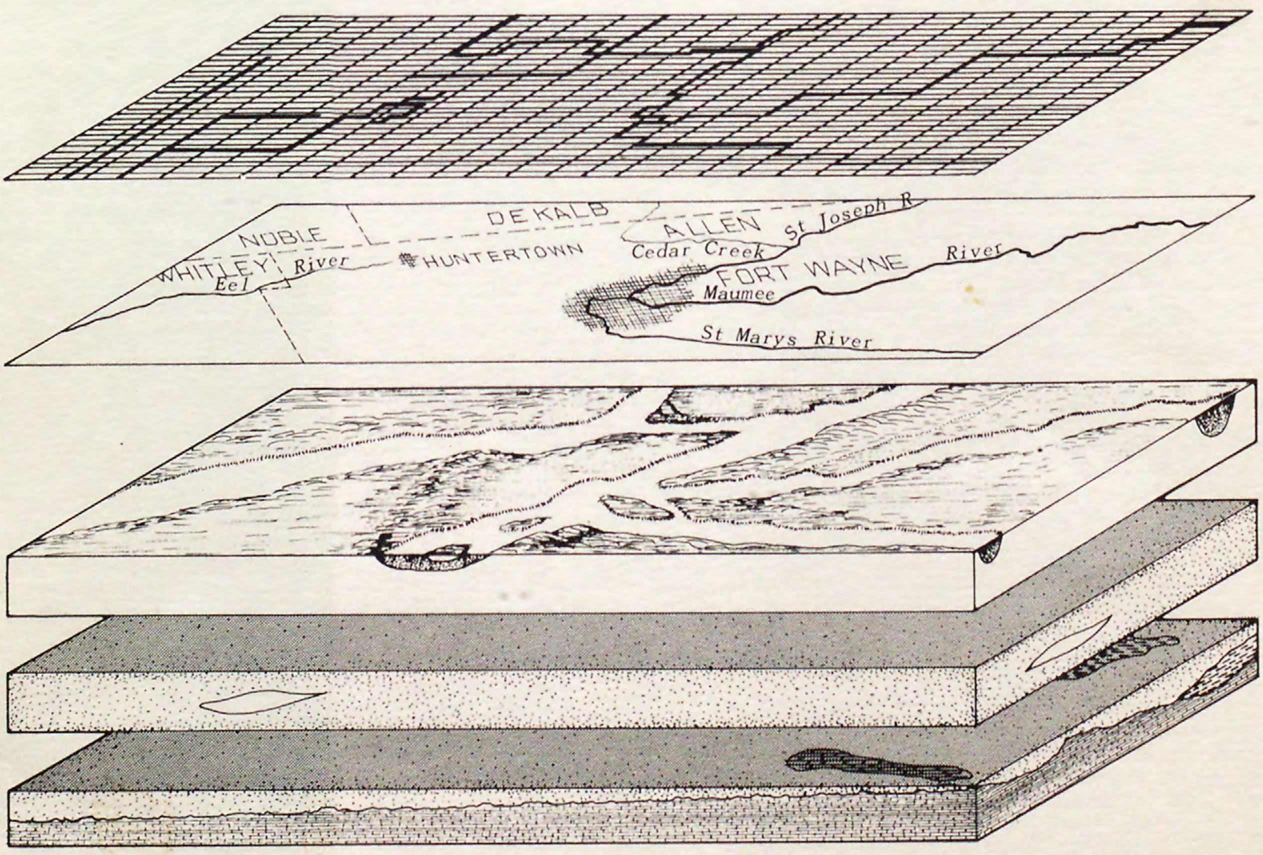
GROUND-WATER AVAILABILITY NEAR FORT WAYNE, ALLEN COUNTY, INDIANA

U. S. GEOLOGICAL SURVEY

WATER-RESOURCES INVESTIGATIONS 80-34



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H. William Menard, Director

For additional information write to:

U.S. Geological Survey
1819 N. Meridian Street
Indianapolis, Indiana 46202

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METRIC CONVERSION FACTORS

The inch-pound and other units used in this report can be converted to units used in the metric system as follows:

Multiply inch-pound units	By	To obtain metric units
Flow		
cubic foot per second (ft^3/s)	28.32	liter per second (L/s)
cubic foot per second (ft^3/s)	0.0283	cubic meter per second (m^3/s)
million gallons per day (Mgal/d)	0.0438	cubic meter per second (m^3/s)
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day (m/d)
Length		
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)
Recharge		
inch per year (in./yr)	2.54	centimeter per year (cm/yr)
Transmissivity		
square foot per day (ft^2/d)	0.0929	square meter per day (m^2/d)
Volume		
million gallon (Mgal)	3,785	cubic meter (m^3)

GROUND-WATER AVAILABILITY NEAR FORT WAYNE, ALLEN COUNTY, INDIANA

By Michael Planert

ABSTRACT

A 3-year study to determine the ground-water potential of confined glacial aquifers in a large part of Allen County was begun in July 1974 by the U.S. Geological Survey. Mapping of outwash sands and gravels delineated two distinct layers of sand and gravel in the southern and eastern parts of the county. The two layers merge into one aquifer more than 100 feet thick north and west in the county. The upper layer of sand and gravel is buried beneath 20 to 140 feet of glacial till.

A digital model that simulates ground-water flow in three dimensions was used to integrate the geologic and hydrologic conditions into a suitable analog simulating the ground-water system surrounding Fort Wayne. The model consisted of three layers representing (1) the upper till, (2) the upper sand and gravel, and (3) the lower aquifer. The lower aquifer represents a combination of the lower sand and gravel and a section of limestone aquifer immediately underlying the glacial sediments.

The model, calibrated against water-level and streamflow data collected during September 1976, was used to simulate pumping and to evaluate the effects of pumping on ground-water levels and streamflows.

Two pumping plans were tested. One plan was designed to concentrate pumping in areas of highest transmissivity and to avoid diverting streamflow already used for Fort Wayne's water supply. The results for two different boundary conditions (constant heads and constant flows) indicate that between 35 and 40 million gallons per day could be produced from the 10 simulated pumping sites and that the streamflow losses would be between 17 and 24 million gallons per day. The high percentage of streamflow loss associated with the first plan led to the second plan that tested the amount of stream diversion caused by the pumping at well sites along the streams. Results from the second plan indicated that approximately 30 million gallons per day could be produced from 6 simulated pumping sites and that the streamflow losses would be between 21 and 25 million gallons per day.

INTRODUCTION

From the 1880's, when the Fort Wayne water utility was established, until the 1930's, water from wells completed in limestone was the sole source of water supply in Fort Wayne. During this period, the distribution system grew to 22 pumping stations throughout the city. Because of excessive drawdown of water levels in the limestone by the early 1930's, the utility abandoned the wells and began to use water from the St. Joseph River (Foland, 1972; Paul Fulkerson, oral commun., 1974).

In 1934, a 285-Mgal (million gallon) impounding pool and a filtration plant were built on the St. Joseph River (fig. 1). As demand for water increased, two reservoirs, Cedarville and Hurshtown, were built upstream from the impounding pool. Releases from the reservoirs maintain the impounding pool at a constant level for the filtration plant intake. The Cedarville Reservoir, completed in 1954 with a storage capacity of 500 Mgal, is about 12 mi (miles) northeast of Fort Wayne (fig. 1). The Hurshtown Reservoir (fig. 1), termed a "pumped-storage reservoir," was built in 1969 approximately one-third of a mile east of Cedarville Reservoir. Water is pumped into the Hurshtown Reservoir during high streamflows and is released, when needed, into a creek that flows into Cedarville Reservoir. The storage capacity of Hurshtown Reservoir is 1,885 Mgal.

Purpose and Scope of the Study

In July 1974, the U.S. Geological Survey, in response to the requirements of the Indiana Department of Natural Resources and the city of Fort Wayne, began a 3-yr (year) study to evaluate the ground-water resources in the glacial drift in a large part of Allen County. The purpose of the study was to (1) determine the feasibility of developing ground water, and (2) define the hydrologic impacts of the development. A multilayered digital model was used to (1) simulate ground-water flow, (2) test the feasibility of proposed pumping programs, and (3) evaluate the effects of development.

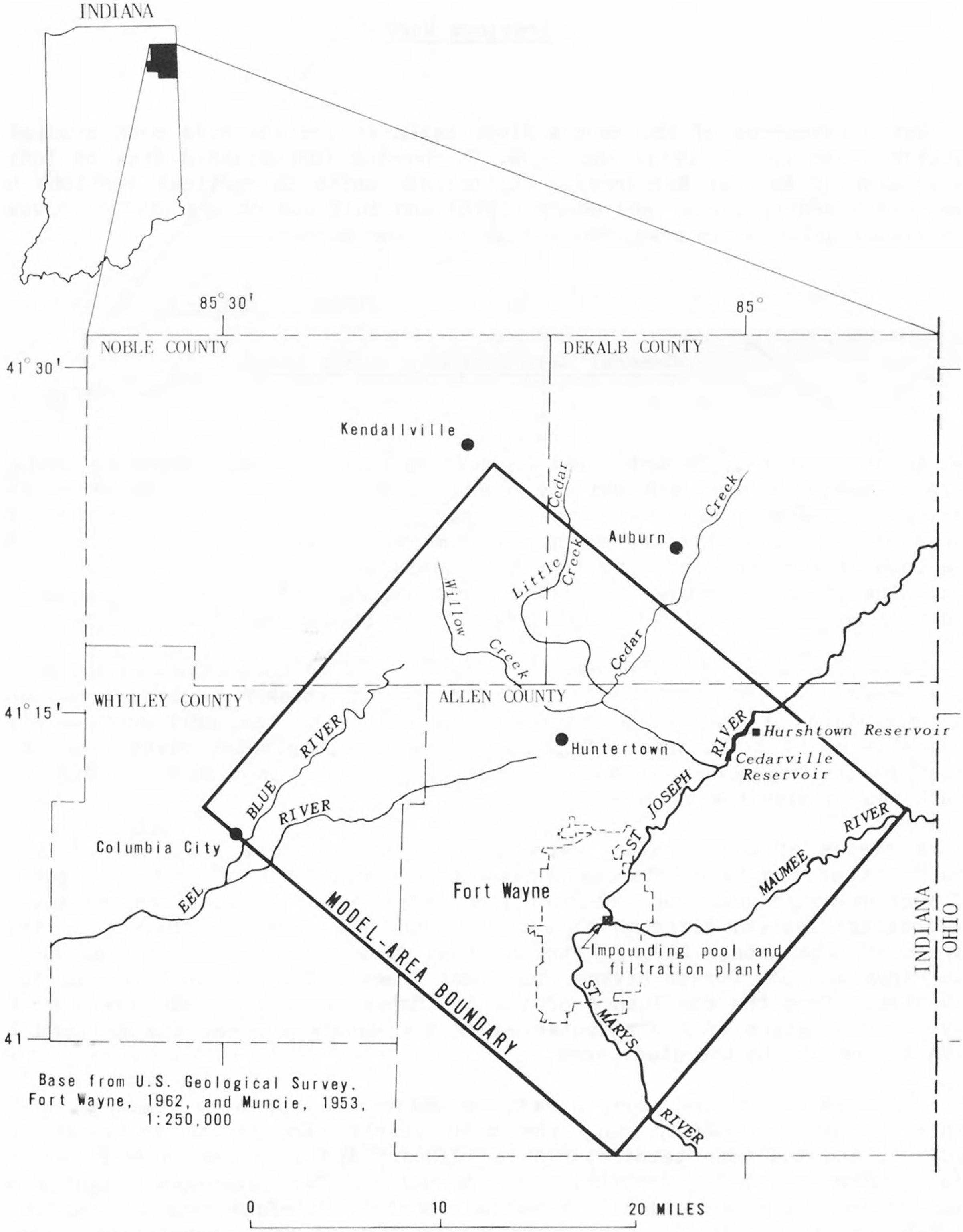


Figure 1.-- Location of study area.

Previous Work

Water resources of the Maumee River basin in Indiana have been studied by Pettijohn and Davis (1973) and by W. C. Herring (unpublished data of Indiana Department of Natural Resources). Lithologic units in vertical sections have been discussed by Bleuer and Moore (1978) and Ault and others (1973); however, individual units on an areal scale have not been mapped.

General Description of Study Area

An approximately 700-mi^2 area in northeastern Indiana, including parts of Allen, Dekalb, Noble, and Whitley Counties, was selected as the study area. The area topography has been shaped primarily by glaciers. The prominent features of the landscape are several end moraines deposited by two ice lobes, one lobe originating from the Lake Erie basin and one from the Lake Huron basin. The topographic high is in the northwest part of the study area along an end moraine, a product of the amassing of material between the ice lobes.

Thickness of glacial sediments ranges from 10 to more than 250 ft; bedrock is exposed only in rock quarries. The sediments thicken to the north, where the elevation of the bedrock surface decreases and the land surface rises. Channels cut by tributaries (fig. 2) to a major preglacial river, the Teays, exist in the bedrock. The main channel of the Teays (not shown in fig. 2) is south and outside the study area.

A series of end moraines deposited by the ice lobe that moved west and south out of the Lake Erie basin have helped shape the present configuration of surface-water drainage. Most of the study area lies within the Maumee River drainage system, although a section in the west part is drained by tributaries of the Wabash River. Streams that flow into the study area are the St. Marys and St. Joseph Rivers and Cedar Creek. The Maumee River is formed downstream from the confluence of the St. Marys and St. Joseph Rivers at Fort Wayne. Headwaters of two tributaries of the Wabash system, the Eel and Blue Rivers, are within the study area.

The climate of the study area is characterized by warm summers and cold winters. In Fort Wayne, Ind., the mean yearly temperature is about 50° F (10° C), and the mean temperatures for January and July are 26.3° F (-3.2° C) and 73.5° F (23.0° C), respectively. Annual rainfall averages slightly more than 34 in. (inches), and the individual monthly rainfall ranges from 2.0 in. in February to 4.1 in. in June. The yearly Class A pan evaporation for the area averages about 42 in. (See Schaal, 1959).

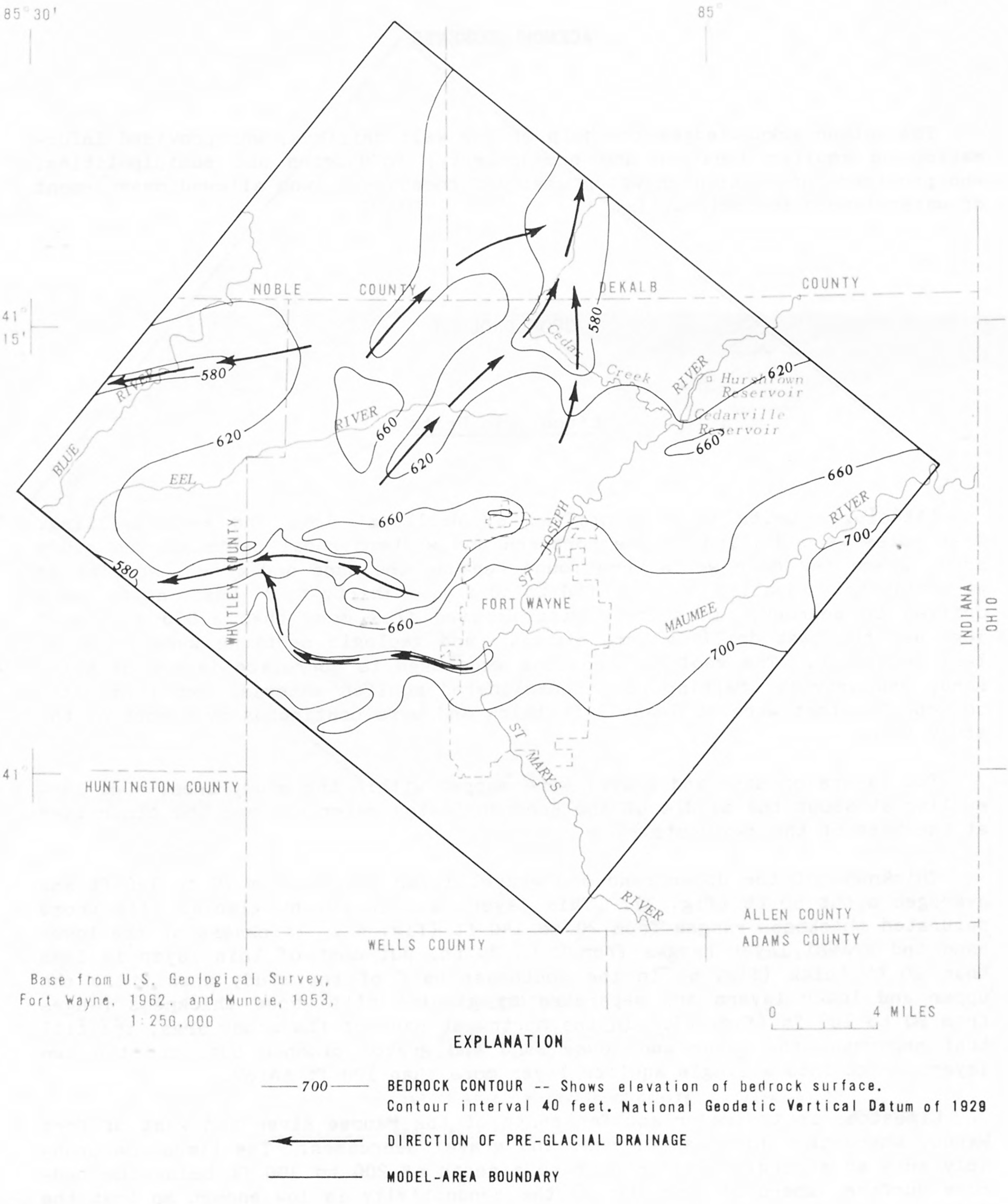


Figure 2.--Elevation of bedrock surface.

ACKNOWLEDGMENTS

The author acknowledges the help of the well drillers, who provided information on aquifer location and productivity; industries and municipalities, who provided information on water use; and homeowners, who allowed measurement of water levels in their wells.

