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EFFECT OF IRRIGATION PUMPING ON THE GROUND-WATER SYSTEM IN NEWTON AND JASPER COUNTIES, INDIANA

U.S. GEOLOGICAL SURVEY
WATER-RESOURCES INVESTIGATIONS 81-38

PREPARED IN COOPERATION WITH THE
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16. Abstract (Limit: 200 words)

Flow in the ground-water system in Newton and Jasper Counties, Indiana, was simulated in a quasi-three-dimensional model in a study of irrigation use of ground water in the two counties. The ground-water system consists of three aquifers: (1) a surficial coarse sand outwash known as the Kankakee aquifer, (2) a limestone and dolomite bedrock aquifer, and (3) a sand and gravel bedrock valley aquifer. Irrigation pumpage, primarily from the bedrock aquifer, was estimated to be 34.8 million gallons per day during peak irrigation in 1977. Acreage irrigated with ground water is estimated to be 6,200 acres. A series of model experiments were used to estimate the effect of increased irrigation on ground-water levels and streamflow. Model analysis indicates that major factors controlling drawdown caused by pumping in the bedrock aquifer are the variations in thickness and in vertical hydraulic conductivity in a semiconfining unit overlying the bedrock. Streamflow was not significantly reduced by hypothetical withdrawals of 12.6 million gallons per day from the bedrock aquifer and 10.3 million gallons per day in the Kankakee aquifer. Simulation of water-level recovery after pumping indicated that a 5-year period of alternating between increasing pumping and recovery after shutdown will cause neither large residual drawdown nor mining of the ground-water system.

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August 1981

UNITED STATES DEPARTMENT OF THE INTERIOR

JAMES G WATT, Secretary

GEOLOGICAL SURVEY

Doyle G. Frederick, Acting Director

For additional information write to:

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

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

Multiply inch-pound unit	By	To obtain SI unit
Length		
inch (in.)	25.40	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
Area		
square mile (mi^2)	2.590	square kilometer (km^2)
acre	0.4047	square hectometer (hm^2)
Volume		
million gallons (Mgal)	3,785	cubic meter (m^3)
Flow		
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
Hydraulic units		
transmissivity, square foot per day (ft^2/d)	0.0929	square meter per day (m^2/d)
hydraulic conductivity, foot per day (ft/d)	0.3048	meter per day (m/d)

DATUM

The altitude datum used in this report is the 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."

EFFECT OF IRRIGATION PUMPING ON THE GROUND-WATER SYSTEM IN

NEWTON AND JASPER COUNTIES, INDIANA

By Marcel P. Bergeron

ABSTRACT

The geometry and the hydraulic characteristics of three major aquifers were defined in a study of irrigation use of ground water in Newton and Jasper Counties, Indiana. A surficial coarse-sand outwash, known as the Kankakee aquifer, is extensive throughout the north half of the study area. Its saturated thickness averages 40 feet and generally ranges from 5 to 60 feet. The aquifer transmissivity averages 10,000 square feet per day and ranges from 250 to 15,000 square feet per day. The uppermost 100 feet of Silurian and Devonian limestone and dolomite constitutes a bedrock aquifer. Its transmissivity generally ranges from 1,000 to 2,000 square feet per day but can range from 10 to 13,500 square feet per day. A confined outwash aquifer of coarse sand and gravel partly fills a buried bedrock valley in southern Newton and Jasper Counties. Its thickness averages 25 feet and generally ranges from 5 to 70 feet. Transmissivity of the aquifer averages 9,400 square feet per day and generally ranges from 400 to 26,000 square feet per day.

The Kankakee aquifer is separated from the bedrock and the bedrock valley outwash aquifers by a fairly continuous semiconfining layer composed primarily of till and lacustrine clay. The vertical hydraulic conductivity of this layer ranges from 0.0006 foot per day in northern Newton County to 0.004 foot per day throughout the rest of the study area.

Irrigation pumping, primarily from the bedrock aquifer, is widespread in the north half of the counties. The area irrigated with ground water is estimated to be 6,200 acres. During the study period (1976-79), pumpage was heavy only in 1977, when, during peak irrigation, it was estimated to be 34.8 million gallons per day.

A quasi-three-dimensional digital model, consisting of two aquifer layers, was used to simulate flow in the aquifer system. The model was calibrated at a steady-state condition in which model-simulated water levels and discharge toward streams were matched with water levels and discharge measured in June 1978 before the irrigation season. These conditions were typical for this time of year.

The effects of irrigation pumping on ground-water levels and streamflow were estimated from a series of model experiments. Model simulation of pumping wells in the bedrock aquifer under various hydrologic conditions indicates that major factors controlling the amount of water-level decline are the variations in thickness and in vertical hydraulic conductivity in the semiconfining unit. Wells, each pumped at the same rate and developed in the Kankakee aquifer in areas of different saturated thickness (25 and 50 feet), produced similar water-level declines. Drawdowns based on pumping simulated in the Kankakee aquifer along one side of streams were insignificant on the opposite side because the Kankakee aquifer and streams are hydraulically well connected.

Because flow in the Kankakee River is usually high during the irrigation season, it is not significantly reduced by simulating pumping of 10.3 million gallons per day from the Kankakee aquifer and 12.6 million gallons per day from the bedrock aquifer in wells along the river. High rates of pumping from the bedrock aquifer produced large drawdowns without significantly reducing streamflow. Under steady-state conditions, pumpage of as much as 112 million gallons per day could cause a 50-percent reduction in the 99.9-percent flow-duration of streamflow (225 million gallons per day) of the Kankakee River at Shelby.

Simulation of water-level recovery after pumping indicated that a 5-year period of alternating between increasing pumping and recovery after shutdown will cause neither large residual drawdown nor mining of the ground-water system.

INTRODUCTION

Problem

Newton and Jasper Counties, in northwestern Indiana, are within the Kankakee River basin, one of the most productive agricultural areas in Indiana. During the past 40 years, irrigation has been used widely in the two counties to supplement inadequate precipitation during the growing season. Although surface water is used, a significant amount of water for irrigation is pumped from the ground-water system. Virtually all the ground water is pumped from a carbonate bedrock aquifer underlying extensive glacial drift.

Seasonal pumping from the aquifer has caused large water-level declines in some observation wells during the growing season in some years. Increasing irrigation pumping in the counties may prevent the recovery of water levels in the bedrock aquifer to prepumping levels between growing seasons.

Purpose and Scope

In 1976, the U.S. Geological Survey, in cooperation with the Indiana Department of Natural Resources, began a study to assess the effect of irrigation on the ground-water system in Newton and Jasper Counties, Ind. The main objectives were to (1) determine the geometry and hydraulic characteristics of the major aquifers and semiconfining beds, (2) define the ground-water-flow system and determine its relationship to the major streams, (3) document historical and current ground-water pumping and its effect on water levels, and (4) evaluate the short-term and long-term effects of increasing pumping on ground-water levels and streamflow.

Location and Setting

Most of the 1,260-mi^2 study area, including the 985-mi^2 model area, is in Newton and Jasper Counties (fig. 1). The area is within the Central Lowland Physiographic Province (Schneider, 1966). The topography is characteristic of the nearly flat landscape of the Kankakee outwash and lacustrine plain of the Northern lake and moraine region of Indiana. Altitude ranges from about 770 ft (feet) south of Goodland, in southeast Newton County, to 620 ft in places along the Kankakee River, in northwest Newton County. Prominent features of the landscape include numerous low-lying sand dunes and the Iroquois end moraine north of the Iroquois River in south-central Newton and Jasper Counties.

The study area is drained by a network of streams and manmade ditches, all tributary to the Kankakee and Iroquois Rivers (fig. 1). The Iroquois River flows into the Kankakee River southeast of Kankakee, Ill., west of Newton County.

The climate is temperate continental. The mean annual temperature, 51.5° F, is based on 30 yr of record at Kentland, Ind., from 1947 to 1977 (National Oceanic and Atmospheric Administration, 1977, p. 6). The monthly mean temperature ranges from 25.6° in January to 74.8° F in July. The mean annual precipitation at Kentland, Ind., averages 36.7 inches (National Oceanic and Atmospheric Administration, 1977, p. 4), and the monthly mean precipitation ranges from 1.8 in February to 4.23 inches in July.

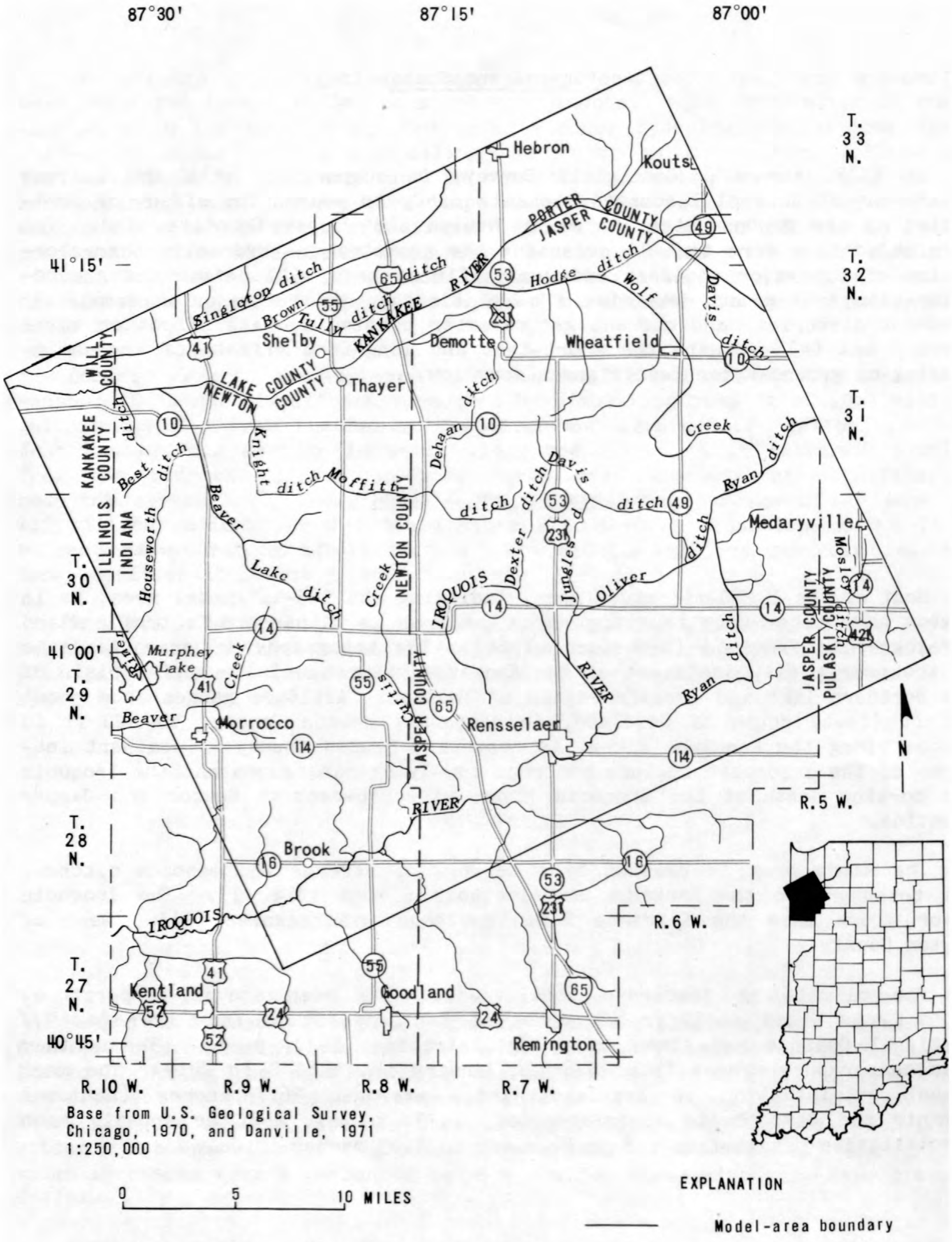


Figure 1.-- Study area, including the model area, in northwest Indiana.

Previous Investigations

Several reports have presented the broader aspects of the ground-water resources of Newton and Jasper Counties, Ind., but none of these studies contain enough quantitative information to assess the problems addressed in the current study. Preliminary investigations of the ground-water resources and the geology of Newton County (Rosenshein and Hunn, 1964b) and Jasper County (Rosenshein and Hunn, 1964a) provided general information on the geology and the major sources of ground water in addition to tabulated records of basic water-well and test-hole data. A report on a fairly detailed reconnaissance of the water and land resources of the Kankakee River basin, which includes Newton and Jasper Counties, by the State of Indiana and others (1976), presents information on general ground-water availability, ground-water flow, bedrock elevation, and the geometry and the areal extent of the surficial unconfined outwash aquifer.

The general nature of statewide irrigation and the potential and the predictions of future irrigation development in Indiana have been summarized in reports by Uhl and Kingsbury (1957), Kemp and others (1967), Kemp (1970), and the Governor's Water Resource Study Commission, State of Indiana (1980, p. 161, 479, 482, and 489). These studies discuss the general nature of irrigation throughout Indiana but provide little detailed information of current acreage irrigated with ground water in Newton and Jasper Counties.

Methods of Investigation

Hydrologic data were collected to define the ground-water-flow system. Drillers' logs were used in mapping the areal extent and the thickness of major aquifers and semiconfining beds. Additional lithologic data were obtained from 138 test holes drilled by the Geological Survey in areas where data were sparse. A 2-in. diameter observation well was installed in each of 119 test holes. Water levels were measured in these observation wells and some domestic and irrigation wells at various times during the study. These data were used to determine the head distributions of major aquifers and the low-flow head difference between the streams and the aquifers. Flow in selected reaches of the major streams was measured during a representative base-flow period to estimate ground-water seepage to the streams.

Historical and current irrigation pumpage were determined. Digital recorders in several observation wells near major pumping centers were used to document seasonal drawdowns.

A digital flow model was used to investigate the short-term and long-term effects of irrigaton pumping on the hydrologic system.

GEOLOGIC SETTING

Bedrock

The study area is underlain by dolomitic limestone, dolomite, shale, sandstone, and minor amounts of siltstone ranging in age from Cambrian to Lower Pennsylvanian. In general, Cambrian and Ordovician rocks lie at depths exceeding 1,300 and 650 ft, respectively. Dolomitic limestone and dolomite of Middle Silurian and Devonian age that subcrop beneath the glacial drift in most of the study area (fig. 2) are major aquifers. Devonian shale and interbedded siltstone and Mississippian shale that subcrop beneath the drift in the south parts of the counties and in northeast Jasper County are minor aquifers. Rocks of Ordovician through Pennsylvanian age are exposed in and near a limestone quarry just east of Kentland.

The axis of the Kankakee Arch (fig. 2) is a major structural feature in the bedrock that extends northwest to southeast. The bedrock dips southwest and northeast away from the arch. A map of the altitude of the upper surface of the bedrock, prepared from drillers' logs and test-drilling information, is shown in figure 3. The map indicates that, because of local bedrock topography, the altitude of the bedrock slopes northwest. The bedrock surface exhibits little relief except in an area along an ancestral bedrock valley, known as the Rensselaer buried valley, in the south parts of Newton and Jasper Counties. This valley, which is younger than the Kankakee Arch, cuts across and disrupts the continuity of the arch north of Rensselaer. Information from drillers' logs and test drilling indicates that the width of the valley generally ranges from 1 to 3 mi (mile) and roughly parallels the Iroquois River. Maximum relief along the valley is 110 ft; bedrock altitudes range from 493 to 603 ft (fig. 3).

Unconsolidated Sediments

A surficial geologic map (fig. 4) indicates successive periods of glaciation during the Pleistocene Epoch. Consequently, the bedrock surface is mantled by drift, which generally ranges in thickness from 5 to 170 ft.

The surficial deposits throughout the north half of the study area are predominantly composed of sand and gravel associated with a broad outwash plain called the Kankakee outwash and lacustrine plain. Regionally, this outwash plain forms an extensive sand and gravel deposit that ranges in width from 15 to 25 mi and extends from South Bend to the Illinois State line. Within the study area, this deposit generally ranges in thickness from 5 to 70 ft.

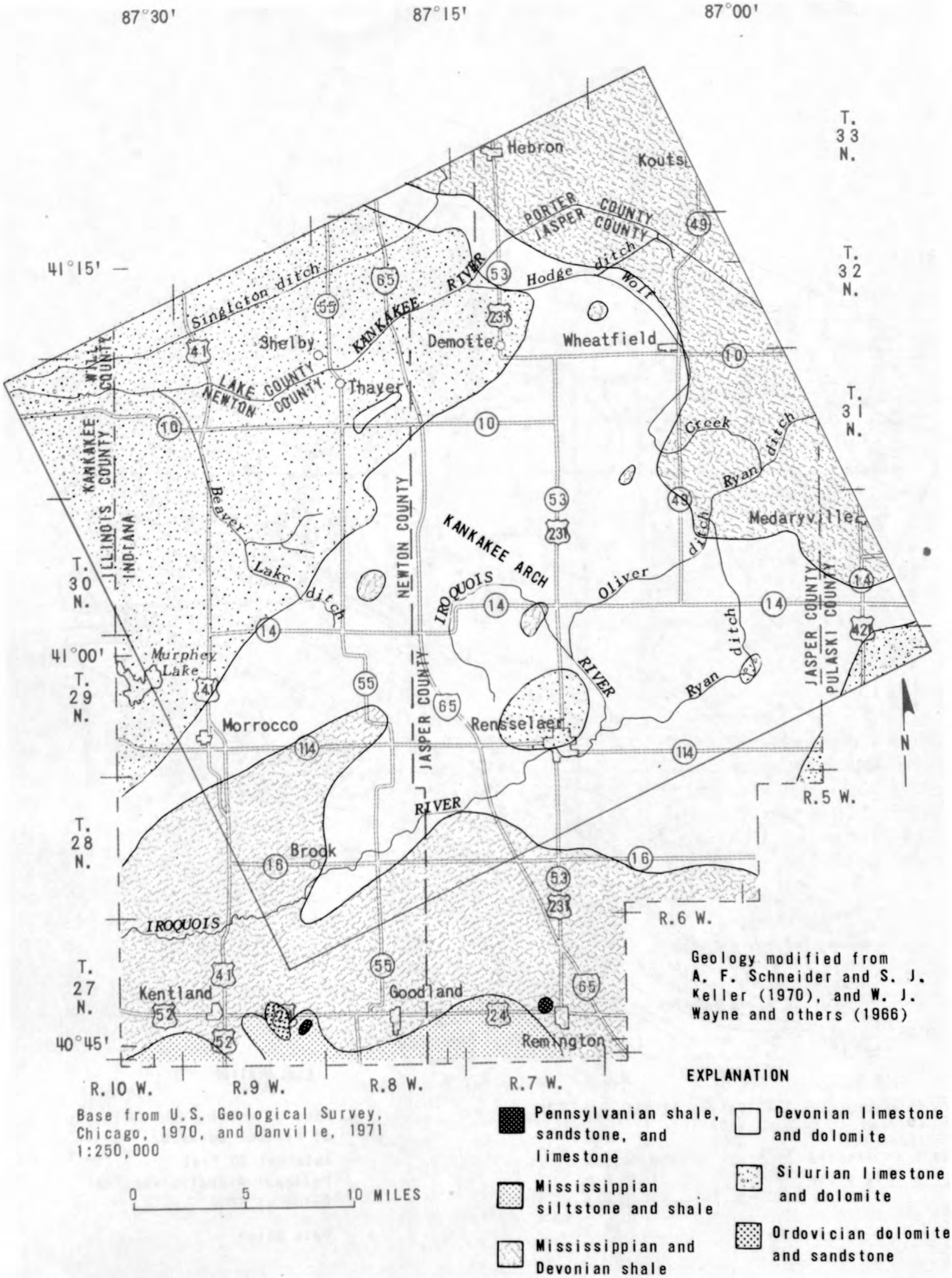


Figure 2.-- Generalized bedrock geology.

Figure 3.-- Altitude of upper surface of the bedrock.

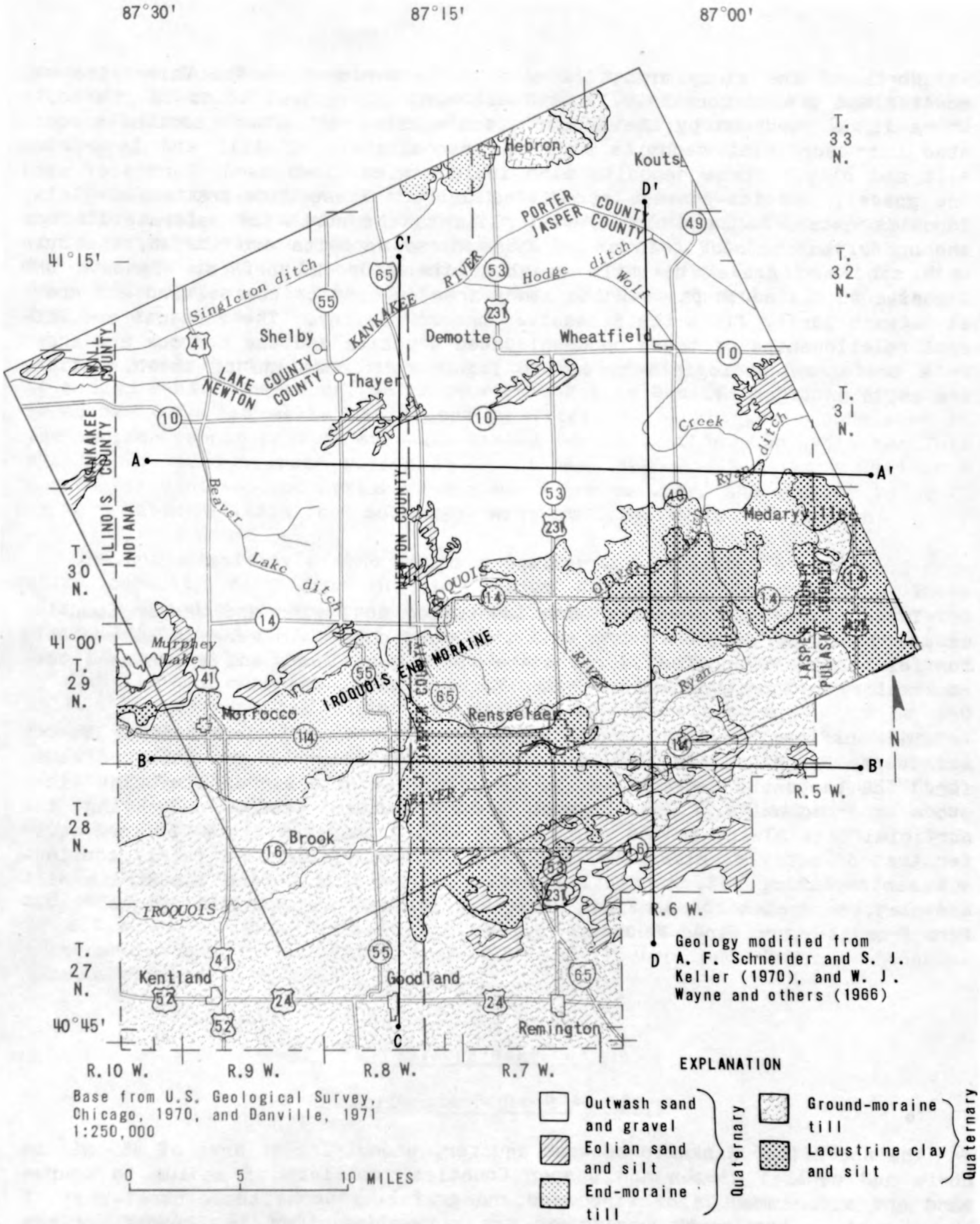


Figure 4.-- Generalized surficial geology.

North of the study area, the outwash is bordered by the Valparaiso end moraine and ground moraine. The outwash sand and gravel thins to the south where it is bordered by the Iroquois end moraine and ground moraine associated with surficial deposits made up predominately of till and lacustrine silt and clay. These deposits also include wind-blown sand, lenses of sand and gravel, and ice-contact stratified drift. These fine-grained surficial deposits extend beneath the outwash plain to the north and separate it from the underlying bedrock. In areas where these deposits are absent, the outwash sand and gravel directly overlies the bedrock surface. Beneath the deposits in the south part of the study area, a body of coarse sand and gravel outwash partly fills the Rensselaer bedrock valley. The vertical and lateral relationships of these unconsolidated deposits and the bedrock are shown as a series of geologic sections in figure 5. Locations of these sections are shown in figure 4.

AQUIFER SYSTEM

Three major aquifers have been identified in Newton and Jasper Counties on the basis of drillers' logs and test-hole data. They are a water-table surficial outwash aquifer, a carbonate bedrock aquifer, and a confined bedrock valley outwash aquifer.

The surficial outwash system throughout north and north-central Newton and Jasper Counties forms a water-table aquifer known as the Kankakee aquifer. The carbonate bedrock aquifer is primarily Silurian and Devonian limestone and dolomite. The bedrock valley outwash body, underlying the surficial till plain, in the south parts of the counties, is a confined aquifer that directly overlies the carbonate bedrock aquifer. A fairly continuous semiconfining bed, consisting predominately of till and lacustrine silt and clay, separates the confined bedrock and the bedrock valley outwash aquifers from the unconfined Kankakee aquifer.

Kankakee Aquifer

The unconfined Kankakee outwash aquifer, underlying an area of 465 mi^2 in north and central Newton and Jasper Counties, consists of medium to coarse sand and minor amounts of fine sand and gravel. Across the central part of the counties, the south border of the outwash aquifer is bounded by the Iroquois end moraine and ground-moraine deposits. North of the study area in Lake County, the aquifer becomes confined where it is overlain by the Valparaiso moraine complex and the associated till plain. The till plain south of the aquifer border, in the south part of the study area, extends beneath the Kankakee aquifer to the north and separates it from the underlying bedrock aquifer. In areas in north Jasper County, where the semiconfining beds are absent, the outwash aquifer directly overlies the bedrock surface.

The Kankakee aquifer is the chief source of ground water for stock and domestic supply and a minor source for irrigation throughout the north and north-central parts of the study area. The saturated thickness of the aquifer averages about 40 ft and generally ranges from a few feet to 60 ft (fig. 6). Wells completed in the aquifer generally produce from 100 to 200 gal/min (gallon per minute).

The poorly sorted Kankakee aquifer contains a significant amount of fine to medium sand. Therefore, wells penetrating this aquifer must use fine-mesh screens. Owing to well loss caused by friction, single wells completed in the aquifer cannot provide adequate yields for irrigation. However, manifold systems of multiple sand points (5 to 7) have been and are being used as a source for center-pivot sprinkler systems in areas where the aquifer is 30 ft or more thick. Yields from such systems range from 700 to 900 gal/min.

The transmissivity of the Kankakee aquifer was estimated by applying specific capacity data from drillers' logs to methods described by Theis (1963). Because of low pump rates and short screen intervals, calculated transmissivities were assumed to reflect the screened interval of the aquifer. Consequently, the hydraulic conductivity of the aquifer was calculated by dividing the computed transmissivity value by the length of screen in the well. The average hydraulic conductivity was then determined to be 250 ft/day (foot per day). The transmissivity for the aquifer can be estimated by multiplying the calculated average hydraulic conductivity by the saturated thickness of the aquifer. For an average hydraulic conductivity of 250 ft/d, the average transmissivity is estimated to be 10,000 ft^2/d (square foot per day) and to range from 250 to 15,000 ft^2/d.

