STATE OF DELAWARE
UNIVERSITY OF DELAWARE

DELAWARE GEOLOGICAL SURVEY

Robert R. Jordan, State Geologist

BULLETIN No. 14

HYDROLOGY OF THE COLUMBIA (PLEISTOCENE) DEPOSITS OF DELAWARE: AN APPRAISAL OF A REGIONAL WATER-TABLE AQUIFER

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PREPARED BY THE UNITED STATES GEOLOGICAL SURVEY
IN COOPERATION WITH THE
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CONTENTS

ILLUSTRATIONS

TABLES

HYDROLOGY OF THE COLUMBIA (PLEISTOCENE) DEPOSITS OF DELAWARE: AN APPRAISAL OF A REGIONAL WATER-TABLE AQUIFER

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Richard H. Johnston

ABSTRACT

The Columbia (Pleistocene) deposits of Delaware form a regional water-table aquifer, which supplies about half the ground water pumped in the State.

The aquifer is composed principally of sands which occur as channel fillings in northern Delaware and as a broad sheet across central and southern Delaware. The saturated thickness of the aquifer ranges from a few feet in many parts of northern Delaware to more than 180 feet in southern Delaware. Throughout 1,500 square miles of central and southern Delaware (75 percent of the State's area), the saturated thickness ranges from 25 to 180 feet and the Columbia deposits compose all or nearly all of the water-table aquifer.

The transmissivity of the aquifer varies greatly reflecting local changes in lithology (from fine sand to coarse sand and gravel) and changes in saturated thickness. However, the hydraulic data indicate that the Columbia deposits effectively act as a medium to coarse sand aquifer. The average transmissivity in central and southern Delaware is about 7,000 ft^2/day (53,000 gpd/ft), and the average hydraulic conductivity is about 90 ft/day. Six areas of above-average transmissivity have been identified in central and southern Delaware, where transmissivity ranges from 10,000 ft^2/day (75,000 gpd/ft) to 22,000 ft^2/day (170,000 gpd/ft). These high transmissivity tracts correspond to "troughs" in a structure contour map on the base of the Columbia deposits, as well as sites of above-average saturated thickness. The highest transmissivity values occur where the saturated section contains primarily "clean" coarse sand and interbedded gravel and not necessarily where the saturated sections are thickest.

The small Coastal Plain streams of central and southern Delaware are incised into the upper part of the Columbia

deposits and derive about three-quarters of their flow from ground-water discharge. Much of the time, the streams simply act as shallow drains from the aquifer. Separation of streamflow hydrographs indicate that the average ground-water runoff from the Columbia deposits is about 800 mgd (million gallons per day). The average rate of recharge to the aquifer (computed from ground-water runoff values during the nongrowing season) is approximately 1 billion gallons per day (equivalent to 13-14 inches annually).

Present pumpage (33 mgd) is small compared to the natural discharge from the aquifer. The aquifer is capable of yielding a much greater quantity of water; however, large withdrawals will be accompanied by corresponding decreases in streamflow (unless the pumped water is returned to the streams or the aquifer).

The specific capacity of large-diameter wells ranges from about 5 to 100 gpm/ft (gallons per minute per foot of drawdown) and averages 28 gpm/ft. Throughout most of central and southern Delaware, it is possible to construct large-diameter wells capable of producing 500 gpm or more for short periods of time. However, the best method of obtaining dependable water supplies in excess of 1 mgd is to locate wells in areas of high transmissivity adjacent to streams characterized by high base flow. Such wells may obtain large amounts of water by diverting the natural discharge from the aquifer to the stream as well as by diverting water already in the stream through the streambed into the aquifer.

Water from the Columbia deposits is generally soft, slightly acidic, and characterized by low dissolved-solid content. High iron content and low pH are the only natural characteristics that may require treatment. However, the position of the Columbia deposits near land surface makes the ground water particularly susceptible to contamination. Instances of contamination reported to date have resulted from salt-water intrusion in a few coastal areas, effluent from septic tanks, leachates from landfills, and accidental spills.

INTRODUCTION

Scope and Purpose of the Investigation

The Columbia deposits of Delaware form a regional water-table aquifer which is the State's most important ground-water resource. The aquifer supplies about half the current ground-water pumpage in the State and is capable of supplying a much greater amount of water.

The purpose of the investigation described in this report was to make a quantitative hydrologic appraisal of the Columbia (Pleistocene) deposits in Delaware. Specifically, it was hoped to better define the hydraulic characteristics of this regional water-table aquifer, to estimate recharge to and discharge from the aquifer, and to ascertain the long-term availability of water from the aquifer. The lithologic character and thickness of the Columbia deposits were reasonably well known from previous studies; however, the aquifer coefficients (transmissivity and storage coefficients) were known at only a few isolated aquifer test sites. A major aim of the study was to obtain additional values of aquifer coefficients by reconnaissance techniques and to compare them with values obtained at a few selected aquifer test sites.

Early in the study it became evident that the relation of ground water to surface water would have to be investigated. It was determined that the streams in central and southern Delaware obtained most of their flow from ground-water discharge (considered equivalent to base flow) rather than from overland runoff. Much of the time the streams simply acted as drains from the aquifer; therefore, estimates of long-term discharge from the aquifer could be determined from base-flow data. Furthermore, a close correlation was established between ground-water levels and base flow in several small basins. Lowering of ground-water levels by pumping water from the aquifer would undoubtedly reduce streamflow. Thus, any statement as to the long-term availability of ground water must consider the effect on streamflow. It was decided to study the relation of ground water to surface water in several small basins and apply the results of these studies to a state-wide appraisal of the shallow aquifer-stream system. The studies of small basins are summarized in a short paper (Johnston, 1971) and will be described in detail in another paper, as yet unpublished.

The transmissivity of the Columbia deposits and, therefore, the yield and specific capacity of wells was known to vary widely throughout the Delaware Coastal Plain. The preparation of a statewide transmissivity map was given high priority during this study. In preparing such a map, it was hoped to make use of reconnaissance techniques (utilizing base-flow recession and ground-water-level recessions) to estimate transmissivity. If the attempt were successful, these techniques could also be useful in future prospecting for large ground-water supplies in Delaware.

The area investigated included all the Coastal Plain of Delaware. However, particular emphasis was placed on central and southern Delaware, where the Columbia deposits form a regional water-table aquifer. This area of emphasis includes all of Sussex County and most of Kent County (Figure 1). The locations of counties and major towns in Delaware, as well as streams, gaging stations, observation wells, production wells, and aquifer test sites referred to in the text are shown in Figure 1.

Acknowledgments

The writer wishes to express his appreciation to Robert R. Jordan, State Geologist and cooperator in this study, for providing much assistance, especially during the field phases of the investigation. Kenneth D. Woodruff and John C. Miller, hydrogeologists with the Delaware Geological Survey, assisted during pumping tests, and Mr. Woodruff geophysically logged several test holes. In addition, exploratory holes were drilled and piezometer tubes were installed using a University of Delaware combination auger and rotary drill rig operated by Boris J. Bilas of the Delaware Geological Survey.

Many individuals aided the writer during the study. It is not possible to single out everyone by name; however, special thanks are given to Mr. and Mrs. James B. Carpenter, who permitted construction of wells on their property for an aquifer test. Thanks are also given to Paul E. White and Sons, Inc. and Delmarva Drilling Company for providing many well logs and pumping-test data.

The writer also wishes to thank his colleagues in the U. S. Geological Survey; Philip Pfannebecker, R. H. Simmons, E. M. Cushing, K. R. Taylor, and I. H. Kantrowitz for suggestions made during the study and for supplying unpublished basic data.

The investigation and report preparation were under the direction of W. F. White, District Chief, U. S. Geological Survey.

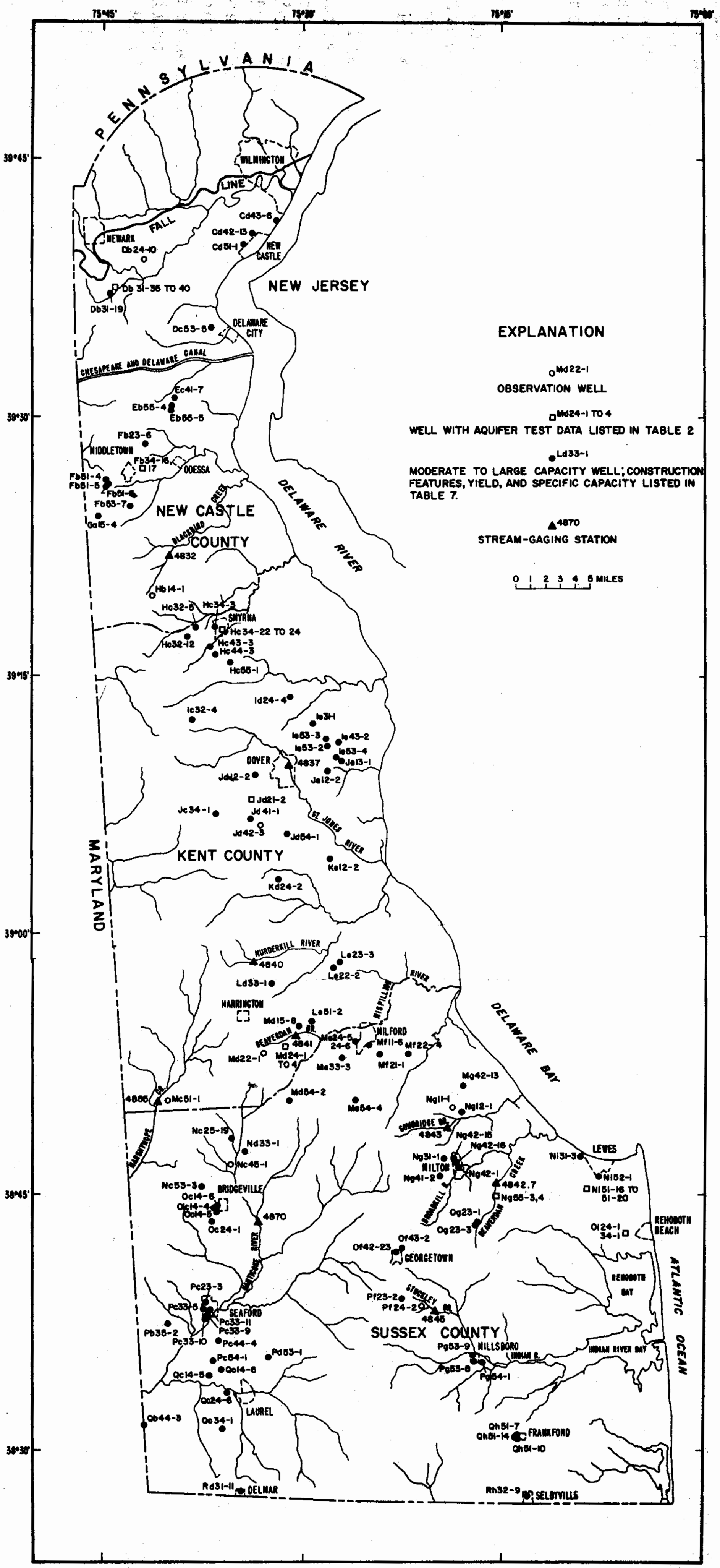

FIGURE I. LOCATIONS OF WELLS AND STREAM-GAGING STATIONS REFERRED TO IN THE TEXT.

Previous Work

Much geologic and hydrologic information bearing on the Columbia deposits has been collected during the past 20 years. Since the formation of the Delaware Geological Survey (DGS) and initiation of a cooperative agreement with the U. S. Geological Survey (USGS) in 1951, the geology and water resources of Delaware have been systematically studied. Early reports, made under joint direction of J. J. Groot (State Geologist) and W. C. Rasmussen and D. R. Rima (District Geologists for the USGS), described the areal occurrence of ground water in Delaware and contain much useful data about the Pleistocene deposits. During the late 1950's and early 1960's, a series of water-table maps was prepared by the USGS and DGS in cooperation with the Delaware State Highway Department. These maps cover the Coastal Plain section of Delaware. They have been very useful in the present study and recent water-level measurements indicate that they are still essentially correct.

During 1967-71, another statewide assessment of ground-water resources was made by the Water Resources Center of the University of Delaware, together with the DGS. The studies have provided a current updating and assessment of the major aquifers in Delaware (see reports by Sundstrom and Pickett listed in references).

The geology of the Pleistocene deposits has been continuously studied by the DGS since its inception. Reports describing the stratigraphy of the Pleistocene deposits by Jordan and Spoljaric (see references) have been particularly useful in understanding the aquifer geometry. In recent years the DGS has applied geological and geophysical techniques in solving local water-supply problems and evaluating Pleistocene channel aquifers. One of the more detailed studies is a recent evaluation of the Columbia deposits in the Middletown-Odessa area by Spoljaric and Woodruff (1970).

In order to make quantitative hydrologic studies of a regional water-table aquifer, such as the Columbia deposits of Delaware, it is essential to have long-term records of streamflow and ground-water levels. Those basic records, collected by the USGS in cooperation with the DGS and the Delaware State Highway Department, provided the basis for estimating discharge from the aquifer as well as that for estimating aquifer coefficients.

HYDROGEOLOGY OF THE COLUMBIA (PLEISTOCENE) DEPOSITS

Lithologic Character and Stratigraphy

Surficial sands of presumed Pleistocene age mantle the older (Cretaceous and Tertiary) sediments nearly everywhere in the Coastal Plain of Delaware. These sands form an unconfined aquifer of regional extent throughout the Delmarva Peninsula. Because of their large areal extent and because several environments of deposition are represented, authors have designated the Pleistocene deposits by several names. Some of the more recent nomenclature is listed in Table 1.

However, for the purpose of this report, a detailed stratigraphic nomenclature is not needed. The general term, "Columbia deposits," is used here and refers to all surficial sand and associated gravel, clay, and silt beds of presumed Pleistocene age in Delaware. The term Columbia deposits, as used in this report, is synonymous with such terms as Pleistocene deposits or Pleistocene aquifer. Use of the term "Columbia" follows Jordan (1962), who proposed the name Columbia Formation for the fluvial Pleistocene deposits of northern and central Delaware and, where the deposits are subdivisible into two formations in southern Delaware, the name Columbia Group (Table 1). A discussion of the historical precedence for use of the term Columbia is given by Jordan (1962). All recent reports of the Delaware Geological Survey have used the terms Columbia Group or Formation as have several recent reports of the Maryland Geological Survey and the USGS (see for example: Back, 1966; Hansen, 1966; and Otten, 1970). However, some objection has been raised to use of the term "Columbia" for all the Pleistocene sediments in Delaware. Based on a recent geologic mapping, J. P. Owens of the USGS (oral communication, Oct. 1972, Table 1) believes that the Pleistocene deposits of northern and southern Delaware are too dissimilar in lithology and genesis for use of one inclusive stratigraphic term such as Columbia Group.

The orange to reddish-brown sand beds which Rasmussen and others (1960) termed the Brandywine Formation of Pliocene (?) age and Hansen (1966) called the red gravelly facies of the Salisbury Formation (Pleistocene), are considered to be part of the Columbia (Pleistocene) deposits in this report. As noted by Jordan (1962), there is no definitive evidence for the age of these beds and recent geologists have considered the unit as part of the Pleistocene deposits. In general, the stratigraphic names listed in Table 1 are not used throughout this report. An exception is the Beaverdam Sand - a readily identifiable white, medium sand found throughout southern Delaware.

<table>
<tr><th colspan="2">RASMUSSEN AND OTHERS (1960)
SUSSEX CO., DELAWARE</th><th colspan="4">JORDAN (1962)
DELAWARE</th><th colspan="3">HANSEN (1966)
SALISBURY, MD. AREA</th><th colspan="3">OWENS (ORAL COMMUNICATION, OCTOBER, 1972)</th><th colspan="2">THIS REPORT
DELAWARE</th></tr>
<tr><th rowspan="2">SERIES</th><th rowspan="2">FORMATION</th><th rowspan="2">SERIES</th><th colspan="3">FORMATION OR GROUP</th><th rowspan="2">SERIES</th><th colspan="2" rowspan="2">FORMATION OR 'FACIES'</th><th rowspan="2">SERIES</th><th colspan="2">FORMATION OR 'FACIES'</th><th rowspan="2">SERIES</th><th rowspan="2">AQUIFER</th></tr>
<tr><th>N. DELAWARE</th><th colspan="2">S. DELAWARE</th><th>N. DELAWARE</th><th>S. DELAWARE</th></tr>
<tr><td rowspan="4">PLEISTOCENE</td><td>PARSONBURG SAND</td><td rowspan="4">PLEISTOCENE</td><td rowspan="4">COLUMBIA FORMATION</td><td rowspan="4">COLUMBIA GROUP</td><td rowspan="3">OMAR FORMATION</td><td rowspan="5">PLEISTOCENE (COLUMBIA GROUP)</td><td colspan="2" rowspan="3">WALSTON CLAY
PARSONBURG SAND</td><td rowspan="5">PLEISTOCENE</td><td rowspan="5">PENSAUKEN FORMATION</td><td rowspan="3">UNNAMED 'BLACK-BARRIER FACIES'</td><td rowspan="5">PLEISTOCENE</td><td rowspan="5">COLUMBIA (PLEISTOCENE) DEPOSITS</td></tr>
<tr><td>PAMLICO FORMATION</td></tr>
<tr><td>WALSTON SILT</td></tr>
<tr><td>BEAVERDAM SAND</td><td>BEAVERDAM FORMATION</td><td rowspan="2">SALISBURY FORMATION</td><td>BEAVERDAM FACIES</td><td rowspan="2">BEAVERDAM FORMATION</td></tr>
<tr><td>PLIOCENE ?</td><td>BRANDYWINE FORMATION</td><td>(?)</td><td colspan="3">BRANDYWINE FORMATION (?)</td><td>"RED GRAVELLY FACIES"</td></tr>
</table>

TABLE 1. CORRELATION CHART SHOWING STRATIGRAPHIC NOMENCLATURE USED FOR THE PLEISTOCENE DEPOSITS OF DELAWARE AND ADJACENT MARYLAND.

The Columbia deposits are composed principally of sands which occur in channel fillings in northern Delaware and as a broad sheet across central and southern Delaware. In addition to sand, the Columbia sediments contain subordinate amounts of gravel, clay, and silt. In central and southern Delaware, where the Columbia deposits constitute a major regional aquifer, the deposits are composed mostly of fine to coarse moderately well sorted quartz sand. As noted by Jordan (1964), the mass of the Pleistocene sediment may be "accurately described as a medium sand." Evidence presented later in the report will show that, hydrologically, the Columbia deposits are effectively acting as a medium to coarse sand aquifer.

In the areas of highest transmissivity (for example at Smyrna and near Houston, as shown in Figure 9), the Columbia deposits consist of well sorted coarse sand with or without bands of gravel. Locally gravel may constitute a sizable fraction of the sediments. Spoljaric and Woodruff (1970) determined that gravel constitutes 30 percent of the sediments in the Middletown-Odessa area. However, the presence of gravel does not necessarily indicate maximum transmissivity. For example, a 110-foot thick section of gravel, with interbedded fine to coarse sand (apparently poorly sorted), near Milton has lower transmissivity than 90-foot sections of predominantly coarse sand at Smyrna and Houston (see Table 2).

The Columbia deposits differ widely in color, ranging from reddish brown and purplish black through shades of brown to tan, yellow, or light gray. The color of the sediments is related to the amount of iron present, as shown by Spoljaric (1971). According to Spoljaric, dark brown sand contains greater than 4 percent ferric iron, whereas the yellow and light gray sand contains less than 1 percent ferric iron. The purplish black color occasionally seen in ironstone beds is probably due to manganese oxides.

The differentiation of the Columbia deposits from underlying units can be made on a lithologic basis in much of northern and central Delaware. However, in southern Delaware, where Miocene sands may directly underlie Pleistocene sands, the differentiation is often difficult. This is particularly true in the case of the Miocene Manokin aquifer, which underlies the Pleistocene throughout a 7-mile-wide belt extending southwest across Delaware from Milton to Laurel (see Figure 2, Sundstrom and Pickett, 1970). The difficulty arises because of the similarity of the white fine to coarse sand of the Beaverdam Sand (Pleistocene) and the gray medium-coarse sand of the underlying Miocene age Manokin aquifer. As pointed out by

Sundstrom and Pickett (1970), the Manokin sands are generally grayer and better sorted than the Pleistocene sands. This rather subjective criterion was used by the writer in identifying the base of the Columbia deposits from well logs in southern Delaware. In central Delaware, the subcropping Miocene beds often consist of gray silty clay or sand with abundant shell material, and the base of the Pleistocene can be identified with more confidence. However, drillers' logs occasionally report thick sections of "white or tannish gray sand." In such cases, the base of the Pleistocene was arbitrarily placed at the uppermost occurrence of a thick gray or "blue" clay bed. Thus, some of the upper Miocene sands, particularly in the Manokin aquifer, may have been included with the Columbia deposits. However, hydrologically the sands are acting as an aquifer unit. Microfossils are often useful in identifying the Miocene-Pleistocene contact; however, geologists of the Delaware Geological Survey report that good index fossils are difficult to obtain from cores and drill cuttings.

Areal Extent and Saturated Thickness

The Columbia (Pleistocene) deposits occur as channel fillings and thin isolated patches in New Castle County and as a broad sheet across most of Kent and Sussex Counties (Figure 3). The Pleistocene sediments are generally considered to be fluvial in origin and, according to Jordan (1964), were deposited by streams entering Delaware from the northeast and spreading south and southeast across Delaware. The narrow channels of northern New Castle County coalesce into a system of braided channels in the southern part of the county (Spoljaric, 1967). From the Kent-New Castle County line south, the Columbia deposits are basically a sheet of sand, which thickens southward across Kent and Sussex Counties. In extreme southern Delaware, these deposits were probably reworked by transgressing-regressing seas (Jordan, 1964). Jordan describes beach, dune, estuarine, offshore bar, and lagoonal facies within the area of marine transgression. These facies of the Columbia sediments, as well as the configuration of the channels, are related to the aquifer transmissivity as will be discussed later.