GEOHYDROLOGY

Lithologic Units

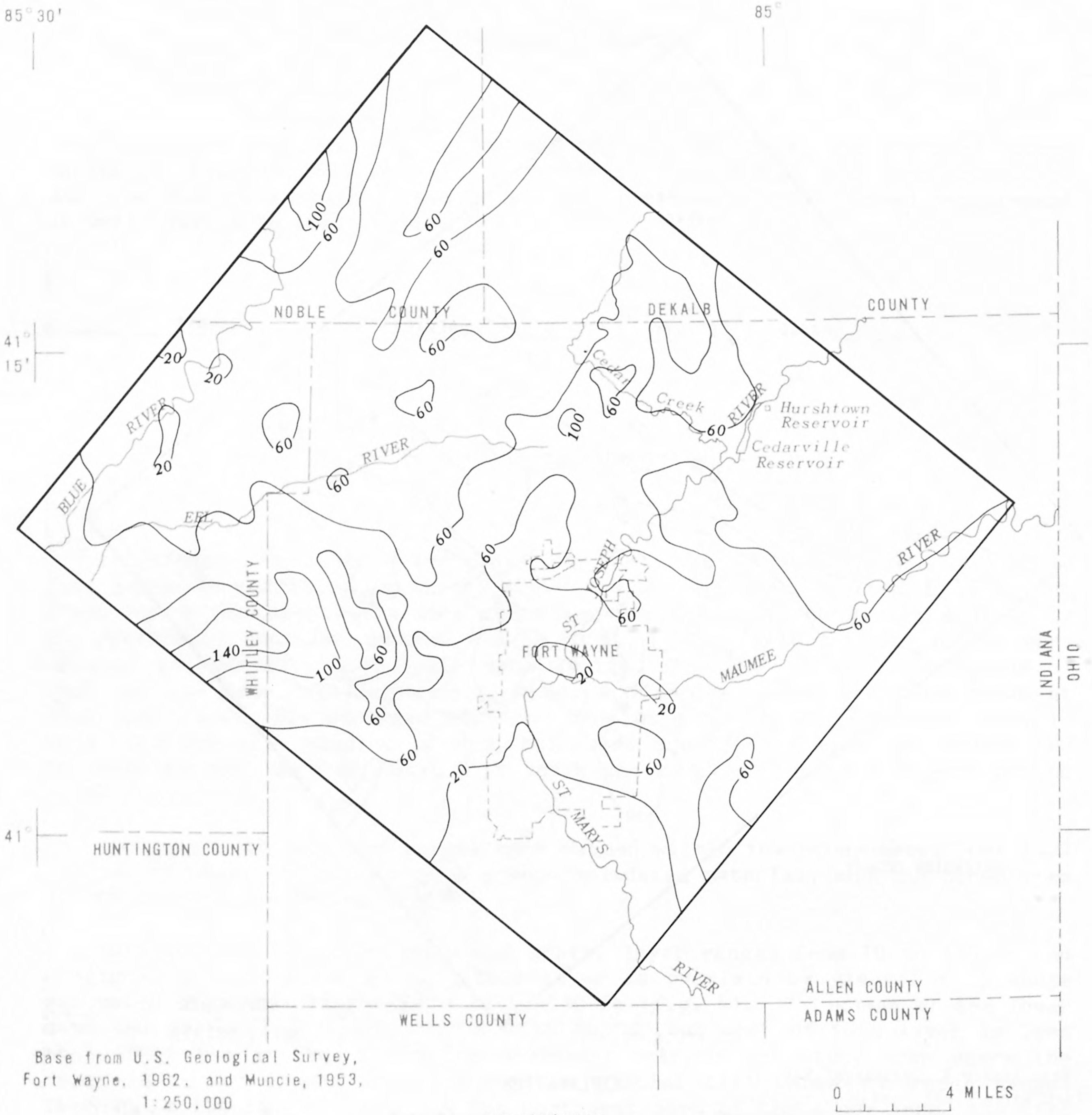
Lithologic units were identified by drillers' logs and test drilling. Test holes were drilled in the northern and western rural parts of the study area, where few deep wells were found because the sand and gravel aquifer is at relatively shallow depths (60 to 80 ft). Thirty-four test holes were drilled to bedrock. Lithologic data derived from more than 2,000 drillers' logs and the test drilling were plotted, and geologic sections were drawn at 1-mi intervals. The geologic sections were used to correlate layers of till, sand, and gravel. Mapping of unconsolidated aquifer material was restricted to deposits that were at least 4 ft thick and were continuous over most of the study area.

Two layers of sand and gravel were mapped within the study area. One layer lies at about the middle of the unconsolidated material, and the other lies at the base of the sediments.

Thickness of the upper sand and gravel layer ranges from 10 to 120 ft and averages about 50 ft (fig. 3). This layer is overlain by glacial till whose saturated thickness ranges from 20 to 140 ft (fig. 4). Thickness of the lower sand and gravel layer ranges from 0 to 80 ft, but most of this layer is less than 20 ft thick (fig. 5) in the southeast half of the study area where the upper and lower layers are separated by glacial till whose thickness ranges from 20 to 100 ft (fig. 6). In the northwest part of the study area, the till that separates the upper and lower sand and gravel pinches out, and the two layers merge into a single aquifer layer more than 100 ft thick.

Limestone is the major aquifer south of the Maumee River and west of Fort Wayne, where the thickness of sand and gravel decreases. The limestone probably acts as a single aquifer unit to a depth of 200 to 300 ft below the bedrock surface, where at such depths the conductivity is low enough so that the remaining limestone may be excluded from the flow system.

Figure 3.--Thickness of the upper sand and gravel aquifer.

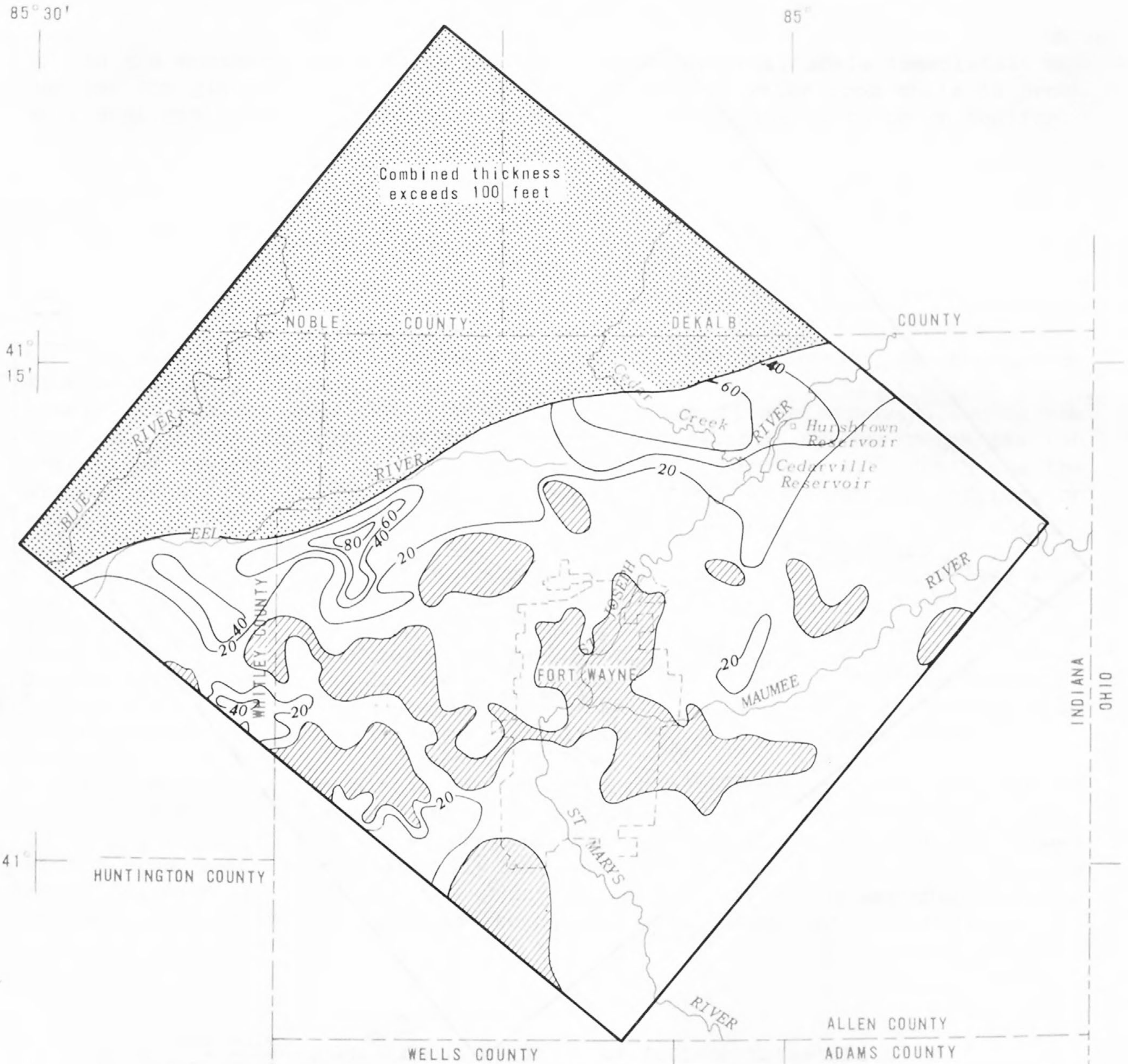


EXPLANATION

—— 60 —— LINE OF EQUAL THICKNESS OF SATURATED MATERIAL -- Intervals 20 and 40 feet

———— MODEL-AREA BOUNDARY

Figure 4.-- Saturated thickness of the upper till unit.



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

0 4 MILES

EXPLANATION

AREA WHERE LOWER AND UPPER SAND AND GRAVEL LAYERS ARE COMBINED

AQUIFER ABSENT

—40— LINE OF EQUAL THICKNESS OF AQUIFER -- Interval 20 feet

——— MODEL-AREA BOUNDARY

Figure 5.--Thickness of the lower sand and gravel aquifer.

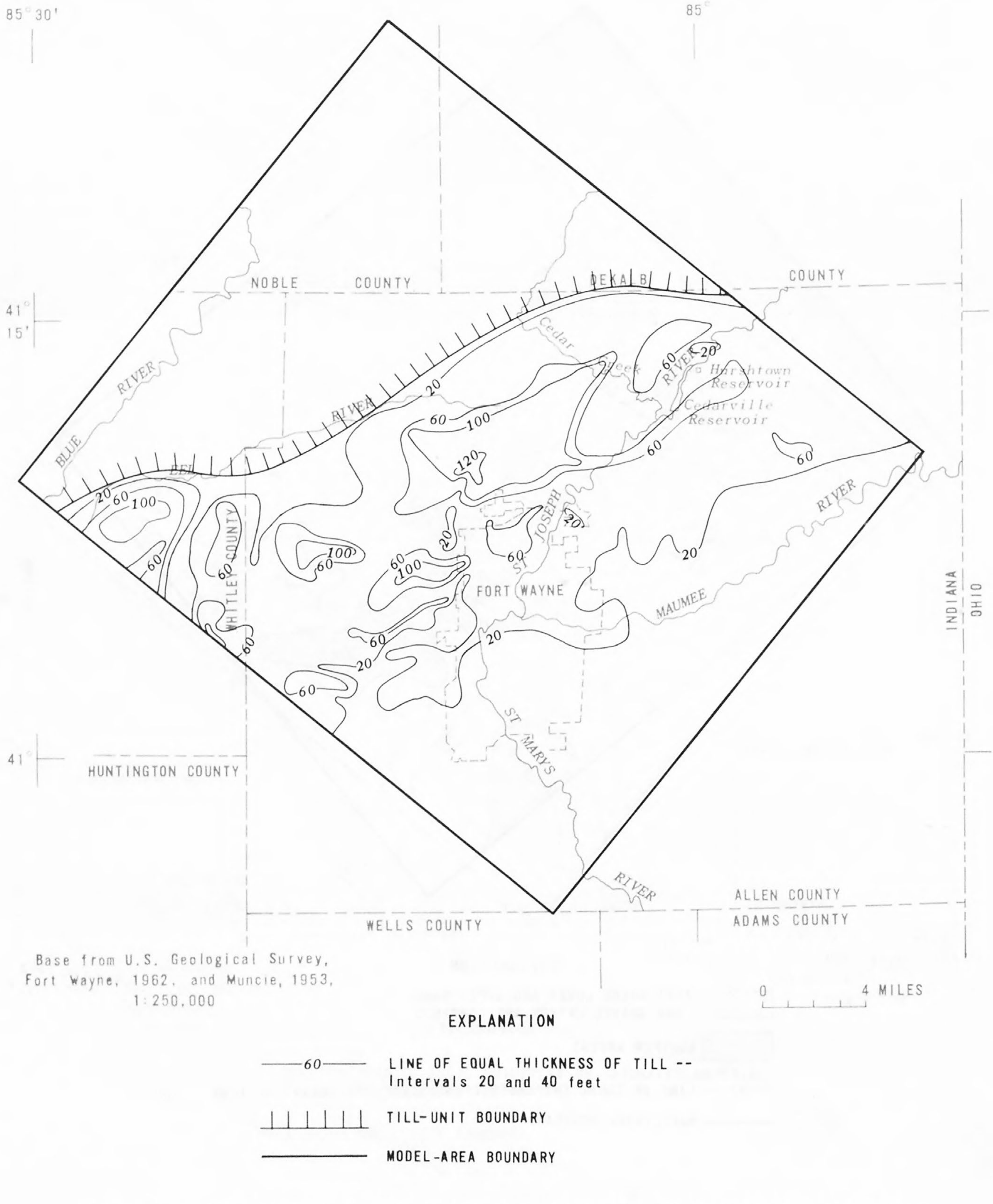


Figure 6. -- Thickness of till that separates the upper and lower sand and gravel aquifers.

In the northern and eastern parts of the study area, shale immediately underlies the glacial drift (fig. 7). Contribution of water from shale is probably negligible, and, therefore, the shale is not considered to be an aquifer.

Ground-Water Flow

The driving force for movement of ground water is gravity, and the direction of movement can be determined by the differences in potentiometric heads within and between aquifers. In a layered system flow is downward, where the head decreases with depth and water in the water-table aquifer recharges the underlying layers, or flow is upward, where head increases with depth and the water discharges from the aquifers to some point such as a stream, spring, or well.

The direction of ground-water movement was mapped by the potentiometric surface in each of the three aquifers--upper and lower sand and gravel aquifers and limestone. A total of 334 water-level measurements was used. The flow pattern in the upper sand and gravel aquifer is shown in figure 8. Water in the aquifer is moving laterally from the higher elevations in Noble and Dekalb Counties into the study area. Part of the ground water discharges into minor streams, and the remaining water discharges into the Maumee River.

A generalized flow pattern in a vertical section (line A-A', fig. 8) is shown in figure 9. In high areas, head decreasing with depth indicates water is moving downward. In the mid-region, between the high areas and the Maumee River, some water moves laterally and some water in the water-table aquifer moves downward and recharges the other aquifers. Near the Maumee River, heads increase with depth, and water in the aquifers discharges into the river.

Hydraulic Characteristics of Aquifer Materials

Hydraulic conductivity of the unconsolidated aquifer materials was determined from specific-capacity data. Specific capacities of domestic wells reported on drillers' logs were used to estimate transmissivities for the screened part of the aquifer by a technique described in Brown (1963, p. 336-338). Hydraulic conductivities were derived by dividing the transmissivity values by the screen length. A discussion of the rationale and the assumptions used with this method are presented by Meyer and others (1975, p. 17-21).

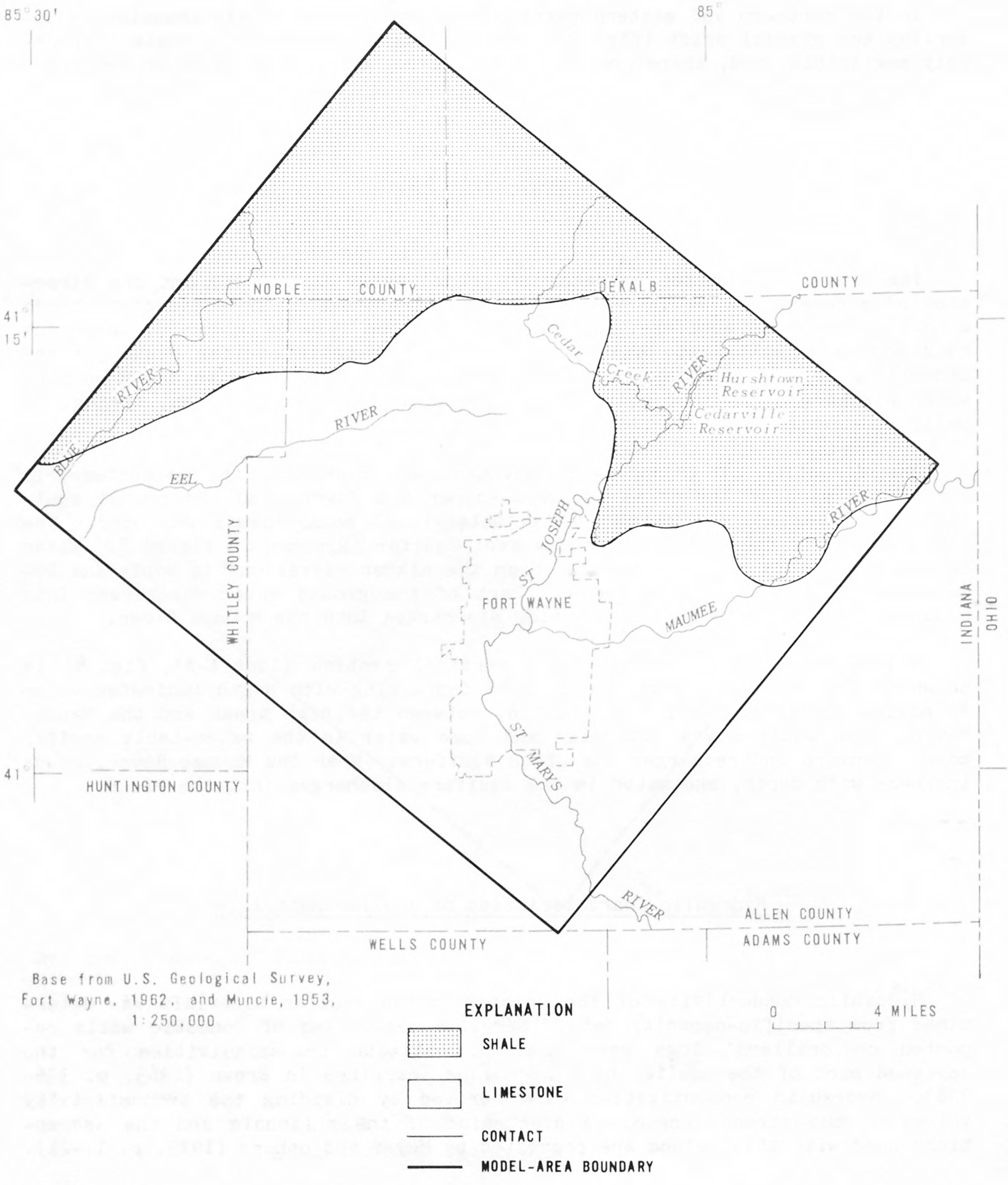
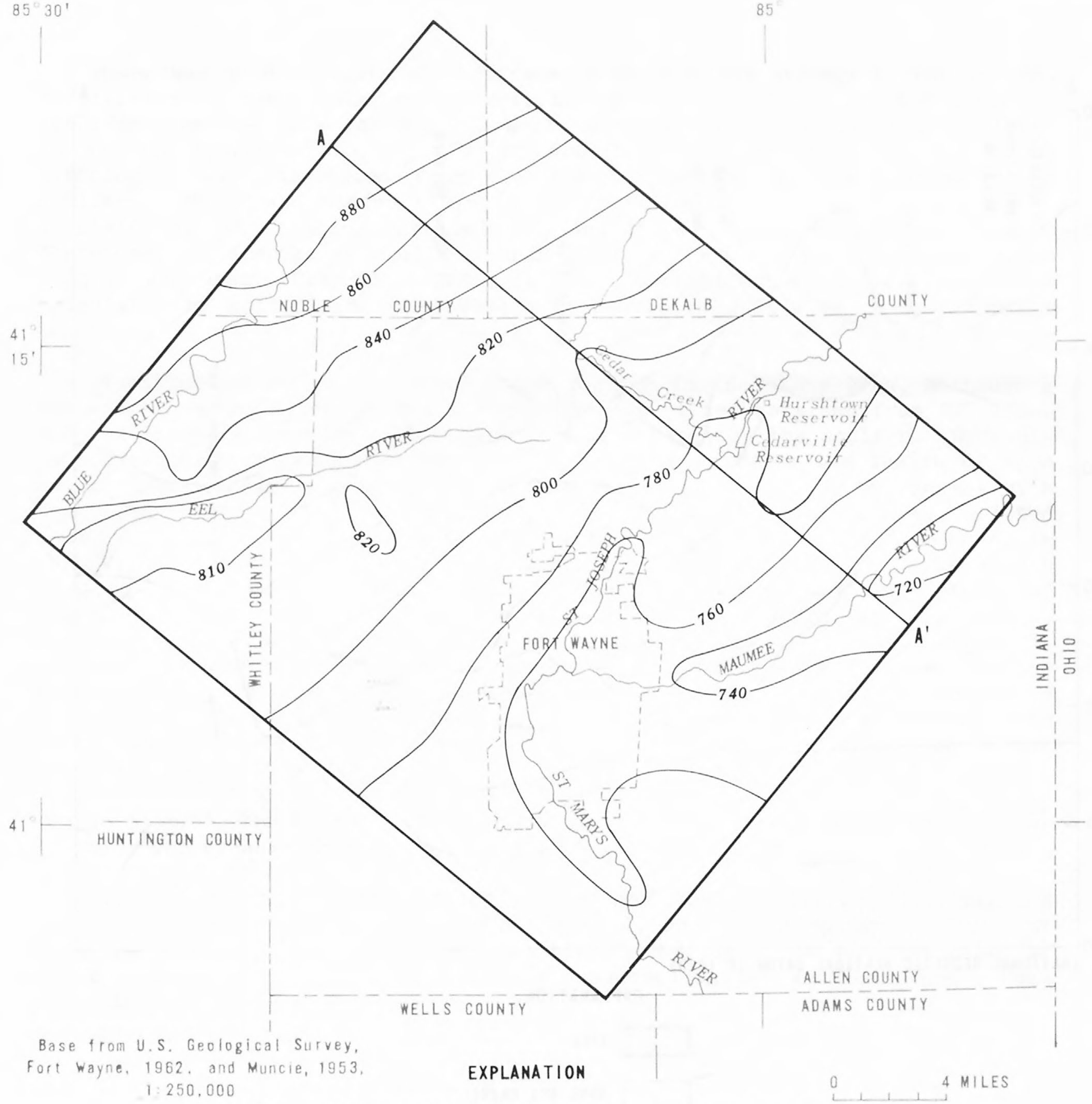


Figure 7.-- Contact line at the bedrock surface between shale and limestone.



