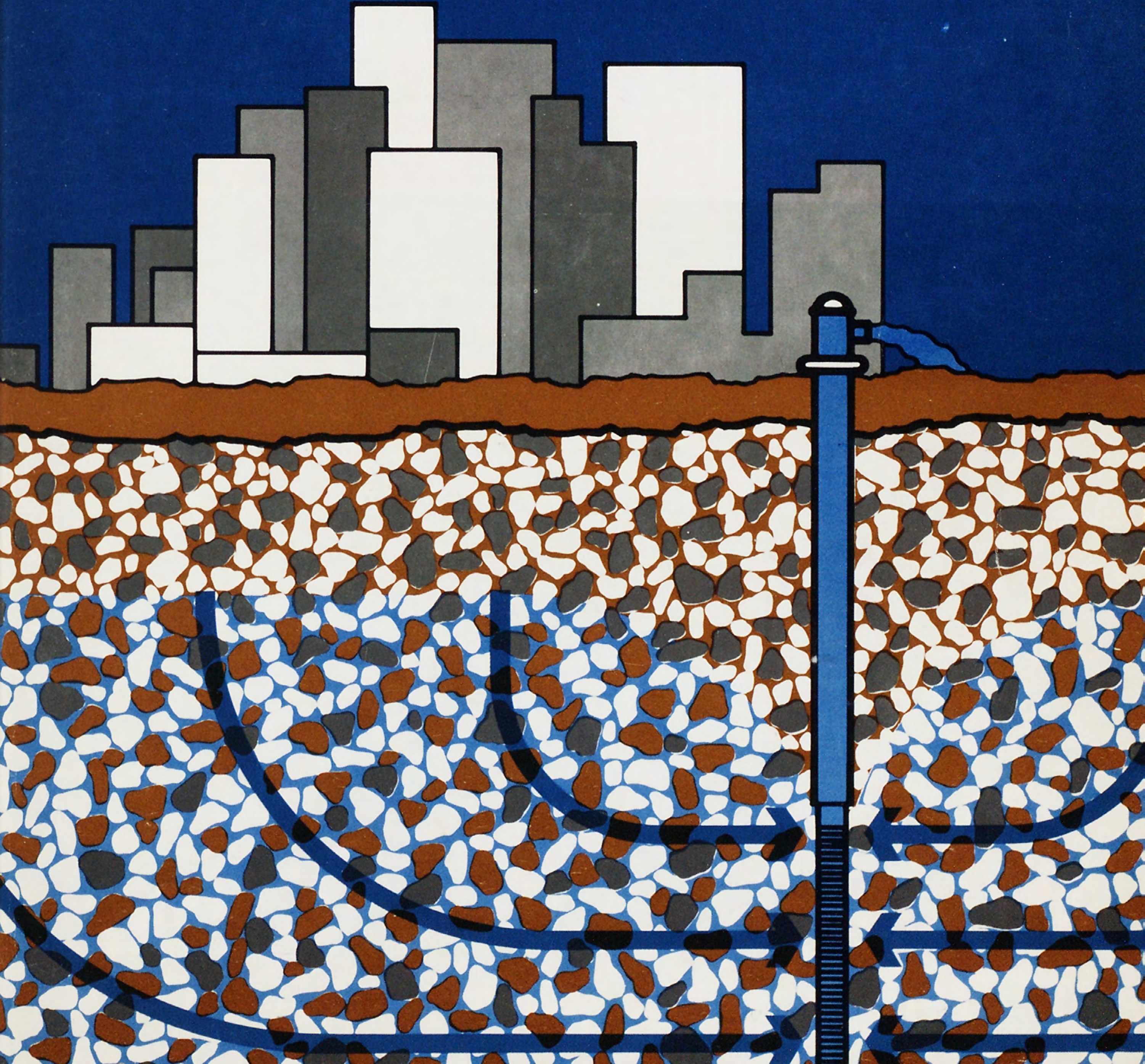
Availability of Ground Water in Marion County, Indiana

U.S. Geological Survey
Open-File Report 75-312



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UNITED STATES
DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY,

AVAILABILITY OF GROUND WATER IN MARION COUNTY, INDIANA

By William Meyer, J. P. Reussow, and D. C. Gillies

With a section on WATER QUALITY by William J. Shampine

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Prepared in cooperation with

Indiana Department of Natural Resources,

Division of Water

Indianapolis, Indiana

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CONTENTS

ILLUSTRATIONS

ILLUSTRATIONS--Continued

ILLUSTRATIONS--Continued

TABLES

METRIC CONVERSION FACTORS

The following factors may be used to convert the English units published herein to the International System of Units (SI)

Multiply English units	By	To obtain SI units
inches (in)	25.4	millimetres (mm)
feet (ft)	.3048	metres (m)
feet per day (ft/d)	.3048	metres (m/d)
feet per year (ft/yr)	.3048	metres per year (m/yr)
cubic feet per day per foot ($ft^3/d/ft$) or feet squared per day (ft^2/d)	.0929	cubic metres per day per metre ($m^3/d/m$) or metres squared per day
gallons per day per foot (gal/d/ft)	.0124	metres squared per day (m^2/d)
cubic feet per second (ft^3/s)	.02832	cubic metres per second (m^3/s)
miles (mi)	1.609	kilometres (km)
miles squared (mi^2)	2.590	kilometres squared (km^2)
gallons per minute (gal/min)	.06309	litres per second (l/s)
million gallons per day (Mgal/d)	.04381	cubic metres per second (m^3/s)

AVAILABILITY OF GROUND WATER IN MARION COUNTY

By William Meyer, J. P. Reussow, and D. C. Gillies

With a section on WATER QUALITY by William J. Shampine

ABSTRACT

The sand and gravel deposits in the glacial drift overlying Marion County constitute the most extensive aquifers in the county. Four areally distinct sand and gravel aquifers were mapped in the drift during the course of this study. The aquifer of greatest economic importance consists of sand and gravel deposits of glacial-outwash origin which coincide with the courses of the White River and Fall Creek. Ground water in this aquifer is generally unconfined. Three thin, areally discontinuous, sheetlike deposits of sand and gravel situated one above the other occur in the till-plain area of the county. These deposits coalesce in some areas, but generally are separated by beds of silt and clay of varying thickness which cause the ground water in the till-plain aquifers to be semiconfined.

The uppermost 100 feet (30 metres) or so of limestone and(or) dolomite underlying the glacial drift constitutes an aquifer of secondary importance in the county. This aquifer is separated from the unconsolidated aquifers either by silt and clay or by shale.

The present movement of ground water through these aquifers is generally toward the White River, Fall Creek, Eagle Creek or Buck Creek. Based on seepage data, water levels in the aquifers in April and May 1974, and subsequent electric-analog-model analysis, discharge of ground water by seepage into the streams in the glacial outwash averages about 42 cubic feet per day per foot (3.9 cubic metres per day per metre) of channel length.

A four-layer electric-analog model incorporating the geometry and hydraulic characteristics of the aquifers, the confining beds and the stream- aquifer connection was constructed to determine the feasibility and hydrologic results of proposed plans for future ground-water development.

Three large well fields were simulated in the model of the glacial-outwash aquifer, with pumping nodes situated so as to take advantage both of high transmissivity areas in the aquifer, and of recharge from the streams.

A series of model experiments were carried out to test the capacity of the aquifers to sustain increases in pumpage. In all of these, a new equilibrium was established within 6 years of simulated pumpage. In two of these experiments, water levels in the discharging wells were allowed to draw down to approximately half of the saturated thickness of the glacial-outwash aquifer. At this drawdown limit, the total discharge of wells in the system was found to be 59 million gallons per day (2.6 cubic

metres per second) using a stream-aquifer connection with conservative hydraulic features, and 81 million gallons per day (3.6 cubic metres per second) assuming that the streams were fully connected to the upper third of the glacial-outwash aquifer. In two other experiments, discharging wells were allowed to drawdown an average of two-thirds of the saturated thickness of the glacial-outwash aquifer. At this limit, the total discharge was found to be 72 million gallons per day (3.2 cubic metres per second) using the conservative stream-aquifer connection, and 103 million gallons per day (4.5 cubic metres per second) assuming a full connection. Some dewatering of the aquifer was associated with the 72 million gallons per day (3.2 cubic metres per second) discharge. In all experiments, the amount that could be pumped from the confined aquifers without disturbing existing domestic wells was found to be small.

INTRODUCTION

The Indianapolis metropolitan area has grown so that in 1975 it encompasses nearly all of Marion County and, to a small extent, parts of the neighboring counties. This growth has created a great demand on the water resources of the area, emphasizing the need for a better understanding of the ground-water system in order to permit proper management of this resource.

Several previous reports (McGuinness, 1943, Roberts and others, 1955, Cable and others, 1971, and Maclay and Heisel, 1972) have dealt with some of the broader aspects of the ground-water system in Marion County, but none of these reports defines the system quantitatively to the extent necessary to answer the questions now of concern. Therefore, at the request of the Indiana Department of Natural Resources, the U.S. Geological Survey entered into a cooperative agreement to conduct a 3-year study, beginning in July 1972, to define and quantify the ground-water system of Marion County. The objectives of this study were to determine the quantity of ground water that could be pumped in the county and to estimate the effects of this pumpage on the ground-water system and on streamflow.

To meet the objectives of the study it was necessary to establish: (1) the areal and vertical extent of the aquifers underlying the county; (2) the magnitude and areal variation of the transmissivity, and storage coefficient of these aquifers; (3) the vertical permeability and thickness of confining beds; (4) recharge, both natural and that which can be induced through ground-water pumpage; (5) discharge, including pumping from wells, seepage of ground water into streams, and direct evapotranspiration from the water table if this occurs; and (6) present water-level and potentiometric surfaces throughout the county.

When these parameters were established, an electric-analog model simulating the ground-water flow system of the county was constructed so that various withdrawals from the ground-water system could be simulated in time and space to estimate the feasibility of these withdrawals and the effects on ground-water levels and streamflow. This report presents the hydrologic data collected and the results of this investigation.

ACKNOWLEDGMENTS

The investigation of the availability of ground water in Marion County was conducted by the U.S. Geological Survey in cooperation with the Indiana Department of Natural Resources, Division of Water. The authors express their thanks to the more than 70 industries and the several municipalities that cooperated fully with efforts made during this study to reconstruct historical pumpage figures in the county. Special thanks are due to the American Aggregates Corp., Carl Verble & Assoc., and the Indianapolis Water Co. for permission to conduct aquifer tests on their property during the course of this study. Permission from Detroit Diesel Allison, Division of General Motors to install observation wells on their property deserves special mention. Finally the technical guidance provided during the course of the study by Messrs. Gordon D. Bennett and Eugene P. Patten of the U.S. Geological Survey is gratefully acknowledged.

REGIONAL SETTING

Marion County is in the central part of Indiana, as shown by figure 1, and encompasses approximately 400 mi^2 (1,040 km^2). The county is generally flat to gently rolling with elevations ranging from 645 to 915 ft (195 to 280 m) above mean sea level.

The entire county is in the White River drainage system, except for the southeastern edge where an area of about 45 mi^2 (117 km^2) is drained by Buck Creek, a tributary of the East Fork White River. The White River enters the county along its northern edge, just east of center, and flows southwesterly to leave the county at its southern border. Fall Creek, a major tributary, enters the county at its northeastern corner and flows southwesterly to the White River. Eagle Creek, the other major tributary in the county, enters at the northwestern corner of the county and flows southeasterly to join the White River several miles below Fall Creek.

Marion County has a temperate climate. Average annual temperature is about 53°F (12°C). The growing season is from April 16 to October 21. The county annually receives about 40 in (1,020 mm) of rainfall, distributed fairly evenly throughout the year. Average monthly rainfall for 1931-60 ranged from about 2.5 to 5 in (63.5 to 127 mm) (Cable and others, 1971; and Roberts and others, 1955).

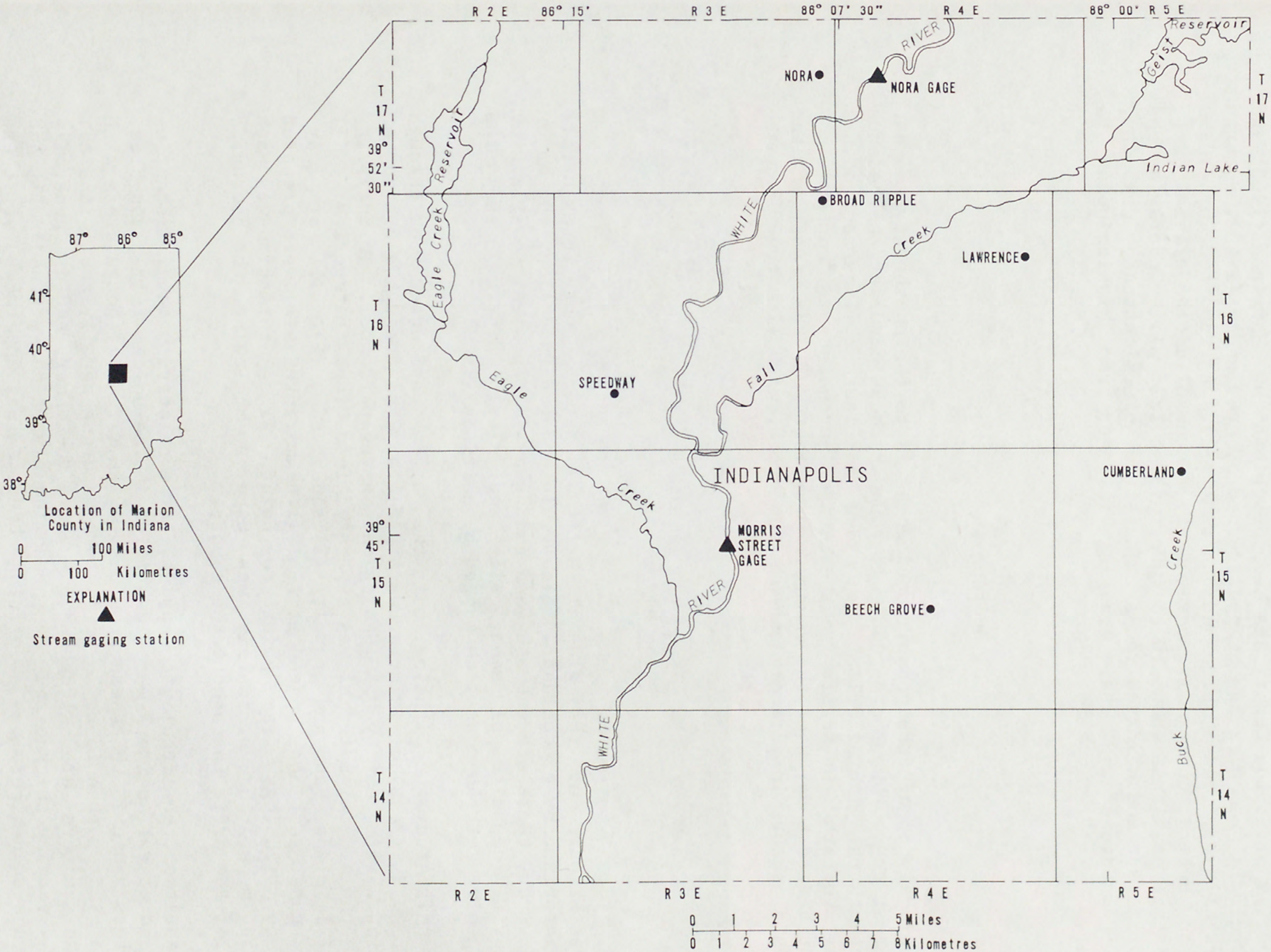

Figure 1.-- Marion County

GEOLOGY

The geology of Marion County has been described by Harrison (1963) and the geohydrologic framework has been discussed by McGuinness (1943), Roberts and others (1955), and Cable and others (1971). As reported by these authors, the entire county is covered by unconsolidated glacial drift ranging in thickness from about 10 to 400 ft (3 to 122 m). Underlying the glacial drift are consolidated rocks consisting of limestone, dolomite, siltstone, sandstone and shale ranging in age from Mississippian to Ordovician.

Bedrock

The regional dip of the consolidated rocks is to the southwest, so that from northeast to southwest progressively younger rocks occur at the bedrock surface beneath the glacial drift. As shown in figure 2, limestone occurs immediately beneath the drift in the entire northeastern, eastern and extreme southeastern parts of the county. Over most of the rest of the county shale occurs immediately beneath the drift, with limestone beneath shale.

Glacial Drift

The greater part of the glacial drift in Marion County forms a till plain characterized by predominantly fine-grained deposits. Along the major streams, however, there are outwash sand and gravel deposits, overlain by recent alluvial deposits of sand, silt and clay. Figure 3 shows the surficial geology of the county as mapped by Harrison (1963).

During this study the lithology of the drift was determined through drillers' records (logs) of existing water wells, and through a program of test drilling carried out by U.S. Geological Survey personnel. In analyzing these data the glacial material was separated into four categories: (1) gravel; (2) mixed sand and gravel; (3) sand; and (4) silt or clay. East-west cross sections, spaced at 1-mi (1.6-km) intervals, were prepared to correlate lithology from point to point. The information was then assembled on maps, and ultimately was used to calculate the hydraulic properties of the drift throughout the county.

As a result of the lithologic mapping, three relatively thin, discontinuous sand and gravel aquifers were identified in the upland till-plain area. These aquifers range in thickness from 0 to 40 ft (13 m) although areas with thicknesses greater than 15 to 20 ft (5 to 6 m) are generally small in areal extent. The aquifers are situated one above the other, and are generally separated by varying thicknesses of predominantly finer-grained silt and clay which act as semipervious confining beds. The uppermost of these

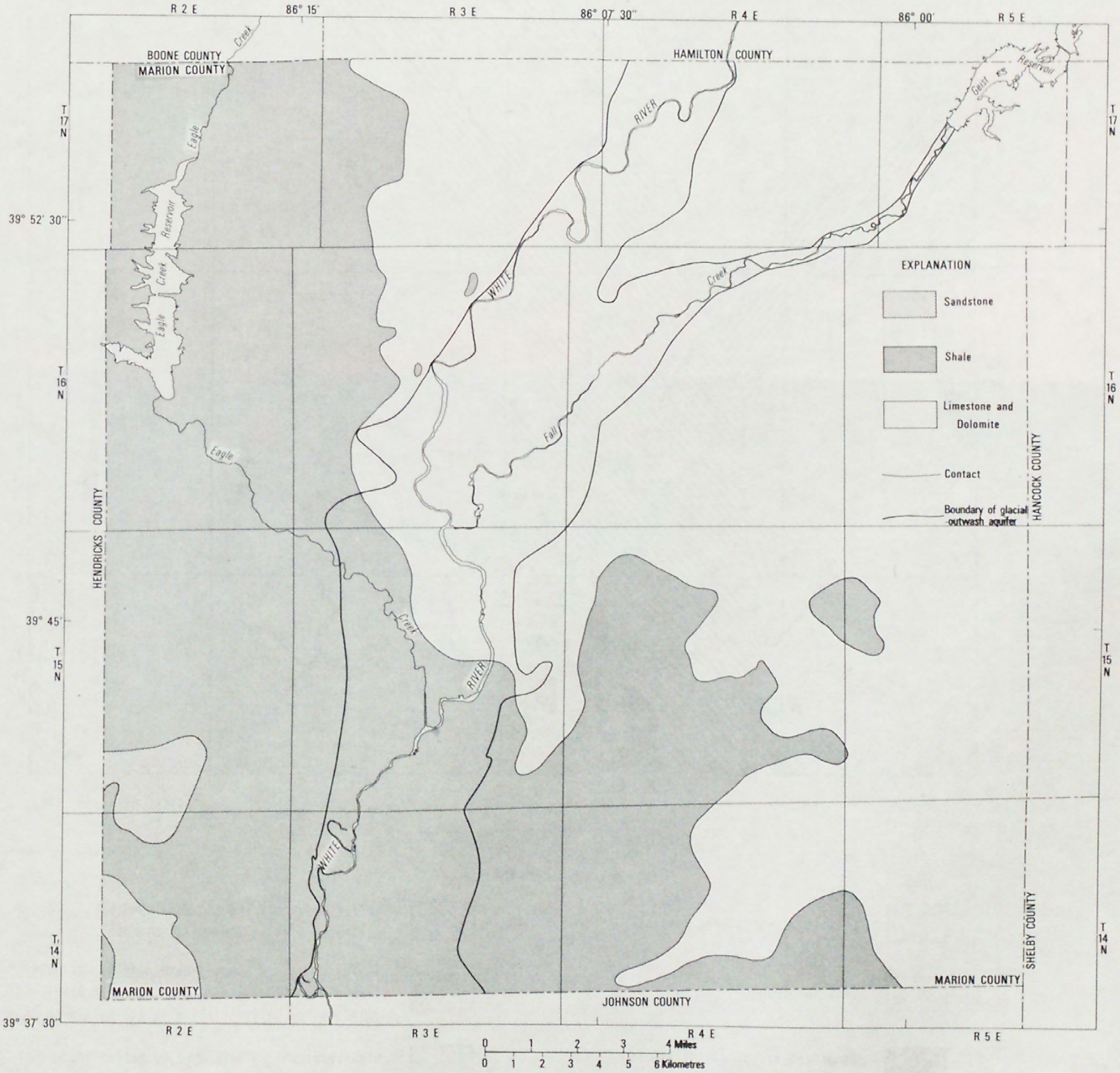

Figure 2.-- Bedrock geology of Marion County

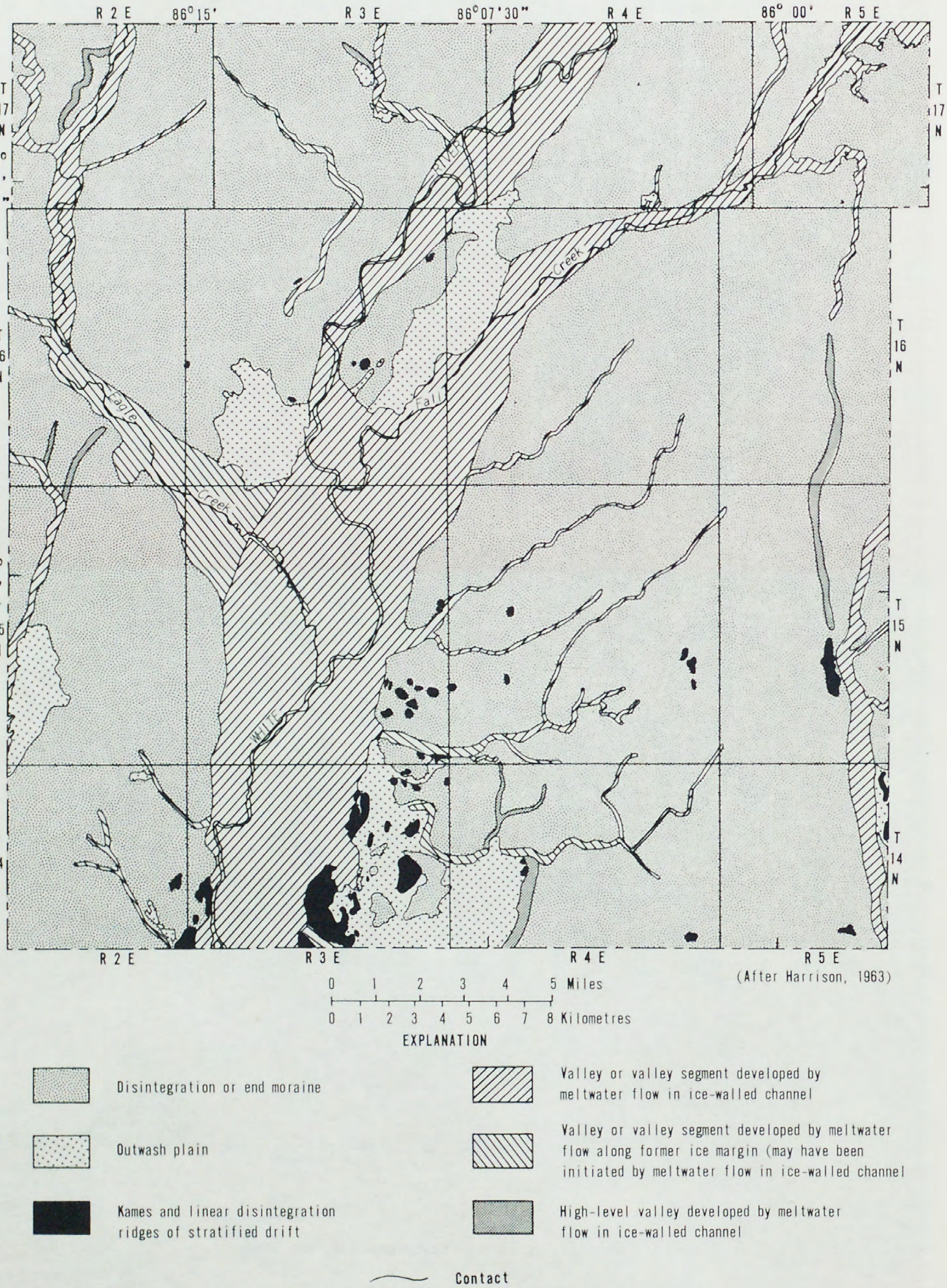

Figure 3.— Surficial geology of Marion County

aquifers is overlain by a varying thickness of this silt and clay; the potentiometric surface in the till-plain area occurs within this uppermost silt and clay. In some areas, two of the sand and gravel aquifers coalesce into one thicker sand and gravel body. In this report, the three sand and gravel aquifers are referred to as the upper, middle, and lower confined aquifers. Figures 4 through 6 show the areal distribution, approximate elevation of the top, and points of known thickness of the three confined aquifers.

As can be seen in figures 4 through 6, the upper, middle and lower confined aquifers in the till do not cover the county as areally continuous deposits. Large areas of silt and clay often separate one area of an aquifer from another. In addition, the upper confined aquifer does not exist in the till plain area on the west side of the glacial-outwash aquifer, nor in that part of the county between White River and Fall Creek. Figures 4 through 6 are based on the interpolation of field data, and because of the areally discontinuous nature of the confined aquifers, the mapping cannot be regarded as completely accurate. At any given point in the till plain it may be possible to find all three aquifers, none at all, or any combination of the three.

