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ANALYTICAL METHODS, NUMERICAL MODELING, AND MONITORING STRATEGIES FOR EVALUATING THE EFFECTS OF GROUND-WATER WITHDRAWALS ON UNCONFINED AQUIFERS IN THE NEW JERSEY COASTAL PLAIN

Water-Resources Investigations Report 98-4003

Prepared in cooperation with the

NEW JERSEY DEPARTMENT OF ENVIRONMENTAL PROTECTION

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U.S. DEPARTMENT OF THE INTERIOR

BRUCE BABBITT, *Secretary*

U.S. GEOLOGICAL SURVEY

Thomas J. Casadevall, *Acting Director*

For additional information
write to:

District Chief
U.S Geological Survey
Mountain View Office Park
810 Bear Tavern Road, Suite 206
West Trenton, NJ 08628

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CONTENTS

ILLUSTRATIONS

ILLUSTRATIONS--Continued

ILLUSTRATIONS--Continued

ILLUSTRATIONS--Continued

TABLES

CONVERSION FACTORS

Multiply	By	To obtain
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
foot per mile (ft/mi)	0.1894	meter per kilometer
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
cubic foot per day (ft^3/d)	0.0283	cubic meter per day
cubic foot per second (ft^3/s)	0.0283	cubic meter per second
inch per year (in/yr)	2.54	centimeter per year
million gallons per day (Mgal/d)	3.785	million liters per day
gallons per minute (gal/min)	3.785	liters per minute

ANALYTICAL METHODS, NUMERICAL MODELING, AND MONITORING STRATEGIES FOR EVALUATING THE EFFECTS OF GROUND-WATER WITHDRAWALS ON UNCONFINED AQUIFERS IN THE NEW JERSEY COASTAL PLAIN

By Edward Modica

ABSTRACT

Analytical and numerical solutions of ground-water withdrawals in the unconfined part of the Kirkwood-Cohansey aquifer system of the Coastal Plain of New Jersey were evaluated for their usefulness in predicting the area of influence of a pumped well and in determining hydraulic characteristics of an aquifer. Additionally, simulations of ground-water withdrawal using a finite-difference model provided information on the ways in which prudent well-location strategies can disperse the local effects of withdrawal over a larger part of an aquifer system. The design of a monitoring network that is sensitive to the ground-water hydraulics of streams and wetlands of the Coastal Plain of New Jersey also was considered for its utility in providing hydrologic data necessary to establish the baseline hydrologic conditions near wetlands and streams and in signaling when ground-water levels are being adversely affected by withdrawals elsewhere in the system.

The application of methods based on the Theis analytical solution to ground-water flow in unconfined aquifers can lead to erroneous estimates of the size of the area of influence generated by ground-water withdrawals. Analysis of time-drawdown data from an unconfined aquifer system are best evaluated by means of the Neuman solution, which accounts for the effects of gravity drainage; however, the pumped well must be far enough from streams so that ground water is not drawn from nearby streams. Time-drawdown data from a test well in Winslow County, N.J., were analyzed by means of the Neuman solution. Results indicate that the aquifer has a relatively high vertical to horizontal anisotropy of 1:198, and a specific yield of 0.028, an indication that the area of influence of a pumped well at the test site would be relatively large.

Results from a finite-difference ground-water-flow model of the northeastern part of the Mullica River Basin near Chesilhurst, N.J., show that the area influenced by a long-term withdrawal is best estimated from a steady-state ground-water-flow analysis that includes the effects of average areal recharge. Withdrawal simulations indicate an order-of-magnitude difference between the size of the area of influence generated from a 3-day (72-hour) withdrawal and the size of the area produced under steady-state conditions. An aquifer characterized by a low specific yield will cause the area of influence to extend farther away from the pumped well.

The contributing area of flow to the pumped well includes areas on the water table that would, under natural conditions, be incorporated into the contributing areas of flow to streams. Ground water that is drawn to a pumped well is diverted from nearby streams; the withdrawal decreases the size of the contributing areas of flow to streams by an amount equal to the contributing area of flow to the well.

Withdrawals made from a well close to a stream divert ground water that would, under natural conditions, flow to the stream. The diverted ground water causes the area of influence of the well to be smaller than it would if the well were far from the stream. Water-table declines caused by withdrawals near streams are, to some degree, mitigated by ground-water diversion from

streams. However, the withdrawals can significantly reduce ground-water seepage to nearby streams, especially along stream reaches and wetlands close to the well. Alternatively, these effects can be dispersed over a large part of the aquifer if wells are located on surface-water divides.

Measurements of seasonal water-level fluctuations in the Mullica River Basin indicate that the greatest fluctuations in water levels are found in upland areas, where the average fluctuation is 3.4 feet. Fluctuations in hydraulic head in the wetland areas averages 1.3 feet. The bimodal average of ranges in water levels show that upland areas are more sensitive to recharge than lowland areas. The pattern of yearly mean water levels fluctuates irregularly about a long-term mean value. Abnormally low or high yearly average values that are brought on by periods of drought or excess recharge are short lived; over time, hydrologic conditions shift back to average levels under natural conditions.

Wetland areas in the New Jersey Coastal Plain are characterized by ground-water seepage into wide, shallow depressions. Periods of inundation are longest in the deepest part of the depression, whereas inundation of areas near the fringes of wetlands due to ground-water seepage is only seasonal. The seepage face in the fringe areas expand and contract in response to seasonal variation in water-table elevation and in response to precipitation.

Values of the aquifer storage coefficient and transmissivity can, in some cases, be determined by use of hydraulic head or streamflow recession analysis as an alternative to aquifer testing. The recession curves developed from hydrographs of Middle Branch and McDonalds Branch in the New Jersey Coastal Plain indicate that the aquifer near McDonalds Branch has about 2.6 times the storage capacity of the aquifer adjacent to Middle Branch; this finding is consistent with the relatively small ranges of water-level changes measured in McDonalds Branch compared to those measured in Middle Branch.

INTRODUCTION

Recent growth in the population of the Coastal Plain of southern New Jersey (fig. 1) has increased the demands on water supply. The Kirkwood-Cohansey aquifer system is a major ground-water reservoir in the Coastal Plain that has been used to meet the water needs of the region. The New Jersey Department of Environmental Protection (NJDEP) permits withdrawals from the aquifer system, but an aquifer test is required to determine the potential effects of withdrawal on water levels in the system. The NJDEP, through its Bureau of Water Allocation (BWA), requires that reports be submitted by prospective water purveyors to support applications for water-supply allocation. The NJDEP has formulated guidelines (Hoffman and others, 1992) that consider the essential aspects of aquifer-test design and test-data analysis. The guidelines do not go into the details of aquifer-test procedures and ground-water hydraulics but do provide a compendium of standard procedures and suggestions for their use.

As part of the BWA protocol to help evaluate the effect of ground-water withdrawals on ground-water levels, the area of influence (defined as drawdown of the water table or the cone-of-depression attributed to well withdrawals) must be evaluated. The BWA considers the area of influence to be an area that undergoes a minimum of 1 ft of drawdown when water levels have completely stabilized in response to sustained withdrawal (Hoffman, and others, 1992). Generally, this area is determined at the end of a 72-hour aquifer-test period. Aquifer-test data are typically analyzed with the Theis solution, which is usually adequate for many aquifer systems in New Jersey and can give reasonable initial estimates of aquifer characteristics and the area of influence.

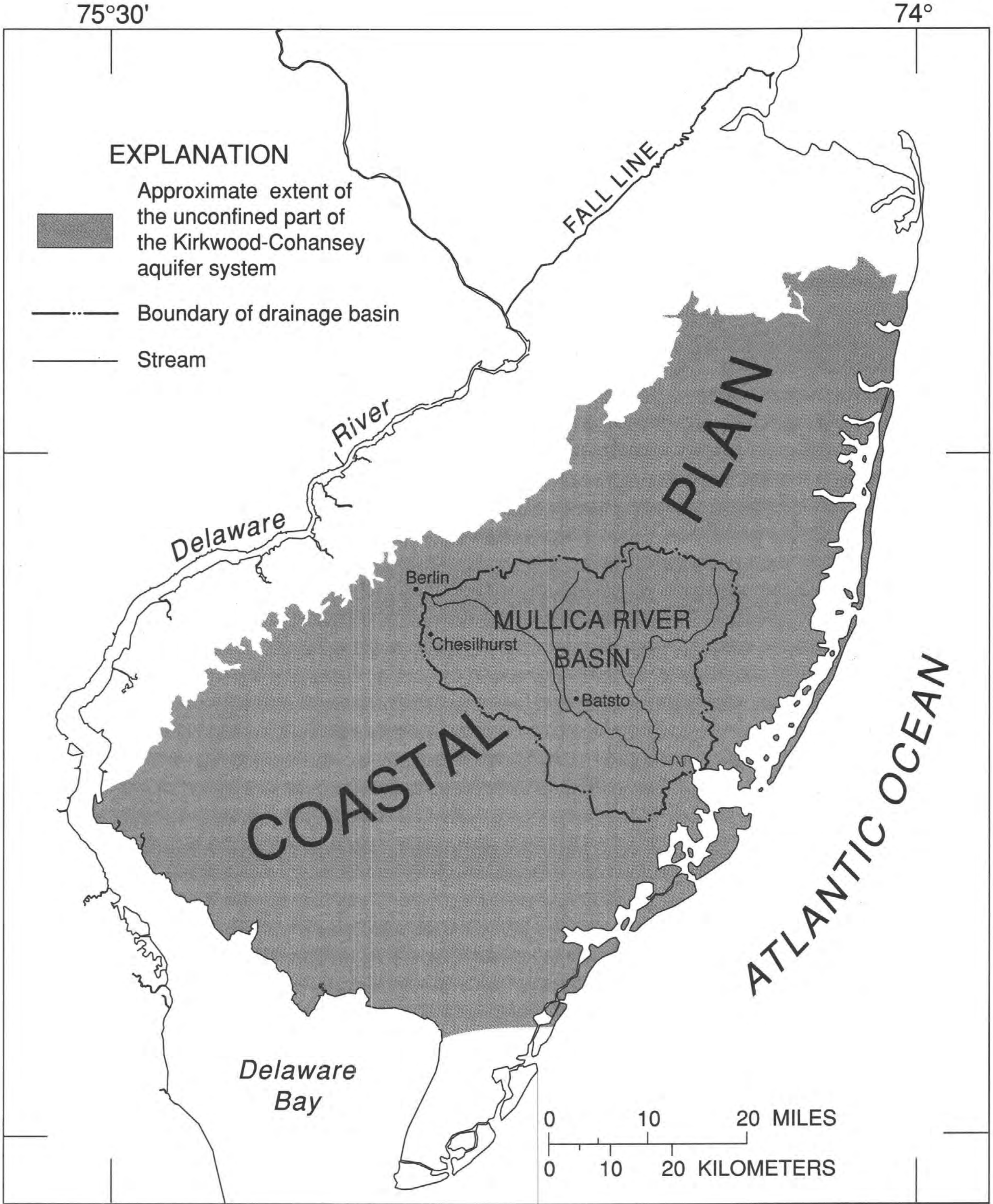


Figure 1. Extent of the unconfined part of the Kirkwood-Cohansey aquifer system in the New Jersey Coastal Plain.

However, several issues related to this standard method of evaluating the area of influence need to be considered. These issues, which are described below, were the subject of a study conducted by the U. S. Geological Survey in cooperation with the NJDEP.

1. **The practicality of using an area of influence generated after only 72 hours of withdrawal to predict the effects of long-term withdrawals.** The 1-ft minimum drawdown and 72-hour aquifer-test duration that are described in the guidelines are necessary and reasonable constraints on an aquifer test. Because drawdown can propagate to great distances from the pumped well, some limit on test duration must be imposed. Furthermore, aquifer tests that last for many days can incur major expenses and create serious water-disposal problems. Nevertheless, the drawdown configuration that results from 72 hours of withdrawal is not in equilibrium with the ground-water-flow system, even though the rate of propagation of the area of influence decreases significantly near the end of that period. Over weeks or months, the area of influence produced by ground-water withdrawal can grow significantly larger than the area produced by 72 hours of withdrawal, depending on the ground-water storage properties of the aquifer. In addition, average areal recharge is a significant component of the hydrologic budget that affects the size of the area of influence when withdrawal occurs over a long period, but this component is unaccounted for in a 72-hour aquifer test. Consequently, an accurate estimate of the configuration of a long-term, stable area of influence cannot be derived from the size of the area of influence that results from a 72-hour aquifer test. The resulting data from such tests are suitable only for estimating the hydraulic characteristics of the aquifer and the size of the area of influence for short periods of withdrawal.

2. **The applicability of the Theis solution to unconfined systems such as those in the Coastal Plain.** An analysis that uses the Theis solution for time-drawdown data derived from an unconfined aquifer can be misleading because the equation on which the Theis solution is based is relevant only to confined flow. The Theis solution is based, in part, on the assumptions that ground water is released from elastic storage during withdrawal and that ground-water flow is horizontal, assumptions that can lead to erroneous estimates of the size of the area of influence when applied to data obtained for an unconfined aquifer.

3. **The effects that ground-water diversion from streams and wetlands in the Coastal Plain have on the area of influence.** Withdrawals made near streams and wetlands can induce ground-water flow to the well that is not accounted for by the Theis solution and may lead to gross errors in the estimates of the areas of influence produced by withdrawal. The consequences of withdrawals near wetlands and streams are important to NJDEP water-allocation permit policy because such policy dictates that withdrawal sites need to be far enough to minimize effects on wetlands and streams.

Purpose and Scope

This report describes the results of a study to evaluate the effects of withdrawals on unconfined aquifers in the Coastal Plain of New Jersey and to evaluate the strengths and limitations of methods used to analyze aquifer-test data. The report also discusses water-level data collected near streams and wetlands in the New Jersey Coastal Plain and suggests strategies for monitoring water levels near wetlands of the Coastal Plain. The study was organized into several components, as follows:

1. The Theis and Neuman analytical methods were evaluated to determine their capability of accurately (a) predicting the size of the area of influence for various aquifer-test durations and (b) determining the hydraulic characteristics of an aquifer from time-drawdown data recorded for an aquifer test near Berlin, N.J.

2. A numerical ground-water-flow model of the Sleeper Branch-Albertson Brook Basins, located in the northwestern part of the Mullica River Basin, was developed and used to (a) determine a configuration of contributing areas of flow to streams under natural conditions and the altered configurations that result from withdrawal, (b) determine the size of the area of influence produced by steady-state withdrawal, 2 weeks of withdrawal, and 3 days (72 hours) of withdrawal for various withdrawal rates, (c) evaluate the effect of withdrawals on ground-water seepage to streams and wetlands, and (d) establish a relation between the area of influence and its distance from a stream or wetland.

3. Water-level data collected during 1992–93 in and around the Mullica River Basin were evaluated to determine the seasonal fluctuations of the water table and the periods of inundation of wetlands.

Work included the installation of piezometers in wetlands of the New Jersey Coastal Plain and the measurement of water levels to establish seasonal changes in the water table. This report discusses findings for October 1992 through June 1994.

Acknowledgments

The assistance of Eric Jacobsen, U.S. Geological Survey, in installing piezometers and measuring water levels is gratefully acknowledged. Thanks also to Timothy Reed, U.S. Geological Survey, for his guidance and assistance in establishing elevations of piezometers in wetlands. The author also wishes to thank John Cervetto and William Haines, Jr., for permission to install piezometers on their properties.

ANALYTICAL METHODS AND NUMERICAL MODELING FOR EVALUATING THE EFFECTS OF GROUND-WATER WITHDRAWALS

Finding a solution to an axisymmetric ground-water-flow problem (radial flow to a well) involves understanding the spatial and temporal scales over which the aquifer system will be affected by pumping. In general, if the scale of the problem is local, and the assumptions regarding the physical characteristics of the aquifer can be simplified, then a solution can usually be derived from an appropriate analytical method, often more quickly and economically than from a numerical ground-water-flow model. Analytical methods can be practical for determining the water-transmitting properties of the aquifer in the area near the pumped well, especially when short pumping times (days) are involved. If ground-water withdrawals pertain to more complex systems, however, and involve factors such as flow to streams, regional-flow gradients, aquifer heterogeneity, or long-term withdrawals at high pumping rates, then flow solutions are best determined by use of numerical models. In this section, both analytical and numerical methods are described and the relative merits of each are evaluated.

Analytical Methods

Analytical methods based on the Theis and Neuman solutions to well hydraulics are discussed in this section. The Theis solution is used widely in standard hydrologic practice and is relatively easy to use. The Neuman solution is more complicated than the Theis solution but is more appropriate to problems that involve withdrawals from water-table aquifers.

Theis Solution to Ground-Water Withdrawal

The Theis nonequilibrium equation (Theis, 1935) is the basis for several time-drawdown analyses that are widely used to determine hydraulic characteristics of an aquifer and to predict the area of influence of a pumped well. The Theis solution, written in terms of drawdown (Fetter, 1994), is

$$s = \frac{Q}{4\pi T}\int_{u}^{\infty}\frac{e^{-u}}{u}du \ . \qquad \text{(1a)}$$

The argument u is given as

$$u = \frac{r^2 S}{4Tt} , \qquad \text{(1b)}$$

where

Q is constant rate of pumping (L^3/T),
s is the drawdown (L),
T is aquifer transmissivity (L^2/T),
t is time since commencement of pumping (T),
r is radial distance from pumped well (L), and
S is storage coefficient (dimensionless).

Typically, an aquifer-test analysis is made by matching time-drawdown data, or a "fit curve" to a Theis curve. The Theis method is based on the assumptions that the aquifer is confined, homogeneous, and of infinite extent. In using the Theis method, one also assumes that the well screen penetrates the entire thickness of the aquifer and that ground water flows horizontally to the well in a radial pattern. Although all these conditions are rarely met in the field, the method can generally be employed under less than ideal conditions with satisfactory results, provided that the aquifer is relatively thick compared to drawdown. In many cases, accepted practice has been to use the Theis method for evaluating time-drawdown data derived from test wells installed in unconfined aquifer.

When ground water is withdrawn from a confined aquifer, water is released from elastic storage; that is, water is made available as a result of minor expansion in water volume and decompression of the aquifer material near the well when hydrostatic pressure is released because of withdrawal. Typically, the coefficient for elastic storage, or storage coefficient (S), is an extremely small number, an indication that the quantity of water released from elastic storage is small and that an area of influence must propagate out over a relatively large aquifer space to satisfy withdrawal.

Neuman Solution to Ground-Water Withdrawal

In an unconfined aquifer, ground water is also released from storage when withdrawals are made from the aquifer. However, in this case, storage refers predominantly to ground water that drains by gravity from the aquifer when hydraulic heads are lowered. The storage term is defined as specific yield, a term that refers to the capacity of the aquifer material to release interstitial water after a dewatering event such as ground-water withdrawal. Values for specific yield are typically much larger than those for the storage coefficient, an indication that the area of influence produced by a pumped well in an unconfined aquifer would be relatively small compared to one produced in a confined aquifer.

To account for the effects of gravity drainage induced by withdrawal, Neuman (1972) developed an analytical solution that predicts time-drawdown relations in unconfined aquifers. Neuman's analytical solution can be represented in simplified form (Fetter, 1994) as

$$s = \frac{Q}{4\pi T} W(u_A, u_B, \Gamma) \quad , \qquad (2a)$$

where Q, s, T are defined as in equation 1 and $W(u_A, u_B, \Gamma)$ is the unconfined well function. The arguments u_A and u_B are defined as

$$u_A = \frac{r^2 S}{4Tt} , \qquad (2b)$$

and

$$u_B = \frac{r^2 S_y}{4Tt} , \qquad (2c)$$

where S_y is the specific yield (dimensionless). The argument Γ is defined as

$$\Gamma = \frac{r^2 K_v}{b^2 K_h} ,$$

where

K_v is the vertical hydraulic conductivity (L/T),
b is the initial saturated thickness of the aquifer (L), and
K_h is the horizontal hydraulic conductivity (L/T).

According to Neuman's model, time-drawdown relations for withdrawal in unconfined aquifers are characterized by three phases (fig. 2). The first phase is commensurate with the initial stage of withdrawal (eq. 2b). The aquifer behaves as an artesian aquifer and contributes water from elastic storage. During this period, flow is horizontal. The time-drawdown data follow the Theis curve for the storage coefficient. During the second phase, water is derived predominantly from gravity drainage (eq. 2c). Declines in the water table are greatest during this phase. Vertical and horizontal components of flow are both induced. The flattening of the Neuman type curve during phase 2 indicates that gravity drainage temporarily recharges the aquifer near the well. In the final phase, flow is again horizontal. Time-drawdown data follow a Theis curve of which the storage coefficient is equal to the specific yield.

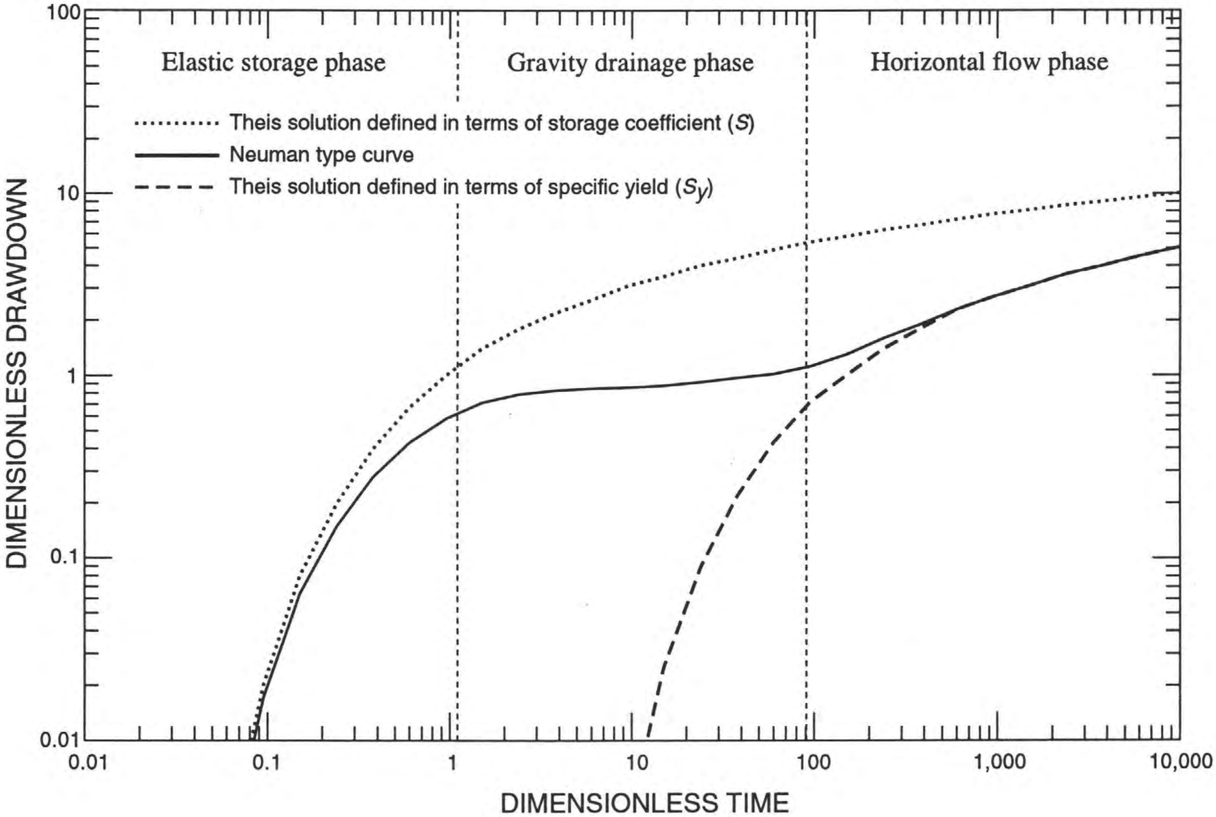


Figure 2. Example of dimensionless time-drawdown curves generated by the Neuman and Theis solutions for an unconfined aquifer and a fully penetrating pumped well.

Neuman had extended his solution to account for hydraulic heads observed in wells with various screen lengths and set at various depths within the saturated thickness of the aquifer (Neuman, 1974). Examples of three time-drawdown curves for an unconfined aquifer with a partially penetrating well are shown in figure 3. Each curve represents the time-drawdown relation that would be recorded in observation wells at the water table, at the middle of the aquifer (half the saturated thickness), and near the bottom of the aquifer. The observation wells are assumed to be 300 ft from the pumped well. The curves show that time-drawdown relations are similar near the bottom and middle of the aquifer after the initial period of withdrawal. However, the curve for the observation well screened near the bottom of the aquifer shows a delayed response. The response at the observation well screened near the water table indicates that hydraulic head decreases at a slower rate during early stages of withdrawal than that observed in the deep observation well because of the lagged effect of gravity drainage at the water table. The curves show that the effects of gravity drainage vary with depth in the aquifer and in time. Consequently, the hydraulic effects produced in an unconfined aquifer by pumpage are not adequately accounted for by the Theis solution, which is predicated on the assumption of horizontal flow.

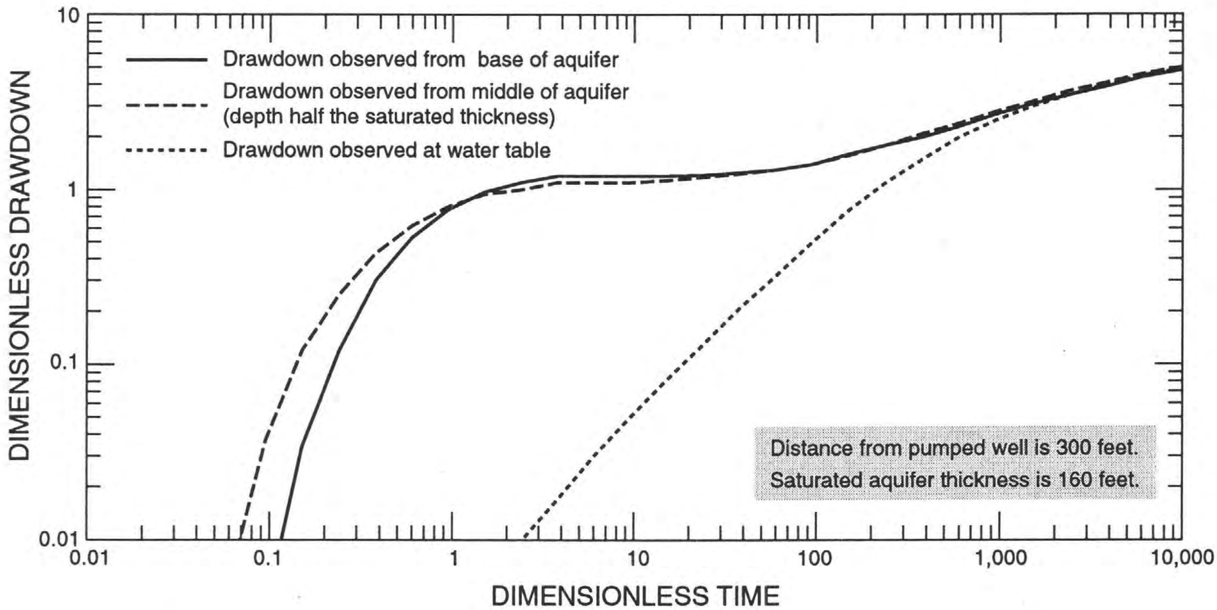


Figure 3. Example of dimensionless time-drawdown curves generated from the Neuman solution for an unconfined aquifer and a partially penetrating pumped well as observed from the water table, middle of aquifer, and base of aquifer.

