A MODEL FOR FLOW THROUGH A GLACIAL OUTWASH AQUIFER IN SOUTHEAST FRANKLIN COUNTY, OHIO

U.S. GEOLOGICAL SURVEY

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UNITED STATES DEPARTMENT OF THE INTERIOR

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CONTENTS

ILLUSTRATIONS

TABLES

CONVERSION FACTORS

Factors for converting inch-pound units to International System units (SI)

Multiply inch-pound units	By	To obtain SI units
foot (ft)	0.3048	meter (m)
foot per second (ft/s)	0.3048	meter per second (m/s)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
inch (in)	25.4	millimeter (mm)
million gallons per day (Mgal/d)	0.0438	$meter^3$ per second (m^3/s)

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ABSTRACT

A glacial aquifer of 70 square miles in the Scioto River valley southeast of Columbus, Ohio, was modeled as a potentially major source of water. The model was constructed from available hydrologic data: Records of precipitation, well hydrographs, well logs, two ground-water level surveys, and analyses of six aquifer tests.

Utilizing this array of data, water levels determined from a series of steady-state simulations of different hydraulic conductivity distributions were calibrated against measured (December 1977) ground-water levels. The simulations that provided the best matches used two hydraulic conductivity distributions: One was an areally varying hydraulic conductivity distribution; the other an areally uniform hydraulic conductivity (40 feet per day) distribution.

After these more probable hydraulic conductivity distributions were found, they were utilized in steady state maximal pumping simulations. The maximal well-field yield for these simulations was 20.5 million gallons per day for the areally varying hydraulic conductivity distribution, and 11.3 million gallons per day for the areally uniform hydraulic conductivity. Sensitivity of well yield to changes in well position and streambed leakance changes was investigated also.

INTRODUCTION

An evaluation of a glacial outwash aquifer in part of the Scioto River basin was made by the U.S. Geological Survey in cooperation with the city of Columbus, Ohio. The purpose of the study was to construct a digital-computer model that can simulate the ground-water system under steady-state conditions. The model developed in this study may be used as the framework for future modeling in which aquifer yields and drawdown effects can be predicted.

Some of the data used as input to the model were provided by the city of Columbus. The city obtained the data from reports by Alden E. Stilson and Associates (1976 and 1977). Also, the authors thank Fred Klaer, Jr. and Associates, Inc. for providing some hydrologic data and first-hand field observations.

DESCRIPTION OF THE AQUIFER SYSTEM

The principal source of ground water in the modeled area is glacial outwash sand and gravel underlying the valleys of the Scioto River and Big Walnut Creek. The outwash ranges from 200 ft to less than 10 ft in thickness, pinching out along the north and west parts of the modeled area. Besides this range in thickness, the degree of heterogeneity within the aquifer is high. Clay lenses and deposits of silt, fine sand, and till occur within the aquifer, locally reducing permeability. The aquifer is overlain generally by 10 to 15 ft of Holocene alluvium.

A top layer of poorly permeable material (till) extends over the aquifer in some areas. The till is thinner along the Scioto River and its tributaries than to the north and to the west of the modeled area. The bedrock east of the Scioto River is relatively impermeable shale; however, along and west of the Scioto River, bedrock is semipermeable limestone. Upward seepage from the limestone contributes water to the overlying glacial sand and gravel (Norris, 1959).

The upland glacial deposits, consisting mostly of till and minor amounts of sand and gravel, which overlie the bedrock west of the modeled area, provide some recharge to the western boundary of the outwash aquifer. For the most part, the upland deposits and the bedrock are relatively unimportant as sources of water except for farm and domestic supplies.

COMPUTER PROGRAM FOR GROUND-WATER FLOW

A finite-difference, two-dimensional ground-water flow simulation program developed by the U.S. Geological Survey (Trescott and others, 1976) was selected for all simulations described herein. The partial differential equations that describe flow through porous media for water table and artesian conditions or any combination thereof can be accommodated by this flexible program. To represent the southeast Franklin County aquifer under steady-state conditions, the combined artesian water-table option was used with river nodes as artesian and all others water table. Specific yield and storage coefficient were set to zero. The numerical algorithm selected for solving the difference equations was the strongly implicit procedure (SIP).

DESCRIPTION OF MODELED AREA AND MODEL GRID

The study area (fig. 1), which includes the southern part of the city of Columbus, is in Franklin County and the northern part of Pickaway County in central Ohio. The rectangular area includes approximately 70 mi^2, extending along the Scioto River from Columbus southward and including the communities of Obetz, Reese, Hamilton Meadows, Shadeville, and Lockbourne, and west from Groveport to approximately 1.5 mi west of the Scioto River. The principal tributaries of the Scioto in the modeled area are Big Walnut and Walnut Creeks.

An irregular rectangular grid network, designed to facilitate the finite-difference calculations, consists of 31 rows and 36 columns, dividing the modeled area into 1116 rectangular cells. The center of each cell is a finite-difference node where hydrologic information is input and aquifer head is evaluated during each simulation. In the well-field area, between the Scioto River and Big Walnut Creek, the irregular grid is more closely spaced. In this area, greater hydrologic detail can be defined and simulated in the model because nodes are more numerous per unit area.

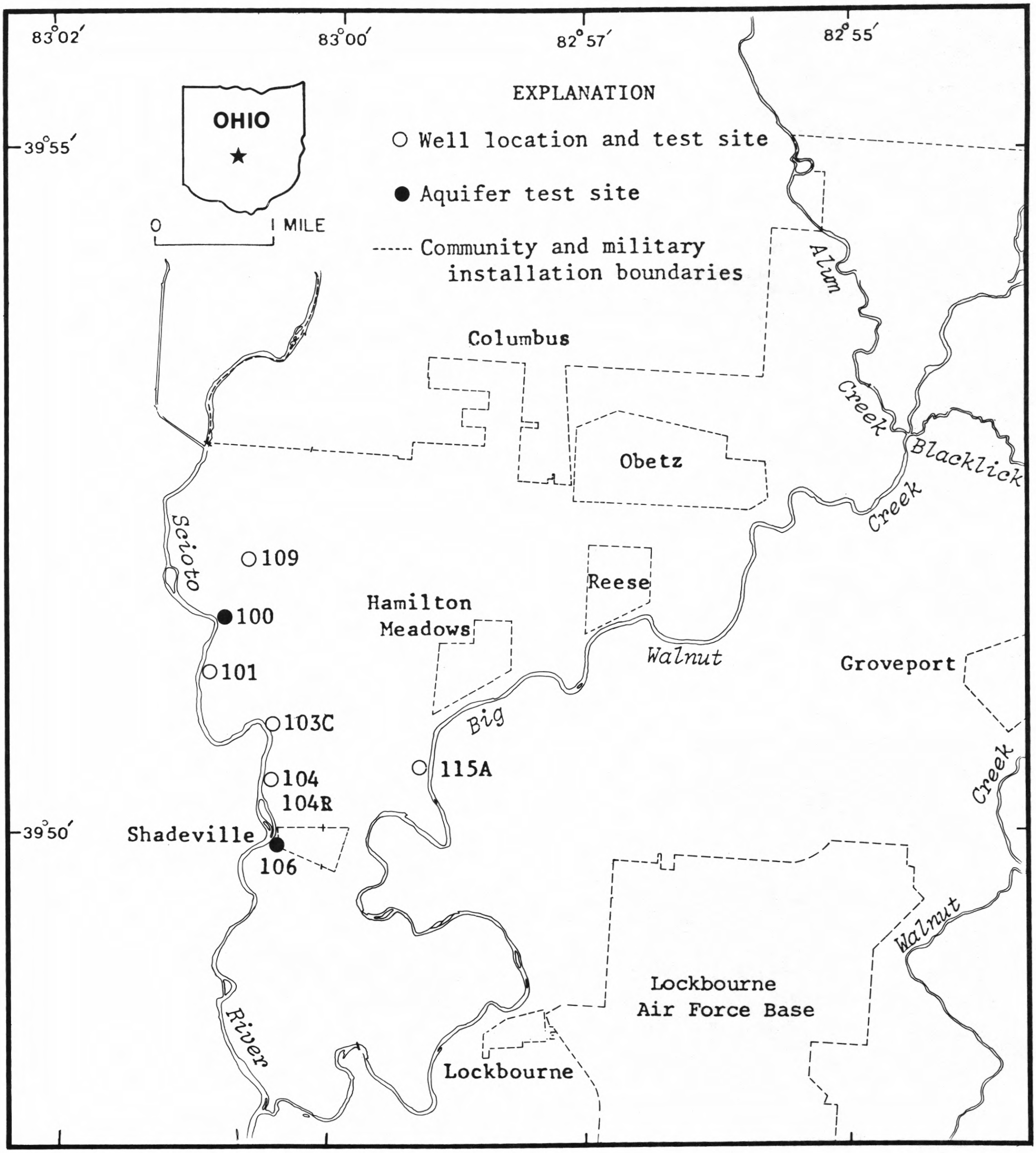


Figure 1.--Modeled area and well location.

