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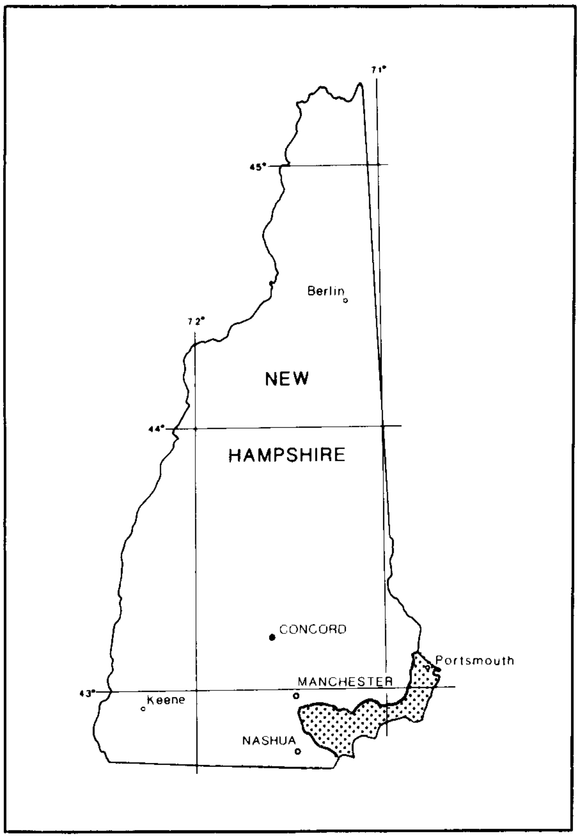
Geohydrology and Water Quality of Stratified-Drift Aquifers in the Lower Merrimack and Coastal River Basins, Southeastern New Hampshire

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 91-4025



Prepared in cooperation with the

NEW HAMPSHIRE DEPARTMENT OF ENVIRONMENTAL SERVICES
WATER RESOURCES DIVISION

GEOHYDROLOGY OF STRATIFIED-DRIFT AQUIFERS

Stratified drift is subdivided on plates 4-6 into four categories on the basis of grain size. These categories are (1) coarse-grained material (sand and gravel) with a median particle diameter predominantly larger than 0.0049 in.; (2) fine-grained material (very fine sand, silt, and clay) with a median particle diameter predominantly smaller than 0.0049 in.; (3) coarse-grained deposits that overlie fine-grained deposits; and (4) fine-grained deposits (including marine) that overlie coarse-grained deposits.

The coarse-grained deposits in the central Powwow River valley, Derry-Island Pond area, the Golden Brook valley, and the isolated kame plains and deltas in Portsmouth-Newington, North Hampton-Hampton, and Kensington (pls. 4-6) constitute some of the most important aquifers in the basin. Of these areas, the aquifers at Powwow River, Golden Brook valley, and the Derry-Island Pond area are the only ones not currently developed at or near their maximum potential yield to wells.

Coarse-grained deposits are part of a valley-train sequence in the Golden Brook valley of Windham. The stratigraphy of these proglacial deposits is characterized by coarse sand and gravel overlying very fine to medium sand (fig. 6). The maximum thickness of these deposits (60 ft) is determined from seismic-refraction surveys in the area. Present flow regulation at Golden Brook causes the stream to go dry during parts of the year, and, therefore, the brook cannot be considered a reliable source of water to recharge the deposits.

Geohydrologic section B-B′ (fig. 7), which extends northwest-southeast across the axis of the Powwow River, shows a striking contrast between the level outwash and the irregular bedrock surface. Seismic-refraction data indicate maximum depths to bedrock of 125 ft. The outwash-plain deposits are generally well-sorted and permeable; therefore, water is believed to flow easily between this aquifer and the Powwow River and nearby Great Pond. The aquifer and surface-water bodies are interdependent and together comprise a water system that is capable of a high yield to wells completed in the aquifer.

Geohydrologic section C-C′ (fig. 8) shows fine- to coarse-grained sand and gravel underlying a kame delta in North Hampton that is interbedded with discontinuous layers of marine sediments. The sand and gravel deposits have excellent water-bearing characteristics and, where saturated thicknesses are large, would yield large supplies of water to wells. Kame-delta deposits penetrated by well NSW-69 consist of exceptionally clean, well-washed coarse-grained sand and pebble gravel that has a 70-ft saturated thickness. Low rates of recharge to the aquifer are considered the primary limit on the productivity of this and many other coastal aquifers. This deposit, like similar kame plains and deltas in the eastern part of the study area, is small in areal extent; thus recharge from precipitation is limited.

Present at the aquifer margins and within the aquifer section are marine deposits of silt and clay that are relatively impermeable and do not yield usable quantities of water to wells. Where the marine sediments border the aquifer perimeter and extend the full saturated thickness, as in the northwestern end of the section, lateral recharge to the aquifer as ground-water discharge from neighboring uplands is restricted. The deposit is not bordered by land significantly higher in elevation and, therefore, recharge from upland runoff is negligible. In addition, the hydraulic connection between Cornelius Brook and the aquifer is poor because of the presence of marine impermeable sediments in the streambed. Aquifers hydraulically isolated from streams can serve as storage reservoirs that can be tapped during late summer periods of drought without causing a serious depletion of streamflow that normally would result if withdrawals are made from aquifers that border a river and are sustained by induced infiltration of river water.

In some localities, sand and gravel is buried beneath the marine sediment. Geohydrologic section D-D′ (fig. 9) shows that Seabrook municipal wells SGW-44 and SGW-65 are completed in coarse-grained sand and gravel, which underlies 20 to 40 ft of fine-grained sand and clay. The contact between the base of the marine deposits and the underlying sand and gravel is gradational at these wells. Contacts are typically gradational where clay overlies ice-marginal deposits, deposited as submarine fans and deltas (Thompson and others, in press) in a high-energy environment that prevailed when meltwater currents flowed from the nearby ice margin. To the northwest, the vertical and horizontal contact between the marine sediments and the sand and gravel is more abrupt and probably is erosional. A 20-ft-thick layer of ice-contact sand and gravel forms an aquifer confined above and below by sediments.

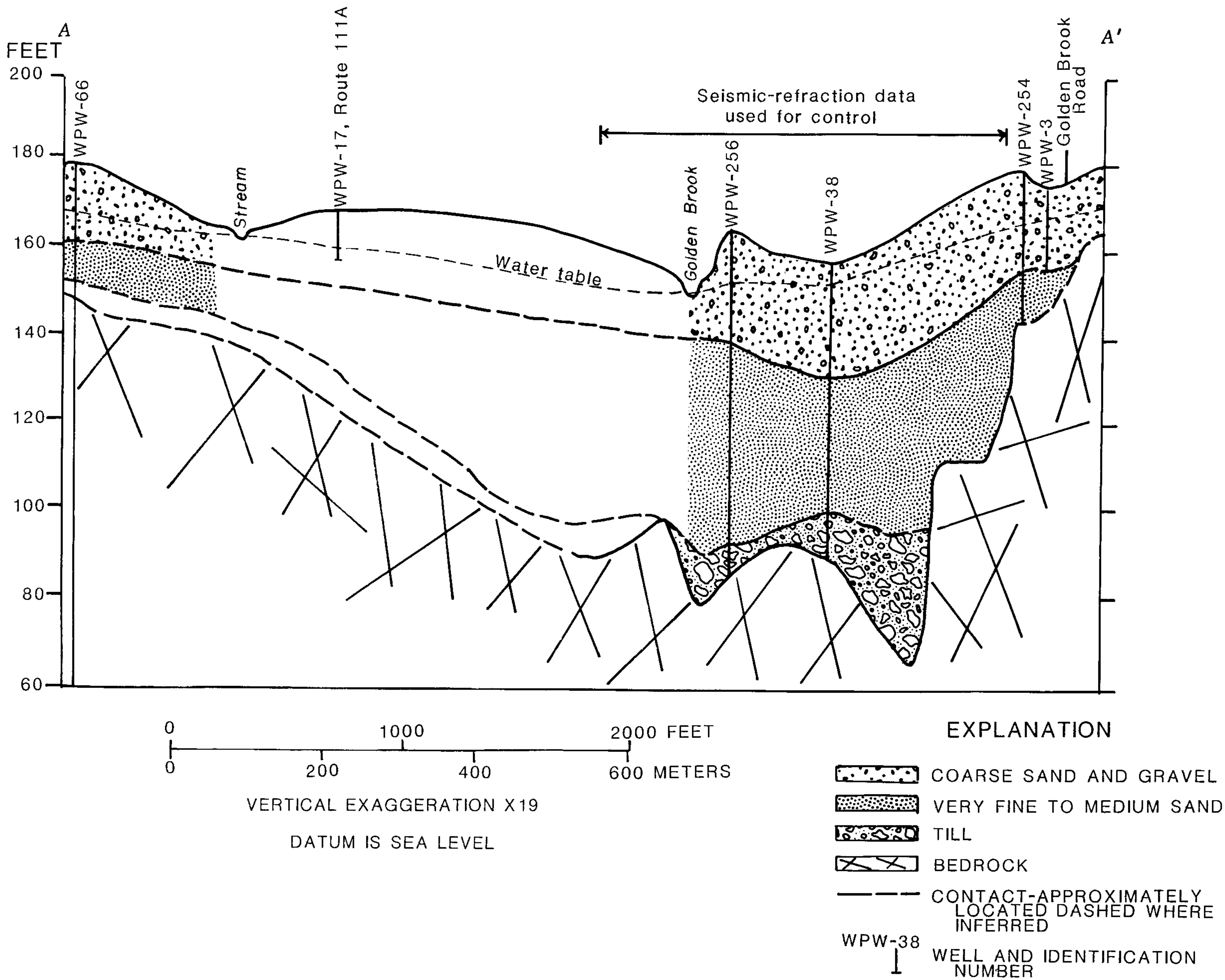


Figure 6.--Geohydrologic section A-A′ showing the valley train in southern Windham. Line of section is shown on plate 1.

