Numerical Model Analysis of the Effects of Ground-Water Withdrawals On Discharge to Streams and Springs in Small Basins Typical of the Puget Sound Lowland, Washington

By David S. Morgan and Joseph L. Jones

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BRUCE BABBITT, Secretary

U.S. GEOLOGICAL SURVEY

Gordon P. Eaton, Director

For additional information write to:

District Chief
U.S. Geological Survey
1201 Pacific Avenue - Suite 600
Tacoma, Washington 98402

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
inch (in)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
acre	0.4047	hectacre
square foot (ft^2)	0.0929	square meter
square mile (mi^2)	2.590	square kilometer
inch per year (in/yr)	2.54	centimeter per year
foot per second (ft/s)	0.3048	meter per second
foot per mile (ft/mi)	0.18943	meter per kilometer
foot per day (ft/d)	0.3048	meter per day
foot per year (ft/yr)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per second per mile ($ft^3/s/mi$)	0.0176	cubic meter per second per kilometer
cubic foot per second per square mile ($ft^3/s/mi^2$)	0.01093	cubic meter per second per square kilometer
cubic foot per year (ft^3/yr)	0.02832	cubic meter per year
million gallons per day (Mgal/d)	0.04381	cubic meters per second
gallon per minute (gal/min)	0.06309	liter per second
gallon per day (gal/d)	0.2642	liter per day

Temperature: To convert temperature given in this report in degrees Fahrenheit (°F) to degrees Celsius (°C), use the following equation: °C = 5/9(°F-32).

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.

Numerical Model Analysis of the Effects of Ground-Water Withdrawals on Discharge to Streams and Springs in Small Basins Typical of the Puget Sound Lowland, Washington

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ABSTRACT

A numerical ground-water flow model of a hypothetical basin was constructed and used to investigate the effects of ground-water withdrawals on rates of natural discharge to streams and springs in small basins of the Puget Sound Lowland. Definitions of the topography, geology, drainage, and climate of the hypothetical basin were based on the features of typical small basins in the Puget Sound Lowland. This information was used to construct a 13-layer numerical ground-water flow model capable of simulating water levels, hydraulic gradients, and discharge to streams and springs. Three sequences of glacial drift and interglacial deposits were simulated in the model; each sequence consisted of recessional outwash, till, advance outwash, and fine-grained interglacial sediments. Alluvial sediments of the major stream valleys and undifferentiated glacial and interglacial deposits were also included in the model. The model was calibrated by comparing simulated hydrologic conditions with expected conditions and making adjustments to values of hydraulic characteristics as needed. The model was calibrated to predevelopment conditions (those prior to pumping), and then used to simulate the effects of pumping on natural discharge to streams and springs. Seven series of simulations were made to investigate the effects of (1) distance from the well to a stream, (2) the presence of confining layers, (3) pumping rate, (4) depth of the pumped aquifer, (5) distance from the well to a bluff, (6) well density, and (7) recharge rate.

The discharge of wells pumping from unconfined outwash aquifers on the drift plains is derived almost entirely from capture of natural discharge to nearby stream reaches. Increasing the lateral distance between the well and stream caused more of the well discharge to be captured from other streams on the drift plain. Pumping from aquifers separated from the stream by one or more confining layers caused a reduction in the effects of pumping on discharge to nearby streams that was offset by an increase in the effects on discharge to more distant streams and springs. The percentage of well discharge captured from springs on the bluff was sensitive to the distance of wells from the bluff. Simulations also showed that increased well density caused greater water-level decline locally, but, at equilibrium, did not affect the extent of the area affected by reduction of natural discharge to streams and springs. Finally, decreased recharge in areas where development had created impervious surfaces had a direct effect on the natural discharge rates to streams and springs. Increased recharge, however, increased natural discharge and offset the effects of well withdrawals. Further analysis of the time-dependent effects of withdrawals would provide additional insights, but would require the development of a transient version of the model.

INTRODUCTION

The effects of ground-water withdrawals on streamflow have become an important issue in the Puget Sound Lowland of western Washington. Surface-water resources are fully allocated in many parts of the region where population growth has increased the demand for water, and future growth will most likely depend on the availability of ground water. The purpose of this study was to gain a better understanding of the relations and interactions between ground-water and surface-water systems in small basins of the Puget Sound Lowland. It was also hoped that this study, conducted in cooperation with the Washington State Department of Ecology (Ecology), would identify some of the important factors controlling the response of the systems to ground-water withdrawals.

Background

In western Washington, as in many areas of the United States, water users, developers, and regulators are confronting questions about the effects of ground-water withdrawals on ground-water levels and streamflow. Regulators, such as Ecology, are charged with the responsibility of limiting these effects to acceptable levels. This task generally requires the regulators to analyze the effects of proposed ground-water withdrawals on streamflow and spring discharge. The site-specific data on geology, ground water, and streamflow to support these analyses are usually not readily available and can be costly to collect. Over the years, many methods have been devised to estimate the response of surface-water systems to ground-water withdrawals, ranging from relatively simple analytical methods, such as the one advanced by Jenkins (1970), to site-specific transient three-dimensional numerical models. The drawbacks of these methods are that they are either too simplistic to be applied in the complex hydrogeologic environments found in the Puget Sound Lowland, or that they are too costly, time consuming, and their results are not transferable between basins (for example, site-specific models). The difficulty in finding a suitable means of estimating ground water-surface water interactions stems in part from the irregular nature of the quasi- layered glacial deposits. The complex assemblage of these deposits makes numerical simulations, analytical solutions, and intuitive assessments difficult to apply and interpret. The inherent difficulty in assimilating the many factors involved in ground water-surface water interactions makes the issue of ground-water rights versus surface-water rights one of the most intractable problems facing water-supply managers and regulators.

A large part of the flow of streams originating in the Puget Sound Lowland consists of ground water discharged from aquifers of unconsolidated Pleistocene glacial outwash deposits. This part of streamflow is termed baseflow. The water in these streams is used for drinking water, irrigation, and industry, and is appropriated (legally "set aside") for water users through a permit system administered by Ecology. Withdrawals from many streams are limited by State regulations that prohibit users from withdrawing water when the stream has receded to a prescribed minimum acceptable flow. The minimum flows were established so that enough water remains in the stream to allow for the passage of anadromous fish (for example, salmon), the dilution of wastes, and other instream uses. In most cases, the total amount of water that has been appropriated from a stream exceeds the amount available (the amount in excess of the minimum acceptable flow) during periods of low flow, and many of these streams have been closed to further appropriation. Nevertheless, the population continues to increase in the Puget Sound Lowland, and in areas where streamflow is no longer available, water managers, developers, and individuals in need of new water supplies are requesting ground-water-withdrawal permits from the State. There is concern that development of ground water as a water supply may lower ground-water levels and consequently the baseflow of streams in some basins. This would reduce the availability of surface water to existing users and could reduce baseflows to levels below the established minimum flow during some periods. In order to allow development of ground-water resources while ensuring acceptable amounts of baseflow in regulated streams, Ecology needs to estimate the potential for a proposed withdrawal to reduce baseflows.

Purpose and Scope

The purpose of this report is to describe the results of a study to improve the fundamental understanding of ground-water flow and the effects of ground-water withdrawals on ground-water discharge to streams and springs in small Puget Sound Lowland basins. The specific objectives of this study were (1) to develop a generalized conceptual model of the hydrogeology of basins in the Puget Sound Lowland; and (2) to demonstrate the effects of various ground-water withdrawal scenarios on the baseflow of streams originating in these basins—specifically, to evaluate the effects of variations in well location, depth of completion, and rate of withdrawal on the rate and distribution of ground water discharged as baseflow. These evaluations will help improve the understanding of ground water-surface water relations in basins with glacial geology typical of western Washington, and provide regulators a means of assessing the soundness and usefulness of existing or proposed regulation or permitting schemes.

Description of the Puget Sound Lowland

The Puget Sound Lowland, as defined by Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993), is an elongate basin that extends approximately 200 mi along its north-south axis from the Fraser River in Canada to the southern extent of Pleistocene glaciation near Centralia, Wash. (fig. 1). The area shown on figure 1 extends from the crest of the Cascade Range on the east and to the Olympic Mountains and the Straits of Juan De Fuca and Georgia on the west, and covers about 17,600 mi^2. The part of the Puget Sound Lowland underlain by Quaternary glacial sediments, which make up the principal aquifers, was the focus of this study. This part of the Puget Sound Lowland ranges in width from about 15 to 80 mi and covers an area of nearly 7,200 mi^2.

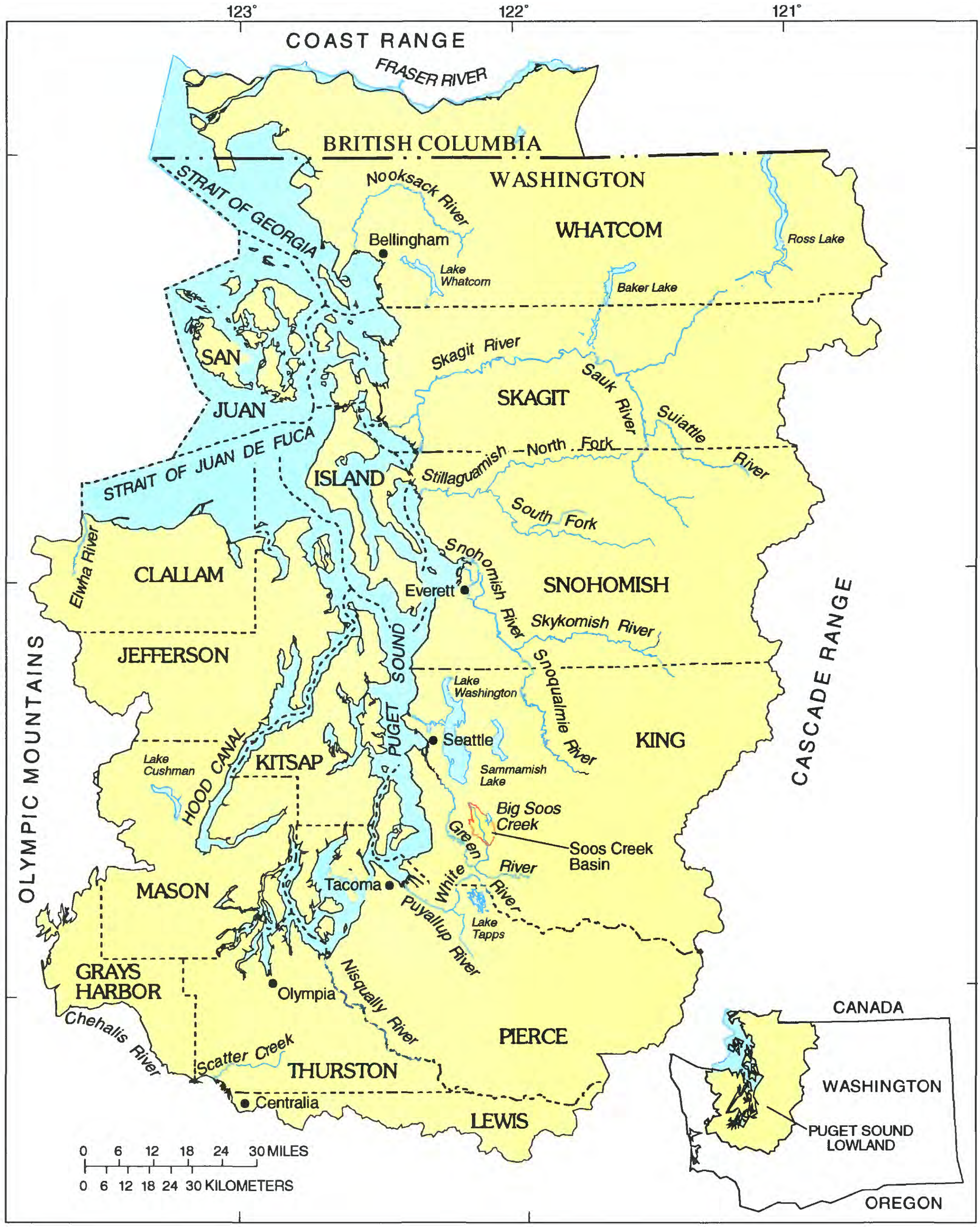

Figure 1.--Location and features of the Puget Sound Lowland, Washington.

During the Pleistocene epoch, southward moving continental glaciers covered the lowland numerous times. Most aquifers and many of the confining layers in the lowland are composed of unconsolidated sedimentary materials deposited as a result of the glaciers' passage. The depositional processes associated with the glaciers produced the layering which is characteristic in the lowland. Periods when the glaciers were advancing or retreating are associated with layered deposits of sand or gravel and till, and periods when glaciers were not in the area, or far removed from it, are associated with fine-grained lacustrine deposits.

The topography of the lowland has been shaped by deposition and erosion that has occurred during the 12,000 to 13,000 years since the last glaciation. The lowland is generally characterized by flat, featureless drift plains that lie at altitudes of 200 to 600 ft above sea level. In places, the drift plains have been incised by major stream valleys; steep bluffs form the boundaries between the drift plains and the major stream valleys below. The effects of continental glaciation on the topography of the lowland are evident in the predominant north-south and northwest-southeast alignment of lakes, ridges, and major stream valleys that were etched by moving ice. As they cross the drift plains, streams have low hydraulic gradient, but the gradient steepens as the streams descend from the plains to the major stream valleys below.

The Puget Sound Lowland has a mid-latitude, Pacific-coast-marine type climate characterized by warm, dry summers and cool, wet winters. Mean annual precipitation ranges from about 25 to 60 in/yr, with a mean of 38 in/yr in 26 drainage basins within the Puget Sound Lowland (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). Nearly 80 percent of annual precipitation falls between October and March. Summer temperatures range from 60°F to 80°F and winter temperatures range from 30°F to 50°F.

Where soils are poorly drained, native vegetation includes fir, cedar, alder, and madrona with an understory of huckleberry, Oregon grape, salal, and blackberry. On well-drained soils underlain by coarse-grained outwash deposits, the dominant vegetation consists of wild grasses, bracken fern, and scotch broom with patches of fir and oak.

In 1990, water use in the Puget Sound Lowland was 810 Mgal/d (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993), 21 percent (174 Mgal/d) of which was ground water supplied by public water systems and 22 percent (178 Mgal/d) of which was ground water from private water systems. The total ground-water withdrawal of 352 Mgal/d in 1990 was approximately three times the amount supplied from ground-water sources in 1965.

Method of Study

Following an extensive review of the literature on the hydrogeology of the lowland, a conceptual model of a small basin was developed that incorporated all of the most significant hydrogeologic characteristics. These characteristics included the glacial stratigraphy unique to the region, the drift plain-bluff-valley topography created by the glaciation and subsequent fluvial erosion and deposition, and the stream networks that provide the surface-water and ground-water drainage for the systems.

References to the many studies of the geology and hydrology of the Puget Sound Lowland can be found in a bibliography compiled by Jones (1991) as part of a regional aquifer system analysis (RASA) carried out by the U.S. Geological Survey. Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) summarizes many of the results and conclusions from the RASA study, including analyses using cross-sectional numerical models in various hydrogeologic settings in the Puget Sound Lowland. This work provided much of the basis for the conceptual model used in the study. The findings of Dion and others (1994) for northern Thurston County and Woodward and others (1995) for southwestern King County also greatly influenced the conceptual model.

The criteria used in developing the conceptual model of the basin were (1) the basin had to be defined in sufficient detail to incorporate the salient features that control ground-water flow, and (2) the definition had to be general enough to be representative of a typical lowland basin. Attaining a balance between simplicity and detail in the conceptual and numerical models was key to producing useful results from the model analysis. Whereas detail was required to provide realistic boundary conditions, simplicity was essential for interpretation of cause and effect relations from model results.

A scale was chosen for the basin that would allow analysis of ground-water development scenarios ranging from single-well, local-scale withdrawals to multiple-well, basin-scale withdrawals. The physical attributes of the hypothetical basin, including topography, geology, and drainage, were then synthesized on the basis of the conceptual model. The conceptual model provided the guidelines such as the altitude and slope of the land surface, stream gradients and tortuosity, and thickness and extent of geologic layers. The spatial data describing the basin were compiled, checked, and stored in digital form using a geographic information system (GIS); this system was later used to create the data files needed by the numerical ground-water model and to store, display, and analyze the results of the model.

Initial estimates of recharge, hydraulic characteristics, and boundary conditions used in the model were based on typical values found by previous investigators in the Puget Sound Lowland. Most of these values were modified during calibration to make the hydrologic conditions simulated by the model more closely match conditions found in small Puget Sound Lowland basins; modifications were generally minor and always left the model parameters well within the range that would be expected in the Puget Sound Lowland for similar conditions or materials. The model was calibrated for predevelopment, steady-state hydrologic conditions. The parameter-adjustment process was completed when the simulated conditions matched expected conditions within tolerable limits.

The hydraulic heads and discharges to streams and springs simulated by the predevelopment, steady-state model represented the baseline hydrologic conditions used in the analysis of the hydrologic response of the hypothetical basin to ground-water development. Seven series of simulations were designed to analyze the effects of specific development variables, such as well depth or distance from a stream, on the response of the ground-water system. The response of the system to each scenario was evaluated by comparing simulated heads and ground-water discharge to streams and springs with those from the baseline model. The location and magnitude of reductions of natural ground-water discharge to streams and springs and the extent and magnitude of water-level declines were compared for each series of simulations. A total of 30 simulations were made in 7 series.

Acknowledgments

Discussions with Linton Wildrick, Robert Garrigues, and Kirk Sinclair of the Washington State Department of Ecology were helpful in development of the scenarios simulated with the numerical model and in determining the best methods of presenting the results of the simulations. Exchanges with members of the Technical Advisory Committee for Hydraulic Continuity, Washington State Water Resources Forum, provided insight into the nature and importance of ground water-surface water issues in the Puget Sound Lowland.

HYDROGEOLOGIC CHARACTERISTICS OF TYPICAL SMALL BASINS IN THE PUGET SOUND LOWLAND

The many factors that control the movement of water within a small basin are, in this report, collectively referred to as the hydrogeologic characteristics of the basin. The most important characteristics are the topography of the land surface, the nature and extent of hydrogeologic layers that ground water must move through, and the quantity of precipitation that percolates through the soil to recharge the ground-water system. These factors play the largest part in determining the direction, rate, and quantity of water moving through the hydrologic system of a basin. This study focused on the effects that these factors have on the exchange of water between the surface and ground-water systems of the basin. In the following sections, these factors are described for basins typical of the Puget Sound Lowland and, more specifically, for the hypothetical basin that was defined for this analysis. The discussion is completed by describing the movement and modes of discharge of ground water in small Puget Sound Lowland basins. Finally, the source of water to pumped wells is reviewed as a preface to discussions of model results.

Topography

The Puget Sound Lowland is characterized by extensive plains of glacial drift, typically 200 to 500 ft above sea level, which have been incised by meltwater from continental and alpine glaciers and by the major streams draining the surrounding mountainous areas. The valleys containing these major streams are typically 10 to 30 mi apart and their floors are at altitudes of less than 100 ft. Smaller tributary streams originating on the drift plain cut smaller valleys into the bluffs that bound the plains. The gradient of these smaller streams is on the order of 80 ft/mi on the upland and 250 ft/mi where they descend to the lower valley at the edges of the drift plains. All of these features were incorporated in the hypothetical basin (fig. 2) of this study.

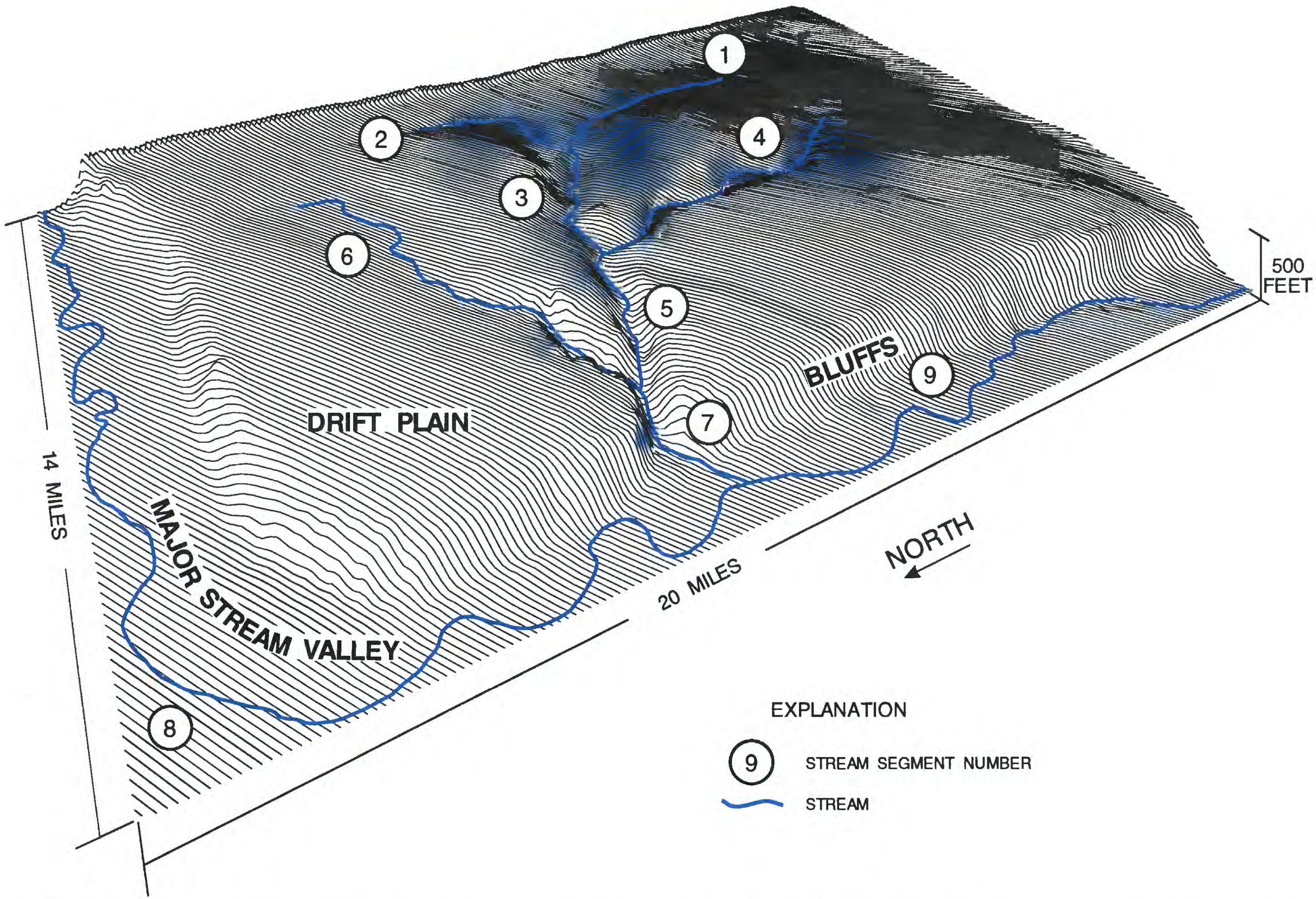

Figure 2.--Three-dimensional perspective view of a hypothetical basin typical of the Puget Sound Lowland, showing topography and streams. View is from the northwest.

Hydrogeologic Layers

The hydrogeology of the hypothetical basin is based on the concept that a glacial episode in the Puget Sound Lowland can be associated with a characteristic sequence of depositional processes. During the long periods between glacial episodes, thick layers of fine-grained sediment were deposited in lakes and by sluggish streams. These deposits are referred to as interglacial fine-grained sediment (Qf). As glaciers advanced southward into the Puget Sound Lowland, coarse debris carried by the glacier was dropped at its leading edge and carried southward by meltwater to form a layer of sand and gravel that is referred to as the advance outwash (Qa), and which overlies the interglacial fine-grained sediments (Qf). As the glacier advanced further southward, it overrode these Qa deposits and additional debris was laid down beneath the glacier and above the advance outwash. This material is typically a highly compacted, unsorted mixture of silt, sand, and pebble-to-boulder debris called till (Qt). Later, as glaciers receded, meltwater carried additional sediment to form recessional outwash (Qr) on top of the till. This sequence of deposition was repeated during ensuing interglacial and glacial periods (Blunt and others, 1987).

These four deposits (recessional outwash, till, advance outwash, and interglacial sediments) form the conceptual depositional sequence of layers that make up a single glacial sequence. Three such glacial sequences are included in the conceptual model of the hypothetical basin (fig. 3).

The youngest sediments in the Puget Sound Lowland are Holocene alluvium (Qal) that has been deposited along valleys by major streams emanating from the Cascade Range and the Olympic Mountains. These broad alluvial valleys have been filled by more than 300 ft of silt, sand, gravel and clay since the last glaciation. Tapped by many high-capacity municipal wells, the alluvium is an important aquifer within the Puget Sound Lowland.

The bedrock (Tb) consist largely of Tertiary claystone, siltstone, sandstone, and some beds of coal (Dion and others, 1994). Because of its low permeability, the bedrock is not an aquifer throughout most of the Puget Sound Lowland and, in the conceptual model of the basin, is considered an impermeable boundary to ground-water flow.

As they were deposited in the Puget Sound Lowland, these sediments (particularly the three glacial sequences) underwent erosion or reworking. Therefore, they are commonly irregular in thickness and composition, and, in some places, they are missing altogether. Geologic maps of the drift plains in the Puget Sound Lowland show large exposures of till (Qt) at the surface with lesser amounts of recessional outwash (Qr) and alluvium (Qal). Advance outwash and interglacial fine-grained sediments are exposed mostly along bluffs at the margin of the drift plain. These hydrogeologic layers and their distribution at the surface are represented in the surficial hydrogeology of the hypothetical basin (fig. 4).

The bedrock structure in the Puget Sound Lowland is roughly trough shaped in a north-south direction. The deepest part of the trough is typically beneath Puget Sound, where the unconsolidated deposits are more than 1,000 ft thick. In the hypothetical model, the maximum thickness of unconsolidated deposits was about 900 ft—where this total thickness exceeds the combined thickness of the three glacial sequences, an underlying layer of undifferentiated deposits (Qu) is part of the conceptual model (fig. 3). Northwest-trending bedrock ridges occur in parts of the Puget Sound Lowland, and this feature was incorporated in the hypothetical basin where bedrock ridges bound the basin on the west and south (fig. 4).

The thicknesses of aquifers and confining layers in the hypothetical basin were chosen by reviewing reports that describe analogous deposits in the Puget Sound Lowland and selecting reasonable values. Table 1 summarizes the findings of the review of literature and lists the values used for the hypothetical basin.

Geologic materials, such as the clays, silts, sands, and gravel that make up the aquifers and confining layers in the Puget Sound Lowland, transmit ground water at rates that are proportional to their hydraulic conductivity. Hydraulic conductivity is defined as the rate that ground water will move through a unit cross section of geologic material under a unit hydraulic gradient. Hydraulic conductivity has dimensions of length per unit time and is commonly expressed in units such as feet per day (ft/d) or centimeters per second. In most geologic materials, hydraulic conductivity varies with direction. In sedimentary deposits, the horizontal and vertical hydraulic conductivity can differ by orders of magnitude. Hydraulic conductivity is typically greatest in the horizontal direction because of the orientation of sediment particles and layers during deposition.

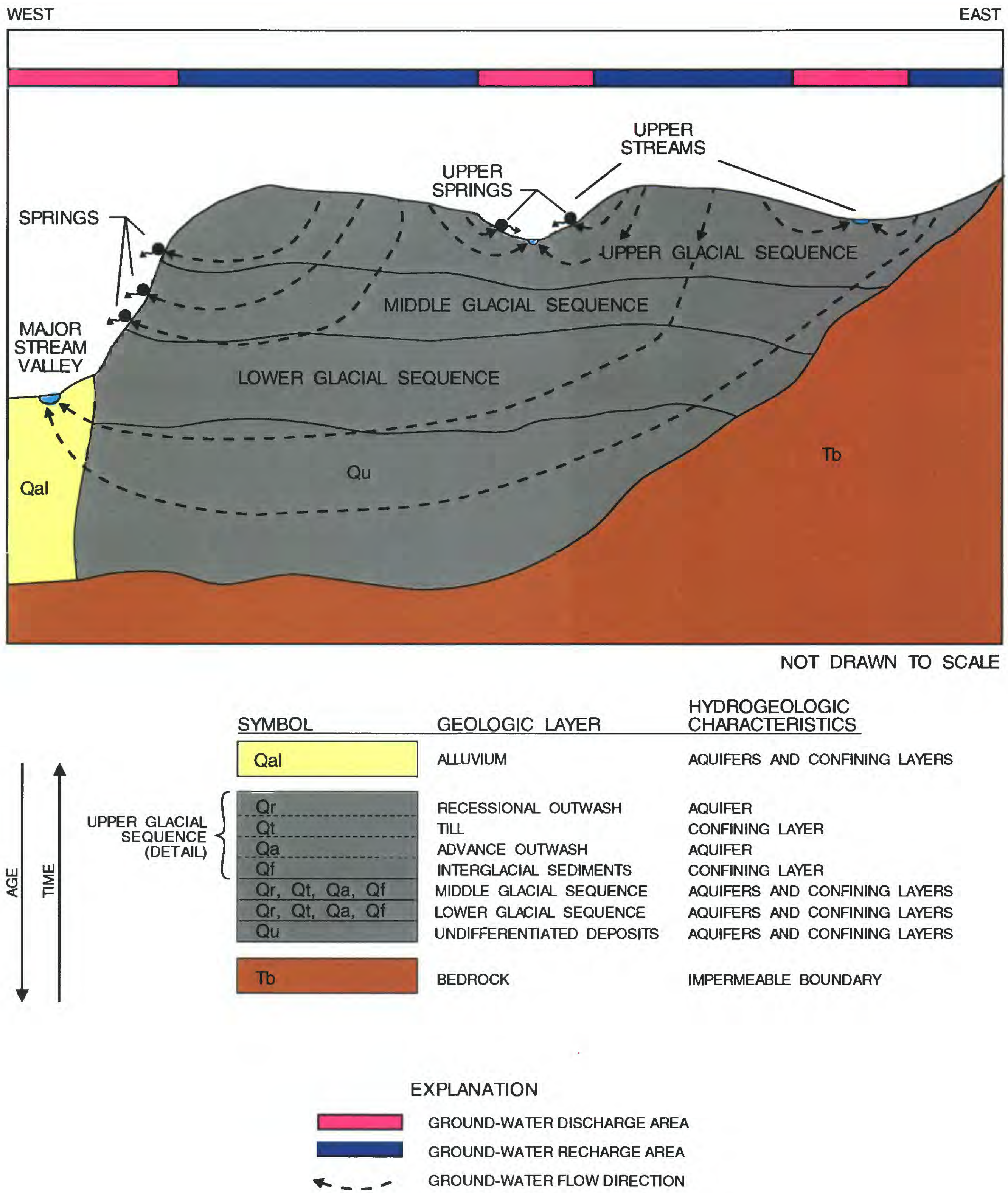

Figure 3.--Hydrogeologic section through a hypothetical basin typical of the Puget Sound Lowland, showing recharge and discharge areas and generalized directions of ground-water flow.

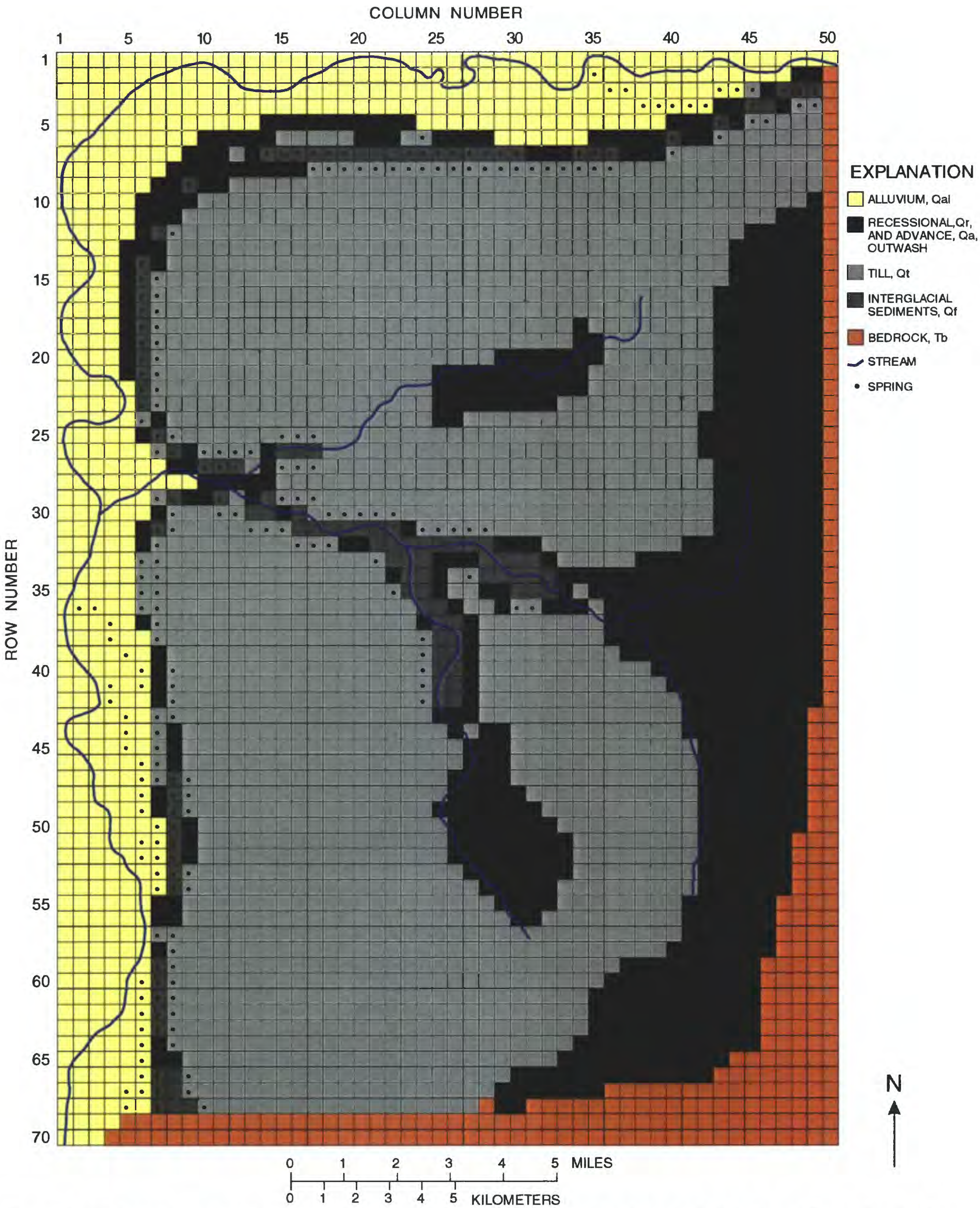

Figure 4.--Hydrogeologic layers exposed at land surface in a hypothetical basin typical of the Puget Sound Lowland. Layers are shown as they were gridded for use in the ground-water flow model.

