SIMULATION OF GROUND-WATER FLOW AND INFILTRATION FROM THE SUSQUEHANNA RIVER TO A SHALLOW AQUIFER AT KIRKWOOD AND CONKLIN, BROOME COUNTY, NEW YORK

By Richard M. Yager

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CONVERSION FACTORS

For the convenience of readers who prefer the metric (International System) units rather than the inch-pound unit used in this report, the following conversion factors may be used:

Multiply Inch-Pound Unit	By	To Obtain Metric Unit
	Length	
inch (in.)	25.40	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
acre	0.405	hectare (ha)
	Flow	
gallon per minute (gal/min)	0.06309	liter per second (L/s)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
inch per year (in/yr)	25.40	millimeter per year (mm/yr)
	Transmissivity	
foot squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
	Temperature	
degrees Fahrenheit (°F)	°C = 5/9 (°F-32)	degrees Celsius (°C)

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Abstract

A four-layer finite-difference model was developed to simulate ground-water flow and induced infiltration to an aquifer underlying the Susquehanna River in the Towns of Kirkwood and Conklin in Broome County. The aquifer consists of sand and gravel deposited in an ancestral river valley during the recession of glacial ice and is in hydraulic connection with the Susquehanna River. The aquifer in 1984 supplied 1.2 million gallons a day to well fields in Kirkwood and Conklin.

Horizontal hydraulic conductivity of the sand and gravel in the calibrated model ranges from 50 to 10,000 feet per day. Vertical hydraulic conductivity ranges from 1.0 to 80 feet per day. The riverbed thickness was estimated from results of piezometer tests to be 2 feet; the hydraulic conductivity of the riverbed was estimated to be 0.2 feet per day. Root-mean-square differences between computed drawdowns and drawdowns measured in observation wells and piezometers during aquifer tests at the Kirkwood well field ranged from 17 to 24 percent.

The sizes of the well-field catchment areas were estimated from a model-generated flow net showing the direction and rate of ground-water flow. The Kirkwood catchment area was estimated to be 250 acres, and the Conklin catchment area was 51 acres.

Ground-water budgets computed by steady-state simulations showed that 58 percent of the ground water withdrawn by the Kirkwood well field is derived from the Susquehanna River during periods of low river stage and low recharge. The factor to which induced-infiltration rate and size of well-field catchment areas are most sensitive is riverbed hydraulic conductivity.

INTRODUCTION

A sand and gravel aquifer system within the Susquehanna River valley in New York State supplies water to more than half the population of Broome County (fig. 1). Ground-water withdrawals from the aquifer in 1980 totaled 16.3 Mgal/d. The aquifer area occupies 21 mi^2 beneath the Susquehanna and Chenango River valleys, which intersect at Binghamton (fig. 1). The aquifer consists largely of unconsolidated deposits of sand and gravel left by melt-water streams draining glacial ice; the most productive deposits are discontinuous and at some sites are considered as separate aquifers (Waller and Finch, 1982, p. 48). The aquifer is in hydraulic connection with the Susquehanna and Chenango River.

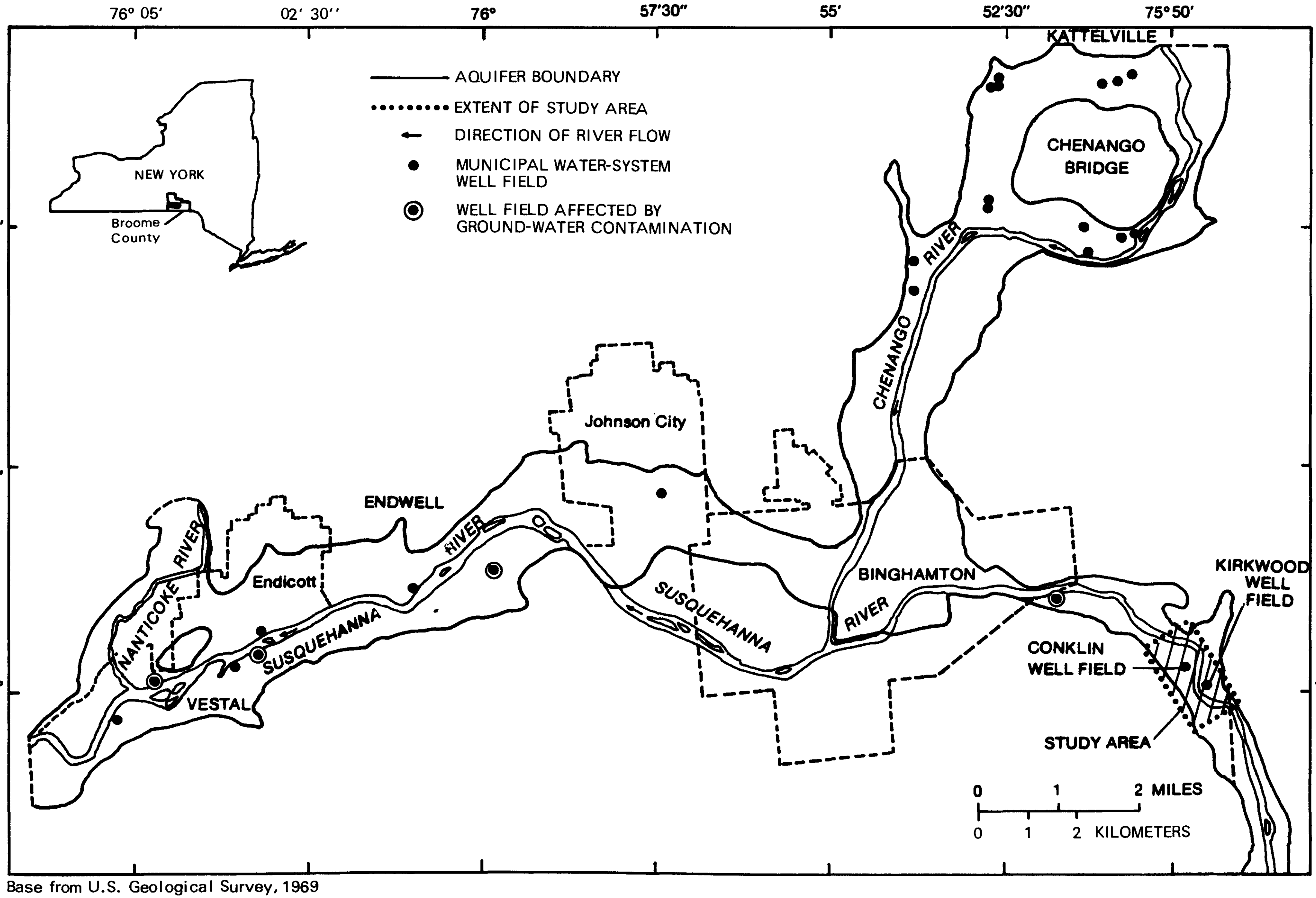

Figure 1.--Location of sand and gravel aquifer in Susquehanna River valley near Binghamton. (Modified from Waller and Finch, 1982.)

Since 1980, three of the five major municipal water systems that tap this aquifer system have shut down production wells because of ground-water contamination by solvents from industrial discharges. Two of these contaminated wells are in Vestal, at the west end of the aquifer, the other is in Conklin, at the east end of the aquifer (fig. 1). In addition, a well in Endicott, across the river from Vestal, requires operation of an additional well to prevent a plume of vinyl chloride in the aquifer from contaminating the public ground-water supply. These incidents of contamination have heightened public concern over the chemical quality of ground water in the aquifer and have drawn attention to the need for a management plan to protect this resource. Such a plan would require that local ground-water recharge areas be defined and amounts of recharge from different sources quantified.

Previous studies have shown that one important source of recharge to the aquifer is infiltration from the Susquehanna River (Randall, 1970, 1977). River water was also a probable source of bacterial contamination in a municipal well in Endicott (Randall, 1970); however, and bottom sediments of the river in several locations in Broome County have been found to contain heavy metals and chlorinated hydrocarbons (McDuffie and others, 1980). These findings indicate that infiltration from the river into the aquifer must be quantified and accounted for in ground-water-protection plans for this area.

In 1983, the U.S. Geological Survey, in cooperation with the Town of Kirkwood, began a study of the aquifer system that supplies water to well fields in the Towns of Kirkwood and Conklin (fig. 1). Objectives of the study were to (1) quantify the hydraulic properties that determine the rate of river infiltration to the aquifer, (2) identify the sources of recharge to the aquifer, and (3) delineate the well-field catchment areas. This information is needed for development of a plan to protect the aquifer against further contamination from surface sources as well as from the Susquehanna River.

Purpose and Scope

This report describes a computer simulation analysis of ground-water flow and river infiltration within the glacial sand and gravel aquifer in the Kirkwood-Conklin area. It includes (1) generalized geologic sections and maps showing the composition of the unconsolidated deposits in the area and the saturated thickness of the aquifer; (2) a discussion of the hydrology of the river and aquifer system that identifies sources of recharge to the aquifer and evidence of the hydraulic connection between the river and aquifer; (3) a description of the procedures to simulate the river and aquifer system, including a summary of results obtained from simulations of drawdowns observed during an aquifer test[1], and (4) maps and diagrams showing well-field catchment areas and sources of recharge to production wells operating within the modeled area. An appendix discusses methods used to estimate the hydraulic conductivity of the aquifer material from aquifer-test and piezometer-test data.

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[1]A related investigation of stream and aquifer interaction, supported by the U.S. Geological Survey's Regional Aquifer System Analysis (RASA) study (Lyford and others, 1984), provided significant data on this relationship.

Method of Investigation

This study was conducted from June 1983 through October 1984 and included the activities described below. Previous reports by Randall (1972, 1985) and Coates (1973) provided additional hydrogeologic data.

Wells and Borings

Split-spoon samples were collected with a hollow-stem auger during installation of 19 observation wells (GS1 through GS19, pl. 1) to identify the lithologic character of the subsurface materials. Geologic logs from 10 observation wells (V01 through V05 and MW1 through MW5m, see pl. 1) and 28 test borings that had been installed in the area by previous investigators were available. The test borings are identified on the maps herein by the latitude and longitude of the boring location.

Ground-Water Temperature and Levels

Ground-water temperatures were monitored monthly from July 1983 through February 1984 in observation wells near the Susquehanna River to establish the extent of river-water infiltration. Temperature was measured to within 0.1°C at 2-ft intervals to the bottom of each well. Ground-water levels were measured monthly in 42 observation wells (pl. 1) to delineate the patterns of ground-water flow and to document the water-table's response to seasonal fluctuations in river stage and to aquifer tests. At five locations, pairs of observation wells were screened at two depths to measure vertical gradients in the aquifer. Five additional observation wells were installed in the Susquehanna River by drive points with 6-inch screens.

Aquifer Tests

Two aquifer tests were conducted in the Kirkwood well field in February and October 1984, and piezometer tests were conducted in several observation wells to estimate the hydraulic characteristics of aquifer and riverbed material. Data were also obtained from an aquifer test conducted in the Conklin well field in October 1983 by other investigators.

Simulation Model

A simulation model was developed from the field data to compute the direction and rate of ground-water movement within the aquifer and the volume of water recharging the aquifer from the major sources, including the Susquehanna River.

Acknowledgments

This investigation was done in cooperation with the Town of Kirkwood, which assisted in the installation of observation wells and in the aquifer tests. Additional support was provided by the U.S. Geological Survey's Regional Aquifer System Analysis (RASA) study (Lyford and others, 1984), which provided suggestions in data collection and model simulations during the investigation. Donald Coates, professor of geology at State University of New York at Binghamton, helped interpret the geology of the area.

DESCRIPTION OF STUDY AREA

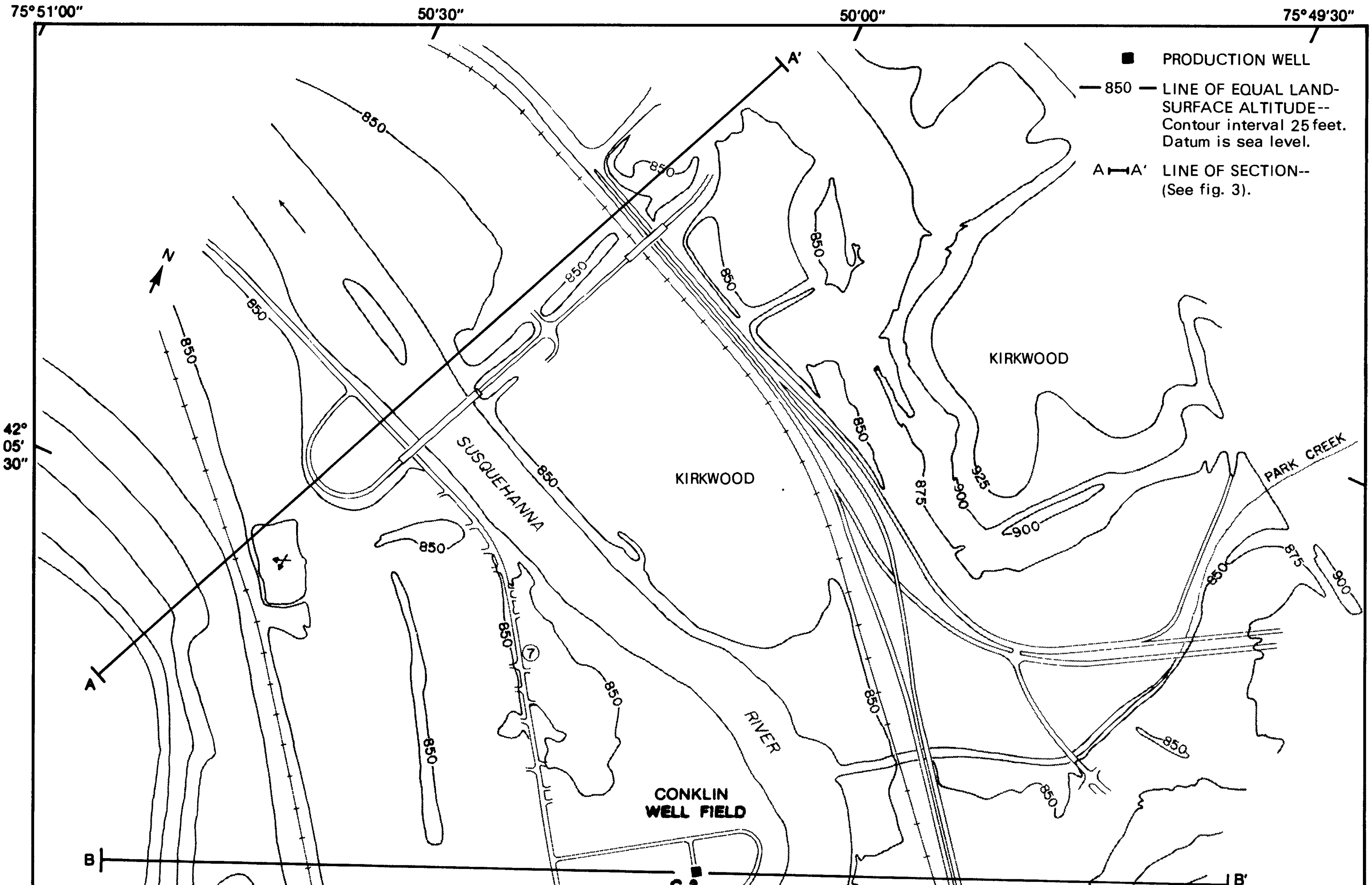
The Kirkwood-Conklin area (fig. 2) is in the Appalachian Plateaus province in south-central New York State, a glaciated plateau of moderate relief. Altitudes range from 1,600 ft above sea level on hilltops along the river to 840 ft on the valley floor. The area studied occupies approximately 1 mi^2 of the valley floor just east of the City of Binghamton near the confluence of Park Creek and the Susquehanna River.

Glacial History

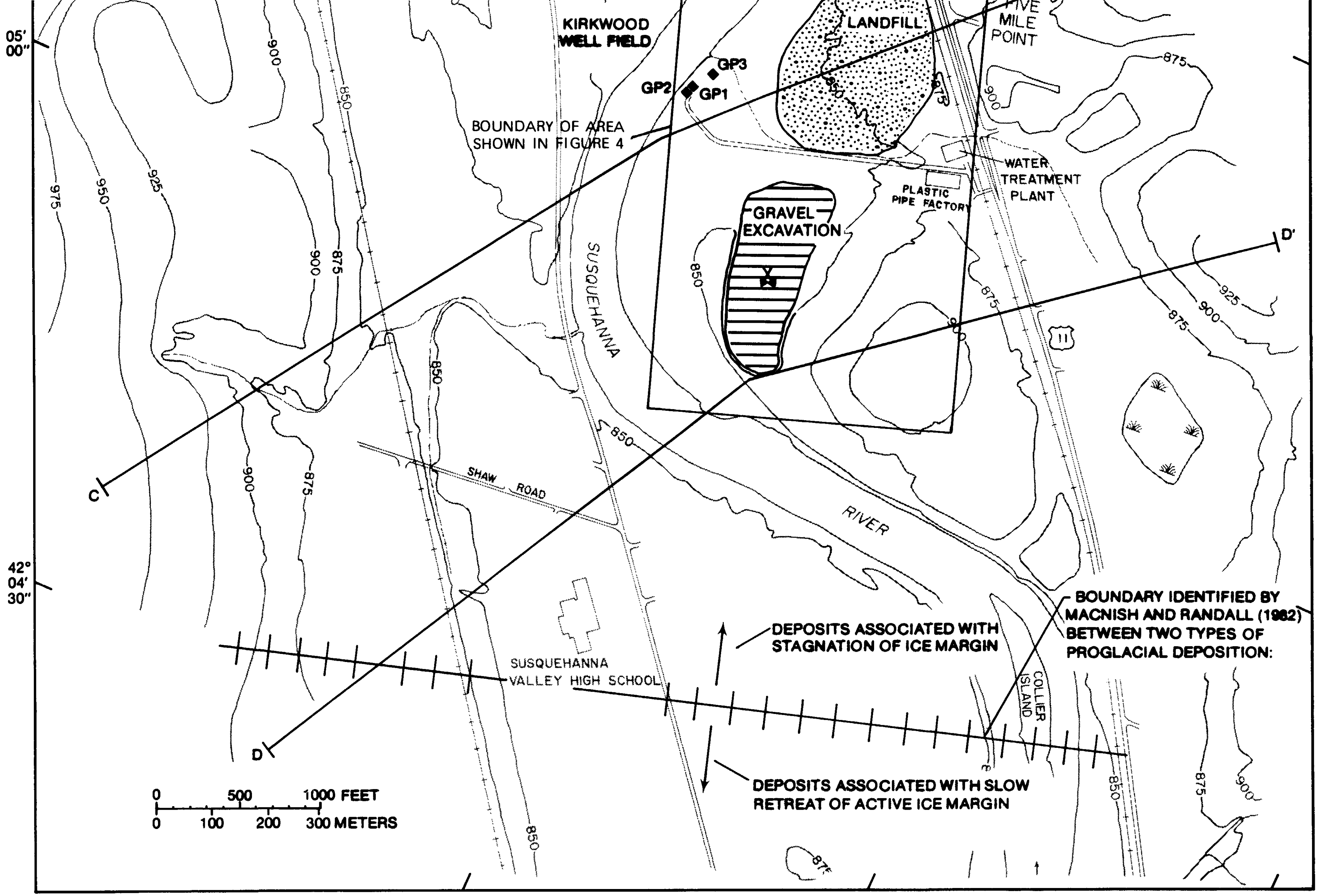
The Susquehanna River valley was carved into the Appalachian Plateau by a preglacial river and was deepened and widened through erosion by glacial ice. As the glacier advanced southward, it carried with it material from the north. During the glacial recession, both the ice and meltwater streams flowing from the ice margin deposited unconsolidated material on the land surface; this material partly filled the deep valley to form valley-fill deposits and thinly covered the adjacent upland areas. Well-sorted materials deposited by water are generally less compact than the poorly sorted materials deposited by the ice. These latter deposits, referred to as till, are found beneath the other unconsolidated deposits and are typically the only deposits covering the bedrock in steep upland areas. The distribution of unconsolidated deposits in this area is uneven because the rates of glacial ice movement and sediment deposition differed from place to place and changed through time.

During the waning stages of glaciation, the receding ice tongue in the Susquehanna valley was bounded on the south by a series of lakes that had formed behind successive sediment dams deposited by earlier meltwater. Meltwater streams emanating from the ice dropped coarse sand and gravel along the stream channels and, where they entered these lakes, they formed channel bars and wide deltas. The finer sand, silt, and clay carried by these streams did not settle immediately but were carried farther from shore, where they settled to the lake bottom to form silt and clay deposits. Ice-contact deposits, sand, and gravel deposited near the ice front range from very well sorted to poorly sorted and may contain lenses of silty sand and gravel or till. Outwash deposits (sand and gravel carried further from the ice front by meltwater streams) generally consist of homogeneous, well-sorted material. Today, the coarse sand and gravel deposits constitute the principal aquifer in the Conklin-Kirkwood area and are generally in good hydraulic contact with the Susquehanna River.

The extent and thickness of coarse deposits in a given area depends on the rate of the glacial retreat. Where the ice margin stagnated, coarse sand and gravel spread across the valley to form thick deposits of outwash. Where the ice margin melted more rapidly, the outwash deposits formed only along the valley walls, where they now remain as terraces above the modern flood plain. In these areas, only a thin layer of coarse sediment was deposited upon the valley floor.



PRODUCTION WELL
850 LINE OF EQUAL LAND-SURFACE ALTITUDE--Contour interval 25 feet. Datum is sea level.
A A' LINE OF SECTION--(See fig. 3).
75°51'00"
50'30"
50'00"
75°49'30"
42° 05' 30"
KIRKWOOD
KIRKWOOD
SUSQUEHANNA
RIVER
PARK CREEK
CONKLIN WELL FIELD
A
A'
B
B'
N
850
875
900
925
7



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 2.--Major geographic features of Kirkwood-Conklin area.

Geologic Setting

Valley-Fill Deposits

The valley-fill deposits in the Kirkwood-Conklin area range from 100 to 130 ft thick. Generalized geologic sections (fig. 3) developed from logs of wells and test borings illustrate the distribution of the valley-fill deposits; the locations of geologic sections, wells, and test borings are shown in plate 1.

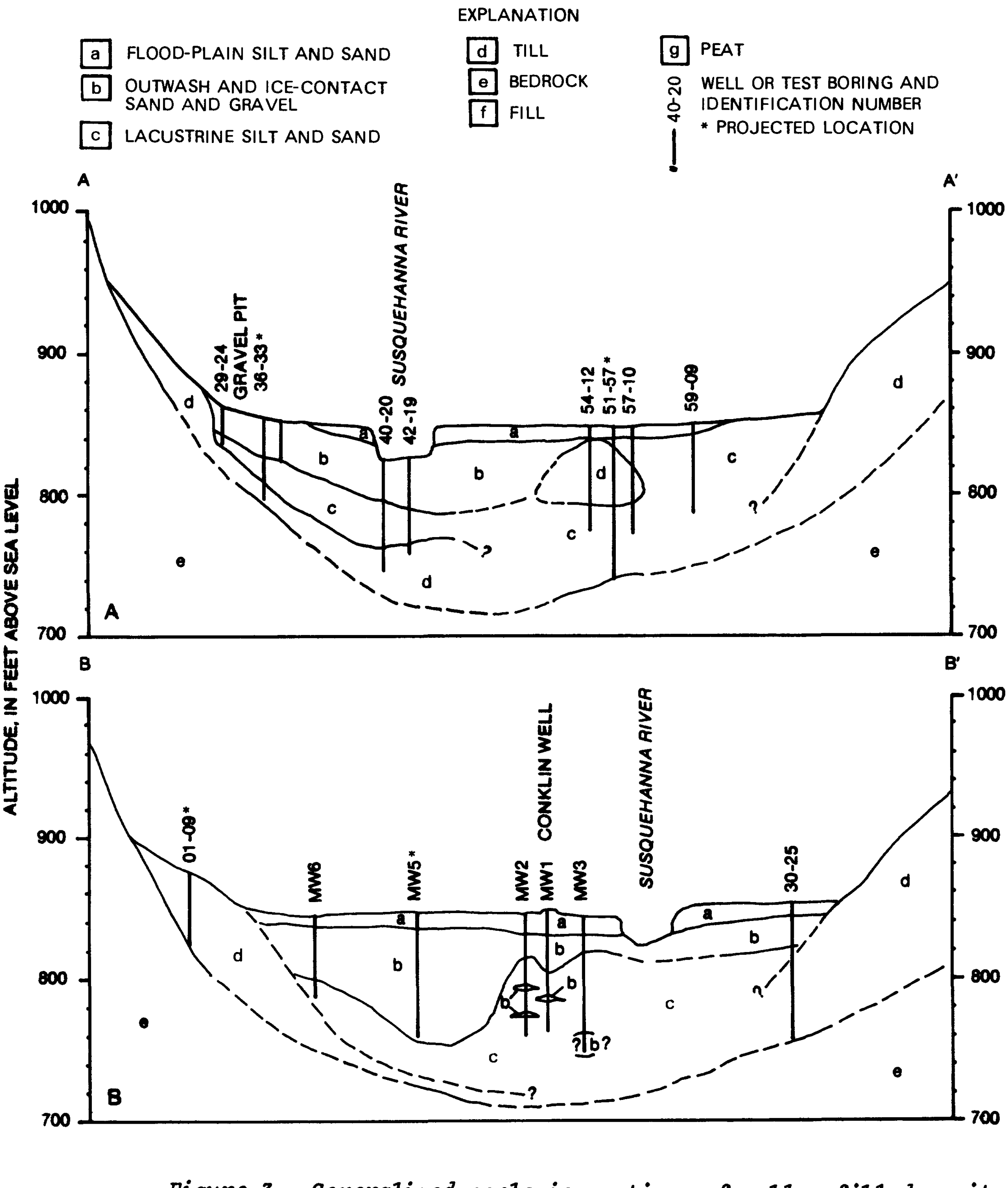
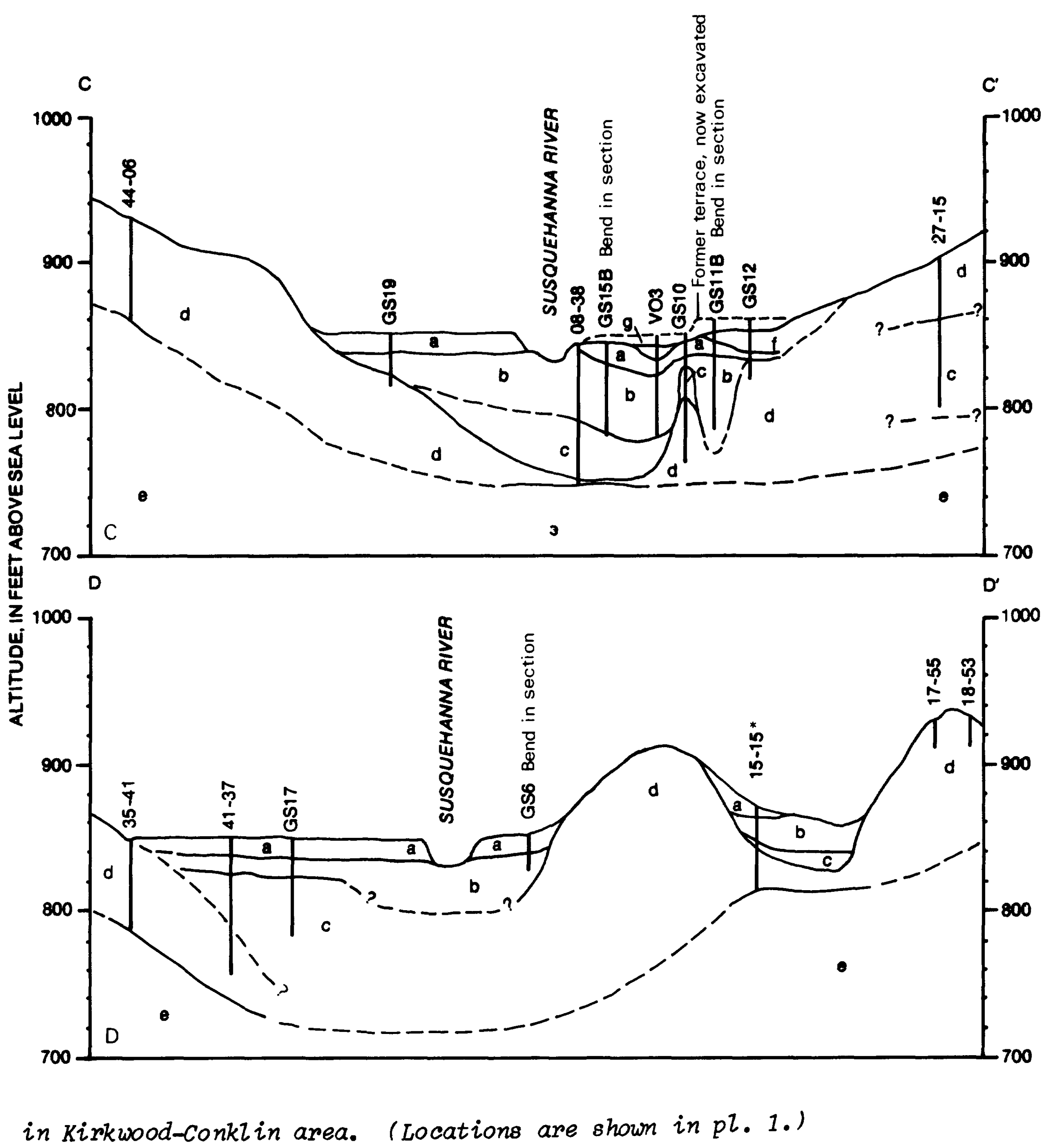


Figure 3.--Generalized geologic sections of valley-fill deposits

The valley-fill deposits vary in composition and correspond to two general types of depositional patterns, as described by MacNish and Randall (1982, p. 16-21). The study area lies where the valley orientation shifts from northwest-southeast to north-south for 1/2 mile; thus, it was probably a transition zone between two patterns of proglacial deposition (fig. 2). The relatively thick layer of sand and gravel in the northern part of the area is typical of valleys in which the ice margin stagnated and is probably a mixture of ice-contact and outwash materials. The relatively thin deposits of sand and gravel beneath the valley floor and terraces along the valley walls in the southern part of the area is typical of a rapid retreat of the ice margin, and these deposits are probably outwash.



in Kirkwood-Conklin area. (Locations are shown in pl. 1.)