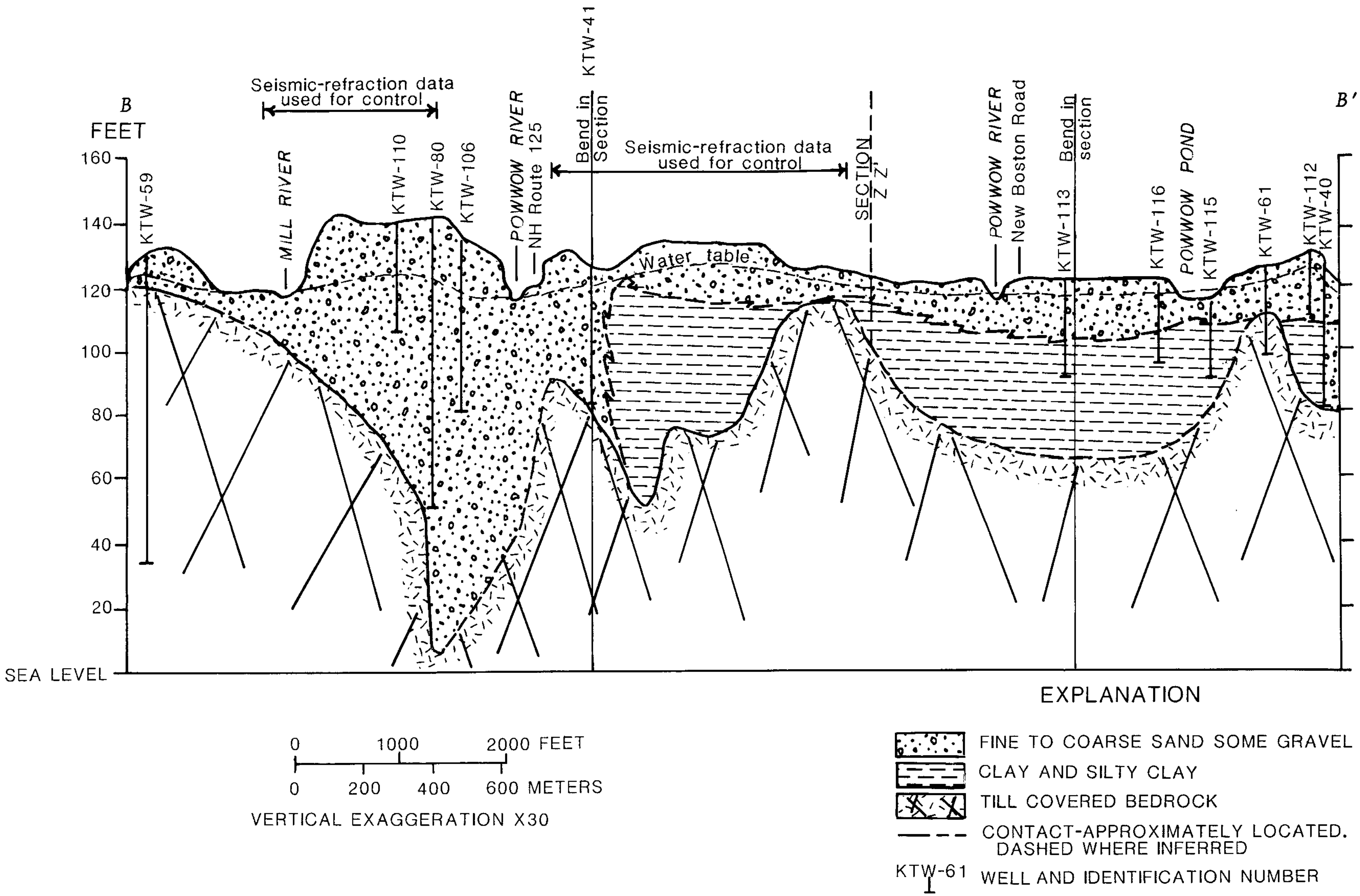


Figure 7.--Geohydrologic section B-B′ showing the outwash plain in Kingston. Line of section is shown on plate 2.

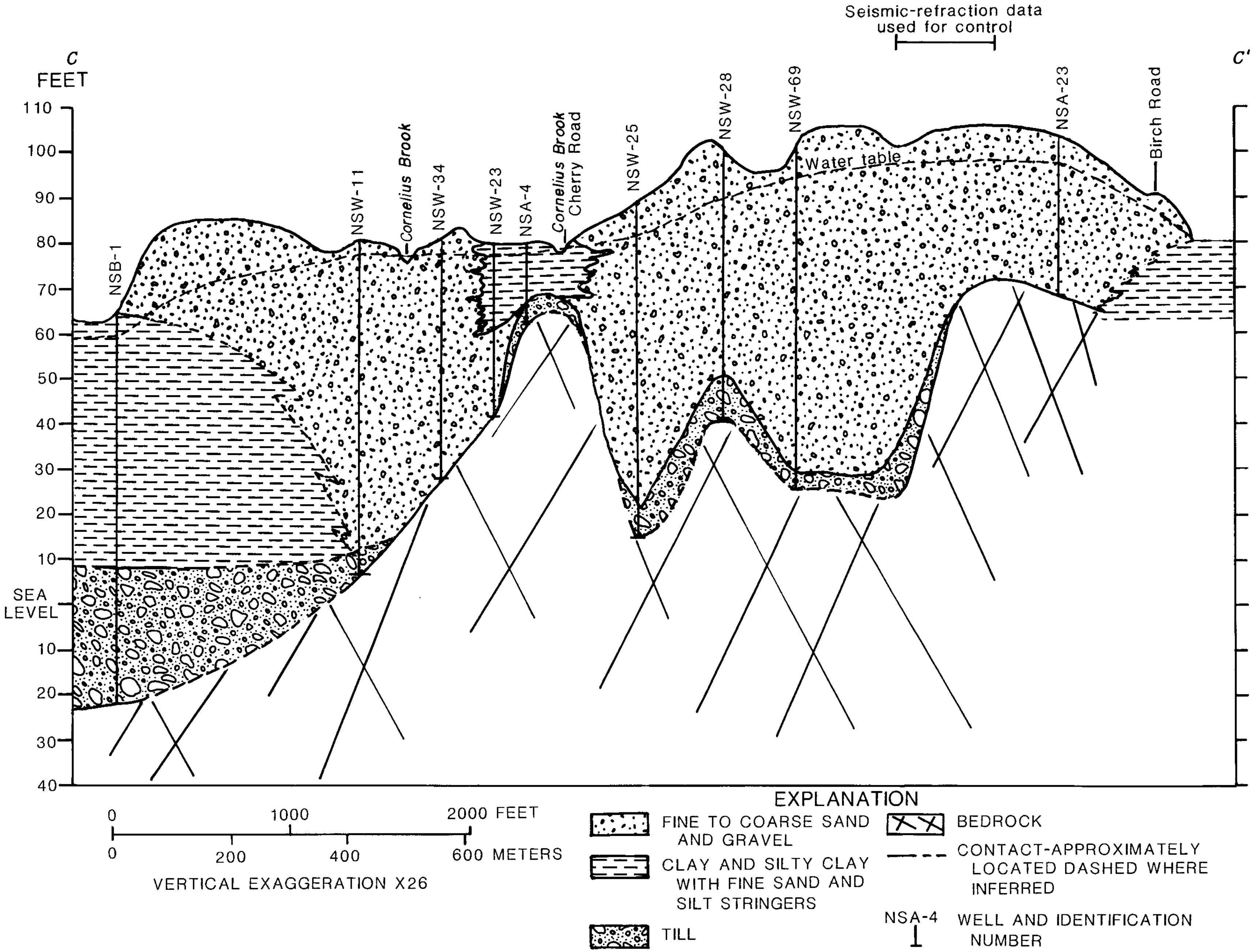


Figure 8.--Geohydrologic section C-C' showing the kame delta in North Hampton. Line of section is shown on plate 3.

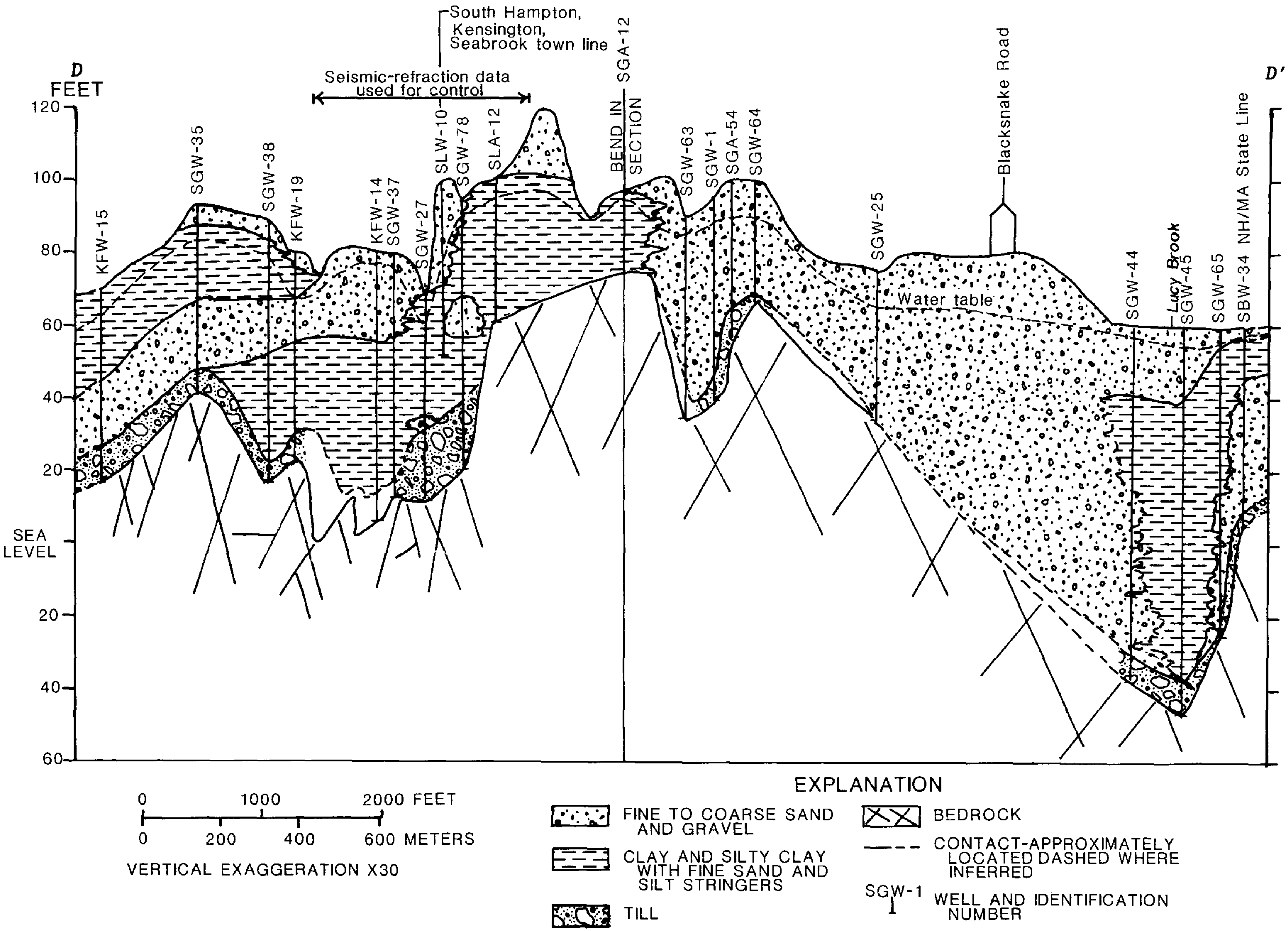


Figure 9.--Geohydrologic section D-D′ showing the kame delta in Kensington and Seabrook. Line of section is shown on plate 3.

Description of Aquifer Boundaries

Areal extents of stratified-drift aquifers are shown on plates 1-3. Also shown on the plates are the locations of marine silt and clay and till-mantled bedrock. The thickness of the solid-line contact on the 1:24,000-scale U.S. Geological Survey map implies ±80-foot horizontal accuracy. Aquifers located within moderate to steeply sloping uplands and valley walls in the western part of the study area are generally mapped to this level of accuracy. Delineation of the aquifer is difficult and contact locations are uncertain in the eastern part of the study area, where a complex geologic history of marine inundation, postglacial erosion, and uplift has produced low scarps and broad swampy lowlands with few ice-contact slopes or meltwater drainageways. As a result, several of the aquifer contacts in the coastal lowlands are inferred and appear as dashed lines, or the contacts are concealed by swamps and marshes and appear as dotted lines.