Table 1.--*Summary of reported thicknesses of hydrogeologic layers in the Puget Sound Lowland, Washington*

[--, no value given]

		Range of thickness, in feet			
Reference	Location	Recess-ional outwash (Qr)	Till (Qt)	Advance outwash (Qa)	Inter-glacial sediments (Qf)
Mundorff and others (1955)	Yelm area	--	[1]15-20	[1]22	--
Sceva (1957)	Kitsap County	--	[2]30	--	[2]100
Liesch and others (1963)	Northwest King County	0-100	[3]150	[2]220	200
Garling and others (1965)	Kitsap County	--	[3]1-50	0-300	1-200
Noble and Wallace (1966)	Thurston County	[1]25	--	[1]30	[1]50
Walters and Kimmel (1968)	Pierce County	[4]60	5-30	25-50	[3]80
Easterbrook (1968)	Island County	0-130	3-175	--	--
Luzier (1969)	Southwest King County	[1]30	0-80	0-300	[3]100
Molenaar and Noble (1970)	Southeast Mason County	[1]30	[1]10	[3]200	[1]15
Livingston (1971)	King County	--	5-20	--	--
Jones (1985)	Island County	[2]45	--	[2]40	--
Dion and others (1988)	Bainbridge Island	10-50	[1]40	[1]30	70-90
Dion and others (1994)	Thurston County	20-50	30-50	25-75	30-120
Vaccaro (written commun., 1995)	Puget Sound Lowland	10-40	20-40	40-50	40-65
This study	Hypothetical Basin	10-50 [1]29	10-50 [1]25	10-115 [1]36	10-150 [1]60

[1]Author's estimate of average thickness.

[2]Estimate of average based on reported values.

[3]Maximum value.

[4]Local (single) observation.

Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) examined estimates of hydraulic characteristics from 17 investigations for the various hydrogeologic layers in the Puget Sound Lowland. He found that the wide range of depositional settings in the region has resulted in an equally wide range in the hydraulic conductivity of the glacial and interglacial deposits.

Coarse-grained alluvium in the major stream valleys can have hydraulic conductivities ranging from 35 to 700 ft/d; however, values of 200 ft/d are more typical. Where finer-grained alluvium occurs, it is typically a fine-sand with silt and clay and has hydraulic conductivity values of about 1 to 15 ft/d (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). Both advance and recessional glacial outwash deposits have hydraulic conductivity values ranging from about 15 to 50 ft/d if they are predominately sand. Values of 100 ft/d are more typical for deposits containing significant fractions of gravel.

The hydraulic conductivity of till in the Puget Sound Lowland varies greatly. Permeameter measurements of hydraulic conductivity range from 0.0002 to 53 ft/d (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). Both this range and the median, 0.12 ft/d, are similar to values reported for till in southern New England, where the range and median were 0.00023 to 96 ft/d and 0.3 ft/d, respectively (Melvin and others, 1992). Little information is available to quantify the hydraulic conductivity of the fine-grained interglacial deposits; however, Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) suggests a range of 0.00001 to 1.0 ft/d, depending on the location and proximity to the source area of the sediments.

Ground-Water Recharge

Most ground-water recharge is derived from infiltration of precipitation and percolation through the soil zone and variably saturated sediments to the water table. In areas where streambeds are above the water table, downward leakage through the streambed and percolation to the water table also contributes to recharge. Finally, in areas where on-site waste disposal systems are used, effluent from these systems also contributes to recharge.

Recharge from infiltration of precipitation has been investigated in several studies in the Puget Sound Lowland. Woodward and others (1995) used a daily water-budget model to estimate ground-water recharge to eight basins in southwest King County. The model, referred to as the Deep Percolation Model (DPM), computes the amount of water that percolates below the root zone after runoff and evapotranspiration are deducted from precipitation (Bauer and Vaccaro, 1987). Dinicola (1990) also used a daily water-budget model to estimate recharge in 33 basins within the Puget Sound Lowland. Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) summarized the ground-water recharge estimates from these studies and the apparent controlling factors on recharge in the Puget Sound Lowland.

Most of the variability in recharge in the Puget Sound Lowland can be attributed to three factors: precipitation, surficial geology, and land use and cover (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). Dinicola (1990) found that where fine-grained glacial till or mudflow deposits were exposed at land surface, recharge rates were much less than where coarse-grained outwash deposits were exposed. After infiltrating through soils in areas underlain by till, most water generally moves laterally along the top of the till until it intercepts a stream channel or land surface. In contrast, where outwash deposits are exposed or immediately underlie the soil, water can freely percolate to the water table. The type of land use and cover controls the amount of precipitation that is lost to evapotranspiration and runoff. Deciduous and conifer forests, grasses, and other types of vegetation have varying water requirements that will affect the amount of water available for recharge. The most important land use or land cover in terms of its effect on recharge is impervious area. Impervious area is generally associated with urban development and includes streets, roofs, driveways and parking lots. Nearly all precipitation that falls on these areas either runs off or evaporates directly. The runoff may be routed either to sewers, ditches, or drywells, and subsequently none, some, or all of the runoff may eventually become recharge.

The mean annual precipitation in the 26 basins whose water budgets are summarized by Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) ranges from 25 to 61 in/yr. Estimated evapotranspiration and runoff ranged from 12 to 21 in/yr and 0 to 22 in/yr, respectively. Baseflow ranged from 0 to 22 in/yr. The range of recharge estimates, 5 to 29 in/yr, reflected the wide ranges of precipitation, evapotranspiration and runoff found in the Puget Sound Lowland.

Vaccaro used regression analysis to determine statistical relations between mean annual precipitation and recharge. The following equations were derived for areas where outwash and till are exposed at land surface:

Outwash areas:

$$R = (0.838P) - 9.77 \quad , \qquad (1)$$

Till areas:

$$R = (0.542P) - 6.06 \quad , \qquad (2)$$

where R is mean annual recharge in inches per year, and P is mean annual precipitation in inches per year. Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) used these equations to estimate recharge within the Puget Sound Lowland. Estimated annual recharge in areas underlain by till and other fine-grained deposits averages 17.5 in. Estimated annual recharge in areas underlain by outwash and other coarse-grained deposits averages 35.9 in; the combined average annual recharge for the entire Puget Sound Lowland is 27 in. These same equations were used in this study to estimate recharge in the hypothetical basin.

Ground-Water Movement and Discharge

In this section, the location, quantity, and modes of discharge from the ground-water system will be discussed, as well as the pathways for ground-water movement between recharge and discharge areas. Figure 3 shows the general directions of ground-water flow in the hypothetical basin.

Ground-water systems have frequently been categorized according to scale, based on the average length of the flow path between recharge and discharge areas (Toth, 1963; Freeze and Cherry, 1979). Woodward and others (1995) have applied the terms local, intermediate, and regional to flow systems in the Puget Sound Lowland and, for consistency, these terms will be used in this report. Local flow systems are characterized by short flow paths within shallow aquifers with small-scale topography usually controlling the location of recharge and discharge areas. At the other extreme, regional flow systems generally have long flow paths within deeper aquifers and are controlled by large-scale topographic features like the Cascade Range and Puget Sound. Intermediate flow systems fall between these extremes. Regional flow systems include the flow paths between the Cascade Range and the Puget Sound that extend mostly through the pre-Quaternary bedrock. Local flow systems generally exist within the upper few hundred feet of Quaternary sediments and recharge is mostly by infiltration of precipitation on the drift plains and discharge is by seepage or springflow onto small streams on the plains, or to larger streams in the adjacent major stream valleys. The major streams act as discharge boundaries to the local flow systems. Intermediate flow systems comprise the flow region above the bedrock and below the deepest part of the local flow system. This arbitrary boundary suggests that there is inter-basin flow above the bedrock under major stream valleys. However, the quantity of ground water underflow between basins defined by local flow systems would be small (W.E. Lum, U.S. Geological Survey, written commun., 1988).

The uppermost recessional outwash aquifer occurs in isolated pockets where it was deposited in topographic lows. It is generally considered to be a water-table aquifer where it is thick and saturated (Woodward and others, 1995). The uppermost till is thought to be saturated where it is overlain by saturated recessional outwash, and at least partially saturated where it is underlain by advance outwash under confined conditions (Woodward and others, 1995). Aquifers beneath the uppermost till are generally confined except near their edges where they have been truncated at bluffs and in canyons by post-Pleistocene erosion. Seepage faces and springs on the bluffs and canyon walls partially dewater the aquifer for some distance from the edge of the aquifer (fig. 3); Woodward and others (1995) suggest that the dewatered zones are typically 0.3 mi wide.

Horizontal Movement

Topography plays an important role in determining the direction of ground-water flow in the Puget Sound Lowland; in fact, the surface of the water table is generally a muted replica of the land surface. A potentiometric surface is an imaginary surface representing the static head of ground water and is defined by the level to which water will rise in tightly cased wells. The water table is a particular potentiometric surface for an unconfined aquifer. The potentiometric surfaces of deeper, confined aquifers exhibit less of the influence of topography with depth, but still are highly controlled by land surface altitude. Ground water moves laterally from topographically high areas toward the major stream valleys and small streams that drain the drift plains. Hydraulic gradient is the change in hydraulic head (water level) per unit of distance in a given direction; the hydraulic gradient generally has both horizontal and vertical components. Typical values for horizontal hydraulic gradients (the slope of the potentiometric surface) range from about 20 ft/mi to 70 ft/mi. Lower gradients of 10 ft/mi occur in very coarse outwash deposits and higher gradients of 100 ft/mi or more occur within very fine-grained sediments or in areas adjacent to steep topography; a regional average for the Puget Sound Lowland of 35 ft/mi has been suggested by J.J. Vaccaro (U.S. Geological Survey, written commun., 1993).

Vertical Movement

Vertical ground-water flow directions vary with location in the basin. In ground-water discharge areas, hydraulic head increases with depth and the direction of the vertical component of hydraulic gradient, and this flow, is upward. Conversely, in recharge areas, hydraulic head decreases with depth and the direction of vertical movement is downward. The upper reaches of small streams on the drift plains are typical of the discharge areas for local flow systems. Springs that issue from outwash aquifers exposed in the stream canyons of the drift plain and on the bluffs above the major stream valley contribute to the flow of streams throughout the year (fig. 3). Most of the upper drift plain, however, is a recharge area and vertical flow is predominately downward. This concept is supported by evidence from many studies that show decreasing hydraulic head with depth below land surface (Woodward and others, 1995; J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). Depths to water in the uppermost confined outwash aquifers range from a few feet or less near streams to 50 ft or more away from streams. Data from wells in the Soos Creek Basin in southwest King County show that head differences between the uppermost confined aquifer and the next deeper aquifer range from approximately 40 to 150 ft and that the larger values tend to occur near the bluffs of the major stream valleys that border the plain (Woodward and others, 1995).

In the major stream valleys, water levels in deeper wells are higher than those in shallow wells, indicating upward flow of ground water and supporting the concept that the major stream valleys are the principal discharge areas for the basin. In wells less than 50 ft deep, water levels are generally a few feet below land surface, but in wells more than 100 ft deep, water levels are above land surface.

Discharge

Ground water leaves (discharges from) the flow system by various means: discharge to streams, discharge to springs, evapotranspiration, and withdrawal by wells. In many areas, a large part of ground-water discharge from springs may flow into streams and indirectly contribute to baseflow. In this report, the contributions to streamflow from direct seepage of ground water through the streambed (baseflow) and from spring discharge that flows into the stream are discussed separately; however, the relative magnitude of the contributions of each is highly variable and difficult to quantify in most field situations.

Baseflow to streams has been estimated by hydrograph separation for several small basins in the Puget Sound Lowland. Woodward and others (1995) estimated baseflow ranging from 4 to 21 in/yr for eight basins; the mean baseflow was 11 in/yr (0.81 $ft^3/s/mi^2$). Baseflow in 26 basins in which recharge estimates were made by Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993) ranged from 0 to 22 in/yr with a mean of about 14 in/yr (1.03 $ft^3/s/mi^2$). Both studies found that baseflow averaged about 36 percent of average annual precipitation. Baseflow for specific stream reaches has been estimated by making gain-loss measurements, but few of these results have been published. Unpublished gain-loss data for the Soos Creek Basin in southwest King County indicate that individual reaches of Soos Creek gain from 0.3 to 3 $ft^3/s/mi$ (D. G. Woodward, U.S. Geological Survey, written commun., 1992). Rates of discharge to major streams are not available.

Spring and seep discharge is difficult to quantify over an entire basin. Individual springs with discharge large enough to measure often do not make up the majority of spring discharge in a basin. More typically, most discharge is to small springs and seeps that cannot be directly measured and are distributed over large areas. Some investigators have attempted to estimate discharge for springs that discharge from the bluffs above major stream valleys in the Puget Sound Lowland. Woodward and others (1995) reported estimates ranging from 0.01 to 0.27 $ft^3/s/mi$, based on spring inventories done by Luzier (1969). These estimates do not include discharge to seepage faces along the bluffs. Woodward and others (1995) suggested that this diffuse discharge could be estimated using potential evapotranspiration (PET) as an index. Phreatophytes are plants whose roots draw water from below the water table. If 25 percent of a 350 ft high bluff is wet or covered by phreatophytes, discharge can occur over an area of 462,000 ft^2/mi. If this area is assumed to transmit water to the atmosphere at the PET rate of 27 in/yr, then the total annual discharge would be 1.04×10^6 $ft^3/yr/mi$, or 0.03 $ft^3/s/mi$. This would be a minimum rate of discharge and is probably much less than the actual rate.

In some areas, a large percentage of spring discharge reaches the stream channel. The contribution of spring discharge to the baseflow of the stream is often indiscernible from direct ground-water discharge to the streambed. Several large (10 to 20 ft^3/s) springs and many smaller (1 ft^3/s) springs contribute to the Nisqually River in Thurston County, and it is estimated that the spring discharge makes up most of the baseflow of the river in some areas (W.E. Lum, U.S. Geological Survey, personal commun., 1992).

Ground water is lost to the atmosphere by evaporation from bare soils and by transpiration from the leaves of phreatophytes. The rate of steady evaporation from bare soil diminishes rapidly with increasing depth to the water table and is negligible for most soils if the water table is more than a few feet below land surface. Transpiration rates are dependent on the type and density of phreatophytes, climatic conditions, quality of water, and depth to water. Evapotranspiration of ground water is an important part of total ground-water discharge in the major stream valleys where the water table is within a few feet of land surface (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). On the upper drift plains the water table generally lies deeper than the roots of phreatophytes can reach except near small streams, where the water table is shallow. In these areas near streams, ground water probably discharges by evapotranspiration at nearly the PET rate.

As many as 30,000 wells in the Puget Sound Lowland withdraw ground water for public supply, domestic, irrigation, commercial, industrial, and institutional purposes (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). According to Vaccaro, ground water supplied 43 percent of all water used in the Puget Sound Lowland in 1990. Typically, ground water makes up an even larger percentage of domestic supplies; for example, in southwestern King County in 1986, the household water needs of 70 percent of the population were supplied by ground water (Woodward and others, 1995). Dion and others (1994) reported that ground water supplied all household water in northern Thurston County in 1988. In northern Thurston County and many other localities in the Puget Sound Lowland, springs are used to supply water for domestic and other uses. Most of the ground-water withdrawals are from the alluvial deposits underlying the major stream valleys and from confined outwash aquifers within the upper 100 to 200 ft on the drift plains. The unconfined recessional outwash deposits on the drift plains are used where they are locally saturated, but they are not an important source in the region.

Source of Water to Wells

In 1940, C.V. Theis published a paper on the hydrologic principles that govern the response of a ground-water system to withdrawals from wells (Theis, 1940). It is worthwhile to review them in order to provide a basis for later discussions of the simulated responses of ground-water systems in the hypothetical basin.

Prior to development of a ground-water system by wells, the system is in a state of equilibrium (steady state) where the natural discharge (D) is exactly equal to the natural recharge (R) when considered over a sufficiently long time period (fig. 5a). Over short time periods, recharge and discharge may not be equal due to normal seasonal variations in climate. However, if average annual climate conditions prevail for successive years, recharge and discharge will be in equilibrium over this period. When recharge and discharge are not equal, ground water is either added to or removed from storage. When recharge exceeds discharge, ground-water storage is increased; conversely when discharge exceeds recharge, ground-water storage is reduced.

When a well begins to withdraw water from a ground-water system, water is removed from storage as the water level drops, forming a cone of depression (fig. 5b). At this stage, the withdrawal (Q) is balanced entirely by a reduction in storage (ΔS):

$$Q = \Delta S \quad . \tag{3}$$

As pumping continues, the cone of depression will expand until it reaches an area where ground water naturally discharges, as to a stream or spring. The cone of depression will reduce the hydraulic gradient toward the discharge area, decreasing the natural discharge to the stream or spring by an amount, ΔD (fig. 5c). The withdrawal will then be balanced by the change in storage, ΔS, and the reduction, or capture, of natural discharge, ΔD:

$$Q = \Delta S + \Delta D \quad . \tag{4}$$

The term "capture" is used in this report to describe the change in location of discharge that occurs when a new stress is imposed on a ground-water system. This term should not be confused with the term "capture area," which is commonly used to describe the contributing area to a well in well-head protection analyses.

The cone of depression will continue to expand as water is removed from storage until it has expanded into a large enough area to capture sufficient natural discharge to completely balance the withdrawal. Once this new balance is achieved, the ground-water system is in a new state of equilibrium ($\Delta S = 0$) and reduced natural discharge ($D - \Delta D$) plus withdrawals (Q) equal natural recharge (R):

$$(D - \Delta D) + Q = R \quad . \tag{5}$$

It is clear from equation 5 that captured natural discharge (ΔD) must be equal to the withdrawal (Q) at equilibrium.

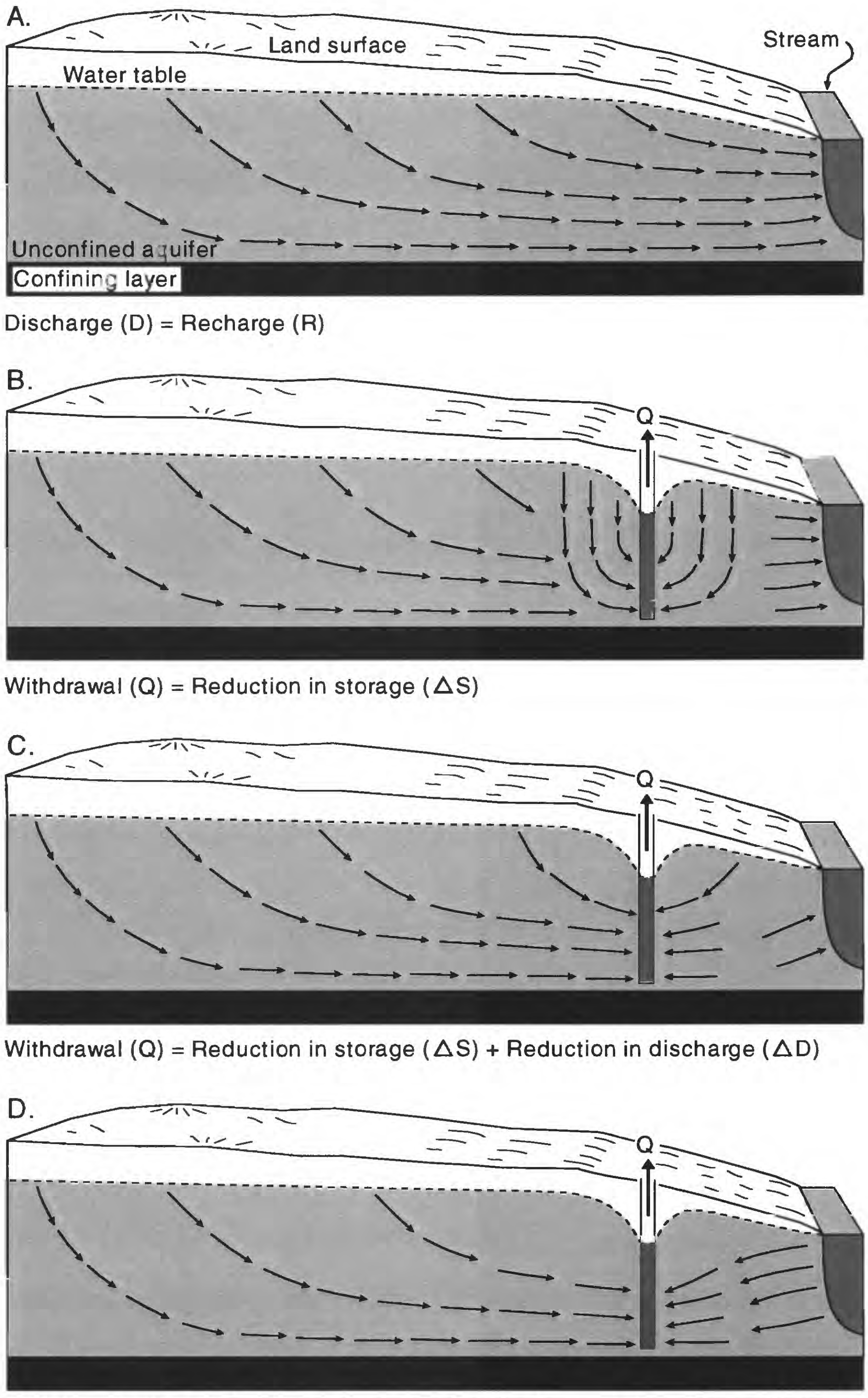

Figure 5.--Source of water to a well (from Heath, 1980).

If the cone of depression expands into a recharge area rather than a discharge area, the hydraulic gradient between the well and the recharge area will be increased. If more water was available than the aquifer could accept as recharge under natural conditions, the increased gradient may allow additional recharge (ΔR) to occur. If and when the increase in recharge plus any decrease in discharge (ΔD) equals the withdrawal, a new equilibrium will be established:

$$(D - \Delta D) + Q = R + \Delta R \quad . \tag{6}$$

In some cases, where pumped wells are located near a stream or the cone of depression expands far enough, the hydraulic gradient can be reversed such that ground-water discharge to the stream stops entirely and water will be induced to move from the stream into the aquifer as additional recharge (fig. 5d).

DESCRIPTION OF THE NUMERICAL MODEL

A three-dimensional numerical model of the hypothetical basin was constructed by selecting boundary conditions and estimating initial values of hydraulic characteristics and recharge. Hydraulic characteristics were adjusted until the model simulated predevelopment conditions within acceptable tolerances. Simulated ground-water levels and discharge to streams and springs agreed with conditions expected in a typical basin.

Approach

Once the conceptual model of the hypothetical basin was defined, the next step was to create a mathematical representation of the basin using a simulation model. To develop the simulation model, first a three-dimensional grid was designed, then it was populated with data on hydraulic characteristics, then boundary conditions were specified, and finally, parameters were adjusted. Model parameters were adjusted until simulated hydrologic conditions were comparable with those that would be expected in the hypothetical basin. The hydrologic conditions considered included the direction and magnitude of hydraulic head gradients (both horizontal and vertical), the rate of seepage to streams, and the rate of discharge to springs. Previous investigators have measured or estimated ranges for these conditions in many parts of the Puget Sound Lowland (as described in previous sections) and these results were relied upon to evaluate how well the numerical model represented the conceptual model. Five parameters were adjusted: horizontal hydraulic conductivity, ratio of horizontal to vertical hydraulic conductivity, streambed hydraulic conductance, hydraulic conductance of springs, and recharge rates. Parameter adjustment continued until the simulated hydrologic conditions were determined to represent typical conditions as defined by previous investigations. Throughout the remainder of the report, the model defined by these parameters and boundary conditions is referred to as the baseline model.

The baseline model was then used to test the effects of one or more discharging wells on ground-water discharge to streams and springs. Several series of simulations were made; each series was designed to show the effect of changes in one variable on ground-water discharge. Variables considered included well depth, distance between well and stream, well discharge rate, and others.

General Features of the Numerical Model

The U.S. Geological Survey's numerical model for simulating ground-water flow, MODFLOW, was used to represent the conceptual model of the hypothetical basin. MODFLOW simulates ground-water flow in three dimensions using finite-difference techniques to solve the partial differential equation describing ground-water movement (McDonald and Harbaugh, 1988). An ancillary program, MODFLOWARC (Orzol and McGrath, 1992), was used to move data directly into MODFLOW from the GIS database.

The MODFLOW program requires that the ground-water system be subdivided, vertically and horizontally, into a finite difference grid of rectangular blocks or cells. The hydraulic properties of the flow system are assumed to be homogeneous within each cell. The saturated flow system of the hypothetical basin was subdivided vertically into 13 layers and horizontally into a 50 column by 70 row grid of 3,500 square cells, each having dimensions of 1,500 ft per side (fig. 6).

Each of the unconsolidated hydrogeologic layers of the conceptual model of the hypothetical basin is represented by an individual layer in the model; bedrock is not explicitly included as a layer in the model, but is represented numerically as a boundary to the system as discussed below. The three glacial sequences, each consisting of four hydrogeologic layers, are represented by the upper 12 model layers, where these layers are present beneath the drift plain. The undifferentiated Qu, are included in the model as the bottom layer (13) beneath the drift plain. The alluvial sediments, Qa, which occur only beneath the major stream valley, were represented in layers 9 through 13 in that part of the model.

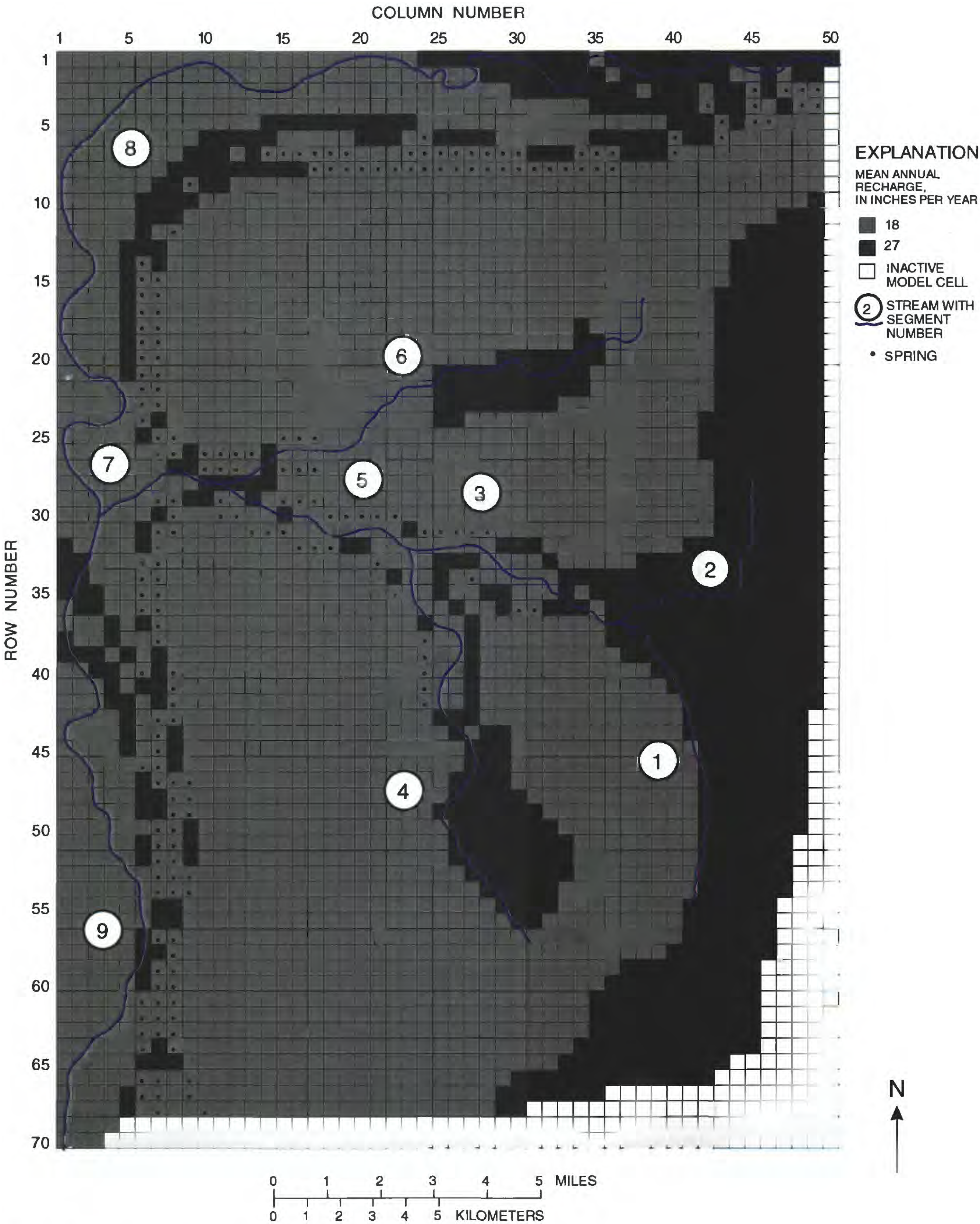

Figure 6.--Model grid, boundary conditions, and mean annual recharge rates.

Transmissivity, which is the product of the horizontal hydraulic conductivity of the aquifer and its saturated thickness, was specified for each cell. Transmissivity was not changed in the simulations because simulated pumping did not significantly reduce the saturated thickness of aquifers.

Boundaries

Boundary conditions were specified in the numerical model according to the concepts of the flow system used to define the hypothetical basin. An important part of the conceptual model of the basin is that the Tertiary bedrock underlying the basin is a low-permeability unit that does not store or transmit significant quantities of ground water. This was the basis for treating the contact between the Tertiary bedrock and the Quaternary glacial and alluvial sediments as a no-flow boundary in the numerical model. This no-flow boundary extends from where the bedrock is exposed along ridges on the eastern and southern boundaries of the basin, beneath the glacial and alluvial deposits, to the northern and western boundaries of the model (figs. 3 and 4).

The northern and western boundaries of the numerical model coincide with the major stream valleys bounding the drift plain. The streams within these valleys are conceptualized as the discharge areas for local- and intermediate-scale ground-water flow systems in the lowlands; all ground water that enters the system as recharge within the hypothetical basin is assumed to leave the system as discharge within the boundaries of the model. The northern and western boundaries of the model were also specified as no-flow boundaries to reflect the assumption that there is no subsurface ground-water flow out of the basin.

Head-dependent flux boundaries were used to represent ground-water discharge from within the basin by seepage to streams and springs (fig. 6). The underlying equations of these boundary conditions are described by McDonald and Harbaugh (1988).

Hydraulic Characteristics

In order to simulate a ground-water flow system with a numerical model, the hydraulic characteristics of the aquifers and confining beds must be specified for each model cell. The hydraulic characteristics normally required to simulate a ground-water system are thickness, hydraulic conductivity, and specific storage. Specific storage was not required for this analysis because all simulations were for steady-state conditions. The thickness of each hydrogeologic layer, specified on the basis of previous investigations, was variable. Ranges in thickness were 10-50 ft for recessional outwash and till, 10-115 ft for advance outwash, and 10-150 ft for interglacial sediments (table 2).

Horizontal Hydraulic Conductivity

The hydraulic conductivity of deposits in the Puget Sound Lowland spans several orders of magnitude. Even within an individual hydrogeologic layer, hydraulic conductivity values show high variability due to small scale erosional and depositional features. This type of spatial variability in hydraulic conductivity was not represented in the model primarily because the scale of the features that cause heterogeneity in hydraulic conductivity are probably not large enough to significantly affect the simulated response of the system on a basin-wide scale. Also, if this heterogeneity were incorporated in the model, it would have made it difficult or impossible to separate effects of variables such as distance between the well and the stream from effects of spatial variability on hydraulic conductivity on the capture of discharge to streams and springs. Therefore, horizontal hydraulic conductivity was assumed to be uniform throughout the basin for each hydrogeologic layer. The recessional, Qr, and advance, Qa, outwash aquifers were assumed to have the same horizontal hydraulic conductivity, 100 ft/d (table 2). The confining layers (till, Qt, and interglacial sediments, Qf) were assigned horizontal hydraulic conductivities of 0.25 and 1.0 ft/d respectively. The undifferentiated deposits, Qu, are composed mainly of fine-grained sediments, but have coarse-grained interbeds; the relatively high hydraulic conductivity of 25 ft/d assigned to the layer reflected the influence of these coarse-grained beds. Transmissivity varied within model layers due to variations in the thickness of layers. The mean transmissivity of the aquifer layers (Qal, Qr, and Qa) ranged from 2,900 ft^2/d for recessional outwash aquifers to 20,000 ft^2/d for the alluvial aquifer.