EXPLANATION

—720— POTENTIOMETRIC CONTOUR -- Shows elevation at which water level would have stood in tightly cased wells. Intervals 10 and 20 feet
National Geodetic Vertical Datum of 1929

A———A' LINE OF SECTION SHOWN IN FIGURE 9

——— MODEL-AREA BOUNDARY

Figure 8.-- Potentiometric surface for the upper sand and gravel aquifer, September 1976.

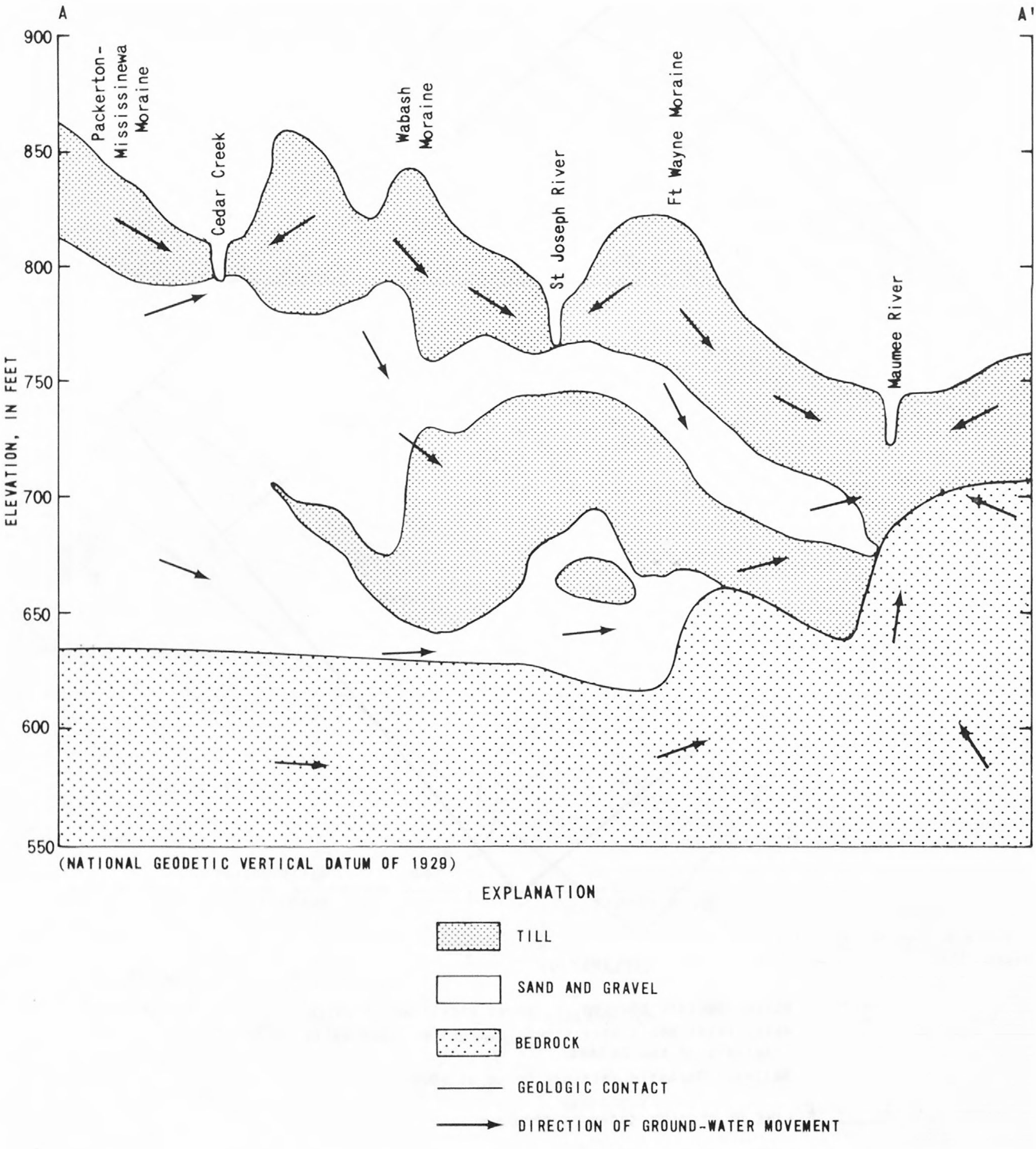


Figure 9.-- Generalized northwest-southeast geologic section showing aquifers and direction of ground-water movement near the confluence of Cedar Creek and the St. Joseph River.

There was no distinguishable difference between the average hydraulic conductivities of sand, sand and gravel, and gravel calculated on the basis of specific-capacity data for the 212 wells screened in unconsolidated aquifers. An average hydraulic conductivity of 400 ft/d (feet per day) for the three lithologies was calculated from all the information on the unconsolidated aquifers. Meyer and others (1975, p. 21) obtained an average hydraulic conductivity of 390 ft/d for confined sand and gravel aquifers in Marion County. Therefore, an average hydraulic conductivity of 400 ft/d for the three lithologies was used. Transmissivities for the unconsolidated aquifers were then calculated by multiplying the average hydraulic conductivity by the aquifer thickness.

Limestone is the major consolidated aquifer in the study area, although a few wells are completed in shale. Specific capacities obtained from 275 limestone wells were used to calculate transmissivities for the aquifer. Hydraulic conductivity was obtained by dividing transmissivity by the length of open hole. The average transmissivity of the limestone, 3,000 ft^2/d, agrees with the value obtained by R. W. Stallman in 1947 (U.S. Geological Survey, written commun., 1947) in a limestone aquifer test done for the U.S. Rubber Corp. in Fort Wayne. A plot of the average hydraulic conductivity against the length of open hole indicates that hydraulic conductivity decreases with depth of limestone penetration (fig. 10). The explanation for the decrease may be that the main water-bearing zone is thin and shallow and that as the well depth increases the hydraulic conductivity decreases. This explanation supports the assumption that a finite thickness of limestone is connected hydraulically to the glacial deposits. The decrease in apparent hydraulic conductivity with depth, attributable to partial penetration alone, is considerably more than one would expect.

No attempt was made to estimate storage coefficient or specific yield of the aquifers because only steady-state flow was considered.

Vertical hydraulic conductivity of the aquifer materials, horizontal and vertical hydraulic conductivities of the till, and hydraulic conductivity of the streambed were not determined directly in the field but were estimated by model calibration. Therefore, they are discussed in the sections "Model Construction" and "Calibration."

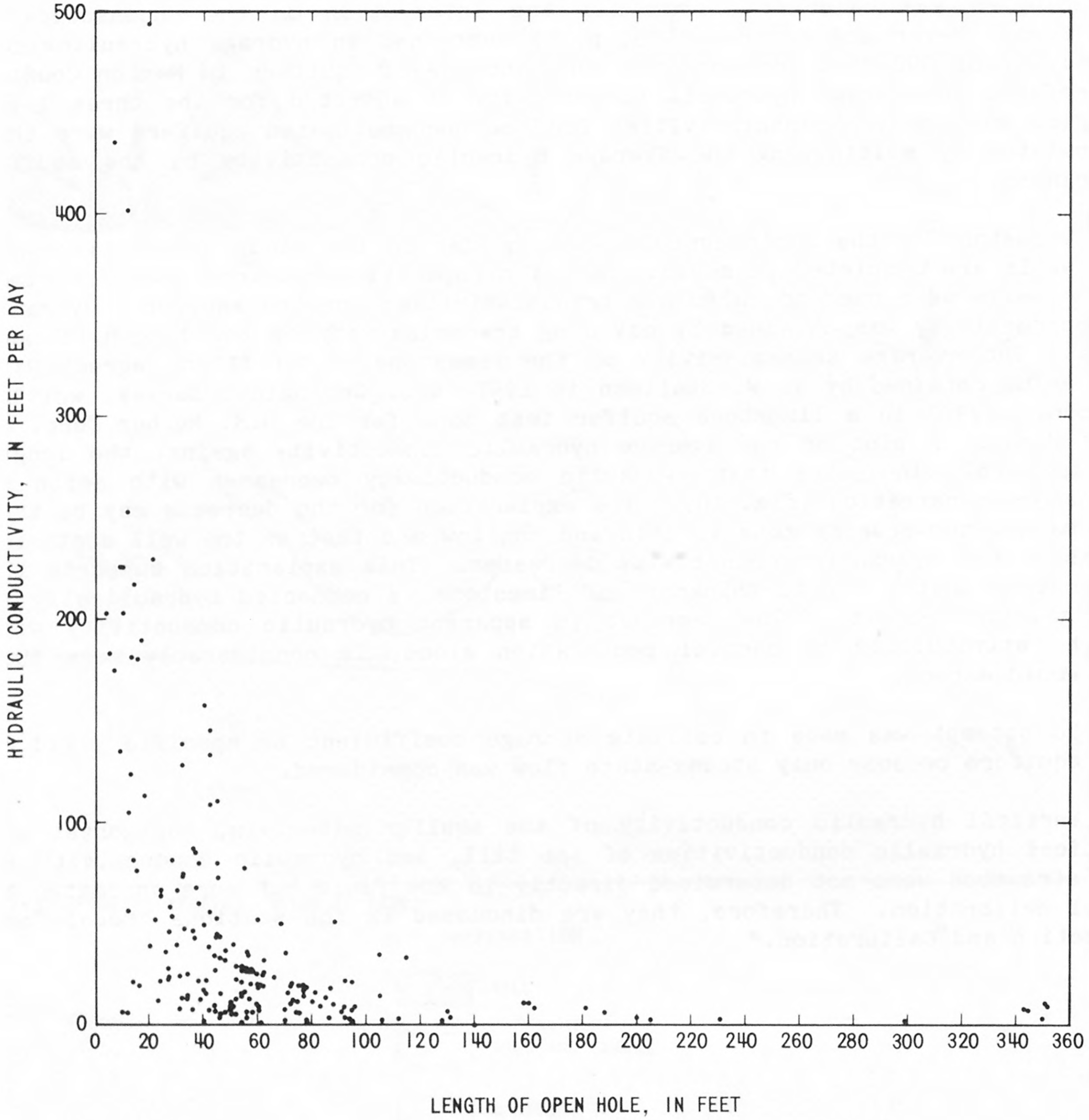


Figure 10.-- Relation of average hydraulic conductivity of the open hole to length of open hole.

The amount of water exchanged between the ground-water system and a stream can be described by the following equation:

$$Q_s = \frac{-K_{SB}}{Z} (h_s - h_a) A$$

where:

Q_s is flow between stream and underlying unit,

K_{SB} streambed hydraulic conductivity,

Z thickness of the streambed,

$(h_s - h_a)$ head difference between stream and underlying unit,

and

A area of streambed.

Flow gained along a reach of stream, head difference between the stream and underlying unit, and area of the streambed were measured in the field. Vertical hydraulic conductivity and thickness of the streambed were not measured directly but were estimated as a combined value termed "the stream-connection factor" (K_{SB}/Z). The head in the aquifer was assumed to be above the streambed bottom.

Streamflow measurements, from September 22 to 24, 1976, at various points and at 90-percent flow duration[1], were used to estimate the ground-water seepage between the points. Discharges at the measuring points are shown in figure 11. At 90-percent flow duration, the gain of flow in the streams was solely ground-water discharge.

Streambed materials, till and sand, whose vertical hydraulic conductivities differ significantly, were examined at each measuring point. At seven measuring points where sand was found, test holes were drilled to determine if there was till at shallow depths or if the stream was in direct connection with the upper aquifer. These test holes were converted to observation wells to determine the head difference between the ground-water system and the stream.

[1]The flow at 90-percent flow duration is exceeded 90 percent of the time.

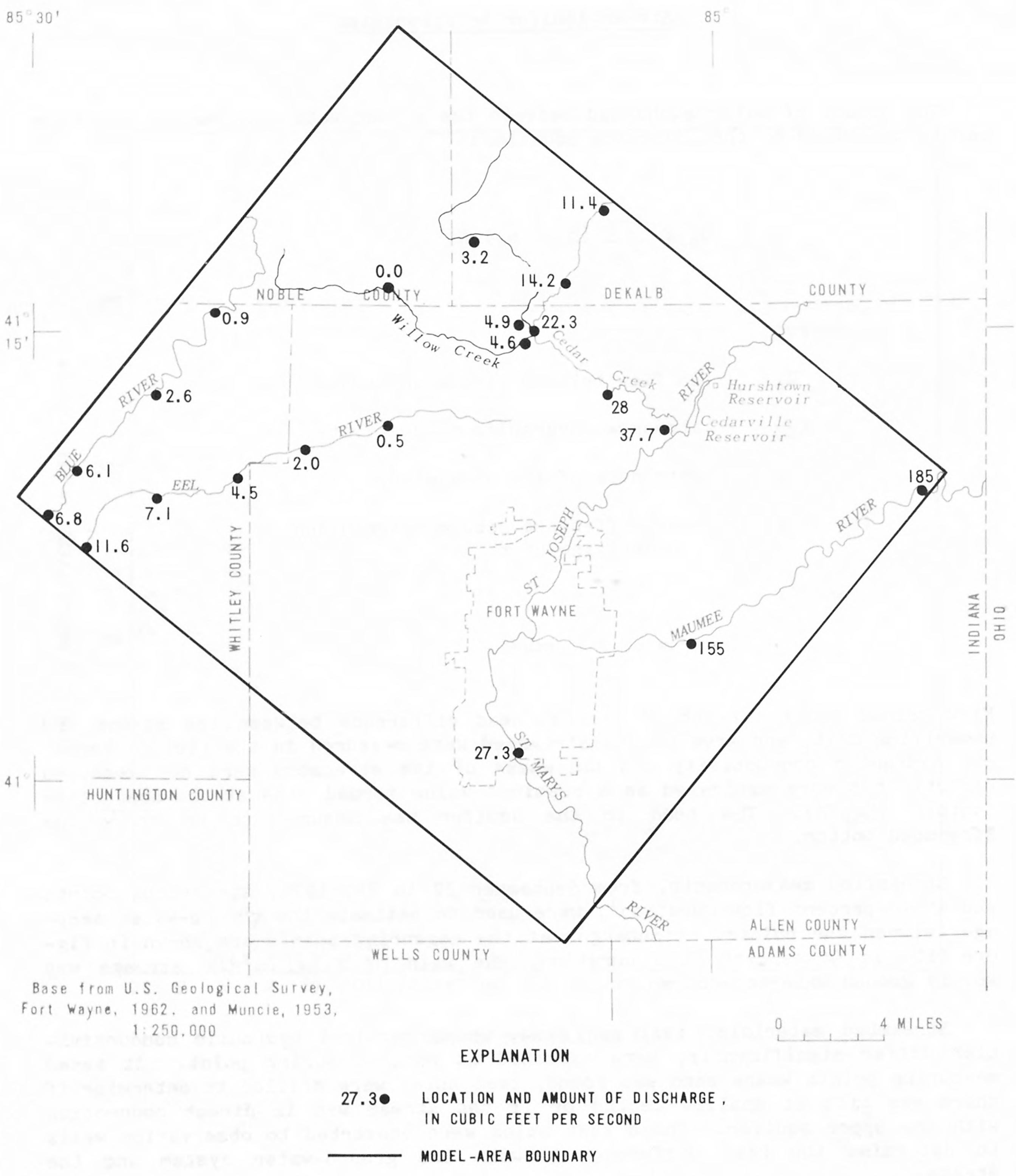


Figure 11.-- Discharge at selected sites, September 22-24, 1976.

Because of low streamflow and of backwater from several dams along the St. Joseph, St. Marys, and Maumee Rivers, streamflow could not be accurately measured. Determination of the hydraulic connection in these streams was based on the head differences between the stream and upper till and adjustment of the "stream-connection" factor during model calibration.

The reaches of a stream that can be used as a check during model calibration account for about 74 ft^3/s of a total discharge of 104 ft^3/s. The total ground-water discharge computed by the model, compared with the total measured discharges, should indicate whether adjustments of the model are acceptable.

Pumpage

All municipalities, industrial plants, and households in the primarily agricultural area outside the area serviced by the Fort Wayne water utility use ground water developed mainly from sand and gravel aquifers west of the St. Joseph River valley and north of Fort Wayne. Limestone is the major aquifer within the valleys of the Maumee and St. Marys Rivers and south and west of the city.

Six municipalities, including two inside the study area, and three private utilities that service housing developments surrounding Fort Wayne are within the drainage systems of the Maumee and Wabash Rivers. These nine suppliers represent the major pumpage (approximately 6.9 Mgal/d) from the sand and gravel aquifers. The largest of these suppliers, Auburn, whose average daily pumpage is 1.9 Mgal, serves a population of 7,400.

An inventory of industrial water usage showed that eight companies in Fort Wayne obtain all or part of their water supply from ground water. The average daily pumpage for these companies is about 7.4 Mgal. This pumpage is from wells completed in limestone, and 7.0 Mgal of it is from the four locations shown in figure 12.

