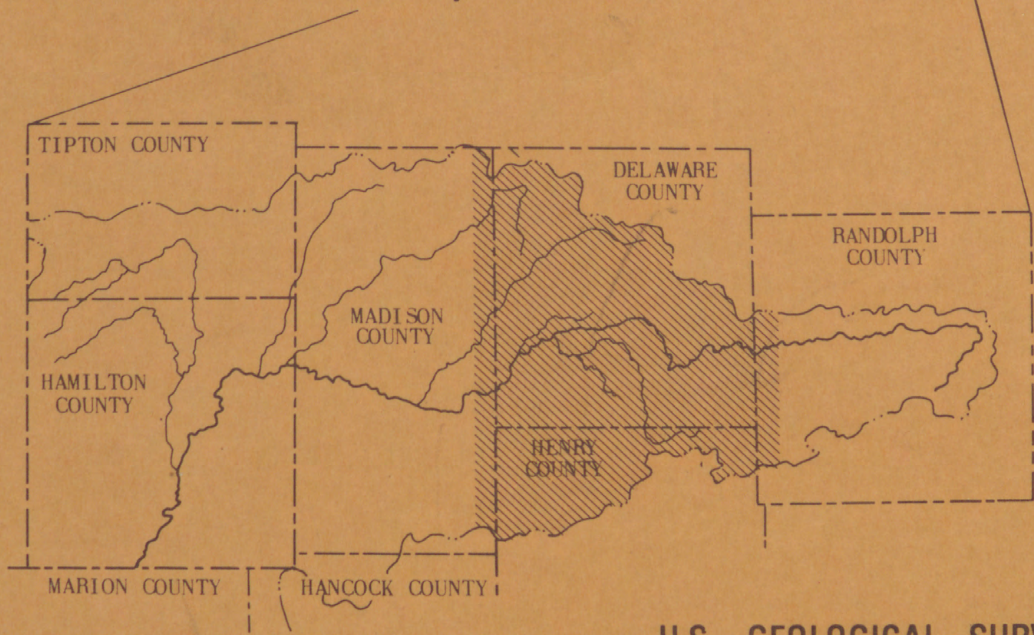
GROUND-WATER RESOURCES OF THE WHITE RIVER BASIN, DELAWARE COUNTY, INDIANA



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WATER-RESOURCES INVESTIGATIONS 82-47

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The ground-water resources of the White River basin in and near Delaware County, Ind., were investigated by mapping the aquifers, calculating their hydraulic properties, determining the distribution of potentiometric head in the aquifers, and determining some of the components of the ground-water budget from data collected in the field. This information was used to construct and calibrate a seven-layer digital ground-water flow model. The model was used to assess ground-water potential in terms of yield, drawdown and depletion in streamflow.

Six confined sand and gravel aquifers interbedded in the drift, an unconfined outwash aquifer, and a bedrock aquifer are the most important aquifer systems in the study area.

Simulations of seven pumping plans provide a general assessment of the water-yielding potential of the three major aquifer systems. Model results indicate that as much as 3 million gallons per day can be developed from well fields where the average drawdown is 20 feet. Model simulations also indicate that 7 million gallons per day is available from a potential well field around Muncie.

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UNITED STATES DEPARTMENT OF THE INTERIOR

JAMES G. WATT, Secretary

GEOLOGICAL SURVEY

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Factors for converting inch-pound units used in this report to international system of units (SI)

Multiply inch-pound unit	By	To obtain SI unit
inch (in.)	25.40	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
foot per day (ft/d)	0.3048	meter per day (m/d)
square foot per day (ft^2/d)	0.0929	square meter per day (m^2/d)
foot per year (ft/yr)	0.3048	meter per year (m/a)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
million gallons per day (Mgal/d)	0.0438	cubic meter per second (m^3/s)

Datum

National Geodetic Vertical Datum of 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "mean sea level."

GROUND-WATER RESOURCES OF THE WHITE RIVER BASIN,

DELAWARE COUNTY, INDIANA

By Leslie D. Arihood and Wayne W. Lapham

ABSTRACT

The ground-water resources of the White River basin in and near Delaware County, Indiana, were investigated by mapping the aquifers, calculating their hydraulic properties, determining the distribution of potentiometric head, and determining some of the components of the ground-water budget from data collected in the field. This informaton was used to construct and calibrate a seven-layer, digital ground-water-flow model. The model, constructed and calibrated to water-level and seepage data collected during the study, simulates conditions for the autumn of 1977. The model was used to assess ground-water potential in terms of yield, drawdown, and streamflow depletion.

Drift covers nearly the entire area and generally ranges in thickness from 0 to 500 feet. Beneath the drift lie Silurian and Ordovician limestone, dolomite, and shale. A bedrock valley trends east-west through the center of the area. Bedrock surface relief is about 400 feet.

Six confined sand and gravel aquifers interbedded in the drift, a bedrock aquifer, and an unconfined outwash aquifer are the three major aquifer systems. The nearly horizontal, areally discontinuous, confined sand and gravel aquifers generally range in thickness from 5 to 40 feet and have an average hydraulic conductivity estimated to be 433 feet per day. The bedrock aquifer underlying all of Delaware County has a permeable thickness estimated to be 150 feet and an average transmissivity of 1,000 square feet per day.

Observation-well records indicate that water-level fluctuation is seasonal and has no long-term trends. Municipal and industrial ground-water pumpage for 1976 was 3.1 million gallons per day (4.8 cubic feet per second). On October 29, 1977, when the flow duration in the White River at Muncie was 80 percent, ground-water seepage to streams was 81 cubic feet per second. The water budget simulated in the model indicated that the rate of inflow to the ground-water system in the modeled area is 102 cubic feet per second: 80 percent from effective areal recharge of precipitation and 20 percent from ground-water flow across the area boundaries. Two percent of the ground-water discharge is pumpage, 66 percent is seepage to streams, and the remaining 32 percent is flow across the area boundaries.

Simulations of seven pumping plans provide a general assessment of the water-yielding potential of the three major aquifer systems. Model results indicate that as much as 3 million gallons per day can be developed from well fields where the average drawdown is 20 feet. Model simulations also indicate that 7 million gallons per day is available from a potential well field around Muncie.

INTRODUCTION

Purpose and Scope

In 1972, the U.S. Geological Survey, in cooperation with the Indiana Department of Natural Resources, began a 3-year study of the ground-water resources of the White River basin in Marion County, Ind. The objectives of that study were to determine (1) the quantity of ground water that could be pumped, and (2) 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 in 1975, a similar cooperative study of the rest of the White River basin upstream from Marion County was started in July 1975. The objective of that study was to assess the ground-water resources in the White River basin upstream from Marion County. The assessment included (1) mapping the aquifers and calculating the hydraulic properties of the aquifers and confining beds, (2) determining the distribution of potentiometric head, (3) measuring ground-water discharge to streams, (4) using a multilayer digital model to determine the water budget of the study area, and (5) using the ground-water flow model to calculate the quantity of water that could be pumped without significant adverse effect on the ground-water system and streamflow.

Location and Setting

The project area, in central and east-central Indiana, covers the 1,500 mi^2 of the White River basin upstream from Marion County (fig. 1). The project area was divided into four study areas, each consisting of the principal county in that area. This division facilitated studying the project area in detail. The four study areas, by county name from west to east (fig. 1), are Hamilton, Madison (Lapham, 1981), Delaware, and Randolph. The subject of this report is the Delaware County study area shown in figure 2. Although parts of Henry, Madison, and Randolph Counties are also included it will be referred to as the Delaware County study area in the remainder of this report.

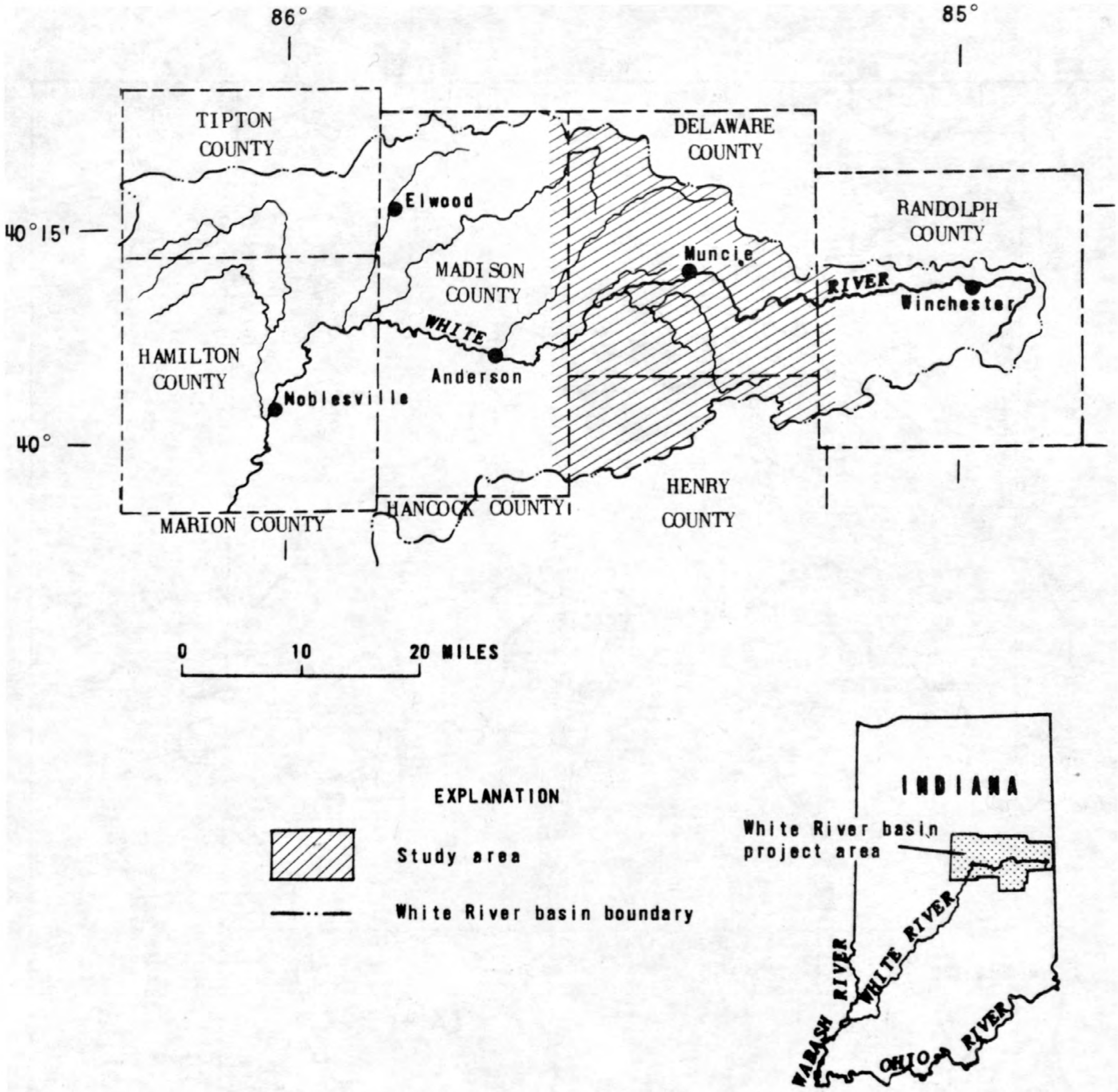


Figure 1.-- Location of Delaware County study area in the White River basin project area.

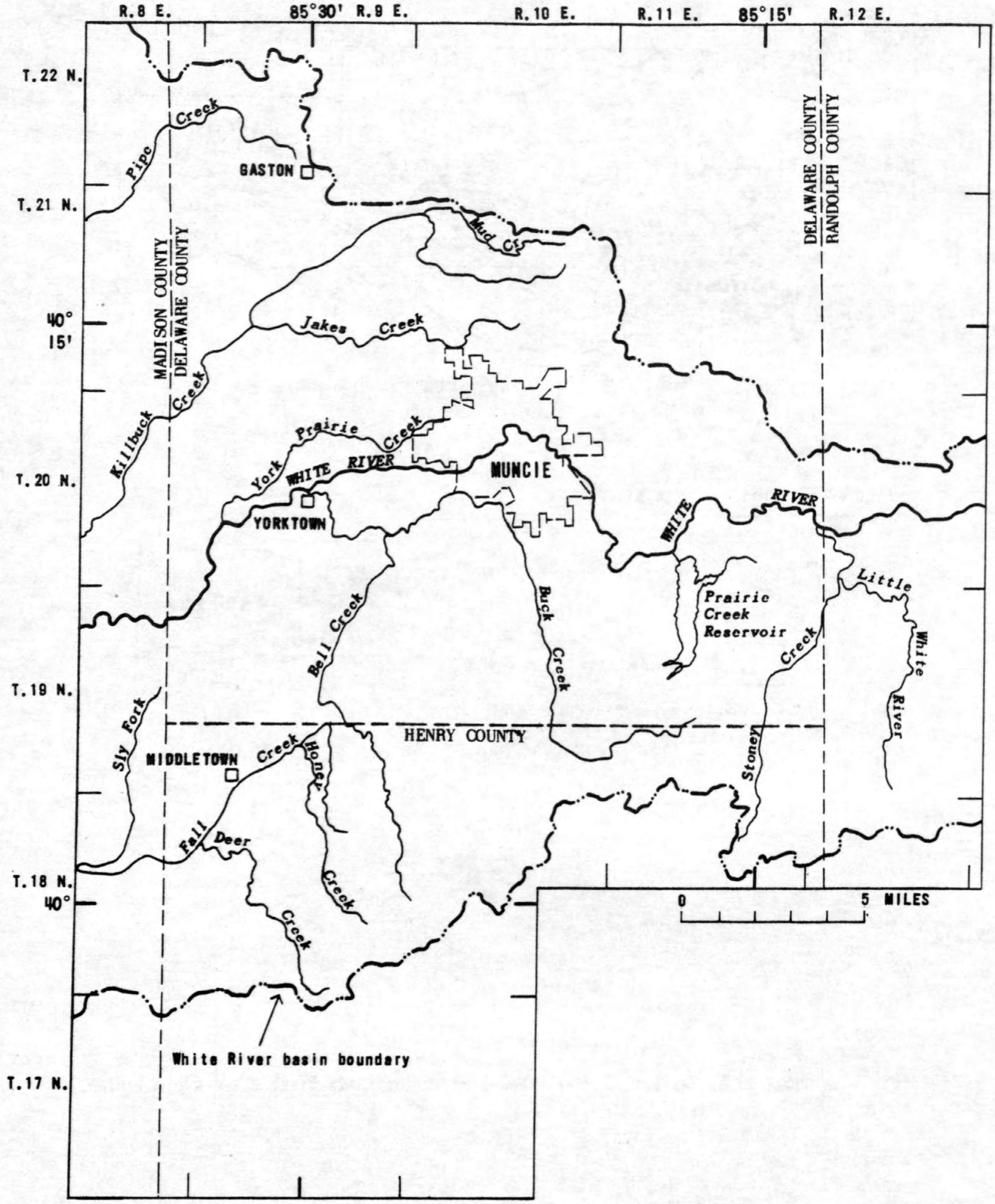


Figure 2.-- Delaware County study area.

The study area (fig. 2) includes about 480 mi^2, but only 270 mi^2 lies within Delaware County. Major cities and towns are Muncie, Yorktown, and Middletown, with populations of 77,216, 3,945, and 2,978, respectively (U.S. Department of Commerce, 1980, p. 28 and 31). Most of the land is used for agriculture, but some is used for industrial complexes, mainly in and near Muncie.

The climate is temperate. During 1941-70, average annual temperature at Muncie was 52.1° F (National Oceanic and Atmospheric Administration, 1941-70). During the same period, monthly average precipitation at Muncie ranged from 2.09 in. in February to 4.47 in. in June, and average annual precipitation was 38.2 in. (National Oceanic and Atmospheric Administration, 1977, p. 5).

The study area lies in the Tipton Till Plain of the Central Lowlands physiographic province (Wayne, 1956, p. 13, 14). Ground surface is flat to gently rolling, and most of the local relief is due to stream incisement. The altitude of the land surface generally ranges from 850 to 1,200 ft.

The largest river is the White River. Other significant streams are Fall, Stoney, Buck, and Killbuck Creeks. Streamflow at the 80-percent flow duration is approximately 18 ft^3/s in the White River at Muncie, 13 ft^3/s in Buck Creek near Muncie, and 5 ft^3/s in Killbuck Creek near Gaston (Horner, 1976, p. 216, 219, and 224).

Previous Investigations

Although the water resources have been studied in general, the ground-water resources have not been evaluated quantitatively.

The water resources of Delaware County are described in a report by Hoggatt and others (1968). Aquifer type, extent, and potential yield are described, and a water budget is presented for the drift and the bedrock systems. However, Hoggatt and others (1968, p. 28, 30) stated that their figures are approximate and that further refinement would be desirable.

The water resources of the upper White River basin are described in Cable and others (1971). This area is basically the project area plus Marion County. The study involved surface water, ground water, and quality of water in the entire upper basin, but no detailed ground-water study of the Delaware County study area was done.

The results of an analog model study of the ground-water system of the upper White River basin are reported in Maclay and Heisel (1972). Nearly all the data used in their study was derived from the investigation of Cable and others (1971). Although Maclay and Heisel quantified the ground-water flow in the basin, the purpose of their study was to investigate the basinwide effects of ground-water development. As a result, that study was not detailed enough locally to provide a quantitative assessment of the ground-water resources.

Meyer and others (1975) determined ground-water availabiltiy in Marion County, just southwest of the study area. Three confined and an outwash aquifer were modeled, but availability from the outwash was emphasized in the pumping plans. An additional 57.5 Mgal/d was simulated from three well fields in the outwash by a pumping plan that used a saturated thickness of half the actual thickness at the pumping well. Drawdowns within half a mile of the pumping wells were generally less than 20 ft.

Methods of Investigation

Hydrologic data collected were used to define the ground-water flow system. These included data for (1) mapping the areal extent, altitudes of the tops, and thicknesses of the aquifers; (2) defining the potentiometric surfaces; (3) calculating the hydraulic properties; and (4) estimating discharge.

Mapping of the areal extent, thickness, and altitude of the top of the sand and gravel units was completed early. Approximately 1,500 rotary and cable-tool drillers' logs of domestic, industrial, and municipal wells previously field located by the Division of Water, Indiana Department of Natural Resources, were used in this mapping. Differentiation of material in each unit into sand, sand and gravel, and gravel was not incorporated in the mapping because of inconsistency in lithologic descriptions on well logs and the predominance of mixed sand and gravel. Consequently, although the units are composed of varying combinations of sand, sand and gravel, and gravel, lithologically they are considered to be mixed sand and gravel units.

The mapping was refined with data collected from 125 test holes drilled by auger by the Geological Survey. In addition to obtaining lithologic and geophysical logs for each test hole, 2-in.-diameter observation wells were installed in most test holes. The additional data did not significantly change the mapping, which indicates that commercial drillers' interpretations generally agree with those of the Survey. In addition to the Survey well logs, about 70 drillers' descriptions were recorded by the Survey during drilling of domestic wells by rotary and cable tools. Water levels in the Survey observation wells and in domestic wells were used to define the potentiometric surfaces of the aquifers. Analysis of specific-capacity data from adjacent Madison County was used to estimate the hydraulic conductivity of the sand and gravel aquifers. Ground-water pumpage was obtained from information furnished by large-scale (>0.1 Mgal/d) users of ground water. Ground-water discharge to streams was estimated by measuring changes in stream discharge throughout the area during low flow and adjusting for other inflows or outflows to or from the stream during the discharge measurements.

A seven-layer digital model was constructed to simulate ground-water flow. The model was used to simulate the effects of various pumping plans on the ground-water system and streamflow and to determine, on a regional basis, the quantity of ground water that could be pumped without adverse effect on the ground-water system and streamflow.

Acknowledgments

The author is grateful to the Division of Water, Indiana Department of Natural Resources for the use of water-well records obtained from their files and to the following municipalities, businesses and industries that provided pumping information: Middletown, Yorktown, Parker City, Gaston, and Mount Summit; and Brooks Foods, Marsh Foods, Warner Gear, Ball Brothers, Chevrolet Motor Division, and Franks Foundries. Appreciation is also expressed to the private well owners in Delaware and Henry Counties who permitted water-level measurements in their wells.

GEOLOGY

The area is mainly underlain by sedimentary rocks of Silurian age covered by varying thicknesses of drift. A generalized west-to-east geologic section is shown in figure 3. This section illustrates the major geologic features: drift overlying the bedrock and composed predominantly of till; thin, generally horizontal discontinuous sand and gravel units interbedded with the till in the drift; and varying thickness of the drift.

Bedrock

Limestone, dolomite, and shale, ranging in age from Silurian to Ordovician and dipping southwest (Indiana Geological Survey, 1956), underlie the study area. Ordovician rocks (mainly shale and limestone) are at the bedrock surface along bedrock valleys through the south half of the study area (fig. 4). Most of the bedrock is overlain by drift, although, in places west of Muncie, bedrock crops out along the White River (Shurig, 1974, p. 15). The bedrock surface is locally incised by steep-walled bedrock valleys; for example, the Anderson Valley (Wayne, 1956, p. 30, 38), which trends west through the study area. Bedrock altitude, caused mainly by erosion, generally ranges from 600 to 1,000 ft (fig. 4). Figure 4 is based on the Indiana Geological Survey's map showing bedrock topography of northern Indiana (Burger and others, 1966). This map was modified by plotting and contouring bedrock-altitude data from well logs on file with the Division of Water, Indiana Department of Natural Resources, and from test holes drilled to bedrock surface by the U.S. Geological Survey. Wayne (1956, p. 30) stated, "Karst phenomena were at least moderately well formed upon the limestone plain prior to it's burial." Harrell (1935, p. 178, 179) stated

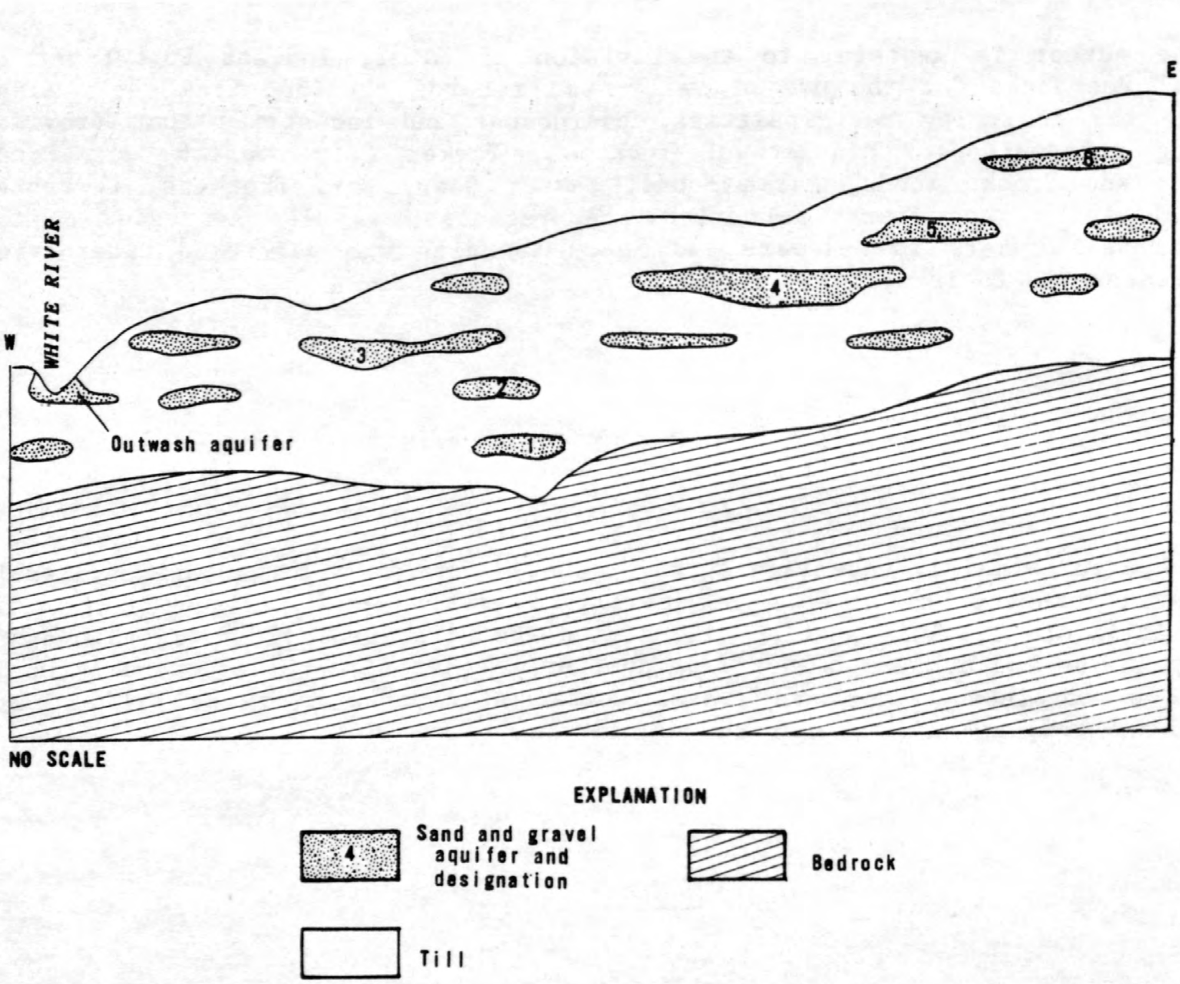


Figure 3.-- Generalized geologic section of study area.