Table 2.--*Thickness and hydraulic characteristics of hydrogeologic layers in the hypothetical basin*

[--, no value given]

Symbol	Geologic layer	Thickness range (feet)	Mean thickness (feet)	Horizontal hydraulic conductivity (feet per day)	Horizontal to vertical hydraulic conductivity (ratio)	Vertical hydraulic conductivity (feet per day)	Model layer(s)
Qal	Alluvium	10-500	400	50	10	5	[1]9-13
Qr	Recessional outwash	10-50	29	100	10	10	1,5,9
Qt	Till	10-50	25	0.25	100	0.0025	2,6,10
Qa	Advance outwash	10-115	36	100	10	10	3,7,11
Qf	Interglacial sediments	10-150	60	1.0	200	0.005	4,8,12
Qu	Undifferentiated deposits	10-480	275	25	150	0.167	13
Tb	Bedrock	--	--	--	--	--	Not in model layer

[1]Occurs only beneath valley floor.

Vertical Hydraulic Conductivity

The heterogeneity of the sedimentary layering in the basin, and particularly of the glacial deposits, imparts considerable anisotropy to their hydraulic conductivity. Anisotropy is the condition of having different properties in different directions. Expressed as the ratio of horizontal to vertical hydraulic conductivities, the anisotropy ratio of unconsolidated sediments is usually greater than one because of the preferred orientation of grains and clasts during deposition. Anisotropy ratios were assigned to each hydrogeologic layer ranging from 10 for the coarse-grained aquifer layers, (Qal, Qr, Qa) to 200 for the mostly fine-grained interglacial layers (Qf). Till layers, Qt, were assigned an intermediate ratio of 100 and the undifferentiated sediments, Qu, were assigned a ratio of 150.

Based on the horizontal hydraulic conductivity value of 0.25 and the anisotropy ratio of 100, the effective vertical hydraulic conductivity of the till was 0.0025 ft/d. This value is within the range of 0.001 to 0.01 ft/d reported by Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). The vertical hydraulic conductivity of the interglacial deposits was about 0.005 ft/d, or twice that of the till layers.

Recharge

Recharge to the ground-water system was specified at rates computed with the regression equations 1 and 2 described in the previous section on ground-water recharge. Based on average annual precipitation of 44 in/yr, rates of 18 in/yr and 27 in/yr were computed for areas where till and outwash are exposed. The distribution of recharge is shown on figure 6. Mean annual recharge for the basin is about 20 in/yr.

Discharge

Ground-water discharge to streams and springs was represented in the numerical model with the MODFLOW packages RIVER and DRAIN (McDonald and Harbaugh, 1988).

Streams

Streams in the hypothetical basin, both on the upper drift plain and in the major stream valley, were simulated as head-dependent flux boundaries with the RIVER package in MODFLOW. Ground-water flux across head-dependent boundaries is calculated as a piecewise-linear function of the difference between the water levels in the aquifer and at the boundary, and the conductance of that boundary (McDonald and Harbaugh, 1988). Sediments between the stream and the aquifer commonly form distinct bed material differing in hydraulic character from the aquifer sediments themselves. This frequently occurs in low-gradient streams where fine sediment may accumulate to form a bed with low hydraulic conductivity. However, in many streams where gradients are sufficient to maintain high flow velocities or where the streambed is periodically flushed of fine-grained sediments by high flows, there may be little or no contrast between the hydraulic properties of the stream bottom and the aquifer itself.

Stream reaches are defined as the lengths of stream contained within one model cell; stream segments are groups of contiguous reaches that have no tributaries. Streams in the basin are identified by segment numbers, with nine stream segments in the basin: segments 1 through 7 on the upper drift plain and segments 8 and 9 in the lower stream valleys (fig. 6).

Head-dependent boundary conductances are analogous to the vertical conductance between model layers. However, boundary conductances are adjusted external to the model to account for the area of a cell that is covered by the boundary, whereas vertical conductances are adjusted internally. The generalized boundary-conductance equation is

$$C = \frac{Kv \times A}{b} \quad , \qquad (7)$$

where C is the boundary conductance, Kv is the vertical hydraulic conductivity of the sediments between the stream and the center of the model cell containing the stream, A is the area of the stream within the model cell, and b is the thickness of sediments between the stream and the center of the model cell containing the stream.

Values of boundary conductance for the streams were estimated by assuming that the vertical hydraulic conductivity (Kv) of sediments between the stream and the cell is equal to 2 percent of the horizontal hydraulic conductivity of the cell. The average stream area in a cell was estimated by measuring the length of each reach and assuming a width of 10 ft. Although no discrete, low-permeability streambed may be present, a thickness (b) must be assumed at each model cell to describe the equivalent thickness of the materials that restrict the flow of water between the stream and the ground-water system. This thickness was specified as 2 ft throughout the model. The conductance values resulting from these assumptions are essentially lumped parameters that represent a proportionality constant that determines the flux between the ground-water system and the stream for a given difference between stream stage and ground-water level. The weighted-mean streambed conductance values used in the model ranged from 3,300 ft^2/d for segment 5 to 14,600 ft^2/d for segment 2; the overall weighted mean conductance was 7,800 ft^2/d.

Springs

The springs that occur on the bluffs above the major stream valley and in the canyons of the streams on the upper drift plain were simulated with the DRAIN package of MODFLOW (McDonald and Harbaugh, 1988). The DRAIN package allows simulation of head-dependent boundary flux in a manner similar to the RIVER package. However, the flux is allowed only in one direction; once the head in the aquifer falls below a specified elevation, discharge stops. Springs were specified at 443 cells in layers 3, 5, 7, 9, and 11 where the outwash aquifers are exposed in canyons and on bluffs (fig. 6). The conductances for the springs in the model were computed with the same general equation as for streams, but the assumptions were somewhat different. The area, A, of the seepage face at each spring was assumed to span the entire width of the cell (1,500 ft) and the height of the seepage face was assumed to be half of the thickness of the cell. The hydraulic conductivity, Kv, of the spring was estimated to be the same as that of the hydrogeologic layer of the cell. The length of the flowpath, b, was the distance from the cell center to the seepage face (750 ft). Using these assumptions, mean spring conductances for spring cells ranged from 1,300 ft^2/d for layer 11 to 2,900 ft^2/d for layer 3, with an overall mean of 2,400 ft^2/d. The discharge altitude for each spring cell was specified as the altitude of the bottom of the aquifer at the point where it is exposed on the bluff or canyon. The 164 spring cells in layer 3 discharged from a mean altitude of 340 ft, and the 25 spring cells in layer 11 discharged from a mean altitude of 66 ft.

Baseline Model Results

About 30 simulations were made in order to obtain a set of hydraulic characteristics and boundary conditions that simulated hydrologic conditions typical of small basins in the Puget Sound Lowland. The process used to obtain these baseline model parameters was trial and error. Following each simulation, simulated hydrologic conditions were compared with the expected conditions and parameters were adjusted to correct errors in the simulated conditions. The final set of model parameters provided the "best fit" between the expected and simulated conditions. Although there are other sets of parameters that could produce equal, and possibly better fits, all of the hydraulic characteristics and boundary conditions in this set are well within ranges expected for the conceptual model of the Puget Sound Lowland.

The baseline model simulates ground-water flow in the hypothetical basin prior to any withdrawal of ground water by wells. The system was assumed to be at equilibrium, or steady state, with discharge balanced by recharge. The expected hydrologic conditions used to calibrate the baseline model are based on data from studies of several areas in the Puget Sound Lowland. Although these data may not reflect steady-state conditions in all cases, the generalized expected conditions are probably typical of the conditions that would be found in an undeveloped basin.

In the following sections, ground-water levels, hydraulic head gradients, and ground-water discharge rates simulated with the baseline model are compared with conditions that would be expected in the typical Puget Sound Lowland basin.

Ground-Water Levels and Hydraulic Gradients

In the baseline model of the hypothetical basin, ground-water flow is generally from the southern, southeastern, and eastern parts of the basin toward the north, northwest, and west. In the shallow part of the flow system, hydraulic gradients and flow directions are strongly controlled by the streams and topography of the upper drift plain (fig. 7). Contours of ground-water altitude show that the shallow aquifers discharge to streams where the contours intersect the streams in V-shapes. Horizontal hydraulic gradients range from less than 15 ft/mi on the northwestern part of the plain to more than 40 ft/mi where they steepen in the uplands on the southeast corner of the plain. Flow directions in the deeper aquifers are more strongly influenced by the discharge areas on the bluffs and in the major stream valleys to the north and west (fig. 8). In the southwest corner of the drift plain, ground water in the shallow aquifer flows from southeast to northwest, whereas deeper ground-water flowpaths are nearly due west toward the springs on the bluff and the major stream. Simulated horizontal hydraulic gradients ranging from about 15 ft/mi to 40 ft/mi, compare closely with typical gradients of 35 ft/mi reported by Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993).

Throughout the drift plain, the simulated vertical component of ground-water flow is downward, except in shallow aquifers near streams where there is upward flow toward the stream. Heads decrease with depth beneath the drift plain, which is a recharge area. Head differences between adjacent aquifers are greatest in the shallow aquifers underlying the plain near the bluff and decrease with depth and to the east and south. Differences in simulated heads are as much as 120 ft between shallow aquifers near the bluff and as little as 20 ft between deep aquifers near the eastern and southern edges of the basin. These simulated differences compare closely with expected differences based on data for southwest King County from Woodward and others (1995), who found differences ranging from about 40 to 150 ft between water levels in the uppermost confined outwash aquifers. Simulated heads increase with depth beneath the major stream valley, which is a discharge area. The maximum simulated vertical head differences between the shallow and deep aquifers beneath the major stream valley were 10 to 20 ft.

Stream and Spring Discharge

Total ground-water recharge to and discharge from the basin, as simulated by the baseline model, was 389 ft^3/s (table 3). Seventy-six percent (295 ft^3/s) of baseline ground-water discharge was by seepage to streams and 24 percent (94 ft^3/s) was by spring discharge. Seventy-three percent (285 ft^3/s) discharged to the major stream valley either by seepage to streams (202 ft^3/s) or by spring discharge on the bluffs (83 ft^3/s). The remaining 27 percent discharged to streams (93 ft^3/s) and springs (11 ft^3/s) on the drift plain. In the following discussion, springs have been grouped with the stream segment to which they would contribute if the spring discharge flowed to the stream; spring discharge has also been summarized this way in table 3. Total ground-water discharge to streams and springs by stream segment is shown in figure 9a.

Annual baseflow averages about 36 percent of precipitation in Puget Sound Lowland basins (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993). The assumed annual precipitation of 44 in/yr in the hypothetical basin would result in baseflow of about 16 in/yr. Over the 262 mi^2 basin this would result in 308 ft^3/s of baseflow. The simulated total seepage to streams in the hypothetical basin is only 295 ft^3/s (table 3); however, this figure does not include any spring discharge that might flow into the stream and contribute to baseflow. Of the 94 ft^3/s of spring discharge, it is likely that at least 13 ft^3/s, which is the difference between the simulated and expected baseflows, contributes to streamflow in the basin.

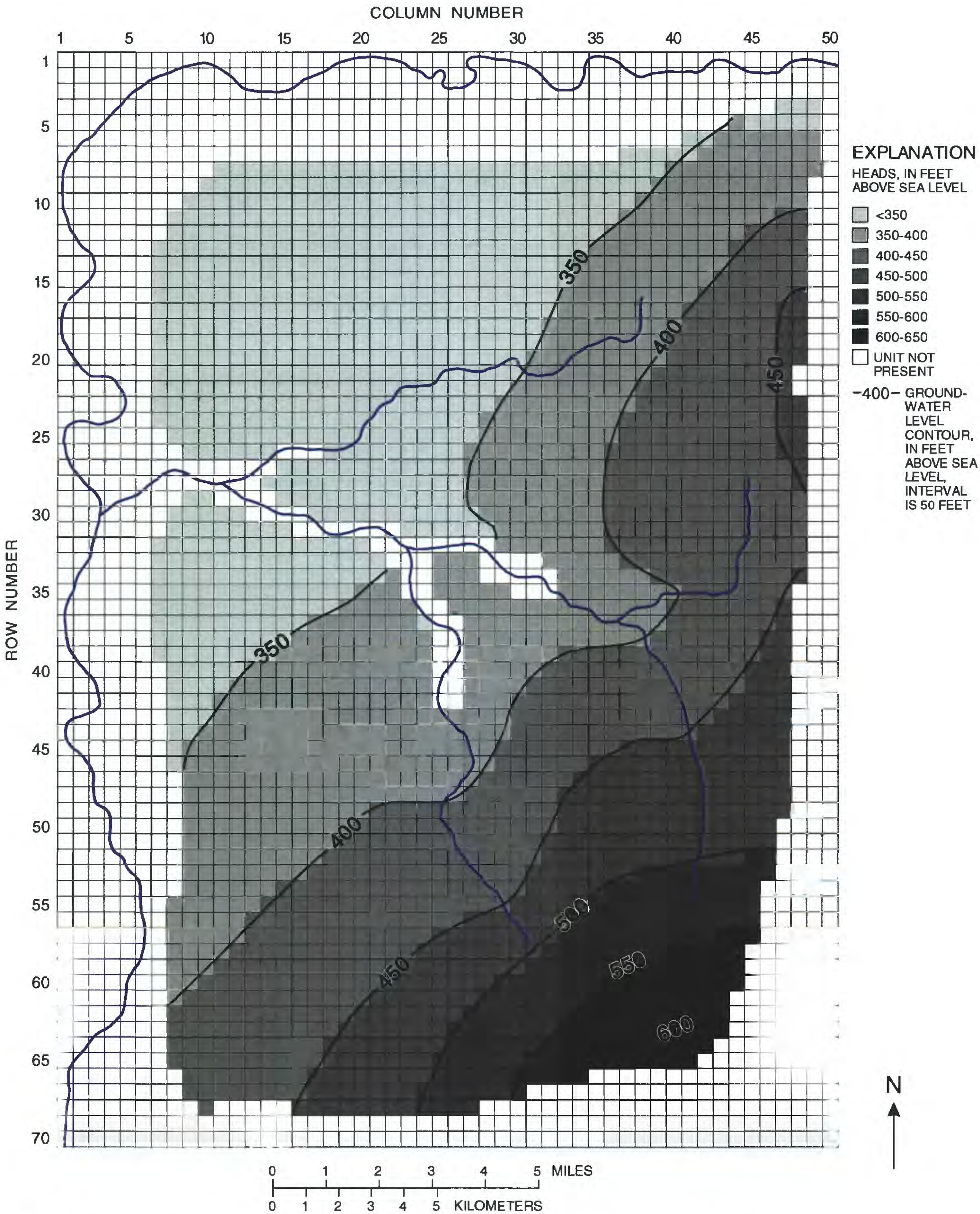

Figure 7.--Simulated water levels in the uppermost confined outwash aquifer (layer 3) for baseline conditions.

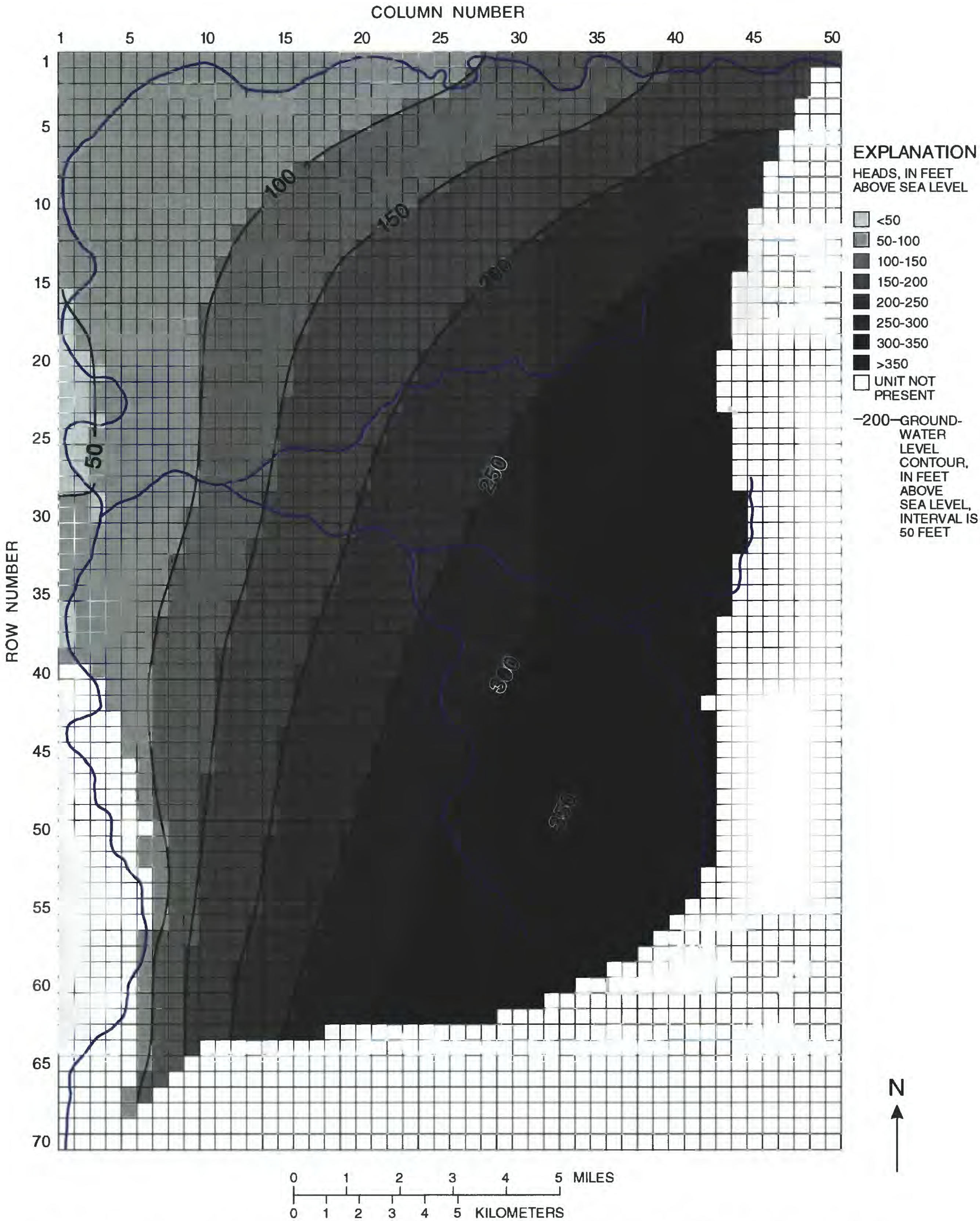

Figure 8.--Simulated water levels in the lowermost confined outwash aquifer (layer 11) for baseline conditions.

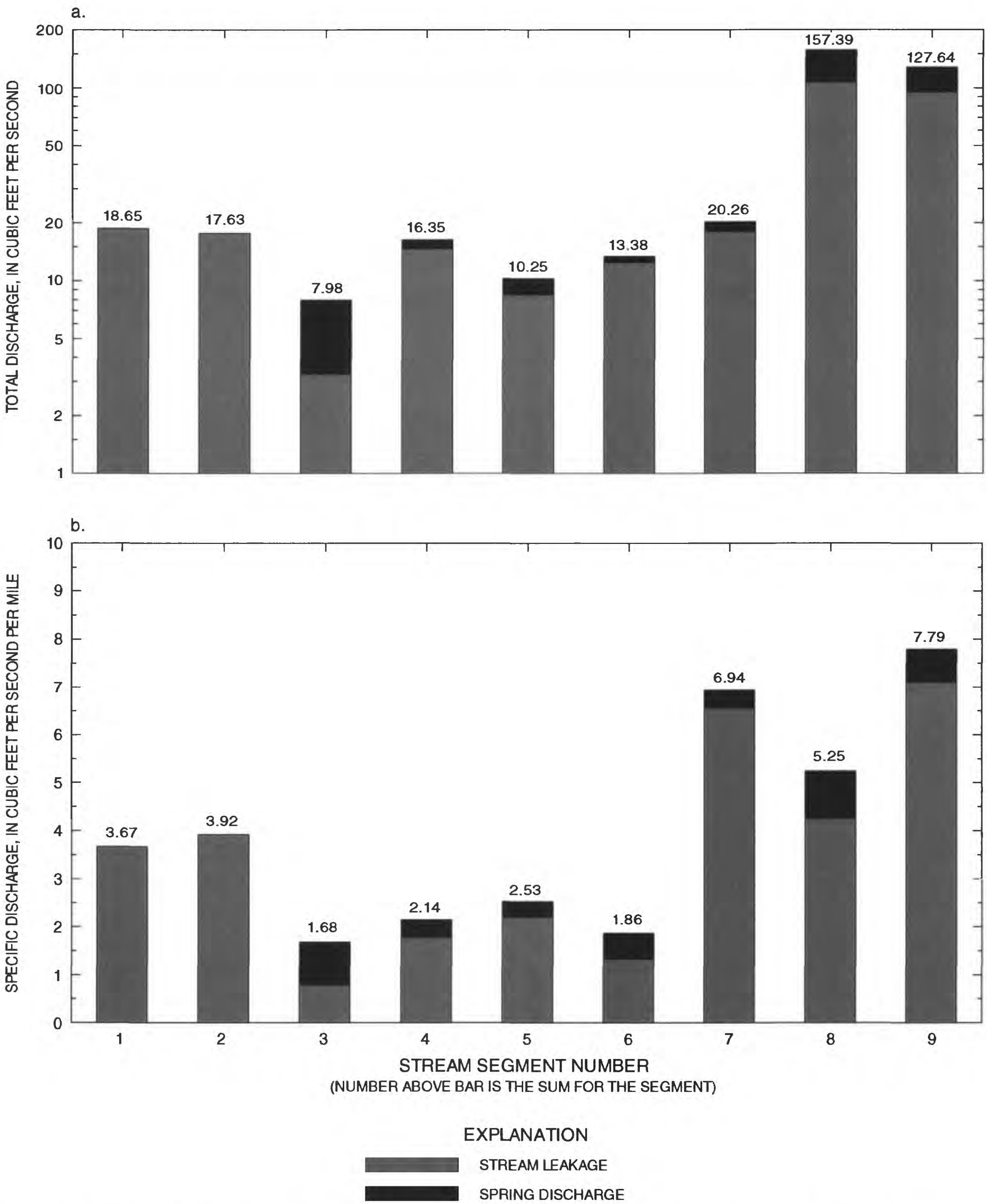

Figure 9.--Simulated ground-water discharge to streams from stream leakage and spring discharge for baseline conditions, by stream segment number, (a) total discharge, and (b) specific discharge.

Table 3.--*Simulated ground-water discharge to streams and springs in the baseline model*

Location	Discharge, in cubic feet per second; percentage of total ground-water discharge in parentheses					
	To streams		To springs		Total	
Drift plain[1]	93	(24)	11	(3)	104	(27)
Stream valley[2]	202	(52)	83	(21)	285	(73)
Total	295	(76)	94	(24)	389	(100)

[1]Includes stream segments 1 through 7 and contributing springs (see figure 6).

[2]Includes stream segments 8 and 9 and contributing springs (see figure 6).

Another assessment of the simulated discharge to streams was made by comparing simulated discharge to streams on the drift plain with estimated baseflow in a small basin in southwest King County. Big Soos Creek (fig. 1) drains an area of 66.7 mi^2 and has a mean monthly flow of about 35 ft^3/s for October (1967-92), a time when flow in streams in the Puget Sound Lowland are supplied almost entirely by baseflow. This discharge represents the best available estimate of minimum annual baseflow to Big Soos Creek, since baseflow is typically greater during winter and spring months when the water table is higher. Based on this estimate, the minimum baseflow per square mile is 0.52 ft^3/s. The drift plain of the hypothetical basin covers about 188 mi^2 and therefore the expected minimum baseflow, based on the Big Soos Creek data, is about 100 ft^3/s. Simulated discharge to streams on the drift plain was 93 ft^3/s by seepage and 11 ft^3/s from spring discharge, for a total of 104 ft^3/s.

According to Vaccaro (J.J. Vaccaro, U.S. Geological Survey, written commun., 1993), estimated annual baseflow in 12 Puget Sound Lowland basins averages about 0.93 $ft^3/s/mi^2$, or nearly twice the minimum baseflow of 0.52 $ft^3/s/mi^2$ estimated above. Using the higher value, the expected annual baseflow from the 188 mi^2 drift plain in the hypothetical basin would be 175 ft^3/s. This value may be a more realistic estimate for comparison with the simulated baseflow, since the model simulates average annual baseflow. On this basis, the simulated baseflow of 104 ft^3/s may be somewhat lower than what would be expected from an area this size. There are several sources of uncertainty in the estimate of annual baseflow. Considering only the uncertainty in determining the area contributing to baseflow, the uncertainty in the estimate of annual baseflow is probably ±25 percent or more. Because simulated baseflow from the drift plain (104 ft^3/s) falls within the range of estimated baseflow (100 to 175 ft^3/s), the distribution of ground-water discharge in the model is a reasonable representation of a typical basin.

The rate of ground-water discharge per mile of stream, or specific discharge, was computed for each stream segment (fig. 9b). Including streams and springs, stream segments 1 through 6 had simulated specific discharges ranging from 1.68 to 3.92 $ft^3/s/mi$, while segments 7 through 9 had discharges ranging from 5.25 to 7.79 $ft^3/s/mi$ (fig. 9b). The large difference is due to the positions of segments 7 through 9 in the lower part of the flow system where they receive discharge from a broader recharge area than do the streams on the drift plain. If spring discharge is not included, specific discharge to streams on the drift plain ranges from 0.8 to 3.9 $ft^3/s/mi$ (segments 1 through 6 only); this range is similar to the range of 0.3 to 3 $ft^3/s/mi$ determined for reaches of Big Soos Creek in southwest King County (Woodward and others, 1995).

The specific discharge of simulated springs was computed by dividing the total discharge of springs contributing to each stream segment by the length of spring discharge area adjacent to that stream segment. Specific discharges ranged from about 0.3 to 1 $ft^3/s/mi$. The values at the lower end of this range are comparable to values at the upper end of the range of 0.01 to 0.27 $ft^3/s/mi$ reported by Woodward and others (1995) for southwest King County. As many as four outwash aquifers are exposed on the bluffs above the major stream valley, and of the 453 spring cells in the model, 78 percent are located on these bluffs. Consequently, some of the highest specific discharges for springs (0.7 and 1.0 $ft^3/s/mi$) are on segments 8 and 9 in the major stream valley below the bluffs. Discharge per spring, however, is not significantly higher; the springs on the bluffs comprise 78 percent of all springs in the model and account for 88 percent of the total spring discharge, whereas springs on the drift plain comprise 22 percent of all springs and discharge 12 percent of total spring discharge (table 3).

SIMULATION OF THE EFFECTS OF GROUND-WATER WITHDRAWALS ON DISCHARGE TO STREAMS AND SPRINGS

An analysis of the effects of ground-water withdrawals on discharge to streams and springs was made using the baseline model. The baseline was used to simulate the effects of ground-water withdrawals under various scenarios in which well depth, location, pumping rate, or other conditions were varied. The effects of withdrawals on ground-water levels, discharge to stream segments, and discharge to individual stream reaches and springs are compared for each series.

Approach

Seven series of simulations were made and each series consisted of from two to six simulations. In most simulations, only one condition was varied from the baseline model. The 30 simulations, and the conditions for each, are summarized in table 4. Each simulation is referred to by an alphanumeric designation, where the letters indicate the simulation series and the integer identifies the simulation within the series. For example, DEPTH.2 is the second simulation in the series in which one condition, the depth of the pumped aquifer, was varied. The locations of wells simulated in each series are shown in figure 10.

The condition varied for each of the seven series of simulations was:

1. Distance of well from stream (pumping unconfined aquifer) (DIST.U.1 to DIST.U.5);
2. Distance of well from stream (pumping confined aquifer) (DIST.C.1 to DIST.C.5;.
3. Pumping rate (PUMP.1 to PUMP.4);
4. Depth of pumped aquifer (DIST.U.5, DIST.C.5, DEPTH.1 to DEPTH.3);
5. Distance of well from bluff (BLUFF.1 to BLUFF.5);
6. Well density (WELL.1 to WELL.2); and
7. Recharge rate (RECH.1 to RECH.6).

The results of each series are discussed as to the simulated effects of pumping on ground-water discharge to streams and springs and on ground-water levels. In the following sections, the effect of the well or wells on discharge to streams and springs is quantified by expressing the reduction in discharge to streams and springs as a percentage of the well discharge. The reduction in, or capture of, discharge to streams and springs is the difference between simulated discharge to streams and springs in the baseline model and simulated discharge in the scenario being considered. The simulated reductions in discharge are grouped by feature (stream or spring) and by the feature's location. Feature type and location are used to describe the source and location of capture; four combinations are used: streams on the upper drift plain, springs on the upper drift plain, streams on the lower valley floor, and springs on the lower bluffs. The term "area of influence" (AOI) describes the area of the pumped aquifer that experienced 0.1 ft or more of water-level decline in response to the pumping stress applied for the scenario.

Effects of Varying Distance From a Stream

The effects of varying distance of a well from a stream on discharge to streams and springs were investigated with two series of simulations. One series (DIST.U) simulated a well pumping from the unconfined aquifer, and the other (DIST.C) simulated a well pumping from the uppermost confined aquifer. Both series comprised five simulations in which the well was moved progressively further from the stream; the well locations and pumping rates were the same for the two series.

Pumping From an Unconfined Aquifer

In the DIST.U series, simulations were made of withdrawals from a single well pumping 300 gal/min from an unconfined aquifer while distance from the well to the stream was varied from less than 1,500 ft to 6,000 ft. The locations of the well in each of the five simulations (DIST.U.1 through DIST.U.5) are shown in figure 10. In the first simulation, DIST.U.1, the well was located in the same model cell as the stream; in each successive simulation, the well was moved one cell farther from the stream until, in simulation DIST.U.5, the well was 6,000 ft from the stream. In each simulation the well pumped from the uppermost recessional outwash aquifer (model layer 1). In this part of the basin the aquifer is 10 to 30 ft thick; the horizontal hydraulic conductivity is 100 ft/d throughout the basin. The bed of stream segment number 4 is in contact with the pumped aquifer in the vicinity of the well; however, the outwash aquifer is limited in extent (fig. 4). The aquifer is bounded laterally and below by a till layer (model layer 2).

Table 4.--*Summary of hydrologic conditions in the simulation series*

[DIST-U, distance of well from stream (pumping from unconfined aquifer); DIST-C, distance of well from stream (pumping from confined aquifer); PUMP, pumping rate; DEPTH, depth of pumped aquifer; BLUFF, distance of well from bluff; WELL, well density; RECH, recharge rate; ✔, indicates condition specified in simulation; <, less than]

		Simulation Series and Number																													
		DIST-U					DIST-C					PUMP				DEPTH			BLUFF					WELL		RECH					
Condition		1	2	3	4	5	1	2	3	4	5	1	2	3	4	1	2	3	1	2	3	4	5	1	2	1	2	3	4	5	6
Aquifer type	Confined	--	--	--	--	--	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔
	Unconfined	✔	✔	✔	✔	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
Model layer	1	✔	✔	✔	✔	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	3	--	--	--	--	--	✔	✔	✔	✔	✔	✔	✔	✔	✔	--	--	--	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔
	5	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	7	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	9	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--
Pumping rate (gallon per minute)	75	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	150	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	300	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	--	--	--	--	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	--	--	--	--	--	--
	600	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	✔	✔	✔	✔	✔	✔
	1,200	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
Distance from stream segment 4 (feet)	<1,500	✔	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	1,500	--	✔	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	3,000	--	--	✔	--	--	--	--	✔	--	--	✔	✔	✔	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	4,500	--	--	--	✔	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
	6,000	--	--	--	--	✔	--	--	--	--	✔	--	--	--	--	✔	✔	✔	--	--	--	--	--	--	--	--	--	--	--	-	---
Distance from bluff (feet)	4,500	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--	--
	9,000	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--	--
	13,500	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--	--
	18,000	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--	--
	22,500	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--	--
Number of pumping wells	1	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	--	--	--	--	--	--	--	--
	8	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--	--
	16	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	✔	✔	✔	✔	✔
	480	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	--	--	--	--	--
Surface geology	Till	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	✔	--	--	✔	--
	Outwash	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	✔	--	✔
Recharge rate	Baseline	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	✔	--	✔	--	✔	--	--
	Reduced	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	--	✔	--	--	--
	Increased	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	✔	✔

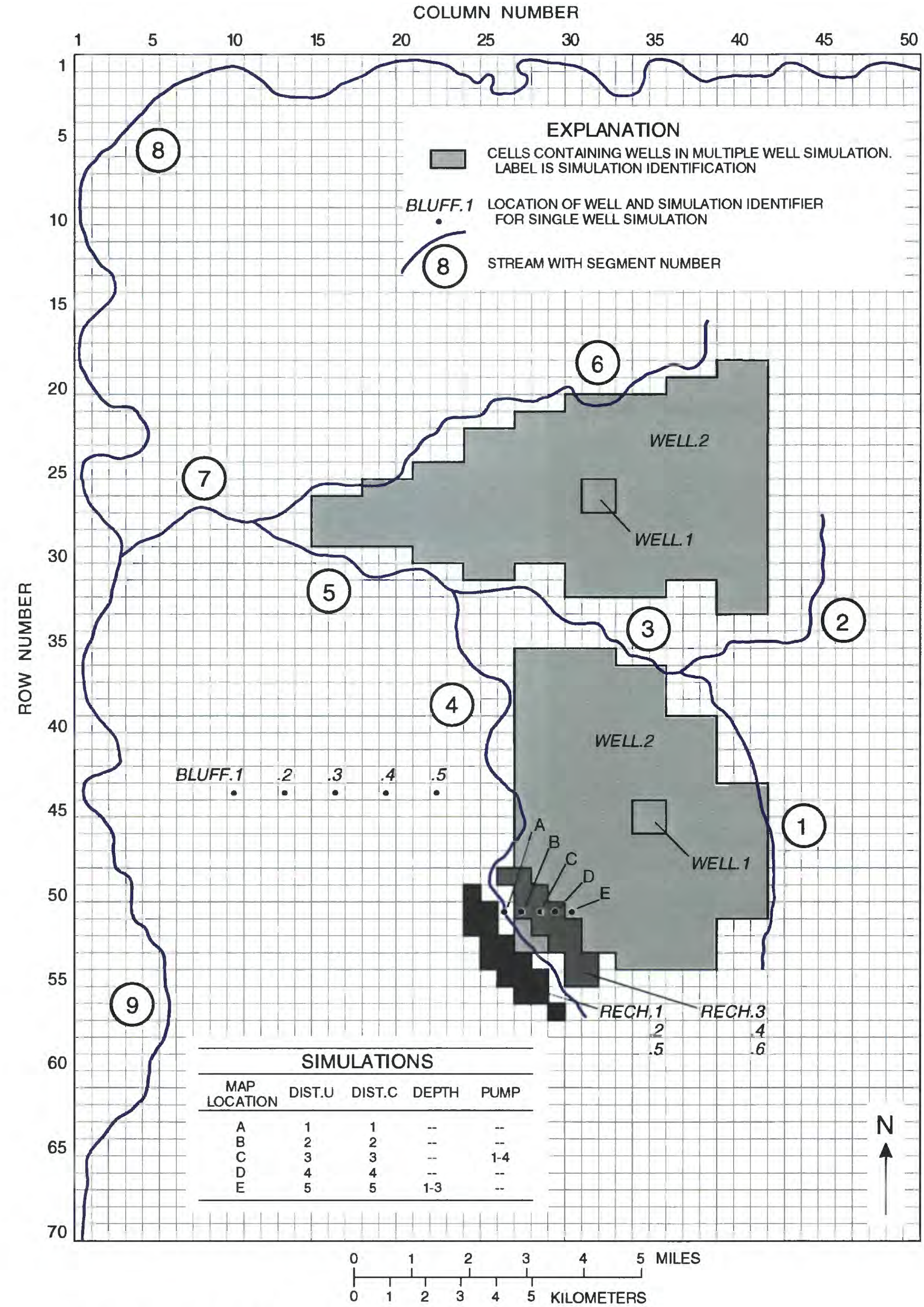

Figure 10.--Locations of wells in simulations.