Figure 2 shows a structure contour map of the base of the Columbia deposits in Delaware (except for northern New Castle County). The map is intended to show the basal configuration of the Pleistocene sediments where they constitute an important regional aquifer - namely central and southern Delaware. For a description of the Pleistocene channels of northern Delaware, see Spoljaric (1967).

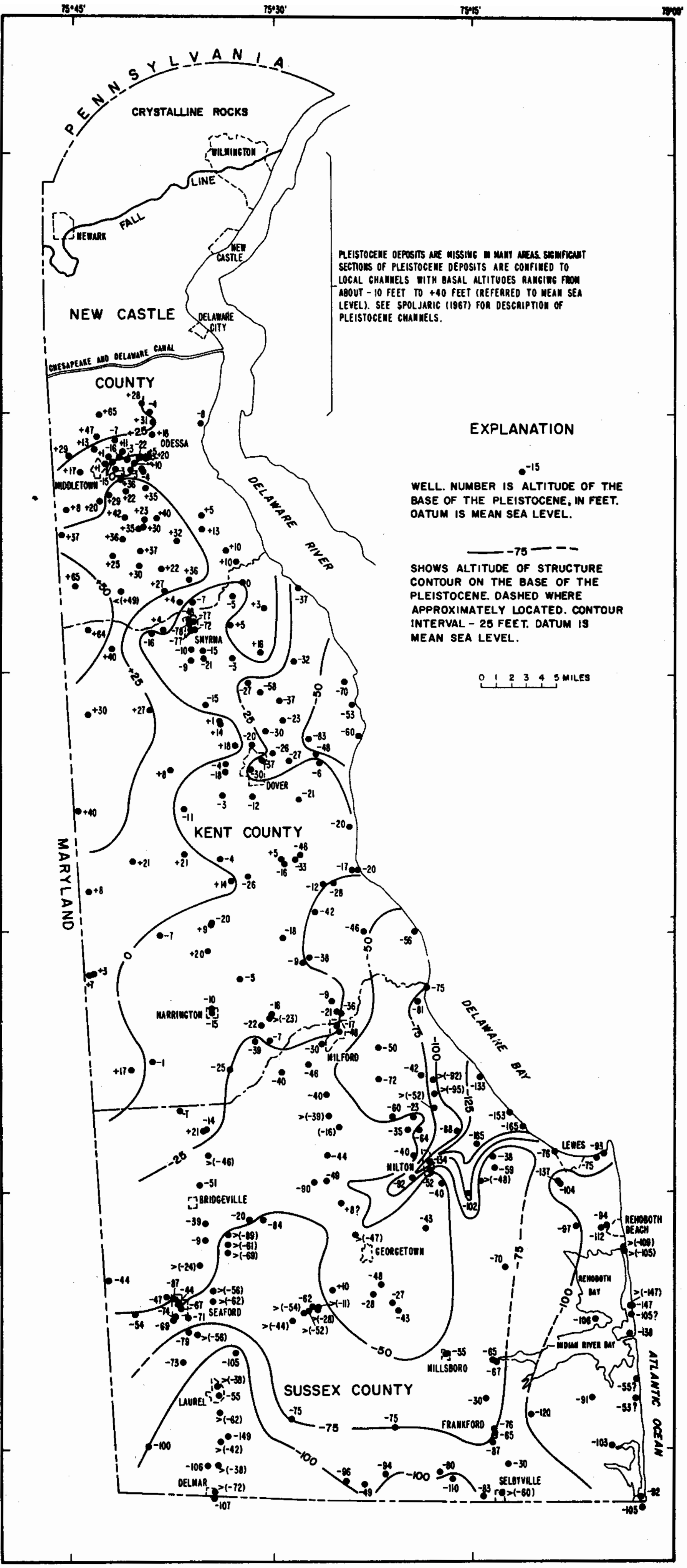

FIGURE 2. STRUCTURE CONTOUR MAP OF THE BASE OF THE COLUMBIA (PLEISTOCENE) DEPOSITS IN DELAWARE.

In general the base of the Pleistocene deposits slopes to the southeast. Altitudes range from about 25 to 50 feet above mean sea level in southern New Castle County to more than 150 feet below sea level in southern Delaware. The 25-foot contours (necessitated by available well data) are insufficient to define the many paleostream channels marking the base of the Pleistocene. However, a few regional troughs (containing former stream channels or groups of channels?) are apparent in Figure 2. A trough extends southeastward from Smyrna and turns eastward toward the Delaware Bay north of Dover. Another trough, possibly containing two major channels, trends northeast from Milton along the present-day course of the Broadkill River. An apparent major trough extends southward from a point midway between Bridgeville and Georgetown through Laurel toward the Delaware-Maryland state line. These troughs are the sites of highly permeable Pleistocene deposits, as will be discussed later in the report.

The saturated thickness of the Columbia deposits ranges from a few feet in many parts of New Castle County and in the northeast corner of Kent County to more than 180 feet in southwestern Sussex County. Figure 3 shows the saturated thickness of the Columbia deposits throughout Delaware (except for northern New Castle County). The north to south increase in saturated thickness is evident from the map. Several regional troughs (corresponding to the troughs in the Pleistocene base map) are characterized by saturated sections thicker than those in the immediately surrounding area.

About 1,500 square miles of Delaware (75 percent of the State's area) are underlain by moderately thick to very thick (25 to 180 feet) sections of saturated aquifer. The areas, by county, underlain by significant saturated sections of aquifer are as follows:

County	Total area (sq mi)	Area underlain by 25 to 180 feet of saturated Columbia deposits (sq mi)	Percentage of area underlain by 25 to 180 feet of saturated Columbia deposits
New Castle	437	21	5
Kent	595	538	90
Sussex	946	946	100
Statewide	1978	1505	75

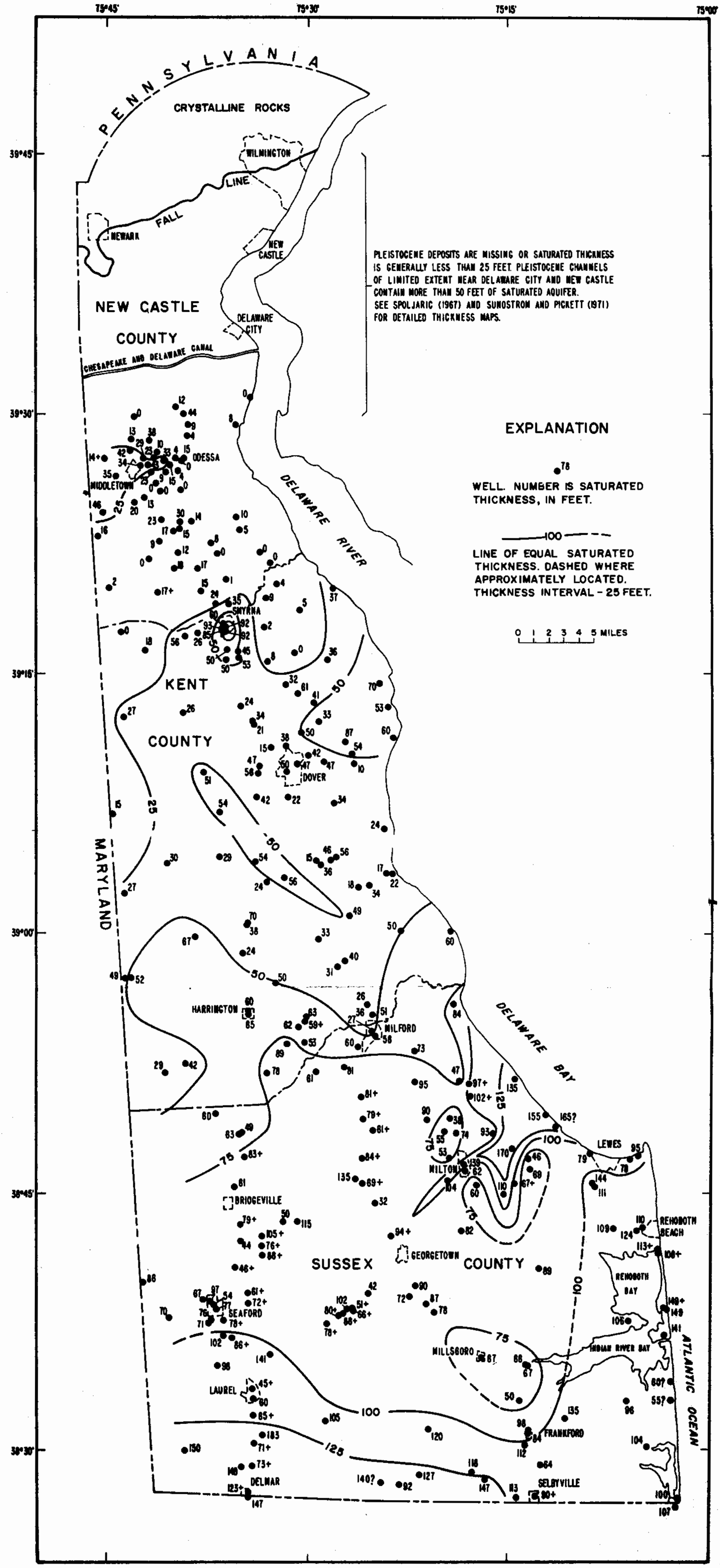

FIGURE 3. SATURATED THICKNESS MAP OF THE COLUMBIA (PLEISTOCENE) DEPOSITS IN DELAWARE.

The thickness values listed for Kent and Sussex Counties were obtained from Figure 3 and the value for New Castle County was estimated from a saturated thickness map presented by Sundstrom and Pickett (1971)

Within the 1,500 square mile area underlain by 25 to 180 feet of saturated section, the Columbia deposits are essentially the water-table aquifer, or represent the most permeable section of the water-table aquifer. In Kent and Sussex Counties, the Columbia deposits are underlain either by Miocene clay beds or one of several Miocene sand aquifers. In general, the Pleistocene sands are thicker and more permeable than the subcropping Miocene sands. With a few local exceptions, the water-table aquifer in central and southern Delaware can be considered as the saturated section of the Columbia deposits.

Hydraulic Characteristics

Transmissivity and Storage Coefficient Determined by Aquifer Tests

Aquifer tests involving a pumping well and nearby observation wells are commonly used to obtain the aquifer coefficients: transmissivity (T), permeability or hydraulic conductivity (K), and storage coefficient (S). However, aquifer tests of the unconfined Pleistocene deposits have many drawbacks. Properly made aquifer tests are relatively expensive and time-consuming, and too often the actual hydrogeologic conditions at the site depart markedly from the idealized conditions required by the equations used to analyze the test results. Close proximity of the pumped well to sources of recharge (streams and lakes), failure to pipe water beyond the cone of influence of the pumped well (to avoid recycling of water), and rain during the tests may give unusable results. On the other hand, recent advances in ground-water hydrology have made it possible to analyze test data that is affected by delayed yield from storage above the water table, differences in horizontal and vertical conductivity (anisotropy), and partial penetration of the aquifer by the pumped well or observation wells.

Logarithmic plots of drawdown versus time for pumping tests of the Columbia deposits typically have an "S" shaped configuration. The early time data (0-5 minutes) often produces a curve that closely matches the nonleaky artesian-type curve of Theis (1935). This early time response is probably caused by depletion of artesian-type storage below the water table. The middle section of the "S" curve is characteristically flat and may result from delayed drainage above the lowered water table and from vertical flow components due to partial penetration of the pumped well and anisotropy in the aquifer. The flat section of the curve may also be produced by nearby sources of recharge or by recycling of pumped water, either of which may render the test data unusable. The late section of the time-drawdown plots (a few hours to a few days after pumping begins) often displays an upward curvature due to completion of delayed drainage. In the past, many hydrologists have simply matched the late data against the Theis artesian type curve to compute T and S assuming that vertical drainage was complete, or nearly so. However, the combined effects of anisotropy in the aquifer and partial penetration of wells plus continued effects of delayed drainage may produce a time-drawdown plot that cannot be realistically matched to the Theis artesian type curve. Recently Boulton (1963) has presented type curves that account for delayed yield from storage, and Stallman (1965) has presented type curves that consider vertical flow components caused by partially penetrating wells and differences in horizontal and vertical conductivity. Boulton's and Stallman's type curves, as presented by Lohman (1972), as well as the Theis artesian type curve, have been used to analyze data from the three tests described in this section. For a thorough discussion of these and other methods of aquifer test analysis, the reader is referred to Stallman (1971) and Lohman (1972).

Table 2 lists aquifer coefficients determined from aquifer tests on the Columbia deposits for which usable data were available. Except for the Houston, Middletown, Milton and, possibly, the Smyrna test, the aquifer coefficients listed should be considered rough approximations. The Houston and Milton tests were made in areas underlain by thick sections (86 and 110 feet) of sand and gravel, and the values of transmissivity (T) and hydraulic conductivity (K_r) obtained are above average for the Columbia deposits. The Middletown test was made in an area underlain by a moderately thick saturated section (42 feet) of medium to coarse sand and the T and K_r values obtained are more typical of the Columbia deposits on a statewide basis.

The values of storage coefficient (S) listed in table 2 (0.01 to 0.07) are within the range generally considered characteristic of unconfined aquifers. However, these S values were obtained from relatively short aquifer tests and do not represent the true specific yield of the aquifer -- which would be effective for long periods of pumping. After a long period of draining (several months), the specific yield is probably close to 0.14 - the average value obtained by a reconnaissance method discussed in a later section.

TABLE 2 - Summary of Aquifer Tests of the Columbia Deposits in Delaware

Location (nearest town)	Owner	Date of Test	Pumping Period (hours)	Wells P=pumped well O=observation Well	Pumping Rate (gpm)	Lithology	Saturated Thickness (b) (feet)	Transmissivity (T)	Hydraulic Conductivity (Kr) (ft/day)	Storage Coefficient (S)	Remarks
Dover	Papen Farms	5/10/68	3½	Jd 21-2 (P) 3 observation wells	750	Fine to coarse sand	42	3,100 ft^2/day (23,000 gpd/ft)	70	?	Distance-drawdown plot gives T=23,000 gpd/ft. Drawdown-time plot matched against Theis artesian type curve gives T=33,000 gpd/ft. Test too short for accurate determination of T and S.
Glasgow	E. I. du Pont de Nemours & Co.	7/12-7/16/69	95	Db 31-35 (P) Db 31-28 (O) Db 31-37 (O) Db 31-39 (O) Db 31-40 (O)	500	Medium (?) sand	75	1,200 ft^2/day (9,000 gpd/ft)	15	?	Complicating factors include partially penetrating pumped well, substantial dewatering of aquifer, and possible recycling of pumped water after 6 hrs. Most probable value of T is 9,000 gpd/ft obtained with early corrected drawdowns and Theis artesian type curve.
Houston	U. S. Geol. Survey and Delaware Geol. Survey	5/17-5/21/71	48	Md 24- 3 (P) 24- 1 (O) 24- 2 (O) 24- 4 (O)	305	Coarse sand with gravel lenses	86	22,000 ft^2/day (165,000 gpd/ft)	250	0.05	See discussion in text.
Lewes	Town of Lewes	12/54	Five tests of variable duration	Ni 51-16 to 51-20 (P) Ni 51- 1 to 51-11 (O)	Different rates in five tests	Coarse sand	130	15,000 ft^2/day (110,000 gpd/ft)	110	.01 to .02	Values of T range from 84,000 to 135,000 gpd/ft for 5 pumping tests, as reported by W. C. Rasmussen in U.S.G.S. files.
Middletown	University of Delaware	5/7-5/8/70	24	Fb 34-16 (P) 34-17 (O) 34-18 (O)	60	Coarse to medium sand	42	4,500 ft^2/day (33,000 gpd/ft)	107	.05	See Spoljaric and Woodruff (1970, p. 96-100) and discussion in text. Conductivity ratio (K_2/Kr)=0.25

TABLE 2 - (continued)

Location (nearest town)	Owner	Date of Test	Pumping Period (hours)	Wells P=pumped well O=observation Well	Pumping Rate (gpm)	Lithology	Saturated Thickness (b) (feet)	Transmissivity (T)	Hydraulic Conductivity (Kr) (ft/day)	Storage Coefficient (S)	Remarks
Milton	U.S. Geol. Survey and Delaware Geol. Survey. Wells are located on property of James B. Carpenter	12/ 8-12/12/70	48	Ng 55- 4 (P) 55- 3 (O)	200	Fine to coarse sand and gravel	110	14,000 ft^2/day (104,000 gpd/ft)	130	0.02	See discussion in text.
Rehoboth Beach	Town of Rehoboth Beach	1952	?	Oi 24- 1 (P) 34- 1 (O)	?	Fine to coarse sand	120	7,300 ft^2/day (55,000 gpd/ft)	60	.01	Reported values of T and S in U.S.G.S. files.
Smyrna	City of Smyrna	3/25/58	13	Hc 34-22 (P) 34-23 (O) 34-24 (O)	1,000	Coarse sand	79	16,000 ft^2/day (120,000 gpd/ft)	200	.02 to .07	Boulton delayed-yield type curve match point gives T=19,000 ft^2/day and S=.02. Stallman partial-penetration type curve match point gives T=12,000 ft^2/day, S=0.07, and conductivity ratio (K_2/Kr)=0.1

Houston aquifer test

The Houston aquifer test site is about midway between Harrington and Milford, Del. as shown by the aquifer test symbol on Figure 1. At the pumped well, the Columbia deposits consist of coarse sand and interbedded gravel beds which extend to a depth of 90 feet and are underlain by Miocene clay which acts as a lower confining bed. When the test was made in May 1971, the water table was about 4 feet below land surface and the saturated thickness (b) was, therefore, 86 feet. A lithologic log at the site and the spacing and construction of wells are shown in Figure 4. As shown in the sketch, the pumped well is screened near the base of the aquifer (70-80 feet), and three observation wells screened near the top, middle and base of the aquifer are located 80-100 feet from the pumped well. Thus, the observation wells are located at a distance from the pumping well about equal to the saturated thickness (b).

Well Md24-3 was pumped at a constant rate of 305 gpm for 48 hours, and recovery measurements were made for 48 hours after cessation of pumping. The orifice method was used to maintain a constant discharge of 305 gpm ($\pm$ 5 percent). Complicating factors included rain during the last hours of recovering, which precludes use of the final recovery measurements, and a slight decline in the discharge rate (to 290-300 gpm) during the final 8 hours of pumping. Pumped water was removed from the site by 600 feet of irrigation pipe and a ditch to Beaverdam Branch a quarter of a mile away.

The drawdown in the three observation wells (after 48 hours of pumping) was relatively small (1.1 to 1.2 feet) compared with the saturated thickness (86 feet). Therefore, it was not necessary to correct observed drawdowns for dewatering of the aquifer, as discussed by Jacob (1963).

Figure 5 shows a logarithmic plot of drawdown (s) against time (t) for the mid-aquifer observation well, Md24-1. The plot has been analyzed by Boulton's (1963) model, which accounts for delayed yield from storage. The plot was superimposed on Boulton's delayed-yield type curve (as presented by Lohman, 1972, plate 8) to obtain the match point and five parameters shown in Figure 5. Transmissivity (T) is calculated as follows:

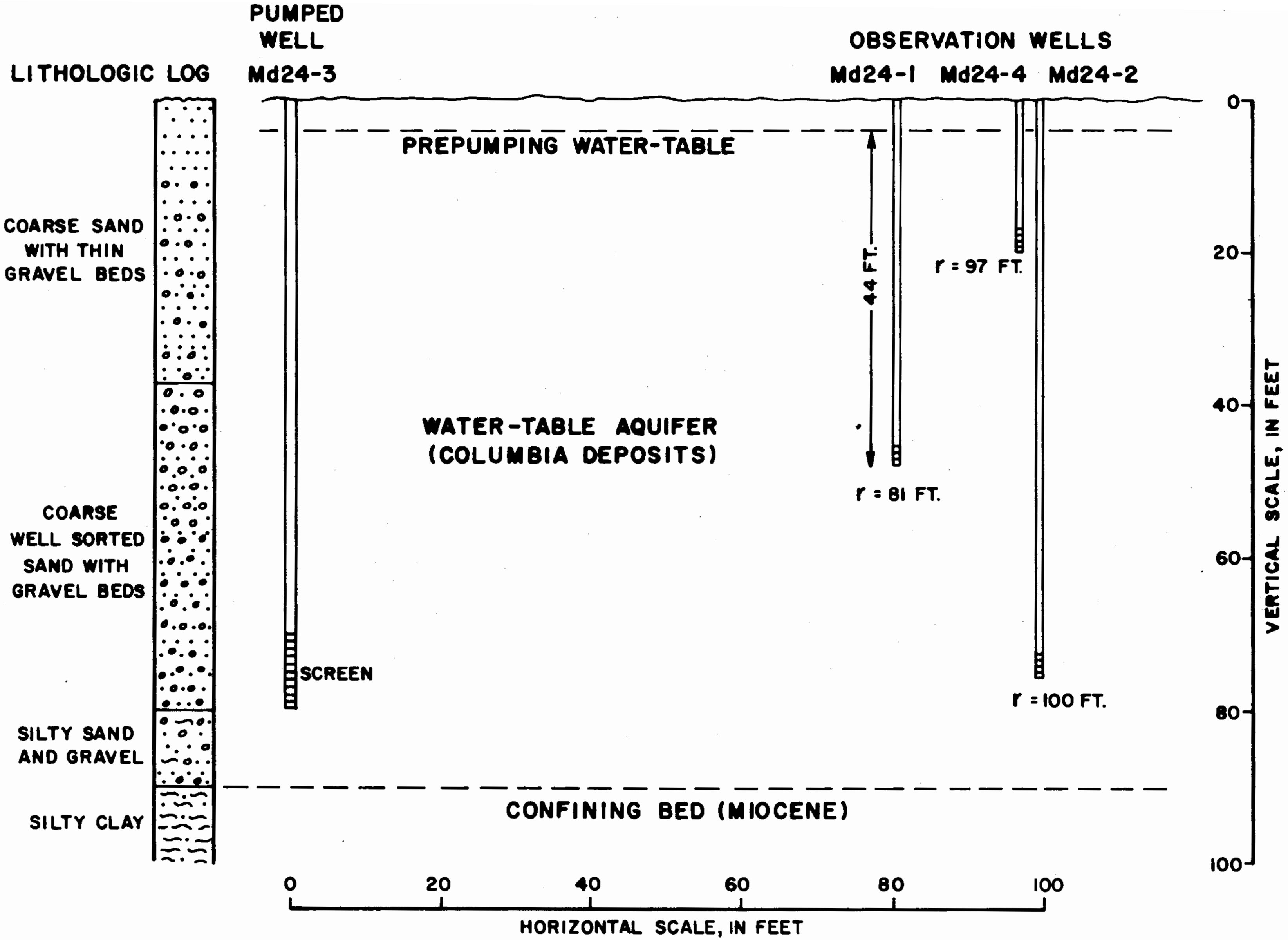

Figure 4. Section through Houston aquifer-test site showing lithology, well construction, and well spacing.

