

WATER RESOURCES OF WILDCAT CREEK AND DEER CREEK BASINS,
HOWARD AND PARTS OF ADJACENT COUNTIES, INDIANA, 1979-82

By Barry S. Smith, Mark A. Hardy, and E. James Crompton

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

<u>Multiply inch-pound unit</u>	<u>By</u>	<u>To obtain Metric units</u>
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
square foot (ft ²)	0.0929	square meter (m ²)
foot per day (ft/d)	0.3048	meter per day (m/d)
square foot per day (ft ² /d)	0.929	square meter per day (m ² /d)
gallon per day per square foot [(gal/d)/ft ²]	0.0410	meter per day (m/d)
mile (mi)	1.609	kilometer (km)
square mile (mi ²)	2.590	square kilometer (km ²)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
cubic foot per second (ft ³ /s)	0.0283	cubic meter per second (m ³ /s)
cubic foot per minute (ft ³ /min)	1.698	cubic meter per minute (m ³ /min)
gallon per day (gal/d)	4.38 x 10 ⁻⁸	cubic meter per second (m ³ /s)
million gallons per day (Mgal/d)	0.4381	cubic meter per second (m ³ /s)
million gallons per year (Mgal/yr)	0.4381	cubic meter per year (m ³ /yr)
cubic foot per day per foot [(ft ³ /d)/ft]	0.929	cubic meter per day per meter [(m ³ /d)/m]
gallon per day per square mile [(gal/d)/mi ²]	9.82x10 ⁻²	cubic meter per day per square kilometer [(m ³ /d)/km ²]
cubic foot per second per square mile [(ft ³ /s)/mi ²]	0.01093	cubic meter per second per square kilometer [(m ³ /s)/km ²]
micromho per centimeter at 25° C (μmho/cm)	1.000	microsiemens per centimeter at 25° C (μS/cm)

To convert degree Fahrenheit (°F) to degree Celsius (°C)

$$^{\circ}\text{C} = (5/9) (^{\circ}\text{F} - 32)$$

ABBREVIATIONS AND SYMBOLS

d	day
°C	degree Celsius
μmho/cm	micromho per centimeter
μS/cm	microsiemens per centimeter
cm	centimeter
mV	millivolt
CaCO ₃	calcium carbonate

WATER RESOURCES OF WILDCAT CREEK AND DEER CREEK BASINS,
HOWARD AND PARTS OF ADJACENT COUNTIES, INDIANA, 1979-82

By Barry S. Smith, Mark A. Hardy, and E. James Crompton

ABSTRACT

The water resources of Wildcat Creek and Deer Creek basins were evaluated by (1) describing the streamflow, (2) defining the geometry and the hydraulic characteristics of aquifers and semiconfining beds, (3) investigating the hydraulic connection between aquifers and major streams, (4) describing ground-water and surface-water quality, and (5) simulating the effects of large-scale pumping on ground-water levels and streamflow.

The area of study is 505 mi² (square miles) of flat to rolling glacial plain. The plain is modified by stream erosion, particularly along Wildcat and Deer Creeks, which are incised. Glacial deposits of the plain, predominantly silty till, range in thickness from zero at bedrock quarries along Wildcat Creek to 200 ft (feet) in buried bedrock valleys. Devonian and Silurian limestone, dolomite, and dolomitic siltstone underlie the glacial deposits.

Seasonal low flows of Wildcat Creek downstream from the mouth of Mud Creek exceeded 0.3 cubic foot per second per square mile, except in and near Kokomo, Indiana. The reaches of Wildcat Creek in and near Kokomo are affected by diversion of surface water, pumping of ground water, storage in reservoirs, treated sewage, and low-head dams.

Thirty-three percent of the 7.1 million gallons per day pumped from the aquifers within the basins was used by Kokomo. A majority of the commercial, industrial, and municipal wells are in and near Kokomo. More than 90 percent of the pumpage was from the bedrock.

Flow in the bedrock aquifer is predominantly through open fractures, joints, bedding planes, and solution channels above the siltstone in the Silurian Mississinewa Shale. Regional transmissivities of the bedrock obtained by ground-water-flow modeling, were 1,250 square feet per day in some areas and 6,250 square feet per day in other areas. Locally, however, the transmissivities may vary considerably from these values. Water levels in the bedrock generally fluctuate plus or minus 2 ft in response to seasonal or periodic changes in recharge, but the trend in water levels is toward steady state. Discontinuous sand and gravel aquifers are interspersed within semiconfining beds of till. Thickness of the sand and gravel aquifers generally ranges from 5 to 40 ft. Horizontal flow predominates in the aquifers because of the contrast in hydraulic conductivity of the sand and gravel aquifers [200 ft/d (feet per day) based on specific-capacity data and flow modeling] and of the semiconfining beds (from 5×10^{-4} to 1×10^{-2} ft/d based on flow modeling). Effective recharge averaging 2.6 inches per year is

attributed to leakage through the semiconfining beds. Hydraulic conductivity of the streambeds, assumed to be 1 ft thick, generally ranged from 2×10^{-2} to 20 ft/d.

Hardness of ground water ranged from 190 to 540 mg/L (milligrams per liter) as calcium carbonate. Types of water included calcium bicarbonate and calcium and magnesium bicarbonate. Concentrations of iron and manganese commonly exceeded 0.3 and 0.05 mg/L. The unconsolidated aquifers had higher mean concentrations of silica and iron and a lower mean concentration of potassium than the limestone aquifer. Ammonium concentration exceeded 0.5 mg/L as nitrogen at many ground-water-sampling sites and reached 22 mg/L as nitrogen at site 15A, July 23, 1981. Possible sources of ammonium and other nutrients in ground water include livestock wastes and fertilizer applied to fields.

Average results of analyses of 10 water samples collected at Indiana State Board of Health stations WC69 and WC63 upstream and downstream from Kokomo on Wildcat Creek from March 10 through December 13, 1982, include: hardness of water, 242 and 307 mg/L; concentration of iron, 1.07 and 1.53 mg/L; concentration of manganese, 0.09 and 0.12 mg/L; and concentration of ammonia, 0.18 and 0.53 mg/L. Average concentration of most constituents was higher at the downstream site than at the upstream site.

A digital flow model was constructed to simulate ground-water flow in the area. Hypothetical well fields were pumped at selected nodes of the model (0.57 mi^2) until water levels declined about 20 ft. Yields from the hypothetical well fields ranged from 1.5 to $4.0 \text{ ft}^3/\text{s}$ (cubic feet per second) but in the typical well field ranged from 2.0 to $2.5 \text{ ft}^3/\text{s}$. Virtual steady state was attained in about 4 years. Reductions in streamflow for selected reaches were generally less than 50 percent of the seasonal low flow where one hypothetical well field was nearby. Compared to 7-day, 10-year low flows, however, reductions in streamflow caused by the typical well field would be great in most reaches.

The combined rate of pumping for well-fields H, I, J, K, L, and M, in the lower sand and gravel and in the bedrock, totaled $13.0 \text{ ft}^3/\text{s}$. Mutual interference of wells was kept to a minimum. Larger concentrations of pumping, as that of wells E, F, and G, which were pumped at $9.0 \text{ ft}^3/\text{s}$ from the middle sand and gravel aquifer, resulted in a greater degree of mutual interference. Regional availability of water at any point in the area of study, therefore, will depend, in large part, on the pumping rates, the degree of mutual interference, the drawdown that is acceptable, and the reduction in flow of nearby streams that is acceptable, as well as characteristics of aquifers.

INTRODUCTION

Purpose and Scope

The reservoir on Wildcat Creek upstream from Kokomo was inadequate as a supplemental water supply for Kokomo during the winter of 1975-76. The city and the Indiana Department of Natural Resources were interested in investigating additional water resources for augmenting the city's water supply. The U.S. Geological Survey, in cooperation with the Indiana Department of Natural Resources, began a study in 1979 to evaluate the water resources of Wildcat Creek and Deer Creek basins in Howard and parts of adjacent counties in north-central Indiana (fig. 1).

The purposes of the study were to (1) describe stream-flow, (2) define the geometry and the hydraulic characteristics of aquifers and semiconfining beds, (3) investigate the hydraulic connection between aquifers and major streams, (4) describe ground-water and surface-water quality, and (5) use a three-dimensional, ground-water flow model to simulate the effects of large-scale pumping on ground-water levels and streamflow in the Wildcat Creek and Deer Creek basins. The report describes the water resources of the area and includes discussions of streamflow, ground-water flow, geology, aquifer properties, interaction between surface water and ground water, and water quality. A digital model is used to predict the effects of ground-water withdrawals on drawdown and streamflow.

Methods of Study

Surface-water resources were evaluated by analysis of flow-duration curves for four long-term gaging stations of the Geological Survey in and downstream from the area of study, by measurements of base flow at 35 sites, and by chemical analyses of water samples collected at two Indiana State Board of Health sampling stations--one downstream from and one upstream from Kokomo on Wildcat Creek. Measurements of base flow were used in estimating discharges from drainage areas. Gains and losses determined from the discharges were used to compare sustained low flows for the 35 reaches of streams defined by the measuring sites. Gains and losses were also used to calibrate a ground-water flow model. Surface-water quality is based on data published by Indiana State Board of Health (1982) and the Geological Survey (Crawford and others, 1979).

Ground-water resources were evaluated by analysis of lithologic logs, seismic-refraction data, gamma logs, water-level measurements, aquifer tests, specific-capacity tests, and chemical analyses of water. Ninety-seven test holes were drilled with an auger rig, and 13 were drilled by the mud-rotary method. Ninety-three observation wells were installed in the test holes. Gamma logs for most of the wells were recorded by the Indiana Geological



Figure 1.-- Location of area of study.

Survey. Seismic-refraction data for use in locating buried bedrock surfaces were obtained from the Indiana Geological Survey. Seismic refraction surveys for defining the water table in the till plain were done by the U.S. Geological Survey. A few split-spoon and Shelby-tube samples from two drilling sites and seven streambed sites were collected by the U.S. Geological Survey and were analyzed by the Indiana Geological Survey.

Ground-water samples for chemical analysis were collected from 30 wells. Samples were collected as close as possible to the well casing to avoid the sample's contacting pipes used for treatment and pumping. Before samples were collected, water was pumped from the wells until temperature, specific conductance, and pH became stable. After they were filtered, the samples were preserved by the methods described in Skougstad and others (1979). Chemical analysis of the samples included determination of concentrations of dissolved major ions, nutrients, and trace elements.

Aquifer maps were drawn; hydraulic characteristics of aquifers and semi-confining beds were estimated; and a ground-water flow model was constructed, calibrated, and used to simulate the effects of hypothetical pumping from the aquifer.

Location and Setting

Wildcat Creek and Deer Creek are tributaries to the Wabash River. The upper parts of the Wildcat and Deer Creek basins (fig. 2) drain an area of 505 mi² centered in Howard County and including parts of Cass, Clinton, Grant, Madison, Miami, and Tipton Counties. The basins are within the Tipton till plain of Indiana (Malott, 1922, p. 104), flat to gently rolling glacial plains slightly modified by stream erosion. Valleys incised in the plain by Wildcat and Deer Creeks are the most prominent topographic features in Howard and adjacent counties.

The climate of the study area is temperate. Mean annual temperature was 52.5° F and average annual precipitation was 37.9 in/yr from 1941 to 1970 at the Kokomo weather station 7 miles southeast of Kokomo (National Oceanic and Atmospheric Administration, 1973). Average discharge of Wildcat Creek at Kokomo was 230 ft³/s or 12.9 (in/yr)/mi² of area drained from 1955 through 1982 (U.S. Geological Survey, 1983, p. 96), and average discharge of Deer Creek near Delphi, 17 miles downstream from Kokomo, was 241 ft³/s or 11.9 (in/yr)/mi² from 1943 through 1982 (U.S. Geological Survey, 1983, p. 88).

Kokomo, whose population was 47,808 in 1980 (Bureau of Census, 1982, p. 16-23), is the only major urban center in the multicounty area. Industries in Kokomo manufacture electronic parts, transmissions, radios, nails, steel wire, fence, rods, and high-performance alloys. Agriculture is the principal use of land in Howard and adjacent counties, and Kokomo is a distribution center for agricultural products.

Kokomo is the major user of water in the area of study. Water is stored in reservoirs owned by the Indiana American Water Co. 3 miles east of Kokomo.

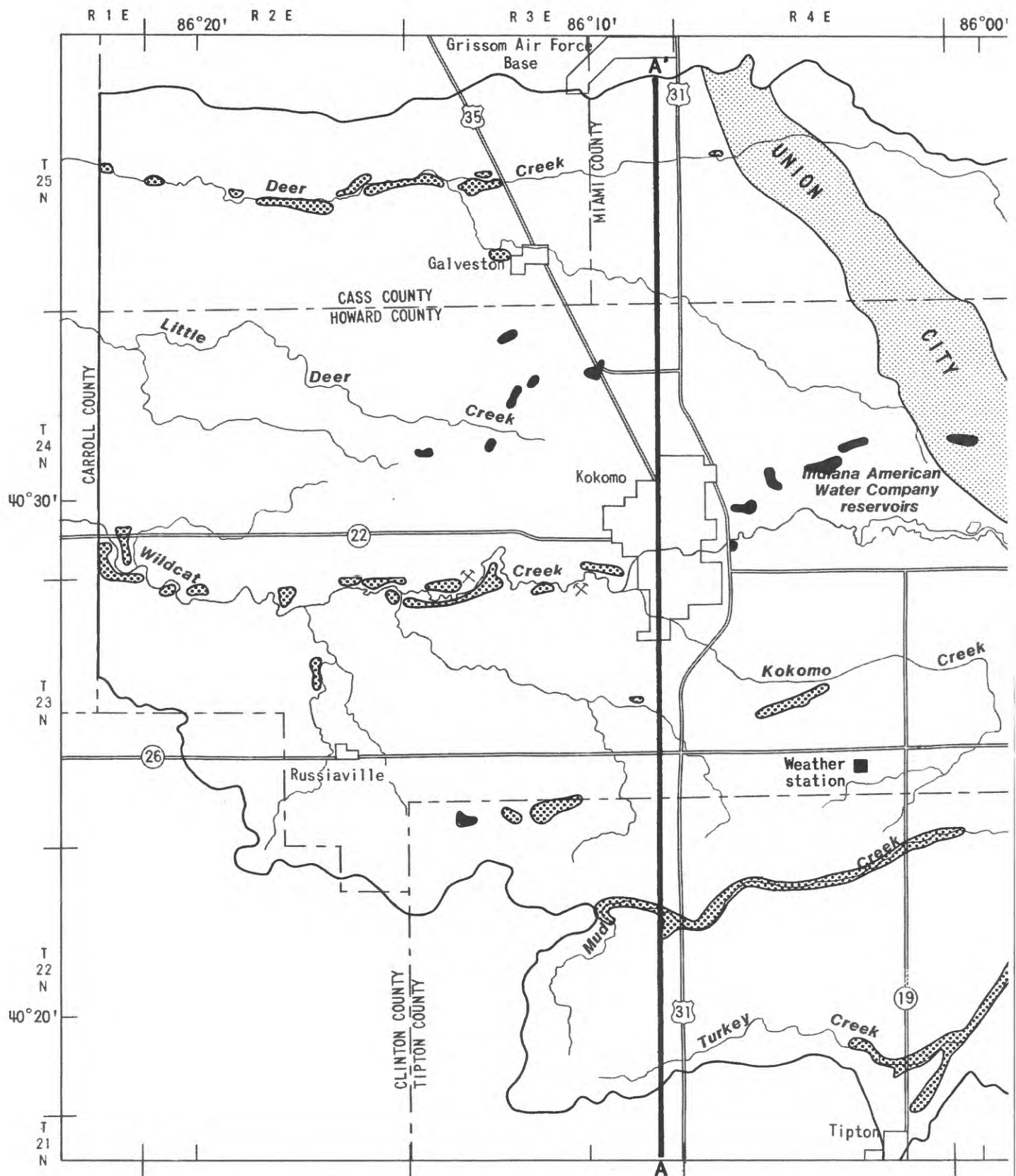
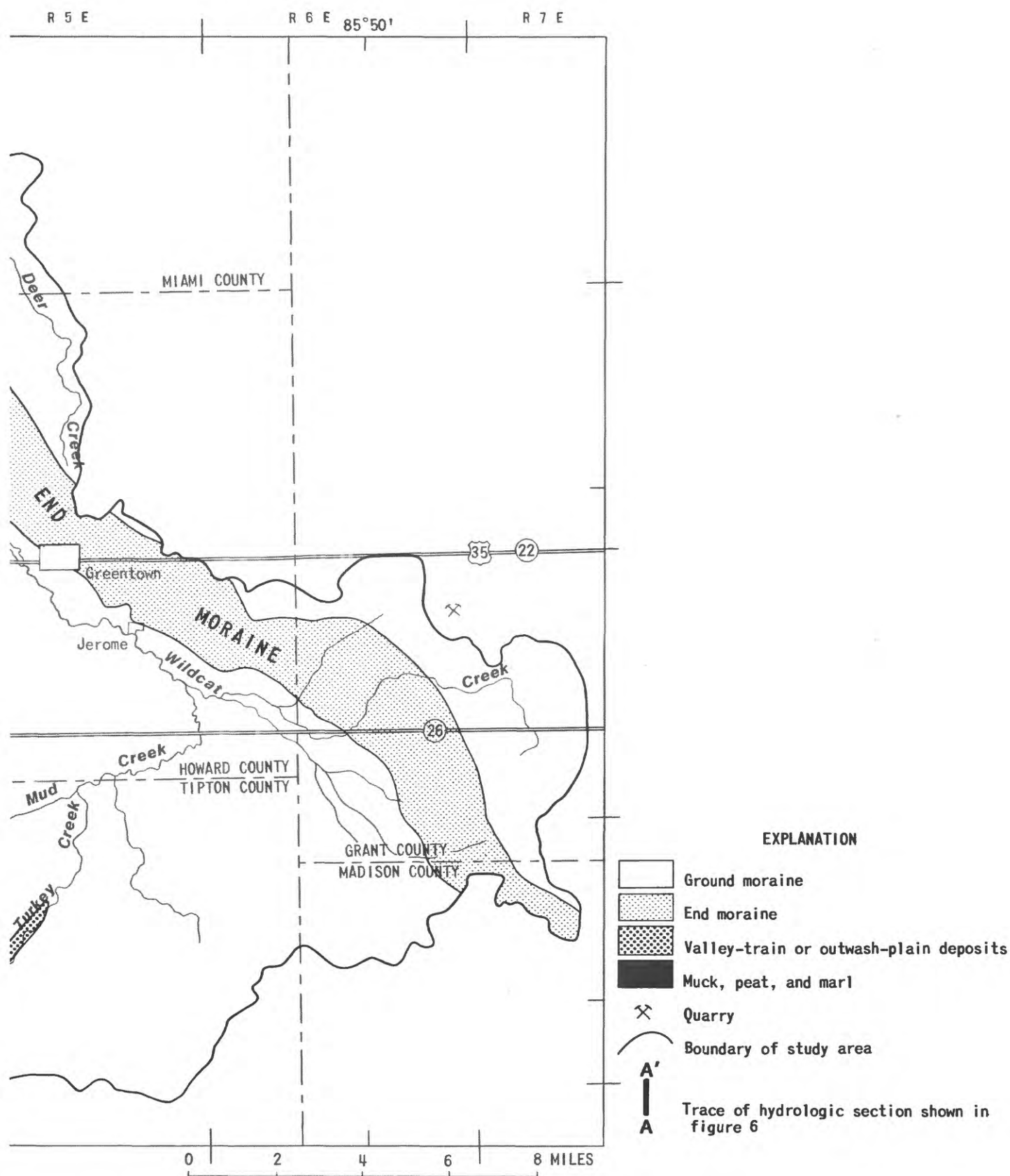


Figure 2.-- Surficial geology.



Surficial geology from W. J. Wayne, G. H. Johnson, S. J. Keller (1966, Part B),
A. M. Burger, J. L. Forsyth, R. S. Nicoll, and W. J. Wayne (1971, Part B).

The water is diverted from Wildcat Creek near State Highway 31. The surface-water supply is supplemented by several well fields, but, before 1981, ground water was a small percentage of water used. Since then, new wells have been installed, but Wildcat Creek remains the principal source of water for Kokomo.

Geology

The surficial geology of north-central Indiana including Howard and adjacent counties in the Tipton till plain was compiled by Wayne and others (1966, part B) and Burger and others (1971, part B). The plain is covered by Wisconsin drift, which includes deposits laid down by melting ice, streams, and ice-dammed lakes (fig. 2). Till--a mixture of unsorted and unstratified boulders, sand, gravel, silt, and clay in a sand to silt matrix--is predominant. Surface of the land is flat in most of the area but is hilly in the Union City moraine that stretches from northwest Miami County to northwest Madison County (fig. 2). The glacial deposits range in thickness from zero in quarries where bedrock is exposed along Wildcat Creek west of Kokomo to more than 200 ft in buried valleys that were eroded in bedrock. More commonly, however, thickness of the deposit ranges from 50 to 150 ft (Marie and Davis, 1974, sheet 2).

The greatest topographic relief is that resulting from incision by streams, particularly along Wildcat and Deer Creeks. Valleys of all the streams are filled with alluvium (silt, sand, and gravel), and some valley-train deposits (predominantly sand and gravel) washed from the receding glaciers of the Wisconsin stage have been mapped along Wildcat and Deer Creeks. The alluvium and valley-train deposits at the surface are generally thin, interspersed with silt, and local in extent. Buried deposits of sand and gravel interspersed in the till, however, are thicker and more extensive than the surficial deposits.

The geology of bedrock in north-central Indiana, including the area of study (fig. 3), was compiled by Wayne and others (1966, part B) and Burger and others (1971, part B). The area of study is near the axis of the Cincinnati arch (Doheney and others, 1975, p. 3), and the bedrock units dip slightly southwest. Devonian limestone and dolomite (Muscatatuck Group, Upper Silurian limestone and dolomite (Salina Formation), and Middle and Lower Silurian limestone, dolomite, and dolomitic siltstone (Wabash formation) underlie glacial deposits. (The stratigraphic nomenclature follows the usage of the Indiana Geological Survey and is not the usage of the U.S. Geological Survey.)

A core of the bedrock at Kokomo collected by the Indiana Geological Survey consisted of 126 ft of cherty and dolomitic limestone near the surface, 20 ft of dolomitic siltstone below the limestone, and 260 ft of limestone and dolomite in the bottom section of the core (Shaver, 1961, p. 13). The limestone near the surface is fractured, creviced, and opened by weathering and solution channeling along joints and bedding planes (Watkins and Rosenshein, 1963, p. B8 and B9). The Mississinewa Shale is a fine-grained, dolomitic siltstone and limestone that is argillaceous and silty (Shaver, 1961, p. 15).

The Silurian rocks contain carbonate reef deposits. A few of these deposits are just outside the study area (Ault and others, 1976). The Silurian rocks, therefore, may contain deposits of open vuggy reef, although none have been documented in the area of study.

Topography of the bedrock (fig. 3) was determined from lithologic descriptions in drillers' records and from seismic-refraction surveys done by the Indiana Geological Survey. The predominant feature of the surface of the bedrock is a valley system cut by streams flowing from east to west (fig. 2). The main channels have steep valley walls that are deepest toward the west.

Previous Studies

The ground-water resources in and near Grissom (Bunker Hill) Air Force Base, a 35-square mile area adjacent to the north-central boundary of the area of study, were reported by Watkins and Rosenshein (1963). Ground-water geology and hydrology, results of aquifer tests and flow-net analyses, and quality of water were discussed in the report. The ground-water resources of part of Tipton County, immediately south of Howard County (figs. 1 and 2), were studied by Arihood (1982), and those of Madison County, immediately southeast, were studied by Lapham (1981). The reports contain maps and discussions of bedrock and sand and gravel aquifers, semi-confining beds, water levels, gains and losses of streamflow, use of water, and ground-water flow analyses by digital models. Water-quality data of Wildcat Creek in and near Kokomo are presented in Crawford and others (1979) and Indiana State Board of Health (1982, p. 103-104).

The current area of study is part of the middle Wabash River basin in Indiana that was studied by Marie and Davis (1974). A regional analysis of availability, use, and quality of water were described on three large sheets. Parts of the study area were also included in two county reports--on Tipton County (Steen, 1968) and Grant County (Heckard, 1968). Availability of ground water is emphasized in the county reports.

Acknowledgments

The authors gratefully acknowledge the services and information provided by many individuals and groups during the field work for the report. Ortman Drilling, Inc¹., Kokomo, Ind., provided storage, gravel, and water for drilling; access to numerous well logs; and information on local hydrogeology. The Indiana American Water Co. provided water and storage for drilling, as well as information on use of water. More than 300 private and public landowners

¹Any use of trade names is for descriptive purposes only and does not imply endorsement by the U.S. Geological Survey.

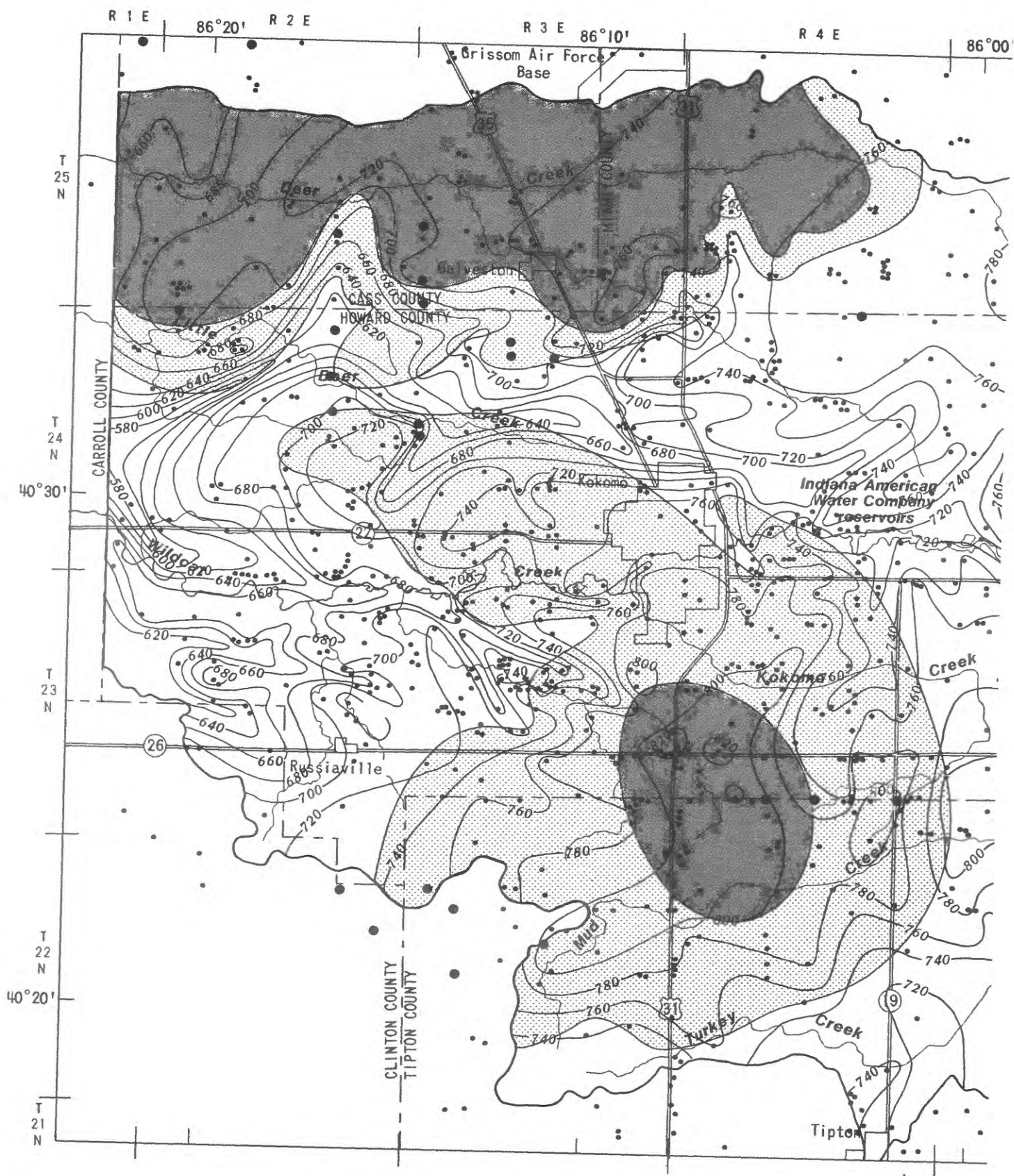
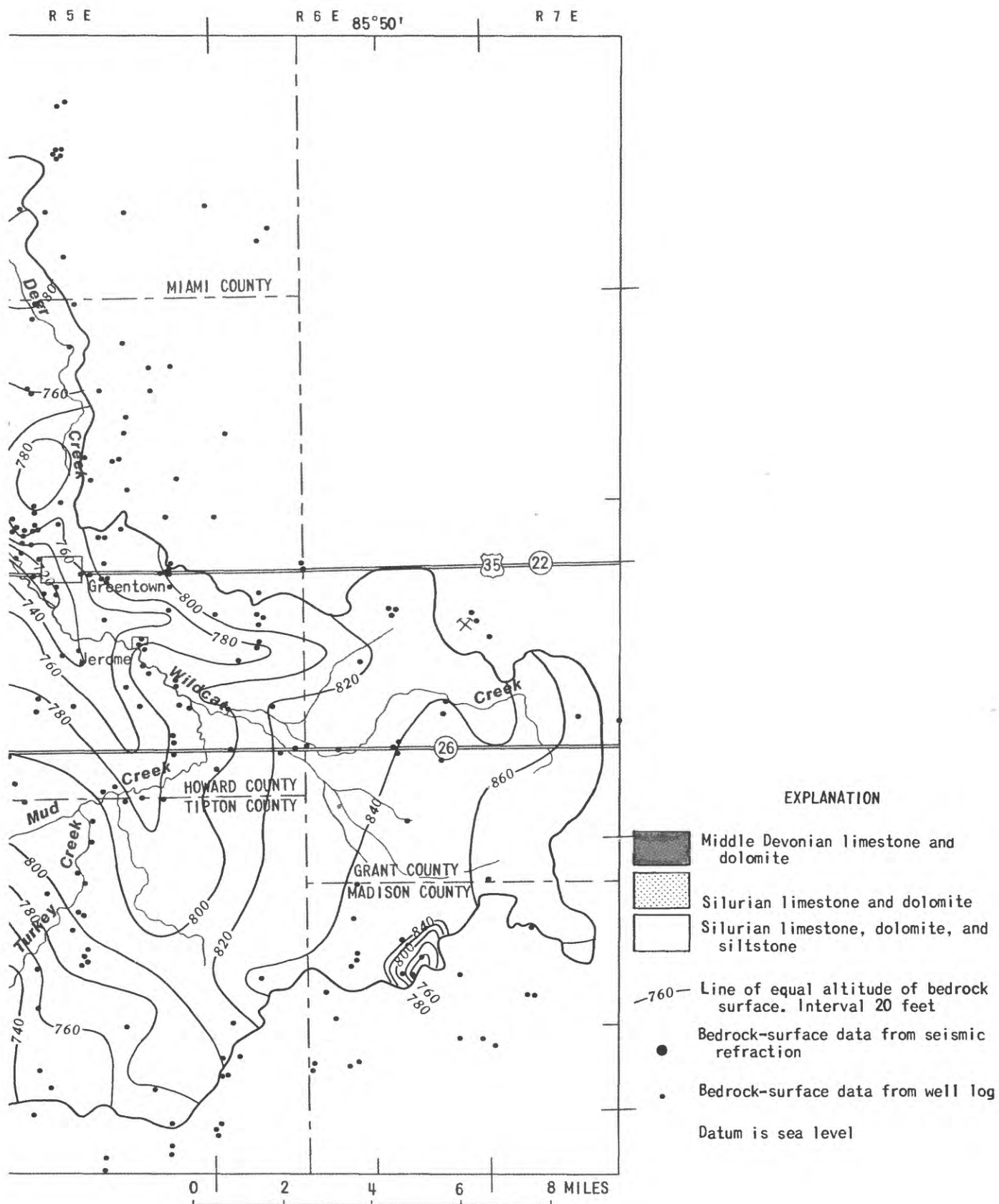


Figure 3.-- Geology and topography of bedrock.



Bedrock geology from W. J. Wayne, G. H. Johnson, and S. J. Keller (1966, Part A), and A. M. Burger, J. L. Forsyth, R. S. Nicoll, and W. J. Wayne (1971, Part A).

allowed access to water wells so that water levels could be measured and water samples could be collected, and 25 private and public organizations provided information on ground-water pumping and use of water.

Employees of three agencies of the State of Indiana provided goods, services, and cooperation essential to the study. Ned Bleuer of the Indiana Geological Survey supplied gamma-log surveys of observation wells, Joe Whaley of the Indiana Geological Survey made gravity and seismic surveys, which aided in the interpretation of bedrock topography, and Sam Frushour of the Indiana Geological Survey made permeameter and seive analyses on split-spoon and Shelby-tube samples. The Indiana Department of Natural Resources, Division of Water, provided copies of numerous well logs, and the Indiana Department of Highways allowed installation of observation wells on public right-of-ways.