These simulations demonstrate that a well pumping from a highly permeable outwash aquifer near a stream will, at equilibrium, capture most of its discharge from ground-water flow that would have discharged to the stream. With the well nearest the stream (less than 1,500 ft; DIST.U.1), about 98 percent of well discharge derives from capture of discharge to the streams and springs in the upper drift plain (fig. 11). Streamflow was reduced by more than 95 percent of well discharge in most of segment 4 and in all of segments 5, 7, and 9 (fig. 12). When the well is moved away from the stream, a larger percentage of pumping is derived from capture of discharge to the lower streams and springs. However, even when the well is at the maximum distance from the stream (6,000 ft; DIST.U.5), 85 percent of well discharge still derives from ground water that would have discharged to streams and springs on the upper drift plain (fig. 11).

Capture of discharge to streams and springs is shown by stream segment in figure 13. These results show that moving the well from less than 1,500 ft to 6,000 ft away from stream segment 4 reduced capture from that segment from 97 percent to 70 percent of well discharge. Converting these percentages to rates shows that the flow in segment 4 was reduced by 291 gal/min (97 percent of 300 gal/min) when the well was located near the stream, but that the reduction in flow was reduced to 210 gal/min (70 percent of 300 gal/min) when the well was located 6,000 ft from the stream. The decrease in capture from segment 4 was offset by increases in capture from all other stream segments, most notably from segments 1 through 3 (to 11.8 percent of well discharge, an increase of 10.9 percent) and segment 9 (to 12.4 percent of well discharge, an increase of 11.2 percent). The increase in capture from the segments 1 through 3 was mostly discharge to streams, whereas capture from segment 9 was mostly discharge to springs.

For many simulations, maps showing simulated water-level declines and capture of discharge to individual springs and stream reaches are presented. On these maps, total capture of ground-water discharge to each stream segment and cumulative downstream capture are listed for each stream segment (see fig. 14a). The segment and cumulative downstream captures are expressed as a percentage of the simulated well discharge. At the downstream end of stream segment 9 (where the stream exits the basin), the cumulative capture should be 100 percent for all simulations. In some cases, the cumulative capture is slightly more or less than 100 percent because of rounding errors in the model.

When the pumped well was located near the stream, the effects of the well on stream discharge was highly localized. When the well was located within 1,500 ft of the stream, nearly all well discharge derived from capture of discharge to one stream segment (number 4), and most of the captured discharge derived from the four stream cells nearest the well (fig. 14a). Simulated drawdown was less than 0.5 ft throughout most of the pumped aquifer (fig. 14a). When the well was located 6,000 ft from the stream, the cone of depression of the well must expand more before it captures enough natural discharge from other springs and streams in the basin to offset the discharge of the well. Figure 14b shows that a maximum drawdown of more than 10 ft was simulated in the cell containing the pumping well and that drawdowns of more than 0.5 ft were simulated in many cells. Figure 14b also shows that discharge to streams and springs was affected over a much larger area when the well was located at a greater distance from the stream. Comparison of the cumulative areas affected by simulated drawdown for the DIST.U.1 and DIST.U.5 simulations (fig. 15) confirms that the AOI was much larger when the well was farther from the stream. Simulated drawdown exceeded 0.1 ft over an area of about 21 mi^2 in simulation DIST.U.5. However, only about 2 mi^2 were affected by drawdowns greater than 0.1 ft when the well was located next to the stream. The stair-step shape of the cumulative-area curves in figure 15 is caused by the discontinuous nature of the uppermost unconfined outwash aquifer.

Pumping From a Confined Aquifer

In the DIST.C series, the effect of distance of the pumped well from the stream were investigated with the well pumping from the uppermost confined outwash aquifer (model layer 3). For the five simulations, the wells were placed in the same cells that were used in the DIST.U-series simulations (fig. 10) at distances ranging from less than 1,500 ft to 6,000 ft. from the stream. The pumping rate of the simulated wells (300 gal/min) were also the same as used in the DIST.U series. The principal difference between this series and the DIST.U series was that the well in the DIST.C series was pumping from a confined aquifer separated from the nearest stream by a low-permeability till layer. The thickness of the till near the well locations averages 25 ft; the vertical hydraulic conductivity of the till is 0.0025 ft/d throughout the basin.

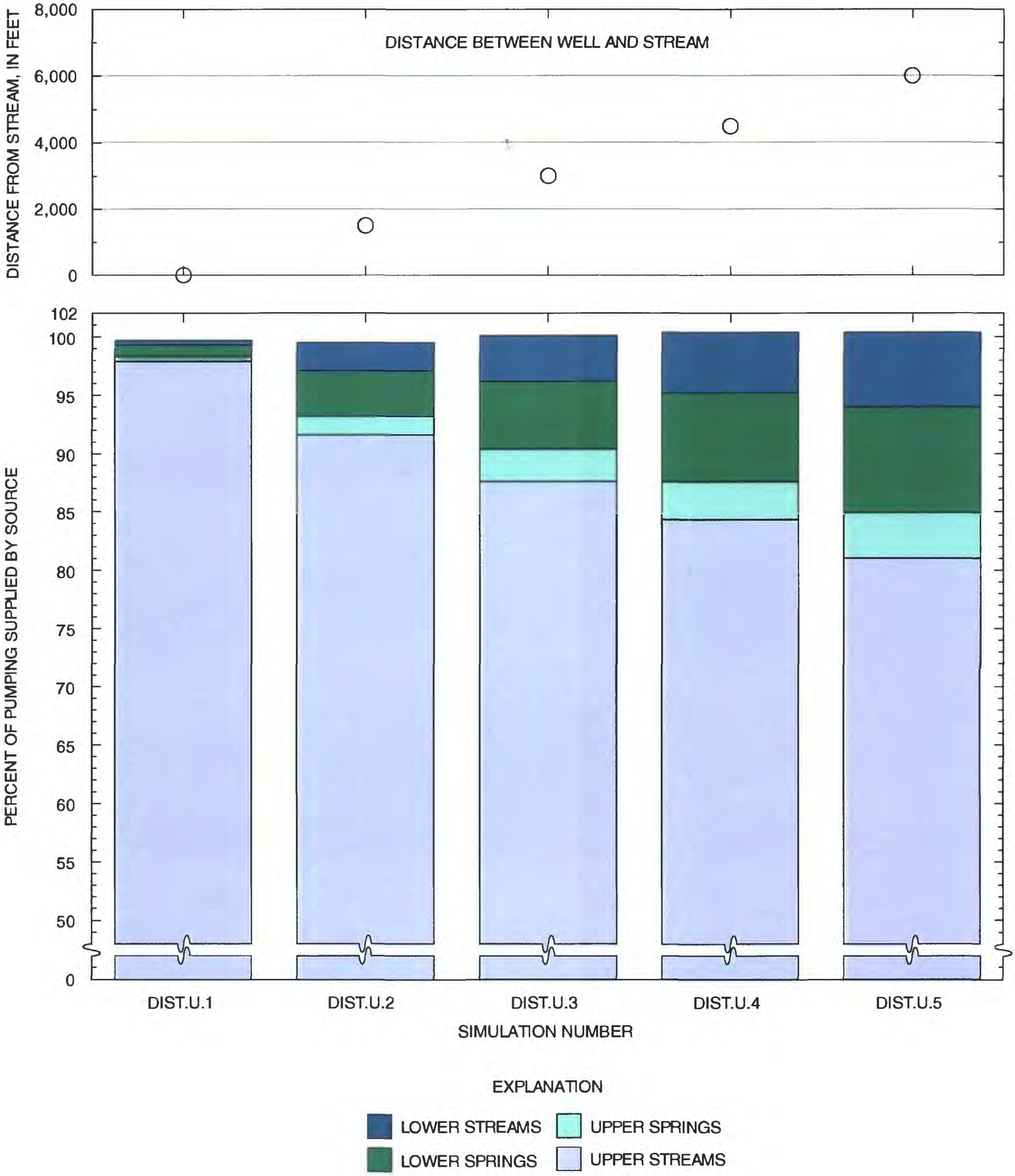

Figure 11.--Sources of water to a simulated well pumping 300 gallons per minute from the unconfined outwash aquifer (layer 1) at various distances from stream segment number 4 (see figure 10).

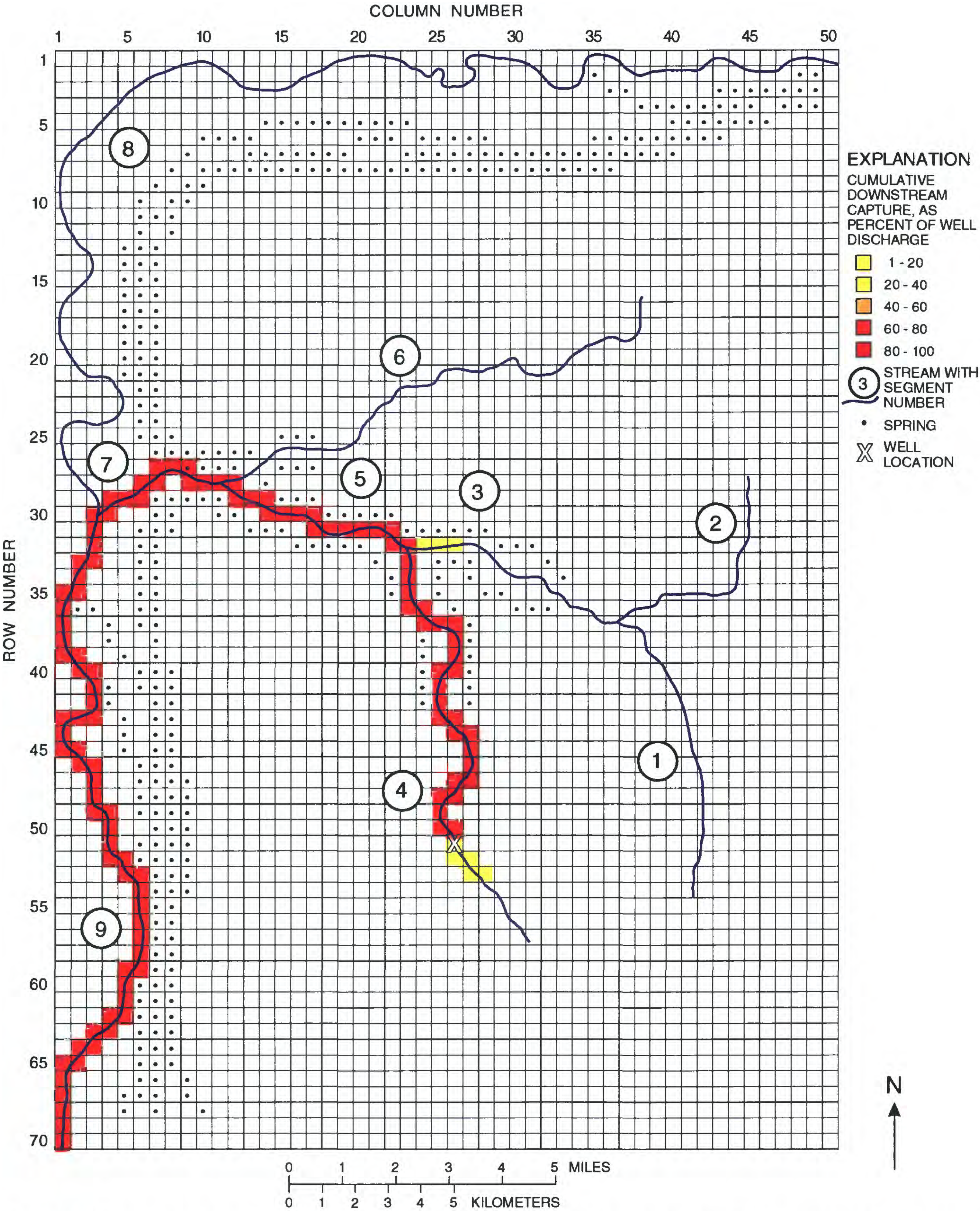

Figure 12.--Cumulative downstream capture of ground-water discharge to streams by a well located near a stream and pumping 300 gallons per minute from an unconfined outwash aquifer (layer 1).

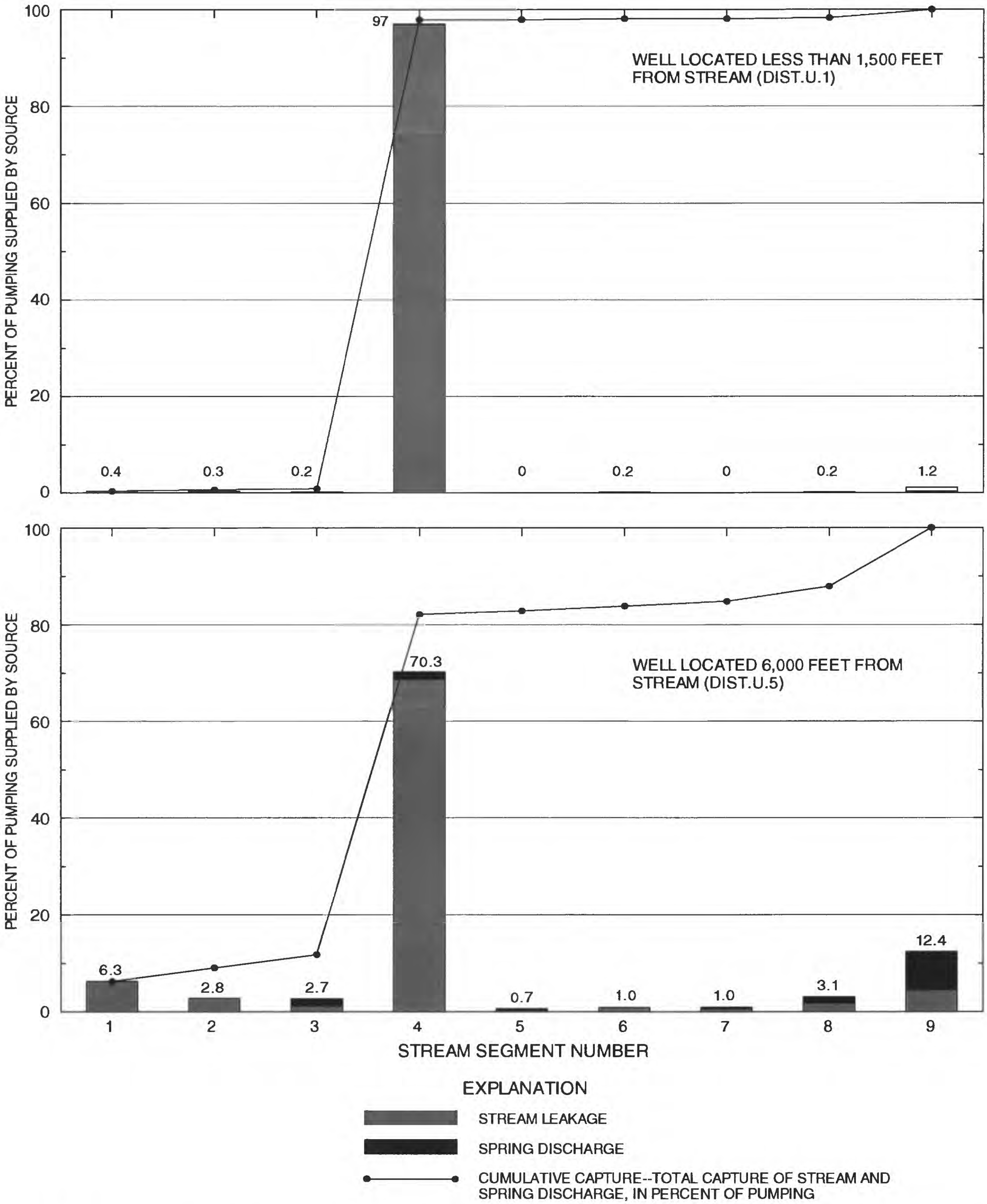

Figure 13.--Simulated capture of ground-water discharge to stream segments by a well pumping 300 gallons per minute from an unconfined outwash aquifer (layer 1).

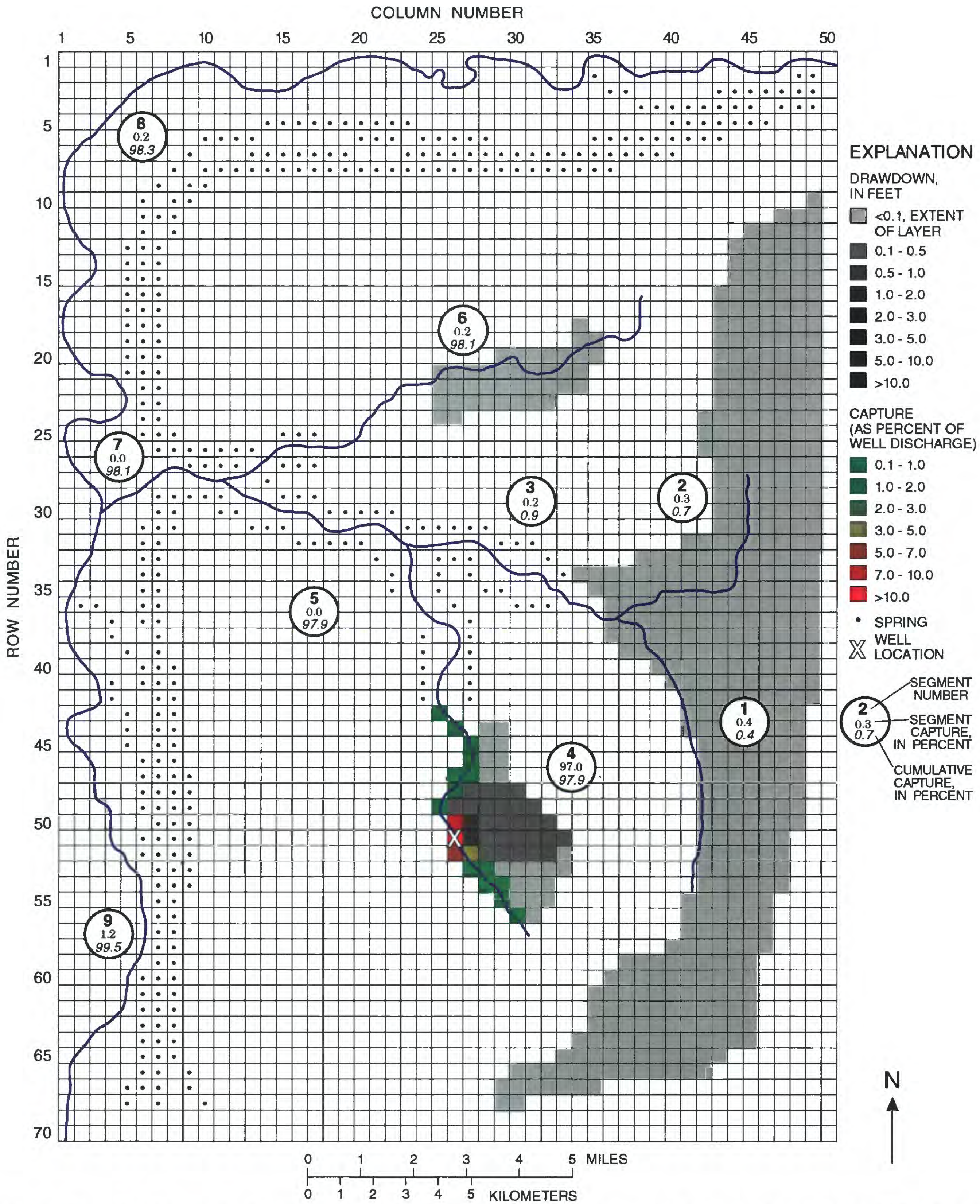

Figure 14a.--Simulated water-level declines and capture of discharge to streams and springs when pumping 300 gallons per minute from an unconfined aquifer (layer 1) with the well located near the stream.

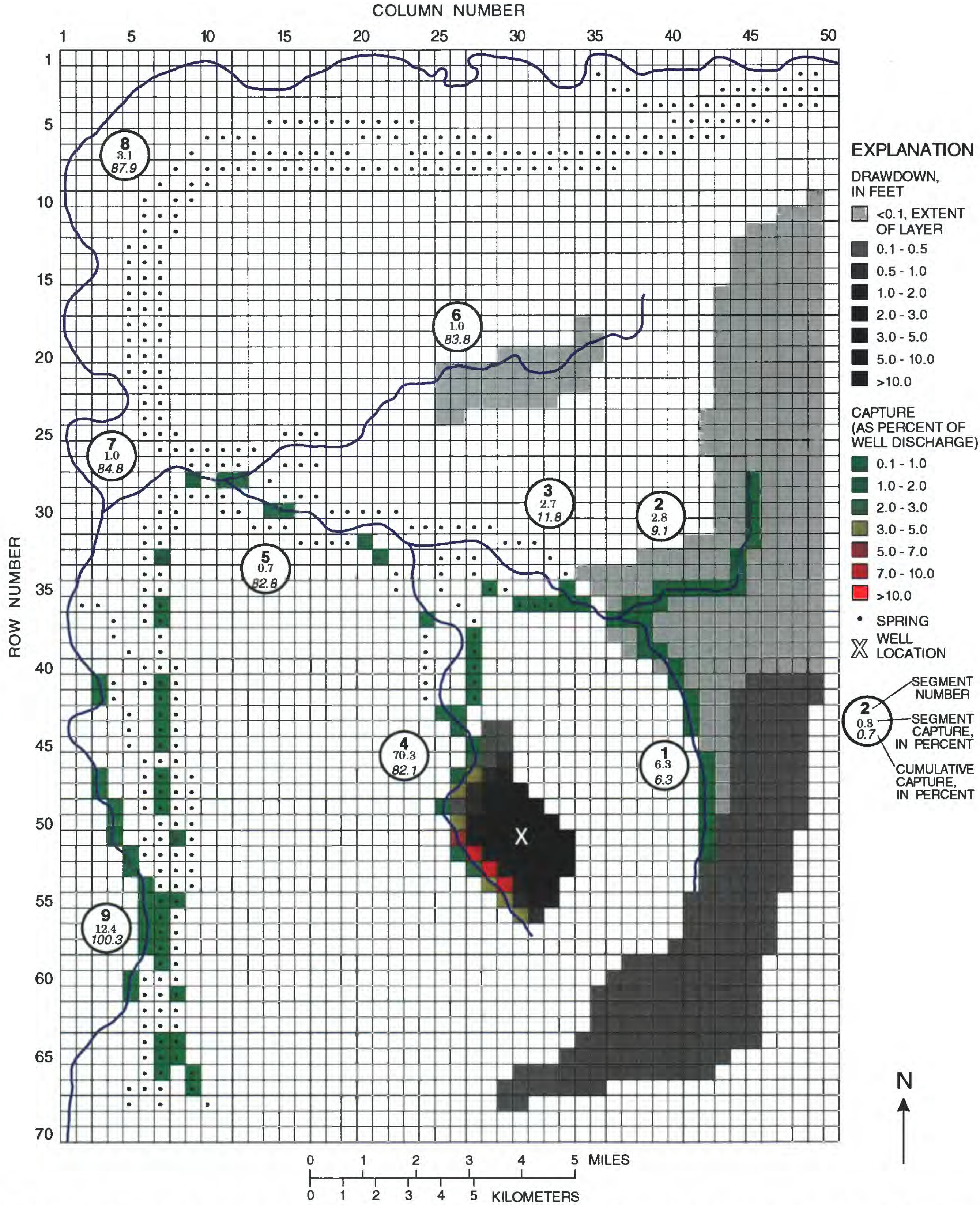

Figure 14b.--Simulated water-level declines and capture of discharge to streams and springs when pumping 300 gallons per minute from an unconfined aquifer (layer 1) with the well located far from the stream.

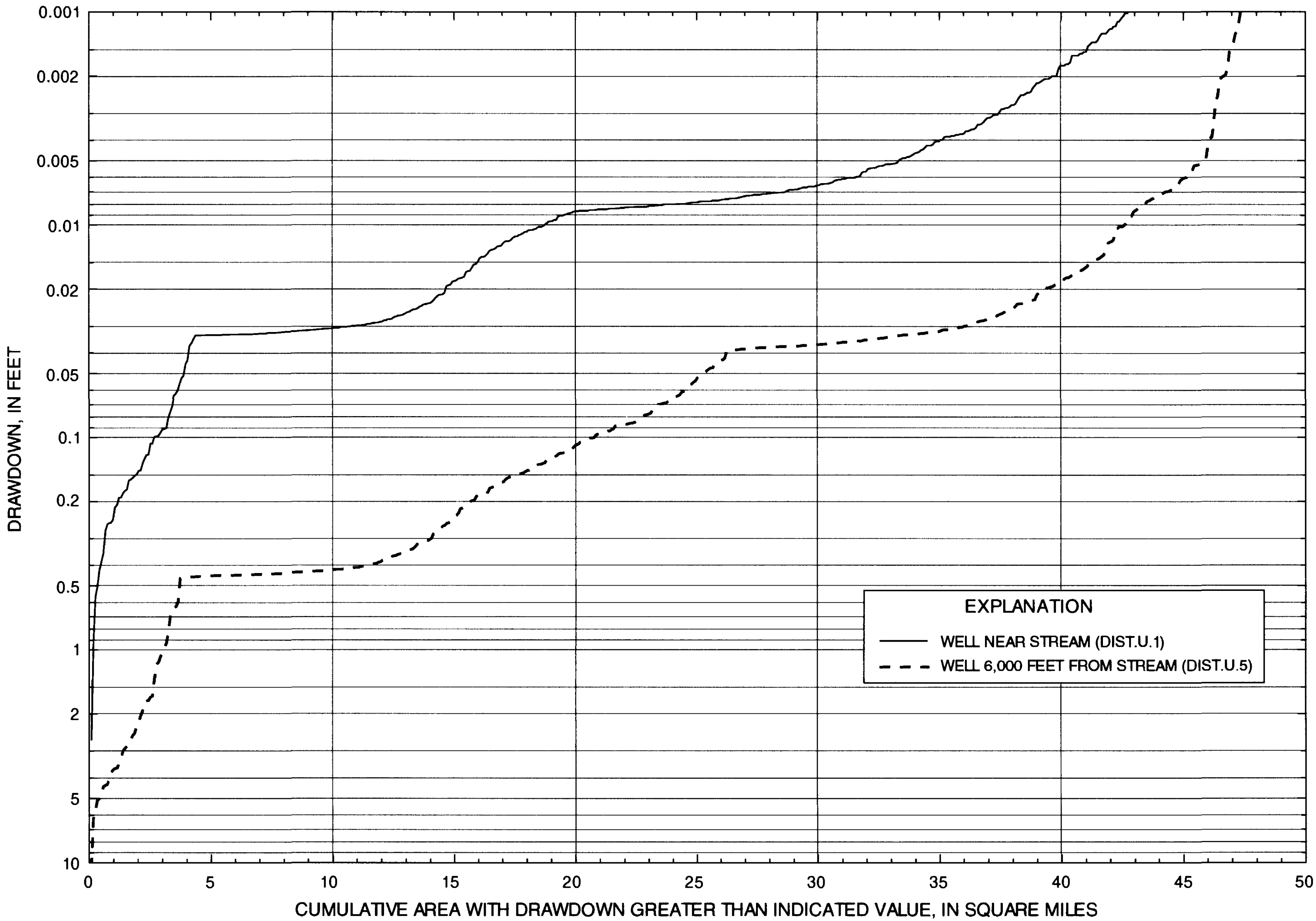

Figure 15.--Cumulative area affected by simulated drawdown in an unconfined outwash aquifer (layer 1) when pumping 300 gallons per minute from a well near the stream (DIST.U.1) and 6,000 feet from the stream (DIST.U.5). Cumulative areas are only shown for simulated drawdown greater than 0.001 feet; total area of aquifer is 47.5 square miles.

The presence of the low-permeability till between the well and the stream had a significant effect on the source of water to the well. More of the well discharge was derived from capture of flow to the streams and springs in the lower valley in these scenarios than in the corresponding scenarios for the unconfined aquifer (DIST.U). The percentage of well discharge captured from the lower valley streams was nearly constant at about 10 percent in all five simulations (fig. 16). Streamflow was reduced by about 50 percent of well discharge at the mouth of segment 4 (fig. 17). The percentage captured from discharge to the springs in the lower valley ranged from 20 percent when the well was nearest the stream (DIST.C.1) to 15 percent when the well was more than a mile to the east of the stream (DIST.C.5) (fig. 16). Spring discharge on the bluffs was more sensitive to the location of the pumped well than was stream discharge because many of the springs discharge directly from the pumped aquifer (model layer 3).

Comparison of simulations DIST.U.1 and DIST.C.1 shows the importance of the till layer in controlling the effect of a well on discharge to streams and springs. When the well pumped from the outwash aquifer directly adjacent to the stream (as in simulation DIST.U.1), about 98 percent of its discharge came from capture of discharge to streams and springs on the upper drift plain. In simulation DIST.C.1, however, a well pumped at the same location from an aquifer separated from the stream by a till layer derived only 70 percent of its discharge from the upper streams and springs (figs. 11 and 16). The effects of the well on the nearest stream reach were even more dramatically attenuated by pumping from the lower aquifer. Comparison of figures 13 and 18 shows that 97 percent of the well discharge in simulation DIST.U.1 was captured from segment 4, while only 51 percent was captured from segment 4 in simulation DIST.C.1. The difference of 46 percent is equivalent to 138 gal/min of the total 300 gal/min of well discharge. The reduction in capture from segment 4 by pumping from the confined aquifer was made up mostly by an increase in capture from segments 1 through 3 (12.3 percent of well discharge) and segment 9 (24.8 percent). Within the range of distances simulated, the distance of the well from stream segment 4 made a small difference in the distribution of capture among stream segments on the upper drift plain (fig. 18). However, because of the direction in which the well was moved, increase in distance from the stream increased capture from segment 1 through 3 and decreased capture from segment 9. Comparison of simulations DIST.U.1 and DIST.C.5 indicates a decrease in capture from segment 4 (44.2 percent of well discharge) offset mostly by increases in capture from segments 1 through 3 (17.3 percent) and segment 9 (18.7 percent).

Lateral distance from the stream to the well is less important than the presence or absence of a confining layer between the pumped well and the stream. This is clearly shown by comparing the maps of simulated drawdown and capture of discharge for simulations DIST.U.1 and DIST.C.1 (figs. 14a and 19). The effects of the pumped well were more broadly distributed within the basin when the well pumps from the confined aquifer. Also, the cumulative areas affected by drawdown were much greater for the simulations of pumping from the confined aquifer. For all five of the DIST.C-series simulations, approximately 95 mi^2 of the confined aquifer was affected by drawdown of more than 0.1 ft (fig. 20); when the well was placed in the unconfined aquifer, the maximum area experiencing more than 0.1 ft of drawdown was 21 mi^2 (fig. 15).