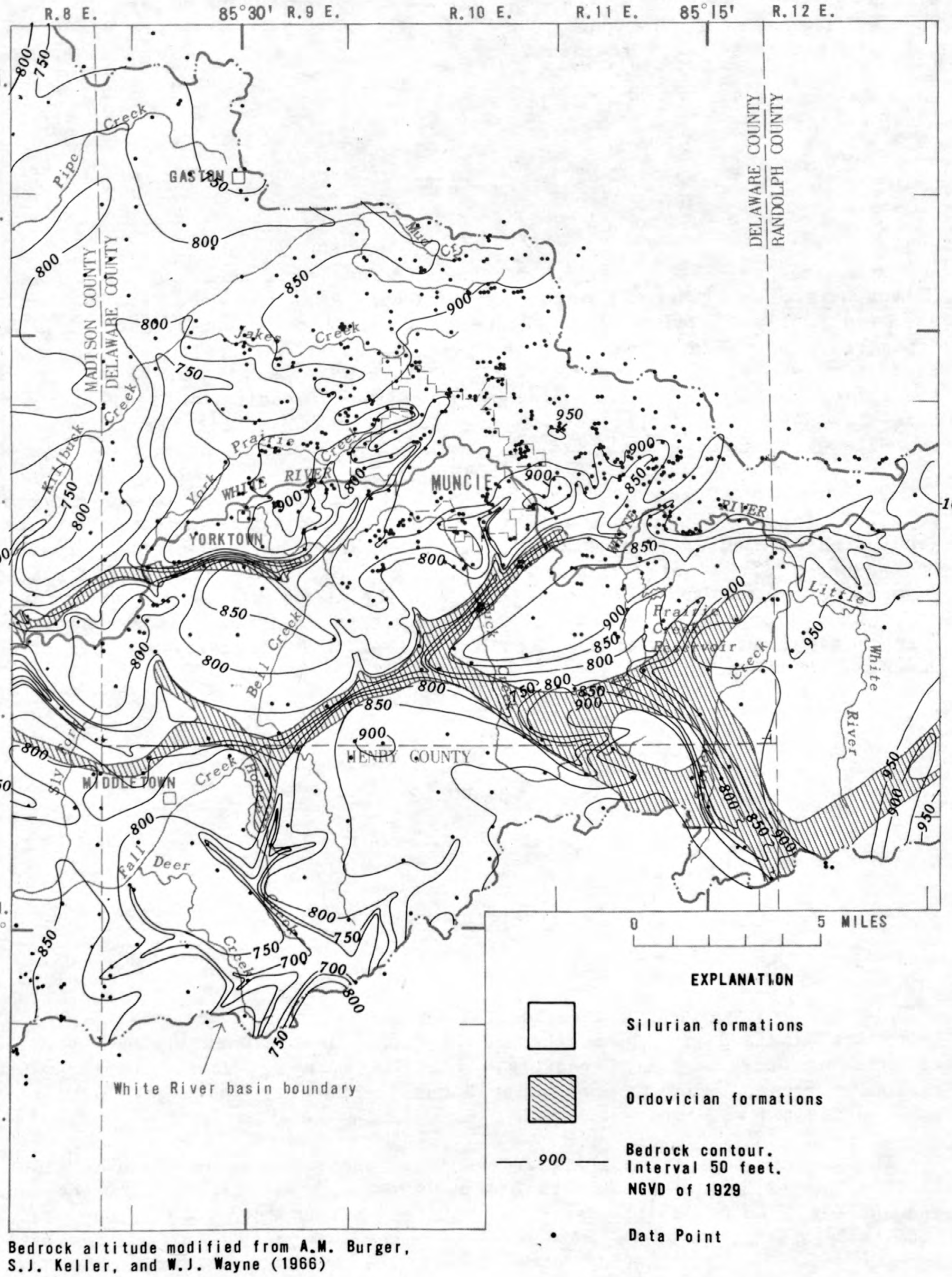


Figure 4.-- Bedrock-surface altitude in the study area.

that the Huntington Dolomite probably underlies the drift along the east boundary of Delaware County, whereas the Liston Creek Formation underlies the drift in the remainder of the county, and that the Mississinewa Shale underlies the dolomite and limestone formations.

Drift

Drift covers most of the study area, increases in thickness from northeast to southwest, and generally ranges in thickness from 0 to 500 ft. The drift is composed mostly of till (poorly sorted clay, silt, and sand) that was probably deposited during at least three glaciations--the Kansan, Illinoian, and Wisconsin (Wayne, 1956, p. 16, 49). Interbedded within the till are thin, sheetlike, and areally discontinuous stratified deposits of sand and gravel. Locally, these nearly horizontal deposits coalesce vertically to form thick deposits of sand and gravel.

The surface of the drift is composed of ground moraine and several other deposits (fig. 5). The Union City moraine, an end moraine, extends east-west just north of Muncie. The moraine forms a low ridge having a relief of about 30 ft (Wayne, 1975, p. 3). Another end moraine lies in the area of Prairie Creek Reservoir. Kames and eskers lie north and east of Muncie and are mined for their sand and gravel. Other surficial deposits include small areas of lacustrine plains, outwash plains, alluvium, and glacial sluiceways. Most of these deposits are shown in figure 5.

HYDROLOGY

Aquifer Geometry

Eight hydrologic units were identified as potential aquifers. Six of these units are sand and gravel interbedded in till. The seventh is the upper 150 ft of predominantly carbonate bedrock, which is more permeable than the deeper carbonate rocks, owing to weathering along fractures. The eighth unit is outwash associated with major streams.

The general stratigraphic relationship of the aquifers is shown in figure 3. The six sand and gravel aquifers are numbered so that aquifer 1 is the deepest and aquifer 6 is the shallowest. The aquifers extend into other study areas and were mapped accordingly. However, a number for each aquifer was assigned independently in each study area during the mapping. Correlation of each aquifer between study areas is given in table 1. For instance, Delaware County aquifer 4 extends into Madison County, where it is called Madison County aquifer 5; into Hamilton County, where it is called Hamilton County aquifer 5; and into Randolph County, where it is called Randolph County aquifer 1.

Table 1.--Correlation of the sand and gravel aquifers between study areas in the White River basin upstream from Marion County, Ind.

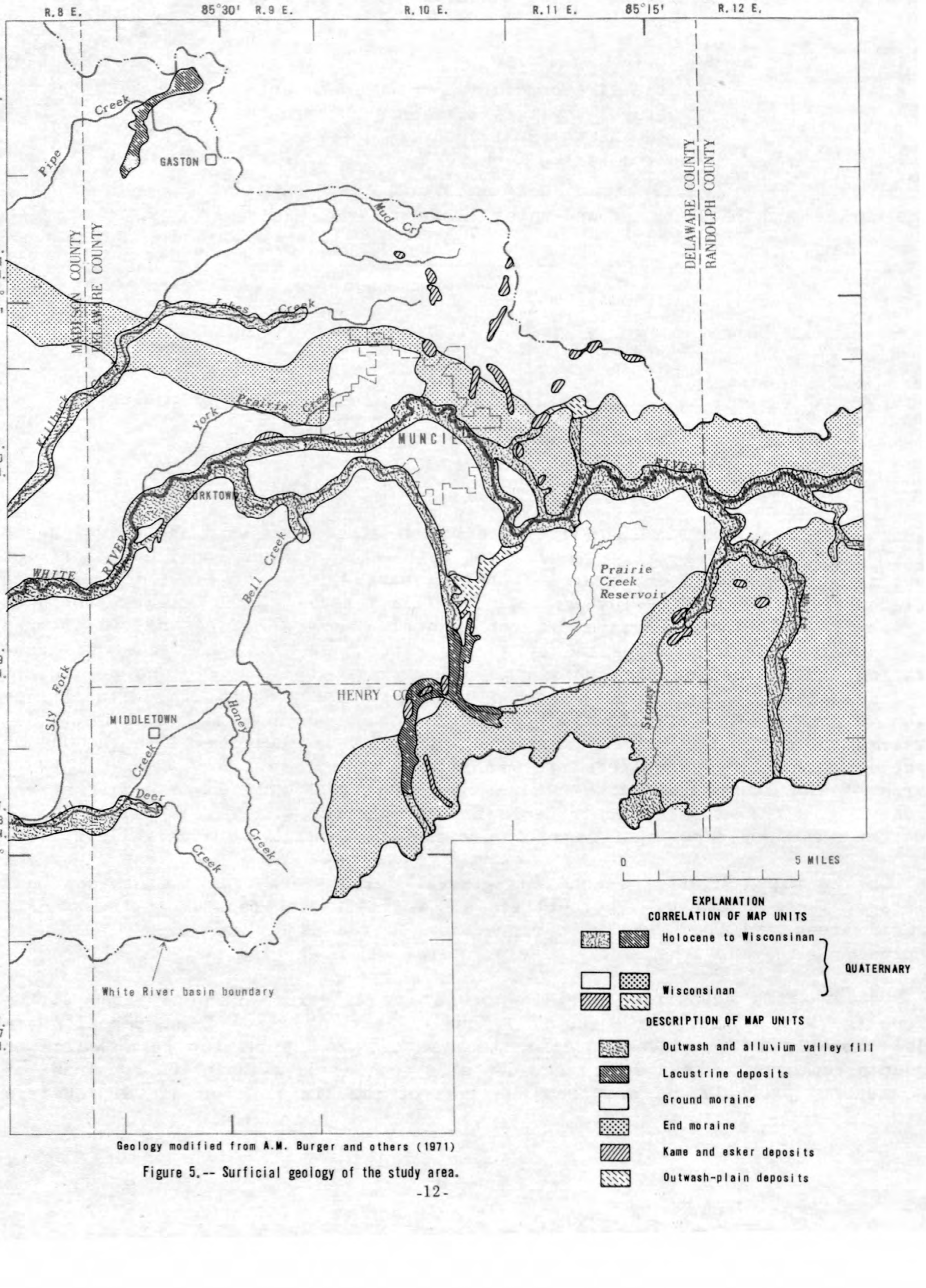
County	Aquifer number							
Hamilton	--	--	--	[1]5	[1]4	3	2	1
Madison	--	--	[1]6	[1]5	4	3	2	1
Delaware	[1]7	6	5	4	3	2	1	--
Randolph	4	3	2	1	--	--	--	--

[1]Insignificant in study area.

Thickness of the six sheetlike, areally discontinuous sand and gravel aquifers (figs. 6-17) generally ranges from 5 to 40 ft. These aquifers represent several units of sand and gravel, within a narrow range of altitude, that are horizontally separated by low-permeability till. Locally, the aquifers coalesce vertically. The aquifer maps are not meant to imply that the individual sand and gravel units were deposited by a single process. But, for ease of illustrating and discussing the individual units, they are grouped into six layers called aquifers 1 through 6. In some places, the mapping may be a simplification of the complex distribution of confined sand and gravel units in a till system. However, the major features of the units are delineated in the illustrations. The upper aquifers are generally found in only the east and south parts of the study area (figs. 15 and 17). Progressively deeper aquifers are found toward the west. Neither Geological Survey nor commercial drillers' logs indicated a major deposit of sand and gravel in the narrow bedrock valleys.

On the maps of all the sand and gravel aquifers are small, isolated, outliers of these aquifers. The outliers are insignificant because of their small areal extent and poor hydraulic connection to the major aquifers. They were included on the maps, however, to help in any future mapping.

Valley-train deposits, overlain locally by alluvium, occupy narrow strips along the White River and Buck, Fall, and Killbuck Creeks. These deposits are also probably a significant aquifer because of their proximity to a source of induced recharge. The age of these deposits (called outwash in the remainder of the report) probably differs from the ages of the six sand and gravel aquifers



Geology modified from A.M. Burger and others (1971)

Figure 5.-- Surficial geology of the study area.

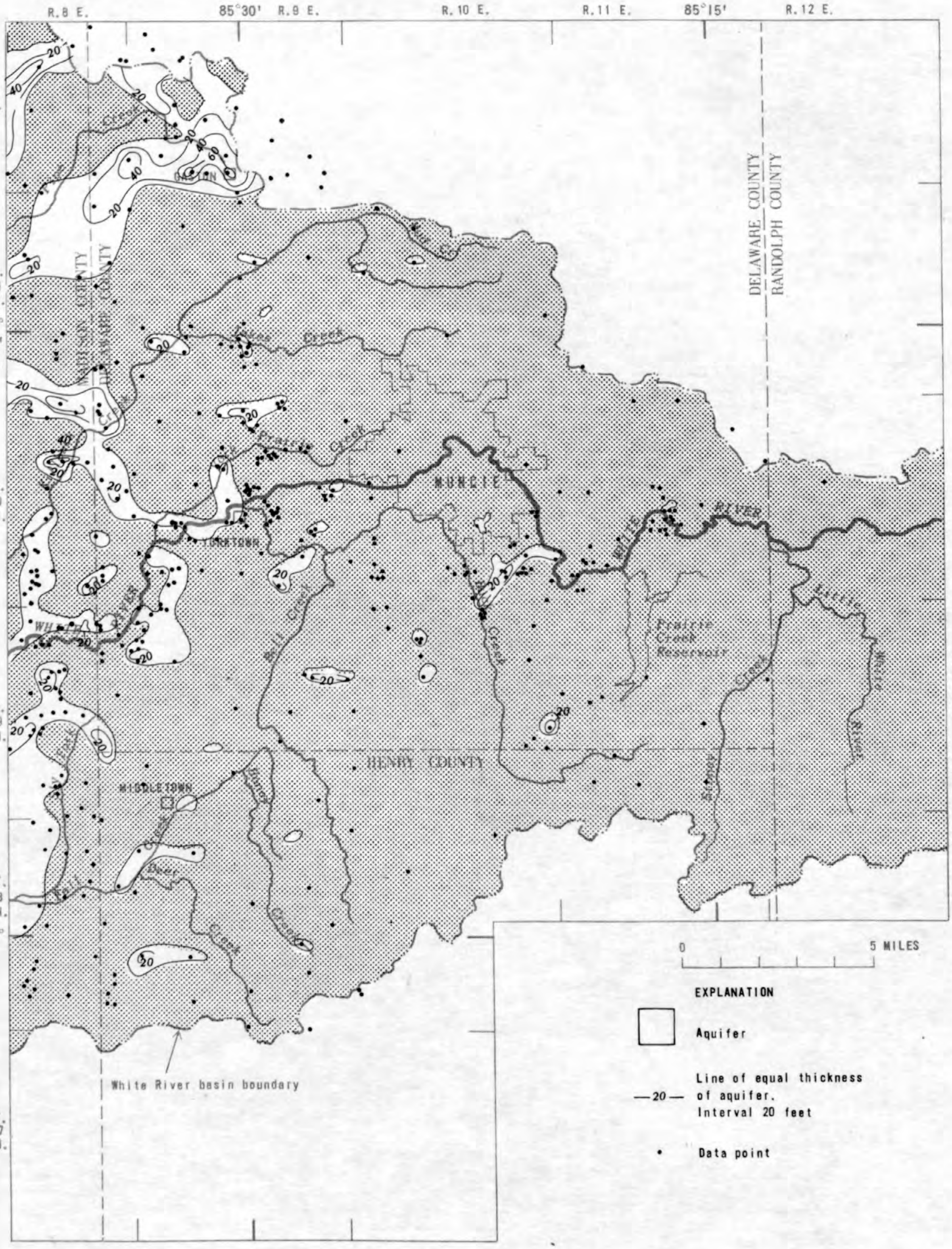


Figure 6.-- Thickness of aquifer 1.

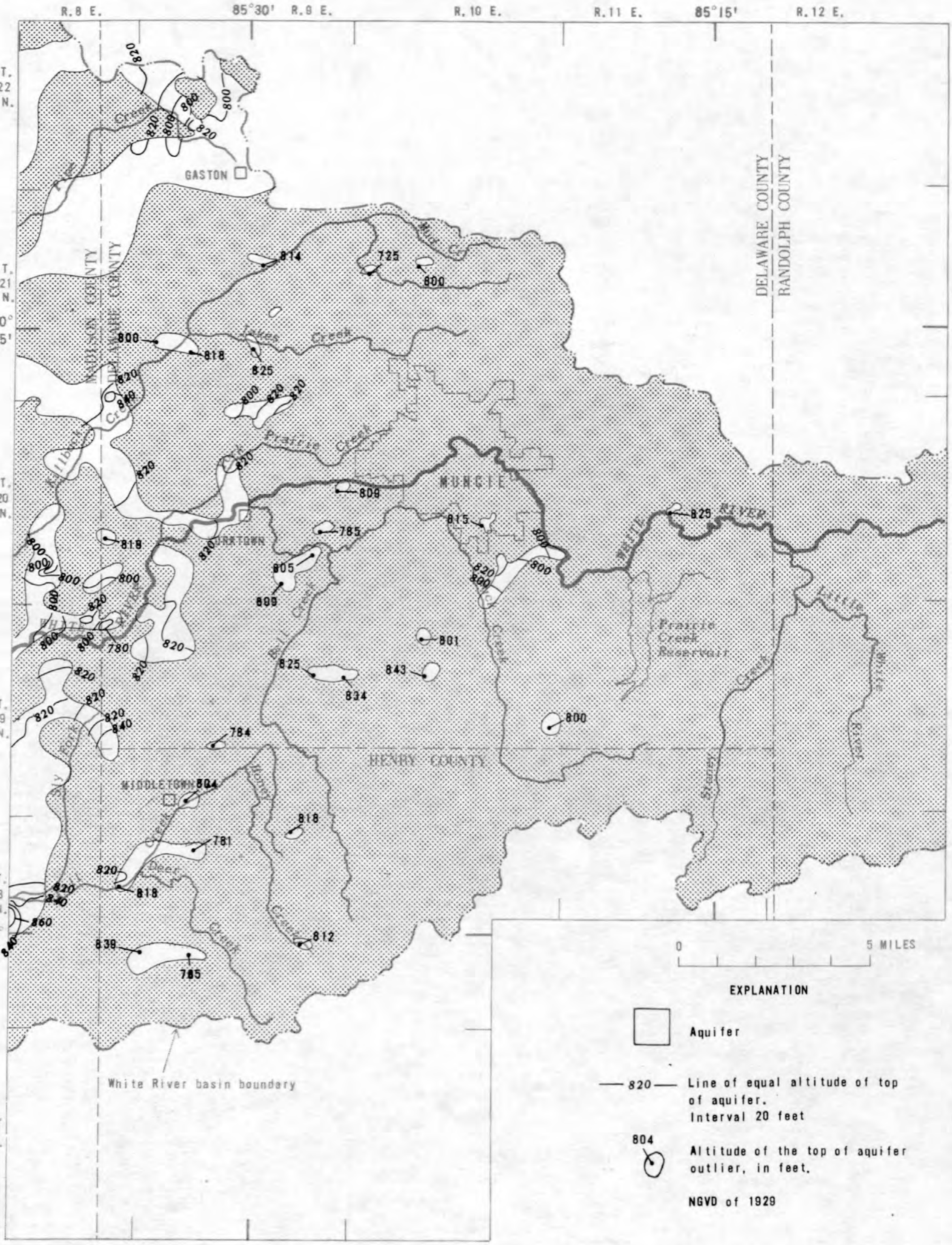


Figure 7.-- Altitude of the top of aquifer 1.

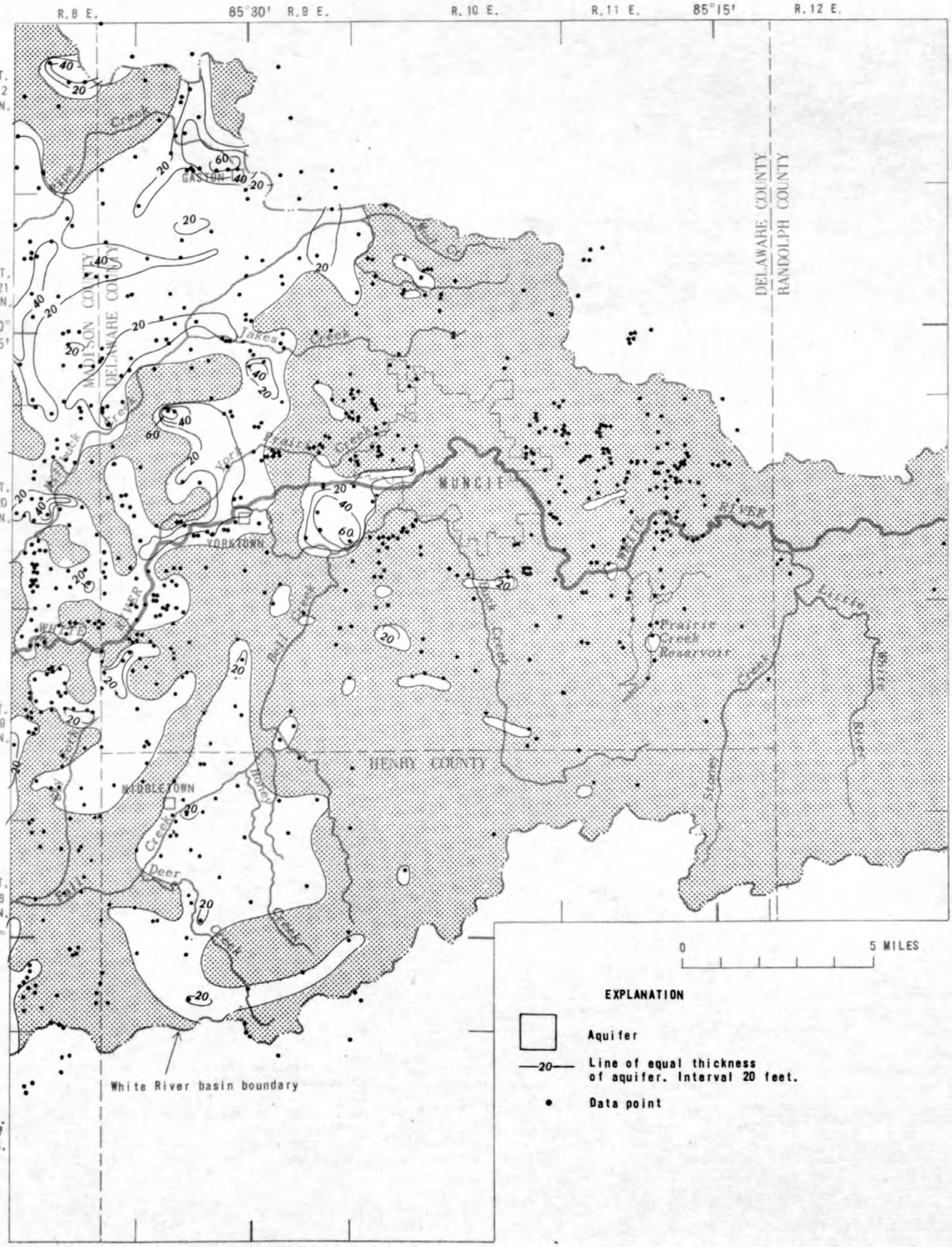


Figure 8.-- Thickness of aquifer 2.

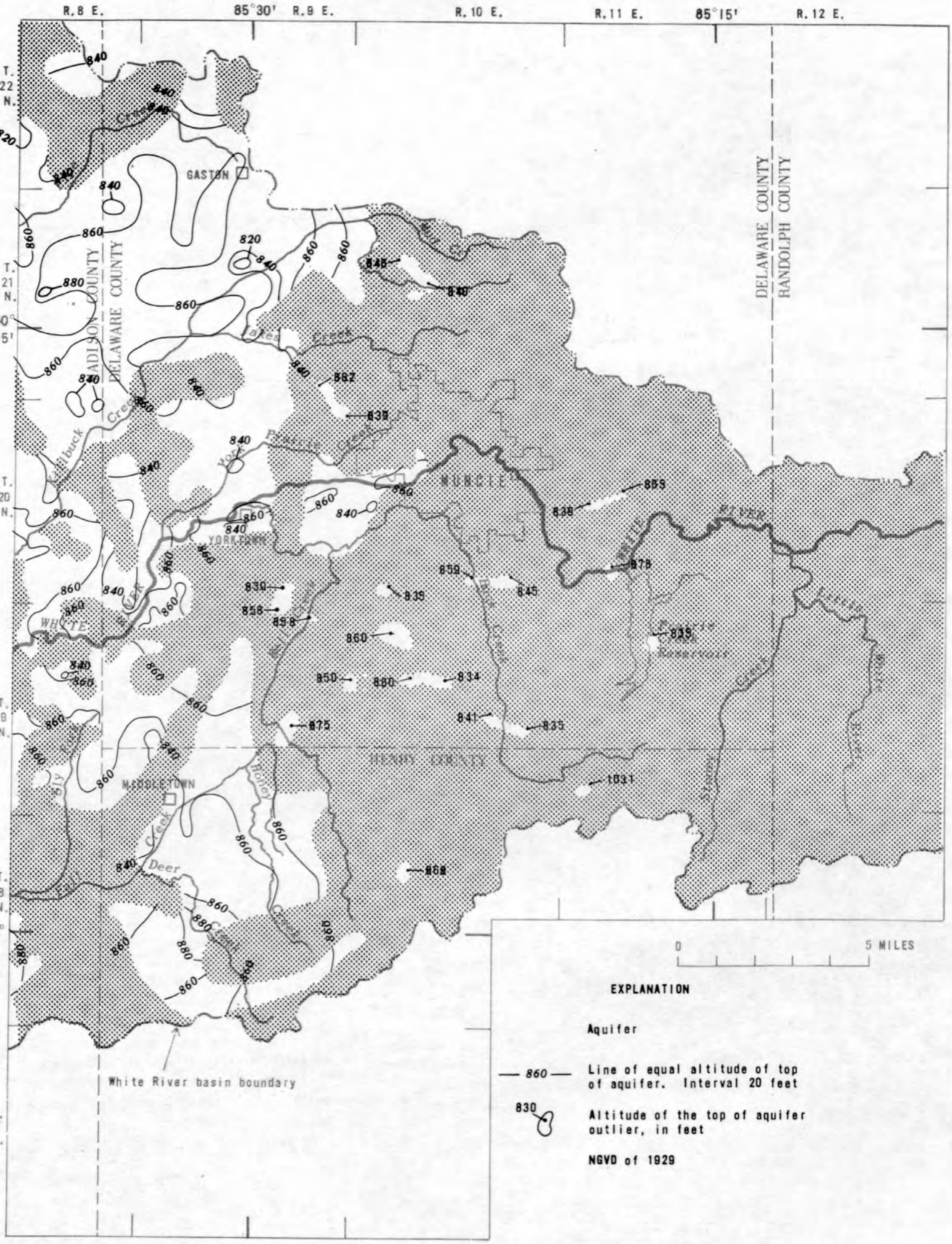


Figure 9.-- Altitude of the top of aquifer 2.

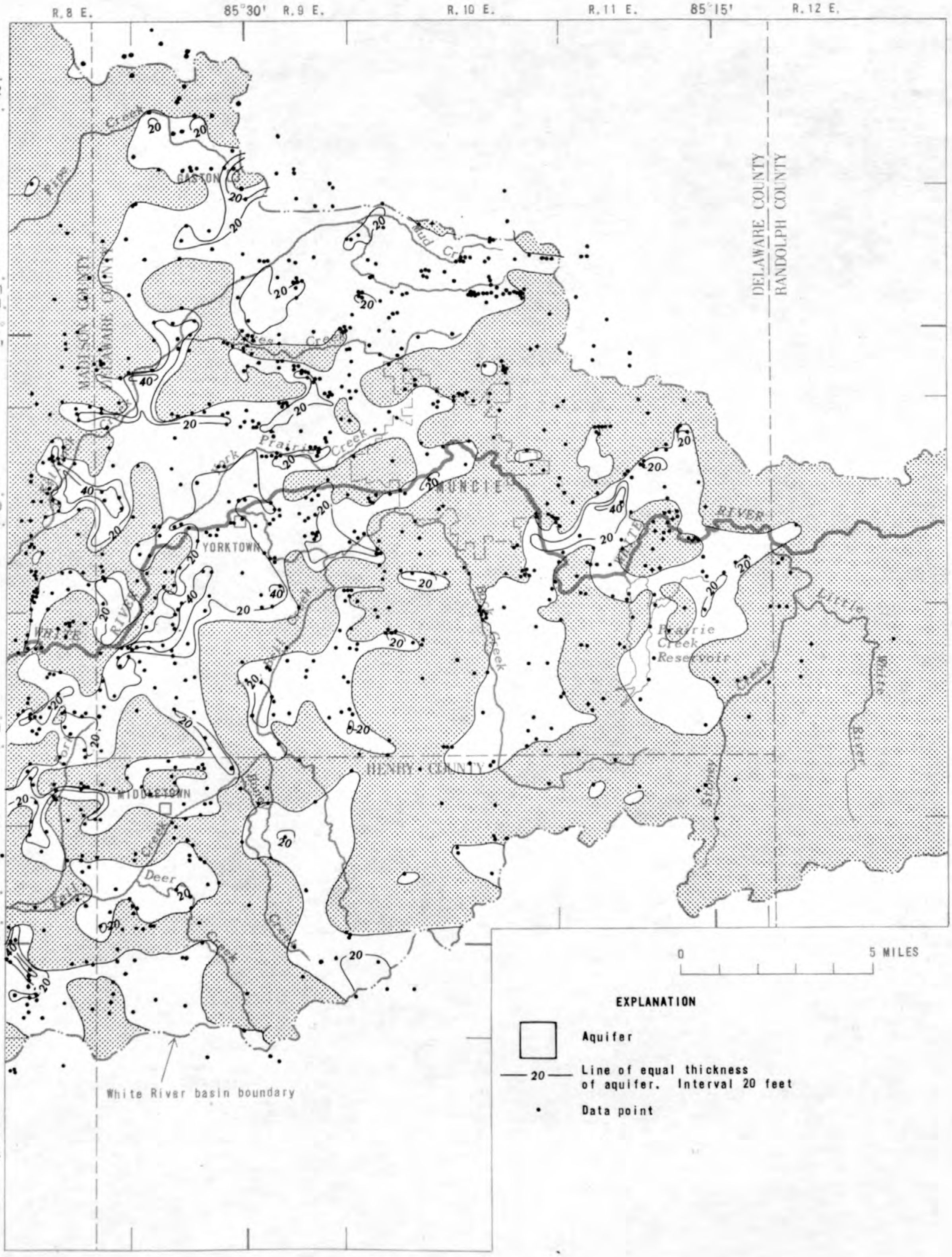


Figure 10.-- Thickness of aquifer 3.