In addition to these three confined aquifers, a major unconfined sand and gravel aquifer was delineated along the White River and Fall Creek (fig. 3). The location of this aquifer generally coincides with the glacial melt water and outwash deposits mapped along these streams by Harrison (1963) as shown in figure 3, although some minor deviations exist. In this report, this aquifer system is referred to as the glacial-outwash aquifer. From an economic point of view it constitutes the most important aquifer in Marion County.

The glacial-outwash aquifer consists predominantly of sand and(or) gravel with discontinuous interbedded layers of finer-grained silt and clay. Well logs that do not show clay layers are rare, but the east-west cross sections indicate that the clay layers generally extend less than 1 mi (1.6 km) horizontally. An exception is the clay lens that occurs below and west of Eagle Creek in section 16 and in parts of sections 15, 17, 20, and 21 in T.15 N., R.3 E. While clay lenses reduce the horizontal transmissivity of the aquifer to some extent, their primary effect is to reduce the vertical hydraulic conductivity; locally, where this reduction is sufficient, semiconfined ground-water conditions may be found in the outwash.

The saturated thickness of the glacial-outwash aquifer was determined by installing 104 small diameter (1.5-in or 40-mm) observation wells throughout the aquifer, contouring the water levels for April 24, 1974, and subtracting the elevation of the top of the underlying bedrock at selected points. Figures 7 and 8 show the contoured water levels and the saturated thickness, respectively of the glacial-outwash aquifer.

In addition to the glacial-outwash aquifer and the three confined aquifers, small outwash deposits of sand and gravel are known to underlie some of the secondary streams in the county, and isolated pockets of sand and gravel are scattered throughout the till-plain area. These smaller deposits are of limited significance for water supply, and could not be mapped in the time available for the present study.

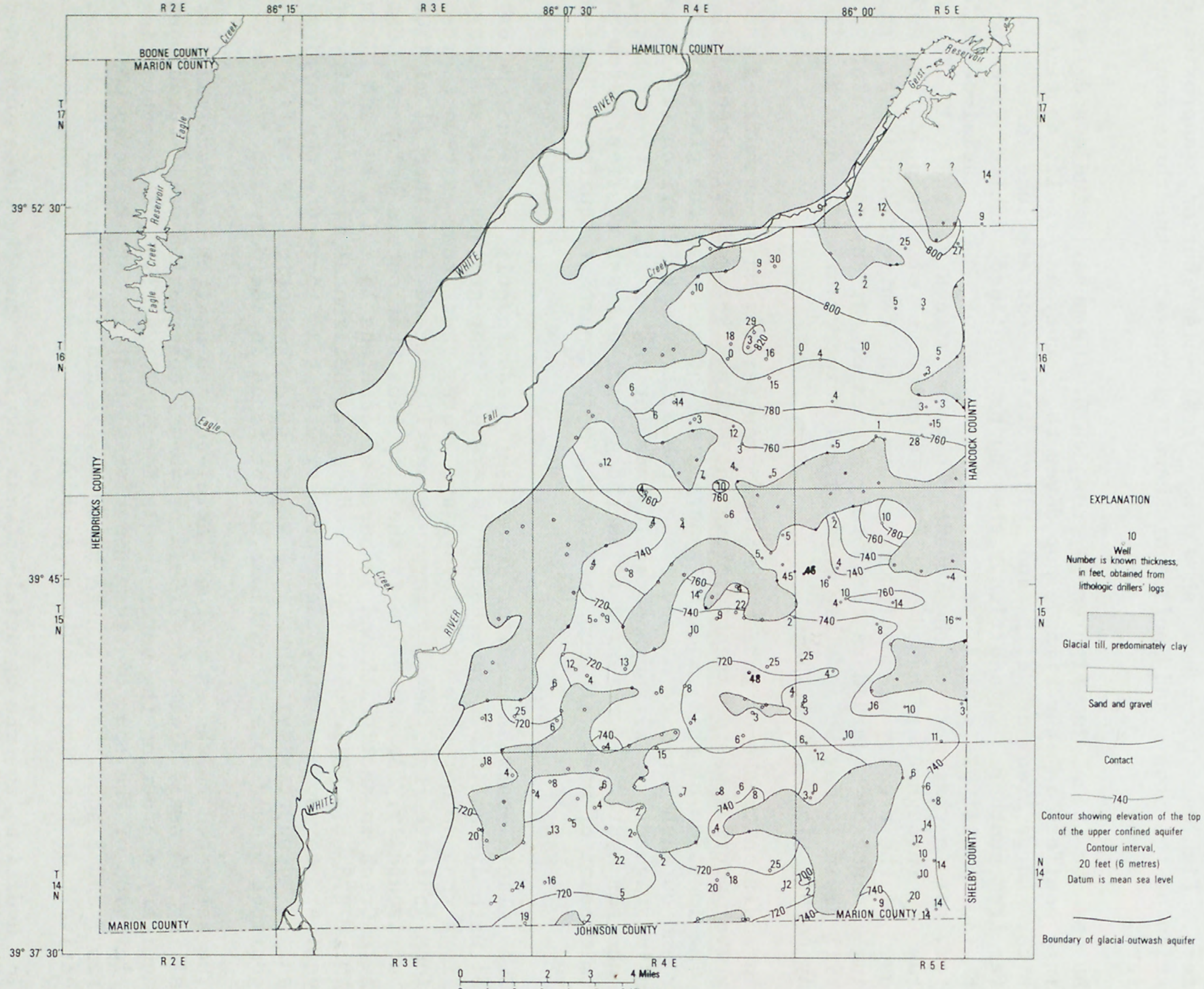

Figure 4. -- Areal distribution, points of known thickness, and the approximate elevation of the surface of the upper confined aquifer

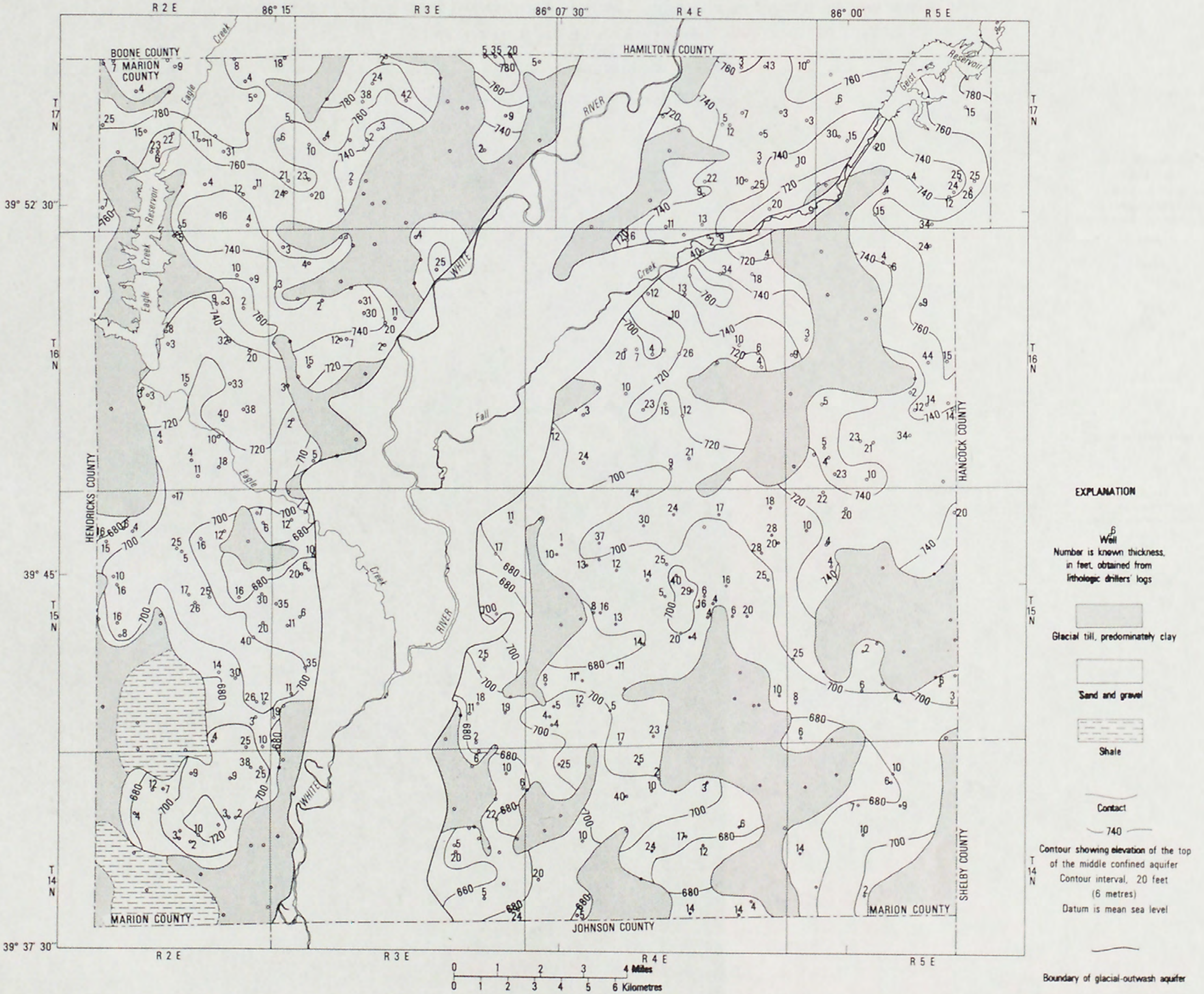

Figure 5. -- Areal distribution, points of known thickness, and the approximate elevation of the surface of the middle confined aquifer

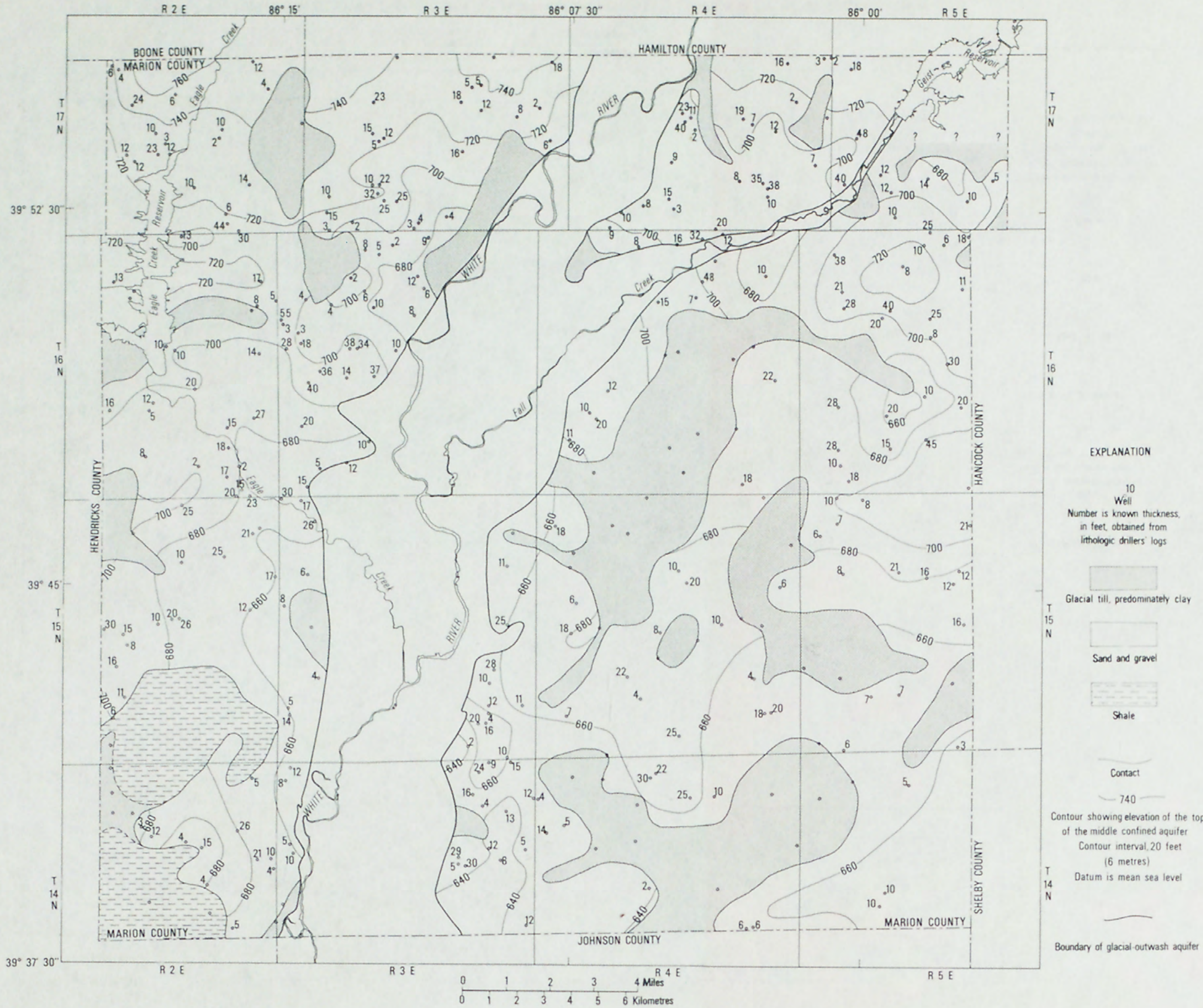

Figure 6. -- Areal distribution, points of known thickness, and the approximate elevation of the surface of the lower confined aquifer

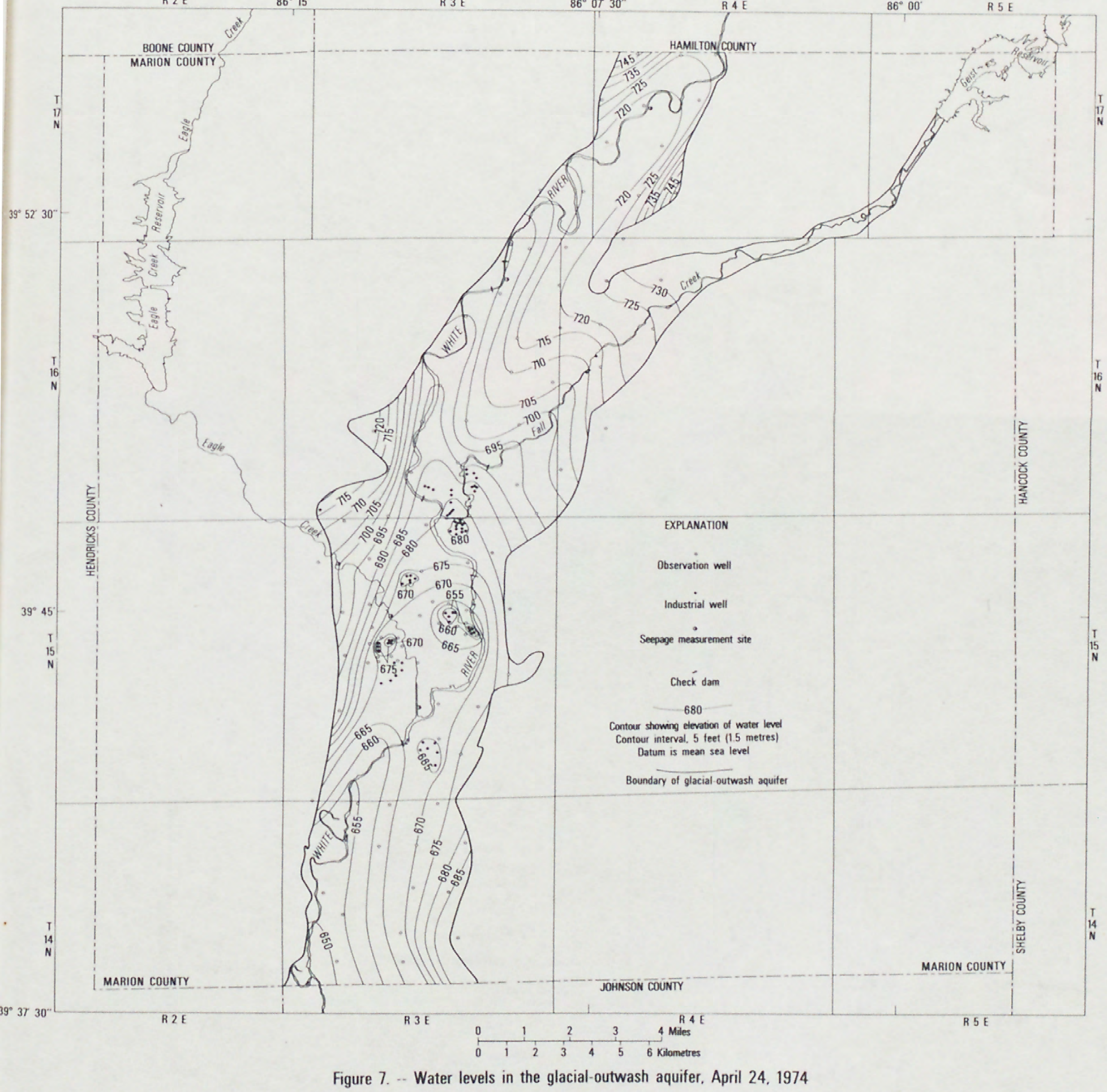

Figure 7. -- Water levels in the glacial-outwash aquifer, April 24, 1974

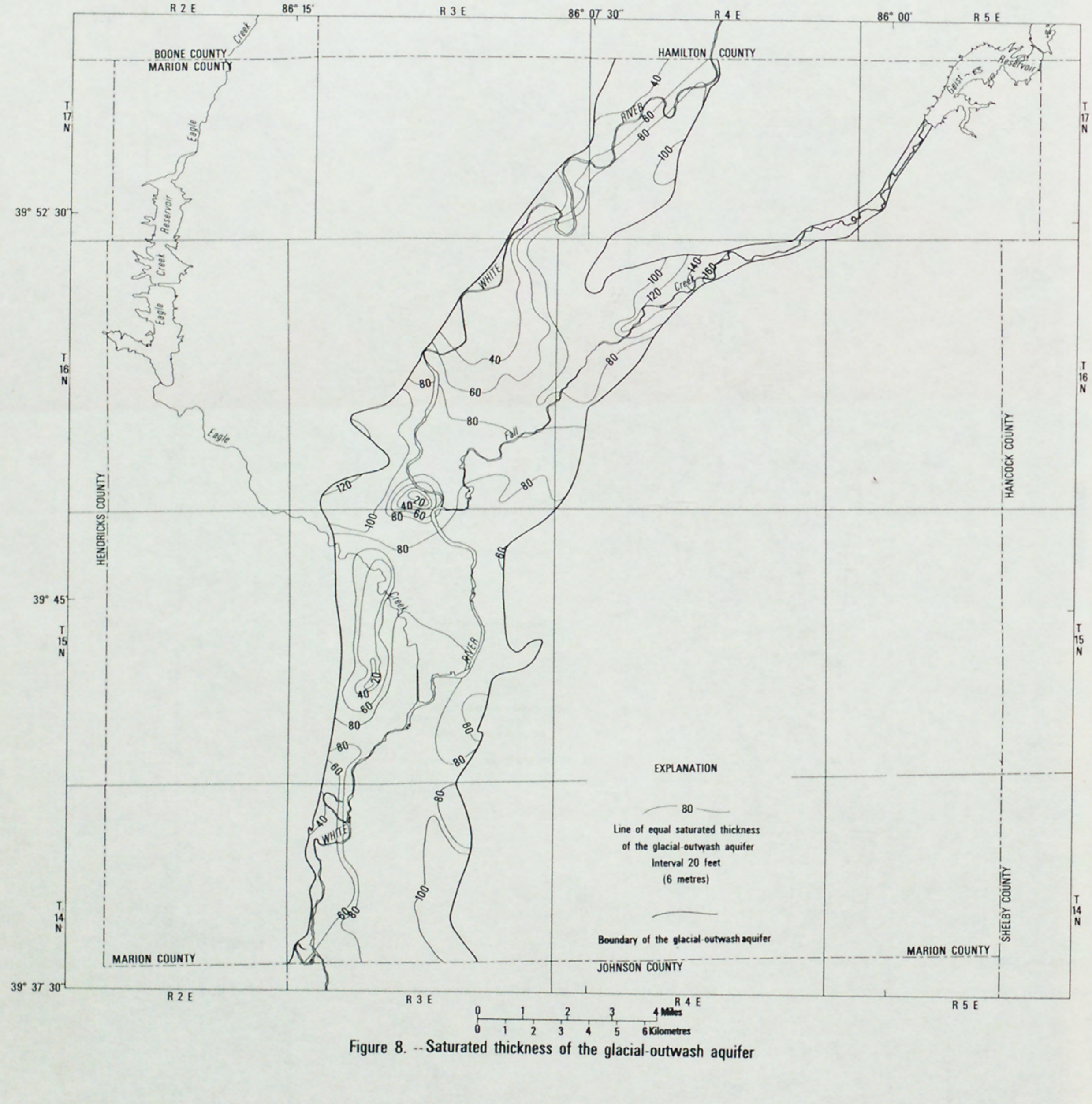

Figure 8. --Saturated thickness of the glacial-outwash aquifer

GROUND-WATER FLOW

The general pattern of ground-water flow in the county is as shown in the cross section of figure 9, which represents conditions in a generalized section taken approximately at right angles to the White River. Part of the precipitation falling on the land surface infiltrates to the water table. In upland areas--that is, on the till plain--some of this water circulates through shallow patterns in the semiconfining material toward small upland streams and springs, where it is discharged to the surface-water system. The rest penetrates to the upper confined aquifer. As shown by the arrows on figure 9, this water circulates downward, laterally and upward toward the major streams, where it is discharged. The pattern of circulation extends through the three confined aquifers and the semiconfining units that separate them, and penetrates to an unknown depth into the underlying bedrock. The pattern of circulation in the outwash aquifer is similar except that most of the water infiltrating to the water table discharges directly to the major streams; little or none is discharged to minor stream channels in local circulation.

In this report, the term effective recharge is used to designate the precipitation that reaches the upper confined aquifer in the till plain, or that reaches the water table in the outwash aquifer.

A certain amount of the ground water, both in the till plain and the outwash, is diverted to wells. The fraction so diverted has varied historically, as will be explained in a later section, but is presently about 50 percent of the effective recharge within Marion County.

Figure 9 illustrates conditions in a vertical plane taken roughly at right angles to the White River. The ground-water flow system, however, is actually three-dimensional and includes components in the downstream direction as well. However, regardless of the relative magnitudes of flow components transverse to or along the streams, the pattern is always one of ground-water discharge into the major stream channels with some diversion to wells. Although this study was concerned solely with Marion County, there is obviously both inflow and outflow of ground water across the boundaries of the county.

At present (1975), the ground-water flow system in the county is in dynamic equilibrium; that is, over a period of several months, the sum of recharge to the system through precipitation and ground-water inflow to the county is equal to the sum of ground-water discharge to the streams, pumpage from wells and ground-water outflow from the county. Over short time periods this balance may be upset, resulting in the temporary accumulation or depletion of ground water in storage.

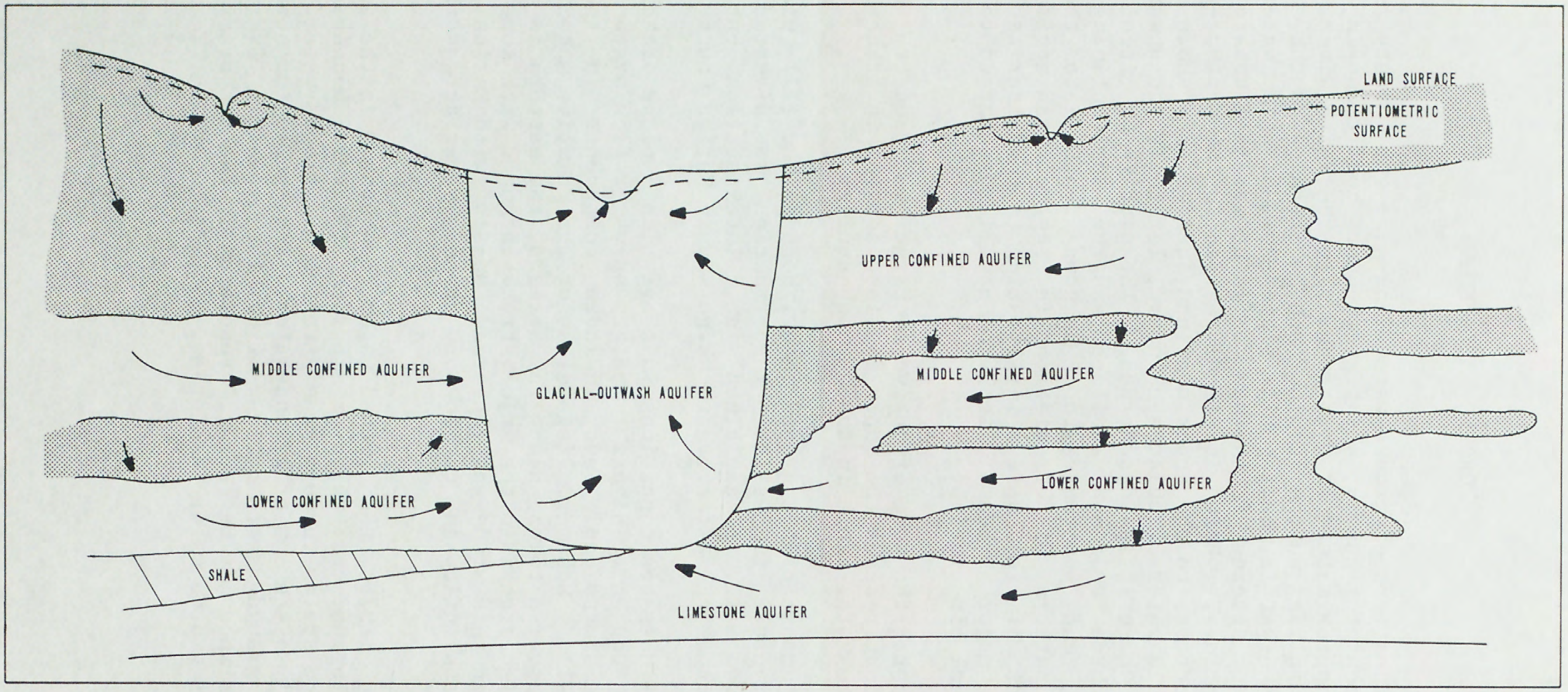

Figure 9.-- Idealized west to east cross section with arrows indicating direction of ground-water flow

HYDRAULIC PROPERTIES OF THE SEDIMENTS

Bedrock

Cable and others (1971) have shown that approximately the first 100 ft (30 m) of the limestone (including dolomite) acts as a single aquifer with an average hydraulic conductivity equal to about 13 ft/d (4 m/d). This system locally yields potable water to wells in relatively large quantity. In general, however, the relatively low hydraulic conductivity of this aquifer compared to that of the overlying sand and gravel in the glacial drift means that it represents a secondary aquifer system in the county.