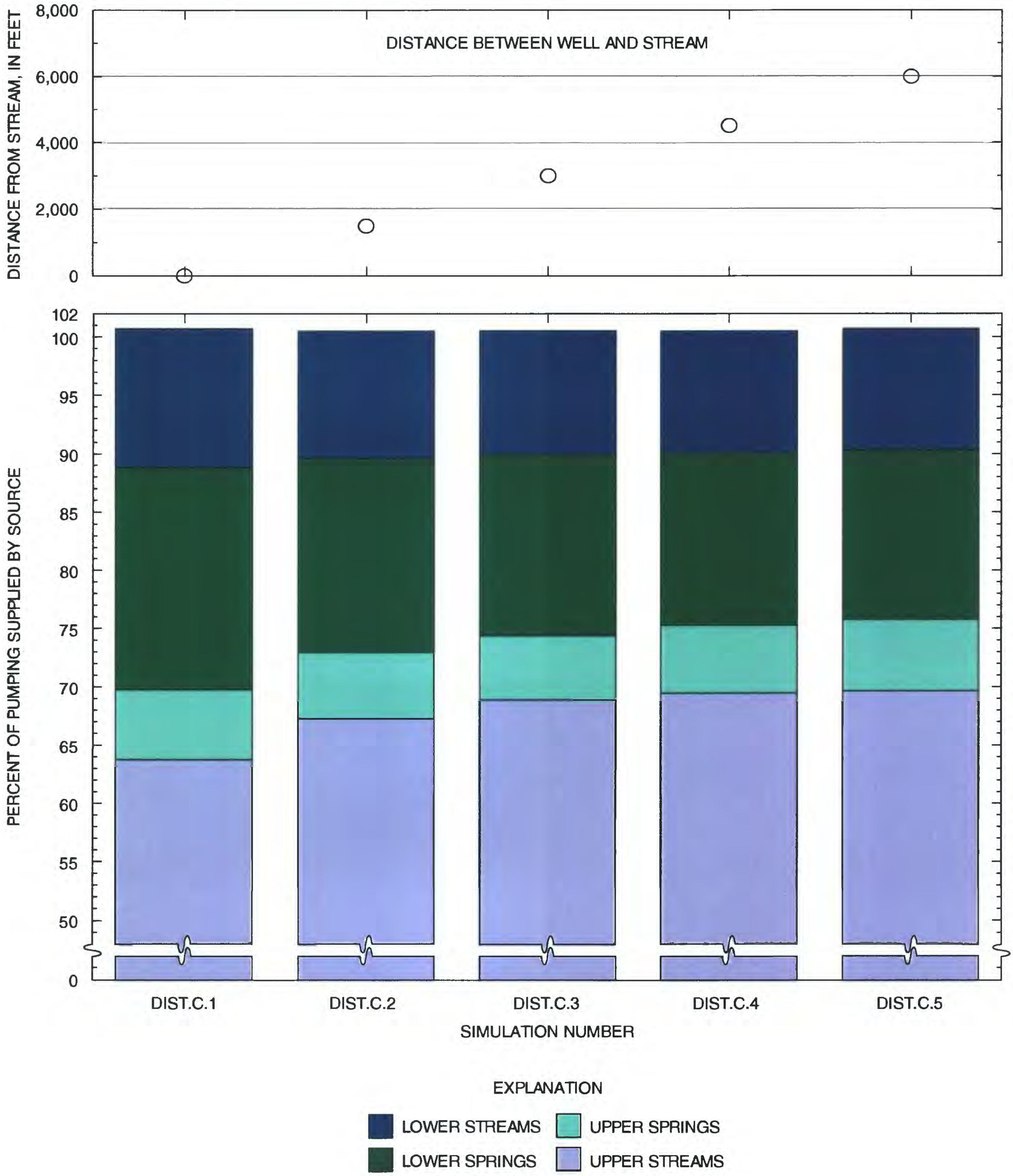

Figure 16.--Sources of water to a simulated well pumping 300 gallons per minute from the uppermost confined outwash aquifer (layer 3) at various distances from stream segment number 4 (see figure 10).

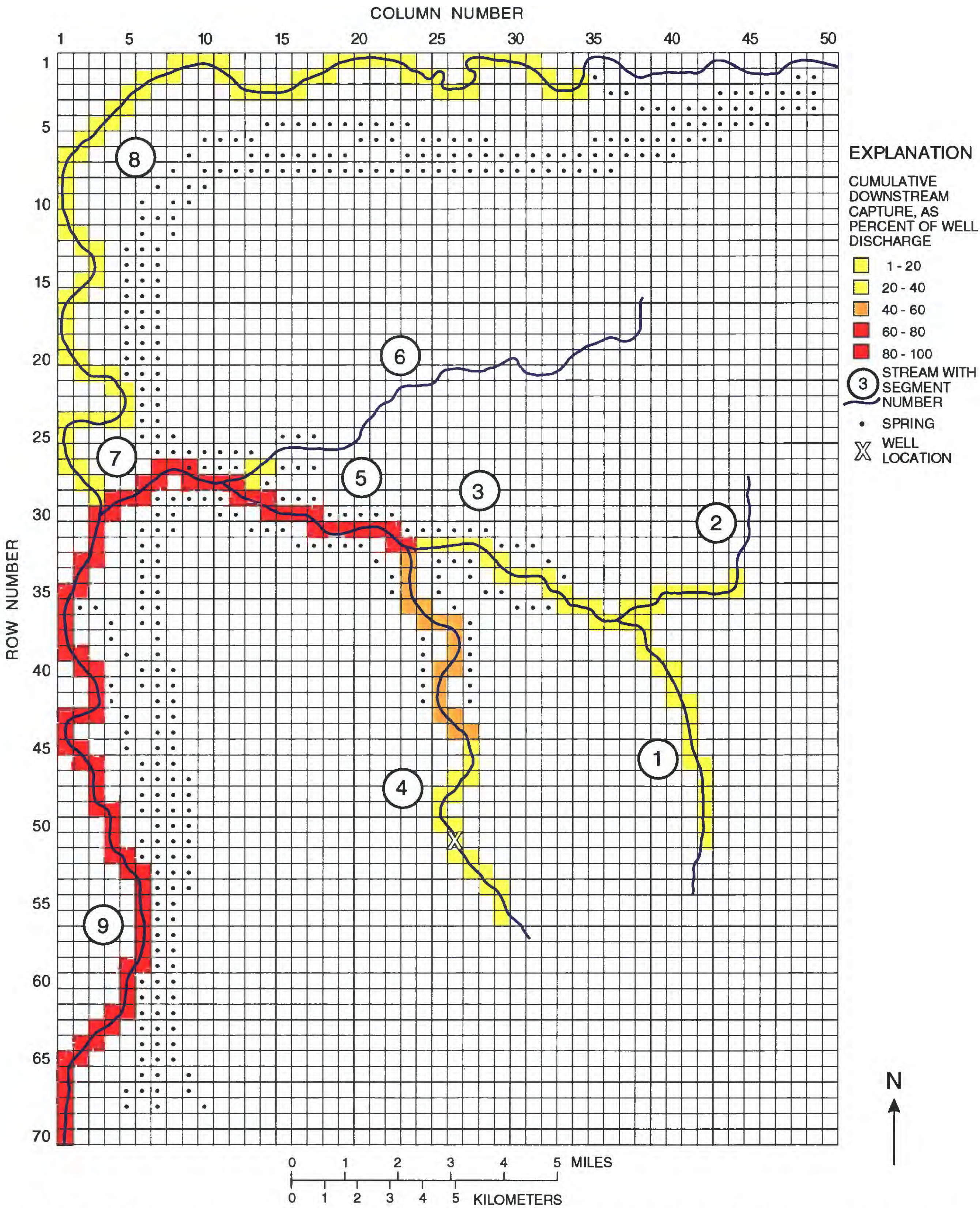

Figure 17.--Cumulative downstream capture of ground-water discharge to streams by a well located near a stream and pumping 300 gallons per minute from a confined outwash aquifer (layer 3).

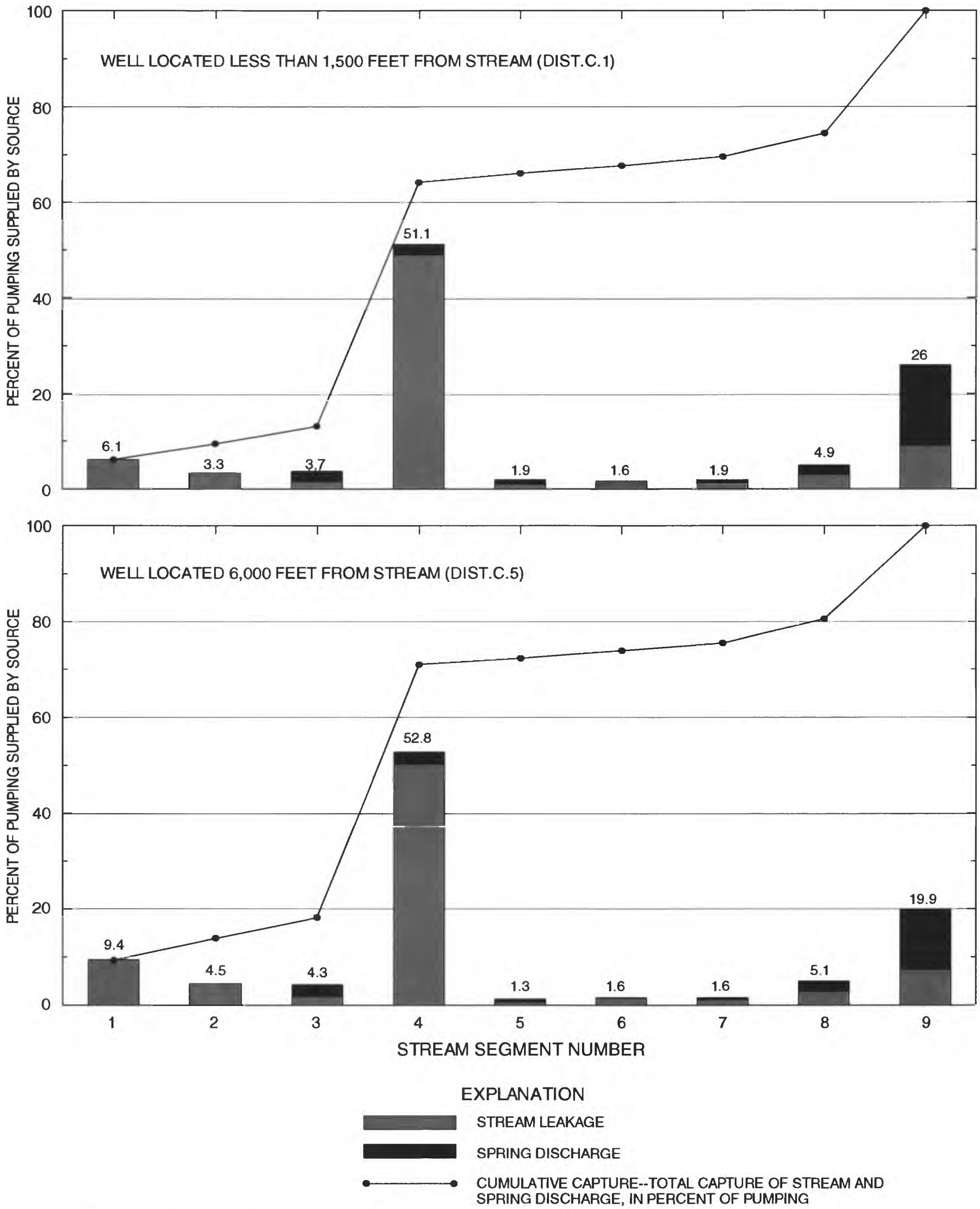

Figure 18.--Simulated capture of ground-water discharge to stream segments by a well pumping 300 gallons per minute from a confined outwash aquifer (layer 3).

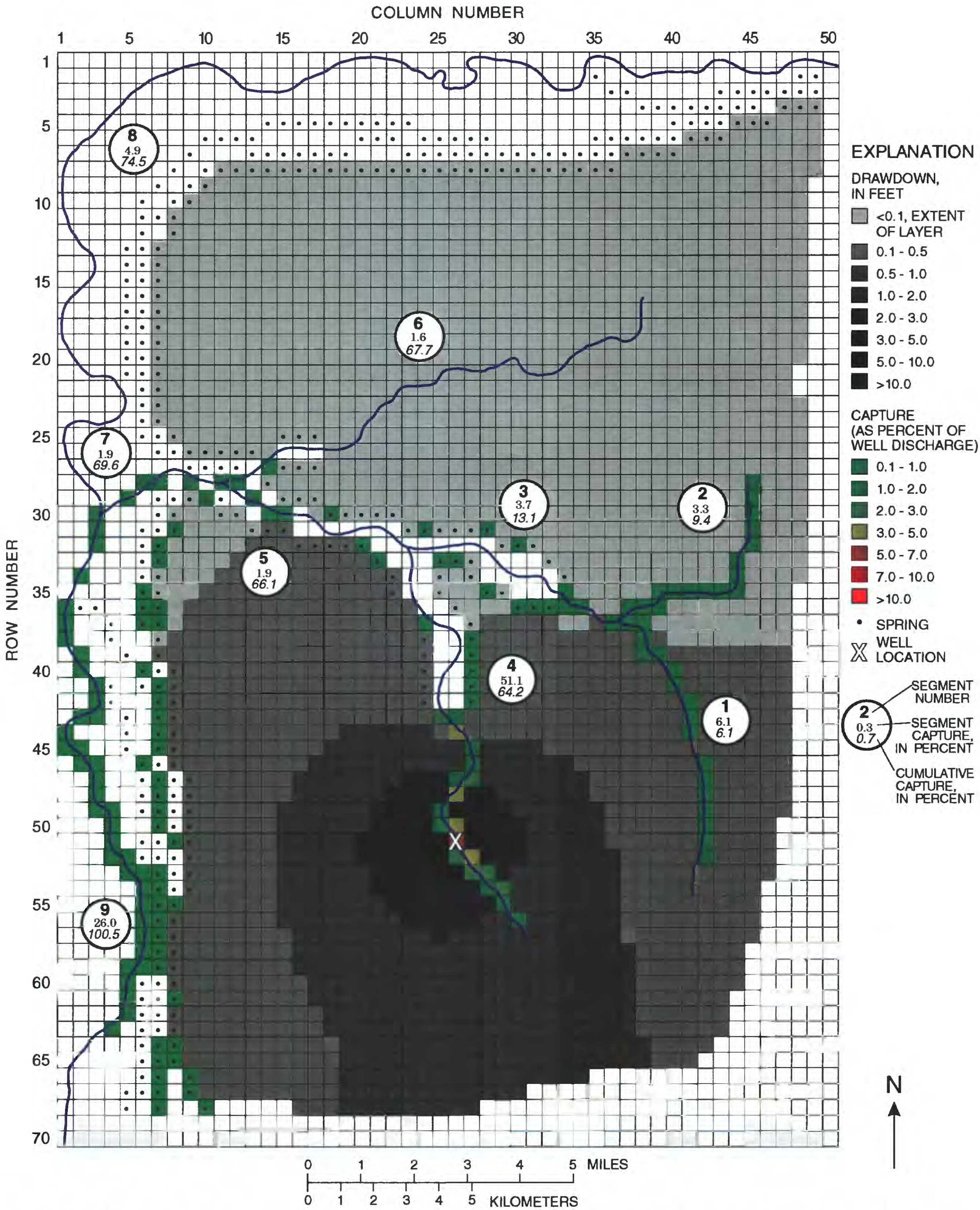

Figure 19.--Simulated water-level declines and capture of discharge to streams and springs when pumping 300 gallons per minute from a confined aquifer (layer 3) with the well located near the stream.

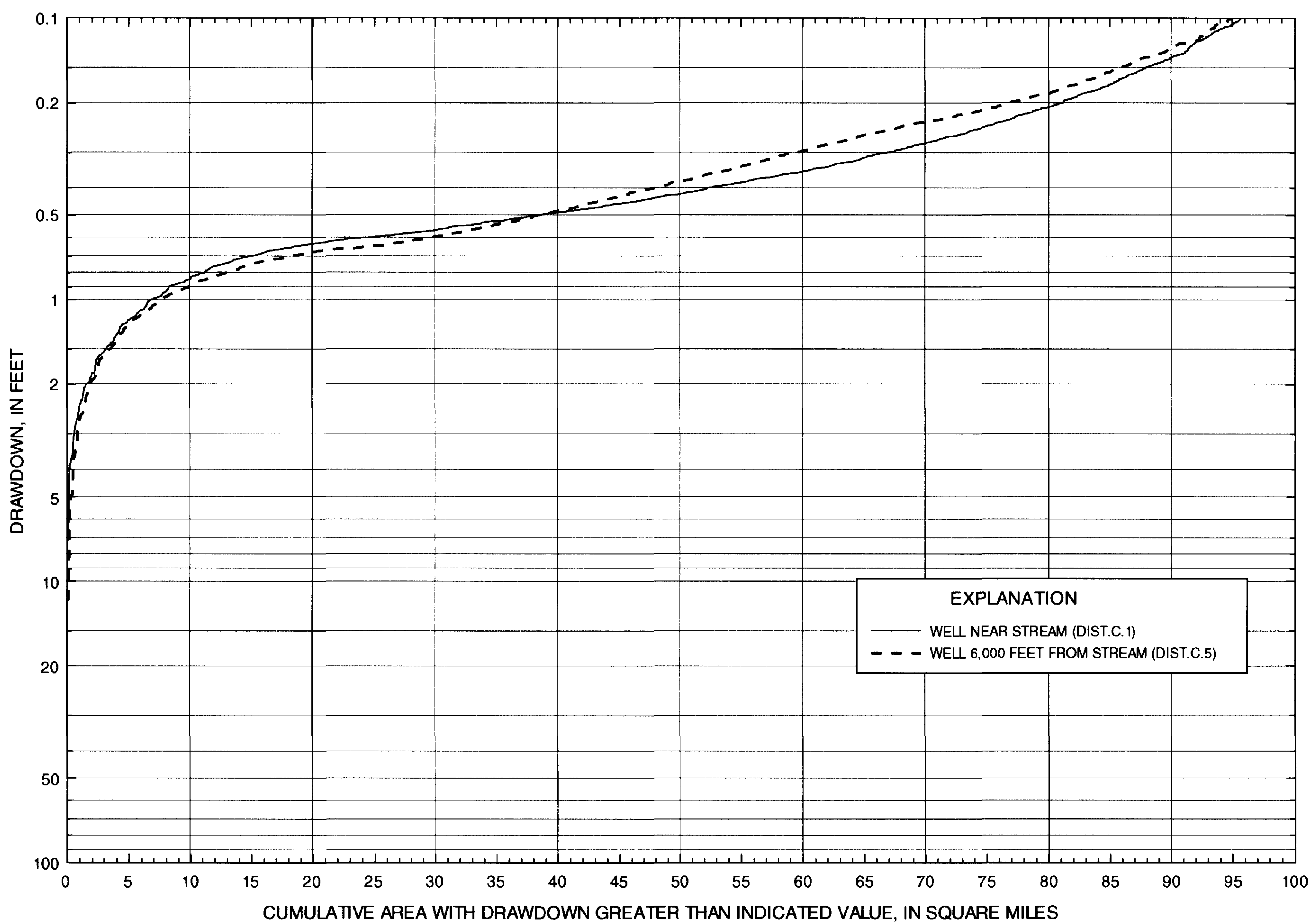

Figure 20.--Cumulative areas affected by simulated drawdown in a confined outwash aquifer (layer 3) when pumping 300 gallons per minute from a well near the stream (DIST.C.1) and 6,000 feet from the stream (DIST.C.5). Cumulative areas are shown only for simulated drawdown greater than 0.1 feet; total area of aquifer is 187 square miles.

Effects of Varying Pumping Rate

The effect of pumping rate on ground-water discharge to streams and springs was analyzed in the PUMP series with five simulations in which the pumping rate was varied from 75 gal/min to 1,200 gal/min while the location and depth of the well were held constant. The well was located approximately 3,000 ft from stream segment number 4 (fig. 10), and pumping was from the uppermost confined outwash aquifer (model layer 3).

Capture of natural discharge, expressed as a percentage of the well discharge, was nearly constant among the five simulations (fig. 21). Thus, simulations in this series showed that capture of natural discharge to streams and springs at any point was directly proportional to the pumping rate of the well. This result is, in part, an artifact of the assumptions used in constructing the model. Specifically, the transmissivity of a hydrogeologic layer does not vary with saturated thickness in this model. This simplification was based on the assumption that the saturated thickness would not change significantly due to any hydrologic stress imposed on the model. If the transmissivity of the aquifer layer were to change, then the response of the system would no longer be dependent only on the pumping rate and would not vary linearly with the pumping rate.

A large number of stream reaches and springs in the lower valley were affected by pumping in scenario PUMP.1 in spite of the low pumping rate of 75 gal/min (fig. 22a). Simulated drawdown in the pumped aquifer was between 2 and 3 ft in the cell containing the well but was less than 0.5 ft over most of the basin.

A few additional stream reaches and springs in the lower valley were affected when the pumping rate was increased to 1,200 gal/min (simulation PUMP.4). Also, the shape of the cone of depression of the well began to show the effects of aquifer boundaries at the higher pumping rate (fig. 22b). The effects are greatest to the south and southeast of the well where the cone of depression reached the bedrock boundary; since no flow can be induced across this boundary, drawdown was more severe in this area. The AOI for pumping rates of 75 and 1,200 gal/min were 46 mi^2 and 143 mi^2, respectively (fig. 23). The effect of boundaries to the aquifer on the AOI are evident in figure 23, which shows a drawdown anomaly at the point where the cumulative area reached 100 mi^2.

Effects of Varying Depth of the Pumped Aquifer

The purpose of the DEPTH series of simulations was to evaluate the effect of the depth of the pumped aquifer on capture of ground-water discharge to streams and springs. In this series of simulations, the well pumped from successively deeper model layers at a location 6,000 ft east of stream segment 4 (fig. 10). The depth to the center of the pumped aquifer ranged from about 20 ft to 250 ft in the five simulations. The pumping rate in each simulation was 300 gal/min. Simulations DIST.U.5 and DIST.C.5 are part of this series because they simulated wells pumped from layers 1 and 3, respectively; simulations DEPTH.1, DEPTH.2, and DEPTH.3 simulated wells pumped from layers 5, 7, and 9 (see table 4).

When the well pumped from the unconfined outwash aquifer (model layer 1), 85 percent of well discharge came from capture of discharge to streams and springs on the upper drift plain. Simulation of pumping from successively deeper aquifers at the same location captures successively greater percentages of discharge to streams and springs in the lower valley (fig. 24). Pumping from layer 9 of the model increased the capture of discharge to lower streams and springs from 15 percent to 46 percent (fig. 24), with the increase about equally divided between capture of stream and spring flows. Capture of spring discharge on the upper drift plain also increased from about 4 percent when the well tapped the shallow aquifer to 6 to 8 percent when the well tapped deeper aquifers.

The effect of well depth on natural discharge to streams and springs is even more marked in individual stream segments. The shallow well (simulation DIST.U.5) captured 70 percent of its discharge from stream segment 4 on the upper drift plain and only 12 percent from segment 9 in the lower valley (fig. 25). The deepest well (pumped from layer 9; simulation DEPTH.3) captured only 22 percent of its discharge from segment 4; the decrease of 48 percent (144 gal/min) was partly offset by increases in the capture from segment 9 of 24 percent (72 gal/min) and from segments 1 through 3 of 12 percent (36 gal/min).

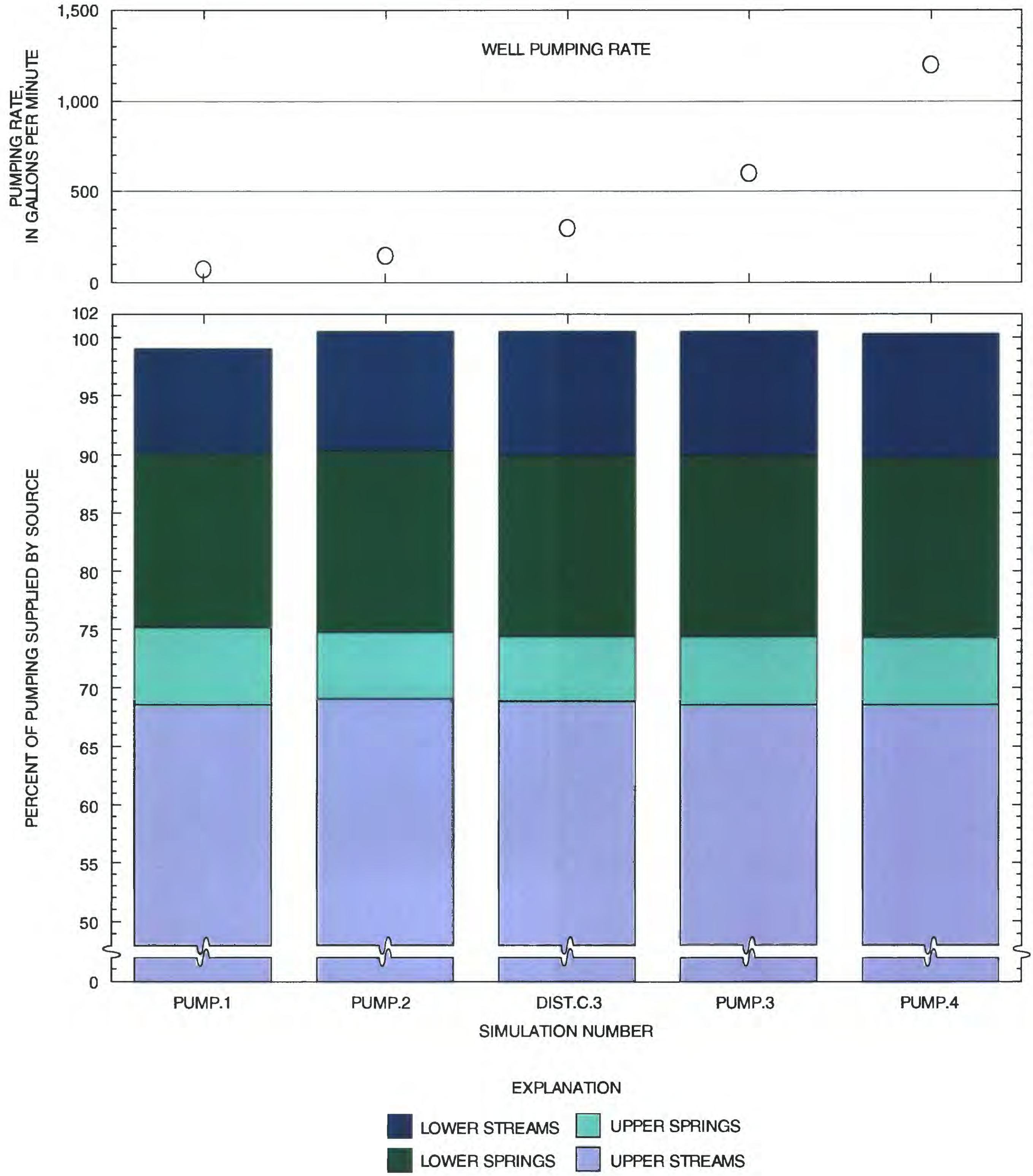

Figure 21.--Sources of water to a simulated well located 3,000 feet from the stream and pumping from a confined aquifer (layer 3) at rates ranging from 75 to 1,200 gallons per minute.

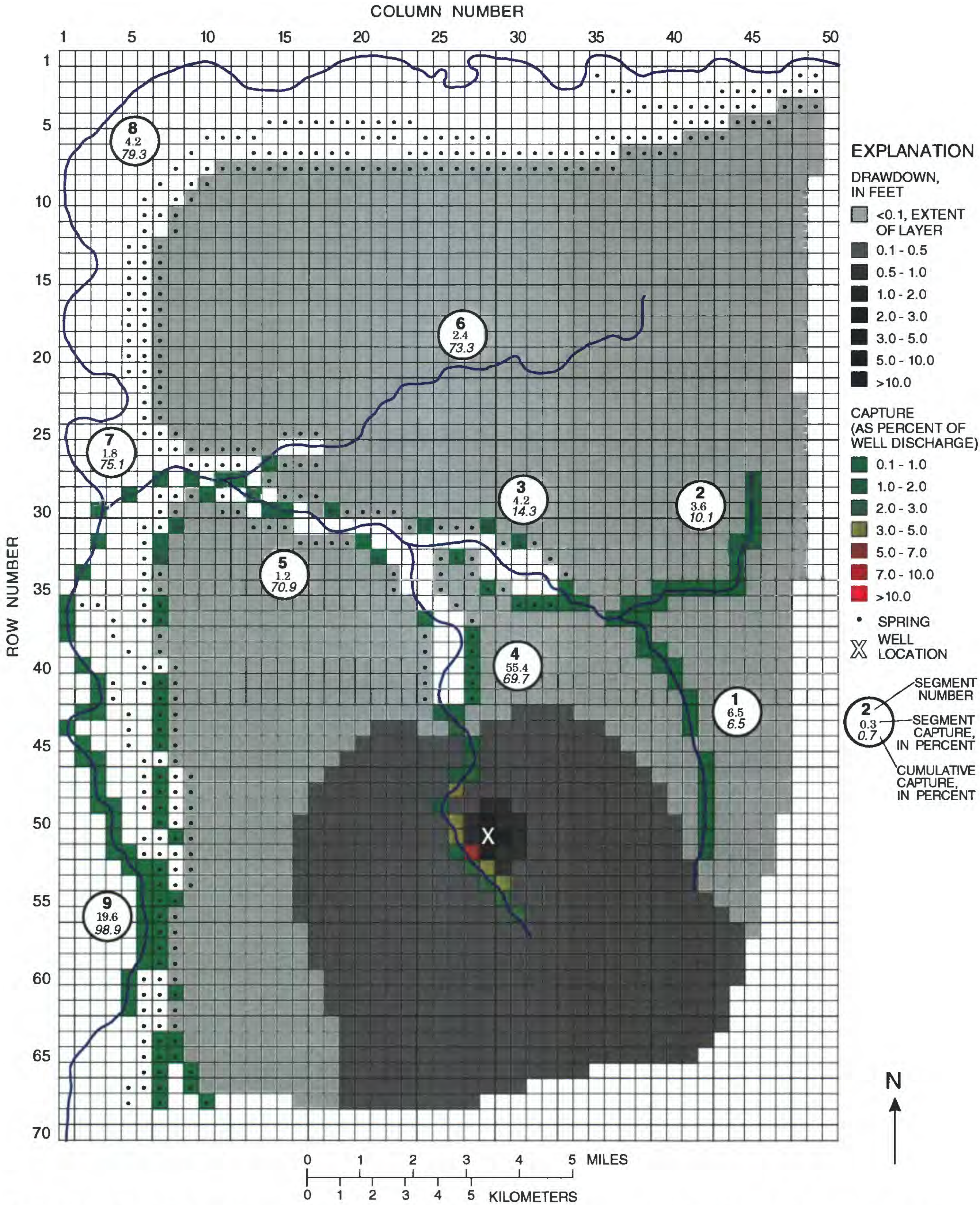

Figure 22a.--Simulated water-level declines and capture of discharge to streams and springs when pumping from a confined aquifer (layer 3) at 75 gallons per minute.

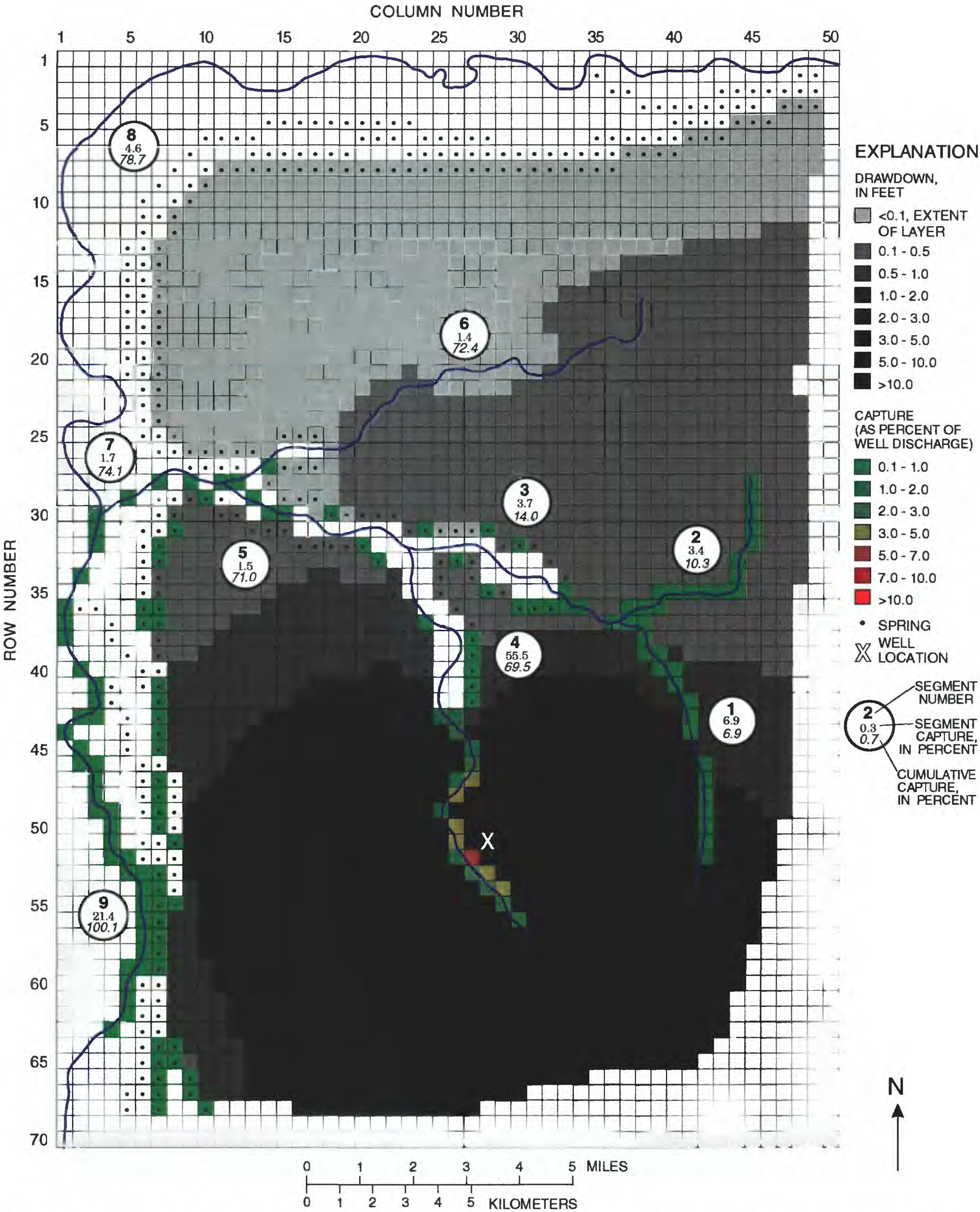

Figure 22b.--Simulated water-level declines and capture of discharge to streams and springs when pumping from a confined aquifer (layer 3) at 1,200 gallons per minute.