The boundaries of an aquifer are of particular importance in describing the response of the ground-water-flow system to withdrawal stress. The ground-water-flow system has two types of flow boundaries--impermeable (no flow) boundaries and recharge and discharge (flow) boundaries (Heath, 1983). Low scarps composed of marine sediments limit the extent of stratified-drift aquifers in the eastern part of the study area. Because these geologic units are virtually impermeable compared to stratified-drift aquifer material, flow across the interface is limited. Wells that tap stratified drift in the proximity of such boundaries yield less water than they would if the aquifer had infinite extent. As shown in figure 10, pumping near an impermeable boundary lowers the water level between the well and the boundary more than pumping does at the opposite side of the well where the aquifer is extensive. Drawdown, therefore, increases with decreasing distance to an impermeable (no flow) boundary at a given pumping rate.

Major perennial streams or large ponds hydraulically connected to stratified-drift aquifers are considered recharge boundaries because they can serve as a source of recharge to the aquifer. When a well near a surface-water body is pumped, the water is initially withdrawn from aquifer storage. As the cone of depression continues to spread, water from an increasingly large area flows toward the well. When the water table in the vicinity of the nearby stream or pond is sufficiently lowered, ground water that would have naturally discharged to the surface water is diverted and captured by the well. Eventually, ground-water levels can be lower than the surface of the surface-water body, at which time surface water can recharge the aquifer by induced infiltration (fig. 11).

An understanding of these basic principles of boundary hydraulics is important for determining the optimum sites for ground-water development. Recharge to pumped wells is greatest and the effects on water levels from pumping are the least when wells are parallel and adjacent to recharge boundaries and at maximum distance from impermeable boundaries.

Recharge, Discharge, and Direction of Ground-Water Flow

Recharge to the stratified-drift aquifers is by infiltration from precipitation, seepage losses from tributary streams, and lateral flow from adjacent till and bedrock. Discharge from the aquifers is by flow to the rivers and ponds in the basin, by evapotranspiration in areas where ground water is near the land surface, and by ground-water withdrawals. The water table marks the top of the saturated zone in the unconsolidated deposits and fluctuates continuously in response to changes in recharge and discharge.

The 10-year hydrograph for well LIW-1, which is completed in stratified drift in the town of Lee, shows the cyclic seasonal variation of water level that is common in stratified-drift aquifers (fig. 12). Ground-water recharge from precipitation exceeds evapotranspiration in late fall and early winter (November-December) and even more in the spring (March-June) when snow melt is an additional component of recharge. Lowest ground-water levels are in late summer or early fall. The depth to water table at this well (30 ft) is generally attained only in the elevated kame plains and deltas of the eastern study area; this depth contrasts with the shallow depths (a few feet) in sand and gravel deposits of lower altitude. The seasonal fluctuation in ground-water levels is typically 3 to 5 ft for wells completed in stratified drift.

Flow

Circulation of ground water is usually confined within topographic basins. The five subbasins that comprise distinct ground-water-flow systems of the report area are shown in figure 1, and include the

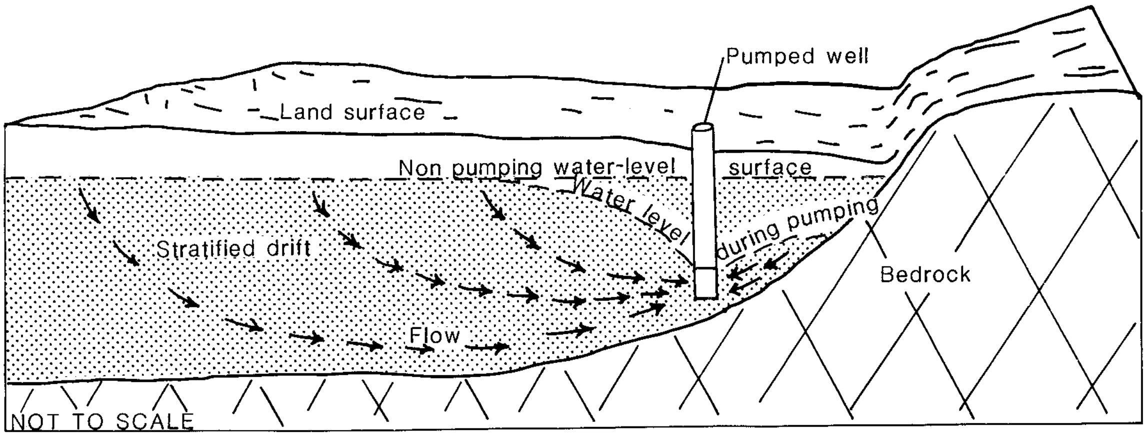


Figure 10.--Ground-water flow and water-level drawdowns at a pumped well near an impermeable boundary.

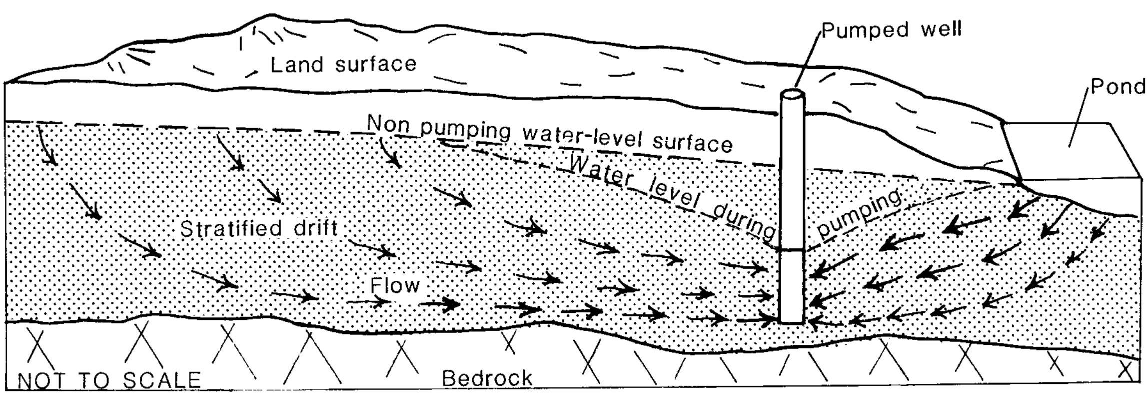


Figure 11.--Ground-water flow and water-level drawdowns at a pumped well near a recharge boundary.

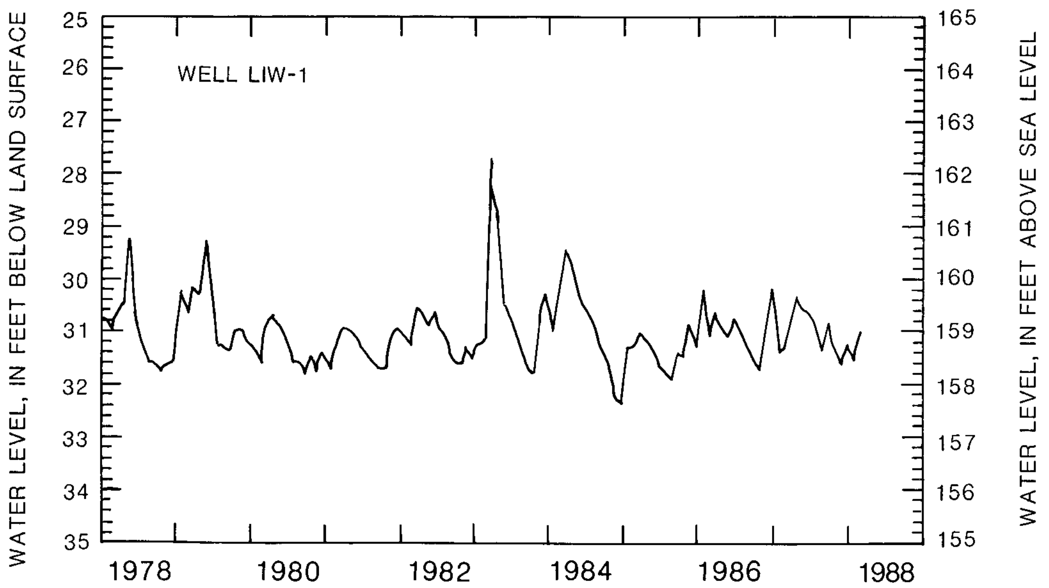


Figure 12.--Water-level fluctuations typical of fluctuations in wells completed in stratified-drift aquifers in elevated kame plains.

Beaver Brook, Spicket River, Little River, Powwow River, and Coastal River basins. The generalized ground-water circulation within a subbasin of the study area is shown in figure 4. The saturated stratified drift within each major flow system can be thought of as a large underground reservoir bounded on top by the water table and on the bottom by fractured crystalline bedrock.

Regional ground-water divides commonly coincide with topographic divides on ridges of till or bedrock. In some areas, major flow systems extend across major drainage divides that separate the lower Merrimack basin from adjacent basins. Areas where this is most likely to occur include the northeast-southwest-trending faults and stress fractures that parallel the regional structural grain and extend beyond the basin boundaries into the Cocheco and Exeter River basins to the north (Lyons and others, 1986).

Ground water in the basin that does not evaporate or transpire or that is withdrawn by wells eventually discharges to lakes, streams, or the Atlantic Ocean. The transit time of ground water in the saturated zone differs spatially, depending on the locations of recharge and discharge zones and the hydrologic characteristics of the aquifer. Generally, the closer to the basin-drainage boundary the water enters the saturated zone, the deeper and greater the distance traveled before it discharges.

Recharge

Recharge is the process by which water is added to the zone of saturation in an aquifer. The amount of water available for development from a stratified-drift aquifer can be limited by the amount of recharge; therefore, when aquifer potential yield is estimated, the amount of water that recharges the system needs to be evaluated. Water pumped from aquifer storage can be replenished by natural and induced recharge, and the contributions from both sources need to be estimated. Neither natural nor induced recharge has been measured directly in the study area. The estimates in this report are based on regional information on ground-water discharge and streamflow.

The sources of recharge to stratified-drift aquifers are infiltration of precipitation that falls directly on the aquifer, surface and subsurface runoff from upland hillslopes adjacent to an aquifer, and leakage from streams that cross the aquifer (fig. 4). Leakage from streams to the ground-water reservoir occurs where the hydraulic head in the aquifer is less than the stream stage.

Under steady-state conditions, the amount of water available for natural recharge from precipitation on stratified drift is roughly equivalent to the average stream runoff (Lyford and Cohen, 1988). Maps showing contours of average annual runoff for the glaciated northeast (Knox and Nordenson, 1955) were adjusted to a more recent period of precipitation (1951-80) to estimate the average annual runoff in southeastern New Hampshire. From this analysis, the approximate recharge to stratified drift from precipitation is 19 in., or 0.9 (Mgal/d)/mi^2.

Natural recharge from tributaries and unchanneled runoff from upland sources can be estimated from streamflow records. The 1966-77 average annual streamflow runoff from till and bedrock uplands was considered a reasonable estimate for the maximum recharge available from all upland sources.

Recharge from unchanneled runoff from upland hillsides has not been extensively documented; however, several investigators (Crain, 1974; MacNish and Randall, 1982; Randall, 1986) assumed that the total runoff from upland hillsides, including surface and subsurface flow, will infiltrate a stratified-drift aquifer at the base of a hillside.