Solving a Hypothetical Withdrawal Problem With the Theis and Neuman Solutions

Consider the problem of determining the sizes of the areas of influence for several pumping durations in a water-table aquifer that has a saturated thickness of 170 ft and that has a hydraulic conductivity of 65 ft/d in the horizontal direction, 2 ft/d in the vertical direction, a specific yield of 0.15, and a storage coefficient of 0.001; values typical of the Kirkwood-Cohansey aquifer system. A hypothetical well pumped at a rate of 1.6 Mgal/d is assumed to have a 40-ft screen, the top of which is set 60 ft beneath the water table. Drawdown is measured at a fully penetrating observation well. Given these constraints and the distances assumed between the pumped well and an observation well listed in the explanation of figure 4, a series of time-drawdown curves was generated by use of the Theis and Neuman solutions.

When analyzing the effects of pumping in an unconfined aquifer with the Theis solution, it is common practice to substitute the specific yield for the storage coefficient to better approximate the area of influence (Freeze and Cherry, 1979). Following this practice, the time-drawdown curves generated by use of the Theis solution (fig. 4A) indicate a 1-ft drawdown at the end of 3 days (72 hours) at a distance of 615 ft from the pumped well. Curves generated by use of the Neuman solution (fig. 4B) indicate the 1-ft drawdown at the end of 3 days would be at 890 ft from the well.

The area of influence after 3 days of pumping predicted by the Neuman solution is more than twice as large as that predicted by the Theis solution. Furthermore, the general shapes of the time-drawdown curves determined by the Neuman solution indicate that drawdown would occur earlier and at greater distances from the pumped well than drawdown indicated by the Theis curves.

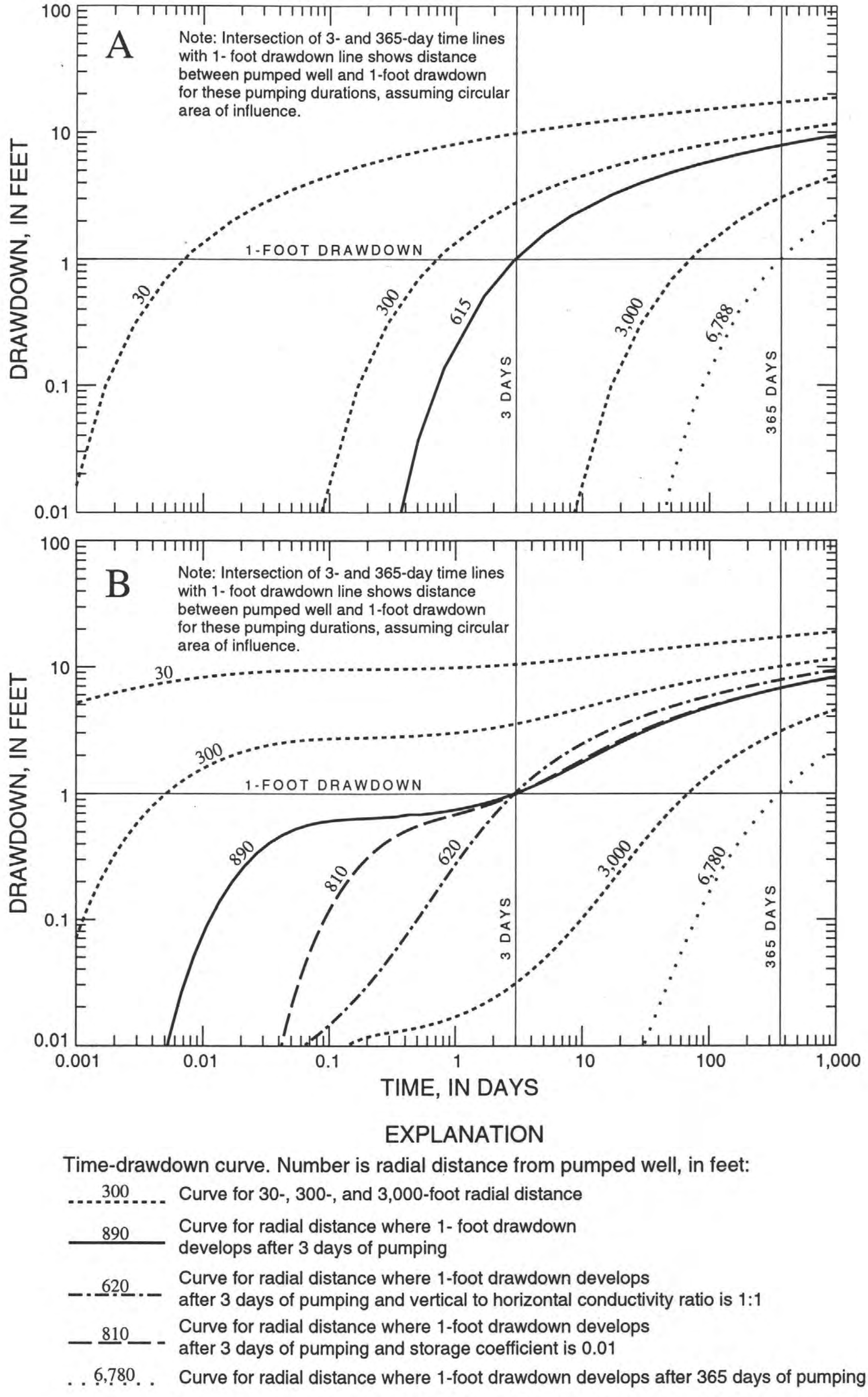


Figure 4. Time-drawdown curves for distances between pumped and observation wells based on (A) Theis and (B) Neuman solutions. (Pumping rate is 1.6 million gallons per day, aquifer transmissivity is 11,050 feet per day, vertical to horizontal conductivity ratio is 1:32.5, and specific yield is 0.15. Specific yield is used for storage coefficient in the Theis solution.)

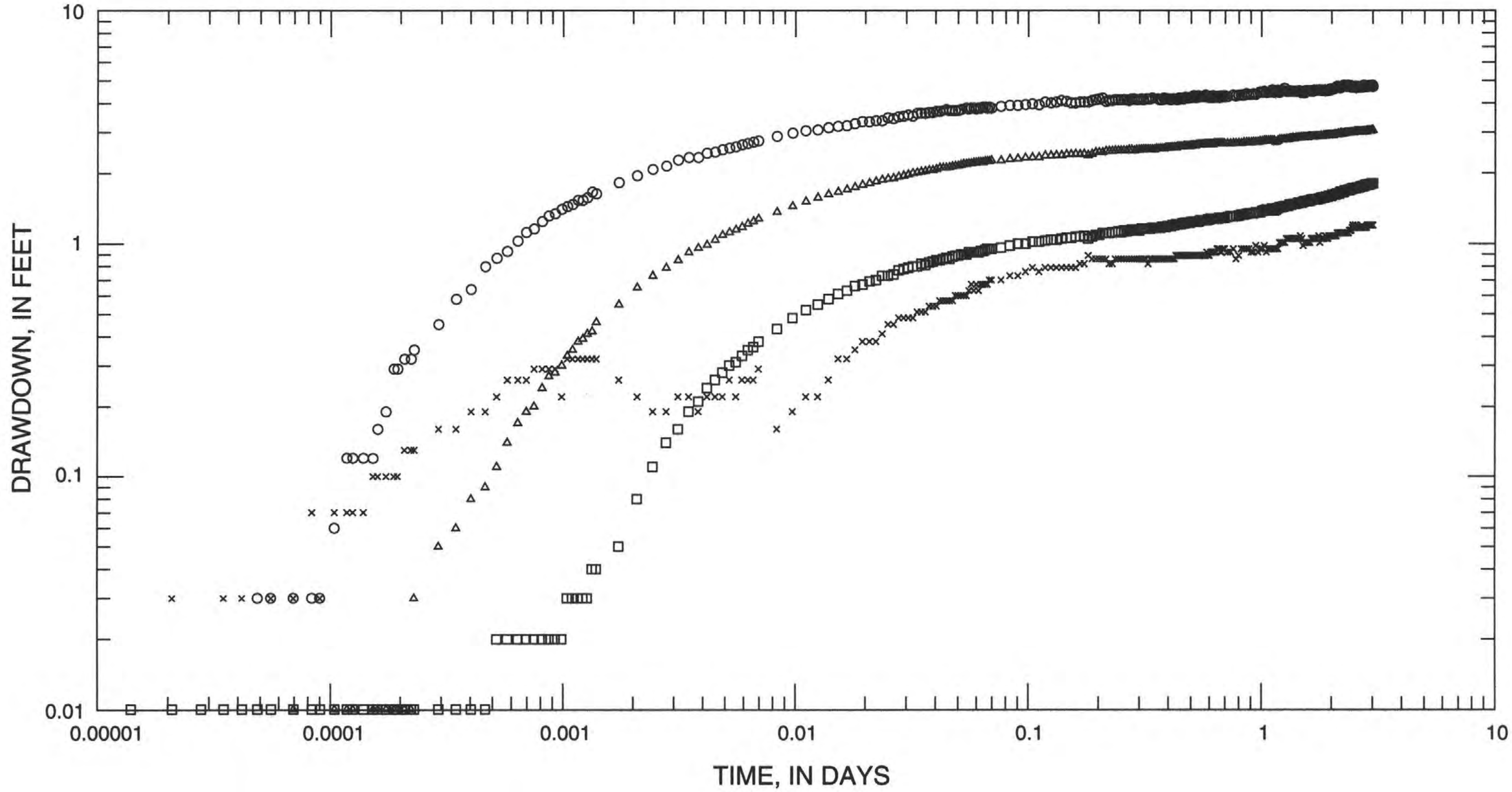
The difference can be attributed largely to aquifer anisotropy (K_v/K_h) and the ratio between elastic and gravity storage (S/S_y), as these effect the timing of gravity drainage. If the aquifer were isotropic (K_v/K_h is 1), the Neuman solution would predict a 620-ft radius for the area of influence, and the gravity-drainage segment of the curve would be short (fig. 4B). The radius and curve from the Neuman solution would then compare well with those predicted by the Theis solution, an indication that horizontal ground-water flow characterizes the time-drawdown relation according to both solutions. Similarly, if S/S_y were larger, the area of influence also would be smaller. For example, if S_y is 0.15 but S is 0.01 (instead of 0.001), the radius of the area of influence predicted by the Neuman solution would be 810 ft, and the gravity-drainage segment of the time-drawdown curve would be shorter than that produced when S/S_y is smaller (fig. 4B). The sizes of the areas of influence after a year of pumping are nearly identical according to both solutions, an indication that the effects of gravity drainage are fully dissipated at advanced time at far distances from the pumped well.

The length of the screen of the pumped well also effects the shape of the time-drawdown curve. The Neuman solution predicts that the radius of the area of influence after 3 days would be about 860 ft, only 30 feet shorter for a fully-penetrating well than for a partially-penetrating well. However, the effects of partial penetration on time-drawdown relations would be more prominent close to the pumped well.

Using the Neuman and Theis Solutions to Evaluate an Aquifer Test

The merits of using the Neuman and Theis solutions for axisymmetric flow problems are best evaluated with real-time data from an aquifer test. The NJDEP has on record the results of an aquifer test at test well TW–7 (New Jersey identification number 07-0831) at Berlin-Cross Keys Rd. in Winslow Township, N.J. ("Berlin" in fig. 1) conducted in spring 1993. The time-drawdown field data for the aquifer test for each of the observation wells are shown in figure 5. During the 3-day aquifer test, the pumping rate was maintained at 324 gal/min. Four observation wells were installed at 50, 150, 251, and 491 ft from the pumped well. The observation well located at 251 ft from the pumped well (OBS-4) is screened at the water table, whereas the tops of the screens of the other wells are approximately 27 ft below the water table. The pumped well was screened in the Kirkwood-Cohansey aquifer system and was equipped with a 35.5-ft screen, the top of which is about 43 ft beneath the water table. The saturated thickness of the aquifer is about 153.5 ft at the test site. Gamma-ray logs from wells drilled in the area indicate that sediments are finer grained from about 90 ft below land surface to the bottom of the aquifer (Zapecza, 1989). Consequently, the water-transmitting properties of the aquifer vary with depth, and the assumption of aquifer homogeneity made in the Neuman or Theis solutions may not strictly apply.

A method of aquifer-test analysis that is based on the Neuman solution but uses LaPlace Transforms to arrive at a solution (Moench, 1993) was used to derive hydraulic characteristics and predict areas of influence by use of data from the aquifer test (fig. 5). According to the method, the time-drawdown data for each observation well are fit to a type curve generated from the Neuman solution by trial-and-error. Data from the earliest time steps are ignored because the effects of well storage are dominant at early time and introduce large errors during the initial stages of drawdown. The data from each observation well are plotted on a single graph and matched to a set of type curves for the observation wells. To evaluate the aquifer-test data by this method, it is necessary to specify the positions of the screens of the observation and pumped wells as input to the flow equations used in the solution. The terms used for the Neuman solution are shown in table 1. The relative positions of the wells to the aquifer and the terms used to describe them are shown in figure 6.



EXPLANATION

Time-drawdown in observation well:

- ○ OBS–1, 50 feet from pumped well
- △ OBS–2, 150 feet from pumped well
- × OBS–3, 491 feet from pumped well
- □ OBS–4, 251 feet from pumped well

Figure 5. Plot of time-drawdown for field data from four observation wells used in an aquifer test at Berlin-Cross Keys Road, Winslow Township, New Jersey.

Table 1. Distances of observation wells from test well and screen-depth parameters used in Neuman's solution to aquifer test at Berlin-Cross Keys Road, Winslow Township, New Jersey

[ft, feet; --, not applicable; spatial relations shown in fig. 6]

Well number	Well name	r (ft)	(r/b)2 [a]	z_1 (ft)	z_2 (ft)	(zd1)[b]	(zd2)[c]
07-0831	TW-7[d]	--	--	--	--	--	--
07-0830	OBS-1	50	0.106	88	128	0.573	0.834
07-0829	OBS-2	150	.955	88	128	.573	.834
07-0827	OBS-3	491	10.232	88	128	.573	.834
07-0828	OBS-4	251	2.674	145	153.5	.945	1.0

a. b = 153.5 ft
b. zd1 = z_1/b
c. zd2 = z_2/b
d. Pumped well; length of screen is 35.5 ft; top of screen 43 ft beneath water table

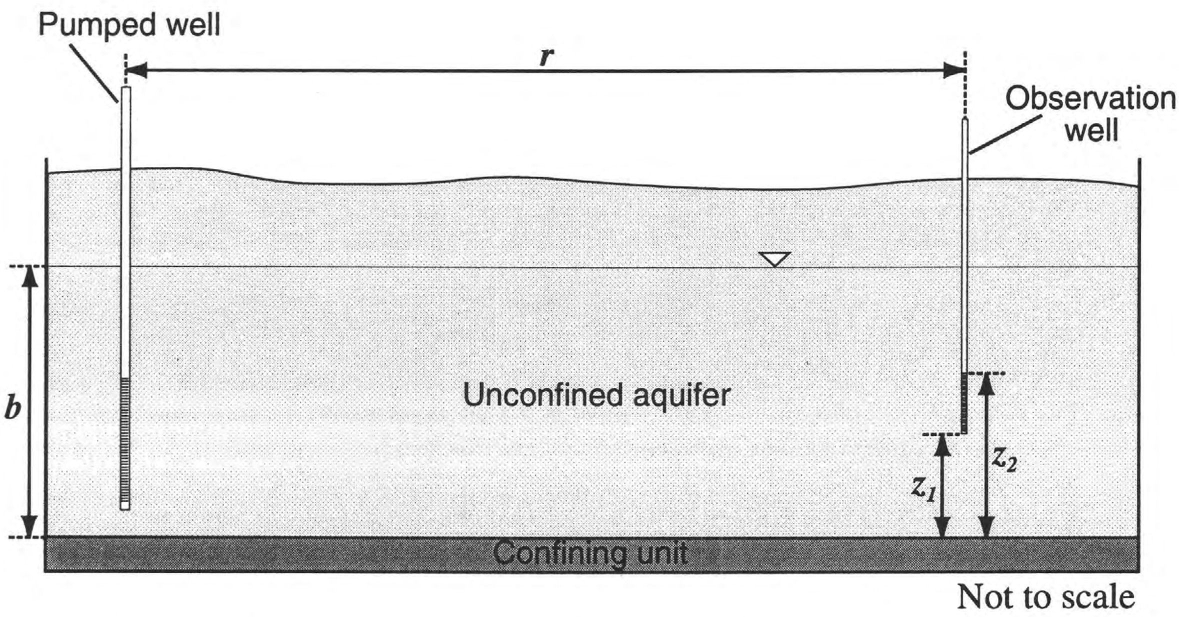


Figure 6. Spatial relations of parameters r, z_1, z_2, and b used in the Neuman solution to ground-water hydraulics near a pumped well completed in an unconfined aquifer.

A composite match of all data and best-fit type curves yields a set of hydraulic characteristics that represent average or integrated values of the water-transmitting properties of the aquifer surrounding the wells at the test site. Data that do not fit the type curves may indicate lack of hydraulic connection between pumped and the observation well, faulty measuring equipment, or simply that well parameters have been inaccurately specified. Data at well Obs–4 (fig. 5) appear to be entirely out of position relative to the type curve and may indicate one of these problems.

The best-fit Neuman curves that are superimposed on a plot of field data are shown in figure 7. The match point also is indicated (but the abscissa that corresponds to the type curves is not shown). The results of the analysis indicate that the aquifer near the test well has a horizontal hydraulic conductivity of 95.1 ft/d and a specific yield of 0.028. Because the Neuman solution relates vertical to horizontal hydraulic conductivities of the aquifer to drawdown, the vertical hydraulic conductivity is also readily derived and was found to be 0.48 ft/d. The high vertical to horizontal anisotropy (1:198) and the relatively low specific yield may indicate that the aquifer is somewhat stratified and interlayered with clay laminae. Time-drawdown curves that correspond to the aquifer characteristics derived from the pumped well (observed from a fully penetrating observation well) are shown in figure 8. The curves indicate that, after 3 days of pumping, the 1-ft drawdown boundary would be 480 ft from the pumped well. This is in close agreement with the 1.2-ft drawdown recorded after 3 days at OBS-3 located 491 ft from well. After a year of pumping, this distance would increase to 4,650 ft.

An analysis of the same aquifer-test data by the Theis solution gives different results from those determined by the Neuman solution. The example provided in figure 9 shows the time-drawdown data as analyzed by Jacob's straight-line method (Cooper and Jacob, 1946), a technique based on the Theis solution that is commonly used in engineering practice. The basis for this method is that drawdown can be expressed by an asymptote when the radius (r) is small and time (t) is large because the argument u (eq. 1b) becomes small and all terms of the solution series after the first two become negligible. The drawdown can then be expressed (Kruseman and deRidder, 1992) as

$$s = \frac{2.3Q}{4\pi T}\left(\log\frac{2.25Tt}{r^2 S}\right), \quad (3a)$$

where s, Q, T, t, r, and S are defined as in equation 1. The plot of s and the log of t is a straight line. If the line is projected to zero drawdown (where s is equal to 0), then t is equal to the initial time (t_o) and expression 3a becomes

$$T = \frac{2.3Q}{4\pi\Delta s} \quad (3c)$$

and

$$S = \frac{2.25Tt_o}{r^2}, \quad (3b)$$

where Δs is the difference in drawdown over a log cycle (fig. 9A). Following this technique, time-drawdown data are plotted on semilog paper. A straight line is fit to one of the field-data curves, and the transmissivity and storage coefficient are determined from the line (Kruseman and

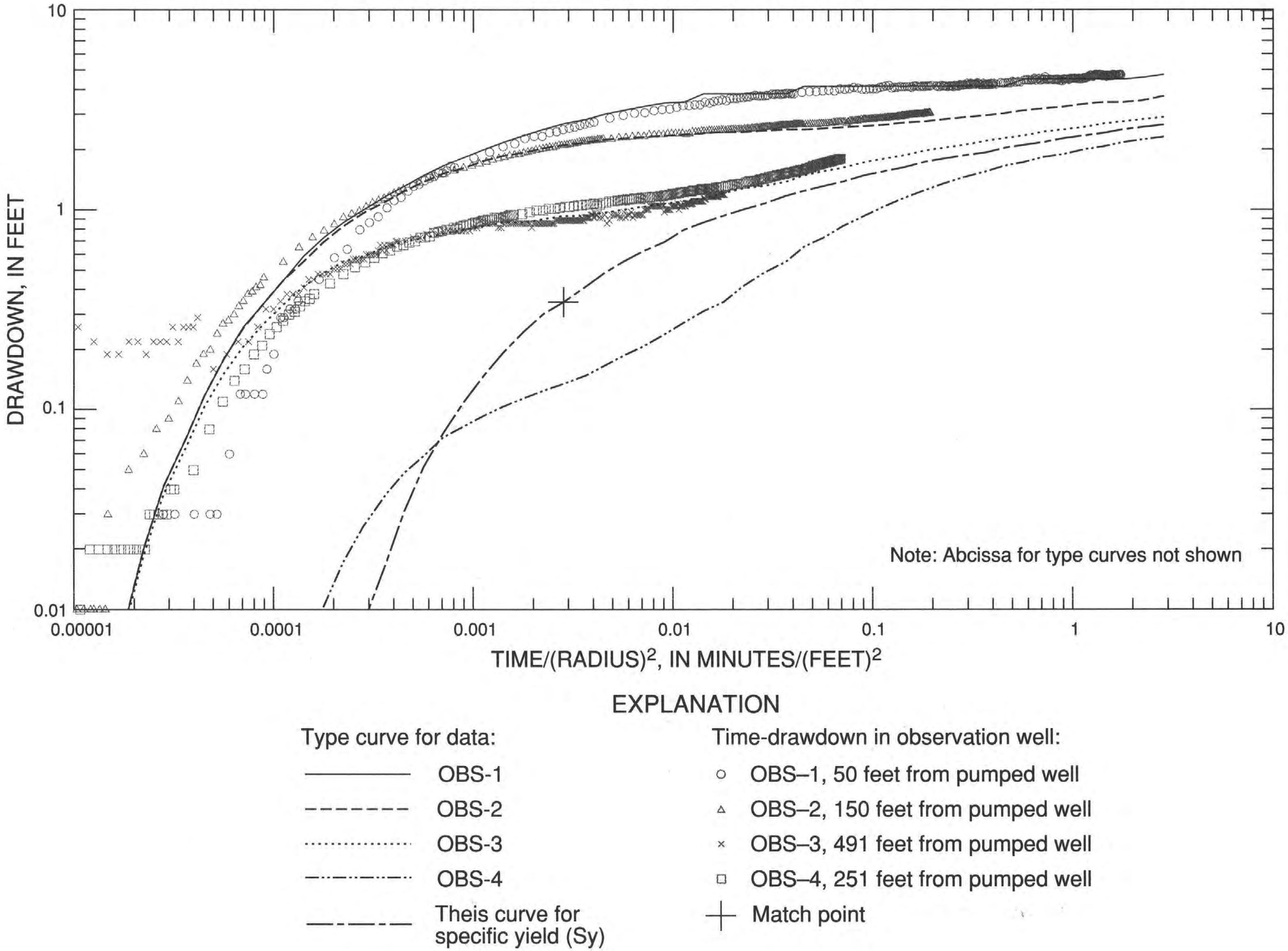


Figure 7. Plot of time/(radius)^2 and drawdown data from four observation wells used in an aquifer test at Berlin-Cross Keys Road, Winslow Township, New Jersey. (The best match curves and match point derived from the Neuman solution are superimposed on plot.)

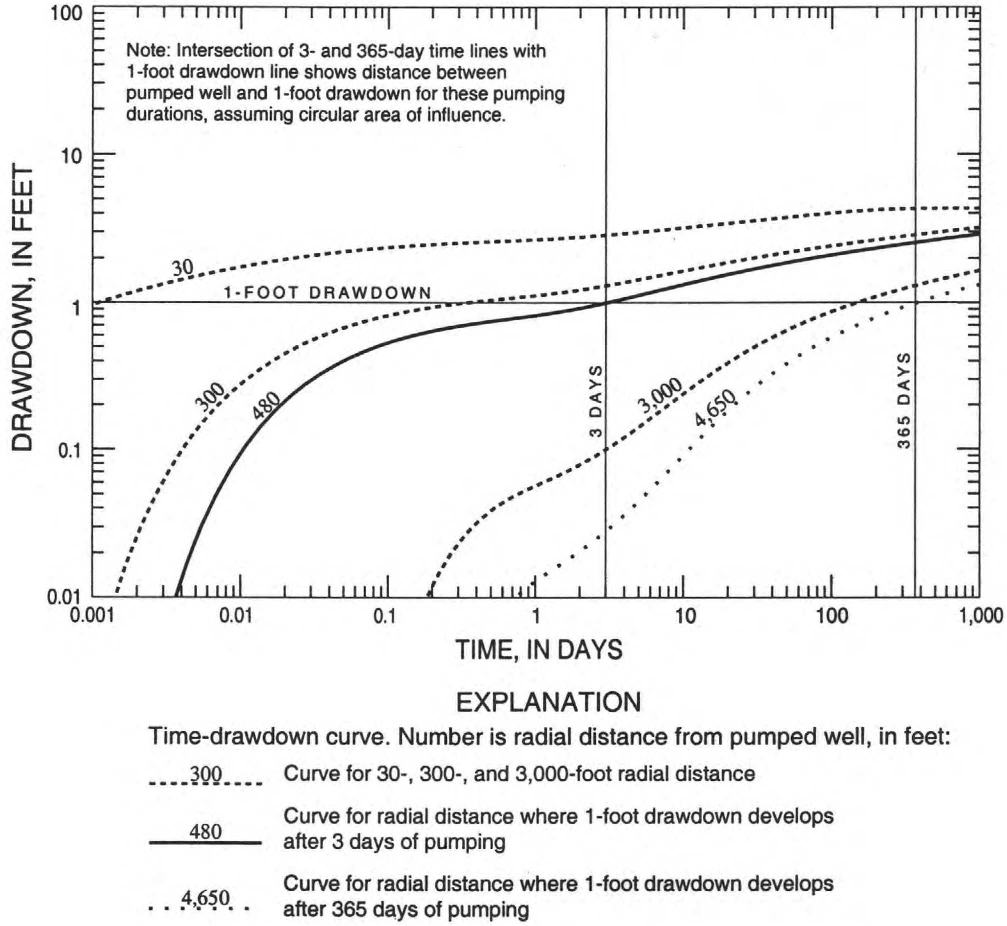


Figure 8. Time-drawdown curves for distances between the pumped and observation wells, derived by use of the Neuman solution for test-well data at Berlin-Cross Keys Road site, Winslow Township, New Jersey.

deRidder, 1992). A transmissivity of 9,677 ft^2/d and a storage coefficient of 0.00058 were obtained from the fit of field data from Obs–2. The calculated transmissivity is nearly 60 percent higher than that derived by use of the Neuman solution (table 2). The storage coefficient is almost 2 orders of magnitude smaller than the specific yield predicted by the Neuman solution. Because the straight-line method would be used to determine aquifer parameters, the specific yield and elastic storage are combined in the estimated storage coefficient. Therefore, the type curves must be determined by using the storage coefficient, an extremely small value. The curves (fig. 9B) indicate that 1 ft of drawdown would develop at 4,181 ft from the well after 3 days of pumping and 46,115 ft from the well after a year of pumping, in contrast with 480 and 4,650 ft for 3 days and 1 year, respectively, as predicted by use of the Neuman solution. Thus, time-drawdown relations calculated using methods based on Theis' equation can lead to grossly inaccurate predictions of the size of the area of influence if values of specific yield used in the calculations are not realistic.