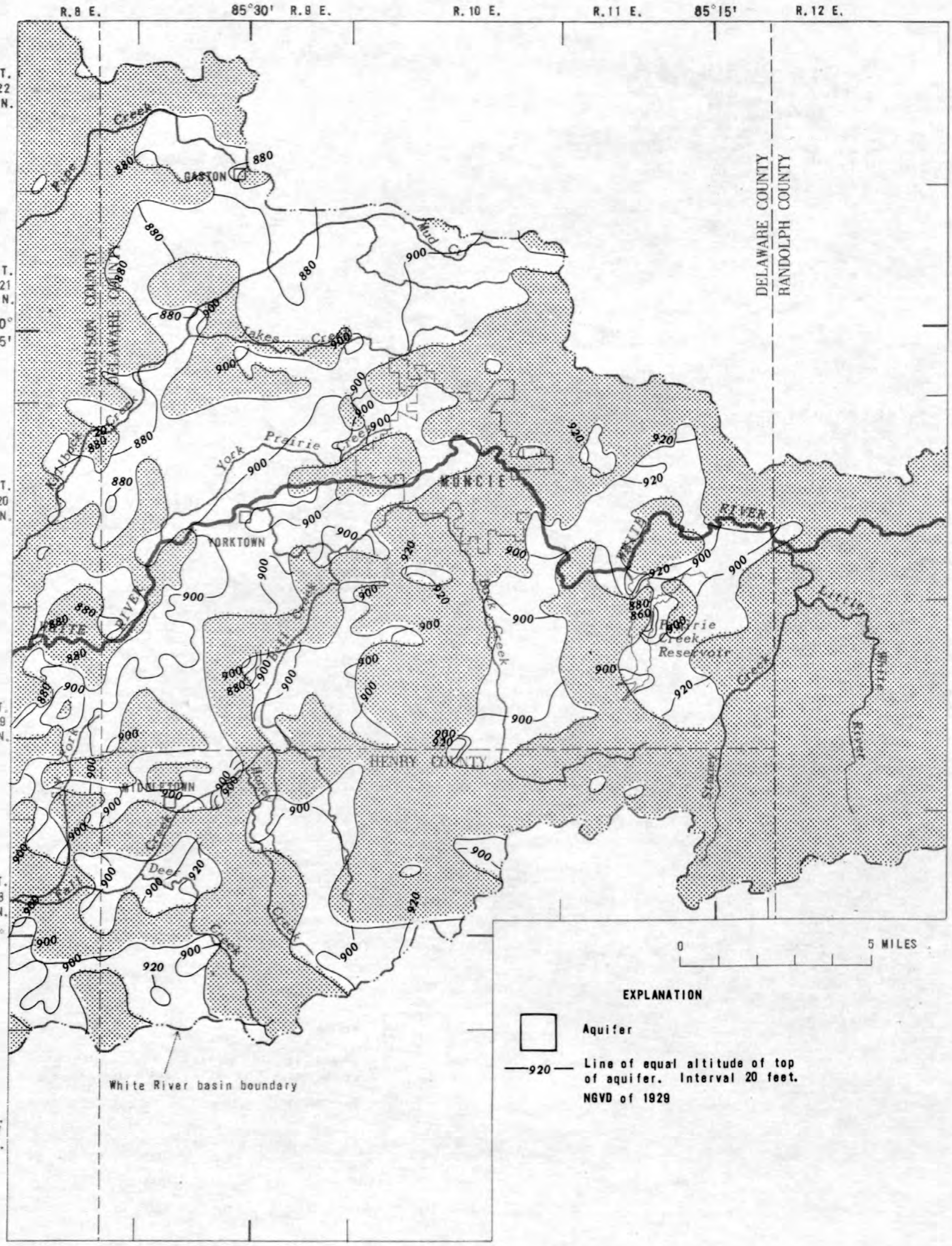


Figure 11.-- Altitude of the top of aquifer 3.

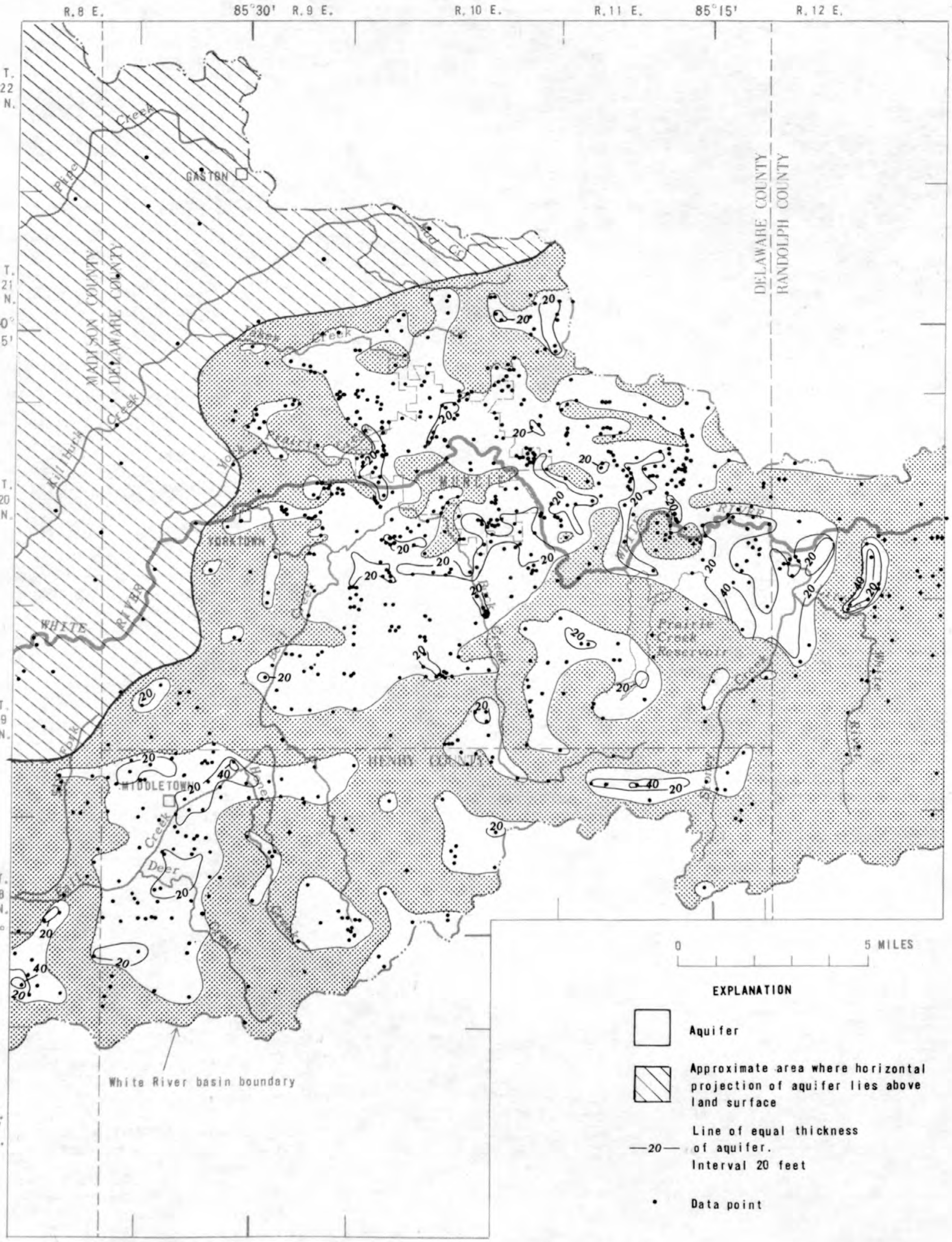


Figure 12.-- Thickness of aquifer 4.

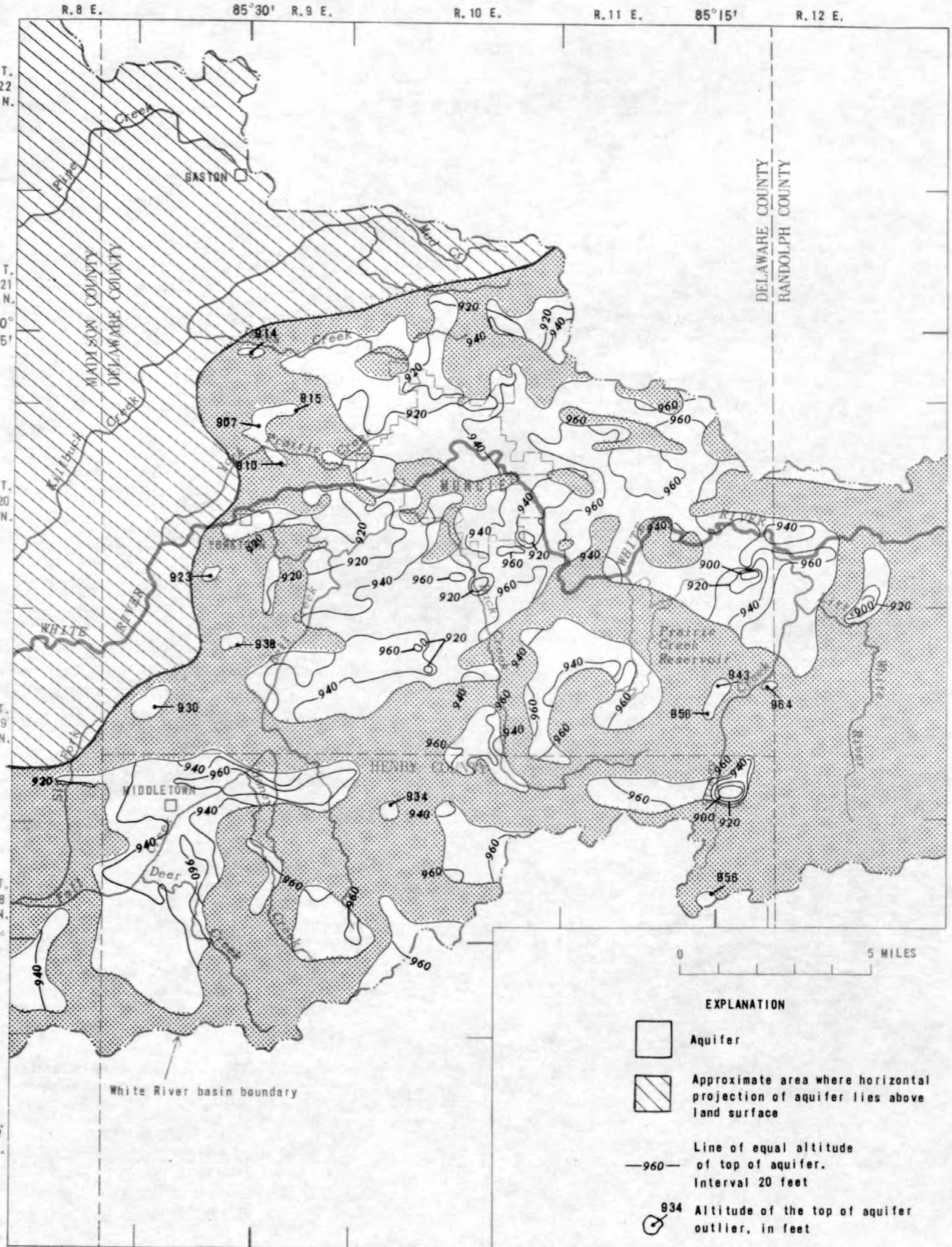


Figure 13.-- Altitude of the top of aquifer 4.

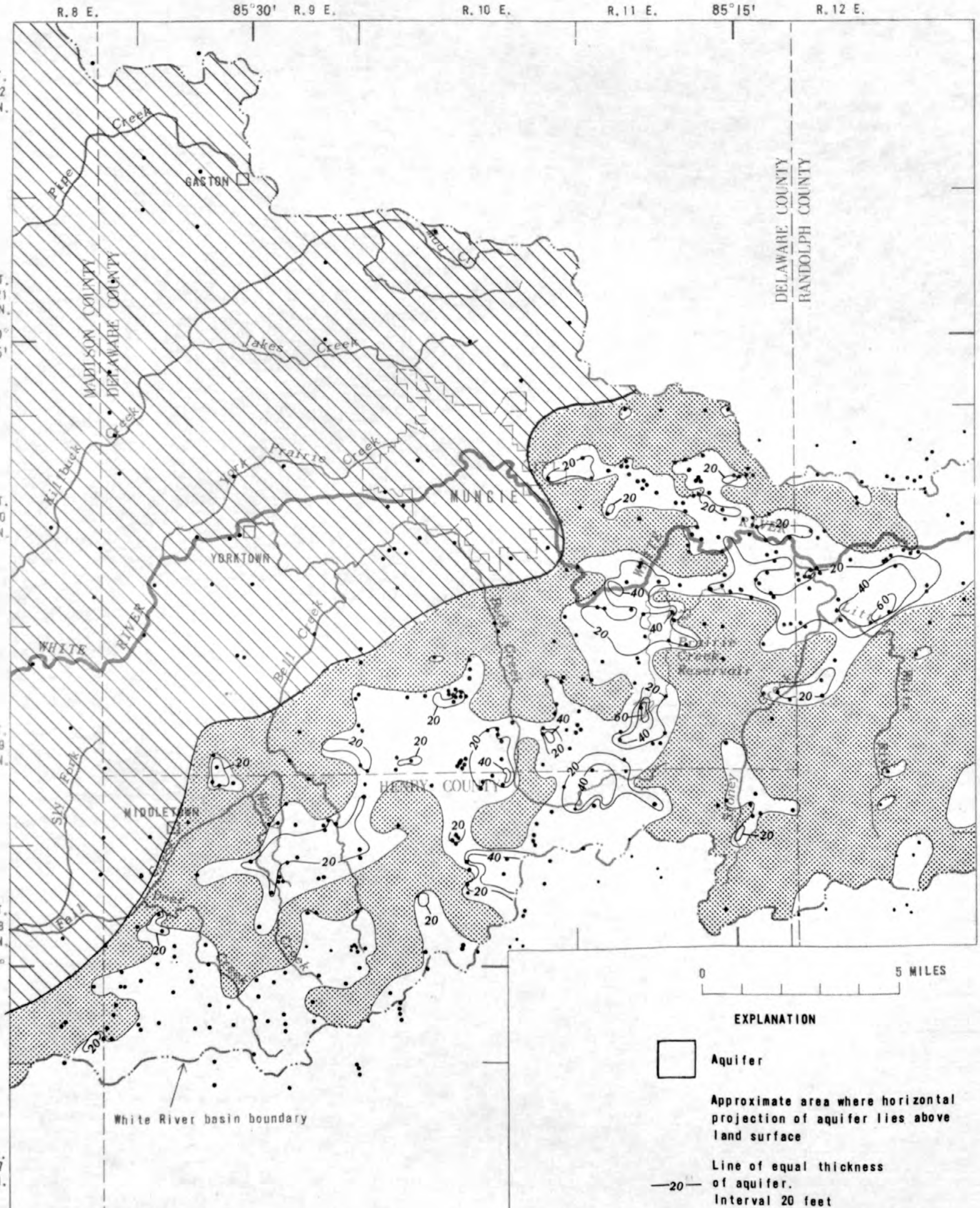


Figure 14.-- Thickness of aquifer 5.

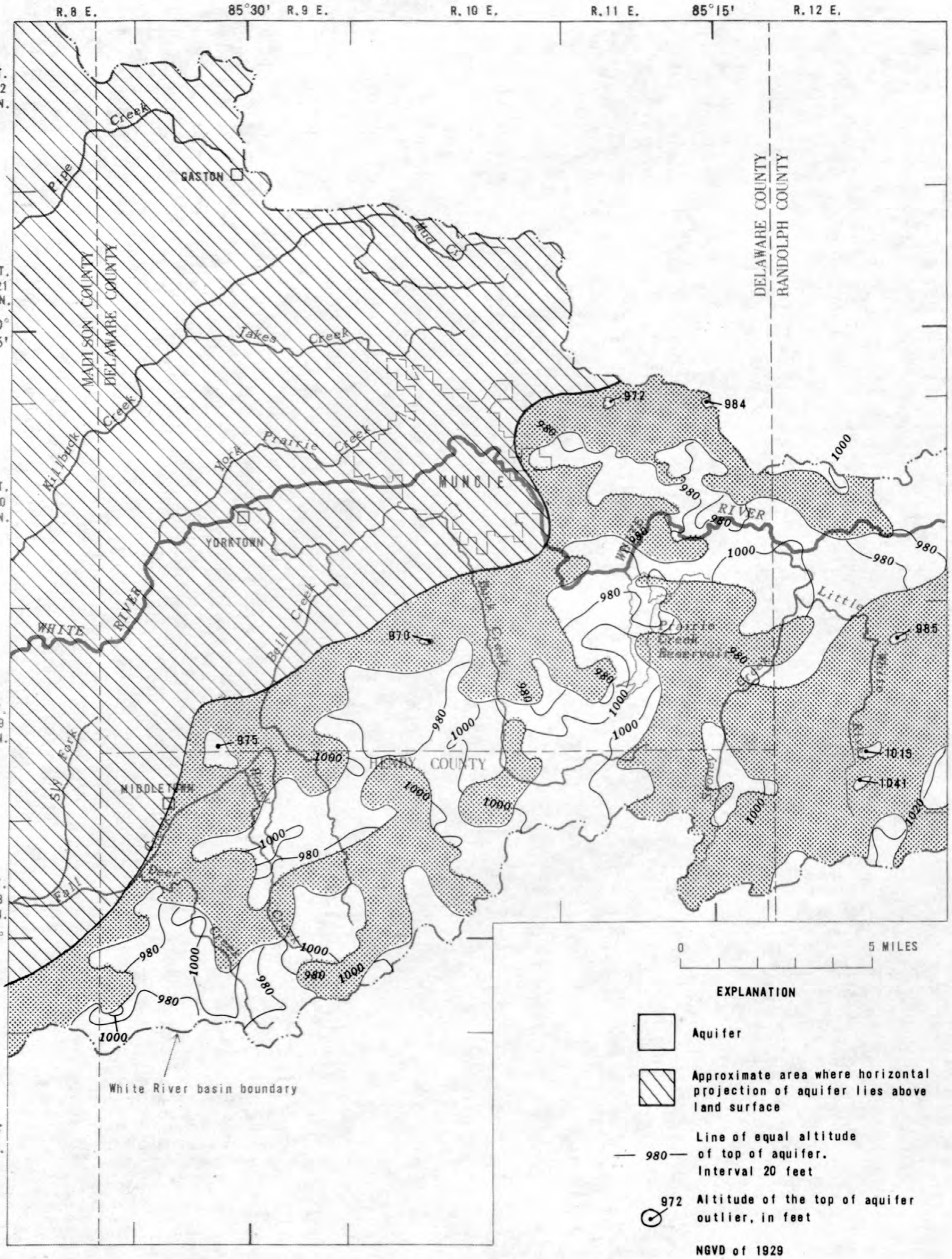


Figure 15.-- Altitude of the top of aquifer 5.

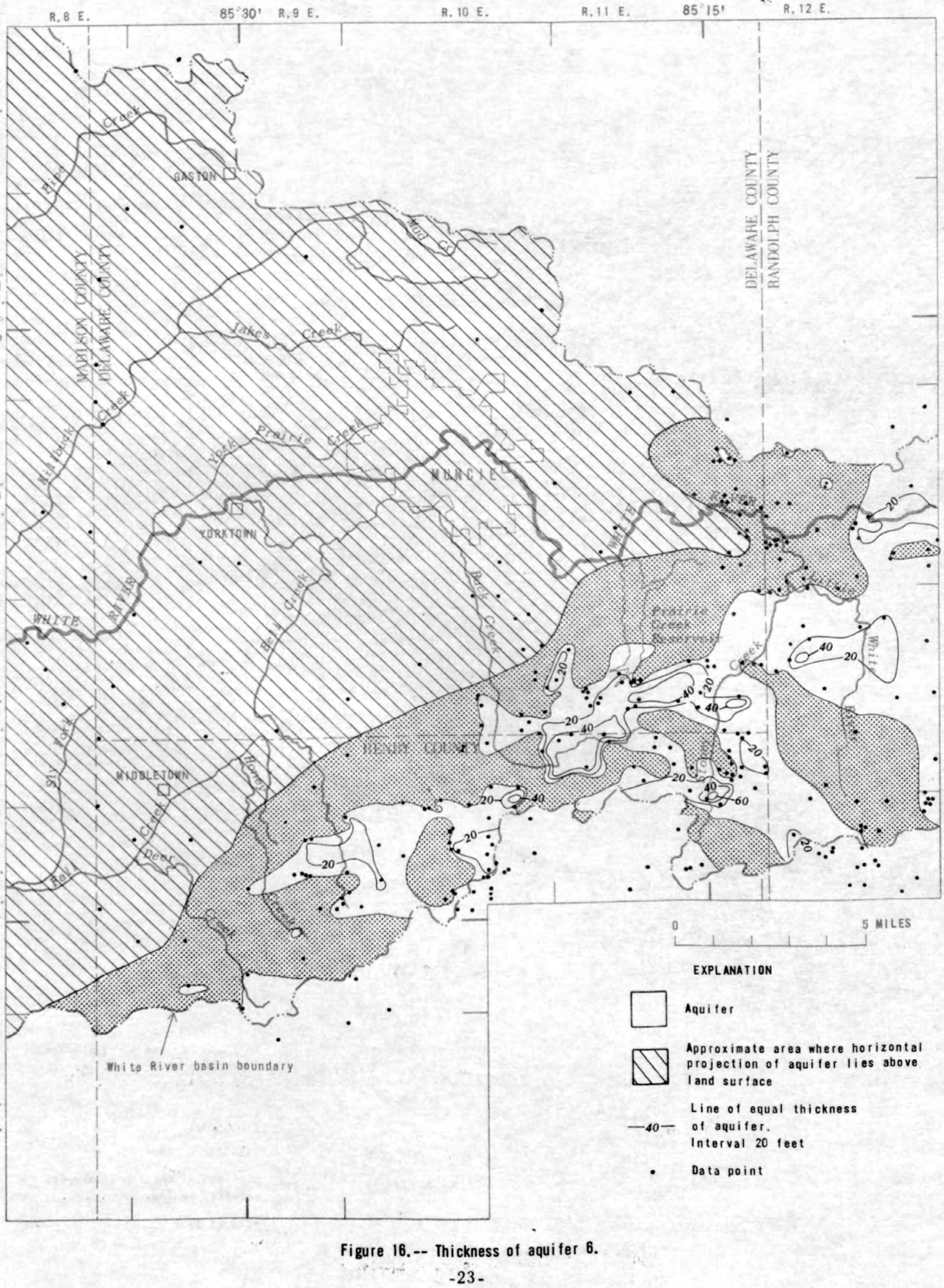
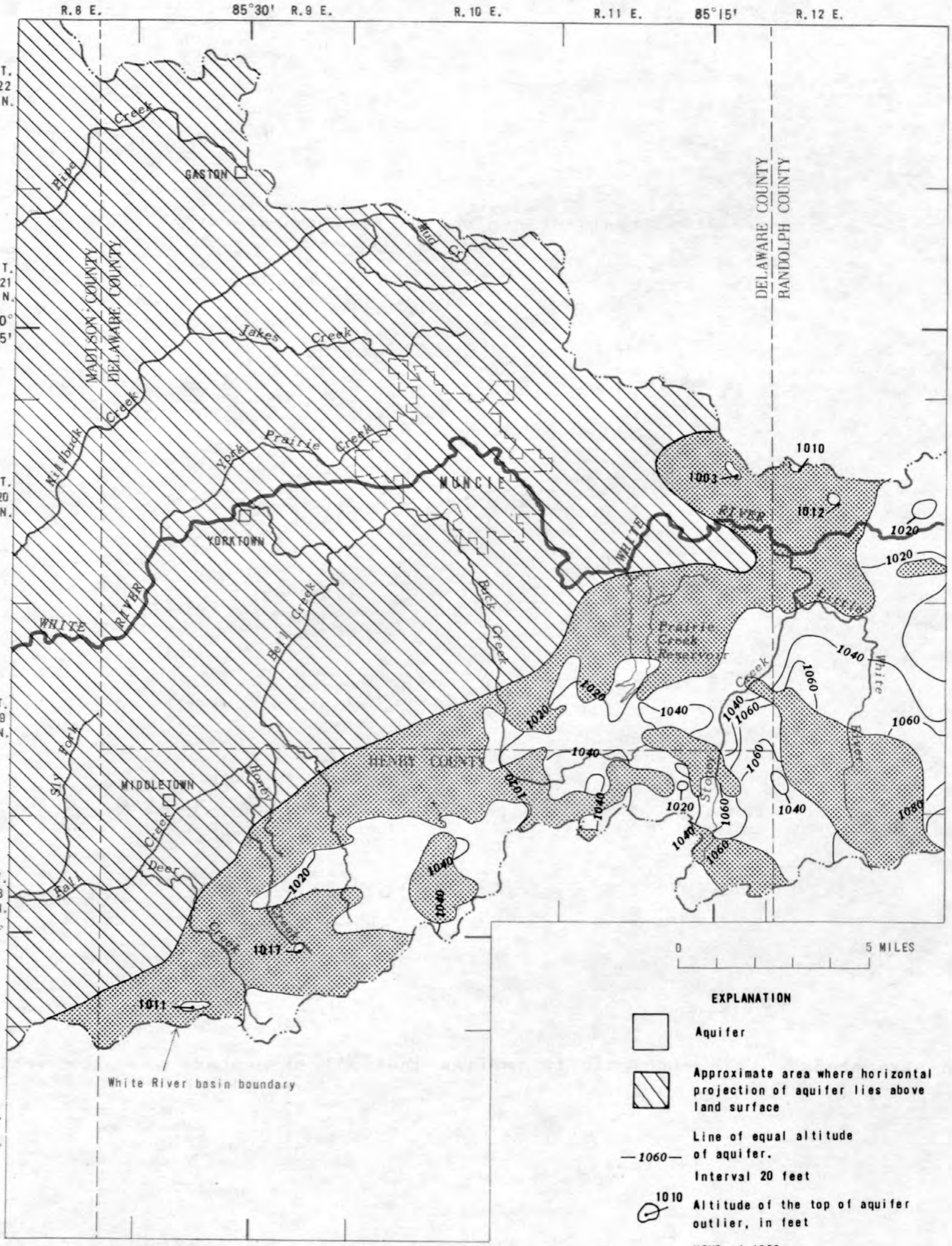


Figure 16.-- Thickness of aquifer 6.



