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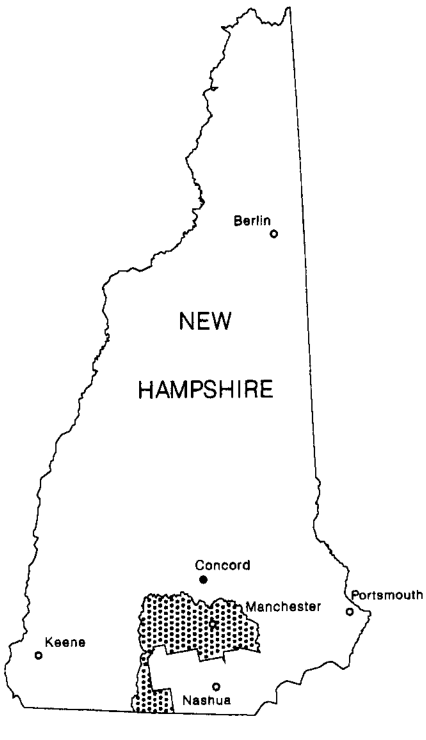
Geohydrology and Water Quality of Stratified-Drift Aquifers in the Middle Merrimack River Basin, South-Central New Hampshire

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 92-4192



Prepared in cooperation with the

NEW HAMPSHIRE DEPARTMENT OF ENVIRONMENTAL SERVICES,
WATER RESOURCES DIVISION

GEOHYDROLOGY AND WATER QUALITY OF STRATIFIED-DRIFT AQUIFERS IN THE MIDDLE MERRIMACK RIVER BASIN, SOUTH-CENTRAL NEW HAMPSHIRE

By Joseph D. Ayotte and Kenneth W. Toppin

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Bow, New Hampshire
1995

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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square mile (mi^2)	2.59	square kilometer
	Volume	
cubic foot (ft^3)	0.02832	cubic meter
gallon (gal)	3.785	liter
million gallons (Mgal)	3,785	cubic meter
	Flow	
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

CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS (continued)

	Transmissivity	
cubic foot per day per square foot times foot of aquifer thickness [$(ft^3/d)/ft^2$]ft or ft^2/d	0.0929	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.

Abbreviated water-quality units used in this report: Chemical concentrations in water are expressed in milligrams per liter (mg/L) or micrograms per liter (μg/L). Milligrams per liter is a unit expressing the concentration of chemical constituents in solution as weight (milligrams) of solute per unit volume (liter) of water; 1,000 μg/L is equivalent to 1 mg/L. Water temperature in degrees Celsius (°C) can be converted to degrees Fahrenheit (°F) by use of the following equation:

$$^{\circ}F = 1.8(^{\circ}C) + 32$$

Geohydrology and Water Quality of Stratified-Drift Aquifers in the Middle Merrimack River Basin, South-Central New Hampshire

By Joseph D. Ayotte and Kenneth W. Toppin

ABSTRACT

The U.S. Geological Survey, in cooperation with the State of New Hampshire, Department of Environmental Services, Water Resources Division has assessed the geohydrology and water quality of stratified-drift aquifers in the middle Merrimack River basin in south-central New Hampshire. The middle Merrimack River basin drains 469 square miles; 98 square miles is underlain by stratified-drift aquifers. Saturated thickness of stratified drift within the study area is generally less than 40 feet but locally greater than 100 feet. Transmissivity of stratified-drift aquifers is generally less than 2,000 feet squared per day but locally exceeds 6,000 feet squared per day. At present (1990), ground-water withdrawals from stratified drift for public supply are about 0.4 million gallons per day within the basin. Many of the stratified-drift aquifers within the study area are not developed to their fullest potential.

The geohydrology of stratified-drift aquifers was investigated by focusing on basic aquifer properties, including aquifer boundaries; recharge, discharge, and direction of ground-water flow; saturated thickness and storage; and transmissivity. Surficial geologic mapping assisted in the determination of aquifer boundaries. Data from 757 wells and test borings were used to produce maps of water-table altitude, saturated thickness, and transmissivity of stratified drift. More than 10 miles of seismic-refraction profiling and 14 miles of seismic-reflection profiling were also used to construct the water table and saturated-thickness maps.

Stratified-drift aquifers in the southern, western, and central parts of the study area are typically small and discontinuous, whereas aquifers in the eastern part along the Merrimack River valley are continuous. The Merrimack River valley aquifers formed in glacial Lakes Merrimack and Hooksett. Many other smaller discontinuous aquifers formed in small temporary ponds during deglaciation.

A stratified-drift aquifer in Goffstown was analyzed for aquifer yield by use of a two-dimensional, finite-difference ground-water-flow model. Yield of the Goffstown aquifer was estimated to be 2.5 million gallons per day. Sensitivity analysis showed that the estimate of aquifer yield was most sensitive to changes in hydraulic conductivity. The amount of water induced into the aquifer from the Piscataquog River was most affected by changes in estimates of streambed conductance.

Results of analysis of water samples from 10 test wells indicate that, with some exceptions, water in the stratified-drift aquifers generally meets U.S. Environmental Protection Agency primary and secondary drinking-water regulations. Water from two wells had elevated sodium concentrations, water from two wells had elevated concentrations of dissolved iron, and water from seven wells had elevated concentrations of manganese. Known areas of contamination were avoided during water-quality sampling.

INTRODUCTION

The population of the 19 southern New Hampshire towns in the middle Merrimack River basin increased by 22 percent between 1980 and 1990 (New Hampshire Office of State Planning, 1985). Economic development has been rapid in south-central New Hampshire, partly because of the area's proximity to metropolitan Boston. This growth has steadily increased demands for water and has stressed the capacity of existing municipal water systems, some of which depend on ground water for part or all of their water supplies. The total withdrawal from stratified-drift aquifers for municipal supply in 1990 was about 0.4 Mgal/d and represents total withdrawal divided by 365 days per year to get average daily use as if the total withdrawal were spread out over a full year (New Hampshire Department of Environmental Services, Water Management Bureau, written commun., 1991). Two of the municipal water systems use ground water seasonally to supplement surface-water supplies. In addition, U.S. Environmental Protection Agency (USEPA) primary and secondary drinking-water regulations on the treatment requirements of surface-water supplies have prompted municipalities to look carefully at their ground-water resources.

Stratified-drift aquifers discontinuously underlie 98 mi^2 of the middle Merrimack River basin, which has a total drainage area of 469 mi^2. Many of the aquifers may be capable of supplying enough water to meet domestic, community, and municipal water needs.

The U.S. Geological Survey (USGS), in cooperation with the New Hampshire Department of Environmental Services, Water Resources Division (NHDES-WRD), has done a series of ground-water studies in New Hampshire to provide detailed geohydrologic information necessary for planning for optimal use of existing water supplies and for the development of new water supplies. The study described in this report encompasses the middle Merrimack River basin and its subbasins, which include the Piscataquog River basin and part of the Souhegan River basin (fig. 1). Major watershed divides were selected as study areas because they are the natural subdivision of the hydrologic system; only a few stratified-drift aquifers, in south-central New Hampshire, extend across major surface-water divides.

Purpose and Scope

The purpose of this report is to (1) describe the geohydrologic characteristics of the stratified-drift aquifers in the middle Merrimack River basin, including the areal extent of the aquifers, water-table altitudes, general directions of ground-water flow, saturated thickness, and transmissivity; (2) present a technique for evaluating the yield of an aquifer; and (3) describe the quality of ground water in the stratified-drift aquifers.

The study was limited to the collection, compilation, and the evaluation of data from the stratified-drift aquifers in the study area. The yield of a stratified-drift aquifer in Goffstown, currently used to augment a public surface-water supply, was evaluated using a numerical model. The modeling technique could be used to evaluate similar aquifers in New Hampshire.

Previous Investigations

Products of previous investigations include a reconnaissance map at a scale of 1:125,000 that shows the availability of ground water in the middle Merrimack River basin (Cotton, 1977). Surficial geology maps for parts of the study area are being produced at a scale of 1:24,000 as part of the Cooperative Geologic Mapping (COGEOMAP) program, a cooperative program between various states and the USGS. In New Hampshire, the Department of Environmental Services, Office of the State Geologist, is the cooperator for this program. Published 7.5-minute quadrangle maps within this study area include Candia (Gephart, 1985a), Derry (Gephart, 1985b), and Townsend, Mass. (Koteff and Stone, 1990). Four unpublished geologic quadrangles include Manchester North and Manchester South (Carl Koteff, U.S. Geological Survey, written commun., 1990), Greenfield, and Greenville (Carol Hildreth, Office of the New Hampshire State Geologist, written commun., 1990) and the New Hampshire parts of the Ashby, Mass.-N.H. and Ashburnham, Mass.-N.H. (Carol Hildreth, Office of the New Hampshire State Geologist, written commun., 1990). Koteff (1970) published the surficial geology of the Milford quadrangle at a scale of 1:62,500. Koteff and others (1984) discuss the surficial geology of the Merrimack Valley and the processes that led to the deposition of lacustrine and deltaic deposits. Numerous other studies were done by private consultants for local concerns.

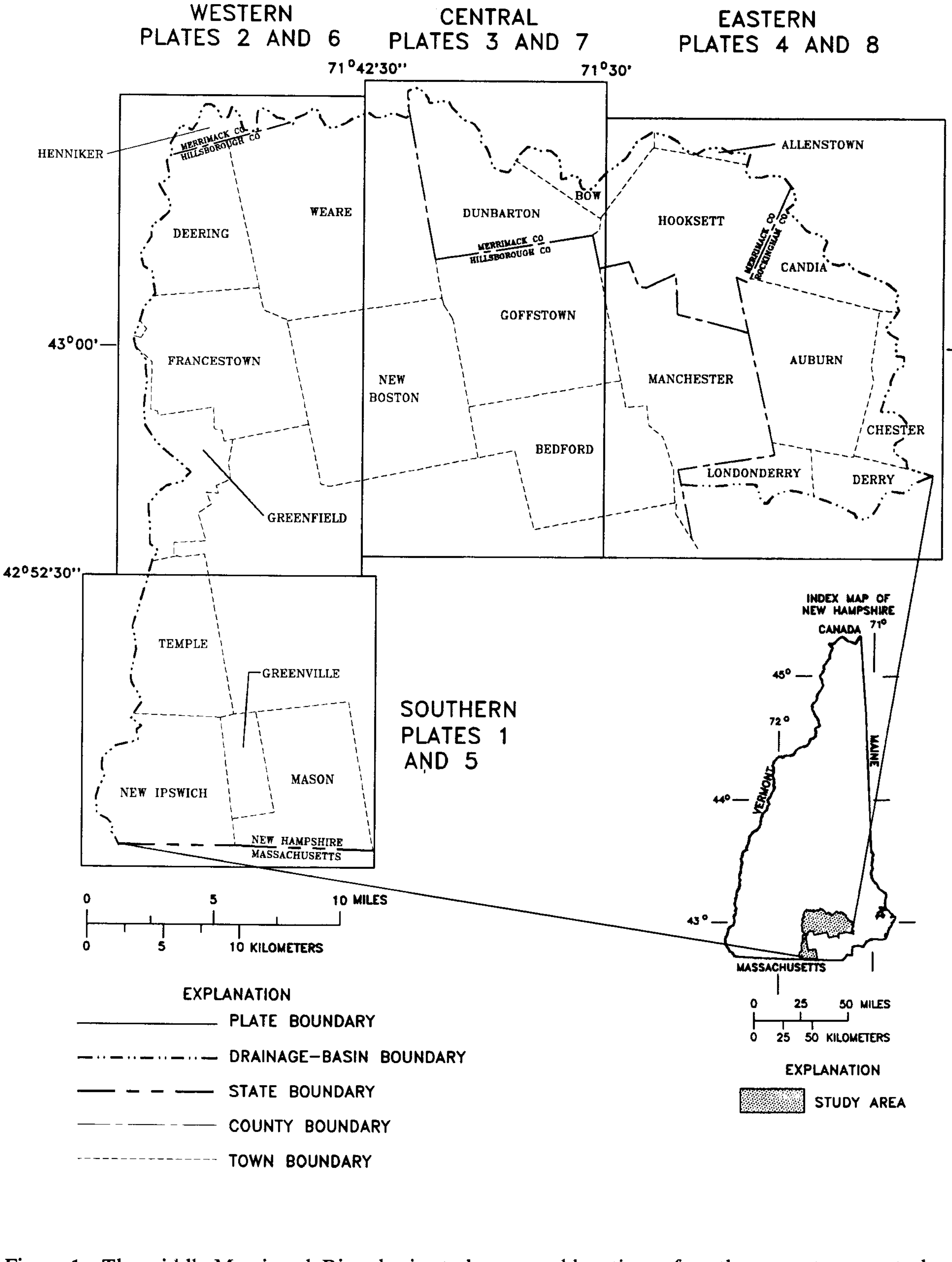


Figure 1.--The middle Merrimack River basin study area and locations of southern, western, central, and eastern plates in south-central New Hampshire.

Methods of Study

The following methods were used in this study:

1. Areal extent of the stratified-drift aquifers was mapped with the aid of soils maps from the U.S. Soil Conservation Service, surficial geologic maps, and data from the COGEOMAP program. Where no data were available, areal extent of stratified-drift aquifers was mapped by USGS personnel.
2. Published and unpublished subsurface data on ground-water levels, saturated thickness, and stratigraphy of the aquifers from the USGS, NHDES-WRD, and the New Hampshire Department of Transportation were compiled. Additional data were obtained from municipalities, local residents, well-drilling contractors, and geohydrologic consulting firms. The locations of wells, borings, and seismic lines were plotted on base maps, and pertinent well and boring data were added to the Ground Water Site Inventory (GWSI) data base maintained by the USGS. Each data point is cross-referenced to a site-identification number and to any other pertinent information about the site.
3. Seismic-refraction profiling, a surface geophysical technique, was used to determine depths to the water table and depths to the bedrock surface. (Locations of these profiles are shown on plates 1-4.) The seismic data were interpreted by using a time-delay, ray-tracing computer program developed by Scott and others (1972). Data from nearby wells and test holes were used to verify the interpretations. Actual depths to the bedrock surface are within 10 percent of the estimates from seismic-refraction profiling. Till is not identified in these interpretations because it is generally thin and cannot be distinguished from stratified drift by use of seismic-refraction methods. Where till is present but is not identified in the interpretation, the computed depth to bedrock is slightly less than the actual depth.
4. Seismic-reflection profiling, another surface-geophysical method, was used to determine depths to bedrock and to infer the sediment type of the aquifers that lie beneath water bodies. Haeni (1986, 1988b) outlines the methods for collecting seismic-reflection data. Seismic-reflection results differ from seismic-refraction results in that information about the texture of the subsurface can sometimes be inferred from the reflection records.
5. Test borings were made at more than 60 locations to improve definition of the thicknesses and geohydrologic characteristics of the stratified-drift aquifers. (Locations of test borings are shown on plates 1-4.) Split-spoon samples of the subsurface materials were collected at 5- to 10-ft intervals to estimate the horizontal hydraulic conductivity at those depths and to determine the stratigraphic sequence of materials comprising the aquifers. Where test borings were made in relatively productive aquifer materials, a 2-in.-diameter well with a polyvinyl chloride (PVC) casing and a slotted well screen was installed. Water levels were measured periodically in these wells, and water samples were collected from selected wells.
6. Data from items 2, 3, 4, and 5 were used to construct maps showing the water-table altitude and saturated thickness of the stratified-drift aquifers.
7. Hydraulic conductivities of the aquifer materials were estimated from grain-size distribution data from 454 samples of aquifer material collected during the completion of test borings and wells in southern New Hampshire. Transmissivities were estimated from logs of test borings and wells by assigning horizontal hydraulic conductivities to each interval sampled, multiplying the hydraulic conductivities by the saturated thickness of the interval, and summing these results. Additional transmissivities were obtained from reports by geohydrologic consultants and from analysis of aquifer-test data. This information was used to prepare maps showing the transmissivity distribution of the stratified-drift aquifers (pls. 5-8).
8. Flow-duration data from a long-term (1941-78) streamflow-gaging station on the South Branch Piscataquog River, in the middle of the study area, were analyzed and used to correlate miscellaneous low-flow measurements on ungaged streams. Streamflows measured where the stream flowed into and

out of major aquifers in the area during periods of low flow can be used to estimate potential aquifer yields.

9. An aquifer in Goffstown was selected to demonstrate a technique for estimating yield on the basis of a two-dimensional numerical model that simulates ground-water flow. The computer program, developed by McDonald and Harbaugh (1988), is a three-dimensional model that can be used to simulate flow in two dimensions. This model was used to estimate the potential yield and the sources of water to wells in the modeled area.
10. Samples of ground water from 10 wells constructed during this study were collected and analyzed. Selected physical properties (specific conductance, pH, temperature) were measured, and concentrations of inorganic constituents were determined. The data provided by these analyses were used to assess the general quality of water from the stratified-drift aquifers.

Numbering System for Wells and Borings

Local numbers assigned to wells and borings entered into GWSI consist of a two-letter town designation (table 1), a supplemental letter designation ("A" for borings done for hydrologic purposes, "B" for borings done primarily for construction purposes, and "W" for all wells in which a casing was set), and a sequential number within each town. For example, the first well in the town of Goffstown is GNW-1.

Acknowledgments

The authors thank the many State and Federal agencies, municipalities, residents, consulting firms, well-drilling companies, and private companies who provided data for this study.

GEOHYDROLOGIC SETTING

Three types of aquifers are present in the study area: (1) stratified drift, which can be a major source of ground water for municipalities; (2) till, which locally can supply minor amounts of water for domestic use; and (3) bedrock, which supplies water to most households in the study area that are not connected to a municipal supply.

Stratified Drift

Coarse-grained stratified drift, the focus of this study, consists of sorted, mostly coarse-grained sediments (sands and gravels) deposited by glacial meltwater at the time of deglaciation. Hydrologic characteristics of these sediments that affect ground-water storage and movement are related to the glaciofluvial environment in which the sediments

Table 1.--*Town codes used in the numbering system for wells and borings*

Town	Two-letter code	Town	Two-letter code
Allenstown	AF	Greenfield	GS
Auburn	AU	Greenville	GV
Bedford	BI	Hooksett	HT
Bow	BU	Manchester	MC
Candia	CD	Mason	MG
Deering	DE	New Boston	NC
Derry	DF	New Ipswich	NJ
Dunbarton	DN	Temple	TM
Francestown	FC	Weare	WG
Goffstown	GN		

were deposited. Stratified-drift deposits are composed of distinct layers of sediments with different grain-size distributions, sorted according to the depositional environment. For example, fast-moving meltwater streams deposit coarse-grained sediments with large pore spaces between grains while fine-grained sediments are washed downstream and deposited in slow moving meltwater. If saturated, these sediments will store and transmit water readily. Fine-grained deposits, which include very fine sand, silt, and clay, were deposited in lacustrine (lake) environments characterized by slow-moving and (or) ponded meltwater; these fine-grained deposits do not transmit water as readily as do the coarse-grained deposits.