Agencies of Kokomo, Ind., provided much information for the study. Thomas High of the Kokomo wastewater treatment plant provided information on treated sewage and maintained records of precipitation. The Parks and Recreation Department and the Board of Public Works and Safety, allowed installation of observation wells on public property.

The Commissioners of Cass, Clinton, Grant, Howard, Madison, and Tipton Counties allowed installation of observation wells on public right-of-ways.

HYDROLOGY

Surface Water

Streamflow

Wildcat Creek and Deer Creek flow west toward the Wabash River (fig. 4). Wildcat Creek drains 355 mi² or 70 percent of the study area, and Deer Creek and Little Deer Creek drain 150 mi² or 30 percent. Flow of Wildcat Creek in and near Kokomo is affected by two reservoirs with a capacity of 1.1 billion gallons (150 million ft³), by diversion of surface-water averaging 9.6 Mgal/d or 15 ft³/s (Indiana Department of Natural Resources, 1982, p. 33), by four low-head dams, and by disposal of treated sewage.

Surface water is readily available, but flow rates are variable at the five gaging stations listed in table 1. Four of the stations were in the area of study, but one has been discontinued. Another, the only gaging station on Deer Creek, is 17 miles downstream from the Cass and Carrol Counties line near Delphi. Although the mean annual discharge differ from station to station, the unit mean annual discharge is only 0.03 from the mean, ± 0.91 (ft³/s)/mi², for data from all five gaging stations (table 1). The controlling factor for mean annual flow is size of the drainage basin; however, this factor does not apply for low flow. A standard deviation of 0.020 from a mean of 0.022 (ft³/s)/mi²

for data from all five gaging stations is an indication of the wide variation in low flow in the study area. Thus, low flow is the result of effects of local hydrogeology and not the size of the drainage basin.

Flow characteristics of discharges at the four gaging stations are shown by flow-duration curves (fig. 5). Flow duration is the percentage of time that a specific discharge has been equaled or exceeded during a given period without regard to the sequence of occurrence. Effects of climate are equalized by using a common reference period (in this case 1962 through 1982), and size of basin is equalized by dividing the discharges by the drainage area. The mean annual unit discharge of the four gaging stations, $0.90 \text{ (ft}^3/\text{s)}/\text{mi}^2$, was equaled or exceeded only 23 percent of the time from 1962 through 1982, and flow was less than $0.90 \text{ (ft}^3/\text{s)}/\text{mi}^2$ 77 percent of the time.

The shape of a flow-duration curve is related to the hydrologic and the geologic characteristics of the drainage area (Searcy, 1959, p. 22).

Duration curves for the gaging stations are identical for high flows (33 percent duration or less) because flow is proportional to size of basin at high flow. At lower flows, however, the duration curves are divergent, and the slopes are variable at low unit discharges. Unit discharges of Wildcat Creek at Kokomo and Deer Creek near Delphi are higher at the low end of the curves than those of Kokomo Creek at Kokomo or Wildcat Creek near Jerome. The higher unit discharges are an indication of higher sustained low flows. Flow of Wildcat Creek at Kokomo is affected, in part, by treated sewage, diversion of surface water, low-head dams, and the reservoirs on Wildcat Creek, but the duration curve of Wildcat Creek at Kokomo is similar to that of Deer Creek near Delphi. Man has influenced flow of Wildcat Creek at Kokomo since the record began and will probably continue to influence the flow. The record of Deer Creek near Delphi is not affected by man, but the gage is 17 mi downstream from the area of study.

Flow-duration curves of Kokomo Creek near Kokomo and Wildcat Creek near Jerome are steep and similar. Unit discharges at the low end of the curves of Kokomo Creek near Kokomo and Wildcat Creek near Jerome are small compared with unit discharges at other sites, which is an indication of small to negligible low flows at Kokomo and near Jerome.

Stream-Aquifer Connection

The hydraulic connection between streambed and underlying aquifer was evaluated by measuring the gain or loss of flow in all perennial streams. Gains and losses of streamflow were measured in 34 reaches, July 11-13, 1981, and in 35 reaches, August 20 and 21, 1981 (U.S. Geological Survey, 1981, p. 384-387) during a base-flow period (fig. 4). One less reach was measured in July because of backwater at the measuring site between reaches 16 and 17. Thus, the gains of streamflow in reaches 16 and 17 are added to the measurements in July. Flow duration averaged 66 percent during July and 70 percent during August. On the basis of a comparison of the rates of flow at the time of measurement at the four gaged sites and magnitude and frequency of

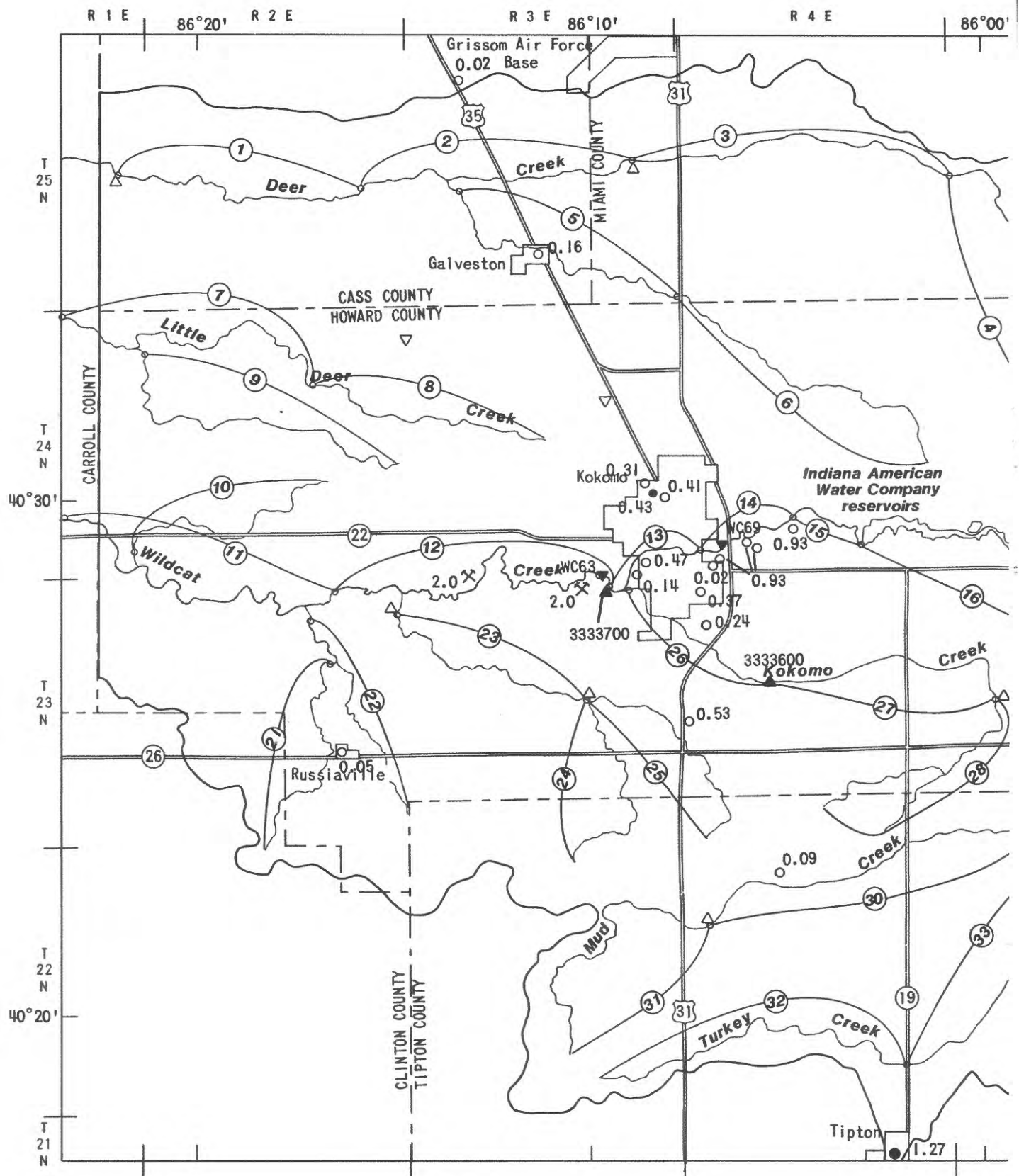
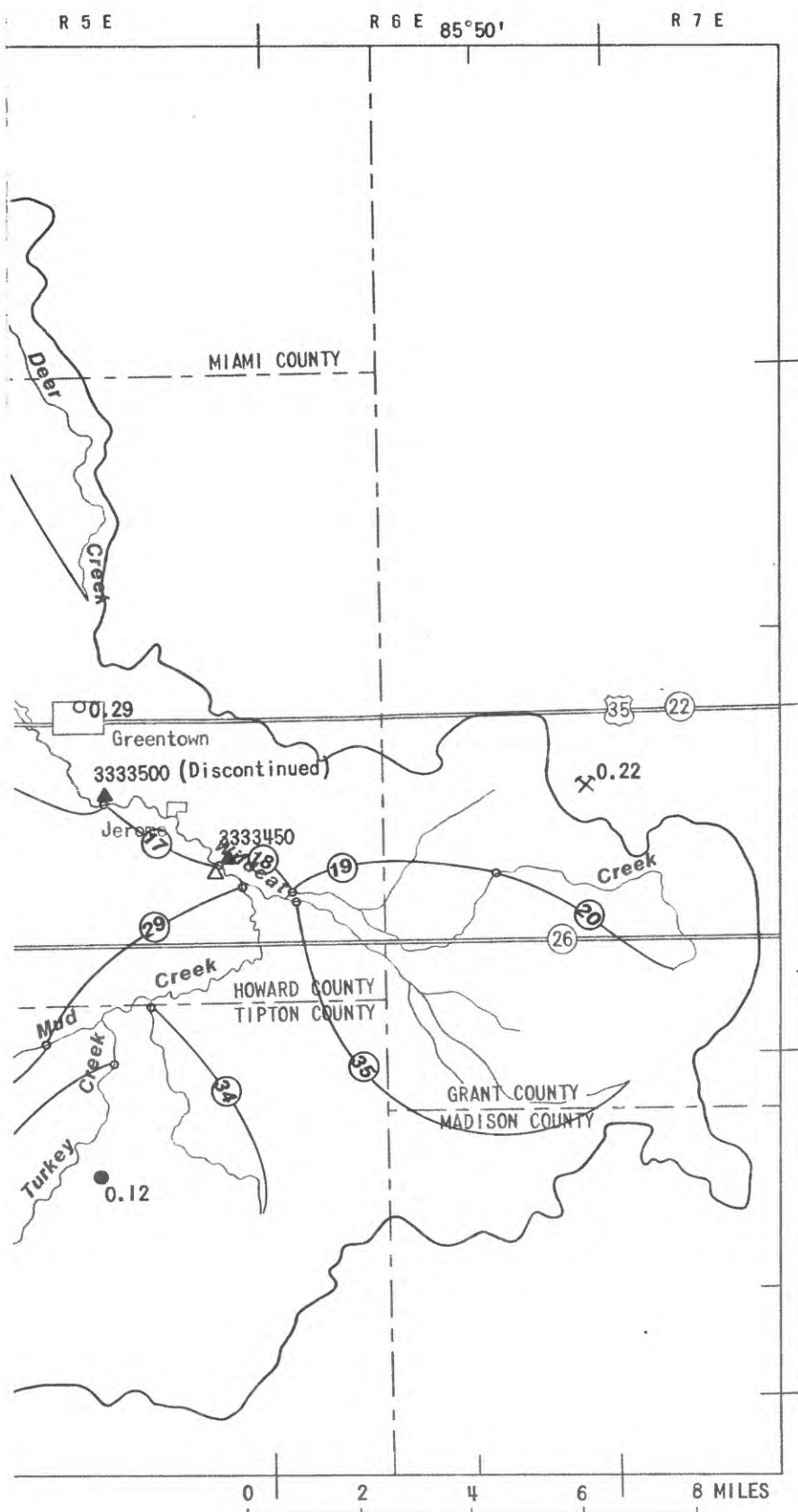


Figure 4.-- Surface-water-data and ground-water pumping sites, 1981.



EXPLANATION

- Streamflow measurement site
- 3333600 ▲ Gaging station and downstream-order number
- WC63 ▼ Indiana State Board of Health stream-monitoring station and number
- ⑮ Number of reach
- 1.27 ● Pumping, in cubic feet per second, and source
From bedrock, and sand and gravel
- 0.02 ○ From bedrock
- 0.43 ● From sand and gravel
- 2.0 ✕ Dewatering, in cubic feet per second
- Samples collected for testing in a permeameter
 - △ Streambed materials
 - ▽ Unconsolidated materials from drill hole
- Dam

Table 1.--Discharge at gaging stations

Number	Name	Drainage area (mi ²)	Period of record (water years)	Mean annual dis- charge ¹ (ft ³ /s)	Unit mean annual discharge [(ft ³ /s)/mi ²]	7-day, 10-year low flow ² (ft ³ /s)	Unit 7-day, 10-year low flow [(ft ³ /s)/mi ²]
3329700	Deer Creek near Delphi ³	274	1944-82	241	0.88	10	0.036
3333450	Wildcat Creek near Jerome	146	1962-82	131	.90	1.3	.009
3333500	Wildcat Creek at Greentown	162	1944-60	150	.93	1.4	.009
3333600	Kokomo Creek near Kokomo	24.7	1960-82	21.6	.87	.10	.004
3333700	Wildcat Creek at Kokomo	242	1956-82	230	.95	12	.050
Mean					0.91		0.022
Standard deviation					.03		.020

¹From U.S. Geological Survey (1983, p. 88, 93, 95, and 96) and (1961, p. 47).²From Stewart (1983, p. 65, 73, 74, 75, and 76).³Outside the area of study.

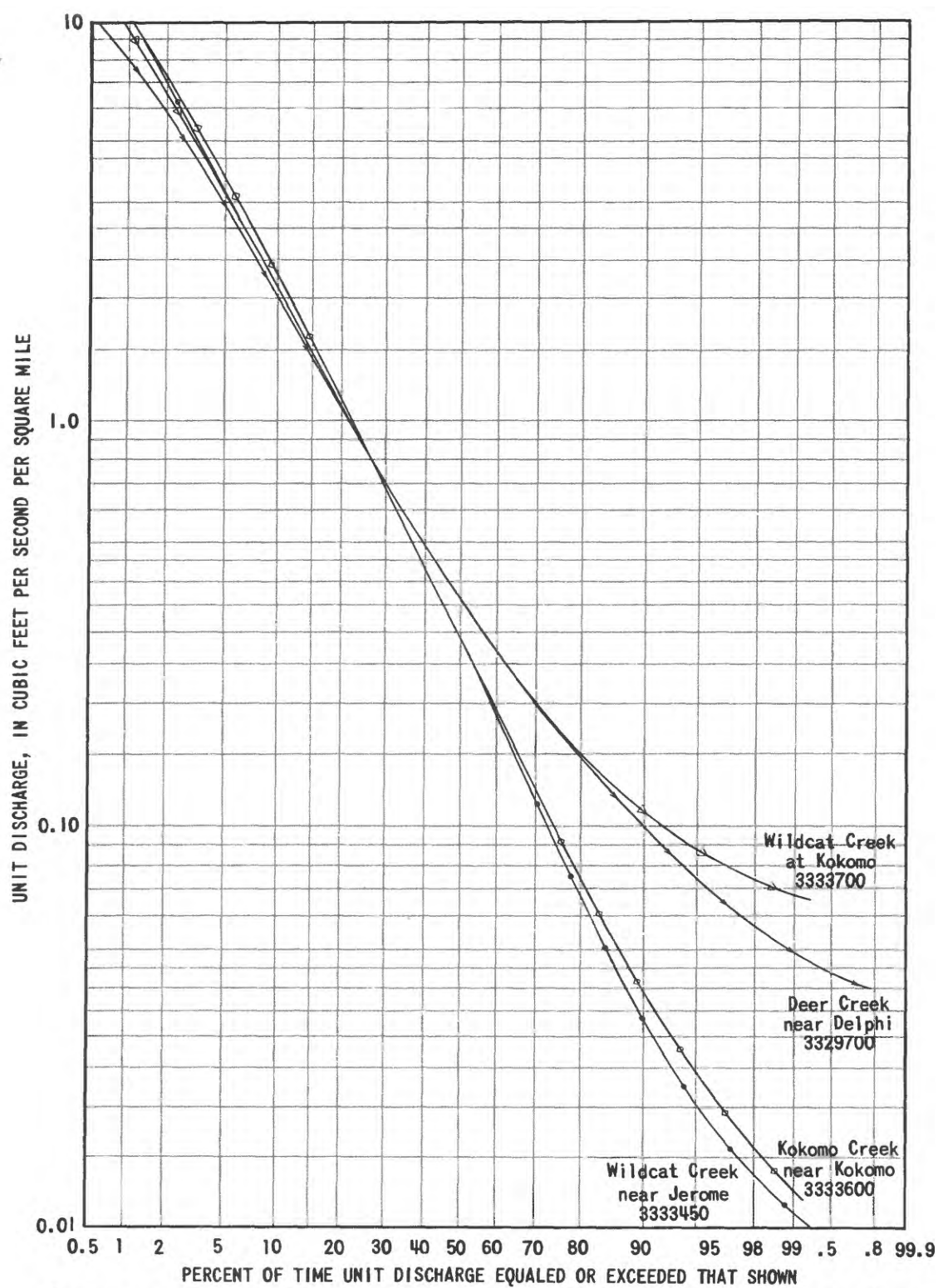


Figure 5.-- Flow-duration curves for selected gaging stations in north-central Indiana, water years 1962-82.

annual low flows for the sites (Stewart, 1983, p. 65, 73, 75, and 76), the authors concluded that the measurements were made at a time approximately equal to the 120-day, 2-year low flow. In other words, a seasonal (120-day) base flow of that magnitude should occur every other year under normal climatic conditions.

Gains or losses for each reach (table 2) were determined by subtracting the discharge measured at the upstream end from the discharge measured at the downstream end of the reach. Diversions of streams were then added to this value, and stream inputs were subtracted. A confidence or error interval is attributed to the person measuring the discharge around each gain or loss. The error is generally ± 5 percent. The gain or loss in each reach was then divided by the difference in drainage area between the downstream site of measurement and the upstream site to allow comparison of these values on a unit basis.

Gains were observed at most of the reaches, which is an indication of ground-water seepage to the streams. Loss in streamflow was observed, however, for one or both sets of measurements in reaches 12, 13, and 15, and flow was reduced in reach 14 because of large-scale ground-water pumping, periodic dewatering of quarries, and storage in reservoirs in or near Kokomo. A loss was also observed for reach 18, the smallest reach, in July; however this reach was gaining in August.

Discharge is proportional to size of basin at high streamflow (less than 33-percent flow duration), but discharge is affected by local hydrogeology at lower flow. For both periods of measurement, the highest unit gains greater than $0.30 \text{ (ft}^3/\text{s)}/\text{mi}^2$ are for reaches 11, 16, and 17, which are the reaches of Wildcat Creek downstream from the mouth of Mud Creek, except for those reaches affected by human activities near Kokomo, reaches 12, 13, 14, and 15. The next highest gains--from $0.16 \text{ (ft}^3/\text{s)}/\text{mi}^2$, the mean value of each period of measurement, to $0.30 \text{ (ft}^3/\text{s)}/\text{mi}^2$ --are for reaches 5, 9, 26, and 29. Gains at the remaining reaches are less than the mean value during one or both of the periods of measurement.

Hydraulic conductivity of the streambeds, which range from 2×10^{-3} to 8 ft/d, was estimated from falling-head permeameter tests done by the Indiana Geological Survey on 9 split-spoon and shelly-type samples from 7 sites (Sam Frushour, Indiana Geological Survey, written commun. 1982,). The samples generally contained a mixture of sand, silt, and gravel. Shell fragments were in two of the samples, and wood fragments were in one. Average hydraulic conductivity of the nine samples was 2 ft/d. The small sampling may not be representative of the hundreds of miles of streambed within Howard and adjacent counties, and the samples may not be representative of the entire range of streambed materials; however, the conductivities are within the range of hydraulic conductivities reported for silt and silty sand-- 7.5×10^{-3} to $7.5 \times 10^3 \text{ (gal/d)}/\text{ft}^2$ or 1×10^{-3} to $1 \times 10^3 \text{ ft/d}$ --by Freeze and Cherry (1979, p. 29). Conductivities of streambed, obtained by use of the ground-water flow model, generally ranged from 2×10^{-2} to 20 ft/d for a streambed assumed to be 1 ft thick.

Table 2.--Measured and modeled gains and losses in streamflow

Reach	July 11, 12, and 13, 1981		August 20 and 21, 1981	
	Measured ¹ (ft ³ /s)	Per area drained [(ft ³ /s)/mi ²]	Measured ¹ (ft ³ /s)	Per area drained [(ft ³ /s)/mi ²]
1	1.6	0.14	2.1	0.18
2	4.5	.22	2.1	.10
3	4.3	.15	7.7	.26
4	1.9	.13	2.7	.19
5	2.9	.19	3.8	.25
6	2.7	.11	4.0	.16
7	1.9	.16	1.6	.13
8	1.9	.13	2.0	.14
9	2.0	.22	1.8	.20
10	² .7	.07	² .4	.04
11	21.6	.94	13.6	.59
12	³ -5.0	-.22	5.3	.24
13	² -2.5	-.33	² -10.0	-1.36
14	⁴ 6.5	.43	⁴ 1.5	.10
15	- .6	-.15	7.6	1.90
16	² 14.9	.33	² 12.2	.37
17 ⁵	--	--	8.4	.70
18	- .4	-.22	- 1.7	-.84
19	1.7	.16	1.2	.12
20	² 1.6	.11	² 1.4	.10
21	1.6	.12	1.2	.09
22	1.4	.15	1.1	.12
23	1.0	.07	1.2	.06
24	.6	.07	1.0	.11
25	1.0	.11	1.2	.13
26	4.7	.40	3.2	.27
27	1.8	.15	1.9	.16
28	.9	.07	.3	.03
29	3.8	.29	2.8	.22
30	1.9	.14	2.0	.15
31	.7	.04	1.4	.09
32	1.3	.12	1.9	.19
33	4.7	.14	3.4	.10
34	2.0	.27	1.0	.11
35	4.0	.17	2.8	.11
Total	96.6	5.32	94.1	5.51
Mean	2.8	.16	2.7	.16

¹Error interval is ± 5 to ± 8 percent around the measured streamflow gains and losses.

²Adjusted for surface-water disposal.

³Negative sign indicates streamflow loss.

⁴Adjusted for surface-water diversion.

⁵Gain for reach 17 is added to reach 16 for the July measurement because of backwater at the measurement site.

Ground Water

Flow

Ground-water flow is strongly influenced by the deposits of till plain. The predominance of till compared to sand and gravel in the unconsolidated deposits is indicated in the cross section south to north through Kokomo (fig. 6). The till is deposited in continuous layers, whereas the sand and gravel is deposited in discontinuous sheets interspersed within the till. The layers of till are semiconfining beds, and ground water is confined in the sand and gravel aquifers. Ground water is also confined within open fractures and solution channels in the upper beds of limestone and dolomite that underlie the glacial deposits. A fine-grained limestone in the Mississinewa Shale lies 680 ft above sea level and 126 ft below the surface in the vicinity of Kokomo (Shaver, 1961, p. 13). Indurated silt and clay in the interstices between the grains of the fine-grained limestone results in a zone of reduced transmissivity. The base of the bedrock aquifer is roughly defined by the top of the silty limestone (Watkins and Rosenshein, 1963, p. B7).

Regional flow directions indicated by arrows in figure 6 were determined by differences in hydraulic pressure (water levels) at various points in the basins. The vertical exaggeration of the geologic section (32 times), however, does not indicate precisely the geometry of ground-water flow. Movement of ground water is related to the geometry and the geology of the basin, and the ground-water flow system generally follows the laws governing discontinuous, layered, heterogeneous systems (Freeze and Cherry, 1979, p. 30, 172, and 197). Because of the contrast in hydraulic conductivity between the layers of till and the sand and gravel aquifers, horizontal flow predominates in the aquifers and is negligible in the semiconfining till. Recharge to the aquifers, however, is by downward vertical leakage through the semiconfining layers, and discharge is by upward vertical leakage through the semiconfining beds and the streambeds. The surface area in the vertical direction of flow through the semiconfining layers is much smaller than the surface area in the horizontal direction.

The principal aquifer is the confined limestone-dolomite, and thousands of domestic wells pump water from the this aquifer (unpublished well records). Water levels in the bedrock (fig. 7) were measured in approximately 150 domestic and commercial wells in and near the area of study from May through July 1980 and in two continuous-record observation wells beginning in 1966 and 1967 and continuing through 1981. Regional directions of flow in the bedrock aquifer are perpendicular to the lines of equal water levels. Flow is from areas of high water levels toward the streams, particularly toward Wildcat Creek and Deer Creek. Water is recharged over the entire area of study except at the stream where the regional ground-water-flow system is discharging. The pattern of flow in the aquifer is variable because flow in bedrock is within open fractures and channels and aquifer characteristics at any point in the bedrock depend on the number and the size of openings penetrated by the well in which the characteristics were measured.

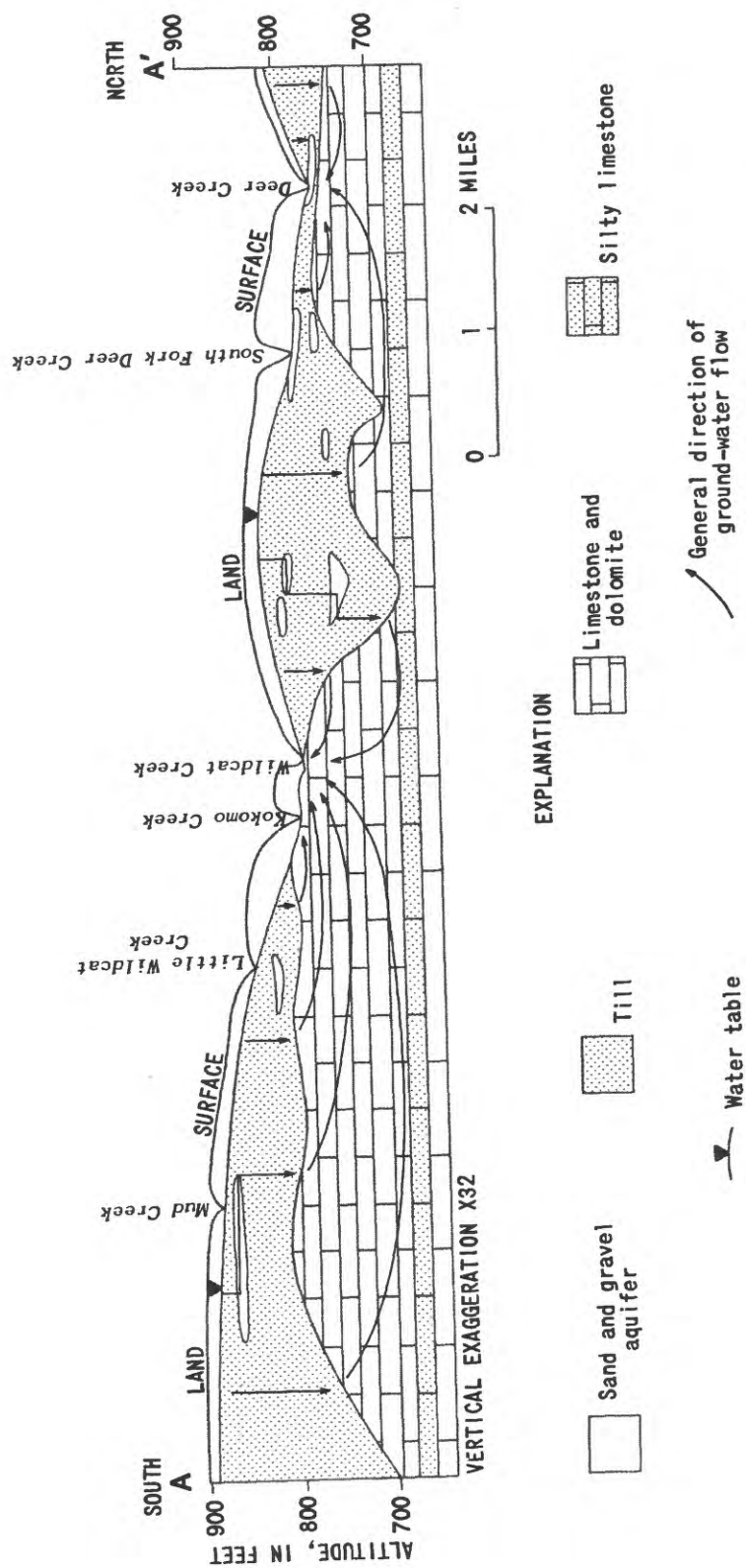


Figure 6.-- Geologic section through Kokomo.

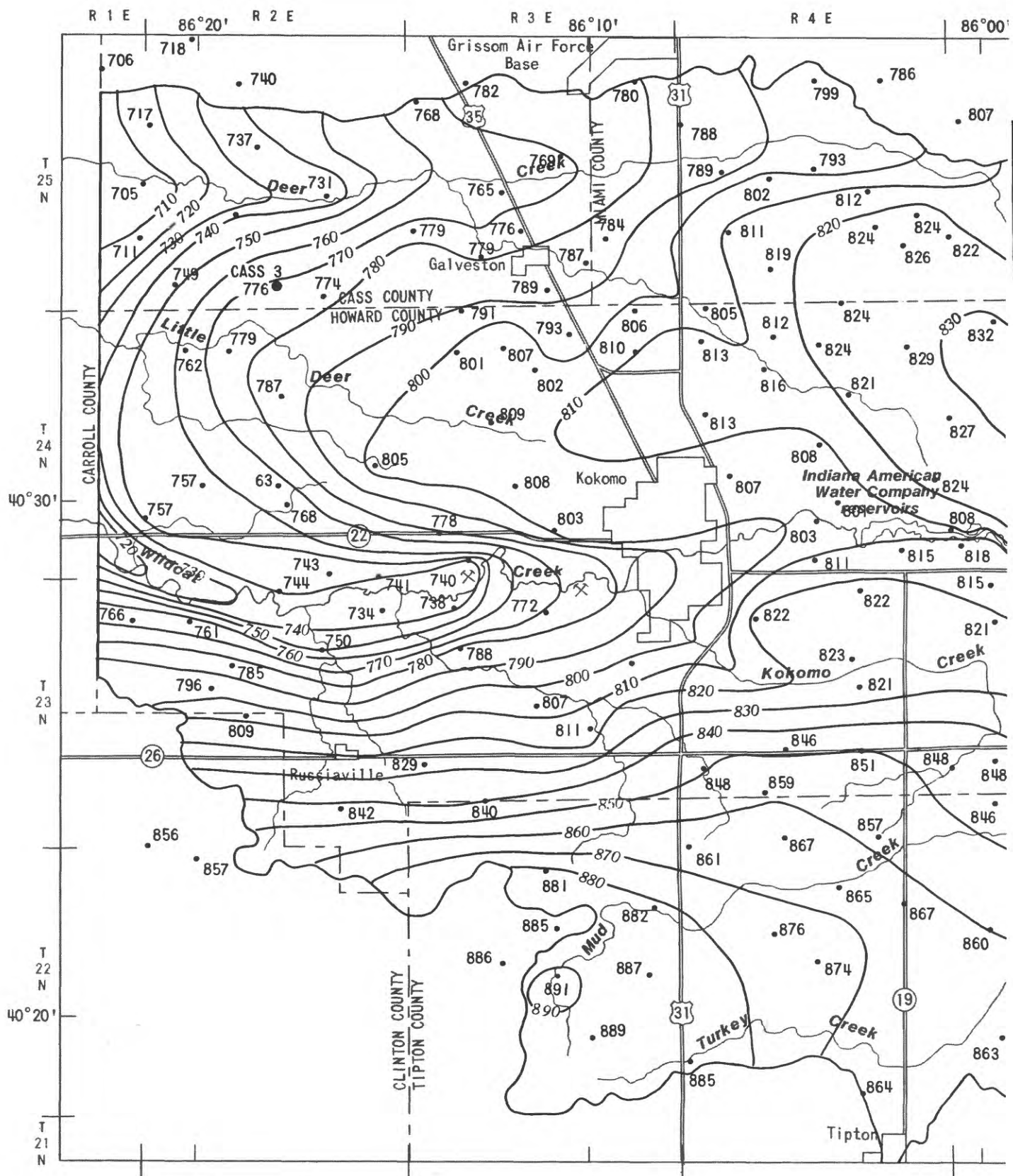
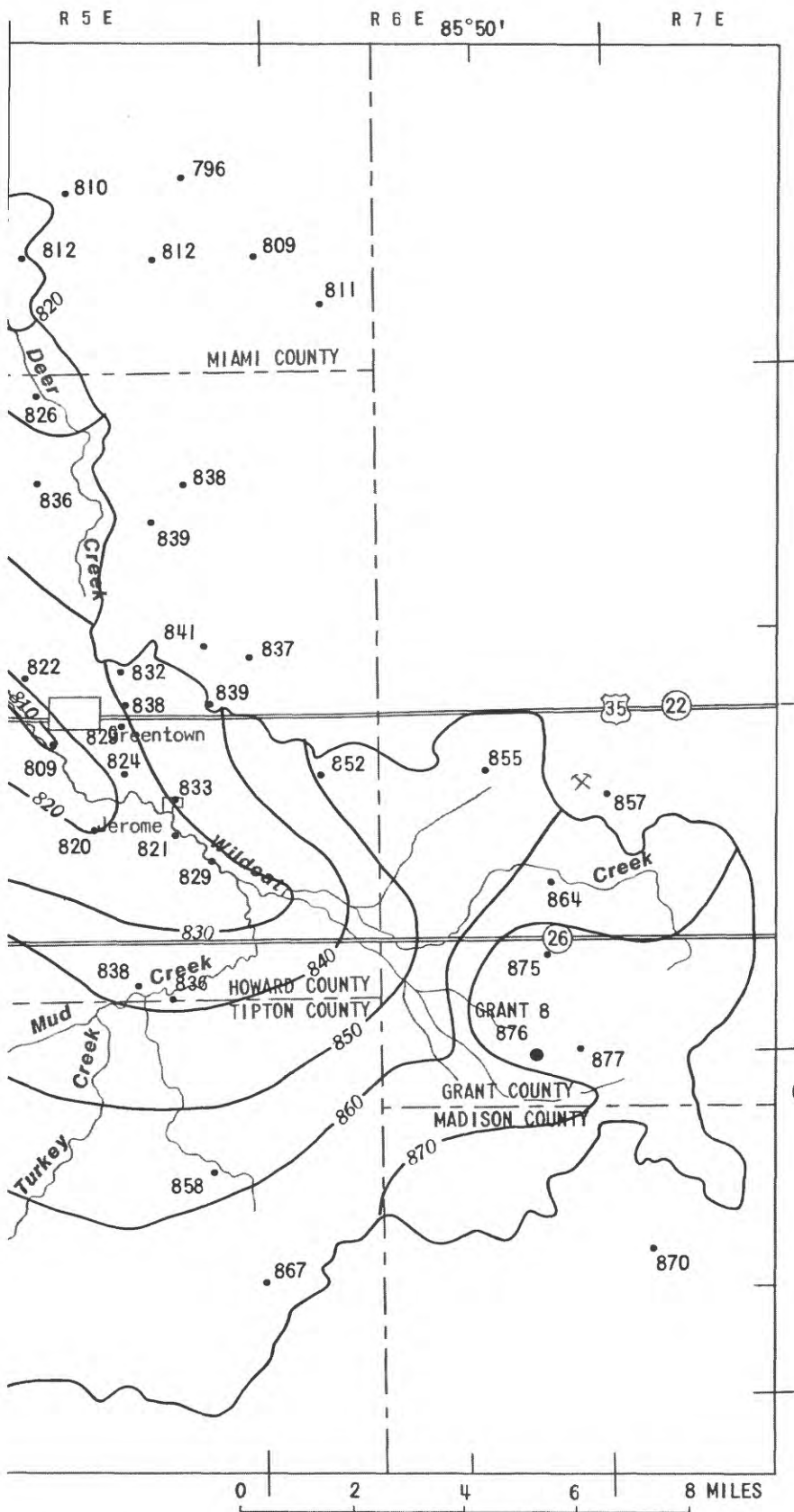


Figure 7.-- Water levels in the bedrock.



