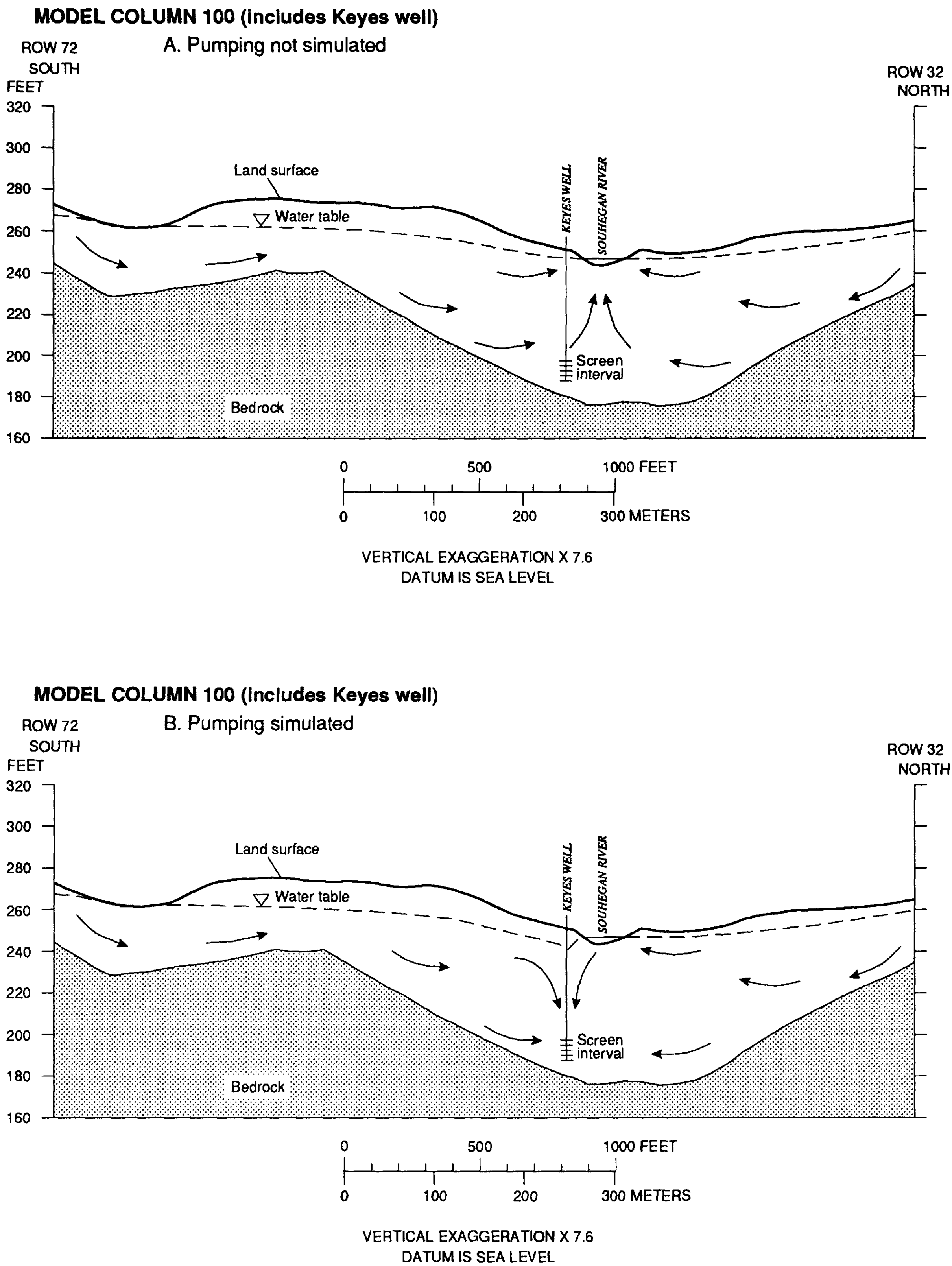


Figure 29.--Generalized hydrogeologic section showing simulated ground-water flow paths through model column 100 at the Keyes well.

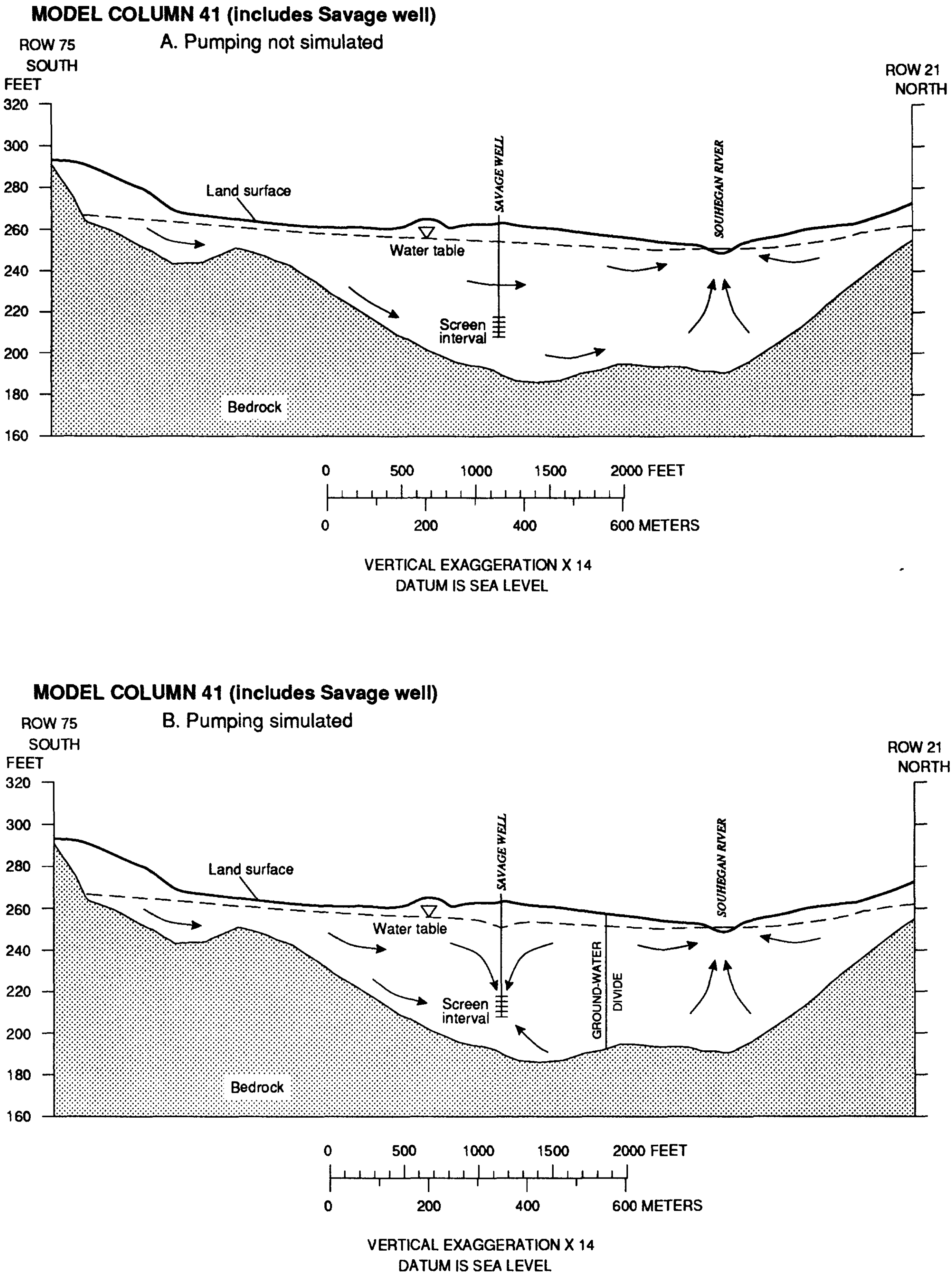


Figure 30.--Generalized hydrogeologic section showing simulated ground-water flow paths through model column 41 at the Savage well.

The extent and possible range in sizes of contributing areas for the Savage and Keyes wells were determined by means of 10 simulations. The first two simulations were done to determine the effects of variation of pumping rates. In the first simulation, the two discontinued public-supply wells were pumped at daily average rates, and all hydraulic parameters were the same as calibrated-model values. In the second simulation, parameter values remained the same but the two wells were pumped at a maximum daily rate. For the remaining eight simulations, pumping at the Savage and Keyes wells was held constant at rates representing daily means. The wells were normally pumped for only a part of a day (8 hours); therefore, the typical pumping rate (the instantaneous rate) was averaged over 24 hours to obtain a daily mean pumping rate for steady-state simulation. The latter eight simulations can be compared to the first simulation (hereafter termed base simulation) to examine the effects of varying hydrologic parameters independently.

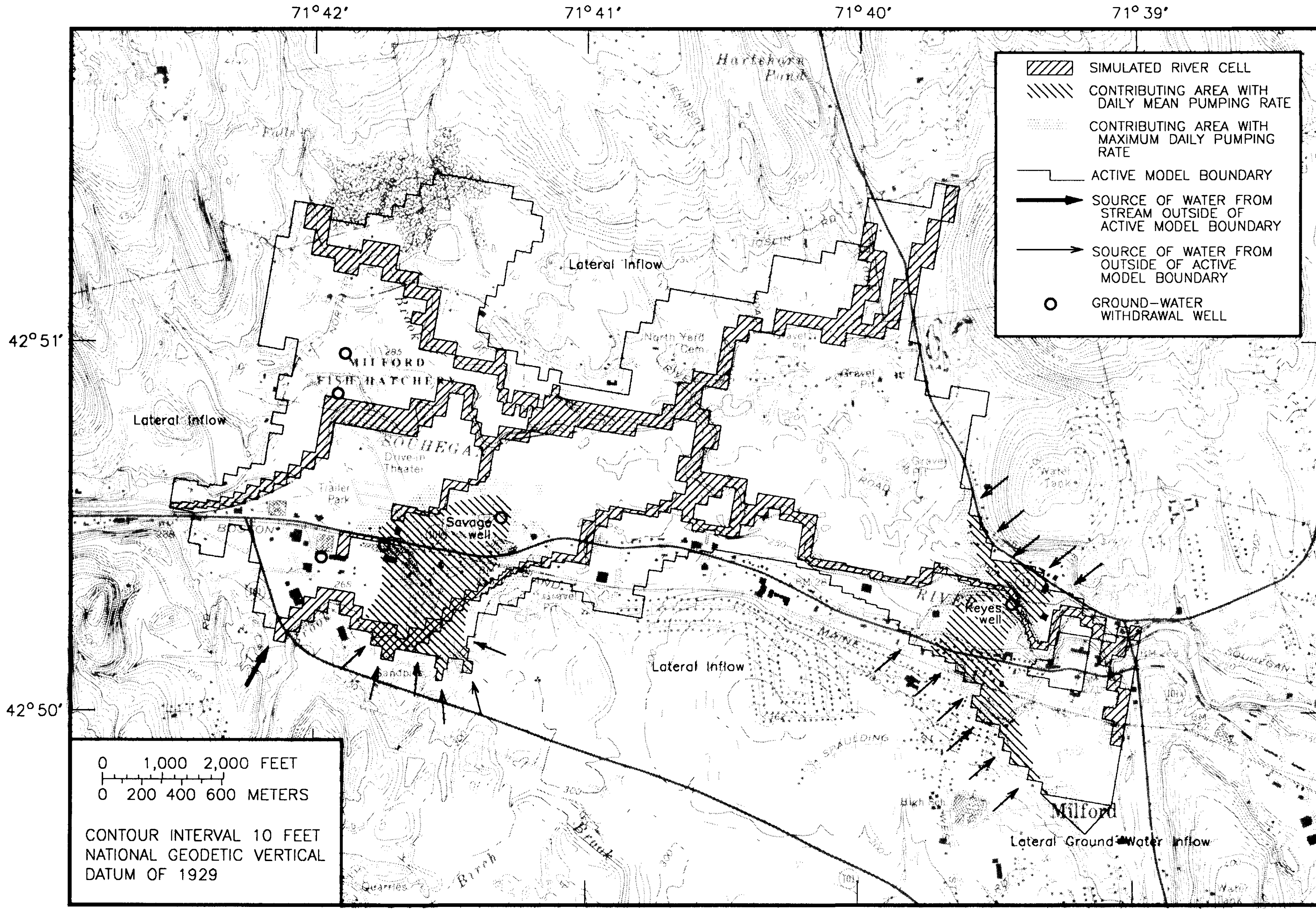
The simulations were similar to those done during the sensitivity analysis of the calibrated model, with the exception of the pumping at the Savage and Keyes wells. Recharge, horizontal hydraulic conductivity, and streambed conductance were increased and decreased by 50 percent of the calibrated-model values. Vertical hydraulic conductivity was increased and decreased by one order of magnitude.

Results of the two initial simulations and the effect of varying four parameters individually are shown in figures 31-35. The areal extent of the contributing areas produced by varying each parameter are listed in table 16. The areas shown (figs. 31-35), and listed in table 16, represent only the contributing areas within the simulated aquifer. Contributing areas should be considered with caution where they are in contact with the model boundary (figs. 31-35). At these locations the contributing areas extend into uplands or adjacent aquifers, which drain to that section of the model boundary. A small part of the contributing areas of the Keyes well extends past the model boundary into the Great Brook aquifer. Contributing areas also extend past delineated areas if the contributing area intersects streams; under these conditions, the drainage area to that stream reach contributes some of the water that flows to the well. Examination of such areas, although they contribute water to wells, was outside the scope of this investigation.

The sources of water to and extent of the contributing areas of the Savage and Keyes wells are governed by aquifer geometry and characteristics and proximity of the well to hydrologic boundaries, such as streams and whether the well is in a ground-water recharge or discharge area. The southwestern extent of the contributing area of the Savage well is controlled by ground-water withdrawals from wells 47 and 49.

