1986

Application of a modular three-dimensional finite difference ground-water flow model to a glacial valley fill stream-aquifer system in the Rockaway drainage basin, New Jersey /

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APPLICATION OF A MODULAR THREE-DIMENSIONAL FINITE DIFFERENCE GROUND-WATER FLOW MODEL TO A GLACIAL VALLEY FILL STREAM-AQUIFER SYSTEM IN THE ROCKAWAY DRAINAGE BASIN, NEW JERSEY

by

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A Thesis

Presented to the Graduate Committee

of Lehigh University

in Candidacy for the Degree of

Master of Science

in

Geological Sciences

Lehigh University

1986

This thesis is accepted and approved in partial fulfillment of the degree of Master of Science

September 23, 1986
(date)

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[illegible signature]
Chairman of Department

Acknowledgements

I extend my sincere appreciation and gratitude to my thesis advisor, Dr. Paul Myers, and my thesis committee members, Dr. Gerard P. Lennon, and Mr. Robert Canace for their suggestions and assistance toward the completion of the thesis. A special thanks to Scott Stanford who generously provided me with much of the neccessary geological data base. Also, I would like to thank Phil Harte for supplying me with much of the necessary hydrologic information.

In addition, I thank my fellow graduate students for their support and encouragement throughout my graduate career. I would also like to extend my gratitude to all the faculty members of Lehigh University who have been influential to my education.

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1.0 ABSTRACT

A hydrogeological study of a valley fill aquifer system situated in an integrated pre-glacial and post-glacial drainage network was conducted. The unconsolidated glacial deposits consist of various proportions of pre-Late and Late Wisconsinan sediments which directly overlie Precambrian crystalline bedrock. The thickness of the deposits range from a thin mantle of till in the uplands to over 91 meters (300 feet) in the main channel of the Ancestral Rockaway River.

Compilation and subsequent interpretation of existing geological and hydrological data indicated that the buried valley deposits constitute a complex multi-aquifer system in which stream-aquifer interactions play an integral role. The Modular Three-Dimensional Finite Difference Ground-Water Flow Model (Mcdonald and Harbaugh, 1984) was used to simulate probable hydrogeological configurations and to determine possible values for the aquifer parameters of the extremely inhomogeneous sediments. Precise delineation of the various stratigraphic units into hydrostratigraphic layers was problematic. Distinctions between the various hydrostratigraphic units were based on both the horizontal and vertical hydraulic conductivity contrasts of adjacent layers.

The calibration process was treated as an inverse problem in which hydraulic conductivity, transmissivity, and leakage were varied. Steady-state simulations with constant areal recharge were conducted. Model calibration was achieved by adjusting aquifer parameters of the model until they corresponded to paramaters representative of the physical system. Matching calculated river fluxes between successive reaches of the Rockaway River with river fluxes observed in the field was used as a guideline in the development of a model capable of simulating actual stream-aquifer flow dynamics.

Overall model calibration was thwarted by the existance of too many hydrologic and geologic unknowns in the system. However, computer modeling was successful in defining constraints on the hydrogeological system and in constructing a realistic and supportive model that can be expanded upon as more field data become available.

2.0 INTRODUCTION

2.1 Purpose and Scope of Investigation

The purpose of this study was to define the hydrogeologic system operating within the Pleistocene valley fill deposits located in a portion of the Rockaway Valley in northeastern Morris County, New Jersey. Included in such an investigation is the characterization of the geology, the groundwater and surface water hydrology, and the climate of the area. Successful description of the dynamics of the groundwater flow regime depends upon how completely these factors can be determined and how thoroughly their interrelationships can be assessed.

The Rockaway Valley aquifer system is believed to consist of a network of tributary buried valleys which coalesce into one main extensive buried valley (Canace et al., 1981; Geonics, 1979). Aquifer parameters in such an environment commonly exhibit extreme variability. The labyrinth of buried valleys and the hummocky topography of the area implies the existence of a complex hydrogeological system. According to Freeze and Witherspoon (1967), hummocky topography, like that of the Rockaway Valley, may foster the development of numerous sub-basins which are superposed on the regional system. In addition, stratigraphic variability such as the presence of a highly permeable unit interfingered with more clayey (less permeable) units alters conditions that would normally be predicted in a more homogeneous system. For example, a permeable lens may affect the regional ground water flow by acting as a conduit, transmitting water to a principal discharge area and thereby altering the position of an anticipated recharge area. Also, the volume of water flowing to and from a local system to the regional system might inadvertantly be determined or affected by the presence of an unknown permeable lens.

Other variables, besides heterogeneity, that affect a system include the number of permeable and non-permeable units, the relative position of these layers, the geometry and lateral extent of the various strata, the existing boundary conditions, the ratio of depth to lateral extent of the basin, and the topographic relief. The interaction of these geological variables with the hydrology results in the evolution of existing discharge and recharge areas within the system.

Since it is apparent that many factors contribute to the configuration of the steady-state groundwater flow patterns of a basin, identifying the flow patterns of each of the components and documenting the existence of each sub-system is a pre-requisite and was paramount to this hydrogeological study. Determining the dynamic interaction of the local, intermediate, and regional flow patterns on the absolute flow regime of a basin is also crucial to a hydrogeological study of this type (Toth, 1963; Freeze and Witherspoon, 1967).

This study was divided into two phases. The first phase involved the construction of a conceptual model. Acquisition of data that accurately depicts the geology and hydrology is critical to the formulation of any hydrogeological model. The second phase of the study was the conversion of the conceptual model into data sets that represent the physical system. These data sets were used to initialize the conditions for a digital or computer model that simulates groundwater flow.

Selection of the appropriate computer model is usually based upon both economic considerations and applicability to the study area. In this case, a model capable of handling surface water-groundwater interactions was required. During the calibration of the computer model initial assumptions and observations are tested, and adjustments are incorporated into both the conceptual and the computer model. Ultimately, calibration of the computer model allows the groundwater flow system that exists in the 'real' physical world to be simulated. The strength of the model is reflected by its accuracy in determining actual or predicted behavior of the stream-aquifer system when tested under different sets of conditions.

The groundwater and surface water hydrologic data base for the portion of the Rockaway Valley in which the study was conducted is scanty. Therefore, much extrapolation and interpolation of the existing data base was necessary in order to fill in data gaps. A primary goal of this study has been to identify unknowns and problem areas and to construct the framework essential to the thorough description of the hydrogeological system. Although calibration of the model has been limited by the sparse data, the model can be used to acquire further insight into the hydrogeological system of the Rockaway Valley.

2.2 Sources of Information

The New Jersey Geological Survey (NJGS) was instrumental in providing information pertaining to the geology of the Rockaway Valley. Stanford's report _The Surficial Geology of the Boonton Quadrangle_ (Stanford, 1985) was invaluable in the development of the conceptual hydrogeological model. It includes descriptions of the thickness and physical nature of the various glacial deposits and cross sections depicting the geometry and juxtaposition of the individual units contained therein. Well logs, a bedrock map, and preliminary data from seismic profiles were also provided by the NJGS.

Streamflow measurements and concurrent water level data were supplied by the United States Geological Survey (USGS), Trenton Office. These data proved to be essential to the calibration of the computer model. Water level data from selected wells, with a brief synopsis of each well's history including its original yield, was also supplied by the USGS. A preliminary water budget for the Upper Rockaway Valley Drainage Basin was also made available.

Pumping schedules from public wells in the study area that withdraw a significant amount of water from the underlying aquifers were made available by the appropriate Water Authorities. These include the Denville, Boonton, Mountain Lakes, and Parsipanny Troy Hills Municipalities.

2.3 Location

The study area is located in northeastern New Jersey (See Figure 1). It includes sections of Denville, Boonton, Mountain Lakes, and Parsippany Troy Hills townships and is located in northeastern Morris County. The study area lies almost completely within the Boonton 7.5 minute Quadrangle with the exception of the south-southwest corner of the study area where it extends slightly into the Dover 7.5 minute Quadrangle. Precambrian bedrock ridges, from 183 (600 feet) to 244 meters (800 feet) in elevation, delineate the natural surface water divide of this sub-basin of the Rockaway River basin and were chosen as the northeast-southwest boundaries of the study area. The other boundaries to the study area were determined by the availability of data. Within practical limits, boundaries to the study area were chosen in accordance with natural

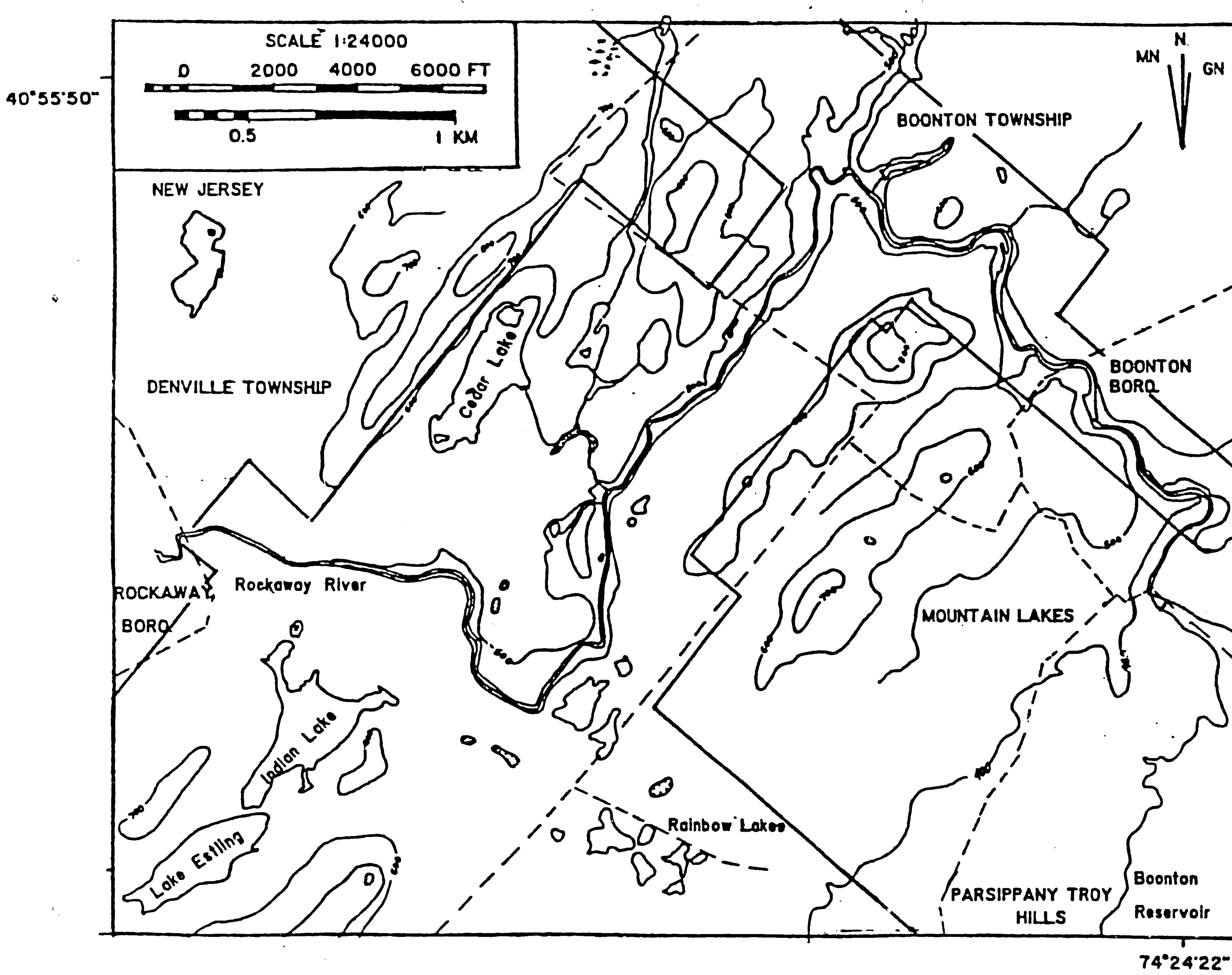


Figure 1 The location of the study area. The contoured elevations are in feet and the datum is mean sea level. The boldlines denote the boundaries to the study area.

drainage patterns. The Rockaway Drainage Basin is part of the more extensive Passaic Drainage Basin and is one of the primary tributaries to the Passaic River.

2.4 Physical Geography

Most of the Rockaway Valley lies within the New Jersey Highlands, one of many physiographic provinces of New Jersey (Figure 2). The New Jersey Highlands are situated between the folded and faulted Paleozoic strata of the Kittatiny Valley to the northwest and the bounding fault of the moderately dipping Triassic and Jurasic rocks of the Piedmont Lowlands to the southeast. The Green Pond Fault and the Ramapo Fault are aligned parallel to the general structural trend, and, in effect, structurally isolate the block of Precambrian metasedimentary to igneous rocks.

These Precambrian rocks of the New Jersey Highlands consist of northeast-southwest trending ridges which are separated by deep narrow valleys commonly underlain by Paleozoic sedimentary and metasedimentary rocks lying 61 (200 feet) to 91 meters (300 feet) below the ridge crests. The structural and lithologic character of the bedrock and subsequent stream erosion have worked together to influence the existing relief (Sims,1958). Locally, glacial processes have reshaped the original landforms. Hummocky topography with occasional depressions reflects the impact that glaciation has had on the geomorphology of the study area.

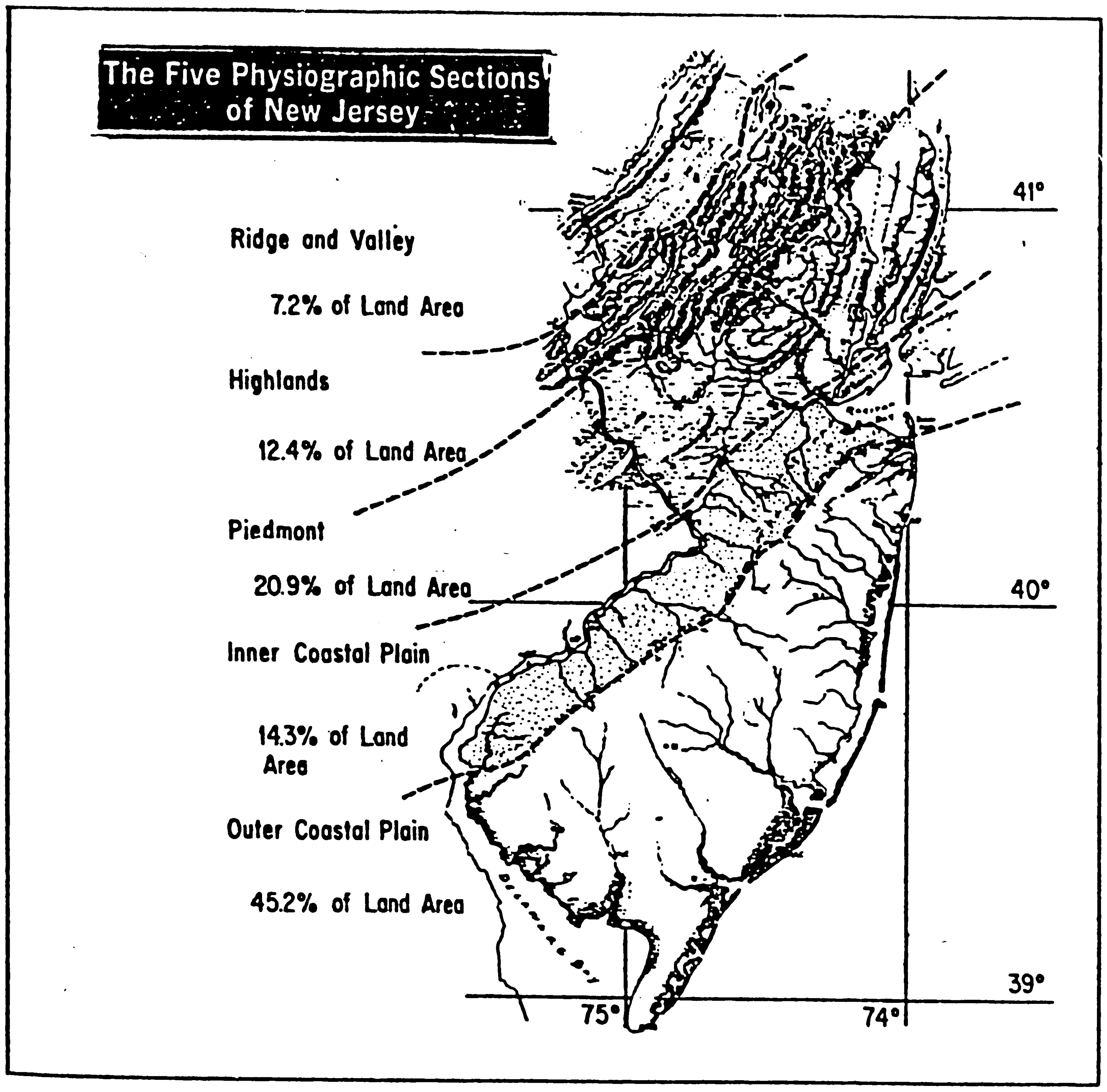


Figure 2 The location of the Physiographic Provinces of New Jersey (map taken from Robichund and Buell, 1973).

The majority of the river valleys within the Highlands parallel the northeast-southwest trending structures. Exceptions to this include segments of the Rockaway River. For example, near Boonton, New Jersey, the river has cut a transverse valley through the resistant crystalline ridge resulting in an abrupt change in its course and causing the river to flow nearly perpendicular to the general structural trend. Such anomalies in the surface drainage patterns are the result of several inter-related processes (Davis and Wood, 1890). The combined effects of glacial processes and tectonic activity have worked together in altering pre-existing drainage patterns.

2.5 Climate

New Jersey's climate is classified as continental which means temperatures fluctuate considerably from the summer to winter months. Daily temperatures and day-to-day temperatures also vary considerably. For example in January, the average temperature reported by northern New Jersey stations is -1^{o} C (30^{o}F) and for July it is 23^{o}C (74^{o}F). Table 1 lists the monthly average temperatures and the annual average temperatures in Northern New Jersey for the period of record from 1931 to 1980. The average for each month for the entire 49 years is included in Table 1. Temperature variations directly affect the amount of water evaporated back into the atmosphere and the length of the

Table 1 Northern New Jersey average temperature, in ^{o}F (table from Halasi-Kun, 1981).

Yr	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual
1931	29.8	31.9	38.8	49.4	60.4	69.2	76.0	72.8	69.6	57.4	49.3	37.6	53.5
1932	40.7	34.1	35.8	47.6	60.8	68.8	73.3	73.1	65.9	55.4	40.4	34.6	52.5
1933	37.0	31.9	37.0	50.1	62.5	70.6	73.3	72.5	67.0	52.7	39.2	30.3	52.0
1934	32.1	17.6	35.8	49.1	62.2	72.0	75.4	69.4	66.5	51.7	46.8	27.8	50.8
1935	26.5	28.7	41.6	48.4	57.2	68.5	75.7	71.8	62.2	54.1	46.8	27.8	50.8
1936	25.7	23.1	44.2	47.2	62.8	68.6	74.3	73.8	66.1	54.2	39.6	36.0	51.3
1937	37.3	23.1	44.2	47.2	62.8	68.6	74.3	74.7	62.8	52.3	42.5	31.9	52.0
1938	29.0	33.9	42.2	52.3	58.8	68.3	74.1	75.0	62.3	55.6	44.3	33.7	52.5
1939	29.4	34.5	37.8	47.0	63.0	70.3	73.0	74.5	65.3	53.7	40.6	33.8	51.9
1940	21.6	30.5	32.8	44.9	59.3	67.5	73.2	68.2	62.3	47.4	42.2	34.8	48.7
1941	27.5	28.4	33.2	55.6	61.9	69.1	73.4	70.0	66.1	57.4	45.7	35.6	52.0
1942	27.2	28.3	41.2	52.8	63.5	68.8	73.6	70.5	64.8	54.7	43.3	28.3	51.4
1943	28.4	30.8	37.6	44.6	61.2	73.5	73.9	72.6	64.0	52.4	40.8	29.0	50.7
1944	30.9	30.5	35.5	46.9	65.0	69.9	75.2	74.1	65.9	52.4	42.6	29.0	51.5
1945	22.3	30.6	48.8	54.0	57.6	68.8	72.7	70.6	68.1	52.8	44.3	27.6	51.5
1946	30.8	30.1	47.2	49.3	59.7	67.6	72.8	68.2	66.5	57.9	46.9	34.6	52.6
1947	34.4	27.3	35.7	49.4	59.1	66.7	73.0	74.1	65.9	59.1	40.8	30.8	51.4
1948	22.1	27.2	40.5	50.2	59.2	68.0	74.2	72.8	65.9	52.8	48.7	33.8	51.3
1949	36.1	36.5	41.5	51.3	60.9	71.8	77.4	74.0	63.1	59.3	42.5	35.6	54.2
1950	38.3	29.7	35.1	47.3	58.3	68.0	72.2	70.7	61.9	56.5	45.1	30.7	51.2
1951	32.8	33.4	40.2	50.6	61.1	68.0	73.3	71.2	64.3	55.8	39.5	34.3	52.0
1952	33.4	33.6	38.2	52.6	57.9	70.9	76.2	71.6	65.1	50.7	44.0	35.1	52.4
1953	34.2	36.0	41.4	50.0	62.1	70.1	74.0	71.5	66.0	55.8	44.2	37.1	53.5
1954	27.2	37.0	39.9	52.1	57.3	69.3	72.9	70.3	64.2	59.9	42.0	33.0	51.9
1955	28.2	31.4	39.8	52.7	63.1	66.9	78.9	76.1	64.3	56.5	42.3	27.4	52.3
1956	30.4	34.0	35.2	46.6	57.7	69.8	70.9	71.2	62.2	54.6	43.8	37.6	51.2
1957	25.3	34.4	39.3	51.7	60.4	71.9	73.3	69.5	65.9	52.0	45.3	36.4	52.1
1958	28.8	25.2	38.1	51.1	57.0	64.6	74.2	71.7	64.1	52.0	44.3	25.6	49.7
1959	28.5	29.1	38.0	51.9	63.4	69.4	74.2	74.1	68.2	56.9	42.9	34.6	52.6
1960	31.0	33.9	30.5	53.4	59.5	68.5	71.3	71.9	64.5	53.1	44.8	25.3	50.6
1961	21.8	31.5	38.8	46.1	57.4	68.9	73.6	72.1	70.8	56.0	44.8	31.1	51.1
1962	28.0	27.4	38.6	50.4	61.6	69.8	70.6	70.1	61.0	54.0	39.9	27.9	49.9
1963	25.3	23.0	39.2	50.0	57.9	68.2	73.3	69.4	60.5	57.5	47.1	26.4	49.8
1964	29.9	28.3	39.8	47.4	61.9	67.1	73.5	69.2	65.3	51.1	45.4	32.2	50.9
1965	25.2	29.1	36.4	46.7	64.0	67.8	71.4	71.9	65.5	51.0	41.7	34.2	50.3
1966	27.3	29.4	39.3	45.6	56.2	69.8	75.2	72.6	62.2	51.3	44.4	32.1	50.4
1967	32.3	25.5	34.3	48.2	52.3	70.1	72.1	70.2	62.2	52.3	38.6	33.7	49.3
1968	22.6	26.3	40.5	51.9	56.5	66.7	73.9	72.2	66.0	55.8	42.6	29.3	50.4
1969	27.1	29.7	36.3	51.8	60.4	69.5	71.2	72.7	64.8	52.5	42.1	29.4	50.6
1970	20.8	29.1	35.0	48.3	60.8	67.2	73.8	73.0	66.4	55.0	44.6	31.7	50.5
1971	23.8	30.0	36.9	47.3	56.6	69.6	72.3	70.7	67.3	58.7	41.5	37.0	51.0
1972	31.0	26.9	36.4	46.0	59.9	65.1	74.0	70.9	65.5	49.0	40.2	35.0	50.0
1973	32.1	29.3	44.2	50.8	56.7	71.5	74.1	74.0	65.4	55.6	44.9	35.6	52.9
1974	31.1	28.5	39.7	52.0	58.7	66.0	73.2	72.1	63.4	49.3	43.9	35.2	51.1
1975	32.7	32.0	36.8	44.4	62.9	68.6	73.0	71.7	60.6	56.6	48.3	33.1	51.7
1976	24.1	35.1	41.7	52.2	58.2	71.0	71.1	70.7	62.1	49.1	37.4	26.1	49.9
1977	18.3	29.7	43.5	51.2	62.0	66.2	73.5	71.0	64.8	51.3	44.7	30.3	50.5
1978	25.0	21.5	35.2	47.9	57.4	67.4	70.7	72.4	61.9	51.6	43.6	33.4	49.0
1979	29.0	19.8	41.9	48.7	60.5	65.6	71.8	74.4	63.7	52.0	46.7	34.6	50.5
1980	29.4	26.5	36.4	50.5	61.8	65.9	74.7	74.4	66.8	50.7	39.6	27.4	50.3
AVE.	29.5	29.7	38.6	49.5	59.9	68.8	73.5	71.9	64.8	53.9	43.3	32.2	51.2

growing season. Both of these factors, in turn, affect the water budget of an area. The series of three contoured maps in Figure 3 illustrate the duration of the growing season.

Precipitation is also a very important climatic factor to consider. Table 2 lists the average annual precipitation and the monthly average precipitation for northern New Jersey from 1931 to 1980. Included in this Table are the average monthly and the annual precipitation for the 49 years of record.

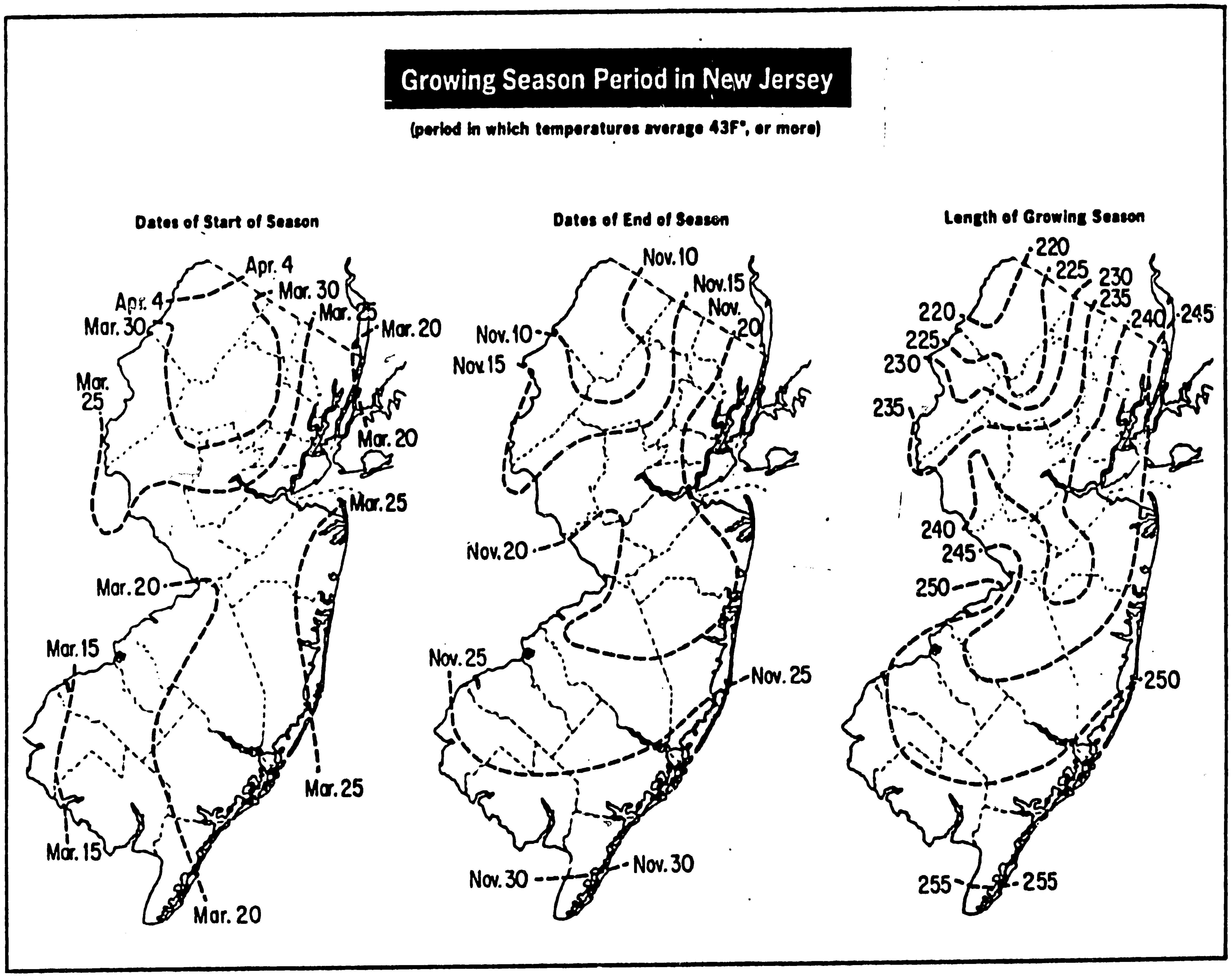


Figure 3 Maps depicting the start, end, and length of the growing seasons for counties in New Jersey. Map taken from Robichaud and Buell, 1973.

Table 2   Northern New Jersey precipitation, in inches (table from Halasi-Kun, 1981).