Recharge from tributary seepage losses was not measured directly; however, results from several investigations indicate that rates may vary considerably depending on the location within the basin and streambed characteristics (Morrissey and others, 1988). Measurements of tributary losses in south-central New York (Randall, 1978) indicate that losses were small near the margins of the main valley (average loss of 0.13 ft^3/s per 1,000 ft of channel) and increased to 1.0 ft^3/s or more per 1,000 ft of channel farther downstream. Rates of seepage loss from streams can differ spatially and temporally, and upland streams can gain rather than lose water, making predictions of upland recharge rates difficult. Estimates of the maximum upland tributary losses to aquifers in the study area were assumed to equal the average annual (1966-77) stream runoff for streams draining till and bedrock uplands.

Few data are available on the amount of runoff from till-covered uplands because of the absence of long-term streamflow records. Recharge from upland runoff was calculated from streamflow data collected at a long-term station in the nearby Mohawk River basin in the town of Strafford. Average annual runoff from till and bedrock in this basin is 1.4 (ft^3/s)/mi^2 or 0.9 (Mgal/d)/mi^2. Recharge to the stratified-drift aquifer from upland runoff and tributary losses was calculated by applying the rate of runoff per unit area of upland to the upland areas bordering the aquifer.

Induced recharge to an aquifer can also occur from the pumping of wells near a surface-water body. As previously mentioned, sustained withdrawal of wells that tap stratified drift can lower the water table below an adjacent stream or lake, thereby inducing recharge from the surface-water body to the aquifer. The quantity of water potentially available to the aquifer through induced recharge during dry periods is limited by the streamflow. Excessive pumping near streams can result in undesirable reductions in streamflow and perhaps even the drying up of sections during periods of low flow.

The 7-day, 10-year estimate of low flow ($Q_{7,10}$) can be used as a measure of the amount of induced recharge available during dry periods. Streamflow statistics from several streams in New Hampshire indicate that the ($Q_{7,10}$) is about the 99-percent flow duration, which represents the amount of streamflow equaled or exceeded 99 percent of the time. If all the streamflow at 99-percent duration became induced recharge, the stream would be dry 1 percent of the time on the average. The ($Q_{7,10}$) was selected because (1) flows of this magnitude are a reasonable estimate of the minimum natural ground-water discharge to streams during a 180-day period of no recharge and (2) limiting the induced recharge to this amount reduces the effect of pumping on streamflow, particularly during dry years.

Discharge

Natural ground-water discharge from the aquifers consists of seepage to streams and ponds and ground-water evapotranspiration.

Ground-water discharge to streams sustains their flow in dry weather. After long periods of little or no precipitation, the rate of discharge per square mile from coarse-grained stratified drift is many times greater than that from till (Thomas, 1966). Low flow--and, therefore, ground-water discharge--can be estimated for any site on a stream in New England by determining the upstream drainage area underlain by coarse-grained stratified drift and till-mantled bedrock (Cervione and others, 1982) and by considering certain other watershed properties (Wandle, 1988).

Streamflow was measured at several sites on unregulated streams in 1986 and 1987 during periods when flow was relatively low and consisted entirely of ground-water discharge. Low-flow measurements are summarized in table 3; locations of measurement sites are shown on plates 1-3.

Low-flow measurements at five separate sites on Beaver Brook indicate that for the 2 years of measurements, streamflows increased progressively downstream in direct proportion to the upstream drainage area. No unusual gains or losses were found between measured segments for a 4.2-mi reach of the river. Comparison of low-flow yields (discharge per square mile of drainage area) indicates that similar proportional relations apply to the other streams measured during the study. Anomalous sources or sinks of ground water were not found between measured segments, and most streamflows increased proportionally to upstream drainage area.

Ground-water evapotranspiration is another significant source of discharge from the aquifers and is greatest during the growing season (April-October) when plants use a large amount of water, temperatures are above freezing, and the days are long. Ground-water evapotranspiration has been estimated to range from 1 to 9 inches per year in the northeastern United States (Lyford and others, 1984). Large amounts of moisture are lost to the atmosphere by evapotranspiration from wetlands and marshes where the water table is within 5 ft of the land surface. Streamflow can be reduced where such areas are extensive.

Water Table and Direction of Ground-Water Flow

Generalized water-table altitudes for aquifers are shown on plates 1-3. These maps are based on available water-level data measurements of depth to water at various times during 1986-88 at all observation wells, and surface-water altitudes on topographic maps. Water-table contours represent "average" water levels for a 3-year period (1986-88) and reflect the adjustments made to water-table altitudes during high- and low-water seasons. Because the seasonal fluctuation in water-table altitude is typically 3 to 5 ft in the stratified-drift aquifers (fig. 12), an approximate average water-table altitude was estimated by taking the average of high and low water-table altitudes measured during 1986-88. In addition, topographic maps were used for vertical control, such that the uncertainty of water levels shown on the plates is approximately one-half of the contour interval (10-20 ft) shown on the map.

Arrows drawn at right angles to the water-table contour lines are shown on plates 1-3 to indicate the

Table 3.--Low-flow measurements at miscellaneous sites

[01073830, station identification number, a unique 8 digit station number assigned by the U.S. Geological Survey to stream-gaging stations or single-flow measurement sites with numbers ascending in the downstream direction; mi, mile; mi^2, square miles; ft^3/s, cubic feet per second; (ft^3/s)/mi^2, cubic feet per second per square mile; --, no data]

Location (plate and site number)	Station identification number	Stream	Tributary to	Location	Drainage area (mi^2)	Measurements Date	Measurements Discharge (ft^3/s)	Basin yield ((ft^3/s)/mi^2)
				PICATAQUA RIVER BASIN				
Plate 3 1	01073830	Bailey Brook	Atlantic Ocean	Lat 42°59′25″, long 70°47′48″, Rockingham County, downstream side of bridge at culvert on West Road, 0.15 mi south of intersection with Garland Road, 0.36 mi north of intersection with South Road, 1.82 mi southwest of Rye, N.H. (plate 3).	0.5	10-21-86 8-26-87	0.15 No flow	0.30 --
2	01073835	Bailey Brook	Atlantic Ocean	Lat 42°59′20″, long 70°46′37″, Rockingham County, downstream side of bridge at culvert on Love Lane, 0.22 mi southwest of intersection with Central Road, 0.60 mi northwest of intersection with South Road, 1.7 mi south of Rye, N.H. (plate 3).	1.73	10-21-86 8-26-87	.35 No flow	.20 --
				MERRIMACK RIVER BASIN				
Plate 1 3	010965844	Beaver Brook	Merrimack River	Lat 42°50′21″, long 71°21′00″, Rockingham County, downstream side of Kendall Pond outlet exactly on the Windham-Londonderry town line, 0.01 mi south of the intersection of South Road and Kendall Pond Road, 3.45 mi northwest of Windham, N.H. (plate 1).	30.8	10-21-86 8-25-87	6.71 .97	.22 .031
4	010965846	Beaver Brook	Merrimack River	Lat 42°49′40″, long 71°20′51″, Rockingham County, 50 ft behind house numbered 16 Pleasant Drive, 0.06 mi east of intersection of Pleasant Drive and Tranquil Road, measurment site is also on Windham-Londonderry town line, 2.9 mi northwest of Windham, N.H. (plate 1).	37.7	10-21-86 8-25-87	8.88 1.33	.24 .035

Table 3.--Low-flow measurements at miscellaneous sites--Continued

Location (plate and site number)	Station identification number	Stream	Tributary to	Location	Drainage area (mi^2)	Measurements Date	Measurements Discharge (ft^3/s)	Basin yield ($(ft^3/s)/mi^2$)
MERRIMACK RIVER BASIN--Continued								
Plate 1 5	010965848	Beaver Brook tributary	Beaver Brook	Lat 42°49'02", long 71°20'41", Rockingham County, 50 ft upstream from mouth of tributary to Beaver Brook, 0.07 mi north of Sirod Road, 0.15 mi west of intersection between tributary and Kendall Pond Road, 2.45 mi northwest of Windham, N.H. (plate 1).	--	10-21-86 8-25-87	.99 .06	-- --
6	01096585	Beaver Brook	Merrimack River	Lat 42°48'23", long 71°21'12", Rockingham County, 20 ft upstream from bridge at the intersection of State Route 128 and Anderson Road, 0.28 mi north of the intersection between State Routes 128 and 111, 2.73 mi west of Windham, N.H. (plate 1).	41.8	10-21-86 8-25-87	11.3 1.26	0.27 .030
7	010965851	Beaver Brook	Merrimack River	Lat 42°47'25", long 71°21'53", Rockingham County, upstream from side of bridge on Bridle Bridge Road, at the Windham-Hudson town line, 0.45 mi west of State Route 128, 3.6 mi southwest of Windham, N.H. (plate 1).	43.6	10-21-86 8-25-87	11.5 1.42	.26 .032
8	[1]010965852	Beaver Brook	Merrimack River	Lat 42°46'59", long 71°21'14", on the Rockingham-Hillsborough county line, 100 ft fownstream from bridge on State Route 128 (Mammoth Rd.)., 0.23 mi south of the intersection with Glance Road, 1.4 mi south of West Windham, N.H. (plate 1).	47.8	10-21-86 8-26-87	13.0 1.42	.27 .030
9	010965905	Golden Brook	Beaver Brook	Lat 42°47'32", long 71°18'16", Rockingham County, upstream from side of bridge on Golden Brook Road, 0.5 mi northwest of intersection with State Route 111A, 1.6 mi south of Windham, N.H. (plate 1).	--	8-26-87	.09	[2].023
Plate 2 10	011005034	Taylor Brook	Spicket River	Lat 42°52'20", long 71°13'47", Rockingham County, 50 ft upstream from bridge on Island Pond Road, 0.3 mi northwest of intersection with North Shore Road, 5.42 mi east of Derry, N.H. (plate 2).	4.8	10-20-86 8-26-87	1.07 .45	.22 .09

Table 3.--Low-flow measurements at miscellaneous sites--Continued

Location (plate and site number)	Station identification number	Stream	Tributary to	Location	Drainage area (mi^2)	Measurements Date	Measurements Discharge (ft^3/s)	Basin yield ($(ft^3/s)/mi^2$)
				MERRIMACK RIVER BASIN--Continued				
Plate 2 11	011005038	Taylor Brook	Spicket River	Lat 42°52'10", long 71°13'27", Rockingham County, upstream from side of culvert on North Shore Road, 0.12 mi east of intersection with Island Pond Road, 5.75 mi east of Derry, N.H. (plate 2).	5.0	10-20-86	.71	.14
						8-26-87	.48	.09
12	01100530	Hittytity Brook	Widow Harris Brook	Lat 42°48'18", long 71°13'07", Rockingham County, downstream 100 ft from culvert on Bluff Road, 0.07 mi west of intersection with Zion's Hill Road, 0.46 mi east of intersection with Scotland Avenue, 1.42 mi northwest of Salem, N.H. (plate 2).	9.4	10-20-86	0.94	0.10
						8-27-87	No flow	--
13	01100535	Widow Harris Brook	Spicket River	Lat 42°47'58", long 71°11'58", Rockingham County, at culvert on Bridge Street, 0.23 mi southeast of intersection with Bluff Street, 0.74 mi north of Salem, N.H. (plate 2).	10.8	10-20-86	4.17	.39
Plate 2 14	01100674	Little River Tributary	Little River	Lat°42 51'12", long 71°04'55", Rockingham County, at culvert on Boston and Maine railroad track, 0.38 mi southwest of intersection with Whittier Street Extension, 2.7 mi southwest of Newton, N.H. (plate 2).	7.95	8-25-87	.42	.18
15	01100675	Kelly Brook	Little River	Lat 42°51'15", long 71°06'03", Rockingham County, at culvert on State Route 125, 0.18 mi southwest of intersection with Old County Road and State Route 125, 1.26 mi northwest of Plaistow, N.H. (plate 2).	1.9	10-20-86	.81	.43
16	01100676	Little River	Merrimack River	Lat 42°50'37", long 71°06'07", Rockingham County, downstream side of bridge on North Main Street, 0.32 mi southeast of intersection with State Route 125, 0.6 mi northwest of Plaistow, N.H. (plate 2).	8.8	10-20-86	3.35	.38
						8-25-87	.54	.06

[1]U.S. Geological Survey streamflow-gaging station.

