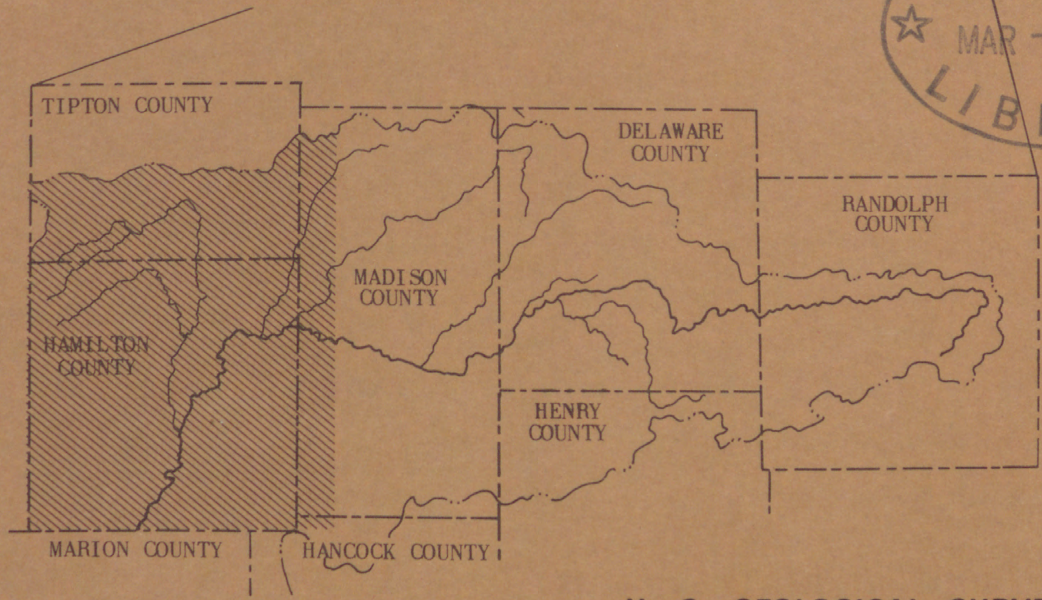
GROUND-WATER RESOURCES OF THE WHITE RIVER BASIN, HAMILTON AND TIPTON COUNTIES, INDIANA



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WATER-RESOURCES INVESTIGATIONS 82-48

Prepared in cooperation with the

INDIANA DEPARTMENT OF NATURAL RESOURCES

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16. Abstract (Limit: 200 words)

The ground-water resources of the White River basin in Hamilton and Tipton Counties were investigated by first collecting driller logs, ground-water levels, streamflow measurements, and pumping data. These data were used to map five confined sand and gravel aquifers interbedded in till, a bedrock aquifer, and an unconfined outwash aquifer. Both two- and three-dimensional ground-water flow models were constructed which provided estimates of the ground-water potential in terms of yield, drawdown and streamflow depletion.

A 39-million-gallons-per-day pumpage was simulated from the outwash aquifer by the two-dimensional model. Pumping was simulated at a rate that reduced the modeled saturated thickness of the aquifer at the wells by half. Pumping in the confined sand and gravel and bedrock aquifers was simulated by the three-dimensional model. Pumpage was simulated over smaller areas than in the outwash aquifer, and pumping plans assessed specific types of conditions, such as extent or thickness of a sand and gravel unit. Simulated pumpages ranged from 1.9 to 6.7 million gallons per day. The highest pumpages were closest to major discharge areas.

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UNITED STATES DEPARTMENT OF THE INTERIOR

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FACTORS FOR CONVERTING INCH-POUND UNITS TO INTERNATIONAL SYSTEM OF UNITS (SI)

Multiply inch-pound unit	By	To obtain SI 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)
	Volume	
acre-foot (acre-ft)	1,233	cubic meter (m^3)
	Flow	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per square mile [$(ft^3/s)/mi^2$]	0.01093	cubic meter per second per square kilometer [$(m^3/s)/km^2$]
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/s)
inch per year (in./yr)	25.40	millimeter per year (mm/yr)
	Transmissivity	
square foot per day (ft^2/d)	0.0929	square meter per day (m^2/d)
	Hydraulic Conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)

DATUM

National Geodetic Vertical Datum of 1929 (NGVD) is a geodetic datum derived from a general adjustment of the first order level nets of both the United States and Canada. It was formerly called "Sea Level Datum of 1929" or "mean sea level" in this series of reports. Although the datum was derived from the average sea level over a period of many years at 26 tide stations along the Atlantic, Gulf of Mexico, and Pacific Coasts, it does not necessarily represent local mean sea level at any particular place.

GROUND-WATER RESOURCES OF THE WHITE RIVER BASIN, HAMILTON AND TIPTON COUNTIES, INDIANA

By Leslie D. Arihood

ABSTRACT

An analysis of the ground-water resources of the Hamilton and Tipton Counties was based on data from about 1,900 well logs, 125 water levels in 125 wells, streamflow measurements at 57 sites, and municipal- and industrial-pumpage records. These data were used to map the flow system and construct a three-dimensional model of the study area and a more detailed two-dimensional model of the outwash deposits along the White River. The models were used to determine the pumpage from several pumping plans and the effect of these pumpages on streamflow and ground-water levels.

Model results indicate that 39 million gallons per day could be obtained from the outwash aquifer. This quantity of pumping in areas of high transmissivity near and parallel to the White River would reduce the saturated thickness of the aquifers at the wells by half. The models also indicated that 0.18 to 6.7 million gallons per day could be pumped from the confined sand and gravel and the bedrock aquifers that have high transmissivities and are near favorable discharge areas.

Drift covers most of the study area and ranges in thickness from 0 to about 400 feet. The drift consists mainly of till and outwash deposits. Beneath the drift lie Silurian and Devonian limestone, dolomite, and shale having a surface relief of about 300 feet.

The study area contains five discontinuous, confined sand and gravel aquifers within the till, an outwash aquifer associated with the White River, and a bedrock aquifer. Of these aquifers, the south half of the outwash aquifer, having a saturated thickness averaging 70 feet and a width ranging from 2 to 3 miles, has the greatest potential for water supply. The general ranges of measurements for the aquifers are: thin and discontinuous confined aquifer, thickness from 5 to 20 feet and transmissivity from 1,000 to 20,000 square feet per day; outwash, transmissivity from 1,000 to 28,000 square feet per day; and bedrock, transmissivity from 500 to 10,000 square feet per day. Vertical hydraulic conductivity of the confining beds between the confined aquifers ranges from 7×10^{-4} to 7×10^{-2} feet per day and averages near 7×10^{-3} feet per day.

INTRODUCTION

Purpose and Scope

In 1972, the U.S. Geological Survey, in cooperation with the Indiana Department of Natural Resources, began a 3-yr study of the ground-water resources of the White River basin in Marion County, Ind. The objectives of that study completed in June 1975 were to (1) determine the quantity of ground water that could be pumped in the county and (2) estimate the effects of this pumpage on the ground-water system and on streamflow (Meyer and others, 1975, p. 2).

After completion of the Marion County study, a similar cooperative study of the rest of the White River basin upstream from Marion County began in July 1975. The objective of this study was to assess the ground-water resources in the White River basin upstream from Marion County. The assessment includes a discussion of the physical and hydraulic characteristics of the aquifers and confining beds and an analysis of the effects of five hypothetical pumping plans on the hydrologic system.

This report describes the ground-water resources in Hamilton and Tipton Counties. It contains maps of the extent, altitude, and thickness of five major sand and gravel aquifers within the till and the altitude of the bedrock surface; transmissivities and hydraulic conductivities of the aquifers and the vertical hydraulic conductivities of confining beds, including the streambeds; ground-water discharges to streams; water levels in the aquifers; and major pumpage from the ground-water system. On the basis of model simulations of five pumping plans, the report indicates the quantity of water that could be pumped without significant adverse effect on the ground-water system and streamflow. The effects are presented in several tables and figures as pumpage, streamflow-depletion rates, and water-level declines.

Methods

Hydrologic data were collected to define the ground-water flow system of Hamilton and Tipton Counties and to simulate the system by two models. Thickness, areal extent, and stratigraphic setting of the major aquifers were mapped from about 1,800 well logs on file with the Indiana Department of Natural Resources, Division of Water. These data were supplemented by data from nearly 80 test holes drilled by the Geological Survey. The hydraulic conductivity of the major aquifers was estimated from specific-capacity data. About 125 ground-water-level and 57 streamflow measurements and a survey of municipal and industrial pumpage helped to define the ground-water flow system and to determine its relation to the streams.

The hydrologic data were also used to construct two digital models of the ground-water flow system. These models were used to estimate the effect of withdrawal on the aquifer system and streamflow.

Location and Setting

The project area, in central and east-central Indiana, covers the 1,500 mi^2 of the basin upstream from Marion County (fig. 1). This area was divided into four study areas, each named for the principal county or counties in these areas. The division simplified studying the project area in detail. The four study areas, by county name from west to east (fig. 1), are Hamilton (but including half of Tipton County), Madison, Delaware, and Randolph. The subject of this report is the Hamilton County study area (fig. 2). The area discussed in this report is further divided into two units. The first unit, about 80 mi^2, includes the White River outwash aquifer in Hamilton County (fig. 3). The second unit, all of Hamilton and half of Tipton Counties, includes the first unit.

Hamilton and Tipton Counties have a temperate climate and an average annual precipitation of 37.11 in. at Noblesville and 35.22 in. at Tipton (National Oceanic and Atmospheric Administration, 1977, p. 4; G. E. Nell, written commun., 1980). The distribution of this precipitation is fairly uniform through the year. Monthly average precipitation for 1941-70 ranged from an average of 2.22 in. during February to 4.18 in. during May (National Oceanic and Atmospheric Administration, 1977, p. 4).

Several streams drain the basin. The White River, the main stream, has an average discharge of 817 ft^3/s (U.S. Geological Survey, 1977, p. 146) and an 80-percent flow duration of 163 ft^3/s (Horner, 1976, p. 232) at Noblesville. Cicero, Fall, Duck, and Stony Creeks drain into the White River.

The study area lies in the Tipton Till Plain physiographic unit (Wayne, 1956, p. 13, 14). The land surface ranges from flat to slightly rolling, and the altitude ranges from 725 ft at the White River on the Marion County line to about 960 ft near Sheridan (U.S. Geological Survey topographic map, Fishers and Sheridan quads). Most surface relief results from the incising of rivers, but a low, broad moraine runs north-south just within the west edge of Hamilton County (Capps, 1910, p. 129).

Hamilton County's 30-percent increase in population from 1970 to 1976 makes it the fastest growing county in Indiana. As of 1976, Carmel had a population of 15,418; Noblesville, 10,499; and Westfield, 2,490. Tipton, Ind., in Tipton County had a population of 4,703. (See U.S. Department of Commerce, 1976.)

Increasing population has affected land use. Land used for growing grain, cattle, and hogs is being converted to urban and suburban developments, particularly in the south part of Hamilton County. Some industrial development is evident also in Hamilton and Tipton Counties, particularly around Noblesville.

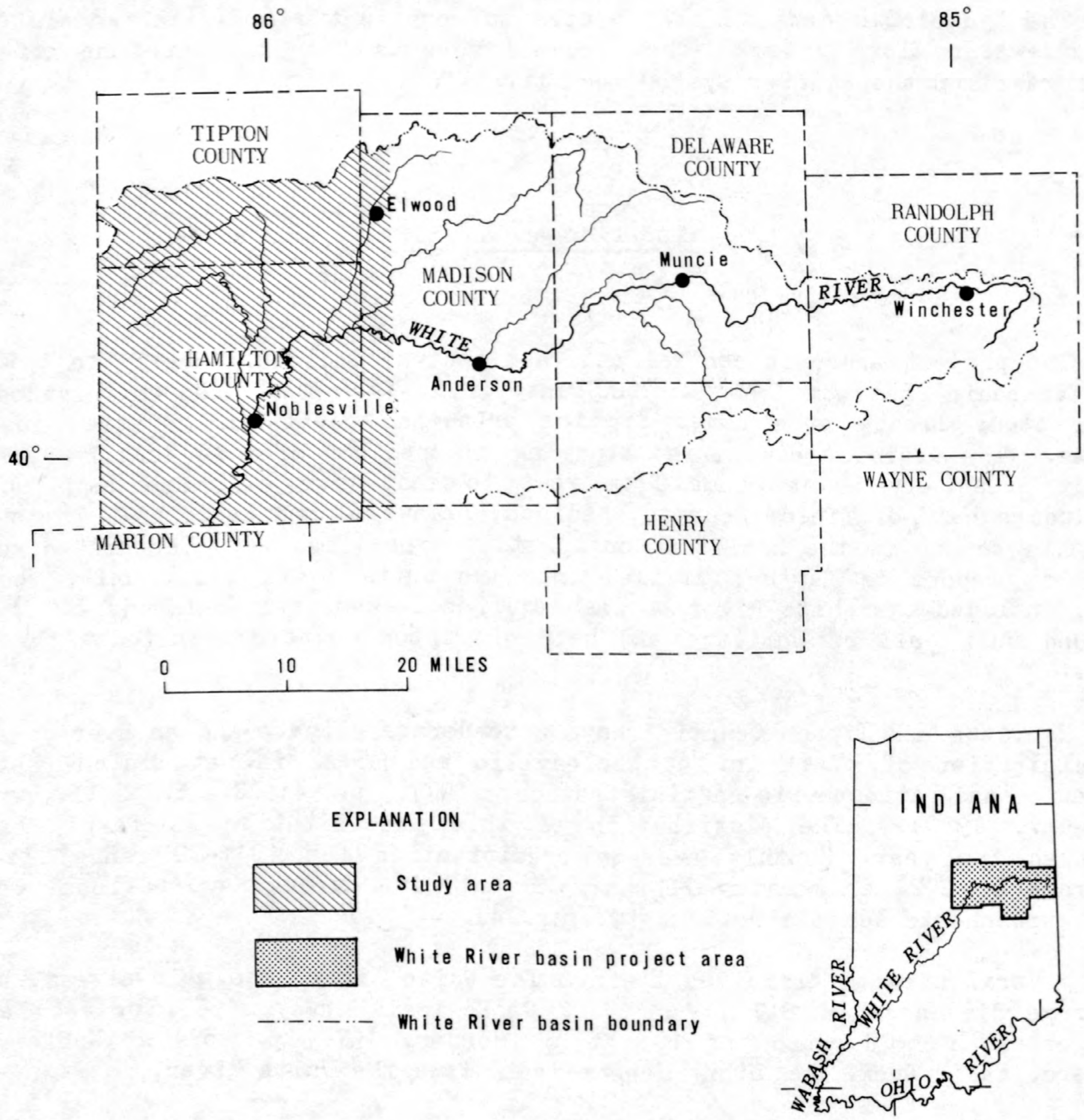


Figure 1.-- Hamilton County study area in the White River basin project.

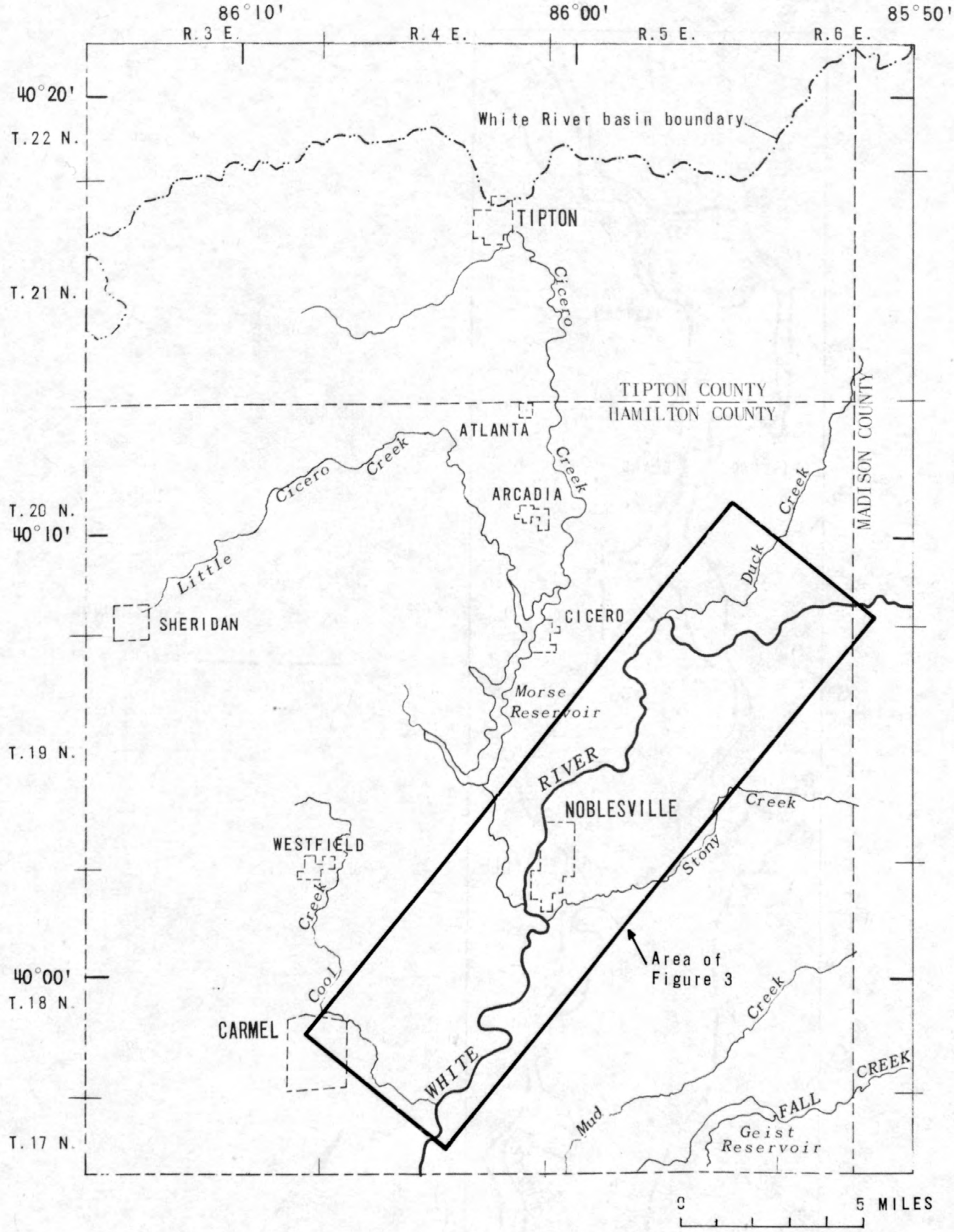


Figure 2.-- Hamilton County study area.

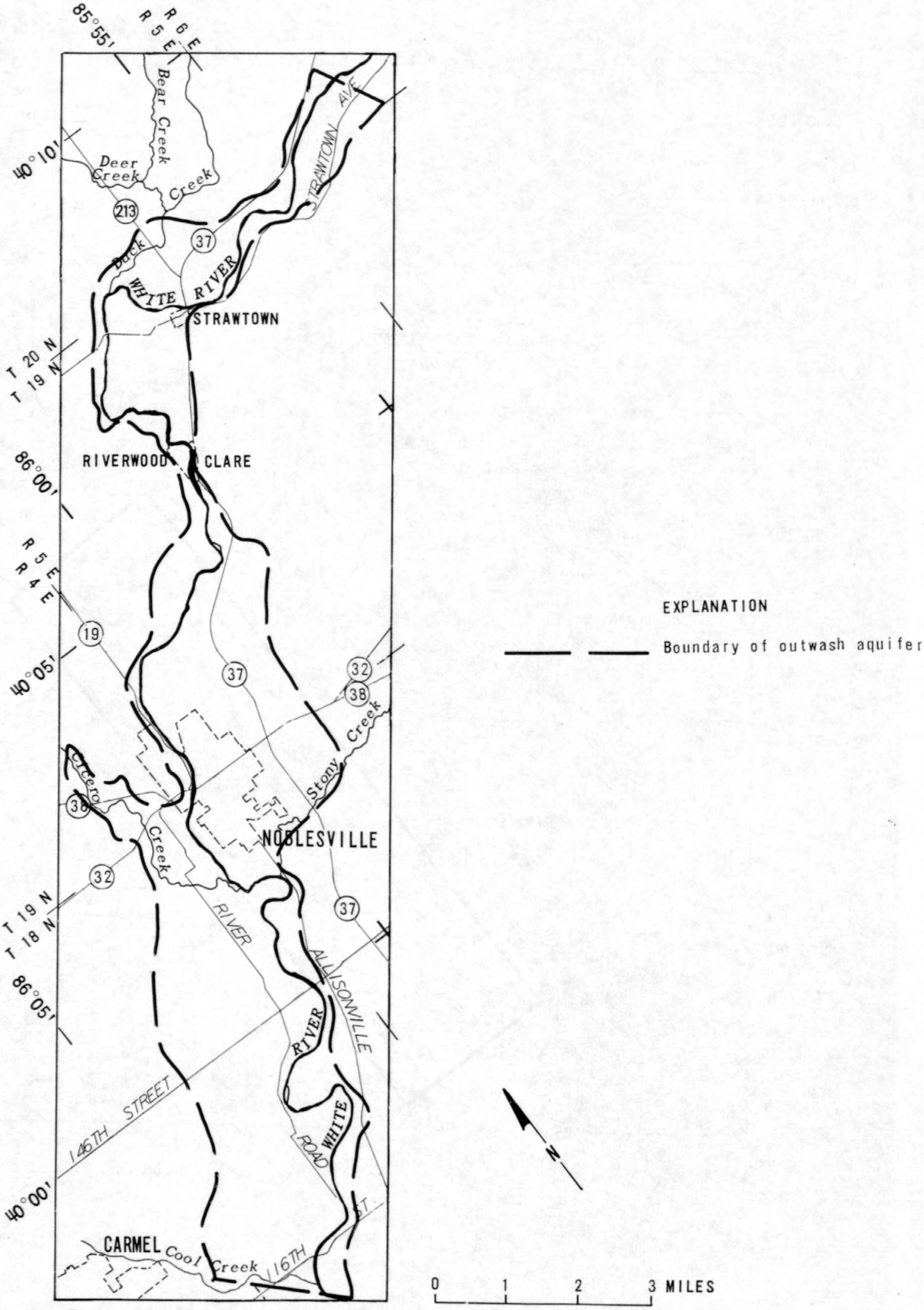


Figure 3.-- The outwash aquifer.

Types of industries in the two counties include metal casting, rubber products, canned foods, agricultural chemicals, and sand, gravel, and limestone quarries.

Previous Investigations

Several studies contain information on the ground-water resources in the general area of this study. Gillies (1976, p. 22) stated that 21.3 Mgal/d of additional water could be withdrawn from the outwash aquifer along the White River from the south edge of Hamilton County to 146th Street. Maclay and Heisel (1972) modeled a single aquifer in all or parts of 12 counties in the upper White River basin, including Hamilton and Tipton Counties. The purpose of their study was to construct a model that could make preliminary estimates of the effects of proposed pumping plans on ground-water levels and on streamflow. In another report, Cable and others (1971) gave qualitative descriptions of the potential yield of water resources in the upper White River basin. They found that the average ground-water discharge in the basin is between 0.6 and 0.8 $(ft^3/s)/mi^2$ (Cable and others, 1971, p. C37). Confined and unconfined aquifers were identified by Cable and others (1971, p. C9), but no specific aquifers were mapped for thickness or areal extent. Meyer and others (1975) determined ground-water availability in Marion County, just south of the study area. Three confined and an outwash aquifer were modeled, but pumpage from the outwash was emphasized in the modeled pumping plans. An additional 57.5 Mgal/d was simulated from three well fields in the outwash by reducing the original saturated thickness at the pumping wells by half. Finally, Herring (1971) and Steen (1968) prepared ground-water availability maps that divide Hamilton and Tipton Counties into areas of maximum well yield.

Acknowledgments

The author is grateful to the Division of Water, Indiana Department of Natural Resources, who provided water-well records from their files, and to the municipalities and industries, who provided pumping information. Appreciation is also expressed to the private well owners in Hamilton and Tipton Counties, who permitted the Geological Survey to measure the water levels in their wells.

GEOLOGY

The study area is underlain by limestone, dolomite, and shale bedrock ranging from Devonian to Silurian in age. Devonian limestone underlies the drift in the southwest half of Hamilton County. Silurian limestone underlies the northeast half of Hamilton County and nearly all of Tipton County (Indiana Geological Survey, 1956). The regional slope of the bedrock surface is from east to west, and local bedrock relief is about 300 ft (fig. 4). Structural dip is southwest (Indiana Geological Survey, 1956). The bedrock Anderson Valley crosses east-west through the north half of the study area. In Tipton and Hamilton Counties, the valley loses its steep-sided, narrow-gorge appearance that it has in counties farther east (Wayne, 1956, p. 38, 39).

The following tables based on Harrell (1935, p. 236 and 457) list the formations and their approximate depths in Hamilton and Tipton Counties.

Hamilton County

Period and epoch	Stratigraphic unit	Thickness (ft)
Quaternary		
Holocene	Chiefly alluvium	0-40
Pleistocene	Glacial deposits	0-352
Devonian	Devonian limestone	40
Silurian	Liston Creek Limestone	Unknown.