MODEL PARAMETERS

Hydraulic Conductivity

The hydraulic conductivity at any point in the aquifer is dependent on the texture, porosity, structure, and grain shape of the material composing the aquifer. Hydraulic conductivity even of a homogeneous, isotropic aquifer is difficult to determine from sampling the aquifer material. Aquifer testing is a preferred way to determine hydraulic conductivity. However, even analysis of aquifer tests for highly heterogeneous aquifers, such as the southeast Columbus aquifer, can give only an average value of hydraulic conductivity in the vicinity of the tests. Because only a few aquifer tests were made in the modeled area (Alden E. Stilson and Associates, 1976, 1977), much of the aquifer remains untested, and extrapolation of known hydraulic conductivities to untested areas, although perhaps helpful, is uncertain.

The results of analyses of aquifer tests (Alden E. Stilson and Associates, 1976, 1977) were compiled and averaged (table 1). The tests from which these hydraulic conductivities were calculated were difficult to analyze. Many of the assumptions necessary to the analyses were not fulfilled by the circumstances of the tests, such as a river that fully penetrated the aquifer and a homogeneous, isotropic aquifer. Also, a time-drawdown technique that was used to calculate aquifer hydraulic conductivity required all delayed drainage to be ended and before boundary effects occur, such as recharge from the river. However, the data used for calculations were taken when induced infiltration was occurring and before delayed drainage had ended (Alden E. Stilson and Associates, 1976, p. 10-19). Perhaps because of these difficulties with the tests and analyses, the results for aquifer hydraulic conductivity are questionable and are higher than those obtained from analyses of aquifer tests along the Scioto River at Piketon, Ohio. For these aquifer tests at Piketon, made in what is considered to be a better aquifer, an average value of hydraulic conductivity of about 480 ft/d was calculated (Norris and Fidler, 1969). More recently, also in the Piketon area, tests by the Geological Survey showed an average hydraulic conductivity of about 450 ft/d. On this basis, most values in table 1 calculated from aquifer tests in the modeled area are probably high.

Table 1.--Hydraulic conductivity, saturated thickness, and depth to bedrock determined at aquifer test sites. (Values from report by Alden E. Stilson and Associates, 1976, 1977).

Site	Hydraulic conductivity (ft/d)	Saturated thickness (ft)	Depth from land surface to bedrock (ft)
100	590	70	88
101	560	70	80
103C	490	86	116
104	470	63	98
104R	570	63	97
106	340	60	116
115A	690	[1] 50	165
Average	530	66	109

[1] At site 115A the ground water level was 4 ft below the river surface. This low level was undoubtly due to unknown pumpage from the quarry directly across Big Walnut Creek. Also, 80 ft of clay and till between bedrock and the well screen were not considered in this determination.

Trial-and-error adjustments during calibration of the model resulted in a 12 percent reduction in each hydraulic conductivity value listed in table 1, and these adjusted values are close to those calculated at Piketon. An examination of well logs in each area showed that almost all the aquifer where well yields are high is described as sand and gravel. Hence, the saturated sand and gravel zones described in the well logs are taken as the total thickness of the aquifer. Typically, several test holes were drilled to bedrock, but the pumped well was screened in the lowest sand and gravel layer consistent with what was thought to be the most productive location. Due to the high permeability of the sand and gravel, the value of hydraulic conductivity for these zones for all well logs in the test areas seems to be approximately 1,070 ft/d.

This value of hydraulic conductivity for the sand and gravel parts of the aquifer was then used to calculate a value of average hydraulic conductivity for each well log in the modeled area that recorded lithology to bedrock (Schmidt, 1958). For those well logs that recorded only poor aquifer material, an average value of hydraulic conductivity of 0.1 ft/d was assigned. This value corresponds to clay or till (Walton, 1970) which constituted the whole section penetrated by poor wells. Hydraulic conductivity values from many logs were then grouped and averaged to obtain an average hydraulic conductivity for an area. Account was also taken of the percentage of well logs in each area that recorded drilling to bedrock. A higher percentage of wells drilled to bedrock in one area relative to another was taken as an indication of poorer aquifer materials.

The modeled area could be conveniently divided into three types of aquifer hydraulic conductivity, high, intermediate, and low (fig. 2). The low area in the northwest part was drawn on the basis of many logs that indicated no good aquifer material and low calculated average hydraulic conductivities. The same procedure was followed for the low areas on the north and west borders. The area of intermediate hydraulic conductivity, between the two areas of low hydraulic conductivity on the west boundary is the site of a buried river valley. The high area in the center of the map between the Scioto River and Big Walnut Creek coincides with a topographic high that is a kame. The already adjusted value of 470 ft/d was further reduced by trial-and-error adjustment to 400 ft/d to represent the high areas of the map (fig. 2). The intermediate values on that map were either an average determined from well logs or, lacking well logs, an intermediate value was assigned, as seemed reasonable.

