Evaluation of Methods for Delineating Areas That Contribute Water to Wells Completed in Valley-Fill Aquifers in Pennsylvania

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CONTENTS

CONTENTS--Continued

Page

ILLUSTRATIONS

Page

TABLES

CONVERSION FACTORS, VERTICAL DATUM, AND SYMBOLS USED IN EQUATIONS

Multiply	By	To obtain
	Length	
inch	25.4	millimeter
foot	0.3048	meter
mile	1.609	kilometer
	Area	
square foot	0.0929	square meter
square mile	2.590	square kilometer
	Volume	
gallon	3.785	liter
cubic foot	0.02832	cubic meter
	Flow	
cubic foot per second	0.02832	cubic meter per second
gallon per minute	0.06309	liter per second
	Other Conversions	
foot per mile	0.1894	meter per kilometer
foot per day	0.3048	meter per day
foot squared per day	0.0929	meter squared per day
cubic foot per day per square foot times foot of aquifer thickness	0.0929	cubic meter per day per square meter times meter of aquifer thickness

Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

SYMBOLS USED IN EQUATIONS

Symbol	Dimensions	Definition
A	L	Twice the distance between the real well and impermeable boundary
B	L	Twice the distance between the real well and stream boundary
b	L	Aquifer thickness
d	L	Distance from well to stream
H	L	Aquifer width
h	L	Hydraulic head
h_s	L	Hydraulic head computed using superposition of drawdown on prepumping hydraulic head
i	--	Slope of potentiometric surface
I	--	Image well
K	L/T	Hydraulic conductivity
L	L	Width between the asymptotic limits separating ground-water flow to well
P	L	Distance from well to stagnation point
Q	L^3/T	Pumping rate from production well
Q_i	L^3/T	Pumping rate from image well
R	L	Radius of time-of-travel area
r	L	Well radius
S	--	Storage coefficient
s	L	Drawdown
s_w	L	Drawdown in pumping well
T	L^2/T	Transmissivity of aquifer
T_i	L^2/T	Transmissivity beyond hydrologic boundary, in the vicinity of the image well
t	T	Duration of pumping or traveltime of interest
V_a	L^3	Volume of water in aquifer pore space
V_p	L^3	Volume of water withdrawn by pumping
V_w	L^3	Volume of leakage or recharge
w	L/T	Recharge or leakage rate per unit surface area
∞	--	Infinity
τ	--	Dimensionless time
θ	--	Porosity

Evaluation of Methods for Delineating Areas That Contribute Water to Wells Completed in Valley-Fill Aquifers in Pennsylvania

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ABSTRACT

Valley-fill aquifers in Pennsylvania are the source of drinking water for many wells in the glaciated parts of the State and along major river valleys. These aquifers are subject to contamination because of their shallow water-table depth and highly transmissive sediments. The possibility for contamination of water-supply wells in valley-fill aquifers can be minimized by excluding activities that could contaminate areas that contribute water to supply wells.

An area that contributes water to a well is identified in this report as either an area of diversion, time-of-travel area, or contributing area. The area of diversion is a projection to land surface of the valley-fill aquifer volume through which water is diverted to a well and the time-of-travel area is that fraction of the area of diversion through which water moves to the well in a specified time. The contributing area, the largest of the three areas, includes the area of diversion but also incorporates bedrock uplands and other areas that contribute water.

Methods for delineating areas of diversion and contributing areas in valley-fill aquifers, described and compared in order of increasing complexity, include fixed radius, uniform flow, analytical, semianalytical, and numerical modeling. Delineated areas are considered approximations because the hydraulic properties and boundary conditions of the real ground-water system are simplified even in the most complex numerical methods.

Successful application of any of these methods depends on the investigator's understanding of the hydrologic system in and near the well field, and the limitations of the method. The hydrologic system includes not only the valley-fill aquifer but also the regional surface-water and ground-water flow systems within which the valley is situated. As shown by numerical flow simulations of a well field in a valley-fill aquifer along Marsh Creek Valley near Asaph, Pa., water from upland bedrock sources can provide nearly all the water contributed to the wells.

INTRODUCTION

More than 2 million people in Pennsylvania depend on ground water as a source of potable water (Solley and others, 1983, p. 10). Unfortunately, contamination of community-supply wells and springs from point and nonpoint sources is a problem throughout the State (Barker, 1988). Cleanup of contaminated ground water can be difficult and costly. The chance of ground-water contamination near public-supply wells and springs can be minimized by protecting the area surrounding these sources from activities that can adversely affect ground-water quality. This strategy is termed wellhead protection.

The 1986 Amendments to the Safe Drinking Water Act established the Wellhead Protection Program to protect ground water used for public drinking supplies from possible contamination (U.S. Environmental Protection Agency, 1989). The Amendments require each state to develop a wellhead-protection program. An effective program includes several essential elements, one of which is the delineation of wellhead-protection zones. These zones are defined in the Safe Drinking Water Act Amendments as "the surface and subsurface area surrounding a water well or well field, supplying a public water system, through which contaminants are reasonably likely to move toward and reach such water well or well field" (U.S. Environmental Protection Agency, 1987, p. 1-2). This definition is similar to the definition used by many hydrologists for the "contributing area" to a well.

The Pennsylvania Department of Environmental Resources is developing a wellhead-protection program. In cooperation with the Department, the U.S. Geological Survey (USGS) evaluated methods for delineating the contributing area to wells throughout the State. Several methods can be used to estimate the size and shape of a well's contributing area (U.S. Environmental Protection Agency, 1987). For example, a contributing area can be estimated by a simple circle of arbitrary radius drawn around the well or by a complex computer simulation of the ground-water-flow system.

Purpose and Scope

This report describes an evaluation and comparison of methods for delineating the contributing area to wells in valley-fill aquifers in Pennsylvania. The methods evaluated (fixed-radius, uniform-flow, analytical, semianalytical, and numerical methods) are limited to those that can be used for delineating the area of diversion, contributing area, and time-of-travel area of wells completed in unconsolidated valley-fill sediments.

The hydrology of an ideal valley-fill aquifer similar to many hydrogeologic settings in Pennsylvania is described, and the effects of pumping on this aquifer are discussed. Then, results of each method used in the ideal valley-fill setting and in a valley-fill aquifer along Marsh Creek Valley near Asaph, Pa., are compared.

Previous Investigations

Factors that control the response of an aquifer system to withdrawals of ground water were discussed by Theis (1940), who showed that withdrawals by wells must be balanced by an increase in natural recharge, decrease in natural discharge, decline in storage, or a combination of these effects. Other investigators have restated and expanded Theis' discussion. Brown (1963), in a discussion of the source of water to wells, showed the response of an aquifer to ground-water withdrawals. He emphasized the difference between the area of diversion of a well and its area of influence. Bredehoeft and Young (1970), Lohman (1972, p. 62), and Bredehoeft and others (1982) summarized Theis' ideas by means of case studies and numerical simulations.

Determination of the contributing area of a well has been investigated by use of analytical, semianalytical, and numerical-modeling techniques. Jacob (1950, p. 344) and Bear (1979, p. 282) present the fundamental analytical equations for flow to a well in a uniform-flow field. Horsley (1983) used the uniform flow-field equation, aquifer geometry, and vertical hydraulic properties to delineate contributing areas. A semianalytical model of transient ground-water flow was used by Nelson (1978) to compute streamlines and time-of-travel positions for injected water. Keely and Tsang (1983) used a similar approach to illustrate the movement of injected wastewater. Examples of the use of numerical computer models to delineate contributing areas include those by Camp, Dresser, and McKee, Inc. (1982) and Reiter (1985).

At least two investigators evaluated and summarized some of the techniques being used to delineate wellhead-protection areas. Morrissey (1989) discussed factors that control the contributing area to wells in typical valley-fill aquifers in New England. He evaluated and compared methods that are commonly used to determine contributing areas and source of water to wells. The U.S. Environmental Protection Agency (1987) provided guidelines for delineating wellhead-protection areas. These guidelines include an evaluation of assumptions, data requirements, and technical merits of methods that can be used to estimate contributing areas of wells.

DESCRIPTION OF VALLEY-FILL AQUIFERS

The typical unconsolidated aquifer used for public supply in Pennsylvania consists of sediments deposited as valley fill by glaciers, lakes, and streams. Although unconsolidated sediments of the Coastal Plain in Bucks, Philadelphia, and Delaware Counties commonly yield greater than 1,000 gallons per minute to wells, substantial quantities of ground water from these sediments are not being withdrawn for public supply (Joseph Lee, Pennsylvania Department of Environmental Resources, Bureau of Community Environmental Control, written commun., 1990). Therefore, unconsolidated aquifers in glaciated and unglaciated valleys are the focus of this study.

Location and Distribution

Of the nearly 3,000 listings of public-supply wells in the USGS ground-water site-inventory (GWSI) data base for Pennsylvania, less than 300 were completed in valley-fill aquifers. Although total statewide withdrawals from these aquifers are small, they are an important local source of public-supply water that can be easily contaminated because of their shallow water-table depth and their large transmissivity.

Stratified deposits that comprise the major valley-fill aquifers in Pennsylvania are located in the glaciated 30 percent of the State and along major rivers such as the Allegheny, Susquehanna, Ohio, and Delaware. The distribution of sediments that comprise major aquifers is shown in figure 1.

Physical Properties

Valley-fill aquifers extend along valley axes, either partly filling a valley or completely filling a preglacial drainage way. They range in width from several hundred feet to more than 2 miles in Erie and Crawford Counties, although widths of about 4,000 feet are typical throughout the State. The thickness of valley-fill sediments is as much as 500 feet in Crawford County (Schiner and Gallaher, 1979, p. 9), but throughout the State is usually less than 200 feet. The depth of wells drilled in valley-fill sediments can be used to estimate minimum sediment thickness. These depths will be less than the actual thickness because not all wells are drilled to bedrock. Depths of 150 wells drilled in valley-fill sediments range from 10 to 217 feet; median depth is 50 feet (fig. 2A).

Typically, the valley fill is a complex assemblage of unstratified glacial drift, stratified glacial drift, and alluvium. The unstratified drift (till) is composed of an unsorted mixture of clay, sand, gravel, and boulders that was deposited beneath and at the margins of the glacial ice. Till generally yields only small quantities of water to wells. Stratified drift is composed of sediments that have been sorted by water. The sorting may have occurred in lakes (lacustrine deposits) or flowing water (glaciofluvial deposits). Lacustrine deposits, chiefly silt and clay, are not major water-yielding units; in contrast, glaciofluvial deposits are highly transmissive and commonly yield several thousand gallons of water per minute to wells. Glaciofluvial deposits include sediments left by water flowing within or under the ice (ice-contact deposits) or by streams carrying sediment-laden meltwater from the glacial terminus (outwash). Alluvium is gravel, sand, silt, and clay deposited by a stream and its tributaries. In glacial valleys, much of the alluvium consists of reworked glacial deposits.

The internal structure of valley-fill sediments is controlled by the topographic setting of the valley in relation to the position of glacial ice and direction of meltwater flow. Several hydrophysiographic terranes in Pennsylvania that are based on these characteristics (fig. 3) were outlined by Randall and Johnson (1988, p. 5). A typical section across the Shenango River in northern Mercer County in a glaciated valley where meltwater drained away from the ice sheet (hydrophysiographic terrane 1) illustrates the internal complexity within valley-fill sediments (fig. 4). The discontinuous nature of transmissive ice-contact deposits and outwash shown in figure 4 results in heterogeneity and anisotropy with respect to the hydraulic properties of this valley-fill aquifer.

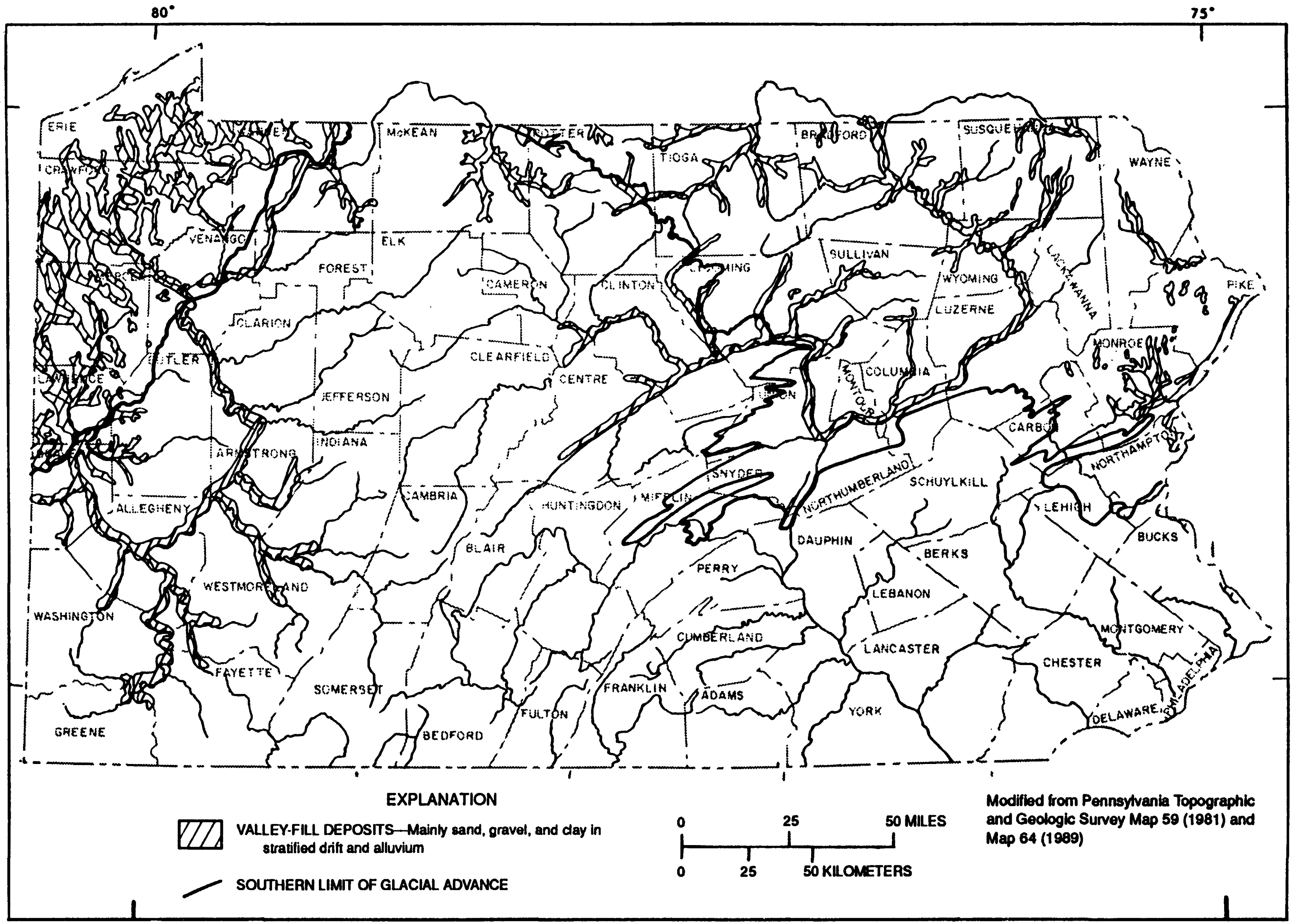


Figure 1. Distribution of major valley-fill aquifers in Pennsylvania.

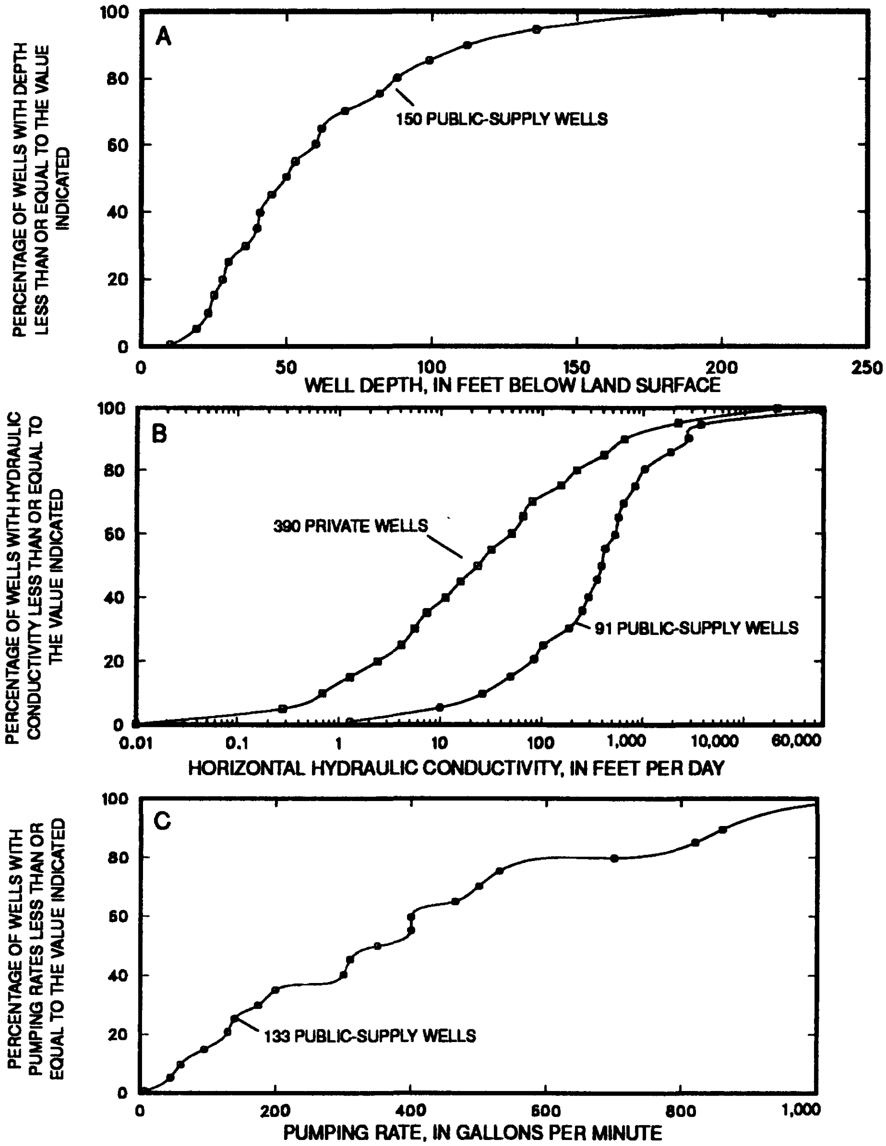
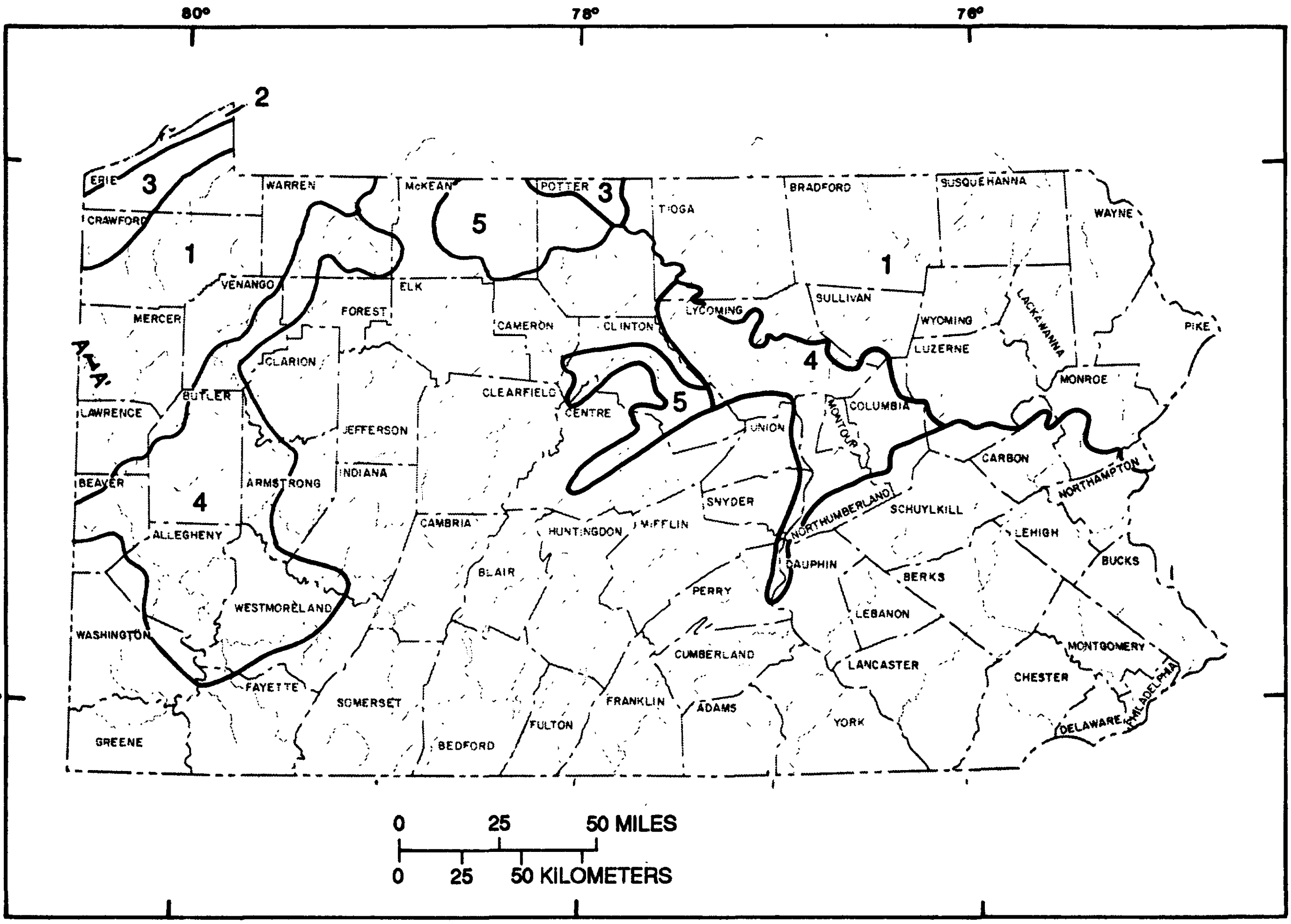


Figure 2. Cumulative frequency of (A) well depth, (B) horizontal hydraulic conductivity, and (C) pumping rate for selected wells completed in valley-fill sediments.

Hydraulic Properties

The hydraulic properties of valley-fill sediments that describe their ability to transmit and store ground water include hydraulic conductivity, specific yield, and specific storage. Hydraulic conductivity differs widely within the valley-fill deposits in Pennsylvania (fig. 2B). Lyford and others (1984, p. 12) report that glacial valley-fill aquifers in the northeastern United States have horizontal hydraulic conductivities of 1 to 13,300 feet per day.

Horizontal hydraulic conductivities of valley-fill sediments in Pennsylvania are estimated to range from 0.01 to 58,000 feet per day on the basis of data from 91 public-supply and 390 private wells (fig. 2B). The values were computed from specific-capacity data and estimates of saturated aquifer thickness described by Theis (1963) and applied by Bradbury and Rothschild (1985). Extremely large hydraulic conductivities (greater than about 2,000 feet per day) probably were caused by the effects of nearby recharge boundaries. The median horizontal hydraulic conductivity is 388 feet per day at public-supply wells and 23 feet per day at private wells. In addition, variability in hydraulic conductivity is less at



EXPLANATION

BOUNDARY OF HYDROPHYSIOGRAPHIC TERRANE

1 GLACIATED VALLEYS THAT SLOPED OR DRAINED FROM THE ICE SHEET—Contain complex sequence of stratified sediments

2 GLACIATED LOWLANDS FORMERLY INUNDATED BY LAKE ERIE—Contain some productive ice-contact deposits within fine-grained lacustrine sediments

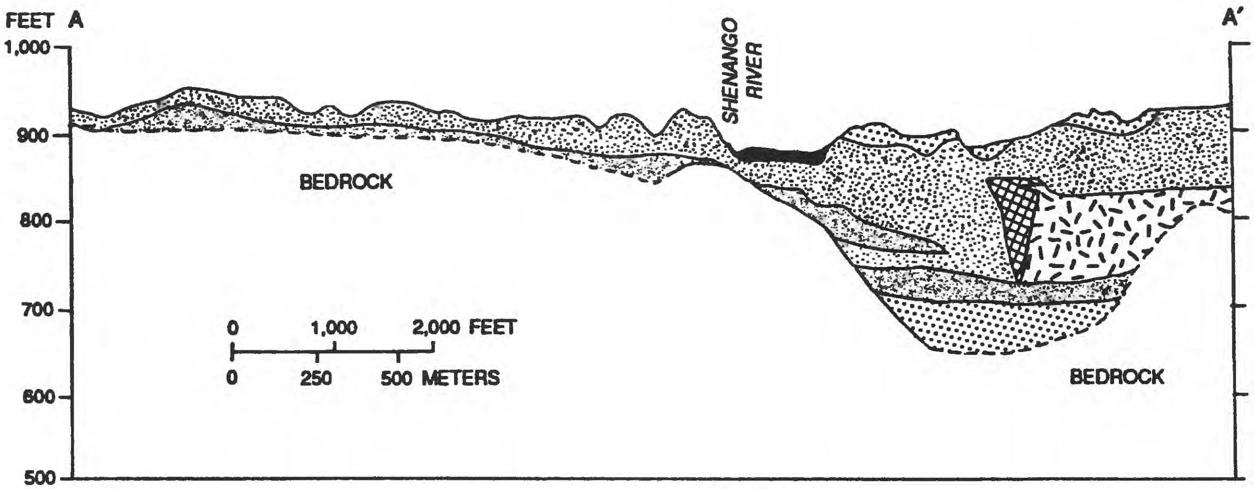
3 GLACIATED VALLEYS THAT DRAINED TOWARD THE ICE SHEET—Contain fine-grained sediments, till, and extensive sandy delta deposits

4 VALLEYS NOT GLACIATED IN MOST RECENT ICE ADVANCE THAT DRAINED AWAY FROM ICE—Contain chiefly sandy outwash

5 VALLEYS NOT GLACIATED IN MOST RECENT ICE ADVANCE THAT DRAINED TOWARD ICE—Contain chiefly fine sediments

A—A' LOCATION OF GEOLOGIC SECTION SHOWN IN FIGURE 4

Figure 3. Hydrophysiographic terranes in the glacial deposits of Pennsylvania. (Modified from Randall and Johnson, 1988, fig. 2.)



VERTICAL EXAGGERATION × 10
DATUM IS SEA LEVEL

EXPLANATION

ALLUVIUM—Chiefly sand and gravel deposited by the Shenango River. May include some glacial outwash

OUTWASH—Chiefly sand and gravel deposited by glacial meltwater

ICE-CONTACT DEPOSITS (?)—Sand and gravel; mode of deposition uncertain

TILL—Unsorted glacial deposits of clay, sand, and boulders

LACUSTRINE DEPOSITS—Laminated silt and clay

DELTAIC DEPOSITS (?)—Clay, silt, and fine sand from streams flowing into a lake

CONTACT BETWEEN LITHOLOGIES—Dashed where approximate

Figure 4. Complex internal structure of valley-fill deposits in the Shenango River Valley, Mercer County.

public-supply wells than at private wells. The reason for these differences is that public-supply wells are sited, drilled, completed, and developed to yield large water supplies; therefore, aquifer tests at these wells probably measure the hydraulic conductivity of the most permeable parts of the valley fill. In contrast, private wells probably represent a fairly random sampling of valley-fill materials because many of these wells were drilled for domestic supply, where proximity to the dwelling was more important than a large yield.

Specific yields of unconsolidated sediments range from almost zero to about 50 percent (Davis and Dewiest, 1966, p. 376); however, typical specific yields for valley-fill sediments in Pennsylvania are probably close to the range of 5 to 35 percent reported by Lyford and others (1984, p. 12).

Specific storage is the volume of water released from or taken into storage per unit volume of aquifer per unit change in head. Specific storage typically is about 1×10^{-6} times the thickness of the confined aquifer (Lohman and others, 1972, p. 13).

Recharge and Discharge

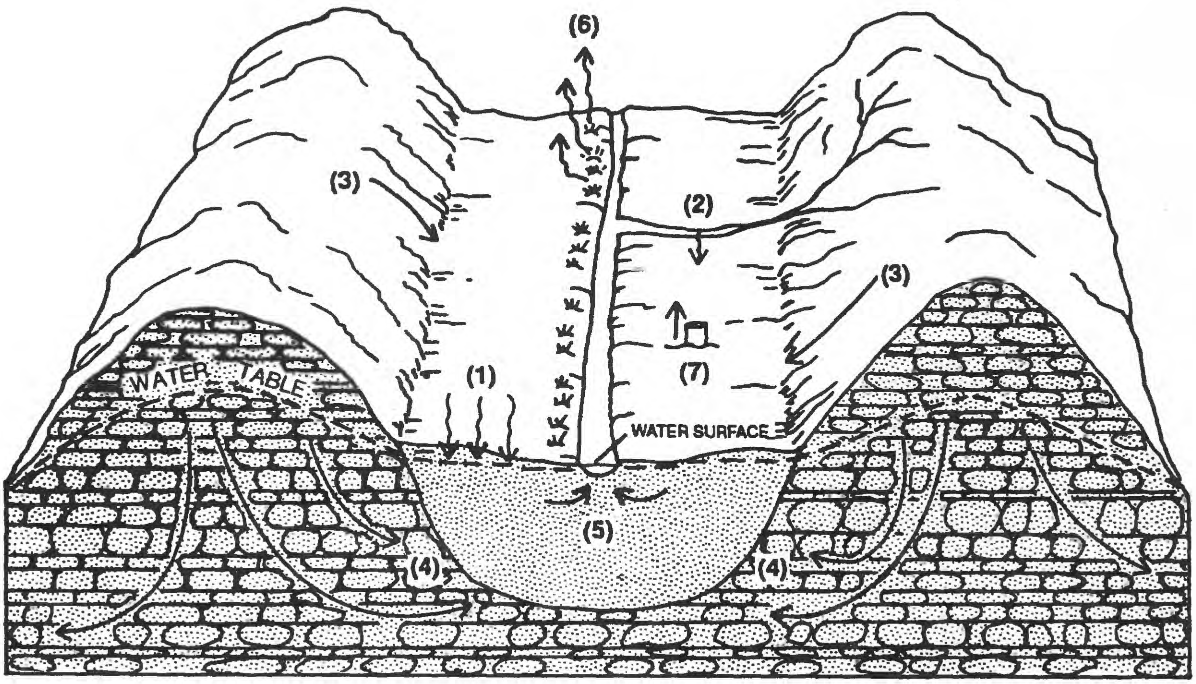
Sources of recharge to valley-fill aquifers include: (1) direct infiltration of precipitation on the valley floor; (2) seepage from streams that drain upland areas where they cross the valley; (3) unchanneled overland runoff and interflow from upland areas; and (4) regional ground-water flow from bedrock aquifers. Ground water discharges to (1) streams and springs, (2) the atmosphere by evapotranspiration, and (3) wells. The relative importance of each component of the ground-water budget depends on the geometry of the valley-fill aquifer and topography of surrounding bedrock uplands. The major recharge and discharge components of an idealized valley-fill aquifer are illustrated in figure 5. Individual components are discussed in detail in the following paragraphs.

Direct precipitation on valley floor.--Recharge from direct infiltration of precipitation on the valley surface (fig. 5, component 1) is the difference between precipitation and sum of overland runoff and evapotranspiration. Overland runoff is probably negligible where the valley floor is directly underlain by sand and gravel and the water table is not at land surface (Lyford and Cohen, 1988). Precipitation ranges from 34 to 46 inches per year in areas of Pennsylvania that contain major unconsolidated aquifers. If about 50 percent of the precipitation is lost to evapotranspiration, as has been estimated for the upper Susquehanna River Basin (Taylor, 1984, p. 11), then recharge by direct infiltration ranges from 17 to 23 inches per year. Recharge could be less than 17 inches per year in areas where the valley floor is directly underlain by clayey sediments or where the water table is at land surface.

Tributary-stream seepage.--Recharge by seepage from streams that cross the valley floor but drain upland areas (fig. 5, component 2) typically contributes 60 to 75 percent of the total recharge to valley-fill aquifers in regions of high topographic relief (Morrissey and others, 1988). Most of Pennsylvania's valley-fill aquifers, except those in the northwestern corner of the State, are in areas of high topographic relief. The rate of recharge depends on streamflow, but typical rates average more than 2.5 cubic feet per second per linear mile of valley (Morrissey and others, 1988).

Overland runoff and interflow.--Much of the precipitation falling on bedrock uplands adjacent to valley-fill aquifers either runs down the steep-sided valley walls as overland runoff or interflow (fig. 5, component 3). Interflow refers to that water flowing downslope beneath the land surface and above the water table. Recharge from this source is difficult to quantify, but estimates indicate that it is a major component of the water budget of valley-fill aquifers. Data compiled by Gebert and others (1987) indicate that 16 to 25 inches of water per year are available for recharge from unchanneled upland areas. Other estimates range from 13 to 21 inches per year (Morrissey, 1989; John Williams, Larry Taylor, and Dennis Low, U.S. Geological Survey, written commun., 1989).

Regional flow of ground water from bedrock.--Ground-water flow from adjacent bedrock aquifers (fig. 5, component 4) can recharge valley-fill sediments, although the amount is difficult to quantify. Differences in chemistry of water in bedrock and in valley-fill sediments have been used to estimate contributions



Not to scale

EXPLANATION

BEDROCK

VALLEY-FILL SEDIMENTS

DIRECTION OF GROUND-WATER FLOW

GROUND-WATER RECHARGE TO AND DISCHARGE FROM THE VALLEY-FILL AQUIFER

RECHARGE

(1) Direct precipitation on valley floor
(2) Tributary-stream seepage
(3) Overland runoff and interflow
(4) Regional ground water from bedrock

DISCHARGE

(5) Streams and springs
(6) Evapotranspiration
(7) Well

Figure 5. The major components of recharge and discharge in an idealized valley-fill aquifer.

from regional flow. Poth (1963, p. 76) stated that the wide range in ionic composition of ground water in glacial and other sediments in parts of Mercer, Butler, and Lawrence Counties resulted from local additions of water from underlying bedrock.

Discharge to streams and springs.--Ground water in valley-fill sediments discharges naturally to streams and springs (fig. 5, component 5). Most ground water discharges to a major stream that flows along the valley axis. Underflow beneath the stream is probably negligible in narrow valleys but, according to Randall and Johnson (1988), can be several cubic feet per second in broad valleys.

Evapotranspiration.--Evapotranspiration of ground water (fig. 5, component 6) can account for 1 to 9 inches of discharge annually (Lyford and others, 1984, p. 12). The importance of evapotranspiration changes seasonally as size of wetlands and uptake by vegetation vary (Hewlett and Nutter, 1970).

Discharge to wells.--Ground-water withdrawal from wells (fig. 5, component 7) can be an additional discharge from valley-fill aquifers. Wells drilled for public supply in valley-fill aquifers in Pennsylvania yield from 7 to 3,000 gallons per minute; median yield is 350 gallons per minute (fig. 2C).

DEFINITIONS OF CONTRIBUTING AREA AND RELATED TERMS

When an aquifer is stressed by pumping, a three-dimensional cone of depression is created around the well. This volume, sometimes referred to as the cone of influence, is defined by the extent of drawdown caused by the pumping (Theis, 1938, p. 891) as shown in figure 6A. The surface area around the well where drawdown is measurable is the well's area of influence (fig. 6B).

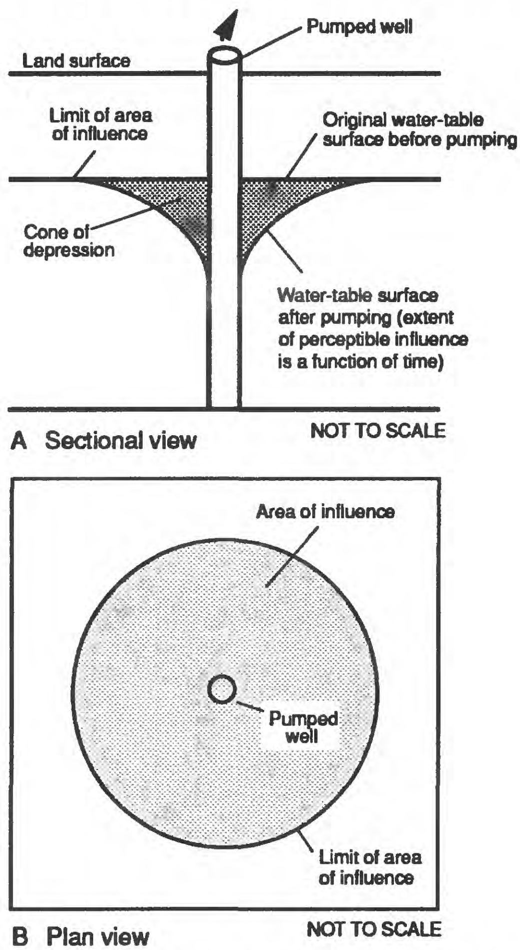


Figure 6. Diagrams of a pumped well showing (A) sectional view of the cone of depression in an unconfined aquifer and (B) plan view of the area of influence (from Morrissey, 1989, fig. 6).