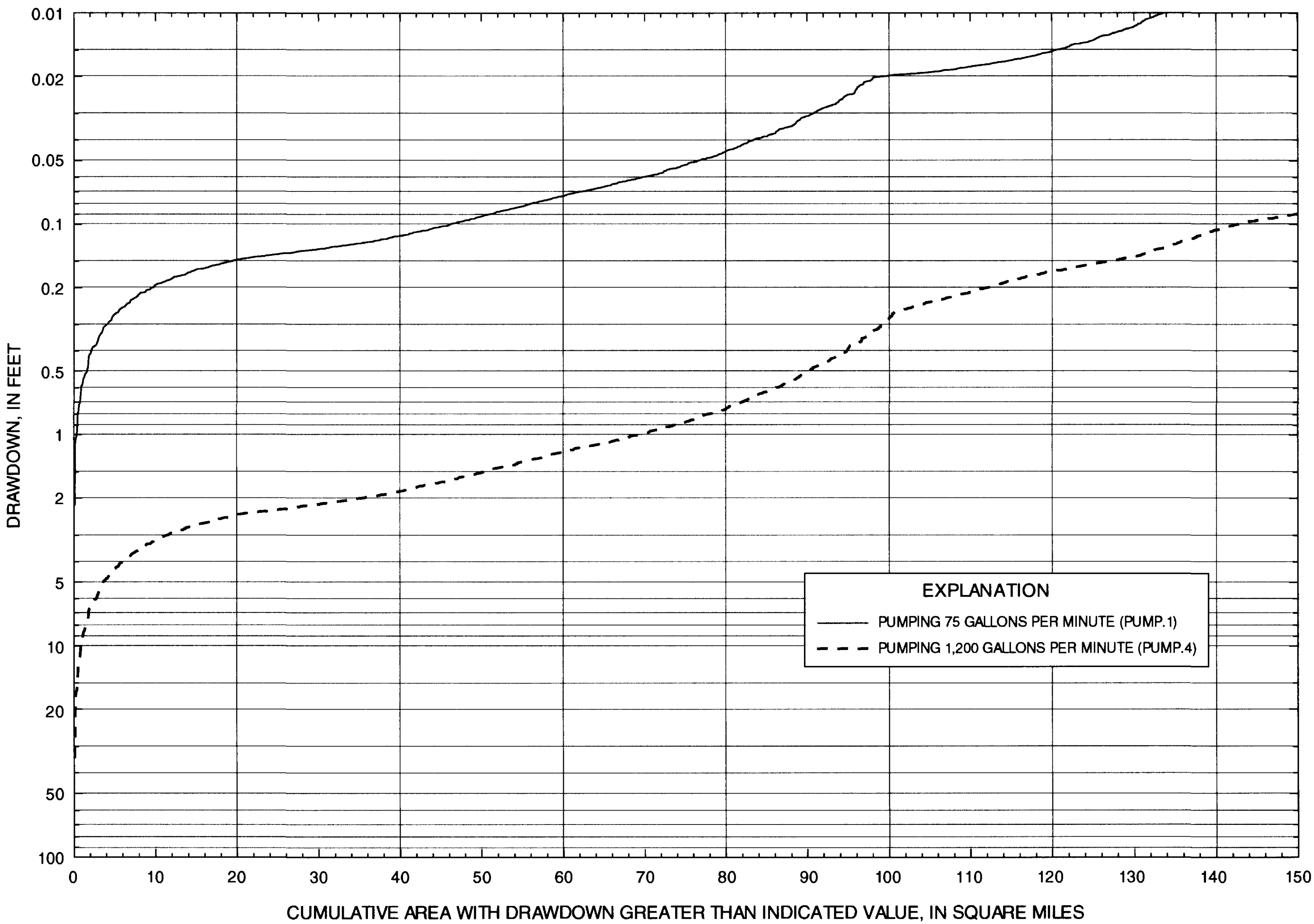

Figure 23.--Cumulative areas affected by simulated drawdown in a confined outwash aquifer (layer 3) when pumping from a well 3,000 feet from the stream (figure 10) at rates of 75 and 1,200 gallons per minute. Cumulative areas are shown only for simulated drawdown greater than 0.01 feet; total area of aquifer is 187 square miles.

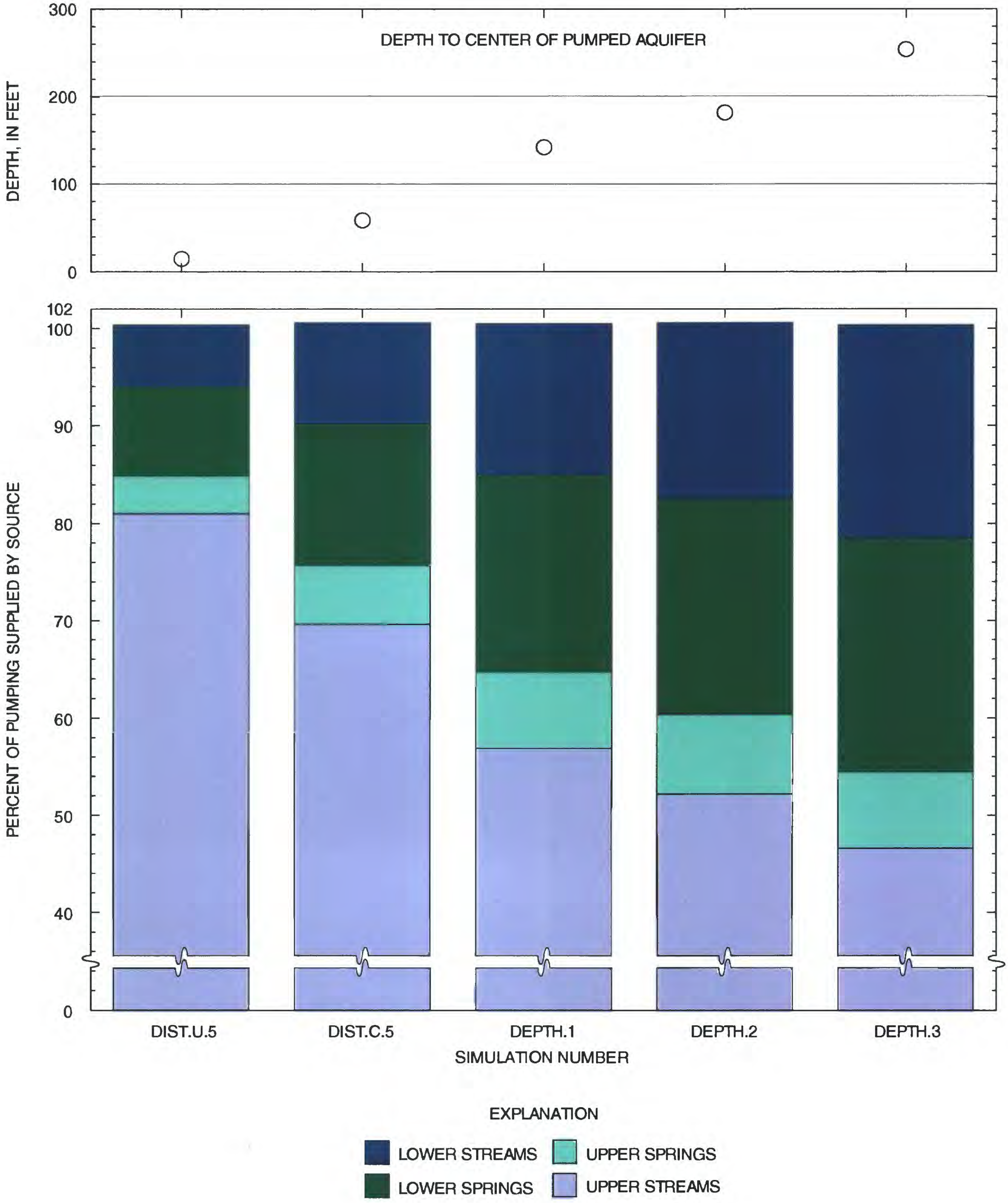

Figure 24.--Sources of water to a simulated well pumping 300 gallons per minute from aquifers at various depths.

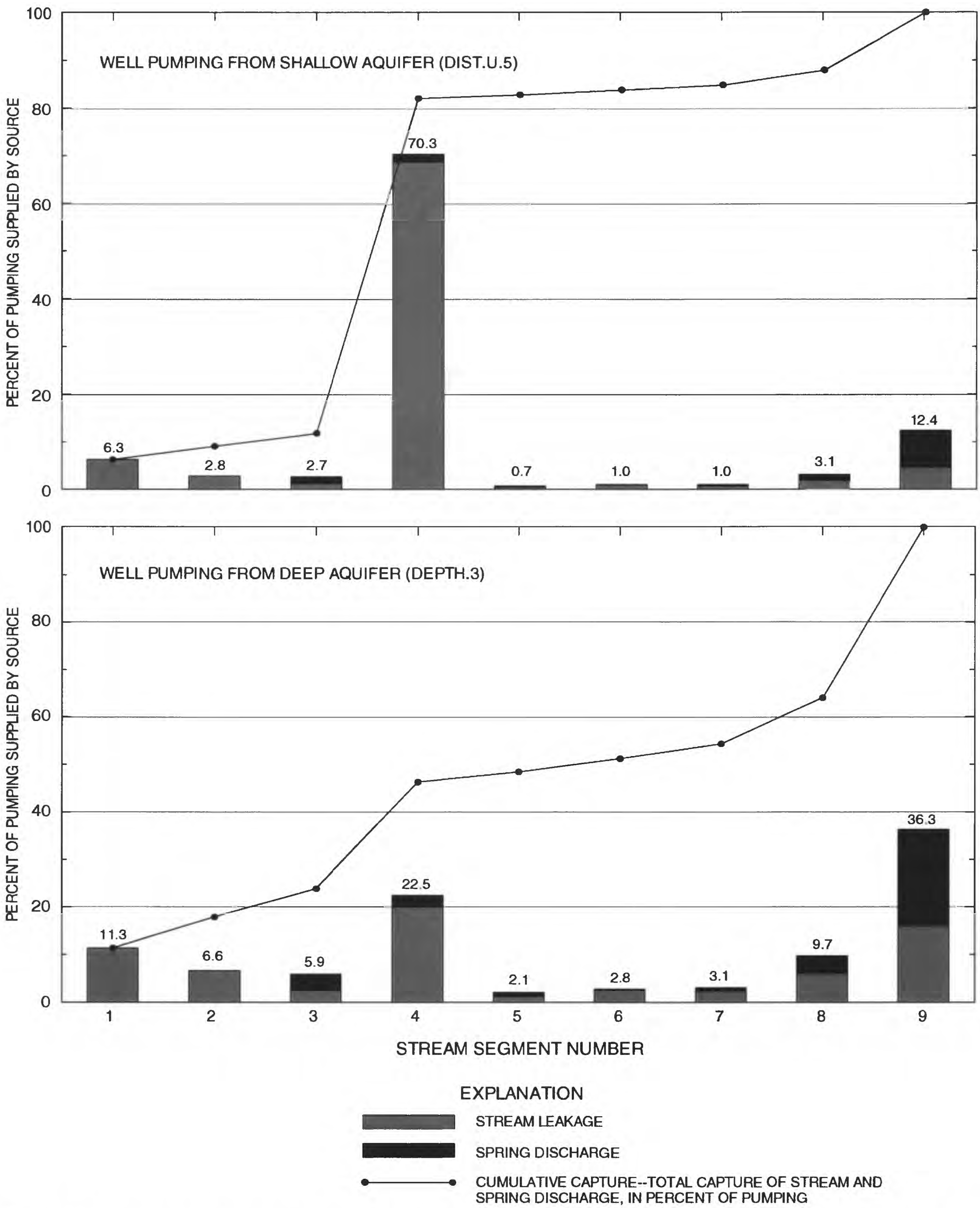

Figure 25.--Simulated capture of ground-water discharge to stream segments by a well pumping 300 gallons per minute from shallow (layer 1) and deep (layer 9) outwash aquifers.

The maps of capture and drawdown for pumping from layer 1 (simulation DIST.U.5, figure 14b), layer 3 (simulation DIST.C.5, figure 26a), and layer 9 (simulation DEPTH.3, figure 26b) show that deepening the well spreads the effects of pumping over a larger area and reduces the magnitude of effects on individual stream reaches and springs. Very few stream reaches or springs contributed more than 1 percent of the water discharged from the pumped well in simulation DEPTH.3, and none contributed more than 2 percent (fig. 26b). The maximum drawdown and cumulative areas of drawdown for wells pumped from layers 3 and 9 are similar; drawdown in the pumped cells was about 10 to 11 ft for each simulation and the AOI ranged from 95 to 116 mi^2 (fig. 27).

Effect of Varying Distance from a Bluff

The bluffs that form the boundary between the upper drift plain and the lower valley are important hydrologic, as well as physiographic, features. The purpose of the BLUFF series of simulations was to determine the effect that proximity of a well to the bluff has on capture of ground-water discharge to streams and springs. In the five simulations made for this series, the well was placed at successively greater distances from the bluff along a line perpendicular to the bluff. Well locations for the five simulations are shown in figure 10. As the distance between the well and the bluff increased, the distance between the well and stream segment 4 decreased. The pumping rate was 300 gal/min for each simulation. The pumped aquifer was the uppermost confined outwash aquifer (model layer 3).

The most obvious effects were on the capture of natural discharge to the springs on the bluff and on the discharge to streams on the upper drift plain. A well placed within 3,000 ft of the bluff (fig. 10) captured 50 percent of its discharge (150 gal/min) from springs on the bluff while capturing only 26 percent from the upper streams (simulation BLUFF.1, figure 28). As the well was moved east from the bluff in successive simulations, the capture from springs on the bluff was replaced by capture from the upper streams. At the location farthest from the bluff (closest to stream segment 4), the percentages captured from springs and from the upper streams were reversed, with about 25 percent coming from the springs on the bluff and 51 percent from the upper streams (simulation BLUFF.5, figure 28). These simulations show that the proximity of the well to discharge areas controls the amount of well discharge derived from capture of ground water that would have discharged to those areas. The percentages of well discharge captured from the lower streams and the upper springs changed little with the distance between the bluff and the well.

The stream segments most affected in this series were segment 4 on the upper drift plain and segment 9 in the lower valley. Only 30 percent (90 gal/min) of well discharge came from capture from segment 9 when the well was located over 4 mi east of the bluff, but 60 percent (180 gal/min) came from this segment when the well was nearest the bluff (fig. 29). Most of the 30-percent increase in discharge to segment 9 that came from moving the well east was offset by the 26-percent decrease in discharge to segment 4 (fig. 29). Had the well been placed in a deeper layer, it is likely that the effects on discharge to springs on the bluff and to streams in the lower valley would have been even more pronounced.

The effect of the distance between the well and the bluff on the cone of depression of the well was minor for this series of simulations. Comparison of the cones of depression for simulations BLUFF.1 and BLUFF.5 (figs. 30a and 30b) indicates that their shapes are affected by the aquifer boundaries, but the overall area and the magnitude of drawdown are very similar for these simulations. The cumulative areas of drawdown shown on figure 31 indicate that simulated drawdown changes little with the distance of the well from the bluff. The AOI for these simulations ranged from 87 to 96 mi^2.

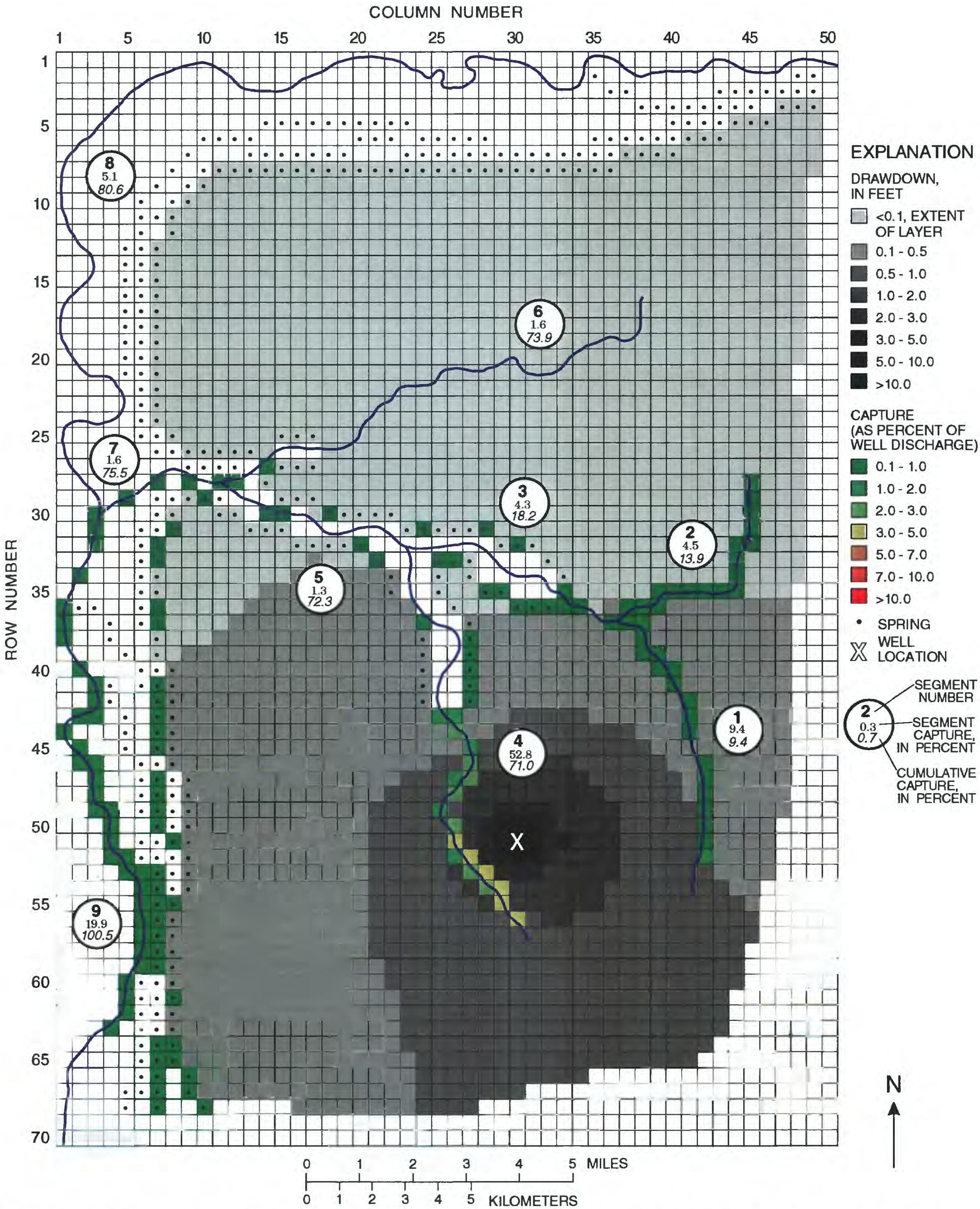

Figure 26a.--Simulated water-level declines and capture of discharge to streams and springs when pumping from a shallow (59 feet, layer 3) aquifer.

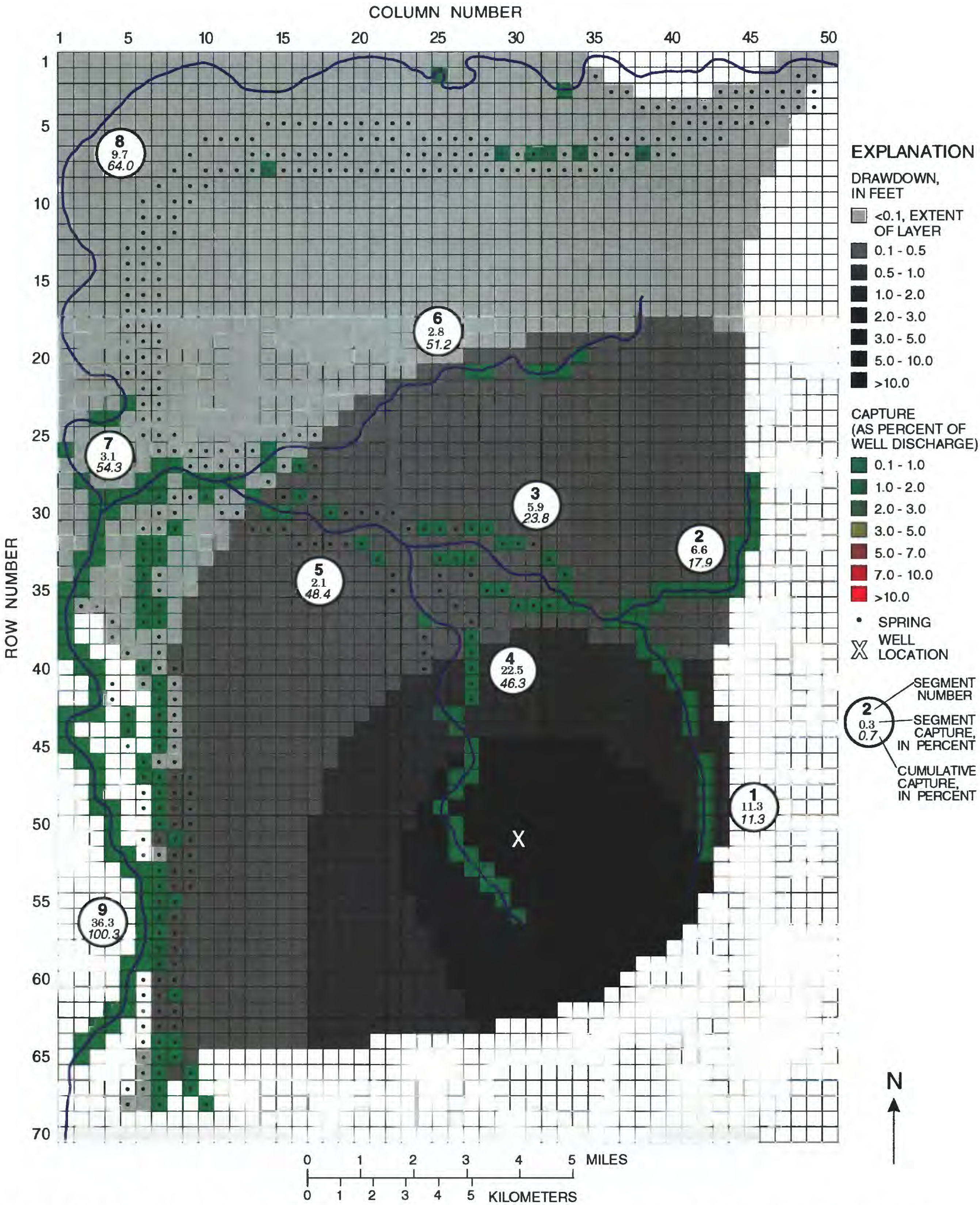

Figure 26b.--Simulated water-level declines and capture of discharge to streams and springs when pumping from a deep (254 feet, layer 9) aquifer.

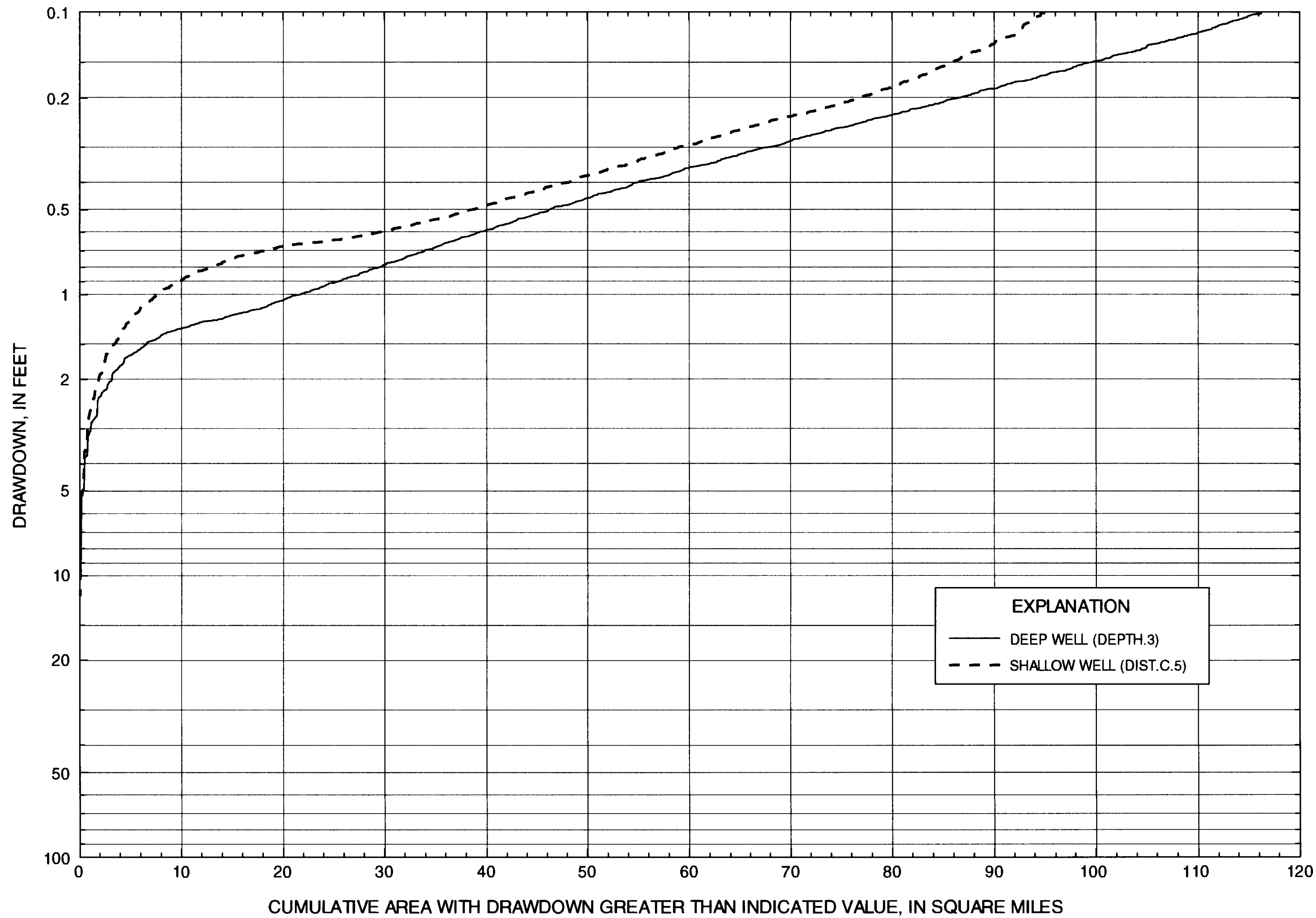

Figure 27.--Cumulative areas affected by simulated drawdown when pumping 300 gallons per minute from shallow (layer 3) and deep (layer 9) wells 6,000 feet from the stream. Cumulative areas for layers 3 and 9 are only shown for simulated drawdown greater than 0.1 feet; total area of layer 3 aquifer is 187 square miles, area of layer 9 aquifer is 207 square miles.

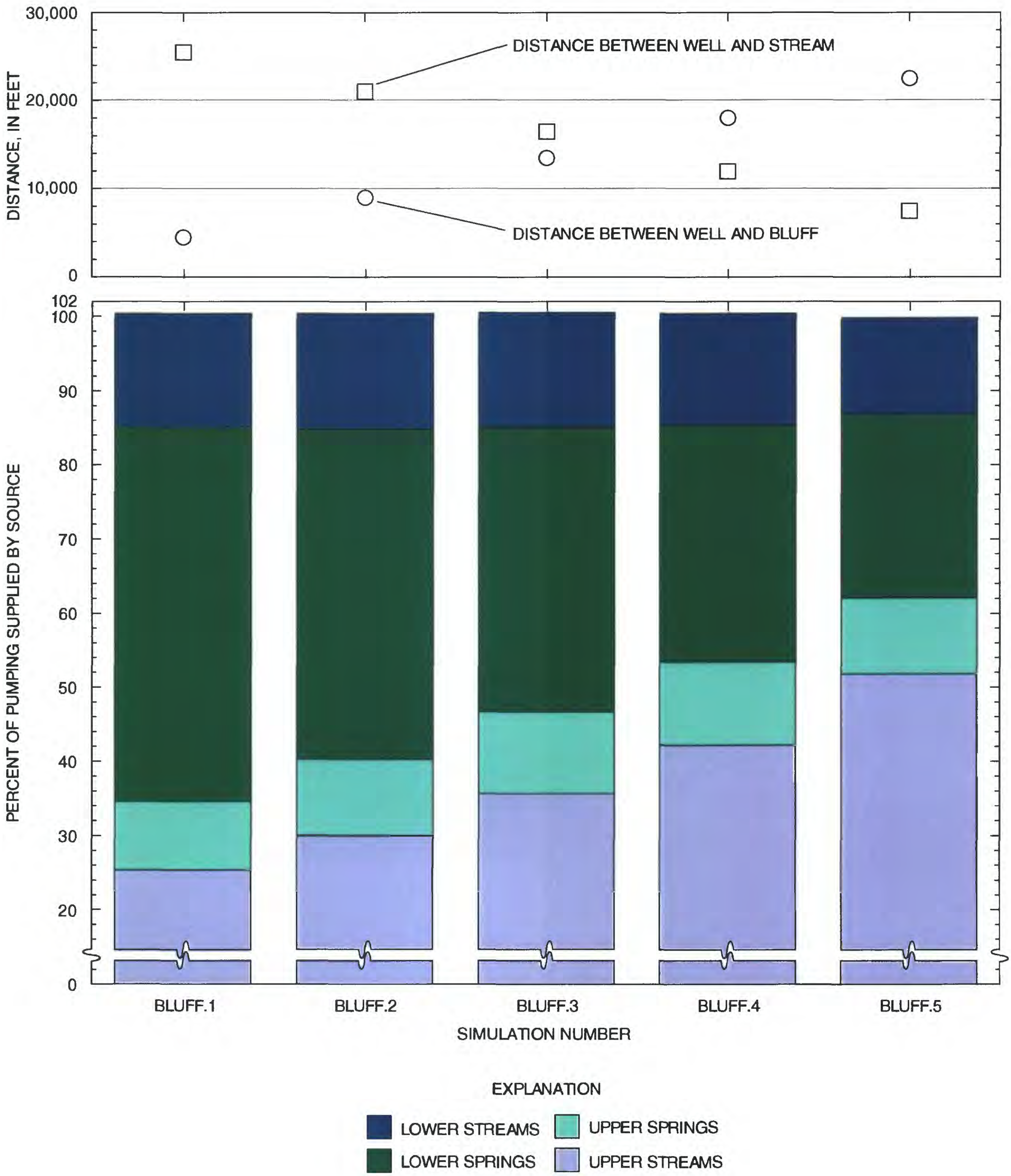

Figure 28.--Sources of water to a simulated well pumping 300 gallons per minute from the uppermost confined outwash aquifer (layer 3) at various locations between the bluff and stream segment number 4 (see figure 10).

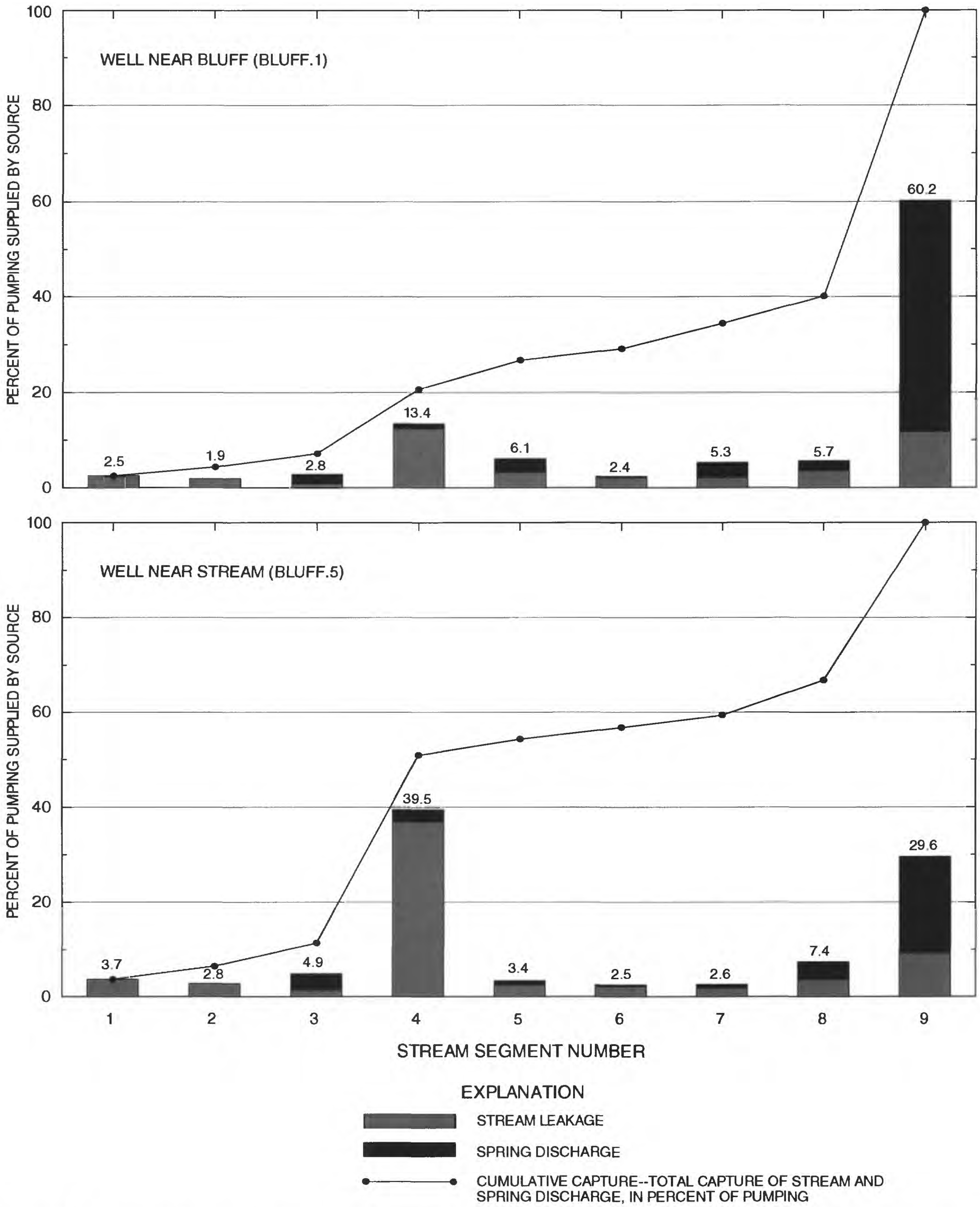

Figure 29.--Simulated capture of ground-water discharge to stream segments by a well pumping 300 gallons per minute near the bluff and near the stream.

