HYDROGEOLOGY, WATER QUALITY, AND GROUND-WATER DEVELOPMENT ALTERNATIVES IN THE BEAVER-PASQUISET GROUND-WATER RESERVOIR, RHODE ISLAND

By David C. Dickerman and Melih M. Ozbilgin

U.S. GEOLOGICAL SURVEY

Water Resources Investigations Report 85-4190

Prepared in cooperation with the

RHODE ISLAND WATER RESOURCES BOARD

Providence, Rhode Island

1985

UNITED STATES DEPARTMENT OF THE INTERIOR

DONALD PAUL HODEL, Secretary

GEOLOGICAL SURVEY

Dallas L. Peck, Director

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CONVERSION FACTORS

For use of those readers who may prefer to use metric (International System) units rather than the inch-pound units used in this report, conversion factors are listed below.

Multiply inch-pound unit	By	To obtain metric (SI) unit
	LENGTH	
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
inch (in)	25.40	millimeter (mm)
	AREA	
acre	0.4047	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
	VOLUME	
gallon (gal)	3.785	liter (L)
cubic foot (ft^3)	0.02832	cubic meter (m^3)
	FLOW	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
gallons per minute per foot [(gal/min)/ft]	0.2070	liters per second per meter [(L/s)/m]
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
	TRANSMISSIVITY	
foot squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)
	HYDRAULIC CONDUCTIVITY	
foot per day (ft/d)	0.3048	meter per day (m/d)

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By David C. Dickerman* and Melih M. Ozbilgin**

ABSTRACT

The 23-square-mile study area is located within the Pawcatuck River basin in southern Rhode Island. Stratified drift is the only principal geologic unit capable of producing yields greater than 350 gal/min (gallons per minute). The stratified-drift aquifer consists of interbedded lenses of sand and gravel, with lesser amounts of silt, silty sand, and clay. Transmissivity of the aquifer ranges from 7,200 to 24,300 feet squared per day. Water-table conditions prevail in the aquifer, which is in good hydraulic connection with perennial streams and ponds.

A digital model of two-dimensional ground-water flow was used to simulate the interaction between surface water and ground water and to evaluate the impact of alternative schemes of ground-water development on ground-water levels, pond levels, and streamflow in the Beaver-Pasquiset ground-water reservoir. Transient simulations of theoretical pumpage were made for a drought period (1963-66) and a wet period (1976-78).

The areas most favorable for development of high-capacity wells (350 gal/min or more) are along the Beaver River and near Pasquiset Pond. Most water withdrawn from wells will be derived from induced recharge from surface-water sources.

The chemical quality of water in the study area is suitable for most purposes. The water is soft and generally contains less than 100 mg/L (milligrams per liter) dissolved solids. Locally, ground water contains elevated concentrations of iron and manganese (7.5 and 3.7 mg/L, respectively), southeast of Pasquiset Pond, and will require treatment if used for public supply.

The ground-water reservoir was simulated with a two-dimensional finite-difference model using a block-centered grid consisting of 33 rows and 75 columns. Model boundaries were treated as constant-flux, leaky constant source head, or leaky. The Beaver River, Pasquiset Brook, and Pasquiset Pond were simulated as leaky boundaries to represent the interaction of the stream-pond-aquifer system. The numerical method used in the model is the strongly implicit procedure.

Long-term historical records were not available for the study area, and it was necessary to utilize estimated long-term average annual ground-water recharge as a stress during the steady state calibration. Differences between measured and simulated water-table altitudes for the final steady state run for 21 selected observation wells averaged +0.07 feet. Combined pumping rates for simulation of ground-water development alternatives at eight sites ranged from 3.25 to 7.00 Mgal/d. Pumping rates for individual wells ranged from 0.25 to 1.50 Mgal/d.

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The model was used to estimate amounts of ground-water pumpage derived from ground-water runoff, induced recharge, and the combination of aquifer storage and reduced evapotranspiration monthly for the 1976-78 period.

Transient simulations suggest that the Beaver-Pasquiset ground-water reservoir is capable of sustaining a pumping rate of 4.25 Mgal/d during years of average ground-water recharge with minimal impact on ground-water levels, pond levels, and streamflow. During extreme drought periods (1965 and 1966) it would be necessary to reduce pumpage below 3.25 Mgal/d to maintain flow in both the Beaver River and Pasquiset Brook.

INTRODUCTION

Background

The Beaver-Pasquiset ground-water reservoir underlies an area of approximately 5 square miles in the valleys drained by the Beaver River and Pasquiset Brook. The ground-water reservoir includes only that part of the study area underlain by stratified drift within the area of the ground-water model shown in figure 1. The ground-water reservoir is located within the Pawcatuck River basin, in southern Rhode Island, and includes parts of the towns of Charlestown, Richmond, and South Kingstown. The outline of the area shown in figure 1 of the ground-water flow model approximates the area of the ground-water reservoir. It is one of nine major ground-water reservoirs in the Pawcatuck River basin (208 Water Quality Management Plan for Rhode Island, 1979) and is one of five in which the Rhode Island Water Resources Board (RIWRB) has done exploratory drilling and aquifer testing.

The RIWRB, which is responsible for implementing development of the State's major water resources, needed to identify sites at which high-capacity wells could be developed that would yield water of suitable quality for municipal-supply use. This need led to the development of a jointly funded study, in 1975, between the RIWRB and the U.S. Geological Survey which involved the collection and analysis of geohydrologic data in each of five ground-water reservoirs. In the Beaver-Pasquiset ground-water reservoir, the RIWRB goal was to identify sites from which an average daily yield of 3 Mgal/d and a maximum pumping capacity of 6 Mgal/d could be obtained. This study also involved cooperation with the University of Rhode Island Civil and Environmental Engineering Department in the development of a ground-water flow model.

The RIWRB proposes (1) to encourage development and management of ground-water resources so as to minimize streamflow depletion during low-flow periods, and (2) to preserve selected favorable sites for future development.

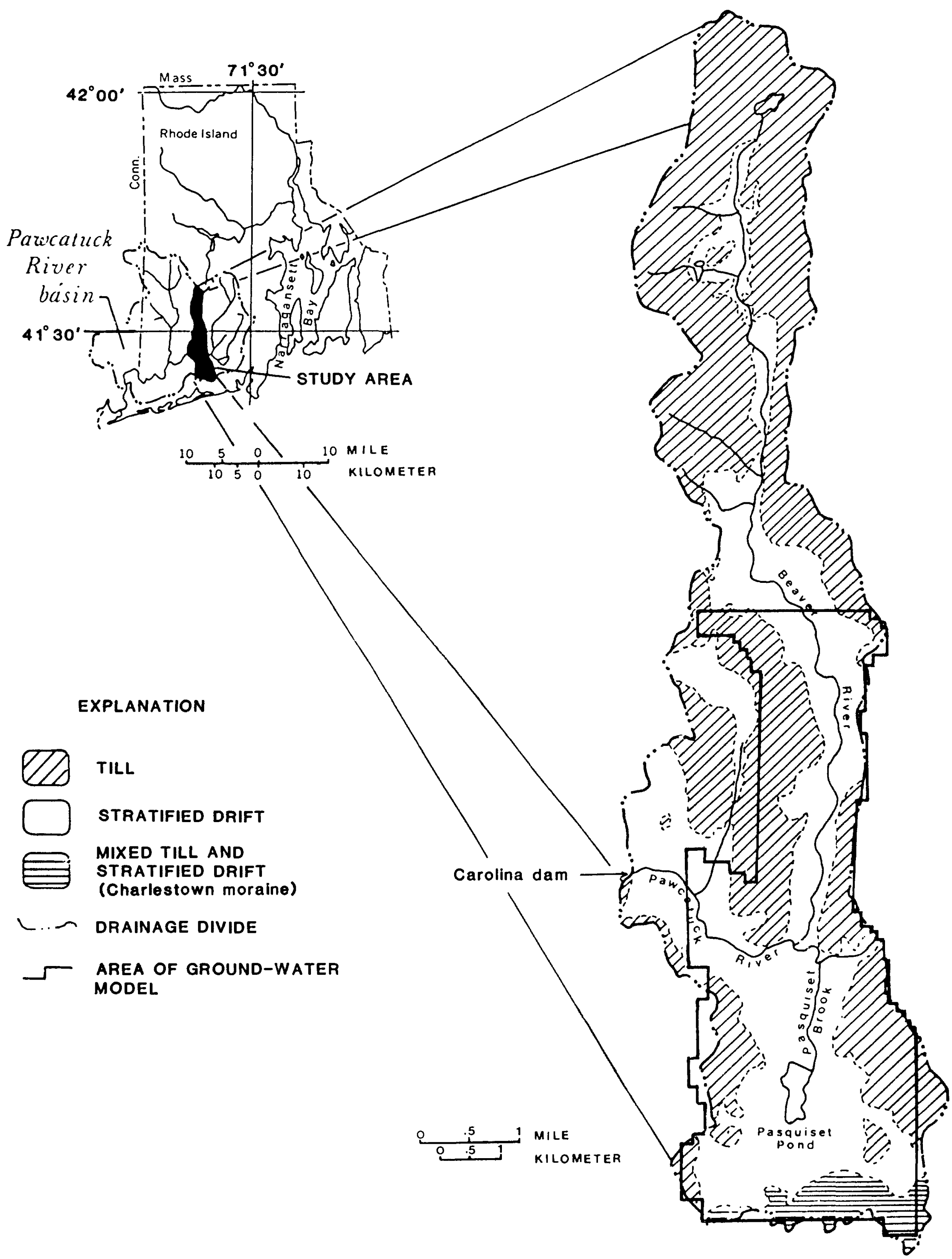

Figure 1.--Location and generalized hydrogeology of the Beaver-Pasquiset study area, and the area covered by the ground-water model.

Purpose and Scope

The report describes the hydrogeologic setting of the study area and the hydrogeology of the Beaver-Pasquiset ground-water reservoir. The objectives of the study were: (1) to determine recharge, stream-aquifer interconnection in, and hydraulic properties of the principal aquifer--the stratified-drift aquifer, (2) to assess the chemical quality of ground water and surface water, and (3) to evaluate, on the basis of ground-water model analysis, the impact of alternative schemes of ground-water development on ground-water levels, pond levels, and streamflow depletion. Hydrogeologic interpretations in this report were based on data collected from September 1974 through September 1976; these data were supplemented by unpublished data collected in previous investigations. The report discusses development of the model and presents results of steady-state and transient simulations of ground-water flow.

Previous and Concurrent Studies

Substantial geohydrologic information is available from earlier studies that include part or all of the study area. Surficial and bedrock geology have been mapped by Feininger (1962), Kaye (1960), Moore Jr. (1959, 1964), and Powers (1957, 1959). Reconnaissance studies on the availability of ground water were done by Bierschenk (1956), Bierschenk and Hahn (1959), Hahn (1959), Lang (1961), and La Sala and Hahn (1960). A comprehensive quantitative study on the availability of ground water in the lower Pawcatuck River basin, Rhode Island, which includes the Beaver-Pasquiset ground-water reservoir, was completed by Gonthier and others (1974). Most of the data on which the present report is based are contained in geohydrologic data reports by Allen and others (1963), and Dickerman and Johnston (1977).

Additional hydrologic data for the Beaver-Pasquiset ground-water reservoir area are being collected by the U.S. Geological Survey as part of the ongoing Pawcatuck River basin study. These data are contained in annual reports titled, "Water Resources Data for Massachusetts, New Hampshire, Rhode Island, and Vermont" (U.S. Geological Survey, 1974), and "Water Resources Data for Massachusetts and Rhode Island" (U.S. Geological Survey, 1975 to present). These data include records of discharge (1974 to present), temperature and specific conductance (June 1979 to present) of the Beaver River near Usquepaug, Rhode Island; discharge (August 1975 to December 1978), temperature and specific conductance (October 1975 to November 1978) of the Beaver River near Kenyon, Rhode Island; measurements of low streamflow at miscellaneous sites, and records of water level fluctuations in observation wells.

Description and Location of Study Area

The Beaver-Pasquiset study area (fig. 1) is in the New England Upland section of the New England physiographic province (Fenneman, 1938, pl. I). The northern three-fourths of the area is characterized by gently rolling topography with rounded hills, several depressions and kettle-hole ponds (Bailey, Long, and No Bottom Ponds), and the narrow, southward-trending Beaver River valley (Feininger, 1962). The southern one-fourth of the area also is characterized by gently rolling topography with rounded hills, but the northward-trending Pasquiset Brook valley is much broader than the Beaver River valley and contains a relatively large swampy area adjacent to Pasquiset Pond. An estimated 85 to 90 percent of the study area is woodland, mostly hardwood, or abandoned pastureland that has become densely overgrown (Moore, Jr., 1959).

The highest topographic point is the summit of Black Plain hill (altitude slightly more than 560 feet above sea level) at the northwest corner of the study area, and the lowest point is at the Carolina dam (altitude 70 feet above sea level) along the western boundary of the study area at the Pawcatuck River outlet. Maximum relief within the 22.7 mi^2 surface-water drainage area is 490 feet. The ground-water and surface-water drainage areas are identical, except along the southern boundary. Here, the water-table divide forming the southern boundary of the 20.8 mi^2 ground-water drainage area is approximately 3,000 feet north of the surface-water divide.

Water Use

Water pumped from wells and streams during 1979 in the Beaver-Pasquiset study area averaged 0.57 Mgal/d. Of this amount, 0.45 Mgal/d (79 percent) was derived from ground water and 0.12 Mgal/d (21 percent) from surface water. Public water supply systems are not available within the study area, and individual home owners rely on well water. Domestic wells accounted for approximately 23 percent (0.13 Mgal/d) of the total water withdrawn during 1979.

The largest water user in the study area is for industry; seventy-five percent (0.43 Mgal/d) of the water withdrawn in 1979 was for industrial use at Kenyon Piece Dye Works Inc. Of this amount, 0.32 Mgal/d was withdrawn from 3 gravel-packed wells (CHW 337, 349, and 410; see fig. 2 for locations) and 0.11 Mgal/d was pumped from the Pawcatuck River. Most of this water was used in industrial processing, with a small amount used for drinking and sanitary needs.

On an average annual basis, water pumped from the Beaver River for irrigation was estimated to be only 0.01 Mgal/d during 1979. However, irrigation water was actually withdrawn at a rate of 0.16 Mgal/d over a four-week period. The withdrawal rate remains the same from year to year, except during average years water is normally withdrawn over an eight-week period. During drought years the withdrawal period is typically extended to 20 weeks.

Eighty-one percent (0.46 Mgal/d) of the water withdrawn from the study area is available for reuse downstream in the Pawcatuck River basin. The remaining 19 percent (0.11 Mgal/d) is lost through consumption and evapotranspiration.

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Acknowledgments

The authors express appreciation to the well drillers and private citizens who provided information and helpful discussion. Special acknowledgment is made to L. R. Andre, Norbert L. Botka, George P. Clark, Mildred L. Godden, Philip M. Green, and Louis E. Pereault who allowed aquifer tests to be conducted on their property. The authors are particularly grateful to William B. Allen, RIWRB; and William E.Kelly, University of Rhode Island Civil and Environmental Engineering Department, for insights gained during many discussions concerning the hydrology and ground-water model of the Beaver-Pasquiset ground-water reservoir.

HYDROLOGIC SETTING

General Geology

The Beaver-Pasquiset study area is underlain by four principal geologic units--bedrock, till, mixed till and stratified drift (Charlestown moraine), and stratified drift. These units differ significantly in geologic origin and in water-yielding characteristics. Of these materials, only those composed predominantly of stratified sand and gravel are sufficiently permeable to yield large quantities (greater than 350 gal/min) of water for development. The stratified drift forms the Beaver-Pasquiset ground-water reservoir, an irregularly shaped body of stratified drift that extends from State Highway 138 on the north to the Charlestown moraine on the south. It ranges in width from 4,000 feet along the Beaver River to 7,000 feet near Pasquiset Pond and has an areal extent of 5 mi^2.

Ground Water

The stratified-drift aquifer along the Beaver River, Pawcatuck River, and Pasquiset Brook is unconfined, and ground-water flow is predominantly horizontal. Within the aquifer there are no known areally extensive layers of impervious sediment to suggest confined conditions at any sites where lithologic data were collected. Locally, however, some parts of the stratified-drift aquifer may be semi-confined where fine-grained sediments were shown in earlier geologic sections.

A map showing the configuration and altitude of the water table in the stratified-drift aquifer was constructed from water levels measured primarily during August to November, 1975 (fig. 2). The direction of ground-water flow in the aquifer is from the till uplands toward the Beaver River, Pawcatuck River, Pasquiset Brook, and Pasquiset Pond.

Surface water

Streamflow records and a knowledge of streamflow characteristics were used to help determine water potentially available for development in the Beaver-Pasquiset ground-water reservoir. Continuous records of streamflow were collected during the study at two U.S. Geological Survey gaging stations on the Beaver River (fig. 2). Data were available for water years 1976-79 for the Beaver River near Usquepaug, R.I. (upper gage), and water years 1976-78 for the Beaver River near Kenyon, R.I. (lower gage). However, the stage-discharge relationship for the lower gage was considered unreliable, except during low-flow periods, because of poor natural stream channel control. Streamflow data for the Pasquiset Brook drainage area was limited to several miscellaneous low-flow discharge measurements. Records of streamflow measurements are published in a report by Dickerman and Johnston (1977) and in annual water-resources data reports of the Geological Survey.

Continuous records of streamflow were not available for the Beaver River during the 1963-66 drought period. However, in order to simulate drought conditions later during ground-water model analysis, these data were necessary. The monthly runoff of the Beaver River was estimated from a correlation curve relating its flow with that of the Pawcatuck River (fig. 3). The relationship was developed using mean monthly runoff data for 1976-78 from the upper gage, and the long-term gaging station on the Pawcatuck River at Wood River Junction, Rhode Island. The long-term gage is located 1.3 miles downstream from the Carolina dam. The reliability of the above relationship to accurately reproduce observed mean monthly runoff from estimated values was tested with data from the upper gage for January 1975 through September 1980. Results of this comparison, shown in figure 4, indicate that estimated values closely approximate observed values.

The duration of streamflow can be shown by a cumulative frequency curve called a flow-duration curve. A flow-duration curve based on short-term records is unreliable for predicting future flows, however, it can be made reliable by adjusting the record to longer periods. The record for the Beaver River near Usquepaug, R.I. (short-term record) was adjusted, based on a method using discharge of equal percent duration (Searcy, 1959), by correlation with the Pawcatuck River at Wood River Junction (long-term record) for the 1942-79 water year reference period. Flow-duration curves, adjusted to the reference period, are shown in figure 5 for the Beaver and Pawcatuck Rivers. The streamflow represented by the high-discharge part of the curves is largely overland runoff and that represented by the low-discharge part is mainly ground-water runoff.

In Rhode Island, the minimum average daily flow for seven consecutive days that can be expected to occur on the average once in ten years is the minimum flow to which water-quality standards for streams apply (R.I Statewide Planning Program and R.I. Dept. of Health, p. A-7, 1976). A relationship developed by Johnston and Dickerman (table 2, 1984) for streams in the Pawcatuck River basin equates the 7-day low flow with a 10 year recurrence interval to the 99-percent flow duration.

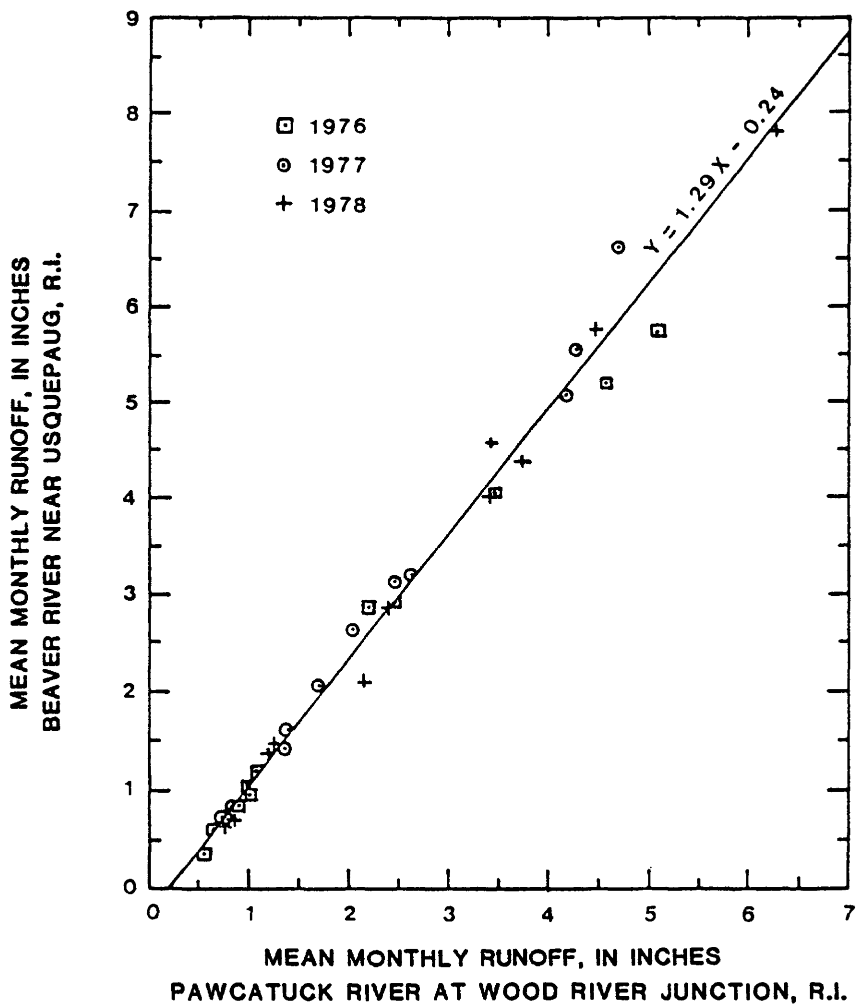

Figure 3.--Correlation between mean monthly runoff at the Beaver River upper gage near Usquepaug, R.I. and the Pawcatuck River at Wood River Junction, R.I.

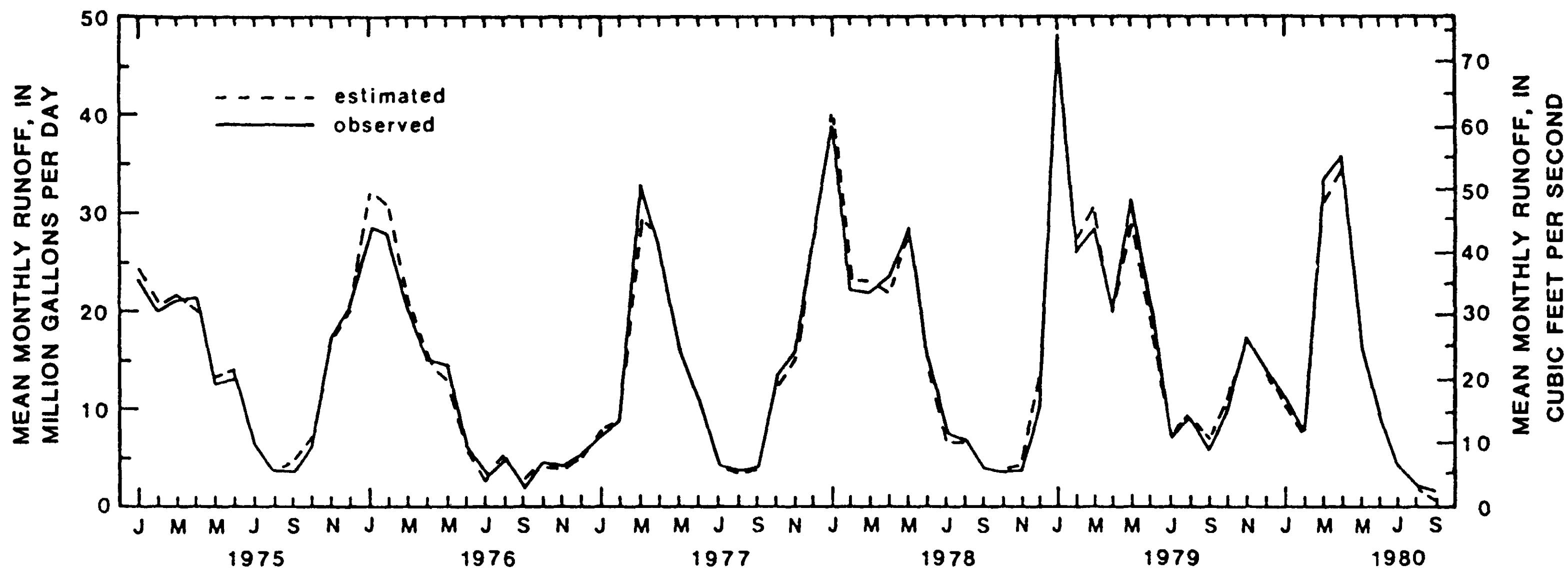

Figure 4.--Comparison between observed and estimated mean monthly runoff at the Beaver River upper gage near Usquepaug, R.I.

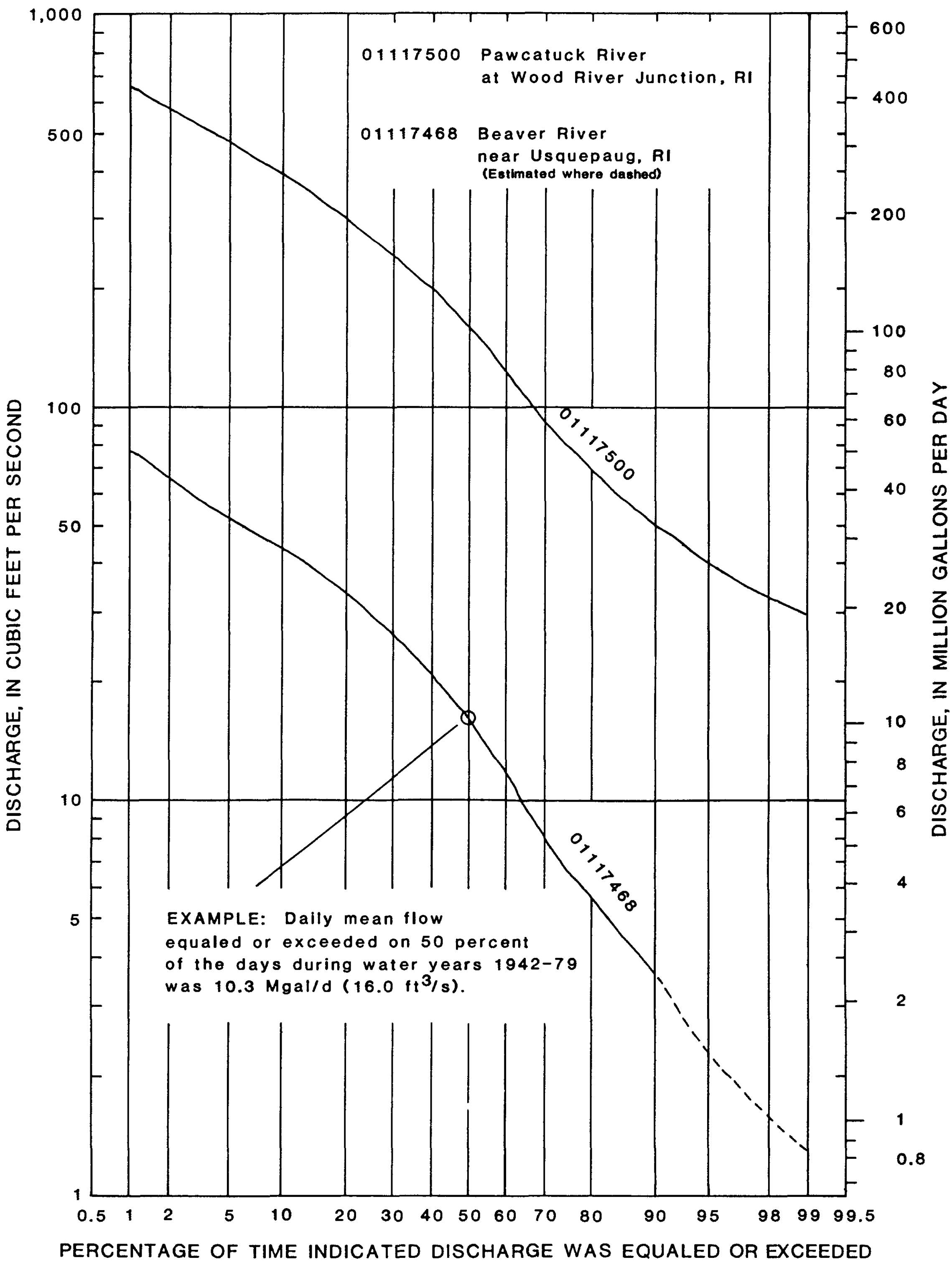

Figure 5.--Duration of daily mean stream discharge at the Beaver River upper gage near Usquepaug, R.I. and the Pawcatuck River gage at Wood River Junction, R.I., 1942-79 water years.

General Water Budget

Water in the Beaver-Pasquiset study area is derived from precipitation, surface runoff, and ground-water underflow from the upper Pawcatuck River basin. Underflow from the upper Pawcatuck River basin, however, was considered negligible (see footnotes 3 and 4, table 1). Water leaves the study area as (1) surface and subsurface outflow at the Carolina dam, (2) ground-water outflow through the Charlestown moraine along the southern edge of the study area, and (3) evaporation and transpiration. Movement of water in the study area may be quantitatively expressed in terms of a water budget, where inflow is equal to outflow plus or minus changes in storage. Averaged over many years of record, however, net changes in storage tend to be small, and are considered negligible. The water budget equation for the study area is expressed as:

$$\underbrace{P + R_t + U}_{\text{INFLOW}} = \underbrace{R_t + U + ET \pm \quad S}_{\text{OUTFLOW}}$$

where P is the precipitation, R_t the total runoff, U the underflow, ET the evaporation and transpiration, and S the change in storage. Components of the long-term average water budget for the Beaver-Pasquiset study area are summarized in table 1.

Annual precipitation at the National Weather Service Station at Kingston, R.I., from 1941-80 ranged from 30.69 (1965) to 68.48 (1972) inches and averaged 46.17 inches (fig. 6). Two climatic conditions, one a wet period and the other a drought period, were selected for ground-water model analysis of alternative schemes of ground-water development. Average annual precipitation was 49.99 inches during the 1976-78 wet period and 37.98 inches during the 1963-66 drought period (fig. 6).

Table 1.--Long-term average water budget for the Beaver-Pasquiset study area, Rhode Island (1941-80). [Mgal/d: million gallons per day]

INFLOW	Mgal/d	OUTFLOW	Mgal/d
Precipitation[1]	51	Total runoff at Carolina (from 95 mi^2)[2]	118
Total runoff from upper Pawcatuck River basin (72 mi^2)[2]	90	Underflow at Carolina	negligible[3]
Underflow from upper Pawcatuck River basin	negligible[4]	Underflow along Charlestown moraine[2]	1
		Evaporation and transpiration[5]	22
		Change in storage	negligible
	141		141

1 Based on long-term average precipitation (46.2 inches) at Kingston, R.I., 1941-80.

2 Estimate based on long-term average runoff (1.25 $Mgal/d/mi^2$) of the Pawcatuck River at Wood River Junction, R.I., 1941-80.

3 Calculated underflow of 0.15 Mgal/d was considered negligible.

4 Calculated underflow of 0.07 Mgal/d was considered negligible.

5 Difference between precipitation at Kingston, R.I. and total runoff of Pawcatuck River at Wood River Junction, R.I.

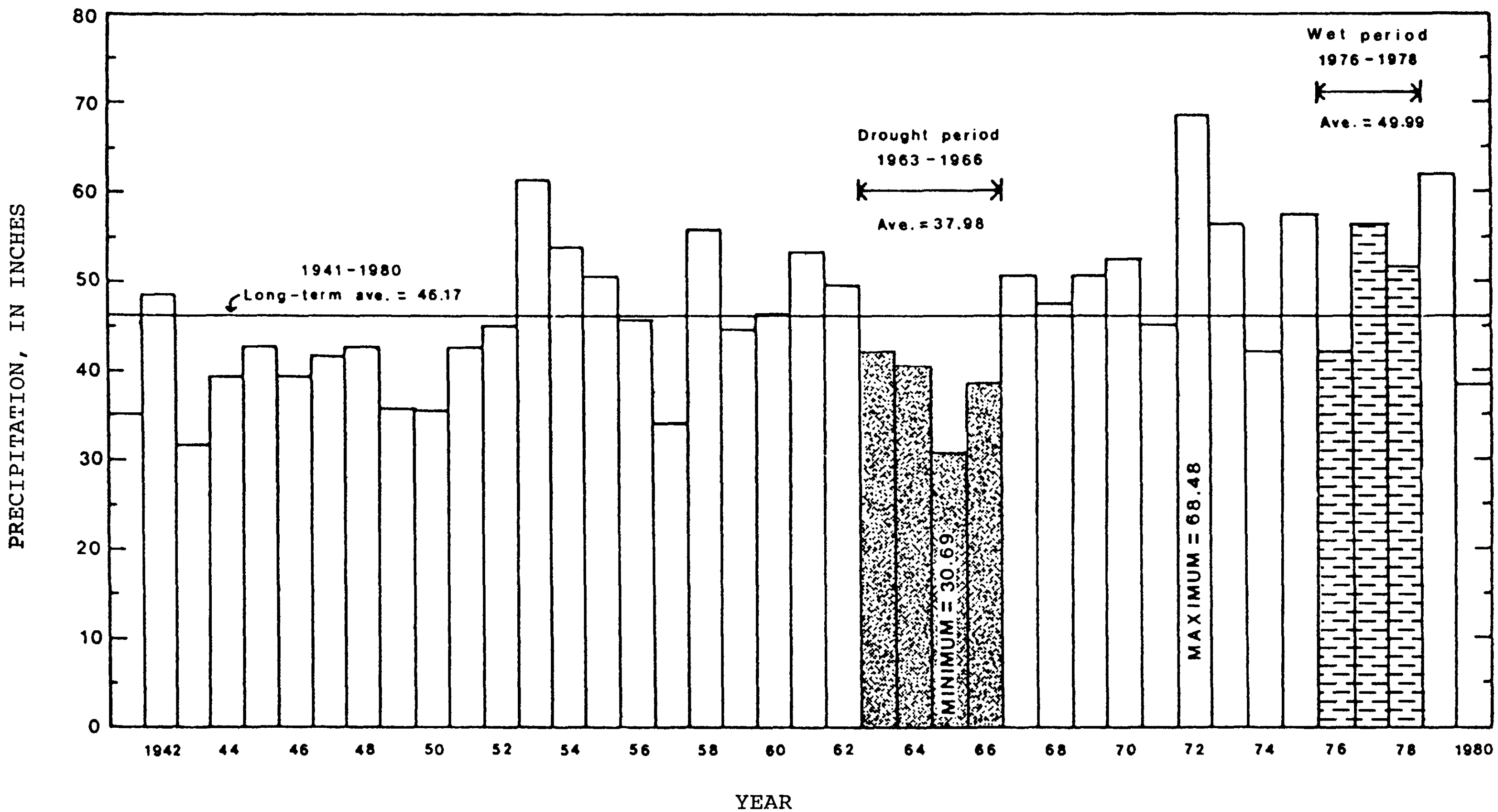

Figure 6.--Annual precipitation at Kingston, R.I., 1941-80.

HYDROGEOLOGY OF THE BEAVER-PASQUISET GROUND-WATER RESERVOIR

Characteristics of the Stratified-Drift Aquifer

Stratified drift covers 44-percent of the study area, and comprises the major aquifer. The drift consists of layers of sorted gravel, sand, silt, and clay that were transported by water from melting glacial ice. The lithologic heterogeneity of this complexly interbedded aquifer system can be seen in longitudinal geologic section A-A' (see fig. 2 for line of section) shown in figure 7. The stream-aquifer system, consisting of the Beaver River, Pasquiset Brook, and the adjacent stratified-drift aquifer, is the principal subject of this report.

Meltwater streams flowing southward from retreating glaciers to the north deposited stratified sand and gravel in the Beaver River valley. This material was named by Feininger (1962) as the eastern valley deposits. The vertical thickness and lithology of these stratified-drift deposits is shown in generalized geologic sections of the northern (B-B', C-C'; fig. 8) and southern (D-D', E-E'; fig. 9) parts of the Beaver River valley (see fig. 2 for line of section). The stratified drift reaches a maximum known thickness of 177 feet at seismic shot point D in geologic section C-C'. The highest yield (670 gal/min) obtained in the study area was from 8-inch test well RIW 395, 400 feet south of geologic section C-C'.

Stratified drift in most parts of the Pasquiset Brook valley is overlain by swamp deposits of Holocene age. This large expanse of swamp made testing drilling difficult to impractical in some areas. The vertical distribution and lithology of the stratified drift is shown in generalized geologic sections north (F-F', fig. 10) and south (G-G', fig. 11) of Pasquiset Pond (see fig. 2 for line of section). In the Pasquiset Brook valley, the maximum aquifer thickness penetrated (164 feet) was at test well CHW 424 shown in geologic section F-F'. The highest tested well yield in the southern part of the Beaver-Pasquiset study area was 610 gal/min. This rate was pumped from test well CHW 400, 800 feet northeast of geologic section G-G'.

All well and seismic shot point locations for the geologic sections are shown in figure 2, along with the water table map of the stratified-drift aquifer.

The only aquifer capable of producing well yields greater than 350 gal/min in the Beaver-Pasquiset ground-water reservoir is the stratified-drift aquifer. The aquifer underlies the Beaver River and Pasquiset Brook valleys and is composed predominantly of sand and gravel, with small amounts of silt and clay. Unconfined conditions prevail in the aquifer, which is hydraulically connected with perennial streams and ponds. However, locally semi-confined conditions exist within the aquifer. The altitude of the water-table, August-November, 1975, is shown in figure 2. The saturated thickness of the stratified-drift aquifer averages 60 to 80 feet and reaches a maximum known saturated thickness of 120 feet (fig. 12). Well yields in the stratified-drift aquifer depend on the natural recharge to the aquifer, the degree of stream-aquifer interconnection, and the hydraulic properties of the aquifer.