The deglaciation process had a pronounced effect in determining the type of stratified-drift deposit that was formed. Deglaciation of the study area is believed to have been a systematic process of stagnation-zone retreat (Koteff and Pessl, 1981). As the active glacial ice receded to the north, zones of stagnant ice remained in contact with the active ice margin that was in the valleys. As the ice continued to recede, new sediment was continuously being brought forward to the active margin and was available for deposition. Most of the stratified-drift aquifers in the study area are valley-fill deposits that can be identified as lacustrine deposits, eskers, kames, kame terraces, kame deltas, outwash, and outwash deltas.

Stratified drift deposited during the deglaciation of the Merrimack River valley was affected by the presence of two large glacial lakes--glacial Lake Merrimack and glacial Lake Hooksett. Thick layers of fine-grained sediment accumulated in the lakes while meltwater from the glacier deposited relatively coarse-grained deltaic sediments within the same lacustrine environment. The maximum probable extent of glacial-lake sediments associated with glacial Lake Merrimack and glacial Lake Hooksett is shown in figure 2. Elevations of glacial-lake levels were projected from measured altitudes of the contact between topset and foreset beds in remnant deltas in the Merrimack River valley. The contacts represent the maximum probable level of the glacial lakes in that area (Carl Koteff, U.S. Geological Survey, written commun., 1990).

Glacial-Lake Deposits

Coarse sand and gravel near the Merrimack River was deposited as deltas and outwash plains, ice-contact deltas or kame deltas, or other fluvial deposits that graded to what was once glacial Lake Merrimack and glacial Lake Hooksett. These two large glacial lakes occupied the present day Merrimack Valley and extended into numerous tributary valleys including those now occupied by the Piscataquog River and Cohas Brook (Koteff and others, 1984). Lacustrine silt and clay do not transmit water readily and impede ground-water flow. The coarse-grained deltaic deposits that are found within the fine-grained lacustrine deposits form coarse-grained aquifers that store and transmit water readily. A large delta in the town of Hooksett, a good example of this type of aquifer, is currently used to supply water for the town. A block diagram of the development of a typical glaciolacustrine deltaic aquifer is shown in figure 3.

Upland Valley-Fill Deposits

The various types of deposits that comprise valley-fill aquifers in the upland parts of the study area are shown in a block diagram in figure 4. The best examples of these aquifers are in Weare, New Boston, and Goffstown, where they are associated with the drainage of the Piscataquog River and the South Branch Piscataquog River. Valley-fill aquifers in this area can be characterized by morphosequence deposition of kames and eskers grading into deltas and outwash plains that are associated with the retreat of the ice margin in an ice-stagnation zone; such deposition commonly resulted in formation of small, temporary glacial lakes near the ice margin (Koteff and Pessl, 1981). The aquifers are coarse grained at the ice-contact end and become progressively fine grained and better sorted where the sediment-laden meltwater lost energy downstream from the ice margin and within the glacial lakes. Where stratified drift is found in small discontinuous deposits, aquifers tend to have consistent texture.

Till

Till is an unsorted mixture of clay, silt, sand, gravel, and rock fragments deposited directly by glacial ice. In this study, till covers most of the bedrock surface and is overlain locally by stratified drift and recent stream deposits. The thickness of till in southern New Hampshire is commonly less than 15 ft but locally can be as much as several tens of feet thick (Bradley, 1964, p. 21). In south-central New Hampshire, till can be divided into an upper till and a lower till (Koteff, 1970, Goldthwait and others,

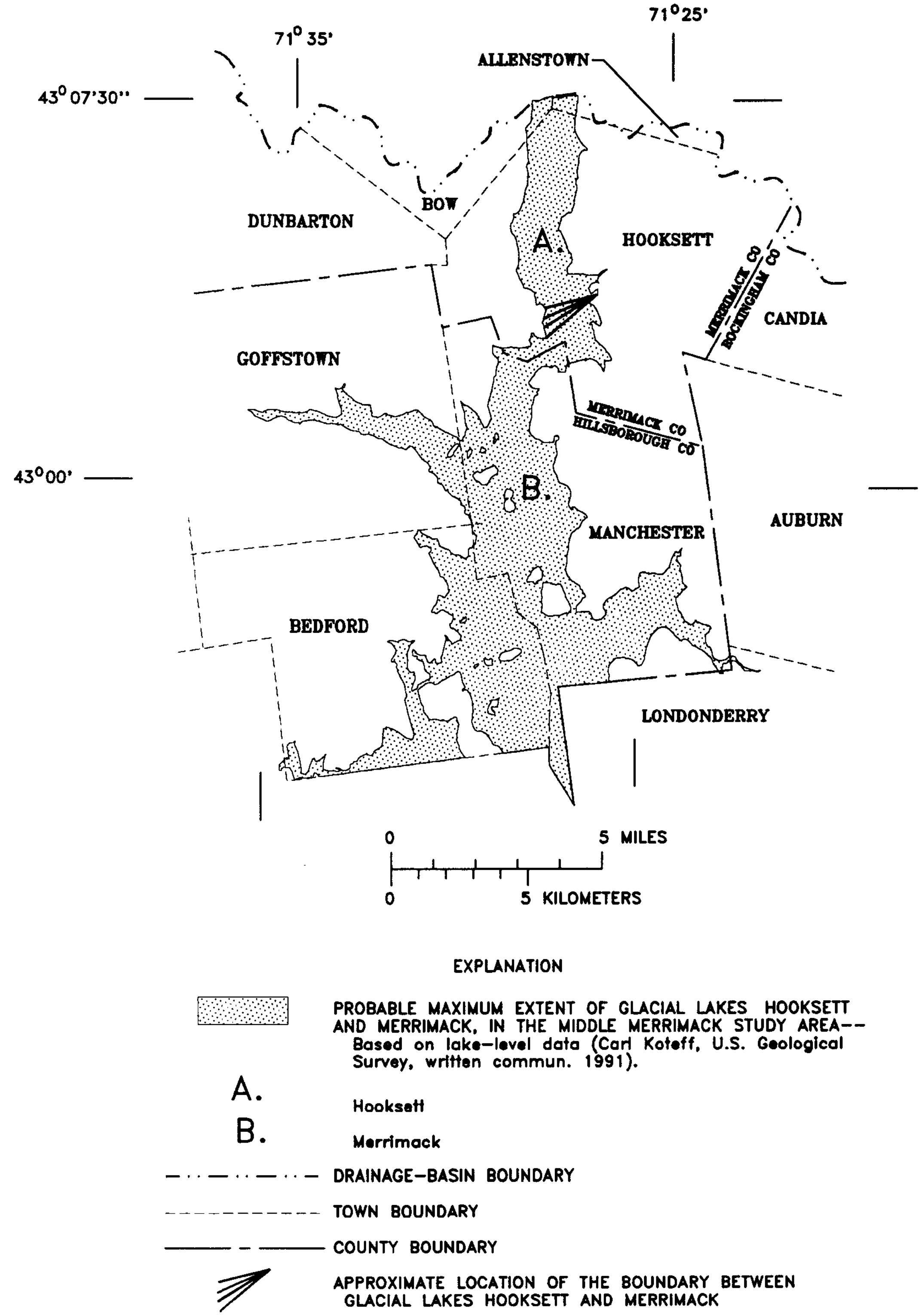


Figure 2.--Probable maximum extent of glacial Lakes Merrimack and Hooksett.

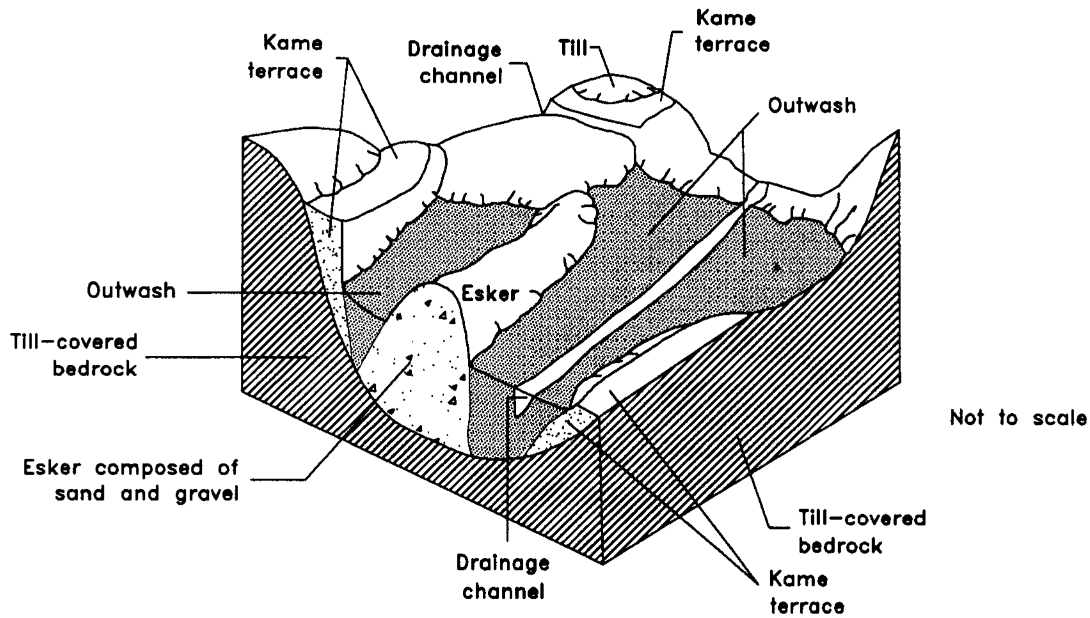


Figure 3.--A glaciolacustrine deltaic aquifer.

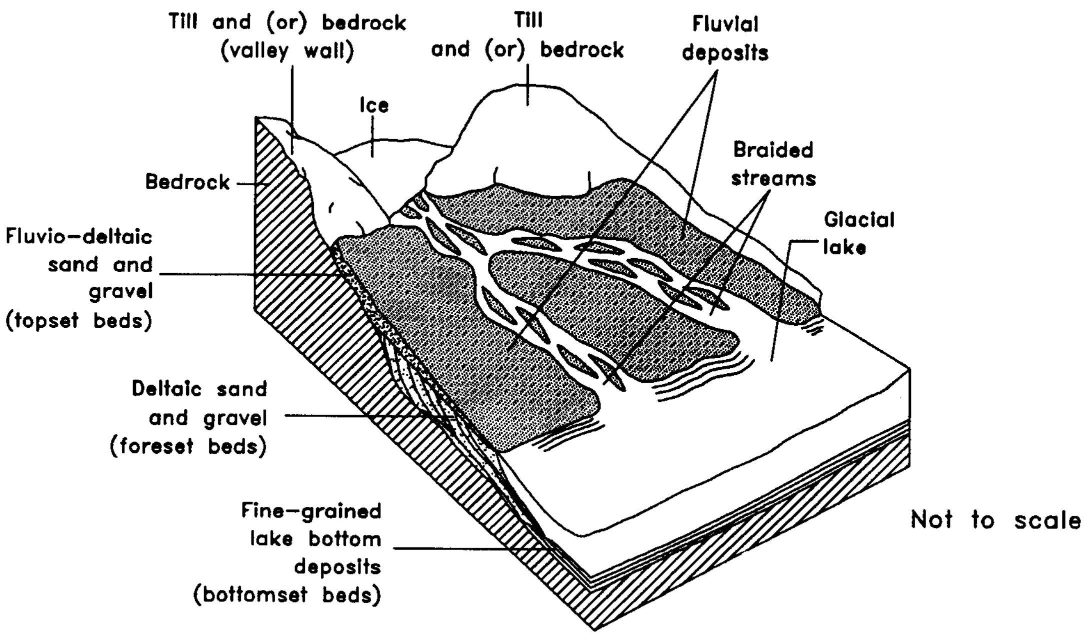


Figure 4.--A valley-fill aquifer composed of outwash and ice-contact deposits.

1951). The two tills are thought to represent two separate major ice advances over the area (Koteff, 1970).

Till is generally not considered to be a major source of ground water because of its low hydraulic conductivity. Large-diameter dug wells completed in till can provide modest amounts of water (commonly less than 3 gal/min) for household needs, but water-level fluctuations within till can be large enough to make these wells unreliable during dry seasons.

Bedrock

Bedrock in the southeastern part of the study area, southeast of the Campbell Hill Fault, consists primarily of metamorphic rocks of pre-Silurian and Precambrian age, including gneiss, slate, schist, quartzite, and metavolcanic rocks (Lyons and others, 1986). Towns or municipalities in this area are Auburn, Bedford, Candia, Manchester, and southeastern parts of Goffstown, Hooksett, and Mason (fig. 1). Bedrock in the northwestern part, northwest of the Campbell Hill Fault, consists primarily of Ordovician to Silurian schist, gneiss, and quartzite. In both sections of the study area, these rocks were intruded by granite, granodorite, syenite, monzonite, and diorite of Devonian and Silurian age (Lyons and others, 1986). The rocks trend in northeasterly belts that parallel the region's structural grain (Lyons and others, 1986). Major fault zones trend northeasterly and are parallel to regional structure. Secondary fractures cut across the primary fractures.

Ground water enters wells completed in bedrock through fractures that are intersected by the well. The yields of these wells depend on the number, size, and degree of interconnection of the fractures. Wells that tap bedrock commonly yield small supplies of water suitable for drinking and other domestic uses. The yields of bedrock wells inventoried for this study ranged from 0.25 to 150 gal/min; the median yield was 7 gal/min. Bedrock wells are capable of yields sufficient for municipal supply where fractures are large and numerous (Cotton, 1985).

GEOHYDROLOGY OF STRATIFIED-DRIFT AQUIFERS

The geohydrology of stratified-drift aquifers was described by identifying (1) aquifer boundaries, (2) direction of ground-water flow from recharge to discharge areas, (3) aquifer thickness and storage, and (4) aquifer transmissivity. Data sources in this investigation included surficial geology maps, records of wells and test borings, and seismic-refraction and seismic-reflection data. Results of the geohydrologic investigation are presented on plates 1-8 and in the text that follows.

Delineation of Aquifer Boundaries

Stratified-drift aquifers in the study area are composed of fine- to coarse-grained sands or sands and gravels deposited by glacial meltwaters; these deposits, in part, are now sufficiently saturated to yield significant quantities of water to wells and springs. The lateral boundaries of the aquifers are defined as the contacts between the stratified drift and till or bedrock valley walls. The position of the contact was determined by use of surficial geology maps, soil maps, test-boring logs, and field mapping done specifically for this study. The bottom boundary is the till and (or) bedrock surface and was determined from analysis of data from seismic-refraction and seismic-reflection surveys, test borings, and domestic water wells. The upper boundary is the water table.

Areal Extent of Aquifers

The areal extent of the stratified-drift aquifers is shown on plates 1-8. Because of the regional scale of this investigation, aquifer boundaries are approximate. The approximate limits of lacustrine deposits associated with glacial Lakes Merrimack and Hooksett are shown in figure 2. Coarse-grained stratified-drift aquifers may be present beneath fine-grained lacustrine deposits but may not have been identified because of the complexity of the stratigraphy and the lack of data. Available data for coarse sediment underlying fine-grained sediment are discussed in the section "Descriptions of Selected Stratified-Drift Aquifers." Although the lacustrine clay, silts and very fine sands are not capable of supplying adequate amounts of water for domestic and community use, the coarse-grained deposits that may lie below could be productive aquifers.

All aquifer boundaries are shown as solid lines. In the explanation on the plates, solid lines represent "approximately located" boundaries. The lateral boundaries of stratified-drift aquifers were delineated from the previously cited published and unpublished surficial geology maps and by interpretation of soil maps of Rockingham, Hillsborough, and Merrimack Counties (Koteff, 1970; Gephart, 1985a,b; Carol Hildreth, Office of the New Hampshire State Geologist, written commun., 1990; Carl Koteff, U.S. Geological Survey, written commun., 1990; Koteff and Stone, 1990).

Stratigraphy of Geohydrologic Units

Data used to define the stratigraphy of geohydrologic units were obtained from existing records of subsurface exploration within the project area. Other test drilling and surface geophysical exploration (seismic refraction and marine seismic reflection) were done to delineate texturally different geohydrologic units within the stratified drift.

Well and boring data

Subsurface data from wells and borings were inventoried, and data locations within the stratified-drift aquifers are plotted on plates 1-4. Geohydrologic data for approximately 3,350 sites have been added to the GWSI data base and checked for accuracy. Data for approximately 2,600 of the 3,350 sites were transferred to GWSI from the NHDES-WRD well-inventory data base. Approximately 420 of the 2,600 NHDES-WRD sites are within stratified-drift aquifer areas. Approximately 750 sites of the 3,350 total sites added to the data base are located in the stratified-drift aquifer areas. Appendix A contains selected data from the GWSI data base for wells and borings within the stratified-drift aquifer areas that were used to construct the accompanying map plates. These data include an identification number for the well, latitude and longitude, depth of the well, water level, and yield of the well. Appendix B contains stratigraphic logs of selected wells and borings in stratified drift. These data were used primarily for estimating the transmissivity of the aquifers where no aquifer-test data or grain-size data were available.

Seismic-refraction data

Seismic-refraction profiles, totaling over 10 mi, were completed at 76 locations to determine depths to the water table (pls. 1-4) and depths to the bedrock surface (pls. 5-8). A 12-channel, signal-enhancing seismograph was used to record arrival times of compressional wave energy generated by a sound source. The data were collected and interpreted according to methods described by Haeni (1988a). The interpretations, made with the aid of a computer program developed by Scott and others (1972), are shown in appendix C. Estimated depths to the water table and to the bedrock surface are generally compared with control data, such as nearby well or boring logs and water-table and bedrock-outcrop observations. The accuracy of the depths to water table and bedrock are within 10 percent of the true depth, as determined from test borings made along selected profiles.

Seismic-reflection data

High-resolution, continuous seismic-reflection data were collected according to methods described by Haeni (1986) along the approximately 14-mi-long reach of the Merrimack River within the study area. Data were also collected on navigable reaches of the Piscataquog River from near its mouth upstream to the town of Goffstown. These data were used to map depths to the bedrock surface beneath the two rivers. During data collection, an array of receivers was towed behind a boat that traveled slowly up or down the river. Compressional waves, generated from a sound source, penetrated the river bottom and were reflected back to the surface in response to the physical differences in the geologic strata. The reflected sound waves were received at the water surface and converted to an electrical signal displayed on a graphic recorder. Data-collection was often affected by the presence of strong reflectors at the water bottom causing multiples of the water-bottom record to obscure any data below. This technique is discussed in detail by Haeni (1988b) and by Morrissey and others (1985).