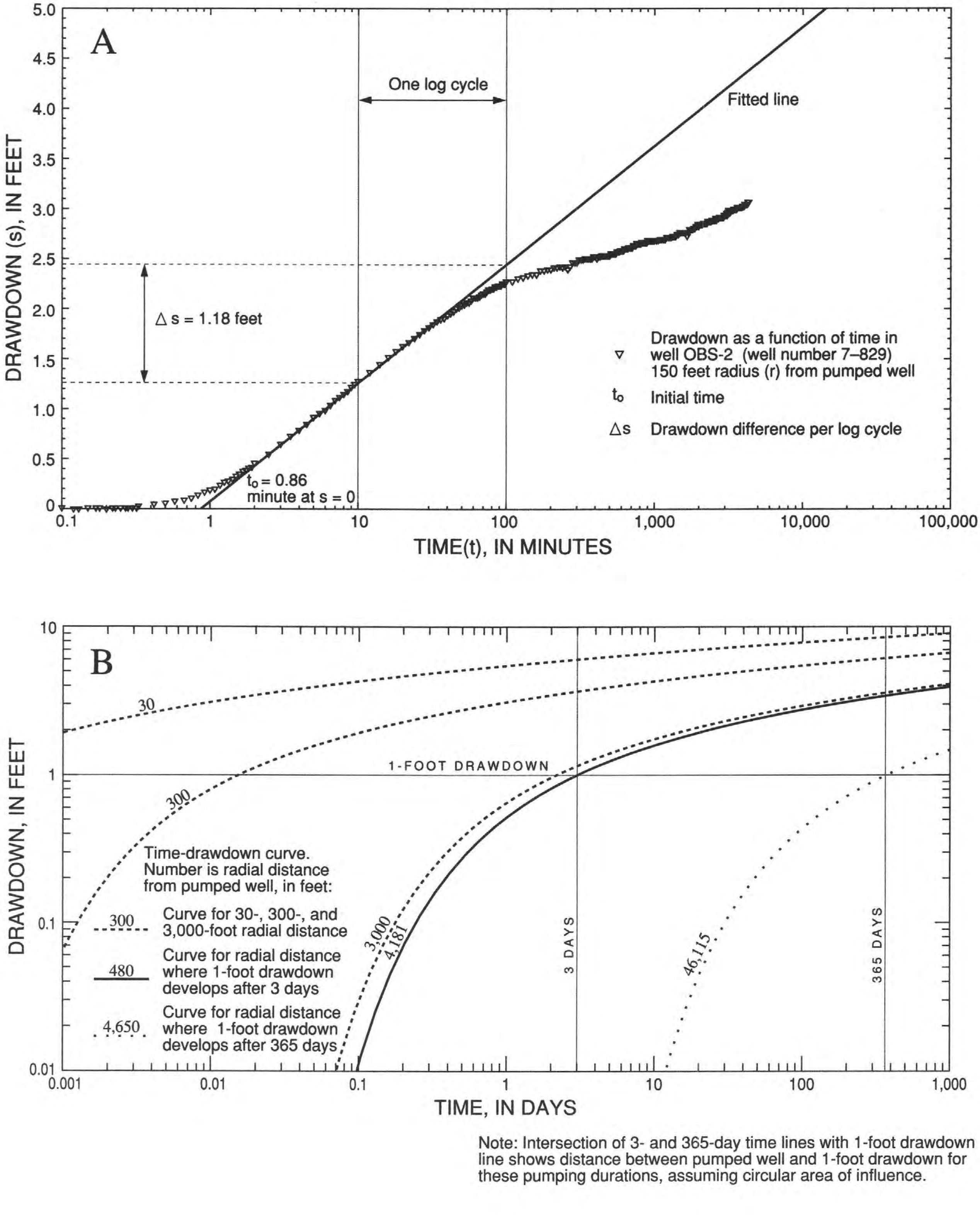


Figure 9. (A) Plot of time-drawdown data and best-fit line using Jacob's straight-line method for the Berlin-Cross Keys Road test well, Winslow Township, New Jersey, and (B) time-drawdown curves derived from the Theis solution.

Table 2. Summary of hydraulic characteristics derived from three analytical methods of aquifer-test-data evaluation for Berlin-Cross Keys Road site, Winslow Township, New Jersey
[T, transmissivity; K_h horizontal hydraulic conductivity; K_v, vertical hydraulic conductivity; S_y, specific yield; S, storage coefficient; ft^2/d, feet squared per day; ft/d, feet per day; ft, feet; --, not applicable]

Method	T (ft^2/d)	K_v (ft/d)	K_h (ft/d)	S_y	S	Radius of influence (ft)
Neuman	14,600	0.48	95.1	0.028	0.0022	480
Jacob's straight-line (time drawdown)	9,677	--	63.0	--	.00058	4,181
Distance-drawdown (Boulton)	6,486	--	42.3	--	.0360	532

Another method that is commonly used to determine the distance between the pumped well and the area of influence at the end of a specified period is referred to as the "distance-drawdown" method (Boulton, 1970). In the method, a single value of drawdown is recorded concurrently in each of the observation wells, usually the final drawdown of the test. A drawdown value for each well and the distances between the observation wells and the pumped well are plotted on semilog paper and fitted with a straight line. The slope of the line is used to determine aquifer characteristics in a manner analogous to the Jacob straight-line method. The drawdown at three observation wells that was recorded at the end of 3 days was plotted according to the distance-drawdown method as shown in figure 10 (data from well Obs–4 were not used). A transmissivity of 6,486 ft^2/d and a storage coefficient of 0.036 were obtained by use of this method. Based on the fitted line, 1ft of drawdown would develop at 532 ft from the well after 3 days of pumping; this radius is nearly an order of magnitude smaller than that obtained by the Jacob method but much closer to the radius predicted by the Neuman solution (table 2).

These results appear to indicate that the distance-drawdown method can predict drawdowns similar to those generated from the Neuman solution for unconfined aquifers, and with less calculation. Boulton (1970) believed that the effect of delayed yield on distance-drawdown curves was unimportant compared with the effect on time-drawdown curves. However, Neuman (1975) cautioned that distance-drawdown curves can give the erroneous impression that they are not affected by delayed yield and are, therefore, amenable to the Theis solution. Neuman warned that the effects of delayed gravity drainage are not fully dissipated from the aquifer even after a few days of withdrawal and that these effects can lead to misleading results if the Theis method is employed. Consequently, caution should be exercised when applying the distance-drawdown method to unconfined systems.

Results from the Neuman and Theis solutions show that the area of influence does not stabilize during withdrawal as long as ground water is derived exclusively from aquifer storage and is not replenished by changes in recharge from boundary conditions. The equations governing ground-water flow to a well (eqs. 1 and 2) predict the development of an increasing area of influence with time (fig. 4), although the rate of propagation would decrease over time as ground-water

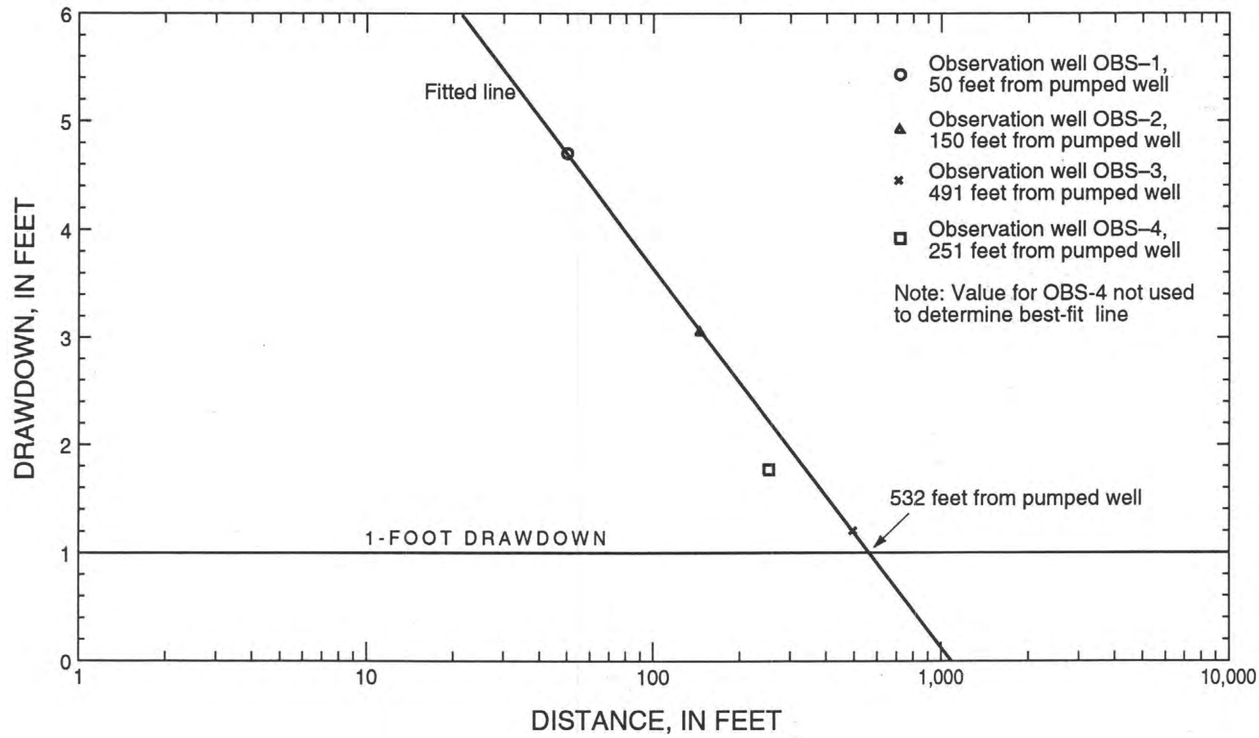


Figure 10. Plot of distance-drawdown for three observation wells at the end of 3 days of withdrawal at Berlin-Cross Keys Road test site, Winslow Township, New Jersey.

storage is drawn from an increasingly larger aquifer volume. Transient analysis of time-drawdown data by methods of Neuman and Theis are best suited for estimating the hydraulic characteristics of an aquifer and for assessing short-term effects of withdrawal close to the well.

Numerical Modeling of Ground-Water Flow Near Headwater Streams in the New Jersey Coastal Plain

In this section, the development and use of a numerical ground-water flow model of the northwest part of the Mullica River Basin are described (fig. 11). The area, henceforth referred to as the “modeled area,” was selected because it typifies an area of residential and commercial development where increased ground-water withdrawal in the future potentially could affect ground-water seepage to an extensive system of streams and wetlands, depending on well location.

Description of Modeled Area

The modeled area encompasses the headwaters of the Sleeper Branch and the Albertson Brook of the Mullica River Basin (fig. 11). Much of the eastern, undeveloped part of the modeled area is within Wharton State Forest. Developed areas are west of Wharton State Forest and include the Waterford Township sewage treatment facility near Chesilhurst.

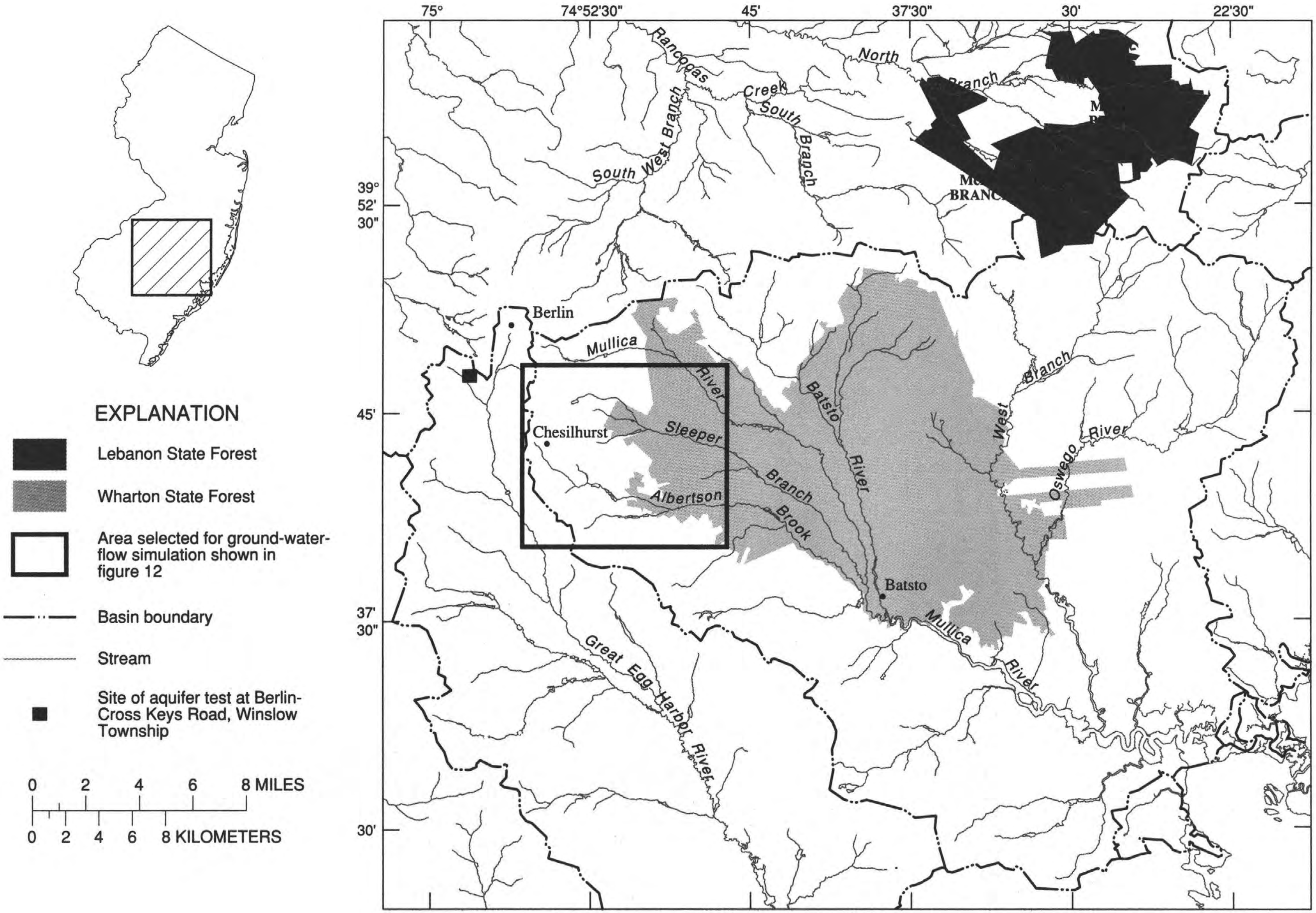


Figure 11. Mullica River Basin and adjacent areas in the New Jersey Coastal Plain.

The Mullica River Basin encompasses part of the Kirkwood-Cohansey aquifer system. The aquifer consists of a layered sequence of unconsolidated marine deposits of Tertiary and Upper Cretaceous age that dip to the east-southeast. The Cohansey sand and Kirkwood Formation compose the Kirkwood-Cohansey aquifer system. The Cohansey sand is predominantly a quartz sand with minor amounts of pebbles, silty and clayey sand, and interbedded clay. The upper layers of the Kirkwood Formation form the lower part of the aquifer system and consist of fine- to medium-grained sand and silty sand. The aquifer system extends from the land surface to a regionally extensive clay at the base of the Kirkwood Formation (Zapecza, 1989).

In this part of the aquifer system, ground water is derived almost entirely from areal recharge. In general, ground water flows laterally in the shallow aquifer from its source area on the land surface to streams and broad, trough-like depressions overlain by wetlands. The direction of ground-water flow is to the east-southeast. Recharge that enters the system at the drainage divide moves deep into the aquifer, then flows laterally beneath the shallow-flow subsystems and eventually exits the system through discharge outlets further downgradient.

Design of Numerical Ground-Water-Flow Model

A three-dimensional ground-water-flow model of unconfined flow was developed for the modeled area by use of the finite-difference computer program MODFLOW (McDonald and Harbaugh, 1988). The hydraulic heads calculated in steady-state simulations were used as initial conditions to calculate drawdowns in transient simulations that involved withdrawal. The model was developed to evaluate the effects that the location of pumped wells can have on ground-water seepage to streams and, conversely, the effects that ground water that normally flows to streams and wetlands can have on the configuration of the area of influence. In addition, withdrawal simulations were used to determine the size of an area of influence produced by long-term ground-water withdrawal under assumed conditions of steady-state areal recharge. Only a numerical flow model can adequately simulate ground-water flow in a system with these complexities.

The model was also used to determine contributing areas of flow to streams and to pumped wells. These results are used to show the relation between the contributing area of flow produced by withdrawal and the configuration of the area of influence. In this report, "contributing area of flow to a stream" refers to an area on the water table in which recharge moves into the aquifer and ultimately seeps into the stream channel. The "contributing area of flow to a well" is similarly defined, except that ground water discharges to a pumped well. Analysis of contributing areas serves to help depict the quantity of ground water that flows to streams under natural conditions and show how withdrawals modify the distribution of ground-water seepage to streams. Contributing areas were generated by use of the particle-tracking computer program MODPATH (Pollock, 1989).

The intended use of the model is to demonstrate hydrologic concepts related to withdrawals in unconfined aquifers near streams and wetlands that are typical of the New Jersey Coastal Plain. The model was not designed to make accurate predictions regarding ground-water budgets, and was not calibrated in the rigorous manner that is required for this purpose. However, results of simulations made with the model can be used to estimate the effects of withdrawals on water levels in the system and show how alternative withdrawal-location strategies can help mitigate the effects of a local withdrawal by distributing these effects throughout the system.

Boundary conditions

The southern, northern, and western boundaries of the model are no-flow boundaries (fig. 12); the surface-water drainage divides are assumed to coincide with ground-water divides. The base of the model also is represented by a no-flow boundary condition because it coincides with the top of the basal Kirkwood, a poorly permeable confining unit. The boundary on the eastern edge of the modeled area is an arbitrary cutoff on the downgradient end of the system and was chosen so that it was far enough away from headwaters of the Sleeper and Pump Branches (where withdrawal scenarios were simulated) so as not to affect simulation results. This boundary was assigned constant values of hydraulic head that were estimated from water-table measurements. The top surface of the model was assigned a specified flux of 0.0035 ft/d (15.33 in/yr). It represents a steady-state, net recharge rate and was chosen by estimating the amount of recharge needed to generate the base flow to streams determined from streamflow measurements at low-flow gaging stations in the modeled area.

The hydraulic interaction between stream and aquifer was simulated by use of a head-dependent flux condition in model cells that overlie the streams and wetlands in the area (fig. 12). Under these conditions, ground water seeps into the streambed in proportion to the difference between hydraulic heads in the aquifer and the streambed elevation. When hydraulic head in the aquifer is higher than the streambed elevation, ground water seeps into the stream. This condition is characteristic of gaining streams and is common in Coastal Plain drainage basins. In places where hydraulic head is lower than the streambed, such as upstream from the start of flow, ground water cannot seep into the stream. These relations are represented by the equations

$$q_s = C_s(h_{aq} - h_s), \; h_{aq} > h_s \quad \text{(gaining stream)} \tag{4a}$$

and

$$q_s = 0, \; h_{aq} \leq h_s \quad \text{(no seepage to stream),} \tag{4b}$$

where

q_s is the ground-water seepage into the stream channel (L^3/T),
C_s is the coefficient of proportionality, termed the "streambed conductance" (L/T),
h_{aq} is the hydraulic head in the aquifer under the streambed (L), and
h_s is the streambed elevation (L).

The streambed conductance, C_s, is a hydraulic parameter that is used in the model to simulate impedance to ground-water seepage. Sediments in the streambed, ground-water flow that converges on the stream channel, and the water-transmitting properties of the aquifer beneath the stream are factors that collectively create resistance to ground-water seepage, the composite effects of which are represented by C_s. In general, increasingly higher streambed conductances were assigned to stream cells with distance downstream; values were determined by trial and error.

Model discretization

Model cells are variably spaced and range in size from 1,500 ft^2 near the boundaries of the model to 300 ft^2 in much of the interior. A plan view of the discretized system is shown in figure 12. Discretization was finer in the central part of the model near headwaters of the Sleeper and Pump Branches, where withdrawal simulations were planned. Smaller cells are needed to simulate pumped wells, and fine discretization is needed around cells that simulate pumpage to better estimate axisymmetric ground-water flow.

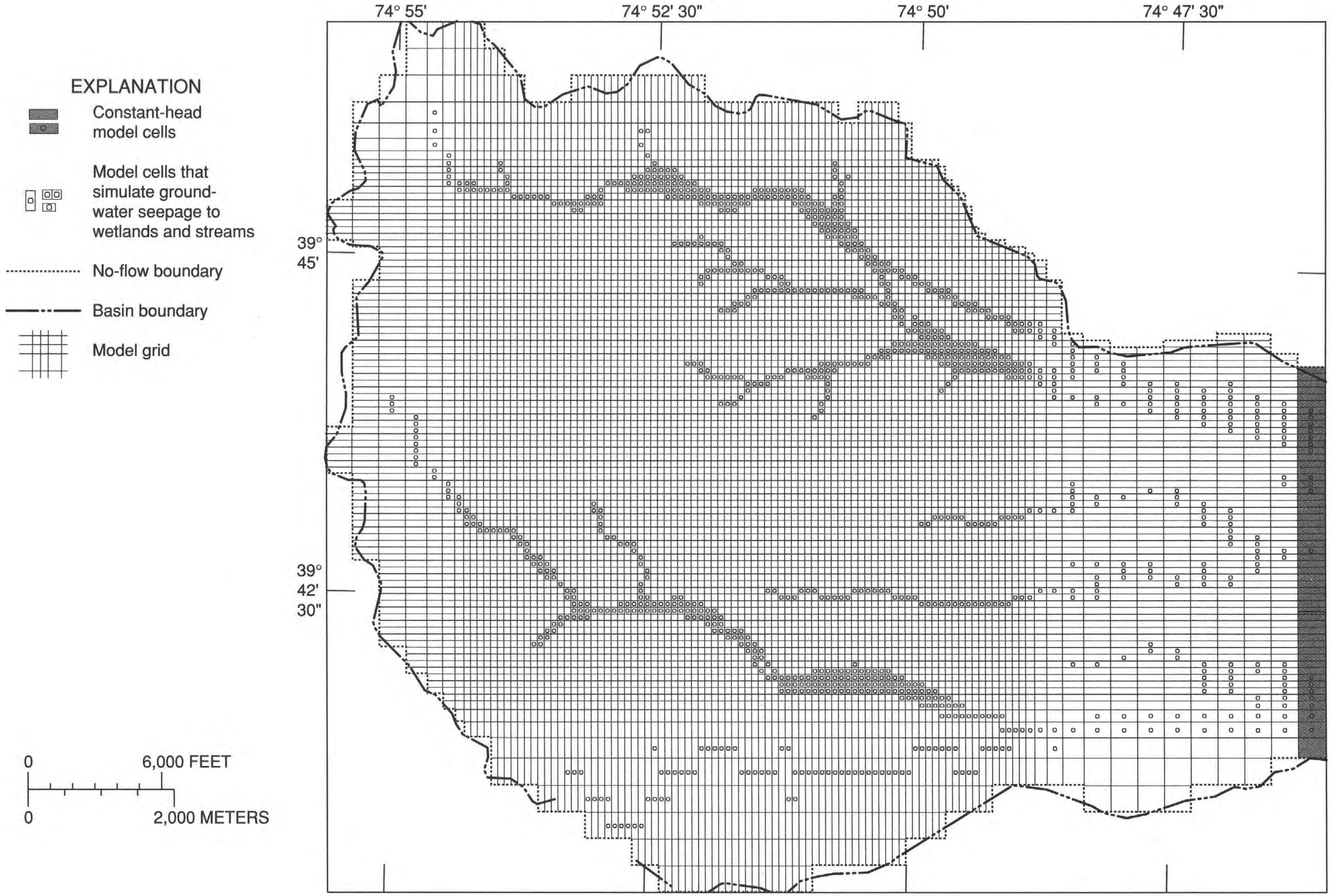


Figure 12. Horizontal model discretization of aquifer and stream-wetland system, Mullica River Basin, New Jersey.

The aquifer system was divided into three layers. The model layers are used to simulate the vertical flow of water in a single unconfined aquifer; they do not represent different hydrologic units. Multiple model layers also allow for the assignment of contrasting horizontal and vertical hydraulic conductivities. In withdrawal simulations discussed later in this report, it was assumed that wells had 40-ft-long screens that were set 60 to 80 ft below the water table. This condition was accommodated by assigning the middle layer a thickness of 40 ft. Because the model simulated flow conditions in an unconfined system, the position of the water table was calculated by MODFLOW as part of the flow solution. Total saturated thickness of the simulated aquifer ranges from 160 to 200 ft.

Hydraulic parameters

Data on hydraulic conductivity of the aquifer in the modeled area are sparse. Horizontal conductivities used for the model range from 100 ft/d in the southwestern part of the model to 65 ft/d in the northeastern part. These values are constrained, in part, by results of aquifer tests in the region. A transmissivity of 20,000 ft^2/d was derived from an aquifer test near Batsto, N.J. (fig. 11), east of the modeled area (Lang and Rhodehamel, 1963). Because the saturated thickness of the aquifer is about 200 ft near the site of the test, the average horizontal hydraulic conductivity is 100 ft/d. Lang and Rhodehamel (1963) also calculated a specific yield of 0.16. Aquifer-test data near Berlin-Cross Keys Road in Winslow Township, just out of the northwest corner of the modeled area (fig.11), indicate a hydraulic conductivity of 95.1ft/d and a specific yield of 0.028.

Calibration

Water levels were measured during November 1992 – December 1993 in eight observation wells (fig. 13) screened in the Cohansey sand. The observation wells are widely scattered throughout the modeled area in undeveloped forest and wetland areas where access is limited. A rigorous calibration criterion for hydraulic head was not specified except that simulated hydraulic heads be within 10 ft of measured water levels. Although depth to water from land surface was measured accurately to within 0.1 ft, the water-level elevations could only be estimated because benchmark elevations were not available for the wells. A list of measured water levels and simulated heads for corresponding cells are listed in table 3. The results show a reasonable match; all simulated hydraulic heads met the specified calibration criterion. The absolute mean error of the difference between simulated heads and estimated water levels is 3.5 ft.

In addition, the resemblance between the configuration of the simulated and interpreted water tables was expected to be reasonably close. To enable this comparison, a water-table map of the modeled area was constructed (fig. 13). In areas adjacent to streams and wetlands, the water table is assumed to be close to the elevation of the streambed–generally a valid assumption where the hydraulic conductivity between the stream and the shallow aquifer is relatively high. The interpreted and simulated water tables are shown in figure 13. In general, the match is reasonable; the two sets of contour lines are close. The contour lines tend to match less well near the southern and western no-flow boundaries of the model, an indication that the assumption that no-flow boundaries coincide with surface drainage divides is not necessarily valid in these areas. Consequently, simulated results near model boundaries must be viewed with caution.