The hydraulic conductivity of the shale, and the rocks below the first 100 ft (30 m) or so of limestone and dolomite is so low that, for the purposes of this report, this material can be neglected.

Glacial-Outwash Aquifer

The transmissivity (T) of the glacial-outwash aquifer at any point is a function of the type and saturated thickness of the material (gravel, sand and gravel, sand, or silt and clay) underlying the point. A knowledge of this information allows aquifer transmissivity to be estimated from the relationship:

$$T = \sum_{i=1}^{3} m_i K_i \qquad (1)$$

where:

m_1, K_1 = the total saturated thickness and hydraulic conductivity, respectively, of gravel

m_2, K_2 = the total saturated thickness and hydraulic conductivity of mixed sand and gravel

m_3, K_3 = the total saturated thickness and hydraulic conductivity of sand

The hydraulic conductivity of the silt and clay is so low relative to the other material that it can be neglected.

Hydraulic conductivity values for gravel, mixed sand and gravel, and sand were estimated using specific capacity data from 550 small domestic wells throughout the glacial-outwash aquifer. As a rule, these wells are equipped with short screens, tapping a limited section of the aquifer within which the lithology is consistent and can be clearly identified as gravel, sand and gravel, or sand. The wells were divided into three groups according to the lithologic material reported in the screened interval, and their specific-capacity data were used to obtain an average hydraulic conductivity for each type of material.

The specific capacities of the wells in each group were converted to transmissivity values for the individual screened intervals using the non-steady technique for unconfined aquifers given by Theis (1963). This method makes use of the relation:

$$T' = Q/s\ (K - 264 \log 5S + 264 \log t) \qquad (2)$$

where Q is the well discharge, in gallons per minute; s is the drawdown in the well after pumping for time t in days; S is the aquifer specific yield, dimensionless; and T' is a value related to transmissivity by means of a graph given by Theis (1963, p. 334). K is a constant for a given well of radius, r, which, for an unconfined aquifer, is given by the relationship:

$$K = -66-264 \log (3.74r^2 x\ 10^{-6}) \qquad (3)$$

The value of T' was converted to transmissivity, T, for the screened interval using a graph presented by Theis. The value of screened-interval transmissivity thus obtained was divided by the screen length of the well to obtain the hydraulic conductivity. Individual values within each group--that is corresponding to each lithologic type--were then averaged to obtain the following average hydraulic conductivities:

Material	Hydraulic conductivity (ft/d)	Hydraulic conductivity (m/d)
Gravel	415	126
Sand and gravel	240	73
Sand	40	12

This method of hydraulic conductivity determination requires the use of an assumed value of storage coefficient and depends on several assumptions. Well entrance losses are assumed negligible, and the flow into the well is assumed to be horizontal, radial, confined to the depth interval penetrated by the well screen, and sustained by withdrawal from aquifer storage within the interval.

The assumption regarding entrance losses is probably justified, as discharge rates among the domestic wells were generally low. The errors generated by deviation from the remaining assumptions, and by incorrect estimation of storage coefficient, were assumed to be random and, therefore, to be minimized in the process of averaging the hydraulic conductivity values for each lithologic type. In fact, however, it is possible that the hydraulic conductivities contain a small systematic error, as the vertical span of the flow pattern toward the well screen would always exceed the screened interval of the well. This would cause the calculated hydraulic conductivities to be slightly high.

Both the specific capacity data and the lithologic information utilized in the hydraulic conductivity calculations were obtained from drillers' records. The methods of specific capacity testing varied from one driller to another, and the identification of lithologic type is of necessity somewhat subjective. Thus, the data were of variable quality; again, however, errors in the results due to inaccurate data were assumed to be random and, therefore, to be minimized in the averaging process.

The average values of hydraulic conductivity for each lithologic type were utilized in equation 1 for each point in the glacial-outwash aquifer at which a lithologic well log was available, to obtain an approximate aquifer transmissivity at that point.

The final hydrologic analyses in this study utilized a four-layer electric-analog model of the aquifer system in Marion County. The transmissivity values obtained through equation 1 were utilized for the glacial-outwash aquifer in the initial configuration of this model. During model calibration, these transmissivity values were increased by 11 percent throughout the system, and were further adjusted in a few local areas, to achieve closer agreement between model results and field data. This process is described in greater detail in a later section. The final transmissivity distribution, following this calibration process, is shown in figure 10.

The long-term storage coefficient of the glacial-outwash aquifer was estimated to be 0.11 based on the results of a 3-day aquifer test of a well in the outwash, and on an electric-analog representation of the nonsteady decay of water levels in the southern one-third of the aquifer during a 55-day period of drought from June 10 to August 5, 1974. This is discussed further in a later section.

An aquifer test in the glacial outwash, analyzed using type curves presented by Stallman (1965), yielded a ratio of vertical to lateral hydraulic conductivity of 1:10. As no other field evidence was available, and as the value obtained is typical of glacial-outwash material, it was assumed for purposes of this study that the 1:10 ratio prevailed throughout the glacial-outwash aquifer.

The vertical hydraulic conductivity of the extensive clay layer within the outwash, described on page 9, was initially assumed to be 6.7×10^{-4} ft/d (2.04×10^{-4} m/d). During calibration of the analog model this value was progressively adjusted to achieve closer agreement with field data. Ultimately, an areal distribution of vertical hydraulic conductivities ranging from 10^{-2} to 7×10^{-2} ft/d (3×10^{-3} to 2×10^{-2} m/d) was used to represent this clay.

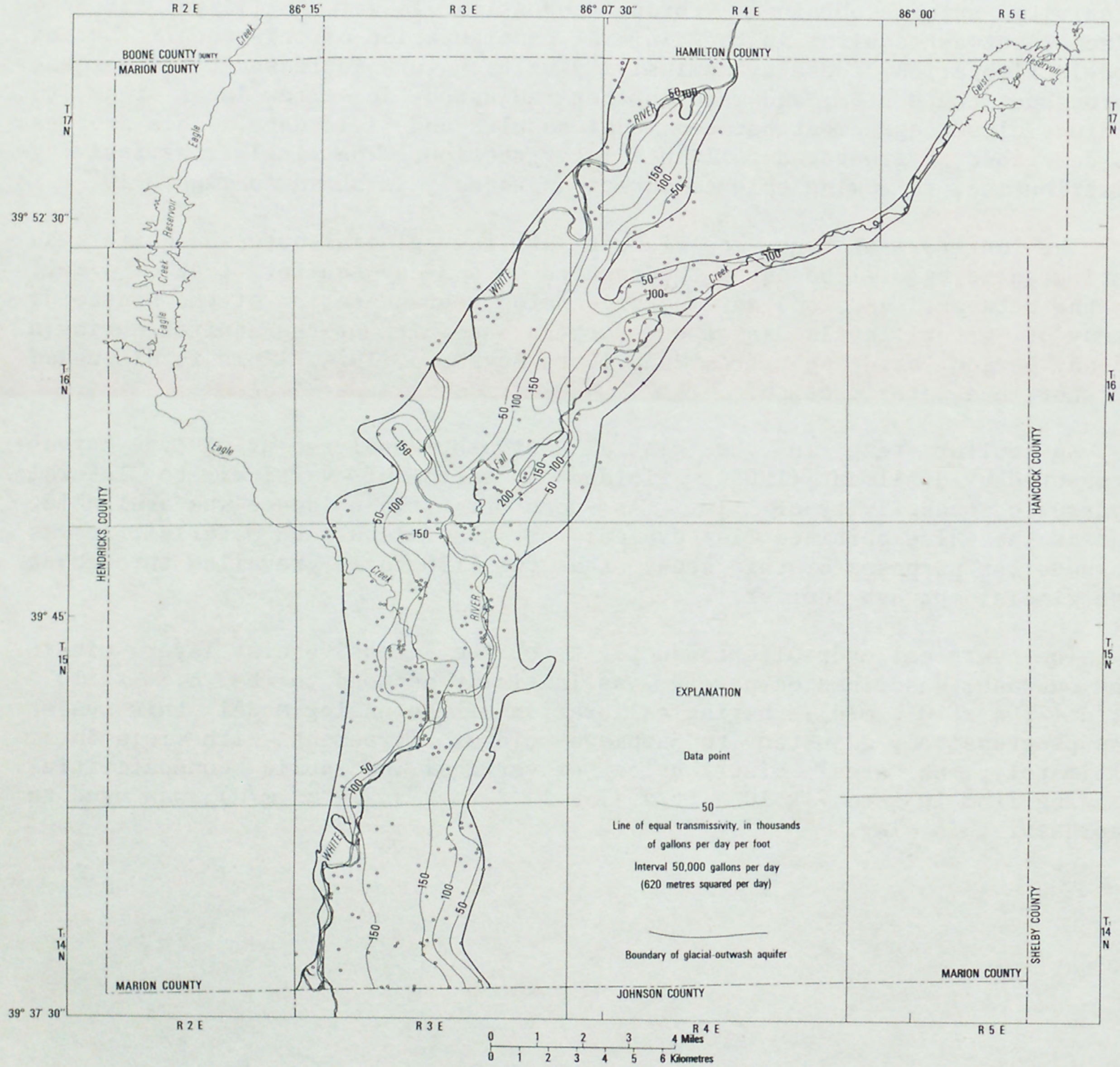

Figure 10.-- Transmissivity of the glacial-outwash aquifer

The Confined Aquifers

An analysis of the specific capacities of domestic wells, similar to that used for the glacial-outwash aquifer, was employed to determine hydraulic conductivities for the material of the confined aquifers. The equation used to calculate transmissivities of the screened intervals were those given by Brown (1963):

$$T' = Q/s\ [K-264 \log_{10} (5\ S \times 10^3) + 264 \log_{10} t] \qquad (4)$$

which is similar to equation 2 except for the factor 10^3, which is inserted for confined aquifers; and

$$K = -66 -264 \log_{10} (3.73\ r^2 \times 10^{-9}) \qquad (5)$$

which is similar to equation 3 except that the factor 10^{-6} is replaced by 10^{-9} for confined conditions. The terms used in equations 4 and 5 are identical to those in equations 2 and 3 except that S represents confined storage coefficient, rather than specific yield. Screened interval transmissivities calculated by equations 4 and 5 were divided by screen length to obtain hydraulic conductivity values, as in the preceding analysis. The assumptions underlying equations 4 and 5 are equivalent to those underlying equations 2 and 3, and the method is subject to the same uncertainties. The shortcomings in the data for the confined aquifers, moreover, were at least as great as those for the glacial-outwash aquifer.

Initially, the hydraulic conductivity data was classified into groups according to lithology. When this was done, the average hydraulic conductivity values for gravel, mixed sand and gravel, and sand were found to be very close. It could not be determined whether this result actually indicated a more uniform composition of the confined material, or simply reflected inadequacies in the data. However, as the individual average values of hydraulic conductivity for each lithologic type were so close, the transmissivity of the confined aquifers was estimated by assuming a single average hydraulic conductivity of 390 ft/d (120 m/d) for the aquifer material. The thickness of each aquifer was contoured and the thickness values were multiplied by this average hydraulic conductivity to obtain transmissivity contours for each aquifer. Again, these values were used in the initial configuration of the analog model; they were uniformly increased by 11 percent, and further adjusted in individual areas, during the process of model calibration. The final transmissivity distributions for the three layers, following model calibration, are shown in figures 11, 12 and 13.

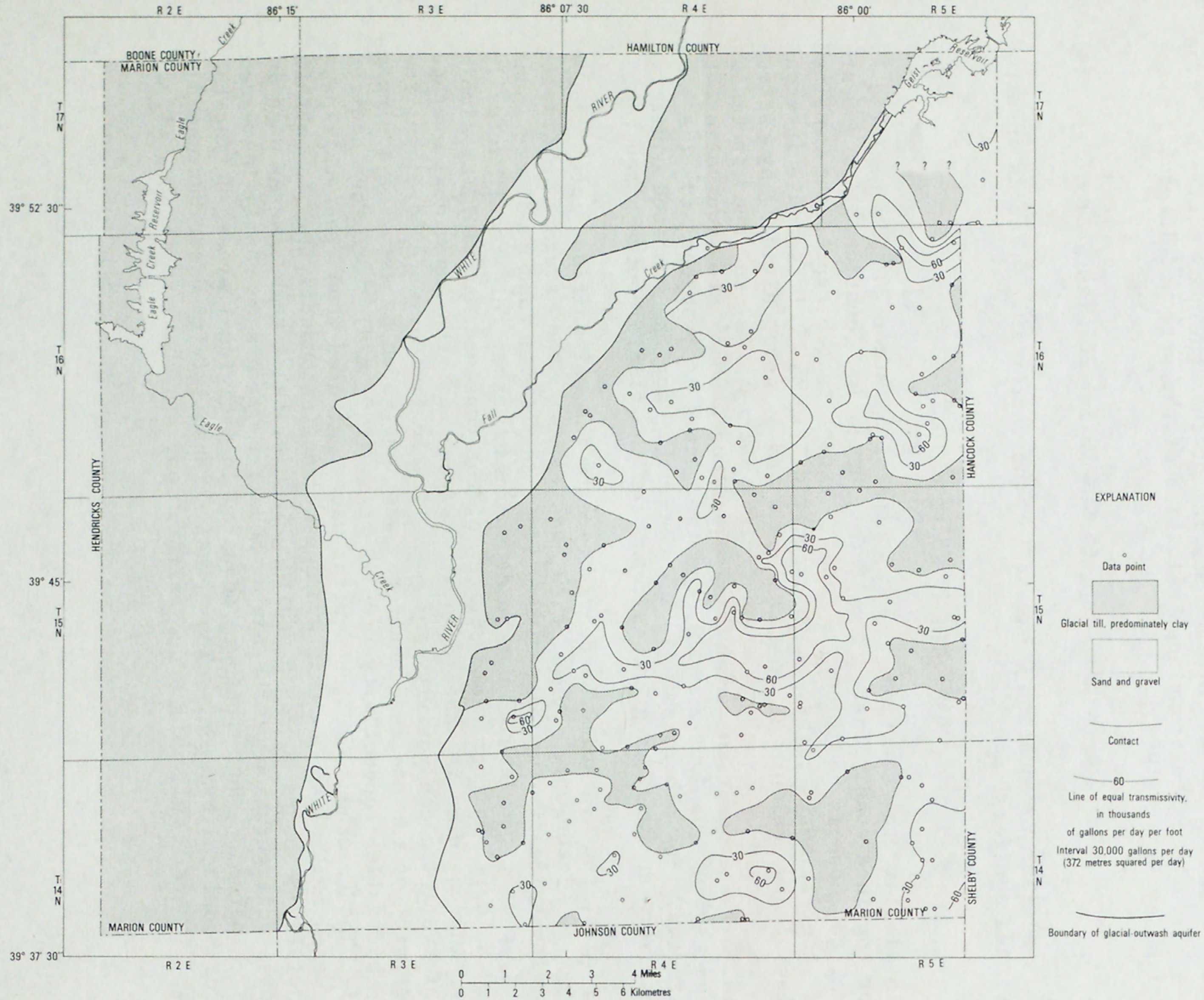

Figure 11.-- Transmissivity of the upper confined aquifer

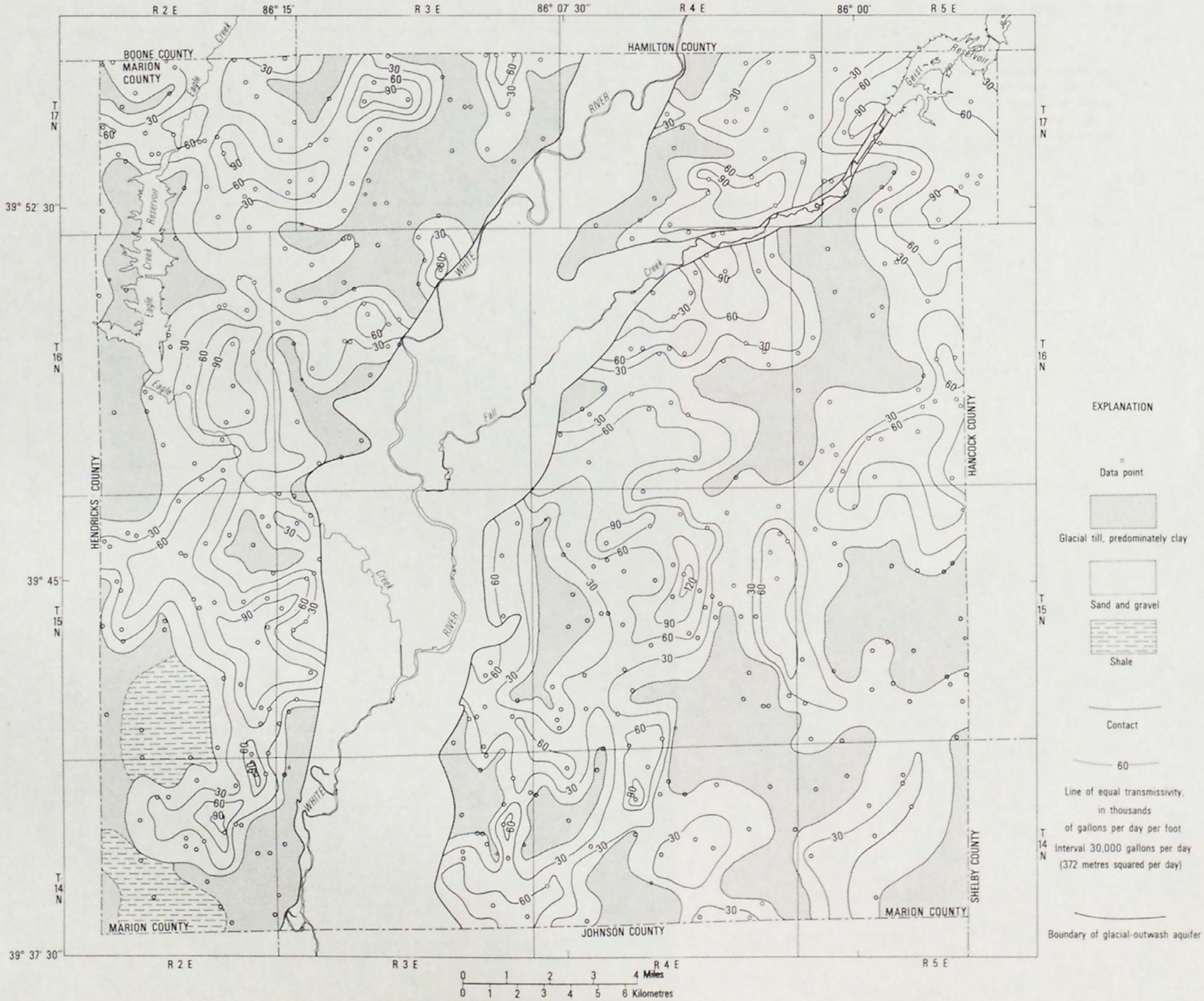

Figure 12.-- Transmissivity of the middle confined aquifer

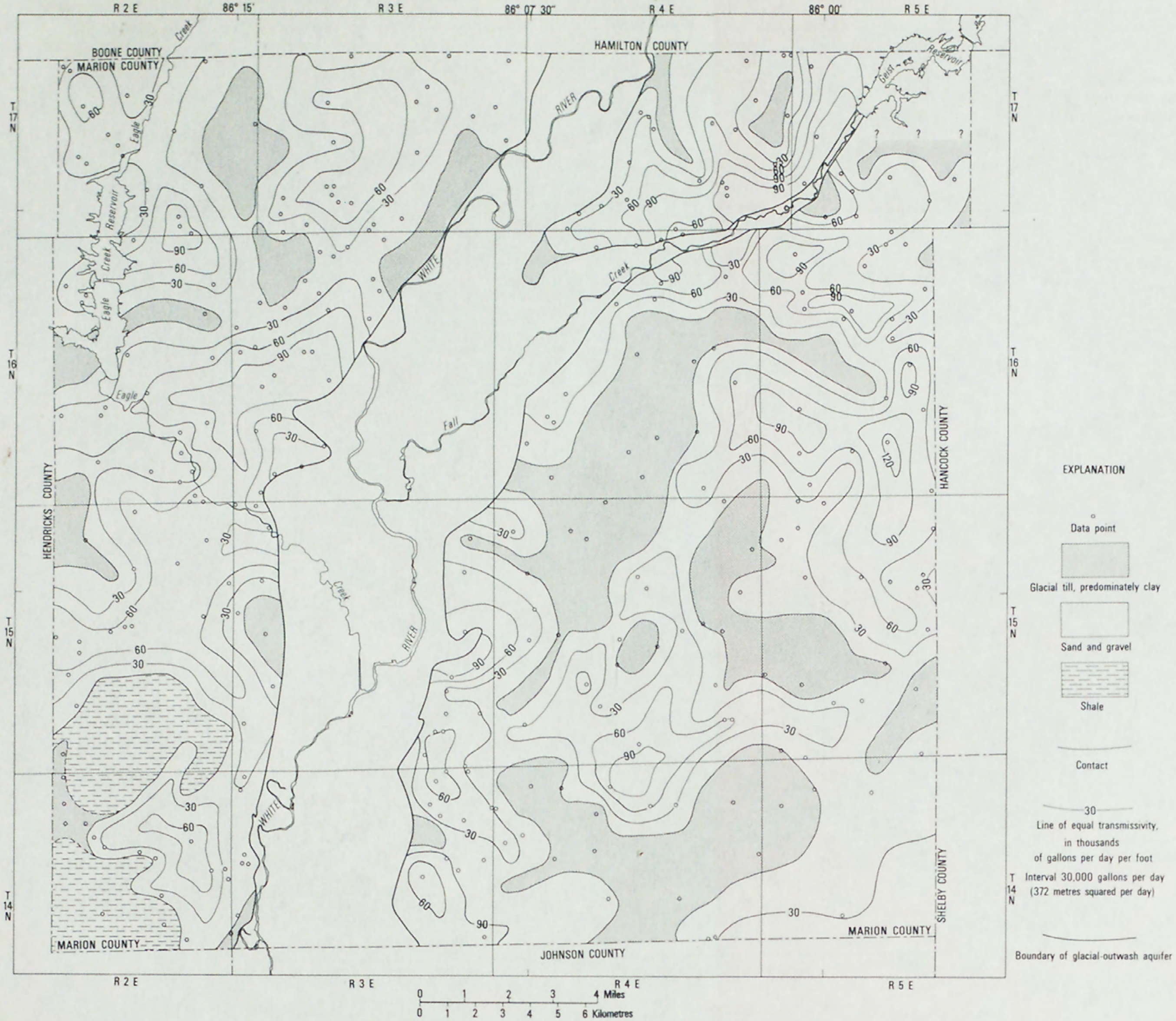

Figure 13.-- Transmissivity of the lower confined aquifer

The storage coefficient of the confined aquifers was estimated from the equation given by Jacob (1940) whereby:

$$S = S_s m = \theta \gamma m (\beta + \frac{\alpha}{\gamma}) \tag{6}$$

for

γ = specific weight of water

β = reciprocal of the bulk modulus of elasticity of water

α = reciprocal of the bulk modulus of elasticity of the solid skeleton of the aquifer

m = thickness of aquifer

S_s = specific storage of the aquifer

θ = aquifer porosity

The confined sand and gravel aquifers were assumed to be twice as compressive as water and θ was assumed to be 0.30. The specific storage thus calculated for the three confined aquifers was 3.3×10^{-6}. Because of the uncertainties inherent in this technique, no attempt was made to contour storage coefficient by multiplying this specific storage by individual values of aquifer thickness. Rather, an average thickness of 20 ft (6.1 m) was assumed for each aquifer, and an average storage coefficient of 6.6×10^{-5} was thus calculated for each aquifer. The storage coefficient of the limestone aquifer was estimated as 10^{-4}.

The Semi-permeable Confining Beds

As indicated in the cross section of figure 9, vertical flow through the semi-permeable confining beds in the till-plain area is an important feature of the ground-water flow system. The property that controls this flow is the vertical hydraulic conductivity of the semiconfining units. An initial estimate of this vertical hydraulic conductivity was obtained through experiments with a small steady-state electric-analog model representing flow in a cross section, such as shown in figure 9. The model consisted of a simple resistor network, one edge of which represented the water table. Head values--that is, model voltages--along this edge were controlled to represent steady-state water-table heads across the line of section. The vertical resistances in the model were then progressively adjusted until heads throughout the model were brought into agreement with head values observed in the field at corresponding depths in the section. The final resistance values corresponded to a vertical hydraulic conductivity of 6.7×10^{-4} ft/d (2.04×10^{-4} m/d).

For the semiconfining layers below the uppermost--that is, for those separating one confined aquifer from another, or from the bedrock--this estimate was used as the vertical hydraulic conductivity in the initial configuration of the four-layer analog. During model calibration, the conductivity was adjusted progressively to improve the match between model results and field data. The final distribution of hydraulic conductivity obtained for the semiconfining layers below the uppermost ranged from 10^{-4} to 1.3×10^{-3} ft/d (3.3×10^{-4} to 4.3×10^{-3} m/d).