Measuring the limits of the area of influence is difficult. In theory, the initial influence of the pumping expands at the speed of sound and rapidly reaches the physical boundaries of the aquifer. In practice, a long period of pumping is generally required to produce measurable drawdown at a significant distance from the well.

The area of influence is not a delineation of the area through which water is contributed to a well. The terms "area of diversion," "contributing area," and "time-of-travel area" are used in this report to define such areas. Several authors have used other terms to define similar areas or volumes around the well. Because of the subtle differences in these terms and their inconsistent usage in wellhead-protection studies, a brief discussion of the terms as used in this report follows.

The zone of diversion is the three-dimensional aquifer volume through which water is diverted to the well. This volume also has been termed "capture zone" (Keely and Tsang, 1983). The projection of this volume to land surface defines the well's area of diversion (Brown, 1963). The contributing area is the area of diversion plus any adjacent surface areas that provide recharge to the aquifer within the zone of diversion (Morrissey, 1989). Because this definition includes surface areas adjacent to the area of diversion, the contributing area can be much larger than the well's area of diversion.

The zone of diversion and area of diversion are restricted to the extent of the valley-fill aquifer itself. The contributing area includes adjacent upland bedrock areas and any other areas that provide recharge to the well but are not part of the valley-fill aquifer. The differences in these terms is important, because most methods used to delineate the contributing area to a well actually provide only a delineation of its area of diversion. Delineation of the areas contributing water from adjacent bedrock uplands usually must be estimated indirectly.

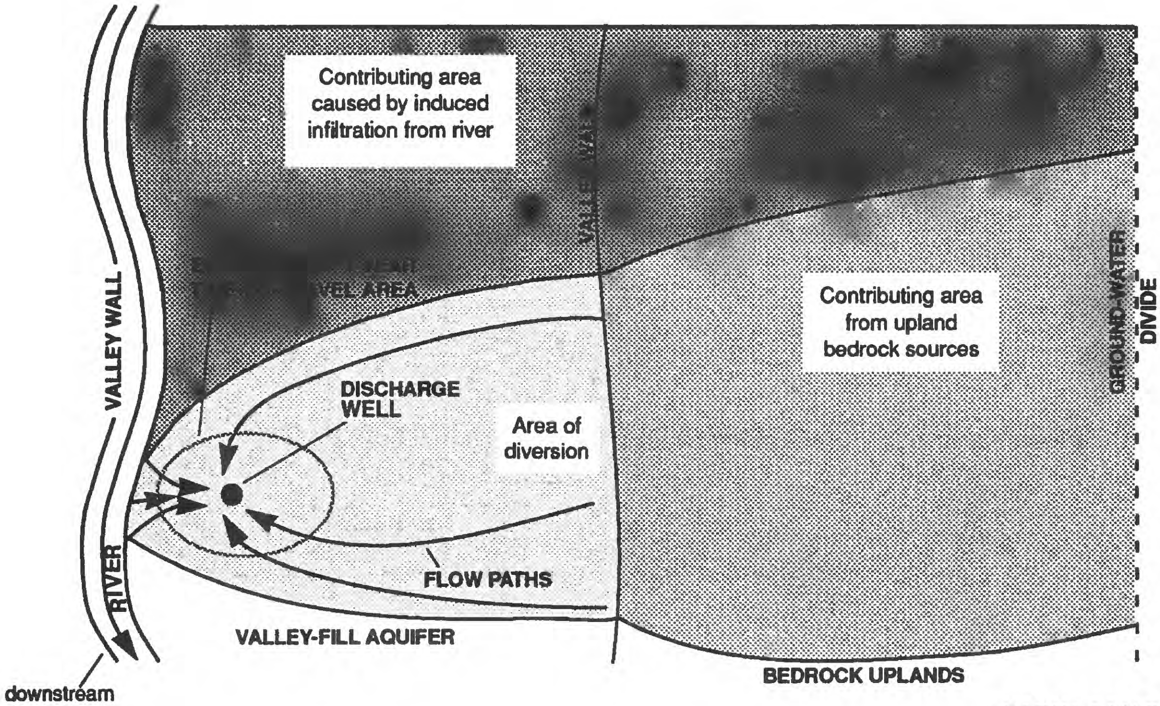
The difference between the area of diversion and contributing area of water to a well in a typical valley-fill aquifer is shown in figure 7. The area of diversion includes only the surface projection of that part of the valley-fill aquifer through which water is diverted to the well. This area would also be equal to the contributing area if precipitation on the aquifer surface were the only source of water contributed to the well; however, because upland runoff and interflow from bedrock commonly provide large amounts of recharge to valley-fill aquifers, the well's contributing area includes parts of the adjacent upland surface shown in figure 7. Similarly, if river water is induced to the well (as is indicated in figure 7), the river's entire watershed upstream from the well is included in the contributing area. Therefore, an understanding of all sources of water entering the well's area of diversion is needed for a proper delineation of its contributing area.

At times, delineation of only a fraction of the area of diversion is desirable; for example, the fraction from which water will reach a well within a specified time. Thus, if a particular contaminant can be shown to be fully attenuated within the aquifer after a period of 1 year, only that part of the aquifer within a 1-year traveltime of water to the well might need to be protected. A hypothetical 1-year time-of-travel area is shown schematically in figure 7.

EFFECTS OF PUMPING ON VALLEY-FILL AQUIFERS

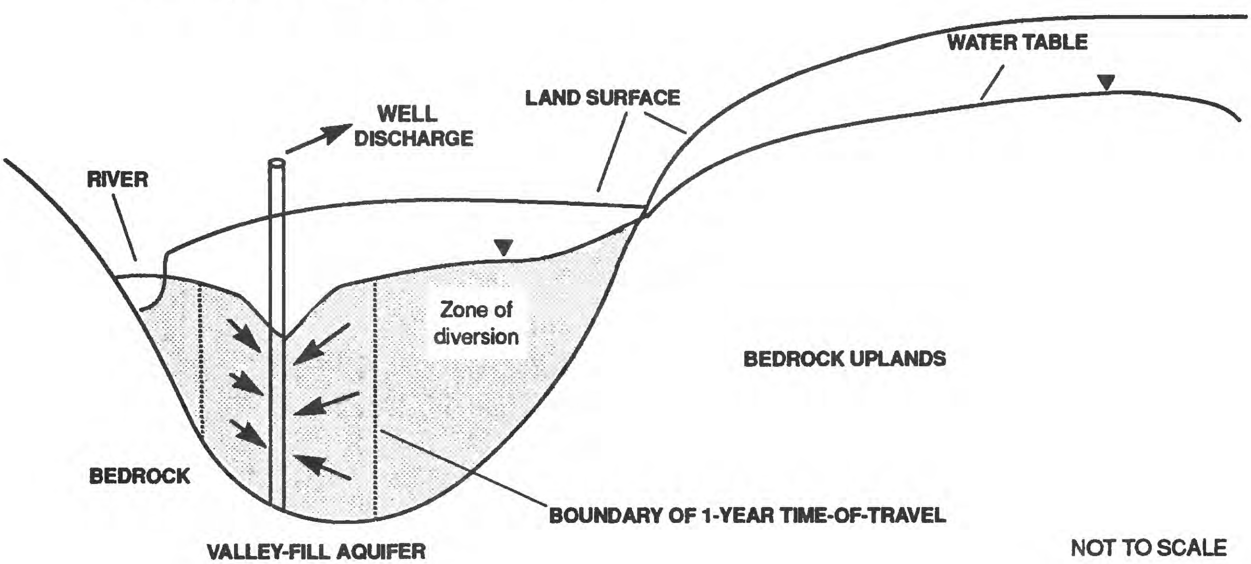
The effects of pumping on the natural flow regime are commonly misunderstood because the areal extent of the hydraulic response of the aquifer (area of influence) is not distinguished from the areal extent of diversion of water toward the well (contributing area). Also, the time required from the start of pumping to a measurable hydraulic response commonly is not distinguished from the traveltime of water to the well. A clear understanding of the differences between the area of influence and the contributing area is a prerequisite for identification of the sources of water diverted to a well.

Before ground-water development, valley-fill aquifers are in a state of dynamic equilibrium (also termed "steady state") such that natural recharge is balanced by an equal quantity of natural discharge. The term "dynamic" means that although at any specific moment the rates of natural recharge and discharge are probably not equal, given enough time, the amounts will balance. Valley-fill aquifers in Pennsylvania are small enough that recharge and discharge amounts balance within the year; water levels return to about the same position each autumn. Pumping from the aquifer disrupts the equilibrium by imposing a new discharge on the system.



PLAN VIEW

NOTE: The sum of the three shaded areas equals the total contributing area of water to the well.



CROSS SECTION

NOTE: Diagram does not refer to idealized aquifer shown in figure 5.

Figure 7. Diagram showing the area of diversion, contributing area, and time-of-travel area.

When a well is pumped, water levels near the well are lowered and the withdrawals initially are balanced by a reduction in ground-water storage. Natural rates of recharge or discharge in the aquifer are not altered until the area of influence resulting from pumping reaches locations of natural discharge or rejected recharge. As Theis (1938) stated: "Under this concept the cone of depression may be considered a pirating agent created by the well to procure water for it, first robbing the aquifer of stored water and finally robbing surface water or areas of transpiration in the localities of recharge or natural discharge." A new equilibrium can be achieved only when ground-water withdrawals are balanced by an increase in natural recharge, decrease in natural discharge, or a combination of both.

Time is required to achieve a new equilibrium after pumping has begun. The amount of time needed is a function of aquifer properties and the proximity of pumping to areas where natural recharge and discharge can be altered. In valley-fill aquifers, areas where natural recharge can be increased are generally small; thus, nearly all withdrawals are balanced by a reduction of natural discharge (streamflow depletion and reduction of evapotranspiration from wetlands).

Hydraulic Response

An idealized valley-fill aquifer was simulated by use of a three-dimensional finite-difference ground-water flow model (McDonald and Harbaugh, 1988) to illustrate aquifer response to pumping stress. The general hydrology of the idealized aquifer was described previously and is illustrated in figure 5. The prepumping, steady-state water budget for the idealized aquifer is shown below. This budget is typical of valley-fill aquifers in Pennsylvania. Physical properties and boundary conditions of the idealized aquifer simulated by the model are shown in figure 8.

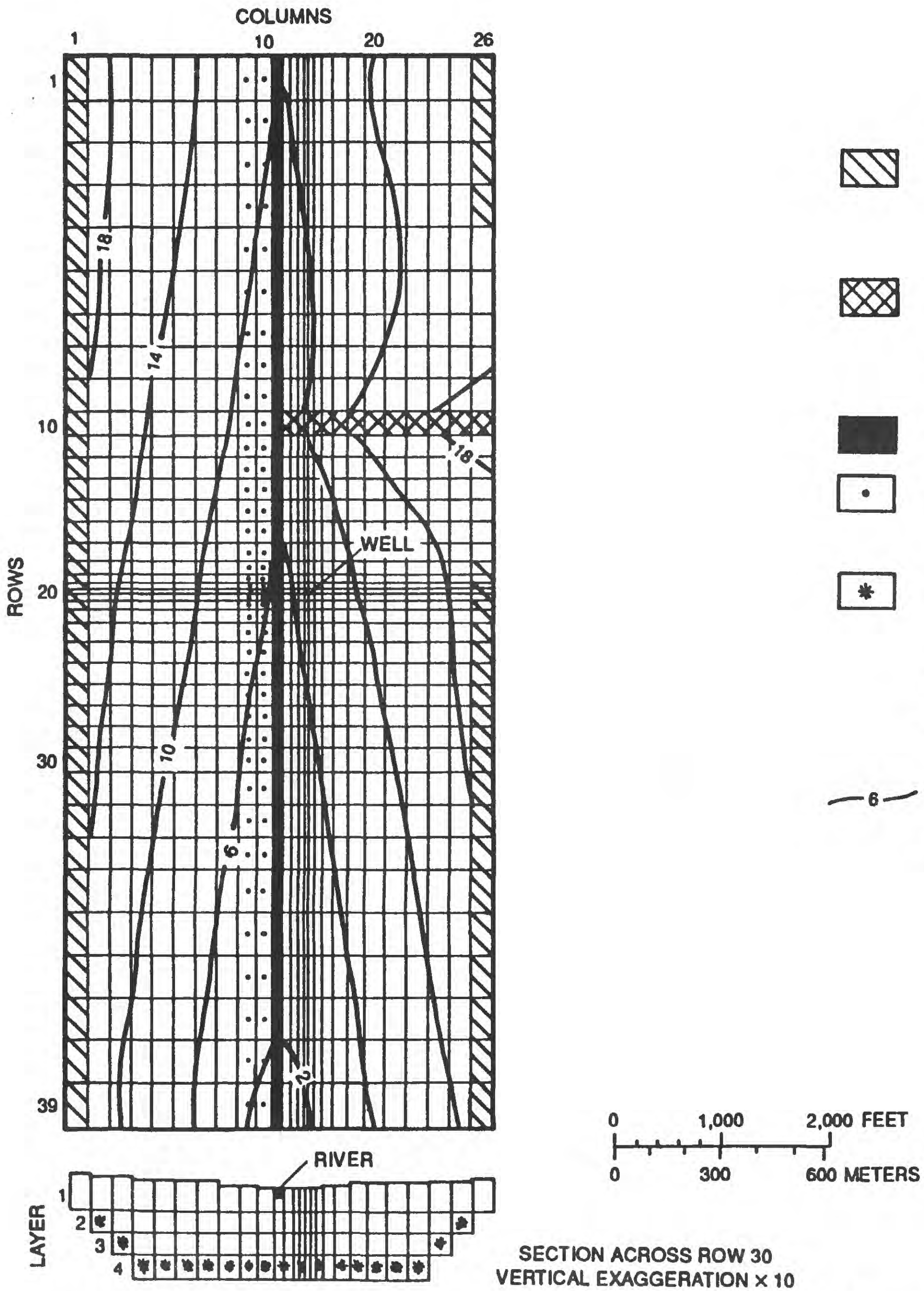
Prepumping steady-state water budget for idealized valley-fill aquifer

Source of recharge (number identifies source of recharge in figure 5)	Rate of recharge, in cubic feet per second	Percentage of total recharge
1) Direct infiltration of precipitation on 1.43 square miles of valley-fill aquifer surface	2.10	30
2) Seepage from an upland tributary stream that drains 1 square mile	.93	14
3) Infiltration to valley-fill aquifer of overland runoff and interflow from 3.3 square miles of bedrock upland area	3.06	44
4) Regional flow of ground water from bedrock to valley-fill aquifer	.81	12
Total	6.90	100

Discharge (number identifies source of discharge in figure 5)	Rate of discharge, in cubic feet per second	Percentage of total discharge
5) Base flow to the river	6.60	96
6) Evapotranspiration from 0.15-square-mile wetlands area	.30	4
Total	6.90	100

In the simulation, the aquifer is stressed by withdrawing water from the bottom 60 feet of saturated sediments (layers 2-4) at a well 220 feet east of the river (row 20, column 15). The well is pumped at 500 gallons per minute (1.114 cubic feet per second) for a period of 3 years. After about 1 year of pumping, the system reaches a new equilibrium (steady state) at which the withdrawals are balanced by a 497-gallon-per-minute reduction in natural discharge and a 3-gallon-per-minute increase in natural recharge.

In the simulation of the ideal aquifer, water levels decline in response to pumping, and the area of influence expands through time (fig. 9). Drawdown is largest near the well and east of the river. The partly penetrating river decreases, but does not prevent, drawdown on the side of the river opposite the well. Theoretically, the only limits to the expansion of the well's area of influence are the valley walls bounding the aquifer; simulated drawdown is at least 0.001 foot everywhere in the valley after 1 year of pumping. Because these small drawdowns are impossible to separate from natural water-level fluctuations, the true extent of the well's influence is not commonly realized.



EXPLANATION

BOUNDARY CONDITIONS

Unchanneled runoff and interflow—Specified flow of 3.06 cubic feet per second distributed along valley margin in layer 1 to simulate unchanneled runoff and interflow from 3.33 square miles of upland area. (1 square mile of upland is assumed to yield 0.93 cubic foot per second of recharge to valley fill).

Tributary stream—Specific flow of 0.93 cubic foot per second in layer 1, row 10, columns 12-26 simulates infiltration from a tributary that drains a 1-square mile upland area.

Well—A well being pumped at 1.114 cubic feet per second is simulated with a specified flow from the model at row 20, column 15 in layers 2-4.

River boundary—River partly penetrates column 11 in layer 1. The river is simulated as having a permeable bed; its properties are described below.

Evapotranspiration—Evapotranspiration is simulated from the river and wetland area 220 feet wide along its west bank. Land elevation of the wetland area is 2 feet above the adjacent river stage. Maximum rate is 28 inches per year and extinction depth is 10 feet.

Recharge from ground-water flow from bedrock is simulated by a general-head boundary adjacent to valley fill. Computes recharge based on horizontal hydraulic conductivity (0.2 foot per day) and hydraulic head (50 feet) of bedrock aquifer and average flow-path length (4,000 feet).

Recharge of 20 inches per year is applied uniformly as a specified flow to layer 1.

No flow—No flow is permitted across the upvalley (north) and downvalley (south) ends of the model. (All discharge is to the river, wetlands, and well).

—6— WATER-TABLE CONTOUR—Shows altitude of simulated prepumping water-table surface, in feet above an arbitrary datum. Contour interval 4 feet.

PHYSICAL PROPERTIES

Aquifer width is 4,000 feet

Aquifer length is 10,000 feet

Aquifer depth is 91 feet maximum (layers 2-4 are each 20 feet thick). Thickness of layer 1 is variable. Bottom of layer 1 is a constant value for each row-- 10 feet less than the river altitude in that row.

Horizontal hydraulic conductivity is 50 feet per day

Anisotropy is 1 (vertical and horizontal)

Specific yield is 0.20

Storage coefficient is 0.00002 (layers 2-4)

River slope is 0.001 (River stage ranges from an altitude of 9.8 feet at north end of model to 0.2 foot at south end).

River bottom is 1 foot thick. Vertical hydraulic conductivity of river bottom is 50 feet per day.

River width is 100 feet (entire width of column 11)

Figure 8. Finite-difference grid, boundary conditions, and physical properties for a model of an idealized valley-fill aquifer.

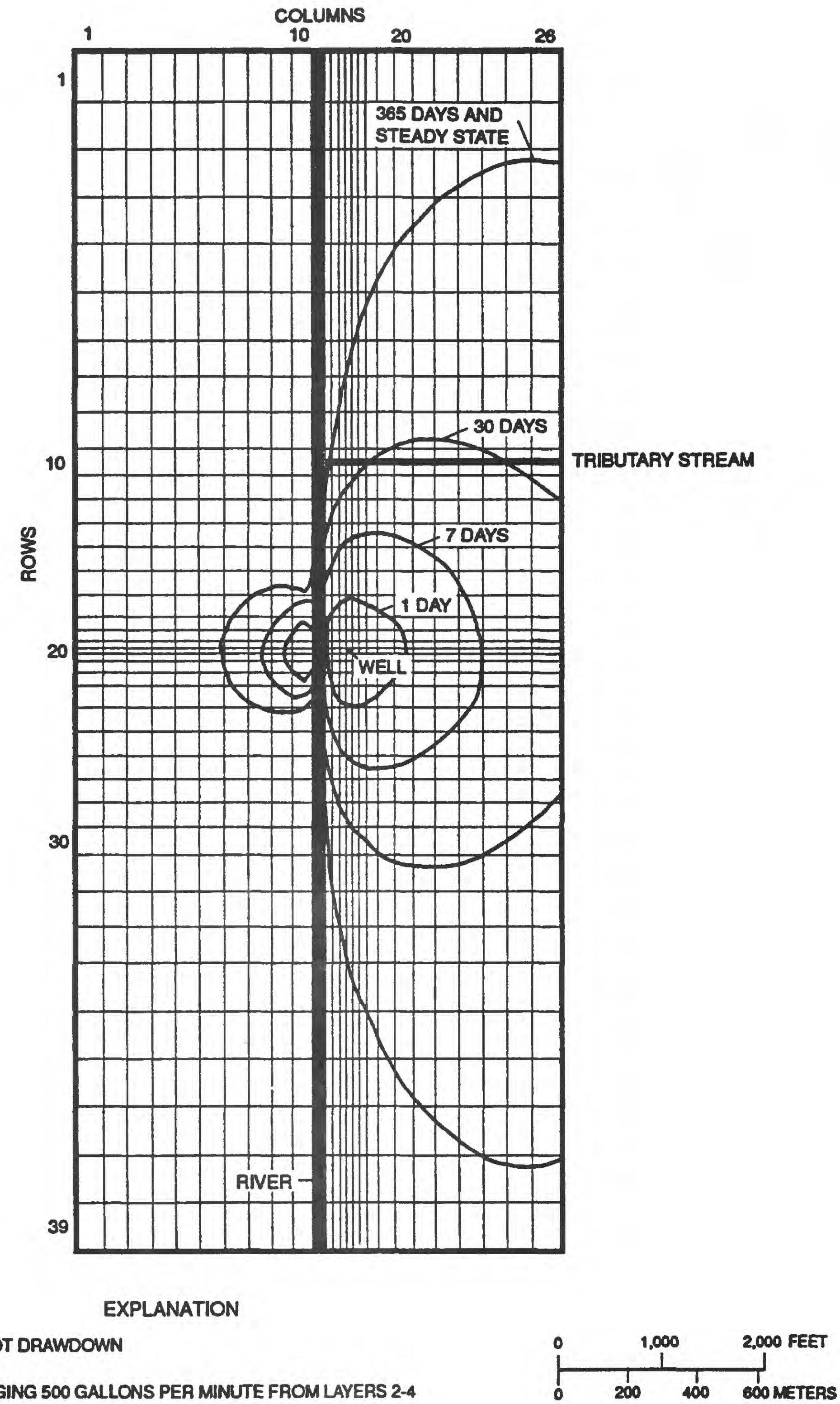


Figure 9. Expansion of the area of influence illustrated by the changing position of the 0.1-foot drawdown contour in an idealized valley-fill aquifer.

Effects of pumping on recharge, discharge, and aquifer storage are shown in figure 10. During the first few hours of pumping, the system is balanced by a nearly equivalent reduction in ground-water storage. As the well's area of influence expands and the hydraulic gradient at the river decreases, ground-water seepage to the river decreases. After 1 day of pumping, this decrease in ground-water flow to the river amounts to about 20 percent of the pumping rate. As pumping continues, the decrease in ground-water flow to the river reaches a maximum of 82 percent of the pumping rate. When drawdown from the well is sufficient to reverse the hydraulic gradient at the stream, water is induced directly from the river. Induced infiltration from the river begins about 2 days after pumping starts and increases to a maximum of 17 percent of the pumping rate.

Although insignificant compared to the reduction in natural streamflow, ground-water withdrawals also affect evapotranspiration and regional flow of ground water into the valley fill. After about 1 year, the lowered water table causes evapotranspiration rates to decrease about 0.25 gallons per minute. The rate of ground-water flow from bedrock aquifers increases by about 3 gallons per minute. The increase in ground-water inflow can be thought of as a small increase in the natural recharge to the valley-fill aquifer.

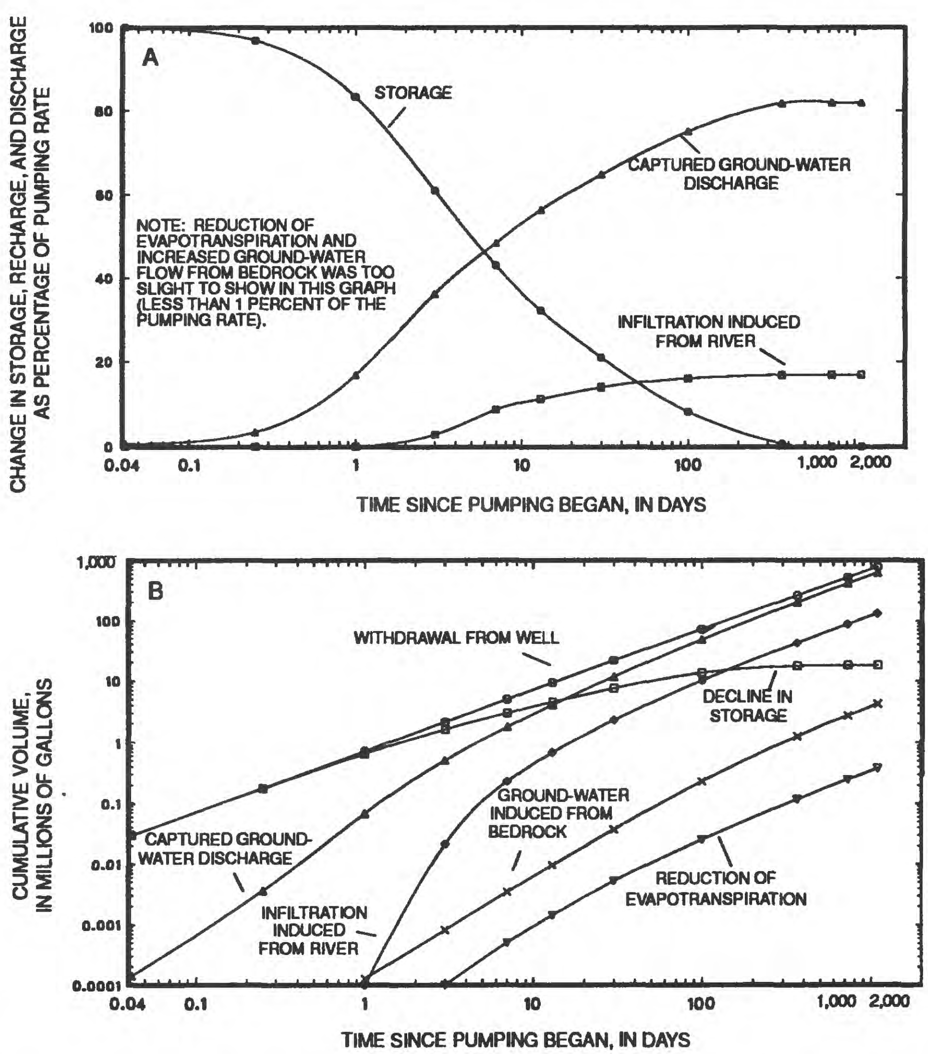


Figure 10. The (A) instantaneous and (B) cumulative effects on recharge, discharge, and storage from pumping 500 gallons per minute for 3 years from an idealized valley-fill aquifer (Aquifer properties specified on figure 8.)

Cumulative effects on the hydraulic system because of pumping are shown in figure 10. A logarithmic scale is used to illustrate the significance of changes in evapotranspiration rate and regional ground-water flow from bedrock. A total of 17.8 million gallons of water must be removed from storage before a new equilibrium can be established in the aquifer. After pumping is stopped, water levels will recover and the 17.8 million gallons removed from storage will be balanced by an equivalent decrease in natural discharge from the system.

Sources of Water to Wells

The idealized valley-fill aquifer described previously (fig. 7) is used to illustrate the complexity of identifying areas of diversion, contributing areas, and time-of-travel areas of wells. Eight sources supply water to a well pumped at 500 gallons per minute after equilibrium conditions are established (fig. 11). Only 24 percent of the well's discharge is from precipitation on the valley floor near the well. The remaining 76 percent is from infiltration induced directly from the river and upland sources that extend beyond the boundaries of the valley-fill aquifer.

The area of diversion and contributing area to the well are shown in figure 12. Although the aquifer geometry and hydraulic properties are simple for this ideal example, the shapes are complex. The area of diversion on the surface of the valley floor encompasses 0.20 square mile extending on both sides of the river. The contributing area to the well from uplands adjacent to the valley is about seven times larger than the area of diversion. Runoff and interflow from adjacent bedrock uplands contribute 50 percent of the well discharge from a 1.38-square-mile watershed area. Regional ground-water flow from bedrock uplands contributes an additional 9 percent. Water induced from the river (17 percent) is contributed from its watershed upvalley from the well. This upvalley watershed could represent a large area for a well adjacent to a major river.

The area of influence (fig. 9), area of diversion, and contributing area (fig. 12) to this well are quite different. Significant drawdown was simulated in areas where water is not diverted to the well; yet, a 1.38-square-mile area outside of the valley-fill aquifer contributes water to the well. Also, because water is induced from the river, the entire upstream watershed is included in the contributing area.

Steady-state time-of-travel areas of 30 days, 100 days, 1 year, and 2 years in the idealized valley-fill aquifer are shown in figure 13. These areas, based on the distance that water has traveled in a given time after steady-state pumping conditions have been established, are all smaller than the area of diversion; however, regardless of the traveltime being considered, the sources of water for the well are the same because the aquifer is in equilibrium. For example, the sources of water indicated in figure 11 remain the same for all traveltimes shown in figure 13.

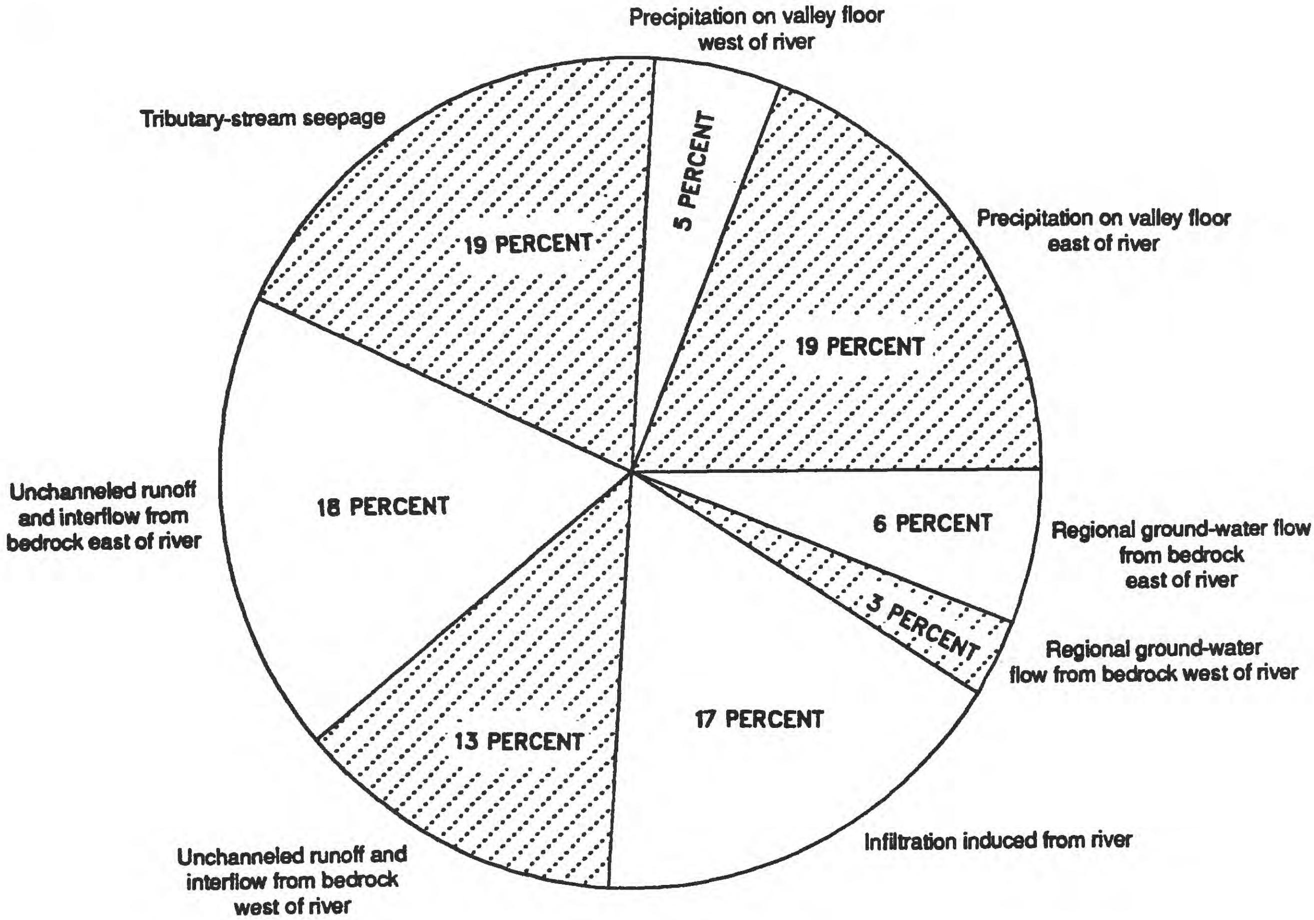


Figure 11. Sources of water to a well, pumped at 500 gallons per minute, in an idealized valley-fill aquifer at steady state. (Aquifer properties specified on figure 8.)

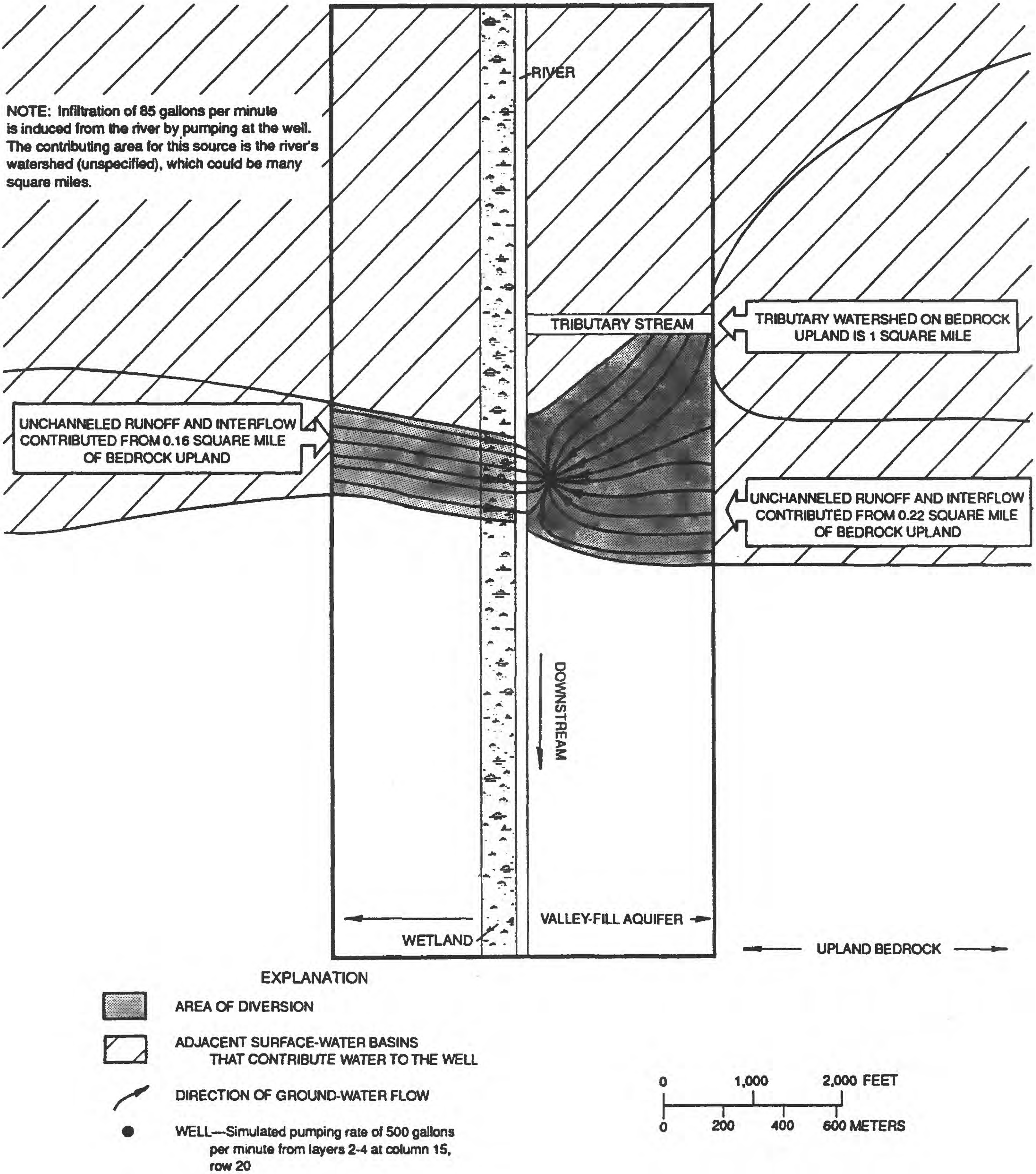


Figure 12. Area of diversion and contributing area of a well discharging from an idealized valley-fill aquifer.