Flood-plain deposits.--Flood-plain deposits of silt and fine sand 8 to 15 ft thick cover most of the valley floor. A soils survey of Broome County indicates the hydraulic conductivity of these deposits to range between 1.3 and 13 ft/d (Giddings and others, 1971, p. 37). However, most of these deposits are above the river and are therefore largely unsaturated and do not serve as conduits for ground water. Only during periods of high river stage do these deposits become saturated.

Sand and gravel.--The layer of sand and gravel beneath the flood-plain deposits ranges in thickness from 10 to 70 ft. Results of analyses of aquifer-test data from the Conklin and Kirkwood well fields (described in the appendix) indicate the hydraulic conductivity of the sand and gravel to range from 300 to 4,900 ft/d within the study area. The higher value was computed from data recorded at the Kirkwood well field, where the sand and gravel deposit is 40 to 50 ft thick and provides 1.4 Mgal/d to production wells. The hydraulic conductivity is probably lower near the Conklin well field, where the deposit is 10 to 20 ft thick and provides 0.3 Mgal/d to the production well. These hydraulic-conductivity values are within the range of values reported by Lyford and others (1984) for outwash materials. Sand and gravel layers beneath the terraces along the valley wall, although less extensive than the ice-contact and outwash deposits, provide sufficient quantities of ground water for domestic use (2 to 5 gal/min).

Lacustrine deposits.--Beneath the outwash sand and gravel are lacustrine deposits of sand and silt that range from 20 to 100 ft thick. Slug and bail tests (described in the appendix) indicate the hydraulic conductivity of the sand and silt to range between 0.2 and 1.8 ft/d. These deposits are 2 to 3 orders of magnitude less permeable than the sand and gravel and do not yield sufficient quantities of ground water for domestic use.

Till.--The outwash and lacustrine deposits in most places are underlain by till, a mixture of silt, sand, and gravel. Till also thinly covers the top of the bedrock hills that form the valley wall and may reach a thickness of 100 ft in small mounds on the valley floor. These till mounds border both sides of the valley in the southern part of the study area; elsewhere they lie near the center of the valley but are buried beneath lacustrine and outwash deposits. (See section A-A', fig. 3.)

Hydraulic conductivity of till generally ranges from 10^{-4} to 10^{-3} ft/d (Freeze and Cherry, 1979, p. 29). Where exposed at land surface, till may yield small quantities of water to shallow wells through fractures in the till matrix, but at depth, although saturated, it does not transmit sufficient quantities of water for domestic use.

Bedrock

The upper 400 ft of bedrock underlying the Kirkwood-Conklin area is predominantly shale. The bedrock has been uplifted and dips to the south with a gradient of about 40 ft/mi. Fractures and bedding planes form a small part of the rock volume and provide the only significant void spaces in which water can be stored and transmitted. Bedrock wells generally supply quantities of water that are sufficient for individual households.

Landfill and Gravel Excavation

Excavation of sand and gravel has altered the land surface and drainage channels to the Susquehanna River near the Kirkwood well field (fig. 4). Section C-C' (fig. 3) indicates the former location of the sand and gravel terrace on the east side of the river valley. The previous extent of this terrace is indicated in figure 5 along with the present land use. Ten to 15 ft of sand and gravel have been removed from the terrace by excavation since the 1950's. The excavations exposed till to the north of the access road to the Kirkwood well field and medium to fine sand to the south.

Excavated areas of the terrace have been filled with various materials, mainly backfill and construction debris. Excavations in the southern part of the terrace were recently terminated, and the excavated area was backfilled with 5 ft of till to restrict infiltration of surface runoff. Backfill was also used to build the access road to the Kirkwood well field and to build part of the foundation for a plastic-pipe factory (fig. 4). A landfill is still operating in the northern part of the terrace (fig. 4).

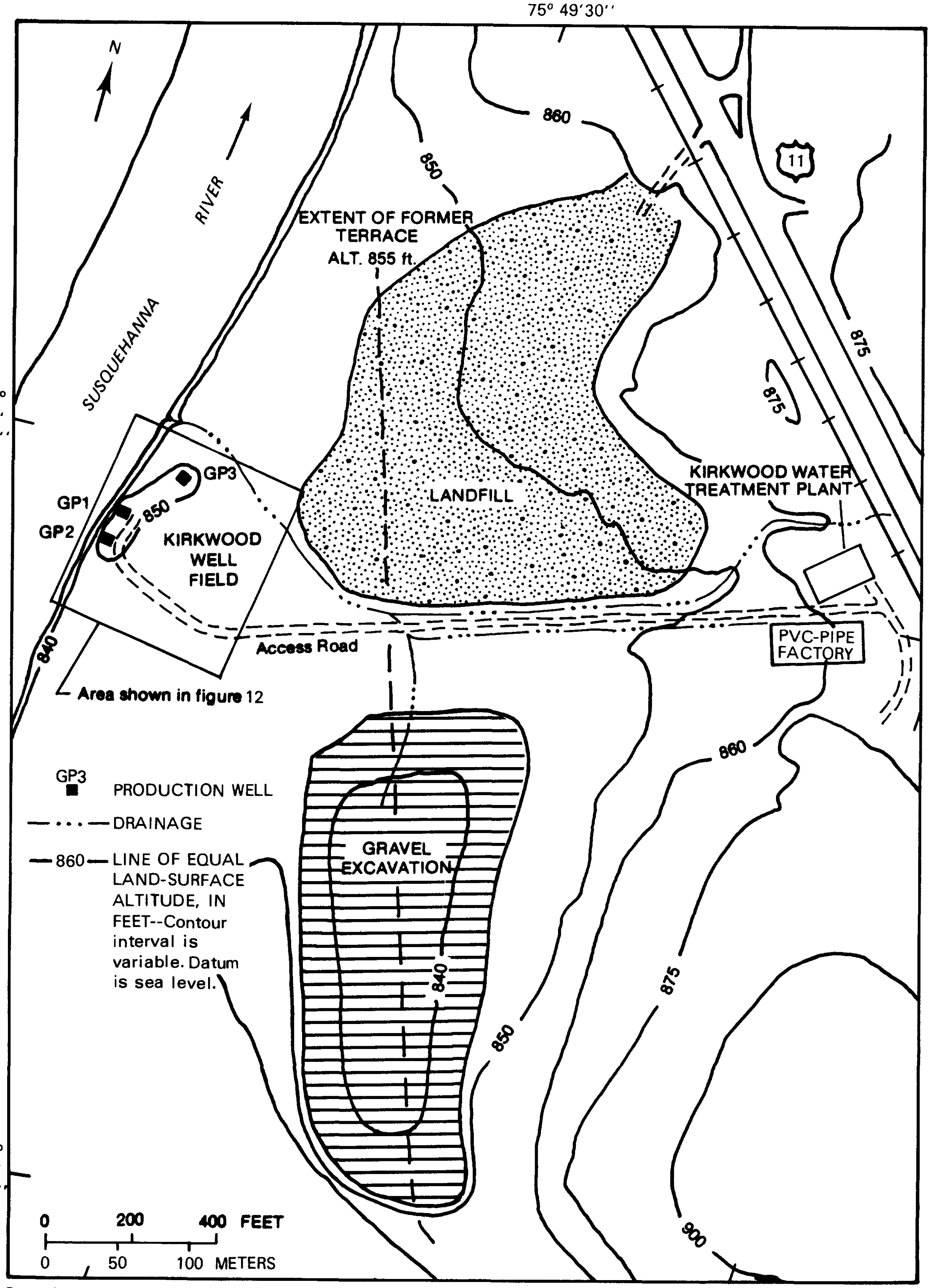
Fill materials used in the excavated areas generally have a lower hydraulic conductivity than the sand and gravel they replace and thus limit ground-water flow through the area formerly occupied by the terrace. The material used to backfill the excavation in the southern part of the former terrace has the lowest hydraulic conductivity, about 0.03 ft/d as estimated from compaction of the material. Backfill used in construction of the access road and the factory foundation probably has low to intermediate permeability (0.1 to 10 ft/d). The landfill materials in the northern part of the excavated area consist mostly of wood, metal, plastic, concrete, and other debris obtained through building demolition. These materials are the most permeable of any fill in the area and may approach the permeability of the original sand and gravel deposits.

Runoff and ground water from the excavated parts of the terrace flow into drainage ditches. A natural channel that carried runoff from upland areas east of U.S. Route 11 has been rerouted to the drainage system along the south edge of the landfill. Discharge measurements indicated that about 90 percent of the discharge of the drainage system (0.10 to 0.15 ft^3/s) during dry weather is derived from floor drains in the plastic-pipe factory; the remainder is ground-water seepage from tile drains beneath the factory and ground water from along the south border of the landfill.

HYDROLOGY

Ground Water

Although ground water can be obtained in all parts of the Kirkwood-Conklin area, only the sand and gravel deposit that constitutes the principal aquifer provides sufficient quantities (at least 200 gal/min) for municipal water supplies. The hydraulic conductivity of the sand and gravel is much higher than that of adjacent materials.



Base from Susquehanna River Basin
Commission, 1974, 1:2400

Figure 4.--Landfill, gravel excavation, and extent of sand and gravel terrace before excavation near Kirkwood well field. (Location is shown in fig. 2.)

Saturated Thickness

The availability of ground water in the sand and gravel aquifer is related to the saturated thickness of the aquifer. The area of maximum saturated thickness (30 to 50 ft) forms an elliptical pattern that extends from the Kirkwood well field westward to just south of the Conklin well field (fig. 5). The saturated thickness decreases to less than 10 ft just northeast and southeast of the zone of maximum thickness in areas where till deposits lie near the land surface, and also along the valley walls (fig. 5).

No subsurface information is available between Route 7 and the Susquehanna River south of the Kirkwood well field nor near the confluence of Park Creek with the Susquehanna River to the north. The aquifer thickness in these areas is estimated to be 10 to 30 ft.

Flow Patterns

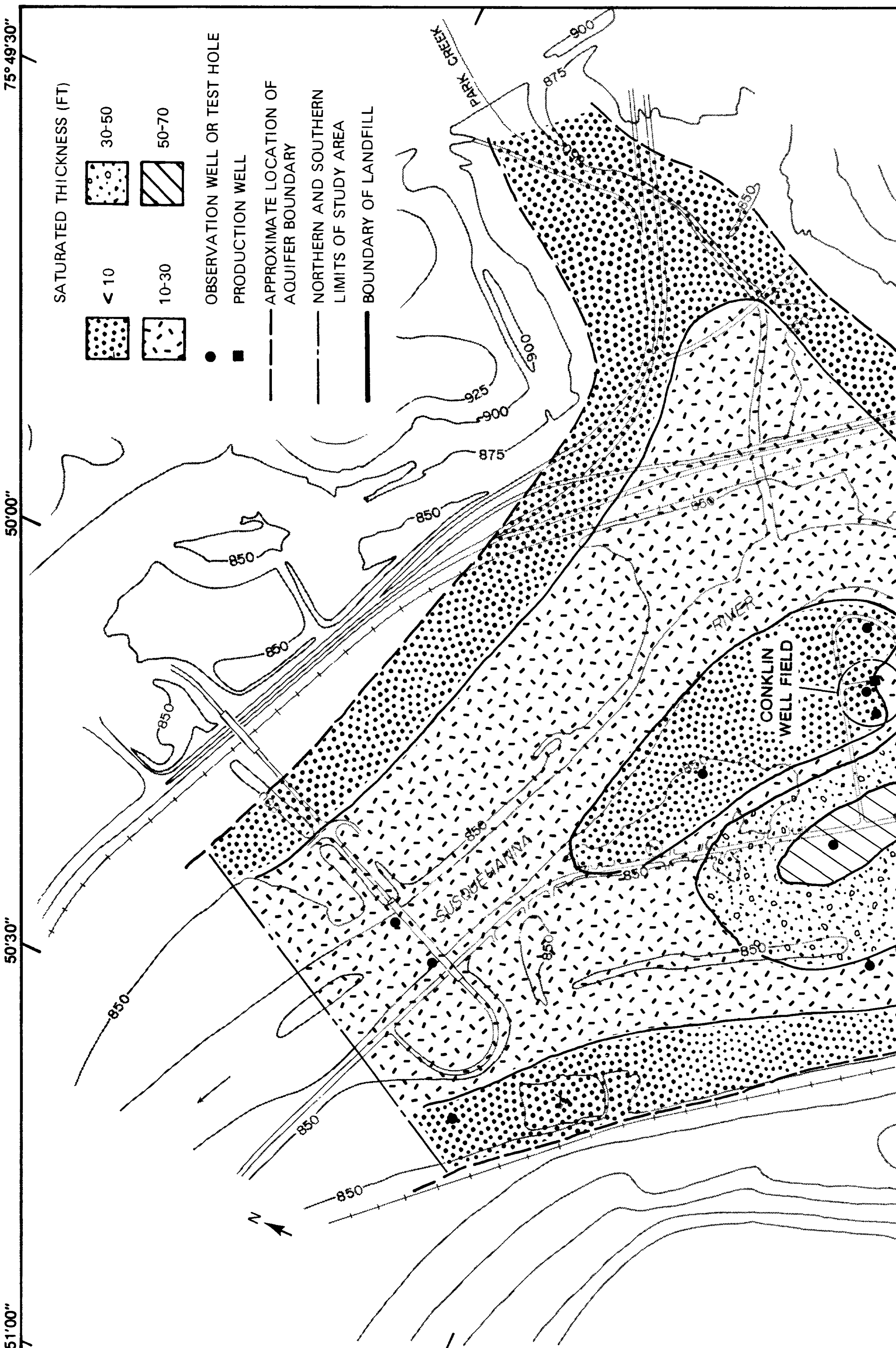
Ground water in the sand and gravel aquifer generally flows from recharge areas along the valley walls toward the Susquehanna River. Approximate directions of ground-water flow through the aquifer, based on ground-water altitudes and river stage measured in April 1984 and October 1984, are shown in figure 6 for periods when recharge and river discharge were near a seasonal maximum and minimum, respectively.

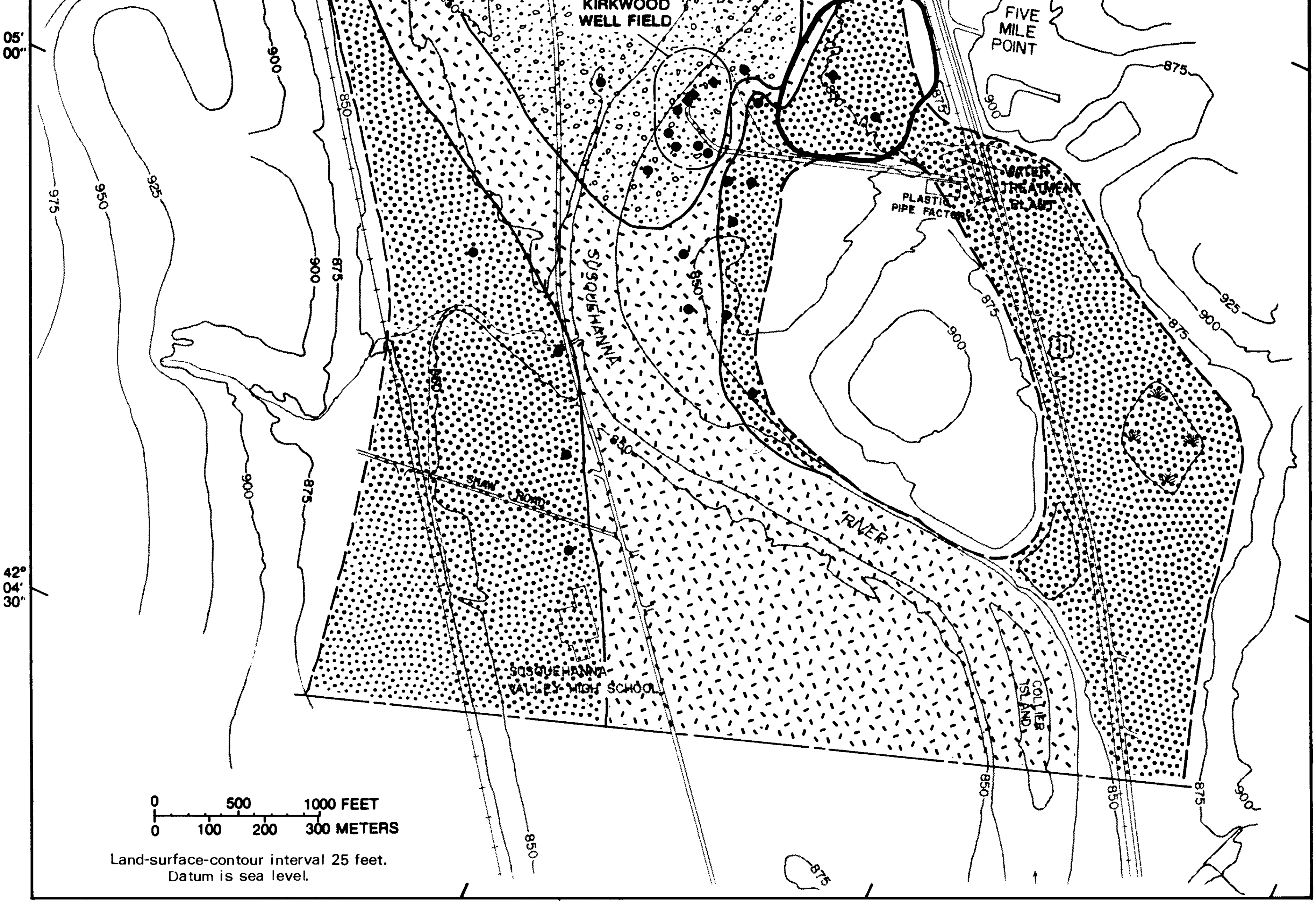
Pumping has altered the natural flow pattern in the vicinity of the Kirkwood and Conklin well fields such that ground water that previously discharged to the river is now captured by production wells. The influence of these pumping centers on the regional ground-water flow patterns is most evident from the contours of average hydraulic head of October 1984 (fig. 6B). The cone of depression associated with each pumping center induces infiltration of river water into the aquifer. The river stages shown in figures 6A and 6B indicate a vertical hydraulic gradient from the river and to the aquifer where the cones of depression reach the river. Flow of ground water to the production wells is discussed in detail in the section on model simulation of ground-water withdrawals.

Ground water beyond the landfill northeast of the Kirkwood well field flows westward through the landfill and into the outwash sand and gravel near the Susquehanna River. The primary source of the flow is ground water in the terrace deposit east of Route 11. The thickness of saturated material in the landfill is about 5 ft.

Recharge and Discharge

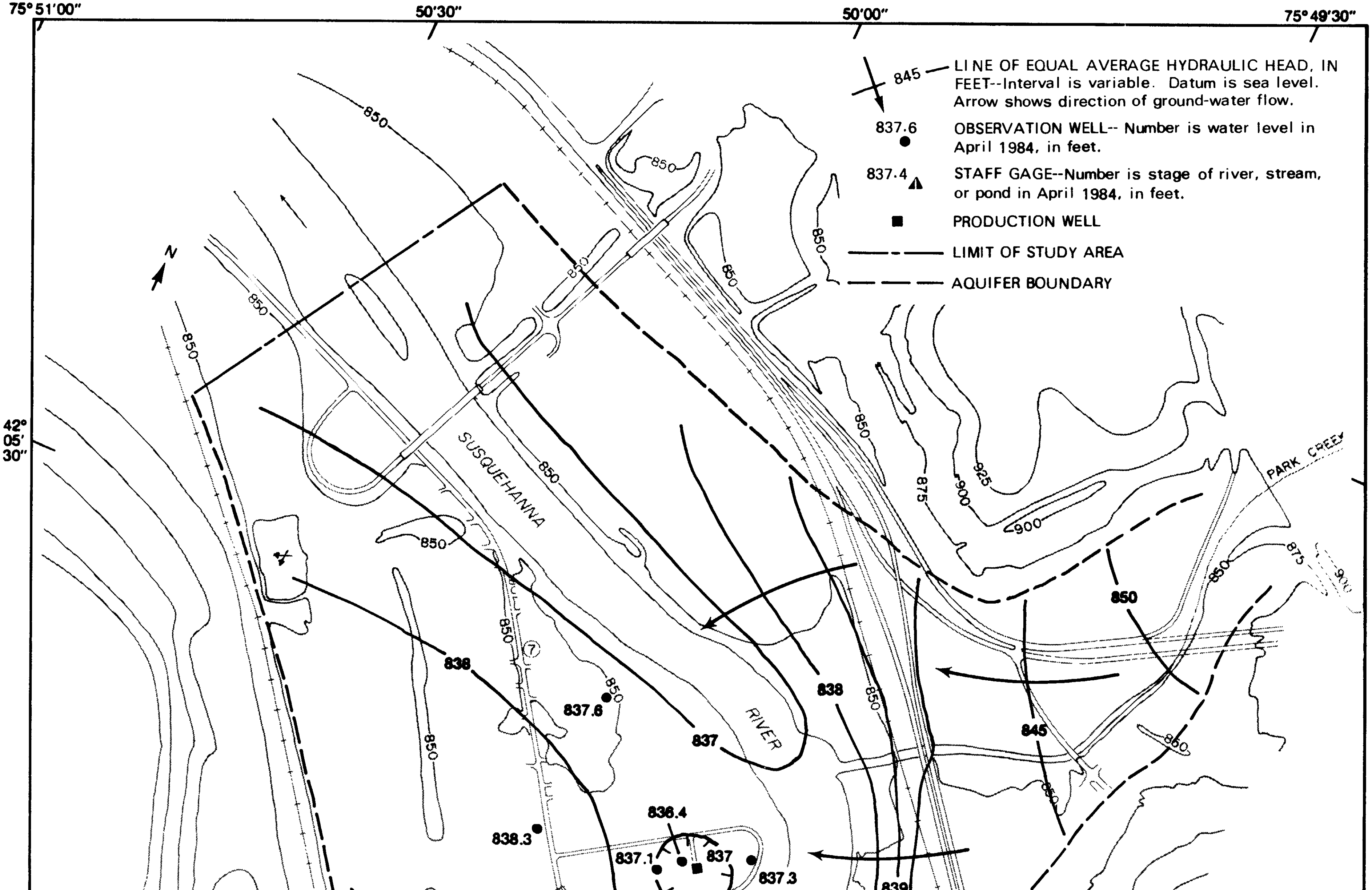
The primary source of recharge to the sand and gravel aquifer is precipitation in the valley and uplands. Recharge to the aquifer includes (1) precipitation infiltrating the flood-plain alluvium, (2) runoff from upland areas infiltrating from tributary streams and along valley wells, (3) underflow from the part of the aquifer that lies upstream, beyond the boundary of the study area, (4) upward leakage from underlying deposits, and (5) infiltration from the Susquehanna River near pumping production wells. Ground water discharges


SATURATED THICKNESS (FT)
< 10
10-30
30-50
50-70
OBSERVATION WELL OR TEST HOLE
PRODUCTION WELL
APPROXIMATE LOCATION OF AQUIFER BOUNDARY
NORTHERN AND SOUTHERN LIMITS OF STUDY AREA
BOUNDARY OF LANDFILL
PARK CREEK
CONKLIN WELL FIELD
SUSQUEHANNA
RIVER
900
875
850
925
N
75°49'30"
50'00"
50'30"
75°51'00"
42°05'30"

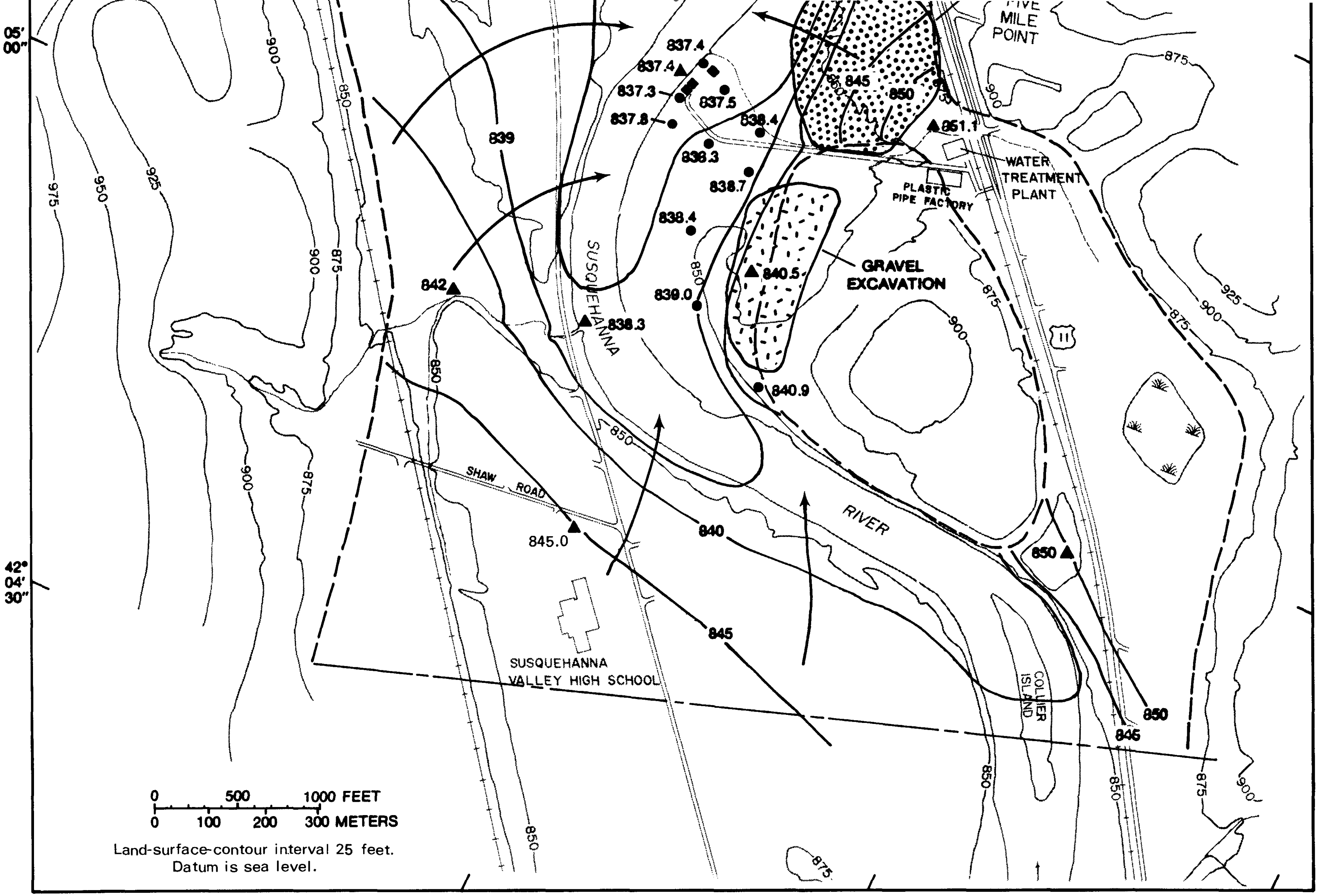


Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 5.--Saturated thickness of aquifer in Kirkwood-Conklin area.