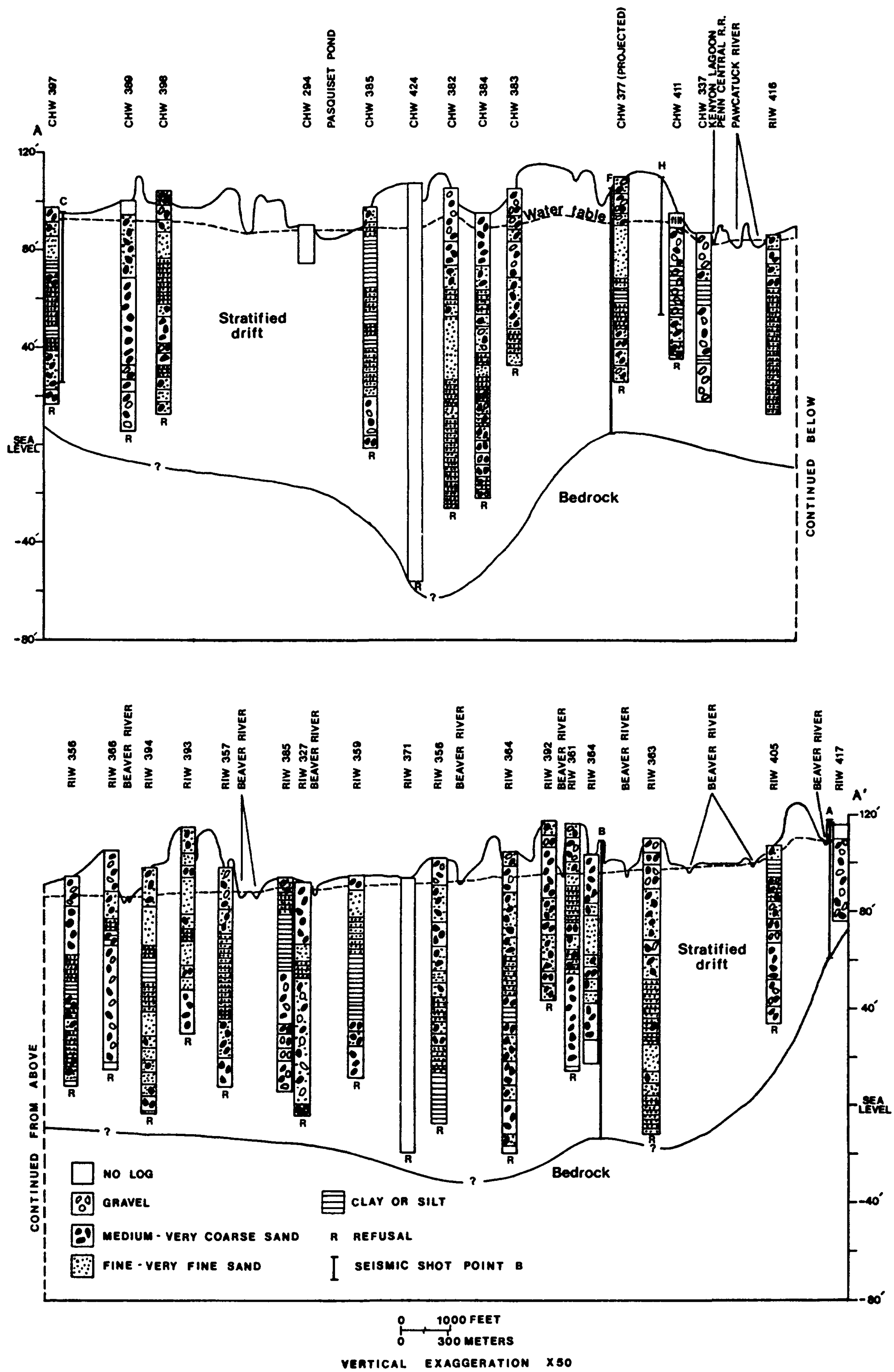

Figure 7.--Longitudinal geologic section of the Beaver-Pasquiset ground-water reservoir showing the complexly interbedded stratified-drift aquifer.

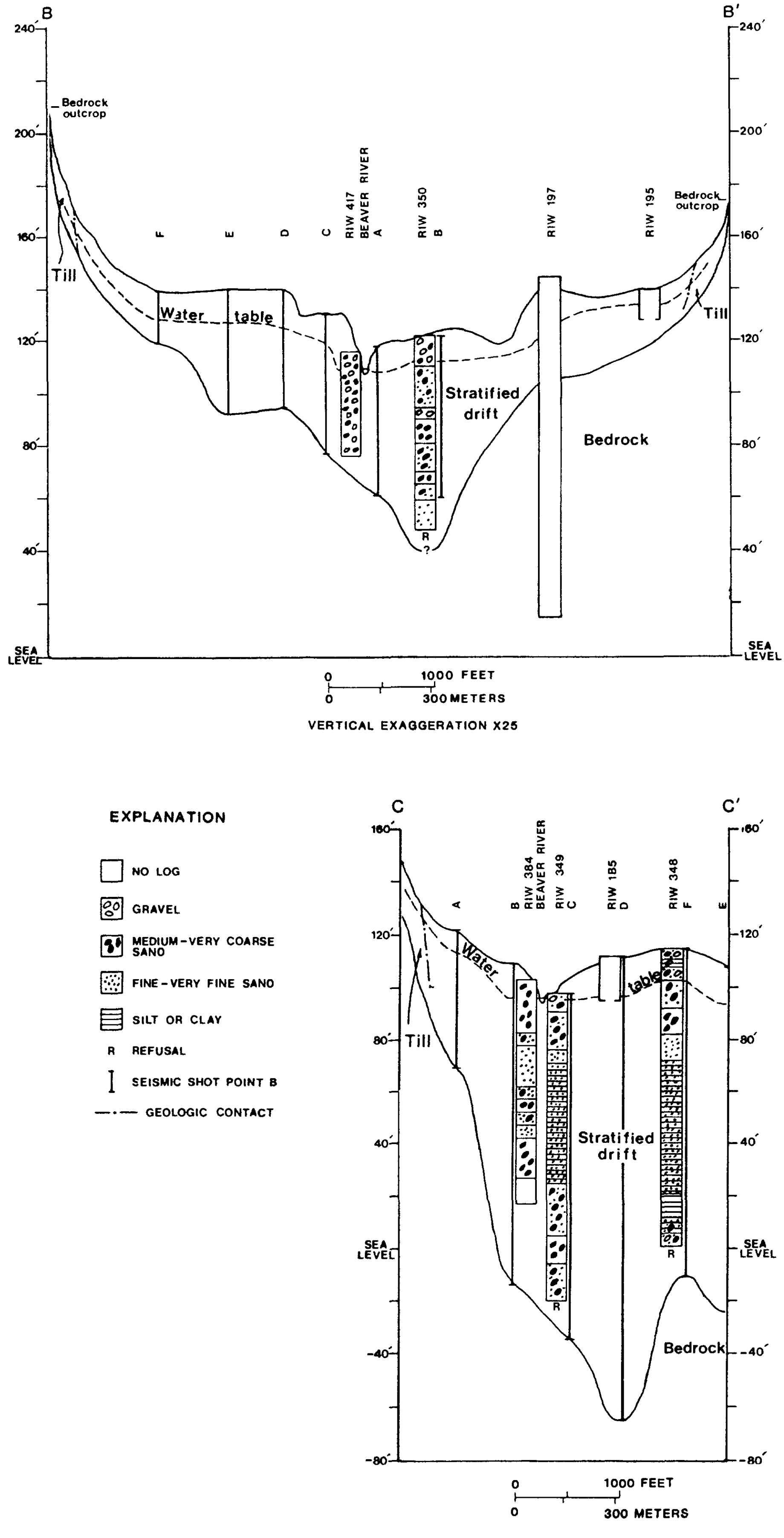

Figure 8.--Generalized geologic sections of the northern part of the Beaver River Valley.

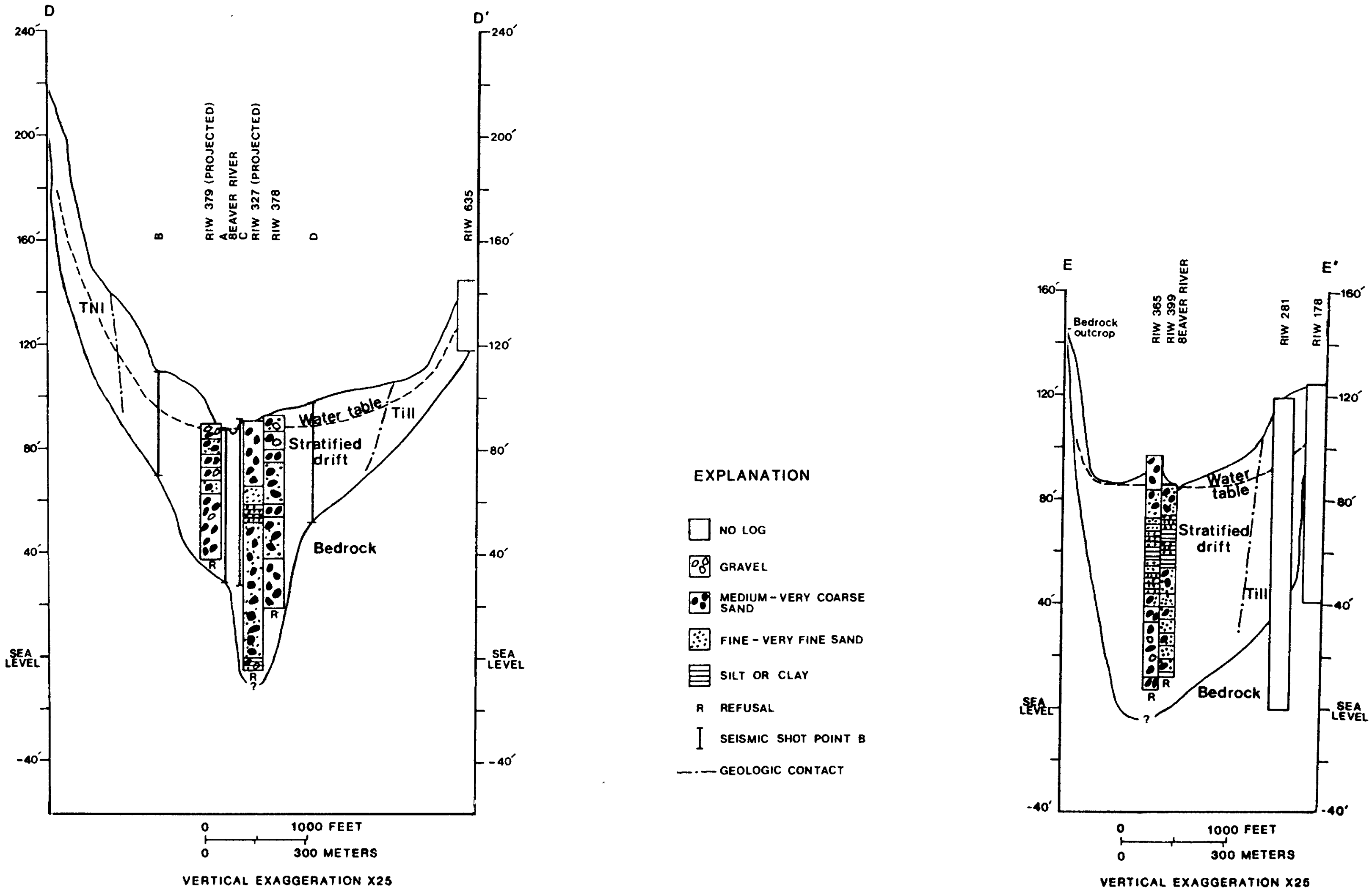

Figure 9.--Generalized geologic sections of the southern part of the Beaver River Valley.

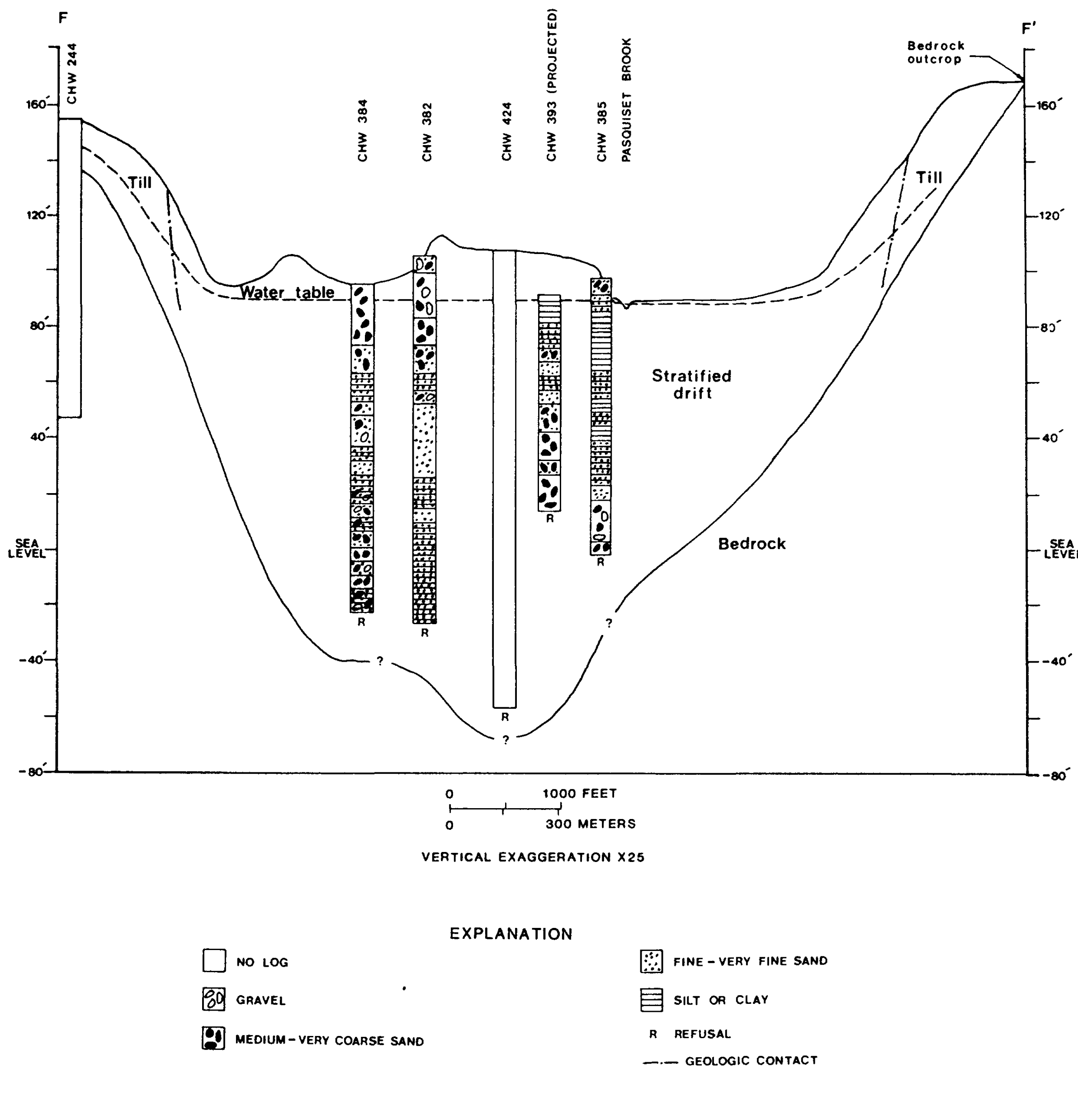

Figure 10.--Generalized geologic section north of Pasquiset Pond.

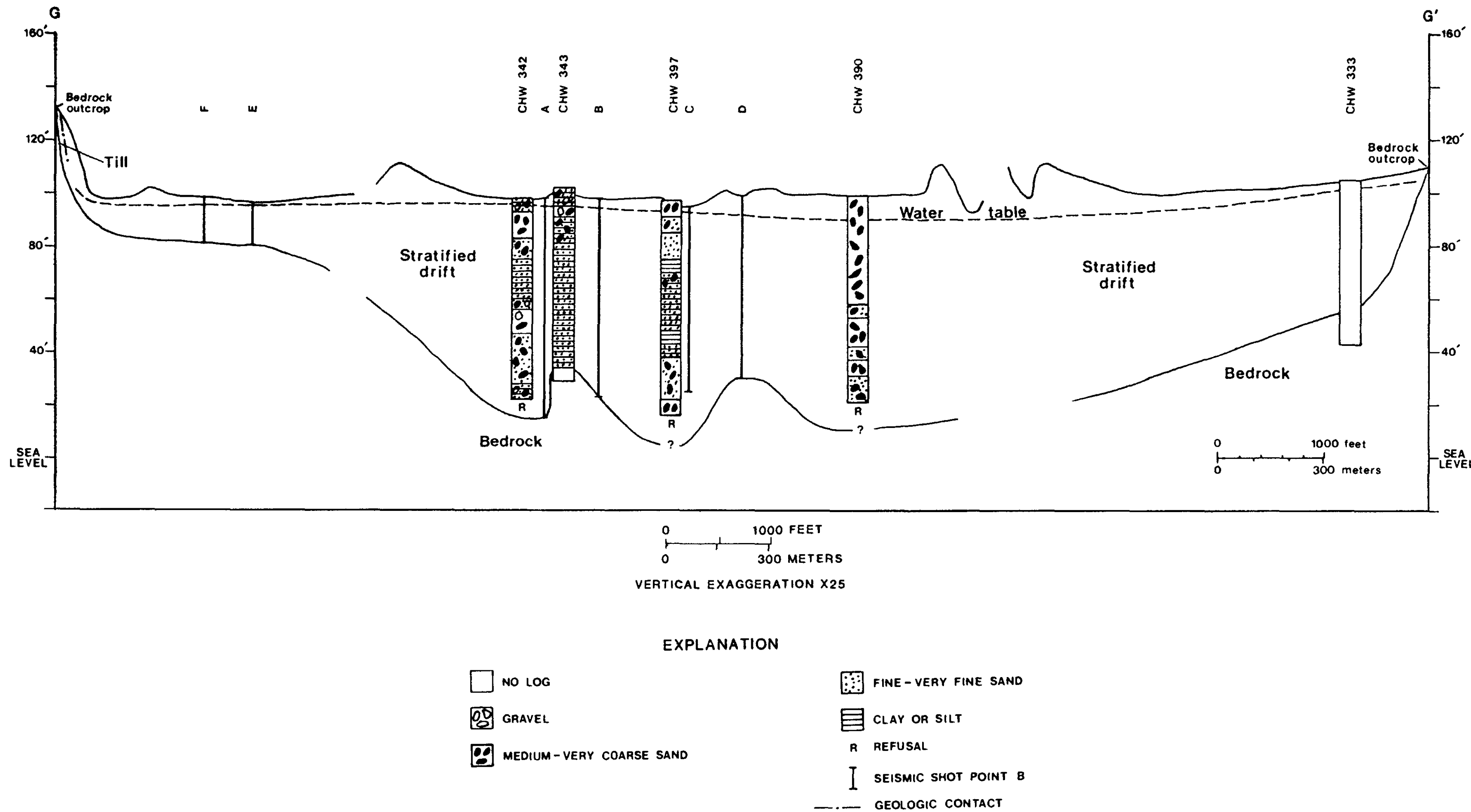

Figure 11.--Generalized geologic section south of Pasquiset Pond.

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Natural Recharge

Under natural conditions, the major source of recharge to the aquifer is precipitation directly on the stratified drift. However, some recharge is also derived from subsurface inflow from adjacent till and bedrock uplands. Variations in precipitation cause changes in recharge. The amount of precipitation that recharges the stratified-drift aquifer cannot be measured directly but can be estimated by determining ground-water discharge and changes in ground-water storage. Ground-water discharge under natural conditions consists of ground-water runoff, ground-water evapotranspiration, and ground-water underflow. Most recharge to the stratified-drift aquifer is eventually released to streams as ground-water runoff. However, some recharge is discharged directly from the water table as ground-water evapotranspiration and as underflow. Ground-water evapotranspiration was not determined in this study and estimated underflow was negligible (less than 0.1 Mgal/d). For purposes of this report, estimates of ground-water recharge are based on ground-water runoff plus or minus changes in ground-water storage. Estimates of ground-water recharge are slightly low, because they do not include ground-water evapotranspiration.

Studies in Connecticut (Randall and others, 1966) and on Long Island, N.Y. (Pluhowsaki and Kantrowitz, 1964, p. 35) show a direct relationship between the amount of till and stratified drift underlying a basin and the percentage of average annual runoff that constitutes ground-water runoff for each type of material. These studies indicate that ground-water runoff constitutes about 35 and 95 percent of average annual runoff from areas underlain by till and stratified drift, respectively. These percentages were used to estimate long-term average annual ground-water runoff. Because net changes in ground-water storage over many years tend to be small, they were considered negligible and ground-water runoff was assumed to equal ground-water recharge over the long term.

Long-term average annual ground-water recharge in the Beaver-Pasquiset study area is estimated to be 16.6 inches (table 2), which is equivalent to a ground-water recharge rate of 9 inches in till areas and 25 inches in stratified drift areas. Table 2 shows a comparison of the long-term average annual water budget for 1941-80 and annual water budgets for 1976-78. Annual ground-water recharge during 1976-78 ranged from 16.56 inches (1976) to 26.22 inches (1977) and averaged 21.9 inches. Average annual recharge during 1976-78 was 5.3 inches above long-term average recharge.

Table 2.--Comparison of the long-term average annual water budget for 1941-80 with annual water budgets for 1976-78.
[All data in inches]

	Pawcatuck River at Wood River Junction, R.I. Long-term average annual 1941-80	Beaver River near Usquepaug, R.I.		
		1976	1977	1978
(a) Precipitation[1]	46.2	42.1	56.1	51.7
(b) Total runoff	26.2	26.6	33.6	36.4
(c) Water loss (a-b)	20.0	15.5	22.6	15.4
(d) Ground-water runoff	16.6	20.5	22.4	25.3
(e) Surface-water runoff (b-d)	9.6	6.1	11.2	11.1
(f) Change in ground-water storage	0.0	-4.0	3.8	-2.4
(g) Ground-water recharge (e+f)	16.6	16.6	26.2	22.9

1 Precipitation measured at the National Weather Service station at Kingston, R.I.

Monthly water budgets for the Beaver-Pasquiset ground-water reservoir are given in table 3 for 1976-78 and table 4 for 1963-66. For the 1976-78 wet period, ground-water recharge was considered equal to ground-water runoff plus or minus the average change in ground-water storage. Ground-water runoff was determined by hydrograph separation using a ground-water rating curve developed from precipitation, streamflow, and ground-water level data. Data points for the rating curve were selected on days when streamflow at the Beaver River upper gage was considered to consist mostly of ground-water runoff. It was assumed that streamflow consisted mostly of ground-water runoff 10 days after precipitation, if both streamflow and ground-water levels were declining. Stream discharge and water level data points on days selected by this procedure were plotted to develop the ground-water rating curve. On days when streamflow did not consist mostly of ground-water runoff, ground-water runoff was estimated from the rating curve using water level data measured on the same day.

For the 1963-66 drought period, monthly ground-water recharge was estimated using a relationship (fig. 13) developed between mean monthly runoff at the Beaver River upper gage and estimated monthly ground-water recharge for 1976-78. Ground-water recharge values determined for 1963-66 using this relationship are approximate, but considered best estimates based on available data. The standard error of estimate for determinations of ground-water recharge using figure 13 is $\pm$ 0.77 inches. Although estimated monthly recharge in table 4 for 1963-66 may vary $\pm$ 0.77 inches, the authors believe that the recharge pattern and amounts are representative of similar drought conditions.

Hydraulic Properties

Transmissivity and specific yield are the hydraulic properties that determine the capacity of an aquifer to transmit, store, and yield water. The product of the horizontal hydraulic conductivity and saturated thickness of the aquifer is transmissivity. Also important to the water-yielding potential of the aquifer is the vertical hydraulic conductivity of the streambed or aquifer, whichever is lower. It is the lower vertical hydraulic conductivity that controls the rate at which streamflow moves into the aquifer and toward the well screen as induced recharge during pumping.

Geohydrologic data collected as part of this study were published in an earlier report titled, "Geohydrologic data for the Beaver-Pasquiset ground-water reservoir, Rhode Island" (Dickerman and Johnston, 1977). Detailed lithologic logs of 110 wells and test holes and drawdown/recovery data for eight aquifer tests obtained between September 1974 and September 1976 were analyzed to determine aquifer hydraulic properties.

The hydraulic properties of the stratified-drift aquifer were determined by analyses of unadjusted drawdown and recovery data by one or more of the following methods: (1) Stallman (1963, 1965) method for vertical movement in an unconfined, anisotropic aquifer, (2) Cooper (1963) method for nonsteady radial flow in a leaky confined aquifer, (3) Walton (1960) method

Table 3.--Summary of transient monthly water budgets, in inches, for 1976-78.

(a) Month	(b) Precip-itation[1]	(c) Water loss (b-d)	(d) Total runoff[2]	(e) Surface-water runoff (d-f)	(f) Ground-water runoff[3]	(g) Change in ground-water storage[4]	(h) Ground-water recharge (f+g)
				1976			
January	6.75	1.03	5.72	1.57	4.15	0.96	5.11
February	3.16	-2.04	5.20	0.94	4.26	-0.96	3.30
March	3.66	-0.41	4.07	0.51	3.56	-0.24	3.32
April	1.69	-1.21	2.90	0.33	2.57	-1.20	1.37
May	3.15	0.29	2.86	0.84	2.02	-0.55	1.47
June	0.78	-0.42	1.20	0.11	1.09	-0.96	0.13
July	2.28	1.67	0.61	0.06	0.55	-0.98	-0.43
August	7.49	6.53	0.96	0.43	0.53	-0.05	0.48
September	2.48	2.12	0.36	0.09	0.27	-0.43	-0.16
October	6.13	5.27	0.86	0.35	0.51	0.55	1.06
November	0.76	-0.03	0.79	0.31	0.48	-0.46	0.02
December	3.76	2.71	1.05	0.52	0.53	0.36	0.89
Total	42.09	15.51	26.58	6.06	20.52	-3.96	16.56
				1977			
January	3.53	2.12	1.41	0.52	0.89	1.20	2.09
February	2.91	1.31	1.60	0.80	0.80	0.91	1.71
March	5.92	-0.70	6.62	2.97	3.65	2.54	6.19
April	4.40	-0.68	5.08	0.46	4.62	-1.13	3.49
May	3.05	-0.15	3.20	0.62	2.58	-0.91	1.67
June	5.41	3.33	2.08	0.47	1.61	-1.10	0.51
July	1.46	0.61	0.85	0.10	0.75	-1.08	-0.33
August	4.78	4.05	0.73	0.31	0.42	-0.31	0.11
September	5.26	4.55	0.71	0.36	0.35	0.17	0.52
October	8.02	5.41	2.61	1.30	1.31	1.44	2.75
November	3.61	0.49	3.12	1.30	1.82	0.53	2.35
December	7.79	2.24	5.55	1.95	3.60	1.56	5.16
Total	56.14	22.58	33.56	11.16	22.40	3.82	26.22
				1978			
January	9.48	1.68	7.80	2.58	5.22	1.39	6.61
February	1.78	-2.22	4.00	0.40	3.60	-2.54	1.06
March	3.24	-1.13	4.37	1.56	2.81	1.39	4.20
April	2.17	-2.37	4.54	1.29	3.25	-1.08	2.17
May	9.60	3.84	5.76	1.99	3.77	1.37	5.14
June	1.44	-1.41	2.85	-0.03	2.88	-2.30	0.58
July	3.14	1.68	1.46	0.39	1.07	-1.34	-0.27
August	4.82	3.45	1.37	0.74	0.63	-0.38	0.25
September	2.62	1.87	0.75	0.41	0.34	-0.58	-0.24
October	3.81	3.15	0.66	0.27	0.39	0.02	0.41
November	4.02	3.32	0.70	0.34	0.36	0.48	0.84
December	5.61	3.51	2.10	1.12	0.98	1.18	2.16
Total	51.73	15.37	36.36	11.06	25.30	-2.39	22.91

1 Precipitation measured at the National Weather Service station at Kingston, R.I.
2 Measured at the upper gaging station on the Beaver River near Usquepaug, R.I.
3 Determined by hydrograph separation using a ground-water rating curve developed from observation well RIW 417.
4 Average changes in ground-water storage determined from long term water-level observation wells in the Pawcatuck River basin.

Table 4.--Summary of transient monthly water budgets, in inches, for 1963-66.

Month	Precipitation[1]	Water loss[2]	Total runoff[3]	Ground-water recharge[4]	Precipitation[1]	Water loss[2]	Total runoff[3]	Ground-water recharge[4]
		1963				1964		
January	3.65	0.14	3.51	2.64	5.53	1.75	3.78	2.89
February	3.31	-0.37	3.68	2.80	3.47	-0.57	4.04	3.13
March	3.60	-1.71	5.31	4.28	2.67	-1.48	4.15	3.23
April	2.17	-0.53	2.70	1.91	7.95	0.73	7.22	6.02
May	4.82	1.58	3.24	2.40	0.67	-2.51	3.18	2.34
June	1.45	-0.71	2.16	1.42	0.86	-0.32	1.18	0.52
July	4.69	3.73	0.96	0.32	4.46	3.67	0.79	0.17
August	2.81	2.02	0.79	0.17	0.88	0.49	0.39	-0.20
September	3.93	3.47	0.46	-0.13	4.18	3.89	0.29	-0.29
October	1.94	1.44	0.50	-0.10	2.49	1.94	0.55	-0.05
November	6.89	5.63	1.26	0.60	3.04	2.48	0.56	-0.04
December	2.78	0.75	2.03	1.30	4.45	2.74	1.71	1.01
Total	42.04	15.44	26.60	17.61	40.65	12.81	27.84	18.73
		1965				1966		
January	3.37	1.30	2.07	1.33	2.72	2.07	0.65	0.04
February	3.65	0.56	3.09	2.26	3.71	2.04	1.67	0.97
March	1.73	-2.30	4.03	3.12	2.85	-0.59	3.44	2.58
April	3.69	0.60	3.09	2.26	1.40	-0.23	1.63	0.93
May	1.94	-0.37	2.31	1.55	6.61	4.36	2.25	1.50
June	1.55	0.31	1.24	0.58	2.28	0.39	1.89	1.17
July	2.30	1.84	0.46	-0.13	2.13	1.42	0.71	0.10
August	2.83	2.50	0.33	-0.25	1.47	1.09	0.38	-0.20
September	2.99	2.71	0.28	-0.30	4.92	4.49	0.43	-0.16
October	3.14	2.72	0.42	-0.17	2.71	2.11	0.60	0.00
November	1.68	1.31	0.37	-0.21	4.82	3.54	1.28	0.61
December	1.82	1.32	0.50	-0.10	2.92	1.79	1.13	0.48
Total	30.69	12.50	18.19	9.94	38.54	22.48	16.06	8.02

1 Precipitation measured at the National Weather Service station at Kingston, R.I.
2 Precipitation minus total runoff.
3 Estimated using relationship shown in figure 3.
4 Estimated using relationship shown in figure 13.

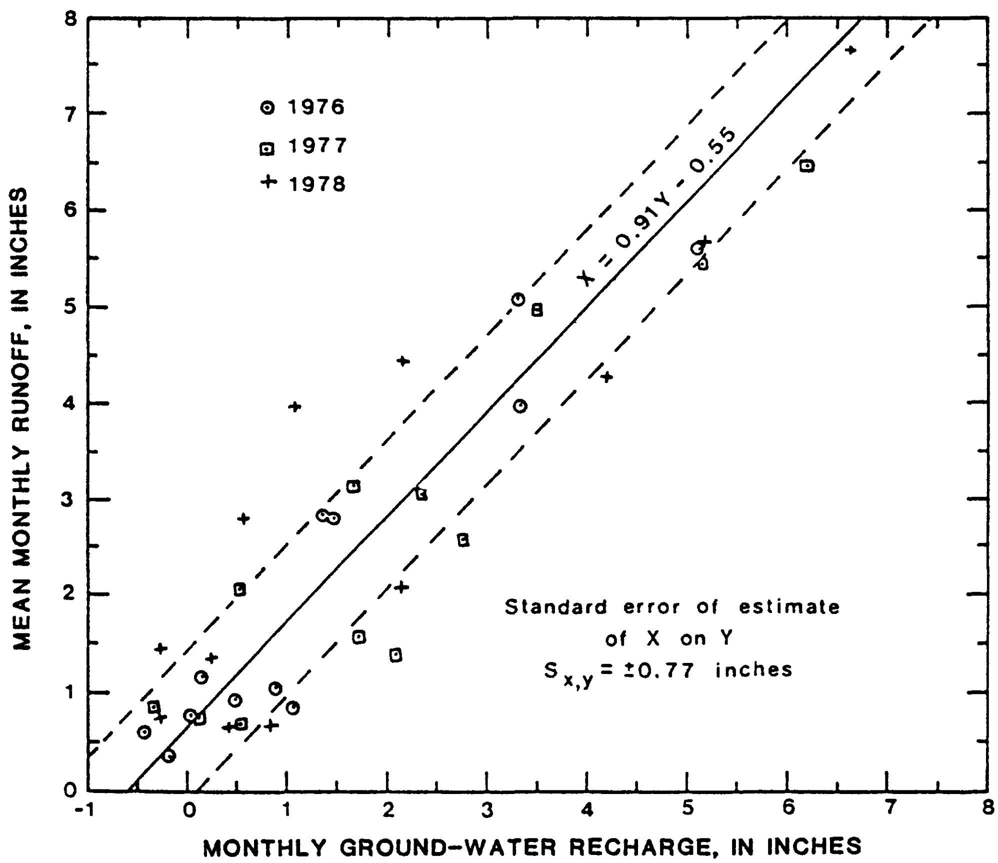


Figure 13.--Correlation between mean monthly runoff and monthly ground-water recharge at the Beaver River upper gage near Usquepaug, R.I.

for modified nonsteady state leaky confined aquifer conditions, (4) Jacob (1946) method for steady state leaky confined aquifer conditions, (5) Cooper and Jacob (1946) method for graphical solution to the modified nonleaky confined formula, and (6) Theis (1935) nonequilibrium formula for a nonleaky confined aquifer. Methods 1 and 2, are described in Lohman (1979, p. 34-38, 31-32), and methods 3-6 are described in Walton (1962, p. 5-6, 9). The assumptions inherent in each analytical method are summarized in table 5.

Examples of solutions for two of the above analytical methods are shown in figures 14 and 15. Figures show flow equations, curve match points, and calculations used to determine aquifer hydraulic properties of transmissivity, horizontal and vertical hydraulic conductivity, and storage coefficient. The Stallman and Cooper methods are shown because they were the primary means of determining vertical hydraulic conductivity.

Estimates of transmissivity also were made from lithologic logs and adjusted specific-capacity data as an additional means of checking hydraulic properties obtained by analysis of aquifer-test data. Table 6 summarizes well construction and aquifer test data, transmissivity estimates, methods of data analysis, and results of analyses.

Aquifer tests were made in the thick, permeable parts of the stratified-drift. In each of the tests, large-diameter (8- to 24-inch) wells were pumped at constant rates that ranged from 345 to 670 gal/min for 20 to 71.5 hours. Depth to the water table at test sites ranged from 3.3 to 21.7 feet below land surface. Pumped wells ranged in depth from 42 to 95 feet, and all had 10 to 16 feet of screen exposed near the bottom of the well. The transmissivity of the stratified-drift aquifer (fig. 16) determined from these tests ranges from 7,200 to 24,300 ft^2/d, and averages 15,300 ft^2/d. Horizontal hydraulic conductivity ranges from 135 to 330 ft/d and averages 200 ft/d; vertical hydraulic conductivity ranges from 0.18 to 48 ft/d, and averages 7.2 ft/d.

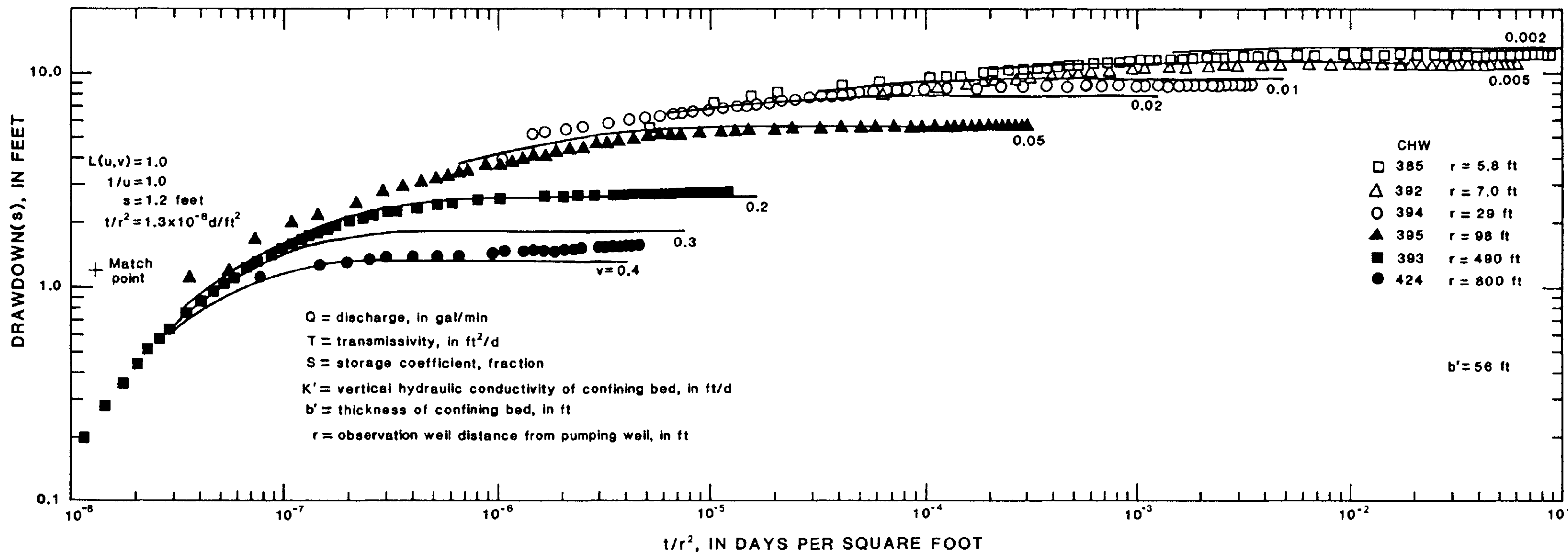
The stratified-drift aquifer is hydraulically anisotropic because the vertical and horizontal hydraulic conductivities differ. The anisotropy is due in part to the interbedding of coarser and finer materials and in part to the orientation of the plate-shaped grains. This causes the hydraulic conductivity of the aquifer to be lower in the vertical direction. The ratio of vertical to horizontal hydraulic conductivity of the stratified-drift aquifer ranges from 1:12 to 1:260, with a median of 1:60.