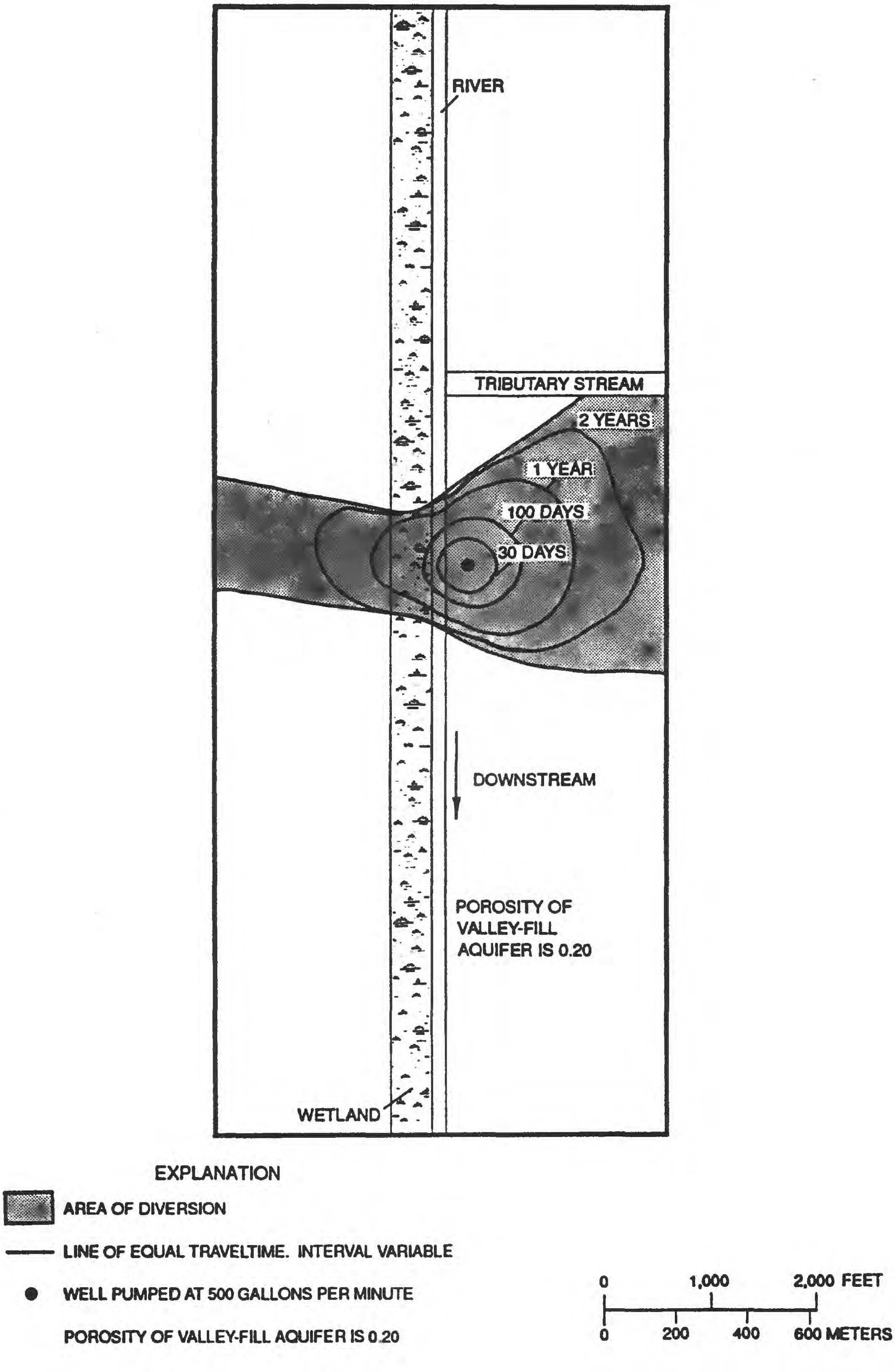


Figure 13. Time-of-travel areas for specific traveltimes in an idealized valley-fill aquifer.

FACTORS AFFECTING THE CONTRIBUTING AREA

The best possible estimate of the contributing area (and area of diversion or time-of-travel area) to a well is based on an awareness of all factors that can affect the sources of water to that well. The size and the shape of the areas are affected by nearly all properties of the flow system. "Flow system" refers not only to ground water in the valley-fill aquifer near the well but also to the regional flow systems in bedrock aquifers underlying the valley-fill aquifer. Factors that affect the size and shape of the contributing area can be grouped into categories of (1) initial and boundary conditions, (2) hydraulic properties, and (3) well characteristics. The model of the idealized valley-fill aquifer shown in figure 8 was used to simulate an example of how boundary conditions, hydraulic properties, and well characteristics can affect the contributing area of water to a well.

Initial and Boundary Conditions

Initial and boundary conditions define physical extent and flux of water into and out of the aquifer. Incomplete knowledge of or erroneous assumptions concerning these conditions can lead to errors in delineation of contributing areas. Initial and boundary conditions include (1) aquifer geometry (thickness and width of valley fill), (2) prepumping hydraulic-head distribution in three dimensions, (3) recharge (flow across the water-table surface, seepage from tributary streams, infiltration from unchanneled upland runoff and interflow, and regional ground-water flow from adjacent bedrock), and (4) discharge (location and hydraulic characteristics of streams that drain the aquifer and evapotranspiration from wetlands).

The sensitivity of the area of diversion to a single change in boundary conditions was illustrated using the model of the idealized valley-fill aquifer. The river was changed from one that partially penetrates in the top 20 feet of the aquifer (fig. 14A) to one that fully penetrates the entire aquifer (fig. 14B). The sources of water for the well are listed in table 1.

A river that fully penetrates the aquifer restricts the area of diversion to the side of the river where the well is completed. The lack of contribution from the valley across the river is made up mostly by an increase in infiltration induced from the river.

Hydraulic Properties

The hydraulic properties of valley-fill sediments in Pennsylvania are quite variable as indicated by the more than six-order-of-magnitude range in hydraulic conductivity (fig. 2B). A huge investment in aquifer testing and observation wells would be required for a full evaluation of hydraulic properties in three dimensions. Even then, much of the heterogeneity within the aquifer would not be defined. Incomplete evaluation of the following hydraulic properties could lead to errors in determination of contributing areas: (1) hydraulic conductivity of aquifer—horizontal and vertical, (2) specific yield and storage coefficient, (3) vertical hydraulic conductivity of riverbed sediments, and (4) saturated thickness.

The sensitivity of the area of diversion to changes in a single hydraulic property was evaluated by addition of a thin clay layer about 20 feet below the water-table altitude in the simulation of the idealized valley-fill aquifer. The clay is only 1 foot thick, and its hydraulic conductivity is 0.002 foot per day. The addition of a clay layer increases the area of diversion but allows precipitation from only a small part of the valley surface to be diverted to the well (fig. 14C). Almost all of the water contributed to the well is from upland-bedrock sources such as overland runoff, interflow, and the tributary stream (table 1).

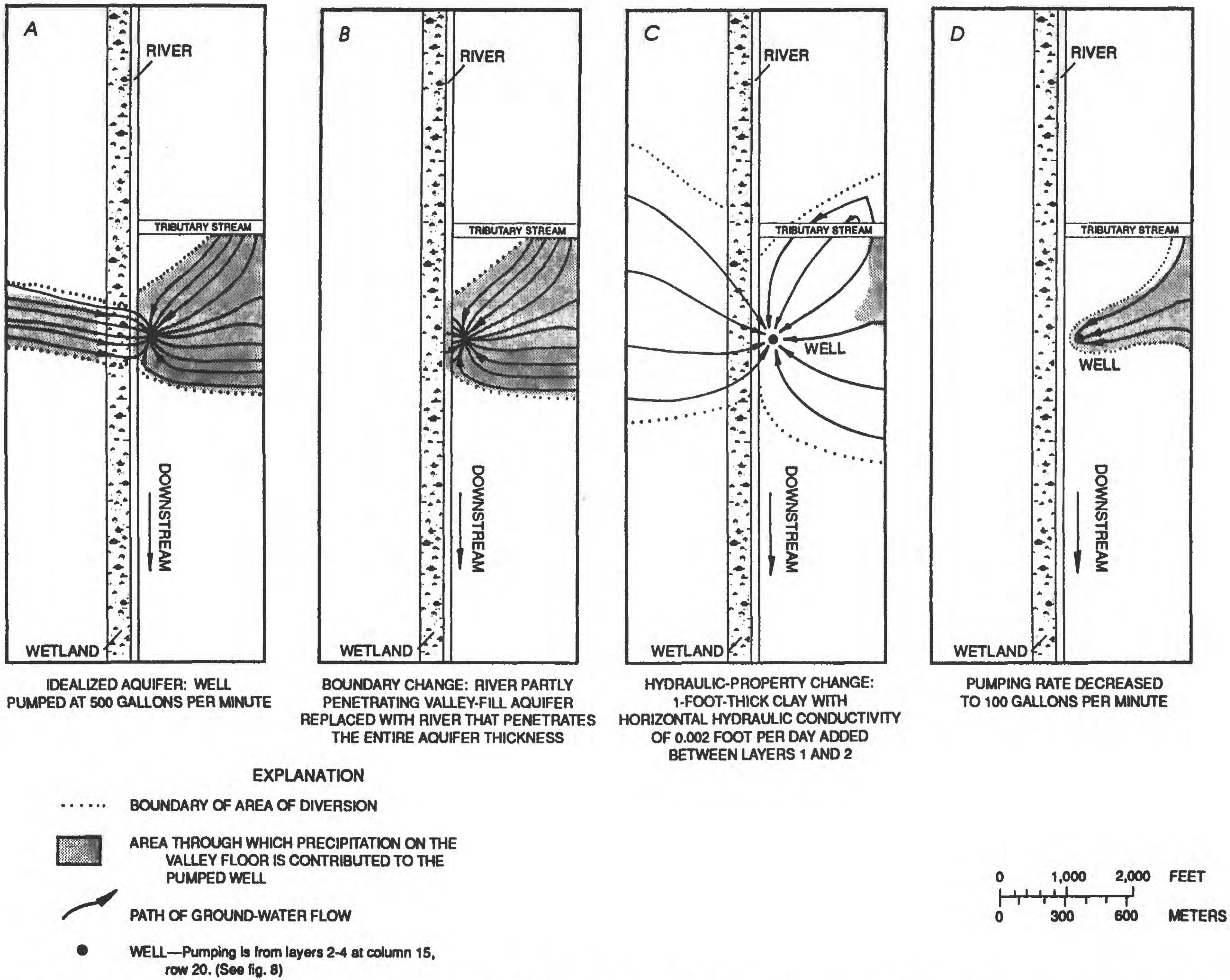


Figure 14. Changes in the area of diversion in an idealized valley-fill aquifer as a result of different boundary conditions, hydraulic properties, and pumping rate.

Table 1. Relations of sources of water to a pumped well and differing boundary conditions, hydraulic properties, and pumping rates

Source of water	Relative contribution to well, as percentage of pumping rate, for idealized aquifer			
	Pumping rate (500 gallons per minute)	River fully penetrating the aquifer	Clay between model layers 1 and 2	Pumping rate (100 gallons per minute)
Direct infiltration of precipitation on valley floor--west of river	5	0	0	0
Direct infiltration of precipitation on valley floor--east of river	19	19	3	38
Seepage from upland tributary stream	19	19	27	20
Infiltration of overland runoff and interflow--western side of valley	13	0	25	0
Infiltration of overland runoff and interflow--eastern side of valley	18	17	25	28
Infiltration induced from the river	17	39	0	0
Regional ground-water flow from bedrock aquifers --west of river	3	0	10	0
Regional ground-water flow from bedrock aquifers--east of river	6	6	10	14
Total	100	100	100	100

Well Characteristics

Some characteristics of the pumped well that affect the contributing area are (1) pumping rate, (2) depth and length of screened interval, and (3) location of pumped well relative to impervious or recharge boundaries.

The effects of well characteristics on the area of diversion and sources of water to the well were evaluated in part by changing the pumping rate in the simulation of the idealized aquifer from 500 to 100 gallons per minute. Decreasing the pumping rate to 100 gallons per minute restricts the area of diversion to the eastern side of the valley (fig. 14D). Drawdown caused by the pumping is not large enough to reverse the hydraulic gradient at the river; thus, no water is induced from the river to the well or from beneath the river (table 1).

EVALUATION OF METHODS TO DELINEATE THE CONTRIBUTING AREA

Most methods used to delineate the contributing area of wells do not explicitly delineate upland bedrock areas that contribute water. Rather, the methods are usually used to delineate an area of diversion or time-of-travel area on the surface of the valley-fill aquifer. The contributing area can then be estimated by sketching the upland bedrock areas that contribute water and adding them to the area of diversion that was delineated on the valley-fill aquifer. All methods for delineation of the area of diversion, contributing area, and time-of-travel area of a well require that the aquifer system be simplified. Even complex numerical models represent only approximations of the hydrologic processes operating in a real aquifer. Therefore, accuracy of the methods will largely be determined by the hydrologic experience and intuition of the investigator.

Methods presented herein are based on the assumption of steady state. At steady state, the area of influence of a well is as large as possible for a given set of conditions; therefore, the contributing area also is at its maximum extent. The contributing area of a well can be delineated before steady state is attained, but the significance of the area is questionable because (1) the velocity distribution around the well is not steady with time (thus, continued pumping will enlarge the contributing area), and (2) some of the water withdrawn by the well is from ground-water storage, whose source is unknown.

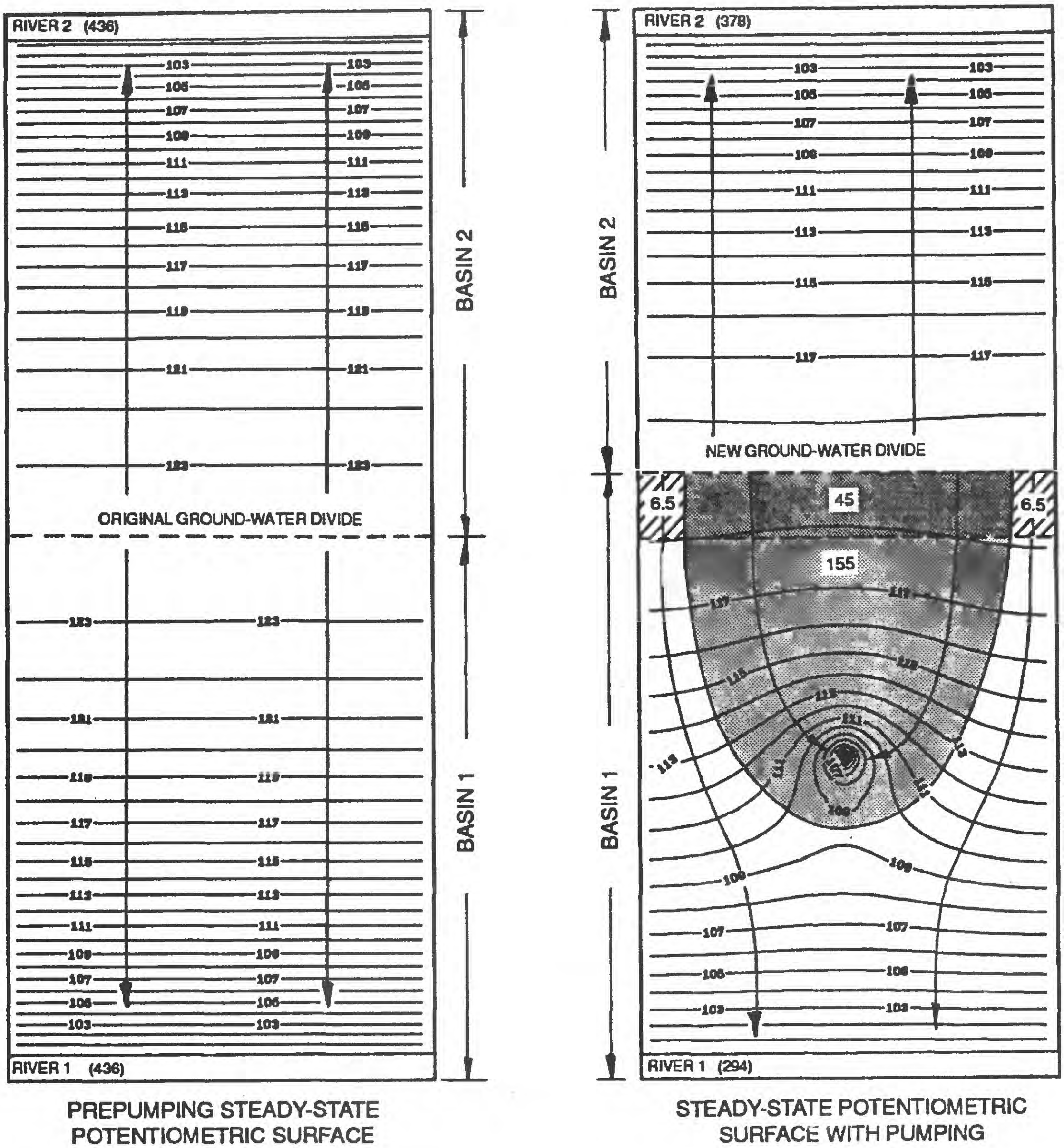
Although not explicitly mentioned in this report as a method of delineating contributing areas, hydrogeologic mapping is an integral part of every investigation. The mapping can involve geophysical investigations such as seismic, electric, and gravity surveys. Mapping helps to identify boundaries and heterogeneities in hydraulic properties, which can greatly influence contributing areas. Mapped boundaries can be used as limits for the area contributing water to the well (U.S. Environmental Protection Agency, 1987, p. 4-14); however, water can pass beneath or across mapped boundaries as shown in figure 15. For example, a water-table divide is frequently used as the upgradient boundary for a well's contributing area, although pumping (even at great distances from the divide) can shift the divide so that ground water is diverted from an adjacent basin. This situation is illustrated in figure 15, where a well pumping 200 gallons per minute from basin 1 at steady state derives 45 gallons per minute of its discharge from precipitation that was contributed to the river in basin 2 before pumping began.

After delineation of the contributing area to a well, a water balance should be computed to determine if the recharge rate to the aquifer throughout the area of diversion equals the pumping rate from the well. In computing this balance, all sources of recharge need to be carefully considered. Identification of water sources are especially important in valley-fill aquifers, where much of the recharge is likely to be from upland sources and induced infiltration from nearby streams, sources that are not always easily quantified.

Description, Application, and Limitations of Methods

Many methods are available for estimation of the area of diversion, contributing area, and time-of-travel area of a well. The methods are grouped here in five major categories in order of increasing complexity: (1) fixed radius, (2) uniform flow, (3) analytical, (4) semianalytical, and (5) numerical flow (U.S. Environmental Protection Agency, 1987; Javandel and others, 1984). A different amount of information concerning aquifer properties, boundary conditions, and well characteristics is required for each method. Therefore, each approach has inherent limitations that affect its use.

Each method provides either an estimate of the area of diversion, source of water to wells, or time-of-travel area. The contributing area of the well is estimated by analysis of sources of water to the valley-fill aquifer within the well's area of diversion. These sources include induced infiltration from streams and water from bedrock uplands.



EXPLANATION

DIVERSION AREA TO WELL THAT PREVIOUSLY WAS A SOURCE OF WATER TO RIVER 1

DIVERSION AREA TO WELL THAT PREVIOUSLY WAS A SOURCE OF WATER TO RIVER 2

DIVERSION AREA TO RIVER 1 THAT PREVIOUSLY WAS A SOURCE OF WATER TO RIVER 2

45 GROUND-WATER RECHARGE (DISCHARGE IF IN PARENTHESES) FOR SELECTED AREAS, IN GALLONS PER MINUTE

155 LINE OF EQUAL HYDRAULIC HEAD, IN FEET ABOVE AN ARBITRARY DATUM

PATH OF GROUND-WATER FLOW

WELL PUMPED AT 200 GALLONS PER MINUTE

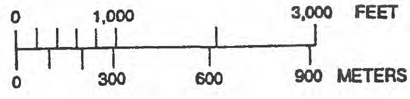


Figure 15. Shift in position of ground-water divide caused by pumping.

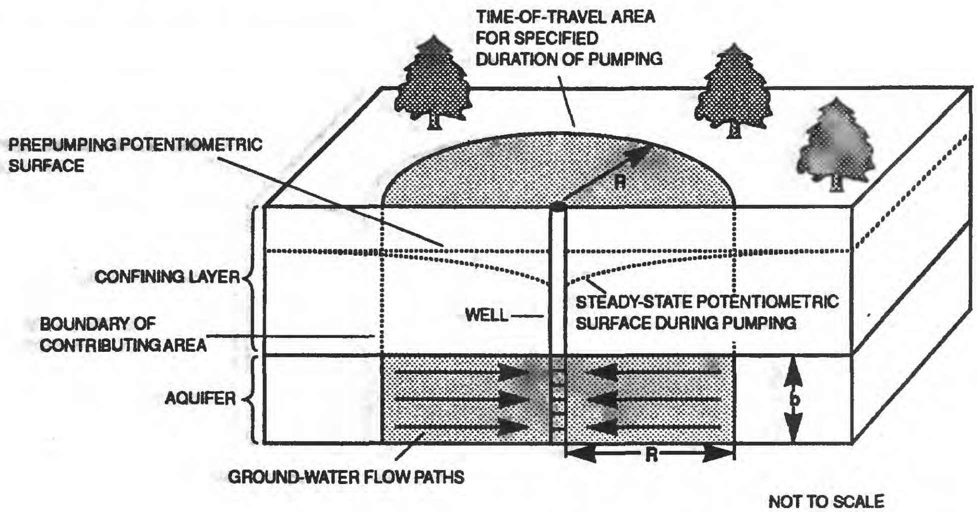
Fixed-Radius Method

The fixed-radius method can be used to estimate a time-of-travel area of the well. The method described here is based on estimation of the cylindrical aquifer volume through which ground water moves towards a well for a given magnitude and duration of pumping (U.S. Environmental Protection Agency, 1987, p. 4-6). The time-of-travel area is the circular projection of this volume at land surface (fig. 16).

In the fixed-radius method, the aquifer is assumed to be confined,[1] of constant thickness, homogeneous, and isotropic; a flat prepumping potentiometric surface and fully penetrating pumped well also are assumed. Boundaries must be sufficiently distant so that the flow field in the vicinity of the well is not significantly affected by their presence. Pumping is assumed to have resulted in two-dimensional steady-state flow in the aquifer.

Application

The fixed-radius method requires estimates of aquifer thickness, aquifer porosity, pumping rate, and pumping duration (traveltime of interest). The equations used are simple and can be solved easily with a hand-held calculator. Special hydrologic knowledge is not needed to compute a contributing area with this method but may be needed to evaluate the validity of the computation.



EXPLANATION

EQUATION USED TO COMPUTE RADIUS OF CONTRIBUTING AREA:

$$R = [(Q \times t)/(\pi \times b \times \theta)]^{1/2},$$

where R is radius of circular time-of-travel area, in feet;
Q is pumping rate, in cubic feet per day;
t is duration of pumping (traveltime of interest), in days;
b is aquifer thickness, in feet; and
θ is porosity of aquifer.

Figure 16. Block diagram illustrating fixed-radius method for estimating a time-of-travel area in a confined aquifer.

[1] The method can be used for an unconfined aquifer if drawdown from pumping is less than about 10 percent of the aquifer's saturated thickness (Reilly and others, 1987).

The time-of-travel area is computed by equating the volume of water withdrawn in a given time to an equal cylindrical volume of pore space in the aquifer around the well. The volume of water withdrawn in a given time is

$$V_p = Qt, \tag{1}$$

where V_p is volume of water pumped, in cubic feet;
Q is pumping rate, in cubic feet per day; and
t is duration of pumping (traveltime of interest), in days.

The cylindrical aquifer volume around the well is

$$V_a = \pi R^2 b\theta, \tag{2}$$

where V_a is pore space aquifer volume, in cubic feet;
R is radius of the time-of-travel area, in feet;
b is aquifer thickness, in feet; and
θ is aquifer porosity.

Equating the volume of water pumped with pore space in the aquifer volume around the well and solving for the radius of the time-of-travel area gives

$$R= [(Qt)/\pi b\theta)]^{1/2} \tag{3}$$

If the aquifer receives a constant flux of recharge or leakage, the following equations apply:

$$V_p = V_a - V_w , \tag{4}$$

and

$$R = [(Qt)/((\pi b\theta) + wt)]^{1/2} \tag{5}$$

where V_w is volume of recharge or leakage, in cubic feet; and
w is recharge or leakage rate per unit surface area, in feet per day.

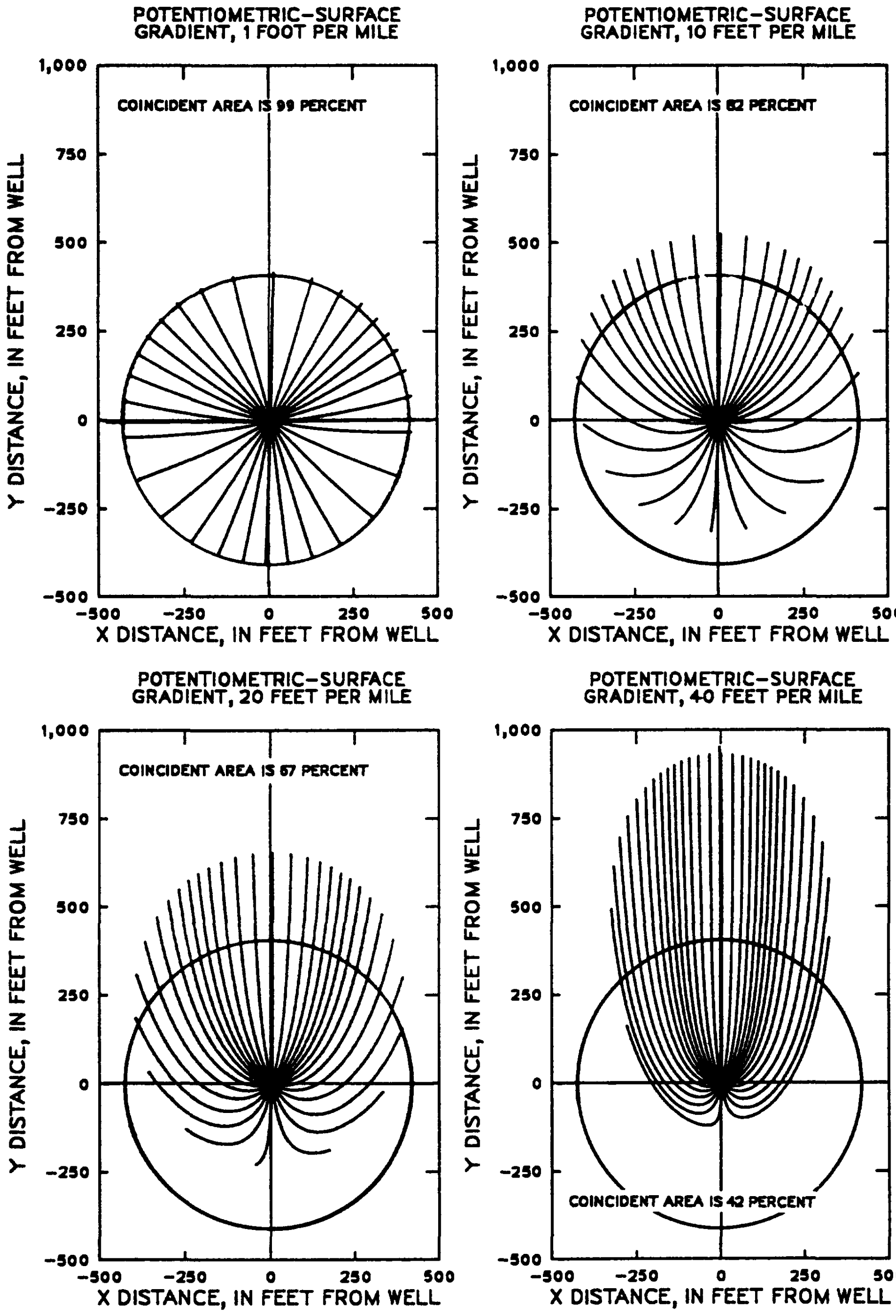
Example.--The radius of the circular 1-year time-of-travel area is computed for the following conditions: pumping rate, 500 gallons per minute (96,244 cubic feet per day); aquifer thickness, 60 feet; aquifer porosity, 0.20; steady, uniform leakage of 0.0055 foot per day through an overlying confining unit. By use of equation 5,

$$R = \left[\frac{(96,244 \text{ cubic feet per day})\ (365 \text{ days})}{[(3.1416)\ (60 \text{ feet})\ (0.20)] + [(0.0055 \text{ foot per day})\ (365 \text{ days})]}\right]^{1/2} = 941 \text{ feet}$$

Limitations

The fixed-radius method is based on a restrictive set of hydrologic assumptions that are not likely to be met in valley-fill aquifers in Pennsylvania. The most restrictive assumptions are a flat prepumping potentiometric surface and distant boundaries. In fact, the assumption of steady-state radial flow to a well in a nonleaky, confined aquifer with a flat prepumping potentiometric surface is not possible unless the well is at the center of a circular island. Nevertheless, estimates of time-of-travel areas obtained by use of the fixed-radius method may be reasonable under certain conditions.

Effects of slope of potentiometric surface.--The effect of a uniformly sloping, prepumping, potentiometric surface on the time-of-travel area to a well is illustrated for several selected gradients in figure 17. Shapes of time-of-travel areas were computed by use of a semianalytical method in which a uniformly sloping water-table surface is assumed (Javandel and others, 1984, p. 40). Time-of-travel areas computed by use of the fixed-radius method are shown in figure 17. For gentle slopes, the area is nearly circular; for steeper potentiometric surfaces, the area is skewed upgradient. The results are closest where the potentiometric surface is nearly flat because the fixed-radius method does not account for a sloping surface.



EXPLANATION

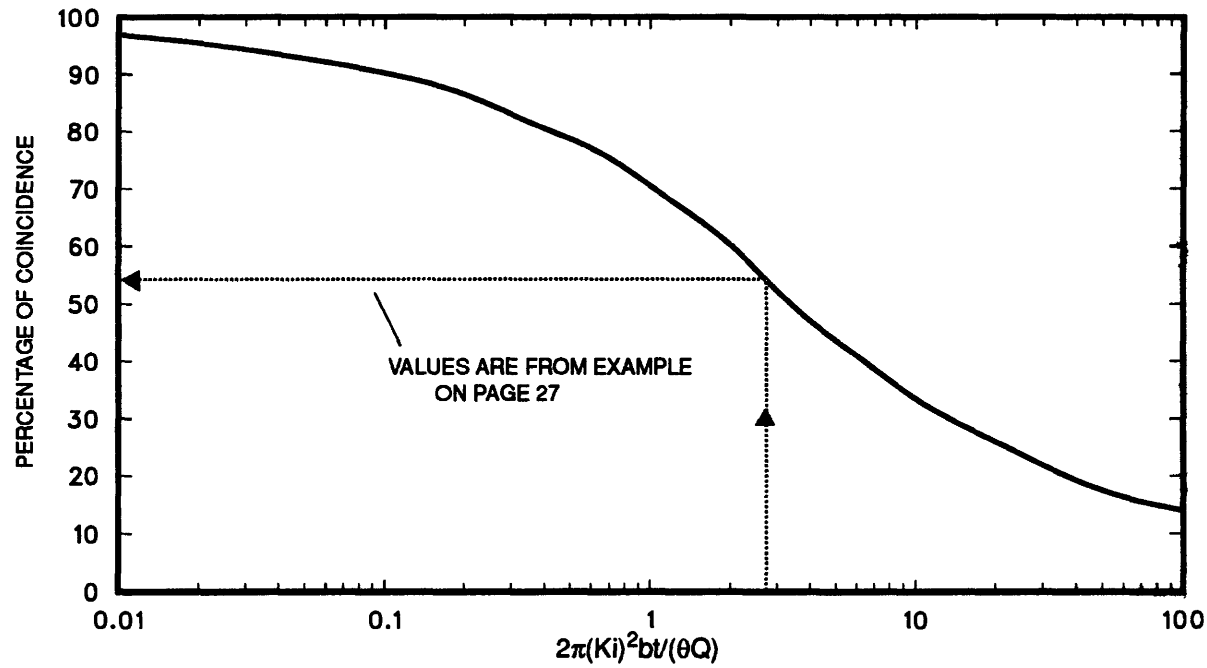
LARGE CIRCLE REPRESENTS TIME-OF-TRAVEL AREA FOR A 1-YEAR TRAVELTIME COMPUTED BY USE OF THE FIXED-RADIUS METHOD

FLOW PATHS WITHIN THE TIME-OF-TRAVEL AREA FOR A 1-YEAR TRAVELTIME COMPUTED BY USE OF A METHOD THAT INCORPORATES THE SLOPE OF THE WATER TABLE (Javandel and others, 1984)

FOR ALL SIMULATIONS—Aquifer is homogeneous and isotropic.
Potentiometric surface slopes uniformly, parallel to the y axis
Aquifer thickness is 60 feet
Hydraulic conductivity is 50 feet per day
Pumping duration (traveltime) is 1 year
Aquifer porosity is 0.20
Pumping rate is 100 gallons per minute

Figure 17. Relation of 1-year time-of-travel areas computed by use of the fixed-radius method to those computed by use of a method that incorporates the slope of the water table.

A sloping potentiometric surface does not preclude use of the fixed-radius method. The degree to which a uniformly sloping surface affects the time-of-travel area is a function of pumping rate, pumping duration, porosity, aquifer thickness, and hydraulic conductivity (Bear, 1979, p. 283). By use of the dimensionless parameter $[2\pi(Ki)^2 bt/(Q\theta)]$, the percentage of coincidence can be determined for the area delineated by the fixed-radius method and by the method that includes the potentiometric-surface slope (fig. 18). For example, consider a well from which 250 gallons per minute of water is withdrawn for 600 days from an unbounded, confined aquifer that is 100 feet thick. The horizontal hydraulic conductivity of the aquifer is 80 feet per day, aquifer porosity is 0.25, and prepumping potentiometric-surface slope is 20 feet per mile. For this example, the dimensionless parameter equals 2.9. As indicated in figure 18, the fixed-radius method delineates only about 54 percent of the time-of-travel area that would be delineated if the potentiometric-surface slope were considered. In this comparison, only 46 percent of the area delineated by the fixed-radius method contributes water to the well, whereas an equal amount of area not delineated contributes to the well. If a traveltime criterion of 30 days is used, the dimensionless parameter equals 0.14 and the fixed-radius method is more accurate; about 89 percent of the contributing area coincides with that from the method that includes the potentiometric-surface slope.



EXPLANATION

K is horizontal hydraulic conductivity, in feet per day;
i is potentiometric-surface slope, in feet per foot;
b is aquifer thickness, in feet;
t is travel time, in days;
θ is porosity; and
Q is pumping rate, in cubic feet per day.

Assumptions

(1) Flow is two-dimensional and steady
(2) Aquifer is homogeneous, isotropic, and boundaries are distant
(3) Potentiometric-surface slope is uniform

Figure 18. Percentage of coincidence of time-of-travel area delineated by use of fixed-radius method and method that includes consideration of the slope of the potentiometric surface.

Boundaries.--The effect of boundaries on the flow field around wells in valley-fill aquifers in Pennsylvania can be a major limitation on use of this method for delineating time-of-travel areas. The effect of a fully penetrating stream on the shape of the contributing area for a well in an aquifer with no recharge depends on distance from stream to well, aquifer porosity, aquifer thickness, pumping duration, and rate of pumping (Muskat, 1937, p. 475). The area that will be overestimated or underestimated by the fixed-radius method can be determined by use of a dimensionless-time parameter, τ (fig. 19). When τ exceeds about 0.3, time-of-travel areas obtained from the fixed-radius method are greatly overestimated if pumping is near a fully-penetrating stream. For example, if a stream is 200 feet from the well described in the previous paragraph, τ for a 600-day time of travel is 4.6; the fixed-radius method delineates all the actual contributing area and an additional area equal to about 130 percent of the actual area. If a traveltime of 30 days is used, however, τ is 0.23; the fixed-radius method equally overestimates and underestimates small areas, each of which are about 6 percent of the actual time-of-travel area.