Altitude of the Water Table

The approximate altitude of the water table within the stratified drift is shown on plates 1-4. These maps were constructed from (1) altitudes of streams, ponds, rivers, and lakes as shown on 1:24,000-scale USGS topographic maps; (2) water-level data from wells stored in GWSI; and (3) analysis of seismic-refraction data. Ground-water altitudes in fine-grained lacustrine deposits represent the ground-water altitude in those deposits only. Saturated coarse-grained stratified drift may be present below fine-grained material in some areas, and a second, deeper potentiometric surface (in confined aquifers) may be present.

Water-level measurements were made seasonally at selected wells in the study area during 1988 and 1989 and were stored in GWSI. Long-term hydrographs showing water levels in two representative wells (MOW-36 and CVW-4) near the study area are shown in figure 5. Well MOW-36 represents water-level fluctuations in a medium- to coarse-grained stratified-drift aquifer. Well CVW-4 represents water-level fluctuations in fine sands, silts, and clay. The data from these wells support the conclusion reached for other parts of New Hampshire (Cotton, 1987; Toppin, 1987; Moore, 1990; Mack and Lawlor, 1992; Moore and others, in press) that natural water-level fluctuations in coarse-grained stratified drift are usually less than 5 ft but can be as much as 10 ft; therefore, a 20-ft contour interval for water-table altitudes under natural conditions is reasonable for a generalized water-table map constructed from water-level measurements made at different times.

Recharge, Discharge, and Direction of Ground-Water Flow

Ground-water recharge includes natural recharge from precipitation that falls directly on the aquifer and infiltrates the water table, lateral inflow from adjacent till and bedrock areas, and, in some places, leakage from streams that traverse the aquifer. Natural recharge is the difference between precipitation and the amount of water lost to evapotranspiration and to surface runoff.

Recharge to stratified-drift aquifers in this study can be estimated from stream-discharge measurements made during periods in which there is no change in ground-water storage, as indicated by the position of the water table. Such estimates require the assumption that the ground-water discharge consists of mostly ground-water runoff. During periods of low flow and after several days without precipita-

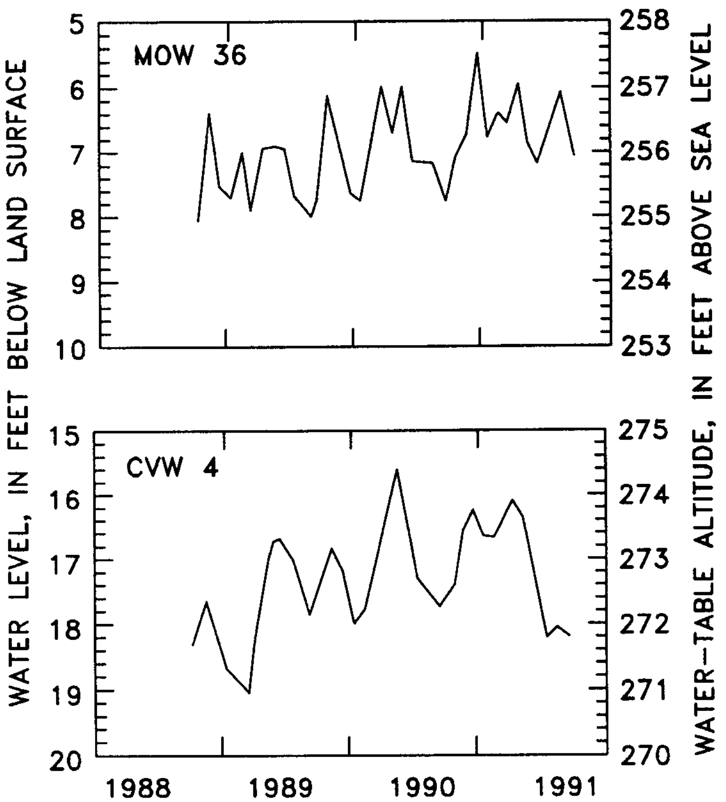


Figure 5.--Water-level hydrographs of observation wells MOW-36 and CVW-4.

tion, the assumption is reasonable. This method probably gives conservative estimates of natural recharge to aquifers.

Streamflow-gaging station 01091000, on the South Branch Piscataquog River in the central part of the study area (pl. 3), was used to monitor flow conditions in the basin. On September 6, 1989, streamflow at this site was at a rate equaled or exceeded 93 percent of the time after 5 days without precipitation. Under these conditions, flow within the basin was low, and ground-water discharge was assumed to be natural recharge from ground-water runoff.

Recharge to stratified-drift aquifers from streams that lose water to the aquifer through permeable streambeds was documented by Randall (1978) and by Morrissey and others (1989). This type of recharge was not observed in any of the base-flow measurements made in this study, although it probably occurs on a small scale within the ±5-percent error associated with base-flow measurements. Such tributary-stream infiltration is common where the tributary streams flow into aquifers that have a deep water table at the stratified-drift, till, and (or) bedrock contact relative to the streambed altitude (D.J. Morrissey, U.S. Geological Survey, written commun., 1989).

Recharge to the stratified-drift aquifers comes, in part, from adjacent till and (or) bedrock uplands. Lateral inflow from upland areas not drained by perennial streams recharges the stratified-drift aquifer at the till and (or) bedrock contact. Recharge to stratified-drift aquifers from upland areas not drained by streams can be estimated by measuring ground-water discharge from till and (or) bedrock uplands that are drained by streams. For a stream in Maine, the estimated average annual lateral inflow of ground water from upland areas to a stratified-drift aquifer was 0.5 $(ft^3/s)/mi^2$ (Morrissey, 1983). Upland areas not drained by streams are generally small but may contribute a significant amount of recharge to aquifers.

Ground-water discharge includes natural leakage into streams, lakes, and wetlands; ground-water evapotranspiration; and withdrawal from wells. During periods of low streamflow, usually in late summer and early fall and after several days without rainfall, streamflow consists almost entirely of ground-water discharge. Streamflow measurements were made during such periods in October 1988 and September 1989 (appendix D). These measurements represent approximately 90-percent flow duration and 93-percent flow duration, respectively. Most of this discharge is assumed to be ground-water runoff, and, thus, it can be used as an estimate of recharge to aquifers in the study area. Further discussion of these measurements is found in the section of the report titled "Description of Selected Stratified-Drift Aquifers."

Direction of ground-water flow in an unconfined aquifer is determined by the water-table gradient. Water-table gradients differed throughout the study area depending on topography and hydraulic conductivity of the stratified-drift deposits. Water-table gradients in fine-grained stratified drift commonly exceeded 5 percent in areas of high topographic relief. Water-table gradients in coarse-grained stratified drift in areas of low topographic relief were less than 0.1 percent. Potentiometric surfaces within confined aquifers (coarse-grained deposits beneath fine-grained deposits) were not contoured because of insufficient data.

Aquifer Characteristics

The geohydrology of stratified-drift aquifers shown on plates 5-8 is based partly on aquifer characteristics that include saturated thickness, storage, and hydraulic conductivity. Estimates of saturated thickness and hydraulic conductivity were used to calculate transmissivity (pls. 5-8). These properties can be used to assess the water-supply potential of stratified-drift aquifers. Values of aquifer storage can be used to estimate aquifer yield.

Saturated Thickness and Storage

Saturated thickness of an unconfined stratified-drift aquifer is the vertical distance between the water table and the base of the aquifer. For many stratified-drift aquifers, the bottom is the till or bedrock surface; for other aquifers, the bottom is the contact between the upper coarse-grained deposits and the underlying fine-grained lacustrine deposits. Saturated thicknesses depicted on plates 5-8 include these fine-grained deposits. Saturated-thickness contours were constructed from test-boring and well data and seismic-refraction and seismic-reflection profiles. The saturated thickness multiplied by the specific yield of an unconfined aquifer determines the amount of ground water that can be released from storage.

The storage coefficient of an aquifer is defined as the volume of water released from or taken into storage per unit surface area of aquifer per unit change in head (Lohman and others, 1972). In unconfined aquifers, the storage coefficient is approximately equal to the specific yield--the amount of water released by gravity drainage from a unit volume of aquifer per unit decrease in hydraulic head. A value of 0.2 is commonly used for specific yield for stratified-drift aquifers in New England (Moore, 1990) and for unconsolidated deposits in other areas (Freeze and Cherry, 1979). Specific yields of 13 samples of stratified drift from southern New Hampshire ranged from 0.14 to 0.34, with an average of 0.26 (Weigle and Kranes, 1966). On the basis of data collected during a 3-day aquifer test done for this study in the Goffstown aquifer, the specific yield ranged from 0.21 to 0.29.

Water released from storage in confined aquifers results from expansion of water and compression of the aquifer as hydraulic head declines. Storage coefficients for confined aquifers, which are significantly smaller than specific yields for unconfined aquifers, range from 0.00005 to 0.005. Small storage coefficients indicate that the amount of water derived from expansion and aquifer compression is much less than that from dewatering by gravity drainage.

Saturated-thickness maps can be used to estimate the amount of ground water stored in an aquifer that is available for use. The saturated volume of an unconfined aquifer is approximately equal to the sum of the products of the areas between successive pairs of saturated-thickness contours multiplied by the average saturated thickness for each area. The actual volume of ground water stored in the aquifer is the product of the saturated volume multiplied by the porosity.

Saturated-thickness maps (pls. 5-8) were constructed from data obtained from surficial geologic maps, seismic-refraction and seismic-reflection profiles, and records of well and test borings. Saturated thicknesses exceeded 120 ft in places. The values calculated for saturated thicknesses included the thickness of all stratified drift regardless of grain size. Layers of clay, silt, and fine sand that overlie, underlie, or are interfingered with the aquifer deposits are included in the thicknesses depicted on plates 5-8. This inclusion of fine material is important to note where glacial Lake Merrimack and glacial Lake Hooksett deposits are present along the Merrimack River and associated tributaries (fig. 2).

Transmissivity and Hydraulic Conductivity

Aquifer transmissivity is defined as the rate at which water at the prevailing kinematic viscosity can be transmitted through a unit width of an aquifer under a unit hydraulic gradient (Lohman and others, 1972). The transmissivity (T) of an aquifer is equal to the saturated thickness (b) multiplied by the horizontal hydraulic conductivity (K, a directional measure of the permeability) and is expressed in feet squared per day; thus,

$$T = Kb, \tag{1}$$

Aquifer transmissivity at a specific site was derived from estimates of hydraulic conductivity of lithologic units in the aquifers. Hydraulic conductivity, in turn, was estimated from grain-size distributions of samples of aquifer materials by use of the regression equation developed by Olney (1983). Hydraulic conductivity, however, which has a vertical and a horizontal vector component, is not accounted for by this equation. In this relation, an effective grain size (D_{10}, in Phi units) was used to estimate hydraulic conductivity (K, in feet squared per day) with the following equation:

$$K = 2{,}100 \times 10^{-0.655(D_{10})}, \tag{2}$$

The effective grain size (D_{10}) is a controlling factor for the hydraulic conductivity of stratified drift in New Hampshire. Effective grain size is defined as that grain size where 10 percent of the sample is finer than the effective grain size and the remaining 90 percent is coarser than the effective grain size. Olney (1983) developed this relation from the results of permeameter tests of stratified-drift samples from Massachusetts. Moore (1990) found that this relation yielded results that fall within the range of results from other relations that have been developed between grain size-size distribution and hydraulic conductivity (Krumbein and Monk, 1942; Bedinger, 1961; and Masch and Denney, 1966). Comparisons with aquifer-test data, however, indicate that equation 2 may not give accurate results for very coarse-grained sand and (or) gravel. Estimates of hydraulic conductivity for aquifers with coarse sands and gravels were, in part, based on comparisons to aquifer-test data for similar deposits. Hydraulic conductivity (and transmissivity) based on grain-size relations are only estimates and may differ significantly from results of aquifer tests.

Hydraulic conductivity was estimated for 454 samples of stratified drift from southern New Hampshire by means of equation 2. The samples were collected in the Exeter and Lamprey River basins (Moore, 1990); in the seacoast area and the lower Merrimack River basin (Flanagan and Stekl, 1990); in the Bellamy, Cocheco, Salmon Falls River basins (Mack and Lawlor, 1992); in the lower Connecticut River basin (Moore and others, in press); in the Contoocook River basin (P.T. Harte and others, U.S. Geological Survey, written commun., 1991), and for this study. The grain-size distribution and the effective grain size (D_{10}) were determined for these 454 samples.

Hydraulic conductivities calculated from equation 2 were plotted against median grain size in phi groups, and the resulting plot was divided into three categories of degree of sorting (fig. 6). These categories are strictly relative and are used to describe the types of stratified-drift aquifer deposits found in New Hampshire. The degree of sorting was based on the standard deviation of each individual sample. These relative categories are described in the following paragraph.

If standard deviations were large (greater than 1.75 phi), the samples were considered "poorly sorted"; if standard deviations were intermediate (1.25 phi to 1.75 phi), the samples were considered "moderately sorted"; and if standard deviations were small (less than 1.25 phi), the samples were considered "well sorted." A regression equation was developed for each of the three categories to determine the relation between hydraulic conductivity and median grain size (fig. 6). The coefficient of determination (R^2) was 0.93 for the "well sorted" samples, 0.72 for the "moderately sorted" samples, and 0.54 for the "poorly sorted" samples. The calculated hydraulic conductivity, grouped by ranges of median grain size and by ranges of standard deviation (degree of sorting), is shown in table 2.

Hydraulic conductivities were calculated for each median phi group and were averaged to determine a mean hydraulic conductivity per group. For example, the mean hydraulic conductivity of sediment samples whose median grain size was described as medium sand and "well sorted" was 38 ft/d (the average of 25 and 51 ft/d; table 2).

Very fine sand, silt, and clay deposits in the study area were not analyzed for grain-size distribution because their hydraulic conductivities are typically low (less than 4 ft/d) and, therefore, considered insignificant (Todd, 1980). Estimates of horizontal hydraulic conductivity for coarse sand and gravel were determined by analysis of aquifer-test data from municipal wells in Goffstown and Hooksett. Such data were not available elsewhere.

The values in table 2 were used to estimate hydraulic conductivities from lithologic descriptions given in logs from test borings and wells. For example, for a lithologic description of 10 ft of

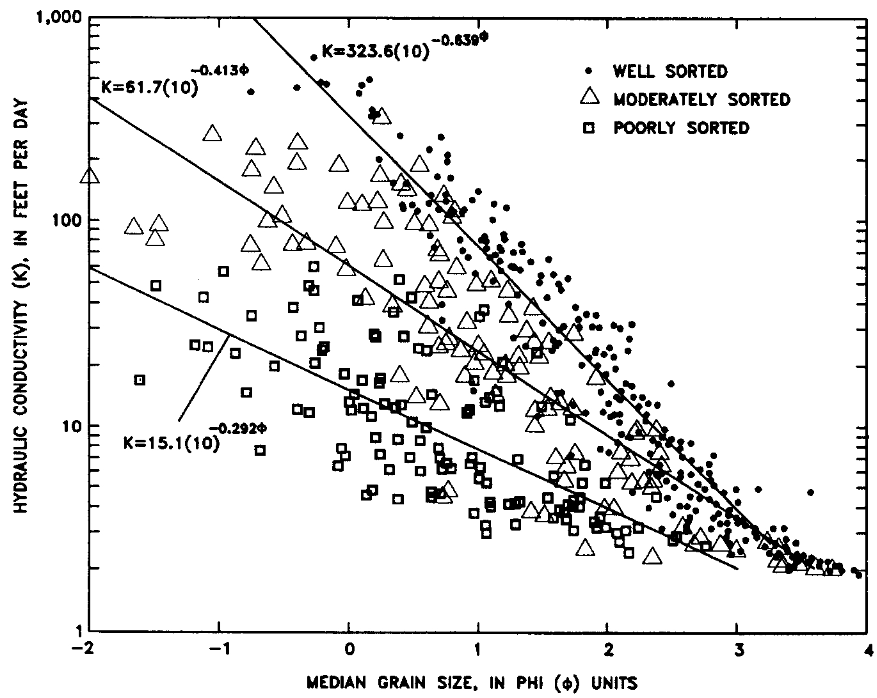


Figure 6.--Relation between estimated hydraulic conductivity, median grain size, and degree of sorting.

Table 2.--*Relation of mean hydraulic conductivity to median grain size and degree of sorting of aquifer material*

[<, less than; >, greater than; --, no data]

Median grain size (phi units)	Median grain description	"Well sorted" standard deviation <1.25 phi	"Moderately sorted" standard deviation 1.25 phi to 1.75 phi	"Poorly sorted" standard deviation >1.75 phi
		Mean hydraulic conductivity (K), in feet per day		
-1.75	granules	--	320	49
-1.25	granules	--	200	35
-.75	very coarse sand	970	120	25
-.25	very coarse sand	470	78	18
.25	coarse sand	220	48	13
.75	coarse sand	110	30	9
1.25	medium sand	51	19	7
1.75	medium sand	25	12	5
2.25	fine sand	12	7	3
2.75	fine sand	6	4	2
3.25	very fine sand	3	3	--
3.75	very fine sand	2	2	--

"moderately sorted" coarse sand overlying 20 ft of "well sorted" fine sand overlying bedrock, the hydraulic conductivities assigned would be 39 ft/d (the average of 30 and 48 ft/d) and 9 ft/d (the average of 12 and 6 ft/d), respectively. The estimate of transmissivity, based on the same description, would be (10 ft $\times$ 39 ft/d) + (20 ft $\times$ 9 ft/d) or 570 ft^2/d.

Descriptions of Selected Stratified-Drift Aquifers

The most extensive and most productive stratified-drift aquifers in the study area (fig. 7) are discussed in this section. Aquifers are discussed from south to north and from west to east, beginning at the southwestern part of the area.

Stratified-drift aquifers that are shown on the southern plate (pl. 1) are characteristically thin and discontinuous and may contain fine-grained glaciolacustrine sediment. Whereas these aquifers may not be useful for a municipal supply, they may be adequate for domestic supply. Accordingly, many old homes in the area have shallow dug wells. The most productive aquifers are described below.

Gould Mill Brook Aquifer

The Gould Mill Brook aquifer, largely in the eastern part of the town of Mason and partly in the town of Brookline (pl. 1, fig. 7), is composed of sand, gravel, and minor silt deposited in a glaciolacustrine environment. The aquifer can be as much as 70 ft thick (Koteff and Stone, 1990). Test drilling of well MGW-1 (pl. 1) revealed 40 ft of saturated sand and a transmissivity greater than 1,000 ft^2/d. This deposit is thinner south of MGW-1.