No data for determining the specific yield of the Kankakee aquifer were collected in this study. The specific yield was assumed to be 0.12, the value estimated for the Kankakee aquifer in previous studies in Lake County (Rosenshein and Hunn, 1968a, p. 21) and Porter and LaPorte Counties (Rosenshein and Hunn, 1968b, p. 8).

Carbonate Bedrock Aquifer

In areas where the outwash is thin or absent or the permeability of the glacial deposits is low, wells derive water from a bedrock aquifer composed primarily of Silurian and Devonian dolomitic limestone and dolomite. These carbonate rocks are used in most of the study area as a source of water for stock, homes, public supply, and irrigation. Wells completed in these rocks are generally less than 300 ft deep, and their yields are variable. Yields

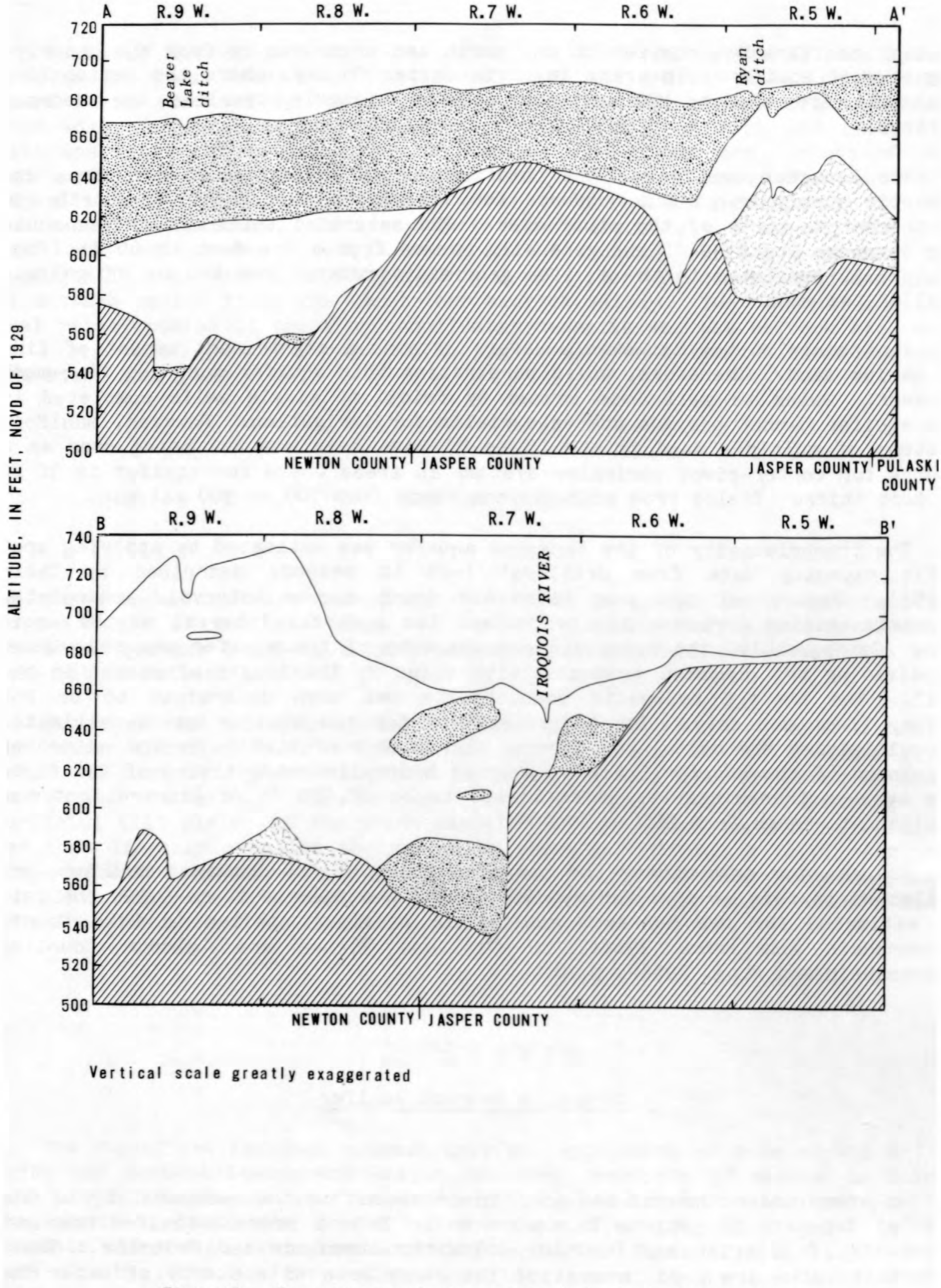
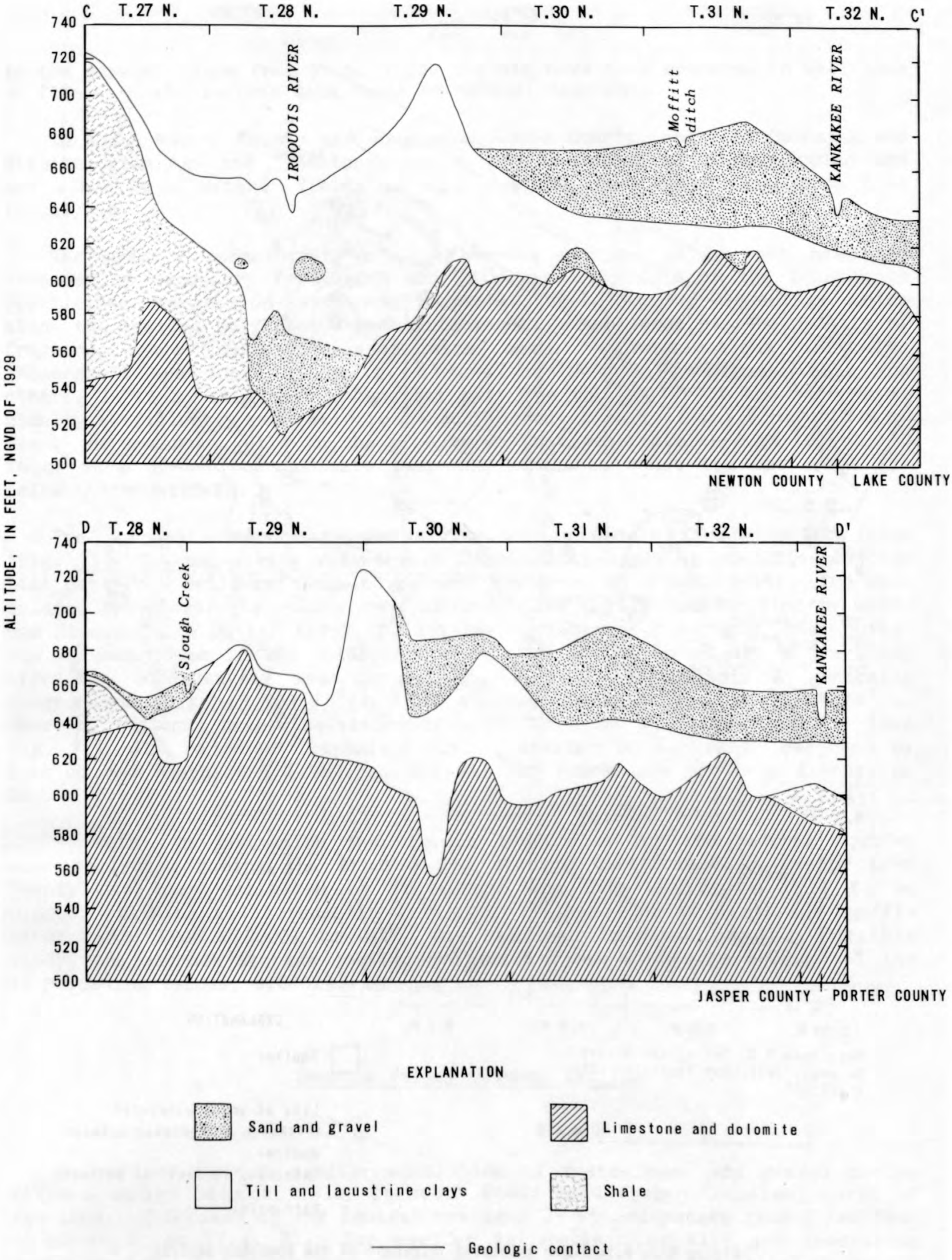
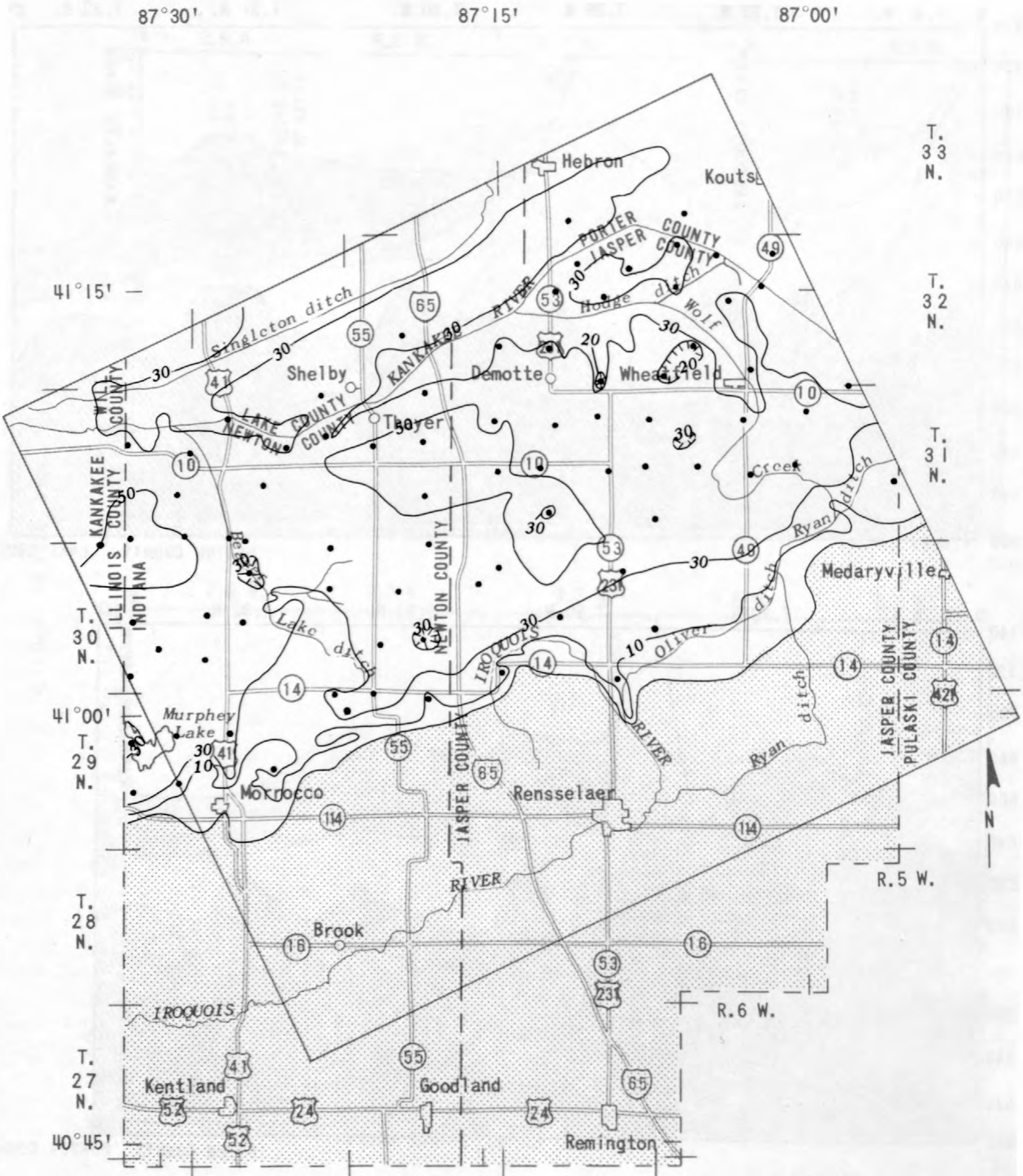


Figure 5.-- Generalized geologic sections A-A', B-B', C-C', and D-D'.



C
T.27 N.
T.28 N.
T.29 N.
T.30 N.
T.31 N.
T.32 N.
C'
IROQUOIS RIVER
Moffitt ditch
KANKAKEE RIVER
NEWTON COUNTY
LAKE COUNTY
D
T.28 N.
T.29 N.
T.30 N.
T.31 N.
T.32 N.
D'
Slough Creek
KANKAKEE RIVER
JASPER COUNTY
PORTER COUNTY
ALTITUDE, IN FEET, NGVD OF 1929
740
720
700
680
660
640
620
600
580
560
540
520
500
EXPLANATION
Sand and gravel
Limestone and dolomite
Till and lacustrine clays
Shale
Geologic contact



Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971
1:250,000

0 5 10 MILES

EXPLANATION

Aquifer

Line of equal saturated thickness of Kankakee outwash aquifer. Interval, in feet, is variable

• Data point

Figure 6.-- Extent and saturated thickness of the Kankakee aquifer.

in the general range from 10 to 2,200 gal/min have been reported in well logs on file with the Indiana Department of Natural Resources.

In south Newton County and southwest Jasper County, shale of Devonian and Mississippian age and limestone, shale, and sandstone of Mississippian age are a source of water. Yields of wells tapping these rocks range from 1 to 15 gal/min.

The primary permeability of the bedrock aquifer is low but has been enhanced by secondary fracturing and jointing. Exposure of the bedrock to pre-Pleistocene erosion has probably enlarged these secondary openings, which allow the aquifer to yield large quantities of water locally. The density of fractures and joints decreases with depth. Investigations in north (Rosenshein and Hunn, 1968a, p. 8) and north-central Indiana (Cable and others, 1971, p. C10) have shown that only the top 100 ft of the bedrock is usually sufficiently permeable to be considered as part of the aquifer. However, the numerous wells in the study area that penetrate through 100 to 150 ft or more of bedrock indicate that the permeable zones in some areas are below the top 100 ft.

The carbonate-bedrock transmissivity varies widely throughout the area (fig. 7). Transmissivity values were computed by applying specific-capacity test data from drillers' logs to methods discussed by Brown (1963). The calculated transmissivity values were corrected for partial penetration by methods discussed in Butler (1957, p. 157-162). For the correction, the aquifer was assumed to be 100 ft thick, and the ratio of the horizontal to vertical hydraulic conductivity was assumed to be 1:1. Transmissivity generally ranges from 10 to 13,500 ft^2/d. The average transmissivity is 3,475 ft^2/d. However, regional transmissivities of 1,000 to 2,000 ft^2/d are common. (See fig. 7.) The range of transmissivity is similar to the range reported by Watkins and Rosenshein (1963, p. B11-14) and Rosenshein and Hunn (1968a, p. 10).

No data for estimating the storage coefficient of the bedrock aquifer were collected during the study. The storage coefficient in nearby Lake County has been estimated by Rosenshein and Hunn (1968a, p. 11) to be 0.0008. Watkins and Rosenshein (1963, p. 11-14) reported a storage-coefficient range from 0.00001 to 0.002 for 9 values in Miami County. For this study, the bedrock storage coefficient was assumed to be the average of the 10 preceding values, 0.00013.

Bedrock Valley Outwash Aquifer

A confined outwash aquifer consisting of coarse sand and gravel partly fills a buried bedrock valley in south Newton and Jasper Counties, north of Kentland. Thickness of the aquifer averages 25 ft and ranges from a few feet to about 70 ft (fig. 8). The aquifer is overlain by till and lacustrine deposits whose thickness generally ranges from 20 to 100 ft. Altitude of the

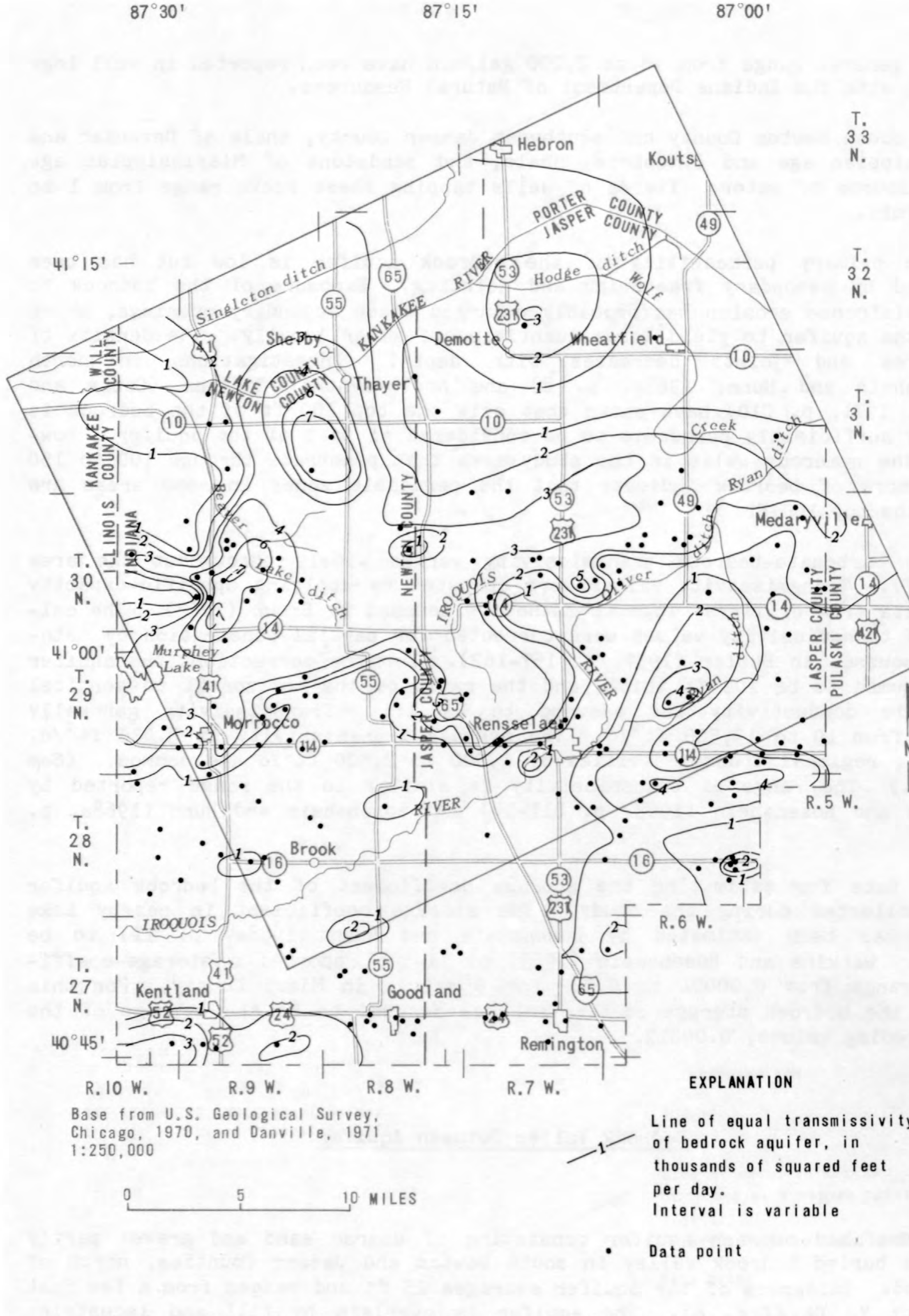


Figure 7.-- Transmissivity distribution of the bedrock aquifer.

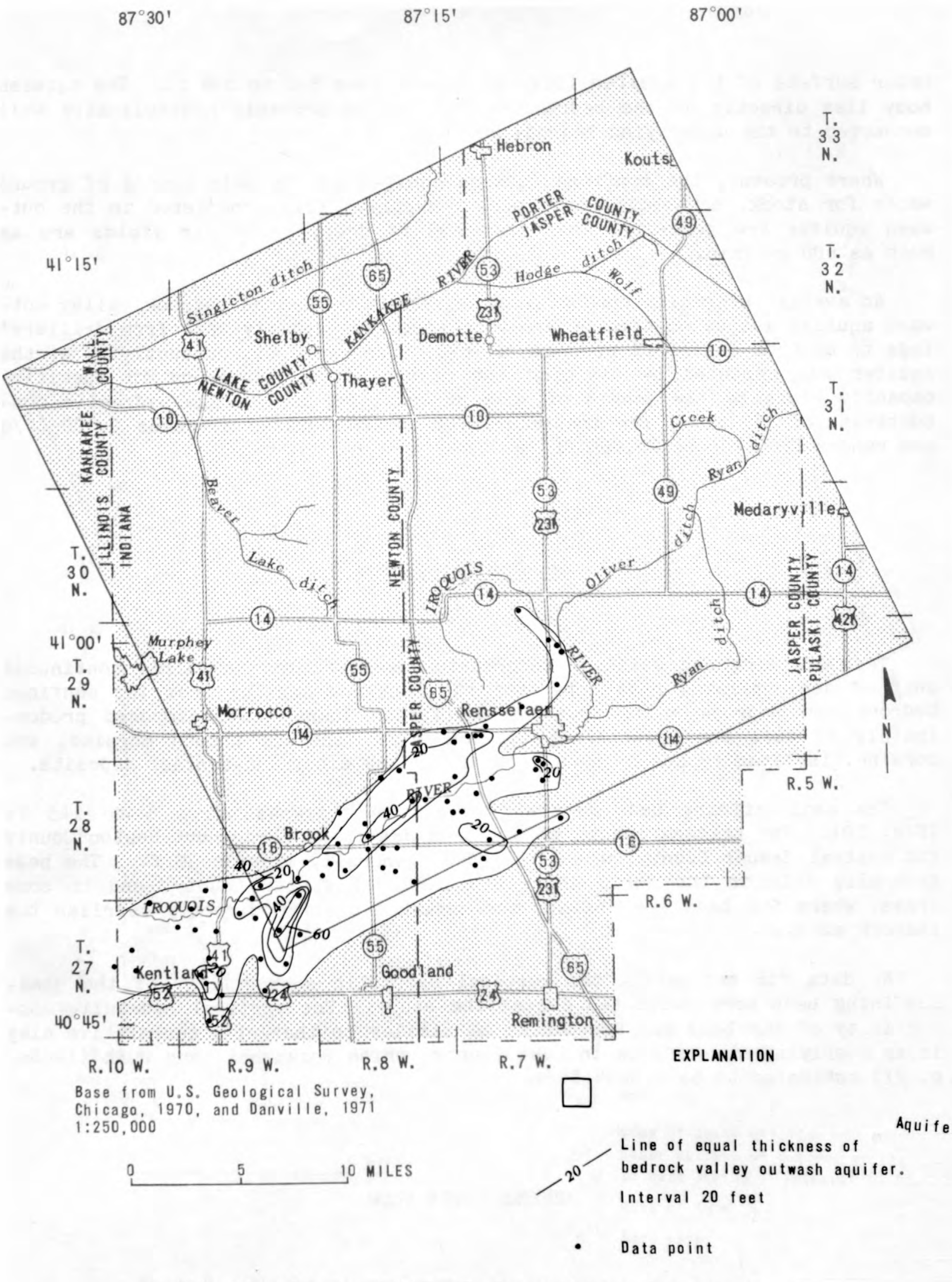


Figure 8.-- Extent and thickness of the bedrock valley outwash aquifer.

upper surface of the aquifer (fig. 9) ranges from 550 to 648 ft. The outwash body lies directly on the bedrock. Thus, it is probably hydraulically well connected to the underlying bedrock aquifer.

Where present, the confined outwash aquifer is the main source of ground water for stock, households, and public supply. Wells completed in the outwash aquifer are generally less than 135 ft deep, and their yields are as much as 600 gal/min.

An average hydraulic conductivity of 375 ft/d for the bedrock valley outwash aquifer was calculated by applying specific capacity data from drillers' logs to methods described by Brown (1963). The hydraulic conductivity of the aquifer was computed by dividing the transmissivity obtained in specific-capacity tests by the length of well screen. For an average hydraulic conductivity of 375 ft/d, the transmissivity of the aquifer averages 9,400 t^2/d and ranges from 400 to 26,000 ft^2/d.

Semiconfining Beds

Test drilling and well-log information indicate that a fairly continuous unit of semiconfining beds separates the Kankakee aquifer from the confined bedrock and bedrock valley outwash aquifers. These beds, composed predominately of till and lacustrine silt and clay, include ground moraine, end moraine, isolated lenses of sand and gravel, and minor ice-contact deposits.

The semiconfining beds generally range in thickness from 5 to 145 ft (fig. 10). The thickest parts of the beds are found throughout Newton County and central Jasper County, where thickness generally exceeds 35 ft. The beds generally thin in the north and south parts of Jasper County, and in some areas, where the beds are absent, the Kankakee aquifer directly overlies the bedrock surface.

No data for estimating the vertical hydraulic conductivity of the semiconfining beds were collected during the study. The vertical hydraulic conductivity of the beds was assumed to be similar to that of a correlative clay layer overlying the bedrock in Lake County, which Rosenshein and Hunn (1968a, p. 21) estimated to be 0.0004 ft/d.

GROUND-WATER FLOW

An observation-well network was established in Newton and Jasper Counties to monitor water levels in the three major aquifers. The water levels were periodically measured to aid in defining the ground-water flow system.

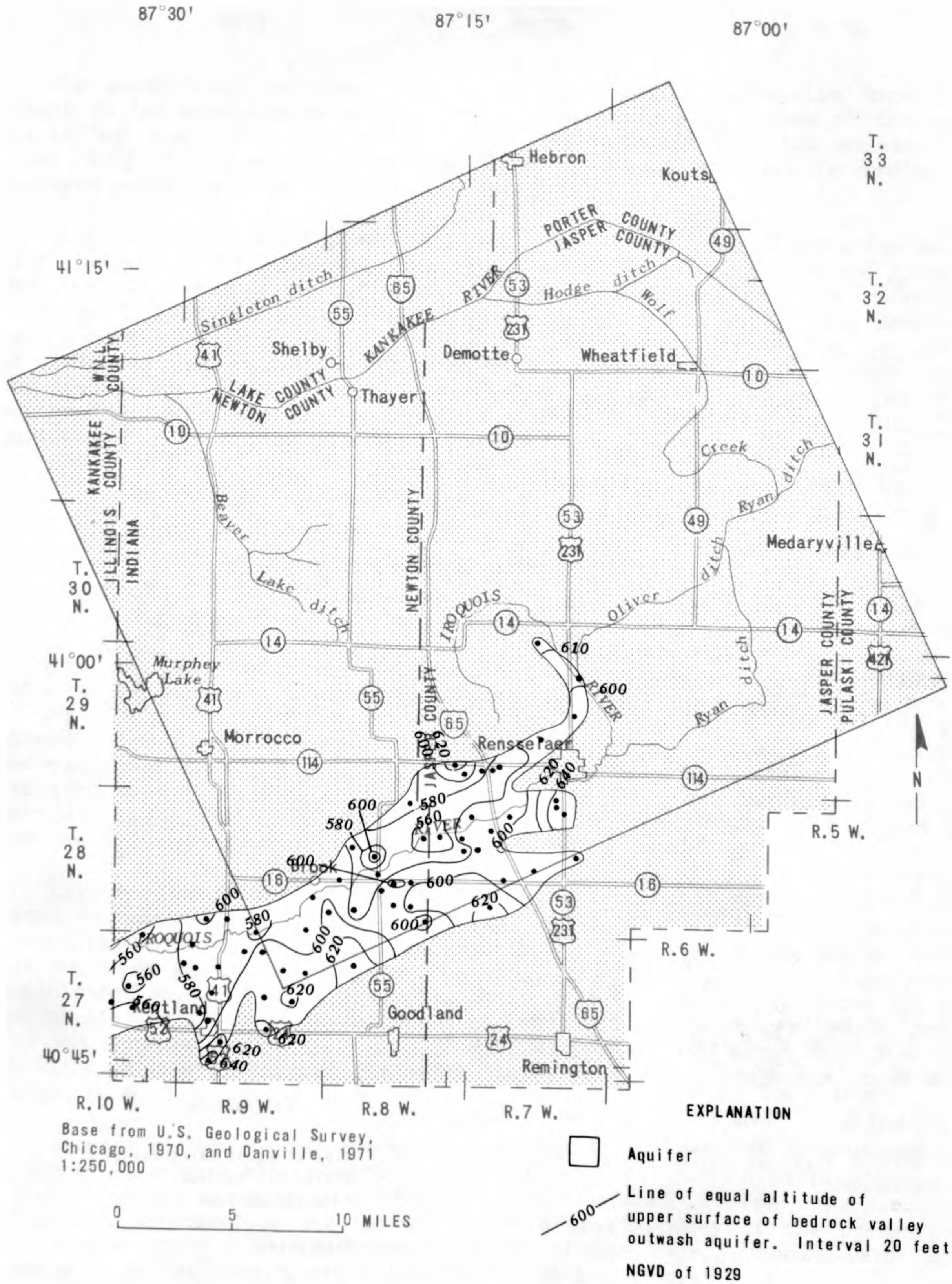


Figure 9.-- Altitude of upper surface of the bedrock valley outwash aquifer.