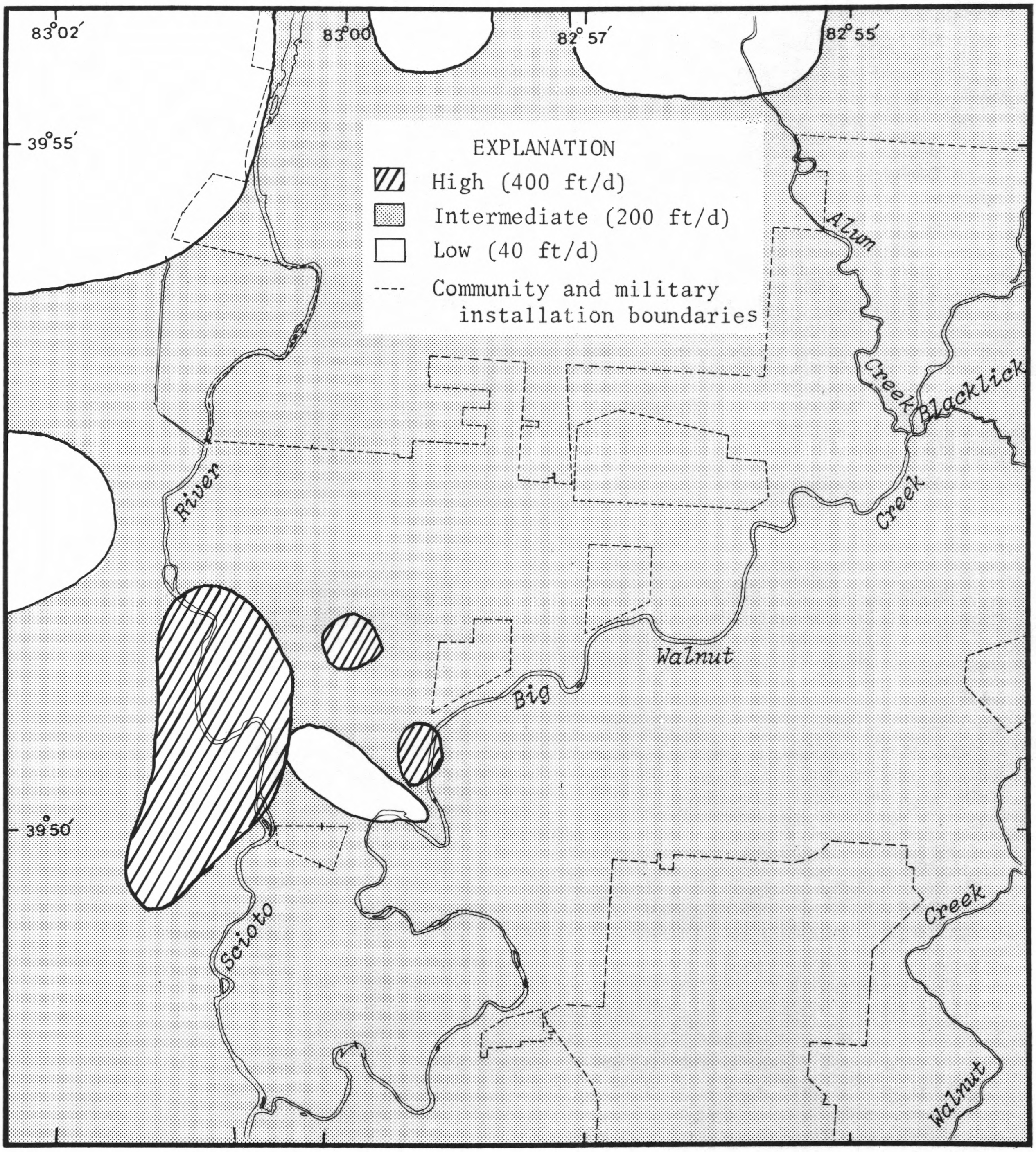


Figure 2.--Areal distribution of hydraulic conductivity used in some simulations, called K-map.

The low hydraulic conductivity area between the Scioto River and Big Walnut Creek in the vicinity of the well field was drawn on the basis of a relatively impermeable boundary determined from analysis of the test at site 115A, (Alden E. Stilson and Associates, 1977, p. 36 and 46). Also, logs of wells M13 and M4, drilled by the Geological Survey at the ends of that oblong area recorded all clay, further supporting the hypothesis of a low hydraulic conductivity area.

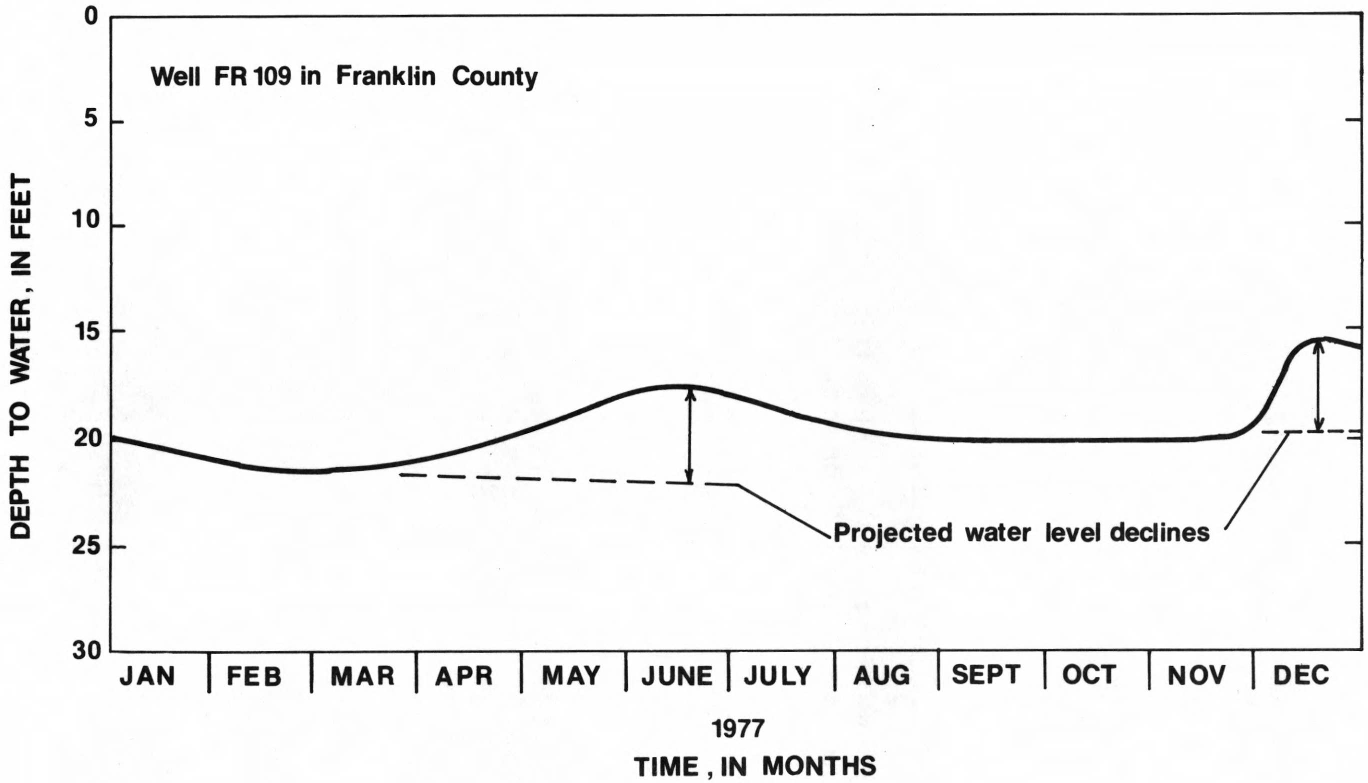
Adjusted aquifer-test results have been extrapolated to untested areas of the aquifer with the aid of well logs. As such, the map is an approximation of the possible areal variation of hydraulic conductivity in the modeled area. Uniform hydraulic conductivity distributions were also considered in some simulations, as will be discussed later.

Recharge From Precipitation

The hydrograph taken from well FR 109 was used to estimate the average recharge from precipitation for the year preceding the December 29, 1977, water-level survey. Although well FR 109 is only three quarters of a mile from the Scioto River, the hydrograph from FR 109 shows no influence of river stage, so the rises in the hydrograph are caused by recharge from precipitation. By summing those water-level rises for the year (in this case, taken as the two major rises) and multiplying that sum by an estimated specific yield (0.1), an estimate of 12 inches of ground-water recharge can be made for 1977 (fig. 3). The 12 inches of recharge is probably close to the average annual rate because the 37.0 inches of precipitation that occurred in 1977 is equal to the average annual precipitation rate of approximately 37 inches per year (U.S. Dept. Commerce, NOAA, 1977). This recharge rate of 12 inches per year was used in the model while simulating the ground-water levels measured in December 1977.

Recharge Across Boundaries

An estimate of the western boundary flow was based on an estimate of precipitation infiltration to the ground-water basin west of the modeled area. The basin, approximately 80 mi^2 in area, receives about 4 inches per year of ground-water recharge. Calculations based on these figures show that 15.5 Mgal/d of ground water flows across the western model boundary from this source.



Recharge = (sum of water-level rises) x (assumed specific yield)
= (5.0 + 5.0) ft x (0.1)
= 1 ft of recharge/yr

Figure 3.--Observation-well hydrograph illustrating a calculation of ground-water recharge for 1977. (See fig. 1 for location of FR 109.)

The limestone underlying the basin also contributes to this boundary flow (Norris, 1956). Estimated flow to the aquifer through the limestone (by Darcy's Law) is less than the flow through the overlying glacial sand and gravel at the boundary. Also, the water of poor quality in the limestone apparently does not find its way to the overlying sand and gravel in any substantial quantities.

From these estimates of flow along cross sections perpendicular to the water-level contours (fig. 4), the total maximum flow from the west boundary is estimated to be 25 Mgal/d. To simulate flow toward the Scioto River along the western boundary, constant-head nodes are set at 720 ft of head for all simulations.