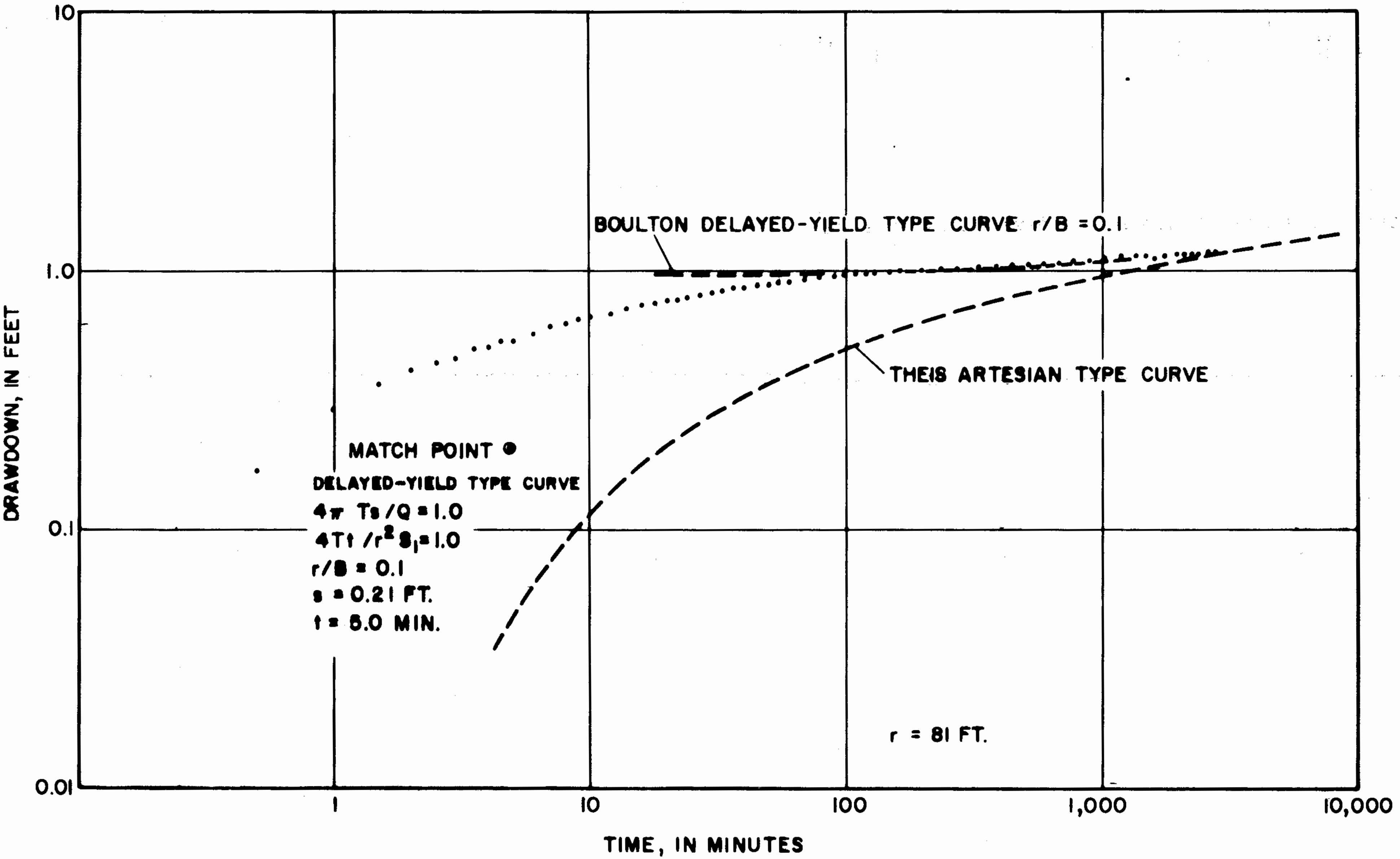

Figure 5. Logarithmic plot of drawdown against time for mid-aquifer observation well (Md24-1) at Houston aquifer test, May 17-19, 1971.

$$T = \frac{1.0Q}{4\pi s}$$

$$T = \frac{(1.0)\ (305\ \text{gal/min})\ (1{,}440\ \text{min/day})}{(4)\ (3.14)\ (0.21\ \text{ft})\ (7.48\ \text{gal/ft}^3)}$$

$$T = 22{,}000\ \text{ft}^2/\text{day} \quad \text{or} \quad 165{,}000\ \text{gpd/ft}$$

Storage coefficient is calculated as follows:

$$S = \frac{4Tt}{r^2\ (1.0)}$$

$$S = \frac{(4)\ (22{,}000\ \text{ft}^2/\text{day})\ (5\ \text{min})}{(6{,}600\ \text{ft}^2)\ (1{,}440\ \text{min/day})}$$

$$S = 0.05$$

This value for storage coefficient is low compared with the value of S = 0.13 obtained by the hydrologic budget method described in the following section. It seems probable that drainage from above the lowered water table was incomplete after the 48 hours of pumping, as attested to by the fact that the very late data on the time-drawdown plot is just approaching the configuration of the Theis artesian type curve (Figure 5).

The aquifer at the Houston site is undoubtedly anisotropic, as shown by the lithologic log and also by the fact that delayed drainage persisted for at least 48 hours.

Figure 6 shows a logarithmic plot of drawdown against time for observation well Md24-2, which is screened near the base of the aquifer. Superposition of this plot on Boulton's delayed-yield type curve (r/B = 0.1) produces the match point and associated parameters shown in Figure 6. The calculated value of T = 23,000 ft^2/day (175,000 gpd/ft) is about the same as obtained for observation well Md24-1. However, the value of S = 0.03 seems anomalously low.

Weeks (1969) has presented a method for determining the ratio of horizontal (K_r) to vertical conductivity (K_z) from drawdowns observed in partially penetrating observation wells. Further, Weeks (written commun., April 1972) suggested using the observed drawdown in the shallow and deep observation wells (Md24-2 and 4) to determine the

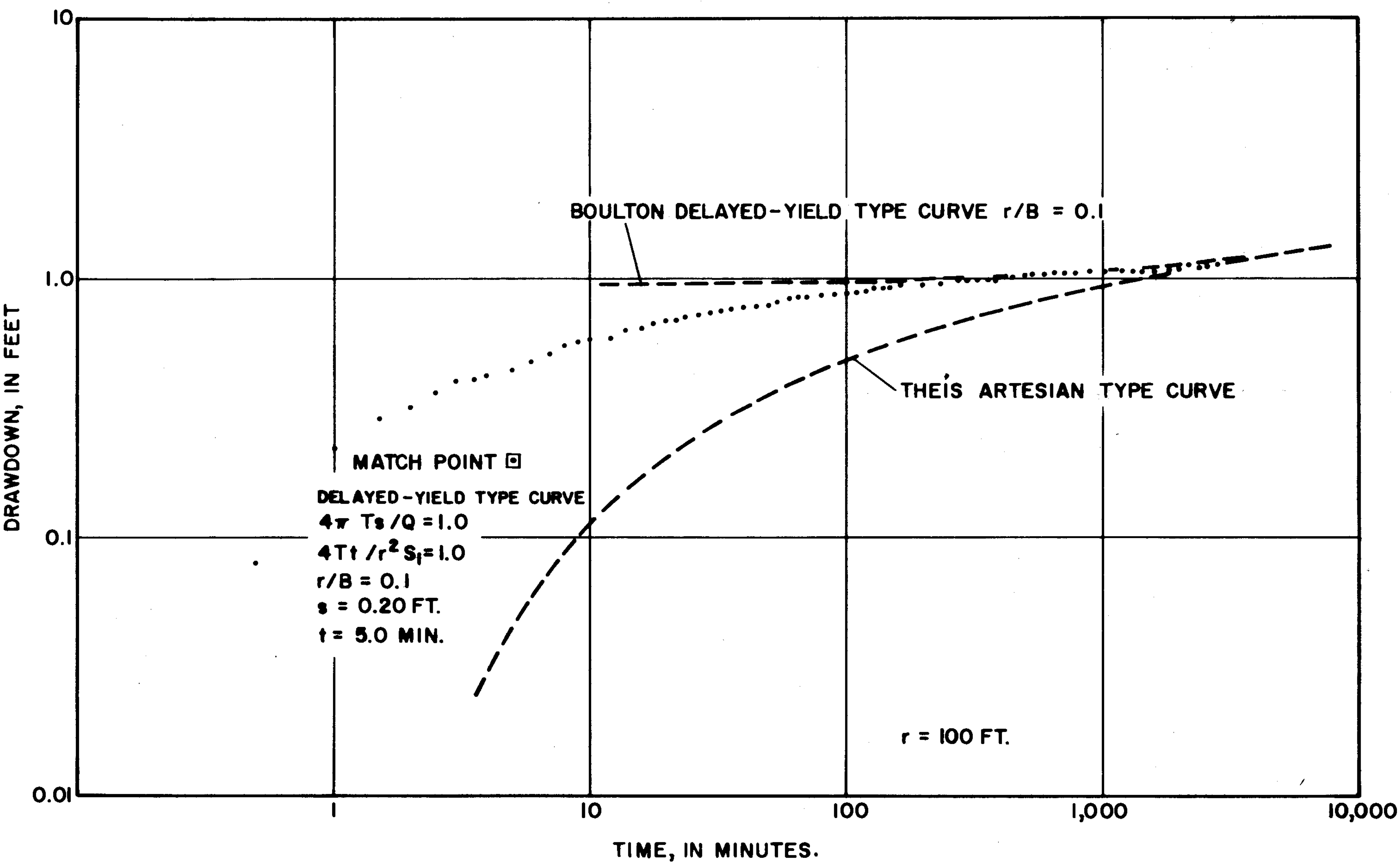

Figure 6. Logarithmic plot of drawdown against time for observation well Md24-2 (near base of aquifer) at Houston aquifer test, May 17-19, 1971.

distance at which the observed drawdowns would have occurred if the aquifer were isotropic. From this distance the ratio of horizontal to vertical conductivity may be calculated. However, due to a poorly defined prepumping water-level trend, accurate drawdowns could not be calculated in the shallow well (Md24-4), and the method could not be used.

Weeks (1969, Table 1) has listed drawdown correction factors for partially penetrating observation wells of different depths near a pumping well that penetrates various percentages of aquifer thickness. For a mid-aquifer observation well at a distance of lb from a pumped well that taps the lower 10 percent of an aquifer, the drawdown correction is almost zero. This situation describes the relation of observation well Md24-1 to the pumping well (Figure 4). Thus, the value of T = 22,000 ft^2/day as calculated at well Md24-1, is probably the most realistic value for transmissivity of the aquifer at the Houston site. Based on a saturated thickness of 86 feet, the horizontal conductivity (K_r) is about 250 ft/day -- similar to that expected for a "clean" coarse sand or coarse sand with gravel lenses (Lohman, 1972, Table 17).

Middletown aquifer test

The Middletown aquifer test site is about half a mile east of Middletown, Del., as shown on Figure 1. The test is described in detail in a report on the geology and hydrology of the Columbia sediments in the Middletown-Odessa area (Spoljaric and Woodruff,(1970, p. 96-100). At the pumped well, Pleistocene sand and interbedded gravel extend to a depth of 76 feet. The saturated section (34 to 76 feet) is reportedly coarse to medium sand. Underlying the Pleistocene deposits are "greensands" of the Rancocas Formation (Paleocene and Eocene Age), which probably act as a leaky confining bed. The test was made by Kenneth Woodruff and other members of the DGS with participation by the writer.

The pumped well (Fb34-16) is screened in the lower half of the aquifer (54-74 ft). Two observation wells (Fb34-17 and 18), both screened near the base of the aquifer (64-74 ft), are located, respectively, 26 and 103 feet from the pumped well.

Well Fb34-16 was pumped for 24 hours, and recovery measurements were made for 22 hours. Discharge was maintained constant at 60 gpm using the orifice method of measurement.

The measured drawdown in the near observation well (Fb34-17) was 1.5 feet at the end of pumping. As the saturated thickness (b) was 42 feet, no correction was applied for dewatering of the aquifer. The water level in the distant observation well (Fb34-18) responded sluggishly to pumping and the drawdown data could not be analyzed.

Figure 7 shows a logarithmic plot of drawdown against time for observation well Fb34-17 (r = 26.5 feet). Stallman (1965) has prepared type curves for analyzing drawdowns observed in partially penetrating observation wells near a pumped well penetrating the lower three-tenths of an aquifer. At the Middletown site, the pumped well is screened in the lower half of the aquifer and observation well Fb34-17 is screened near the base of the aquifer -- approximating the conditions of one of Stallman's models. As can be seen in Figure 7, the time-drawdown plot for observation well Fb34-17 exactly matches Stallman's type curve for $\psi = 0.327$ (Plate 7E in Lohman, 1972) from about 15 minutes to 1,440 minutes (end of the test). Using the match point and associated parameters for Stallman's type curve, transmissivity is calculated as follows:

$$T = 1.0 \frac{Q}{s}$$

$$T = \frac{(1.0)\ (60 \text{ gpm})\ (1{,}440 \text{ min/day})}{(2.6 \text{ ft})\ (7.48 \text{ gal/ft}^3)}$$

$$T = 4{,}500 \text{ ft}^2\text{/day} \quad \text{or} \quad 33{,}000 \text{ gpd/ft}$$

Using the same match point, storage coefficient is:

$$S = \frac{Tt}{1.0\ r^2} = \frac{(4{,}500 \text{ ft}^2\text{/day})\ (10 \text{ min})}{(702 \text{ ft}^2)\ (1{,}440 \text{ min/day})}$$

$$S = 0.05 \text{ (rounded)}$$

Spoljaric and Woodruff (1970, Figure 18) calculated a transmissivity of 37,000 gpd by matching the later part of the time-drawdown plot to the Theis artesian type curve. This value is not significantly different from the T calculated with the Stallman type curve. On the other hand, Woodruff calculated a storage coefficient value of 0.006 which is less realistic for an unconfined aquifer than the S value noted above.

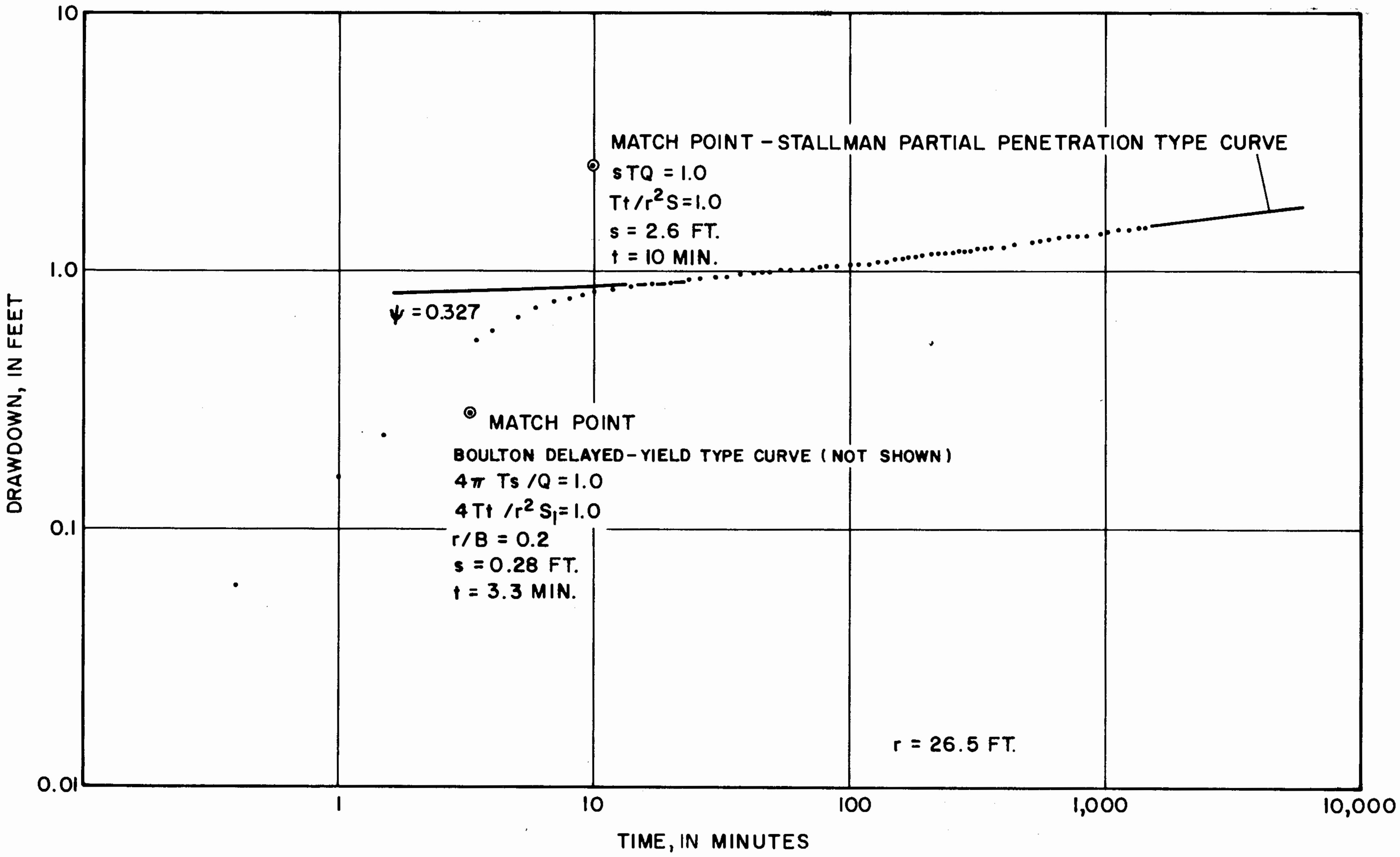

Figure 7. Logarithmic plot of drawdown against time for observation well Fb34-17 (base of aquifer) at Middletown aquifer test, May 7-8, 1970.

The time-drawdown plot was also matched against Boulton's delayed-yield type curve (lower match point in Figure 7). The resulting value of transmissivity was somewhat lower (25,000 gpd/ft) and S was about the same (0.04), as obtained from the Stallman type curve. Inasmuch as observation well Fb34-17 is only a short distance from the pumping well (about 0.6b), vertical flow components due to partial penetration and the effects of anisotropy in the aquifer are probably considerable. For this reason and because of the very close curve match, the values of T and S obtained in the Stallman type curve are considered to be the most valid aquifer coefficients at the Middletown site.

Based on a saturated thickness of 42 feet and transmissivity of 4,500 ft^2/day, the horizontal conductivity (K_r) is 107 ft/day. Such a value is typical of medium to coarse sand. The ratio of vertical to horizontal conductivity (K_z/K_r) may be calculated with the parameters obtained from the match point on Stallman's type curve (Figure 6) as follows:

$$\psi = \frac{r}{b}\sqrt{\frac{K_z}{K_r}} \quad \text{or} \quad \frac{K_z}{K_r} = \left[\frac{(\psi)\ \ (b)}{r}\right]^2$$

$$\frac{K_z}{K_r} = \left[\frac{(0.327)\ \ (42\ \text{ft})}{26.5\ \text{ft}}\right]^2 = 0.25$$

Thus, the horizontal conductivity (K_r) is four times as great as the vertical conductivity (K_z), which is therefore 27 ft/day. By comparison, at the Smyrna aquifer test site (Table 2), K_r is 10 times as great as K_z. At the Middletown site, the aquifer is basically a coarse to medium sand in the zone of saturation (see log in Spoljaric and Woodruff, 1970, p. 118). However, at the Smyrna site, the aquifer is a coarse sand of high conductivity (200 ft/day) which is overlain by beds of silty medium sand. Thus, the greater ratio of K_r to K_z at Smyrna seems realistic. Conductivity ratios are generally not available in published reports on the Coastal Plain sand aquifers. However, for comparison, Weeks (1969) reported conductivity ratios of 2 to 20 for five aquifer tests of glacial outwash in Wisconsin.