Tipton County

Period and epoch	Stratigraphic unit	Thickness (ft)
Quaternary		
Holocene	Sand, clay, and gravel	0-40
Pleistocene	Sand, clay, and gravel	0-150
Devonian	Devonian limestone	0-60
Silurian	Silurian limestone	300

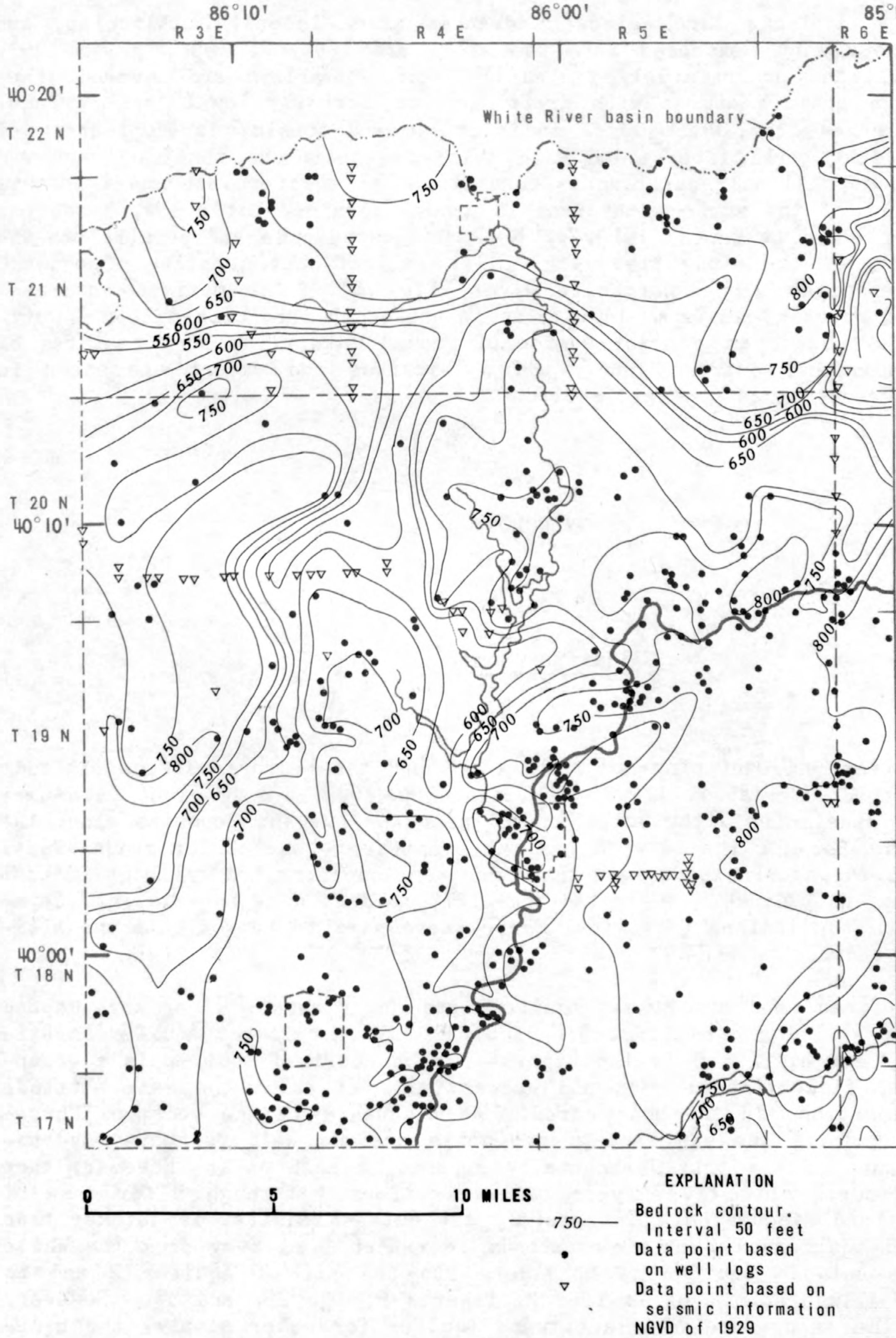


Figure 4.-- Bedrock surface altitude in study area.

Probably at least three glacial advances (the Wisconsin, Illinoian, and probably the Kansan) extended into the study area (Wayne, 1956, p. 16, 49). These glaciations progressively filled the bedrock valleys and covered other bedrock topographic features with drift to form a nearly level land surface. The drift ranges from 0 to 400 ft in thickness and consists predominantly of clayey and silty till interbedded with thin, discontinuous zones of sand and gravel. These sand and gravel zones constitute the aquifers discussed in this report. Most of the surface material is ground moraine, but a low, broad end moraine, mentioned by Capps (1910, p. 129), is just inside and parallel to the west boundary of the study area (fig. 5). A significant quantity of outwash has been deposited along the White River (fig. 6). The maximum saturated thickness of outwash, at least 140 ft, is in the south part of Hamilton County. Overall, the outwash is a major source of ground water. Typical features of bedrock topography, outwash, and aquifer thickness and extent are shown in figure 5, a generalized geologic section.

HYDROLOGY

Aquifer Geometry

The aquifer and confining-bed thicknesses and the bedrock-surface altitude were determined from about 1,900 well logs. However, insufficient data were available for defining lithologies in and near the outwash deposits along the White River. Because the outwash has the greatest potential for ground-water development, the available data in this area were supplemented by about 80 test holes drilled to bedrock. Where bedrock-altitude data were nonexistent, seismic data from the Indiana Geological Survey were used to approximate the altitude of the bedrock surface.

Five confined sand and gravel aquifers and an outwash aquifer were mapped from all the well-log data (fig. 5). Aquifer 1 is the lowest aquifer stratigraphically, and aquifer 5 is the highest. Each confined aquifer is a grouping of thin, discontinuous sand and gravel units at about the same altitude at random locations in the study area. As discussed in the section "Three-Dimensional Model," the sand and gravel units are not well connected hydraulically because of the till horizontally separating each unit. However, they have been grouped into five layers called aquifers 1 through 5 for ease of illustration and discussion. In general, the outwash aquifer is thicker than the confined aquifers but at most extends less than 3 mi away from the White River. The outwash aquifer is considered to be part of aquifer 2 and is included in illustrations of aquifer 2, figures 9, 10, 29, and 35. However, because of the importance of the outwash aquifer for water supply, the aquifer is also discussed and illustrated separately from the confined aquifers in other parts of the report.

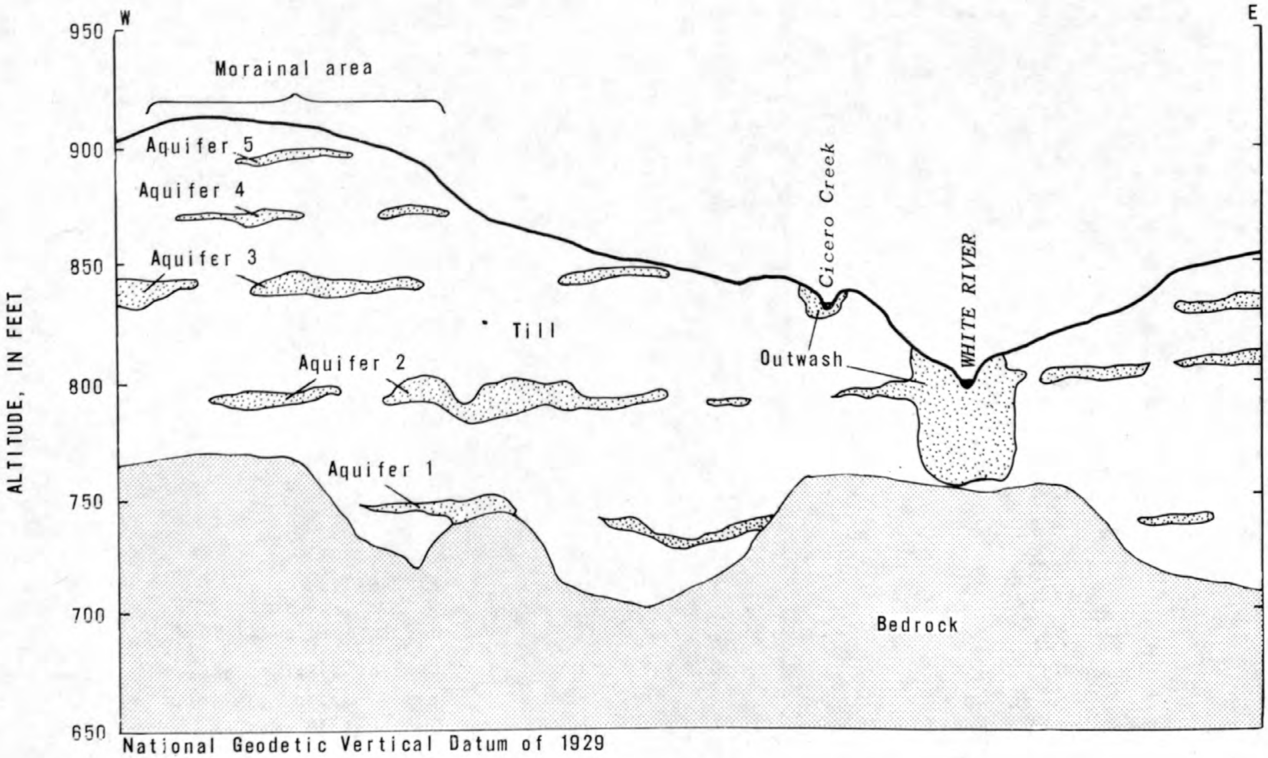


Figure 5.-- Generalized geologic section of the study area.

The saturated thickness of the unconfined outwash aquifer (fig. 6) extends vertically to either bedrock or a clay layer overlying the bedrock. The aquifer is covered by a 5- to 15-ft clay and silt layer and is interbedded with discontinuous 1- to 4-ft-thick clay lenses. In some places, a 1- to 10-ft-thick clay layer separates the aquifer from the bedrock. The areas of unsaturated outwash in figure 6 indicate that the ground-water level is below the bedrock surface and that the outwash above the bedrock surface is dry. The unsaturated outwash could be thought of as nonaquifer areas, or areas that could change because of conditions that cause a different water-table level. In general, the extent and the thickness of the outwash aquifer decreases from south to north. The saturated thickness of the 2- to 3-mi-wide south half of the aquifer averages about 70 ft. The north half of the aquifer thins both vertically and horizontally in places. In the north half, width of the aquifer averages about 1 mile, and saturated thickness generally ranges from 20 to 80 ft.

Thickness, areal extent, and upper-surface altitude of the confined sand and gravel aquifers are shown in figures 7-16. Aquifers 4 and 5 are found in only a low moraine along the west boundary of Hamilton County, whereas aquifer 2, the most extensive, underlies all the study area. Aquifers 1-5 in Hamilton and Tipton Counties correspond to aquifers 1-5 of the Madison County ground-water study (Lapham, 1981) just east of the study area. Range in thickness of

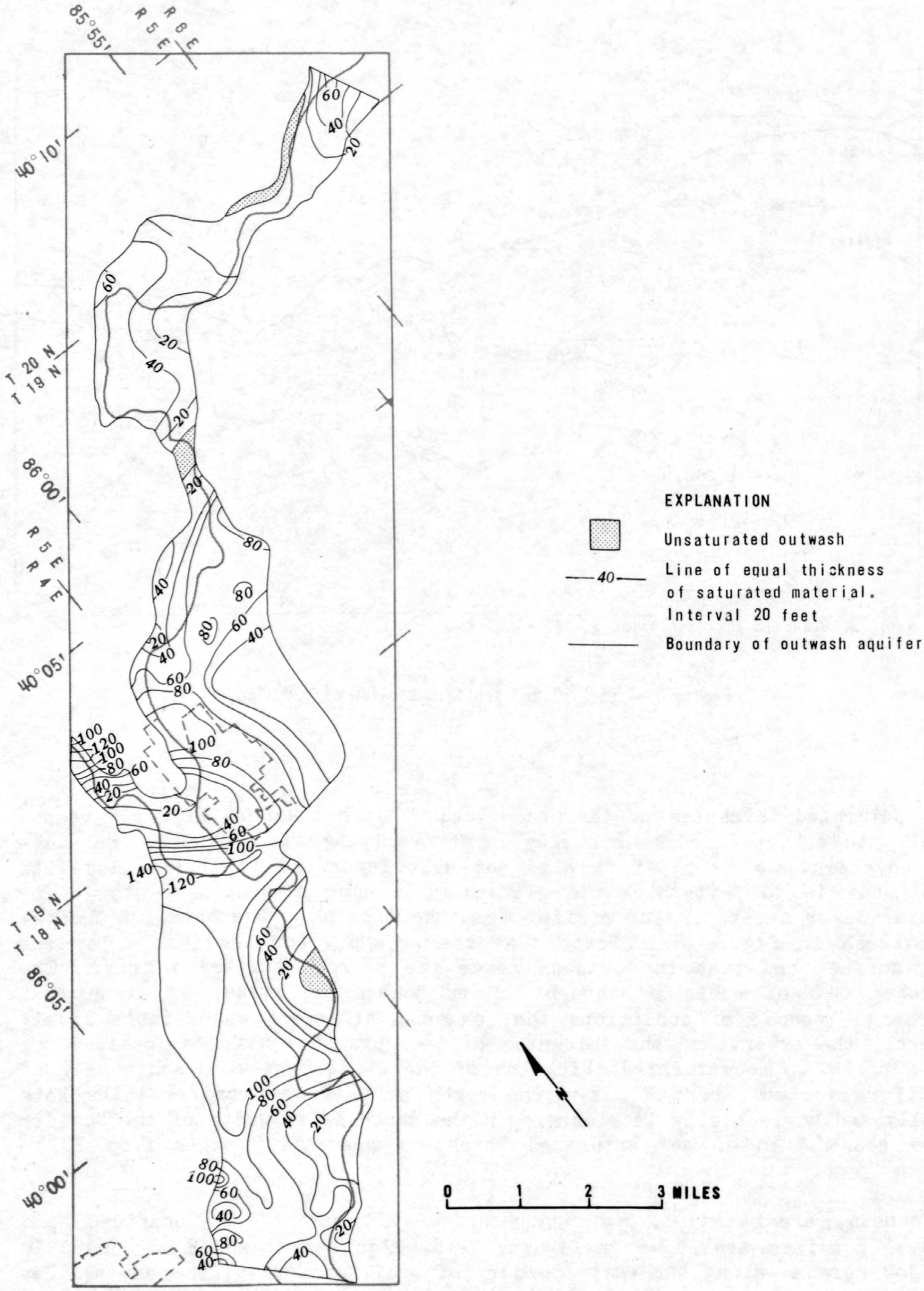
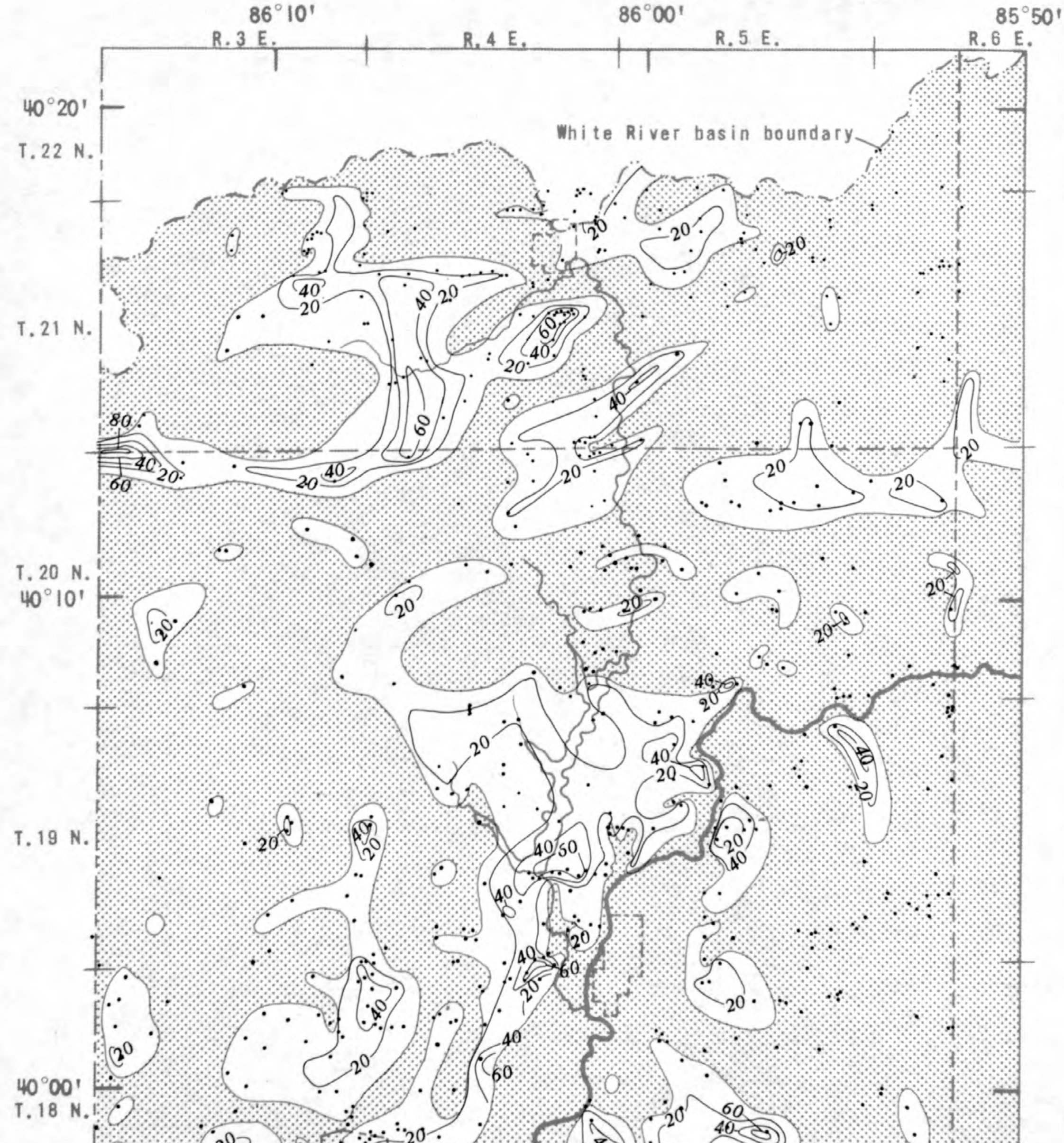


Figure 6.-- Saturated thickness of the outwash aquifer.



EXPLANATION

Aquifer —20— Line of equal thickness of aquifer. Interval 20 feet . Data point

Figure 7.-- Thickness of aquifer 1.

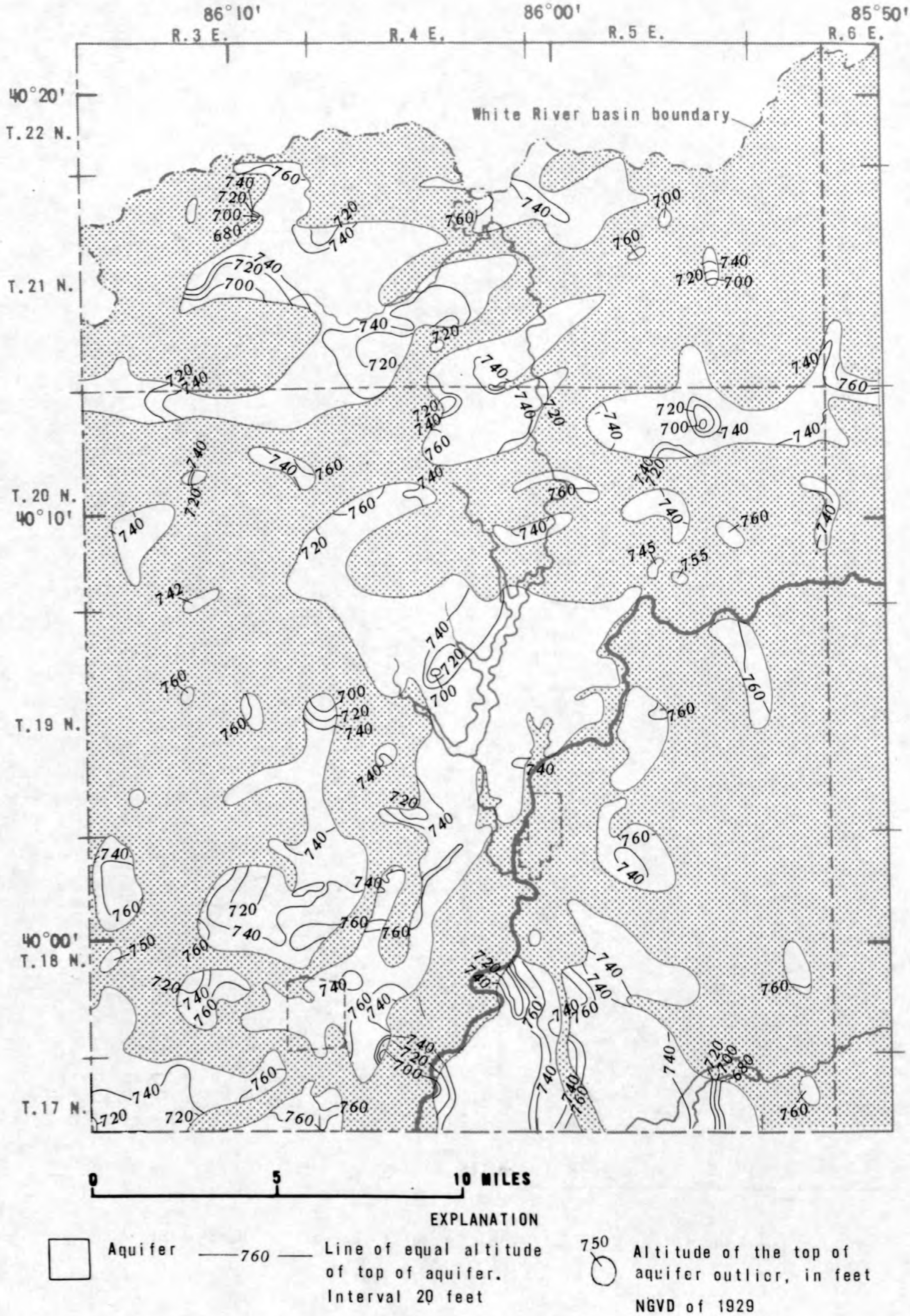


Figure 8.-- Altitude of the top of aquifer 1.

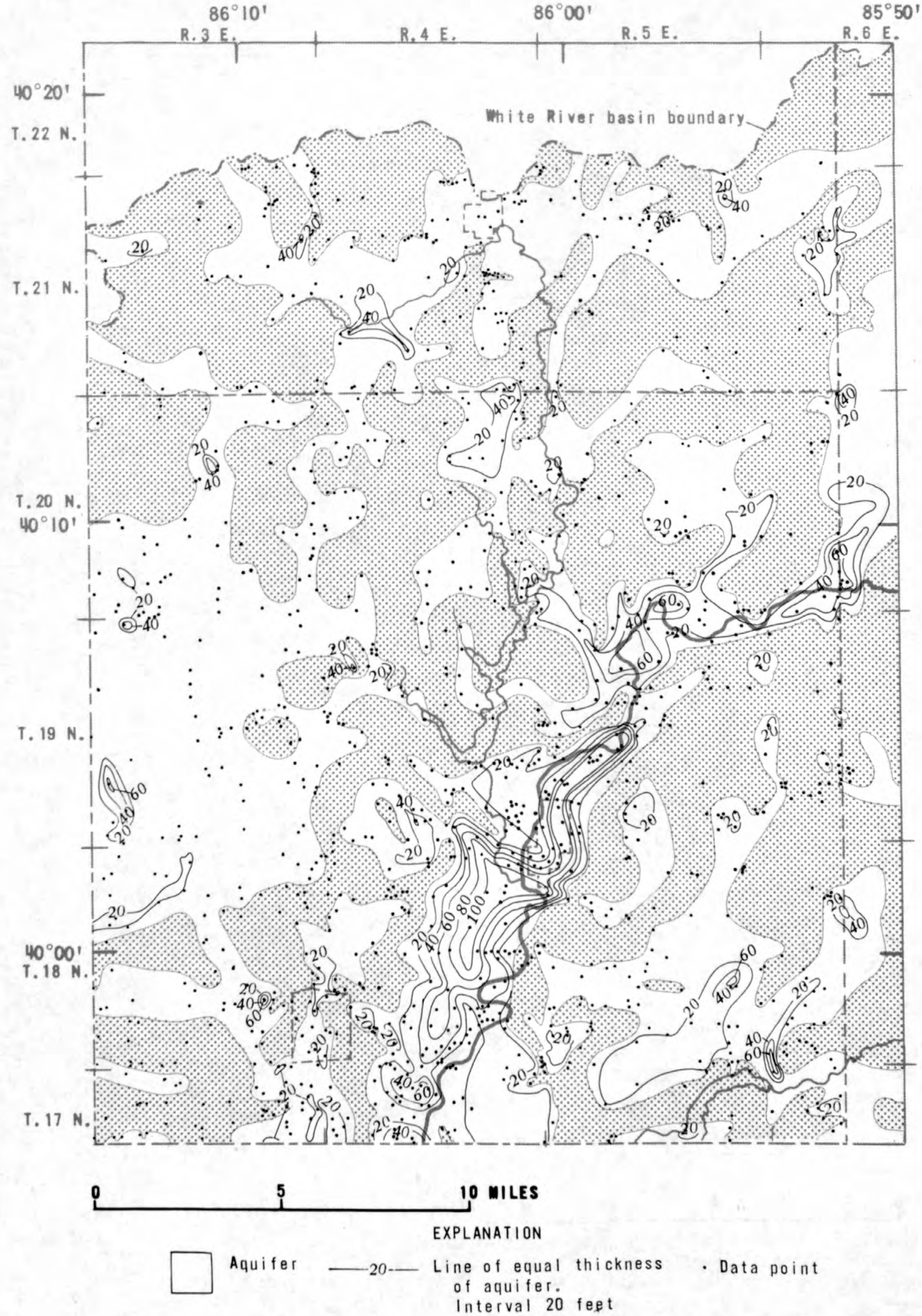
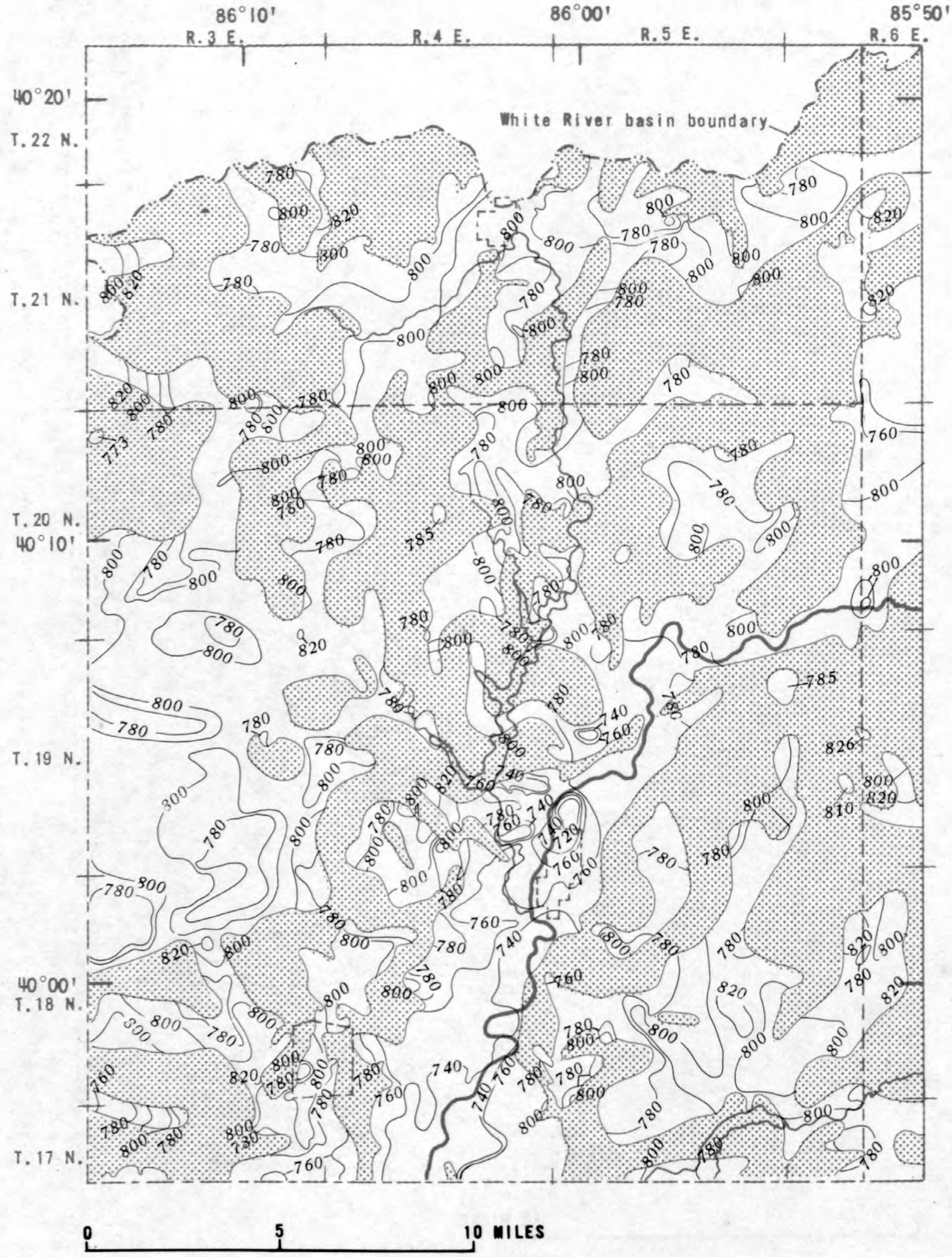


Figure 9.-- Thickness of aquifer 2.