The northern and eastern boundaries of the modeled area are much farther away from the well field, so flow from these boundaries is expected to have an extremely small influence on well-field yield. Near the northern boundary, the aquifer pinches out, and the surface is urbanized. For these reasons, little boundary flow is expected from the northern boundary between the Scioto River and Alum Creek. To simulate flow from this border, constant-head nodes ranging from 715 ft of head near the Scioto River to 735 ft farther east, to 715 ft near Alum Creek are used. No-flow nodes make up the rest of the northern boundary.

The eastern boundary is simulated with 730-ft constant-head nodes between Blacklick Creek and Walnut Creek, with no-flow nodes making up the rest of the eastern boundary. A no-flow boundary is used to simulate the southern boundary because all water-level contours are approximately perpendicular to the border.

Total areal recharge to the modeled area is approximately 39 Mgal/d (based on 12 inches of recharge per year). Recharge is expected to be small from the northern, eastern, and southern boundaries, and the maximal western boundary recharge is estimated at 25 Mgal/d. Thus, the total boundary recharge to the aquifer is expected to be equal to or less than areal recharge.

Bedrock Altitude

Two sources of information about bedrock altitude were used: Well logs at the aquifer test sites, and a bedrock altitude map (Schmidt, 1958). When these two sources disagree, the well logs from the aquifer test sites were used as the basis for model input.

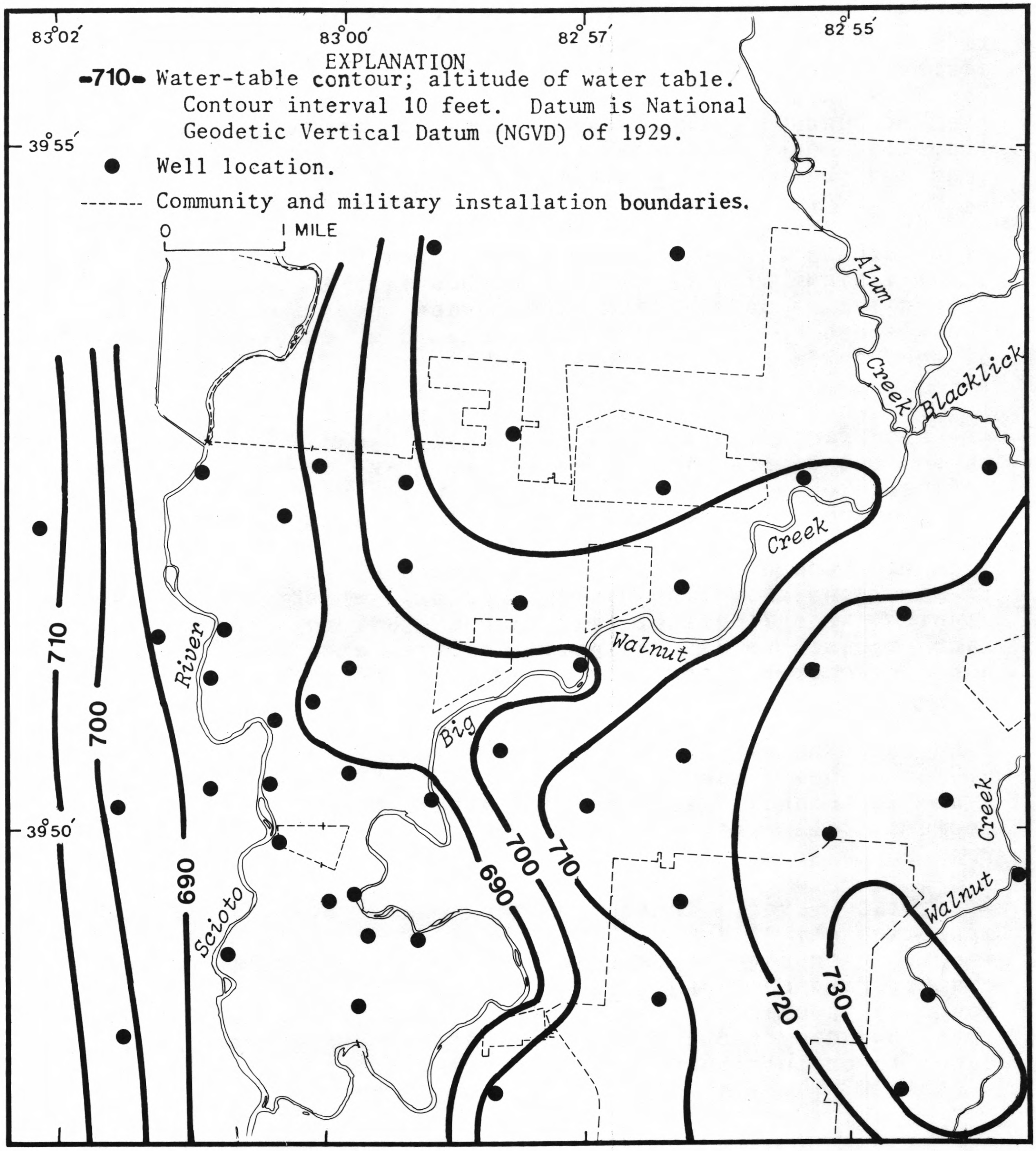


Figure 4.--Contours of December 1977 water-table survey.

Stream Parameters

Stream elevations were taken directly from the ground-water contours shown on the December 1977 ground-water level survey map (fig. 4). A linear interpolation was used between contour crossings to determine stream elevation. A stream depth of 5 ft was used in all simulations for all parts of each stream.

Streambed Parameters

Streambed vertical hydraulic conductivity has been estimated by analytical analyses and simulation of aquifer tests near the streams. The average streambed vertical hydraulic conductivity along the Scioto River at five aquifer test sites was calculated to be 4.36 x $(10)^{-6}$ ft/s (Alden E. Stilson and Associates, 1976, p. 15-2). Simulations of aquifer tests and of nonpumping steady-state conditions at site 101 (fig. 1) were best achieved with a vertical hydraulic conductivity of approximately 3 x $(10)^{-6}$ ft/s, which is similar to the aquifer test results. The vertical hydraulic conductivity of 4.36 x $(10)^{-6}$ ft/s was used in all simulations, along with a variable streambed thickness map. Although a uniform streambed thickness of 1 ft was used to define a streambed leakance (streambed vertical hydraulic conductivity divided by streambed thickness) of 4.36 x 10^{-6} sec^{-1}, a variable streambed thickness was used to weight the area covered by the stream. That is, if the area of the stream was 1 percent of the area of the corresponding node, then the streambed thickness would be increased to 100 ft. Likewise, if the area of the stream was equal to the area of the node, the streambed thickness would be 1 ft. There is not necessarily a one-to-one relationship between the reduction in streambed leakance and the resulting reduction in infiltration because of aquifer head differences. However, because the aquifer head varies little, the existence of the one-to-one relationship can be assumed. Also the area of infiltration is still over the entire area of the finite-difference cell and not just over the area of the stream.

Pumpage in the Modeled Area

Pumpage in the area modeled is small. For 1976 the municipal users and their pumping rates were:

Obetz	0.35
Groveport	0.3
Lockbourne Air Force Base	1.5
Hamilton Meadows	0.27
Total	2.42 Mgal/d

Because areal distribution of discharge rates was simulated with insignificant effects on water-table contours, only the largest rate, that at Lockbourne Air Force Base, was used in all simulations. Also, there is unknown ground-water pumpage from sand and gravel quarries along Big Walnut Creek and these rates have not been simulated.

CHARACTERISTICS OF THE MODEL

The prime value of a model lies in its quality as a simplified abstraction. Features of the hydrologic system that have a weak interaction with more important features are ignored or approximated in the model. Nevertheless, an array of crucial features that have a complicated dependence are accurately related. Within this context, the basic assumptions in the southeast Franklin County aquifer model used to create the simulations are as follows:

(1) The glacial-outwash aquifer is a single unconfined aquifer except beneath the Scioto River, Big Walnut Creek, and Walnut Creek where a semipermeable streambed separates the stream from the mostly confined aquifer below.

(2) Flow in the aquifer is horizontal and two dimensional.

(3) Bedrock beneath the aquifer forms an impermeable boundary.