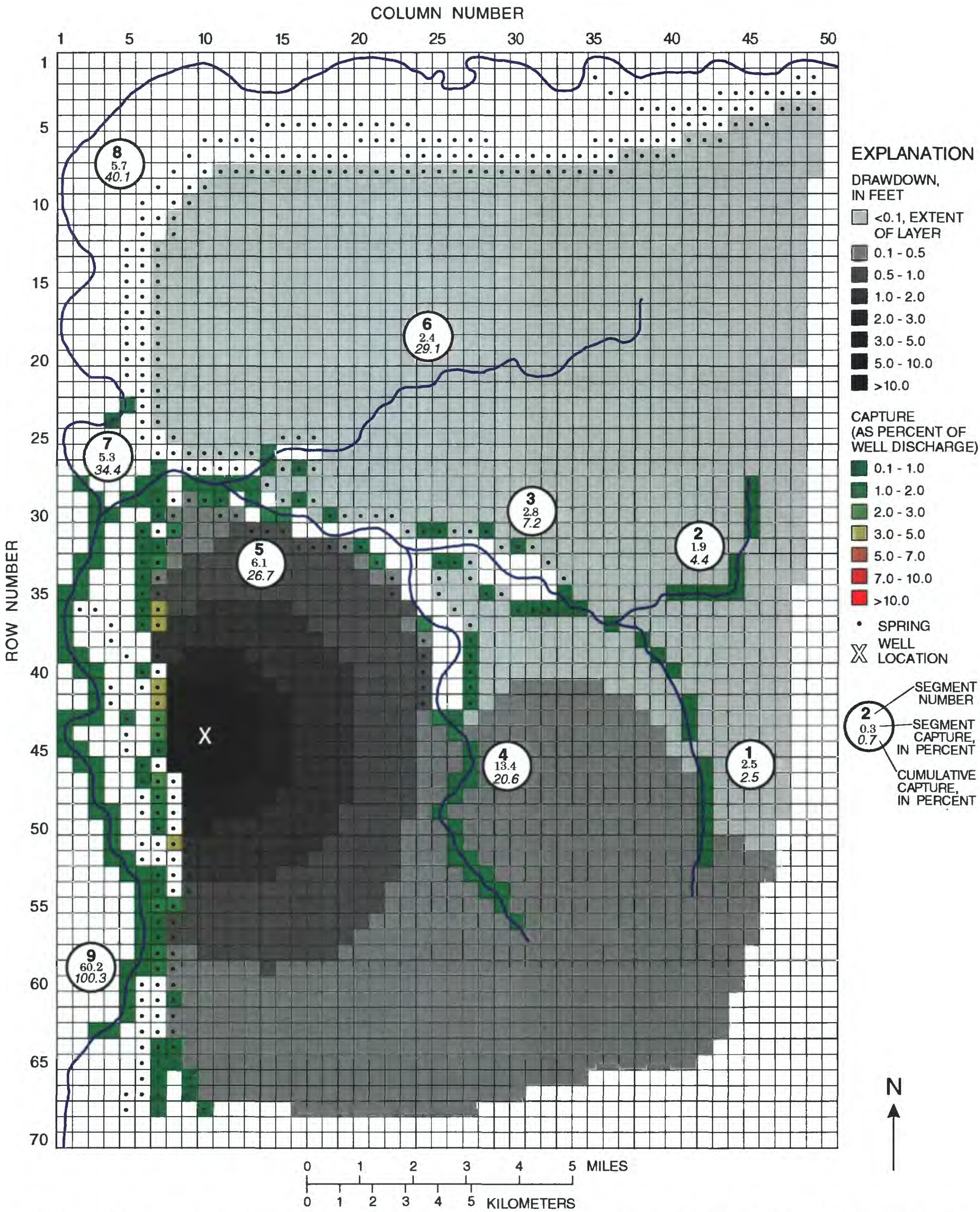

Figure 30a.--Simulated water-level declines and capture of discharge to streams and springs with the well located near the bluff.

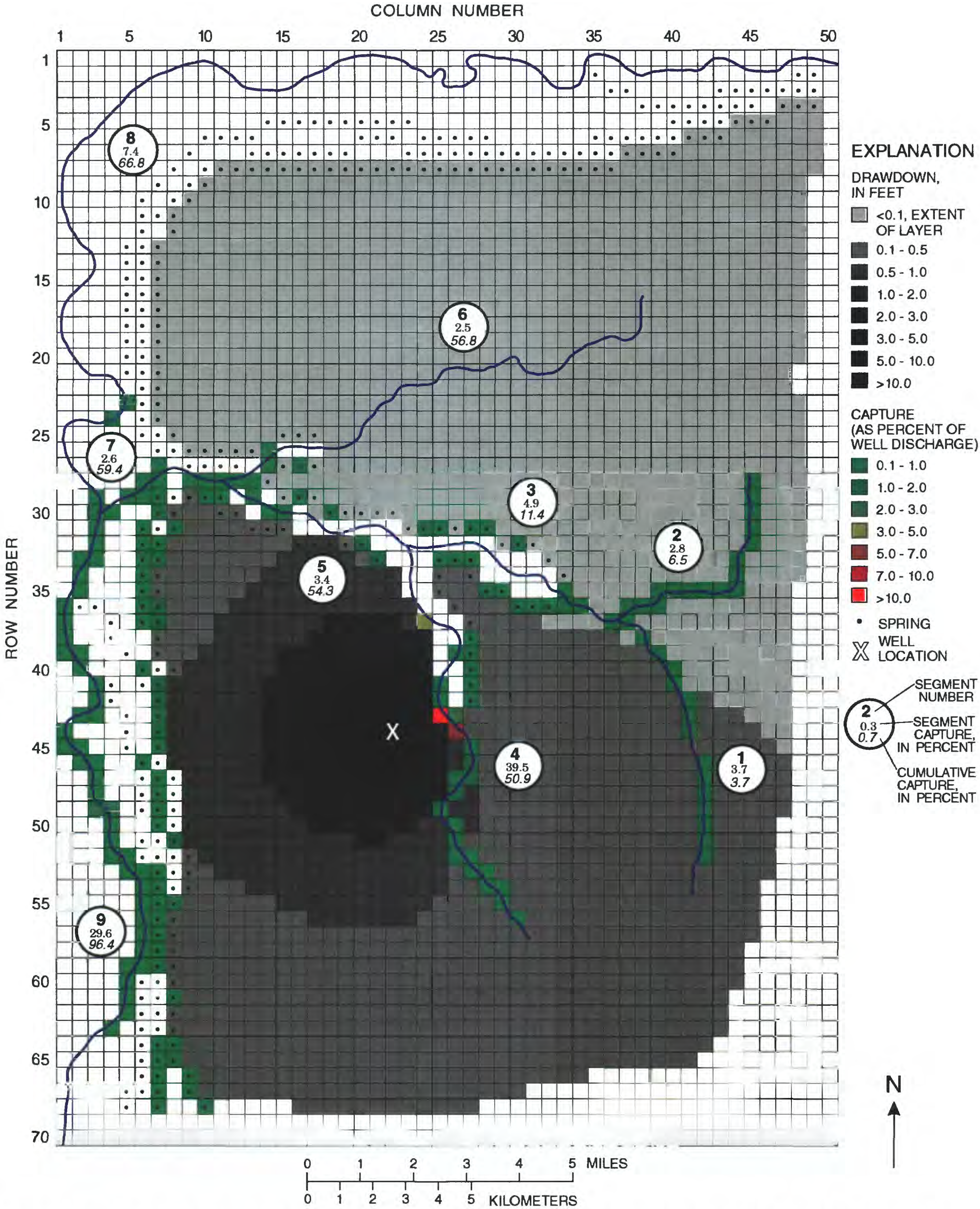

Figure 30b.--Simulated water-level declines and capture of discharge to streams and springs with the well located far from the bluff.

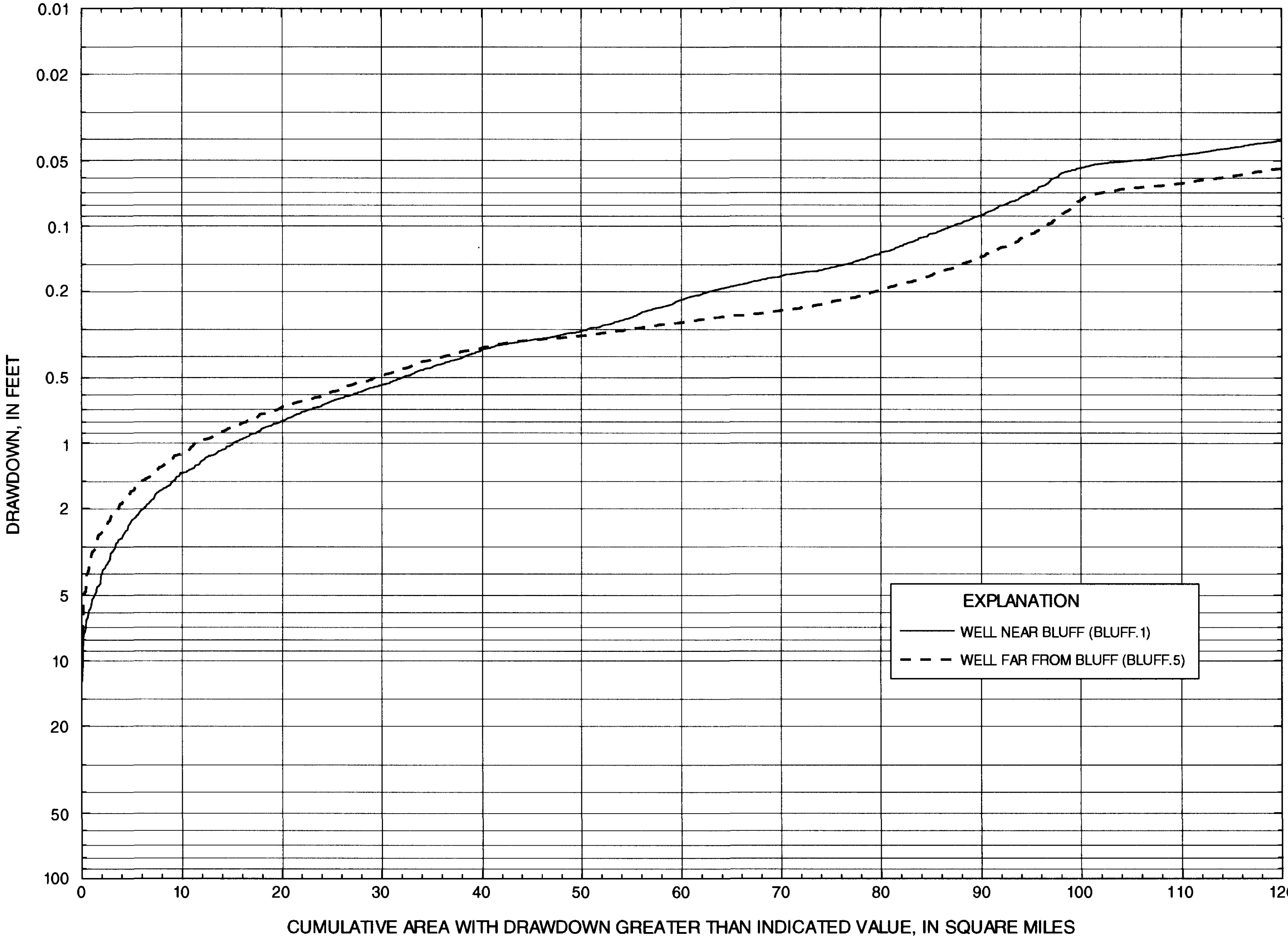

Figure 31.--Cumulative areas affected by simulated drawdown in a confined outwash aquifer (layer 3) when pumping 300 gallons per minute at varying distances from the bluff. Cumulative areas are shown only for simulated drawdown greater than 0.01 feet; total area of layer 3 aquifer is 187 square miles.

Effects of Varying Well Density

The effects of well density on capture of ground-water discharge to streams and springs were analyzed with two simulations (WELL.1 and WELL.2).

In each scenario, it was assumed that 960 new homes were to be built in the basin and that their water needs would be supplied by ground water. The average household water use was assumed to be 450 gal/d (0.312 gal/min). This resulted in an average annual demand of 300 gal/min from ground water. Each model cell covers approximately 52 acres. Twelve acres of each developed cell was assumed to be used for roads and utility right-of-ways, leaving 40 acres available for homes.

The first simulation (WELL.1) assumed a moderate development density of 3 homes per acre. Under this scenario, the 960 homes could be developed within 8 model cells. Each of the 8 model cells contained a separate 40-acre development of 120 homes and each development (cell) had its own well. With a pumping rate of 37.5 gal/min for each of the eight wells, the total pumping in the scenario was 300 gal/min. Four of the developments were clustered in the northern part of the basin and four were clustered in the central part of the basin (fig. 10).

The second simulation (WELL.2) assumed a low development density of 1 home per 20-acre parcel. Under this scenario, the 960 homes were spread over 480 model cells. The homes in this simulation were also divided equally between the northern and central parts of the basin for comparison with simulation WELL.1 (fig. 10). At this density each home would have an individual well for domestic water supply, with no significant pumping for any other water use. The pumping rate for each model cell was 0.625 gal/min and the total pumping for the basin was 300 gal/min. The aquifer developed in each simulation was the uppermost confined outwash aquifer (model layer 3).

At steady-state, the effect of well density on the capture of discharge to streams and springs was minimal. In both the high and low density pumping scenarios, most well discharge was captured from streams on the upper drift plain (table 5). Streams and springs in the rest of the basin contributed approximately equally, with percentages of captured discharge ranging from 11 to 21 percent. The differences between the sources for these two scenarios was remarkably small; the maximum difference was only 4 percent. Since the centroids of pumping for both simulations were essentially the same, the cone of depression expanded to approximately the same areas in each simulation in order to divert the discharge required to offset the pumping withdrawal.

In both simulations, the effects of pumping on natural ground-water discharge rates were spread over broad areas, minimizing the effects on discharge to individual stream reaches and springs. Only a few stream reaches and spring cells contributed more than 1 percent of the total well discharge, and these were all on the upper drift plain with most on stream segment 3 (figs. 32a,b).

The most noteworthy difference in the effects produced by these two pumping simulations was in the distribution of simulated drawdown (figs. 32a,b). Drawdown of 0.5 ft or more was simulated over an area of 25 mi^2 in both scenarios. However, with low density development, the maximum drawdown was only 0.8 ft, compared with a maximum of over 4 ft for high density development. The areas affected by drawdown were essentially equal for drawdown greater than 0.5 ft and the AOI for both simulations was about 145 mi^2 (fig. 33).

Table 5.--*Sources of water to wells for simulations of the effects of well density on ground-water discharge to streams and springs*

[Values are expressed as percentages of total well discharge]

Simulation	Upper streams	Upper springs	Lower streams	Lower springs
		High density		
WELL.1	56	17	12	15
		Low density		
WELL.2	53	21	11	15
Difference	3	-4	1	0

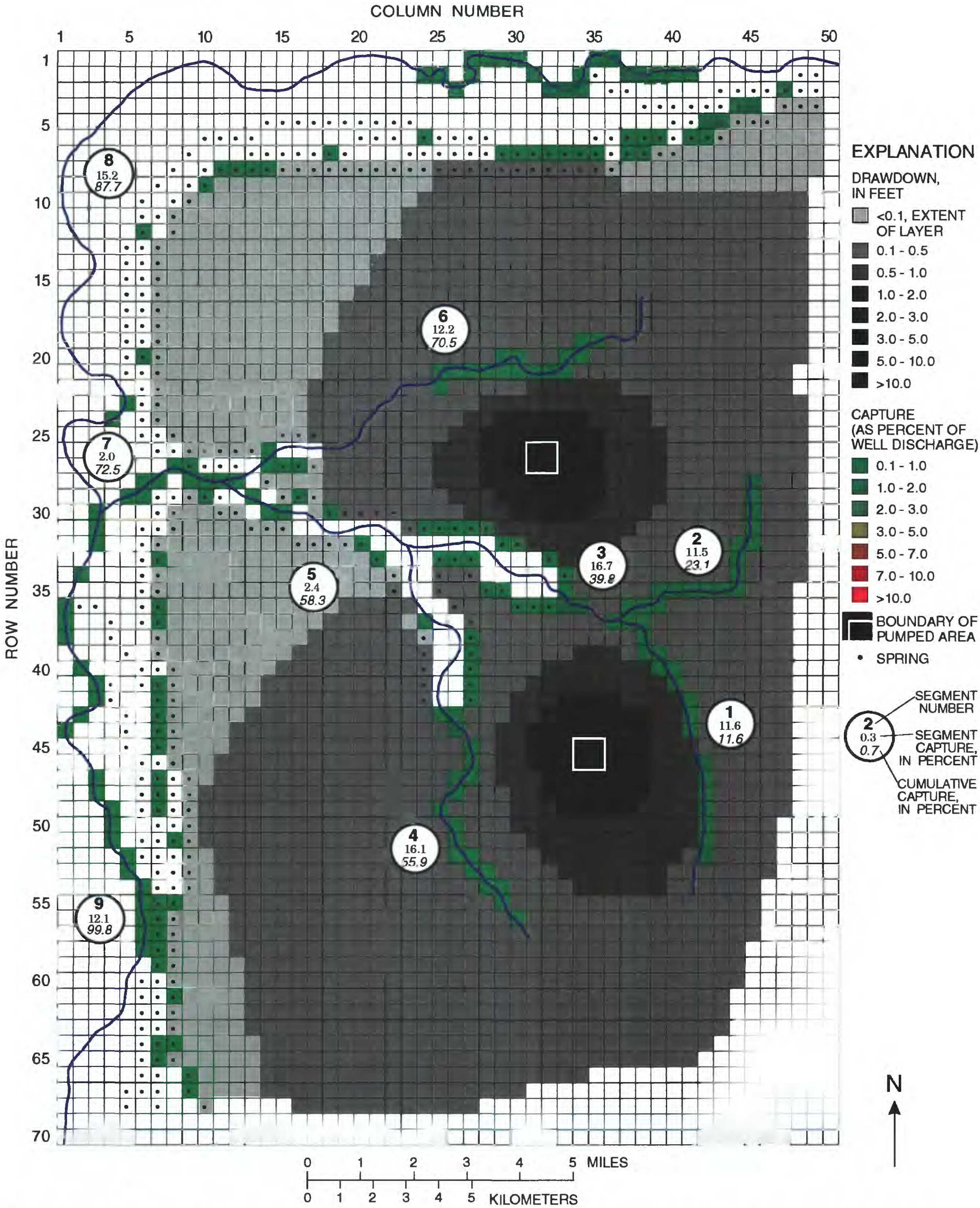

Figure 32a.--Simulated water-level declines and capture of discharge to streams and springs with a high well density.

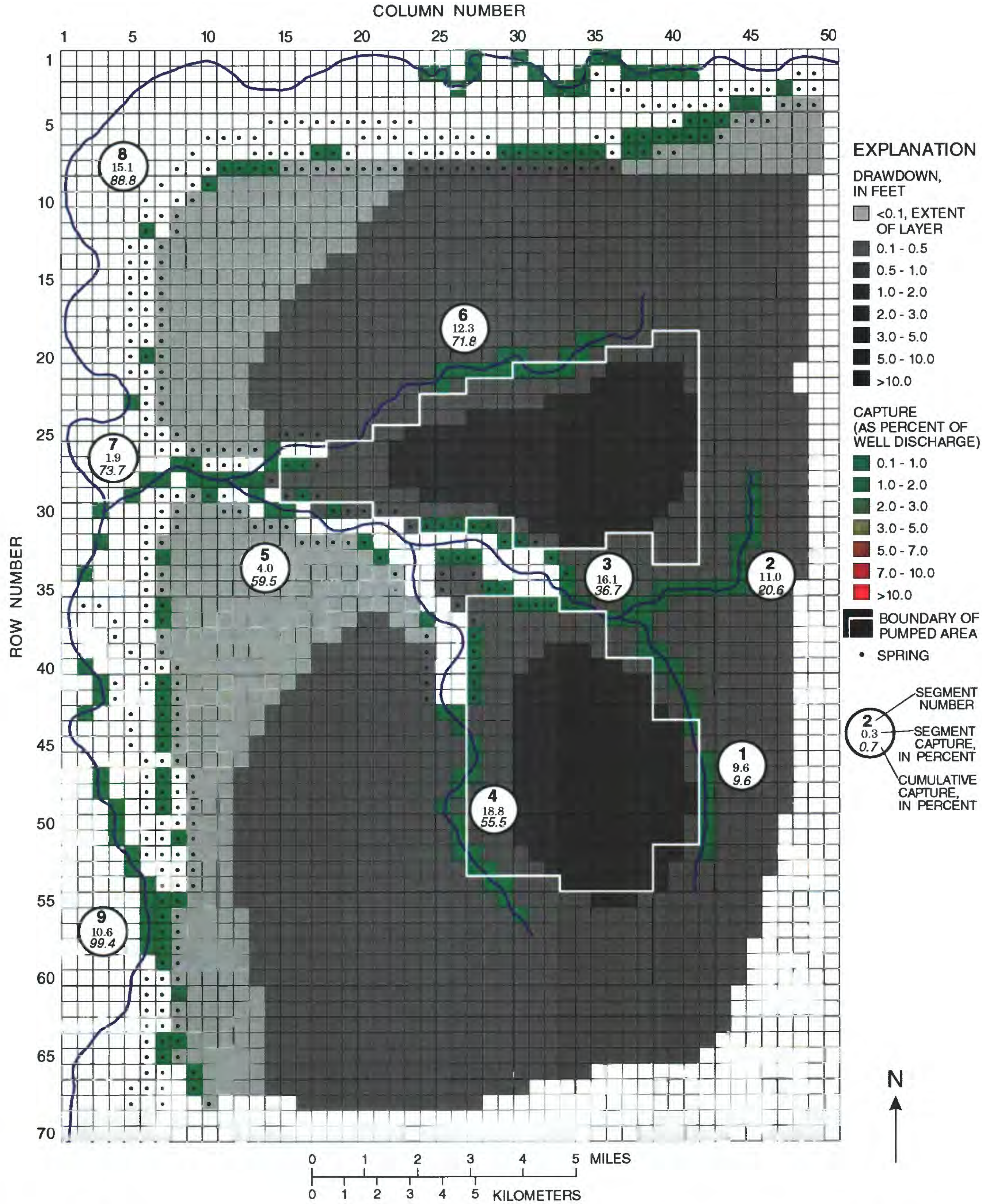

Figure 32b.--Simulated water-level declines and capture of discharge to streams and springs with a low well density.

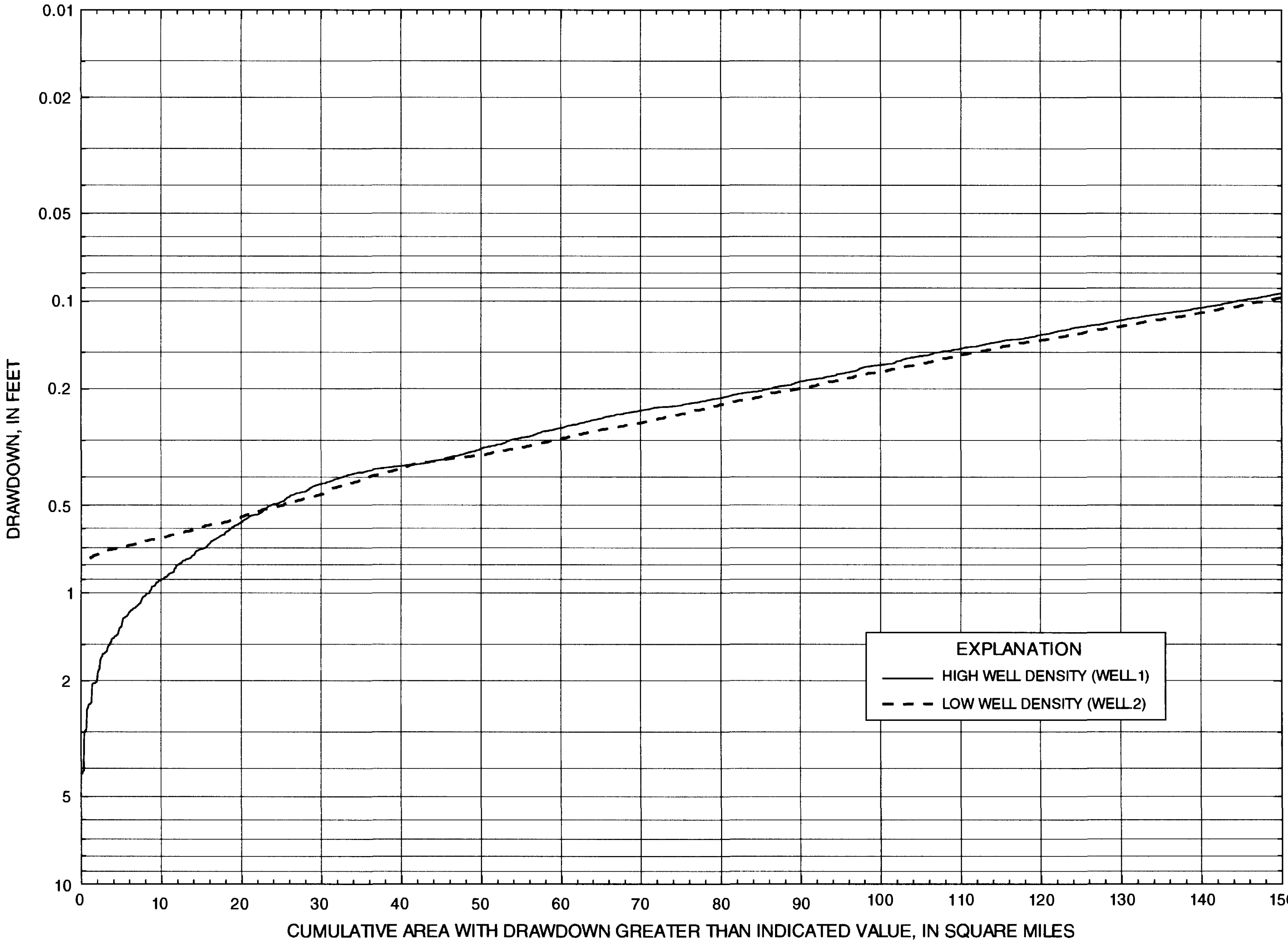

Figure 33.--Cumulative areas affected by simulated drawdown in a confined outwash aquifer (layer 3) when pumping 300 gallons per minute at varying well densities. Cumulative areas are shown only for simulated drawdown greater than 0.01 feet; total area of layer 3 aquifer is 187 square miles.

Effects of Varying Recharge Rate

Six simulations were made in the RECH series to analyze the effect of changes in ground-water recharge on capture of ground-water discharge to streams and springs. Development in the basin changes the surface of the land, affecting recharge. Generally, development reduces recharge. Under natural conditions, a significant amount of precipitation falls on and infiltrates pervious soils, but under developed conditions it falls on impervious surfaces such as streets and roofs and is routed directly to surface drains.

In some instances, however, natural recharge rates are maintained or even exceeded in developed areas. This generally occurs where household waste water is disposed of in septic systems and cesspools. Another means of increasing recharge rates is to route runoff from impervious surfaces directly into permeable soils through drains, or "drywells", as they are sometimes called. Both on-site waste systems and drywells are efficient means of recharging the ground-water system because water does not have to percolate through the soil zone where it is subject to losses due to plant transpiration and evaporation. Unfortunately, recharge from these sources is often of poor quality and may not be a desirable addition to the ground-water reservoir.

Each of these conditions, reduced and increased recharge, would have an effect on the capture of ground-water discharge to streams and springs by pumping wells. The purpose of the RECH series of simulations was to determine the extent and magnitude of these effects in the hypothetical basin. The term "effective discharge" describes the net withdrawal from the ground-water system resulting from both well discharge and changes in recharge rate caused by development. As an example, if development occurs and a new well withdraws 600 gal/min and recharge is reduced by 300 gal/min because of the increases in impervious surfaces, the effective discharge is 900 gal/min. The term is useful because it allows comparison of scenarios that include both well withdrawals and changes in recharge.

In the RECH series of simulations, it was assumed that moderate density (3 homes per acre) development of 1,920 homes covers an area equivalent to 16 model cells near stream segment 4 (fig. 10). The area is underlain by recessional outwash deposits, and the average annual recharge is estimated to be 27 in/yr. The per-household water use is the same as was assumed for previous scenarios (450 gal/day), and the average annual pumping rate for the development is 600 gal/min; all ground water is withdrawn from the uppermost confined outwash aquifer (model layer 3).

In simulation RECH.3, streets and other impervious surfaces cover 27 percent of the area, and all runoff from these surfaces is routed directly to ditches and is not allowed to recharge the ground-water system. The area has sanitary sewers, and no other measures are taken to enhance recharge, so total recharge is effectively reduced by 27 percent, or the equivalent of about 300 gal/min, from natural conditions. The effective discharge for simulation RECH.3 was approximately 900 gal/min. In simulation RECH.4, the area is served by a sanitary sewer system, but measures are taken to enhance recharge (such as drywells and retention basins) and recharge rates are the same as under natural conditions. The effective discharge for this simulation was 600 gal/min. In simulation RECH.6, recharge enhancement measures are taken and the area does not have sewers. It was assumed that 50 percent of the water requirements of the household move through the septic system to recharge the shallow ground-water system. Thus, for simulation RECH.6, the effective discharge rate was equal to the pumping rate, 600 gal/min, less the 50 percent that returned to the ground-water system, 300 gal/min.

Because the natural recharge rate is highly dependent on the hydraulic characteristics of the surface hydrogeologic layer, three additional simulations (RECH.1, RECH.2, and RECH.5) were made for an area adjacent to stream segment number 4 that is underlain by till (fig. 10) so that comparisons could be made of the effect of the surface hydrogeologic layer. The locations of the wells in the till and outwash scenarios could not be the same; however, they were adjacent and had nearly identical proximity to nearby boundaries (fig. 10). The same assumptions were used regarding the percentages of impervious area and recharge reductions as well as the percentage of domestic water use available for recharge through the septic systems. The conditions for each simulation are summarized in table 4.

In the till-covered area, reducing the recharge of 18 in/yr by 27 percent over the 1.3 mi^2 area of development effectively reduced recharge by about 200 gal/min, or one-third of the 600 gal/min well discharge for the development. For simulation RECH.1, the effective discharge was approximately 800 gal/min. In simulation RECH.2, recharge was maintained at the natural rate, so that the effective discharge was equal to the pumping rate of 600 gal/min. The third simulation in the till covered area (RECH.5) had recharge increased by an amount equal to 50 percent of the pumping rate or 300 gal/min; thus for simulation RECH.5, the effective discharge rate was equal to 600 gal/min less 300 gal/min, or 300 gal/min.

The percentages of effective discharge supplied by capture of ground water that would have discharged to streams and springs are shown in figure 34. Within the till area, for the cases of reduced recharge and natural recharge (RECH.1 and RECH.2; fig. 34), the percentages of capture from various sources are identical. Although these are percentages of the effective discharge, equal percentages do not represent equal rates of capture. For example, 57 percent of the effective discharge rate was captured from the upper streams in both the reduced (RECH.1) and natural (RECH.2) recharge simulations; because of the difference in effective discharge rates, this represents capture of 456 gal/min from the upper streams in the reduced recharge simulation and capture of only 342 gal/min in the natural recharge simulation (fig. 34). By increasing recharge, as was simulated in RECH.5, an essentially new source of water becomes available to offset discharge from the pumped wells. This additional recharge was added to the water-table layer (layer 2 in this case) while the pumped well discharged from layer 3. The additional recharge resulted in additional downward leakage and additional water available to the wells, streams, and springs at equilibrium.

The simulated drawdown distribution in figure 35a reflects the effects of both the well discharge and the reduction in recharge in scenario RECH.1. The cone of depression is elongated parallel to the stream segment with its shape strongly controlled by the boundaries to the aquifer. There are several stream reaches in segment 4 with baseflow reduced by 3 to 5 percent of the effective discharge rate of 800 gal/min. Additionally there are several springs on the bluff above stream segment 9 that have had discharge reduced by 1 to 2 percent of the effective discharge. In contrast, increasing recharge to the till had a noticeable effect on the drawdown distribution and on the percentage of well discharge derived from captured baseflow and spring discharge (fig. 35b). The cumulative areas of drawdown (fig. 36) reveal that, at equilibrium, water levels increase as recharge increases, and accordingly, the size of the AOI decreases from about 133 mi^2 for decreased recharge to 97 mi^2 for increased recharge.

In simulations RECH.3, RECH.4, and RECH.6, the wells pumped from the uppermost confined aquifer where it is overlain by a till layer and a recessional outwash aquifer. Reducing the natural recharge rate of 27 in/yr by 27 percent over the 1.3 mi^2 area of development resulted in an effective reduction of recharge of about 300 gal/min. Combined with the well pumping rate of 600 gal/min, the effective discharge from the system was 900 gal/min. Because of their locations, these wells captured a large percentage of their discharge from baseflow to stream segment 4. The primary reason was that the outwash aquifer provided discharge to the stream, and pumping the wells induced a stronger downward vertical hydraulic gradient and greater flux from the surficial outwash aquifer, through the till, to the pumped aquifer. Figure 34 shows that there are slight differences between the reduced and natural recharge simulations, but the relative sources of water to the wells are similar. Due to the difference in effective discharge rates for the reduced and natural recharge simulations, the rates of capture are not the same even though the percentages are similar.