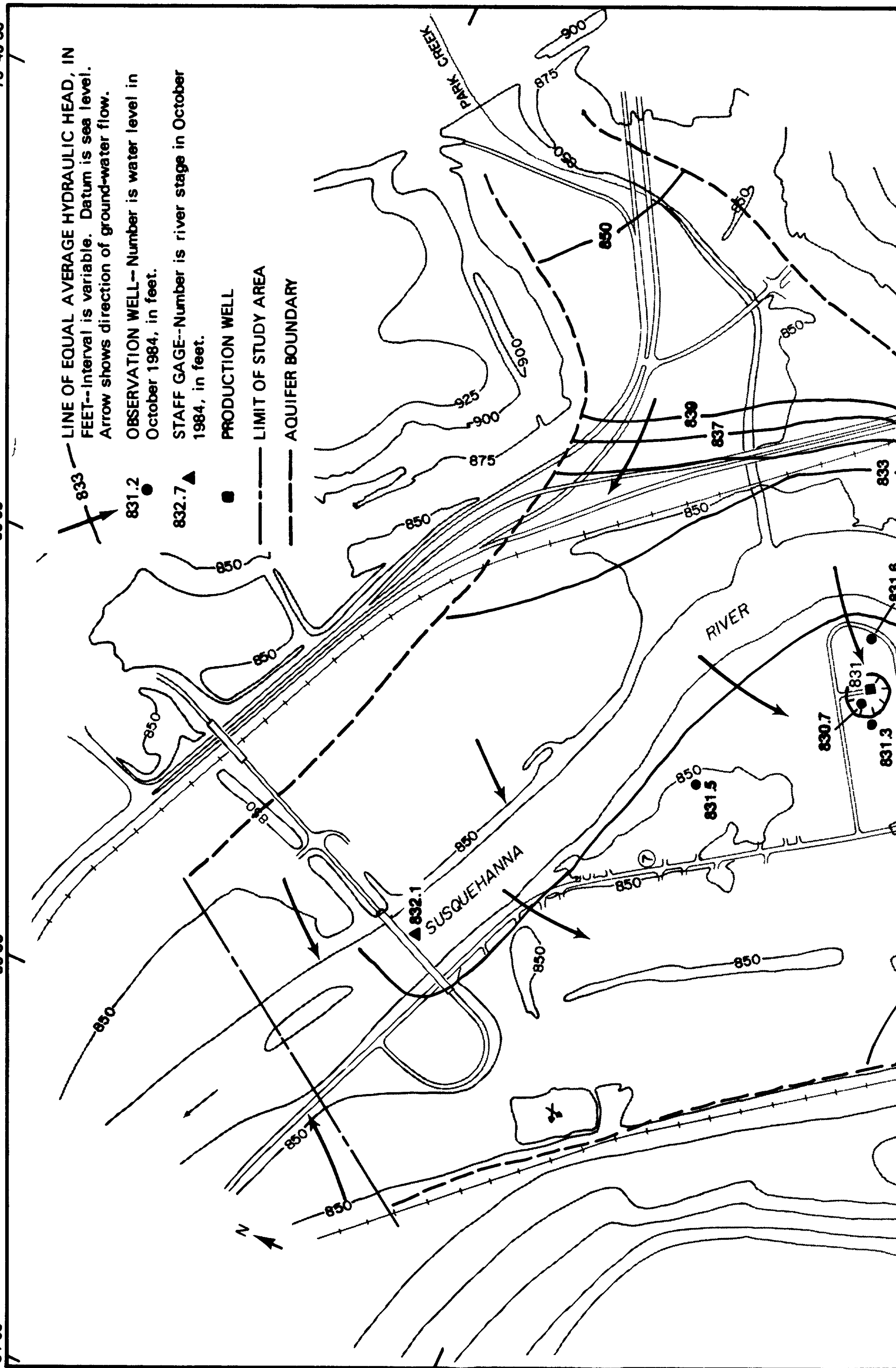


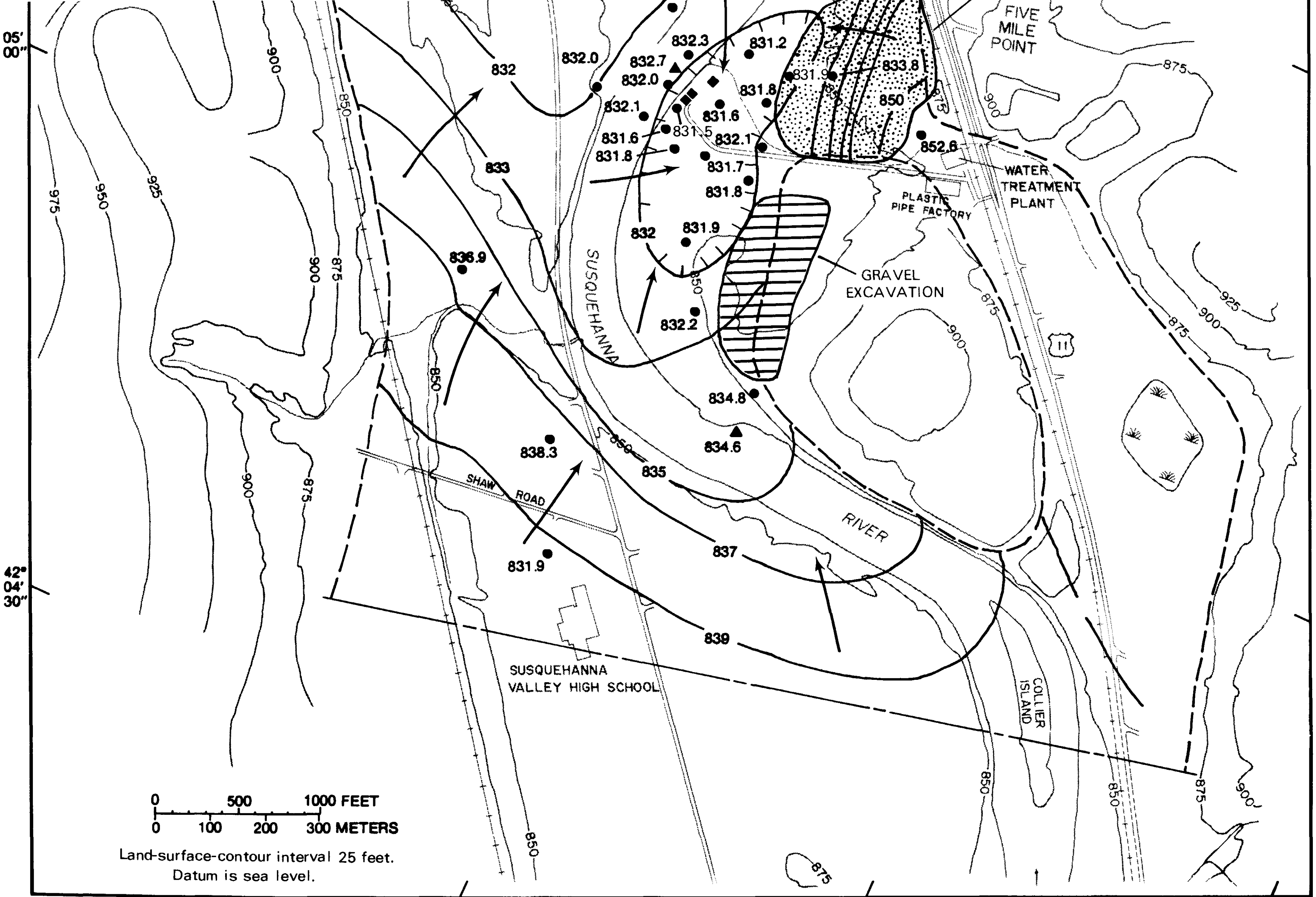
75°51'00"
50'30"
50'00"
75°49'30"
42°
05'
30"
845 — LINE OF EQUAL AVERAGE HYDRAULIC HEAD, IN FEET--Interval is variable. Datum is sea level. Arrow shows direction of ground-water flow.
837.6 OBSERVATION WELL-- Number is water level in April 1984, in feet.
837.4 STAFF GAGE--Number is stage of river, stream, or pond in April 1984, in feet.
PRODUCTION WELL
LIMIT OF STUDY AREA
AQUIFER BOUNDARY
SUSQUEHANNA
RIVER
PARK CREEK
N
850
875
900
925
845
838
837
839
837.6
838.3
836.4
837.1
837.3



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 6A.--Average hydraulic head and direction of ground-water flow with production wells GP1 and C-2 in operation in April 1984, a period of maximum recharge and river discharge.


833 LINE OF EQUAL AVERAGE HYDRAULIC HEAD, IN FEET--Interval is variable. Datum is sea level. Arrow shows direction of ground-water flow.
831.2 OBSERVATION WELL--Number is water level in October 1984, in feet.
832.7 STAFF GAGE--Number is river stage in October 1984, in feet.
PRODUCTION WELL
LIMIT OF STUDY AREA
AQUIFER BOUNDARY
PARK CREEK
SUSQUEHANNA
RIVER
N
75°49'30"
50'00"
50'30"
75°51'00"
42°05'30"
832.1
831.5
830.7
831.3
831
839
837
833
850
875
900
925



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 6B.--Average hydraulic head and direction of ground-water flow with production wells GP1 and C-2 in operation in October 1984, a period of minimum recharge and river discharge.

from the aquifer to (1) production wells in Kirkwood and Conklin, (2) the northern (downvalley) part of the aquifer beyond the boundary of the study area, (3) the Susquehanna River, (4) tributary streams, and (5) the atmosphere through evapotranspiration. Losses into tributary streams and through evapotranspiration are assumed to be only a minor percentage of the total ground-water discharge because the water table is generally 8 to 12 ft below land surface, which is deeper than most stream channels and the rooting depth of most vegetation.

The annual volumes of ground-water recharge and discharge estimated from information obtained in previous studies and calculated directly from field measurements are summarized in table 1 for a section of the aquifer within the 1-mi^2 study area shown in figure 6. Recharge from precipitation on the flood plain was estimated to be 1.0 Mgal/d; this was based on a recharge rate of 22 in/yr reported by Randall (1977, p. 17).

Infiltration from tributary streams that drain upland areas was estimated by a method presented in MacNish and Randall (1982, p. 37). The method estimates the potential recharge rate from tributary streams by multiplying the length of channel crossing the aquifer with a gradient of less than 1 percent by 650 (gal/d)/ft. This potential rate is compared with the long-term flow duration of the stream taken from a plot of average flow duration for upland basins within the Susquehanna River basin, expressed per unit area (MacNish and Randall, 1982, fig. 15). Where streamflow is estimated to fall below the potential recharge rate, the estimate of average recharge is reduced to allow for periods of deficient flow. About 90 percent of the estimated 1.1 Mgal/d of infiltration from tributary streams is from the Park Creek valley (fig. 2). Infiltration along valley walls in areas not drained by streams was assumed to be equal to the 90-percent flow duration discharge for upland basins.

Upward leakage through the lacustrine sand and silt deposits that underlie the aquifer is extremely small because these deposits have low hydraulic conductivity. Randall (1985, pl. 3), using a ground-water-flow model, obtained vertical hydraulic-conductivity values of 1×10^{-5} to 5×10^{-5} ft/d for a silty sand layer beneath an aquifer in Johnson City, 7 mi downstream (fig. 1). Upward leakage in the Kirkwood-Conklin area was estimated from Darcy's law:

$$Q = KAi \tag{1}$$

where: Q is the volume of upward leakage, ft^3/d
K is the hydraulic conductivity, ft/d
A is the cross-sectional area of flow, ft^2
i is the hydraulic gradient, ft/ft (the difference between head in observation wells screened in the aquifer and in the underlying lacustrine deposit).

Using the estimate of vertical hydraulic conductivity reported by Randall (1985) and an average hydraulic gradient of 1.0 ft/ft yields an upward leakage value of 0.006 Mgal/d--negligible relative to the other recharge sources.

Underflow into and out of the study area was estimated by Darcy's law from saturated-thickness values shown in figure 5, an average hydraulic conductivity of 1,000 ft/d (calculated from aquifer-test data in the appendix), and hydraulic gradients shown in figure 6A. About 4.9 Mgal/d is estimated to enter the section of the aquifer within the study area across the south (upstream) boundary under a gradient of 0.010 ft/ft. About 0.8 Mgal/d leaves across the north (downstream) boundary of the study area.

The net ground-water discharge to the Susquehanna River (5.0 Mgal/d, table 1) represents the volume of ground water discharged from the aquifer to the river minus the volume of infiltration from the river to the aquifer. The estimated net gain in river discharge of about 7 ft^3/s within the study area was too small to calculate directly from discharge measurements; a measurement error of 5 percent at the lowest river discharge recorded during the study (200 ft^3/s) would be 10 ft^3/s. Consequently, neither the infiltration from the river to the aquifer nor the discharge from the aquifer to the river could be measured directly, but both were estimated from model simulations described further on.

Table 1.--Estimated volumes of ground-water recharge and discharge in Kirkwood-Conklin aquifer system.

[Volumes are in millions of gallons per day.]

SOURCES	Volume	Percentage	DISCHARGES	Volume	Percentage
Recharge from precipitation on the flood plain	1.0	14	Production wells	1.2	17
Infiltration from upland areas	1.1	16	Net ground-water discharge to river	5.0	72
Underflow into study area	4.9	70	Underflow out of study area	.8	11
Total	7.0	100		7.0	100

Surface Water

The Susquehanna River meanders through the study area in a channel incised 10 ft into the flood-plain deposits. It drains an area of 2,232 mi^2. Discharge is generally greatest during snowmelt periods in the spring and is lowest just before the growing season ends in the fall. The mean daily discharge of the Susquehanna River during 1983-84 at a gaging station in Conklin, 4.5 mi upstream from the study area, and the median mean daily discharge during 1932-80, are plotted in figure 7. Low flows in the fall are typically near 500 ft^3/s, whereas high flows in the spring may exceed 10,000 ft^3/s. High river stages frequently cause flooding of the valley floor near the river. Peak discharge of a 100-year flood at the Kirkwood-Conklin area is estimated to be 63,900 ft^3/s, whereas peak discharge of a 10-year flood is estimated to be 46,500 ft^3/s (Zembrzuski and Dunn, 1979, p. 48). A 100-year flood would cause a river stage of 854 ft above sea level in the Kirkwood-Conklin area. (See altitude contours in fig. 6 to estimate extent of flooding.)

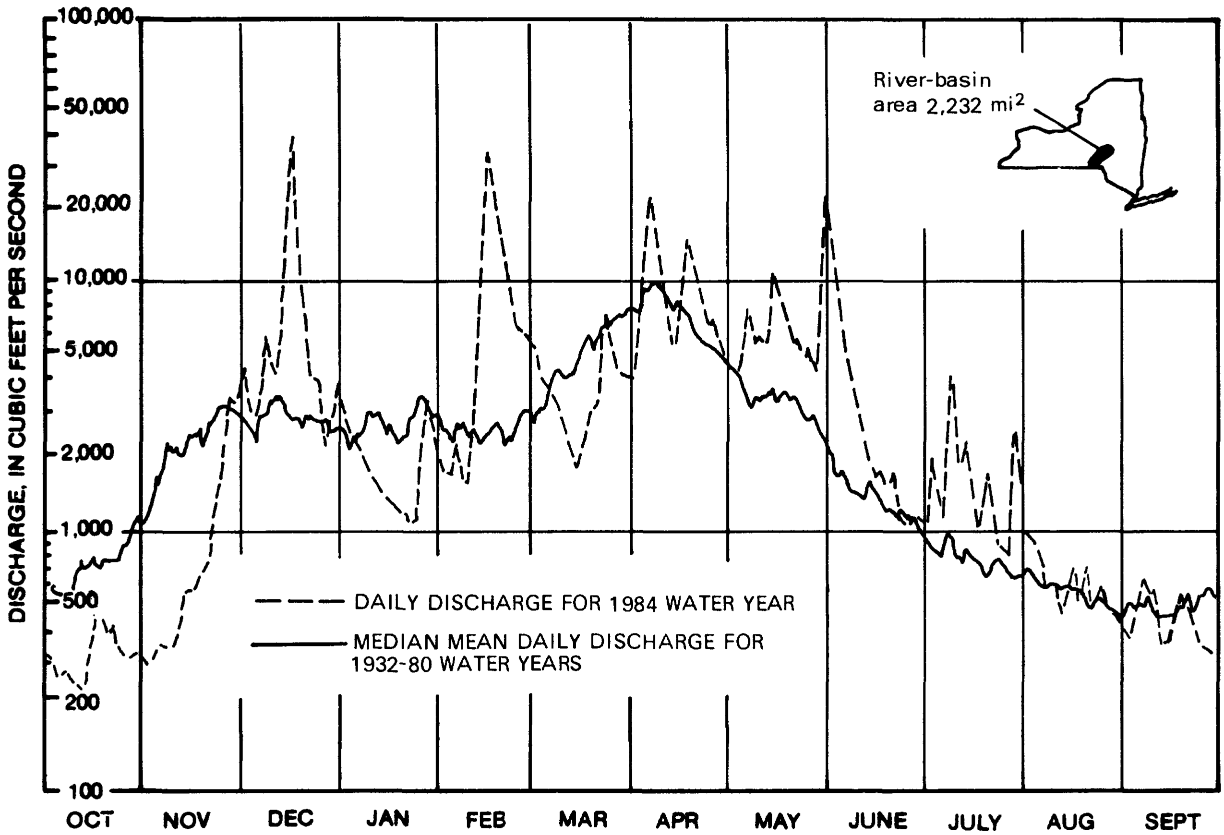


Figure 7.--Mean daily discharge of the Susquehanna River at Conklin, October 1983 through September 1984, and median mean daily discharge during 1932-80 water years.

Relationship Between Ground Water and Surface Water

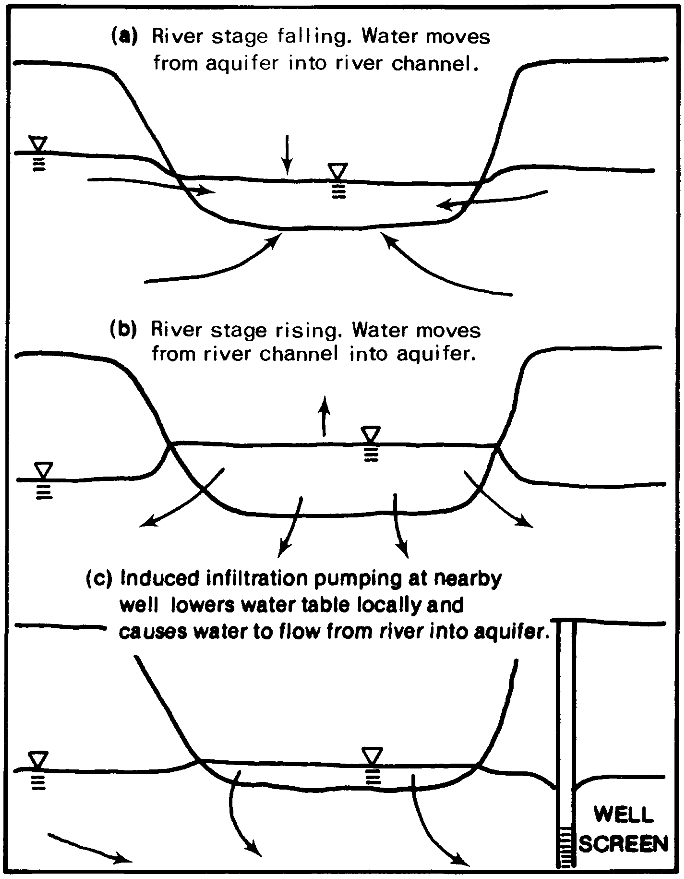
Ground water in the Kirkwood-Conklin area is in hydraulic connection with the Susquehanna River. During periods when the river stage declines below the water-table altitude (fig. 8A), ground water discharges into the river and thus increases the river flow. When the river stage is above the water table (fig. 8B) river water infiltrates through the channel bottom and banks into the aquifer. Where production wells lower the water-table altitude below river stage (fig. 8C), river water also infiltrates into the aquifer.

Riverbed Hydraulic Conductivity

The riverbed of the Susquehanna River is heavily armored with cobbles and boulders. Below the cobble armor is a layer of silt and organic material. Installing four drive-point wells in the river indicated that this layer is about 2 ft thick. The drive-point wells were placed in riffles, where the current is strongest; pools in the river would tend to collect more sediment and increase the riverbed thickness. Slug tests performed on these four wells indicate the horizontal hydraulic conductivity of the layer to range from 1 to 6 ft/d. The hydraulic conductivity of the riverbed probably varies along the channel according to sedimentation and scour patterns.

Figure 8.

Relationship between ground-water level and river stage with resulting direction of water movement.



River Stage and Ground-Water Levels

The stage of the Susquehanna River is the primary control on ground-water levels in the aquifer throughout the study area. Ground-water levels change with river stage, declining in late summer and early fall and rising in the fall and winter (fig. 9). The highest ground-water levels were recorded during peak flows. A 3-day rainfall in December 1983 caused the Susquehanna River to rise about 14 ft and inundate part of the flood plain, during which time river water infiltrated into the aquifer and caused ground-water levels to rise 6 ft. These peak flows represent a significant source of recharge to the aquifer. The rise and decline of ground-water levels during this period (fig. 9) indicates that the aquifer system responds within a few days to changes in river stage.

Ground-Water Temperature

Vertical profiles of ground-water temperatures measured at monthly intervals during the study indicate the effect of infiltrating river water on ground-water temperatures. In areas not influenced by the river, ground-water temperatures change gradually throughout the year, and the extremes do not differ by more than 10°C. The response to seasonal variations in air temperature diminishes with depth, and ground-water temperature fluctuates less than 1°C below 30 or 40 ft. Infiltration of river water can cause temperatures to change much more rapidly because heat gain and loss in ground water is transmitted through convection by the infiltrating river water.

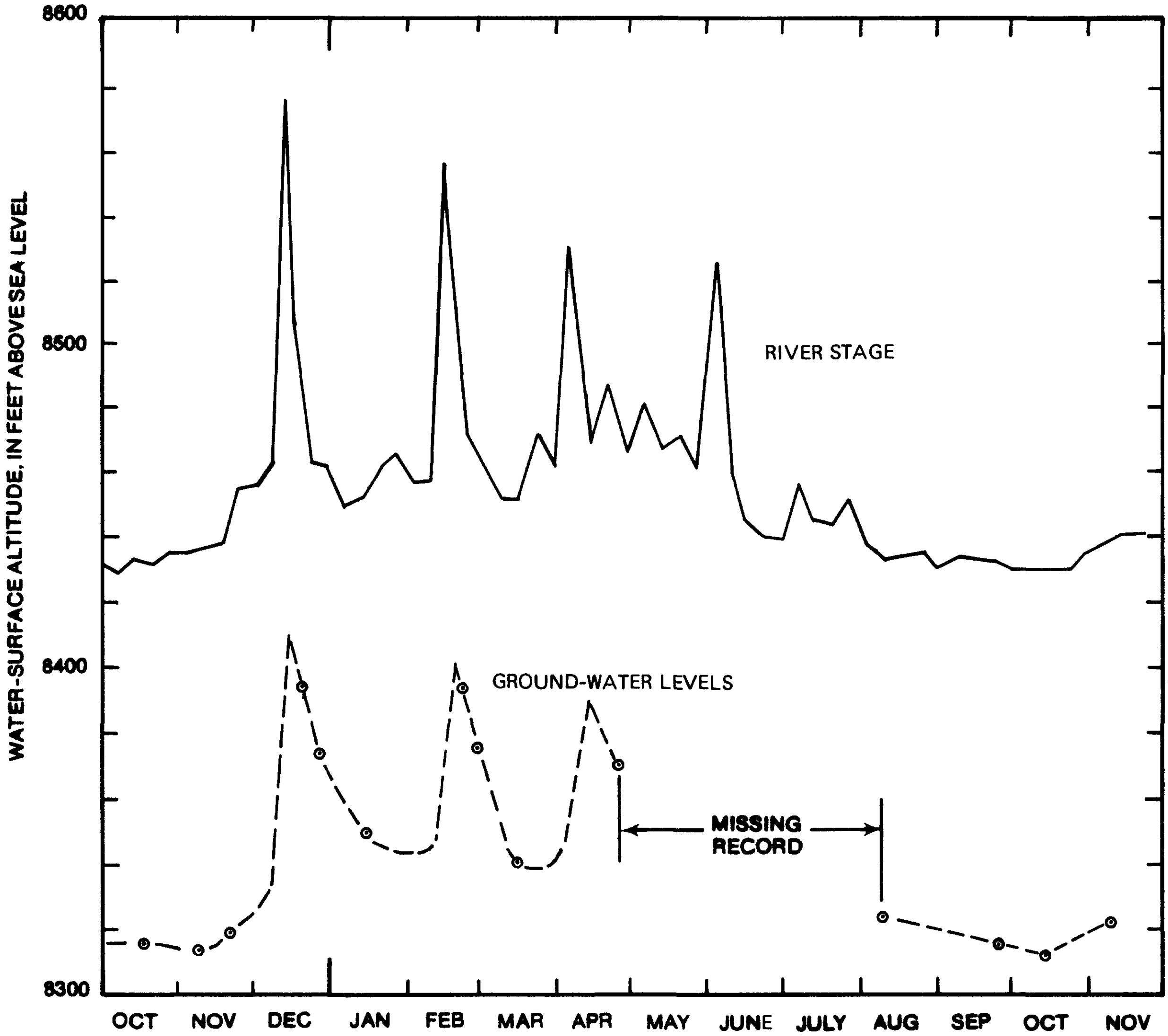


Figure 9.--Ground-water levels in observation well V 03 and stage of the Susquehanna River at Conklin, October 1983 through November 1984. (Well location is shown on pl. 1.)

The seasonal variation of temperature with depth in observation well GP1B is shown in figure 10 along with temperatures recorded in the Susquehanna River. Maximum ground-water temperatures of more than 22°C were recorded in September-October 1983. Minimum temperatures of less than 3°C were subsequently recorded in April-May 1985. This large range in temperature is caused by the infiltration of river water, which warms the aquifer during late summer and fall and cools it during winter and spring. The largest range of temperature (3°-22°C) was recorded at a depth of 15 to 20 ft below the riverbed, an area that probably lies along a flow path from the river to production well GP1.

Observation wells in which a large range in temperature was observed indicate infiltration from the river. The profiles of maximum temperature from July 1983 through March 1984 observed in well GP1B (fig. 11) show a range

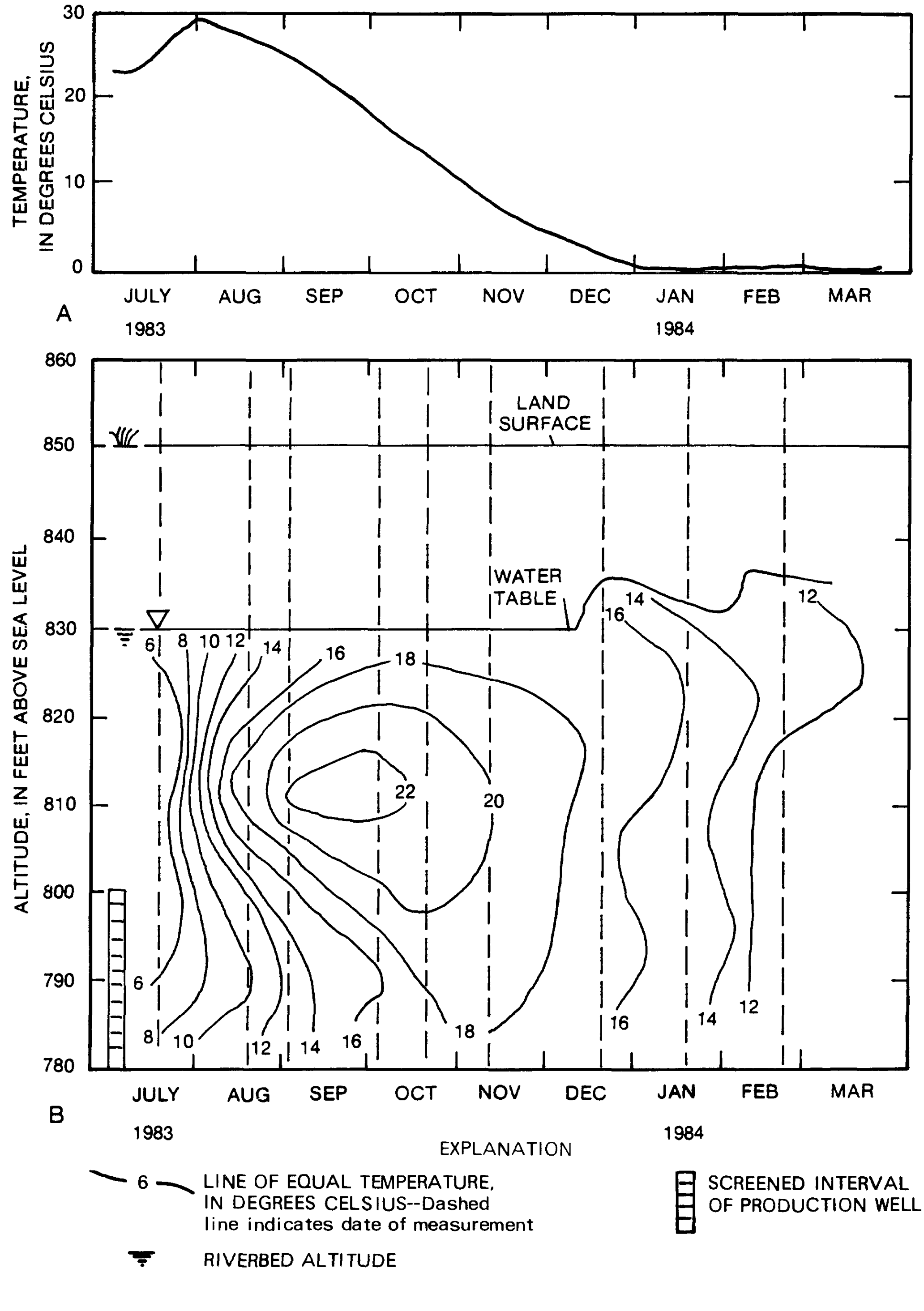


Figure 10.--Seasonal variation of water temperature July 1983 to March 1984:
A. Water temperature of Susquehanna River near well GP1B.
B. Ground-water-temperature depth profiles for observation well GP1B, near production well GP1.
(Well locations are shown in pl. 1.)

of 17°C at a depth of 20 ft. Profiles of maximum and minimum temperature in observation wells GP2B, V01, and V02, all within 55 ft of the river, are similar to those at well GP1B. In contrast, the profile for observation well V05, 260 ft upgradient from production well GP1 in the Kirkwood well field and 90 ft from the river, shows a range of only 2°C at a depth of 20 ft. Profiles recorded in other observation wells upgradient and away from the river were similar to those at well V05. Figure 12 shows the distribution of wells in which river water influences the recorded temperature profiles. The map indicates that the highest rate of river infiltration is in a small area near the production wells.