EXPLANATION

- 810— Line of equal water-level altitude. Interval 10 feet
- GRANT 8 Continuous-record well.
- 876 Water level in feet
- 870 Water level in feet. Measured May through July 1980
- Datum is sea level

Water levels were measured once in approximately 40 domestic and commercial wells in the confined sand and gravel aquifers from May through July 1980 and periodically in approximately 90 observation wells in the confined sand and gravel aquifers and semiconfined layers from December 1980 through May 1982. On the basis of these data, regional directions of flow in the confined sand and gravel aquifers are similar to those for the bedrock. Local and intermediate patterns of flow in the sand and gravel aquifers, however, are superimposed on the regional patterns.

Pumpage and Water-Level Fluctuation

Water was pumped at 7.1 Mgal/d from aquifers in the basins from 1970 to 1981. More than 90 percent of the pumping was from the bedrock. Thirty-three percent of the ground-water pumpage was used by Kokomo, and the majority of all commercial, industrial, and municipal wells are in and around Kokomo (fig. 4). Seasonal and periodic fluctuations of less than 2 ft superimposed on a long-term trend are shown in hydrographs of the continuous-record wells (fig. 8). Minor gains and losses in aquifer storage are indicated by seasonal and periodic fluctuation; however, the long-term trend is approximately steady state. Water levels in the bedrock aquifer will not differ much from those measured in 1980 without major changes in pumpage. The changes in ground-water levels caused by intermittent dewatering of the limestone quarries west of Kokomo is an example of how water levels in small areas may differ from those measured in 1980.

Effective Recharge

Effective recharge is the rate of water per unit area contributed to the water table from precipitation. Seasonal and periodical variation in effective recharge results in fluctuation of water levels (fig. 8) and, consequently, of storage. The short-term changes, however, are superimposed over a constant long-term trend.

Recharge rates can be estimated by streamflow analysis. Flow into the basin is assumed to be equal to flow out of the basin. In large basins, such as the area of study, this assumption is valid because the surface area recharged from precipitation is large compared to the surface area where ground-water flow might cross a surface-water boundary. Also, Wildcat and Deer Creeks are incised in the till plain, and regional ground-water flow is discharging into the streams. The discharges, therefore, are estimates of the water budgets of the basins.

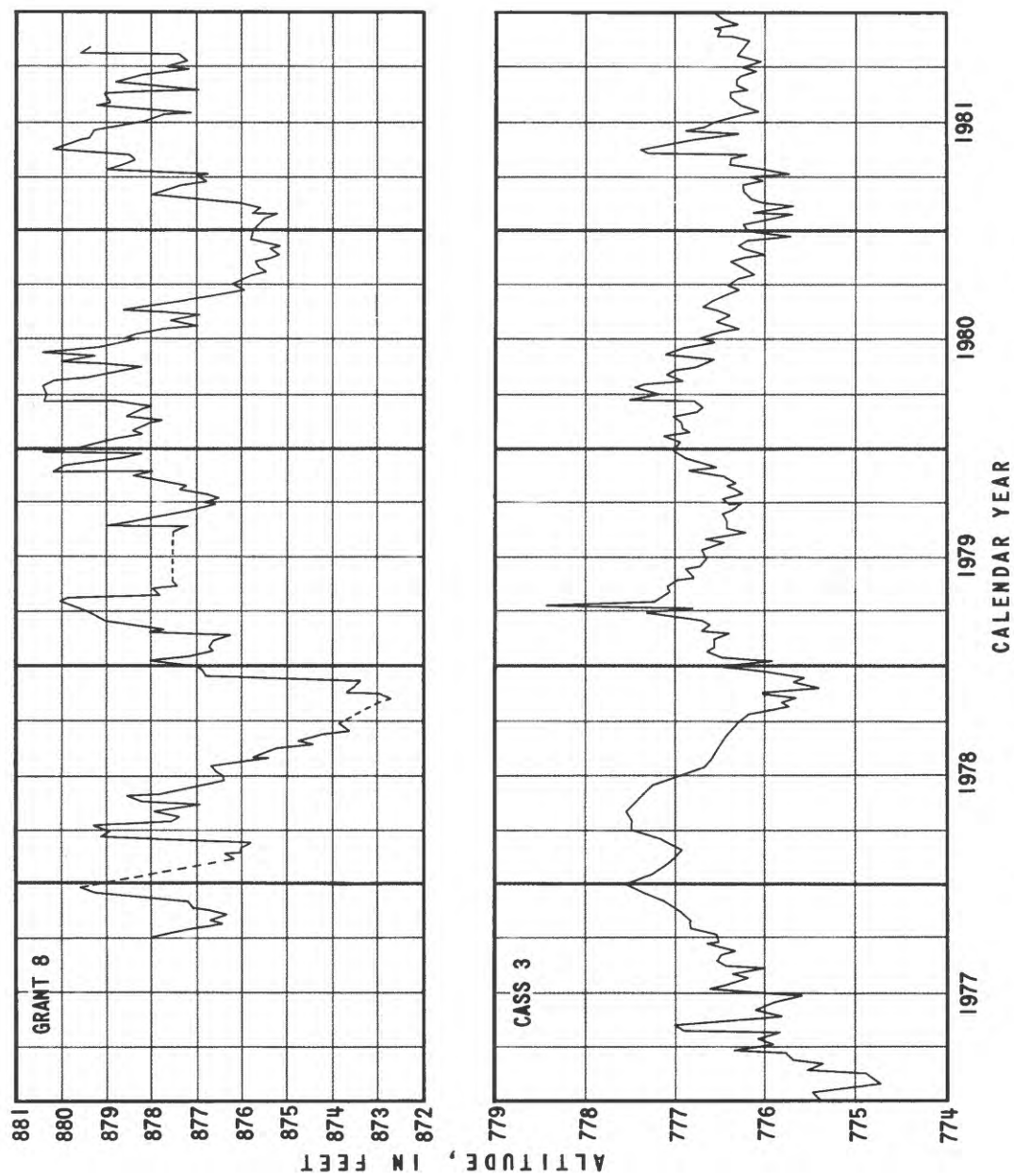


Figure 8.-- Water-level fluctuations in observation wells.

Flow was measured at the Carroll County line for Wildcat Creek, Deer Creek, and Little Deer Creek, which together at that line drain 505 mi², a good approximation of the area of study. For convenience of expression, flow, in cubic feet per second per area drained, was converted to inches per year.

Four separate measurements of gain and loss of streamflow were used to estimate recharge (U.S. Geological Survey, 1981, p. 357-363, and 1982, p. 384-387). The streamflows were adjusted for diversion of surface water and disposal of sewage at the time of the measurements. Seepage and the estimates of recharge rates are listed in the following table:

Date of measurement		Flow duration	Seepage (ft ³ /s)	Estimate of recharge (in/yr)
May	15-16, 1980	57	139.5	3.7
July	11-13, 1981	65	96.6	2.6
Aug.	20-21, 1981	70	94.1	2.5
Sept.	11-13, 1980	83	<u>55.0</u>	<u>1.5</u>
Mean			96.3	2.6

The range in recharge rates, which are based on periods of recession in streamflow at various flow durations, represents seasonal and periodic fluctuations; however, the range is not great. The mean recharge rate, 2.6 in/yr, agrees with the recharge rate calculated for the bedrock immediately north of the area of study in the vicinity of Grissom Air Force Base where 2 Mgal/d was estimated for a 16-square-mile area (a recharge rate of 2.6 in/yr) by flow-net analysis (Watkins and Rosenshein, 1963, p. B15).

HYDROGEOLOGY

Geometry of the Sand and Gravel Aquifers

Discontinuous, confined sand and gravel aquifers, separated vertically by semiconfining beds, were defined from analysis of gamma logs and lithologic descriptions in drillers' logs. To simplify mapping, the authors selected level planes through the semiconfining beds that separated the aquifers and for convenience labeled them upper, middle, and lower. Then thicknesses of aquifers between the semiconfining beds were plotted, were contoured, and were used to determine the extent of the aquifers. Thicknesses less than 3 ft were considered to be insignificant except for defining a thinning edge of aquifer. The authors are not implying that there are three distinct sand and gravel aquifers. The sand and gravel aquifers are discontinuous and interspersed in the till. The method allows an accounting of all the important sand and gravel

deposits. The altitudes of the tops of the aquifers were also contoured, and the areas where the aquifers were connected were determined by overlaying top and thickness maps.

Altitude of the top of the upper sand and gravel aquifer (fig. 9) ranges from 775 to 880 ft. By definition, the upper sand and gravel aquifer is above an altitude of 775 ft. The upper aquifer is capped by semiconfining beds generally ranging in thickness from 5 to 70 ft. The upper aquifer may be connected to the middle sand and gravel aquifer in three locations near the Indiana American Water Co. reservoirs and at one location 4 mi north of Tipton. The upper aquifer may also be connected to the bedrock 7 mi northeast of Tipton.

Thickness of the upper aquifer, generally 10 to 20 ft, is about 40 ft immediately south and west of Russiaville, north of the intersection of State Routes 26 and 19, and southeast of Greentown (fig. 10). The thickness is 40 ft also where the upper and middle aquifers are connected northeast of the Indiana American Water Co. reservoirs. Thickness is greater than 50 ft, however, where the upper and the middle aquifers are connected at the Indiana American Water Co. reservoirs and 3 mi northwest of the reservoirs.

Altitude of the top of the middle aquifer (fig. 11) generally ranges from 710 to 770 ft. The middle aquifer is in the north, west, and south parts of the area but is not widely distributed in the southeast where the projected plane of the aquifer terminates at the bedrock subcrop. Thickness of the semiconfining beds that separate the upper and the middle aquifers generally ranges from 5 ft along the lower half of Deer Creek to 100 ft along the southwest boundary of the area of study. The middle aquifer is not connected to the lower sand and gravel aquifer. However, it may be connected to the bedrock at three locations--7 mi north of Kokomo, 8 mi northeast of Kokomo, and 3 mi east of Kokomo.

The middle sand and gravel aquifer is generally from 10 to 20 ft thick and, in a few areas, is 30 to 40 ft thick (fig. 12). It is more than 40 ft thick, however, below the east end of the Indiana American Water Co. reservoirs.

Altitude of the top of the lower sand and gravel aquifer (fig. 13), in the buried bedrock valley system west of Kokomo, generally ranges from 660 to 700 ft. The lower aquifer is separated from the middle aquifer by a semiconfining bed ranging in thickness from 5 to 80 ft and is separated from the bedrock below by a semiconfining bed generally ranging in thickness from 5 to 100 ft. The lower sand and gravel aquifer may be connected to the bedrock aquifer in a few small areas shown in figure 13.

Thickness of the lower aquifer, which generally ranges from 10 to 20 ft (fig. 14), is greater than 30 ft in two areas--below the upstream reach of Little Deer Creek and 2 mi west of Russiaville in Clinton County.

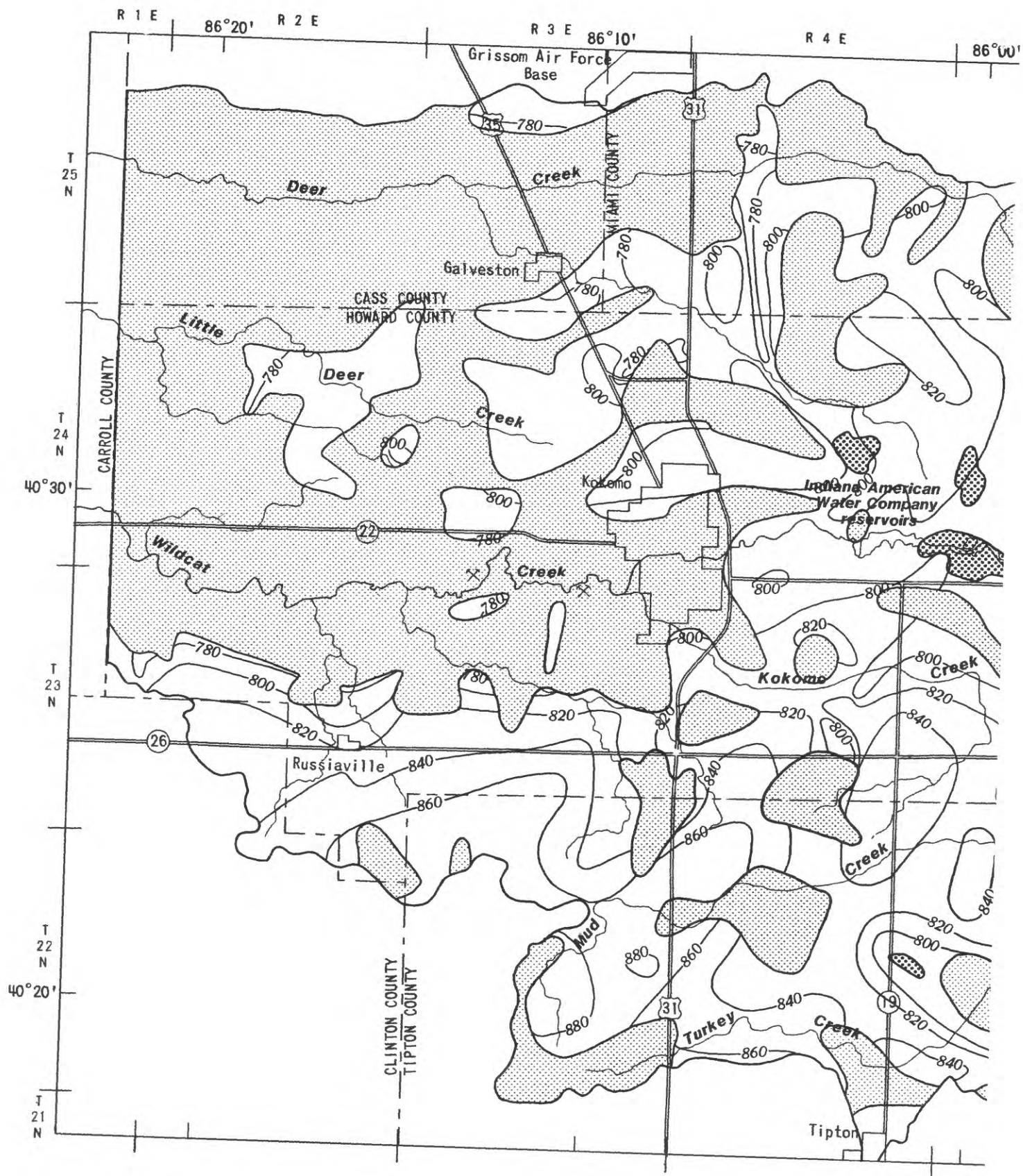


Figure 9.-- Altitude of top of the upper sand and gravel aquifer.

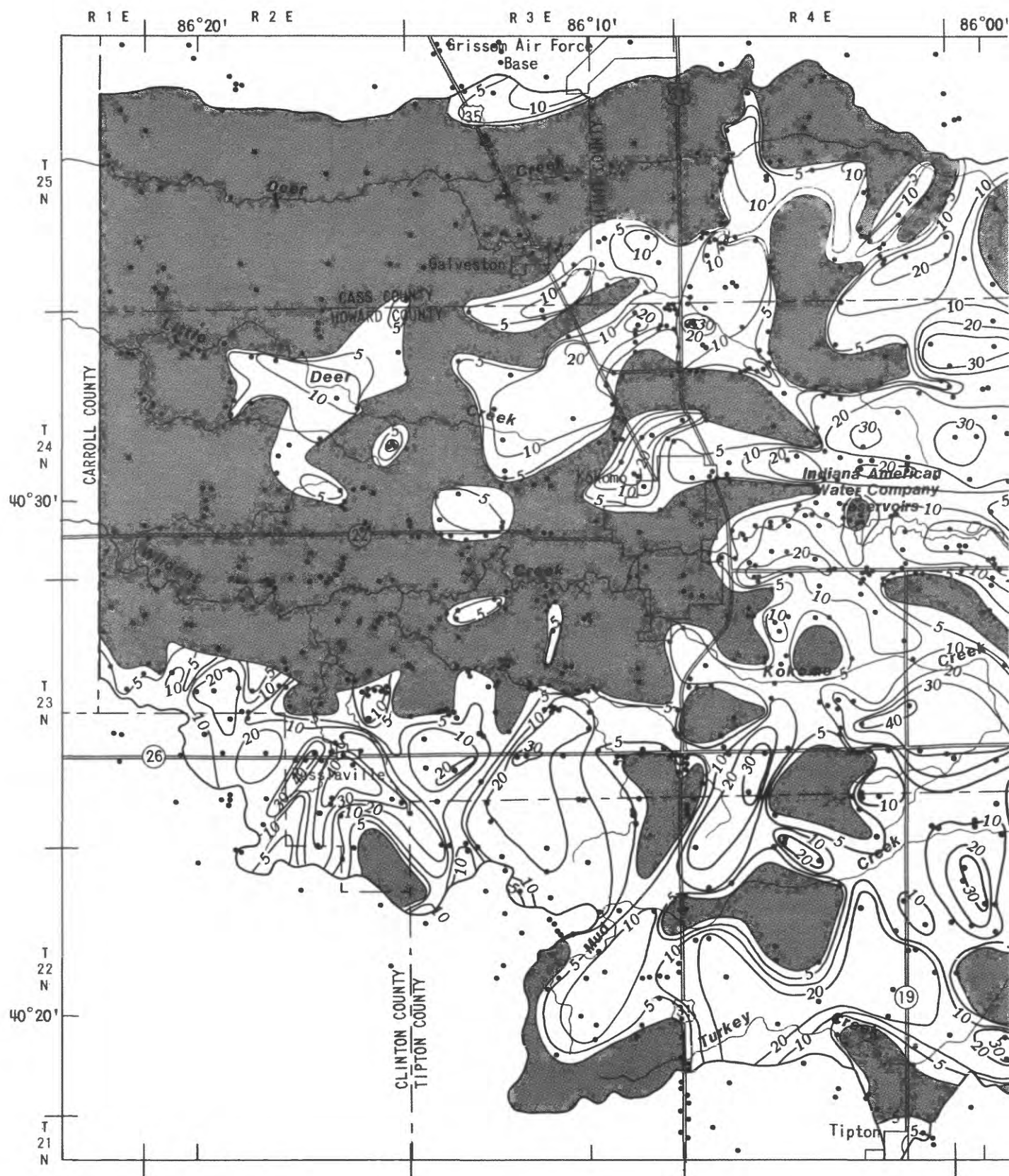
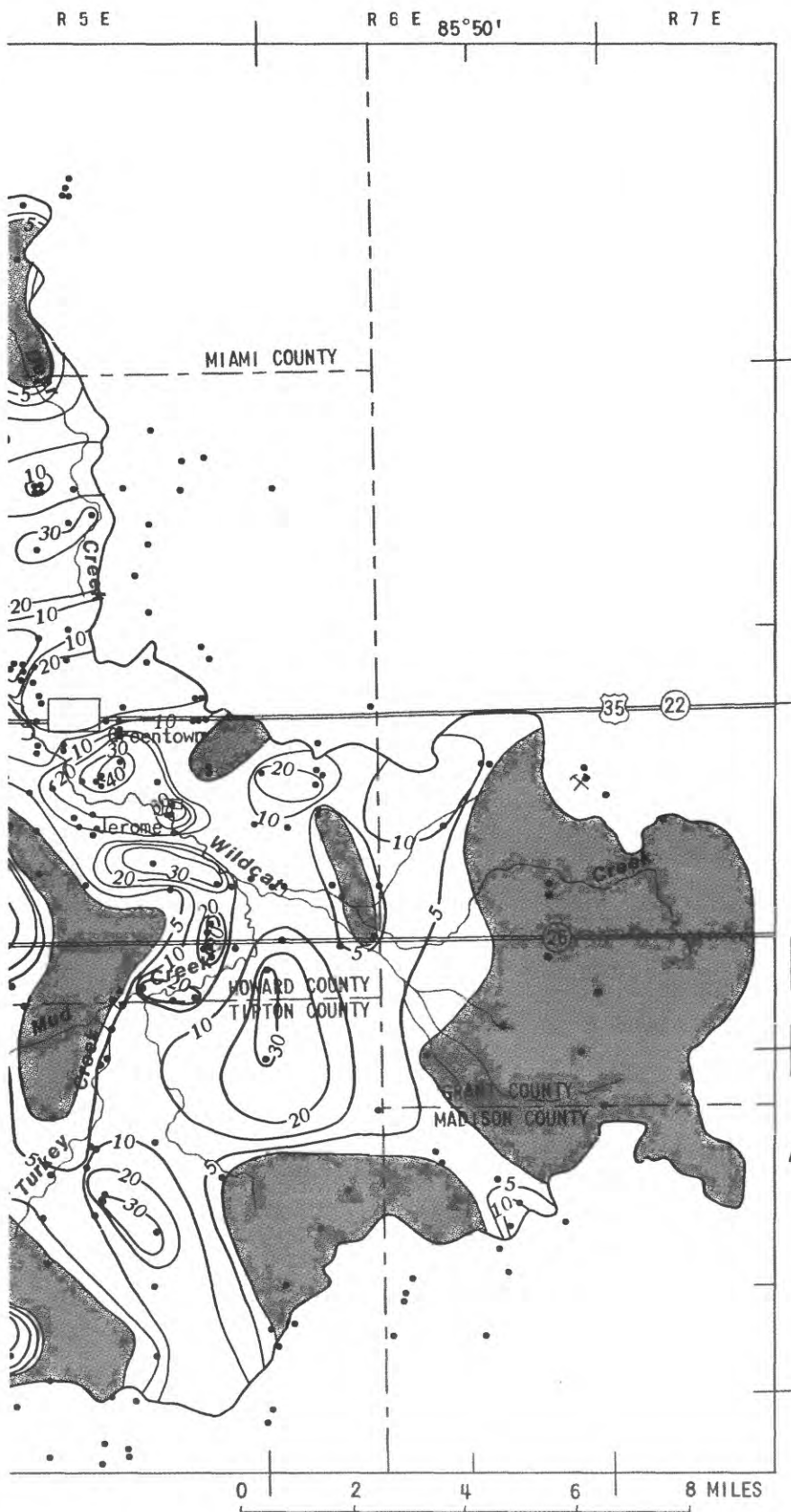


Figure 10.-- Thickness of the upper sand and gravel aquifer.



EXPLANATION



Upper sand and gravel aquifer



Till, silt, or clay.
May contain sand or gravel lenses
less than 3 feet thick



Line of equal thickness.
Intervals 5 and 10 feet.
Thicknesses of aquifer were obtained
from drillers' well logs.

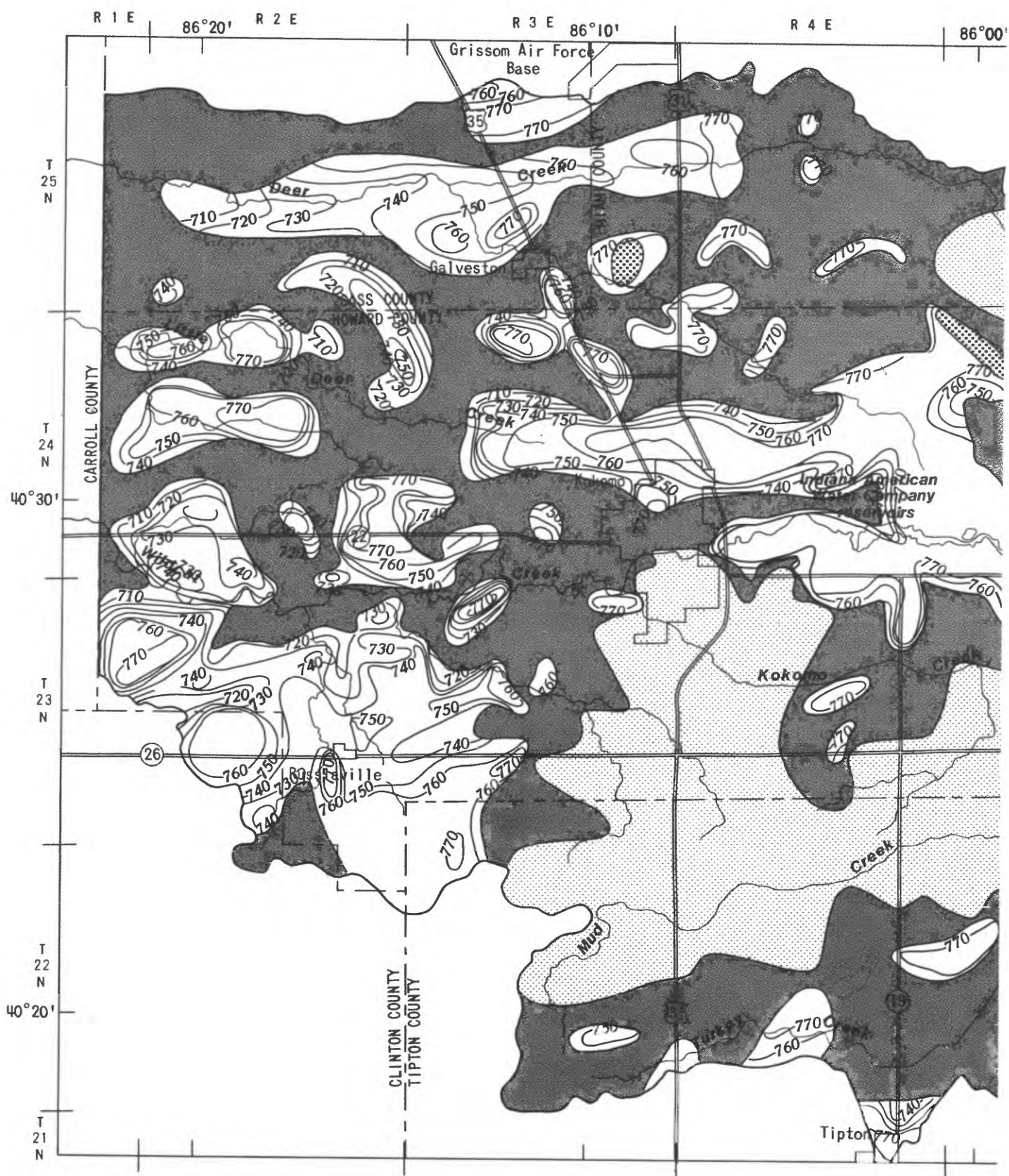
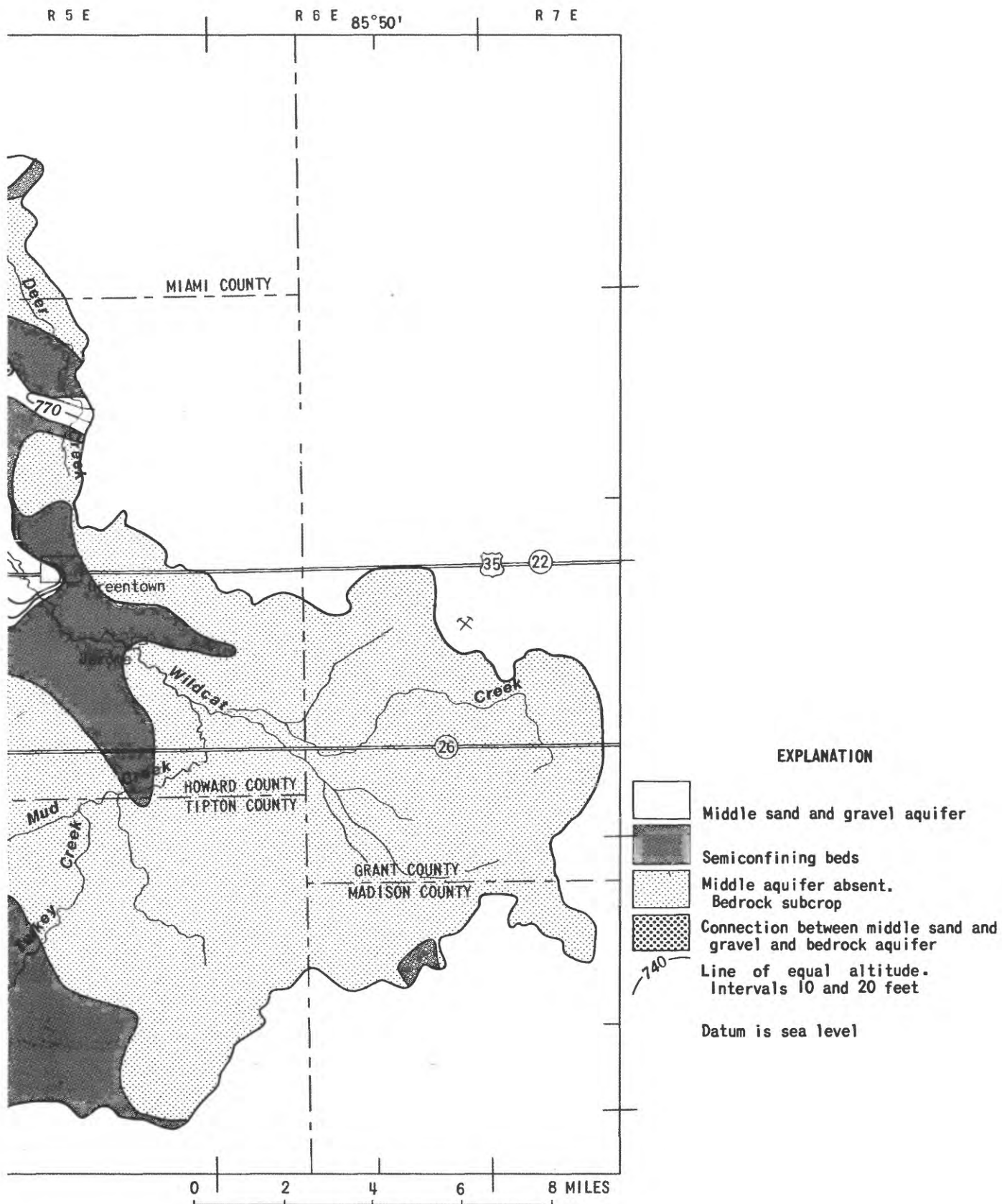


Figure 11.-- Altitude of top of the middle sand and gravel aquifer.