With reduced recharge (RECH.3, fig. 37a), pumping in the area underlain by outwash resulted in less simulated drawdown than pumping in the area underlain by till (RECH.1, fig. 35a). Pumping from the aquifer underlying the unconfined aquifer adjacent to the stream had a significant effect on the upper reaches of segment 4; several reaches had baseflow reduced by 5 to 10 percent of the effective discharge rate of 900 gal/min and a total of 61 percent, or 550 gal/min, was captured from baseflow to segment 4 alone. Comparison of figures 37a and 37b shows that increasing recharge, and thus decreasing the effective discharge from 900 to 300 gal/min, greatly attenuated the effects of pumping on individual stream reaches and springs. The effect of increased recharge on the drawdown distribution is illustrated by the cumulative drawdown areas shown on figure 36. Increasing recharge in the area covered by outwash resulted in greater reductions in the effective discharge rate; the rate was reduced from 900 to 300 gal/min in the outwash area, compared with a reduction from 800 gal/min to 300 gal/min in the till area. In spite of this, the simulation results show that effects on discharge to streams and springs were attenuated when recharge was increased in the till area. As an index of the change in drawdown caused by increasing recharge in each area, the AOI for pumping in the outwash area decreased by 20 mi^2, compared with the 36 mi^2 decrease for pumping from the till area (fig. 36).

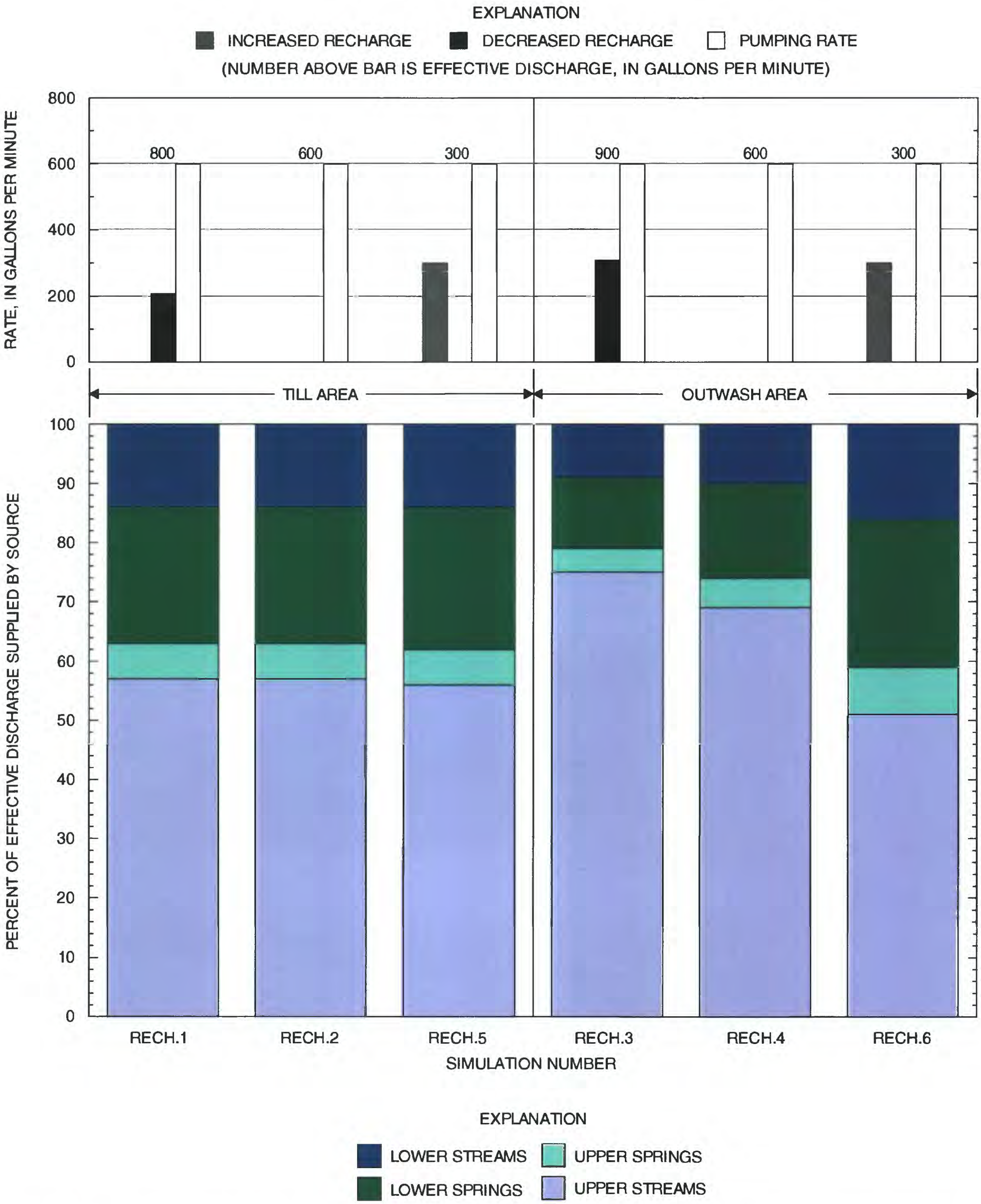

Figure 34.--Sources of water to simulated wells pumping from a confined outwash aquifer (layer 3) in areas overlain by till and outwash under conditions of increased and decreased recharge. Locations of wells are shown in figure 10.

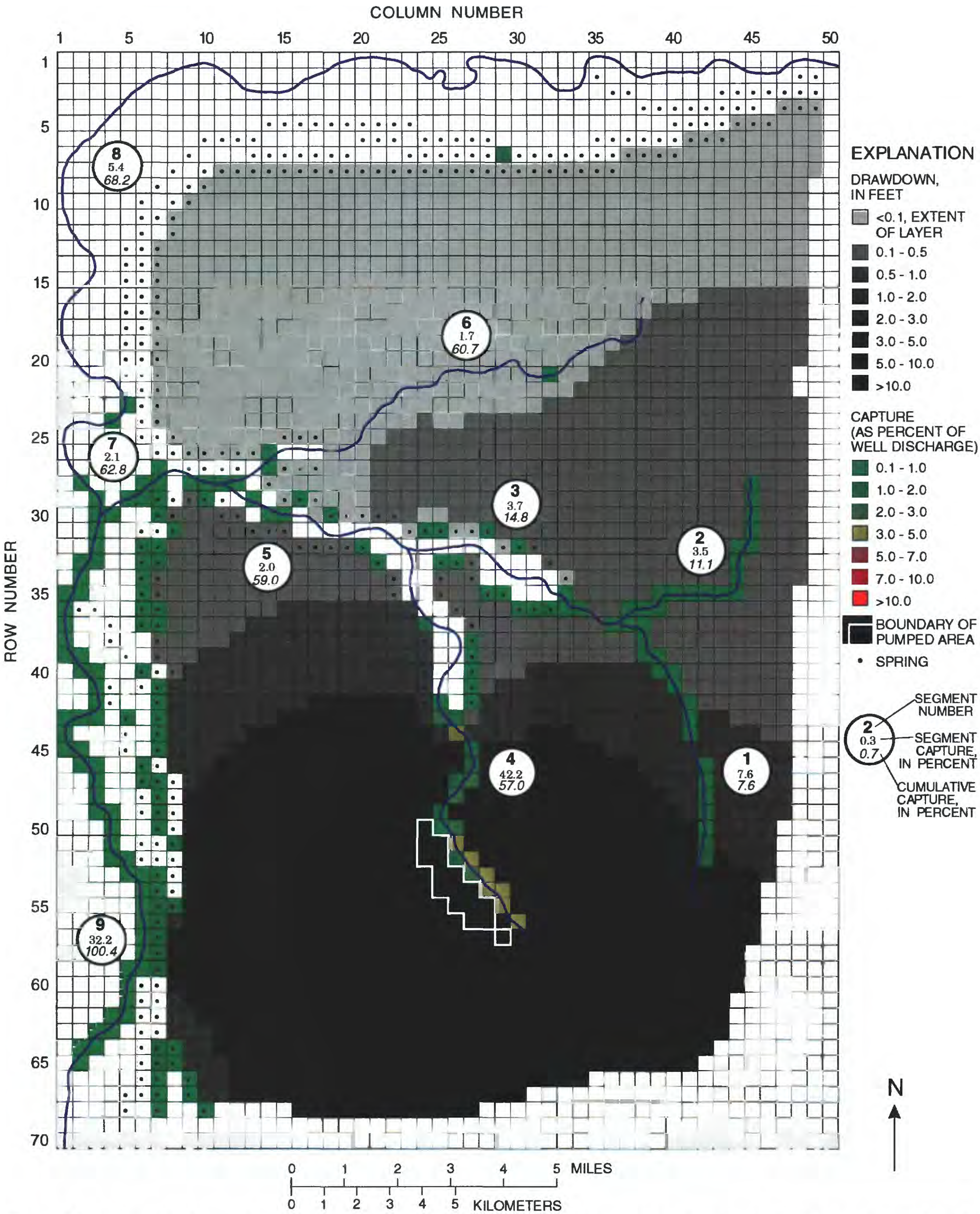

Figure 35a.--Simulated water-level declines and capture of discharge to streams and springs with simulated pumping of 600 gallons per minute from the uppermost confined aquifer (layer 3) where till is at land surface and recharge is decreased.

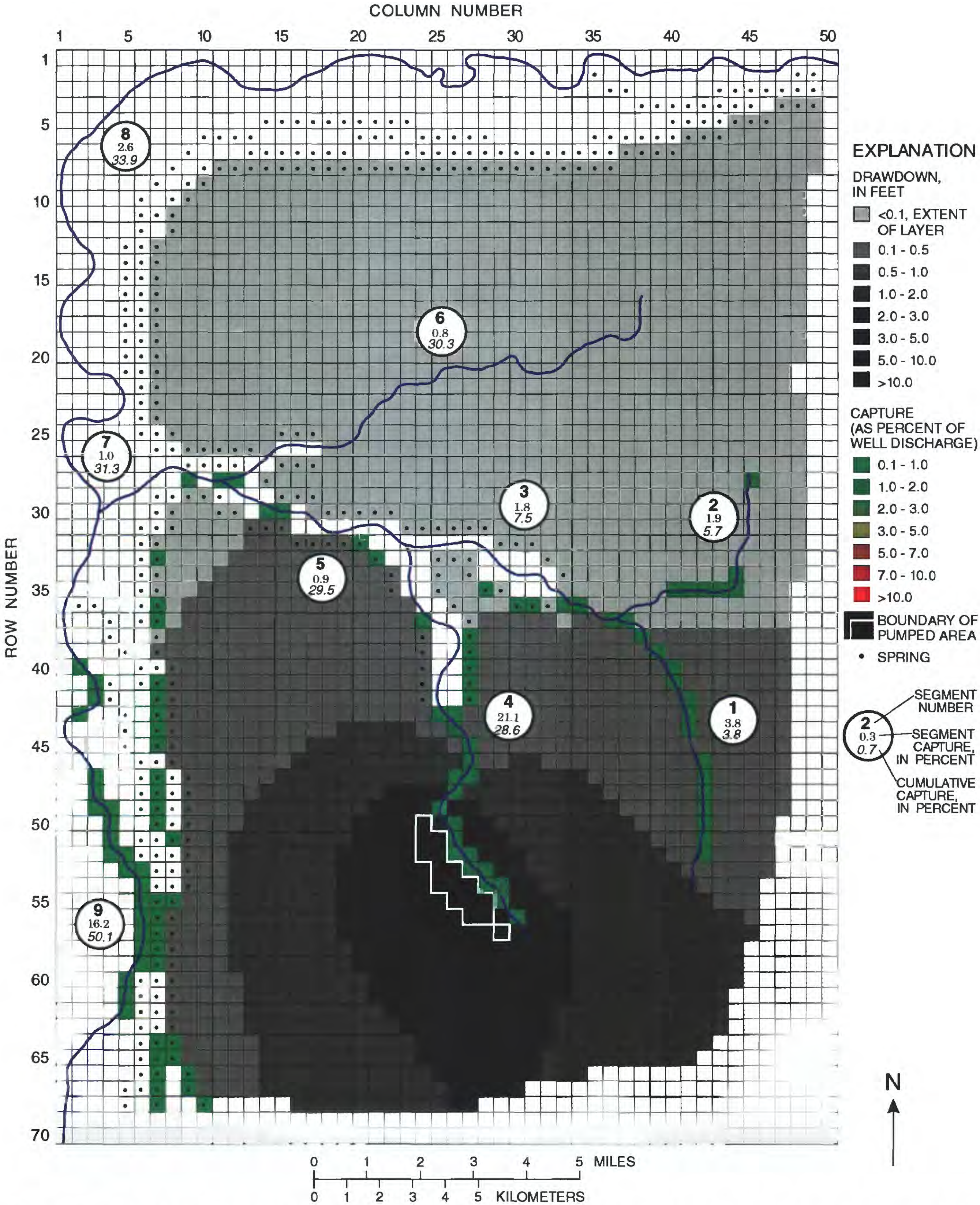

Figure 35b.--Simulated water-level declines and capture of discharge to streams and springs with simulated pumping of 600 gallons per minute from the uppermost confined aquifer (layer 3) where till is at land surface and recharge is increased.

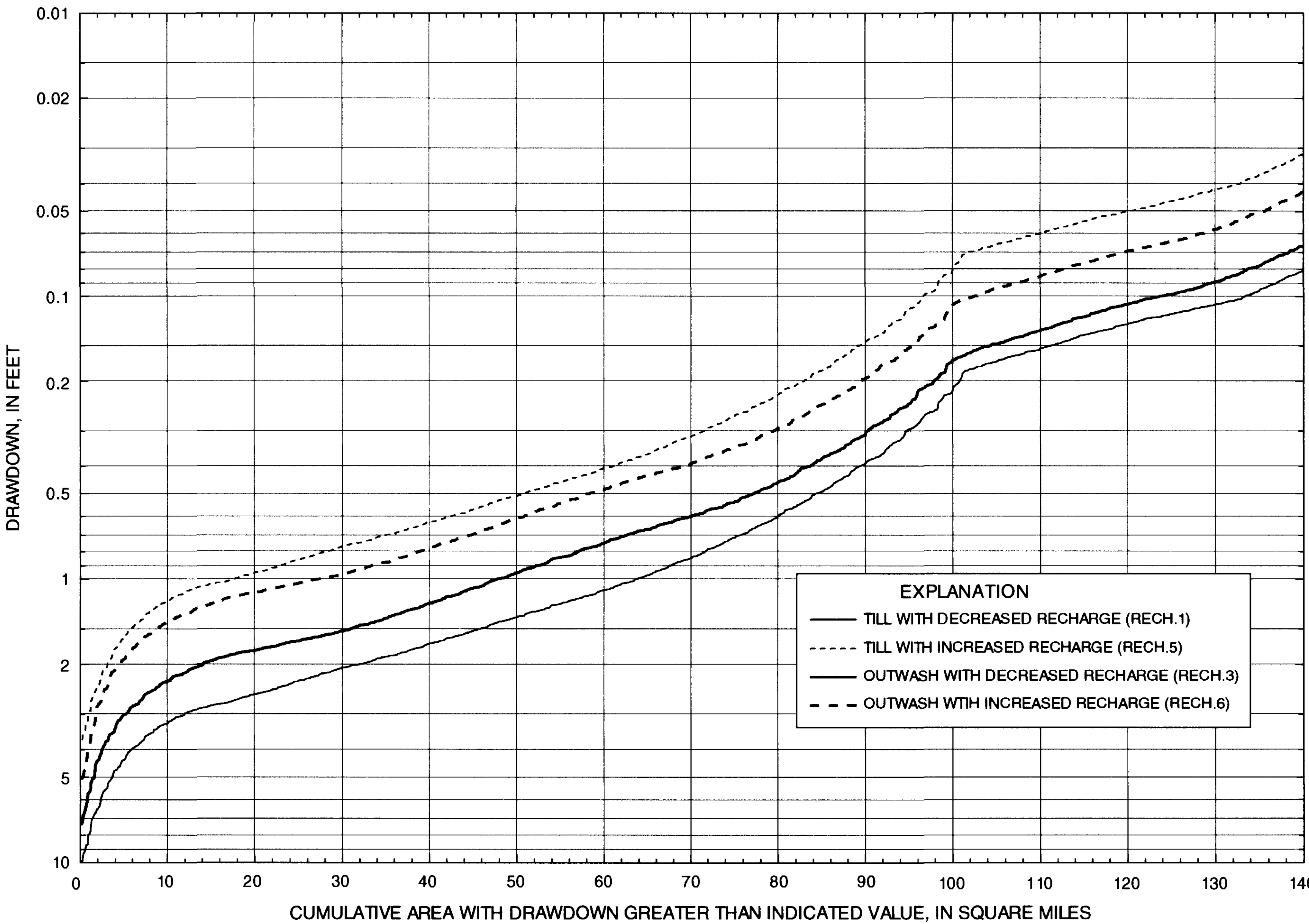

Figure 36.--Cumulative areas affected by simulated drawdown in a confined outwash aquifer (layer 3) when pumping 600 gallons per minute in areas underlain by till and outwash and under conditions of increased and decreased recharge. Cumulative areas are shown only for simulated drawdown greater than 0.01 feet; total area of layer 3 aquifer is 187 square miles.

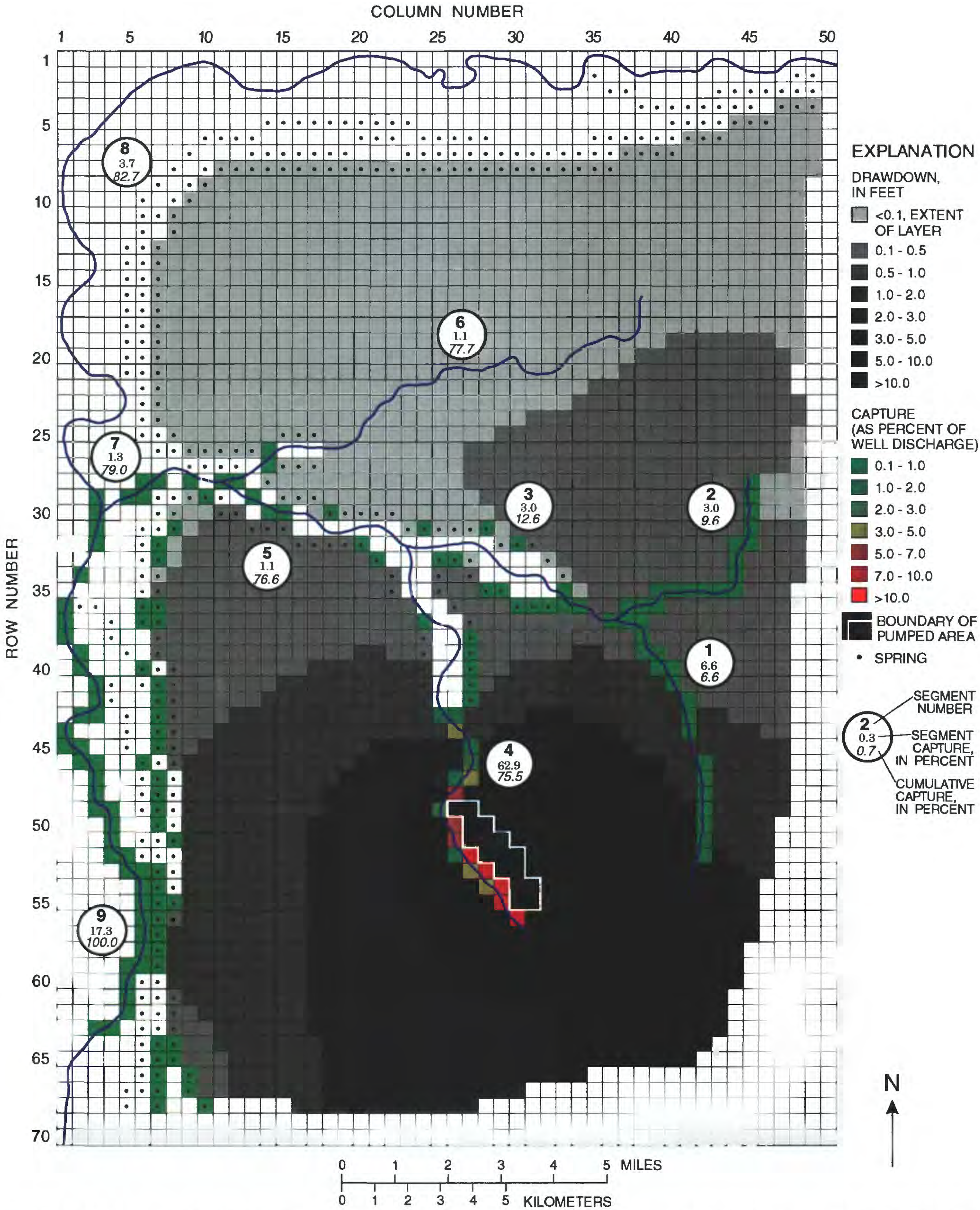

Figure 37a.--Simulated water-level declines and capture of discharge to streams and springs with simulated pumping of 600 gallons per minute from the uppermost confined aquifer (layer 3) where outwash is at land surface and recharge is decreased.

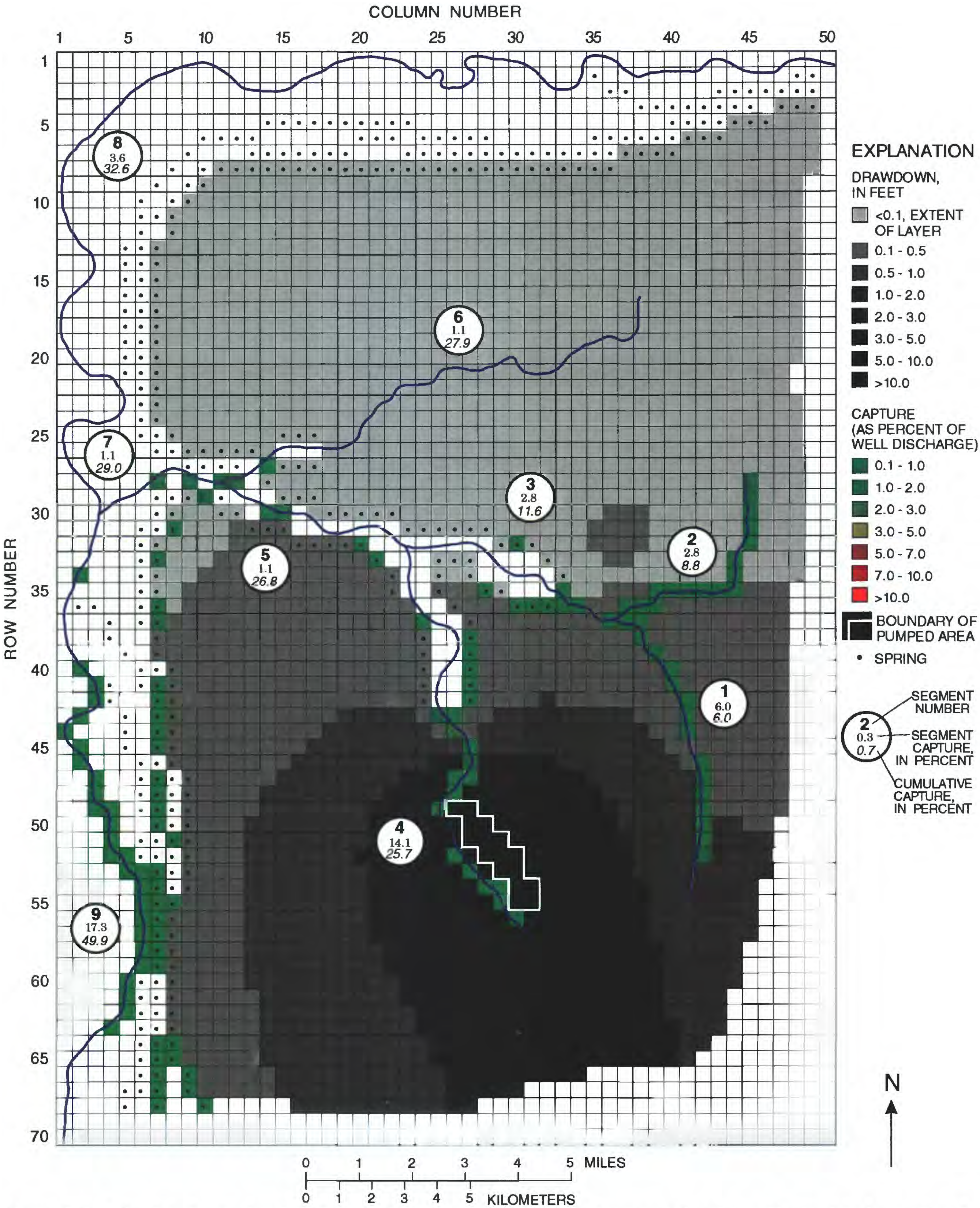

Figure 37b.--Simulated water-level declines and capture of discharge to streams and springs with simulated pumping of 600 gallons per minute from the uppermost confined aquifer (layer 3) where outwash is at land surface and recharge is increased.

SUMMARY AND CONCLUSIONS

The effects of ground-water withdrawals on streamflow have become an issue of major concern in the Puget Sound Lowland of western Washington as continuing population growth increases the demand for water. Surface-water resources are completely allocated in important segments of the region, and future growth there will most likely depend on the availability of ground water. Though the basic nature of interactions between the ground-water and surface-water systems is well known, the details of these interactions in small basins of the Puget Sound Lowland are not well understood. Ground-water development will, in most cases, affect the baseflow to streams. The lack of understanding about the details of the hydrologic system is due to a number of factors, but the most important may be the complexity of the Quaternary geology of the region. Repeated series of glacial advances and retreats, punctuated by interglacial periods of deposition and erosion, created an extremely complex system of aquifers and confining layers through which ground water passes as it moves from recharge areas to discharge points at streams and springs. The lack of understanding is compounded by the difficulty and expense of collecting the data necessary to characterize these systems. Finally, many of the traditional tools for assessing the interactions between ground water and surface water are either too simplistic to be useful (analytical models) or too complex and expensive (numerical models) to be practical for solving site-specific problems that face regulators. Nonetheless, regulators must maintain minimum streamflow and protect the interests of surface-water rights, so they need a more detailed conceptual understanding of the effects of ground-water withdrawals on streamflow to help guide decisions.

The purpose of this study was to provide a better understanding of relations and interactions between the ground-water and surface-water systems in small basins of the Puget Sound Lowland and, particularly, to identify some of the important factors controlling the response of the systems to ground-water withdrawals. The primary tool in this investigation was a numerical ground-water-flow model. The model was developed for a hypothetical basin in the Puget Sound Lowland and was based on a conceptual model synthesized from the work of many previous investigators in the region. Topography, geology, drainage, and climate were defined for the 262 square mile hypothetical basin. Hydrologic conditions simulated by the numerical model, such as ground-water levels and discharge to streams and springs, were compared with conditions typical of the region, and model parameters were adjusted until simulated and typical conditions agreed closely. The calibrated, or baseline, model was then used to simulate the effects of ground-water withdrawals on discharge to streams and springs under a variety of scenarios. Seven series of scenarios were simulated in which the effects of 1) distance from the well to a stream, 2) the presence of a confining layer, 3) pumping rate, 4) depth of the pumped aquifer, 5) distance from the well to a bluff, 6) well density, and 7) recharge rate were evaluated.

The results of each simulation were compared with the baseline model results to compute the percentage of the well discharge that was derived, or captured, by diverting flow that otherwise would have discharged to streams and springs. All simulations were of equilibrium, or steady-state conditions. That is, they simulated conditions after water levels had adjusted to the pumping stress and no changes in ground-water storage were occurring.

The central part of the hypothetical basin is a drift plain composed of layered Pleistocene glacial drift and interglacial sediments bounded on the east and south by low-permeability Tertiary bedrock and on the west and north by steep bluffs. At the base of the bluffs, 200 to 600 feet below the drift plain, lies a broad valley drained by a major stream. The valley contains up to 500 feet of alluvium consisting of a heterogeneous mixture of gravel, sand, silt and clay. The drift plain has relatively low relief and the streams that drain the plain have low gradients until they descend the bluff to the major stream valley; where the stream crosses the bluff it has incised a deep canyon, exposing the drift deposits. Recharge to the ground-water system depends on annual precipitation and the permeability of the geologic layer at the surface. The mean annual precipitation in the basin is 44 inches per year and the recharge rate in areas where the more permeable outwash deposits are exposed is 27 inches per year compared to recharge of only 18 inches per year in areas where the less permeable till is exposed; till covers most of the basin and the average recharge is 20 inches per year (389 cubic feet per second).

A three-dimensional numerical model of the ground-water-flow system of the hypothetical basin was constructed using the U.S.Geological Survey's MODFLOW model. The ground-water system was subdivided horizontally into a regular grid of cells, each having dimensions of 1,500 feet per side; 50 columns and 70 rows were included in the grid. The vertical dimension was subdivided using 13 layers of cells. Three glacial sequences, each consisting of recessional outwash, till, advance outwash, and interglacial deposits, were part of the conceptual model of the hypothetical basin. Each hydrogeologic layer was simulated using a separate model layer and, therefore, the three glacial sequences made up the upper 12 layers of the model. Beneath the drift plain, the 13th (bottom) layer represented undifferentiated glacial and interglacial deposits. The Quaternary alluvium underlying the major stream valley was represented in layers 9 through 13. The lower boundary of the model represented the contact between the Quaternary unconsolidated sediments and the consolidated Tertiary siltstones and mudstones that form a low-permeability (no-flow) boundary to the model.

Thickness and hydraulic characteristics of the hydrogeologic layers were initially assigned on the basis of values published from previous investigations in the Puget Sound Lowland. Values of hydraulic characteristics were modified during model calibration to provide a better fit to expected hydrologic conditions in the hypothetical basin. The horizontal hydraulic conductivity of the glacial sequences ranged from 0.25 foot per day and 1.0 foot per day for the till and interglacial confining layers to 100 feet per day for the outwash aquifers. The alluvial deposits of the major stream valleys were assigned a value of 50 feet per day and the undifferentiated deposits a value of 25 feet per day. Ratios of horizontal to vertical hydraulic conductivity ranged from 10 for outwash and alluvial aquifers to 100 and 200 for till and interglacial confining layers. Each layer was assumed to be homogeneous.

Ground water generally flows downward beneath the principal recharge area on the drift plain and then flows laterally from the south and east toward the primary discharge areas, where it flows upward. The primary discharge areas are the major stream valley and the springs that discharge on the bluffs to the north and west; however, shallower, local flow systems also discharge to streams and springs on the drift plain. In the baseline model, 73 percent (285 cubic feet per second) of the ground water discharged to the major stream valley and springs on the bluffs; the remaining 27 percent (104 cubic feet per second) discharged to streams and springs on the drift plain. The proportions of discharge to the major stream valley and the drift plain were reasonable on the basis of expected baseflow to streams on the drift plain of 100 to 175 cubic feet per second. The simulated range in specific discharge to streams on the drift plain of 0.3 to 3 cubic feet per second per mile also compared well with the expected range of 0.8 to 3.9 cubic feet per second per mile based on gain-loss data for a small watershed in southwest King County.

The following principal conclusions were drawn from the simulation of various pumping scenarios.

- A well pumped from an unconfined outwash aquifer that is in contact with a streambed will capture nearly all of its discharge by diverting flow from the nearest reaches of the stream. Increasing the distance between the well and the stream allows the well to capture some discharge from other streams on the drift plain, but does not affect discharge to springs on the bluffs or to the stream in the lower valley.
- When a confining layer separates the nearest stream from the pumped aquifer, the effects of pumping spread over a much larger area. The low-permeability confining layer forces the cone of depression of the well to extend to greater distances to divert the natural discharge required to offset pumping.
- The presence of a confining layer between the well and the stream is more important than the distance between the well and the stream in determining the distribution of capture of natural discharge throughout the basin.
- At equilibrium, the magnitude of drawdown and capture at any point are a function of the pumping rate.
- As the depth of a well and the number of confining layers between it and discharge areas increases, capture of discharge to streams and springs is distributed over increasingly larger areas.
- The bluffs are important hydrogeologic boundaries. Discharge to springs on the bluffs is very sensitive to the distance of wells from the bluffs.
- The density of wells does not have a significant effect on the equilibrium distribution of capture of natural discharge; however, higher well densities result in greater local drawdown effects.
- Impervious areas associated with development can reduce ground-water recharge. Natural discharge to streams and springs will be reduced by an equivalent amount, in addition to the reduction due to capture by wells. Artificially enhancing recharge to exceed natural recharge rates will increase natural discharge to streams and springs and can offset capture of natural discharge by wells.

These conclusions are based on the simulated equilibrium response of the ground-water flow system in the hypothetical basin to the various pumping scenarios. The results of the steady-state (equilibrium) model are a very simplified representation of a system that, in reality, changes temporally in very complex ways. The equilibrium model allows evaluation of scenarios based on their long-term (equilibrium) effects, but simulation of the transient response of the system to seasonal variations in pumping or to long-term climatic changes (drought) would allow greater insight as to the time required to reach equilibrium and to the short-term as opposed to long-term response in different parts of the system. For example, when pumping from a confined aquifer near a stream on the drift plain, drawdown in the confined aquifer may be transmitted very quickly to the bluffs where it captures discharge to springs, whereas capture from the nearby stream may take much longer because of the time required for the drawdowns to be transmitted across the confining layer. Development of a transient version of the hypothetical basin model would involve (1) estimating values of storage coefficient for each hydrogeologic layer, (2) estimating the seasonal distribution of recharge in the basin, and (3) calibrating the model to expected conditions.

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