Simulated base flow to streams was compared to the averaged stream discharge measurements made at eight low-flow streamflow stations (fig. 13) for water year 1993. Stream measurements were made approximately six times during the year. Because of the good control on stream discharge, the calibration criterion for discharge was made more stringent than for hydraulic heads;

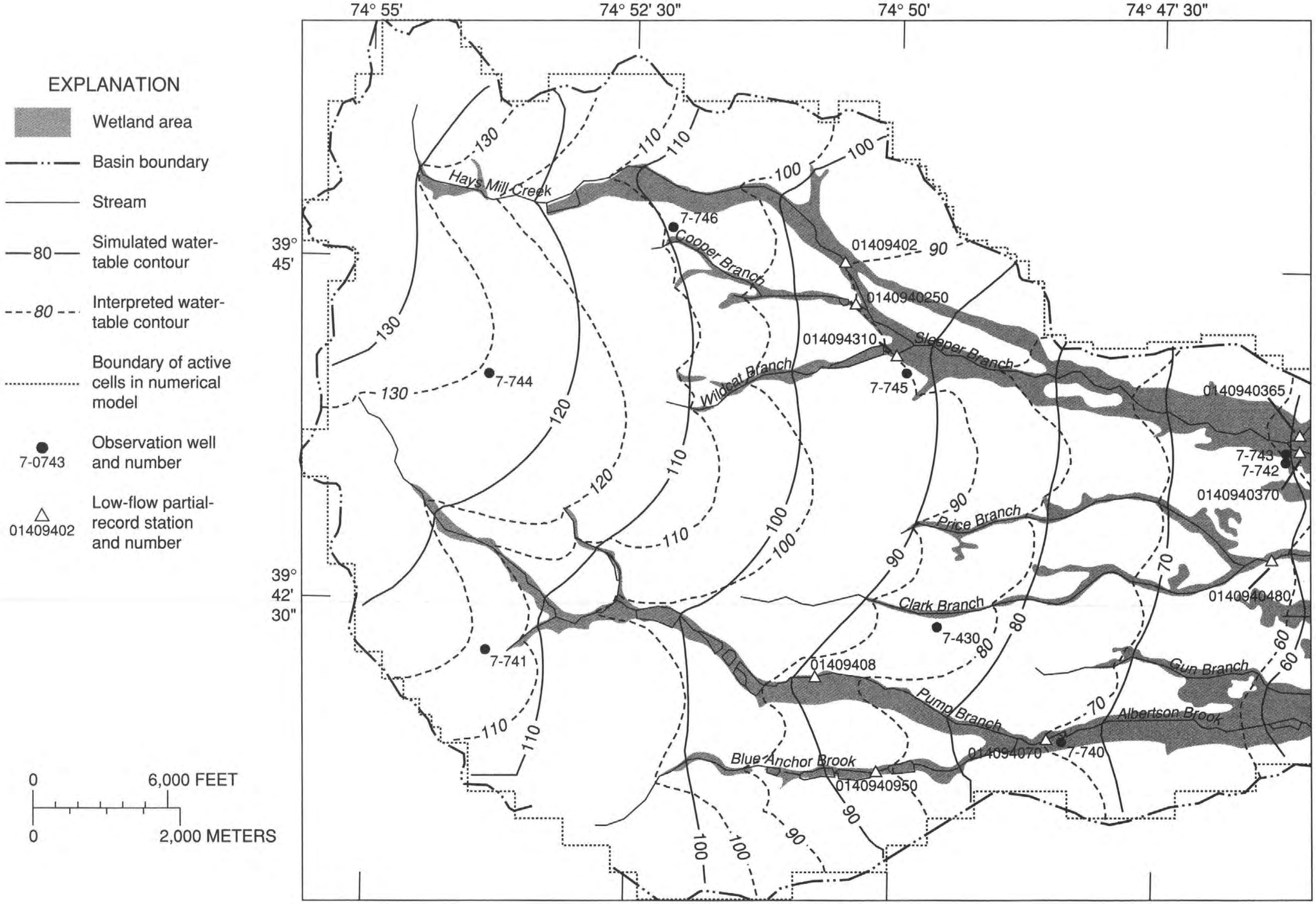


Figure 13. Locations of low-flow gaging stations, observation wells, stream-wetland systems, and simulated and interpreted water-table contours in the northwestern part of the Mullica River Basin, New Jersey.

Table 3. Average measured water levels and simulated hydraulic heads for corresponding cells in ground-water-flow model of Mullica River Basin, New Jersey

Well number	Measured depth to water[a] (feet)	Estimated altitude of water[b] table (feet)	Simulated water level[c] (feet)	Difference[d]
07–0744	28.7	128.5	122.8	5.7
07–0746	10.0	106.3	107.9	-1.6
07–0741	15.2	117.1	112.1	5.0
07–0745	10.9	81.9	89.7	-7.8
07–0430	14.9	81.6	83.5	-1.9
07–0740	- .1[e]	67.1	70.5	-3.4
07–0743	2.0	61.4	59.5	1.9
07–0742	2.8	60.6	60.0	.6

a. Measurements made from land surface between November 1992 and December 1993.
b. Estimated elevation of measuring point minus measured depth to water; datum is sea level.
c. Based on weighed values of the nearest surrounding model cells; datum is sea level.
d. Mean absolute error = 3.5.
e. Negative number indicates water level above land surface.

specifically, simulated stream discharge had to match measured stream discharge to within 5 percent. A list of simulated and measured flows is given in table 4. Simulated stream discharge for all but one station easily met the specified calibration criterion. The simulated base flow for Wildcat Branch was only 0.11 ft^3/s less than the measured value and matched the measured value to within 6.4 percent, still a reasonably good match.

Hydraulic conductivities and stream conductances were adjusted by trial and error until simulated hydraulic heads and base flow were within the calibration criteria. No constraints were placed on streambed conductance because it is a composite parameter. However, upper and lower limits were placed on hydraulic conductivities to keep the model from becoming an unrealistic representation of the aquifer system. According to Rhodehamel (1973), the maximum horizontal hydraulic conductivity of the Cohansey sand is 150 ft/d. Harbaugh and Tilley (1984) used 15 ft/d as a lower limiting value for the Mullica River Basin to account for lower conductivity material in the aquifer. Hydraulic conductivities used for the ground-water-flow model fell well within these limiting values.

Table 4. Measured streamflow at low-flow gaging stations and simulated ground-water seepage to streams, Mullica River Basin, New Jersey
[ft^3/s, cubic feet per second]

Low-flow gaging station[a] name and site number	Average measured streamflow[b] (ft^3/s)	Simulated ground-water seepage (ft^3/s)	Difference[c] (ft^3/s)	Difference / average measured streamflow (percent)
Hays Mill 01409402	8.02	7.95	0.07	0.9
Cooper Branch 0140940250	1.06	1.05	.01	.9
Wildcat Branch 014094310	1.73	1.62	.11	6.4
Sleeper Branch 0140940365	6.53 [d]	6.34	.19	2.9
Clark Branch 0140940480	2.46	2.43	.03	1.2
Pump Branch 01409408	7.86	8.16	- .30	3.8
Blue Anchor Branch 0140940950	2.00	1.98	.02	1.0
Albertson Brook 0140940970	6.37 [e]	6.46	- .09	1.4

a. Location of gaging stations shown in figure 13.
b. Measurements made between November 1992 and December 1993.
c. Mean absolute error = 0.10.
d. Does not include streamflow measured at Wildcat Branch, Cooper Branch, and Hays Mills Creek gaging stations.
e. Does not include streamflow measured at Blue Anchor Brook and Pump Branch gaging stations.

Ground-Water Flow to Streams and Wetlands

The contributing areas of flow to streams and the water-table configuration of the simulated flow system are shown in figure 14. The contributing areas represent source areas of recharge that flows to segments of the stream network. The term "segment" is defined as a reach of stream between two gaging stations or between the start of flow and the closest downstream gaging station. In general, the shape of the contributing areas tapers in the upgradient direction, to the west. The long axis of a contributing area is always normal to lines of equal hydraulic head. The water-table gradient indicates that ground water moves from west to east. Contributing areas of flow to the upstream segments are totally or partly surrounded by the contributing areas of flow to stream segments further downstream. This pattern indicates that some recharge captured near the headwaters of the drainage basin discharges to stream reaches that are relatively far down the drainage basin. The contributing areas of ground-water flow to the constant-head boundary on the east side of the model represent contributing areas of flow to streams further downgradient, outside the boundary of the model. Because all incoming water in the flow model originates as areal recharge at the water table, the size of each contributing area divided by the total plan area of the model indicates the fraction of total recharge that discharges to a stream segment. Consequently, contributing areas provide a means of visualizing the ground-water budget for the modeled system.

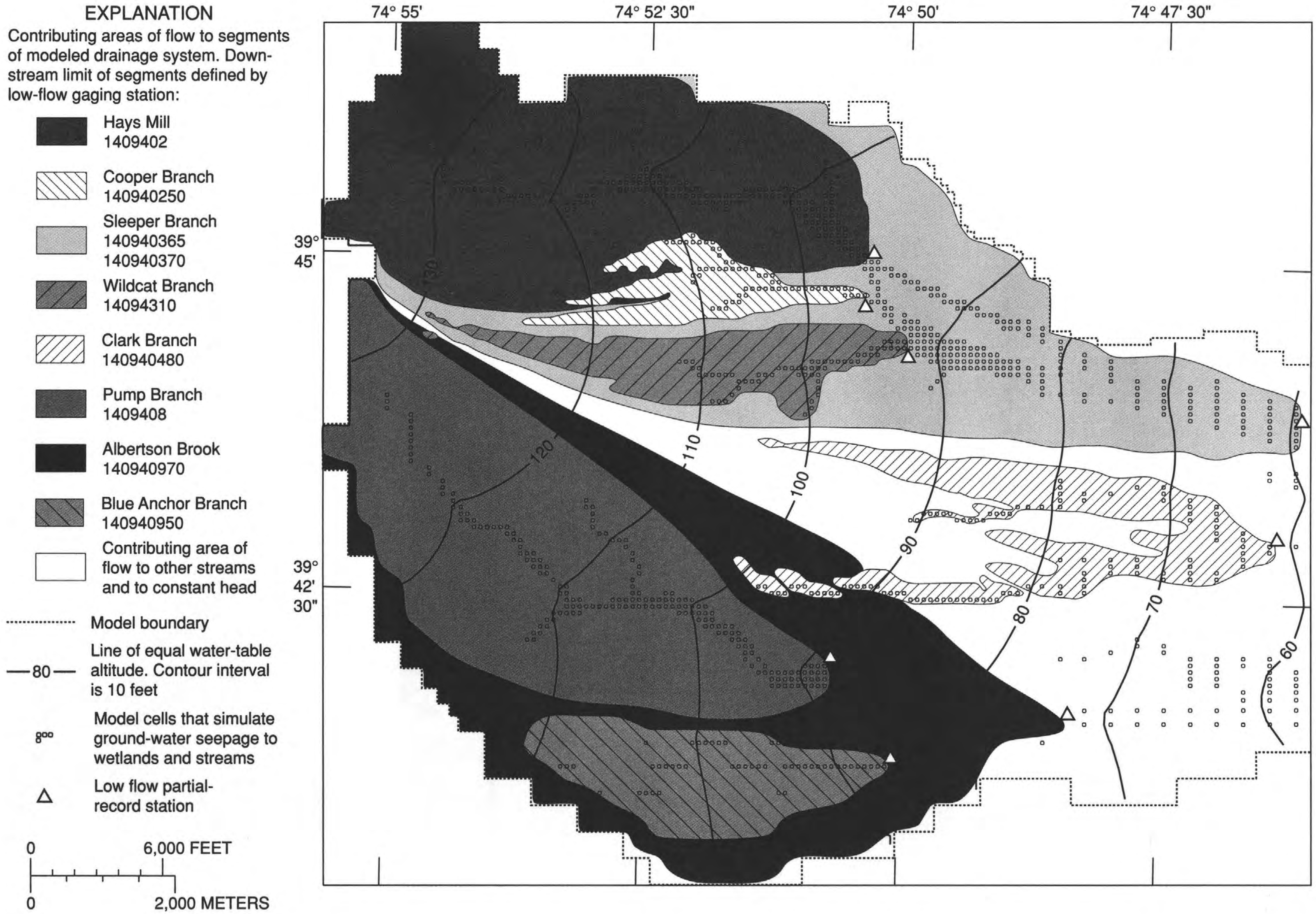


Figure 14. Simulated contributing areas of flow to stream-wetland system and configuration of water table under steady-state, natural conditions, Mullica River Basin, New Jersey.

For comparison with the preceding description of natural conditions, consider a scenario in which a pumped well is placed near the center of the modeled area and is pumped at a rate of 1.6 Mgal/d under steady-state conditions. The resulting configuration of the contributing area of flow to streams and to the well are shown in figure 15. The tapered end of the contributing area of flow to the well extends to the west, upgradient from the well, to the boundary of the system. The contributing area of flow to the well includes areas on the water table that would, under natural conditions, be incorporated into the contributing areas of flow to streams. For example, the upstream end of the contributing area of flow to Clark Branch (fig. 12) is truncated by the contributing area of flow to a well. In addition, the contributing area of flow to Albertson Brook that develops under natural conditions is reduced in size under withdrawal conditions. Because recharge equals discharge in this steady-state ground-water-flow system, ground water discharges to the pumped well at the expense of seepage to streams. Consequently, the withdrawal must result in a reduction in the size of the contributing areas to streams by an amount equal to the contributing area of the withdrawal.

Similarly, ground-water seepage to streams is reduced by an amount that is equivalent to the pumpage. The overall distribution of the reduction in seepage to streams and wetlands that results from ground-water withdrawal in the modeled system is shown in figure 16. The reductions in ground-water seepage for each stream segment that is attributed to withdrawal are listed in table 5. The total simulated reduction is 2.46 ft^3/s and is approximately equal to the pumping rate of 2.48 ft^3/s (1.6 Mgal/d). In general, the amount of reduction of ground-water seepage to streams increases with the proximity of the well to a stream. Streams that border on or that are within the area of influence of a pumped well (fig. 16) lose most of the ground-water seepage to the well. However, these streams are minor, headwaters streams characterized by low rates of seepage under natural conditions compared to larger tributaries downstream. Therefore, reductions in seepage to major streams are actually small or negligible.

The lines of equal drawdown shown in figures 15 and 16 represent a steady-state cone of depression on the water table that results from constant, long-term withdrawal. Under steady-state conditions, recharge and discharge are in equilibrium, and ground-water storage does not change. Although figures 15 and 16 show only the drawdown that is equal to or greater than 1 ft, the water table is everywhere decreased slightly from its natural level, even near the boundaries. The configuration of the area of influence is affected predominantly by the water-table gradient, pumping rate, areal recharge, and the relative location of streams. Water-table gradients indicate the magnitude of ground-water flow in the direction of decreasing hydraulic head. Pumpage in such a flow field results in an area of influence that is elongated in the direction of flow and shorter on the downgradient side of the well than on the upgradient side. The maximum depth of drawdown is 13.9 ft. The area of influence extends over the headwaters of the Wildcat and Clark Branches (fig. 16). The lines of equal drawdown are inflected toward the well where they cross the stream, an indication that streams affect the area of influence. Although stream stages are not simulated, they would also decrease in response to the reduced ground-water seepage to streams in a system in which long-term withdrawals are made.

The contributing area of flow to a pumped well and its area of influence are different hydraulic phenomena that do not necessarily affect the same aquifer space (fig. 15). This is particularly true in systems with regional flow gradients, where the area of influence may only partly coincide with the contributing area. The area of influence is a measure of ground water displaced from the aquifer by pumpage. The contributing area of flow to the well is an area on the water table in which recharge moves into the ground-water-flow system and eventually discharges to the well.

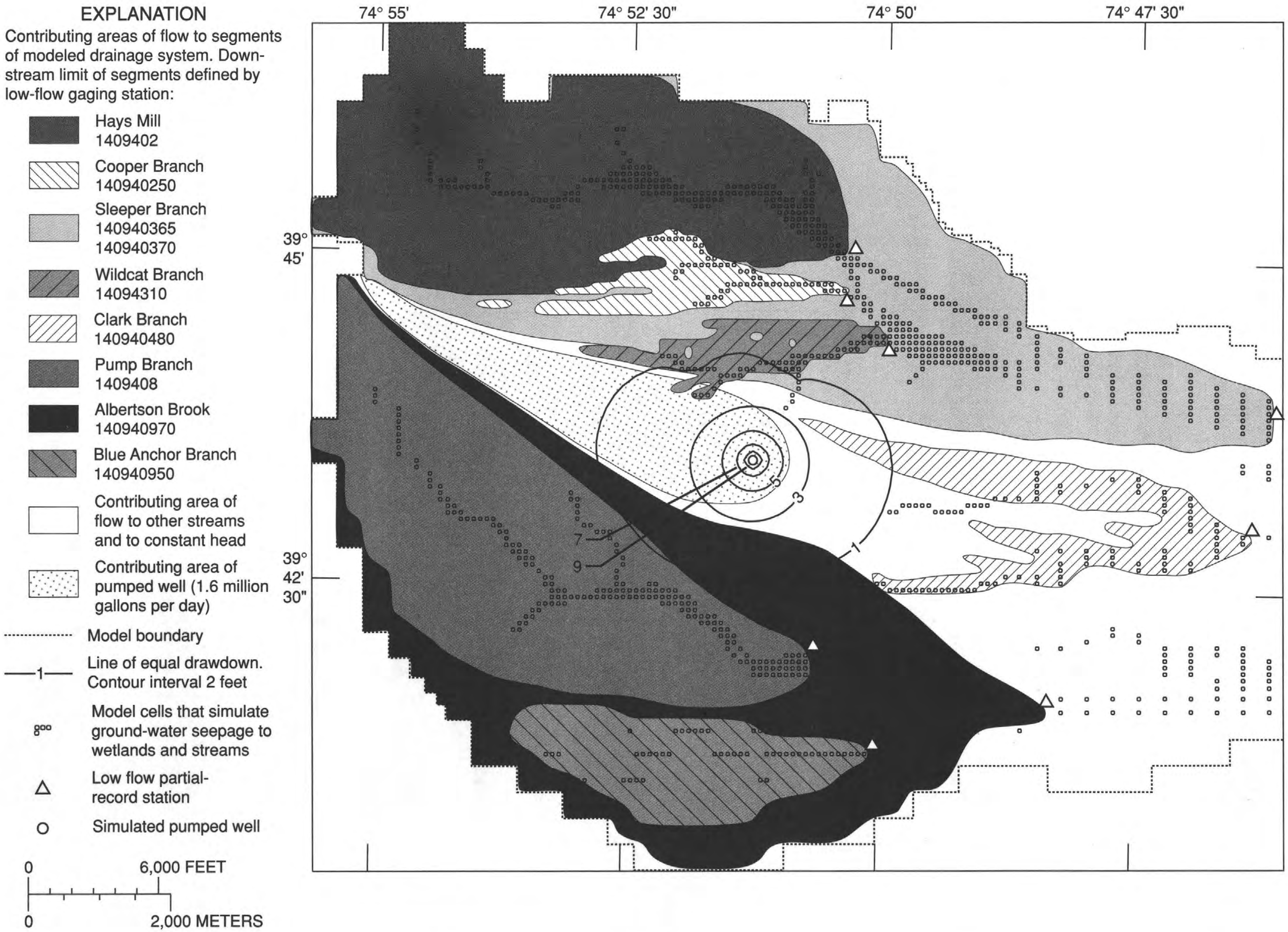


Figure 15. Simulated contributing areas of flow to stream-wetland system and to a pumped well under steady-state conditions, Mullica River Basin, New Jersey. (Contour lines show water-table drawdown from initial, unstressed level.)

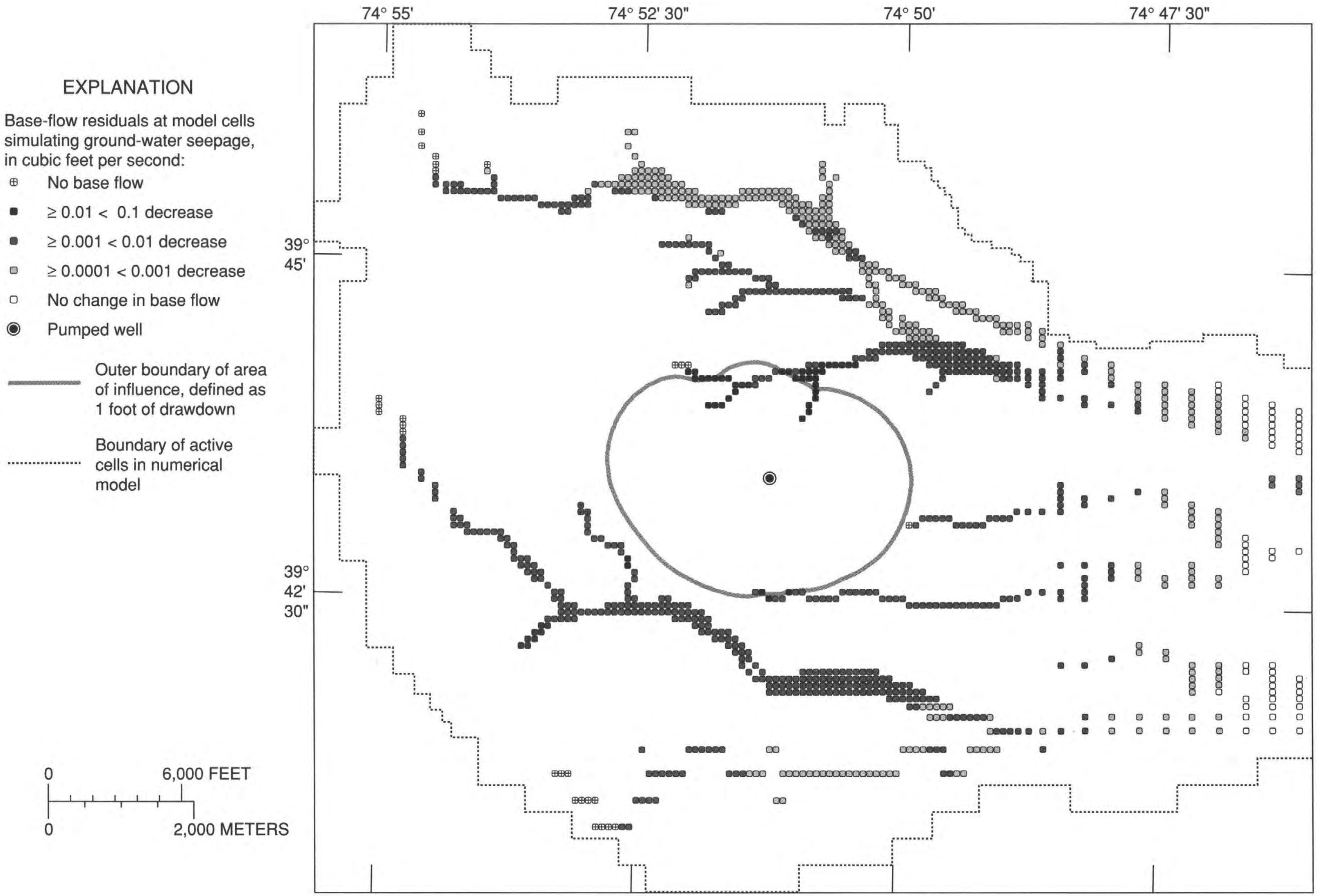


Figure 16. Residuals of ground-water seepage to streams for ground-water flow simulation of modeled system for steady-state, unstressed conditions and steady-state withdrawal conditions, Mullica River Basin, New Jersey. (Pumping rate is 1.6 million gallons per day.)

Table 5. Simulated ground-water seepage to streams and wetlands under unstressed and withdrawal conditions, Mullica River Basin, New Jersey
[ft^3/s, cubic feet per second]

Simulated streams	Stream discharge under unstressed conditions (ft^3/s)	Stream discharge under withdrawal conditions[a] (ft^3/s)	Residual stream seepage[b] (ft^3/s)	Seepage reduction (in percent)
Hays Mill	7.95	7.80	0.15	1.9
Cooper Branch	1.05	.92	.13	12.2
Wildcat Branch	1.62	.89	.73	45.2
Sleeper Branch	6.34	6.10	.24	3.7
Clark Branch	2.43	2.05	.38	15.6
Pump Branch	8.16	7.54	.62	7.6
Blue Anchor Branch	1.98	1.93	.05	2.7
Albertson Brook	6.46	6.32	.14	2.2
Other simulated streams	6.00	5.92	.02	.3

a. Simulated withdrawal rate 2.48 ft^3/s (1.6 million gallons per day).
b. Total residual ground-water seepage to streams = 2.46 ft^3/s.

A contributing area of flow is calculated from ground-water velocities and is not directly measurable as is drawdown (Javandel, 1986). The ramification of this is that all recharge that falls within the area of influence does not necessarily flow to the well even though drawdown gradients are directed toward the well.

Effects of Aquifer Storage and Variations in Pumpage on the Area of Influence

Consider the effect of 3 days of withdrawal on the distribution of ground water near the well. The simulated area of influence that results from a 3-day withdrawal is shown in figure 17A. The well location and pumping rates are the same as those described for the steady-state withdrawal simulation. For the transient simulation, areal recharge was not simulated for the 3-day period; a specific yield of 0.15 and storage coefficient of 0.001 were assigned to the aquifer. Because the modeled system is unconfined, the flow simulation was more sensitive to specific yield than to the storage coefficient. The resulting area of influence (fig. 17A) has a radius of about 560 ft and does not impinge on nearby tributaries. The area of influence that was generated under steady-state conditions (fig. 15) is reproduced in figure 17B to facilitate comparison. The distance between the pumped well and the southern boundary of the area of influence is about 5,200 ft; the distance between the well and the downgradient boundary to the east is about 6,200 ft, and between the well and the upgradient boundary to the west, about 7,400 ft. The size of the area of influence produced by steady-state pumpage is nearly an order of magnitude larger than that resulting from a 3-day withdrawal.

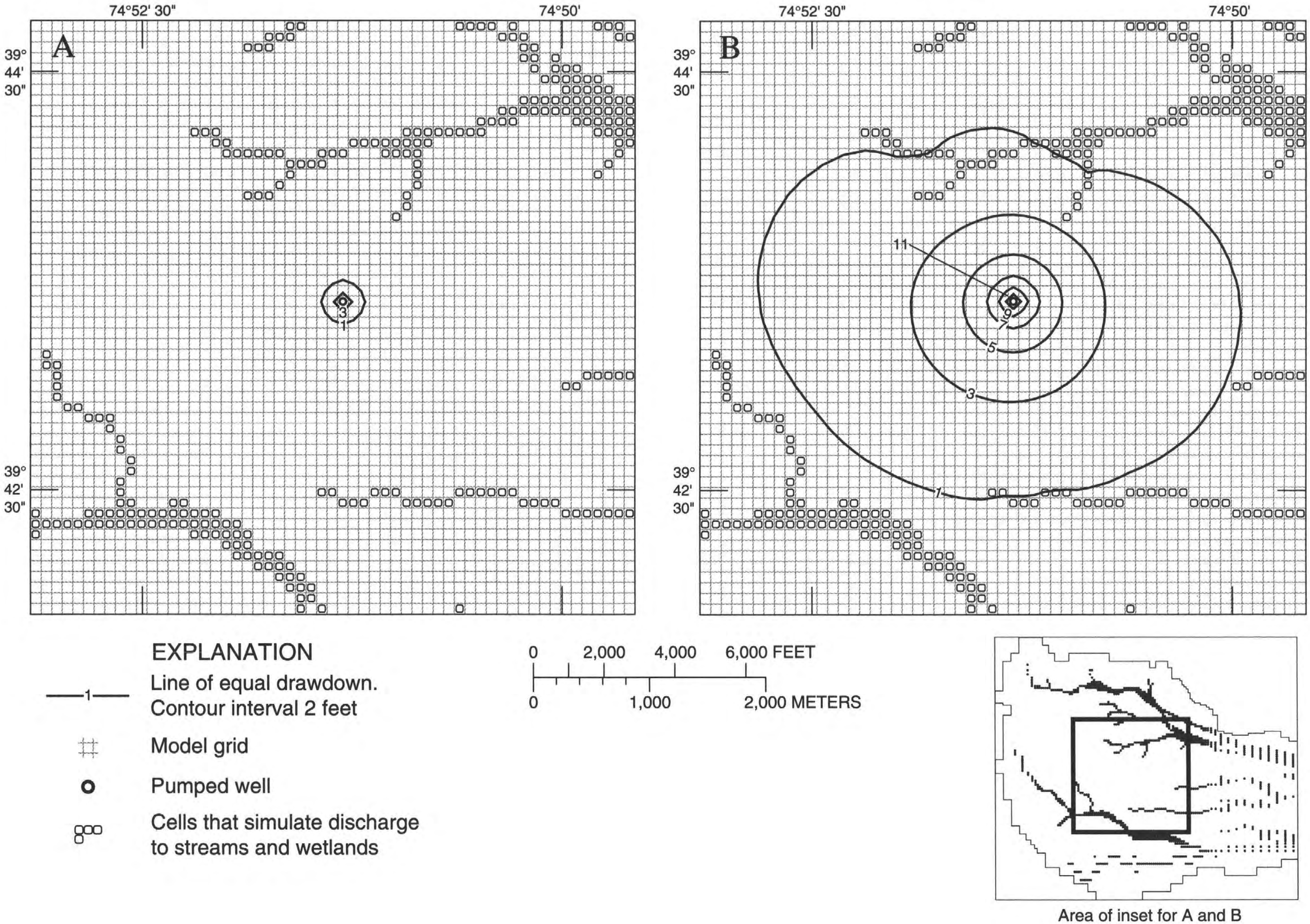


Figure 17. Subarea of simulated aquifer system, Mullica River Basin, New Jersey, after (A) 3 days of ground-water withdrawal, and (B) during steady-state withdrawal. (Area of influence is defined by 1-foot drawdown contour. Pumping rate is 1.6 million gallons per day for each simulation.)