EXPLANATION

Aquifer	—780— Line of equal altitude of top of aquifer. Interval 20 feet	785 Altitude of top of aquifer outlier, in feet

NGVD of 1929

Figure 10.-- Altitude of the top of aquifer 2.

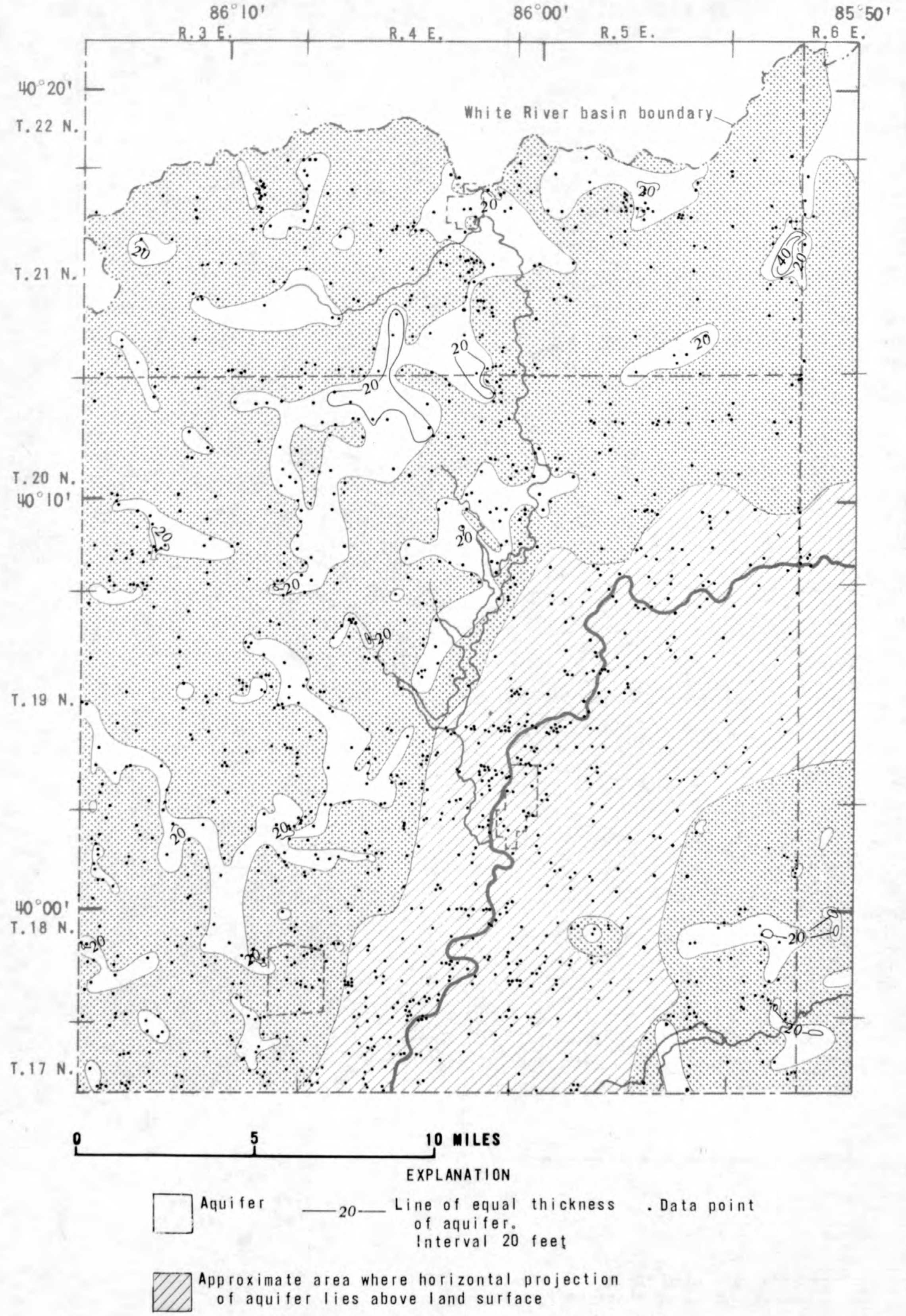
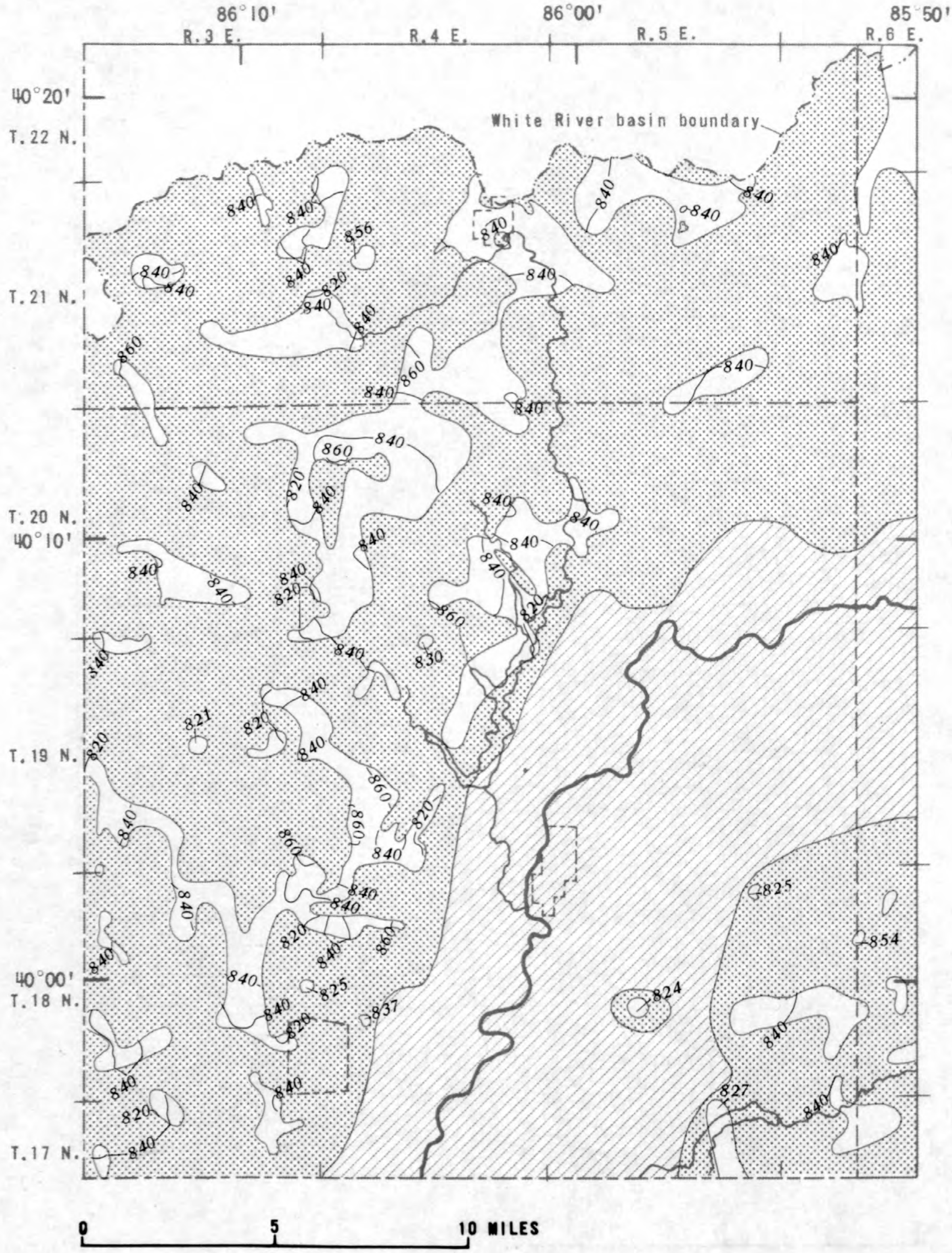


Figure 11.-- Thickness of aquifer 3.



EXPLANATION

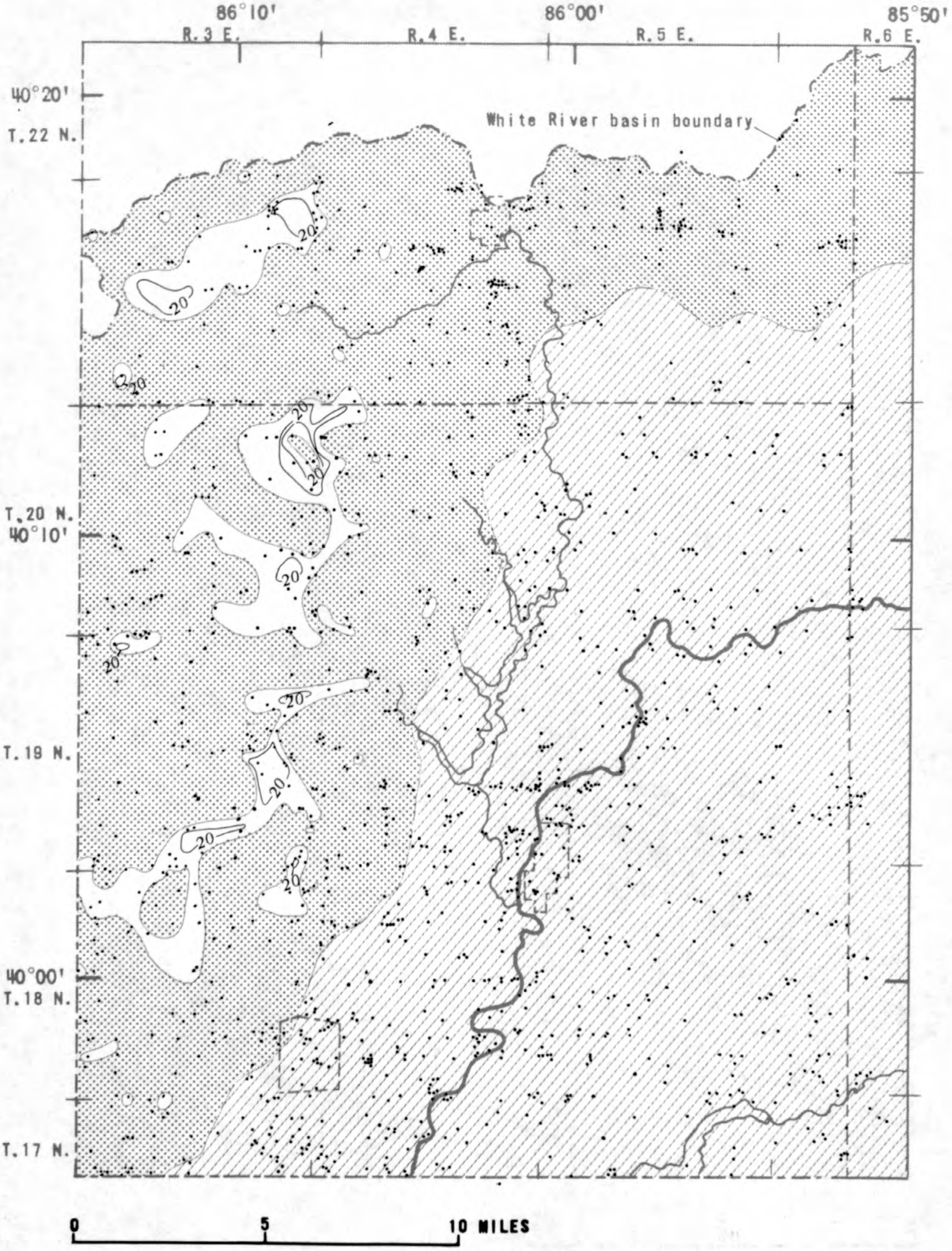
Aquifer

860 Line of equal altitude of top of aquifer. Interval 20 feet

825 Altitude of top of aquifer outlier, in feet

Approximate area where horizontal projection of aquifer lies above land surface

Figure 12.-- Altitude of the top of aquifer 3.



EXPLANATION

Aquifer

—20— Line of equal thickness of aquifer. Interval 20 feet

. Data point

Approximate area where horizontal projection of aquifer lies above land surface

Figure 13.-- Thickness of aquifer 4.

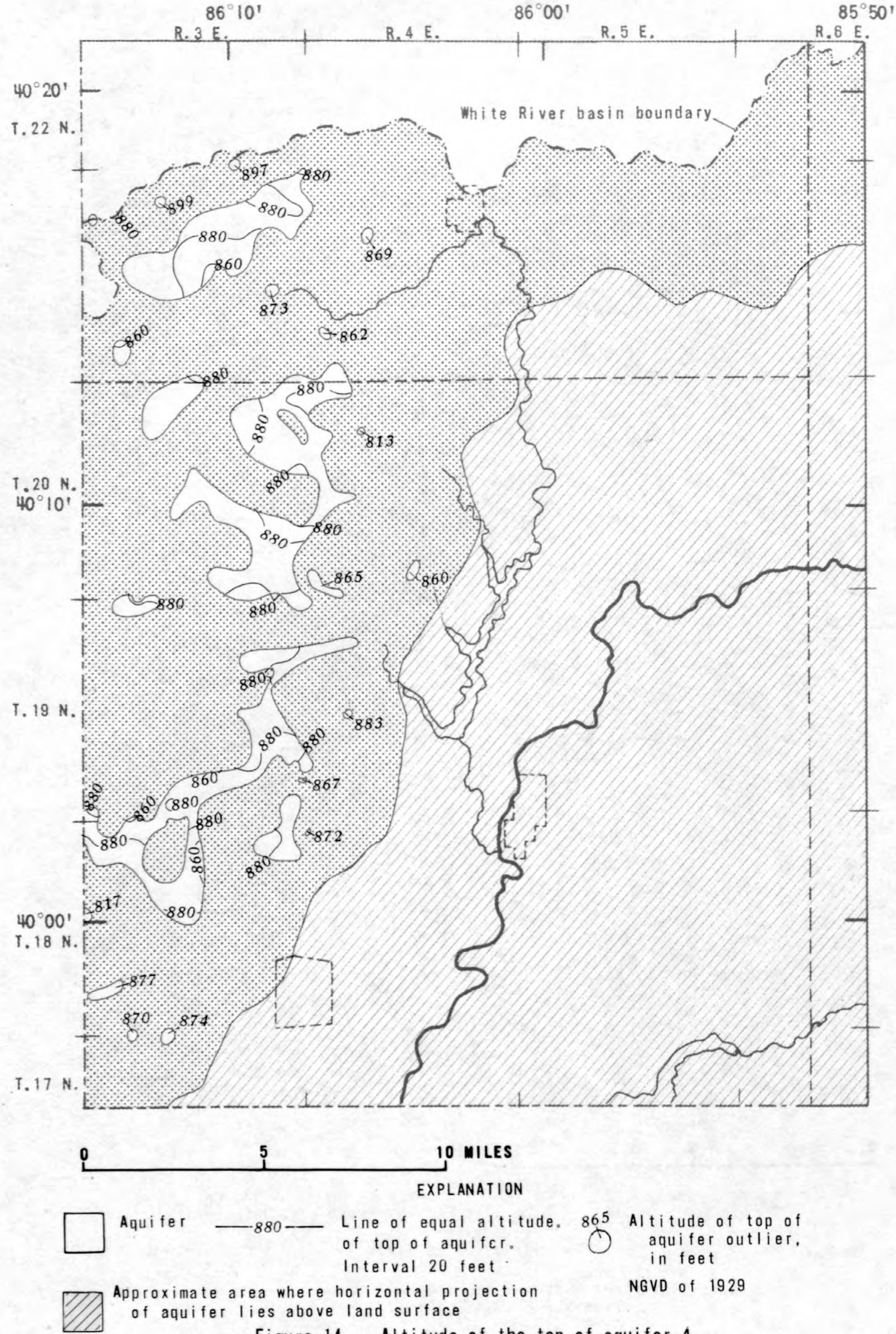
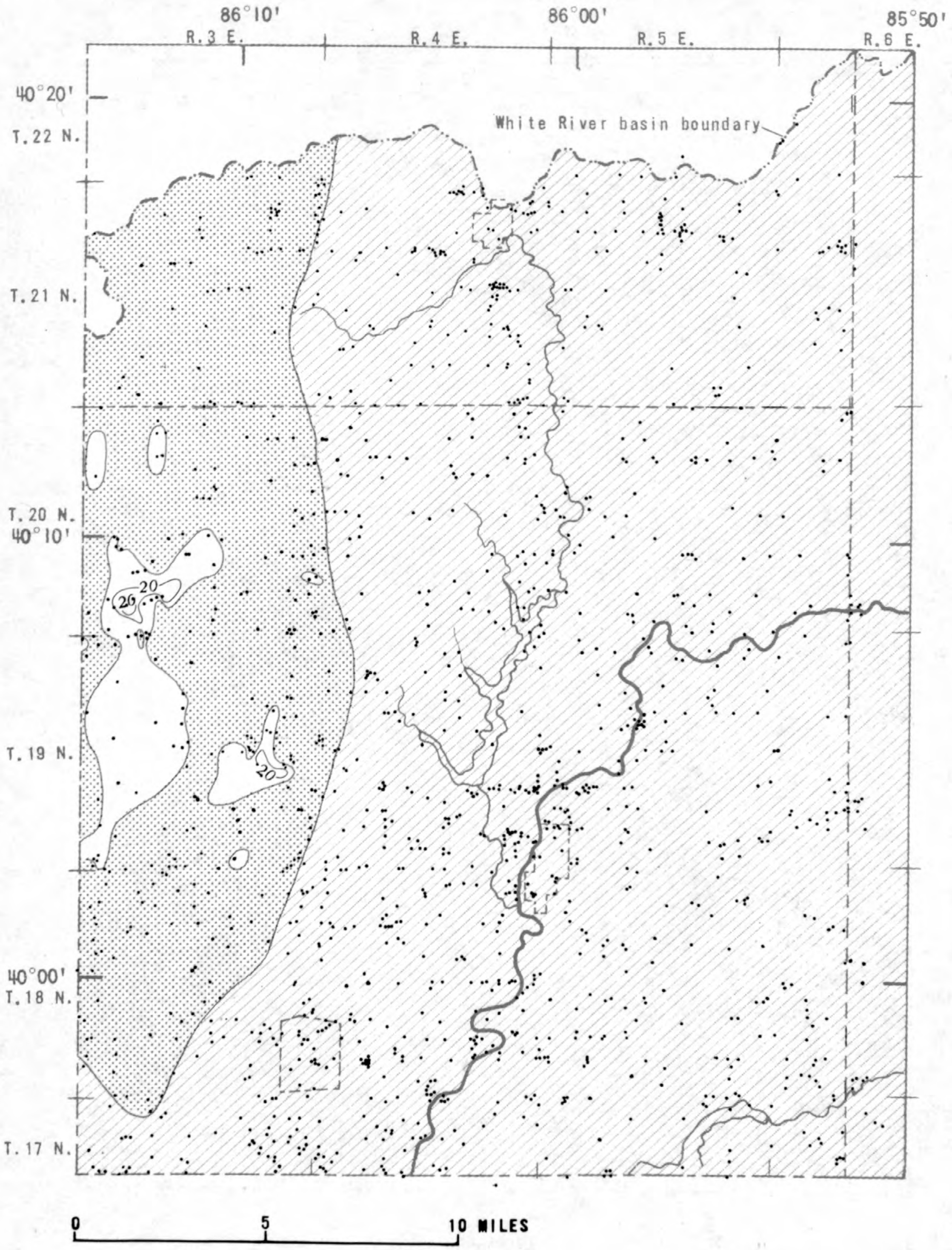


Figure 14.-- Altitude of the top of aquifer 4.



EXPLANATION

Aquifer

—20— Line of equal thickness of aquifer. Interval 20 feet

· Data pcint

Approximate area where horizontal projection of aquifer lies above land surface

Figure 15.-- Thickness of aquifer 5.

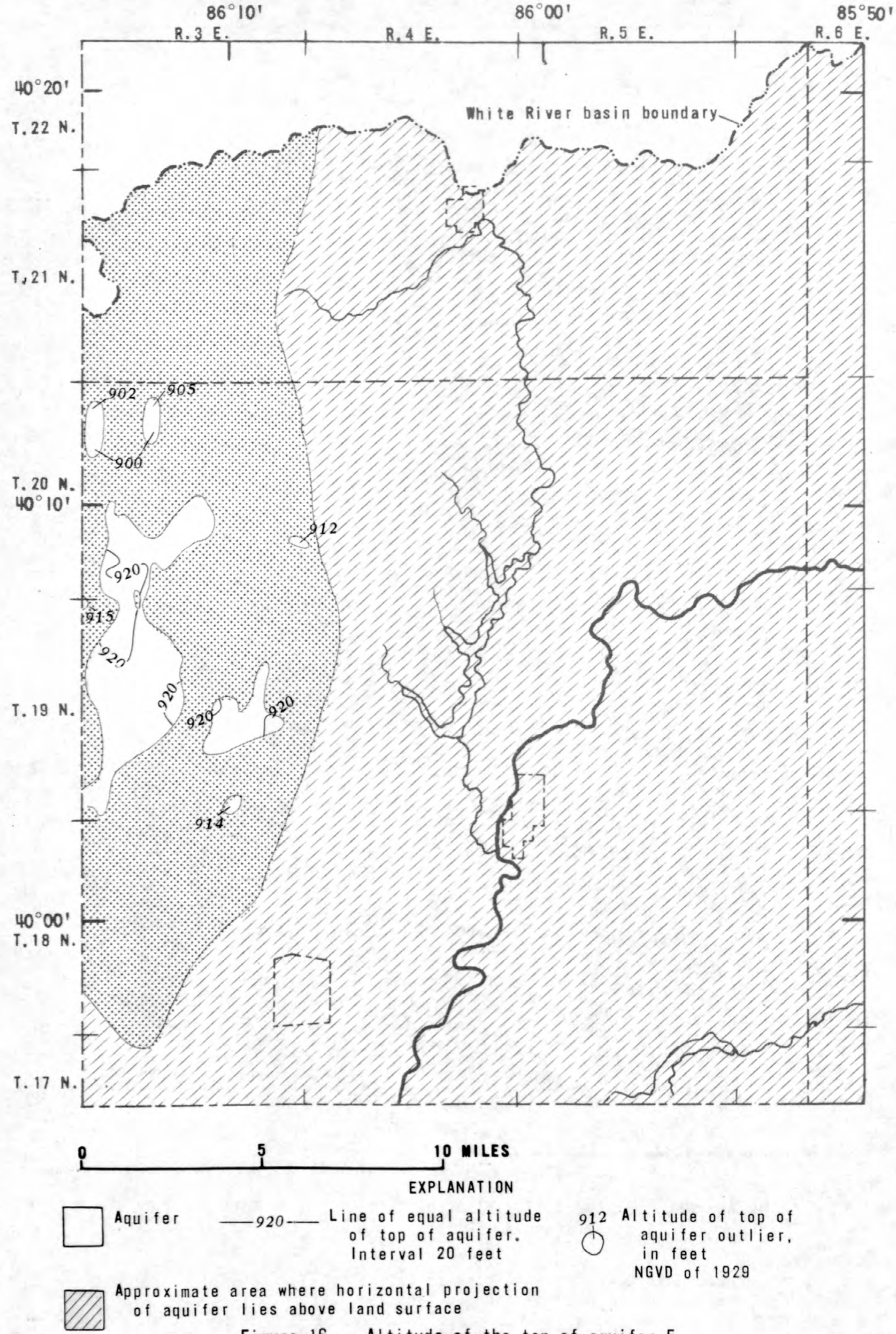


Figure 16.-- Altitude of the top of aquifer 5.

the aquifers is generally from 5 to 20 ft, but in small areas outside the outwash aquifer the thickness may be as much as 90 ft. Though not evident in the maps, some vertically adjacent aquifers join to form a sand and gravel unit that is thicker than either adjacent aquifer shown in the maps.

The areal extent of some sand and gravel aquifers is too small to be mapped. Scattered, insignificant lenses of sand and gravel between aquifers 1 through 3 were neither mapped nor modeled. Also, because only a few wells extend to bedrock in some areas, the extent and the thickness of aquifer material below aquifer 1 is not well known, particularly in buried bedrock valleys. No test holes were drilled beyond 1 mile from the White River outwash area. Although no potentially significant buried-valley aquifer was indicated by available data, buried-bedrock valley aquifers in Madison and Delaware Counties were investigated to determine their water potential.

Bedrock is a potentially significant aquifer for municipal, industrial, and domestic water supplies in areas where sand and gravel aquifers are absent. Although Cable and others (1971, p. C10) and Meyer and others (1975, p. 17) assumed an average permeable thickness of 100 ft for the bedrock, 150 ft was assumed in the Hamilton County study area because many wells penetrate as much as 150 ft of bedrock. However, thickness of the bedrock aquifer may exceed 150 ft. The consequences of assuming an average 150-ft thickness are discussed in the section "Bedrock Aquifer."

Hydraulic Characteristics of the Ground-Water Flow System

Three properties of the aquifers and confining beds are discussed here: hydraulic conductivity, vertical hydraulic conductivity, and transmissivity. Another common property discussed in model studies, storage coefficient, was not used in this study because all model runs simulated steady state. However, the storage coefficients for aquifer types in and near the study area are given in the table that follows.

Aquifer type	Storage coefficient	Source
Outwash	[1]0.11	Gillies (1976, p. 10).
Confined sand and gravel	[2]$3.3x10^{-6}$	Meyer and others (1975, p. 25).
Bedrock	$1x10^{-4}$	Do.

[1]Specific yield.
[2]Specific storage.

Outwash Aquifer

The hydraulic conductivity of the outwash aquifer was determined by calculating weighted averages of hydraulic conductivity at each well site. Weighted averages were used because the Geological Survey's borings by power auger indicated as many as three types of aquifer material at a site. The conductivities used were 40 ft/d for sand, 240 ft/d for sand and gravel, and 415 ft/d for gravel, from Meyer and others (1975, p. 18). Clay was not included in the calculation of conductivity because its contribution is small (about 0.1 ft/d) relative to other materials.