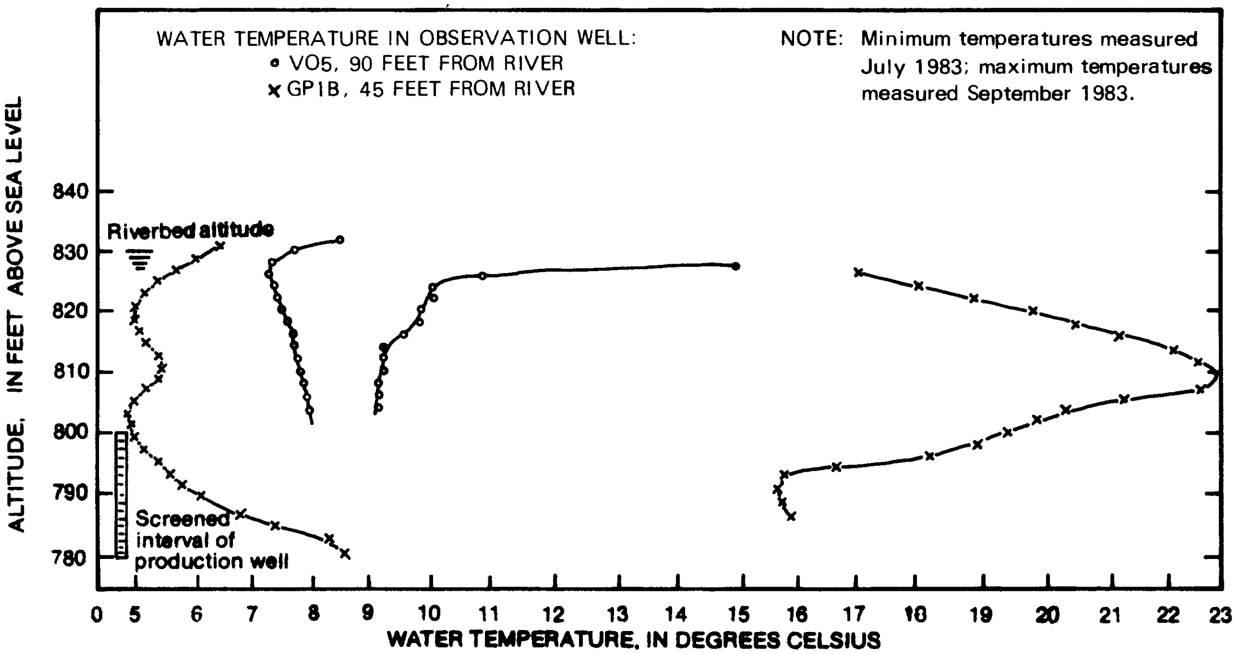


Figure 11.--Maximum and minimum temperature profiles recorded in observation wells V05, 90 ft from river, and GP1B, 30 ft from river, July 1983 to March 1984.

SIMULATION OF GROUND-WATER FLOW AND INFILTRATION FROM SUSQUEHANNA RIVER

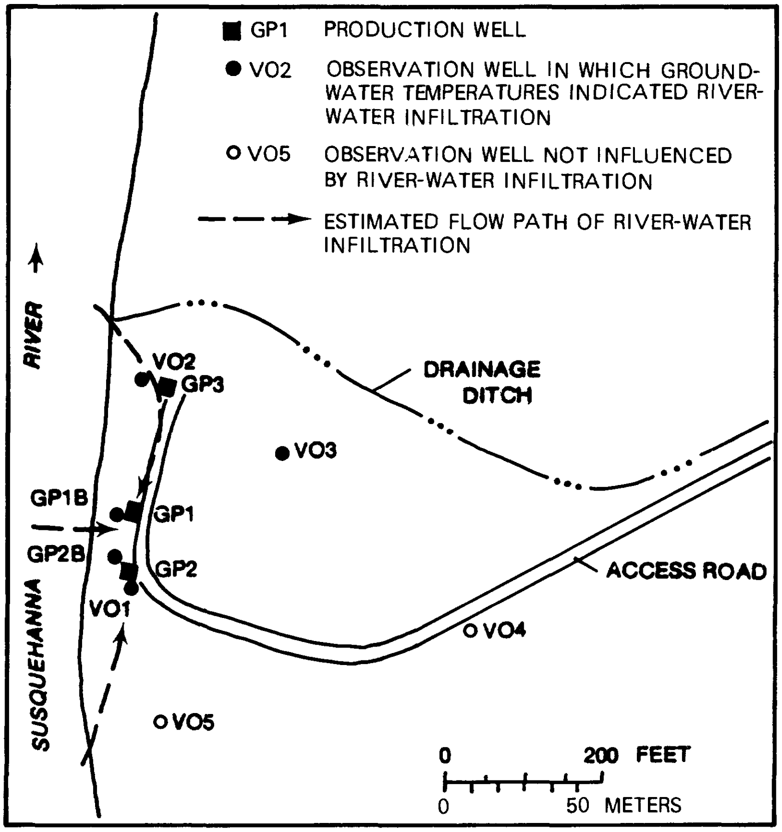
A computer model that simulates ground-water flow in three dimensions was used to quantify hydraulic properties of the riverbed and aquifer material and to estimate the quantity of river water entering the aquifer. The flow paths generated by the model were used to locate the sources of recharge within the catchment areas contributing recharge to the well fields.

MODEL DESCRIPTION

A finite-difference model developed by McDonald and Harbaugh (1984) to simulate ground-water flow in three dimensions was chosen for the study to simulate the magnitude and direction of horizontal flow and vertical hydraulic

Figure 12.

Extent of area in the Kirkwood well field in which ground-water temperatures were affected by river-water infiltration. (Locations are shown in fig. 4, p. 12.)



Base from Susquehanna River Basin Commission, 1974, 1:2400

gradients observed near the river during aquifer tests. The model solves a finite-difference approximation to the partial-differential equation governing the movement of ground water through porous material, given in McDonald and Harbaugh (1984):

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\frac{\partial}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right) - W = S_s\frac{\partial h}{\partial t} \qquad (2)$$

where:

x, y, and z are cartesian coordinates aligned along major axes of hydraulic conductivity, K_{xx}, K_{yy}, and K_{zz};
h is the hydraulic head (L);
W is a volumetric flux per unit volume representing sources and(or) discharges of ground water (t^{-1});
S_s is specific storage of the porous material (L^{-1}); and
t is time (t).

The solution of the equation yields the hydraulic head, h, equivalent to ground-water levels measured in observation wells. Values for the variables defined in equation 2 were specified in the model, together with flow and(or) head conditions at the boundaries of the aquifer system and the initial distribution of hydraulic head. The hydraulic head distribution computed by the model was used to estimate the direction and rate of ground-water flow between points within the aquifer.

MODEL DESIGN

Model Grid

The sand and gravel aquifer is represented in the model by rectangular blocks. The blocks are formed by a grid that divides the aquifer-surface area into rows and columns, and the valley-fill material into four layers. Each rectangular block is assumed to represent homogeneous material.

The grid contains 50 rows and 42 columns, and active blocks within the grid together represent a 0.68-mi^2 area. The rows, columns, and layers (pl. 2, figs. 13, 14) are spaced such that smaller blocks are near the pumping centers and larger blocks near the aquifer boundaries; this allows the model to simulate the steep hydraulic gradient near the pumping centers in detail while minimizing the total number of blocks in the model.

The vertical layers in the model were specified according to the general aquifer composition in the area and to the placement of the screens in the production and observation wells (fig. 14). Changes in saturated thickness or type of material within a layer are accounted for by the distribution of transmissivity values of the layer. Model layer 1 represents the fine-grained flood-plain deposits and the upper 10 to 15 ft of silty sand and gravel that is probably outwash material. This layer overlies lacustrine sand and silt and till where the saturated thickness is less than 10 ft (fig. 5) and in other areas overlies coarser sand and gravel that probably is ice-contact material. Layer 1 also represents the construction landfill near the Kirkwood well field.

The remainder of the sand and gravel deposit is represented in three model layers. Layer 2 represents the upper 15 ft of the ice-contact deposit, which contains lenses of silty sand and gravel and overlies lacustrine sand and silt where the saturated thickness is less than 30 ft. Layers 3 and 4 represents the deeper aquifer material where the saturated thickness exceeds 30 ft. Layer 3 was chosen to be 20 ft thick to correspond to the depths of the production well screens in the Kirkwood well field and represents the coarse layer of sand and gravel. Layer 4 represents the lower 10 ft of the silty sand and gravel overlying the lacustrine deposit.

The screens of production wells GP1, GP2, and GP3 in the Kirkwood well field and of 28 observation wells monitored during the aquifer tests corresponded to either layer 1, 2 or 3. The Conklin production well C-2 and the remaining observation wells fully penetrate the aquifer and pass through two, three, or four model layers. These screen locations are listed in tables 4 and 5 and are discussed in the section on simulation of the aquifer tests, further on.

Boundaries

Boundaries are specified for the top, bottom, and sides of the modeled area. The upper boundary includes the river and a water table with recharge from precipitation and increased recharge along valley walls from runoff (fig. 14). The contact between the aquifer and surrounding till is assumed to be the bottom, no-flow boundary. Lateral boundaries are no-flow where they represent the till-and-aquifer contact and specified-head across the valley where ground water flows to or from the simulated area (pl. 2).

Underflow.--Movement of ground water as underflow to and from parts of the aquifer beyond the model boundary is simulated as leakage through a specified head boundary (pl. 2). This allows the flow rate at the boundary to vary during simulations; thus, the effect of the boundary on the simulated cone of depression extending from the pumping centers can be compared among different simulations by the changes in flow rate and head at the boundary. The head is not specified at the specified-head boundary for blocks in layer 1 in which river leakage is simulated.

Flow across the specified head boundary is simulated by the following equation:

$$Q = C\ (H - h) \tag{3}$$

where: Q is the rate at which water enters or is discharged at a block along the boundary (L^3t^{-1});
C is the conductance of the block (L^2t^{-1});
H is the head at the boundary (L); and
h is the head in the block.

The conductance term, C, for each block along the boundary is given by

$$C = \frac{KA}{L} \tag{4}$$

where: K is the horizontal hydraulic conductivity of the block (Lt^{-1}); (the term "hydraulic conductivity" used hereafter in the report refers to the horizontal hydraulic conductivity)
A is the cross-sectional flow area (L^2); and
L is the flow length (L).

The head along the specified head boundary was extrapolated from the hydraulic gradient measured near the boundary. Hydraulic conductivity at the specified-head boundary was estimated from analysis of drawdown data collected during the aquifer tests and adjusted during model calibration.

River.--Infiltration of river water into the aquifer and discharge of ground water to the river were simulated as leakage through a semiconfining layer representing the riverbed. Flow through the riverbed was calculated from equation 3 with the conductance, C, defined as the vertical hydraulic conductivity of the riverbed, K_R, and the flow length, L, defined as the thickness of the riverbed.

The riverbed thickness was assumed to be 2 ft, and the horizontal hydraulic conductivity of the riverbed was estimated as 3.5 ft/d from results of the piezometer tests described in the section on hydrologic properties. The vertical hydraulic conductivity of the riverbed was at first assumed equal to the horizontal hydraulic conductivity but was later decreased during model calibration. Riverbed altitudes were taken from channel cross sections (pl. 1) measured by the U.S. Army Corps of Engineers (1982, pl. A-2). River stages were determined from these cross sections and adjusted in accordance with the measured stage at staff gage R2 (pl. 1).

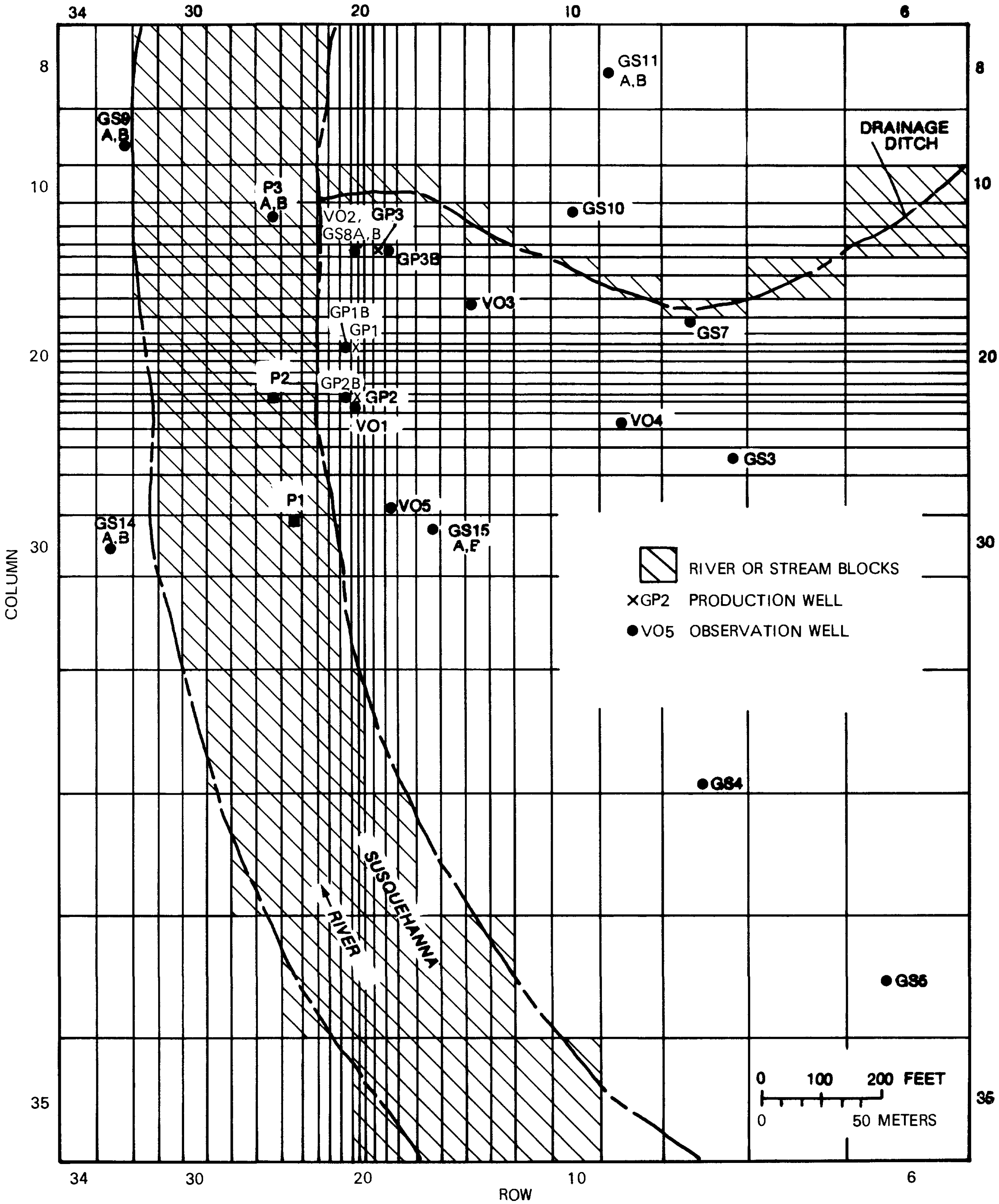


Figure 13.--Detail of model grid near Kirkwood well field. (Location is shown in pl. 2.)

Tributary streams.--Ground-water discharge to tributary streams was simulated in a manner similar to that used for ground-water discharge to and from the river. The hydraulic conductivity of the streambed was assumed equal to that of the riverbed, and the thickness of the streambed was assumed to be 1 ft. Streambed altitudes were measured or scaled from topographic maps.

Recharge.--Recharge from precipitation was assumed to enter all blocks in layer 1 that do not represent the Susquehanna River. Recharge through infiltration from tributary streams that drain upland areas in the Town of Conklin was represented by increasing the recharge rate in blocks near the boundary where the stream channels enter the valley floor and along the valley wall, as shown in plate 2. The volume of recharge in these areas was calculated by the method of MacNish and Randall (1982, p. 37). Drainage from upland areas in the Town of Kirkwood enters the aquifer as ground-water flow through the construction-material landfill. This flow was represented by a specified-head boundary upgradient of the landfill. (See pl. 2.)

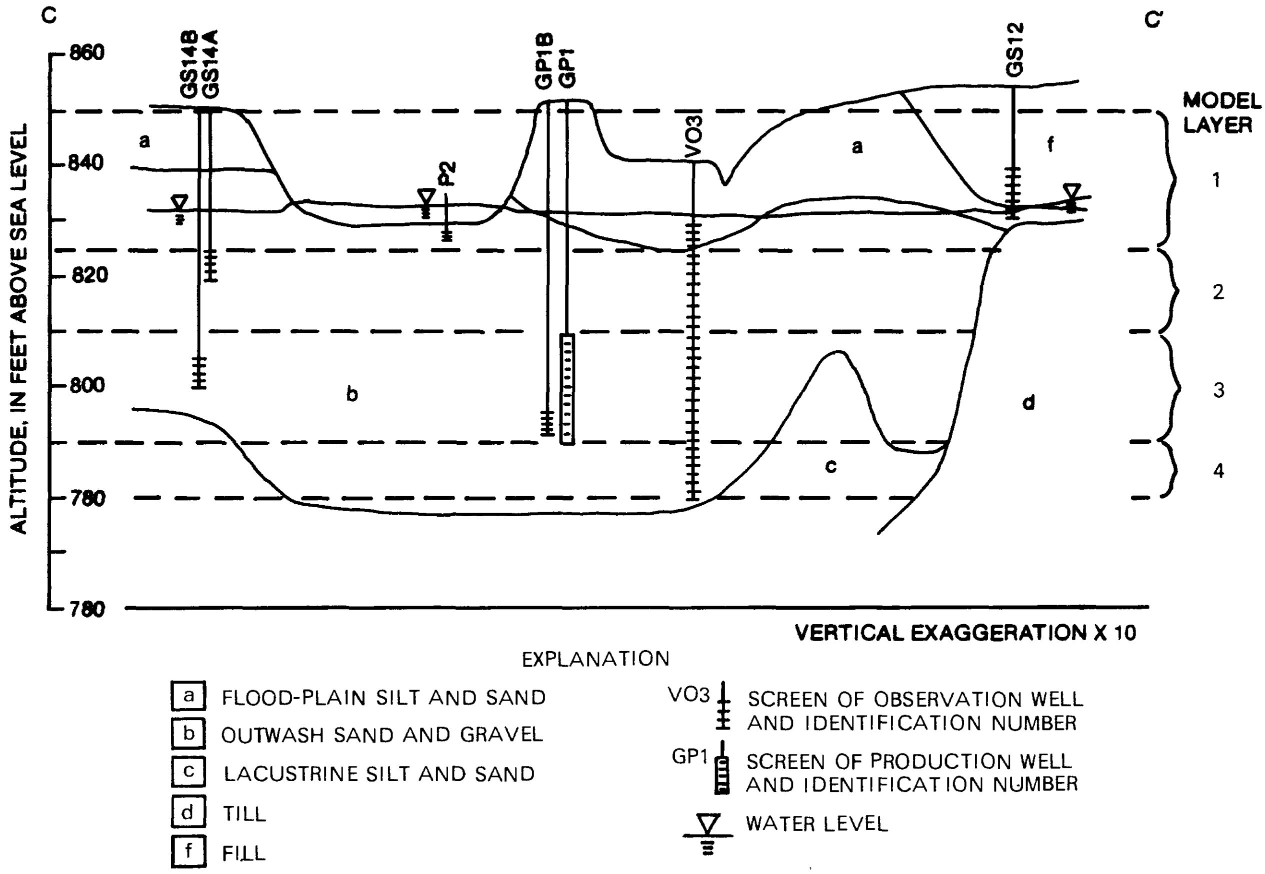


Figure 14.--Schematic vertical profile of aquifer along cross-section C-C' showing layers and boundaries used in model. (Location is shown in pl. 1.)

Hydraulic Properties

Blocks representing sand and gravel or sand and silt deposits were assigned initial horizontal and vertical hydraulic conductivity and storage values. Hydraulic conductivity was assumed equal along both horizontal axes; that is, each block was assumed to be isotropic.

Hydraulic conductivity.--The hydraulic conductivity of the sand and gravel aquifer was initially estimated to range from 200 to 2,000 ft/d from an analysis of drawdown data from aquifer tests, and hydraulic conductivity of the sand and silt deposits bordering the aquifer was assumed to be 1 ft/d from results of slug and bail tests. Results of the aquifer-test analysis and the slug and bail tests are described in the appendix. Hydraulic conductivity of the landfill material was initially assumed to be 10 ft/d, equivalent to that measured for silty sand and gravel (Freeze and Cherry, 1979, p. 29). These values are summarized in table 2.

Hydraulic-conductivity values of the sand and gravel and the landfill were adjusted during model calibration to match drawdowns measured during the aquifer tests; the value for the sand and silt was left unchanged because these areas have little effect on the ground-water flow through the sand and gravel.

The final distribution of hydraulic conductivity used in the model reflected the heterogeneity of the aquifer materials inferred from the aquifer-test analysis. The highest value of hydraulic conductivity used was 10,000 ft/d, which was in layer 3 and corresponded to the most productive layer of sand and gravel tapped by the Kirkwood production wells. Hydraulic conductivity in layers 2 and 4 was 2,000 ft/d near the Kirkwood well field and 330 ft/d near the Conklin well field; these values are close to the values obtained from the aquifer-test analysis. Hydraulic conductivity of layer 1, which represented the outwash, ranged from 50 to 500 ft/d.

The range of calibrated values for the sand and gravel of 50 to 10,000 ft/d in this model (table 2) is within the range of values reported by Lyford and others (1984, p. 2) for glacial aquifers. However, the upper value of the range is high, so values ranging from 1,000 to 10,000 ft/d were tested in a sensitivity analysis to verify the high value. Results of these simulations are discussed in the section on simulation of aquifer tests.

Table 2.--Hydraulic-conductivity values used in calibrated model.

[Values are in feet per day.]

Geologic unit	Initial range	Calibrated range	Values found in other glacial aquifers (from Lyford and others, 1984)
Outwash sand and gravel	1,000-2,000	50-10,000	1-13,300
Lacustrine sand and silt	1	1	10^{-4} - 1
Landfill debris	10	50	--

The transmissivity distribution in each model layer is calculated by multiplying the hydraulic conductivity of each model block by the saturated thickness of the layer at each block. The layer transmissivities were then summed to obtain the total transmissivity of the aquifer. The distribution of the total transmissivity is shown in figure 15.