EXPLANATION

Aquifer

Approximate area where horizontal projection of aquifer lies above land surface

—1060— Line of equal altitude of aquifer. Interval 20 feet

1010 Altitude of the top of aquifer outlier, in feet

NGVD of 1929

Figure 17.-- Altitude of the top of aquifer 6.

interbedded in the till. However, the deposits are lithologically similar and generally grade laterally into the aquifers. Therefore, the outwash deposits were mapped as continuations of one or more of the six sand and gravel aquifers. Although the areal extent of the outwash deposits is not distinguishable from the confined sand and gravel aquifers, outwash thickness is generally less than 20 ft.

In areas without sand and gravel aquifers, the Silurian carbonate aquifer is commonly used for municipal, industrial, and domestic water supplies and, therefore, is a potentially significant aquifer. Although Cable and others (1971) and Meyer and others (1975, p. 17) assumed that the average permeable thickness of the carbonate aquifer is 100 ft, 150 ft was assumed in this study because many wells penetrate 150 ft of bedrock. However, the thickness of the bedrock aquifer may exceed 150 ft. The consequences of this assumption are discussed in the section "Bedrock Aquifer," which follows this section.

Hydraulic Characteristics of the Ground-Water System

Sand and Gravel Aquifers

The two types of sand and gravel aquifers are the confined aquifers and the outwash aquifer. However, because of a similarity of aquifer materials and a lack of data, the authors assumed that the hydraulic conductivity of the unconfined areas of the aquifer is the same as that of the confined areas.

The hydraulic conductivity of the confined sand and gravel aquifers was calculated from specific-capacity data obtained for thirteen 2-in.-diameter observation wells in Madison County. Depths of the wells ranged from 23 to 275 ft. A description of the aquifer tests used to determine conductivity is given in Lapham (1981, p. 17). The hydraulic conductivities calculated with specific-capacity data for the 13 observation wells are given by Lapham (1981, p. 30). The average hydraulic conductivity of the sand aquifers was 156 ft/d, and that of the sand and gravel aquifers was 710 ft/d. Because most aquifer material is indicated in well logs to be mixed sand and gravel, a representative hydraulic conductivity for the six sand and gravel aquifers (figs. 6-17) was calculated by averaging the hydraulic conductivity of the sand aquifers with the hydraulic conductivity of the sand and gravel aquifers. The average hydraulic conductivity calculated in this manner is 433 ft/d. The range in hydraulic conductivity (24-1,633 ft/d) of the sand and gravel aquifers demonstrates the extent of the variability of the hydraulic conductivity of the aquifers.

The hydraulic conductivity of the sand and gravel aquifers has also been calculated in previous studies of the White River basin. Using specific-capacity data for wells, Cable and others (1971, p. C11) calculated a hydraulic conductivity of 200 ft/d for the confined sand and gravel aquifers and 334 ft/d for the unconfined sand and gravel aquifers. Meyer and others (1975, p. 21) determined a calibrated hydraulic conductivity of 433 ft/d for the confined sand and gravel aquifers in Marion County.

Transmissivities of the sand and gravel aquifers were calculated from the specific-capacity data for each of the 13 observation wells by the method described by Brown (1963). Because only the bottom 3 ft of the aquifer was screened, the measured drawdowns were adjusted for partial penetration by the method described by Butler (1957, p. 159-162). On the basis of a study by Meyer and others (1975, p. 19), the ratio of horizontal to vertical hydraulic conductivity of the sand and gravel aquifers was assumed to be 10:1. Hydraulic conductivities were calculated by dividing the calculated transmissivities by the total thicknesses of the aquifers.

Transmissivity distributions for each of the six sand and gravel aquifers can be determined by multiplying the aquifer thickness shown in figures 6, 8, 10, 12, 14, or 16 by 433 ft/d, the average hydraulic conductivity of the confined sand and gravel aquifers. Because of the large range in hydraulic conductivity calculated from the specific-capacity data, the transmissivity of an aquifer calculated by multiplying the thickness of the aquifer by 433 ft/d results in an estimate of the transmissivity of the aquifer. However, these transmissivities are adequate in a regional analysis of the ground-water flow system such as that done in this study.

Neither the storage coefficient nor the specific yield of the sand and gravel aquifers was calculated because only steady-state conditions were used.

Bedrock Aquifer

The transmissivity distribution of the carbonate bedrock aquifer (fig. 18) was calculated with specific-capacity data for approximately 300 domestic and municipal bedrock wells by the method described in Brown (1963). Transmissivity averaged 1,000 ft^2/d and generally ranged from 100 to 10,000 ft^2/d. The specific-capacity data for wells that did not fully penetrate the 150 ft of permeable bedrock assumed in the study were corrected for partial penetration by the method described in Butler (1957, p. 160). In this correction, the ratio of horizontal to vertical bedrock hydraulic conductivity was assumed to be 1:1. This ratio was assumed because much of the bedrock permeability probably results from vertical fractures. Model analysis of the Madison County study area indicated that the difference in the areal distribution of bedrock potentiometric head for ratios of 1:1 and 100:1 was insignificant and that the maximum difference in transmissivities corrected for partial penetration for ratios of 1:1 and 100:1 at a well was 50 percent but for most wells was much less than 50 percent. Areally, however, the difference in the transmissivity of the 150 ft of bedrock was as much as two orders of magnitude. Because the ratio of bedrock permeabilities does not account for much of the variability in bedrock transmissivity, the assumption of a 1:1 ratio should not significantly affect the transmissivity distribution.

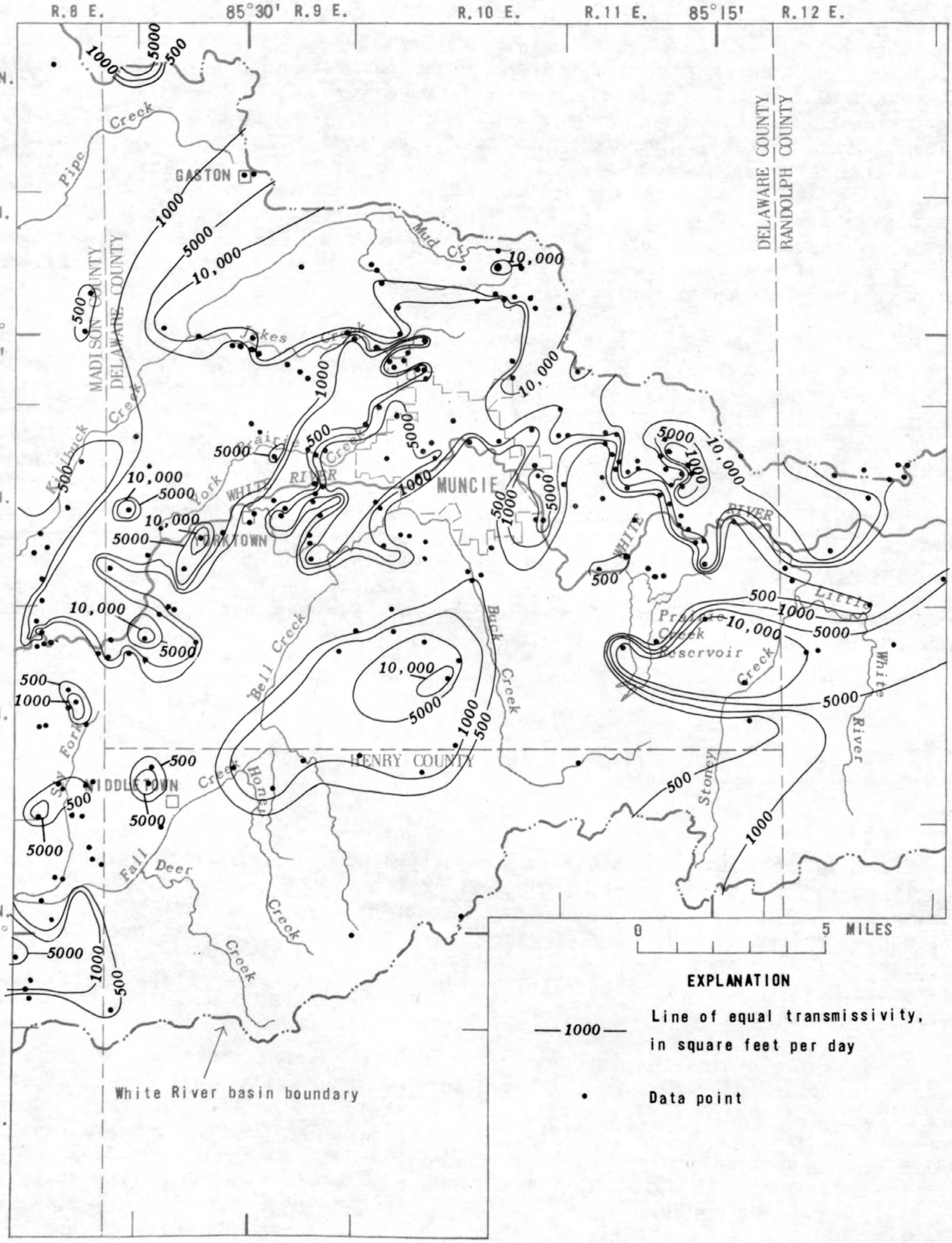


Figure 18.-- Transmissivity of bedrock aquifer calculated from specfic-capacity data.

The secondary permeability of the bedrock is probably highly variable, both horizontally and vertically. Therefore, the average thickness assumed for the permeable bedrock, 150 ft, is approximate. For the same reason, the areal variation in transmissivity of the bedrock is an estimate of the areal variability in transmissivity. The potential error in assuming only a 150-ft bedrock aquifer is in the calculations that correct for partial penetration. The calculated bedrock transmissivity increases directly with aquifer thickness. Therefore, the calculated transmissivities are an estimate of the upper part of the bedrock, where data was available and where the bedrock was probably most permeable. Even though transmissivity may double, depending on the depth of bedrock chosen, transmissivity varies areally by as much as two orders of magnitude. Therefore, the possible error in assuming an incorrect thickness is small compared to the range in transmissivity.

From bedrock well-log data, Cable and others (1971) determined that the thickness of the permeable bedrock in the White River basin is approximately 100 ft and estimated that the hydraulic conductivity of the bedrock is 13.4 ft/d. The transmissivity of the bedrock aquifer calculated from these data is 1,340 ft^2/d. The three preceding values were used in the ground-water study of Marion County by Meyer and others (1975, p. 17). Cable and others (1971, p. C12) noted considerable areal variation in bedrock transmissivity. Although a large variation in the bedrock transmissivity is shown in figure 18, the average transmissivity, 1,000 ft^2/d, is close to the average transmissivity of the bedrock, 1,340 ft^2/d, calculated by Cable and others (1971).

The storage coefficient of the bedrock aquifer was not calculated because only steady-state conditions were used.

Semipermeable Confining Beds

The till in which the aquifers are interbedded is composed primarily of a poorly sorted mixture of clay, silt, and sand that acts as a semipermeable confining bed. Although vertical hydraulic conductivity of the till is less than horizontal hydraulic conductivity, the area for vertical flow is three orders of magnitude greater than the area for horizontal flow. For this reason, the till transmits little water horizontally, and vertical flow through the till between aquifers dominates.

No data for estimating the vertical hydraulic conductivity of the confining beds were collected. The vertical hydraulic conductivity of the confining bed that Meyer and others (1975, p. 26) used in their analog model of Marion County ranges from 10^{-4} to 1.3×10^{-3} ft/d. In the Delaware County study, an average value of the vertical hydraulic conductivity obtained in the Marion County study (Meyer and others, 1975), 7×10^{-4} ft/d, was assumed. During calibration, the initial value was changed where necessary to obtain a better match of field conditions. As in the Marion County study, the range of vertical hydraulic conductivity of the confining beds in the Delaware County study area was large.

Ground-Water Flow

Adjacent shallow and deep wells were installed at about 40 locations. Water-level measurements in these wells show that flow is generally downward in upland areas, mainly horizontal near streams, and upward at the streams. The flow is primarily vertical through the till and horizontal through the aquifers. Flow patterns and horizontal gradients are illustrated by the potentiometric surfaces shown in the section, "Model calibration." The flow patterns indicate that regional flow is northwest, then west at the county line (figs. 24-30). In addition, discharge to streams is indicated. The flow pattern in the Delaware County study area is in transition from the mostly regional flow, as in Randolph County, to flow that discharges locally, as in Madison and Hamilton Counties. Horizontal gradient shown in the figures varies but is commonly about 10 ft/mi. Horizontal and vertical gradients indicate the need to consider three-dimensional flow in an analysis of the ground-water flow system.

Water-Level Fluctuations

Water levels change in response to variations in precipitation, evapotranspiration, major pumpage, and ground-water discharge and recharge. Some of these changes are illustrated in the hydrograph of observation well Delaware 4 (fig. 19), a 91-ft-deep well in aquifer 3. Annual water-level fluctuation in this well is generally about 3 ft. More important, however, the hydrograph indicates that the average annual water level did not change significantly throughout the period of record (1967-80).

Water levels in Geological Survey and domestic wells screened in both the bedrock and in the sand and gravel aquifers throughout the basin also support the assumption that water levels do not change significantly. Water levels in 28 wells in the basin were measured during at least two of the three autumns of 1976, 1977, and 1978 for comparison. The deviation in water levels from autumn to autumn ranged from 0.1 to 4.8 ft and averaged 2.2 ft. The average does not represent maximum annual water-level deviation because all measurements were made in the autumn. Although ground-water levels fluctuate in response to seasonal variations in recharge, the ground-water system is in dynamic equilibrium and approximates steady state. If this were not true, the trend of the average annual water level in wells throughout the basin would either rise or fall.

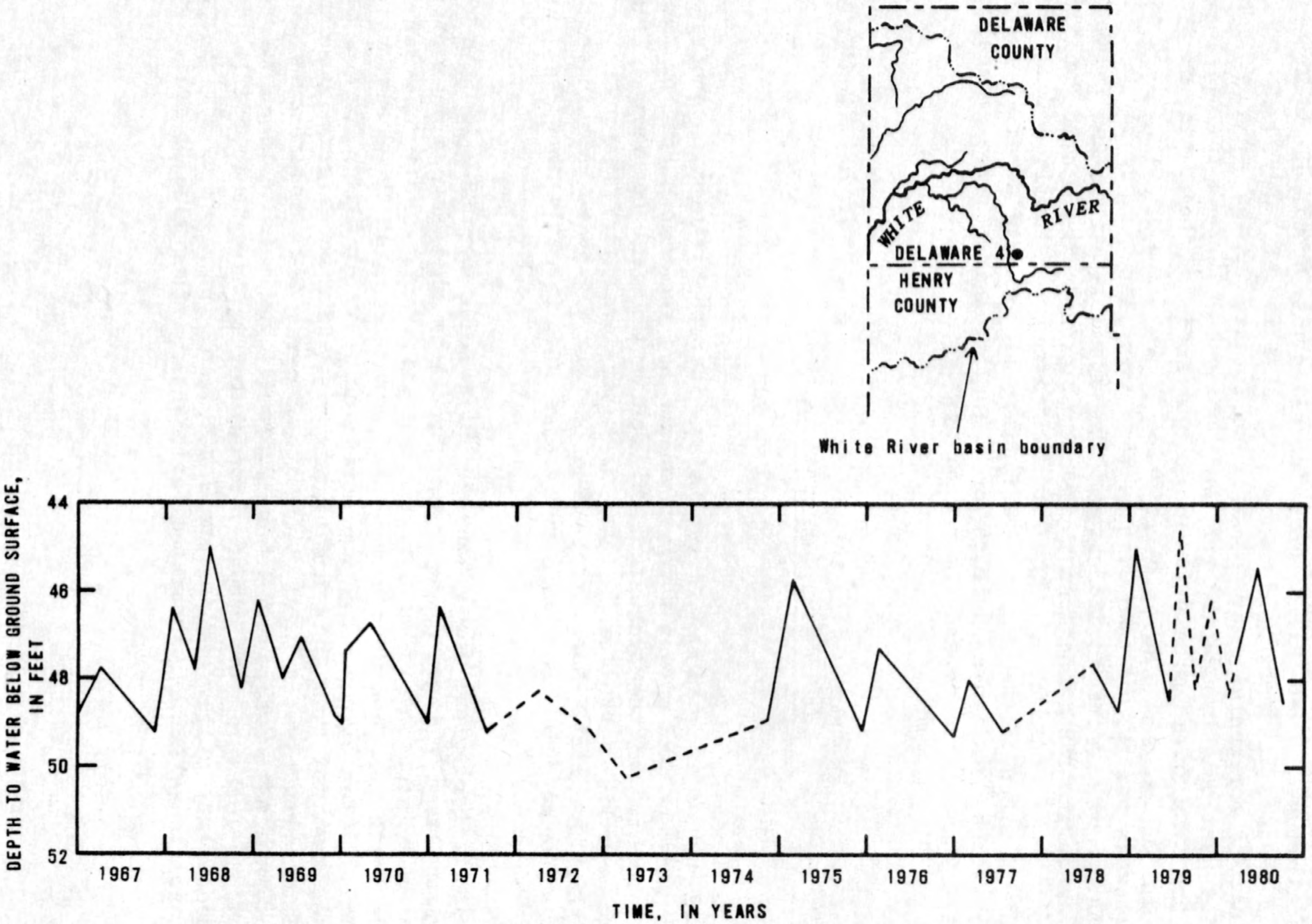


Figure 19.-- Water-level fluctuation in well Delaware 4 tapping aquifer 3.

Areal Recharge

Recharge is only a small part of total precipitation because of losses to surface runoff and evapotranspiration. For the average base period simulated in the model of Marion County by Meyer and others (1975, p. 48), the effective recharge to the till averaged approximately 2 in./yr. On the basis of this recharge, only about 5 percent of the 38.2 in. annual precipitation is recharged to the regional ground-water system. If recharge over the 425-mi^2 Delaware County study area were 5 percent, base flow would be 63 ft^3/s.

The average quantity of precipitation that enters the ground-water system as effective recharge can be assumed to be constant from year to year. This assumption is supported by the water-level trend in observation well Delaware 2 and water-level fluctuations in the domestic and Geological Survey observation wells.

Evapotranspiration in the White River basin north of Johnson County is approximately 65 percent of the annual precipitation (Cable and others, 1971, p. C8). Capture of evapotranspiration by lowering of the water table is possible. However, Meyer and others (1975, p. 38) concluded that the capture in Marion County is small. In Marion County, the water table was sufficiently deep that direct evaporation or transpiration from the water table was minimal (Meyer and others, 1975, p. 38). Transpiration was further assumed to be minimal because most of the vegetation in Marion County derives its water from the unsaturated zone (Meyer and others, 1975, p. 38). No data were collected to verify these assumptions in the Delaware County study area. The capture of evapotranspiration by lowering the water table was not considered in this study.

Ground-Water Seepage to Streams

Local streams generally receive ground-water discharge, except perhaps during short-term high flows, when streams discharge into ground-water systems. The rate of ground-water seepage to streams was determined by simultaneously measuring stream discharge at points along the streams during low flow. Except in sections having simultaneous surface inflows and (or) outflows, the difference in discharge between two adjacent measurements was equal to the ground-water seepage. Seepage within a section was adjusted for surface inflows and (or) outflows. Discharge was measured on the White River at Muncie on October 29, 1977, during base flow and at 80-percent flow duration. Discharge at this flow duration is approximately 18 ft^3/s in the White River at Muncie, 13 ft^3/s in Buck Creek near Muncie, and 5 ft^3/s in Killbuck Creek near Gaston (Horner, 1976, p. 216, 219, and 224). Any increase in streamflow within a section of stream at 80-percent flow duration can reasonably be assumed to be ground-water seepage to the stream, after correction for man-induced inflows and outflows. Ground-water seepage to the streams at 80-percent flow duration on October 29, 1977, was 81 ft^3/s. The stream sections measured are shown in figure 20.

Because of standard error in measuring stream discharge, a range of seepage was calculated by assuming a ± 5-percent error in all discharge measurements. These minimum and maximum seepages are given in table 2. Some of the ground-water seepage rates are approximate (footnote 4, table 2). The rate for these stream sections was estimated by multiplying the ratio of the stream-section length to gaged length times the discharge at the site of the measured streamflow (fig. 19).

Ground-Water Pumpage

As pumpage can be a significant part of total ground-water discharge, historical and current pumpage was surveyed. Historical and current static and pumping water-levels in and near the pumped wells were also surveyed; but little information was obtained. In determinating pumpage, less than 0.1 Mgal/d (0.16 ft^3/s) was not considered significant unless it was part of local pumpage totaling more than 0.1 Mgal/d (0.16 ft^3/s). Locations of significant pumpage during 1976 are shown in figure 21.

From the data collected, the rate of significant ground-water pumping for 1976 was 3.1 Mgal/d (4.8 ft^3/s). This rate is also reasonable for the pumping rate during the past 5 to 10 yr because historical data indicate that the rate has not changed significantly during this time. In fact, the small, scattered pumping would not significantly affect regional flow patterns, even for nonsteady-state conditions. Therefore, because the total pumpage consists of scattered pumpage and because this pumpage has not changed significantly during the past 5 to 10 yr, the ground-water system is assumed to be in equilibrium with the pumpage. This assumption is supported by observation of water-level declines near a large well field tapping aquifer 1 northwest of Anderson in Madison County. Distance of wells from the White River ranged from 2 to 7 mi. According to a study done for the well-field operator, water levels generally had stabilized within 2 yr after pumping about 4 Mgal/d began (Donald Davis, oral commun., 1980).

Although the Madison County well-field drawdown should be fairly representative of how quickly the ground-water flow system stabilizes to the pumping simulations described in this report, specific factors control the response at any one site. Water-level stabilization is a function of (1) distance to the boundary from where recharge is derived, (2) the hydraulic properties of the medium between the pumpage and the source of recharge, and (3) the storage coefficient of the medium.

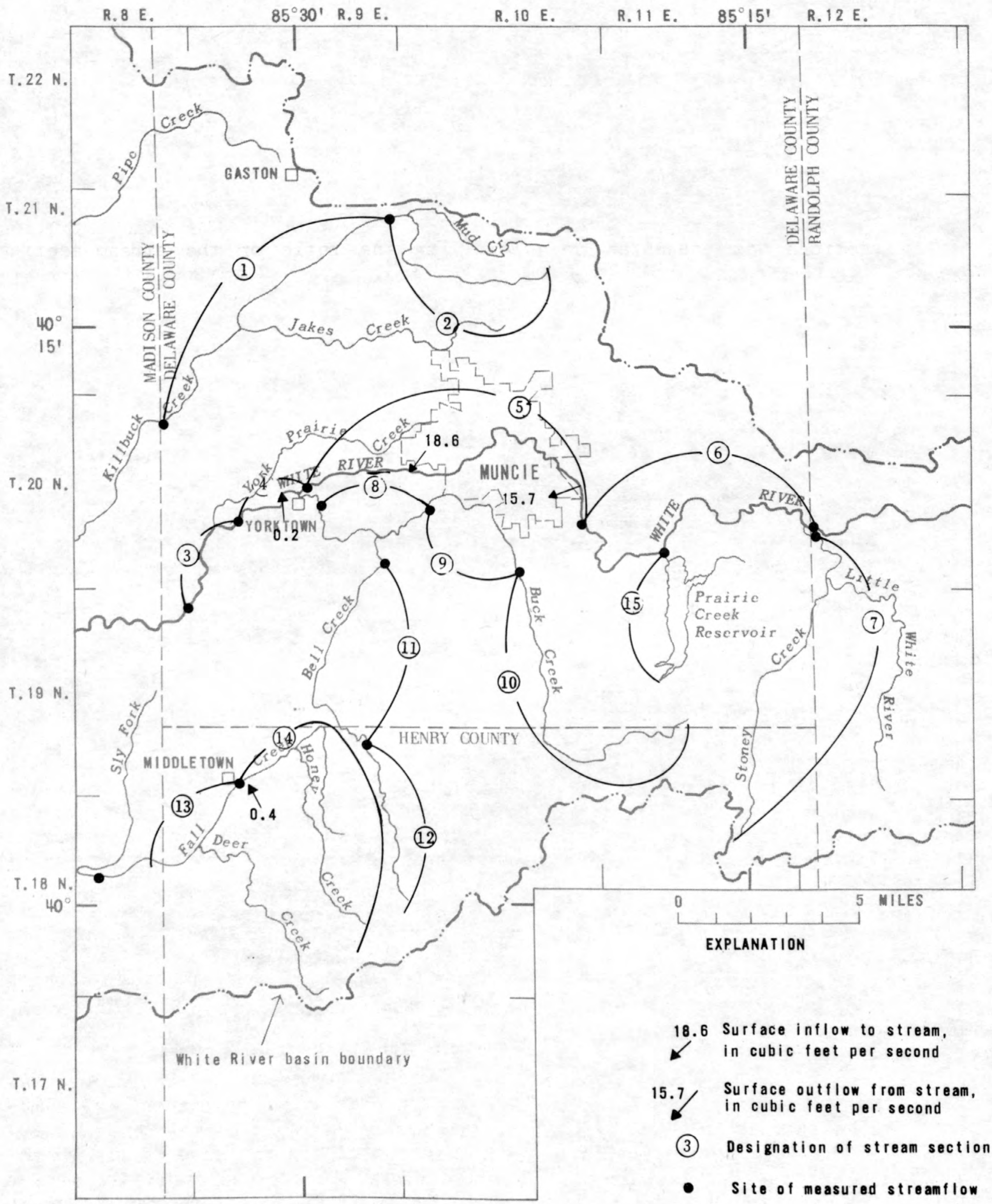


Figure 20.-- Stream sections where ground-water seepage was calculated, October 29, 1977.