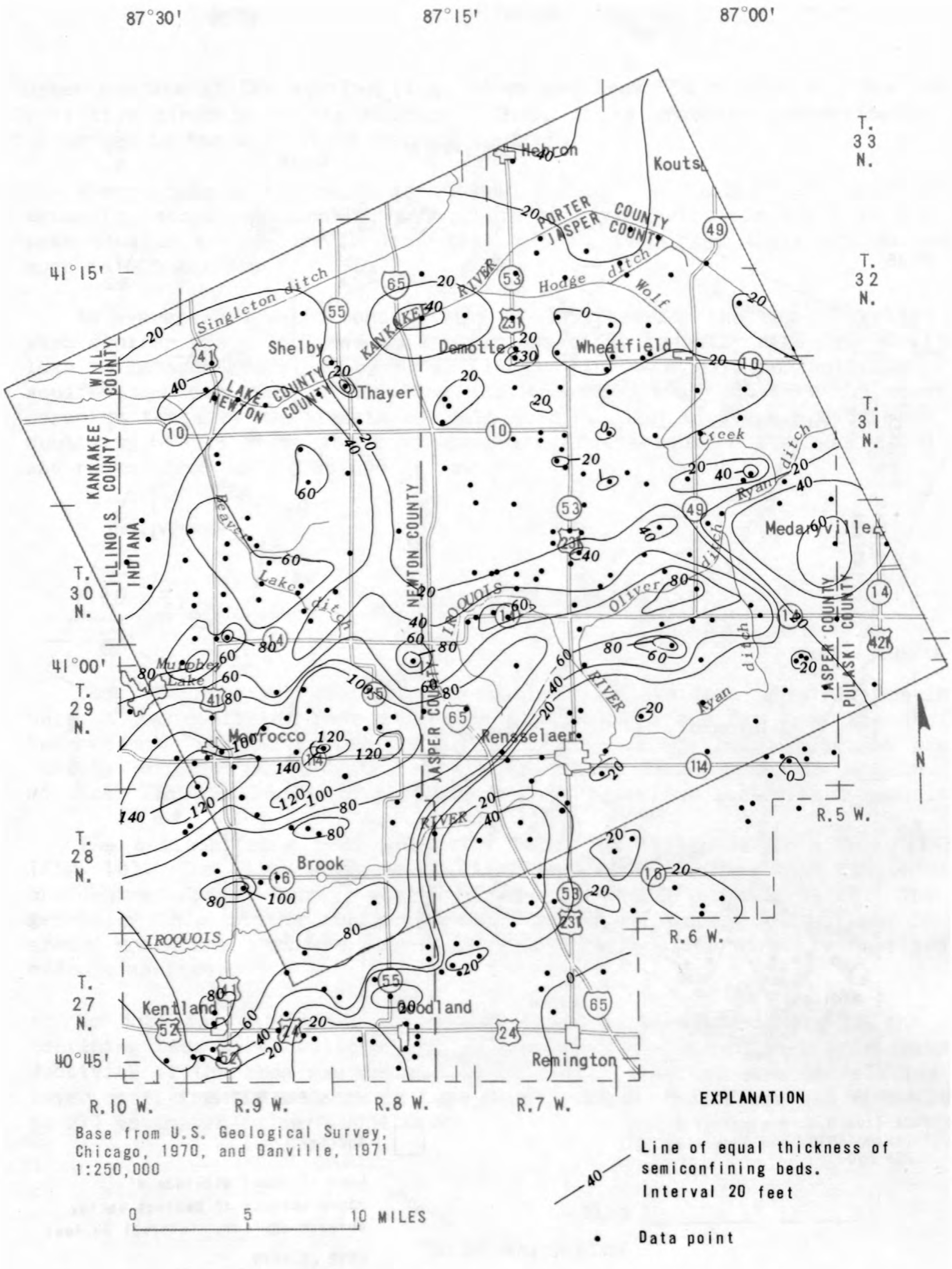


Figure 10.-- Thickness of semiconfining beds.

The generalized geologic section in figure 11 schematically shows the nature of the ground-water flow system in the study area. Some of the precipitation that falls on the ground surface infiltrates the unsaturated zone. Part of this water is removed by transpiration, and part is removed by downward percolation as recharge into the drift.

Within the saturated zone, the water flows laterally from topographically high recharge areas to lower discharge areas along major streams and rivers. The confined bedrock and bedrock valley outwash aquifers are generally recharged by downward flow from the water-table aquifer through the semiconfining beds. In the vicinity of major streams, flow in the confined aquifers is primarily vertical into the streams, where ground water is removed from the study area. The ground-water system is also replenished by lateral flow across the boundaries of the counties under hydraulic gradients within the aquifers.

CONNECTION BETWEEN AQUIFER AND STREAMS

The hydraulic connection between the Kankakee River, Iroquois River, and other major streams and the aquifers is important when evaluating the amount of water that the aquifers will yield to pumping wells. These rivers and streams are major discharge areas for water flowing through the aquifers. Heavy pumping near the streams can reduce the amount of ground water discharging to the streams, or induce infiltration from the streams through the streambeds if the hydraulic gradient in the vicinity of the stream is reversed by pumping.

In the study, the relationship between the streams and the aquifers was based on measurements on June 19 and 20, 1978, during a baseflow recession. Precipitation from the most recent storm, 1.3 in., before the measurements, was recorded at Shelby on June 7, 1978 (National Oceanic and Atmospheric Administration, 1978, p. 1). On the basis of the flow-duration curve for the gaging station on the Kankakee River at Dunn's Bridge (Horner, 1976, p. 496), the measured streamflow represented a 47.3 percent flow duration. The flow duration of the streamflow for other gaging stations within the study area ranged from 42.6 to 65 percent.

Because streamflow was measured approximately 2 weeks after a storm, the measured flow probably consisted of surficial underflow of basin storage in addition to ground-water discharge from the aquifers. However, the component of surficial underflow was assumed to be a small entity of the streamflow, and the measured flow was representative of the average ground-water discharge to the streams during a typical irrigation season.

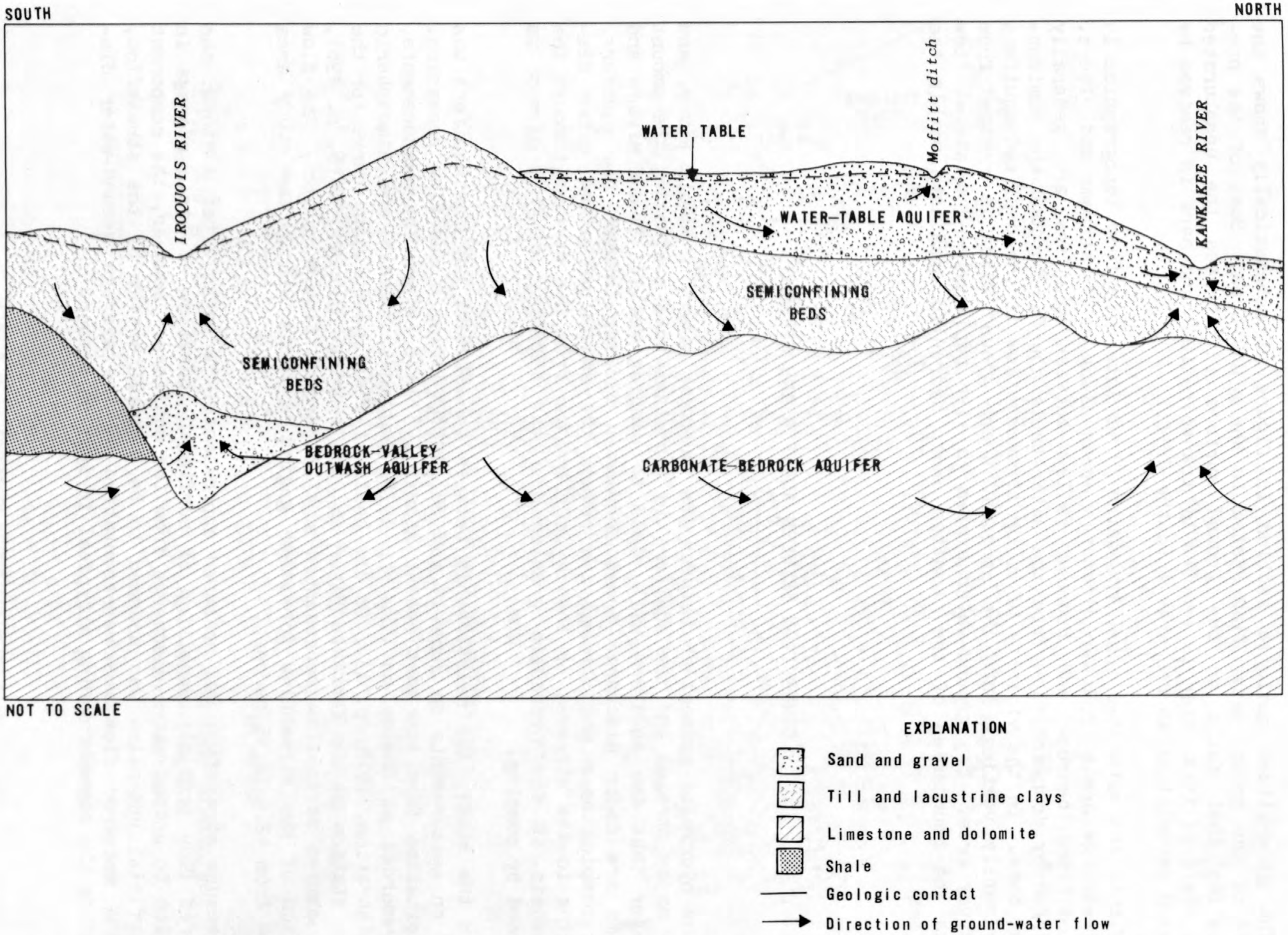


Figure 11.-- Generalized geologic section showing aquifers and direction of ground-water flow in Newton and Jasper Counties, Ind.

Calculations of the ground-water discharge to the streams are based on simultaneous measurements of streamflow at various locations within the study area. Flow was measured at upstream and downstream points on various reaches of the streams. The discharge to each reach was estimated by balancing inflow against outflow along the reach. Estimates of ground-water discharge to the various reaches of streams measured are shown in figure 12. Total discharge in the study area was estimated to be 530 ft^3/s (cubic foot per second) or 342 Mgal/d (million gallons per day).

WATER-LEVEL FLUCTUATIONS

Water levels fluctuate daily and seasonally in wells that penetrate bedrock and outwash aquifers in Newton and Jasper Counties in response to variations in recharge, discharge, and barometric pressure. Hydrographs of selected observation wells completed in the aquifers show a seasonal relationship (fig. 13). Water levels are normally highest during March and April and lowest during July and August.

Seasonal fluctuations in water levels are the result of variation in recharge to and discharge from the aquifer. Examination of rainfall data indicates that variation in recharge is not entirely a function of variation in precipitation. Monthly precipitation, fairly constant throughout the year, ranges from a mean of 1.8 in. in February to 4.23 in. in July (National Oceanic and Atmospheric Administration, 1977, p. 4). The variation in recharge is largely a function of evapotranspiration. During the growing season, transpiration by shallow-rooted crops and direct evaporation effectively reduce the amount of water from precipitation infiltrating downward through the unsaturated zone.

Depths to water throughout the study area are sufficient to minimize the amount of direct evaporation or transpiration from the water table, except from small areas along streams. Consequently, no additional water lost to evapotranspiration will be recovered by the lowering of water levels by pumping. The effect of evapotranspiration on the flow system is to reduce the recharge to the aquifer system.

Another cause of water-level fluctuations within the study area is agricultural activities. Water-level declines of more than 30 ft in some observation wells (Jasper 4, Newton 7, and Newton 8) that are near major irrigation centers have coincided with periods of peak pumping.

Examination of the hydrographs (fig. 13) shows that although water levels fluctuate, owing to seasonal stress to the flow system, the water levels return to approximately the same level every spring. The magnitude of water-level change recorded in observation wells during the study is an indication that the aquifer system is in dynamic equilibrium. The average water-level changes in 5 bedrock-aquifer wells and 62 outwash-aquifer wells in April 1977 and in April 1978 were +1.9 and +1.6 ft, respectively.

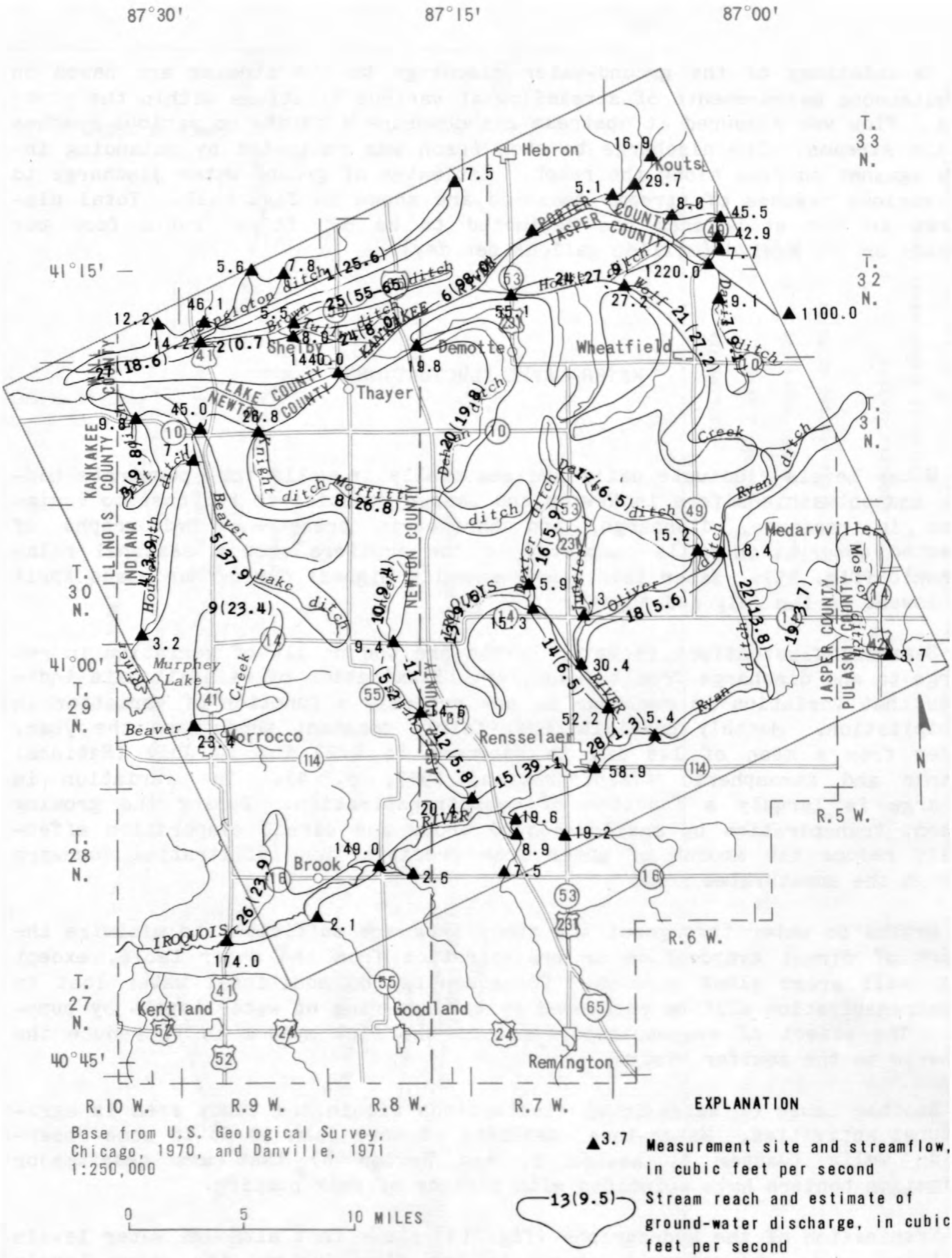


Figure 12.-- Ground-water discharge to streams on June 19-20, 1978.

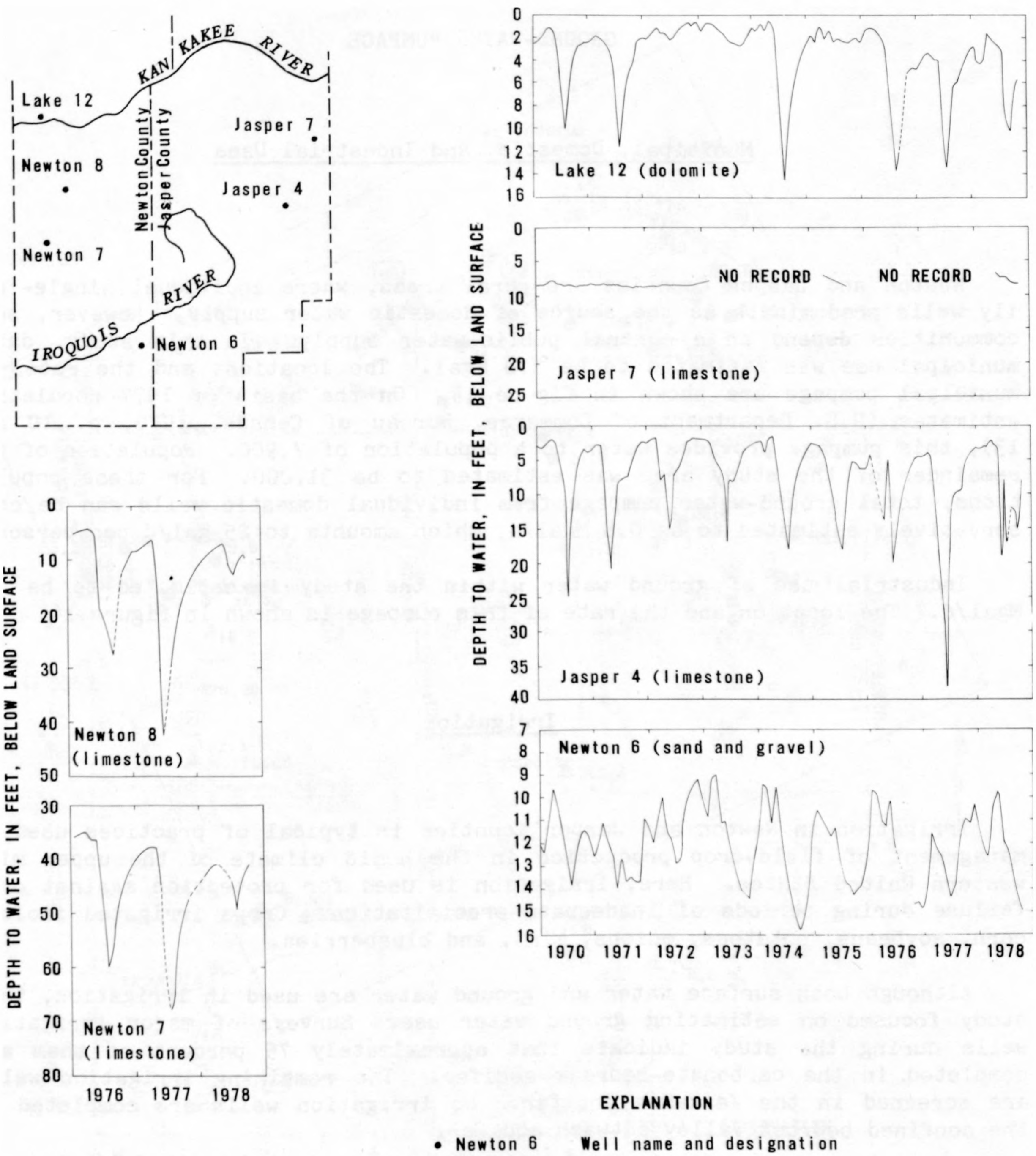


Figure 13.-- Water-level fluctuations in and locations of Geological Survey observation wells in Newton and Jasper Counties, Ind.

GROUND-WATER PUMPAGE

Municipal, Domestic, and Industrial Uses

Newton and Jasper Counties are rural areas, where individual single-family wells predominate as the source of domestic water supply. However, some communities depend on a central public-water supply. In this study, daily municipal use was estimated to be 1.0 Mgal. The locations and the rates of municipal pumpage are shown in figure 14. On the basis of 1977 population estimates (U.S. Department of Commerce, Bureau of Census, 1979, p. 10 and 13), this pumpage provides water to a population of 7,900. Population of the remainder of the study area was estimated to be 31,000. For these populations, total ground-water pumpage from individual domestic wells can be conservatively estimated to be 0.8 Mgal/d, which amounts to 25 gal/d per person.

Industrial use of ground water within the study is estimated to be 0.4 Mgal/d. The location and the rate of this pumpage is shown in figure 14.

Irrigation

Irrigation in Newton and Jasper Counties is typical of practices used in management of field-crop production in the humid climate of the upper midwestern United States. Here, irrigation is used for protection against crop failure during periods of inadequate precipitation. Crops irrigated include corn, soybeans, potatoes, onions, mint, and blueberries.

Although both surface water and ground water are used in irrigation, this study focused on estimating ground water use. Surveys of major irrigation wells during the study indicate that approximately 75 percent of them are completed in the carbonate-bedrock aquifer. The remaining irrigation wells are screened in the Kankakee aquifer. No irrigation wells are completed in the confined bedrock valley outwash aquifer.

The distribution of irrigation, shown in figure 14, indicates that pumping is restricted to the north half of the study area. Farming in this area requires irrigation because the soils have developed on coarse outwash deposits and windblown sands that are easily drained. During periods of inadequate precipitation and high temperature, water levels in these deposits are readily lowered. Total acreage irrigated from ground-water sources was estimated to be 6,200 acres (9.69 mi^2): 2,340 acres (3.66 mi^2) in Newton County and 3,860 acres (6.03 mi^2) in Jasper County. Soils in the south half of the study area developed on till and lacustrine deposits, and the high clay content of the soil inhibits drainage. Few, if any, crops are irrigated in this area.

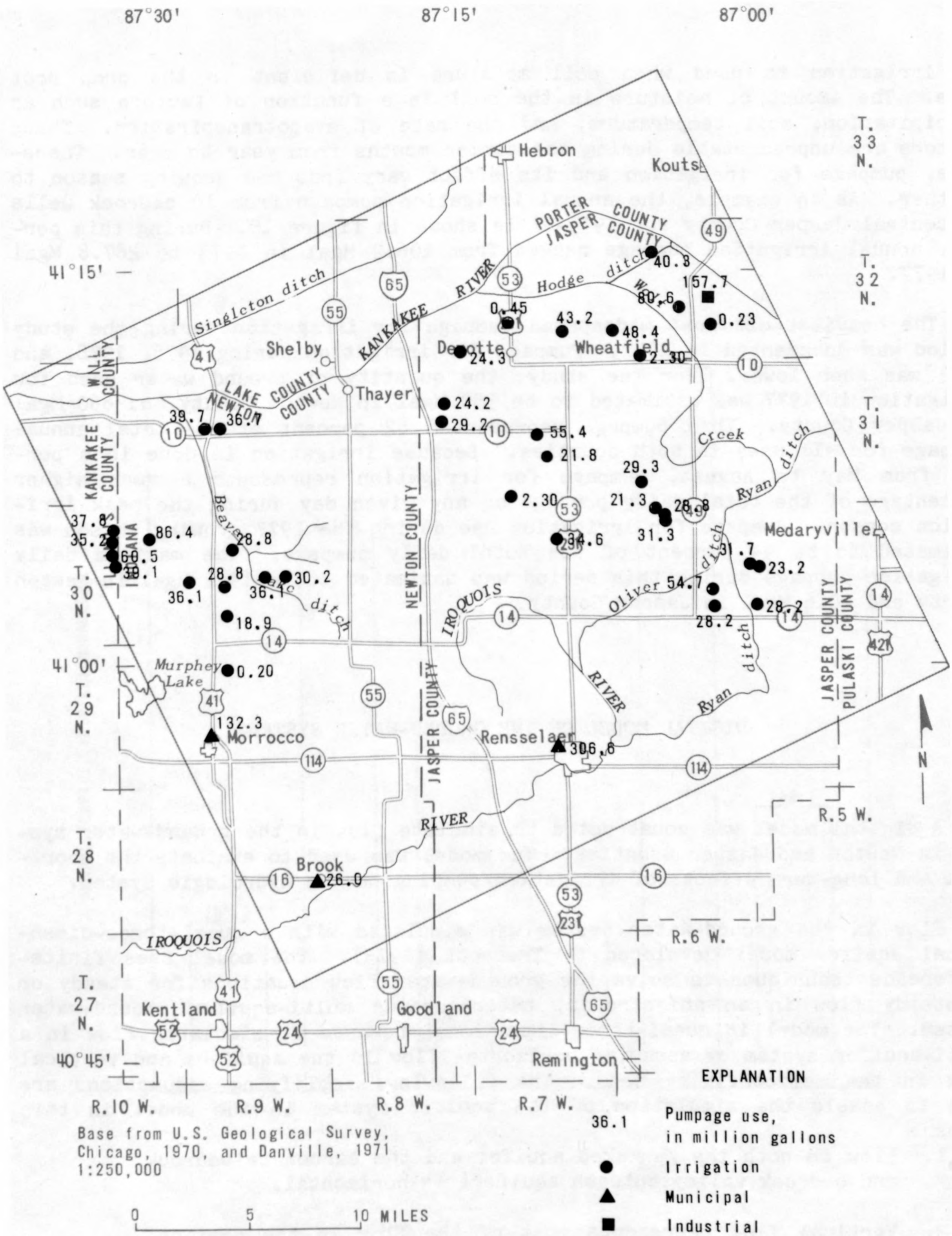


Figure 14.-- Distribution and rate of major municipal, industrial, and irrigation pumpage in study area during calendar year 1977.