Milton aquifer test

The Milton aquifer test was made on the property of Mr. and Mrs. James H. Carpenter, about 3 miles southeast of Milton, Del., as shown in Figure 1. The pumping well taps one of the thickest known sand and gravel sections in

Delaware (see brief lithologic log in Figure 15). Furthermore, as will be discussed in a later section, the draining stream, Beaverdam Creek, has the highest base flow in Delaware. At the pumped well, Pleistocene fine to coarse sand extends to a depth of 50 feet and poorly sorted fine to coarse sand and gravel beds to a depth of 132 feet. A lower confining bed of clayey silt occurs from 132 to 150 feet; however, the Miocene Manokin aquifer (medium to coarse sand) occurs below 163 feet. At the time of the aquifer test, the water table was 22 feet below land surface and the saturated thickness of the Pleistocene deposits was, therefore, 110 feet. The pumped well (Ng55-4) is screened near the middle of the aquifer (from 60 to 70 feet). One observation well (Ng55-3), 98 feet away, is also screened near the middle of the aquifer (65 to 68 feet).

Well Ng55-4 was pumped for 48 hours and recovery measurements were made for 48 hours after the pumping period. A constant discharge of 198 gpm was maintained for 48 hours using the orifice method (variation less than 5 percent). The pumped water was discharged by irrigation pipe and a ditch to Beaverdam Creek, about 1,200 feet from the site.

The measured drawdown in observation well Ng55-3 was small (1.32 feet) compared to the saturated thickness (110 feet). Thus, no correction was needed to account for dewatering of the aquifer.

Figure 8 shows a logarithmic plot of drawdown against time for observation well Ng55-3. This plot can be closely matched with Boulton's delayed-yield type curve $r/B = .2$ (Plate 8 in Lohman, 1972). As shown on Figure 8, the time-drawdown plot follows the delayed-yield type curve from 10 to 1,000 minutes, and the later points closely follow the Theis artesian type curve. Using the match point and associated parameters given in Figure 8, transmissivity (T) was calculated to be 14,000 ft^2/day (or 104,000 gpd/ft). Based on a saturated thickness of 110 feet, horizontal conductivity (K_r) is about 130 ft/day. This value is representative of coarse to medium sand (Lohman, 1972, Table 17) or "dirty" gravel.

The storage coefficient (S) was calculated to be 0.02, which is rather low for an unconfined aquifer. However, the fact that the late time-drawdown data closely match the Theis type curve suggests that delayed drainage was nearly complete and that the calculated S is valid.

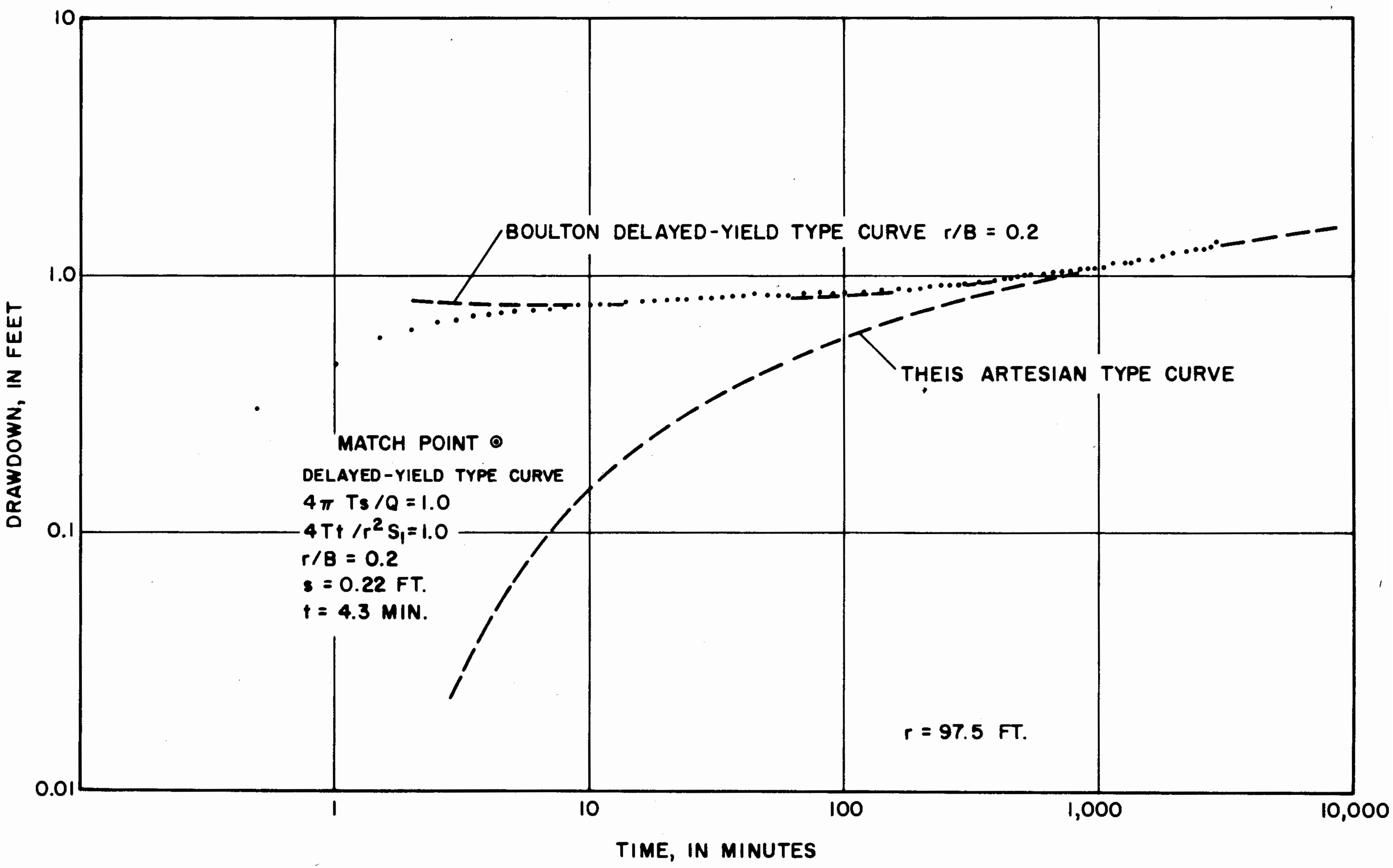

Figure 8. Logarithmic plot of drawdown against time for mid-aquifer observation well (Ng55-3) at Milton aquifer test, Dec. 8-10, 1970.

Aquifer Coefficients Estimated by Reconnaissance Methods

Aquifer coefficients (transmissivity: T; storage coefficient: S; and hydraulic diffusivity: T/S) were estimated by several reconnaissance methods. Transmissivity (T) was estimated from the specific capacity of wells based on drillers' reports. Transmissivity values also were estimated from well logs by estimating values of hydraulic conductivity (K) for the materials penetrated. Storage coefficient (S) was estimated by a hydrologic-budget technique involving ground-water level changes as compared to runoff and precipitation. Hydraulic diffusivity (T/S) was estimated from ground-water level fluctuations using methods presented by Rorabaugh (1960, 1966) and Stallman and Papadopulos (1966).

The advantages and disadvantages of these methods and their applicability to the Delaware Coastal Plain will be discussed in a report, as yet unpublished, by the writer. The methods were used in the present study primarily to prepare a statewide map of the transmissivity of the Columbia deposits. The applicability of these methods for this purpose is discussed briefly here.

Most of the values of transmissivity used to prepare the transmissivity map (Figure 9) were obtained from specific-capacity data. The specific-capacity data are based on reported drawdowns in wells after short periods of pumping (generally 2 to 8 hours). Specific capacity was converted to transmissivity by equations and a chart presented by Theis (1963) for water-table aquifers. Conversion factors used to compute T from specific capacity ranged from 1,400 to 1,800 depending upon the well diameter and period of pumping. For example, specific capacity expressed in gallons per minute per foot of drawdown, times 1,500 equals transmissivity expressed in gallons per day per foot of aquifer; gpd/ft. Transmissivity values obtained by this method, and by aquifer test analysis, are listed in Table 7. Transmissivities calculated from specific capacity must be considered rough estimates. In particular, the length of well screen and well efficiency were not considered in these calculations. However, more than half the T values listed in Table 7, and shown in Figure 9, were obtained from fully penetrating large-diameter wells, and the efficiency of such wells would be expected to be high initially. Where well efficiency is much less than 100 percent and the well screen penetrates a small part of the aquifer, the transmissivity values calculated from specific-capacity data will be erroneously low. However, no wells constructed with very short screens or wells with casing diameters less than 8 inches were used to compute transmissivity. Specific capacity is discussed further in the later section on the availability of water, and a frequency graph (Figure 13) of specific capacity for wells tapping the Columbia deposits is presented.

Transmissivity has been estimated from geologic logs of several test holes. In using this method, an average hydraulic conductivity (K) is assigned to each lithologic interval penetrated by the hole. The average K is multiplied by the thickness of the interval and the sum of these values provides an estimated T for the well. The values of K used for this method are average values obtained from the aquifer tests listed in Table 2 as follows:

coarse sand with gravel beds. . .	250	ft/day
silty ("dirty") gravel.	125 - 150	ft/day
coarse sand	200	ft/day
coarse to medium sand	100	ft/day
fine to coarse sand	50 - 75	ft/day
medium sand	50	ft/day
fine sand	10 - 20	ft/day

These values of K are generally in agreement with K values listed in Lohman (1972, Table 17). Estimating T by this method is subject to considerable error because the method involves the geologist's judgment of average effective grain size. For this reason, only a few values of T estimated from well logs were used to prepare the statewide transmissivity map (Figure 9).

The values of transmissivity obtained from pumping-test data, specific-capacity data, and estimates from geologic logs are local values applying to a small segment of the aquifer. On the other hand, values of T obtained by methods that consider discharge from the aquifer to draining streams, as measured by water-level declines in wells, are average values for a larger sample of the aquifer. The areal values of T are, of course, highly useful in preparing a statewide transmissivity map.

Rorabaugh (1960) has shown that ground-water levels will decline exponentially with time (after a critical time period elapses), according to the hydraulic diffusivity (T/S) of the aquifer and the square of the distance (a^2) from the draining stream to the ground-water divide. Values of T/S obtained by Rorabaugh's method are listed for several wells in Table 3. The applicability of Rorabaugh's method to the Coastal Plain of Delaware and a graphical example of the solution for T/S will be presented in a companion report by the writer, as yet unpublished. In the case of well Md24-1, listed in Table 3, there is close agreement between the T/S value (175,000 ft^2/day) obtained with Rorabaugh's method and the T/S value (170,000 ft^2/day) obtained using the transmissivity from the pumping test and long-term specific yield (S = 0.13) listed in Table 3.

Stallman and Papadopulos (1966) have presented a method for determining T/S from the decline in water level at a well tapping a wedge-shaped aquifer drained by two streams. Values of T/S for two wells obtained with this method are listed in Table 3. This method should be highly useful for evaluating the Pleistocene water-table aquifer. However, the method requires a continuous record of water-level fluctuations in a well; whereas observation well data available in Delaware are generally limited to monthly measurements.

Storage coefficient was estimated at several wells by means of simplified water-budget approach. The storage coefficient (S) is considered to be equivalent to the long-term specific yield. Specific yield is generally defined as the change in the amount of water in storage per unit area of aquifer that occurs as a result of a unit change in head. However, the change in storage is gradual after a change in head in a water-table aquifer. This delayed drainage causes specific yield calculated after a day or two of drainage to be less than after several weeks or months drainage. For this reason, the S values computed from short aquifer tests (Table 2) are not representative of the specific yield effective during long periods of aquifer drainage (particularly the summer-fall period). The long-term specific yield was calculated for periods of soil-moisture surplus (winter periods), based on the assumption that all precipitation during such periods either entered the aquifer as recharge or left the area as stream runoff. Evapotranspiration is thus considered to be negligible. Actually some evapotranspiration does occur during the winter months but the rates are less than 0.25 inch per month during December, January, and February (Mather, 1969).

The simplified equation for computing specific yield is as follows:

$$\text{Specific yield} = \frac{\text{Precipitation - runoff}}{\text{Net change of water table}}$$

Runoff includes both direct runoff and ground-water runoff. In applying the equation, periods of 5 to 10 days following a heavy rain were selected. The values of specific yield calculated from the equation assume that delayed yield is complete at the end of the period. Also it assumed that the net water-level change measured in an observation well is indicating the average net change in aquifer storage in the

TABLE 3 - Aquifer Coefficients Estimated by Reconnaissance Methods

Well No. and Location	Hydraulic Diffusivity (T/S) (ft^2/day)	Transmissivity (T)	Storage Coef-ficient (S)	Hydraulic Conductivity (K)	Remarks
Db24-10 (near Christiana, Del.)	2,700	-	-	-	Average T/S calculated by Rorabaugh's method using water-level recessions in 1964 and 1968. Thin saturated section (12 ft).
Hb14-1 (near Blackbird, Del.)	7,100	-	-	-	Average T/S calculated by Rorabaugh's method using water-level recessions in 1963, 1964, and 1968. Thin saturated section.
Md22-1 (near Williams-ville, Del.)	65,000	7,100 ft^2/day 54,000 gpd/day	0.11	95 ft/day 700 gpd/ft^2	Average T/S calculated by Rorabaugh's method using water-level recessions in 1963, 1964, 1965, and 1968. S calculated by hydrologic-budget method, using 1967-69 water-level data.
Md24-1 near Houston, Del.)	150,000 175,000	19,000 ft^2/day 145,000 gpd/ft 21,000 ft^2/day 160,000 gpd/ft	0.13	Average K = 230 ft/day 1,700 gpd/ft^2	T/S=150,000 ft^2/day by Stallman-Papadopulos method using 1970 water-level data. T/S=175,000 ft^2/day by Rorabaugh's method using 1970 water-level data. S=0.13 using hydrologic-budget method
Ng11-1 (near Milton, Del.)	24,000	3,400 ft^2/day 25,000 gpd/ft	0.14	70 ft/day 500 gpd/ft^2	Average T/S calculated by Rorabaugh's method using water-level recessions in 1963, 1964, 1965, and 1968. S calculated by hydrologic-budget method using 1969 water-level data.
Nc45-1 (near Greenwood, Del.)	4,500	780 ft^2/day 5,800 gpd/ft	0.17	10? ft/day	Average T/S calculated by Rorabaugh's method using water-level recessions in 1964 and 1968. S calculated by hydro-logic-budget method using 1963-64 and 1966-67 water-level data. Anomalously low "T" con-firmed by driller's report.
Pf24-2 (near Stockley, Del.)	32,000	4,800 ft^2/day 36,000 gpd/ft	0.15	64 ft/day 540 gpd/ft^2	T/S calculated by Stallman-Papadopulos method using 1970 water-level data. S calculated by hydrologic-budget method using 1970 water-level data.

area. The values of S obtained by this method range from 0.11 to 0.17 and average 0.14. Values of S calculated by this method as well as values of T/S and T calculated with the Rorabaugh and Stallman-Papadopulos methods are listed in Table 3.

Areal Distribution of Transmissivity and Hydraulic Conductivity

Figure 9 shows the distribution of transmissivity (T) in the Columbia deposits throughout Delaware. The map is based primarily on transmissivity values estimated from the specific capacities of large-diameter wells. Accurate control points were provided by the T values obtained from the aquifer tests and the T values obtained by analyses of water-level recessions (using the methods of Rorabaugh, 1960 and Stallman-Papadopulos, 1966). A few transmissivity values estimated from geologic logs are shown on the map. However, because these estimates were highly subjective, only a few such T values are shown.

Because of the spotty nature of the well data, a relatively large interval of 5,000 ft^2/day is shown for the lines of equal transmissivity in Figure 9. It is readily apparent from the map that large changes in T occur in relatively short distances. These changes in T reflect changes in lithology (from fine to medium sand to coarse sand and gravel, for example) as well as changes in saturated thickness. Comparison of the transmissivity map with the saturated aquifer thickness map (Figure 3) indicates a general increase in transmissivity and saturated thickness from north to south across Delaware.

Six areas have been outlined on Figure 9 where the transmissivity is known to exceed 10,000 ft^2/day (75,000 gpd/ft). Within these areas, transmissivities are as much as 170,000 gpd/ft (22,000 ft^2/day). The highest transmissivities do not necessarily correspond to the sites of thickest saturated section.

Four of the high transmissivity areas occur in the fluvial facies of the Pleistocene aquifer -- north of a presumed former shoreline, which Jordan (1964) postulated as extending along "a west-southwest axis running roughly from Lewes through Seaford." These high T areas are most likely underlain by stream channel deposits and the transmissivities indicate that the average effective grain size is about that of a coarse sand. Comparison of the structure contour map on the base of the Pleistocene (Figure 2), the saturated thickness map (Figure 3), and the transmissivity map (Figure 9) suggests that:

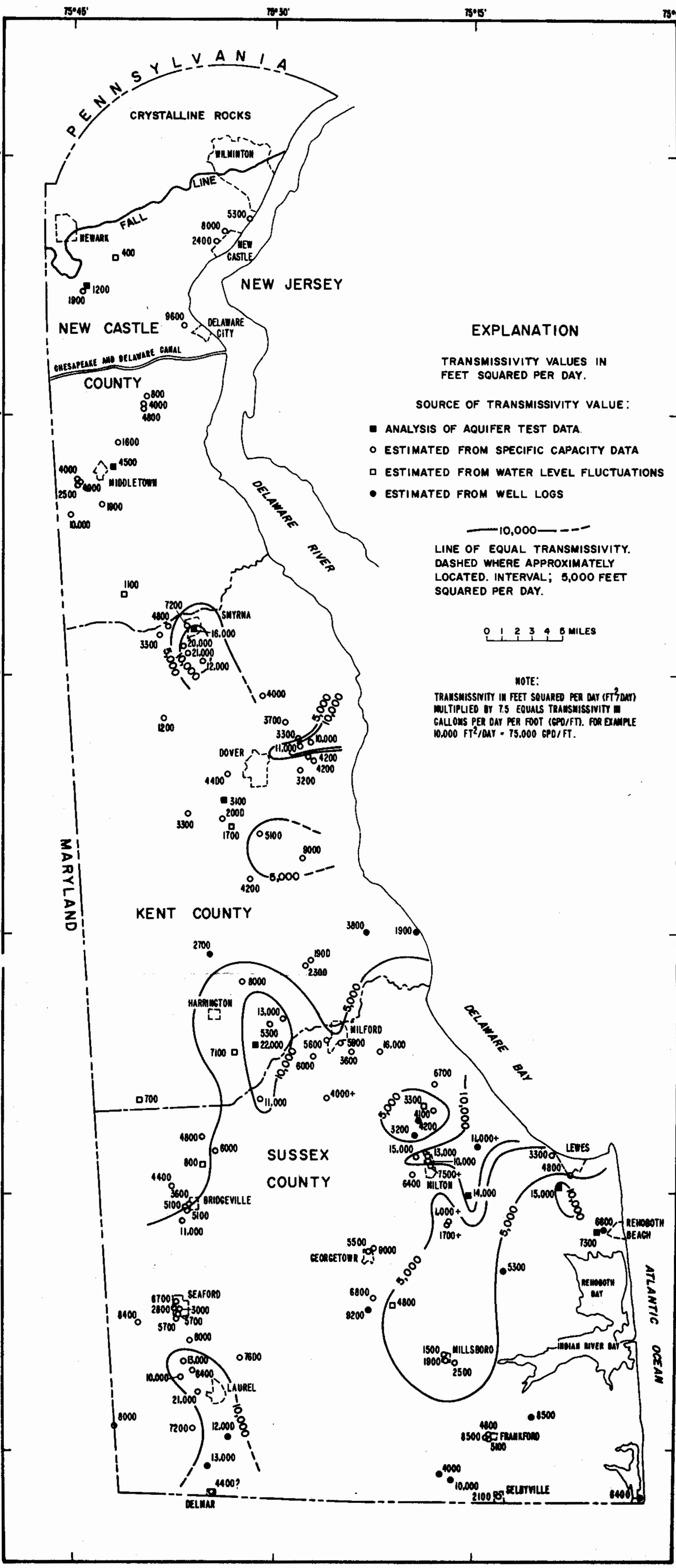

FIGURE 9. TRANSMISSIVITY MAP OF THE COLUMBIA (PLEISTOCENE) DEPOSITS IN DELAWARE.

(1) The high transmissivity area at Smyrna is part of a southeast trending channel which may turn eastward and include the high T area northeast of Dover.

(2) The high transmissivity area at Milton may include two channels which coalesce and head northeast along the present-day course of the Broadkill River.

(3) The high transmissivity area between Harrington and Milford may possibly be part of a southeast trending channel (or channels).

Two of the high transmissivity areas occur in the near-shore (beach, estuarine, neritic, and lagoonal) facies of the Columbia deposits. Correlation of the high T areas with the individual facies is rather tenuous. The high transmissivity area south of Laurel may be related to very thick Pleistocene channel fill deposits just south of the Delaware state line -- about 2 miles north of Salisbury, Maryland. As mapped by Hansen (1966), this so-called paleochannel contains sand and gravel attaining a maximum thickness of 220 feet. More recently, Weigle (1972) has mapped a tributary channel extending northward toward the Delaware line just east of Delmar. If extended, this channel would intersect the 183-foot value shown southeast of Laurel on the saturated thickness map (Figure 3). Actually all the area south of Seaford (including Laurel) to the Maryland line is characterized by a very thick saturated section and above average transmissivity. The area is probably a complex mixture of beach sand and stream channel deposits which can only be delineated by much exploratory drilling.