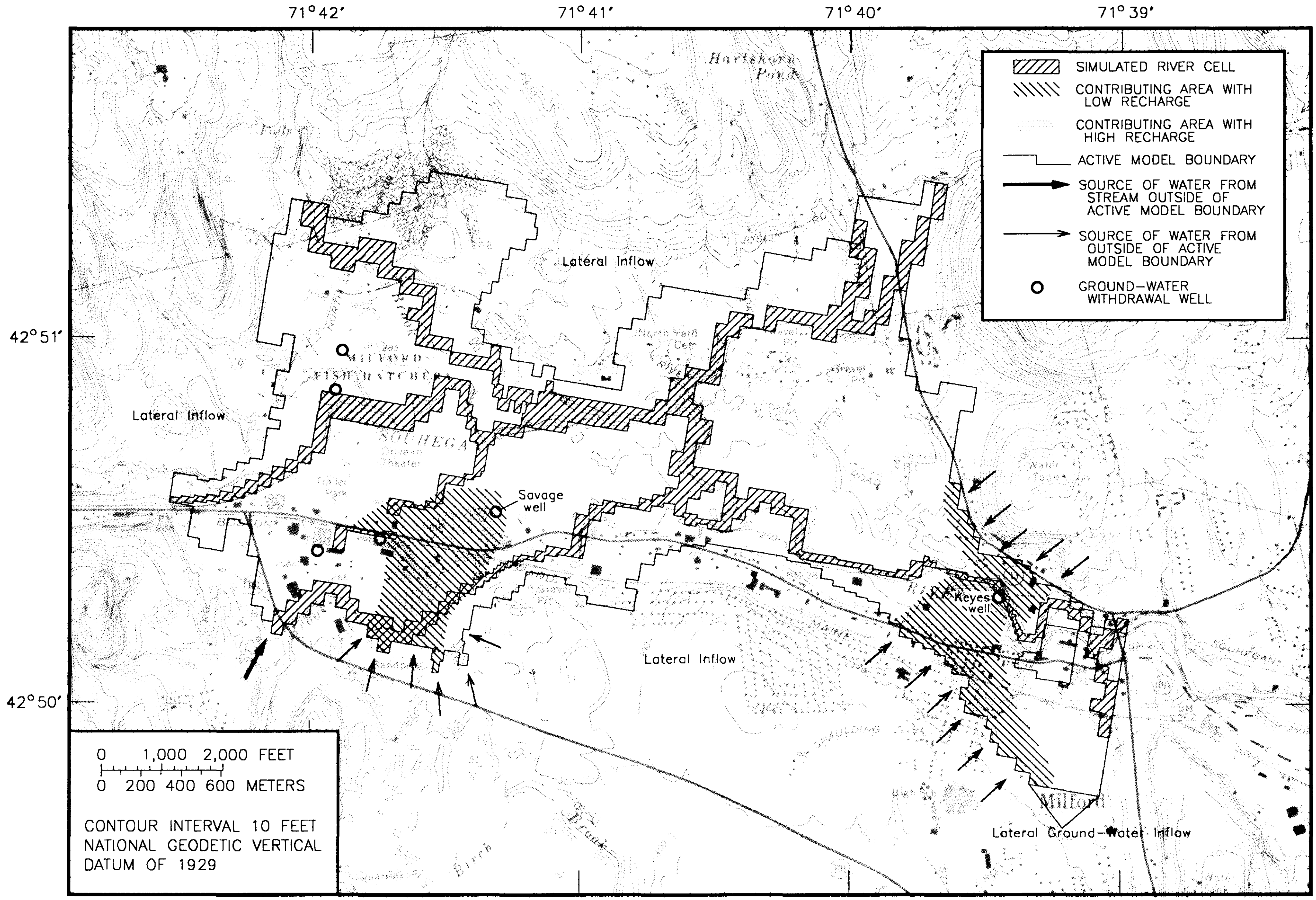
At mean pumping rates (0.323 ft^3/s or 145.0 gal/min), the contributing area of the Savage well covers 0.148 mi^2 (table 16) and is confined to the area between the discharge ditch, Tucker Brook, and the southern model boundary (fig. 31). The configuration of this area indicates that the discharge ditch and Tucker Brook, because they lose water, contribute water to the Savage well. The upgradient, western limit of the contributing area of the Savage well is confined by the downgradient limit of the contributing area to wells 47 and 49. At mean pumping rates (0.223 ft^3/s or 100.1 gal/min), the contributing areas to the Keyes well occupies a narrow band, trending north-south across the aquifer to the valley-wall model boundaries (fig. 31). The contributing area of the Keyes well is not bounded by the effects of the Souhegan River. Size and configuration of the Keyes contributing area indicates that the rate of induced infiltration is less than the rate of recharge from other sources to the well.

Simulated variations in pumping rate had an expected and significant effect on the size of the simulated contributing areas (fig. 31). When simulated pumpage was increased to the maximum rates (0.97 ft^3/s or 435.4 gal/min) the contributing area of the Savage well increased to 174 percent of its original size to cover 0.258 mi^2 within the modeled area (table 16). This contributing area extends beyond Tucker Brook and the discharge ditch, and crosses the Souhegan River at the western edge of the modeled area. At maximum pumpage rates, ground-water recharge at the western model boundary follows a deep flow path and passes underneath the Souhegan River and contributes water to the Savage well. The aquifer underlying the Souhegan River, near the western model boundary, is thick and enables ground water from the western model boundary to flow underneath the Souhegan River instead of discharging to it. Model results of maximum pumpage at the Savage well indicate that seepage losses from the Souhegan River do not directly contribute water to the well. Because the model is not calibrated to maximum pumping rates of the Savage well, further calibration is suggested to evaluate model predictability. The contributing area of the Keyes well is also significantly increased at the maximum pumping rate to 181 percent of its original size to an area of 0.186 mi^2 (table 16). At



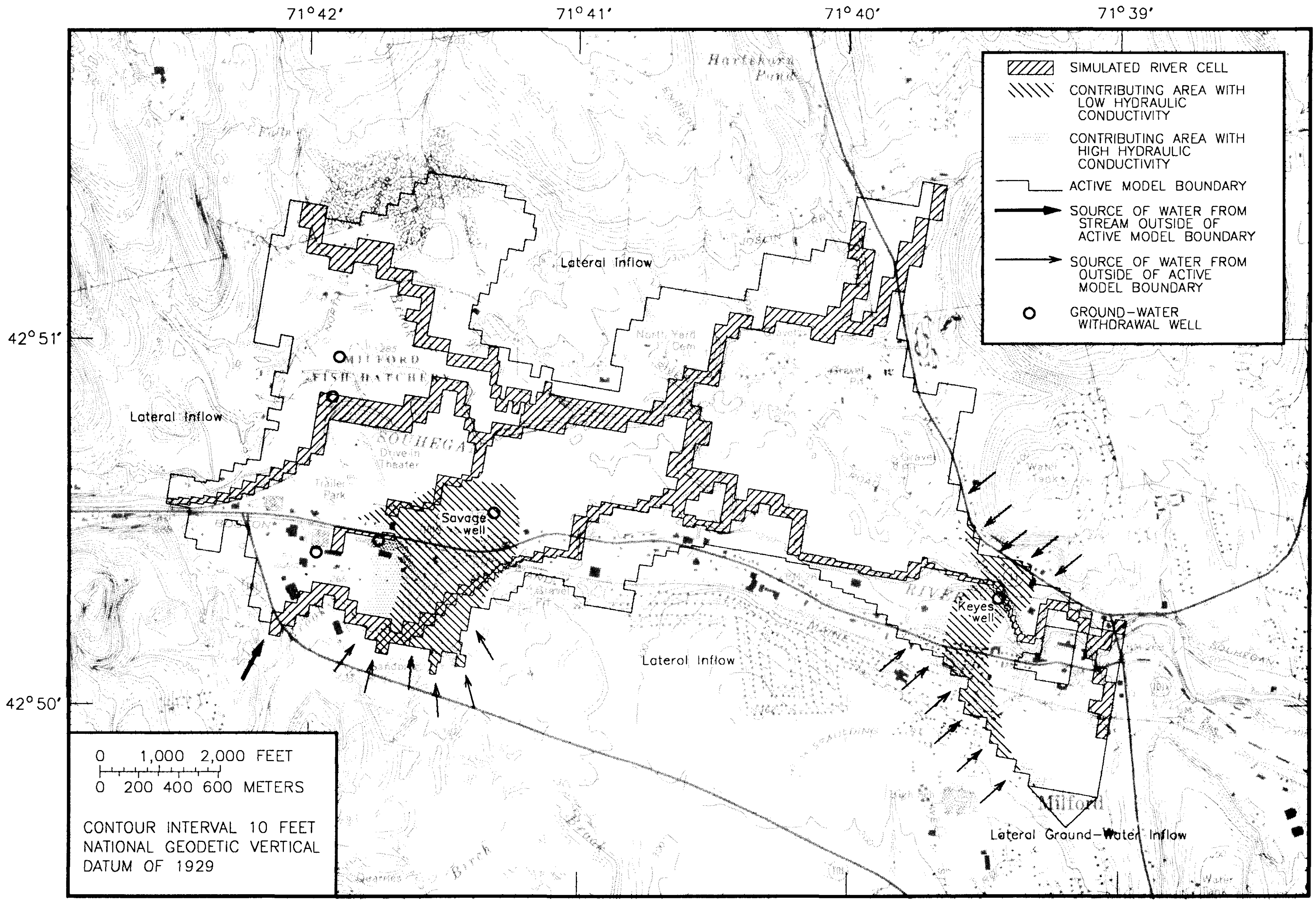
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 31.--Possible ranges in contributing areas of the Savage and Keyes wells in relation to variation in pumping rate.



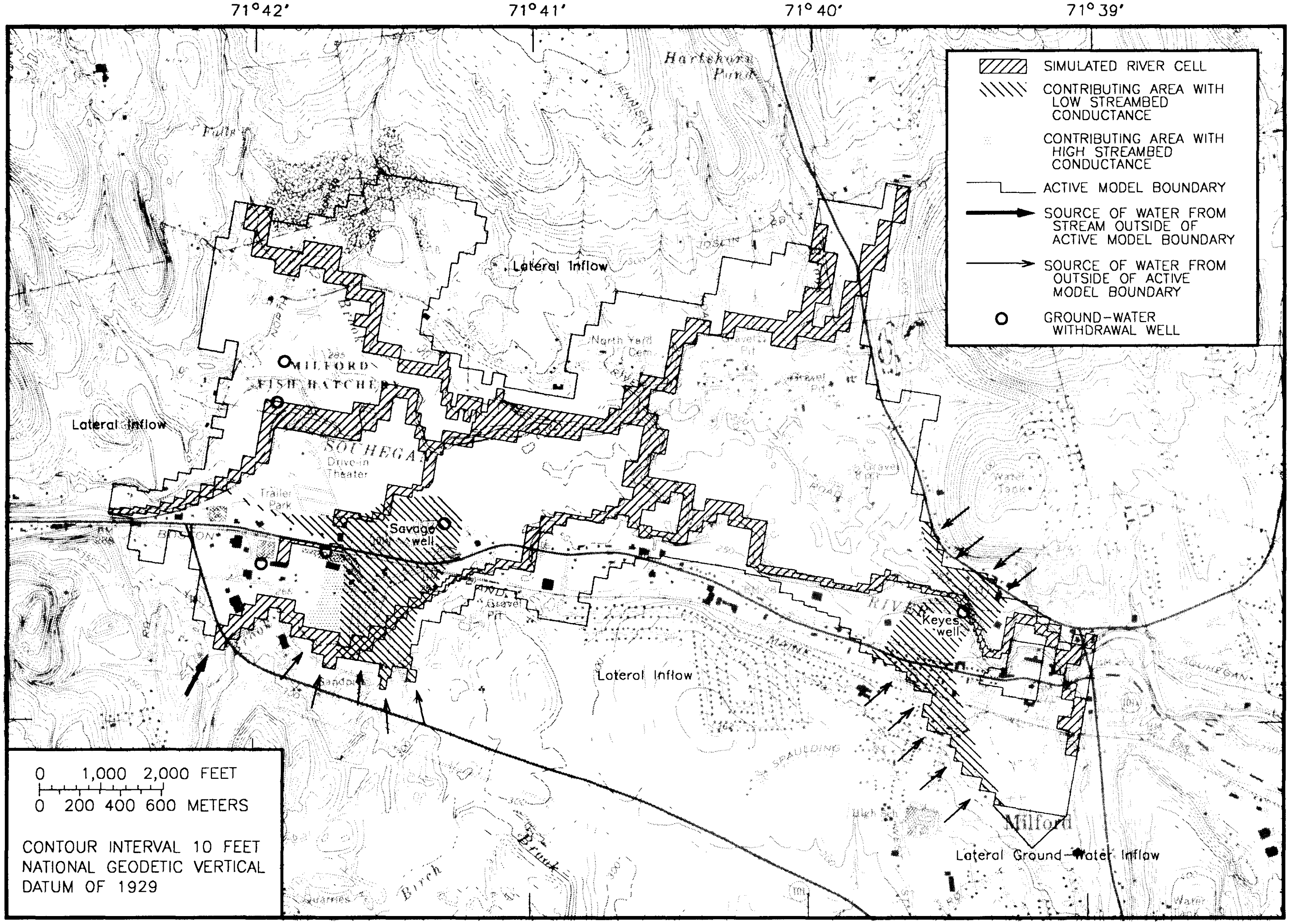
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 32.--Possible ranges in contributing areas of the Savage and Keyes wells in relation to variation in recharge.