Irrigation is used when soil moisture is deficient in the crop root zone. The amount of moisture in the soil is a function of factors such as precipitation, soil temperature, and the rate of evapotranspiration. These factors are unpredictable during the summer months from year to year. Therefore, pumpage for irrigation and its effect vary from one growing season to another. As an example, the annual irrigation pumpage from 10 bedrock wells in central Jasper County for 1973-78 is shown in figure 15. During this period, annual irrigation pumpage ranged from 106.2 Mgal in 1973 to 287.8 Mgal in 1977.

The heaviest and most widespread pumpage for irrigation during the study period was documented in 1977. Pumpage for irrigation during 1976, 1978, and 1979 was much lower. For the study, the quantity of ground water used for irrigation in 1977 was estimated to be 350 Mgal in Newton County and 660 Mgal in Jasper County. This pumpage represented 62 percent of the total annual pumpage for all uses in both counties. Because irrigation is done in a period from May to August, pumpage for irrigation represents a much higher percentage of the total daily pumpage on any given day during the peak irrigation season. Pumpage for irrigation use during the 1977 growing season was estimated to be 94 percent of the total daily pumpage. The maximum daily irrigation pumpage during this period was estimated to be 14.2 Mgal in Newton County and 20.6 Mgal in Jasper County.

DIGITAL MODEL OF THE GROUND-WATER SYSTEM

A digital model was constructed to simulate flow in the ground-water system in Newton and Jasper Counties. The model was used to evaluate the short-term and long-term effects of irrigation pumping on the hydrologic system.

Flow in the ground-water system was simulated with a quasi-three-dimensional aquifer model developed by Trescott (1975). The model uses finite-difference techniques to solve the ground-water flow equations for steady or nonsteady flow in an anisotropic, heterogeneous multi-aquifer ground-water system. The model is quasi-three-dimensional because it simulates flow in a multi-aquifer system by assuming horizontal flow in the aquifers and vertical flow in the semiconfining beds. The following simplifying assumptions are made to enable the simulation of the aquifer system in the model in this manner:

1. Flow in both the Kankakee aquifer and the carbonate bedrock and bedrock valley outwash aquifers is horizontal.

2. Vertical flow represents most of the flow in the semiconfining layer even though horizontal flow exists. The effects of storage in the semiconfining bed is ignored.

3. The Kankakee aquifer is homogeneous and isotropic. The hydraulic conductivity of the aquifer is uniform, vertically and horizontally.

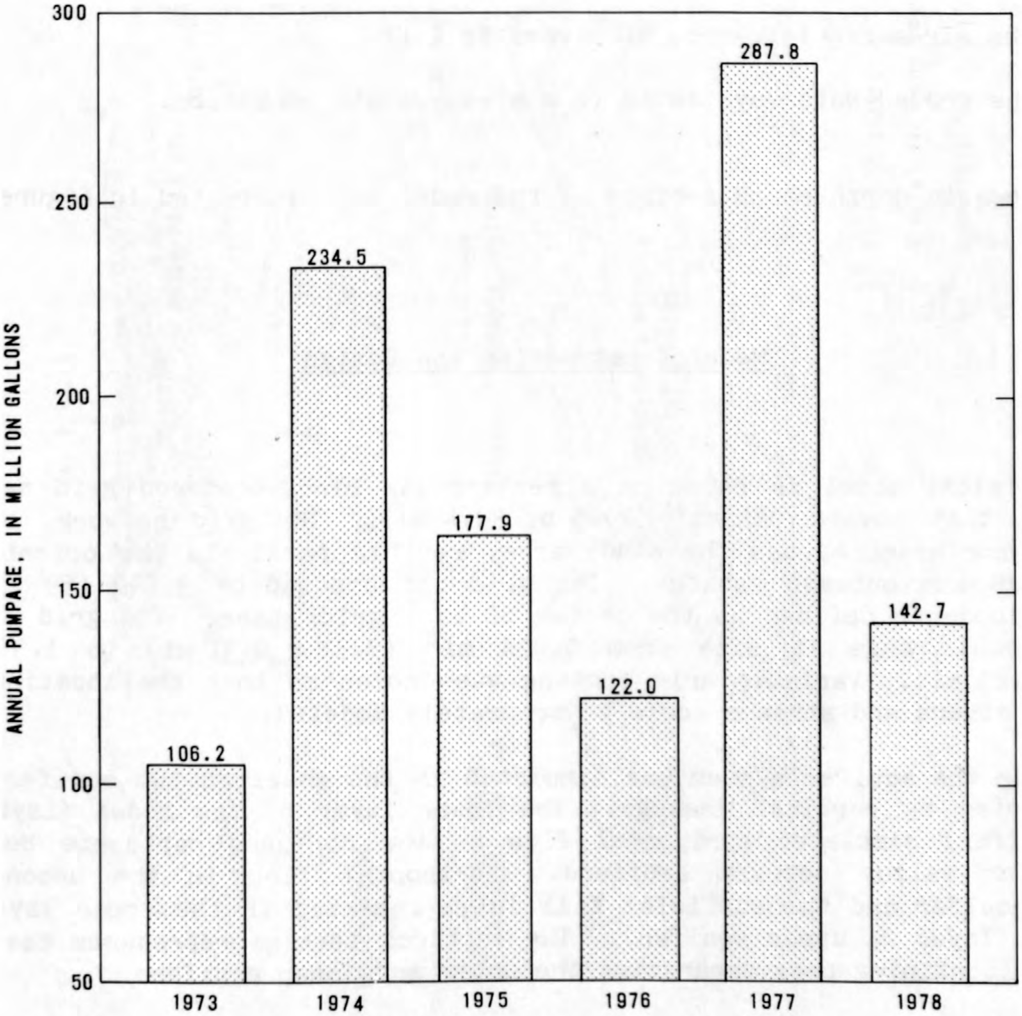


Figure 15.-- Annual variation in irrigation pumpage from 10 bedrock wells in central Jasper County, Ind.

4. The bedrock aquifer is heterogeneous and isotropic. The hydraulic conductivity is uniform vertically but not horizontally. Only the top 100 ft of the bedrock is an aquifer.

5. The confined bedrock valley outwash aquifer is homogeneous and isotropic, and the hydraulic conductivity of the outwash is uniform, vertically and horizontally. Because no semiconfining bed separates this outwash aquifer from the underlying bedrock aquifer, these aquifers represent a single unit in the model.

6. The streambed thickness of rivers is 1 ft.

7. The ground-water system is in a steady-state condition.

A schematic north-south section of the model is illustrated in figure 16.

Model Construction and Design

The digital model is based on a rectangular block-centered grid network (fig. 17) that covers 985 mi^2 (29.5 by 33.4 mi). The grid network, alined southwest-northeast across the study area, roughly parallels the orientation of the Kankakee outwash aquifer. The grid is composed of 3,540 (59 by 60) nodes. A node is defined as the center of each grid space. The grid spaces in the model range in size from 0.096 mi^2 (0.31 x 0.31 mi) to 1.33 mi^2 (1.08 x 1.23 mi). Variable grid spacing was chosen so that the locations of the major rivers and streams could be accurately modeled.

Flow in the aquifer system was simulated in the model as two aquifer layers connected by vertical leakage. The lower layer of the model (layer 1, lower aquifer) simulates horizontal flow in the confined carbonate bedrock and bedrock valley outwash aquifers. Horizontal flow in the unconfined Kankakee aquifer and the surficial till is represented in the upper layer of the model (layer 2, upper aquifer). The vertical leakage represents the flow in the semiconfining beds separating the upper and lower aquifers.

Input for layer 1 of the model included initial estimates of the water-level distribution and the transmissivity of the bedrock aquifer. The initial water-level distribution was derived from a potentiometric map of the bedrock aquifer based on water levels measured in June 1978. On the basis of water-level fluctuations, these water levels were assumed to be in a steady-state condition preceding seasonal irrigation pumping. In areas where the confined bedrock outwash aquifer overlies the bedrock, head information from the outwash aquifer was entered into the model. The transmissivity distribution of layer 1 was based on the estimates of transmissivity computed for the bedrock and bedrock valley outwash aquifer from specific-capacity data. The

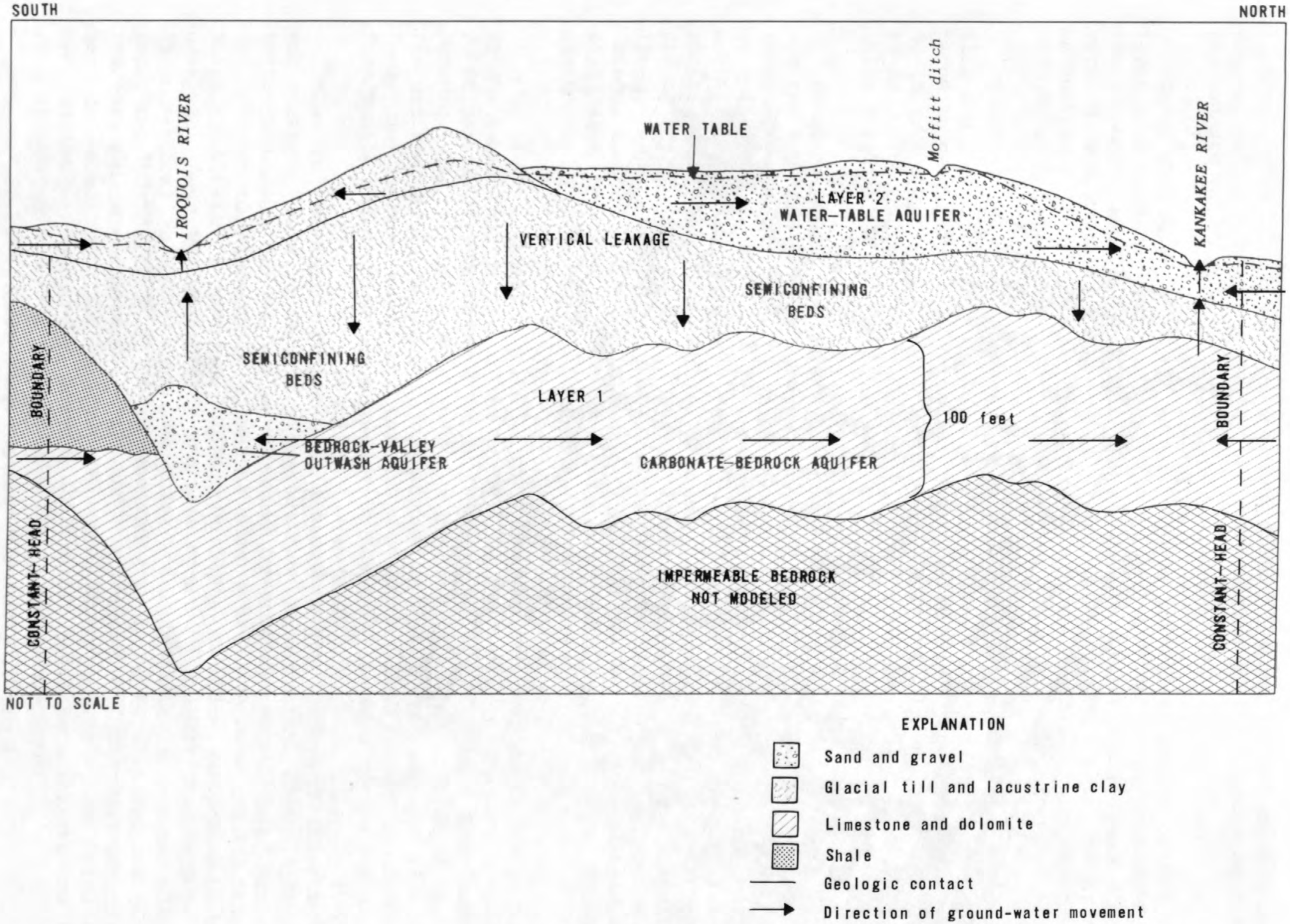


Figure 16.-- Generalized geologic section showing aquifers and direction of ground-water flow in the digital model.

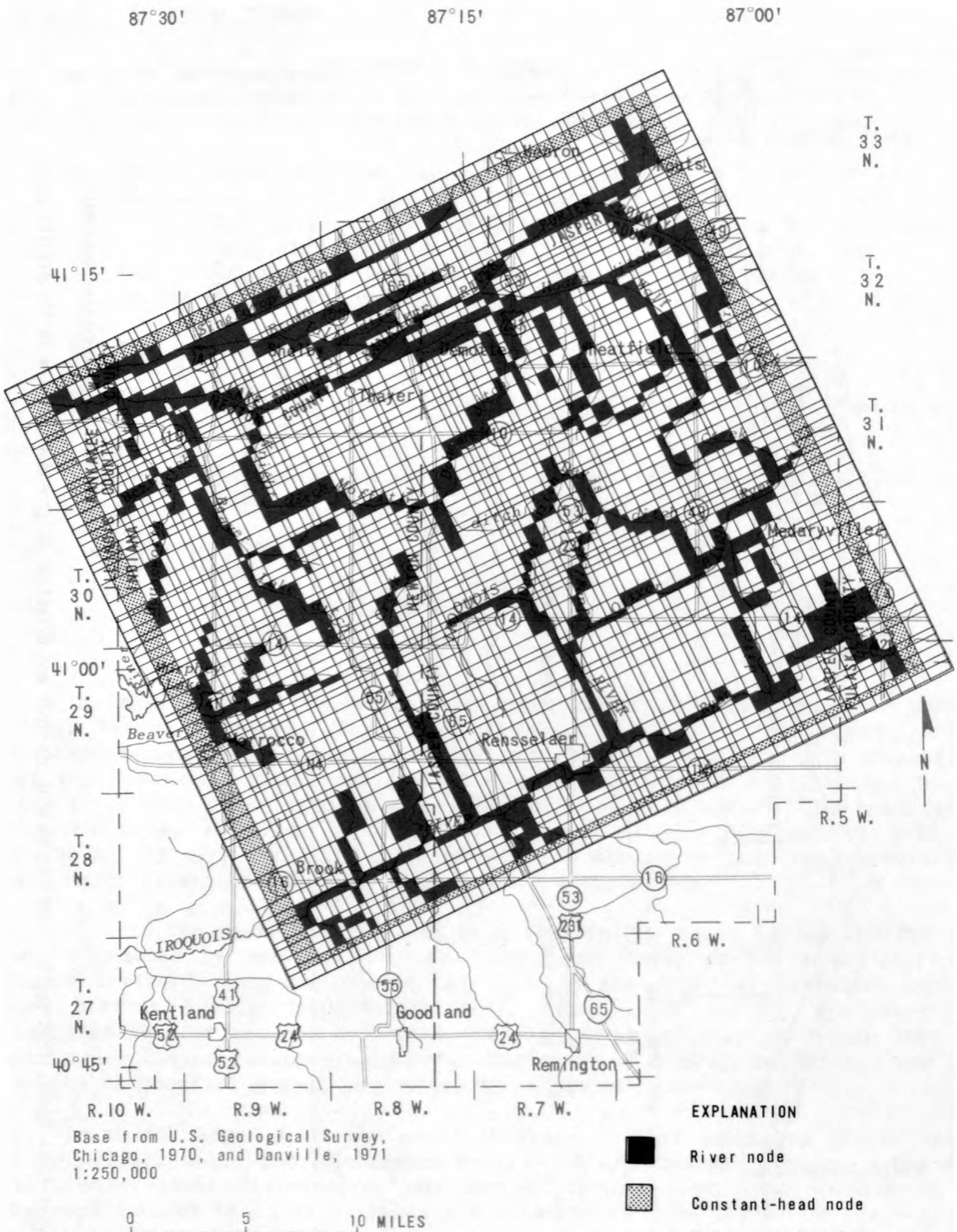


Figure 17.-- Finite-difference grid used in model.

transmissivity of layer 1 in the area of the bedrock valley outwash aquifer was assumed to be the sum of the transmissivities of the bedrock aquifer and the bedrock valley outwash aquifer.

Input for layer 2 of the model included initial estimates of the water-level distribution, the bottom altitude, and the hydraulic conductivity of the Kankakee aquifer. The initial water-level distribution was based on water levels measured in June 1978. The initial estimate of the hydraulic conductivity of the Kankakee aquifer, 250 ft/d, was based on specific-capacity data.

South of the Kankakee aquifer, layer 2 represented flow in the surficial till and lacustrine deposits. However, because this area is of minor importance to development of irrigation, the flow in these deposits was not modeled in great detail. The hydraulic conductivity of the deposits was assumed to be 1 ft/d. The bottom altitude of the layer was set at 20 ft below the initial estimate of water level.

Boundary conditions in aquifer models must be selected so that the type and the location of the boundary will have a minimal effect on the result of stresses imposed within the model. Therefore, the north boundary of the model grid was extended north of the Kankakee River into south Lake County, the south boundary was placed just beyond the Iroquois River in the till plain south of the Kankakee aquifer boundary, and east and west boundaries were placed approximately along the east Jasper County and the west Newton County lines, respectively. Constant-head boundary conditions were assigned at each node on the periphery of the model. (See fig. 17.) The constant-head values were taken from the potentiometric-surface maps of the aquifers, which were constructed from water levels measured in June 1978.

One other boundary condition was considered in addition to the peripheral constant-head boundaries of the model. This boundary concerns the hydraulic relationship of the top 100 ft of the bedrock aquifer with the underlying bedrock. The transmissivity of the bedrock below the top 100 ft of the aquifer was assumed to be small as compared with that of the top 100 ft. Therefore, the bedrock below the top 100 ft of the aquifer was assumed to be impermeable and represented a no-flow boundary at the base of the model.

Flow in the semiconfining beds separating the upper and lower aquifers was simulated as a vertical leakage between the upper and lower layers. The vertical leakage is defined as effective vertical permeability of the material between two vertically adjacent nodes divided by the thickness of the material between two nodes. This resistance to flow was approximated by using the vertical permeability and thickness of the least permeable material in a vertical column between two nodes. For instance, where two nodes are separated vertically by 25 ft of sand and gravel, 60 ft of clay, and 50 ft of bedrock, the vertical permeability and the thickness of the clay material are used in calculating the vertical leakage. Where the upper sand and gravel unit directly overlies the bedrock aquifer, vertical permeability and thickness of bedrock are used to calculate the vertical leakage.

A total of 711 nodes were chosen to simulate a selected set of rivers, streams, and ditches within the modeled area. Grid spaces that most closely delineated the course of each river were used to represent the locations of the rivers and streams (fig. 17). All river nodes were simulated in the upper layer of the model.

Input data for each river node included the altitudes of the river head and the bottom of the riverbed as well as the leakage coefficient of the river. The leakage coefficient, L_R, is defined by the following equation:

$$L_R = \frac{K_v \times A_s}{m \times A_n}$$

where

K_v is the vertical permeability of the streambed,
m the thickness of the streambed,
A_s the area of the stream, and
A_n the area of the node.

The streambed permeability of streams that drain areas of outwash deposits throughout the north and north-central parts of Newton and Jasper Counties was initially assumed to be 25 ft/d or one-tenth of the permeability of the Kankakee outwash aquifer. For streams draining the surficial till and lacustrine deposits, the initial streambed permeability was chosen to be the same as the initial estimate of the vertical permeability of the unit, 0.0004 ft/d. The streambed thickness for all streams was assumed to be 1 ft. The leakage coefficient of the river is combined with the head difference between the stream and the aquifer in Darcy's equation to calculate the discharge to or from the modeled stream.

Steady-State Calibration

A series of steady-state model simulations was used to match simulated water levels and ground-water discharge to streams to water levels and base-flow measured during June 1978. These conditions were assumed to represent the average steady-state condition in the aquifer system before the start of a typical irrigation season.

Model calibration is a procedure of adjusting input variables so that model results closely represent real conditions. The input variables adjusted in the model calibration included areal recharge, vertical leakage in the semiconfining bed, transmissivity of the confined bedrock and bedrock valley outwash aquifers, hydraulic conductivity of the Kankakee aquifer, and the hydraulic connection between the aquifer system and the modeled streams.

The least known input variables are the vertical leakage and the areal recharge. Consequently, the calibration process began with a series of simulations to investigate these two variables. Several values of areal recharge ranging from 0.42 to 1.17 ft/yr were used on the Kankakee aquifer in layer 2. Only the top layer of the model, layer 2, was recharged. In general, an areal recharge rate of 1.08 ft/yr best approximated the measured water levels and the ground-water discharge to various stretches of the modeled streams. However, in an area along the south border of the Kankakee aquifer, where the outwash deposits grade into finer grained material associated with the till plain, a recharge value of 0.42 ft/yr best approximated the measured water levels and discharges to the streams. A recharge rate of 0.08 ft/yr was selected to produce the saturated level in the surficial-till deposits south of the Kankakee aquifer boundary although the model-simulated discharges to the streams were much lower than the measured ground-water discharge in this area.

In the next series of model simulations, values of vertical leakage (TK) of the semiconfining clay unit were investigated. Vertical hydraulic conductivity values of the semiconfining unit ranging from 0.000009 to 0.09 ft/d were simulated. Within the resolution of the model, values ranging from 0.0004 to 0.009 ft/d provided an adequate match to the vertical-head gradients measured between the upper and the lower aquifers. However, the best approximation of the measured data was provided by a vertical hydraulic conductivity value of 0.004 ft/d.

During the calibration process, changes were made to the initial distributions of the transmissivity in layer 1 and the hydraulic conductivity in layer 2 to evaluate their effect on model-simulated water levels. Changes in these variables resulted in only minor changes in the model-simulated water-level distribution. Therefore, the initial distribution of the layer 1 transmissivity and the layer 2 hydraulic conductivity remained unchanged in the calibration process.

In conjunction with the model simulations, the vertical hydraulic conductivity of the modeled streambeds was also adjusted until the measured ground-water discharge to streams was adequately matched. Because of possible error in streamflow measurements, model-simulated discharge to streams was matched within a range of values that represented a 5-percent measurement error in streamflow. Final acceptable values of vertical streambed permeability for streams modeled on the Kankakee outwash plain ranged from 0.04 to 43 ft/d. Model-simulated discharge to streams on the till plain south of the outwash aquifer did not adequately match the measured data. However, for consistency, values of the vertical hydraulic conductivity for streambeds modeled on the till plain were lowered to the vertical hydraulic conductivity of the underlying semiconfining clay unit.

After the steady-state model simulations, most of the simulated water levels matched the measured ones within 10 ft, and model-simulated discharge to streams adequately approximated the measured discharge data.

Transient Calibration

After the first phase of steady-state simulations were completed, a series of transient model simulations was done to investigate the ability of the model to approximate water-level declines caused by irrigation pumping. Model-simulated water-level declines were compared to drawdowns recorded in Geological Survey observation wells, Lake 12, Newton 8, and Jasper 4 thirty days after the beginning of pumping during the irrigation season in 1977.

Before the simulations, the model was modified so that it could be used in the transient analysis. Constant-head boundaries used in the steady-state model were changed to constant-flux boundaries. These constant fluxes were derived from the constant-head boundary fluxes computed in the final acceptable solution during the first set of steady-state solutions. This condition has one of two effects: (1) If there is no development beyond the boundary, then the condition represents the minimum condition of development in the study area because an increase in gradient caused by pumping will induce more flow through the boundaries. (2) If there is a development beyond the boundaries, then the condition indicates the maximum condition of development in the study area because there would be less flow across the boundaries, owing to other area development.

As explained previously, the storage coefficient of the bedrock aquifer was assumed to be 0.00013. The storage coefficient of the bedrock valley outwash aquifer was also assumed to be 0.00013. A specific yield of 0.12 was assigned to the Kankakee aquifer in layer 2. On the basis of estimates of specific yield for clay material by Johnson (1967, p. 14-49), a lower specific yield, 0.01, was selected for the surficial till south of the Kankakee outwash aquifer in layer 2.

Pumpage used in the model simulations was based on estimates of irrigation pumpage during the spring and summer of 1977. This pumpage amounted to 34.2 Mgal/d. The transient analysis indicated that, when the 1977 irrigation pumpage was incorporated into the model, the model-simulated drawdowns were lower than the water-level declines measured in the selected observation wells.

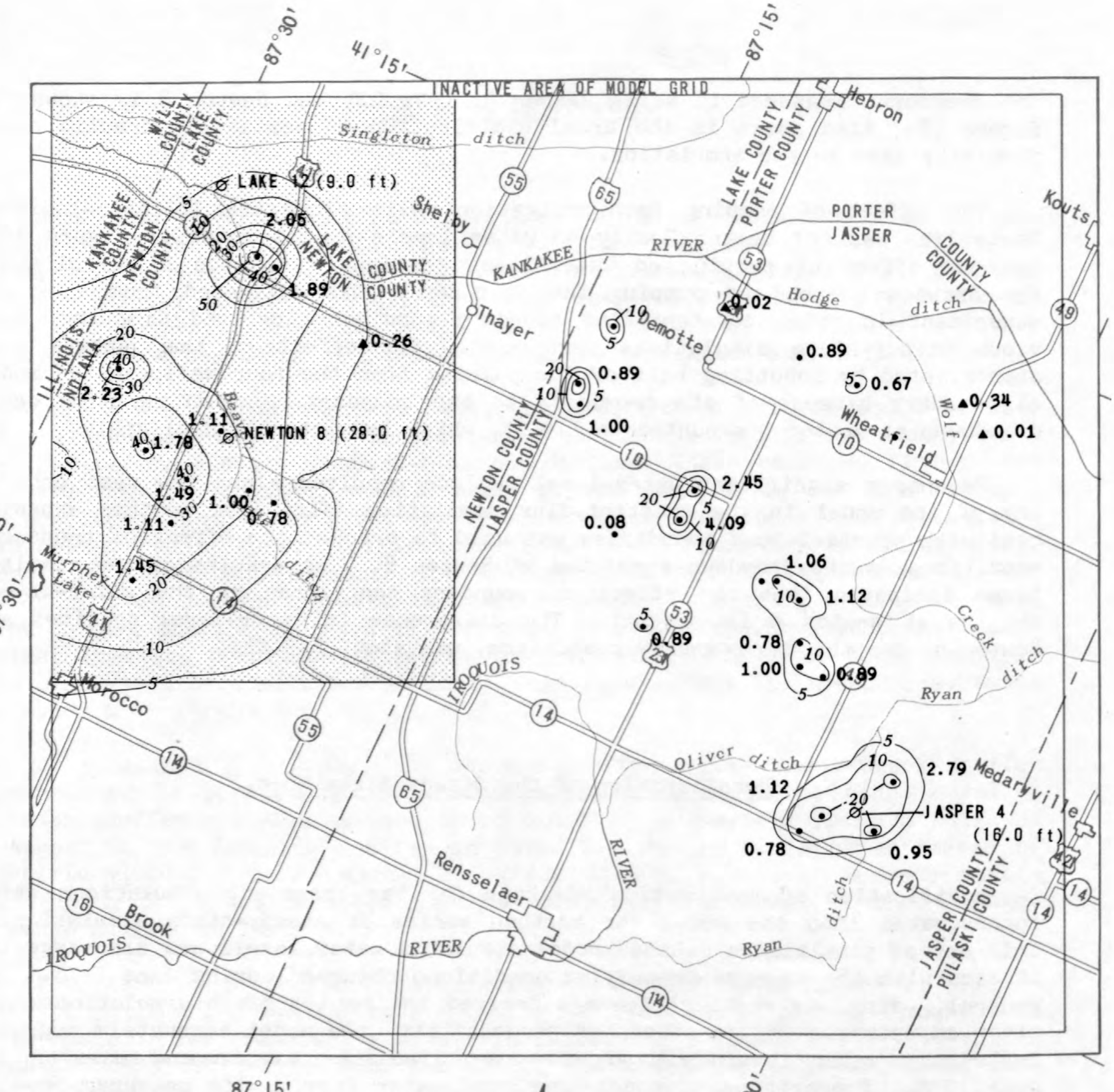
The effect of varying several hydrologic parameters, such as the transmissivity of the bedrock aquifer, the storage coefficient of the bedrock aquifer, the streambed permeability of various streams, and the vertical hydraulic conductivity of the semiconfining bed, was investigated. The most signigicant changes in water-level declines resulted from varying the vertical hydraulic conductivity of the semiconfining beds. Reductions in the vertical hydraulic conductivity caused the water-level declines to increase and to spread over a wider area than that computed in previous transient simulations. The value of vertical hydraulic conductivity that best approximated the drawdown recorded in well Newton 4 was 0.004 ft/d. A lower value, 0.0006 ft/d, provided the best match to water-level declines measured in wells Lake 12 and Newton 8. Comparison of the model-simulated drawdowns in layer 1 with the drawdowns measured in wells Jasper 4, Lake 12, and Newton 8 is shown in figure 18. Also shown is the areal distribution of vertical hydraulic conductivity used in the simulation.