Vertical hydraulic conductivity.--Vertical movement of ground water was simulated in the model as leakage between adjacent layers. Vertical leakage between blocks was calculated from:

$$Q_v = C \Delta h \tag{5}$$

where: Q_v is vertical flow (L^3t^{-1}),
Δh is the difference in head between the upper and lower blocks (L), and
C is the hydraulic conductance defined as:

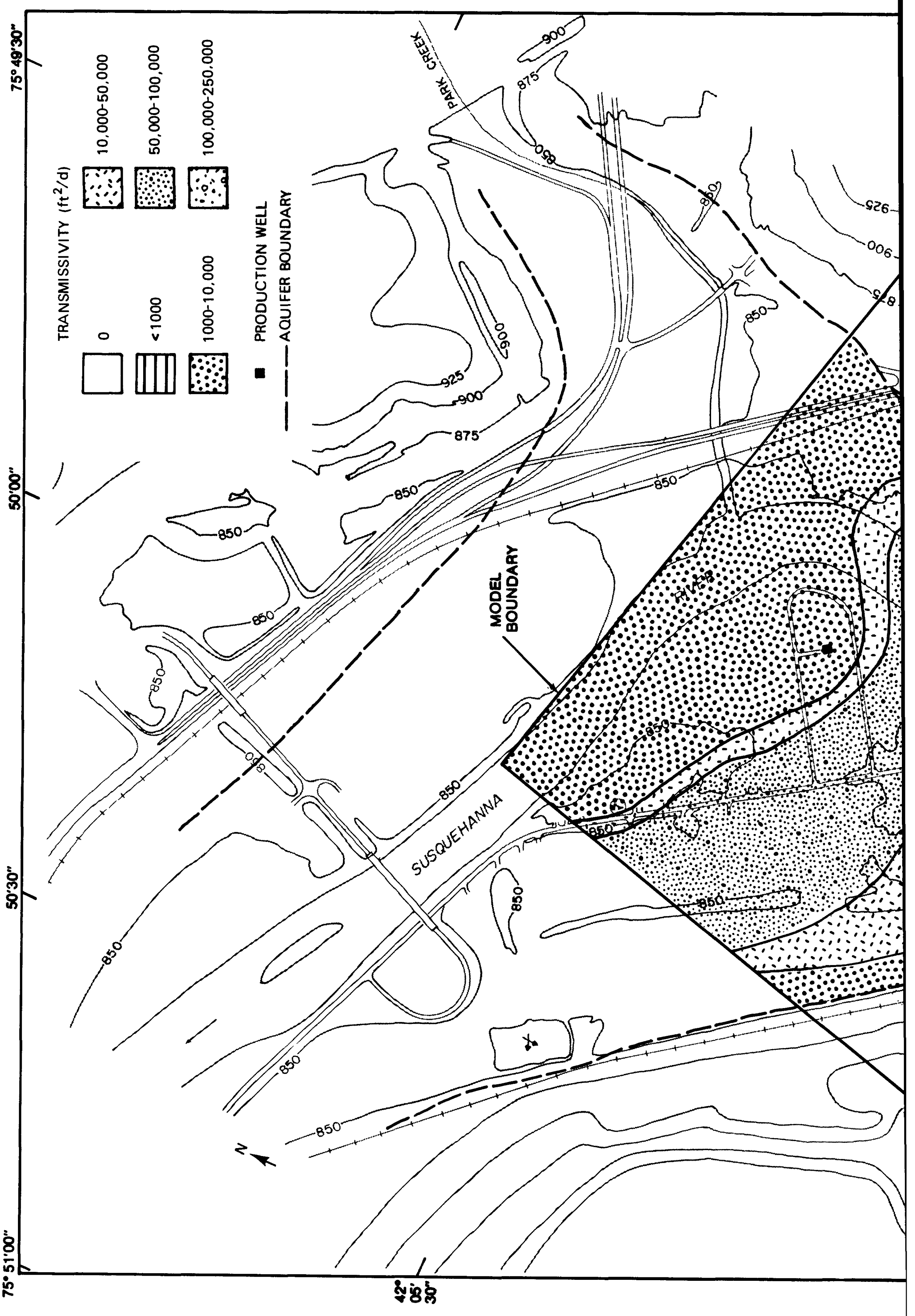
$$C = \frac{K_v A}{L}$$

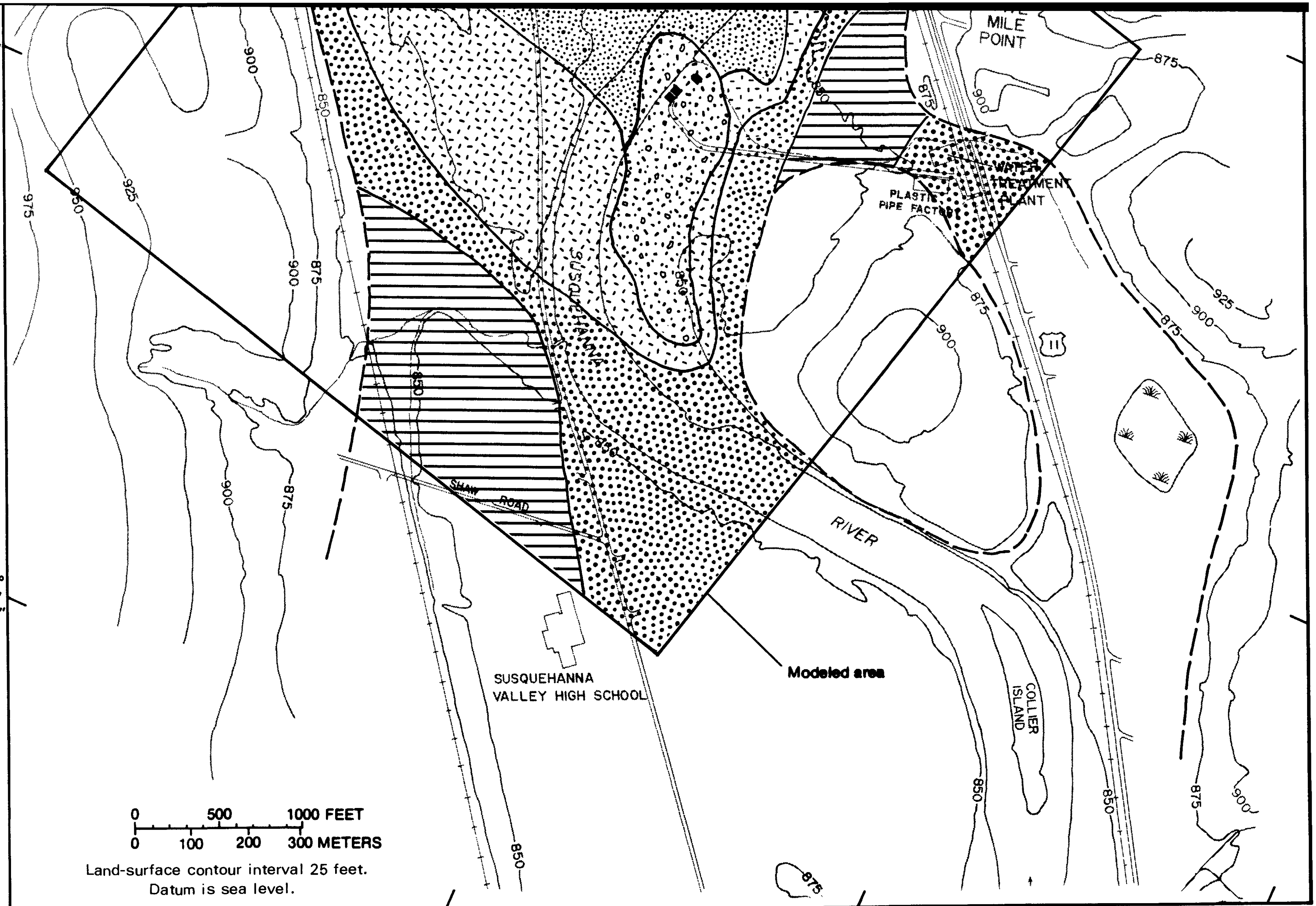
where: K_v is the vertical hydraulic conductivity (LT^{-1}),
A is the area of the block (L^2), and
L is the distance between block centers of adjacent layers.

The horizontal hydraulic conductivity of the sand and gravel aquifer is much greater than the vertical hydraulic conductivity. This anisotropy results from lenses of silty sand and gravel that restrict the vertical movement of ground water without significantly decreasing the horizontal transmissivity of the material. For preliminary model runs, an initial estimate for the ratio of horizontal to vertical hydraulic conductivity of 55:1 was assumed from an analysis of drawdowns recorded during aquifer tests. Through subsequent model calibration, this value was increased to between 125:1 and 250:1 to match drawdowns observed during the aquifer tests.

Storage coefficient and specific yield.--The volume of ground water released when water levels are lowered depends on whether the aquifer is confined or unconfined. Release of water from confined material is caused by the elastic response of the aquifer material as the pressure upon it is decreased. The volume of water released is small and is described by the storage coefficient, which for confined sand and gravel aquifers is typically between 10^{-3} and 10^{-4} (Lyford and others, 1984, p. 12). In contrast, water released from unconfined materials results from drainage of pore spaces. The volume of water released is much larger than that released by confined materials and is described by the specific yield, which for sand and gravel aquifers ranges from 0.05 to 0.35 (Lyford and others, 1984, p. 12).

All layers in the model except layer 1 are confined by the overlying model layers, and a storage coefficient of 10^{-3} was specified for each on the basis of results of the aquifer test analysis. Layer 1 represents the upper part of the sand and gravel, which is unconfined except where it is overlain by riverbed deposits. (See fig. 14.) A specific yield of 0.25 was specified for the unconfined areas in layer 1, and a storage coefficient of 10^{-3} was used beneath the river.


TRANSMISSIVITY (ft²/d)
0
<1000
1000-10,000
10,000-50,000
50,000-100,000
100,000-250,000
PRODUCTION WELL
AQUIFER BOUNDARY
MODEL BOUNDARY
PARK CREEK
SUSQUEHANNA
RIVER
75°49'30"
50'00"
50'30"
75°51'00"
42°05'30"



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 15.--Aquifer transmissivity values used in model simulations.

MODEL CALIBRATION

The model was calibrated by simulating (1) a short-term equilibrium condition in October 1984 when production wells were off and river stages were stable, and (2) transient conditions during an aquifer test performed in October 1984 in the Kirkwood well field. Water levels recorded in observation wells were compared with head and drawdown distributions obtained in the steady-state and transient-state simulations to assess the model's ability to accurately represent the ground-water system.

Calibration Procedure

The hydraulic properties of the aquifer and boundary conditions that were adjusted during model calibration are listed in table 3. Each of these elements was adjusted through trial and error. Computed heads and drawdowns were then compared with those measured in observation wells, and if the adjustments improved the computed values, they were retained. During calibration, a range of values for hydraulic properties was tested to investigate the sensitivity of model results to variations in these values.

Improvements in model results were determined by comparing residuals or differences between the computed and observed hydraulic heads or drawdowns. Improvements were also measured by computing the root-mean-square (RMS) difference, defined as:

$$RMS = \left[\frac{\Sigma(AE)^2}{n}\right]^{1/2} \tag{8}$$

where: AE is the |(observed value - predicted value)| for each observation well, and
n is the number of observation wells.

The difference between observed and computed drawdowns, AE, was divided by the observed drawdown to express the RMS difference as a percentage of the observed drawdown. To calculate the RMS difference for observation wells that were screened in more than one model layer, the composite hydraulic head or drawdown in the well was found by weighting the value from each model layer by the layer thickness.

Table 3.--Elements adjusted during model calibration.

Boundary conditions	Hydraulic properties
Specified-head boundary	Hydraulic conductivity
River stage	Vertical anisotropy
Recharge	Vertical hydraulic conductivity of riverbed
	Specific yield
	Storage coefficient

Steady-State Simulation

Hydraulic heads computed by steady-state simulations of nonpumping conditions were compared with ground-water levels measured in 35 observation wells in October 1984 during a period when the production wells had not been in operation for 8 hours. At this time heads in the aquifer had nearly recovered to prepumping levels. The use of steady-state simulations assumed that the system was at a short-term equilibrium and that water levels in the aquifer were not changing. Because the aquifer responds rapidly to changes in river stage, usually within a few days, ground-water levels fluctuate throughout the year. However, river stage was relatively constant during the late summer and early fall of 1984. (See fig. 9.) Conditions in the aquifer during this period of the year best approximated a short-term equilibrium.

Differences between the observed heads and those computed by the calibrated model are listed in table 4. The RMS difference was 0.33 ft, with a maximum difference of -1.2 ft between computed and observed heads at observation well GS17; computed heads were within 0.5 ft of those observed in all but seven wells. The greatest differences between computed and measured heads are near model boundaries and near the contact between the landfill and the sand and gravel aquifer. Simulated water-table contours for the nonpumping condition in October 1984, equivalent to the hydraulic head distribution for layer 1, are shown in figure 17 (p. 40).

The average recharge rate of 22 in/yr reported by Randall (1977) was reduced to 9 in/yr during steady-state simulations to match water levels in the aquifer that were at a minimum in October 1984. Recharge did not occur during the aquifer test and was not included in transient-state simulations discussed below.

Transient-State Simulation

Transient-state simulations were run to duplicate the aquifer test in Kirkwood in October 1984. The head distribution used as the initial conditions for transient-state simulations of the aquifer test was that provided by the steady-state simulation described above. No precipitation was recorded at the Binghamton airport, 7 miles north of the study area, in the 10 days preceding the aquifer test, and river stage varied less than 0.2 ft during the same period (fig. 16). Water levels in observation wells unaffected by pumping from the production well varied less than 0.1 ft during the 24-hr test. The short duration of the test and the relatively small fluctuations in water

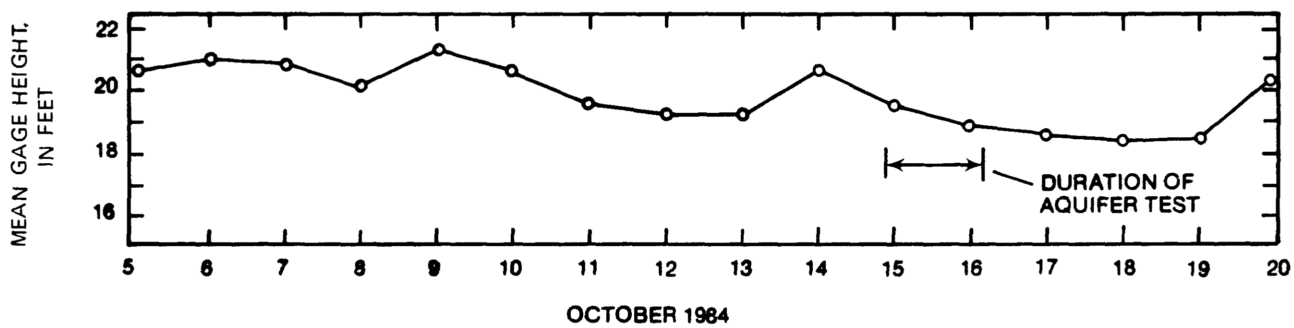


Figure 16.--Mean daily gage height of the Susquehanna River at Conklin, October 1984.

levels justify the assumption that the aquifer system was at a short-term equilibrium before and during the aquifer test.

Simulations represented two phases--a 23-hour period when wells GP1 and GP2 were both pumped at 1,000 gal/min, and the succeeding 5-hour period when well GP1 was pumped at 1,000 gal/min and GP2 was idle. Drawdowns computed by transient-state simulations of the aquifer test were compared with drawdowns measured in 31 observation wells (table 5). The RMS difference in computed drawdowns was 16 percent for phase 1 and 24 percent for phase 2. Maximum differences between computed and observed drawdowns in phase 1 were -2.78 ft at

Table 4.--Ground-water levels during nonpumping conditions in October 1984 as computed by steady-state simulation.

[Altitudes are in feet above sea level. Differences are in feet. Well locations are shown in pl. 1.]

Observation well	Model layer	Water level Observed	Water level Computed[1]	Percent
GS3	2-3	832.70	832.92	0.22
GS4	3	832.70	832.95	.25
GS5	2	832.90	833.23	.33
GS6	1	834.80	834.71	-.09
GS7	2	833.00	832.93	-.07
GS9A	2	832.70	832.73	.03
GS9B	3	832.70	832.73	.03
GS10	2	832.70	832.90	.20
GS11A	2	832.70	832.96	.26
GS11B	3	832.70	832.89	.19
GS12	1	833.80	834.91	1.11
GS13	1	850.65	849.96	-.69
GS14A	2	832.90	832.78	-.12
GS14B	3	832.80	832.77	-.03
GS15A	2	832.80	832.89	.09
GS15B	3	832.60	832.89	.29
GS16A	2	831.90	832.86	.96
GS16B	3	832.00	832.84	.84
GS17	1	839.40	840.58	1.18
GS18	1	838.50	837.66	-.84
GS19	1	837 10	836.81	-.29
V01	2-4	832.50	832.86	.36
V03	1-4	832.70	832.87	.17
V04	2-4	832.70	832.91	.21
V05	1-3	832.70	832.89	.19
MW1	3	832.00	832.46	.46
MW2	2-3	832.00	832.48	.48
MW3	3-4	832.10	832.52	.42
MW4	1	831.80	831.77	-.03
MW5	2-3	832.80	831.21	-.59
MW6	2-3	832.40	832.15	-.25
P1	1	832.90	832.84	-.06
P2	1	832.80	832.83	.03
P3A	1	832.70	832.81	.11
P3B	3	832.70	832.81	.11
			RMS difference = 0.33 ft	

[1]at nearest model node

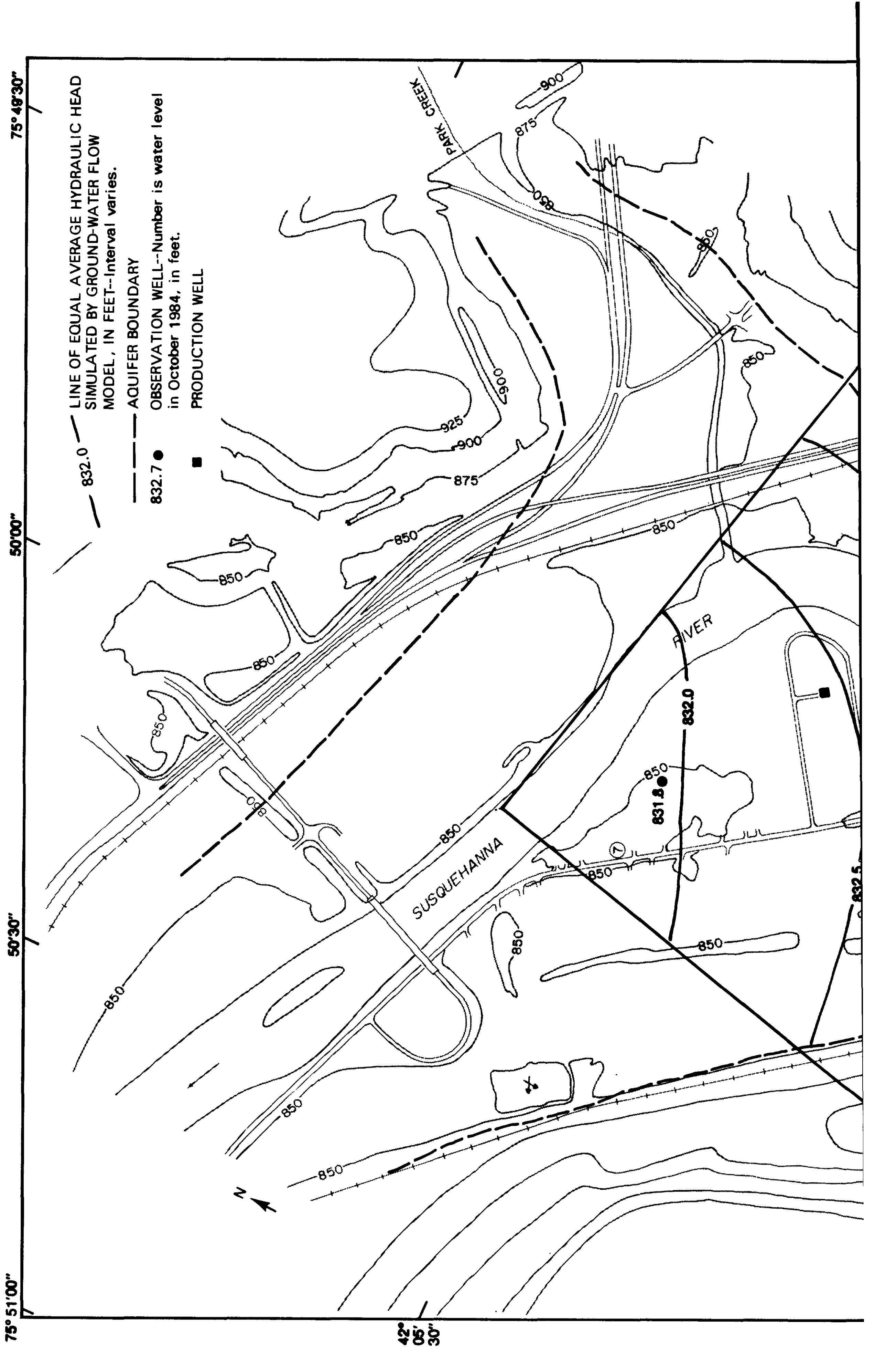
well GP2B and in phase 2 were +0.3 ft at wells GP1B, GP2B, P3A, and P3B. Computed draw-downs were within 0.3 ft for all but six wells in the first phase and four wells in the second.

Computed distributions corresponded closely to actual drawdowns measured in observation wells. The distribution of computed drawdown in layers 2 and 3 at the end of the first phase are shown in figures 18A and 18B, respectively; vertical profiles of the computed drawdown distribution along cross sections E-E' and F-F' are presented in figure 19. Measured drawdowns of at least 1.0 ft extended more than 400 ft laterally beneath the Susquehanna River to the west of the well field and more than 800 ft toward the gravel excavation site to the south. The smaller drawdowns across the river indicate that the river acts as a leaky recharge boundary that supplies water to the well field.

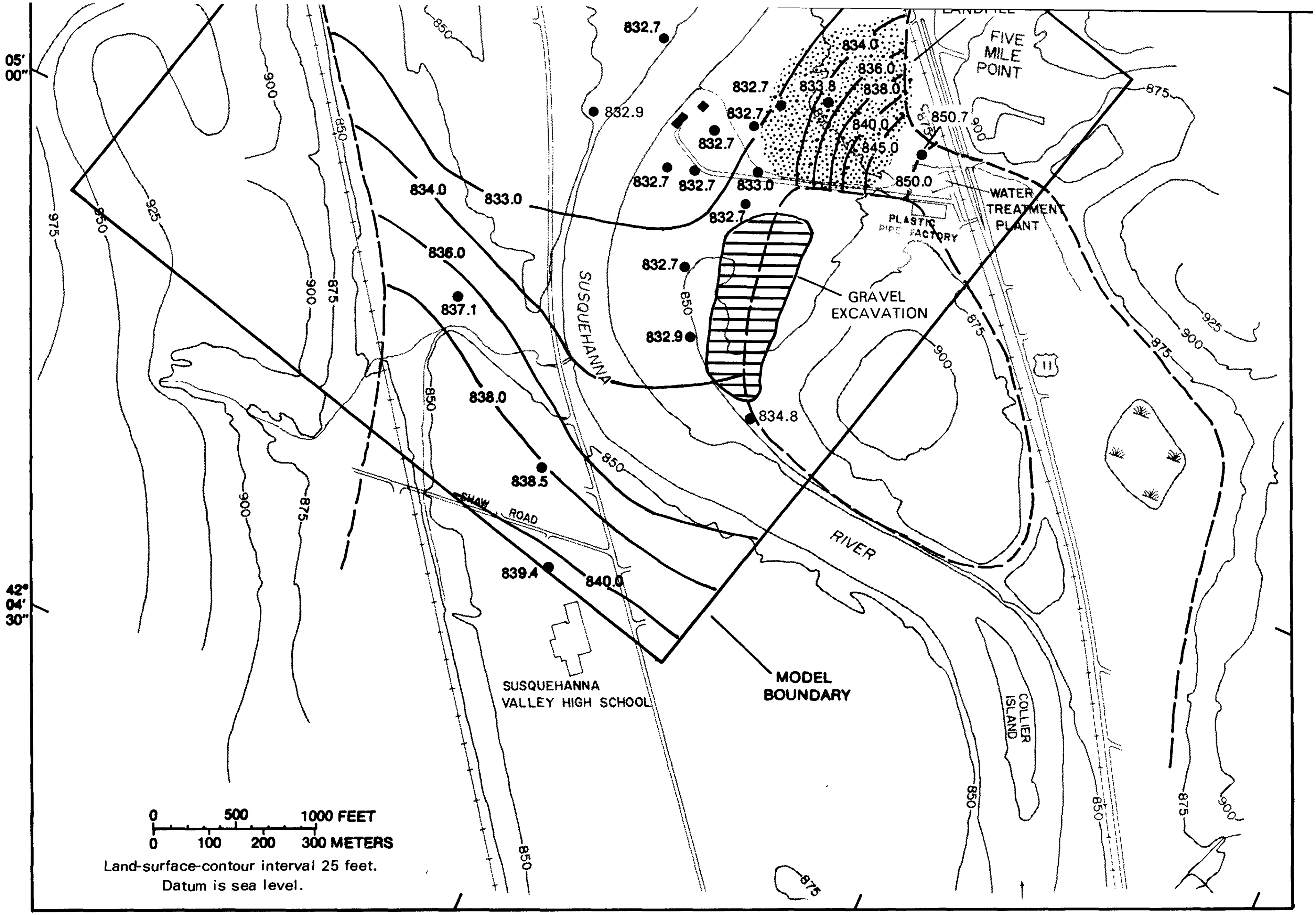
Table 5.--Drawdowns computed by transient-state simulations of aquifer tests in October 1984 at Kirkwood well field.

[All values are in feet. Locations are shown in pl. 1.]

		Wells GP1 and GP2 pumped at 1,000 gal/min for 23 hours			Well GP1 pumped at 1,000 gal/min for 5 hours		
Observation well	Model layer	Observed	Computed	Percent difference	Observed	Computed	Percent difference
GS3	2-3	1.37	1.39	1.4	0.96	0.91	-5.2
GS4	3	1.42	1.37	-3.5	.79	.88	-11.4
GS5	2	.74	.63	-14.9	.67	.48	-28.4
GS6	1	.00	0.00	0.0	.00	.00	0.0
GS7	2	1.18	1.12	-5.1	.84	.80	-4.8
GS9A	2	.85	.89	4.7	.64	.66	3.1
GS9B	3	1.23	1.11	-9.8	.73	.76	4.1
GS10	2	1.56	1.15	-26.3	.98	.84	-14.3
GS11A	2	1.20	.90	-25.0	.84	.67	-20.2
GS11B	3	1.08	1.15	6.5	.73	.81	11.0
GS12	1	.00	.01	0.0	.00	.02	0.0
GS13	1	.00	.00	0.0	.00	.00	0.0
GS14A	2	1.16	.80	-31.0	.83	.59	-28.9
GS14B	3	1.16	.99	-14.7	.77	.67	-13.0
GS15A	2	1.07	1.33	24.3	.96	.91	-5.2
GS15B	3	1.63	1.86	14.1	.97	1.11	14.4
GS16A	2	1.49	1.20	-19.5	.90	.87	-3.3
GS16B	3	1.53	1.59	3.9	.76	1.05	38.2
GS17	1	.00	.00	0.0	.00	.00	0.0
GS18	1	.00	.00	0.0	.00	.00	0.0
GS19	1	.00	.00	0.0	.00	.00	0.0
VO1	2-4	2.57	2.53	-1.6	.99	1.17	18.2
VO3	1-4	1.76	1.64	-7.8	1.04	1.09	4.8
VO4	2-4	1.53	1.54	0.1	.95	.99	4.2
VO5	1-3	1.70	1.39	-18.2	1.06	.93	-12.3
GP1B	3	2.34	2.65	13.2	1.40	1.74	24.3
GP2B	3	6.73	3.95	-41.3	1.08	1.38	27.8
P1B	1	1.30	1.06	-18.5	.76	.79	3.9
P2	1	1.26	1.02	-19.0	.75	.75	0.0
P3A	1	1.06	1.01	-4.7	.40	.74	85.0
P3B	3	1.30	1.62	24.6	.71	1.06	49.3
	RMS difference			17%			24%

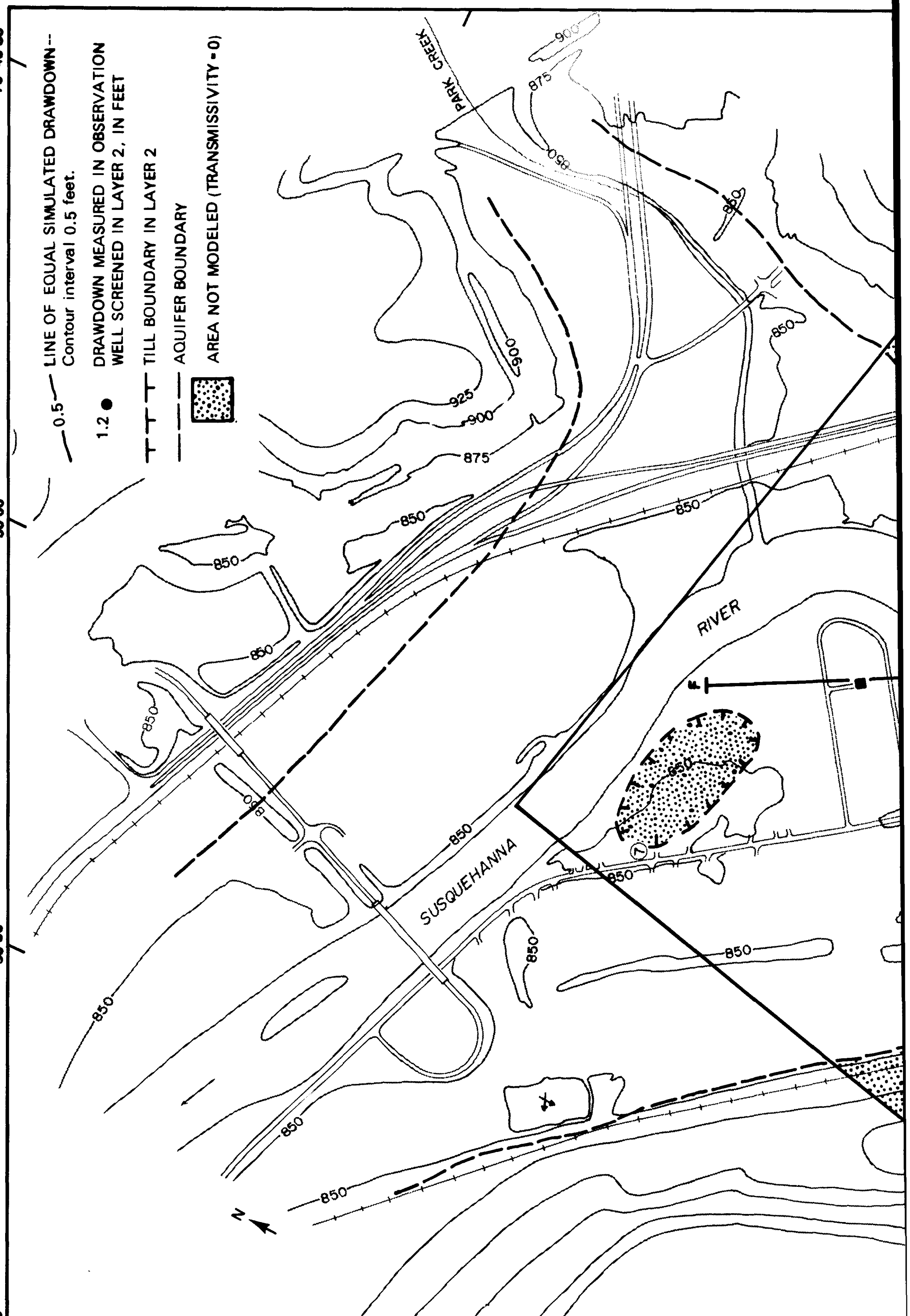


75°49'30"
50'00"
50'30"
75°51'00"
42°05'30"
832.0 LINE OF EQUAL AVERAGE HYDRAULIC HEAD SIMULATED BY GROUND-WATER FLOW MODEL, IN FEET--Interval varies.
AQUIFER BOUNDARY
832.7 ● OBSERVATION WELL--Number is water level in October 1984, in feet.
■ PRODUCTION WELL
PARK CREEK
SUSQUEHANNA
RIVER
N
832.0
831.8
832.5
850
875
900
925

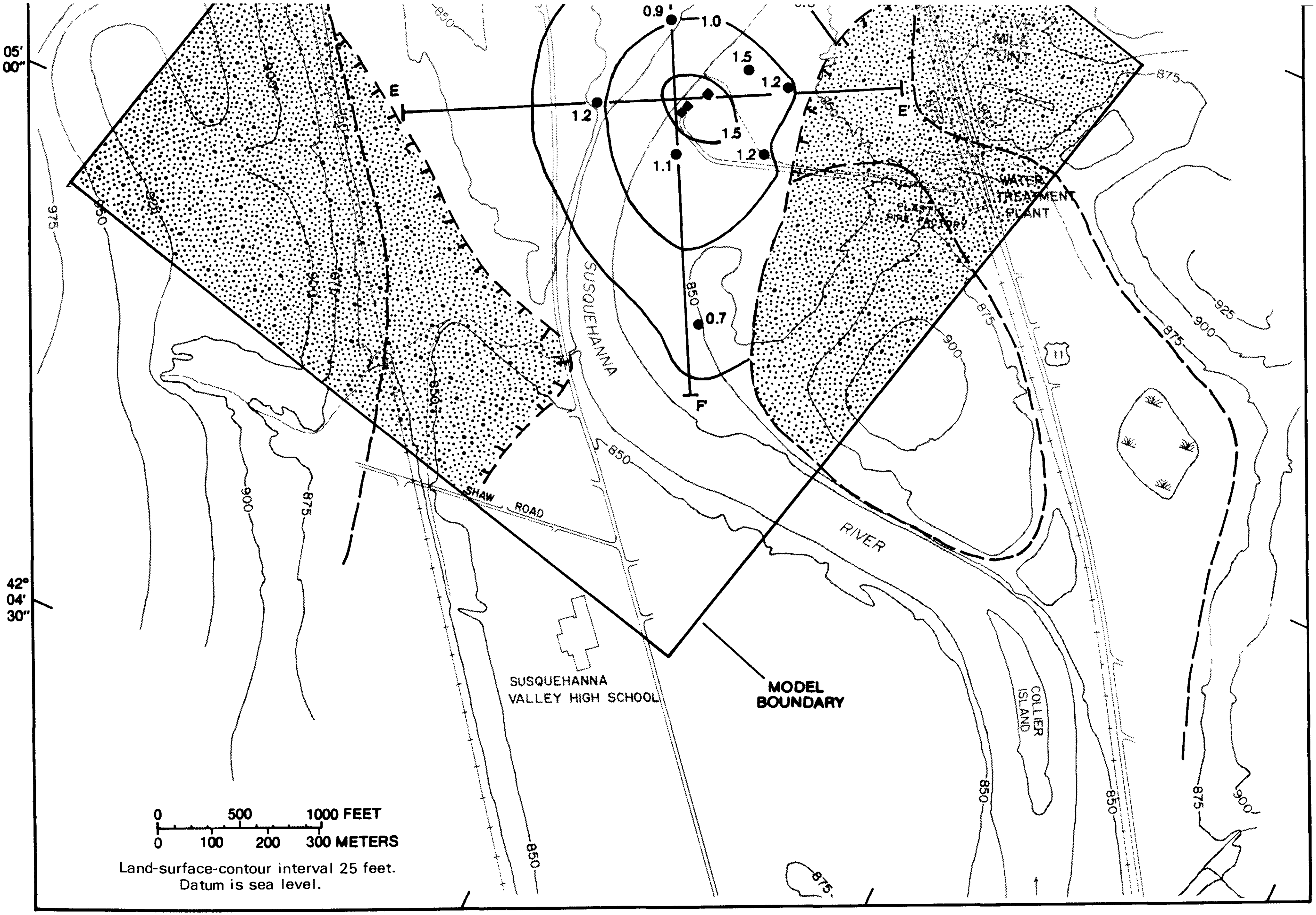


Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 17.--Water-table altitudes predicted by steady-state simulations of October 1984 after cessation of pumping.