Smithville Aquifer

The Smithville aquifer is in the town of New Ipswich, 0.4 mi west of Smithville at the Smithville Flood Control dam (pl. 1, fig. 7). Saturated thickness of this aquifer exceeds 60 ft, and the transmissivity ranges from 1,000 to 2,000 ft^2/d. Water retained behind an earthen dam may enhance recharge and storage in the aquifer and increase its potential for ground-water resources.

Upper Stony Brook Aquifer

The upper Stony Brook aquifer, at Russell Station Road in the town of Greenfield (pl. 2, fig. 7), is 2.6 mi southeast of Greenfield Village center at the headwaters of Stony Brook. This small, isolated aquifer is mainly composed of coarse-grained stratified drift. The saturated thickness is approximately 20 ft throughout much of the aquifer and is greater than 40 ft in the center of it. Although the aquifer is small, the coarse-grained, uniform sediment make this a potentially productive aquifer for a domestic or small community water supply. Estimated transmissivities exceeded 4,000 ft^2/d in the deep, central zone. A small part of the aquifer is within the town of Lyndeborough (Toppin, 1987; pls. 1-2).

Upper South Branch Piscataquog River Aquifer

The upper South Branch Piscataquog River aquifer, which begins in Francestown near the headwaters of the South Branch Piscataquog River (pl. 2, fig. 7), extends along the river valley south into Lyndeborough and heads northeast from there into New Boston. The saturated thickness of this aquifer is generally greater than 20 ft but exceeds 80 ft in the deeper sections near the Lyndeborough-New Boston town line. Part of the aquifer fills an overdeepened channel scoured by glacial ice along the Francestown Turnpike. Saturated thicknesses range from 40 to 60 ft (pl. 6). Test borings south of Francestown along the Francestown Turnpike indicate that fine-grained lacustrine deposits are present throughout the entire saturated thickness of the aquifer. Similar fine-grained sediment is also present at the southernmost part of the aquifer near the confluence of Cold Brook. To the northeast, test borings (NCW-186) and surface observations indicate coarse-grained deposits and saturated thicknesses that exceed 40 ft. Cobble gravel at the surface prevented drilling and sampling in the area adjacent to Lyndeborough Road, 1 mi southwest of New Boston. Kames and eskers in this area indicate the likelihood of coarse-grained deposits below the surface. Transmissivity in this part of the aquifer is probably less than 2,000 ft^2/d, but exceeds 1,000 ft^2/d, indicating the potential for domestic or small community water supply.

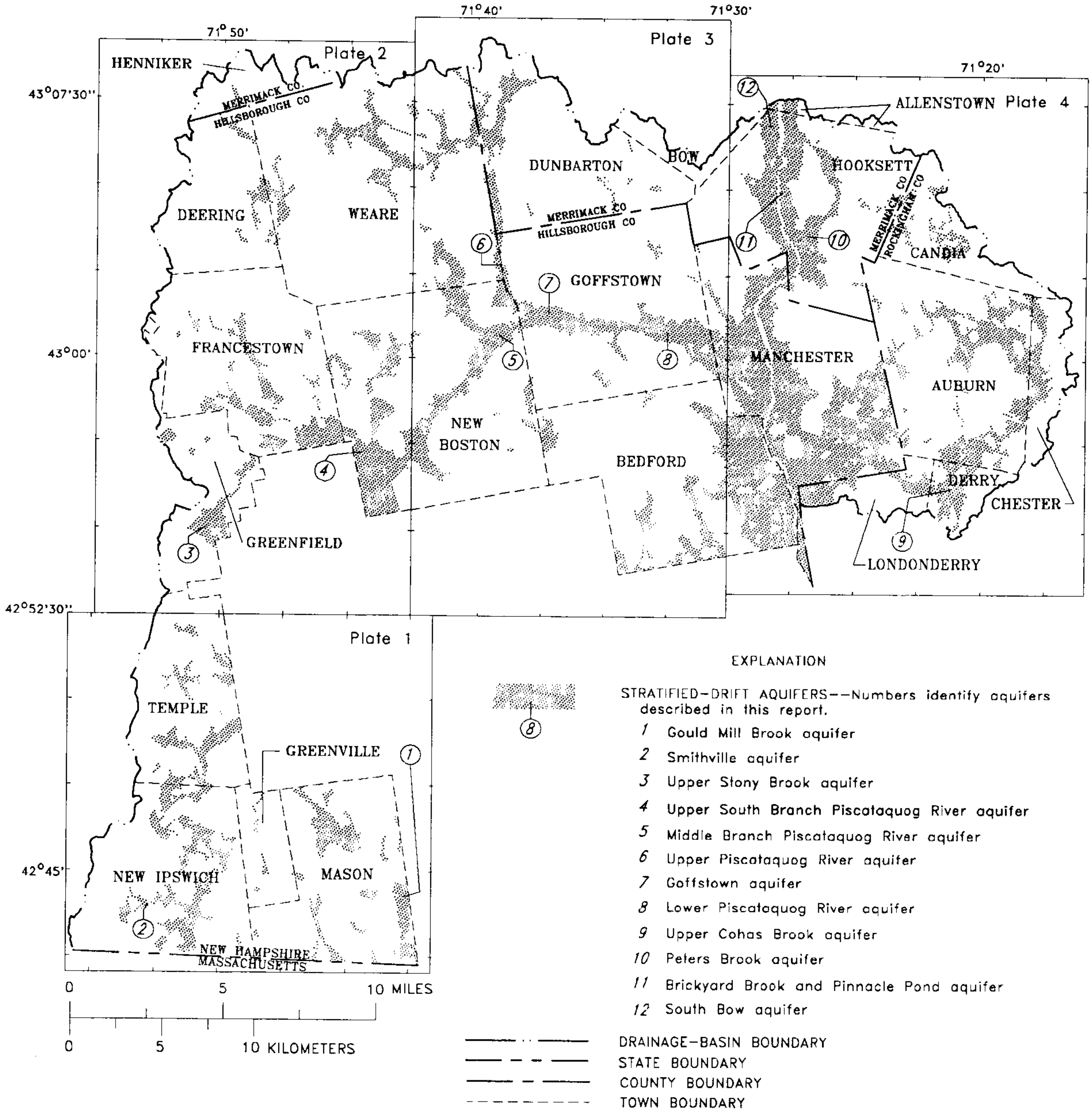


Figure 7.--Locations of aquifers.

Middle Branch Piscataquog River Aquifer

The Middle Branch Piscataquog River aquifer (pl. 3, fig. 7) extends north up the valley of the Middle Branch Piscataquog River from the confluence of the Middle Branch Piscataquog River and the South Branch Piscataquog River in New Boston. An extensive delta at the confluence is thought to be coarse grained and a potentially productive area within the aquifer; however, no data are available because of lack of access to the property. Saturated thickness near the delta may exceed 60 ft. Saturated thicknesses are greater than 20 ft in the middle and upper parts of the stratified-drift aquifer and exceed 40 ft in the northern part. Estimated transmissivity for the northern area exceeds 1,000 ft^2/d. Several year-round and seasonal residents currently withdraw water from shallow dug wells.

Upper Piscataquog River Aquifer

The upper Piscataquog River aquifer in Weare (pl. 3, fig. 7) below Everett Lake is bounded by Everett Dam to the north and by Riverdale Dam to the south (pl. 2). This narrow, north-south valley-fill aquifer has an average width of 1,500 ft and consists of less than 10 ft of medium to coarse-grained saturated valley-fill deposits overlying thick fine-grained sands and silt. The saturated thickness of this deposit is greater than 90 ft in places, despite the narrowness of the valley. The estimated transmissivity of fine-grained deposits is less than 2,000 ft^2/d. An aquifer with this transmissivity may not be adequate to support a municipal water supply, but it is suitable for a domestic or small community water supply. Currently (1992), the Kuncanowet Hills Mobile Home Park withdraws approximately 5,000 gal/d (M.A. Horn, U.S. Geological Survey, written commun., 1992) from two shallow dug wells in this aquifer to supply water to about 80 people (New Hampshire Water Supply and Pollution Control Division, written commun., 1991).

Flow measurements made on the Piscataquog River at sites 11 and 13 (appendix D) on September 6, 1989, indicate an approximate increase in flow of 1.13 ft^3/s (500 gal/min) in 3.6 mi. These flow measurements can be used as an index of aquifer yield.

Goffstown Aquifer

The Goffstown aquifer is immediately west of the Goffstown town center and underlies an area of approximately 1.2 mi^2 (pl. 3, fig. 7). The aquifer is situated at the confluence of the Piscataquog River and the South Branch Piscataquog River. Saturated thicknesses in this aquifer exceed 60 ft in places and average 40 ft (fig. 8). Transmissivities for the coarse-grained sediments exceed 8,000 ft^2/d in the center of the aquifer and average greater than 2,000 ft^2/d for most of the aquifer. The aquifer is thought to be composed of deltaic sands and gravels that were deposited in glacial Lake Merrimack when it occupied this part of the Piscataquog River valley.

A 3-day aquifer test was done by the USGS by pumping well GNW-1 at a well field owned by the Goffstown Water Precinct. Time, distance, and drawdown data obtained during the test were analyzed according to a method by Neuman (1974) that accounts for partially penetrating wells. Horizontal hydraulic conductivities near well GNW-1 ranged from 250 to 350 ft/d. The average ratio of horizontal to vertical hydraulic conductivity was 10:1. Transmissivities were determined to be greater than 9,000 ft^2/d at GNW-1. Two wells at this site are currently used to augment water withdrawals from a surface-water reservoir and average daily withdrawals in 1990 were about 0.071 Mgal/d for August, September, and October (New Hampshire Department of Environmental Services, Water Management Bureau, written commun., 1991). This small aquifer may be capable of supplying large amounts of water for municipal use. The yield of this aquifer is discussed in detail in the section "Estimation of Aquifer Yield for the Goffstown Aquifer."

Lower Piscataquog River Aquifer

The lower Piscataquog River aquifer is downstream from Goffstown village and is almost entirely within the area formerly occupied by glacial Lake Merrimack (pl. 3, fig. 7). Generally, this aquifer has low potential for ground-water withdrawal because of the extent of fine-grained lacustrine deposits associated with the former glacial lake. Most of the fine-grained stratified drift is overlain by thin, medium to coarse-grained delta and (or) outwash sands and gravel. Throughout the aquifer, the Piscataquog River has eroded to the bedrock surface, which defines the aquifer bottom in this area. Most of the lower Piscataquog River stratified-drift

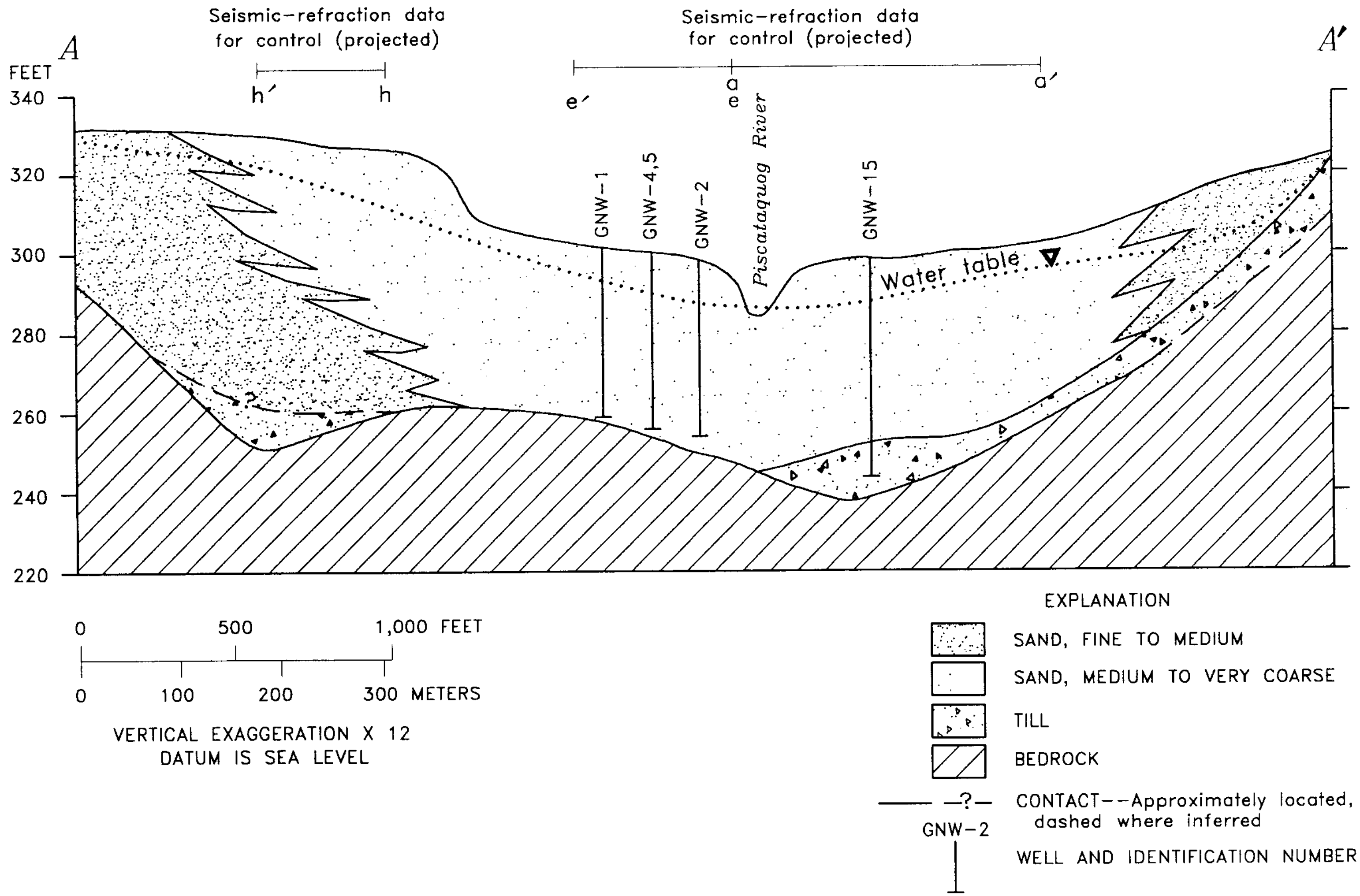


Figure 8.--Geologic section through the Goffstown aquifer.

deposits are terraced and are composed of unsaturated thin coarse-grained sediments overlying saturated fine-grained sediments. Water-table gradients of 5 percent are found in this area. Where saturated, however, the coarse-grained stratified drift may be a valuable aquifer for a domestic or small community water supply.

Upper Cohas Brook Aquifer

The upper Cohas Brook aquifer is in the northeastern corner of Londonderry (pl. 4, fig. 7). The aquifer is composed of coarse-grained ice-contact sand and gravel overlying glaciolacustrine fine sands. Transmissivities exceeded 2,500 ft^2/d in parts of the aquifer. Contamination from buried drums containing volatile organic carbons (VOC's) may limit the use of this aquifer as a potable water supply. This site is on the National Priority List of hazardous-waste sites (U.S. Environmental Protection Agency, 1986) and the contamination problem is discussed in a report by Stekl and Flanagan (1992).

Peters Brook Aquifer

The Peters Brook aquifer is approximately 2 mi south of Hooksett Village in Hooksett along the Merrimack River valley (pl. 4, fig. 7). Meltwater, flowing south along what is now the Peters Brook drainage, deposited deltaic sediments that built across glacial Lake Merrimack and eventually closed off the valley. This delta became a stratified-drift dam that created glacial Lake Hooksett (Koteff and others, 1984). The water surface of glacial Lake Hooksett was approximately 15 ft higher than the water surface of glacial Lake Merrimack to the south. Deltas that formed later, to the north of this delta, were graded to the level of glacial Lake Hooksett (Koteff and others, 1984). Test borings HTW-18, HTW-19, and HTW-20 drilled in the delta show that coarse-grained sand and gravel overlies fine sand and silt; however, most of the coarse-grained deposits have been excavated. The borings also show that the bedrock surface is relatively flat but deepens sharply near the present day Merrimack River (fig. 9). For example, test borings at wells HTW-18 and HTW-19 (pl. 4) indicate that bedrock is present 50 ft below the surface. Test boring HTA-3 (500 ft to the west) indicates bedrock is present 114 ft below the surface. Additional data pertaining to the surficial geology and hydrogeology can be found in a report by BCI Geonetics and Caswell (1980).

The saturated thickness of the Peters Brook aquifer increases from less than 20 ft in the east to greater than 100 ft near the Merrimack River. The large saturated thickness near the river does not contribute significantly to the transmissivity of the aquifer because most of the thickness consists of fine-grained lacustrine sediments. The zones of highest transmissivity (up to 4,000 ft^3/s) are where the saturated ice-contact, deltaic sand and gravel is thickest. This area is immediately east and west of N.H. Route 3 near well HTW-268 (pl. 4). The average saturated thickness of coarse-grained sand in this area ranged from 20 to 40 ft.

The Central Hooksett Water Precinct currently (1992) withdraws ground water from wells HTW-1 and HTW-2 on the eastern side of N.H. Route 3 and from well HTW-268 on the western side of N.H. Route 3. Wells HTW-2 and HTW-268 are close to Peters Brook, and pumping may induce flow from the brook into the aquifer. Base-flow measurements of Peters Brook in 1988 and 1989 indicate that ground-water discharge was about 0.8 ft^3/s or 550,000 gal/d (appendix D, site 3). Water withdrawn from well HTW-1 comes primarily from storage, because no surface-water sources are nearby. The average daily withdrawal from the three wells in 1989 was approximately 389,900 gal/d. The combined average daily withdrawal from the three wells was approximately 75,900 gal/d in January, February, and March 1990 and from HTW-1 was 18,200 gal/d in July 1990 (New Hampshire Water Management Bureau, written commun., 1990).