Uniform-Flow Method

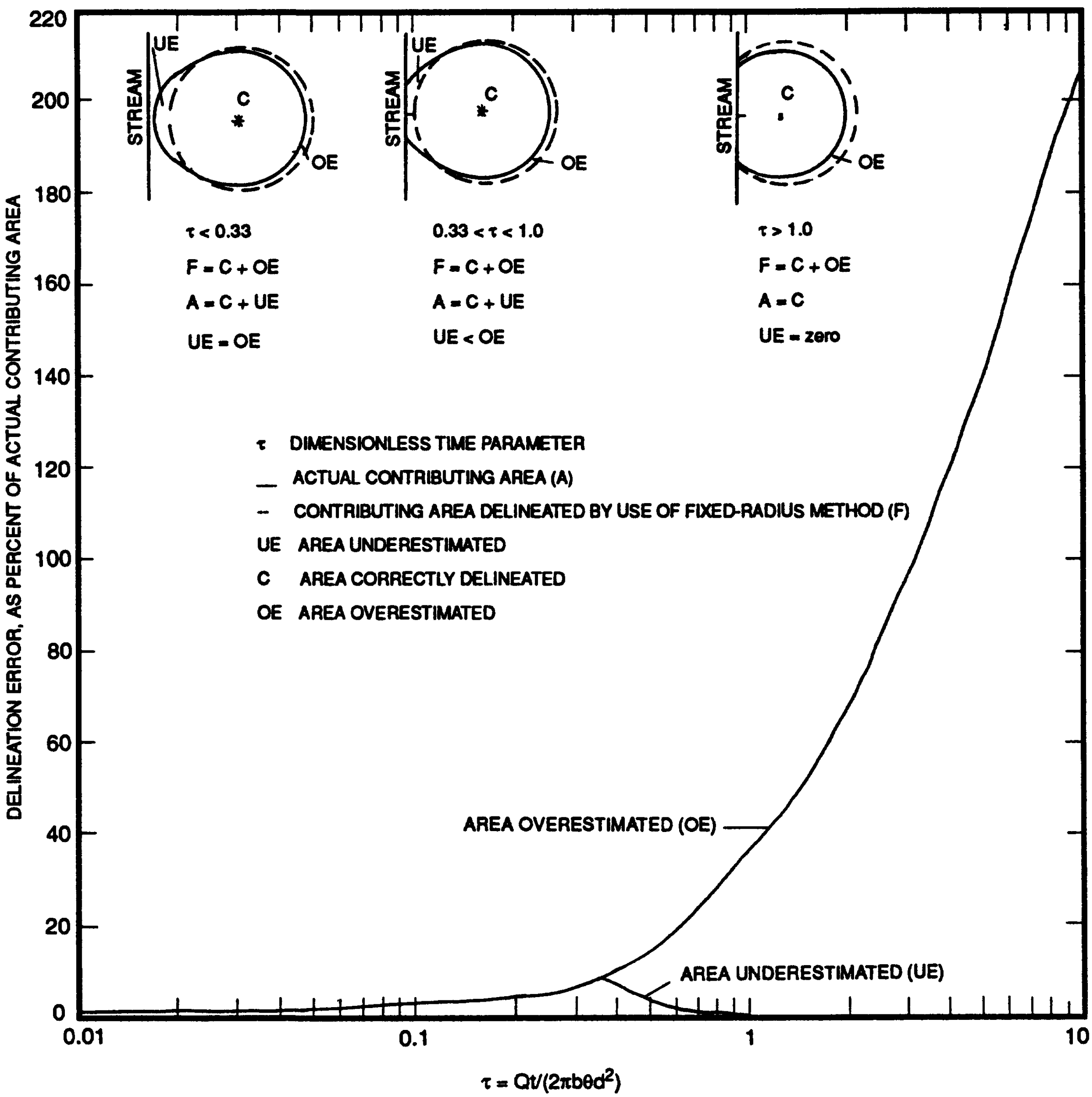
The uniform-flow method is an analytical solution that can be used to estimate the contributing area for steady flow to a well (Bear, 1979, p. 282; Todd, 1980, p. 121). The method is a means of estimating the area of diversion or time-of-travel area of a well. Adjacent contributing areas that can provide recharge to the aquifer within the zone of diversion are not explicitly delineated but can be estimated by analysis of the position of the area of diversion with respect to different recharge sources.

The uniform-flow equations are derived by superposition of the Dupuit equation for radial flow around a well with the one-dimensional, uniform, prepumping flow field. Superposition applies to linear systems where the solution of a problem that includes several inputs is equal to the sum of the solutions to the simpler individual problems. Because the differential equations that describe ground-water flow in confined aquifers are linear, the superposition approach is rigorously correct. Unconfined aquifers, described by nonlinear equations, may be analyzed by superposition only approximately; however, if regional drawdown caused by pumping in an unconfined aquifer is less than about 10 percent of the prepumping saturated thickness, errors caused by nonlinearity will be minor (Reilly and others, 1987, p. 19).

In the uniform-flow method, the aquifer is assumed to be confined (or unconfined if drawdowns are small), of constant thickness, homogeneous, and isotropic. Additionally, the prepumping water table is assumed to be uniformly sloping, and pumping from a fully-penetrating well is assumed to have resulted in a steady state. The assumptions of steady-state flow and a uniformly sloping potentiometric surface are not theoretically possible in an unbounded aquifer; but, if boundaries are distant, results from this method may be satisfactory.

Application

Information on pumping rate, hydraulic conductivity, aquifer thickness, and prepumping potentiometric surface is essential for use of the uniform-flow method. The equations can be solved by use of a handheld calculator, although the trial-and-error approximations required by one of the equations can be tedious. As with all methods, hydrologic judgment is needed to evaluate the validity of areas delineated by use of uniform-flow equations.



EXPLANATION

Q is pumping rate, in cubic feet per day
t is pumping duration (traveltime), in days
b is aquifer thickness, in feet
θ is porosity
d is distance from well to stream, in feet

Figure 19. Error in delineating a time-of-travel area for a well near a fully-penetrating stream by use of the fixed-radius method.

The area of diversion is delineated by calculating the position of the flow lines that separate ground-water flow to the well from flow that bypasses the well (fig. 20). For computation purposes, the well is located at the origin of an x-y cartesian coordinate grid in which the prepumping potentiometric surface parallels the x-axis and slopes in the negative direction on that grid. The following equations are needed to compute the flow-line positions:

$$P = -Q/(2\pi Kbi), \tag{6}$$

$$L = 2\pi P, \text{ and} \tag{7}$$

$$x = -y/[\tan(-y/P)], \tag{8}$$

where P is distance from the well to the stagnation point, in feet;

L is the width between the asymptotic limits separating flow to the well from flow that bypasses the well, in feet;

Q is pumping rate, in cubic feet per day;

K is hydraulic conductivity, in feet per day;

b is aquifer thickness, in feet;

i is uniform slope of the prepumping potentiometric surface, in feet per foot;

x is coordinate distance of a point on the limiting flow line parallel to the uniform flow; and

y is coordinate distance of a point on the limiting flow line perpendicular to the uniform flow.

All angles are in radians.

First, the distance to the stagnation point (P) is determined. This distance will always be negative, indicating that the stagnation point is downgradient from the well. Second, the asymptotic limit of the flow line (L) that separates water moving to the well from flow that bypasses the well is computed. Finally, the location of the flow line between the stagnation point and the asymptotic limit is determined by substitution of values from 0 to 1/2 L for y in equation 8 and solving for x. The solution is symmetric about the x-axis; thus only half of the area of diversion boundary needs to be computed. Note that when x equals 0, y equals 1/4 L.

A time-of-travel area can be estimated by use of a more general form of the uniform-flow equation (Bear and Jacobs, 1965). Additional data needs are aquifer porosity, time-of-travel criteria, and x-y coordinates of observation points. The time of travel (t) for any point within the limits of flow to the well (except on the x- or y-axes) is computed by use of equation 9.

$$t = \left(\frac{x\theta}{Ki}\right) + \left[\ln\left(\frac{\sin(\arctan(y/x))}{\sin((-y/P) + \arctan(y/x))}\right)\left(\frac{P\theta}{Ki}\right)\right] \tag{9}$$

where K, i, b, P, and Q are as previously defined;

θ is aquifer porosity;

y is distance from well normal to the uniform flow, in feet; and

x is distance from well parallel to the uniform flow, in feet.

All angles are in radians.

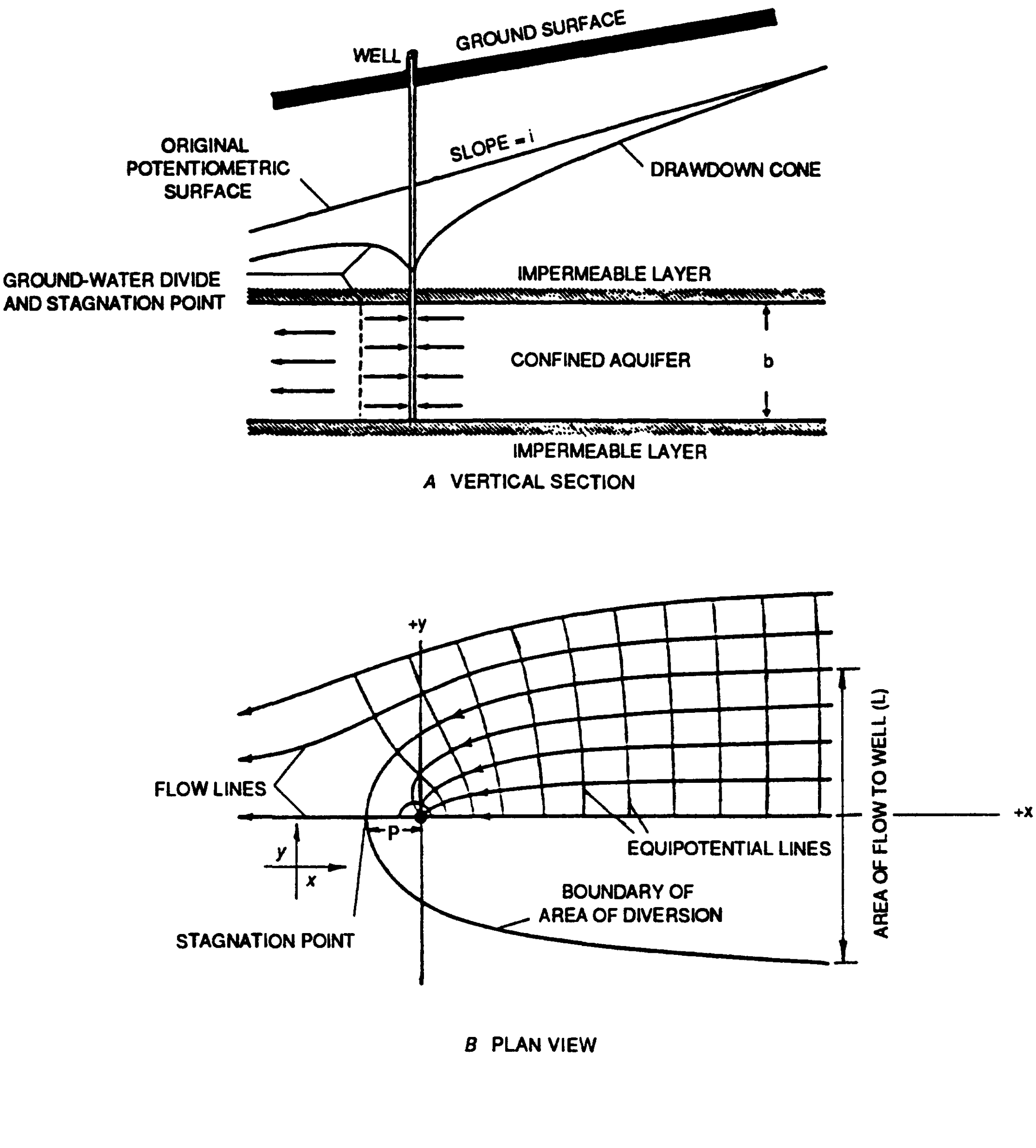


Figure 20. Steady-state flow to a well penetrating a confined aquifer having a uniformly sloping prepumping potentiometric surface: (A) vertical section and (B) plan view. (Modified from U.S. Environmental Protection Agency, 1987, p. 4-15.)

To solve for the position of a time-of-travel area from equation 9, one must compute t for several x-y coordinates until the bounding line of the area can be drawn. The solution is symmetrical about the x-axis so only positive values of y need to be computed. The equation works for any x-y coordinate except for those points along the x-axis parallel to the regional uniform flow where y = 0. To compute the upgradient position of any time-of-travel area on the x-axis, one must solve the following equation by trial and error:

$$x = (Kit/\theta) + [\ln (1-(x/P))/(-P)]. \qquad (10)$$

Examples.--The area of diversion is computed by use of equations 6-8 for the following conditions: pumping rate, 500 gallons per minute; aquifer thickness, 80 feet; hydraulic conductivity of the aquifer, 100 feet per day; and prepumping water-table slope, 20 feet per mile.

Q = 96,244 cubic feet per day (500 gallons per minute)

K =100 feet per day

b = 80 feet

i = 0.00379

$$P = \frac{-96,244 \text{ feet per day}}{2\,(3.1416)\ (100 \text{ feet per day})\ (80 \text{ feet})\ (0.00379)} = -505 \text{ feet}$$

L = 2 (3.1416) (-505 feet) = 3,173 feet

The area of diversion is sketched in figure 21 from determinations of the x and y coordinates of its boundary between the stagnation point (P) and limiting value (L) by use of equation 8. An example of the calculation of x for a given y value of 1,000 feet is:

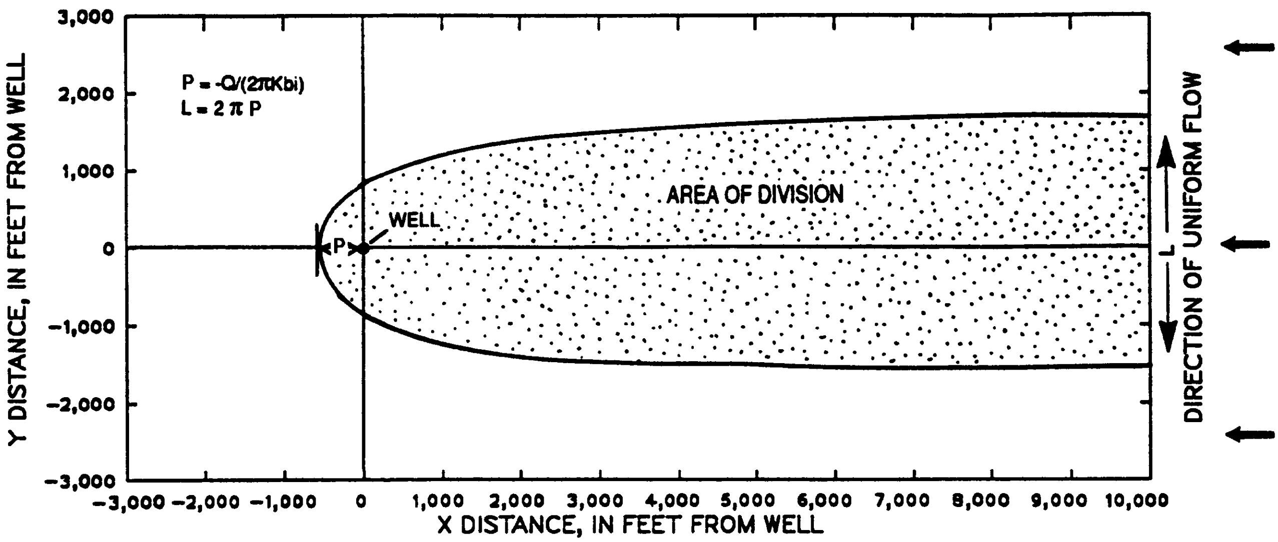
$$x = \left[\frac{-1{,}000 \text{ feet}}{\tan\left(\dfrac{-1{,}000 \text{ feet}}{-505 \text{ feet}}\right)}\right] = 434 \text{ feet}$$

The upgradient boundary of the area of diversion is not defined by this method. Theoretically, it would extend to the upgradient source of water to the aquifer.

The time-of-travel areas for several selected traveltimes are shown in figure 22. These areas of the aquifer were computed by use of equation 9, well characteristics described in the previous example (fig. 21), and a porosity of 0.15.

Limitations

For valley-fill aquifers in Pennsylvania, areas of diversion and time-of-travel areas estimated by use of the uniform-flow method may not be accurate because of the effects of boundary conditions and aquifer heterogeneities. The inability of the equation to account for boundary conditions is especially important because most valley-fill aquifers are crossed by through-flowing streams and are restricted by bedrock walls whose permeabilities are less than those of the aquifers. The effect of a nearby boundary is illustrated in figure 23, which shows a well being pumped near a fully penetrating stream where uniform flow is perpendicular to the stream (Jacob, 1950, p. 349). Water will be induced from the stream if Q is greater than $\pi dKbi$, where d is the distance from the well to the stream.



EXPLANATION

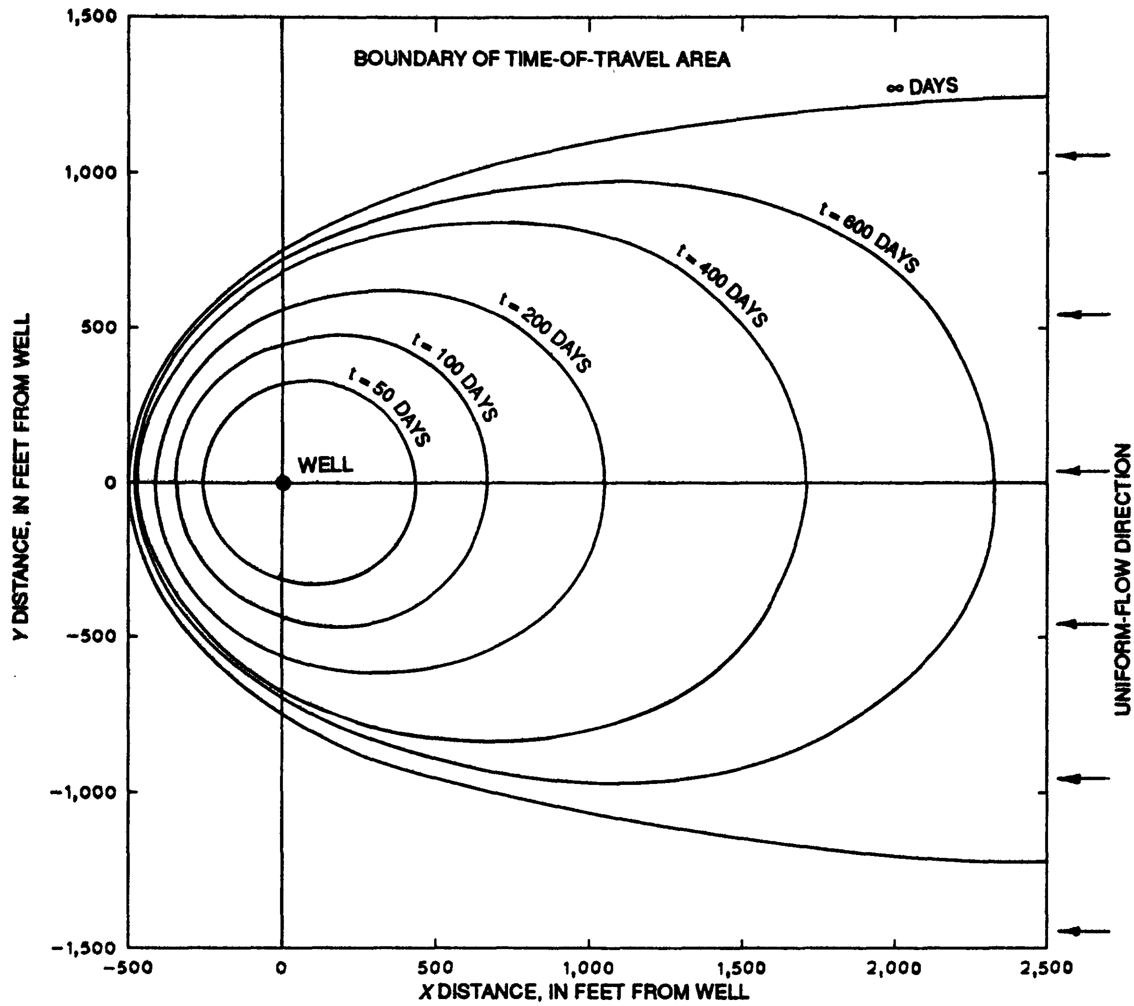
AQUIFER AND WELL CHARACTERISTICS

Hydraulic conductivity (K) is 100 feet per day
Aquifer thckness (b) is 80 feet
Slope of prepumping poteniometric surface (i) is 20 feet per mile
Pumping rate (Q) is 500 gallons per minute

DISTANCE FROM WELL TO STAGNATION POINT (P) IS -505 FEET

WIDTH BETWEEN ASYMPTOTIC LIMITS OF BOUNDARY SEPARATING FLOW TO WELL FROM THAT BYPASSIING IT (L) IS ± 3,173 FEET

Figure 21. Estimation of the area of diversion by use of the uniform-flow method.



EXPLANATION

AQUIFER PROPERTIES AND WELL CHARACTERISTICS

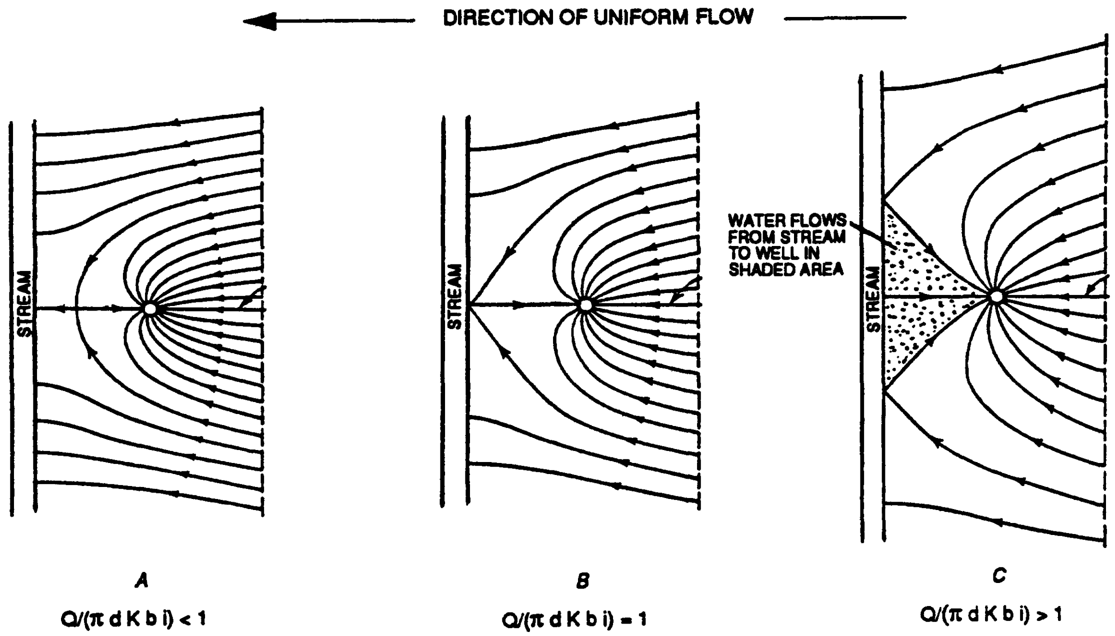
Hydraulic conductivity (K) is 100 feet per day
Aquifer thickness (b) is 80 feet
Slope of prepumping potentiometric surface (i) is 20 feet per mile
Pumping rate (Q) is 500 gallons per minute
Porosity (θ) is 0.15

Travel times (t) are computed using the equation:

$$t = \frac{x\theta}{Ki} + \left[\ln\left(\frac{\sin\left(\arctan\left(\frac{y}{x}\right)\right)}{\sin\left(\left(\frac{-y}{P}\right) + \arctan\left(\frac{y}{x}\right)\right)} \right) \left(\frac{P\theta}{Ki}\right) \right] ,$$

where y is distance from well normal to uniform flow, in feet;
x is distance from well parallel to uniform flow, in feet;
P is distance from well to the stagnation point; and
all angles are in radians.

Figure 22. Estimation of time-of-travel areas for selected traveltimes by use of the uniform-flow method.



EXPLANATION

Q is pumping rate, in cubic feet per day;

d is distance from well to stream, in feet;

K is hydraulic conductivity, in feet per day;

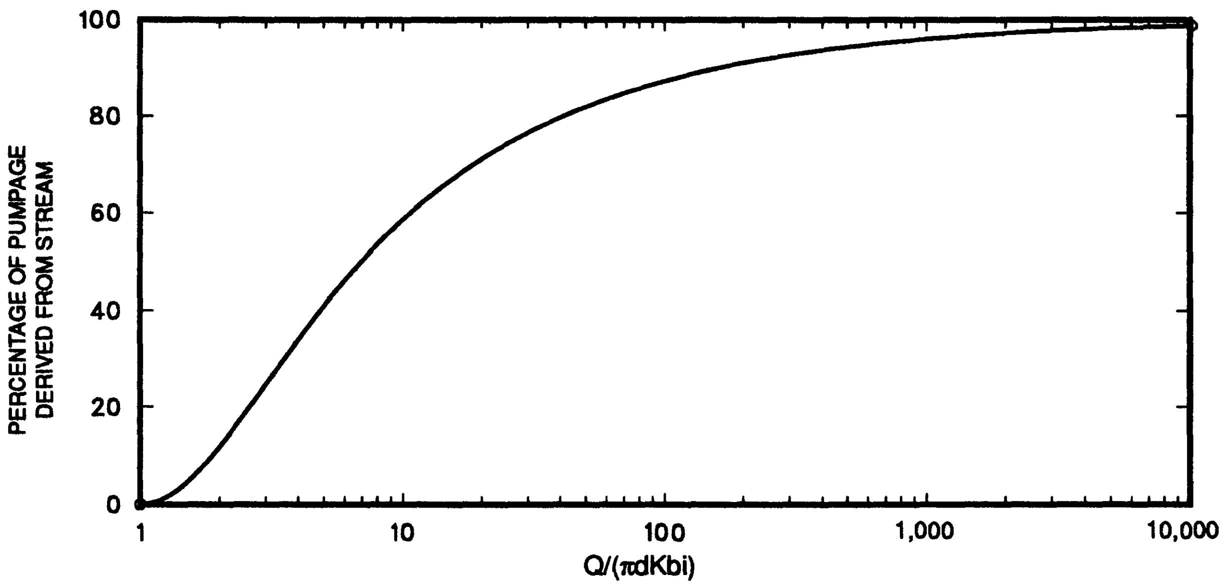
i is potentiometric-surface slope, consistent units; and

b is aquifer thickness, in feet.

Figure 23. A pumped well in a uniform-flow field perpendicular to a fully penetrating stream: (A) no flow induced from the stream, (B) limiting case, and (C) flow induced from the stream to the well.

The percentage of water derived from the stream can be estimated from figure 24. The position of the asymptotic limit of the flowlines separating water that flows to the well from water that bypasses the well would be reduced in proportion to the amount of water derived from the stream. For example, if the well described in figure 21 was 1,200 feet from a fully-penetrating stream, the factor Q/(πdKbi) would be less than 1; therefore, no water would be derived from the stream, and the asymptotic limit of the area-of-diversion boundary and stagnation point shown in figure 21 would not be affected. If the well was only 300 feet from the stream, however, Q/(πdKbi) would be 3.4, and about 28 percent of the water flowing to the well would be water derived from the stream. The limit (L) to which the area-of-diversion boundary approaches (1,588 feet for this example) should be decreased 28 percent, to about 1,143 feet. If the value of Q/(πdKbi) was less than 1, the stream could still affect the contributing-area boundary near the well although the limit (L) to which the boundary asymptotically approaches would remain unaffected. These effects become negligible for a dimensionless parameter of less than about 0.1.

Although the previous example is for a restrictive case, it illustrates how the uniform-flow field can be affected by nearby boundaries. The effect of a partially penetrating stream on the area of diversion would be less than that shown in figure 24. Effects of other boundary conditions and various uniform-flow directions are described in Bear (1979) and in Dacosta and Bennett (1960).



EXPLANATION

Q is pumping rate, in cubic feet per day;
d is distance from well to fully penetrating stream, in feet;
K is hydraulic conductivity, in feet per day;
b is aquifer thickness, in feet; and
i is potentiometric-surface slope, in consistent units.

Figure 24. Percentage of pumpage from a well derived from a fully penetrating stream where uniform flow is perpendicular to the stream.

Analytical Methods

Analytical methods can be used to estimate the area of diversion of a well in a bounded valley-fill aquifer by application of superposition (Reilly and others, 1987) and image-well theory (Ferris and others, 1962). Adjacent contributing areas that can provide recharge to the aquifer within the zone of diversion are not explicitly delineated but can be estimated by analysis of the position of the area of diversion with respect to various recharge sources. The superposition procedure involves three steps: (1) computing drawdown values by use of an analytical equation that best fits the hydrogeologic situation; (2) subtracting drawdown values from the prepumping potentiometric surface; and (3) sketching the area of diversion from the resultant potentiometric surface. An example of this procedure is shown in figure 25. If the well is affected by either a linear recharge or a discharge boundary such as a river or valley wall, drawdown values in step 1 are computed by use of image wells. In such instances, image wells are positioned on the basis of the aquifer geometry, and the drawdown or water-level increase caused by each image well at the location of interest is computed separately. The resultant drawdown is the algebraic sum of drawdowns from all individual image wells (fig. 26). Use of image wells in this way allows flexibility in representing boundary conditions and the choice of many possible analytical equations for computing drawdown.

A variety of analytical equations can be used to calculate the drawdown near a pumped well. The best known of these are used to solve for restrictive conditions of steady flow (Thiem, 1906) and nonsteady flow (Theis, 1935) to a homogeneous, ideal aquifer. More complex solutions are available for leaky aquifers (Hantush and Jacob, 1955; Hantush, 1960), anisotropy in the horizontal plane (Papadopulos, 1965; Hantush, 1966), multiple aquifers (Neuman and Witherspoon, 1969), water-table aquifers (Boulton, 1963; Neuman, 1975), partially penetrating wells (Hantush, 1961; Neuman, 1974), and finite-diameter wells (Papadopulos and Cooper, 1967; Papadopulos, 1967). Other solutions are summarized in Bear (1979), Kruseman and de Ridder (1976), Lohman (1972), and Walton (1988). Computation of drawdown is greatly facilitated by computer programs such as those found in Reed (1980), Walton (1988), and van der Heijde and Beljin (1988).

In image-well theory, ground-water flow is assumed to be two-dimensional, and boundaries are assumed to be fully penetrating and linear. The aquifer is assumed to be confined (unconfined if drawdowns are small) and homogeneous. Other assumptions depend on the analytical equation used. Various equations account for leakage, horizontal anisotropy, partially penetrating wells, casing storage, and many other aquifer and well characteristics. Data needed depend on the analytical equation selected, but all require estimates of distance from the production well to boundaries, pumping rate, hydraulic conductivity, and saturated aquifer thickness. If the solution is for nonsteady flow, data on the duration of pumping and aquifer storage will be needed.

Application

Aquifer boundaries are replaced by image wells to prevent drawdown along the river boundary and flow across the impermeable boundary. The image is placed across the boundary so that the distance from the image to the boundary equals that between the real well and the boundary. If the boundary is a river, the image well is an injection well. If the boundary is an impermeable valley wall, the image well is a withdrawal well. The magnitude of injection or withdrawal at each image well is equal to the rate of pumping from the real production well.

Where multiple boundaries are simulated, a network of image wells is needed. In figure 27, for example, a discharging image well (I_2) is used to simulate the impermeable boundary; however, drawdown caused by this image is affected by the perennial river boundary. Thus, a recharging image well (I_3) is added. This pattern repeats (I_2, I_3, I_6, I_7) across both boundaries to infinity. The same logic applies to image wells used to simulate the perennial river (I_1, I_4, I_5, I_8). In practice, image wells are added until the effect of an additional well is insignificant.

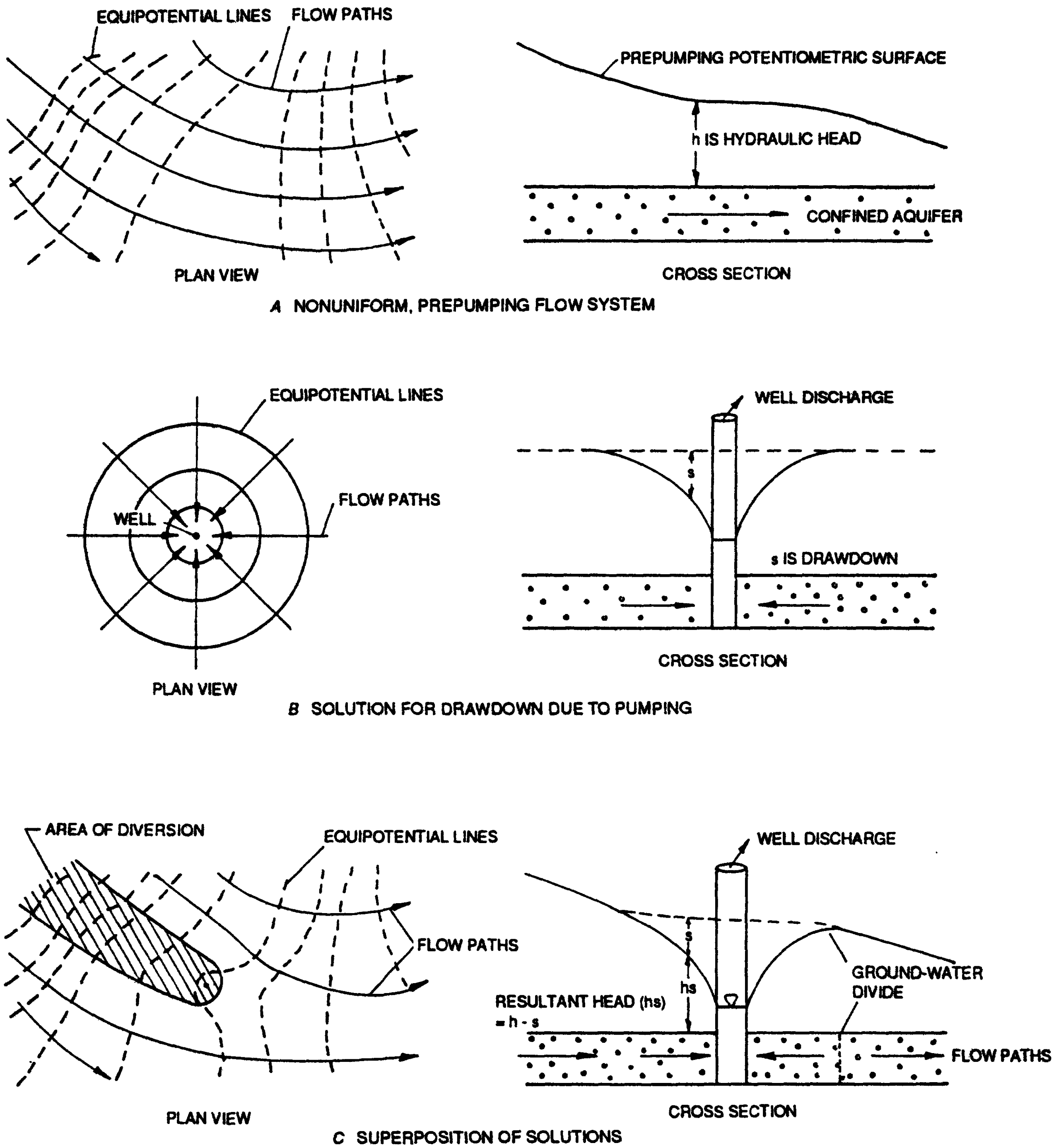
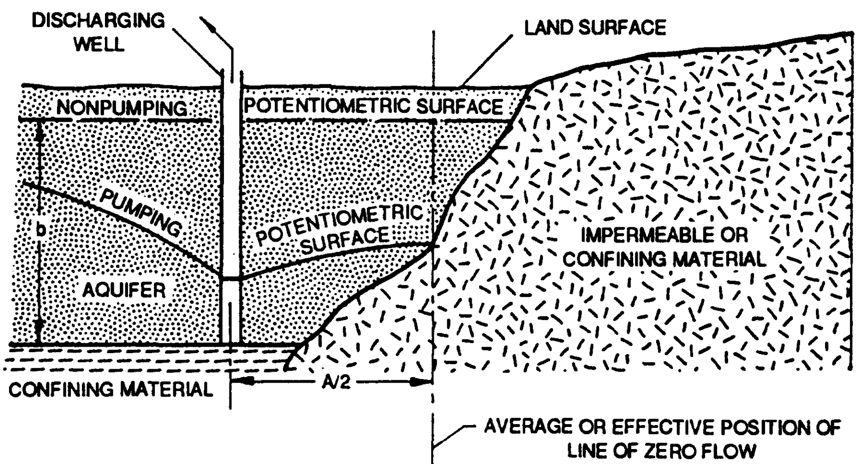
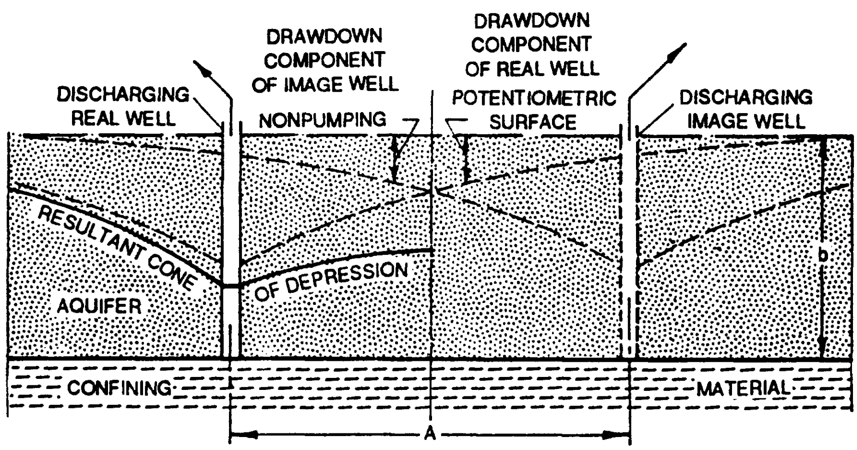


Figure 25. Diagram showing (A) prepumping potentiometric surface, (B) drawdown due to pumping, and (C) potentiometric surface and area of diversion from superposition of (A) and (B).