Year	Jan.	Feb.	Mar	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
1931	2.09	2.34	3.69	2.99	4.50	5.30	5.18	4.69	2.22	2.71	0.82	2.28	37.81
1932	4.17	2.40	4.86	2.33	2.55	3.52	2.45	2.96	2.16	6.52	8.70	2.32	44.94
1933	1.82	3.17	5.07	5.11	4.67	2.79	3.66	9.96	7.59	2.35	0.62	2.89	49.70
1934	3.15	2.26	3.48	4.31	4.73	4.15	3.68	3.23	10.14	2.63	3.14	2.79	47.69
1935	3.49	2.55	2.48	1.81	1.20	4.34	6.28	2.07	4.60	5.55	4.40	1.43	39.21
1936	6.66	2.83	6.05	3.78	3.12	5.85	1.95	4.50	2.88	3.83	4.41	6.36	49.22
1937	6.29	2.47	2.80	4.67	3.34	5.11	3.59	7.38	2.78	4.41	4.63	2.18	49.65
1938	3.93	2.21	2.42	3.21	3.46	7.74	9.01	2.77	8.99	2.26	3.27	1.18	52.75
1939	3.63	5.21	4.47	5.06	1.51	3.65	2.21	4.54	1.76	3.98	1.79	1.93	39.74
1940	2.27	2.86	5.73	5.87	6.39	4.04	3.08	5.30	4.12	2.41	4.57	3.48	50.12
1941	3.04	2.44	2.69	2.23	1.90	4.82	6.07	4.85	0.46	2.00	3.27	3.98	37.75
1942	2.92	2.73	5.63	1.44	3.19	3.91	7.61	9.49	6.13	3.27	4.66	5.05	56.03
1943	3.06	2.09	2.85	2.67	5.71	3.56	4.50	3.32	1.60	7.58	3.37	1.18	41.49
1944	3.37	2.43	5.71	5.36	1.83	4.53	1.83	2.48	7.17	1.75	6.24	3.32	46.02
1945	3.53	3.19	2.73	3.87	6.11	4.93	10.23	4.76	5.65	2.46	5.94	4.39	57.79
1946	1.57	2.22	3.09	1.35	7.53	5.24	3.23	4.92	3.94	1.79	1.32	2.10	41.30
1947	3.48	2.06	3.19	5.02	7.67	4.55	4.94	3.60	3.22	2.15	7.21	3.36	50.45
1948	4.19	2.34	3.73	3.92	7.80	6.02	4.09	4.71	1.03	1.89	4.00	6.73	50.45
1949	6.28	3.05	2.16	4.38	4.86	0.24	4.49	3.43	4.29	1.92	1.31	3.43	39.84
1950	2.80	4.45	4.52	2.34	3.77	3.09	5.38	4.84	2.29	1.87	5.73	5.07	46.15
1951	3.67	4.47	6.41	3.33	4.28	4.13	6.47	3.56	1.99	4.59	7.09	5.21	55.20
1952	4.95	2.38	5.39	7.36	5.98	5.11	5.52	6.15	5.20	0.89	5.06	4.40	58.39
1953	5.97	2.34	7.43	6.26	4.92	2.49	4.13	2.18	2.06	3.64	2.12	5.11	48.65
1954	1.98	2.10	3.75	3.09	4.53	1.06	1.69	7.47	5.73	2.21	5.94	3.70	43.25
1955	0.93	3.23	4.39	2.65	1.60	3.76	1.10	14.36	2.79	8.88	3.02	0.37	47.08
1956	1.86	4.76	5.06	3.78	2.69	4.10	5.61	3.61	4.16	3.02	3.77	4.36	46.78
1957	2.10	2.49	2.72	5.81	2.71	2.21	1.92	2.18	2.83	3.18	3.72	6.97	38.84
1958	5.07	4.23	4.25	6.59	3.94	2.71	4.78	3.32	4.51	5.95	3.13	1.36	49.84
1959	2.51	2.02	3.78	3.02	1.45	5.10	4.00	5.77	2.43	5.86	3.90	4.66	44.50
1960	3.29	4.24	2.36	3.72	4.22	2.35	7.92	5.92	7.17	2.10	2.31	2.86	48.46
1961	2.88	3.40	4.91	5.68	3.58	2.58	6.44	4.42	1.96	1.92	3.22	3.36	44.35
1962	2.79	4.14	3.10	4.21	1.49	3.63	2.56	5.58	3.52	3.58	4.43	2.65	41.68
1963	2.53	2.35	3.45	0.86	2.91	2.59	3.23	2.21	4.40	0.37	6.94	2.11	33.95
1964	4.65	2.83	2.24	5.74	1.19	3.33	4.40	0.64	1.59	1.21	2.43	4.40	34.65
1965	2.64	3.08	2.83	2.43	1.54	1.34	2.37	3.96	2.82	3.41	2.03	2.01	30.46
1966	3.05	3.72	2.43	2.75	3.88	2.58	6.44	4.42	1.96	1.92	3.22	3.36	44.35
1967	1.47	2.04	6.17	2.68	3.90	3.55	6.45	7.11	2.11	2.78	2.43	5.53	46.22
1968	2.28	0.98	4.48	2.76	7.29	5.52	1.51	3.20	2.32	2.51	4.99	3.86	41.70
1969	1.69	2.17	3.30	3.52	2.87	3.75	8.56	4.54	5.39	1.81	3.54	5.75	46.89
1970	0.81	3.47	3.58	4.05	3.29	2.73	3.45	4.45	2.29	4.54	5.65	2.74	41.05
1971	2.73	5.10	3.45	2.55	4.54	1.56	5.36	10.28	6.52	4.23	5.39	1.88	53.59
1972	2.45	4.55	3.68	3.37	6.63	10.87	4.01	1.89	2.38	4.67	9.70	6.06	60.36
1973	4.60	3.67	3.69	7.09	5.43	6.49	4.31	4.96	3.88	4.62	1.84	8.93	59.51
1974	3.48	1.81	5.45	3.99	4.02	4.32	2.16	6.78	7.62	2.15	1.86	5.03	48.67
1975	5.43	3.67	4.21	2.95	4.75	6.99	10.83	4.37	9.26	4.33	3.96	2.84	63.59
1976	5.32	2.66	2.30	2.80	4.08	4.70	3.81	4.14	3.40	6.22	0.64	2.41	42.48
1977	1.62	2.76	7.22	4.69	1.83	7.70	2.31	4.29	5.21	4.50	7.30	5.51	51.94
1978	8.12	1.61	3.84	2.19	7.30	2.79	3.74	6.63	3.84	1.75	2.60	4.69	49.10
1979	10.51	4.40	3.58	4.09	6.50	3.25	3.37	4.86	7.54	4.15	6.35	2.27	58.17
1980	1.99	1.07	7.33	7.27	2.59	3.34	3.34	1.77	1.69	4.41	3.38	0.66	
Ave.	6.86	2.94	4.08	3.86	4.04	3.99	4.47	4.72	4.14	3.39	3.92	3.67	46.73

3.0 GEOLOGY

Basically, the study area consists of two contrasting geological environments: unconsolidated Pleistocene to Holocene valley fill deposits and foliated Precambrian crystalline rocks comprising the surrounding flat-topped ridges.

3.1 Precambrian Geology

The Precambrian rocks of the New Jersey Highlands had originally been mapped as belonging to one of the three following formations- Byram Gneiss, Losee Gneiss, and the Pochuck Gneiss. The Byram Gneiss Formation is described as a gray granitoid gneiss composed of microcline, microperthite, quartz, hornblende or pyroxene with occasional mica. The Losee Gneiss is a white granitoid gneiss composed of oligoclase and quartz. Orthoclase, pyroxene, hornblende, and biotite may be present in varying but generally minor amounts. The Pochuck Gneiss is a dark granular gneiss composed of pyroxene, hornblende, oligoclase, and magnetite. However, according to Sims (1958), the gneisses grade into each other through intermediate rock types, making the distinction between each of the formations somewhat arbitrary. Therefore, he redefined the classification of these rocks by using mineralogical adjectives instead of separating them into one of the three previously accepted formations.

It is these northeast-southwest striking gneissic rocks dipping steeply to the southeast that make up the resistant bedrock ridges characteristic of the New Jersey Highland. The steep ridges and narrow valleys parallel the northeast trending isoclinal to open folds of the area. The axes of the folds generally plunge northeast at shallow to moderate angles.

The Precambrian rocks have a well developed foliation. The foliation was produced by the dimensional orientation of platy and tabular minerals and by the parallelism of the relict sedimentary layering (Sims,1958). In addition, many of the rocks are marked by a lineation, parallel to the fold axes.

Hydrogeologically, the most significant characteristic of these crystalline rocks is the fracture pattern imprinted on them by later brittle deformation. Faulting, jointing, and related fracturing has increased the potential of otherwise relatively impermeable rocks to store and transmit water.

3.1.1 Fault Geometries

Figure 4 illustrates the locations and trends of faults that may exist in the area. These were extracted from the most recent and the most structurally detailed geologic maps

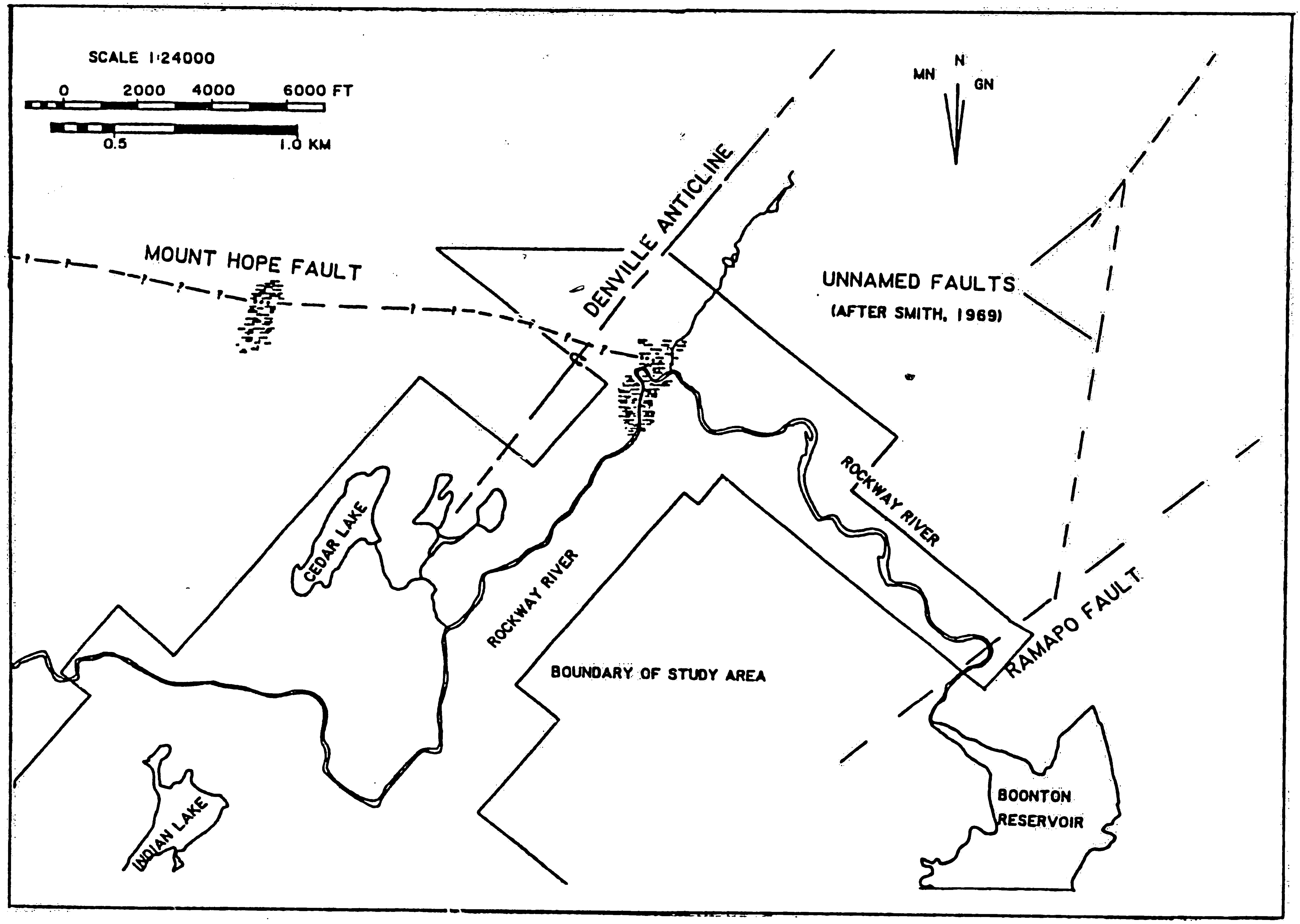


Figure 4 Location of primary structural features in the Boonton Quadrangle (map compiled from Sims, 1958 and Smith, 1969).

of the Boonton Quadrangle. The Mount Hope Fault strikes N 78^{o} W and has an average dip of 60^{o} SW. This high angle normal fault has a horizontal separation of about 91 meters (300 feet). From exposures in mines, it can be seen that there is little difference in its attitude and displacement between the surface and a depth of 640 meters (100 feet), (Sims, 1958). Also, exposures of the fault in mine workings indicate that there is a 6 (20 feet) to 9 meter (30 feet) wide brecciated and fragmented zone. Most of the dislocation surfaces are marked by gouge zones that range in width from 5.1 centimeters (6 inches) to 60.9 centimeters (2 feet) and are marked by abundant secondary chlorite. According to Sims (1958), "Water circulates through the zone; when first intersected on any level the flow is several hundred gallons per minute, but after the fault is tapped at a lower level, the flow abruptly diminishes." The most eastern segment of the Mount Hope Fault on Sims' Geological Map of the Dover Magnetite District Morris County, New Jersey, (1958) is shown in Figure 4. It appears that the fault extends into the northern corner of the study area. Here the fault surface is obscured by the glacial valley fill deposits. Since this geological map terminates at the Denville Anticline which is the northwest bordering ridge of the hydrogeologic study, it is uncertain how far east into the study area the fault may extend. The hydrologic significance of this fault is discussed in Section 12.0.

In addition to the Mount Hope Fault there are three other prominent faults in close proximity to the study area. Smith (1958) presents a very general geological map of the Precambrian Geology of Central and Northern New Jersey Highlands which includes the Boonton Quadrangle. According to this map two obliquely intersecting faults lie within the Boonton Quadrangle. The projection of these proposed faults is shown on the structural map of the study area (Figure 4). Anomalous patterns in the topography were used as guidelines to determine the locations of faults within and adjacent to the study area.

Finally, the extreme northeastern corner of the study area is bounded by the Ramapo Fault. Here, the Rockaway River, about one half mile upstream from the Boonton Reservoir, turns abruptly and flows approximately parallel to the northeast-southwest trending fault. The fault marks the juxtaposition of Precambrian basement rock with Jurassic(Triassic?) sedimentary rocks of the Newark Group. This geological feature is generally referred to as the 'Highland Escarpment'.

3.1.2 Joint Patterns

Rocks of the New Jersey Highlands contain three principal joint sets which are observed in nearly every rock exposure (Sims, 1958). The jointing patterns are classified as transverse, longitudinal, and diagonal. The transverse joint sets are the most abundant and pervasive. They exhibit two maxima of N 35° W and N 50° W; both display dips of nearly 90°. Spacing of the transverse joints ranges from a few inches to several feet. The joints are evenly spaced in a single outcrop. Exposures in mines suggest that transverse joints exist as deep as 1,000 feet beneath the surface. Longitudinal joints are second in abundance and generally have smooth plane surfaces with highly variable spacing. These joint sets are oriented nearly parallel to the fold axes with a strike of N 45° E. They are inclined to the northwest with an average dip of 45°. Lastly, the diagonal joints are steep fractures which trend slightly east of north.

3.2 Pleistocene and Holocene Geology

The unconsolidated deposits of Pleistocene and Holocene Age consist of gravel, sand, and clay sediments in various proportions. These sediments are of glacial, glaciofluvial, fluvial, and lacustrine origin (Meisler,1976). The deposits directly overlie the crystalline Precambrian bedrock and have filled in previously existing stream valleys. The areal extent, thickness, and spatial configuration of these deposits were of primary importance to this study since these materials are the major groundwater producers of the area (Meisler, 1976; Canace et al., 1983).

Knowledge of the physical nature and the delineation of the valley fill deposits into stratigraphic units is necessary in order to understand the hydrogeology of the study area. However, it is not the author's intent, nor the purpose of this investigation to discuss the glacial history of the Rockaway Valley. For a more detailed and highly enlightening discussion of the subdivision of the unconsolidated deposits into stratigraphic units, the reader is referred to Stanford's report on the Surficial Geology of the Boonton Quadrangle (Stanford,1985).

Traditionally, deposits of glacial origin have been sub-divided into stratified drift and till. Stratified drift encompasses all materials deposited by flowing water emanating from the glacier. In the Rockaway Valley, these sorted and layered deposits include valley fill materials of outwash sands and gravels, and lacustrine fine sands. Tills or nonstratified drift consist of heterogeneous mixtures of boulders, gravels, sands, and clays. In the Rockaway Valley, the Wisconsinian Terminal Moraine comprises the primary deposits of this nature, although there are mapped deposits of other tills, including mixtures of ground moraine and colluvium. Figure 5, a surficial map of the area, shows the areal distribution of the unconsolidated

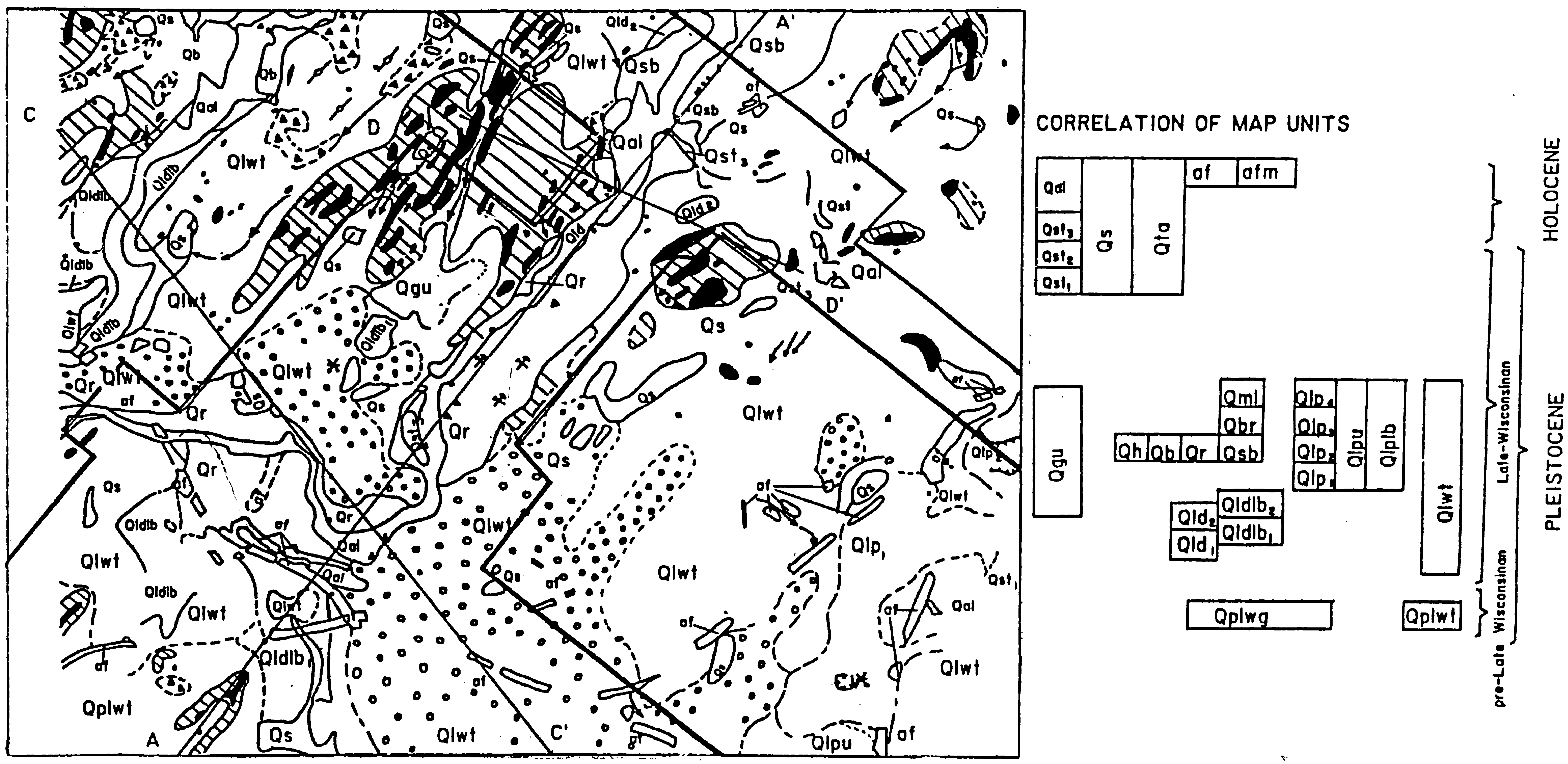
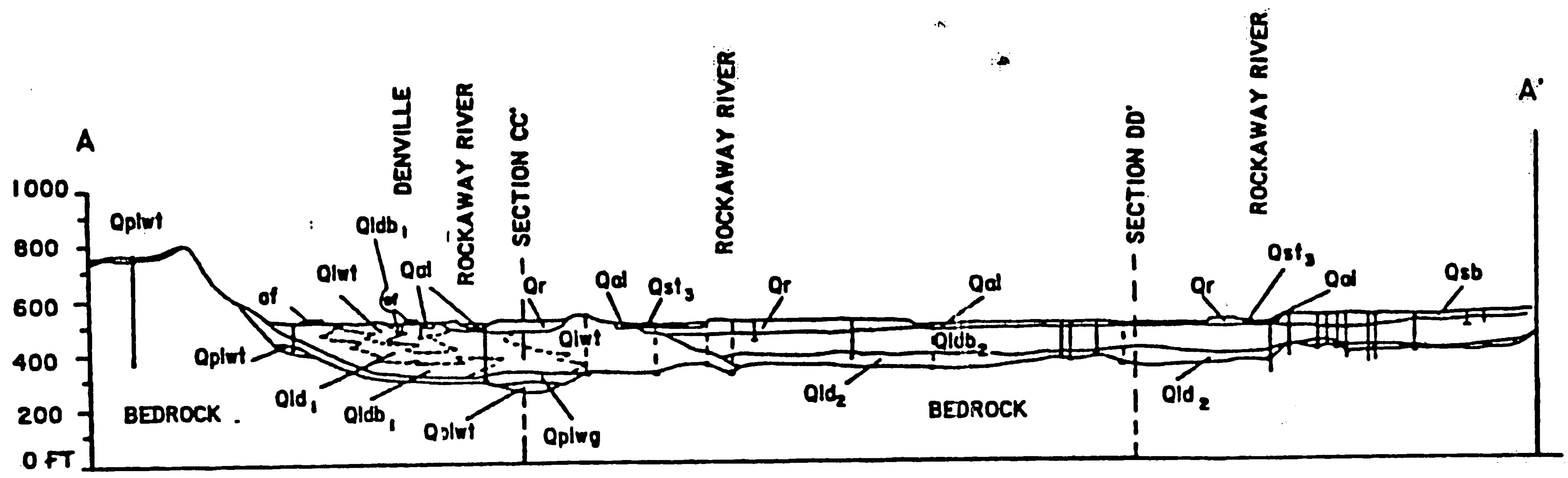
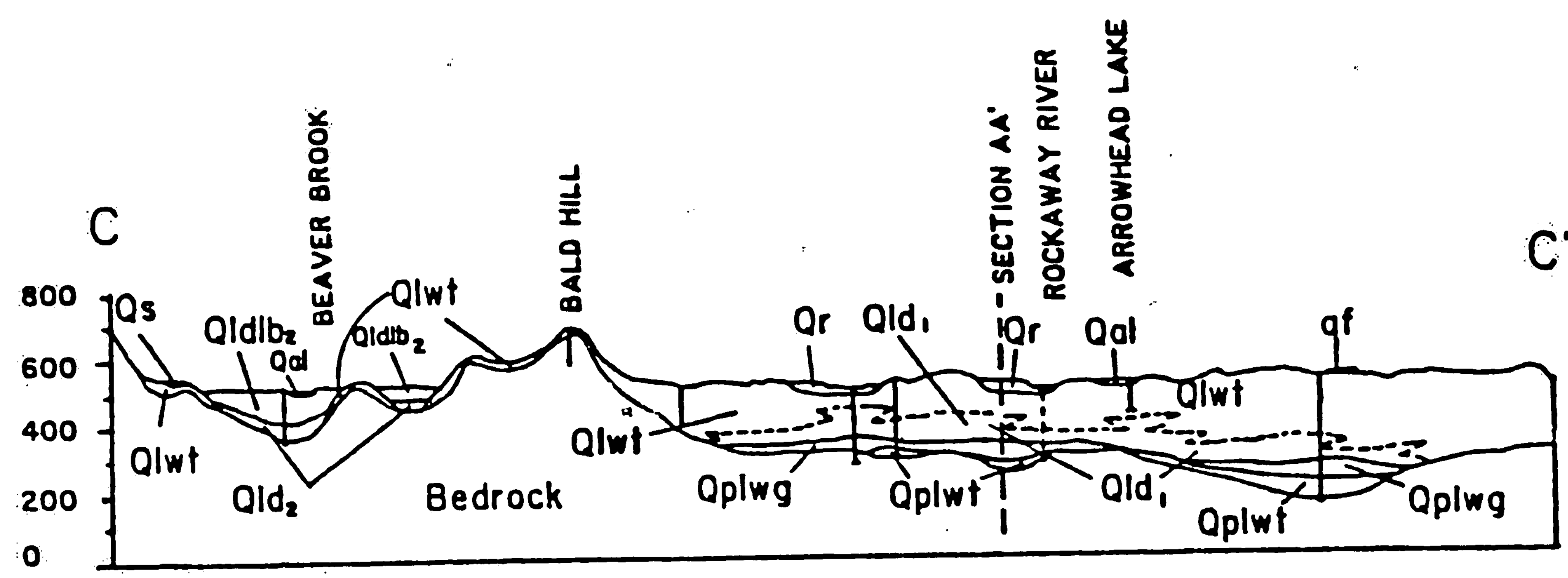


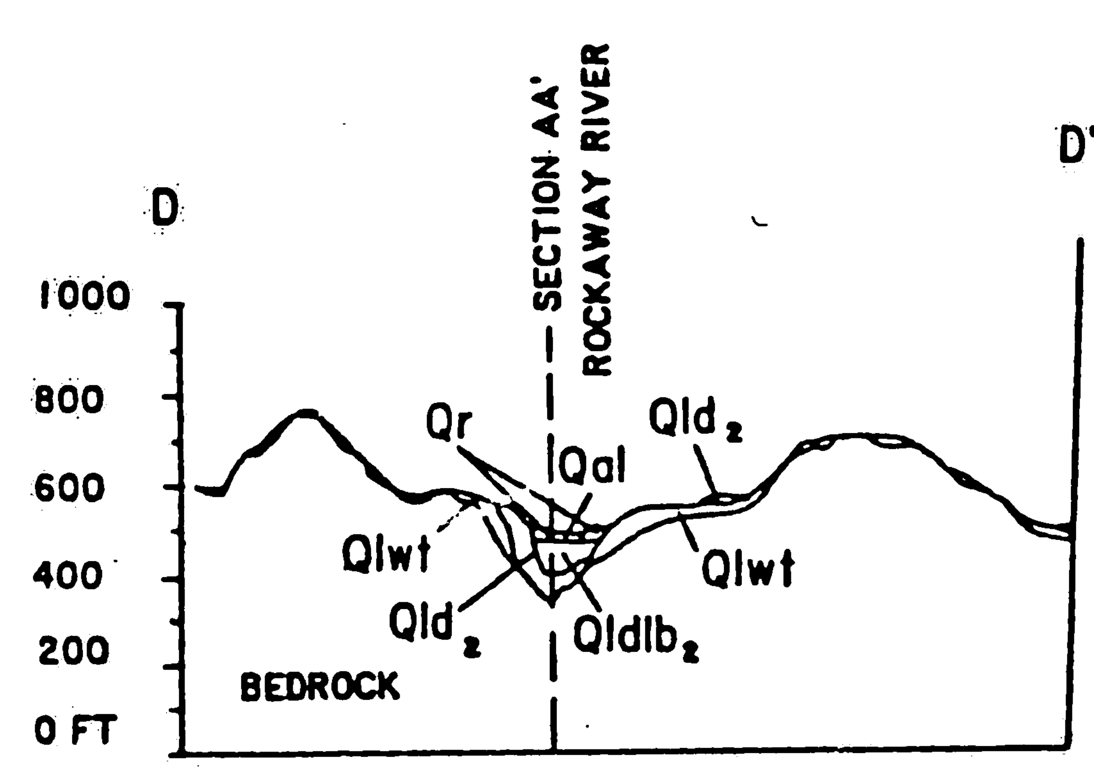
Figure 5 Surficial Geology of the study area
(Geology by Stanford, 1985).

Pleistocene to Holocene sediments as well as outcrops of Precambrian bedrock.

The stratified deposits in the area have been further subdivided by Stanford (1985) on the basis of depositional environments, which indirectly reflect their relative properties and stratigraphic position with respect to one another. For example, the classifications of outwash deposits and of lacustrine deposits indirectly relate to the respective compositions and physical characteristics of these stratified materials. The classification also reflects orderly depositional processes that imply a certain geometry and position in the stratigraphic column. The generalized cross sections in figures 6,7, and 8 and the accompanying geological key (see Appendix A) summarizes the stratigraphic subdivisions of both the stratified and non-stratified drift deposits and serves to demonstrate the spacial configuration of each deposit. The significance of this classification is apparent when the relationship between the mode of deposition, i.e., by glacial melt water or by the glacier itself is coupled with the resulting physical nature of the respective deposits. In other words, there is a direct relationship between the hydrogeological properties of a deposit and its depositional history. Of course, the rock from which these deposits were derived also plays a role in the degree of permeability that the sediments will have. For example, the greater the







HORIZONTAL SCALE 1" = 2000 FT
VERTICAL SCALE 1" = 400 FT
VERTICAL EXAGGERATION = 5X

Figures 6, 7, and 8 Geological cross sections depicting the stratigraphy and geometry of the unconsolidated sediments. See Figure 4 for Line of Section (Cross Sections Constructed by Stanford, 1985).

percentage of clay minerals in the matrix of the source rock the less permeable the sediments will be.