For the uppermost semiconfining layer--that is for the interval between the water table and the highest confined aquifer present at a given location--a slightly different procedure was used. During the calibration of the four-layer model, the recharge flow crossing this uppermost confining layer was adjusted progressively to achieve the best possible agreement between heads in the model and those in the aquifer. A more detailed description of this work is given in a later section. The analysis resulted in a set of recharge figures giving the flow actually penetrating to the uppermost confined aquifer in each part of the till plain. Letting v represent the recharge flow per unit area determined in this way, the vertical conductivity of the uppermost confining bed was calculated for each section of the modeled area from the relation:

$$K_v = \frac{v\,m'}{h_w - h_c} \qquad (7)$$

where K_v represents vertical hydraulic conductivity; m' represents the interval between the water table and the uppermost confining bed; h_w represents the water-table head; and h_c represents the head in the uppermost confined aquifer. Because the exact water-table elevation within the upper semiconfining unit was generally unknown, it was assumed arbitrarily to be 10 ft (3 m) below land surface throughout the till-plain area for purposes of this calculation.

The values of vertical hydraulic conductivity obtained in this way for the uppermost semiconfining layer ranged from 2×10^{-4} to 1.5×10^{-2} ft/d (6×10^{-5} to 4.6×10^{-3} m/d). One possible source of error in this method of calculation lies in the fact that the actual flow through the upper semiconfining bed includes a large fraction that discharges to upland streams and springs, whereas in the analog analysis only the fraction that penetrates to the uppermost confined aquifer was represented. The resultant error would tend to make the calculated values of vertical hydraulic conductivity somewhat low.

THE WHITE RIVER

The average flow of the White River at the gaging station near Nora, in the northern part of the county, was 1,050 ft^3/s (30 m^3/s) for the period between October 1929 and September 1967 (Rohne, 1972).

Flow in the White River is not regulated, although small check dams at points along its course in the county (fig. 7) cause the stream velocity in their immediate vicinity to be reduced from that in unaffected sections of the stream. The flow of both Fall Creek and Eagle Creek is regulated by reservoirs within the county. Cicero Creek, a major tributary entering the White River north of Marion County, also is regulated. Thus the flow of the White River within Marion County is indirectly controlled. A major diversion of flow from the White River occurs at Broad Ripple, just below the Nora gaging station, where water is diverted into the Indianapolis Water Company's Canal. Water is also diverted from Fall Creek at the Indianapolis Water Company's Fall Creek filtration plant. Treated sewage effluent represents a major addition to river flow at the junction of Eagle Creek and the White River and near Southport Road in the southern part of the county.

The 7-day, 10-year low flow of the White River at the Nora gaging station, calculated for the same period of record as above, is 70 ft^3/s (2 m^3/s). The 7-day, 10-year low flow of the White River at the Morris Street gaging station, which is 17.6 river mi (28.3 km) below the Nora gage and 2.5 mi (4 km) below the junction of Fall Creek and White River, is only 49 ft^3/s (1.4 m^3/s) for the period of record from March 1904 to July 1906 and April 1930 to September 1967. The minimum daily discharge recorded for the White River at the Nora gaging station was 49 ft^3/s (1.4 m^3/s) in September 1941. At the Morris Street gaging station the minimum daily discharge of the White River recorded was 8 ft^3/s (0.23 m^3/s), also in September 1941 (Rohne, 1972). The lower discharge at the downstream gaging station reflects the effects of the diversions from the White River at the Indianapolis Water Company's canal and the diversion from Fall Creek at the Indianapolis Water Company's Fall Creek filtration plant.

GROUND-WATER SEEPAGE TO STREAMS

The observation-well system installed in the glacial-outwash aquifer was designed in part to establish the relationship of the ground-water system to the major streams in the county. Contour maps of water levels in the glacial-outwash aquifer prepared for various times of the year (April 24, July 30, November 4, 1974, March 20, 1975) indicate that the streams are gaining water from the ground-water system in Marion County throughout the year.

The rate at which a stream gains ground water along any given segment of its reach depends upon the hydraulic properties of the aquifer, the vertical hydraulic conductivity of the streambed, the recharge to the aquifer, and the stream stage. Inasmuch as recharge and stream stage vary with time, seepage of ground water into a stream must also vary with time. The check dams along the White River (fig. 7) influence the rate at which the White River and Fall Creek gain ground water, both by modifying the stream stage and by altering the hydraulic conductivity of the streambed as a result of silting.

Attempts were made to measure ground-water seepage into several reaches along the major stream channels by measuring streamflow simultaneously at the upstream and downstream ends of the reach. With one exception, these reaches were located within the municipal area. For all the reaches within the municipal area, the seepage results were obscured by uncertainties in other components of inflow or outflow, such as water diversion, tributary inflow, sewage effluent disposal, cooling water disposal, return of unused diversions, and channel storage behind check dams. The single reach outside the municipal area was unaffected by tributary inflow or by any disruption of the flow regime due to human activity. This reach was on the White River north of the metropolitan area, and actually extended beyond the northern border of Marion County. The measured reach was approximately 6.6 mi (10.6 km) in length; its lower end was approximately 3 mi (4.8 km) downstream from the county line.

The seepage measurement was made on October 20, 1973, during the season when flow should have been close to its lowest value for the year. The upstream measurement was 272 ft^3/s (7.7 m^3/s) and the downstream measurement was 289 ft^3/s (8.2 m^3/s). Taking the difference as ground-water inflow, seepage into the channel is calculated as 42 ft^3/d (3.9 m^3/d) per linear foot (metre) of channel. This result agrees closely with seepage measurements made farther north on the White River, in connection with a study in the Carmel area.

It is hazardous to extrapolate a single seepage measurement into other reaches of the system, where hydrologic conditions may differ substantially. This method was nevertheless utilized, as there was no other basis upon which to estimate seepages throughout the area. Figure 19 shows five reaches along the White River, one along Eagle Creek and one along Fall Creek. Reach 1 along the White River coincides with the southern half of the reach in which the seepage measurement was made--that is, with the part of the seepage measurement reach lying within the county. The reaches along

Eagle Creek and Fall Creek include only the segments of those creeks lying within the area of the glacial-outwash aquifer. No estimates of seepage were made for Eagle Creek or for reach 4 on the White River, as heavy pumping occurs near these reaches, and diversion of ground water to the well fields would almost certainly cause the seepage rates to differ markedly from the figure of 42 ft^3/d per linear foot (3.9 m^3/d per linear metre) determined in the measurement. Table 1 presents calculated total seepages for the five remaining reaches--Fall reaches 1, 2, 3 and 5 on the White River--assuming that the seepage rate of 42 ft^3/d per linear foot (3.9 m^3/d per linear metre) prevailed in each of these reaches.

Table 1.--Calculated seepages

	Reach	Calculated Seepage assuming an inflow of 42 ft^3/day per foot of channel (Mgal/d)
White River	1	5
	2	12
	3	12
	5	10
Fall Creek		23

Because of the uncertainties inherent in extrapolating a single seepage measurement throughout the area, the figures in table 1 can be taken only as a general indication of the magnitudes of seepage to be expected in the county. Because the measurement was made under low-flow conditions, it seems probable that the calculated results are conservative. It should be noted that the figures refer only to direct ground-water seepage into the major channels, and do not in any way constitute the total contribution of ground water to streamflow within the county. The total discharge of ground water into the stream system would include all spring flow and seepage into all minor streams throughout the county, particularly in the till plain.

DEVELOPMENT OF GROUND WATER IN MARION COUNTY

Ground-water pumpage within Marion County for various time periods has been discussed in reports by McGuinness (1943), Roberts and others (1955), and Carter (written commun., Indiana State Board of Health, 1969). During this investigation, new efforts were made to extend and refine the existing data on historical rates of ground-water withdrawal in the county, and on the water levels associated with these withdrawals.

Municipal and Industrial Use

According to McGuinness (1943), the first ground-water pumpage of any significance in Marion County began in the early 1880's. At that time and up to the early 1900's, the principal producer of ground water in the county was the Indianapolis Water Company, a privately-owned utility.

The utility initially obtained their water from two well fields, one located on the east side of the White River at Washigton Station and the other located at Riverside. The pumping capacity of these plants grew to 22 Mgal/d (1 m^3/s) and 12 Mgal/d (0.5 m^3/s), respectively.

In 1905, the utility began treating water from the White River, and as the demand for the treated water increased, the utility relied increasingly on surface water instead of ground water. By 1930, less than 30 percent of the utility's processed water was from ground water; this trend continued until by 1941, less than 15 percent of the utility's 37 Mgal/d (1.6 m^3/s) distribution was ground water. (In 1973, less than 4 percent of the utility's total distribution was ground water.) As a result of this use of surface water by the utility, the principal pumpage of ground water shifted from the utility to private industry.

McGuinness reported that in 1941 an average of 51.9 Mgal/d (2.3 m^3/s) of ground water was pumped within the county, of which 39.4 Mgal/d (1.7 m^3/s) was pumped by private industry, 4.8 Mgal/d (0.2 m^3/s) by public water utilities (including Speedway and Lawrence), 4.7 Mgal/d (0.2 m^3/s) by agriculture, and 3.0 Mgal/d (0.1 m^3/s) by private domestic users. McGuinness estimated that about 55 percent of the industrial and municipal ground-water pumpage was from sand and gravel wells and 45 percent was from limestone wells.

The amount of ground-water pumpage during the war years, 1943 to 1945, a period of large industrial expansion and water demand, was not measured but estimates by Roberts and others (1955) suggest withdrawal rates between 77 and 80 Mgal/d (3.4 to 3.5 m^3/s). In a survey made by Roberts and others (1955) for 1952, the total ground-water pumpage in the county was estimated to be 56.3 Mgal/d (2.5 m^3/s). Of this, 39.3 Mgal/d (1.7 m^3/s) was pumped by private industry, 8.7 Mgal/d (0.4 m^3/s) by agriculture, 2.8 Mgal/d (0.1 m^3/s) by public water utilities, and 5.5 Mgal/d (0.2 m^3/s) by private domestic users. During this same year the Indianapolis Water Company distributed an average of 58.1 Mgal/d (2.5 m^3/s) of processed water, of which only a negligible amount (less than 1 percent) was ground water.

Carter (written commun., Indiana State Board of Health, 1969) reported that ground-water withdrawal in Marion County in 1967 for industrial and municipal usage was 32.2 Mgal/d (1.03 m^3/s).

For the current study, approximately 80 industrial, commercial, institutional, and military establishments were contacted in Marion County, as were the four municipal water utilities (Indianapolis Water Company, City of Lawrence, City of Speedway, and Town of Cumberland) in an attempt to fill in the gaps in the existing data. Each establishment was requested to supply a yearly ground-water pumpage for 1930 through 1973, and semiannual water levels in their wells. The replies to the survey were incomplete and the quality of the data ranged widely.

Few water levels were obtained from the recent survey and the quality of much of the records is poor. Only two companies make regular (weekly or daily) checks of water levels in their wells. A few others, who have had well-maintenance contracts, have records of water levels measured at the time of servicing, which is every 2 to 5 years. Most companies, however, have kept no records at all. Therefore, insufficient historical water-level data exist over most of the study area.

Annual ground-water pumpage data obtained from the survey are on the whole, of much better quality. Even so, much information has been lost, as many companies have pumpage records for the past 10 to 15 years but have discarded records for prior years. In addition, other companies that once pumped ground water have ceased to exist and their records have been lost. From the data on pumpage collected, the total annual ground water pumped was calculated for each year since 1930 and then compared with the few published estimates available. The data obtained from the recent survey of industrial and municipal ground- water pumpage could account for only 51 percent of the figure published for 1952 by Roberts and others (1955). For 1967, however, the pumpage data obtained from the recent survey indicated 9 percent more ground-water pumpage in Marion County than had been estimated by Carter (written commun., Indiana State Board of Health, 1969).

Data collected by the present survey were considered acceptable back to about 1960. Table 2 gives the average annual industrial and municipal ground- water pumpage recorded for the entire county from 1960 through 1973. Figure 14 shows the location of wells or well fields that had an average annual pumpage from 1960 through 1973 of 0.1 Mgal/d (0.004 m^3/s) or greater. Despite the limitation of 0.1 Mgal/d, approximately 97 percent of the industrial and municipal pumpage recorded for each year is shown. It is noteworthy that of the pumpage recorded and plotted for 1973, less than 20 percent is from the bedrock. Further, the survey indicated that ground-water pumpage outside the unconfined glacial-outwash aquifer represents less than 20 percent of the total pumpage for the county.

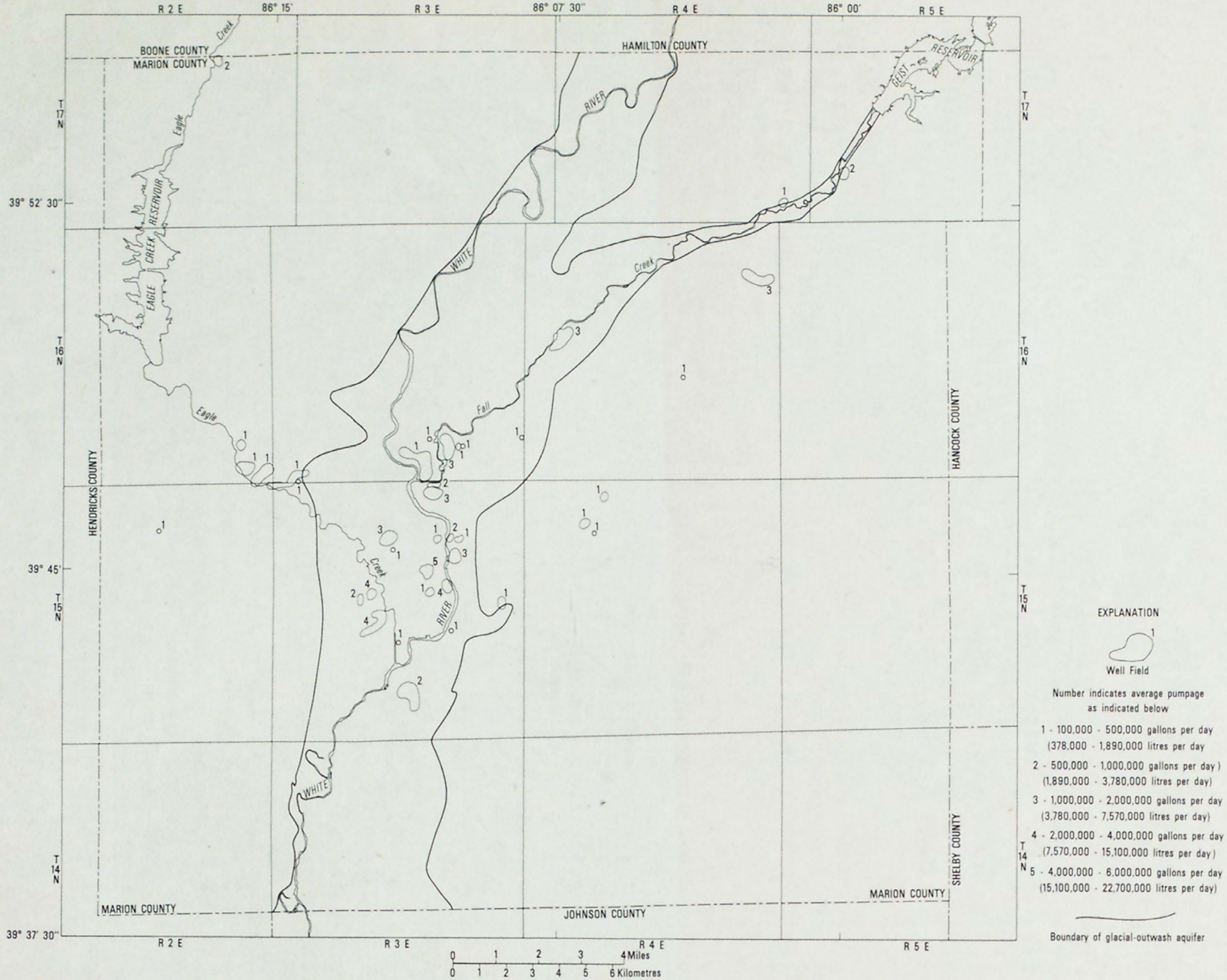

Figure 14. -- Distribution of major industrial and municipal pumpage in Marion County, 1960-74

The survey also indicated that the major centers of ground-water pumpage all occurred within a relatively small area of the glacial-outwash aquifer in Indianapolis, rather than being scattered throughout the county. In addition, the major pumping centers historically and presently are generally located within 1 mi (1.6 km) of one of the major streams.

Table 2.--Average daily industrial and municipal ground-water pumpage in Marion County, in millions of gallons per day, 1960 through 1973

Year	Pumpage	Year	Pumpage
1960	26.8	1967	35.1
1961	28.0	1968	36.9
1962	27.2	1969	39.7
1963	27.9	1970	37.8
1964	31.6	1971	40.2
1965	30.0	1972	36.3
1966	33.5	1973	35.0

Domestic use

The 1973 population of Marion County, according to the Indianapolis Chamber of Commerce, is 839,000. Because the total population served by public water supply is about 739,000, approximately 100,000 people use water from domestic wells. Using acceptable estimates of per capita daily use, the total domestic ground-water pumpage from private wells can be reasonably estimated at between 9.0 and 11.0 Mgal/d (0.4 to 0.5 m^3/s).

Total use

Total ground-water pumpage in Marion County from all sources for 1973 can thus be estimated as from 46 to 48 Mgal/d (2 to 2.1 m^3/s). From these statistics and those published previously, it is apparent that although fluctuations in total ground-water pumpage in Marion County have occurred, the total ground-water pumpage has not changed substantially since the late 1940's. The increase in total demand for water in Marion County has been met instead through increased treatment and distribution of surface water by the Indianapolis Water Company.

Well construction

High capacity water wells in Marion County are of three basic types: tubular bedrock, tubular sand and gravel, and gravel-packed. The tubular bedrock wells range from 6 to 16 in (150 to 410 mm) in diameter and up to

400 ft (122 m) in depth. They are cased from the land surface to the bedrock and then have another 200 to 300 ft (61 to 91 m) of open hole in the bedrock. These bedrock wells are not equipped with conventional well screens and usually have only a strainer around the intake. Even when good aquifer material is present above the bedrock, these wells are rarely screened in it, but are instead cased on through to the bedrock.

The tubular sand and gravel wells range from 6 to 18 in (150 to 460 mm) in diameter and up to about 200 ft (61 m) in depth. They are usually equipped with a conventional well screen 10 to 30 ft (3 to 9 m) long, and are rarely screened in more than one formation or in more than one section of a thick water-bearing formation. Even in the unconfined glacial-outwash aquifer where the saturated thickness of sand and gravel above the bedrock may reach 75 to 100 ft (23 to 30 m), these wells are screened only in the lowermost 10 to 30 ft (3 to 9 m), just above the bedrock, and are cased with blank pipe above the screen.

The gravel-packed wells generally range from 21 to 72 in (530 to 1,830 mm) in diameter. Some of these wells contain two casings with gravel packed in the annular space, whereas others have a smaller diameter casing set in the larger diameter hole with the annular space being packed with gravel. As with the tubular wells, these gravel-packed wells are screened only in the bottom of the aquifer with a screen 10 to 30 ft (3 to 9 m) in length.

WATER-LEVEL CHANGES

The Geological Survey and the Indiana Department of Natural Resources have conducted a cooperative program of measuring water levels in selected observation wells in Marion County since the middle 1930's. The number of wells in the program has fluctuated with time; most of the wells in the observation-well network were measured from the early or middle 1940's into the 1950's. At the present time (1975), there is only one well, Marion-32, being measured on a continuous basis in the county. The location of wells that at one time or another have been in the observation well network is shown in figure 15.

The water levels in these observation wells exhibit a pronounced seasonal fluctuation as shown by the selected hydrographs in figure 16. As can be seen in the figure, the water levels fluctuate in a sinusoidal fashion, with highs occurring in March or April and lows in September or October.

In addition to the seasonal fluctuations, wells for which data are available show an overall water-level decline from the late 1930's to about the middle of 1941 (fig. 16). A slight recovery occurs in most of the data from 1941 to 1942, no change from 1942 to 1944, and then a period of overall recovery beginning in August 1944. As noted by McGuinness (1943), the 12-year period 1930-41 represented the driest period recorded to that time (and actually up to the present) in Indianapolis. The 2 years, 1940 and 1941, are the 2 driest years in succession recorded up to the present day. The water-level decline from the late 1930's to 1941 probably represents the effects of this dry period. The recovery between 1941 and 1942 correlates with the end of the dry period. This recovery trend was apparently interrupted by the heavy pumping during the war, causing a resultant stabilization of water levels through 1944, when the general recovery trend resumed.

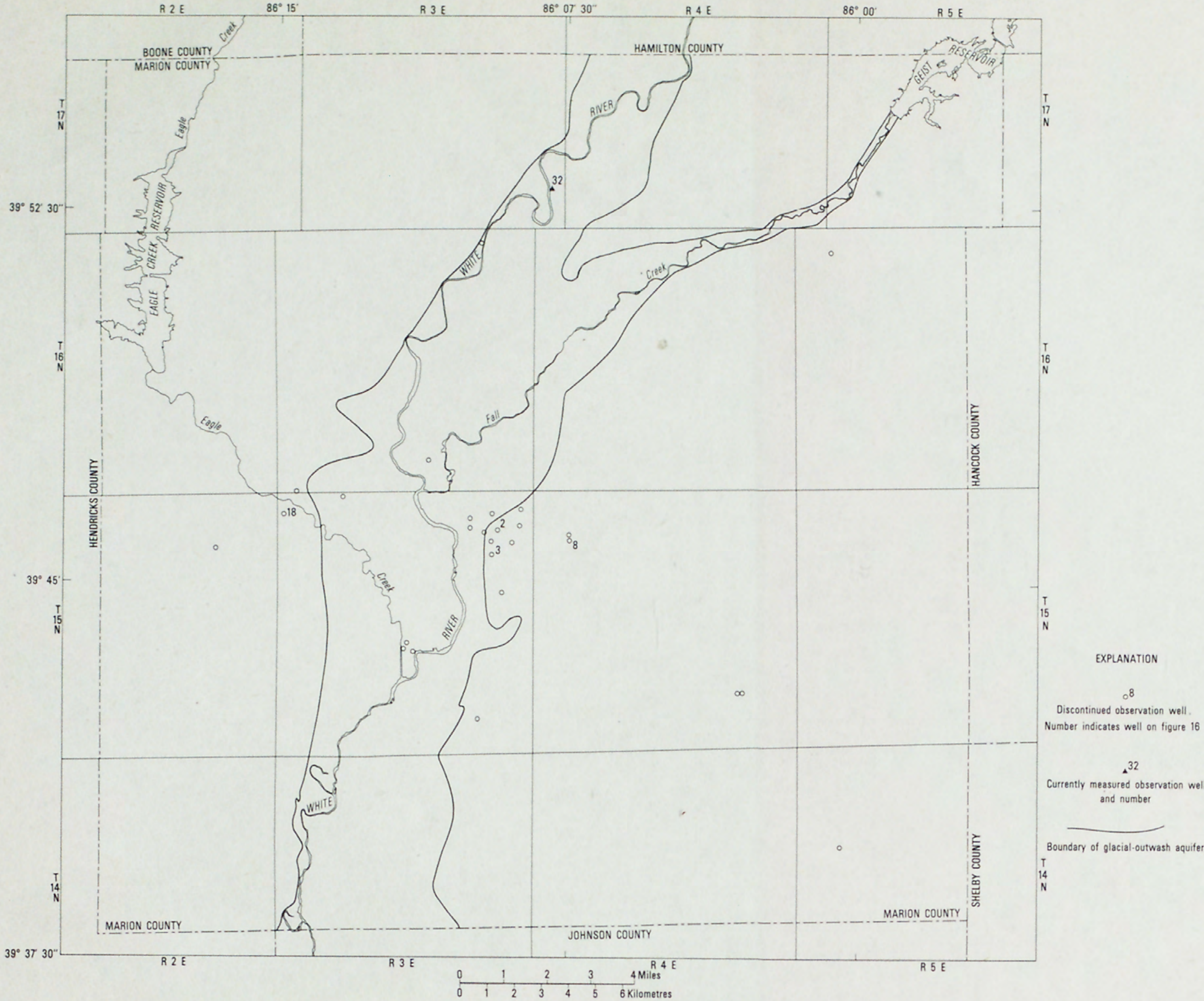

Figure 15. -- Location of observation wells in Marion County

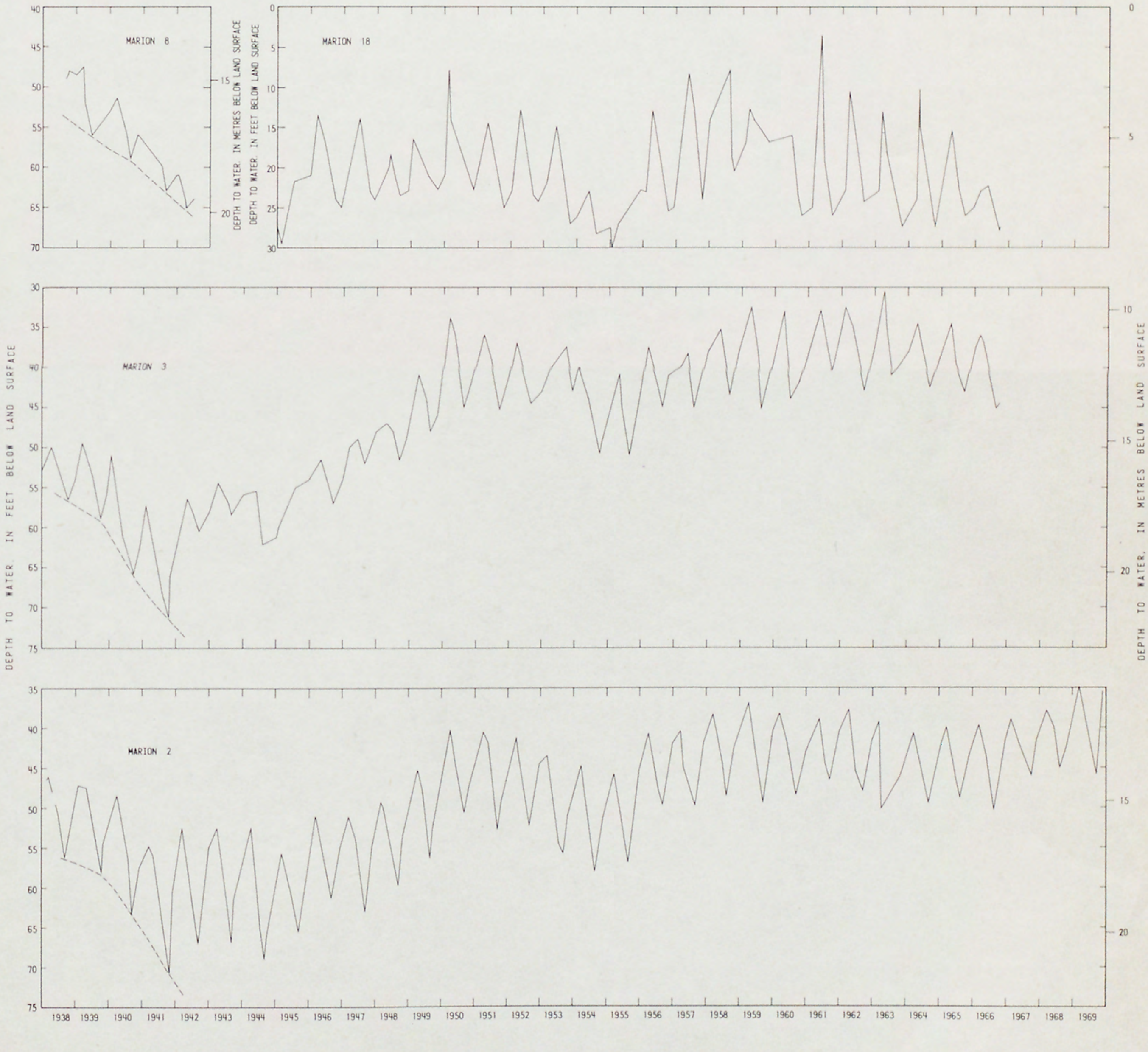

Figure 16.-- Water-level fluctuations in selected observation wells in Marion County

EVAPOTRANSPIRATION

Evapotranspiration is a major element in the water budget of Marion County. However, the plants responsible for the bulk of the transpiration are shallow rooted, and draw water predominantly from the unsaturated zone. Moreover, depths to water throughout the county are sufficient to minimize direct evaporation or transpiration from the water table, except possibly in a narrow zone along the major streams. Figure 17 shows depth to water in the glacial-outwash aquifer in April 1974. Depths to the water table in the till plain are difficult to measure; usually, however, in a hydrologic setting of this type, the depth to water in upland areas exceeds that in the stream valleys. Thus it appears that most of the evapotranspiration in the county is from the unsaturated zone. The significance of this in terms of ground-water development is that no major retrieval of water now lost through evapotranspiration will occur if the water table is lowered due to pumping.