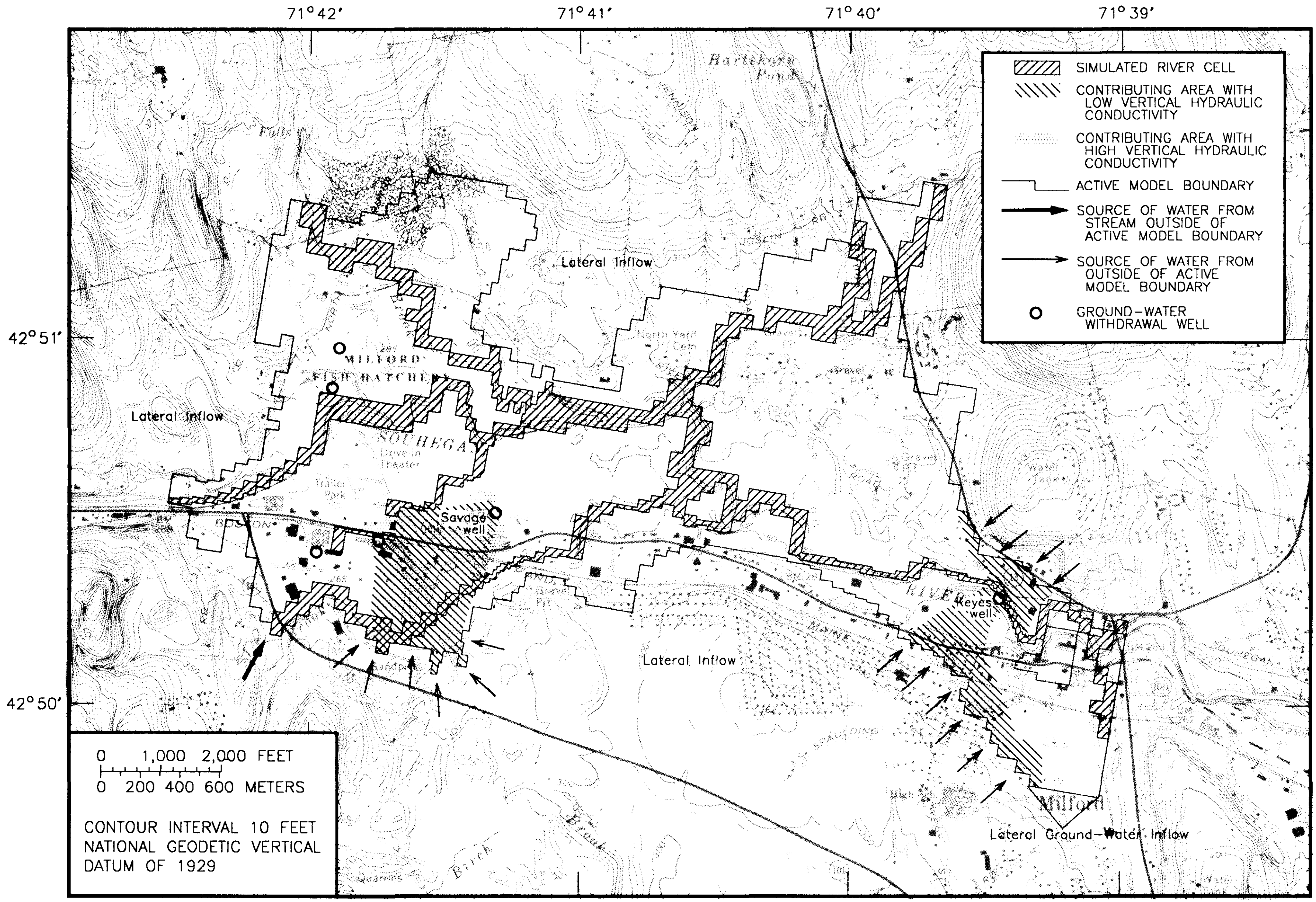
Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 33.--Possible ranges in contributing areas of the Savage and Keyes wells in relation to variation in horizontal hydraulic conductivity.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 34.--Possible ranges in contributing areas of the Savage and Keyes wells in relation to variation in streambed conductance.



Base from U.S. Geological Survey
Milford, N.H. 1:24,000, 1968, photorevised 1985

Figure 35.--Possible ranges in contributing areas of the Savage and Keyes wells in relation to variation in vertical hydraulic conductivity.

Table 16.--*Range of contributing areas of the Savage and Keyes wells produced by varying pumping rate and other model parameters*

[mi^2, square miles]

Parameter	Contributing area: Savage well, Area (mi^2)	Savage well, Percentage of base simulation area	Keyes well, Area (mi^2)	Keyes well, Percentage of base simulation area
Base simulation pumping rate (0.323 ft^3/s (145.0 gal/min) at Savage well, 0.223 ft^3/s (100.1 gal/min) at Keyes well)	0.148	100	0.103	100
Maximum pumpage (0.97 ft^3/s (435.3 gal/min) at Savage well, 0.67 ft^3/s (300.7 gal/min) at Keyes well)	.258	174	.186	181
Recharge × 0.5	.159	107	.163	175
Recharge × 1.5	.141	95	.076	74
Horizontal hydraulic conductivity × 0.5	.167	113	.085	82
Horizontal hydraulic conductivity × 1.5	.129	87	.112	108
Streambed hydraulic conductance × 0.5	.157	106	.111	108
Streambed hydraulic conductance × 1.5	.152	103	.100	97
Vertical hydraulic conductivity × 0.1	.144	97	.110	107
Vertical hydraulic conductivity × 10	.150	109	.086	83

increased pumping rates, the contributing areas of both wells included more area representing ground-water inflow at the model boundary and additional parts of stream drainage basins in the adjacent uplands.

Varying aquifer recharge by ±50 percent of the calibrated-model values resulted in substantially different contributing areas at the Keyes well but only minor differences at the Savage well (fig. 32, table 15). Decreasing recharge by 50 percent caused the contributing area of the Keyes well to increase in size by 75 percent. For the same simulations, the contributing area to the Savage well increased by only 7 percent. This is probably because recharge sources to the ground-water system are markedly different near the two wells. The southwest part of the aquifer is recharged primarily by infiltration of the Souhegan River, Tucker Brook, and the discharge ditch; thus, the Savage well is less affected by variations in recharge from precipitation. Streamflow losses from the industrial discharge ditch and Tucker Brook help sustain withdrawals at the Savage well; streamflow losses from the Souhegan River help sustain withdrawals from production wells 47 and 49 and production well 87 (fig. 3 and 27). The east part of the aquifer, however, is largely a ground-water discharge area dominated by gaining stream reaches, and primarily recharged by infiltrating precipitation and lateral inflow at model boundaries; thus, the Keyes well is affected by variations in precipitation recharge.

Variations in horizontal hydraulic conductivity affected the contributing areas of the Savage and Keyes wells differently (fig. 33). Adjustments in the horizontal hydraulic conductivity produced an atypical response for the Savage well with an increase in hydraulic conductivity, and for the Keyes well with a decrease in hydraulic conductivity. By increasing hydraulic conductivity 50 percent from the calibrated model, a 13-percent decrease in contributing area resulted at the Savage well but an 8-percent increase in contributing area resulted at the Keyes well (table 16). The decrease in size of the Savage contributing area is caused by an increase in seepage losses along the discharge ditch in drainage basin 8 and Tucker Brook in drainage basin 9 (fig. 25). A similar increase in simulated seepage losses from basin 8 and 9 was noted during the sensitivity analysis of the calibrated model (table 15). Decreasing hydraulic conductivity by 50 percent conversely affected the size of the contributing areas (table 16). The decrease in size of the Keyes contributing area is caused by an increase in induced infiltration from the Souhegan River.

Changes in streambed conductance produced slight changes in contributing areas of the two wells (fig. 34). Varying streambed conductance by 50 percent caused the contributing area to vary from +6 to +3 percent at the Savage well and from -3 to +8 percent at the Keyes well (table 16). It is not apparent why the Savage contributing area increased in size with a 50-percent increase in streambed hydraulic conductance.

Changes in vertical hydraulic conductivity had a greater effect on contributing areas than did streambed conductance. Variation of vertical hydraulic conductivity affected the Keyes contributing area more than the Savage contributing area (table 16). The contributing area to the Keyes well (fig. 35) increased in size by 7 percent, to cover 0.110 mi^2 (table 15) when vertical hydraulic conductivity was decreased by an order of magnitude and decreased in size by 17 percent, when vertical hydraulic conductivity was increased by an order of magnitude. This is probably because the aquifer in the Keyes well area is narrow and very limited areally; therefore, this well must receive a large part of its pumped water by way of vertical flow.

In summary, pumping rate had the greatest effect on the size and shape of contributing areas to the two discontinued public-supply wells. Variations of the other model parameters examined--recharge, horizontal hydraulic conductivity, vertical hydraulic conductivity, and streambed hydraulic conductance--had different effects on the contributing areas of the two wells. Recharge did not have as substantial an effect on the contributing area of the Savage well as at the Keyes well; this indicates that the Savage well receives much of its water from streamflow losses. At the Keyes well, in contrast, all parameters investigated had substantial effects on the contributing area. The most notable effects on the Keyes well contributing area were produced by varying recharge and horizontal hydraulic conductivity. Streambed conductance was more influential at the Keyes well than at the Savage well because of the different ground-water-flow systems in the two areas. At the Keyes well, infiltration is induced through fewer river cells than at the Savage well; therefore, streambed conductance has a greater influence near the Keyes well than near the Savage well. The Savage well, however, is in an area where the aquifer is recharged by the Souhegan River and its tributaries. The aquifer parameters and boundary conditions that affect the size and shape of the contributing area of a pumped well are highly dependent on the nature of the ground-water-flow system (particularly aquifer geometry), the proximity of the well to aquifer boundaries, and whether the well is in an area of ground-water recharge or discharge.

Further research is suggested to evaluate the transient dynamics of contributing areas caused by seasonal variations in recharge. The contributing areas to the Savage and Keyes wells may be particularly sensitive to variations in seasonal recharge. Further research could be directed toward investigation of contaminant transport in the aquifer, in as much as attenuation of contaminants by dispersion and chemical reactions was not addressed in this study.

SUMMARY AND CONCLUSIONS

The Milford-Souhegan aquifer consists of as much as 114 ft of unconsolidated glacial sediments in a buried pre-Pleistocene valley, and has a saturated thickness of more than 100 ft. The aquifer is composed predominantly of sand and gravel interbedded with silt; deposits generally are finer in the eastern part than in the western part. Horizontal hydraulic conductivity of stratified-drift deposits ranges from approximately 1 to 1,000 ft/d.

Ground-water flow is controlled by stream-aquifer interactions because of the close hydraulic connection between the Souhegan River, its tributaries, and the aquifer. In the western reaches of the Souhegan River, the river recharges the aquifer and ground-water flow is away from the river. In the eastern reaches of the Souhegan River, ground water discharges to the river and ground-water flow is towards the river.

Total recharge to the 3.3-mi^2 aquifer, based on October 1988 streamflow data, is estimated to have been 5.31 ft^3/s, or the equivalent of 21.8 in/yr. Estimates of the major sources of recharge to the aquifer are: 3.19 ft^3/s from infiltration of precipitation, 1.44 ft^3/s from surface-water infiltration, and 0.64 ft^3/s from lateral inflow from upland areas. The recharge rate in the till-covered upland areas is estimated to have been 0.205 $(ft^3/s)/mi^2$ during low flow in October 1988.

Ground-water withdrawals in the Milford-Souhegan aquifer were approximately 5 ft^3/s in 1988. Most withdrawals are in the western part of the aquifer. A major component of the withdrawals are for the Milford Fish Hatchery and are sustained primarily by induced infiltration from streamflow.

A variable-grid, 5-layer, finite-difference model of the Milford-Souhegan aquifer was constructed to simulate three-dimensional ground-water flow. The ground-water-flow model was calibrated to hydrologic conditions in October 1988, which are assumed to be at steady-state. The ground-water-flow model was used to simulate ground-water heads, stream-aquifer fluxes, and ground-water-flow directions and rates in the Milford-Souhegan aquifer. A semianalytical particle-tracking program, which incorporates the flow-model results, was used to delineate contributing areas of two discontinued public-supply wells. Ground-water withdrawals from these wells were discontinued after volatile organic compounds (TCE and PCE) were found in concentrations exceeding USEPA recommended levels.

Simulations of October 1988 conditions suggest that ground-water flow is primarily horizontal except within 200 ft of major production wells. Regional flow (flow within the aquifer) is affected by variations in recharge and horizontal hydraulic conductivity. Local flow, flow near ground-water withdrawal wells, is affected by variations in vertical hydraulic conductivity, whereas regional flow in the aquifer is not.

Simulated pumping of the Savage and Keyes public-supply wells revealed that the effects of pumping on the ground-water-flow system are highly dependent on the nature of the flow system and the characteristics of the aquifer at each well site. Simulated pumpage at the public-supply wells indicates that (1) the area of influence at the Savage well is larger than that at the Keyes well, (2) the Savage well captures 47 percent of its pumped water from surface-water infiltration along the discharge ditch and Tucker Brook, (3) the Savage well receives 53 percent of its pumped water from infiltrating precipitation and lateral inflow at model boundaries from till-bedrock uplands, and (4) the Keyes well receives 70 percent of its recharge from infiltrating precipitation and lateral inflow.

Contributing areas of the simulated Savage and Keyes wells vary with changes in pumping rate, recharge rates, and hydraulic properties of the aquifer and streambeds. Aside from variations caused by the pumping rate, estimates of the contributing area of the Savage well ranged from 0.129 mi^2 for a 50-percent increase in the estimated value of horizontal hydraulic conductivity to 0.162 mi^2 for a 50-percent decrease in the estimated value of horizontal hydraulic conductivity. Estimates of contributing area of the Keyes well ranged from 0.076 mi^2 for a 50-percent increase in estimated value of recharge to 0.163 mi^2 for recharge reduced by 50 percent of estimated values. Sensitivity analyses showed that the importance of stream hydraulic characteristics on the recharge area to a well is inversely related to the amount of stream length in contact with the contributing areas.

Variations in pumping rate had a substantial effect on the contributing area of each well. Increasing the simulated pumping rate to maximum short-term rates (0.97 ft^3/s (435.4 gal/min) at Savage and 0.67 ft^3/s (300.7 gal/min) at Keyes) produced even larger contributing areas, 0.258 and 0.186 mi^2 respectively, for the Savage and Keyes wells, than did variations in the other parameters.

In summary, the parameters and boundary conditions that affect the size and shape of the contributing area of a pumped well are dependent on the nature of the ground-water-flow system. Aquifer geometry, proximity to boundaries, and location of the pumped well in relation to ground-water recharge or discharge areas are important factors in determining the size and extent of the contributing area of a well.

SELECTED REFERENCES

Cooper, H.H., and Jacob, C.E., 1946, A generalized graphical method for evaluating formation constants and summarizing well field history: American Geophysical Union Transcript, v. 27, p. 526-534.

Cotton, J.E., 1977, Availability of ground water in the lower Merrimack River basin, southern New Hampshire: U.S. Geological Survey Water-Resources Investigations Report 77-69, 1 sheet, scale 1:125,000.

D.L. Maher Company, 1985, Ground water exploration investigation at the Milford Fish Hatchery, Milford, New Hampshire: North Reading, Mass., 7 p.

----- 1988, Construction report for Well No. 5, Milford Fish Hatchery, Milford, New Hampshire: North Reading, Mass., 4 p.

Goldberg, Zoino & Associates, Inc., 1986, Hydrogeologic study and transformer inventory, Milford Elm Street trust site, Milford, New Hampshire: Manchester, N.H., File no. D-5224, 12 p.

HMM Associates, Inc., 1989, Draft remedial investigation, Savage well site, Milford, New Hampshire: Concord, Mass., no. 2176/HAZ/2880, 218 p.

Howard, Needles, Tammen & Bergendoff, 1973, Comprehensive regional water quality management plan: Nashua Regional Planning Commission, 165 p.

Jacob, C.E., 1946, Radial flow in a leaky artesian aquifer: American Geophysical Union Transcript, v. 27, no. 2, p. 198-205.

Koteff, Carl, 1970, Surficial geologic map of the Milford quadrangle, Hillsborough County, New Hampshire: U.S. Geological Survey Geologic Quadrangle Map GQ-881, scale 1:62,50.