Within the Beaver River valley, the highest values of hydraulic properties were determined at 8-inch test well RIW (Richmond well) 395. Transmissivity values computed by different analytical methods (see table 6) range from 14,200 to 24,300 ft^2/d, and average 20,900 ft^2/d. Horizontal hydraulic conductivity ranges from 175 to 280 ft/d, and averages 250 ft/d; while vertical hydraulic conductivity ranges from 5.3 to 25.4 ft/d, and averages 15.2 ft/d.

Table 5.--Assumptions on which equations used to analyze aquifer-test data in the Beaver-Pasquiset ground-water reservoir are based (x, condition treated in this report; adapted from Stallman, 1971).

Assumption	Stallman (1963, 1965)	Walton (1960)	Cooper (1963)	Jacob (1946)	Cooper & Jacob (1946)	Theis (1935)
A. Control-well characteristics:						
Full penetration	x	x	x	x	x	x
Partial penetration	x					
Diameter infinitesimal		x	x	x	x	x
Diameter finite	x					
B. Conductivity and flow conditions:						
Homogeneous, isotropic	x	x	x	x	x	x
Homogeneous, anisotropic	x					
Areally infinite	x	x	x	x	x	x
Dewatering negligible	x	x	x	x	x	x
Flow radial		x	x	x	x	x
Flow radial and vertical	x					
Nonsteady flow	x	x	x	x	x	x
Steady flow				x		
Horizontal flow in aquifer	x	x	x	x	x	x
Vertical flow through confining bed		x	x	x		
C. Storage relation:						
Water released from storage instantaneously	x	x	x	x	x	x
Storage in confining bed neglected		x	x	x		
Confined (artesian)			x	x	x	x
Leaky confined		x	x	x		
Unconfined (water table)	x	x				

COOPER (1963) SOLUTION FOR NONSTEADY RADIAL FLOW IN A LEAKY CONFINED AQUIFER AT CONSTANT r



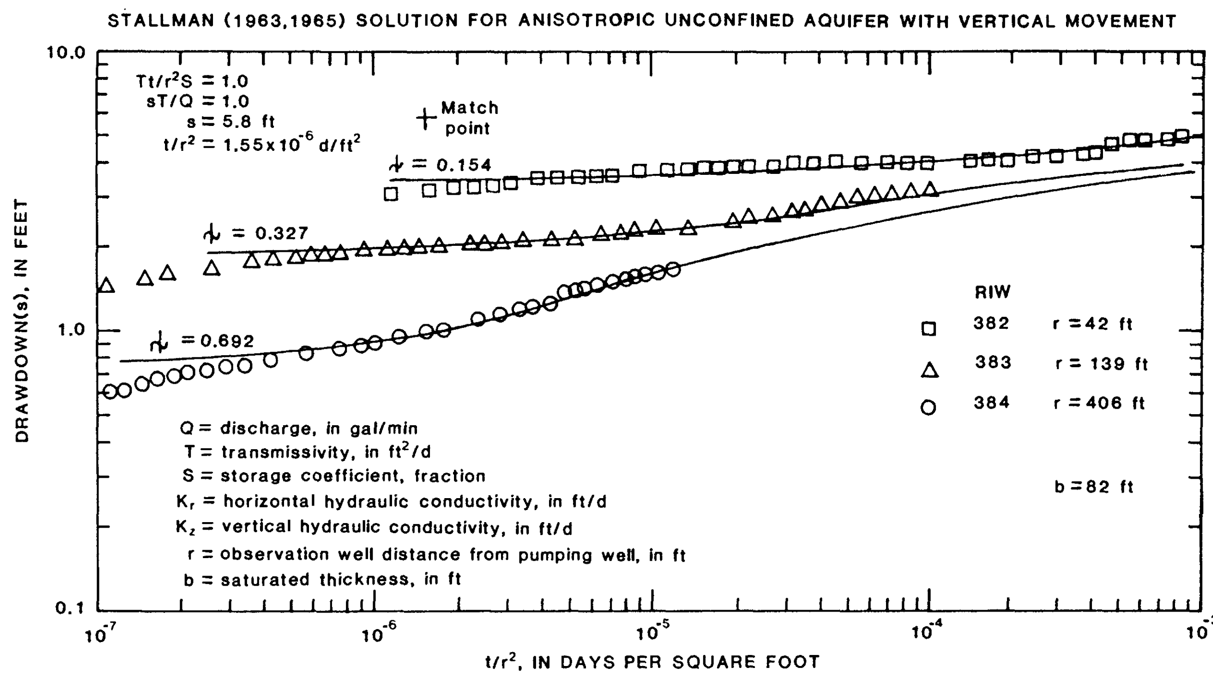
$$T = \frac{QL(u,v)}{4\pi s} = \frac{(560\ \text{gal/min})(1.0)(1440\ \text{min/day})}{4\pi(1.2\ \text{ft})(7.48\ \text{gal/ft}^3)} = 7{,}200\ \text{ft}^2/\text{d}$$

$$S = 4T\frac{t/r^2}{1/u} = (4)(7{,}200\ \text{ft}^2/\text{d})\left(\frac{1.3\times10^{-8}\ \text{d/ft}^2}{1.0}\right) = 0.0004$$

$$\frac{K'}{b'} = 4T\frac{v^2}{r^2} = 0.0034\ \text{to}\ 0.0146$$

$$K' = b'\left(4T\frac{v^2}{r^2}\right) = 0.19\ \text{to}\ 0.82\ \text{ft/d}\ ; \quad \text{average}\ K' = 0.40\ \text{ft/d}$$

Figure 14.--Composite logarithmic plot of s versus t/r^2 in observation wells due to pumping well Charlestown 396 (CHW 396) near Pasquiset Pond.



$$T = 1.0 \frac{Q}{s} = \frac{(1.0)(670 \text{ gal/min})(1440 \text{ min/day})}{(5.8 \text{ ft})(7.48 \text{ gal/ft}^3)} = 22{,}000 \text{ ft}^2/\text{d}$$

$$S = \frac{Tt}{1.0r^2} = (22{,}000 \text{ ft}^2/\text{d})(1.55 \times 10^{-6} \text{d/ft}^2) = 0.03$$

$$\frac{K_z}{K_r} = \left[\frac{(\psi)(b)}{(r)}\right]^2 = 0.0195 \text{ to } 0.0904 \;;\quad K_r = \frac{T}{b} = \frac{22{,}000 \text{ ft}^2/\text{d}}{82 \text{ ft}} = 270 \text{ ft/d}$$

$$K_z = \left(\frac{K_z}{K_r}\right)(270 \text{ ft/d}) = 5.3 \text{ to } 24.3 \text{ ft/d} \;;\ \text{average } K_z = 13.2 \text{ ft/d}$$

Figure 15.--Composite logarithmic plot of s versus t/r^2 in observation wells due to pumping well Richmond 395 (RIW 395) near the Beaver River.

Table 6.--Summary of hydraulic properties determined from aquifer tests
(in: inches; ft: feet; hrs: hours; gal/min: gallons per minute;

Well construction and pump test data for pumping well									Transmissivity estimated from	
Pumping well[1]	Diameter[2] (in)	Screened Interval[3,4] (ft)	Static water level[4] (ft)	Date of aquifer test	Length of test (hrs)	Pumping rate (gal/min)	Draw-down (ft)	Specific capacity (gal/min /ft)	Litho-logic log (ft^2/d)	Adjusted specific capacity[5] (ft^2/d)
TOWN OF CHARLESTOWN										
337	24 x 12	60-70	6.00	12-08-58	20	490	34.00	14	16.200	7,000
349	18 x 12	32-42	3.50	05-18-53	24	345	15.80	22	7,200	11,600
396	8	79-94	8.35	10-28-75	71	555	46.98	12	6,600	8,800
400	8	76-90	8.38	12-08-75	71	610	15.80	39	13,800	--
410	18 x 12	39-55	5.00	07-07-69	48	400	28.33	14	12,100	5,400
TOWN OF RICHMOND										
385	8	70-85	4.35	08-13-75	48	550	40.99	13	11,800	17,300
395	8	80-95	21.68	08-27-75	48	670	14.97	45	8,000	15,600
400	8	70-85	11.77	09-15-75	71	610	18.90	32	8,500	24,300
405	8	58-72	3.30	10-06-75	71.5	560	38.20	15	12,200	11,600
415	8	73-88	18.80	11-18-75	71	585	33.95	17	13,000	15,800

1 Local well number based on the town in which it is located. See figure 2 for location of pumping well.
2 The smaller number or single number is the diameter of the well casing and screen, and the larger number is the diameter of the drilled hole. The space between the drilled hole and screen is filled with a highly permeable material, called the gravel pack.
3 Bottom of screened interval is well depth.
4 Feet below land-surface datum.
5 Drawdown in Charlestown wells 337,396,and 400,and Richmond wells 385,395,400,405,and 415, was adjustedforwell loss,partial penetration,and dewatering. Drawdown in Charlestown well349was adjustedforwellloss and partial penetration,and Charlestown well 410wasadjustedforpartial penetration and dewatering.

for the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir. ft²/d: square feet per day; ft/d: feet per day)

Hydraulic properties determined by analytical methods							
Method[6]	Observation well no.[1,7]	Distance from pumping well (ft)	Transmissivity (ft^2/d)	Hydraulic conductivity: Horizontal[8] (ft/d)	Hydraulic conductivity: Vertical (ft/d)	Storage coefficient[9]	Pumping well[1]
			TOWN OF CHARLESTOWN				
--	--	--	--	250	--	--	337
--	--	--	--	185	--	--	349
c	Composite of 385,392, 394,395,393,424	6, 7, 29,98,490,800	7,200	85	0.19-0.82	0.0004	396
d	do.	do.	8,200	95	.18	--	
e	do.	do.	8,400	100	--	.02	
f	394	29	9,000	105	--	.0001	
f	[10]395	98	9,900	115	--	.003	
f	[11]395	98	9,900	115	--	.20	
f	[10]393	490	9,200	105	--	.0004	
f	[11]393	490	9,200	105	--	.009	
		site average	8,900	105	0.40		
a	403	31	23,500	285	48	.06	400
a	404	102	20,200	245	17	.05	
a	[12]404	102	21,700	265	18	.03	
a	402	190	11,200	135	12	.12	
a	[12]402	190	11,200	135	12	.12	
a	398	650	15,900	195	2.5	.02	
e	Composite of 403, 404,402,398	31, 102,190,650	19,300	235	--	.007	
e	do.	do.	14,800	180	--	.05	
		site average	17,200	210	18		
--	--	--	--	240	--	--	410
			TOWN OF RICHMOND				
b	372	149	11,100	135	3.2	.0004	385
c	Composite of 374, 372,376,355	101, 149,287,548	15,000	190	1.0-3.2	.0002	
e	do.; 357	do.; 1200	14,700	185	--	.05	
f	[12]374	101	20,000	250	--	.009	
f	[10]372	149	11,100	140	--	.0004	
f	[11]372	149	12,000	150	--	.005	
f	376	287	17,400	220	--	.002	
		site average	14,500	180	1.98		
a	Composite of 382, 383, 384	42 139, 406	22,000	270	5.3-24.3	.03	395
a	382	42	23,000	280	25.4	.03	
a	383	139	24,300	280	11.0	.02	
e	Composite of 382, 386,387,383,384	42, 47,96,139,406	14,200	175	--	.04	
		site average	20,900	250	15.2		
a	Composite of 396, 398,399,394,366	30, 113,150,415,560	13,700	170	--	.03	400
a	396	30	17,300	235	7.5	.05	
a	398	113	16,800	230	2.2	.01	
a	399	150	10,000	135	3.5	.03	
a	394	415	11,000	150	2.2	.01	
a	366	560	18,100	260	1.9	.01	
e	365	5	15,400	190	--	.0001	
e	396	30	15,400	190	--	.0005	
e	398	113	22,700	285	--	.000006	
		site average	15,600	205	3.5		
a	402	47	14,200	205	11.0	.04	405
a	404	159	15,400	220	4.6	.03	
e	Composite of 402, 404,407,409	47, 159,390,1185	14,000	200	--	.05	
		site average	14,500	210	7.8		
a	Composite of 410, 411, 396, 398	72, 101,590,657	18,800	260	1.5-6.4	.02	415
e	Composite of 410, 411, 412, 357	72, 101,195,2350	12,400	170	--	.01	
		site average	15,600	215	3.2		

6 (a) Vertical movement (Stallman, 1963, 1965) described in Lohman (1972) p. 34-38; (b) modified nonsteady state leaky confined (Walton, 1960), described in Walton (1962) p. 5; (c) nonsteady radial flow leaky confined (Cooper, 1963), described in Lohman (1972) p. 31-32; (d) steady state leaky confined (Jacob, 1946), described in Walton (1962), p. 5-6; (e) modified nonleaky confined (Cooper and Jacob, 1946), described in Walton (1962), p. 9; and (f) non-equilibrium formula (Theis, 1935), described in Walton 1962), p. 6.

7 Well or wells used in analysis.

8 Determined by dividing transmissivity by distance from static water level to bottom of screen in pumped well.

9 Most values smaller than 0.12 were determined from early drawdown dataand are not indicative ofthetrue storagecoefficientoftheaquifer. Values greater than 0.12 are believed toapproachthetruestorage coefficient (specific yield) of the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir.

10 Hydraulic properties determined by analysis of early time-drawdown data.

11 Hydraulic properties determined by analysis of late time-drawdown data.

12 Hydraulic properties determined by analysis of recovery data.

page 35 follows

The lowest values of hydraulic properties in the Beaver River valley were determined at 8-inch test well RIW 385 where lithologic logs revealed a discontinuous 34 foot layer of semi-confining material composed of very fine sand, silt and clay. Transmissivity ranges from 11,100 to 20,000 ft^2/d, and averages 14,500 ft^2/d whereas horizontal hydraulic conductivity ranges from 135 to 250 ft/d, and averages 180 ft/d. Vertical hydraulic conductivityof the semi-confining material ranges from 1.0 to 3.2 ft/d, and averages 1.98 ft/d.

Within the Pasquiset Brook valley, the highest values of hydraulic properties were determined at 8-inch test well CHW (Charlestown well) 400, where values of transmissivity computed by different analytical methods (see table 6) range from 11,200 to 23,500 ft^2/d, and average 17,200 ft^2/d. Horizontal hydraulic conductivity ranges from 135 to 285 ft/d, and averages 210 ft/d whereas vertical hydraulic conductivity ranges from 2.5 to 48 ft/d, and averages 18 ft/d.

The lowest values of hydraulic properties in the Pasquiset Brook valley were determined at 8-inch test well CHW 396 where lithologic logs show a 56 foot layer of semiconfining material composed primarily of silt and clay. Transmissivity ranges from 7,200 to 9,900 ft^2/d, and averages 8,900 ft^2/d whereas horizontal hydraulic conductivity ranges from 85 to 115 ft/d, and averages 105 ft/d. Vertical hydraulic conductivity of the semi-confining material ranges from 0.19 to 0.82 ft/d, and averages 0.40 ft/d.

Storage coefficients were also determined by analysis of aquifer test data and results are shown in table 6. The storage coefficient in an unconfined water-table aquifer is virtually equal to the specific yield (Lohman and others, 1972, p. 13). In water-table aquifers, the storage coefficient or specific yield may range from about 0.05 to 0.30 (Ferris and others, 1962, p. 78). Areally extensive confining layers were not found at any of the sites tested in the Beaver-Pasquiset ground-water reservoir area.

Most storage coefficient values shown in table 6 smaller than 0.12 were determined from early drawdown data and are not indicative of the true storage coefficient of the stratified-drift aquifer. Williams and Lohman (1949, p. 213, 220) state that the true value of specific yield is obtained only after the saturated material has been drained for a long time. They conclude that, even for sand-size materials, 2 months to more than 1 year would be required for the drainage to reach equilibrium and, thus, give the maximum specific yield.

Storage coefficients shown in table 6 are affected by delayed yield and values more reflect the duration period of the test than the true specific yield of the aquifer. To obtain direct measurements of the maximum specific yield would probably require a longer period of observation than the relatively short time period of most aquifer tests. Therefore, based on laboratory analysis of sediment samples in the adjacent upper Pawcatuck River basin (Allen and others, 1963) in materials similar to this study, it was assumed that the average specific yield of the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir is about 0.20.

Site averages for hydraulic properties shown in table 6 are probably higher than those for much of the stratified drift in the Beaver-Pasquiset study area. This is because aquifer-test sites were located in geologically promising areas selected after extensive 2-1/2-inch exploratory test drilling. Therefore, the test results and reported well yields in the report are indicative of what may be expected from properly constructed wells that tap the stratified drift in the more productive parts of the Beaver-Pasquiset ground-water reservoir.

Stream-Aquifer Interconnection

Under natural conditions, the water-table gradient slopes toward the river (fig. 2), and ground-water discharges from the stratified-drift aquifer into the river. However, under pumping conditions the water-table gradient decreases and ground-water runoff to the stream is reduced. If pumping is of sufficient volume and duration, the gradient may be reversed and water from the stream will move by induced infiltration through the streambed into the stratified-drift aquifer. The areas most favorable for development of high-capacity wells (350 gal/min or more) are along the Beaver River and near Pasquiset Pond. Most of the water withdrawn from wells in the stratified-drift aquifer will be derived from induced recharge from the Beaver River and Pasquiset Pond.

The amount of induced recharge diverted to wells is governed by the vertical hydraulic conductivity of the streambed and underlying aquifer, streambed thickness, area of streambed through which infiltration occurs, viscosity of the water (which is temperature dependent), average head difference between the stream and aquifer within the streambed area of infiltration, and quantity of water in the stream.

The beds of the Beaver River and Pasquiset Brook are generally composed of loosely packed sand and gravel. The streambeds are assumed to have a higher vertical hydraulic conductivities than that of the underlying stratified-drift aquifer which typically contains layers of fine silt or silty sand. The average vertical hydraulic conductivity of the underlying aquifer was assumed to be the "effective streambed" hydraulic conductivity, as it is called in this report, which now becomes a controlling factor in the infiltration of streamflow to pumping wells. The vertical hydraulic conductivity of the stratified-drift aquifer was determined from data collected during controlled aquifer tests at five sites along the Beaver River and at one site along Pasquiset Brook. Average values of vertical hydraulic conductivity range from 1.98 to 15.2 ft/d along the Beaver River to 0.40 ft/d along Pasquiset Brook near the north end of Pasquiset Pond. The width of the streambeds of the Beaver River and Pasquiset Brook average about 15 feet.

If stream reaches are to maintain some flow at all times, then the quantity of streamflow during low-flow periods limits the amount of water available for induced recharge to the stratified-drift aquifer. For this study the flow equaled or exceeded 98-percent of the time was considered the index of streamflow available for induced recharge to the aquifer. The 98-percent flow duration at the mouth of the Beaver River is estimated to be 2.0 ft^3/s.

Water-Bearing Characteristics of Bedrock and Till

Each of the geologic units are generally capable of yielding usable quantities of water to wells and, therefore, constitute aquifers. Crystalline or metasedimentary bedrock underlies the entire basin. The bedrock aquifer should yield at least a small quantity of water to a well at almost any locality. The Hope Valley Alaskite Gneiss (crystalline) and the Blackstone Group (metasedimentary), a quartz-feldspar-biotite gneiss, all of late Precambrian age, are the predominant bedrock units in the basin. Water-bearing fractures in the bedrock decrease in size and frequency with depth and become sparse below 300 feet. Reported yields of wells in these bedrock units range from 2 to 30 gal/min, with a median yield of 6 gal/min (Dickerman and Johnston, 1977, table 1). The higher yielding wells were drilled in the Blackstone Group, or near its geologic contact with the Hope Valley Alaskite Gneiss. Data from bedrock wells reported by Dickerman and Johnston (1977) show that well depths ranged from 65 to 315 feet, with a median depth of 125 feet.

During the Pleistocene Epoch continental glaciers advanced from the north and covered this area several times. These glacial ice sheets deposited rock debris, called "drift", which includes till, stratified drift, and scattered rock fragments. In the Beaver-Pasquiset basin, most of the drift was deposited during the advance and retreat of the last ice sheet during the Wisconsin glacial age. Drift is subdivided into two distinct kinds: till or nonstratified drift, and stratified drift. However, there is no sharp dividing line between till and stratified drift, and one grades into the other.

Till, locally called "hardpan", forms a generally thin discontinuous mantle over the bedrock and usually reflects the topography of the underlying bedrock. It is not sorted or stratified by water action and consists of a mixture of material ranging in size from boulders to clay. Till covers about 53-percent of the study area, and has an average thickness of 25 feet. It reaches a maximum known thickness of 80 feet, along the east side of the Beaver River valley, approximately 3,000 feet north of Kenyon.

Although till is usually a poor water-bearing material, it does constitute an aquifer capable of yielding small, but in places unreliable, supplies for domestic and agricultural use. Generally, till will not yield more than 5 gal/min to large-diameter wells (Bierschenk and Hahn , 1959; LaSala and Hahn, 1960). Wells in till typically go dry during drought periods and may go dry annually during late summer or early fall.

Mixed deposits of till and stratified drift constitute the Charlestown moraine, which forms the southern boundary of the Beaver-Pasquiset study area. The moraine is characterized by its elongated hummocky belt of morainic ridges and hills, and numerous dry kettle holes. East of the study area, many of these depressions contain perennnial ponds. Although these mixed deposits cover only three-percent of the basin, they are, nevertheless, a source of water for homes within the Charlestown moraine. Wells that penetrate and are screened or open ended in the stratified sand and gravel layers, should be capable of yielding small to moderate (5 to 40 gal/min) supplies for domestic use.

QUALITY OF GROUND WATER AND SURFACE WATER

The chemical quality of water in the Beaver-Pasquiset ground-water reservoir is suitable for most purposes. The water is soft and generally contains less than 100 mg/L dissolved solids. The principal cations of calcium, magnesium, sodium, and potassium are generally present in concentrations less than 8 mg/L. The principal anions of bicarbonate, sulfate, and chloride are generally present in concentrations less than 21 mg/L. The hydrogen ion concentration, or pH, ranges from 5.0 to 7.1. Water in this pH range is somewhat corrosive.

Iron and manganese generally occur in concentrations less than the recommended U.S. Environmental Protection Agency (USEPA) maximum limit (table 7) for public water systems. However, southeast of Pasquiset Pond, concentrations of iron and manganese of 7.5 mg/L and 3.7 mg/L have been measured in 8-inch test well CHW 400. Ground-water in this area will require treatment for iron and manganese removal if used for a public drinking supply.

All wells sampled for nitrate (NO_3 as N) met the recommended USEPA maximum limit of 10 mg/L for public water systems. However, two wells, RIW 363 and 364, had slightly elevated nitrate levels of 9.6 and 8.4 mg/L respectively. These wells are located along the East side of the Beaver River (fig. 2) downgradient from adjacent agricultural fields. Fertilizers applied to these fields may be, in part, responsible for nitrate values in these wells being above the median value of 0.20 mg/L.

Analyses of water samples collected from streams at 13 sites in the study area (Dickerman and Johnston, table 16, 1977) during periods of low flow, when streamflow was composed primarily of ground-water runoff, were similar to analyses of ground water from wells in the stratified-drift aquifer.

Continuous records of specific conductance were collected at the U.S. Geological Survey lower gage on the Beaver River near Kenyon, R.I. during water years 1976-78. From October 1975 through September 1978, specific conductance ranged from 24 to 111 microsiemens/cm at 25°C, and averaged 72 microsiemens/cm at 25°C. The low mean specific conductance of 72 microsiemens/cm at 25°C is indicative of the overall excellent quality of streamflow in the Beaver River.

Table 7 summarizes the chemical and physical properties of ground water and surface water in the Beaver-Pasquiset study area.

Table 7.--Summary of chemical and physical properties of ground water and surface water in the Beaver-Pasquiset study area.

[Concentration, in milligrams per liter, except as noted.]

Constituent or property	Maximum limit for drinking water	Ground water				Surface water			
		Samples	Low	Median	High	Samples	Low	Median	High
Silica (SiO_2)	--	5	9.0	15	17	12	4.0	9.3	13
Iron (Fe)	[1] 0.3	62	0.00	0.08	11	17	0.08	0.20	1.00
Manganese (Mn)	[1] .05	58	.00	.02	3.70	13	.01	.02	.09
Calcium (Ca)	--	12	3.2	7.8	12	12	2.6	4.6	9.3
Magnesium (Mg)	--	12	.9	2.3	4.9	12	.9	1.5	2.0
Sodium (Na)	--	10	4.1	5.6	12	10	4.0	5.6	11
Potassium (K)	--	10	.7	1.3	2.0	10	.8	1.0	1.5
Bicarbonate (HCO_3)	--	6	7	21	38	12	7	11	18
Sulfate (SO_4)	[1] 250	12	1.0	12	16	17	2.0	6.1	14
Chloride (Cl)	[1] 250	53	4.0	10	41	17	4.4	9.0	18
Fluoride (F)	[2] 2.0	15	.0	.2	.9	8	.2	.2	.3
Nitrate (NO_3) as N	[2,3] 10	48	.00	.20	9.60	17	.00	.20	.79
Dissolved solids (residue at 105°C)	[1] 500	9	55	76	125	--	--	--	--
Dissolved solids (residue at 180°C)	[1] 500	--	--	--	--	10	36	54	70
Dissolved oxygen	--	55	.3	6.8	10.3	10	5.2	9.2	10.2
Hardness as ($CaCO_3$)	--	32	14	24	94	12	11	18	31
Alkalinity as ($CaCO_3$)	--	26	3	12	31	5	13	14	15
Specific conductance (microsiemens/cm at 25°C)	--	69	43	87	310	19	48	76	150
pH (Units)	[1] 6.5-8.5	25	5.0	6.1	7.0	18	5.2	6.4	7.1
Color (Platinum cobalt units)	[1] 15	12	0	0	5	12	10	31	110

1 Secondary maximum contaminant level established for public water systems by the U.S. Environmental Protection Agency (1979).
2 Maximum contaminant level for inorganic chemicals established for public water systems by the Rhode Island Department of Health, Division of Water Supply (1977).
3 Maximum contaminant level for inorganic chemicals established for public water systems by the U.S. Environmental Protection Agency (1975).

SIMULATION OF GROUND-WATER-DEVELOPMENT ALTERNATIVES

Conceptual Model

A conceptual model of the stream-aquifer system in the Beaver-Pasquiset ground-water reservoir was developed from geohydrologic data contained in reports by Allen and others (1963), and Dickerman and Johnston (1977). The Beaver-Pasquiset model includes simplifying assumptions which form the basis of a conceptual model of the flow system. It is through this conceptualization and simplification process which makes it possible to simulate the system mathematically. These basic assumptions are as follows:

(1) Ground-water flow in the stratified-drift aquifer is horizontal, and the aquifer is isotropic.
(2) Recharge to the aquifer from precipitation is uniformly distributed and varies monthly.
(3) Flow rates at the southern drainage divide, and ground-water inflow and outflow remain constant with time.
(4) All pumping wells are considered to be screened the full saturated thickness of the aquifer and are 100-percent efficient. To compensate for these idealized well-construction characteristics, maximum allowable drawdown under pumping conditions is limited to 25-percent of the initial saturated thickness.
(5) Dewatering of the unconfined aquifer and subsequent decline in transmissivity is negligible.
(6) Ground-water evapotranspiration decreases linearly with depth of water table from a maximum at land surface to zero at 5 feet or more below land surface.
(7) The rate of head loss at a stream node in a given section in the Beaver River and Pasquiset Brook is defined by the Manning formula (Ozbilgin and Dickerman, 1980, p. 3).

Although these basic assumptions do not always represent actual field conditions of the stream-pond-aquifer system, the authors believe that any deviations from them probably do not introduce large errors in conceptualization of the system nor in the model simulations.

The part of the Beaver-Pasquiset study area selected for simulation is outlined in figure 17. The stratified drift is essentially one aquifer layer underlain by low permeability till or bedrock (assumed to be impermeable) and is simulated using a two-dimensional model instead of a three-dimensional model. The northern half of the Beaver River drainage basin was not included in the model because it is composed of till and thinly saturated stratified drift with little potential for development of large quantities (greater than 350 gal/min) of water from wells. The area selected for the model covers 10.2 mi^2, of which 5.3 mi^2 is stratified drift, 4.1 mi^2 is till, and 0.8 mi^2 is mixed till and stratified drift.

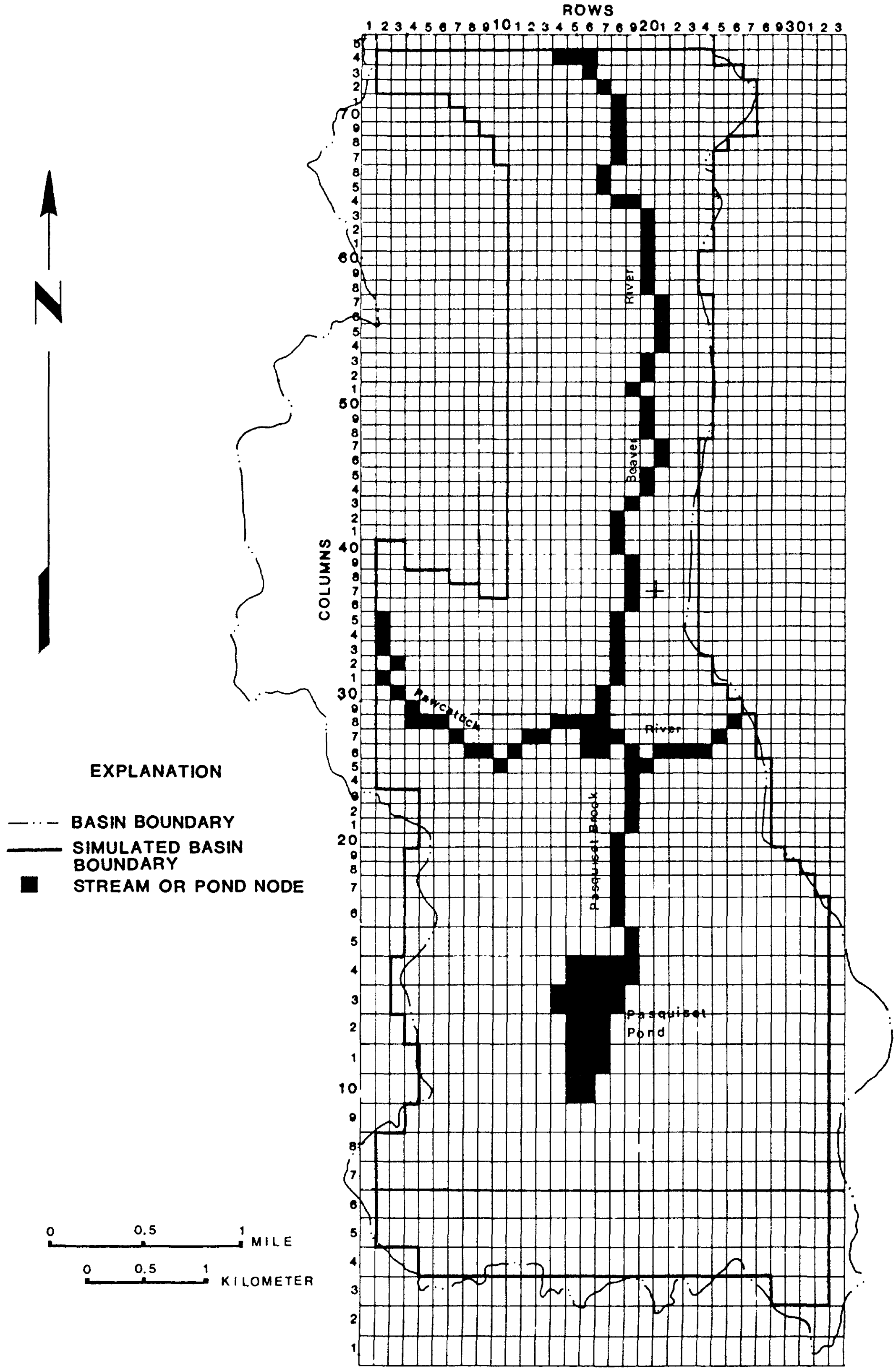


Figure 17.--Finite-difference grid of the Beaver-Pasquiset ground-water model.

Digital Model

Digital models that simulate ground-water flow are widely used in the management of ground-water resources in order to assess the impact of withdrawals on streamflow and ground-water levels. A two-dimensional model developed by Trescott and others (1976) was modified to enable more detailed simulation of the interaction between surface water and ground water. For detailed information on theoretical development of the modified model, data deck instructions, and complete program listing, the reader is referred to U.S. Geological Survey Water Resources Investigations Report 83-4251 by Melih M. Ozbilgin and David C. Dickerman (1984).

A digital model of a stream-aquifer system is an approximation of the simplified mathematical representation of a complex geohydrologic system. The digital model is a computer program designed to solve equations that govern ground-water flow and can be used to evaluate the effects of different stresses imposed on the stream-aquifer system. Flexibility, speed, and accuracy are three of the model's greatest assets.

The model, developed by Trescott and others (1976) uses a finite-difference method to numerically approximate differential equations that describe the two-dimensional flow of ground water. Solution of these equations requires subdivision of the modeled area into rectangular blocks (fig. 17). The center of each block in the grid is called a node. Active nodes are represented in illustrations by a dot or plus sign (fig. 18). Dots represent nodes in stratified drift, and plus signs represent nodes in till. The grid network for the model consists of 33 rows and 75 columns, and defines 2,475 nodes, of which 1,008 nodes are outside the model boundaries (nonactive nodes). The simulated blocks are 400 feet on a side, except in the area near Pasquiset Pond where they are 400 feet by 800 feet.

A finite-difference equation that approximates flow in the block is evaluated at each node of the grid, and the set of equations is solved simultaneously. The solution technique used in the model is the strongly implicit procedure (SIP) developed by Stone (1968).

The flow equations require that the hydraulic properties of the aquifer and other hydrologic parameters be defined for the entire model area. The grid network is superimposed over appropriate maps, and average parameter values are assigned to respective nodes. In the Beaver-Pasquiset model, values were assigned to each node for transmissivity, confining bed hydraulic conductivity, river head, river node identification, recharge, and land surface elevation.

Average transmissivities used in the model range from 240 to 16,500 ft^2/d in stratified drift areas (fig. 16), and 400 to 1,100 ft^2/d in mixed stratified drift and till areas. A uniform transmissivity of 60 ft^2/d was used in all till areas.

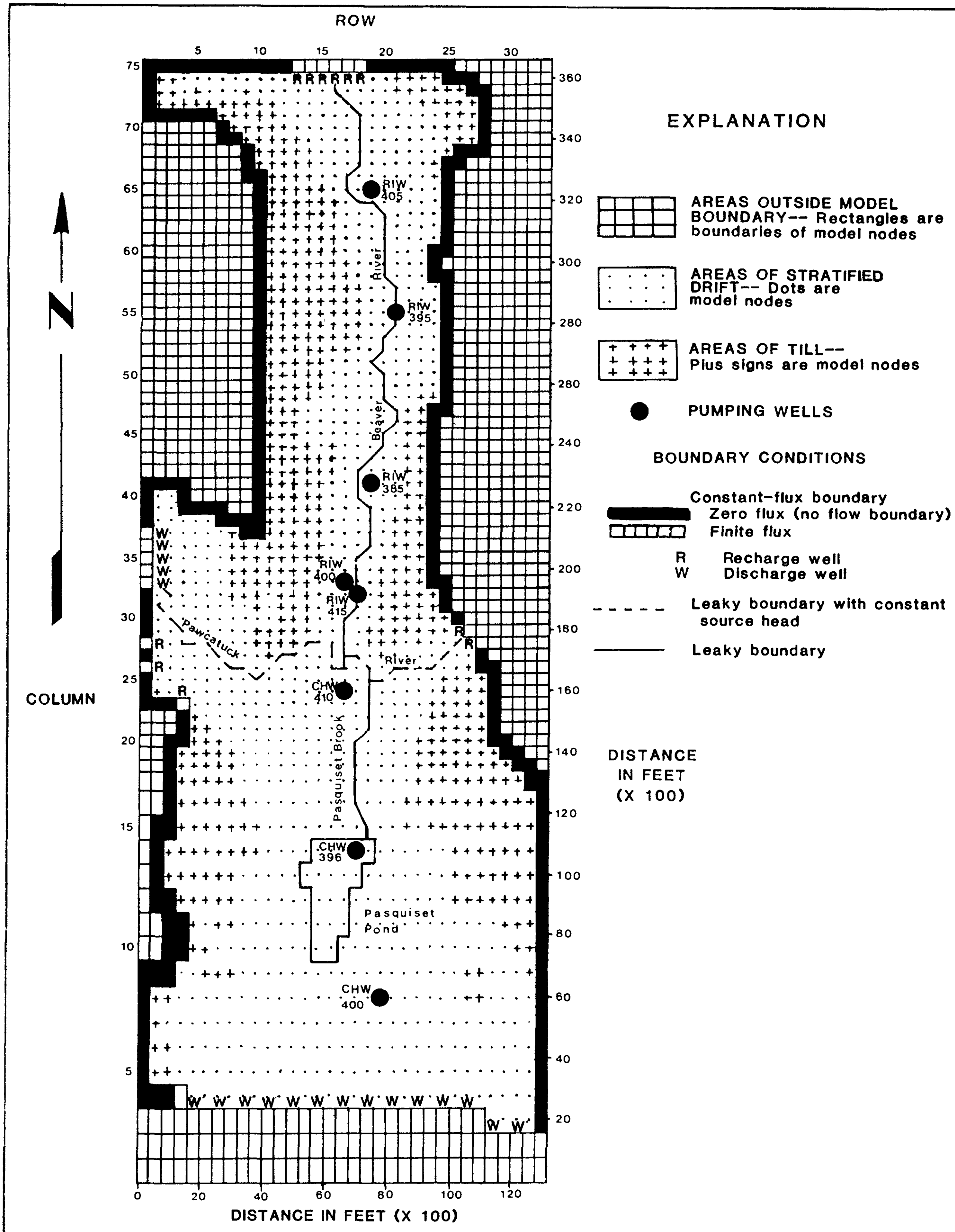


Figure 18.--Beaver-Pasquiset ground-water model boundary conditions and location of pumping wells.