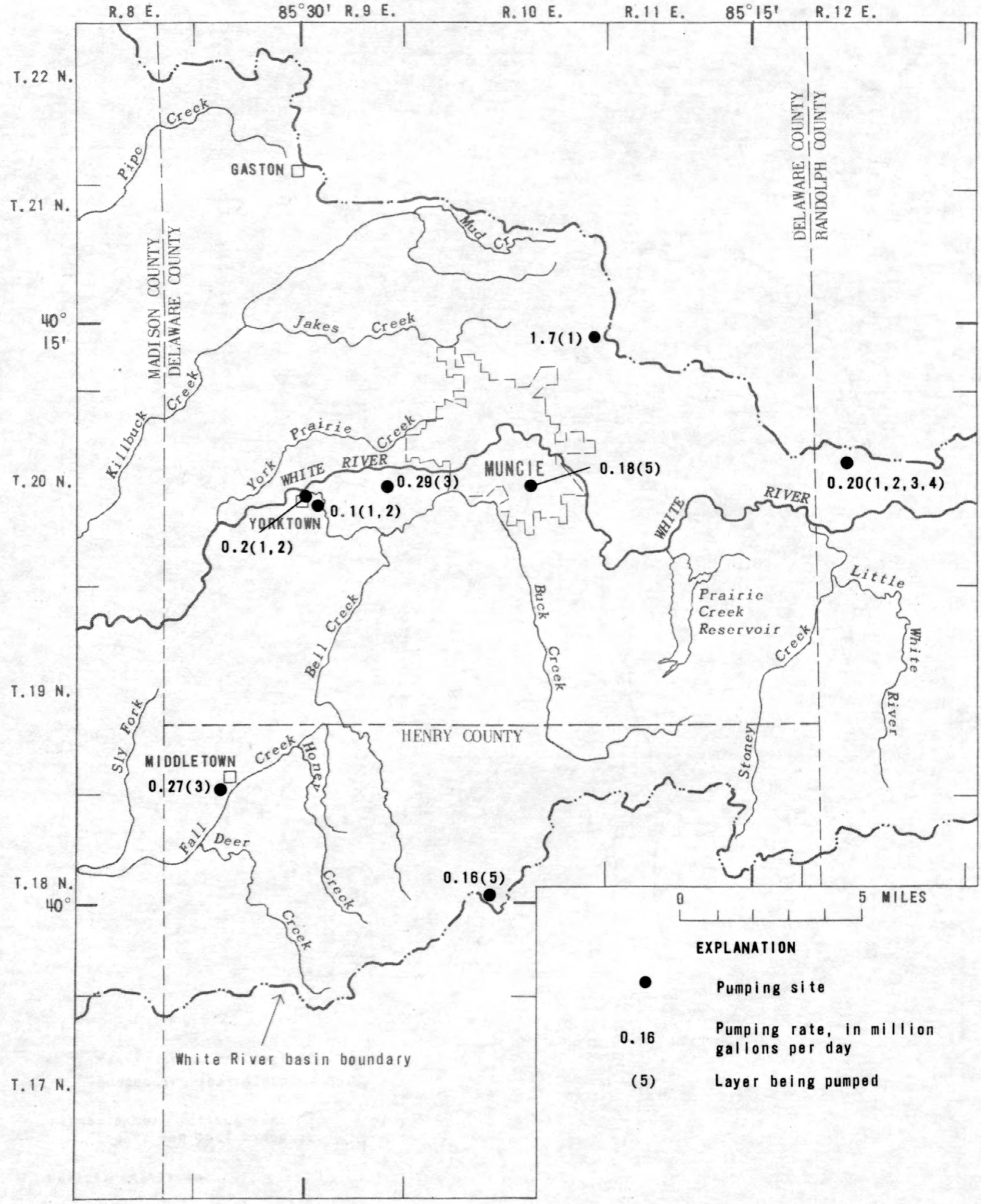


Figure 21.-- Significant ground-water pumping in Delaware County study area, Indiana, 1976.

During 1976, municipal pumpage averaged 0.77 Mgal/d (1.2 ft^3/s), and industrial pumpage averaged 2.3 Mgal/d (3.6 ft^3/s). The rate of domestic pumpage was not determined because a count of private wells was not available. Total domestic pumpage may be a significant amount relative to municipal and industrial pumpage. However, domestic pumpage is scattered and individually is too small to be counted. Even in heavily populated areas, domestic pumpage is small because these areas depend on municipally supplied water.

MODEL SIMULATION OF GROUND-WATER FLOW

Model simulation of ground-water flow provides flow rates of the components of the ground-water budget that were not directly obtained from field data or previous studies. In the present study, data have been collected for calculating the rate of ground-water seepage to streams and the average annual rate of pumping. At 80-percent flow duration in October 1977, rate of ground-water seepage to streams was 81 ft^3/s. Annual average rate of pumping for 1977 was 4.8 ft^3/s (3.1 Mgal/d). Rates of effective recharge to the ground-water system at 80-percent flow duration and ground-water flow across study-area boundaries were not determined. Thus, a complete ground-water budget based on field-collected data was not determined. However, model simulation of the ground-water flow yields flow rates for all the components of the ground-water budget and calculates the effects of pumping on ground-water levels and on streamflow.

Simplifying Assumptions for Model Simulations

Simplifying assumptions for the geometry and the hydraulic characteristics of the ground-water system were made to simulate the ground-water flow. These assumptions result in a model that simplifies the actual ground-water system (fig. 22) while retaining all pertinent characteristics. The assumptions made in developing the model are:

1. Flow in the drift is quasi-three-dimensional, horizontal in and between the sand and gravel aquifers and vertical (leakage) between layers.

2. Flow in the bedrock is horizontal, except where the bedrock occupies more than one layer. (See the next section.)

3. The six confined sand and gravel aquifers are homogenous and horizontally isotropic, and their hydraulic conductivity is 433 ft/d.

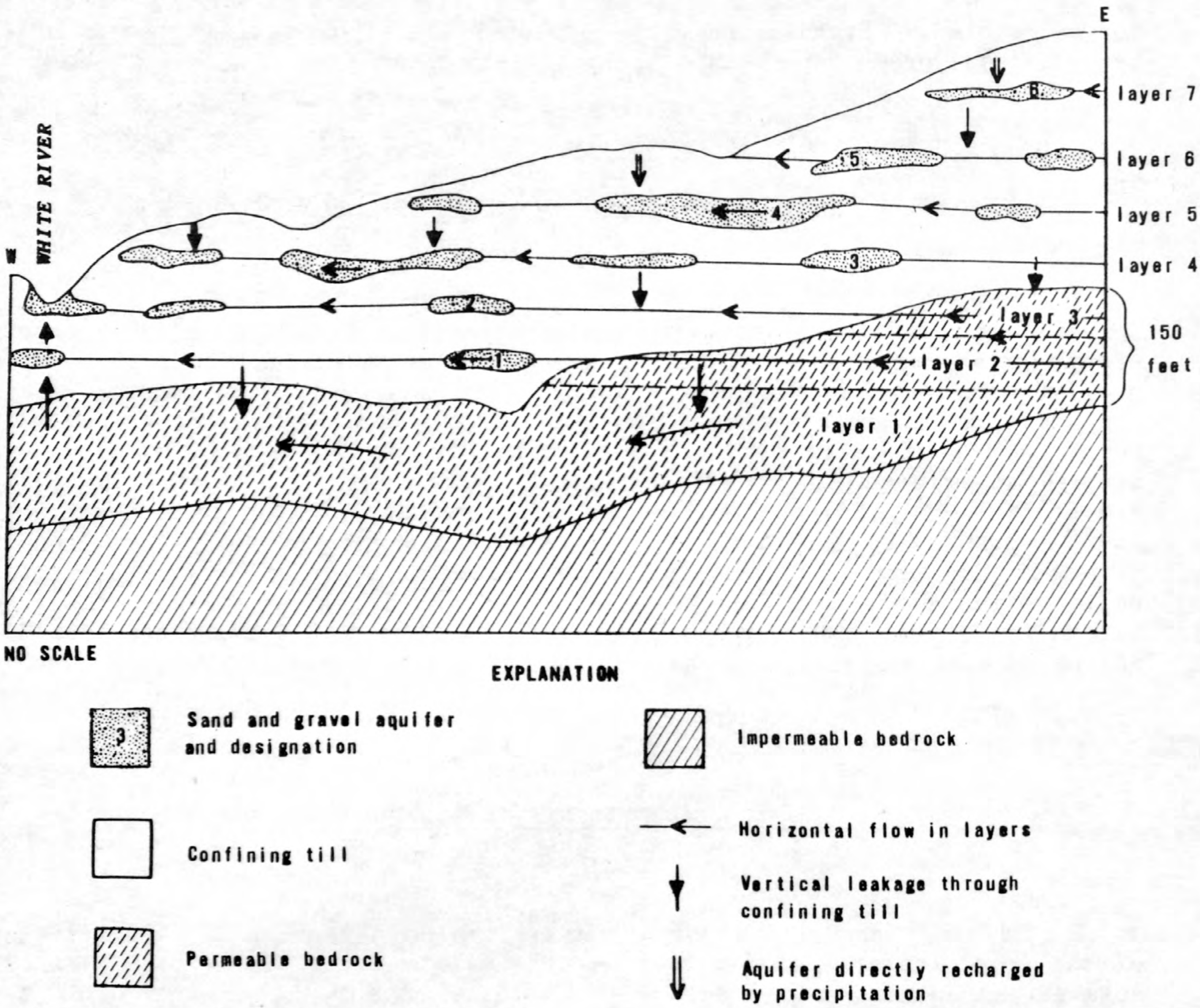


Figure 22.-- Generalized geologic section of study area showing the design of the model.

4. The ratio of horizontal to vertical hydraulic conductivity of the bedrock aquifer is 1:1. The hydraulic conductivity of the bedrock aquifer is constant with depth but is areally variable.

5. Only the upper 150 ft of the bedrock aquifer is permeable.

6. Streambeds are 1 ft thick and are composed of a material of lower vertical hydraulic conductivity than that of the aquifers.

7. Some minor streams are insignificant discharge points for the ground-water system and can be ignored. (This simplification eliminates some shallow ground-water circulation.)

8. The ground-water system is in steady state.

9. Domestic pumpage does not significantly affect the flow system at any one point and is accounted for by a reduction in recharge.

Model Selection, Design, and Construction

A slightly altered version (D. G. Gillies, written commun., 1978) of the quasi-three-dimensional finite-difference model of Trescott (1975) was used to simulate ground-water flow. This model achieves a more realistic approximation of the ground-water flow system than the original model does. This alteration included the flexibility to simulate recharge to any model layer and to simulate streams in any model layer.

The finite-difference grid used is shown in figure 23. The grid consists of 1,798 (31 x 58) block-centered nodes covering 410 mi^2. Grid spacing ranges from 1,000 to 4,000 ft, and grid-block area ranges from 0.14 to 0.57 mi^2.

A variable grid was designed to provide additional detail in areas of special interest. The small grid spacing around the White River provided more resolution for simulating and detailing the effects of future pumping on the river. The variable-grid spacing also provided detail in some areas of large head gradients and thus allowed these areas to be simulated more accurately in the model.

The area was divided into seven layers (fig. 22) for modeling the ground-water flow. Layer 1 generally represents the bedrock aquifer. Layers 2 through 7 generally represent either sand and gravel aquifers 1 through 6, or till. However, the lower layers sometimes represent the bedrock aquifer.

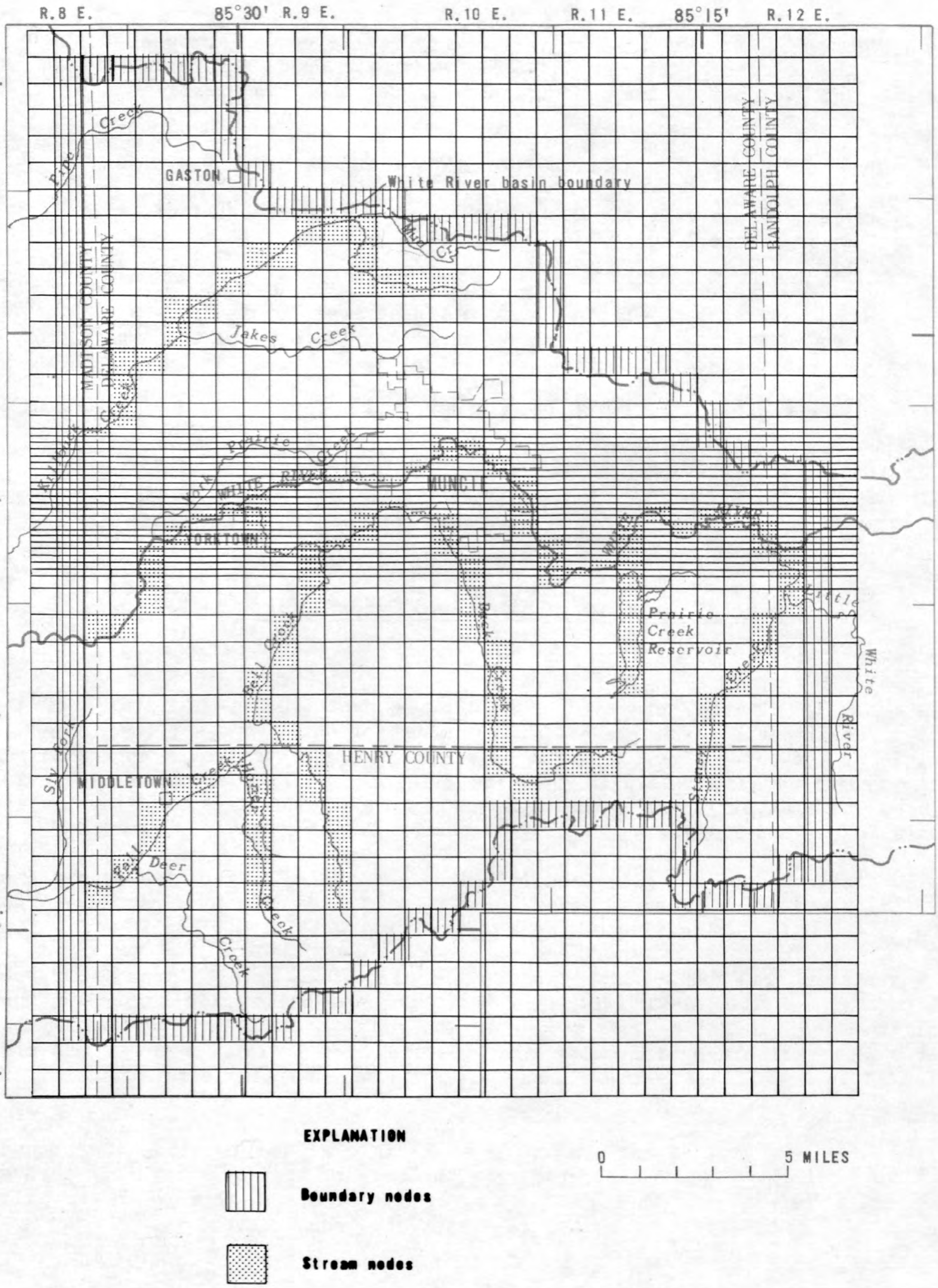


Figure 23.-- Finite-difference grid used to simulate ground-water flow in the study area.

The transmissivity of the bedrock aquifer for each node (center point of grid block) in the model in layer 1 was assigned by overlaying the grid on the bedrock transmissivity map (fig. 18) and estimating the average transmissivity within each grid block. Some of the upper parts of the 150 ft of permeable bedrock were laterally within one or more of the layers above layer 1 (fig. 22). For this condition, the total bedrock transmissivity (normally assigned completely to layer 1) was divided among each of the layers containing bedrock in proportion to the thickness of bedrock in each layer.

A grid overlay was also used to assign transmissivity values to aquifers 1 to 6. However, the transmissivities could not be estimated directly. Figures 6, 8, 10, 12, 14, and 16 were used first to estimate the thickness of aquifers 1 through 6 for each node in layers 2 through 7, respectively. These thicknesses were then multiplied by 433 ft/d, the average conductivity for confined sand and gravel aquifers, to calculate the appropriate transmissivity.

Because the six sand and gravel aquifers are areally discontinuous, some areas in layers 2 to 7 consist of till or bedrock, rather than sand and gravel (fig. 22). In these areas, transmissivities were assigned as follows: Till, for example, was assigned a transmissivity of 2.8 ft^2/d. This transmissivity was based on a hydraulic conductivity equal to 0.14 ft/d (1 gallon per day per square foot). The value is within the range of hydraulic conductivities for till given by Freeze and Cherry (1979, p. 29) and Todd (1959, p. 53) and is assumed to represent the average hydraulic conductivity of the till. The conductivity was then multiplied by 20 ft, the average thickness of till separating the laterally discontinuous sand and gravel aquifers. This procedure provided a small but finite transmissive connection laterally between the discontinuous parts of each aquifer. Zero transmissivities were imposed in each layer beyond the point where the layer intercepted land surface.

Vertical flow in the ground-water system was simulated as leakage between model layers. The leakage coefficients for each model node were calculated by first determining the thickness of the least permeable material (till, bedrock, or sand and gravel) between model layers and then dividing the vertical hydraulic conductivity estimated for that material by the thickness. The vertical hydraulic conductivity of the bedrock was calculated individually for each bedrock-leakage coefficient. In this calculation, the ratio of vertical to horizontal hydraulic conductivity of the bedrock was assumed to be 1:1. The vertical hydraulic conductivity was then calculated by estimating the average bedrock transmissivity at each node (fig. 18) and dividing that transmissivity by the assumed thickness of permeable bedrock, 150 ft. The vertical hydraulic conductivity of sand and gravel was based on the assumption that the ratio of horizontal to vertical hydraulic conductivity of the sand and gravel was 10:1. Because the average hydraulic conductivity of the sand and gravel was 433 ft/d, the first calculation of the vertical hydraulic conductivity of the sand and gravel was 43 ft/d. The first calculation of the vertical hydraulic conductivity of the confining beds of till, 7×10^{-4} ft/d, was the average of the range for vertical hydraulic conductivity for the confining beds reported by Meyer and others (1975, p. 26).

The areal distribution and the rate of ground-water seepage to streams were simulated. All streams were modeled as leakage boundaries (fig. 23). The head of the stream remained at the stream-surface altitude measured October 29, 1977, for each stream node.

The first streambed-leakage coefficients were calculated by dividing an assumed streambed thickness, 1 ft, into the assumed streambed vertical hydraulic conductivity, 4×10^{-2} ft/d, an average value for a clay layer in the outwash reported by Meyer and others (1975, p. 19). Although thickness and conductivity vary, adjustments to leakage coefficients were assumed to be related to a change in conductivity.

Ground-water pumpage simulated as part of the steady-state ground-water system was 1.25 Mgal/d (1.93 ft^3/s) and was distributed as shown in figure 21. The model did not simulate all this pumpage because the modeled area did not include all the pumping sites.

Lateral model boundaries were simulated as boundaries of constant potentiometric head in the perimeter of all seven layers containing aquifer or till. The potentiometric head at each boundary node in each layer was based on water-level maps for each aquifer. Using these specified heads, the model computed ground-water flow across the model boundaries. Because this boundary flow (or flux) at nodes representing till was very small, parts of the boundaries containing till could probably have been modeled as no-flow boundaries. Almost all the boundary flux crosses boundaries at aquifers. The upper boundary was always an active model layer that received recharge. At the lower boundary, vertical flow into layer 1 was zero.

Model Calibration

The steady-state flow model was calibrated to a set of potentiometric heads and seepage rates to streams. The model was assumed to be calibrated when model-simulated potentiometric heads and seepage rates sufficiently matched corresponding measured potentiometric heads and estimated seepage rates.

Model-simulated potentiometric heads were matched to heads measured during autumns 1976, 1977, or 1978. Whether heads measured in a 2-yr range can be used as one set of water levels may be questionable, even if they are measured at the same time of year. Variation in the head in each observation well measured during at least two of the three autumns averaged only 2.2 ft for the entire project area. On the basis of the Marion County study by Meyer and others (1975, p. 48), the author expected a maximum difference of about 10 ft between model-simulated and field-measured heads. Because the average difference in head measured during the three autumns was well within the match of model-simulated and measured water levels expected by the author, all measurements made in the autumns 1976-78 were used collectively as one set of water-level measurements.

Model-simulated, ground-water seepages to streams were matched to seepages for various sections of stream (table 2). The seepage was determined from stream discharge measured on October 29, 1977, when flow duration of all streams was about 80 percent.

Calibration consisted of adjusting hydrologic variables in the model until model-simulated heads and seepages matched the field measurements. The transmissivity distributions of the aquifers and boundary heads were based on data collected in the field. Vertical hydraulic conductivity of the till, areal recharge to the ground-water system, and the streambed-leakage coefficient, however, were not based on field data and were less well defined. Therefore, most changes during model calibration involved the last three variables.

Final calibrated values of the vertical hydraulic conductivity of the till generally ranged from 7×10^{-4} to 7×10^{-2} ft/d and averaged 4×10^{-2} ft/d.

Effective recharge in the calibrated model ranged from 0.04 to 0.4 ft/yr and averaged 0.15 ft/yr. The range in recharge from precipitation used in the calibrated model is consistent with the 0.08 to 0.67 ft/yr range that Meyer and others (1975, p. 48) used for the till in their calibrated model of Marion County. The range is probably due, at least partly, to areal differences in the slope of the land surface and the vertical hydraulic conductivity of the till above the top aquifer, which cause variations in the rate of infiltration. The effective recharge from precipitation represents the recharge to the regional flow system and does not include recharge that circulates in the shallow ground-water system and discharges locally to streams that were not modeled. However, the recharge that discharges locally as shallow ground-water circulation is probably insignificant relative to the recharge that enters the regional flow. The effect of a different rate of recharge to the outwash aquifer was investigated in the Madison County model (Lapham, 1981, p. 49) because large differences in recharge between the outwash aquifer and the confined sand and gravel aquifers were simulated in Hamilton (L. D. Arihood, written commun., 1980) and Marion Counties (Meyer and others, 1975, p. 48). In Madison County however, model results indicated that a difference in the effective recharge to the outwash from that used as effective recharge to the confined aquifer produced neither any significant difference in the distribution of potentiometric head nor ground-water seepage. The probable reason for the small differences is that the outwash deposits cover only a small percentage of the modeled area. Outwash deposits were not considered to be extensive in Delaware County either. Therefore, no difference in recharge between the outwash and the confined sand and gravel aquifers was simulated in the Delaware County model.

A streambed thickness of 1 ft and a vertical hydraulic conductivity of 4×10^{-2} ft/d were assumed in calculating the first vertical-leakage coefficient of the streambeds. This conductivity implies that all streambeds were composed of material having a low vertical hydraulic conductivity. Adjustments during calibration resulted in hydraulic conductivities ranging from 7×10^{-4} to 80 ft/d for a 1-ft thick streambed. This thickness was chosen for convenience. Other streambed thicknesses could have been assumed, but the resulting change in streambed conductivity would be small relative to the range of vertical hydraulic conductivity of the streambeds.

Final transmissivity distributions for the bedrock and the confined sand and gravel aquifers after calibration differed only slightly from the original ones. Because calculation of the transmissivities for the sand and gravel aquifers were based on an average hydraulic conductivity, the transmissivity distributions used in the model are estimates of the transmissivities.

Model-simulated, steady-state potentiometric surfaces for the bedrock and the six sand and gravel aquifers are shown in figures 24-30. Generally, the match between model-simulated heads and the corresponding measured heads was within 10 ft. Because model calibration involved measuring and matching heads in the aquifers only, how well the model simulates heads in the non-aquifer material (dashed lines in figs. 24-30) is not known. The dashed lines are included to show the regional flow pattern and the distribution of head in the till and (or) bedrock connecting the discontinuous parts of each aquifer. The model-simulated ground-water flow pattern in the bedrock aquifer indicates that ground water discharges locally to streams in the study area (fig. 24).

Model-simulated and field-measured seepages for streams are shown in table 2. Where possible, the model-simulated rates were matched to rates within the maximums and minimums of field-measured rates. Maximums and minimums were determined by assuming a ±5-percent error in all the discharge measurements from which the seepage to the stream section was calculated. Otherwise the rates were matched as close to the measured rates as possible. With few exceptions, model-simulated rates matched the field-estimated rates within ±5 percent. Lack of a match in some sections, even within ±5 percent, is probably due to the inability of the model to simulate the flow system in enough detail in those sections. However, these differences should not seriously limit the usefulness of the model in simulating the effect of future stresses on ground-water levels and streamflow. The apparent loss of flow in stream section 3 is probably due to measurement error, and the model-simulated seepage is probably closer to the true seepage than the measured one.

The water budget for the calibrated model is shown in table 3. This tabulation represents the distribution of ground-water inflow and outflow during calibration. The data indicate that 80 percent of the inflow to the modeled ground-water system is effective recharge from precipitation and 20 percent is flow across boundaries. Outflow consists of 2 percent pumpage, 66 percent seepage to streams, and 32 percent is flow across boundaries. The 81.9-ft^3/s precipitation recharge results in a 2.7-in./yr average areal recharge. This water budget corresponds to low flow (80-percent flow duration) and consequently should provide conservative estimates of the effects of pumpage on ground-water levels and streamflow.

Distribution of the rates of inflow and outflow in the calibrated model is consistent with field conditions discussed in the section, "Ground-Water Flow." Both regional- and local-flow patterns are evident in Delaware County. Thirty-two percent of flow from Delaware County is across boundaries. Eight percent of outflow from Madison County, where discharge is mostly to local streams (Lapham, 1981, p. 60), is across boundaries. Outflow across the Randolph County boundary is 69 percent of the total outflow (W. W. Lapham, written commun., 1980), which shows regional flow out of that study area is dominant.

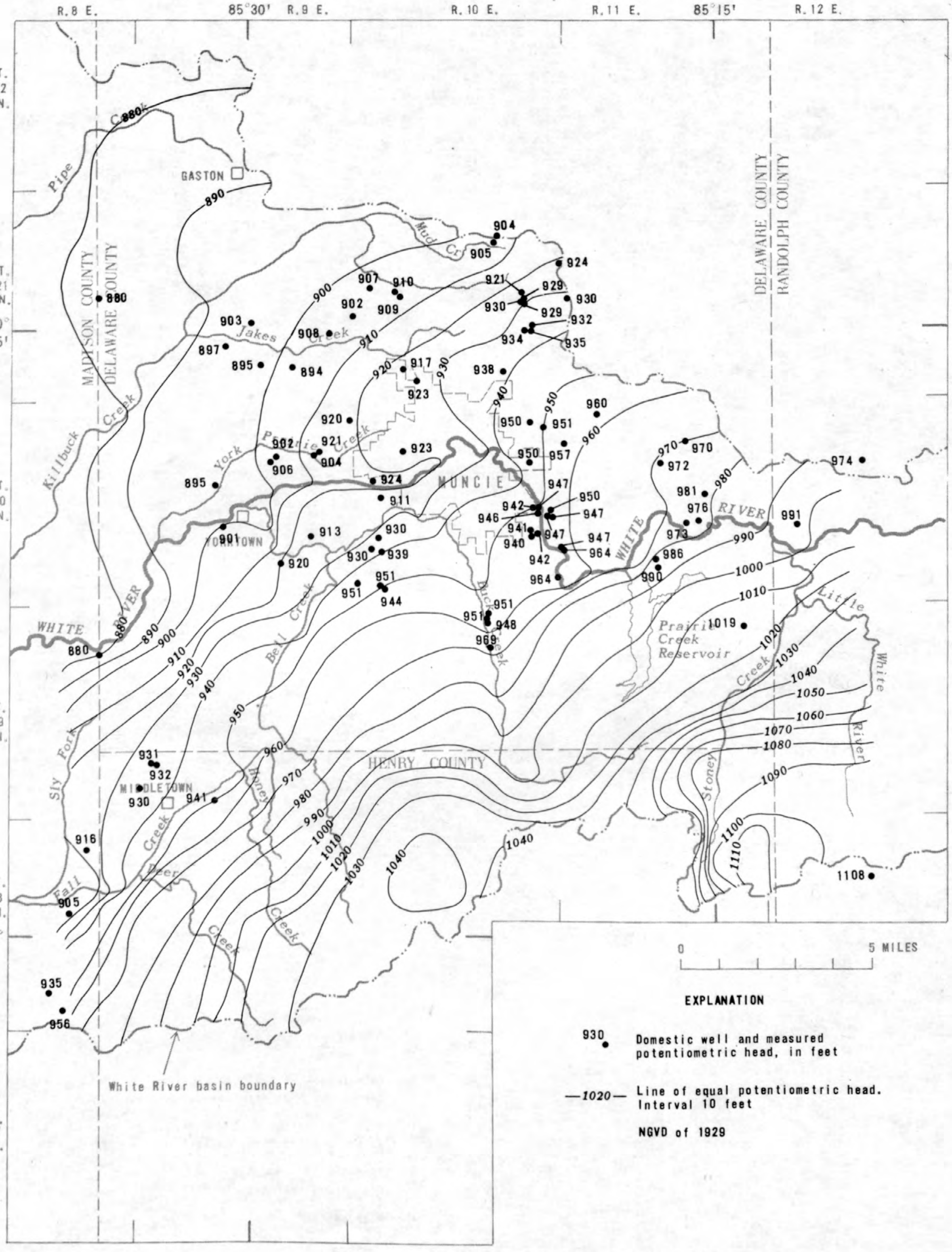


Figure 24.-- Model-simulated, steady-state potentiometric surface of the bedrock aquifer.