Kruseman, G.P., and de Ridder, N.A., (5th ed.) 1983, Analysis and evaluation of pumping test data: Wageningen, The Netherlands, International Institute for Land Reclamation and Improvements/ILRI, 189 p.

Lohman, S.W., 1979, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Lyons, J.B., and Bothner, W.A., 1989, Transect through New England Appalachians: 28th Annual International Geological Congress, Guidebook T162, 64 p.

Lyford, F.P., and Cohen, A.J., 1988, Estimation of water available for recharge to sand and gravel aquifers in the glaciated northeastern United States, *in* Regional Aquifer Systems of the United States: the Northeast Glacial Aquifer Series, American Water Resources Association Monograph Series no. 11, p. 37-54.

MacNish, R.D., and Randall, A.D., 1982, Stratified-drift aquifers in the Susquehanna River basin, New York: New York Department of Environmental Conservation Bulletin 75, 68 p.

McClymonds, N.E., and Franke, D.L., 1972, Water-transmitting properties of aquifers on Long Island, New York: U.S. Geological Survey Professional Paper 627-E, 24 p.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water-flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 6, Chapter A1, 586 p.

Meinzer, O.E., 1923, Outline of ground-water hydrology with definitions: U.S. Geological Survey, Water-Supply Paper 494, 71 p.

Meyer, Rex, 1963, A chart relating well diameter, specific capacity, and the coefficients of transmissibility and storage, *in* Bentall, Ray, Methods of determining permeability, transmissibility, and drawdown: U.S. Geological Survey Water-Supply Paper 1536-I, p. 338.

Morrissey, D.J., 1989, Estimation of the recharge area contributing water to a pumped well in a glacial-drift, river-valley aquifer: U.S. Geological Survey Water-Supply Paper 2338, 41 p.

Myette, C.F., Olimpio, J.C., and Johnson, D.C., 1987, Area of influence and zone of contribution to superfund-site wells G and H, Woburn, Massachusetts: U.S. Geological Survey Water-Resources Investigations Report 87-4100, 21 p.

Nashua Regional Planning Commission, 1980, Update to Water Quality Inventory: 55 p.

National Oceanic and Atmospheric Administration, 1988, Climatological data annual summary, New England, 1988: Ashville, N.C., v. 100, no. 13, 52 p.

----- 1989, Hourly precipitation data, New England: U.S. Department of Commerce, 2 p.

Neuman, S.P., 1974, Effect of partial penetration on flow in unconfined aquifers considering delayed gravity response: Water Resources Research, v. 10, no. 2, p. 303-312.

New Hampshire Water Supply and Pollution Control Division, 1985, Hydrogeological investigation of the Savage well site, Milford, New Hampshire: Report 145, 91 p.

NUS Corporation, 1987, Final site-inspection report: Milford, N.H., Hampshire Paper Company, 23 p.

Olney, S.L., 1983, An investigation of the relationship between the coefficient of permeability and effective grain size of unconsolidated sands: Boston Mass., Boston University, unpublished Master's thesis, 61 p.

Pollock, D.W., 1988, Semianalytical computation of path lines for finite-difference models: Ground Water, v. 26, no. 6, p. 743-750.

----- 1989, Documentation of computer programs to compute and display pathlines using results from the U.S. Geological Survey modular three-dimensional finite-difference ground-water-flow model: U.S. Geological Survey Open File Report 89-391, 188 p.

R.E. Chapman Company, 1957a, Log of pump test, Mount Vernon Street, Town of Milford: Oakdale, Mass., 11 p.

----- 1957b, Log of pump test, Savage Farm, town of Milford: Oakdale, Mass., 13 p.

----- 1960, Log of pump test, Savage Farm, town of Milford: Oakdale, Mass., 6 p.

----- 1968, Log of pump test, Ford Garage, town of Milford: Oakdale, Mass., 4 p.

----- 1972, Log of pump test, Keyes Field, town of Milford: Oakdale, Mass., 4 p.

----- 1981, Log of pump test, Savage Farm, town of Milford: Oakdale, Mass., 6 p.

Roy F. Weston, Inc., 1984, Report of findings, hydrogeologic and water quality investigations, Hitchiner Manufacturing Co., Milford, New Hampshire: Concord, N.H., 33 p.

----- 1987, Report of findings, hydrogeologic investigation, AMP Incorporated, Milford, New Hampshire: Concord, N.H., 31 p.

Toppin, K.W., 1987, Hydrogeology of stratified-drift aquifers and water quality in the Nashua Regional Planning Commission area of south-central New Hampshire: U.S. Geological Survey Water-Resources Investigations Report 86-4358, 45 p.

Trescott, P.C., Pinder, G.F., and Larson, S.P., 1976, Finite-Difference Model for aquifer simulation in two dimensions with results of numerical experiments: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 7, Chapter C1, 116 p.

U.S. Environmental Protection Agency, 1986, Amendment to national oil and hazardous substance contingency plan; national priorities list; final rule and proposed rules: Federal Register, v. 51, no. 111, p. 21053-21112.

U.S. Weather Bureau, 1964, Dicennial census of United States climate, climatic summary of the United States, supplement for 1951 through 1960: U.S. Department of Commerce, 141 p.

Walton, W.C., 1962, Selected analytical methods for well and aquifer evaluation: Illinois State Water Survey Bulletin v. 49, 81 p.

APPENDIX A

Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire

Appendix A.--*Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire*

[ft, feet; --, no data. Till thickness (not shown) is the difference between depth to base of stratified drift and the depth to bedrock]

Well number	Local identifier	Land-surface elevation (ft)	Data code[a]	Base of stratified drift (ft below land surface)	Depth to bedrock (ft)	Depth drilled (ft)	Depth of screen Top (ft)	Depth of screen Bottom (ft)	Saturated thickness (ft)
1	KEYES 1	248.7	1	78.0	85.0	85.0	53.0	55.0	72.9
2	KEYES 2D	246.6	3	--	--	57.0	54.5	56.5	--
3	KEYES 3D	244.8	3	--	--	55.0	48.7	50.7	--
4	KEYES 4D	243.3	3	--	--	53.0	49.9	51.9	--
5	LW-01D	264.8	7	85.0	114.0	124.3	100.0	110.0	91.2
6	LW-02D	243.1	1	62.0	62.0	73.5	45.0	55.0	56.6
7	LW-03D	247.3	1	80.0	80.0	90.2	44.5	54.5	69.7
8	LW-04D	243.4	1	80.0	80.0	90.0	40.0	50.0	73.4
9	M0W-33	260.0	4	--	--	52.0	--	--	--
10	GW-02D	255.4	1	34.0	34.0	44.0	19.0	29.0	--
11	GW-03D	252.4	1	23.0	23.0	38.0	28.0	38.0	--
12	GW-04D	255.6	1	19.0	19.0	31.5	21.5	31.5	--
13	GW-05D	261.0	7	33.0	33.0	48.0	23.0	33.0	22.0
14	RFW-1	256.0	1	28.0	28.0	28.0	8.0	28.0	26.3
15	RFW-2	254.2	1	35.0	35.0	35.0	10.0	35.0	30.9
16	RFW-3	254.5	1	43.0	43.0	43.0	13.0	43.0	38.1
17	RFW-4	252.1	1	16.0	16.0	16.0	6.0	16.0	13.6
18	PA-1	258.3	4	--	--	11.5	--	8.7	--
19	PA-2	255.5	4	--	--	11.0	--	8.7	--
20	PA-3	259.1	4	--	--	11.5	--	7.8	--
21	MI-7	255.4	0	--	--	--	--	--	--
22	MI-8	261.9	0	--	--	--	--	--	--
23	MI-10	252.1	1	59.0	59.0	59.0	44.0	47.0	54.3
24	MI-11	252.9	1	63.0	63.0	63.0	40.0	56.0	57.1
25	MI-12	251.6	1	50.0	50.0	50.3	43.0	49.0	43.7
26	MI-15	266.5	0	--	--	--	--	--	--
27	MI-16	269.1	0	--	--	--	--	--	--
29	M0W-36	260.0	3	--	--	14.6	--	--	--
30	MI-19	275.6	1	25.0	62.0	82.5	65.0	80.0	51.2
31	MI-20	275.6	3	--	--	82.5	10.0	40.0	--
32	MI-20A	274.7	4	--	--	14.8	--	--	--
33	MI-21	273.0	6	30.0	--	53.0	15.0	40.0	--
34	MI-21A	270.0	0	--	--	--	--	--	--

Appendix A.--*Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire*--Continued

Well number	Local identifier	Land-surface elevation (ft)	Data code[a]	Base of stratified drift (ft below land surface)	Depth to bedrock (ft)	Depth drilled (ft)	Depth of screen Top (ft)	Depth of screen Bottom (ft)	Saturated thickness (ft)
35	MI-22	270.0	1	75.0	94.0	112.5	99.0	114.0	84.2
36	MI-22A	270.1	4	--	--	11.7	--	--	--
37	MI-23	270.0	1	75.0	94.0	112.5	10.0	75.0	84.9
38	MI-24	270.6	1	77.0	96.0	101.5	10.0	85.0	85.8
39	MI-24A	271.7	4	--	--	14.0	--	--	--
40	MI-25	270.6	1	57.0	104.0	110.0	101.8	111.0	93.7
41	MI-26	270.6	1	57.0	104.0	110.0	8.0	88.0	93.7
42	MI-27	270.7	2	57.0	86.0	92.0	13.0	78.0	75.5
43	MI-28	270.3	2	38.0	56.0	56.0	35.0	55.0	46.2
44	MI-30	265.4	6	70.0	--	75.0	27.0	72.0	--
45	MI-31	266.0	3	--	--	60.0	36.0	54.0	--
46	MI-32	270.2	3	--	--	95.0	30.0	75.0	--
47	MI-33	268.2	0	--	--	--	--	--	--
49	MI-35	265.9	0	--	--	--	--	--	--
50	MI-36	270.0	0	--	--	--	--	--	--
51	MI-37	270.6	0	--	--	--	--	--	--
52	MI-38	270.0	0	--	--	--	--	--	--
54	MI-41	258.6	4	--	--	20.0	--	--	--
55	MI-42	257.4	4	--	--	20.0	--	--	--
56	MI-43	257.2	4	--	--	20.0	--	--	--
57	M0W-63	270.0	2	65.0	65.0	69.0	53.0	62.0	53.6
58	MI-44	259.8	4	--	--	20.0	--	--	--
59	MI-45	264.9	0	--	--	--	--	--	--
60	MI-46	267.3	0	--	--	--	--	--	--
61	MI-47	270.0	0	--	--	--	--	--	--
62	MI-48	260.3	0	--	--	--	--	--	--
64	--	265.3	0	--	--	--	--	--	--
65	--	260.0	0	--	--	--	--	--	--
66	--	270.0	0	--	--	--	--	--	--
67	--	250.0	0	--	--	--	--	--	--
68	--	267.9	0	--	--	--	--	--	--
69	--	266.3	0	--	--	--	--	--	--
70	--	264.1	0	--	--	--	--	--	--
71	--	264.0	0	--	--	--	--	--	--
72	MI-62	260.0	2	58.0	60.7	60.7	17.0	58.0	55.0

Appendix A.--*Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire*--Continued

Well number	Local identifier	Land-surface elevation (ft)	Data code[a]	Base of stratified drift (ft below land surface)	Depth to bedrock (ft)	Depth drilled (ft)	Depth of screen Top (ft)	Depth of screen Bottom (ft)	Saturated thickness (ft)
73	MI-64	259.9	0	--	--	--	--	--	--
74	M0W-35	260.0	2	59.0	59.0	60.0	--	--	56.0
75	M0A-1	239.5	2	74.0	74.0	74.0	--	--	64.2
76	M0A-2	244.6	4	--	--	13.0	--	--	--
77	M0A-3	241.1	2	52.0	52.0	52.0	--	--	46.3
78	M0A-4	249.5	2	43.0	54.0	54.0	33.0	38.0	46.5
84	FH-1	268.0	3	--	--	66.0	51.0	66.0	--
85	FH-2	262.4	0	--	--	--	--	--	--
86	FH-3	260.0	0	--	--	--	33.0	43.0	--
87	FH-4	262.2	0	--	--	--	--	--	--
88	FH 85-1	261.0	4	--	--	26.0	--	--	--
89	FH 85-2	250.0	4	--	--	41.0	34.0	39.0	--
90	FH 85-3	252.8	4	--	--	31.0	24.0	29.0	--
91	FH 85-4	251.6	4	--	--	31.0	24.0	29.0	--
92	FH 85-5	252.3	4	--	--	31.0	24.0	29.0	--
93	FH 85-6	252.0	4	--	--	26.0	22.0	25.0	--
94	FH 85-7	253.5	4	--	--	31.0	21.0	26.0	--
95	FH 85-8A	260.0	4	--	--	26.0	20.0	26.0	--
96	FH 1974	254.5	0	--	--	--	--	--	--
97	B1	269.9	5	31.0	--	43.0	--	--	--
98	B3	269.3	5	34.0	--	34.0	--	--	--
99	B4	270.0	5	39.0	--	54.5	--	--	--
100	B6	269.0	1	26.2	26.2	26.2	--	--	24.8
101	B8	269.7	3	--	--	26.0	--	--	--
102	B9	275.3	5	36.0	--	40.3	--	--	--
103	B11	275.0	5	37.0	--	38.0	--	--	--
104	B12	275.4	5	42.0	--	48.4	--	--	--
122	WW-125	269.0	0	--	--	--	--	--	--
123	GW-01S	256.1	3	--	--	20.0	6.0	16.0	--
124	GW-01D	256.5	1	40.0	56.0	76.4	60.0	70.0	51.2
125	GW-01M	256.7	5	40.0	--	41.0	30.0	40.0	--
126	KEYES	240.1	3	--	--	60.0	50.0	60.0	--
127	HAYWOOD	256.3	0	--	--	--	--	--	--
128	SAVAGE	261.0	3	--	--	52.0	42.0	52.0	--
129	KEYES 1-	241.7	1 2	50.0	50.0	50.0	41.0	50.0	41.5
130	KEYES 2-	240.5	1 2	65.0	65.0	65.0	52.0	60.0	55.9

Appendix A.--*Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire*--Continued