Some of the transmissivity values shown for southeastern Delaware (Figure 9) seem to be anomalously low in view of the fact that the saturated thickness in that area generally exceeds 75 feet. However, as mapped by Jordan (1962), a large part of southeastern Sussex County is underlain by fine sand and silt which Jordan termed the Omar Formation. These sand and silt beds could be expected to have hydraulic conductivities of less than 10 to 20 ft/day. Rasmussen and others (1960) applied the terms "Pamlico and Talbot Formations" and "Walston silt" to describe these fine sands and silts. Owens (oral communication, Oct. 1972) considers these fine-grained sediments as part of a "back barrier facies" which occur in a continuous narrow belt extending from southeastern Delaware to Virginia parallel to the coastline. In any case, below average values of transmissivity can be expected in parts of southeastern Delaware where these fine-grained deposits make up a sizable part of the saturated section of the Pleistocene deposits.

Laboratory determinations of hydraulic conductivity, as reported by Rasmussen and others (1960), indicate large variations in values. Samples collected from shallow pits had conductivities ranging from less than 1 to about 200 ft/day. Samples collected from well Qd21-2 at Laurel (in the high transmissivity area) had conductivities ranging from 125 to 500 ft/day.

The average transmissivity (T) of the Columbia deposits is about 7,000 ft^2/day (53,000 gpd/ft) in central and southern Delaware (based on T values shown in Figure 9). Within this area, the average saturated thickness of the aquifer is about 75 feet and, therefore, the average hydraulic conductivity (K) is about 90 ft/day. Such a K value is characteristic of medium to coarse sand. Jordan (1964) also concluded that the Columbia deposits are mainly medium sand, based on geologic studies. Therefore, both the geologic studies and hydraulic data support the conclusion that the Columbia deposits are effectively a medium to coarse sand aquifer.

In those parts of Delaware where the Columbia deposits are known to consist of medium or medium to coarse sand, a hydraulic conductivity (K) of 50 to 100 ft/day may be assumed. In areas where the aquifer is mostly "clean" coarse sand and interbedded gravel (as at the Houston and Smyrna aquifer test sites), K can be assumed to be 200-250 ft/day. In areas where the aquifer is predominantly coarse sand (as at Lewes) or "dirty" gravel (as at the Milton test site), the average K can be assumed to be 100-150 ft/day.

GROUND-WATER HYDROLOGY

The Columbia deposits are part of an interdependent stream-aquifer system in the Delaware Coastal Plain. The Columbia deposits comprise the uppermost and most permeable section of saturated sand in the water-table aquifer throughout most of Delaware. In fact, the water-table aquifer is composed essentially of Columbia (Pleistocene) deposits in about 75 percent of Delaware (1,500 square miles). The Pleistocene deposits receive most recharge reaching the water table and are the outlet for most ground-water discharge.

The small Coastal Plain streams are incised into the upper part of the Columbia deposits and derive much of their flow (50-90 percent) from ground-water runoff. Except for periods of overland runoff, these streams act as shallow drains from the aquifer. Precipitation and evapotranspiration rates determine the recharge rate to the aquifer and, in turn, changes in storage in the aquifer closely control base flow of the draining streams. The rate at which the aquifer discharges water to the draining streams is also influenced by the aquifer's hydraulic characteristics (Johnston, 1971).

Underlying the Columbia deposits of central and southern Delaware are Miocene deposits which contain several sand aquifers. In much of the area, the Miocene sands are separated from the Columbia deposits by confining beds of sandy clay. Locally there are small differences in hydraulic head between the Columbia and Miocene aquifers. However, the quantity of leakage is considered to be very small relative to the recharge and discharge to streams from the Columbia deposits.

The following section will consider changes in storage in the Columbia deposits, and by analysis of base flow data, present estimates of recharge to and discharge from the Columbia deposits.

Changes in Storage (Water-level Fluctuations)

The amount of water in storage in the Columbia deposits is constantly changing. The changes in storage are indicated by rising or declining water levels in wells. In general, water levels in the aquifer display a seasonal pattern of fluctuations. The period from mid-October to early April (nongrowing season) is typically a period of soil moisture surplus and ground-water recharge and water levels generally rise. The period from mid-April to mid-October (growing season) is, for the most part, characterized by a soil moisture deficit and declining water levels. During the growing season, ground-water recharge is infrequent. Ground-water storage gradually decreases as the aquifer discharges to draining streams and evapotranspiration becomes operative (where the water table is shallow).

Figure 10 shows water-level fluctuations in a shallow observation well (Md22-1) as compared to precipitation over a 10-year period. This pattern of fluctuations can be considered typical for the shallow water table in the Columbia deposits (where not affected by man). The low water levels during 1962-66 and 1968 reflect deficient precipitation during those years (sometimes referred to as the mid-1960's drought). The years 1967, 1969, and 1971 were characterized by above-normal precipitation and water levels are correspondingly high. The seasonal fluctuation is evident throughout the 10-year period, but is more sharply defined during the dry years. The highest water levels occur during either spring or summer in the wet years and the lowest levels occur

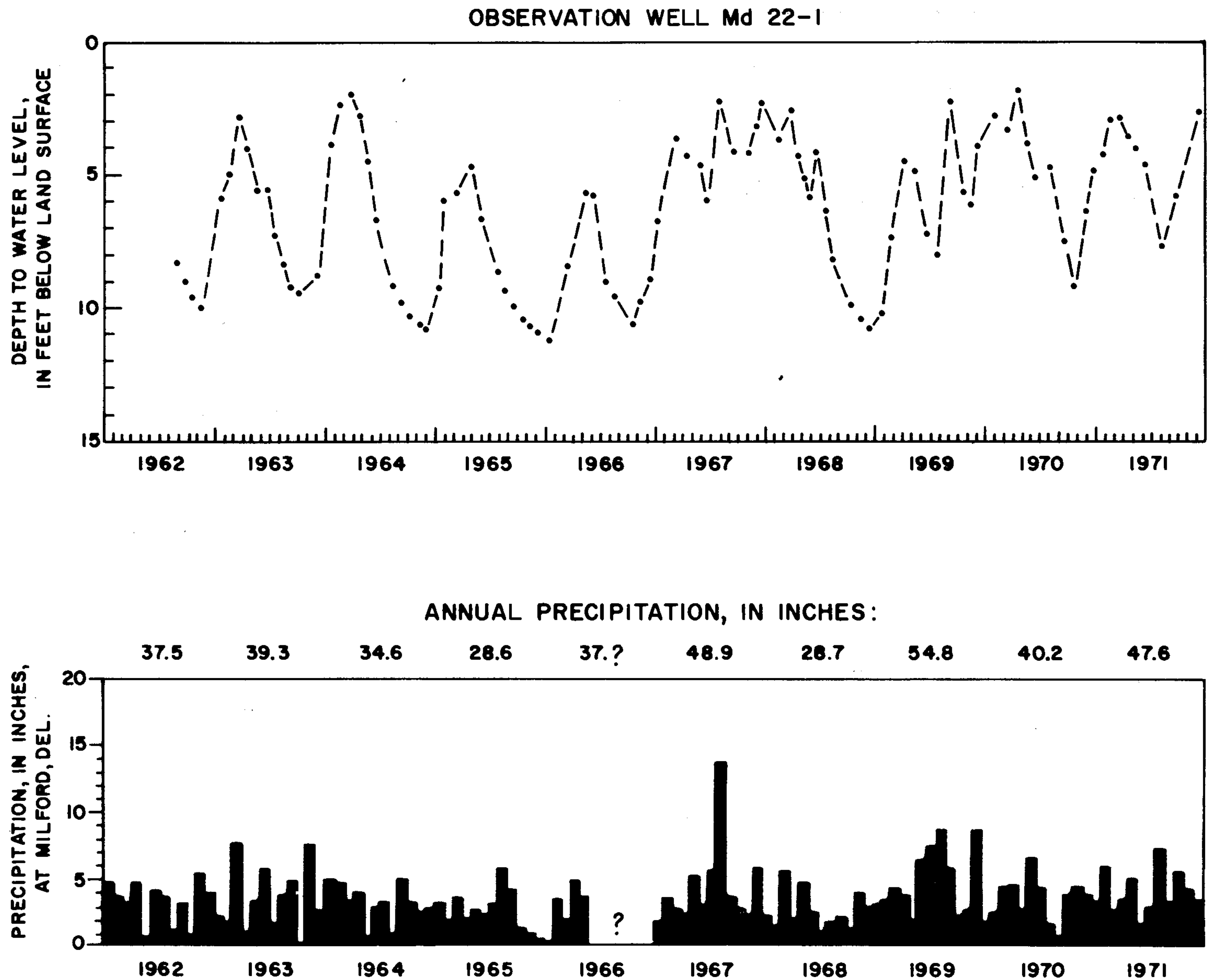

Figure 10. Hydrograph of water level in shallow observation well Md22-1 and precipitation at Milford, Del., 1962-1971.

in the fall of the dry years (1965 and 1968, for example). However, no long-term decline or rise in the water table is apparent.

The saturated thickness of the aquifer at well Md22-1 is 70 to 80 feet. Thus, the seasonal water-level fluctuations indicate a change of about 10 percent of the water in storage in the aquifer. In northernmost Kent County and most of New Castle County, where the saturated thickness is less than 25 feet, seasonal fluctuations of the water table change the amount of water in storage by a greater percentage. In eastern and southern Sussex County, where the saturated thickness exceeds 100 feet, seasonal water-level fluctuations change the amount of water in storage by approximately 5 percent.

Changes in storage in the aquifer, as measured by water levels in a specific well, are caused by several factors including:

(1) Location of the well with respect to the ground-water divide and the draining stream (wells close to the divide show larger water-level fluctuations and wells close to streams show smaller fluctuations).

(2) Hydraulic characteristics of the aquifer, particularly the hydraulic diffusivity (T/S).

(3) Thickness and character of the unsaturated zone (which influences the rate and timing of recharge).

(4) Relation of water-table altitude to artesian heads in underlying aquifers.

Figure 11 shows hydrographs of water levels in four observation wells, representing different combinations of the hydrologic factors cited above. The locations of these wells are shown in Figure 1.

The uppermost hydrograph (Md24-1) shows water-level fluctuations in an area characterized by a relatively shallow water table and a sand and gravel aquifer of very high transmissivity and diffusivity ($T = 22,000\ ft^2/day$ and $T/S = 180,000\ ft^2/day$). Infiltration is apparently rapid and, during the winter and spring, water levels begin rising within a few hours after the start of a heavy rain.

Well Md24-1 is relatively close to the draining stream (1,200 feet away), and part of the rise may be due to bank-

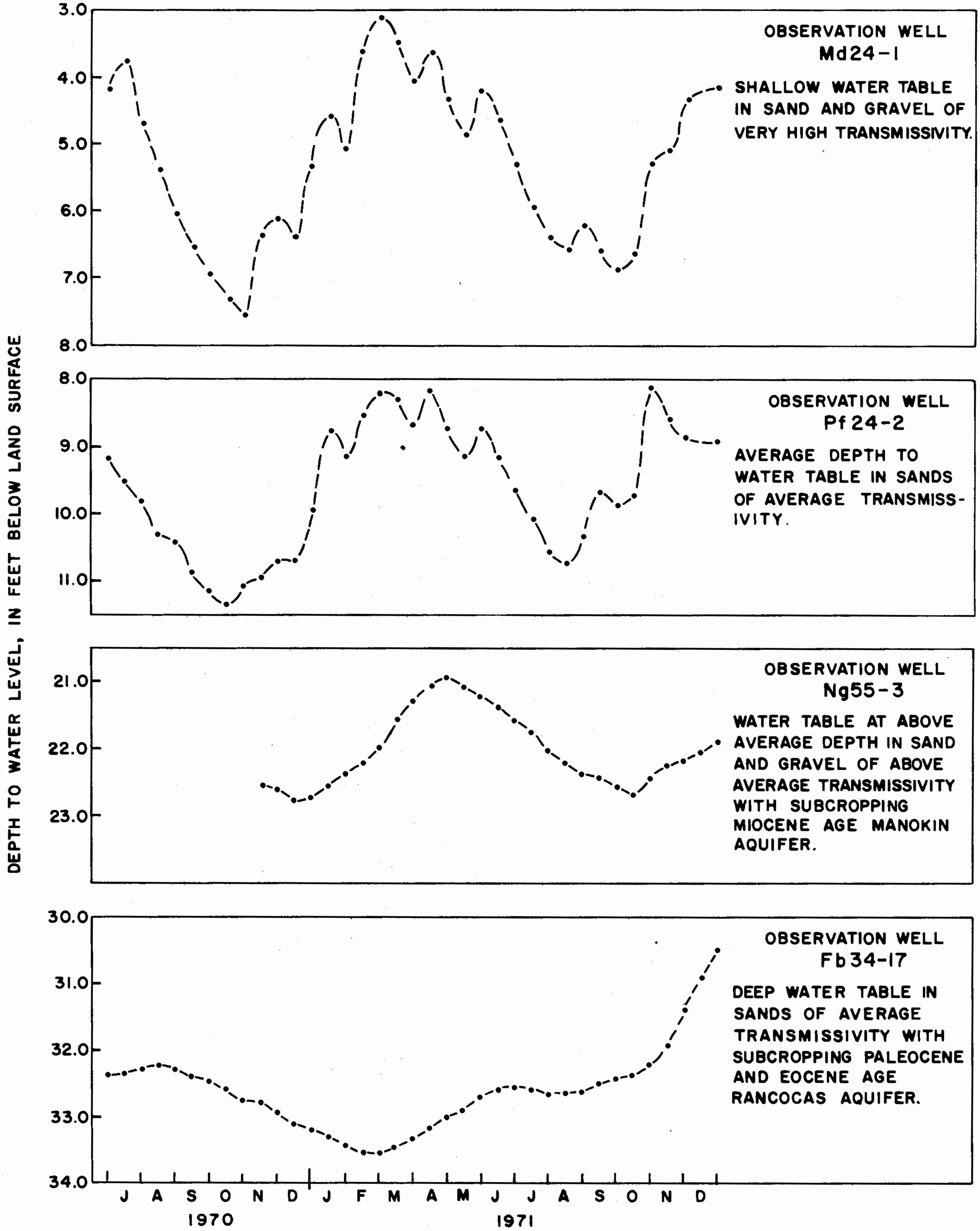

Figure 11. Hydrographs of water levels in four observation wells tapping the Columbia deposits under different hydrogeologic conditions.

storage effects. The bank-storage effects are complicated by the fact that the regional water-table frequently exhibits a rise nearly equivalent to that of the stream during heavy storms. As a result, there is rarely a reversal of the hydraulic gradient between the stream and the aquifer; in most instances there is only a reduction in this gradient. Thus, one can identify two conditions in connection with bank-storage effects: (1) the condition resulting from a small rise in stream stage, which temporarily reduces the hydraulic gradient to the stream and thereby decreases aquifer discharge to the stream -- the so-called "ground-water back-water effect," and (2) the condition where the stream stage rises above the stage of the water table, causing water to move from the stream into the aquifer. The writer made studies that involved piezometer tube measurements near several streams, and these measurements indicated that the "ground-water back-water effect" is apparently common in the Pleistocene deposits. However, the bank-storage condition in which the stream stage rises above the water-table stage is a rarity. At well Md24-1, a small part of the water-level rise after heavy rains is probably due to the "ground-water back-water" effect.

The hydrograph of well Pf24-2 shown in Figure 11 is typical of many wells in the Columbia deposits of central and southern Delaware. The well is near the ground-water divide but relatively close (1,500 feet) to the two draining streams. The well penetrates sand of average saturated thickness (75 ft) and average transmissivity (T = 4,800 ft^2/day and T/S = 32,000 ft^2/day). The mean depth to the water table is about twice that observed in well Md24-1; however, response to recharge is rapid in well Pf24-2 also. Generally water levels begin rising within a few hours after a heavy rain. At well Pf24-2, as in well Md24-1, the close proximity of the draining stream suggests that bank storage ("ground-water back-water") is a contributing factor during recharge after heavy rains.

The hydrograph of well Ng55-3 (third from the top in Figure 11) shows water-level fluctuations in an area underlain by a thick saturated section (102 ft) of "dirty" sand and gravel of moderatley high transmissivity (T = 14,000 ft^2/day). The water table is relatively deep (21-23 ft) compared with most of Delaware and the annual water-level fluctuation is small (about 2 ft). The water level does not respond to individual storms and the spring recharge peak occurs about 2 months later in well Ng55-3 than in wells Md24-1 and Pf24-2 (Figure 11). The Miocene Manokin aquifer underlies well Ng55-3 at depth and is separated from the Pleistocene deposits by an 18-foot silt bed which probably acts as a leaky confining layer. The artesian head in the

Manokin is unknown at well Ng55-3, but the head is probably higher than the water-table altitude (8-10 ft above mean sea level). Probably there is substantial leakage from the Manokin (which obtains recharge some distance away) into the Pleistocene sand, thus explaining the 2-month delay in recharge response. Furthermore, the water level in well Ng55-3 displays barometric fluctuations (more typical of an artesian aquifer than the unconfined Pleistocene sand), suggesting hydraulic connection with the Manokin aquifer.

The lower hydrograph in Figure 11 shows water-level fluctuations in well Fb34-17 which taps a 42-foot thick saturated section of Pleistocene sand of average transmissivity ($T = 4,500$ ft^2/day). The water table at the well is exceptionally deep (30-33 ft) for the Delaware Coastal Plain. Directly underlying the Pleistocene sands are "greensands" of the Paleocene and Eocene Rancocas aquifer and the two aquifers are probably hydraulically connected. As shown in Figure 11, the normal October water-level rise (as seen in wells Md24-1 and Pf24-2) did not occur in well Fb34-17 until March, 1971. Furthermore, during the wet year of 1971, the normal summer water-level decline is almost nonexistent, and the water level shows a generally rising trend throughout the year. The combination of a thick unsaturated zone (which delays recharge) and hydraulic connection with the Rancocas aquifer (which may be receiving recharge at some distance) is probably responsible for the water-level trend seen in Figure 11.

Well Fb34-17 does not show the normal seasonal water-level fluctuations but, rather, is responding to long-term changes in precipitation. The year 1970 was characterized by below-normal precipitation (34 inches at Middletown), whereas 1971 was characterized by exceptionally high precipitation (56 inches), and the changes in ground-water storage observed in well Fb34-17 simply reflect this year-to-year difference.

Ground-Water Discharge

Ground-water discharges continuously from the Columbia deposits to the small streams draining the Delaware Coastal Plain as well as directly to the Delaware Bay and Atlantic Ocean. A small amount of ground water also evaporates and transpires. That part of the streamflow supplied by ground-water discharge is referred to as ground-water runoff. During periods when a stream is supplied largely by ground-water discharge, it is said to be at base flow. Actually, the small nontidal streams in the Delaware Coastal Plain are supplied entirely by ground-water seepage during base-flow conditions (Johnston, 1971). After a rainless period of a few days, base-flow conditions prevail and the total discharge of a stream represents ground-water runoff. However, ground-water runoff persists during periods of heavy rainfall and may, in fact, be higher than during later periods of base flow (Figure 12).

At streams where continuous gaging stations are operated, it has been possible to determine ground-water discharge by separation of the streamflow hydrographs into the components of ground-water runoff and overland runoff. Determination of ground-water runoff from streamflow hydrographs assumes that the gaging station is measuring the total runoff from the basin. In most cases this is true. However, discrepancies may occur between the measured runoff and the actual runoff because of either man-made effects such as withdrawal of water and operation of dams, or natural effects such as movement of ground water into or out of deep artesian aquifers. By selecting basins where the man-made effects are known to be minimal and where inter-aquifer movement of water is negligible, a fairly accurate estimate of ground-water runoff can be made by hydrograph separation. Assuming that the rate of ground-water discharge is the same in the gaged areas as in the ungaged areas (including tidal streams and areas discharging directly to the Delaware Bay and the ocean), it is possible to estimate the total ground-water runoff from the water-table aquifer.

As mentioned in an earlier section, the water-table aquifer consists largely or entirely of the Columbia deposits throughout an area of about 1,500 square miles in Delaware. The ground-water runoff of the streams draining this 1,500 square mile area, therefore, represents the ground-water discharge minus evapotranspiration from the Columbia deposits in Delaware.

The streamflow hydrographs were separated into the ground-water and overland-runoff components by using a master base-flow recession curve as described by Riggs (1963). Preparation of the base-flow recession curves and the assumptions upon which they are based, are discussed in an as yet unpublished report by the writer.