(4) Recharge to the aquifer is from riverbed leakage (mostly by induced infiltration from pumping), boundary flow, and precipitation that is uniformly distributed.

(5) Ground water is discharged by leakage to rivers, and by pumping from wells.

(6) Evapotranspiration from the ground-water reservoir is considered in the recharge estimate, and is not considered as a separate parameter.

(7) Stream stages remain constant in all streams throughout each simulation.

(8) Water is neither added to nor taken out of storage (specific yield, storage coefficient, and storage of the confining bed have all been set equal to zero).

The above assumptions were used in all simulations in two stages. In the first stage, hydrologic data were defined for the model and trial simulations were made by adjusting the hydraulic conductivity. The resultant potentiometric contours and water budgets were compared to field measurements and estimates to determine the best simulations. In the second stage, each of the best sets of hydrologic characteristics were used to predict the maximal yield of the well field. (Maximal yield of each well is such that by increasing the yield of any well by 1 ft^3/s will cause the well to go dry.)

MODEL CALIBRATION OF AQUIFER HYDRAULIC CONDUCTIVITY

Different hydraulic conductivity distributions were used in attempts to simulate the water levels measured in the modeled area in December 1977 (fig. 4). Areally variable hydraulic conductivity distributions were tried while keeping the areal recharge constant at 12 inches per year. These were the hydraulic conductivity map distribution (fig. 2), a uniform reduction of that distribution by 1/2, called 1/2 K-map, and similarly others called 1/5 K-map, and 1/10 K-map. A comparison of water-level contours generated by these simulations (figs. 5 to 8) with the water levels of the December 1977 (fig. 4) survey showed the 1/5 K-map simulation to be the best overall match. The K-map, and 1/2 K-map simulations produced, in general, hydraulic gradients that are too flat, while the 1/10 K-map simulations produced hydraulic gradients that are steeper than the conditions observed in December 1977. Also, the 730-ft contour of the 1/10 K-map simulation extends half-way across Lockbourne AFB in disagreement with 1977 field measurements. Another unsatisfactory feature of the K-map simulation is the total absence of a 730-ft water-level contour.

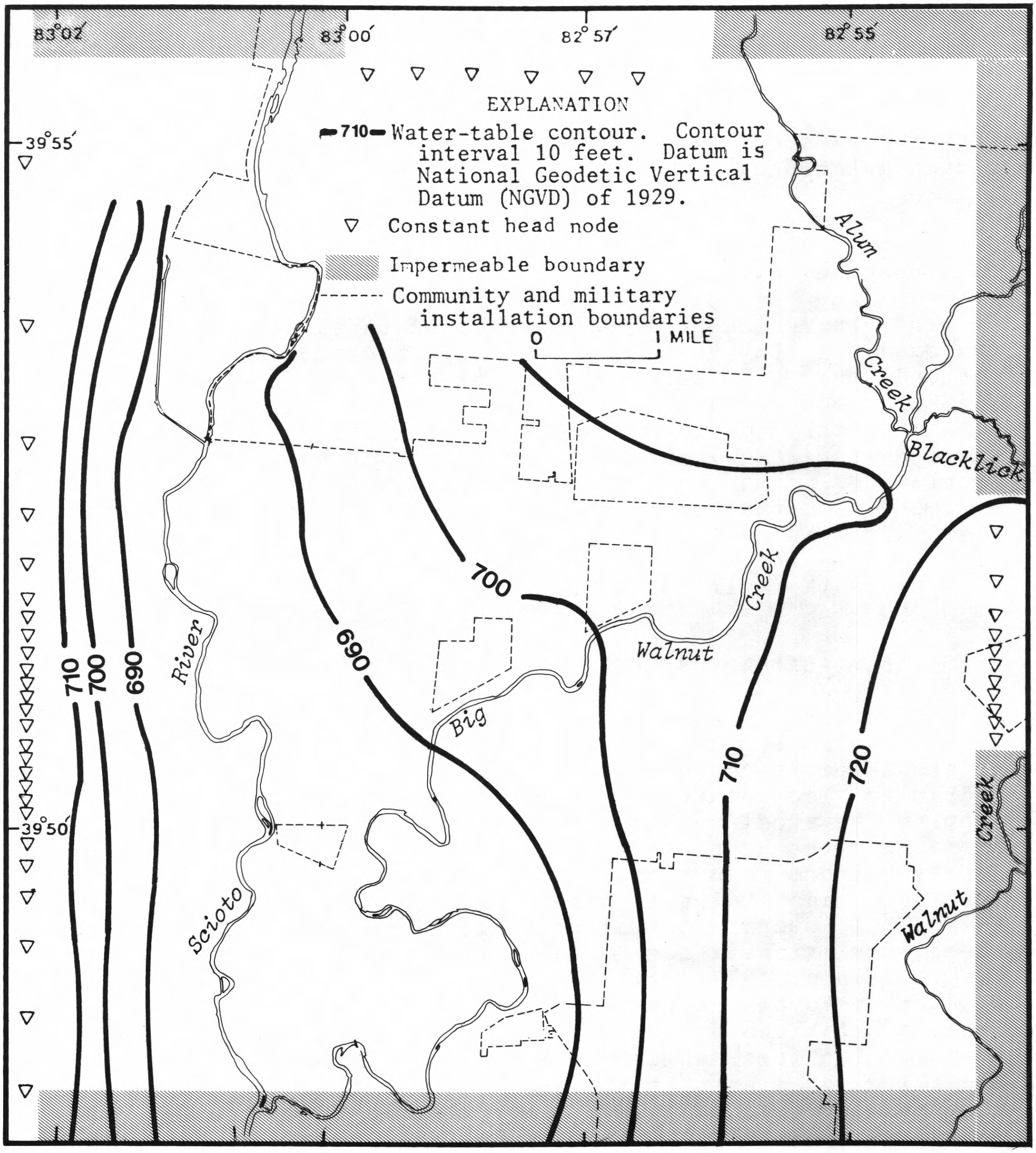


Figure 5.--Simulated contours of water levels generated from an areal distribution of hydraulic conductivity equal to K-map.