The transmissivity of the outwash aquifer (fig. 17) is the product of its saturated thickness and its average hydraulic conductivity. Outwash transmissivity generally ranged from 1,000 to 28,000 ft^2/d.

Confined Sand and Gravel Aquifers

Data for calculating hydraulic conductivity were available from Geological Survey specific-capacity data and results from calibration of the ground-water flow model. Conductivity was first determined from 13 specific-capacity tests in Madison County (Lapham, 1981, p. 17). Initially, transmissivities of the aquifers were calculated from the specific-capacity data at each of the 13 sites by a method described by Brown (1963). Because only the bottom 3 ft of the aquifer in each well was screened, the calculations were adjusted for partial penetration by the method described by Butler (1957, p. 159-160). In the adjustment, the ratio of horizontal to vertical hydraulic conductivity of the sand and gravel aquifers was assumed to be 10:1 on the basis of work by Meyer and others (1975, p. 19). Hydraulic conductivities were then calculated by dividing the transmissivities by the aquifer thickness.

The average of the 13 tests, 433 ft/d, was used in model analysis as an initial confined-aquifer conductivity. During model calibration, however, the conductivity was reduced to 216 ft/d. This reduction produced closer agreement between measured and calculated values of water levels and seepage to streams than the higher value. More details on this reduction are discussed in the section "Three-Dimensional Model." The 216 ft/d is probably the most accurate average hydraulic conductivity for the five aquifers.

Although the first hydraulic conductivity was higher than the final calibrated value, 216 ft/d was within the range of conductivity available from other sources. For example, Lapham's average conductivity from the 13 specific-capacity tests is 433 ft/d, but the range of conductivities that he calculated is from 24 to 1,600 ft/d (Lapham, 1981, p. 30). Maclay and Heisel (1972, p. 5) used about 200 ft/d in their analog model of the upper White River basin, which includes the project area. Meyer and others (1975 p. 21) used

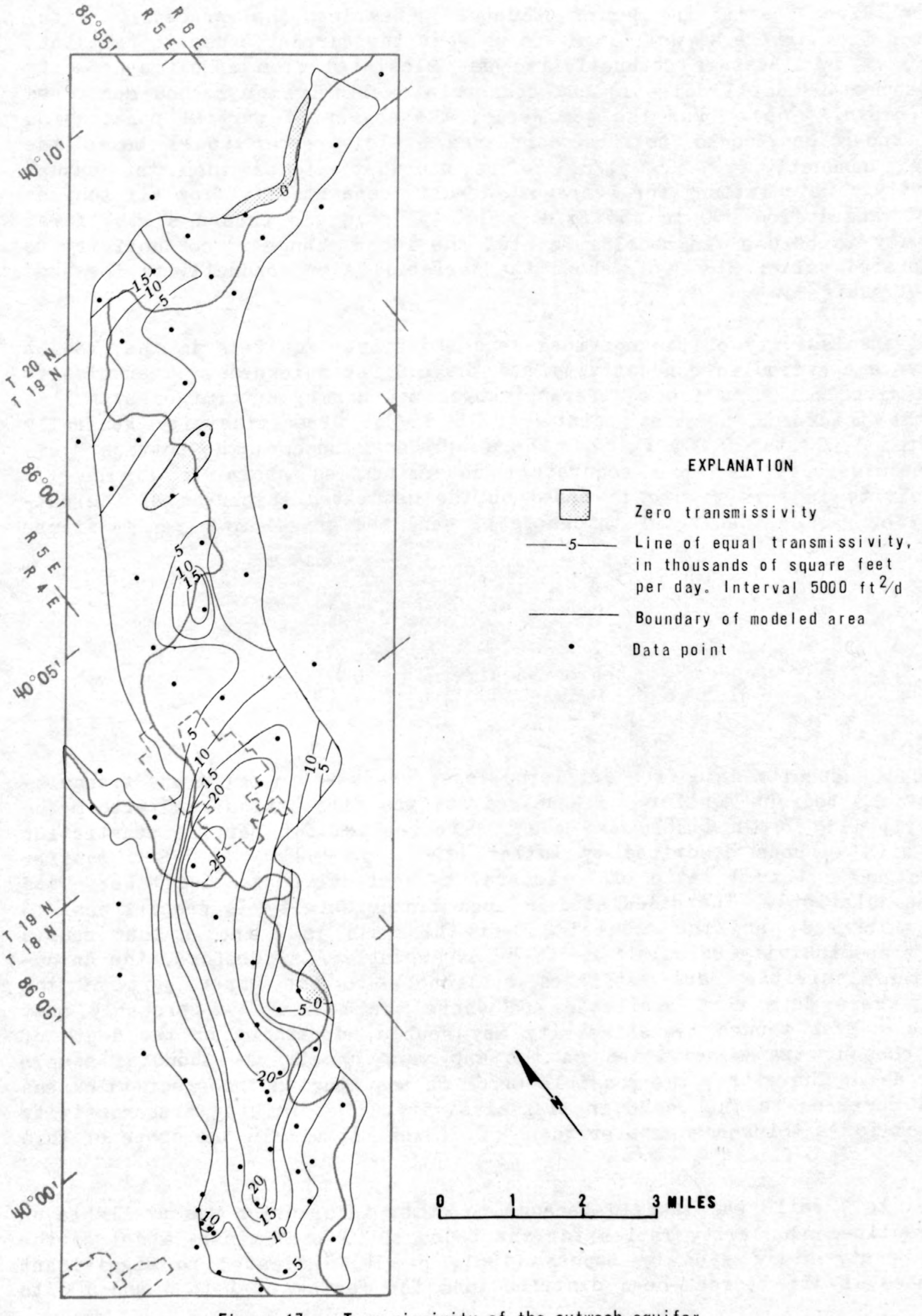


Figure 17.-- Transmissivity of the outwash aquifer.

their calibrated value, 433 ft/d, as an average conductivity for aquifers in the White River basin in Marion County. Besides the preceding data, information from drillers' well logs was used in the current study to calculate conductivity. An average conductivity was calculated from specific-capacity data of about 400 small domestic and commercial wells by the method mentioned earlier (Brown, 1963). For the 400 wells, the degree of partial penetration was not known and could not be corrected. This error would cause the calculated conductivity, 305 ft/d, to be somewhat higher than the actual conductivity. Calculations for average hydraulic conductivity from all sources generally ranged from 200 to 433 ft/d. The 433 ft/d was chosen as the first conductivity to be used in model runs, but the later change in conductivity to the calibrated value, 216 ft/d, shows the variability of conductivity for sand and gravel aquifers.

The transmissivity of the confined sand and gravel aquifers is the product of the average hydraulic conductivity and the aquifer thickness. Transmissivity at a given point on the aquifer-thickness map can be determined by multiplying the thickness at that point by 216 ft/d. Transmissivity generally ranged from 1,000 to 20,000 ft^2/d. Where aquifer 2 becomes the outwash aquifer, transmissivity is more accurately determined as shown in figure 17. Transmissivity in this figure is based on the saturated thickness of the outwash aquifer rather than total thickness of sand and gravel as given in figure 9.

Bedrock Aquifer

Specific-capacity data from drillers' logs were used to calculate transmissivity of the bedrock aquifer. Transmissivity was first calculated with methods described by Brown (1963) and then was corrected for partial penetration effects with methods described by Butler (1957, p. 159). A 150-ft aquifer thickness and a 1 to 1 ratio of horizontal to vertical conductivity were used in the calculations. The calculated bedrock transmissivity is proportional to aquifer thickness, and the assumed 150-ft thickness is a factor that causes bedrock transmissivity calculations to be conservative. Therefore, the calculated transmissivities are estimates applicable to the upper part of the bedrock, where data were available and where the bedrock was probably most permeable. Even though transmissivity may double, depending on the depth of bedrock chosen, transmissivities on the map vary by several thousand square feet per day. Therefore, the possible error in assuming an incorrect thickness is small compared to the range in transmissivity. Determining transmissivity of bedrock for a thickness greater than 150 ft was not within the scope of this study.

The 1 to 1 ratio was assumed because no other information was available on the hydraulic-conductivity ratios for the bedrock. A preliminary model of the Madison County study area by Lapham (1981, p. 31) indicated no significant differences in the bedrock-head distributions for ratios of 1 to 1 and 100 to

1. To correct the transmissivities for partial penetration, the author used anisotrophy ratios of 1 to 1 and 100 to 1. These ratios changed the calculated value at a well by at most 50 percent, and for most data, much less than 50 percent. However, areally the difference in the transmissivity of the bedrock varied by several thousand square feet per day. Therefore, as in the error of choosing thickness, the possible error in assuming an incorrect ratio is small compared to the range in transmissivity. Calculated transmissivities were plotted at their respective locations on a map (fig. 18) that represents the areal variation in transmissivity. Besides the transmissivity calculations, the map also contains some transmissivities determined during model calibration. Transmissivities generally ranged from 500 to 10,000 ft^2/d.

Confining Beds

The aquifers are separated by varying thicknesses of confining beds, which are composed primarily of till. Although vertical hydraulic conductivity of the till is less than horizontal conductivity, the area of vertical flow is several orders of magnitude greater than the area of horizontal flow. Thus, the till transmits little water horizontally, and vertical flow through the till between aquifers dominates. Vertical flow is controlled in part by the vertical hydraulic conductivity of the confining layers.

The only source of information available for vertical hydraulic conductivity of the till near the study area was Meyer and others (1975, p. 25-26). During calibration of their Marion County analog model, they found that conductivity ranged from 1×10^{-4} to 13×10^{-4} ft/d . The average of these two values, 7×10^{-4} ft/d, was chosen as the first vertical hydraulic conductivity for the model simulations. Final conductivity ranged from 7×10^{-4} to 7×10^{-2} ft/d.

Streambed Leakance

The only significant flow through streambeds is water transferred vertically through the streambed to and from the underlying aquifer. This flow is controlled in part by the streambed leakance--the vertical hydraulic conductivity of the streambed divided by the streambed thickness. The streambed thickness was assumed to be 1 ft because no data for determining the thickness were available. Although 1 ft may not be the actual thickness, the assumption does not cause error. If the ratio of streambed conductivity to thickness allows the correct seepage, the thickness can be any reasonable value. Then the vertical hydraulic conductivity is adjusted to allow the field-measured flow through the streambed in model simulations.

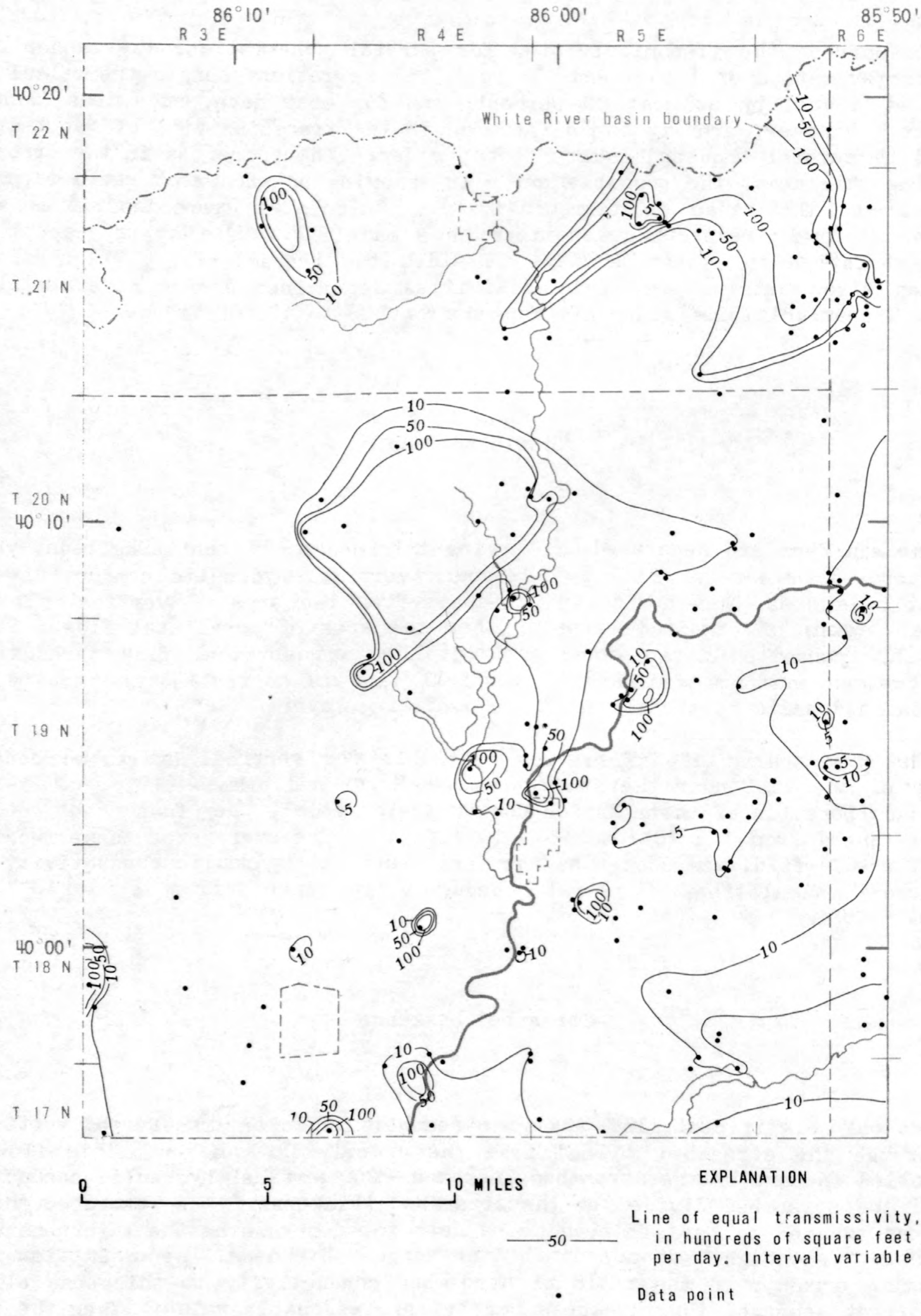


Figure 18.-- Transmissivity of bedrock aquifer estimated from specific-capacity data.

The initial vertical hydraulic conductivity of the streambed used was 4×10^{-2} ft/d. This conductivity is the average that Meyer and others (1975, p. 19) used for a clay layer in the outwash aquifer in Marion County. Because the clay investigated by Meyer and others had a vertical conductivity two orders of magnitude times that of the till, the clay layer possibly originated as a less compacted fluvial sediment. Therefore, as a first vertical conductivity, the hydraulic properties of the streambed material, also a fluvial sediment, were assumed to be similar to that of the clay layer. Most of the streambed conductivities resulting from calibration were within the range from 4×10^{-2} to 4.0 ft/d, and more than one-third were 4.0 ft/d.

Ground-water Flow

The general flow pattern in the drift and the bedrock can be seen by comparing water levels from observation wells. Adjacent shallow and deep wells were installed at several points in the project area. Water-level measurements showed that flow is generally downward in upland areas, mainly horizontal near major streams, and then upward at major streams. This is also the pattern of discharge to small streams. The flow pattern is established by water flowing mainly vertically through the till, then mainly horizontally through the aquifers.

Water-level data from the first 150 ft of the bedrock aquifer indicate that flow through the bedrock was to the major streams. Data also indicate that vertical flow extended deeper than the 150 ft bedrock layer, but this flow was probably small and also was not within the scope of the study.

The flow pattern and the horizontal gradients are well represented by the calibrated potentiometric surfaces shown in the section "Model Calibration." The flow pattern indicates that most ground-water discharge was to streams within the study area. This pattern is typical in downstream counties within the project area. However, upstream in Randolph County, most ground-water flow is out of that county. The large outward flow indicates that Randolph County is basically a recharge area.

The horizontal gradient shown in figures 27-30 is commonly about 10 ft/mi. Vertical gradients are upward or downward depending on the location relative to divides and streams. The vertical gradient generally ranges from ± 4 ft between the bedrock and two lowest sand and gravel aquifers to as much as $\overline{3}0$ ft between aquifers 2 and 3. Data were insufficient to determine vertical gradients between aquifers 3, 4, and 5. The vertical and horizontal gradients indicate the need to consider three-dimensional flow in an analysis of the ground-water flow system.

Water-Level Changes

Water levels change in response to variations in evapotranspiration, major pumping, and natural ground-water discharge and recharge. Some of these changes are illustrated in the water-level data from three Geological Survey continuous-record observation wells. The first well, designated Hamilton 5, is 2.5 mi southwest of Noblesville and 1 mi west of the White River. The well is screened at about 80 ft within the outwash aquifer (fig. 22, well 32A). Water-level data collected at the well for 14 years are shown in figure 19. Depth to water at the well ranges from 8.5 to 11.7 ft. A seasonal cycle is noticeable, but no general trend in water levels is obvious. Rather, water levels fluctuate around the 10-foot level. The trends of hydrographs for observation wells Madison 7 and 8, near Anderson in Madison County, were similar to the trends of the hydrograph for Hamilton 5 (Lapham, 1981, p. 36). The two wells are in aquifers whose hydraulic characteristics are similar to those of aquifers in Hamilton County. Madison 7 and 8, operated from about 1950 to 1970, have depths of 20 and 415 ft. Madison 8 penetrates bedrock. Amplitudes in Madison 7 and 8 were 7 and 10 ft, respectively, and the average water levels were nearly constant. The seasonal fluctuation indicates losses to evapotranspiration and subsequent recharge by precipitation. The constant range of fluctuations result from two factors, insignificant pumpage increases and nearly uniform precipitation.

Effects of evapotranspiration on water levels during the growing season is demonstrated by declining water levels. The loss of water to evapotranspiration is by interception of recharge and possibly by direct depletion from the water table. In their study of ground water in Marion County, Meyer and others (1975, p. 38) stated that soil moisture is almost the only source of water for growing plants and that the depth to the water table is generally too great for significant evapotranspiration directly from the water table. The water loss from interception of recharge and direct depletion from the water table is not known, and determining these numbers was not within the scope of this report. Evapotranspiration was accounted for in modeling by a reduction of recharge to the aquifer.

Besides seasonal changes, long-term trends in the average water level were also studied. However, no trends were apparent, either in the observation-well hydrographs or water-level data from counties adjacent to Hamilton. Water levels in observation wells throughout Madison County returned to about the same level in each autumn of 1976-78 (Lapham, 1981, p. 35). For the same till system and similar pumping, water levels in Hamilton County probably respond similarly. Data from Marion County also suggest that water levels in Hamilton County are at steady state. Even in Marion County, where pumping exceeds that in Hamilton County, a graph by Meyer and others (1975, fig. 16) shows that, during normal precipitation, the general water-level surface in the till fluctuated only slightly from about 1956 to 1966. The 11-yr record was considered to be in a normal precipitation period because average precipitation for those years is only 0.11 in. higher than the 30-yr average (National Oceanic and Atmospheric Administration, 1956-66). The steady water levels indicate that pumpage increases in the outwash and till did not cause a downward water-level

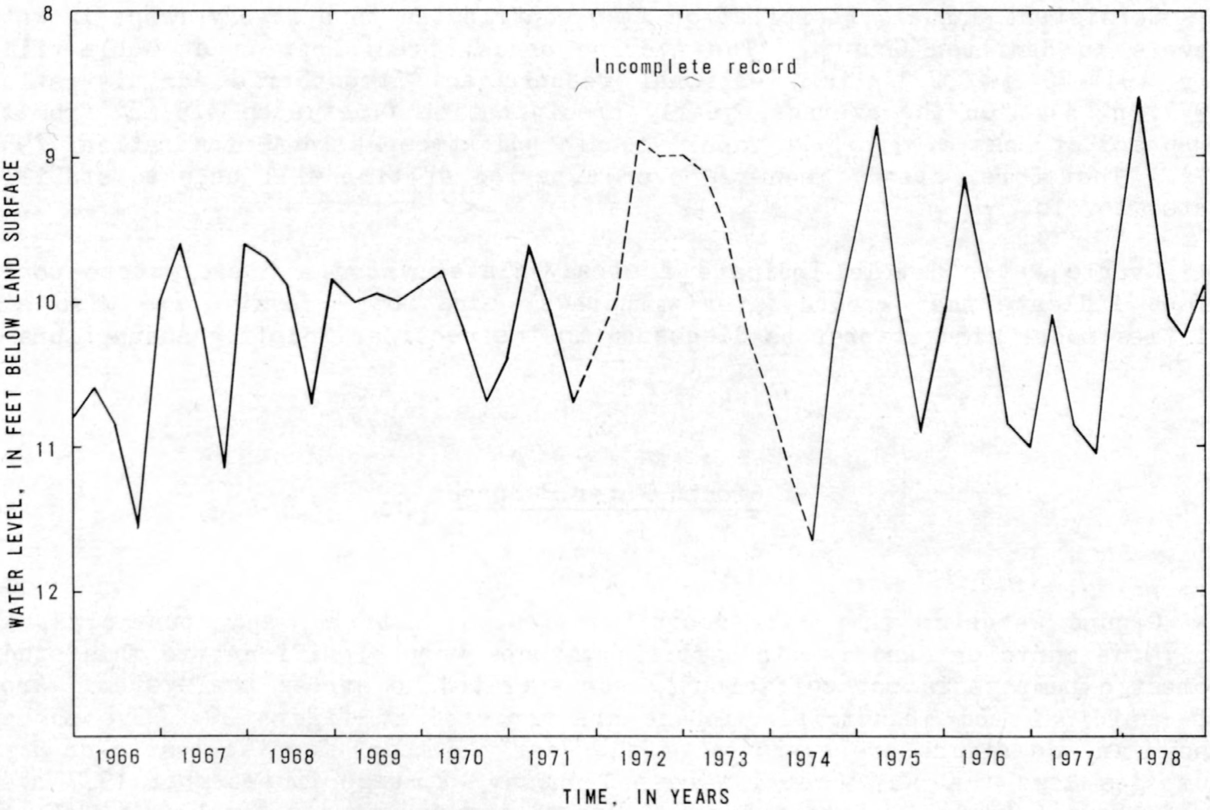


Figure 19.-- Water-level fluctuations in observation well Hamilton 5.

trend. If the ground water system responds without noticeable water-level declines in Marion County, then the system response in Hamilton County would probably be similar.

Consistent annual precipitation also contributes to a steady range in water levels in Hamilton County. The average annual precipitation at Noblesville, for 1941-70 is 37.11 in. (National Oceanic and Atmospheric Administration, 1977, p. 4). On the average, yearly precipitation fluctuates 3.5 in. from the mean and at most 8.8 in. (National Oceanic and Atmospheric Adminstration, 1956-77). Therefore, steady recharge over a period of time will help to stabilize water levels.

Stable water levels indicate a steady-state system. Steady-state conditions indicate that ground-water withdrawals are not excessive and also simplifies model simulations, as discussed in the section "Modeling Assumptions."

Ground-Water Pumpage

Ground water is pumped by municipalities, industries, and households, but only the municipal and the industrial pumpages were significant to this study. Domestic pumpage is not sufficiently concentrated to stress the system. Areas of municipal and industrial pumpage are depicted in figure 20. The numbers shown in the figure are averages of the yearly pumpage for the last 4 or 5 yr. Pumpages less than 0.1 Mgal/d were not shown. Pumpage in December 1977 averaged about 6 Mgal/d. Except for 0.4 Mgal/d pumpage from a gravel pit, all the rest is municipal.

Ground-Water Interaction with Streams

Ground-water discharge to streams was quantified by gain-and-loss studies along several streams. The increase in streamflow (ground-water seepage), about 70 ft^3/s on October 30, 1977, is shown in figure 21. Discharge was measured during low flow. At Noblesville, on the White River, discharge was 173 ft^3/s, a 78-percent flow duration (Horner, 1976, p. 232).