As a means of independently verifying the time-drawdown relations predicted by the numerical model, the parameters used to simulate withdrawal in the numerical model were evaluated with the Neuman solution. A plot of simulated time-drawdown from the numerical solution and the type curves from the Neuman solution are shown in figure 18. For the Neuman solution, two observation wells were assumed to be 300 ft from the pumped well, one completed at the water table and another at 130 ft below the water table. The simulated hydraulic heads at model cells in layers 1 and 2 on the right side of the cell that simulates withdrawal in figure 17 were used to compare the analytical results of the Neuman solution.

In general, the numerical-model results are consistent with predictions made by the Neuman solution. This is expected because the numerical model is designed to simulate ground-water flow under unconfined conditions. The model results are sensitive to values of specific yield but are relatively insensitive to values of the storage coefficient. For times greater than 0.08 day (115 minutes) but less than 30 days, numerical results compare well with the curve generated from the Neuman solution at the water table. Numerical results tend to underpredict drawdown for early times near the bottom of the aquifer. After 30 days, the simulated drawdown of the numerical model is affected by the proximity of no-flow boundaries of the model to the well and is greater than the drawdown from the Neuman solution. The model boundary effect indicates that the effects of withdrawal after 30 days would extend beyond the modeled domain and into adjacent basins. Thus, the limits of the model are reached in approximately 30 days given the pumping conditions of this simulation.

The size of the area of influence is affected, to a large degree, by pumping rate, pumping duration, and storage properties of the aquifer. To get a sense of how these factors work to control the area of influence, 21 ground-water flow simulations of withdrawal from a single well were made. The simulations incorporated seven pumping rates that increased from 0.4 to 1.6 Mgal/d by increments of 0.2 Mgal/d. In seven simulations, the aquifer was assigned a specific yield of 0.15. In another seven simulations, the seven pumping rates were repeated, but the specific yield of the aquifer was reduced to 0.075. An additional seven simulations were made under steady-state conditions using the pumping rates previously specified and with the areal-recharge rate used under natural conditions (fig. 14). The location of the pumped well is the same as that shown in figures 15 and 16.

The model results are summarized in a plot (fig. 19) that shows the relation between the size of the area of influence and pumping rate for each of the specific-yield values used in the simulations for 3 days, 14 days, and steady-state conditions. Because the shapes of the areas of influence are ellipsoidal, the average distance between the well and the 1-ft-drawdown boundary is used to indicate the size of the area of influence and, for convenience, is referred to as the "radius."

The plots indicate that withdrawals that last for 3 days result in considerably smaller areas of influence than those produced under steady-state conditions. As pumping rates increase, the contrast in size of the area of influence between transient, short-term withdrawal and steady-state, long-term withdrawal increases significantly. However, the radii approach an asymptotic limit with increasing pumping rate. A comparison of the areas of influence generated in 3 days to those generated under steady-state conditions shows that the sizes differ by nearly an order of magnitude. For example, for a pumping rate of 0.6 Mgal/d, the radius of the area of influence is 300 ft after 3 days of withdrawal but increases to about 2,900 ft under steady-state withdrawal. Similarly, for a pumping rate of 1.2 Mgal/d, the radius is 500 ft after 3 days of withdrawal but increases to about 4,800 ft under steady-state withdrawal.

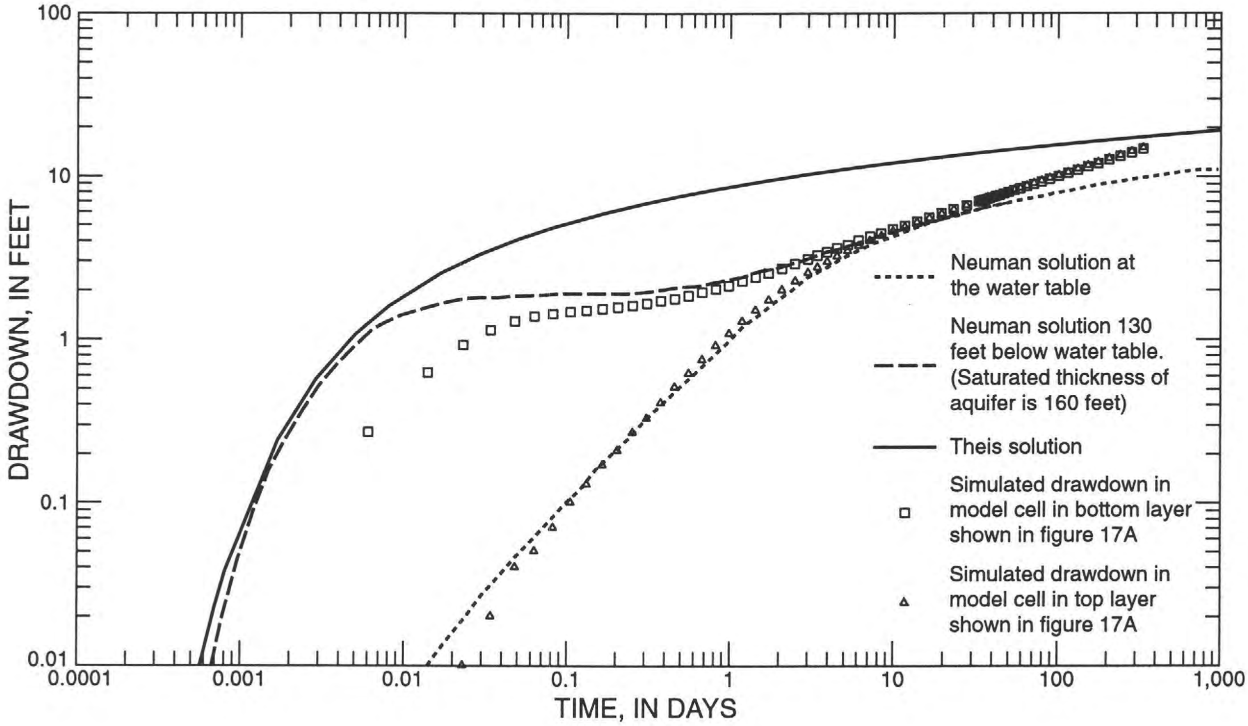


Figure 18. Time-drawdown relations derived from the Neuman solution and the numerical solution as observed 300 feet from the pumped well, at the water table, and near the bottom of the aquifer.

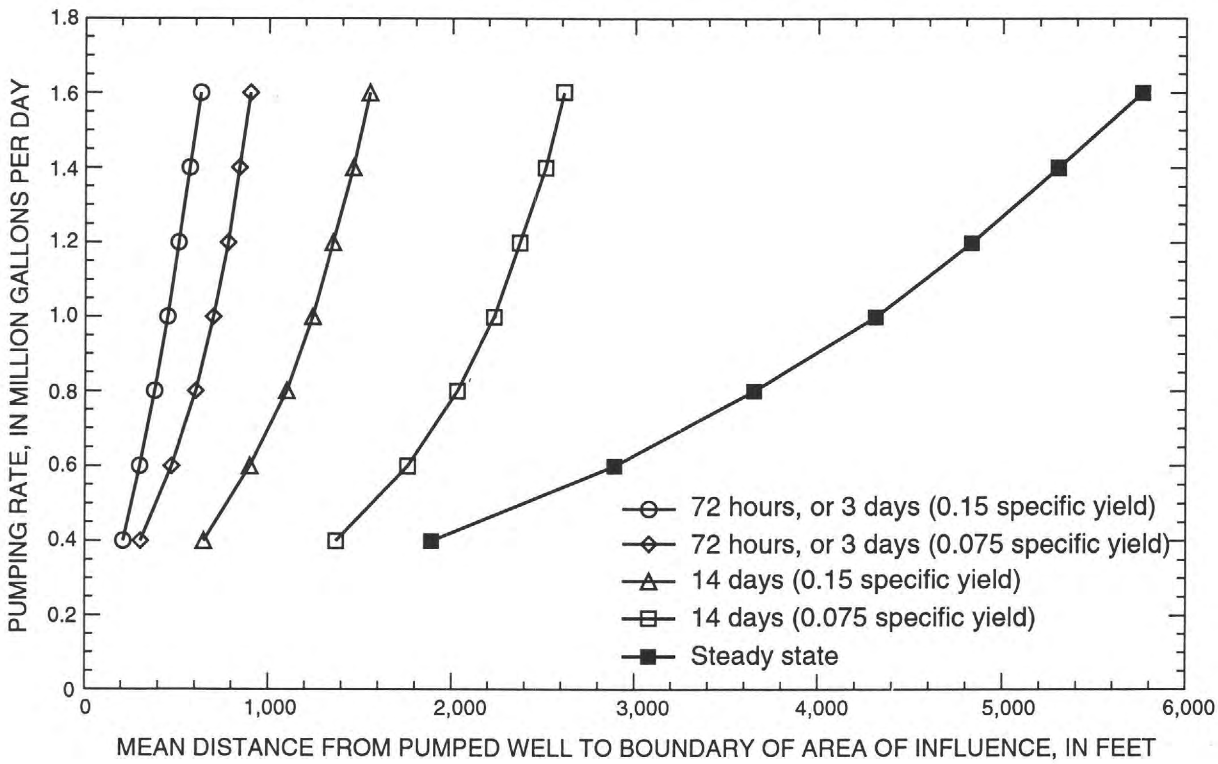


Figure 19. Mean distance between pumped well and area of influence for simulated pumpage for various pumping rates at 3 days, 14 days, and steady state.

The specific yield of the aquifer also affects the size of the area of influence. The plots of area sizes that are produced after 3 days and 14 days indicate that the areas of influence were larger in simulations that used the smaller value of specific yield. Aquifers with small specific yields have small storage capacity; consequently, ground-water withdrawal forces the area of influence to propagate relatively far out from the pumped well. Because specific yield varies considerably in real aquifer systems, the area of influence that results from pumpage in these systems also must vary.

Effects of Pumped-Well Location on the Area of Influence

When a pumped well is near gaining streams and wetlands, the withdrawal can cause ground water that would normally flow to streams to flow to wells instead and effectively create a recharge boundary between the well and the stream. In this report, the component of ground water that is rerouted by withdrawal from a stream to the well is referred to as "diverted" ground water. The diverted ground water tends to reduce the size of the area of influence compared to the area of influence produced by a well far from a stream. Time-drawdown data from wells near streams, if evaluated by the analytical methods discussed previously, would lead to incorrect estimates of aquifer parameters and propagation rates of the area of influence because these analyses are predicated on the assumption that the aquifer is boundless. Consequently, the analytical methods cannot evaluate the effects of the recharge boundary created by streams. (Analytical solutions to axisymmetric flow problems that account for the effects of recharge boundaries on well withdrawal are available, but their application is restricted because of other limiting assumptions; see Ferris and others, 1962). Yet, an understanding of the effects of ground-water diversion is an important hydrologic factor in withdrawals made in the New Jersey Coastal Plain because of the numerous streams, wetlands, and impoundments in the region.

Pumped wells near streams and other water bodies can also induce streamwater infiltration; that is, cause streamwater to recharge the aquifer along the channel because water levels in the aquifer are drawn below the streambed. This effect is important in areas where streams are influent (losing streams). However, in the Coastal Plain where streams are effluent (gaining streams), ground-water diversion is the most significant source of boundary recharge to the pumped well and is the focus of this discussion.

A withdrawal made near a stream generally produces an area of influence that is noncircular and asymmetrically distributed with respect to the pumped well (fig. 15) because ground water is diverted preferentially to the well. An example is shown by two transient simulations, one of a pumped well relatively near a reach of Pump Branch and one relatively far from it (fig. 20). A pumping rate of 0.47 Mgal/d was used for each simulation, and the effects of withdrawal that are produced at the water table after 30 days are shown. In the simulation depicted in figure 20, a model cell simulating withdrawal is next to Pump Branch (area of influence labeled "A" in fig. 20). The resulting area of influence is asymmetrical with respect to the pumped well. The distance between the pumped well (that is, the center of the model cell that simulates withdrawal) and the nearest boundary of the area of influence (southwest of the well toward Pump Branch) is about 800 ft, whereas the distance between the pumped well and farthest boundary of the area of influence is about 1,700 ft (to the northeast, away from Pump Branch). The shorter distance on the stream side of the well is attributed largely to ground water diverted from the stream. However, in the simulation depicted in figure 20, the area of influence of a pumped well far from Pump Branch (area of influence labeled "B" in fig. 20) is larger than that produced by the well near the stream. At this increased distance from Pump Branch, the effects of ground-water diversion from nearby streams on the area of influence diminish, as indicated by the larger size and nearly circular shape of the area of influence.

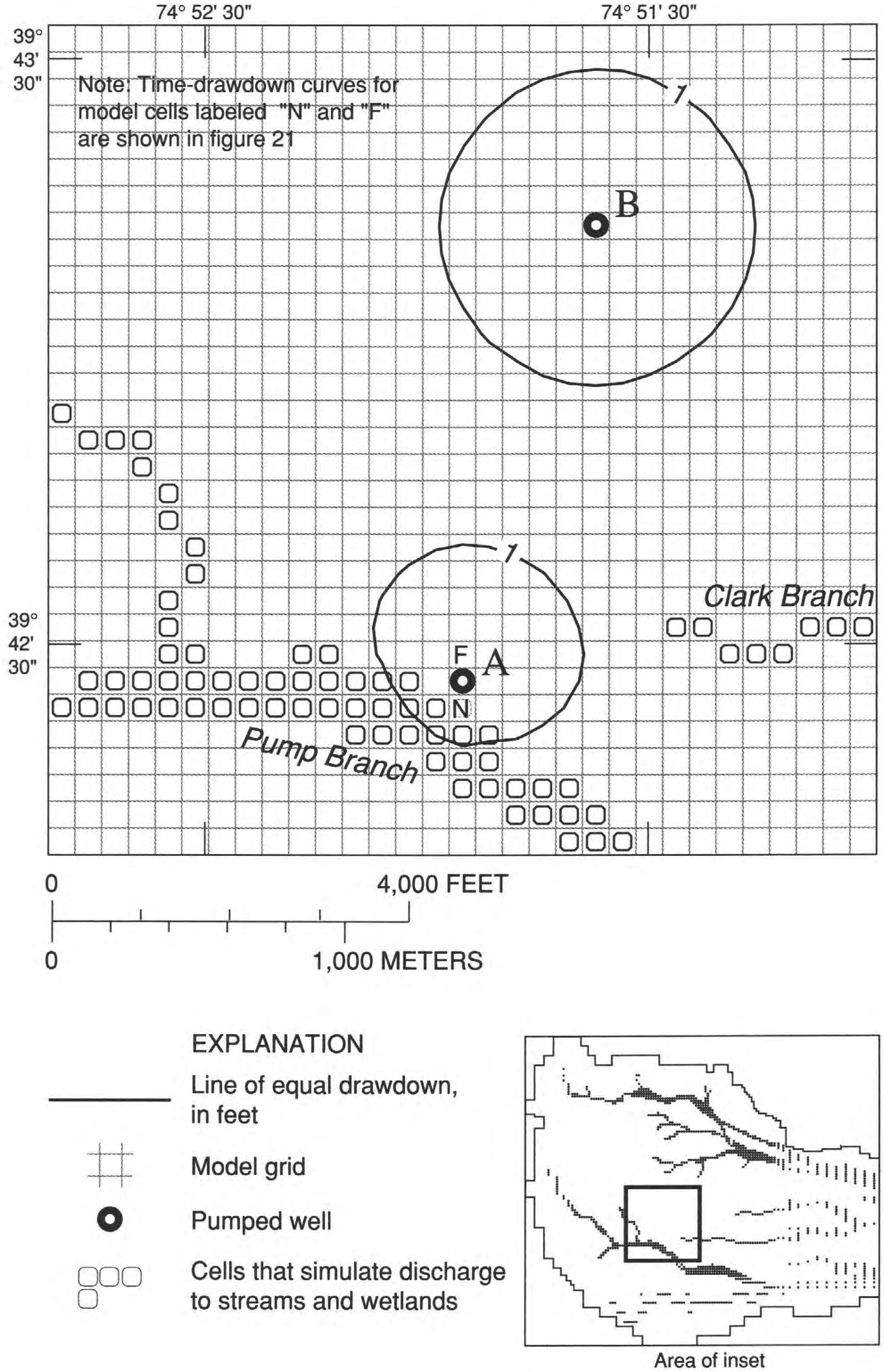


Figure 20. Areas of influence generated after 30 days by two transient pumped-well simulations, A and B, superimposed on common area of model, Mullica River Basin, New Jersey. (Well A is 300 feet from Pump Branch and well B is 5,200 feet from Pump Branch. Pumping rate is 0.47 million gallons per day for each simulation.)

The effect of ground-water diversion also can be observed within a single area of influence by comparing the drawdown produced on the stream side of the well to the drawdown produced on the opposite side of the well. The drawdown at the water table that results during the first 12 days of withdrawal at model cells labeled "N" and "F" (fig. 20) are plotted in figure 21. Cell N, located closer to Pump Branch, shows less drawdown for the same time period than does cell F, located on the opposite side of the well. The simulated time-drawdown curves for N and F diverge with time. The relative difference in drawdown at N and F indicates that ground water is diverted largely from a source southwest of the well.

A type curve generated by the Neuman solution also is shown in figure 21. The Neuman solution was set up to solve for the aquifer conditions specified in the numerical model at the location of the pumped well. An observation well was assumed to be 300 ft from the pumped well. The time-drawdown curve derived from the Neuman solution plots between the drawdowns of cells N and F of the numerical model. Although the trends for numerical and analytical solutions are similar for early time, the drawdowns generated with the numerical solution diverge from the analytical curve at later times. Divergence of time-drawdown curves would increase to an even greater extent for longer withdrawal periods and at higher pumping rates.

Using these concepts and the model, a ground-water flow experiment can be designed to determine how the size of the area of influence and ground-water seepage to streams are affected by pumped wells that are placed at various distances from the stream and pumped at various rates. This information can be used to provide an indication of the distance that a pumped well should be placed from a stream to disperse the effects of withdrawal over a larger part of the aquifer and to minimize effects on specific streams. The numerical model developed for this study was used to simulate the areas of influence produced in 27 hypothetical steady-state withdrawals from 9 different model cells between Pump and Wildcat Branches along transect A-A', shown in figure 22. The nine withdrawal simulations were made using the pumping rates of 0.2, 0.4, and 0.8 Mgal/d.

The relation between the size of the area of influence and the location of the pumped well between Pump and Wildcat Branches is shown in figure 23. The average distance between the pumped well and the 1-ft drawdown boundary is used to represent the size of the area of influence. In general, the size of the area of influence decreases with proximity to the stream and is smallest where the well is close to stream and largest where the well is close to the inter-stream divide. However, because ground-water seepage to Pump Branch is greater than seepage to Wildcat Branch, the area of influence produced by the well near Pump Branch (well 1 of figures 22 and 23) is smaller than that produced by the well near Wildcat Branch (well 9). This is because more ground water is available to divert from Pump Branch than from Wildcat Branch. A well near Wildcat Branch must tap into other sources of ground water to satisfy its pumpage requirements because relatively little water is available to divert from this stream. If both streams received identical ground-water seepage, then a well at the inter-stream divide would produce the largest area of influence on the transect. However, because of the different seepage rates to the streams, the largest area of influence is produced in a well located off the inter-stream divide and closer to Wildcat Branch.

The reduction of ground-water seepage to streams in the area surrounding transect A-A' that results from withdrawals is shown in figure 24. The figure shows the proportion of seepage that each stream or group of streams loses to a well pumped at 0.8 Mgal/d from each of the nine locations along transect A-A' (fig. 22). Because the withdrawal simulation is steady state, the total loss in ground-water seepage is equal to the pumping rate irrespective of the well location.

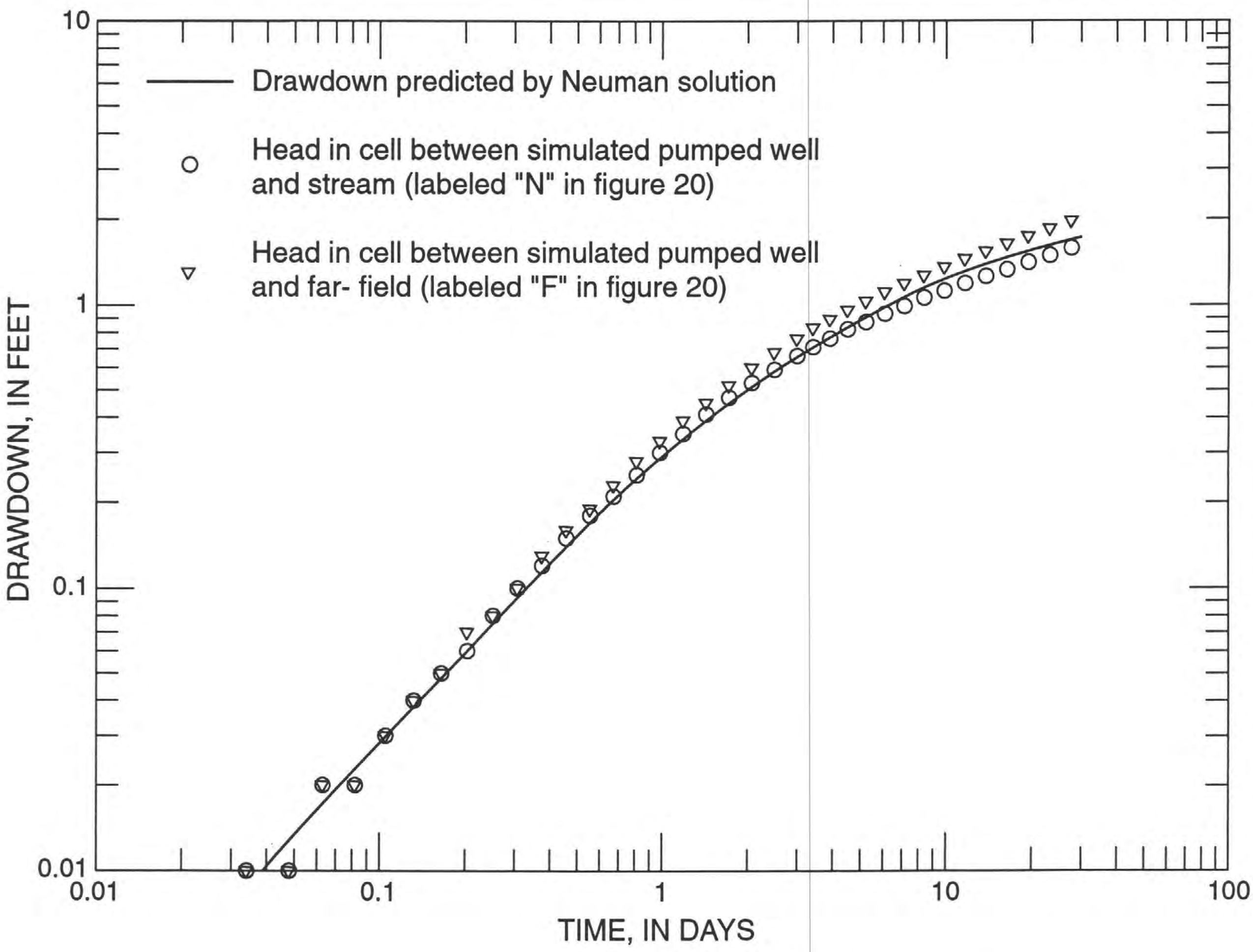


Figure 21. Plot of simulated time-drawdown at a model cell between the pumped well and Pump Branch, New Jersey, and a model cell between the pumped well and the far-field. (Neuman's analytical solution for simulated aquifer conditions also is shown.)

However, the relative proportion of seepage lost by each stream depends on well location. Most losses of ground-water seepage are distributed among streams that are closest to the pumped well. The headwaters of Wildcat, Pump, and Clark Branches are closest to the well locations along transect A-A' and account for the greatest reduction in ground-water seepage. Smaller losses are also noted for streams that are relatively far from the pumped well.

When well 1 is pumped at 0.8 Mgal/d, 75 percent of pumpage is supplied by diverted ground water from Pump Branch and 25 percent is supplied by diverted ground water from other streams. When well 9 is pumped instead, 55 percent of pumpage is supplied by Wildcat Branch and 45 percent is derived from other streams. In both scenarios, either Wildcat Branch or Pump Branch sustains the brunt of the seepage loss. However, when well 6 is pumped, 25 percent of pumpage is supplied by Pump Branch, 35 percent by Wildcat Branch, and 35 percent from other streams. Consequently, the relatively large reductions in ground-water seepage to a stream that results when a well or wells are close to the stream can be minimized if the well is located farthest from all streams, near the inter-stream divide.

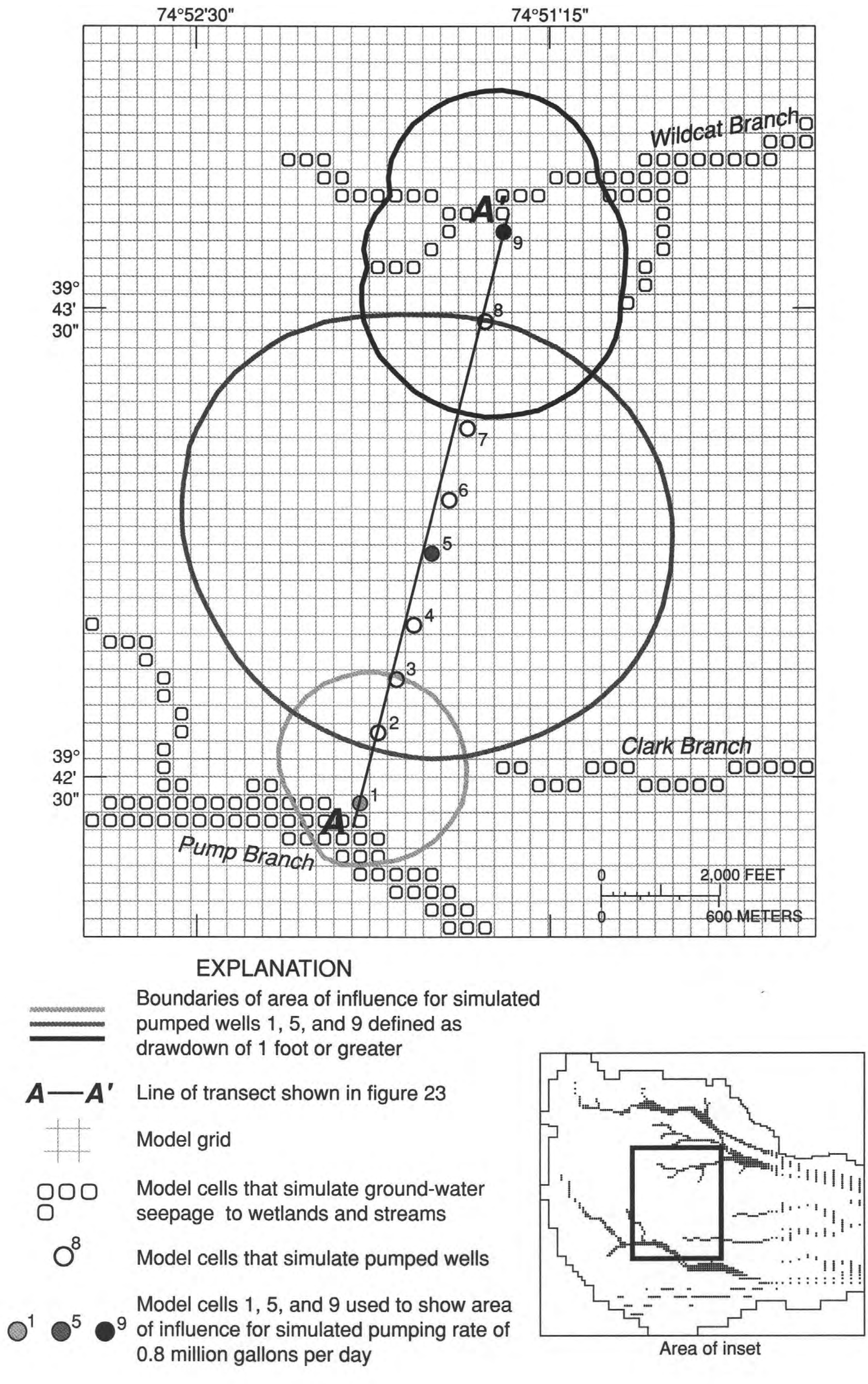


Figure 22. Areas of influence generated by three steady-state pumped-well simulations superimposed on common area of model, Mullica River Basin, New Jersey. (Pumping rate is 0.8 millon gallons per day for each simulation.)