Surface water bodies simulated in the model are identified in figure 17 as stream or pond nodes. Because aquifer nodes in the model are wider than natural streams, leakage to or from stream nodes was reduced to compensate for actual stream width. No adjustments were necessary for Pasquiset Pond because the area of a pond node equals the area of the underlying aquifer node. Flow between surface water bodies and the stratified-drift aquifer is calculated in the model using "effective streambed" hydraulic conductivity values.

Boundary Conditions

An important role of the model is to represent conditions at its boundaries accurately. Within the gridded area shown in figure 17, most model boundaries (simulated drainage divide) were selected to coincide as closely as possible with hydrologic boundaries (drainage divide). Model boundaries that do not coincide with hydrologic boundaries were located far enough from hypothetical wells to have negligible effect on drawdown during stress periods. Boundaries in the Beaver-Pasquiset ground-water model were treated as constant-flux, leaky constant source head, or leaky.

A constant flux may be zero or have a finite value. Under natural conditions, a drainage divide is a no-flow boundary, and water from one side of the divide does not move across the boundary to the other side. In the model a zero-flux boundary is used to simulate no flow across a boundary and is treated by assigning a value of zero transmissivity to nodes outside the boundary. However, where flow is known to move across a boundary, such as ground-water inflow or outflow, a finite-flux boundary is used. This type of boundary is treated by assigning a fixed value of volumetric flow to recharge or discharge wells at appropriate nodes.

Streams and ponds were modeled as leaky constant source head or leaky boundaries. Constant source head boundaries are given a fixed value of static head, while leaky boundaries relate boundary flux to boundary head.

Boundary conditions in the Beaver-Pasquiset ground-water model and location of hypothetical pumping wells are shown in figure 18.

The eastern boundary was treated as a zero-flux (no flow) boundary at the drainage divide, except where the Pawcatuck River enters the study area. Here ground-water inflow to the model area was simulated with a finite-flux boundary using two recharge wells near the Pawcatuck River.

The western boundary was treated as a no-flow boundary at the basin drainage divide south of the Pawcatuck River and at a subbasin drainage divide north of the Pawcatuck River, except where the river leaves the study area and at three nodes adjacent to a till area. Ground-water outflow from the model area was simulated with a finite-flux boundary using five discharge wells near the Pawcatuck River. A finite-flux boundary was also used to simulate ground-water inflow from a small till area outside the model boundary using three recharge wells.

The northern boundary was treated primarily as a no-flow boundary, except at six nodes near the center of the boundary where a finite-flux boundary was used. Six recharge wells were placed at these six nodes to simulate ground-water inflow from that part of the Beaver River drainage area not included in the model.

In the southern part of the model, the water-table map (fig. 2) shows a ground-water divide approximately 3,000 feet north of the surface-water divide. This ground-water divide could have been treated as a no-flow boundary. However, this would have resulted in erroneous model simulations because of the effect nearby pumping (hypothetical well CHW 400) would have on the position of the ground-water divide. In order to minimize the effect during pumping stress periods, the southern boundary of the model was located along the surface-water divide in the Charlestown moraine. To simulate ground-water outflow southward through the moraine, a specified finite-flux boundary consisting of 14 discharge wells was used. Pumping rates at these 14 wells were varied during steady-state calibration in order to simulate the position of the ground-water divide shown in figure 2.

The top of the stratified-drift aquifer was simulated as the water table, and the bottom of the aquifer was assumed to be a no-flow boundary. Transmissivities assigned to till and bedrock nodes are higher than that of till alone to account for any bedrock flow contributions to the stratified drift. There are no known regional flow contributions from the bedrock aquifer, and any bedrock contributions to the stratified drift are probably derived from precipitation that has infiltrated the till and bedrock and discharges to the stratified-drift aquifer as ground-water runoff.

The Beaver River, Pasquiset Brook, and Pasquiset Pond were treated as leaky boundaries to represent the interaction of the stream-pond-aquifer system. The Pawcatuck River, however, was modeled as a constant source head boundary since water available for induced recharge based on minimum daily streamflow (9.69 Mgal/d), substantially exceeds nearby ground-water withdrawals (0.25 Mgal/d).

Calibration

Steady State

Before a model can be used reliably to simulate the effects of future imposed stresses, it should be capable of duplicating the response of the stream-pond-aquifer system to known historical stresses within acceptable limits. Long-term records, however, were not available for the Beaver-Pasquiset ground-water reservoir area and it was necessary to utilize estimated long-term average annual ground-water recharge to the aquifer as a stress. The acceptability of the steady-state model was determined by comparing measured water levels in the stratified-drift aquifer and estimated ground-water runoff between the upper and lower gaging stations along the Beaver River during August-November 1975 with those values predicted by the model. Average water levels measured during August-November 1975, in long-term observation wells in the Pawcatuck River basin, indicate that the fall of 1975 represents long-term average annual conditions in the aquifer. Differences between predicted and measured water levels were considered acceptable if they were within two feet of each other in most places. However, discrepancies of up to 10 feet were considered acceptable along the stratified drift/till contact where data are sparse.

An areally uniform recharge rate was used for all steady-state and transient simulations because it was not possible to distinguish between recharge rates for till and stratified drift from total monthly recharge rates. During steady state simulations a recharge rate of 16.5 inches (table 2) was distributed evenly over the entire model. This rate of recharge represents the long-term (1941-80) average (table 2). Use of a uniform recharge rate of 16.5 inches, however, caused water levels around Pasquiset Pond to be unacceptably high in till areas and low in stratified drift areas. The rate of ground-water flow across the till-stratified drift contact had to be adjusted to approximate fall 1975 water levels. Adjustments were made by varying the transmissivity at nodes where control data was sparse, along the till-stratified drift contact, until water levels were considered to be within acceptable limits.

In order for the model to accurately simulate the position of a ground-water divide south of Pasquiset Pond under long-term average conditions, it was necessary to estimate ground-water outflow through the Charlestown moraine. Since there is no surface runoff from the moraine in the study area, ground-water outflow is assumed equal to total runoff. Using long-term average annual runoff of 26.2 inches (table 2), ground-water outflow southward through the Charlestown moraine was estimated to be 2.1 Mgal/d. Several steady state runs were made varying the distribution of ground-water outflow across 14 hypothetical wells simulating the southern boundary of the model until the position and water table altitude of the ground-water divide duplicated that shown in figure 7.

Initially, ground-water evapotranspiration losses were not taken into account in the model. This resulted in simulated heads being too high where the water table was close to land surface. Also, the difference in estimated and simulated ground-water runoff between the upper and lower gaging stations along the Beaver River was too high. These two problems indicated that the model had to be recalibrated incorporating ground-water evapotranspiration.

The steady-state calibration was repeated incorporating ground-water evapotranspiration at all nodes where the water table was within 5 feet of land surface. Ground-water evapotranspiration was not measured directly for this study. However, long-term potential evapotranspiration was calculated from the Thornthwaite (Thornthwaite and Mather, 1957) equation using the 40-year period from 1941-1980. The resulting long-term potential evapotranspiration rate of 24.6 inches produced acceptable results for both water levels and ground-water runoff. The effective ground-water evapotranspiration rate for the stratified drift in the model is 5.36 inches annually, and was determined by dividing the total volume of evapotranspiration removed by the model by the total area of stratified drift in the model.

Figure 19 shows the comparison between the estimated water table and the simulated water table for average conditions during the period August-November 1975 for the final steady state model run. Measured and model-simulated water-table altitudes for the final steady-state run are given in table 9 for 21 selected observation wells in the modeled area. Differences between measured and simulated water-table altitudes ranged from +2.6 to -2.5 feet, and average +0.07 feet. During the same period, average ground-water runoff between the upper and lower gaging stations along the Beaver River was estimated to be 2.29 Mgal/d, while the model predicted 2.28 Mgal/d. Table 8 shows the simulated steady-state water budget for the aquifer for the Fall of 1975. The model was considered calibrated, and heads from the final steady-state run were used as the starting condition for transient-model simulations.

Table 8.--Simulated steady-state water budget for the two-dimensional model for the fall of 1975.

Sources	(cubic feet per second)	Discharges	(cubic feet per second)
Recharge	13.16	Ground-water runoff	8.60
Ground-water inflow	.32	Evapotranspiration	2.74
		Ground-water outflow	2.14
Total sources	13.48	Total discharge	13.48

Transient

Transient-model analysis was utilized to determine impacts of alternative schemes of ground-water development on ground-water levels, pond levels, and streamflow. For transient simulations in the Beaver-Pasquiset ground-water reservoir, two climatic periods (see fig. 6) were chosen; (1) one representative of wet conditions, 1976-78, and (2) one representative of drought conditions, 1963-66. The acceptability of the transient model was determined by comparing (1) measured and predicted average monthly water levels for nonpumping simulations, and (2) estimated ground-water runoff, between the upper and lower gaging stations, to the Beaver River for the 1976-78 period with predicted ground-water runoff in the model during the calibration phase.

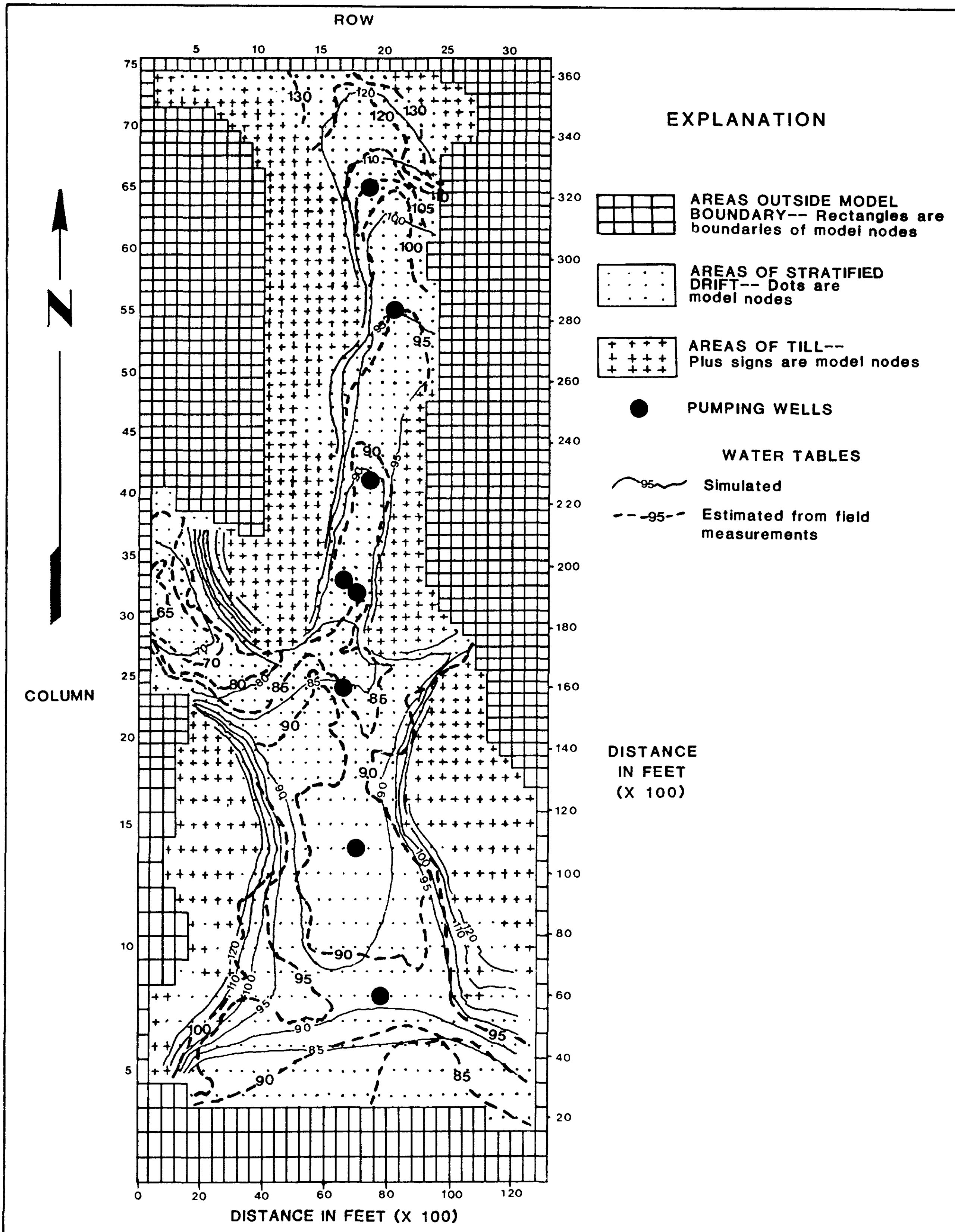


Figure 19.--Comparison of water table estimated from field data and simulated steady-state water table, August-November 1975.

Table 9.--Measured and model-simulated water table altitudes, August-November, 1975, for the final steady state run for selected observation wells in the modeled area.

Well location[1] Node (Column)	(Row)	Well number	Field-measured water-table altitude (feet above sea level)	Model-simulated water-table altitude (feet above sea level)	Water-table altitude difference (feet)
7	22	CHW 390	90.0	89.6	-0.4
8	20	CHW 400	91.7	90.3	-1.4
9	19	CHW 398	91.7	90.3	-1.4
10	22	CHW 401	90.0	90.9	+0.9
14	18	CHW 396	88.5	88.0	-0.5
16	14	CHW 384	89.0	88.9	-0.1
17	14	CHW 383	91.0	88.5	-2.5
24	17	CHW 410	84.0	85.2	+1.2
28	16	RIW 416	83.9	84.0	+0.1
32	18	RIW 415	85.8	85.8	0.0
33	17	RIW 400	86.9	86.6	-0.3
36	18	RIW 393	86.5	87.5	+1.0
37	19	RIW 547	90.0	87.7	-2.3
38	19	RIW 357	87.9	88.3	+0.4
41	19	RIW 385	88.8	89.7	+0.9
47	21	RIW 371	90.7	92.0	+1.3
55	21	RIW 395	95.0	95.0	0.0
56	22	RIW 370	96.4	95.8	-0.6
65	19	RIW 405	104.3	105.3	+1.2
66	19	RIW 407	106.1	107.5	+1.4
68	18	RIW 417	109.0	111.6	+2.6
				Average difference	+0.07

1 See figure 2 for well location.

1976-78 Wet Period

In the transient model, computed hydraulic head depends on starting conditions and length of the simulation time. Therefore, storage properties of the aquifer must be specified in the model. All aquifer parameters used in the steady-state model remained unchanged during transient simulations, and starting heads were taken from the final steady-state run. In an unconfined aquifer the storage coefficient is virtually equal to the specific yield. Specific yields for the model were initially taken as 0.2 for stratified drift and 0.1 for till.

Heads in the aquifer should be in a state of dynamic equilibrium (steady state) to start transient analysis (Rushton and Wedderburn, 1973). The Beaver-Pasquiset transient model was adjusted for 1976 conditions before simulating the 1976-1978 period. Inflow from the Beaver River along the upstream model boundary for 1976, the 1976 recharge rate (16.56 inches), and the 40-year average evapotranspiration rate (24.58 inches) were input to the model on a monthly basis. Using the computed August-November 1975 heads from the final steady-state model run, the transient model was run until the aquifer equilibrated under the new recharge condition and steady-state was reached. The model was considered to be at steady state when monthly cyclic variations in ground-water runoff remained constant with time. Figure 20 shows that the transient model reached steady state during the third year under the new recharge condition. However, model results indicated that ground-water runoff to the Beaver River was being overestimated during low-flow periods. Specific yield was adjusted to 0.15 for stratified drift and 0.05 for till, while all other aquifer parameters remained constant. The model was run again until a steady state was reached, and measured average monthly water level fluctuations and estimated ground-water runoff matched simulated values reasonably well. Richmond observation well 370 (fig. 21) shows that measured average monthly water levels for the 1976 nonpumping simulation were within 0.7 to 1.7 feet of predicted model values.

Using the ground-water heads obtained from the steady state response, ground-water recharge shown in table 3 was applied to simulate the 1976 to 1978 period under nonpumping conditions. Figure 21 compares measured and predicted average monthly water levels in observation wells RIW 370 and RIW 417 for nonpumping transient simulations. Comparison of estimated and predicted monthly ground-water runoff between the upper and lower gaging stations along the Beaver River, January 1976 to December 1978, is shown in figure 22 for nonpumping conditions. Figure 22 shows that the model under predicts ground-water runoff during periods of high streamflow. This is because estimated high peaks of ground-water runoff contain some overland runoff that could not be separated using the ground-water rating technique explained earlier. The model does not simulate overland runoff and no attempt was made to match high peaks. During periods of low flow most streamflow is derived from ground-water runoff.

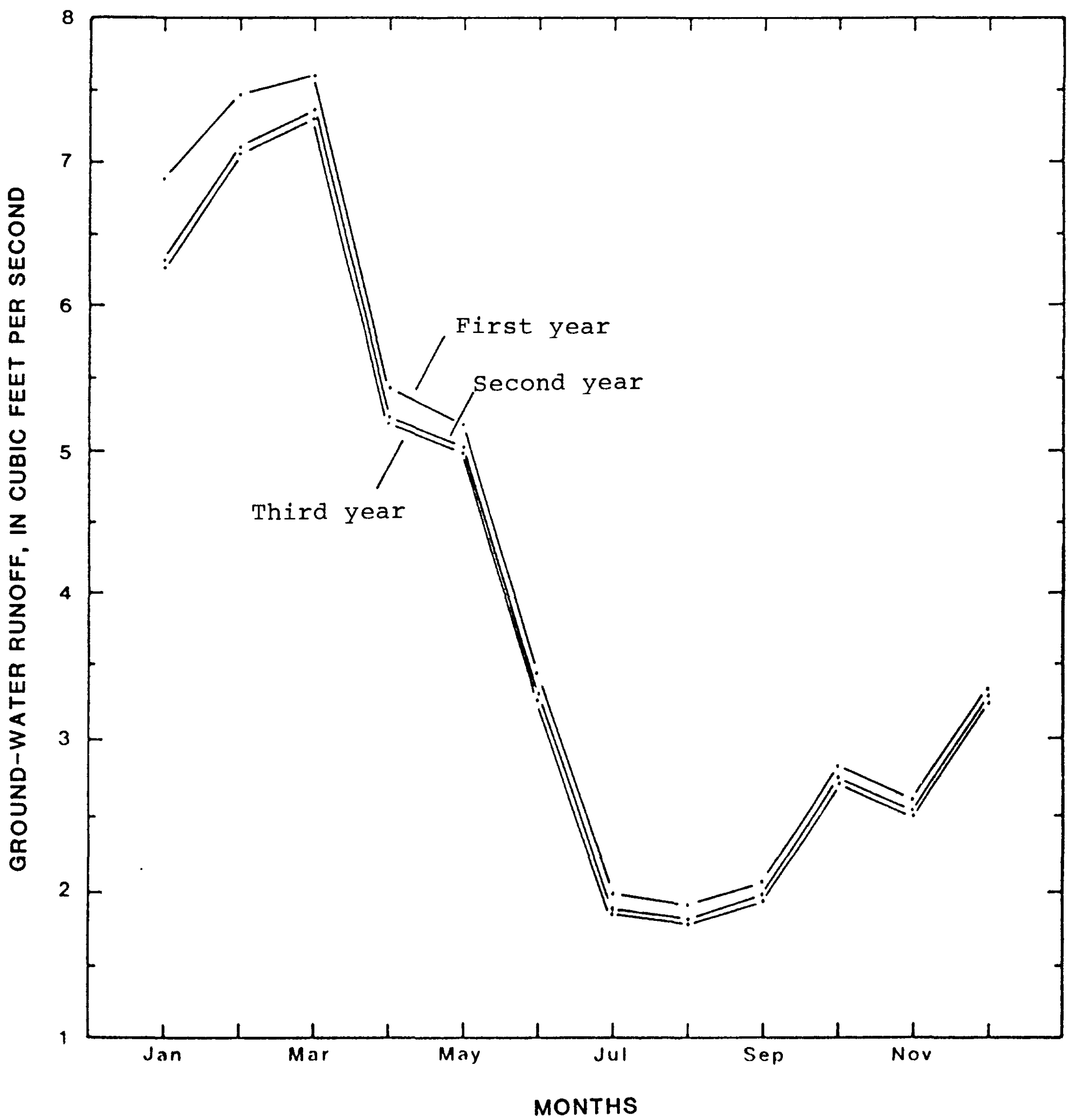


Figure 20.--Comparison of the monthly cyclic variation in ground-water runoff at the mouth of the Beaver River showing that a dynamic equilibrium was reached during the third year.

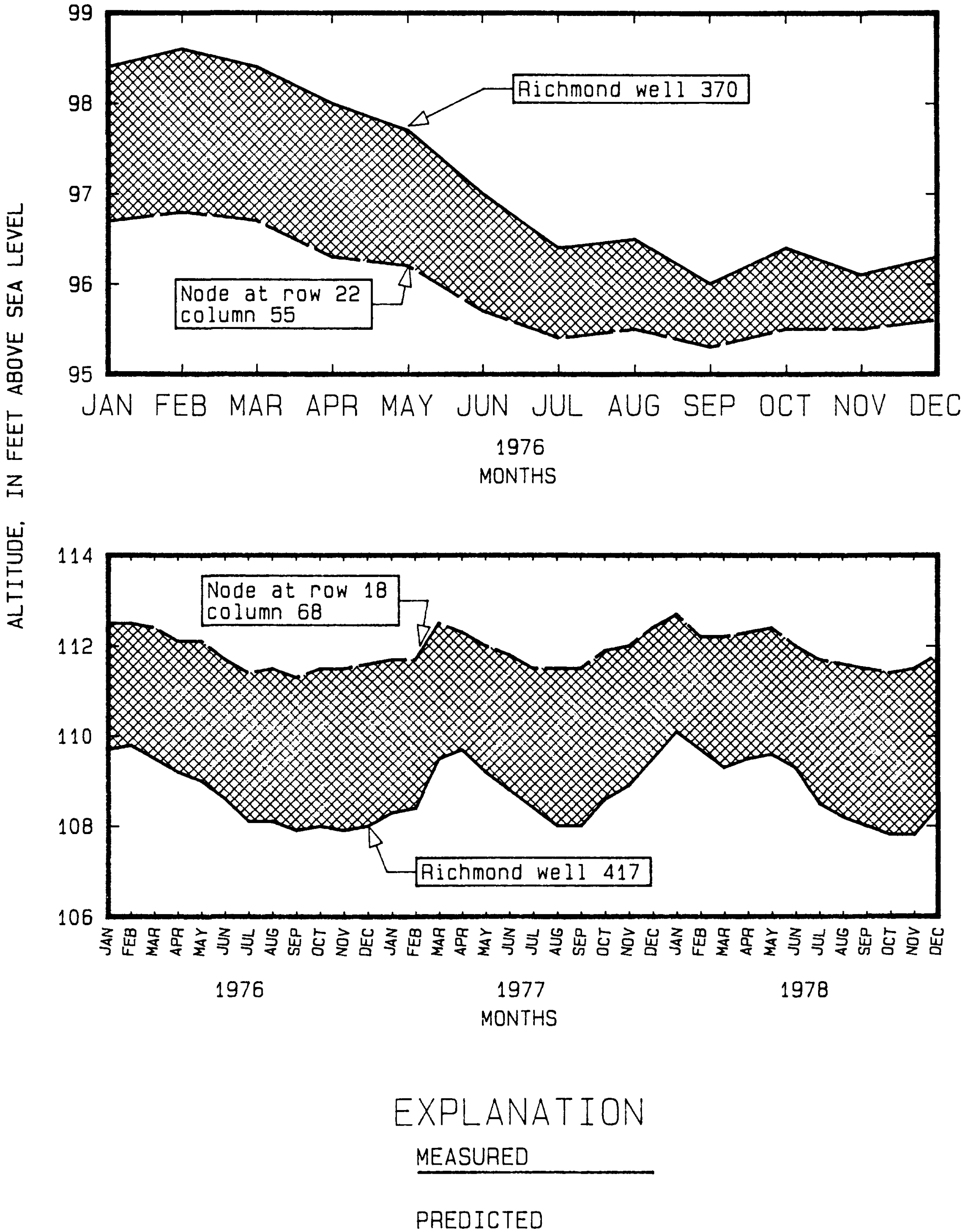


Figure 21.--Comparison of measured and predicted average monthly water levels for nonpumping transient simulations.

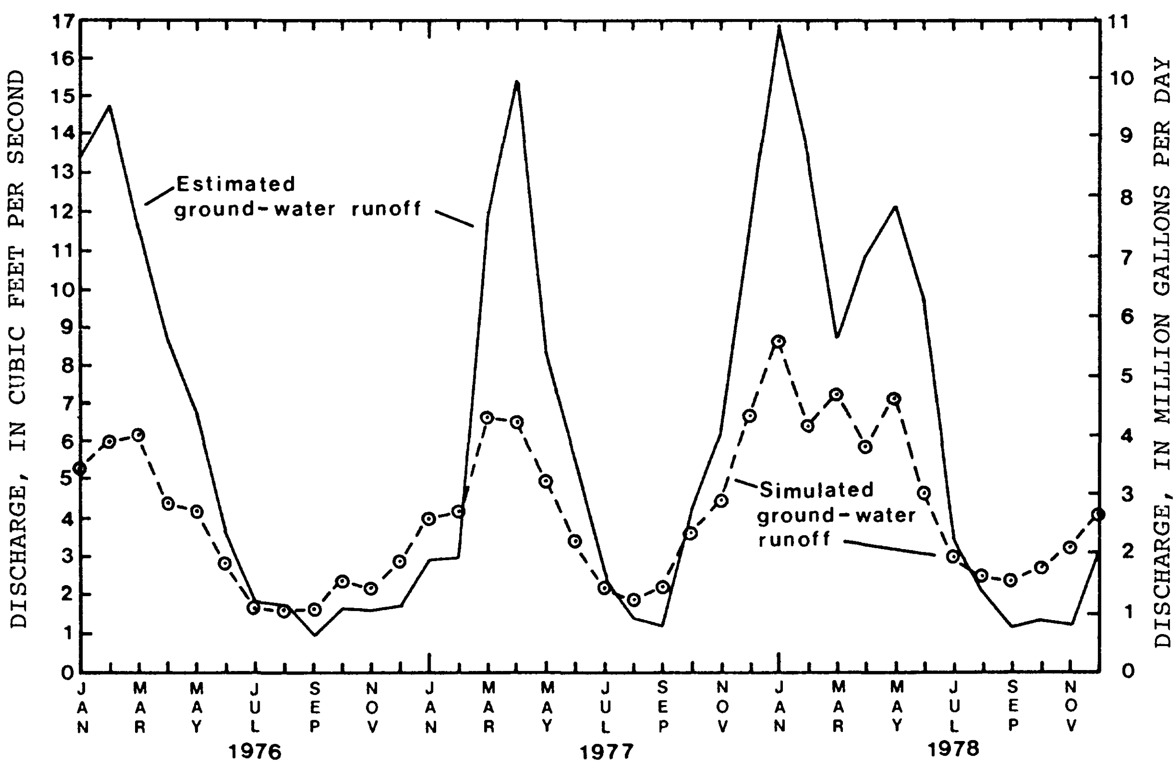


Figure 22.--Comparison of estimated and simulated monthly ground-water runoff between the upper and lower gaging stations along the Beaver River, January 1976 to December 1978, for non-pumping conditions.

Figure 22 also shows that during periods of low streamflow that the model over predicts ground-water runoff, but that differences between predicted and simulated monthly ground-water runoff, nevertheless, are within 1 ft^3/S. Based on the comparisons shown in figures 21 and 22, the transient model for the 1976-78 wet period was considered calibrated and acceptable to use for pumping simulations.

Figure 23 shows simulated transient water-table contours for December 1976 for nonpumping conditions. This figure is given to compare changes in ground-water head during an average year for development alternatives tested in ground-water pumping simulations.

Sensitivity Analysis

As part of this study, an analysis was made to determine the sensitivity of the entire model to changes in streambed hydraulic conductivity, specific yield, and transmissivity. Values for each parameter were varied while changes in ground-water runoff and ground-water head were observed at different nodes throughout the model. Results of selected sensitivity analysis and the amount that each parameter was varied from model input values are shown in figures 24 and 25. Figure 24 shows deviations in ground-water runoff upstream of the mouth of the Beaver River and figure 25 shows deviations in ground-water head at Charlestown well 400. Model input values are represented in these figures by the zero line on the vertical axis for each parameter tested.

The sensitivity of each parameter was determined by first running the model for a simulated period of several years until the water table reached equilibrium, after which cyclic variations remained constant. This analysis was done using input parameters for transient simulations and the recharge distribution for 1976. Streambed hydraulic conductivity, specific yield, and transmissivity values used in transient simulations were then varied by amounts outside the expected range of values to test extremes. The model was run for two years using heads from the final steady-state calibration as starting heads. The periods of greatest concern in the sensitivity analysis are the months of June-November when streamflow is usually low. These months are of concern because during low flow periods ground water withdrawals will have the greatest effect on reducing streamflow and lowering head in the aquifer.

Sensitivity analysis shown in figure 24 indicate that two-fold changes in streambed hydraulic conductivity and transmissivity have little effect on ground-water runoff. Order of magnitude decreases in streambed hydraulic conductivity, however, would cause an average increase in ground-water runoff of about 0.5 ft^3/s from April-September. Decreasing model specific yield from 0.15 to 0.10 has little effect on ground-water runoff during low flow periods. Although increasing model specific yield from 0.15 to 0.25 had some effect, it was minimal in that ground-water runoff only increased on the average about 0.25 ft^3/s during low flow periods.

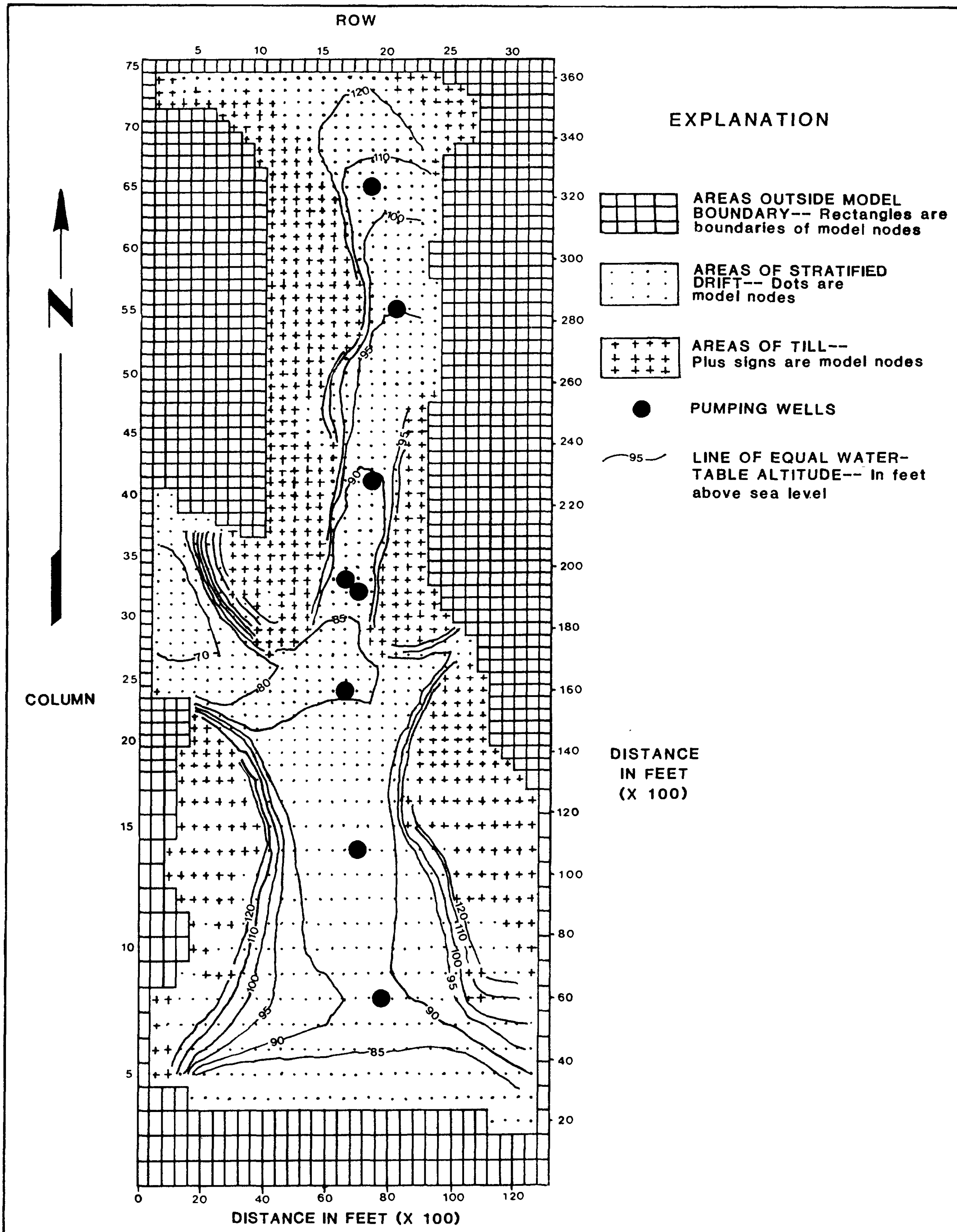


Figure 23.--Simulated water table at the end of December 1976 for nonpumping conditions.

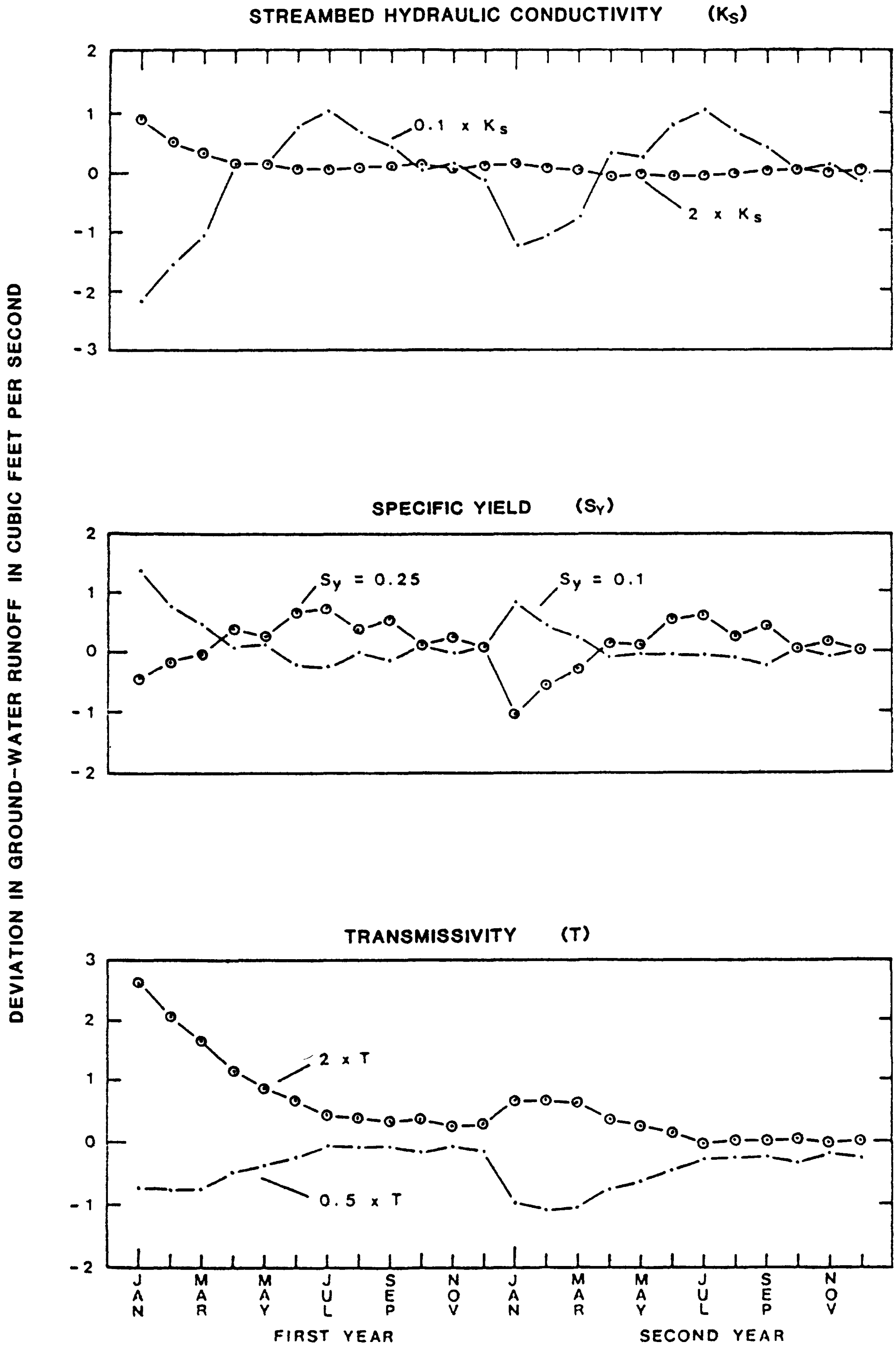


Figure 24.--Sensitivity analyses showing change in ground-water runoff at the mouth of the Beaver River.