Although evapotranspiration occurs predominantly in the unsaturated zone, a transient effect is nevertheless transmitted to the water table. The seasonal change in the evapotranspiration draft on the unsaturated zone produces an opposite seasonal change in the recharge available for infiltration to the water table. This is at least partly responsible for the seasonal fluctuations in water level indicated in figure 16.

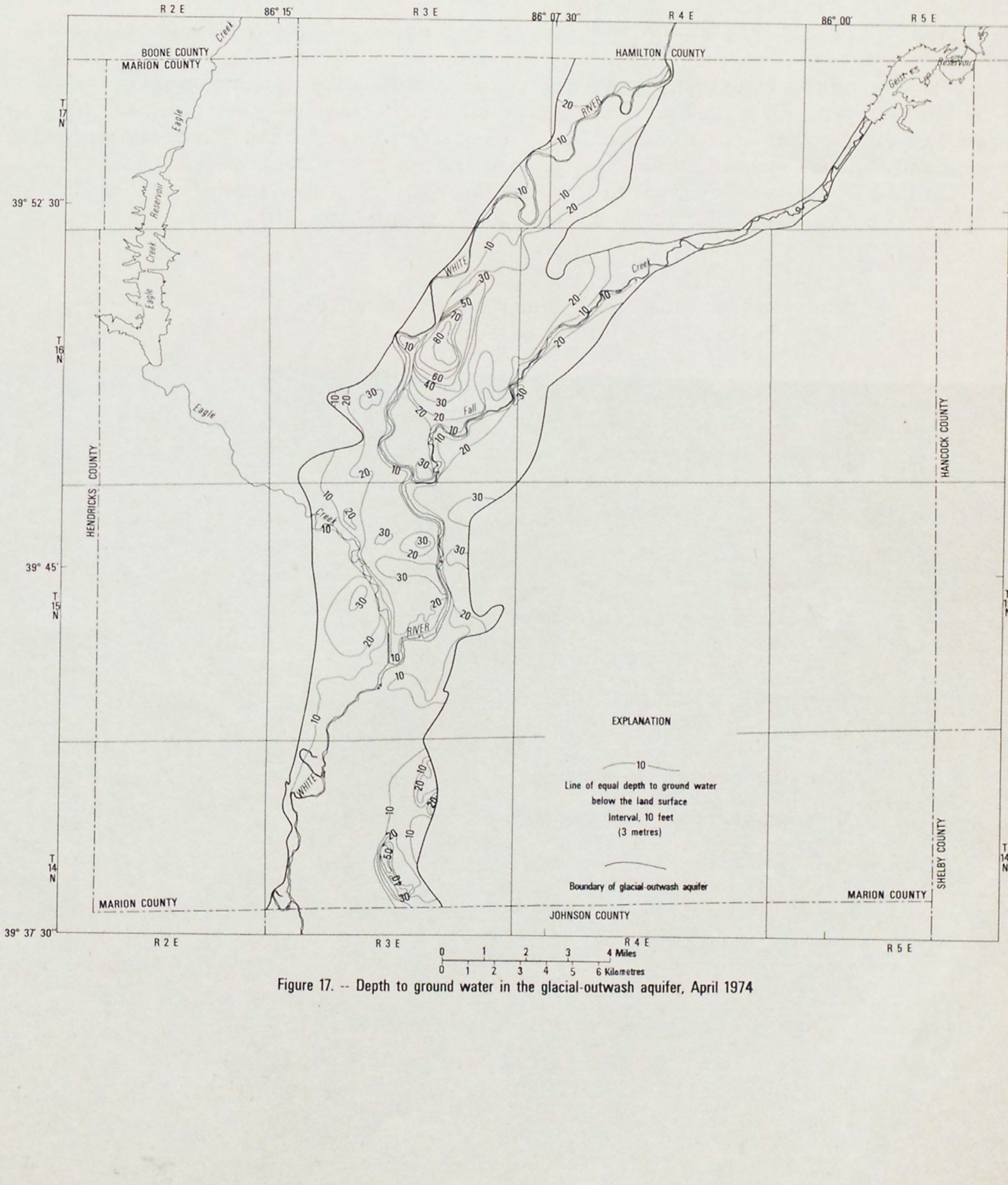

Figure 17. -- Depth to ground water in the glacial-outwash aquifer, April 1974

THE ELECTRIC-ANALOG MODEL

The electric-analog model of the Indianapolis area simulates the regional movement of ground water in Marion County through the various aquifers of the glacial drift and the underlying limestone. The model was developed to evaluate the effects of large-scale ground-water withdrawals from the aquifers. The use of electric analog techniques for the solution of hydrologic problems has been well documented (Skibitzke, 1961, Walton and Prickett, 1963, and Patten, 1965) and is not recounted here. In the interest of completeness, however, a brief description of the analogy is included, particularly as it applies to the model of the Indianapolis area.

The analogy between the movement of ground water in an aquifer and the flow of current in an electrical conductor is based on the similarity between the partial differential equation describing two-dimensional, non-steady ground-water flow in a homogeneous, isotropic, confined aquifer (equation 8a) and an equivalent equation for a two-dimensional diffusion field in electricity (Karplus, 1958, p. 33) (equation 8b).

$$\nabla^2 h = \frac{S}{T} \frac{\partial h}{\partial t} \tag{8a}$$

$$\nabla^2 V = RC \frac{\partial V}{\partial t} \tag{8b}$$

where:

$$\nabla^2 = \left(\frac{\partial^2}{\partial \chi^2} + \frac{\partial^2}{\partial y^2}\right)$$

S = storage coefficient of the aquifer

T = transmissivity of the aquifer

h = head of water

t = time

R = electrical resistance

C = electrical capacitance

V = voltage

The similarity of the two equations is readily apparent in the one-to-one correspondence between the respective elements. In practice, the finite difference forms of equations 8a and 8b are used. A grid, generally square, is superimposed on the aquifer. The head of any given internal intersection (node) of the grid is then related to heads at the surrounding nodes by the finite difference expression (Stallman, 1956):

$$T\left(\sum_{i=2}^{5} h_i - 4h_1\right) = a^2 S \frac{\partial h}{\partial t} \qquad (9)$$

where h_1 represents the head at a given node

$h_2 .. h_5$ represent the heads at the four surrounding nodes

a is the grid spacing

and the remaining terms are as previously defined

If resistors and capacitors are joined electrically in a square mesh array, the voltage at any given internal node is related to the voltage at the surrounding nodes by the relationship (Skibitzke, 1961):

$$\frac{1}{R}\left(\sum_{i=2}^{5} V_i - 4\,V_1\right) = C\,\frac{\partial V}{\partial t} \qquad (10)$$

where V_1 represents the voltage at a given node

$V_2 ... V_5$ represent the voltages at the four surrounding nodes

and the remaining terms are as previously defined

Thus, an array of resistors and capacitors can be utilized to obtain approximate solutions to ground-water flow problems. The analogy between the two systems is formalized by the use of appropriate scale factors: time in the aquifer, t_w (days), is related to time in the model t_e (seconds); the potential for ground-water flow in the aquifer, h (head), is related to the potential for flow of electricity in the model, V (volts); the rate of flow of water in the aquifer, q_w (gallons per day); is related to the rate of flow of electricity in the model, q_e (amperes); and mass in the aquifer, Q_w (gallons), is related to energy, Q_e (coulombs), in the model.

These terms can be related such that, for the model used in this study:

$$\frac{h}{V} = k_2 = 9 \ \frac{\text{feet}}{\text{volt}}$$

$$\frac{q_w}{q_e} = k_3 = 3 \times 10^9 \ \frac{\text{gallons per day}}{\text{ampere}}$$

$$\frac{t_w}{t_e} = k_4 = 3.65 \times 10^5 \ \frac{\text{days}}{\text{seconds}}$$

$$\frac{Q_w}{Q_e} = k_1 = k_3 \ \ k_4$$

Equations 9 and 10 show that resistance in the model is inversely proportional to aquifer transmissivity, and that capacitance is directly related to the aquifer storage coefficient.

Model Construction

The transmissivities and storage coefficients of the upper, middle, and lower confined aquifers, and the glacial-outwash aquifer were modeled as three separate resistor-capacitor (R-C) networks or layers. The limestone aquifer was modeled as a fourth layer. The nodal spacing between resistors representing the confined aquifers is 1 in (25.4 mm) equal to 2,640 ft (805 m) in the aquifer. The limestone aquifer was modeled at a nodal spacing of 1 in (25.4 mm) on the model equal to 5,280 ft (1,610 m) in the aquifer. The glacial-outwash aquifer was modeled with a node spacing of 0.5 in (12.7 mm) equal to 1,320 ft (402 m). Vertical flow of water between the aquifers was modeled by connecting a resistor between corresponding nodes of the layers. The value of these resistors, R_L, was inversely proportional to the vertical hydraulic conductivity, K', of the silt and clay, and directly proportional to the clay thickness, m':

$$R_L = \frac{k_3}{k_2 (K'/m') a^2} \tag{11}$$

In areas where these aquifers were not separated by clay, the resistance between layers was also determined by equation 11, but in that case the quantity (K'/m') referred to properties of the aquifers rather than of the confining beds. In that situation K' is the average vertical hydraulic conductivity of the sands, and m' is the average flow line distance between the modeled aquifers. These relationships are shown in figure 18, where typical model nodes are superimposed on the idealized regional aquifer system.

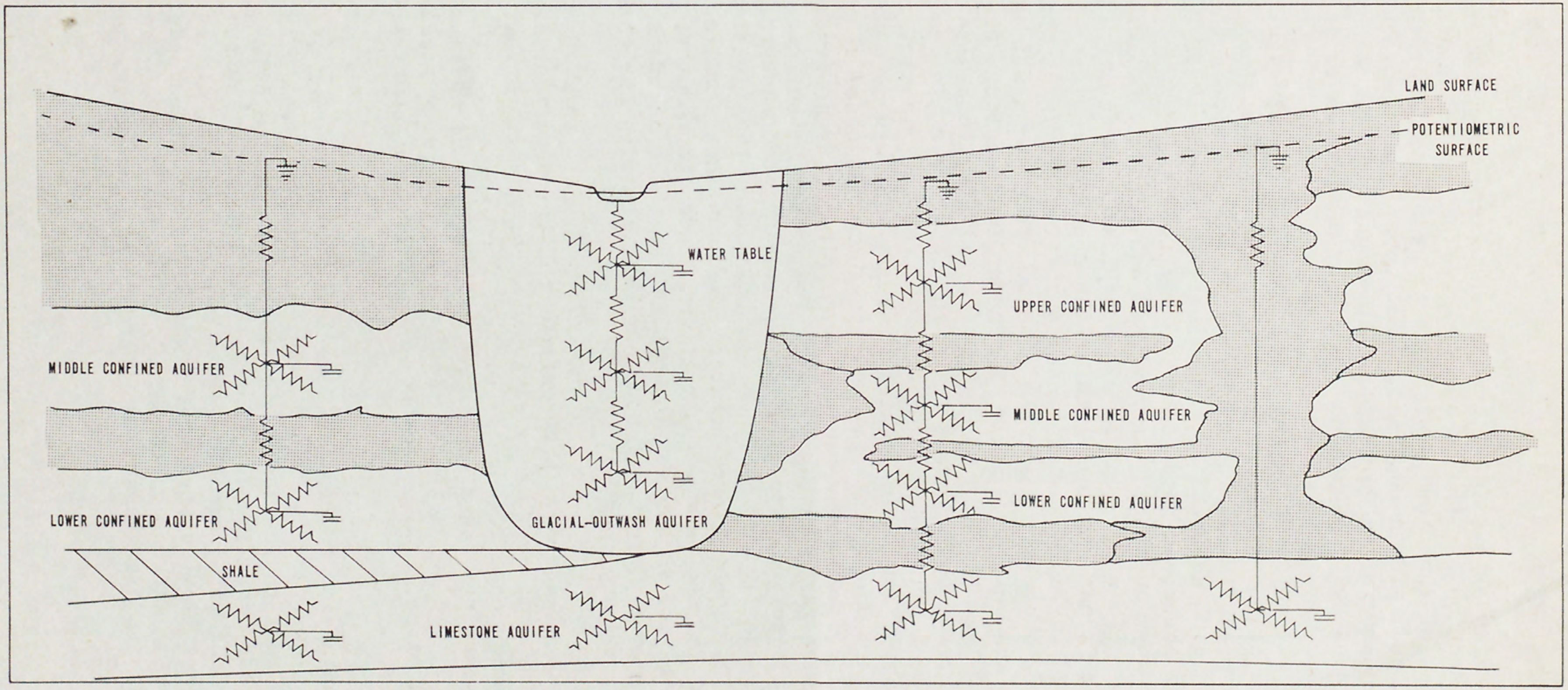

Figure 18.-- Idealized west to east cross section showing typical nodal configurations on analog model

Each series of resistors on figure 18 shows the nodal connections of a typical vertical section of the flow system. The limestone aquifer is the lowermost component of the flow system and is overlain, in ascending order, by the lower confined, middle confined, and upper confined aquifers. The glacial-outwash aquifer also overlies the limestone aquifer, and is shown in this cross section as being separated from it by a shale layer (bedrock).

The R-C network representing the transmissivity and storage coefficient of the limestone aquifer is shown as the lowermost electrical components. Above the limestone aquifer in the confined part of the system there may be as many as three R-C networks representing the transmissivity and storage coefficient of the three confined aquifers or the glacial-outwash aquifer. Locally, however, any or all of the confined aquifers may be absent because of a concentration of clay lenses. The right side of figure 18 shows two possible extremes; in one extreme, each of the confined aquifers is present, and is represented by an R-C network and a vertical resistor that connects it to the aquifer above or below; the other extreme shows all the confined aquifers missing, and only a single resistor representing the aggregate thickness of the clay. The values of the vertical resistors were determined from equation 11, and the uppermost vertical resistor is connected to electrical ground. The ground connection simulates the initial position of the top of the zone of saturation (water table) and was used only for the nonsteady analysis. During the steady-state analysis, these resistors were not grounded, but were connected to electrical current generators that provided effective recharge to the system. The nodes to the left of figure 18 show that the upper confined aquifer is missing, and that there is no connection between the lower confined aquifer and the limestone in this area. The lack of connections between the confined aquifer and the limestone is based on the very low vertical hydraulic conductivity of the shale. The glacial-outwash aquifer is modeled as three layers that, except where clays intervene, are continuous horizontally with the corresponding layers of the confined aquifers.

The glacial-outwash aquifer is unconfined and the uppermost layer of the network representing this aquifer is equipped with capacitors that represent a water table storage coefficient of 0.11. The nodes of the confined aquifers have capacitors based on a storage coefficient determined from equation 4, as do the nodes in the lower two layers representing the glacial-outwash aquifer. The storage coefficient of the limestone is modeled as 10^{-4}.

In summary, the analog model is a four-layer approximation of the actual three-dimensional flow system that underlies Marion County. It is a comprehensive synthesis of the available geologic and hydrologic data, and represents the most detailed description of the flow system that is consistent with hydrologic judgment. It is not, however, either a literal description of the flow system nor necessarily a correct description; accordingly, measures were taken to establish the validity of the model and to demonstrate that it is capable of reproducing known characteristics of the flow system.

Model Verification

Verification of the analog model was based on an analysis of the flow system as it was documented by water levels measured during April and May 1974. At that time, steady or near steady conditions prevailed in the aquifer system, and discharge to the streams and to the existing industrial wells was approximately balanced by recharge from rainfall. The objectives of the steady-state model analysis were: (1) to duplicate the three-dimensional head distribution observed in the flow system; (2) to synthesize the rates of effective recharge from rainfall; and (3) to synthesize the hydraulic connection between the major streams and the glacial-outwash aquifer. The steady-state analysis was a trial- and-error procedure where effective recharge and the hydraulic connection of the streams to the aquifer were systematically changed until the head distribution in the aquifers and the rate of ground-water seepage to the streams were judged satisfactory.

The effective recharge to the model is defined as the rate of flow into either the upper layer of the glacial-outwash aquifer or the upper confined aquifer. It is significantly less than the total recharge to the ground-water system because it does not account for shallow circulation to springs and upland streams in the silts and clays that overlie the confined aquifers. The effective recharge is only that water that enters the regional flow system, and which is ultimately drained directly by the major streams, discharged by wells, or leaves the basin as underflow. Recoverable evapotranspiration from this sytem is considered negligible.

Only the principal streams of the system were modeled. The omission of numerous smaller streams entering the glacial-outwash aquifer may result in a small error in the total quantity of ground-water seepage to streams measured on the model. It is felt, however, that this error is negligible owing to the intermittent flow of these streams and the fact that they penetrate only the underlying clay.

With the exception of the northern and southern boundaries of the glacial-outwash aquifer, and of certain isolated intervals around the periphery of the till plain, all external boundaries of the flow system were modeled as impermeable. The northern boundary of the glacial-outwash aquifer corresponds to an equipotential with an elevation of 745 ft (227 m), and provides to the model the equivalent of underflow from the alluvium north of the county line. The southern boundary of the glacial-outwash aquifer corresponds to an equipotential with an elevation of 650 ft (198 m), and allows underflow from the modeled area to be measured.

Effective recharge to the model was simulated by electrical currents that were imposed at each node of the upper confined aquifer, and at every other node of the glacial-outwash aquifer. The input currents were arbitrarily adjusted in a trial-and-error fashion until the head distribution in the model approximated the known heads in the aquifer. Where the approximation was not satisfactory, or where the quantity of ground-water seepage to the rivers was judged to be in error, the model was modified and the trial-and-error procedure of adjusting the input currents was repeated. During this analysis, it was necessary also to modify the flow rates between aquifers by changing the value of some of the vertical connecting resistors.

The hydraulic connection between the stream and the glacial-outwash aquifer was adjusted concurrently with the manipulation of effective recharge. The iterative progression of these adjustments "relaxed" the model to a final acceptable solution based on the synthesized values of these two variables. The criteria used for determining the hydrologic connection were the total probable seepage gain by the streams and the regional pattern of the water table in the outwash aquifer. The previous estimates of total seepage gain of the streams averaged 42 ft^3/d per foot of stream (3.9 m^3/d per metre) or about 50 to 70 Mgal/d (2.2 to 3.1 m^3/s) of total direct seepage to the major stream channels within the county.

Seepage to the rivers was simulated using resistors at each node along the course of the streams. The value of each resistor was systematically adjusted until the ground-water heads at that point were duplicated. The resulting current flow through the resistor was then a measure of ground-water seepage into the stream in that node area. Table 3 shows the quantitites of ground water flowing into selected stream reaches, as computed by the model, and compares those data with seepage figures (table 1) determined by uniform application of the 42 ft^3/d per foot of stream (3.9 m^3/d per metre) seepage rate over each selected reach. Figure 19 shows the locations of the stream reaches utilized in this analysis. The similarity between the two sets of figures is remarkable, with only Fall Creek showing a discrepancy of about 35 percent less seepage than the model value. Considering that the seepage measurement on which the computed value was based was made on the White River, it seems proper to attribute this difference to the hydraulic characteristics of Fall Creek and to its significantly narrower channel width.

Table 3.--Comparison of model-derived seepage data with computed seepages (in million gallons per day)

	Fall Creek	Eagle Creek	White River: Reaches 1	2	3	4	5
Model	14.9	1.5	5.4	11	11	2.2	12.8
Computed	23	*	5	12	12	*	10

Streams and measurement reaches as shown on figure 19.
*No computation made because of effects of Indianapolis pumping.

During the course of this analysis it became apparent that a simultaneous solution for the appropriate ground-water heads and the total seepage to the streams could not be achieved with the original transmissivity distribution. Accordingly, all transmissivities on the model

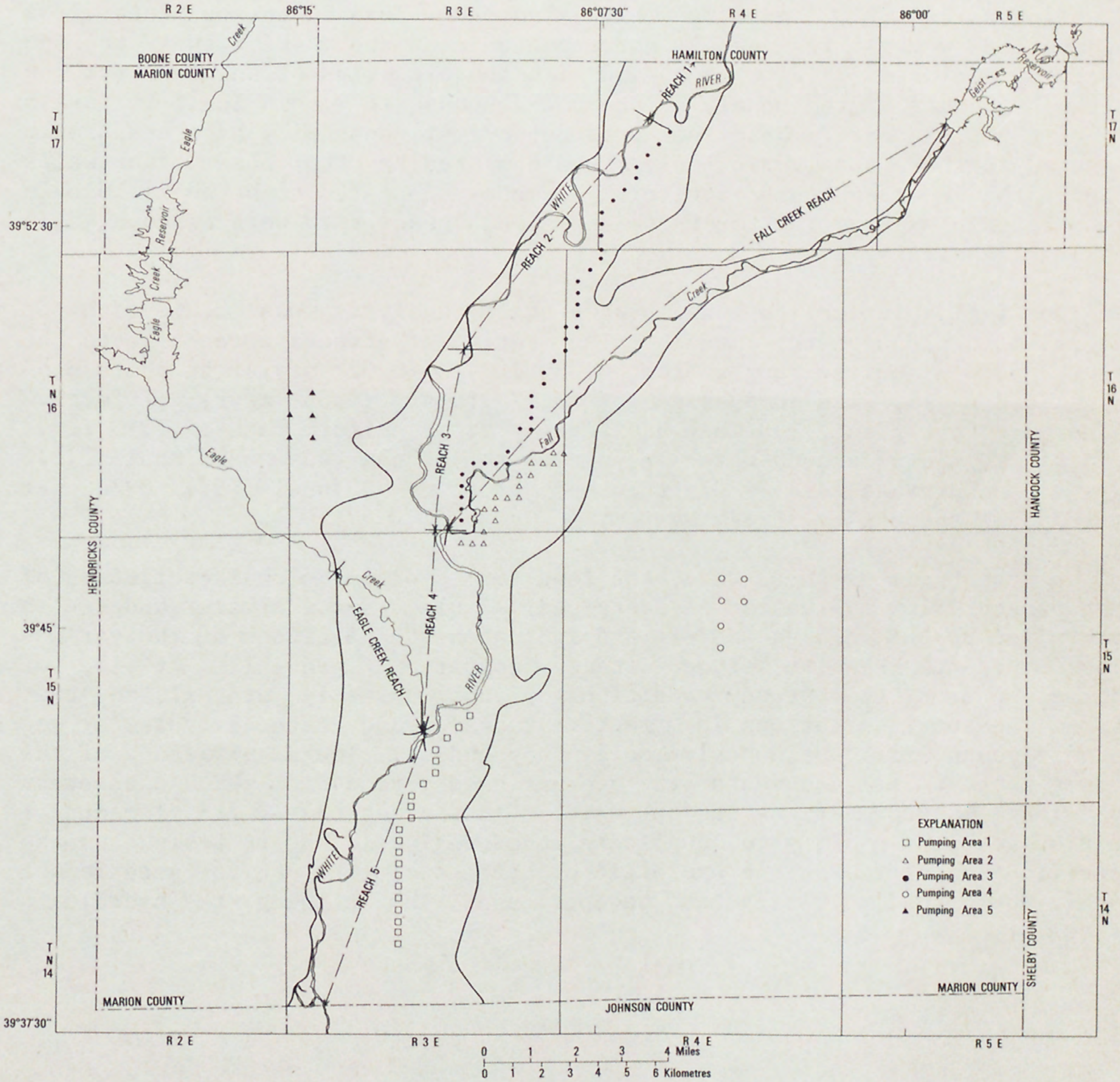

Figure 19.-- Location of areas of simulated pumpage

were increased by a uniform 11 percent, and in addition, the transmissivities of two areas in the glacial outwash were increased by 100 percent. In the latter case, the modeled transmissivity was increased from 50,000 to 100,000 gal/d/ft (620 to 1,240 m^2/d), implying that the alluvial material was coarser than had been inferred from interpolation between test holes. Six additional test holes were subsequently drilled in the area and the presence of an appreciable thickness of sand and gravel substantiated the inference drawn from the model. The final transmissivity distribution used in the model is shown in figures 10-13.