Brickyard Brook and Pinnacle Pond Aquifer

The Brickyard Brook and Pinnacle Pond aquifer is just west and south of Hooksett Village on the west side of the Merrimack River. A segmented esker, in place before the deposition of glacial-lake sediment, is traceable south of Pinnacle Pond and then again south and west of interchange 11 on Interstate 93 (pl. 4). The Brickyard Brook part of the aquifer was probably found in the waters of glacial Lake Hooksett, which was dammed by the Peters Brook delta. The elevation of the top of the Brickyard Brook delta is approximately 300 ft above sea level. Test borings at wells HTW-15 and HTW-17 (pl. 4) indicate coarse-grained ice-contact sand and gravel buried beneath the deltaic sands and associated lacustrine sediments (appendix B). Medium sand was found at about 186 ft above sea level in samples from well HTW-15, and coarse to very coarse sand was found at an altitude of 130 to 180 ft in well HTW-17 (appendix

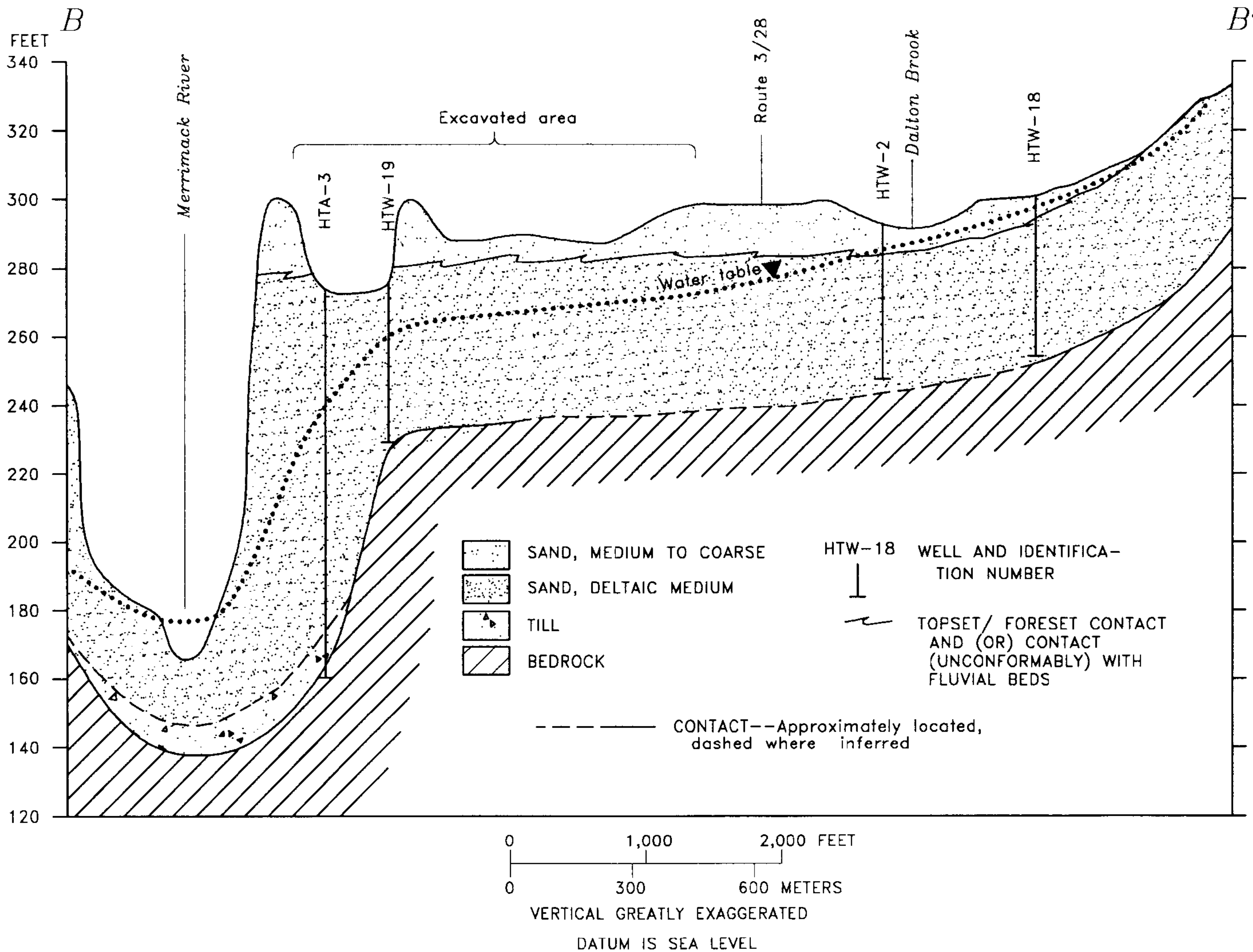


Figure 9.--Geologic section through the Peters Brook aquifer.

B). These data indicate the presence of a buried esker that may be hydraulically connected to the Merrimack River. The saturated thickness of the aquifer exceeds 62 ft at well HTW-15 and is 47 ft at well HTW-17. Further exploration is needed to determine if this deposit is a viable aquifer for water supply.

The Pinnacle Pond part of the aquifer (0.4 mi to the north) may also contain coarse-grained esker deposits; it is currently (1992) used by the Hooksett Village Water Precinct for water supply. In 1991, approximately 143,460 gal/d was withdrawn from well HTW-265 at the northern end of the pond. Approximately 95,438 gal/d was withdrawn from well HTW-10 at the southern end of the pond in 1990 (New Hampshire Water Management Bureau, written commun., 1991).

South Bow Aquifer

The South Bow aquifer is located along the northern boundary of the study area, which is adjacent to the Merrimack River in Bow and in part of Hooksett. Well BUW-8 (pl. 4) penetrated 50 ft of lacustrine very fine to medium sand overlying 18 ft of coarse-grained ice-contact sand and gravel (appendix B). The log of well BUW-9 (1,000 ft to the east) also showed that 20 ft of medium to very coarse ice-contact sand underlies lacustrine fine sand, which is underlain by 46 ft of very fine to medium sand. Aquifer transmissivity in this area ranges from 2,000 to 4,000 ft^2/d. These deposits are associated with a large delta that formed at the edge of glacial Lake Hooksett. Most of this delta deposit is outside of the study area.

Estimation of Yield for the Goffstown Aquifer

A two-dimensional ground-water-flow model was used to evaluate estimates of aquifer yield and to delineate drawdown due to ground-water withdrawal from a stratified-drift aquifer in Goffstown, N.H., west of the town center (pl. 3). The hydraulic characteristics of this aquifer are described in the section entitled "Description of Selected Aquifers," and a geologic section of the aquifer is shown in figure 8. The model is a numerical representation of the ground-water-flow system defined by a system of equations governing ground-water flow. The computer program used was developed by McDonald and Harbaugh (1988).

Conceptual and Numerical Model

In order to represent the ground-water-flow system in numerical form, one must describe the system by a conceptual model. Conceptualization of the system accounts for the processes involved in, and factors that influence, ground-water flow such as recharge, horizontal hydraulic conductivity of the aquifer, and stream-aquifer interaction. A conceptual model of the ground-water-flow system is shown in figure 10.

The saturated part of the stratified-drift aquifer consists of unconsolidated sand and gravel bounded by till and bedrock on the sides and bottom and by the water table on the top. Recharge to the stratified-drift aquifer is by infiltration of precipitation and lateral ground-water inflow from the upland till and (or) bedrock areas. The aquifer may also receive recharge from infiltration of stream water to the aquifer depending on the relative altitude of the water table and the stream surface.

The variability of aquifer properties and the difficulty of accurately measuring them results in a simplified representation of the stratified-drift aquifer. Several simplifying assumptions used to construct the model of the Goffstown aquifer are as follows:

1. **Two-dimensional horizontal flow is adequate to represent the flow system.** In the real system, ground-water flow is horizontal and vertical, but predominantly horizontal. Flow in the aquifer is always in response to a hydraulic gradient--a difference in hydraulic head divided by the distance between the heads. Vertical hydraulic gradients are generally downward in areas of ground-water recharge and upward in areas of ground-water discharge, such as to rivers. Strong vertical gradients also are present near discharging wells. The magnitude of vertical-flow gradients near a pumped well diminishes rapidly with distance from the well. The margin of error associated with simulating ground-water heads by considering only two-dimensional horizontal flow in the aquifer is small except in the area near the pumped wells.

2. **Ground water is withdrawn from wells that are fully penetrating and 100 percent efficient.** Wells used for supply are generally not fully penetrating but are commonly screened in the bottom 25 percent of the

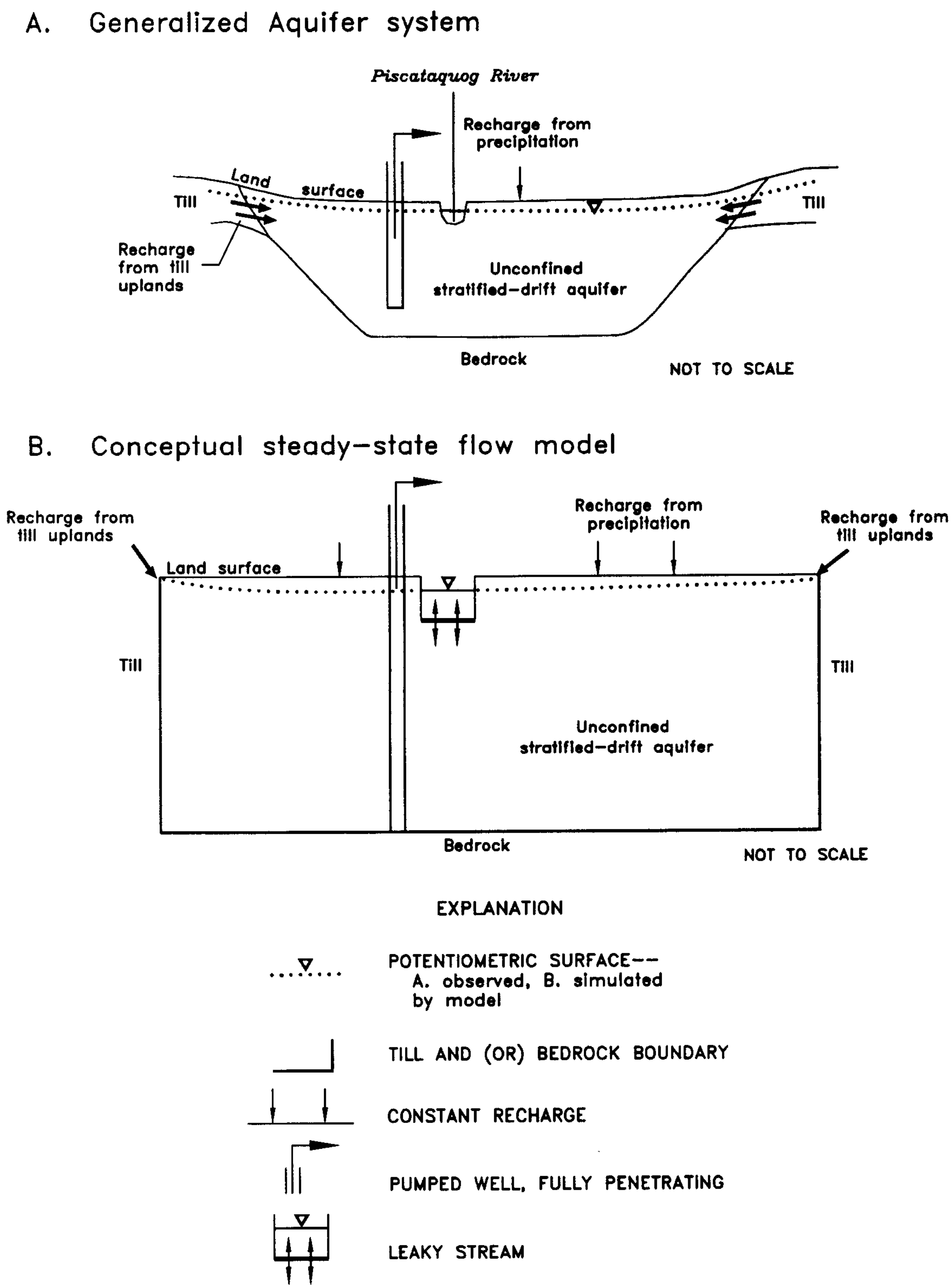


Figure 10.--Generalized aquifer system and conceptual model of steady-state ground-water flow in the Goffstown aquifer, Goffstown, New Hampshire.

aquifer. In addition, these wells are not 100 percent efficient. Increased drawdown in the well results from energy loss between the aquifer and the well, which is a function of well design and construction. The net effect of applying the simplifying assumptions related to fully penetrating wells to well efficiency is that slightly less drawdown is simulated than would occur in the natural system.

3. **There is no flow of ground water between till and (or) bedrock and stratified drift.** The model area is a valley-fill, stratified-drift aquifer that occupies a till-covered bedrock valley. In an aquifer where horizontal and vertical gradients are found in stratified drift and the till and (or) bedrock below, ground-water may flow between the aquifer and the surrounding geologic units. Measurements of aquifer discharge reflect the amount of water assumed to be recharging the aquifer from natural recharge and do not indicate that any additional water is flowing into or out of the aquifer from the bedrock. Because of this observation and the lack of vertical-gradient data, it was assumed that ground water did not flow between the stratified-drift and till and (or) bedrock boundary. Lateral inflow from till uplands adjacent to the edge of the stratified drift was simulated and applied to the edge cells of the model, as discussed in the section "Selection of Input Parameters."

4. **Finite-difference approximation of the nonlinear, partial differential equations governing two-dimensional flow result in reasonable values of head at any given site within the aquifer.** Flow in the conceptual model is described by differential equations that are solved numerically by use of a finite-difference approximation. The aquifer is discretized in space or divided into discrete blocks (cells) and hydraulic properties are assumed to be constant within each cell. For unconfined systems, the linear equations are not strictly applicable because changes in the potentiometric surface affect the transmissivity, and changes in the transmissivity with time result in nonlinear aquifer response. Because the changes in transmissivity are small throughout most of the aquifer, inaccuracies that result from this approximation are minimal. Exact solutions to the linear equations are impossible; therefore, the solutions are determined by solving the series of linear equations, through the process of iteration, until the greatest change in the solution (greatest change in the head) is less than some stated limit. A limit of 0.01 ft was used to end the iteration process.

   The analysis was performed in steady-state which was sufficient for the purposes of this yield estimate. Transient analyses were beyond the scope of this model and might be appropriate for management purposes.

Model grid

The model grid for the Goffstown aquifer is composed of 45 rows and 75 columns. This grid is superimposed on the long axis of the Piscataquog River valley. A variable-size grid was used in this model whereby the cell dimensions ranged from 200 ft by 200 ft (over most of the model) to 50 ft by 50 ft in areas where wells were to be simulated because of the high density of the data in these areas.

Boundary conditions

The Goffstown aquifer is bounded by till-covered bedrock valley walls on the north and south sides and by till and (or) constrictions in the valley where the Piscataquog River and South Branch Piscataquog River flow into the aquifer system on the western boundary at Goffstown center (pl. 3, fig. 7). In the numerical model, these physical features that limit the aquifer are represented by constant-flux boundaries and account for the small amounts of inflow into the stratified-drift aquifer from adjacent till-covered uplands.

Rivers and streams within the model are simulated as head-dependent flux boundaries. A head-dependent flux boundary specifies the amount of water allowed to move from the river into the aquifer or from the aquifer into the river as dependent on the head in the river and the head in the aquifer for any given model cell (Franke and others, 1987). Water can flow either into the stream from the aquifer or from the stream into the aquifer depending on the hydraulic gradient between the stream and aquifer. Stream widths, lengths, and stages were obtained from 1:24,000-scale topographic maps and from field measurements. Streambed thickness was estimated to average 2 ft based on measurements made near the well field and streambed hydraulic conductivity was estimated to be 3 ft^3/s by grain-size analysis of sediment cores collected at selected locations along the stream. The water table is simulated as a free-surface boundary and is free to move up and down in response to changes in head at any given cell (Franke and others, 1987).

Selection of input parameters

Input parameters consist of (1) position of the potentiometric surface, (2) gains in streamflow, (3) recharge to the aquifer, (4) hydraulic conductivity of the aquifer, and (5) aquifer saturated thickness. Input parameters for the numerical model were based on field measurements made between March 1988 and October 1989.

Annual water-table fluctuations were generally small (1-2 ft) near the center of the aquifer where the water-table gradient was low (0.2 percent) and somewhat greater (3-4 ft) toward the edge of the aquifer where the water-table gradient was steeper (2.0 percent). Water-table altitudes measured during September 1989 define the potentiometric surface used as initial heads in the model (table 3).

Streamflow measurements were made concurrently with water-table measurements to quantify stream-aquifer interaction. Flows were measured in all streams as they entered and left the model area and at points in between. Gains in streamflow over the 1.2-mi length of the model area ranged from 2.19 to 24.9 ft^3/s. No streamflow losses were measured for any of the reaches, although streamflow losses are likely in some reaches.

Recharge to the aquifer was based on measured stream gains during a period of low flow on September 6, 1989. These flows represented 93-percent flow duration on the basis of long-term records (1940-78) from a USGS streamflow-gaging station (01091000) on the South Branch Piscataquog River. (Use of low-flow measurements is a means of conservatively estimating recharge to the aquifer and is not representative of long-term average conditions.) The gain of 2.19 ft^3/s over the 1.2-mi^2 model area is equal to a recharge rate of 19.8 in/yr. This gain is approximately equal to the $\pm$5-percent error associated with the measurements used to compute the value of 2.19 ft^3/s. The recharge rate of 19.8 in/yr is approximately one-half of the average-annual precipitation recorded in this area (National Oceanic and Atmospheric Administration, 1987).

Lateral recharge is ground-water recharge from upland till and (or) bedrock that does not discharge to a stream before it flows into the aquifer. This ground water effectively recharges the edges of the modeled aquifer. Lateral inflow to the aquifer at the stratified-drift and till and (or) bedrock boundary was estimated by measuring ground-water discharge to several small tributaries that drain till-covered bedrock uplands. The flow values were divided by the drainage areas to estimate the ground-water discharge per square mile of upland drainage basin. Ground-water discharge ranged from 0.065 to 0.27 $(ft^3/s)/mi^2$. The average discharge of 0.09 $(ft^3/s)/mi^2$ was used in the model. This is relatively consistent with the average discharge of 0.205 $(ft^3/s)/mi^2$ used by Harte and Mack (1992).

The estimated horizontal hydraulic conductivity of the aquifer ranged from 3 to 150 ft/d. The lowest hydraulic conductivity is near the edges of the aquifer at the stratified drift and till and (or) bedrock contact. The highest hydraulic conductivity, at the center of the stratified-drift aquifer, reflects the presence of coarse-grained kame and delta deposits. These hydraulic conductivities were largely estimated from relations between hydraulic conductivity and grain size distribution of aquifer sediments that are described in the section of the report on "Transmissivity and Hydraulic Conductivity." Hydraulic conductivity in the coarse-grained deposits was estimated from results of a 3-day aquifer test done by the USGS at a well (GNW-1, pl. 3) owned by the Goffstown Water Precinct.

Saturated thickness was determined from test drilling and from extensive seismic-refraction profiling over most of the model area (pl. 3). The saturated thickness averages 30-40 ft and exceeds 60 ft locally.