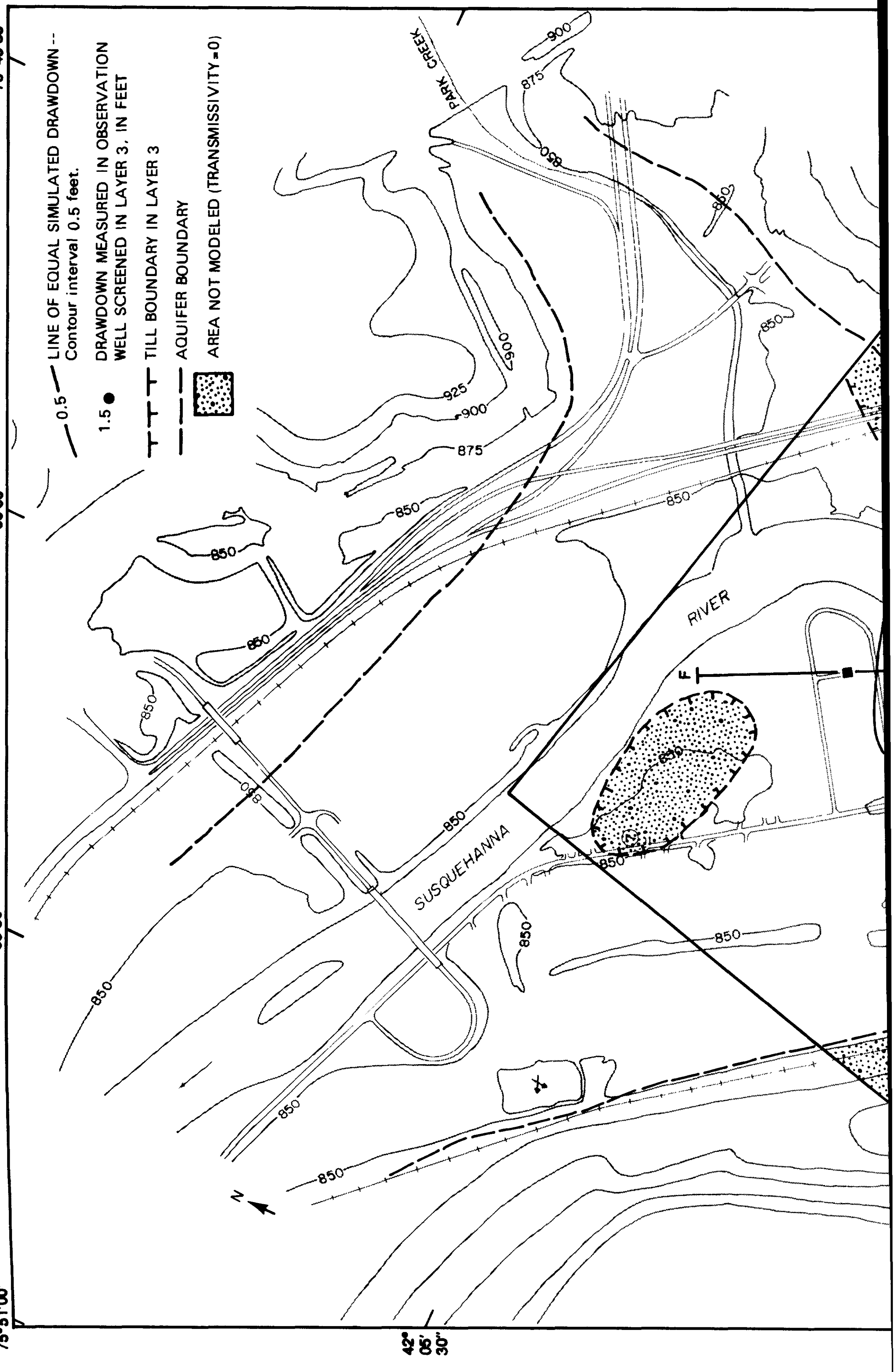


0.5 LINE OF EQUAL SIMULATED DRAWDOWN-- Contour interval 0.5 feet.
1.2 ● DRAWDOWN MEASURED IN OBSERVATION WELL SCREENED IN LAYER 2, IN FEET
TILL BOUNDARY IN LAYER 2
AQUIFER BOUNDARY
AREA NOT MODELED (TRANSMISSIVITY = 0)
PARK CREEK
SUSQUEHANNA
RIVER
900
875
850
925
75°49'30"
50'00"
50'30"
75°51'00"
42° 05' 30"
N

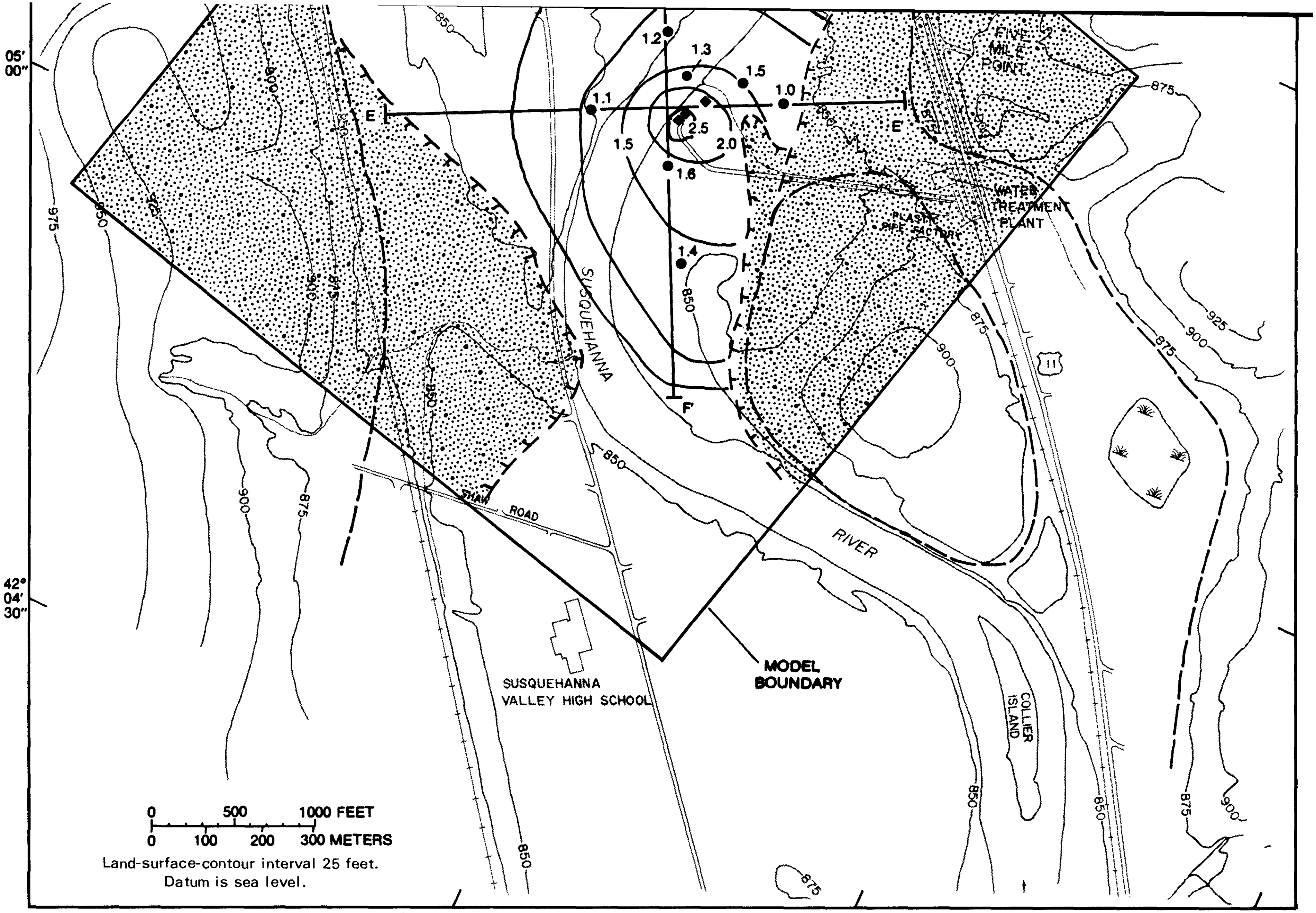


Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 18A.--Distribution of drawdowns in model layer 2 computed by transient-state simulation of October 1984 aquifer test at Kirkwood.



0.5 LINE OF EQUAL SIMULATED DRAWDOWN-- Contour interval 0.5 feet.
1.5 DRAWDOWN MEASURED IN OBSERVATION WELL SCREENED IN LAYER 3, IN FEET
TILL BOUNDARY IN LAYER 3
AQUIFER BOUNDARY
AREA NOT MODELED (TRANSMISSIVITY=0)
75°49'30"
50'00"
50'30"
75°51'00"
42°05'30"
PARK CREEK
SUSQUEHANNA
RIVER
F
N
900
875
850
925



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 18B.--Distribution of drawdowns in model layer 3 computed by transient-state simulation of October 1984 aquifer test at Kirkwood.

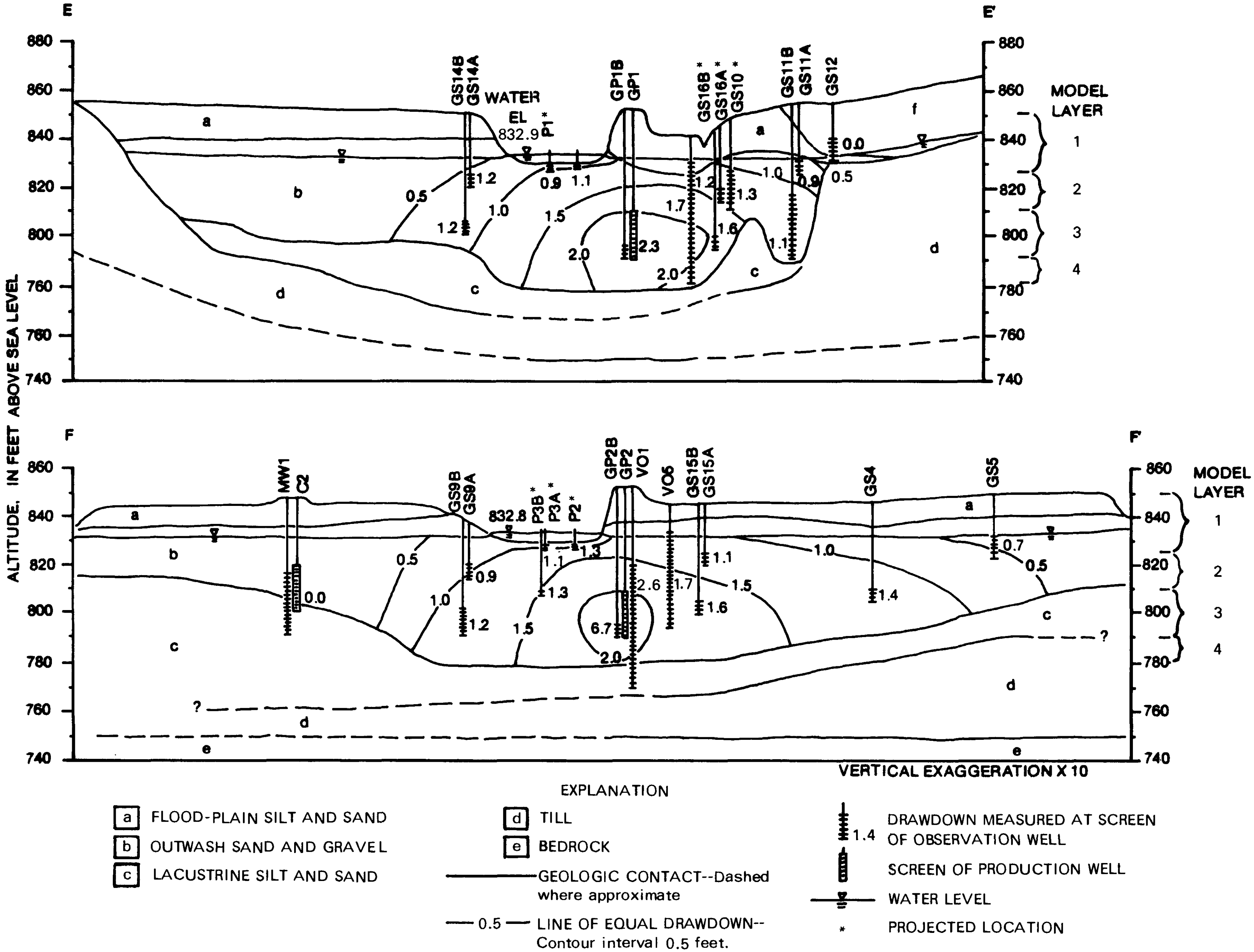


Figure 19.--Generalized sections E-E' and F-F' showing vertical distribution of drawdown

Calibrated Values

The optimum values obtained in the calibrated model (table 6) were those which accurately simulated the distribution of the observed drawdown and produced the lowest root-mean-square difference between simulated and observed drawdowns. Horizontal and vertical hydraulic conductivity values were first assigned as a single value for each model layer and were then adjusted within each layer to match measured drawdowns. The other properties were adjusted uniformly at all model blocks. Adjustments of horizontal hydraulic conductivity, storage coefficient, and specific yield were made through comparison of simulated drawdowns with those observed during the aquifer tests. Vertical hydraulic conductivity was adjusted through comparisons of simulated and observed drawdowns at paired wells screened in the upper and lower parts of the aquifer (for example, wells GS9A and GS9B, which represent layers 2 and 3; see figs. 18 and 19). Vertical hydraulic conductivity of the riverbed was adjusted through comparison of simulated and observed drawdowns in observation wells installed in the Susquehanna River.

Observed drawdowns were greater in the deeper part of the aquifer than in the shallow part. Simulations of this pattern were improved by increasing the hydraulic conductivity of layer 3 from 2,000 to 10,000 ft/d and increasing the anisotropy in layers 3 and 4 to 125:1 and in layers 1 and 2 to 250:1. These changes are consistent with observations of silty lenses throughout the aquifer, which are more prevalent in the upper part of the sand and gravel deposits. The high values of hydraulic conductivity were required to simulate the 2.34-ft drawdown in well GP1B (near the production well) and the 0.74 ft drawdown in well GS5 (800 ft upgradient of the production well).

To improve the match to observed drawdowns, hydraulic conductivity was decreased near well GP2 (fig. 19). Although both wells GP1 and GP2 were pumped at the same rate, 2.3 ft of drawdown was measured in observation well GP1B, and 6.7 ft of drawdown was measured in observation well GP2B. Both are the same distance from the respective production wells and screened at the

Table 6.--Optimum values obtained from calibrated model for hydraulic properties of aquifer materials.

Term	Range
Horizontal hydraulic conductivity (ft/d)	
Layer 1	50-500
Layer 2	330-2,000
Layer 3	750-10,000
Layer 4	200-2,000
Vertical hydraulic conductivity (ft/d)	
Layer 1	2
Layer 2	1-8
Layer 3	6-80
Layer 4	8-16
Specific yield (layer 1)	0.25
Storage coefficient (all layers)	10^{-3}
Vertical hydraulic conductivity of riverbed (ft/d)	0.2

same depth. The difference in hydraulic conductivity in these two areas (less than 100 ft apart) may be due to precipitation of iron oxide on the aquifer material near well GP2; water produced from this well has a much higher iron concentration than that produced by GP1 (Ground Water Associates, Inc., 1982).

Vertical hydraulic conductivity of the riverbed was decreased to match drawdowns measured in observation wells in the river that were screened beneath the riverbed (wells P1 to P3B). Decreasing the vertical hydraulic conductivity also improved the match between simulated and observed drawdowns in wells GS14 and GS9, across the river from the production wells. These calibrated values of the vertical hydraulic conductivity are about 5 percent of the values for horizontal hydraulic conductivity determined by piezometer tests. This indicates an anisotropy of 20:1 in the riverbed, which is within the range reported for unconsolidated materials (Todd, 1980, p. 81).

SENSITIVITY OF SIMULATED DRAWDOWNS

Transient-state simulations were run to investigate the sensitivity of simulated drawdowns to changes in the maximum hydraulic conductivity (K_{max}) of layer 3 in the model. Two simulations of the aquifer test were run, in which (1) K_{max} was reduced to the value of 2,000 ft/d estimated from analysis of aquifer-test data (run A, table 7); and (2) K_{max} was reduced to 7,500 ft/d while vertical hydraulic conductivity of the aquifer material (K_v) and the riverbed (K_r) were increased by 100 percent (run B, table 7). Maximum hydraulic conductivity was varied from 1,000 to 10,000 ft/d in an additional series of transient-state simulations.

The maximum hydraulic conductivity value of 10,000 ft/d used in model layer 3 gave the best fit between observed and simulated drawdowns. As indicated in table 7, root-mean-square (RMS) differences in drawdown obtained with a maximum hydraulic conductivity of 2,000 ft/d (run A) were much greater than those obtained from the calibrated value of 10,000 ft/d. The greatest differences in drawdowns resulting from the lower values of hydraulic conductivity are near the edges of the drawdown cone at wells GS4 and GS5 and near the production well GP1. (See table 7 and fig. 18.) Lower values of hydraulic conductivity produced more drawdown near the production well than was observed and limited the extent of drawdowns to an area smaller than was actually affected by the aquifer test.

The effect of increasing the maximum hydraulic conductivity (K_{max}) from 1,000 to 10,000 ft/d is illustrated in figure 20 as the RMS difference in computed drawdowns at the Kirkwood well field. Increasing hydraulic conductivity from 2,000 to 5,000 ft/d reduced the RMS difference by about 30 percent; a further increase from 5,000 to 10,000 ft/d reduced the RMS difference by another 16 percent.

The match between simulated and observed drawdowns can also be improved with slightly lower values of hydraulic conductivity if the vertical anisotropy and vertical hydraulic conductivity of the riverbed are increased. Run B (table 7) used a maximum hydraulic conductivity of 7,500 ft/d and produced the same average RMS difference as did the calibrated values. However, differences between computed and observed drawdowns were greater in run B near the of the drawdown cone at wells GS4 and GS5 and across the river from the pumping wells at wells GS9B and GS14B.

RMS differences in drawdown computed either from calibrated values or the values in run B are small enough to be explained by lateral variability in the anisotropy of the aquifer material or by variability of vertical hydraulic conductivity along the riverbed. Other combinations of values for the hydraulic properties of the aquifer material may also produce RMS differences as low as those attained during model calibration. However, the sensitivity analysis indicates that the maximum hydraulic conductivity of the aquifer material is in the range of 5,000 to 10,000 ft/d because the use of lower values significantly increased RMS differences in computed drawdowns.

Table 7.--Differences between drawdowns observed during October 1984 aquifer test at Kirkwood and drawdowns computed by transient-state simulations after calibrated model values were changed. Run A: K_{max} = 2,000 ft/d; Run B: K_{max} = 7,500 ft/d, and K_v and K_R increased 100 percent.

[All values are in feet. Locations are shown in pl. 1.]

Well number	Model layer	Wells GP1 and GP2 pumped at 1,000 gal/min for 23 hours: Drawdown	Difference Run A	Difference Run B	Well GP1 pumped at 1,000 gal/min for 5 hours: Drawdown	Difference Run A	Difference Run B
GS3	2-3	1.37	-0.08	-0.30	0.96	-0.10	-0.18
GS4	3	1.42	-.46	-.49	.79	-.13	-.11
GS5	2	.74	-.43	-.32	.67	-.41	-.30
GS6	1	.00	.00	.00	.00	.00	.00
GS7	2	1.18	.00	-.27	.84	.01	-.11
GS9A	2	.85	.13	-.18	.64	.09	-.10
GS9B	3	1.23	-.02	-.45	.73	.10	-.13
GS10	2	1.56	-.29	-.60	.98	-.05	-.20
GS11A	2	1.20	-.25	-.54	.84	-.11	-.28
GS11B	3	1.08	.13	-.29	.73	.14	-.08
GS12	1	.00	.02	.02	.00	.02	.02
GS13	1	.00	.00	.00	.00	.00	.00
GS14A	2	1.16	-.24	-.54	.83	-.17	-.33
GS14B	3	1.16	-.01	-.45	.77	-.01	-.23
GS15A	2	1.07	.42	.15	.96	.24	-.28
GS15B	3	1.63	.50	-.11	.97	.25	-.01
GS16A	2	1.49	-.19	-.49	.90	.06	-.10
GS16B	3	1.53	.13	-.30	.76	.36	.14
GS17	1	.00	.00	.00	.00	.00	.00
GS18	1	.00	.00	.00	.00	.00	.00
GS19	1	.00	.00	.00	.00	.00	.00
V01	2-4	2.57	1.29	-.21	.99	.57	.11
V03	1-4	1.76	.19	-.21	1.04	.26	.00
V04	2-4	1.53	.00	-.31	.95	.05	-.09
V05	1-3	1.70	-.22	-.51	1.06	-.09	-.21
GP1B	3	2.34	2.22	.20	1.40	1.69	.36
GP2B	3	6.73	-.78	-2.98	1.08	.87	.22
P1B	1	1.30	.02	-.27	.76	.20	.27
P2	1	1.26	.00	-.30	.75	.14	-.05
P3A	1	1.06	.10	-.18	.40	.44	.26
P3B	3	1.30	.49	-.01	.71	.49	.20
	RMS difference		28%	27%		43%	23%

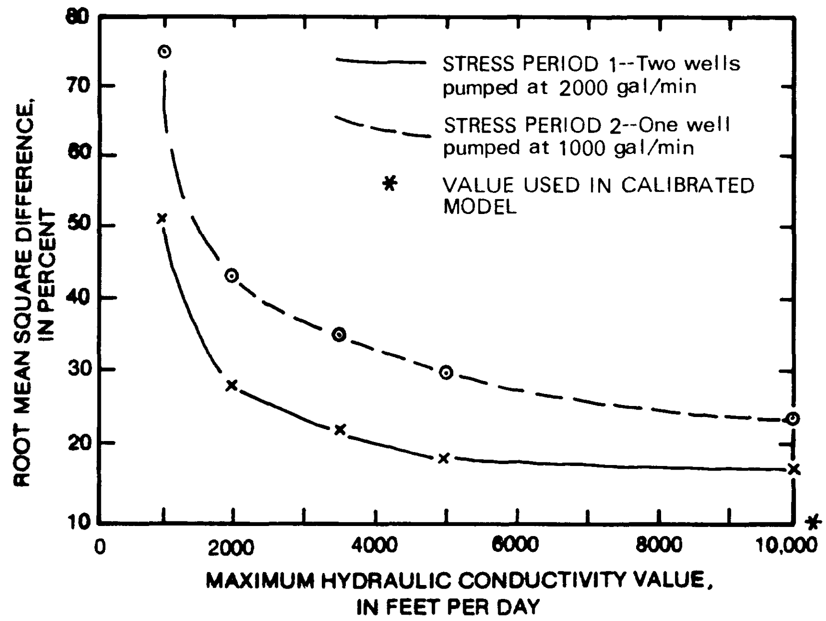


Figure 20.

Root-mean-square-differences in simulated drawdowns over range of maximum values of hydraulic conductivity used in model simulations.

SIMULATION OF GROUND-WATER WITHDRAWALS

The calibrated model was used to simulate ground-water flow to production wells in the Kirkwood and Conklin well fields. The ground-water flow paths generated by the model were used to delineate the catchment area associated with each well field. Computed rates of ground-water flow were then used to estimate the quantity of flow from various sources of recharge within the Kirkwood catchment area.

Kirkwood and Conklin Well-Field Catchment Areas

The well-field catchment areas were delineated by the model calibrated to simulate steady-state conditions of October 1984. Because ground-water levels and recharge are lowest during this period, the rate of river-water infiltration to the aquifer estimated by the model represents a maximum potential rate. To account for the intermittent pumping of the well fields, Kirkwood well GP1 in the simulation was pumped at 75 percent of its capacity, or 1.08 Mgal/d, and Conklin well C2 was pumped at 50 percent of capacity, or 0.16 Mgal/d.

Simulated flow direction and sources of ground-water flow to production wells are shown by the flow net in figure 22 (p. 52), and hydraulic-head contours for layer 1, which represents the water table, are shown separately in figure 23 (p. 54). The simulated contours agree closely with the water-table contours interpolated from water-level measurements (fig. 6).

The flow net is based on streamlines constructed from flow rates simulated through each model block. Ground-water flow is simulated by the model as flow between adjacent model blocks, as shown schematically in figure 21.