The computed head distribution in the ground-water flow system is shown in figures 20, 21, and 22, which also show head data obtained from observation wells. In general, correspondence of the model results to the observed head data is quite good, and with the exception of an area near the northwest border of the model, the computed heads are within 10-15 ft (3-4.5 m) of the field data. Near the northwest border, unusually high heads were measured in three observation wells when compared to other observation wells tapping the lower confined aquifer. As there appeared to be no lithologic or hydrologic explanation for these anomalous heads they were discounted in the model analysis.

The final solution to the steady-state analysis was achieved after several iterations on the model. The rates of effective recharge to the upper confined aquifer ranged from 0.08 ft/yr (0.02 m/yr) to about 0.67 ft/yr (0.2 m/yr), with an average of 0.17 ft/yr (0.05 m/yr). Effective recharge to the glacial-outwash aquifer was at a uniform rate of 1.13 ft/yr (.34 m/yr). Total ground-water seepage to the stream was 58.8 Mgal/d (2.6 m^3/s) or an average rate of 37 ft^3/d per foot of channel (3.4 m^3/d per metre) over the entire stream system.

On the basis of these data, the long term hydrologic characteristics of the present flow system can be described. Table 4 shows a water budget as determined by the analog model, and indicates the magnitude of the various flow components when the assumed steady conditions prevail. It is, of course, apparent that steady conditions do not normally prevail, and that natural seasonal variations in effective recharge and natural cycles of wet and dry years will continuously modify this budget. The significance of the budget is that it documents an average base period from which consistent hydrologic interpretations can be made. In the aggregate, the elements of this budget, the ability to duplicate the regional water table, and the specification of reasonable hydraulic properties of the aquifer used in the model, would tend to verify the concepts used in defining the hydrologic system.

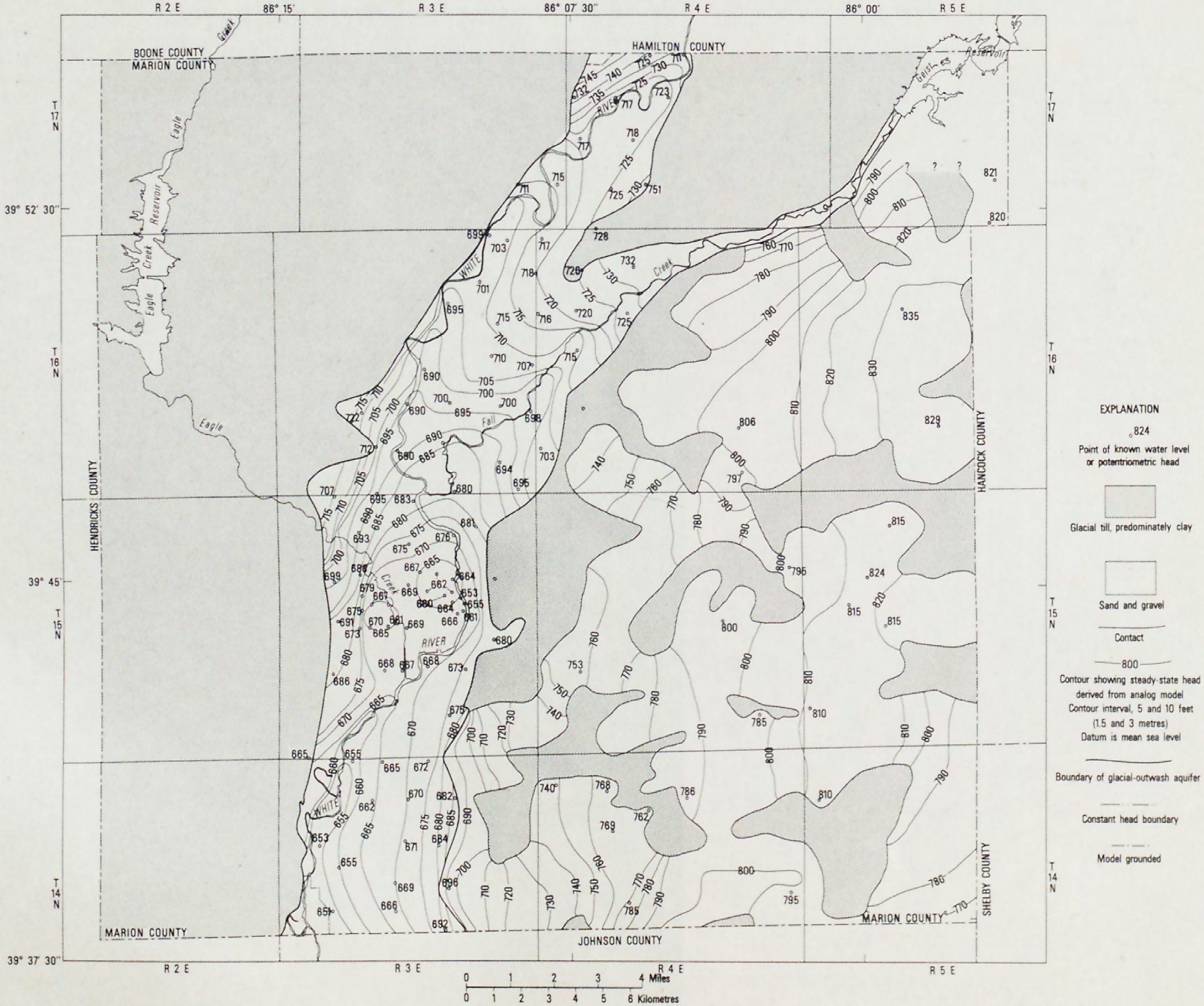

Figure 20. -- Comparison of the final steady-state heads derived from the electric-analog model with known water levels in the glacial-outwash aquifer and with known potentiometric heads in the upper confined aquifer

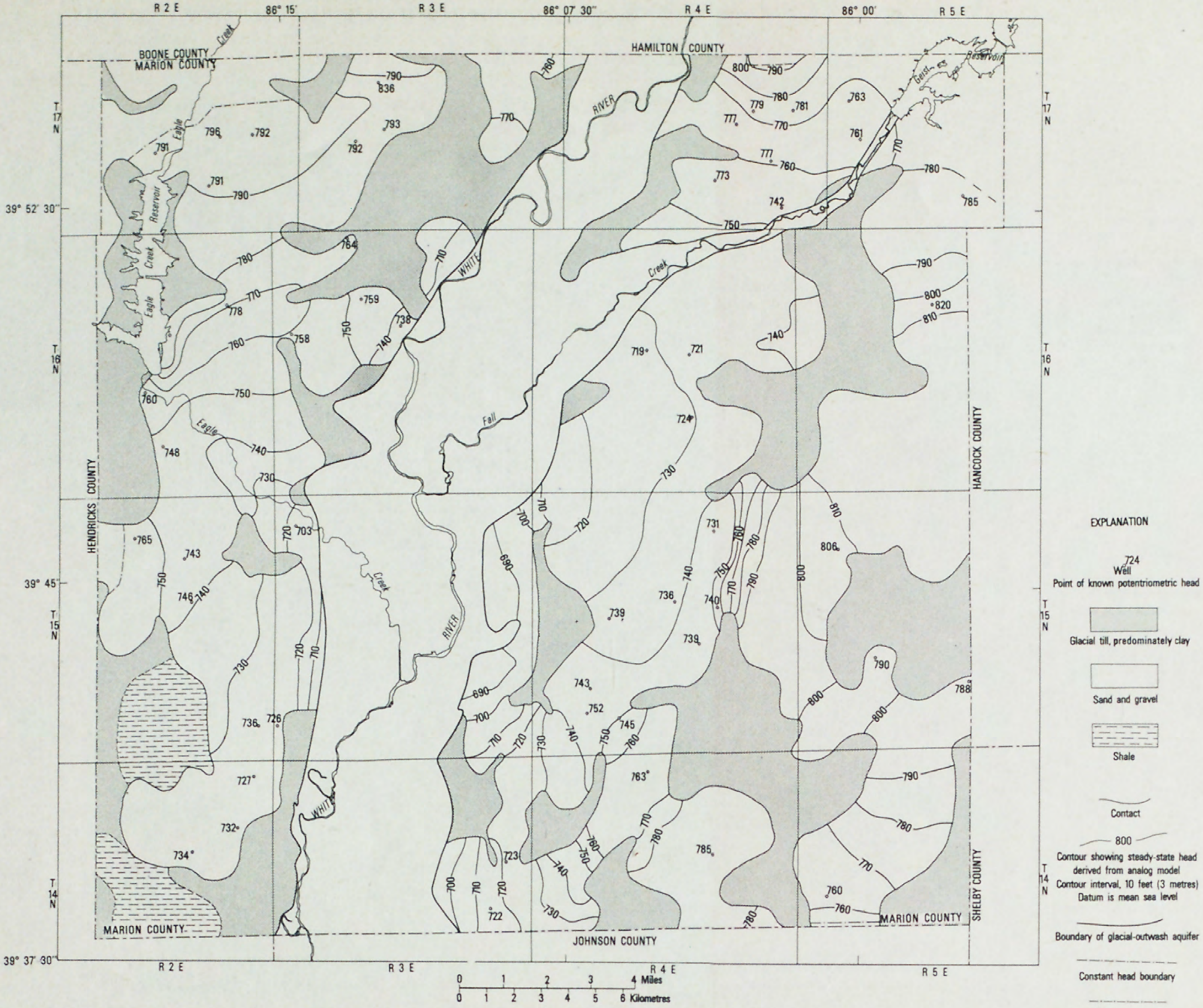

Figure 21. -- Comparison of the final steady-state heads derived from the electric-analog model with known potentiometric heads in the middle confined aquifer

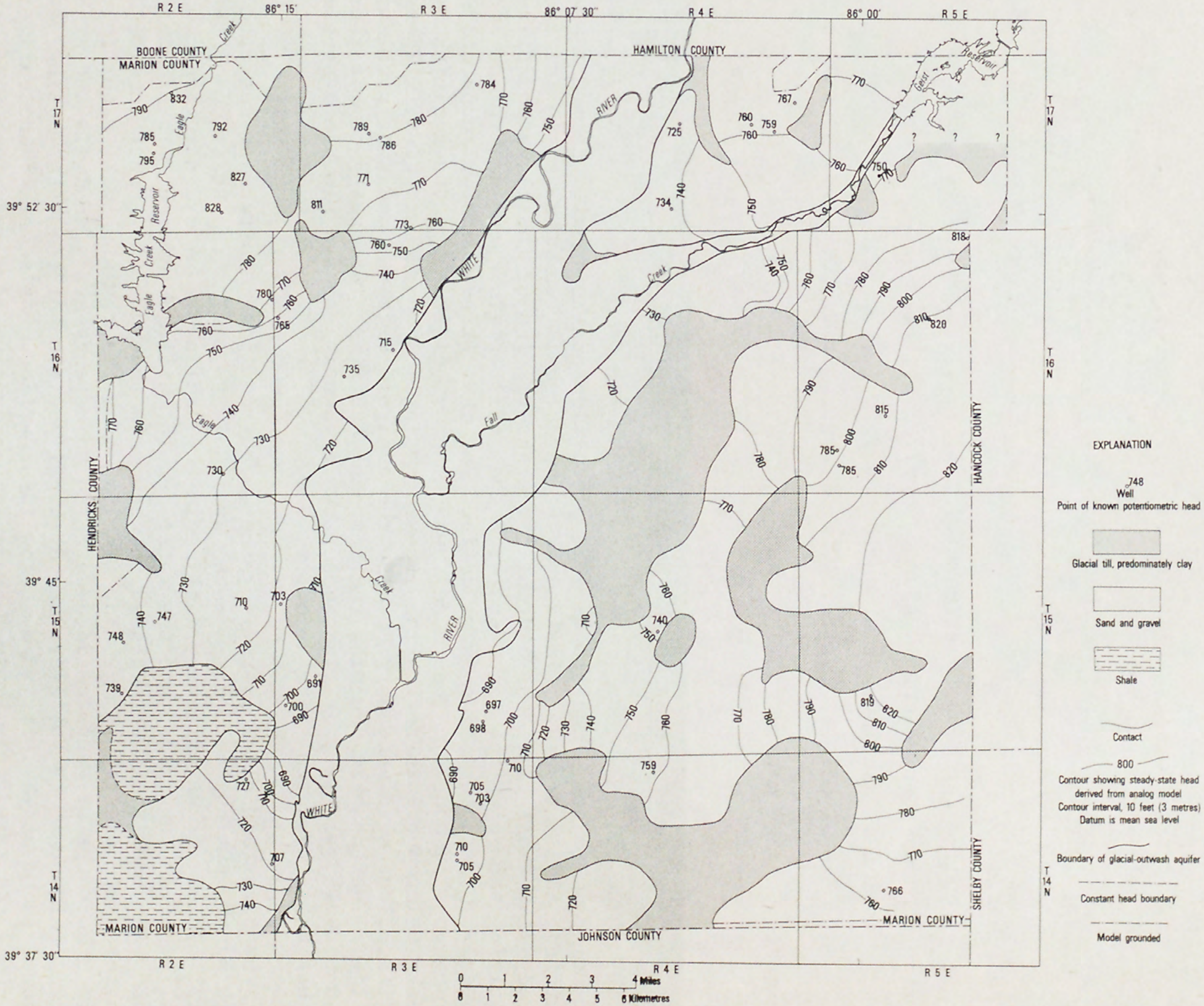

Figure 22. -- Comparison of the final steady-state heads derived from the electric-analog model with known potentiometric heads in the lower confined aquifer

Table 4.--Marion County ground-water budget as determined from the electric-analog model (average daily quantities in million gallons per day)

I. Recharge		
A) Effective recharge		
1) Glacial-outwash aquifer	48.4	
2) Upper confined aquifer	32.5	
B) Ground-water underflow into county	20.0	
Total		100.9
II. Discharge		
A) Ground-water flow to streams	58.2	
B) Ground-water underflow from county	8.2	
C) Municipal and industrial pumpage*	32.6	
Total		99.0

*Domestic pumpage not included in model

Modification to the Model for Transient Analyses

Prior to imposing additional well production on the model, the model was modified for analyzing time-dependent problems. On the basis of the steady-state analysis, the hydraulic characteristics of the clays and silts overlying the combined aquifers in the till plain were determined and hydraulic characteristics were defined for the stream-aquifer connection throughout the model. The model was then modified by the addition of electrical components reflecting these hydraulic characteristics. These modifications were necessary because the analysis of time-dependent problems yield solutions that describe the changes occurring in the system, rather than actual values; for instance, the change in seepage flow to a stream rather than the actual total flow.

The clays and silts overlying the upper confined aquifer are saturated with water, but no field data were available to define the position of the water table. If water levels decline in the confined aquifer, downward flow will occur in the clays, and the water table will be lowered. To simulate such flow, resistors were placed between each node of the upper confined aquifer and electrical ground. The value of the resistor was determined from the rates of recharge documented in the steady-state analysis and the known thickness of the clay; the original position of the water table was assumed to be about 10 ft (3 m) below land surface. As noted in an earlier section, the resultant vertical hydraulic conductivities were obtained by applying equation 7 to these data. The vertical hydraulic conductivities resulting from this analysis ranged from 2×10^{-4} to 1.5×10^{-2} ft/d (6×10^{-5} to 4.6×10^{-3} m/d), in generally good agreement with the initial value of 7×10^{-4} ft/d (2×10^{-4}m/d) assigned to the lower semiconfining layers.

This technique of simulation in which the water table is represented by grounded connections, is inexact in that it does not explicitly simulate withdrawal from water-table storage in the clays. It is an acceptable approximation, however, when the rates of flow are reasonably low. As will be described subsequently, one model run was deemed unacceptable because vertical flow rates through these resistors were high, and indicated that the hypothetical water table in the clays would decline at a rate between 2 and 3 ft/yr (0.6 and 0.9 m/yr). While such flow rates are possible, the described model simulation is inadequate, and electrical capacitors would have to be added to the vertical resistors for an appropriate simulation.

As discussed previously, the resistors representing the hydraulic connection of the stream to the glacial-outwash aquifer were adjusted until the model heads corresponded to field measurements; the resultant electrical current flowing through each resistor was proportional to the ground-water seepage into the stream reach represented by the model node. For the transient analysis it was necessary to replace those resistors with a new set whose values represent the actual hydraulic characteristics of the stream-aquifer connection.

A lumped characteristic satisfying this requirement can be defined simply as:

$$R = \frac{E}{I} \propto \frac{h_a - h_s}{Q_s} \tag{12}$$

where h_a is the head in the aquifer, h_s is the head of the stream, q_s is the flow determined from the steady-state analysis, the appropriate scale factors are implied by the proportionality sign. The equation relating resistance to flow between a stream and an aquifer can be written as:

$$R \propto \frac{\overline{\Delta z}}{K_z(WL)} \tag{13}$$

where $\overline{\Delta z}$ is the length of the average flow path, K_z is the average vertical hydraulic conductivity over that flow path, W is the width of the stream, and L is the length of the stream represented by the node. Combining (12) and (13) and solving for K_z,

$$K_z = \frac{\overline{\Delta z}\, q_s}{(h_a - h_s)(WL)} \tag{14}$$

where all quantities on the right side of the equation are known with the exception of $\overline{\Delta z}$, which is assumed to be approximately 5 ft (1.5 m).

Equation 14 allows an approximate solution for K_z, with only a minimal uncertainty in specifying $\overline{\Delta z}$. The values of K_z thus determined ranged from 0.03 to 67 ft/d (0.01 to 20 m/d), with more than 80 percent of the values falling in the range of 0.03 to 40 ft/d (12 m/d). This range compares with the 30 ft/d (9 m/d) vertical hydraulic conductivity of the alluvium determined by the cross-sectional model and as modified by the present study. In general, this range can be regarded as the minimum vertical hydraulic conductivity consistent with the regional water levels and effective recharge rates. As will be shown in the subsequent transient analyses, large increases in these conductivities do not greatly affect the response of the hydrologic system to increased pumping.

MODEL PREDICTIONS

The objectives of this study were to determine the quantity of ground water that could be pumped in the county and to estimate the effects of that pumpage on the ground-water system and streamflow. The following analyses address that problem, and develop a range of probable maximum drafts on the ground-water system that could be sustained indefinitely under present hydrologic conditions. The hypothetical well fields used in these analyses were "designed" so as to conform to areas of maximum transmissivity, maximum saturated thickness, and proximity to the streams.

Individual wells on the model were assumed to be fully penetrating and to be of sufficient diameter and construction that well losses would be minimal. The density of wells on the model was such that individual discharges ranged from 200 to 3,000 gal/min (13 to 190 l/s) and no consideration was given either to the economics of wells with small discharge, or to the hydraulic problems attendant to wells with large yields. In practice, pumpage from large capacity model wells could be obtained from several smaller wells distributed over the area represented by the node; however, in the analyses that follow, it was assumed that the discharge from each node would occur from a single well located at the center of the node. It is explicitly stated that while the well fields used in the following model runs are located at apparently productive sites, no attempt was made to optimize well yields, either through relocation of wells or by varying the density of wells.

The predictive analyses were carried out in a series of experiments utilizing the analog model. In four of these experiments, three extensive well fields were simulated in the glacial-outwash aquifer and a fourth well field was simulated in the middle confined aquifer east of the White River. Pumpage from a fifth well field, tapping the lower confined aquifer west of the river, was simulated in a separate experiment. Figure 19 shows the location of the simulated wells, and their grouping into the five well fields.

Two additional experiments were carried out. In one of these, well field 1 in the glacial outwash was pumped at a high rate for 30 days; in the second, this well field was pumped at a somewhat lower rate for 90 days. These experiments were intended to illustrate the consequence of high pumpage rates over short periods. Before presenting the experimental results, brief discussions will be given of the factors limiting ground-water development and of the corrections applied to the analog data.

Factors Limiting Ground-Water Development

Any discussion of pumpage must logically be phrased in terms of its consequences, and of the constraints and limiting factors that govern the system. The factors limiting ground-water development, from a purely hydraulic point of view, are drawdown of water levels and availability of recharge from the streams. These two factors are interrelated, and both come into play in the evaluation of any pumping scheme.

In an unconfined aquifer such as the glacial outwash, the maximum discharge from an individual well theoretically occurs when the water level in the well is drawn down to the base of the aquifer; however, it is normally impractical to operate a well at full drawdown. Applying the Depuit assumptions in an analysis of flow to an isolated well in an unconfined aquifer, it is easily shown that about 90 percent of the theoretical maximum well discharge is obtained when the drawdown in the well is two-thirds of the original saturated thickness of the aquifer. Thus it is reasonable to set two-thirds of the saturated thickness as a drawdown limit; most of the available discharge should be obtained at this drawdown, but at the same time water levels in the discharging wells need to be high enough to avoid chronic operational problems.

Other considerations may arise, however, in connection with the establishment of drawdown criteria; water levels elsewhere in the aquifer than at the discharging wells may be critical. The analog experiments simulate only the proposed additional pumping, not the pumpage already occurring in the aquifer. The combined effect of existing and proposed pumpage is obtained by superposing the two individual effects, and is an analysis extraneous to the analog work. The proposed new pumpage will cause drawdowns throughout the aquifer, and will thus affect water levels in existing wells. If the effect is too severe, operational problems in these existing wells will result. Thus it is reasonable to establish criteria regarding drawdowns, as indicated by the analog, at points occupied by existing wells.

Recharge from the streams acts in conjunction with water-level drawdown as a constraint on available pumpage. During the initial phases of any expansion in pumping, the increase in ground-water withdrawal is supplied from aquifer storage, which implies water-level drawdown. As drawdown continues, the natural discharge of ground water to streams is reduced, and may be reversed, so that seepage from the stream to the aquifer takes place. The reduction in natural outflow and(or) the induced seepage increase progressively as drawdown continues. Eventually, they may together balance the increase in withdrawal; when this occurs the system has attained a new equilibrium, and further drawdown of water level ceases. If this occurs before drawdown constraints have been violated, the increase in pumpage could presumably be sustained indefinitely without undesirable effects.

However, the ability of streams to generate a new hydrologic equilibrium is limited. The reduction in natural discharge to a stream cannot exceed the original value of the natural discharge, and there is an upper limit to the seepage which can be induced from a stream. This upper limit is reached when the regional water table falls well below the streambed, so that the connection between the stream and aquifer consists of a narrow saturated band beneath the stream, in which flow is virtually vertical and occurs under unit gradient of head. The flow which can be sustained under these conditions depends upon the vertical hydraulic conductivity of the streambed, and of the aquifer itself. The point to be stressed, however, is that this flow is a maximum; further lowering of the water table produces no increase in the hydraulic gradient, and hence, no increase in seepage. An additional restriction on seepage from a stream is sometimes imposed by the flow characteristics of the stream. If, for any appreciable percentage of the time, the flow of the stream falls below the maximum seepage which streambed and aquifer properties can sustain, then the flow of the stream, rather than the hydraulic conductivity distribution in the earth materials, dictates the maximum possible seepage rate.

If the original natural discharge to a stream, taken together with the maximum seepage that can be induced from it, are less than a proposed increase in pumpage, then a new equilibrium cannot be achieved as a result of stream seepage during pumping at the proposed increased rate. If other sources of recharge are ruled out, sustained pumping at the proposed rate will produce sustained withdrawal from aquifer storage, which in turn implies continuing drawdown, ultimately to levels which would violate whatever drawdown constraints have been established.

Correction of Analog Data

To evaluate analog model results in terms of constraints on the water levels in the discharging wells themselves, two types of correction must first be applied. For the glacial-outwash aquifer, drawdowns as given by the analog model actually represent changes in water level that would be observed in a confined aquifer equal in thickness to the original saturated thickness of the outwash. The simulation does not account for the effects of dewatering of the aquifer. A method suggested by Jacob (1963) was utilized to correct the model drawdowns for the dewatering effect. The drawdown as indicated by the model at a given node, moreover, is a regional or average value for the block of aquifer represented by that node--it does not reflect the actual water level within a discharging well. A method suggested by Prickett (1967) was utilized to calculate the drawdown within discharging wells, based on the model results.

In Jacob's method of correction for the dewatering effect, the model drawdown, s', is related to the actual drawdown in the water-table aquifer, s, by the relation:

$$s' = s - s^2/2m \tag{15}$$

Regional drawdown values--that is, drawdowns elsewhere than in the discharging wells--were calculated by solving this equation for s at representative points throughout the aquifer. It was found that where the model draw- down, s', was 15 ft (4.5 m) or less, variation of the saturated thickness, m, within the limits shown in figure 8, has no appreciable effect on the correction; a constant value of m was used, therefore, in the equation for values of s' less than or equal to 15 ft (4.5 m). For higher values of s', the individual value of m as indicated in figure 8 was used in calculating s. Owing to inherent errors in a finite difference approximation of the continuous flow medium, well drawdown at a pumping node is always underestimated on a model.

In calculating drawdowns within discharging wells, the correction described by Prickett (1967) was first applied to the model drawdown, and Jacob's correction was subsequently applied to the result. In Prickett's correction the head loss in the radial cone of depression around an individual well is computed and added to the regional drawdown indicated by the model. Again using s' to represent the model drawdown, the drawdown in a well, s_ω, is given by Prickett's correction:

$$s_\omega = s' + 0.3665 \frac{Q}{T} \log \left(\frac{a}{4.81\ r_w}\right) \tag{16}$$

where Q is the well discharge, T is aquifer transmissivity, a is the analog mesh spacing and r_w is the well radius. The technique assumes that the discharge, Q, occurs through a single well located at the center of the aquifer block represented by the node. Prickett's method was developed for artesian conditions, and the well drawdown obtained in the calculation actually represents the drawdown that would be observed in a well tapping a confined aquifer of thickness equal to the original saturated thickness of the glacial outwash. Jacob's correction therefore, was applied to the result to obtain an estimate of the actual well drawdown in the unconfined outwash.