The thickness of the valley fill materials varies considerably from place to place (Figure 9). Nonuniform thickness is a result of several dynamic processes that shaped the Pleistocene environment. Erosional activity responsible for the deep scouring of channels by the pre-glacial Rockaway River drainage system produced convenient paths along which lobes of the glacier could travel and continue downcutting. These deepened troughs were then available for subsequent deposition of unconsolidated materials. Climatic influences helped to dictate the maximum advance of the glacier during Wisconsinan times. This in turn resulted in the deposition of thick layers of materials within the main pre-glacial channel of the Rockaway River and produced the thick sediment pile at the terminus of the glacier, which comprises the Wisconsinan Terminal Moraine. Lastly, the transporting potential of the glacier and its meltwaters determined the volume, character, and assorted sediment sizes characteristic of these Pleistocene deposits thus leaving behind a lasting imprint of the complex and highly energetic glacial and peri-glacial environment.

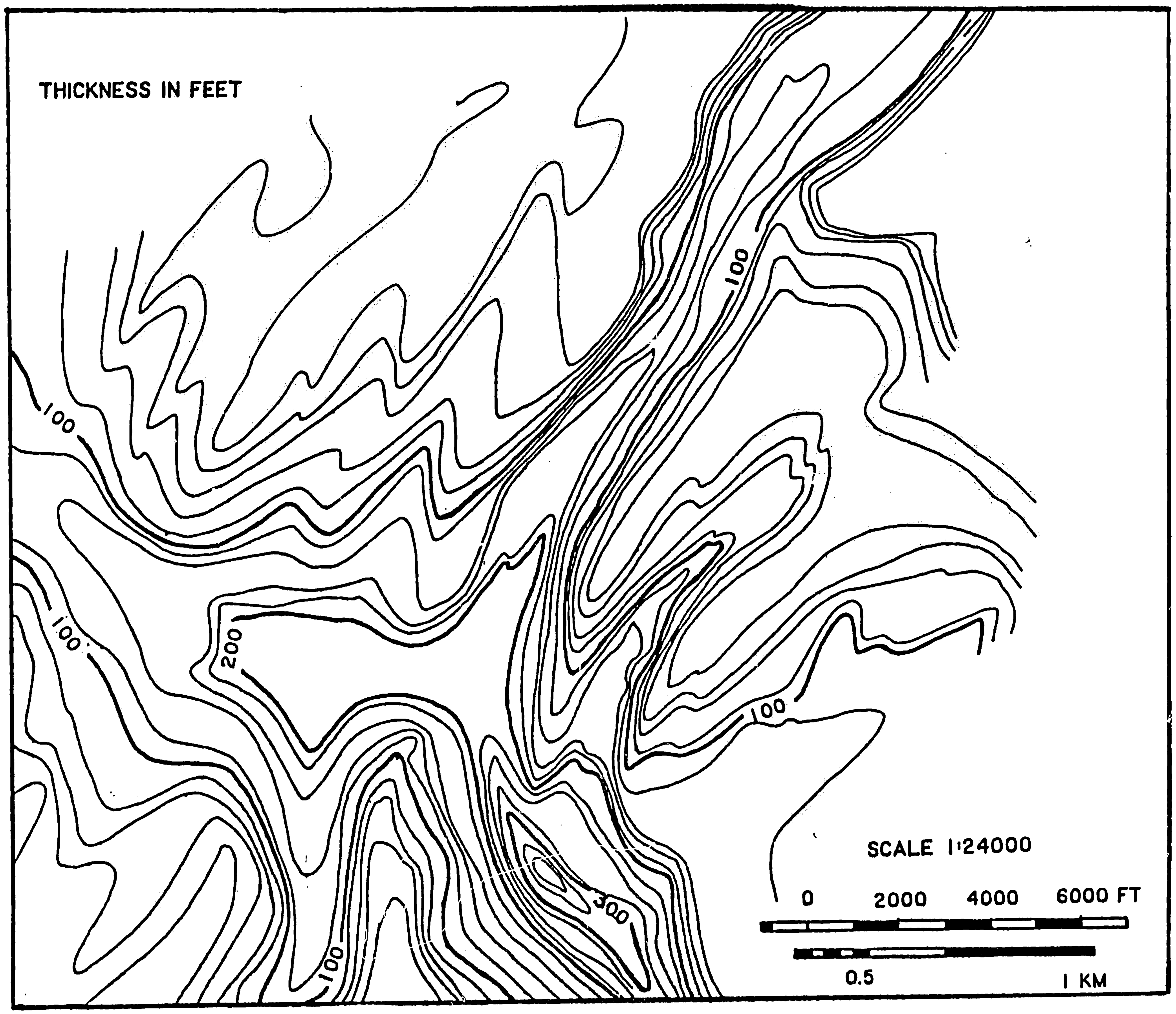


Figure 9 Isopach map of unconsolidated sediments of Pleistocene and Holocene Age, existing in the study area.

As previously mentioned, the pre-glacial Rockaway River, southeast of Denville, flowed within a valley trending nearly transverse to the regional trend of the Highland Escarpment. It is within this pre-glacial valley that the maximum thickness of glacial sediments accumulated. Modification of flow patterns became necessary as ice and sediment blocked off the existing route of the river. The development of the pro-glacial drainage system coupled with the infilling of parts of the pre-glacial river valley has resulted in an integrated network of buried tributary valleys which coalesce at the pre-glacial Rockaway River channel (Canace, et al.,1983; Geonics,1979).

4.0 HYDROGEOLOGY

Converting the stratigraphic column of unconsolidated sediments into hydrogeological units that are both representative of the actual geology yet more manageable in the development of a numerical model required an involved hierarchical and systematic approach. The first step toward the separation of the thick sequence of unconsolidated materials into hydrogeological layers included an analysis of the literature containing documentation of water bearing properties of the glacially derived sediments. The most applicable information obtained indicated, as one would expect, that stratified deposits are the most likely candidates for groundwater exploitation. The wide range of hydraulic conductivity and porosity of glacial materials is demonstrated in Tables 3 and 4, respectively. In addition, determination of how and to what extent the geometrical configuration of the various geologic layers influences the groundwater flow patterns was also necessary before a decision as to how the various layers should be grouped into workable hydrogeological units could be made.

Knowledge of the way in which the sediments were deposited allows inferences to be made as to the grain size, degree of sorting, and other properties of the material. This in turn provides insight into the ability of the sediments to store and to transmit fluids. Therefore, an integrated analysis of the depositional mode and of the

Table 3

Porosity Ranges for Unconsolidated Sediments

Material	n(%)
Well sorted sand or gravel	25-50
Sand and gravel, mixed	20-35
Glacial till	10-20
Silt	35-50
Clay	33-60

Representative Values of Porosity

Material	n(%)
Gravel, coarse	28
Gravel, medium	32
Gravel, fine	34
Sand, coarse	39
Sand, medium	39
Sand, fine	43
Silt	46
Clay	42
Till (predominantly silt)	34
Till (predominantly sand)	31

TABLE 4

Ranges of Hydraulic Conductivity for Unconsolidated Sediments (after Fetter, 1980)

Material	Conductivity (m/d)
Clay	10^{-6} - 10^{-3}
Silt, sandy silts, clayey sands, till	10^{-3} - 10^{-1}
Silty sands, fine sands	10^{-2} - 10
Well-Sorted sands, Glacial outwash	1 - 10^{2}
Well-sorted gravel	10 - 10^{3}

Representative Values of Hydraulic Conductivity
(afer Todd, 1980)

Material	Conductivity	Type of Measurement*
Gravel, coarse	150	R
Gravel, medium	270	R
Gravel, fine	450	R
Sand, coarse	45	R
Sand, medium	12	R
Sand, fine	2.5	R
Silt	0.08	H
Clay	0.0002	H
Till, predominantly sand	0.49	R
Till, predominantly gravel	30	R

* H is horizontal hydraulic conductivity, R is a repacked sample

calculated aquifer parameters allows the subdivision of the sequence of glacial deposits according to hydrogeological properties. This subdivision can be markedly different than a subdivision based soley on the recognition of geological units. A quick study of the geologic cross-sections reveals that there are numerous map units. However, in hydrogeological terms, and for modeling purposes, it is commonly more practical and expedient to group those units exhibiting similar hydrogeological properties into a single unit. Exactly how to regroup the unconsolidated deposits into a few manageable, yet representative layers is the dilemma. The existence of at least a three layer, multi-aquifer system is supported by 1975 pump test data conducted at the Boonton well field by the Moretrench American Corporation. However, the results strongly suggest that such a three layer aquifer system may exist only in the vicinity of the Boonton well field. The interfingering nature of the various valley fill deposits toward and within the main tributary (pre-glacial river channel) hints at the complexity of the system and the task of reducing the geological units to meaningful hydrogeological layers (see figures 6, 7, and 8). Computer modeling was used as a tool to help unravel the problem of how to regroup the stratigraphic layers into more manageable but realistic hydrostratigraphic units by testing different hydrogeological configurations.

Preliminary delineation of the units was necessary to initialize the data sets for the first computer simulations. Thus, the uppermost layer, a water table aquifer, with an initial hydraulic conductivity estimated to be 3.05×10^{-5} meters/sec (10^{-4} ft/sec), was chosen as layer 1. Within the valley of the Rockaway River, at approximately 152 meters (500 feet) in elevation, the hydraulic conductivity of the sediments is slightly greater. Here the deposits are mainly outwash materials consisting of pebble gravel and sand. Toward the outer limit of the Terminal moraine and also toward the bedrock ridges, the hydraulic conductivity of the sediments decreases. This is because nonstratified drift, namely till, comprises the material of the water table aquifer beyond this point. However, delineating the boundaries of the water table aquifer is a rather straightforward procedure up to the point where it is juxtaposed against the front of the Wisconsinian Terminal Moraine. Here, the subdivision of a thick pile of glacial till with interfingering stratified deposits into meaningful hydrogeological units of similar hydrogeological properties is not obvious. The problem is complicated by a lack of detailed information on the local ground water flow patterns. Therefore, until more data is collected, the discretization of the complex pile of unconsolidated sediments along the terminal moraine and slightly beyond where till, lake bottom sediments and glaciofluvial sands and gravels interfinger is based mainly on hydraulic conductivity. It is recognized that this initial gross lumping of different strata, based mainly on scanty hydraulic conductivity data, will have to be adjusted either as new data become available or as calibration of the model proves impossible because of these preliminary assumptions.

Knowing that there is supporting evidence for the existence of three layers, each of which exhibit contrasting hydraulic conductivities, makes it appealing to readily assume that this is the case throughout the study area. However, the fact that the geological layers (or stratigraphic units) believed to make up the valley fill deposits do not coincide with a three aquifer system and the fact that there is no supporting data for three separate aquifers (there may be more or there may be less) existing in the vicinity of the terminal moraine precludes such a hasty conclusion. It appears that the fundamental hydrogeological difference between layer 2 and layer 3 in the locality of the Boonton well field is that layer 2 is much thicker and exhibits a greater variability in both horizontal and vertical flow. Both aquifers are thought to be confined, although it is quite possible that the layers acting as confining units are not continuous. Also, the pump test data indicate that the confining units are actually quite permeable. This strongly implies that in certain locations semi-confining conditions exist. Once again, the validity of this initial assumption was tested during model calibration.

5.0 AQUIFER PARAMETERS

An aquifer or an aquifer system is described most accurately by its characteristic parameters. Aquifer parameters include Transmissivity (T), Hydraulic Conductivity (K), Leakance (Vcon), and Saturated thickness(b). Representative values of these parameters for each modeled layer are prerequisites for model calibration.

Values of T, K , Vcon,and b for this study were determined from a combination of sources. These included extrapolation of parameters from previous studies conducted in similar valley fill aquifers (Meisler,1976; Canace et al., 1981), extrapolation of specific capacity data obtained from records of the first drilling of wells in the area, and interpretation of existing well logs.

Table 5 lists the pertinent data extracted from previous studies that were used to help determine possible values for some of the aquifer characteristics of the Rockaway Valley aquifer system. The various ranges of values indicates that, in general, valley fill aquifers are heterogeneous. Included in Table 5 is a brief discussion of the means by which the various data were obtained.

5.1 Determination of Tranmissivities

Very few transmissivity values have been determined from the analysis of pump tests and field investigations conducted in the Rockaway Valley. Therefore, estimates of

Table 5 Selected Values of Transmissivity and Storativity from Previous Studies in Glacial Valley Fill Environments

Source	Transmissivity (m^2/d) T_{low}	T_{high}	$T_{ave.}$	Storativity/ Specific Yield $S_{ave.}$	Methodology
Vecchiolo and Nichols (1966)	1,198	3,985	2,592	No Data	Analysis of pump test data from Morris Co., NJ
Meisler (1976)	482	3,211	1,828	$4x10^{-6}$/0.16	Results of calibrated computer model
Moretrench American Corp., 1975	125	1,046	685	No Data	Analysis of pump-down test s/t plots by the Jacob Non-Equilibrium Method
	(deep aquifer)				
	598	2,352	1,475		
	(intermediate aquifer)				
Tetra Tech, Inc., 1978	36	2,389	566	No Data	Based on the relationships K=2,000 Q/S(L) and T = Kb
Canace and others (1981)	210	1,817	721	$2.6x10^{-3}$	Analysis of aquifer pump test data by methods developed by Pricket (1965), Jacob (1963), Cooper and Jacob, (1946), Theis (1935), and Theim (1906)

transmissivity had to be determined through the application of the Theis equation to existing specific capacity data (Heath, 1983). Specific capacity is the pumping rate of the well divided by the observed drawdown (Q/s). Modifications to the Theis equations result in the following equation:

$$(1) \quad T = \frac{W(u)}{4\pi} \times \frac{Q}{s}$$

where $W(u)$ = the well function of u

$$(2) \quad u = r^2 s / 4Tt$$

T = Transmissivity

S = Storage coefficient

Q = Pumping rate

s = Drawdown

t = Length of the pumping period

r = effective radius of the well

For convenience, $W(u)/4\pi$ can be expressed as a constant. This is accomplished by calculating values for u (equation 2). The initial estimate of u is obtained by substituting representative values of S, T, r, t, determined from field observations, into equation 2. Once a value of u is determined then reference to a table of selected values of W(u) that correspond to the respective 1/u values

(Table 6) enables a constant for $W(u)/4\pi$ to be used in equation 1. Table 7 lists the calculated transmissivity values and the layer to which the values correspond. Transmissivities, calculated from previous studies in similar valley fill environments were used as the initial value of T in equation 1. A range of possible transmissivity values are calculated based on the highest and lowest value of transmissivity determined from previous investigations. The density of transmissivity values for the area was limited by well records with specific capacity data. There was no systematic way to judge the validity of the specific capacity data included in the well records, therefore it was assumed that all values were applicable. Calculated transmissivity values were plotted on the base map and then contoured (Figure 10). This allowed an interpolation of transmissivity to be made where no dat existed.

The use of the Theis equation to calculate transmissivity is not without certain limitations. Assumptions which accompany Theis type solutions include, but are not limited to the following (Heath, 1983):

1. The Transmissivity of the aquifer tapped by the pumping well is constant but values obtained are only applicable within the confines of the steady state cone of depression.

Table 6 Selected values of W(u) for values of 1/u (from Heath, 1983).

1/u	10	7.69	5.88	5.00	4.00	3.33	2.86	2.5	2.22	2.00	1.67	1.43	1.25	1.11
10^{-1}	0.219	0.135	0.075	0.049	0.025	0.013	0.007	0.004	0.002	0.001	0.000	0.000	0.000	0.000
1	1.82	1.59	1.36	1.22	1.04	.91	.79	.70	.63	.56	.45	.37	.31	.26
10	4.04	3.78	3.51	3.35	3.14	2.96	2.81	2.68	2.57	2.47	2.30	2.15	2.03	1.92
10^2	6.33	6.07	5.80	5.64	5.42	5.23	5.08	4.95	4.83	4.73	4.54	4.39	4.26	4.14
10^3	8.63	8.37	8.10	7.94	7.72	7.53	7.38	7.25	7.13	7.02	6.84	6.69	6.55	6.44
10^4	10.94	10.67	10.41	10.24	10.02	9.84	9.68	9.55	9.43	9.33	9.14	8.99	8.86	8.74
10^5	13.24	12.98	12.71	12.55	12.32	12.14	11.99	11.85	11.73	11.63	11.45	11.29	11.16	11.04
10^6	15.54	15.28	15.01	14.85	14.62	14.44	14.29	14.15	14.04	13.93	13.75	13.60	13.46	13.34
10^7	17.84	17.58	17.31	17.15	16.93	16.74	16.59	16.46	16.34	16.23	16.05	15.90	15.76	15.65
10^8	20.15	19.88	19.62	19.45	19.23	19.05	18.89	18.76	18.64	18.54	18.35	18.20	18.07	17.95
10^9	22.45	22.19	21.92	21.76	21.53	21.35	21.20	21.06	20.94	20.84	20.66	20.50	20.37	20.25
10^{10}	24.75	24.49	24.22	24.06	23.83	23.65	23.50	23.36	23.25	23.14	22.96	22.81	22.67	22.55
10^{11}	27.05	26.79	26.52	26.36	26.14	25.96	25.80	25.67	25.55	25.44	25.26	25.11	24.97	24.86
10^{12}	29.36	20.09	28.83	28.66	28.44	28.26	28.10	27.97	27.85	27.75	27.56	27.41	27.28	27.16
10^{13}	31.66	31.40	31.13	30.97	30.74	30.56	30.41	30.27	30.15	30.05	29.87	29.71	29.58	29.46
10^{14}	33.96	33.70	33.43	33.27	33.05	32.86	32.71	32.58	32.46	32.35	32.17	32.02	31.88	31.76

Examples: When $1/u = 10 \times 10^{-1}$, $W(u) = 0.219$; when $1/u = 3.33 \times 10^2$, $W(u) = 5.23$.

TABLE 7

Estimates of Transmissivity obtained from Specific Capacity Values and Transmissivity Data from Previous Investigations and Modification of the Theis Equation

Units from which Specific Capacity Data was obtained	Transmissivity (m^2/d) T_{high}	T_{low}
Shallow till (Well W-4, Q/s = 20 m^2/d)	23	20
Shallow stratified drift		
(Well B-3, Q/s = 268 m^2/d)	305	275
(Wells B-1 and B-2, Q/s = 536 m^2/d)	616	546
Intermediate sand and gravel		
(Well B-6), Q/s = 2,682 m^2/d)	3,957	3,612
Deep sand and gravel		
(Well D-8, Q/s = 116 m^2/d)	172	157
(Wells B-4 and B-5, Q/s = 322 m^2/d)	474	433
Deep till		
(Well D-5, Q/s = 125 m^2/d)	185	169
(Well D-7, Q/s = 27 m^2/d)	41	37
(Well W-11, Q/s = 200 m^2/d)	295	270

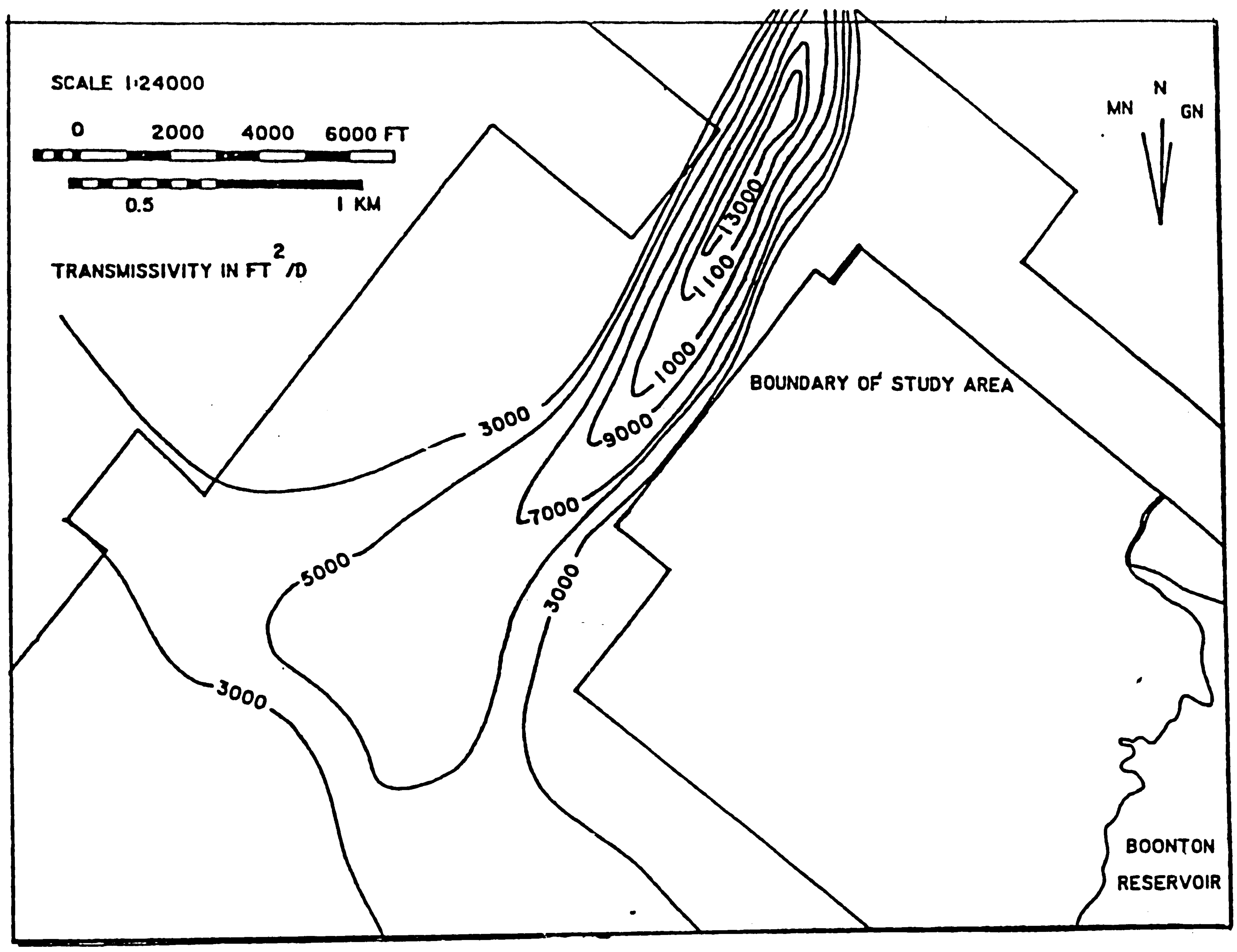


Figure 10 Preliminary transmissivity contour map of the lower aquifer layer in the study area.

2. Water withdrawn from the aquifer is derived entirely from storage and is discharged instantaneously with a corresponding decline in head.

3. The discharging well penetrates the entire thickness of the aquifer and its diameter is extremely small in comparison with the pumping rate, so that storage in the well is negligible.

5.2 Determination of Saturated Thickness and Hydraulic Conductivity

An isopach map of the entire sequence of unconsolidated deposits is shown in Figure 9. This map was constructed from an unpublished bedrock topography map (Stanford,1985) and the Boonton 7.5 minute quadrangle map. The difference between the intersection of a surface elevation contour and a bedrock elevation contour was used to determine the total thickness of the unconsolidated materials at that point. All calculated points and control points (actual thickness determined from wells that have achieved bedrock) were subsequently contoured.

Isopach maps depicting the thickness of each of the stratigraphic layers were constructed from geological cross-sections (Stanford,1985), and data from well logs. Hydraulic conductivity values for each of the layers were calculated once the best estimates of the thickness of each unit at selected localities was determined. However, for this particular study, the model required a hydraulic conductivity map for only the top unconfined aquifer. The equation $K = T/b$ is the basis of the calculations used to construct the necessary hydraulic conductivity map. In this case, b is the saturated thickness of the water table aquifer. The actual saturated thickness of the water table aquifer was not known but it was assumed that the saturated thickness was only slightly less than that of the entire water table aquifer thickness. This assumption seems reasonable because the water table in the Rockaway Valley is close to the surface and the thickness of the unsaturated zone (the difference between the saturated and the entire thickness) is negligible relative to the total thickness. The saturated thickness of the water table aquifer was determined by superimposing the water level contour map over the water table aquifer isopach map. The transmissivity values calculated as outlined in the previous section were the transmissivities used in the solution of the hydraulic conductivity equation. Calculated hydraulic conductivity values were subsequently contoured (Figure 11).

5.3 Determination of Leakance

Determination of the Leakance factors that accurately quantify the flux of water from one

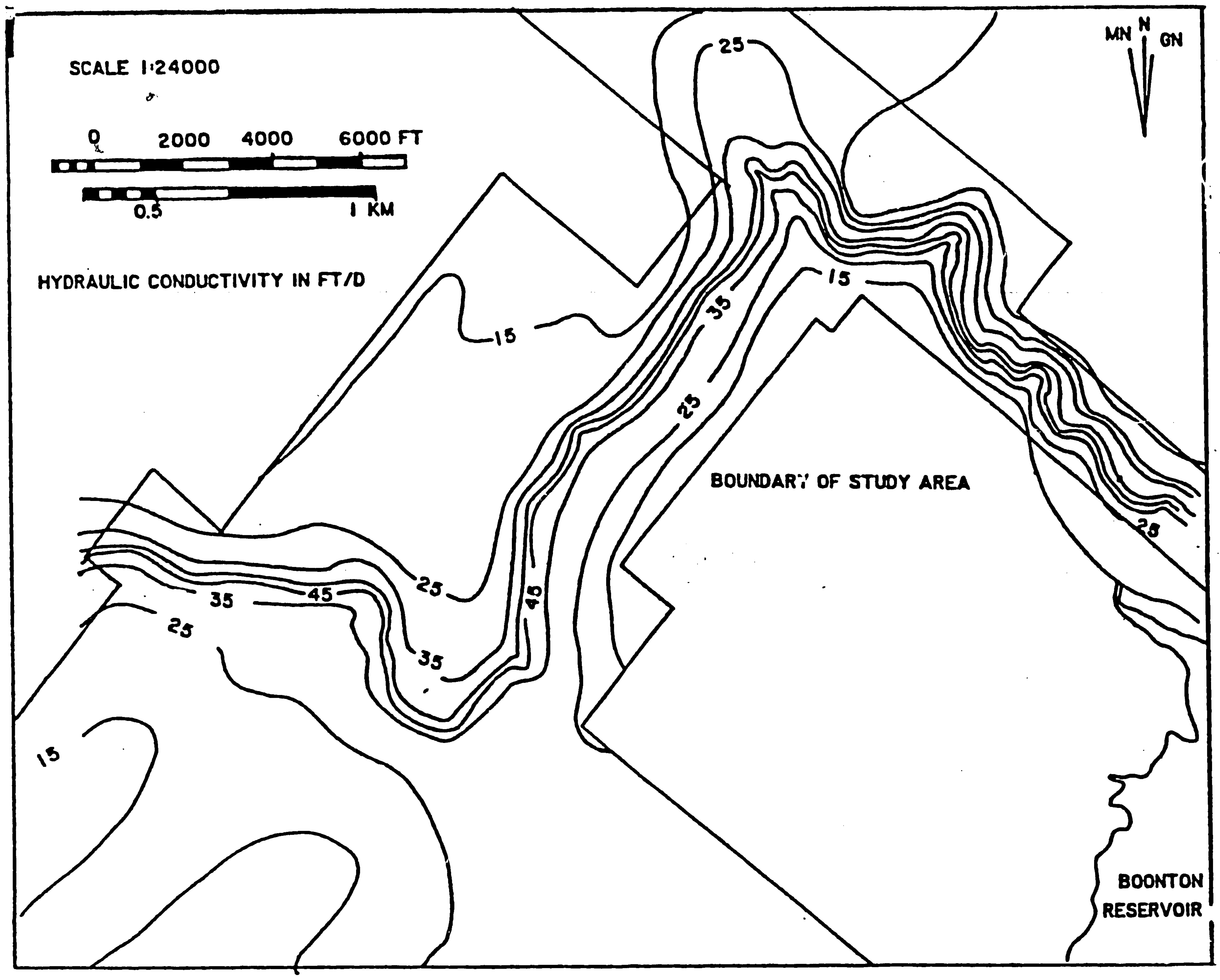


Figure 11 Preliminary hydraulic conductivity map of the water table aquifer in the study area.

glaciostratigraphic unit to the next is not possible from the existing data base. However, the identification of feasible hydrologic relationships between the various units allows qualification of possible leakage occurring between hydrostratigraphic units in the Rockaway Valley multi-aquifer system. Validity of these qualified estimates can be tested by means of the behavior of the model.

According to tests conducted by the Moretrench American Corporation (MTA) at Boonton's well field, there is significant vertical leakage between the various aquifer strata. A pump-down test was conducted at the well field on wells 5 and 6. Well 6 is the only one from the well field which taps the intermediate unit. The intermediate aquifer is believed to be the most productive aquifer in the multi-aquifer system. s/t (drawdown over time) plots and modified s/r (drawdown over distance) plots were constructed by MTA for eleven observation wells. According to MTA:

> After five days of pumping a state of equilibrium had not been achieved and drawdowns continued to increase with time. A comparison of drawdown in three observation wells at approximately equivalent radius, with sensing elements in the shallow, intermediate, and deep aquifer strata indicates that the greatest drawdown occured in the intermediate aquifer. Vertical gradients were established between the deep and intermediate, and the shallow and intermediate aquifers. Pumping at 200 gpm caused recharge water to move from the deep and shallow into the intermediate aquifer stratum. However, in view of the relatively large thickness of the intermediate aquifer it is likely that the principal direction of recharge water is horizontal. Drawdowns in the shallow

aquifer stratum were not as great as those in the intermediate stratum but indicated a fair degree of vertical communication between these two strata (Moretrench American Corporation,1975).