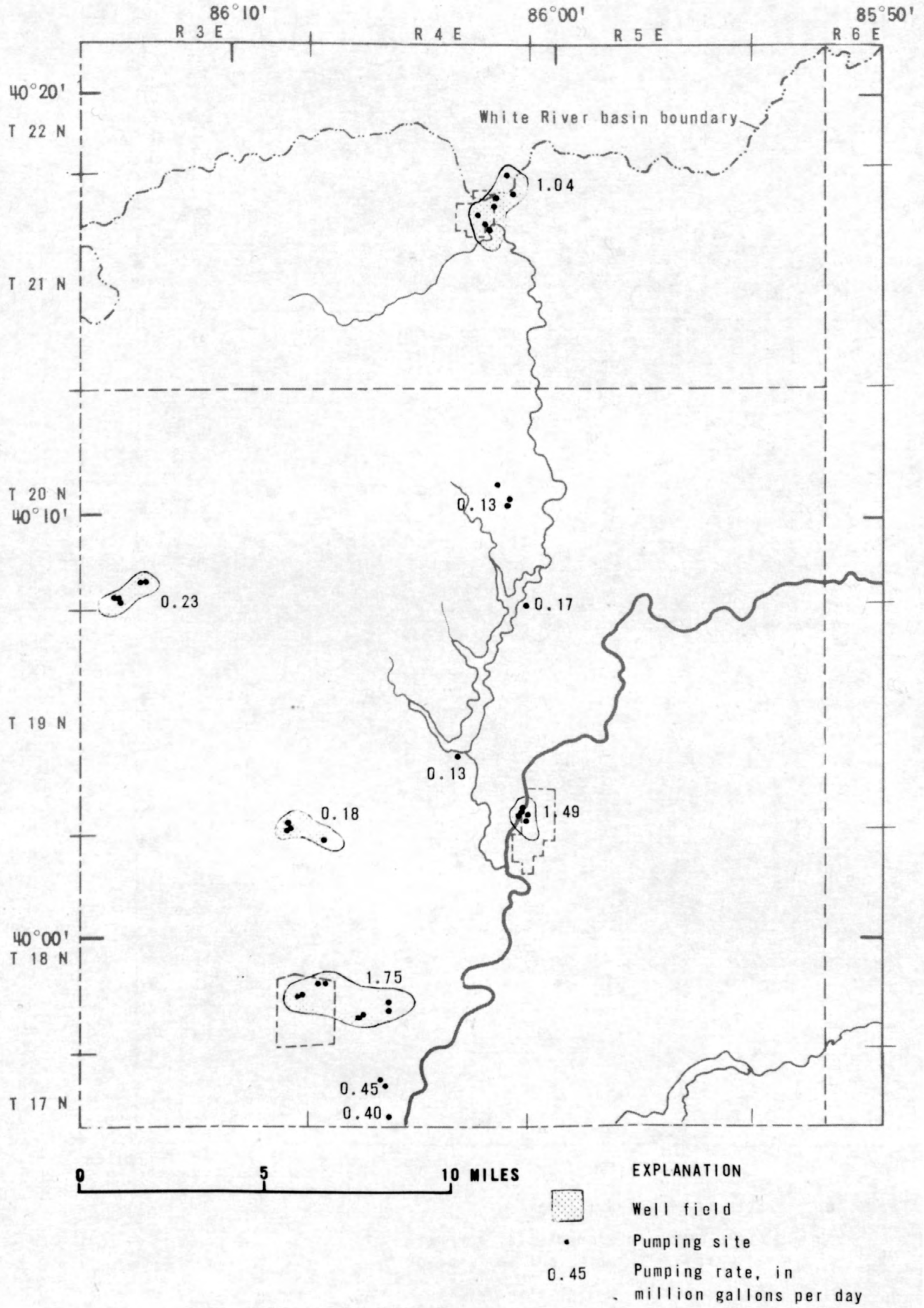
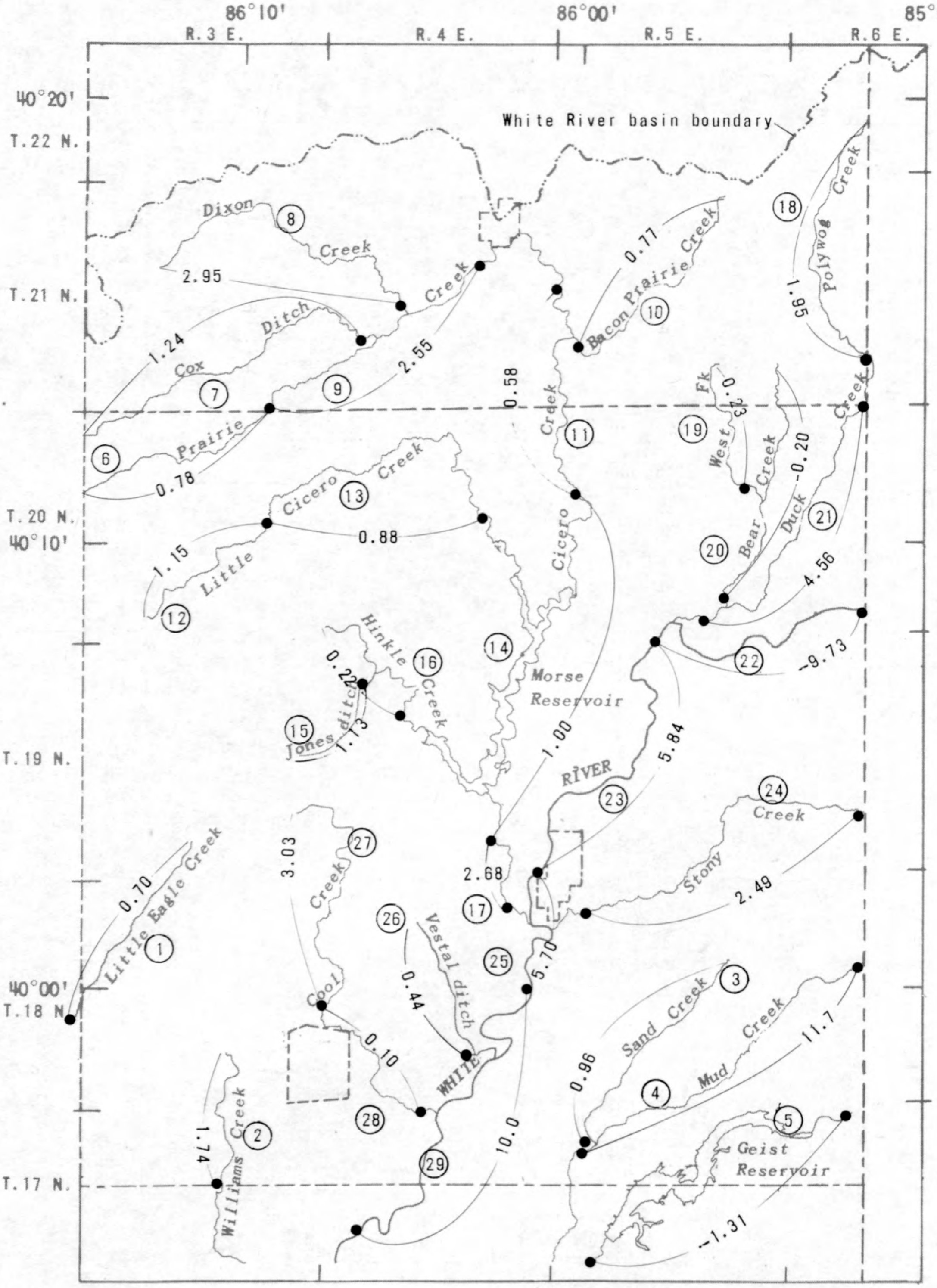


Figure 20.-- Areal distribution of ground-water pumpage.



EXPLANATION

- ● Site of measured streamflow
- 6.9 Field-measured ground-water seepage to stream, in cubic feet per second
- ① Stream section designation

Figure 21.-- Stream sections where ground-water-seepage rates were estimated, October 30, 1977.

SIMULATION OF GROUND-WATER FLOW

Description of the Model

Three major aquifer systems underlie the study area: confined sand and gravel, bedrock, and outwash. The till, containing five aquifers separated vertically by till-confining beds, is underlain by a bedrock aquifer. The outwash aquifer consists of one sand and gravel unit containing minor lenses of clay.

One objective of the study was to determine the pumping that is possible and the effect of this pumping on the ground-water system. Pumping is most likely to increase in the outwash aquifer because of its development potential and its location near population centers. Therefore, a two-dimensional model was constructed to simulate flow in the outwash aquifer enclosed by the rectangle in figure 2. This model excluded the bedrock aquifer, except for a few bedrock highs, because the contribution of flow from the bedrock to the outwash aquifer was insignificant. The first modeling of the outwash was done by Gillies (1976) in the south part of the outwash aquifer in Hamilton County. The two-dimensional model overlaps some of this area and extends northeast to the east boundary of Hamilton County (fig. 22). The area modeled three-dimensionally is also shown in figure 22. This model simulates flow in the outwash, confined sand and gravel, and bedrock aquifers and estimates the effects of pumping on the system. In the three-dimensional model, the bedrock is modeled because bedrock transmissivity is significant relative to that of the materials above it.

Construction of the models was based on the author's understanding of the flow system. Incorporating every detail of the drift into the model was not possible. Therefore, assumptions and generalizations were made that did not significantly affect the accuracy of results but that did reduce the complexity of the model. Because the three-dimensional model was calibrated before the two-dimensional model, the three-dimensional model is discussed first in the sections that follow.

Modeling Assumptions

Assumptions for the three-dimensional model simulating the outwash, drift, and bedrock aquifer systems follow. Some of the assumptions are illustrated in figure 23.

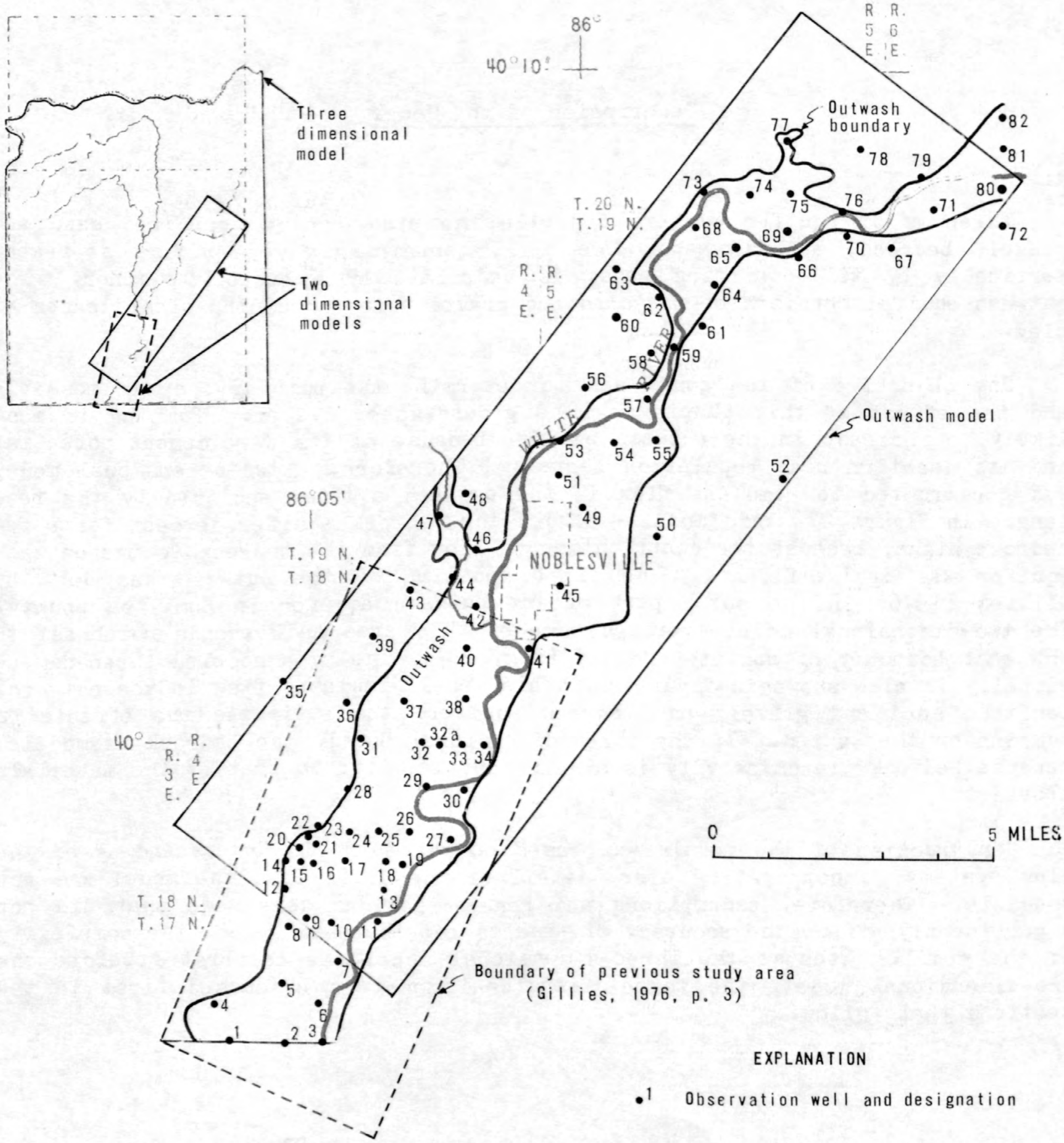


Figure 22.-- Locations of previous and current modeling studies and of observation wells.

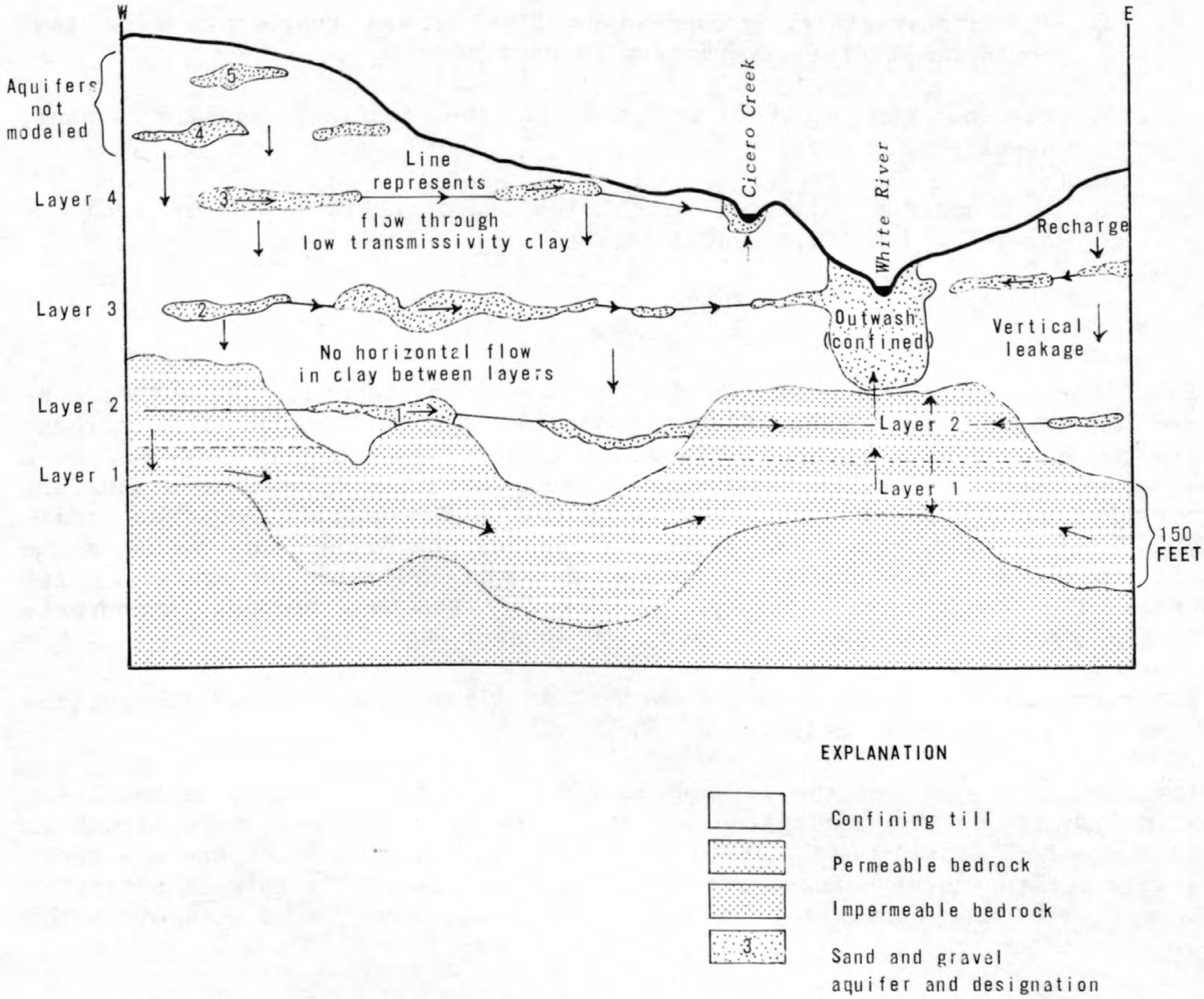


Figure 23.-- Generalized geohydrologic section showing design of the three-dimensional model.

1. Flow in the ground-water system is three dimensional. (a) Flow in the aquifers and the till separating two horizontal adjacent units of the same aquifer is horizontal. The confining beds vertically separating two different aquifers contain no horizontal flow. (b) Flow through the confining beds between the aquifers is vertical.

2. The sand and gravel aquifers are homogenous and horizontally isotropic.

3. (a) The upper 150 ft of bedrock is permeable. (b) The hydraulic conductivity of the bedrock at a point is uniform with depth but varies areally. (c) The ratio of vertical to horizontal hydraulic conductivity in the bedrock is 1:1.

4. The steady-state, ground-water flow system represents flow that could be obtained by pumping in most years.

5. The outwash aquifer is part of the confined aquifer system, specifically aquifer 2.

6. All aquifer material, except the bedrock, has a uniform sand and gravel hydraulic conductivity.

7. Streambeds are 1 ft thick.

Even though some flow in the confining beds is horizontal, the author considered the flow to be insignificant, compared with horizontal flow through aquifer material, and assumed that horizontal flow was absent. However, in a layer where till is the only material between two units of the same aquifer, horizontal flow through that till was included in the model. For that condition, 2.7 ft^2/d was used as the transmissivity for the till. This value is the product of 0.134 ft/d and 20 ft assumed for the hydraulic conductivity and thickness of the till, respectively. Twenty feet was used because it approximated the average thickness of sand and gravel aquifers.

The significance of assumption 4 is that it eliminates the need for a storage coefficient in the model.

Assumption 5 says that the outwash aquifer is confined, yet it is mostly an unconfined aquifer. The assumption was made, however, to simplify construction of the three-dimensional model. Even with the assumption, predictions are realistic unless water-level changes are 10 percent or more of the outwash saturated thickness. For that condition, transmissivity would have to be changed in the model.

Five assumptions for modeling the outwash aquifer follow. Some of the assumptions are illustrated in figure 24.

1. Flow in the outwash aquifer is horizontal only.
2. The outwash aquifer is isotropic.
3. The only flow in bedrock is in bedrock highs that intercept ground-water flow from the outwash.
4. Streambeds are 1 ft thick.
5. The ground-water flow system is in steady state.

The predominance of horizontal flow (item 1) can be demonstrated at well site 23. Shallow and deep wells at the site are in the outwash aquifer. One well is about 40 ft deeper than the other, and water levels are commonly within 0.02 ft of each other. However, some water flows vertically throughout the system. The aquifers are naturally recharged from above by precipitation, and water is discharged from the aquifers into streams.

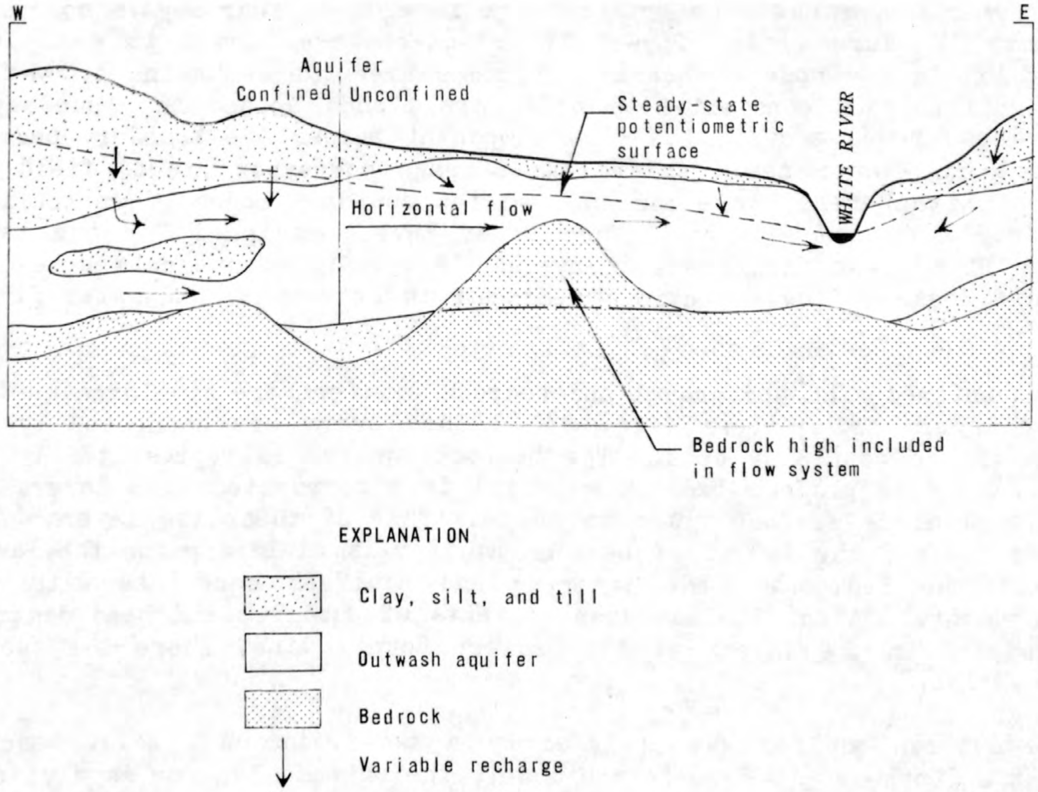


Figure 24.-- Generalized geohydrologic section showing design of the two-dimensional model.

Assumption 3 was made because of the low bedrock transmissivity and variable bedrock configuration. The bedrock aquifer was assumed to be impermeable in most of the modeled area because bedrock transmissivity is generally about one-tenth that of the outwash aquifer. However, some bedrock highs were within a few feet of the land surface or were covered with only thin layers of sand and gravel. In these examples, most of the flow is through the bedrock, and, therefore, bedrock transmissivity must be modeled. The north half of the modeled outwash aquifer contains bedrock highs of this type. A notable example of a bedrock high can be seen at Riverton (fig. 3), where bedrock is exposed and rises above the water-level surface.

Model Selections and Designs

The outwash, drift, and bedrock aquifer systems were analyzed with a three-dimensional model described by Trescott (1975). Two modifications of the computer code were made to allow more flexibility in model simulation. Streams and recharge can be simulated in any layer of the model exposed to the surface.

The three-dimensional flow system is divided horizontally by a grid and vertically by layering. The grid network in each of four layers consists of 37 rows and 53 columns (fig. 25)--1,961 block-centered nodes in each layer and 7,844 nodes in the model. Nearly all nodes are squares having 3,000-ft sides. During calibration, constant-head nodes are placed around the boundary of the grid. These nodes provide the flow component across the boundary necessary to sustain ground-water heads and stream seepage corresponding to field measurements. Constant heads were assigned to the boundary nodes by contouring known water levels for each layer. Then water levels assigned for one layer were compared with those for adjacent layers. Necessary adjustments were then made to insure logical flow directions near and in between ground-water divides and streams.

The confined sand and gravel and bedrock aquifers are simulated in the model as four layers or aquifers (fig. 23). The layers are connected by vertical leakage in the confining beds. The bedrock aquifer is represented by layer 1. However, in some places, bedrock material is incorporated into layers 2 and 3, where the bedrock surface rises to the altitude of these two layers (fig. 23). In those places, the 150 ft of bedrock aquifer is divided among the layers that intersect the bedrock. The layering and aquifers correlate with those in Madison County. Also, similar areal extents of aquifers and head distributions were modeled in the layers at the Madison County line, where the two modeled areas overlap.

The outwash aquifer was analyzed by a two-dimensional model described by Trescott and others (1976). In the model, the outwash aquifer is divided into a grid network consisting of 41 rows and 123 columns, for a total of 5,043 block-centered nodes (fig. 26). A variable-grid spacing was used to provide a finer grid network in the south half of the modeled area, where most new pumpage was simulated, than in the north. The square nodes of this area are 600 ft on a side, whereas node dimensions in the north half of the modeled area are 600 by 1,200 ft. The model grid was extended to 3 mi south of 146th Street (fig. 3), or 3 mi into the area previously studied by Gillies (1976). In that way, model ed pumping just north of 146th Street would not be affected significantly by the south boundary of the modeled area.

Nodes that simulate special conditions were used at several grid locations. Constant ground-water-flux nodes were placed on the boundary of the modeled area. The flux values were determined by calculating the amount of flux at the two-dimensional model boundary from layers 2, 3, and 4 of the calibrated three-dimensional model. The same constant-flux values were used during calibration and pumping simulations. Therefore, additional flux across the boundary was not possible in response to additional pumping during pumping simulations. Pumping nodes were placed in areas of significant pumping (fig. 20). The simulated wells fully penetrated the aquifer and had no well-entrance loss. The White River and Cicero Creek were simulated by river nodes that allowed appropriate leakage through a confining streambed.

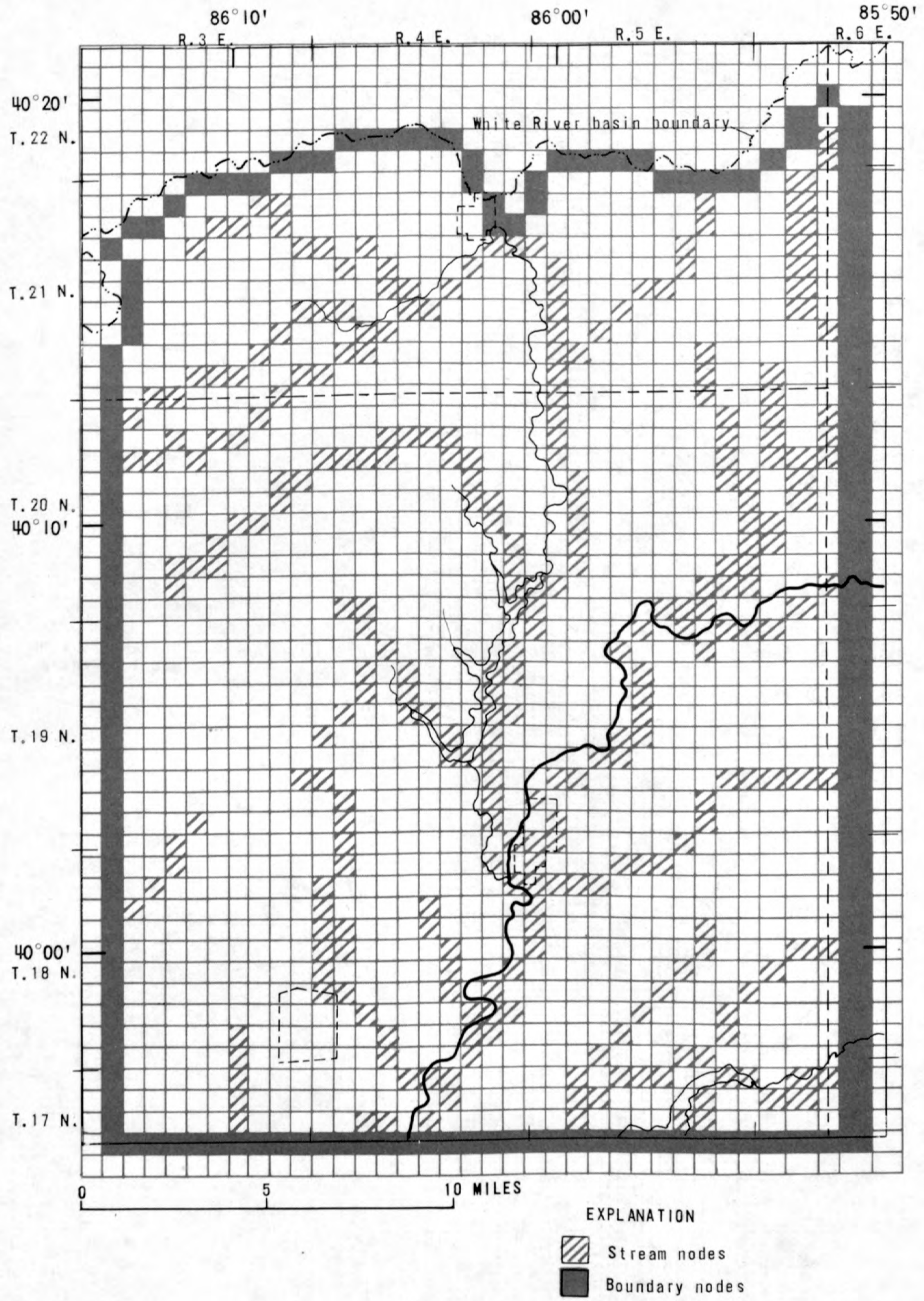


Figure 25.-- Finite-difference grid for the three-dimensional model.