[2]Low flow is affected by regulation such that the discharge per square mile may not be representative.

direction of the horizontal component of ground-water flow. Flow may have a vertical as well as horizontal component in some localities, such as along the basin-drainage divides in and adjacent to the marine-silt and -clay confining layers that are in the vicinity of pumped wells and beneath ponds and streams of the basin. Because of the scale of these maps and the presence of local lenses of fine-grained material, flow directions at specific sites could differ from the regional flow directions shown on the plates.

Aquifer Characteristics

The capacity of a stratified-drift aquifer to store, transmit, and yield water can be partly described by its hydraulic characteristics--saturated thickness, transmissivity, and storage coefficient--and by its boundary conditions. The estimated yield to pumped wells and the resulting drawdowns throughout an aquifer also can be determined if these aquifer properties are known.

Saturated Thickness

Saturated thickness of an unconfined stratified-drift aquifer is the vertical distance between the water table and the bottom of the aquifer. In confined aquifers, saturated thickness is the distance from the top or overlying confining layer of the aquifer to the bottom of the aquifer. The stratigraphy of an unconfined aquifer above a confined aquifer and water levels in these aquifers is shown in figure 5. In the western part of the study area, where extensive fine-grained deposits are absent, the bottom of the stratified-drift aquifer is the till or bedrock surface. Some flow is expected to occur between stratified drift and bedrock wherever open fractures in bedrock are in contact with the stratified drift and there is a hydraulic head difference between these units. Where these conditions exist, the drift and bedrock form a single water-bearing unit. Given the number of high capacity bedrock municipal-supply wells in the study area (8 wells with an average yield of 328 gal/min) recharge from bedrock to stratified-drift aquifers may be significant in some areas. East of the towns of Kingston, Plaistow, and Newton, the bottom of the aquifer typically is the top of the fine-grained marine deposits.

Saturated thickness was mapped separately for unconfined and confined aquifers (pls. 4-6). Saturated thickness of unconfined aquifers was determined by plotting the difference between the water table and the depth to marine sediments, till, or bedrock from interpretation of seismic-refraction profiles (figs. A1-A17, Appendix), seismic-reflection profiles (figs. 13 and 14), logs of wells and test holes, and bedrock outcrops. It should be emphasized that the contours on the saturated-thickness maps apply only to saturated stratified-drift deposits.

The maps of saturated thickness can be used in conjunction with other hydrologic data to indicate favorable areas for the placement of high-yielding production wells. Where all other hydrologic characteristics are equal, a thick aquifer will produce more than a thin aquifer. Stratified drift having saturated thicknesses less than 20 ft cannot usually provide large water supplies, even where the deposits are coarse grained. Such thinly saturated, low-yield areas are found along valley margins and in upland areas where most of the section of stratified drift is dry; examples include kame-terrace deposits high on valley walls in the western part of the study area and elevated kame plains and kame deltas above the marine lowlands in the coastal areas. The saturated thickness is less than 20 ft in any areas of stratified drift where no saturated-thickness lines are shown on plates 4-6. In some of these areas, such as between the Powwow River south of Great Pond and the till contact east of Powwow Pond, the saturated thickness may exceed 20 ft because of discontinuity in the fine-grained marine sediments.

Saturated thickness ranges from 20 to 100 ft in outwash in a long north-south-trending valley that extends from Greenwood Pond to Country Pond in Kingston (pl. 5). This outwash represents the most extensive, thickly saturated aquifer in the study area. Other aquifers that consist of thickly saturated (60 ft or more) coarse-grained material include: the delta along the western shore of Island Pond in Derry; the ice-contact deposits in Newton, south of Amesbury Road; the kame delta that underlies Pease Air Force Base in Portsmouth, specifically the center of the kame delta near the Haven supply well (PXW-2); the kame delta that crosses Route 95 in Greenland; and the kame delta, which includes Knowles Pond, in North Hampton. These aquifers contain the largest stored volume of water within stratified-drift aquifers in the lower Merrimack and coastal river basins.

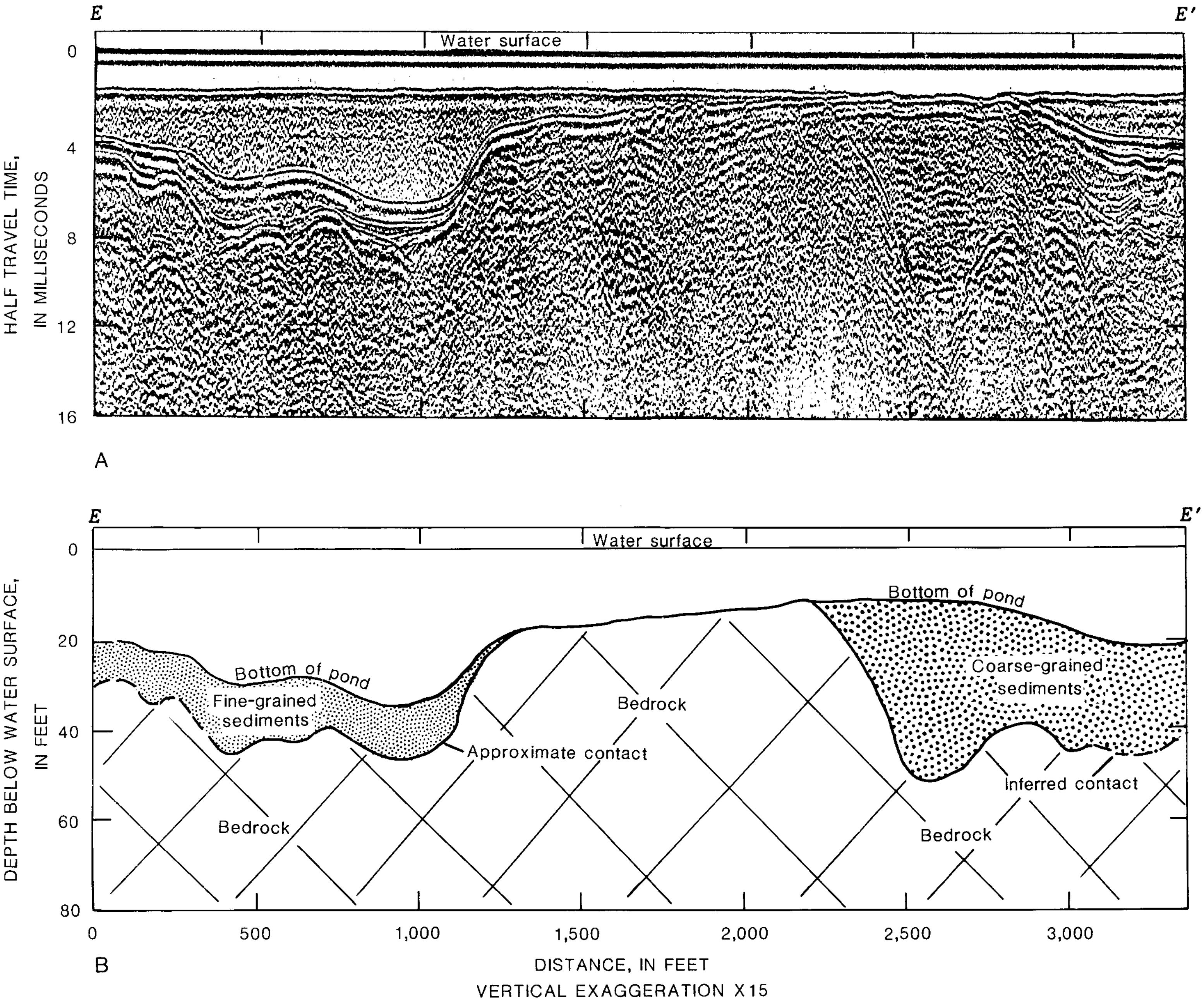


Figure 13.--Seismic-reflection profile (A) and cross section (B) of Cobbetts Pond, Windham, interpreted from seismic-reflection data. Line of section shown on plate 1.

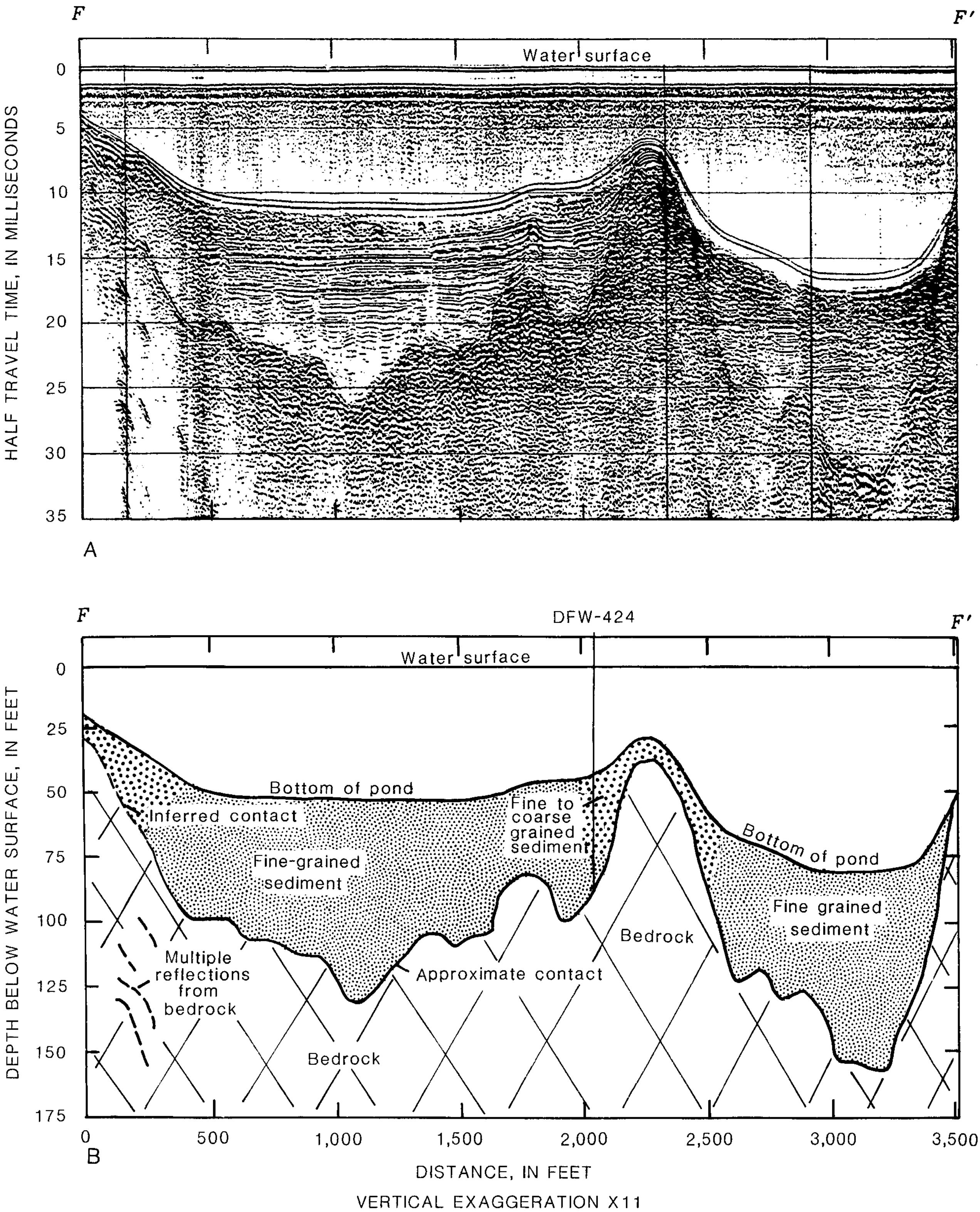


Figure 14.--Seismic-reflection profile (A) and cross section (B) of Island Pond, Derry, interpreted from seismic-reflection data. Line of section shown on plate 2.