MTA did not quantify the amount of leakage occuring between the different layers. It appears that MTA is suggesting that both of the confining units are sufficiently permeable to transmit water from the overlying and underlying aquifers into the intermediate aquifer. Another possibility is that the source of recharge is the release of water from storage in the confining beds. Either interpretation implies the existence of semiconfining conditions in the intermediate aquifer and possibly in the deep aquifer. Unfortunately, the copy of the MTA report available to the author did not include any of the distance-drawdown plots or the data used to construct such plots. Therefore, the resultant plots from the Boonton well field test could not be compared to the family of type curves which reflect the many various leaky artesian conditions that may exist. It is important to note that the well field is in close proximity to the river and that this recharge boundary must have some impact on the results of any aquifer tests if the pump test is conducted over a long enough period of time.

Since model calibration was conducted under only steady state conditions determination of storativity values for the hydrostratigraphic units was not necessary.

6.0 SURFACE AND GROUNDWATER HYDROLOGY

6.1 Drainage Basin Characteristics

The Rockaway River lies within the larger, Passaic Drainage basin. The Passaic River Basin is a 90 kilometer (56 miles) long by 42 kilometers (26 miles) wide watershed with an area of 2,422 square kilometers (935 square miles). About 15.5 % of the Passaic River Basin is occupied by the Rockaway watershed. The Rockaway Drainage Basin is 351 square kilometers (135.7 square miles) with 300 square kilometers (116 square miles) comprising the Upper Rockaway and 51 square kilometers (19.7 square miles) comprising the Lower Rockaway Basin. The Rockaway River, with its headwaters originating in the Highlands region, is one of the major tributaries to the Passiac River. Within the New Jersey Highlands, the stream gradient is extremely high , ranging from 1:3 to 1:100, and the course of the river generally follows the northeast-southwest trending ridges of the Highlands (Summers et al., 1978). Narrow stream valleys with steeply dipping valley walls are the general rule. Streamflow is to the southwest in the northern reaches of the drainage basin and to the northeast in the southern reaches. The study area lies within the Upper Rockaway Drainage Basin and covers approximately 30 square kilometers (11.5 square miles). The Rockaway River, where it cuts transverse to the bedrock ridges and dissects the

Wisconsinan Terminal Moraine in Denville to the outlet of the river into the Boonton Reservoir, is a total of 13 kilometers (eight miles) long. The channel gradient is about 2.5 meters per kilometer (13 feet per mile) along its entire flow path within the study area but steepens abruptly from the town of Boonton to the Boonton Reservoir, and achieves a maximum gradient just upstream from the Ramapo Fault, with a channel slope of approximately 133 meters per kilometer (700 feet per mile). The USGS maintains a stream gaging station 2.9 kilometers (1.8 miles) upstream from the Boonton Reservoir. The average discharge, calculated from 45 years of record (1937-1982), is 6.4 cubic meters per second (228 cubic feet per second), with extremes of a maximum discharge of 158 cubic meters per second (5,590 cfs), on April 5,1984 and a minimum discharge of 0.28 cubic meters per second (10 cfs) on August 10, 1966 (Bauersfeld et al., 1982).

The Rockaway Watershed is stippled with numerous lakes, ponds, and reservoirs. Within the study area 5.2 percent of land is occupied by surface water bodies. However, locally the primary source of potable water is from ground-water reservoirs.

6.2 Stream-Aquifer Interaction

The flow dynamics between the river and the underlying aquifer is best determined by seepage runs which are determined from the analysis of stream discharge measurements taken at selected intervals along reaches of the river. Figure 12 shows the location of the USGS stream-gaging stations, the distance between each measuring site, and the number assigned by the USGS to each station. Table 8 lists the discharge for each location and the date on which the data were collected. There are three separate sets of seepage runs that were conducted by the USGS and another set of discharge measurements collected by personnel from Tetra Tech, Inc., (Summers et al., 1978). Comparison of the total net flux of water from station 30B (Tetra Tech's station 4) to station 68 (Tetra Tech's station 5), approximately 13 kilometers (8 miles) downstream, indicates that there is a significant fluctuation in streamflow among the various sampling dates. Some of the disparity in the 1977 data presented in the report by Tetra Tech, Inc. is due, according to the report, to differences in irrigation demands. Pumping would be greater during the late summer months. By early November, the effects of the summer pumping is negligible on streamflow in the Rockaway River. More importantly, however, is that the discrepency between stream discharge measured for each site may also be attributed to the fact that error is introduced during field measurements. According to records and personnel from the USGS, a deviation of plus or minus five percent from observed streamflow is, under the best of conditions, the

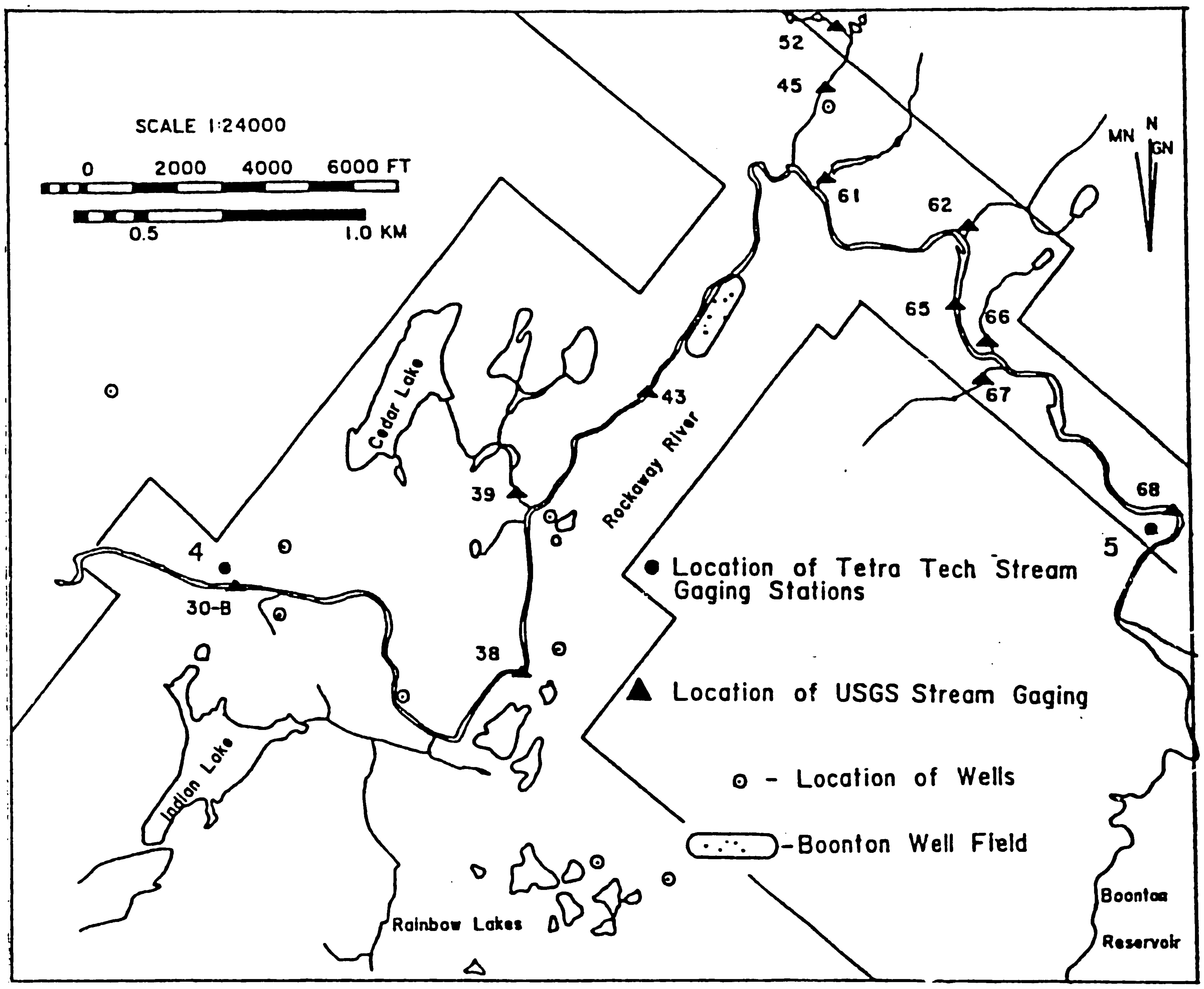


Figure 12 Map depicting location of stream-gaging stations and location of wells that were monitored for water levels during seepage runs.

Table 8 Stream Discharge Data

USGS Station	Discharge (L^3/s) 9/19/85	4/12/85	10/16/84	Net Seepage Between Selected Stations 9/19/85	4/12/85	10/6/84
	Date Streamflow Data was Collected					
30 B	27.5 ft^3/s	157.0 ft^3/s	27.8 ft^3/s			
	(0.778 m^3/s)	(4.443 m^3/s)	(1.786m^3/s)			
38	39.6 ft^3/s	162.0 ft^3/s	30.8 ft^3/s	+12.1 ft^3/s	+5.0 ft^3/s	+3.0 ft^3/s
	(1.721 m^3/s	(4.585 m^3/s)	(0.872 m^3/s)	+0.342 m^3/s	+0.142 m^3/s	+0.085 m^3/s
				(From Station 30 B to 38)		
39(T)	0.34 ft^3/s	1.80 ft^3/s	0.113 ft^3/s			
	(0.010 m^3/s)	(0.051 m^3/s)	(0.003 m^3/s)			
43	41.7 ft^3/s	165.0 ft^3/s	30.8 ft^3/s	+1.75 ft^3/s	+1.2 ft^3/s	+0.113 ft^3/s
	(1.180 m^3/s)	(4.669 m^3/s)	(0.872 m^3/s)	+0.049 m^3/s	+0.034 m^3/s	+0.003 m^3/s
45(T)	0.0 ft^3/s	13.5 ft^3/s	0.0 ft^3/s	(From Station 38 to 43)		
		(0.382 m^3/s)				
61(T)	0.03 ft^3/s	0.1 ft^3/s	0.0 ft^3/s			
	(849 cm^3/s)	(0.003 m^3/s)				
62(T)	0.021 ft^3/s	1.20 ft^3/s	0.035 ft^3/s			
	(594 cm^3/s)	(0.034 m^3/s)	(99.1 cm^3/s)			
65	36.7 ft^3/s	174.0 ft^3/s	35.9 ft^3/s	-4.95 ft^3/s	-5.8 ft^3/s	+5.07 ft^3/s
	(1.038 m^3/s)	(4.924 m^3/s)	(1.016 m^3/s)	-0.14 m^3/s	-0.164 m^3/s	+1.056 m^3/s
66(T)	0.0 ft^3/s	0.13 ft^3/s	0.002 ft^3/s	(From Station 43 to 65)		
		(0.004 m^3/s)	(56.6 cm^3/s)			
67(T)	0.057 ft^3/s	1.40 ft^3/s	0.069 ft^3/s			
	(0.002 m^3/s)	(0.039 m^3/s)	(0.002m^3/s)			
				(From Statio 65 to 68)		
68	39.6 ft^3/s	175.0 ft^3/s	37.0 ft^3/s	+2.84 ft^3/s	-0.53 ft^3/s	+1.03 ft^3/s
	(1.21 m^3/s)	(4.95 m^3/s)	(1.047 m^3/s)	+0.08 m^3/s	-0.015 m^3/s	+0.029 m^3/s
Net Seepage from Station 30 B to Station 68				+11.18ft^3/s	-0.13 ft^3/s	+9.213 ft^3/s
				+0.316 m^3/s	-0.004 m^3/s	+0.260 m^3/s
Seepage Rate				+1.4 cfs/mi	-0.016 cfs/mi	+1.15 cfs/mi
				+24.4 ccs/m	-0.311 ccs/m	+20.199 ccs/m
Net Seepage Rate from Tetra Tech's Station 4 to Station 5 (Nov. 18, 1977)				+4.08 cfs/mi		
				+71.75 ccs/m		

lowest possible error. Regardless of the cause for flow differences, it seems apparent, based on the high variability in the discharge measurements taken during various baseflow conditions, that the volume of streamflow in the river is probably influenced by fluctuations in the water table levels of the underlying aquifer. Therefore, the initial assumptions that the stream-aquifer system possesses a strong hydrologic continuity and that the stresses imposed upon the aquifer are reflected by the net flux of water between the two systems were used as guidelines toward the construction of a hydrogeological model.

One of the most apparent differences between the seepage runs taken in September 1985 and October 1984 is the net flux observed between stations 30B and 38 and between stations 43 and 65. Although the two different seepage runs were not done at the exact same time of the year it is assumed that they both adequately reflect baseflow conditions because they were both undertaken during a period of low seasonal recharge. Therefore, it is highly unlikely that the difference of 0.25 m^3/s or 9 ft^3/s (the difference between the October 1984 data and the September 1985 data for the reach between stations 30 B and 38) can be accounted for by seasonal fluctuations. Although explanations such as increase in water demands via evapotranspiration or unusually high base flow in September 1985 cannot be ruled out, it is more likely that this discrepancy is primarily a result of human induced stresses onto the system or that the error introduced during stream gaging produced the noticeable differences between the net flow for the two separate dates. For example, when a range of plus or minus 5 % error is considered in the calculations, the net flux of the September 1985 data between station 30 B to 38 ranges from +16.0 to +9.0 ft^3/s while the October 1984 net flux between these stations ranges from +6.0 to 0.0 ft^3/s. Consideration of the error margins indicates that the discrepancy between the two seepage runs conducted on different days may not be as great ($9.0 - 6.0 = +3.0$ ft^3/s) or may be greater ($16.0 - 0.0 = +16.0$ ft^3/s) than that shown by calculations in which this error has been disregarded. Dividing this troublesome reach into smaller reaches with more stream gaging stations would help to identify the source of this increased flux of water along the river. It would also eliminate the possibility that the influx is due to an ungaged tributary along that portion of the reach. The other major difference between the two sets of base flow data (net flow between stations 43 and 65) presents a curious problem. It is interesting that the absolute value of the net flow is nearly the same for both dates (about 5 cfs). However, there is a sign difference, indicating that on September 19, 1985 this reach is losing and that on October 16, 1984 it is gaining. To add to the mystery, analysis of the April 12, 1985 data (high flow) for this reach shows that the absolute value of the flux is once again very nearly the same as for the two separate low base flow values. However, when the error range of plus or minus 5 % of the measured discharge is considered the net seepage between these stations for the three separate dates falls in the range of -9.0 to -1.0 ft^3/s (September, 1985), +26.0 to -8.0 ft^3/s (April, 1985), and -9.0 to +2.0 ft^3/s (October 1984). Despite this range in seepage values, the water level contour map of the unconfined aquifer and the scattered water level measurements indicate that the hydraulic gradient is toward the river throughout the study area. Thus there should be no losing reaches unless some other factors are influencing the hydraulics between the stream and the water table aquifer. In this particular reach (from Station 43 to 65) there are no water level measurments available that would confirm the possibilty of a losing reach of the river. Plausible explanations for these observed anomalies in the streamflow data are more thoroughly discussed in a later section where the interrelationship between flux in the river, groundwater levels, and the impact of the geology on the hydraulics of the system is evaluated. Overall, comparison of the total net flux, when considered as a net discharge per mile, for the entire eight miles of the river that flows through the study area, renders this disparity along a specific reach less conspicuous (Table 8). Therefore, when the system is viewed as a whole local differences tend to be minimized.

6.3 Water Level and Potentiometric Contour Maps

In September 1985, the USGS, as part of their data base acquisition for the Rockaway Valley Drainage Basin, monitored water levels in selected wells while conducting concurrent seepage runs. Location of wells monitored on September 19, 1985 are shown in Figure 12. The water level data were plotted and then contoured by the author. Since the valley fill materials make up a multi-aquifer system, judgments as to which layer the water level measurements corresponded were based on the information supplied by well records. One can infer the layer from which the well is screened in by analysis of data on the casing length and the screening interval of each of the wells. Due to the low density of wells in which water levels were measured during September, considerable extrapolation was necessary to complete the contour maps. Two separate water level maps were drawn. Figure 13 represents the altitude of the water table in the unconfined aquifer. The principle that the water level in water table aquifers commonly mimics the topography was applied during the construction of this map. The hydraulics of the aquifer-stream system and the influence of pumping from shallow wells were also considered. Figure 14 illustrates the attitude of the

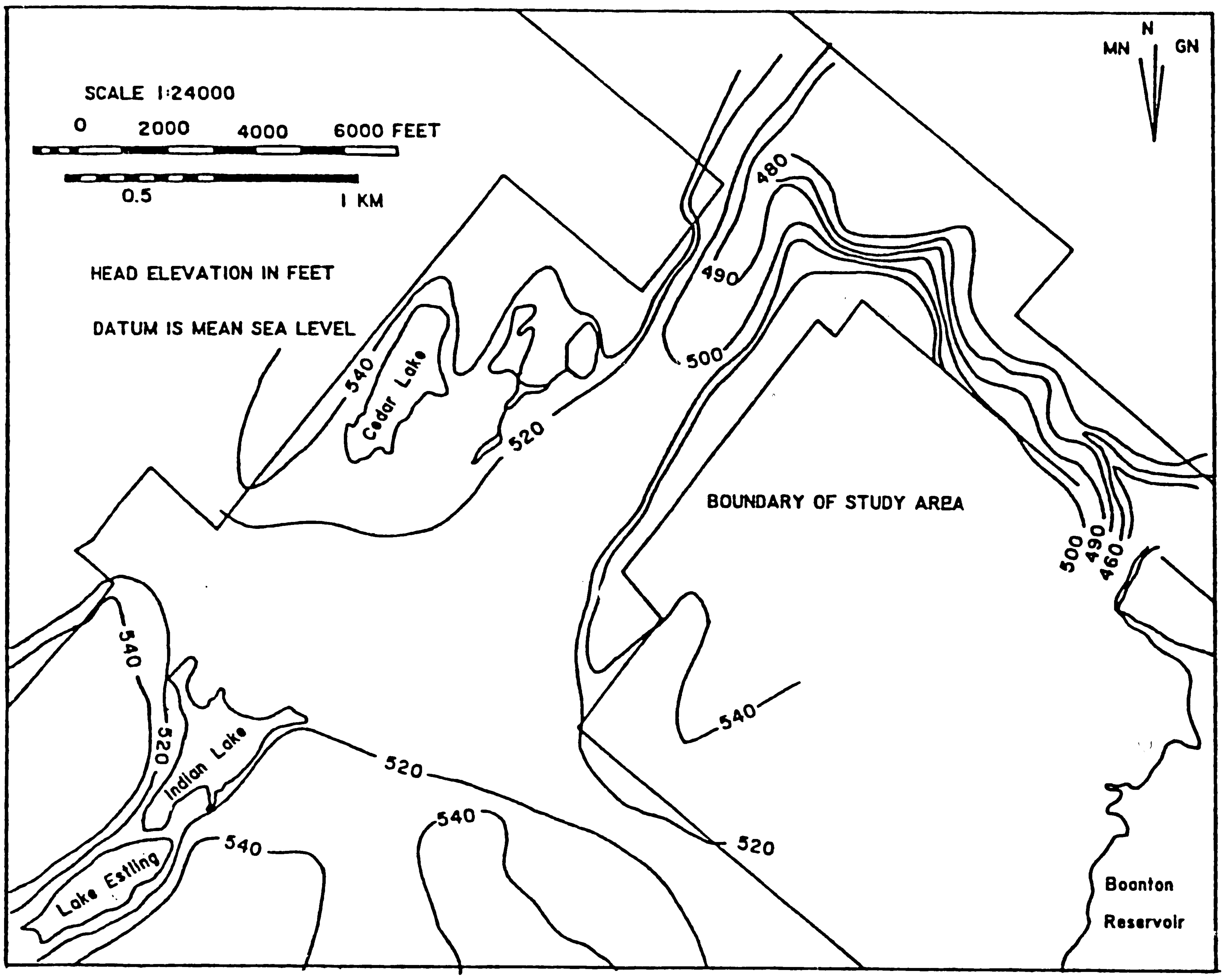


Figure 13 Water table elevation contour map constructed from water level measurements taken concurrently with streamflow data.

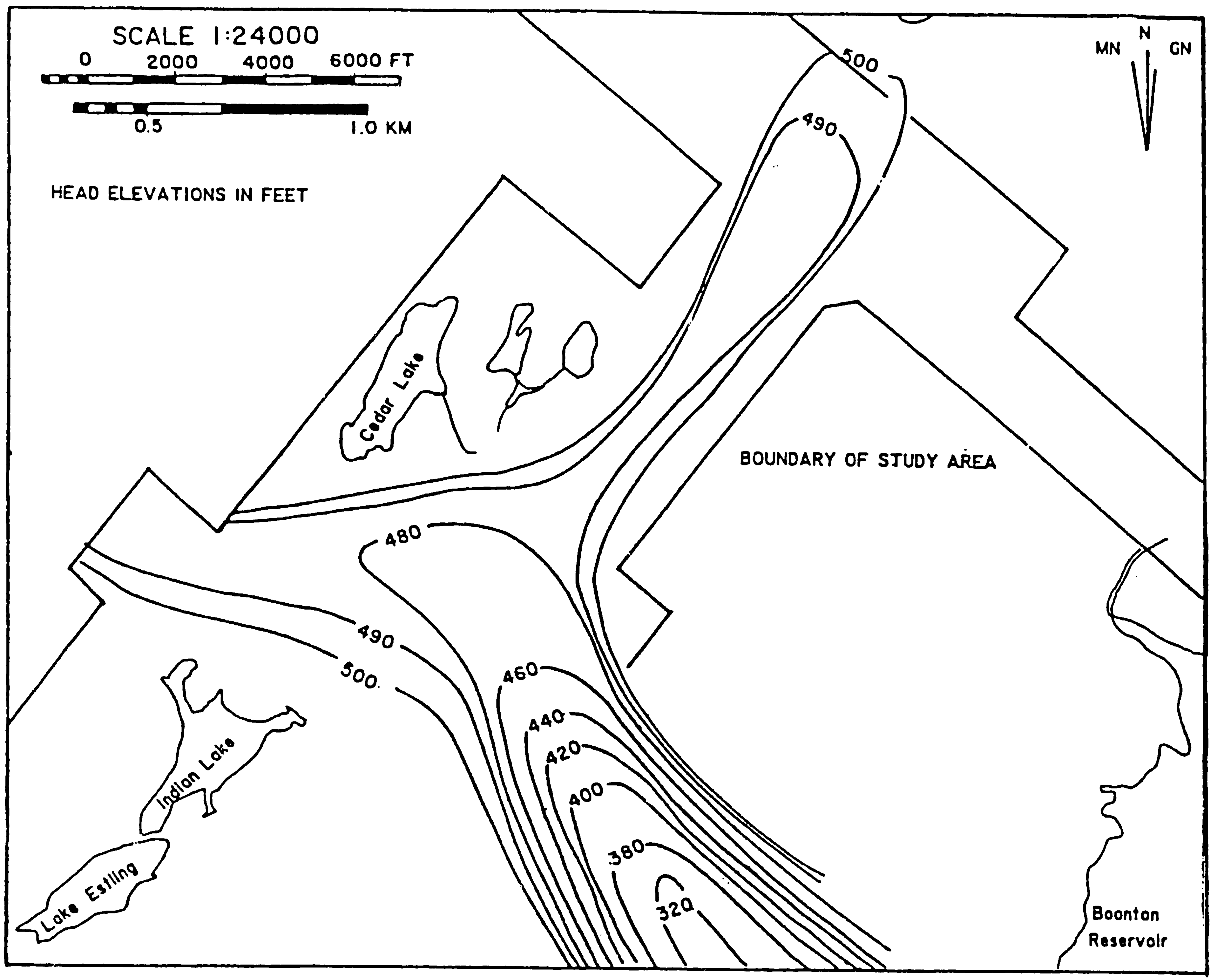


Figure 14 Potentiometric surface contour map constructed from water level measurements taken concurrently with streamflow data.

potentiometric surface for the layers modeled as confined. Because data base acquisition has just begun it was decided that not enough reliable data existed to support the construction of two separate potentiometric maps for the two potential confined aquifers. Therefore, the data was combined and one map drawn that most closely represents the groundwater flow patterns in all possible underlying aquifer units. Supporting evidence for this generalization is given by the fact that it is not totally certain that two confined aquifers exist throughout the study area. Also, the delineation of the sediments into stratigraphic units indicates the existence of one confined aquifer within the present day valley of the Rockaway River and the possible existence of many individual lens-like aquifers in the southeast-southwest region of the study area. These many unknowns do not warrant a complex subdivision of the stratigraphic units into hydrostratigraphic units. The main distinction between the various units is probably only discernable by the respective conductivities of the units. If this is the case, then modeling this thick pile of sediments as heterogeneous materials should accomodate this grouping of all possible lower aquifer units into one basal aquifer.

6.4 Water Budget Calculations

A hydrologic water budget for the Upper Rockaway River basin was calculated for three different time periods from preliminary data by Phil Harte (USGS,1985). A long term average (1921-1950), a dry year (1965), and a wet year (1984) are presented in Table 9. A water budget for 1985 was not included with these preliminary calculations. However, a 1985 budget was estimated by the author based on 1985 precipitation data, measured runoff, and the three water budgets.

An annual hydrologic budget can be expressed as the following:

$$P = Q + E + dS/dt$$

where P = Average Annual Precipitation

Q = Average Annual Stream Discharge

E = Average Annual Evapotranspiration

dS/dT = Change in Storage of both ground-water and surface-water in the basin

The above is a very generalized hydrologic equation. More explicitly Q can be separated into its surface run-off component (Q_s) and its ground-water discharge component (Q_g). In addition, when data such as yearly consumption and groundwater inflows or outflows (for basins in which the surface water and groundwater divides do not coincide or when the boundaries of a study area do not coincide exactly with the surface drainage divide) are available, they are included in the hydrologic budget. It was assumed that the

Table 9: Preliminary hydrologic budget for the Upper Rockaway River Basin

Budget	Components	Subcomponents	Cm in Water Over Area (300 km^3)	Percentage of Total Budget
Long Term Average (1921-1950)	Precipitation		122	100
	Water Loss		58	48
	Runoff		64	52
Water Year 1965 (Dry Year)	Precipitation		87	100
	Water Loss		66	70
		Evapotranspiration	58	
		Consumed	8	
		Ground-water outflow	0.1	
	Runoff		26	33
		Ground	18	
		Surface	8	
Water Year 1984 (Wet Year)	Precipitation		172	100
	Water Loss		60	35
		Evapotranspiration	52	30
		Consumed	8	4
		Ground-water outflow	0.1	0.1
	Runoff		112	65
		Ground	71	
		Surface	41	

Above preliminary water budget calculations by Harte (USGS, 1985)

Budget	Components	Subcomponents	Cm in Water Over Area (300 km^3)	Percentage of Total Budget
Water Year 1985	Precipitation		104	100
	Water Loss		72	70
		Evapotranspiration	65	
		Consumed	8	
		Ground-water outflow	0.1	
	Runoff		31	30
		Ground	22	
		Surface	9	

1985 water budget calculations based on extrapolated percentages from Harte's calculations, 1985 precipitation data, and the average of maximum and minimum stream discharge data.

*change in storage of the basin over a yearly period is zero. However, for such short intervals the assumption that storage of the basin does not change can be erroneous (Viessman and et al., 1977). Historical records of water levels and streamflow data would help to indicate the accuracy of this assumption. The equation then becomes:

$$P = Q_s + Q_g + E + Q_{con,} + Q_{out}$$

where
- P = Average Annual Precipitation
- Q_s = Average Annual Surface Run-off
- Q_g = Average Annual Baseflow
- E = Average Annual Evapotranspiration
- $Q_{con.}$ = Total Annual Amount of Water Consumed from Basin
- Q_{out} = Net Flow of Groundwater into or out of the Basin

An estimate of the 1985 hydrologic budget was based mainly on the percentage of the components that comprise the total 1965 water budget. For example, in 1965, 70 % and 30 % of the yearly precipitation was accounted for by water loss ($E + Q_{con} + Q_{out}$) and by run-off ($Q_s + Q_g$), respectively. These percentages were maintained during the calculation of the 1985 water budget. The 1965 water year was labeled as a dry year with an annual precipitation of 86.7 centimeters (34.12 inches). The 1985 water year with an annual precipitation of 104 centimeters (40.76 inches) is believed to resemble conditions of a "dry" year much more closely than a "wet" year such as 1984 where the annual precipitation is 172 centimeters (67.68 inches) (Table 9). Table 9 also lists the calculated water budget and the respective percentages of the individual components for the 1985 water budget. Many assumptions accompany this estimate. It was presupposed that inflows to the system equal all outflows from the system. Antecedent conditions imposed by the 1984 wet year on the water budget for 1985 were not taken into consideration during these calculations. Total runoff was determined by averaging discharge values for a baseflow period and for a high flow period. A recharge rate of 7.32×10^{-9} meters/second (2.4×10^{-8} feet/second) or an annual total of 22.1 centimeters/year (8.7 inches/year) for the Rockaway River basin was estimated from this 1985 budget and proved to be a reasonable recharge value when simulated by the model.

7.0 MODEL ORGANIZATION

The computer modeling was accomplished using A Modular Three-Dimensional Finite-Difference Ground-water Flow Model (Modflow) written by Michael G. McDonald and Arlen W. Harbaugh (1984). Modflow is capable of simulating groundwater flow in three dimensions. A block-centered finite-difference method is used to solve the system of simultaneous linear algebraic difference equations that are used to describe the head distribution within a particular system. This approach allows an approximate numerical solution to the partial differential equation that describes the three-dimensional movement of groundwater through porous materials.

7.1 Discretization Convention

Prior to model simulations the aquifer system needs to be discretized into individual blocks or cells. This is accomplished by designing a grid which optimally divides the horizontal projection of the study area into separate rows and columns. Both the design and the orientation of the grid with respect to the physical system should be such that it accurately and efficiently represents the geology and hydraulics of the area to be modeled. Vertical discretization of the study area is accomplished by delineating the hydrogeological system into discrete layers which may be modeled as aquifers, aquicludes, or aquitards.