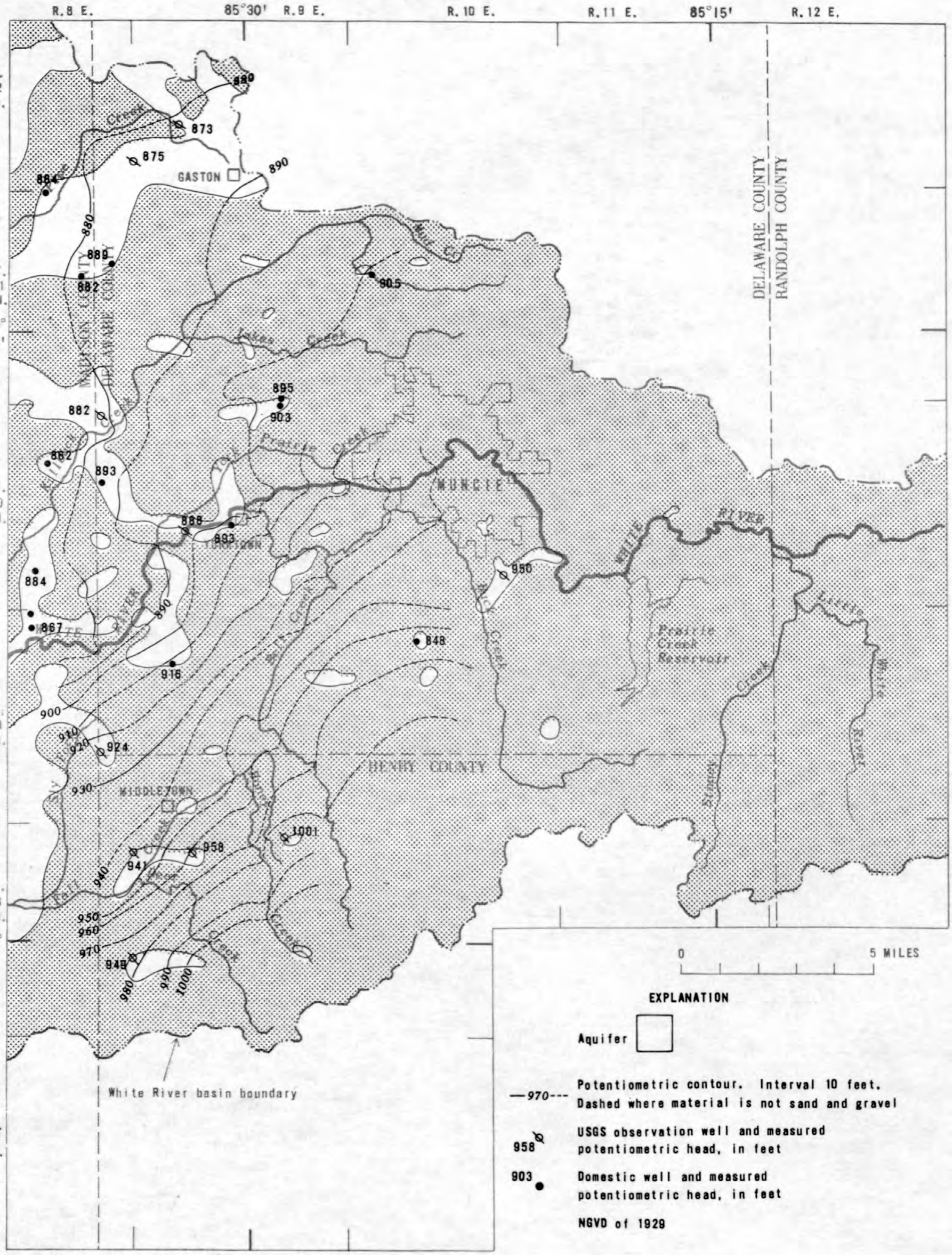


Figure 25.-- Model-simulated, steady-state potentiometric surface of aquifer 1.

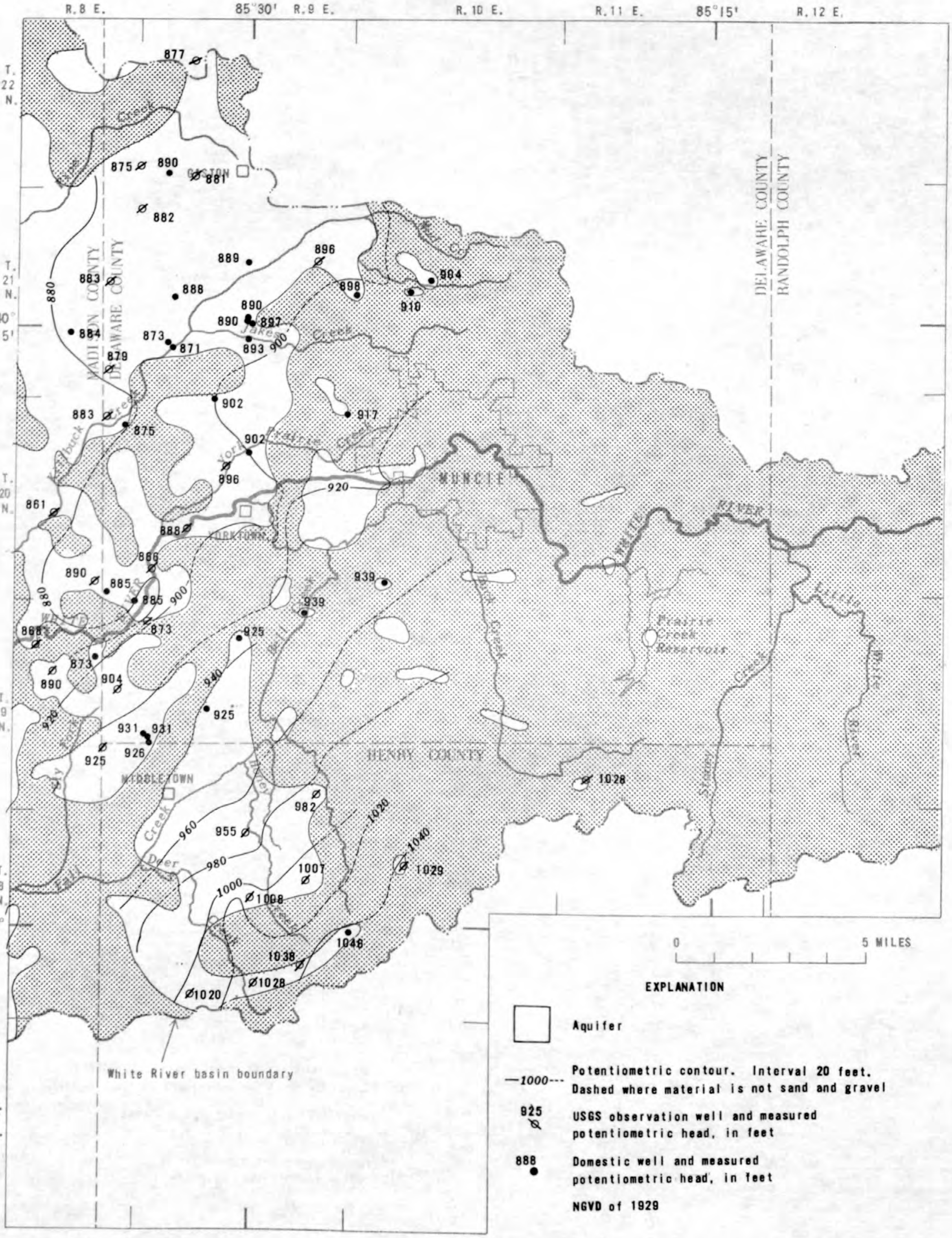


Figure 26.-- Model-simulated, steady-state potentiometric surface of aquifer 2.

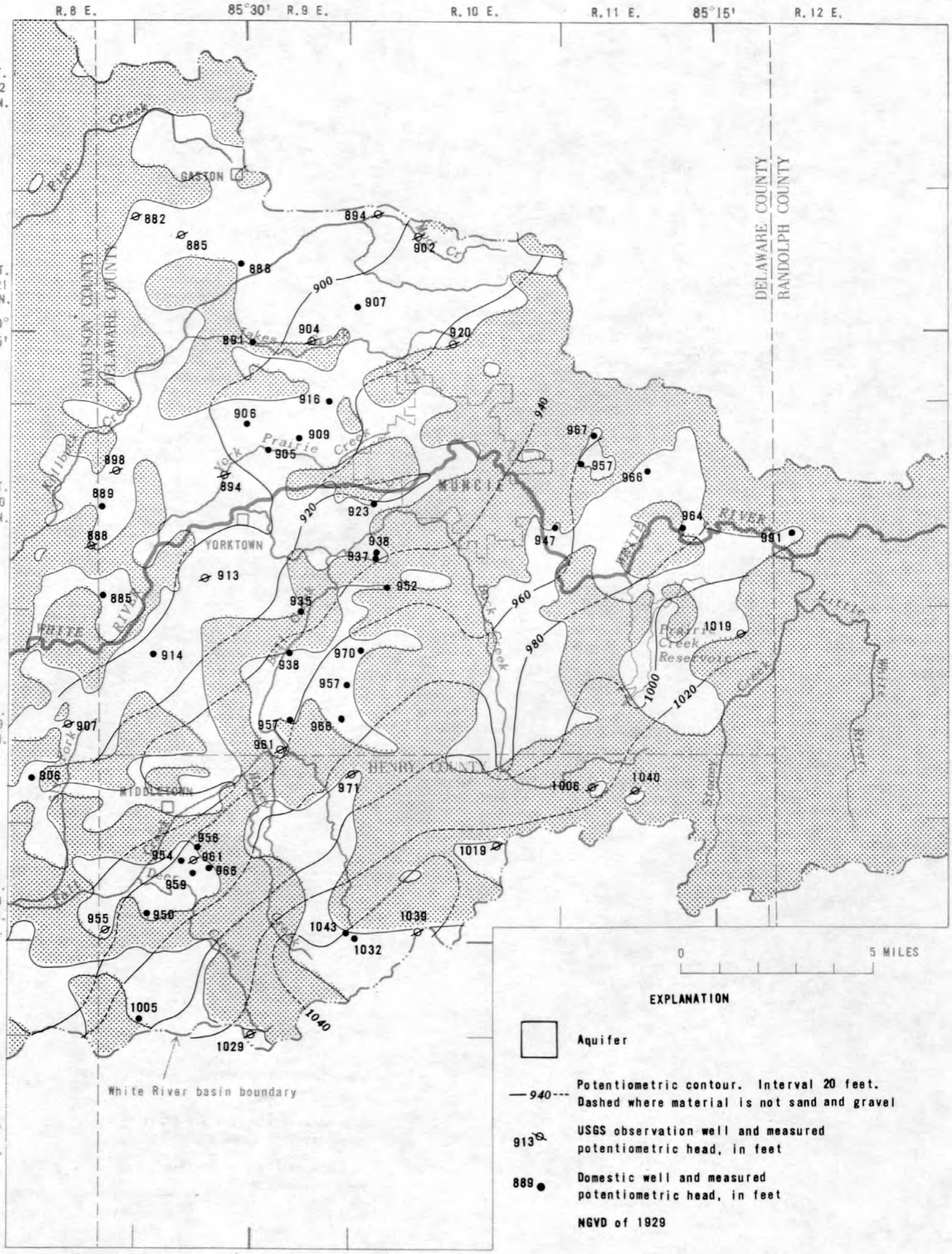


Figure 27.-- Model-simulated, steady-state potentiometric surface of aquifer 3.

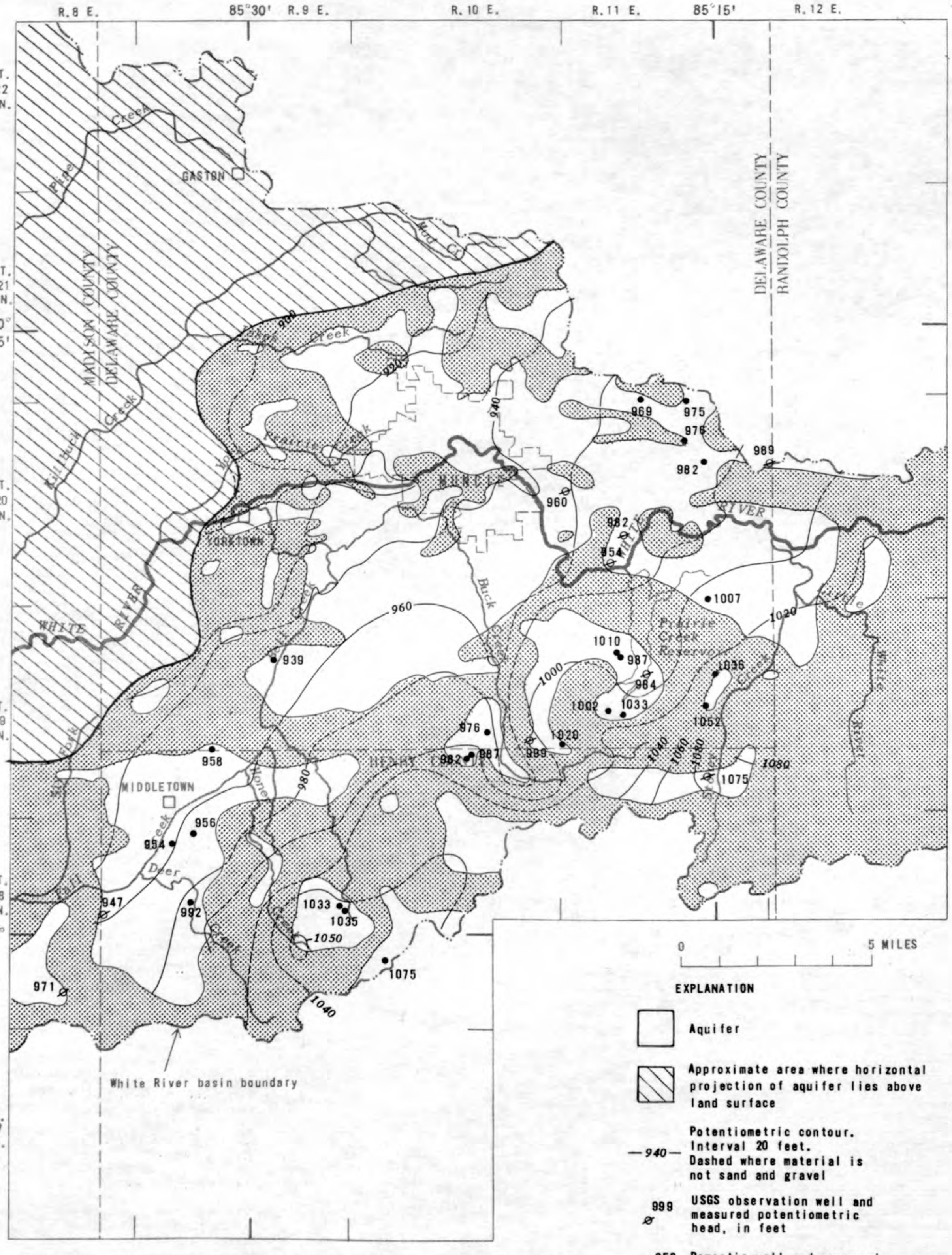


Figure 28.-- Model-simulated, steady-state potentiometric surface of aquifer 4.

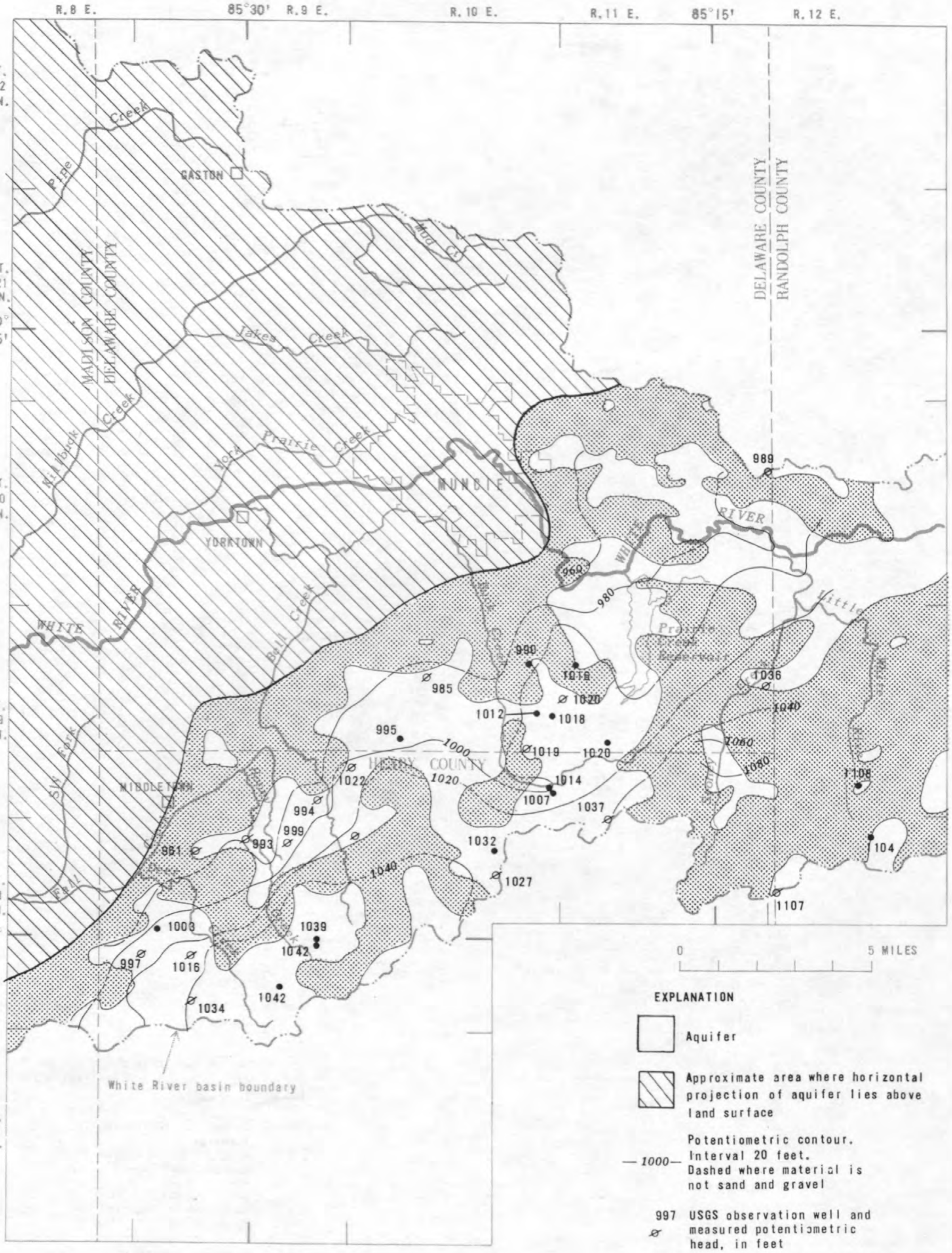


Figure 29.-- Model-simulated, steady-state potentiometric surface of aquifer 5.

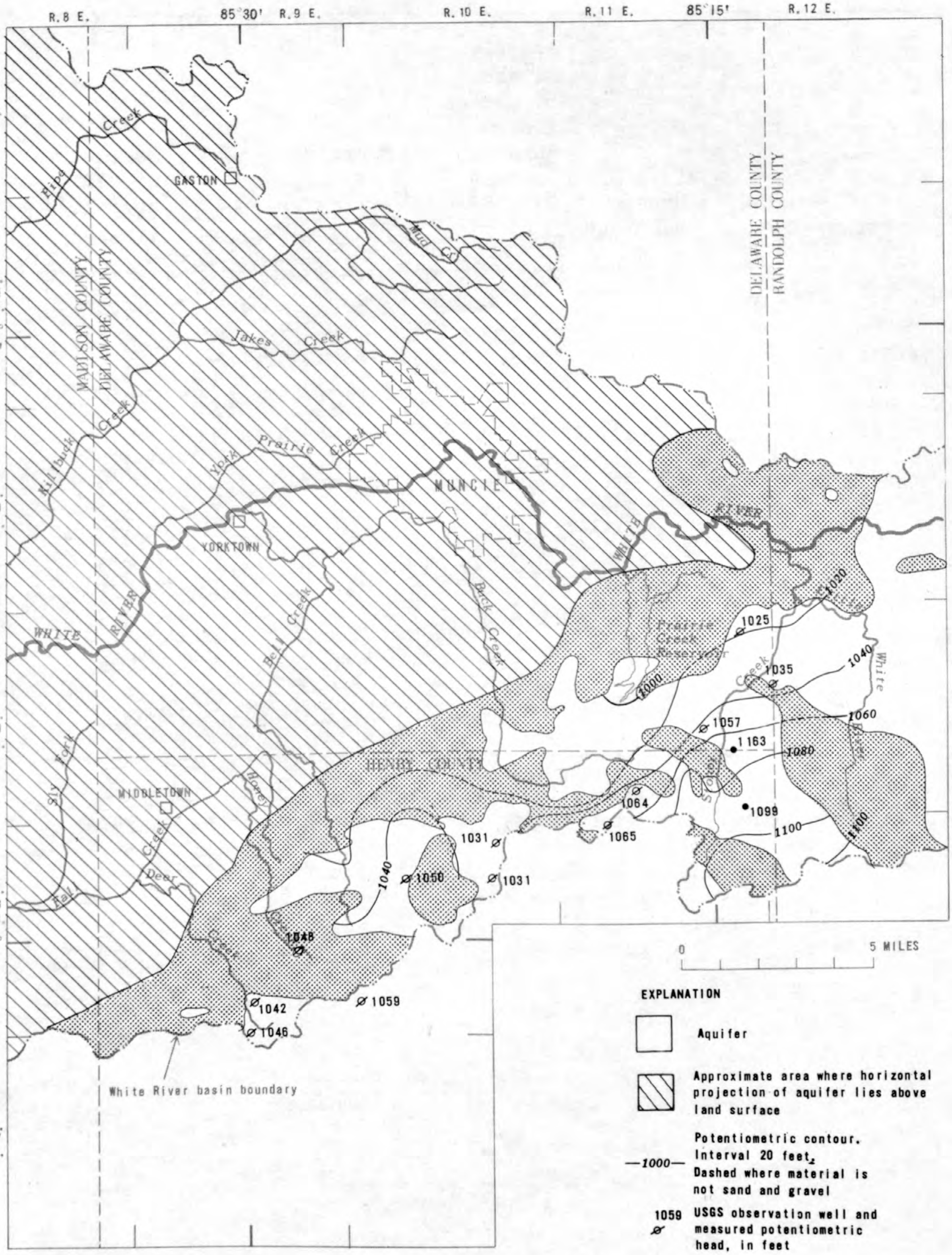


Figure 30.-- Model-simulated, steady-state potentiometric surface of aquifer 6.

Table 2.--Model-simulated and measured ground-water seepage for streams modeled in the study area

Stream name	Stream section[1]	Seepage at about 80-percent flow duration, October 29, 1977 (ft^3/s)	Measured rates of seepage[2] (ft^3/s)		Model-simulated seepage[3] (ft^3/s)
			Minimum	Maximum	
Killbuck Creek	1	4.1	3.7	4.5	4.3
Do.	2	.9	.8	1.0	.95
White River	3	-8.5	-16	-1.1	2.6
Do.	4	.7	-6.2	8.0	1.5
Do.	5	4.3	.5	8.1	6.8
Do.	6	10.9	8.1	14	9.0
Stoney Creek	7	5.1	4.8	5.4	4.9
Buck Creek	8	.2	-2.9	3.3	.71
Do.	9	2.0	.3	3.7	3.2
Do.	10	15.5	15	16	10
Bell Creek	11	3.9	3.2	4.7	4.3
Do.	12	1.9	1.8	2.0	1.1
Fall Creek	13	[4]2.5	-----	----	2.9
Fall Creek and Honey Creek	14	5.4	5.1	5.6	4.3
Prairie Creek and Reservoir	15	[4]4.0	-----	----	3.0

[1]For location of stream sections, see figure 19.
[2]Negative number indicates seepage is into the ground-water system from the stream.
[3]Total model-simulated seepage in this table differs slightly from ground-water seepage in streams in table 3 because of round-off error.
[4]Estimated.

Table 3.--Steady-state water budget of the calibrated model at 80-percent flow duration

	Recharge (ft^3/s)		Discharge (ft^3/s)
Effective recharge from precipitation	81.9	Ground-water pumpage	1.93
		Ground-water seepage to streams	66.7
Ground-water flow across model boundaries into study area	20.5	Ground-water flow across model boundaries out of study area	32.7
Total recharge	102	Total discharge	101

ASSESSMENT OF GROUND-WATER AVAILABILITY

Several pumping plans (table 4) were simulated in the three major aquifer systems to determine their effect on water levels and streamflow. The three systems are the confined sand and gravel aquifers within the till, the bedrock aquifer, and the outwash aquifer associated with the major streams. The model used in the simulations was designed to simulate stresses and determine system response on a regional scale, but it does not provide the detail of drawdown needed, for example, in a pumping plan having only one well that causes a drawdown cone covering only one or two model nodes.

Constant-flux boundaries were used for all layers in all pumping simulations, and all simulations were run to steady state. These conditions produced maximum water-level declines and streamflow depletion for a given pumping simulation. The constant-flux boundaries allow no inducement of additional ground-water flow across the boundaries, and steady state allow no water to be derived from storage. Although no transient simulations were made, steady state is probably reached within 5 yr or less for the pumpage rates simulated. This conclusion is based on the time for water levels to stabilize in and near the well field northwest of Anderson. (See the section, "Ground-Water Pumpage.") The water levels seemed to stabilize within 2 yr after pumpage of 4 Mgal/d began.

To ensure that the constant-flux boundary was not significantly affecting the results of the pumping simulations, the authors also simulated constant-head boundaries. A constant-head-boundary simulation was made if the water-level decline at any boundary for the constant-flux boundary simulation was more than about 1 ft. This boundary condition tends to minimize water-level changes attributable to pumpage by maintaining a constant head and, thus, zero drawdown at the model boundaries. If the results of a simulated constant-flux boundary are virtually identical with those of a simulated constant-head boundary, the results are not being affected by the model boundaries. The results of both simulations are included in the report if they differ.

The model-simulated drawdowns and streamflow depletions discussed in this report are caused only by the simulated pumpage. The simulations do not account for any other stresses on the ground-water system within the modeled area except those simulated during model calibration. For example, if pumpage of the ground-water system outside the modeled area increases, the boundary fluxes simulated in the calibrated model may change. This change may alter the distribution of flow within the study area from that established during calibration.