Description of Experiments and Results

In four of the analog experiments the simulation was continued until a new equilibrium was established. These four experiments are designated A, B, C and D for purposes of discussion. In experiments A and B the discharges from wells in the glacial-outwash aquifer were adjusted by trial and error to values that would produce drawdowns within the wells, after attainment of a new equilibrium, of approximately one-half the original saturated thickness of the aquifer. This criterion was chosen arbitrarily to achieve a conservative result. In experiments C and D, discharges from wells in the glacial outwash were adjusted to values that would produce drawdowns within the wells, again at a new equilibrium, of approximately two-thirds the original saturated thickness of the aquifer. This drawdown, as noted previously, should provide close to the maximum practical discharge from individual wells. Because of operational difficulties and time limitations, the drawdown criteria were not achieved exactly in all wells on every run, but rather were satisfied as nearly as possible in an average sense.

Two experiments were carried out for each drawdown criterion because of uncertainty regarding the efficiency of the stream-aquifer connection. In experiments A and C, the streams were modeled as separated from the unconfined aquifer by streambed materials ranging in hydraulic conductivity from 0.03 to 67 ft/d (0.01 to 20 m/d), as determined in the steady-state analysis. In experiments B and D, the streams were simulated as fully penetrating the upper one-third of the glacial outwash, with no intervening streambed material. It was felt that these two methods of simulation would bracket the actual field situation.

In each of these four experiments pumpage was also simulated from well field 4 (fig. 19) tapping the middle confined aquifer. The drawdown criterion for the confined aquifer was not applied within the discharging wells themselves, but rather over the surrounding region. The regional drawdown, as indicated by the uncorrected analog result for the node from which pumping occurred, was limited to 20 ft (6.1 m). This limit was set because greater drawdowns could interfere with the operation of domestic wells by lowering water levels below the pump intakes. As in the glacial-outwash aquifer, the drawdown criterion was applied to the results following attainment of a new equilibrium.

The pumpage simulated in these experiments actually represents increased pumpage above present levels, and the drawdowns represent water-level changes from present levels. Seepage rates from the streams similarly represent changes from present values. In each of the four experiments, the new equilibrium was achieved within 6 years after the start of pumping.

Table 5 summarizes the pumpages measured in the four experiments. The first column gives the total pumpage, and the succeeding columns give the pumpage breakdown according to the well fields. Table 6 summarizes the seepages, at equilibrium, from the various stream reaches (fig. 19) in each of the experiments, and, for purposes of comparison, the seepage into those reaches as determined in the steady-state analysis. Table 6 also gives the inflow from constant head boundaries, which represents ground-water inflow from outside the boundaries of the model, and inflow from "leakage" to the confined aquifers, which represents either contributions from water-table storage in the till plain, or contributions from induced recharge or reduced natural outflow in the till plain.

Table 5.--Pumpage rates by well field for selected model simulations, in million gallons per day

Experiment	Total pumping	Well field 1	Well field 2	Well field 3	Well field 4
A	59	21.0	10.7	25.8	1.5
B	81	26.9	17.5	35.3	1.0
C	72	24.5	14.1	32.4	1.1
D	103	37.5	19.2	45.2	1.1

Locations of well fields and wells are given in figure 19.

Table 6.--Rates of Induced Recharge, in million gallons per day, for experiments A-D

Experiment	Total Pumping	Fall Creek	Eagle Creek	White River Reach 1	2	3	4	5	Constant head boundaries	Leakage flow to combined aquifers
Steady state		14.9	1.51	5.37	11.04	10.98	2.20	12.78	-------	-------
A	59	19.8	1.8	2.7	8.4	1.5	6.6	12.4	4.2	2.7
B	81	30	.6	2.4	14.4	3.6	12.6	12.9	3.5	1.8
C	72	24.0	2.25	3	12.3	4.2	6.0	13.5	6.0	3.0
D	103	35.3	.6	9.6	12.0	4.8	17.9	15.9	4.3	2.4

Locations of river reaches are shown in figure 19.

Experiment A provides a conservative estimate of the potential of the glacial outwash, as it is based on drawdowns of only 50 percent of the saturated thickness in the pumping wells, and assumes an imperfect connection between the streams and the aquifer. In experiment A, the total discharge from wells in the glacial outwash was found to be 57.5 Mgal/d (2.5 m^3/s); an additional 1.5 Mgal/d (0.07 m^3/s) was withdrawn from well field 4 in the middle confined aquifer. The drawdown distributions resulting from this pumpage are shown in figures 23 through 25. These figures show that the regional drawdown in the immediate vicinity of the discharging wells in the glacial outwash would be about 20 ft (6.1 m). In general, however, the drawdowns throughout the glacial outwash would not be such as to cause problems in existing wells, or to cause seepage from any stream reach to approach its limiting value. The seepage rates for Fall Creek, Eagle Creek and reach 4 of the White River (fig. 19) exceed the steady-state seepage into those reaches, implying that there would be a net flow out of these reaches under the new equilibrium. The seepage rates for reaches 1, 2, 3 and 5 on the White River are less than the corresponding steady-state seepages, indicating that the river would still gain flow in these reaches. Well field 4, in the middle confined aquifer, produced a total of 1.5 Mgal/d (0.07 m^3/s) in experiment A. This is a small fraction of the discharge from the glacial outwash; pumpage from the confined aquifer was limited to this level by the constraint that regional drawdowns not exceed 20 ft (6.1 m).

As noted previously, the water table in the till plain was not simulated using capacitance, but rather was treated as a constant head surface through the use of grounded connections. If flow from the water table in this experiment is assumed to represent only storage depletion, it would correspond to a maximum water-table decline in the till plain of approximately 0.1 ft/yr (0.03 m/yr). Simulation of the water table as a constant head boundary thus appears to have been acceptable for this well field.

In experiment B, the same drawdown criteria were applied as in A, although the drawdown values within the wells, after final adjustment of the discharges, were for the most part slightly greater than in experiment A. This was the result of practical difficulties in the simulation and not of any hydrologic consideration. In experiment B, the streams were assumed to be perfectly connected to the upper third of the glacial outwash. The results of this simulation are therefore optimistic and represent the highest possible estimate of the yield that the aquifer might sustain under the limitation that drawdowns in the pumping wells equal approximately 50 percent of present saturated thickness.

The total discharge of the well fields in the glacial outwash in experiment B was 80 Mgal/d (3.5 m^3/s); an additional 1 Mgal/d (0.04 m^3/s) was obtained from well field 4 in the middle confined aquifer. Eagle Creek and reaches 1 and 3 on the White River show changes in seepage less than their steady-state inflow values, and so would continue to gain flow under the new equilibrium; the remaining reaches would all lose flow under the new regime. Flow losses from Fall Creek would be particularly heavy, amounting to approximately 16 Mgal/d (6.7 m^3/s), and there is doubt as to whether the flow of Fall Creek would be sufficient to sustain this loss under drought conditions. Drawdowns at the new equilibrium in experiment B are shown in figures 26 through 28; in general, these drawdowns would not be such as to generate problems.

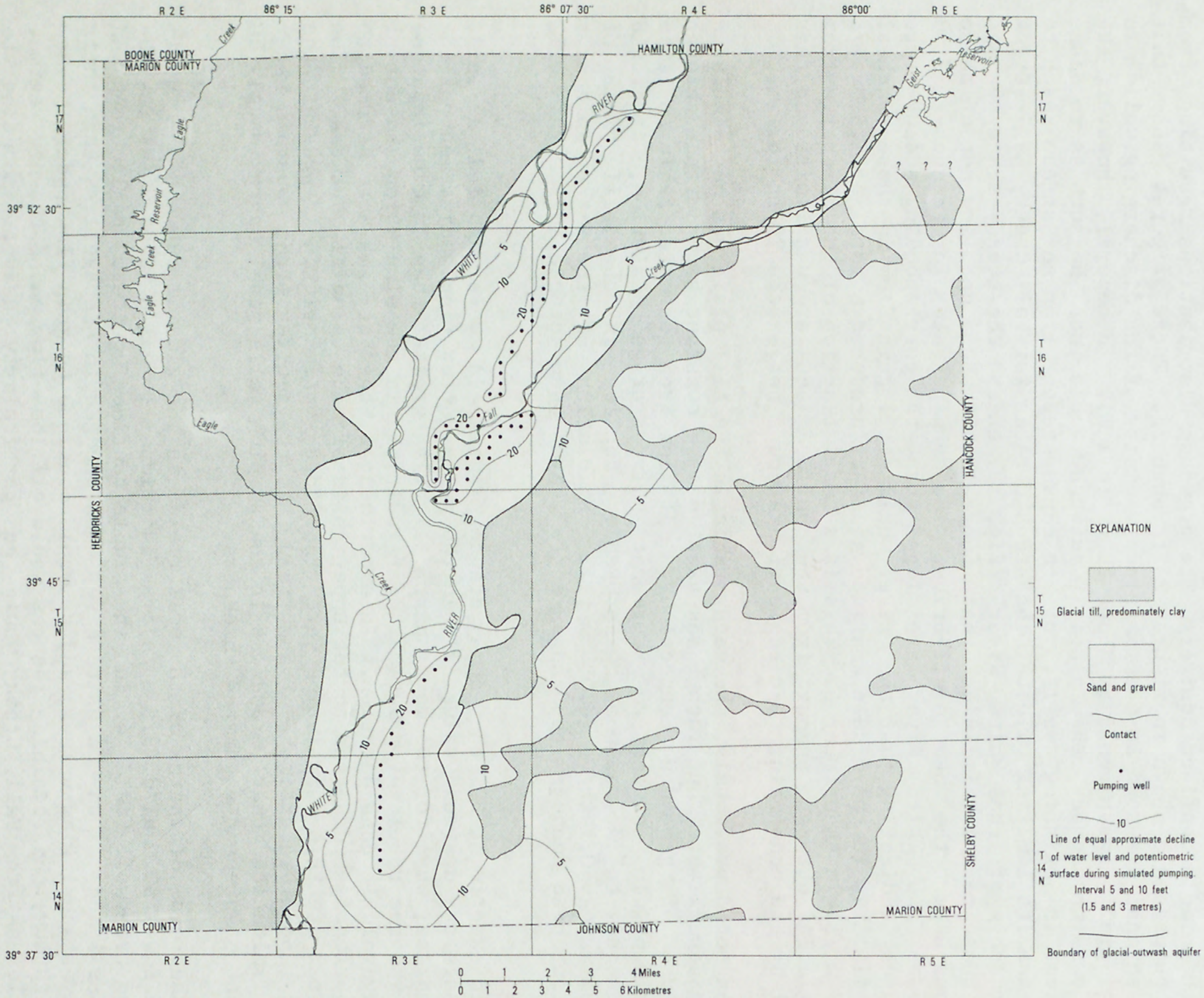

Figure 23. -- Computed declines in the water level in the glacial-outwash aquifer and in the potentiometric surface of the upper confined aquifer for simulated ground-water pumpage equal to 59 million gallons per day (experiment A)

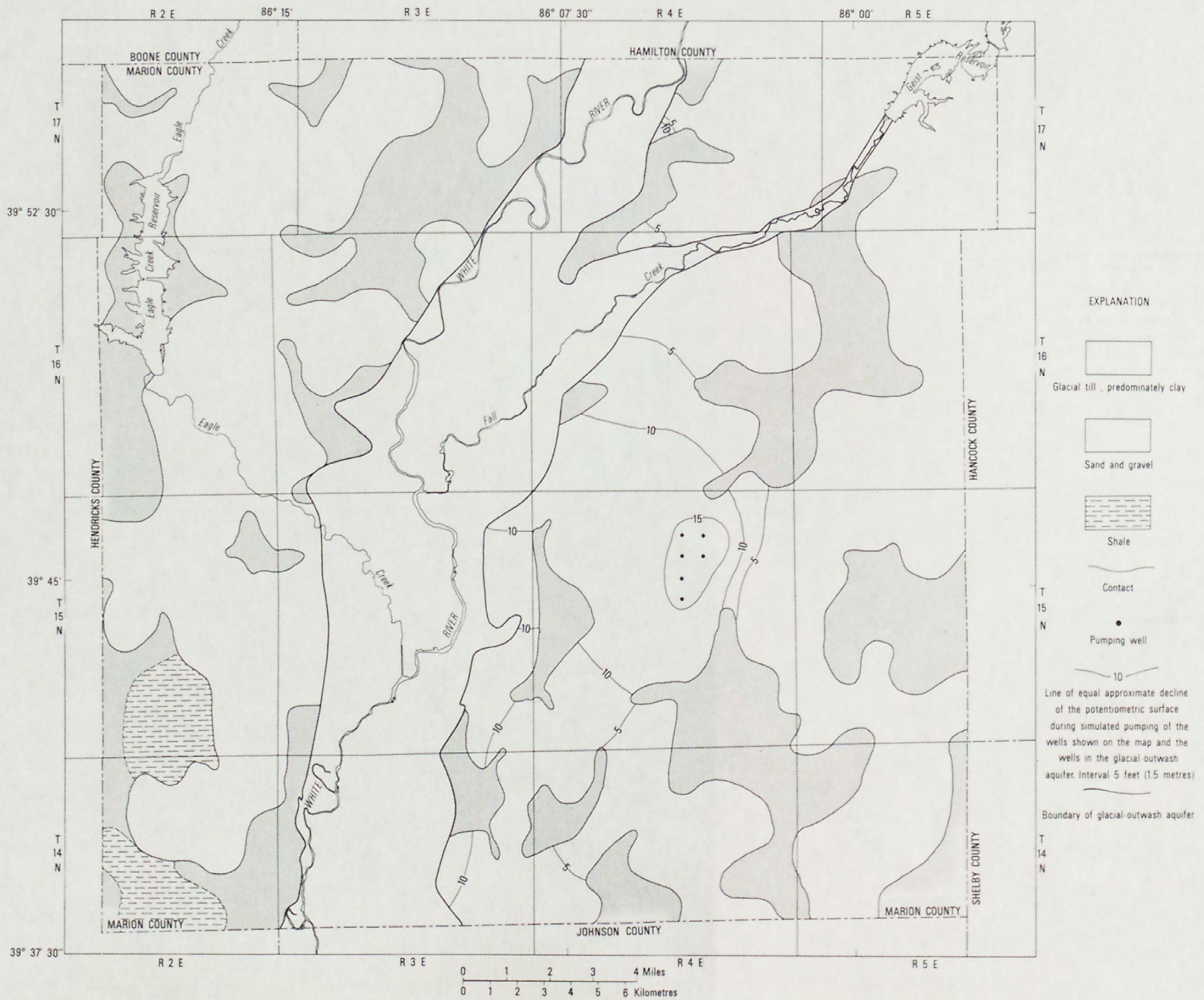

Figure 24. -- Computed declines in the potentiometric surface of the middle confined aquifer for simulated ground-water pumpage equal to 59 million gallons per day (experiment A)

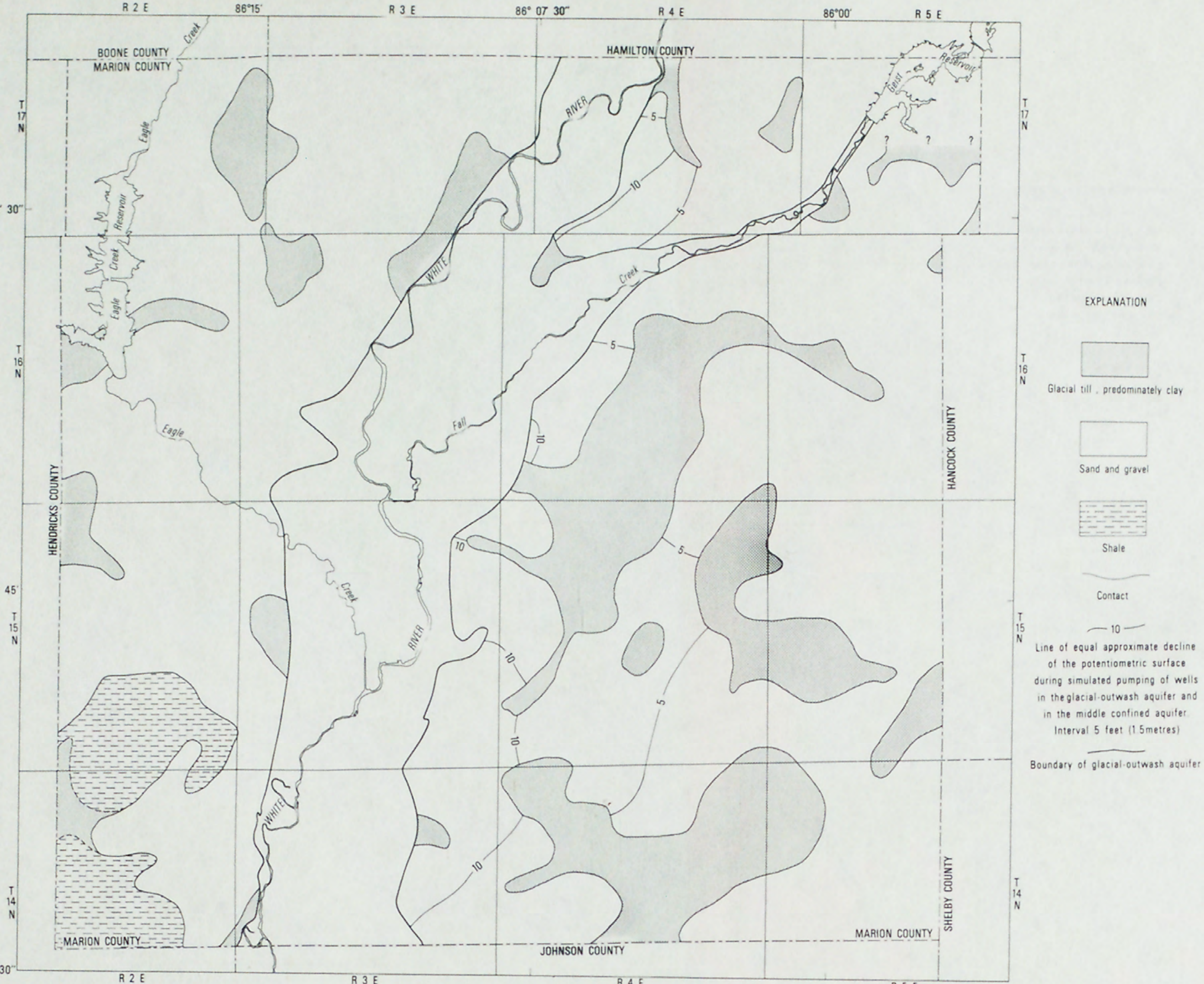

Figure 25. -- Computed declines in the potentiometric surface of the lower confined aquifer for simulated ground-water pumpage equal to 59 million *gallons per day* (*experiment* A)

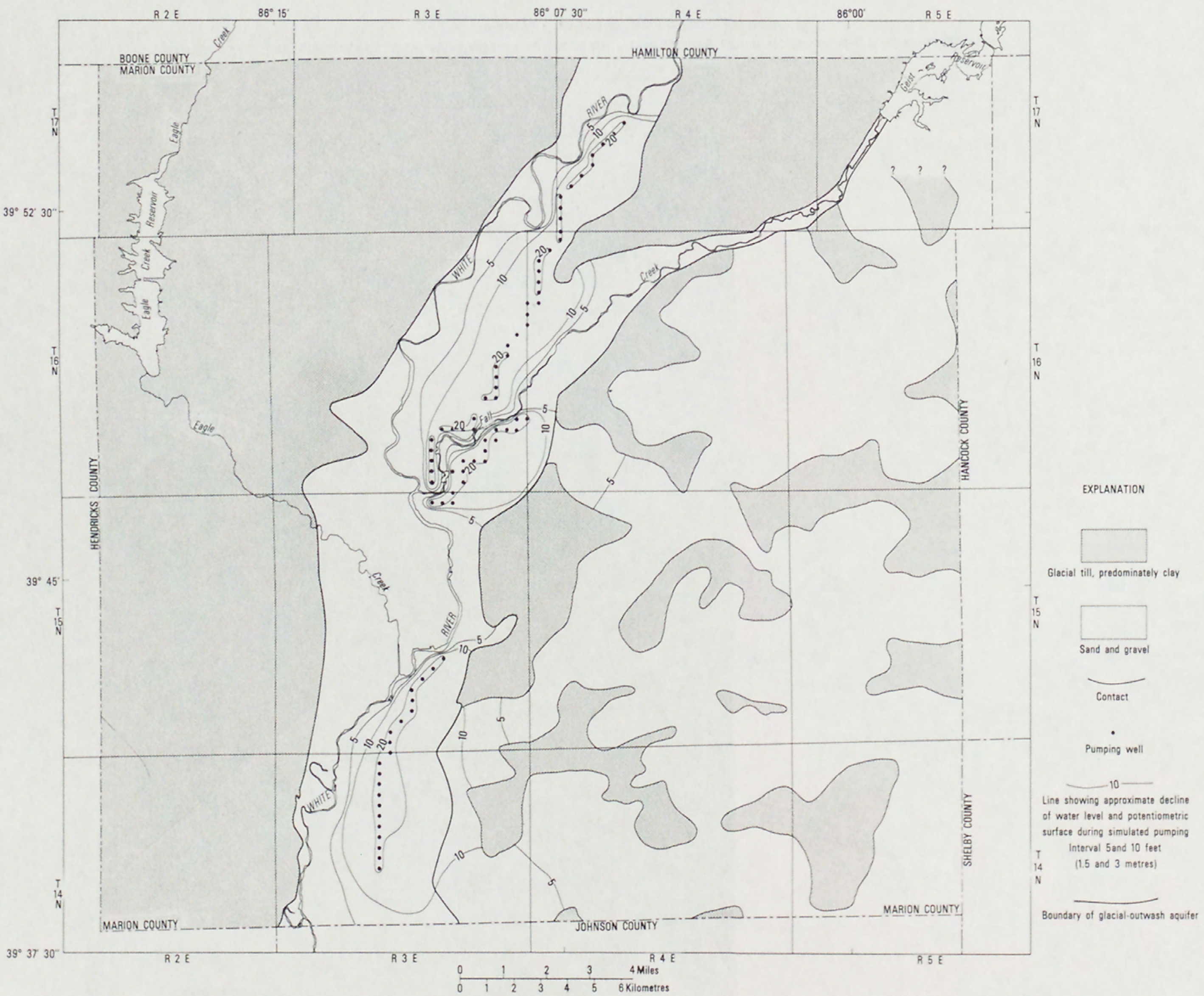

Figure 26. -- Computed declines in the water level in the glacial-outwash aquifer and in the potentiometric surface of the upper confined aquifer for simulated ground-water pumpage equal to 81 million gallons per day (experiment B)

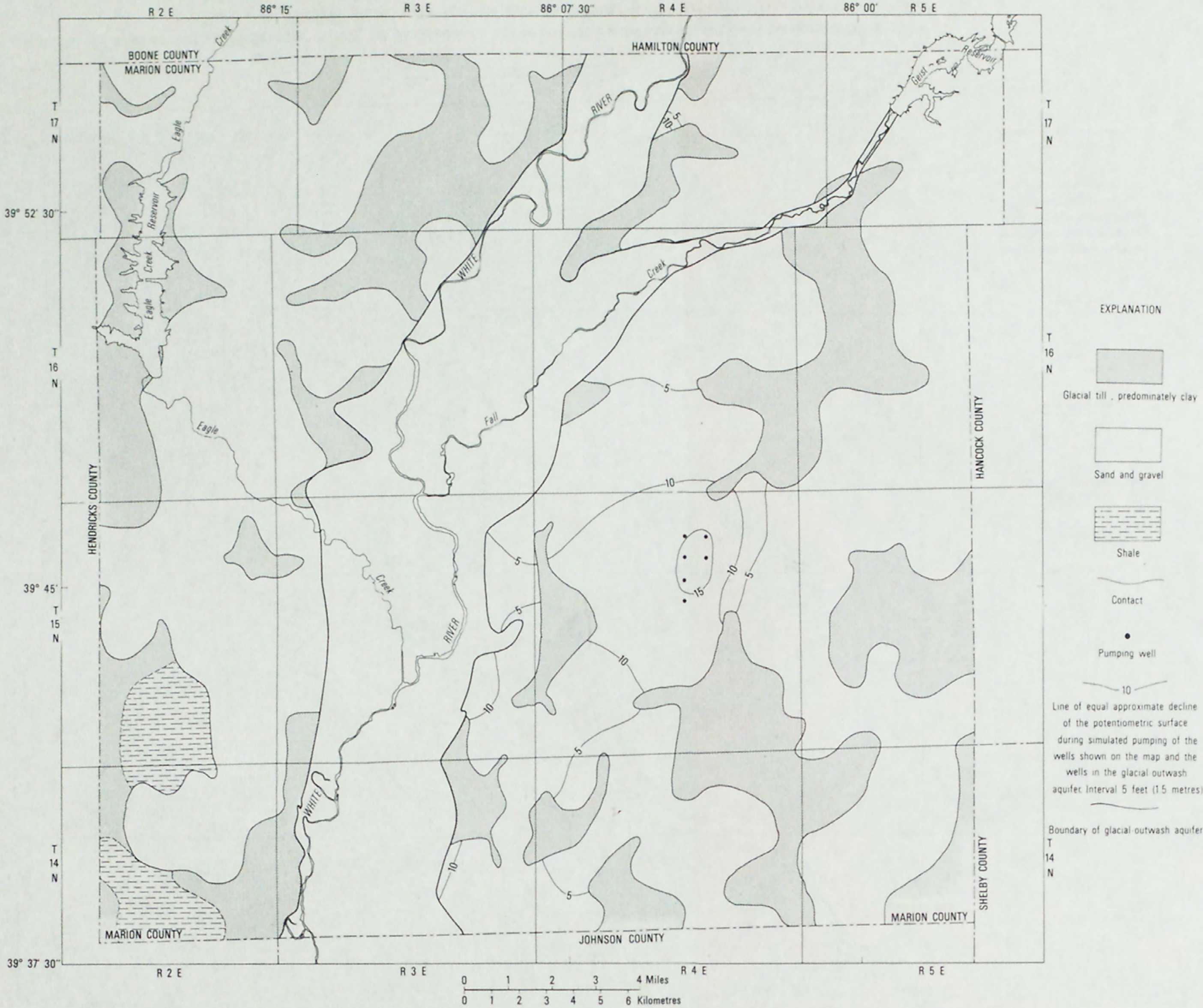

Figure 27. -- Computed declines in the potentiometric surface of the middle confined aquifer for simulated ground-water pumpage equal to 81 million gallons per day (experiment B)