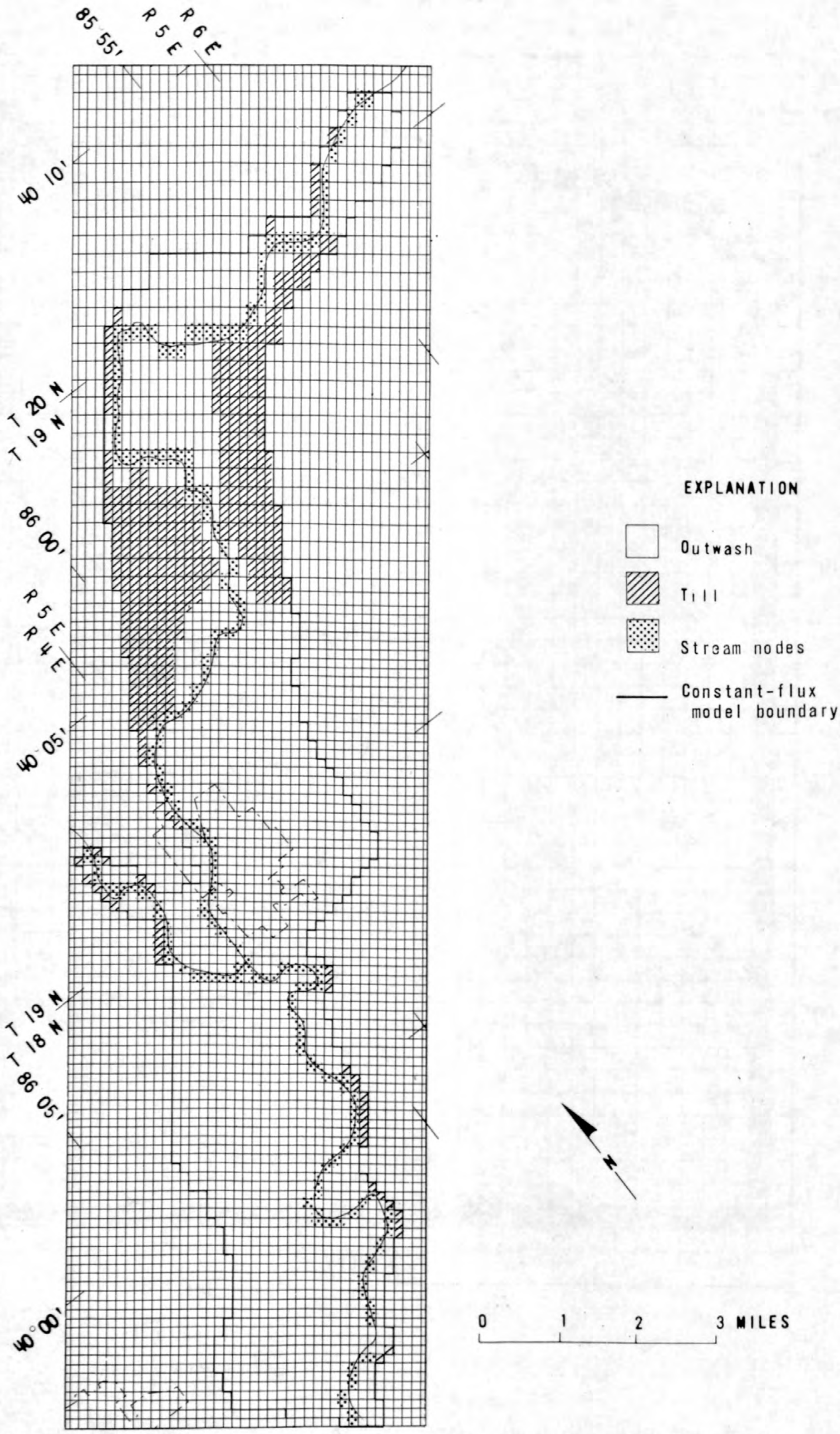


Figure 26.-- Finite-difference grid for the two-dimensional model.

Model Calibration

After the model had been designed, and data for variables had been put into the model, the model was calibrated for estimating the response to different pumping plans. During calibration, model parameters were adjusted to obtain a best match between measured and computed values. Adjustments were made until additional changes did not significantly improve the match. The parameters adjusted first were those having the least information available. Calibration of the three-dimensional model before the two-dimensional model allowed calculation of ground-water flux at the two-dimensional model boundary and provided some of the other input data for the two-dimensional model.

Three-Dimensional Model

A series of steady-state simulations consisted of matching model-simulated water levels and ground-water seepage to measured data. The water levels were measured during November and December 1977, whereas seepage was measured on October 30, 1977.

Model parameters were adjusted throughout a layer to bring simulated water levels within a narrow range of the measured levels. The first parameter that was adjusted in this manner was vertical hydraulic conductivity. Because simulated water levels were generally too high, the vertical conductivities were increased one to two orders of magnitude. Vertical hydraulic conductivity that best approximated measured data ranged from 7×10^{-4} to 7×10^{-2} ft/d and averaged about 7×10^{-3} ft/d. Although the increase caused improvements in almost the entire model, vertical hydraulic conductivity adjustments did not solve all problems. Agreement between measured and computed water levels and stream seepage was improved by adjusting aquifer hydraulic conductivity.

In the calibration, hydraulic conductivity in the sand and gravel aquifers was decreased from 433 ft/d to 216 ft/d. Two benefits resulted from the reduction. The match between measured and simulated water levels near the White River improved, and inflow from the boundary to the White River decreased to a more acceptable value. At the end of calibration, 433 ft/d was used in a model run to determine if the system response had improved. With this conductivity, the simulated water levels in several nodes were too low, seepages into upland streams were too low, and seepages into the White River were too high.

Adjusting the clay transmissivity had little effect on the system. For example, neither water level nor seepage varied perceptibly in the range from 0 to 25.9 ft^2/d. Therefore, the first transmissivity used, 2.59 ft^2/d, was kept for future model simulations.

In a few areas, bedrock transmissivity was adjusted to match measured water levels and water-level gradients. Transmissivity distribution based on transmissivity calculations and calibration adjustments are shown in figure 18.

Recharge adjustments were useful in selectively changing water levels and ground-water discharge to streams. Recharge to the till was varied from 2 to 4.5 in./yr, whereas recharge to the outwash, mainly along the White River, was 6 in./yr. The outwash recharge is about half that used by Gillies (1976, p. 18) and Meyer and others (1975, p. 48) and may seem to be too low. However, the 6-in. outwash recharge provided a closer match between computed and measured seepage than the recharge used by Gillies and by Meyer and others. Reasons for higher outwash recharge used by these investigators are that their model calibrations were in conjunction with a higher streamflow duration than the one used in this (Hamilton County) study and that some tributary stream discharges were included in the seepage figures. The additional water included in the seepage measurements was produced by the additional recharge. Therefore, because field data were collected during a lower flow duration in the present study than that in the 1975 and 1976 studies and did not include White River tributary discharge in the measured streamflow increase, the outwash should receive significantly less recharge than it received in past studies.

Recharge from precipitation represents recharge to the regional flow system and does not include recharge that circulates in the shallow ground-water system and discharges locally to small streams that were not modeled. However, the amount of recharge that discharges locally as shallow ground-water circulation is probably insignificant relative to the quantity that enters the regional flow.

The last model parameter adjusted in the calibration was the streambed hydraulic conductivity. For a streambed thickness of 1 ft, streambed conductivity generally ranged from 4×10^{-2} to 4.0 ft/d. However, for more than one-third of the stream sections, 4 ft/d produced the best match. In many simulations, a range of streambed conductivities for a specific stream section produced similar quantities of seepage, but the lowest conductivity was used.

Measured and calibrated stream seepages associated with streambed hydraulic conductivity are given in table 1. Although measured seepages are given, the true seepages at the time of measurement may be different because of measurement error. Therefore, possible ranges of seepage based on measurement error are given in the table. During calibration, the author attempted to match computed and measured seepages as closely as possible. But, if the computed seepage was at least within the range of seepage for a section, then the model was considered to be matching field conditions.

Computed seepages for several stream sections are outside the possible range of seepages. However, no attempt to improve the match between computed and measured seepages was made because that would conflict with other information known about the system. For example, seepage into streams typically increases per unit length downstream. Variation from this pattern was apparently caused by local and temporary variations in recharge. One example of this variation is Mud Creek, where discharge increased as much as 12 ft^3/s between measurement sites compared with an average for the study area of about 2.5 ft^3/s. Because seepage in Mud Creek was significantly different from seepages at nearby sections, no attempt was made to get a complete match of the anomalous seepage caused by the temporary and local difference in recharge. Also, in some sections, further adjustments of seepage would cause errors in nearby ground-water levels. Therefore, the measured water levels were used for calibrating

Table 1.--Measured and model-simulated seepages for stream sections in the study area

Stream name	Stream section[1]	Field-measured seepage at about 78 percent flow duration[2] (ft^3/s)	Calculated range of field-measured seepage (ft^3/s) Minimum	Calculated range of field-measured seepage (ft^3/s) Maximum	Model-simulated seepage (ft^3/s)
Little Eagle Creek	1	0.70	0.64	0.91	0.39
Williams Creek	2	1.7	1.6	1.9	1.4
Sand Creek	3	.96	.77	1.2	.61
Mud Creek	4	12	11	13	7.0
Geist Reservoir	5	-1.3	-4.2	1.6	1.4
Prairie Creek	6	.78	.62	.94	.42
Cox ditch	7	1.2	.85	1.6	1.1
Dixon Creek	8	3.0	2.6	3.2	1.7
Cicero Creek	9	2.6	1.2	4.3	1.4
Bacon Prairie Creek	10	.77	.62	.92	.74
Cicero Creek	11	.58	-1.3	2.5	3.0
Little Cicero Creek	12	1.2	1.0	1.3	.44
Do.	13	.88	.49	1.3	1.0
Morse Reservoir	14	1.0	-4.0	5.8	2.9
Jones ditch	15	1.1	1.0	1.2	.61
Hinkle Creek	16	.22	- .03	.44	.22
Cicero Creek	17	2.7	- .71	4.6	2.1
Polywog Creek	18	2.0	1.7	2.9	1.5
West Fork	19	.23	.18	.28	.39
Bear Creek	20	- .20	- .29	.13	.65
Duck Creek	21	4.6	3.1	6.1	4.7
White River	22	-9.7	-27	8.1	3.8
Do.	23	5.8	-12	23	14
Stony Creek	24	2.5	1.5	3.6	2.2
White River	25	5.7	-15	26	8.3
Vestal ditch	26	.44	.35	.53	.10
Cool Creek	27	3.0	2.7	3.3	2.8
Do.	28	.10	- .80	1.0	.46
White River	29	10	-7.9	31	12.4

[1]Location of stream section shown in figure 21.
[2]Negative number indicates flow into the ground-water system.

seepages. The difficulty in matching seepages in some of the stream sections might indicate the inability of the model to simulate the flow system in those sections. Yet, the differences in seepages should not limit the usefulness of the model in estimating the effect of stresses on the ground-water flow system.

The three-dimensional, steady-state calibration of potentiometric surfaces of each aquifer and measured water levels are shown in figures 27 through 30. Nearly all water levels matched within 10 ft, and about half matched within 5 ft. This degree of match was comparable with those in models of other parts of the White River basin project area and the match in drift deposits of Marion County obtained by Meyer and others (1975, p. 48). Computed and measured stream seepages are shown in figure 21. Finally, table 2 is a water budget that represents the model-simulated distribution of ground-water flow for stream seepages at a 78-percent flow duration.

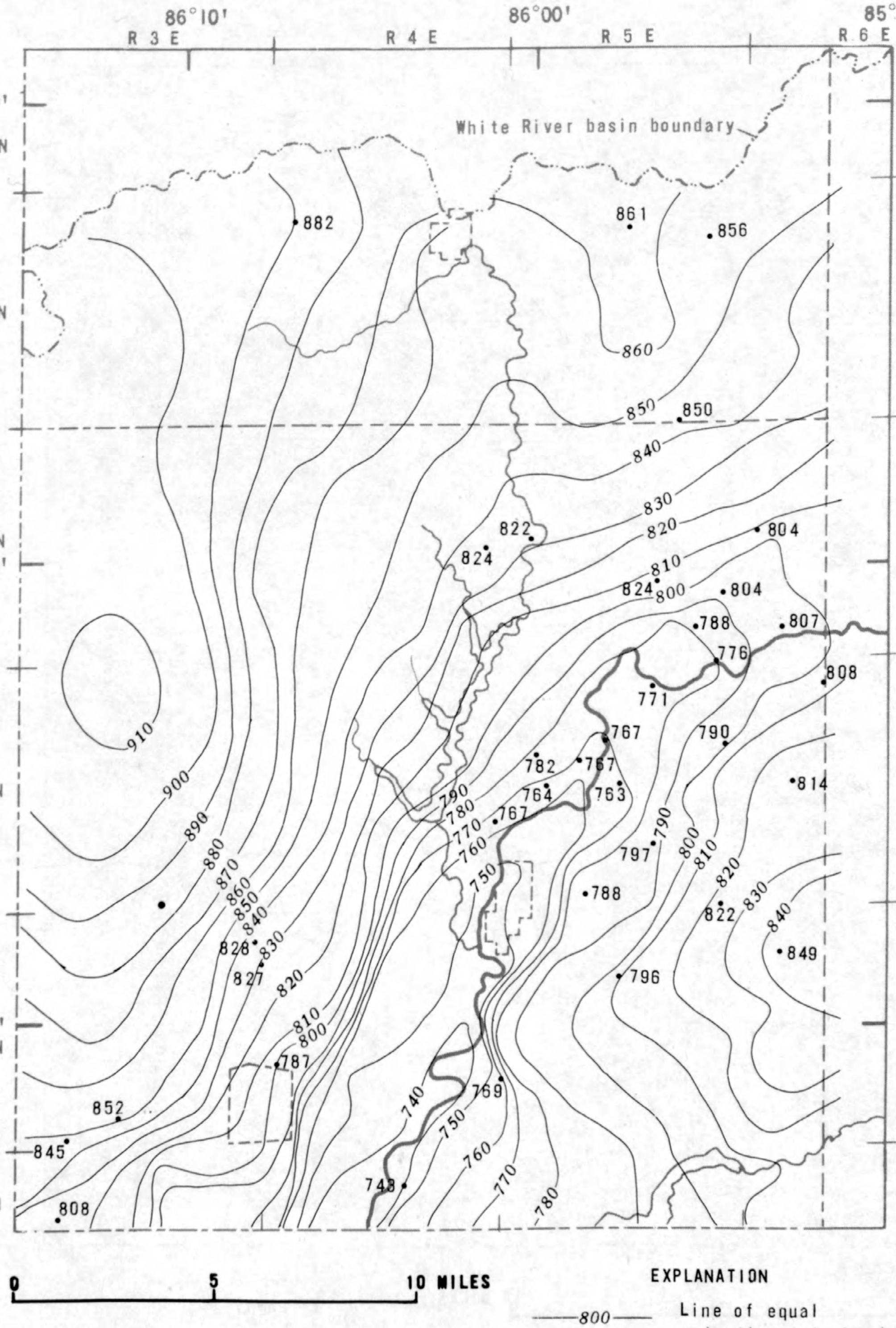
Table 2.--Steady-state water budget for the three-dimensional model

Recharge	(ft^3/s)	Discharge	(ft^3/s)
Precipitation	94.4	Ground-water pumpage	7.25
Ground-water underflow into study area	12.4	Ground-water discharge to streams	82.6
		Ground-water underflow out of study area	19.8
Total recharge	106.8	Total discharge	109.6

Two-Dimensional Model

The calibration of the two-dimensional model of the outwash aquifer was similar to that of the three-dimensional model, except that the variables determined in the three-dimensional model were used as estimates for variables in the outwash model. The variables were boundary flux, stream seepage, streambed hydraulic conductivity, and recharge. Only boundary flux was not readily available from the three-dimensional model. Fluxes for all layers in the drift were calculated by applying the water-level gradient and harmonic mean of transmissivity at the outwash model boundary to Darcy's law.

The only variable requiring a significant change from the three-dimensional model result to that used in the outwash model was recharge. Recharge to the outwash model had to be increased to supply the additional water coming from the



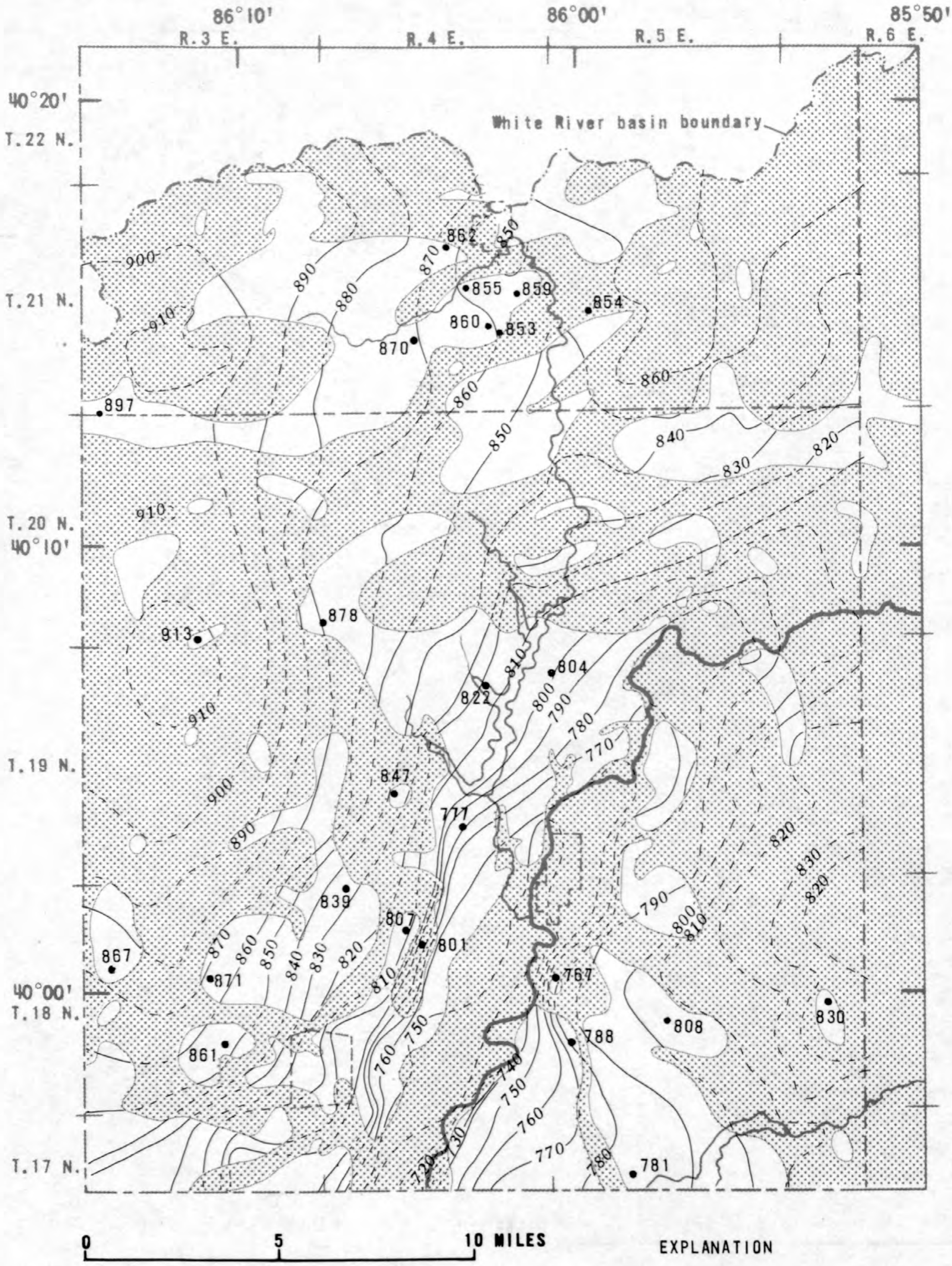
EXPLANATION

—800— Line of equal potentiometric head. Interval 10 feet NGVD of 1929

845 Measured potentiometric head, in feet

• Data point

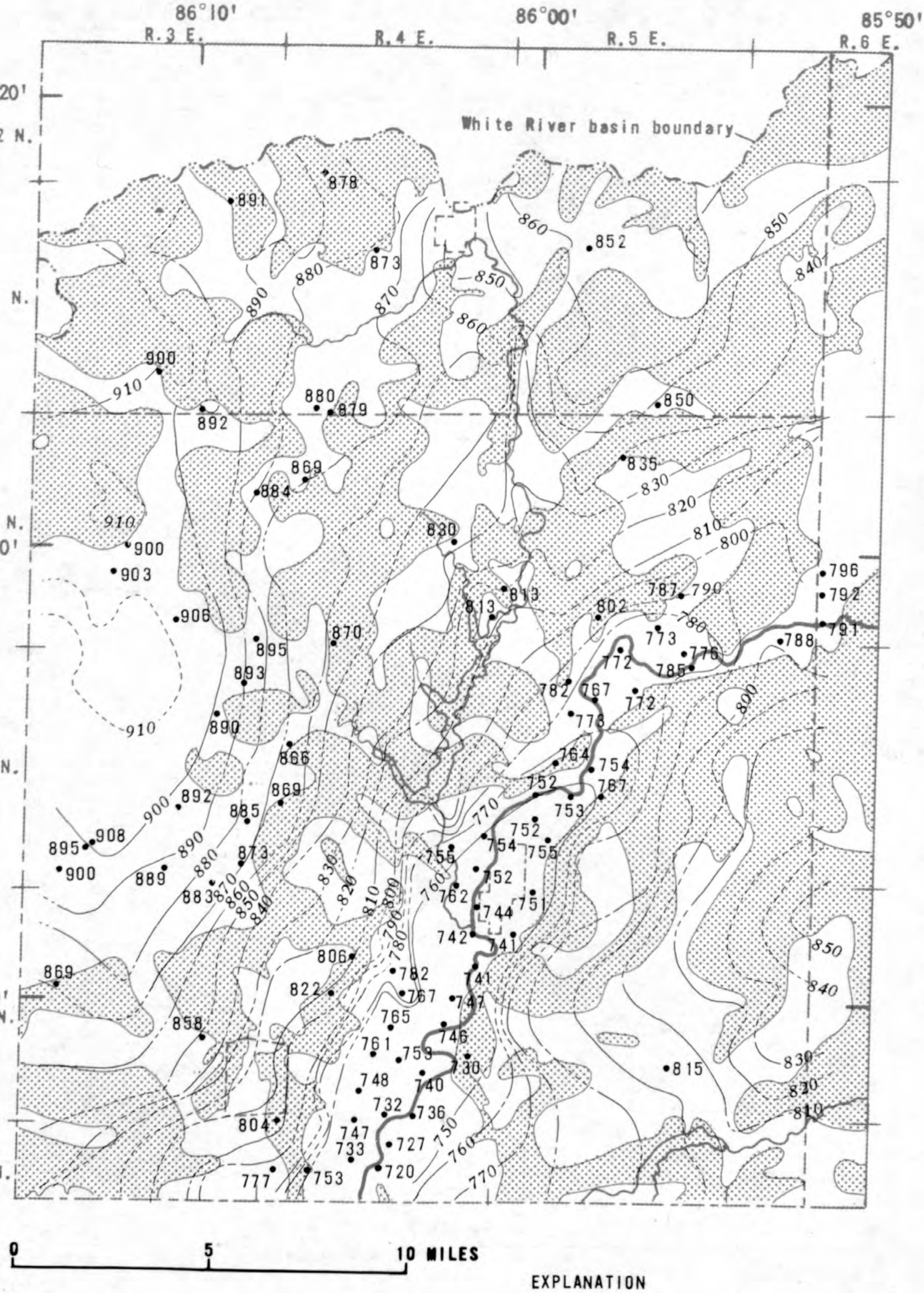
Figure 27.-- Model-simulated steady-state potentiometric surface of the bedrock aquifer.



EXPLANATION

☐ Aquifer	——800---- Potentiometric contour. Interval 10 feet. Dashed where material is not sand and gravel	808	Measured potentiometric head, in feet
		•	Data point
			NGVD of 1929

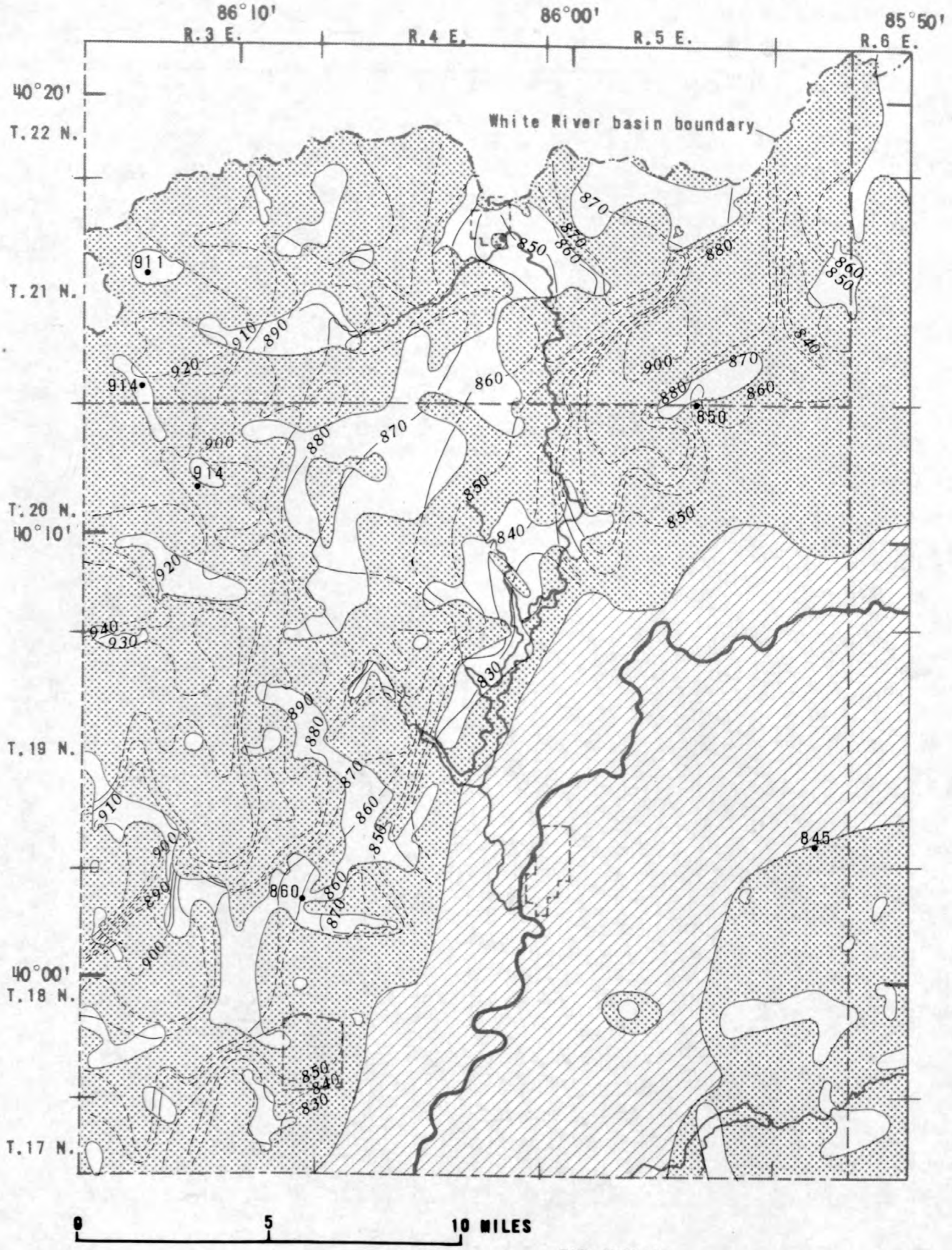
Figure 28.-- Model-simulated steady-state potentiometric surface of aquifer 1.