Well number	Local identifier	Land-surface elevation (ft)	Data code[a]	Base of stratified drift (ft below land surface)	Depth to bedrock (ft)	Depth drilled (ft)	Depth of screen Top (ft)	Depth of screen Bottom (ft)	Saturated thickness (ft)
131	KEYES 3-	240.3	1 2	52.0	52.0	52.0	42.0	50.0	45.4
132	POTTER 1D	251.8	1	67.0	80.0	80.0	55.0	57.0	63.3
133	POTTER 2D	253.8	3	--	--	60.0	56.0	58.0	--
134	POTTER 3D	253.7	3	--	--	60.0	56.0	58.0	--
135	FORD 34	241.4	2	50.0	50.0	50.0	40.0	50.0	40.0
136	FORD OBS	247.1	3 1	46.0	46.0	46.0	--	--	33.0
137	FORD 33	240.0	2	40.0	40.0	40.0	--	--	27.0
138	FORD 32	240.0	2	42.0	42.0	42.0	32.0	42.0	29.0
139	FORD 1	239.8	2	47.0	50.0	50.0	35.0	50.0	37.3
140	FORD 5	241.7	2	35.0	35.0	35.0	--	--	23.2
141	FORD 4	245.3	2	47.0	47.0	47.0	--	--	34.7
142	KEYES 2S	246.1	3	--	--	57.0	18.0	20.0	--
143	KEYES 3S	246.0	3	--	--	55.0	16.6	18.6	--
144	KEYES 4S	244.3	3	--	--	53.0	14.4	16.4	--
145	POTTER 1S	252.0	1	67.0	80.0	80.0	16.0	18.0	63.1
146	POTTER 2S	253.7	3	--	--	60.0	18.0	20.0	--
147	POTTER 3S	253.7	3	--	--	60.0	17.0	19.0	--
148	LW-01M	265.1	3	--	--	60.0	42.6	52.6	--
149	LW-01S	265.2	3	--	--	40.0	25.6	35.6	--
150	LW-02S	243.4	3	--	--	17.0	4.0	14.0	--
151	LW-03S	250.0	3	--	--	25.0	9.0	19.0	--
152	LW-04S	244.8	3	--	--	20.0	5.0	15.0	--
153	M0W-38	262.7	4	--	--	16.0	--	--	--
154	M0W-32	261.8	4	--	--	54.5	--	--	--
155	GW-02S	255.2	3	--	--	17.0	6.0	16.0	--
156	GW-03S	252.4	3	--	--	20.0	8.4	18.4	--
157	GW-04S	255.6	3	--	--	15.4	5.4	15.4	--
158	GW-05S	264.2	3	--	--	19.0	7.0	17.0	--
160	HAMP B1	266.3	4	--	--	21.5	10.0	20.0	--
162	HAMP B3	258.9	4	--	--	30.0	20.0	30.0	--
163	MI-2	258.9	4	--	--	49.0	37.0	47.0	--
164	MI-3	260.0	4	--	--	49.0	44.0	49.0	--
165	MI-4	259.6	4	--	--	49.0	39.0	49.0	--
166	MI-5	260.0	4	--	--	49.0	39.0	49.0	--

Appendix A.--*Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire*--Continued

Well number	Local identifier	Land-surface elevation (ft)	Data code[a]	Base of stratified drift (ft below land surface)	Depth to bedrock (ft)	Depth drilled (ft)	Depth of screen Top (ft)	Depth of screen Bottom (ft)	Saturated thickness (ft)
167	MI-6	259.2	0	--	--	--	--	--	--
168	MI-6A	259.5	0	--	--	--	--	--	--
169	MI-9	262.2	0	--	--	--	--	--	--
170	MI-14	260.0	0	--	--	--	--	--	--
171	MI-29	269.9	2	49.0	51.5	51.5	31.5	51.5	40.7
172	MI-40	259.8	4	--	--	17.0	--	--	--
173	H12-71	250.0	2	28.0	36.0	36.0	--	--	31.1
174	H11-71	241.6	2	35.0	39.0	39.0	25.0	35.0	33.7
175	H9-71	250.8	2	25.0	28.5	28.5	20.0	25.0	21.6
176	H8-71	250.0	2	25.0	32.0	32.0	20.0	25.0	24.7
177	H6-71	249.5	2	11.0	16.0	16.0	--	--	--
178	H7-71	246.9	2	12.0	15.0	15.0	--	--	--
179	H10-71	250.9	2	28.0	34.0	34.0	18.0	28.0	26.0
180	H5-71	250.5	2	28.0	31.0	31.0	23.0	28.0	22.3
183	B-61	239.9	2	23.0	23.0	23.0	--	--	--
188	M0A-25	262.0	2	60.0	72.0	72.0	50.0	60.0	61.9
189	M0A-35	265.2	2	12.0	12.0	12.0	--	--	--
190	M0A-37	260.0	2	13.0	13.0	13.0	--	--	--
191	M0A-38	270.0	2	14.0	14.0	14.0	--	--	--
193	M0W-15	260.0	4	--	--	55.0	--	--	--
194	M0W-58	268.7	2	76.0	76.0	76.0	54.0	63.0	65.7
195	M0W-64	260.0	2	76.0	76.0	76.0	41.0	49.0	70.4
196	M0W-65	260.0	2	73.0	73.0	73.0	54.0	62.0	67.5
197	M0W-66	252.8	2	37.0	37.0	37.0	27.0	33.0	34.3
198	M0W-67	249.8	2	45.0	45.0	45.0	37.0	43.0	42.2
199	M0W-68	245.0	2	53.0	53.0	53.0	36.0	42.0	49.0
200	M0W-25	259.7	2	4.0	4.0	4.0	--	--	--
201	M0W-26	260.0	2	14.0	14.0	14.0	--	--	--
202	M0W-19	260.8	4	--	--	52.0	--	--	--
203	MI-63	270.0	4	--	--	67.0	24.0	64.0	--
204	MI-13	249.6	6	33.0	--	33.0	12.0	18.0	--
205	HAMP GW 4	270.5	0	--	--	--	--	--	--
207	RB-38	259.7	4	--	--	13.0	--	--	--
208	FH-5	267.9	3	--	--	65.0	50.0	65.0	--
209	HMM 1C	275.5	2	62.0	62.0	71.0	51.0	61.0	51.1

Appendix A.--*Data for selected wells and boreholes in the Milford-Souhegan aquifer, Milford, New Hampshire*--Continued

Well number	Local identifier	Land-surface elevation (ft)	Data code[a]	Base of stratified drift (ft below land surface)	Depth to bedrock (ft)	Depth drilled (ft)	Depth of screen Top (ft)	Depth of screen Bottom (ft)	Saturated thickness (ft)
210	HMM 2B	270.0	2	79.0	115.0	164.0	71.0	81.0	112.1
212	HMM 4B	270.1	2	45.0	45.0	98.0	46.0	56.0	39.3
213	HMM 5B	269.3	2	62.0	62.0	69.0	49.0	59.0	51.0
214	HMM 6B	270.0	2	71.0	71.0	80.0	56.0	65.0	61.1
215	HMM 7B	266.4	2	55.5	58.5	69.0	45.0	56.0	53.4
216	HMM 8B	265.0	2	67.0	90.0	94.0	57.0	67.0	83.4
217	HMM 9C	262.2	2	91.0	91.0	105.0	79.0	90.0	80.6
218	HMM 10C	266.5	2	91.6	91.6	101.0	81.0	91.0	86.3
219	HMM 11R	261.0	2	59.0	65.0	115.0	52.0	64.0	59.4
220	HMM 12A	262.4	2	64.0	66.0	78.0	25.0	35.0	56.4
221	HMM 13B	260.0	2	58.0	64.0	76.0	48.0	58.0	61.4
222	HMM 14R	253.7	2	60.0	60.0	110.0	50.0	60.0	57.1
223	HMM 15A	250.8	2	27.5	27.5	39.0	11.0	27.0	13.2

[a]Data code:

1 Drilled to bedrock, elevation determined by surveying.
2 Drilled to bedrock, elevation estimated from topographic map.
3 Not drilled to bedrock or till, elevation determined by surveying.
4 Not drilled to bedrock or till, elevation estimated from topographic map.
5 Not drilled to bedrock but drilled into till, elevation determined by surveying.
6 Not drilled to bedrock but drilled into till, elevation estimated from topographic map.
7 Water table estimated from adjacent wells.

APPENDIX B

Water levels in selected wells during the October 17, 1988 Keyes aquifer test

Appendix B.--Water levels in selected wells during the Keyes aquifer test, October 17, 1988

[R, distance from pumping well in feet; time, in minutes from commencement of pumping; level, water level in feet above sea level; --, no data]

Well 126 (Keyes well) R = 0		Well 2 R = 237		Well 142 R = 236		Well 3 R = 251		Well 143 R = 259		Well 4 R = 237		Well 144 R = 237	
Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level
0.0	232.47	0.0	235.77	0.0	235.89	0.0	235.58	0.0	235.52	0.0	235.63	0.0	235.62
1.0	222.47	.5	235.77	.5	235.89	.5	235.58	.5	235.52	.5	235.63	.5	235.62
1.5	217.97	1.0	235.77	1.0	235.89	1.0	235.57	1.0	235.52	1.0	235.63	1.0	235.62
2.0	214.57	1.5	235.77	1.5	235.89	1.5	235.57	1.5	235.52	1.5	235.62	1.5	235.62
3.0	212.47	2.0	235.77	2.0	235.89	2.0	235.56	2.0	235.52	2.0	235.60	2.0	235.62
4.0	212.47	3.0	235.76	3.0	235.89	3.0	235.50	3.0	235.52	3.0	235.51	3.0	235.59
6.5	210.97	4.0	235.74	4.0	235.88	4.0	235.41	4.0	235.52	4.0	235.38	4.0	235.55
10.0	209.97	5.0	235.70	5.0	235.87	5.0	235.29	5.0	235.52	5.0	235.22	5.0	235.53
11.5	210.97	6.0	235.66	6.0	235.86	6.0	235.18	6.0	235.52	6.0	235.07	6.0	235.45
18.5	210.97	7.0	235.61	7.0	235.86	7.0	235.09	7.0	235.52	7.0	234.93	7.0	235.43
23.0	210.67	8.0	235.56	8.0	235.85	8.0	234.99	8.0	235.51	8.0	234.81	8.0	235.41
25.5	210.57	9.0	235.50	9.0	235.85	9.0	234.92	9.0	235.51	9.0	234.71	9.0	235.38
36.0	210.47	10.0	235.44	10.0	235.85	10.0	234.86	10.0	235.51	10.0	234.61	10.0	235.37
40.0	210.47	12.0	235.32	12.0	235.84	12.0	234.77	12.0	235.51	12.0	234.48	12.0	235.35
50.0	210.47	14.0	235.22	14.0	235.83	14.0	234.72	14.0	235.49	14.0	234.41	14.0	235.33
56.0	210.47	16.0	235.11	16.0	235.83	16.0	234.68	16.0	235.49	16.0	234.33	16.0	235.30
60.0	210.47	18.0	235.02	18.0	235.83	18.0	234.66	18.0	235.49	18.0	234.31	18.0	235.29
70.0	210.47	20.0	234.93	20.0	235.82	20.0	234.63	20.0	235.49	20.0	234.26	20.0	235.29
80.0	210.47	22.0	234.86	22.0	235.82	22.0	234.62	22.0	235.49	22.0	234.23	22.0	235.28
90.0	210.47	24.0	234.79	24.0	235.82	24.0	234.61	24.0	235.49	24.0	234.22	24.0	235.28
100.0	210.47	26.0	234.73	26.0	235.82	26.0	234.60	26.0	235.49	26.0	234.20	26.0	235.27
120.0	210.47	28.0	234.67	28.0	235.82	28.0	234.60	28.0	235.49	28.0	234.18	28.0	235.25
140.0	210.47	30.0	234.62	30.0	235.82	30.0	234.58	30.0	235.49	30.0	234.16	30.0	235.25
160.0	210.47	35.0	234.53	35.0	235.81	35.0	234.57	35.0	235.49	35.0	234.16	35.0	235.25
180.0	210.57	40.0	234.46	40.0	235.80	40.0	234.56	40.0	235.48	40.0	234.14	40.0	235.24
200.0	210.97	45.0	234.41	45.0	235.80	45.0	234.55	45.0	235.48	45.0	234.13	45.0	235.24
240.0	211.47	50.0	234.37	50.0	235.80	50.0	234.55	50.0	235.48	50.0	234.12	50.0	235.23
300.0	211.47	55.0	234.34	55.0	235.79	55.0	234.55	55.0	235.48	55.0	234.12	55.0	235.23
590.0	211.17	60.0	234.32	60.0	235.79	60.0	234.55	60.0	235.48	60.0	234.11	60.0	235.23
700.0	211.47	70.0	234.28	70.0	235.78	70.0	234.54	70.0	235.48	70.0	234.10	70.0	235.22
815.0	211.47	80.0	234.26	80.0	235.78	80.0	234.54	80.0	235.48	80.0	234.09	80.0	235.22
1080.0	211.47	90.0	234.24	90.0	235.78	90.0	234.53	90.0	235.47	90.0	234.08	90.0	235.21
1285.0	211.47	100.0	234.22	100.0	235.76	100.0	234.52	100.0	235.47	100.0	234.07	100.0	235.20
1405.0	211.97	120.0	234.21	120.0	235.76	120.0	234.52	120.0	235.47	120.0	234.06	120.0	235.19
1545.0	211.97	140.0	234.20	140.0	235.75	140.0	234.52	140.0	235.46	140.0	234.06	140.0	235.18
1675.0	211.97	160.0	234.20	160.0	235.74	160.0	234.52	160.0	235.45	160.0	234.05	160.0	235.17
1672.0	211.97	180.0	234.19	180.0	235.73	180.0	234.52	180.0	235.45	180.0	234.05	180.0	235.16
1799.0	211.97	200.0	234.19	200.0	235.72	200.0	234.52	200.0	235.44	200.0	234.04	200.0	235.15
1902.0	211.47	240.0	234.19	240.0	235.71	240.0	234.52	240.0	235.44	240.0	234.06	240.0	235.13