Because model calibration involved matching heads in the aquifers to simulated heads, the model's effectiveness in simulating drawdown in areas not shown as aquifer is not known. Also, the results of the model simulations should not be accepted as precise predictions of what will happen in the field, but rather an estimate of what will happen.

Potential for Ground-Water Development in the Three Major Aquifer Systems

To investigate the potential for ground-water development in the three major aquifer systems, the author used seven pumping plans, A through G. (See fig. 31 and table 4.) Although other locations have potential that is probably equal to that of the locations investigated by plans A through G, the location chosen for each pumping plan represents one of the areas of greatest potential for the particular aquifer system investigated.

Each pumping plan was designed to investigate the potential yield from an aquifer system (bedrock, confined sand and gravel, or outwash) having a specific geometry under set conditions such as proximity to a major discharge area. The yields of each plan could then be compared. Plan A was designed for determining the hydrologic response to pumping from permeable bedrock. Plan B was used to investigate the response of a thick sand and gravel aquifer for comparison with plan A. Plan D was chosen to investigate the hydrologic response of an aquifer similar to the one in plan B, except that in plan D the aquifer was distant from the White River. In plans C and E, pumping was from one location; but in plan E pumping was in the White River outwash, whereas in plan C pumping was in the next lower aquifer. Plan F was designed to investigate the effects of pumping in a sand and gravel aquifer having a reservoir as a nearby discharge area. Finally, plan G was used to determine the effects of developing a larger quantity of water by pumping over a larger area than in the other plans.

The pumping plans were assessed in terms of the quantity of ground water pumped and the associated drawdowns and streamflow depletions. The pumpage in the plans was limited to the quantity that would cause and maintain a drawdown of 20 ft in the node in which pumping was simulated. In practical terms, a limit of 20 ft of drawdown in the ground-water system alleviates the potential problem of pumping the ground-water system at rates that significantly affect nearby wells already in use.

The distribution of drawdown due to the pumping plans, for the aquifer pumped in each plan, is shown in figures 32-40. For a given plan, drawdowns in aquifers other than the aquifer or aquifers in which pumpage was simulated are less than the drawdowns shown for that plan. Because model calibration involved measuring and matching ground-water heads in the aquifers only, the model's effectiveness in simulating drawdown in the nonaquifer material, shown as dashed lines in figures 32-40, is not known.

The pumping required to maintain the drawdown specified for each pumping plan, model-simulated streamflow depletion, and percent reduction in streamflow at 80-percent flow duration for each pumping plan, are listed in table 5. Only the stream sections with a streamflow depletion greater than or equal to 1 percent are listed for each plan. Streamflow depletion is cumulative. However, stream discharge is generally large compared with the simulated streamflow depletion attributable to pumping. Therefore, not including upstream loss in downstream flow probably does not significantly affect the results shown in table 5.

Simulated pumpage resulting from plans A-F ranged from 2.1 to 3.8 Mgal/d (3.3 to 5.9 ft^3/s). The 3.8 Mgal/d is an average of the pumpages resulting from constant-head and constant-flux boundaries in plan C. Comparison of the results of the first six pumping plans simulating pumpage in the three major aquifer systems (table 5) shows that the pumpage simulated in the bedrock and the outwash are within the range of pumpages from the confined sand and gravel aquifer system in the drift. Differences in the areal extent of the cones of depression resulting from pumpage in the three aquifer systems are generally no more than a few miles. In comparing and contrasting the results of the six plans, the reader should be aware that some of the differences may not be solely due to the differences that were outlined in table 4. Variations in aquifer transmissivity, vertical hydraulic conductivity of the confining beds, and the connection between streams and aquifers near the simulated pumpage contribute to the differences in simulated pumpage and drawdown between simulations. Although the pumping plans were designed to minimize the effect of these variations, the variations can not be eliminated. Development of the ground water in any of the three aquifer systems, as simulated in the study, will generally result in withdrawals of as much as 3 Mgal/d and a drawdown of more than 5 ft extending 1.5 to 3 mi from the pumping centers.

The pumpage simulated in plan G was 7.2 Mgal/d (11 ft^3/s). However, the area pumped in plan G was greater than the areas in the other plans. Therefore, the pumpage of plan G, although significantly greater than that for the other plans, is based on pumping a larger area than in the other plans and can not be compared with the pumpage of those plans.

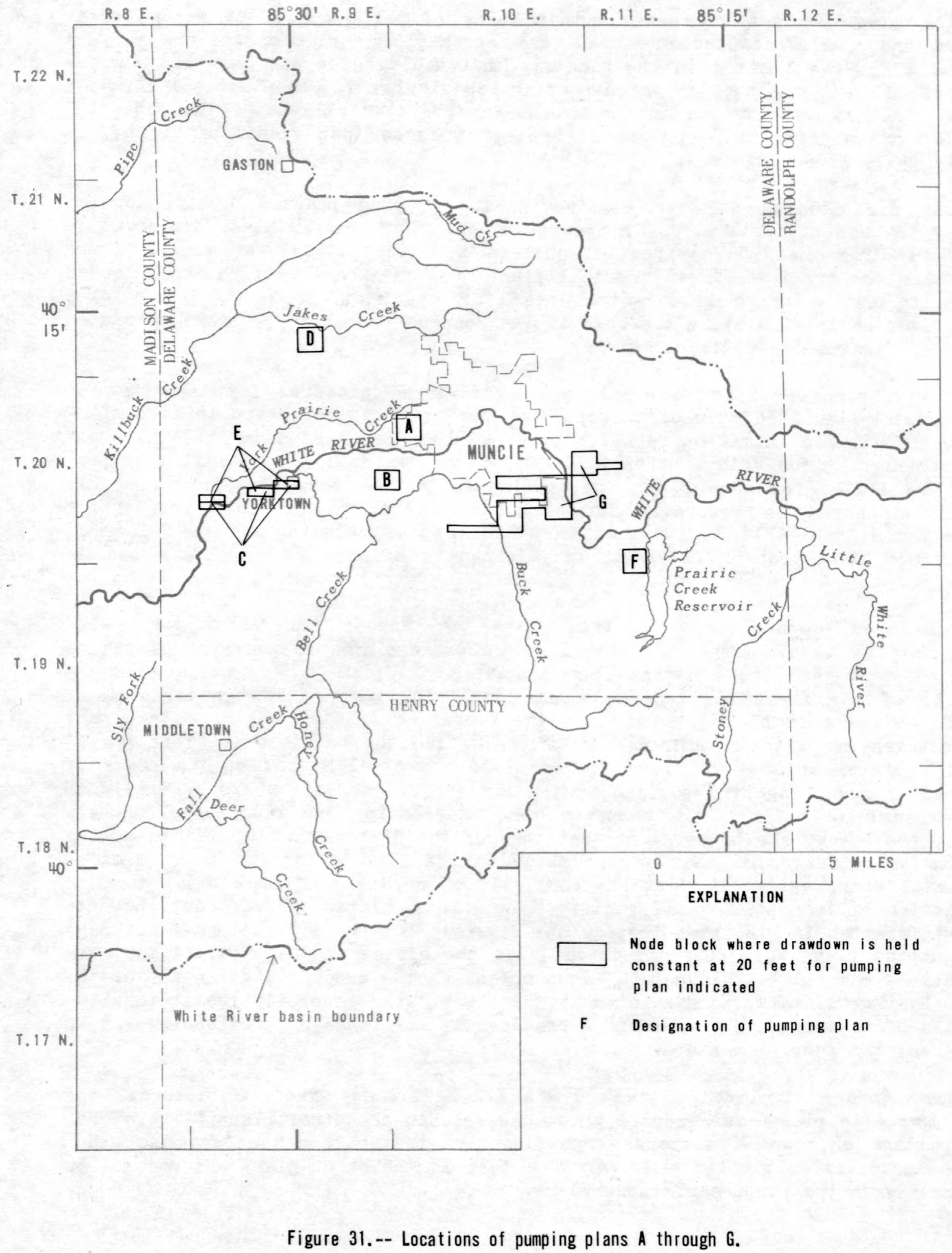


Figure 31.-- Locations of pumping plans A through G.

Table 4.--Characteristics of the seven pumping plans used to assess the potential for ground-water development

Pumping plan	Aquifer in which pumpage is simulated[1]	Geohydrologic characteristics of aquifer system being pumped and location of source of induced recharge
A	Bedrock	Moderately high bedrock transmissivity; pumping near the White River.
B	2	Thick, but not extensive aquifer; pumping near the White River.
C	2	Moderately thick, long, narrow aquifer; pumping from aquifer below White River outwash aquifer.
D	2	Thick, narrow, extensive aquifer; pumping distant from the White River.
E	3	Moderately thick, extensive aquifer associated with outwash at White River; pumping at White River at same location as plan C.
F	5	Thick extensive aquifer around Prairie Creek Reservoir; pumping in aquifer at or near reservoir bottom.
G	4	Moderately thick and extensive aquifer; pumping over a wide area near White River.

[1]Locations of aquifers are given in figure 3.

Results of plans A through F were similar. Yields were slightly greater near the large discharge areas, such as White River and Prairie Creek Reservoir, than elsewhere. The yield for plan D, distant from the White River, was the least of all the plans, and the extent of the cone of depression was one of the least also. Plans C and E were designed to compare the yield of the outwash aquifer with that of a deeper aquifer at the same location. Yields from both plans were the same for constant-flux boundaries. But the yield was lower for plan E (pumping from the outwash) having constant-head boundary. Because the true boundary is between constant flux and constant head, then the true yield is probably slightly lower in plan E than in plan C. Although the yields for plans C and E differed, the shape and the extent of the cone of depression for the constant-flux and constant-head boundary simulations were virtually the same.

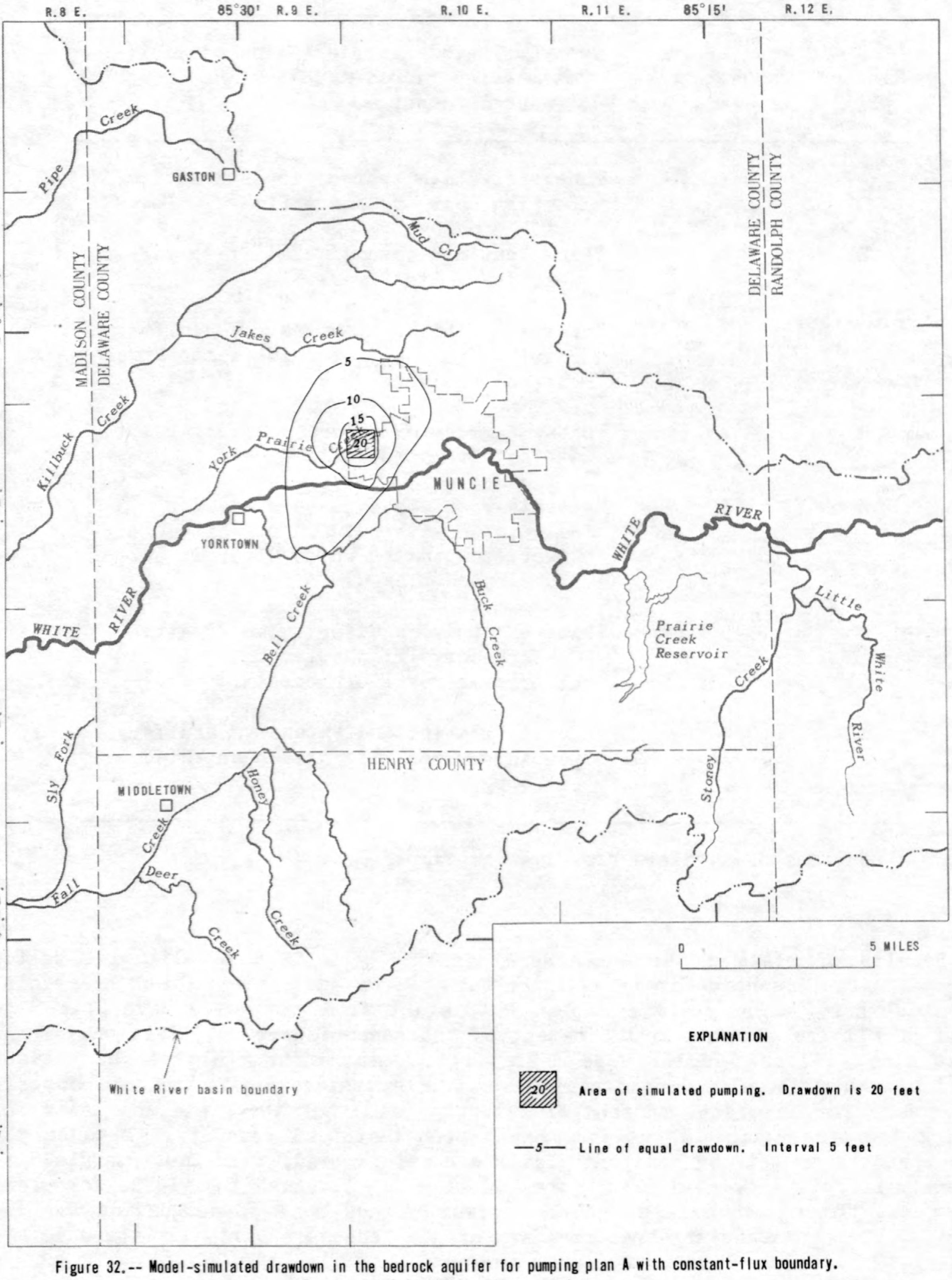


Figure 32.-- Model-simulated drawdown in the bedrock aquifer for pumping plan A with constant-flux boundary. Pumping was 2.2 million gallons per day.

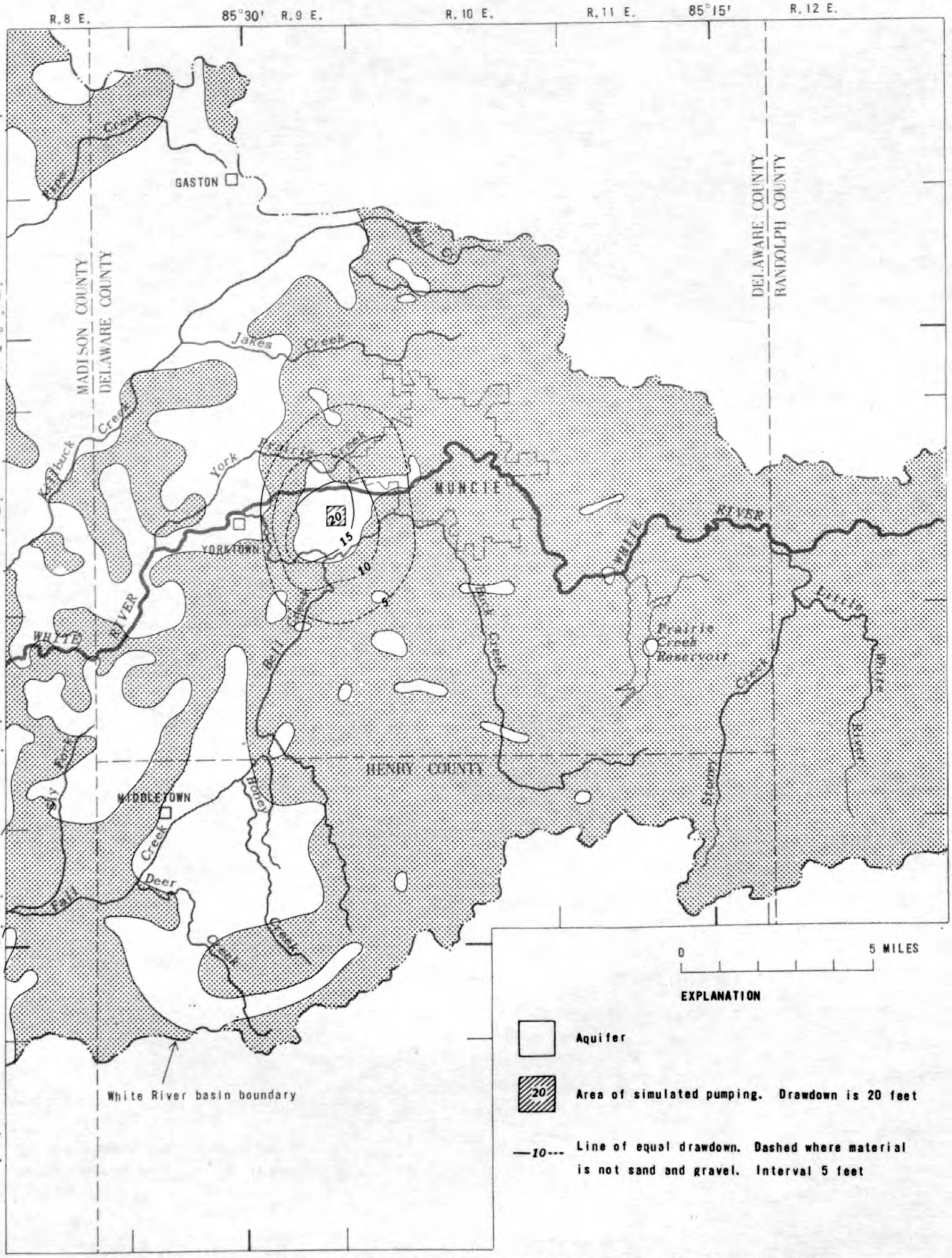


Figure 33.-- Model-simulated drawdown in aquifer 2 for pumping plan B with constant-flux boundary. Pumping was 2.7 million gallons per day.

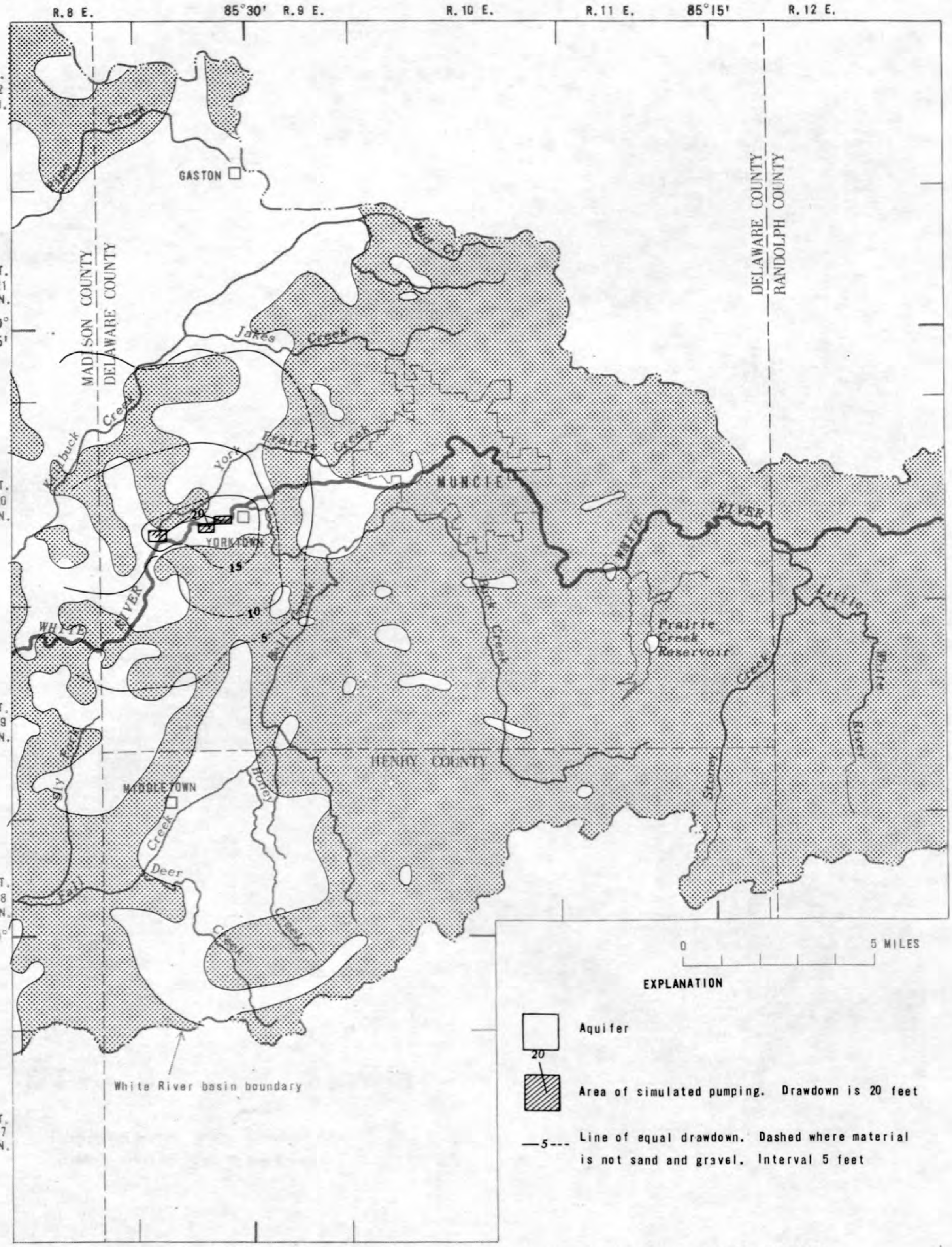


Figure 34.-- Model-simulated drawdown in aquifer 2 for pumping plan C with constant-flux boundary. Pumping was 2.7 million gallons per day.

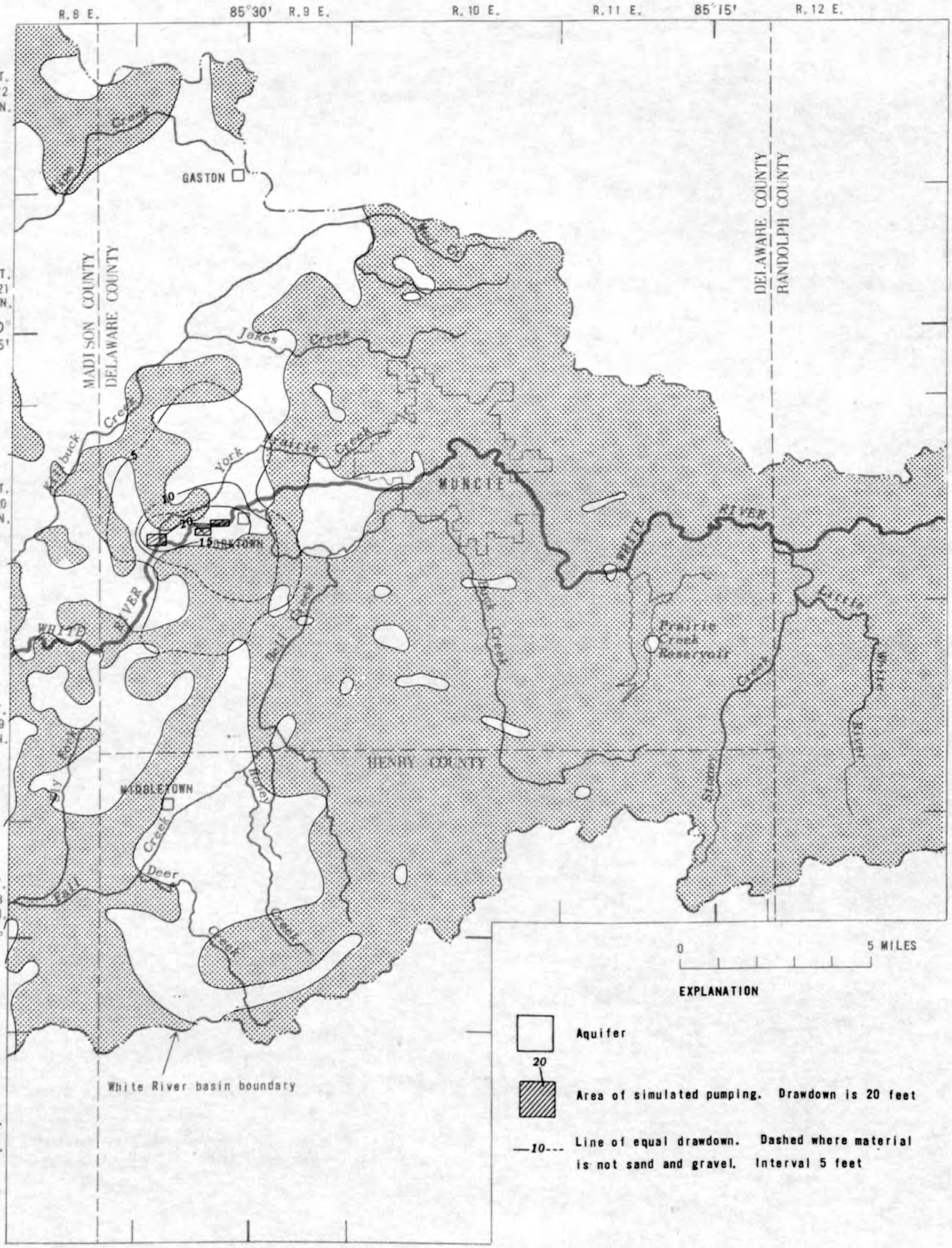


Figure 35.-- Model-simulated drawdown in aquifer 2 for pumping plan C with constant-head boundary. Pumping was 4.9 million gallons per day.

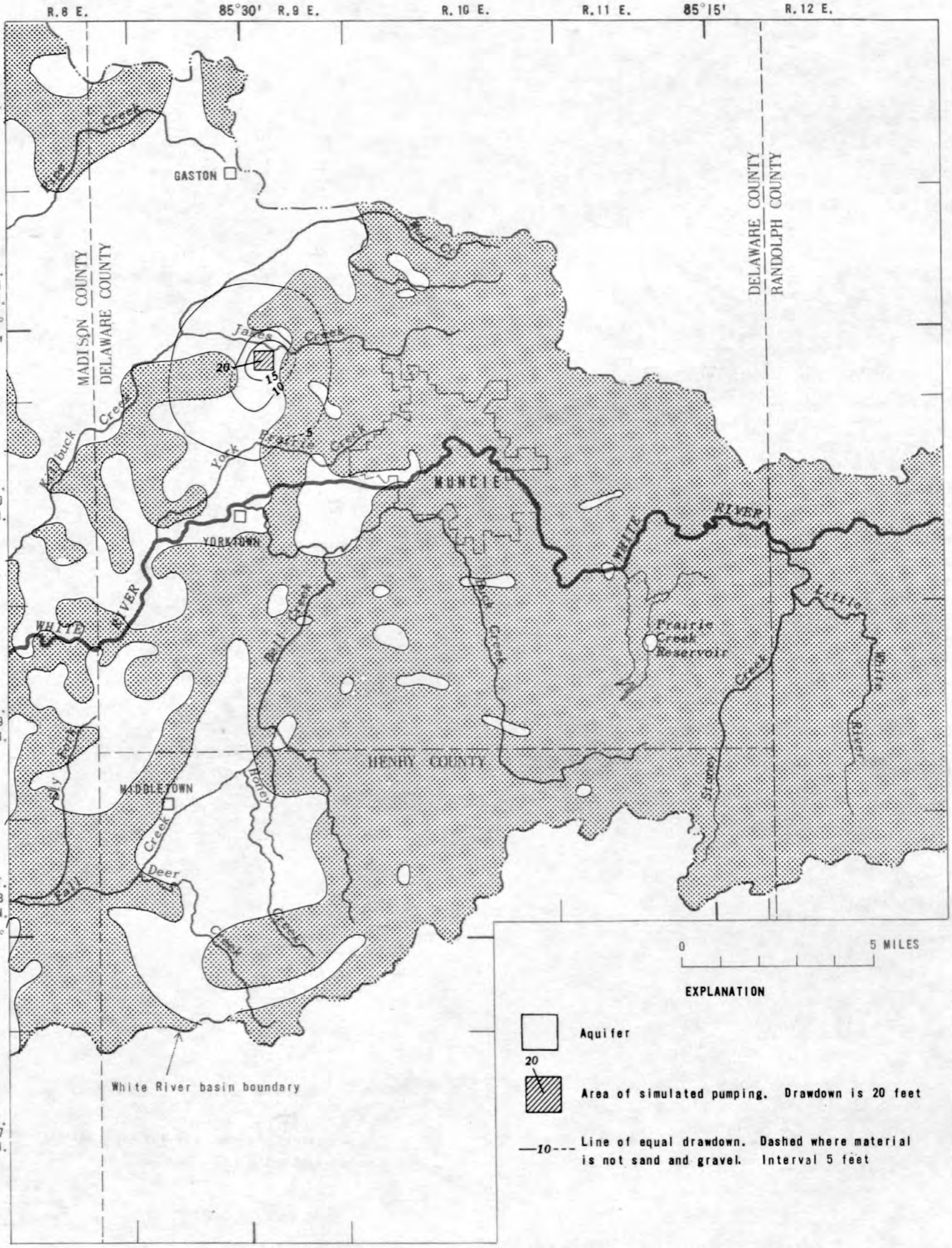


Figure 36.-- Model-simulated drawdown in aquifer 2 for pumping plan D with constant-flux boundary. Pumping was 2.1 million gallons per day.