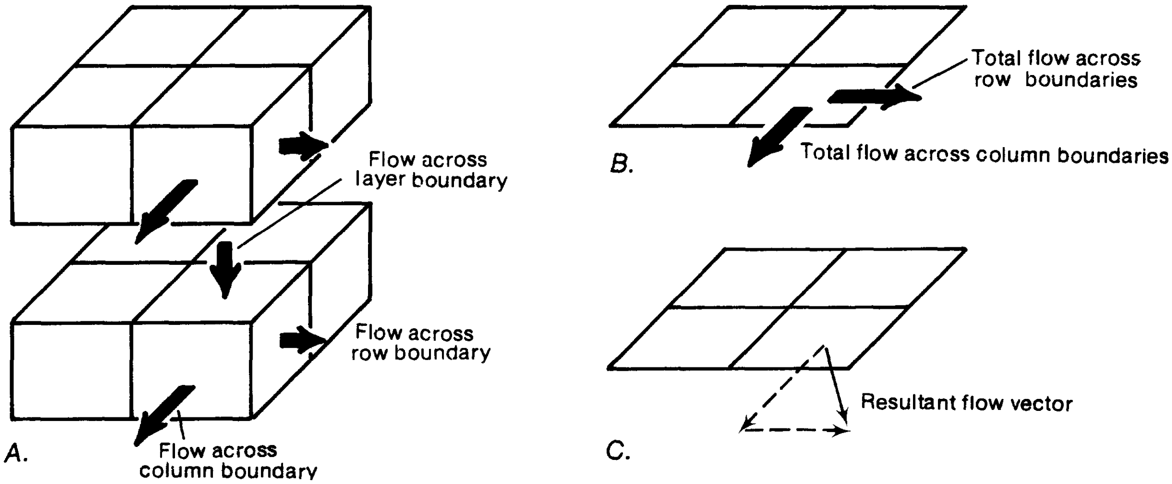
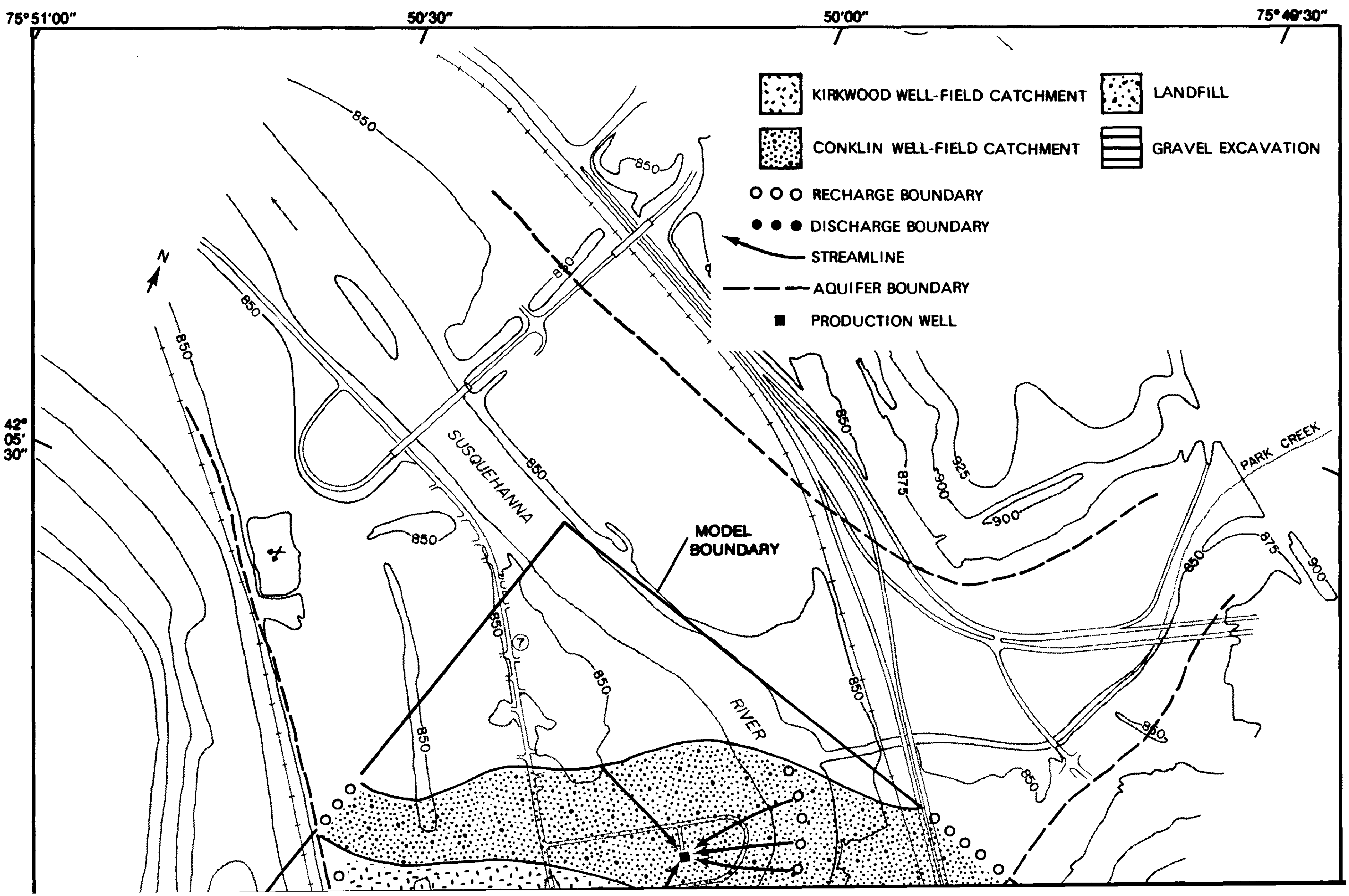


Figure 21.--Schematic diagram of flow vectors for model blocks. A. Flow across block boundaries simulated by model. B. Simulation of flows across row and column boundaries along a vertical section. C. Vector addition to obtain resultant flow vector through block.

Flows across row and column boundaries were summed for each model layer to obtain a two-dimensional representation of the flow system (fig. 21b). These flows through the block boundaries were represented by vectors that were added to obtain the resultant flow vector through the block (fig. 21c). Boundaries of well-field catchment areas (coincident with ground-water divides) were drawn to include all vectors toward the well field. The size of the well-field catchment areas defined by the flow net within the modeled area is 250 acres (0.38 mi^2) for well GP1 in Kirkwood and 51 acres (0.08 mi^2) for well C2 in Conklin.

Streamlines representing flow paths from recharge to discharge boundaries were also derived from the map of flow vectors. The spacing of the streamlines around the pumping wells in figure 22 was chosen such that the rate of ground-water flow between any two streamlines is the same. The area between two adjacent streamlines is termed a stream tube. Ten streamlines were drawn to the Kirkwood well field; thus, the flow within each stream tube represents 10 percent of the total volume produced by the well. Similarly, five streamlines were drawn to the Conklin well, and the volume of flow within each stream tube represents 20 percent of the ground-water produced.

The flow net (fig. 23) can be used to indicate where pumped water originates. Interpretation of the flow net is complicated, however, because some streamlines originate at the river, indicating infiltration from the river, and others terminate at the river, indicating ground-water discharge to the river. Still others begin at the model's southern boundary and continue northward beneath the river to the well field. These flow patterns are illustrated in cross section in figure 24, which indicates that some of the ground water produced at well GP1 originates in the Conklin area and flows eastward beneath the river to well GP1. To identify the amount of water derived from various sources required analysis of the flownet and of recharge to blocks in the model during steady-state simulation.


75°51'00"
50'30"
50'00"
75°49'30"
42°
05'
30"
KIRKWOOD WELL-FIELD CATCHMENT
CONKLIN WELL-FIELD CATCHMENT
LANDFILL
GRAVEL EXCAVATION
RECHARGE BOUNDARY
DISCHARGE BOUNDARY
STREAMLINE
AQUIFER BOUNDARY
PRODUCTION WELL
MODEL BOUNDARY
SUSQUEHANNA
RIVER
PARK CREEK
N
850
875
900
925

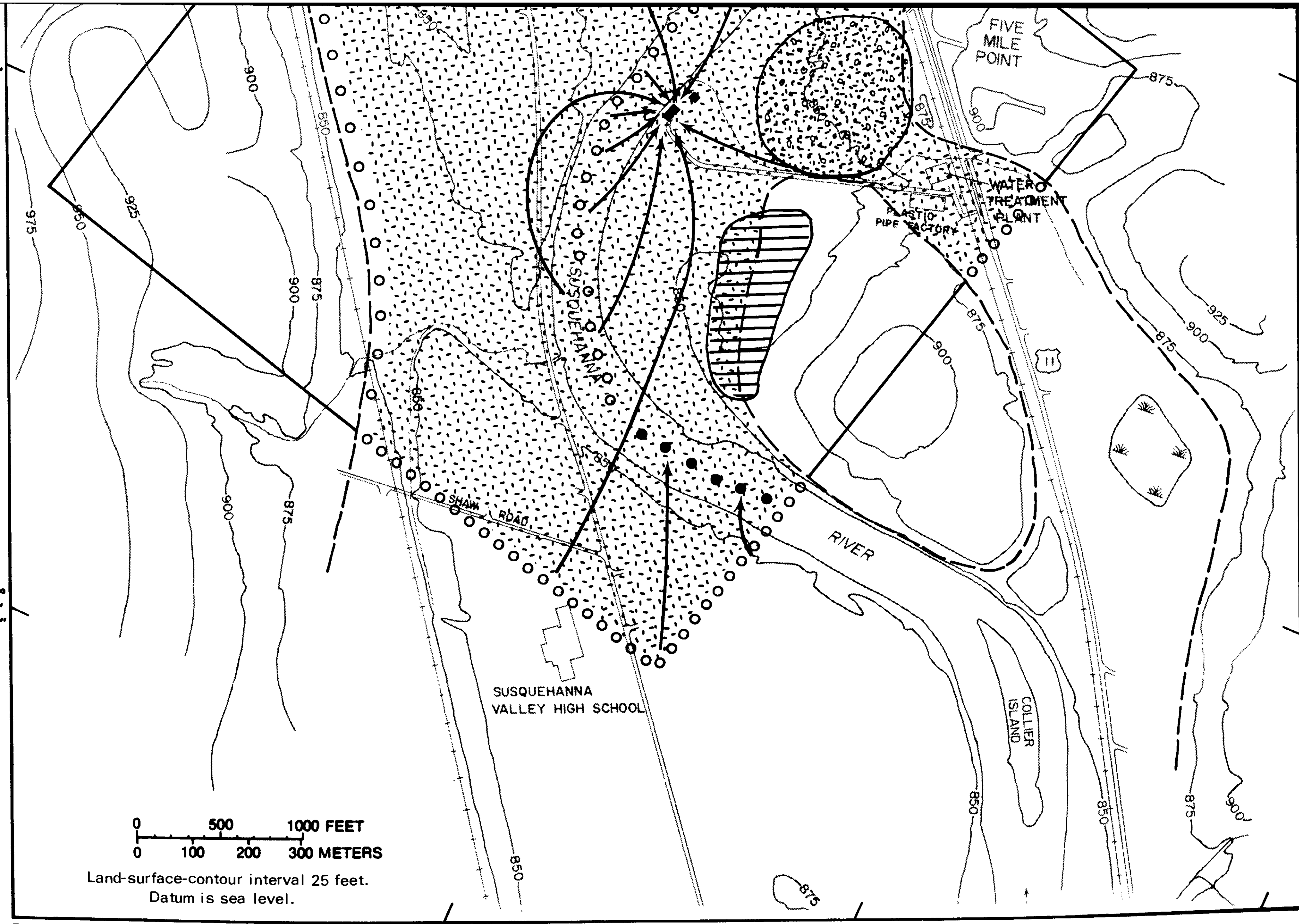
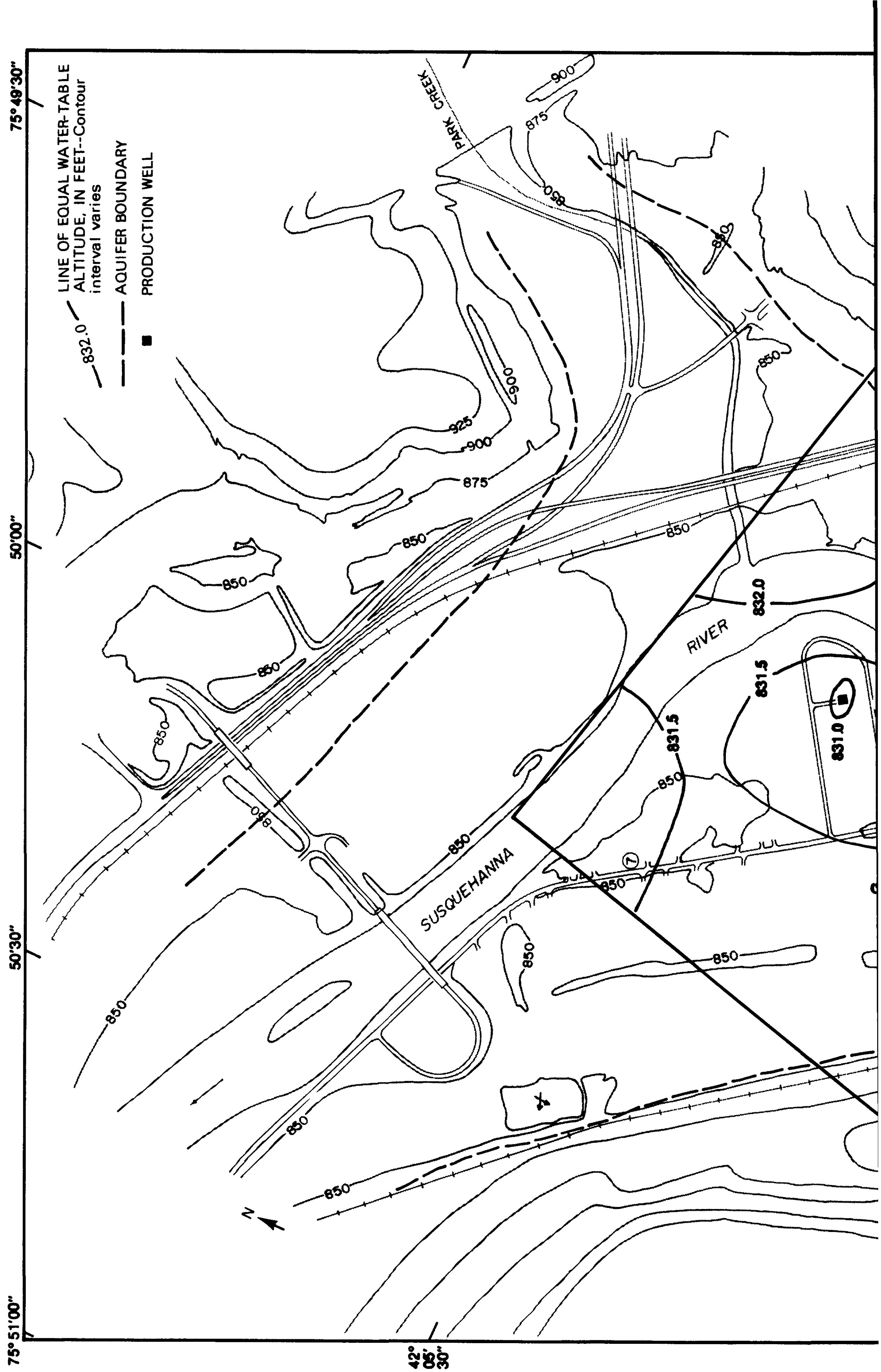
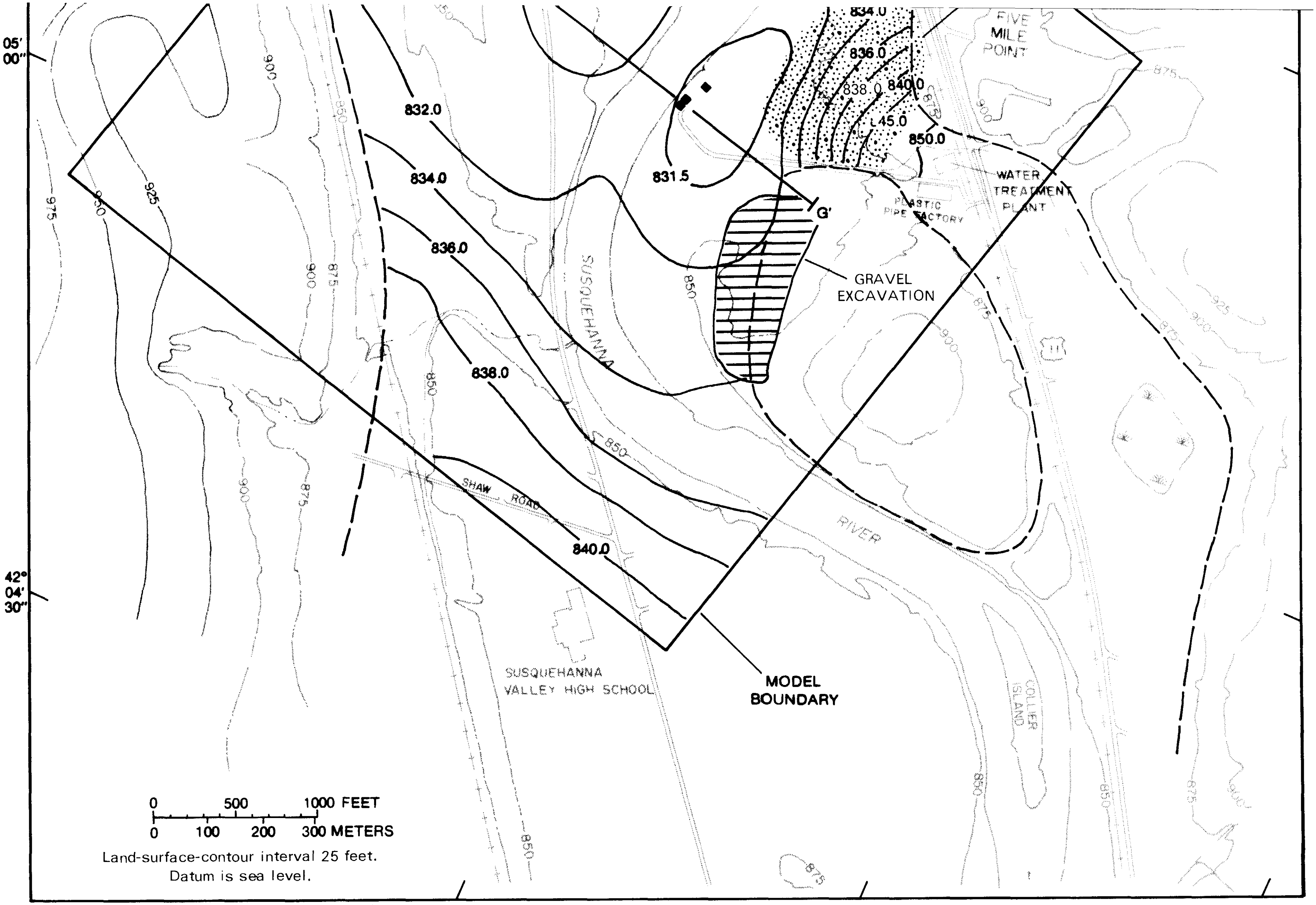


Figure 22.--Catchment areas for production wells and direction of ground-water flow derived from steady-state simulation of ground-water withdrawals.



832.0 LINE OF EQUAL WATER-TABLE ALTITUDE, IN FEET--Contour interval varies
AQUIFER BOUNDARY
PRODUCTION WELL
PARK CREEK
SUSQUEHANNA
RIVER
900
875
850
925
832.0
831.5
831.0
N
75°49'30"
50'00"
50'30"
75°51'00"
42° 05' 30"



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 23.--Water-table contours derived from steady-state simulation of ground-water withdrawals.

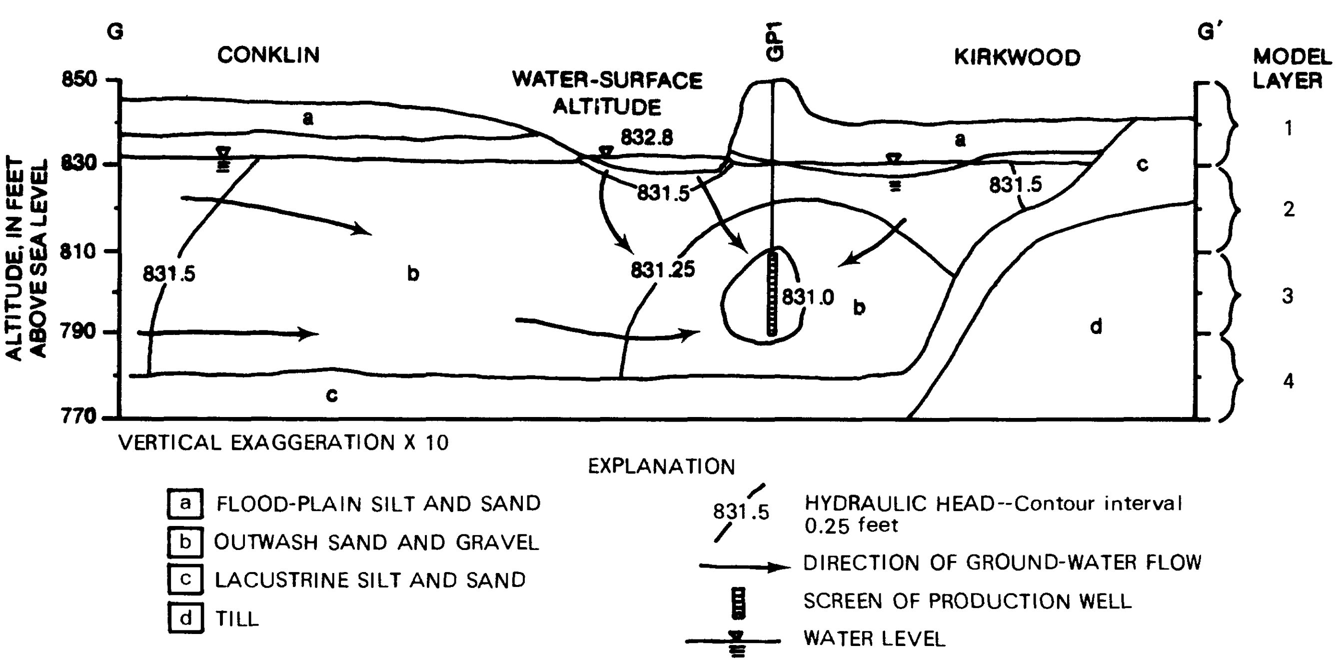


Figure 24.--Generalized section showing vertical distribution of hydraulic head resulting from steady-state ground-water withdrawals.

Sources and Areas of Recharge in Kirkwood Well Field

Recharge within the catchment area of the Kirkwood well field was estimated from the ground-water budget computed by the steady-state simulation of ground-water withdrawals in October 1984. Recharge includes (1) direct recharge from precipitation, (2) infiltration from tributary streams, (3) induced infiltration from the Susquehanna River, and (4) underflow from model boundaries. The relative contributions of these sources to the total well-field production are summarized in table 8, which indicates that 58 percent of recharge was infiltration from the river.

Recharge derived from specific parts of the Kirkwood well-field catchment area was computed from the flow net (fig. 22) and from the sources of recharge listed in table 8. The Kirkwood well-field catchment area was divided into four areas: the Susquehanna River, the Kirkwood side of the river, the Conklin side of the river, and the landfill. Recharge from the Kirkwood side of the river consists of infiltration from precipitation; recharge from the Conklin side includes precipitation, underflow from outside the modeled area, and infiltration from tributary streams. Recharge from the landfill includes infiltration from precipitation and underflow. The percentage of recharge derived from each of these areas is illustrated in figure 25. Most of the ground water produced by the well field originates from the Susquehanna River and the Conklin side of the river.

The relative contribution of these areas to the Kirkwood well field is likely to change in response to seasonal variations in precipitation and evapotranspiration. From November through May, most vegetation is dormant and evapotranspiration is at minimum. During these months, infiltration from precipitation increases, which results in higher river discharge and ground-water levels. (See figs. 7 and 9.) Increased recharge to the Kirkwood well field from precipitation would cause the amount of recharge induced from the Susquehanna River to decrease. Recharge from the Susquehanna River is greatest from June through September and reaches a seasonal maximum at the end of the growing season.

Table 8.--Recharge to Kirkwood well-field catchment area from major sources during October 1984.

[Recharge values are in Mgal/d]

Source of recharge	Volume	Percentage of total
Direct recharge from precipitation	0.16	15
Infiltration from tributary streams	.03	3
Infiltration from the Susquehanna River	.63	58
Underflow across model boundaries	.26	24
Total	1.08	100
Size of catchment area (mi^2)	0.39	

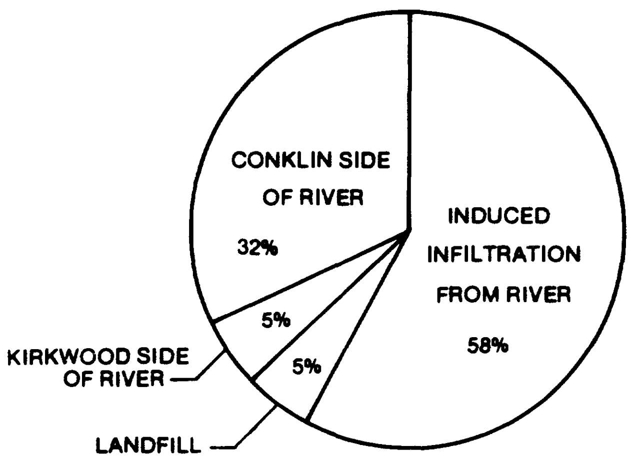


Figure 25.--Relative contribution to Kirkwood well field from major recharge areas within the well-field catchment area during low river stage.

Sensitivity of Simulations

Steady-state simulations were used to assess the effect of uncertainty in calibrated values of vertical hydraulic conductivity of the riverbed (K_R) on the well-field catchment areas and sources of recharge predicted by the model. The sensitivity analyis consisted of varying K_R over a range of 0.005 to 2.0 ft/d.

Increasing the vertical hydraulic conductivity of the riverbed (K_R) to 2.0 ft/d decreased the size of the well-field catchment areas, as shown in figure 28A (p. 60), so that the Kirkwood catchment area does not extend beyond the river. Decreasing the K_R to 0.02 ft/d increased the size of the catchment area to include nearly the entire modeled area (fig. 28, p. 62). The relationship of predicted catchment area to K_R is plotted in figure 26. Infiltration from the river increases as K_R increases, and vice-versa (table 9). For values of K_R above 2.0 ft/d, infiltration from the river is so large that changes in K_R have little effect on the size of the catchment area. Similarly, for values below 0.2 ft/d, infiltration from the river becomes less significant and no longer affects the catchment area. Recharge from three sources corresponding to different values of K_R is depicted in figure 27. This plot indicates that, at K_R values smaller than about 10^{-2} ft/d, the river contributes little recharge to the aquifer.

Steady-state simulations indicate that the size of the predicted well-field catchment area is sensitive to the vertical hydraulic conductivity of the riverbed. Transient-state simulations indicate that the set of calibrated values used in the model is not unique. The value of K_R can be decreased by 100 percent without significantly affecting the difference between simulated and observed drawdowns. Uncertainty in the value of K_R is therefore the most limiting factor in the interpretation of model results. This uncertainty could be reduced by using other methods to estimate the hydraulic conductivity or to verify model results, such as (1) permeameter tests on core samples of the riverbed, (2) analysis of water-quality data to estimate the contribution of river water to a production well, or (3) modeling of ground-water temperatures to estimate the volume of river-water infiltration into the aquifer.

Figure 26.

Size of catchment area as a function of riverbed hydraulic conductivity.

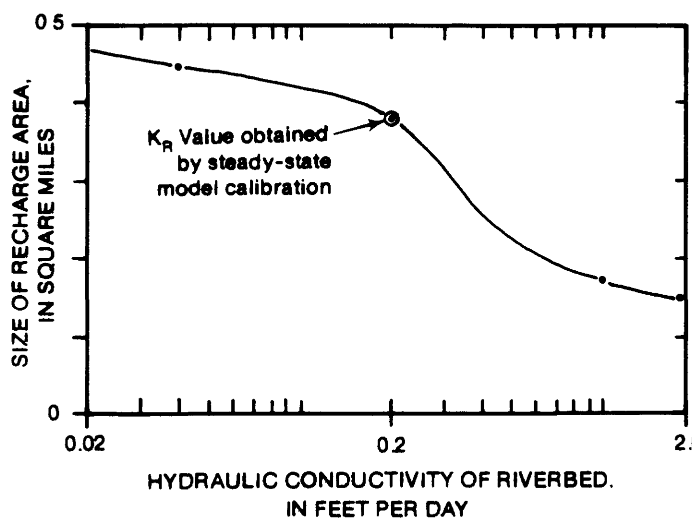


Table 9.--Effect of changes in vertical hydraulic conductivity of the riverbed on recharge to Kirkwood well-field catchment area from major sources, for steady-state conditions October 1984.