Saturated thickness was also mapped for confined aquifers wherever well data were sufficient to locate saturated coarse-grained deposits beneath confining layers. Significant confined aquifers having saturated thickness of 20 ft or more include: a buried ice-contact deposit tapped by the Sherburne municipal well field in south Portsmouth; the aquifer tapped by wells PXW-27, PXW-28, PXW-29, and PXW-39; and the buried ice-contact material at the border of Seabrook, N.H. and Salisbury, Mass., which is tapped by three municipal wells (SGW-44, SGW-45, and SBW-34).

Storage Coefficient

The storage coefficient of an aquifer is defined as the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head (Theis, 1938). The storage coefficient of an unconfined aquifer is equal to specific yield--the volume of water that can be obtained by complete gravity drainage from a unit volume of the aquifer. Laboratory tests done on 13 unconsolidated samples from southern New Hampshire, that ranged from fine-grained lacustrine sands to coarse-grained sands and gravels, indicate that specific yields range from 0.14 to 0.34 and average 0.26 (Weigel and Kranes, 1966). A value of 0.2 is commonly used to estimate specific yield in unconfined aquifers in New England.

The storage coefficient for the confined aquifers was not measured in this investigation. Typical values for confined aquifers are in the range of 0.00005 to 0.005 (Todd, 1980). This range indicates that much less water is released from storage per unit decline in head from confined aquifers than from unconfined aquifers until pumping lowers the head in confined aquifers to the point that they are no longer confined.

Transmissivity and Hydraulic Conductivity

Transmissivity, a measure of the ability of an aquifer to transmit a fluid, is calculated by multiplying the horizontal hydraulic conductivity (the volume of water at the existing kinomatic viscosity that moves in unit time through a unit area of aquifer under a unit hydraulic gradient) by the saturated thickness (Heath, 1983). The transmissivity distribution in an aquifer reflects the combined effects of variations in both of these factors. An aquifer composed of well-sorted, coarse-grained material will have a much higher transmissivity than one with the same saturated thickness but composed of fine-grained material. For example, although the aquifer in Londonderry immediately south of the Manchester/Grenier airport and the aquifer at Pease Air Force Base in Portsmouth have similar saturated thicknesses, the transmissivities of the Portsmouth aquifer are more than double those of the Londonderry aquifer because of differences in texture (grain-size distribution).

Hydraulic conductivities were estimated from stratigraphic logs of wells and test holes for sites for which reliable logs were available. These estimates are based on an empirical relation developed by Olney (1983) that was derived from regression of hydraulic conductivity data. The relation expresses hydraulic conductivity (K, in ft/d) as a function of effective grain size determined from grain-size analyses (D_{10} in phi units)--

$$K = 2{,}100\,(10)^{-0.655D}10, \qquad (1)$$

Effective size is defined as the grain-size diameter at which 10 percent of the sample consists of small grains and 90 percent consists of large grains. A relation between hydraulic conductivity and median grain size (table 4) was developed by applying equation 1 to the results of grain-size analyses of 175 samples obtained from drilling in southeastern New Hampshire. Hydraulic conductivity values were assigned to each interval of a stratigraphic log by applying equation 1 if grain-size analyses were available, or by carefully comparing material descriptions in the logs to table 4 if grain-size distribution curves were unavailable. Estimates for horizontal hydraulic conductivity were made for 48 test holes by applying equation 1 and for 556 wells on the basis of lithologic descriptions and the data in table 4. The latter method is more subjective and less accurate then the method based on grain-size analysis; however, there is agreement between estimates derived from descriptive logs and those derived from quantitative techniques such as specific-capacity and aquifer tests.

Transmissivity at each well and test hole was determined by multiplying the horizontal hydraulic conductivity by the saturated thickness of the corresponding interval of the stratigraphic log and summing the products. Specific-capacity tests and (or) aquifer tests at nine wells also were used to estimate transmissivity by methods described by Ferris and others (1962) and Theis (1963). Although specific-capacity and aquifer tests provide the most accurate estimates of transmissivity, reliable records of such

Table 4.--*Hydraulic conductivity estimated for lithologies in stratified drift in southeastern New Hampshire*

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

Sample description	Median grain size (mm)	Hydraulic conductivity[1] (ft/d)
Sand		
Very fine	0.06 – 0.125	3
Fine	.125 – .25	10
Medium	.25 – .50	30
Coarse	.50 – 1.00	130
Very Coarse	1.00 – 2.00	190
Gravel		
Fine	2.00 – 4.00	250
Coarse	4.00 – 16.00	300 or greater

[1] Estimates based on empirical relations assuming isotropic conditions in the sample [From Olney, 1983].

tests are available for only a few municipal wells or test wells.

Transmissivity zones on plates 4-6 are based on interpolation and extrapolation of transmissivity from individual wells and auger holes, with consideration of saturated thickness and horizontal hydraulic conductivity of the surficial geologic units. Transmissivity can be extremely variable over short distances because of the heterogeneity of stratified-drift deposits. Because the distribution and quality of the transmissivity data differ from place to place and subjective judgement was involved in drawing the zones of equal transmissivity, the zones shown on plates 4-6 should be considered general estimates of transmissivity.

Evaluation of Water Availability and Simulation of Ground-Water Flow

Ground-water availability in the shallow stratified-drift aquifers is generally enhanced where surface-water bodies are in direct hydraulic connection with the aquifer. Recharge induced from a surface-water body augments well yields, especially where large freshwater ponds or lakes are close to pumping centers in coarse stratified drift. In the lower Merrimack River and coastal river basins, however, most public-water-supply wells are in stratified-drift aquifers that are several tens of thousands of feet distant from freshwater bodies; therefore, induced recharge from non-saline surface-water sources is expected to have little to no effect on well yields for most gravel-pack production wells in the study. The pond-aquifer system is an underutilized public-water-supply source that has potential for further exploration and development. Although it is an untapped water-supply resource, if pumpage of the pond (and stream) aquifer system is much greater than natural recharge, streams could show reduced flow or dry up and pond levels could be lowered to a point that water resources are adversely affected.

Two pond-aquifer systems, the one at Cobbetts Pond in Windham and another at Powwow River valley in Kingston (pls. 1 and 2), were selected for detailed hydrologic evaluation because of their potential importance as public water supplies. Water from the ponds is an important source of recharge because of the large storage volume of the ponds and hydraulic connection to the aquifers. Results from drilling and seismic reflection indicate that the ponds generally have a well-sorted, medium-grained sand bottom in the nearshore areas and a dense organic mud in the deep parts of the ponds. Maximum exchange between surface water and ground water is expected where the sandy bottom sediments predominate. The objectives of the evaluation were to (1) estimate the potential yield of the stratified-drift aquifer, (2) delineate the contributing areas for pumped wells withdrawing water from the pond-aquifer system, and (3) discuss selected aquifer areas in terms of favorability for future withdrawal centers.

Description and Conceptualization of Model

The aquifer evaluation involved application of a method that is analagous to the principle of superposition and incorporates the USGS modular three-dimensional finite-difference ground-water-flow model (McDonald and Harbaugh, 1988). The procedure for evaluation involved construction of a water-table map representing a natural (unstressed) aquifer and a numerical model of the aquifer. The model was then used to calculate drawdown that would result from hypothetical pumping stress. Finally, the composite water-table that would result from the pumping of production wells was generated by superimposing the drawdowns calculated with the model on the map of the unstressed water table. The aquifers near Cobbetts Pond and the Powwow River were simulated as two-dimensional ground-water systems, and ground-water withdrawals were simulated for 180 days without recharge. A stress period of 180 days approximately represents the maximum duration of annual periods when evapotranspiration is high and recharge is small. It is assumed that recharge to the aquifers for the remainder of the year would supply sufficient water to sustain pumping during periods of negligible recharge.

Potential yield was determined from the model by analyzing the effect of pumping stress on water levels in the aquifer. The area contributing water to the pumped wells was determined from the final water-table configuration.

The principle of superposition states that, for systems governed by linear equations, the solutions to individual parts of a problem can be added to solve composite problems (Reilly and others, 1987). Superposition reduces many ground-water problems to simpler terms. The most important constraint is that the governing differential equation and boundary conditions must be linear.

The pond aquifers in Windham and Kingston are unconfined (saturated thickness changes in response to pumping); therefore, the governing differential equation is nonlinear and superposition is not strictly applicable. Models can be formulated, however, so that changes in saturated thickness caused by pumping and the corollary changes in transmissivity are taken into account in the simulation. This enables the problem to be formulated by accounting only for changes in stress on the system and calculating drawdowns, which is analagous to superposition. Such models calculate drawdown on the basis of stated initial thicknesses of the aquifer. The water table is specified to start at zero drawdown, and the bottom of the aquifer is set equal to the negative of the saturated thickness. The drawdown solution is dependent on the initial thickness and the amount of change in thickness due to the applied stress. An example of the application of this method of prediction of aquifer response to pumping stress is described by Moore (1990) for similar stratified-drift aquifers in southeastern New Hampshire.

The general procedure for estimating the potential yield of the aquifers was to introduce production wells into the two-dimensional numerical model and vary their discharges. The locations of hypothetical production wells in the model were selected on the basis of hydrologic and practical considerations. The maximum well yields were at aquifer locations of maximum transmissivity and near ponds where induced infiltration could occur. In addition, well sites surrounded by an adequate ground-water-protection area were selected. It was assumed that open space, such as farms, parks, and wetlands, would be the most desirable types of land use to have in proximity to the wells. Selecting sites in open areas was done to satisfy the requirement that wells be surrounded by a minimum 400-foot protection radius, as specified in the New Hampshire Water Supply and Pollution Control Commission rules WS 309.04 and 309.5 of the Drinking-Water Regulations (1984). This protection radius was not always possible for the final well site; however, the chosen area was generally free from conflicting land uses. The maximum possible amounts of ground water that could be withdrawn were determined by adjusting the number of hypothetical pumped wells, the well locations, and well-discharge rates until maximum drawdown within the model approached, but did not exceed, 50 percent of the initial saturated thickness of the aquifer. A limitation of the numerical model is the inability to simulate the decline in pond-stage that would result from pumping. The model assumes the ponds are at constant stage and are unresponsive to changes in pumping stress.