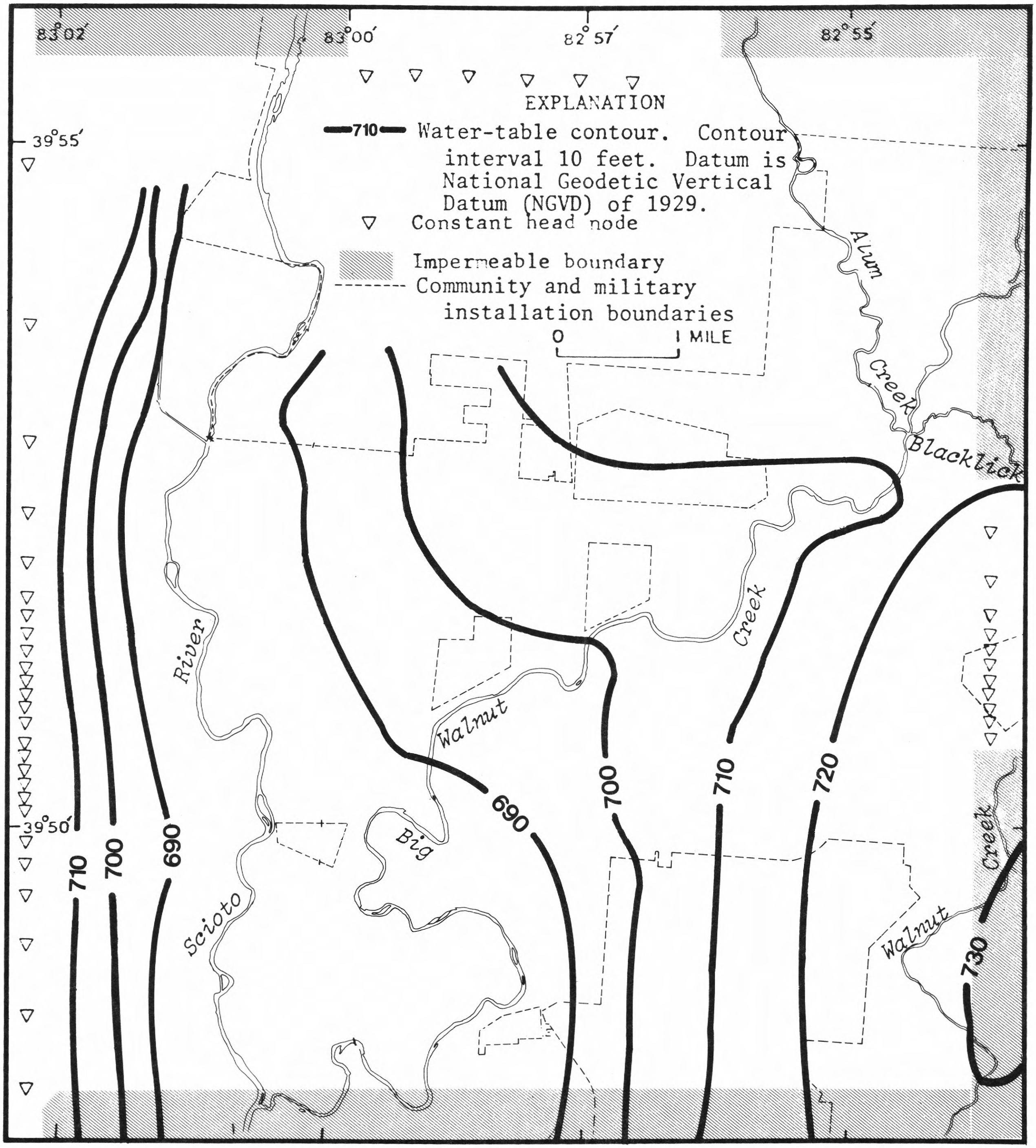


Figure 6.--Simulated contours of water levels generated from an areal distribution of hydraulic conductivity equal to 1/2 K-map.

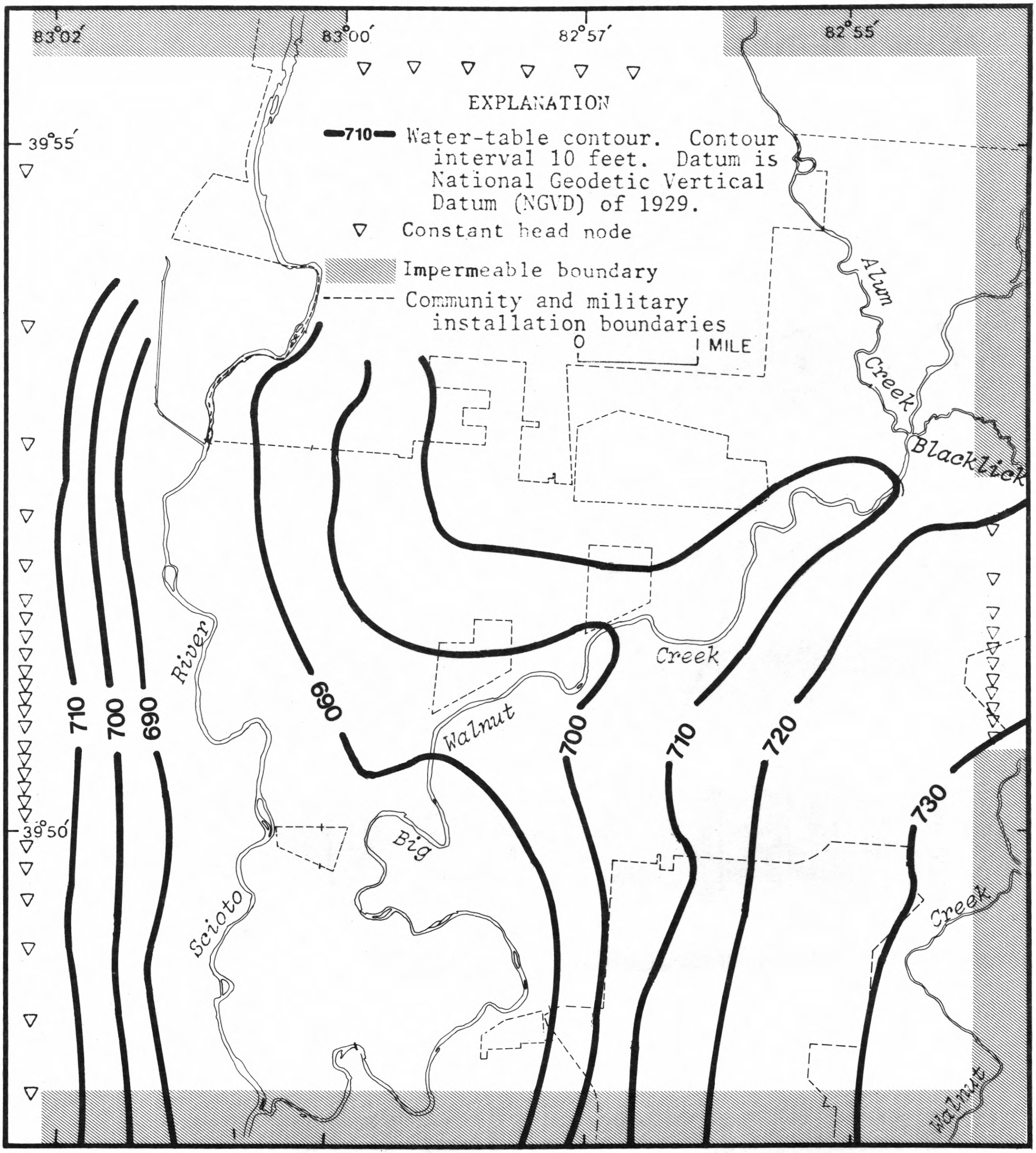


Figure 7.--Simulated contours of water levels generated from an areal distribution of hydraulic conductivity equal to 1/5 K-map.