The effect of pumping four irrigation wells along the Illinois-Indiana State line west of Newton County on water levels is not shown in figure 18, but the effect of pumping on water-level declines at Newton 8 was slight. The drawdown caused by pumping the four wells was estimated from a model experiment in which constant-flux boundary conditions derived from the previous steady-state simulations were used. The effect of the pumping was approximated by inputting half of the pumping from the four wells at the model boundary because of its proximity to the boundary. Resulting additional drawdowns at Newton 8 amounted to 1.4 ft, which is virtually negligible.

Because a significant water-level decline resulted along the west boundary of the model in the constant-flux simulation (fig. 18), another experiment with constant-head boundaries was used to bracket the effect of boundary conditions on the drawdown simulated at Newton 8. The results of the simulations indicated that the effect of boundary conditions on the water-level decline at Newton 8 is minimal. The difference in water-level declines at Newton 8, for the two boundary conditions, was about 1.5 ft.

Recalibration of the Steady-State Model

Modification of the vertical leakage for the transient simulations were incorporated into the model for another series of steady-state simulations. This set of simulations matched model-simulated water levels and discharge to streams with the average prepumping conditions observed during June 1978. In general, using the vertical leakage derived in the transient simulations and minor adjustments in the streambed permeability, the model adequately approximated the water levels and ground-water discharge to streams measured in June 1978. Comparisons of model-simulated water levels with measured levels for layer 1 and layer 2 for the final calibrated steady-state model are shown in figures 19 and 20, respectively. Most of the simulated water levels matched measured water levels within 10 ft. A comparison of the model-simulated ground-water discharge to streams with the range of measured discharge to streams measured on June 19 and 20, 1978, representing a 5-percent measurement error in streamflow, is shown in table 1.



EXPLANATION

Vertical hydraulic conductivity of semi-confining beds equals 0.0043 ft/d

Vertical hydraulic conductivity of semi-confining beds equals 0.00065 ft/d

—10— Line of equal water-level decline in layer 1. Interval, in feet, is variable

• Simulated pumping well in layer 1

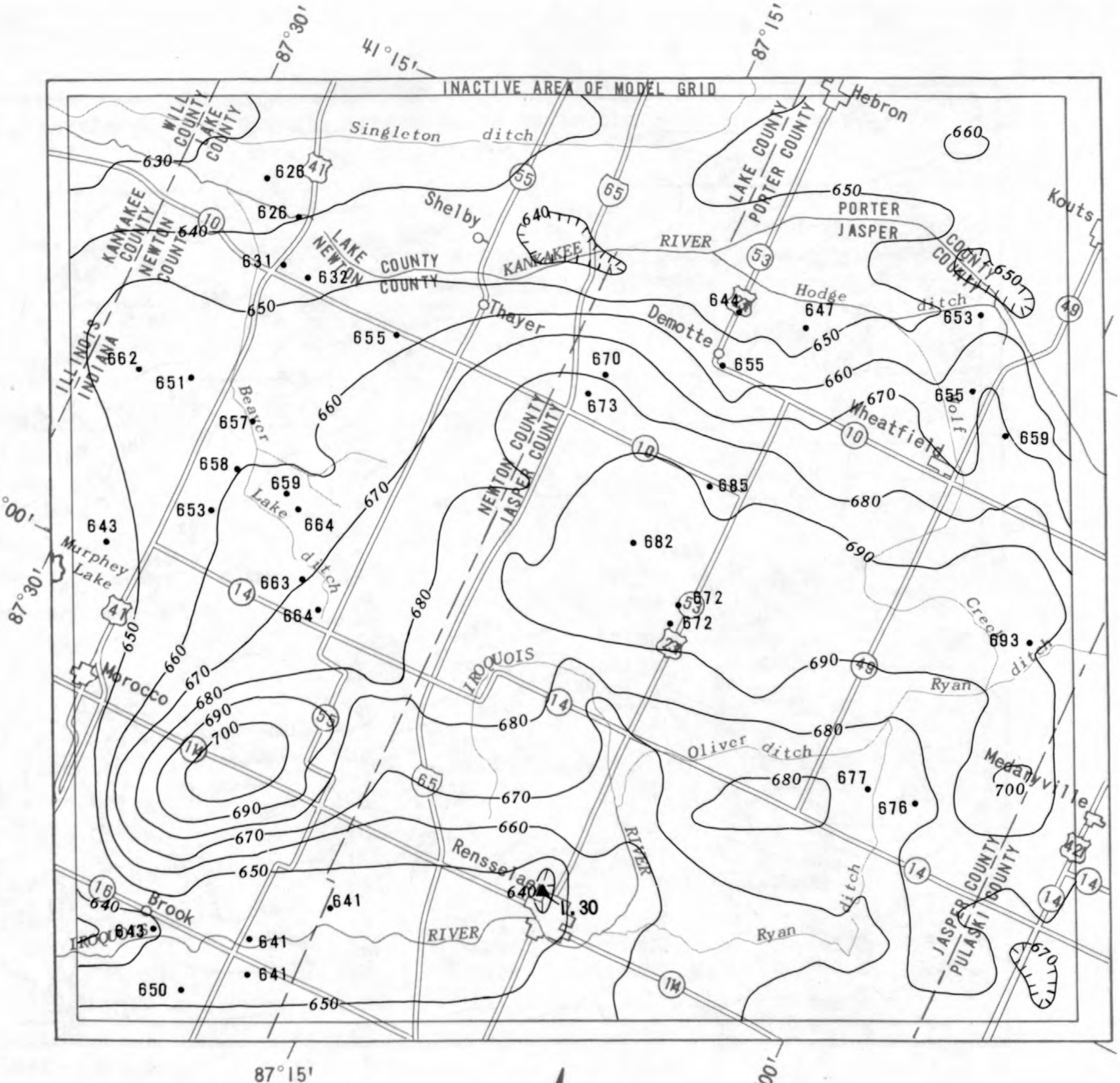
▲ Simulated pumping well in layer 2

1.12 Pumping rate, in cubic feet per second

Ø JASPER 4 Observation well and designation

(16.0) Measured drawdown, in feet

Figure 18.-- Relation of model-simulated drawdown in layer 1 and drawdowns measured in wells Jasper 4, Lake 12, and Newton 8, for the final transient calibration simulation.



Base from U.S. Geological Survey,
Chicago, 1970, and Danville, 1971
1:250,000

N

0 5 10 MILES

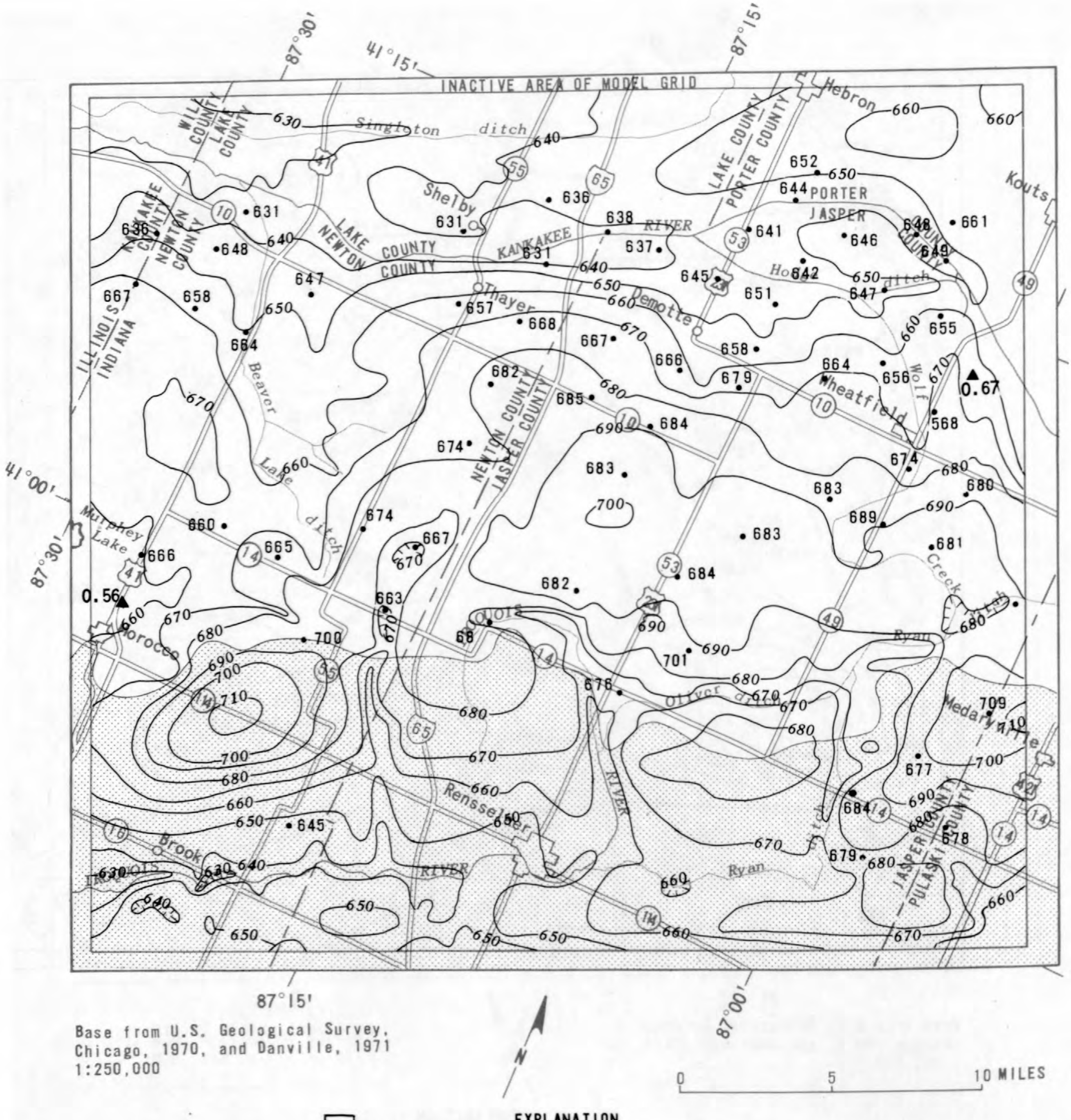
EXPLANATION

—660— Model-simulated water-level contour in layer 1. Shows altitude of simulated water level. Interval 10 feet NGVD of 1929

•664 Point of measured water level, in feet

▲1.30 Simulated pumping well and pumping rate, in cubic feet per second

Figure 19.-- Relation of calibrated, steady-state water levels in layer 1 and water levels measured on June 19-20, 1978.



Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971 1:250,000

0 5 10 MILES

EXPLANATION

Aquifer

—660— Model-simulated water-level contour in layer 2. Shows altitude of simulated water levels. Interval 10 feet NGVD of 1929

665• Point of measured water level, in feet

▲ 0.67 Simulated pumping well and pumping rate, in cubic feet per second

Figure 20.-- Relation of calibrated, steady-state water levels in layer 2 and water levels measured on June 19-20, 1978.

Table 1.--Calibrated, steady-state ground-water discharges in selected stream reaches and discharges measured in the reaches on June 19-20, 1978

Stream	Stream reach[1]	Model-simulated discharge to reach (ft^3/s)	Measured discharge to reach (ft^3/s)	Range of discharge that represents a 5-percent measurement error in streamflow (ft^3/s)
Singleton ditch	1	24.63	25.59	21.51 - 28.24
Do.	[2]27	11.53	10-26	----------------
Brown ditch	2	1.98	0.66	-0.73 - 2.05
Do.	23	6.88	5.55	5.27 - 5.83
Best ditch	3	8.24	9.77	9.28 - 10.26
Riner Houseworth ditch	4	4.57	3.85	3.33 - 4.37
Beaver Lake ditch	5	36.79	37.87	35.26 - 40.48
Kankakee River	6	76.62	98.02	-40.10 - 237.61
Do.	[2]25	38.92	40-80	----------------
Davis ditch	7	9.37	9.08	8.63 - 9.53
Knight and Moffitt ditches	8	32.89	26.8	25.46 - 27.34
Beaver Creek	9	25.79	23.4	22.23 - 24.57
Curtis Creek	10	13.08	9.37	8.90 - 9.84
Do.	11	3.25	5.23	4.03 - 6.43
Do.	12	.83	5.80	4.08 - 7.55
Iroquois River	13	7.13	9.52	8.34 - 10.46
Do.	14	3.45	6.50	1.60 - 11.40
Do.	15	2.98	39.09	24.72 - 50.51
Do.	[2]26	1.21	3-19	----------------
Do.	28	1.52	1.34	-0.99 - 7.47
Dexter ditch	16	5.15	5.9	5.61 - 6.20
Jungles and Davis ditches	17	15.48	16.45	15.60 - 17.27
Oliver ditch	18	9.00	5.55	2.79 - 8.34
Mosley ditch	19	4.86	3.71	3.52 - 3.90
Dehaan ditch	20	23.80	19.8	18.81 - 20.79
Wolf Creek	21	37.29	27.2	25.84 - 28.55
Ryan ditch	22	28.6	13.76	13.07 - 14.45
Tully ditch	24	5.01	7.99	7.59 - 8.39
Hodge ditch	29	30.70	27.9	23.79 - 32.02

[1]See figure 12 for location of stream reach.
[2]Acceptable range of ground-water discharge was not computed. because the amount of discharge to the reach is estimated.

The sources and discharges derived from the calibrated steady-state solution for the prepumping conditions simulated on June 19-20, 1978, are summarized in table 2. The magnitude of the flow components in the water budget indicates the general nature of the regional flow in the aquifer system from one year to another before an irrigation season.

Table 2.--Water budget of the calibrated steady-state model simulating conditions on June 19-20, 1978

Sources (ft^3/s)		Discharges (ft^3/s)	
Direct recharge	528.24	Historical pumpage	2.53
Boundary flux	11.85	Discharge to streams	502.41
		Boundary flux	33.06
Totals	540.09		538.00

ASSESSMENT OF THE EFFECTS OF IRRIGATION PUMPING

A main objective of the study was to assess the effect of irrigation pumping on the ground-water system in Newton and Jasper Counties. The purpose of the digital flow model was to evaluate this problem. The effects of potential increases in irrigation pumping on the ground-water system were studied. Subjects of the study included water-level decline, well interference, streamflow reduction, and water-level recovery after the end of irrigation pumping.

The recalibrated steady-state model was used to simulate seasonal irrigation pumping. The boundary conditions used in the steady-state simulations (constant head) was changed to constant flux for the simulations. Because boundaries simulated in this manner effectively represent a no-flow boundary condition, if development outside of the model does not affect flow across the boundaries, drawdowns and streamflow reductions caused by the hypothetical pumpage in the model experiments represent maximum effects of the pumpage.

The storage terms used during the transient calibration were assigned to the aquifers for the transient model experiments. A storage coefficient of 0.00013 was selected for the bedrock and the bedrock valley outwash aquifers. The Kankakee outwash aquifer was assigned a specific yield of 0.12, and the surficial till modeled in layer 2 south of the outwash aquifer was assigned a specific yield of 0.01.

The areal recharge used in the model experiments was assumed to be the same as the areal recharge derived from the steady-state calibrated model. This recharge rate was held constant throughout all the experiments.

The pumping simulated in the model experiments was used to demonstrate the effect of irrigation pumping during the growing season. A pumping rate of 800 gal/min used for each pumping node simulated represents an irrigation system that could irrigate a field of 320 acres (0.5 mi^2). The simulated wells were pumped continuously over a 4-month period to simulate a season of severe drought.

All pumping experiments simulate the effect of hypothetical pumping and existing pumping included in the steady-state calibrated model. The effects of pumping other than pumping simulated in the steady-state calibration are not considered in the analysis.

Water-Level Decline and Well Interference

Water-level decline and well interference caused by irrigation pumping are of primary importance when the effect of irrigation on ground-water availability within the study area is considered. Large-scale ground-water withdrawals for irrigation, primarily from the bedrock aquifer, has resulted in substantial water-level declines during the late summer. Water-level declines of as much as 30 ft have been recorded in some observation wells near major pumping centers. Further irrigation development will result in additional drawdown.

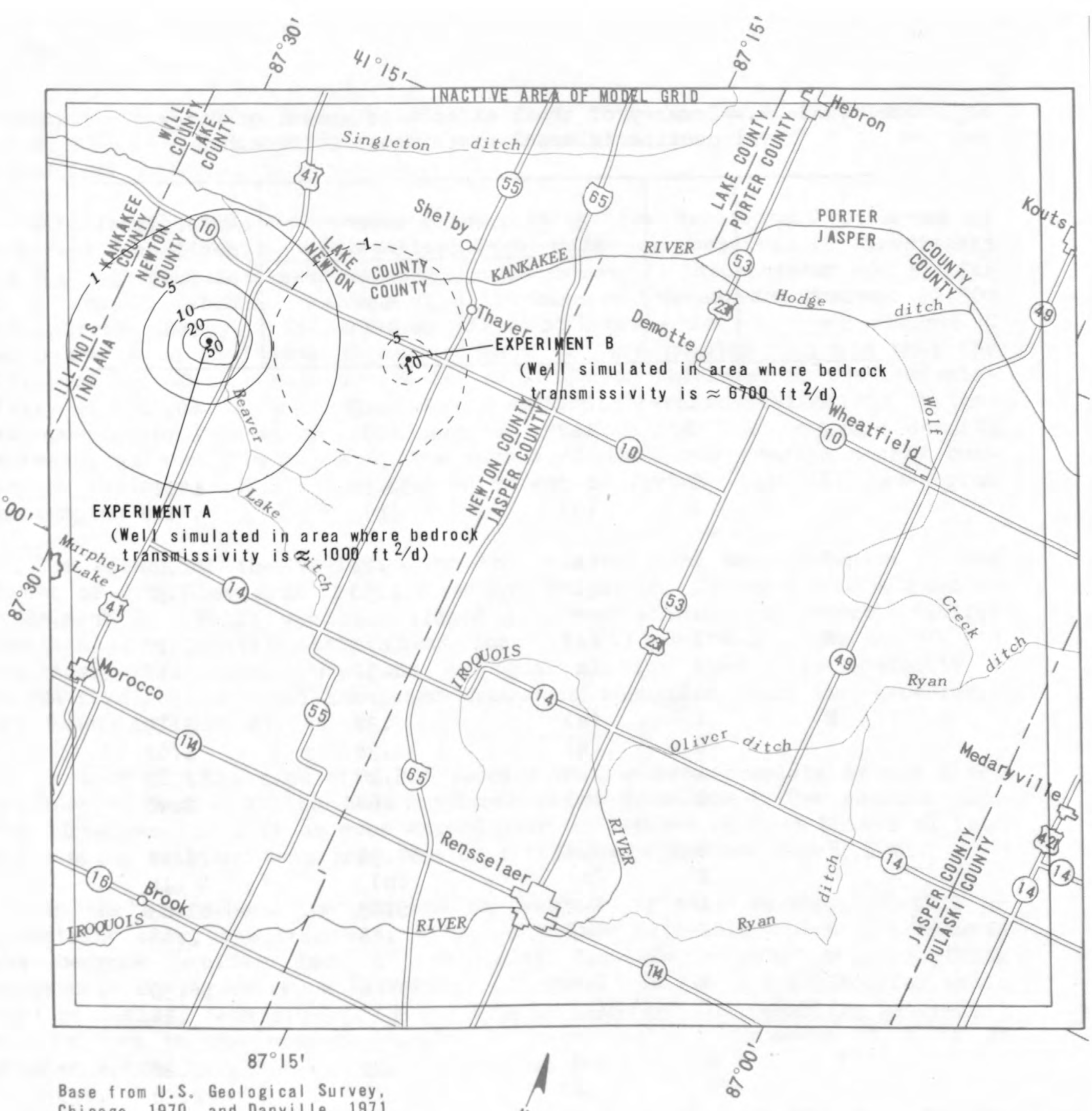
The effect of irrigation development on water levels in areas of varying hydrologic conditions were simulated in a series of eight experiments with the model. These experiments were assigned the letters A through H to simplify the discussion of the results. A brief description of the hydrologic conditions of the area in each experiment is given in table 3.

All simulations demonstrate the effects of development of irrigation wells pumping at a specified rate at various locations within the study area. The development of additional irrigation wells pumping at the same rate in areas where the simulated pumpage is located would probably produce a similar amount of drawdown as the modeled pumpage. No attempt was made in the analysis to use multiple pumping-well schemes to optimize the spacing of irrigation wells. The general effect of a system of multiple irrigation wells can be evaluated by using the law of superposition of solutions. The cumulative drawdown for pumpage from the bedrock aquifer caused by more than one well can be calculated by summing up the individual effect of each well. For pumping from the unconfined Kankakee outwash aquifer, the transmissivity is reduced as water levels decline. Consequently, the individual effect of multiple-well pumping cannot be superposed. However, if the transmissivity of the outwash aquifer is not significantly reduced by the pumping, the general effect of multiple-well pumping can be approximated by the superposition principle.

Table 3.--General description of the hydrologic conditions of areas pumped in model experiments A through H

Model experiment	Description of the hydrologic conditions of area being pumped
A	Bedrock aquifer. Pumping in an area of low transmissivity and distant from induced recharge source. Bedrock overlain by 50-60 ft of semiconfining beds.
B	Bedrock aquifer. Pumping in an area of high transmissivity and distant from induced recharge source. Bedrock overlain by 50-60 ft of semiconfining beds.
C	Kankakee aquifer. Pumping in an area of 25 ft of saturated thickness and distant from induced recharge source.
D	Kankakee aquifer. Pumping in an area of 50 ft of saturated thickness and distant from induced recharge source.
E	Bedrock aquifer. Areas pumped overlain by less than 10-ft or 35- to 40-ft thicknesses of semiconfining bed.
F	Bedrock aquifer. Pumping in areas overlain by semiconfining beds of low (0.0006 ft/d) and high (0.004 ft/d) vertical hydraulic conductivity.
G	Bedrock aquifer. Pumping in an area of high transmissivity near induced recharge source (Kankakee River).
H	Kankakee aquifer. Pumping in an area of high saturated thickness near induced recharge source (Kankakee River).

The effect of variations in transmissivity on water-level declines within the bedrock aquifer was demonstrated in model experiments A and B. Well sites were selected for experiments A and B so that the simulated pumpage was derived from areas of low transmissivity (1,000 ft^2/d) and high transmissivity (6,700 ft^2/d). Thickness (50 to 60 ft) and vertical permeability (0.0006 ft/d) of the overlying semiconfining unit were similar at both well locations. The model-simulated water-level declines resulting from the experiments are shown in figure 21 and are summarized in table 4.



Base from U.S. Geological Survey,
Chicago, 1970, and Danville, 1971
1:250,000

EXPLANATION

—— 5 —— Line of equal water-level decline, experiment A, in feet. Interval is variable

- - - 5 - - - Line of equal water-level decline, experiment B, in feet. Interval is variable

• Simulated pumping well, layer 1. Pumping rate for each well is 1.78 ft^3/s (800 gal/min)

Figure 21.-- Model-simulated water-level declines in model experiments A and B.

Table 4.--Summary of areal extent and amount of water-level decline in model experiments A through H

Model experiment	Layer	Area (mi^2) where water-level declines are greater than or equal to 5 ft	Area (mi^2) where water-level declines are greater than or equal to 1 ft	Water-level declines at pumping node (ft)
A	1	12.78	59.83	70.32
	2	(a)	(b)	.28
B	1	1.85	32.65	14.05
	2	(a)	(b)	.28
C	1	(a)	.21	1.29
	2	.08	1.30	7.97
D	1	(a)	(b)	.28
	2	(a)	1.32	4.67
E	1	(a)	.38	2.19
	2	(a)	.38	1.62
	1	1.47	5.17	82.83
	2	(a)	.64	2.48
F	1	9.91	57.35	28.07
	2	(a)	(b)	.16
	1	1.45	5.03	82.83
	2	(a)	.64	2.48
G	1	2.42	33.46	16.50
	2	(a)	(b)	.14
H	1	(a)	.09	1.28
	2	.17	.74	2.58

a Indicates water-level decline of less than 5 ft.
b Indicates water-level decline of less than 1 ft.

The results indicate that development of irrigation pumping from the bedrock aquifer in areas of varying transmissivities will produce significantly different patterns of water-level declines in these areas. Water-level declines from experiment B indicate that drawdowns of 5 ft or more extend 0.75 to 1 mi from the pumping well and cover an area of approximately 2 mi^2 and that drawdowns of 1 ft or more extend to a distance of 3 to 4 mi from the well and spread over 30 mi^2. Water-level declines were greatest in experiment A, where the simulated pumpage is derived from an area of low bedrock transmissivity. These results indicate that drawdowns of 5 ft extend 2 to 2.5 mi from the pumping well and that water-level declines of 1 ft or more extend from 4 to 5 mi from the well.

Water-level declines resulting from irrigation development in areas of different transmissivity in the Kankakee aquifer are presented in experiments C and D. The wells are in areas where the saturated thickness of the aquifer is 25 and 50 ft (fig. 22) and the transmissivities are 6,250 and 12,500 ft^2/d, respectively. The model-simulated drawdowns for both experiments C and D are shown in figure 22 and in table 4. The results indicate that the effect of the 50-percent decrease in the saturated thickness or the transmissivity of the outwash aquifer on the drawdown pattern resulting from irrigation pumping will be minor. Although the drawdown near the pumping node will increase, as shown in table 4, the radius of influence remains nearly constant. Drawdowns of 1 ft or greater extend no further than 1 mi away from the pumping well.

The effect of the variation of the semiconfining bed thickness on the amount of water-level declines caused by irrigation pumping was simulated in experiment E. Wells were positioned in areas within the bedrock aquifer overlain by relatively thin (less than 10 ft) or thick (35 to 40 ft) sequences of semiconfining beds of equal vertical hydraulic conductivity (0.004 ft/d). The model-simulated drawdowns resulting from the experiment are shown in figure 23.

In general, the simulation of pumping in the area overlain by the thick sequence of semiconfining beds produced large drawdowns. The results show that drawdowns of 5 ft or more extend over a distance of less than 1 mi from the pumping well and that drawdowns of 1 ft or more stretch over 1.5 mi.