Figure 12 shows an example of hydrograph separation for Stockley Branch -- a small stream in southern Delaware (Figure 1). In making the separation, the master recession curve is superimposed on the streamflow hydrograph at the best fit; all streamflow above the curve is considered overland runoff, whereas all flow below the curve is ground-water runoff. The segments of the master recession curve used to

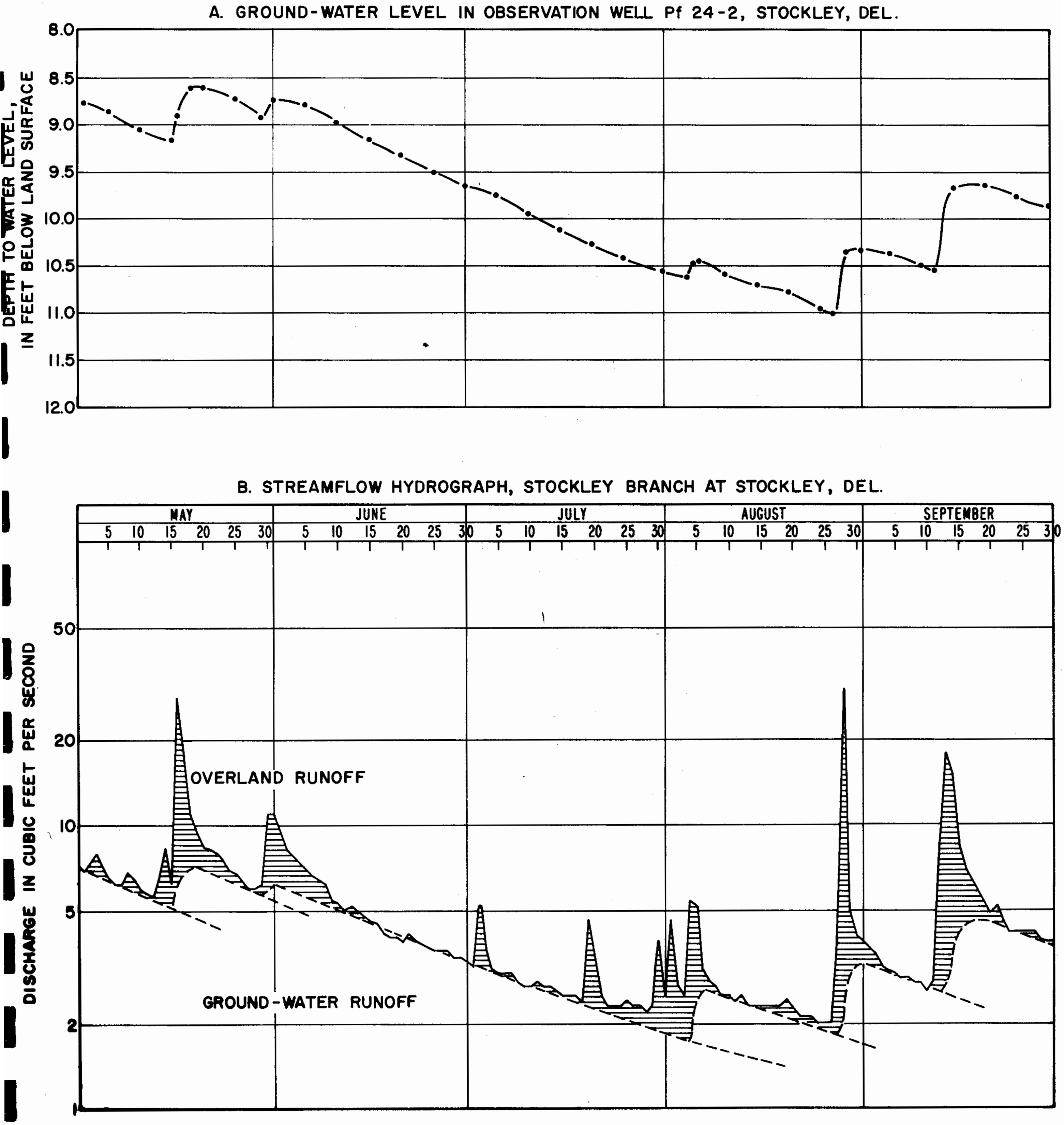

FIGURE 12. GROUND-WATER RUNOFF AND OVERLAND RUNOFF AT STOCKLEY BRANCH COMPARED TO GROUND-WATER LEVELS, MAY-SEPTEMBER 1971.

make the separation are shown by dashed lines in the streamflow hydrograph in Figure 12(B). The master curve is moved sideways on the streamflow hydrograph after each ground-water recharge event. These events are indicated by periods of rising ground-water levels, as shown in the observation well hydrograph in the upper part of Figure 12. The peak ground-water stage during such recharge events is assumed to coincide with a ground-water-runoff peak. The observation well in this case is about 1,500 ft from Stockley Branch and is a good indicator of changes in ground-water storage.

As discussed earlier, rises in ground-water levels may result not only from recharge but also from bank storage (the so-called "ground-water back-water effect"). However, bank-storage effects are probably minimal at well Pf24-2, which is about 1,500 ft from the stream. This is indicated by the fact that during some of the overland-runoff peaks seen in Figure 12 (July 2 and 19, 1971, for example), there is no corresponding rise in ground-water level.

During periods of heavy overland runoff and frequent recharge to the water-table aquifer (particularly during the winter and spring), rating curves of base flow against ground-water stage were used to estimate ground-water runoff. Such curves are presented and discussed in another report by the writer, as yet unpublished, on ground water-surface water relations. The relationship between ground-water stage and base flow is variable, however, being dependent upon the current evapotranspiration rate and time since last recharge (see papers by Rorabaugh, 1960 and 1966). For example, as shown in Figure 12, on July 5, a ground-water level of 9.75 feet was equivalent to a base flow of 3 cfs (cubic feet per second), whereas on September 25 a similar stage of 9.75 feet was equivalent to a base flow of about 4 cfs. Presumably the higher evapotranspiration rate in July is responsible for the lower base flow occurring at the same ground-water stage as the higher base flow in September. Also, the base flow on September 25 closely followed a time of recharge and ground-water levels were declining at a faster rate than on July 5, a month after the last recharge (see Rorabaugh, 1960 and 1966). Both the master recession curves and ground-water stage base flow rating curves are approximations. However, the estimates of ground-water runoff obtained by use of these curves are probably accurate within 10-15 percent.

Ground-water runoff was determined for six streams draining the Pleistocene aquifer by the same method described for Stockley Branch. The drainage areas, average discharges, and ground-water runoff of these six streams are listed in Table 4. The 3-year period (1968-70) was selected for determinations of ground-water discharge because precipitation was

TABLE 4 - Streamflow and Ground-Water Runoff from Six Basins in Central and Southern Delaware

Stream and location of gaging station	Drainage Areas		Streamflow		Ground-Water Runoff (1968-70)	
	Surface drainage area (mi^2)	Ground-water basin area (mi^2)	Average discharge (long term) (cfs)1/	Average discharge (1968-70) (cfs)	Average ground-water runoff	Average ground-water runoff (winter period)
Beaverdam Branch at Houston, Del.	2.82	2.85	3.49 (12 years)	3.72	2.91 cfs (1.02 cfsm)2/	3.50 cfs (1.23 cfsm)
Murderkill River near Felton, Del.	13.6	15.6	17.4 (12 years)	20.3	10.7 cfs (0.69 cfsm)	13.7 cfs (0.88 cfsm)
Nanticoke River near Bridgeville, Del.	75.4	72.5	90.6 (27 years)	91.6	76 cfs (1.05 cfsm)	86 cfs (1.19 cfsm)
St. Jones River at Dover, Del.	31.9	31.5	29.8 (12 years)	31.5	16.1 cfs (0.51 cfsm)	23.3 cfs (0.74 cfsm)
Sowbridge Branch near Milton, Del.	7.08	8.0	9.55 (14 years)	8.38	6.57 cfs (0.82 cfsm)	7.37 cfs (0.92 cfsm)
Stockley Branch at Stockley, Del.	5.24	5.6	6.89 (27 years)	6.09	4.84 cfs (0.86 cfsm)	5.90 cfs (1.05 cfsm)
Totals	136.0	136.0	157.7	161.6	117.0 cfs	139.8 cfs
Mean	-	-	-	1.19 cfsm	0.86 cfsm	1.03 cfsm

1/ Cubic feet per second

2/ Cubic feet per second per square mile near normal and streamflow was about average during the 3 years. As shown in Table 4, the combined average discharge from the six basins was 161.6 cfs during 1968-70 compared with 157.7 cfs for the period of record at the six gaging stations.

Observation well data were available in four of the six basins listed in Table 4, and it was possible to obtain a reasonably accurate determination of ground-water runoff from the streamflow hydrographs. However, for two of the six streams (St. Jones River and Murderkill River), observation well data were unavailable and the base flow values listed in Table 4 are estimates. During the long summer-fall base flow recession, these estimates are reasonable accurate. However, during periods of considerable overland runoff and frequent ground-water recharge (winter and spring), the ground-water runoff values for these two streams are only rough estimates.

Note that the drainage areas used to calculate ground-water runoff per square mile are the areas of the ground-water basin (as determined from water-table maps) and not the surface-drainage areas. However, although the surface- and ground-water drainage areas of individual basins may vary considerably, the total surface- and ground-water drainage areas for the six streams are the same (136 square miles).

The average ground-water runoff of the six streams listed in Table 4 is 0.86 cfsm (cubic feet per second per square mile), or 550,000 gal/day/sq mi (gallons per day per square mile), as compared with the average discharge (1.19 cfsm). Ground-water discharge, therefore, constitutes about 72 percent of the total flow of these streams. Using the average ground-water runoff rate of 550,000 gal/day/sq mi, the total ground-water discharge from the Columbia deposits has been estimated. This estimate is based on the following assumptions:

(1) Ground-water runoff occurs at the same rate in the ungaged areas as in the six drainage basins listed in Table 6.

(2) Direct outflow to the Delaware Bay and Atlantic Ocean, which could not be determined directly, also occurs at the same rate as ground-water runoff in the six basins.

These assumptions are considered plausible because long-term rates of precipitation and evapotranspiration are similar throughout the 1,500 square miles where the Columbia deposits constitute the water-table aquifer. Also, the aquifer characteristics in the six basins are typical of the 1,500 square mile extent of the Columbia deposits in Delaware. Based on these assumptions, the average ground-water discharge from the Columbia deposits in Delaware is about 800,000,000 gal/day (or 800 mgd).

After reaching the water table and before discharge from the Columbia deposits, some water moves into and out of the underlying Tertiary artesian aquifers. Locally, the amount of this leakage may be considerable (see, for example, the discussion of Beaverdam Creek basin in the later section on the availability of large water supplies). However, such leakage does not change the estimate of discharge given above because the Columbia deposits are the final discharge outlet.

Ground-Water Recharge

Recharge to the Columbia deposits takes place primarily during the nongrowing season from about mid-October to early April. During this period, frequent pulses of recharge gradually raise ground-water levels as well as the base flow of streams. Throughout the growing season, from April to mid-October, recharge is less frequent. A soil-moisture deficit exists much of the time and ground-water levels gradually decline. Heavy rains resulting from thundershowers and tropical storms are occasionally sufficient to overcome the soil moisture deficiency and to provide surplus water which infiltrates to the water table. Figure 10, showing the 10-year hydrograph of ground-water levels in shallow observation well Md22-1, illustrates the typical pattern of recharge. Except for the wet summers of 1967 and 1969, nearly all recharge occurred during the winter.

Over the long term, there has been no change in the amount of water in storage in the Columbia deposits, as noted previously. Therefore, the long-term average recharge rate is equal to the discharge from the aquifer. Inland from the coast, the discharge is the sum of the ground-water runoff and ground-water evapotranspiration.

The average rate of recharge is difficult to determine at a given site. Some investigators have determined recharge by multiplying the average specific yield of an aquifer by the net water-table rise during a long period of time (see, for example, Rasmussen and Andreasen, 1959). However, this method requires a large number of observation wells located in areas characterized by different infiltration rates. Furthermore, specific yield is dependent upon time and other variables and estimates of the average effective specific yield are problematical. The effective specific yield during a period of recharge will also depend upon the moisture content above the water table at the time of recharge, which varies according to the time since the previous period of recharge. This time factor in the effective specific yield is influenced by the particle size and other variables. In addition, the rise in ground-water levels observed in a well may include bank-storage components as well as ground-water recharge.

Although recharge occurs primarily during the winter, there is little difference between the mean ground-water stage during the summer and winter. A hydrograph showing the average depth to water level in 13 water-table wells in Delaware has been presented by Boggess and others (1964) for the 12-year period, mid 1950 to July, 1962. During the summer (May-September), the mean depth to water level is slightly lower (0.46 foot) than during the winter (November-March). This small difference is probably caused by evapotranspiration directly from the aquifer. Shallow observation wells in the Beaverdam Branch, Nanticoke River, and Stockley Branch basins (listed in Table 4) are characterized by summer water levels averaging 0.2 to 0.4 foot lower than in the November-March period. However, a shallow well just outside Sowbridge Branch basin (also listed in Table 4) is characterized by a mean summer water level about 0.6 foot higher than the mean winter water level. The reason for the higher summer water level may be the relatively low hydraulic diffusivity (T/S) in the basin, which results in slow drainage of the Pleistocene deposits in the summer and perhaps is due also to upward leakage from the Miocene deposits during the summer. These explanations are conjectural and, in any case, the observation well near the Sowbridge basin is atypical of those tapping the Columbia deposits.

Inasmuch as the ground-water discharge varies directly with ground-water stage, the ground-water discharge during the summer (ground-water runoff plus ground-water evapotranspiration) is not substantially less than the ground-water discharge during the winter (all ground-water runoff). Because there has been no long-term change in ground-water storage, the long-term rates of ground-water discharge and recharge are equal. Based on the premise that the winter and summer rates of ground-water discharge are similar, the ground-water runoff during winter is approximately equal to the average year-round recharge rate. Note that a large part of the recharge during the winter is carried over in aquifer storage and later discharged during the summer. The small amount of summer recharge provides only part of the summer ground-water runoff and evapotranspiration.

The average rate of recharge in the 1,500 square mile area where the Columbia deposits constitute the water-table aquifer, is assumed to be the same as in the 136 square mile area drained by the six streams listed in Table 4. Within this 136 square mile area (where the difference between the mean summer and winter ground-water levels is less than 0.4 foot), the average rate of ground-water recharge is approximately equivalent to the ground-water runoff during the winter. The average recharge is, therefore, approximately 1 cfs per square mile or 650,000 gallons per day per square mile. The total recharge to the Columbia deposits throughout its 1,500 square mile extent in Delaware is, thus, about 1 billion gallons per day.

GROUND-WATER DEVELOPMENT

Yield and Specific Capacity of Wells

The yields of wells tapping the Columbia deposits vary widely. Reported yields range from a few gallons per minute for small diameter 2 to 4 inch shallow wells sufficient for domestic use to more than 1,000 gpm for some large diameter irrigation and public supply wells.

The specific capacity of a well is more meaningful than a simple statement of yield. Specific capacity is the yield per unit of drawdown and is generally expressed as gallons per minute per foot of drawdown (gpm/ft). Specific capacity depends upon well construction and the transmissivity of the aquifer. Construction features such as casing diameter, length of well screen or slotted casing, slot size, type of gravel pack, and effectiveness of well development have a marked effect on specific capacity. Drilling mud may temporarily clog the voids in the Columbia sand and the openings in the well screen or slotted casing. To obtain maximum specific capacity, drilling mud as well as silt and fine sand in the aquifer immediately outside the well is removed by well development. Pumping, surging, and back washing are effective methods of developing wells in the Columbia deposits. Note that, whereas transmissivity determines the maximum specific capacity that may be obtained at a given well site, the construction and development of the well determine the specific capacity actually obtained.

The yield, specific capacity, and construction features of selected large diameter wells tapping the Columbia (Pleistocene) deposits are listed in Table 7 and the well locations are shown in Figure 1. Table 7 indicates the short-term well yields and specific capacities which may be expected in many locations in Delaware assuming similar well construction.

The probable range of specific capacities of wells can be estimated from a frequency graph. Figure 13 is a frequency graph for most of the wells listed in Table 7. The graph shows specific capacity plotted against percentage of wells. The specific capacities were calculated from drillers' reports of yield and drawdown at the end of relatively short periods of pumping (generally 2 to 8 hours). All the wells considered have casings larger than 8 inches in diameter, and about 50 percent of the wells exceed 12 inches in diameter. Data from wells with very short well screens or short slotted casings were not considered in preparing Figure 13. It is assumed that the wells were properly developed before testing.

Figure 13 shows that specific capacities range from about 3 to 100 gpm/ft. The median specific capacity is 23 gpm/ft; that is 50 percent of the wells have specific capacities exceeding 23 gpm/ft. The median specific capacity is less than the mean; in fact, 63 percent of the wells have a lower specific capacity than the mean value of 28 gpm/ft. This distribution of values indicates that a small number of wells of high specific capacity are affecting the mean. Considering the areal distribution of transmissivity in the Columbia deposits, this is not surprising. The concept of a few troughs of high transmissivity surrounded by larger areas of more moderate transmissivity (as shown in Figure 9), indicates that a few wells of high specific capacity compared with many wells of low to moderate specific capacity could be expected.

Most of the wells used to prepare Figure 13 are located in central and southern Delaware. Figure 13 shows that the chances of obtaining a well with specific capacity of more than 10 gpm/ft are very good throughout the area. Actually 80 percent of the wells may be expected to have specific capacities in the range of 10 to 55 gpm/ft.

Current Withdrawals

Approximately 33 million gallons of water per day was pumped from the Columbia deposits in 1970 - the latest year for which pumpage figures are available. Table 5 shows the distribution of pumpage by county and use. The table indicates that water is used about equally for industrial, municipal, and rural-agricultural purposes. Pumpage is greatest in Sussex County where the Columbia deposits constitute the major aquifer. Pumpage is least in New Castle County where the Columbia deposits are confined to local channels and where surface sources and Cretaceous-Tertiary aquifers supply most of the water. The 33 mgd withdrawn from the Columbia deposits represented about half of total groundwater pumpage (64 mgd) in Delaware during 1970.

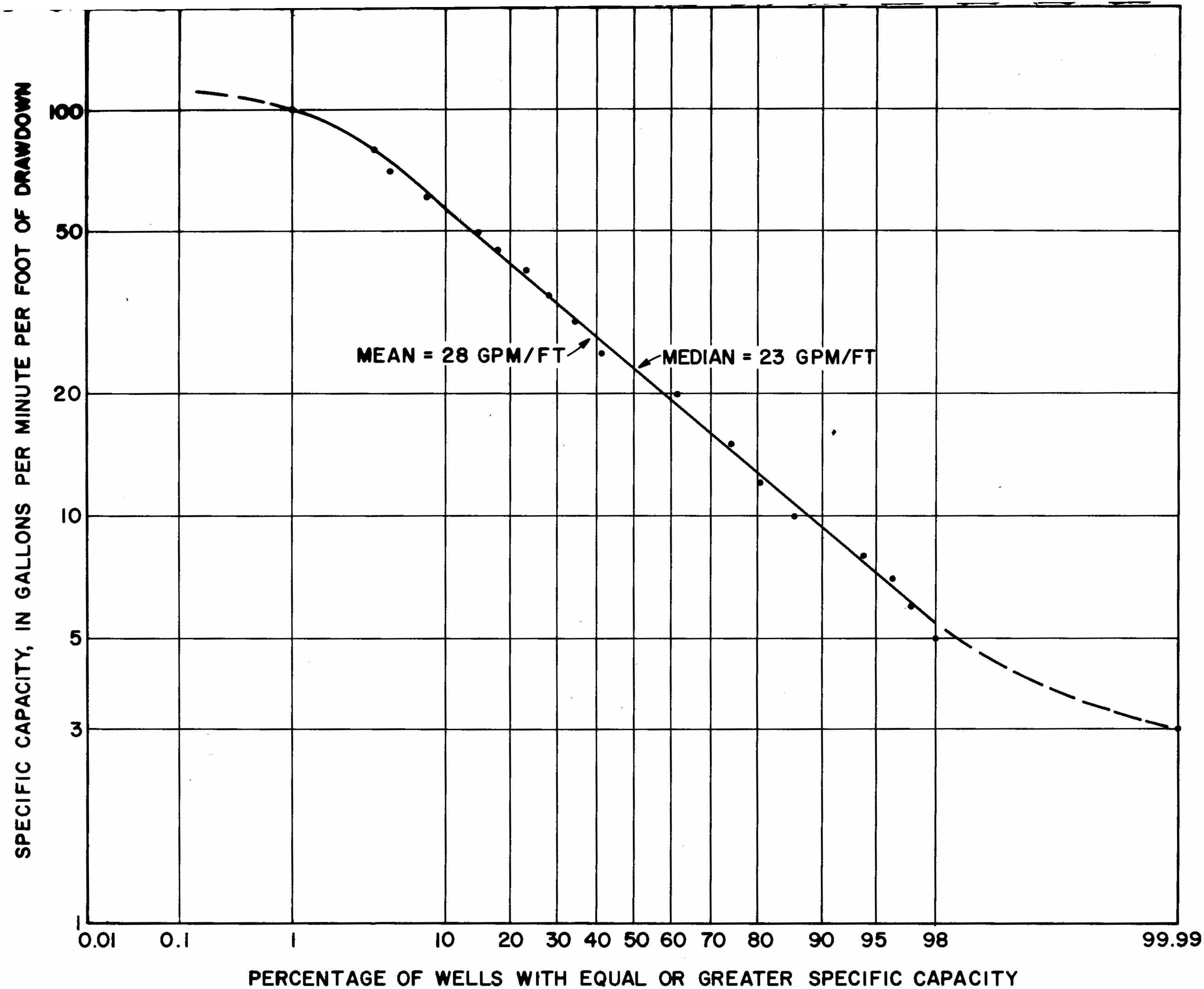

FIGURE 13. SPECIFIC-CAPACITY FREQUENCY GRAPH FOR LARGE-DIAMETER WELLS TAPPING THE COLUMBIA DEPOSITS.

TABLE 5 - Estimated Pumpage of Water from the Columbia Deposits in Delaware during 1970 (million gallons per day)

Use	New Castle County	Kent County	Sussex County	Total
Municipal Supply	5.0	0.9	4.5	10.4
Industrial	-	2.0	9.0	11.0
Rural (domestic)	0.6	3.2	4.0	7.8
Livestock	0.2	0.2	1.1	1.5
Irrigation	0.1	1.1	1.4	2.6
Totals	5.9	7.4	20.0	33.3

Source: Water use inventory made by S. W. McKenzie for the U. S. Geological Survey in 1971 and for the University of Delaware in 1966. Assignment of pumpage figures to the Columbia deposits and collection of some municipal pumpages was made by R. H. Johnston.

Pumpage from the Columbia (Pleistocene) deposits in Delaware was 16.1 mgd in 1953, according to Marine and Rasmussen (1955, Table 7). Thus, pumpage has more than doubled in the 17-year interval from 1953 to 1970.