MODELING THE GROUND-WATER SYSTEM

A computer program that simulates three-dimensional ground-water flow (Trescott, 1975), was used to model the Fort Wayne area. The program uses finite-difference techniques to solve the ground-water flow equation for three-dimensional, steady or nonsteady flow in an anisotropic, heterogeneous ground-water system. The finite-difference techniques require that the ground-water system be divided into rectangular blocks and that average values of each parameter be assigned to every block in the finite-difference grid. The horizontal finite-difference grid used in the model is shown in figure 13. The grid was designed to fit the major streams in order to minimize lateral displacement of the streams in the model. However, there is some displacement along the Eel, St. Marys, and Maumee Rivers. The grid was expanded by progressively multiplying each block width away from the streams by a maximum factor of two. The smallest dimensions for one block was 1,056 ft on a side and the largest was 8,446 ft by 16,896 ft. The total size of the network was 57 by 69 blocks.

Model Construction

The model was constructed to evaluate a steady-state system without using specific yield of the upper till and the storage coefficients of the aquifers and the lower till. The author considered the system to be at steady state on the basis of 19 yr of water-level records. Most of the water levels measured in September 1976 were within a few feet of the water levels recorded at the time the wells were drilled. Also, the assumed steady state is supported by the records of two observation wells outside but near the study area. Fluctuation of water levels in the two wells (one in Whitley County and one in Noble County) was less than 4 ft during 9 yr of records.

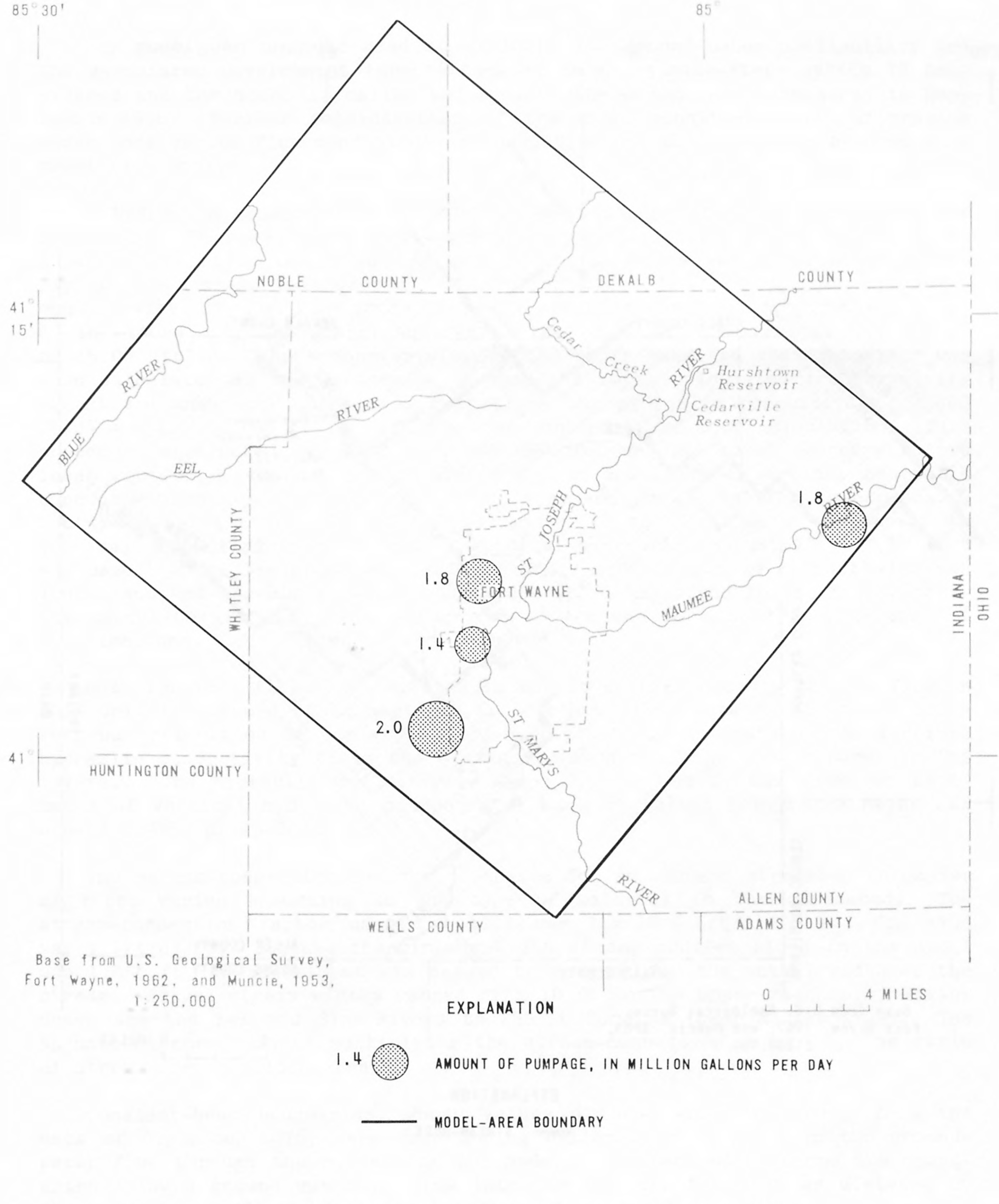


Figure 12.--Large-scale ground-water pumpage, September 1976.

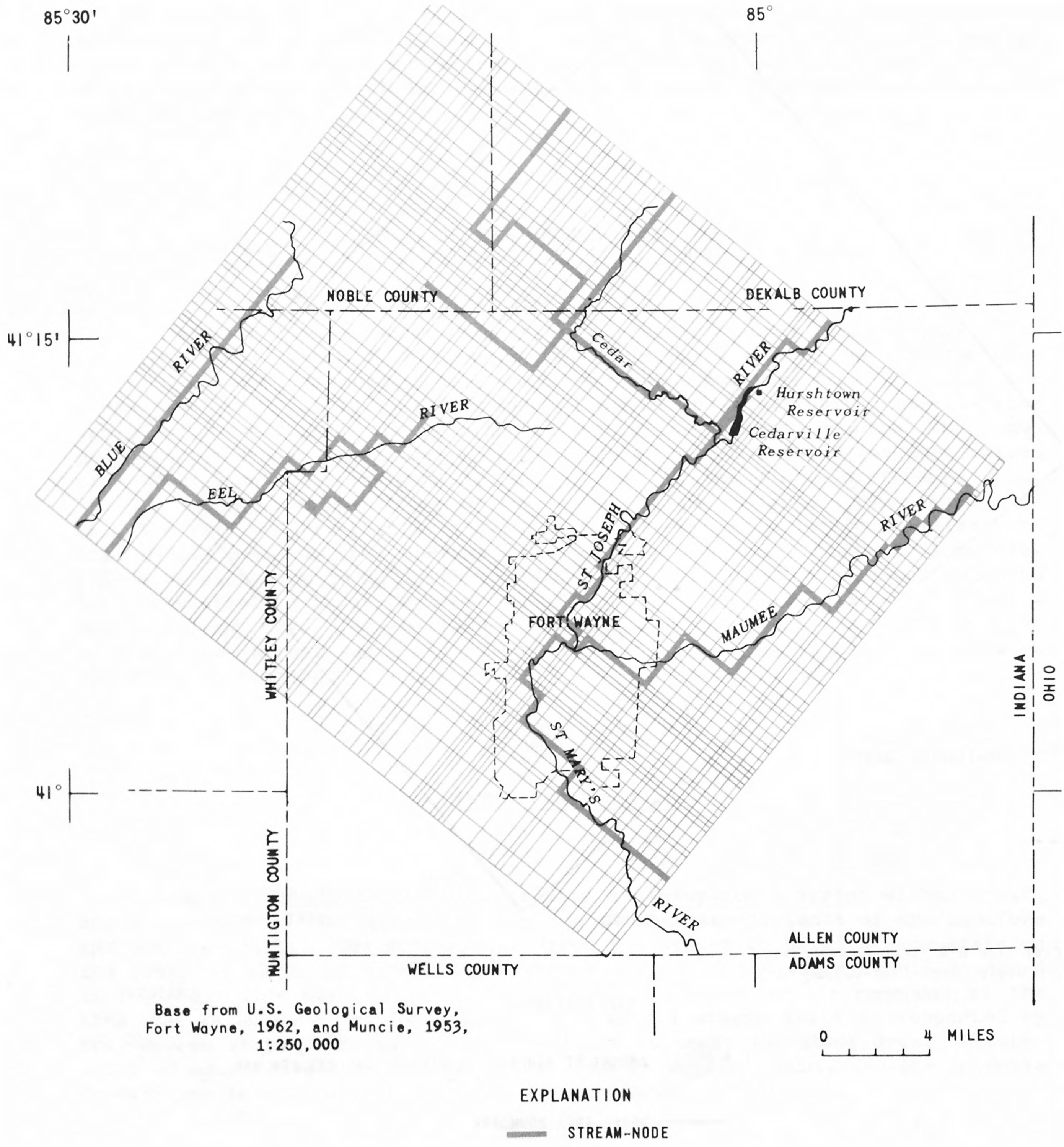


Figure 13 -- Horizontal finite-difference grid used in the model.

The model can only be used to evaluate the ground-water availability and the associated development impacts because only a steady-state system is considered and the model is calibrated against the water levels measured in September 1976. Further consideration of transient conditions and calibration under more varied flow conditions are needed before the model can be used as a predictive tool.

A hydraulic conductivity of 400 ft/d was assigned to sand and gravel for the study. In areas where the upper and lower sand and gravel aquifers are separated by till, the transmissivity of the aquifers was computed by multiplying the average thickness of an aquifer block by the assigned hydraulic conductivity; however, in areas where the till is missing, the transmissivity of the lower sand and gravel aquifer in the model was arbitrarily assumed to be 15,000 ft^2/d. The transmissivity of the upper sand and gravel aquifer was then calculated as the difference between the total calculated transmissivity (total thickness of sand and gravel times the hydraulic conductivity 400 ft/d) and the 15,000 ft^2/d to insure the continuity of model simulation. This procedure eliminated any need to compensate for an artificial boundary in the lower aquifer during pumping simulations. In addition, a vertical hydraulic conductivity had to be assigned to the sand and gravel aquifer to compensate for the missing till. The author assumed a 10 to 1 ratio of horizontal to vertical conductivity; therefore, a vertical hydraulic conductivity of 40 ft/d was used. Where limestone was at the bedrock surface and in contact with the lower sand and gravel, an additional 3,000 ft^2/d was added to the transmissivity calculated for the sand and gravel to account for the transmissivity of the limestone.

Only vertical flow was simulated in the lower till because lateral flow in this unit is assumed to be negligible. Vertical flow between all three aquifers was calculated for each block by multiplying an estimate of the vertical hydraulic conductivity times the average thickness of the till separating the layers. The hydraulic conductivity used, 2×10^{-4} ft/d, was based on estimates of vertical hydraulic conductivity made by Walton (1965) and Meyer and others (1975, p. 25-26).

The stream-connection factor, calcuated for an assumed streambed thickness of 1 ft, varied according to the type of material in the streambed. The stream-connection factor used for till was 1×10^{-2} (ft/d)/ft and for sand was 7 (ft/d)/ft. Because the minimum width of any aquifer block in the model was 1,056 ft, an adjustment was needed to account for the actual width of the stream. Actual stream widths ranged from 10 ft in the upper reaches of Willow Creek and the Eel and Blue Rivers to 250 ft for the lower Maumee River. The adjustment consisted of multiplying the stream-connection factor by the ratio of stream area to block area.

Constant-head boundaries, whose values of head were determined from the data of September 1976, were used during calibration to preserve the ground-water flow through the borders of the model. Constant heads along the boundaries allowed ground water to flow into, or out of, the area as dictated by the hydraulic gradients from the interior of the model.

The base of the model was considered to be impermeable. The reduction of hydraulic conductivity with depth in the limestone and the low conductivity of the shale should severely limit the contribution of water through this boundary when compared to the amount of water available from the lateral boundaries and the streams.

Ground-water discharge in the study area is a summation of the net ground-water inflow through the boundaries of the study area and the recharge from precipitation. An initial value of 2.0 in./yr was used for recharge from precipitation. This recharge rate, calculated from the increase in streamflow due to increase in drainage area, equaled 77 percent of the ground-water discharge to streams in the modeled areas. The recharge rate is subject to change because the relationship between the amount of ground water entering the area through the borders and by precipitation is not known. After calibration begins, a mass balance produced by the program determines the amount of flow into the model from the constant-head boundaries. The rate of recharge from precipitation can then be altered, if necessary.

Ground-water pumpage for the four locations shown in figure 12 was included in the model. The rate of pumpage has been constant for more than 10 yr; therefore, the ground-water system is probably at steady state.

Model Calibration

The model was calibrated to hydrologic conditions of September 1976. Water levels and seepage to streams were simulated by the model. To match the model results with the field values, the author adjusted the vertical hydraulic conductivity of the till layers, the vertical connection between the streams and the aquifer, and the transmissivity of the upper sand and gravel in one area.

The first adjustment that had to be made was to increase the vertical hydraulic conductivity of the upper till because the small amount of downward flow was causing unreasonably high heads in the upper aquifer. The vertical conductivity in the upper till was increased from the initial 2.0×10^{-4} to 2.3×10^{-2} ft/d. This change produced higher heads in the lower aquifer, so it was necessary to reduce the vertical conductivity of the lower till to 1.5×10^{-4} ft/d. This reduction for the lower till also means that the vertical conductivity for the sand and gravel in the area, where the lower till is missing, was also changed. The vertical hydraulic conductivity of the sand and gravel was reduced from 40 to 30 ft/d.

Along the lower reach of the St. Joseph River and the entire Maumee River, the vertical resistance to flow through the till layers had to be decreased to roughly the value of the stream-connection factor. Use of the values originally calculated prevented discharge to these streams, raised the water levels, and forced discharge through the borders of the model. The St. Joseph and Maumee Rivers have probably downcut and refilled their valleys because depths to bedrock in these areas are generally less than 50 ft. This depositional history would explain the higher vertical hydraulic conductivity of the till underlying these streams than that elsewhere.

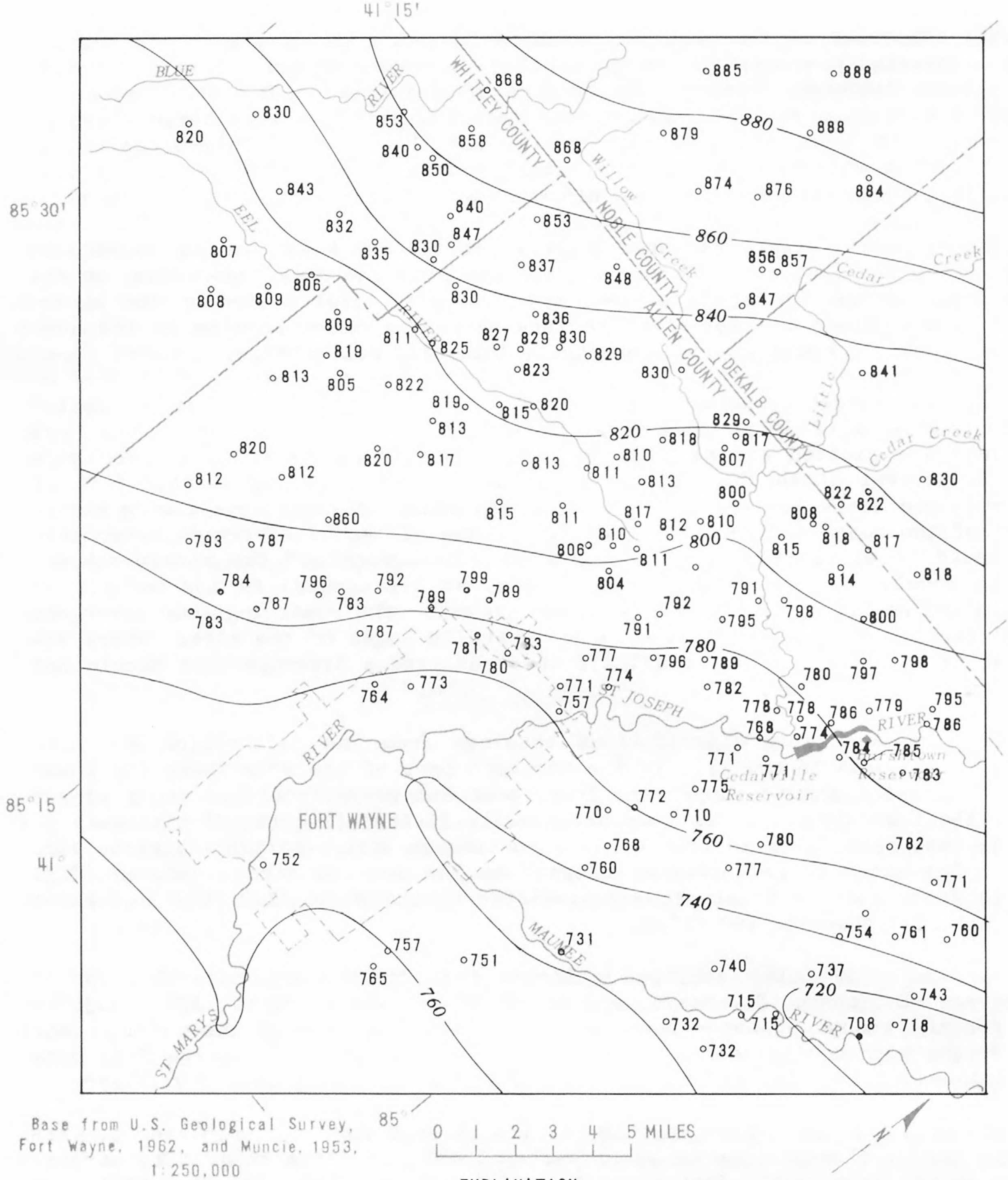
The transmissivity in an area surrounding the Eel River had to be reduced from 15,000 ft^2/d to 3,000 ft^2/d to lower heads in the area, and, thus, to reduce ground-water discharge to the river. Specific information on the lithologic nature along the Eel River is lacking, and the end moraine in the area could extend to depths greater than those initially estimated.

The potentiometric surfaces produced by the calibrated model are presented in contour maps of 20-ft intervals for each aquifer. Points where head was measured are plotted on the maps for comparison (figs. 14 and 15). For more than 60 percent of the blocks, the calculated water level was within 5 ft of the measured water level. All but 11 of 334 measured water levels were within 10 ft of the model-derived water levels. Seven of the 11 divergent water levels could be attributed to the artificial displacement of the stream courses in the model. At some points, the stream, as represented in the model, had been moved nearly a mile from its actual course. The remaining four divergent water levels were in two isolated areas near the edges of the model, where detailed information was not available; however, these discrepancies should not influence the analysis.

The transmissivity distributions obtained from the calibration are presented in figures 16 and 17. In the northern part of the area where the lower till is missing and the upper and lower sand and gravel aquifers are a single unit, the transmissivity has been arbitrarily divided into two fictitious layers so that each layer is treated as a continuous unit over the entire modeled area. The match of ground-water seepage to the modeled stream reaches (fig. 18) is good, and the total seepage simulated by the model (109 ft^3/s) compares well with the observed 104 ft^3/s.