A REAL SYSTEM



NOTE: Aquifer thickness (b) should be very large compared to resultant drawdown near real well
"A" equals twice the distance between real well and impermeable boundary

B HYDRAULIC COUNTERPART OF REAL SYSTEM

Figure 26. Simulation of drawdown around a well near an impermeable valley wall by superposition of image-well effects. (Modified from Ferris and others, 1962, p. 149.)

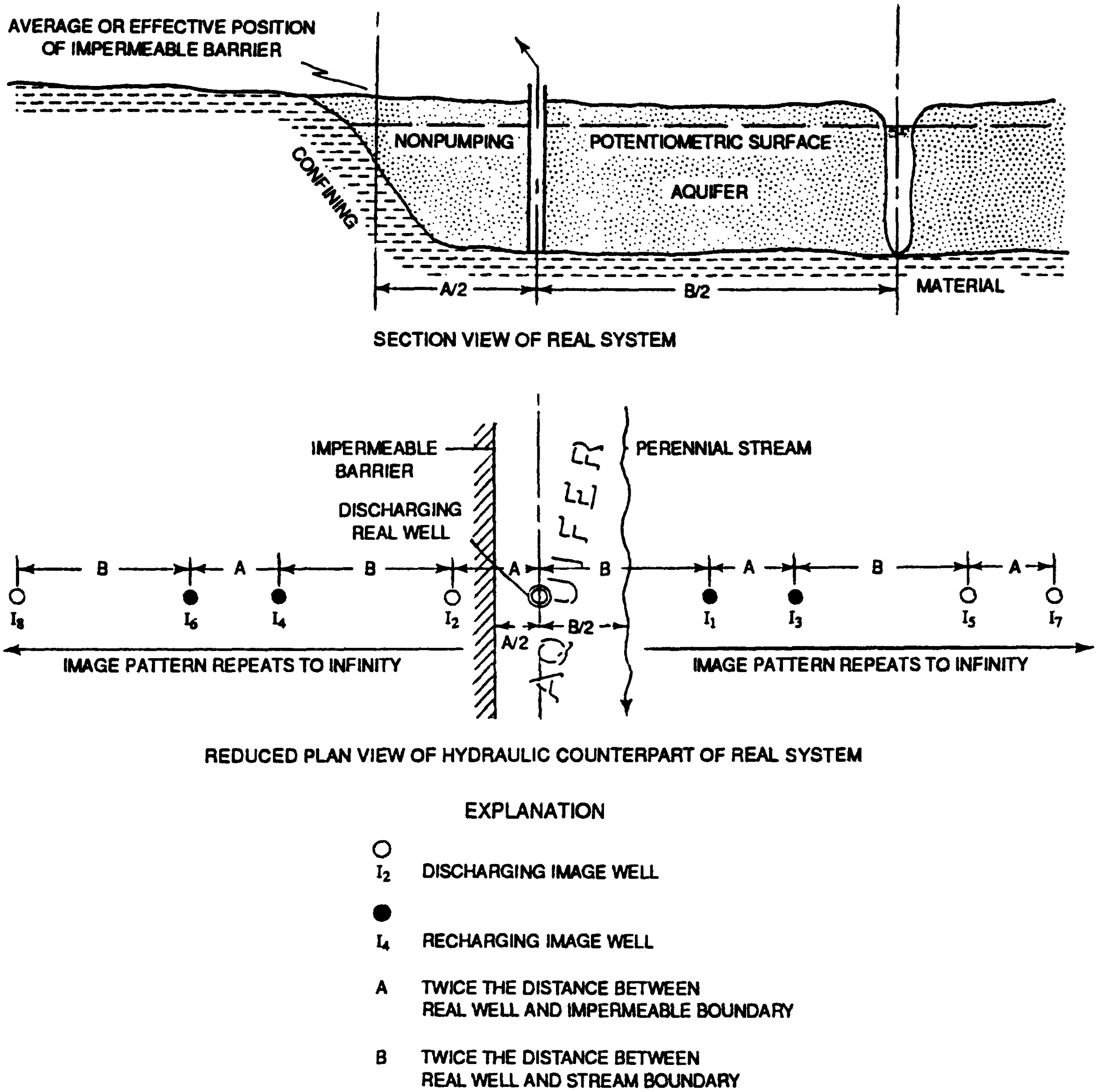


Figure 27. Simulation of parallel boundaries by use of an infinite array of image wells. (Modified from Ferris and others, 1962, p. 157.)

Special Case, Strip Aquifers.--If the boundary geometry is fairly simple, the effects of image wells can be included in a closed-form solution such as that for a strip aquifer. Use of the strip-aquifer method requires that an aquifer be bounded by a fully-penetrating linear stream of infinite length on one side and by a parallel infinite, linear, impermeable boundary or stream on the other side. This geometry results in a strip aquifer of infinite length as shown by the strip aquifer bounded by a stream and valley wall in figure 28. The geometry assumed in the strip-aquifer method should be applicable to some situations in valley-fill aquifers in Pennsylvania. Similar closed-form solutions are available for other boundary configurations such as semi-infinite aquifers, wedge-shaped aquifers, and bounded quadrants (Rorabaugh, 1956; Hydrologisch Colloquium, 1964; Kruseman and de Ridder, 1976; Bear, 1979).

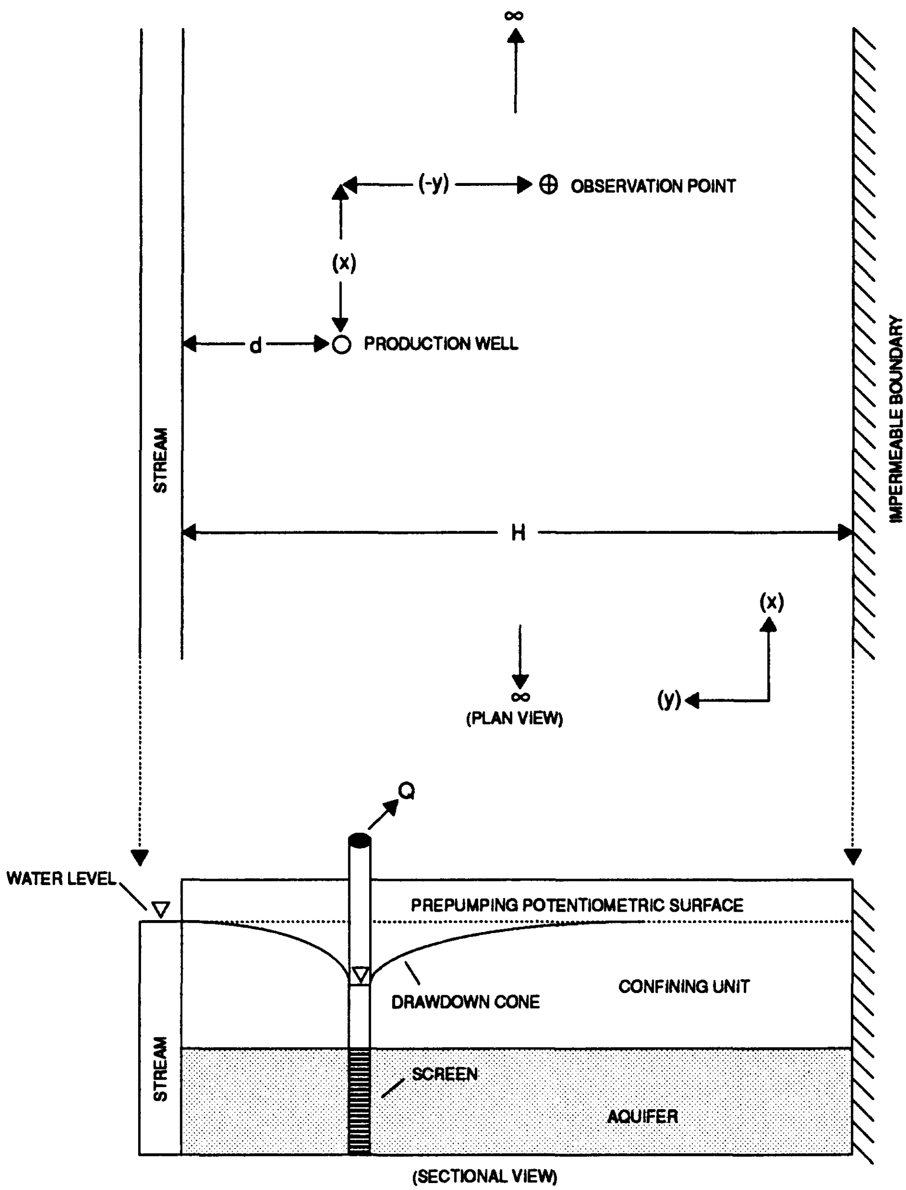


Figure 28. Geometry of a strip aquifer.

The strip-aquifer method is applied to the strip aquifer bounded by a stream and parallel valley wall by use of the following equations to compute drawdown (Kirkham, 1949):

Drawdown in the pumped well (s_w) (angles are in radians)

$$s_w = \frac{Q}{(2\pi T)} \ln \left[\frac{\tan\left(\frac{2\pi d}{4H}\right)}{\tan\left(\frac{\pi r}{4H}\right)} \right]. \tag{11}$$

Drawdown at any observation point (s) (angles are in radians)

$$s = \frac{Q}{(4\pi T)} \ln \left[\left(\frac{\cosh\left(\frac{\pi x}{2H}\right) + \cos\left(\frac{\pi y}{2H}\right)}{\cosh\left(\frac{\pi x}{2H}\right) - \cos\left(\frac{\pi y}{2H}\right)} \right) \left(\frac{\cosh\left(\frac{\pi x}{2H}\right) - \cos\left[\frac{\pi (y-2d)}{2H}\right]}{\cosh\left(\frac{\pi x}{2H}\right) + \cos\left[\frac{\pi (y-2d)}{2H}\right]} \right) \right], \tag{12}$$

where s_w is drawdown in the production well, in feet;
s is drawdown in the observation point, in feet;
T is aquifer transmissivity, in feet squared per day;
Q is pumping rate, in cubic feet per day;
H is aquifer width, in feet;
d is distance from the production well to the stream, in feet;
r is well radius, in feet;
x is distance along the stream from the production to the observation well, in feet; and
y is distance normal to the stream from the production to the observation well, in feet.

The distance to observation points are all referenced in x-y coordinates from the pumped well (fig. 28). The solution for drawdown is symmetrical about the well in the x direction; thus, computations can be reduced by solving for only half of the strip aquifer. The potentiometric surface that results from the pumping is estimated by subtracting the drawdown values from the prepumping potentiometric surface. The area of diversion on the aquifer surface can then be sketched from the resulting potentiometric surface.

Example of Use of Image Wells.--Image wells were used to compute drawdown near a stream in a confined, valley-fill aquifer 2,000 ft wide. The well is 220 feet from a fully penetrating linear stream and 1,780 ft from the valley wall. The boundaries are simulated by use of recharging image wells placed 220 feet across the stream and valley wall. Discharge (Q) of the real well is 500 gallons per minute; therefore, the image wells inject water at the same rate. The duration of pumping (t) is 1 year, storage coefficient (S) is 0.20, and transmissivity is 4,000 square feet per day. Drawdown values were computed by use of the Theis (1935) equation (fig. 29B). The steady-state area of diversion (fig. 29C) is determined by superposition of the computed drawdown values (fig. 29B) from the prepumping water-table surface (fig. 29A).

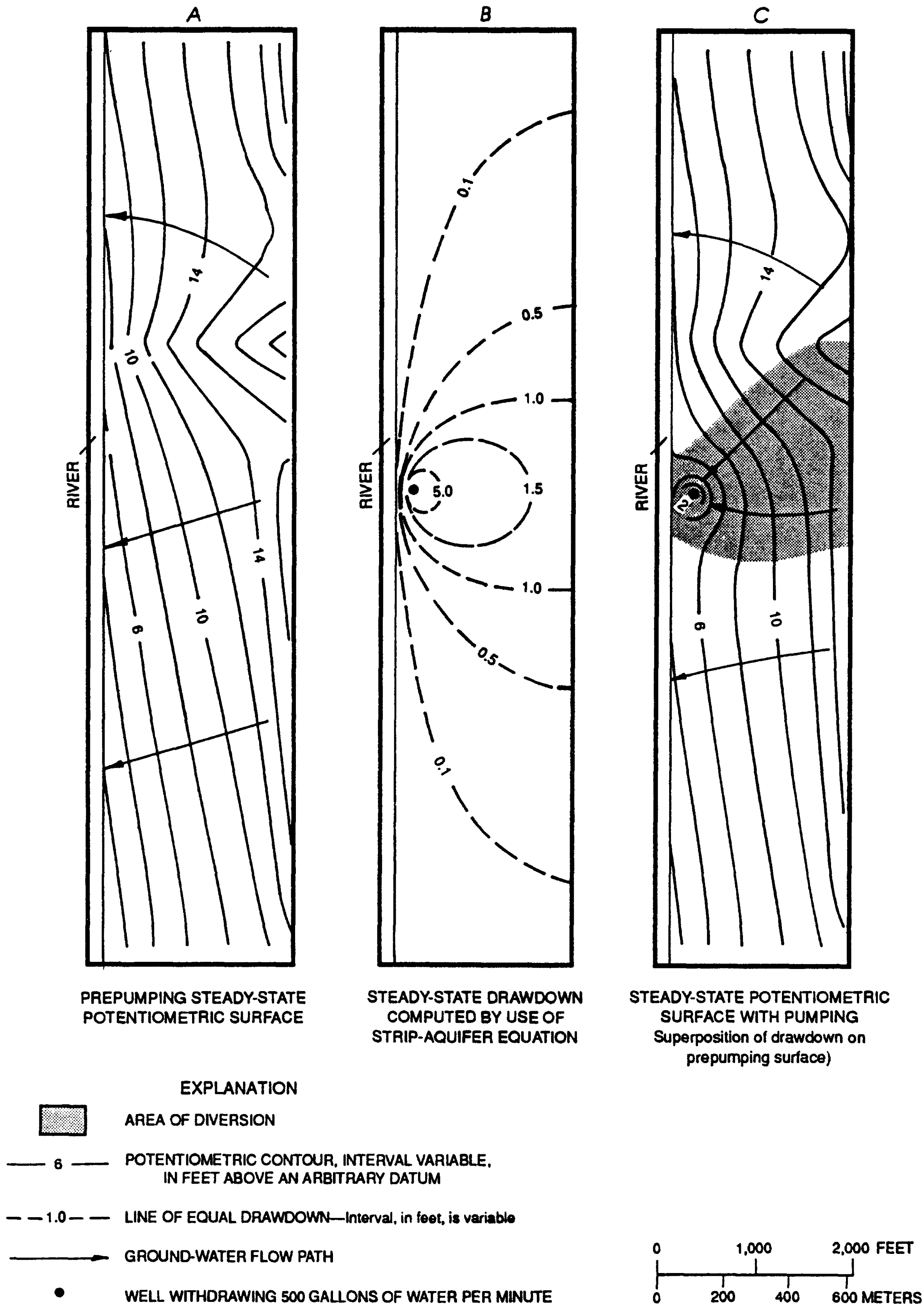


Figure 29. Estimation of the area of diversion by use of the analytical method and image wells.

Example of Use of the Strip-Aquifer Method.—The same problem described above is solved by use of the strip-aquifer method. Computed drawdown values are the same as the ones shown in figure 29. An example of the computation from equation 12 for an observation point 100 feet toward the stream (+y direction) and 40 feet upvalley or downvalley is

$$s = \left[\frac{96,257 \text{ cubic feet per day}}{(4\pi)\,(4,000 \text{ square feet per day})}\right]$$

$$\ln\left[\left(\frac{\cosh\left[\frac{(\pi)\,(40 \text{ feet})}{(4,000 \text{ feet})}\right] + \cos\left[\frac{(\pi)\,(100 \text{ feet})}{(4,000 \text{ feet})}\right]}{\cosh\left[\frac{(\pi)\,(40 \text{ feet})}{(4,000 \text{ feet})}\right] - \cos\left[\frac{(\pi)\,(100 \text{ feet})}{(4,000 \text{ feet})}\right]}\right)\left(\frac{\cosh\left[(\pi)\left(\frac{40 \text{ feet}}{4,000 \text{ feet}}\right)\right] - \cos\left[(\pi)\frac{((100 \text{ feet}) - (440 \text{ feet}))}{4,000 \text{ feet}}\right]}{\cosh\left[(\pi)\left(\frac{40 \text{ feet}}{4,000 \text{ feet}}\right)\right] + \cos\left[(\pi)\frac{((100 \text{ feet}) - (440 \text{ feet}))}{4,000 \text{ feet}}\right]}\right)\right]$$

= 4.45 feet.

The drawdown distribution computed by use of equation 12 was subtracted (superposed) from the prepumping potentiometric surface. The area of diversion was sketched from the resulting potentiometric surface with the same results as shown in figure 29.

Limitations

The advantage in the use of image wells as illustrated is that one of many analytical equations may be used. The method, however, is limited to treating boundaries in an ideal manner. Hydrologic judgement is needed to determine whether the ideal boundaries specified by the model reasonably represent the field situation of interest. A linear, impermeable boundary will probably be adequate where valley walls are steep and the hydraulic conductivity of the valley fill is at least 10 times that of the bedrock upland. Near bounding valley walls that are not virtually impermeable, image-well withdrawal rates can be adjusted (Walton, 1988, p. 241) by use of the equation

$$Qi = Q\,(T - Ti)/(T + Ti), \qquad (13)$$

where Qi is pumping rate of image well, in gallons per minute;

Q is pumping rate of real well, in gallons per minute;

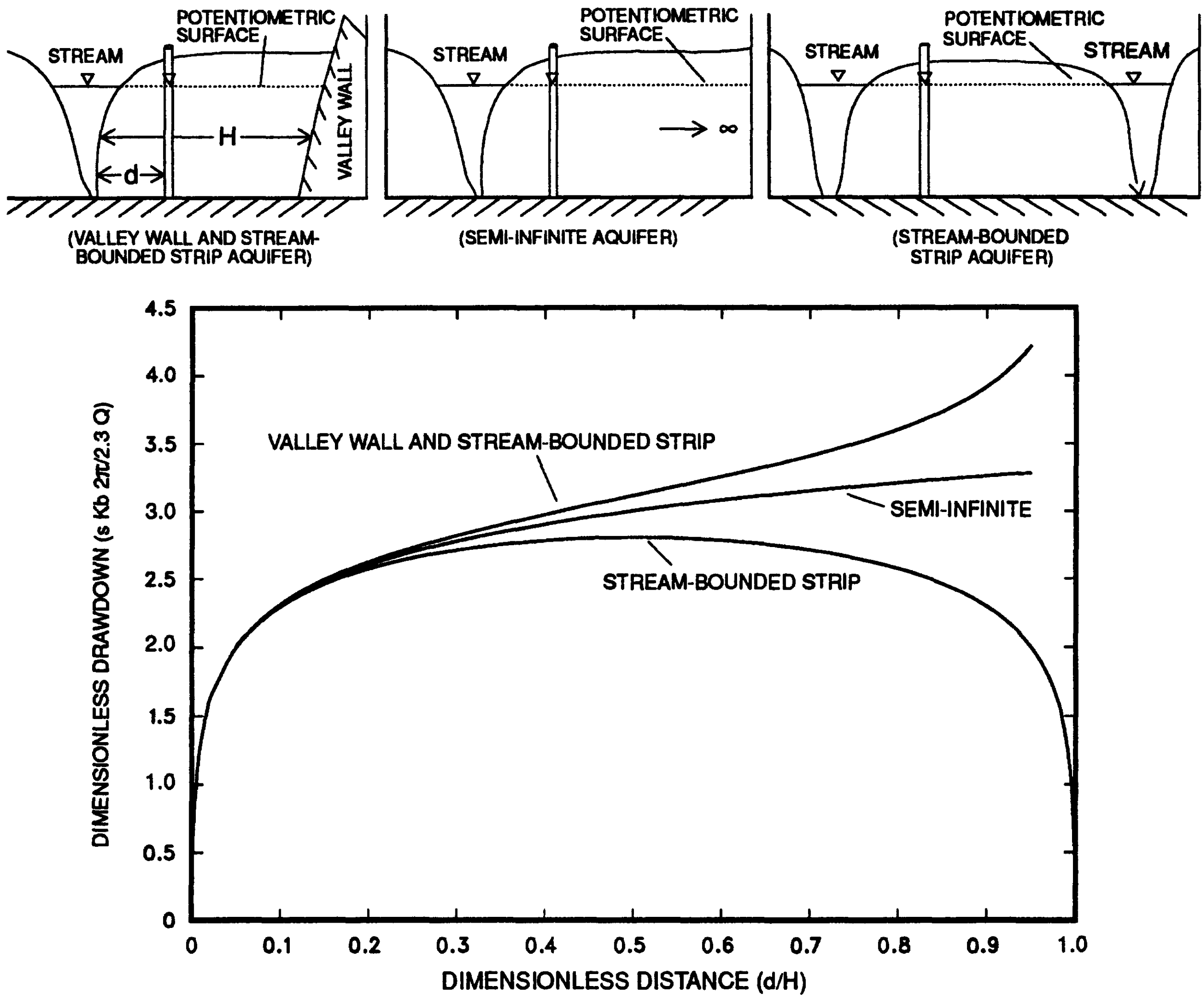
T is transmissivity near the real well, in feet squared per day; and

Ti is transmissivity beyond the boundary near the image well, in feet squared per day.

The treatment of streams as fully penetrating boundaries in valley-fill aquifers is not a realistic portrayal of most streams in Pennsylvania, with the possible exception of sections of large rivers such as the Allegheny, Monongahela, Susquehanna, and Ohio. The propagation of effects of pumping beneath partially penetrating streams violates boundary conditions assumed in image-well theory.

In the example shown in figure 29, a nonequilibrium equation for drawdown (Theis, 1935) was used to estimate the steady-state area of diversion to a well. In theory, a strict steady-state flow cannot be represented by this method; however, a long duration of pumping will approximate that condition adequately. For pumping from a well near a stream in a homogeneous, isotropic aquifer, about 90 percent of the water discharged at the well will be balanced by a depletion of streamflow if $t = [d^2S/(0.3T)]$, where t is duration of pumping, d is distance from the well to stream, S is storage coefficient or specific yield, and T is transmissivity (Theis, 1941). A pumping duration longer than t will provide an approximate steady-state drawdown solution.

In places where the valley wall is distant or the strip aquifer is bounded by two parallel streams, the closed-form equations for a semi-infinite aquifer (Rorabaugh, 1956) or stream-bounded strip (Kruseman and de Ridder, 1976, p. 112) can be used in place of the valley wall and stream bounded strip-aquifer method described previously. The sensitivity of computed drawdown to these boundary conditions is illustrated in figure 30. Solutions are similar where the distance from the stream to production well (d) is less than about 30 percent of the aquifer width (H) but are increasingly divergent as the well is moved further from the stream.



EXPLANATION

s is drawdown in observation well, in feet;

K is hydraulic conductivity, in feet per day;

b is aquifer thickness, in feet;

Q is pumping rate, in cubic feet per day;

d is distance from observation well to stream, in feet; and

H is aquifer width, in feet. (Here, (H) in the semi-infinite aquifer is the width throughout which drawdown is computed.)

Figure 30. Effects of boundaries on drawdown for three bounded-aquifer methods applicable to valley-fill aquifers.

Semianalytical Models

Semianalytical models are popular for estimating the area of diversion to a well. The popularity of these models is derived from their ease of application and ability to provide graphical plots of ground-water flow paths and traveltime fronts. The models are called semianalytical because an analytical equation is used to compute the hydraulic-head distribution around a well and a numerical technique is used to compute traveltime fronts. To compute traveltime fronts, one must know the velocity distribution everywhere in the flow field. Several techniques are available to determine the velocity distribution; the general steps involved in one approach are outlined below:

1. Drawdown because of pumping is computed from an analytical equation. Image wells can be used if linear boundaries are present.
2. Drawdown is superposed upon a prepumping uniform-flow field to derive the hydraulic-head distribution around the well.
3. Equipotentials, streamlines, and velocity distribution are computed from the hydraulic-head distribution.
4. "Particles" are tracked along streamlines to establish traveltime fronts by use of a numerical technique.

Because numerical techniques involve a large number of calculations, semianalytical methods require the use of a computer. Examples of some semianalytical computer codes include PATHS[2] (Nelson and Schur, 1980), RESSQ[2] (Javandel and others, 1984), DREAM[2] (Bonn and Rounds, 1989) and WHPA[2] (Blandford and Huyakorn, 1989). Other codes are listed in van der Heijde and Beljin (1988).

Most semianalytical computer codes were written for two-dimensional, steady-state flow in a homogeneous, isotropic, confined aquifer. If drawdown is small relative to the saturated thickness, the method also applies to an unconfined aquifer. If drawdowns are influenced by linear, fully penetrating boundaries, image wells can be used to simulate their effects. In equations used to evaluate the velocity distribution around the well, the prepumping flow field is assumed to be a uniformly sloping, planar surface.

Application

Semianalytical methods require use of a computer and information on pumping rate, porosity, aquifer thickness, pore velocity of uniform flow, well radius, and distance to boundaries. Some model codes require an estimate of hydraulic conductivity or transmissivity, depending on how the velocity field and streamlines are formulated. The software generally is menu driven and easy to use.

Semianalytical methods can be used to estimate the area of diversion, time-of-travel areas, and quantitative streamline positions. Adjacent contributing areas that can provide recharge to the aquifer within the zone of diversion are not explicitly delineated but can be estimated from analysis of the position of the area of diversion with respect to different recharge sources.

Examples.--The semianalytical model RESSQ (Javandel and others, 1984, p. 35) was used to illustrate how a semianalytical model can compute streamlines and time-of-travel areas. For this example, a 6-inch diameter well being pumped at 100 gallons per minute is 500 feet from a fully penetrating stream. A uniform prepumping flow velocity of 164 feet per year is directed normal to the stream. The aquifer thickness is 50 feet, and porosity is 0.20.

Streamlines and time-of-travel areas are shown in figure 31. The streamlines toward the well are quantitative. They indicate that one-third of the pumpage is induced from the stream. The remainder of the pumpage is captured ground water that would have discharged to the stream. Time-of-travel areas were plotted for traveltimes of 30 days, 2 years, and 5 years. At 30 days, the area of diversion is nearly circular; for longer time periods, the effects of the stream and uniform-flow field distort the radial flow near the well.

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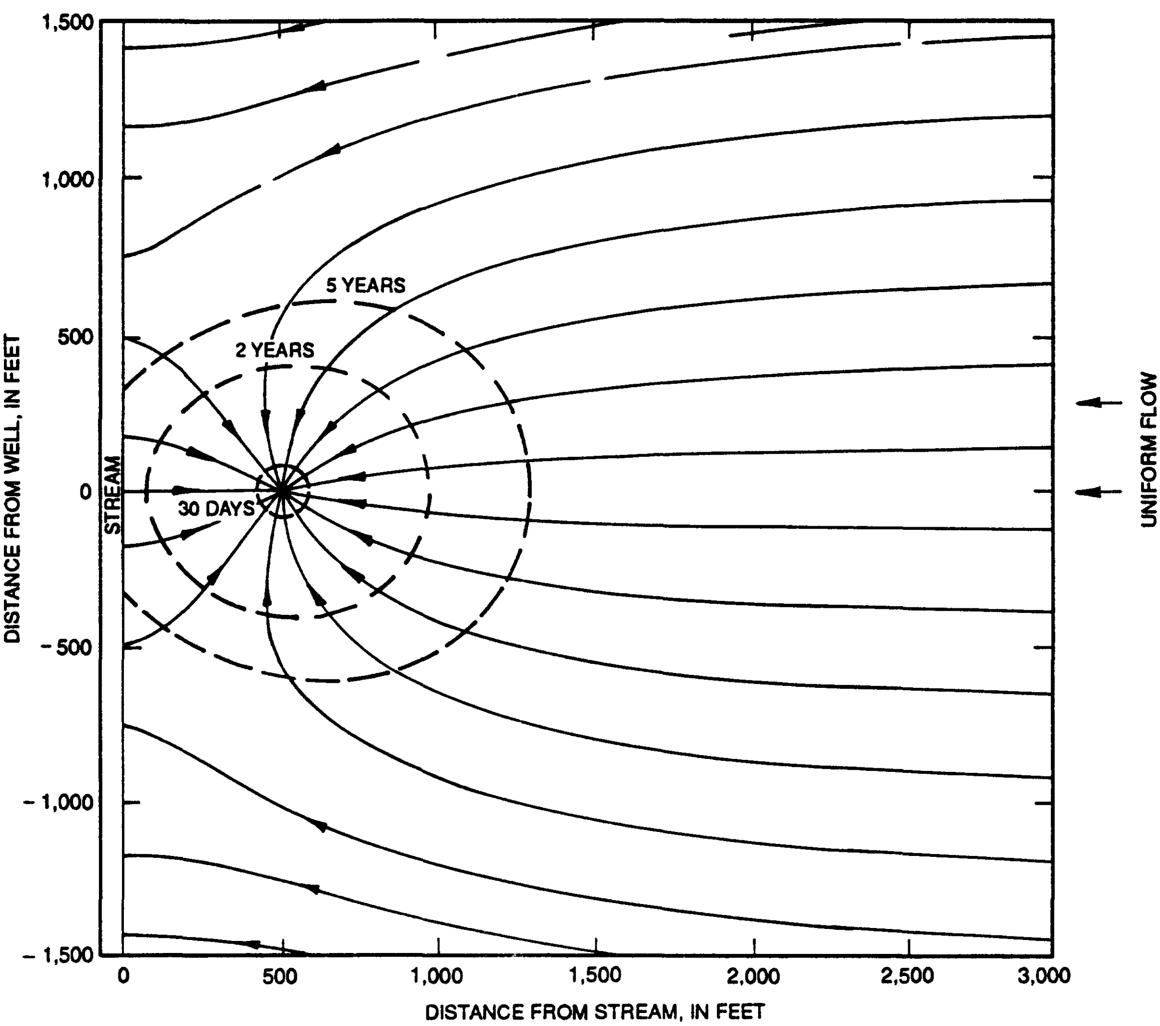
[2] The use of names of proprietary software in this report is for identification only and does not constitute endorsement by the U.S. Geological Survey.



EXPLANATION

— — — BOUNDARY OF TIME-OF-TRAVEL AREA

GROUND-WATER FLOW PATH

● WELL PUMPED AT 100 GALLONS PER MINUTE

AQUIFER AND WELL CHARACTERISTICS:

Aquifer thickness is 50 feet
Porosity is 0.20
Pore velocity of prepumping uniform flow is 164 feet per year
Well radius is 0.25 foot
Distance from well to stream is 500 feet

Figure 31. Time-of-travel areas for selected traveltimes computed by use of a semianalytical method for a well pumping near a stream.

Semianalytical methods can be used to delineate time-of-travel areas for well fields where pumping from one well interferes with another. The RESSQ simulation (fig. 32) shows the distortion of 1-year time-of-travel areas for three closely spaced wells, each being pumped 100 gallons per minute from an aquifer described with uniform flow and distant boundaries.

Limitations

The semianalytical method is a powerful and flexible way to determine effects of pumping in a uniform-flow field. A major advantage of this method is the ease with which time-of-travel areas can be computed. Also, because streamlines can be plotted, quantitative information about the source of water to wells can be obtained by this method.

The same restrictions that apply to the uniform-flow method hold for the semianalytical method, with the exception that well interference can be simulated and boundaries can be approximated by use of image wells. Image wells must be used with caution, however, because some semianalytical computer codes allow only restrictive geometries to be simulated. For example, the RESSQ code (Javandal and others, 1984) allows uniform flow with boundaries only if the flow is perpendicular to a recharge boundary or parallel to an impermeable boundary. Another important restriction is that of a uniform prepumping flow field. In real aquifers, water-table surfaces are irregularly shaped; which can alter contributing areas considerably from those that are based on a uniformly sloping surface.

Numerical Flow Models

Numerical flow modeling is a widely used and powerful method for studying the effects of pumping on a hydrologic system. The power of this method is derived from its ability to simulate most factors that affect the contributing area of a well, including (1) nonlinear, nonfully penetrating boundary conditions, (2) complex patterns of recharge and discharge, and (3) spatial heterogeneity of hydraulic properties.

Numerical models approximate the partial differential equation of ground-water flow by means of a matrix of algebraic equations that can be solved simultaneously with a computer. Many numerical models based on various techniques to approximate and to solve the ground-water flow equation have been documented. Van der Heijde and Beljin (1988) list 27 fully documented flow models that could be used for delineation of contributing areas to wells. Examples of widely used models include those described by Prickett and Lonnquist (1971), Trescott (1975), and McDonald and Harbaugh (1988).