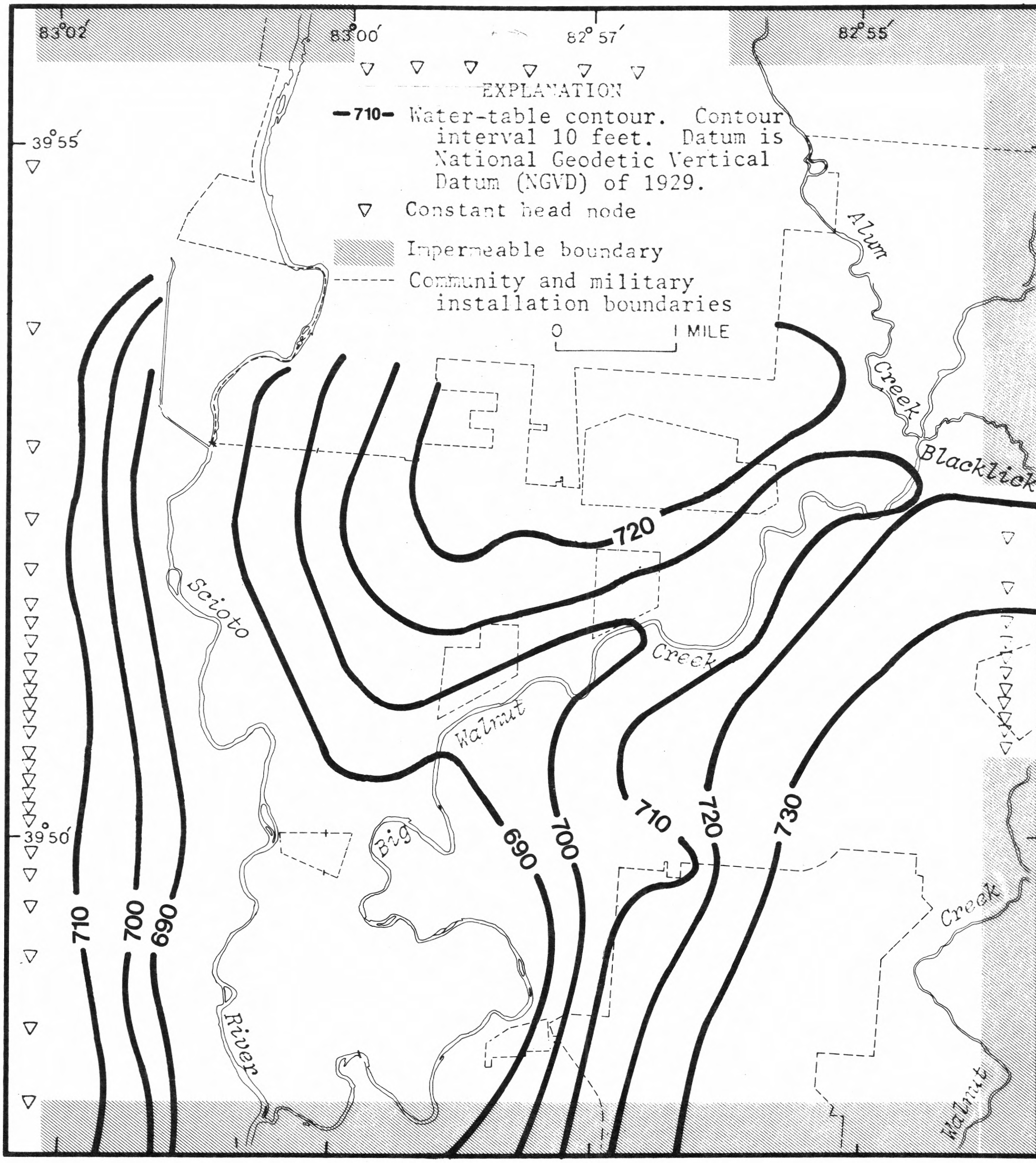


Figure 8.--Simulated contours of water levels generated from an areal distribution of hydraulic conductivity equal to 1/10 K-map.

Uniform distributions of hydraulic conductivity also were used to simulate water-level contours. The uniform distributions have the value of the intermediate hydraulic conductivity zone of 1/2 K-map and 1/5 K-map. That is, hydraulic conductivities of 100 ft/d and 40 ft/d, respectively. The contours generated by these areally uniform hydraulic conductivity simulations are identical to the corresponding contours generated by the 1/2 K-map and 1/5 K-map simulations, except in the area of the well field. In the area of the well field, the 690-ft contour is shifted about 0.1 mile in some places toward, and in some places away from the direction of ground-water flow. The water budgets of the uniform hydraulic conductivity simulations (table 2) are also very similar to the 1/2 K-map and 1/5 K-map water budgets. An examination of net inward boundary flow (table 2) shows that the K-map, 1/2 K-map, and uniform 100 ft/d map simulations exceed the estimate of less than 39 Mgal/d of boundary flow developed earlier. The only simulations satisfactory in all respects are the 1/5 K-map simulation and the corresponding uniform hydraulic conductivity (40 ft/d) simulation.

SIMULATION OF THE WELLS

To simulate the pumping of the well field, well nodes were assigned in the model to well sites 101, 103C, 104, and 115A. Simulated vertical wells have an average estimated radius of 124 ft (table 3). This radius was calculated by averaging the internodal distances between the pumping node and its nearest node on the south and on the north, then dividing by 4.81 (Trescott and others, 1976, p. 9). The irregular grid, the heterogeneous aquifer, and the presence of the nearby stream make an exact simulation of the well radius difficult. The modeled wells are located 270 to 300 ft from the center of the adjacent stream. Internodal distance between a stream node and the adjacent pumping node is 273 ft at sites 101 and 115A and 297ft at sites 103C and 104. To test the sensitivity of well position on maximal yield, wells were moved to the middle of the stream at each test site. This increased maximal yield approximately 10 percent for the uniform hydraulic conductivity simulation with K = 100 ft/d.

Table 2.--Water budget for nonpumping simulations (a positive number indicates recharge to the ground-water system, and a negative number indicates discharge)(ft^3/s).

	1/10 K-map	K = 4- ft/d	1/5 K-map	K = 100 ft/d	1/2 K-map	K-map
Precipitation recharge --------	60.1	60.1	60.1	60.1	60.1	60.1
Pumpage at Lockbourne AFB -----	-2.3	-2.3	-2.3	-2.3	-2.3	-2.3
Net inward boundary flow ------	8.0	22.7	21.9	64.8	62.1	126.0
Inward boundary flow from the west --------------------	2.5	9.2	8.3	27.8	25.2	52.4
Flow from aquifer to streams ---------------------	-65.7	-80.5	-79.7	-122.5	-120.0	-183.9

Table 3.--Yield and drawdown for maximal yield simulations[1].

Site	Saturated thickness (ft)	Aquifer hydraulic conductivity of simulation								Estimated radius of simulated well
		Poor simulations				Good simulations				
		1/2 K-map		100 ft/d		1/5 K-map		40 ft/d		
		Yield (ft^3/s)	Drawdown (ft)	Yield (ft^3/s)	Drawdown (ft)	Yield (ft^3/s)	Drawdown (ft)	Yield (ft^3/s)	Drawdown (ft)	
101	78	14.3	50	8.3	50	6.3	37	3.3	35	147
103C	98	14.4	57	8.4	50	7.4	59	3.4	39	112
104	96	15.4	52	9.4	58	7.4	56	3.4	40	104
115A	130	22.4	92	16.4	77	10.4	70	6.4	65	134
Total		66.5 (43.0 Mgal/d)		42.5 (27.5 Mgal/d)		31.5 (20.4 Mgal/d)		17.2 (11.1 Mgal/d)		Avg 124

[1] Maximal yield of each well is such that by increasing the yield by 1 ft^3/s of any well will cause the well to go dry.

STEADY-STATE MAXIMAL PUMPING SIMULATIONS

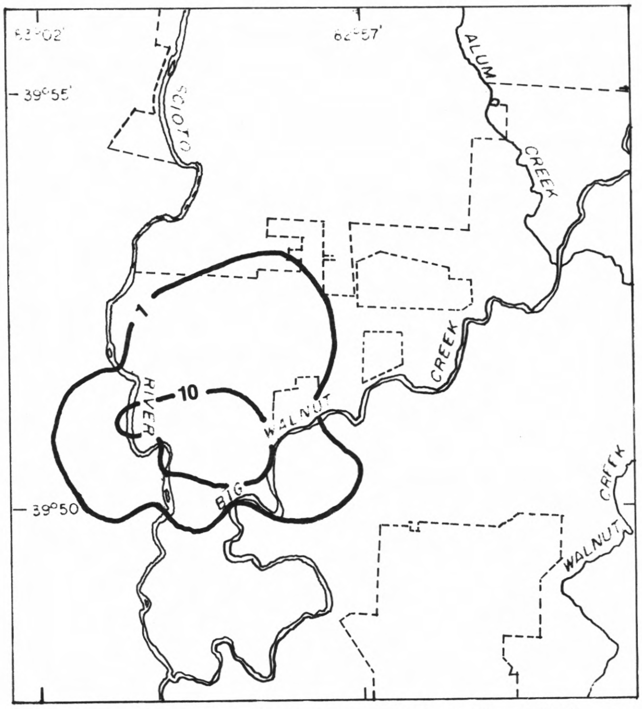
Pumping with maximal yield from the four wells was simulated for four hydraulic conductivity distributions. Each of the simulations started with the initial head distribution of its corresponding steady-state nonpumping simulation and proceeded to a steady-state maximal yield drawdown distribution (fig. 9). For each of these simulations initial aquifer saturated thickness, well yield, and drawdown was calculated (table 3).

The water budgets for each maximal pumping simulation show that approximately 70 percent of well yield is coming from induced infiltration from nearby streams (table 4). A comparison of water budgets for pumping and nonpumping simulations indicates that net boundary flow sometimes increased slightly (up to about 1 percent) because of pumping.