Well 126 (Keyes well) R = 0		Well 2 R = 237		Well 142 R = 236		Well 3 R = 251		Well 143 R = 259		Well 4 R = 237		Well 144 R = 237	
Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level
1920.0	211.47	300.0	234.19	300.0	235.69	290.0	234.52	290.0	235.42	300.0	234.03	300.0	235.06
2040.0	211.47	480.0	234.12	480.0	235.62	460.0	234.48	460.0	235.39	474.0	233.96	474.0	235.01
2200.0	211.47	610.0	234.11	610.0	235.59	603.0	234.47	603.0	235.38	612.0	233.94	612.0	235.00
2605.0	211.47	720.0	234.09	720.0	235.58	708.0	234.48	708.0	235.39	716.0	233.95	716.0	234.99
2720.0	211.47	832.0	234.08	832.0	235.53	827.0	234.48	827.0	235.39	836.0	233.95	836.0	234.97
2840.0	211.47	1160.0	234.07	1160.0	235.51	1160.0	234.51	1160.0	235.42	1150.0	233.96	1150.0	234.94
2965.0	211.47	1285.0	234.07	1285.0	235.49	1285.0	234.52	1285.0	235.42	1291.0	233.99	1291.0	234.92
3370.0	211.67	1350.0	234.08	1350.0	235.47	1407.0	234.55	1407.0	235.44	1413.0	233.99	1413.0	234.93
3520.0	211.97	1560.0	234.09	1560.0	235.45	1560.0	234.59	1560.0	235.50	1555.0	234.03	1555.0	234.95
3665.0	211.97	1560.0	234.07	1560.0	235.43	1684.0	234.59	1684.0	235.50	1675.0	234.02	1675.0	234.94
4080.0	211.47	1818.0	234.03	1818.0	235.42	1832.0	234.55	1832.0	235.48	1825.0	233.94	1825.0	234.90
4200.0	211.47	1918.0	234.01	1918.0	235.41	1922.0	234.51	1922.0	235.44	1915.0	233.91	1915.0	234.86
4320.0	211.47	2054.0	233.99	2054.0	235.39	2064.0	234.50	2064.0	235.43	2081.0	233.89	2081.0	234.85
4440.0	211.47	2176.0	233.98	2176.0	235.38	2182.0	234.50	2182.0	235.43	2171.0	233.89	2171.0	234.84
4620.0	211.47	2335.0	233.97	2335.0	235.36	2340.0	234.48	2340.0	235.41	2331.0	233.87	2331.0	234.81
4805.0	210.67	2618.0	233.94	2618.0	235.34	2615.0	234.47	2615.0	235.40	2620.0	233.84	2620.0	234.78
4895.0	210.57	2726.0	233.94	2726.0	235.32	2723.0	234.47	2723.0	235.40	2728.0	233.84	2728.0	234.77
5010.0	210.57	2847.0	233.93	2847.0	235.31	2843.0	234.47	2843.0	235.40	2847.0	233.83	2847.0	234.76
5115.0	210.57	2967.0	233.92	2967.0	235.30	2963.0	234.47	2963.0	235.40	2969.0	233.83	2969.0	234.76
5495.0	210.47	3400.0	233.89	3400.0	235.29	3390.0	234.45	3390.0	235.39	3404.0	233.80	3404.0	234.73
5840.0	210.97	3535.0	233.89	3535.0	235.28	3531.0	234.46	3531.0	235.38	3539.0	233.79	3539.0	234.72
6070.0	211.17	3680.0	233.88	3680.0	235.26	3681.0	234.46	3681.0	235.37	3689.0	233.79	3689.0	234.71
6165.0	210.97	4094.0	233.83	4094.0	235.25	4090.0	234.37	4090.0	235.31	4097.0	233.69	4097.0	234.62
6340.0	210.97	4200.0	233.82	4200.0	235.24	4198.0	234.37	4198.0	235.30	4203.0	233.69	4203.0	234.62
6440.0	211.17	4320.0	233.82	4320.0	235.23	4327.0	234.35	4327.0	235.29	4322.0	233.68	4322.0	234.61
7075.0	211.67	4440.0	233.82	4440.0	235.23	4438.0	234.37	4438.0	235.29	4442.0	233.72	4442.0	234.63
7675.0	211.97	4620.0	233.82	4620.0	235.22	4628.0	234.39	4628.0	235.32	4620.0	233.71	4620.0	234.63
8535.0	212.37	4820.0	233.76	4820.0	235.19	4823.0	234.32	4823.0	235.28	4816.0	233.61	4816.0	234.58
9160.0	212.27	4912.0	233.73	4912.0	235.19	4915.0	234.31	4915.0	235.27	4909.0	233.60	4909.0	234.57
10005.0	212.27	5030.0	233.74	5030.0	235.17	5032.0	234.32	5032.0	235.27	5024.0	233.62	5024.0	234.57
--	--	5140.0	233.73	5140.0	235.17	5143.0	234.31	5143.0	235.26	5134.0	233.60	5134.0	234.55
--	--	5508.0	233.70	5508.0	235.16	5502.0	234.28	5502.0	235.24	5511.0	233.57	5511.0	234.52
--	--	5847.0	233.70	5847.0	235.15	5845.0	234.28	5845.0	235.24	5851.0	233.57	5851.0	234.51
--	--	6086.0	233.71	6086.0	235.13	6084.0	234.32	6084.0	235.27	6088.0	233.61	6088.0	234.54
--	--	6461.0	233.70	6461.0	235.11	6455.0	234.30	6455.0	235.25	6465.0	234.57	6465.0	234.57
--	--	7093.0	233.81	7093.0	235.11	7089.0	234.59	7089.0	235.44	7090.0	233.83	7090.0	234.81
--	--	7690.0	234.13	7690.0	235.23	7695.0	234.97	7695.0	235.94	7686.0	234.33	7686.0	235.15
--	--	8560.0	234.20	8560.0	235.36	8656.0	235.00	8656.0	236.05	8563.0	234.29	8563.0	235.18
--	--	9173.0	234.24	9173.0	235.42	9177.0	235.06	9177.0	236.07	9169.0	234.32	9169.0	235.21
--	--	10014.0	234.22	10014.0	235.47	10011.0	235.01	10011.0	236.00	10016.0	234.25	10016.0	235.15
--	--	10116.0	234.23	10116.0	235.48	10116.0	235.01	10116.0	235.00	10116.0	234.24	10116.0	235.15

Appendix B.--Water levels in selected wells during the Keyes aquifer test, October 17, 1988--Continued

Well 1 (Keyes well) R = 699		Well 132 R = 176		Well 145 R = 176		Well 133 R = 208		Well 146 R = 208		Well 134 R = 276		Well 147 R = 276	
Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level
0.0	235.52	0.0	235.65	0.0	235.67	0.0	236.90	0.0	236.79	0.0	235.91	0.0	235.97
.5	235.52	.5	235.65	.5	235.67	.5	236.77	.5	236.79	.5	235.91	.5	235.97
1.5	235.52	1.0	235.65	1.0	235.67	1.0	236.65	1.0	236.79	1.0	235.91	1.0	235.97
2.0	235.48	1.5	235.64	1.5	235.67	1.5	236.65	1.5	236.79	1.5	235.91	1.5	235.97
3.0	235.50	2.0	235.63	2.0	235.67	2.0	236.65	2.0	236.79	2.0	235.91	2.0	235.97
4.0	235.49	3.0	235.57	3.0	235.66	3.0	236.55	3.0	236.79	6.0	235.88	6.0	235.97
5.0	235.48	4.0	235.47	4.0	235.66	4.0	236.55	4.0	236.79	7.0	235.88	7.0	235.97
6.0	235.49	5.0	235.36	5.0	235.66	5.0	236.55	5.0	236.79	9.0	235.84	9.0	235.97
7.0	235.49	6.0	235.20	6.0	235.66	6.0	236.55	6.0	236.79	10.0	235.83	10.0	235.97
8.0	235.48	7.0	235.04	7.0	235.66	7.0	236.15	7.0	236.79	14.0	235.71	14.0	235.97
9.0	235.48	8.0	234.88	8.0	235.66	8.0	235.95	8.0	236.79	16.0	235.66	16.0	235.95
10.0	235.49	9.0	234.68	9.0	235.66	9.0	235.75	9.0	236.79	18.0	235.57	18.0	235.95
12.0	235.49	10.0	234.57	10.0	235.66	12.0	235.30	12.0	236.79	19.0	235.54	19.0	235.95
14.0	235.48	12.0	234.27	12.0	235.66	14.0	235.15	14.0	236.79	20.0	235.51	20.0	235.95
16.0	235.49	14.0	234.01	14.0	235.65	16.0	234.15	16.0	236.79	22.0	235.45	22.0	235.95
18.0	235.48	16.0	233.80	16.0	235.65	18.0	234.00	18.0	236.79	24.0	235.39	24.0	235.95
20.0	235.47	18.0	233.62	18.0	235.65	20.0	233.95	20.0	236.79	26.0	235.33	26.0	235.95
22.0	235.47	20.0	233.46	20.0	235.65	22.0	234.00	22.0	236.79	28.0	235.26	28.0	235.95
24.0	235.48	22.0	233.34	22.0	235.65	24.0	233.85	24.0	236.79	35.0	235.11	35.0	235.96
26.0	235.49	24.0	233.22	24.0	235.65	26.0	233.83	26.0	236.79	40.0	235.04	40.0	235.96
28.0	235.48	26.0	233.13	26.0	235.65	28.0	233.79	28.0	236.79	45.0	234.98	45.0	235.97
30.0	235.48	28.0	233.06	28.0	235.65	30.0	233.81	30.0	236.79	50.0	234.94	50.0	235.97
35.0	235.47	30.0	232.99	30.0	235.65	35.0	233.77	35.0	236.79	55.0	234.93	55.0	235.97
40.0	235.49	35.0	232.89	35.0	235.65	40.0	233.74	40.0	236.79	60.0	234.92	60.0	235.97
45.0	235.47	40.0	232.83	40.0	235.65	45.0	233.70	45.0	236.79	70.0	234.89	70.0	235.97
50.0	235.48	45.0	232.78	45.0	235.65	50.0	233.67	50.0	236.79	80.0	234.84	80.0	235.97
55.0	235.49	50.0	232.76	50.0	235.65	55.0	233.67	55.0	236.79	90.0	234.84	90.0	235.97
60.0	235.47	55.0	232.73	55.0	235.65	60.0	233.68	60.0	236.79	96.0	234.84	96.0	235.99
70.0	235.47	60.0	232.72	60.0	235.65	70.0	233.67	70.0	236.79	100.0	234.84	100.0	235.97
80.0	235.46	70.0	232.72	70.0	235.64	80.0	233.65	80.0	236.79	120.0	234.84	120.0	235.97
90.0	235.47	80.0	232.69	80.0	235.64	90.0	233.65	90.0	236.79	125.0	234.84	125.0	235.96
100.0	235.49	90.0	232.68	90.0	235.64	100.0	233.62	100.0	236.79	140.0	234.84	140.0	235.97
120.0	235.47	100.0	232.66	100.0	235.64	120.0	233.66	120.0	236.79	160.0	234.84	160.0	235.97
140.0	235.47	120.0	232.66	120.0	235.64	140.0	233.63	140.0	236.79	180.0	234.84	180.0	235.97
160.0	235.47	140.0	232.66	140.0	235.64	160.0	233.65	160.0	236.79	200.0	234.83	200.0	235.97
180.0	235.46	160.0	232.66	160.0	235.64	180.0	233.65	180.0	236.79	205.0	234.84	205.0	235.97
200.0	235.46	180.0	232.67	180.0	235.63	200.0	233.64	200.0	236.79	240.0	234.85	240.0	235.97
240.0	235.45	200.0	232.68	200.0	235.63	240.0	233.68	240.0	236.79	365.0	234.86	365.0	235.96
334.0	235.41	240.0	232.74	240.0	235.63	300.0	233.65	300.0	236.79	507.0	234.84	507.0	235.97
642.0	235.42	340.0	232.74	340.0	235.60	349.0	233.65	349.0	236.63	645.0	234.83	645.0	235.96

Appendix B.--Water levels in selected wells during the Keyes aquifer test, October 17, 1988--Continued

Well 1 (Keyes well) R = 699		Well 132 R = 176		Well 145 R = 176		Well 133 R = 208		Well 146 R = 208		Well 134 R = 276		Well 147 R = 276	
Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level	Time	Level
738.0	235.45	501.0	232.71	501.0	235.59	496.0	233.62	496.0	236.62	751.0	234.82	751.0	235.96
944.0	235.46	642.0	232.68	642.0	235.59	649.0	233.60	649.0	236.61	858.0	234.82	858.0	235.95
1200.0	235.52	747.0	232.68	747.0	235.55	744.0	233.59	744.0	236.55	1188.0	234.84	1188.0	235.94
1323.0	235.51	854.0	232.72	854.0	235.58	850.0	233.61	850.0	236.58	1320.0	234.84	1320.0	235.93
1447.0	235.58	1182.0	232.73	1182.0	235.54	1195.0	233.62	1195.0	236.56	1443.0	234.86	1443.0	235.93
1605.0	235.63	1315.0	232.77	1315.0	235.54	1318.0	233.84	1318.0	236.56	1590.0	234.87	1590.0	235.93
1715.0	235.63	1438.0	232.80	1438.0	235.54	1440.0	233.68	1440.0	236.55	1711.0	234.86	1711.0	235.92
1868.0	235.58	1592.0	232.81	1592.0	235.55	1595.0	233.69	1595.0	236.54	1865.0	234.82	1865.0	235.91
1983.0	235.56	1709.0	232.80	1709.0	235.54	1710.0	233.68	1710.0	236.54	1967.0	234.81	1967.0	235.91
2090.0	235.53	1860.0	232.72	1860.0	235.53	1853.0	233.62	1853.0	236.53	2115.0	234.79	2115.0	235.90
2211.0	235.53	1975.0	232.69	1975.0	235.52	1971.0	233.59	1971.0	236.53	2231.0	234.80	2231.0	235.90
2372.0	235.52	2095.0	232.69	2095.0	235.52	2101.0	233.59	2101.0	236.52	2390.0	234.79	2390.0	235.88
2653.0	235.51	2227.0	232.69	2227.0	235.51	2224.0	233.59	2224.0	236.51	2649.0	234.78	2649.0	235.88
2757.0	235.51	2373.0	232.67	2373.0	235.50	2375.0	233.58	2375.0	236.50	2754.0	234.76	2754.0	235.86
2878.0	235.51	2645.0	232.65	2645.0	235.49	2647.0	233.56	2647.0	236.48	2872.0	234.76	2872.0	235.86
2998.0	235.51	2748.0	232.64	2748.0	235.49	2751.0	233.54	2751.0	236.47	2991.0	234.76	2991.0	235.85
3439.0	235.50	2868.0	232.63	2868.0	235.48	2865.0	233.54	2865.0	236.47	3439.0	234.74	3439.0	235.84
3574.0	235.50	2987.0	232.65	2987.0	235.47	2985.0	233.54	2985.0	236.47	3585.0	234.74	3585.0	235.83
3693.0	235.49	3429.0	232.62	3429.0	235.44	3437.0	233.53	3437.0	236.44	3700.0	234.74	3700.0	235.82
4128.0	235.39	3580.0	232.62	3580.0	235.45	3575.0	233.53	3575.0	236.44	4124.0	234.69	4124.0	235.80
4231.0	235.39	3699.0	232.62	3699.0	235.44	3702.0	233.52	3702.0	236.42	4228.0	234.69	4228.0	235.80
4350.0	235.38	4122.0	232.56	4122.0	235.42	4125.0	233.45	4125.0	236.41	4345.0	234.69	4345.0	235.79
4469.0	235.44	4223.0	232.55	4223.0	235.40	4225.0	233.44	4225.0	236.39	4467.0	234.70	4467.0	235.80
4657.0	235.43	4341.0	232.55	4341.0	235.40	4343.0	233.45	4343.0	236.39	4654.0	234.69	4654.0	235.79
4796.0	235.39	4463.0	232.58	4463.0	235.39	4465.0	233.48	4465.0	236.39	4790.0	234.68	4790.0	235.78
4951.0	235.38	4650.0	232.56	4650.0	235.39	4652.0	233.46	4652.0	236.38	4945.0	234.62	4945.0	235.77
5077.0	235.38	4786.0	232.54	4786.0	235.38	4789.0	233.43	4789.0	236.36	5064.0	234.62	5064.0	235.77
5176.0	235.37	4942.0	232.41	4942.0	235.36	4938.0	233.34	4938.0	236.34	5176.0	234.62	5176.0	235.77
5539.0	235.36	5062.0	232.42	5062.0	235.36	5059.0	233.35	5059.0	236.34	5535.0	234.60	5535.0	235.76
5875.0	235.37	5173.0	232.41	5173.0	235.36	5175.0	233.34	5175.0	236.34	5872.0	234.60	5872.0	235.75
6136.0	235.41	5531.0	232.39	5531.0	235.34	5532.0	233.33	5532.0	236.33	6132.0	234.61	6132.0	235.74
6472.0	235.38	5869.0	232.42	5869.0	235.33	5871.0	233.34	5871.0	236.32	6478.0	234.60	6478.0	235.72
7150.0	235.69	6127.0	232.45	6127.0	235.31	6123.0	233.37	6123.0	236.31	7141.0	234.69	7141.0	235.70
7731.0	236.21	6481.0	232.43	6481.0	235.30	6487.0	233.35	6487.0	236.29	7726.0	234.86	7726.0	235.69
8597.0	236.20	7139.0	232.64	7139.0	235.34	7134.0	233.56	7134.0	236.34	8597.0	234.91	8597.0	235.73
9189.0	236.25	7724.0	233.01	7724.0	235.54	7724.0	233.88	7724.0	236.54	9204.0	234.94	9204.0	235.76
10032.0	236.17	8591.0	233.03	8591.0	235.70	8591.0	233.90	8591.0	236.70	10030.0	234.93	10030.0	235.80
--	--	9192.0	233.06	9192.0	235.73	9192.0	233.92	9192.0	236.72	--	--	--	--
--	--	10025.0	233.01	10025.0	235.74	10025.0	233.88	10025.0	236.72	--	--	--	--
--	--	10116.0	232.99	10116.0	235.72	10116.0	233.97	10116.0	236.69	--	--	--	--