The match of total seepage indicates that the stream-connection factor along the St. Joseph, St. Marys, and Maumee Rivers was correct. After adjusting for the amount of water supplied by the net flow through the model boundaries, the author obtained a final value of 1.6 in./yr for recharge from precipitation. This value is slightly less than the initially used 2.0 in./yr.

In September 1976, flow duration of the streams was low (90 percent). Although yearly fluctuations in ground-water levels are less than 5 ft, streamflows change drastically during the year. Selection of the low-flow period as a base for the simulation of pumpage restricts the amount of stream water available for infiltration and the amount of water that can be induced by pumping.



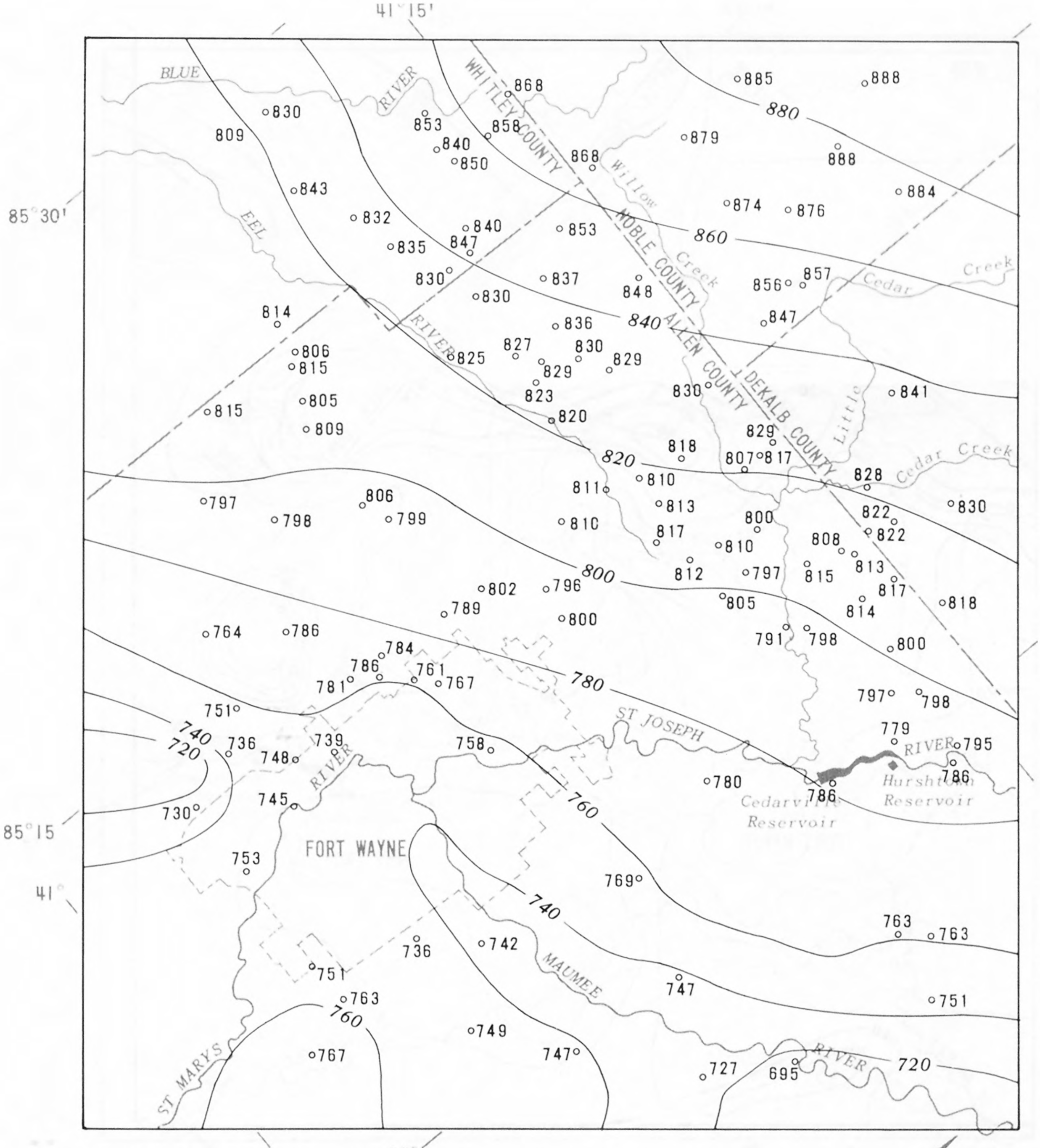
EXPLANATION

— *780* — POTENTIOMETRIC CONTOUR-- Shows elevation at which water level would have stood in tightly cased wells. Contour interval 20 feet. National Geodetic Vertical Datum of 1929

○751 POINT OF MEASURED WATER LEVEL, IN FEET

———— MODEL-AREA BOUNDARY

Figure 14.-- Comparison of model-derived potentiometric surface with measured water levels in the upper aquifer, September 1976.



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

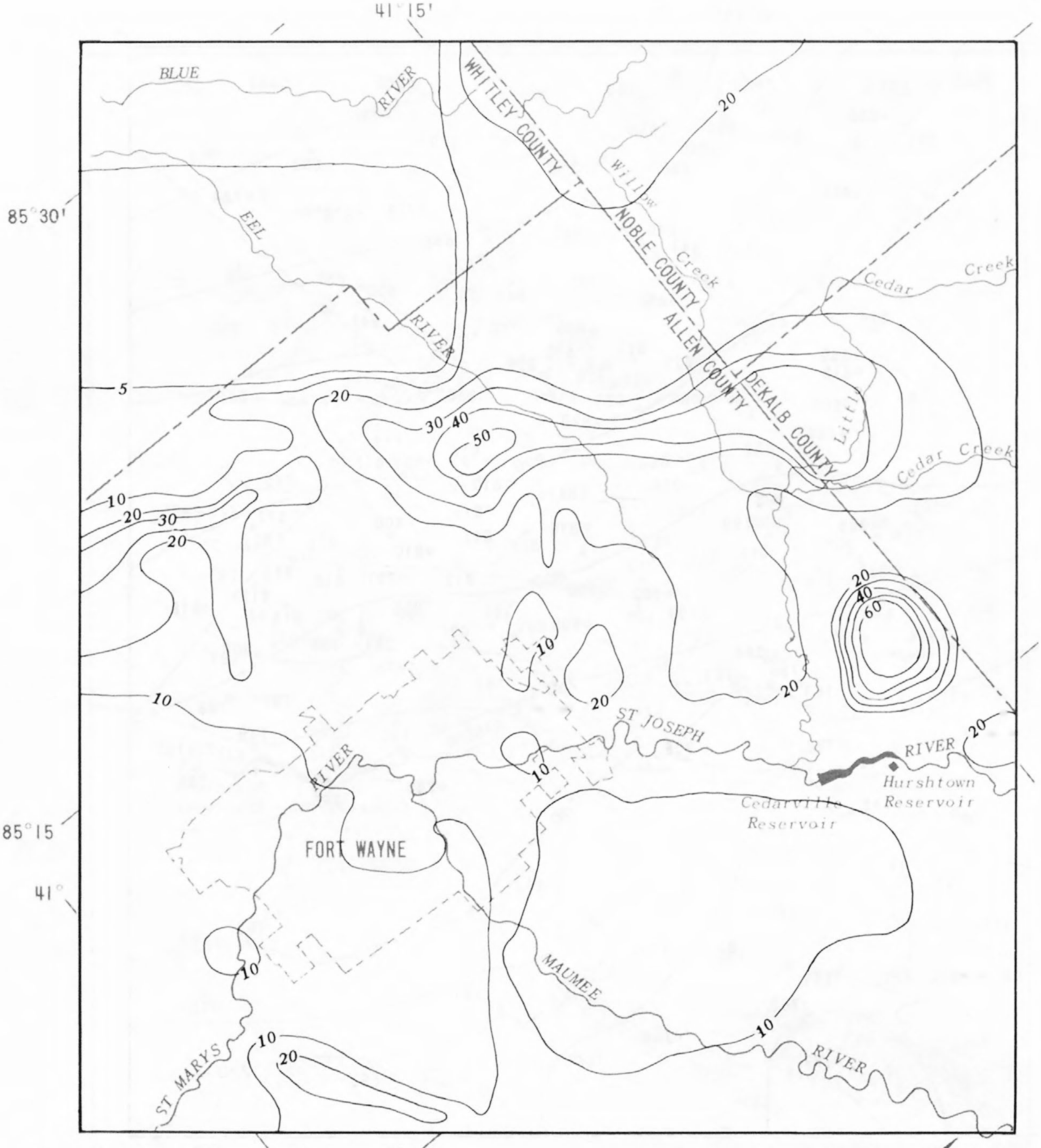
EXPLANATION

—720— POTENTIOMETRIC CONTOUR -- Shows elevation at which water level would have stood in tightly cased wells

747∘ POINT OF MEASURED WATER LEVEL, IN FEET National Geodetic Vertical Datum of 1929

——— MODEL-AREA BOUNDARY

Figure 15.-- Comparison of model-derived potentiometric surface with measured water levels in the lower aquifer, September 1976.



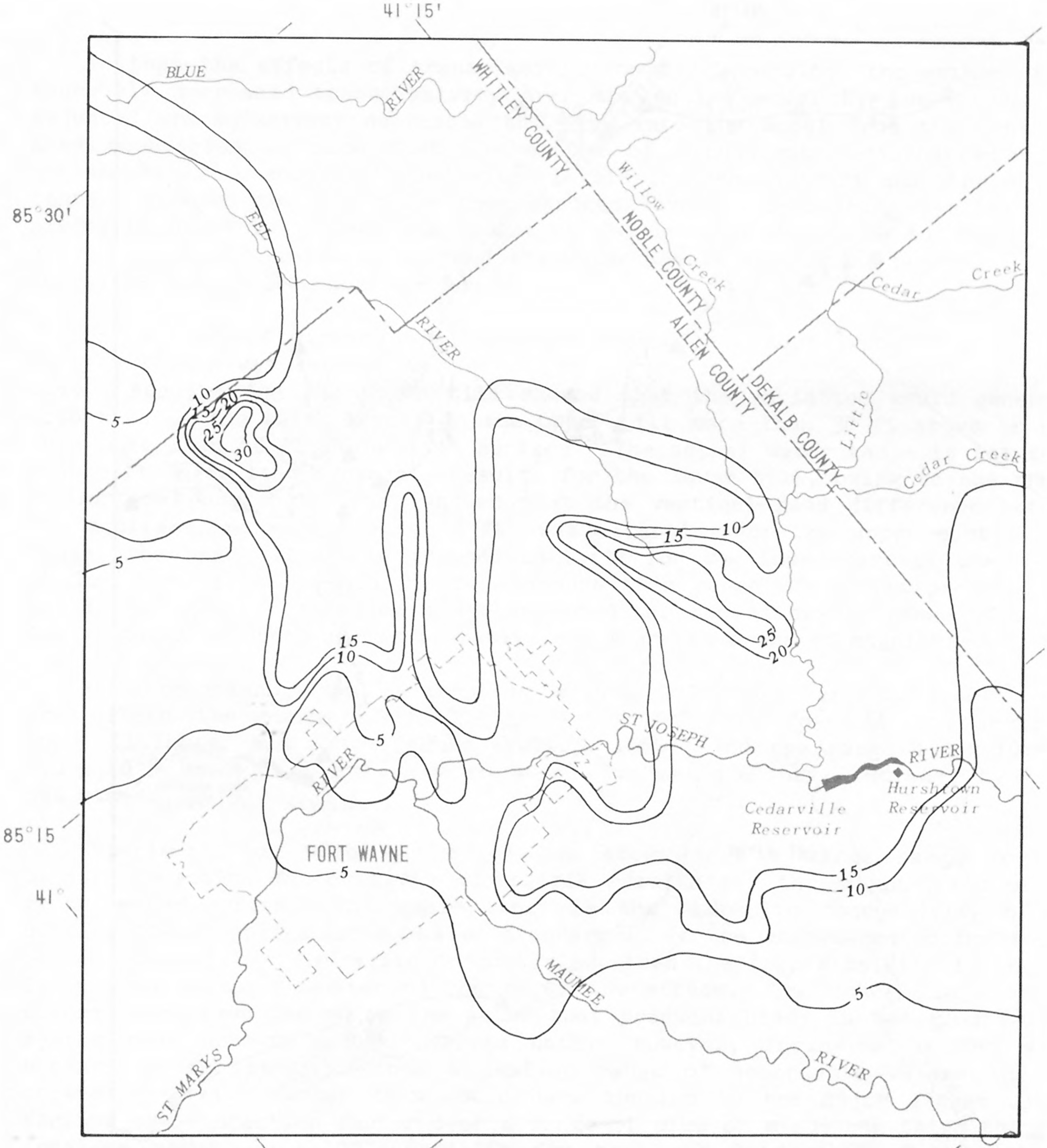
Base from U.S. Geological Survey,
Fort Wayne, 1962, and Muncie, 1953,
1:250,000

EXPLANATION

—10— LINE OF EQUAL TRANSMISSIVITY -- Intervals 5,000 and 10,000 feet squared per day

—— MODEL-AREA BOUNDARY

Figure 16.-- Transmissivity of the upper aquifer.



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

0 1 2 3 4 5 MILES

EXPLANATION

—10— LINE OF EQUAL TRANSMISSIVITY -- Interval 5,000 feet squared per day

——— MODEL-AREA BOUNDARY

Figure 17.-- Transmissivity of the lower aquifer.

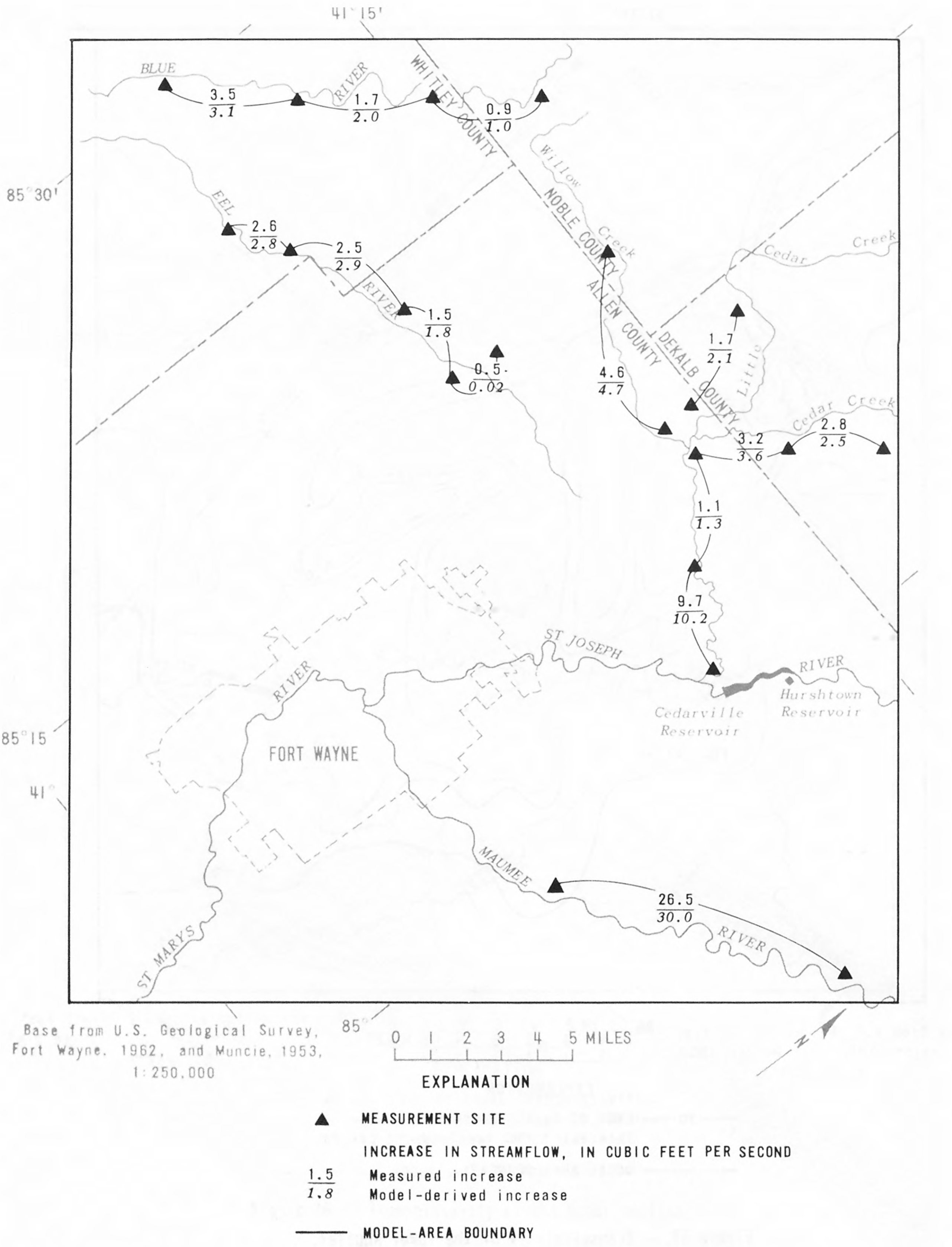


Figure 18.-- Comparison of model-derived increase with measured increase in streamflow, September 1976.

To test the effects of transmissivity on model results, the author simultaneously decreased transmissivity over the entire model by one-third. The adjusted transmissivity decreased the flow into the model from the constant-head boundaries so much that the amount of ground-water discharge to the streams became unacceptable and water levels in both aquifers near the pumping centers dropped to well below the measured levels. Increasing the transmissivity by one-third raised the amount of ground-water discharge to the streams above acceptable limits and raised the water levels around the pumping centers to values well above those measured.

The effect of varying the vertical hydraulic conductivity of all tills simultaneously was tested over a range of plus or minus one order of magnitude. Results for the upper till showed that the variation would generally either raise the water levels in the upper till more than 30 ft above or drop them more than 40 ft below land surface. The actual water table is between 5 and 20 ft below land surface. Results for the lower till, owing to the change in vertical conductivities, showed that the vertical head difference between the aquifers, normally 10 to 15 ft, was altered. For the upper part of the range, the head difference dropped to 1 ft; for the lower part of the range, it increased to more than 25 ft. Because this range of values produced results that greatly exceeded the observed data, the allowable range of values for vertical conductivity is probably less than one order of magnitude.

For comparison, the calibration values 2.3×10^{-2} and 1.5×10^{-4} ft/d are within the range 8.4×10^{-2} to 4.0×10^{-5} ft/d for till in areas of Ohio, Illinois, and South Dakota (Walton, 1965), and the range 1.5×10^{-2} to 1.0×10^{-4} used in a ground-water model in Marion County, Indiana, by Meyer and others (1975).

Experimentation with the model has shown that a large range for the stream-connection factor gives acceptable results for the major drain of the ground-water system. The upper limit is the hydraulic conductivity of the aquifer that supplies ground-water discharge. If the stream-connection factor exceeds the aquifer hydraulic conductivity, then the transmissivity is the only control on ground-water discharge to the stream. The lower limit of the stream-connection factor is the value that produces heads in the ground-water system that do not match observed data. However, inclusion of the minor streams in the study defined a smaller range of acceptable values for the stream-connection factor than would have applied to the major stream alone. Varying the connection factor over a range of plus or minus one-third the calibration values [1×10^{-2} (ft/d)/ft for till and 7 (ft/d)/ft for sand] altered the computed ground-water discharge to the minor streams considerably; however, discharge to the main drain and most water levels remained within the limits of the match.