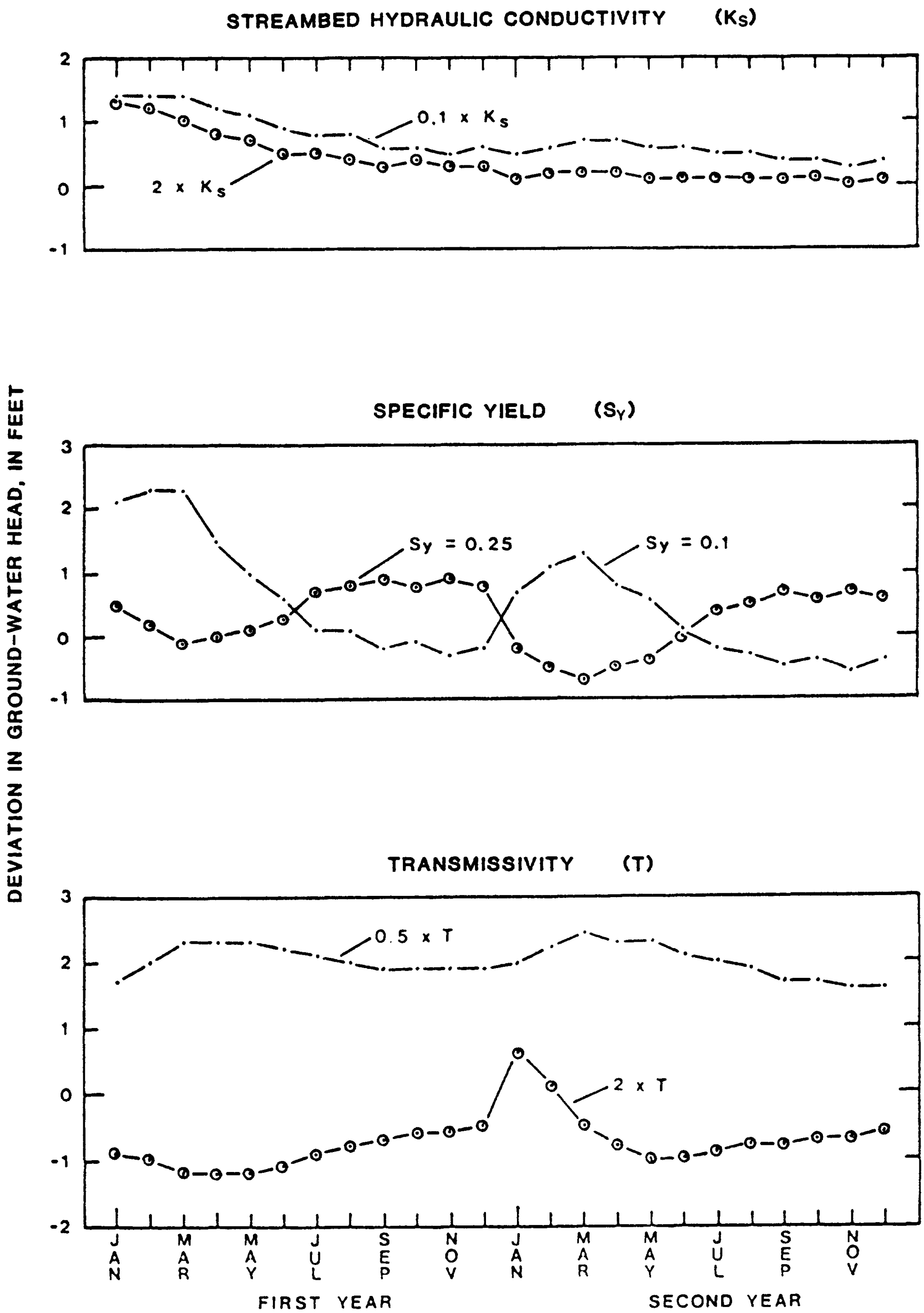


Figure 25.--Sensitivity analyses showing change in aquifer head at Charlestown well 400.

A site (Charlestown well 400) 2100 feet from the closest surface water body was used for sensitivity analysis of ground-water head. This site was selected to maximize the effect that changes in transmissivity would have on deviations in ground-water head while minimizing the effects of induced recharge from a stream. Figure 25 shows that ground-water head is most sensitive to decreases in transmissivity at Charlestown well 400. Increasing specific yield from 0.15 to 0.25 or decreasing specific yield from 0.15 to 0.10 causes deviations in ground-water head of plus or minus 0.5 foot. Figure 25 also shows that changes in streambed hydraulic conductivity have minimal effect on deviations in ground-water head when distance to the surface water source is 2100 feet or more.

In summary, sensitivity analysis show that ground-water runoff is most sensitive to large decreases in streambed hydraulic conductivity and that head in the stratified-drift aquifer is most sensitive to decreases in transmissivity. Differences between calibrated model values and the values input for the sensitivity analysis illustrate the range in simulated response associated with what is believed to be the maximum possibility for error for each parameter.

Simulated Effects of Ground-water Development

Hypothetical Ground-water Pumpage during 1976-78 Wet Period

Four development alternatives were simulated for the 1976-78 wet period to test the ability of the Beaver-Pasquiset ground-water reservoir to sustain an average daily yield of 3 Mgal/d, with a maximum short-term pumping capacity of 6 Mgal/d. The objectives of the pumping simulations were to withdraw as much water as possible from the Beaver-Pasquiset ground-water reservoir without (1) lowering the flow of the Beaver River below the estimated 98-percent flow duration, (2) causing flow in Pasquiset Brook to cease, (3) lowering the level of Pasquiset Pond 4 feet below an altitude of 89 feet above sea level, and (4) depleting initial aquifer saturated thickness 25 percent. Table 10 gives the estimated maximum pumping rate and simulated pumping rates for individual wells during various development alternatives. Estimated maximum pumping rates for individual wells (table 10) range from 0.65 to 2.50 Mgal/d. Pumping rates are not cumulative because additional drawdown caused by interference between wells was not considered. Nevertheless, they do provide some measure of potential short-term emergency well yields. Combined pumping rates for simulations ranged from 3.25 to 7.00 Mgal/d, with 2.50 to 5.00 Mgal/d from the Beaver River valley and 0.75 to 2.00 Mgal/d from the Pasquiset Brook valley. Pumping rates for individual wells ranged from 0.25 to 1.50 Mgal/d. All pumpage is assumed to be exported from the study area.

Table 10.--Ground-water development alternatives tested for the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir.
(ft: feet; in: inches; Mgal/d: million gallons per day)

Pumping site	Model node	Well number	Data from 8-inch test well: Screen interval (ft)	Screen slot size (in)	Pumping rate (Mgal/d)	Water level above screen (ft)	Estimated maximum pumping rate[1] (Mgal/d)	Pumping periods simulated[2]: 1976-1978 Development alternative (pumpage, in Mgal/d) 1	2	3	4	1963-1966 5
BEAVER RIVER VALLEY												
Andre	19, 65	RIW 405	58-63 63-73	0.160 .060	0.80	54	1.00	0.25	0.25	0.50	0.50	0.25
Perreault	21, 55	RIW 395	80-85 85-95	.160 .120	.96	58	2.50	.75	1.00	1.00	1.50	.75
Godden	19, 41	RIW 385	70-80 80-85	.120 .160	.79	65	1.50	.50	.50	.50	1.00	.50
Clark I	17, 33	RIW 400	70-75 75-85	.160 .120	.87	58	2.00	.75	1.00	.75	1.50	.75
Clark II	18, 32	RIW 415	73-83 83-88	.120 .160	.84	54	1.00	.25	.50	.25	.50	.25
							SUBTOTAL[3]	2.50	3.25	3.00	5.00	2.50
PASQUISET BROOK VALLEY												
Kenyon	17, 24	CHW 410[4]	39-55	.120	.57	34	.65	.25	.25	.25	.50	.25
Botka	18, 14	CHW 396	79-84 84-94	.160 .120	.79	70	1.00	.25	.25	.50	.50	.25
Green	20, 8	CHW 400	75-80 80-90	.160 .120	.87	66	2.50	.25	.50	.50	1.00	.25
							SUBTOTAL[5]	0.75	1.00	1.25	2.00	0.75
							TOTAL	3.25	4.25	4.25	7.00	3.25

1 Estimated maximum pumping rates are not cumulative because interference between wells was not considered. Estimates are based on unadjusted specific capacity and water available above the top of the well screen. Optimum size well casing required to pump estimated maximum ranges from a minimum of 12 inches to a maximum of 24 inches.
2 Wells were pumped continuously during all simulations.
3 Total ground-water pumpage upstream of the mouth of the Beaver River.
4 Kenyon Piece Dye Works, Inc. 18 x 12 inch gravel-packed well.
5 Total ground-water pumpage upstream of the mouth of the Pasquiset Brook.

Ground-water development alternatives were simulated at eight pumping sites consisting of seven hypothetical wells and one real well. Hypothetical well sites were pretested during 8-inch aquifer tests, conducted as part of this study, at rates ranging from 0.79 to 0.96 Mgal/d. The real well, an 18- x 12-inch gravel-packed well owned by Kenyon Piece Dye Works, Inc., was pretested at 0.57 Mgal/d during its acceptance test. Locations of hypothetical and real pumping wells are shown in figure 18.

Comparisons between changes in stream discharge, ground-water runoff and induced recharge were prepared from monthly model output values for nonpumping and pumping simulations (figs. 26 to 33). Monthly total cumulative flow was used to prepare illustrations showing changes in stream discharge (figs. 26, 28, 30, and 32). The monthly total leakage rate was used to prepare illustrations showing changes in ground-water runoff [leakage into (-) the river] and induced recharge [leakage out of (+) the river] (figs. 27, 29, 31, and 33). An example of monthly flow is shown later in table 12.

The impact of simulated ground-water withdrawals of 2.50 Mgal/d on streamflow depletion upstream of the mouth of the Beaver River for development alternative one is shown in figures 26 and 27. Figure 26 shows that withdrawal of 2.50 Mgal/d causes an average monthly decline in stream discharge of about 3 ft^3/s, but that average monthly stream discharge only dropped below the estimated 98-percent flow duration during September 1976. At no time during this simulation period did the model indicate any reach of the Beaver River drying up. Stream discharge declines under pumping conditions because part of the water pumped from the well comes from ground-water runoff intercepted before it discharges into the stream and part is derived from induced recharge of streamflow. The rest of ground-water pumpage is derived from aquifer storage and reduced evapotranspiration.

Figure 27 shows that total ground-water runoff available upstream of the mouth of the Beaver River was being diverted to wells from June through December 1976, July through September 1977, and July through October 1978, while pumping 2.50 Mgal/d. The amount of ground-water pumpage derived from induced recharge from the river during these months is shown by the stippled pattern in figure 27 .

During testing of development alternative two, ground-water withdrawals in the Beaver River valley were increased from 2.50 to 3.25 Mgal/d (table 10). Figure 28 shows that withdrawal of 3.25 Mgal/d causes an average monthly decline in stream discharge of about 4 ft^3/s, and that average monthly stream discharge dropped below the estimated 98-percent flow duration during July and September 1976. At no time during this simulation period did the model indicate any reach of the Beaver River drying up.

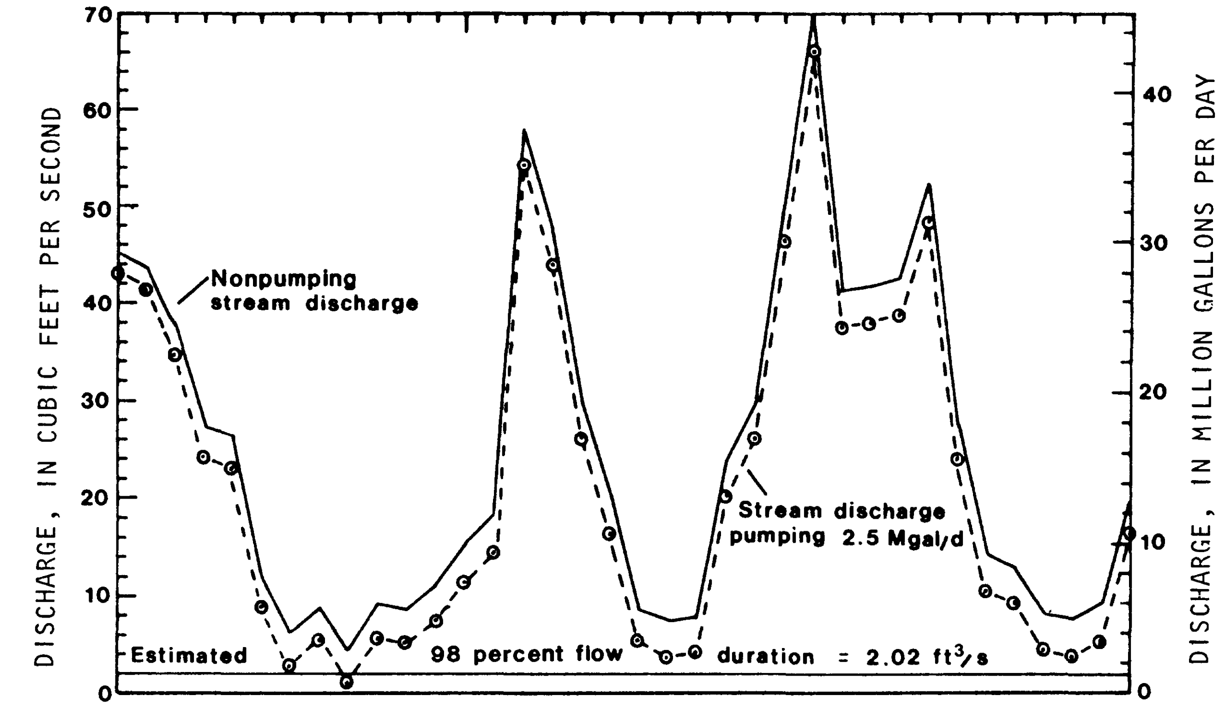


Figure 26.--Comparison of stream discharge at the mouth of the Beaver River for nonpumping and pumping (2.5 Mgal/d) simulations for 1976-78.

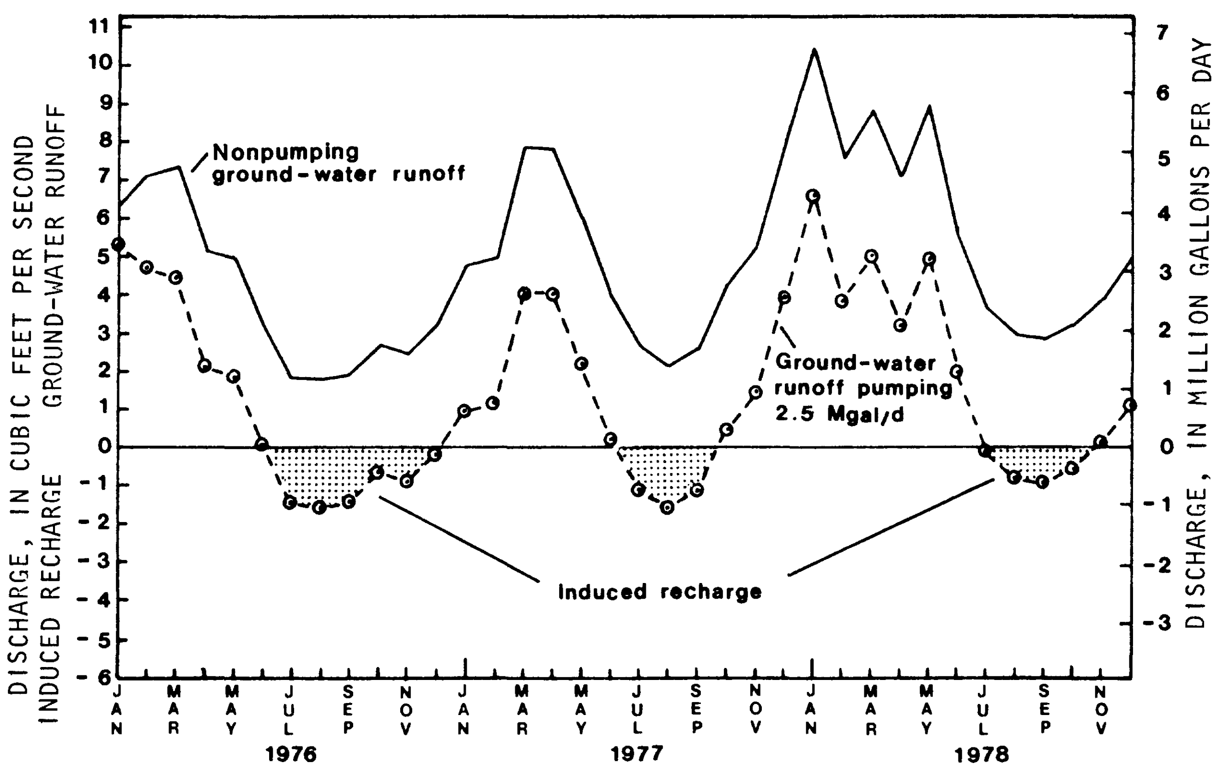


Figure 27.--Comparison of ground-water runoff and induced recharge at the mouth of the Beaver River for nonpumping and pumping (2.5 Mgal/d) simulations for 1976-78.

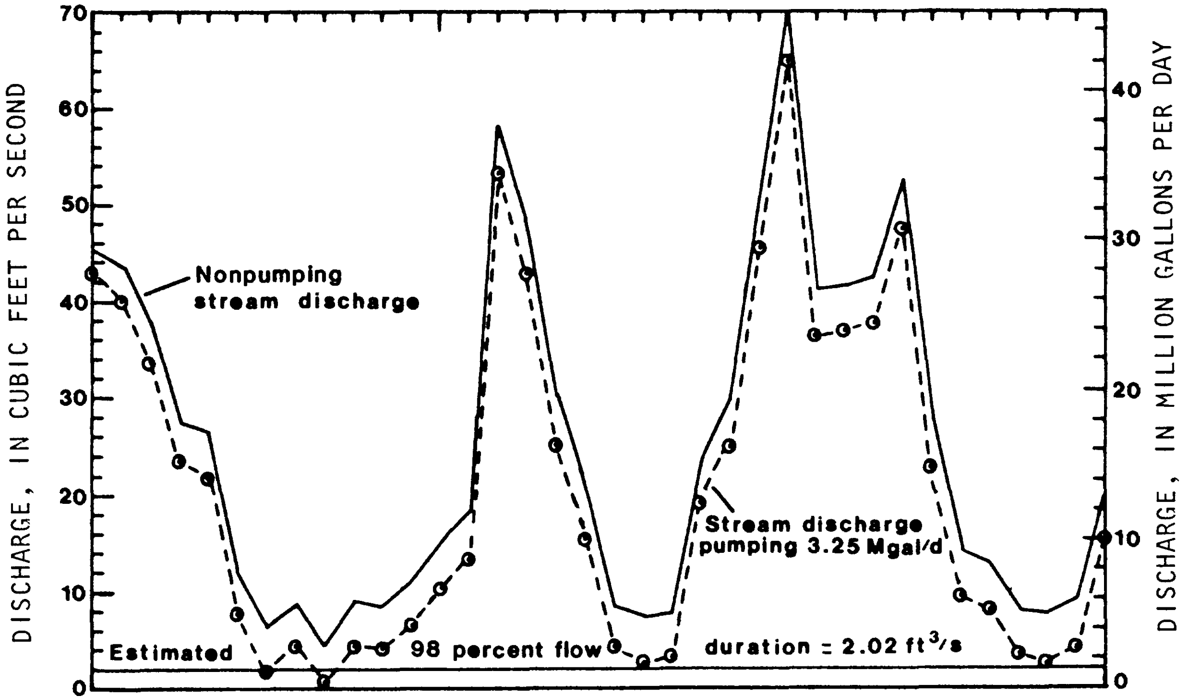


Figure 28.--Comparison of stream discharge at the mouth of the Beaver River for nonpumping and pumping (3.25 Mgal/d) simulations for 1976-78.

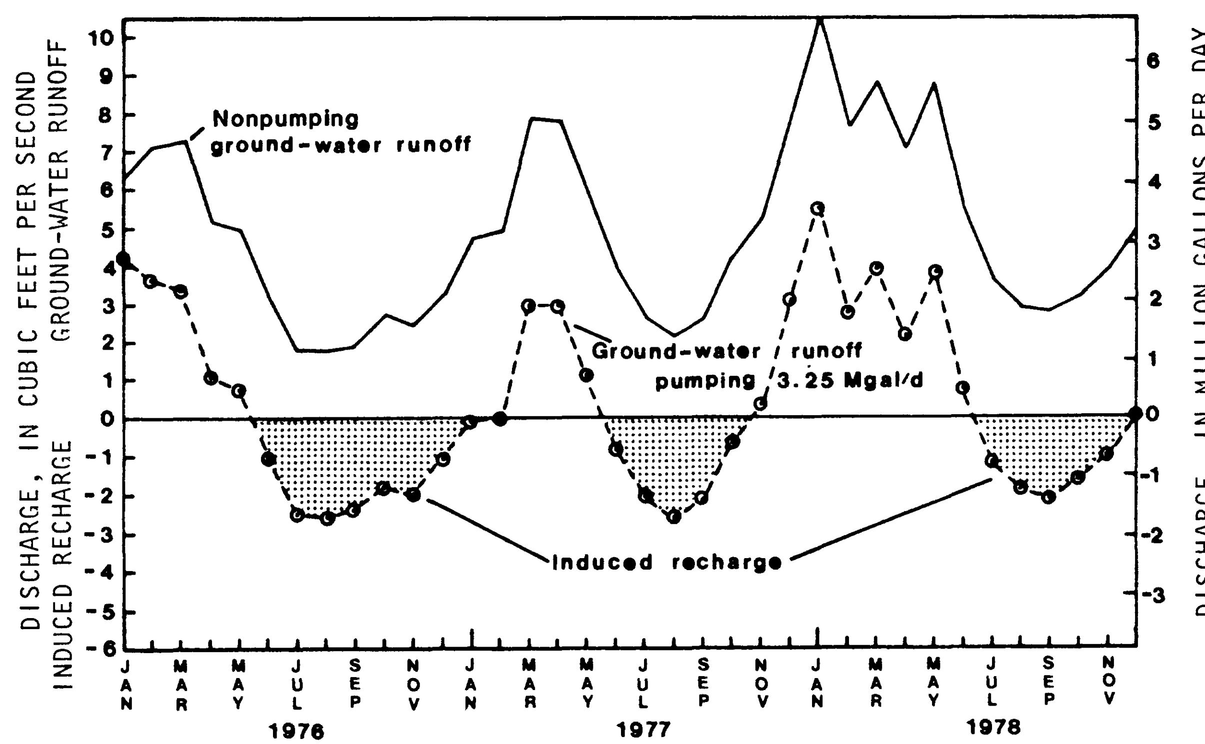


Figure 29.--Comparison of ground-water runoff and induced recharge at the mouth of the Beaver River for nonpumping and pumping (3.25 Mgal/d) simulations for 1976-78.

Figure 29 shows that total ground-water runoff available upstream of the mouth of the Beaver River was being diverted to wells from June 1976 through February 1977, June through October 1977, and July through December 1978. The amount of ground-water pumpage derived from induced recharge during these months is shown by the stippled pattern. Analysis of ground-water pumpage derived from induced recharge shown in figures 27 and 29 indicates that at the higher pumping rate, there will be 6 more months of induced river infiltration.

During development alternative three, ground-water pumpage of 0.25 Mgal/d was shifted from the Beaver River valley to the Pasquiset Brook valley to increase flow at the mouth of the Beaver River during July and September 1976. Average monthly stream discharge was increased during this simulation above the estimated 98-percent flow duration for July 1976, but not for September 1976. Flow during September increased from 0.09 ft^3/s to 0.44 ft^3/s. The impact of simulated ground-water withdrawals of 3.00 Mgal/d on stream discharge, and on ground-water runoff and induced recharge is shown in figures 30 and 31.

To test the maximum pumping capacity proposed by the Water Resources Board, ground-water withdrawals in the Beaver River valley were doubled from 2.50 Mgal/d during development alternative one to 5.00 Mgal/d for development alternative four. Figure 32 shows that withdrawal of 5.00 Mgal/d causes average monthly stream discharge to drop below the estimated 98-percent flow duration during 5 months in 1976, and 3 months in 1977 and 1978. This figure also demonstrates that the model can simulate intermittent drying and recovery of stream reaches by showing the Beaver River going dry at its mouth in July and September 1976 and recovering flow in August and October 1976.

Model results pumping 5.00 Mgal/d also show that all available ground-water runoff is being diverted to wells during 26 out of 36 months, and that most pumpage is derived from induced recharge at this simulated pumping rate (fig. 33). A sharp decrease in induced recharge during September 1976 (fig. 33) was caused by a 4,000-foot reach of the Beaver River going dry and no water available for induced recharge. However, average monthly induced recharge is not computed as zero by the model because flow was available in the stream during some days of the month.

The model did not indicate any reaches of the Pasquiset Brook going dry during development alternatives one through three when ground-water withdrawals varied from 0.75 to 1.25 Mgal/d. However, this was not the case with simulated ground-water withdrawals of 2.00 Mgal/d during development alternative four. Monthly output similar to table 12 indicated that pumping at a rate of 2.00 Mgal/d caused Pasquiset Brook, at the outlet of Pasquiset Pond, to go dry during some months in each of the three years simulated. An 800-foot reach (1 model node) of Pasquiset Brook went dry in July, August, and December 1976; July and October 1977; and September, October, and November 1978. A 1,600-foot reach (2 model nodes) of the brook went dry in September, October, and November 1976, and August and September 1977.

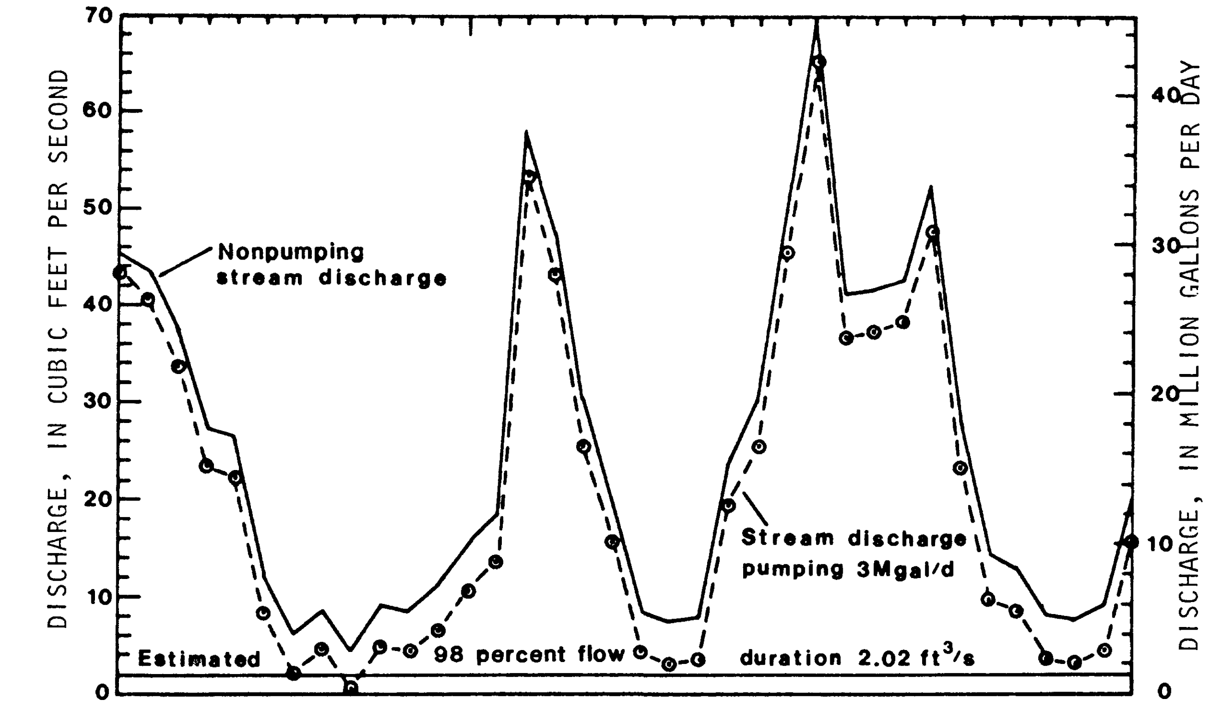


Figure 30.--Comparison of stream discharge at the mouth of the Beaver River for nonpumping and pumping (3 Mgal/d) simulations for 1976-78.

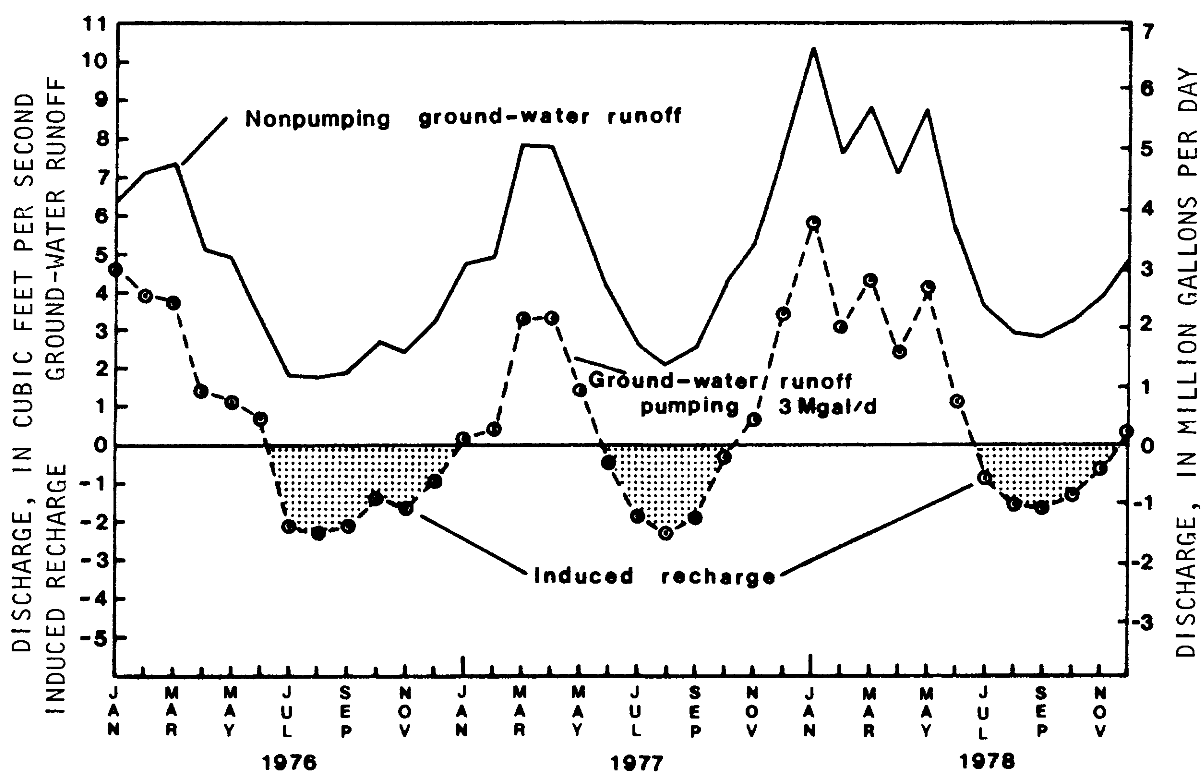


Figure 31.--Comparison of ground-water runoff and induced recharge at the mouth of the Beaver River for nonpumping and pumping (3 Mgal/d) simulations for 1976-78.

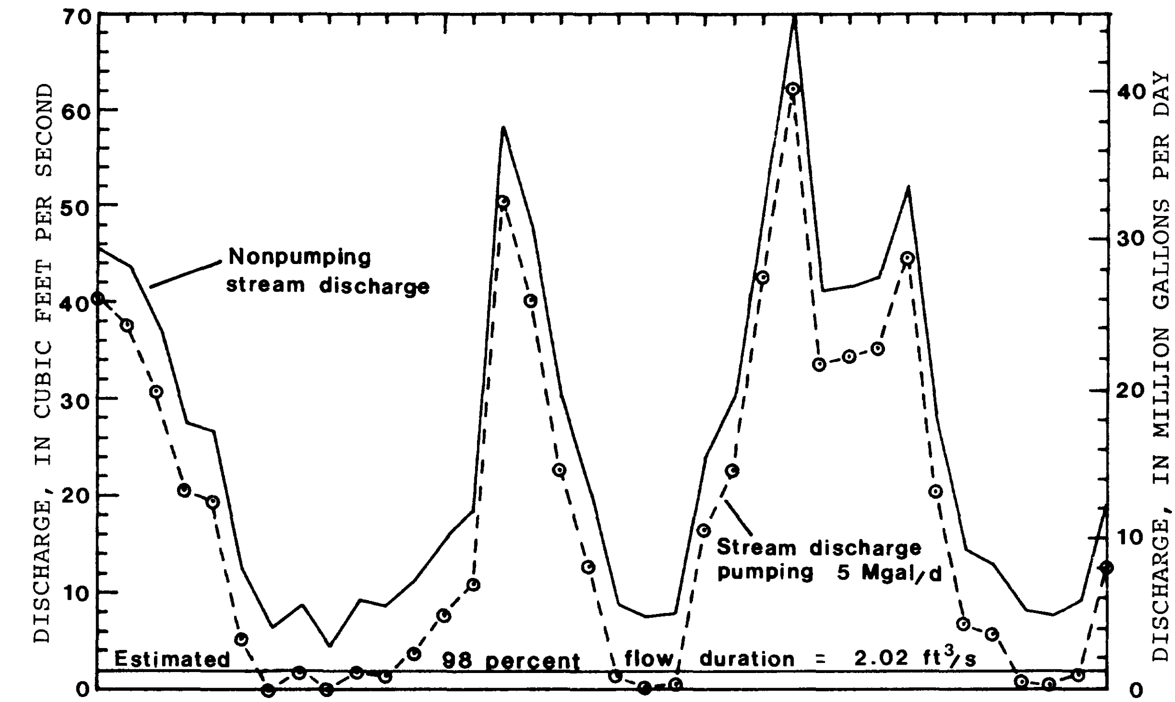


Figure 32.--Comparison of stream discharge at the mouth of the Beaver River for nonpumping and pumping (5 Mgal/d) simulations for 1976-1978.

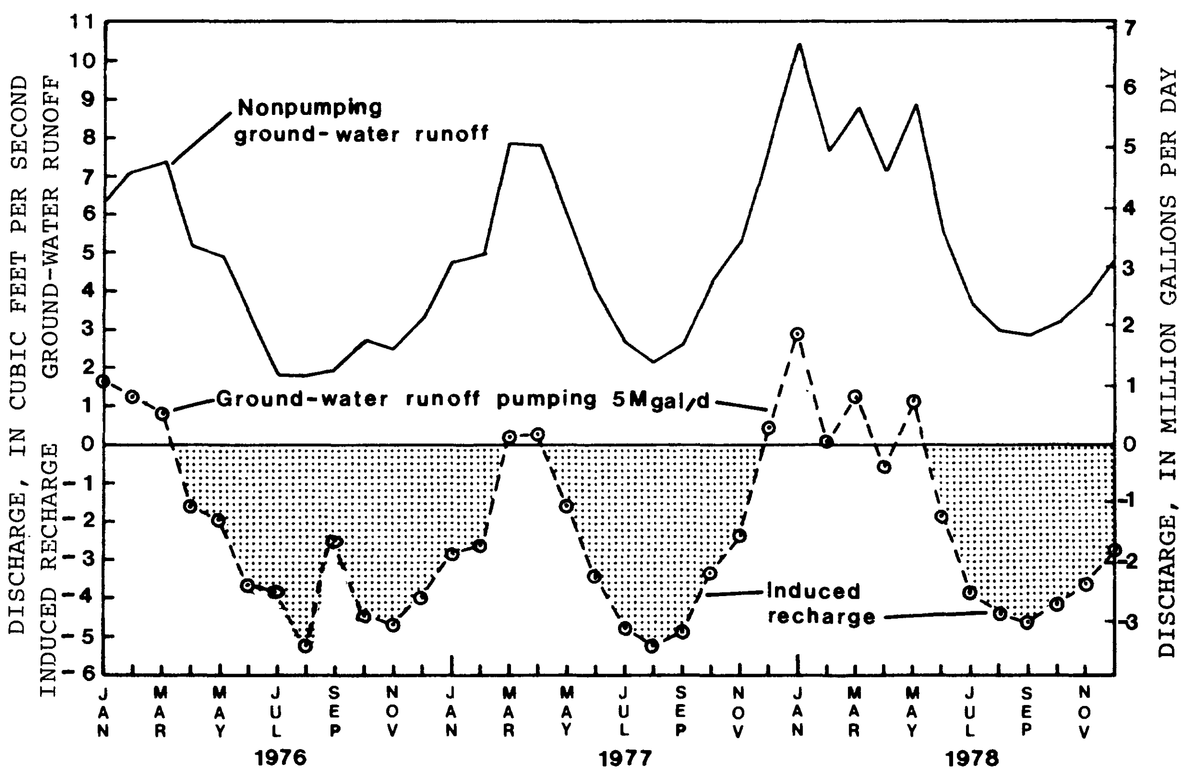


Figure 33.--Comparison of ground-water runoff and induced recharge at the mouth of the Beaver River for nonpumping and pumping (5 Mgal/d) simulations for 1976-1978.