The numerical models developed for the Windham-Cobbetts Pond and Kingston-Powwow River aquifers represent approximations of complex natural systems. Aquifer yields estimated by use of the models are based on several assumptions; therefore, yields should be considered not as exact quantities but rather as reasonable estimates of the maximum amount of water available from wells having specific locations and construction characteristics and tapping a stratified-drift aquifer with specific hydraulic characteristics and boundaries. Certain basic assumptions and limitations of the model simulations are as follows:

1. The aquifer characteristics of transmissivity, saturated thickness, and the water-table configuration shown on plates 1-6 are the source of input values used in the model simulations and are assumed to be representative of the natural system.

2. Prevailing ground-water flow in the stratified drift is predominantly horizontal, there is no change in flow to or from the underlying bedrock as a result of pumping, and the aquifer is isotropic. Vertical-flow components, which could be present near streams or ponds or near partially penetrating production wells, could cause local deviations from simulated water levels.

3. The variable characteristics of the aquifer can be represented by a finite number of blocks or cells, within which each of the aquifer characteristics is assumed to be uniform. Each block represents a discrete area of the aquifer with a single value for saturated thickness, horizontal hydraulic conductivity, and specific yield. This generalization introduces some error because the variability of aquifer characteristics in a given area can be greater than what is represented by the corresponding parameters for a discretized cell of the model.

4. The use of a numerical model to solve the ground-water-flow equation provides only an approximation of the true solution. The numerical model arrives at a solution indirectly through successive iterations of solving the ground-water-flow equation until the maximum head difference between successive iterations is less then a set tolerance level. Errors introduced are minimal as compared with errors involved in estimating aquifer characteristics.

5. All hypothesized production wells are considered to be screened through the full saturated thickness of the aquifer and to be 100-percent efficient. To compensate for the fact that few wells actually meet these ideal well-construction criteria, the maximum allowable drawdown at wells was 50 percent of the initial saturated thickness.

   Because drawdowns in the model represent the average drawdown over the area of a model cell, an adjustment is necessary to estimate drawdown at a pumped well. The total drawdown at well locations was estimated by adding the average drawdown computed by the model for the cell to the additional drawdown in the well calculated from solving the following equation (Trescott and others, 1976):

$$S_A = \frac{2.3Q}{2\pi T} \log \frac{a}{4.81(r_w)}, \qquad (2)$$

   where S_A is an adjustment to drawdown calculated in the cell containing the pumped well;
   Q is discharge at the well, in cubic feet per day;
   T is transmissivity, in ft^2/d;
   a is length of a model cell, in ft; and
   r_w is radius of pumped well, in ft.

   The equation for estimating the total drawdown at a well location is

$$S_T = S_m + S_A, \qquad (3)$$

   where S_T is the theoretical total drawdown, in feet in the production well; and
   S_m is drawdown, in feet, computed by the finite-difference flow model in the model cell containing the pumped production well.

   The theoretical total drawdown represented by this sum, which is always more than the average drawdown computed by the model, more accurately reflects the actual drawdown at the production well than the drawdown for the model cell.

6. The simulation of ponds as head-dependent flux boundaries allows a reasonable estimate for the potential yield of a pond-aquifer system. In the model simulation, the ponds are treated as constant-head boundaries such that surface-water elevations are arbitrarily held at a constant level. In actuality, ground-water withdrawals in an adjacent aquifer may lower pond levels. If the

amount of discharge at the production wells were greater than the amount of water that naturally flows out of the system, pond levels as well as ground-water levels would be lowered throughout the modeled area. Because the model could not simulate this occurrence, an independent calculation of the amount of water that naturally flows out of the system was needed to check the reasonableness of the model solution.

Description of Analytical Method

An analytical method was applied to check the reasonableness of the total amount of water withdrawn from pumped wells compared with the availability of water in the aquifer. Estimates of the amount of water available to an aquifer in hydraulic connection with a surface-water body require information on the quantity of water obtained from the surface reservoir. The low flow of streams, specifically the 7-day, 10-year low flow ($Q_{7,10}$), can be used as an index for evaluating basin-wide potential yield in stream-aquifer systems. The $Q_{7,10}$ is a statistically derived value; the minimum mean discharge for 7 consecutive days in a given year will be equal to or less than the $Q_{7,10}$ average for once in 10 years. This index was chosen as a reasonable approximation of aquifer potential yield after considering the following:

1. Any withdrawal from wells will be balanced by an equal reduction in discharge to streams or ground-water evapotranspiration. As long as the amount of water withdrawn from wells is small in relation to the natural ground-water discharge to streams, the average decline in the water table from pumping should not differ from what would occur naturally during a non-pumping period.
2. The $Q_{7,10}$ is a reasonable estimate of the minimum natural ground-water discharge to streams during a 180-day period without recharge.
3. Verification that the simulated model discharge does not exceed the $Q_{7,10}$ is important because the model is based on the assumption of constant stage in several ponds. If the average decline in the water table during a stress period was no greater than would occur naturally, pond level would not decline significantly during the stress period. Typically, pond levels in these aquifers remain relatively constant throughout the summer; therefore, the model would be a close approximation to reality. If a significant decline in water table were allowed, the model-simulated recharge from ponds would be too large and the results would be invalid.

The aquifer at Cobbetts Pond is drained by a small intermittent stream regulated by sluice gates at the western shore of Cobbetts Pond. Because of this regulation, an accurate measure of the range in natural streamflow is not possible; however, the $Q_{7,10}$ can be estimated by applying an analytical method that was developed for ungaged basins (Cervione and others, 1982). The method, based on the assumption that low flows are sustained by the discharge of water from adjacent aquifers, has been successfully applied to estimate potential yield for stratified-drift aquifers in Connecticut and Massachusetts. The analytical method, developed by Thomas (1966) and later modified by Cervione and others (1982), makes use of a regression equation to relate the $Q_{7,10}$ at any site on a stream to the proportion of upstream drainage area underlain by stratified drift and the proportion underlain by till-mantled bedrock. This regression equation is

$$Q_{7,10} = 0.67A_{sd} + 0.01A_{till}\,, \qquad (4)$$

where $Q_{7,10}$ is the 7-day, 10-year low flow, in ft^3/s;

A_{sd} is the drainage area underlain by coarse-grained stratified drift, in mi^2; and

A_{till} is the drainage area underlain by till-mantled bedrock, in mi^2.

Application of the method is as follows:

1. The basin drainage divide upstream from a site on the master stream of the basin is drawn on the map by use of the topographic contours.
2. The area enclosed by the drainage divide is measured in square miles.
3. The area of stratified drift contained within the drainage divide is measured in square miles. (The area of till-mantled bedrock is equal to the total drainage area less the area of stratified drift).
4. $Q_{7,10}$ is estimated by use of equation 4.

Windham-Cobbetts Pond Aquifer

Water-laid deposits of sand and gravel fill a narrow, linear valley immediately west and to the south of Cobbetts Pond in Windham. This stratified-drift aquifer is currently undeveloped and was selected for evaluation because of its potential importance for public water supply. The location of this valley aquifer is shown in figure 16. The finite-difference grid used to model this area consisted of 67 columns and 26 rows with uniform cell dimensions of 200 ft on a side (fig. 15). Only the cells on the aquifer were considered "active" and were involved in the numerical computations. The total area of aquifer represented by active cells was 1.14 mi^2.

Boundary conditions

The eastern and western borders of the aquifer are defined by the contact between till or bedrock and stratified drift. This geologic contact was represented in the model by a "no-flow" boundary. Although termed a "no-flow" boundary, some flow does occur between the bedrock and stratified-drift aquifers and it may be significant in places. Results from an inventory of public-supply wells indicate that bedrock wells account for approximately one-fifth of the total water pumped from the study area (table 2). Because the bedrock aquifer can be a significant water producer, the flux across the bedrock stratified-drift boundary can be an important component of the total recharge to the stratified-drift aquifer. The exact location, direction, and quantity of flow between these aquifers will vary depending, in part, on the prevalence of water-bearing fractures in the rock and the absence of till semi-confining layers between aquifers. Because this investigation focused on the geohydrology of stratified-drift aquifers little is known about the hydraulic properties of bedrock and till and the effect on water availability in the stratified-drift aquifer. In a superposition model, the no-flow boundary assumes that pumping will not create additional flow across the boundary. This condition was not strictly met in the model because slight (3 ft or less) drawdowns produced by pumping reached the model "no-flow" boundaries in places. The model results, therefore, are slightly conservative in the estimate of aquifer potential yield and the calculated contributing area is slightly larger than what would be estimated if the model were simulated such that pumping could create additional flow across the boundaries.

Natural flux across the till or bedrock, stratified-drift no-flow boundary was accounted for in the model because the map of initial water-table altitudes (fig. 16), upon which drawdowns are superimposed, indicates the natural lateral flow from the till-covered valley walls. The model boundary was arbitrarily terminated southwest of Simpson Pond. This termination corresponds to a southeastern edge of the aquifer that is beyond the reach of drawdowns calculated by the model and, therefore, was simulated as a no-flow boundary.

Three ponds, Cobbetts Pond in the northeastern corner and Rock and Simpson Ponds in the southeastern corner of the model, overlie part of the aquifer. The ponds were simulated in the model as head-dependent flux cells. This type of simulation allows leakage across a semipermeable pond bottom in response to a head gradient between water in the pond and in the aquifer.

The underlying till-bedrock surface was simulated as a no-flow boundary. Some leakage probably occurs across this boundary, but it was assumed to be small enough to be considered negligible in a water balance for the aquifer.

Aquifer parameters

The parameters of hydraulic conductivity, water-table altitude, aquifer-bottom elevation, and specific yield were assigned to each cell in the model. Appropriate values of hydraulic conductivity were determined from the maps of transmissivity and saturated thickness (pl. 4). The grid network is superimposed over these maps, and parameter values were calculated for the respective cells. For example, hydraulic conductivity was computed by dividing transmissivity by saturated thickness. Altitudes of the water-table and pond surface were set to an arbitrary initial value of zero feet, which created a continuous flat boundary suitable for analysis by superposition. Aquifer-bottom elevation was equivalent to the negative of the saturated thickness at each cell.