In the area where the bedrock is overlain by thin deposits of till or lacustrine clay, the thin clay allows a better hydraulic connection between the bedrock aquifer and the overlying Kankakee aquifer than a thick sequence. Consequently, a large part of water from the bedrock-aquifer wells will be derived from storage in the outwash aquifer, and resulting water-level declines in the bedrock aquifer will be small. Drawdowns of 1 ft or greater extend only 0.4 mi from the pumping well.

Model experiment F (fig. 24) demonstrates the effect of hydraulic conductivity of a semiconfining bed on water-level declines. Wells were placed in areas of similar semiconfining-bed thicknesses but of different vertical hydraulic conductivities (0.0006 and 0.004 ft/d).

Pumping in the area of low vertical permeability (0.0006 ft/d) produced a large radius of influence. Results of the simulation are shown in figure 24 and are summarized in table 4. Pumping in the area of low permeability created drawdowns of 5 ft or more within 2.5 to 3.0 mi from the pumping well. Drawdowns of 1 ft or more extend to a distance of 6 mi from the pumping well. In comparison, pumping in the area of high vertical permeability (0.004 ft/d) produced drawdowns of 5 ft or more as far as 1 mi from the pumping well. Drawdowns of 1 ft or more extend over a distance of 1.5 mi from the well. The magnitude of the difference in the areal extent of water-level

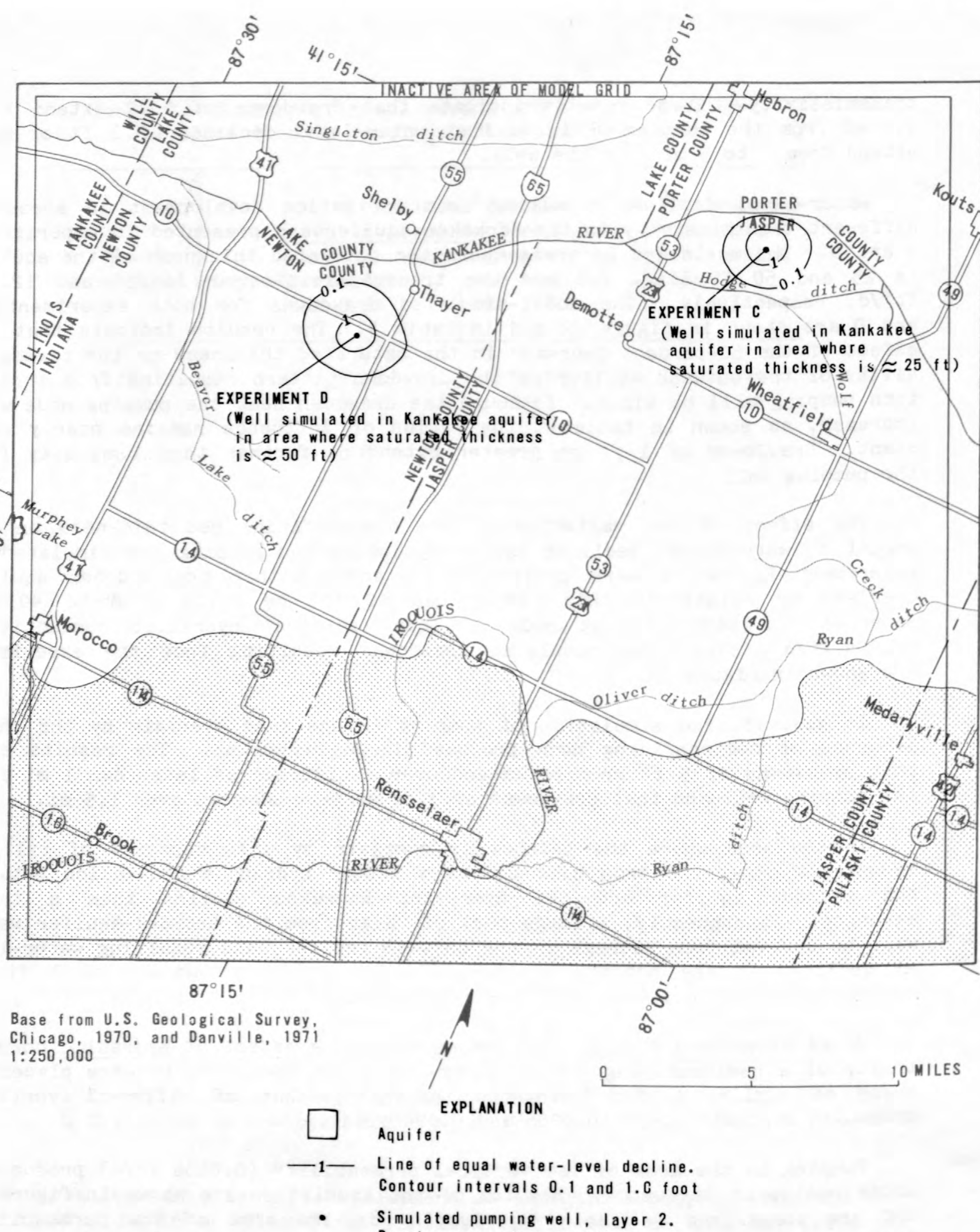
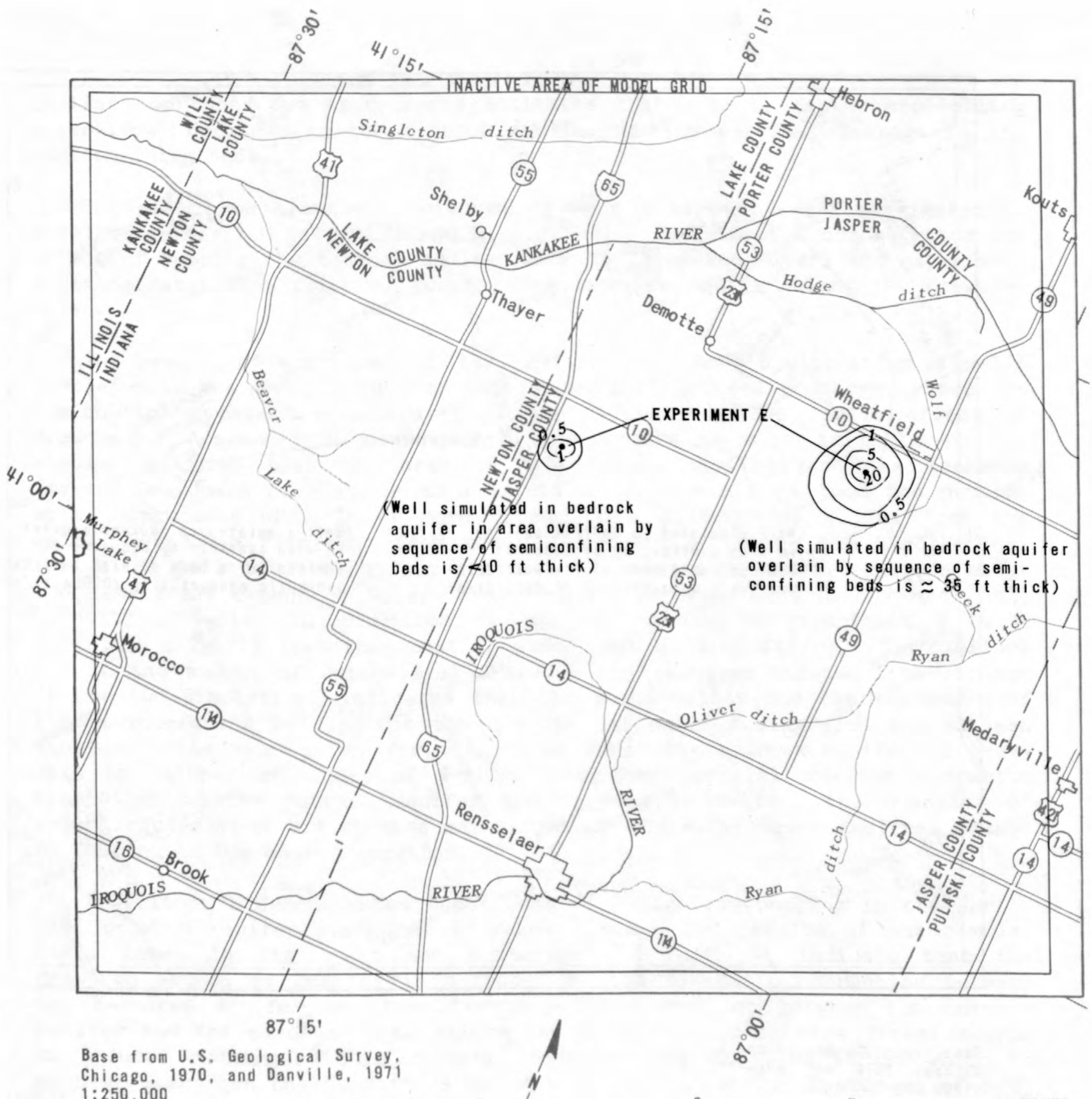


Figure 22.-- Model-simulated water-level declines in model experiments C and D.



Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971 1:250,000

EXPLANATION

—5— Line of equal water-level decline, in feet. Interval is variable

• Simulated pumping well, layer 1. Pumping rate of each well is 1.78 ft^3/s (800 gal/min)

Figure 23.-- Model-simulated water-level declines in model experiment E.

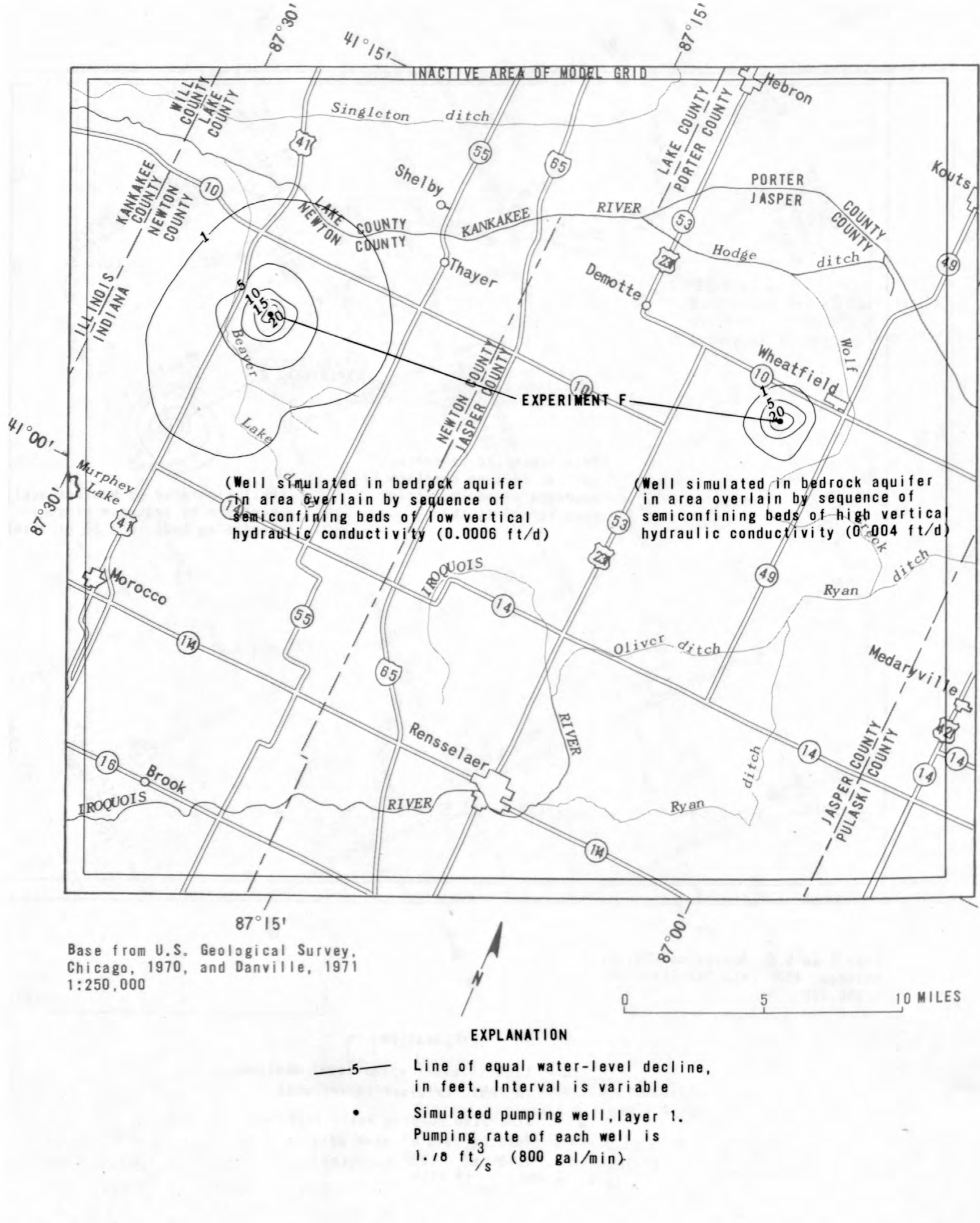


Figure 24.-- Model-simulated water-level declines in model experiment F.

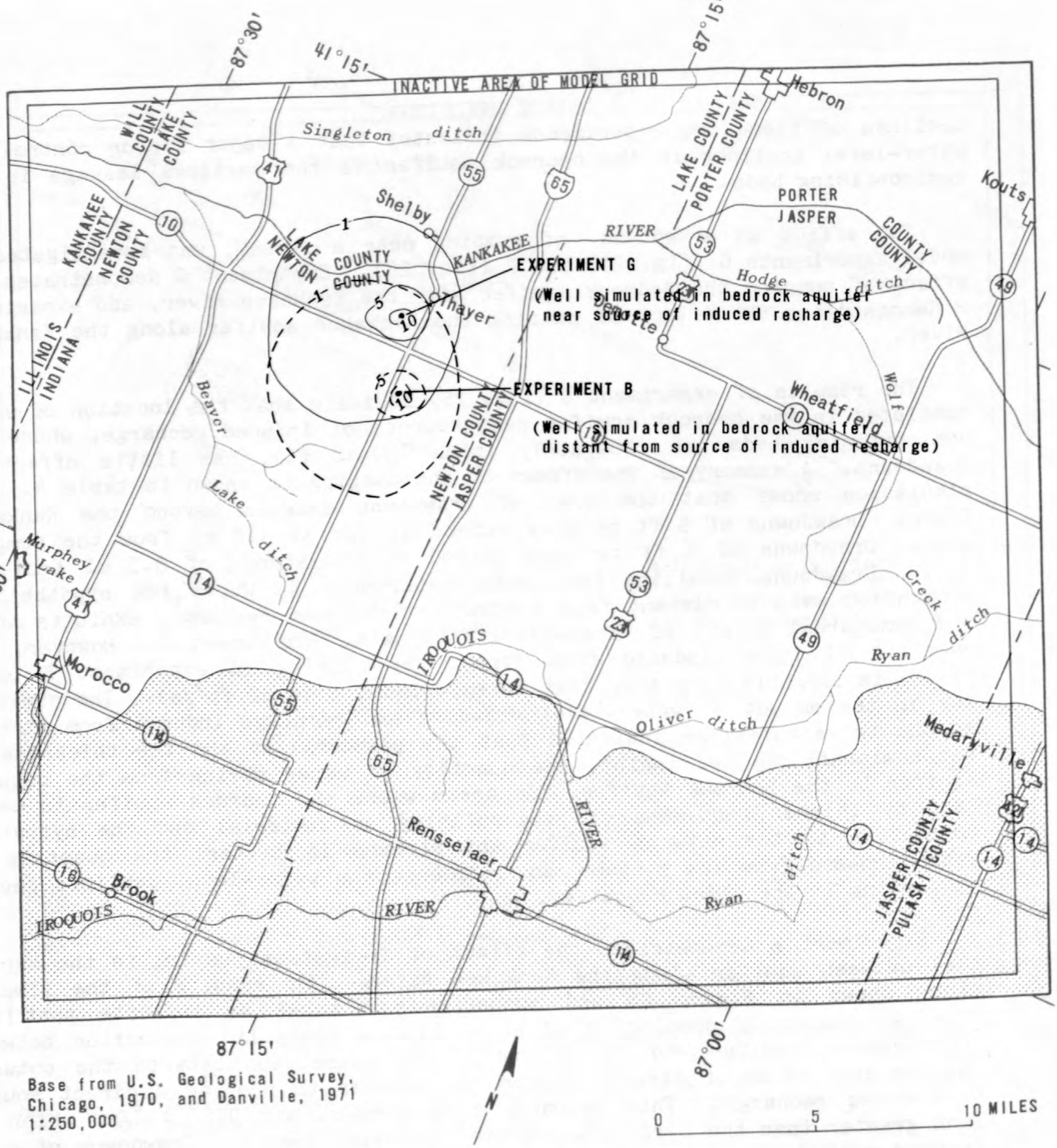
declines of these two experiments indicates that a major factor controlling water-level declines in the bedrock aquifer is the vertical leakage in the semiconfining beds.

The effect on drawdowns, of pumping near a stream, was investigated in model experiments G (fig. 25) and H (fig. 26). Experiment G demonstrates the effect of pumping the bedrock aquifer near the Kankakee River, and experiment H demonstrates the effect of pumping the Kankakee aquifer along the Kankakee River.

The results of experiment G (fig. 25) indicate that the location of wells completed in the bedrock aquifer near sources of induced recharge, where the semiconfining beds are moderately thick (30-40 ft), has little effect on drawdowns. A summary of the effect of the pumping is shown in table 4. The simulation shows that the area of drawdowns extends beyond the Kankakee River. Drawdowns of 5 ft or more extend as far as 1.5 mi from the pumping well. Drawdowns of 1 ft or more extend to a distance of 3.5 mi from the well. Drawdowns resulting from model experiment B, where the hypothetical irrigation well is distant from a source of induced recharge, exhibits about the same areal extent of water-level decline as experiment G. However, the quantity of water induced from streams by pumping in experiment B (0.42 ft^3/s) is slightly less than that in experiment G (0.62 ft^3/s). The similarity in the amount of water-level drawdown and recharge induced from streams in the two simulations indicates that the permeability and the thickness of the semiconfining bed inhibit the quantity of water moving from the outwash aquifer to the bedrock aquifer. In areas where the bedrock aquifer is overlain by thinner sequences of semiconfining bed material and the hydraulic connection to the outwash aquifer and streams is better, the quantity of water induced from the streams may influence the water-level declines caused by pumping in the bedrock aquifer.

Experiment H demonstrates the effect of irrigation pumping in the surficial outwash aquifer along the Kankakee River. The results of the simulation, shown in figure 26 and summarized in table 4, indicate that the drawdown caused by pumping is affected by the hydraulic connection between the Kankakee aquifer and the streams. The connection between the outwash aquifer and the major streams allows the streams to provide a direct source of induced recharge. This recharge from streams was 0.62 ft^3/s, which is much greater than the 0.12 ft^3/s derived in experiment D. Drawdowns of 1 ft or more extend approximately 0.5 mi away from the pumping well. Water-level declines near and beyond the stream in the vicinity of pumping are small.

In summary, model experiments A through H demonstrate the variability and the range of water-level declines that development of irrigation wells under a variety of hydrologic conditions might cause. Results of the simulations indicate that the areal extent of drawdowns of 1 ft or more, caused by the hypothetical pumping in the bedrock aquifer, can range from as little as 0.4 mi^2 to more than 60 mi^2. The variation in water-level declines resulting from development of irrigation pumping in the Kankakee aquifer, in general, is minimal. The areal extent of drawdowns of 1 ft or more cover an area of about 1 mi^2 and extend as far as 1 mi from the pumping well.



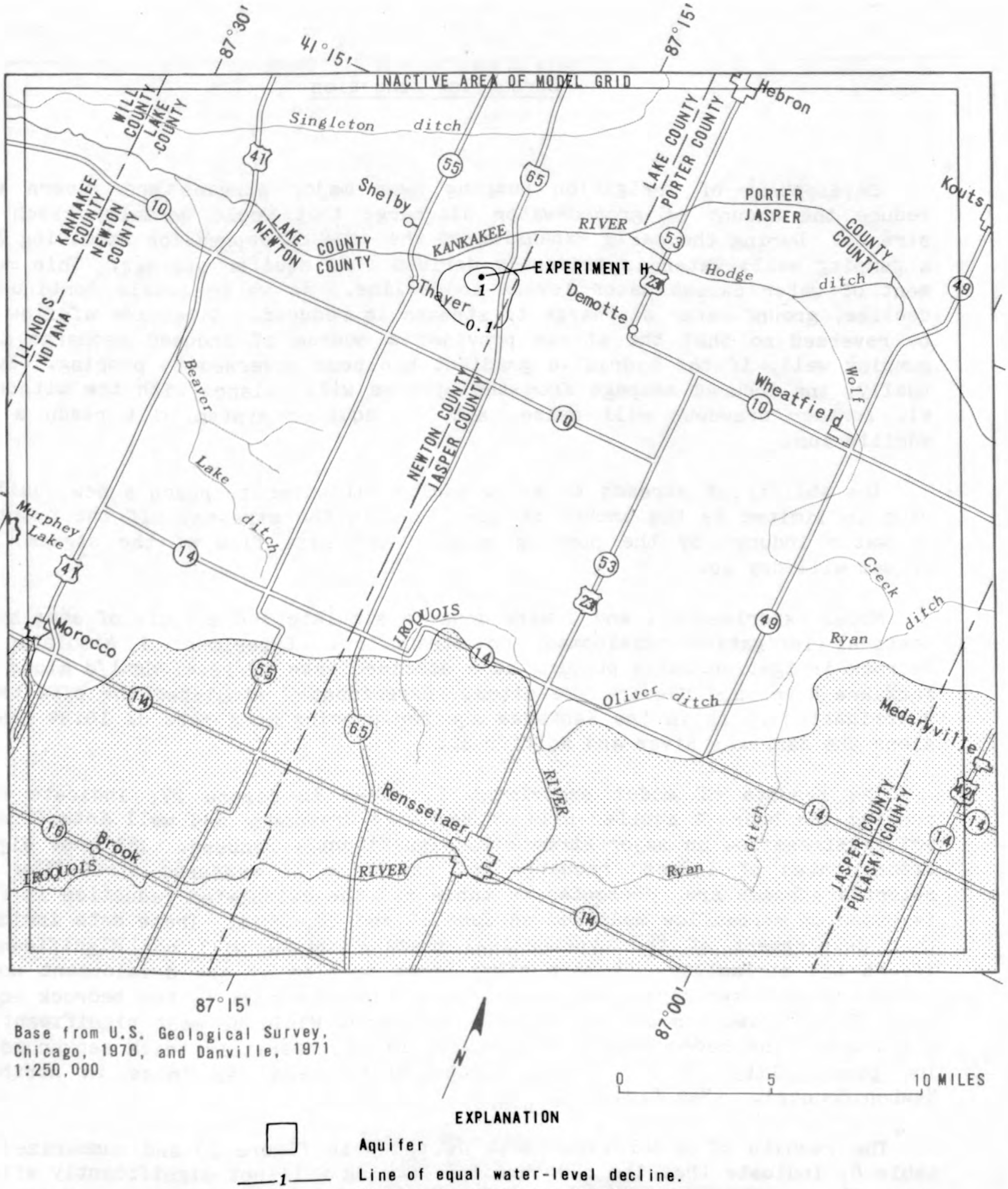
EXPLANATION

—5— Line of equal water-level decline resulting from experiment G, in feet. Contour intervals 1, 5, and 10 feet

---5--- Line of equal water-level decline resulting from experiment B, in feet. Contour intervals 1, 5, and 10 feet

• Simulated pumping well, layer 1. Pumping rate of each well is 1.78 ft^3/s (800 gal/min)

Figure 25.-- Model-simulated water-level declines in experiments B and G.



EXPLANATION

Aquifer

Line of equal water-level decline. Contour intervals 0.1 and 1.0 foot

Simulated pumping well, layer 2. Pumping rate of well is 1.78 ft^3/s (800 gal/min)

Figure 26.-- Model-simulated water-level declines in model experiment H.

Streamflow Reduction

Development of irrigation pumping near major streams and rivers will reduce the amount of ground-water discharge that would normally reach the streams. During the early expansion of the cone of depression resulting from a pumping well, water is primarily derived from aquifer storage. This movement of water causes water levels to decline. As water levels continue to decline, ground-water discharge to streams is reduced. Direction of flow may be reversed so that the stream provides a source of induced recharge to a pumping well, if the hydraulic gradient has been reversed by pumping. Eventually, the induced seepage from the streams will balance with the withdrawal, further drawdown will cease, and the aquifer system will reach a new equilibrium.

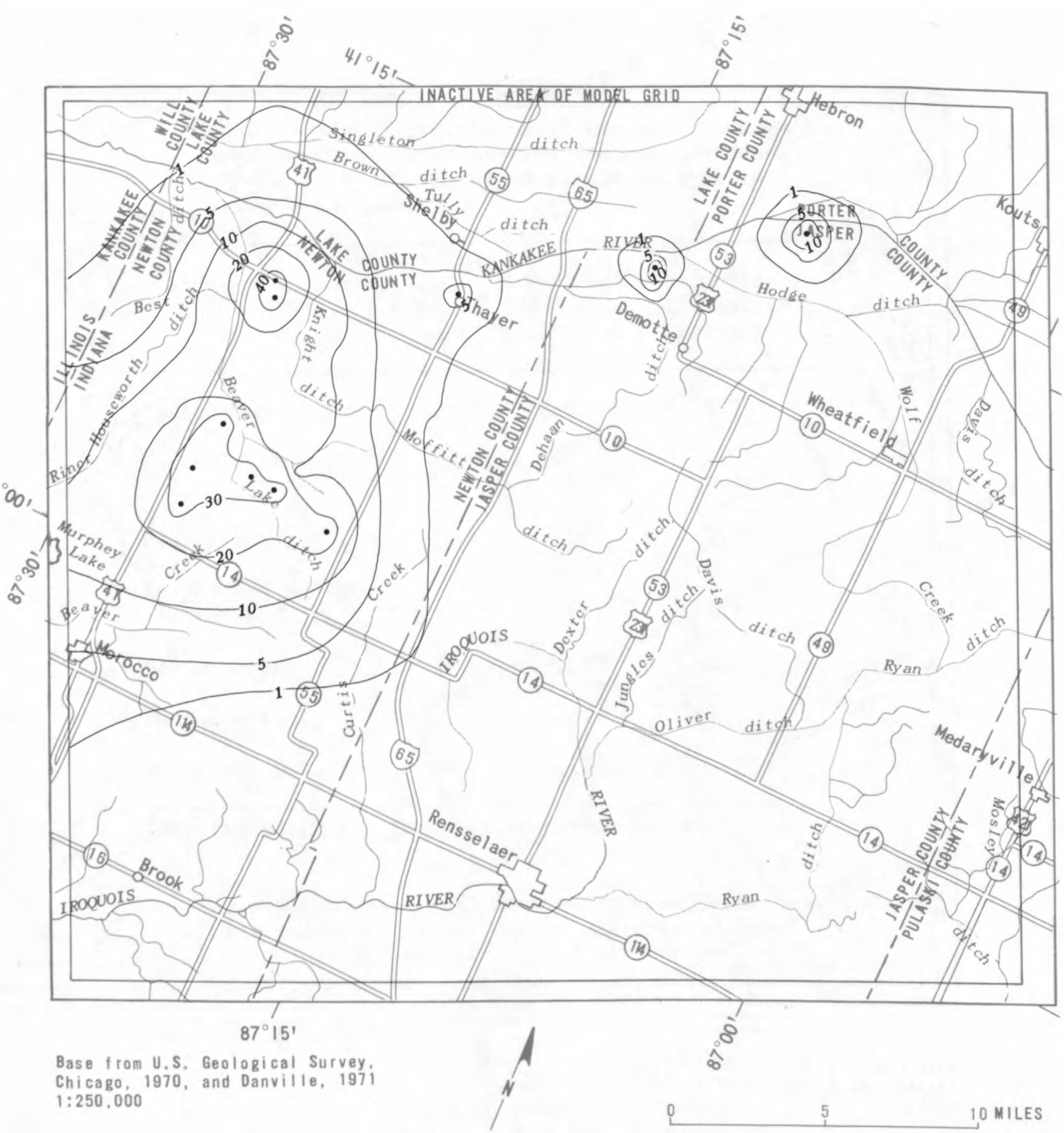
The ability of streams to allow additional water to reach a new equilibrium is limited by the amount of base flow in the streams. If the quantity of water induced by the pumping exceeds the base flow of the stream, the stream will dry up.