APPENDIX C

Ground-water-level data for wells in the Milford-Souhegan aquifer

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer

[--, no data]

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
1	248.71	11.20	237.51	09/28/88	236.64
		11.30	237.41	10/05/88	
		11.45	237.26	10/11/88	
		11.42	237.29	10/31/88	
		12.58	236.13	12/01/88	
		12.98	235.73	12/28/88	
		13.53	235.18	02/02/89	
2	246.61	10.70	235.91	09/28/88	235.31
		10.53	236.08	10/05/88	
		10.82	235.79	10/13/88	
		10.67	235.94	10/31/88	
		11.68	234.93	12/01/88	
		12.22	234.39	12/28/88	
		12.49	234.12	02/02/89	
3	244.84	9.10	235.74	09/28/88	235.07
		9.29	235.55	10/05/88	
		9.25	235.59	10/13/88	
		9.11	235.73	10/31/88	
		10.16	234.68	12/01/88	
		10.65	234.19	12/28/88	
		10.85	233.99	02/02/89	
4	243.31	7.63	235.68	09/28/88	235.04
		7.78	235.53	10/05/88	
		7.70	235.61	10/13/88	
		7.56	235.75	10/31/88	
		8.73	234.58	12/01/88	
		9.15	234.16	12/28/88	
		9.32	233.99	02/02/89	
6	243.10	4.97	238.13	10/13/88	237.69
		4.94	238.16	10/18/88	
		4.94	238.16	10/20/88	
		4.75	238.35	10/22/88	
		4.55	238.55	10/23/88	
		4.46	238.64	10/24/88	
		4.43	238.67	10/25/88	
		4.55	238.55	10/31/88	
		6.79	236.31	12/07/88	
		7.85	235.25	02/02/89	
		7.30	235.80	12/28/88	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
7	247.30	10.00	237.30	10/13/88	237.03
		9.92	237.38	10/18/88	
		9.95	237.35	10/20/88	
		9.70	237.60	10/22/88	
		9.42	237.88	10/23/88	
		9.32	237.98	10/24/88	
		9.28	238.02	10/25/88	
		9.52	237.78	10/31/88	
		11.48	235.82	12/07/88	
		11.95	235.35	12/28/88	
		12.43	234.87	02/02/89	
8	243.40	6.27	237.13	10/13/88	236.78
		6.23	237.17	10/18/88	
		6.28	237.12	10/20/88	
		6.04	237.36	10/22/88	
		5.80	237.60	10/23/88	
		5.71	237.69	10/24/88	
		5.67	237.73	10/25/88	
		5.84	237.56	10/31/88	
		7.74	235.66	12/07/88	
		8.31	235.09	12/28/88	
		8.88	234.52	02/02/89	
9	260.00	4.00	256.00	10/28/55	--
14	256.03	2.21	253.82	11/19/86	254.27
		0.72	255.31	12/08/86	
		1.07	254.96	03/25/87	
		2.39	253.64	06/10/87	
		2.41	253.62	09/18/87	
15	254.18	4.29	249.89	11/19/86	250.12
		2.72	251.46	12/08/86	
		3.01	251.17	03/25/87	
		4.60	249.58	06/10/87	
		5.66	248.52	09/18/87	
16	254.51	4.90	249.61	11/19/86	249.62
		3.45	251.06	12/08/86	
		3.79	250.72	03/25/87	
		5.30	249.21	06/10/87	
		6.21	248.30	09/18/87	
		5.70	248.81	10/13/88	
17	252.12	2.37	249.75	11/19/86	249.68
		1.12	251.00	12/08/86	
		1.46	250.66	03/25/87	
		2.73	249.39	06/10/87	
		3.67	248.45	09/18/87	
		3.29	248.83	10/13/88	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
18	258.32	8.02	250.30	11/19/86	250.43
		6.16	252.16	12/08/86	
		8.18	250.14	06/10/87	
		9.19	249.13	09/18/87	
19	255.49	5.42	250.07	11/19/86	249.91
		5.19	250.30	06/10/87	
		6.14	249.35	09/18/87	
20	259.12	8.96	250.16	11/19/86	250.19
		7.17	251.95	12/08/86	
		9.25	249.87	06/10/87	
		10.33	248.79	09/18/87	
21	255.42	6.55	248.87	09/12/84	249.57
		5.74	249.68	11/09/84	
		5.26	250.16	10/13/88	
22	261.92	3.30	258.62	04/18/83	256.07
		7.11	254.81	09/27/83	
		6.32	255.60	09/12/84	
		6.66	255.26	11/09/84	
23	252.06	5.23	246.83	09/12/84	--
		4.13	247.93	11/09/84	
24	252.95	6.57	246.38	09/12/84	247.10
		5.60	247.35	11/09/84	
		4.22	247.56	10/13/88	
25	251.58	5.80	245.78	09/12/84	245.32
		4.98	246.60	11/09/84	
		8.17	243.59	10/13/88	
26	266.50	5.85	260.65	04/18/83	257.85
		9.74	256.76	09/12/84	
		10.36	256.14	11/09/84	
28	273.35	12.01	261.34	09/12/84	--
		10.33	263.02	09/12/84	
29	260.00	7.86	252.14	1/21/88	252.60
		7.01	252.99	2/23/88	
		7.42	252.58	3/23/88	
		7.60	252.40	4/21/88	
		6.90	253.10	5/23/88	
		7.50	252.50	6/25/88	
		6.95	253.05	7/23/88	
		7.96	252.04	8/23/88	
		7.60	252.40	9/24/88	
		8.08	251.92	10/21/88	
		6.40	253.60	11/21/88	
		7.57	252.43	12/22/88	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
30	275.60	11.41	264.19	09/29/83	264.81
		11.74	263.86	10/06/83	
		11.62	263.98	10/21/83	
		11.01	264.59	11/07/83	
		9.10	266.50	02/10/84	
		8.29	267.31	05/11/84	
		10.77	264.83	08/22/84	
		11.22	264.38	09/12/84	
		11.58	264.02	11/09/84	
		11.17	264.43	10/13/88	
31	275.60	12.00	263.60	09/29/83	265.06
		11.22	264.38	10/06/83	
		11.19	264.41	10/21/83	
		10.44	265.16	11/07/83	
		8.98	266.62	02/10/84	
		8.22	267.38	05/11/84	
		10.76	264.84	08/22/84	
		11.08	264.52	09/12/84	
		10.87	264.73	11/09/84	
		10.65	264.95	10/13/88	
32	274.70	11.40	263.30	09/23/83	264.01
		11.42	263.28	09/27/83	
		11.33	263.37	10/06/83	
		11.24	263.46	10/21/83	
		10.62	264.08	11/07/83	
		8.16	266.54	05/11/84	
		10.62	264.08	08/22/84	
		10.85	263.85	09/12/84	
		11.08	263.62	11/09/84	
		10.23	264.47	10/13/88	
33	273.00	8.40	264.60	10/03/83	265.09
		8.55	264.45	10/06/83	
		8.50	264.50	10/21/83	
		7.50	265.50	11/07/83	
		6.61	266.39	02/10/84	
		6.75	266.25	05/11/84	
		8.32	264.68	08/22/84	
		8.31	264.69	09/12/84	
		8.26	264.74	11/09/84	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
34	269.98	8.47	261.51	09/27/83	262.39
		8.45	261.53	10/06/83	
		8.40	261.58	10/21/83	
		7.63	262.35	11/07/83	
		4.72	265.26	05/11/84	
		7.31	262.67	08/22/84	
		7.67	262.31	09/12/84	
		8.10	261.88	11/09/84	
35	270.00	12.76	257.24	10/06/83	260.21
		10.66	259.34	10/21/83	
		9.56	260.44	11/07/83	
		7.82	262.18	02/10/84	
		7.03	262.97	05/11/84	
		9.80	260.20	08/22/84	
		10.03	259.97	09/12/84	
		10.40	259.60	11/09/84	
		10.07	259.93	10/13/88	
36	270.10	8.40	261.70	09/23/83	261.86
		9.02	261.08	09/27/83	
		8.87	261.23	10/06/83	
		8.93	261.17	10/21/83	
		7.61	262.49	11/07/83	
		6.53	263.57	05/11/84	
		8.74	261.36	08/22/84	
		7.38	262.72	09/12/84	
		8.68	261.42	11/09/84	
37	270.00	9.96	260.04	10/06/83	260.87
		10.11	259.89	10/21/83	
		9.03	260.97	11/07/83	
		7.61	262.39	02/10/84	
		6.86	263.14	05/11/84	
		9.20	260.80	08/22/84	
		9.72	260.28	09/12/84	
		10.05	259.95	11/09/84	
		9.61	260.39	10/13/88	
38	270.60	11.56	259.04	10/21/83	260.37
		10.64	259.96	11/07/83	
		8.59	262.01	02/10/84	
		7.58	263.02	05/11/84	
		10.56	260.04	08/22/84	
		10.81	259.79	09/12/84	
		11.39	259.21	11/09/84	
		10.74	259.86	10/13/88	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
39	271.70	11.22	260.48	09/27/83	261.44
		11.24	260.46	10/06/83	
		11.13	260.57	10/21/83	
		10.23	261.47	11/07/83	
		6.63	265.07	05/11/84	
		9.75	261.95	08/22/84	
		11.60	260.10	09/12/84	
40	270.10	9.20	260.90	10/10/83	260.27
		11.32	258.78	10/21/83	
		10.52	259.58	11/07/83	
		8.20	261.90	02/10/84	
		10.12	259.98	08/22/84	
		10.37	259.73	09/12/84	
		7.15	262.95	11/05/84	
		11.21	258.89	11/09/84	
		10.37	259.73	10/13/88	
41	270.10	8.90	261.20	10/10/83	260.27
		11.53	258.57	10/21/83	
		10.65	259.45	11/07/83	
		8.18	261.92	02/10/84	
		7.04	263.06	05/11/84	
		10.14	259.96	08/22/84	
		10.48	259.62	09/12/84	
		11.25	258.85	11/09/84	
		10.26	259.84	10/13/88	
42	270.70	12.13	258.57	10/21/83	260.19
		11.28	259.42	11/07/83	
		8.71	261.99	02/10/84	
		7.54	263.16	05/11/84	
		10.63	260.07	08/22/84	
		11.08	259.62	09/12/84	
		11.97	258.73	11/09/84	
		10.72	259.98	10/13/88	
43	270.30	9.50	260.80	07/26/83	260.50
		10.02	260.24	08/05/83	
		10.54	259.72	08/17/83	
		11.40	258.86	09/27/83	
		10.90	259.36	10/20/83	
		7.08	263.18	11/28/83	
		7.58	262.68	01/26/84	
		10.03	260.23	09/12/84	
		10.92	259.34	11/09/84	
		9.70	260.56	10/13/88	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
44	265.40	10.00	255.40	07/28/83	256.21
		9.16	256.25	08/05/83	
		9.85	255.56	08/17/83	
		11.10	254.31	09/27/83	
		9.83	255.58	10/20/83	
		6.01	259.40	11/28/83	
		7.18	258.23	01/26/84	
		10.00	255.41	09/12/84	
		10.33	255.08	11/09/84	
		8.61	256.85	10/13/88	
45	266.00	10.00	256.00	08/03/83	256.22
		9.20	256.79	08/05/83	
		9.85	256.14	08/17/83	
		10.90	255.09	09/27/83	
		10.39	255.60	10/20/83	
		6.10	259.89	11/28/83	
		10.93	255.06	01/26/84	
		9.09	256.90	09/12/84	
		10.20	255.79	11/09/84	
		9.77	256.22	10/13/88	
46	270.19	12.00	258.20	08/05/83	259.16
		11.55	258.64	08/17/83	
		12.19	258.00	09/27/83	
		11.64	258.55	10/20/83	
		7.89	262.30	11/28/83	
		5.10	265.09	01/26/84	
		11.24	258.95	09/12/84	
		11.78	258.41	11/09/84	
		11.03	259.16	10/13/88	
47	268.23	16.02	252.21	09/27/83	252.62
		13.56	254.67	11/28/83	
		15.04	253.19	01/26/84	
		15.91	252.32	09/12/84	
		17.52	250.71	11/09/84	
48	279.68	15.26	264.42	09/12/84	264.33
		15.21	264.47	11/09/84	
		13.56	264.11	10/13/88	
49	265.92	8.25	257.67	04/18/83	252.91
		12.35	253.57	09/27/83	
		13.41	252.51	09/12/84	
		18.03	247.89	11/09/84	
50	270.01	4.24	265.77	04/18/83	262.99
		7.79	262.22	09/12/84	
		8.58	261.43	11/09/84	
		7.23	262.55	10/13/88	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
51	270.63	6.25	264.38	04/18/83	261.63
		10.13	260.50	09/12/84	
		10.62	260.01	11/09/84	
52	270.00	5.87	264.13	04/18/83	--
		9.67	260.33	09/27/83	
53	286.28	9.92	276.36	04/18/83	--
54	258.60	9.00	249.60	11/15/80	252.23
		6.11	252.54	09/12/84	
		5.68	252.97	11/09/84	
		4.85	253.80	10/13/88	
55	257.40	8.50	248.90	11/15/80	251.87
		5.09	252.29	09/12/84	
		4.65	252.73	09/12/84	
		3.84	253.54	10/13/88	
56	257.20	7.50	249.70	11/15/80	252.99
		5.36	251.88	09/12/84	
		4.76	252.48	11/09/84	
		4.25	252.99	10/13/88	
57	269.99	11.40	258.59	11/01/78	--
58	259.79	8.00	251.79	09/12/84	--
		8.37	251.42	11/09/84	
59	264.93	3.97	260.96	04/18/83	--
60	267.28	6.26	261.02	04/18/83	--
61	270.00	9.09	260.91	04/18/83	--
62	260.33	5.00	255.33	09/12/84	--
		5.39	254.94	11/09/84	
72	259.97	5.63	254.34	10/13/88	--
74	259.99	3.00	256.99	01/01/60	
75	239.47	9.80	229.67	05/19/80	
77	241.11	5.75	235.36	05/21/80	
78	249.48	7.50	241.98	05/22/80	--
84	267.98	34.00	233.98	10/06/88	235.26
		32.78	235.20	10/13/88	
		36.25	231.73	10/21/88	
		33.17	234.81	10/24/88	
		27.41	240.57	02/02/89	
85	262.42	12.54	249.88	12/20/88	--
		12.78	249.64	02/02/89	
86	260.02	15.97	244.05	12/20/88	--
		15.81	244.21	02/02/89	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
87	262.16	12.00	250.16	09/01/88	244.58
		21.00	241.16	10/06/88	
		20.51	241.65	10/13/88	
		22.25	239.91	10/21/88	
		21.00	241.16	11/01/88	
		13.22	247.79	12/20/88	
		12.91	249.25	02/02/89	
		16.58	245.58	10/24/89	
89	249.96	5.64	244.36	04/22/85	--
90	252.85	5.50	247.35	04/22/85	--
91	251.57	5.39	246.17	04/23/85	--
92	252.30	5.78	246.52	04/23/85	245.89
		7.10	245.23	10/28/88	
		6.40	245.93	12/07/88	
93	252.00	5.78	246.20	04/23/85	--
94	253.49	5.57	247.89	04/24/85	--
95	260.00	3.70	256.30	05/02/85	--
97	269.89	5.90	263.99	01/21/88	--
98	269.33	8.40	260.93	02/01/88	--
99	269.96	8.30	261.66	02/10/88	--
100	269.05	1.40	267.65	01/27/88	--
101	269.71	0.40	269.31	01/29/88	--
102	275.30	2.80	272.50	01/29/88	--
103	275.01	1.80	273.21	01/22/88	--
104	275.37	2.50	272.87	02/09/88	--
123	256.13	4.80	251.33	09/28/88	250.79
		4.90	251.23	10/05/88	
		4.89	251.24	10/13/88	
		4.43	251.70	10/22/88	
		4.48	251.65	10/23/88	
		4.50	251.63	10/24/88	
		4.56	251.57	10/25/88	
		4.71	251.42	10/31/88	
		6.93	249.20	12/07/88	
		7.14	248.99	12/28/88	
		7.38	248.75	02/02/89	
124	256.54	4.15	252.39	09/28/88	251.68
		4.24	252.30	10/05/88	
		4.28	252.26	10/13/88	
		4.03	252.51	10/22/88	
		4.02	252.52	10/23/88	
		3.99	252.55	10/24/88	
		4.04	252.50	10/25/88	
		4.18	252.36	10/31/88	
		6.58	249.96	12/07/88	
		6.77	249.77	12/28/88	
		7.21	249.33	02/02/89	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
125	256.71	5.00	251.71	09/28/88	251.11
		5.00	251.71	10/05/88	
		5.11	251.60	10/13/88	
		4.76	251.95	10/22/88	
		4.78	251.93	10/23/88	
		4.76	251.95	10/24/88	
		4.83	251.88	10/25/88	
		4.97	251.74	10/31/88	
		7.22	249.49	12/07/88	
		7.43	249.28	12/28/88	
		7.76	248.95	02/02/89	
126	240.10	8.41	231.69	05/10/72	234.89
		4.40	235.74	09/28/88	
		4.56	235.58	10/05/88	
		4.48	235.66	10/17/88	
		4.38	235.76	10/31/88	
128	261.00	4.00	257.00	03/21/57	257.57
		3.21	257.79	05/01/60	
		3.10	257.90	04/13/81	
129	241.70	8.50	233.20	10/28/71	--
130	240.52	9.10	231.42	10/28/71	--
131	240.32	6.60	233.72	10/29/71	--
132	251.75	15.98	235.77	10/05/88	235.08
		16.10	235.65	10/12/88	
		15.95	235.80	10/31/88	
		16.95	234.80	12/01/88	
		17.41	234.34	12/28/88	
		17.65	234.10	02/02/89	
133	253.77	18.03	235.74	10/05/88	235.11
		18.05	235.72	10/13/88	
		17.97	235.80	10/31/88	
		18.90	234.87	12/01/88	
		19.37	234.40	12/28/88	
		19.66	234.11	02/02/89	
134	253.67	17.65	236.02	10/05/88	235.41
		17.70	235.97	10/13/88	
		17.64	236.03	10/31/88	
		18.39	235.28	12/01/88	
		18.90	234.77	12/28/88	
		19.31	234.36	02/02/89	
135	241.38	10.00	231.38	08/28/56	--