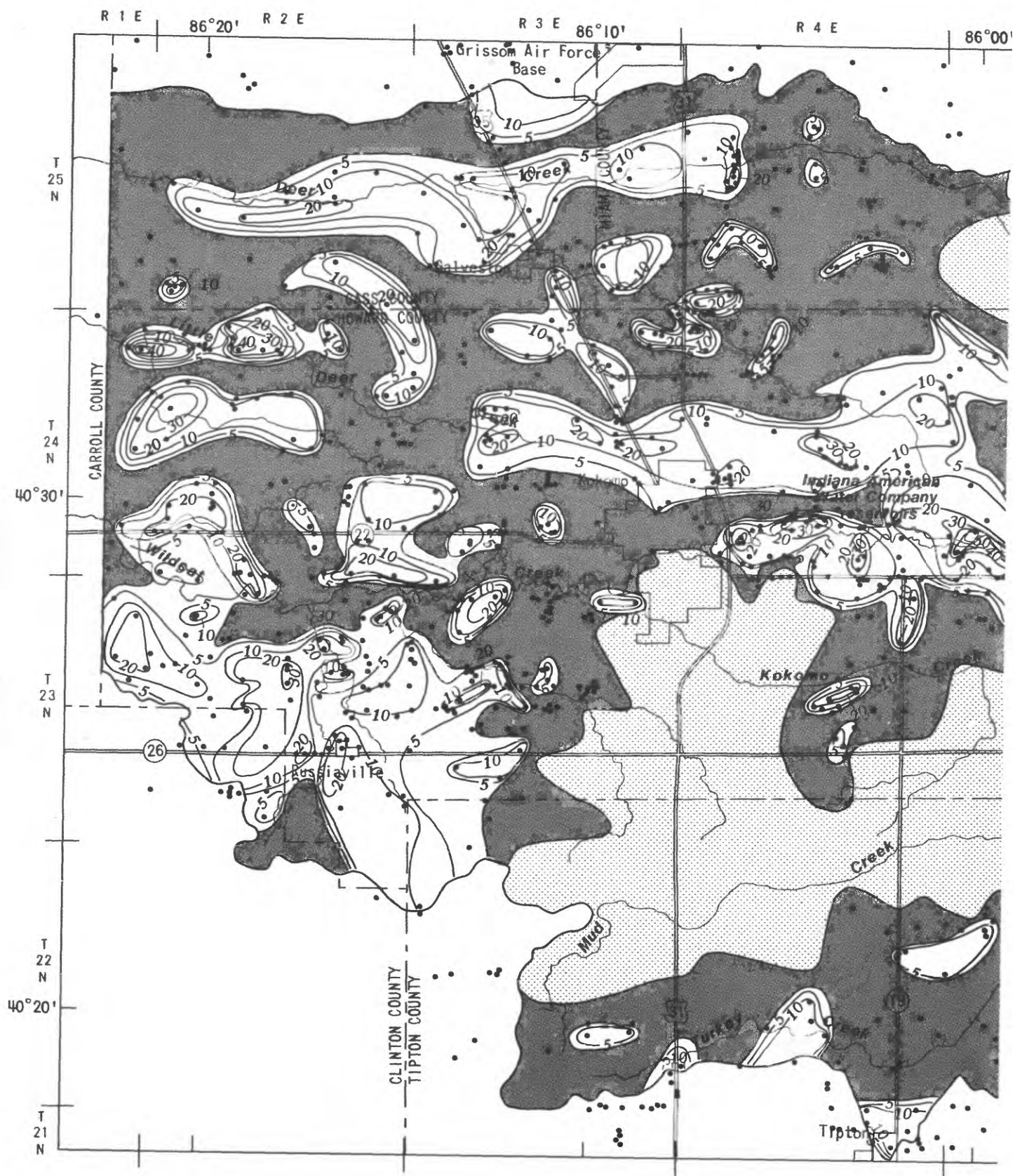
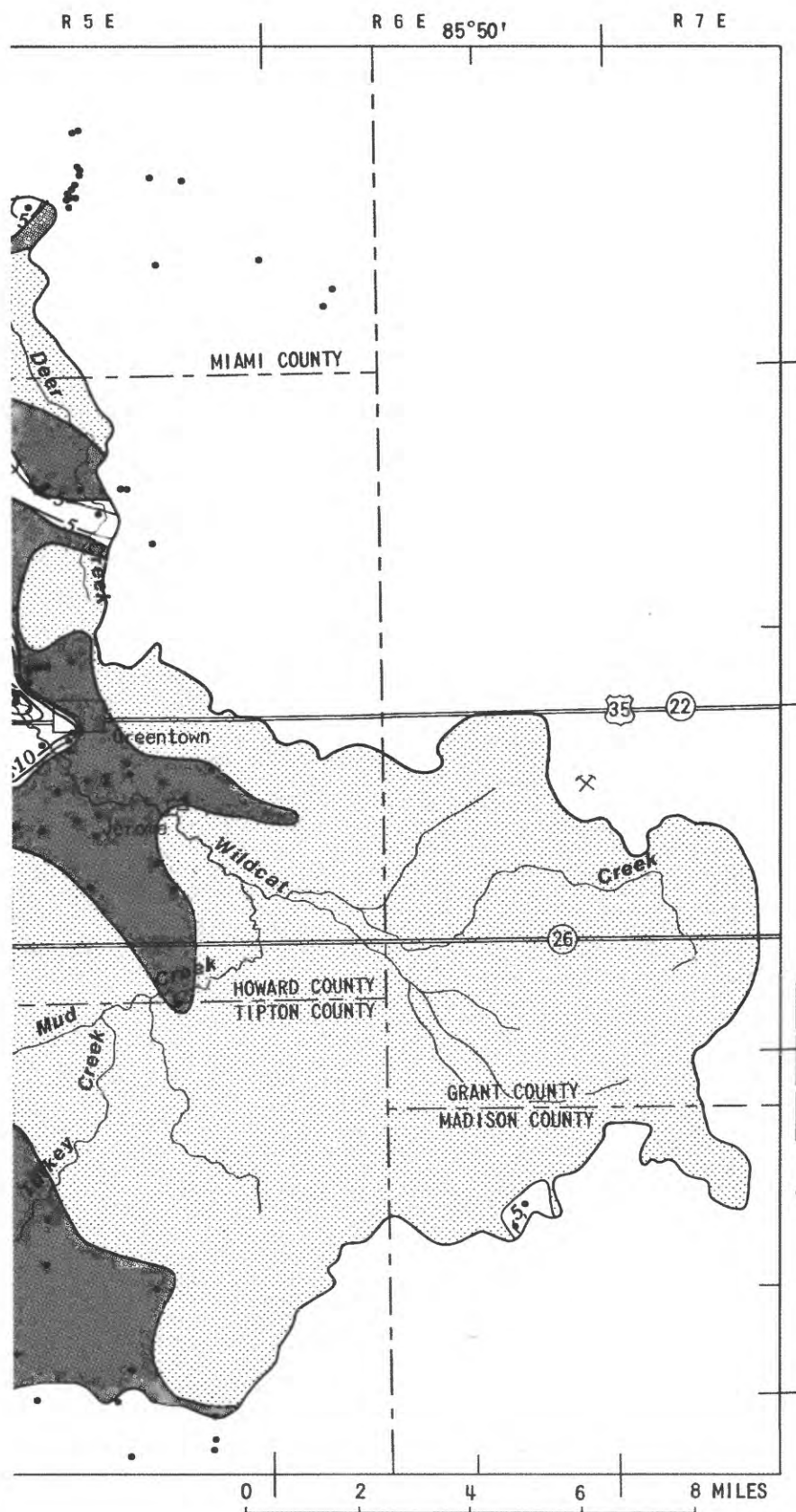


Figure 12.-- Thickness of the middle sand and gravel aquifer.



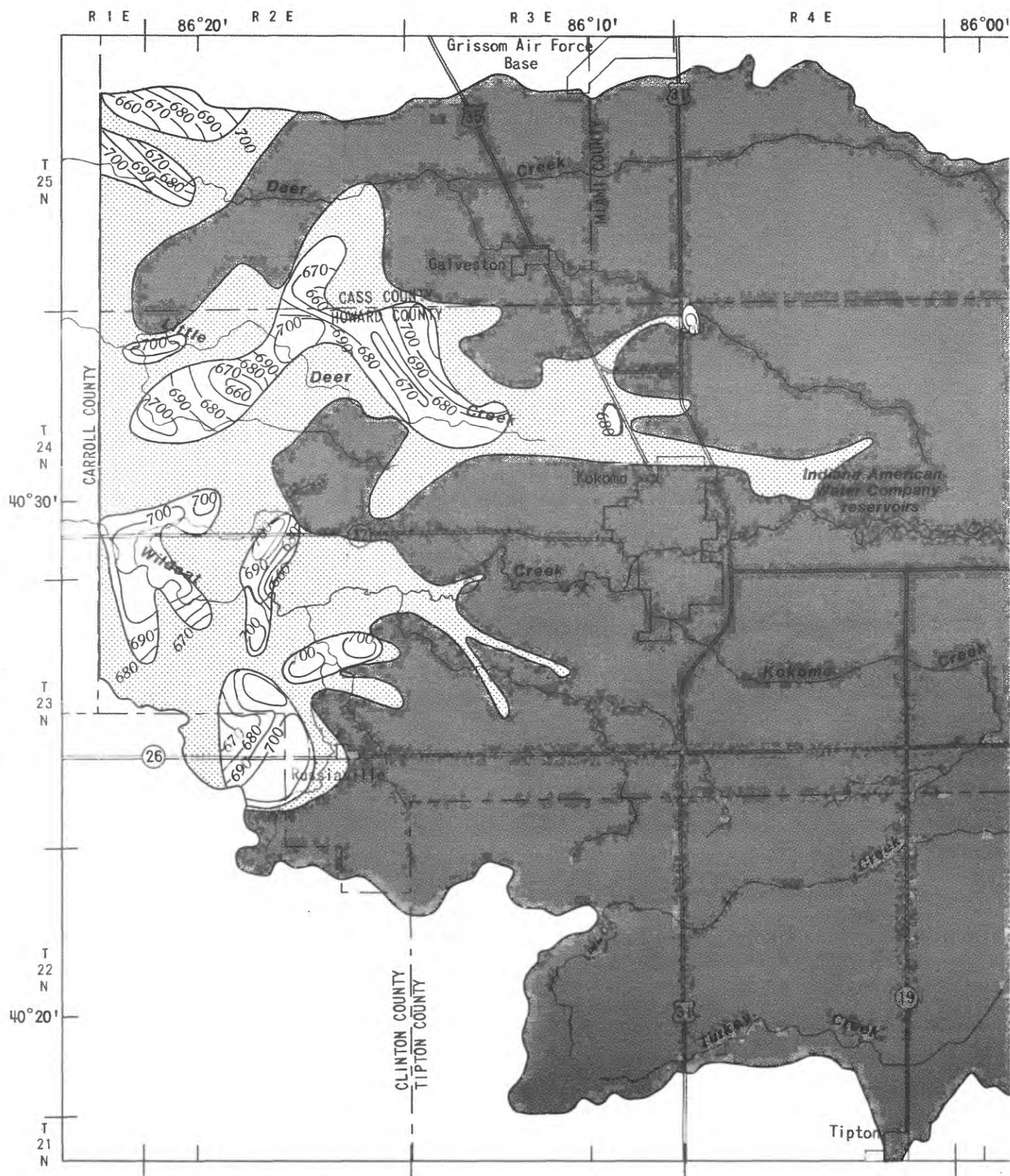
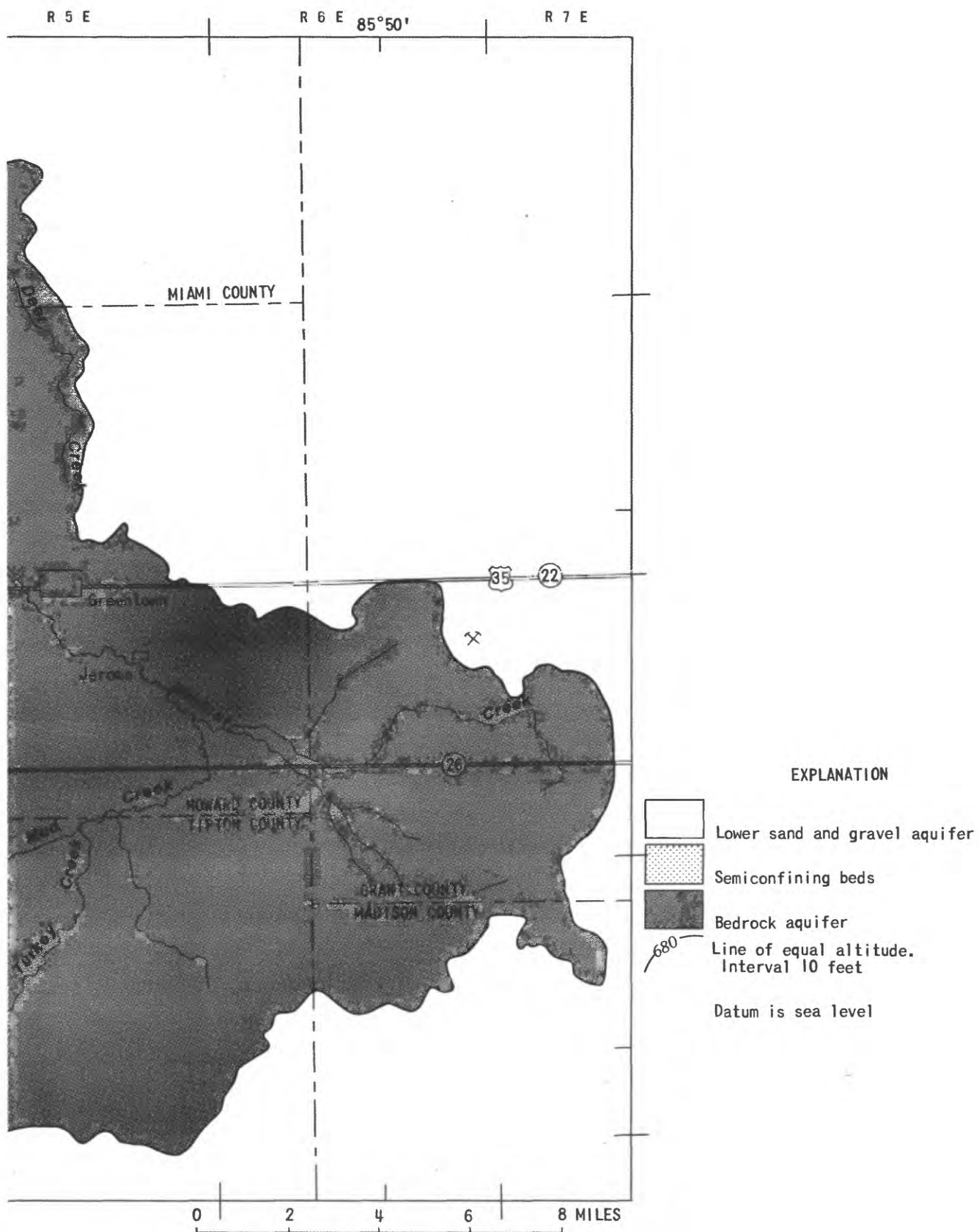


Figure 13.-- Altitude of top of the lower sand and gravel aquifer.



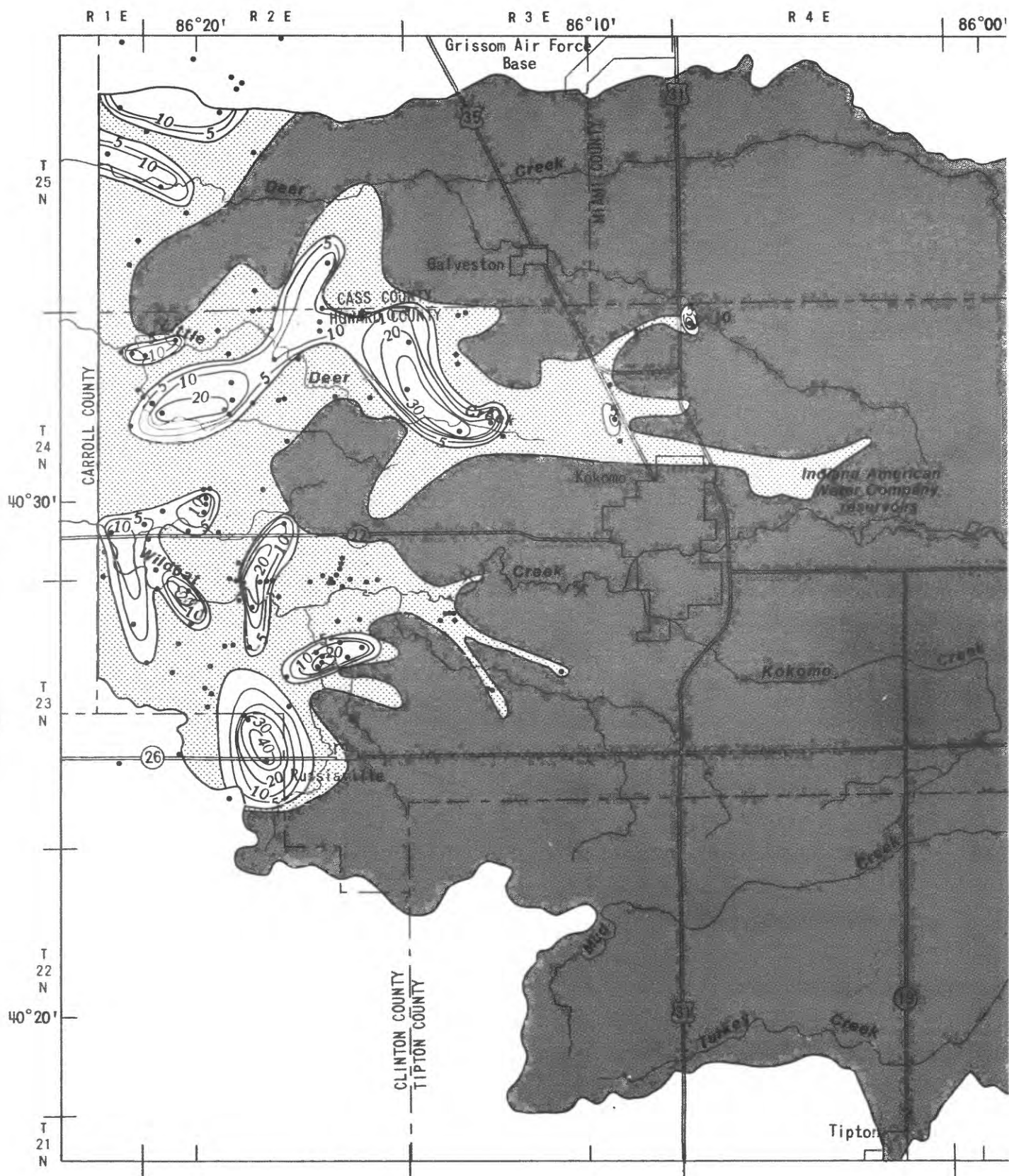


Figure 14.-- Thickness of the lower sand and gravel aquifer.

Hydraulic Characteristics of Aquifers

Transmissivity of the bedrock aquifer (fig. 15) was estimated at more than 200 sites from specific-capacity or aquifer-test data that were collected and analyzed by Angel Martin, Jr. (U.S. Geological Survey, oral and written commun., 1982). Transmissivities based on specific-capacity data applied to the method described in Brown (1963) were corrected for partial penetration of the aquifer by the method described by Butler (1957, p. 160). The ratio of horizontal to vertical hydraulic conductivity was assumed to be 1 in the calculations (Angel Martin, Jr., U.S. Geological Survey, oral and written commun., 1982). Transmissivities calculated from aquifer tests or specific-capacity tests pertain to local conditions. Depending on the number and the size of openings in the bedrock intercepted by pumping and observation wells, local transmissivities may differ over short distances.

For the bedrock immediately north of the area of study at Grissom Air Force Base, transmissivities ranging from 13,000 to 46,000 (gal/d)/ft or 1,700 to 6,200 ft²/d were estimated from controlled aquifer tests based on the Theis nonequilibrium equation (Watkins and Rosenshein, 1963, p. B12 and B23). A regional or long-term transmissivity of 9,000 (gal/d)/ft or 1,200 ft²/d was also calculated at Grissom from analysis of flow nets by use of Darcy's equation (Watkins and Rosenshein, 1963, p. B13 and B23).

Virtually no data were available for determining storage coefficient in the area of study. Storage coefficients for the bedrock, ranging from 1×10^{-5} to 2×10^{-3} , however, were estimated from controlled aquifer tests at Grissom Air Force Base (Watkins and Rosenshein, 1963, p. B23). Those values are probably applicable to the area of study.

Transmissivities of the confined sand and gravel aquifers were estimated by applying specific-capacity data to the method described in Brown (1963) and were corrected for partial penetration of the aquifer by use of the method described by Butler (1957, p. 160). The ratio of horizontal to vertical hydraulic conductivity was assumed to be 1 in the calculations (Angel Martin, Jr., U.S. Geological Survey, oral and written commun., 1982). Horizontal hydraulic conductivities of the confined sand and gravel aquifers were then determined by dividing transmissivities by the thickness of the aquifer penetrated. The average of 54 hydraulic conductivities, 200 ft/d, was used as initial input to the ground-water flow model. Hydraulic conductivities reported for similar sand and gravel aquifers in Hamilton and Tipton Counties (Arihood, 1982, p. 24) and in several counties in the upper White River basin (Cable and others, 1971, p. C- 11) were 216 and 200 ft/d. The horizontal transmissivity of the sand and gravel aquifers in Howard and adjacent counties can be calculated by multiplying the thickness of the sand and gravel aquifers (figs. 10, 12, and 14) by the hydraulic conductivity, 200 ft/d.

Hydraulic Characteristics of Semiconfining Beds

Hydraulic conductivities ranging from 3×10^{-4} to 1×10^{-3} ft/d were obtained by falling-head permeameter tests of four split-spoon samples of unconsolidated materials from two drill-hole sites shown in figure 4 (Sam Frushour, Indiana Geological Survey, written commun., 1982). Four samples from two sites are probably unrepresentative of the area of study, and hydraulic conductivity from laboratory analysis of the split-spoon samples may be unrepresentative of field conditions.

Vertical hydraulic conductivities of the semiconfining layers, ranging from 5×10^{-4} to 1×10^{-2} ft/d, were obtained by use of a three-dimensional ground-water flow model. Hydraulic conductivities from flow models represent regional conductivities. The local and regional hydraulic conductivities are within the range of conductivities [9×10^{-1} to 3×10^{-4} (gal/d)/ft² or 4×10^{-5} to 1×10^{-1} ft/d] reported for tills at various locations in Ohio, Illinois, and South Dakota (Norris, 1962).

The horizontal hydraulic conductivity of the semiconfining layers was not determined. However, on the basis of tests with the ground-water flow model, horizontal flow through the semiconfining layers was negligible in comparison with horizontal flow in the aquifers.

WATER QUALITY

Ground Water

Ground-water samples were collected from 30 wells (fig. 16). All observation wells except D4 and D11 were household water-supply wells. Thirteen of the wells were open to a sand and gravel aquifer, and the other 17 wells were open to limestone. Samples were analyzed for selected dissolved ions, nutrients, trace elements, and properties. National Interim Primary Drinking Water Regulations and National Secondary Drinking Water Regulations are cited for reference in table 3 (U.S. Environmental Protection Agency, 1982 and 1979).

Concentrations of selected dissolved ions and properties are shown in table 4. Hardness of the ground water ranged from 190 to 540 mg/L as calcium carbonate, an indication of extremely hard water (Durfor and Becker, 1962, p. 27). The type of water at all sites was calcium bicarbonate or calcium and magnesium bicarbonate, but chemical quality of water at sites 1A, 1B, 2B, and 9B is different from that at the other sites (fig. 17). Dissolved-solids concentrations at sites 1A, 1B, and 9B were 582 mg/L or greater (table 4). The primary ions causing the high dissolved-solids concentrations included calcium and sulfate at site 1A; calcium, sodium, chloride, and sulfate at site 1B; and

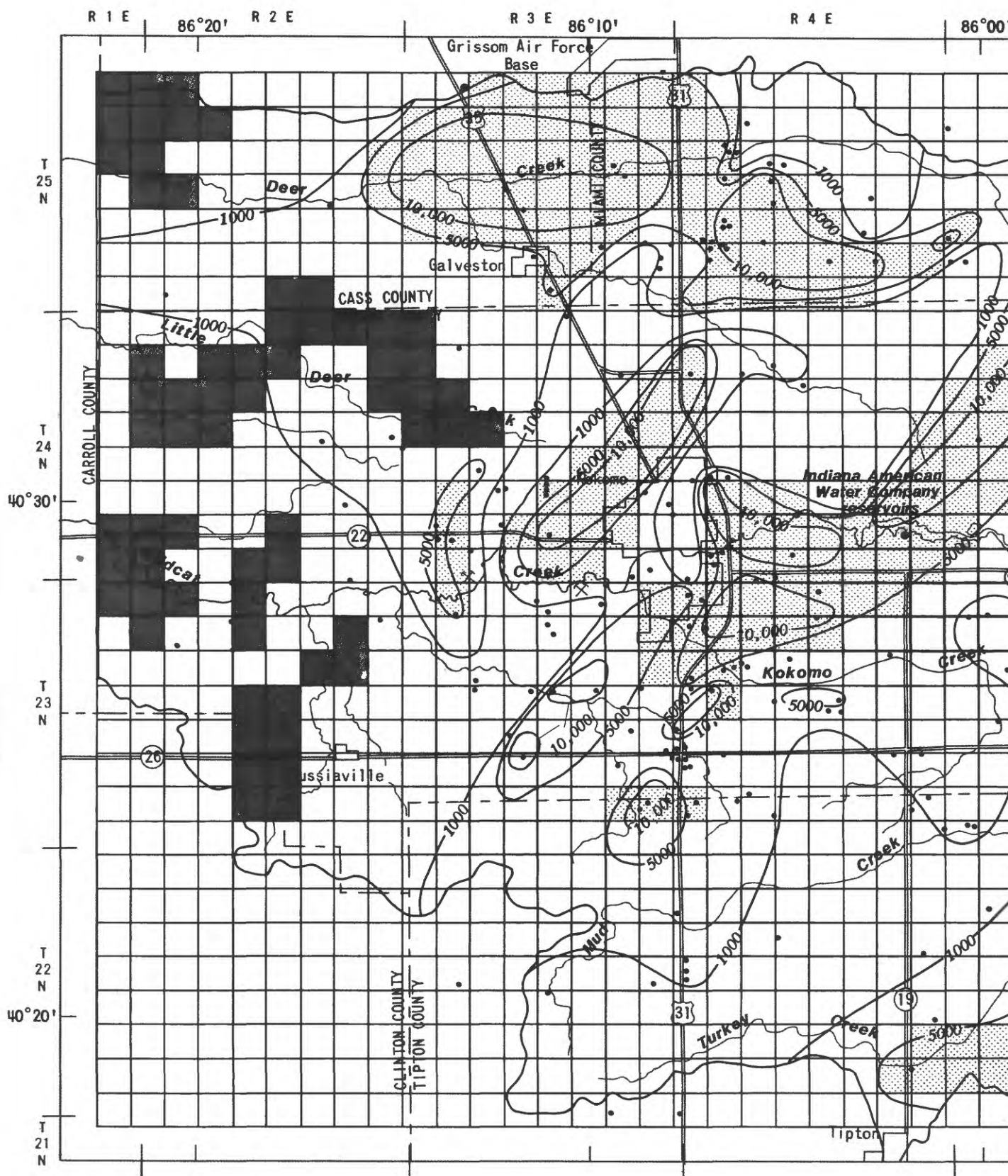


Figure 15.-- Transmissivity of the bedrock.

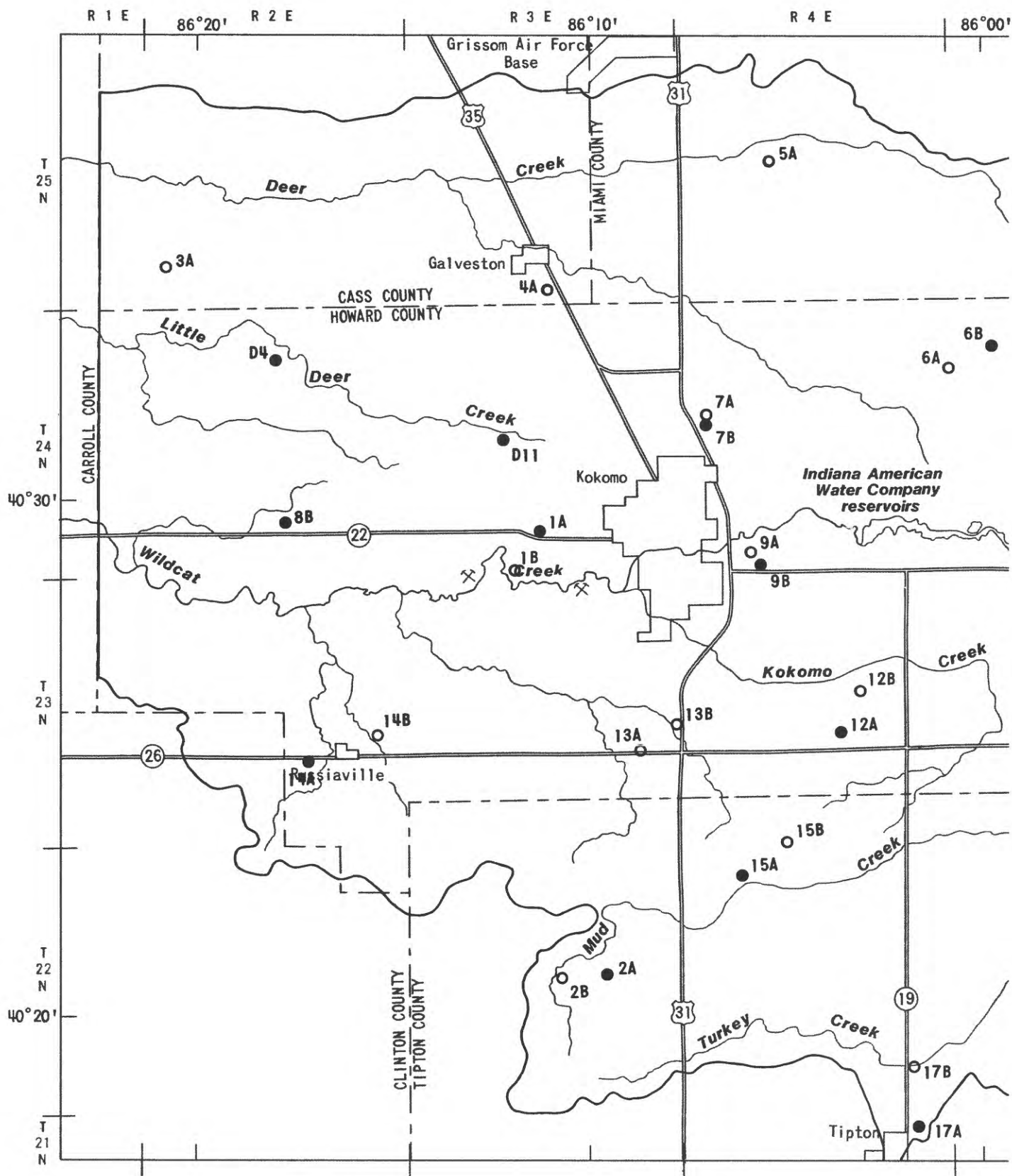
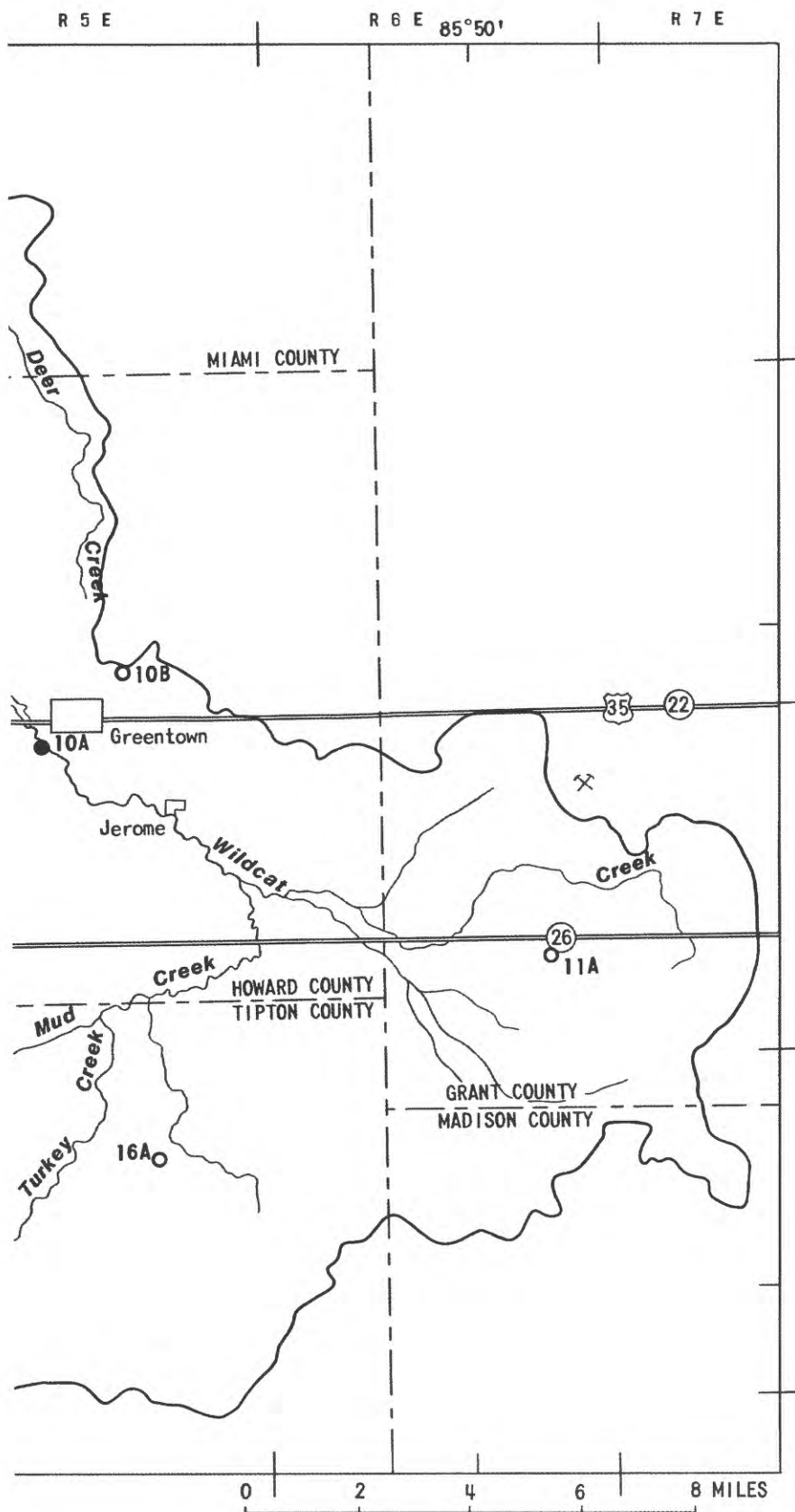


Figure 16.-- Ground-water-quality sampling sites.