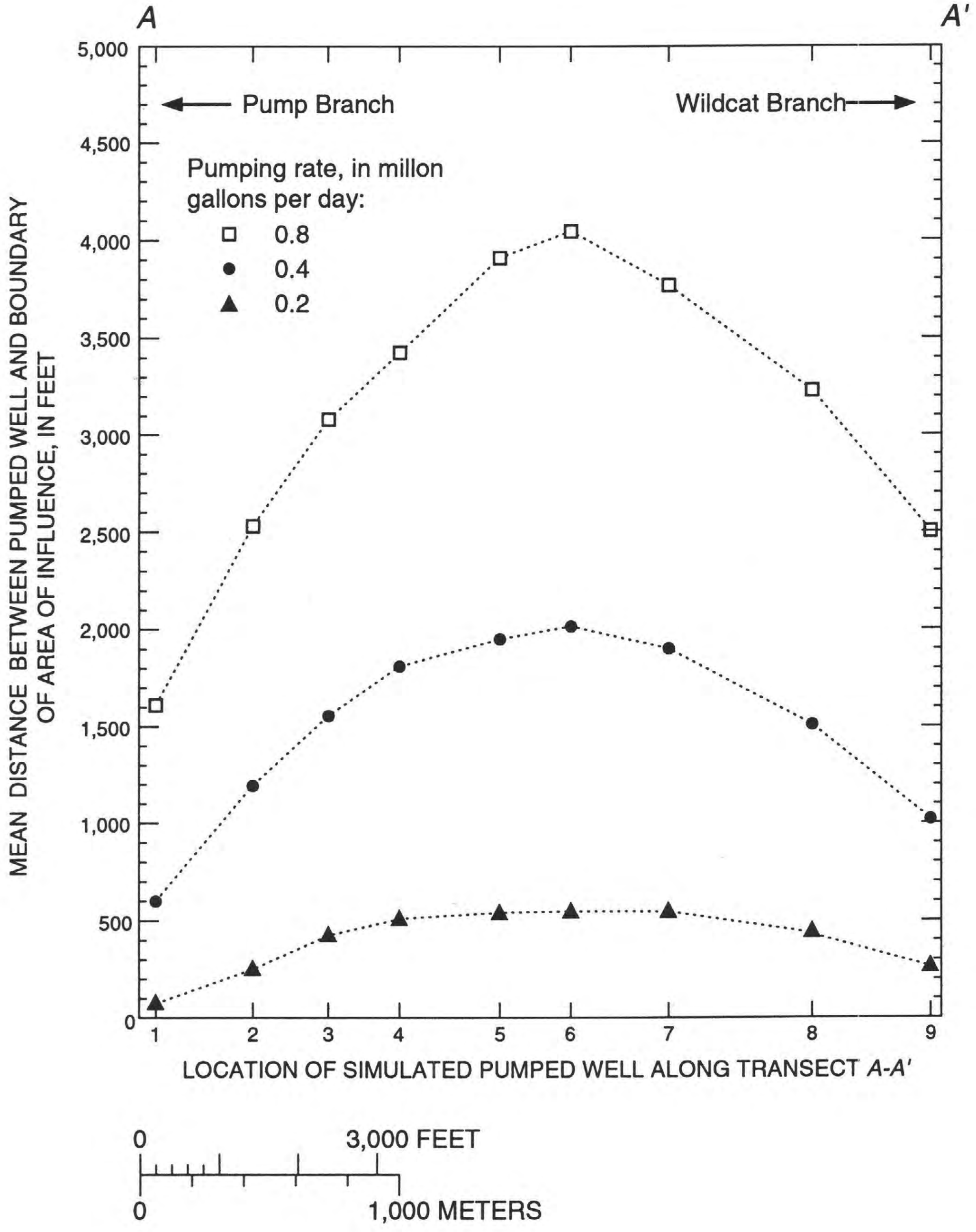


Figure 23. Distance between simulated pumped well and 1-foot boundary of the area of influence, Mullica River Basin, New Jersey. (Location of transect *A-A'* shown in fig. 22)

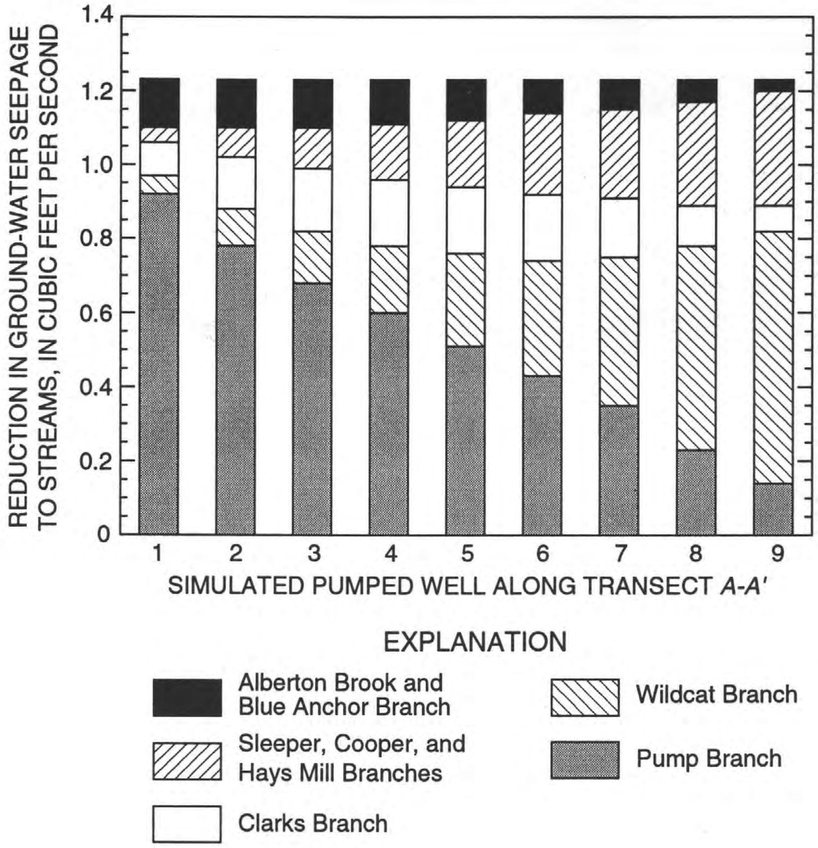


Figure 24. Simulated reduction of ground-water seepage to streams in the Mullica River Basin, New Jersey, caused by withdrawal of 1.23 cubic feet per second from wells 1 through 9. (Location of transect A-A' shown in fig. 22)

The size the area of influence is also affected by the pumping rate. For a pumping rate of 0.8 Mgal/d, the areas of influence generated from pumped wells 1 and 9 intersect parts of Wildcat and Pump Branches (fig. 22). The simulations indicate that a pumped well needs to be at least 1,800 ft from Pump Branch (between wells 2 and 3) in order to produce an area of influence that would not impinge on Pump Branch and adjacent wetlands (although, at this distance, the area of influence would impinge on the tip of Clark Branch). Similarly, the pumped well needs to be at least 3,500 ft from the Wildcat Branch and its tributaries (between well 6 and 7) in order to produce an area of influence that would not impinge on Wildcat Branch and adjacent wetlands.

The simulated results also show that, for low pumping rates, a well can be relatively close to a stream and still produce an area of influence that does not impinge on the stream. At a pumping rate of 0.4 Mgal/d, the well could be as close as 600 ft from Pump Branch and as close as 1,100 ft from Wildcat Branch without causing the area of influence to impinge on the streams (compared to the 1,800- and 3,500-ft respective distances resulting from a 0.8 Mgal/d pumping rate). Similarly, at a pumping rate of 0.2 Mgal/d, the pumped well could be as close as 150 ft from Pump Branch and as close as 300 ft from Wildcat Branch without causing the area of influence to impinge on the streams.

The results of simulations depicting withdrawal near streams indicate that the area of influence will be smaller than predicted by methods that assume a boundless aquifer if withdrawals are made near streams. Although smaller areas of influence appear to affect the ground-water system less, the small size of the area of influence is the result of a significant loss of ground-water seepage from a nearby stream. In contrast, the relatively large area of influence that results from withdrawals made at wells far from streams on divides signifies that seepage losses to local streams are minimized by dispersing withdrawal effects among many streams in the ground-water system.

MONITORING STRATEGIES FOR STREAMS AND WETLANDS IN THE NEW JERSEY COASTAL PLAIN

A monitoring strategy is an essential complement to an aquifer-testing protocol. Monitoring strategies involve judicious placement of devices that measure hydrologic conditions of an aquifer system, such as water levels and streamflow. The devices generally consist of piezometers and stream gages, which can be instrumented with recorders that track water levels over small time increments.

Hydrologic Considerations

The hydrologic data collected from a carefully designed monitoring network increases the knowledge of (1) the baseline hydraulic conditions in an aquifer system–that is, the range of seasonal and long-term water-level fluctuations in an aquifer and variations in ground-water seepage to streams and wetlands–and (2) the response of an aquifer system to natural stresses. An understanding of the short- and long-term effects of ground-water-related development on water levels in the aquifer, such as withdrawals or artificial recharge, is predicated on the knowledge of baseline hydraulic conditions. Such knowledge can alert resource managers to potential water-supply problems, such as persistent downward trends in water levels from normal levels, particularly near streams and wetlands. Furthermore, an evaluation of the hydraulic responses of an aquifer to natural stresses can provide a relatively inexpensive means of determining the storage and water transmitting properties of an aquifer.

Water-Level Fluctuations

A monitoring network (fig. 25) was developed in part of the Kirkwood-Cohansey aquifer system to implement the above-stated objectives. The network serves as a prototype water-level monitoring network used to gage the hydrologic condition in a representative part of the Kirkwood-Cohansey aquifer system. The area monitored was restricted to the Mullica River Basin for this investigation, but it could be expanded to include a larger region in order to provide a more comprehensive water-management tool. The network consists largely of observation wells that had already been installed by Federal, State, and county agencies. Many of these observation wells served previous hydrologic investigations and currently remain intact. The wells selected represent a range of land elevations and depths in the unconfined aquifer system. Well casings are made of steel and PVC; well diameters range from 2 to 8 in. Because there were few wells in wetland areas at the commencement of this investigation, 12 drive-point piezometers were installed in wetlands in the Mullica River Basin to augment the network.

Water levels were measured monthly during November 1992 – October 1993 to determine seasonal fluctuations in the water table in the Mullica River Basin area. The yearly ranges and mean values of depth to water from land surface measured in observation wells of the monitoring

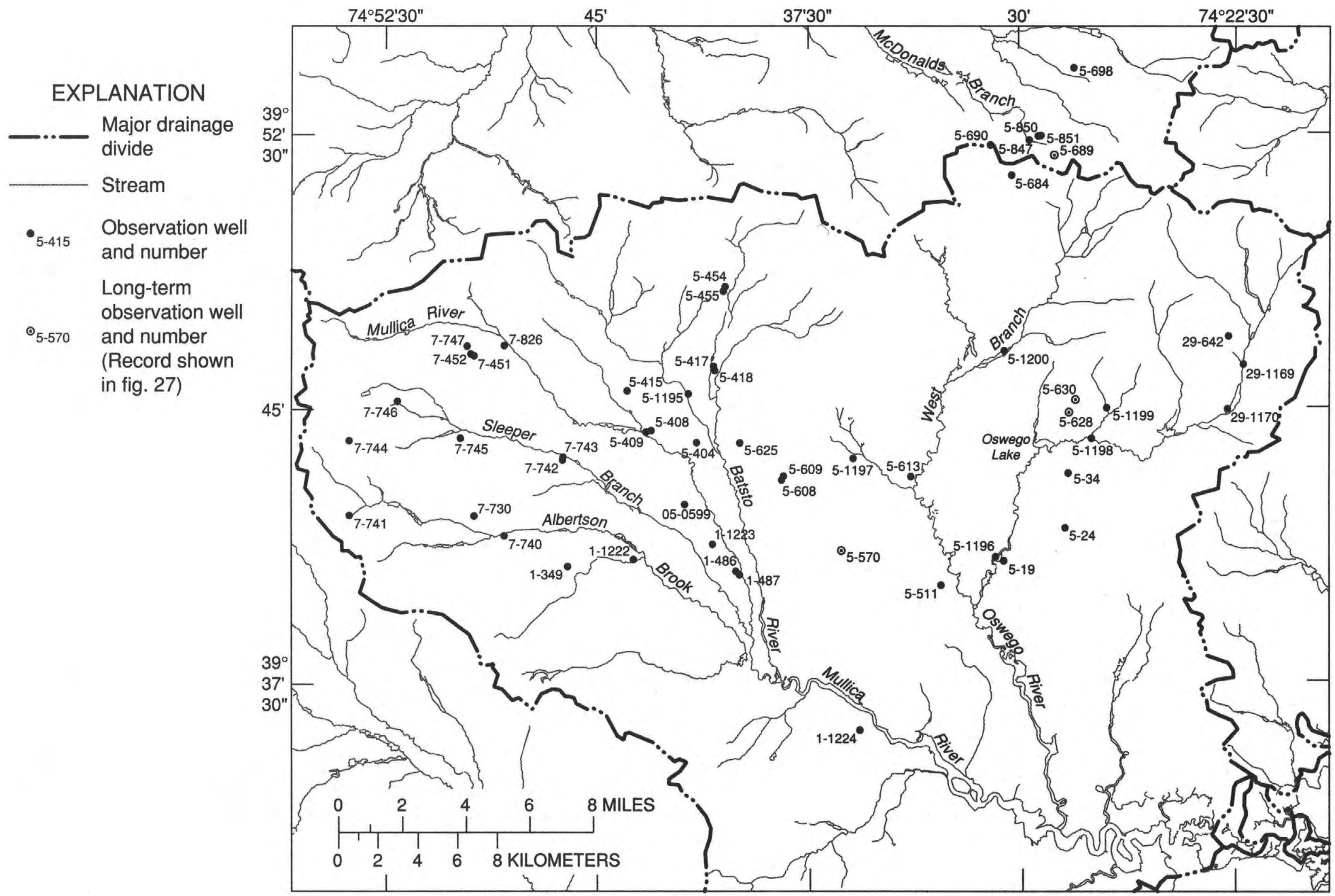


Figure 25. Observation-well network used to record water-level fluctuations in the Mullica River Basin, New Jersey, November 1992–October 1993.

network are shown in figure 26. Water levels were measured at wells in upland and lowland areas. Upland areas are defined as areas where the water level is below land surface and is a true water table. Lowland areas are defined as areas where hydraulic head is above land surface, usually in wetlands, where the hydraulic heads are consistently higher than the base of the wetland. (In this report, the base of the wetland is considered to be the interface between the sandy material and the muck.)

In general, the scatter of yearly mean values of depth water among wells in upland areas increases with increasing land-surface elevation. The plot shows that the mean values of depth to water fall in a 9-ft interval, from 2 to 11 ft below land surface, in wells below elevations of 60 ft. However, in wells above elevations of 60 ft, the mean values of depth to water fall within a 30-ft interval, from 2to 32 ft below land surface. The plot indicates an abrupt increase in the scatter of mean values of depth to water in wells at land surfaces higher than 60 ft above datum; this increased scatter may simply relate to the greater variation in land-surface elevations attributed to steeper land gradients at elevations higher than 60 ft.

Yearly ranges of depth to water among wells in upland areas were from 1.2 to 5.5 ft (wells 07-0740 and 05-0570, respectively), and the average range was 3.4 ft. There appears to be little correlation between yearly range of depth to water and land-surface elevation; yearly ranges of as much as 4.7 ft are found in low-lying land areas (well 05-0404), whereas ranges of as little as 2.7 ft are found in relatively high land areas (well 07-0741). The variation in the ranges of depth to water measured in wells is most likely related to variations in grain sizes in the aquifer near the well screen. The ranges in hydraulic heads measured in wells in wetlands were from 0.5 to 2.6 ft, and the average range was 1.3 ft; all heads were above land surface. This range of heads is narrow compared to those in upland areas and shows that hydraulic heads in lowland areas are less responsive to recharge events than are heads measured in upland areas. Wetlands are found at any land-surface elevation and are not restricted to low-lying areas.

To gain a perspective of what yearly measurements represent, it is necessary to understand water levels in terms of long-term data. The yearly range and mean values of depth to water for water years 1951–93 are shown in figure 27. The water levels were measured at four wells in Mullica River Basin and Lebanon State Forest (fig. 25). In figure 27, the hydrograph of yearly mean heads is superimposed on to long-term mean range and average yearly mean for each well over the period of record. For the three lower wells in the plot, yearly ranges and yearly mean depth to water appear to correlate somewhat; that is, ranges in water-level fluctuations are greater when mean values of water levels are higher. For well 05-628, a well screened close to land surface, this relation is reversed, an indication that fluctuations in the water table near land surface are suppressed. The ranges of depth to water are different for each well, an indication that the wells are differentially sensitive to hydraulic heads. However, the long-term pattern of fluctuations that result from seasonal variations in recharge is reflected in all wells; lows and peaks are synchronous on all yearly mean hydrographs.

Yearly mean water levels tend to oscillate irregularly about a long-term mean value. Under natural conditions, yearly means and ranges of depth to water do not depart from long-term average values for sustained periods. Even during extreme drought or exceedingly wet years, departures from average values tend to be short lived. For example, during 1967 the water level of well 05-570 reached a low of 18 ft, 5.8 ft lower than the long-term average. However, by 1969, the water level climbed back to within the mean range. Similarly, in 1972, the water level reached a high of 8.5 ft, 3.5 ft higher than the long-term average. By 1975, the water level was back to within the

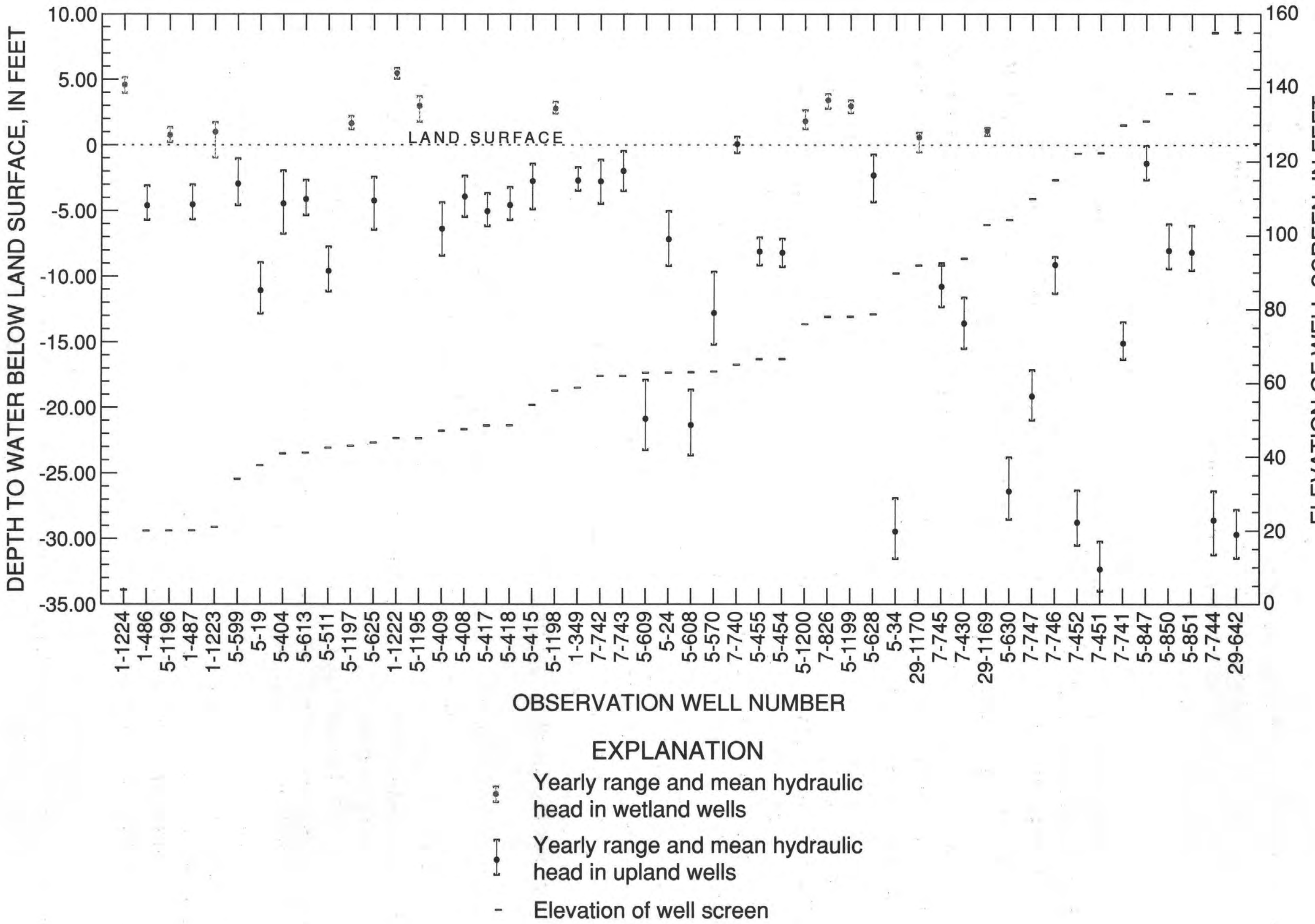
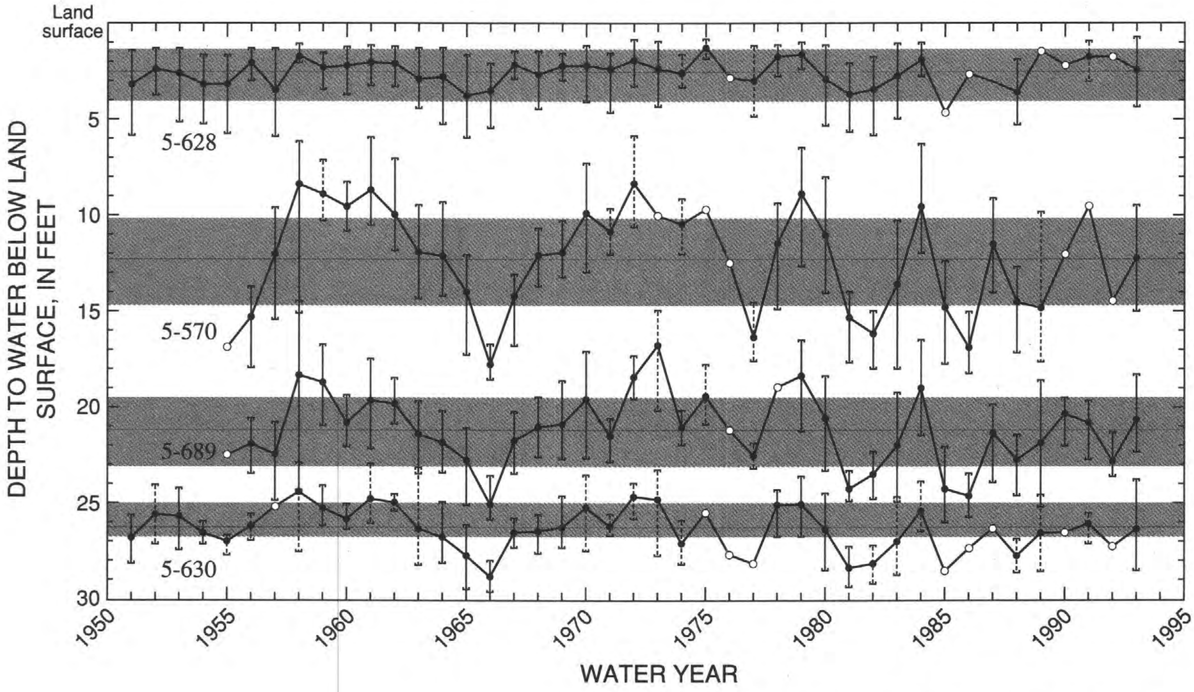


Figure 26. Maximum and minimum depth to water from land surface in observation wells of monitoring network, Mullica River Basin, New Jersey, arranged in order of increasing land-surface elevation.



Note: Water year is defined as the 12-month period from October 1 through September 30. It is designated by the calendar year in which it ends.

EXPLANATION

Mean range of hydraulic head for water years 1951–93

Mean of mean hydraulic head for water years 1951–93

Trend of mean hydraulic head for water years 1951–93

Range of hydraulic head for water year, based on 10 or more monthy measurements

Range of hydraulic head for water year, based on fewer than 10 monthy measurements with at least 1 measurement made in late winter and 1 made in late summer

Mean hydraulic head for water year, based on fewer than 10 monthy measurements with no measurement made in late winter and (or) late summer

Figure 27. Range and mean of water levels recorded in four observation wells in the Mullica River Basin, New Jersey, water years 1951–93.

mean range. Consequently, departures from long-term averages, such as occur during droughts, are not necessarily a concern to long-term water supply. However, water levels that show sustained or continuing downtrends from baseline levels would indicate that a new stress on the system, such as recently installed pumping facilities, may alter the natural ground-water flow in the system and potentially affect water supply.

Ground-Water Seepage to Streams and Wetlands

The New Jersey Coastal Plain is a relatively flat region where land-surface gradients typically range from 2 to 10 ft/mi. Highest elevations at the major drainage divides are about 200 ft. Because of the low relief, ground-water seepage collects readily as surface-water bodies in lowlands. In general, many of the wetlands are located in shallow, troughlike topographic depressions or channels and contain streams that meander through the mucks that accumulate in these channels. A conceptual cross section through a stream and riparian wetland that is typical of the New Jersey Coastal Plain is illustrated in figure 28. Ground water flows from the upland areas to the channel in the lowland. Ground water seeps to the land surface at all points below where the water table intersects the land surface. Because the channels are filled with an accumulation of organic deposits and muck, ground water that seeps into the channel disperses into these deposits and ultimately discharges to the stream.