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
136	247.10	12.92	234.18	09/20/68	234.09
		12.73	234.33	12/01/88	
		13.09	233.97	12/28/88	
		13.19	233.87	02/02/89	
137	239.96	13.00	226.96	08/27/56	--
138	239.97	13.00	226.97	08/27/56	--
139	239.83	12.67	227.13	09/23/68	--
140	241.75	11.75	229.95	09/23/68	--
141	245.25	12.25	232.95	09/23/68	--
142	246.45	10.40	236.05	09/28/88	235.35
		10.76	235.69	10/05/88	
		10.55	235.90	10/13/88	
		10.37	236.08	10/31/88	
		11.29	235.16	12/01/88	
		11.93	234.52	12/28/88	
		12.38	234.07	02/02/89	
143	245.62	9.95	235.67	09/28/88	234.97
		10.15	235.47	10/05/88	
		10.10	235.52	10/13/88	
		9.96	235.66	10/31/88	
		11.10	234.52	12/01/88	
		11.58	234.04	12/28/88	
		11.74	233.88	02/02/89	
144	243.28	7.57	235.71	09/28/88	235.04
		7.75	235.53	10/05/88	
		7.70	235.58	10/13/88	
		7.53	235.75	10/31/88	
		8.66	234.62	12/01/88	
		9.13	234.15	12/28/88	
		9.31	233.97	02/02/89	
145	251.76	16.05	235.71	10/05/88	235.10
		16.07	235.69	10/13/88	
		15.95	235.81	10/31/88	
		16.85	234.91	12/01/88	
		17.37	234.39	12/28/88	
		17.68	234.08	02/02/89	
146	253.79	18.05	235.74	10/05/88	235.12
		18.10	235.69	10/13/88	
		17.97	235.82	10/31/88	
		18.85	234.94	12/01/88	
		19.36	234.43	12/28/88	
		19.70	234.09	02/02/89	

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
147	253.66	17.54	236.12	10/05/88	235.48
		17.65	236.01	10/13/88	
		17.58	236.08	10/31/88	
		18.25	235.41	12/01/88	
		18.80	234.86	12/28/88	
		19.29	234.37	02/02/89	
150	242.80	4.65	238.15	10/13/88	237.78
		4.64	238.16	10/18/88	
		4.64	238.16	10/20/88	
		4.43	238.37	10/22/88	
		4.23	238.57	10/23/88	
		4.15	238.65	10/24/88	
		4.11	238.69	10/25/88	
		4.25	238.55	10/31/88	
		6.17	236.63	12/07/88	
		6.70	236.10	12/28/88	
		7.27	235.53	02/02/89	
151	247.60	10.36	237.24	10/13/88	237.00
		10.30	237.30	10/18/88	
		10.31	237.29	10/20/88	
		10.11	237.49	10/22/88	
		9.81	237.79	10/23/88	
		9.70	237.90	10/24/88	
		9.64	237.96	10/25/88	
		9.88	237.72	10/31/88	
		11.67	235.93	12/07/88	
		12.15	235.45	12/28/88	
		12.62	234.98	02/02/89	
152	243.50	4.76	238.74	10/13/88	238.69
		4.81	238.69	10/18/88	
		4.85	238.65	10/20/88	
		4.17	239.33	10/22/88	
		3.82	239.68	10/23/88	
		3.71	239.79	10/24/88	
		3.67	239.83	10/25/88	
		3.94	239.56	10/31/88	
		5.51	237.99	12/07/88	
		6.44	237.06	12/28/88	
		7.23	236.27	02/02/89	
153	262.74	14.60	248.14	06/28/63	--
154	261.76	3.00	258.76	10/27/55	--
160	266.33	11.00	255.33	04/01/85	--

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
161	271.14	14.50	256.64	04/04/87	--
162	258.90	25.40	233.50	04/04/87	--
163	258.90	2.40	256.50	04/13/81	252.01
		9.59	249.31	09/12/84	
		8.68	250.22	11/09/84	
164	260.00	2.75	257.25	04/13/81	252.28
		10.66	249.34	09/12/84	
		9.76	250.24	11/09/84	
165	259.57	10.02	249.55	09/12/84	--
		9.21	250.36	11/09/84	
166	260.00	10.17	249.83	09/12/84	--
		9.42	250.58	11/09/84	
167	259.17	9.73	249.44	09/12/84	--
		8.89	250.28	11/09/84	
168	259.47	10.03	249.44	09/12/84	--
		9.18	250.29	11/09/84	
169	262.22	3.90	258.32	04/18/83	256.44
		6.37	255.85	09/27/83	
		6.56	255.66	09/12/84	
		6.30	255.92	11/09/84	
170	259.97	4.19	255.78	04/18/83	254.05
		6.98	252.99	09/27/83	
		6.44	253.53	09/12/84	
		6.09	253.88	11/09/84	
171	269.90	9.00	260.90	07/28/83	259.24
		11.74	258.12	08/05/83	
		11.47	258.39	08/17/83	
		12.85	257.01	09/27/83	
		12.47	257.39	10/20/83	
		7.81	262.05	11/28/83	
		8.60	261.26	01/26/84	
		10.51	259.35	09/12/84	
		11.95	257.91	11/09/84	
		9.53	259.99	10/13/88	
172	259.82	6.39	253.43	09/12/84	--
		6.23	253.59	11/09/84	
173	250.00	4.92	245.10	11/03/71	--
174	241.59	5.33	236.29	11/04/71	--
175	250.82	6.94	243.92	11/03/71	--
176	249.98	7.29	242.68	11/02/71	--
179	250.85	8.04	242.85	11/03/71	--
180	250.48	8.67	241.78	11/02/71	--

Appendix C.--Ground-water-level data for wells in the Milford-Souhegan aquifer--Continued

Well number	Land-surface elevation (feet)	Depth to water (feet)	Water-table elevation (feet)	Measurement date (mm/dd/yy)	Observed mean (feet)
188	261.97	10.10	251.87	08/30/78	--
194	268.71	10.30	258.41	08/28/78	--
195	260.00	5.60	254.40	11/02/78	--
196	260.03	5.50	254.53	11/03/78	--
197	252.80	2.70	250.10	03/23/83	--
198	249.82	2.75	247.02	03/24/83	--
199	244.98	4.00	240.98	03/31/83	--
202	260.84	7.33	253.54	10/10/55	--
204	249.62	4.50	245.12	04/12/83	--
206	262.47	10.00	252.47	--	--
208	267.89	37.59	230.30	09/02/88	228.66
		37.59	230.30	09/09/88	
		42.59	225.30	09/21/88	
		42.59	225.30	10/06/88	
		43.29	224.60	10/13/88	
		44.00	223.89	10/21/88	
		42.00	225.89	10/24/88	
		42.59	225.30	11/01/88	
		37.00	230.89	11/04/88	
		28.06	234.62	12/20/88	
		29.02	238.87	02/02/89	
209	275.46	10.85	264.56	11/10/88	--
210	270.01	2.89	267.11	11/28/88	--
211	266.14	7.50	258.64	12/08/88	--
212	270.06	5.72	264.36	10/25/88	--
213	269.29	10.97	258.29	10/19/88	--
214	270.00	9.86	260.10	10/26/88	--
215	266.37	5.12	261.27	10/06/88	--
216	264.99	6.55	258.39	10/12/88	--
217	262.19	10.40	251.79	10/31/88	--
218	266.49	5.30	261.19	11/17/88	--
219	260.97	5.62	255.37	11/12/88	--
220	262.38	9.60	252.78	12/05/88	--
221	260.00	2.60	257.40	11/29/88	--
222	253.72	2.90	250.82	11/22/88	--
223	250.81	14.30	236.51	12/07/88	--

☆U.S. GOVERNMENT PRINTING OFFICE: 1992 -601-933/60029