Calibration of steady-state model

Model calibration is the process of adjusting input parameters until model-computed heads closely agree with observed heads (water levels). In the model, wells used to simulate actual observation wells represented the average head for the entire cell. Water levels observed in 10 wells during late August and early September 1989 were used as the reference heads in calibrating the model. Streamflow into and out of the aquifer was measured simultaneously during a period of 93-percent flow duration.

Aquifer horizontal hydraulic conductivities, streambed conductances, and river stage were varied in model zones, based on reasonable ranges of uncertainty, around values of aquifer characteristics observed in the field. The process was continued until the absolute difference between observed head and computed head at each of the 10 observation wells was less than 3 ft. The absolute differences between observed and computed heads in the calibrated model ranged from 0.11 to 2.95 ft, and the average absolute difference was 0.37 ft (table 3). The recharge to the model area from precipitation and from lateral seepage from till uplands was not varied during calibration because recharge values reflected streamflow measurements made during periods of base flow. Ground-water discharge from the aquifer to the stream was equal to the amount of recharge applied to the model and was set at 2.19 ft^3/s, the discharge measured during base flow.

The steady-state water-table configuration computed by the model is shown in figure 11. This head distribution was adopted as the starting-head array in the model simulations used to estimate the yield of the Goffstown aquifer. The steady-state water budget computed by the model is shown in table 4.

Sensitivity analysis of nonstressed steady-state model

Sensitivity analysis shows the effect of variations in parameters on model results. The analysis indicates which parameters have the most effect on ground-water-flow simulations and where future data-collection efforts can be concentrated.

Principal input parameters (recharge, streambed conductance, and horizontal hydraulic conductivity) were increased and decreased independently throughout the model to observe the effect on computed water levels. Input parameters were varied over a reasonable range of values that reflects the uncertainty of correctly estimating them. Observed heads minus the computed heads were analyzed statistically, and the results are shown in figure 12 as a series of boxplots.

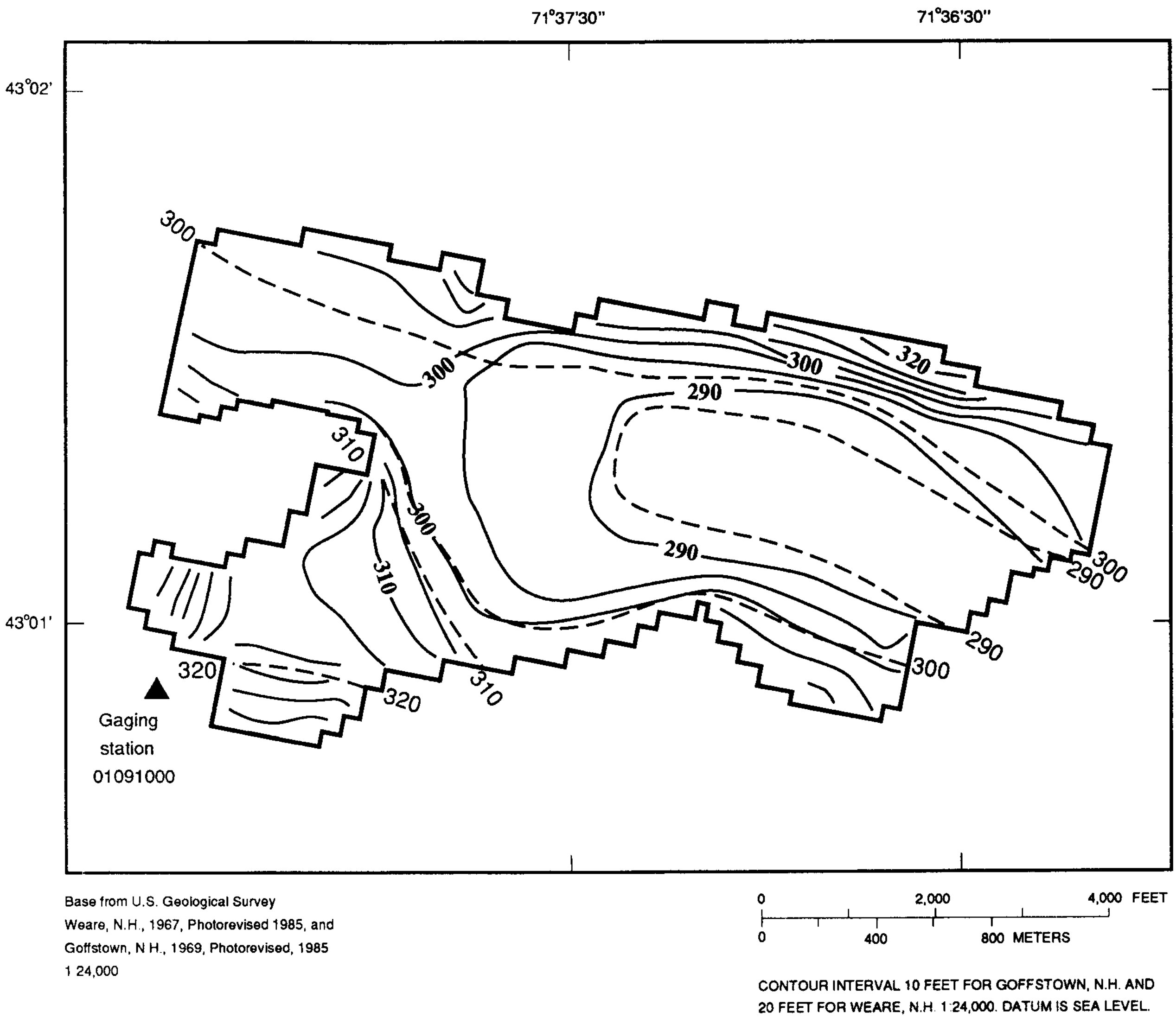
The boxplots show the interquartile range (IQR), which is the range of the central 50 percent of data as well as the position of extreme values. A comparison of boxplot 1 (calibrated heads) with boxplot

Table 3.--*Differences between observed and computed heads in the calibrated steady-state model for the Goffstown aquifer*

[ft, feet]

Well location in model row, column	Local well number (plate 3)[1]	Observed water level (ft)	Computed water level (ft)	Observed minus computed (ft)
04,07	GNW-17	299.80	302.75	-2.95
14,43	GNW-1	285.50	286.20	-.70
15,47	GNW-7	285.60	286.05	-.45
16,45	GNW-4	285.60	286.01	-.41
18,41	GNW-9	285.50	285.91	-.41
20,35	GNW-8	285.80	285.69	.11
20,45	GNW-2	285.40	285.73	-.33
26,20	GNW-14	295.00	292.25	2.75
27,52	GNW-15	286.00	286.11	-.11
31,31	GNW-16	289.10	290.25	-1.15

[1] Wells are shown on plate 3.



EXPLANATION

MODEL BOUNDARY

—— 290 —— SIMULATED POTENTIOMETRIC SURFACE CONTOUR--Shows simulated altitude of water table under steady-state conditions. Contour interval 5 feet.

290 – – – – 290 INTERPRETED POTENTIOMETRIC CONTOUR--Contour interval 10 feet. National Geodetic Vertical Datum 1929.

Figure 11.--Water-table configuration for the simulation of the Goffstown aquifer.

Table 4.--*Model-calculated steady-state water budget for the Goffstown aquifer*

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

Sources of water	Inflow (ft^3/s)	Outflow (ft^3/s)
Natural recharge;		
Precipitation	2.2	0
Lateral recharge from till and (or) uplands	.2	0
River leakage	1.8	4.2
Total	4.2	4.2

5 (hydraulic conductivity x 0.6) indicates that reasonable decreases in the hydraulic conductivity tend to elevate the overall potentiometric surface and produce more variation in the water levels (fig. 12). The greatest variations are on the edges of the model, where the water-table gradient is the steepest. In addition, a comparison of boxplot 1 with boxplot 3 (recharge times 1.5) indicates that heads increase proportionally and vary over a wider range by increasing the amount of recharge to the model. Nonstressed model results are shown to be less sensitive to changes in streambed conductance (boxplots 4 and 5) than to changes in other parameters.

Estimate of aquifer yield

Aquifer yield was estimated by simulation of the total ground-water withdrawal that could be simultaneously pumped from a series of wells distributed throughout the aquifer. The amount of water discharged from a given well was limited by hydrologic constraints discussed in the following paragraph. The total aquifer yield was the sum of the withdrawals from the individual wells. Hypothetical withdrawal wells, located on town-owned and undeveloped and (or) agricultural land, were restricted to zones of high transmissivity.

Natural recharge and induced infiltration from streams are the two main sources of water to the hypothetical wells. The total water available from natural recharge was 2.4 ft^3/s (table 4). Water available to the wells from the stream was limited by a conservative withdrawal scheme to maintain a minimum streamflow. The minimum-streamflow scheme is only one of many possible withdrawal schemes and was used only as an example. This scheme allows the streamflow that is equaled or exceeded 99 percent of the time (99-percent-flow duration) to flow in the stream and uses the streamflow that is equalled or exceeded 95 percent of the time (95-percent-flow duration) minus the flow that is equalled or exceeded 99 percent of the time. For the Goffstown aquifer, the 95-percent-flow duration minus the 99-percent-duration flow is equal to 6.4 ft^3/s for all streams flowing into the aquifer.

The total water potentially available to wells is equal to the total natural recharge (2.4 ft^3/s) and the available streamflow (6.4 ft^3/s) or 8.8 ft^3/s, which seems high for the approximately 1.2-mi^2 aquifer. The available streamflow is probably high because the aquifer is at the confluence of two major river drainages entering the area from the west and north.

For each hypothetical well in the aquifer, simulated drawdown produced by pumping was limited to 50 percent of the saturated thickness at the well. For a given cell containing a well, the final computed head is the average for the entire cell; the head in the well is less than this average value. The actual drawdown in the cell was calculated by a method described by Trescott and others (1976).

Ground-water-withdrawal wells were simulated at four locations that met the previously discussed well-location criteria. Steady-state simulations were run to determine the rate that water could be pumped simultaneously from all four wells without reducing the saturated thickness by more than 50 percent at any well and without inducing infiltration of more than 6.4 ft^3/s of streamflow. The four wells yielded 0.8 ft^3/s (0.52 Mgal/d), 1.1 ft^3/s (0.71 Mgal/d), 0.9 ft^3/s

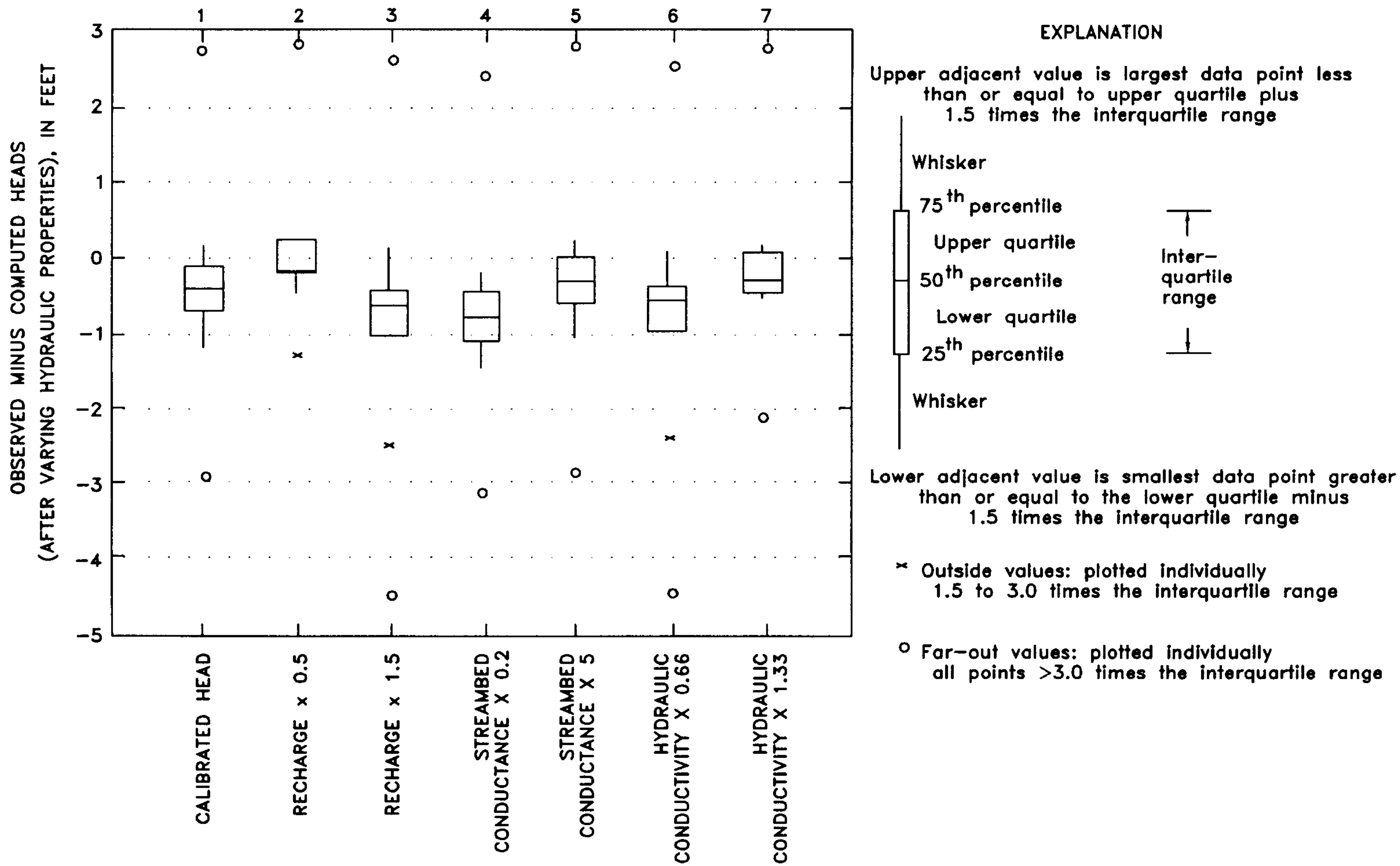


Figure 12.--The statistical distribution of the difference between observed and computed heads for sensitivity tests of the nonstressed steady-state flow model.

(0.58 Mgal/d), and 1.1 ft^3/s (0.71 Mgal/d) for a total of 3.9 ft^3/s (2.5 Mgal/d). Lines of equal drawdown for this simulation are shown in figure 13.

The estimated yield of the Goffstown aquifer was not limited by the water available for induced infiltration but by the limit for drawdown of 50 percent of the saturated thickness and by the number of hypothetical wells in the model. Additional wells would likely increase the yield. By adding more wells in different locations or in different configurations, additional streamflow could possibly be induced to infiltrate the aquifer. The simulation results in this report are considered to be conservative and only one of several options that can be used for estimating aquifer yield. The final steady-state potentiometric surface under the yield-estimate simulation during pumping is shown in figure 14. The final water budget for the yield estimate is shown in table 5.

The estimated aquifer yield is 3.9 ft^3/s; 1.5 ft^3/s is from induced recharge from the river and 2.4 ft^3/s is from natural recharge (table 5). Water captured by the wells (aquifer yield) is equal to the decrease in ground-water discharge plus the increase in recharge (Lohman and others, 1972, p. 3).

Sensitivity analysis of stressed steady-state model

A second sensitivity analysis was done to the Goffstown model to test the effects of the yield estimate to changes in input parameters. The ranges in parameter values tested were similar to those used in the steady-state calibrated nonstressed model. The results of the analysis of each change in input value are shown in figure 15.

Heads at the 10 observation wells (table 3) were compared for each simulation in the steady-state stressed sensitivity analysis. The results are summarized in figure 15. The data from which the boxplots were constructed represent the difference between the heads computed for the final yield-estimate simulation and the heads computed for each change in the input-parameter values. Three parameters--recharge, streambed conductance, and horizontal hydraulic conductivity--were varied over the same range as in the sensitivity analysis of the steady-state nonstressed model. Large reductions in streambed conductance appear to have a small but noticeable effect on the magnitude of the head differences (boxplot 3, fig. 15) and reduced recharge produces lower heads than were expected. Variations of horizontal hydraulic conductivity cause noticeable changes in the magnitude of the head differences; for example, when horizontal hydraulic conductivity is decreased, the magnitude of the heads increases.

The changes in input-parameter values also affected the estimated aquifer yields. The parameter changes and resulting changes in estimated yield are shown in table 6. The changes in recharge and streambed-conductance values have little effect on the total yield of the aquifer compared to changes in the horizontal hydraulic conductivity (table 6).

Changes in the horizontal hydraulic conductivity have the largest effect on estimated yield. For example, a 33-percent decrease in horizontal hydraulic conductivity results in a 36-percent decrease in the yield. Similarly, a 33-percent increase in horizontal hydraulic conductivity causes a 22-percent increase in yield.

Varying the input parameters also changes the percentage of water coming from each source. As recharge is decreased, more water must come from the stream to produce the same yield. As the streambed conductance is decreased, flows between the stream and aquifer decrease, whereas the total estimated yield drops only to 3.7 ft^3/s.

This sensitivity analysis indicates that, within the selected ranges of parameter values, uncertainty in the values of horizontal hydraulic conductivity assigned in the model causes the greatest range of estimated aquifer yields. Future data-collection to refine the distribution of horizontal hydraulic conductivity could substantially improve the model-derived estimates of aquifer yield. The effect on aquifer yield of streambed conductance, a parameter for which there is little field data, is highly variable throughout the model and not well defined. Additional streambed-conductance data, in combination with base-flow measurements of streams, would increase confidence in the model-derived water budget and the sources of water to pumped wells.

Appraisal of Yield Estimate

The estimate of aquifer yield and its sensitivity to changes in the values of selected input parameters for one particular withdrawal scheme is summarized in tables 5 and 6. Several other alternative withdrawal plans were not considered. For example, the estimated yield of 3.9 ft^3/s (2.5 Mgal/d) was shown to be constrained by limiting drawdown to 50 percent of the saturated thickness. The number and location of wells simulated in the model may also affect yield.