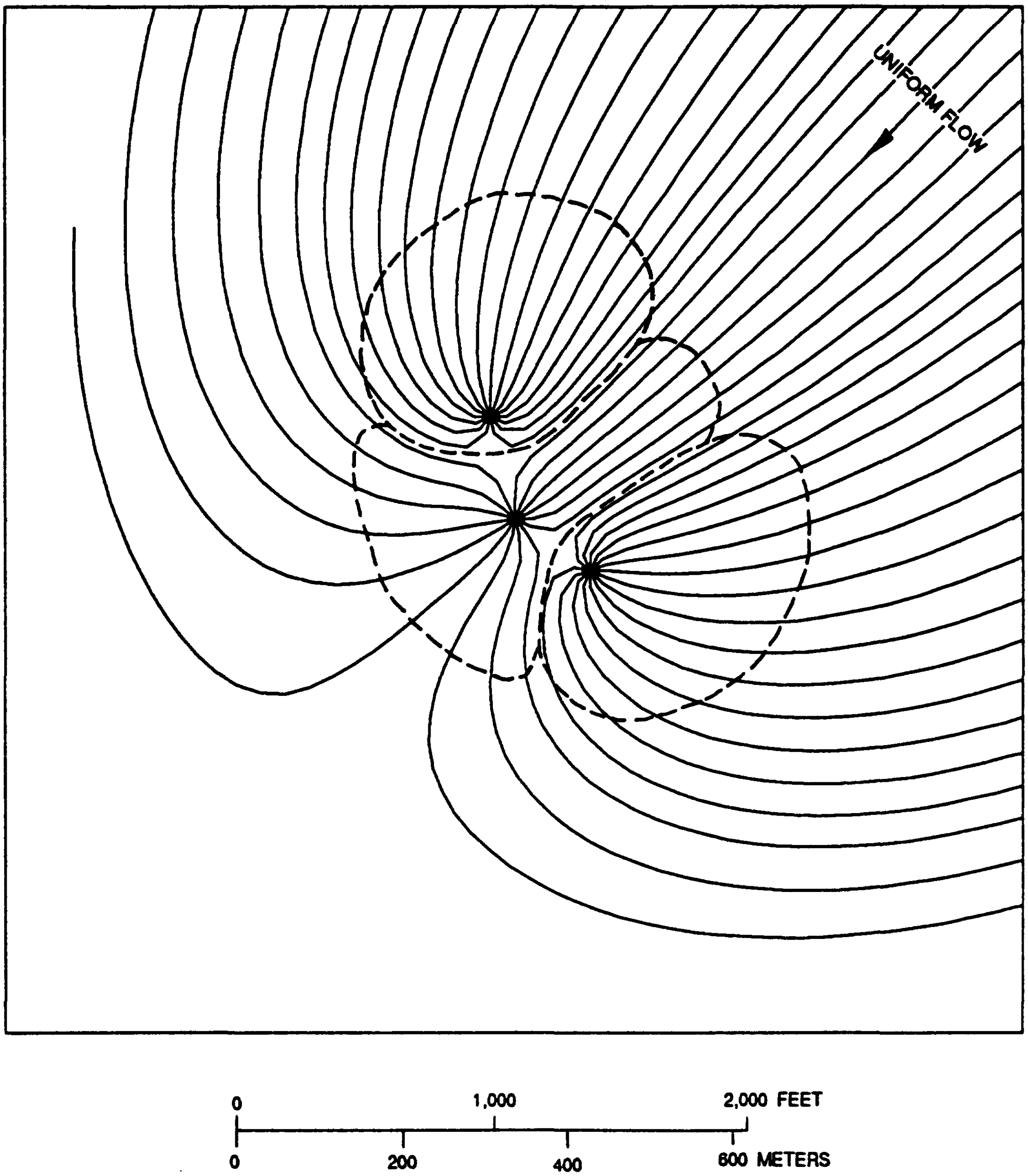
Data needed to delineate contributing areas depend more on the hydrologic system being simulated than on the actual model used. The data fall into the following categories:

Boundaries and initial conditions:

1. Aquifer geometry–thickness and internal structure of valley fill.
2. Prepumping hydraulic-head distribution in three dimensions.
3. Recharge–flow across the water-table surface, seepage from tributary streams, infiltration from unchanneled upland runoff and interflow, and regional ground-water flow from adjacent bedrock.
4. Discharge–location of streams and springs that drain the aquifer, evapotranspiration, and underflow.

Hydraulic properties:

1. Hydraulic conductivity–horizontal and vertical spatial variations.
2. Specific yield and storage coefficient–for transient simulations.
3. Hydraulic conductivity of riverbed sediments.
4. Porosity–needed for computing time of travel.



EXPLANATION

BOUNDARY OF TIME-OF-TRAVEL AREA

GROUND-WATER FLOW PATH

WELL PUMPED AT 100 GALLONS PER MINUTE

AQUIFER AND WELL CHARACTERISTICS:
Aquifer thickness is 33 feet
Porosity is 0.25
Pore velocity of prepumping uniform flow is 262 feet per year
Well radius is 0.25 foot

Figure 32. One-year time-of-travel areas computed by use of a semianalytical method for three closely spaced wells.

Well characteristics:

1. Pumping rate.
2. Position (depth and length) of screened interval.
3. Location of well relative to impervious or recharge boundaries.

Data needs listed here include the major factors that affect the contributing area. The data requirements for simulating an aquifer by use of a numerical model are more extensive than for the other methods.

Use of numerical models requires a computer and the judgement of an experienced hydrologist. Results from models can be erroneous when the conditions in the real hydrologic system are poorly conceptualized. Poor conceptualization can result if either the real system is not fully understood or limitations of the numerical method are not considered.

Application

Numerical flow models are used to simulate the hydraulic-head distribution throughout the modeled area. If the hydraulic properties of an aquifer are uniform, the area of diversion can be approximated from the simulated potentiometric surface by a sketch of the limiting flowline position. Alternatively, particle-tracking programs have been developed (Shafer, 1987; Zheng, 1989; and Pollock, 1989) to compute streamlines and time-of-travel areas from the hydraulic-head and flux values simulated by the flow model.

Application of a numerical flow model involves several steps and feedback loops (fig. 33). The major steps are conceptualization, data collection, model construction, model adjustment, and prediction. First, the flow system is conceptualized so the essential elements are identified (at least in a semi-quantitative

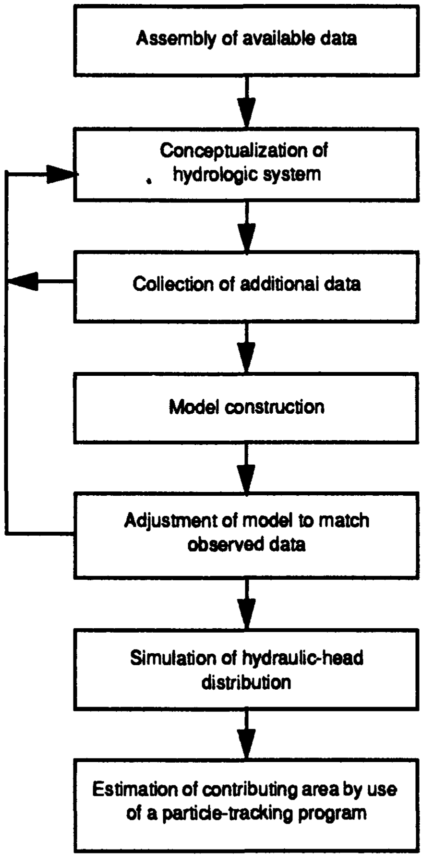


Figure 33. General procedure for delineating a contributing area by use of a numerical model.

manner). Additional data may be needed to refine the conceptualization. Next, an appropriate model is selected and constructed to represent the conceptualized hydrologic system. The study area is then divided into cells that represent boundary conditions and define the flow of water into and out of the modeled area. Hydraulic properties and recharge rates are adjusted until the model adequately simulates observed data. Finally, the simulated potentiometric surface and the flow budget can be used with a particle-tracking program to estimate the contributing area to a well. Several authors provide detailed information on modeling applications and philosophy (Konikow, 1978; Mercer and Faust, 1980; van der Heijde and others, 1988; Walton, 1988; Wang and Anderson, 1982; Anderson and Woessner, 1992).

Commonly, numerical flow models are constructed to simulate ground-water flow in two dimensions for systems where vertical flow components are insignificant. Morrissey (1989) found that, for a hypothetical valley-fill aquifer less than 100 feet thick, contributing areas delineated by two- and three-dimensional models are similar if the ratio of horizontal to vertical hydraulic conductivity is less than 10; however, vertical gradients can be high if (1) aquifers are thick, (2) vertical variations in aquifer properties are significant, and (3) wells do not fully penetrate the aquifer. In these cases, a three-dimensional model would provide a more complete approximation of the hydrologic system.

Three examples of the use of numerical models are presented in the following section. The examples illustrate the flexibility of the modeling approach in simulation of complex boundary conditions, heterogeneity in aquifer properties, and partial penetration of wells.

Example 1, complex boundaries.--A three-dimensional, finite-difference model (McDonald and Harbaugh, 1988) and particle tracker (Pollock, 1989) were used to simulate the area of diversion and the contributing area for the ideal aquifer described in the section "Effects of Pumping on Valley-Fill Aquifers." This example illustrates the complex boundary conditions that can be simulated by use of a numerical model. Aquifer properties and boundary conditions assigned to the four-layer model are shown in figure 8. Boundary conditions simulated by the model include (1) recharge from precipitation, (2) recharge from a partly penetrating tributary stream, (3) recharge from ground water in bedrock, (4) recharge from upland runoff and interflow, (5) discharge to a partly penetrating river with a sloping bottom, and (6) discharge to a wetland.

The area of diversion, contributing area, and sources of water to a well being pumped at 500 gallons per minute are shown in figures 11 and 12. The complexity of the model allows eight separate sources of water to be identified and their contributing areas to be approximated. More than three-fourths of the water that moved directly to the well is from sources other than precipitation on the valley fill.

Example 2, heterogeneity of hydraulic conductivity.--A two-dimensional model is used to simulate an aquifer in which hydraulic conductivity is heterogeneous. This distribution of hydraulic conductivity (fig. 34A) represents a valley where a river has meandered across its width and has left behind permeable sediments in abandoned oxbows. The aquifer is recharged by infiltration of 20 inches of precipitation per year. All discharge is to a fully penetrating river and the well. The prepumping potentiometric surface in this example does not clearly indicate the differences in hydraulic conductivity within the aquifer; however, when the well is pumped at 200 gallons per minute, preferential zones of ground-water flow become apparent as illustrated by the area through which water moved to the well for a traveltime of 2 years (fig. 34B). This area, which is primarily within zones where hydraulic conductivity is 200 feet per day, is the 2-year time-of-travel area.

Example 3, vertical anisotropy with a partially penetrating well.--Differences in hydraulic properties in vertical and horizontal directions also can affect the area of diversion to a well. The effect can be especially pronounced if the well is partially penetrating. In this example, the three-dimensional simulation of the ideal aquifer (fig. 8) is modified so that the ratio of horizontal to vertical hydraulic conductivity is 40. A well is simulated to pump water from the bottom 20 feet of the aquifer at a rate of 200 gallons per minute. The area of diversion and the area in which precipitation is captured are shown in figure 35. Owing to anisotropy, precipitation that infiltrates near the well bypasses the well and discharges to the river; thus, the area of diversion expands further than it would if this recharge were available to the well. This

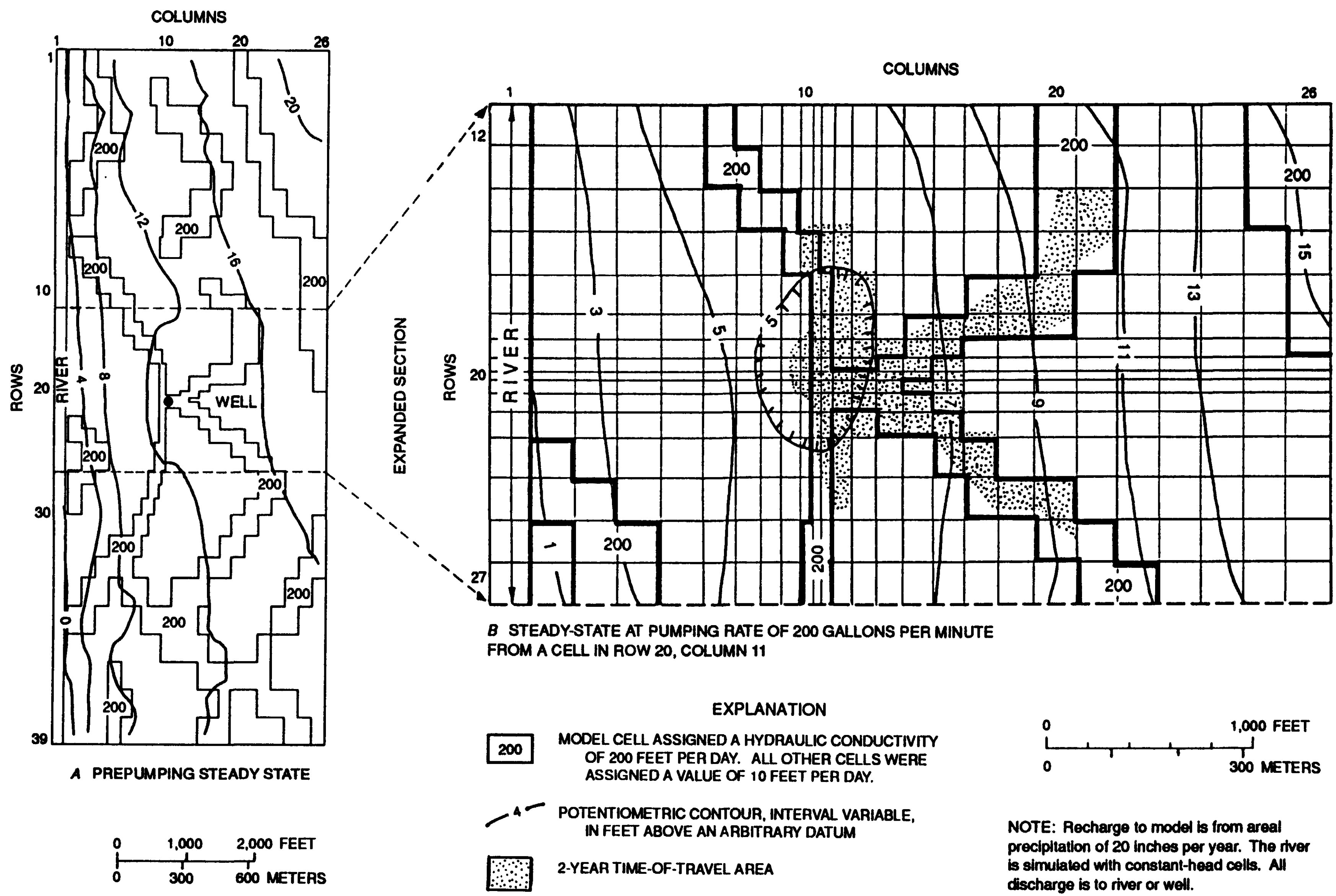
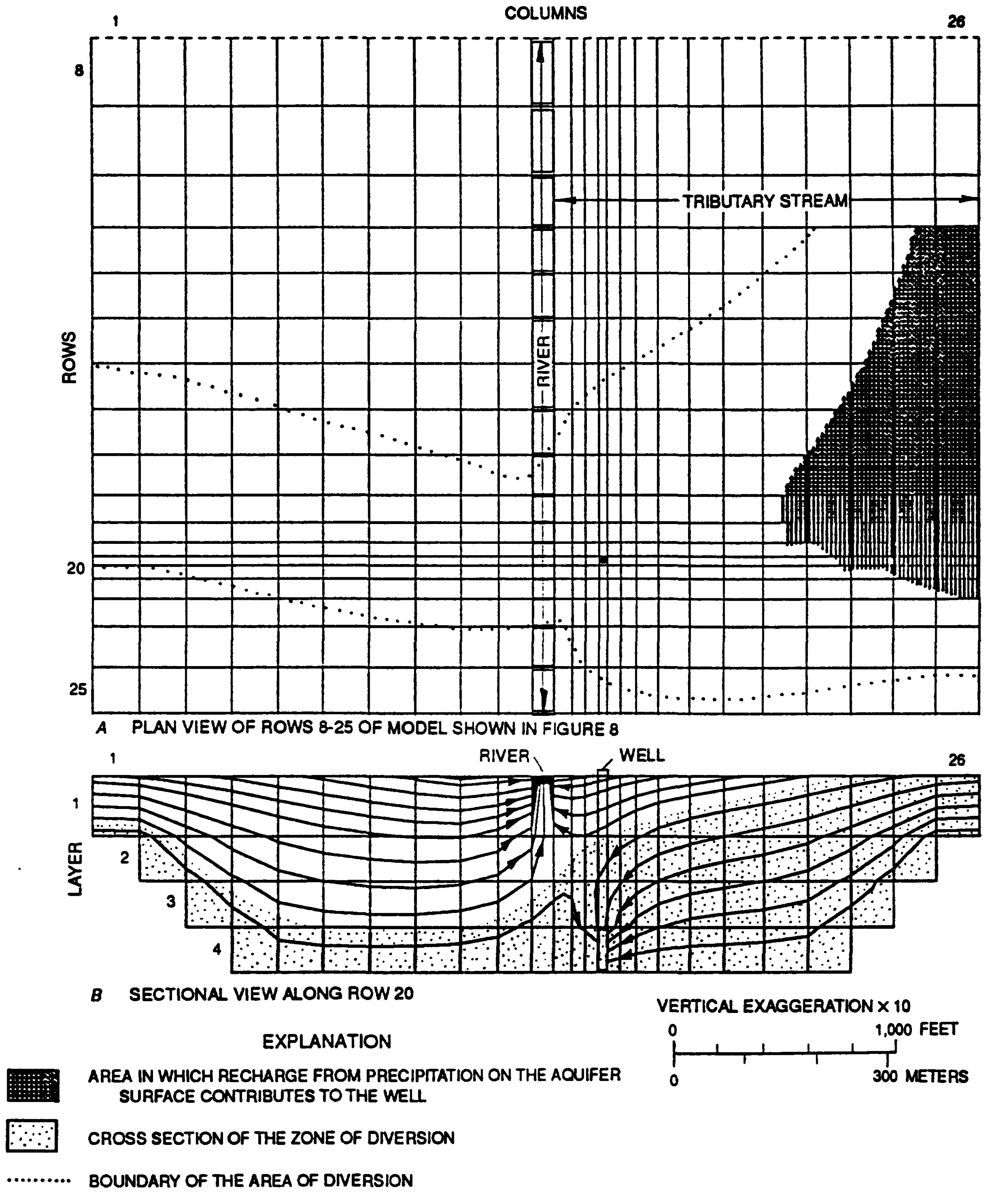


Figure 34. Two-dimensional-model simulation of an aquifer having a heterogeneous hydraulic-conductivity distribution (example 2).



NOTE: Model is same as that shown in figure 8 except that horizontal hydraulic conductivity is 40 times greater than in the vertical direction.

Figure 35. Three-dimensional numerical simulation for an aquifer with a horizontal-to-vertical hydraulic-conductivity ratio of 40 (example 3).

example illustrates a limitation of the techniques that do not account for vertical anisotropy (such as analytical methods and areal two-dimensional numerical models) and that will result in underestimation of the area of diversion to the well.

Limitations

As illustrated in the previous examples, the area of diversion to a well can be extremely dependent on adequate identification of vertical and horizontal variations in hydraulic properties. Unfortunately, the heterogeneity in hydraulic properties can be difficult to quantify. Field investigations such as aquifer tests are useful; but such efforts are costly, and results are not always definitive.

The significance of heterogeneities in hydraulic conductivity on the shape of the area of diversion is not always apparent because their effect is scale dependent. The two-dimensional numerical simulation (fig. 36) illustrates this phenomenon, where the shape of the 100-day time-of-travel area is influenced by small zones of virtually impermeable clay near the well. The area of diversion, which involves longer travel paths than the time-of-travel area does, is not affected.

The accuracy of numerical models is dependent upon adequate discretization of the study area. Grid cells need to be small enough that the aquifer geometry, hydraulic properties, and potentiometric surface are adequately simulated; but not so small as to make computations more difficult than necessary. Generally, small cells are needed near wells to define the steeply sloping potentiometric surface, whereas larger cells may be acceptable further from the well. Adequacy of the grid spacing can be determined only by trial and error. If the simulated shape of the area of diversion does not change when a grid with additional cells is used, then the discretization is sufficiently small.

Although numerical methods can simulate much of the complexity within an aquifer system, even the most elaborate models are a simplification of the real system. A numerical model can never be shown to be a uniquely correct simulation of the hydraulic system; other models can always be constructed that also will adequately represent the measured characteristics of the system. Therefore, even though numerical simulation offers the best possible representation of an aquifer system, an area of diversion delineated by this method must be viewed as an approximation.

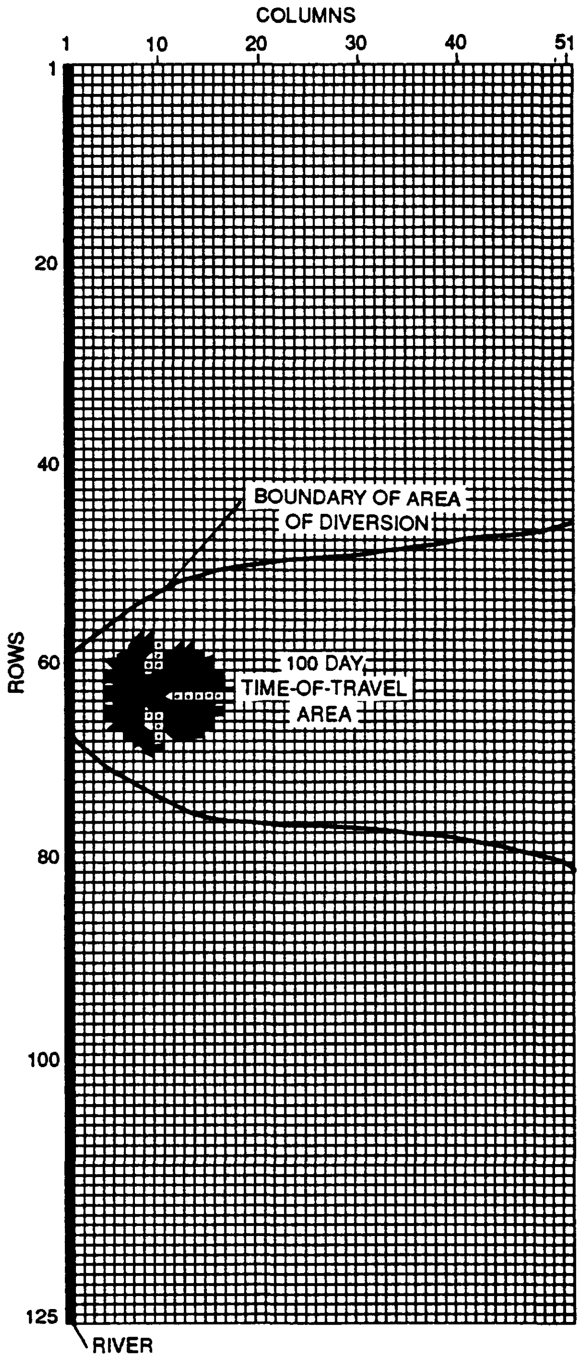
Comparison of Methods

Methods to delineate the area of diversion, contributing area, and time-of-travel area are compared for two aquifer systems. First, the methods are compared for the idealized aquifer shown in figures 8 through 13. As previously discussed, this setting includes many of the complexities of valley-fill aquifers in Pennsylvania. The methods also are compared for a real well field in the valley-fill aquifer along Marsh Creek near Asaph, Pa. At the Marsh Creek site, large annual fluctuations in natural recharge and ground-water withdrawals make delineation of contributing areas difficult.

Idealized Valley-Fill Aquifer

As previously discussed, a three-dimensional, numerical flow model was constructed to simulate steady-state flow in an ideal valley-fill aquifer (fig. 8). That aquifer includes a partially penetrating river and a tributary stream that loses water to the aquifer at a constant rate within the surrounding bedrock valley walls. The area of diversion, contributing area, and time-of-travel areas (30 days, 100 days, 1 year, and 2 years) were delineated by use of a numerical model and particle-tracking program (McDonald and Harbaugh, 1988; Pollock, 1989) for a well being pumped at 100 and 500 gallons per minute. Because this method represents the most rigorous delineation technique, it is used as the standard against which delineations by the fixed-radius, uniform-flow, analytical, and semianalytical methods are compared.

The following parameters were used for delineations made by the fixed-radius, uniform-flow, analytical, and semianalytical methods: (1) horizontal hydraulic conductivity, 50 feet per day; (2) porosity, 0.20; (3) saturated thickness, 80 feet; (4) water-table slope, 0.005; and (5) distance from well to river, 220 feet. The hydraulic conductivity, porosity, and distance to the river are the same as those used in the three-dimensional model. The water-table slope and the saturated thickness were estimated from the



EXPLANATION

CELLS REPRESENTING CLAY, HORIZONTAL HYDRAULIC CONDUCTIVITY IS 0.02 FOOT PER DAY

CELLS REPRESENTING SAND, HORIZONTAL HYDRAULIC CONDUCTIVITY IS 20 FEET PER DAY

NOTE: Model recharge is from uniform infiltration of 20 inches of precipitation per year. All discharge is to the fully penetrating river in column 1 and to the well pumped at 200 gallons per minute at row 63, column 9.

0 1,000 3,000 FEET

0 300 600 900 METERS

Figure 36. Effect of small-scale heterogeneities in hydraulic conductivity on the two-dimensional numerical simulation of 100-day time-of-travel area and area of diversion.

prepumping potentiometric surface (fig. 8). The potentiometric surface is not uniform. It slopes toward the river from both sides and away from the valley walls and tributary stream as shown in figure 8. The approximate average slope is about 0.005 on the side of the river where the well is located. The saturated thickness is variable but is about 80 feet near the pumped well.

Fixed-radius method

Time-of-travel areas for the given pumping rates (100 and 500 gallons per minute) were delineated by the use of the fixed-radius method. The areas were compared with those delineated with the three-dimensional model in figure 37. In general, the fixed-radius method best approximates time-of-travel areas for large pumping rates and small traveltimes as indicated by the similarity of the 30- and 100-day time-of-travel area for a pumping rate of 500 gallons per minute. At a rate of 100 gallons per minute, pumping does not create a strong radial-flow pattern relative to the natural flow field; consequently, the shapes of all time-of-travel areas are elongated. For longer traveltimes, the main stream, valley walls, and tributary stream affect the shapes of the time-of-travel areas for both pumping rates.

The poor agreement between fixed-radius and three-dimensional simulation methods could have been predicted before the analysis was done. To determine if the fixed-radius method will be a poor delineation technique, one must first estimate the effect of the sloping potentiometric surface by use of the graph in figure 18. In this case, the dimensionless parameter $[2\pi(Ki)^2bt]/(\theta Q)$ was approximately 0.1 or less for only the 30-day and the 100-day simulations at 500 gallons per minute; thus, the potentiometric-surface slope should significantly distort all the time-of-travel areas but these two.

The effect of the nearby river on time-of-travel areas computed by use of the fixed-radius method also can be roughly estimated by inspection of the graph and diagrams in figure 19. The dimensionless parameter $Qt/(2\pi b\theta d^2)$ was greater than 0.3 for all but the 30-day simulation at 100 gallons per minute. Therefore, the fully-penetrating stream could affect all of the other delineations significantly. In this example, however, the stream was partly penetrating, so the effect of the stream was less than that predicted from figure 19.

Uniform-flow method

Areas of diversion delineated by use of the uniform-flow method and the three-dimensional model are shown in figure 38. At a pumping rate of 100 gallons per minute, the shapes of the areas are similar because flow was not induced from or beneath the stream. Differences near the tributary stream are caused by the nonuniform slope of the water table in that area. At a pumping rate of 500 gallons per minute, however, the shapes of the areas differ considerably because boundaries (especially the partially penetrating river) are important at this larger pumping rate and cannot be simulated by use of the uniform-flow method. Because the well captures flow from nearly opposite directions, the assumption of a single, uniform flow direction is violated.

The possible effect of a fully penetrating river on the area of diversion computed by use of the uniform-flow method can be estimated in advance from inspection of figure 24. Because the river in the idealized aquifer is only partly penetrating, its effect will be somewhat less than that indicated by the graph. At a pumping rate of 100 gallons per minute, the dimensionless parameter ($Q/\pi dKbi$ in fig. 24) of 1.4 indicates that less than 10 percent of the water is induced from the stream. Because very little water is induced from the stream, the asymptotic limit of the area of diversion boundary should be only slightly affected by the stream. The effect of the stream could be great enough (dimensionless parameter greater than 0.1), however, to affect the shape of the area of diversion near the well. At a pumping rate of 500 gallons per minute, the dimensionless parameter in figure 24 is 7.0. Consequently, even though not fully penetrating, the stream is likely a significant source of water to the well. Therefore, the area of diversion delineated by the uniform-flow method is too large because it does not account for flow from or beneath the river.

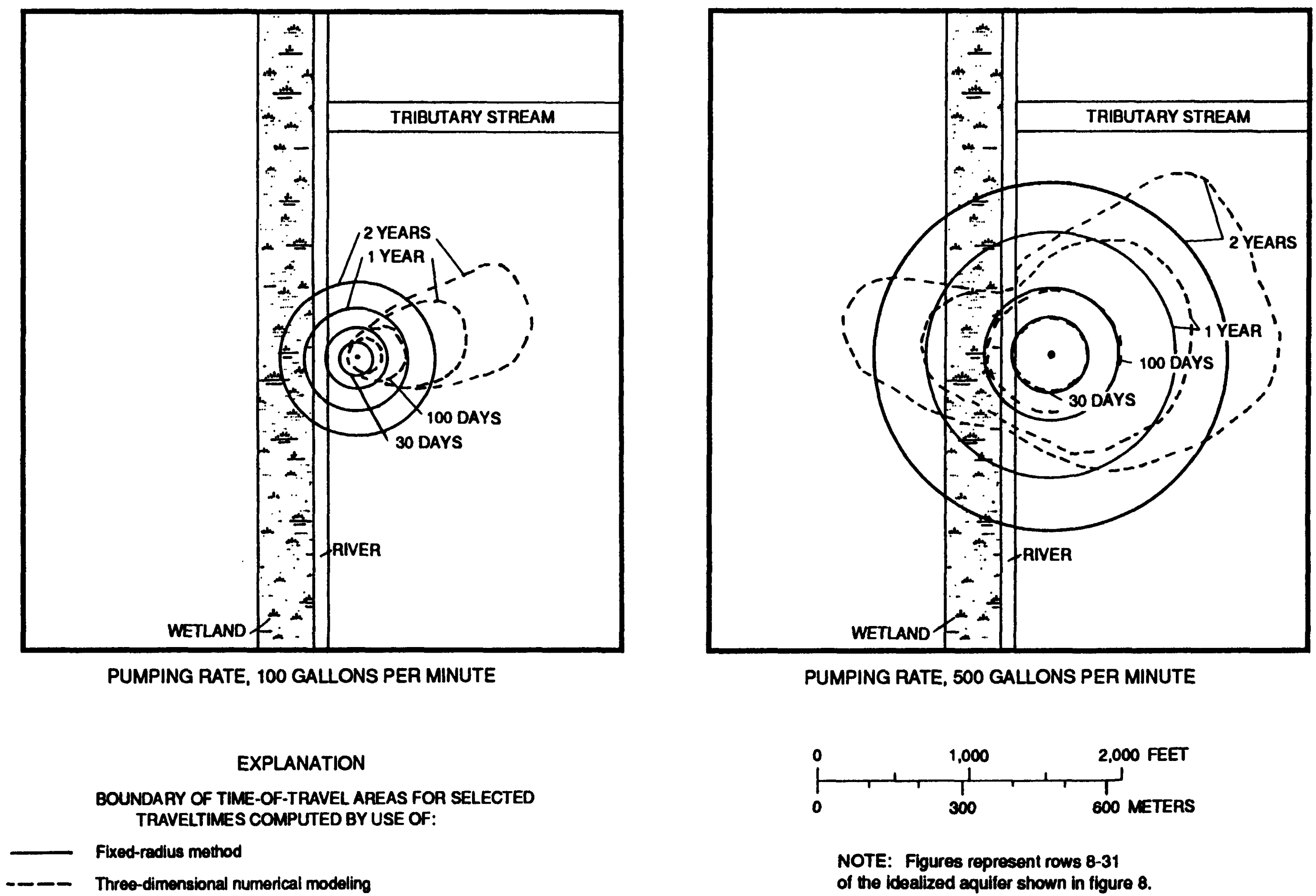


Figure 37. Time-of-travel areas in an idealized aquifer computed by three-dimensional numerical modeling and by the fixed-radius method for selected pumping rates.

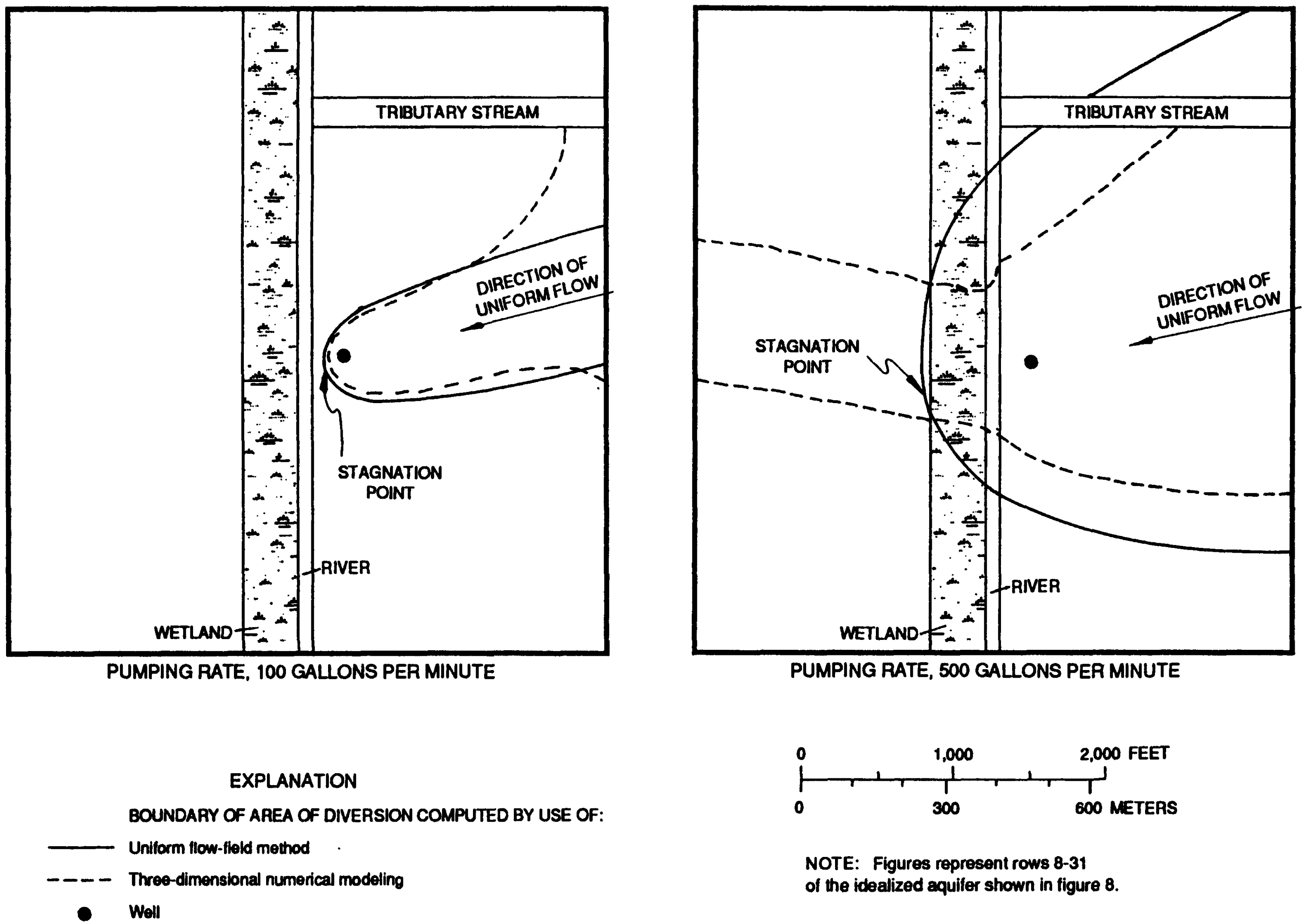


Figure 38. Area of diversion in an idealized aquifer computed by three-dimensional numerical modeling and by the uniform-flow method for selected pumping rates.

Analytical method

The area of diversion delineated by use of the image-well method and that delineated by three-dimensional modeling are shown in figure 39. The image-well method includes the effects of a nonuniform water-table surface and boundary conditions; however, the river and bounding valley walls are treated as linear and fully penetrating.

.Shapes of the areas of diversion are similar at a pumping rate of 100 gallons per minute (fig. 39A). The nonuniform water-table slope near the tributary stream, which was not properly simulated with the uniform-flow and semianalytical methods, is taken into account when drawdowns are superposed on the prepumping water-table surface. Shapes of the areas differ considerably at a pumping rate of 500 gallons per minute (fig. 39B). As with the other methods, simulation of the river as fully penetrating precludes any contributions of water beyond that boundary, which, in this instance, is a significant amount of water.

Semianalytical method

The area of diversion and time-of-travel areas for specified traveltimes were delineated by use of the semianalytical method, and are shown with those from the three-dimensional model in figure 40. The semianalytical method is based on the assumption of a uniformly sloping water table, as is the uniform-flow method; therefore, where the water-table slope is not uniform near the tributary stream (fig. 40A) and across the river (fig. 39B), the computed areas of diversion differ.

The semianalytical method is more powerful than the uniform-flow method because boundaries can be simulated. By use of image wells to represent the river at a pumping rate of 500 gallons per minute, the areas of diversion and time-of-travel areas outlined on the well side of the river are similar in size and shape (fig. 40B); however, water contributed from the aquifer beneath the river is neglected because image wells represent a fully penetrating stream.

The contributing areas and time-of-travel areas are similar at a pumping rate of 100 gallons per minute (fig. 40A). In this instance, the effect of the river is slight, and the water table can be reasonably approximated as uniform; however, as traveltimes increase, time-of-travel areas will differ increasingly because ground water will flow through areas where the water-table slope is not uniform, and an assumption of the method will be violated.