Parameters describing the hydrologic properties of the pond bottom were assigned to cells in the model that correspond to locations of Cobbetts and Simpson Ponds. Grid cells in which leakage between the pond and aquifer was simulated are shown in figure 16. Depths of the ponds range from 10 to 20 ft and were determined from bathymetric data obtained from seismic-reflection traverses across the

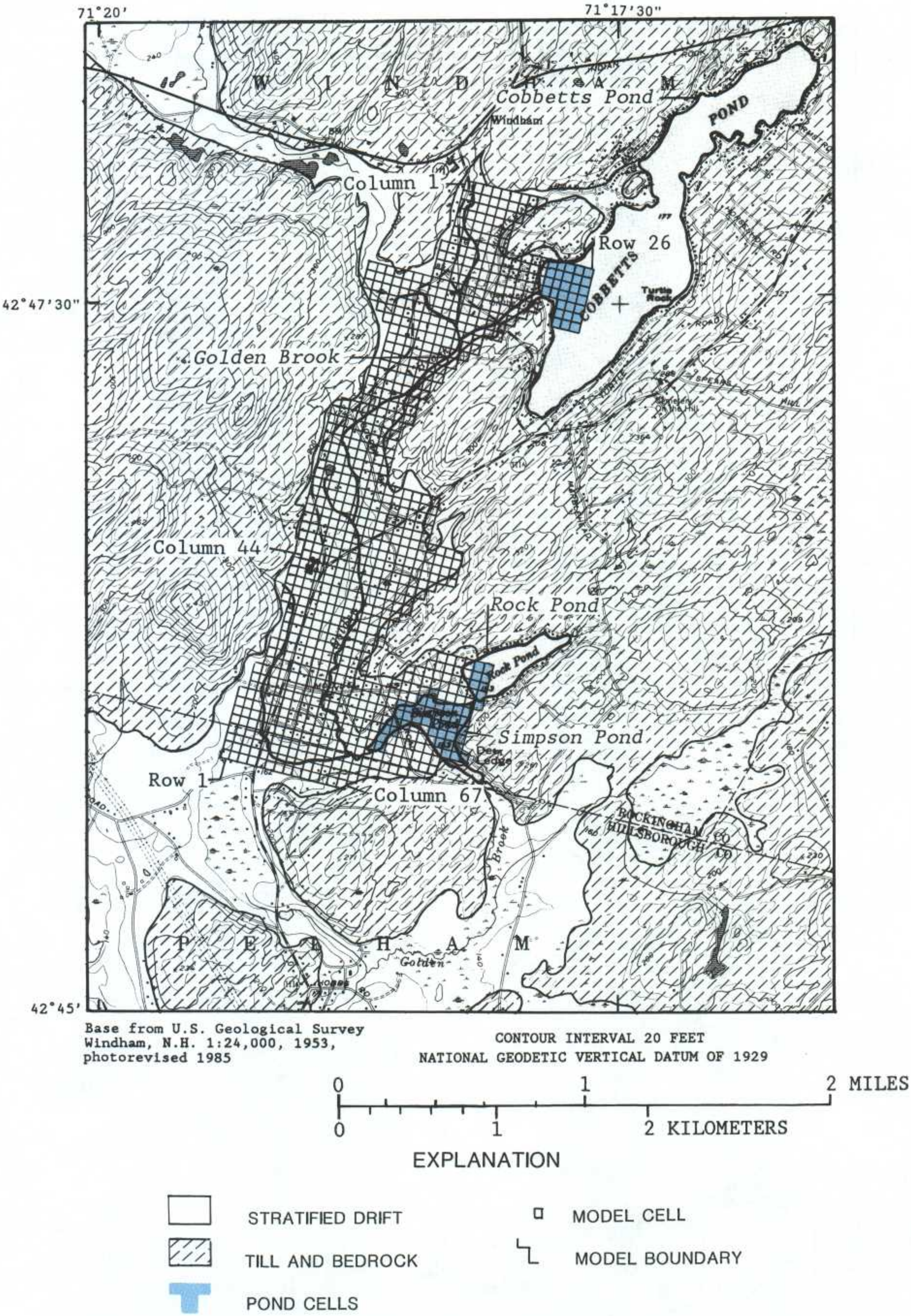
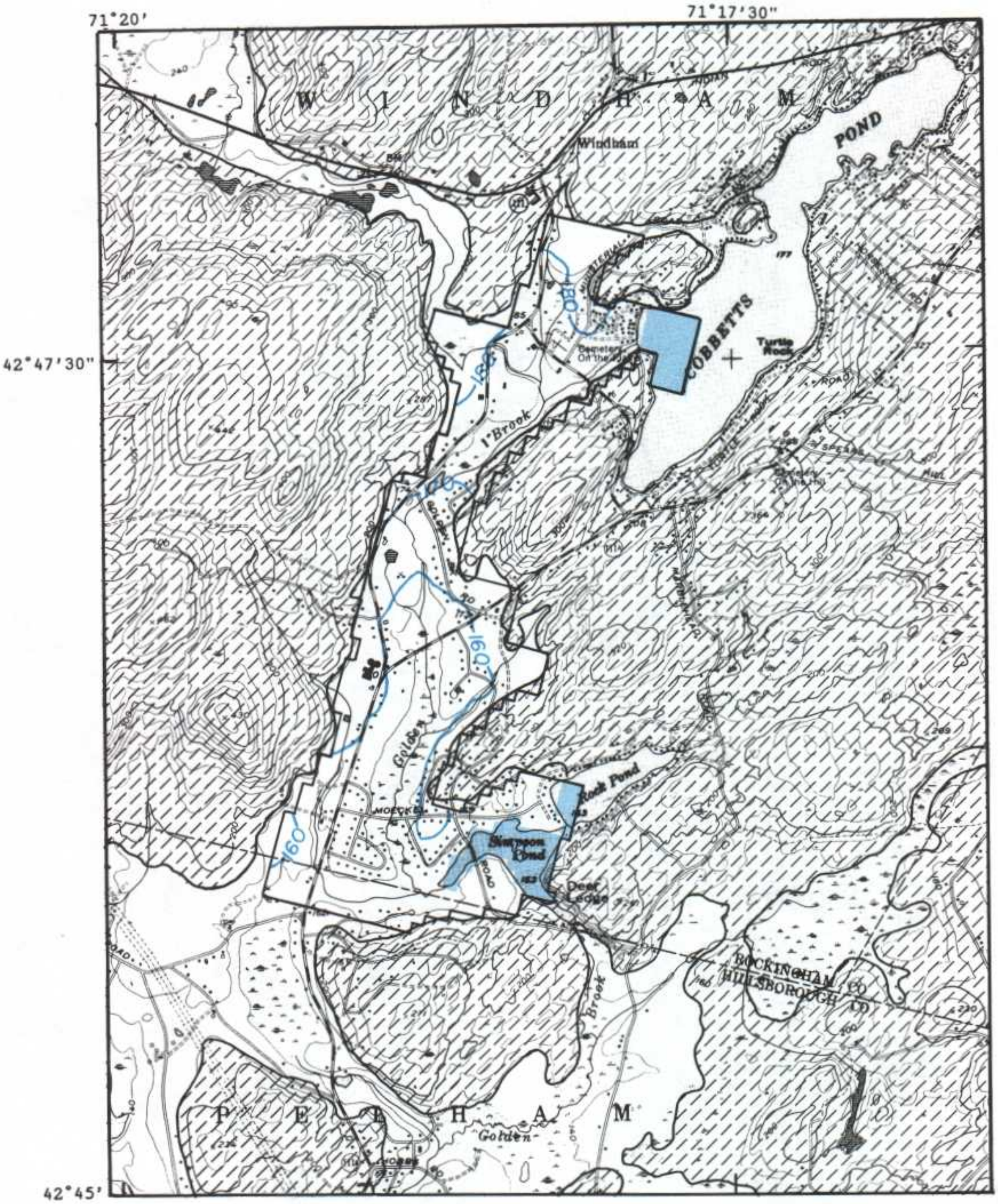


Figure 15.--Finite-difference grid used to discretize the Windham-Cobbetts Pond aquifer.



Base from U.S. Geological Survey
Windham, N.H. 1:24,000, 1953,
photorevised 1985

CONTOUR INTERVAL 20 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

0 1 2 MILES

0 1 2 KILOMETERS

EXPLANATION

STRATIFIED DRIFT

TILL AND BEDROCK

POND CELLS

MODEL BOUNDARY

—170— WATER-TABLE CONTOUR--Shows approximate average altitude of the water table under natural conditions in 1987. Contour interval 10 feet. National Geodetic Vertical Datum of 1929

Figure 16.--Initial water table for the Windham-Cobbetts Pond simulation.

ponds. Figure 13 shows a typical seismic-reflection profile for a traverse across the western shore of Cobbetts Pond in a region of hydrologic connection between pond and aquifer. Hydraulic conductivity of the pond bottom was assumed to be 2 ft/d), and the thickness of the material was assumed to be 2 ft. These estimates were based on values calculated for riverbed deposits (Haeni, 1978; de Lima, 1989), and were assumed to be representative of the pond-bottom hydrology.

The final parameter, specific yield, was estimated to be 0.2 for this unconfined aquifer and was assigned to the entire model area.

Application of numerical model

Five wells were simulated in the model and pumped at various rates until the limits for drawdown were attained at the end of 180 days. Five wells were needed because the model cells that contained the pumped wells went dry if the total discharge simulated was distributed among fewer wells. The locations of the wells and resultant drawdowns are shown in figure 17. The combined pumping rate of the five wells, 0.64 Mgal/d, represents the estimated potential yield of the basin. The potential yield as used in this report describes the rate at which water can be withdrawn from a basin under conditions of pumping for 180 days without recharge and without having adverse impacts on the water resource (see glossary).

Sensitivity analysis

A sensitivity analysis of some of the model parameters was done to show the effect these parameters have on estimates of potential yield. The analysis determined the relative importance of input data values on calculations of potential yield and provided a basis for assessing uncertainty in the simulations given the likely range in each value.

The principal input parameters of aquifer hydraulic conductivity, pond-bottom conductance, specific yield, and duration of pumping were independently increased and decreased by a constant factor throughout the modeled area while other parameters were left unchanged. A reference simulation was selected to represent the best estimate of the hydrologic properties of the aquifer. This reference serves as a standard from which subsequent simulations with different input values can be compared. The amount of adjustment of each parameter differed according to the likely range of each parameter.

The results of the analysis of each change in parameter value are shown in figures 18 and 19 for an east-west profile along column 44 (fig. 15). The profile includes the hypothetical production well of largest simulated discharge (0.135 Mgal/d) and represents an area of the model (column 44, row 11) where the greatest difference in drawdowns occurred between simulations. The computed percentage changes in storage and leakage and the drawdowns for each simulation are listed for comparison in table 5. It should be emphasized that these data represent changes in two of three components of recharge, aquifer storage, and pond and river leakage. Not represented is the change in recharge from till and bedrock uplands, which is commonly simulated in numerical ground-water flow models. The superposition technique, however, simplifies the flow regime by simulating the till-aquifer boundary as "no flow", where additional flow is not induced across this boundary. The tabulated results, therefore, are intended only to show relative differences in the simulations.

Aquifer hydraulic-conductivity (K_a) values were multiplied by factors of 1 (model standard), 2, and 0.5 in three separate simulations. Doubling the hydraulic conductivity of the aquifer resulted in a 32-percent overall reduction in drawdown at production wells and a decrease in the natural ground-water flow to the ponds. The reduction in drawdown extended generally to a distance of 1,500 ft from well locations, beyond which drawdowns were slightly greater than at corresponding cells in the model standard.

Reducing the hydraulic conductivity by 0.5 had the greatest effect on drawdown at well locations for these simulations. Overall, drawdown increased by 57 percent at production wells, and a greater percentage of water came from aquifer storage. Drawdowns decreased slightly relative to the model standard near the till-aquifer boundaries.

Pond-bottom conductance (K_b) values were multiplied by factors of 1 (model standard), 10, and 0.1 in three separate simulations. These changes resulted in computed heads that were identical to those of the model standard. No change in the proportions of water derived from the two sources (table 5) was observed. The insensitivity of the model shows that pond leakage is not the primary control on the availability of water to the simulated production wells, even though one simulated production well was within 600 ft of Cobbetts Pond. The results are probably a reflection of (1) limited