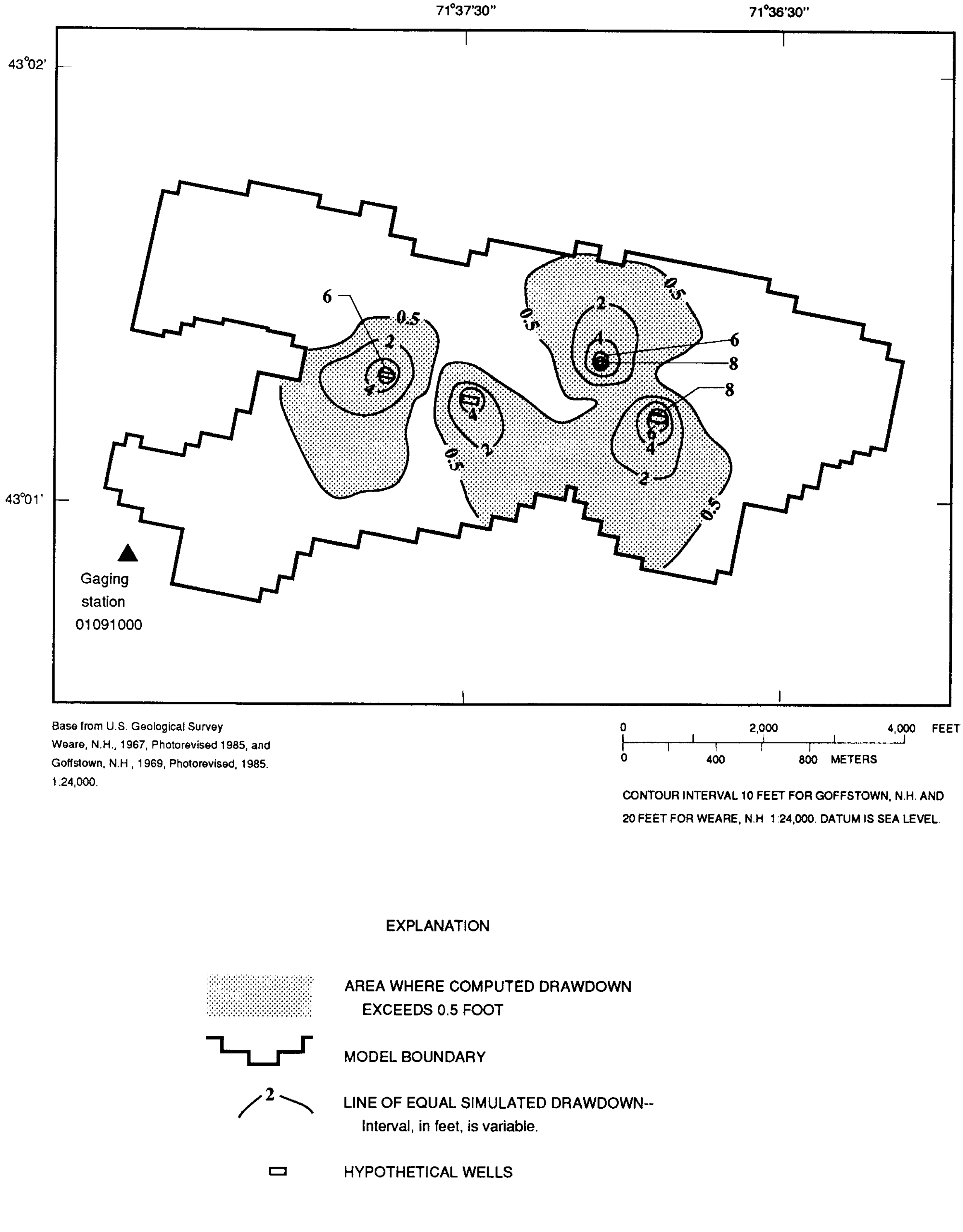


Figure 13.--Simulated drawdown near four hypothetical wells in the Goffstown aquifer.

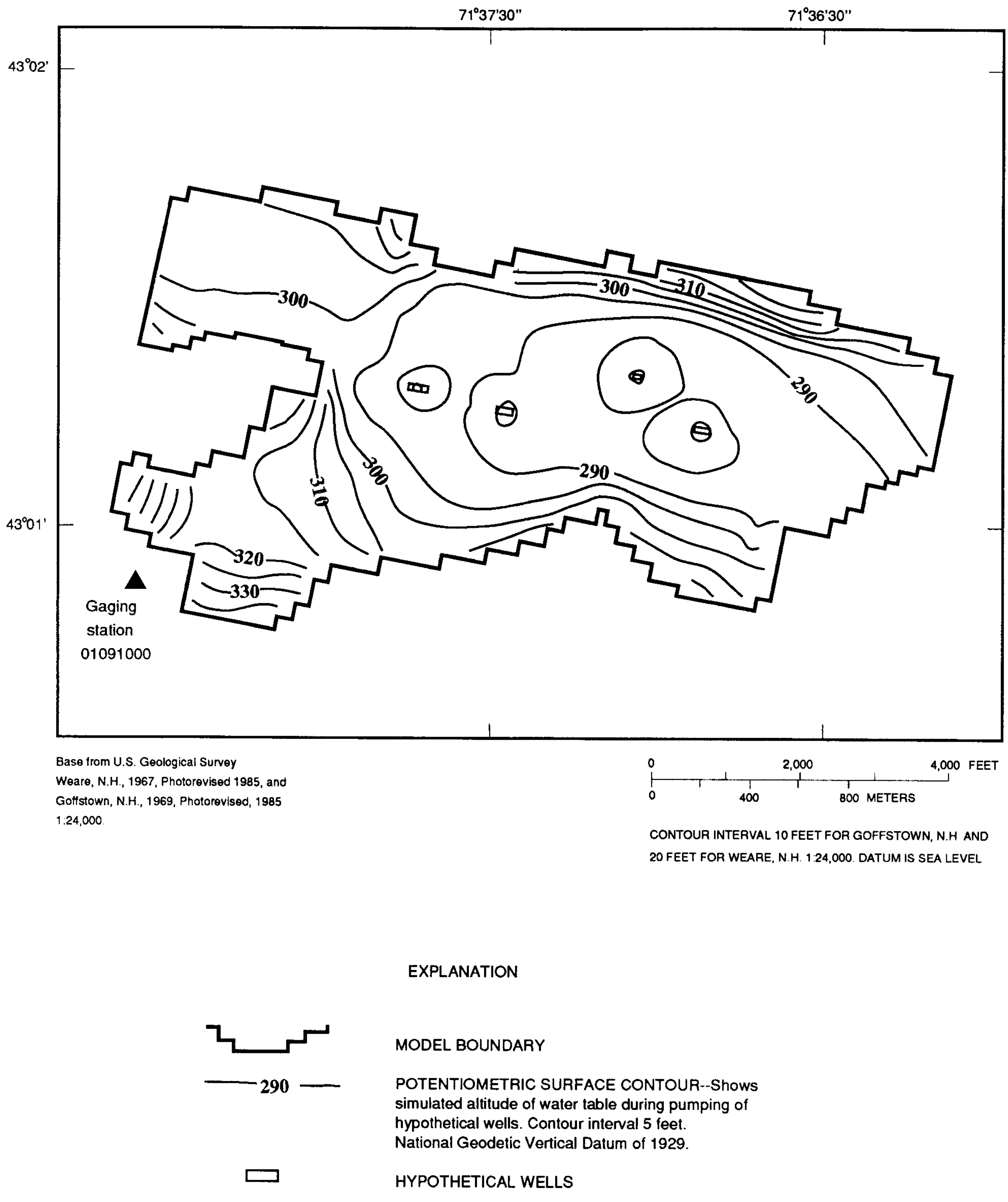


Figure 14.--Steady-state water-table configuration, during pumping, for the Goffstown aquifer.

Table 5.--*Model-calculated steady-state water budget for the yield estimate with simulated withdrawal of 3.9 cubic feet per second*

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

Source of water	Inflow (ft^3/s)	Outflow (ft^3/s)
Natural recharge:	2.2	0
Lateral recharge from till and (or) bedrock uplands	.2	0
River leakage		
captured by wells	1.5	0
not captured by wells	2.5	2.5
Wells	0	3.9
Total	6.4	6.4

Sensitivity analysis of the model indicates that the estimates of yield can range from 2.5 to 5.0 ft^3/s (1.6 to 3.2 Mgal/d) primarily because of uncertainty in the horizontal hydraulic conductivities used in the model. Whereas changes in the input data for recharge and streambed conductance do not significantly affect the yield, changes in these parameters could affect not only the percentage of water that comes from each source but also the total water budget. If recharge from precipitation is decreased, additional water to sustain withdrawals could come from decreases in ground-water discharge to streams and from induced infiltration from streams. As streambed conductance was decreased, the total flow between the stream and aquifer decreased, but enough water was induced into the aquifer to sustain the estimated yield.

Only 3.9 ft^3/s of the 8.8 ft^3/s can be withdrawn under the modeled conditions. Other withdrawal schemes may allow for more or less water to be withdrawn. This model considered only two sources of water available to wells: (1) areal and lateral recharge and (2) intercepted and induced ground-water discharge (streamflow). Additional yield from aquifer storage may be available to wells over short periods of time (high-demand times) with the assumption that the water will be replaced during times of low demand and (or) times of greater than average recharge to the aquifer (such as periods of high rainfall).

WATER QUALITY

Water samples from 10 wells were collected, from June to November 1988, and analyzed for inorganic and organic compounds. The results were used to evaluate the background water quality of the stratified-drift aquifers in the middle Merrimack River basin area. During the sampling phase of this study, areas where ground water was known to be contaminated (CERCLA sites) were avoided.

The choice of sampling procedure depended on the source of the water sampled. All the sampled wells were developed either with compressed air or with a centrifugal pump to remove water introduced during drilling, foreign material, and sediment and to improve the hydraulic connection with the aquifer. Wells were allowed to stabilize for at least 1 month before sampling. Just before sampling, the wells were pumped until temperature and specific conductance stabilized and at least three times the volume of water in the well was evacuated. This procedure helped ensure that the sampled water represented water from within the aquifer.

Results of the chemical analyses are presented and compared with the USEPA (1992) primary and secondary drinking-water regulations and the New Hampshire Department of Environmental Services, Water Supply Engineering Bureau drinking-water recommendations (New Hampshire Department of Environmental Services, Water Supply Engineering

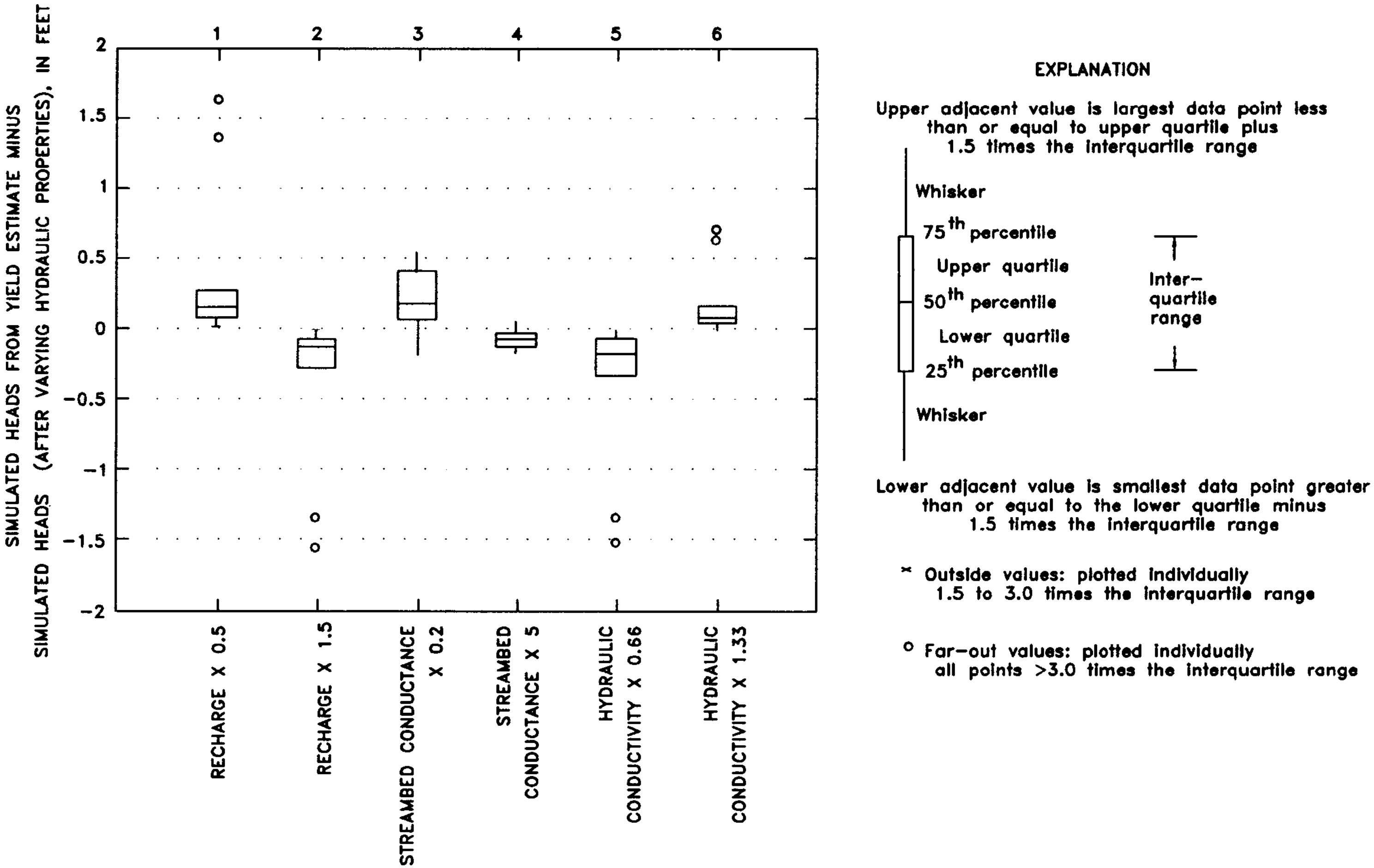


Figure 15.--The statistical distribution of differences between simulated heads during pumping and simulated heads during pumping after varying hydraulic properties in the model.

Table 6.--*Changes in the model-calculated yield estimate resulting from variation of hydraulic properties*

[Individual and total well yields in cubic feet per second]

Model run	Yield of hypothetical well				
	1	2	3	4	Total
Calibrated model	0.8	1.1	0.9	1.1	3.9
Recharge × 0.5	.8	1.0	.9	1.1	3.8
Recharge × 1.5	.8	1.1	1.0	1.1	4.0
Streambed conductance × 0.2	.8	1.0	.9	1.0	3.7
Streambed conductance × 5.0	.8	1.1	1.0	1.1	4.0
Hydraulic conductivity × 0.66	.5	.7	.6	.7	2.5
Hydraulic conductivity × 1.33	1.0	1.4	1.2	1.4	5.0

Bureau, written commun., 1990) in table 7. Naturally occurring constituents that have no recommended limits, but whose concentrations are generally less than a few micrograms per liter, also are included in table 7. Many of the constituents listed in table 7 were not detectable in water samples from the stratified-drift aquifers in the study area.

Results of the sample analyses indicate that water from the stratified-drift aquifers is generally suitable for drinking and other domestic uses. Water from two wells (GSW-101 and WGW-19) had sodium concentrations that exceeded 20 mg/L (milligrams per liter), water from two wells (FCW-3 and WGW-19) had concentrations of dissolved iron that exceeded 300 μg/L (micrograms per liter), and water from eight wells (FCW-3, GNW-14, GSW-101, NCW-8, NJW-1, NJW-4, NJW-5, WGW-19) had concentrations of manganese greater than 50 μg/L. Individual constituents and properties are discussed in the following paragraphs.

Specific conductance--a measure of the ability of water to conduct electrical current--ranged from 53 μS/cm (microsiemens per centimeter at 25 degrees Celsius) in water from well NJW-1 to 410 μS/cm in water from well WGW-19. The median (84 μS/cm) for all water samples was less than the median (132 μS/cm) of the entire State for public supply wells completed in stratified-drift aquifers (Morrissey and Regan, 1987).

Total dissolved-solids (solids residue, table 7) concentrations in water include all ionized and un-ionized dissolved solids in solution. The total dissolved-solids concentrations of all water samples from stratified-drift aquifers ranged from 39 (well NJW-1) to 216 mg/L (well WGW-19) and were less than the maximum recommended limit for drinking water of 500 mg/L established by the New Hampshire Water Supply Engineering Bureau (1990). The low concentration of dissolved solids in these stratified-drift aquifers can be attributed to the low solubility of the aquifer matrix and the relatively short time that the water is in contact with the aquifer (Morrissey and Regan, 1987).