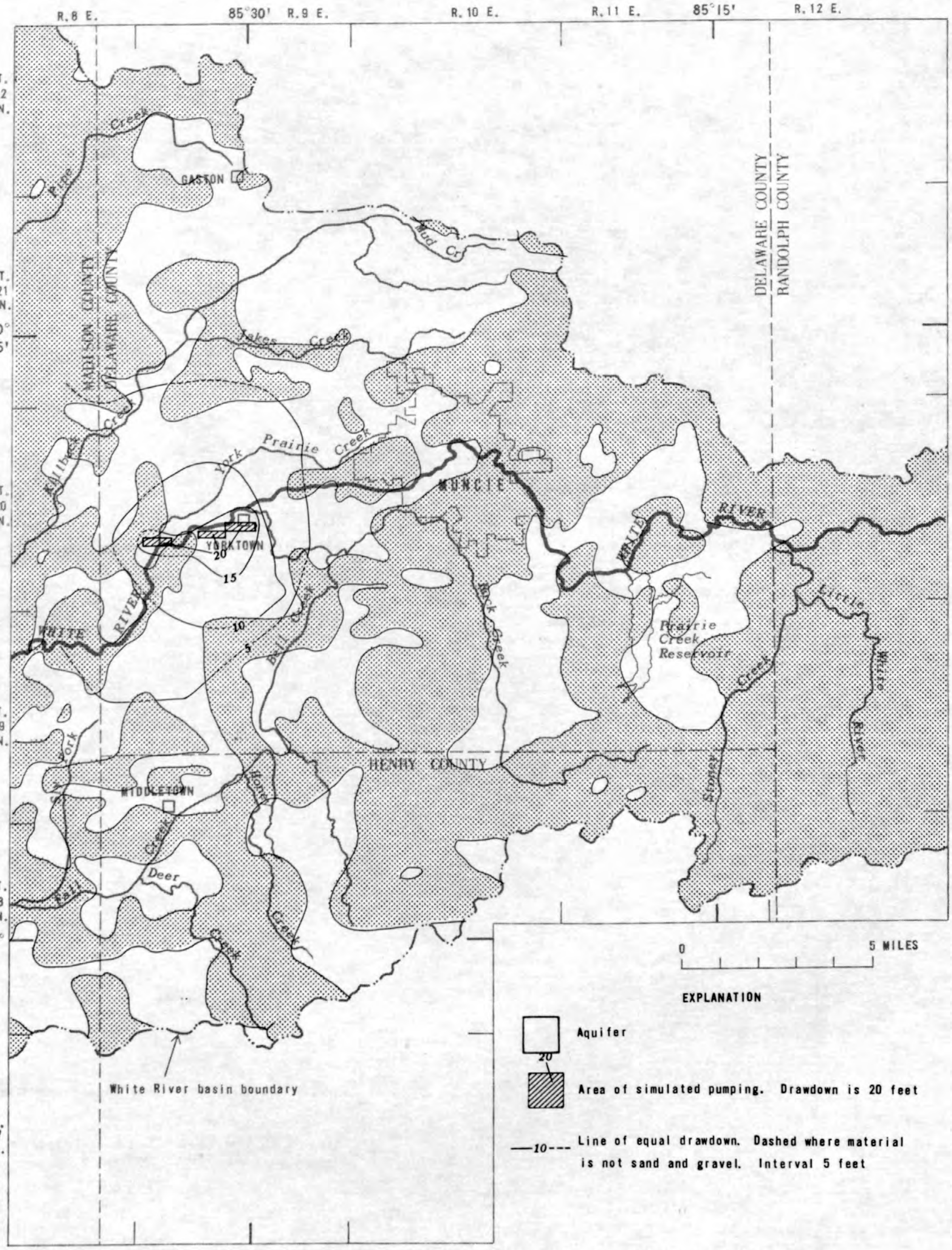


Figure 37.-- Model-simulated drawdown in aquifer 3 for pumping plan E with constant-flux boundary. Pumping was 2.6 million gallons per day.

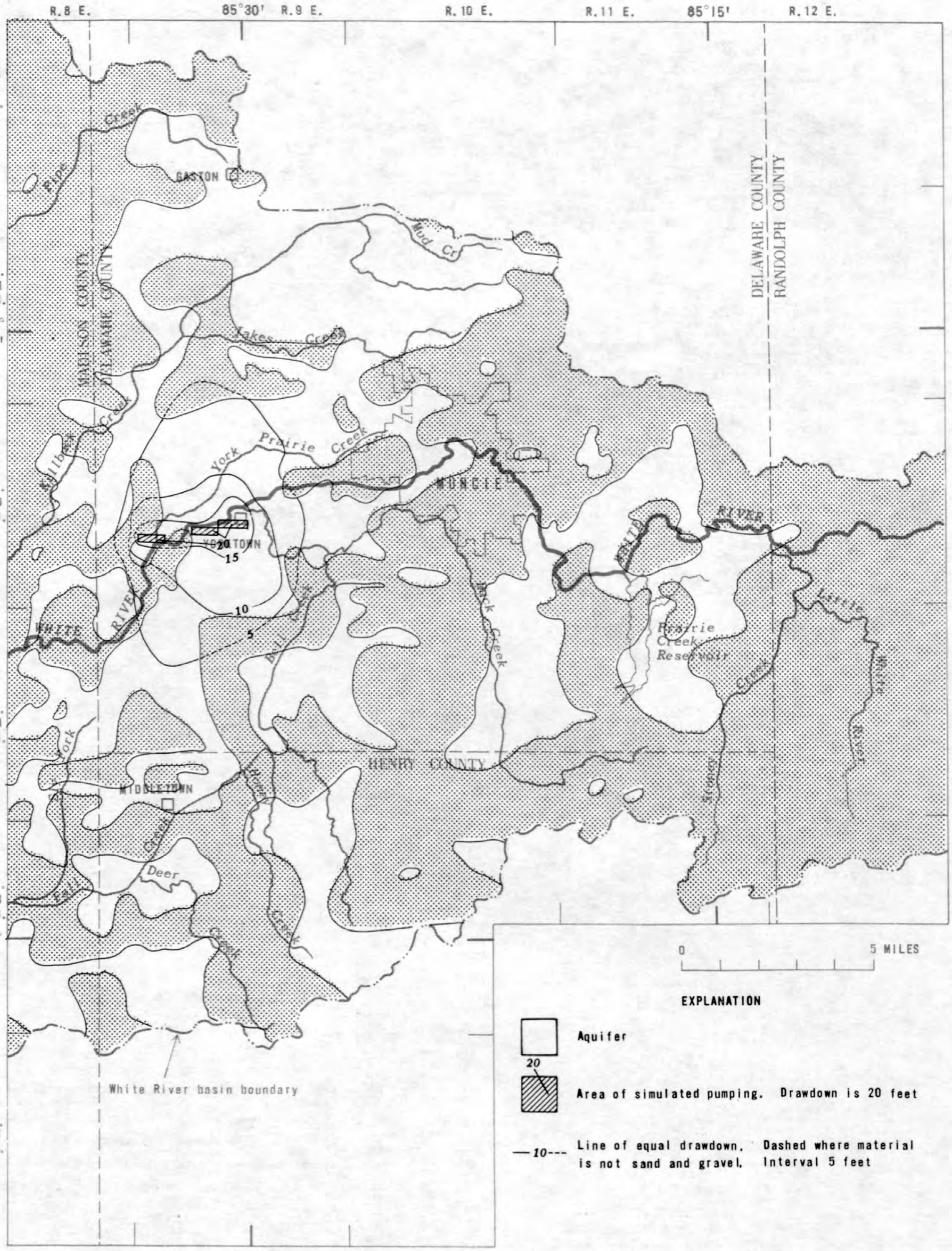


Figure 38.-- Model-simulated drawdown in aquifer 3 for pumping plan E with constant-head boundary. Pumping was 3.1 million gallons per day.

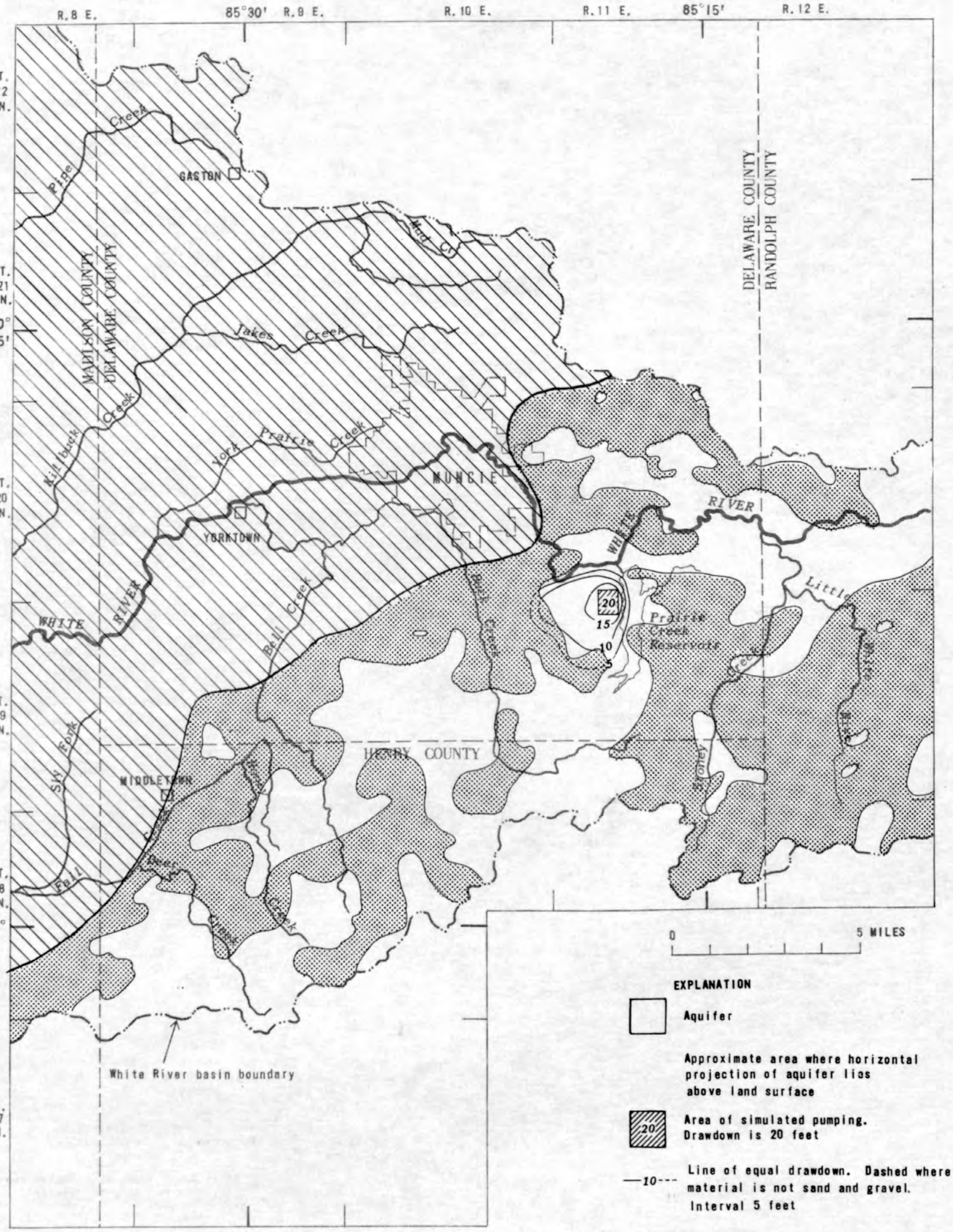


Figure 39.-- Model-simulated drawdown in aquifer 5 for pumping plan F with constant-flux boundary. Pumping was 3.3 million gallons per day.

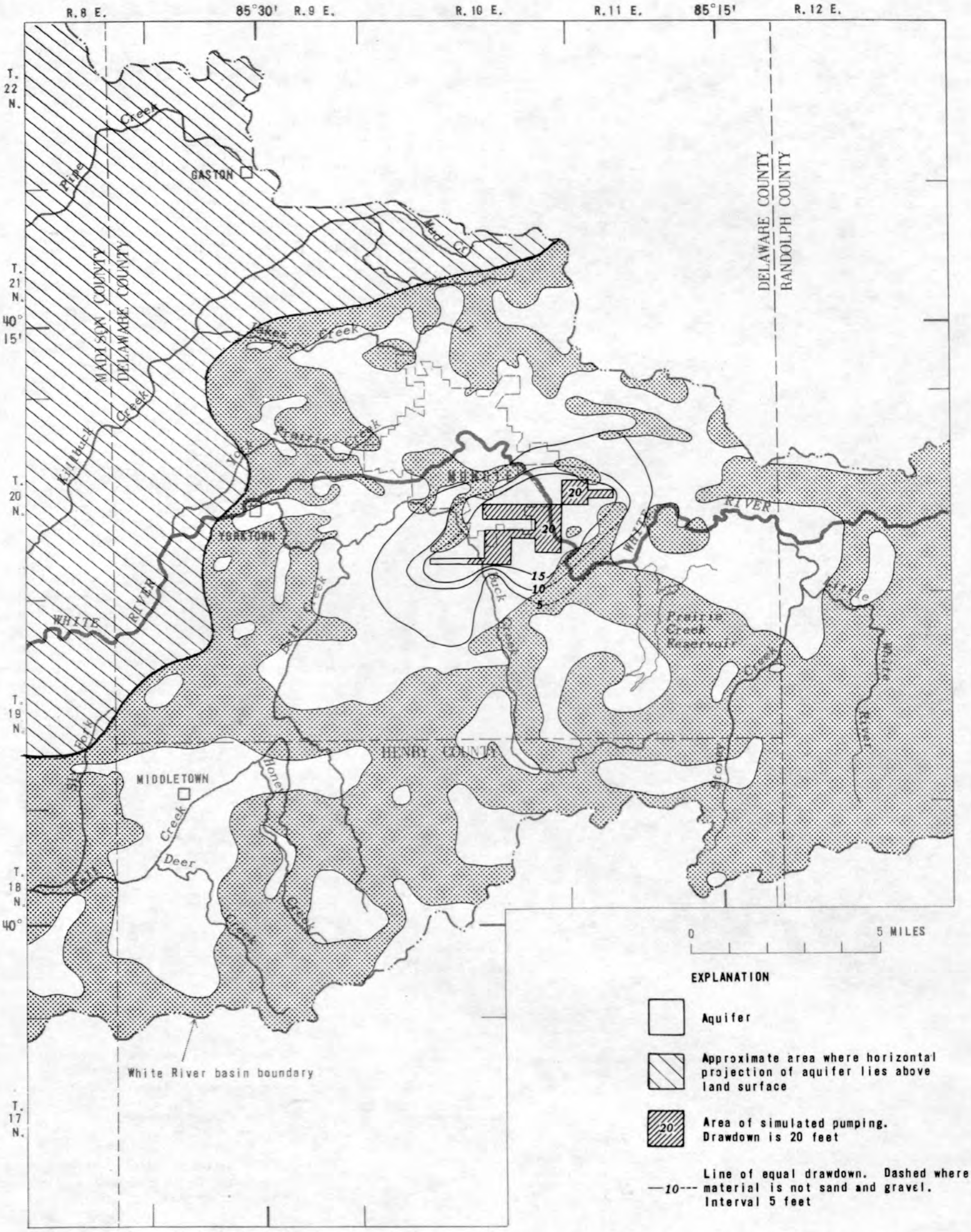


Figure 40.-- Model-simulated drawdown in aquifer 4 for pumping plan G with constant-flux boundary. Pumping was 7.2 million gallons per day.

Table 5.--Results of pumping plans

Pumping plan[1]	Model-simulated pumping (ft³/s) / (Mgal/d)	Stream section most affected by pumping[2]	Stream name	Steady-state ground-water seepage to stream section (ft³/s)	Approximate streamflow at downstream end of section and at 80-percent flow duration (ft³/s)	Model-simulated streamflow depletion in stream section[3] (ft³/s)	Approximate reduction in streamflow in stream section[3] (percent)
		1	Killbuck Creek	4.1	5.0	0.2	4
A		2	do.	.9	1.0	.4	40
(Constant-	3.4	5	White River	4.3	42	1.1	3
flux	2.2	8	Buck Creek	.2	30	.23	1
boundary)		9	do.	2.0	18	.42	2
		11	Bell Creek	3.9	13	.16	1
		1	Killbuck Creek	4.1	5	0.2	4
		2	do.	.9	1	.3	30
B		5	White River	4.3	42	1.1	3
(Constant-	4.1	8	Buck Creek	.2	30	.2	1
flux	2.7	9	do.	2.0	18	.7	3
boundary)		10	do.	16	16	.1	1
		11	Bell Creek	3.9	13	.4	3
		12	do.	1.9	2	.1	5
C		1	Killbuck Creek	4.1	5.0	0.8/.4	16/8
		2	do.	.9	1.0	.5/.1	50/10
(Constant-	4.2	3	White River	-8.5	70	.4/1.1	1/2
flux	2.7	4	do.	.7	73	.3/1.3	<1/2
boundary)		5	do.	4.3	42	.7/1.0	2/2
		8	Buck Creek	.2	30	.2/.4	1/1
(Constant-	7.6	9	do.	2.0	18	.2/.1	1/1
head	4.9	11	Bell Creek	3.9	13	.3/.2	2/2
boundary)		14	Fall and Honey Creeks	5.4	6	.2/.1	3/2
D							
(Constant-	3.3	1	Killbuck Creek	4.1	5.0	0.8	16
flux	2.1	2	do.	.9	1.0	1.0	100
boundary)		5	White River	4.3	42	.5	1
E		1	Killbuck Creek	4.1	5.0	0.8/.4	16/8
(Constant-	4.1	2	do.	.9	1.0	.4/.1	40/10
flux	2.6	3	White River	-8.5	70	.8/.4	1/1
boundary)		5	do.	4.3	42	.7/.6	2/1
		8	Buck Creek	.2	30	.2/.2	1/1
		9	do.	2.0	18	.2/.1	1/1
(Constant-	4.8	11	Bell Creek	3.9	13	.3/.3	2/2
head	3.1	14	Fall and Honey Creeks				
boundary)				5.4	6	.2/.1	3/2
F		6	White River	11	34	0.8	2
(Constant-	5.1	9	Buck Creek	2.0	18	.1	1
flux	3.3	10	do.	16	16	.2	1
boundary)		15	Prairie Creek	.3	.3	3.8	100
		2	Killbuck Creek	0.9	1.0	0.2	20
		5	White River	4.3	42	.9	2
G		6	do.	11	34	1.0	3
(Constant-	11	8	Buck Creek	.2	30	.2	1
flux	7.2	9	do.	2.0	18	6.7	37
boundary)		10	do.	16	16	.4	3
		11	Bell Creek	3.9	13	.3	2
		12	do.	1.9	2	.1	5
		15	Prairie Creek	.3	.3	.3	100

[1]For description of pumping plan, see table 4.
[2]Stream section locations are shown in figure 20.
[3]Single number can be obtained with either constant-flux or constant-head boundaries; where two numbers are given, top number is based on constant-flux boundary and bottom number is based on constant-head boundary.

Plan F produced one of the highest yields (3.3 Mgal/d or 5.1 ft^3/s) and a small cone of depression. Of this yield, 3.8 ft^3/s was from Prairie Creek, whereas only 0.3 ft^3/s discharged into the creek and reservoir. This condition implies that the simulated pumpage depletes the reservoir at a rate of 3.5 ft^3/s.

The purpose of plan G was to determine the maximum yield that could be obtained if most potential pumpage sites around Muncie were developed. For this condition, 7.2 Mgal/d (11 ft^3/s) could be pumped. In plan G, the 15-ft contour is more extensive than the other contours because of the spread of pumpage. Yet the 5-ft contour has about the same areal extent as in other plans. Streamflow reduction was greatest in Buck Creek, which lost 6.7 ft^3/s, or 37 percent of the flow leaving section 9.

Results for the pumping plans are similar to those in other counties in the project area. Simulated pumpage for plans in all counties generally ranged from 0.6 to 3.3 Mgal/d (1 to 5 ft^3/s). The diameters of the 5-ft-drawdown contours are about 5 mi. The simulated pumpages indicate that the flow reduction in streams discharging more than 2 ft^3/s is usually no more than 10 percent of total streamflow.

An investigation of the effect of a more intensive, larger scale development of the ground-water system than that of the seven individual pumping plans might include using the principal of superposition. For instance, the drawdown in the bedrock aquifer, caused by simultaneous pumping of plans A and B, can be estimated by adding the drawdown in the bedrock aquifer, caused by plan A, to the drawdown in the bedrock aquifer, caused by plan B at, any point. Addition of drawdowns in this manner may result in drawdowns near the area of simulated pumpage that are greater than the maximum 20-ft-drawdown criterion used in the individual pumping plans. In the same manner, streamflow depletions for each stream section can also be estimated, except where depletion exceeds streamflow. In the superposition of drawdowns, the drawdown at the boundary may become greater than 1 ft. For this condition, the influence of the boundary may be much more significant than when pumping plans are simulated individually.

SUMMARY AND CONCLUSIONS

The ground-water resources of the White River basin in and near Delaware County, Ind., were investigated by mapping the aquifers, calculating their hydraulic properties, determining the distribution of potentiometric head, and determining some of the components of the ground-water budget. This information was used in constructing and calibrating a seven-layer, digital ground-water-flow model. The flow model, constructed and calibrated to water-level and seepage data, simulated conditions for the autumn of 1977. The model was used to assess the ground-water potential in terms of yield, drawdown, and streamflow depletion.

Drift, generally ranging in thickness from 0 to 500 ft, covers nearly the entire study area. Beneath the drift lie limestone, dolomite, and shale of Silurian and Ordovician age. The Anderson Valley, a tributary of the Teays Valley system, trends east-west through the center of the area. Relief of the bedrock surface is about 400 ft.

The three major aquifer systems are (1) six confined sand and gravel aquifers within the glacial drift, (2) the bedrock aquifer, and (3) an unconfined outwash aquifer associated with the major streams. The sand and gravel aquifers within the drift generally range in thickness from 5 to 40 ft, and the aquifers are nearly horizontal and areally discontinuous. Locally, they may coalesce vertically to form one thick deposit. On the basis of specific-capacity data from adjacent Madison County, the average hydraulic conductivity of the sand and gravel aquifers was calculated to be 433 ft/d. The bedrock aquifer underlying all of Delaware County has an average permeable thickness estimated to be 150 ft and an average transmissivity of about 1,000 ft^2/d. The outwash aquifer is hydraulically similar to the six confined sand and gravel aquifers. Generally, the outwash aquifer is neither areally extensive nor thick but is important because of its proximity to potential sources of induced recharge and connection with areally extensive aquifers.

Water-level fluctuations in observation wells indicate that the ground-water system is generally in dynamic equilibrium. Ground-water pumpage for 1976 was 3.1 Mgal/d (4.8 ft^3/s). Ground-water seepage to streams at 80-percent flow duration on October 29, 1977, was 81 ft^3/s.

The water budget simulated in the model indicates that the rate of flow into the ground-water system is 102 ft^3/s. Of this, 80 percent is due to areal recharge of precipitation, and 20 percent is due to ground-water flow across the boundaries. Two percent of the ground-water outflow is pumpage, 66 percent is seepage to streams, and the remaining 32 percent is ground-water flow across the boundaries.

Model analysis was used to assess the potential for ground-water development of the three major aquifer systems. Results indicate that (1) as much as 3 Mgal/d can be developed at some locations, (2) the 5-ft-drawdown contour was about 5 mi in diameter, and (3) the streamflow reduction caused by pumpage was 10 percent or less for streams discharging more than 2 ft^3/s. A 7-Mgal/d pumpage was obtained by a simulated pumping plan that included most potential pumping sites around Muncie. Although detailed well hydraulics and pumping constraints were not considered in the model simulations, their consideration in the evaluation of the results of these and any future simulations would be advantageous.

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. I-336-338.

Burger, A. M., Forsyth, J. L., Nicoll, R. S., Wayne, W. J., 1971, Geologic map of the 1° x 2° Muncie quadrangle, Indiana and Ohio, showing bedrock and unconsolidated deposits: Indiana Geological Survey, Regional Geologic Map no. 5, parts A and B.

Burger, A. M., Keller, S. J., and Wayne, W. J., 1966, Bedrock topography of Northern Indiana: Indiana Department of Natural Resources, Geological Survey, Miscellaneous Map no. 12.

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.

Freeze, R. A., and Cherry, J. A., 1979, Ground water: Englewood Cliffs, N.J., Prentice-Hall, Inc., 604 p.

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

Hoggatt, R. E., Hunn, J. D., and Steen, W. J., 1968, Water resources of Delaware County: Indiana Department of Natural Resources, Division of Water, Bulletin 37, 57 p.

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

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, 1941-1970, Climatological data annual summaries, Indiana: Asheville N.C., National Oceanic and Atmospheric Administration, Environmental Data Service, v. 46-75, no. 13.

______ 1977, Climatological data annual summary, Indiana: Asheville, N.C., National Oceanic and Atmospheric Administration, Environmental Data Service, v. 82, no. 13, 13 p.

Shurig, D. G., 1974, Engineering soils map of Delaware County: West Lafayette, Indiana, Purdue University, 57 p.

Todd, D. K., 1959, Ground-water hydrology: New York, John Wiley and Sons, Inc., 336 p.

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.

U.S. Department of Commerce, 1980, 1980 Census population and housing (Indiana): U.S. Department of Commerce, Bureau of Census, Advanced Report PHC80-V-16, 32 p.

Wayne, W. J., 1956, Thickness of drift and bedrock physiography of Indiana north of the Wisconsin glacial boundary: Indiana Geological Survey Report of Progress no. 7, 70 p.

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