[Recharge values are in percent.]

Source of recharge	Vertical hydraulic conductivity of riverbed		
	0.02	0.20*	2.0
Direct recharge from precipitation	18	15	6
Infiltration from tributary streams	3	3	--
Infiltration from Susquehanna River	15	58	74
Underflow across model boundaries	64	24	20
Total	100	100	100
Size of catchment area within modeled area (mi^2)	0.47	0.39	0.15

*value obtained by steady-state model calibration

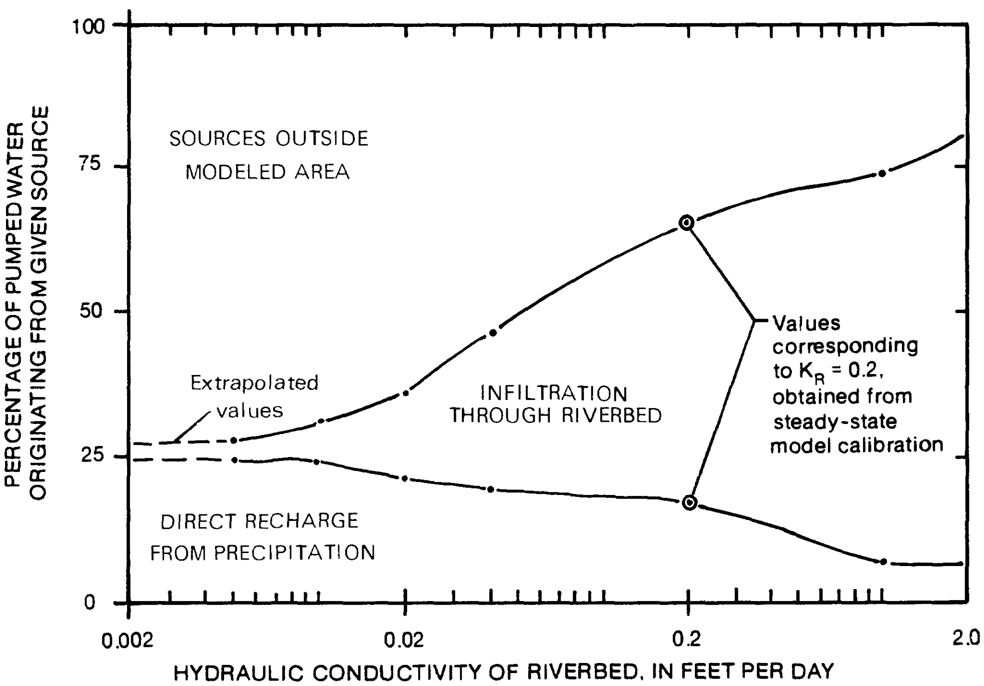
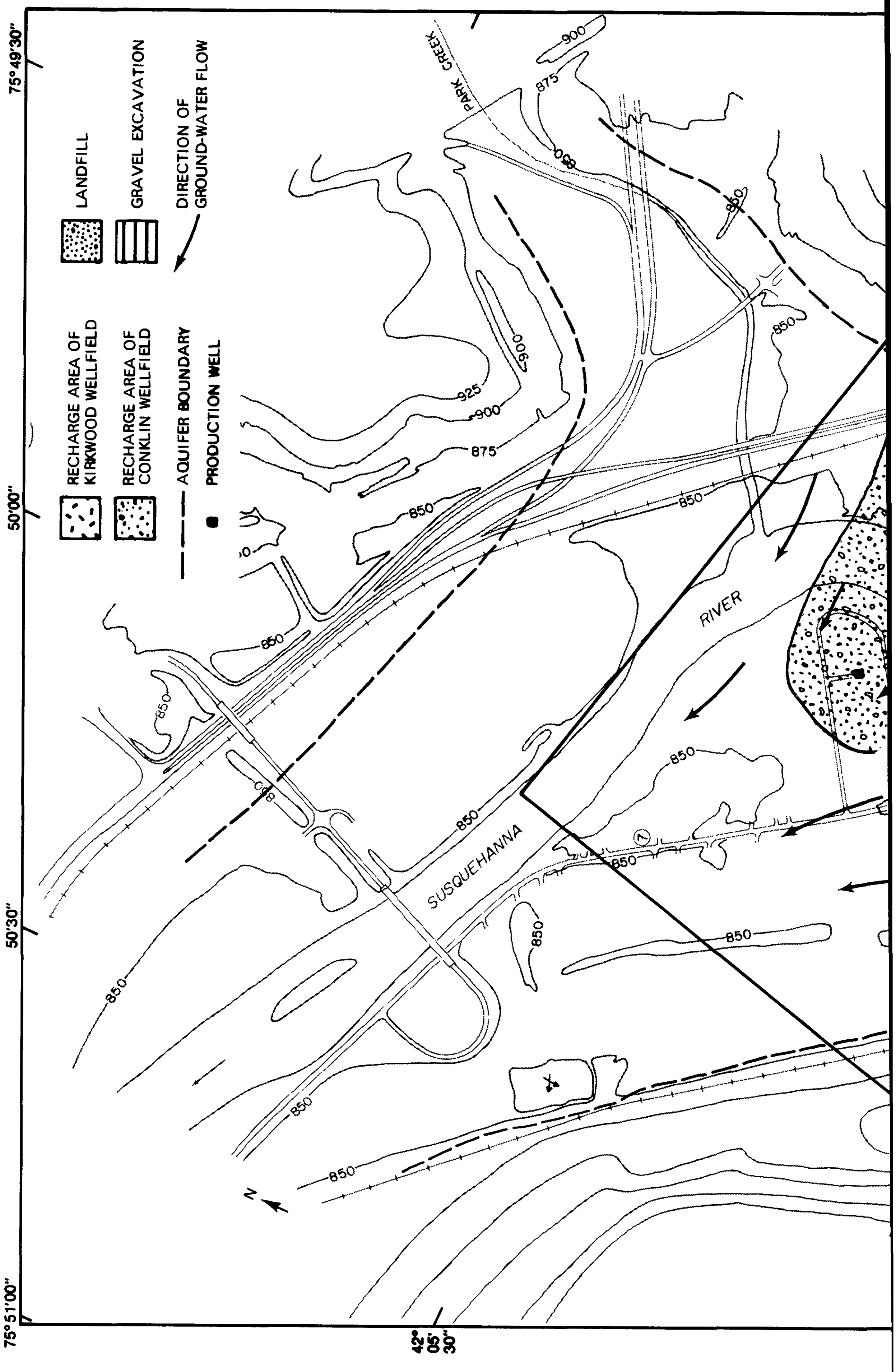
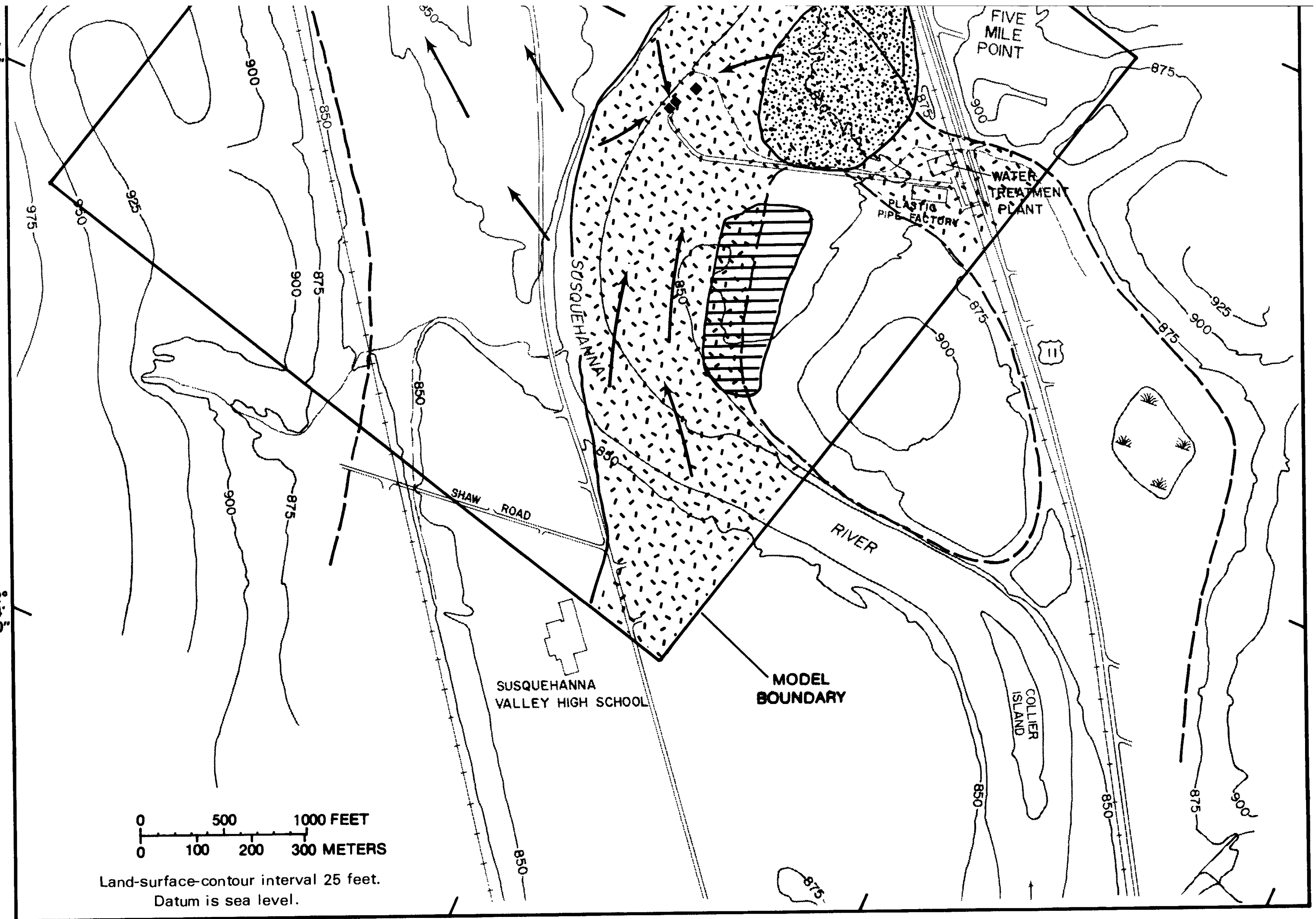


Figure 27.--Predicted recharge from major sources as a function of riverbed hydraulic conductivity.

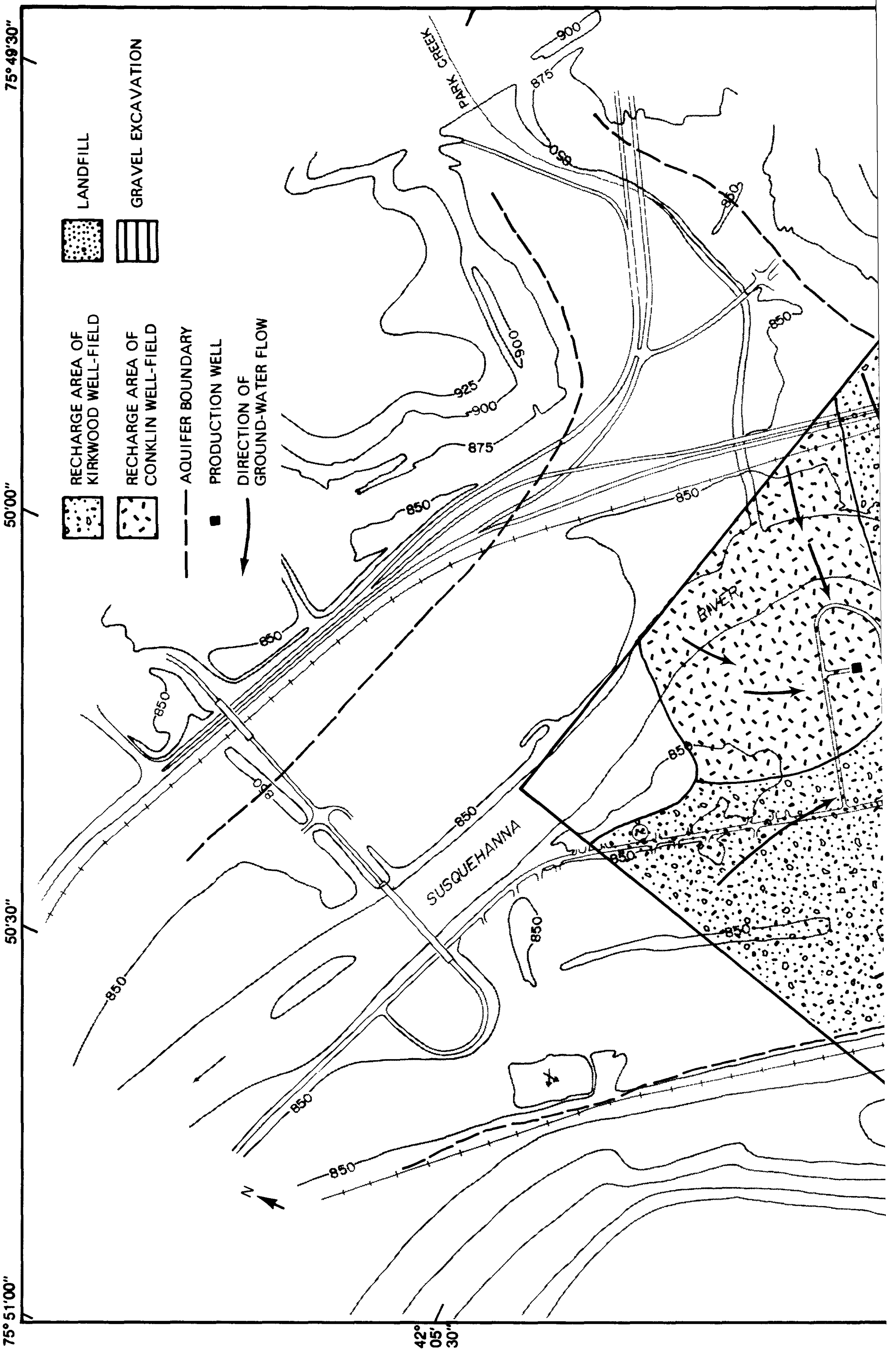


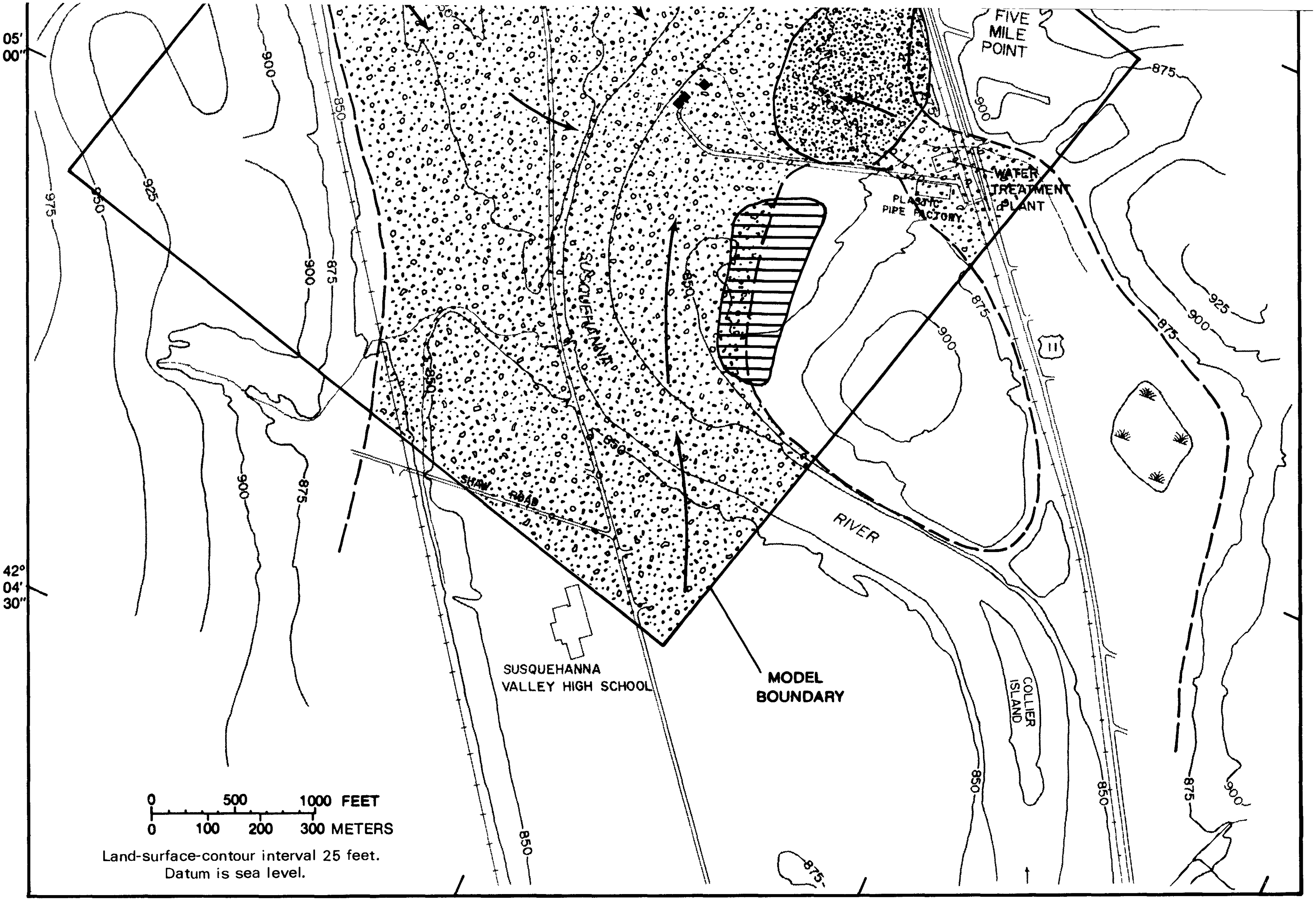
RECHARGE AREA OF KIRKWOOD WELLFIELD
RECHARGE AREA OF CONKLIN WELLFIELD
AQUIFER BOUNDARY
PRODUCTION WELL
LANDFILL
GRAVEL EXCAVATION
DIRECTION OF GROUND-WATER FLOW
PARK CREEK
SUSQUEHANNA
RIVER
N
75°49'30"
50'00"
50'30"
75°51'00"
42°05'30"
925
900
875
850



Base from Susquehanna River Basin Commission, 1974 1:2400

Figure 28A.--Well-field catchment area when riverbed hydraulic conductivity is 2.0 ft/d.


75°49'30"
50'00"
50'30"
75°51'00"
42°
05'
30"
LANDFILL
GRAVEL EXCAVATION
RECHARGE AREA OF KIRKWOOD WELL-FIELD
RECHARGE AREA OF CONKLIN WELL-FIELD
AQUIFER BOUNDARY
PRODUCTION WELL
DIRECTION OF GROUND-WATER FLOW
PARK CREEK
SUSQUEHANNA
RIVER
N
900
875
850
925



Base from Susquehanna River Basin Commission,1974 1:2400

Figure 28B.--*Well-field catchment area when riverbed hydraulic conductivity is 0.02 ft/d.*

SUMMARY

A three-dimensional finite-difference model was developed to simulate ground-water flow and infiltration from the Susquehanna River near well fields in the towns of Kirkwood and Conklin. The aquifer system consists of sand and gravel deposited in a preglacial river valley during the recession of glacial ice. The saturated thickness of the aquifer ranges from 10 to 50 ft and is greatest in an elliptical area extending westward from the Kirkwood (east bank) well field. Ground water in the aquifer is in good hydraulic connection with the Susquehanna River, and ground-water levels respond to changes in river stage. Vertical profiles of ground-water temperatures measured in wells near the river throughout the year indicate that river water influences ground-water temperatures and confirm that river water infiltrates into the aquifer.

The calibrated model used horizontal hydraulic conductivity values of 50 to 10,000 ft/d for the sand and gravel, and vertical hydraulic conductivity values of 0.1 to 40 ft/d. The riverbed thickness was calculated from results of piezometer tests to be 2 ft, and the vertical hydraulic conductivity of the riverbed in the calibrated model was 0.2 ft/d.

Model accuracy was estimated through comparison of simulated and observed drawdowns during an aquifer test at the Kirkwood well field. The root-mean-square difference in drawdown ranged from 17 to 24 percent.

Model simulations were used to prepare a flow net showing the direction and rate of ground-water flow to the production wells. The sizes of the Kirkwood well-field catchment area was estimated to be 250 acres (0.39 mi^2); that of the Conklin well field was 51 acres (0.08 mi^2). The ground-water budget during steady-state simulations showed that 58 percent of the ground water withdrawn by the Kirkwood well field during a period of low river stage in October 1984 was derived from the Susquehanna River; 32 percent was derived from the Town of Conklin; 5 percent was derived from the landfill near Kirkwood, and 5 percent was derived from other areas in the Kirkwood side of the river.

The hydraulic property to which recharge and well-field catchment areas are most sensitive is vertical hydraulic conductivity of the riverbed. Uncertainty in the value of this term is the most limiting factor in the interpretation of model results.

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APPENDIX

ESTIMATION OF HYDRAULIC CONDUCTIVITY BY ANALYTICAL METHODS

Estimates of hydraulic conductivity that were used in the simulation model were developed from analyses of drawdown and recovery data recorded during aquifer and slug and bail tests. The aquifer tests consisted of pumping one or more production well at a constant rate for at least 24 hours while recording water-level changes in several observation wells. The piezometer tests consisted of bailing or slugging an observation well and recording changes in the water level until the initial water level was reached.

Aquifer Tests

Drawdown data were analyzed from tests made in February 1984 at production wells GP1 and GP3 in the Kirkwood well field and in October 1983 at production well C2 in the Conklin well field. Production wells ceased pumping at least 8 hours before each test, and ground-water levels were monitored to ensure that the aquifer system had recovered before the test was begun. Water levels were not affected by pumping from neighboring well fields except those recorded at observation well MW3 in the Conklin test. These data were corrected by subtracting the additional drawdown caused by production well GP1. The locations of observation wells used in each test are shown in plate 1, and the screened intervals are listed in table 10, which summarizes the results of the aquifer-test analysis.

Drawdown data were analyzed by the method of Neuman (1975) for unconfined aquifers with delayed yield. Type curves developed by Neuman (1975) were identified for each observation well through a curve-matching procedure that used logarithmic plots of time vs. drawdown. The type curves were generated for each observation well from a computer program by Neuman (1975) for cases in which the production and observation wells partially penetrate the aquifer thickness.

The Neuman method assumes a homogeneous aquifer of infinite lateral extent--conditions that were not met in aquifer tests at the Kirkwood and Conklin well fields. Type curves were matched to the drawdown observed during the first 100 minutes of each test because the later drawdowns were influenced by the proximity of the recharge boundary corresponding to the river and by impermeable boundaries that limit the lateral extent of the aquifer.

The relatively high value of hydraulic conductivity suggests that the sand and gravel in this area may have originated as ice-contact material. This conclusion is supported by the abundance of silty layers in boreholes through the upper 20 to 30 ft of the deposit. This material becomes coarser with depth, and 8-in-diameter cobbles were encountered at 40 to 45 ft as the casing for production well GP3 was being driven.

The degree of vertical anisotropy of the aquifer material is illustrated by plots of time as a function of drawdown for three observation wells in the Kirkwood well field (fig. 2). The wells are 30 ft from production well GP3 and screened at differing intervals (pl. 1). Observation wells GS8A and GS8B

Table 10.--Estimates of hydraulic coefficients.

[Well locations are shown in pl. 1.]

A. Analysis of aquifer test data by method of Neuman (1975)

Observation well	Screened interval (ft above sea level)	Distance from production well (ft)	Hydraulic variables[1] b	T	K	K_v	S	S_y	β
Aquifer test at Kirkwood well field with GP3 pumping 2,000 gal/min, February 1984									
GS8A	815-820	30	45	64,000	1,400	26	.0110	--	0.01
GS8B	795-800	30	45	220,000	4,900	11	.0270	--	.001
VO1	770-820	260	45	73,000	1,600	4.7	.0010	--	.06
VO2	770-820	30	45	100,000	2,300	52	.0320	--	.01
VO3	780-830	160	45	91,000	2,000	1.6	.0010	--	.01
Aquifer test at Conklin well field with C2 pumping 220 gal/min, October 1983									
MW1	795-815	27	30	7,200	230	1.7	.0040	.21	.06
MW2	798-818	170	30	7,200	230	1.7	.0006	.01	.20
MW3	783-813	280	10	3,200	320	.2	.0003	--	.06

B. Analysis of specific capacity data by method of Walton (1970, fig.5.7, p. 317)

Production well	Diameter (ft)	Screened interval (ft above sea level)	Saturated thickness (ft)	Discharge (gal/min)	Drawdown[2] (ft)	Specific capacity[3] [(gal/min)/ft]	Hydraulic variables[1] T	K
GP3	2.0	790-810	45	2,000	3.0	960	[4]200,000	4,500
C2	1.0	798-818	30	350	8.0	88	17,000	580

[1] r = distance from production well (ft);
b = saturated thickness (ft);
T = transmissivity (ft^2/d);
K = horizontal hydraulic conductivity;
K_v = vertical hydraulic conductivity;
S = storage coeffieient;
S_y = specific yield;
β = $\frac{K_v}{K}\frac{r^2}{b^2}$; and

[2] Drawdown after original installation

[3] Specific capacity has been divided by productivity factor to account for partial penetration as described by Walton (1970, fig. 5.11, p. 320).

[4] Transmissivity has been multiplied by factor of 0.9 to account for larger well radius as described by Walton (1970, fig. 5.16, p. 315).

have 5-ft screens that are set above and opposite the production well screen, respectively. As expected, drawdowns in GS8B were greater than those observed in GS8A. Because observation well V02 is screened through the entire saturated thickness, drawdowns observed in V02 should represent the average drawdown in the aquifer and should fall between those observed in GS8A and GS8B. However, drawdowns observed in V02 were only slightly larger than those in GS8A. This indicates that GS8B is screened in a relatively narrow layer with high hydraulic conductivity that contributes most of the flow to the production well. The aquifer material above and below this layer has a lower hydraulic conductivity and contributes flow toward the production well screen through vertical leakage to the more productive layer. Drawdowns are propagated farther in the productive layer than in the aquifer material above and below.

The degree of anisotropy of the aquifer is reflected by the small value of β (0.001 to 0.06 from table 10) that identifies the type curve along which the drawdown data fall. The vertical hydraulic conductivity, K_v, is therefore much less than the horizontal hydraulic conductivity, K. The aquifer material was initially assumed to have a transmissivity of 100,000 ft^2/d with an anisotropy ratio of 55:1, from drawdowns recorded in the fully penetrating observation well V02. However, model simulations indicated that the aquifer transmissivity was closer to the estimate of 200,000 ft^2/d obtained from well G8B and that anisotropy varied from 250:1 to 125:1. The value of the storage coefficient, S, was assumed to be 10^{-3}.

Analysis of aquifer test data from the Conklin well field indicates that the aquifer material here is more isotropic (β = 0.06 to 0.20) and has a much lower transmissivity (7,200 ft^2/d). The storage coefficient was assumed to be 6×10^{-4}, and the specific yield was assumed to be 0.21.

Estimates of transmissivity were also computed from specific capacity data from the two production wells that were used in the aquifer tests. (See table 10.) Specific capacity was calculated from original installation data because drawdowns recorded after extended operation of the wells increased as a result of clogging at the well screen. The specific capacity was adjusted by a productivity factor to account for partial penetration, as described by Walton (1970, fig. 5.11, p. 320). Transmissivities estimated from figure 5.7 in Walton (1970, p. 317) are higher than those computed from the aquifer-test data but are close to the values obtained through model calibration.

Slug and Bail Tests

Slug and bail tests were run in several observation wells to estimate the hydraulic conductivity of the finer grained valley-fill deposits. The recovery data were analysed by the methods of Hvorslev (1951) and Cooper and others (1967); a comparison of results of the two methods is given in table 11. The estimates computed by the two methods were generally in close agreement. The hydraulic conductivity of the riverbed was assumed to range from 1.0 to 6.0 ft/d, and that of the sand and silt underlying the sand and gravel aquifer was assumed to be 0.2 to 1.8 ft/d.

Table 11.--Comparison of hydraulic conductivity values calculated by two methods from piezometer-test data.

[Values are in feet per day. Well locations are shown in pl. 1.]

Geologic unit	Observation well	Hvorslev (1951) method	Cooper and others (1967) method
Sand	GS7	0.70	1.8
	GS10	.46	.23
Sand and silt	GS1	.02	.05
Riverbed	P1A	1.0	6.0
Backfill in gravel excavation	P4	.04	.03