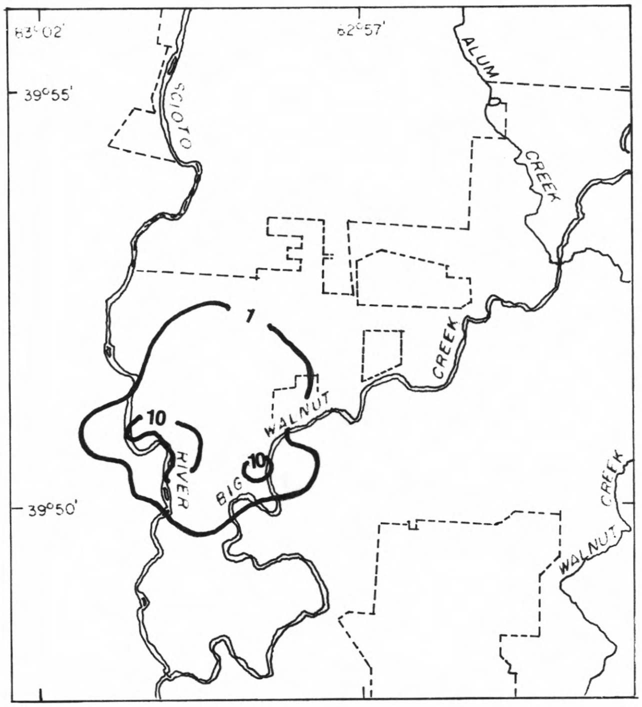
The Effect of Streambed Leakance Changes

Natural scouring or filling of stream channels that occurs with changes in streamflow and sediment load will affect maximal yield from the well field. Sometimes to increase induced infiltration of surface water, stream dredging or the construction of ponds near the well field is tried. All these changes can be modeled.

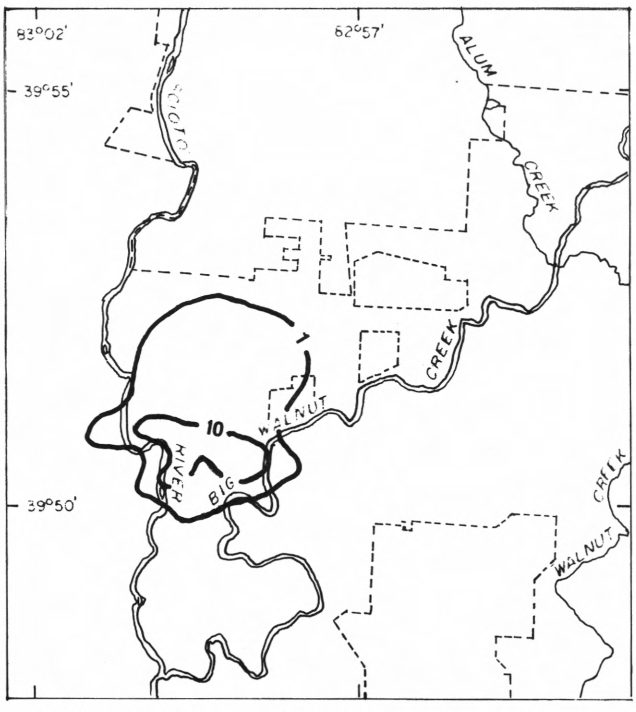
To provide an idea of the effect of a change of this sort on maximal well-field yield, streambed leakance was varied from $4.36 \times 10^{-6} sec^{-1}$ to $10^{-5} sec^{-1}$, and a maximal pumpage simulation was run for the uniform hydraulic conductivity simulation: K = 100 ft/d. Yield increased approximately 13 percent. Physically this change in streambed leakance represents a change in streambed thickness to less than half of its original thickness. For the uniform hydraulic conductivity simulation, K = 40 ft/d, the same change in streambed leakance results in a smaller percentage increase in yield.



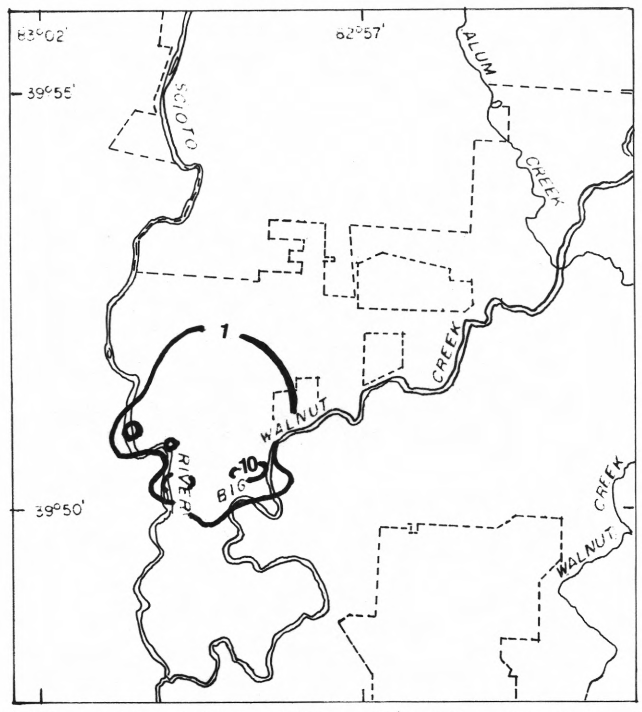
1/2 K - map (yield = 43.2 Mgal/d)



Uniform K - 750 gpd/ft (Yield = 27.6 Mgal/d)



1/5 K - map (Yield = 20.5 Mgal/d)



Uniform K - 300 gpd/ft (Yield = 11.3 Mgal/d)

EXPLANATION

—10— Drawdown contour. Contour **interval** 9 feet.

------- Community and military installation boundaries.

0 1 miles

Figure 9.--Maximum pumpage drawdown contours with several possible areal distributions of hydraulic conductivity (K).

Table 4.--Water budget for pumping simulations (a positive number indicates recharge to the ground water system, and a negative number indicates discharge)(ft^3/s).

	K = 40 ft/d	1/5 K-map	K = 100 ft/d	1/2 K-map
Precipitation recharge ----	60.1	60.1	60.1	60.1
Maximum pumpage of four collectors ---------	-17.2 (-11.1 Mgal/d)	-31.5 (-20.4 Mgal/d)	-42.5 (-27.5 Mgal/d)	-66.5 (-43.0 Mgal/d)
Percent of collector pumpage coming from streams -----------	68.5%	74.7%	72.5%	73.6%
Pumpage at Lockbourne AFB --------------------	-2.3	-2.3	-2.3	-2.3
Net inward boundary flow --	22.7	22.0	65.1	62.8
Inward boundary flow from the west ----------	9.2	8.4	27.9	25.6
Flow from aquifer to streams -------------	-63.1	-48.3	-80.3	-54.1

SUMMARY AND CONCLUSIONS

Initially, available hydrologic data were examined to determine possible values of hydrologic characteristics for input to a ground-water flow model. These data included records of precipitation, well hydrographs, well logs, two ground-water level surveys, and data analyses of six aquifer tests. Values used were drawn from these data to simulate measured December 1977 ground-water levels.

A hydraulic conductivity map (K-map) was constructed for the modeled area from aquifer test analyses and well logs. Considerable variation exists for the areal distribution of hydraulic conductivity but, generally, the highest hydraulic conductivity occurs at and near the aquifer test sites.

Steady-state simulations using the K-map distribution and uniform reductions of that distribution called 1/2 K-map, 1/5 K-map, and 1/10 K-map were tried. Corresponding uniform distributions were also tried and showed only small differences from the simulations using areally varying hydraulic conductivity. Of the distributions tried, hydraulic conductivity that best simulated December 1977 water levels and satisfied all other criteria for a good simulation was the 1/5 K-map distribution, and its corresponding uniform hydraulic conductivity distribution (K = 40 ft/d.

Maximal pumpage was simulated for the two most probable nonpumping steady-state simulations. Maximal steady-state yield for the four wells was 20.5 Mgal/d for the 1/5 K-map simulation, and 11.3 Mgal/d for the corresponding uniform hydraulic conductivity, K = 40 ft/d, simulation. For the simulations tried, these two well-field yields are considered to be the most probable yields. The choice of which one best simulates field conditions depends on whether higher aquifer hydraulic conductivity exists in the well-field area than elsewhere, or whether the aquifer hydraulic conductivity is better approximated by a uniform distribution.

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