EXPLANATION

☐ Aquifer	—*800*--- Potentiometric contour. Interval 10 feet. Dashed where material is not sand and gravel	806	Measured potentiometric head, in feet
		•	Data point
			NGVD of 1929

Figure 29.-- Model-simulated steady-state potentiometric surface of aquifer 2.



EXPLANATION

Aquifer

——900——— Potentiometric contour. Interval 10 feet. Dashed where material is not sand and gravel

Approximate area where horizontal projection of aquifer lies above land surface

914 Measured potentiometric head, in feet

• Data point

NGVD of 1929

Figure 30.-- Model-simulated steady-state potentiometric surface of aquifer 3.

nonmodeled bedrock. Boundary flux and streambed conductivity in the two-dimensional model differed slightly from calibrated results in the three-dimensional model. These differences were due to differences in scale, which allowed more detail to be incorporated in the outwash model.

The match of water levels was generally improved by increasing recharge in a downstream direction from 8 to 12 in./yr. The 12-in./yr at the south end, where the study model and Gillies' model (1976) overlap, agrees closely with Gillies' 11.9 in./yr.

As recharge was increased, water levels near the stream in the south two-thirds of the model increased, and, therefore, streambed hydraulic conductivity increased in a pattern similar to that of recharge. Streambed conductivity downstream increased from 0.04 ft/d, the calibrated three-dimensional result, to 0.4 ft/d at the south boundary of the model, where Gillies (1976, p. 18) reported 0.67 ft/d.

Local discrepancies between computed and measured water levels were resolved by adjusting transmissivity and boundary flux. Transmissivities in a few places were adjusted downward so that they more accurately indicated the lower hydraulic conductivity of the silty sand and gravel. Where water levels were low in an area of clean sand and gravel, boundary flux was increased to match heads. Few of the boundary fluxes were altered, and most of the necessary changes were in the southwest area of the model boundary.

The adjustments resulted in a water-level match generally within 3 ft of the measured value. The steady-state, calibrated, potentiometric surface and the measured heads are shown in figure 31. Ground-water seepage in the outwash model fairly closely matched the calibrated seepages used in the three-dimensional model. The two sets of calibrated seepage and the range of calibrated seepage attributible to measurement error are given in the following table. The stream sections correspond to those in figure 21.

Stream section	Two-dimensional model (ft^3/s)	Three-dimensional model (ft^3/s)	Potential measured seepages (ft^3/s)
22	3.02	3.83	-26.6 to 8.14
23	12.6	14.5	-12.3 to 23.0
25	9.74	8.30	-14.8 to 26.2
29	13.1	12.4	- 7.89 to 31.5

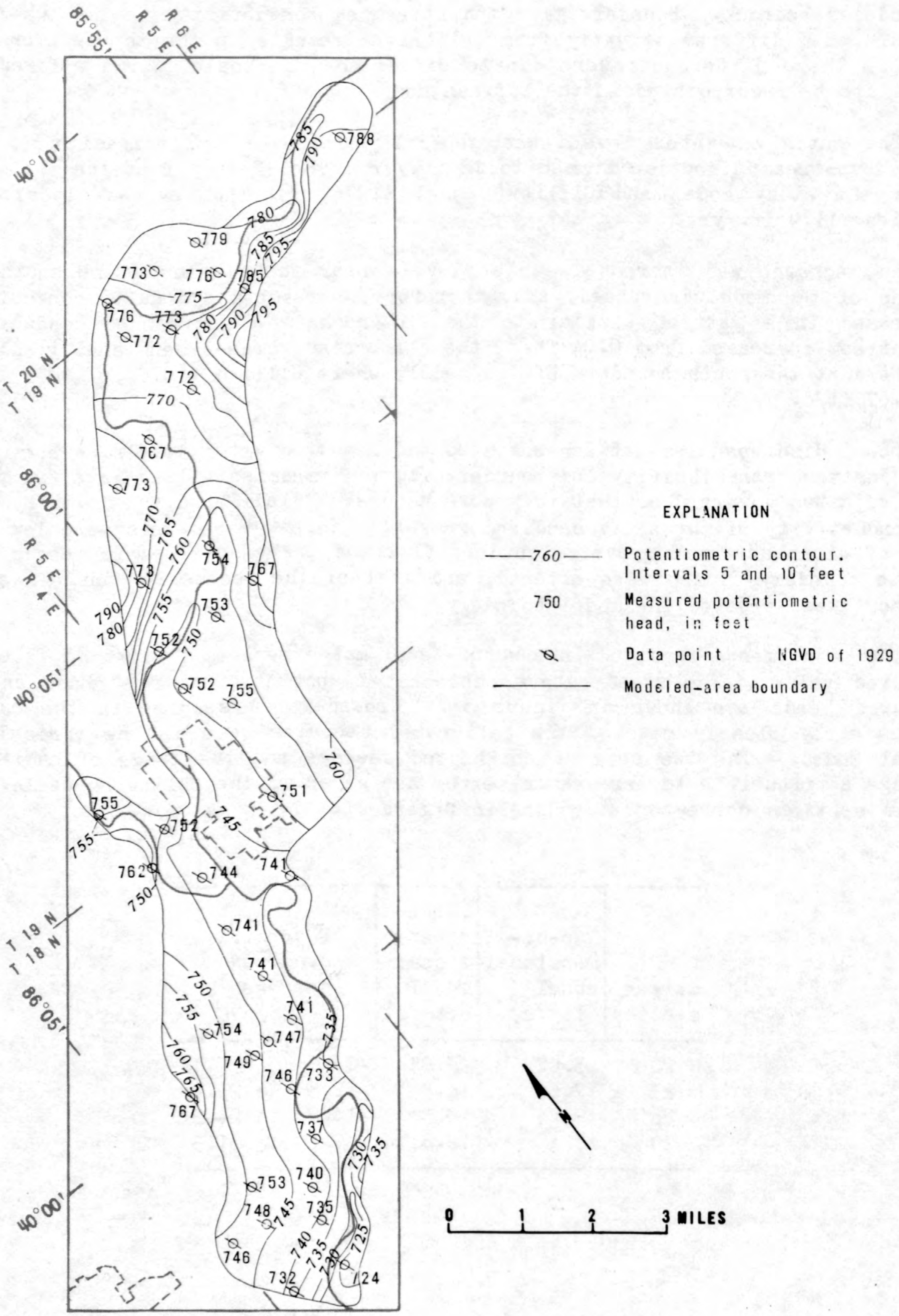


Figure 31.-- Model-simulated steady-state potentiometric surface of the outwash aquifer.

Finally, a ground-water budget that quantifies the distribution of ground-water flow is given in table 3.

Table 3.--Steady-state water budget for the two-dimensional model

Recharge	ft^3/s	Discharge	ft^3/s
Precipitation	21.5	Ground-water pumpage	3.91
Ground-water underflow into study area	22.5	Ground-water discharge to streams	39.5
		Ground-water underflow out of study area	.65
Total recharge	44.0	Total discharge	44.1

Simulation of Additional Ground-Water Pumpage

One of the objectives of this study was to analyze the ground-water pumpage available from selected pumping plans. The three-dimensional model simulated pumpage in the confined sand and gravel and the bedrock aquifers, and the two-dimensional model simulated pumpage in the outwash aquifer. The pumpage, streamflow depletion, and water-level declines associated with each pumping plan are discussed and compared in the next two sections of the report.

Pumpage from the Confined Sand and Gravel and the Bedrock Aquifers

One approach to evaluating pumping from aquifers in the drift and the bedrock is to develop several pumping plans that individually cover the entire study area. Yields from the plans can be compared to determine which plan gives the maximum yield from the system. However, this approach has disadvantages. Because the area of the multilayered aquifer system is about 520 mi^2, several possible combinations of pumping sites could be developed, and many plans would have to be tested. Also, pumping wells throughout the study area may not be practical or of interest to an individual water user.

The approach that was used in pumping simulations was to develop pumping plans in areas having high yields and to pump specific areas in the bedrock and the sand and gravel aquifers. Pumpage, drawdown, and streamflow depletion were observed for four pumping plans under controlled conditions. This information indicated how much pumpage a given area or an aquifer could yield and how one pumping plan compared with another. Therefore, on the basis of these pumping plans, the model user can devise a pumping plan that fits a particular need and can estimate the hydologic response to that plan. In addition to plans describing the response to typical pumping, the possibility of increasing pumping near Westfield was also investigated as a fifth plan.

Pumping plans were evaluated at steady state to observe the maximum effects of time on streamflow depletion and drawdown. The precise time of the maximum effects was not determined, but experience at a well field near Anderson provides some information on this subject. After a month of operation, pumpage from the well field had reached 4 Mgal/d. According to a study done for the well-field operator, water levels from this pumpage stabilized within about 2 years (Donald Davis, well-field operator, oral commun., 1981).

Most of the pumping plans included separate simulations of constant-head and constant-flux boundaries. However, the difference in system reponse to the two boundaries was insignificant because the distance between the simulated pumpage and the model boundary minimized boundary effects. Therefore, only the constant-flux-boundary drawdowns are illustrated.

To minimize the variation of factors, the author used an average drawdown of 20 ft in the pumping node in the first four pumpage simulations. In this way, the risk of pumping domestic wells dry is lessened. Also, an identical number of equal-sized nodes was used in each pumping plan so that areas of drawdown would be equal. Three of the first four plans simulated pumping in one general area, and two plans simulated pumping in adjacent nodes.

Locations of all the pumping plans are shown in figure 32. Plan A evaluated pumping from the bedrock aquifer (layer 1 in fig. 23) in an area of high transmissivity. Plan B evaluated pumping from an areally extensive and thick sand and gravel aquifer (aquifer 1). Plan C evaluated pumping from three vertically adjacent aquifers, aquifers 1, 2, and 3 (layers 2, 3, and 4 in fig. 23) at the same location as plan B. Plan C was used to determine whether pumping from a well in each of three aquifers at one location yielded more water than pumping from one aquifer in three locations (plan B). Plan D evaluated the effects of pumping near a reservoir. Pumping in this plan is from aquifer 1, which extends beneath the reservoir. A confining till lies between the aquifer and the reservoir. Two of the widely separated pumping nodes are at real pumping centers. All the pumping nodes of plan D represent pumpage increases from possible increased development around Morse Reservoir. Plan E doubles the Westfield pumpage (pumping from aquifer 1 in an area southeast of the town). The results of simulated pumpage from plans A-D are presented in four drawdown maps (figs. 33-36) and table 4. Each map shows the aquifer that is pumped.

Each of the four pumping plans affected the ground-water flow system differently. Plan A pumped 4.3 Mgal/d from the bedrock. About three-fourths of this pumpage was ground water that normally discharges into Cicero Creek (Morse

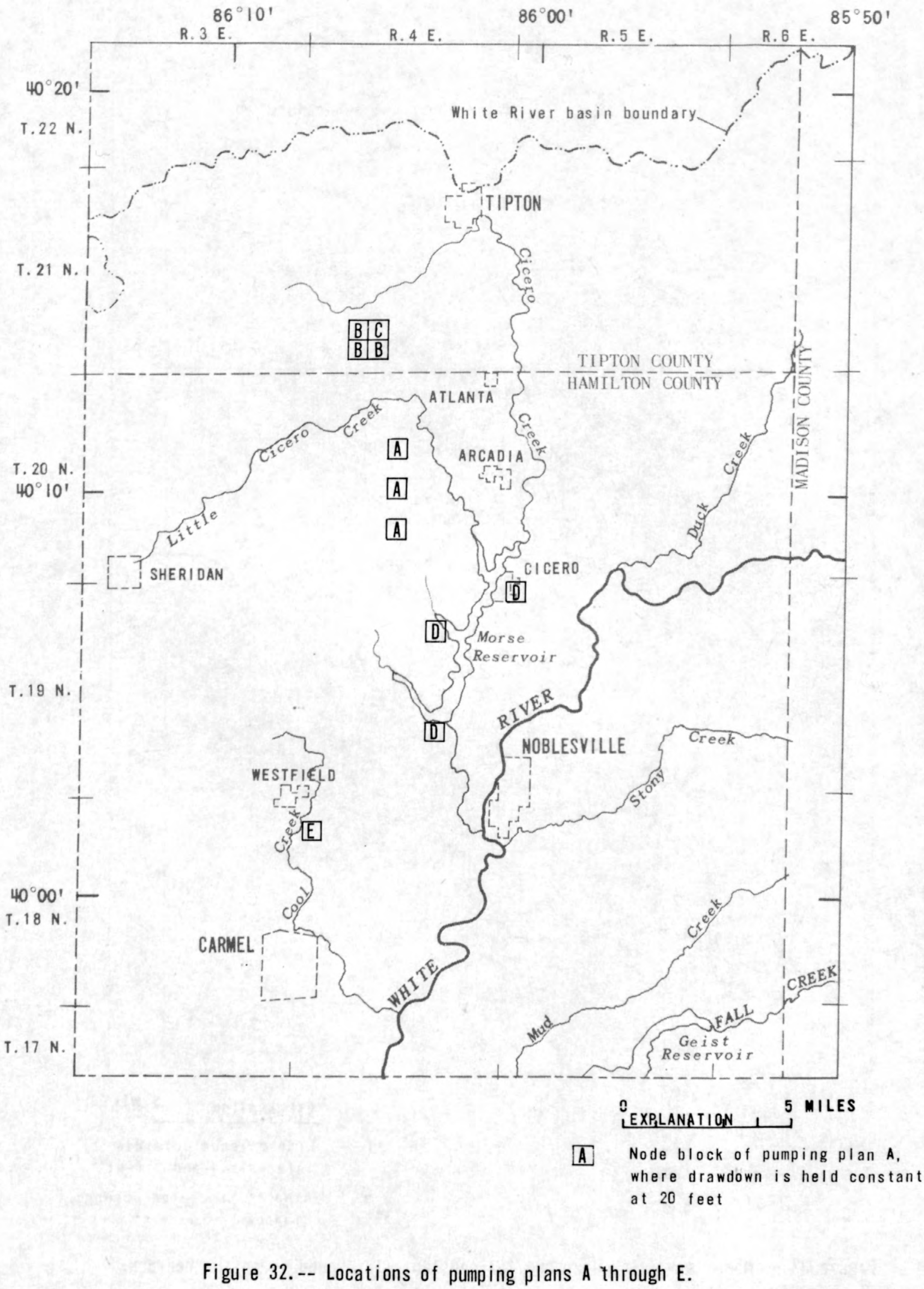


Figure 32.-- Locations of pumping plans A through E.

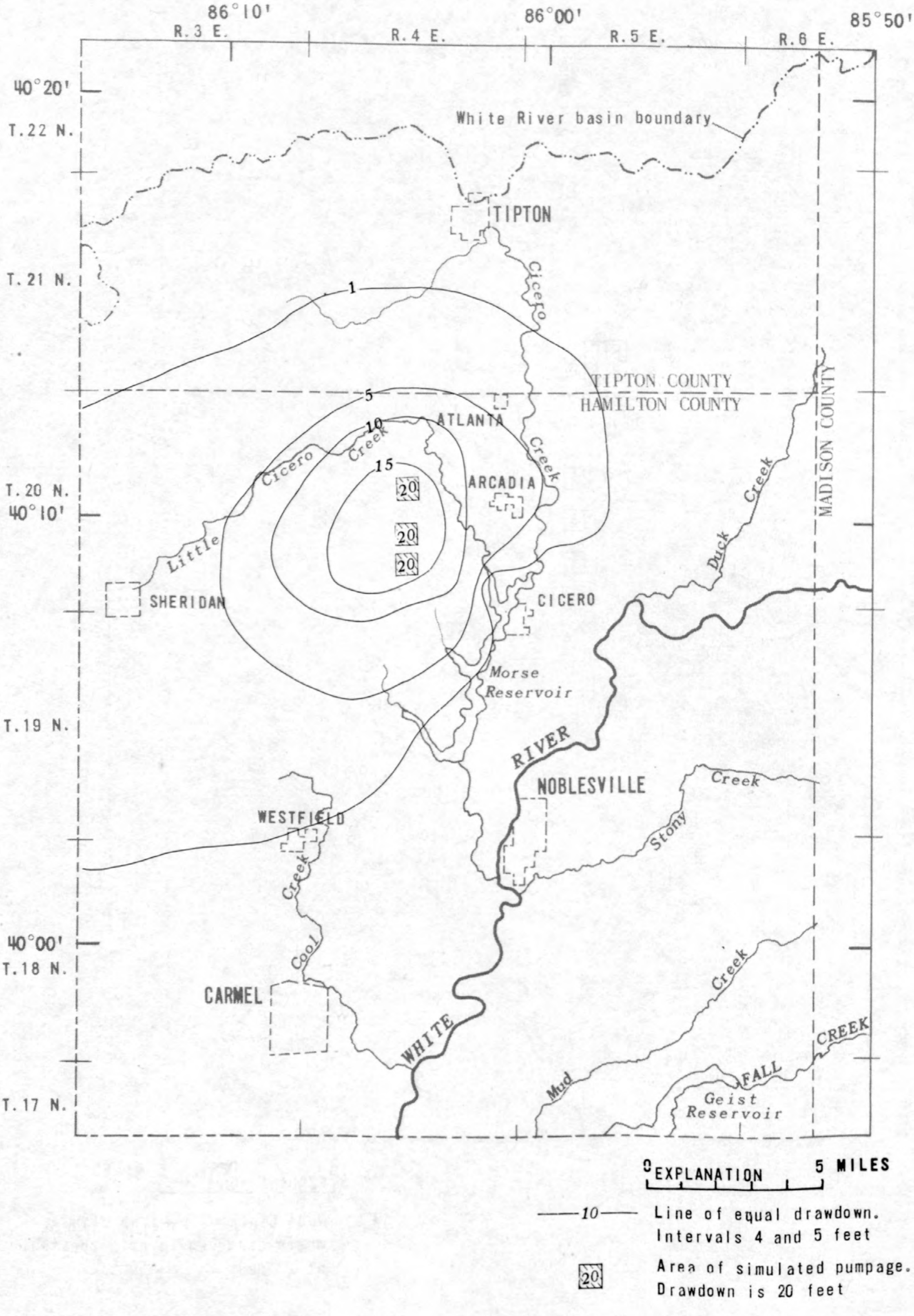


Figure 33.-- Model-simulated drawdown for constant-flux boundary in the bedrock aquifer, for pumping plan A.

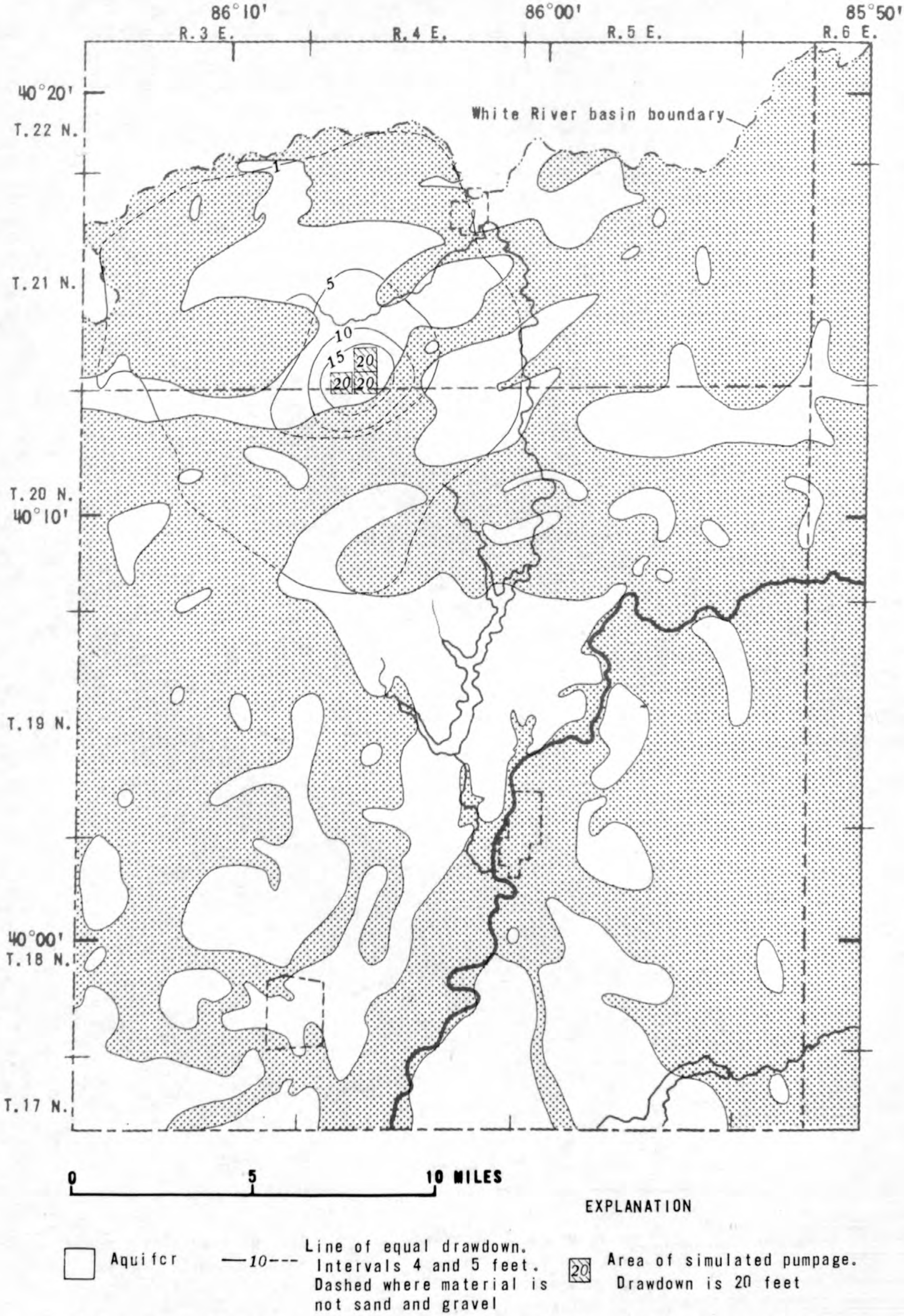
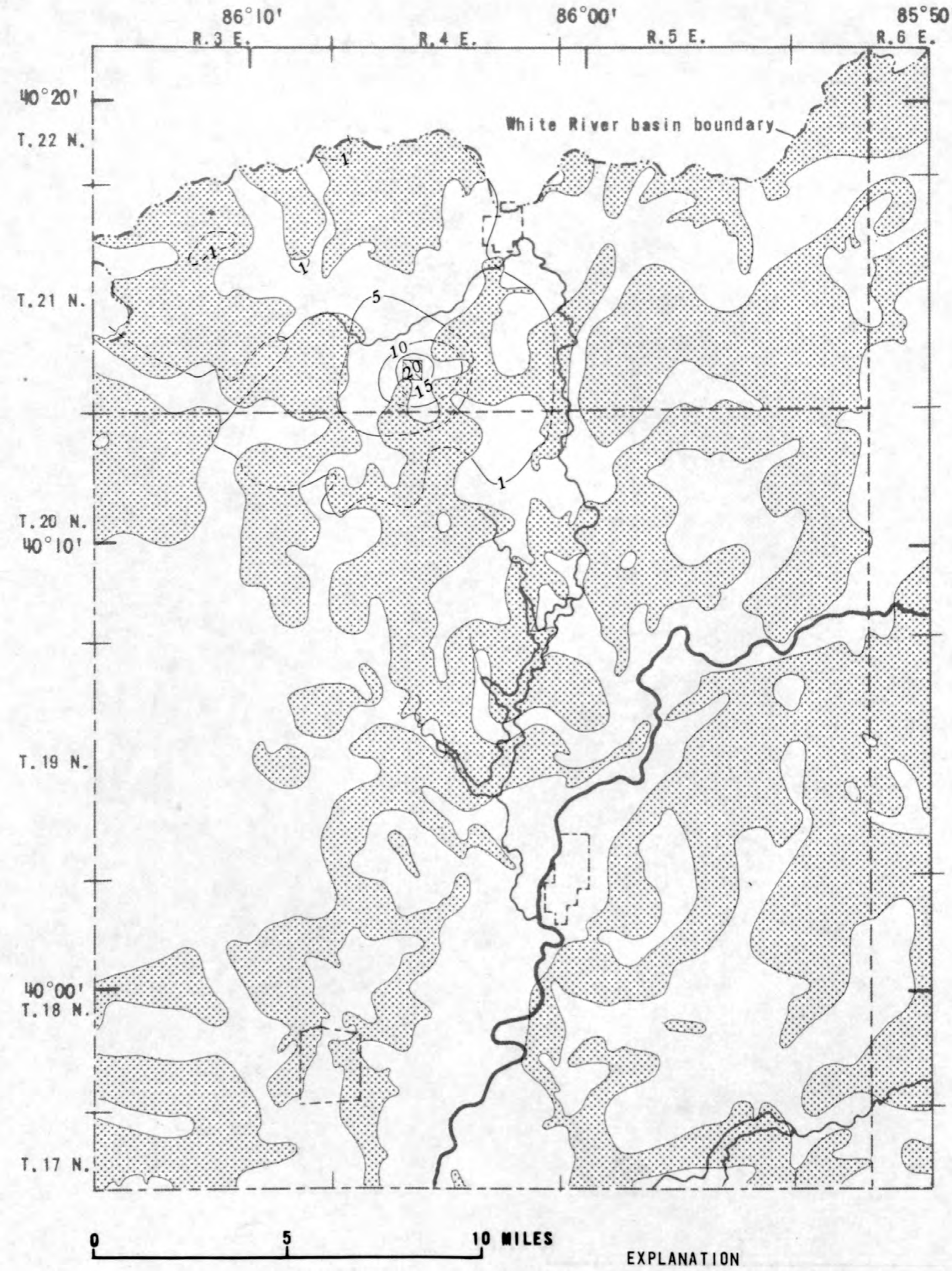


Figure 34.-- Model-simulated drawdown for constant-flux boundary in aquifer 1, for pumping plan B.



EXPLANATION

Aquifer

—10— — — Line of equal drawdown. Intervals 4 and 5 feet. Dashed where material is not sand and gravel

20 Area of simulated pumpage. Drawdown is 20 feet

Figure 35.-- Model-simulated drawdown for constant-flux boundary in aquifer 2, for pumping plan C.