EXPLANATION

- Well finished in unconsolidated aquifer.
- Well finished in bedrock aquifer
- 6B Well designation

Table 3.--Drinking-water regulations (maximums)

[mg/L, milligram per liter; N, nitrogen; TU, turbidity unit; pH, the negative logarithm of the effective hydrogen-iron concentration or hydrogen-ion activity, in gram equivalents per liter, used in expressing acidity and alkalinity on a scale from 0 to 14]

National Interim Primary Drinking-Water Regulations, amended through July 1, 1982
(U.S. Environmental Protection Agency, 1982)

Arsenic (mg/L)	Barium (mg/L)	Cadmium (mg/L)	Chromium (mg/L)	Fecal Coliform (counts/100 mL)	Fluoride (mg/L)	Lead (mg/L)	Mercury (mg/L)	Nitrate as N (mg/L)	Selenium (mg/L)	Silver (mg/L)	Turbidity (TU)
0.05	1	0.010	0.05	31 and 4	11.4 to 2.4	0.05	0.002	10	0.01	0.05	21 and 5

National Secondary Drinking-Water Regulations
(U.S. Environmental Protection Agency, 1979, p.2)

Chloride (mg/L)	Copper (mg/L)	Iron (mg/L)	Manganese (mg/L)	pH	Sulfate (mg/L)	Zinc (mg/L)	Total dissolved solids (mg/L)
250	1	0.3	0.05	6.5-8.5	250	5	500

- ¹Concentration varies with temperature: 1.4 mg/L in the range 79.3° to 90.5° F; 2.4 in the range 53.7° F and below.
- ²The maximum concentrations of turbidity are applicable to both community water systems and noncommunity water systems using surface water sources in whole or in part. The maximum concentrations of turbidity in drinking water, measured at a representative entry point(s) to the distribution system, are:
- (a) One turbidity unit (TU), as determined by a monthly average pursuant to section 141.22, except that five or fewer turbidity units may be allowed if the supplier of water can demonstrate to the State that the higher turbidity does not do any of the following:
- (1) Interfere with disinfection;
 - (2) Prevent maintenance of an effective disinfectant agent throughout the distribution system; or
 - (3) Interfere with microbiological determinations.
- (b) Five turbidity units based on an average for two consecutive days pursuant to section 141.22, turbidity sampling and analytical requirements.
- ³ The maximum counts of coliform bacteria, applicable to community water systems and noncommunity water systems, are as follows:
- (a) When the membrane filter technique pursuant to section 141.21(a) is used, the number of coliform bacteria shall not exceed any of the following:
- (1) One per 100 milliliters as the arithmetic mean of all samples examined per compliance period pursuant to section 141.21(b) or (c)...
 - (2) Four per 100 milliliters in more than one sample when less than 20 are examined per month; or
 - (3) Four per 100 milliliters in more than five percent of the samples when 20 or more are examined per month.
- [The reference for footnotes ¹, ², and ³ is U.S. Environmental Protection Agency (EPA), 1982].

Table 4.--Chemical analyses of ground-water samples
[Concentrations in milligrams per liter except for pH or as otherwise noted; mV, millivolt; CaCO₃, calcium carbonate]

Site	Date	Time ¹	Temp. of water (°C)	pH	Specific conductance (µmho/cm at 25° C)	Oxida- tion reduc- tion poten- tial (mV)	Alka- linity as CaCO ₃	Sil- ica	Iron ²	Manga- nese ²	Cal- cium	Magne- sium	Sol- dium	Potas- sium	Sul- fate ²	Chlo- ride ²	Fluo- ride ²	Total dis- solved solids ² (Resi- due on evap- at 180° C)	Hardness as CaCO ₃
1A	7-21-81	1045	12.8	7.0	910	-435	340	19	1.70	0.180	110	40	11	1.4	95	25	0.6	582	440
1B	7-21-81	1215	13.5	7.0	1,450	-435	330	16	2.00	.050	140	45	80	2.9	130	200	.3	902	540
2A	7-23-81	1045	13.0	7.6	740	35	340	16	2.70	.040	65	27	29	1.7	.0	2.7	1.2	353	270
2B	7-23-81	1220	12.7	7.4	690	81	350	10	.18	.006	59	33	46	4.6	1.6	14	1.5	417	280
3A	7-30-81	1150	12.5	7.2	550	115	280	18	1.10	.020	70	28	8.7	1.1	25	3.3	.6	341	290
4A	7-21-81	1400	13.0	7.2	638	-565	330	16	1.00	.030	73	29	10	1.5	.7	1.6	.8	337	300
5A	7-30-81	1410	13.1	7.1	630	141	330	13	.65	.020	76	31	12	2.7	22	2.5	1.0	373	320
6A	7-28-81	1605	12.8	7.4	600	90	300	18	1.40	.030	79	29	10	1.7	21	1.9	1.1	378	320
6B	7-29-81	1105	13.5	7.2	670	110	340	19	1.20	.040	81	31	11	1.9	23	2.2	1.0	382	330
7A	7-21-81	1135	12.7	7.4	590	125	310	16	1.60	.030	78	27	8.9	1.5	5.9	2.2	.6	342	310
7B	7-27-81	1415	15.8	7.5	600	111	270	18	2.10	.030	72	29	7.4	1.5	34	4.5	.7	355	300
8B	7-22-81	1155	12.9	7.4	569	-5	320	14	2.90	.040	59	30	20	2.0	.0	1.8	1.4	338	270
9A	7-24-81	1355	12.1	7.2	710	148	310	16	2.10	.030	89	30	15	1.9	36	24	.9	432	350
9B	7-27-81	1620	12.7	7.3	940	115	310	16	2.10	.050	97	37	33	2.5	38	80	.8	589	390
10A	7-28-81	1245	13.3	7.3	685	88	300	15	2.20	.060	85	30	6.9	1.4	49	5.4	.5	413	340
10B	7-28-81	1450	12.4	7.2	650	131	330	12	.50	.010	79	31	12	3.9	24	4.0	1.2	381	330
11A	7-29-81	1255	12.4	7.2	695	100	320	13	1.70	.030	74	37	17	2.5	54	2.5	1.2	430	340
12A	7-24-81	1120	12.3	7.4	595	121	330	16	1.10	.008	70	27	24	2.4	.0	6.4	.8	356	290
12B	7-24-81	1235	12.5	7.3	570	154	300	.1	.01	.001	68	28	22	2.5	6.0	7.3	1.2	345	290
13A	7-20-81	1600	14.0	7.2	560	37	300	13	1.30	.040	62	28	19	2.2	.0	7.2	1.2	330	270
13B	7-24-81	1010	12.8	7.3	670	38	310	13	.01	.040	74	32	15	3.0	14	20	1.2	380	320
14A	7-22-81	1400	12.0	7.3	655	-15	300	18	2.80	.070	72	26	15	1.0	24	7.3	.9	358	290
14B	7-22-81	1535	12.5	7.3	570	-75	330	12	.32	.008	61	33	33	4.0	2.3	17	1.5	339	290
15A	7-23-81	1615	13.4	6.8	970	200	510	32	9.40	.120	100	33	18	2.4	<1.0	2.6	.7	511	390
15B	7-29-81	1615	13.2	7.1	665	157	370	13	.36	.040	68	31	28	4.1	.0	4.9	.9	391	300
16A	7-29-81	1415	12.1	7.3	670	94	360	14	1.30	.050	67	31	31	2.5	18	1.7	1.2	395	300
17A	7-23-81	1400	12.1	7.5	770	140	340	21	1.60	.020	95	34	14	1.2	55	15	.7	482	380
17B	7-28-81	1500	12.5	7.5	630	140	350	20	1.30	.001	74	27	20	1.4	.0	1.6	1.0	371	300
D4	7-30-81	1030	13.3	7.1	620	100	350	18	2.00	.330	67	27	19	1.5	.3	1.7	.8	343	280
D11	7-17-81	1500	14.4	7.5	580	106	330	13	1.60	.080	49	17	5.7	1.3	1.3	1.4	.6	358	190

¹ Twenty-four-hour time. For example, 1415 hours is 2:15 p.m.

² See table 3 for maximum allowable concentrations in drinking water recommended by EPA (1979 and 1982).

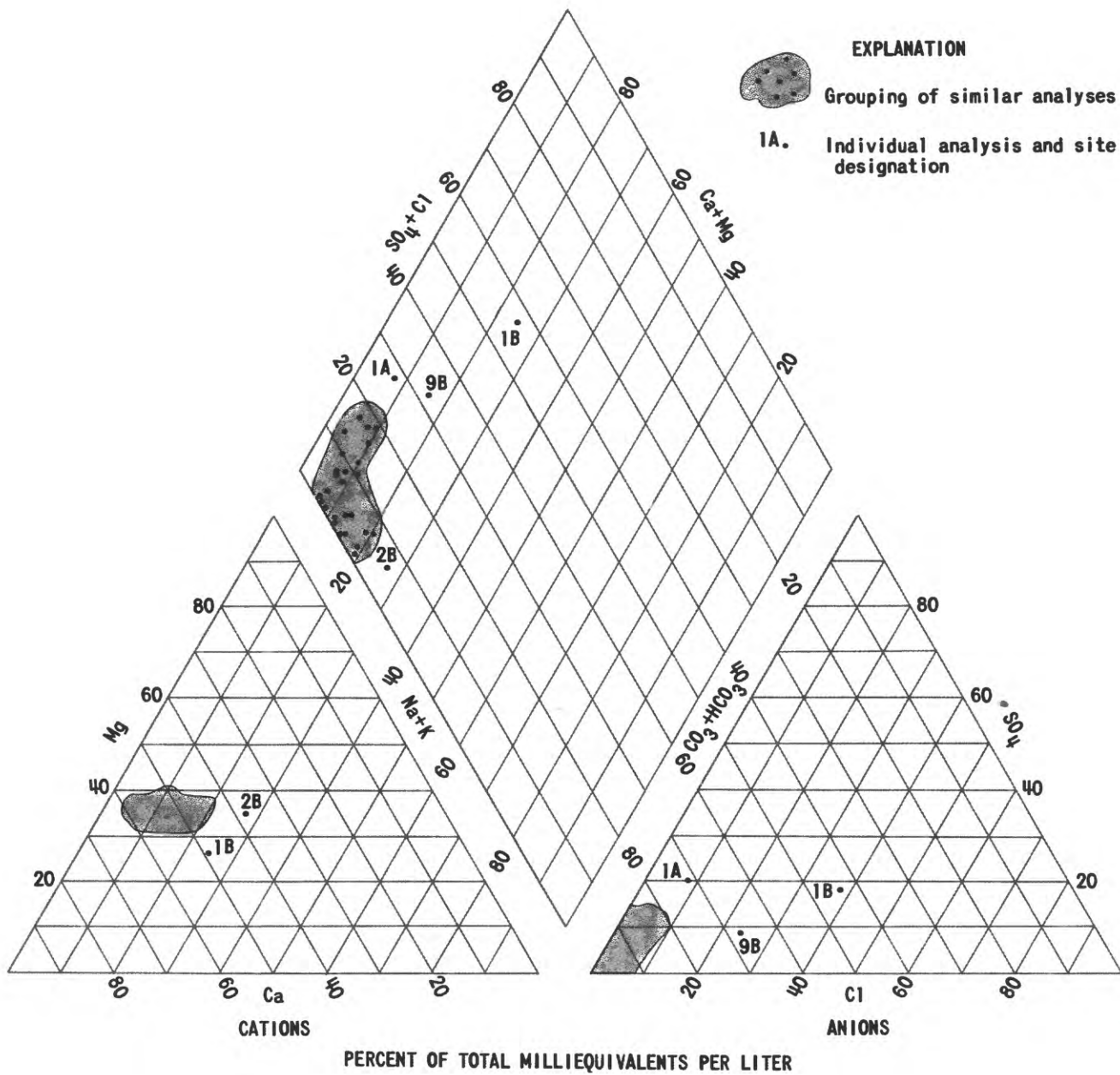


Figure 17.-- Ground-water analyses represented by trilinear diagram.

calcium, magnesium, and chloride at site 9B. Although the concentration of sodium at site 2B was one of the highest (46 mg/L), concentrations of calcium and sulfate ranked among the lowest (59 and 1.6 mg/L).

Data are insufficient for determining the causes of the water-quality differences. Locally, sulfide, sulfate, and (or) organic materials in aquifers may be the causes of a high concentration of sulfate. Site 9B is in a crowded residential area where septic tanks and salt used on roads might be sources of chloride.

Data were examined statistically to determine if there were water-quality differences between the two types of aquifers. Statistical tests were done on normalized data at 95-percent confidence. Mean concentration of dissolved silica (18 mg/L) was higher, and mean concentration of dissolved potassium (1.6 mg/L) was lower (fig. 18) in water from the unconsolidated aquifer (sand and clay) than in water from the limestone aquifer (13 and 2.5 mg/L). Contact time of water in the unconsolidated aquifer was longer than contact time in the limestone. The sands are probably the major sources of silica, whereas some clays commonly adsorb potassium from solution.

Concentration of dissolved iron commonly exceeded 0.3 mg/L. Concentrations greater than 0.3 mg/L result in taste and staining problems in domestic water supplies (U.S. Environmental Protection Agency, 1979, p. 2 and 24). Mean concentration of iron in water from the unconsolidated aquifers was significantly higher (2.2 mg/L) than in water from the limestone (0.50 mg/L). Some well screens could be plugged by precipitation of iron hydroxide in the unconsolidated aquifers.

Some concentrations of manganese exceeded 0.05 mg/L. Concentrations greater than 0.05 mg/L result in taste and staining problems in domestic water supplies (U.S. Environmental Protection Agency, 1979, p. 2 and 26). Mean concentrations of manganese in the two types of aquifers were not significantly different.

Concentrations of dissolved nutrients are shown in table 5. Concentrations were highest south and west of Kokomo. At sites 2A, 2B, 8A, 12B, 13A, 14A, 14B, 15A, 15B, 17A, 17B, and D4, ammonium concentration exceeded 0.5 mg/L as nitrogen, the maximum concentration recommended for public water supplies (National Academy of Science and the National Academy of Engineering, 1972, p. 55). In addition to having the highest concentration of ammonium (22 mg/L as nitrogen), site 15A had the highest concentrations of organic nitrogen (7 mg/L), phosphorus (0.28 mg/L), and organic carbon (12 mg/L). The source of high concentrations of organic components is possibly waste disposal at local livestock farms. The high ammonium concentrations at the other sites might have been caused by disposal of livestock wastes and (or) applications of ammonia and nitrification inhibitors to agricultural fields. Generally, high concentrations of ammonium in ground water are associated with disposal of livestock wastes rather than with applications of ammonia to fields (Stewart and others, 1967; Chichester, 1976). The counts of fecal coliform bacteria at sites 2B, 15A, 15B, and 17B in 1982 and 1983 were zero.

Concentrations of trace elements are shown in table 6. None of the concentrations exceeded maximum concentration recommended by the U.S. Environmental Protection Agency (1979 and 1982), and no trends in concentrations were evident for sites or aquifers. (See table 3.)

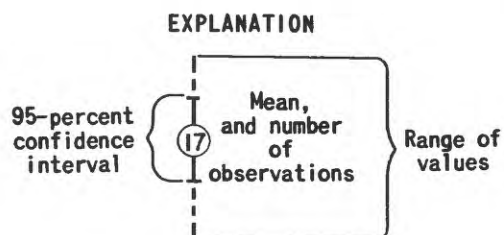
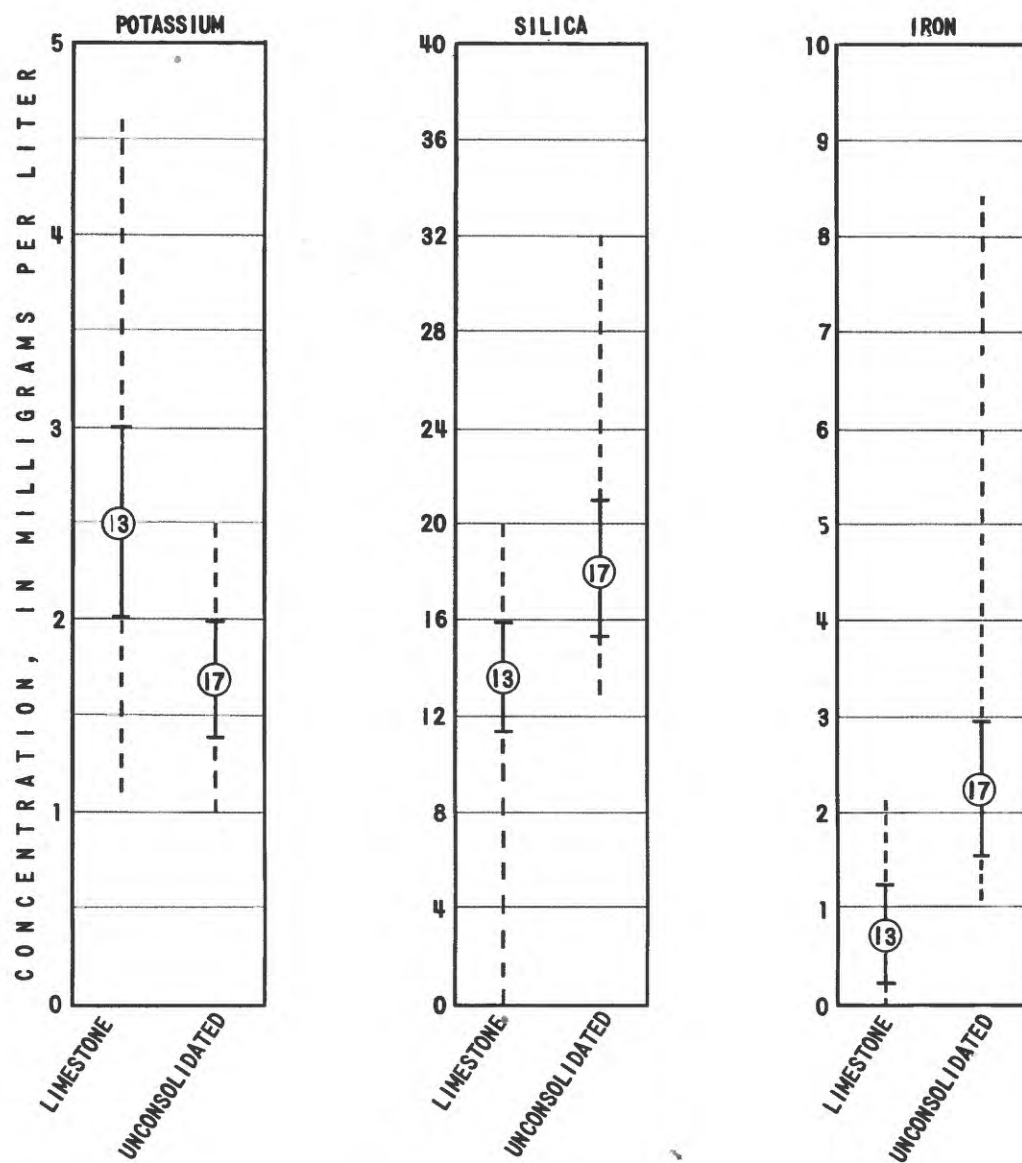


Figure 18.-- Concentrations of potassium, silica, and iron in limestone and unconsolidated aquifers.

Table 5.--Nutrients dissolved in ground-water samples
[All concentrations in milligrams per liter]

Site	Date	Time ¹	Nitrite as nitrogen	Nitrite plus nitrate ² as nitrogen	Ammonium as nitrogen	Organic nitrogen	Total nitrogen	Ortho- phosphate as phosphorus	Total phosphorus	Dissolved organic carbon
1A	7-21-81	1045	<0.01	<0.01	0.19	0.00	0.20	<0.01	<0.01	3.8
1B	7-21-81	1215	< .01	< .01	.15	.15	.30	< .01	< .01	1.1
2A	7-23-81	1045	< .01	.01	1.70	.10	1.80	.11	.11	2.8
2B	7-23-81	1220	< .01	< .01	.70	.30	1.00	< .01	< .01	5.7
3A	7-30-81	1150	< .01	.01	.08	.07	.16	< .01	< .01	.9
4A	7-21-81	1400	< .01	.01	.27	.23	.51	< .01	< .01	1.2
5A	7-30-81	1410	< .01	< .01	.24	.01	.25	< .01	< .01	.9
6A	7-28-81	1605	< .01	.01	.15	.01	.17	< .01	< .01	1.3
6B	7-29-81	1105	< .01	.01	.21	.17	.39	< .01	< .01	1.1
7A	7-21-81	1135	< .01	.01	.23	.06	.30	< .01	.03	1.3
7B	7-27-81	1415	< .01	.01	.36	.15	.52	< .01	< .01	1.1
8A	7-22-81	1155	< .01	< .01	.61	.07	.69	.05	.06	3.7
9A	7-24-81	1355	< .01	.01	.19	.26	.46	< .01	< .01	1.1
9B	7-27-81	1620	< .01	< .01	.33	.14	.48	< .01	< .01	.9
10A	7-28-81	1245	< .01	.01	.12	.32	.45	< .01	< .01	.9
10B	7-28-81	1450	< .01	< .01	.31	.00	.32	< .01	< .01	1.4
11A	7-29-81	1255	< .01	.01	.30	.05	.36	< .01	< .01	.9
12A	7-24-81	1120	< .01	.01	.44	.15	.60	< .01	.03	3.6
12B	7-24-81	1235	< .01	.02	.70	.00	.72	< .01	.13	1.8
13A	7-02-81	1600	< .01	.01	1.60	.10	1.70	.05	.05	1.6
13B	7-24-81	1010	< .01	.01	.50	.00	.51	.04	.04	.8
14A	7-22-81	1400	< .01	< .01	1.90	.00	1.90	.04	.07	1.4
14B	7-22-81	1535	< .01	.01	.87	.08	.96	< .01	.01	2.0
15A	7-23-81	1615	< .01	.01	22	7.00	29	.21	.28	12.0
15B	7-29-81	1615	< .01	< .01	2.20	.00	2.20	.01	.03	4.1
16A	7-29-81	1415	< .01	< .01	.50	.27	.78	< .01	.02	2.3
17A	7-28-81	1400	< .01	< .01	1.40	.30	1.70	.06	.06	1.5
17B	7-28-81	1500	< .01	< .01	1.50	.00	1.50	< .01	.03	3.9
D4	7-30-81	1030	< .01	.01	3.70	.00	3.70	3.70	.03	2.5
D11	7-17-81	1500	< .01	< .01	.33	.00	.33	< .01	< .01	2.0

¹Twenty-four hour time. For example, 1415 hours is 2:15 p.m.

²See table 3 for maximum allowable concentration of nitrate as nitrogen in drinking water recommended by U.S. Environmental Protection Agency (1982).

Table 6.--Trace elements dissolved in ground-water samples
[All concentrations in micrograms per liter]

Site	Date	Time ¹	Arsenic	Barium	Cadmium	Chromium	Copper	Lead	Mercury	Selenium	Silver	Zinc
1A	7-21-81	1045	8	200	1	<10	1	1	<0.1	<1	<1	8
1B	7-21-81	1215	0	300	1	10	1	2	.1	<1	<1	10
2A	7-23-81	1045	6	400	1	<10	1	5	.1	<1	<1	70
2B	7-23-81	1220	1	300	1	<10	1	2	.2	<1	<1	5
3A	7-30-81	1150	8	400	1	10	1	2	<.1	<1	<1	10
4A	7-21-81	1400	5	300	2	10	1	1	<.1	<1	<1	6
5A	7-30-81	1410	3	100	1	<10	<1	2	<.1	<1	<1	20
6A	7-28-81	1605	7	200	1	<10	1	2	<.1	<1	<1	5
6B	7-29-81	1105	9	300	1	<10	<1	1	<.1	<1	<1	40
7A	7-21-81	1135	2	200	1	10	1	2	<.1	<1	<1	10
7B	7-27-81	1415	2	200	1	<10	1	<1	<.1	<1	<1	30
8B	7-22-81	1155	0	400	1	10	1	<1	.1	<1	<1	4
9A	7-24-81	1355	2	300	2	10	1	2	.1	<1	<1	20
9B	7-27-81	1620	1	300	1	<10	1	1	<.1	<1	<1	5
10A	7-28-81	1245	4	200	1	<10	1	<1	<.1	<1	<1	70
10B	7-28-81	1450	2	100	1	10	1	<1	<.1	<1	<1	4
11A	7-29-81	1255	9	100	1	<10	1	4	<.1	<1	<1	8
12A	7-24-81	1120	0	500	1	<10	1	<1	<.1	<1	<1	4
12B	7-24-81	1235	1	200	1	10	<1	<1	.2	<1	<1	4
13A	7-20-81	1600	4	200	1	10	1	<1	<.1	<1	<1	8
13B	7-24-81	1010	1	200	1	<10	1	2	.1	<1	<1	4
14A	7-22-81	1400	3	100	1	10	1	1	<.1	<1	<1	4
14B	7-22-81	1535	1	300	1	10	1	2	.2	<1	<1	4
15A	7-23-81	1615	1	300	2	10	1	1	<.1	<1	<1	4
15B	7-29-81	1615	0	100	1	10	1	<1	<.1	<1	<1	4
16A	7-29-81	1415	9	200	1	10	1	<1	<.1	<1	<1	10
17A	7-23-81	1400	1	500	1	<10	<1	<1	<.1	1	<1	10
17B	7-23-81	1500	14	400	2	10	1	1	<.1	<1	<1	50
D4	7-30-81	1030	13	200	1	10	<1	1	<.1	<1	<1	20
D11	7-17-81	1500	12	200	1	10	1	5	<.1	<1	<1	9

¹Twenty-four hour time. For example, 1415 hours is 2:15 p.m.

Surface Water

Urban development is probably the greatest factor affecting surface-water quality in the area around Kokomo. The Indiana State Board of Health measured water quality of Wildcat Creek at two sites monthly from March through December 1982--site WC 69 at the upstream (east) edge of Kokomo and site WC 63 1 mile downstream from the west edge, (table 7 and fig. 4). This stream and its tributaries drain the area in and around Kokomo. Average concentrations of most of the constituents measured at the downstream site were higher than those measured at the upstream site (Indiana State Board of Health, 1982, p. 103-104). The higher concentrations at the downstream site can be attributed to urban runoff and industrial and municipal wastewater discharged into Wildcat Creek.

In a study of the dissolved-oxygen concentration of Wildcat Creek in and downstream from Kokomo, Crawford and others (1979) concluded that the major factor influencing the concentration of oxygen dissolved in the creek was benthic-oxygen demand (BOD)¹--oxygen consumed during the decay of organic deposits in the streambed. Benthic-oxygen demand was greatest downstream from the wastewater treatment plant in Kokomo.

Personnel of Indiana Department of Natural Resources measured water quality of Wildcat Creek during a survey of fish population downstream from Kokomo before, during, and after a 2-inch rain. On the basis of the measurements, the investigators concluded that urban runoff and combined-sewer overflow containing high concentrations of sediment and dissolved constituents can temporarily degrade surface-water quality (Edward Braun, Indiana Department of Natural Resources, oral commun., 1984).

Specific conductance of Wildcat Creek was reported to be less than 800 $\mu\text{mho}/\text{cm}$ for the entire creek, except for an approximately 5-mile reach west of Kokomo, where the conductance was greater than 800 $\mu\text{mho}/\text{cm}$ (Marie and Davis, 1974). The dissolved-solids concentration (in milligrams per liter) of streams in the area around Kokomo is about 68 percent of the specific conductance (in micromho per centimeter), according to Marie and Davis (1974). Specific conductance reported by the Indiana State Board of Health (1982, p. 103-104) was 488 $\mu\text{mho}/\text{cm}$ at upstream site WC 69 and 656 $\mu\text{mho}/\text{cm}$ at downstream site WC 63. Dissolved-solids concentrations at the two sites were 311 and 403 mg/L (Indiana State Board of Health (1982, p. 103-104). Some idea of the water quality of Wildcat Creek is evident by a comparison of data in table 7 with the standards in table 3. For example, concentrations of turbidity, iron, and manganese, as well as the counts of fecal coliform, exceed the drinking-water standards given in table 3; but the other concentrations for which standards are given are less than those of the standards.

¹In this report, the acronym BOD is used to abbreviate the term "benthic-oxygen demand"; this is different usage from the ordinary definition of BOD as "biochemical oxygen demand".

Table 7.--Average water quality of Wildcat Creek near Kokomo, March 10 through December 13, 1982

[Data from Indiana State Board of Health (ISBH), Division of Water Pollution Control, 1982 Water Quality Annual Report, p. 103-104; mg/L, milligram per liter; µg/L, microgram per liter; see table 3 for standards of variables that are underlined; BOD, benthic-oxygen demand; COD, chemical-oxygen demand; mL, milliliter; DO, dissolved oxygen; NO₂+NO₃, nitrite plus nitrate as nitrogen; PCB, polychlorinated biphenyl; pH, the negative logarithm of the effective hydrogen-ion concentration or hydrogen-ion activity, in gram equivalents per liter, used in expressing acidity and alkalinity on a scale from 0 to 14; µmho/cm, micromho per centimeter; spec. cond., specific conductance; Temp °C, temperature in degrees Celsius; TOC, total organic carbon; NTU, national turbidity unit; CaCO₃, calcium carbonate; and MgCO₃, magnesium carbonate]

ISBH Station ¹	Alkalinity (mg/L)	Ammonia (mg/L)	Arsenic (µg/L)	BOD (mg/L)	COD (mg/L)	Cadmium (µg/L)	Chloride (mg/L)	Chromium, total (µg/L)	Copper (µg/L)	Cyanide (mg/L)	DO (mg/L)
WC63	186	0.53	4	3.8	21	2.0	50	13	15	0.008	8.9
WC69	166	.18	2	3.1	17	2.0	30	10	5	.006	9.8

ISBH station ¹	Fluoride (mg/L)	Hardness (mg/L)	Iron (µg/L)	Lead (µg/L)	Manganese (µg/L)	Mercury (µg/L)	Nickel (µg/L)	Nitrite plus Nitrate (mg/L)	Nitrogen (mg/L)	Oil and grease (mg/L)	pH field	pH lab	Phosphorus (mg/L)	Potassium (mg/L)	Sodium (mg/L)
WC63	0.4	307	1,527	15	117	0.100	11	4.4	1.3	3.1	7.7	7.8	0.12	3.4	28.3
WC69	.4	242	1,066	11	93	.100	10	3.9	.9	2.9	7.8	7.8	.10	2.2	10.9

ISBH station ¹	Suspended solids (mg/L)	Spec cond (µmho/cm)	Sulfate (mg/L)	Temp (°C)	TOC (mg/L)	Turbidity (NTU)	Zinc (µg/L)	Calcium as CaCO ₃ (mg/L)	Magnesium as MgCO ₃ (mg/L)	Dissolved solids (mg/L)	Fecal streptococci (count/100 mL)	Fecal coliform (count/100 mL)
WC63	25	656	85.0	16	6.3	36	138	204	103	403	1,405	6,382
WC69	22	488	42.2	15	5.7	31	26	160	82	311	723	3,322

¹Station WC 63 is downstream from Kokomo. Each value is the average of 10 measurements.
Station WC 69 is upstream from Kokomo. Each value is the average of 10 measurements.

Although urban development is a major factor in the degrading of the water quality of Wildcat Creek, the quality of the creek improves with distance downstream from Kokomo, owing to an increasing assimilative capacity and dilution of the stream. Wildcat Creek is generally potable just upstream from its confluence with the Wabash River (Stephen Boswell, Indiana State Board of Health, oral commun., 1984).