Very often confining units need not be modeled, although the effects of a confining layer may be simulated. A discretization scheme that is adaptable to both a cartesian coordinate system (x,y,z) and to normal computer array convention (i,j,k) is used in Modflow. With this convention, any point in the system is assigned to a particular node or cell. A node is easily located by referring to its i,j,k coordinates or by its row, column, and layer. A node or a cell represents a three-dimensional piece of the actual aquifer. Each node is modeled as either a constant head cell, a no-flow cell, or a variable head cell. It is important to note that the hydraulic properties assigned to a particular cell in the model are treated by Modflow as an average of the range of properties that commonly exist in that simulated portion of an aquifer. Therefore, during model computations the initial hydraulic parameters assigned to the cell are constant throughout the cell and the resultant calculated head is also an average representation for the actual distribution of heads that may exist in the cell. The importance of designing a grid (assignment of rows, columns, and layers) that most accurately depicts the geology and hydrology of a system cannot be overemphasized. Effort and success in this area optimizes the practicality, efficiency, and accuracy of the model.

7.2 Modflow's Package Concept

Modflow is organized into discrete Packages. Each Package is implemented to help simulate a specific part of the system to be modeled. The Basic (BAS) Package and the Block-Centered Flow (BCF) Package are the essential Packages of Modflow. In the BAS package the following information is specified: grid dimensions, number of active layers, number of time steps, and the length of the stress period for each simulation. It is also the package in which starting heads for each time step and for each layer are specified. Coded arrays that denote whether a cell is to be modeled as a no-flow node or a constant head node or a node whose head varies with time are specified in the BAS package for each active layer. These data sets act as the fundamental framework which identifies the uniqueness of each model.

How each modeled layer is to be simulated is stated by the user in the BCF Package. Layers can be modeled as confined, unconfined, or convertible between confined and unconfined. Arrays depicting the hydraulic conductivity (for unconfined layers) or the transmissivity (for confined layers) are necessary input into the BCF package. The bottom elevations of simulated unconfined aquifers and the vertical conductance between layers other than the very bottom layer are also specified here. Calculation of the conductance components of the finite-difference equation and of the terms used to designate the flux of water from storage (if it is a transient simulation) are handled in the BCF Package. This information allows the volume and the rate of water flowing between adjacent cells to be computed.

Most of the other packages included in Modflow are employed when additional external stresses need to be simulated. These include a River Package, a Recharge Package, a Well Package, a Drain Package, and a Evapotranspiration Package. Each package offers numerous options whose functions are to make the simulation representative of the conditions that actually exist in the aquifer system. Oftentimes, lakes are modeled in the River Package, mostly because this allows the surface water body to be more realistically simulated. All Packages, except Evapotranspiration, were used at some stage throughout the development of the Rockaway Valley model. Also, two different iterative techniques, the Strongly Implicit Procedure and the Slice-Successive Overrelaxation method are available to the user. The Strongly Implicit Procedure was used as the iterative method for solving the linear equations which describe the flow system.

8.0 Goals and Guidelines of the Model Calibration Phase

The primary goal of the modeling phase was to adequately discretize the aquifer system and to obtain reasonable values for the various aquifer parameters. The geology of the study area is complex and the delineation of the deposits into stratigraphic units does not coincide exactly with the implied existence of hydrostratigraphic units. Therefore, before attempts at fine tuning the model can be made it is the author's conviction that it is first necessary to discretize the multi-aquifer system such that both the geology and hydrology of the study area are accurately represented. Also, the many remaining unknowns associated with the necessarily thorough description of the physical properties of the various units greatly hinders the attempt to achieve a unique solution (or an accurate and realistic model) of the hydrogeological regime. Manipulation of the various parameters may result in matching observed water levels with calculated water levels and may result in achieving identical fluxes in river flow for both observed and calculated situations but during the process one may inadvertantly create a model that in other respects has little resemblance to the real system. Hence, it is crucial to maintain controls on those parameters that are known with certainty and to set limits for those parameters that are inferred.

Assumptions which had an effect on the direction of the modeling phase should also be noted. In particular, during model simulations it was assumed that the underlying bedrock is impermeable and has no impact on the hydrogeological system of the valley fill materials. When a detailed water budget is calculated for the area then the validity of this and other assumptions can be assessed.

The following list itemizes those principles that were adhered to during the trial and error calibration of the model.

1. All model runs were conducted under steady-state conditions.

2. The aquifer system was modeled as heterogeneous and horizontally isotropic (hydraulic conductivity in the row direction equals the hydraulic conductivity in the column direction.

3. The starting head elevations used to initialize the steady-state configuration of heads for the computer model adequately represented the head distribution existing in the study area.

9.0 GRID SELECTION

A series of grids, each one slightly more detailed than its predecessor, were used during model simulations. Grid design was modified as more field data became available and as more insight into the system was gained from the previous simulation.

9.1 Grid One

A preliminary design with fourteen rows, ten columns, and two layers was used as the starting grid. Figure 15 illustrates the important aspects of the grid, however, it is not presented in its entirety. Each cell was a square with a row and column length of 762.3 meters (2,500 feet) and each represented 0.57 km^2 (0.22 mi^2) of the entire 30 km^2 (11.5 mi^2) study area. The top layer was modeled as unconfined with thickness varying from 1.5 meters (5 feet) along model boundaries where Precambrian rock lies close the surface to 9.1 meters (30 feet) in the valley. Layer 1 was limited to the valley where stratified drift has been mapped on the surface. Its edges were terminated toward the southwest abutment formed by the Wisconsinan Terminal Moraine and at the northeast edge of the study area where the river turns sharply toward the northeast. The bottom layer was modeled as a layer that is partially convertible between confined and unconfined. Maximum thickness existed

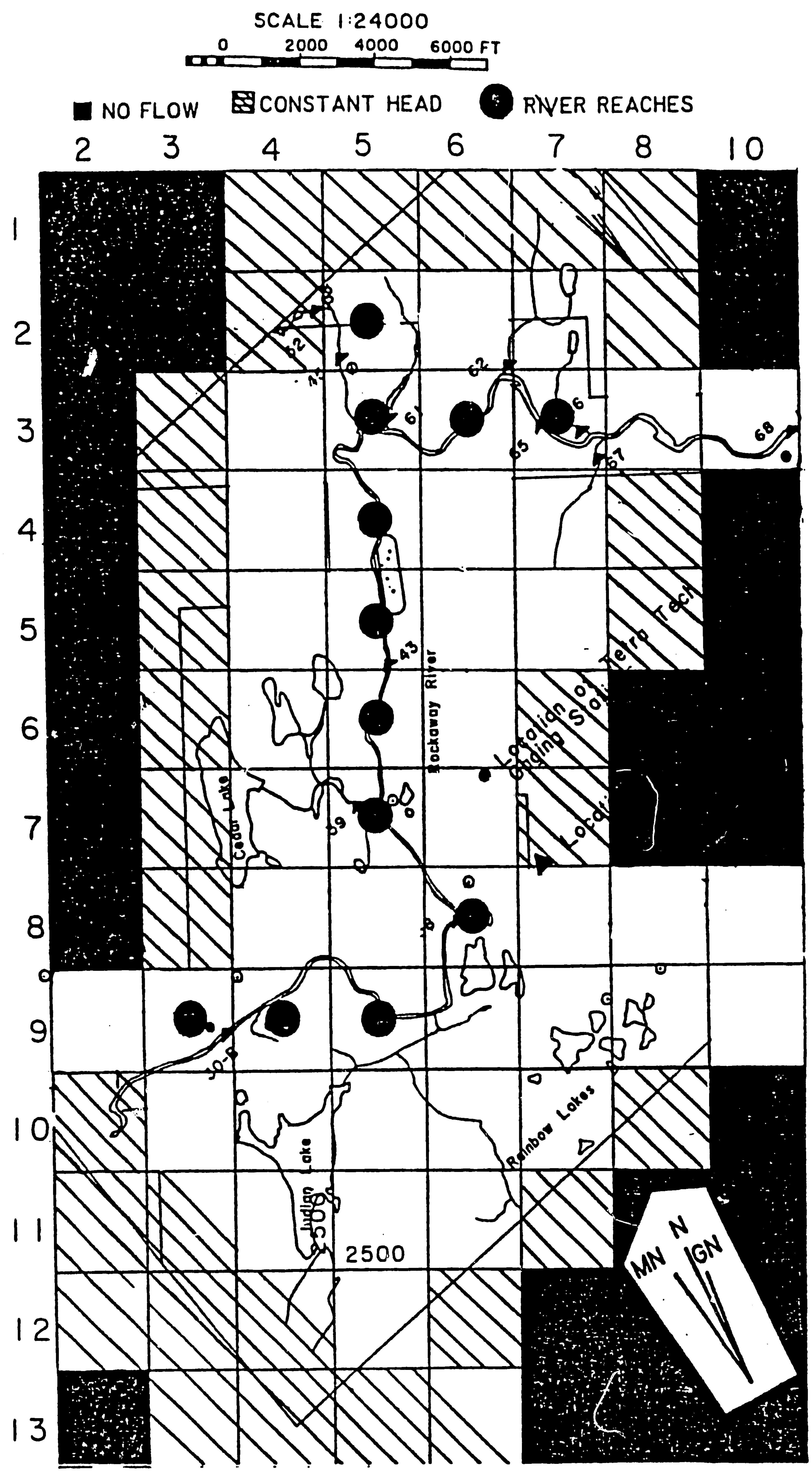


Figure 15 Preliminary grid design (Grid 1), each constant area cell represents .57 km^2 of the entire 29.8 km^2 study area.

where the valley opens up in the southwest corner of the study area along the limits of the terminal moraine. The bottom layer (layer 2) continues down the valley beneath layer 1 and terminates in the north-northeastern corner of the study area. Layer thickness was based on isopach maps constructed by the author with cross sections constructed by Stanford (1985).

All active nodes were modeled as cells whose head may vary with time, although the outermost cells along the edges of the study area constituted no-flow boundaries. The Rockaway River was modeled through the use of the River Package. Data needed for the simulation of the river included river head elevations, values for the conductance of the riverbed material, and the bottom elevation of the river bed. The data input was organized by selecting those nodes whose area encompassed a segment of the river and then averaging the river head and river bottom elevation values for each of the cells. Conductance was calculated separately for each of the eleven river reaches because this value is a function of the percentage of the river that lies within the cell. Field measurements of river heads were not available and thus these values were obtained from a topographic map. The effects of pumping wells on the stream-aquifer system were also modeled, although pumping schedules for the day the USGS conducted seepage runs had not yet been obtained. The pumping rates of wells modeled during this phase were therefore estimated from previous data. An initial estimate of a combined loss of approximately 5 cfs from both layers, at the Boonton well field, was used during simulations with grid number one. This estimate was derived by recognizing that the Boonton well field lies within the only losing reach of the river in the study area. It was originally assumed, therefore, that the loss of water from the river to the underlying aquifer system was due to pumpage at the well field. However, review of pumping schedules indicates that the loss of 0.14 m^3/s (5 ft^3/s) over the length of this reach cannot possibly be accounted for by withdrawal at the well field. No other pumping wells were simulated while using the discretization convention of grid one.

Hydraulic conductivity values of the unconfined aquifer used in model simulations were obtained by superposing the grid over the previously constructed hydraulic conductivity contour map and calculating the weighted average of the hydraulic conductivity values which fall within each cell. Transmissivity values of the basal aquifer and the saturated thickness of the water table aquifer used during model calibration were determined by the same technique.

9.1.1 Results of Simulations Using Grid One

Preliminary model runs indicated that the boundaries chosen for the model corresponded well with the natural drainage patterns. The orientation of the grid, with a sequence of cells falling in series up the valley proved to be the most optimal grid arrangement. This particular configuration also permitted another series of cells to fall along the trend of the main buried valley of the ancestral Rockaway River where maximum deposition occurred. One particularly noteworthy response of the model indicated the necessity to modify subsequent models. Problems with the convention of assigning no-flow boundaries to the entire system occurs when a modeled boundary of a study area does not correspond to drainage divide but actually intersects a suspected regional groundwater flow path where water is moving out of the local system. Extremely high heads were calculated by the model for those computational nodes in close proximity to the synthetic no-flow boundary. For example, in the south-southeast corner of the study area, head differentials are forcing groundwater to move at a relatively high velocity from the northwest edge of the area to the southeast edge. However, this flux of water out of the study area is difficult to simulate because of the barrier to flow imposed by the arbitrary assignment of no-flow to model boundaries. Two different solutions to this problem were possible. One solution would have been to increase the edge of the study area (the proposed extension applies only to this troublesome region) until it coincided with the natural drainage divide. This approach was ruled out because it would have made the study area much larger than would have been practical or manageable. A second approach involving the installation of "drains" into the model by means of the Drain Package was tested as a possible solution. Three drains, arranged in series along three cells, were used to simulate the flow of water out of the confines of the study area. Adjustments to drain elevations were necessary before the proper flux of water from the study area was simulated. With the drains installed, heads along this boundary no longer achieved unreasonably high values.

Comparison of observed heads to calculated heads indicated that the values used in the hydraulic conductivity matrix and in the vertical conductivity matrix were of the correct order of magnitude. River fluxes along the respective reaches were compared to river fluxes observed from field measurements. The initial values used to depict the riverbed conductance were decreased by an order of magnitude before calculated river fluxes matched observed river fluxes. After the initial assumptions of the assorted aquifer parameters were tested grid one became of minimal use. It became apparent, as more field data became available, that the grid should be redesigned so that it could accommodate more detail and would more realistically account for the complicated geology.

9.2 Grid Two

The design of grid two was based on the insight gained from model runs using grid one. In order to properly model the geometry and the variable thickness of the hydrogeological units a variable width grid was introduced. By reducing the areal coverage that each cell represented, fewer generalizations and more detail became encompassed in the model. Also, higher resolution of the aquifer-stream system could be attained. The design of grid two was a multi-stage procedure (Figure 16). Initially, grid two was set up in an array of eighteen rows, eighteen columns, and only one layer. This isolated the unconfined aquifer and enabled one to more effectively discern the impact of the river package on the model. Simulating the stream-aquifer system proved to be the most efficient means of determining local variations in hydraulic conductivity in the unconfined aquifer. Once this goal was accomplished within reasonable limits, the second and third layers were added to the system. The addition of a third layer to the model was the result of the subdivision of the basal aquifer modeled with grid 1. A more detailed discussion of the results obtained modeling with one layer versus three layers is presented in the following section. With the configuration shown in Figure 16, the impact of local variations in the stream-aquifer system was more pronounced and more clearly defined.

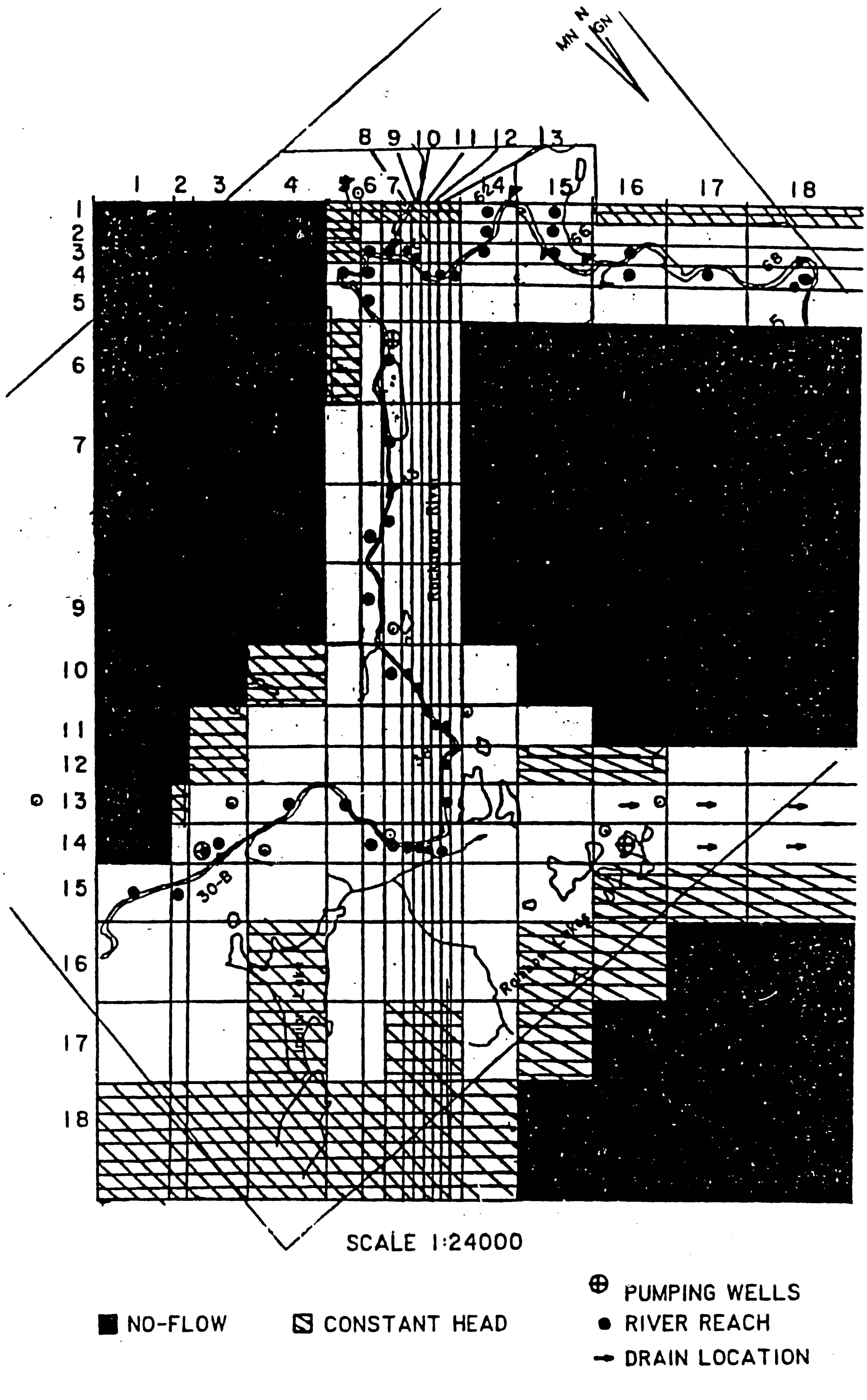


Figure 16 Grid configuration of variable width grid 2 showing location of pumping wells, constant head nodes and drains.

The top layer was modeled as unconfined and received constant areal recharge along all cells modeled as variable head. The average thickness of the uppermost layer was changed from 7.6 meters (25 feet) to about 12.1 meters (40 feet). The second and third layers were modeled as confined. Additional data, including seismic and pump test results prompted these adjustments to the original model. The number of cells containing river reaches was increased to forty-five. By the time the grid two model was being developed, pumping schedules had been obtained from the various water authorities and incorporated into the model. This additional information allowed an assessment to be made of the possible influences the pumping wells might be having on the multi-aquifer system. Five wells were used to simulate the withdrawal of groundwater from the various layers. The location of the pumping wells and the layer from which each well is withdrawing water is shown in Figure 16. As was the case with grid one, drains were used to offset the effects that no-flow cells have on the heads calculated by the model in cells adjacent to regions which are actually part of the regional flow system where water is transported away from the local system. Lakes were modeled as constant head nodes, although during the final phase of model calibration they were modeled as 'river' reaches. The River Package allows a more realistic simulation of the hydraulics existing between surface water bodies and the underlying aquifer. Additional constant head nodes were used at selected locations along the surface drainage divide and proved essential to maintaining saturated thickness along active nodes in close proximity to the bedrock ridges.

9.2.1 Results of Simulation Using Grid Two

The trial and error calibration of the hydrogeological model using grid two was particularly successful along reaches of the valley fill aquifers which underlie the present valley occupied by the Rockaway River. Comparison of the net river leakage obtained from field observation with the net river leakage calculated by the model indicates that a reasonable representation of the aquifer parameters existing in the valley fill aquifers was being simulated via the computer model. However, this applies only to certain segments of the river valley and only to the upper, unconfined aquifer.

The addition of two more layers to the hydrogeological model did not significantly alter the head distribution map of the uppermost unconfined aquifer. Maximum head differences between the one layer and the three layer model were about 1.5 meters (5 foot) but in general the head difference fell within a 0.61 meter (2 foot) range. This observation has various possible implications. First, it could be that the values used for the vertical conductance maps, which represent the relative ease water may pass between layers one and two and between layers two and three, were in error. If the values chosen for the Vcon arrays represented much greater confining or even impermeable conditions than those that actually exists in the aquifer system, then little or no hydraulic connection between the layers would be modeled. This situation (relatively impermeable confining units) is not predicted by the results of the pump test data. Therefore, it is unlikely that there would be little or no leakage between the two layers. Possibly the amount of vertical leakage from one layer to another is dependent upon the precise location examined in the geologically complex system. Because of the nature of depositional processes in glacial systems, it is highly probable that layers acting as confining units thin, thicken, or pinch out and generally do not form continuous layers throughout the system. Therefore, a variable vertical conductivity map was developed for the model. Modflow offers several methods that can be used to calculate the vertical conductance. The choice as to which approach appeared optimal was based on the geological conditions of the area. In this case, it was assumed that the amount and rate of water flowing vertically in the Rockaway Valley multi-aquifer system is dictated primarily by the thickness and vertical conductivity of the semi-confining to confining units. It is assumed that there is a significant difference in the hydraulic conductivities of the "aquifers" and "aquitards". Therefore, the calculation of vertical conductance is given by estimates of the vertical hydraulic conductivity of the confining unit divided by the thickness of the confining unit. Because the shape of the confining units or clay beds are highly variable this value fluctuates considerably throughout the study area. More field study that gathers data on the head distribution in each of the layers is needed to establish whether or not there is an intimate hydraulic connection between the various layers. Also, a thoroughly planned pump test, complete with strategically located observation wells, would indicate whether or not leaky confining conditions exist.

The most vexing problem encountered with both the grids and the models was the tendency for many of the nodes along the steeper valley walls to dry up during the first iteration of a simulation. These cells represent the unconfined aquifer as it occurs at the boundary between bedrock ridges forming the topographic highs and the valley floors underlain by unconsolidated materials. An excessive drop in hydraulic head from adjacent nodes, which lie in a perpendicular cross section to the long axis of the valley, was identified as the cause of the problem. In some instances, the difference in driving head from one node to the next was as much as 30.5 meters (100 feet). It appears that for relatively shallow aquifers, the model cannot support such a drastic change in head between adjacent cells. It is possible that in the Rockaway Valley aquifer system these problematic areas in the water table aquifer are not saturated enough to provide a significant contribution to the groundwater system. If this is the case perhaps the nodes representing these areas should be inactivated. However, a straightforward answer to this problem is thwarted by a lack of data in critical areas. To date, measurements of the water levels in the aquifer along the transition zone from the saturated valley fill materials to the bare bedrock ridges are sparse. Because of this data deficiency, the initial heads in the water table aquifer where field data does not exist were established according to surface elevations. This assignment of heads is based on the principle that water table elevations commonly mimic the topography. The drying up of cells along these boundaries implies that either the assumption above is in error or the unconsolidated sediments near topographic highs are not fully saturated. In order to identify the cause for many of these cells drying out it was necessary to redesign the grid so that large changes in head between cells adjacent cells would be avoided. The use of a finer grid eliminates the possibility that the computer model might not be able to handle such an excessive plummeting of heads between adjacent cells.

9.3 Final Grid Selection

9.3.1 Grid Dimensions and Packages Implemented

The final grid consisted of 49 rows, 28 columns, and two layers. The grid orientation and dimensions of the rows and columns are shown in Figure 17. The finer grid mesh allows smoother transitions from highs along the bedrock ridges of 213 meters (700 feet) to valley lows of 149 meters (490 feet). It can also better accomodate the variability of the hydrogeological system and allows the system to be more accurately simulated by the model. One hundred and twelve river reaches were used to model the interaction between the river and the aquifer system and between the numerous lakes and the underlying aquifer. Sixteen drain cells were incorporated into the model and served as a conduit for flow out of the system at the southeast corner of the study area. Four pumping wells are used to simulate the effects of public water demands on the aquifer system. One well is withdrawing groundwater from the water table aquifer, the rest are tapping the lower aquifer for water. Location of the 66 cells with river reaches, the 46 cells with lakes, the 16 cells with drains, and the four nodes with pumping wells are shown in Figure 17. Figure 17 also shows the assignment of constant head, computational, and no-flow nodes for the top modeled layer. The location and amount of constant head cells employed were determined from the models response to many different combinations.

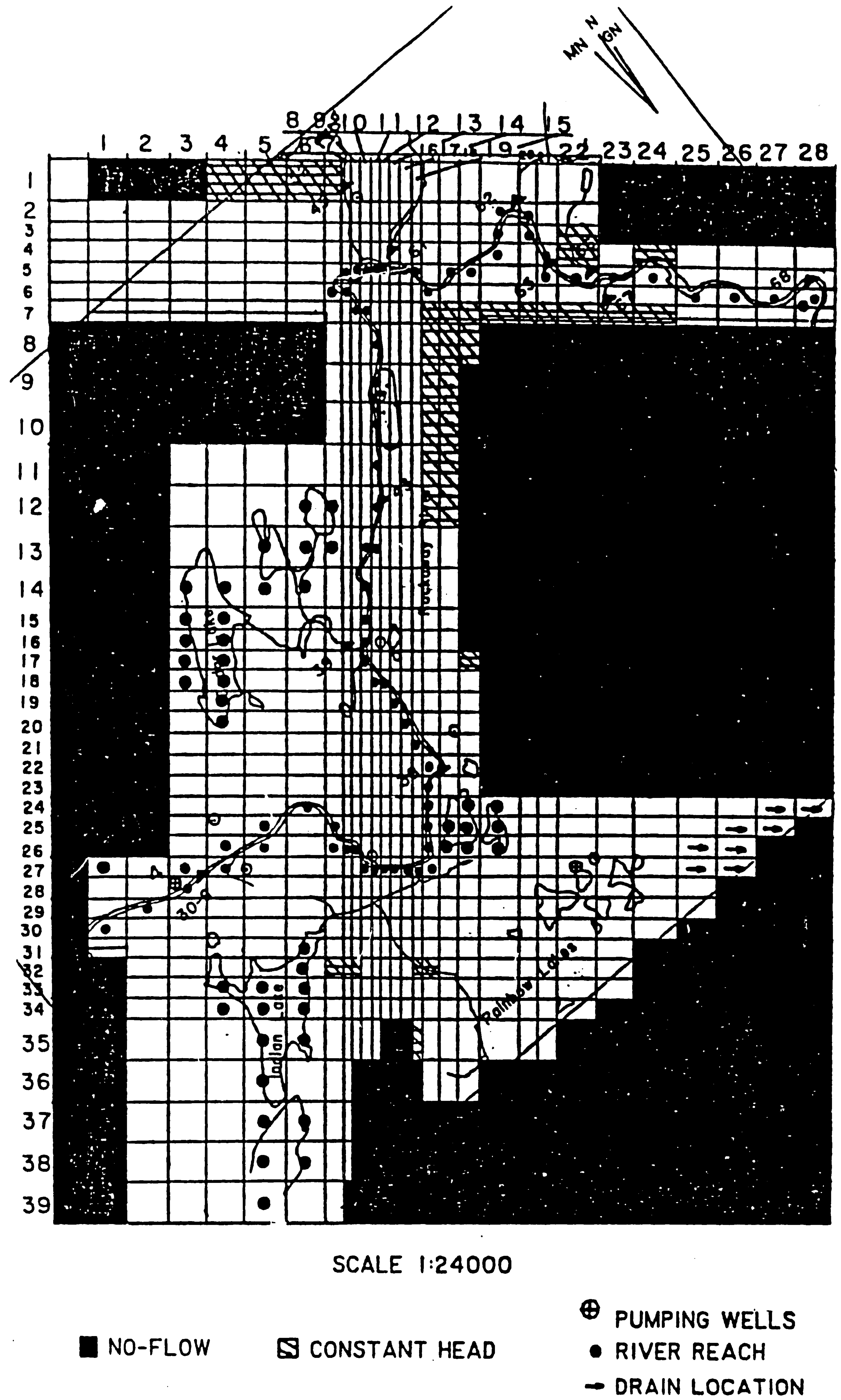


Figure 17 Final grid design consisting of 28 columns, 40 rows, and 2 aquifers modeled. Location of constant head nodes, pumping wells, and drains are shown.

The final selection of constant head nodes was based on the applicability of the calculated water budget data, the volume of streamflow calculated for each river cell, and the amount of water introduced into the system through these constant head cells to actual field observations and conditions.

9.3.2 Vertical Discretization

The top layer is modeled as an unconfined aquifer with a river penetrating into its surface. Such a configuration implies that stresses imposed on the aquifer will influence the behavior of the river, and vice versa.

The underlying hydrogeological units are grouped together and modeled as one confined layer. Considering the data base, it was decided that this approach was the most efficient and practical decision at this time. The lack of data with which the possible occurrence of vertical gradients could be located, the gaps in head data for the underlying units, and the limited number of known aquifer parameters precludes a more detailed simulation of the underlying system. Therefore, all data believed to represent hydrogeological conditions of the water producing units below the water table aquifer were compiled into one underlying layer. Until more data become avaliable, there is little justification for a more refined model.

10.0 FINAL SELECTION OF AQUIFER PARAMETERS

A brief synopsis of how the aquifer parameters were chosen and the values of these parameters that produced the most accurate representation of the real system is presented in the following subsections.

10.1 Saturated Thickness of the Water Table Aquifer

Thickness of the unconfined aquifer ranged from 1.5 meters (5 feet) to 14 meters (45 feet). However, most of the valley-fill unconfined aquifer system is believed to be about 35 feet thick. The isopach map of the water table aquifer provided the basis for selection of aquifer thickness. Little modification to this parameter was necessary or justified during model runs except where data gaps existed. Problem areas were limited mainly to computational nodes proximal to topographic highs. A three-dimensional representation of the bottom elevation of the water table aquifer is given in Figure 18.