The minor streams are more sensitive to the stream-connection factor than the major stream because only a part of the ground-water flow discharges into them and the remainder flows downgradient to the major stream. The amount of ground water that discharges into small streams depends on the hydraulic connection between those streams and the ground-water system. For the major stream, however, all ground-water flow must discharge into the stream. Gradients in the ground-water system depend on the distribution of transmissivity and stream-aquifer connection. Because the stream-connection factor is the vertical conductivity divided by the thickness of the streambed, no calibration values can be assigned to the two individual parameters.

The results of the preceding tests show that the values obtained for the three parameters that were undefined before model calibration are constrained; therefore, depending on future stresses under steady-state conditions, the model can probably be used to describe the response.

Pumping Evaluation

The purpose of this study was to determine the feasibility of developing ground-water supplies from various parts of the study area and to evaluate the hydrologic impacts of the development. The simulated steady-state conditions, where no water is derived from storage, are probably conservative and yield maximum drawdowns.

The model was calibrated with data collected in September 1976, which may not represent the average aquifer conditions. Calibration was done at low streamflow, so the simulated results are probably conservative. For steady-state conditions, simulated effects on the streams and boundaries are immediate, and there is no time delay in drawdown caused by water taken from storage as in the actual system.

Because the aquifer extends beyond the limits of the study area, the constant-head model boundaries could influence the model results in determining availability of water and impact of development. The proposed pumpage combined with present pumpage could affect water levels beyond the model boundaries; however, the assumed constant-head boundaries do not allow such changes in water levels beyond the boundaries. The effects of change of water levels and flow across the boundaries depend on characteristics of the aquifer in the area beyond the modeled boundaries and the distance between pumping centers and the boundaries. Two simulations for determining the effect of boundaries on model results were made for each pumping distribution. Constant-head boundaries were used to give a maximum flow across the model boundaries and, thus, less drawdown within the model area. Constant-flux boundaries were used to give a minimum flow across the model boundaries and to yield large drawdown. The values used for the constant-flux boundaries were determined by the model during calibration and were equal to the flow rates through the boundary nodes under constant-head conditions.

Pumping evaluations included those for the four pumping centers identified in figure 12. Representatives of local industries have projected little change in their pumping demands during the time being considered, so pumping rates should remain constant during the period of interest.

Model experiments 1 and 2 were designed to demonstrate the effects of pumping on the ground-water system and minimize diversion of flow from Fort Wayne's surface-water supply. Ten hypothetical pumping sites were chosen for the two experiments, six in the upper aquifer and four in the lower aquifer (fig. 19). The sites are in areas of high transmissivity. Rate of pumping was adjusted to draw the head down to the top of the aquifer at each block representing a pumping site. This procedure prevented errors in the calculation of flow because the model does not correct transmissivity if the computed water level in a confined aquifer drops below the top of the aquifer. The production available at each site does not necessarily represent one well. Furthermore, using the average head over a block as the constraint on dewatering does not imply that dewatering would not occur in the immediate vicinity of pumping wells in the real system. The number of wells and their deployment would determine the possibility and severity of aquifer dewatering.

Constant-head boundaries were used in model experiment 1. The pumpage simulated for this experiment was 38.9 Mgal/d. The resultant lines of equal drawdown are shown in figure 20, and streamfow losses are shown in table 1. The stream reaches identified in table 1 correspond to those shown in figure 21.

The amount of water diverted from Willow Creek by pumping wells (fig. 19) exceeds the flow measured in September 1976 (4.6 ft^3/s) by 1 ft^3/s; this diverted water would introduce a slight error in the drawdowns by providing a nonexistent source of water. A second simulation was run to test the magnitude of error caused by the overdraft; in this simulation the flow from Willow Creek was removed from the experiment. The results of the second simulation showed no appreciable change in drawdown (less than 0.1 ft). Cedar Creek would provide the extra 1 ft^3/s of water. So, if actual pumping would cause Willow Creek to go dry, the flow in Cedar Creek could easily supply the additional flow needed without significantly affecting drawdown.

In experiment 2, the constant heads at the boundaries of the model were converted to constant fluxes. The pumpage produced for this experiment was 34.7 Mgal/d. The resulting drawdowns for the two aquifers associated with the pumpage are shown in figure 22, and the streamflow losses are shown in table 1. The flow induced from Willow Creek in experiment 2 exceeded the flow measured in September 1976 by about 2 ft^3/s, but, as in experiment 1, this deficit could be supplied by Cedar Creek without significantly affecting drawdown.

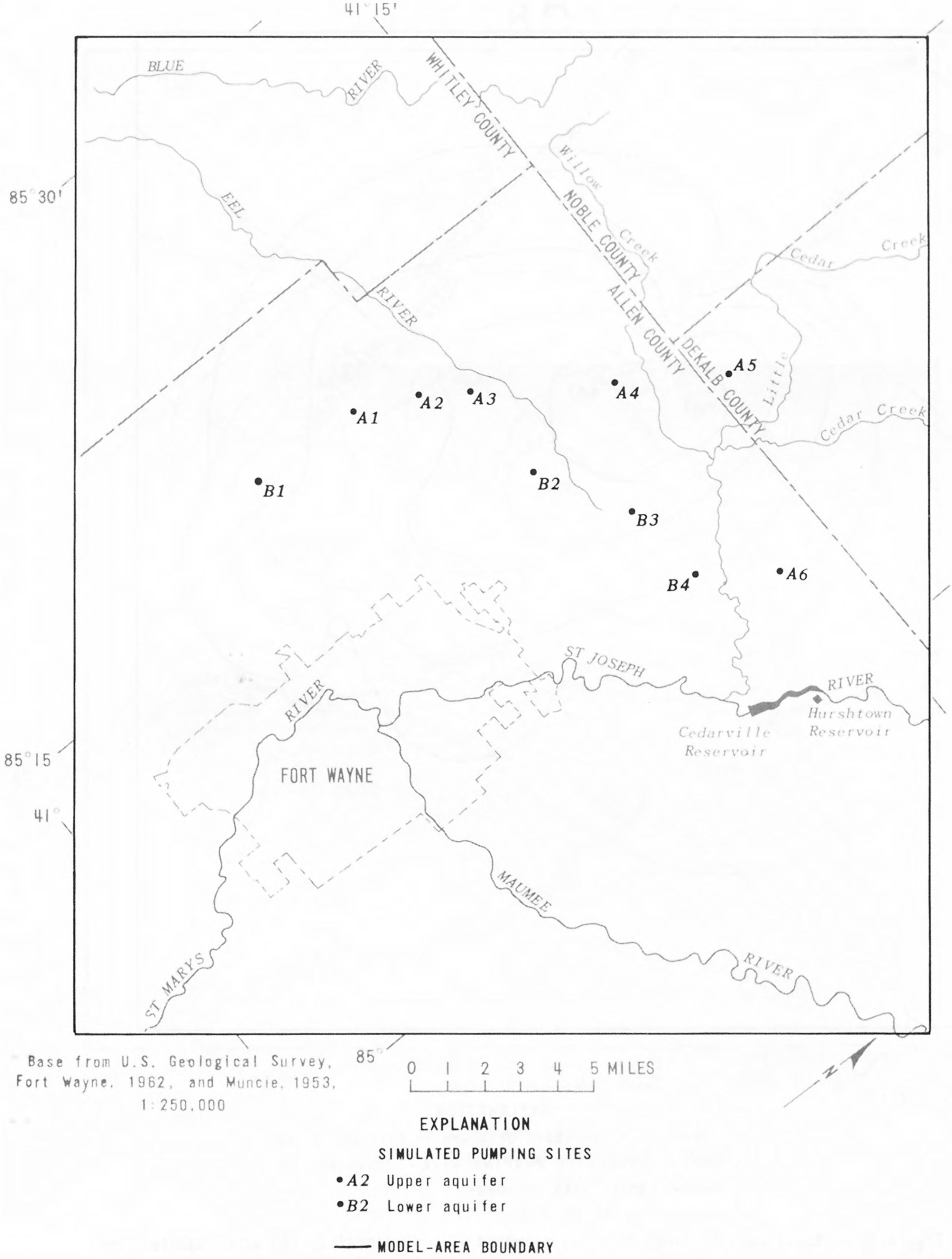
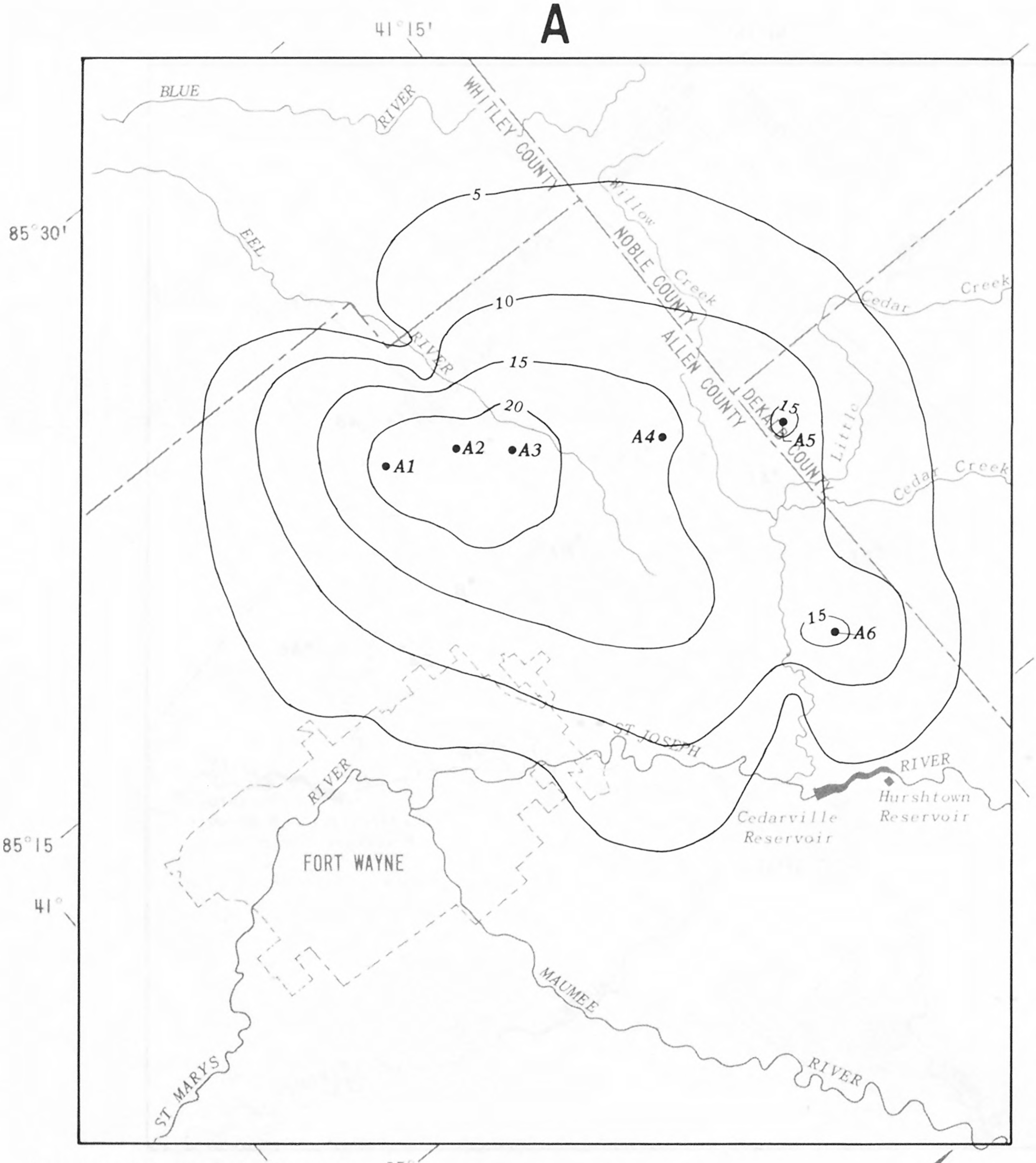


Figure 19.-- Locations of simulated pumping sites for model experiments 1 and 2, near Fort Wayne, Ind.



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

0 1 2 3 4 5 MILES

EXPLANATION

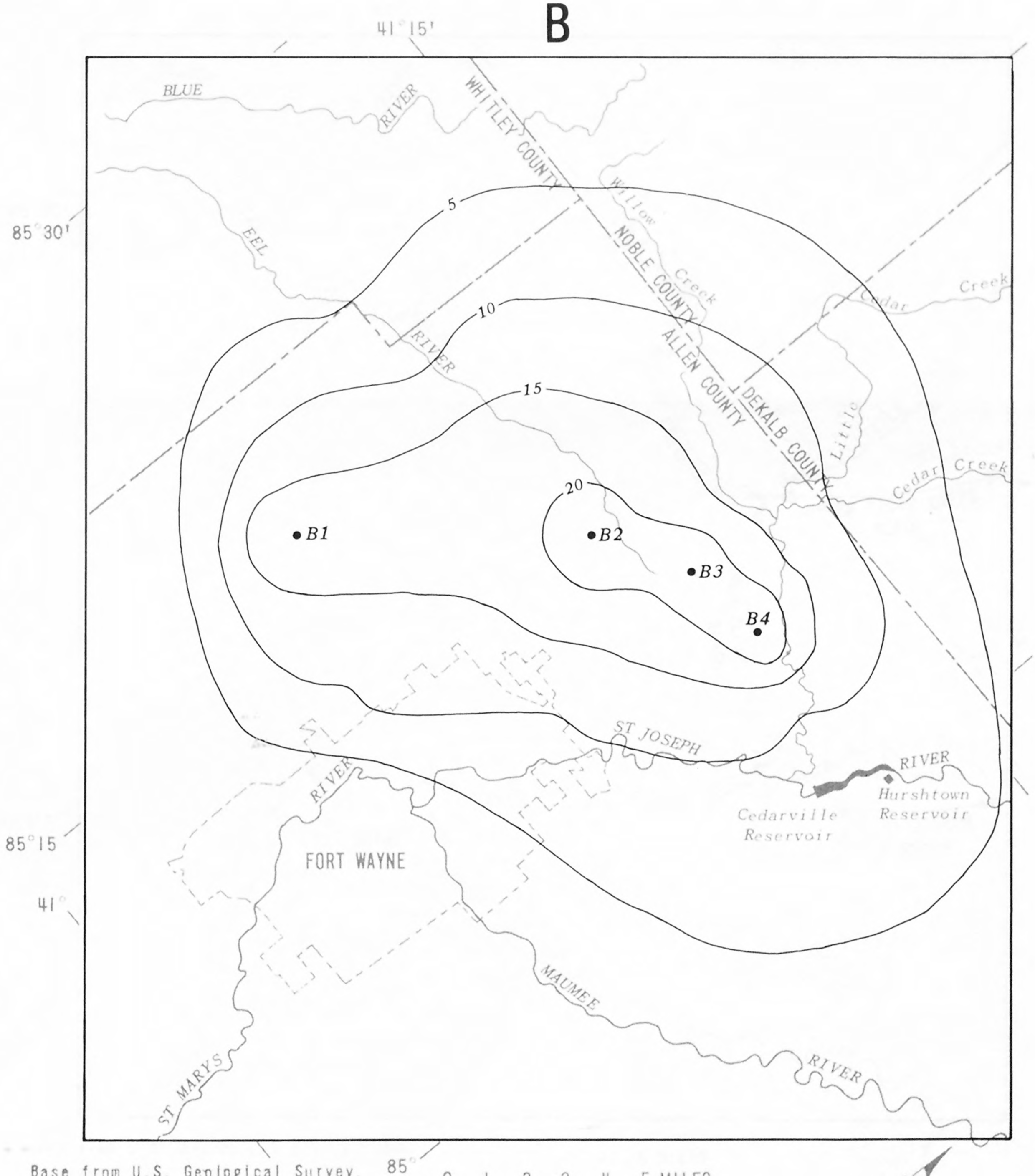
—15— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

•A1 SIMULATED PUMPING SITE -- Upper aquifer

—— MODEL-AREA BOUNDARY

Figure 20.-- Model-derived lines of equal drawdown for experiment 1, (A) upper aquifer and

B



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

EXPLANATION

—15— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

• B4 SIMULATED PUMPING SITE -- Lower aquifer

—— MODEL-AREA BOUNDARY

(B) lower aquifer, where the total pumpage for constant-head boundaries equaled 38.9 million gallons per day.