SIMULATION OF GROUND-WATER FLOW

Design and Construction of Model

A three-dimensional, finite-difference model (McDonald and Harbaugh, 1984) was used to simulate the ground-water flow system of upper Wildcat Creek and Deer Creek basins. In this model, the ground-water flow system is divided into discrete, interconnecting blocks, and the variables of the general equation for ground-water flow (hydraulic characteristics, dimensions of blocks, rates of recharge and pumping, boundary, and initial water level) are assigned to the center (node) of each block. Blocks for streams were also designated, and the appropriate characteristics and dimensions were assigned.

One of the mathematical algorithms in the model (the strongly implicit procedure) was chosen for solving the interconnected system of block-centered flow equations. The system of equations containing unknown water levels is solved simultaneously, and final water level is calculated for the center of each block (node) on the basis of the variables in and around the node. Technically, the variable that is solved in the system of equations is the water level at each node. Practically, however, the validity of the variables (including the water level at any node) depends on the credibility and the availability of data in and around the block. Therefore, with the model, the data are assimilated into one interrelated system, and the modeler can test and adjust the less credible variables against the more credible ones.

Upper Wildcat Creek and Deer Creek basins were divided areally into squares (blocks) 4,000 ft on a side (fig. 19). Boundaries of the model are approximately the same as boundaries of the basin, except for the west boundary. Constant-head boundaries were used in the nodes of the perimeter during calibration. Flow from the boundaries and from effective recharge calculated by the model from the constant heads were subsequently used as constant-flux boundaries in sensitivity analysis and in simulations of hypothetical well fields.

Locations of blocks on streams were approximately the same as the locations of streams and ditches, and locations of blocks at pumping sites were approximately the same as the locations of major ground-water-pumping sites. Vertical leakage through each block on a stream is calculated by the model on the basis of the difference between stage of stream and the water level of the aquifer at the node. Hydraulic conductivity and the dimensions of the stream are lumped together in the model.

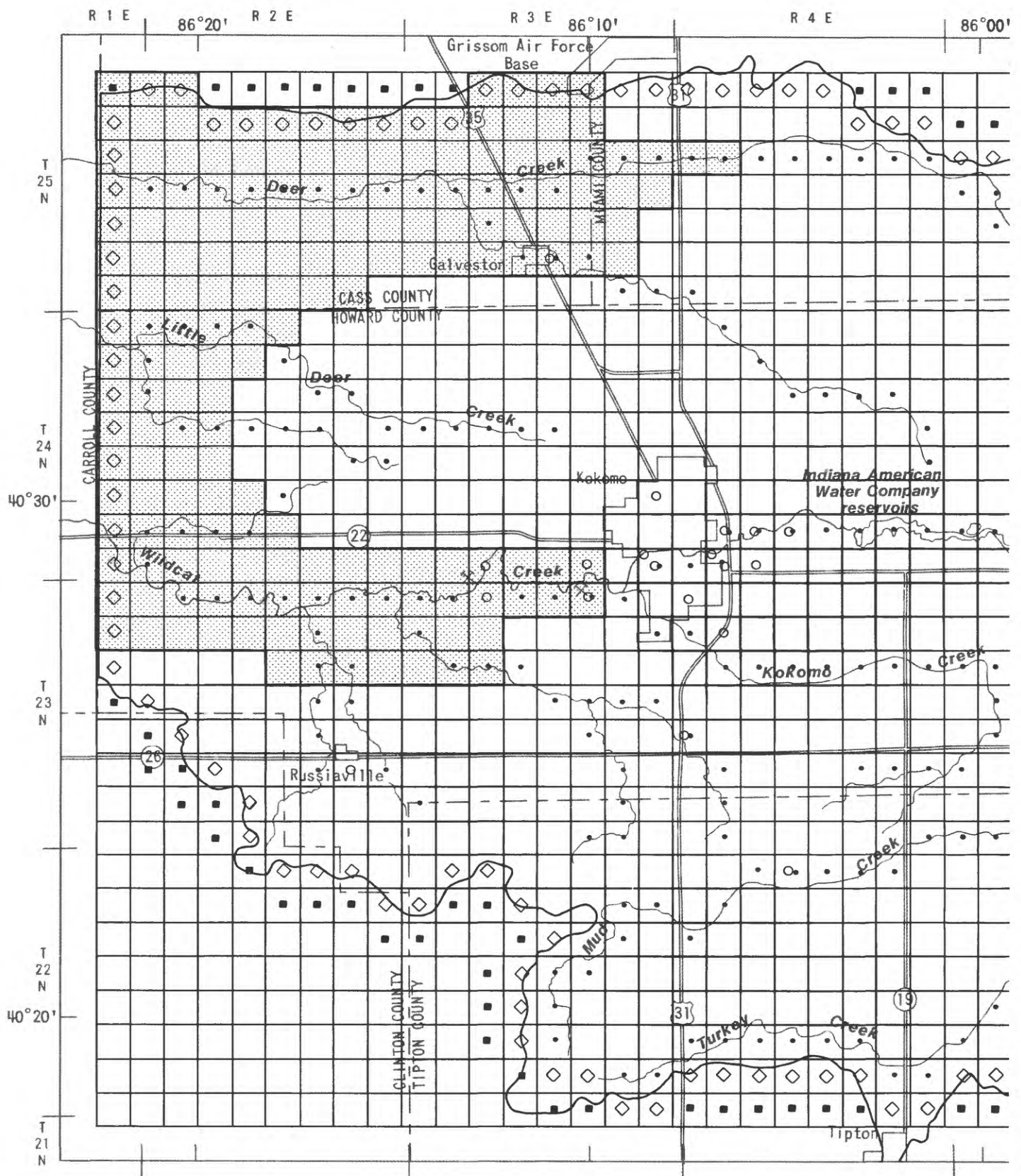
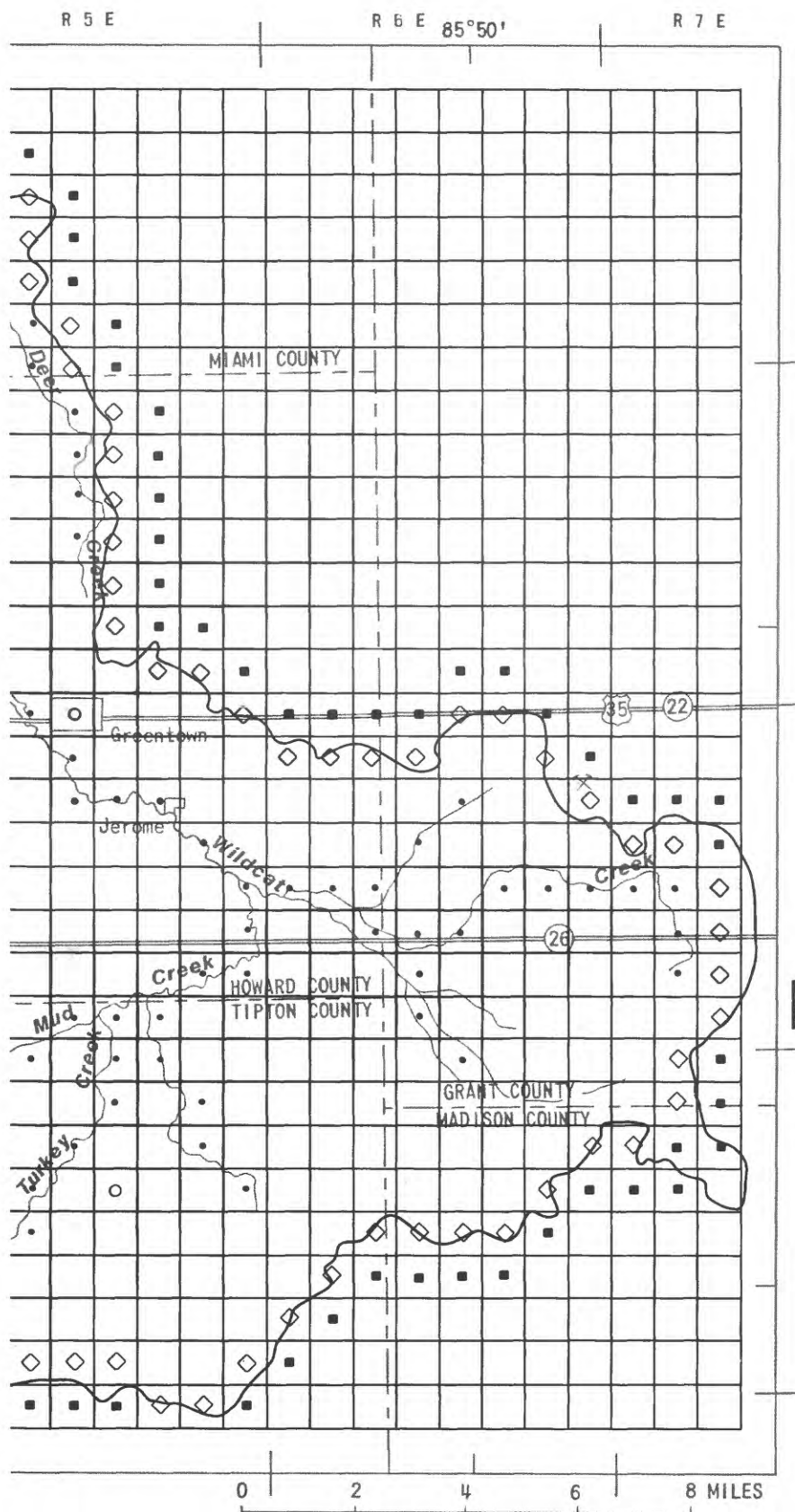







Figure 19.-- Finite-difference grid of model used to simulate ground-water flow.



EXPLANATION

-  Layer 1 absent
-  Active model boundary. Layers 2, 3, and 4
-  Inactive model boundary
-  Stream node
-  Pumping node

The area of study was divided vertically into four layers of varying dimensions separated by three vertical-leakage planes of varying dimensions. Water-table (unconfined) flow in the topmost semiconfining beds was generally simulated in the top layer, horizontal flow in the upper and middle sand and gravel aquifers was simulated in the second and third layers, and horizontal flow in the lower sand and gravel aquifer and in the bedrock aquifer was simulated in the fourth layer. Horizontal conductivities of less than 1 ft/d were simulated wherever the aquifer was missing in the layers, but the model was not sensitive to this variable because horizontal flow through the semiconfining beds is negligible.

Vertical leakage through the semiconfining beds and the aquifers between the four horizontal layers was simulated in three vertical-leakage planes. Vertical leakage between the layers is calculated in the model on the basis of the difference between the water levels in each node. The vertical conductivity and the thickness of the semiconfining bed (or aquifer) between layers are lumped together in the model; therefore, calculation of the vertical conductivity on the basis of thickness is, extraneous to the model.

Only the major properties of the ground-water flow system are simulated in the model. The properties are simulated on the basis of the data, the method available, and the problem posed. Thus, assumptions that will permit the simulation of the most important aspects of the flow system are needed. In simulating ground-water flow and in evaluating water resources, the authors made the following assumptions:

1. Ground water is incompressible, and its density is constant from one location to another.
2. Ground-water flow is laminar.
3. The principal components of hydraulic conductivity are aligned with the three-dimensional, coordinate axis of the finite-difference grid.
4. Hydraulic characteristics do not change with the direction of measurement.
5. Sand and gravel aquifers are confined.
6. The bedrock is a confined aquifer whose transmissivity varies areally.
7. The vertical hydraulic conductivity of semiconfining layers varies areally.
8. Ground-water storage in the semiconfining layers is negligible compared with that in the aquifers.
9. Flow into or out of storage in the aquifers is instantaneous with changes in water level.

10. Recharge to or discharge from a node is instantaneous through the thickness of the layer.
11. Evapotranspiration will not be decreased by lowering water levels during pumping, and, therefore, effective recharge will not be increased during pumping.

Calibration and Sensitivity of Model

The model was calibrated to water levels measured in domestic and commercial wells from May through July 1980, water levels measured in observation wells, and water levels determined from seismic data from June through July 1981. Natural ground-water levels do not fluctuate more than plus or minus 2 ft in Howard and adjacent counties, and the water levels are virtually at steady state. Measured and model-simulated water levels are shown in figures 20 through 23. The model was also calibrated to streamflow gains and losses measured July 11-13, 1981, and again August 20-21, 1981 (table 8). Because of errors inherent in the measurement of streamflow, minimum and maximum limits for the measured gains and losses (table 2) were used.

The model was calibrated by use of the root mean square error (RMSE) of the differences between 309 measured and corresponding model-simulated water levels. The RMSE is the square root taken after the sum of differences between measured and modeled water levels are squared and divided by the number of measurements. Therefore, general agreement between the measured and model-simulated water levels is tested by the RMSE. The lower the RMSE, the better the overall match.

The model was calibrated in a series of tests and adjustments to variables. Variables that held the least confidence because of a lack of data (vertical hydraulic conductivity of the semiconfining layers, hydraulic conductivity of streambed, and effective recharge) were tested first and were adjusted on a regional scale.

In a series of tests, vertical conductivity of the semiconfining layers was adjusted concurrently through a range of values while all other variables were held constant. Effective recharge was assumed to be 2.6 in/yr, and, for a 1 ft thick streambed, hydraulic conductivity was assumed to be 1 ft/d. RMSE's were recorded for the various values of vertical conductivity tested. Next, recharge rates were adjusted through a range of values, and RMSE's were recorded. Several values of vertical conductivity of the semiconfining beds were then tested with the new recharge rates. RMSE's were plotted and were contoured on a grid whose axes were recharge rates and vertical conductivity. Thus, ranges of values that produced the lowest RMSE's for both variables were determined. The ranges were from 1×10^{-2} to 1×10^{-1} ft/d for the vertical conductivity of the semiconfining beds and from 2.5 to 3.5 in/yr for the recharge rate. Hydraulic conductivities of the streambeds in the range from 0.5 to 5.0 ft/d were subsequently tested in unison in the model while all other variables were held constant. There was no significant improvement in RMSE's or streamflow seepages in the range.

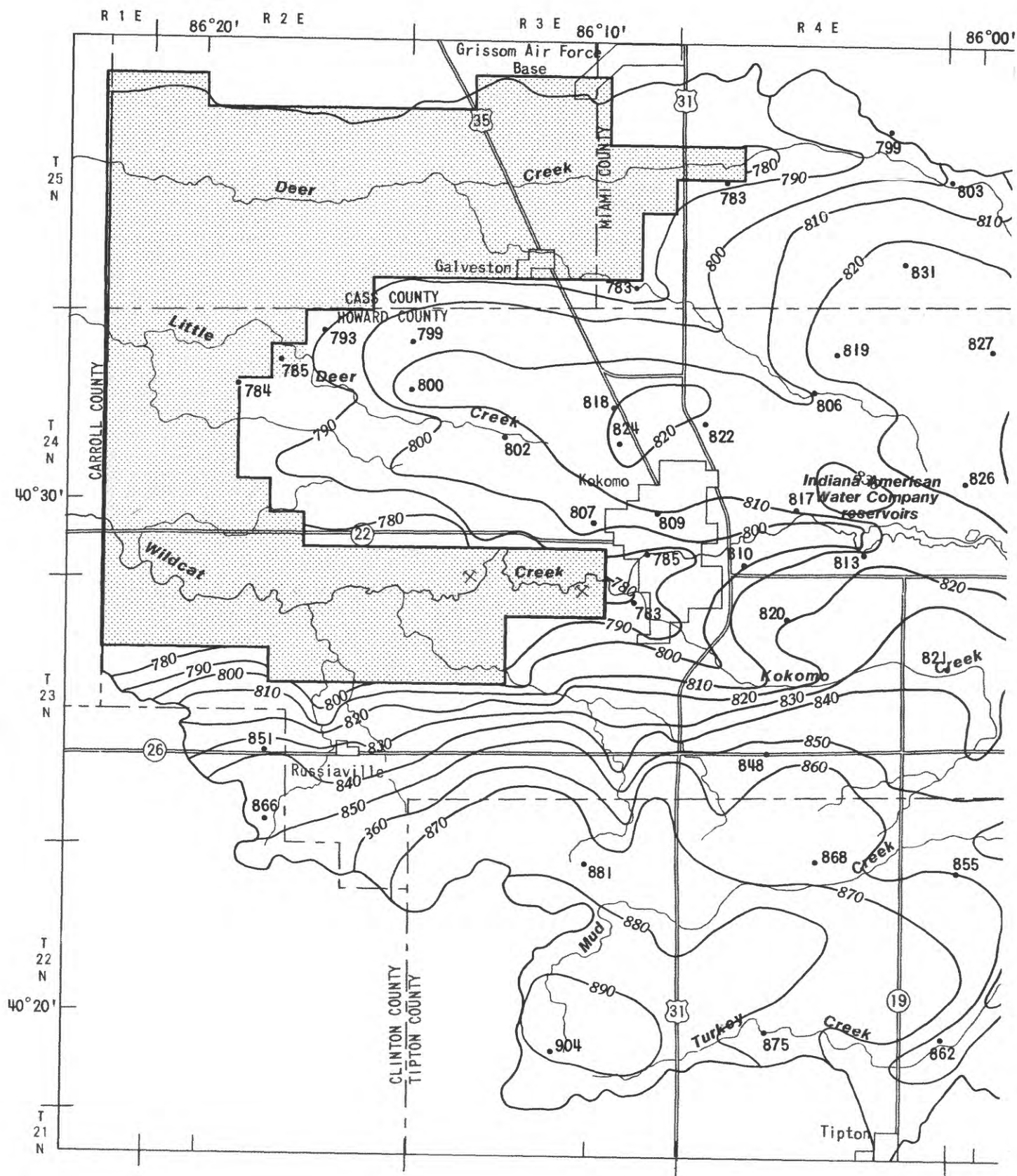
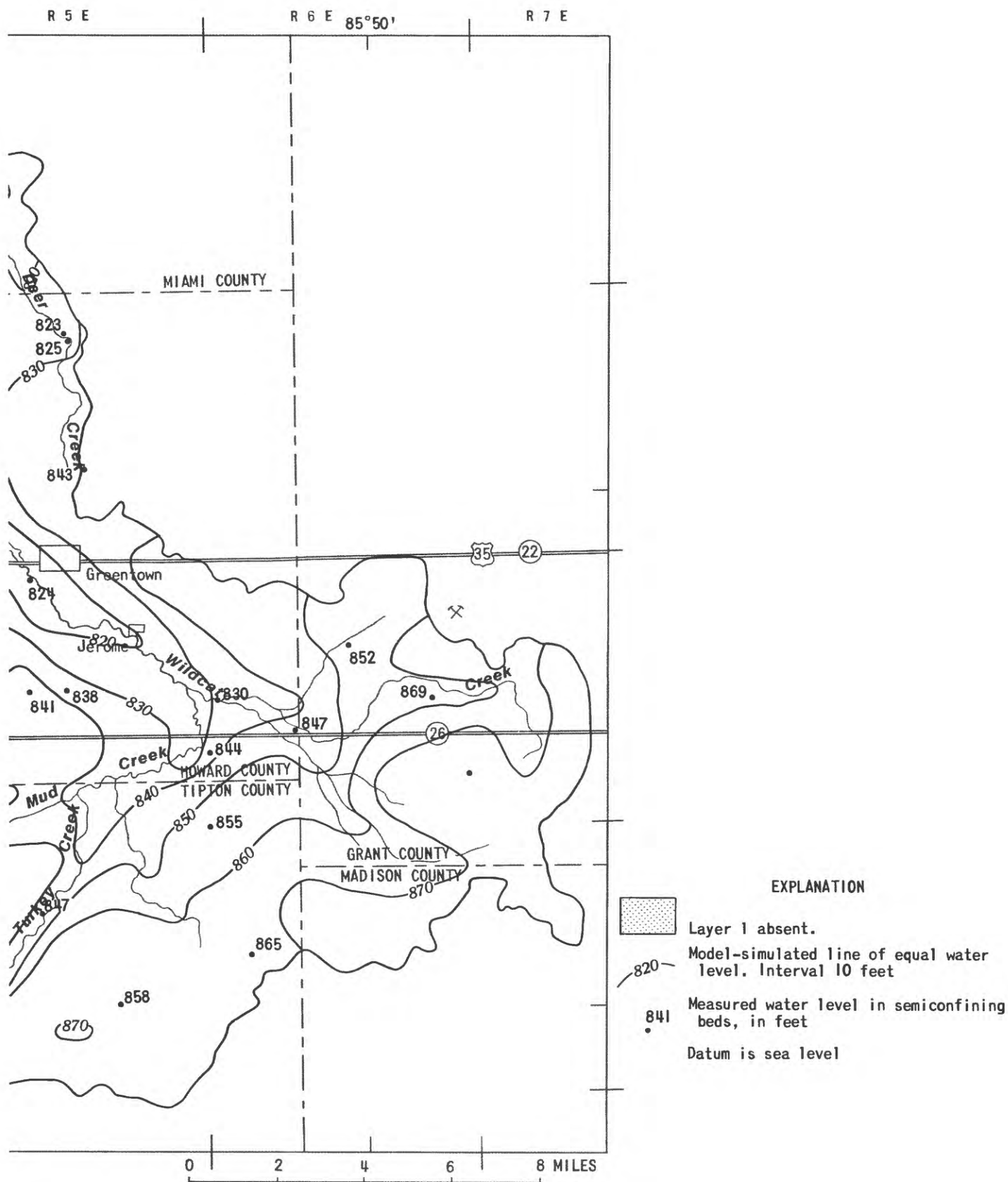


Figure 20.-- Measured and model-simulated water levels in layer 1.



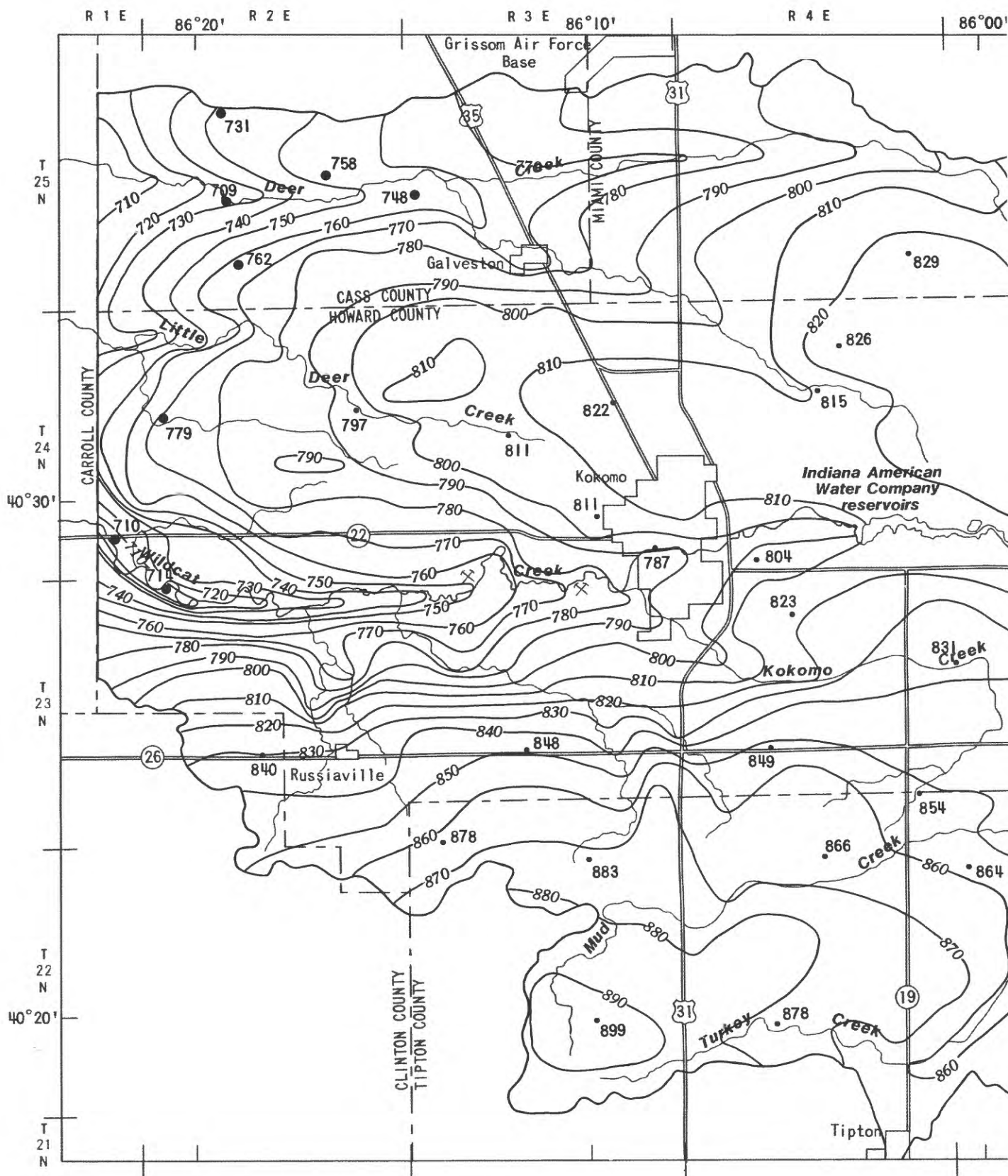
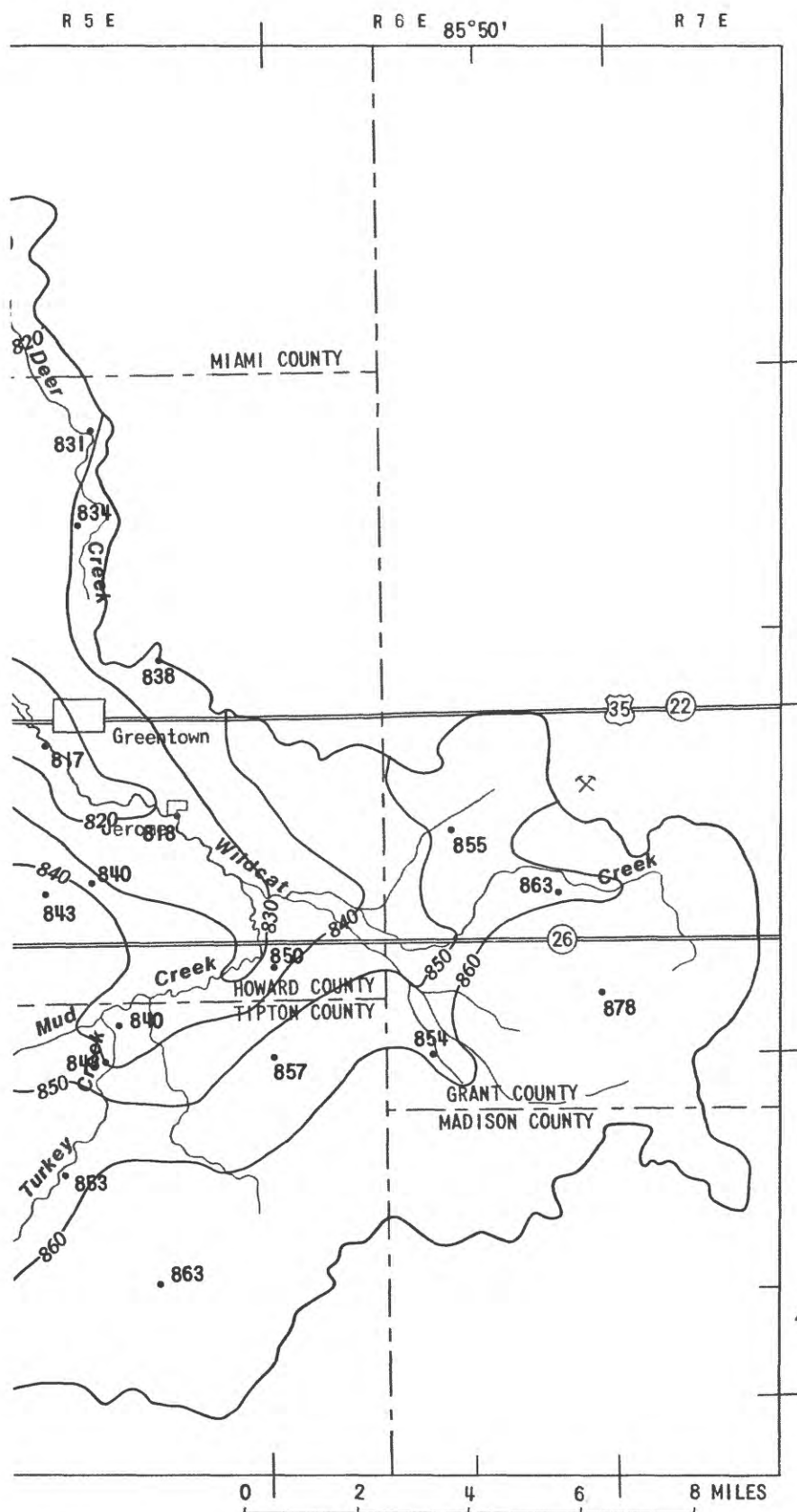


Figure 21.-- Measured and model-simulated water levels in layer 2.



EXPLANATION

- 830 — Model-simulated line of equal water level. Interval 10 feet
- 779 • Measured water level in semiconfining beds, in feet
- 818 • Measured water level in upper sand and gravel aquifer, in feet
- Datum is sea level

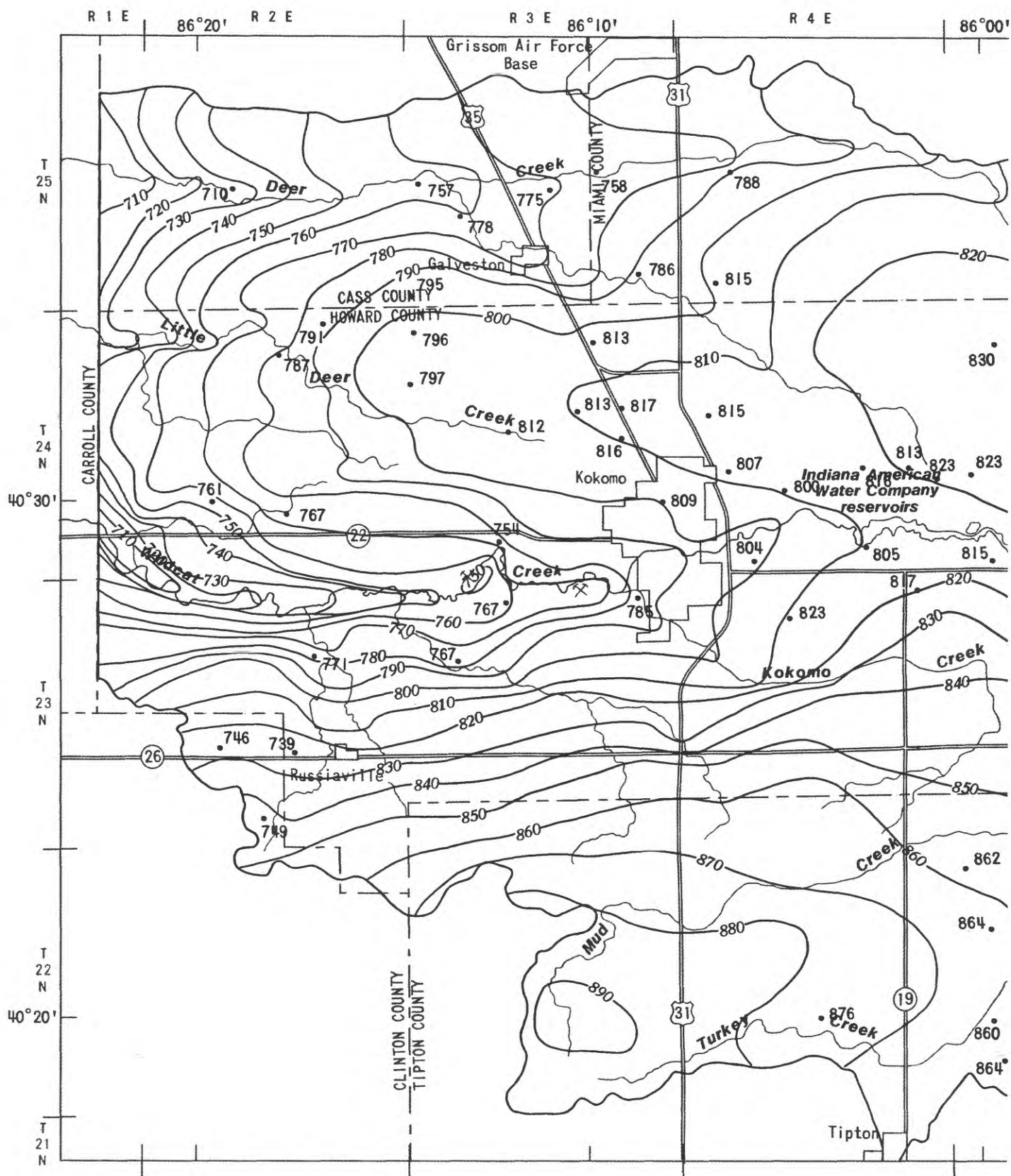
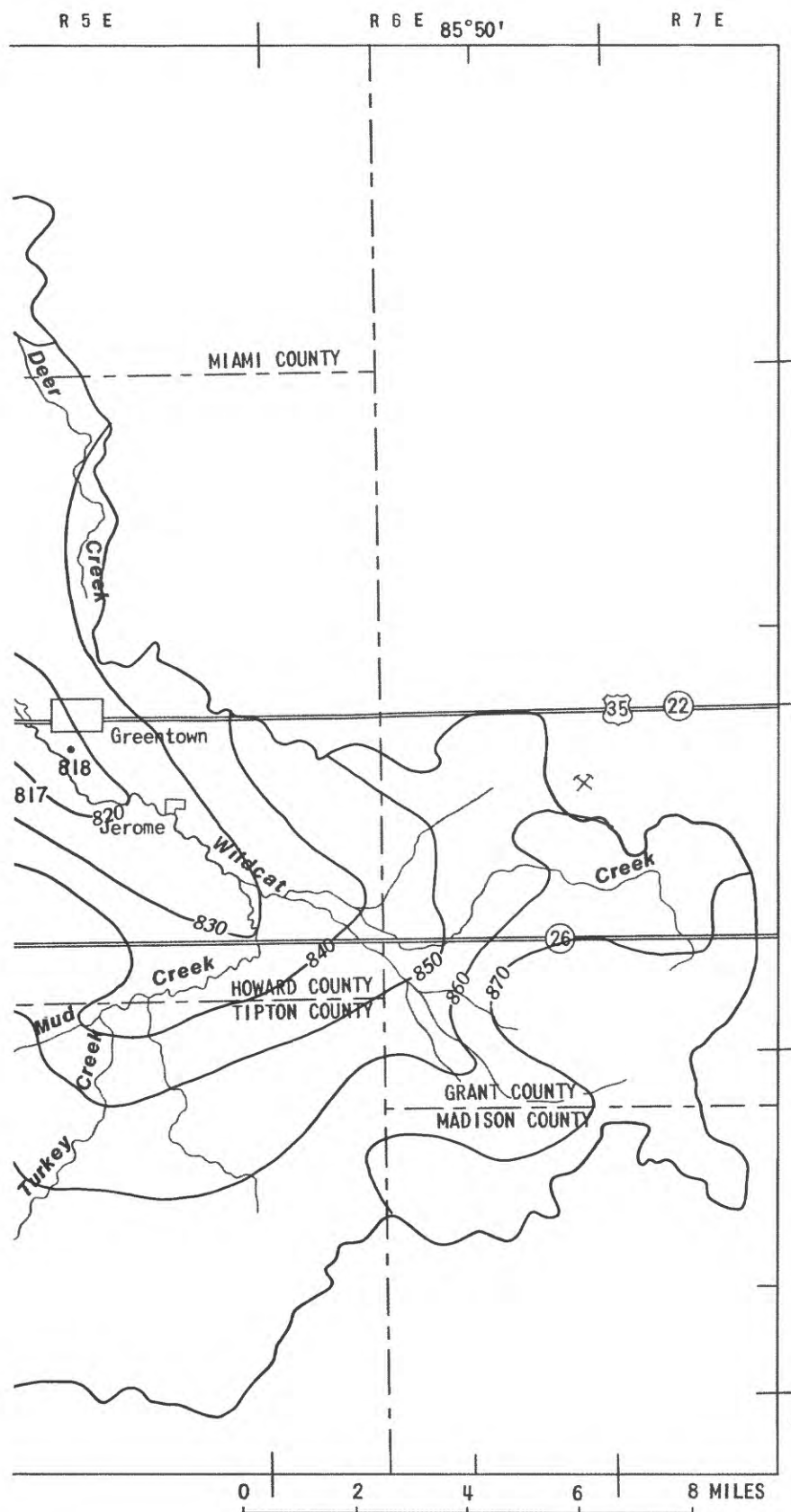




Figure 22.-- Measured and model-simulated water levels in layer 3.