10.2 Hydraulic Conductivity of the Water Table Aquifer

Because very few field measurements of this parameter exist, a trial and error approach was used to determine representative hydraulic conductivity values for the unconfined aquifer. Also, difficulty in assigning representative hydraulic conductivities was encountered because of the complex network of variable sediments present

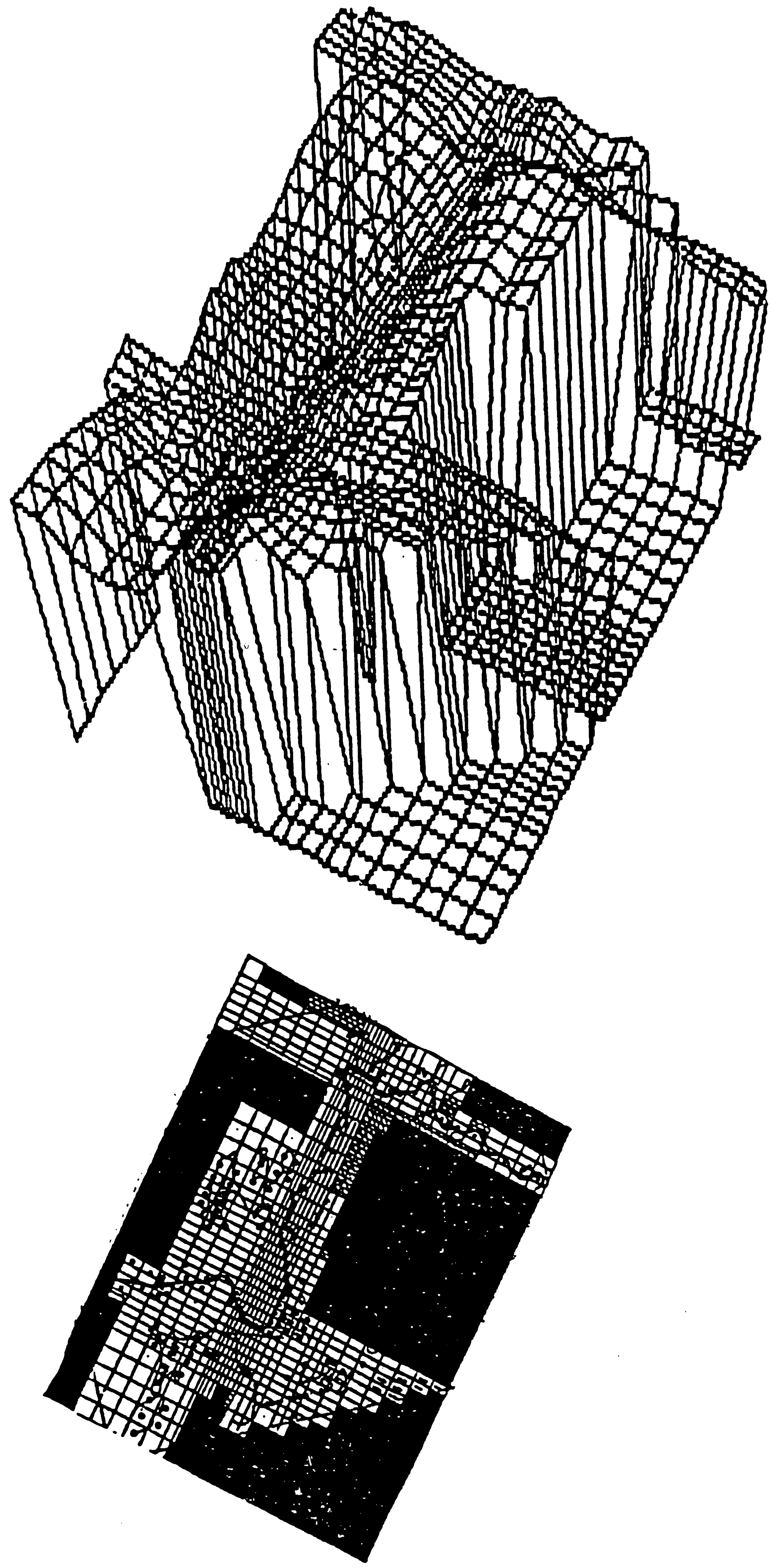


Figure 18 A three-dimensional representation of the bottom elevation of the water table aquifer. The grid inset provides a frame of reference.

in this unit. An illustration of the three dimensional distribution of hydraulic conductivity values finally chosen for this model is shown in Figure 19.

The final selection of hydraulic conductivity values for layer 1 was made on the basis of two separate lines of evidence. For those cells located in the immediate vicinity of the river and for those cells containing reaches of the river the determination of hydraulic conductivity values was based on the ability of the model to simulate the river leakage that was observed in the field. Trial and error runs indicated that the net flux of water to and from the aquifer and stream is predominantly a function of hydraulic conductivity. This observation is to be expected considering the major role conductance plays in the groundwater flow equation.

The other method used to select possible hydraulic conductivity values involves analysis of water budgets. Comparison of water budgets calculated from model runs using tentative hydraulic conductivities to those estimated from field observations proved to be a useful criterion on which to eliminate values that produced unreasonable results.

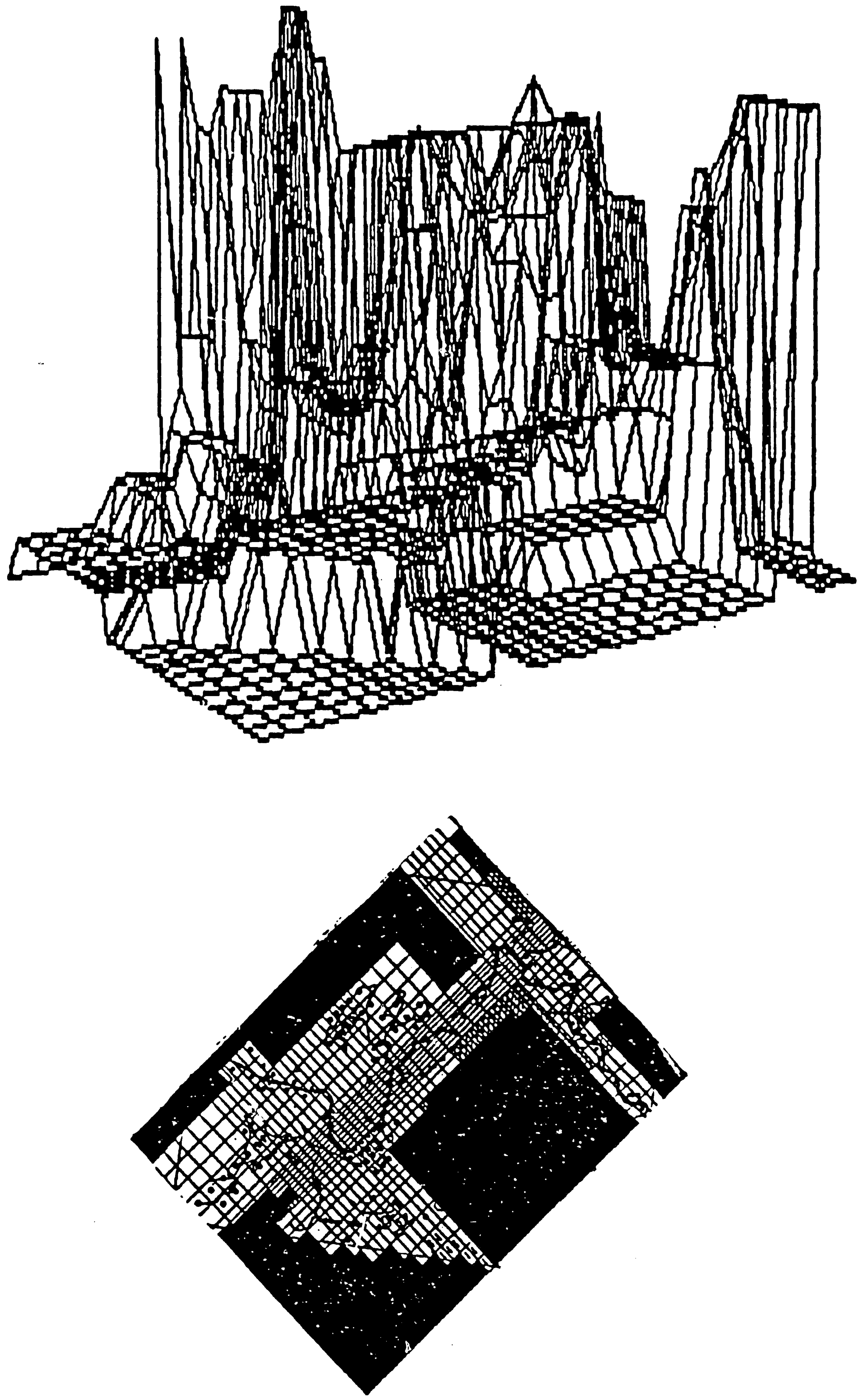


Figure 19 A three-dimensional representation of the hydraulic conductivity of the water table aquifer. The grid inset provides a frame of reference.

The assignment of incorrect hydraulic conductivity values was particularly noticeable at certain constant head nodes. The exorbitant flux of water into the system, generated at that node, clearly indicated that the hydraulic conductivity assigned to that node, and possibly to adjacent nodes, was much too high. Adjustments were incorporated into the model based on these observations.

10.3 Vertical Conductance Map

Calculation of the vertical conductance (Vcon) component used to describe the flux of water to and from vertically adjacent cells was partially based on horizontal conductivity values pertaining to the less permeable 'confining units' that were determined in previous hydrogeological investigations conducted in similar geological environments in the same general area. In addition, calculations of Vcon were based on the thickness of the confining unit as it is depicted in the cross sections and isopach maps, and from published ratios of horizontal to vertical conductance calculated for similar materials. Combination of these data enabled an a Vcon map to be constructed based on an educated calculation of the Vcon values. Adjustments were made to the original values as simulation runs indicated modifications were necessary. A three-dimensional map of the final Vcon array used during model calibration is shown in Figure 20.

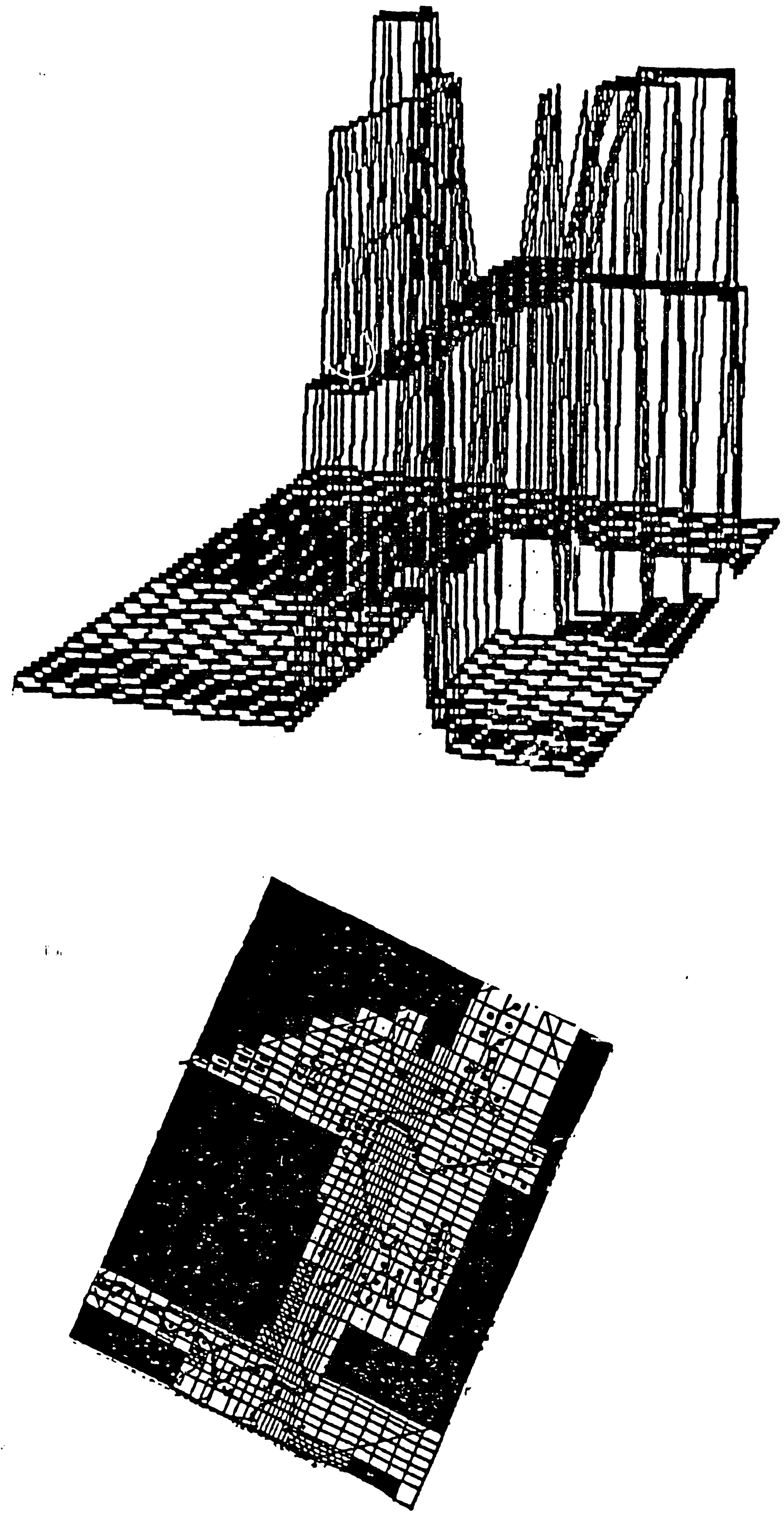


Figure 20 A three-dimensional representation of the vertical conductivity between layer 1 and layer 2. The grid inset provides a frame of reference.

10.4 Transmissivity of the Confined Aquifer

Construction of a transmissivity map for the bottom layer was based on extrapolation of scanty field data and application of numerous assumptions, namely those implicit in the conclusion that the transmissivity values calculated by the modified Theis equations (equation 1 from Section 5.1) are applicable to the study area. There are many combinations of thickness and hydraulic conductivity values that could be used to determine the distribution of this parameter. For the most part, the isopach map of the total unconsolidated sediments (Figure 9) and the contoured map of transmissivities (Figure 10) calculated using the Theis solution were used as guidelines. Because there were many unknowns associated with this parameter it was important to first achieve reasonable simulations by varying the order of magnitude. Matching of observed water level elevations in the lower layer with water level elevations calculated by the model allowed the construction of a more clearly defined transmissivity map. Once fairly good results were obtained then a fine tuning of the map became justified.

A three-dimensional illustration of the final transmissivity map selected for model calibration is shown in Figure 21.

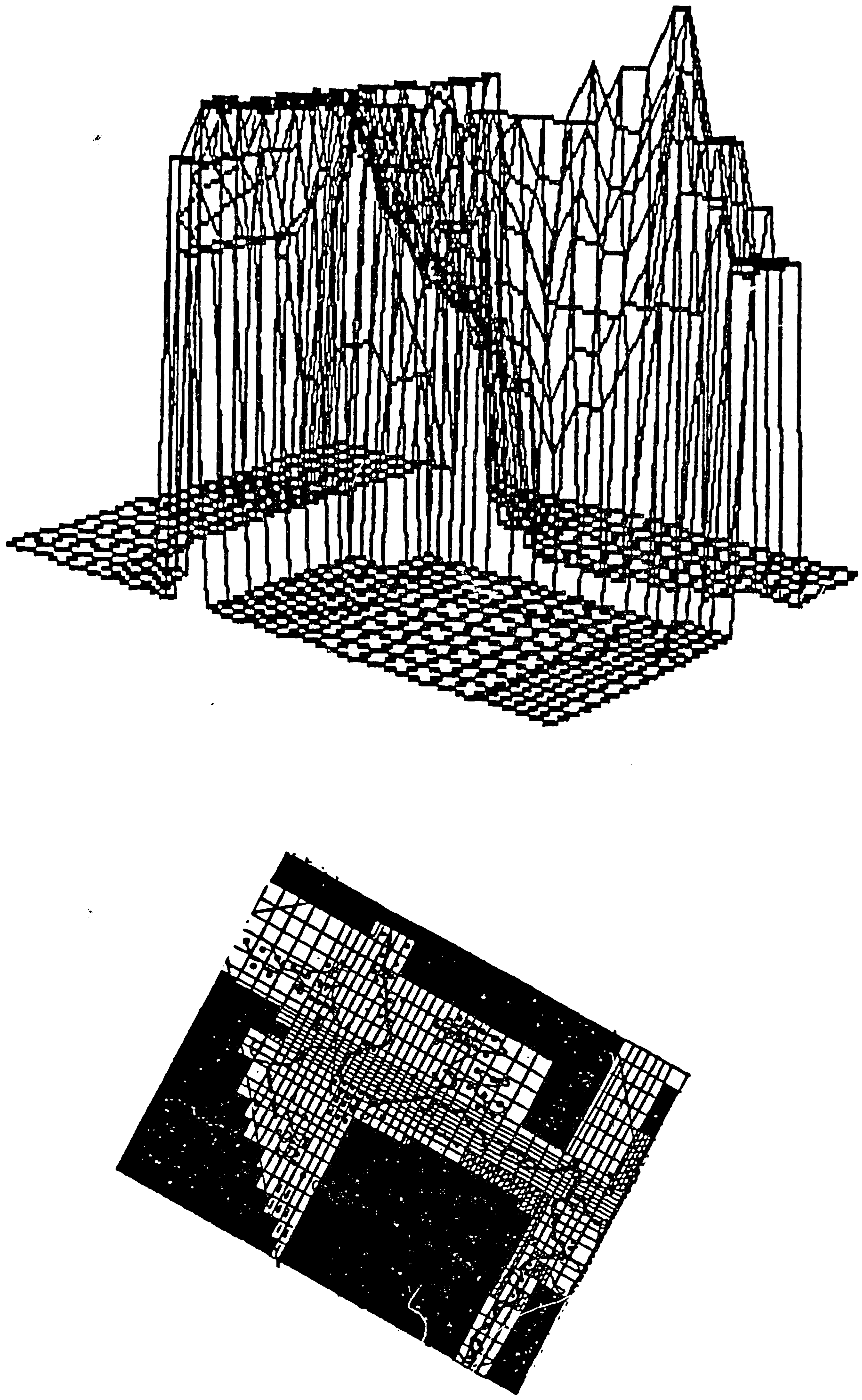


Figure 21 A three-dimensional representation of the transmissivity of the lower aquifer. The grid inset provides a frame of reference.

11.0 ANALYSIS OF MODEL CALIBRATION

Trial and error adjustments of aquifer parameters was the only practical way to approach the calibration of the model. Furthermore, because little field data exists for this particular study and because the steady-state configuration of starting heads is uncertain, many unknowns exist and a unique solution seems unlikely. However, calibration of the model based on adjusting aquifer parameters within reasonable limits and then comparing the steady-state solution of the calculated stream flux between reaches to the flux actually observed proved to be successful in constraining the numerous combinations of aquifer parameters. Therefore, initial efforts to refine the model were focused on matching observed river fluxes obtained during base flow conditions with the river fluxes calculated by the model. Future efforts can then be focused on matching observed water level elevations with those predicted by the model. Adjustments of vertical conductivity and possibly transmissivity of the bottom layer to obtain good matches between calculated and observed heads should be the next step in subsequent calibration.

In this study calibration of the model was terminated after reasonable values of river fluxes were obtained by the model. During this phase calculated head elevations were comparable to observed elevations in only a few cells. Continued calibration may have resulted in a much better match of head elevations but it is the author's opinion that such attempts are futile until more data become available and until the existing data base is shown to be reliable.

11.1 Calibration Based on Comparison of River Leakage

Evaluation of model calibration using streamflow leakage could only be done for river reaches 3, 4, and 5 (See Figure 22). The discrepancy between the net leakage observed for the 1985 and the 1984 streamflow data for reach 2 (both data sets were obtained during baseflow conditions) is too great and precludes attempts at calibration of this region of the study area based on net river leakage. Therefore, primary efforts were focused on calibration of the model in the region from the tributary valley fill aquifer to the point where the river intersects the Ramapo Fault. Table 10 lists the data input used to simulate the river reaches in the model.

Table 11 lists the river fluxes calculated by the model for each node that contains a river reach. These fluxes were determined using the aquifer parameters as outlined in the last few sections. Itemization of the leakage for each node was done to show that reasonable fluxes of water were calculated for each cell and to prove that the net calculated seepages for a reach were not based on the summation of unrealistic values.

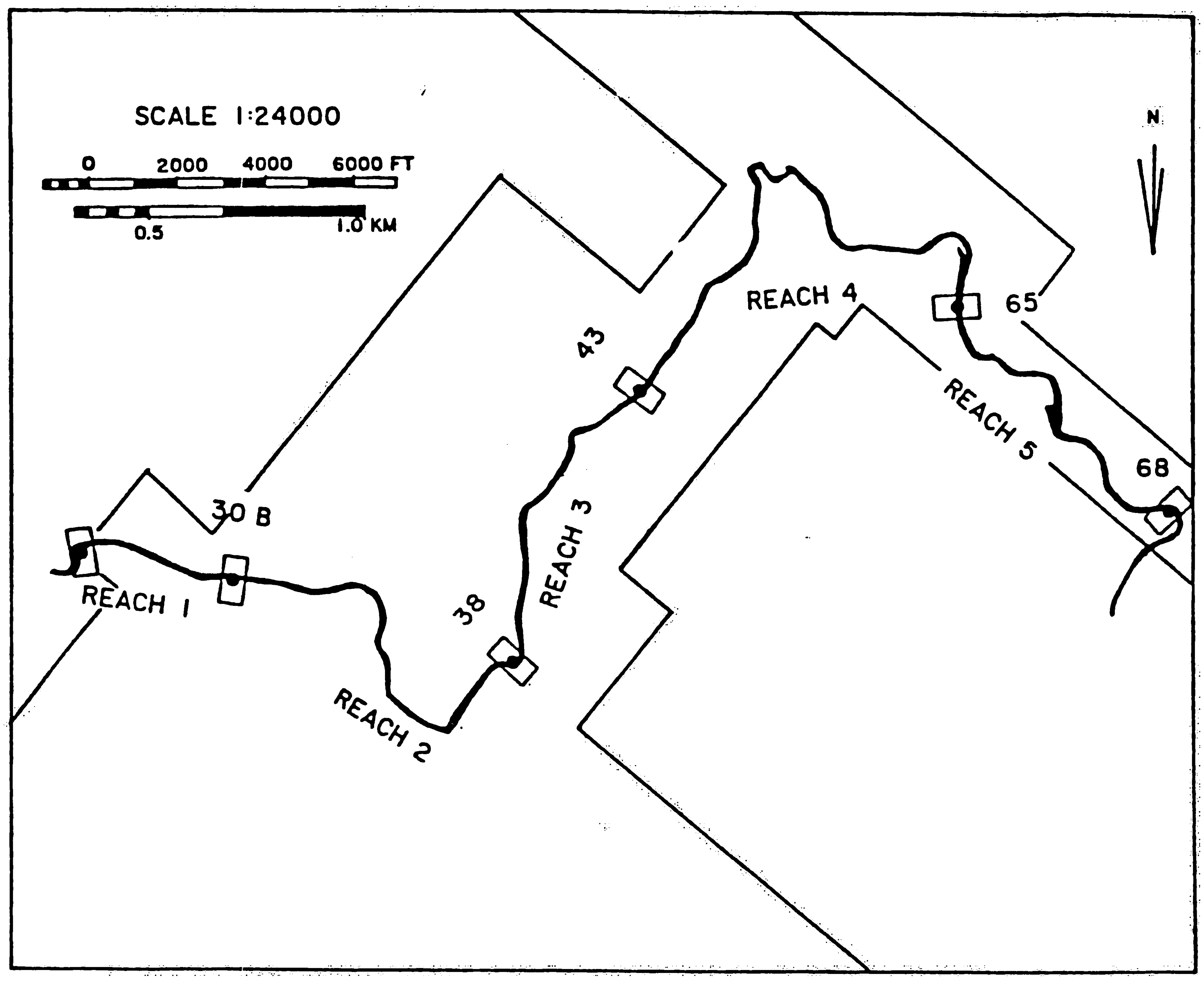


Figure 22 Location of river reaches in the study area and corresponding stream gaging stations for September 1985 data.

Table 10 Data set used to simulate effects of the river

Layer,Row,Column	Stage	Conductance	Bottom Elevation	River Reach
1,26,19	508.	1.30	485.	1
1,25,19	508.	2.40	485.	2
1,30,1	517.	0.40	512.	3
1,29,2	516.	0.40	511.	4
1,28,3	515.	0.40	510.	5
1,27,4	514.	0.40	509.	6
1,26,5	513.	0.40	508.	7
1,25,5	512.	0.40	507.	8
1,24,6	511.	0.40	506.	9
1,25,27	510.	0.30	505.	10
1,26,7	509.	0.40	504.	11
1,26,8	508.	0.40	503.	12
1,26,9	507.	0.30	502.	13
1,27,10	506.	0.30	501.	14
1,27,11	506.	0.30	501.	15
1,27,12	506.	0.20	500.	16
1,27,13	505.	0.20	500.	17
1,27,14	505.	0.20	500.	18
1,27,15	505.	0.10	500.	19
1,27,16	505.	0.10	499.	20
1,26,16	504.	0.10	499.	21
1,25,16	504.	0.10	499.	22
1,24,16	504.	0.10	498.	23
1,23,16	503.	0.10	497.	24
1,22,16	503.	0.10	496.	25
1,21,15	502.	0.10	495.	26
1,20,14	501.	0.10	495.	27
1,19,13	500.	0.10	494.	28
1,18,12	498.	0.10	493.	29
1,18,11	497.	0.10	492.	30
1,17,10	496.	0.10	491.	31
1,16,10	495.	0.10	490.	32
1,15,10	494.	0.10	489.	33
1,14,10	493.	0.10	488.	34
1,13,10	492.	0.10	487.	35
1,13,11	491.	0.10	486.	36
1,12,11	490	0.10	485.	37
1,11,11	489.	0.20	484.	38
1,10,11	488.	0.20	483.	39
1, 9,11	487.	0.10	482.	40
1, 8,11	486.	0.10	481.	41
1, 7,10	485.	0.10	480.	42
1, 7, 9	484.	0.10	479.	43
1, 6, 8	483.	0.10	478.	44

Table 10 (continued)

Layer,Row,Column	Stage	Conductance	Bottom	River
1, 5, 9	482.	0.10	477.	45
1, 5,10	481.	0.10	476.	46
1, 5,11	480.	0.10	475.	47
1, 5,12	479.	0.10	474.	48
1, 5,13	478.	0.10	473.	49
1, 5,14	477.	0.20	472.	50
1, 5,15	476.	0.20	471.	51
1, 6,16	475.	0.20	470.	52
1, 5,17	474.	0.20	469.	53
1, 5,18	473.	0.20	468.	54
1, 4,19	472.	0.40	467.	55
1, 3,19	469.	0.40	464.	56
1, 2,19	468.	0.40	463.	57
1, 2,20	464.	0.70	459.	58
1, 3,20	460.	0.80	455.	59
1, 4,21	459.	0.80	454.	60
1, 5,21	457.	0.90	452.	61
1, 5,22	455.	0.80	450.	62
1, 5,23	452.	0.80	447.	63
1, 5,24	450.	0.80	445.	64
1, 6,25	449.	0.90	443.	65
1, 6,26	444.	0.80	437.	66
1, 6,27	439.	0.90	433.	67
1, 6,28	434.	0.80	427.	68
1,38, 5	515.	7.50	495.	69
1,38, 6	515.	5.00	495.	70
1,39, 5	515.	7.50	495.	71
1,37, 5	515.	2.50	495.	72
1,37, 6	515.	5.00	495.	73
1,36, 5	513.	3.30	493.	74
1,35, 5	513.	6.60	493.	75
1,34, 5	513.	0.60	493.	76
1,33, 5	513.	0.60	493.	77
1,31, 6	513.	2.50	493.	78
1,32, 6	513.	3.70	493.	79
1,33, 6	513.	3.70	493.	80
1,34, 6	513.	3.70	493.	81
1,35, 6	513.	3.30	493.	82
1,34, 4	513.	0.30	493.	83
1,33, 4	513.	2.50	493.	84
1,32, 4	513.	0.63	493.	85
1,20, 3	533.	0.32	518.	86
1,18, 4	533.	3.80	518.	87
1,17, 4	533.	3.80	518.	88
1,16, 4	533.	4.00	518.	89
1,15, 4	533.	3.30	518.	90
1,14, 4	533.	3.30	518.	91

Table 10 (continued)

Layer,Row,Column	Stage	Conductance	Bottom Elevation	River Reach
1,14, 3	533.	2.50	518.	92
1,15, 3	533.	2.50	518.	93
1,16, 3	533.	2.50	518.	94
1,17, 3	533.	2.50	518.	95
1,18, 3	533.	0.30	518.	96
1,20, 4	533.	2.50	518.	97
1,19, 4	533.	5.00	518.	98
1,14, 5	530.	2.50	515.	99
1,13, 5	530.	4.50	515.	100
1,14, 6	525.	2.50	510.	101
1,12, 6	525.	4.00	505.	102
1,13, 7	525.	2.00	505.	103
1,13, 6	525.	2.00	505.	104
1,12, 7	525.	2.50	505.	105
1,26,17	509.	1.30	490.	106
1,26,18	509.	2.30	490.	107
1,25,17	509.	1.30	490.	108
1,25,18	509.	2.20	490.	109
1,24,18	509.	1.30	490.	110
1,24,19	508.	0.90	485.	111
1,27, 1	528.	0.70	500.	112

Table 11 Calculated River Leakage for each Node Containing a Segment of the Rockaway River and the Net Flux Between the USGS Stream Gaging Stations as Calculated by the Model. Leakage is in cubic feet per second.