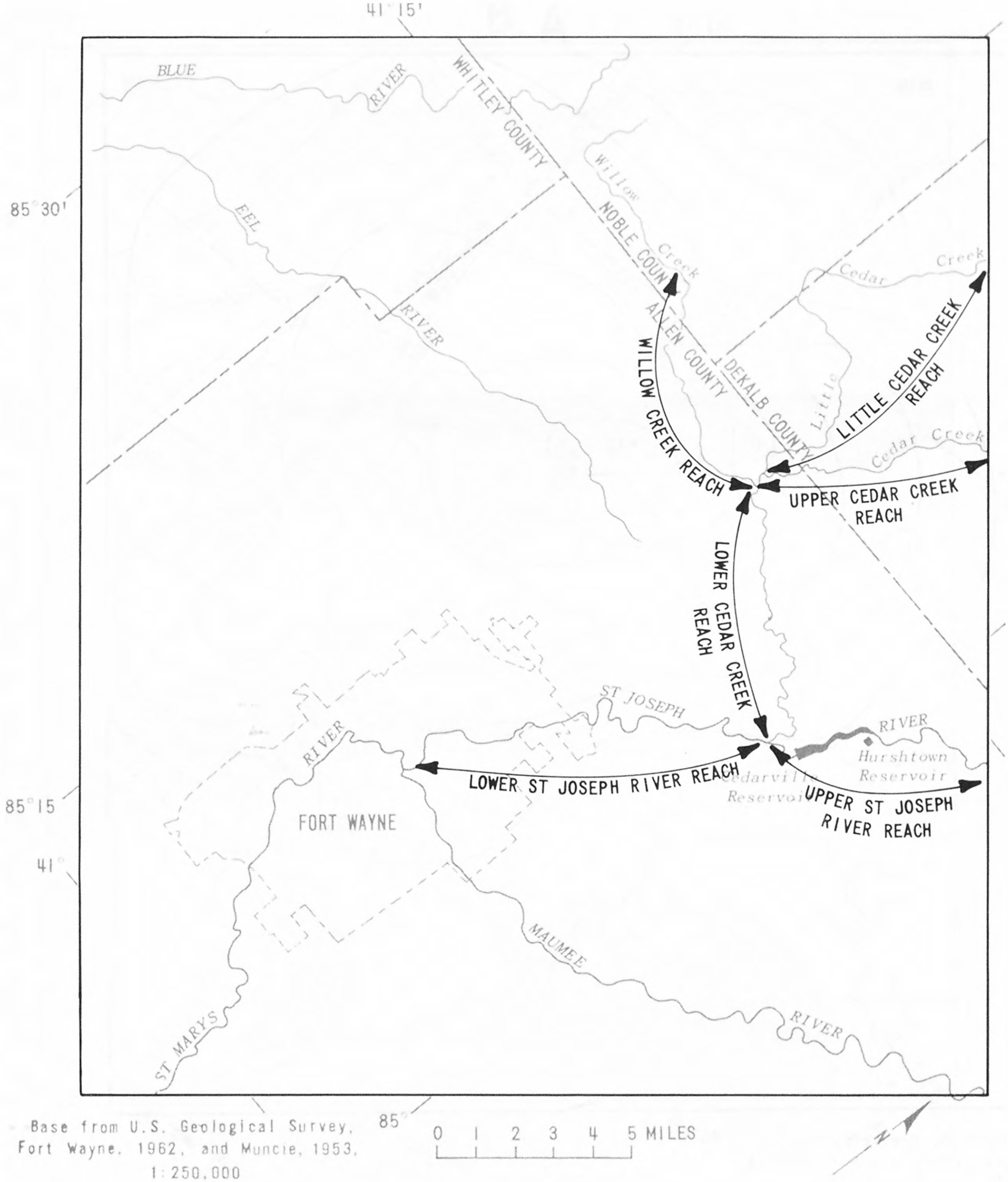
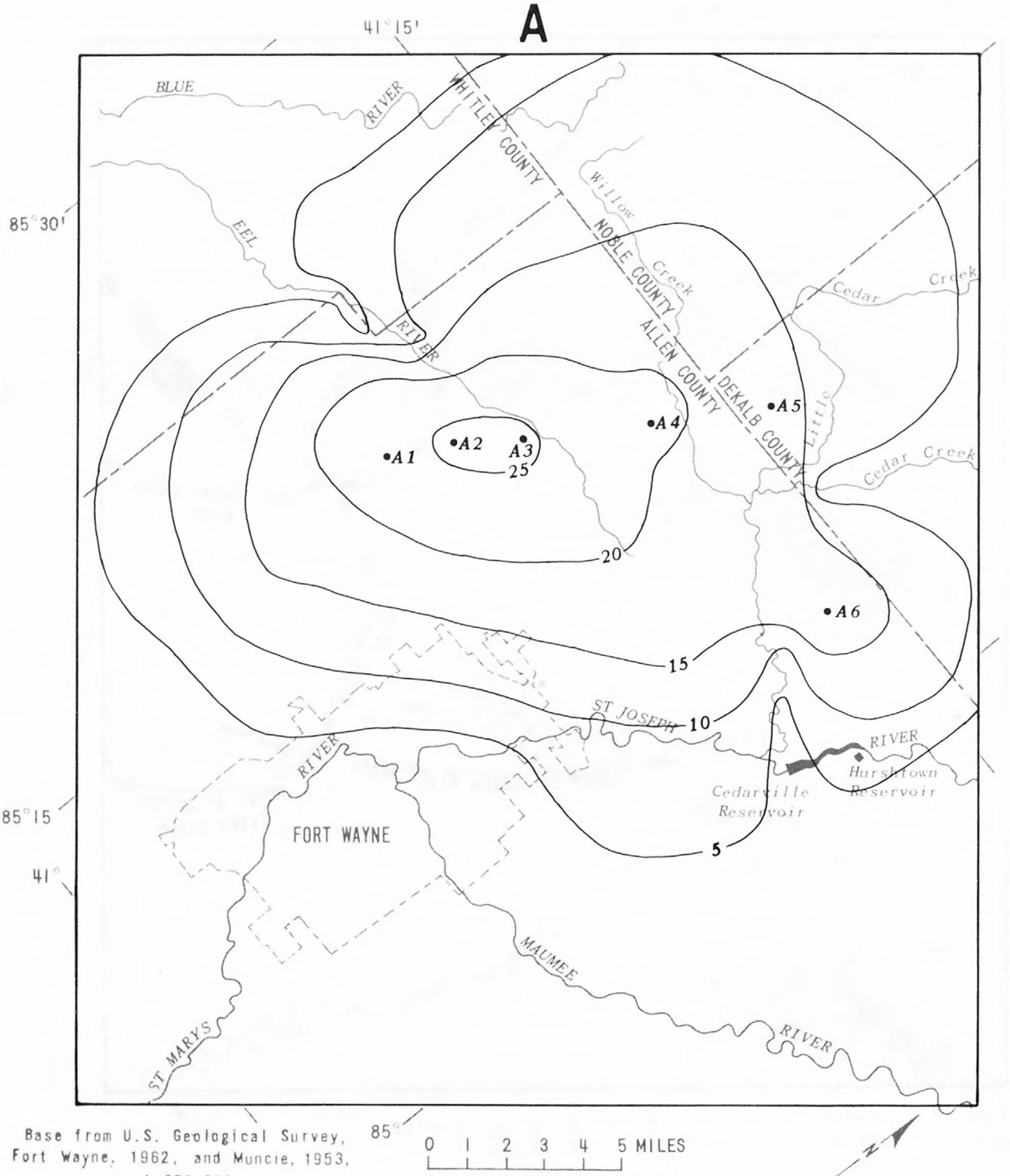


Figure 21.-- Locations of stream reaches used to determine streamflow loss due to pumping.

A



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

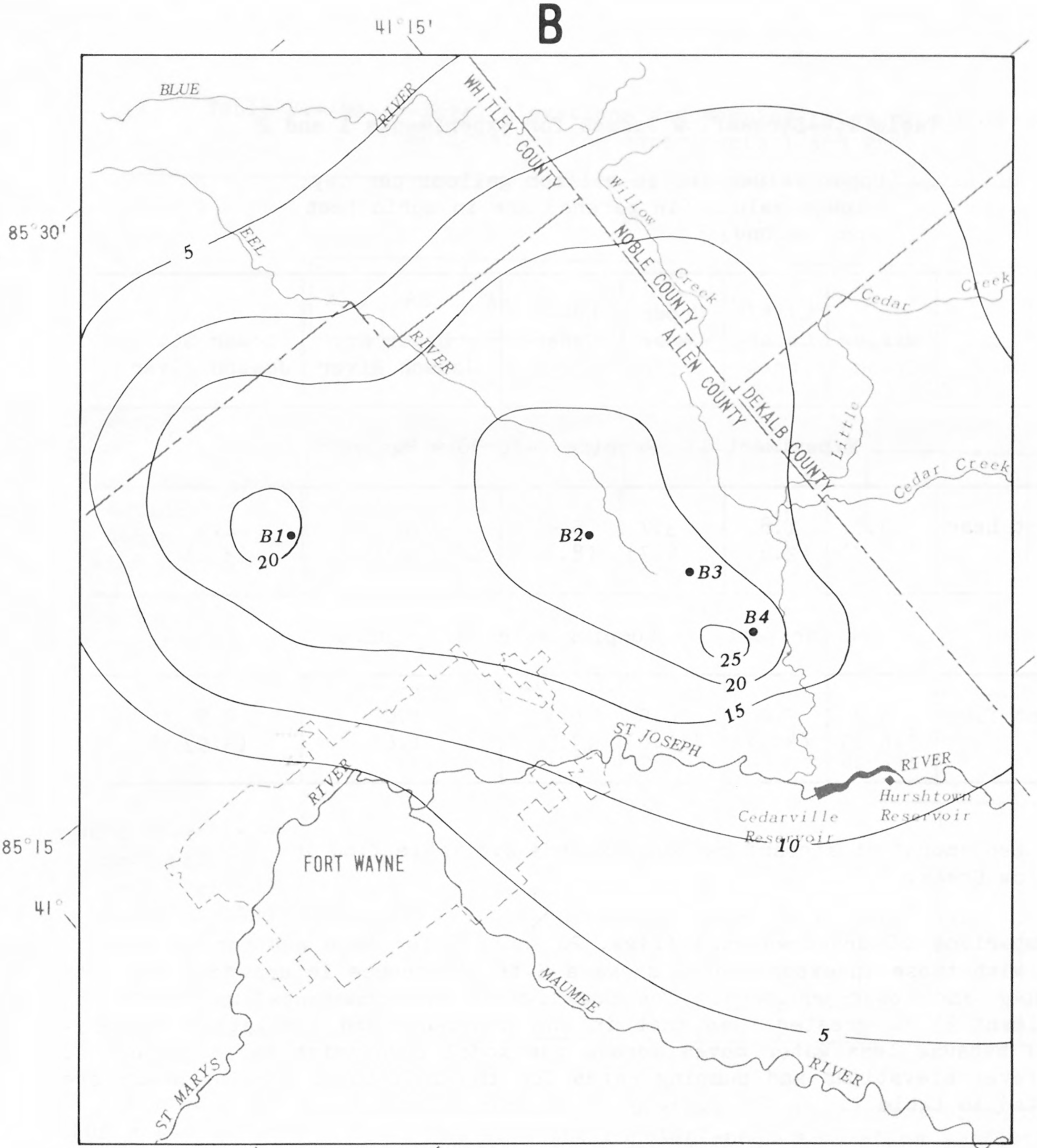
EXPLANATION

—20— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

•A2 SIMULATED PUMPING SITE -- Upper aquifer

——— MODEL-AREA BOUNDARY

Figure 22.-- Model-derived lines of equal drawdown for experiment 2, (A) upper aquifer and



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

EXPLANATION

—20— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

•B4 SIMULATED PUMPING SITE -- Lower aquifer

—— MODEL-AREA BOUNDARY

(B) lower aquifer, where the total pumpage for constant-flux boundaries equaled 34.7 million gallons per day.

Table 1.--Streamflow losses for experiments 1 and 2

[Upper values are in million gallons per day; lower values (in parens) are in cubic feet per second]

Boundary	Willow Creek	Little Cedar Creek	Upper Cedar Creek	Lower Cedar Creek	Upper St. Joseph River	Lower St. Joseph River
Experiment 1: Pumping rate 38.9 Mgal/d						
Constant head	3.7 [1](5.7)	1.8 (2.8)	3.7 (5.7)	5.4 (8.3)	0.6 (.9)	2.0 (3.1)
Experiment 2: Pumping rate 34.7 Mgal/d						
Constant flux	4.3 [1](6.6)	3.2 (4.9)	6.5 (10.0)	6.1 (9.5)	1.0 (1.6)	2.5 (3.9)

[1]Predicted amount of streamflow loss exceeds available flow in Willow Creek.

Comparison of drawdown maps (figs. 20 and 22) for each aquifer in experiment 1 with those in experiment 2 shows a 5-ft difference in drawdown for both the upper and lower aquifers. The drawdown in the constant-flux simulation (experiment 2) is greater than that in the constant-head simulation (experiment 1) because less water moves across the model boundaries in experiment 2. Water-level elevations and pumping rates for the individual pumping areas are presented in table 2.

Comparison of the streamflow losses shown in table 1 for experiments 1 and 2 shows that the rate of water loss in each reach was at least 0.7 ft^3/s more in the constant-flux simulation than in the constant-head simulation. The largest differences in streamflow loss are along reaches nearest the heaviest concentration of pumpage. Because the only source of water for the constant-flux simulation is the streams, the amount of streamflow loss predicted would be the maximum value for the same pumping and distribution in the aquifer system. The streamflow loss from the supply at Fort Wayne due to this pumping distribution would be between 17 and 24 Mgal/d (27 and 37 ft^3/s) or a maximum of about 70 percent of the pumpage.

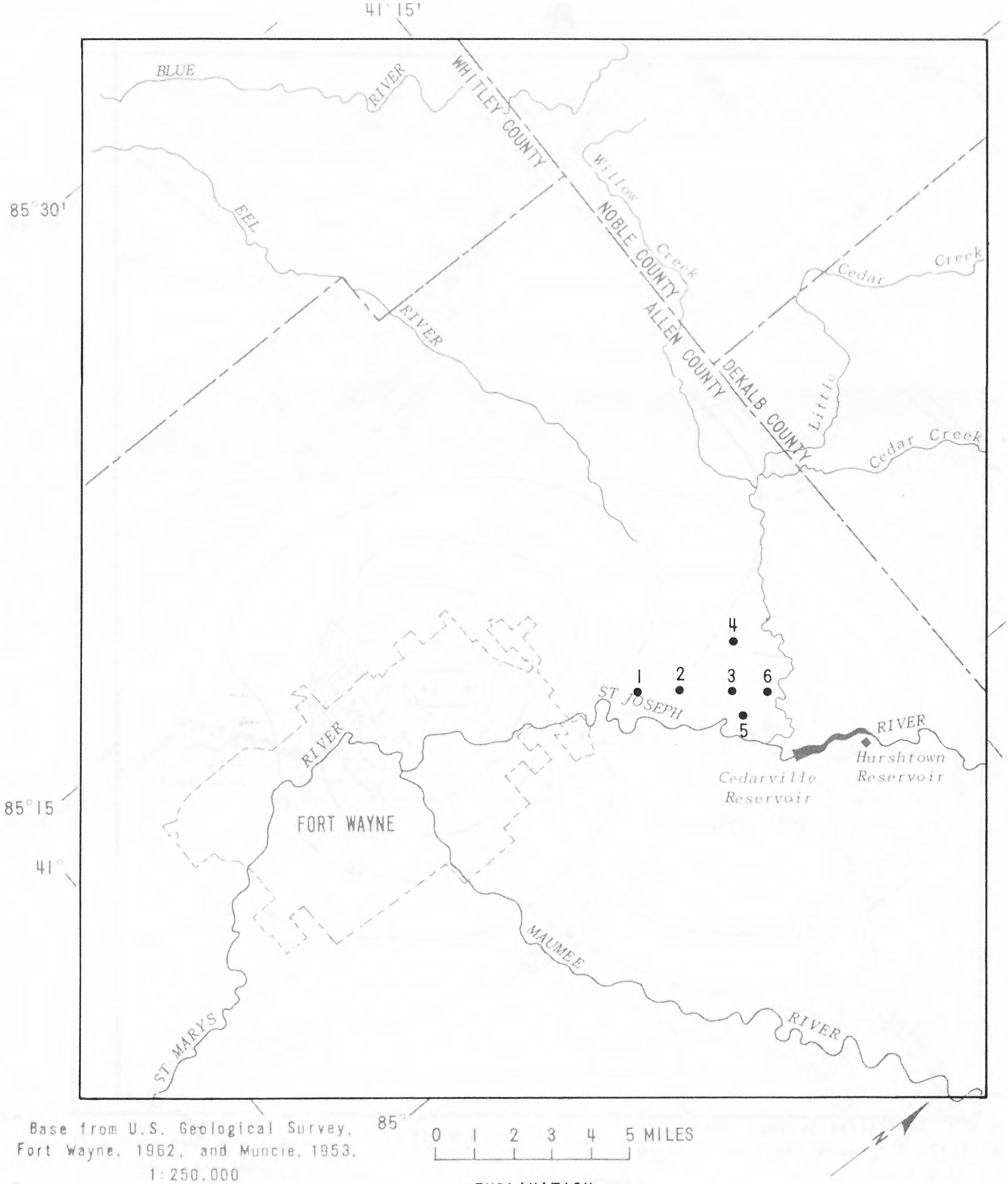
Table 2.--Water-level elevations and pumping rates of simulated pumping sites for experiments 1 and 2

	Pumping sites									
	A1	A2	A3	A4	A5	A6	B1	B2	B3	B4
Experiment 1. Pumping rate 38.9 Mgal/d for constant-head boundaries										
Head (feet above National Geodetic Vertical Datum of 1929)	788	790	791	807	808	785	769	792	791	786
Pumping rate (Mgal/d)	2.7	3.3	4.7	1.9	5.8	5.8	3.3	2.3	3.3	5.8
Experiment 2. Pumping rate 34.7 Mgal/d for constant-flux boundaries										
Head (feet above National Geodetic Vertical Datum of 1929)	786	789	789	801	804	782	771	790	788	783
Pumping rate (Mgal/d)	1.7	1.9	3.7	3.9	5.8	5.8	1.7	1.6	2.8	5.8

A production of 35 Mgal/d should be obtained for the distribution of pumping sites in experiments 1 and 2. The constant-head simulation predicted that approximately 55 percent of the pumpage would be contributed from sources other than the St. Joseph River and that 40 percent would be contributed by flow induced across the model boundaries. The constant heads can provide an unlimited amount of water, whereas the constant-flux boundaries were set so that proposed pumping would provide no additional water into the model. Therefore, 40 percent should represent the maximum value for flow across the model boundaries.

The pumping distribution used in experiments 1 and 2 tested areas of highest transmissivity in the upper and lower aquifers. In those experiments, the pumping sites were placed to minimize the diversion from the St. Joseph River, Fort Wayne's water supply. Because the streamflow losses were great in experiments 1 and 2, two additional model experiments (3 and 4) were run to determine the effects of pumping from areas of high transmissivity along the St. Joseph River; the pumping sites (1 to 6) are plotted in figure 23. Constant-flux boundaries were used in experiment 3, and constant-head boundaries were used in experiment 4. As in experiments 1 and 2, production rates were limited by drawing the head in the blocks representing the pumping sites down to the top of the aquifer. Because the pumping sites were along streams, little water was diverted from other sources. Thus, the pumping rate used in experiment 3 (29.9 Mgal/d) did not require adjustment when the boundary conditions were changed for experiment 4. Drawdowns in experiments 3 and 4 are shown in figures 24 and 25, respectively. Water-level elevations and rates at which pumping was simulated at each of the six pumping sites in experiments 3 and 4 are shown in table 3. Streamflow losses for both experiments (table 4) were greatest in lower Cedar Creek, but none of the reaches would go dry at the pumping rates used. The amount of streamflow loss equaled 70 percent of the pumpage for the constant-head boundaries (experiment 4) and 83 percent for the constant-flux boundaries (experiment 3).

In evaluating the experiments, the author concluded that the differences between the two boundary conditions are within acceptable limits and that the distances from the boundaries to the pumping centers do not unduly influence the calculated drawdowns. There was some drawdown at the boundaries in the constant-flux simulations where ideally none would be preferable. Any conclusions drawn from the pumping distributions should probably be based on experiments 2 and 3, in which constant-flux boundaries were used.