Hydroperiod and seasonally intermittent ground-water seepage

During a typical seasonal cycle, recharge events are more frequent around the late autumn and winter than during the late spring and summer; thus, relative excesses or deficits in ground-water availability occur according to the season. In response, water levels in the aquifer fluctuate between maximum levels in the early spring and minimum levels in late summer. An observable effect of these water-level changes in the aquifer is the upstream and downstream migration of the stream start of flow. In areas of low land-surface gradients, such as in the Coastal Plain of New Jersey, this migration can extend over thousands of feet of streambed. Another effect of these water-level changes is that the width of the seepage area in a channel expands in the winter and contracts in the summer in response to the fluctuations of the water table. As a result, ground water seeps through a fringe area along the edges of the channel during only a part of the year. The width of the fringe area is determined by the slope of land surface at the edge of the channel and the range of water-table movement during the seasonal cycle. The average seasonal range of hydraulic heads in wetlands of the Mullica River Basin was found to be 1.3 ft. Assuming that the slope across the fringe area is about 2 degrees (a 1-ft drop over 30 ft horizontal distance), a 1.3-ft increase in the water table would result in a 34-ft-wide expansion of the wetland on either side.

The fluctuation in the size of the ground-water seepage area during the year is related to the length of time that the wetland is flooded as a result of ground-water seepage and is described as a "hydroperiod," or period of inundation (Duever, 1988). The hydroperiod is generally continuous throughout the year in the deepest parts of the channel at elevations that are lower than the lowest seasonal level of the stream start of flow. The hydroperiod decreases to progressively shorter periods within the fringe area of a wide channel and upstream near headwaters where the start of flow moves up and down the length of a stream.

Examples of water levels in wetlands that were measured monthly during 1992–93 in the Mullica River Basin are shown in figure 29. The plots shown in figure 29A were made from measurements in five observation wells that were installed in different wetlands in the Mullica River Basin (fig. 25). The hydraulic heads in wells screened beneath the base of the wetland indicate that

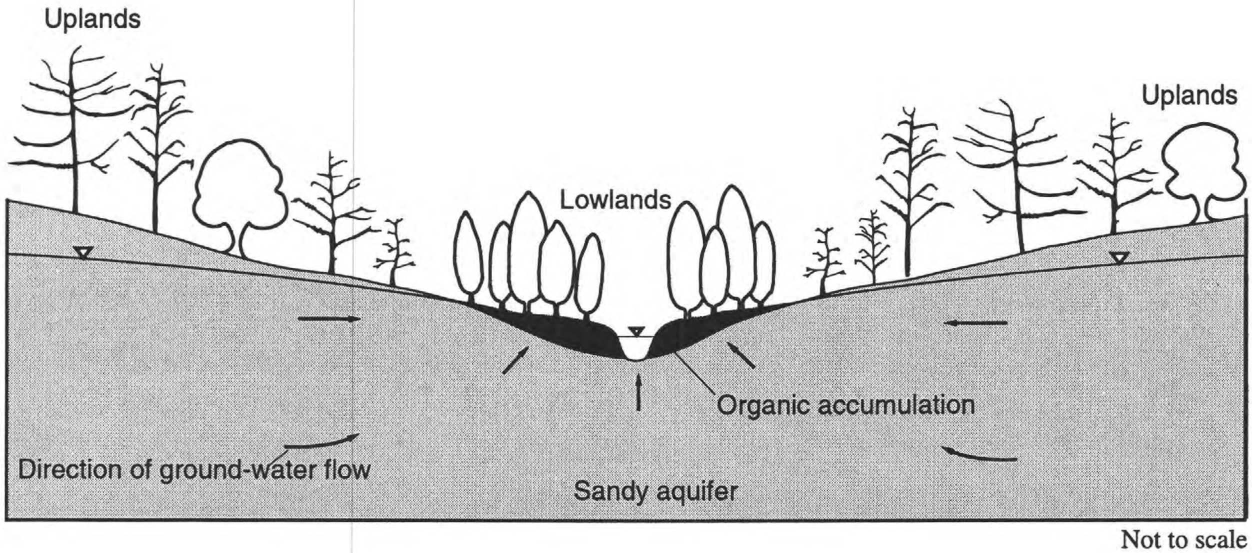


Figure 28. Cross section through typical Pinelands wetland driven by ground water seeping into shallow land depression.

ground water seeps into the wetland. In general, seepage is proportional to the length of the column of water above the base of the wetlands. Positive values of head in figure 29A indicate that the hydraulic heads are above the base of the wetland. Negative values of head indicate that the water level is below the base of the wetland and that no ground water seeps into the channel. For the most part, ground water seeps into the channel and floods the wetlands at the sites of the observation wells, and the hydroperiods were continuous in areas near the observation wells during the period of record. Ground-water seepage was highest in March and lowest in July. However, records from observation well 01-1223 indicate that the water level fell below the base of the wetland between early June and mid-July (fig. 29A).

The ground-water hydrographs shown in figure 29B were made from measurements made at three observation wells near the Oswego Lake area (fig. 25). The measurements from well 29-1170, located near a minor tributary northwest of the Papoose Branch, show almost constant water levels of less than 1 ft. Between early June and mid-July, water levels fell below the base of the wetland. Water levels measured in well 05-1198 near Papoose Branch near the Oswego River remained near the 3-ft level for the period of record and indicate significantly higher ground-water seepage at this site than at the 29-1170 site. Observation well 05–0628 (fig. 27) is in an upland area. Water levels measured at this site show a seasonal fluctuation pattern that is typical of water levels observed in wells screened in the shallow Kirkwood-Cohansey aquifer system in the Mullica River Basin.

Hydraulic head and ground-water seepage distribution along transects in headwaters wetlands

A transect of drive-point piezometers was installed across the upper reaches of Middle Branch and McDonalds Branch in Lebanon State Forest, N.J. (figs. 30 and 31). The transects were established to measure seasonal water-level fluctuations and to determine hydraulic-head distribu-

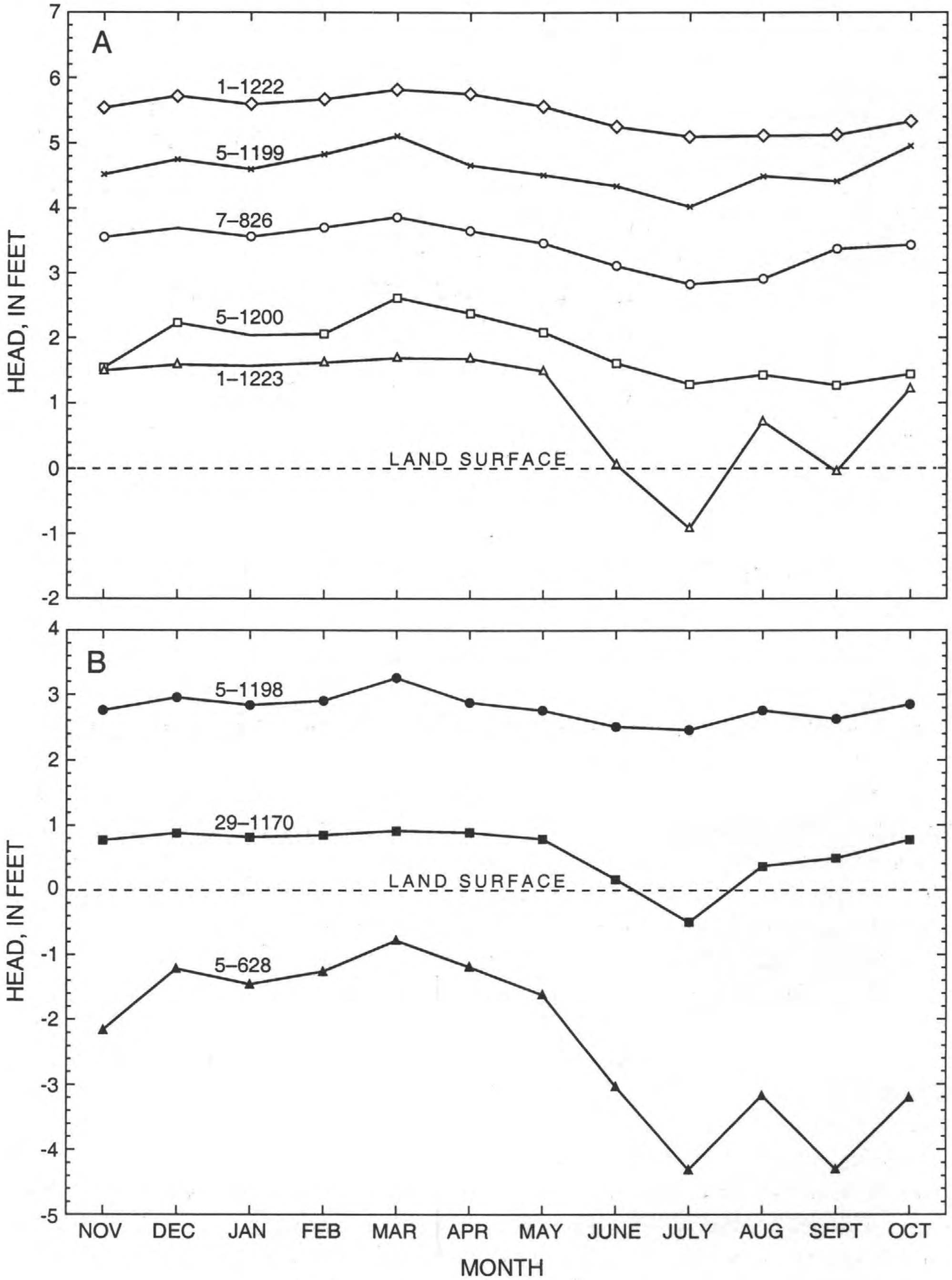


Figure 29. (A) Hydraulic heads recorded during November 1992–October 1993 in wetlands of the Mullica River Basin, New Jersey, and (B) hydraulic heads in and near wetlands near the upper Oswego River area.

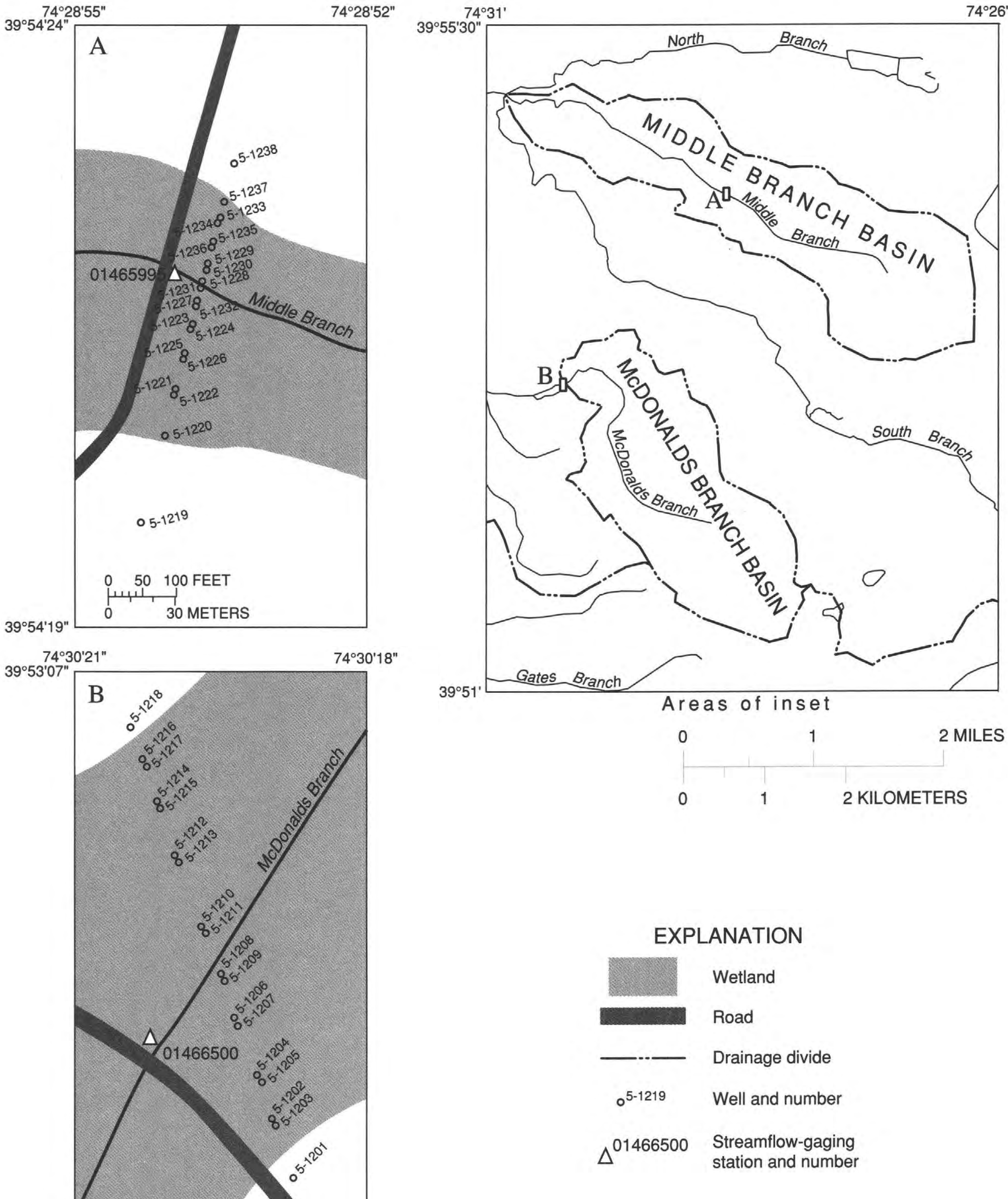


Figure 30. Locations of piezometer transects and streamflow-gaging stations at (A) Middle Branch and (B) McDonalds Branch in Lebanon State Forest, New Jersey.

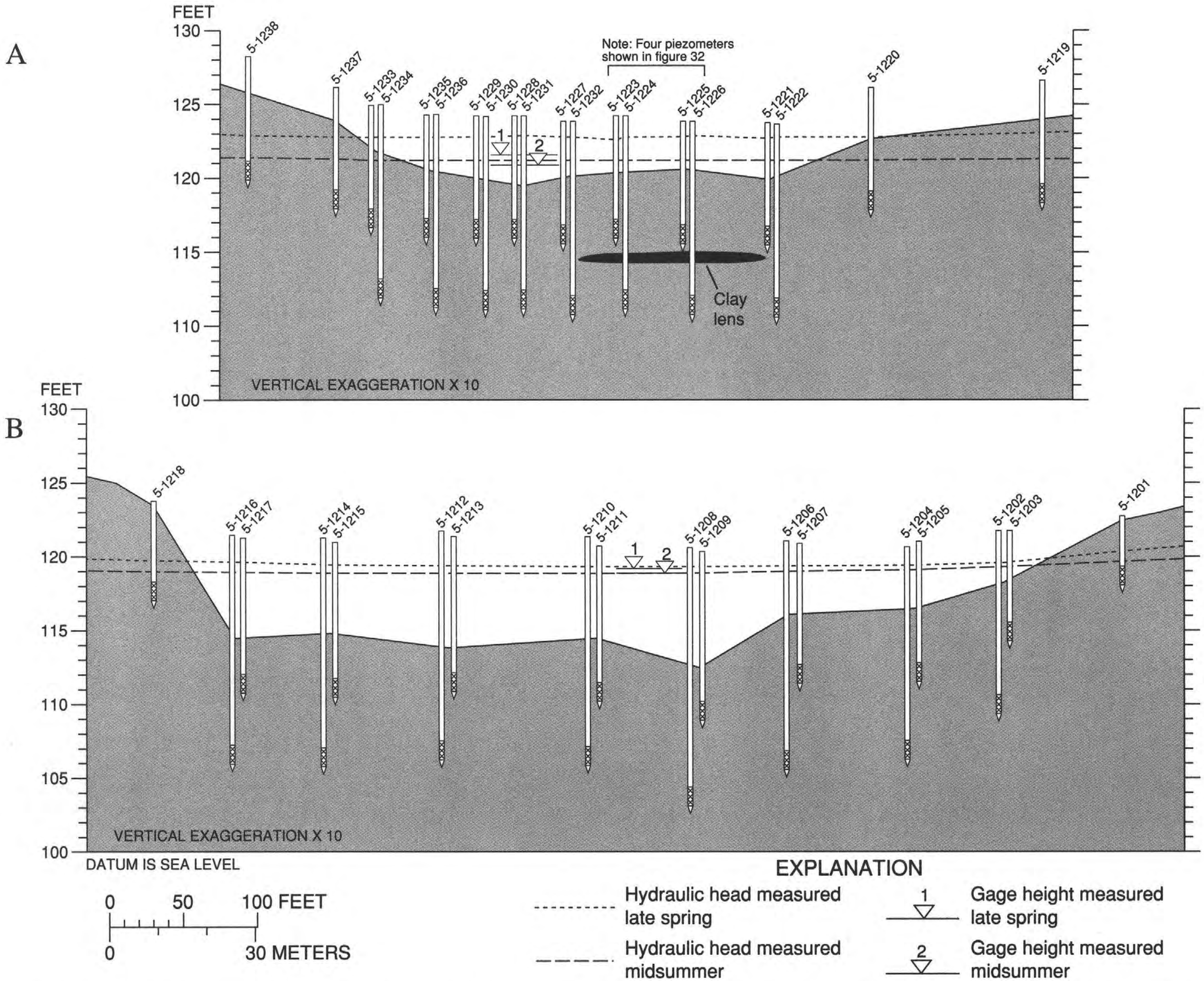


Figure 31. Transects showing locations of drive-point piezometers and water levels for late spring and midsummer 1993 across (A) Middle Branch and (B) McDonalds Branch in Lebanon State Forest, New Jersey.

tions beneath headwaters wetlands at a fine spatial scale. Piezometer doublets were installed in the wetland part of the transects; the screens were set to approximately 3.5 and 9 ft below the base of the wetland. Single piezometers were installed in the upland area at the edges of the transects. The piezometers were installed so that screens were at least 3 ft below the wetland to minimize any hyporheic effects from surface water. The Middle Branch transect spans a 550-ft section that includes the wetland on either side of Middle Branch and a narrow band of the adjacent upland (fig. 31A). Similarly, the McDonalds Branch transect spans a 700-ft section of wetlands and a small upland area near its edges (fig. 31B). Transect sites were located next to streamflow gaging stations so that streamwater stage and discharge could be measured concurrently with hydraulic heads in the piezometers. In addition, a tipping bucket rain gage was installed near the Middle Branch site.

In general, hydraulic heads measured in each pair of shallow and deep piezometers were nearly identical for all measurements made throughout the season (except for heads measured in piezometers 05-1223 and 05-1225 at the Middle Branch site), an indication that lines of equal potential are nearly vertical in the shallow aquifer and that ground water flows laterally under the wetlands. However, the fact that these heads are higher than the heads at the surface of the streambed indicates that ground water seeps into the channel and that vertical hydraulic head gradients do exist close to the base of the wetland. The 1.3-ft screen interval used in the piezometers is too wide to detect small vertical differences in hydraulic head near the base of the wetland. Furthermore, by positioning screens 3 ft below the base of the wetland, detection of differences in vertical hydraulic head was precluded. The changes in hydraulic head can be measured near the base of the wetland by use of a relatively fine sampling interval. For example, the difference in hydraulic heads between the top of the streambed and points beneath the streambed in Middle Branch were measured with a minipiezometer designed for a 1-in. sampling interval and equipped with a differential manometer. Measurements showed that head differentials graded from 0.05 to 0.4 ft between the depths of 1.5 and 2.1 ft below the streambed.

During periods of drought, ground water is released from aquifer storage. Because the size and arrangement of grains that make up the aquifer vary spatially and because the storage properties of an aquifer are inherently linked to these, aquifer storage also varies spatially. Consider the hydraulic-head changes along the Middle Branch and McDonalds Branch transects between late spring and midsummer 1993. Water-level measurements made on June 1 and July 30, 1993, along the Middle Branch transect are shown in figure 31A. During this period, hydraulic heads decreased across the transect by about 1.7 ft. This decline was a result of the nearly total lack of rainfall during this period. The decline in hydraulic heads indicates that ground-water seepage also decreased, as shown by the difference between the stage and head in each piezometer. In addition, a 35-ft-wide fringe area on the left side and 45-ft-wide area on the right side of the wetland went dry during this period (fig. 31A), effectively reducing the total width of the seepage area by about 80 ft near the area of the transect.

Water-level measurements made on June 1 and August 2, 1993, along the McDonalds Branch transect are shown in figure 31B. Differences in heads between these dates vary from 0.4 ft near the center to about 0.7 ft near the edges of the transect and are significantly less than those of the Middle Branch transect. A fringe area less than 10 ft wide on either side of the wetland went dry during this period (fig. 31B), effectively reducing the total width of the seepage area by less than 20 ft near the area of the transect. The two transects are about 2.2 mi apart (fig. 30), so both sites would have received similar precipitation. Differences in the hydraulic responses indicate differences in local storage and water-transmitting properties of the aquifer.

It has been observed that ground-water seepage rates are greater near the edges of lakes or wide channels than at the centers (McBride and Pfannkuch, 1975). Although seepage rates across the wetland were not measured directly, differences between hydraulic heads and stage can be used to show differences in seepage rate as long as the aquifer material is relatively homogeneous. The material in the shallow aquifer along the transect at McDonalds Branch was considered to be relatively homogeneous on the basis of cores and the consistent number of hammer blows needed to install piezometers along the transect. The differences between the surface water and the hydraulic heads measured on June 1 in each piezometer on the transect show a progressive increase in head difference from the center to the edges (from 1.61 ft at center of the transect to 2.02 ft at the northern end and 2.65 ft at the southern end of the transect). In contrast, the material in the shallow aquifer along the transect at Middle Branch contains clay layers near the surface, as determined from inspection of cores and from the many hammer blows needed to install piezometers along the transect. The differences between the stage and the hydraulic heads measured on June 1 in each piezometer on the transect vary little from 1.16 ft (fig. 31).

The ubiquitous clay layers in the Cohansey sand can create local perturbations in the head distribution that are not representative of heads on a larger scale. Clay layers typically found in the aquifer generally consist of poorly permeable or "tight" material several feet thick, and they may extend hundreds of feet laterally in the shallow aquifer. Where a clay layer is several feet from the land surface, the hydraulic heads measured in piezometers screened at depths above and below the layer may show dissimilar patterns of hydraulic heads over time. In general, hydraulic heads measured in a well screened below a clay layer will reflect the prevalent hydrologic character of the area, whereas the heads measured in a well screened above a clay layer will reflect the hydrologic character of the part of the aquifer that is hydraulically disconnected from the aquifer by the clay layer.

As an example, consider the four piezometers installed above and below the clay layer shown in figure 31A. The piezometer pairs are about 35 ft apart. The clay layer is approximately 5 ft from the land surface and is about 2 ft thick. Four ground-water hydrographs, based on bimonthly head measurements made in the four piezometers between January and September of 1993, are shown in figure 32. Hydraulic heads in all piezometers are higher than the base of the wetland during this period, an indication that ground water seeped into the channel at this site even during the driest part of the year. The patterns of the hydrographs plotted from heads measured in piezometers screened below the clay layer (wells 05-1224 and 05-1226) are nearly identical for the period over which measurements were made. The pattern of these hydrographs are nearly identical to hydrographs of all other piezometers of the transect (not shown) and indicate that, for the period, the hydraulic heads have a signature or characterizing pattern indigenous to the area.

Hydrographs for piezometers screened above the clay, however (wells 05-1223 and 05-1225), show different patterns than those for piezometers screened below the clay and, moreover, are different from each other. The clay layers function as a barrier and isolate the shallow aquifer above the clay from the rest of the aquifer. During the late winter and early spring, the hydraulic heads in the piezometers screened beneath the clay are higher than those screened above the clay. This anomaly may be caused by increased pressure head that developed beneath the clay during the spring when ground-water flow to the wetland increased. During late summer, the pressure head beneath the clay dropped as ground-water flow decreased.

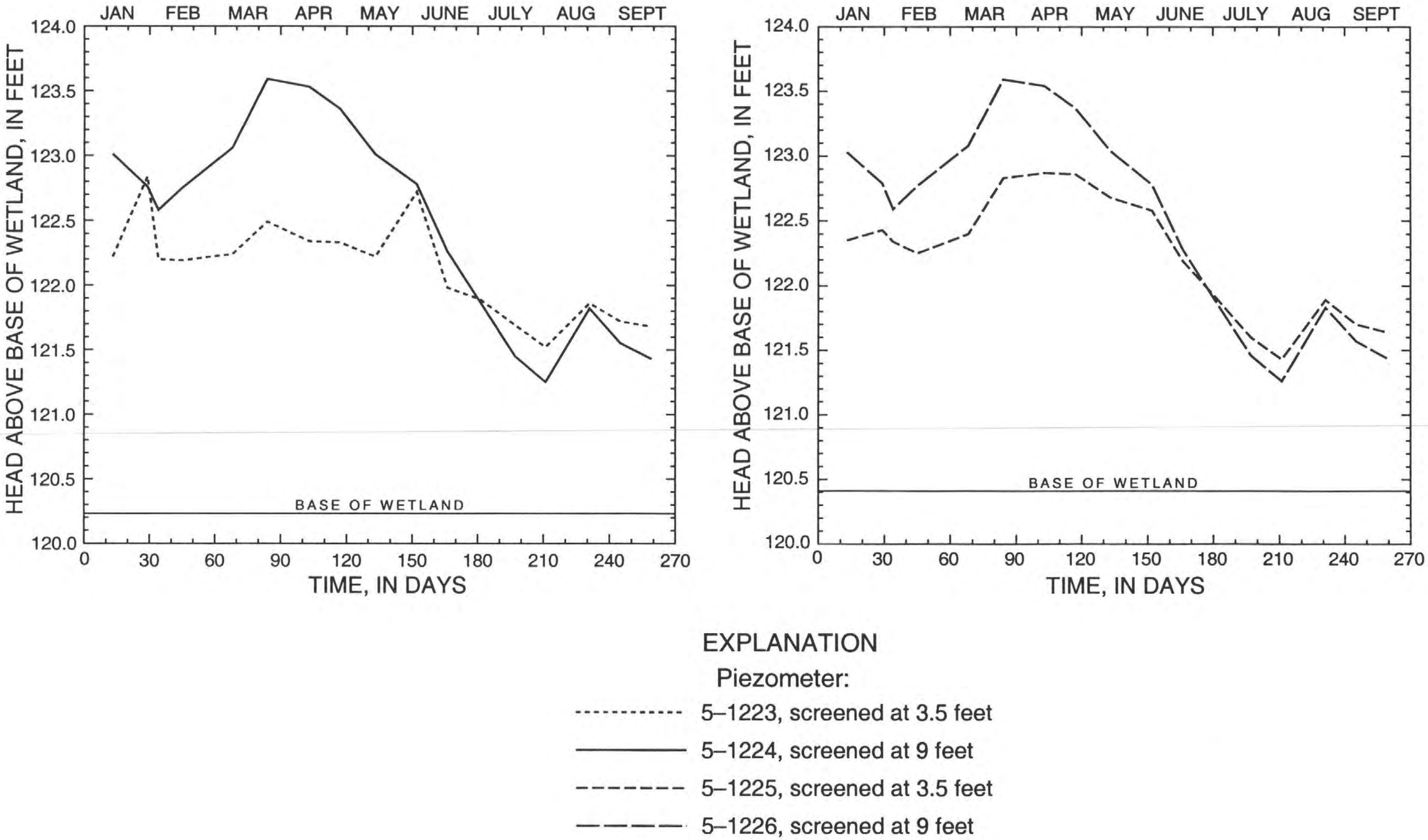


Figure 32. Water levels recorded during January–September 1993 in piezometers at two sites 35 feet apart along the Middle Branch transect, Lebanon State Forest, New Jersey. (Piezometer locations are shown in figs. 30 and 31.)

Aquifer response to recharge

The hydraulic response of the shallow aquifer to recharge provides useful information on aquifer storage and water transmitting-properties and on rates of water-level decline during periods of drought or rise during periods of recharge. The aquifer storage coefficient (*S*) and transmissivity (*T*) have been determined, in some cases, by recession analysis of hydraulic head or streamflow. Because recession analysis is based on aquifer response to naturally occurring input that is readily available, it can often be employed more conveniently and economically than aquifer testing to determine the water-transmitting properties of an aquifer.