Row	Column	Calculated River Leakage	Reach ID	Net Flux
27	4	$+0.3910 \times 10^{-1}$		
26	5	-0.3972×10^{-1}		
25	5	$+0.6015 \times 10^{-1}$		
24	6	-0.4351×10^{-1}		
25	7	-0.5799×10^{-1}		
26	7	-0.1252×10^{-1}		
26	8	-0.1998×10^{-1}		
26	9	-0.1379×10^{-1}		
27	10	$+0.1301 \times 10^{-1}$		
27	11	-0.3744×10^{-1}		
27	12	-0.7604×10^{-1}		
27	13	-0.1779×10^{-1}		
27	14	-0.6804×10^{-1}		
27	15	-0.6368×10^{-1}	1	-0.7937
27	16	-0.9442×10^{-1}		
26	16	-0.3212×10^{-1}		
25	16	-0.5067×10^{-1}		
24	16	-0.9294×10^{-1}		
23	16	-0.8311×10^{-1}		
22	16	-0.1021		
21	15	-0.5417×10		
20	14	-0.1592×10^{-1}		
19	13	$+0.1058 \times 10^{-1}$		
18	12	$+0.4502 \times 10^{-1}$		
18	11	+0.1084		
17	10	+0.1296	2	+1.7472
16	10	+0.1384		
15	10	+0.1577		
14	10	+0.2572		
13	10	+0.2393		
13	11	+0.2534		
12	11	+0.4778		

Table 11 (Continued)

Row	Column	Calculated River Leakage	Reach ID	Net Flux
11	11	+0.4416		
10	11	-0.4035×10^{-1}		
9	11	+0.2553		
8	11	+0.1352		
7	10	-0.7876×10^{-2}		
7	9	$+0.5087 \times 10^{-1}$		
6	8	$+0.7868 \times 10^{-1}$		
5	9	$+0.7926 \times 10^{-1}$	3	+1.9536
5	10	$+0.4619 \times 10^{-1}$		
5	11	$+0.2625 \times 10^{-1}$		
5	12	$+0.2178 \times 10^{-1}$		
5	13	$+0.2601 \times 10^{-1}$		
5	14	$+0.3989 \times 10^{-1}$		
5	15	$+0.5454 \times 10^{-1}$		
6	16	+0.1807		
5	17	$+0.7587 \times 10^{-1}$		
5	18	+0.1130		
4	19	-0.5588×10^{-1}		
3	19	$+0.4589 \times 10^{-2}$		
2	19	$+0.7464 \times 10^{-1}$		
2	20	$+0.6329 \times 10^{-1}$		
3	20	+0.1755		
4	21	+0.1134		
5	21	+0.2089		
5	22	+0.9371		
5	23	+0.5837		
5	24	+0.7186	4	+2.8358
6	25	+0.1976		
6	26	$+0.5396 \times 10^{-1}$		
6	27	$+0.3027 \times 10^{-1}$		
6	28	+0.1056		

Excellent matches of observed and calculated fluxes for reaches 3 and 5 were achieved. A gain of 1.75 ft^3/s was calculated for reach 3 by the model which is precisely the net seepage observed between stations during the September 1985 seepage runs (see Table 8). A gain of 2.84 ft^3/s was calculated for reach 5 which is identical to field observations of September 1985. A gain of 1.95 ft^3/s was calculated for the middle reach, an absolute difference of 60 % from the observed leakage. Data collected from the three separate seepage runs (see Table 8) indicate that something odd may be occurring along reach 4. This is evident mainly by analysis of the stream discharge data (from September and April 1985) which suggest that this reach is losing water to the underlying aquifer. However, when the data is viewed in terms of the possible ranges in error associated with stream gaging techniques then only the September data falls soley within values indicative of a losing reach. The range of values for possible net seepage for this reach when an error margin of plus or minus 5% for each data collecting site is considered is -0.028 m^3/s (-1 ft^3/s) to -2.55 m^3/s (-9.0 ft^3/s) for the September 1985 data, +0.736 m^3/s (26 ft^3/s) to -0.226 m^3/s (-8.0 ft^3/s) for the April 1985 data, and -0.255 m^3/s (-9.0 ft^3/s) to +0.057 m^3/s (+2.0 ft^3/s). Although the field data corresponding to the respective seepage runs were collected during different climatic conditions the net flux observed in reach 4 is similar among all three data collection episodes. Pumping schedules were not obtained for the 1984 date and no pumpage took place during the April 1985 seepage runs (Harte, personal communications) but it is a safe assumption that pumpage at the Boonton well did not exceed September 1985 water withdrawals and, more importantly, there is no way the observed gain can be accounted for by either pumping or non-pumping during the seepage runs. Morover, the 1985 high flow seepage data shows a net loss of 5.8 ft^3/s . This is hardly the expected river leakage for high flow conditions. Possible explanations for these erratic observations are given in the next section.

Calibration of the model using only streamflow data is not feasible because it excludes careful scrutiny of other sections of the study area and neglects the interaction of the bottom layer. Comparison of water levels calculated to water levels observed for both layers supports the above statement (See Table 12) . Overall, head difference between the data obtained from the model and data measured in the field are reasonably similar, particularly in those monitored wells that are in close proximity to reaches 3 through 5 where primary calibration efforts were focused.

Table 12 Comparison of calculated versus observed water levels

Grid Location	Layer	Observed Water Level	Calculated Water Level	Difference in percent
10,12	2	148.8 m (491 ft)	150.6 m (494 ft)	1.2
10,12	1	149.7 m (491 ft)	150.0 m (492 ft)	0.2
10,11	1	150.0 m (492 ft)	149.1 m (490 ft)	0.6
9,12	2	149.4 m (490 ft)	150.9 m (495 ft)	1.0
10,11	2	148.5 m (487 ft)	150.0 m (492 ft)	1.0
9,12	1	149.4 m (490 ft)	150.0 m (492 ft)	0.4
1, 8	2	150.9 m (495 ft)	151.5 m (497 ft)	0.4
16,11	2	149.1 m (489 ft)	152.5 m (500 ft)	2.2
27,11	2	150.3 m (493 ft)	153.0 m (502 ft)	1.8
25,25	2	101.5 m (333 ft)	115.9 m (353 ft)	14.1
27,22	2	118.9 m (390 ft)	128.4 m (421 ft)	7.9

Comparison of Observed and Calculated Streamflow

Reach ID	Observed Seepage	Calculated Seepage	Difference in percent
1	+0.342 m^3/s +12.1 ft^3/s	+0.022 m^3/s (+0.79 ft^3/s)	93
2	+0.049 m^3/s +1.75 ft^3/s	+0.049 m^3/s (+1.75 ft^3/s)	0
3	-0.14 m^3/s -4.95 ft^3/s	+0.055 m^3/s (+1.95 ft^3/s)	61
4	+0.08 m^3/s +2.84 ft^3/s	+0.08 m^3/s (+2.84 ft^3/s)	0

12.0 Discussion

Comparison of the contoured water table and the potentiometric maps calculated by the model (Figures 23 and 24 respectively) to the water table and potentiometric maps constructed by the author (Figures 13 and 14, respectively) and the three-dimensional representation of the two aquifers (Figures 25, 26, and 27) indicates that similar patterns in the head distribution have been achieved with the chosen aquifer parameters and with the selected configuration of hydrostratigraphic units. Although the absolute values of head elevations are not identical, the similarity in their distribution is believed to reflect the validity of many initial assumptions which have accompanied the development of this model.

Modflow is capable of calculating volumetric water budgets for each model simulation. The resultant water budgets from successful runs indicate that net inflows and outflows fall within reasonable values (Table 13). However, it was necessary that a considerable volume of water be input to the system by constant head nodes located at the topographic drainage divide. The relatively large flux of water originating from these necessary constant head cells could imply the existence of one or more of several different conditions. First of all, if the constant head cells were switched to computational nodes many cells

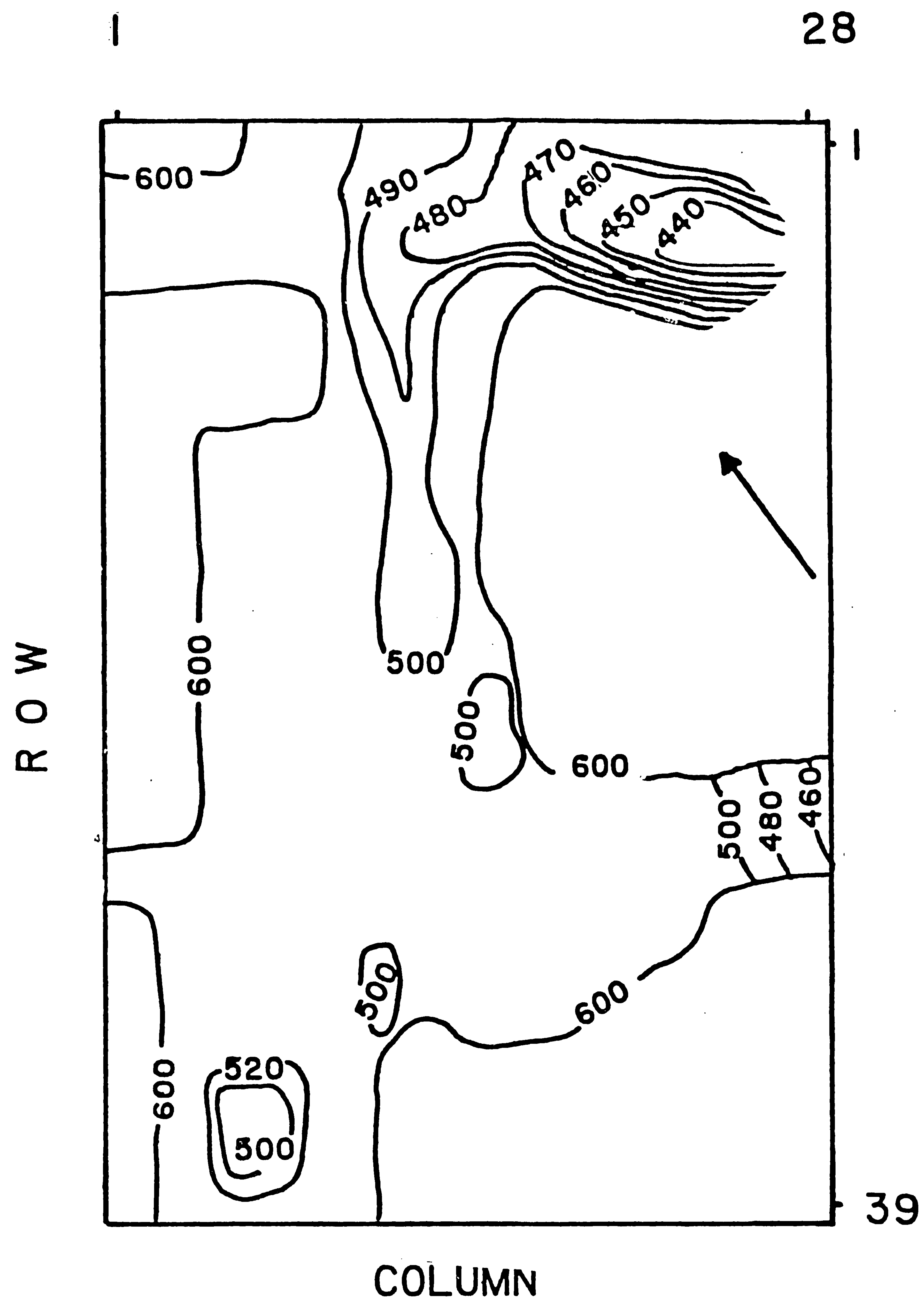


Figure 23 Contour map of water levels calculated by the model for the water table aquifer. Head elevations in feet above mean sea level.

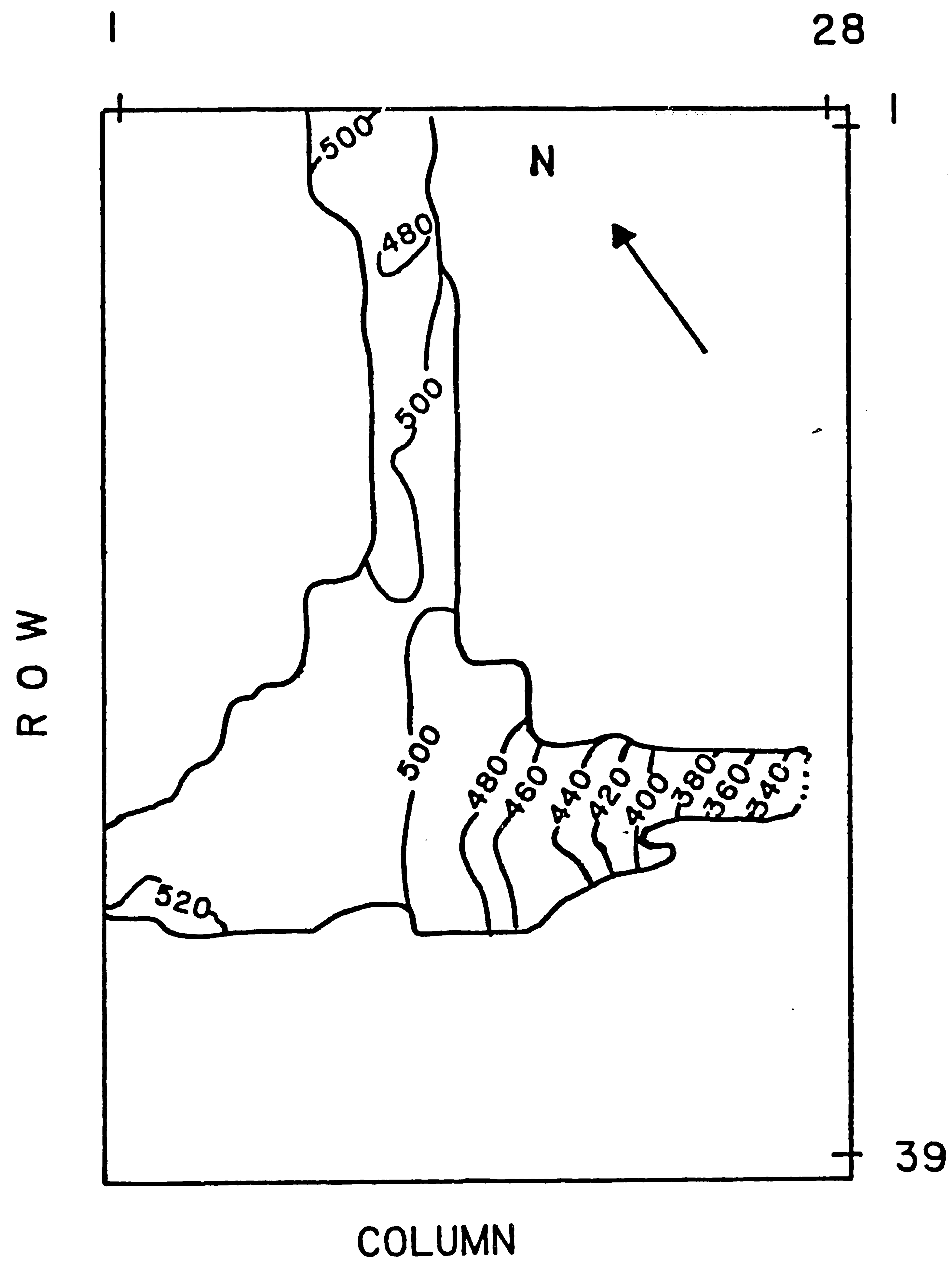
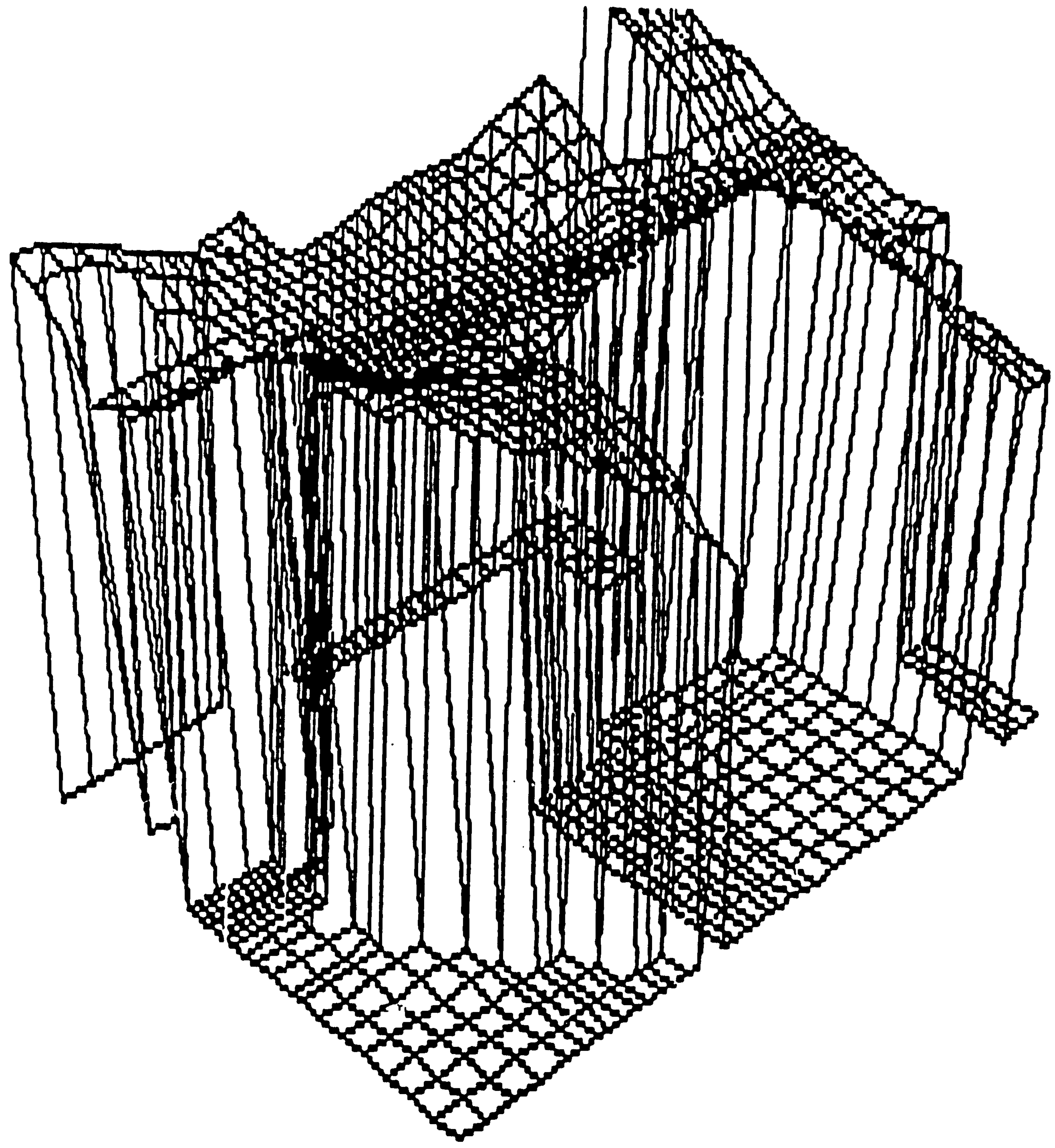


Figure 24 Contour map of the potentiometric surface of the lower aquifer. Head elevations in feet above mean sea level.



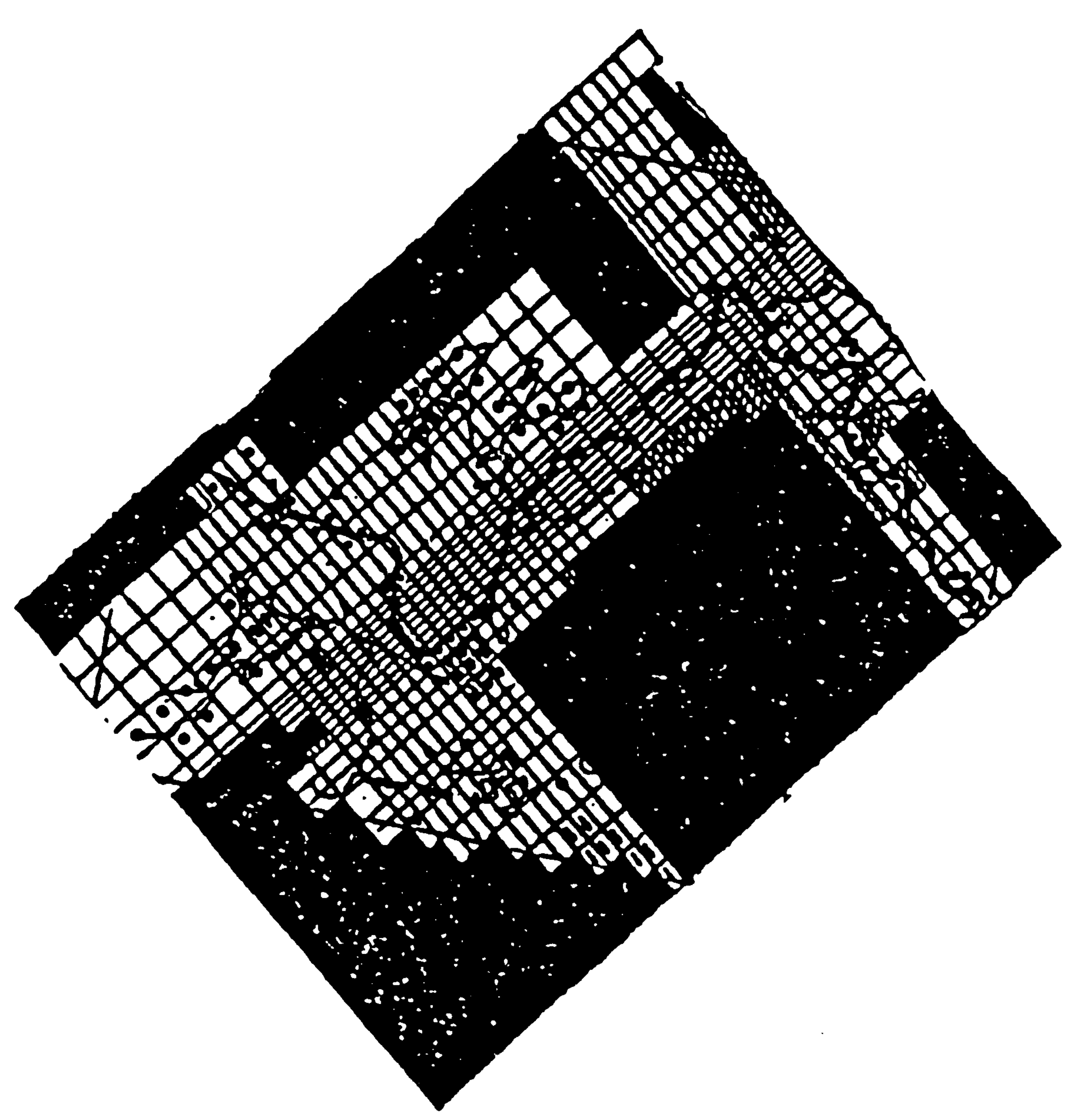


Figure 25 A three-dimensional representation of the head elevations of the water table aquifer calculated by the model.

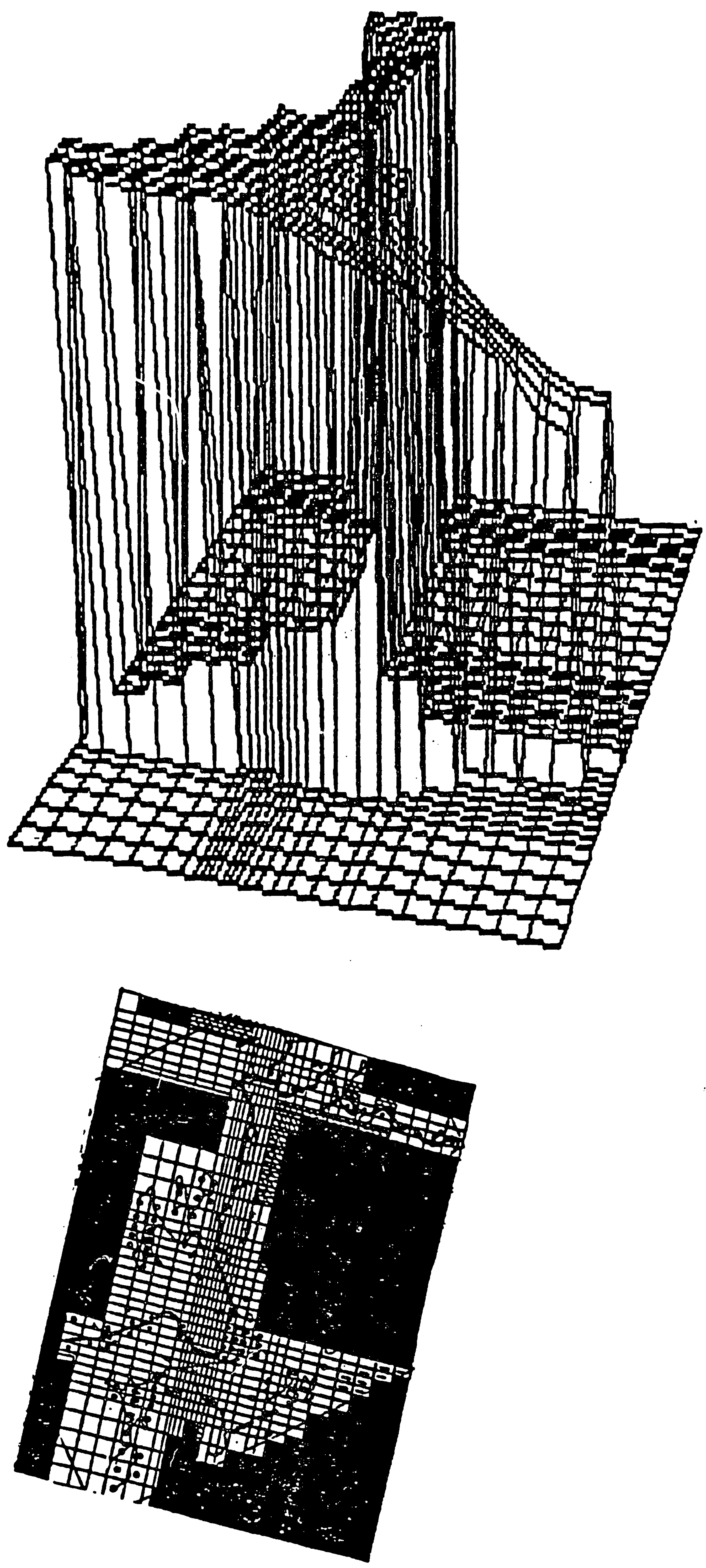


Figure 26 A three-dimensional representation of the head elevations of the lower aquifer calculated by the model. This view highlights the effect of drains which act to transmit water out of the system.

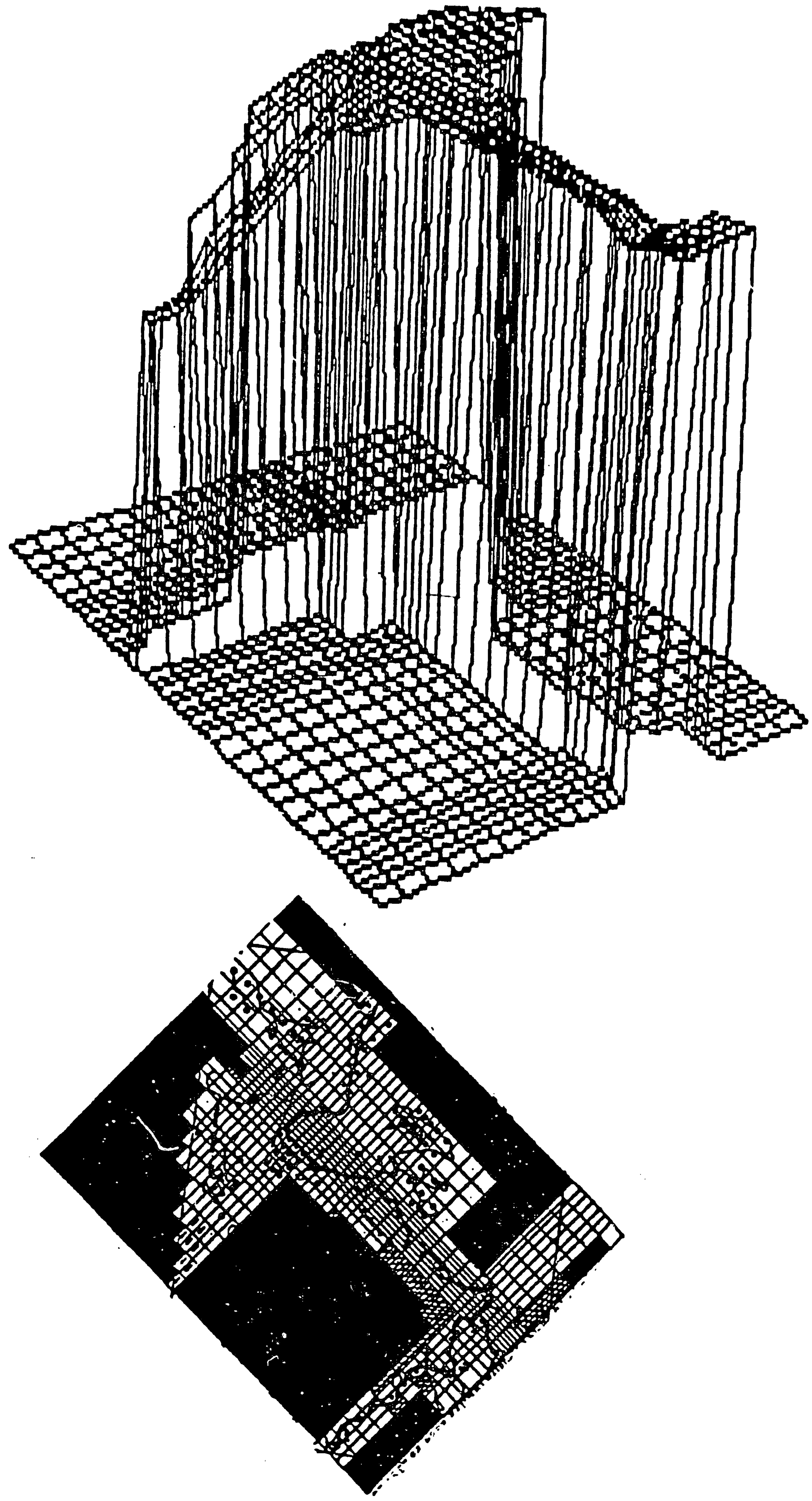


Figure 27 A three-dimensional representation of the head elevations of the lower aquifer calculated by the model. This view accentuates the effect of pumping wells at the Boonton Well Field.