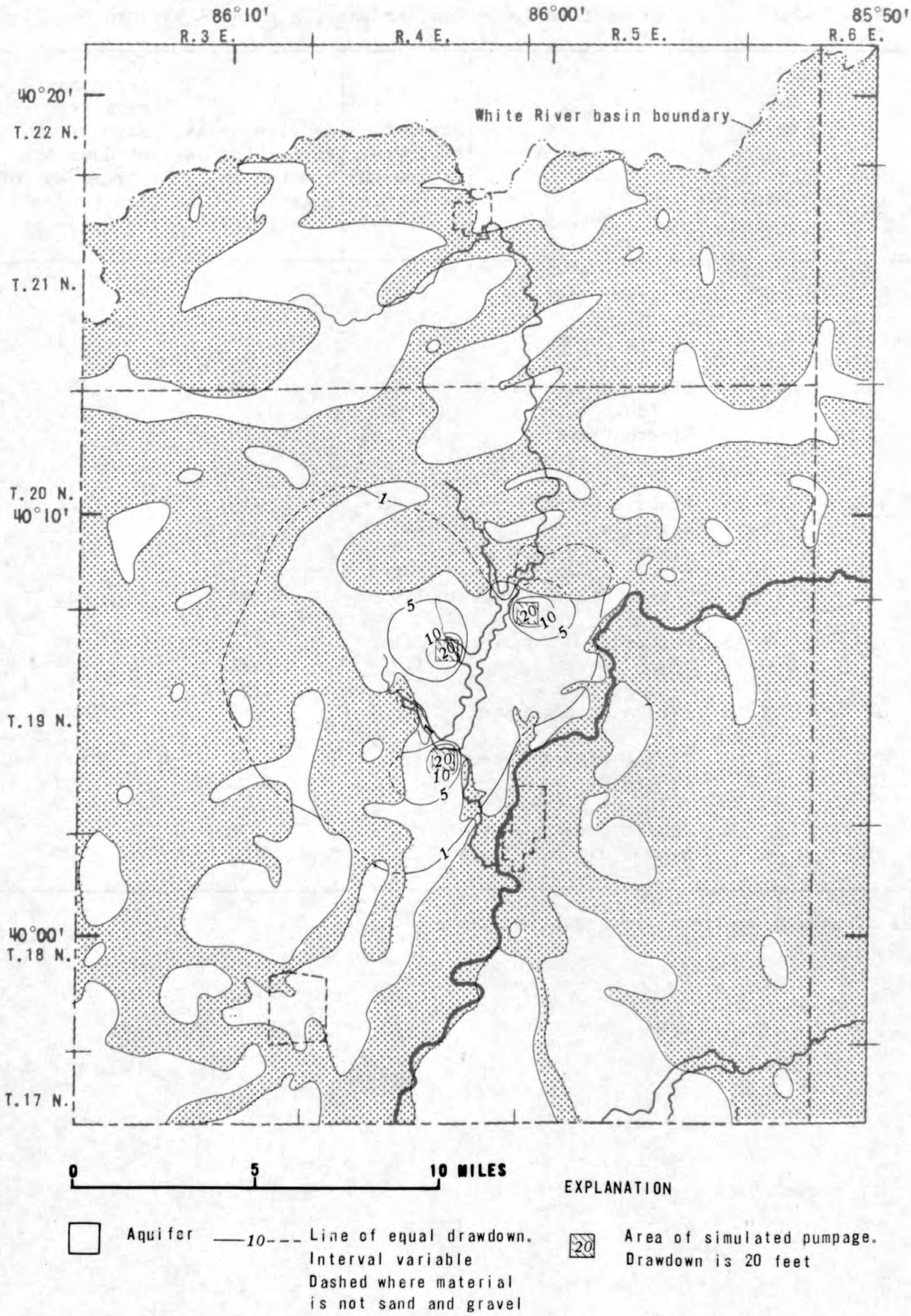


Figure 36.-- Model-simulated drawdown for constant-flux boundary in aquifer 1, for pumping plan D.

Table 4.--Pumpage and streamflow depletion for pumping plans A through D

Pumping plan	Pumpage (Mgal/d) / (ft3/s)	Stream sections most affected by pumping[1]	Stream	Streamflow at downstream end of section (ft^3/s)	Model-simulated streamflow depletion in section (ft^3/s)	Approximate reduction in streamflow at downstream end of section (percent)
A	4.3 / 6.6	6	Prairie Creek	0.42	0.10	24
		7	Cox ditch	1.1	.13	11
		9	Cicero Creek	4.5	.18	4
		11	do.	8.6	.45	5
		12	Little Cicero Creek	.44	.13	30
		13	do.	1.3	1.5	100
		14	Cicero Creek (Morse Reservoir)	12	3.2	27
B	2.3 / 3.6	6	Prairie Creek	.42	.10	24
		7	Cox ditch	1.1	.50	44
		8	Dixon Creek	1.7	.50	29
		9	Cicero Creek	3.6	.93	26
		13	Little Cicero Creek	1.5	.70	47
C	1.9 / 2.9	7	Cox ditch	1.1	.39	35
		8	Dixon Creek	1.7	.44	25
		9	Cicero Creek	3.9	.75	19
		13	Little Cicero Creek	1.5	.56	37
D	6.7 / 10	13	Little Cicero Creek	1.5	.10	7
		14	Cicero Creek (Morse Reservoir)	14	8.0	57
		17	Cicero Creek	8.2	.38	5
		26	Vestal ditch	.10	.10	100

[1]Locations of stream sections are shown in figure 21.

Reservoir) and Little Cicero Creek stream sections. Plan A induced ground-water recharge from both sections. However, the reservoir limited the spread of drawdown. Plan B pumped about 2.6 Mgal/d from an areally extensive aquifer. The cone of depression of plan B was smaller than that of plan A, and the plan did not induce recharge from streams. Plan C pumped 1.9 Mgal/d from three vertically adjacent aquifers. Its pumpage was the least of plans A through D, and its areal drawdowns and streamflow depletions were smaller as well. The low pumpage for plan C is related to well placement. The three superimposed wells of plan C would interfere more among themselves than with the horizontally separated wells of plan B. Therefore, for equal drawdowns, less pumpage could be developed in plan C than in plan B. Pumping plan D near the reservoir, was the most successful. The 6.7-Mgal/d pumpage of this plan caused the least drawdown of the four plans. The success of this plan is explained by the proximity of the pumping to a source of recharge in an aquifer that extends beneath the recharge source. Drawdowns were less than 10 ft within 1 mi of the pumping center because of the proximity of Morse Reservoir and because the pumping centers were widely separated. Seventy-eight percent of the pumpage is from the combination of induced recharge and diversion of water discharging to the reservoir. The simulated pumpage is 6.4 Mgal/d more than the pumpage in 1977 around Morse Reservoir.

All the pumping plans can produce more water than any one community was using in 1977. However, the plans differ in the amount that can be pumped and the effect on streams and drawdown. The volume of water pumped from an aquifer for a given drawdown is related to the hydraulic connection between the aquifer and the area of discharge, a stream or a reservoir. If an aquifer is separated from a stream by till, horizontally and vertically, then pumpage and streamflow depletion are smaller than they would be if till did not separate the stream and aquifer. This condition is exibited in plans B and C. Where pumpage is from high-transmissivity bedrock, the aquifer is only vertically separated from a stream by till. Vertical flow through till is not as limiting as horizontal flow because of the greater area where vertical flow is possible and because of the greater vertical gradient that can be developed. Therefore, plan A, compared with other plans, developed a high pumpage and even caused recharge from the reservoir. Plan D had the best of all conditions in that pumping was from a sand and gravel aquifer that extended beneath the reservoir, the pumping centers were widely separated, and pumping was near a source of recharge.

Results from the pumping plans are similar to those in other study areas of the project area. Simulated pumpage for plans in all study areas generally ranged from 1 to 5 ft^3/s (0.6 to 3.3 Mgal/d). Most of the 5-ft drawdown contours are about 5 mi in diameter. As a result of the simulated pumpages, flow reduction is usually no more than 10 percent in streams discharging more than 2 ft^3/s.

A fifth pumping plan investigated the effects of doubling the pumpage at Westfield by adding another well in aquifer 1 southeast of town (plan E, fig. 32) and pumping 0.18 Mgal/d. The additional pumpage caused more than a 1-ft drawdown in 1.3 mi^2 and diverted 0.22 ft^3/s (0.15 Mgal/d) from Cool Creek. Drawdown was 4.5 ft in a simulated pumping well gravel packed to a radius of 3 ft. Well logs used in the study showed aquifer 1 to be as thick as 40 ft in the pumped area.

Pumpage from the Outwash Aquifer

The potential of the outwash as a water supply is greater than that of either the confined sand and gravel or the bedrock aquifers. Two major reasons for the high potential are the high transmissivities, many near 20,000 ft^2/d, and a nearby major discharge area, the White River. Because of the high potential and the small area of the aquifer, the entire outwash aquifer was analyzed for pumping sites to determine well yield.

Two of the conditions applied during the pumping simulation were a steady-state system, which achieved the maximum effect from time on the flow system, and a constant ground-water-flux boundary. This boundary prevented an increase in flow and also maximized stream depletion and drawdown responses.

Other conditions imposed were similar to those used by Gillies (1976, p. 22) and Meyer and others (1975, p. 55-56). Because of similarities in pumping simulations, responses of ground-water systems in the Hamilton County study area and nearby areas can be compared. Pumping wells were simulated in areas of high transmissivity and parallel to the White River to minimize drawdown. Pumping was increased until drawdown in the well equaled 50 percent of the saturated thickness of the original aquifer at the well. First estimates of pumpage were made for each well, and pumpages were balanced by a trial-and-error process until the 50-percent drawdown criterion was reasonably met.

As a result of the outwash pumping plan, 39 Mgal/d was obtained from the outwash. Individual well pumpage generally ranged from 315 to 1,250 gal/min and averaged 860 gal/min. This pumpage caused a 39-Mgal/d decrease in streamflow, which consists of a 14-Mgal/d induced recharge from the White River and a 25-Mgal/d capture of water enroute to the river. Streamflow reduction could be less, depending on how much of the pumped water is returned to the river. The drawdowns caused by the pumping are shown in figure 37, where drawdowns near the well are as much as 50 ft near 146th Street. The associated water-level surface caused by the pumping is shown in figure 38, and the water budget for the pumping simulation is given in table 5.

The results in table 5 are based on a constant-flux boundary, which does not allow additional flux across the model boundary in response to pumpage. However, additional flux can be induced from discharge areas outside the modeled area. True drawdowns and inducement from the White River would be less and pumpage more than those reported in the table. The results presented in the illustrations and the table represent an upper limit of drawdown and stream loss.

Results from Gillies (1976, p. 22) indicated that 21.3 Mgal/d of water could be pumped if several wells were pumped down to 50 percent of the original saturated thickness in the well. This pumpage is smaller than that of the Hamilton County study. Drawdowns near pumping wells were only 35 ft, and flow induced from the river was only 0.68 Mgal/d in the study by Gillies (1976). If the 14-Mgal/d induced flow predicted by the Hamilton County study were reduced to 2.3 Mgal/d, as in one of the trial-and-error simulations, 28 Mgal/d pumpage could be

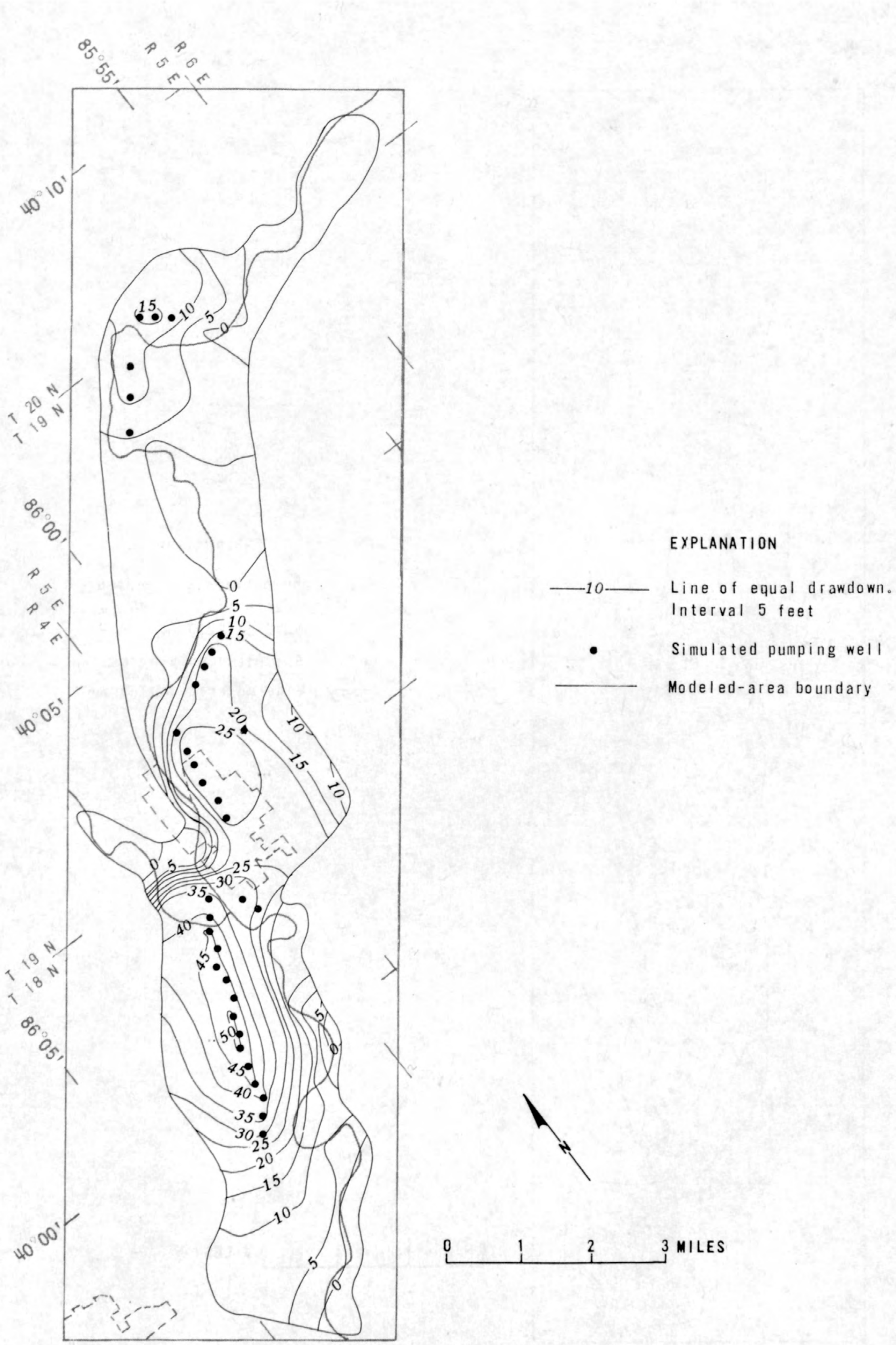


Figure 37.-- Model-simulated drawdown for constant-flux boundary in the outwash aquifer.

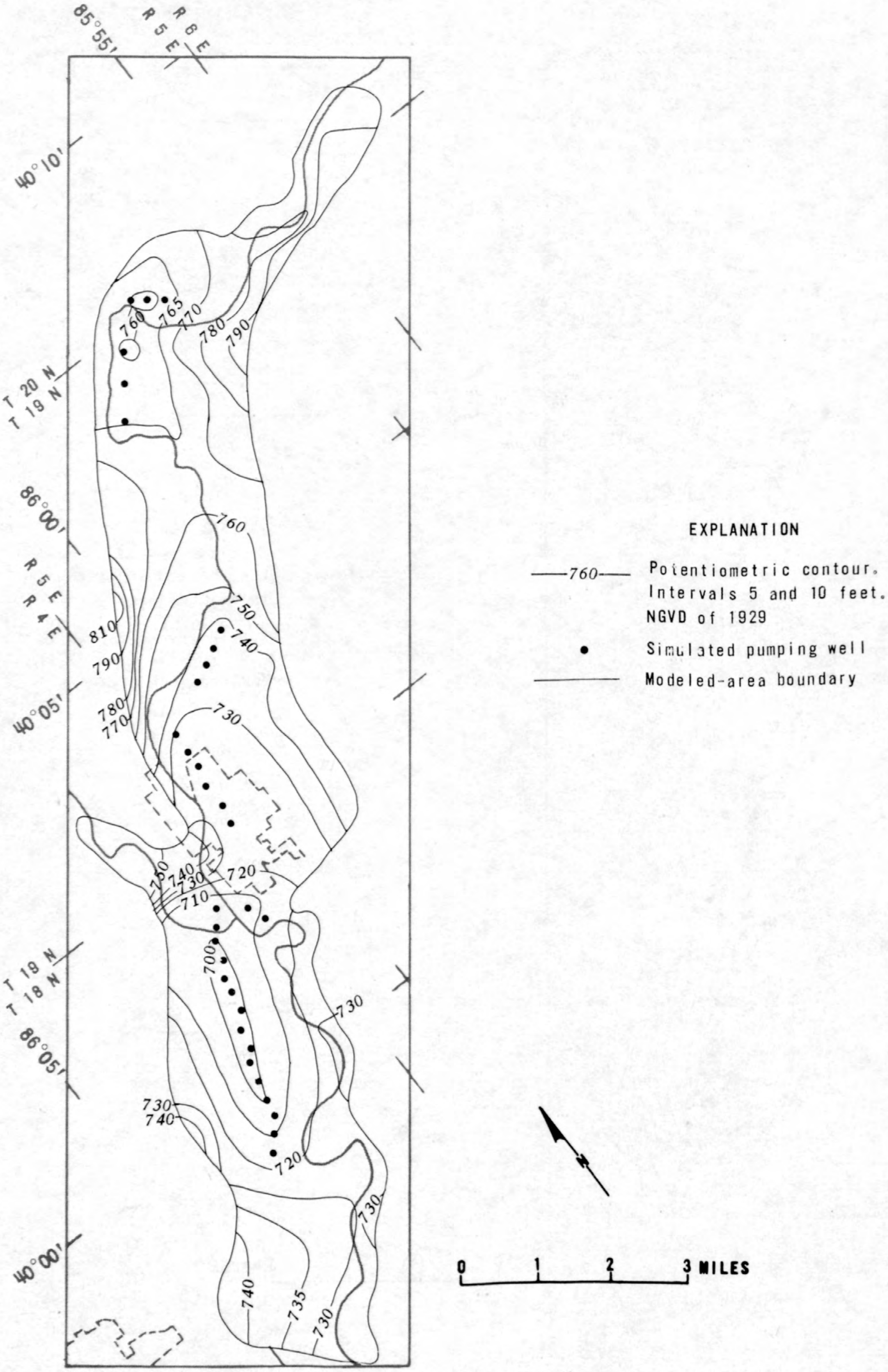


Figure 38.-- Model-simulated steady-state potentiometric surface from pumping the outwash aquifer.

Table 5.--Steady-state water budget for pumpage from the outwash aquifer[1]

Recharge	ft^3/s	Discharge	ft^3/s
Precipitation	21.4	Ground-water pumpage	64.6
Ground-water underflow into study area	22.5	Ground-water underflow out of study area	.65
Ground-water recharge from streams	21.2		
Total recharge	65.1	Total discharge	65.2

[1]Constant-flux boundary.

pumped. The reasons for the higher pumpage of the Hamilton County study, even with little induced flow, are that about 10 more wells were pumping in the simulation and that the wells were pumping in an area greater than that in Gillies' study (1976).

SUMMARY

The Hamilton and Tipton Counties study area of the White River basin contains five confined sand and gravel aquifers, a large outwash aquifer associated with the White River, and a bedrock aquifer. Thickness of the thin and discontinuous confined aquifers generally ranges from 5 to 20 ft. Vertical confining beds of till separate these aquifers. The average saturated thickness of the south half of the outwash aquifer is 70 ft, and the width is 2 to 3 mi. The north half of the outwash aquifer thins both vertically and horizontally at different points, averages about 1 mi in width, and generally ranges from 20 to 80 ft in saturated thickness.

Pumping simulations for the three-dimensional model were chosen so that the most favorable locations were stressed in the confined sand and gravel and bedrock aquifers. The two types of aquifers were chosen for pumping in specific locations near a major stream. Simulated pumping ranged from 1.8 to 6.7 Mgal/d. The best pumping plan tested was the one near Morse Reservoir in an aquifer underlying the reservoir. The 6.7-Mgal/d pumping from that plan caused drawdowns generally less than 10 ft within 1 mile of the pumping. The least productive pumping plans were those pumping from aquifers that did not extend to or beneath a major stream or a reservoir but, rather, pumped water near smaller streams. Two pumping plans having these characteristics had pumpages of 1.8 and 2.3 Mgal/d. Although these plans had lower simulated pumpages than the other plans, the spread of the cone of depression and streamflow depletion were less also.

One pumping plan in the confined sand and gravel doubled Westfield's current pumpage in aquifer 1 southeast of town. A 3-ft-radius well pumped 0.18 Mgal/d, caused 4.5 ft of drawdown, and diverted 0.22 ft^3/s from Cool Creek.

With the constant-flux boundary, simulation of 39-Mgal/d pumpage from the outwash aquifer induced 14-Mgal/d recharge from the White River and caused as much as 50 ft of drawdown near some pumping wells. The 39 Mgal/d resulted from pumping until the saturated thickness at the pumping wells was reduced by half. Pumpage at individual wells ranged from 315 to 1,250 gal/min and averaged 860 gal/min.

A three-dimensional model of confined sand and gravel and bedrock aquifers and a two-dimensional model of outwash simulated the effects of pumping on the ground-water systems. Computed water levels were generally within 10 ft of measured levels in the three-dimensional model and within 3 ft in the two-dimensional model. The three-dimensional model was used to investigate pumping potential in the confined sand and gravel and bedrock, whereas the more detailed two-dimensional model was used to investigate the potential of the outwash aquifer. Model-parameter values determined during the three-dimensional model calibration and used in the two-dimensional model calibration required little change except for an increase in outwash recharge from 6 in. in the three-dimensional model to either 8 or 12 in. in the two-dimensional model. This increase enabled the two-dimensional model to make up for the flow contribution from bedrock, which was not modeled.

The characteristics of the ground-water flow system discussed in the Hamilton County study are hydraulic conductivity, vertical hydraulic conductivity, transmissivity, and streambed hydraulic conductivity. Data from 13 specific-capacity tests in adjoining Madison County were used to estimate a hydraulic conductivity of 433 ft/d for the confined sand and gravel aquifers. However, during model calibration, a lower conductivity, 216 ft/d, produced closer agreement between measured and model-simulated conductivities. On the basis of a 216-ft/d hydraulic conductivity, the author estimated that the transmissivity for the confined aquifers ranges from 1,000 to 20,000 ft^2/d. Hydraulic conductivity of the outwash for the two-dimensional model was determined in a manner similar to that in previous studies of the outwash aquifer. The 216-ft/d hydraulic conductivity used for the confined aquifer is within the range of conductivity used for the outwash aquifer. Outwash transmissivity was estimated to range from about 1,000 to 28,000 ft^2/d. Bedrock transmissivity, determined by specific-capacity tests and model calibration, generally ranges from 500 to 10,000 ft^2/d. The vertical hydraulic conductivity of the confining beds ranges from 7×10^{-4} to 7×10^{-2} ft/d and averages 7×10^{-3} ft/d. Streambed hydraulic conductivity generally ranges from 4×10^{-2} to 4.0, and more than one-third of them were near 4.0. The White River streambed conductivity ranged from 0.04 to 4.0 ft/d.

Water levels in and near the study area indicate that the ground-water flow system is in steady state. However, a yearly cycle of water-level fluctuations is caused by annual fluctuations in evapotranspiration. Otherwise, yearly inflows and outflows of the system are in balance. Precipitation recharge for the till ranged from 2.0 to 4.5 in./yr. Precipitation recharge for the outwash was 6.0 in./yr. The sum of municipal and industrial pumpage during 1977 was about 6 Mgal/d. Nearly all this pumpage was municipal, which may, in turn, supply local industries. Discharge into streams totaled about 70 ft^3/s October 30, 1977.

REFERENCES

Brown, R. H., 1963, Estimating the transmissibility of an artesian aquifer from the specific capacity of a well, in Bental, Ray, Methods of determining permeability, transmissibility, and drawdown: U.S. Geological Survey Water-Supply Paper 1536-I, p. 336-338.

Butler, S. S., 1957, Engineering Hydrology: Englewood Cliffs, N.J., Prentice-Hall, Inc., 356 p.

Cable, L. W., Daniel, J.F., Wolf, R. J., and Tate, C. H., 1971, Water resources of the Upper White River basin, east-central Indiana: U.S. Geological Survey Water-Supply Paper 1999-C, 38 p.

Capps, S. R., 1910, The underground waters of north-central Indiana: U.S. Geological Survey Water-Supply Paper 254, 267 p.

Gillies, D. C., 1976, Availability of ground water near Carmel, Hamilton County, Indiana: U.S. Geological Survey Water-Resources Investigations 76-46, 27 p.

Harrell, M. A., 1935, Ground water in Indiana: Indiana Department of Public Works, Conservation Department, Division of Geology, Publication no. 133, 504 p.

Herring, W. C., 1971, Water resources of Hamilton County with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water.

Horner, R. G., 1976, Statistical summaries of Indiana streamflow data: U.S. Geological Survey Water-Resources Investigations Open-File Report 35-75, 526 p.

Indiana Geological Survey, 1956, Atlas of mineral resources of Indiana, map no. 9: Indiana Department of Conservation, Geological Survey.

Lapham, W. W., 1981, Ground-water resources of the White River basin, Madison County, Indiana: U.S. Geological Survey Water-Resources Investigations 81-35, 112 p.

Maclay, R. W., and Heisel, J. E., 1972, Electric analog model study of the Upper White River basin, Indiana: U.S. Geological Survey open-file report, 27 p.

Meyer, William, Reussow, J. P., and Gillies, D. C., 1975, Availability of ground water in Marion County, Indiana, with a section on water quality by W. J. Shampine: U.S. Geological Survey Open-File Report 75-312, 87 p.

National Oceanic and Atmospheric Administration, 1956-77, Climatological data annual summaries, Indiana: Asheville, N.C., National Oceanic and Atmospheric Administration, Environmental Data Service, v. 61-80, 82.

Steen, W. J., 1968, Water resources of Tipton County with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water.

Trescott, P. C., 1975, Documentation of finite difference model for simulation of three-dimensional ground-water flow: U.S. Geological Survey Open-File Report 75-438, 102 p.

Trescott, P. C., Pinder, G. F., and Larson, S. P., 1976, Finite difference model for aquifer simulation in two dimensions with results of numerical experiments: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 7, Chapter C1, 116 p.

U.S. Department of Commerce, 1976, Current population report: Bureau of the Census, Series P-25, no. 753, 30 p.

U.S. Geological Survey, 1977, Water-resources data for Indiana, 1976: U.S. Geological Survey, Water-Data Report IN-76-1, 357 p.

Wayne, W. J., 1956, Thickness of drift and bedrock physiography of Indiana north of the Wisconsin glacial boundary: Indiana Department of Conservation, Geological Survey Report of Progress no. 7, 69 p.

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