Pumpage for irrigation varies greatly in Delaware from year to year depending upon precipitation during the summer. In 1970, a year of average precipitation, pumpage for irrigation was a modest 2.6 mgd as shown in Table 5. However, in 1966, a very dry year, pumpage for irrigation was considerably higher as reported by Sundstrom and Pickett (1968, 1969, 1970, and 1971).

Potential Development

General Availability of Water from the Columbia Deposits

The current pumpage from the Columbia deposits (33 mgd) represents only about 4 percent of the natural discharge from the aquifer (800 mgd). Therefore, it is apparent that a much greater amount of water can be withdrawn from the aquifer. The question is how much greater can pumpage be without producing undesirable effects? The term "safe yield" has been used in the past to describe the amount of water that can be withdrawn from an aquifer on an indefinitely long basis. However, the interpretation of what is "safe yield" is controversial among hydrologists. Probably the most practical definition of safe yield has been given by Lohman (1972) as follows: "the amount of water one can withdraw without getting into trouble." What constitutes "trouble" in pumping from the Columbia deposits of Delaware? To answer this question it is necessary to review briefly the hydrologic effects of withdrawing large amounts of water from wells.

In what has become a classic paper in hydrology, Theis (1940) has described the essential factors involved in the response of an aquifer to pumping from wells. As noted by Theis, all water discharged by wells must be balanced by a loss of water somewhere. Initially, there will be a loss of water from storage in the aquifer. Later, as the cone of depression expands, recharge must increase (if recharge was previously rejected), or the natural discharge to streams must decrease in order to balance the loss of water.

Figure 14 shows the manner in which the cone of depression theoretically expands around a well pumping at a continuous rate of 500 gpm from the Columbia deposits. In

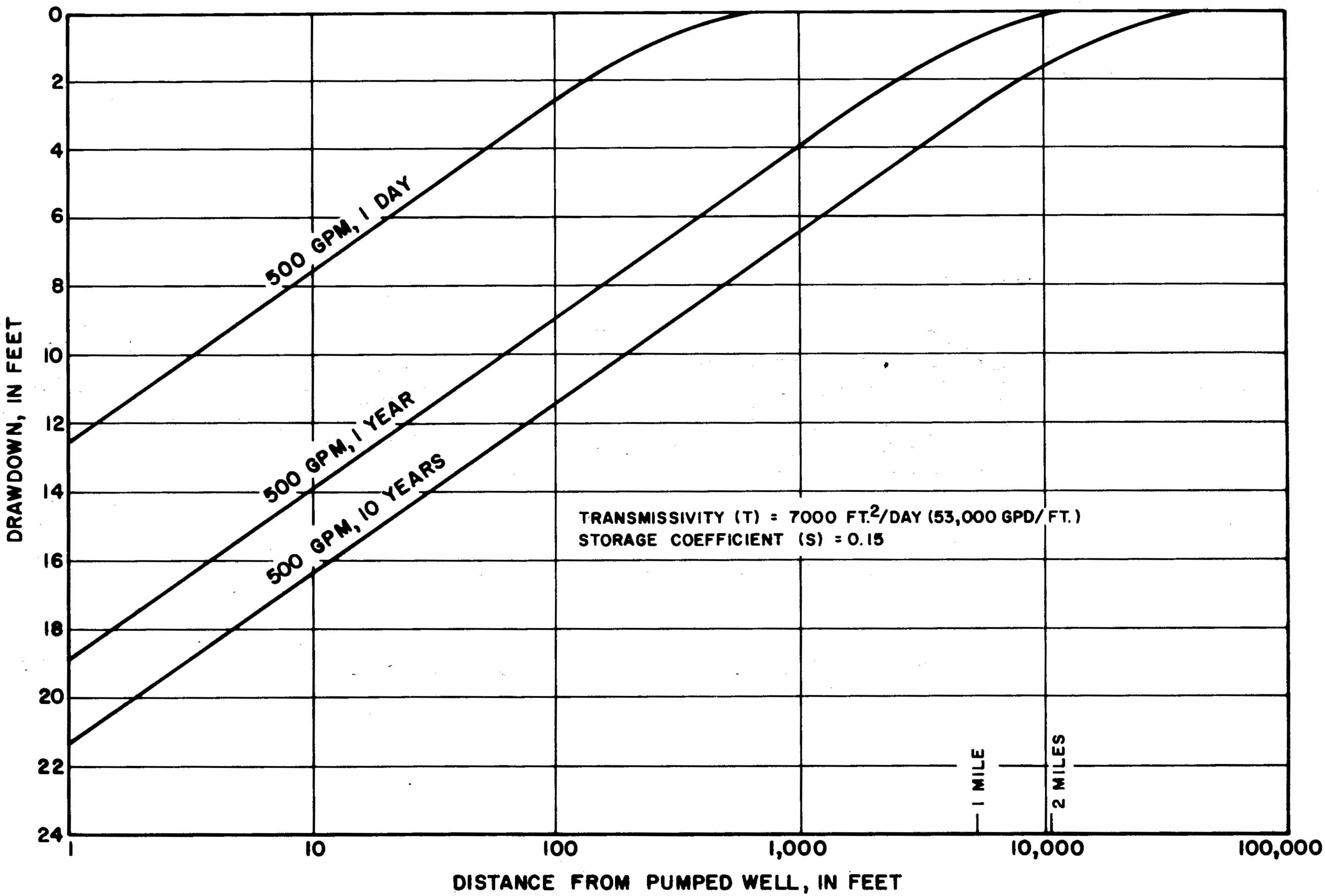

Figure 14. Drawdown around a hypothetical pumped well discharging at 500 gallons per minute from Columbia deposits of average transmissivity.

preparing Figure 14, the average transmissivity (7,000 ft^2/day) and the average (long-term) specific yield (0.15) for the Columbia deposits in central and southern Delaware were applied to the Theis (1935) equation to compute the drawdown shown. It is evident from Figure 14 that the cone of depression will extend about 600 feet from the well after 1 day of pumping and, theoretically, will extend about 2 miles from the well after 1 year of continuous pumping. As observed by the writer, the average distance from perennial streams to the ground-water divide in central and southern Delaware is about 4,000 feet. Therefore, the cone of depression will probably intersect a stream before a year of pumping elapses. Initially some water that would naturally discharge to the stream will be diverted to the well. Later water may percolate through the streambed into the aquifer and be diverted to the well. Although the streams are partially penetrating, vertical head losses within the Columbia aquifer are usually very small, and the streams can be treated as fully penetrating when considering their interaction with well pumpage.

In addition, lowering of the water table may reduce direct evapotranspiration from the aquifer immediately adjacent to the stream (where the water table is normally shallow), and additional water will be diverted to the well.

In some parts of Delaware, upward leakage into the Columbia deposits or downward leakage into the Miocene aquifers is substantial under natural conditions. In these areas, pumping a well in the Columbia deposits will cause either an increase in upward leakage or a decrease in downward leakage. The term capture has been applied to any decrease in natural discharge from an aquifer (direct evapotranspiration, ground-water runoff to streams, or leakage) brought about by pumping (Lohman and others, 1972).

When the amount of the captured water from all sources equals the pumping rate of the well, the system will be in equilibrium and there will be no further expansion of the cone of depression. Pumping from additional wells will disturb the equilibrium and require expansion of the composite cones of depression to capture more water. Another effect produced by pumping is reduction of the saturated thickness of the aquifer which, in turn, reduces the transmissivity. The result of a reduction in transmissivity is that a deepening of the cone of depression is required to maintain the same pumping rate.

The relative importance of each of the sources contributing water to a pumped well should be considered in assessing the availability of water from the Columbia deposits. Reduction in direct ground-water evapotranspiration (ET) is probably not an important source of "captured" water moving toward a pumped well. There is not a great difference between the mean annual base flow (0.86 cfsm) and the mean winter base flow (1.03 cfsm) of streams draining the aquifer (Table 4). There are several factors which could be responsible for this difference, one of which is direct ET from the water table near streams. Even if it assumed that the entire base flow difference is due to direct ET and if the unlikely assumption is made that all this water loss could be captured, there would not be a substantial contribution to a pumping well. Inducing an appreciable amount of water to move upward from the Miocene artesian aquifers, across confining clay beds, is theoretically possible but difficult to achieve in practice. Appreciable upward movement will occur only if there is a substantial head difference (between the water table in the Columbia deposits and the potentiometric surface in the artesian aquifer) over a large area. As shown in Figure 14, the drawdown around a well pumping from the Columbia deposits is relatively small (less than 10 feet) at distances greater than a few hundred feet from the well. In summary, it is probable that most of the water diverted to a well pumping from the Columbia deposits (after the initial period of reduction in aquifer storage) is water that would naturally discharge to streams. Therefore, large long-term withdrawals of water from the Columbia deposits will probably be balanced by nearly equal reductions in the base flow of streams.

Returning to the question of what is the "safe yield" of the aquifer, it is apparent that the "trouble" that must be considered is reduction of streamflow. As noted earlier, ground-water runoff constitutes about 72 percent of the total streamflow. Therefore, if 100 mgd were continuously withdrawn from the Columbia deposits, the total ground-water runoff (800 mgd) would be reduced by 12 percent and the total streamflow would be reduced by 9 percent. In practice, then, the "safe yield" of the Columbia deposits is that amount of reduction in streamflow that is permissible. Of course, if water pumped from the aquifer is returned to the draining stream and not "consumed" or transported out of the basin, there is no stream depletion problem. Surface water is not used for public water supplies in central and southern Delaware. However, uses of surface water include small withdrawals for industrial purposes and a small amount for irrigation plus use of the streams for recreational purposes and sewage dilution. The decision as to the amount of streamflow depletion permissible really becomes a local basin problem. Where a small basin is underlain by a highly transmissive aquifer and the surface water is not used, a large reduction in streamflow might be permissible. On the other hand, if the surface water is used and the aquifer transmissivity is low, it might be undesirable to reduce streamflow; the low transmissivity probably would make it costly and impractical to do so anyway.

Sundstrom and Pickett (1968, 1969, and 1970) estimated that 320 mgd of ground water is available from the water-table aquifer in central and southern Delaware. However, note that withdrawal of 320 mgd, if balanced by an equal reduction in ground-water discharge, would result in a 30 percent decline in total streamflow. On the other hand, if a large part of the 320 mgd were returned to streams directly or by way of the Pleistocene deposits, as would be expected, the 30 percent decline in total streamflow would be reduced proportionately. To minimize the effect on streamflow, Sundstrom and Pickett (1970) suggested planning ground-water withdrawals to capture as much evapotranspiration as possible in areas of high water table or swamps. However, as noted previously, rates of ground-water evapotranspiration appear to be low in most areas, and this method of "capturing" water is probably not feasible except in swampy areas. Sundstrom and Pickett (1970) also suggested using ground-water pumpage for supplementing the base flow of streams during droughts. This could readily be accomplished by returning a large part of the normal ground-water pumpage directly to streams.

In summary, the amount of <u>possible</u> withdrawal of water from the Columbia deposits in central and southern Delaware is limited to the ground-water runoff to the streams (about 800 mgd). However, the <u>desirable</u> or <u>practical</u> rate of withdrawal is less than this amount depending upon: (1) the desired rate of streamflow in any particular area, and (2) the amount of ground-water pumpage returned to the streams or the aquifer.

Availability of Large Ground-Water Supplies

The need for large ground-water supplies in central and southern Delaware falls into two general categories:

(1) High-capacity wells that can be pumped at rates of 500 gpm or more for relatively short periods. Such wells are used by seasonal resorts, particularly during summer weekends and for irrigation.

(2) High-capacity wells that can be pumped continuously to supply the year-round needs of industries and public water supplies.

In the first case, it is desirable to construct a well in a thick saturated section of Columbia sand characterized by high transmissivity. As noted earlier, the average transmissivity of the Columbia deposits is about 7,000 ft^2/day (53,000 gpd/ft) and the average saturated thickness is about 75 feet. Figure 14 shows the drawdown around a pumping well tapping the Columbia deposits, assuming average transmissivity. It is apparent from Figure 14 that wells constructed in such an aquifer can readily yield 500 gpm, with relatively small drawdown for short periods of time. Pumped water will come primarily from storage in the aquifer, which is readily replaced during the nonpumping period (the winter-spring period of ground-water recharge).

Within the areas of high transmissivity shown in Figure 9 (T greater than 10,000 ft^2/day), properly constructed wells should produce 500 gpm with little drawdown. For short periods, 1,000 to 2,000 gpm can be obtained from these wells. Note that 25 wells with reported short-term yields exceeding 1,000 gpm are listed in Table 7.

Actually throughout central and southern Delaware, where the average saturated thickness of the Columbia deposits is 75 feet, properly constructed large-diameter wells should produce 500 gpm or more in most areas. To obtain a yield of 500 gpm from such an aquifer, a specific capacity of about 10 gpm per foot of drawdown is needed. As shown in Figure 13, almost 90 percent of the large-diameter wells tapping the Columbia deposits in Delaware have specific capacities exceeding 10 gpm per foot.

Obtaining high-capacity wells capable of continuously yielding 500 gpm or more requires consideration of factors other than the transmissivity and saturated thickness of the aquifer. This is particularly true if a well field is needed that will continuously yield several million gallons per day. As noted in the previous section, most water obtained from a continuously pumped well in the Columbia deposits is (after a long period) water that would naturally discharge to streams. An ideal method of obtaining very large supplies of water from the Columbia deposits is to locate wells in highly transmissive sand and gravel adjacent to a stream characterized by high base flow. Such wells may obtain large amounts of water by diverting water that naturally discharges from the aquifer to the stream, as well as diverting water already in the stream through the streambed into the aquifer. The hydrologic situation in two areas that meet these requirements will be discussed here. One if a proven area near Salisbury, Md., and the other is an undeveloped area near Milton, Del., which was investigated during this study.

Channel-fill deposits of Pleistocene age, located between Salisbury, Md. and the Delaware-Maryland state line, are ideally situated for development of a large water supply utilizing streambed infiltration. These highly permeable sand and gravel beds attain a maximum thickness of 220 feet (Hansen, 1966) and the transmissivity as determined from a 30-day aquifer test is 400,000 gpd per ft (Mack and Thomas, 1972). Under natural conditions, the Pleistocene deposits discharge about 7 cfs (3,100 gpm) into two streams: Little Burnt Branch and the North Prong Wicomico River. During the aquifer test, one well, near Little Burnt Branch, was pumped continuously at 4,000 gpm. After 30 days of pumping Mack and Thomas (1972) concluded that the well was obtaining some water that had been discharging naturally to the two streams and some water that was already in the streams. The flow in Little Burnt Branch declined sharply and had the test continued 3 or 4 more days, the stream would have gone dry. According to Mack and Thomas (1972), at pumping rates less than 400 gpm, the pumped water would initially come from aquifer storage and later would be derived from water in transit to the stream plus some from local recharge. However, when the pumping rate is increased from about 400 gpm to 4,000 gpm, water in Little Burnt Branch moves through the streambed into the aquifer. Note that the Pleistocene channel aquifer near Salisbury has the highest known transmissivity on the Delmarva Peninsula. However, it is the close proximity of the draining streams and the highly permeable streambed in Little Burnt Branch that make possible the large amounts of water available from wells in this aquifer.

Beaverdam Creek basin near Milton, Del. has hydrologic characteristics favorable for the development of a large water supply by diversion of streamflow. Figure 15 shows pertinent hydrologic features in the downstream part of the basin including water-table contours, ground-water pickup along the stream, and a brief lithologic log. As shown in the log, the Columbia deposits contain 132 feet of sand and gravel (110 feet of which is saturated). The transmissivity of the Columbia deposits is about 100,000 gpd per ft based on a 48-hour aquifer test (see discussion of this test in the section on aquifer coefficients). Beaverdam Creek is characterized by the highest base flow per square mile in Delaware. During 1970, a year of slightly below-average precipitation, the average ground-water runoff was 11.8 cfs (7.6 mgd) which is equivalent to 1.65 cfs per square mile, almost twice the average ground-water runoff (0.86 cfsm) listed for other Delaware streams in Table 4. There is substantial ground-water discharge to the stream in the stretch from Hunter's Mill Pond to the gaging station, as shown by a seepage run made in April, 1969 (Figure 15).

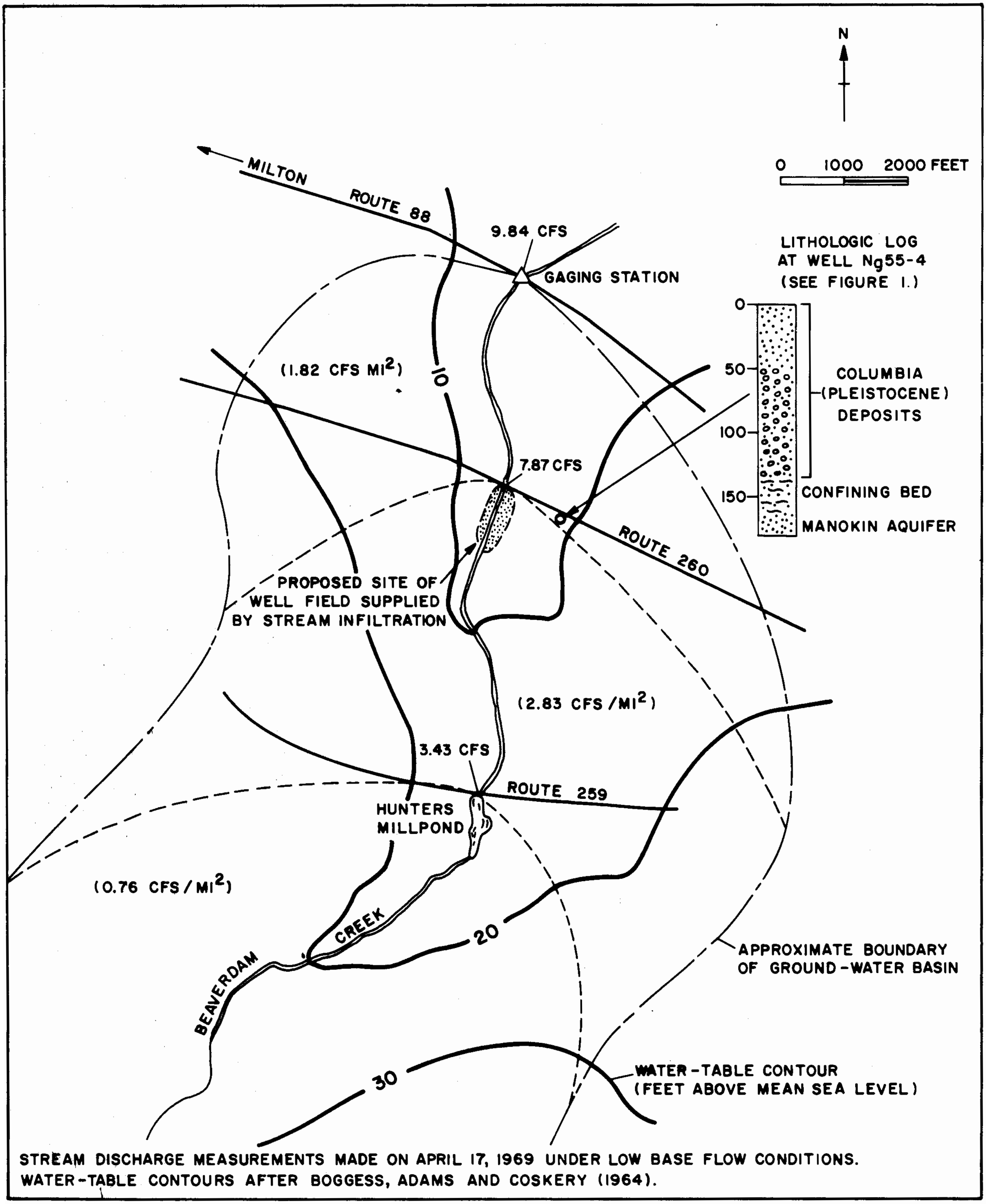

Figure 15. Hydrologic conditions in part of Beaverdam Creek Basin near Milton, Del.

Similar quantities of ground-water inflow were computed using airborne radiometric temperature measurements, as reported by Hollyday (1970).

The very high base flow in Beaverdam Creek (1.65 cfsm) cannot be accounted for by discharge from the Columbia deposits alone. The test drilling disclosed that permeable sands of the Miocene Manokin aquifer underlie the Columbia deposits with only a thin (18-ft) confining bed (silt) separating the two aquifers. It seems probable that upward leakage from the Manokin contributes to the exceptionally high base flow of Beaverdam Creek (Johnston, 1971).

Several factors affect the yield of a well field supplied by water diverted from a stream. These factors have been summarized by Rorabaugh (1963). In Beaverdam Creek basin, the ultimate yield of a well field will depend primarily on:

(1) the ability of the streambed to transmit water (vertical hydraulic conductivity of the sand and silt constituting the streambed),

(2) the transmissivity of the aquifer,

(3) the average base flow of the stream, and

(4) the ability of wells to capture ground water en route to the stream.

Figure 15 indicates that the greatest ground-water pickup occurs between Hunter's Millpond and Route No. 260. Wells in this area would be expected to have the greatest chance of capturing ground water before it discharges to the stream. On the other hand, a well field located downstream near the gaging station would be in a position to divert the entire flow of the stream, if necessary.

The most important factor in selecting the location of wells to be supplied by streambed infiltration may be the hydraulic conductivity of the silty layer on the streambed. A line of shallow observation wells constructed at right angles to a stream is probably the best approach to evaluating the hydraulic nature of the silty layer. If the water table profile has a gradual slope and is continuous with the stream surface, the silty layer is probably not an effective barrier. If the water table is above or below the stream surface immediately adjacent to the stream, the silty layer is an effective seal.