EXPLANATION

-  830 Model-simulated line of equal water level. Interval 10 feet
-  876 Measured water level, in feet
- Datum is sea level

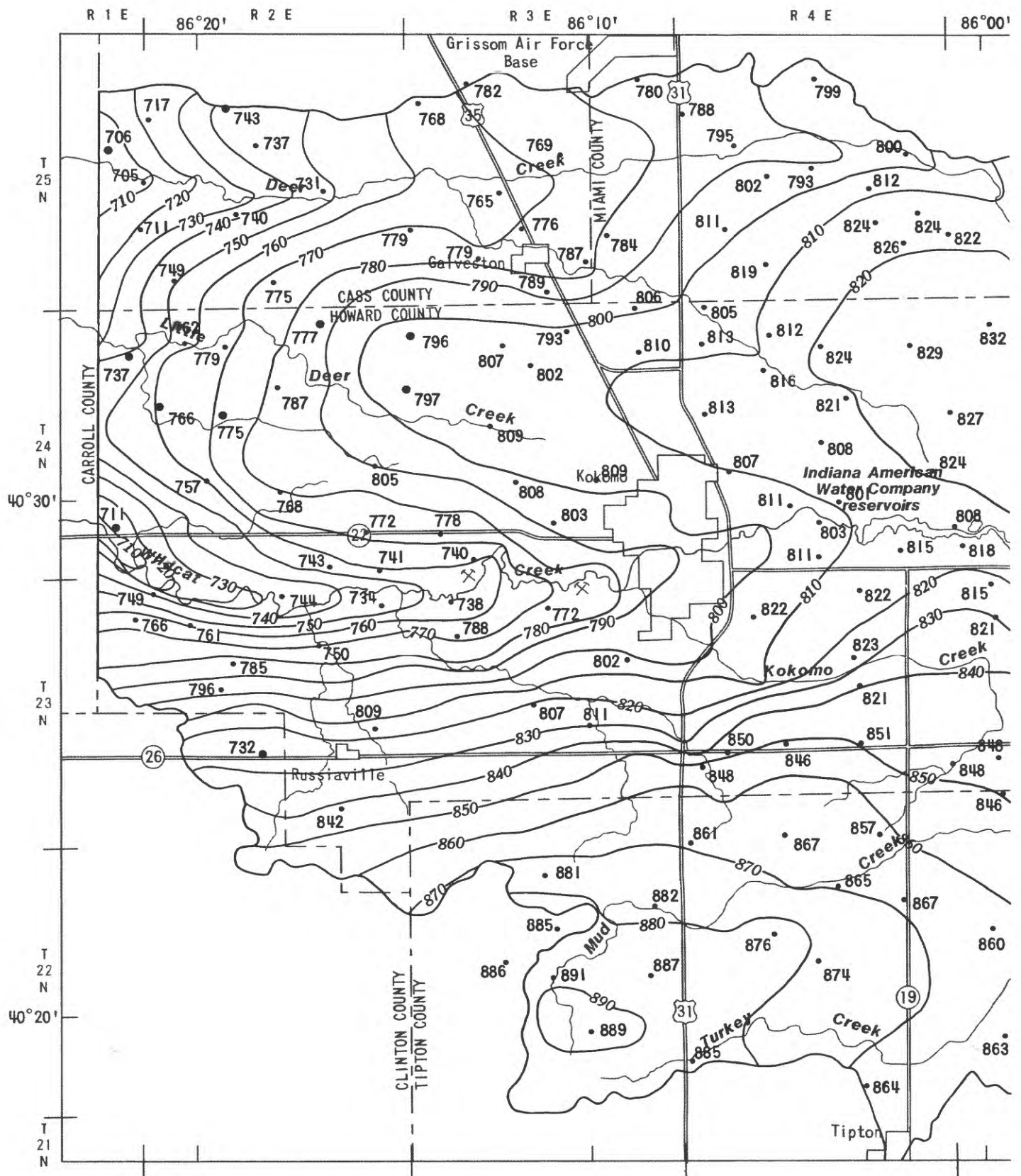


Figure 23.-- Measured and model-simulated water levels in layer 4.

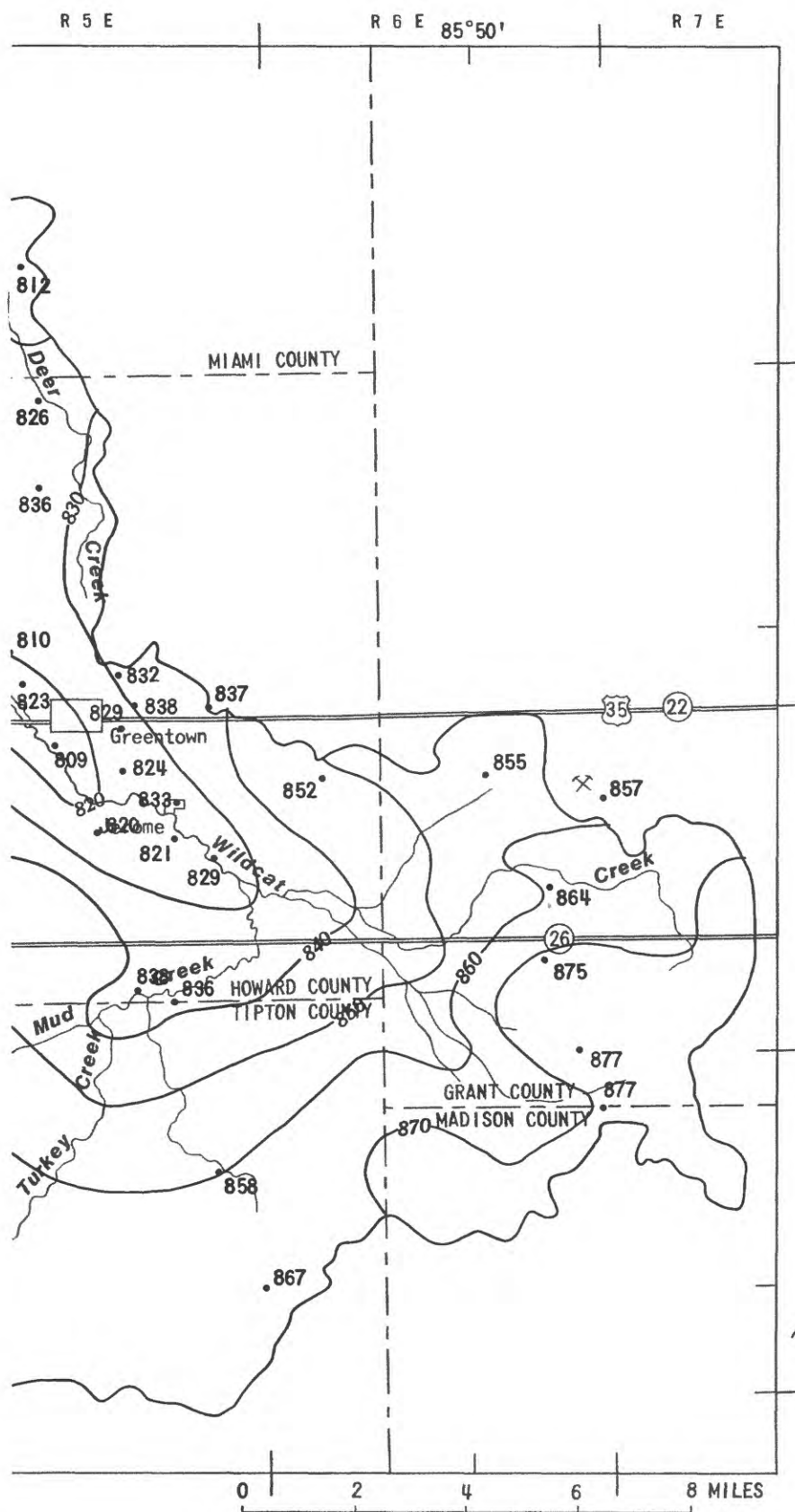


Table 8.--Model-simulated seepages to streams and ranges of measured seepages

Reach ¹	Measured seepage July 11, 12, and 13, 1981		Measured seepage August 20 and 21, 1981		Model-simulated seepage
	Minimum ²	Maximum	Minimum	Maximum	
1	- 0.1	3.3	0.0	4.2	3.6
2	3.1	5.9	.2	4.0	5.2
3	3.9	4.7	7.1	8.4	3.8
4	1.8	2.0	2.6	2.8	1.8
5	2.4	3.3	3.2	4.4	3.1
6	2.6	2.8	3.8	4.2	2.4
7	1.4	2.4	1.1	2.0	1.7
8	1.8	2.0	1.9	2.1	1.7
9	1.9	2.1	1.7	1.9	2.1
10	.6	.7	.4	.5	.9
11	15.3	27.9	7.1	20.0	7.6
12	³ -11.5	1.6	0.1	10.4	3.2
13	- 8.9	4.0	-14.1	-5.4	.1
14	3.4	9.6	- 4.2	7.2	.6
15	- 4.4	3.2	2.1	13.1	1.7
16	14.2	15.7	7.8	16.6	6.7
17	--	--	5.1	11.7	4.8
18	- 2.6	1.8	- 3.4	.0	1.7
19	1.5	2.0	1.0	1.5	1.6
20	1.5	1.6	1.3	1.5	1.4
21	1.5	1.7	1.1	1.2	1.4
22	1.2	1.7	1.0	1.3	1.5
23	.8	1.2	.9	1.5	1.1
24	.6	.7	1.0	1.0	.6
25	.9	1.0	1.1	1.3	.7
26	4.2	5.2	2.8	3.6	3.4
27	1.6	1.9	1.8	2.0	1.5
28	.8	.9	.3	.4	.6
29	3.4	4.2	1.7	3.8	4.0
30	1.8	2.1	1.7	2.2	2.0
31	.7	.7	1.4	1.5	.6
32	1.2	1.3	1.8	2.0	1.1
33	4.4	5.1	3.0	3.7	4.6
34	2.1	2.3	1.0	1.1	2.1
35	4.1	4.6	2.6	2.9	4.0
Total	57.1	133.4	47.4	140.6	84.9
Mean	1.7	3.9	1.4	4.0	2.4

¹Reaches are shown in figure 4.

²Minimum and maximum limits of gains and losses were determined from confidence intervals ranging from +5 to +8 percent of the measured gains and losses (table 2).

³Negative sign indicates loss of streamflow.

Transmissivities of the bedrock were tested and were adjusted by using water-level data and calibration of the flow model from the initial estimates that were based on specific-capacity and aquifer-test data. Four transmissivities taken from the contours of estimates of transmissivities were originally used: 500, 1,000, 5,000, and 10,000 ft²/d. Two transmissivities--1,250 and 6,250 ft²/d--resulted from testing by use of the model (fig. 15). The range in transmissivity was narrowed by use of the model because use of the model tended to regionalize the original values that represented local flows.

The vertical hydraulic conductivity of the semiconfining layers was tested by changing values in each layer individually; however, the changes did not improve the calibration. Much of the calibration involved adjustment of the vertical conductivities of the semiconfining layers areally within individual layers to improve the match with measured water levels. Other tests involved varying the recharge rates for Deer Creek and Wildcat Creek basins, the transmissivities of the aquifers, the streambed conductivities along selected reaches, and the rates of dewatering at the quarries west of Kokomo. The final stages of the calibration involved local adjustments to recharge rates, hydraulic conductivities of streambeds, and hydraulic conductivity of the semiconfining layers to match the measured gains and losses of streamflow.

Values that resulted from the steady-state calibration are listed in the following table:

Transmissivity of sand and gravel aquifers	650 to 8,700 ft ² /d
Transmissivity of the bedrock aquifer	1,250 and 6,250 ft ² /d
Vertical conductivity of semiconfining layers	0.01 to 0.0005 ft/d
Effective recharge rates	1.0 to 5.5 in/yr
Hydraulic conductivity of streambeds (for unit thickness)	0.02 to 20.0 ft/d

The values and ranges generally agree with those published for similar areas of Indiana by Meyer and others (1975), Gillies (1981), Lapham (1981), Arihood and Lapham (1982), and Arihood (1982).

Sensitivity of the calibrated model to changes in selected variables (fig. 24) was tested with RMSE's of the differences between measured and model-simulated water levels. One variable was changed through a range of values while all other variables were held constant. RMSE's resulting from the changes were recorded. Points corresponding to the RMSE's and to the multiples of the changes in the calibrated values were then plotted on semilog paper, and a curve defining the sensitivity of the calibrated model to the changes in variables was drawn. Constant-flux boundaries were used during the simulations. The effects of constant-head boundaries on model sensitivity was not tested because previous analyses had shown that constant-head boundaries tend to buffer the effect of changes in variables by changing flow across a

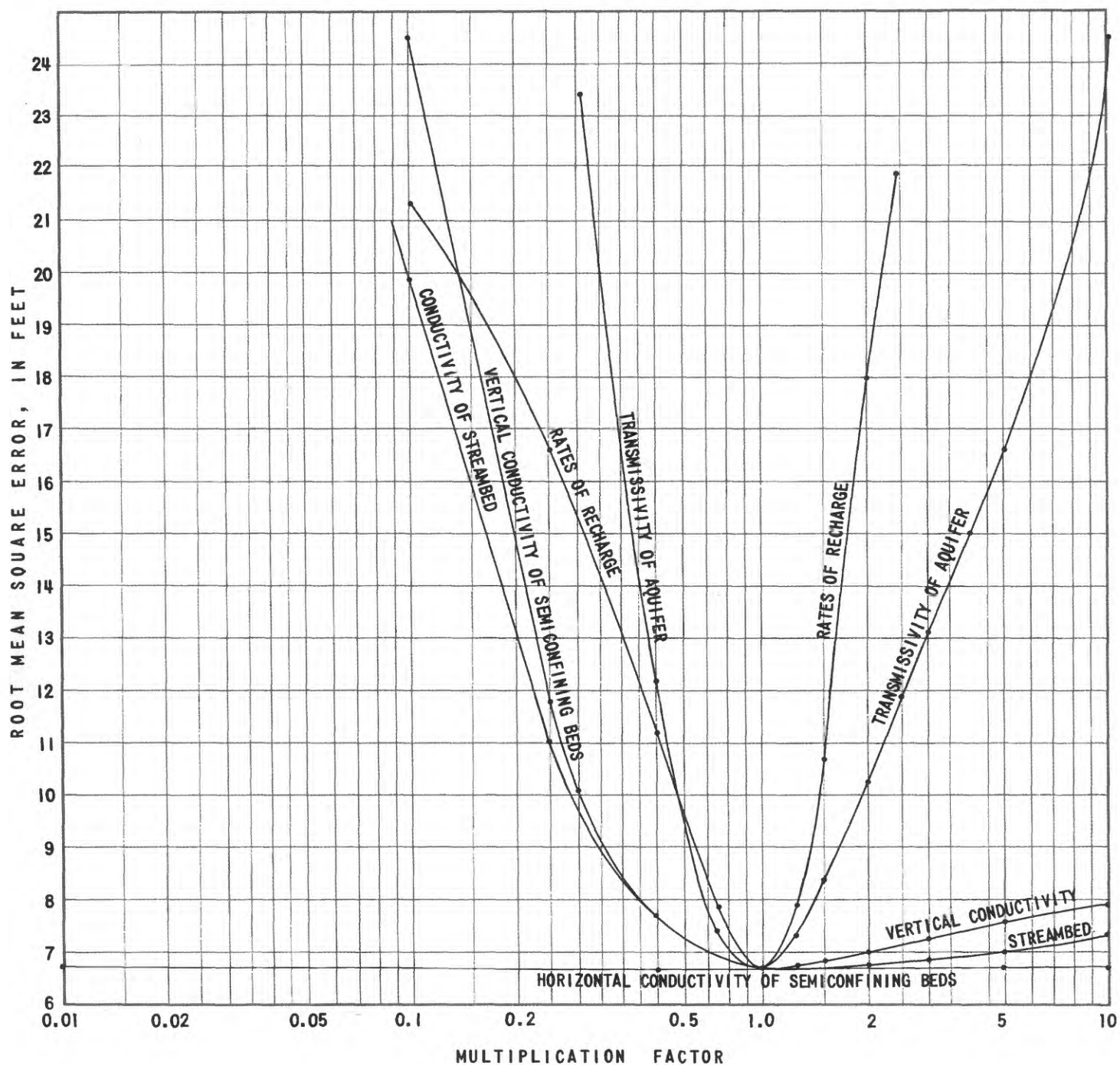


Figure 24.-- Sensitivity of model to selected variables.

boundary. Thus, constant-flux boundaries are the more sensitive of the two kinds of boundaries (Bailey and Imbrigiotta, 1982, p. 45, and Smith, 1983, p. 44).

Sensitivity curves were drawn for rate of recharge, transmissivity of aquifer, vertical conductivity of semiconfining layers, horizontal conductivity of semiconfining layers, and conductivity of streambed (for a unit thickness of streambed). The calibrated model is less tolerant (more sensitive) to changes in rates of recharge and transmissivity of the aquifer than to changes in hydraulic conductivity of the streambed or vertical conductivity of the semiconfining layers. The model is sensitive to decreases in conductivity of the streambed and vertical conductivity of the semiconfining layers but is not sensitive to increases in the two variables. The model is insensitive to any changes in horizontal conductivity of the semiconfining layers because there is virtually no horizontal flow through the semiconfining layers.

Steady-State Water Budget

The volumetric budget of the calibrated, ground-water flow model is representative of the steady-state water budget during a seasonal, base-flow recession that should recur approximately every other year under normal climatic conditions.

The largest part of flow into the system is recharge. Recharge is represented in the model as a flux to the topmost, active interior nodes of the model. Boundary flux represents recharge and flow across the boundaries in the model. Its calculation is based on Darcy's law and the water levels assigned as constant heads at the nodes along the boundary. Effective recharge to the entire ground-water system, therefore, is greater than the recharge applied to just the interior nodes (88 ft³/s) but is less than the recharge applied to the interior nodes and the recharge attributed to the nodes along the boundary (104 ft³/s). For the assumed recharge rate, 2.6 in/yr or 6.8×10^{-9} ft/s, recharge to one topmost node (4,000 ft by 4,000 ft) is about 0.1 ft³/s, and, for the 118 boundary nodes the total effective recharge is about 12 ft³/s. The effective recharge to the area of study, therefore, is 100 ft³/s, and the flow into the area modeled from the boundaries is just 4 ft³/s or 4 percent of all inflow. Thus, mean effective recharge rate over the entire model is 100 ft³/s divided by the area 505 mi² (1.4×10^{10} ft²) or 7.1×10^{-9} ft/s, which is 2.7 in/yr. The water budget for Wildcat Creek and Deer Creek basins is listed in the following table:

IN (cubic feet per second)		OUT (cubic feet per second)	
Effective recharge	100	Rate of pumping	11
Constant-flux boundaries	4	Constant-flux boundaries	8
Streams	<u>1</u>	Streams	<u>86</u>
Total	105	Total	105

The largest part of flow out of the ground-water flow system, about 86 ft³/s, is discharged to streams. Another 11 ft³/s is attributed to major ground-water pumping. The remaining 8 ft³/s, or 8 percent of flow out of the model, is boundary flux. Nearly all that flow leaves the modeled area from the bedrock. Flow out of the modeled area and across the west boundary, which does not coincide with boundaries of the basin, accounts for about 4 ft³/s. Thus, approximately 4 ft³/s or 4 percent of flow out of the model represents ground-water flow out of the basin.

Effects of Hypothetical Pumping

The calibrated model was used to simulate the effects of hypothetical pumping on water levels and streamflows. The hypothetical pumping would be in addition to the pumping in 1981, which is incorporated into the model. Declines in water level and streamflow caused by the hypothetical pumping are subtracted from the water levels and streamflows of the calibrated model, which represents a seasonal low-flow period of about 70-percent flow duration. Steady-state hypothetical pumping was simulated with constant-flux and then constant-head boundaries to test the effects of flow across boundaries on pumping locations. If there were no change in recharge or discharge just beyond the modeled area, the two boundaries would represent maximum (constant-head) and minimum (constant-flux) flow across boundaries. Locations of hypothetical pumping were selected so that effects of pumping on boundaries were minimal. The transient equation and constant-flux boundaries were also used in simulating hypothetical pumping so that the approximate time for the simulation to reach steady state could be determined. Two storage coefficients (1×10^{-5} and 1×10^{-3}) were used for the confined aquifer. However, there was practically no difference in the times for reaching equilibrium, with the two storage coefficients. Leakage through the semiconfining beds and distance to streams were the principal factors controlling time to reach equilibrium.

Pumping locations were selected in areas of high transmissivity where water levels were at least 20 ft above the top of the aquifer. Pumping rates were selected by trial and error until the water level in the 4,000 by 4,000 ft (0.57 mi²) pumping node was lowered about 20 ft. Thus, the rates of pumping represent the amount of water available with a drawdown assumed to be 20 ft at each pumping location. A drawdown of 20 ft was selected so that the results could be compared with those of several previous water-resources investigations in Indiana and because drawdowns greater than 20 ft would generally cause dewatering. The modeled drawdown is an average drawdown over the grid block. Actual drawdown would be much greater at and near the well.

The hypothetical simulations do not include every potential well-field site but rather are representatives of the various hydrogeologic conditions in the basins. Transmissivities at the pumping nodes ranged from 1,090 to 6,510 ft²/d, and rate of pumping representing available water (yield) ranged from 1.0 to 4.0 ft³/s, depending on the local hydrogeology (table 9). At the average pumping node, transmissivity was 4,340 ft²/d, yield was in the range from 2.0 to 2.5 ft³/s, and time for the drawdown in the pumping node to reach 95 percent of the drawdown simulated for a virtual steady state was about 4 years.

Table 9.--Yields of hypothetical well fields

Well field ¹	Aquifer	Yield (ft ³ /s)	Transmissivity (ft ² /d)	Drawdown available ² (ft)	Draw- down at steady state (ft)	Years to attain "virtual" steady state ³
A	Upper sand and gravel	2.5	5,640	51	19	4
B	Upper sand and gravel	2.5	6,510	50	20	4
C	Upper sand and gravel	1.5	4,770	64	18	8
D	Middle sand and gravel	1.5	3,910	38	21	1
E	Middle sand and gravel	2.0	4,340	71	18	5
F	Middle sand and gravel	3.0	4,340	61	20	4
G	Middle sand and gravel	4.0	4,340	65	20	2
H	Lower bedrock	1.0	1,250	106	16	2
I	Lower bedrock	2.0	1,250	148	21	7
J	Lower sand and gravel	1.5	6,500	166	19	10
K	Bedrock	3.5	6,250	130	20	4
L	Lower sand and gravel	1.5	1,090	75	23	<1
M	Bedrock	3.5	6,250	162	20	<1
Average		2.3	4,340	91	20	≈4

¹Locations are shown in figures 25, 26, and 27.

²Drawdowns available above bottom of aquifer before pumping. Altitude of bottom of bedrock aquifer is assumed to be 680 ft.

³Number of years for the drawdown in the pumping node to reach 95 percent of the drawdown simulated at virtual steady state.

Well fields cause water-level declines that coalesce and form an expanding cone of depression around the field (transient state). Expansion continues until the rate of water being pumped is equaled by an increase in the rate of recharge to the ground-water system, a decrease in the rate of discharge from the ground-water system or a combination of the two. If the effective recharge is not significantly increased by the decline in water levels and a decrease in evapotranspiration, the rate of pumping must be equaled by diverting ground water that would have discharged to streams or by inducing surface water from the streams. When the rate of pumping is equaled by flow to the well field, the cone ceases to expand, and an equilibrium or steady state is reached.

Water-level declines at steady state caused by hypothetical pumping from the upper sand and gravel aquifer well-fields A, B, and C are shown in figure 25. Water-level declines in layers above and below the upper sand gravel aquifer are similar but not as large as those in the layer where hypothetical pumping was simulated. The configurations of the potentiometric surface shown in figure 25 are strongly affected by the proximity of the well fields to nearby streams. The spacing of the 1- and 5-foot contours expand away from the streams around all three well fields. The extent of the 5-foot contour around well-field A, pumping $2.5 \text{ ft}^3/\text{s}$, however, is smaller than that around well field C, pumping just $1.5 \text{ ft}^3/\text{s}$, because well-field A is closer to Wildcat Creek and to the reservoirs on Wildcat Creek, where regional ground-water flow converges. For the hypothetical pumping at well-fields A, B, and C, the volume of streamflow reduction from the reaches (fig. 4) would be between the volumes of streamflow reductions calculated for constant-head and constant-flux boundaries (table 10).

Reductions in streamflow are greatest in the reaches closest to the well field. The greatest reductions, from 1.0 to $1.2 \text{ ft}^3/\text{s}$ in reaches 16 and 29, are less than 25 percent of the seasonal low flow. Depending on conditions at boundaries, however, approximately 33 percent of the seasonal low flow could be reduced in the headwaters (reaches 4, 6, 32, and 34). The difference in flow induced at constant-head and constant-flux boundaries is insignificant in most of the reaches, except for reach 4 where as much as $0.4 \text{ ft}^3/\text{s}$ might be contributed across the boundaries.

Water-level declines at steady state caused by hypothetical well-fields D, E, F, and G that withdrew water from the middle sand and gravel aquifer are shown in figure 26. The configuration of the cones of depression are strongly affected by proximity to Wildcat Creek and to the reservoirs on the creek. Transmissivities and available drawdowns of well-fields E, F, and G are similar, but the yields are different (table 9). Well-field G near the flanks of the reservoir has the largest yield ($4.0 \text{ ft}^3/\text{s}$) and the smallest cone of depression. Well-field F, farther from the flanks of the reservoir than well-field G but near Wildcat Creek, yields $3.0 \text{ ft}^3/\text{s}$ and has a larger cone of depression than well-field G. Well-field E, farthest of all the well fields from Wildcat Creek and the reservoirs, yields just $2.0 \text{ ft}^3/\text{s}$ and has the largest cone of depression.

Table 10.--Reduction in streamflow caused by hypothetical well-fields A, B, and C

Reach	Constant-flux boundaries		Constant-head boundaries	
	Reduction (ft ³ /s)	Percent reduction	Reduction (ft ³ /s)	Percent reduction
4	0.6	33	0.2	11
6	.6	33	.5	21
15	.1	6	.1	6
16	1.2	20	1.0	15
17	.1	2	.1	2
18	.2	13	.2	12
19	.2	14	.2	12
28	.1	17	.1	17
29	1.0	25	1.0	25
30	.5	25	.5	25
31	.1	20	.1	17
32	.4	36	.2	18
33	.5	12	.3	16
34	.5	31	.4	19
35	.4	11	.3	8
Total	6.5		5.2	

The configuration of the cones of depression encircling hypothetical well-fields E, F, and G are affected by proximity to each other and to well fields simulated in the model. The line representing the 5-foot decline in water level encircles the three hypothetical well fields and the actual well fields immediately east of Kokomo (fig. 4). Thus, mutual interference is a limiting factor in the availability of ground water, and the combined withdrawal from well-fields E, F, and G (9.0 ft³/s) represents the hypothetical regional availability of ground water.

Reductions in streamflow caused by hypothetical well-fields D, E, F, and G are listed in table 11. Half of the reductions in streamflow (5.1 to 5.2 ft³/s) was surface water induced from the reservoirs (1.5 ft³/s) or ground water that would have flowed to reach 16 (3.6 to 3.7 ft³/s). Large-percentage reductions in seasonal low flows are indicated for reaches 13, 14, and 15 (Wildcat Creek) because those reaches have been affected by real (1981) surface-water diversion and ground-water pumping simulated in the model.

The sum of hypothetical pumping with constant-flux boundaries, 10.5 ft³/s, was equaled by 10.4-ft³/s reductions in streamflow. The difference is probably due to a roundoff error of 0.1 ft³/s or less than 1 percent. Hypothetical pumping with constant-head boundaries caused reductions in streamflows of 9.7 ft³/s. A maximum of 0.8 ft³/s or 9 percent of the pumping, therefore, was equaled by flow induced from the boundaries. No individual reach, however, was significantly affected by the change in type of boundaries.