Output from the modified model also allows the impact of simulated pumping on the altitude of Pasquiset Pond to be assessed (fig. 34). It shows the altitude of the pond declining an average of only 0.10 foot below a nonpumping altitude of 88 feet above sea level when pumping a total of 0.75 Mgal/d from wells CHW 396, CHW 400, and CHW 410 in the Pasquiset Brook valley (see fig. 18 for location of wells). Under simulated pumping of 2.00 Mgal/d, the altitude of the pond declined a maximum of 1.7 feet in November 1976, 2.2 feet in September 1977, and 1.5 feet in October 1978. Figure 35 shows the surface area, average depth, and storage of Pasquiset Pond. The surface altitude of Pasquiset Pond was established as 89 feet above sea level, July 19, 1949, and is published in a pond and lake survey conducted by the Division of Fish and Game, Rhode Island Department of Agriculture and Conservation, June 12, 1957. The 0.10 foot decline in pond level caused by pumping 0.75 Mgal/d had minimal effect on reducing the surface area of the pond. Pumping 2.00 Mgal/d would, on the average, reduce the overall size of Pasquiset Pond by approximately 8 million gallons, the white area (1.5 feet average depth) shown in figure 35.

In the stratified-drift aquifer, maximum drawdown did not exceed 12 percent of initial aquifer saturated thickness at any node during any of the 4 pumping simulations. Maximum simulated drawdown for pumping of 5.00 Mgal/d in the Beaver River valley was 9.2 feet of an available 100 feet. Maximum drawdown predicted by the model pumping of 2.00 Mgal/d in the Pasquiset Brook valley was 11.2 feet of an available 100 feet. Table 11 shows a comparison between measured drawdown and maximum predicted drawdown at model pumping nodes near 8-inch test sites.

Simulated water table contours for December 1976 are given in figures 36 and 37 for development alternative one and four. Comparison of pumping water table contours (figs. 36 and 37) with nonpumping water table contours (fig. 23) show the greatest change occurring around the 85 contour in the Beaver River Valley, and the 90 contour in the Pasquiset Brook Valley, where pumping during development alternative 1 caused water level declines of about 2 feet. Development alternative 4, where pumpage was approximately doubled, caused water level declines of about 3 feet around the 85 contour in the Beaver river Valley. However, the most significant change under development alternative 4 occurred in the Pasquiset Brook Valley, where the 85 contour moved approximately 3,600 feet to the north. This caused the ground-water divide to shift about 5,000 feet north of its original position shown on the water-table map in figure 2.

September 1976 was chosen to illustrate the effectiveness of the model in simulating surface water and ground water interactions. Any month could have been chosen, however, September 1976 was selected because it is the month that shows the greatest impact of ground-water withdrawals on the stream-pond-aquifer system. Profiles of stream discharge along the Beaver River (fig.38a&b), from the northernmost stream node to the mouth, show simulated reduction in streamflow due to ground-water withdrawals for the 4 development alternatives for September 1976.

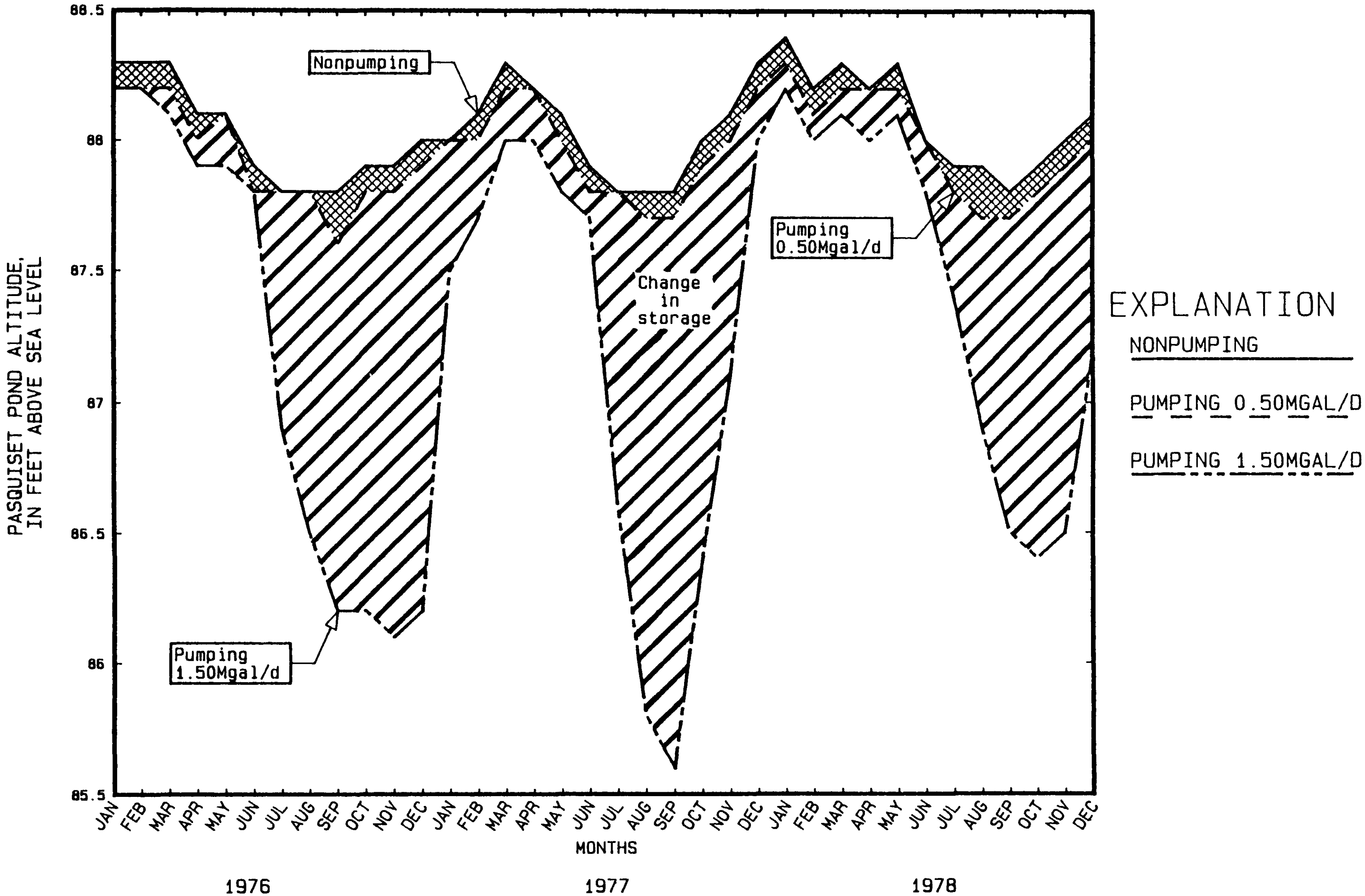
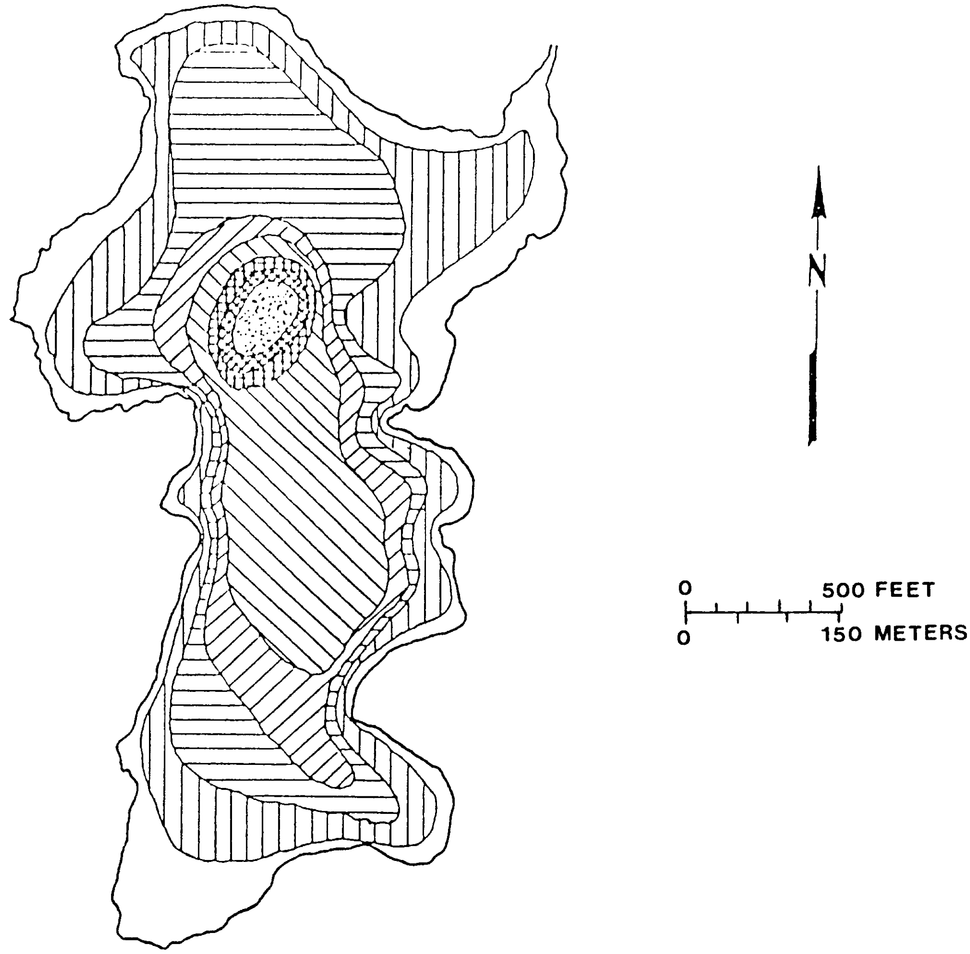


Figure 34.--Comparison of changes in Pasquiset Pond altitude for nonpumping and pumping simulations for 1976-78 showing effects of nearby ground-water pumpage on pond storage.



	Average depth (feet)	Area (acres)	Water in storage (gallons)
	1.5	16.6	8,100,000
	4.5	18.3	26,800,000
	7.5	19.7	48,100,000
	10.5	7.6	26,100,000
	13.5	11.1	48,700,000
	16.5	1.8	9,800,000
	19.5	0.9	5,500,000
	21.0	1.3	8,800,000
		77.3	181,900,000

Figure 35.--Map showing surface area, depth, and storage at Pasquiset Pond.

Table 11.--Comparison of drawdowns from aquifer tests with simulated drawdowns during maximum pumpage of 7 million gallons per day, September 1976. (d: days; Mgal/d: million gallons per day; ft: feet)

			AQUIFER TEST			SIMULATION				
Pumping site	Model node	Well number	Time (d)	Pumping rate (Mgal/d)	Drawdown[1] (ft)	Time (d)	Pumping rate (Mgal/d)	Drawdown[1,2] (ft)	Drawdown at pumping node (ft)	Drawdown[3] (ft)
Andre	19, 65	RIW 405	3	0.80	13.8 (4.3)	334	0.50	5.9 (4.3)	3.3	7.2
Perreault	21, 55	RIW 395	2	.96	7.5 (4.6)	334	1.50	8.2 (4.6)	2.0	11.4
Godden	19, 41	RIW 385	2	.79	8.9 (2.1)	334	1.00	11.2 (2.1)	6.0	12.3
Clark I	17, 33	RIW 400	3	.87	10.7 (5.4)	334	1.50	15.0 (5.4)	9.2	18.6
Clark II	18, 32	RIW 415	3	.84	8.5 (6.1)	334	.50	6.2 (6.1)	4.5	7.4
Kenyon	17, 24	CHW 410	2	.57	18.3 (2.0)	334	.50	8.4 (2.0)	4.0	9.2
Botka	18, 14	CHW 396	3	.79	11.5 (6.5)	334	.50	5.2 (6.5)	2.2	7.4
Green	20, 8	CHW 400	2	.87	8.1 (5.8)	334	1.00	15.0 (5.8)	11.2	17.5
				[4]6.46			[5]7.00			

1 Number in parentheses ( ) is distance from pump well to observation well in feet.
2 Drawdown at model pumping node adjusted to radius shown in parenthesis ( ).
3 Drawdown at model pumping node adjusted to 1 foot radius.
4 Wells not pumped simultaneously.
5 Wells pumped simultaneously.

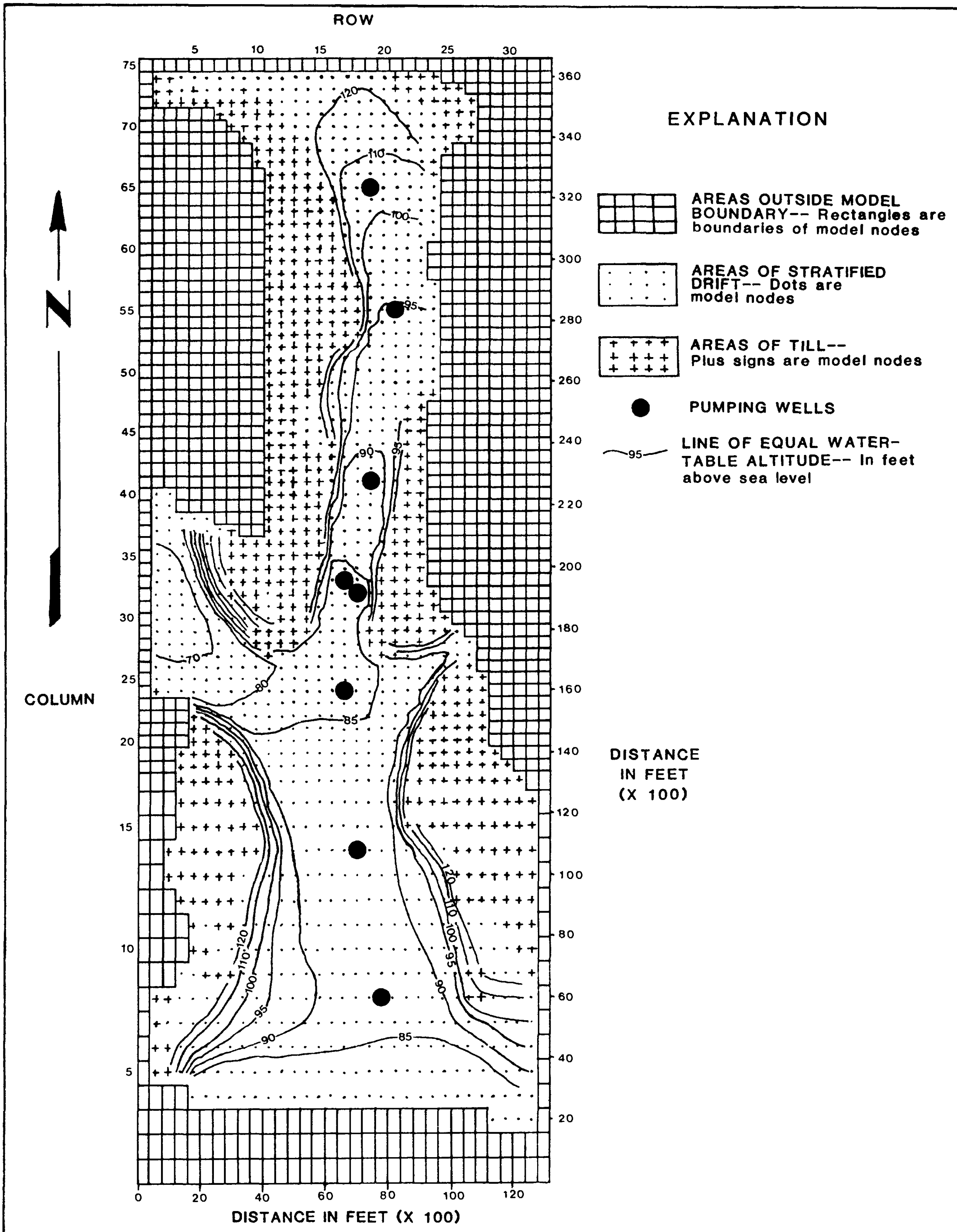


Figure 36.--Simulated water table at the end of December 1976 for development alternative 1 pumping 3.25 Mgal/d.

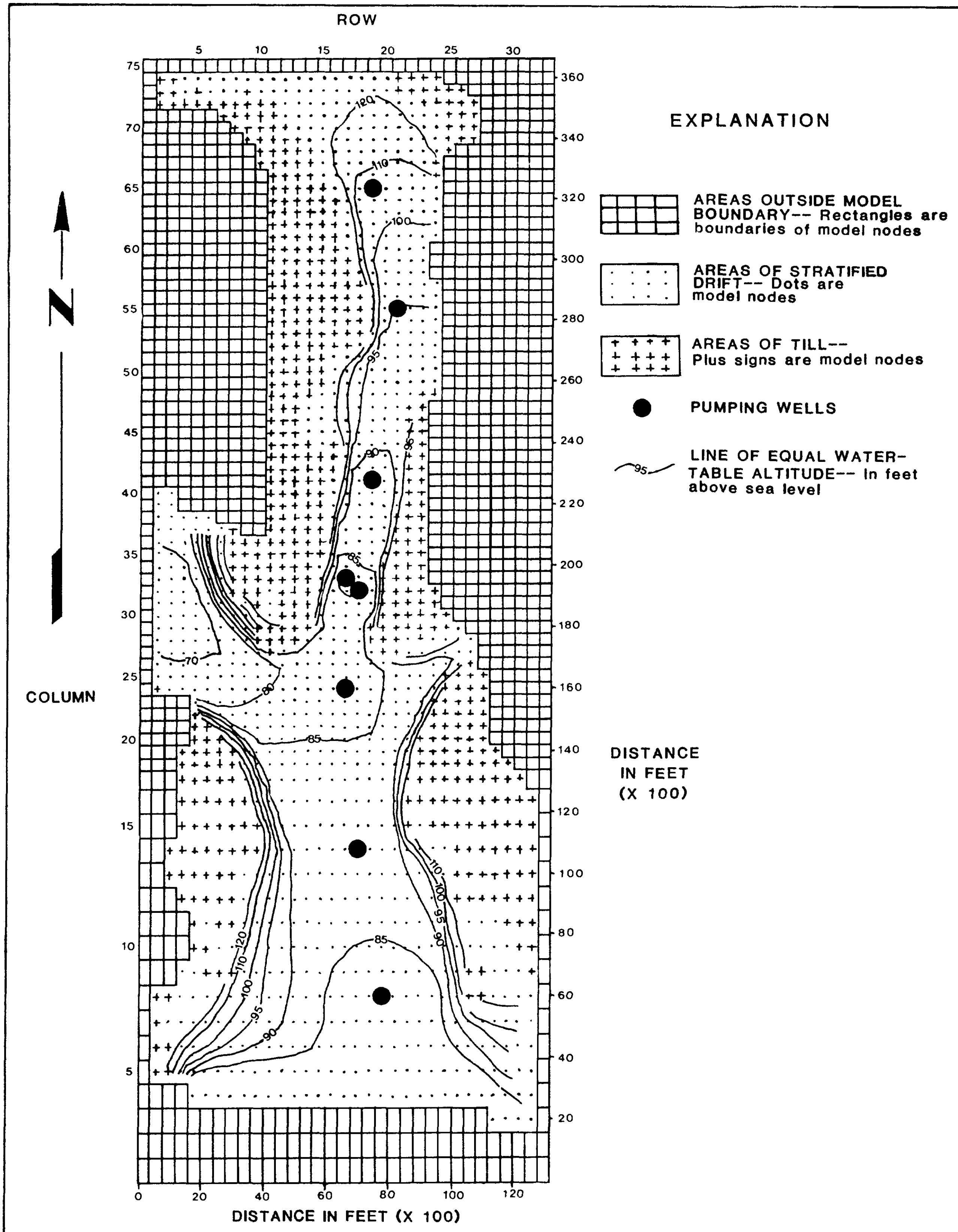


Figure 37.--Simulated water table at the end of December 1976 for development alternative 4 pumping 7 Mgal/d.

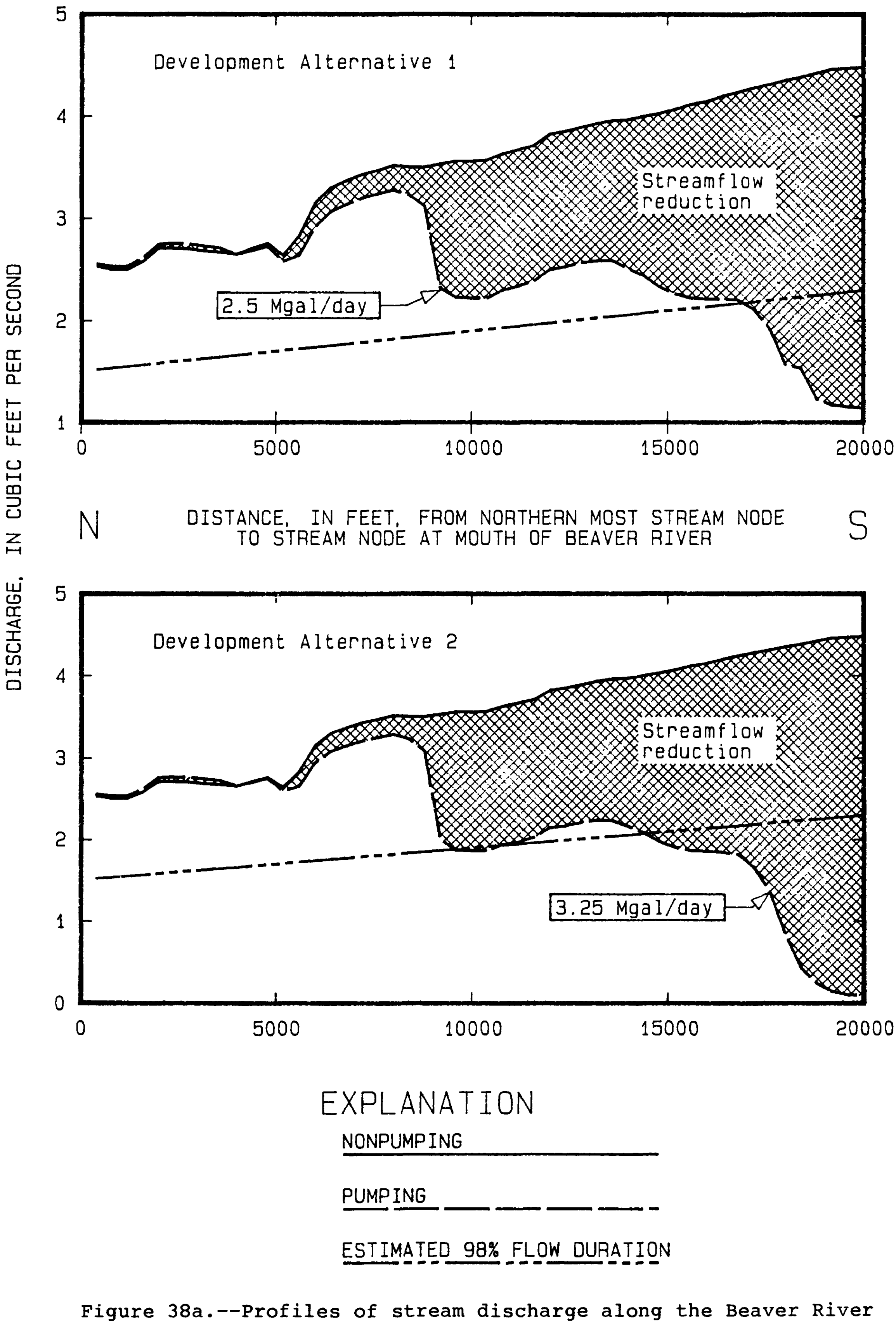


Figure 38a.--Profiles of stream discharge along the Beaver River showing reduction in streamflow due to ground-water pumpage under simulated development alternatives 1 and 2, September 1976.

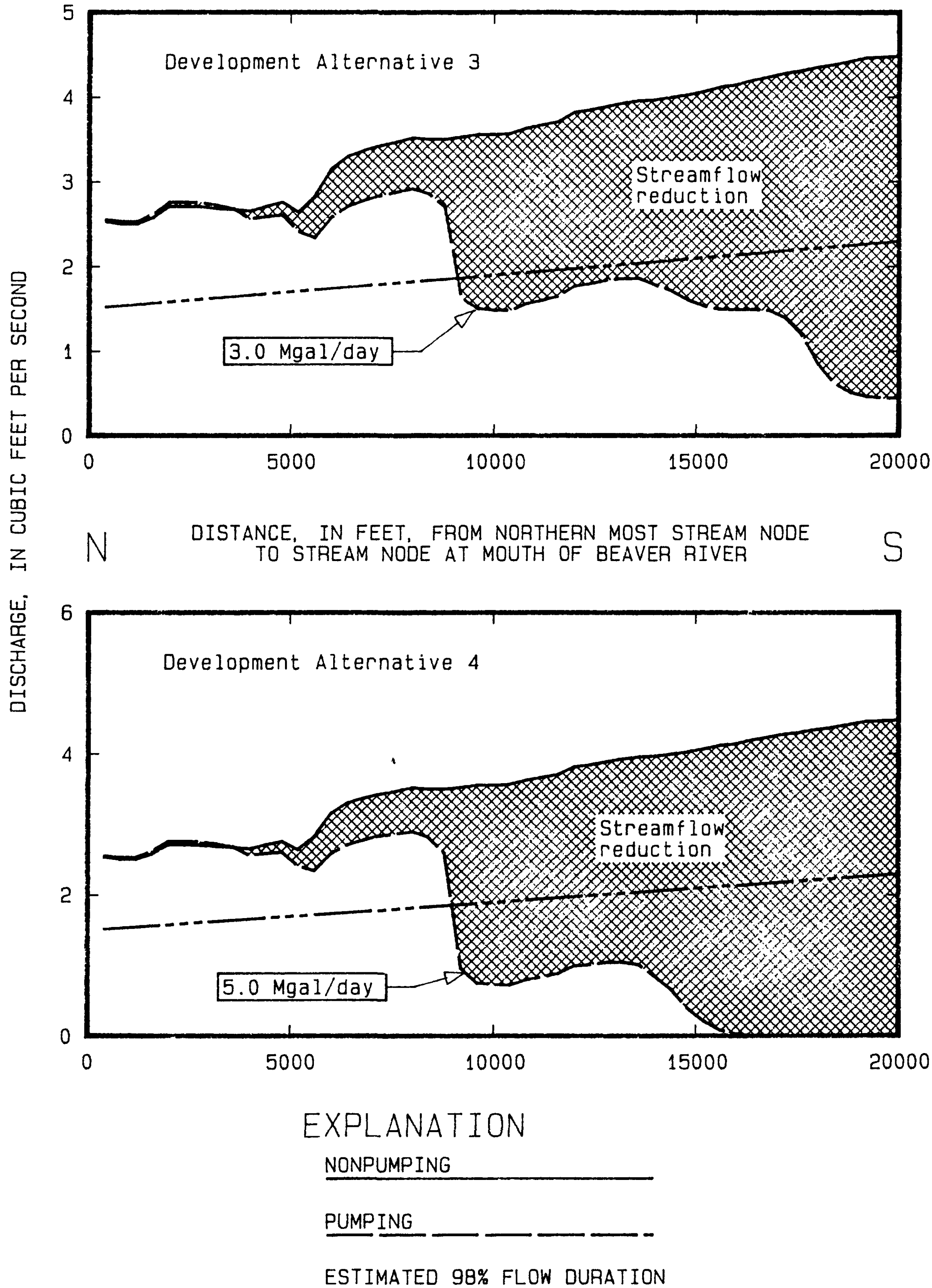


Figure 38b.--Profiles of stream discharge along the Beaver River showing reduction in streamflow due to ground-water pumpage under simulated development alternatives 3 and 4, September 1976.

From model output, ground-water runoff available for diversion to wells is obtained by totaling leakage rates at each node for the nonpumping simulation. For example, table 12 shows that ground-water runoff available for diversion to wells for September 1976 is 1.94 ft^3/s. Ground-water pumpage derived from induced recharge is obtained by totaling leakage rates at each node for the pumping simulation. Table 12 shows that induced recharge accounted for 2.51 ft^3/s of ground-water pumpage for September 1976. The amount of pumpage derived from a combination of aquifer storage and reduced evapotranspiration is a calculated value equal to total ground-water pumpage minus ground-water runoff and induced recharge (see table 13).

The source of water diverted to five hypothetical pumping wells in the Beaver River valley, and the percentage of water derived from each source during the 1976-78 simulation period, is shown in figure 39 for combined pumping rates of 2.50 and 5.00 Mgal/d. Values used in figure 39 were obtained from monthly model output similar to that shown in table 12. A point of interest in figure 39 is the abrupt increase in the percentage of ground-water pumpage derived from combined aquifer storage and reduced evapotranspiration in July and September 1976 for simulated pumping of 5.00 Mgal/d. The increase was caused by the Beaver River going dry at several stream nodes upstream of its mouth, which thereby decreased the availability of streamflow for infiltration. Table 12 shows the last 10 river nodes dry in September 1976 for simulated pumping at a rate of 5.00 Mgal/d. With the model output shown in table 12, and the method of analysis described above the user can quickly and efficiently evaluate stream-aquifer interactions for any selected time period.

The model output in table 12 also shows that pumping 5.00 Mgal/d from wells in the Beaver River valley diverts all ground-water runoff available under nonpumping conditions (1.94 ft^3/s) to wells before it discharges to (leaks into) the river, and that 2.51 ft^3/s of river water is diverted to wells as induced recharge. Although the total amount of ground-water runoff available for development upstream of the mouth of the Beaver River has been diverted to wells, it is important to note that the output in table 12 shows that some stream nodes (those with minus values) are still receiving ground-water runoff from the aquifer. This is because pumpage is spread out along the Beaver River and some stream nodes are either outside individual cones of depression surrounding pumping wells or minimally affected by them. Figure 40 illustrates combined ground-water pumpage effects on stream and pond nodes in September 1976 pumping 7.00 Mgal/d.

Table 12.--Sample of model output of stream leakage and cumulative flow for September 1976 for simulated pumping at 5 million gallons per day from wells in the Beaver River valley. (ft^3/s: cubic feet per second)

				Nonpumping simulation	Pumping simulation	
				Leakage into (-)[1] and out of (+)[2] the river		
Pumping node Row	Pumping node Column	Stream node Row	Stream node Column	Leakage rate ft^3/s	Leakage rate ft^3/s	Cumulative flow ft^3/s
		14	74	0.0092	-0.0140	2.5489
		15	74	0.0227	0.0167	2.5322
		16	74	0.0056	0.0032	2.5290
		16	73	-0.0839	-0.0921	2.6211
		17	72	-0.1258	-0.1341	2.7552
		18	71	-0.0017	-0.0078	2.7631
		18	70	0.0100	0.0093	2.7538
		18	69	0.0195	0.0301	2.7237
		18	68	0.0077	0.0428	2.6810
		18	67	0.0247	0.1228	2.5582
		17	66	-0.0601	-0.0291	2.5873
		17	65	-0.0546	-0.0223	2.6096
19	65	18	64	0.1239	0.1976	2.4121
		19	64	-0.1922	0.0733	2.3388
		20	63	-0.3193	-0.2429	2.5817
		20	62	-0.1528	-0.1320	2.7137
		20	61	-0.0699	-0.0634	2.7771
		20	60	-0.0607	-0.0583	2.8354
		20	59	-0.0388	-0.0356	2.8710
		20	58	-0.0471	-0.0328	2.9038
		21	57	0.0212	0.0794	2.8244
		21	56	-0.0047	0.2373	2.5872
21	55	21	55	-0.0244	1.6286	0.9586
		21	54	-0.0314	0.2055	0.7530
		20	53	-0.0066	0.0235	0.7295
		20	52	-0.0014	0.0068	0.7227
		19	51	-0.0652	-0.0775	0.8002
		20	50	-0.0352	-0.0379	0.8381
		20	49	-0.0472	-0.0522	0.8903
		20	48	-0.1090	-0.1070	0.9973
		21	47	-0.0279	-0.0244	1.0217
		21	46	-0.0368	-0.0274	1.0491
		20	45	-0.0394	-0.0049	1.0541
		20	44	-0.0298	0.0469	1.0072
		19	43	-0.0107	0.1672	0.8400
		18	42	-0.0302	0.1812	0.6588
19	41	18	41	-0.0348	0.2680	0.3908
		18	40	-0.0409	0.1777	0.2131
		19	39	-0.0432	0.1319	0.0812
		19	38	-0.0342	0.0548	0.0264
		19	37	-0.0484	0.0000	0.0000
		19	36	-0.0392	0.0000	0.0000
		18	35	-0.0399	0.0000	0.0000
		18	34	-0.0367	0.0000	0.0000
17	33	18	33	-0.0397	0.0000	0.0000
18	32	18	32	-0.0281	0.0000	0.0000
		18	31	-0.0388	0.0000	0.0000
		17	30	-0.0376	0.0000	0.0000
		17	29	-0.0149	0.0000	0.0000
		17	28	-0.0050	0.0000	0.0000
				TOTAL = [3]-1.94	TOTAL = [4]2.51	

1 Negative values indicate ground-water runoff to the river.
2 Positive values indicate recharge induced from the river.
3 Total ground-water runoff upstream of the mouth of the Beaver River, September 1976, for nonpumping conditions.
4 Total recharge induced upstream of the mouth of the Beaver River, September 1976, for a pumping rate of 5 million gallons per day.

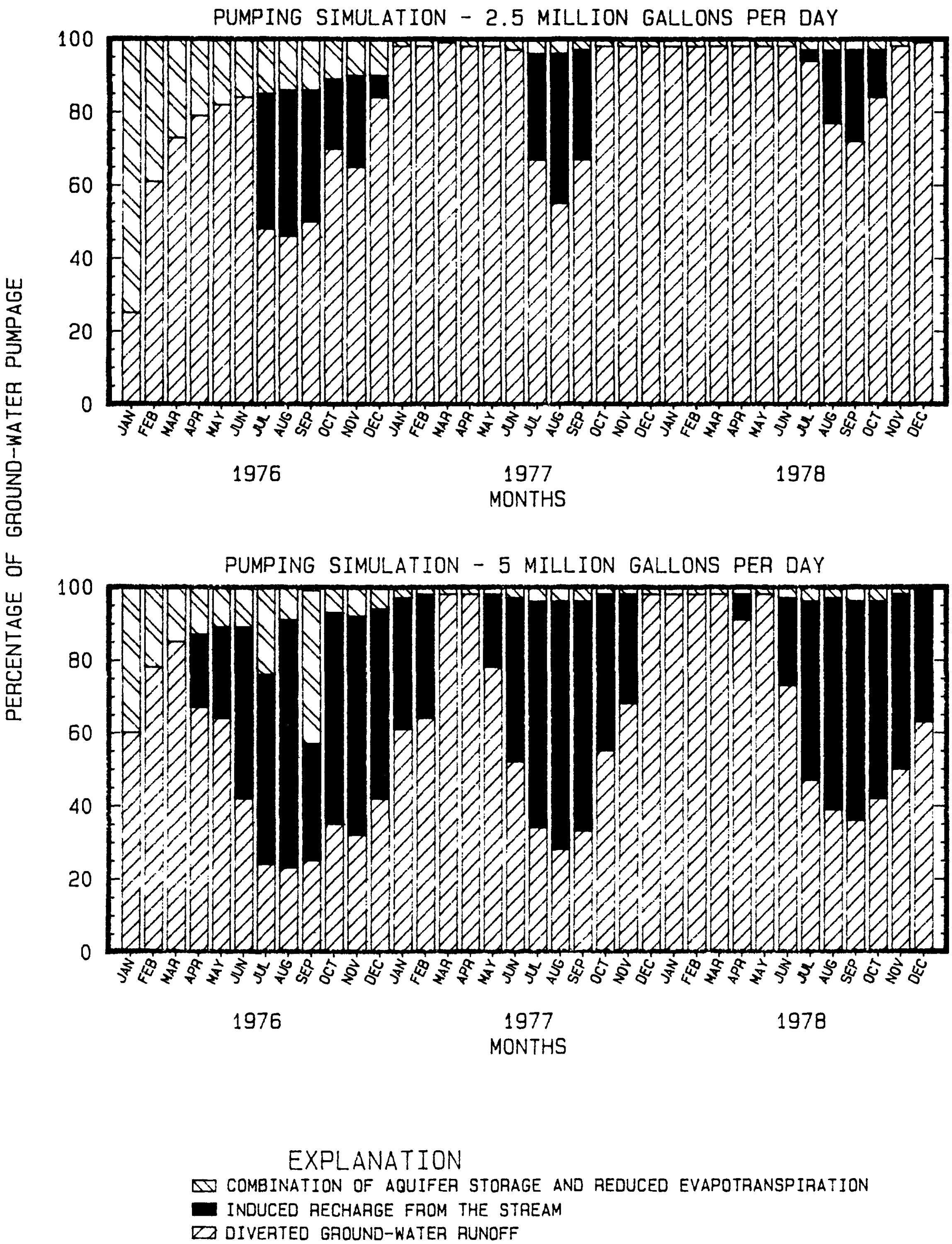


Figure 39.--Percentage of pumpage derived from diverted ground-water runoff, induced stream recharge, and the combination of aquifer storage and reduced evapotranspiration for simulated pumping of 2.5 and 5 Mgal/d from 5 wells upstream of the mouth of the Beaver River.