Sodium (Na) and chloride (Cl) can be introduced into ground water from nonindigenous sources (wet or dry deposition such as sea salt and aerosols) and anthropogenic sources. The major anthropogenic source of both sodium and chloride is NaCl used in road salting. On the basis of limited data, it is estimated that New Hampshire towns and cities used about 33,000 tons per year of NaCl for deicing roads, (Hall, 1975). The highest concentra-

Table 7.-- *Chemical analyses of ground-water samples*

[ft, feet; μS/cm, microsiemens per centimeter at 25 degrees Celsius; °C, degrees Celsius; mg/L, milligrams per liter; μg/L, micrograms per liter; <, less than; --, no data]

Local well number	Date of sample location	Water level in depth below land surface (ft)	Depth of well, total (ft)	Depth to top of sample interval (ft)	Specific conductance, field (μS/cm)	pH, field (standard units)	Temperature of water (°C)	Oxygen, dissolved (mg/L as O_2)	Hardness (mg/L as $CaCO_3$)	Calcium, dissolved (mg/L as Ca)	Magnesium, dissolved (mg/L as Mg)	Sodium, dissolved (mg/L as Na)	Potassium, dissolved (mg/L as K)	Alkalinity, lab (mg/L $CaCO_3$)
FCW-3	11-08-88	11.9	20	17.5	65	6.8	10.5	--	15	4.1	1.2	5.0	1.8	22
GNW-1	06-10-88	--	40	30.0	106	5.9	8.5	4.4	18	5.5	.91	13	1.4	10
GNW-14	11-08-88	.61	55	50.0	92	6.0	9.0	4.9	26	8.6	1.2	6.8	.90	14
GSW-101	11-08-88	6.13	35	32.5	280	5.6	9.0	1.6	15	4.8	.62	32	1.3	9.0
MGW-1	11-07-88	5.91	32	30.0	71	6.2	9.5	8.9	14	4.5	.74	7.4	2.2	13
NCW-8	11-08-88	22.93	50	48.0	75	6.3	8.7	1.4	25	7.7	1.5	4.6	2.0	25
NJW-1	11-07-88	7.01	55	52.5	53	6.7	9.1	2.2	13	4.0	.65	4.1	1.7	15
NJW-4	11-07-88	7.45	40	37.5	110	5.8	9.6	3.6	14	4.4	.76	16	1.4	11
NJW-5	11-07-88	14.26	35	32.5	69	6.5	9.1	--	20	4.2	2.4	4.8	2.0	16
WGW-19	11-08-88	4.46	30	27.5	410	5.4	9.8	.7	16	5.1	.87	75	1.6	10
U.S. Environmental Protection Agency drinking-water regulations for listed property or chemical constituent														
SMCL[1]	--	--	--	--	--	--	--	--	--	--	--	[4]20 - 250	--	--
MCL[2]	--	--	--	--	--	--	--	--	--	--	--	--	--	--

Table 7.-- *Chemical analyses of ground-water samples--Continued*

Local well number	Chloride, dissolved (mg/L as Cl)	Fluoride, dissolved (mg/L as F)	Silica, dissolved (mg/L as SiO_2)	Solids, residue at 180 °C, dissolved (mg/L)	Sulfate, dissolved (mg/L as SO_4)	Nitrogen, nitrate, dissolved (mg/L as N)	Nitrogen, $NO_2 + NO_3$, dissolved (mg/L as N)	Nitrogen, ammonia, dissolved (mg/L as N)	Phosphorus, dissolved (mg/L as P)	Aluminum, dissolved (µg/L as Al)	Arsenic, dissolved (µg/L as As)	Barium, dissolved (µg/L as Ba)	Beryllium, dissolved (µg/L as Be)	Boron, dissolved (µg/L as B)
FCW-3	3.4	0.1	13	44	12	[3]<0.09	<0.10	0.02	0.01	<10	3	5	<0.5	<10
GNW-1	20	.3	10	63	7.0	[3]<.50	.51	.01	<.01	20	<1	10	<.5	<10
GNW-14	12	.1	14	59	5.1	1.09	1.10	<.01	.01	10	<1	8	1	<10
GSW-101	46	<.1	8.3	101	9.4	.67	.69	<.01	.01	40	<1	29	1	<10
MGW-1	8.8	<.1	11	47	4.6	.94	.96	<.01	.01	<10	<1	3	2	<10
NCW-8	6.1	.1	21	58	5.4	[3]<.09	<.10	.02	<.01	<10	3	5	<.5	<10
NJW-1	1.5	.1	15	39	8.8	[3]<.13	.14	.02	.03	<10	6	<2	1	<10
NJW-4	20	<.1	12	63	8.1	[3]<.22	.23	.01	<.01	<10	<1	8	<.5	<10
NJW-5	2.1	<.1	19	43	14	.15	.16	.03	<.01	<10	1	8	<.5	<10
WGW-19	120	.1	9.4	216	8.0	[3]<.09	<.10	.04	<.01	90	1	74	<.5	<10
U.S. Environmental Protection Agency drinking-water regulations for listed property or chemical constituent														
SMCL[1]	[4]250	[4]2.0	--	[4]500	[4]250	--	--	--	--	50-200	--	--	--	--
MCL[2]	250	4.0	--	500	--	10	10	--	--	--	50	2,000	4	--

Table 7.-- *Chemical analyses of ground-water samples*--Continued

Local well number	Cadmium, dissolved (μg/L as Cd)	Cobalt, dissolved (μg/L as Co)	Copper, dissolved (μg/L as Cu)	Iron, dissolved (μg/L as Fe)	Lead, dissolved (μg/L as Pb)	Lithium, dissolved (μg/L as Li)	Manganese, dissolved (μg/L as Mn)	Mercury, dissolved (μg/L as Hg)	Molybdenum, dissolved (μg/L as Mo)	Nickel, dissolved (μg/L as Ni)	Strontium, dissolved (μg/L as Sr)	Vanadium, dissolved (μg/L as V)	Zinc, dissolved (μg/L as Zn)	Carbon, organic, dissolved (mg/L as C)	Dichlorobromomethane, total (μg/L)
FCW-3	<1	<3	<10	1,800	<10	<4	930	<0.1	<10	2	23	<6	<3	1.6	<0.2
GNW-1	<1	<3	<10	19	<10	<4	16	<.1	<10	3	45	<6	21	.8	<.2
GNW-14	<1	<3	<10	12	<10	<4	62	<.1	<10	5	58	<6	5	.3	<.2
GSW-101	<1	<3	<10	16	<10	<4	61	<.1	<10	3	42	<6	4	.7	<.2
MGW-1	<1	<3	<10	9	<10	<4	6	<.1	<10	3	67	<6	6	.3	<.2
NCW-8	<1	<3	<10	20	<10	<4	400	<.1	<10	7	59	<6	<3	.8	<.2
NJW-1	<1	<3	<10	160	<10	<4	630	<.1	<10	3	34	<6	4	.5	<.2
NJW-4	1	<3	<10	7	<10	<4	50	<.1	<10	3	70	<6	<3	.4	<.2
NJW-5	1	3	<10	17	<10	<4	790	<.1	<10	5	28	<6	<3	.6	<.2
WGW-19	2	4	<10	350	<10	<4	310	<.1	<10	3	67	<6	6	1.1	<.2
U.S. Environmental Protection Agency drinking-water regulations for listed property or chemical constituent															
SMCL[1]	--	--	1,000	300	--	--	50	--	--	--	--	--	5,000	--	--
MCL[2]	5	--	--	--	50	--	--	2.0	--	100	--	--	--	--	--

Table 7.-- *Chemical analyses of ground-water samples--Continued*

Local well number	Carbon tetra-chlo-ride, total (µg/L)	1,2-Di-chloro-ethane, total (µg/L)	Bromo-form, total (µg/L)	Chloro-di-bromo-methane, total (µg/L)	Chloro-form, total (µg/L)	Toluene, total (µg/L)	Benzene, total (µg/L)	Chloro-benzene, total (µg/L)	Chloro-ethane, total (µg/L)	Ethyl-benzene, total (µg/L)	Methyl-bromide, total (µg/L)	Methyl-chlo-ride, total (µg/L)	Methyl-ene chlo-ride, total (µg/L)	Tetra-chloro-ethyl-ene, total (µg/L)
FCW-3	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2
GNW-1	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
GNW-14	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
GSW-101	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
MGW-1	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NCW-8	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NJW-1	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NJW-4	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NJW-5	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
WGW-19	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
U.S. Environmental Protection Agency drinking-water regulations for listed property or chemical constituent														
SMCL[1]	--	--	--	--	--	--	--	--	--	--	--	--	--	--
MCL[2]	5.0	5.0	--	--	--	--	5.0	--	--	--	--	--	--	--

Table 7.-- *Chemical analyses of ground-water samples*--Continued

Local well number	Tri-chloro-fluoro-methane, total (μg/L)	1,1-Di-chloro-ethane, total (μg/L)	1,1-Di-chloro-ethyl-ene, total (μg/L)	1,1,1-Tri-chloro-ethane, total (μg/L)	1,1,2-Tri-chloro-ethane, total (μg/L)	1,1,2,2-Tetra-chloro-ethane, total (μg/L)	1,2-Di-chloro-benzene, total (μg/L)	1,2-Di-chloro-propane, total (μg/L)	1,2-Transdi-chloro-ethene, total (μg/L)	1,3-Di-chloro-propene, total (μg/L)	1,3-Di-chloro-benzene, total (μg/L)	1,4-Di-chloro-benzene, total (μg/L)	2-Chloro-ethyl-vinyl-ether, total (μg/L)
FCW-3	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2
GNW-1	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
GNW-14	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
GSW-101	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
MGW-1	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NCW-8	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NJW-1	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NJW-4	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
NJW-5	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
WGW-19	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2	<.2
U.S. Environmental Protection Agency drinking-water regulations for listed property or chemical constituent													
SMCL[1]	--	--	--	--	--	--	--	--	--	--	--	--	--
MCL[2]	--	--	7.0	200	--	--	--	--	--	--	--	--	--

Table 7.-- *Chemical analyses of ground-water samples*--Continued

Local well number	Di-chloro-di-fluoro-methane, total (µg/L)	Trans-1,3-di-chloro-propene, total (µg/L)	Cis-1,3-di-chloro-propene, total (µg/L)	1,2-Dibromo-ethyl-ene, total (µg/L)	Vinyl-chlo-ride, total (µg/L)	Tri-chloro-ethyl-ene, total (µg/L)	Styrene, total (µg/L)	1,2-Dibromo-ethane total (µg/L)	Xylene, total recoverable (µg/L)
FCW-3	<0.2	<0.2	<0.2	--	<0.2	<0.2	<0.2	<0.2	<0.2
GNW-1	<.2	<.2	<.2	<0.2	<.2	<.2	<.2	--	<.2
GNW-14	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
GSW-101	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
MGW-1	.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
NCW-8	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
NJW-1	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
NJW-4	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
NJW-5	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
WGW-19	<.2	<.2	<.2	--	<.2	<.2	<.2	<.2	<.2
	U.S. Environmental Protection Agency drinking-water regulations for listed property or chemical constituent								
SMCL[1]	--	--	--	--	--	--	--	--	--
MCL[2]	--	--	--	--	2.0	5.0	100	--	10,000

[1] SMCL--Secondary Maximum Contaminant Level: Contaminants that affect the aesthetic quality of drinking water. At high concentrations or values, health implications as well as aesthetic degradation may also exist. SMCL's are not Federally enforceable but are intended as guidelines for the States (U.S. Environmental Protection Agency, 1992).

[2] MCL--Maximum Contaminant Level: Enforceable, health-based regulation that is to be set as close as is feasible to the level at which no known or anticipated adverse effects on the health of a person occur. The definition of feasible means the use of the best technology, treatment techniques, and other means that the Administrator of the U.S. Environmental Protection Agency finds, after examination for efficacy under field conditions and not solely under laboratory conditions, are generally available (taking cost into consideration) (U.S. Environmental Protection Agency, 1992).

[3] A "less than" value in this column indicates that a value in either the nitrite or the nitrite plus nitrate analysis was below detection (nitrate was determined by subtracting the value for nitrite from the value for nitrite plus nitrate).

[4] Secondary level set by the New Hampshire Department of Environmental Services, Water Supply Bureau (New Hampshire Department of Environmental Services, Water Supply Bureau, written commun., 1987).

tion of chloride was 120 mg/L from well WGW-19; less than one-half of the USEPA (1992) secondary maximum contaminant level (SMCL[1]) for chloride (250 mg/L,) established as a taste threshold. The water samples from two wells had sodium concentrations (32 mg/L at well GSW-101 and 75 mg/L at well WGW-19) that exceeded the Health Advisory Level for sodium (20 mg/L) established by the USEPA (1992) as a recommended limit for people with heart, hypertension, or kidney problems. The ratio of Na to Cl in water from well WGW-19 was 1 to 1 meq/L (milliequivalents per liter), indicating that NaCl (probably from road salt) is the source of both constituents.

The pH of water is a measure of the hydrogen-ion activity. Water having a pH of 7.0 is neutral, less than 7.0 is acidic, and greater than 7.0 is alkaline. The pH of most ground water in the United States ranges from about 6.0 to 8.5 (Hem, 1985, p. 63-64). The pH of water sampled in the field ranged from 5.4 to 6.8; the median was 6.1. The range of pH in stratified-drift aquifers sampled for previous studies (Moore, 1990; Flanagan and Stekl, 1990; Mack and Lawlor, 1992; Moore and others, in press) in this series (1984-89) ranged from 5.3 to 8.5, and the median was 6.1. The most basic or alkaline ground-water samples came from well FCW-3 (6.8). In this study, the most acidic water was from wells GNW-1 (5.9), GNW-14 (6.0), GSW-101 (5.6), MGW-1 (6.2), NCW-8 (6.3), NJW-4 (5.8), and WGW-19 (5.4). All these samples had pH values that were less than the SMCL of 6.5 established by the USEPA (1992).

The alkalinity of a solution is defined as the capacity for solutes in water to react with and neutralize acid (Hem, 1985, p. 106). It is commonly thought of as an indicator of buffering capacity--the water's ability to resist changes in pH upon addition of an acid. Almost all of the alkalinity in most natural water can be attributed to carbonate and bicarbonate ions. Because stratified-drift aquifers in New Hampshire consist of sediment derived from bedrock with a low carbonate mineral content, alkalinity in New Hampshire ground water is generally low. Alkalinity in samples from this study was determined by incremental titration of unfiltered samples with aliquots of 0.01639N sulfuric acid to an end point of pH 4.5. For all the water samples, alkalinity ranged from 9.0 mg/L as $CaCO_3$ (at well GSW-101) to 25 mg/L as $CaCO_3$ (at well NCW-8). The median alkalinity, 14.5 mg/L as $CaCO_3$, indicates that water from this area has low alkalinity and, therefore, has low buffering capacity.

The predominant form of inorganic nitrogen in natural water is nitrate, an oxidized, highly soluble compound. Excess nitrate in ground water can originate from fertilizer applications, leachate from sewage systems, or agricultural wastes. Nitrate (NO_3 as N) in ground water has been linked to methemoglobinemia, or blue-baby syndrome (Lukens, 1987). For all the samples, the concentration of NO_3 as N was the highest in the water from well GNW-14 (1.09 mg/L). This concentration is less than the maximum contaminant level (MCL[2]) for NO_3 as N (10 mg/L) established by the USEPA (1992). Inorganic nitrogen also can be present as nitrite or ammonium. In all the water samples, nitrogen concentrations as ammonium ranged from less than 0.01 to 0.04 mg/L.

The sulfate (SO_4^{-2}) ion is one of the major anions in natural water. Oxidation of sulfide ores, gypsum, and anhydrite and atmospheric deposition are sources of sulfate, but sulfate-producing minerals generally are not present in stratified-drift aquifers in New Hampshire. Sulfate is reduced by anaerobic bacteria to hydrogen sulfide gas (H_2S), which can be detected by smell at concentrations of only a few tenths of a milligram per liter. The sulfate concentration for all the ground-water samples ranged from 4.6 to 14.0 mg/L, and the median was 8.05 mg/L. The SMCL for sulfate (SO_4^{-2}) in drinking water is 250 mg/L.

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[1] SMCL, Secondary Maximum Contaminant Level: Contaminants that affect the aesthetic quality of drinking water. At high concentrations or values, health implications as well as aesthetic degradation may also exist. SMCL's are not Federally enforceable but are intended as guidelines for the States (U.S. Environmental Protection Agency, 1992).

[2] MCL, Maximum Contaminant Level: Enforceable, health-based regulation that is to be set as close as is feasible to the level at which no known or anticipated adverse effects on the health of a person occur. The term feasible means the use of the best technology, treatment techniques, and other means that the Administrator of the U.S. Environmental Protection Agency determines, after examination for efficacy under field conditions and not solely under laboratory conditions, are generally available (taking cost into consideration) (U.S. Environmental Protection Agency, 1992).

Manganese and iron are common elements in minerals in stratified-drift deposits within this study area. Elevated concentrations of manganese, often accompanied by elevated concentrations of iron, were the most common water-quality problem found during this investigation. Manganese, an abundant metallic element, is an undesirable impurity in water because of its tendency to deposit black oxide stains (Hem, 1985, p. 85). Water from seven wells had manganese concentrations that exceeded the SMCL of 50 μg/L (U.S. Environmental Protection Agency, 1992)--930 μg/L at FCW-3, 62 μg/L at GNW-14, 61 μg/L at GSW-101, 400 μg/L at NCW-8, 630 μg/L at NJW-1, 790 μg/L at NJW-5, and 310 μg/L at WGW-19. Iron, if present in excessive amounts in residential water supplies, forms red oxyhydroxide precipitates that can stain clothes and plumbing fixtures. Concentrations of iron in water from two of the sampled wells, 1,800 μg/L at well FCW-3 and 350 μg/L at well WGW-19, exceeded the SMCL of 300 μg/L (U.S. Environmental Protection Agency, 1992).

Aluminum (Al), the third most abundant element in the Earth's crust, rarely is present in water at concentrations greater than a few tenths or hundredths of a milligram per liter (Hem, 1985, p. 73). Exceptions can be found in highly acidic waters where the Al^{+3} ion is dissolved. Water from well WGW-19 had the highest aluminum concentration, 90 μg/L (0.09 mg/L), and the lowest pH value, 5.4.

Most trace metals are present in the soil as cations that are strongly adsorbed by oxides and hydroxides (particularly aluminum, iron, and manganese) and complexed by organic ligands at near-neutral values of pH (Drever, 1982); the dissolved concentrations are, therefore, usually low. All of the ground-water samples analyzed had trace-metal concentrations that are either below or more than two times the detection limit for the following metals: boron, cadmium, cobalt, copper, lead, lithium, molybdenum, mercury, and vanadium. In addition, the concentrations of the following metals were within the range of values commonly found in natural water (Hem, 1985): dissolved barium, beryllium, nickel, strontium, and zinc.

Detectable concentrations of arsenic were found in the water from three wells; 3 μg/L in water from well FCW-3, and NCW-8, and 6 μg/L in water from well NJW-1. These values were less than the MCL of 50 μg/L.

Water from wells sampled in this study was tested for 37 VOC's. All of the samples tested had concentrations of VOC's that were less than the detection level of 0.2 μg/L.

SUMMARY AND CONCLUSIONS

The middle Merrimack River basin in south-central New Hampshire encompasses an area of 469 mi^2, which is underlain by approximately 98 mi^2 of stratified drift. A 22-percent increase in population from 1980 to 1989 has caused an increased demand on the water resources of this area. At present (1992), ground-water withdrawals from stratified drift for public supply within the basin do not exceed 0.4 Mgal/d. The towns of Goffstown and Hooksett are the primary users of this water. Many of the shallow stratified-drift aquifers within the study area could be valuable sources of domestic and municipal water supplies, but they are not developed to their fullest potential.

Stratified-drift deposits in the basin largely reflect local and regional glacial-lake environments that existed at the time of deposition. Many are deltas deposited into glacial lakes or locally ponded meltwater.

Stratified-drift aquifers in the southwestern part of the study area are generally thin, and much of the stratified drift consists of fine-grained glaciolacustrine sediment. Transmissivities are generally less than 1,000 ft^2/d. Some of these deposits, however, are capable of supplying enough potable water for domestic or small community supply.

Stratified-drift aquifers in the western and central parts of the study area are composed of fine-grained lacustrine and coarse-grained ice-contact deposits. Saturated thicknesses of these stratified-drift deposits exceed 100 ft in places. A total of 14 stratified-drift aquifers have transmissivities greater than 1,000 ft^2/d. Transmissivity in the most productive aquifers exceeded 6,000 ft^2/d.

Stratified-drift aquifers in the eastern part of the study area were formed in regional glacial-lake environments. Glacial Lakes Merrimack and Hooksett had a profound effect on the deposition of stratified drift in the Merrimack River valley. This large river valley contains extensive eskers, kames, and deltas, as well as the fine-grained lacustrine deposits. The influence of these glacial lakes extends into the larger tributary valleys for several miles. Total saturated thicknesses of stratified-drift aquifers in this area are commonly greater than 20 ft and exceed 150 ft in some areas. Transmissivities are locally greater than 2,000 ft^2/d.

Of the potentially valuable aquifers in the middle Merrimack River basin, only the Goffstown aquifer, Peters Brook aquifer, and the Pinnicle Pond