Rorabaugh (1960) showed that the ratio of transmissivity to the storage coefficient (*T/S*) can be computed from slopes of recession at any well. Later (1964), he extended his analysis to streamflow and showed that the slope of a base-flow recession curve is related to the stream-aquifer property T/a^2S, where *a* represents the average distance from the stream to the ground-water divide. According to this relation, base flow declines exponentially with time after recharge ceases. When T/a^2S is greater than 0.2, the plot of the logarithm of streamflow versus time becomes a straight line. The time when this occurs is referred to as the "critical time" (t_c) and is expressed as

$$t_c = \frac{0.2a^2S}{T} . \tag{5a}$$

In practice, the base-flow recession curve is evaluated to determine the time required for streamflow to decline through one log cycle after the critical time. This relation is expressed as

$$\frac{T}{a^2S} = \frac{0.933}{(\Delta t) / (\log\ cycle)} , \tag{5b}$$

where Δt is the recession time through one log cycle.

These relations can be used to gain insight into the stream-aquifer properties near Middle Branch and McDonalds Branch sites. Daily mean streamflow for McDonalds and Middle Branches is plotted in figure 33. The records for each station are concurrent during the period between mid-April and mid-October 1993. The hydrographs are characterized by many small recharge events during the spring. Multiple recharge events that occur over a relatively short period tend to mask out recessive behavior and hamper the determination of a representative base-flow recession rate. Consequently, the extrapolated recession curves that were developed for each stream (fig. 33) are based on a composite of many short recession periods. The recession curve for Middle Branch has a steeper slope than does the curve for McDonalds Branch (40 days per log cycle and 63 days per log cycle, respectively) and indicates that the aquifer adjacent to McDonalds Branch has a greater storage capacity than does the aquifer adjacent to Middle Branch.

Rorabaugh (1964) showed that the volume of ground water remaining in storage (*V*) after the critical time is reached can be determined by the a relation that is expressed in a simplified form as

$$V = \frac{Qt_c\Delta t}{2.3} , \tag{5c}$$

where Qt_c is ground-water seepage, in cubic feet per second, to the stream at the critical time. The amount of water added to ground-water storage following a recharge event can be determined from equation 5c as the difference between the base-flow recessions (Qt_c) at the critical times before and

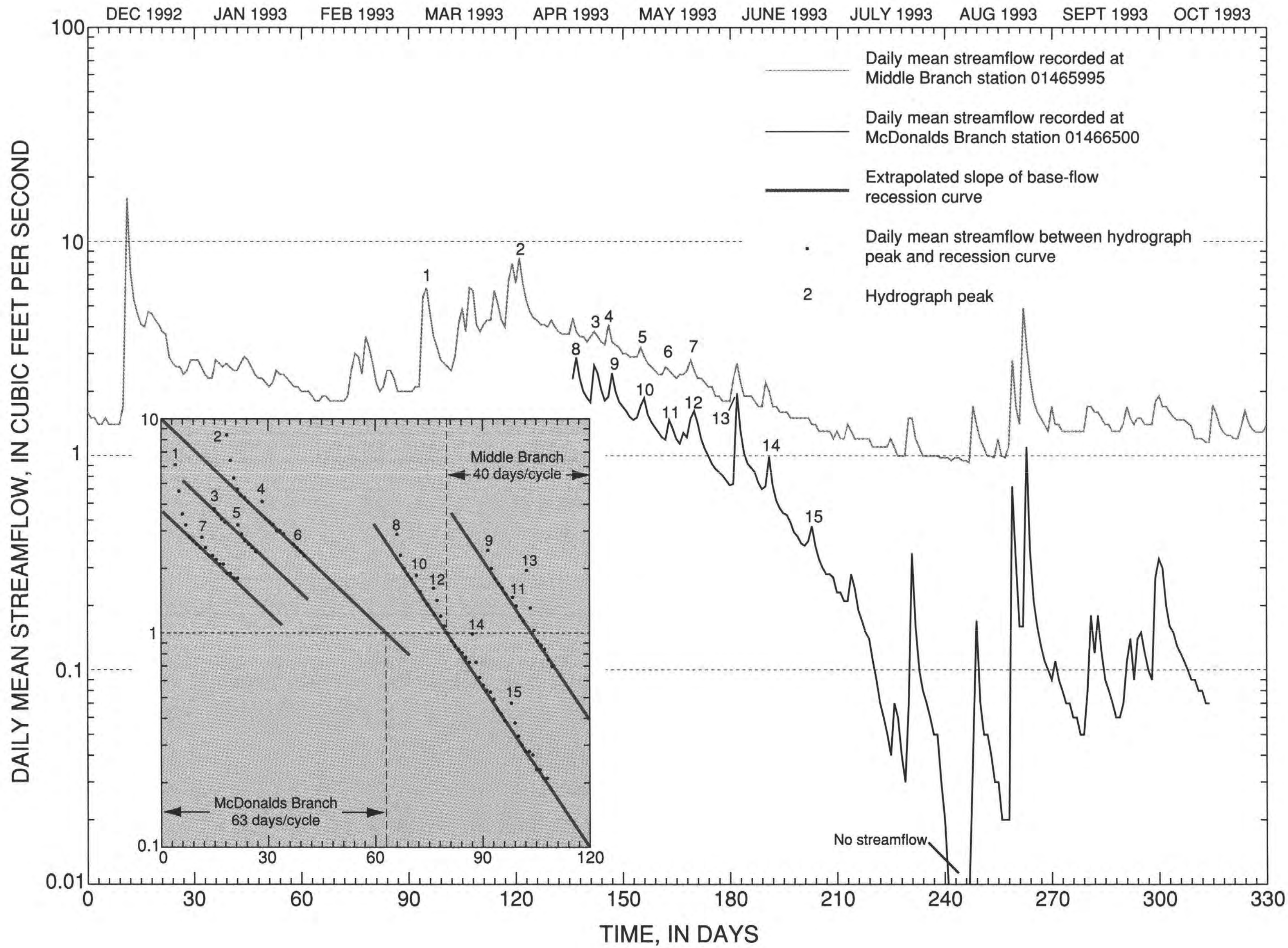


Figure 33. Hydrographs of daily mean streamflow and base-flow recession curves for Middle Branch and McDonalds Branch gaging stations, New Jersey.

after a recharge event. The critical time between peak recharge and the commencement of base-flow recession for the Middle Branch and McDonalds Branch hydrographs is calculated as 8.5 days and 13.5 days, respectively (from eq. 5a). The base-flow recession rate for each stream was determined from the curves after the critical times from the recharge events of May 6 and 20, 1993, (labeled as 5 and 7 on the McDonalds Branch hydrograph and as 10 and 12 on the Middle Branch hydrograph).

The calculated volume of ground water that went into storage after the May 20 recharge event was 5.2 ft^3/d for Middle Branch and 13.7 ft^3/d for McDonalds Branch. Assuming that the extrapolated curves are reasonable representations of base flow in these systems, the calculated volumes suggest that the aquifer near McDonalds Branch has about 2.6 times the storage capacity of the aquifer near Middle Branch.

The higher storage value calculated for McDonalds Branch than for Middle Branch is consistent with the relatively small ranges of water-level changes measured in McDonalds Branch compared to those measured in Middle Branch (fig. 31). Although water levels in Middle Branch seem to be more sensitive to recharge than those in McDonalds Branch, less water is actually stored in the aquifer. This would imply that the area of influence generated by a pumped well would propagate out farther from the well if pumpage took place near Middle Branch rather than near McDonalds Branch. The effects of storage in terms of specific yield on the size of the area of influence were shown in figure 19.

An example of the differences in the aquifer responses to recharge events near Middle Branch and McDonalds Branch is the record of changes in stream stage through time that result from a precipitation event (fig. 34). A total of 2.19 in. of rain was recorded by the Middle Branch rain gage over a 12-hour period during August 6 and 7, 1993. Of the total recorded rainfall, 1.59 in. fell within the first hour. Assuming that the precipitation patterns were similar at the McDonalds Branch site, the stage at McDonalds Branch peaked within 2 hours of the start of precipitation (at 1,000 hours) before going into a decline. The stage at Middle Branch peaked more than 16 hours after the start of precipitation and remained at peak level for 4 more hours before declining.

To determine whether bank storage has an effect on streamflow, hydraulic heads were measured at 15-minute intervals in piezometer 05-1228 at Middle Branch site (fig. 31A). After 15 hours from the start of rainfall, heads in the shallow aquifer increased by nearly 0.6 ft, from 121.06 to 121.64 ft (fig. 34C). However, the increase in streamflow never raised the stream stage to a level greater than the heads in the shallow aquifer. Consequently, streamwater levels never increased significantly enough, compared to hydraulic heads in the aquifer, to generate bank storage, and ground-water seepage conditions persisted in the headwaters of Middle Branch–even during an intense recharge event. The increase in hydraulic heads in the aquifer did, however, result in rewetting of the fringe area of the wetland by an estimated 13 ft on either side. The effects of the August 6 recharge event lasted for 10 days; that is, the water levels in the aquifer had declined to 121.14 ft by August 16, still nearly 0.1 ft higher than the water level 10 days earlier.

Practical Considerations for Installing Observation Wells Near Streams and Wetlands

To be successful, a monitoring strategy needs to include water-level measurements in areas where fluctuations are greatest. Water-level data collected for this study (fig. 26) show that the average water-level fluctuation in upland areas is 3.4 ft and that some water levels fluctuate as much as 5.5 ft. Although wells in upland areas do not provide information on the hydraulic

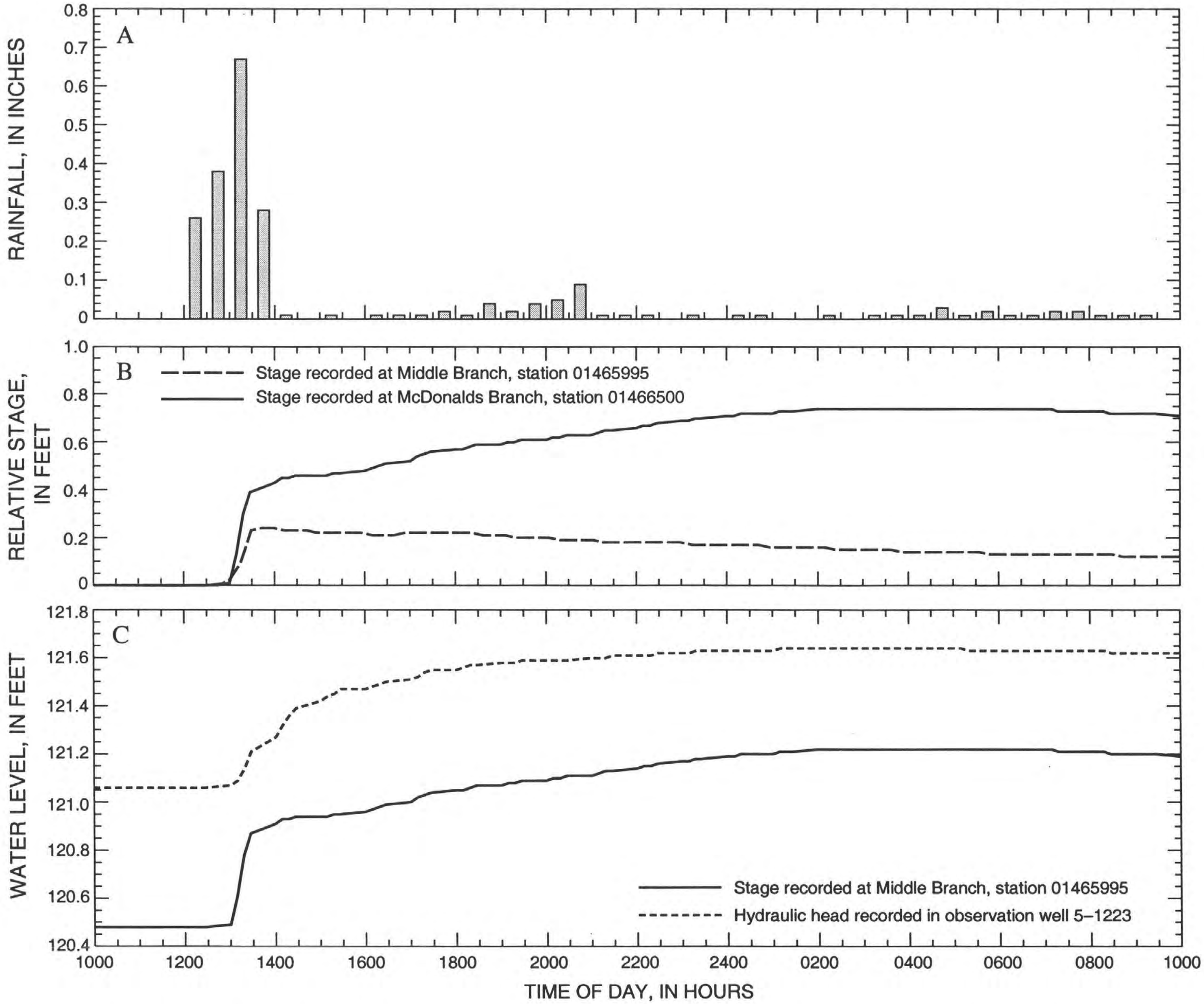


Figure 34. (A) Rainfall during the 24-hour period August 6-7, 1993, (B) responses of stages at Middle Branch and McDonalds Branch, New Jersey, and (C) responses of stage and hydraulic head in the shallow aquifer at Middle Branch.

condition of specific streams and wetlands, the measurements from these wells are more sensitive to natural and artificial stresses of the aquifer system in general than are measurements from lowland wells; hence, upland wells can be an integral part of a monitoring strategy. If heads in upland areas decline to much lower than baseline levels and if they persist or continue downward, then potential shortages in water supply may be signaled.

A successful monitoring strategy also needs to include measurements in wetland areas, especially those wetlands that are close to sites of potential development. Even though fluctuations of hydraulic head are lower than those in uplands, about 1.3 ft on average, the potential effects on specific systems is learned by monitoring these systems. The basic concepts of ground-water seepage and hydroperiods of the stream-wetland systems described in this report could be useful in the design of a monitoring network in streams and wetlands of the Coastal Plain of New Jersey. A conceptual model of stream-wetland system and its hydroperiodic relations is shown in figure 35. The zone adjacent to the streams is an area in which ground-water seepage is continuous during the year. Ground water seeps through the fringe area around the periphery of the system for only part of the year. The hydroperiodic patterns could be mapped for specific systems in areas of concern where it is important to determine the natural ground-water seepage patterns and to monitor any changes over time that may be caused by nearby withdrawal of ground water.

Hydroperiodic data in wetlands can be determined by measuring heads in piezometers screened several feet in the sand beneath the wetlands. Piezometers could also be installed along the stream near its headwaters to track the seasonal fluctuation in the start of flow and headward movement of the seepage face. The centers of wetlands are least sensitive to changes in hydraulic head, and the ground-water seepage rates are generally lower there than seepage near the edges of the channel. Fringe areas are most sensitive to changes in hydraulic conditions. Seepage rates are also higher near the edge in wide channels. Consequently, sampling devices would ideally be installed along the fringe area between the uplands and the wetlands.

To establish the hydroperiods that represent long-term water-level patterns of a wetland system, it is necessary to measure the seasonal maximum and minimum water levels over the year. These data become more representative of average conditions when obtained over many years. Ideally, measurements would be made during the late winter–early spring and during the late summer–early autumn to establish the times when the highest and lowest water levels occur. From these data, average water-level fluctuations in wetlands could then be established.

Generally, a piezometer need only be installed in the sand a few feet under the sand-muck interface to measure the hydraulic head beneath a wetland, provided that aquifer material is relatively homogeneous. For the piezometers installed in this study, no measurable differences in heads were found in the range of 2 and 10 ft below land surface. The screen depth of the piezometer is best determined by the purpose of the well. If the well is to be used for monitoring the water levels, then a screen depth of several feet is probably adequate. If, however, the well is to be installed as part of an aquifer test, then a deeper setting–probably 10 to 30 feet–would be best for measuring the anticipated drawdown. Narrow screened intervals are used to track subtle differences in heads, such as exist beneath seepage channels. However, as a practical matter, wider intervals on the magnitude of 1 ft are better for general purpose because they average out the variation in heads over a vertical length that result from variation in grain size.

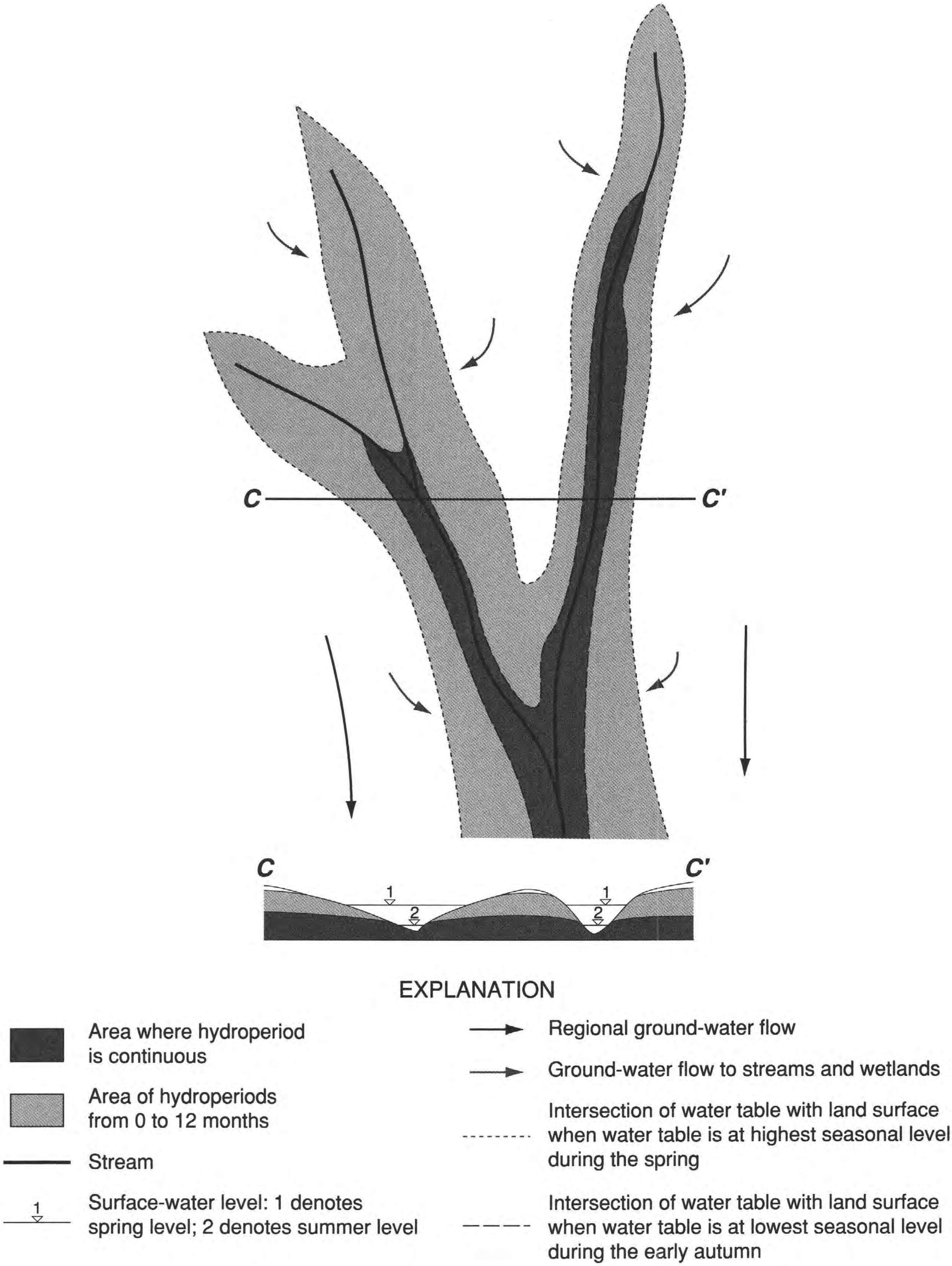


Figure 35. Hydroperiodic relations in a hypothetical stream-wetland system characteristic of the Coastal Plain of New Jersey.

SUMMARY AND CONCLUSIONS

Analytical and numerical solutions to ground-water withdrawals were evaluated for their usefulness in predicting the area of influence of a pumped well and in determining hydraulic characteristics of an aquifer. The hydraulic concepts derived were used to provide supplementary guides regarding withdrawal strategies in unconfined aquifers near streams and wetland systems of the New Jersey Coastal Plain. Results of simulations made with a ground-water flow model of the area near Chesilhurst, in the northeastern part of the Mullica River Basin, were used to furnish information on prudent well-placement strategies and on the distribution of effects of withdrawal over the system.

The design of a monitoring network sensitive to the ground-water hydraulics of streams and wetlands of the Coastal Plain also was considered. A monitoring network, established from preexisting wells and newly installed piezometers in wetland areas was used to determine yearly water-table fluctuations in the Kirkwood-Cohansey aquifer system in part of the Mullica River Basin of the Coastal Plain of New Jersey. Results from the evaluation include the following:

1. Application of methods based on the Theis analytical solution to unconfined aquifers can lead to erroneous estimates in the size of the area of influence generated from ground-water withdrawals. Use of the Theis solution also can lead to the determination of hydraulic characteristics that misrepresent the water-transmitting properties of the aquifer. Because the effects of gravity drainage vary with depth and azimuthal distance from the pumped well in an unconfined aquifer, the total hydraulic effects produced in an unconfined aquifer by withdrawal are not adequately accounted for by the Theis solution, which requires the assumption of horizontal flow. The commonly used distance-drawdown method can give results resembling those produced with the Neuman solution. However, because the effects of delayed yield are not fully dissipated from the system even after 72 hours of pumping, the method may produce results that are misleading.

2. Analysis of time-drawdown data from the Kirkwood-Cohansey unconfined aquifer system is best conducted with the Neuman solution because it accounts for the effects of gravity drainage. Aquifer-test analyses are also sensitive to the position of screens of observation wells in the aquifer, particularly near the water table. The Neuman solution is applicable to pumped wells that are far enough from streams so as not to draw ground-water flow away from the streams. Time-drawdown data from more than one observation well are best analyzed compositely in order to determine aquifer parameters that are spatially averaged and representative of the aquifer surrounding the test site. The composite analysis of pumped-well data also make anomalous data patterns more apparent and help the interpreter avoid erroneous conclusions.

   Time-drawdown data from a test well in Winslow Township, N.J., were analyzed by means of the Neuman solution. The analysis indicates that the aquifer has a relatively high vertical to horizontal anisotropy of 1:198, a horizontal hydraulic conductivity of 95.1 ft/d, and a vertical hydraulic conductivity of 0.48 ft/d; the specific yield of the aquifer is 0.028. The aquifer characteristics indicate that the aquifer is somewhat stratified and can have relatively low yield near the test site.

3. The area of influence for a long-term withdrawal is best estimated from a steady-state ground-water-flow analysis that includes average areal recharge. Ground-water-flow simulations for the northwestern part of the Mullica River Basin indicate an order-of-

magnitude difference between the area of influence produced from a 3-day (72-hour) withdrawal and the area produced under steady-state conditions. Thus, a 3-day aquifer test is not suited for extrapolating a steady-state configuration of the area of influence but is best applied to establishing the hydraulic characteristics of the aquifer and the propagation rates of the area of influence. The simulations also show that the area of influence is sensitive to the specific yield of the aquifer; a decrease in specific yield will cause the area of influence to increase.

4. The contributing area of flow to the pumped well includes areas on the water table that would, under prestress conditions, be incorporated into the contributing areas of flow to streams. Because recharge must equal discharge in a steady-state ground-water-flow system and because all incoming water in the northwestern part of the Mullica River Basin originates as areal recharge at the water table, ground-water discharge to the well in this area is at the expense of seepage to streams, and the withdrawal results in a reduction in the size of the contributing areas to streams by an amount equal to the contributing area of the withdrawal. The contributing area of flow to a well is a distinct hydraulic phenomenon from drawdown, and the two do not necessarily affect the same aquifer space.

5. Withdrawals made from a well close to a stream divert ground water to the well that would, under natural conditions, flow to the stream. The diverted ground water functions as a recharge boundary and produces an area of influence that is smaller than one produced in an aquifer of "infinite extent." Diverted ground water offsets the drawdown more on the stream side of the pumped well than on the far-field side of the well. Results from flow simulations of the northwestern part of the Mullica River Basin show that the reduction in ground-water seepage caused by withdrawal depends on pumping rate and distance between the pumped well and stream. The area of influence increases with increased pumping rate but decreases with proximity to the stream. Although potential water-table declines caused by withdrawal near streams are, to some degree, mitigated by ground-water diversion from streams, withdrawals can reduce ground-water seepage to streams significantly if the pumped well is near streams and wetlands. However, the effects of withdrawal are dispersed over a larger part of the aquifer system whereby more streams are affected but much less significantly if pumped wells are located as far as possible from streams, on surface-water divides.

6. Measurements of seasonal fluctuations in the Mullica River Basin and surrounding areas indicate that the greatest water-level fluctuations are in upland areas, where fluctuations average 3.4 ft and can be as great as 5.5 ft. Smaller fluctuations are measured in wetland areas, where fluctuations average 1.3 ft above land surface. The bimodal average fluctuation in heads in uplands and lowlands indicate that upland areas are more sensitive to recharge than lowland areas. Local variations in fluctuations of water levels are caused by variation in grain size and in distribution of aquifer material. The long-term pattern of yearly mean water levels indicate that heads fluctuate irregularly about a long-term mean value and that abnormally low or high yearly average values, induced by deficits or excesses in recharge, are short lived. Over time, the hydraulic conditions tend to move back to average levels as long as natural conditions are the only factors controlling the hydrologic budget.

Wetland areas in the New Jersey Coastal Plain are characterized by ground water that seeps into topographic depressions. The period of inundation is longest and often perennial in the deepest part of the depression. A fringe area along the borders of wetlands is characterized by seasonally intermittent ground-water seepage. The seepage face of the fringe area expands and contracts in response to seasonal variation in the water table and in response to precipitation.

7. The hydraulic response of the shallow aquifer to recharge provides information on aquifer storage, water-transmitting properties, and rates of water-level change. The determination of the aquifer storage coefficient and transmissivity can be attained by means of hydraulic head or streamflow recession analysis. Such analysis can, in some cases, provide a method of determining aquifer properties that is more economical than aquifer testing. The recession curves developed from hydrographs of Middle Branch and McDonalds Branch indicate that the aquifer near McDonalds Branch has about 2.6 times the storage capacity of the aquifer adjacent to Middle Branch. This finding is consistent with the relatively small ranges of water-level changes measured in McDonalds Branch compared to those measured in Middle Branch.

8. Monitoring networks that consist of piezometers screened in the shallow aquifer can provide hydrologic data necessary to establish the baseline hydrologic conditions in an aquifer system and can signal when flow conditions are affected by ground-water use elsewhere in the system and the degree to which the effects of withdrawal are distributed in the system. A monitoring strategy needs to include water-level measurements in upland areas where the range of water-level fluctuations is greatest. Measurements from upland areas are more sensitive to natural and artificial stresses of the aquifer system in general than are measurements from lowland areas. For monitoring water levels in and adjacent to wetlands, it is generally sufficient to install piezometers in a shallow aquifer in the sand underlying the organic deposits. The level of the water table at fringe of wetland is good indication of hydraulic head and seepage conditions beneath the wetland. However, clay layers that are close to the land surface can create localized hydraulic conditions in the aquifer that may not be representative of hydraulic conditions elsewhere in the aquifer. Monitoring wells installed near the site of withdrawal can be useful for monitoring potential reductions of water levels.

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