Aquifer in Marsh Creek Valley near Asaph, Pennsylvania

Methods that can be used to delineate areas of diversion, contributing areas, and time-of-travel areas were compared for a well field in Tioga County near Asaph, Pa. (fig. 41). This well field has been studied by Williams (1991) as part of the USGS's Northeast Glacial Aquifers project (Lyford and others, 1984). The well field consists of three wells (used by the National Fisheries Research and Development Laboratory) that are completed in the glaciofluvial valley-fill deposits along Marsh Creek Valley. The locations of wells and the approximate prepumping water-table configuration are shown in figure 42. Delineation of contributing areas for wells at this site is difficult because the aquifer is bounded by irregularly shaped bedrock valley walls and the pumping rate and natural recharge vary seasonally.

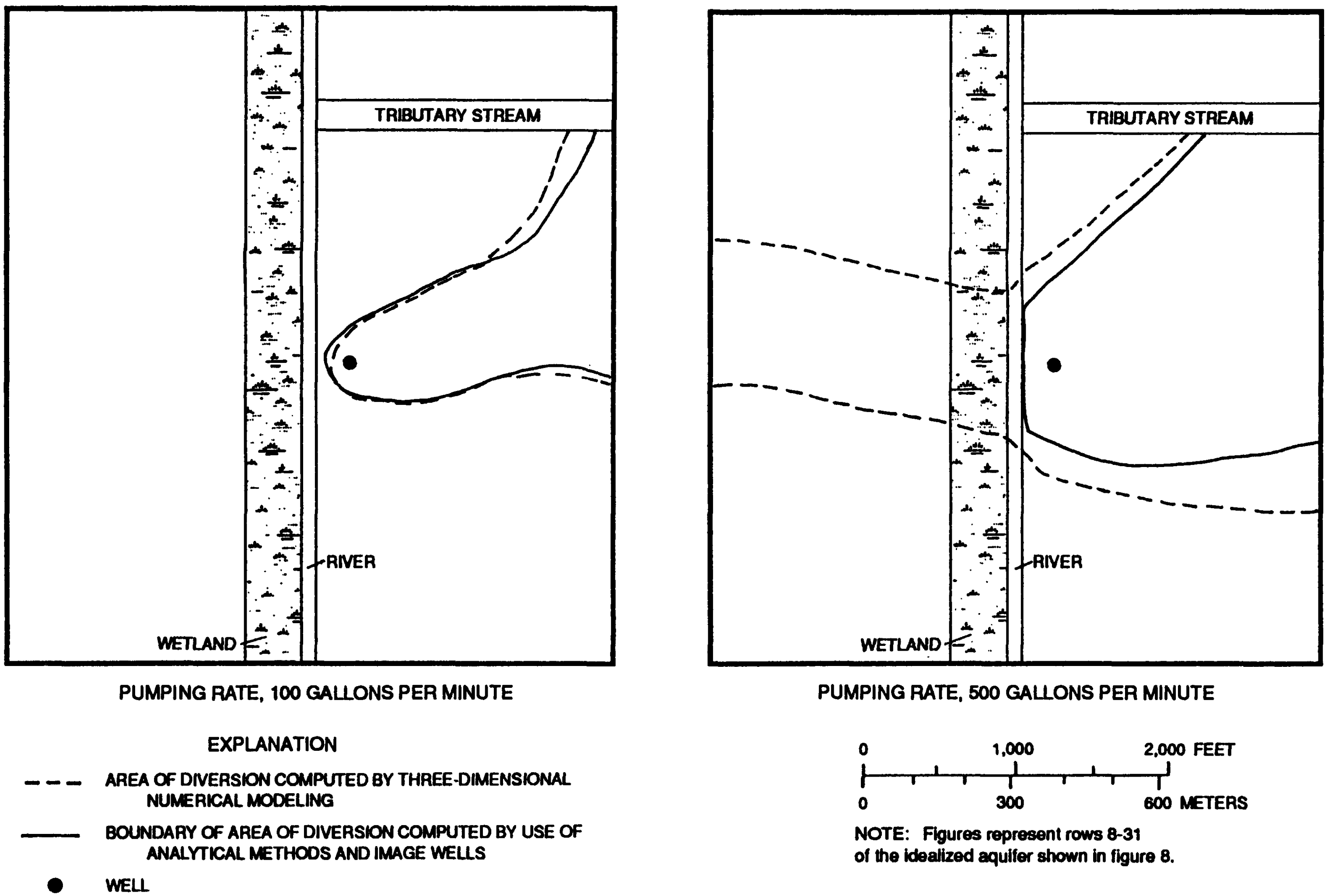
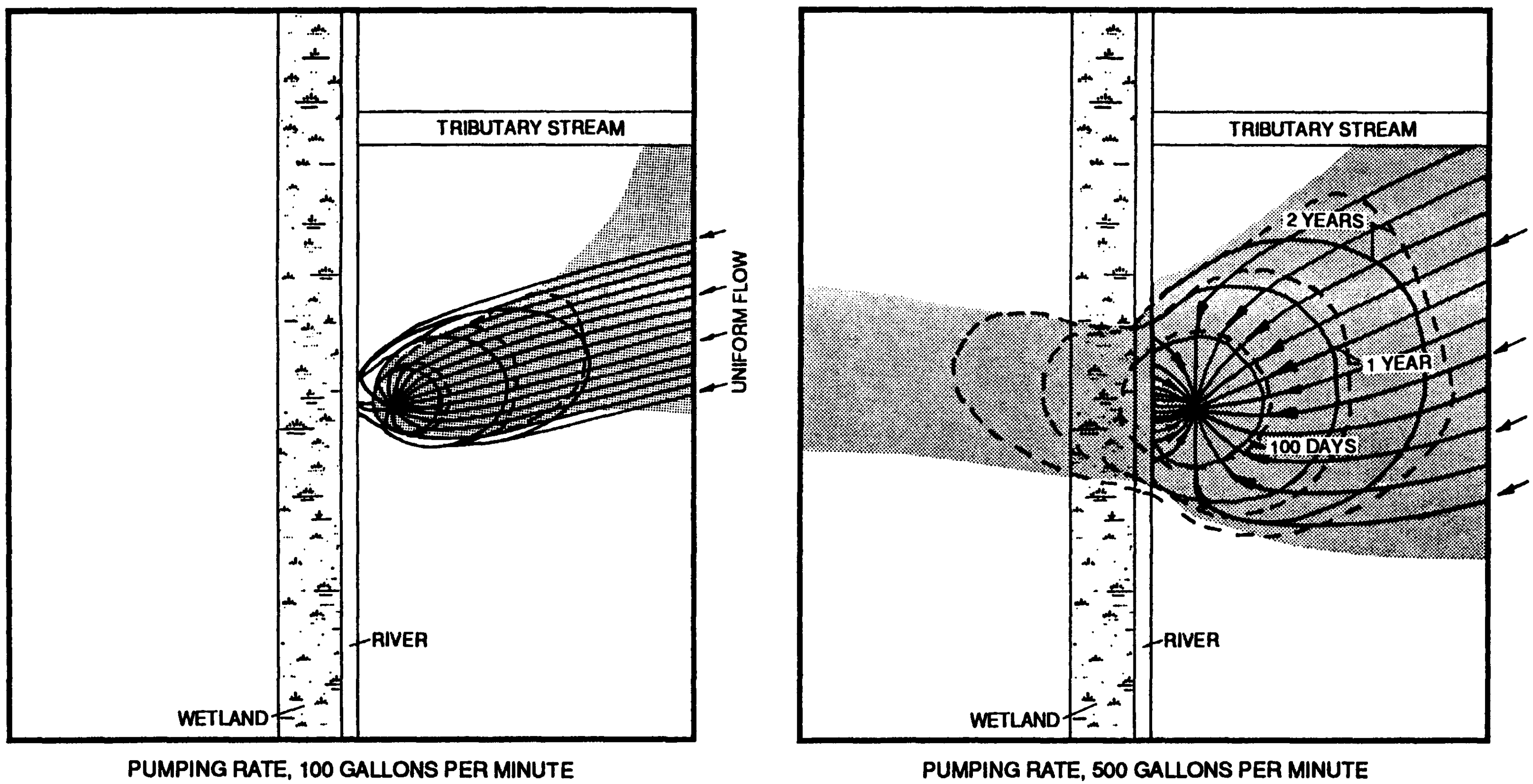


Figure 39. Areas of diversion in an idealized aquifer computed by three-dimensional numerical modeling and by analytical methods for pumping rates of 100 and 500 gallons per minute.



EXPLANATION

AREA OF DIVERSION COMPUTED BY THREE-DIMENSIONAL NUMERICAL MODELING

FLOW PATHS COMPUTED BY USE OF SEMIANALYTICAL METHODS

BOUNDARY OF TIME-OF-TRAVEL AREA FOR SELECTED TRAVELTIMES COMPUTED BY USE OF:
—— Semianalytical method
- - - Three-dimensional numerical modeling

0 1,000 2,000 FEET
0 300 600 METERS

NOTE: Figures represent rows 8-31 of the idealized aquifer shown in figure 8.

Figure 40. Areas of diversion and time-of-travel areas in an idealized aquifer computed by three-dimensional numerical modeling and by analytical methods for pumping rates of 100 and 500 gallons per minute.

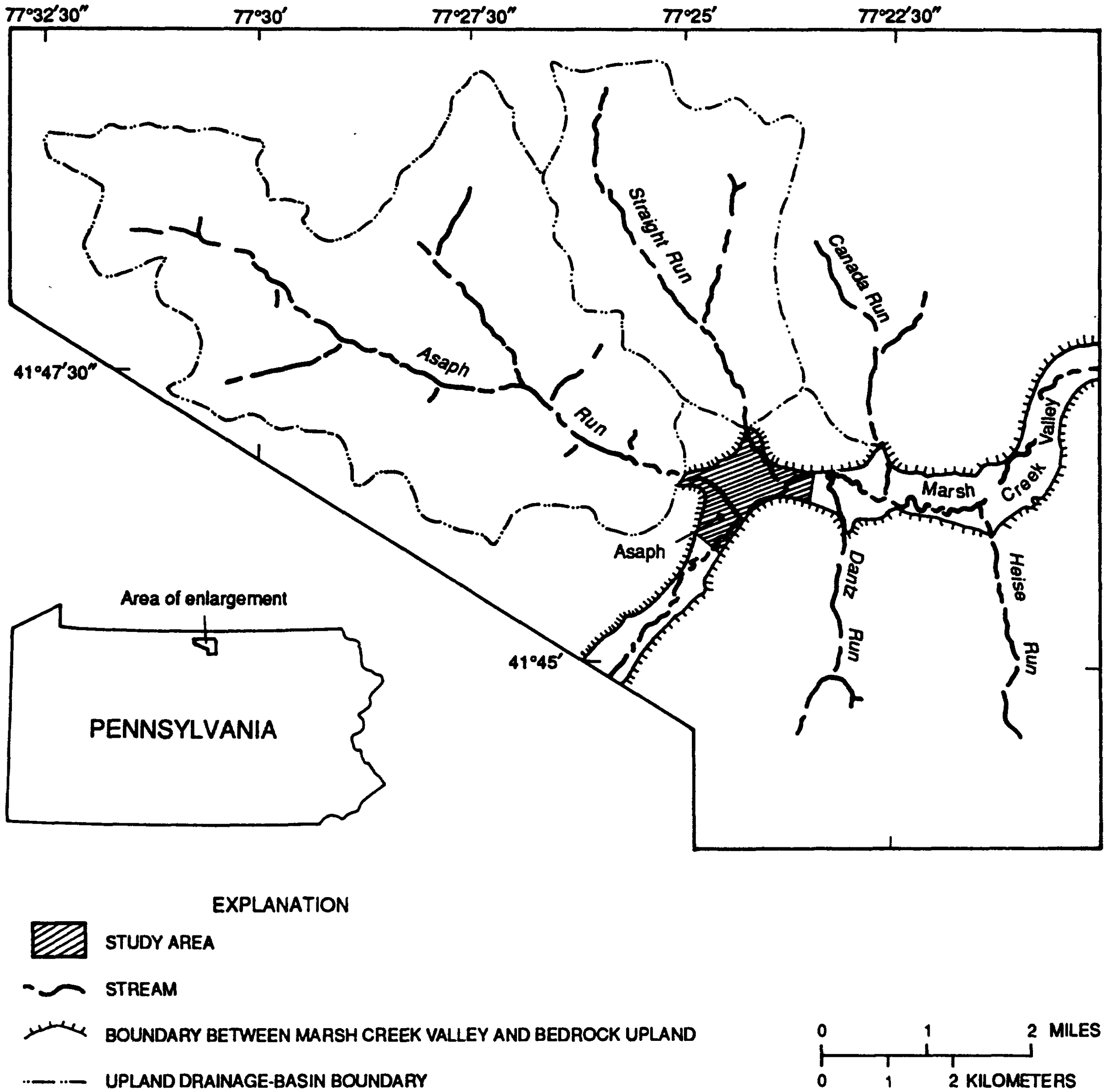


Figure 41. Location of Marsh Creek valley-fill aquifer near Asaph, Pennsylvania.

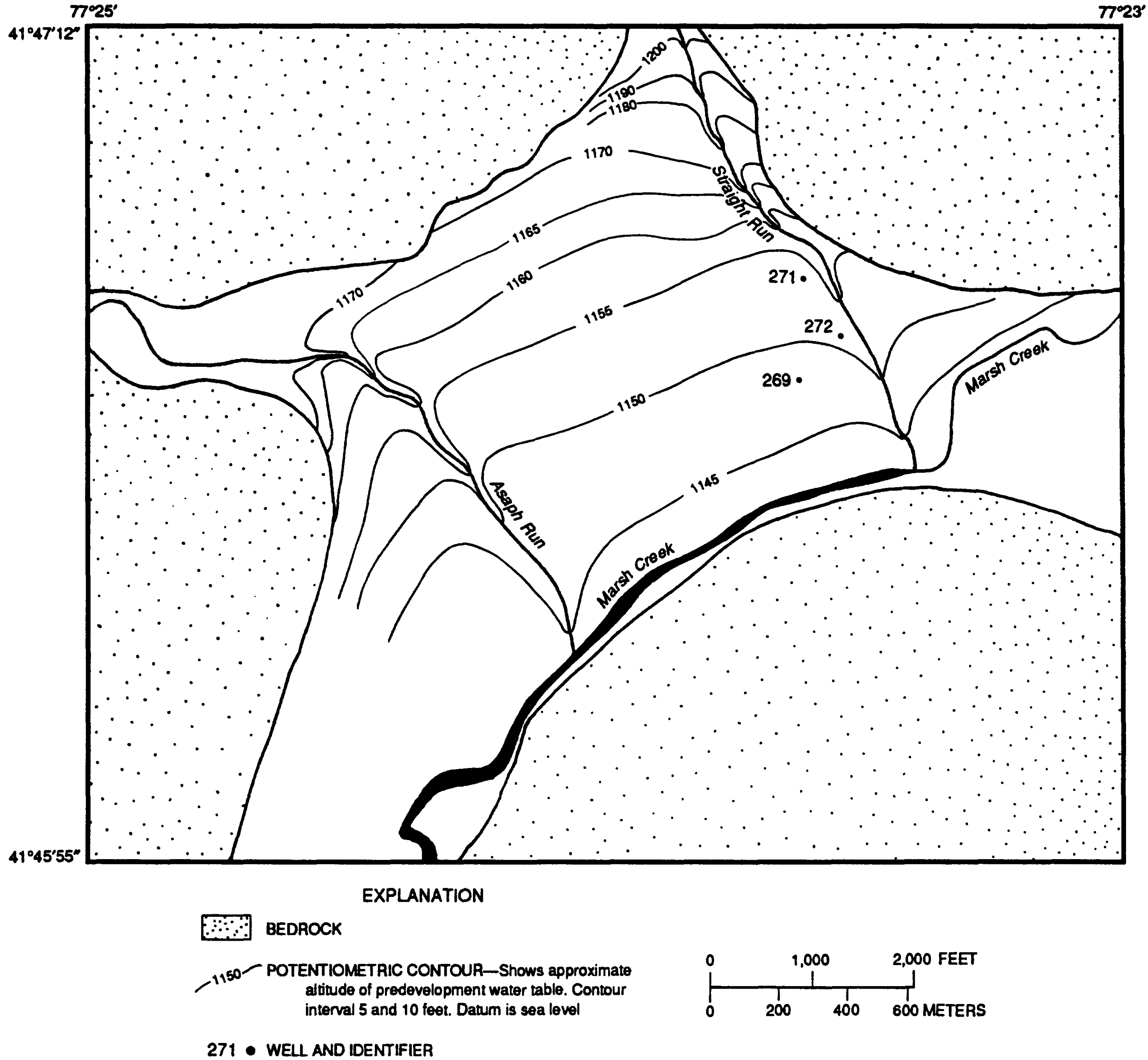


Figure 42. Approximate altitude of predevelopment potentiometric surface. (Modified from Williams, 1991, fig. 10.)

Contributing areas vary throughout the year depending on recharge and pumping rates. The approximate range in size of these areas was delineated by simulation of two extreme conditions: (1) wet periods from about December through May, when recharge is large and pumping is small; and (2) dry periods from June through November, when recharge is small and pumping is large. Water-budget terms for wet and dry periods that could be estimated from measurements by Williams (1991) are shown in the table that follows.

Source of recharge or discharge	Recharge or discharge (cubic feet per second)	
	Wet periods	Dry periods
Infiltration from Straight Run	4.2	2.0
Infiltration from Asaph Run	7.0	2.0
Precipitation on valley surface	1.2	.4
Unchanneled runoff and interflow from bedrock uplands	1.8	.6
Pumping from wells	1.5	3.2

The methods to delineate contributing areas are based on the assumption that steady state has been established in the aquifer during the wet and dry periods. In reality, wet and dry periods of approximately 6 months may not be long enough to establish a steady state. Therefore, the contributing areas delineated represent maximum and minimum positions that could be expected for wet and dry periods.

Areas of diversion, contributing areas, and time-of-travel areas for the Marsh Creek well field were delineated by use of the following methods: (1) numerical modeling; (2) fixed radius; (3) uniform flow; (4) semianalytical; and (5) analytical.

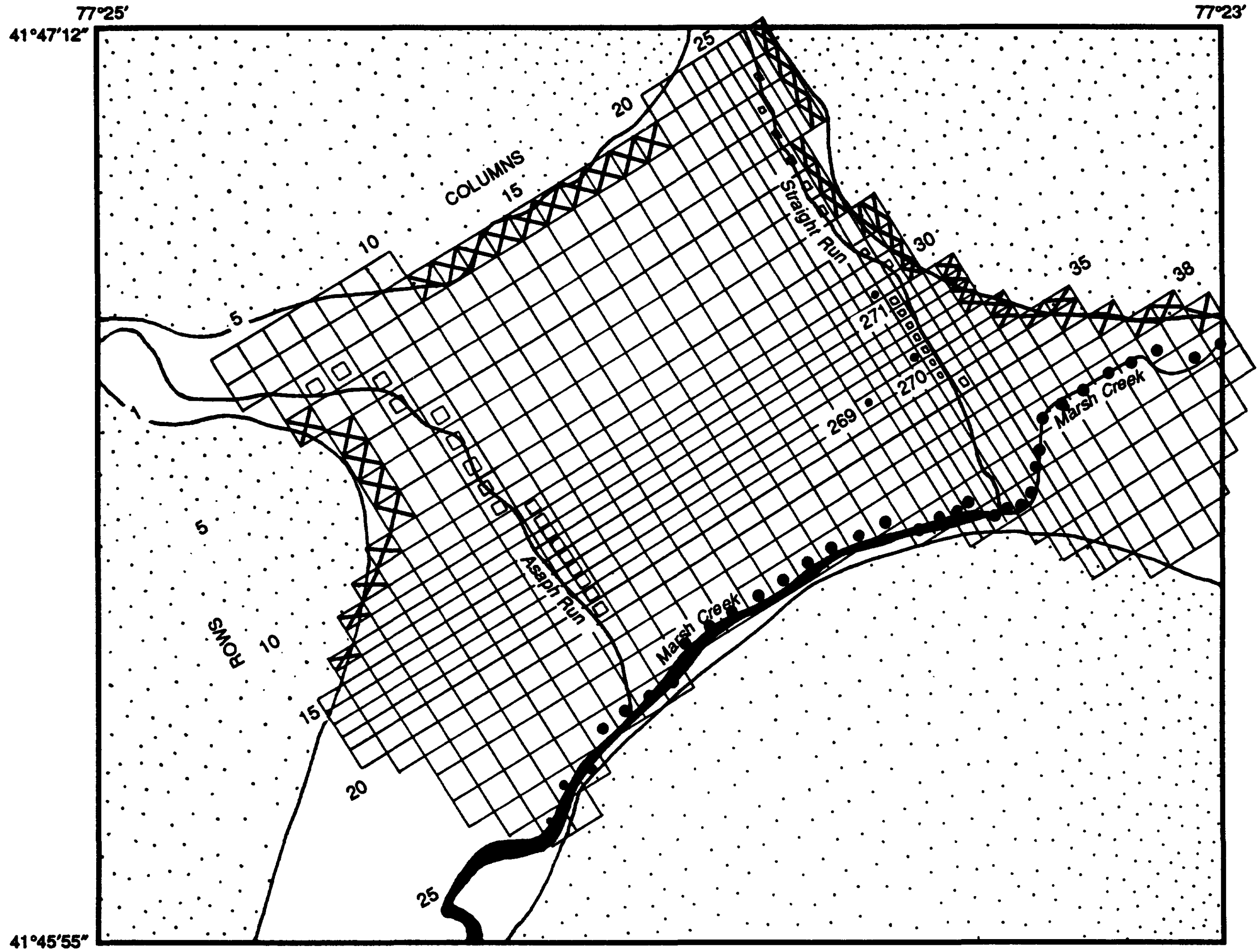
Numerical flow modeling

A two-dimensional, finite-difference numerical flow model (McDonald and Harbaugh, 1988) and particle tracker (Pollock, 1989) were used to delineate the areas of diversion and 100-day time-of-travel area. The finite-difference grid and boundary conditions used to construct the model are shown in figure 43. Horizontal hydraulic conductivities of 100 and 200 feet per day were assigned to the western and the eastern halves of the area, respectively. Recharge and pumping rates for wet and dry simulations are in the water budget shown above. All discharge is to the pumped wells and Marsh Creek.

Areas of diversion delineated by use of the numerical model are shown in figure 44. During wet periods, thin areas of diversion terminate at Straight Run from which 98 percent of the water is derived from infiltration. During dry periods, the areas of diversion extend nearly throughout the entire aquifer to make up the deficit of water from decreased recharge and increased pumpage. Straight Run, Asaph Run, Marsh Creek, and unchanneled upland runoff are the sources of 61, 4, 12, and 15 percent of the pumpage, respectively, during dry periods (table 2). During either wet or dry periods, precipitation on the valley surface contributes a maximum of 8 percent of the water pumped, thus, upland sources are always the source of at least 92 percent of the pumpage.

Contributing areas that provide each source of water also vary greatly between wet and dry periods (table 2). During wet periods, when most of the pumping is derived from Straight Run, the contributing area consists chiefly of its 7-square-mile watershed. During dry periods, when infiltration from Marsh Creek is induced, its 61-square-mile watershed becomes part of the contributing area. For successful wellhead-protection efforts in situations such as this, the possibility of contamination from distant source areas must be considered.

The 100-day time-of-travel areas for wet and dry periods also were delineated by use of the numerical model (fig. 45). During the wet season, the 100-day time-of-travel areas and the area of diversion are virtually the same (See figs. 44A and 45A).



EXPLANATION

BEDROCK

MODEL BOUNDARY CELLS

CONSTANT-FLOW—Represents recharge from unchanneled upland runoff and interflow

CONSTANT-FLOW—Represents recharge from Straight and Asaph Runs

RIVER—Represents Marsh Creek as a partially penetrating stream having a streambed 2 feet thick and a vertical hydraulic conductivity of 0.1 foot per day

271 ● WELL AND IDENTIFIER

0 1,000 2,000 FEET

0 200 400 600 METERS

Figure 43. Finite-difference grid and boundary conditions for the two-dimensional numerical model of a well field near Asaph, Pennsylvania.

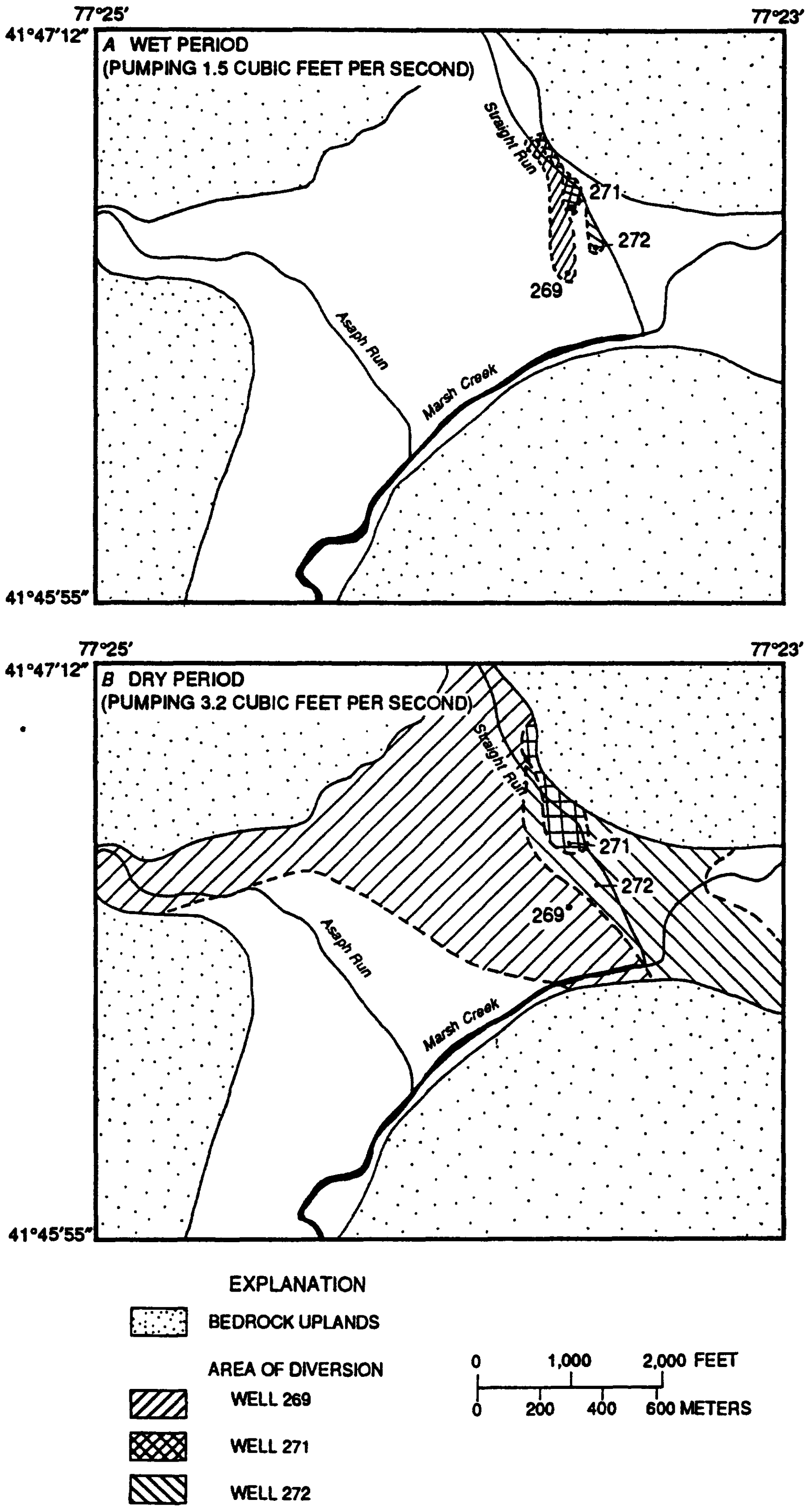


Figure 44. Areas of diversion during (A) wet and (B) dry periods for a well field near Asaph, Pennsylvania, delineated by use of a two-dimensional numerical model.

Table 2. Sources of water and contributing areas of wells in dry and wet seasons in Marsh Creek Valley simulated by use of a two-dimensional numerical model

Source of water in dry season	Percentage of pumpage from each source				Contributing area, in square miles			
	Well 269	Well 271	Well 272	All wells	Well 269	Well 271	Well 272	All wells
Precipitation on valley surface	12	1	9	8	0.03	0.01	0.02	0.06
Unchanneled runoff and interflow from bedrock uplands	20	7	18	15	1	.3	1	2.30
Infiltration from Straight Run	42	92	53	61	7	7	7	21
Infiltration from Asaph Run	11	0	0	4	16	0	0	16
Infiltration from Marsh Creek	15	0	20	12	61	0	61	62
Total	100	100	100	100	85.03	7.31	69.02	161.36
Source of water in wet season	**Percentage of pumpage from each source**				**Contributing area, in square miles**			
	Well 269	Well 271	Well 272	All wells	Well 269	Well 271	Well 272	All wells
Precipitation on valley surface	3	1	2	2	.6	.03	.21	.84
Unchanneled runoff and interflow from bedrock uplands	0	0	0	0	0	0	0	0
Infiltration from Straight Run	97	99	98	98	7	7	7	21
Infiltration from Asaph Run	0	0	0	0	0	0	0	0
Infiltration from Marsh Creek	0	0	0	0	0	0	0	0
Total	100	100	100	100	7.6	7.03	7.21	21.84

Fixed-radius and semianalytical methods

The 100-day time-of-travel area delineated by use of the fixed-radius and semianalytical methods was compared to that area delineated by numerical modeling (figs. 45 and 46). The assumptions inherent in the fixed-radius method make it poorly suited for application at this site. Neither nearby boundaries, well interference, nor sloping water table can be simulated by use of this method.

A semianalytical method documented by Javandal and others (1984) was used to simulate Marsh Creek as a fully penetrating line-source stream boundary and the eastern valley wall as an impermeable boundary by use of image wells (fig. 46B). To simulate the wet and dry periods, the water-table gradient was varied from 0.013 to 0.007. This range approximates the variability in the gradient measured by Williams (1991). The magnitude of the uniform flow was computed from the water-table gradient, a hydraulic conductivity of 200 feet per day, aquifer thickness of 95 feet, and porosity of 0.20. The direction of flow is restricted in the model code and must be perpendicular to Marsh Creek. This restriction affects the time-of-travel areas, especially for wet periods when Straight Run provides large amounts of recharge. For dry periods, the 100-day time-of-travel area is similar to that delineated by numerical modeling (see figs. 45B and 46).

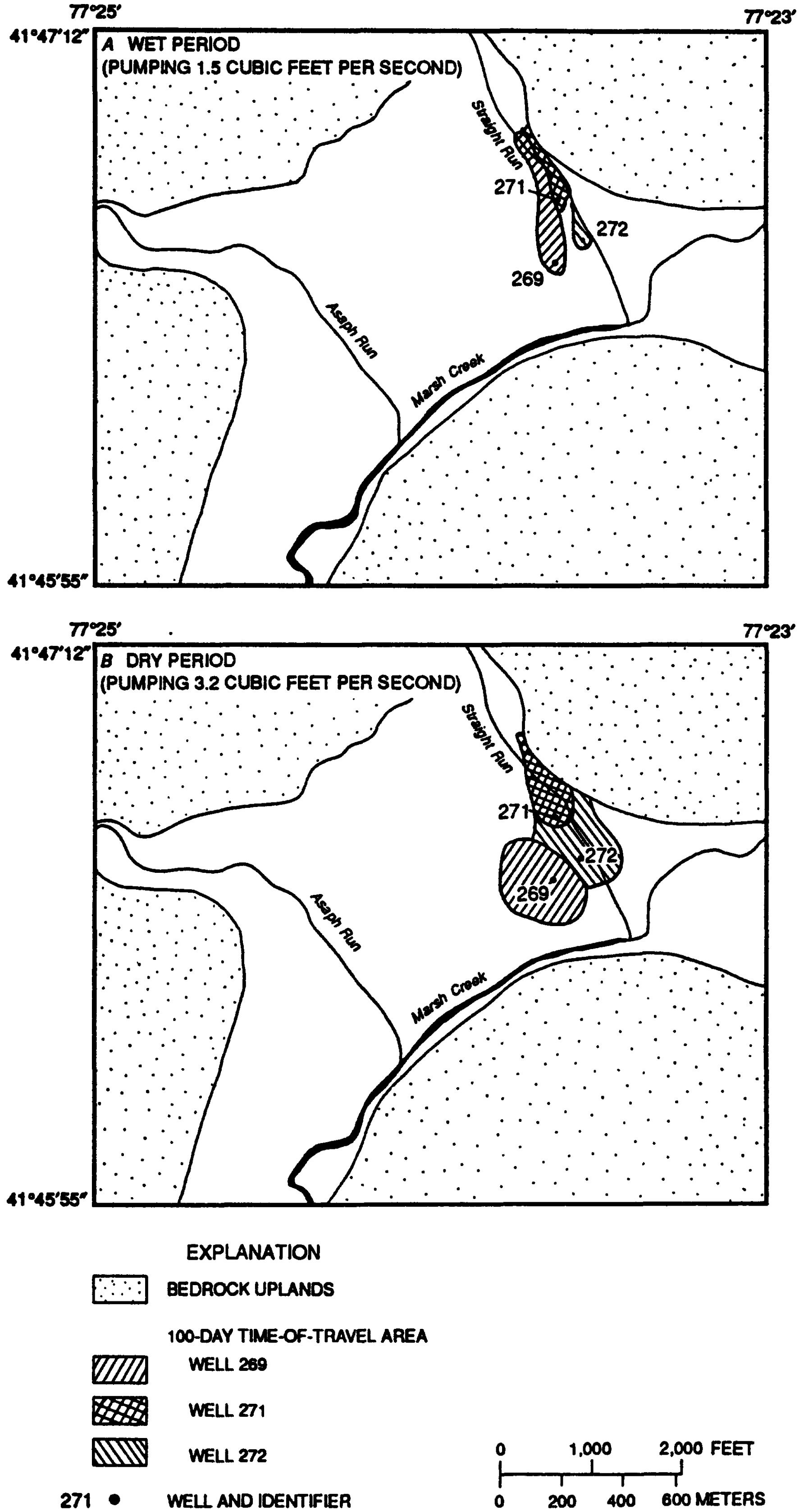


Figure 45. The 100-day time-of-travel areas during (A) wet and (B) dry periods for a well field near Asaph, Pennsylvania, delineated by use of a two-dimensional numerical model.

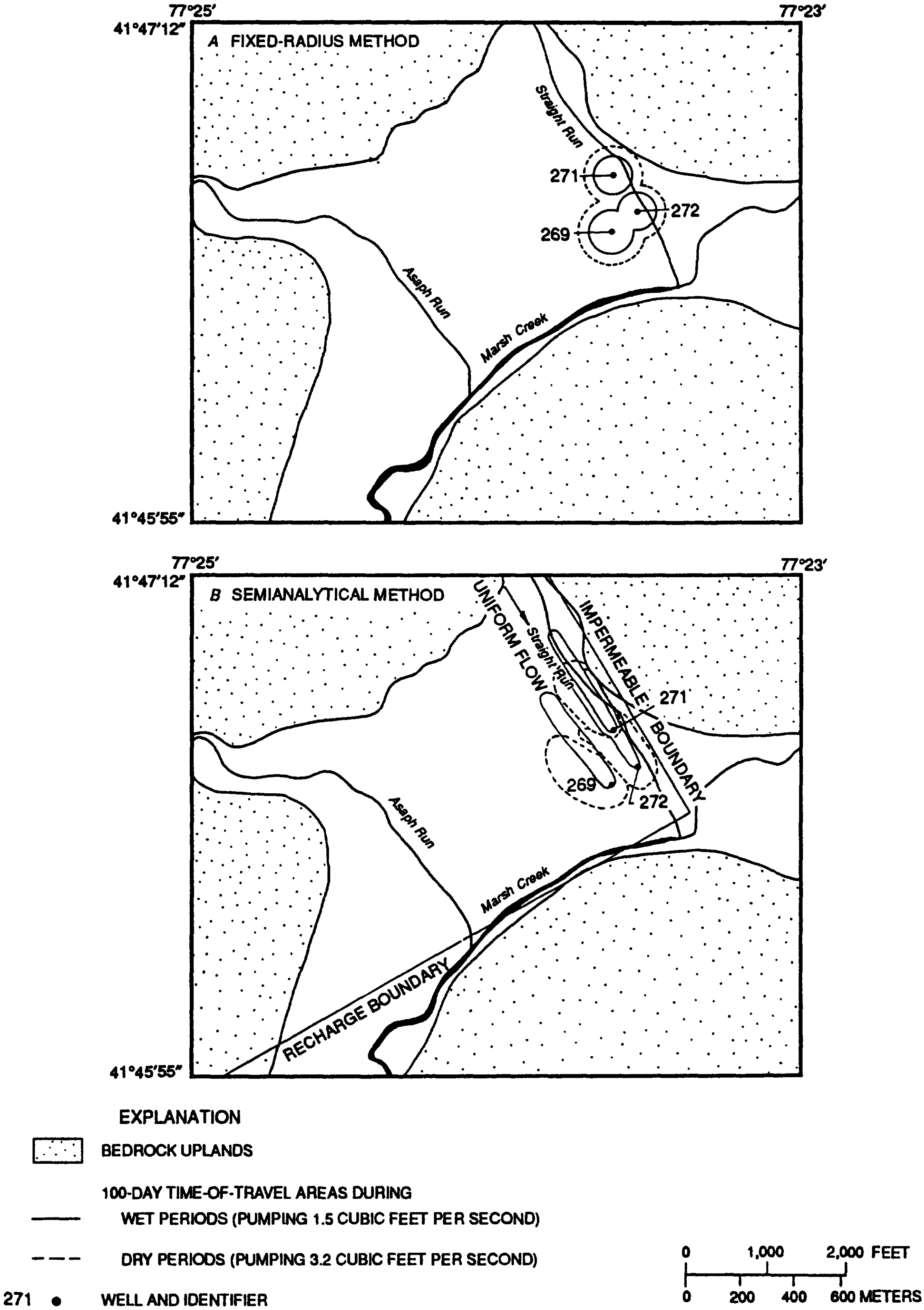


Figure 46. The 100-day time-of-travel areas during wet and dry periods for a well field near Asaph, Pennsylvania, delineated by use of (A) fixed-radius and (B) semianalytical methods.