Model experiments I and J were done to simulate the effects of some hypothetical irrigation development on streamflow. Experiment I simulated 11 bedrock irrigation wells pumping at a combined rate of 12.61 Mgal/d along the Kankakee River and Beaver Lake ditch. Experiment J simulated the effects of 9 irrigation wells in the Kankakee aquifer pumping at a rate of 10.34 Mgal/d along the Kankakee River and Wolf Creek.

The results of model experiment I, shown in figure 27, indicate that pumping the bedrock aquifer will cause large drawdowns and well interference before streamflow on major streams is significantly reduced. The most significant rates of induced recharge and reduction in measured streamflow for selected streams are presented in table 5. The calculated reduction in flow is based on streamflow measured on June 19 and 20, 1978. These data indicate that development of the hypothetical pumping scheme will not significantly affect the streamflow. A more significant problem of the development would be the large water-level declines and well interference in the bedrock aquifer. These drawdowns and the well interference would be most significant in areas where the bedrock aquifer is overlain by moderately thick sequences of low permeability (0.0006 ft/d), semiconfining beds (as those in northern Newton County). (See fig. 27.)

The results of model experiment J, shown in figure 28 and summarized in table 6, indicate that the hypothetical pumping will not significantly affect streamflow. The pattern of water-level decline indicates a significant hydraulic connection between the outwash aquifer and the streams.

The amount of streamflow reduction computed for experiments I and J was based on streamflow measurements on June 19 and 20, 1978. As explained in the section "Connection Between Aquifer and Streams," the streamflow ranged in duration from 42.6 to 65 percent. During a period of drought, streamflow would probably be much lower than the flows measured in June 1978. The flow



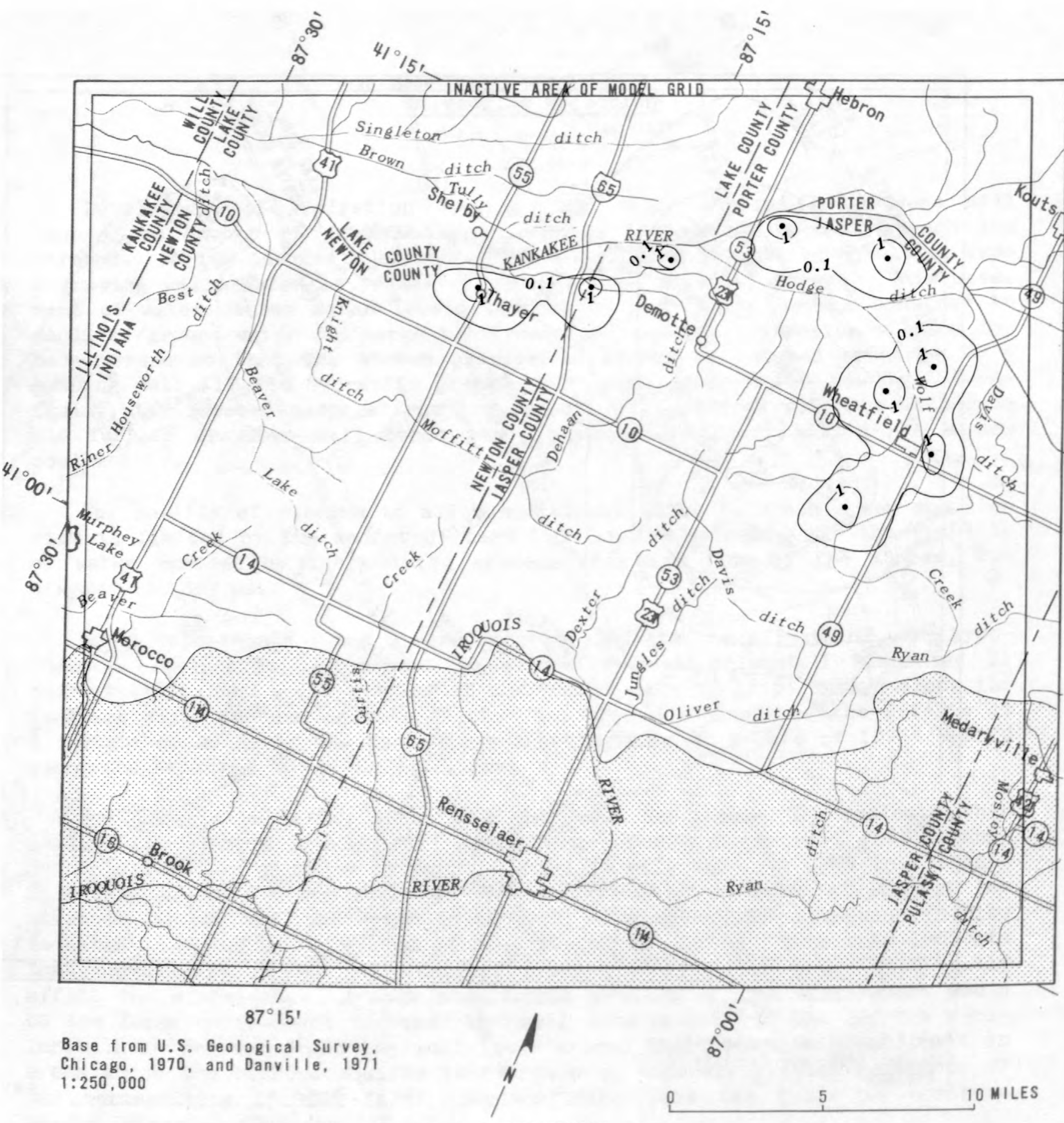
Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971 1:250,000

EXPLANATION

—1— Line of equal water-level decline, layer 1, in feet. Interval is variable

• Simulated pumping wells, layer 1. Pumping rate of each well is 1.78 ft^3/s (800 gal/min)

Figure 27.-- Model-simulated water-level declines in model experiment I.



Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971
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EXPLANATION

☐ Aquifer

——1—— Line of equal water-level decline. Contour intervals 0.1 and 1.0 foot

• Simulated pumping wells, layer 2. Pumping rate of each well is 1.78 ft^3/s (800 gal/min)

Figure 28.-- Model-simulated water-level declines in model experiment J.

Table 5.--Significant rates of induced recharge and reduction in streamflow measured on June 19-20, 1978, for selected stream reaches in model experiment I

Stream	Stream reach[1]	Induced recharge (ft^3/s)	Reduction in streamflow (percent)
Beaver Lake ditch	5	2.82	6.3
Kankakee River	6	1.83	.20
	25	.85	.47
Best ditch	3	.16	1.63
Knight and Moffitt ditches	8	.45	16.9
Beaver Creek	9	.68	2.6
Dehaan ditch	20	.89	4.5
Hodge ditch	29	.14	.25

[1]See figure 12 for location of stream reach.

Table 6.--Significant rates of induced recharge and reduction in streamflow measured on June 19-20, 1978, for selected stream reaches in model experiment J

Stream	Stream reach[1]	Induced recharge (ft^3/s)	Reduction in observed streamflow (percent)
Kankakee River	6	2.645	0.35
Dehaan ditch	20	1.54	7.80
Wolf Creek	21	.70	2.57
Hodge ditch	29	.17	1.57

[1]See figure 12 for location of stream reach.

in all streams except the Kankakee River would probably be very low or nonexistent. Under these conditions, irrigation development along the smaller streams might deplete any remaining flow. Then, the only dependable source of induced recharge would be the Kankakee River.

The magnitude and the frequency of summer low flow on the Kankakee River is shown by long-term streamflow data for the gaging station at Shelby. On the basis of streamflow measurements from October 1922 to September 1967, 7-day, 10-yr low flow is estimated to be 465 ft^3/s or 300 Mgal/d. The 1-day, 50-yr low flow is 362 ft^3/s or 234 Mgal/d (Rohne, 1972, p. 311). These flows represent the 99.4 and 99.9 percent flow duration, respectively (Horner, 1976, p. 496), and indicate that a significant amount of irrigation pumping could be developed before the flow of the Kankakee River would be greatly reduced. Under steady-state conditions, development of a 112-Mgal/d irrigation pumping rate would reduce the 1-day, 50-yr low flow on the Kankakee River at Shelby by no more than 50 percent. Because irrigation pumping is seasonal and steady-state conditions may not be reached, this pumping rate would reduce the streamflow by no more than, and probably less than, 50 percent. This type of extensive development would probably be limited more by a large water-level decline and well interference rather than by streamflow reduction.

Water-Level Recovery After Irrigation

Irrigation pumping in Newton and Jasper Counties is seasonal and is generally confined to a 4 to 6-month period extending from April and May to August and September. After this period, pumping ceases and water levels recover.

The nature of water-level recovery was described by Ferris and others (1962, p. E100-101). They state: "If a well is pumped, or allowed to flow, for a known period of time and then shut down and allowed to recover, the residual drawdown at any instant will be the same as if the discharge of the well had been continued but a recharge well with the same flow had been introduced at the same point at the instant the discharge stopped. The residual drawdown at any time during the recovery period is the difference between the observed water level and the nonpumping water level extrapolated from the observed trend prior to the pumping period." If the pumped well is allowed to recover over the same period of time required for the pumped well to reach steady-state conditions, the recovery curve will be the mirror image of the pumping curve, the residual drawdown will be equal to zero, and water-level recovery will be complete.

Model experiment K was simulated to investigate the long-term water-level recovery after irrigation pumpage ceases during the growing season. The experiment was based on a hypothetical pumping program, outlined in table 7, that might be used during a 5-yr period of high demand for irrigation. This

Table 7.--Summary of the hypothetical irrigation pumping program used in model experiment K

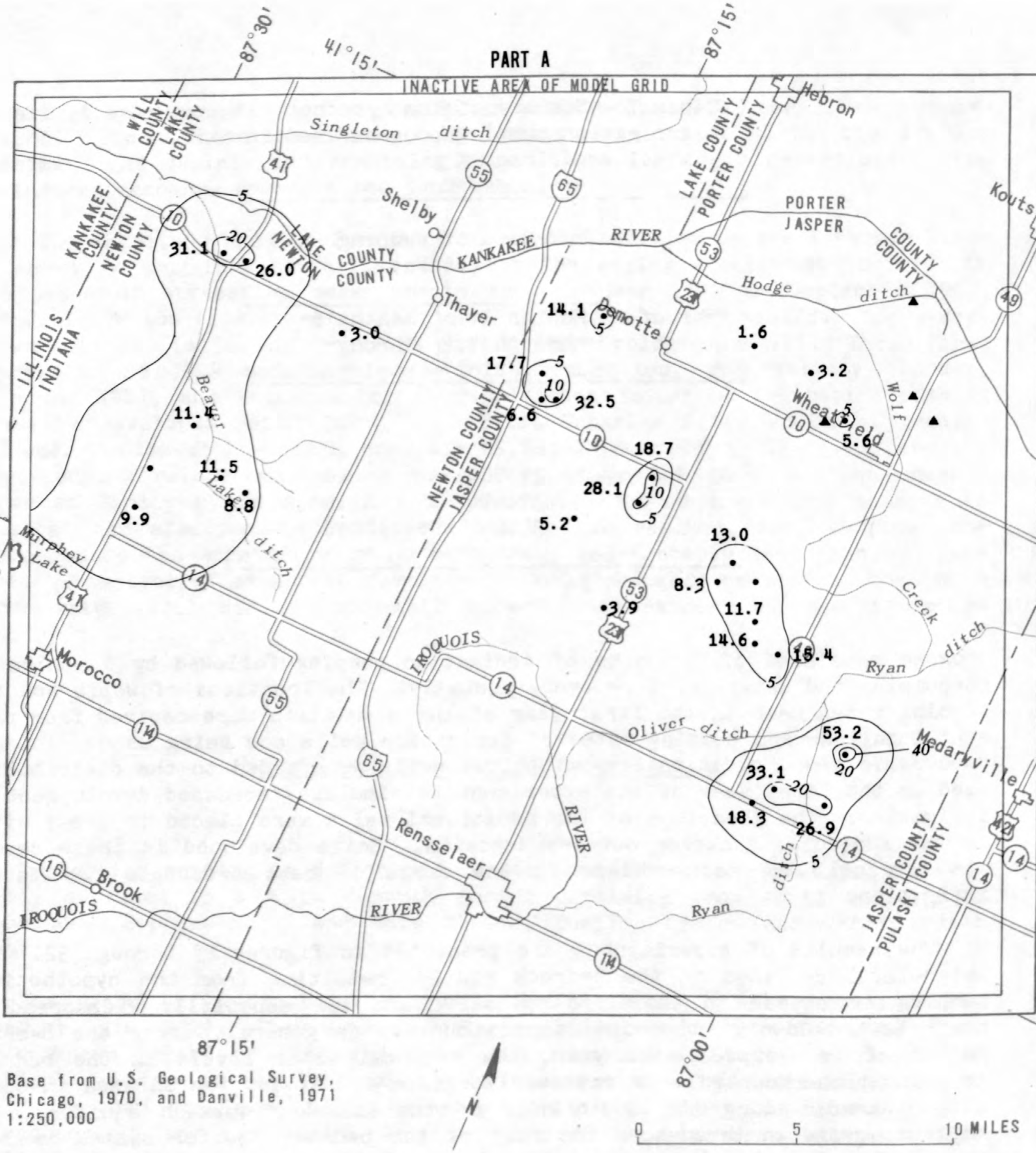
Year of simulation	Rate of pumping (ft^3/s) Months 1 through 4, irrigation	Months 5 through 12, nonirrigation
1	24.56	0.0
2	27.23	.0
3	29.90	.0
4	32.57	.0
5	35.24	.0

program consisted of 4 months of irrigation pumping followed by 8 months of nonpumping and water-level recovery annually. The locations of wells and the pumping rates used in the first year of the simulation were derived from present locations and pumping rates of irrigation wells now being used. In each successive year, additional hypothetical wells were added to the distribution used in the first year of the experiment to simulate increased development of irrigation. The locations of the additional wells were placed in areas within the surficial Kankakee outwash deposits. Soils developed in these deposits are well drained and are in the areas of best prospects for future irrigation.

The results of experiment K are presented in figures 29 through 32. The water-level declines in the bedrock aquifer resulting from the hypothetical pumping from year to year are extensive and are especially widespread in north Newton County and south Jasper County. In general, after the 8-month period of recovery in each year, the regional water levels in the bedrock throughout the modeled area are nearly recovered, except for an area in south Jasper County along the south edge of the Kankakee outwash aquifer. The residual drawdown throughout the rest of the bedrock aquifer system is less than 2 ft after 1 yr and less than 3 ft after 5 yr. The maximum areas of residual drawdown are in the immediate vicinity of the simulated wells in south-central Jasper County. The nature of water-level recovery from the bedrock aquifer is shown by a 5-yr hydrograph for a selected node in layer 1 (fig. 33) near a simulated pumping center in south-central Jasper County.

Water levels declined in the Kankakee aquifer as a result of pumping in both the outwash and the bedrock aquifers. These declines (fig. 30) are less than drawdowns in the bedrock aquifer. In general, residual drawdowns after the recovery period, also shown in figure 32, are less than 1 ft after 1 yr and 2 ft after 5 yr. The nature of water-level decline and recovery for a selected node near a simulated pumping center in south-central Jasper County is shown in figure 33.

PART A



Base from U.S. Geological Survey,
Chicago, 1970, and Danville, 1971
1:250,000

EXPLANATION

—5— Line of equal water-level decline in layer 1. Interval, in feet, is variable

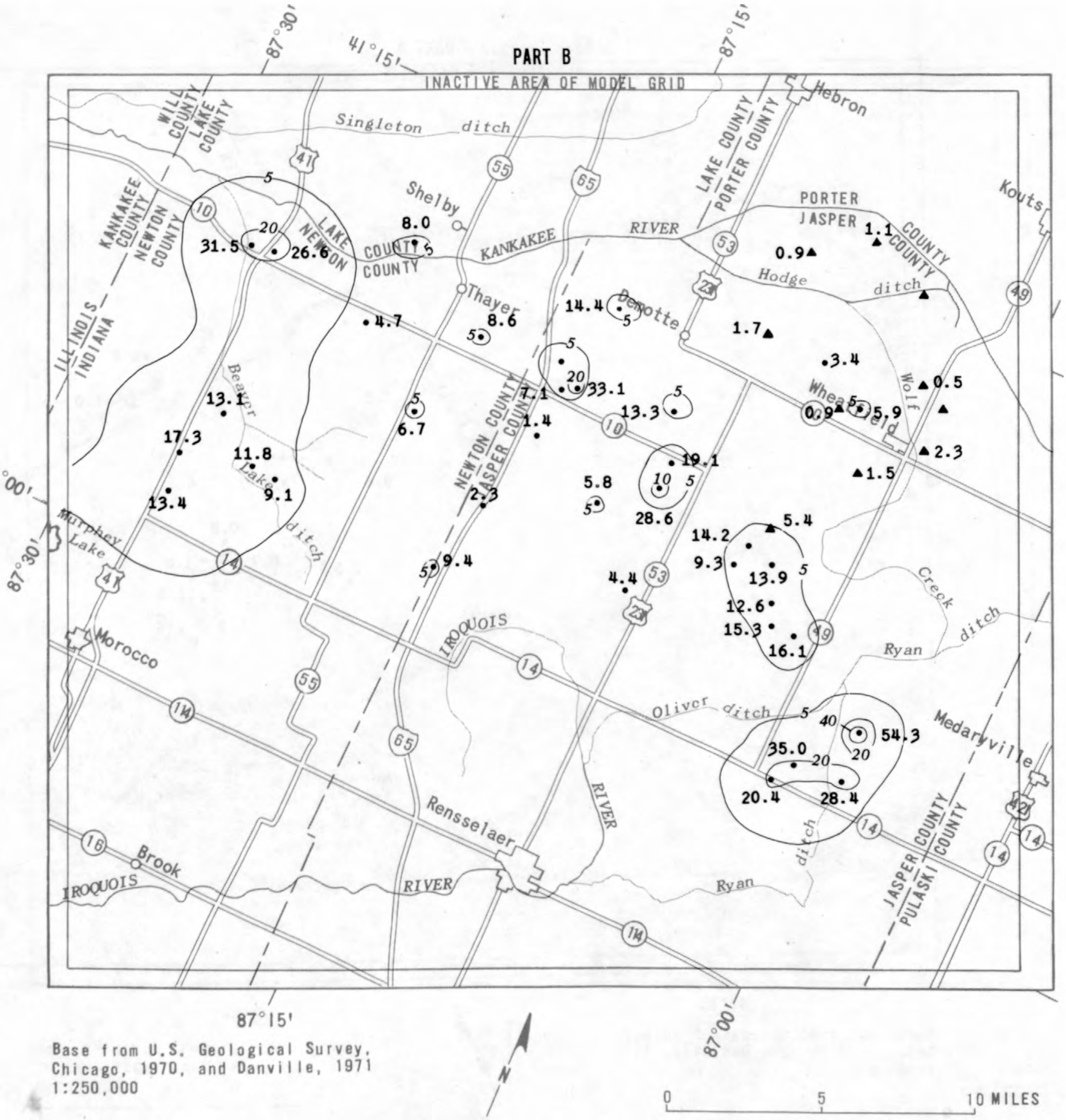
14.6• Water-level decline in simulated pumping well in layer 1, in feet. Declines less than 0.5 ft, not included

• Simulated pumping well in layer 1

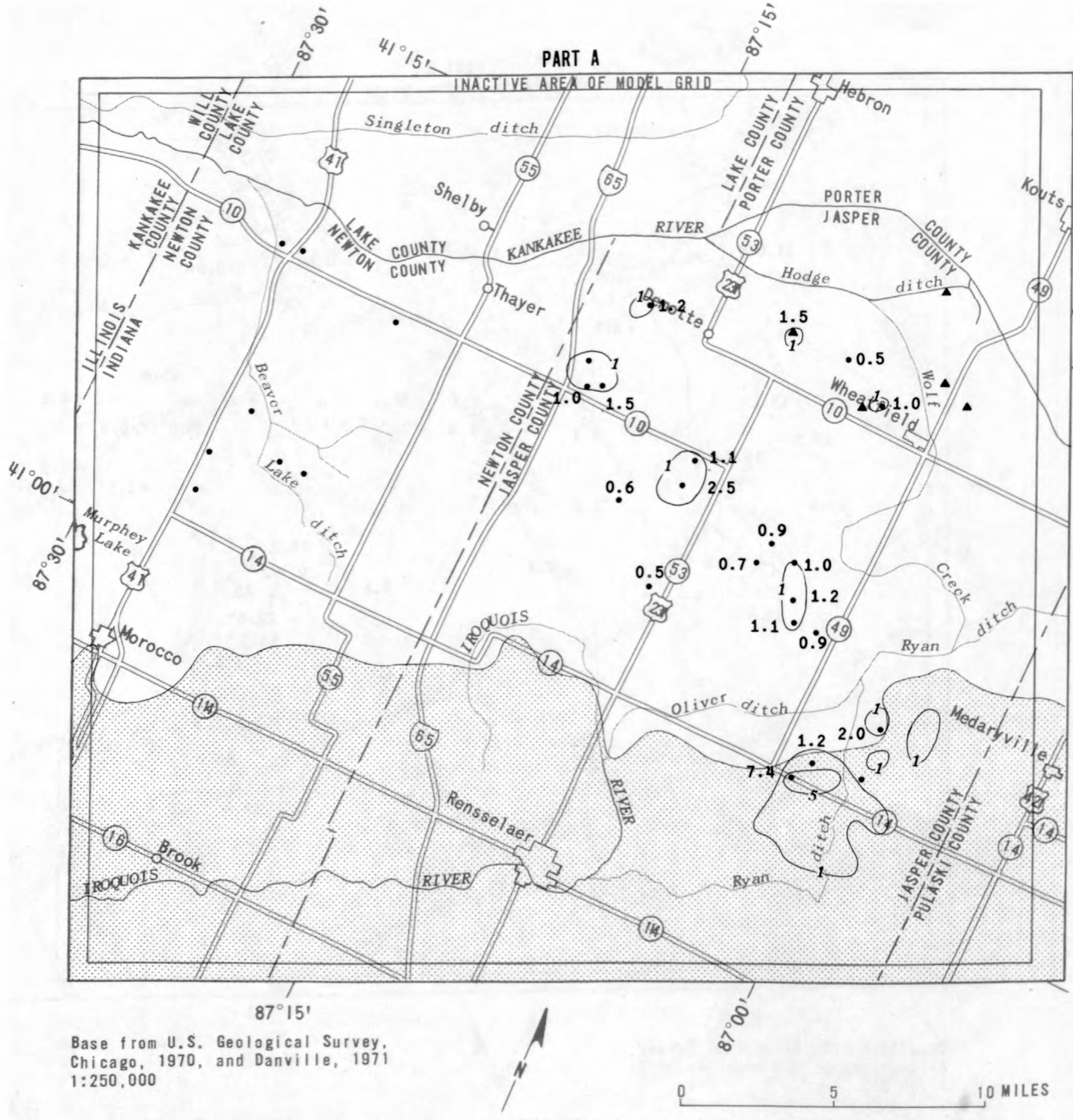
▲ Simulated pumping well in layer 2

Figure 29.-- Model-simulated water-level declines in model experiment K

PART B



in layer 1 (after irrigation): part A after 4 months in first year and part B after 4 months in fifth year.



Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971
1:250,000

EXPLANATION

Aquifer

Line of equal water-level decline in layer 2. Interval, in feet, is variable

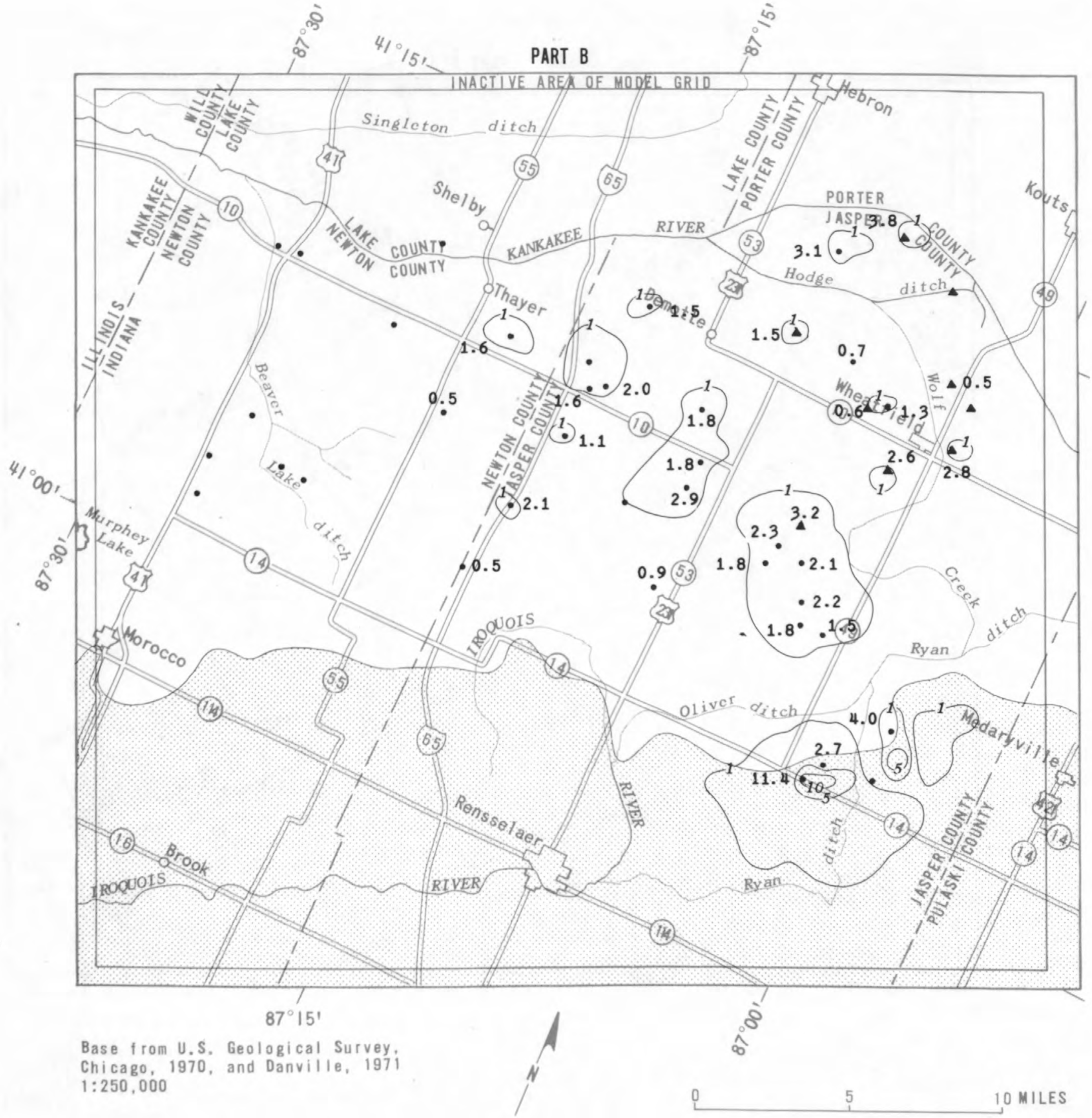
7.4 • Water-level decline in simulated pumping well in layer 2, in feet. Declines less than 0.5 ft, not included

• Simulated pumping well in layer 1

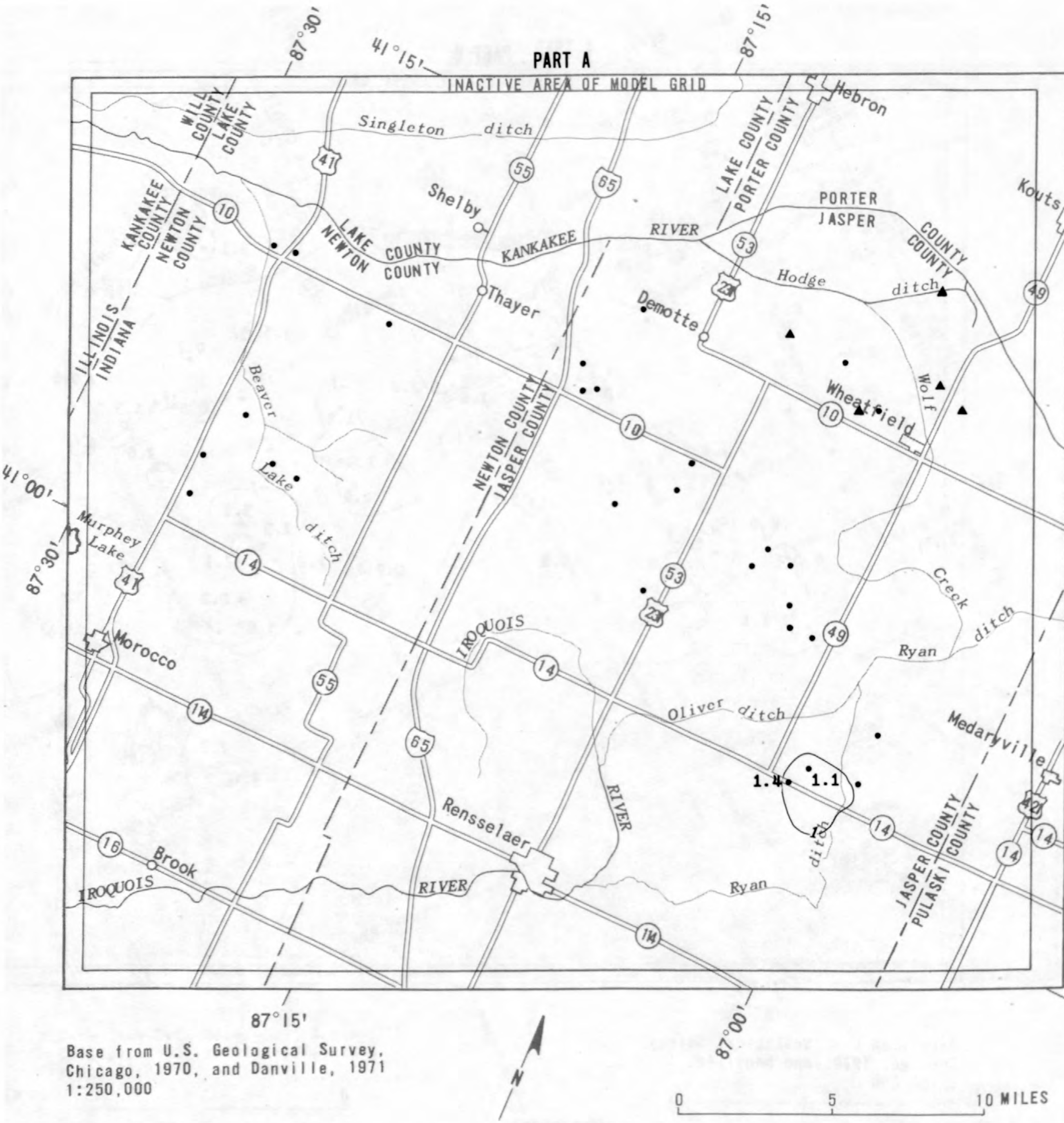
▲ Simulated pumping well in layer 2

Figure 30.-- Model-simulated water-level declines in model experiment K

PART B



in layer 2 (after irrigation): part A after 4 months in first year and part B after 4 months in fifth year.



Base from U.S. Geological Survey, Chicago, 1970, and Danville, 1971 1:250,000

EXPLANATION

—1— Line of equal water-level decline in layer 1, in feet. Interval 1 foot

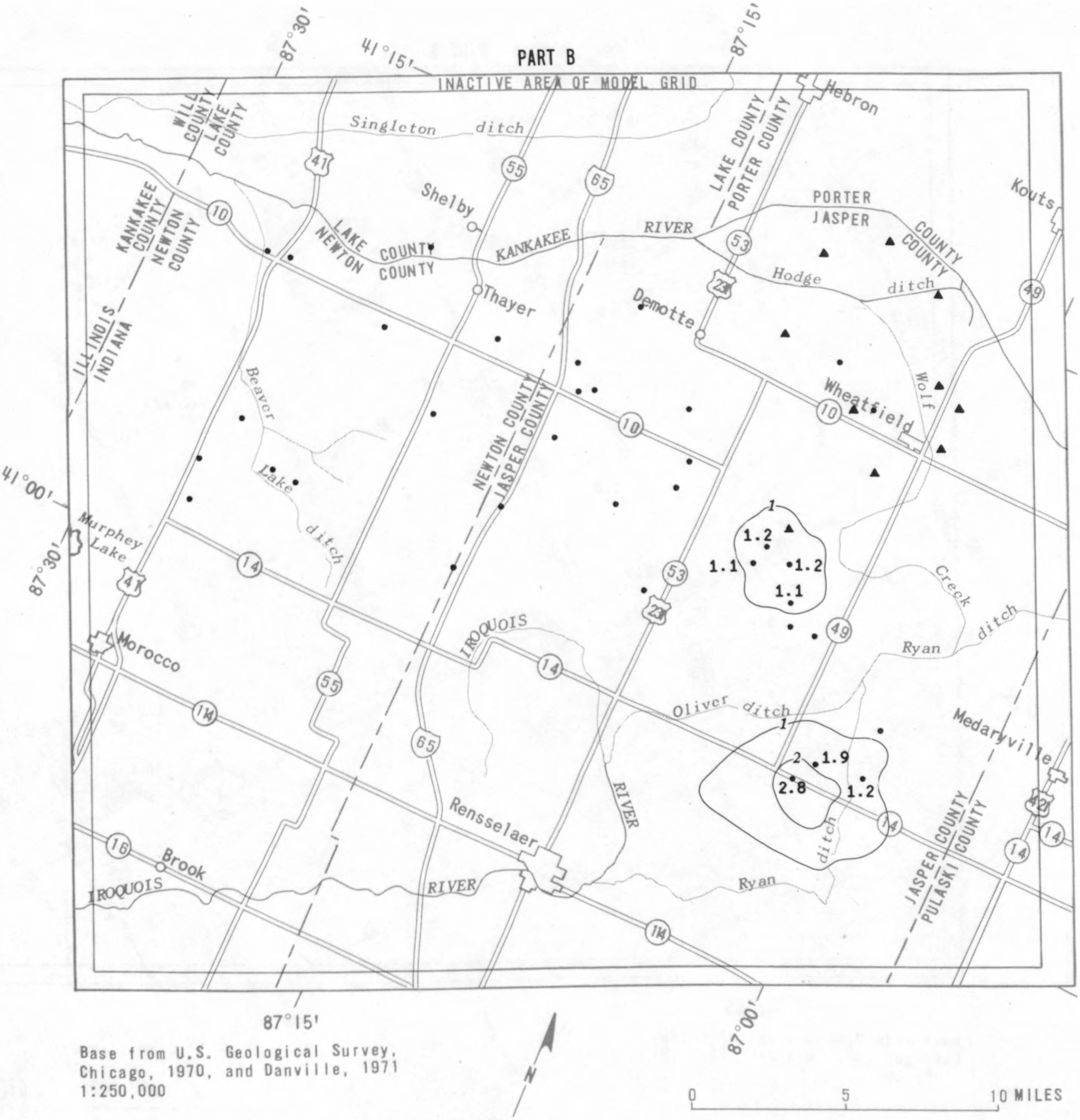
1.4 • Water-level decline in simulated pumping well in layer 1, in feet. Declines less than 0.5 ft, not included

• Simulated pumping well in layer 1

▲ Simulated pumping well in layer 2

Figure 31.-- Model-simulated water-level declines in model experiment K

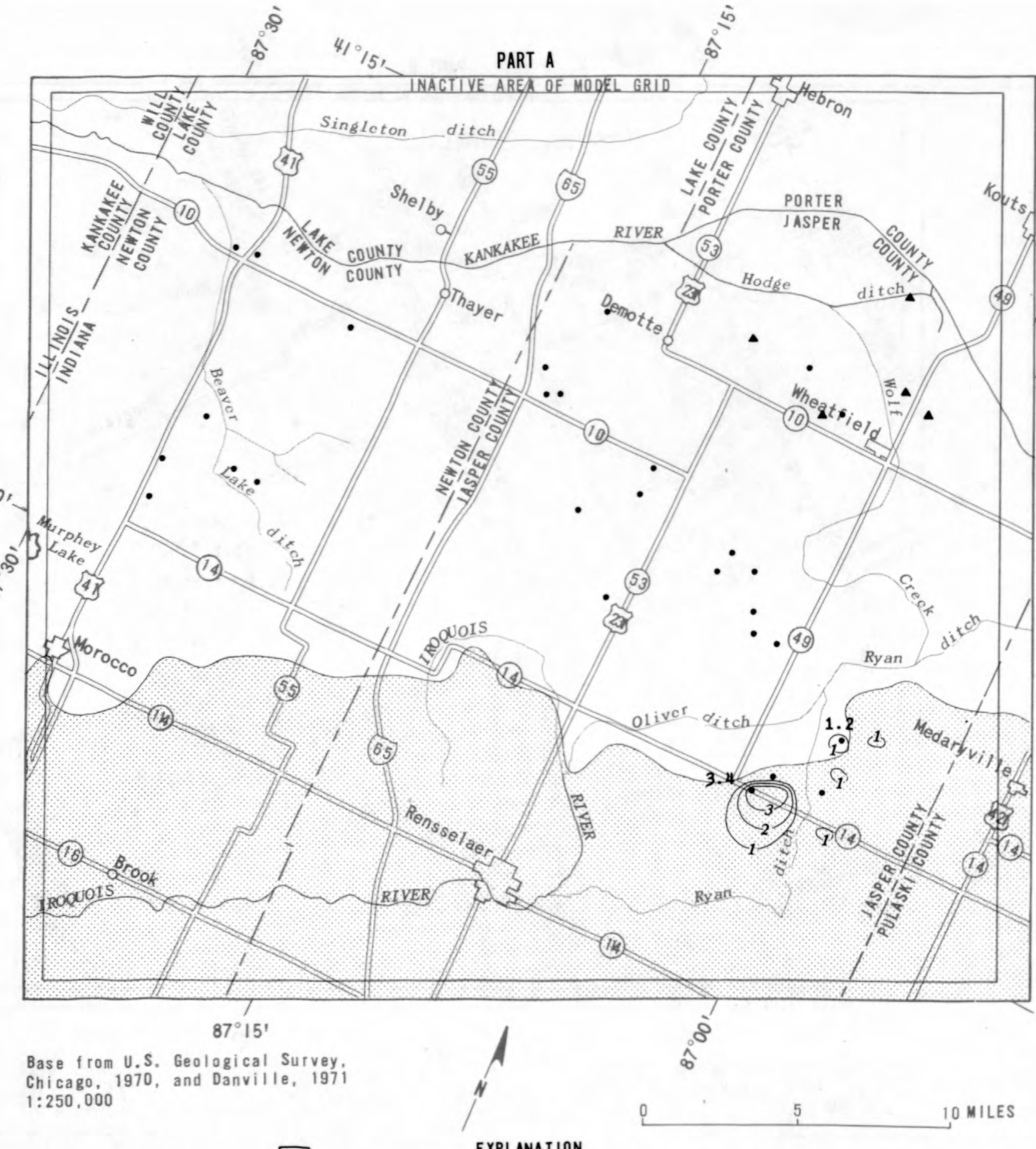
PART B



Base from U.S. Geological Survey,
Chicago, 1970, and Danville, 1971
1:250,000

0 5 10 MILES

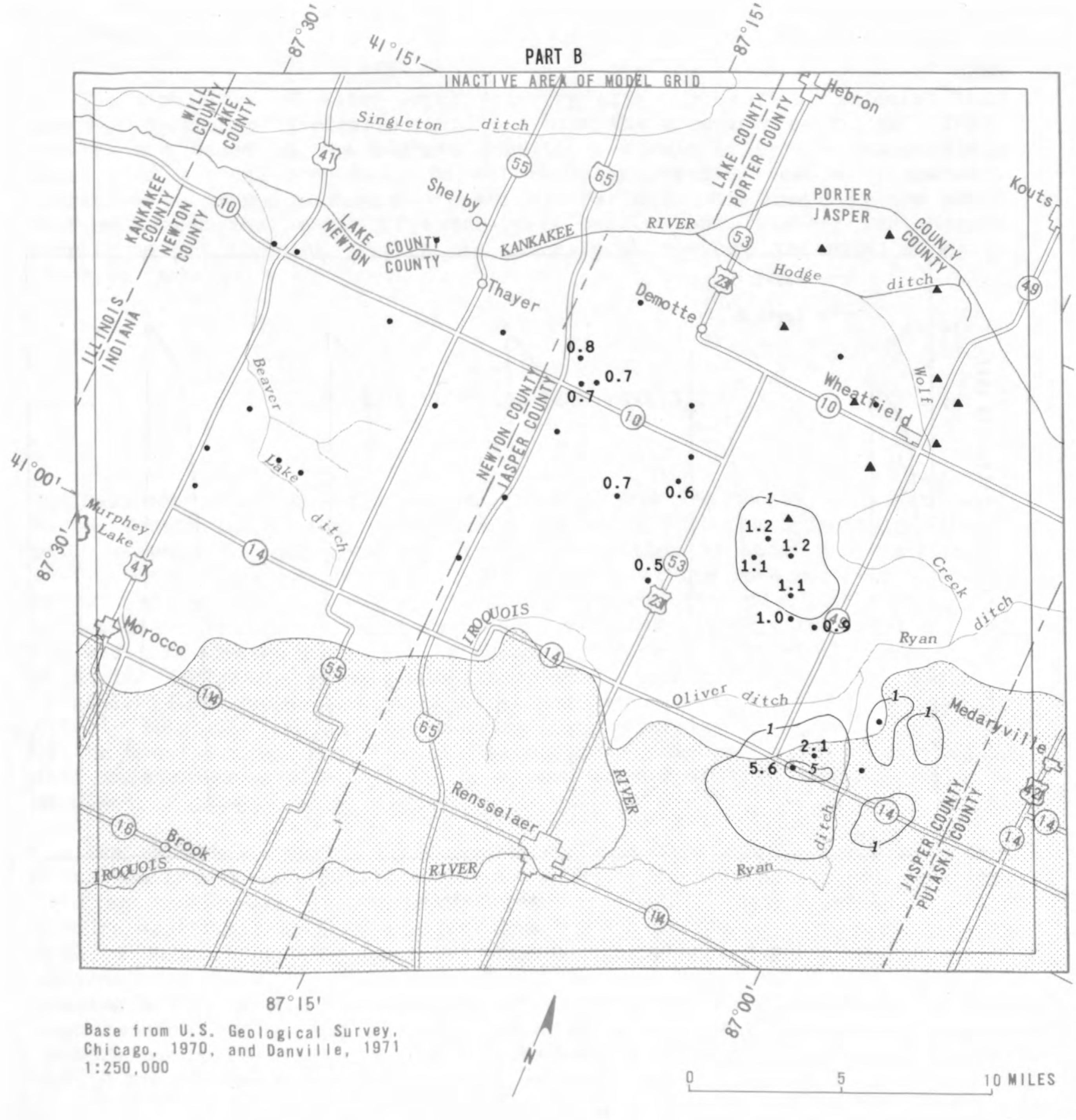
in layer 1 (after recovery): part A after first year and part B after fifth year.



EXPLANATION

- Aquifer
- —— 1 —— Line of equal water-level decline in layer 2. Interval 1 foot
- 3.4 • Water-level decline in simulated pumping well in layer 2, in feet
- • Simulated pumping well in layer 1
- ▲ Simulated pumping well in layer 2

Figure 32.-- Model-simulated water-level declines in model experiment K



in layer 2 (after recovery): part A after first year and part B after fifth year.

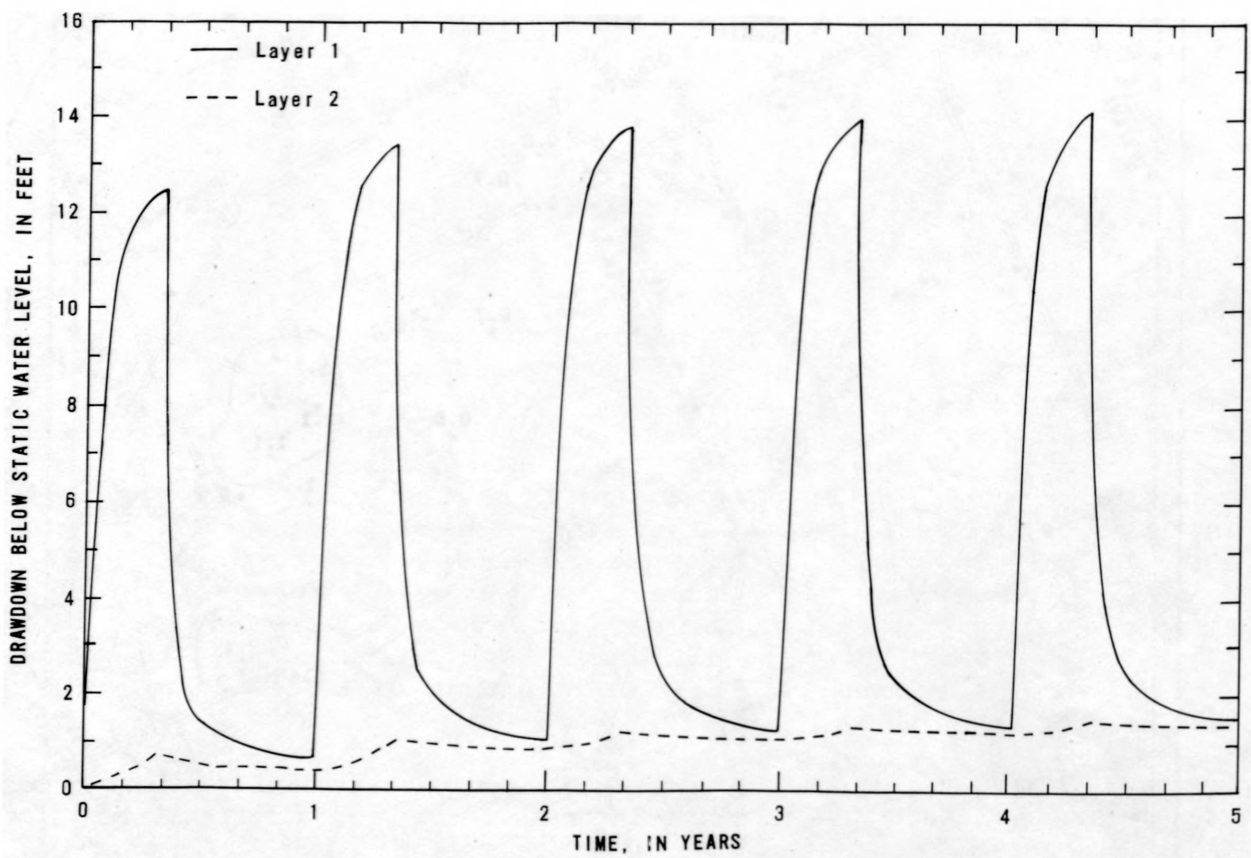


Figure 33.-- Five-year hydrographs for layers 1 and 2 at model node ''47, 48'', in model experiment K.

The simulation of water-level recovery after irrigation indicated that the 5-yr program of irrigation will not mine the ground-water system. These results are based on the average conditions simulated by the steady-state model, but average conditions do not normally prevail because of seasonal variations in recharge during dry and wet periods. Consequently, the model only estimates the amount of water-level decline and recovery for average conditions and does not predict the water-level response to annual fluctuations in recharge.

LIMITATIONS OF MODEL ANALYSIS

Development of irrigation pumping in east Kankakee County, Ill., and west Newton County Ind., along the State boundary affects the hydrologic system. Daily pumpage in this area during peak irrigation is about 6.7 Mgal. The irrigation is near or within the west boundary of the modeled area. Because of the proximity of the pumping to this boundary, the effect of the pumping cannot be accounted for in the model analysis. The predictive capability of the model is limited in scope. A further refinement of the model, incorporating the irrigation pumping along the west boundary of the model, would be helpful. This refinement would entail collection of additional hydrologic data to establish the aquifer geometry and properties along the west border of the modeled area. With the additional data, the model could be extended into east Kankakee County, Ill., and the effect of this pumping could be examined.

The ability of the model to match observed data during transient conditions has been investigated in certain areas. However, additional verification and refinement of the model would be useful. This refinement would require additional data on aquifer geometry and hydraulic characteristics as well as water-level declines resulting from pumping. Ongoing monitoring of irrigation pumping and its effect could be done by annual determination of pumping and by adding bedrock observation wells to the present observation-well network. These wells could be installed in areas near major pumping centers along the Illinois State boundary, in north-central Newton County, and in central Jasper County.

SUMMARY AND CONCLUSIONS

The aquifer system in Newton and Jasper Counties consists of three major aquifers: the unconfined Kankakee outwash aquifer, a confined carbonate-bedrock aquifer, and a confined bedrock valley outwash aquifer.

The Kankakee aquifer is primarily composed of coarse sand but contains small amounts of fine sand and gravel. The aquifer covers an area of 465 mi^2 in the north halves of Newton and Jasper Counties. The saturated thickness of the aquifer averages 40 ft and generally ranges from 5 to 60 ft. The transmissivity of the aquifer averages 10,000 ft^2/d and generally ranges from 250 to 15,000 ft^2/d. The specific yield of the aquifer is estimated to be 0.12.

The upper 100 ft or so of Silurian and Devonian age limestone and dolomite is the carbonate-bedrock aquifer. The highly variable transmissivity of the bedrock generally can range from 10 to 13,500 ft^2/d but generally ranges from 1,000 to 2,000 ft^2/d. The aquifer storage coefficient is estimated to be 0.00013.

A confined outwash aquifer, composed of coarse sand and gravel, partly fills a buried bedrock valley in south Newton and Jasper Counties. The aquifer thickness is 25 ft and generally ranges from 5 to 70 ft. The transmissivity of the aquifer generally ranges from 400 to 26,000 ft^2/d.

A fairly continuous semiconfining unit of till and lacustrine clay and silt, which separates the confined carbonate bedrock and bedrock valley outwash aquifers from the Kankakee outwash aquifer, generally ranges in thickness from 5 to 145 ft. The vertical hydraulic conductivity of the unit is approximately 0.0006 ft/d in north and north-central Newton County and 0.004 ft/d in the rest of the study area.

Irrigation is widespread in the north and north-central parts of the study area, where soils have developed on the surficial outwash deposits. The total irrigated area is approximately 6,200 acres (9.69 mi^2). The primary source of ground water used for irrigation is the carbonate bedrock aquifer, but the Kankakee aquifer is being increasingly used.

During the study period, irrigation pumping reached a peak in 1977. Total daily use of ground water for irrigation at the peak irrigation season was estimated to be 34.8 Mgal, 14.2 Mgal in Newton County and 20.6 Mgal in Jasper County.

A two-layered digital model was constructed to simulate flow in the three major aquifers at a steady-state condition. The model was calibrated to match ground-water levels and ground-water discharge to streams measured in June 1978. Steady-state analysis of the ground-water system indicates that 1.08 ft/yr recharge reaches the Kankakee water-table aquifer and that 528 ft^3/s discharges to the major streams and rivers. The average industrial and municipal pumpage from the aquifer system in 1978 was estimated to be 2.53 ft^3/s. Ground-water inflow and outflow at the model boundaries are 11.85 and 33.06 ft^3/s, respectively.

The effects of irrigation pumping on ground-water levels in the bedrock and Kankakee aquifers and on streamflow were simulated in a series of model experiments. Model simulation of a single pumping well in the bedrock aquifer under various hydrologic conditions indicates that major factors controlling water-level decline are the variations in thickness and in vertical hydraulic conductivity of the semiconfining beds. Irrigation pumping simulated in the bedrock aquifer, in an area of low transmissivity (1,000 ft^2/d) overlain by a thick sequence (50-60 ft) of semiconfining beds with low vertical hydraulic conductivity (0.0006 ft/d), produced drawdowns of 5 ft or more over an area of 13 mi^2 and drawdowns of 1 ft or more in an area of approximately 60 mi^2. In contrast, pumping simulated in an area of moderate bedrock transmissivity (1,340 ft/d) overlain by a thin (<10 ft) semiconfining bed with higher vertical hydraulic conductivity (0.004 ft/d) resulted in drawdowns of 1 ft or more covering an area of 0.4 mi^2.

In two experiments, identical pumpage was used in two saturated thicknesses of the Kankakee aquifer. The area of drawdowns of 1 ft or more caused by pumping in areas of large (50-ft) and small (25-ft) saturated thickness was approximately 1.3 mi^2 in both experiments. Drawdown in the vicinity of the pumping well varied slightly.

Drawdowns caused by pumping in the Kankakee aquifer near streams were insignificant because the Kankakee aquifer and streams are well connected. Because flows in the major streams are large, reduction in flow is not significantly affected by nearby simulated wells producing 12.6 Mgal/d in the bedrock aquifer and 10.3 Mgal/d in the Kankakee aquifer. Pumping in the bedrock aquifer would produce large drawdowns before flow is significantly reduced. Simulated wells in the bedrock aquifer along Beaver Lake ditch produced regional drawdowns ranging from 20 to 35 ft and reduced the flow of Beaver Lake ditch by only about 6 percent.

In a base-flow period, streamflow on the Kankakee River could sustain a significant amount of irrigation development, although large-scale development might deplete streamflow on tributaries of the Kankakee River. Under steady-state conditions, irrigation pumping of 112 Mgal/d would reduce the streamflow at the 99.9-percent flow duration (225 Mgal/d), measured at the gaging station at Shelby, by only 50 percent.

Simulation of water-level recovery after irrigation pumping indicated that a 5-yr period of alternating between increasing pumping and recovery will not cause ground-water mining.

Additional data for expanding the modeled area would have to be collected before the large-scale irrigation development in west Newton County, Ind., and east Kankakee County, Ill., could be assessed.

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