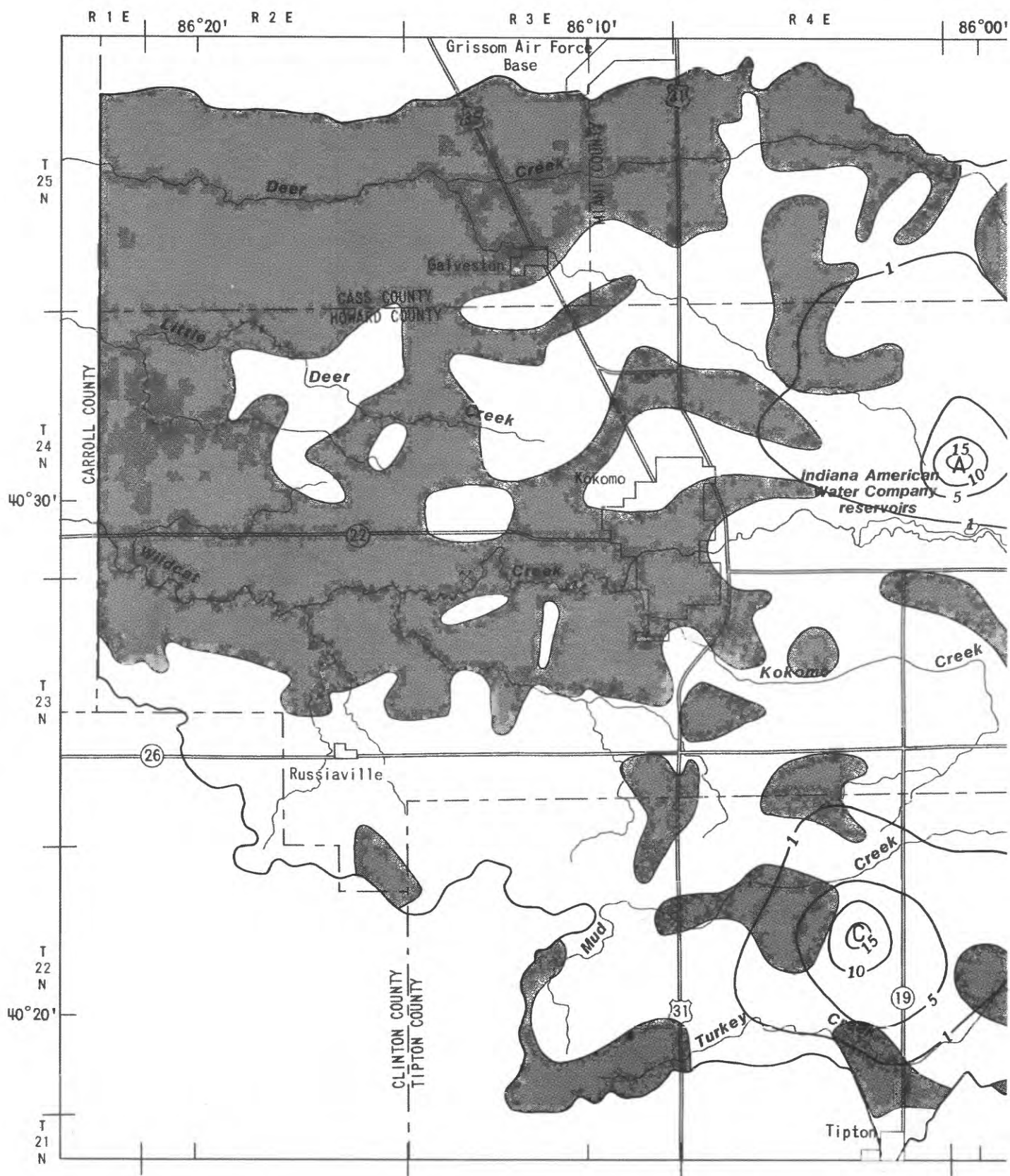
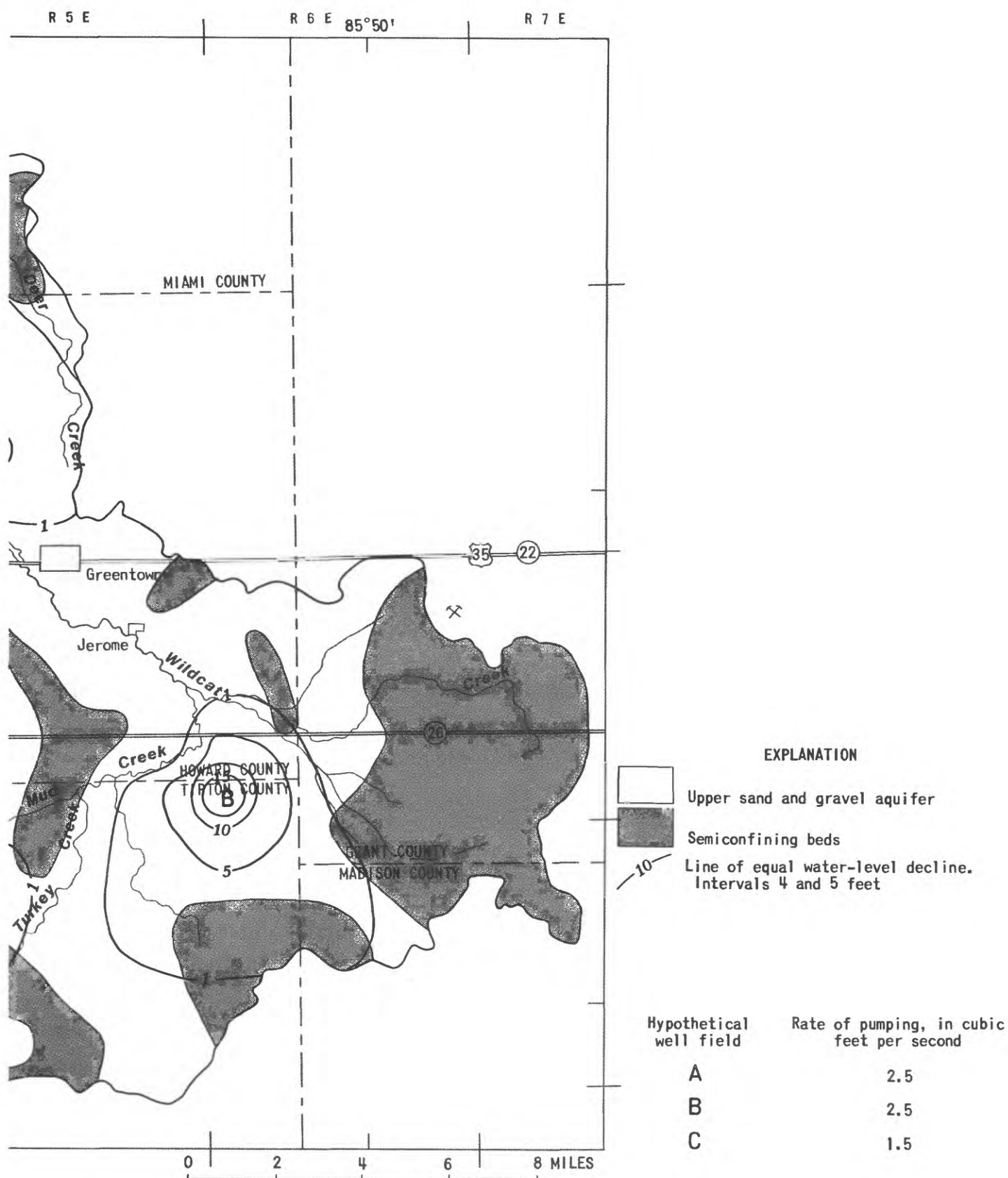


Figure 25.-- Water-level declines caused by hypothetical



withdrawal from the upper sand and gravel aquifer.

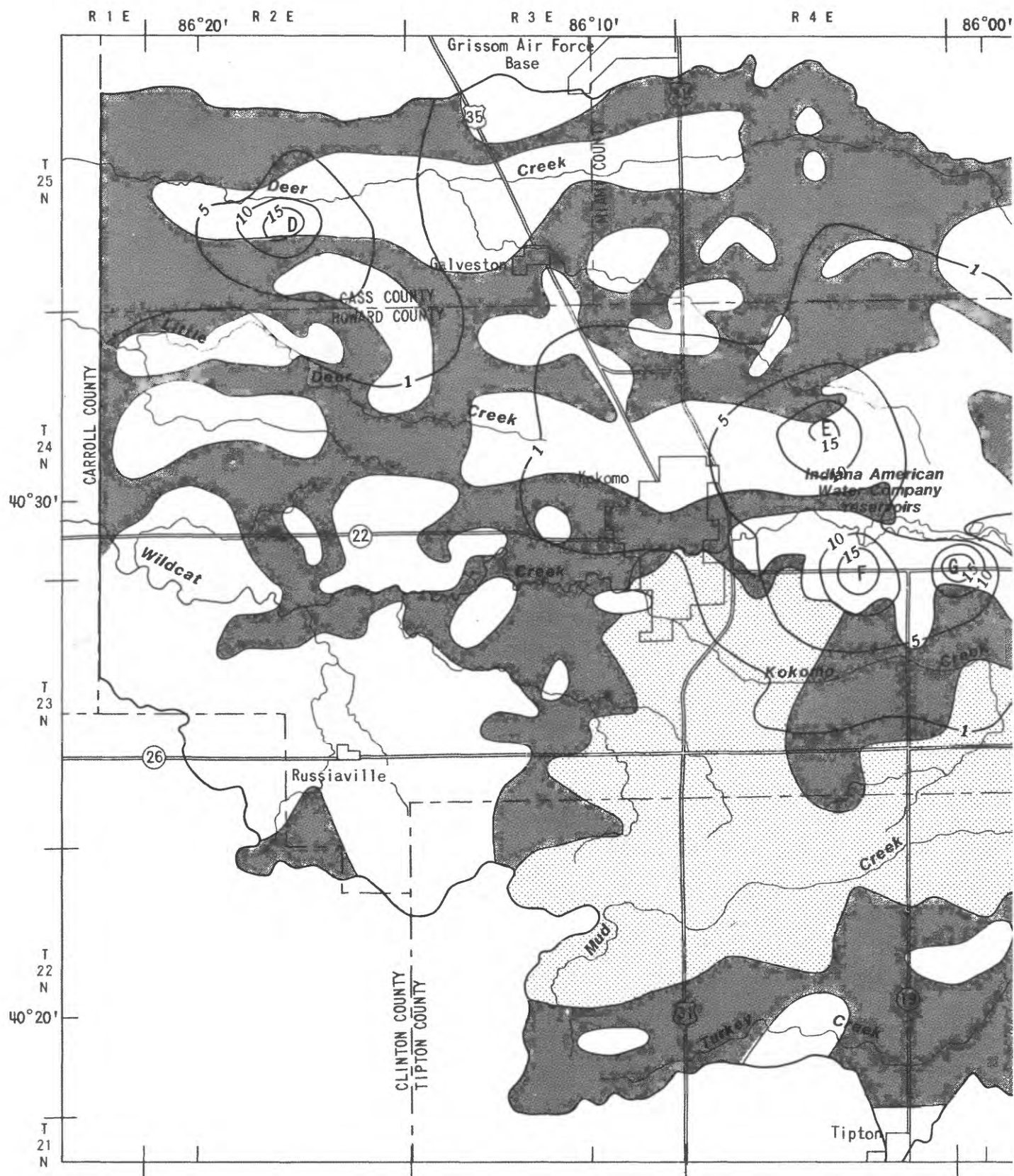
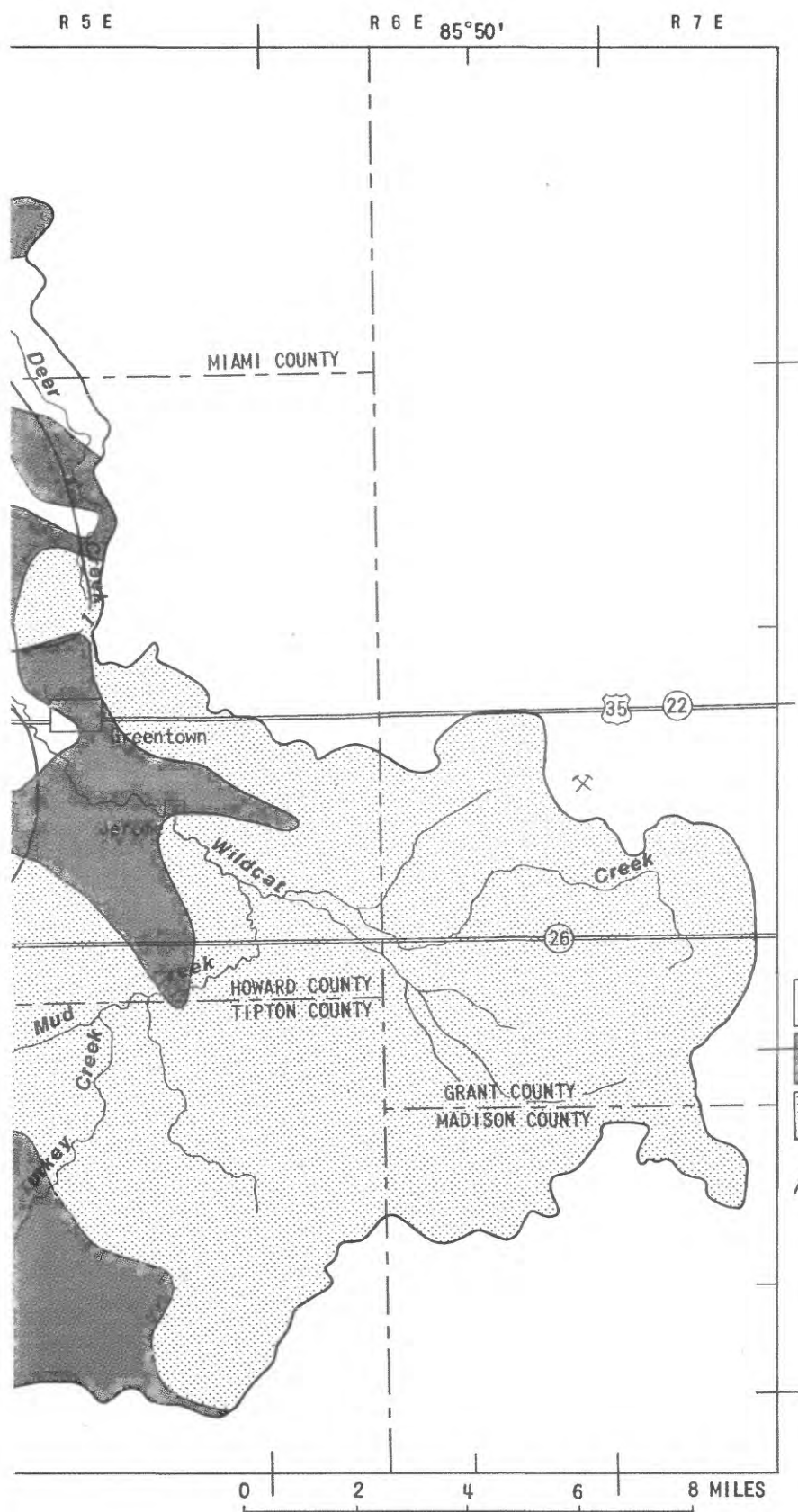






Figure 26.-- Water-level declines caused by hypothetical



EXPLANATION

-  Middle sand and gravel aquifer
-  Semiconfining beds composed of till, silt, or clay
-  Middle aquifer absent. Bedrock subcrop
-  Line of equal water-level decline. Interval 5 feet

Hypothetical well field	Rate of pumping, in cubic feet per second
D	1.5
E	2.0
F	3.0
G	4.0

withdrawal from the middle sand and gravel aquifer.

Table 11.--Reduction in streamflow caused by hypothetical well-fields D, E, F, and G

Reach	Constant-flux boundaries		Constant-head boundaries	
	Reduction (ft ³ /s)	Percent reduction	Reduction (ft ³ /s)	Percent reduction
1	0.9	25	0.8	22
2	.3	6	.2	4
3	.1	3	.1	3
4	.3	17	.1	6
5	.1	3	.1	3
6	1.0	42	1.0	42
7	.2	12	.1	6
8	.1	6	.1	6
12	.1	3	.1	3
13	.1	100	.1	100
14	.3	50	.3	50
15	.7	42	.7	41
16	5.2	78	5.1	76
17	.1	2	.02	.4
26	.4	12	.4	12
27	.3	20	.3	20
28	.2	33	.2	33
Total	10.4		9.72	

Water-level declines caused by withdrawals from the lower sand and gravel aquifer at hypothetical well-fields J and L and withdrawals from bedrock at well-fields H, I, K, and M are shown in figure 27. The transmissivity of the aquifer in well-field L is one-sixth (17 percent) that of the aquifer in well-field J, but the yield of well-field L is equal to that of well-field J because well-field L is bounded on two sides by small streams and on a third side by Wildcat Creek. Well-field J is near the headwaters of one small stream. The effect of transmissivity and proximity to streams is shown most prominently in the times required for well-fields L and J to attain virtual steady state (less than 1 and 10 yr). More than 10 times as much water was withdrawn from aquifer storage in well-field J as in well-field L, and the cone of depression in J was larger than the one in L. If pumping from storage is desirable, perhaps as a backup supply when streamflow is low, then hydrogeologic conditions similar to those in and near well-field J would yield water for short periods with minimal effect on streamflow. However, the locating of well fields near major streams would minimize water-level declines.

The characteristics of well-fields K and M each withdrawing $3.5 \text{ ft}^3/\text{s}$ from the bedrock are similar to each other (table 9). Well-field K, however, attained steady state in 4 yr compared with less than 1 yr for well-field M. The cone of depression in well-field K was larger than that in well-field M because well-field K is near actual pumping centers simulated in the model and near reaches of Wildcat Creek that are affected by surface-water diversions and the reservoirs. Well-field M, in contrast, is not near any actual pumping centers and is near the confluence of two streams.

Well-fields H and I produced less from the bedrock than well-fields K and M (table 9). The reason for the difference in production is that well-fields H and I are in areas where the transmissivity of the bedrock is lower than that in well-fields K and M and well-fields H and I are near the headwaters of Deer Creek basin, whereas well-fields K and M are near larger streams. Well-field H is much closer to a stream than well-field I and, therefore, attained steady state much sooner than well-field I.

Reductions in streamflow at steady state caused by pumping well-fields H, I, J, K, L, and M are listed in table 12.

For constant-head boundaries, the largest reductions in streamflow-- 1.3 and $1.6 \text{ ft}^3/\text{s}$ --were in reaches 16 and 29, those closest to the largest rates of withdrawal. The percentages of reduction in seasonal low flows were large in reaches 13, 14, and 15--the reaches affected by actual pumpage, diversion of surface water, and reservoirs. Percentages of reduction were also large for reaches 4, 6, and 8; the headwaters of Deer Creek basin; and for reaches 29 and 34 near well-field M.

The sum of hypothetical withdrawals with constant-flux boundaries ($13.0 \text{ ft}^3/\text{s}$) was equaled by a reduction in streamflow of $12.9 \text{ ft}^3/\text{s}$. Hypothetical pumping with constant-head rather than constant-flux boundaries resulted in reductions in streamflow of $11.5 \text{ ft}^3/\text{s}$. A maximum of $1.5 \text{ ft}^3/\text{s}$ or 12 percent of the rate of pumping, therefore, was equaled by flow induced from the boundaries. The only reach significantly affected by boundaries, however, was reach 4 where as much as $0.5 \text{ ft}^3/\text{s}$ could be induced.

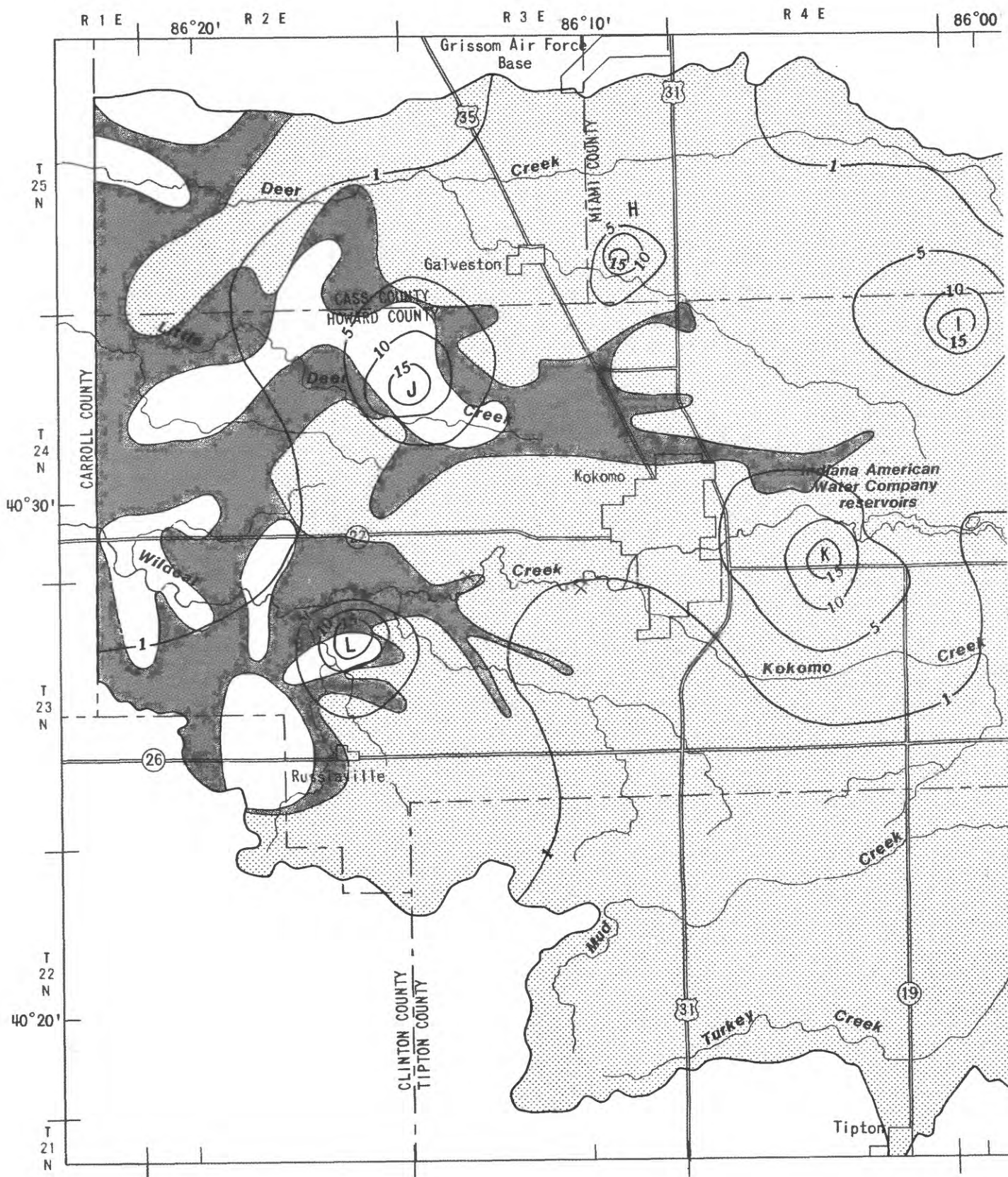
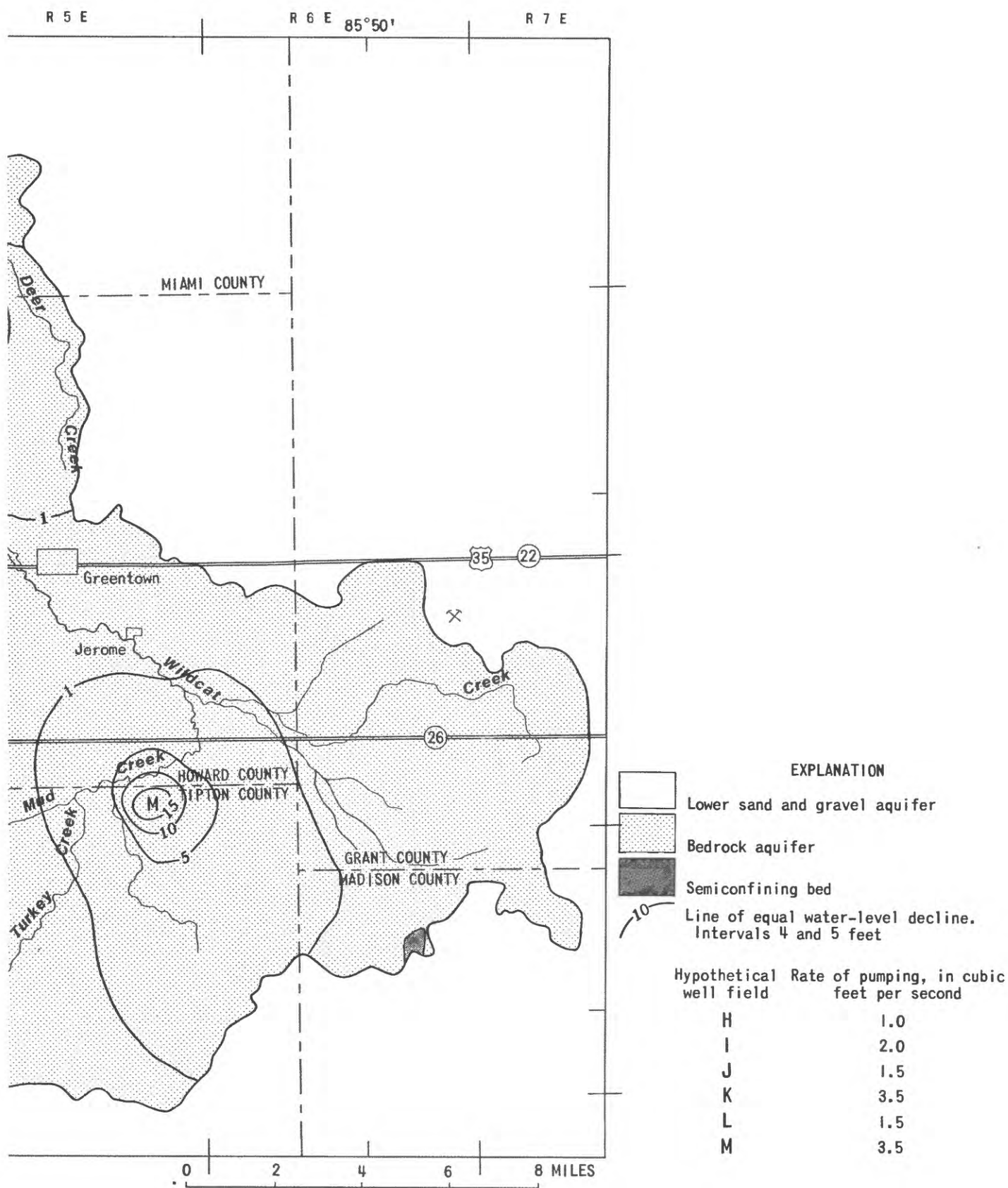


Figure 27.-- Water-level declines caused by hypothetical withdrawal



from the lower sand and gravel and the bedrock aquifers.

Table 12.--Reduction in streamflow caused by hypothetical well fields H, I, J, K, L, and M

Reach	Constant-flux boundaries		Constant-head boundaries	
	Maximum reduction (ft ³ /s)	Percent reduction	Minimum reduction (ft ³ /s)	Percent reduction
1	0.1	3	0.1	3
2	.4	8	.3	6
3	.4	10	.3	8
4	1.0	56	.5	28
5	.5	16	.5	16
6	.8	33	.7	29
7	.2	12	.2	12
8	.9	53	.9	52
9	.1	5	.1	5
11	.3	4	.2	3
12	.6	19	.6	19
13	.1	100	.1	100
14	.4	67	.4	67
15	.7	41	.7	41
16	1.4	21	1.3	19
17	.2	4	.2	4
18	.2	12	.2	12
19	.1	6	.1	6
21	.3	21	.2	14
22	.3	20	.3	20
23	.1	9	.1	9
25	.1	14	.04	6
26	.5	15	.5	15
27	.3	20	.3	20
28	.1	17	.1	17
29	.6	40	1.6	40
33	.3	7	.3	6
34	.7	33	.6	28
35	.2	5	.1	2
Total	12.9		11.54	

SUMMARY AND CONCLUSIONS

The water resources of Wildcat Creek and Deer Creek basins--505 mi² of Howard and parts of Cass, Clinton, Grant, Madison, Miami, and Tipton Counties--were evaluated. Streamflow at stream gages and gain-and-loss measurements were analyzed. Geometry and hydraulic characteristics of sand- and-gravel aquifers, bedrock, and semiconfining beds were defined; hydraulic connection between aquifers and streams was investigated; quality of ground water was analyzed; and a three-dimensional ground-water flow model was constructed and was used to simulate the effects of hypothetical pumping on water levels and streamflow.

Wildcat Creek and Deer Creek basins are within the Tipton till plain. The plain is flat to rolling and is modified by stream erosion, particularly along Wildcat and Deer Creeks, which have incised the till plain. Glacial deposits of the plain--predominantly silty till--range in thickness from zero at quarries along Wildcat Creek to 200 ft in buried valleys cut into the bedrock. Upper Silurian and Devonian limestone and dolomite and Lower and Middle Silurian limestone, dolomite, and dolomitic siltstone underlie the glacial deposits.

During high streamflow (less than 33-percent flow duration), discharge is proportional to size of basin, but at lower flows discharge is affected by local hydrogeology. Discharges greater than 0.3 (ft³/s)/mi² were measured during a seasonal low flow on the reaches of Wildcat Creek downstream from the mouth of Mud Creek, except for the reaches in and near Kokomo. The reaches in and near Kokomo are affected by diversion of surface water, pumping of ground water, discharge of treated sewage, and regulation of reservoirs.

Thirty-three percent of the 7.1 million gal/d (11 ft³/s) pumped from the aquifers within the basins was used by Kokomo. The majority of all commercial, industrial, and municipal wells are in and near Kokomo. More than 90 percent of the water pumped was from the bedrock.

Flow in the bedrock is predominantly through open fractures, joints, bedding planes, and solution channels in the Silurian and Devonian limestone and dolomite above the siltstone in the Silurian Mississippian Shale. Regional transmissivities of the bedrock, obtained by ground-water-flow modeling, were 1,250 and 6,250 ft²/d; however, locally the transmissivities may vary considerably from these values. Water levels in the bedrock generally fluctuate plus or minus 2 ft in response to seasonal or periodic changes in recharge, but the trend in water levels is toward steady state.

Ground-water flow is strongly influenced by the lithology and fabric of the glacial deposits of the till plain. The deposits of till are semiconfining beds, and discontinuous sand and gravel aquifers interspersed in the till are confined. Thickness of the sand and gravel aquifers generally ranges from 5 to 40 ft. Horizontal flow predominates in the aquifers because of the contrast in hydraulic conductivity of the sand and gravel aquifers (200 ft/d based on specific-capacity data and flow modeling) and of the semiconfining beds (from 5×10^{-4} to 1×10^{-2} ft/d based on flow modeling). Effective recharge to the aquifers (averaging 2.6 in/yr on the basis of analysis of streamflow) is by leakage through the semiconfining beds.

Wildcat Creek and Deer Creek are the major discharge drains of the regional ground-water flow system. Hydraulic conductivity of the streambeds, assumed to be of unit thickness, generally ranged from 2.0×10^{-2} to 20 ft/d on the basis of ground-water flow modeling.

Effective recharge to the upper Wildcat Creek and Deer Creek basins was about 100 ft³/s, and boundary flow into the basins based on simulations with the flow model was about 4 ft³/s. About 86 ft³/s was discharged to the streams, 11 ft³/s was pumped from aquifers, and 8 ft³/s was discharged from the aquifers to the areas outside the modeled area. Of the 8 ft³/s flowing out of the modeled area, about 4 ft³/s was out the west boundary of the model, which does not correspond with boundaries of the basin.

Ground-water samples were collected from 17 wells in the limestone aquifer and 13 wells in unconsolidated aquifers. Hardness of water ranged from 190 to 540 mg/L as calcium carbonate. Water types were calcium bicarbonate and calcium and magnesium bicarbonate. Concentrations of iron and manganese commonly exceeded 0.3 and 0.05 mg/L, the maximums recommended by the U.S. Environmental Protection Agency (1979, p. 2) for avoiding taste and staining problems. Mean concentrations of silica and iron were higher in the unconsolidated aquifers (sand and gravel) than in the limestone aquifer, but mean concentration of potassium was higher in the limestone than in the unconsolidated aquifers. Concentrations of ammonium commonly exceeded 0.5 mg/L as nitrogen; maximum concentration of ammonium was 22 mg/L as nitrogen. Possible sources of ammonium and other nutrients in ground water include livestock wastes and fertilizer applied to fields.

Average results of analyses of 10 water samples collected at Indiana State Board of Health station WC69 and WC63 upstream and downstream from Kokomo on Wildcat Creek from March 10 through December 13, 1982, included: Hardness of water, 242 and 307 mg/L; concentration of iron, 1.07 and 1.53 mg/L; concentration of manganese, 0.09 and 0.12 mg/L; and concentration of ammonia, 0.18 and 0.53. Average concentration of most constituents was higher at the downstream site than at the upstream site, probably because of urban runoff.

In addition to actual pumpage, hypothetical pumpages were simulated under steady state and transient conditions. Pumping rates were adjusted at selected nodes until water levels declined about 20 ft over the 0.57 square-mile area represented by the hypothetical well field. Yields from the hypothetical well fields ranged from 1.5 to 4.0 ft³/s (depending on local hydrogeology), but the typical well field yielded from 2.0 to 2.5 ft³/s and attained virtual steady state in about 4 yr. Mutual interference in hypothetical well-fields H, I, J, K, L, and M, in the lower sand and gravel and in the bedrock, whose combined pumping rate was 13.0 ft³/s, was minimal. Mutual interference in hypothetical well-fields E, F, and G, which were closer to each other than well-fields H, I, J, K, L, and M and whose combined pumping rate was 9.0 ft³/s, was greater than in well-fields H, I, J, K, L, and M.

Virtually all the water withdrawn hypothetically from the aquifers would have flowed to reaches of nearby streams. Reductions in individual reaches were generally 1 ft³/s or less, and all reductions except one were less than 1.6 ft³/s per reach. The exception was a reduction of 5.2 ft³/s or about 77 percent of the seasonal low flow of reach 16--the reach of the reservoirs on Wildcat Creek--which was caused by locating three hypothetical well fields

nearby. Reductions in streamflow for individual reaches were generally less than 50 percent of the seasonal low flow where one hypothetical well field was nearby. Exceptions were reaches 13, 14, and 15, in and near Kokomo, and affected by activities of man, and reaches 4 and 8, which are headwaters of Deer Creek basin. Compared to 7-day, 10-year low flows (table 1), however, reductions in streamflow of most reaches caused by the typical well field would be great. Exceptions are reaches of Wildcat Creek downstream from Mud Creek, but excluding reaches in and near Kokomo. The availability of water depends on the number and the proximity of well fields (current and proposed); the pumping rates; the degree of mutual interference and acceptable drawdown; on characteristics of the aquifer, semiconfining bed, and streambed; and the acceptable reduction in streamflows.

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