A line of piezometer tubes constructed at right angles to the stream at the Beaverdam Creek gaging station indicated a steeply sloping water table. At a tube 20 feet from the stream, the water table is 0.5 to 1.0 foot above the stream. In addition, small seeps occur in the stream bank a few feet above the stream surface. All these data suggest that the silt bed is probably an effective barrier near the gaging station.

A suggested location for a well field to be supplied by streambed infiltration is shown in Figure 15. The site is in the downstream part of the reach of greatest ground-water pickup and is within a quarter of a mile of the aquifer test site where high transmissivity has been confirmed. The ground-water runoff of the stream at the proposed well field is about 80 percent of that measured at the gage. Thus, ground-water runoff during 1970 would have been 9.4 cfs or 6.1 mgd. The 6 mgd figure gives an approximate indication of the amount of water that could be pumped continuously at the proposed well field. Lowering of the water table might induce additional water to move up from the Manokin aquifer; however, the amount of such additional water cannot be estimated with present information. It is unlikely that additional water can be obtained by reduction of evapotranspiration as the water table is lowered. The water table is 20 feet below land surface a short distance from the stream and ground-water evapotranspiration is probably very small at present.

The fact that the part of Beaverdam Creek basin shown in Figure 15 is less than 10 miles from Lewes and Rehoboth Beach (see Figure 1) suggests that the area could be a useful future source of water for the fast-developing seashore area. At present, combined pumpage of ground water along the Delaware Atlantic seashore (Lewes, Rehoboth Beach, Bethany Beach, etc.) is about 4 mgd (Miller, 1971). This water is supplied by a large number of wells, and some salt-water contamination has occurred in shallow wells near the beaches. In contrast, three or four properly constructed large-diameter wells located in the Beaverdam Creek basin should readily yield 4 mgd, and the location about 7 miles inland would minimize the possibility of salt-water encroachment.

The advantages of a well field supplied by streambed infiltration over a well field supplied only from aquifer storage may be summarized as follows:

(1) By utilizing induced infiltration, a relatively large supply of water can be obtained from a few wells at one site. In contrast, many wells, more widely spaced, would be needed to supply the same quantity of water from the Columbia deposits if the wells were not near a stream.

(2) Objectionable characteristics of surface water, such as turbidity and inconsistent temperature, are usually removed or attenuated during percolation through the aquifer.

The major disadvantage of an induced-infiltration supply is that streamflow is reduced and the stream can go dry under maximum pumpage.

Wells suppled by river infiltration have been developed in many areas, including the Ohio, Mississippi, and Mohawk River valleys. In some of those areas, very large water supplies have been developed by horizontal-type collector wells, which are effective in increasing rates of infiltration through the streambed. However, large infiltration supplies have also been obtained by conventional vertical wells in Louisville, Ky. and Schenectady, N.Y.

Wells obtaining water by streamflow infiltration are not common in Delaware at present. In the Seaford area, where about 4 mgd is withdrawn from the Columbia deposits, a large part of the pumpage is probably ground water that naturally discharges into the Nanticoke River and some of the water may be obtained from streambed infiltration. At Smyrna, Del., a supply well at the Delaware State Hospital and one of the town supply wells tap a permeable sand aquifer within 200 feet of Lake Como. At high sustained pumping rates, those wells may obtain water by infiltration from the lake.

The potential for developing very large infiltration supplies in the Delaware Coastal Plain is limited because the area is drained by many small streams. The two largest streams, the Nanticoke River and Marshyhope Creek, whose drainage areas are 75 and 44 square miles, respectively, have sufficient base flow to support large supplies derived from stream infiltration. The Columbia deposits have an average thickness of about 50 feet in the Marshyhope basin. However, almost nothing is known of the aquifer transmissivity in this basin. In the Nanticoke basin, the range of transmissivity is wide (6,000 to 80,000 gpd per ft), and the possibility exists that a site suitable for a large infiltration supply can be found.

A few smaller basins probably have the potential for developing infiltration supplies. In each case, a preliminary study, such as was made in the Beaverdam Creek basin, would be needed to assess the area as a possible infiltration supply. The study ideally would include test drilling, seepage runs along the stream, and a preliminary aquifer test.

QUALITY OF WATER AND GROUND-WATER CONTAMINATION

Water from the Columbia deposits is generally "soft," slightly acidic, and characterized by low dissolved-solid content. High iron content and low pH are the only natural characteristics of the water that commonly require treatment. However, local problems of ground-water contamination have occurred in recent years due to salt-water intrusion in a few coastal areas, waste disposal, and accidental spills.

Table 6 lists the major chemical constituents in ground water from the Columbia deposits. The table is based on chemical analyses made by the U. S. Geological Survey on samples from 19 wells. Several additional analyses were not included because of known or suspected contamination by salt-water encroachment or waste disposal. Table 6 is, therefore, an attempt to show the "natural" chemical quality of the ground water. The principal constituents, as shown in Table 6, are calcium, sodium potassium, silica, bicarbonate, sulfate, chloride, and nitrate. Hardness (both the calcium magnesium and noncarbonate varieties) is generally low and treatment is usually not necessary.

High iron content is a common water-quality problem in the Columbia deposits. About one-third of the wells listed in Table 6 produced water containing more than 0.3 mg/l -- the maximum recommended for drinking water by the U. S. Public Health Service (1962). Such water will cause reddish-brown staining and produce iron scale which may clog pipes or well screens. However, iron may be removed by several methods including aeration and filtration and treatment with lime or polyphosphates.

The occurrence of iron in the Columbia deposits is sporadic. The highest reported iron concentrations occur at Rehoboth Beach and Selbyville (Rasmussen and others, 1960), and the lowest occur northwest of Seaford (Woodruff, 1970). The iron-rich water at Selbyville is the subject of a current geochemical investigation by John C. Miller of the DGS.

The position of the Columbia deposits near land surface makes the aquifer particularly susceptible to contamination by leachates from landfills and dumps, effluent from septic tanks, and accidental spills. In a recent survey of ground-water contamination in the northeastern states, 9 of 12 cases cited for Delaware involved the Columbia (Pleistocene) deposits. Contamination of the ground water by effluent from septic tanks or cesspools reportedly has occurred at Dover, St. Georges, and Laurel. In one of those areas, located just south of Dover, Miller (1972) determined that the water in 6 of 13 domestic wells sampled contained nitrate concentrations

TABLE 6 - Chemical Constituents in Water from 19 Wells Tapping the Columbia deposits (Chemical constituents, in milligrams per liter)

Constituent or Chemical Property	Minimum	Maximum	Average
Silica (SiO_2)	9.8	25.	16.
Iron (Fe)	.00	2.1	.33
Calcium (Ca)	1.6	17.	7.6
Magnesium (Mg)	0.4	13.	5.2
Sodium and Potassium (Na + K)	3.7	40.	15.
Bicarbonate (HCO_3)	4.	38.	17.
Sulfate (SO_4)	0.4	40.	13.
Chloride (Cl)	4.	86.	21.
Nitrate (NO_3)	0.	36.	13.
Dissolved Solids	50.	235.	113.
Hardness (as $CaCO_3$):			
Calcium, Magnesium	5.	93.	39.
Noncarbonate	0.	64.	18.
pH	5.4	7.5	6.1

exceeding the maximum level of 45 mg/l recommended for drinking water by the U. S. Public Health Service (1962). Other reported cases of ground-water contamination have been caused by seepage of salty water from a dredging spoil area along the Chesapeake and Delaware Canal, seepage of salty water from an uncovered salt pile near Christiana, and leakage from gasoline storage tanks near Laurel.

Salt-water contamination of the Columbia deposits is a local, but not as yet serious, problem in the coastal areas of Delaware. During the 1940's, salty ground water appeared in wells used by the towns of Lewes, New Castle, and Rehoboth Beach. For Lewes and Rehoboth Beach, it was necessary to drill new wells farther inland from the beach.

The salt-water problem was not investigated in the current hydrologic appraisal of the Columbia deposits. However, any projected large withdrawals of water from this aquifer in the coastal zone ideally would include a local site investigation of salt-water encroachment. The salt-water problem has been reviewed in general terms by Sundstrom and Pickett (1970) and a detailed discussion of salt-water encroachment at Lewes is given by Rasmussen and others (1960).

APPLICATION OF FINDINGS

The findings of this report should be helpful in the development of large ground-water supplies in central and southern Delaware. The important factors to be kept in mind when developing water supplies from the Columbia (Pleistocene) deposits are as follows:

(1) Water in the Columbia deposits and much of the water in streams is, in effect, "one water." This conclusion is based on the fact that discharge of ground water from the Columbia deposits supplies about three-quarters of the streamflow. In contrast, overland runoff supplies only one-quarter of the streamflow. Therefore, if large amounts of ground water are pumped, it must be anticipated that declines in streamflow will occur (unless the pumped ground water is returned to streams or the aquifer). Conversely, if water is to be withdrawn from a stream, ground water cannot also be pumped on a large scale.

(2) Large supplies of ground water (more than 1 mgd) can be most easily and economically obtained by constructing wells in areas of high transmissivity (shown in Figure 9) adjacent to streams characterized by high base flow. Wells at such sites may obtain large amounts of water by diverting ground water that naturally discharges to a stream and, also, by diverting water already in the stream to move through the streambed into the aquifer.

(3) The 800 mgd of water that discharges from the Columbia deposits is not the "safe yield" of the aquifer. This water provides the fair-weather (base flow) of streams in central and southern Delaware as well as ground-water pumpage from the aquifer.

The discussion of Beaverdam Creek basin near Milton provides an example of how to identify an area of high ground-water potential. A gaging station provided the necessary streamflow record from which it was determined that the stream had very high base flow. High transmissivity was suspected based on nearby well logs, geologic data, and base-flow data. Exploratory drilling and a pumping test confirmed the existence of a thick section of Columbia deposits and high transmissivity. The maps of saturated thickness and transmissivity (Figures 3 and 9) presented in the report are intended to help those interested in identifying similar areas in the future.

Note that high transmissivity alone will not insure the permanence of a large ground-water supply. High T means that, initially, well yields and specific capacity will be high. However, unless the draining stream or streams are characterized by high base flow, the wells are dependent primarily on aquifer storage, and as pumping proceeds, yields may decline.

An area southwest of Houston (Figure 1) is a case in point. At this site, the Columbia deposits have the highest transmissivity (22,000 ft^2/day or 165,000 gpd/ft) known to date in Delaware. The site is very near the divide between the drainage areas of the Chesapeake and Delaware Bays. The nearby stream, Beaverdam Branch, drains less than 3 square miles and the total base flow is small. Within this small basin a ground-water supply of about 1.5 mgd can be efficiently and economically developed from one or two high-capacity wells. However, despite impressive yields from one or two wells, this small basin will never provide a very large supply of ground water.

In addition to providing suggestions for the future development of large ground-water supplies from the Columbia deposits, this report is intended to supply technical data for more sophisticated hydrologic studies. Ideally such studies would take the form of digital or electric analog models of the stream-aquifer system. These model studies can provide an understanding of the workings of the system and can also be used as water-management tools.

TABLE 7 - Yield and specific capacity of large-diameter wells tapping the Columbia (Pleistocene) deposists and estimated transmissivity of the aquifer

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity	
							gal/day/ft	ft^2/day
Cd 42-13	Artesian Water Co.	73	17	49 - 73	570	38	60,000	8,000
Cd 43-6	Atlas Chemical Industries, Inc.	71	26	52 - 67	600	27	40,000	5,300
Cd 51-1	Artesian Water Co.	47	17	24 - 47	350	10	18,000	2,400
Db 31-19	E. I. du Pont Co.	65	10	53 - 65	100	8	14,000	1,900
Db 31-35	E. I. du Pont Co.	80	8	70 - 80	500	11	9,000	1,200
Dc 53-5	Getty Oil Co.	90	8	70 - 90	525	48	72,000	9,600
Eb 55-4	Baker Brothers	40	17	16 - 40	600	19	30,000	4,000
Eb 55-5	Warren Baker	45	17	17 - 45	450	24	36,000	4,800
Ec 41-7	Fred Haas	93	17	29 - 89	220 ?	4	6,000	800
Fb 23-6	Fred Wicks	71	8	30 - 70	100	8.3	12,000	1,600
Fb 34-16	University of Delaware	74	8	54 - 74	110	3.5	33,000	4,500
Fb 51-4	George & Sam Brooks	44	17	24 - 44	520	19	29,000	4,000
Fb 51-5	George & Sam Brooks	44	17	16 - 44	620	20	30,000	4,000
Fb 51-6	Norman & Sam Brooks	54	17	--	300	12	19,000	2,500
Fb 53-7	Chris Wicks	80	17	21 - 80	580	9	14,000	1,900
Ga 15-4	Gerald Zeh	63	17	30 - 60	1,120	51	76,000	10,000
Hc 32-5	City Products Corp.	56	10	43 - 56	325	20	36,000	4,800

TABLE 7 - (continued)

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity	
							gal/day/ft	ft²/day
Hc 32-12	W. L. Wheatley, Inc.	77	12	--	780	14	25,000	3,300
Hc 34-3	City of Smyrna	100	10	70 - 100	550	30	54,000	7,200
Hc 34-22	City of Smyrna	96	12	55 - 85	1,100	54	120,000	16,000
Hc 43-3	George Wicks	115	10	--	1,020	82	150,000	20,000
Hc 44-3	George Wicks	133	12	75 - 132	1,050	88	160,000	21,000
Hc 55-1	Walter Gibe	120	10	90 - 120	1,000	52	93,000	12,000
Ic 32-4	Frank Johnson	48	12	--	350 ?	5	9,000	1,200
Id 24-4	Philip Cartanza	157	17	25 - 157	1,050	20	30,000	4,000
Ie 31-1	Joseph Zimmerman	94	17	16 - 94	940	19	28,000	3,700
Ie 43-2	Philip Cartanza	106	17	16 - 103	1,400	50	75,000	10,000
Ie 53-2	Alfred Bilbrough	114	17	21 - 113	700	54	82,000	11,000
Ie 53-3	Alfred Bilbrough	72	17	36 - 72	750	17	25,000	3,300
Ie 53-4	Alfred Bilbrough	70	17	13 - 69	720	21	32,000	4,200
Jc 34-1	Joseph Wild	97	17	20 - 96	900	17	25,000	3,300
Jd 12-2	Eugene Gagan	104	17	40 - 96	1,300	22	33,000	4,400
Jd 21-2	Papen Farms	96	17	40 - 96	760	17	23,000	3,100
Jd 41-1	Libby, McNeil, & Libby, Inc.	125	10	87 - 109	400	9	15,000	2,000

TABLE 7 - (continued)

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity	
							gal/day/ft	ft2/day
Jd 54-1	Joseph Jackewiez	118	17	26 - 118	1,260	21	38,000	5,100
Je 12-2	Jacob Zimmerman	72	10	--	600	16	24,000	3,200
Je 13-1	Alfred Bilbrough	70	17	18 - 66	780	21	31,000	4,200
Kd 24-2	Joseph Kowalski	134	13	38 - 134	1,050	21	31,000	4,200
Ke 12-2	Kenneth Bergold	146	17	30 - 74; 122 - 146	1,600	44	67,000	9,000
Ld 33-1	Walter Winkler	64	17	20 - 56; 60 - 64	1,400	39	60,000	8,000
Le 22-2	Charles West	73	8	13 - 73	400	11	17,000	2,300
Le 23-3	Charles West	74	13	22 - 74	380	9	14,000	1,900
Le 51-2	Floyd Blessing	105	17	--	1,200	64	96,000	13,000
Md 15-8	Libby, McNeil, & Libby, Inc.	67	10	41 - 67	280 +	22	40,000	5,300
Md 24-3	U. S. Geological Survey - Delaware Geological Survey	80	8	70 - 80	300 +	10	165,000	22,000
Md 54-2	John Annet	82	17	30 - 82	1,180	53	80,000	11,000
Me 24-5	City of Milford	80	8	--	400	18	26,000	3,500
Me 24-6	City of Milford	69	12	39 - 59	520	29	42,000	5,600
Me 33-3	Donald Calhoun	74	17	38 - 70	700	30	45,000	6,000
Me 54-5	Delmarva Nursery	90	8	58 - 90	500	20	30,000 +	4,000 +

TABLE 7 - (continued)

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity	
							gal/day/ft	ft2/day
Mf 11-6	City of Milford	67	--	--	220	25	44,000	5,900
Mf 21-1	Diamond State Nurseries	63	8	--	180	15	27,000	3,600
Mf 22-4	Brown Thawley	131	17	44 - 106	1,400	80	120,000	16,000
Mg 42-13	Draper Foods, Inc.	89	8	69 - 89	150 +	35	50,000	6,700
Nc 25-19	Bramble Canning Co.	84	--	--	1,300	24	36,000	4,800
Nc 53-3	O. A. Newton & Sons	100	12	--	1,080	22	33,000	4,400
Nd 33-1	J. Howard Lyons	92	17	--	1,200	30	45,000	6,000
Ng 12-1	Carlton Clifton	61	6	--	300	17	31,000	4,100
Ng 31-1	Willard Workman	122	17	--	900	74	110,000	15,000
Ng 41-2	Willard Workman	112	17	--	1,080	32	48,000	6,400
Ng 42-1	Town of Milton	68	6	--	200	31	56,000 +	7,500 +
Ng 42-15	Draper Canning Co.	---	13	--	1,300	65	100,000	13,000
Ng 42-16	Draper Canning Co.	84	13	--	1,020	49	73,000	10,000
Ng 55-4	U. S. Geological Survey - Delaware Geological Survey	70	8	60 - 70	200 +	7	104,000	14,000
Ni 31-3	Lewes Dairy	60	8	--	100 +	14	25,000	3,300

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity gal/day/ft	Estimated transmissivity ft2/day
Ni 51-16	Town of Lewes	97	10	--	480	16		
Ni 51-17	Town of Lewes	157	10	--	500	11		
Ni 51-18	Town of Lewes	89	10	--	400	11	110,000	15,000
Ni 51-19	Town of Lewes	151	10	--	975	--		
Ni 51-20	Town of Lewes	146	10	--	900	26		
Ni 52-1	Diamond State Poultry Co.	94	8	--	100	20	36,000	4,800
Oc 14-4	H. P. Cannon & Son, Inc.	109	8	--	600	21	38,000	5,100
Oc 14-5	H. P. Cannon & Son, Inc.	116	10	--	800	21	38,000	5,100
Oc 14-6	H. P. Cannon & Son, Inc.	98	8	--	500	15	27,000	3,600
Oc 24-1	H. P. Cannon & Son, Inc.	106	17	--	700	55	83,000	11,000
Of 42-23	Townsend, Inc.	105	16	--	950	27	41,000	5,500
Of 43-2	Swift & Co.	110	10	--	575	44	66,000	9,000
Og 23-1	Paramount Poultry Co.	64	6	--	150	4	7,000 +	1,000 +
Og 23-3	Paramount Poultry Co.	78	8	--	340	8	13,000 +	1,700 +
Oi 24-1	Town of Rehoboth	102	12	73 - 102	380	6		
Oi 34-1	Town of Rehoboth	131	12	69 - 131 (Multiple)	720	23	55,000	7,300
Pb 35-2	Ralph O'Day	89	17	33 - 89	920	42	63,000	8,400

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity	
							gal/day/ft	ft2/day
Pc 23-3	City of Seaford	95	10	--	800	28	50,000	6,700
Pc 33-5	E. I. du Pont Co.	98	10	--	540	12	21,000	2,800
Pc 33-9	E. I. du Pont Co.	83	10	--	640	24	43,000	5,700
Pc 33-10	E. I. du Pont Co.	78	10	--	600	24	43,000	5,700
Pc 33-11	E. I. du Pont Co.	101	10	--	620	13	23,000	3,000
Pc 44-4	Ralph Givens	109	17	65 - 109	300 +	40	60,000	8,000
Pc 54-1	Fred O'Neal	100	17	16 - 100	900	65	100,000	13,000
Pd 53-1	Emory Spicer	164	17	44 - 160	950	38	57,000	7,600
Pf 23-2	H. Kruger, Inc.	180	17	40 - 180	1,200	29	51,000	6,800
Pg 53-8	Town of Millsboro	83	8	--	250	9	14,000	1,900
Pg 53-9	Town of Millsboro	84	8	--	180	7	11,000	1,500
Pg 54-1	Millsboro Poultry Co.	105	8	--	400	12.5	19,000	2,500
Qb 44-5	Howard Rider	111	17	--	1,070	40	60,000	8,000
Qc 14-5	Coleman Wheatley	98	17	--	830	49	74,000	10,000
Qc 14-6	Coleman Wheatley	90	17	--	550	32	48,000	6,400
Qc 24-6	Emory Spicer	121	17	--	1,400	107	160,000	21,000
Qc 34-1	Paul Spear	86	17	78 - 86	1,300	36	54,000	7,200
Qh 51-7	Town of Frankford	100	8	80 - 100	100 +	22.5	36,000	4,800

TABLE 7 - (continued)

Well number	Owner	Depth (feet)	Diameter (inches)	Screened interval (feet)	Yield (gallons per minute)	Specific capacity (gpm/ft)	Estimated transmissivity	
							gal/day/ft	ft^2/day
Qh 51-10	Delmarva Poultry Co.	105	8	--	240	24	38,000	5,100
Qh 51-14	Town of Frankford	101	8	80 - 101	350 +	40	64,000	8,500
Rd 31-11	Town of Delmar	139	8	--	---	18.4	33,000 ?	4,400
Rh 32-9	Town of Selbyville	125	8	105 - 125	300	9	16,000	2,100

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