TABLE 13

Volumetric Budget for Entire Model

Cumulative Volumes (ft^3/s)

	IN:	OUT:
Storage	0.0	0.0
Constant Head	0.65731E+06	15532
Wells	0.0	0.28806E+06
Drains	0.0	0.25488E+06
Recharge	0.48756E+06	0.0
River Leakage	0.23074E+06	0.81714E+06
Totals:	0.13756E+07	0.13756E+07

IN - OUT = 0.63128

PERCENT DISCREPANCY: 0.00

Rates for Time Step (ft^3/s)

	IN:	OUT
Storage	0.0	0.0
Constant Head	7.6078	0.17977
Wells	0.0	3.3340
Drains	0.0	2.9500
Recharge	5.6430	0.0
River Leakage	2.6706	9.4576
Totals:	15.921	15.921

IN - OUT = 0.73064E-05

PERCENT DISCREPANCY: 0.00

adjacent to these changed constant head nodes will dry up. Also, the volume of streamflow will decrease to a value below that considered to be an acceptable simulation of observed flow. Therefore, it is safe to assume that this amount of water is necessary to maintain the model although it is uncertain where the point of origin for this water may be. One possible source (aside from water originating from the drainage divide which flows within the water table aquifer toward the valley), is the introduction of water into the system by the underlying bedrock. A large head differential over a short distance could be created by the contrast in bedrock elevations from the ridge crests to the valley bottom. If the bedrock is sufficiently permeable then this driving head could be inducing recharge from points along the valley walls and possibly along the valley floor to the overlying sediments. Further field work, such as installation of piezometers at various depths within the valley and along the transition zone from saturated materials to the bare bedrock would help to determine whether or not the underlying bedrock is releasing water into the valley fill sediments. The documentation of uniform joint patterns observed in nearly every exposure of the crystalline bedrock (Sims, 1958) suggests that secondary permeability may exist in these rocks. Conversely, the assumption that the unconsolidated deposits proximal to the bare bedrock ridges are saturated could be erroneous.

Adjustments to the model were made by inactivating those cells that encompassed mostly bare bedrock as defined by the surficial geologic map of the area and by maintaining constant head nodes only in cells that seemed to be critical to the causation of drying adjacent nodes. Until more field investigations are conducted there seems to be little evidence to support the modeling of recharge from the underlying bedrock to the valley fill materials.

Another area of concern is presented by the anomalous seepage run between stations 43 and 65. This is the only reach along the river within the study area that exhibits influent conditions (according to the 1985 baseflow data and the 1985 high flow data). Normally, in a stream-aquifer situation where the underlying materials are relatively thick (it is uncertain how thick the deposits are beyond the point where the river turns abruptly at Boonton and flows north-northeast) it would not seem likely that the heads would decrease so suddenly as to cause a reversal of flow from the river to the underlying aquifer system. The fact that this reversal is also observed during high flow conditions (April, 1985) where bank storage should be at its maximum hints that something odd occcurring in this portion of the system. Considering what is known of the geology, there are two plausible explanations. One possibility is that there is a stratigraphic discontinuity within this reach which has created an intensification of the vertical component of flow. If the hydraulic conductivity of the underlying basal aquifer abruptly increases then the resultant effect would be an increase in flow from the overlying less permeable unit to the underlying unit thus creating a conduit for flow that passes under the upper layer (Freeze and Witherspoon, 1967). A configuration such as this is highly possible in a buried aquifer sytem of glacial origin where interfingering and stratigraphic pinchouts as illustrated in the cross sections are common. As noted above, straigraphic units do not necessarily coincide with hydrostratigraphic units, particularly in materials such as these.

The other plausible explanation is that part of the buried valley system within this reach intersects with the proposed projection of the Mount Hope fault (Figure 4). According to Sims (1958), this fault zone is highly brecciated and water has been observed to circulate freely through this conductive zone created along the fault trace. The introduction of a more permeable unit (the fractured bedrock) beneath a less conductive unit would alter anticipated flow patterns as discussed in the previous paragraph. Existing hydraulic conductivity and water level data in this troublesome reach are sparse. Therefore a strong conclusive statement as to the exact reason for the flow reversal would be premature.

Determination of the precise location of discharge and recharge areas and the identification of the suspected local and regional flow systems within the study area is greatly hindered by the complex nature of the buried valley system and the hummocky terrain. It is suspected that the major discharge areas are limited to the valley as evidenced by the maintenance of streamflow in the river whereas recharge probably occurs over the entire upland region. Insufficient data on the actual saturated thickness of the sediments and the low density of readings of elevations of the static water levels for the study area precludes the identification of other major discharge or recharge areas. It appears that there is a definite series of smaller, more local, sub-basins which are defined by the stream-aquifer sytem and by local variations in the permeabilities of the various valley fill deposits. The large hydraulic gradient of the lower layer in the vicinity of the Wisconsinan terminal moraine where maximum thickness of the unconsolidated sediments exists might indicate that the local flow system is interacting with the deeper, regional flow system of the Upper Rockaway Basin. The basal and probably very permeable unit, in this locality, is acting as a conduit transmitting significant quantities of water from the more local system to the regional basin. Further downstream, within the confines of the river valley, the interaction between local and regional systems is probably negligible.

13.0 Preliminary Sensitivity Analysis

A sensitivity analysis of the model to changes in hydraulic conductivity values of the water table aquifer was performed by increasing and decreasing the selected hydraulic conductivity array by 30 % and comparing the computed heads and the calculated river fluxes with the heads and river fluxes observed in the field. Table 14 presents the model's response to a 30 % increase of the entire hydraulic conductivity array and Table 15 shows the affects of a 30 % decrease of the hydraulic conductivity array. During the sensitivity analysis of this parameter all other aquifer parameters of the model were kept constant. Comparison of the heads caculated with the three separate hydraulic conductivity data sets indicate that the model is insensitive to this parameter when considered within a range of plus or minus 30 % of the selected hydraulic conductivity values. However, this statement applies only to the regions of the study area where water level readings are avaliable to compare with the calculated heads. Furthermore, there are only three water level measurements taken from layer 1 and these are all located at the Boonton well field. It is expected that variations in the hydraulic conductivity will have the greatest impact on the heads in the water table aquifer but because of the low density of water level measurements a conclusive statement a

Table 14 Results of the sensitivity analysis of hydraulic conductivity on the heads and streamflow calculated by the model (hydraulic conductivity values used in this simulation were determined by increasing the selected hydraulic conductivity 30 %.

Grid Location	Layer	Observed Water Level	Calculated Water Level	Difference in percent
10,12	2	148.8 m (491 ft)	150.6 m (494 ft)	1.2
10,12	1	149.7 m (491 ft)	150.3 m (493 ft)	0.7
10,11	1	150.0 m (492 ft)	149.4 m (490 ft)	0.4
9,12	2	149.4 m (490 ft)	150.9 m (495 ft)	1.0
10,11	2	148.5 m (487 ft)	150.0 m (492 ft)	1.0
9,12	1	149.4 m (490 ft)	150.0 m (492 ft)	0.4
1, 8	2	150.9 m (495 ft)	151.5 m (497 ft)	0.4
16,11	2	149.1 m (489 ft)	152.5 m (500 ft)	2.3
27,11	2	150.3 m (493 ft)	153.4 m (503 ft)	2.0
25,25	2	101.5 m (333 ft)	107.6 m (353 ft)	6.0
27,22	2	118.9 m (390 ft)	128.7 m (422 ft)	8.2

Comparison of Observed and Calculated Streamflux

Reach ID	Observed Seepage	Calculated Seepage	Difference in percent
1	+0.342 m^3/s +12.1 ft^3/s	+0.017 m^3/s +0.59 ft^3/s	95
2	-0.049 m^3/s +1.75 ft^3/s	+0.061 m^3/s +2.4 ft^3/s	24
3	-0.14 m^3/s -4.95 ft^3/s	+0.079 m^3/s +2.81 ft^3/s	156
4	+0.08 m^3/s +2.84 ft^3/s	+0.102 m^3/s +3.60 ft^3/s	27

Table 15 Results of the sensistivity analysis of hydraulic conductivity on the heads and streamflow calculated by the model (hydraulic conductivity values used in this simulation were determined by decreasing the selected hydraulic conductivity 30 %.

Grid Location	Layer	Observed Water Level	Calculated Water Level	Difference in percent
10,12	2	148.8 m (491 ft)	150.6 m (494 ft)	1.2
10,12	1	149.7 m (491 ft)	150.0 m (492 ft)	0.2
10,11	1	150.0 m (492 ft)	149.0 m (489 ft)	0.6
9,12	2	149.4 m (490 ft)	150.9 m (495 ft)	1.0
10,11	2	148.5 m (487 ft)	150.0 m (492 ft)	1.0
9,12	1	149.4 m (490 ft)	150.0 m (492 ft)	0.4
1, 8	2	150.9 m (495 ft)	151.5 m (497 ft)	0.4
16,11	2	149.1 m (489 ft)	152.5 m (500 ft)	2.3
27,11	2	150.3 m (493 ft)	153.1 m (502 ft)	1.8
25,25	2	101.5 m (333 ft)	107.6 m (353 ft)	6.0
27,22	2	118.9 m (390 ft)	128.4 m (421 ft)	7.6

Comparison of Observed and Calculated Streamflow

Reach ID	Observed Seepage	Calculated Seepage	Difference in percent
1	+0.342 m^3/s +12.1 ft^3/s	+0.024 m^3/s (+0.83 ft^3/s)	93
2	+0.049 m^3/s +1.75 ft^3/s	+0.041 m^3/s (+1.44 ft^3/s)	17
3	-0.14 m^3/s -4.95 ft^3/s	+0.053 m^3/s (+1.89 ft^3/s)	138
4	+0.08 m^3/s +2.84 ft^3/s	+0.058 m^3/s (+2.07 ft^3/s)	27

ddressing the sensitivity of this parameter to the model is not warrented. Alternatively, comparison of heads calculated for each cell of the upper layer with these three scenarios would indicate more clearly the overall sensitivity of the model to this parameter. However, the lack of field data with which to compare these simulated results makes such a comparison unwarrented at this time.

Evaluation of the calculated net river seepage for each of the three different hydraulic conductivity arrays indicates that the model is slightly sensitive to changes in this parameter. The implications in the context of the effect of this parameter on calculated river seepages are not extremely noteworthy, however, since the calculated fluxes fall within the range in error associated with current practices in the collection of stream discharge data.

The results of the preliminary sensitivity analysis suggest that the model is not extremely sensitive to variations in hydraulic conductivity. Therefore, if the hydraulic conductivity value or values used during model calibration vary by about plus or minus 30 % from the actual field conductivity values the impact on the calculated results will not be acute. In other words the model can accomadate certain ranges of error in this parameter without adversely effecting simulations.

14. Recommendations for Future Work

* A sensitivity analysis of the model's response to variations in the transmissivity, vertical conductance and recharge should be conducted prior to fine tuning of the model. The results of such an analysis will help to indicate the degree of accuracy that each of the parameters requires for a successfully calibrated model.

* Installation of piezometers, notably in the vicinity of the Wisconsinan Terminal Moraine, will supply additional data that may help to discern vertical gradients between the various units and will also assist in the eventual delineation of the glacial sediments into hydrostratigraphic units.

* Flow patterns that may exist between the local system and the more regional system of the basin particularly along a traverse from the turn in the Rockaway River (about 2.5 to 3 miles downstream from the point where the River flows within the Boonton Quadrangle) to the south-southeast edge of the study area should be identified.

* Drains used in the model to simulate the effects of groundwater moving out of the system should be replaced with constant head cells. This modification may produce a more

realistic simulation of the large hydraulic gradients existing in this region.

* The extent to which the unconsolidated sediments are saturated in areas other than along the flood plain of the Rockaway River should be determined. More specifically, it is important to know if the sediments toward the bare bedrock ridges are saturated.

* The interaction between the lakes in the study area and the underlying aquifer system should be more thoroughly scrutinized. Most likely the majority of these lakes are relics of previous glacial processes. However, how these lakes are maintaining their volume should be determined.

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Appendix A Description of Map Units

Post Glacial Deposits

af - Artificial till: Excavated till, sand, gravel, and bedrock; construction debris (brick, concrete, asphalt), cinders, and slag; in railroad and highway embankments, dams, and filled land. As much as 20 feet thick; generally less than 10 feet thick. Many small areas of fill in urban areas not mapped.

afm - Mine Tailings: Piles and embakments of waste rock excavated from iron mines and rock quarries. Includes angular boulders, cobbles, and pebbles of bedrock; minor cinders, ash, and slag, as much as 20 feet thick.

Qal - Alluvium: Dark brown to light gray, in places mottled yellow and orange brown, silt, and fine sand; minor clay and pebble to cobble gravel. Contains variable amounts of organic matter. Includes some peat and muck along the lower reaches of Beaver Brook, as much as 10 feet thick

Qs - Swamp Deposits: Typically gray silt and clay (minor sand) overlain by brown peat, in turn overlain by dark brown to black muck and organic silt. Silt and clay may also occur interbedded with the peat. In swamps along larger streams peat may be minor or absent. Generally less than ten feet thick but as much as 25 feet thick (Wakeman and others, 1943).

Qta - Talus: Angular boulders of bedrock with little or no matrix material. Forms steep apron along base of cliffs on Copperas Mountain, as much as 20 fet thick (estimated).

STREAM TERRACE DEPOSITS: Silt and fine sand to cobble gravel in terraces 5 to 80 feet above present Rockaway River.

Qst_3 - Fine sand and silt, minor pebble gravel, as much as 10 feet thick (estimated), in terraces 5 to 10 feet above present floodplain. Probably of Holocene Age.

Qst_2 - Cobble and gravel as much as 20 feet thick in terraces 10 to 40 feet above present floodplain. Formed after complete draining of glacial Lake Passaic.

Qst_1 - Poorly sorted coarse cobble and pebble as much as 10 feet thick in terraces at an elevation of 320 feet on northwest side of Boonton Reservoir, 80 feet above present river level. Deposit may include as much as 50 feet (estimated) of foreset sand beneath the cobble and pebble gravel. Possible delta built into the Great Notch stage of glacial Lake Passaic.

<u>Glacial</u> <u>Deposits</u> (Late Wisconsinan)

Qml - Maple Lake Deposits: Vertical sequence of discontinuous pebble to cobble(?) gravel (as much as 10 feet thick in the Boonton Quadranglew) overlain by silt and fine sand (as much as 100 feet thick in the Boonton Quadrangle), partially filling the valley along Kinnelan Road. Deposited in a glacial lake that drained to the south over a spillway at approximately 730 feet. Deposit thickness and coarsens toward the ice margin north of Maple Lake in the Wanaque Quadrangle.

Qbr - Butler Reservoir Deposits: Cobble and gravel partially filling valley south of Butler Reservoir. Deposited in small glacial lake draining south over a spillway at an elevation of approximately 770 feet. Largely removed by gravel pit operations; original thickness at least 20-30 feet.

ROCKAWAY VALLEY OUTWASH DEPOSITS:

Pebble gravel, sand, and minor cobble gravel deposited as a fluvial sheet on top of lake-bottom deposits (units Qldlb and Qldlb described below) after draining of Glacial Lake Denville. Includes contemporaneous deposits in the Hibernia Brook valley (Qh), Beaver Brook valley (Qb), along the Rockaway River (Qr), and in the Stony Brook valley (Qsb).

Qsb - Stony Brook Outwash: Cobble gravel on north grading to sand and pebble gravel on south, as much as 30 feet thick. Deposited by meltwater from ice margins north of the north edge of the deposit.

Qr - Rockaway River Outwash: Chiefly cobble and pebble gravel on west grading to chiefly sand on east, as much as 40 feet thick. Deposited by meltwater from ice margins to the west and north in the Dover Quadrangle.

Qb - Beaver Brook Outwash: Pebble gravel and sand, minor fine cobble gravel, as much as 20 feet thick (estimated). Deposited by meltwater from ice margins north of Meriden.

Qh - Hibernia Brook Deposits: Pebble gravel and sand, minor fine cobble gravel, as much as 20 feet thick (estimated). Deposited by meltwater from ice margins north of Hibernia.

GLACIAL LAKE PASSAIC DEPOSITS: Fine sand and silt to cobble gravel deposited as deltas in the Moggy Hollow stage of glacial Lake Passaic (Qlp_1, Qlp_2, Qlp_3, and Qlp_4); medium sand to pebble gravel (minor cobble gravel) deposited on the lake bottom, uncorrelated to major deltas (Qlpu); and fine sand, silt, and clay deposited on the lake bottom (Qlplb).

Qlp_1 - Sand and pebble gravel to south; local coarse cobble to boulder gravel to north. As much as 100 feet thick over thick till. Deposited from one or more ice margin positions south of the Rockaway River.

Qlp_2 - Delta of pebble gravel and coarse cobble gravel as much as 20 feet thick overlying silt and fine to medium (minor coarse) sand as much as 60 feet thick (estimated). Deposited from ice margin on northeast edge of deposit.

Qlp_3 - Delta of sand (minor silt and pebble gravel) overlain by pebble to cobblle gravel, and some boulder gravel. Thickens from a feather edge over thick till on the northwest to an estimated 100 feet thick on the southeast. Deposited from ice margin on northeast edge of deposit.

Qlp_4 - Delta of silt and fine to medium sand (minor coarse sand and pebble gravel) as much as 50 feet thick, overlain by sand, pebble gravel, and minor cobble gravel as much as 40 feet thick, in turn overlainb by pebble to cobble gravel as much as 10 feet thick. Entire deposit is approximately 50 feet over rock on the northwest; 100 feet thick on the southeast.

Qlpu - Medium sand to pebble gravel (minor cobble gravel) as much as 50 feet thick (estimated), generally overlying till or bedrock. Occurs in low knolls or as draped sheets over bedrock topography. Deposited on the lake bottom at the ice front, uncorrelated to major deltas.

Qlplb - Yellow-brown, orange-brown, and light gray thinly layered to massive fine sand, silt, and clay: as much as 40 feet thick; gradational to deltaic deposits in places.

GLACIAL LAKE DENVILLE DEPOSITS: Sediment deposited in glacial Lake Denville, which drained through a spillway at an elevation of approximately 525 feet at Mount Tabor in the Morristown quadrangle. South of the Terminal Moraine, fine sand, silt, and minor clay ($Qldlb_1$) is interbedded with sand and gravel (Qld_1) and till (Qlwt, described below). In places, these units are all overlain by medium to coarse sand and pebble gravel(Q_r described above). North of the Terminal Moraine there is a vertical sequence of cobble gravel (Qld_2), overlain by fine sand, silt, and minor clay ($Qldl_2$), in turn overlain in places by medium to coarse sand and pebble gravel (Qb, Qh, Qr, and Qsb; all described above).

$Qldlb_2$- Fine sand, silt, and minor clay as much as 100 feet thick. Deposited on the Lake Bottom as ice receded up valley from the Terminal Moraine.

Qld_2 - Medium to coarse sand and pebble gravel to cobble gravel deposited near the ice front on the lake bottom and as small terraces along the valley walls as ice receeded up-valley from the Terminal Moraine. As much as 100 feet thick (estimated) beneath terrace remnants; 0-30 feet thick in the subsurface on the valley bottom.

$Qldlb_1$- Fine sand, silt, and minor clay as much as 150 feet thick; interbedded with and overlying Qld_1 and Qlwt. Deposited on the lake bottom when ice was south of and at the end of the Terminal Moraine.

Qld_1 - Sand and gravel deposited on the lake bottom near the ice front before and during deposition of the Terminal Moraine. Interfingers and underlies $Qldlb_1$, and in places, Qlwt, as much as 100 feet, occurs in subsurface only.

Qgu - UNCORRELATED STRATIFIED DRIFT DEPOSITS: Chiefly sand and pebble gravel, some cobble gravel; includes some clayey silt in deposit along Stone House Brook in Kinnelon. Generally less than 20 feet thick. Form small terraces and knolls; occurs as valley bottom fill along Stone House Brook. Not correlated to ice margin positions or principal glacial lakes.

Qlwt - TILL: Unstratified and unsorted boulders, cobbles, and pebbles in a yellow-brown to orange-brown (oxidized), gray-brown (non-oxidized) silty fine sand to fine to medium sand matrix. Red-brown matrix common in area of Jurrasic bedrock in southeastern corner of the quadrangle. As much as 150 feet thick in Terminal Moraine (average approximately 50 feet), where it is interbedded with sand and gravel as much as 30 feet thick (especially in the area of glacial Lake Denville). North of the Terminal Moraine the till is as much as 60 feet thick (average about 20 feet thick) in areas of continuous till, and is generally thicker and more continuous on the north-east facing slopes of principal ridges and hills. In places till is underlain by weathered bedrock. South of the Terminal Moraine there is a narrow fringe of Late Wisconsinan till extending as much as one mile beyond the limit of hummocky topography. This till is as much 50 feet thick on the hill southeast of Indian Lake. Within Lake Denville, till south of the Terminal Moraine occurs in patches as much as 50 feet thick interbedded with Qld_1 and $Qldlb_1$.

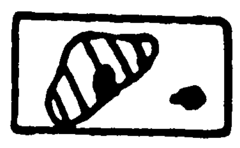
Map unit includes colluvium: mixed till and, in places, angular pebbles, cobbles, and boulders of bedrock debris in a matrix of orange-brown to yellow-brown silty fine sand to coarse sand. Forms discontinuous aprons and fans along bases of steep slopes, and may be as much as 20 feet thick.

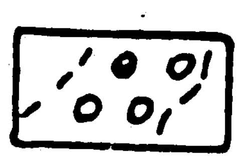
<u>Glacial</u> <u>Deposits</u> (Pre-Late Wisconsinan)

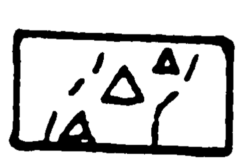
Qplwt - Till and Colluvium: Unstratified and unsorted boulders, cobbles, and pebbles of angular bedrock debris and some rounded erratics in an orange-brown silty fine sand matrix. Erratics much less abundant than in Late Wiscinsinan till. Forms broad gradually sloping aprons along lower portions of slopes. Estimated to be as much as 40 feet thick. Uncolluviated pre-Late Wiscinsinan till occurs in the subsurface beneath Late Wisconsinan deposits in and south of the Terminal Moraine, where it is as much as 20 feet thick.

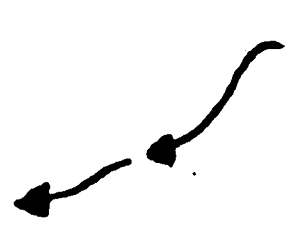
Qplwg - Stratified Drift Deposits: Sand and gravel, minor silt and clay, possibly deposited in a pre-Late Wiscinsinan glacial lake. May include pre-glacial or interglacial fluvial sand and gravel. Present in subsurface only; as much as 60 feet thick.

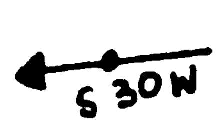
MAP SYMBOLS

 Bedrock Outcrop: Ruled pattern indicates scattered bedrock outcrop; surficial deposits generally less than 10 feet thick. Solid pattern indicates extensive bedrock outcrop; surficial deposits generally absent. Dot indicates isolated bedrock outcrops in areas where surficial deposits are generally greater than 10 feet thick.

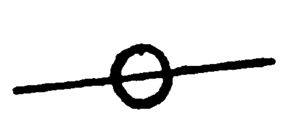
 Terminal Moraine: Belt of hummocky topography, mostly in till

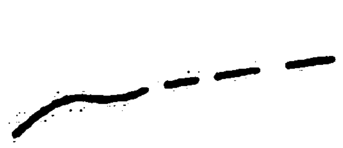
 Boulder Field: Surface accumulations of boulders, generally on terraces and valley bottoms. May be till surfaces washed by subglacial, proglacial, and ice-marginal meltwater.

 Meltwater Channel



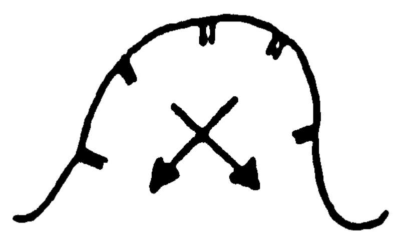
Striations

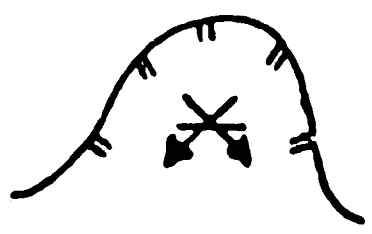
 Axis of Streamlined Till

 Contact: Dashed where Approximately located

 Scarp Eroded by Meltwater

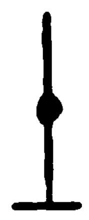
MAP SYMBOLS (Continued)

 Active Gravel Pit

 Inactive Gravel Pit

● Well or test boring to construct cross-sections

▲ Seismic shot point used to construct cross-section

 Well or test boring on cross-sections: dot indicates location of bedrock surface.

 Seismic shot point on cross-sections: dot indicates location of bedrock surface

APPENDIX B

MISCELLANEOUS WELL INFORMATION

MISCELLANEOUS WELL INFORMATION

USGS GRID NO.	270188	270189	270191	270321	270324
LATITUDE/ LONGITUDE	40 53 35 74 26 50	40 54 13 74 27 41	40 54 58 74 25 28	40 54 58 74 25 28	40 53 34 74 28 28
MODEL GRID ROW, COLUMN	17,22	16,12	27,22	21,17	27,10
LOCAL ID	Tower Hill 4	MLWD 4	MLWD 5	GEONICS 2	GEONICS 4
ALTITUDE OF LAND SURFACE	189.0 m	152.5 m	158.5 m	153.9 m	149.4 m
WELL DEPTH	140.9 m	19.5 m	101.2 m	50.9 m	56.4 m
SPECIFIC CAPACITY	8 m2/d	418 m2/d	No Data	No Data	No Data
DATE WELL COMPLETED	1/1/22	8/25/47	8/25/47	9/11/79	No Data
WATER LEVEL DATE	NA	No Data	1/8/69	9/11/79	No Data
WATER LEVEL	NA	4.6 m	37.8 m	10.7 m	No Data
PUMPING WATER LEVEL	NA	11.9 m	No Data	No Data	No Data
ALTITUDE OF WATER LEVEL	NA	147.9 m	120.7 m	143.3 m	No Data
ALTITUDE OF WATER LEVEL FOR SEPT./ OCT., 1985	170.6 m	148.9 m	112.2 m (at 6:10 pm) 119.0 m (at 6:55 am)	141.7 m	150.2 m

MISCELLANEOUS WELL INFORMATION

USGS GRID NO.	270110	270325	270117	270116	270323
LATITUDE/ LONGITUDE	40 55 00 74 26 47	40 55 42 74 26 17	40 52 43 74 31 51	40 54 07 74 28 59	40 53 35 74 27 08
MODEL GRID ROW,COLUMN	9,12	1,8	40,4	25,4	25,24
LOCAL ID	BTWD 3	GEONICS 3	DTWD 6	DTWD 4	GEONICS 1
ALTITUDE OF LAND SURFACE	151.9 m	150.9 m	167.7 m	158.5 m	155.5 m
WELL DEPTH	7.6 m	42.7 m	42.5 m	35.7 m	72.3 m
SPECIFIC CAPACITY	263 m2/d	No Data	691 m2/d	792 m2/d	84 m2/d
DATE WELL COMPLETED	8/28/46	9/24/79	9/6/77	1/13/58	9/11/79
WATER LEVEL DATE	8/64	9/24/79	7/15/60	9/28/61	8/9/60
WATER LEVEL	1.2 m	1.5 m	4.5 m	3.7 m	0.84 m
PUMPING WATER LEVEL	2.4 m	No Data	11.3 m	25.6 m	14.0 m
ALTITUDE OF WATER LEVEL	150.6 m	No Data	163.2 m	146.3 m	154.7m
ALTITUDE OF WATER LEVEL FOR SEPT./ OCT., 1985	149.4 m	150.7 m	No Data	147.0 m	101.5 m

MISCELLANEOUS WELL INFORMATION

USGS GRID NO.	270003	270109	270108	270029	270111
LATITUDE/ LONGITUDE	40 54 51 74 26 41	40 54 53 74 26 55	40 54 56 74 26 54	40 54 59 74 26 50	40 54 59 74 26 52
MODEL GRID ROW, COLUMN	10,12	10,12	10,11	9,12	10,11
LOCAL ID	BTWD 5	BTWD 2	BTWD 1	BTWD 6	BTWD 4
ALTITUDE OF LAND SURFACE	152.5 m	153.5 m	153.9 m	151.1 m	146.9m
WELL DEPTH	32.3 m	11.6 m	13.1 m	16.8 m	31.2m
SPECIFIC CAPACITY	316 m2/d	494 m2/d	514 m2/d	2685 m2/d	NA
DATE WELL COMPLETED	5/30/58	12/10/30	10/20/30	8/64	1/17/57
WATER LEVEL DATE	6/ 6/58	6/ 6/58	5/30/58	8/64	8/64
WATER LEVEL	4.0 m	3.2 m	4.1 m	1.2 m	No Data
PUMPING WATER LEVEL	9.1 m	7.6 m	8.2 m	2.4 m	No Data
ALTITUDE OF WATER LEVEL	148.3 m	146.2 m	150.7 m	149.8 m	No Data
ALTITUDE OF WATER LEVEL FOR SEPT./ OCT., 1985	148.7 m	149.8 m	149.9 m	149.4 m	148.6 m

Vita

Brenda J Sirois, the daughter of Jerome W. and Shirley Sirois, was born on May 30, 1961 in Waterville, Maine. She received a B.S. in Geological Sciences from Lehigh University in 1983.