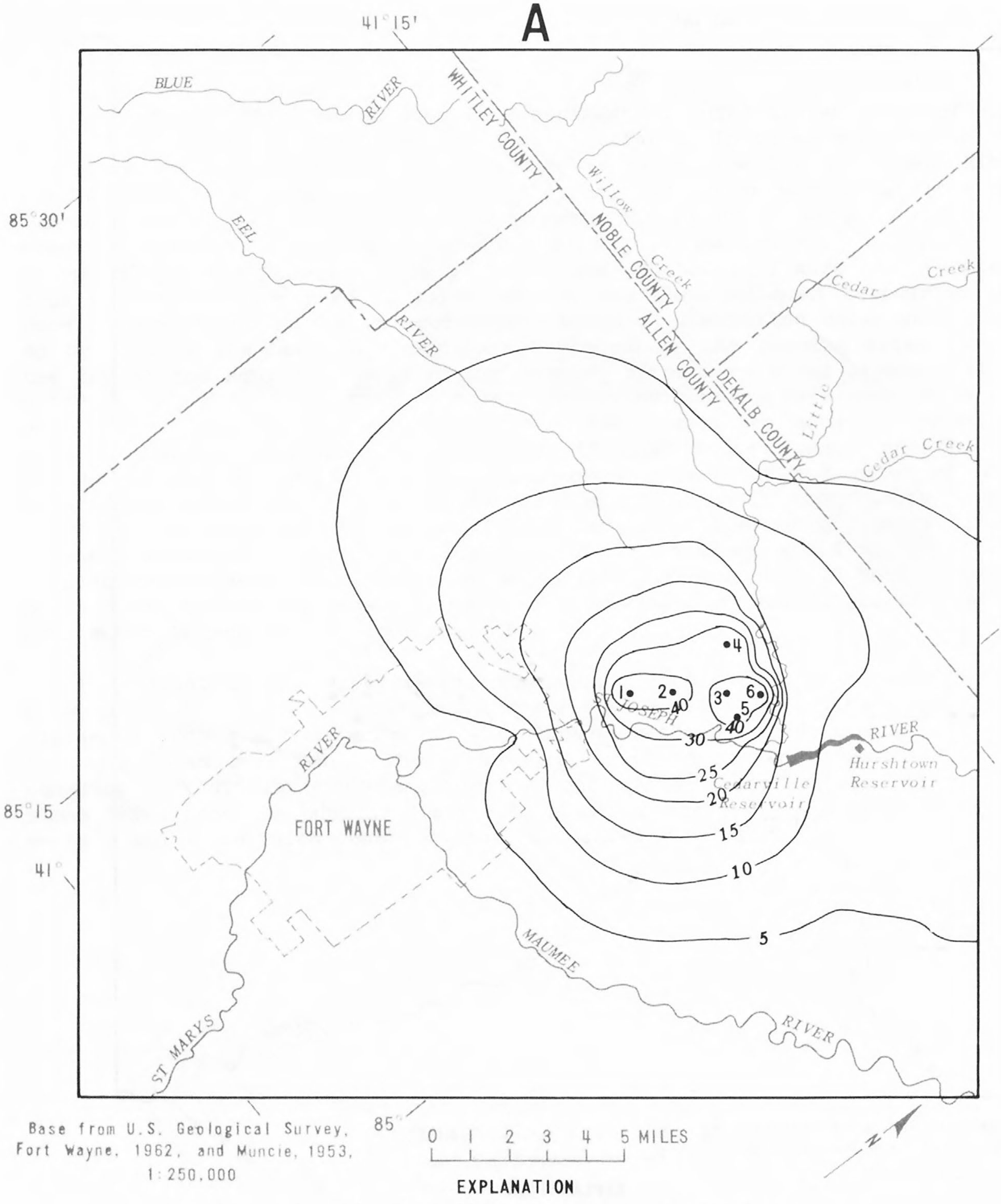


EXPLANATION

2 SIMULATED PUMPING SITE -- Upper aquifer

MODEL-AREA BOUNDARY

Figure 23.-- Locations of simulated pumping sites for model experiments 3 and 4.



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

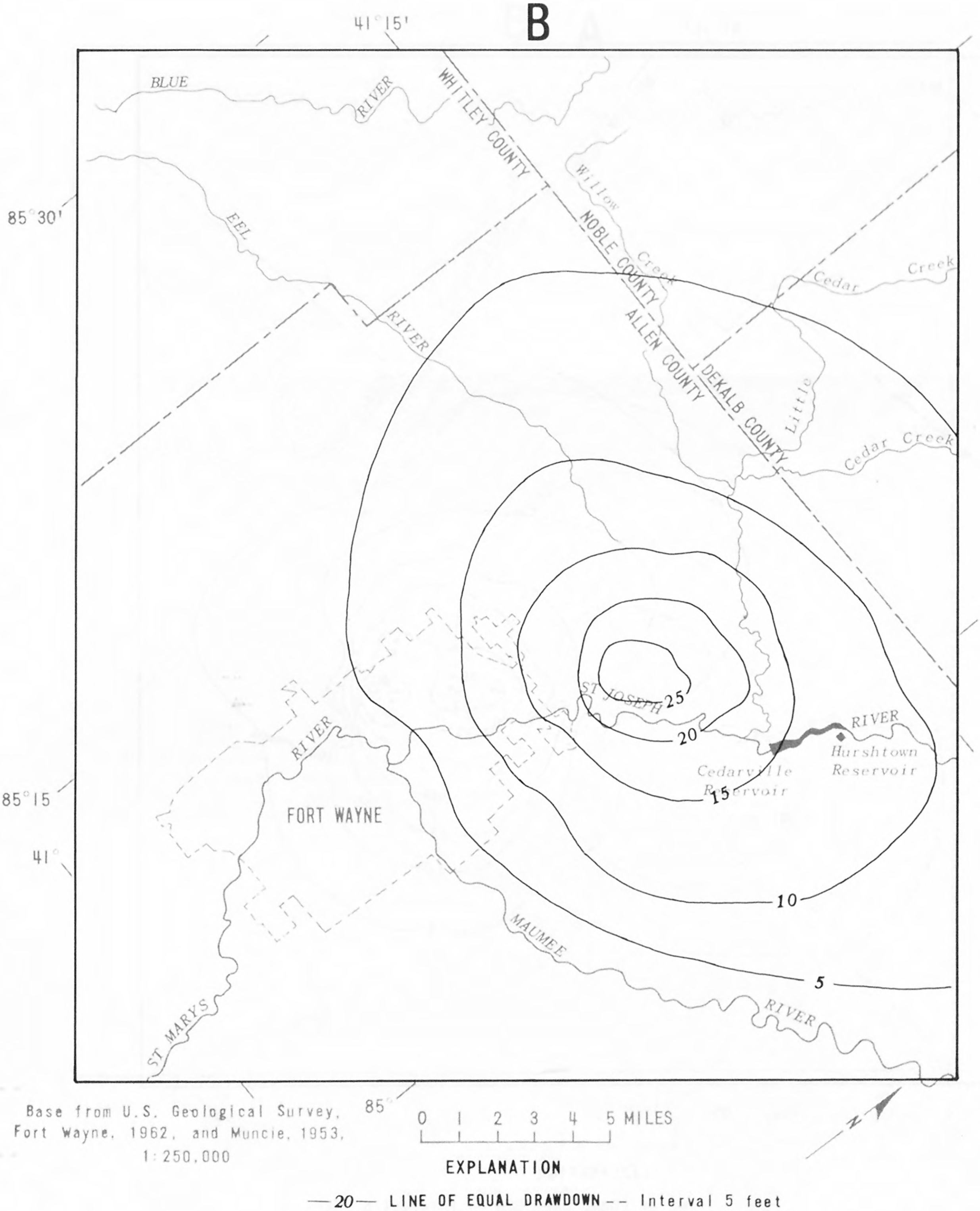
EXPLANATION

—*20*— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

2• SIMULATED PUMPING SITE -- Upper aquifer

——— MODEL-AREA BOUNDARY

Figure 24.-- Model-derived lines of equal drawdown for experiment 3, (A) upper aquifer and



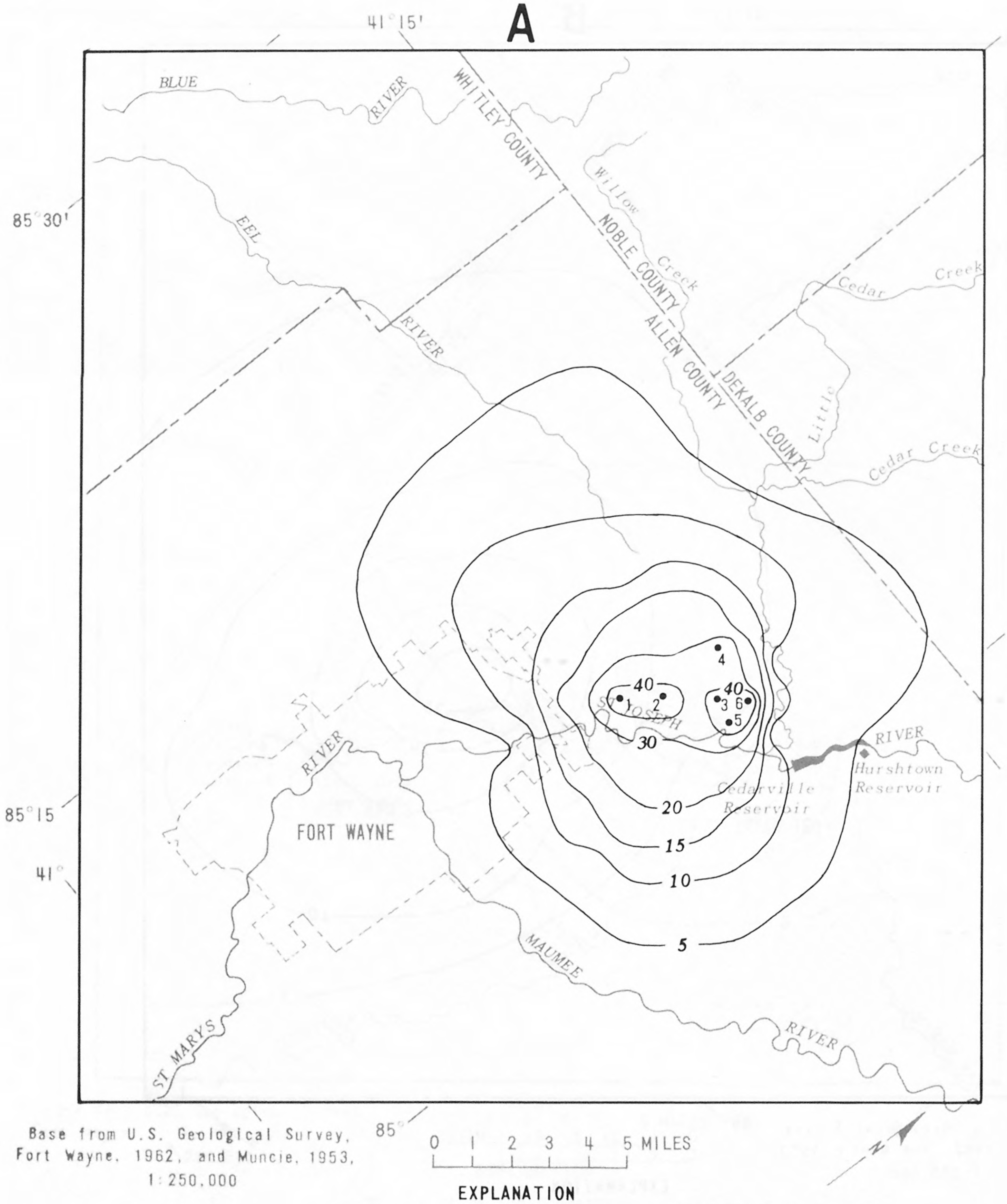
Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

EXPLANATION

—20— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

——— MODEL-AREA BOUNDARY

(B) lower aquifer, where the total pumpage for constant-flux boundaries equaled 29.9 million gallons per day.



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

0 1 2 3 4 5 MILES

EXPLANATION

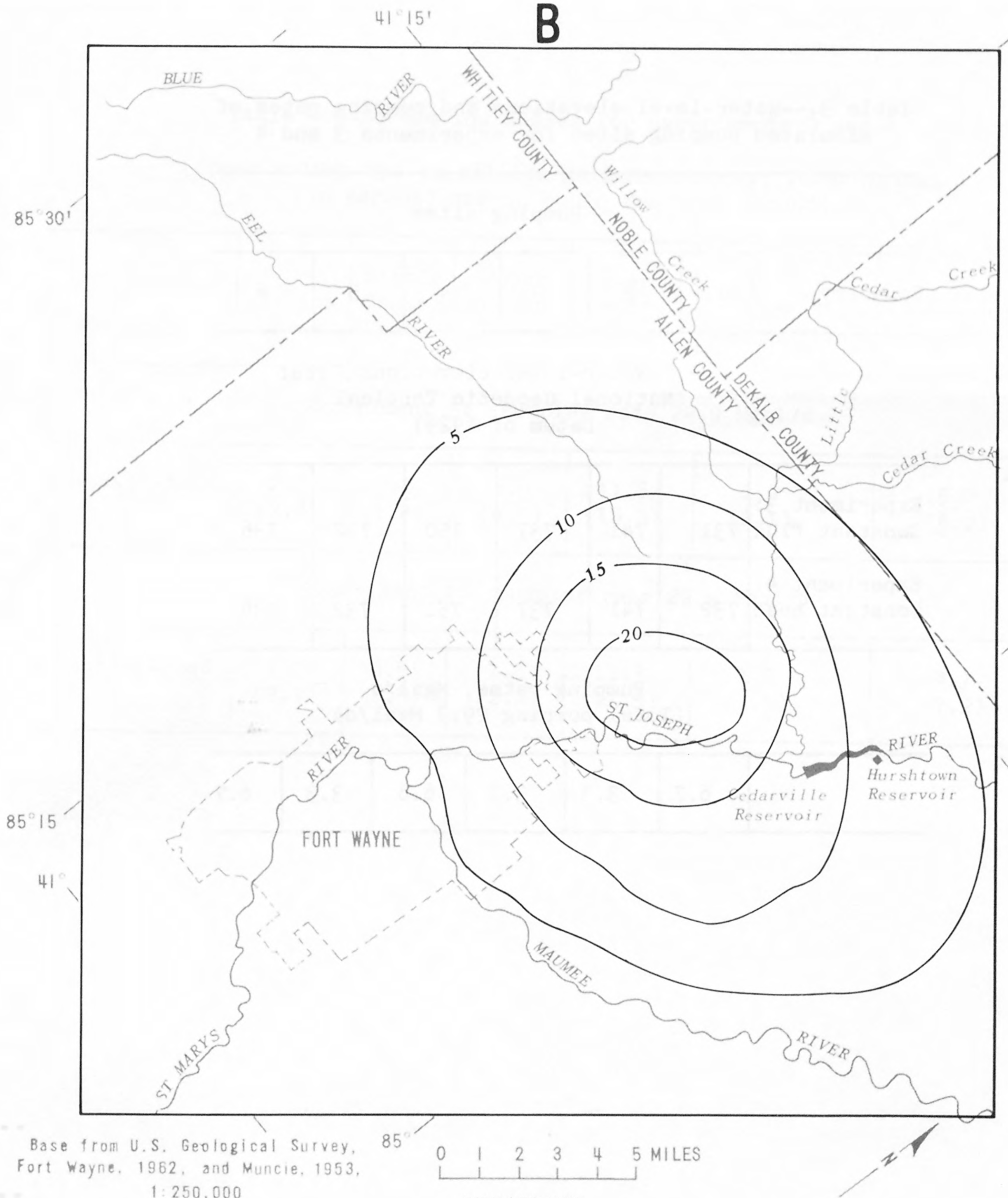
—20— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

•3 SIMULATED PUMPING SITE -- Lower aquifer

—— MODEL-AREA BOUNDARY

Figure 25.-- Model-derived lines of equal drawdown for experiment 4, (A) upper aquifer and

B



Base from U.S. Geological Survey, Fort Wayne, 1962, and Muncie, 1953, 1:250,000

EXPLANATION

—*20*— LINE OF EQUAL DRAWDOWN -- Interval 5 feet

——— MODEL-AREA BOUNDARY

(B) lower aquifer, where the total pumpage for constant-head boundaries equaled 29.9 million gallons per day.

Table 3.--Water-level elevations and pumping rates of simulated pumping sites for experiments 3 and 4

	Pumping sites					
Boundary	1	2	3	4	5	6
	Water-level elevations, feet (National Geodetic Vertical Datum of 1929)					
Experiment 3: Constant flux	731	741	737	750	732	736
Experiment 4: Constant head	732	741	737	751	732	736
	Pumping rates, Mgal/d (Total pumping 29.9 Mgal/d)					
	6.7	3.3	3.3	6.6	3.3	6.7

Table 4.--Streamflow losses for experiments 3 and 4

[Upper values are in million gallons per day; lower values (in parens) are in cubic feet per second]

Boundary	Willow Creek	Little Cedar Creek	Upper Cedar Creek	Lower Cedar Creek	Upper St. Joseph River	Lower St. Joseph River
Experiment 3: Pumping rate 29.9 Mgal/d						
Constant-flux	2.2 (3.4)	1.2 (1.8)	3.0 (4.7)	11.8 (18.3)	1.4 (2.1)	5.2 (8.0)
Experiment 4: Pumping rate 29.9 Mgal/d						
Constant-head	1.7 (2.7)	0.5 (.8)	1.6 (2.4)	11.3 (17.5)	1.0 (1.5)	4.7 (7.2)

SUMMARY AND CONCLUSIONS

In the Fort Wayne study area, glaciation has been responsible for the deposition of sediments whose thickness ranges from 10 to 250 ft. From test drilling and drillers' logs, three aquifers have been defined in the study area. At mid-depth in the glacial drift is a continuous layer of sand and gravel. At the base of the drift in the south and east parts of the study area is a less continuous, generally thinner layer of sand and gravel. To the north and west of Fort Wayne, in Allen County, the lower layer thickens and merges with the upper sand and gravel layer to produce an aquifer more than 100 ft thick. The third aquifer is the limestone that underlies the glacial drift throughout much of the study area.

Most ground water in the study area is derived from the downward percolation of precipitation. However, because the study area does not include the entire flow system associated with the Maumee River, ground water is also moving laterally from outside into the study area. Flow originates in the highest elevations of the northwest part of the study area and moves to the Maumee River, the main drain of the system. Some of the flow is discharged to the minor streams, and some is pumped from wells.

A digital model was constructed to help define the ground-water system and to study the feasibility of using ground water as a source for Fort Wayne's water supply.

Because Fort Wayne's source of water supply is the St. Joseph River, the first simulated pumping plan attempted to minimize the pumping effect on the tributaries to this river. Ten sites for simulated pumping were chosen in areas of high transmissivity. Two boundary conditions, constant head (experiment 1) and constant flux (experiment 2), were used in the simulation. The two conditions were necessary to define the limits of the possible pumpage, associated water-level declines, streamflow losses, and amount of water derived from outside the model boundaries. Simulated pumpage for the two boundary conditions were 38.9 Mgal/d for constant head and 34.7 Mgal/d for constant flux. The predicted streamflow losses for the St. Joseph River at Fort Wayne for constant head and constant flux were 17 and 25 Mgal/d (27 and 37 ft^3/s), respectively. The variation in water level on the regional scale was about 5 ft. Therefore, the pumping distribution for experiments 1 and 2 should be able to produce water at a rate of at least 35 Mgal/d.

A second simulated pumping plan evaluated the effects of pumping close to the St. Joseph River and Cedar Creek. Simulated pumpage from six sites in the upper aquifer was about 30 Mgal/d for each of the two boundary conditions, constant head (experiment 4) and constant flux (experiment 3). The predicted streamflow losses for the St. Joseph River at Fort Wayne for constant head and constant flux were 21 and 25 Mgal/d (32 and 38 ft^3/s), respectively.

This report had two objectives: (1) to determine the feasibility of developing ground-water supplies from various parts of the study area and (2) to evaluate the hydrologic impacts of the development, such as streamflow losses and ground-water-level declines. Model results show that the ground-water system has potential for development; the amount of simulated pumpage from the sand and gravel deposits is substantial, 30 to 39 Mgal/d. However, the effect of ground-water development on the flow of streams that supply Fort Wayne (St. Joseph River and Cedar Creek) is also substantial. The average simulated percentages of water lost from the flow of the St. Joseph River due to pumping were 56 percent (average for experiments 1 and 2) and 77 percent (average for experiments 3 and 4). Any large amount of pumpage from the sand and gravel deposits will decrease streamflow.

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