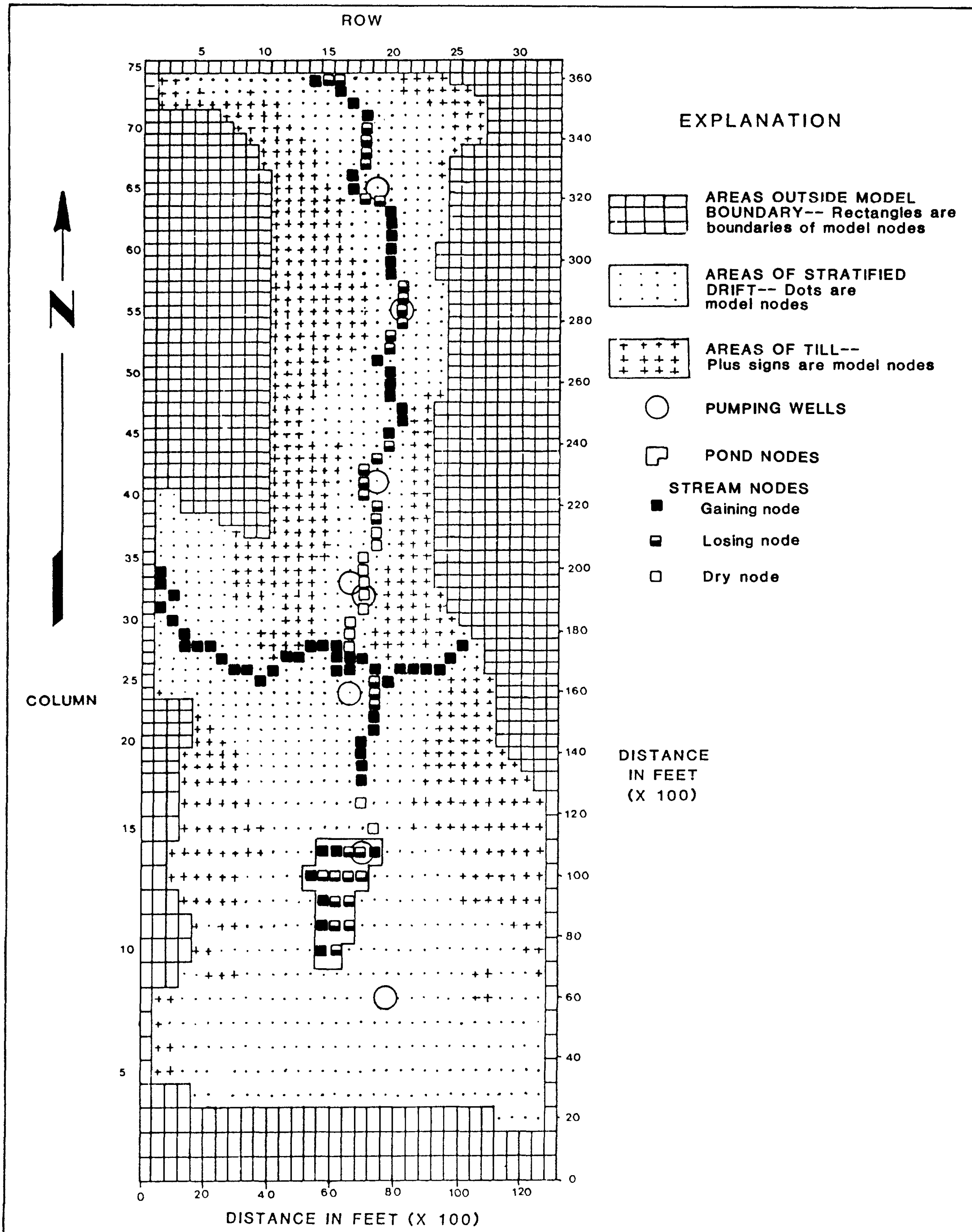


Figure 40.--Ground-water pumpage effect of 7 Mgal/d on stream and pond nodes, September 1976.

Table 13 summarizes the effects of simulating cumulative ground-water pumpage (5.00 Mgal/d) upstream of the mouth of the Beaver River. Table 13 shows that the 5.00 Mgal/d ground-water pumpage for September 1976 was made up of 25 percent ground-water runoff, 32 percent induced recharge, and 42 percent ground-water storage and reduced evapotranspiration. For a detailed breakdown of ground-water pumpage by pumping site see table 14. For each of the four pumping sites listed in table 14, values derived from model output shown in table 12 were totaled for each pumping node within each cone of pumping influence. Figure 41 identified those river nodes that were contributing water to wells. River nodes were used to compute pumpage contributions derived from ground-water runoff, induced recharge, and the combination of aquifer storage and reduced evapotranspiration. Figure 41 shows simulated drawdown over the entire Beaver-Pasquiset model area for September 1976 under maximum combined pumping of 7.00 Mgal/d. Table 14 shows that during September 1976, induced recharge provides from 0 to 94 percent of total pumpage from individual wells, ground-water runoff provides 3 to 25 percent, and the combination of aquifer storage and evapotranspiration provides from 3 to 89 percent.

Table 13.--Summary for the month of September 1976, showing the source of water to wells in the Beaver River valley for simulated pumping at a combined rate of 5 million gallons per day. [Mgal/d: million gallons per day]

GROUND-WATER PUMPAGE

Source	Amount Mgal/d	Percent
(A) Ground-water runoff	1.26	25
(B) Induced recharge	1.62	32
(C) Aquifer storage and reduced evapotranspiration (C=D-A+B)	2.12	42
(D) Total ground-water pumpage	5.00	100

Table 14.--Summary showing source, amount, and percent of water diverted to each pumping site along the Beaver River, pumping 5 million gallons per day, September 1976. [Mgal/d: million gallons per day]

GROUND-WATER PUMPAGE

Pumping site	Model node Row	Model node Column	Source	Amount Mgal/d	Percent
1	19	65	Ground-water runoff		25
			Induced recharge		19
			Aquifer storage and reduced evapo-transpiration		56
			Ground-water pumpage	0.50	
2	21	55	Ground-water runoff		3
			Induced recharge		94
			Aquifer storage and reduced evapo-transpiration		3
			Ground-water pumpage	1.50	
3	19	41	Ground-water runoff		14
			Induced recharge		66
			Aquifer storage and reduced evapo-transpiration		20
			Ground-water pumpage	1.00	
4	17	33	Ground-water runoff		11
	18	32	Induced recharge		0
			Aquifer storage and reduced evapo-transpiration		89
			Ground-water Pumpage	1.50 0.50	
TOTAL COMBINED PUMPAGE				5.00	

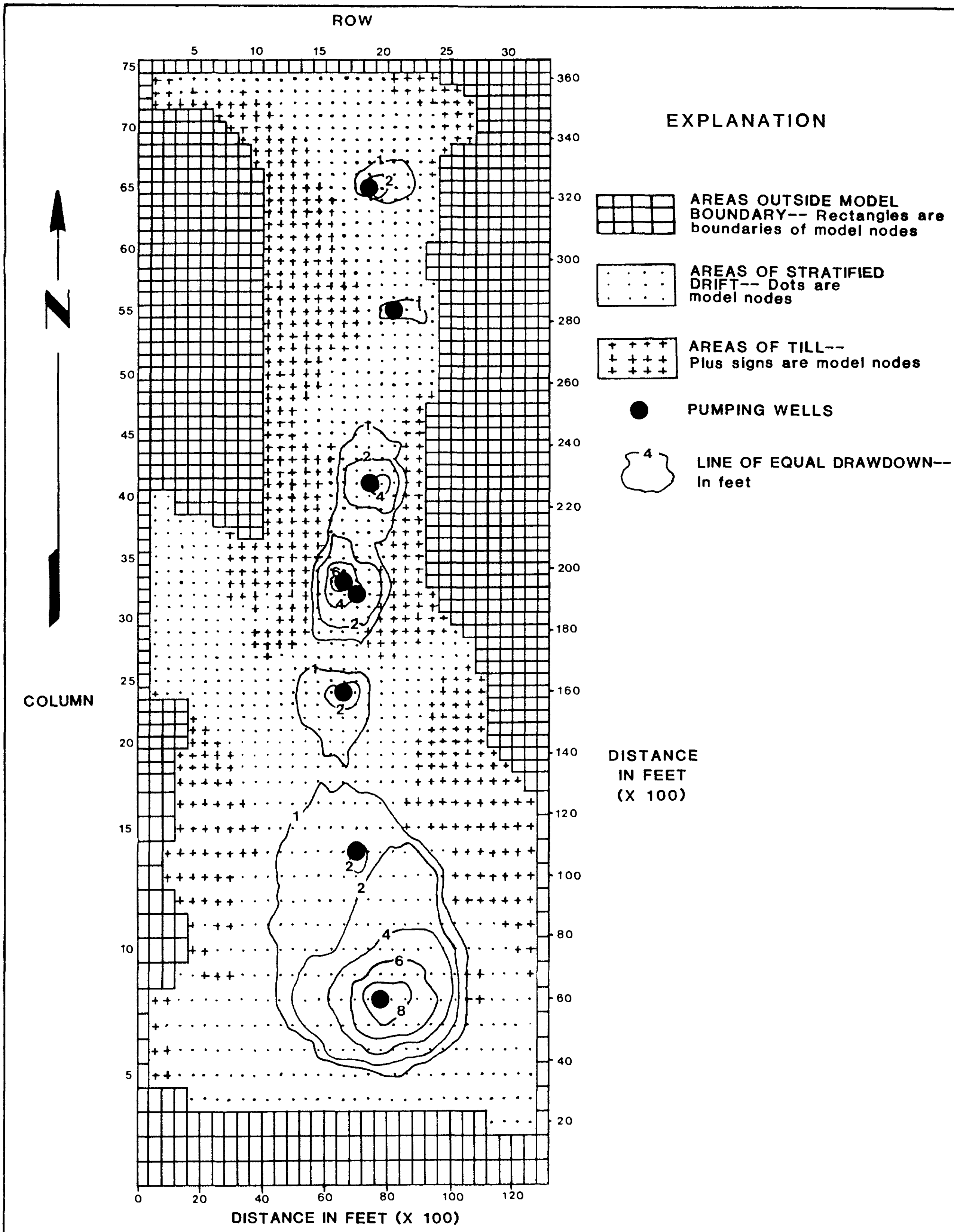


Figure 41.--Simulated drawdown pumping 7 Mgal/d, September 1976.

Based on the four criteria established earlier under the objectives of transient pumping simulations, the Beaver-Pasquiset ground-water reservoir is capable of sustaining a pumping rate of 4.25 Mgal/d during average or above average years of ground-water recharge with minimal impact on ground-water levels, pond levels, and streamflow depletion. However, pumpage of 7.00 Mgal/d probably cannot be sustained on a continuous basis from the Beaver-Pasquiset ground-water reservoir without causing streamflow at the mouth of the Beaver River to drop below the estimated 98-percent duration for several consecutive months during an average year of ground-water recharge. Continuous pumping at 7.00 Mgal/d will also probably cause flow to cease at some stream reaches near the mouth of the Beaver River and along Pasquiset Brook near the pond outlet. Nevertheless, a pumping rate of 7.00 Mgal/d could be maintained for short periods of time (less than 12 hours per day) during average years of recharge with minimal impact on the hydrologic system. During no simulation was the altitude of Pasquiset Pond lowered 4 feet below an altitude of 88 feet above sea level, nor was the initial aquifer saturated thickness depleted 25 percent.

Hypothetical Ground-water Pumpage during 1963-66 Drought Period

The 1963-66 period represents the lowest consecutive four years of precipitation recorded at the National Weather Service Station at Kingston, Rhode Island, since the station began operation in 1889. The 1963-66 period was considered representative of extreme drought conditions. This period was selected to test the Beaver-Pasquiset ground-water reservoir's capacity to sustain an average daily yield of 3.25 Mgal/d during extreme drought conditions.

Ground-water level and streamflow data are not available for the Beaver-Pasquiset area during the period 1963-66. Therefore, it was necessary to utilize estimated ground-water recharge for 1963-66 (see table 4) as a stress and compare estimated monthly total runoff with predicted monthly ground-water runoff between the upper and lower gaging stations along the Beaver River. Ground-water recharge for 1963-66 was estimated using the relationship shown in figure 13 in the natural recharge section of this report. Estimates of monthly total runoff for the Beaver River upper gage were made using the relationship established earlier in figure 3 (see Surface water section).

Heads in the aquifer during the fall of 1962 were not known, and starting heads were derived from steady-state calibration which represent long-term average conditions. Using starting heads from the final steady state run should be a reasonable approach since 1976 and 1963 were average years of measured precipitation.

Estimated inflow from the Beaver River along the upstream model boundary, the 1963 estimated recharge rate (17.61 inches), and the 40-year average evapotranspiration rate (24.58 inches) were input to the model on a monthly basis. The Beaver-Pasquiset model was run until the aquifer equilibrated under the new recharge condition and steady-state was reached. Using the ground-water heads obtained from the steady state response, ground-water recharge shown in table 4 was applied to simulate the 1963-66 period under nonpumping conditions. Comparison of estimated monthly total runoff and predicted monthly ground-water runoff between the upper and lower gaging stations along the Beaver River, January 1963 to December 1966, is shown in figure 42 for nonpumping conditions. The model could not be calibrated for the 1963-66 drought period due to lack of measured streamflow and water level data. Nevertheless, based on the comparison shown in figure 42, the transient model for the 1963-66 drought period was considered acceptable to use for pumping simulations. Simulations for the 1963-66 drought are an approximation of what might happen in the real world system with recharge and withdrawals similar to that used in the model.

Figure 43 and 44 show simulated transient water table contours for December 1963 and 1966 for nonpumping conditions. These figures are given to compare changes in ground-water head for development alternative five.

Pumping objectives and withdrawal sites tested during simulations for the 1963-66 drought period were the same as for the 1976-78 wet period. Only one development alternative was tested for the 1963-66 period (alternative 5 in table 10). Development alternative five (1963-66) involved the same hypothetical pumpage as alternative one (1976-78) with total pumping of 3.25 Mgal/d, 2.50 Mgal/d from the Beaver River valley and 0.75 Mgal/d from the Pasquiset Brook valley.

The impact of simulated ground-water withdrawals of 2.50 Mgal/d on streamflow depletion upstream of the mouth of the Beaver River is shown in figures 45 and 46. Figure 45 shows that withdrawal of 2.50 Mgal/d causes average monthly stream discharge to drop below the estimated 98-percent flow during 2 months in 1964, 6 months in 1965, and 2 months in 1966.

Figure 46 shows that total ground-water runoff available upstream of the mouth of the Beaver River was being diverted to wells 30 out of 48 months. It also shows that a large percentage of pumpage is derived from induced recharge (stippled pattern) during the last two years (1965-66) of the simulation period.

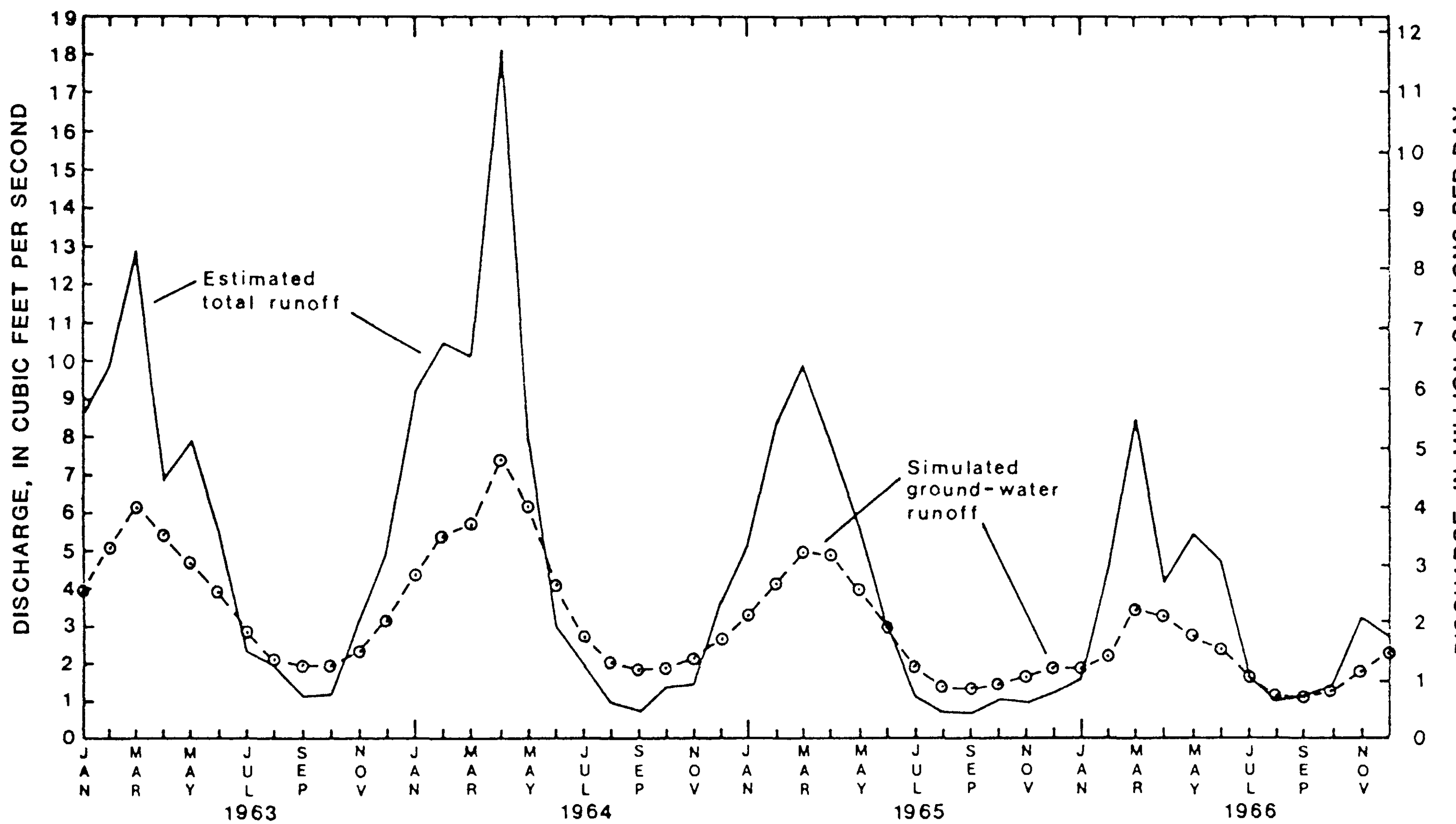


Figure 42.--Comparison of estimated monthly runoff and simulated monthly ground-water runoff between the upper and lower gaging stations along the Beaver River, January 1963-December 1966, for nonpumping conditions.

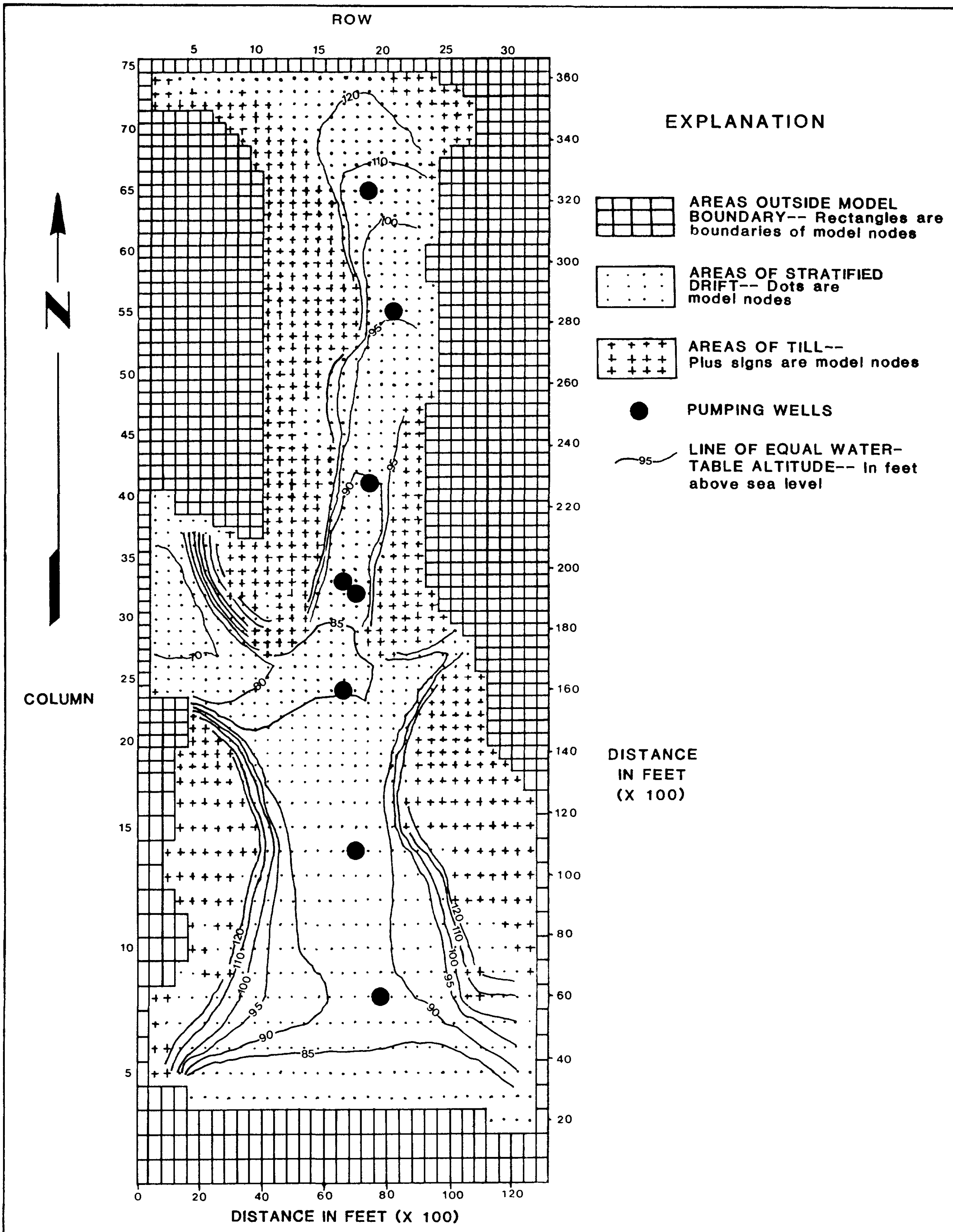


Figure 43.--Simulated water table at the end of December 1963 for nonpumping conditions.

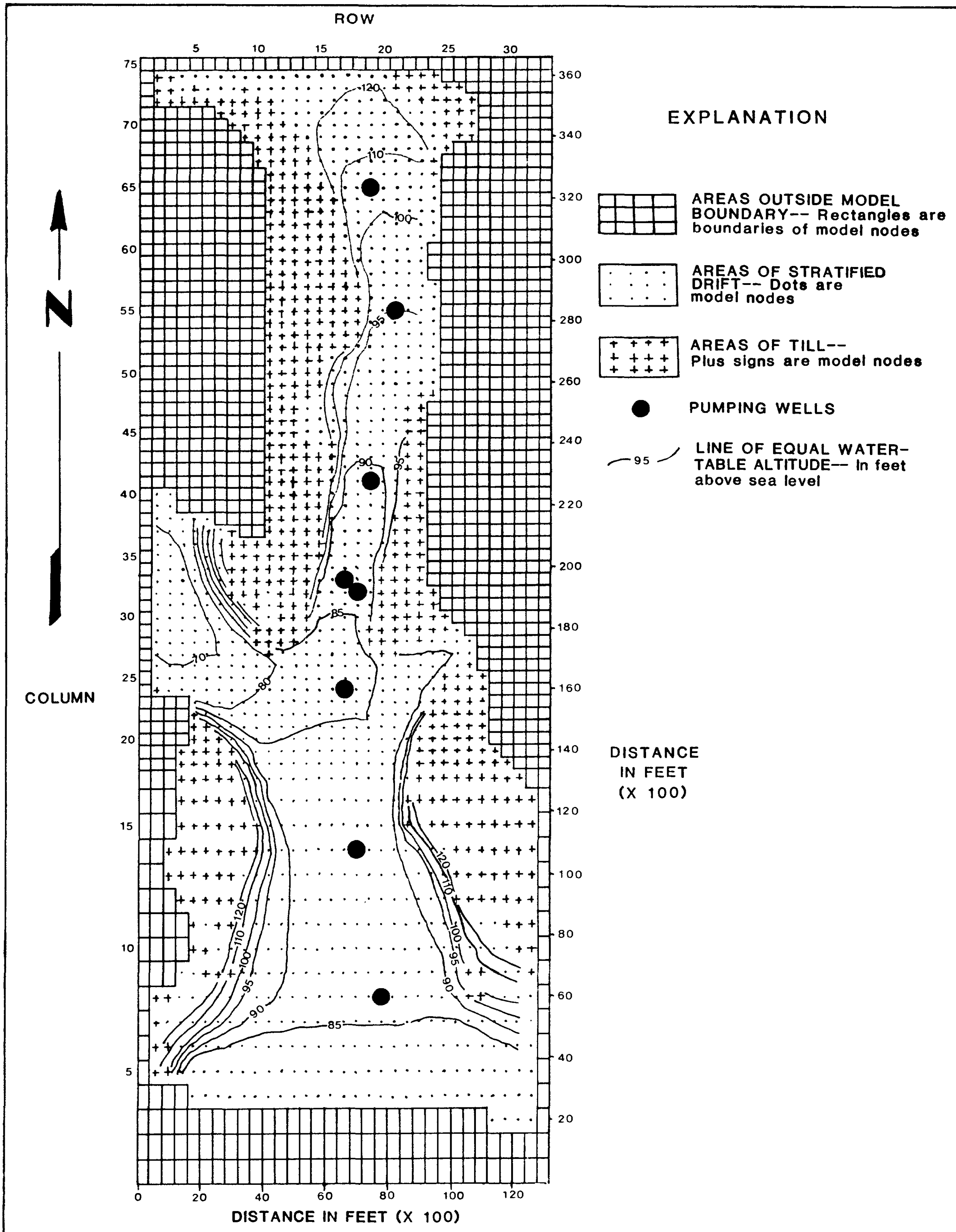


Figure 44.--Simulated water table at the end of December 1966 for nonpumping conditions.

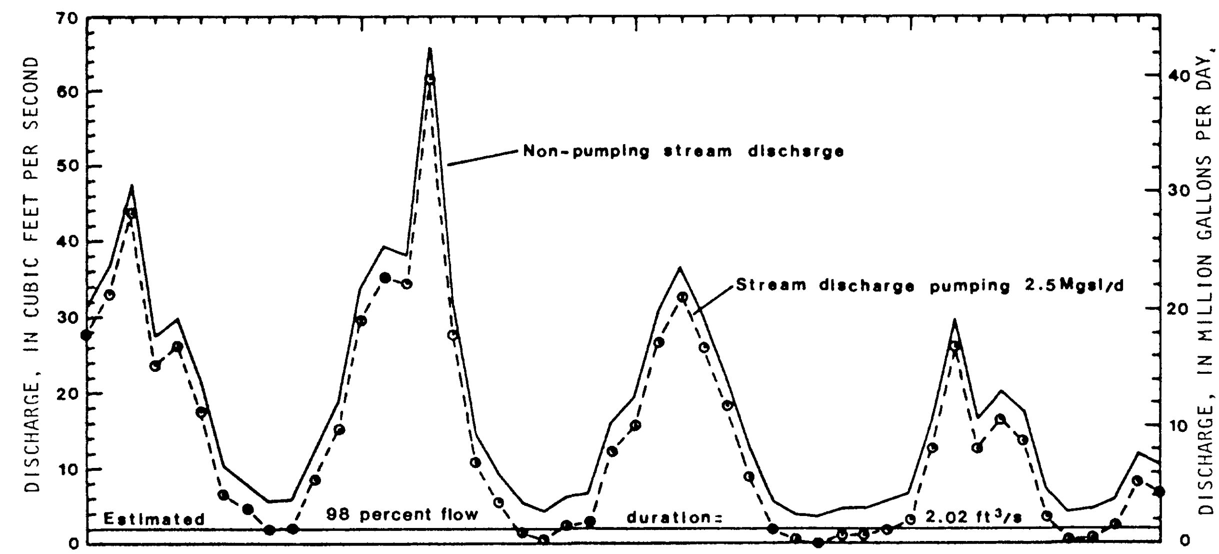


Figure 45.--Comparison of stream discharge at the mouth of the Beaver River for nonpumping and pumping (2.5 Mgal/d) simulations for the 1963-66 drought period.

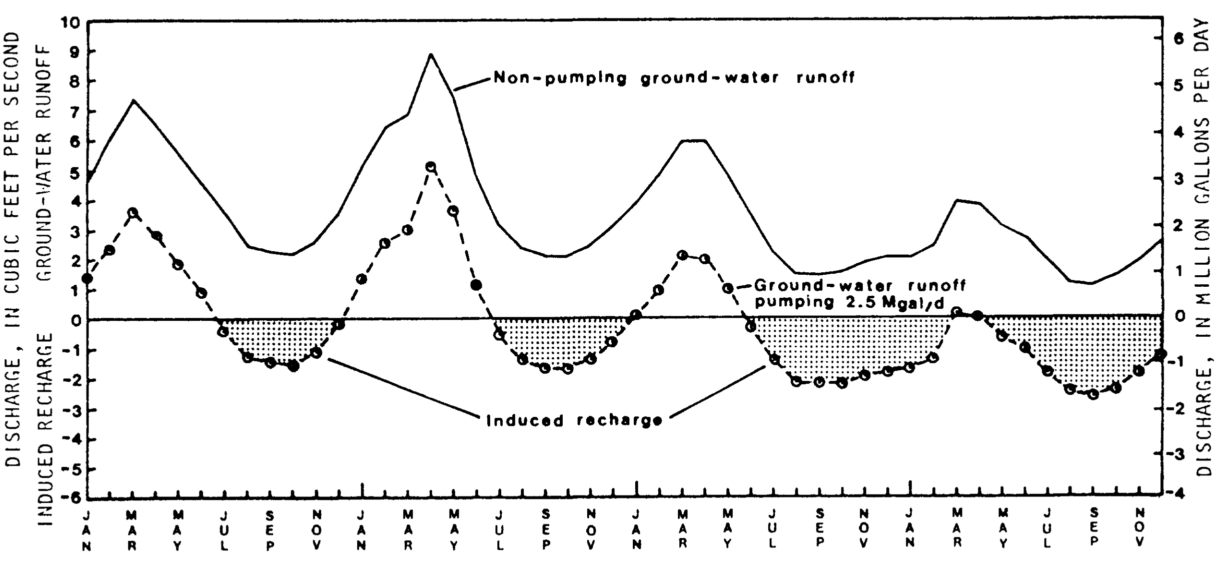


Figure 46.--Comparison of ground-water runoff and induced recharge at the mouth of the Beaver River for nonpumping and pumping (2.5 Mgal/d) simulations for the 1963-66 drought period.

September is the month that shows the greatest impact of ground-water withdrawals on the stream-pond-aquifer system during both wet (September 1976) and drought (September 1965) simulation periods. During the 1963-66 pumping simulation period, the Beaver River went dry over a 1,200-foot reach (2 nodes) at its mouth during September 1965. Figure 47 compares changes in stream discharge profiles along the Beaver River during September 1965 and 1976. The figure shows downstream reduction in total streamflow, and stream sections where flow drops below the estimated 98-percent flow duration due to simulated ground-water pumpage of 2.50 Mgal/d. Predicted streamflow at the mouth of the Beaver River could be increased above the estimated 98-percent flow duration for September 1976 by shifting some pumpage from RIW 400 and RIW 415 to withdrawal sites further upstream. However, increasing streamflow above the estimated 98-percent flow duration for September 1965 probably could not be done alone by shifting pumpage upstream and would require some reduction in pumpage from the Beaver River valley.

Pumpage of 0.50 Mgal/d from the Pasquiset Brook valley, for the 1963-66 simulation period, caused 1,600 (2 nodes) to 2,400-foot (3 nodes) reaches of Pasquiset Brook at the Pasquiset Pond outlet to go dry from August 1965 through January 1966 and July through December 1966. During extreme drought periods it may become acceptable, depending on water supply demand, to dry up some reaches of Pasquiset Brook since they almost go dry naturally during low flow periods. However, if it becomes unacceptable to dry up any reach of Pasquiset Brook during extreme drought periods, like 1965 and 1966, then it will be necessary to reduce pumpage below 3.25 Mgal/d.

It is important to note that streamflows input to and calculated by the model are average monthly values. However, in the actual stream, some days in a month have more flow than other days. Therefore, under pumping conditions streams may go dry only during certain days of the month instead of the whole month as predicted by the model.

The impact of pumping 0.50 Mgal/d from the Pasquiset Brook valley for the 1963-66 simulation period on the altitude of Pasquiset Pond is shown in figure 48. It shows the altitude of the pond declining below an altitude of 88 feet above sea level by an average of only 0.10 foot between January 1963 and July 1965, but declining by as much as 0.80 foot in October through December 1965 and by 1.80 feet in October 1966.

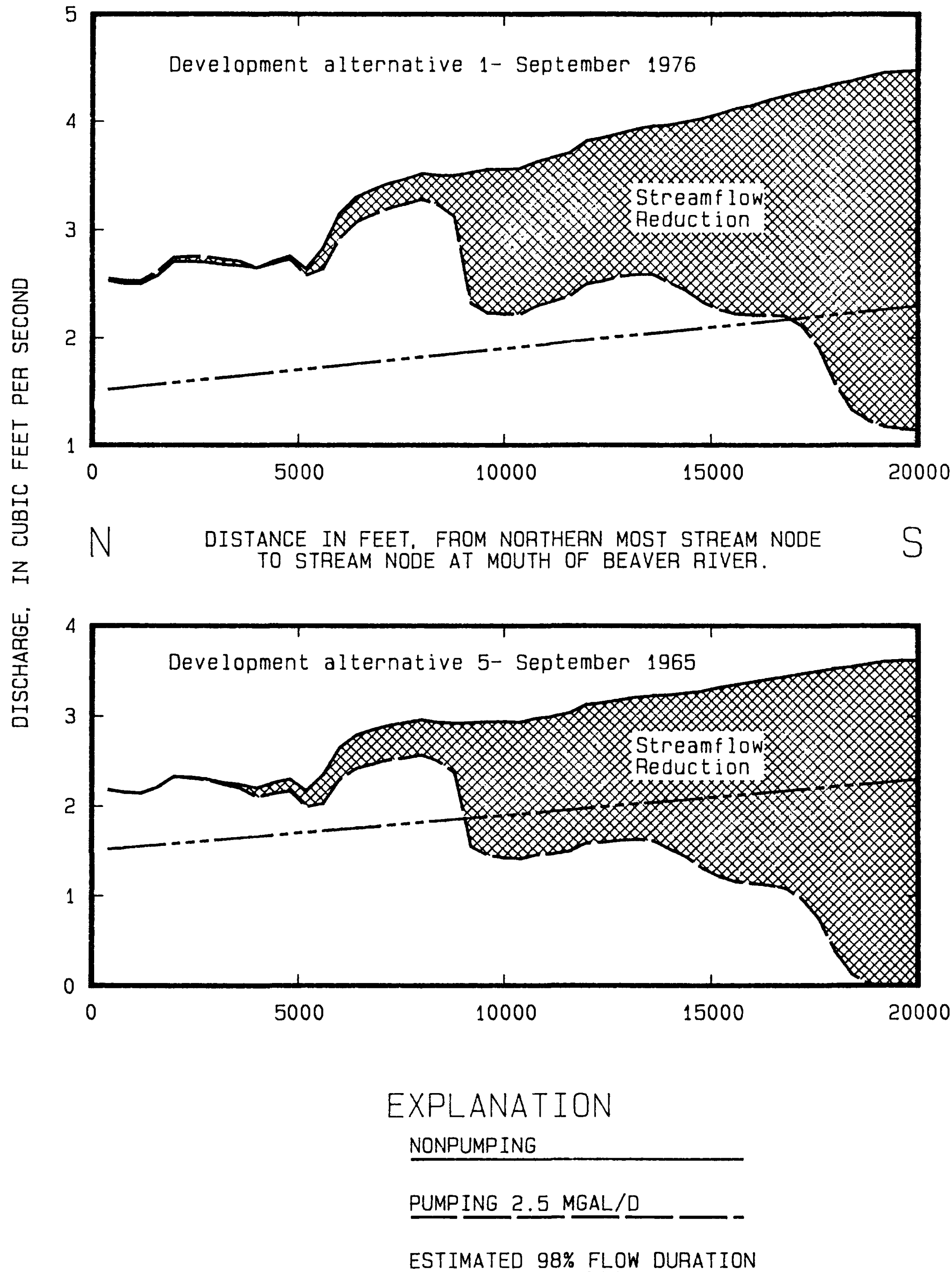


Figure 47.--Comparison of stream discharge profiles along the Beaver River showing reduction in streamflow due to simulated ground-water pumpage for September 1965 and 1976.

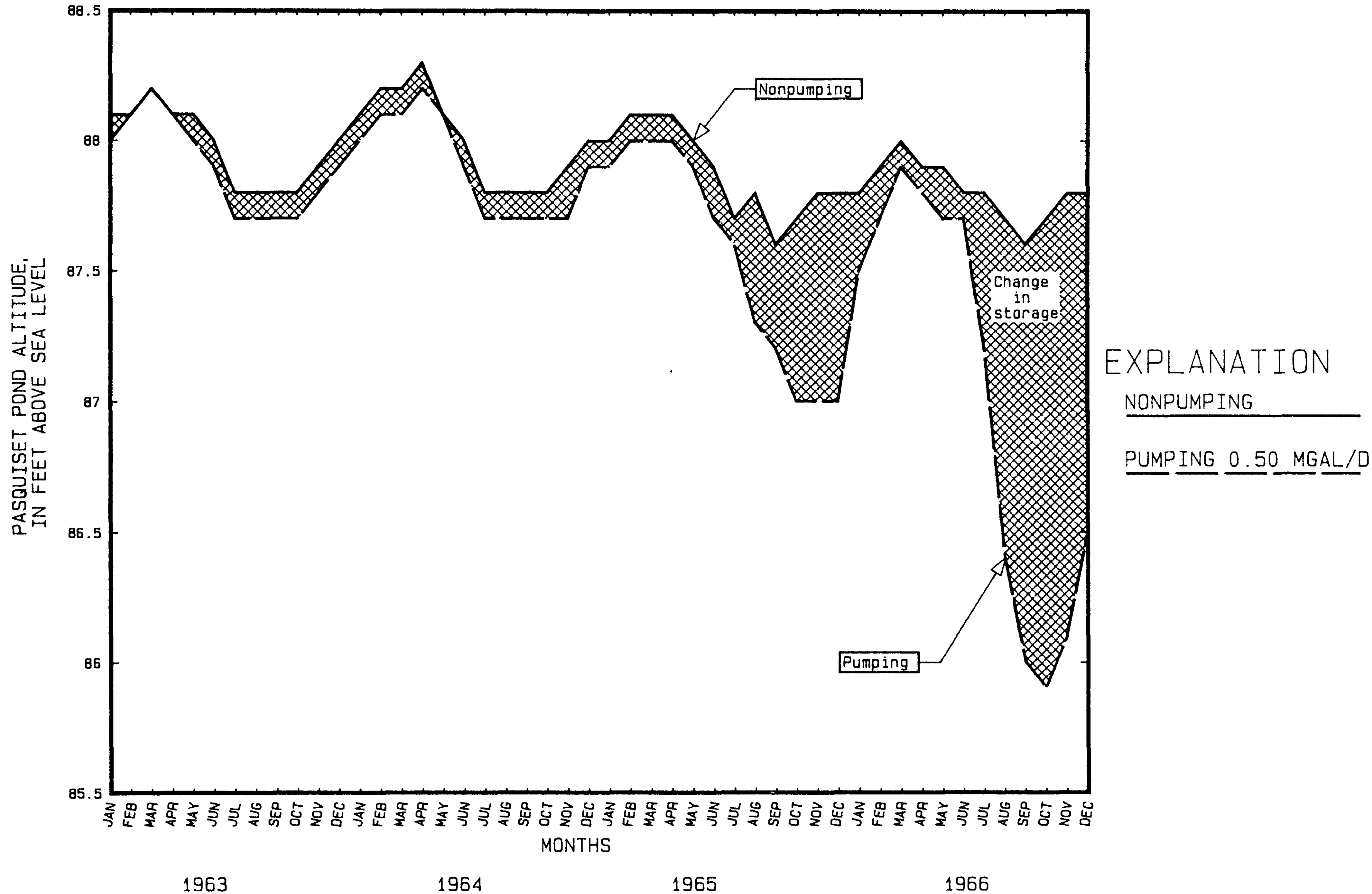


Figure 48.--Comparison of changes in Pasquiset Pond altitude for nonpumping and pumping simulations for 1963-66 drought period showing effects of nearby ground-water pumpage on pond storage.