Uniform-flow, analytical, and semianalytical methods

The uniform-flow, analytical, and semianalytical methods were used to delineate the area of diversion for wet and dry periods. The uniform-flow method does not account for interference of drawdown between wells, so the discharge from each well was summed to represent an equivalent single well (fig. 47A). The inability of this method to account for boundaries limits its usefulness for this site, as illustrated by the large difference between areas of diversion delineated by use of this method and those delineated by use of the numerical model. (Compare figs. 47A and 44.).

The image-well analytical method allowed flexibility in simulation of boundaries and water-table slope. For the wet period, Marsh Creek and Straight Run were simulated as line-source stream boundaries. For the dry period, when streamflow of Straight Run is at or near the annual minimum, the eastern valley wall was simulated as an impermeable boundary. The wet-period areas of diversion are similar to those delineated by use of the numerical model, but the dry-period areas are very different. (Compare figs. 47B and 44.) The differences are probably caused by numerical-model simulation of Marsh Creek as a partially penetrating stream whose bottom sediments are about 1,000 times less permeable than the aquifer sediments. The semianalytical method (fig. 48) allowed simulation of simple boundary conditions and well-interference effects. These areas of diversion differ from the results of the numerical model because boundaries and the water-table configuration could not be accurately simulated. (Compare figs. 48 and 44.)

These simulations indicate that delineation of contributing areas for a real well field is difficult. Because pumping rates and natural recharge vary over a large range, even results of numerical models must be viewed as estimates. The analytical methods were easier to apply than were the other methods but are the least accurate because of their inability to simulate the real aquifer boundaries.

GUIDELINES FOR SELECTION OF METHODS

Selection of the most appropriate method to delineate a contributing area depends on technical considerations such as the data requirements and assumptions inherent in the method and other factors such as cost, time, and computer availability. Methods that are easy to use commonly are restricted by required assumptions about the flow system that reduce their accuracy. Methods that are difficult to use (requiring special training and computer resources) are more flexible and less restricted by assumptions about the flow system than methods that are easy to use. The major assumptions required in the methods previously discussed in this report are summarized in table 3. These assumptions, along with some considerations of the effort required to use each method, are compiled in a flowchart to guide method selection (fig. 49).

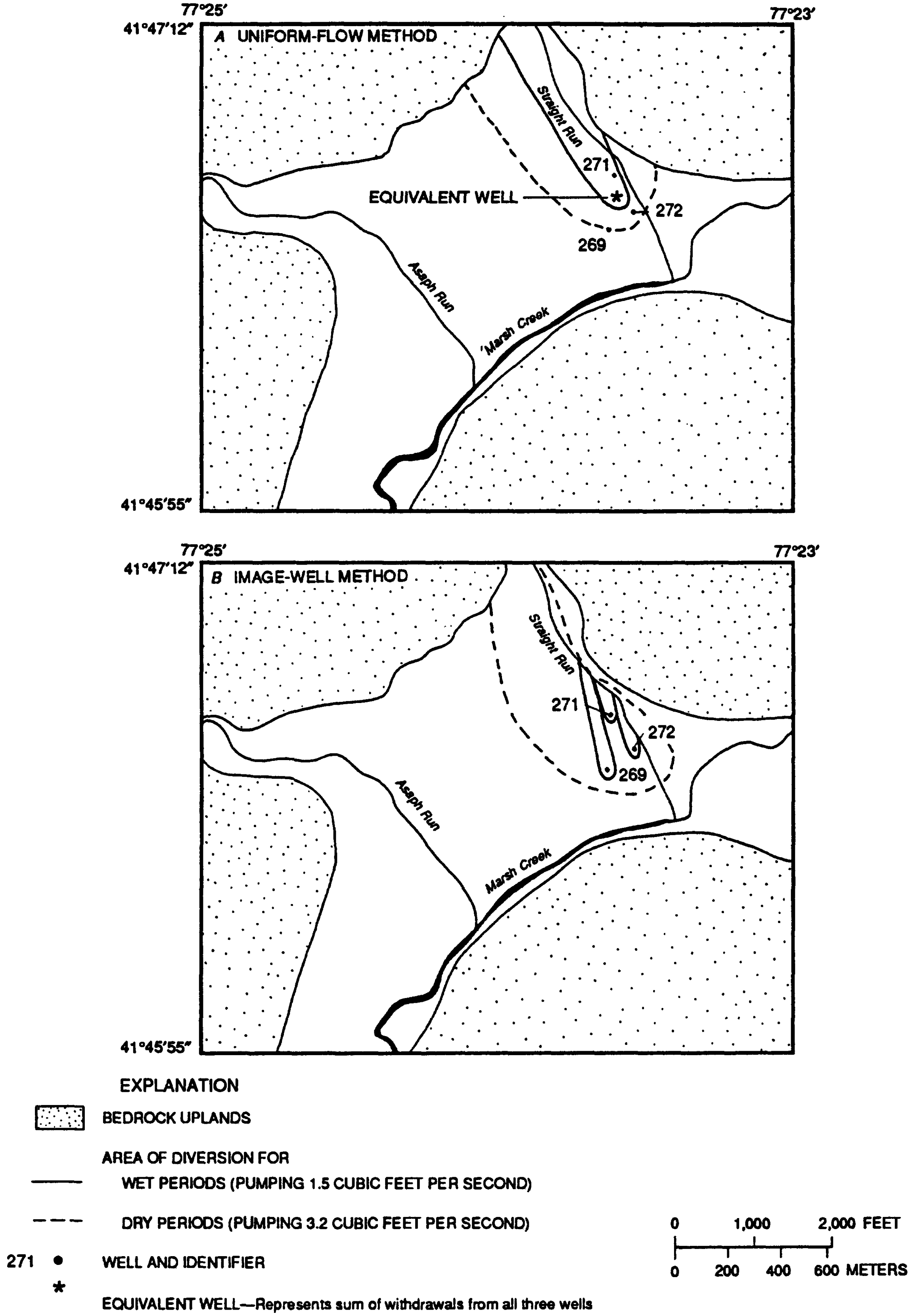


Figure 47. Areas of diversion during wet and dry periods for a well field near Asaph, Pennsylvania, delineated by use of (A) uniform-flow and (B) image-well methods.

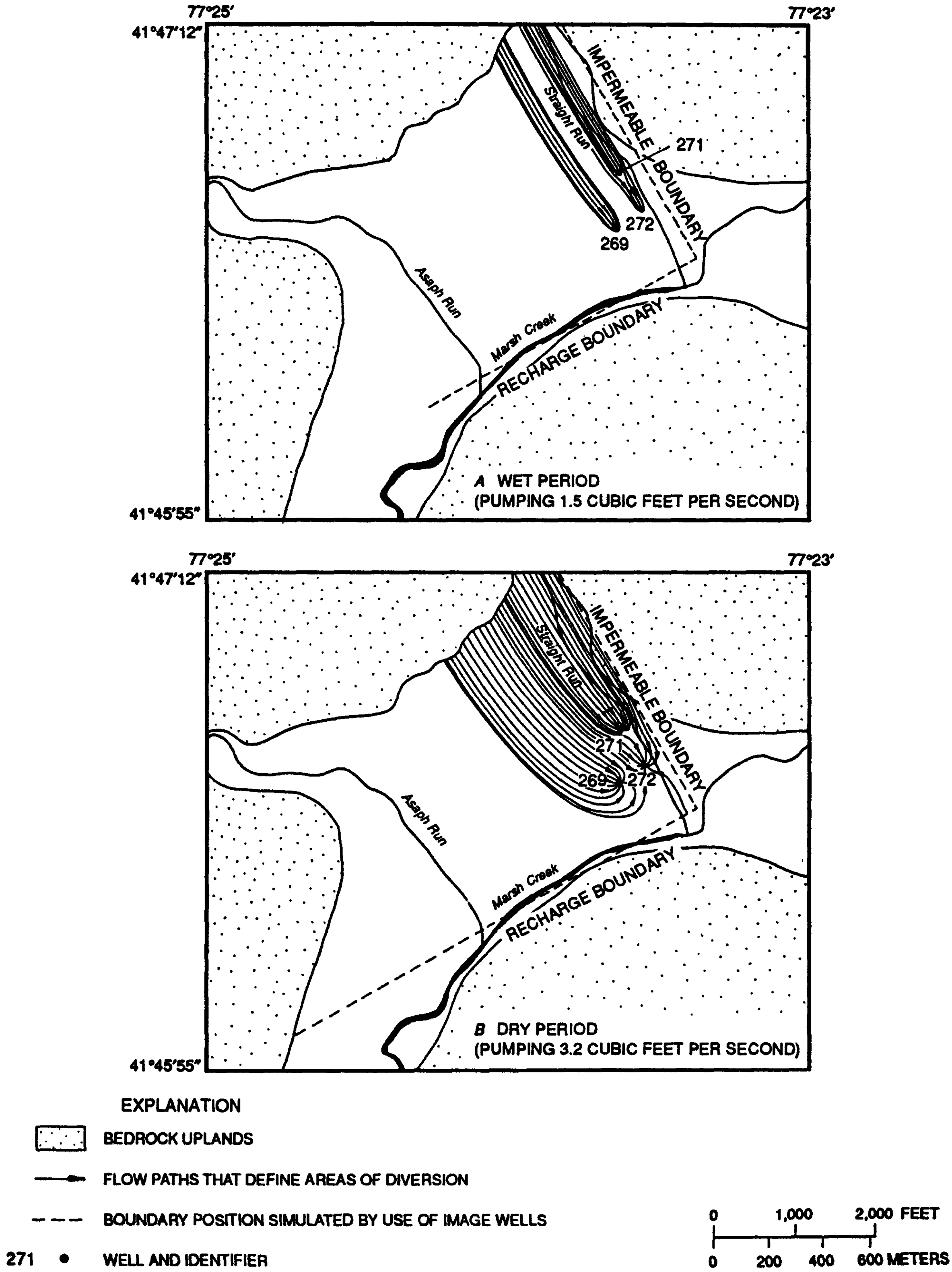


Figure 48. Areas of diversion during (A) wet and (B) dry periods for a well field near Asaph, Pennsylvania, delineated by use of a semianalytical method.

Table 3. Major assumptions inherent in selected methods for delineation of contributing area

[2-D, two-dimensional; 3-D, three-dimensional]

Assumptions about these hydrologic factors	Methods					
	Fixed radius	Uniform flow	Analytical	Two-dimensional semianalytical	Two-dimensional numerical flow modeling	Three-dimensional numerical flow modeling
Aquifer type	Confined [1] 2-D	Confined[1] 2-D	Confined[1] 2-D	Confined[1] 2-D	Confined or unconfined 2-D	Confined or unconfined 2-D
Thickness	Uniform	Uniform	Uniform	Uniform	Variable	Variable
Potentiometric surface	Flat [2]	Uniform[2]	Measured	Uniform	Measured	Measured
Aquifer properties	Homogeneous and isotropic horizontally vertically	Homogeneous and isotropic horizontally and vertically	Homogeneous and isotropic horizontally and vertically [3]	Homogeneous and isotropic horizontally and vertically	Heterogeneous and anisotropic horizontally	Heterogeneous and anisotropic vertically and horizontally
Boundary conditions	None[4]	None[4]	Linear and fully penetrating	Linear and fully penetrating	Irregular geometry, partially penetrating	Irregular geometry, partially penetrating
Recharge	None or uniform	None	None or uniform	None	Variable	Variable
Well characteristics	Fully penetrating single well	Fully penetrating single well	Fully or partially penetrating single well	Fully penetrating single or multiple wells	Fully penetrating single or multiple wells	Fully or partially penetrating single or multiple wells
Type of area delineated	Time-of-travel area	Time-of-travel area or area of diversion	Area of diversion only	Time-of-travel area or area of diversion	Time-of-travel area or area of diversion	Time-of-travel area or area of diversion

[1]Unconfined aquifer can be simulated if drawdown is less than 10 percent of saturated thickness.
[2]Can estimate effect of uniform slope from figure 17.
[3]Depends on analytical equation used.
[4]Effect of nearby recharge boundary illustrated in figure 18 or 23.

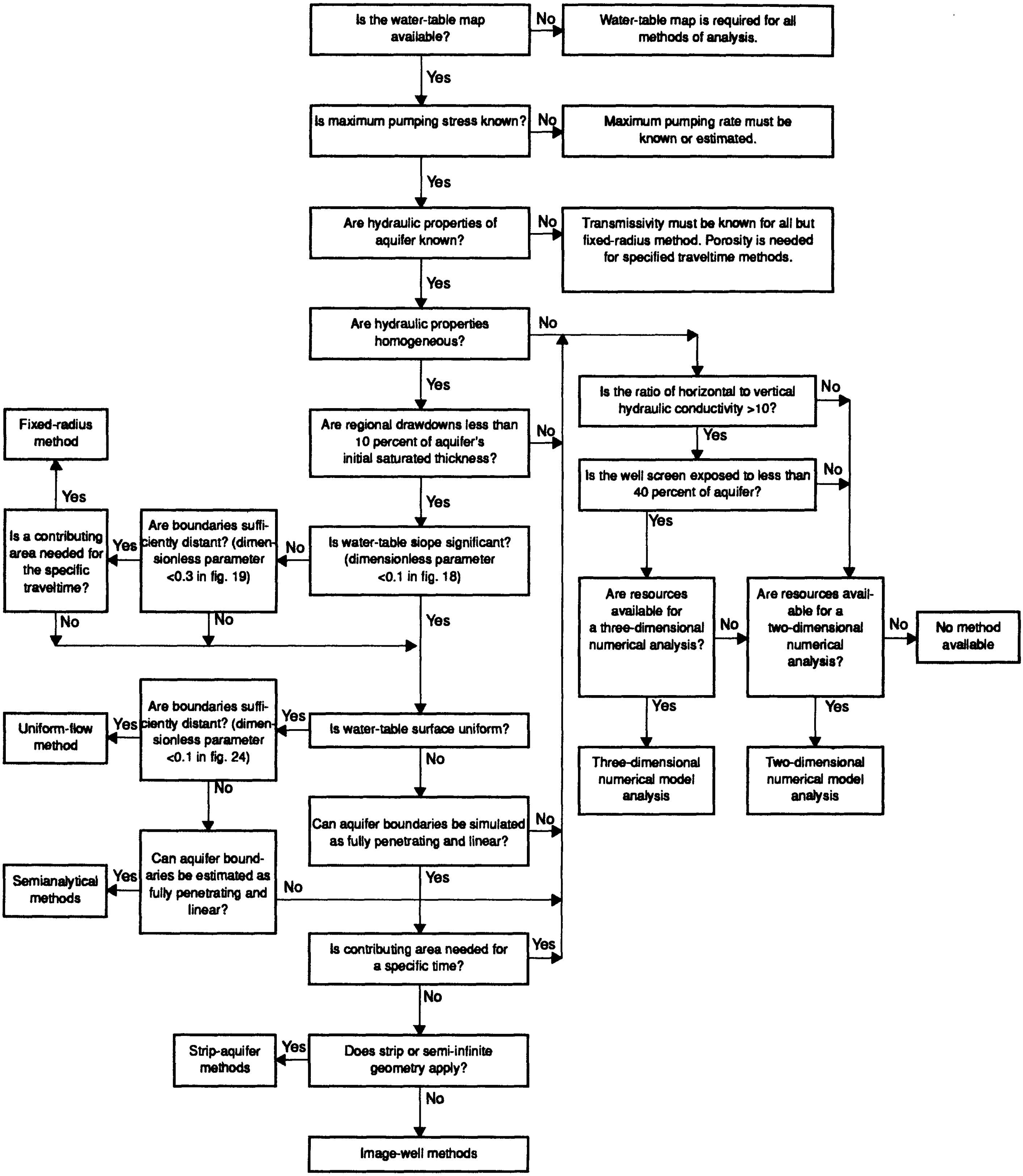


Figure 49. Guidelines for selection of methods for delineation of contributing area.

SUMMARY AND CONCLUSIONS

Unconsolidated aquifers are important as sources of water for public supply throughout the glaciated one-third of Pennsylvania and along major river valleys in the Ohio, Susquehanna, and Delaware River Basins. Protection of wells completed in these permeable, shallow aquifers involves identification of the area throughout which contaminants can move to the wells.

In attempts to delineate contributing areas, one can confuse area of influence of a well with the contributing area of the well. The area of influence of a pumped well is the projection to land surface of the extent of drawdown caused by the well. The contributing area is only that part of the aquifer and adjacent surface areas that provide water to the well. Attempts to delineate contributing areas by use of the area of influence (drawdown criteria) can result in considerable error.

The major sources of water to wells in unconsolidated valley-fill aquifers are (1) precipitation on the aquifer surface, (2) seepage from tributary streams that drain bedrock upland areas, (3) unchanneled runoff and interflow from bedrock uplands, and (4) infiltration induced from streams that drain the valley-fill aquifer. The upland bedrock typically is the source of 60 to 75 percent of recharge to valley-fill aquifers in areas of high topographic relief. Therefore, water pumped from valley-fill aquifers likely will be largely from upland bedrock sources. Delineation of contributing areas for a well field in the glaciofluvial sediments along Marsh Creek in Tioga County indicates that 92 to 98 percent of the pumpage is from upland sources, depending on the pumping rate and the season.

Methods that can be used to delineate contributing areas in unconsolidated aquifers include fixed radius, uniform flow, analytical, semianalytical, and numerical modeling. However, these methods actually do not identify the entire contributing area to a well because they do not explicitly delineate upland bedrock areas that contributed water. The methods (except for certain applications of numerical models) delineate an area of diversion or time-of-travel area on the surface of the valley-fill aquifer. The area of diversion is a projection to land surface of the aquifer volume through which water is diverted to the well. The time-of-travel area is a fraction of that area of diversion through which water is transported to the well in a specified time. Usually the contributing area can be estimated by sketching the upland bedrock areas that contribute water and adding them to the area of diversion delineated on the valley-fill aquifer.

Except for numerical modeling, the methods are based on the assumption of steady-state flow in the aquifer. In reality, recharge and pumping rates in valley-fill aquifers vary considerably throughout the year; thus, contributing areas delineated by use of average rates represent an average position about which the actual contributing area will fluctuate.

Use of the fixed-radius, uniform-flow, analytical, and semianalytical methods is generally restricted to two-dimensional, steady-state flow in aquifers that are (1) homogeneous, (2) confined (or unconfined if drawdown is small compared to the aquifer's saturated thickness), and (3) situated so that ground-water withdrawals are unaffected by nearby boundaries or so that nearby boundaries can be simulated as fully penetrating and linear. Because most unconsolidated aquifers in Pennsylvania consist of a complex assemblage of heterogeneous sediments, bounded by irregularly shaped valley walls and overlain by partly penetrating streams, the assumptions inherent in these methods must be carefully considered in their use.

Numerical flow modeling is by far the most flexible and powerful method to simulate the factors in a hydrologic system that influence the contributing area. Numerical modeling coupled with a particle-tracking program is considered the most rigorous method for delineating areas of diversion, contribution, and time of travel in most instances; however, use of this method requires an experienced hydrologist and significantly more effort than do the fixed-radius, uniform-flow, analytical, and semianalytical methods. As with the other methods, numerical flow modeling requires simplification of the aquifer system; therefore, even areas delineated by use of this method are approximations.

REFERENCES CITED

Anderson, M.P., and Woessner, W.W., 1992, Applied ground water modeling: Simulation of flow and advective transport: San Diego, Calif., Academic Press, 381 p.

Barker, J.L., and Molina, Nicholas, 1988, Pennsylvania ground-water quality, *in* National Water Summary 1986: U.S. Geological Survey Water-Supply Paper 2325, p. 429-436.

Bear, Jacob, 1979, Hydraulics of groundwater: New York, McGraw-Hill, 569 p.

Bear, J., and Jacobs, M., 1965, On the movement of water bodies injected into aquifer: Journal of Hydrology, v. 3, no. 1, p. 37-57.

Blandford, T.N., and Huyakorn, P.S., 1989, An integrated semi-analytical model for the delineation of wellhead protection areas: U.S. Environmental Protection Agency contract 68-08-003, 10 chapters.

Bonn, B.A., and Rounds, S.A., 1989, DREAM--Analytical groundwater flow programs: Chelsea, Michigan, Lewis Publishers, 115 p.

Boulton, N.S., 1963, Analysis of data from non-equilibrium pumping tests allowing for delayed yield from storage: Institute Civil Engineers Proceedings, v. 26, p. 469-482.

Bradbury, K.R., and Rothschild, E.R., 1985, A computerized technique for estimating the hydraulic conductivity of aquifers from specific capacity data: Ground Water, v. 23, no. 2, p. 240-246.

Bredehoeft, J.D., Papadopulos, S.S., and Cooper, H.H., 1982, Groundwater--The water-budget myth, *in* National Research Council, Geophysics Study Committee, Scientific basis of water-resource management: Washington, D.C., National Academy Press, p. 51-57.

Bredehoeft, J.D., and Young, R.A., 1970, A temporal allocation of ground water--a simulation approach: Water Resources Research, v. 6, no. 1, p. 3-21.

Brown, R.H., 1963, The cone of depression and the area of diversion around a discharging well in an infinite strip aquifer subject to uniform recharge *in* Bentall, Ray, compiler, Shortcuts and special problems in aquifer tests: U.S. Geological Survey Water-Supply Paper 1545-C, p. C69-C85.

Camp, Dresser, and McKee, Inc., 1982, Wellfield travel time model for selected wellfields in Dade, Broward, and Palm Beach Counties, Florida: Ft. Lauderdale, Fla., Final Report 9243-110, 165 p.

Dacosta, J.A., and Bennett, R.R., 1960, The pattern of flow in the vicinity of a recharging and discharging pair of wells in an aquifer having areal parallel flow: International Association of Scientific Hydrology Publication 52, p. 524-536.

Davis, S.N., and DeWiest, R.J.M., 1966, Hydrogeology: New York, John Wiley and Sons, 463 p.

Ferris, J.G., Knowles, D.B., Brown, R.H., and Stallman, R.W., 1962, Theory of aquifer tests: U.S. Geological Survey Water-Supply Paper 1536-E, 174 p.

Gebert, W.A., Graczyk, D.J., and Krug, W.R., 1987, Average annual runoff in the United States, 1951-80: U.S. Geological Survey Hydrologic Investigations Atlas HA-710, scale, 1:7,500,000, 1 sheet.

Hantush, M.S., 1960, Modification of the theory of leaky aquifers: Journal of Geophysical Research, v. 65, no. 11, p. 3,713-3,725.

----1961, Drawdowns around a partially penetrating well: American Society Civil Engineers Proceedings, Journal of the Hydraulics Division, v. 87, no. HY4, p. 83-98.

----1966, Wells in homogeneous anisotropic aquifers: Water Resources Research, v. 2, no. 2, p. 273-279.

Hantush, M.S., and Jacob, C.E., 1955, Non-steady radial flow in an infinite leaky aquifer: American Geophysical Union Transactions, v. 36, no. 1, p. 95-100.

Hewlett, J.D., and Nutter, W.L., 1970, The varying source area of streamflow from upland basin, *in* Interdisciplinary aspects of watershed management: Proceedings of American Society of Civil Engineers Symposium, Montana State University, Bozeman, p. 65-93.

Horsley, S.W., 1983, Delineating zones of contribution of public supply wells to protect ground water, *in* Proceedings of the National Water Well Association Eastern Regional Conference, Ground-Water Management, Orlando, Florida, p. 366-392.

Hydrologisch Colloquium, 1964, Steady flow of ground water towards wells: Committee for Hydrological Research, T.N.O., Publ. no. 10, 179 p., The Hague.

Jacob, C.E., 1950, Flow of ground water, *in* Engineering Hydraulics: edited by H. Rouse, John Wiley, chap. 5, p. 321-386.

Javandel, I., Doughty, C., and Tsang, C., 1984, Groundwater transport: Handbook of mathematical models: American Geophysical Union, Water Resources Monograph Series 10, 228 p.

Keely, J.F., and Tsang, C.F., 1983, Velocity plots and capture zone of pumping centers for ground-water investigations: Ground Water, v. 21, no. 6, p. 701-714.

Kirkham, D., 1949, Flow of ponded water into drain tubes in soil overlying an impervious layer: Transactions of the American Geophysical Union, v. 30, no. 3, p. 369-385.

Konikow, L.F., 1978, Calibration of ground-water models, *in* Verification of mathematical and physical models in hydraulic engineering: American Society of Civil Engineers Special Conference, College Park, Maryland, Aug. 9-11, 1978.

Kruseman, G.P., and de Ridder, N.A., 1976, Analysis and evaluation of pumping test data: Wageningen, The Netherlands, International Institute for Land Reclamation and Improvement, Bulletin 11, 200 p.

Lohman, S.W., 1972, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Lohman, S.W., and others, 1972, Definitions of selected ground-water terms-- revisions and conceptual refinements: U.S. Geological Survey Water-Supply Paper 1988, 21 p.

Lyford, F.P., and Cohen, A.J., 1988, Estimation of water available for recharge to sand and gravel aquifers in the glaciated Northeastern United States, *in* Randall, A.D., and Johnson, A.I., eds., Regional aquifer systems of the United States - the Northeast Glacial Aquifers: American Water Resources Association Monograph 11, p. 37-61.

Lyford, F.P., Dysart, J.E., Randall, A.D., and Kontis, A.L., 1984, Glacial aquifer systems in the northeastern United States--a study plan: U.S. Geological Survey Open-File Report 83-928, 33 p.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap. A1, 586 p.

Mercer, J.W., and Faust, C.R., 1980, Ground-water modeling: Mathematical models: Ground Water, v. 18, no. 3, p. 212-227.

Morrissey, D.J., 1989, Estimation of the recharge area contributing water to a pumped well in a glacial-drift, river-valley aquifer: U.S. Geological Survey Water-Supply Paper 2338, 41 p.

Morrissey, D.J., Randall, A.D., and Williams, J.H., 1988, Upland runoff as a major source of recharge to stratified drift in the glaciated Northeast, *in* Randall, A.D., and Johnson, A.I., eds., Regional aquifer systems of the United States--the Northeast Glacial Aquifers: American Water Resources Association Monograph 11, p. 17-36.

Muskat, M., 1937, The flow of homogeneous fluids through porous media: McGraw-Hill, 763 p.

Nelson, R.W., 1978, Evaluating the environmental consequences of groundwater contamination-- 4. Obtaining and utilizing contaminant arrival distributions in transient flow systems: Water Resources Research, v. 14, no. 3, p. 441-450.

Nelson, R.W., and Schur, J.A., 1980, PATHS groundwater hydrologic model: Richland, Wash., Battelle Pacific Northwest Laboratories, PNL-3162.

Neuman, S.P., 1974, Effect of partial penetration on flow in unconfined aquifers considering delayed gravity response: Water Resources Research, v. 10, p. 303-312.

-----1975, Analysis of pumping test data from anisotropic unconfined aquifers considering delayed gravity response: Water Resources Research, v. 11, no. 2, p. 329-342.

Neuman, S.P., and Witherspoon, P.A., 1969, The theory of flow in a confined two aquifer system: Water Resources Research, v. 5, no. 4, p. 803-816.

Papadopulos, I.S., 1965, Nonsteady flow to a well in an infinite anisotropic aquifer: Symposium International Association of Scientific Hydrology, Dubrovnik, p. 21-31.

-----1967, Drawdown distribution around a large-diameter well: National Symposium on Ground-Water Hydrology, San Francisco, Calif., Proceedings, p. 157-168.

Papadopulos, I.S., and Cooper, H.H., Jr., 1967, Drawdown in a well of large diameter: Water Resources Research, v. 3, no. 1, p. 241-244.

Pennsylvania Topographic and Geological Survey, 1981, Glacial deposits of Pennsylvania: Map 59.

-----1989, Surficial materials of Pennsylvania: Map 64.

Pollock, D.W., 1989, Documentation of computer programs to compute and display pathlines using results from the U.S. Geological Survey modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 89-381.

Poth, C.W., 1963, Geology and hydrology of the Mercer Quadrangle, Mercer, Lawrence, and Butler Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 16, 149 p., 5 plates.

Prickett, T.A., and Lonnquist, C.G., 1971, Selected digital computer techniques for groundwater resource evaluation: Illinois State Water Survey Bulletin 55, Urbana, Illinois.

Randall, A.D., and Johnson, A.I., 1988, The Northeast glacial aquifers RASA project--an overview of results through 1987, in Regional Aquifer Systems of the United States—the Northeastern Glacial Aquifers: American Water Resources Association Monograph Series no. 11, p. 1-15.

Reed, J.E., 1980, Type curves for selected problems of flow to wells in confined aquifers: U.S. Geological Survey Techniques of Water-Resources Investigations, book 3, chap. B3, 106 p.

Reilly, T.E., Franke, O.L., and Bennett, G.D., 1987, The principle of superposition and its application in ground-water hydraulics: U.S. Geological Survey Techniques of Water-Resources Investigations, book 3, chap. B6, 28 p.

Reiter, P.F., 1985, Estimating pumping well recharge areas with computer models: *in* Proceedings from the National Water Well Assoc. Conference on Ground-Water Models, August 19-20, 1985, Columbus, Ohio.

Rorabaugh, M.I., 1956, Ground water in northeastern Louisville, Kentucky, with reference to induced infiltration: U.S. Geological Survey Water-Supply Paper 1360-B, p. 117-131.

Shafer, J.M., 1987, GWPATH--interactive ground-water flow path analysis: Illinois State Water Survey Bulletin 69, 41 p.

Schiner, G.R., and Gallaher, J.T., 1979, Geology and groundwater resources of western Crawford County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 46, 103 p., 3 plates.

Schiner, G.R., and Kimmel, G.E., 1976, Geology and ground-water resources of northern Mercer County: Pennsylvania Geological Survey, 4th ser., Water Resource Report 33, 136 p.

Solley, W.B., Chase, E.B., and Mann, W.B., IV, 1983, Estimated use of water in the United States in 1980: U.S. Geological Survey Circular 1001, 56 p.

Taylor, L.E., 1984, Groundwater resources of the upper Susquehanna River Basin, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 58, 136 p., 1 plate.

Thiem, G., 1906, Hydrologische Methoden: Leipzig, Gebhardt, 56 p.

Theis, C.V., 1935, The relation between the lowering of the potentiometric surface and the rate and duration of discharge of a well using groundwater storage: Transactions of the American Geophysical Union, v. 2, p. 519- 524.

-----1938, The significance and nature of the cone of depression in ground-water bodies: Economic Geology, v. 33, p. 889-902.

-----1940, The source of water derived from wells: Civil Engineering, v. 10, no. 5, p. 277-280.

-----1941, The effect of a well on the flow of a nearby stream: American Geophysical Union Transactions, v. 22, pt. 3, p. 734-738.

-----1963, Estimating the transmissibility of a water-table aquifer from the specific capacity of a well, *in* Bentall, Ray, compiler, Methods of Determining Permeability, Transmissibility, and Drawdown: U.S. Geological Survey Water-Supply Paper 1536-I, p. 332-336.

Todd, D.K., 1980, Groundwater hydrology (2nd ed.): New York, John Wiley & Sons, 535 p.

Trescott, P.C., 1975, Documentation of finite-difference model for simulation of three-dimensional groundwater flow: U.S. Geological Survey Open-File Report 75-438, 32 p.

U.S. Environmental Protection Agency, 1987, Guidelines for delineation of wellhead protection areas: EPA 440/6-87-010, 5 chapters.

-----1989, Wellhead protection programs: Tools for local governments: EPA 440/6-89-002, 50 p.

van der Heijde, Paul, and Beljin, M.S., 1988, Model assessment for delineating wellhead protection areas: U.S. Environmental Protection Agency EPA 440/6-88-002.

van der Heijde, Paul, El-kadi, A.I., and Williams, S.A., 1988, Groundwater modeling--an overview and status report: U.S. Environmental Protection Agency contract CR-812603, 259 p.

Walton, W.C., 1988, Practical aspects of groundwater modeling (3rd edition): National Water Well Association, 587 p.

Wang, H.F., and Anderson, M.P., 1982, Introduction to ground water modeling: Finite difference and finite element methods: San Francisco, W.H. Freeman, 256 p.

Williams, J.H., 1991, Tributary-stream infiltration in Marsh Creek Valley, North-Central Pennsylvania: U.S. Geological Survey Water-Resources Investigations Report 90-4052, 39 p.

Zheng, C., 1989, PATH3D, A ground-water path and traveltime simulator, users manual: Rockville, Md., S.S. Papadopulos and Assoc.

GLOSSARY

Aquifer.--"A formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs" (Lohman and others, 1972, p. 2).

Area of diversion.--Surface area of the aquifer that has the same horizontal extent as the volume throughout which water is diverted to the well (Brown, 1963).

Area of influence.--The area throughout which water levels have declined measurably because of discharge from a well. Theoretically, the effects extend to the boundary of the aquifer.

Capture zone.--See zone of diversion.

Contributing area.--Area of diversion along with any adjacent surface areas that provide recharge to the aquifer within the zone of diversion.

Equilibrium.--See steady-state flow.

Homogeneity.--"A material is homogeneous if its hydrologic properties are identical everywhere" (Lohman and others, 1972, p. 8).

Hydraulic conductivity.--"The hydraulic conductivity of the medium is the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow" (Lohman and others, 1972, p. 4).

Induced infiltration.--Seepage to a well from a naturally gaining surface-water source induced by a reversal of the hydraulic gradient due to pumping.

Isotropy.--"That condition in which all significant aquifer properties are independent of direction" (Lohman and others, 1972, p. 9). Properties that are dependent upon direction are said to be anisotropic.

Specific capacity.--"The rate of discharge of water from the well divided by drawdown of water level within the well" (Lohman and others, 1972, p. 11).

Specific yield.--"The volume of water yielded from water-bearing material by gravity drainage, as occurs when the water table declines" (Lohman, 1972, p. 6).

Steady-state flow.--"Steady flow occurs when at any point, the magnitude and direction of the specific discharge are constant with time" (Lohman and others, 1972, p. 6).

Storage coefficient.--"The volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head" (Lohman and others, 1972, p. 13).

Streamflow capture.--The capture by a well of ground water that would have contributed to a gaining stream if the well were not pumped.

Streamflow depletion.--The reduction of streamflow due to streamflow capture or induced infiltration.

Transient flow.--The condition when at any point in the ground-water system, the magnitude or direction of flow changes with time.

Transmissivity.--"The rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient" (Lohman and others, 1972).

Uniform flow.--A characteristic of a flow system where specific discharge has the same magnitude and direction at any point.

Wellhead protection zone.--"The surface and subsurface area surrounding a water well or well field, supplying a public water system, through which contaminants are reasonably likely to move toward and reach such water well or well field" (U.S. Environmental Protection Agency, 1987).

Zone of diversion.--The aquifer volume through which water is diverted to the well.