An important point concerning lowering the level of Pasquiset Pond below an altitude of 88 feet above sea level is that during low precipitation years (1965 and 1966) model predictions are conservative. Predictions are conservative because ground-water outflow from the model's southern boundary were obtained from steady-state calibration for 16.5 inches of long-term average annual ground-water recharge. These values also do not vary with time. For example, predicted water levels for 1966 would be on the low side because 8.0 inches of recharge was applied to the model while 16.5 inches was taken out of the model area along the southern boundary. Therefore, it may be possible to pump more ground water from the Pasquiset Brook valley during drought years similar to 1965 and 1966 with less lowering of the level of Pasquiset Pond than the model predicts. This additional ground-water pumpage could be diverted from the Beaver River valley thereby possibly keeping flow in the Beaver River above the estimated 98-percent flow duration at all times.

Simulated transient water table contours for December 1963 and 1966 are given in figures 49 and 50 for development alternative five. Figure 49 and 50 are given for comparing pumping water table contours with nonpumping water table contours shown earlier in figures 43 and 44.

The user of this report is cautioned that model predictions of streamflow depletion in development alternative five are approximate. Values may be slightly more or less than shown due to input of ground-water recharge to the model with a standard error of estimate of ± 0.77 inches.

SUMMARY AND CONCLUSIONS

This report describes the hydrogeology and water quality of the Beaver-Pasquiset ground-water reservoir, and presents results of transient simulations of ground-water flow. A U.S. Geological Survey two-dimensional model, developed by Trescott and others (1976) and modified by Ozbilgin and Dickerman (1984), was used to simulate interactions between the stream, pond, and aquifer. Transient simulations show the impact of ground-water development alternatives on ground-water levels, pond levels, and streamflow depletion.

The Beaver-Pasquiset study area is in southern Rhode Island and is located within the Pawcatuck River basin. Stratified drift is the only geologic unit capable of producing well yields greater than 350 gal/min. The saturated thickness of the stratified-drift aquifer averages 60 to 80 feet and has a maximum known satuated thickness of 120 feet. Aquifer transmissivity ranges from 7,200 to 24,300 ft^2/d and horizontal hydraulic conductivity ranges from 135 to 300 ft/d. Revised maps of aquifer saturated thickness and transmissivity are provided on figures 12 and 16. Vertical hydraulic conductivity ranges from 0.40 to 15.2 ft/d. Unconfined conditions prevail in the aquifer, which is hydraulically connected with perennial streams and ponds. Annual precipitation from 1941-80 averaged 46.17 inches and long-term average annual ground-water recharge is estimated to be 16.6 inches.

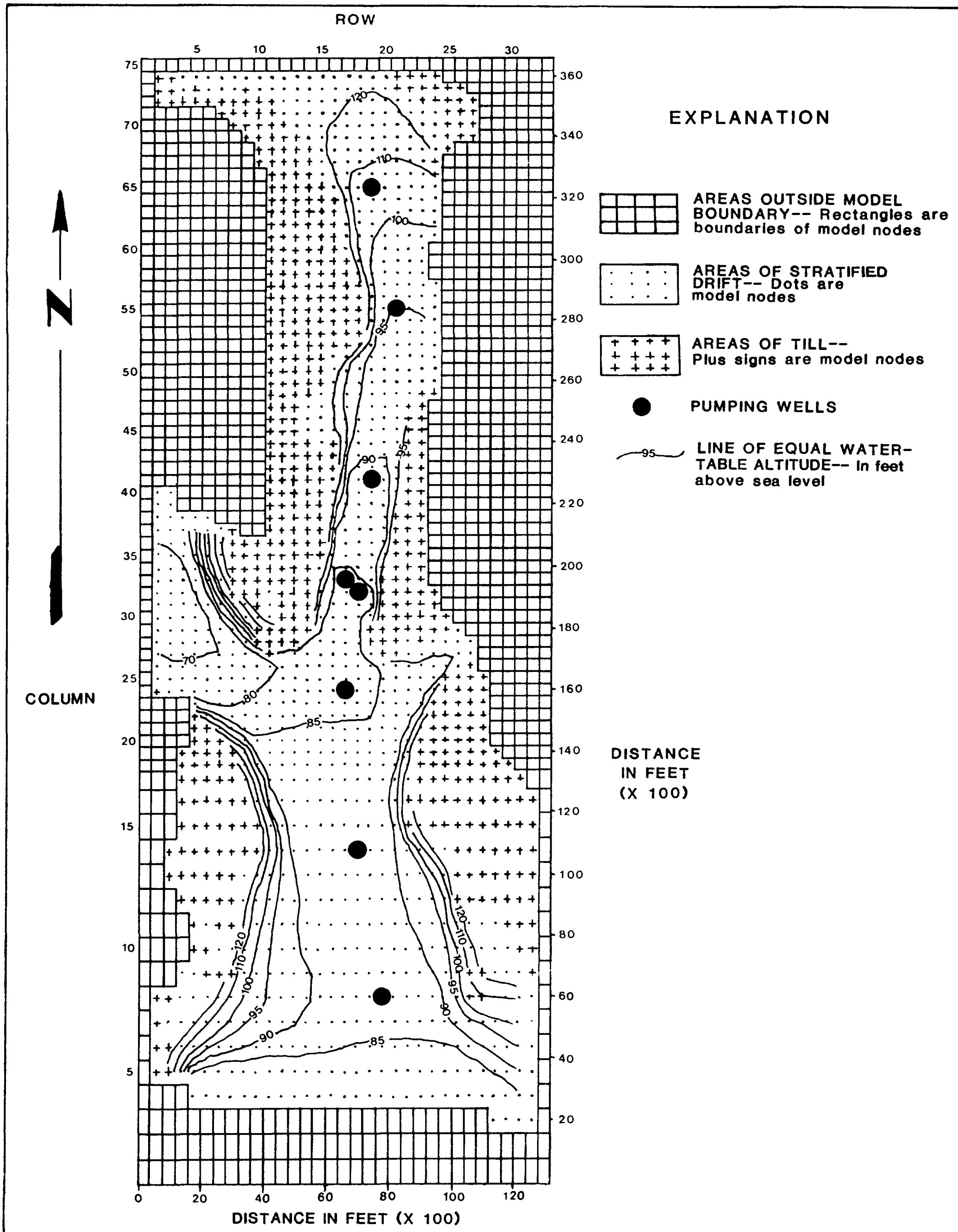


Figure 49.--Simulated water table at the end of December 1963 for development alternative 5 pumping 3.25 Mgal/d.

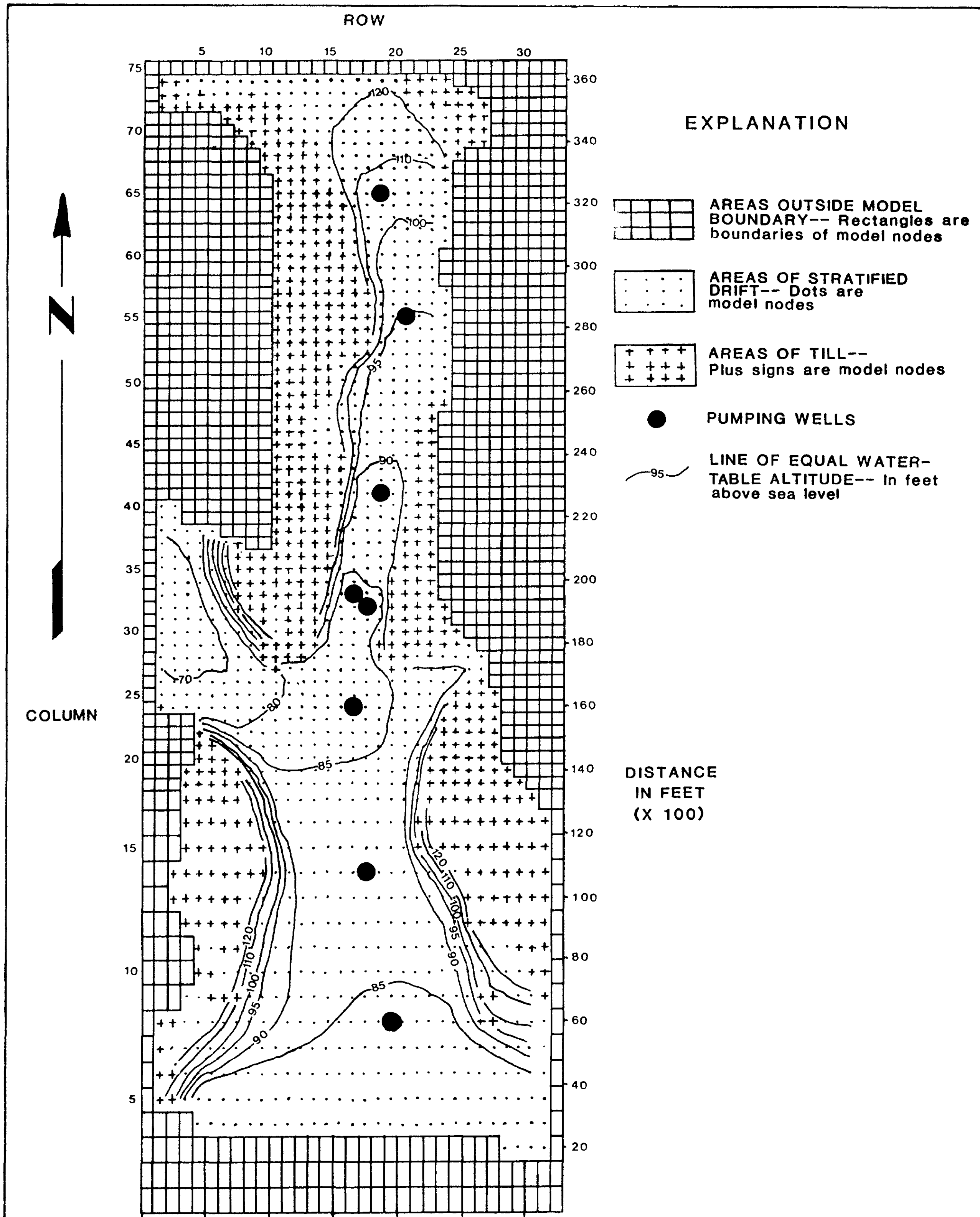


Figure 50.--Simulated water table at the end of December 1966 for development alternative 5 pumping 3.25 Mgal/d.

The areas most favorable for development of high-capacity wells (350 gal/min or more) are along the Beaver River and near Pasquiset Pond. Most water withdrawn from wells in the stratified-drift aquifer will be derived from induced recharge from the Beaver River and Pasquiset Pond. Streambed materials are generally composed of loosely packed sand and gravel that were assumed to have higher vertical hydraulic conductivities than the least permeable layer in the underlying aquifer.

The chemical quality of water in the study area is suitable for most purposes. The water is soft, somewhat corrosive, and generally contains less than 100 mg/L dissolved solids. The only water quality problem appears to be southeast of Pasquiset Pond where iron and manganese concentrations are above the USEPA maximum limits for public water systems. Ground water in this area will require treatment for iron and manganese removal if used for drinking.

The Beaver-Pasquiset ground-water reservoir was simulated with a two-dimensional finite-difference model using a block-centered, rectangular finite-difference grid. The grid network for the model consists of 33 rows and 75 columns, and defines 2,475 nodes. The numerical method used in the model is the strongly implicit procedure. Model boundaries coincide as closely as possible with hydrologic boundaries, and were treated as constant-flux, leaky constant source head, or leaky. The Beaver River, Pasquiset Brook, and Pasquiset Pond were simulated as leaky boundaries to represent the interaction of the stream-pond-aquifer system. Effective streambed hydraulic conductivities used in the model ranged from 0.40 to 15.2 feet per day.

Sensitivity analyses were made to determine the effect of changing streambed hydraulic conductivity, specific yield, and transmissivity in the model. Results show that ground-water runoff is most sensitive to large decreases in streambed hydraulic conductivity and that head in the aquifer is most sensitive to decreases in transmissivity.

Long-term historical records were not available for the study area and it was necessary to utilize estimated long-term average annual ground-water recharge as a stress during the steady state calibration. Heads from the final steady state run were used as the initial condition for transient simulations. Differences between measured and simulated water-table altitudes for the final steady state run for 21 selected observation wells averaged 0.07 feet. Acceptability of the transient model was determined by comparing estimated and predicted ground-water runoff and average monthly water levels for the 1976-78 period. Combined pumping rates for simulation of ground-water development alternatives at eight sites ranged from 3.25 to 7.00 Mgal/d. Pumping rates for individual wells ranged from 0.25 to 1.50 Mgal/d.

Transient simulations suggest that the Beaver-Pasquiset ground-water reservoir is capable of sustaining a pumping rate of 4.25 Mgal/d during average or above average years of ground-water recharge with minimal impact on ground-water levels, pond levels, and streamflow. However, pumpage of 7.00 Mgal/d probably cannot be sustained on a continuous basis without causing streamflow at the mouth of the Beaver River to drop below the estimated 98-percent duration for several months during a year of average recharge. Continuous pumping at 7.00 Mgal/d will also probably cause flow to cease at some stream reaches near the mouth of the Beaver River and along Pasquiset Brook near the pond outlet. However, a pumping rate of 7.00 Mgal/d could be maintained for short periods of time (less than 12 hours) during years of average recharge with minimal impact on the hydrologic system. During extreme drought periods, such as 1965 and 1966, it would be necessary to reduce pumpage below 3.25 Mgal/d in order to maintain continuous flow in both the Beaver River and Pasquiset Brook.

Simulations show the maximum decline in altitude of Pasquiset Pond, below a surface altitude of 88 feet above sea level, to be 2.2 feet in September 1977. It also shows the overall size of the pond declining by an average of 1.5 feet under maximum pumping for the 1976-78 simulation period.

In the stratified-drift aquifer, maximum drawdown did not exceed 12 percent of initial aquifer saturated thickness (100 feet) during any pumping simulation. Maximum simulated drawdown was 9.2 feet in the Beaver River valley and 11.2 feet in the Pasquiset Brook valley.

Transient-model output was used to make monthly determinations of the amount of ground-water pumpage derived from ground-water runoff, induced recharge, and the combination of aquifer storage and reduced evapotranspiration. Monthly determinations were made for pumping simulations of 2.50 and 5.00 Mgal/d upstream of the mouth of the Beaver River for the 1976-78 period. Utilizing transient model output, the model user can readily evaluate stream-aquifer interactions for any selected time period.

Any month could have been chosen to illustrate in detail the effectiveness of the model to simulate surface-ground water interactions. However, September 1976 was chosen because it is the month that shows the greatest impact of ground-water withdrawals on the stream-pond-aquifer system. The model of the Beaver-Pasquiset ground-water reservoir can also be used to simulate intermittent drying and recovery of stream reaches.

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GLOSSARY

ANISOTROPY: That condition in which all significant properties vary with direction.

AQUIFER: A formation, group of formations, or part of a formation that contains enough saturated permeable material to yield significant quantities of water to wells and springs.

AQUIFER TEST: A controlled field experiment wherein the effect of pumping a well is measured in the pumped well and in observation wells for the purpose of determining hydraulic properties of an aquifer.

BEDROCK: The solid rock, commonly called "ledge", that forms the earth's crust.

COLOR: Color is expressed in units of the platinum-cobalt scale proposed by Hazen (1892, p. 427-428). A unit of color is produced by one milligram per liter of platinum in the form of the chloroplatinated ion. The intensity of color is rated numerically from 0 to 500, a color of 5 being equivalent to 1/100 that of the standard. The extent to which a water is colored by material in solution may indicate the presence of organic material that may have some bearing on the dissolved-solids content.

CONCEPTUAL MODEL, of a stream-aquifer system: A general idea or understanding of the concept representing an existing stream-aquifer system, that make it possible to realistically simulate that system mathematically.

CONE OF DEPRESSION: A depression produced in a water table or other potentiometric surface by the withdrawal of water from an aquifer; in cross section, shaped like an inverted cone with its apex at the pumping well.

CONFINED AQUIFER (ARTESIAN AQUIFER): An aquifer in which ground water is confined under pressure significantly greater than atmospheric throughout. See UNCONFINED AQUIFER.

CONSTANT-FLUX BOUNDARY: A constant flux may be zero (impermeable boundary) or have a finite value.

<u>Zero-flux boundary</u>: A model boundary condition that is specified by assigning a value of zero transmissivity to nodes outside the boundary to simulate no flow across the boundary.

<u>Finite-flux boundary</u>: A model boundary condition that is specified by assigning a fixed value of volumetric flow to recharge (or discharge) wells at appropriate nodes to simulate flow across the boundary.

CONTINUOUS-RECORD GAGING STATION: A site on a stream at which continuous measurements of stream stage are made. These records are converted to daily flow after calibration by flow measurements.

DIGITAL MODEL: A simplified mathematical representation of a complex aquifer system. A computer program designed to solve ground-water flow equations.

DISCHARGE: The volume of water that passes a given point within a given period of time.

<u>Mean-discharge</u>: The arithmetic average of individual daily mean discharges during a specific period.

<u>Instantaneous discharge</u>: The discharge at a given time.

DISSOLVED SOLIDS: The residue from a clear sample of water after evaporation and drying for 1 hour at 180oC; consists primarily of dissolved mineral constituents, but may also contain organic matter and water of crystallization.

DRAINAGE AREA: The drainage area of a stream at a specified location is that area, measured in a horizontal plane, which is enclosed by a drainage divide.

DRAINAGE BASIN: A part of the surface of the earth that is occupied by a drainage system, which consists of a surface stream or a body of impounded surface water together with all tributary surface streams and bodies of impounded surface water.

DRAINAGE DIVIDE: The rim of a drainage basin. Drainage divide, or just divide, is used to denote the boundary between one drainage area and another.

DRAWDOWN: The decline of water level in a well after pumping starts. It is the difference between the water level in a well after pumping starts and the water level as it would have been if pumping had not started.

DURATION OF FLOW, of a stream: The percentage of time during which specified daily discharges have been equaled or exceeded in magnitude within a given time period.

EVAPOTRANSPIRATION: Water withdrawn from a land area by evaporation from water surfaces and moist soil and plant transpiration.

GAGING STATION: A particular site on a stream, canal, lake, or reservoir where systematic observations of hydrologic data are obtained.

GAINING STREAM: A stream or reach of a stream whose flow is being increased by inflow of ground water.

GRAVEL PACKED WELL: A well in which filter material is placed in the annular space to increase the effective diameter of the well, and to prevent fine-grained sediments from entering the well.

GROUNDWATER: Water in the ground that is in the zone of saturation, from which wells, springs, and ground-water runoff are supplied.

GROUND-WATER DISCHARGE: The discharge of water from the saturated zone by (1) natural processes such as ground-water runoff and ground-water evapotranspiration and (2) discharge through wells and other man-made structures.

GROUND-WATER DIVIDE: A line on a water table on each side of which the water table slopes downward in a direction away from the line. It is analogous to a divide between two drainage basins on a land surface. Generally a ground-water divide is found nearly below a surface-drainage divide, but in many localities there is no relation between the two.

GROUND-WATER EVAPOTRANSPIRATION: Ground water discharged into the atmosphere in the gaseous state by direct evaporation and by transpiration by plants.

GROUND-WATER OUTFLOW: That part of the discharge from a drainage basin that occurs through the ground. The term "underflow" is often used to describe ground-water outflow.

GROUND-WATER RECHARGE: The amount of water that is added to the saturated zone.

GROUND-WATER RESERVOIR: Parts of the sand and gravel aquifer where ground-water is accumulated under conditions that make it suitable for development and use.

GROUND-WATER RUNOFF: That part of the runoff which has passed into the ground, has become ground water, and has been discharged into a stream channel as spring or seepage water.

HARDNESS: A physical-chemical characteristic that is commonly recognized by the increased quantity of soap required to produce lather. It is attributable to the presence of alkaline earths (principally calcium and magnesium) and is expressed as equivalent calcium carbonate ($CaCO_3$). The following classification is used by the U.S. Geological Survey: soft, 0-60 mg/l; moderately hard, 61-120 mg/l; hard, 121-180 mg/l; very hard, more than 180 mg/l.

HEAD,STATIC: The height above a standard datum of the surface of a column of water (or other liquid) that can be supported by the static pressure at a given point.

HETEROGENEITY: Heterogeneity is synonymous with nonuniformity. A material is heterogeneous if its hydrologic properties vary with position within it.

HYDRAULIC CONDUCTIVITY : The volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow. Expressed herein in feet per day. These values may be converted to gallons per day per square foot by multiplying by 7.48.

HYDRAULIC GRADIENT: The change in static head per unit of distance in a given direction. If not specified, the direction generally is understood to be that of the maximum rate of decrease in head.

INDUCED INFILTRATION: The process by which water moves into an aquifer from an adjacent surface-water body, owing to reversal of the hydraulic gradient, in response to pumping.

INDUCED RECHARGE: The amount of water entering an aquifer from an adjacent surface-water body by the process of induced infiltration.

ISOTROPY: That condition in which all significant properties are independent of direction.

LEAKY BOUNDARY: A boundary condition that relates boundary flux to boundary head. Its most common use is to represent the interaction between a water table aquifer and a stream or river which is separated from the aquifer by a semi-pervious streambed layer.

LITHOLOGIC LOG: Description of geologic material collected during sampling of test wells.

LOSING STREAM: A stream or reach of a stream that is losing water to the ground.

MILLIGRAMS PER LITER (MG/L, mg/l): A unit for expressing the concentration of chemical constituents in solution. Milligrams per liter represents the weight of solute per unit volume of water.

NATIONAL GEODETIC VERTICAL DATUM OF 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called mean sea level. NGVD of 1929 is referred to as sea level in this report.

pH: Symbol denoting relative concentration of hydrogen ions in a solution; pH values range from 0 to 14 -- the lower the value, the more acid the solution; that is, the more hydrogen ions it contains. A value of 7.0 is the neutral point; values greater than 7.0 indicate an alkaline solution; values less than 7.0 indicate an acid condition.

POTENTIAL EVAPOTRANSPIRATION: Water loss that will occur if at no time there is a deficiency of water in the soil for use of vegetation.

PRECIPITATION: The discharge of water from the atmosphere, either in a liquid or solid state.

RUNOFF, TOTAL: Part of precipitation that appears in surface streams. It is the same as streamflow unaffected by artificial diversion, storage, or other works of man in or on stream channels. Includes both surface- and ground-water runoff.

SATURATED THICKNESS: The thickness of an aquifer below the water table. As measured for the stratified-drift aquifer in this report, it is the vertical distance between the water table and the bedrock surface and in places includes till present between the stratified drift and the bedrock surface.

SATURATED ZONE: That part of the water-bearing material in which all voids, large and small, are ideally filled with water under pressure greater than atmospheric.

SPECIFIC CAPACITY: The specific capacity of a well is the rate of discharge of water from the well divided by the drawdown of water level within the well.

SPECIFIC CONDUCTANCE: A measure of the ability of a water to conduct an electrical current, expressed in microsiemens per centimeter at 25°C. Specific conductance is related to the type and concentration of ions in solution and can be used for estimating the dissolved-solids content of the water. Commonly, the concentration of dissolved solids (in milligrams per liter) is about 65 percent of specific conductance (in microsiemens per cm at 25°C). This relation is not constant from stream to stream or from well to well, and it may even vary in the same source with changes in the composition of the water.

SPECIFIC YIELD (Sy): Ratio of the volume of water a fully saturated rock or unconsolidated material will yield by gravity drainage, given sufficient time, to the total volume of rock or unconsolidated material; commonly expressed as percentage.

STEADY STATE: Equilibrium water levels or heads; aquifer storage and water levels do not vary with time.

STORAGE COEFFICIENT: The volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head; commonly expressed as a decimal or percentage. In an unconfined aquifer the storage coefficient is virtually equal to the specific yield.

STRATIFIED DRIFT: Unconsolidated sediment that has been sorted by glacial meltwater and deposited in layers, or strata.

STREAMFLOW: The discharge that occurs in a natural channel. "Streamflow" is more general than "runoff", as streamflow may be applied to discharge whether or not it is affected by diversion or regulation.

TILL: A geologic term for a glacial deposit of predominantly nonsorted, nonstratified material ranging in size from boulders to clay. It is commonly so compact that it is difficult to penetrate with light drilling equipment.

TRANSIENT STATE: Nonequilibrium water levels or heads; water levels and aquifer storage vary with time.

TRANSMISSIVITY: The rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient. It is equal to the product of the hydraulic conductivity and saturated thickness. Expressed herein in feet squared per day.

UNCONFINED AQUIFER (WATER-TABLE AQUIFER): An aquifer in which the upper surface of the saturated zone, the water table, is at atmospheric pressure and is free to rise and fall.

UNDERFLOW: See "GROUND-WATER OUTFLOW".

UNSATURATED ZONE: The zone between the land surface and the water table.

WATER TABLE: The upper surface of the saturated zone.

WATER YEAR: A 12-month period, October 1 through September 30. It is designated by the calendar year in which it ends.

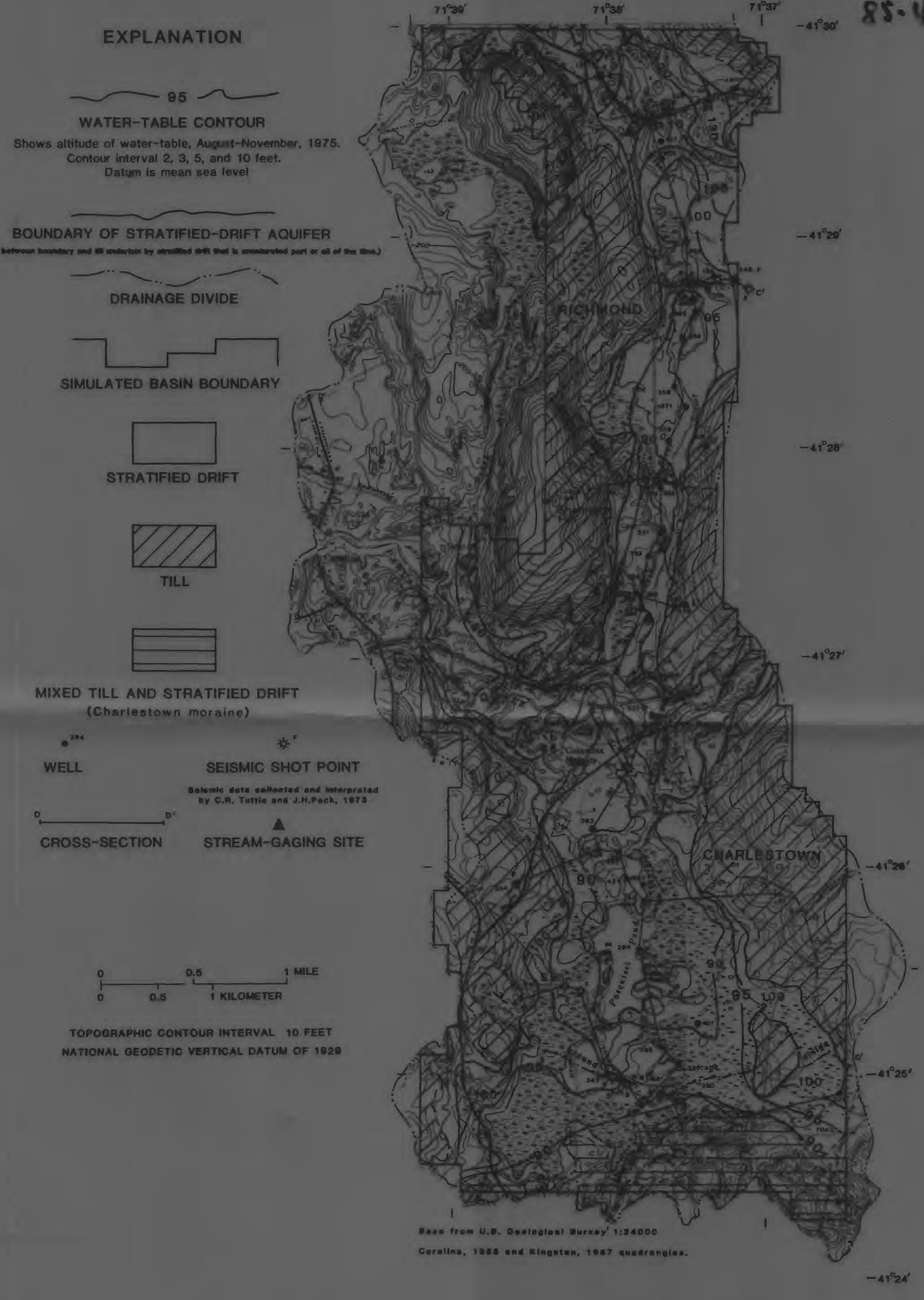

Figure 2.--Map showing configuration of the water table in the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir.

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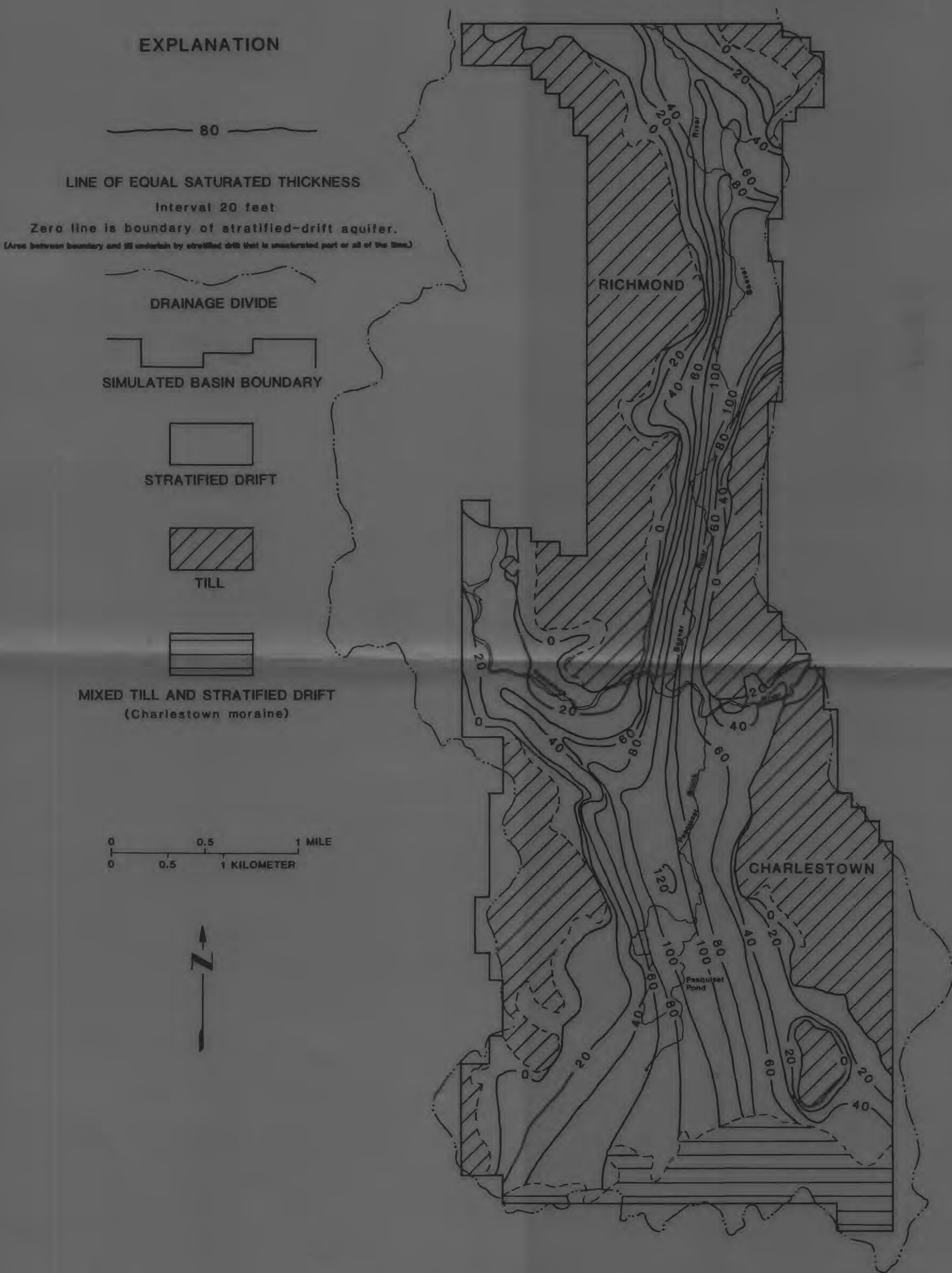

Figure 12.--Map showing configuration of the saturated thickness in the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir .

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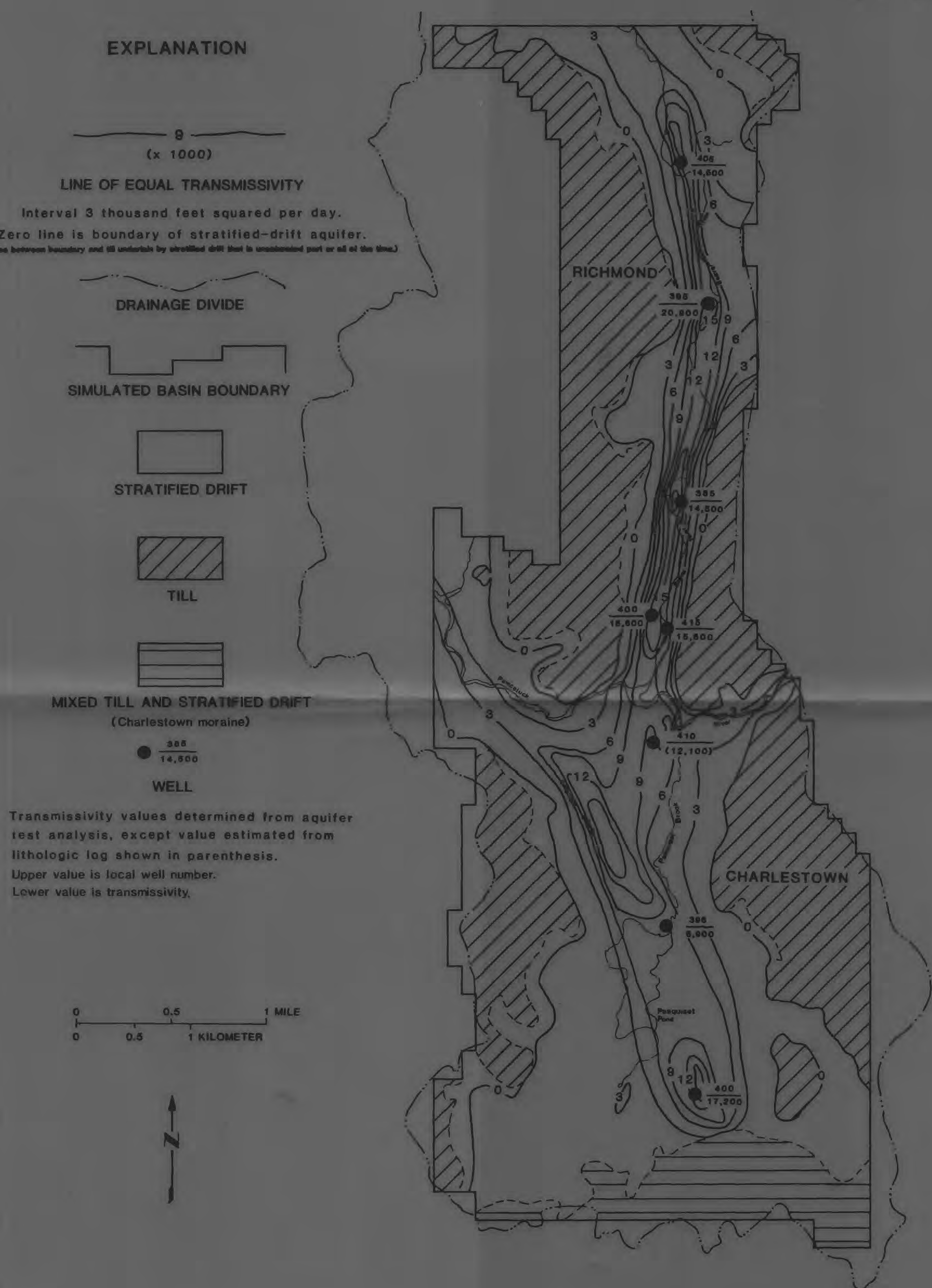

Figure 16.--Map showing configuration of the transmissivity in the stratified-drift aquifer in the Beaver-Pasquiset ground-water reservoir .