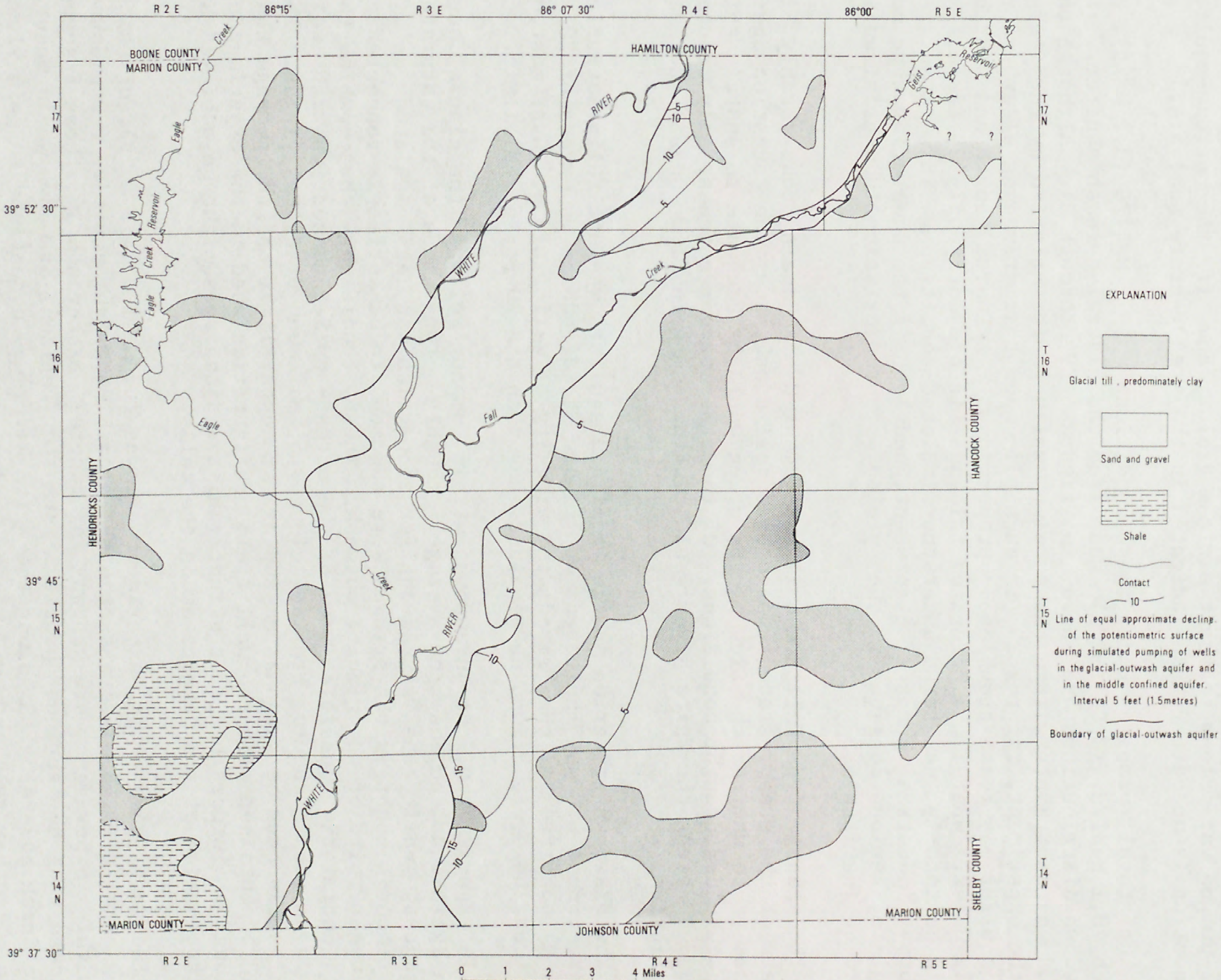

Figure 28. -- Computed declines in the potentiometric surface of the lower confined aquifer for simulated ground-water pumpage equal to 81 million gallons per day (experiment B)

In experiment C, drawdowns in the wells in the glacial-outwash aquifer equalled, on the average, roughly two-thirds of the original saturated thickness of the aquifer. At the same time, the stream-aquifer connection was modeled as in experiment A, using the streambed hydraulic conductivities developed in the steady-state analysis. In a sense, therefore, experiment C is a "worst-case" analysis, giving results that could be expected if wells are pumped at close to their maximum yields, and if the stream-aquifer connection is no better than indicated in the steady-state analysis.

The total pumpage under these conditions was 72 Mgal/d (3.2 m^3/s), of which 70.1 Mgal/d (3.07 m^3/s) was from the glacial-outwash aquifer. The drawdown distributions are shown in figures 29 through 31. Although the drawdown constraints within the pumping wells were satisfied by the design of the experiment, figures 29 through 31 show that drawdowns elsewhere in the aquifer may reach troublesome proportions. An area south and west of Eagle Creek, where the saturated thickness is presently only 20 ft (6.1 m) would be totally dewatered under the simulated conditions. This is an area of heavy industrial pumpage at present, and even if the dewatering were not complete, serious declines in the yields of the existing industrial wells would be inevitable.

The seepage from Fall Creek in experiment C is again considerably in excess of the steady-state seepage into Fall Creek, and the question again arises as to whether the flow of the creek would be adequate to sustain this seepage under drought conditions. Of the other stream reaches, only reaches 1 and 3 along the White River would continue to gain flow under the new equilibrium.

The drawdown criterion in the confined aquifers was not fully satisfied in experiment C; as indicated on figure 30, drawdowns in excess of 20 ft (6.1 m) occurred in the immediate vicinity of well field 4. Over most of the area, however, drawdowns remained less than the 20-ft (6.1-m) limit.

In experiment D, the drawdowns within pumping wells in the glacial outwash again equalled, on the average, roughly two-thirds of the saturated thickness, while the stream-aquifer connection was as simulated as in experiment B--that is the streams were assumed to be perfectly connected to the upper third of the glacial outwash. These conditions thus combine the most optimistic possible view of the stream-aquifer connection with well drawdowns that should provide close to maximum practical yields. The results can thus be taken as an approximate upper limit to the pumpage that could be obtained with the well field configuration used in the experiments, although it is possible that slightly higher discharges might be achieved by redistribution of the pumpage among the wells.

The total pumpage in experiment D, was 103 Mgal/d (4.5 m^3/s), of which 101.9 Mgal/d (4.46 m^3/s) was from the glacial-outwash aquifer. Drawdowns from this experiment are shown in figures 32 through 34; these figures indicate that no serious problems due to regional drawdown would occur. Eagle Creek and Reach 3 along the White River would continue to gain flow, while all other reaches would lose, under the new equilibrium. As in experiments B and C, the losses from Fall Creek would be high, prompting the question as to whether the flow of the creek could sustain the seepage during prolonged drought.

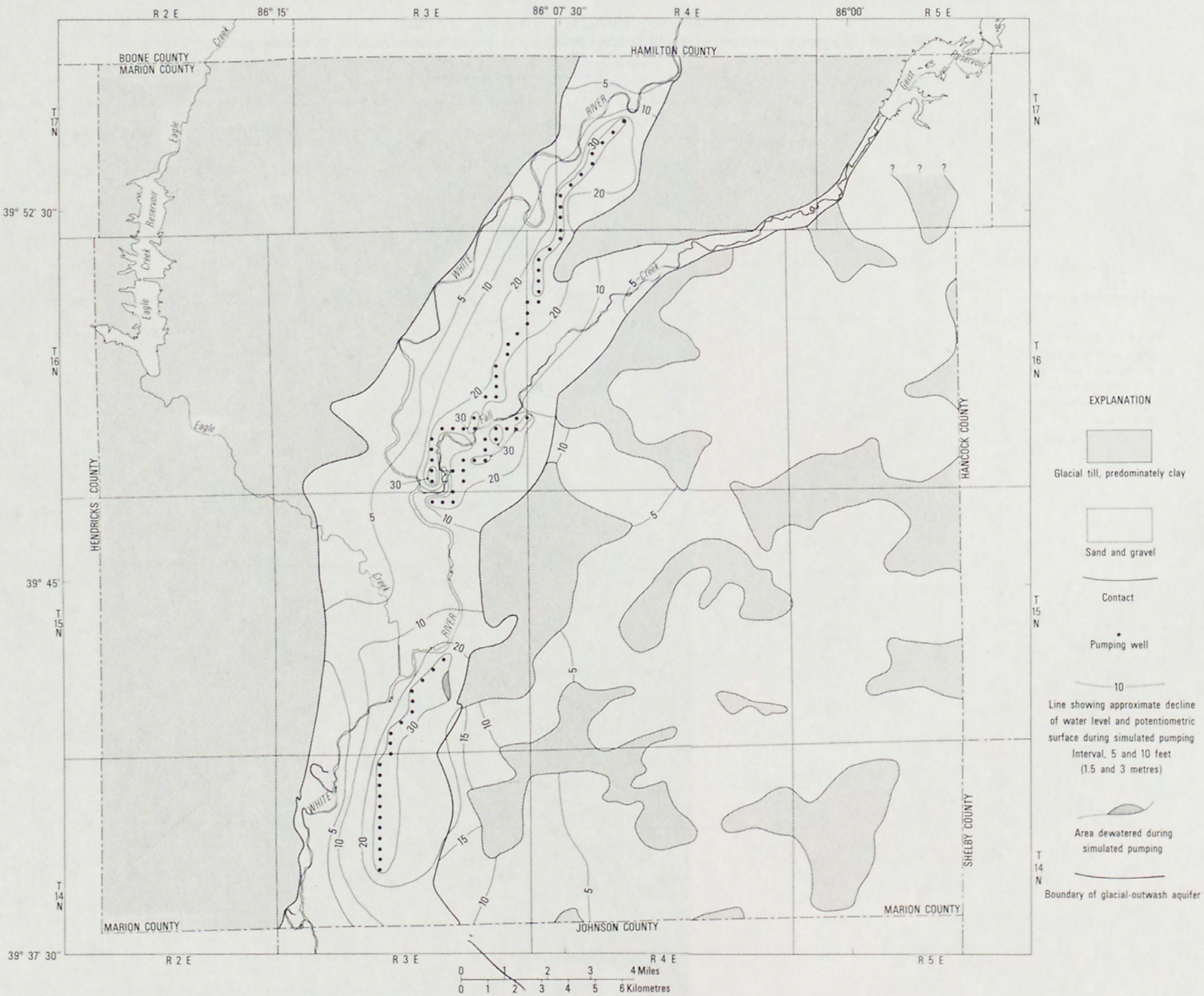

Figure 29. -- Computed declines in the water level in the glacial-outwash aquifer and in the potentiometric surface of the upper confined aquifer for simulated ground-water pumpage equal to 72 million gallons per day (experiment C)

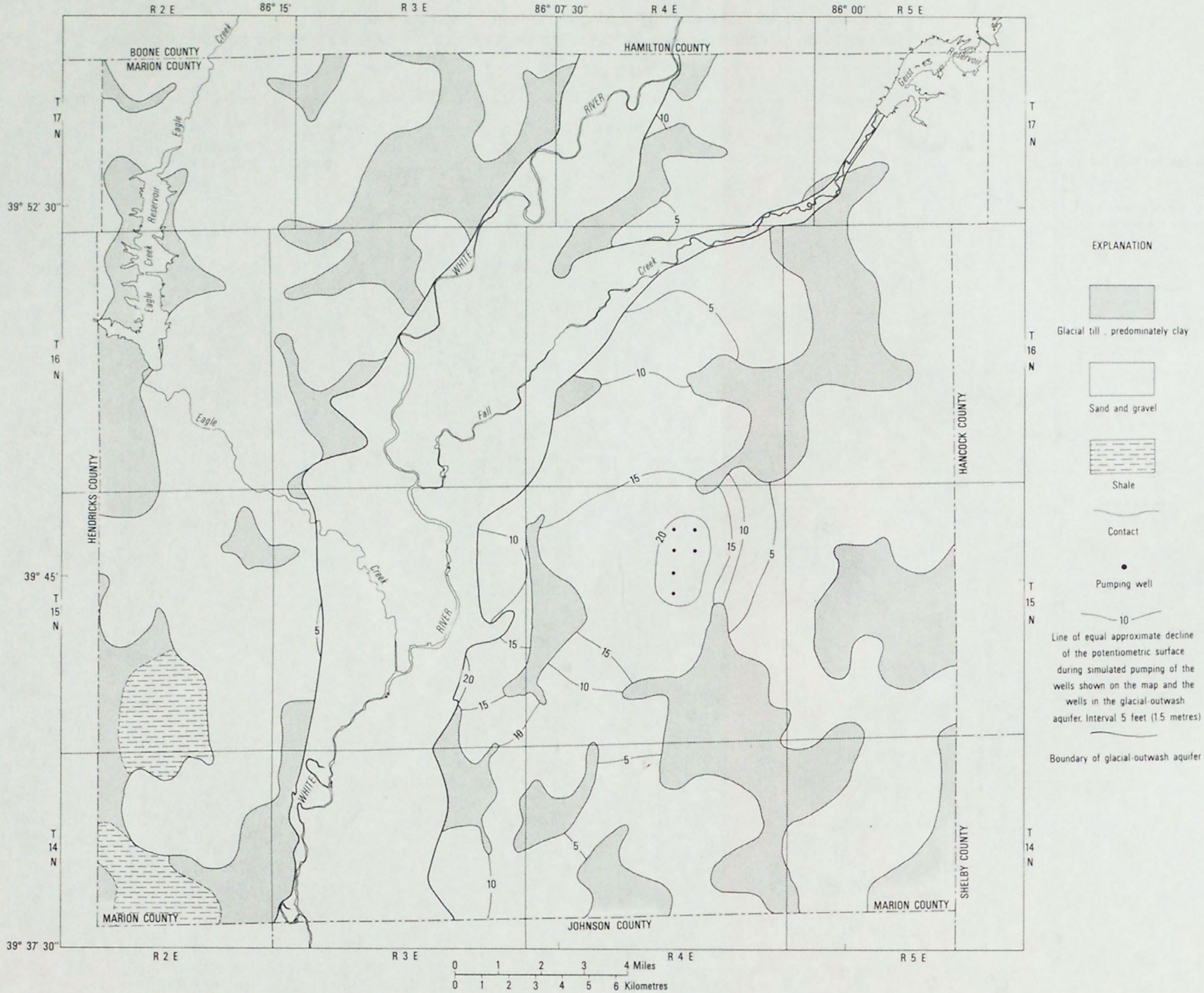

Figure 30. -- Computed declines in the potentiometric surface of the middle confined aquifer for simulated ground-water pumpage equal to 72 million gallons per day (experiment C)

Figure 31 -- Computed declines in the potentiometric surface of the lower confined aquifer for simulated ground-water pumpage equal to 72 million gallons per day (experiment C)

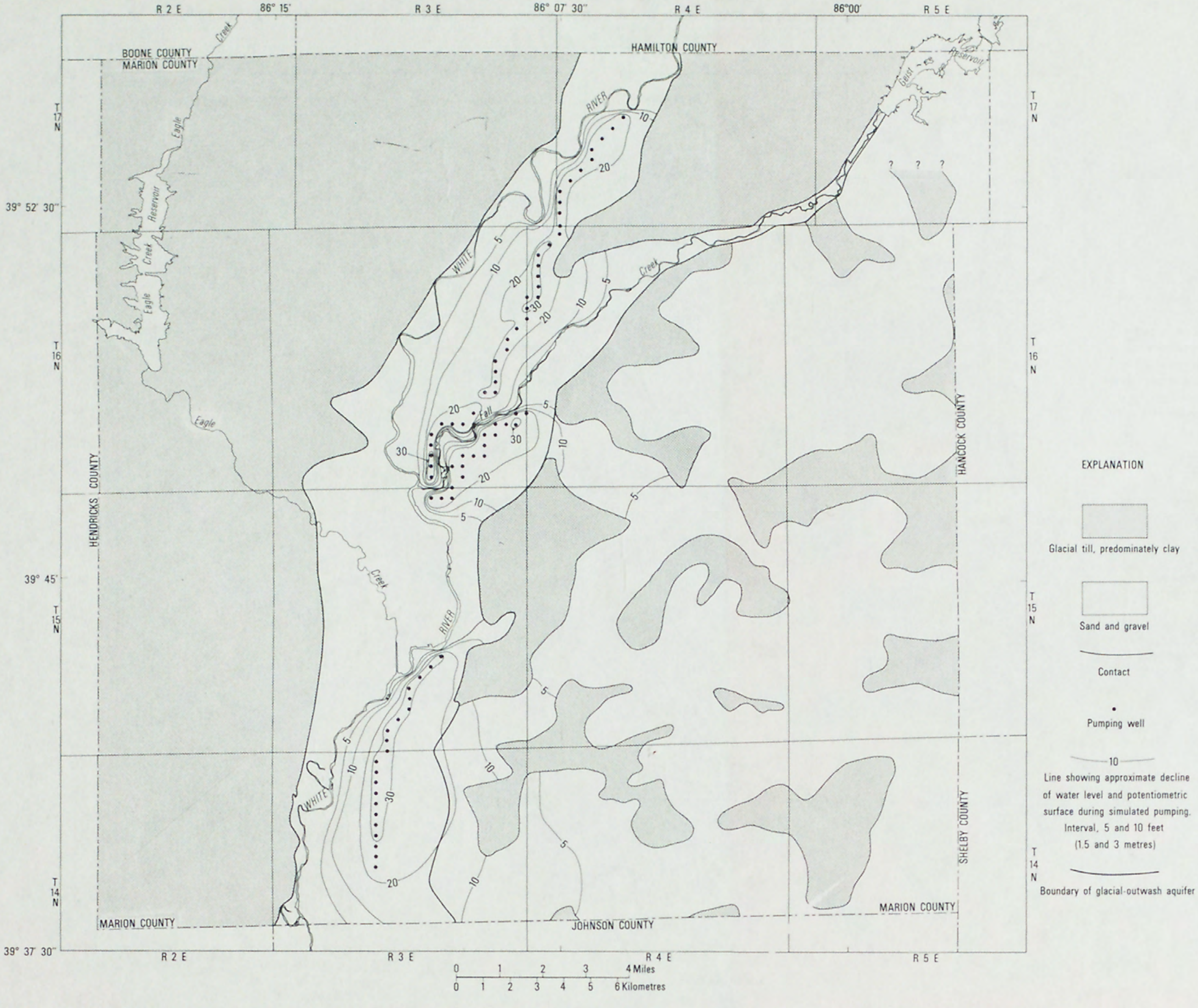

Figure 32. -- Computed declines in the water level in the glacial-outwash aquifer and in the potentiometric surface of the upper confined aquifer for simulated ground-water pumpage equal to 103 million gallons per day (experiment D)

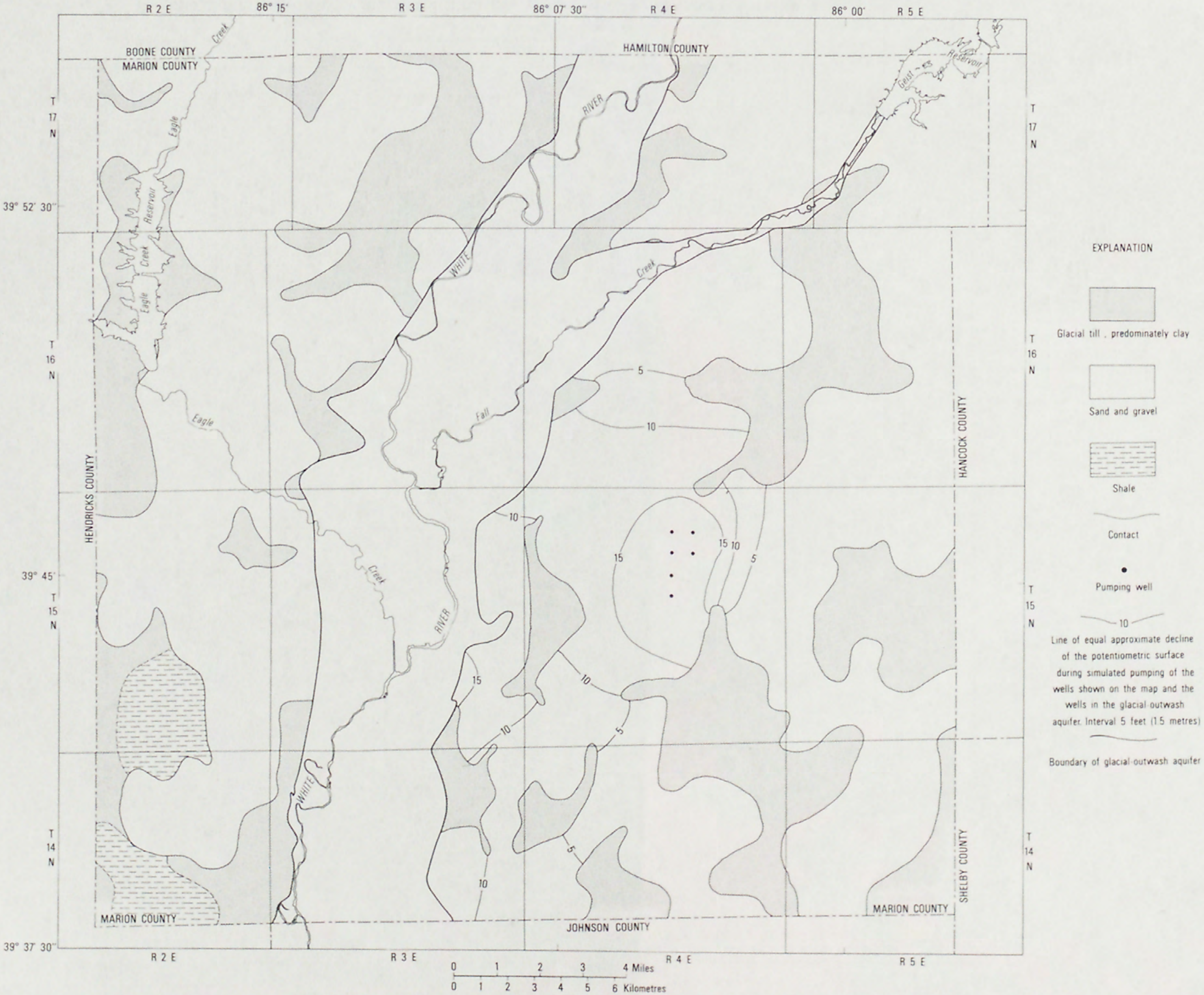

Figure 33. -- Computed declines in the potentriometric surface of the middle confined aquifer for simulated ground-water pumpage equal to 103 million gallons per day (experiment D)

Figure 34. -- Computed declines in the potentiometric surface of the lower confined aquifer for simulated ground-water pumpage equal to 103 million gallons per day (experiment D)

Experiments A through D constituted the most significant phases of the analog-model analysis. However, three other experiments, all designed to answer specific questions, were carried out.

In experiment E, well field 5 of figure 19, tapping the lower confined aquifer west of the White River, was pumped. The drawdown constraint was that employed for well field 4 in experiments A through D--that the regional drawdown should not exceed 20 ft (6.1 m) anywhere in the aquifer. Under this constraint it was found that 4.2 Mgal/d (0.18 m^3/s) could be pumped from well field 5. However, the analog simulation in this case proved to be inadequate. The water table in the till plain, as noted earlier, was simulated as a constant head surface. Flow from this surface, when translated into an equivalent rate of fall of the water table, yielded a decline of 2 to 3 ft/yr (0.6 to 0.9 m/yr). This indicated that treatment of the water table as a constant head boundary was not justified, and that had the water table been modeled using capacitance, the same equilibrium would not have been achieved. Specifically, the yield of the well field, subject to the same constraint, would have been less when equilibrium was reached. Thus, while the simulation was inadequate in experiment E, the results indicate that somewhat less than 4.2 Mgal/d (0.18 m^3/s) could be developed at this site under a new equilibrium, with the 20-ft (6.1-m) drawdown restriction.

Experiments F and G were designed to test the capacity of the aquifer to produce high yields for short periods. In experiment E, well field 1 of figure 19 was pumped at a rate sufficient to bring drawdowns in the aquifer to the levels measured at equilibrium in experiment A, within a period of 30 days. In experiment G, well field 1 was pumped at a rate sufficient to achieve these drawdowns within a period of 90 days.

The discharge from well field 1 in experiment E was 59 Mgal/d (2.6 m^3/s); its discharge in experiment G was 38 Mgal/d (1.66 m^3/s). Figure 35 shows a hydrograph of drawdowns in the central part of well field 1 in the two experiments. During the 30- and 90-day periods, 92 and 76 percent of the respective pumpage is taken from aquifer storage. The results can be taken as a rough indication of the potential of the aquifer to sustain pumpage during periods of drought. However, it must be recognized that the results indicate what would happen due only to an increase in pumpage of 59 or 38 Mgal/d (2.6 or 1.66 m^3/s). If water levels are first drawn down through general ground-water development, the effects indicated in figure 35 would be superposed on the already lowered water levels, with consequences correspondingly more severe.

It should be emphasized that in all of the experiments, the predictive results may tend to be optimistic because well-entrance losses were assumed negligible throughout the analysis. However, this is offset by the fact that a single pumping well was postulated for each node of the model from which discharge was simulated, whereas in practice, the discharge from a node area could be withdrawn through several wells to minimize well losses and local formation losses.

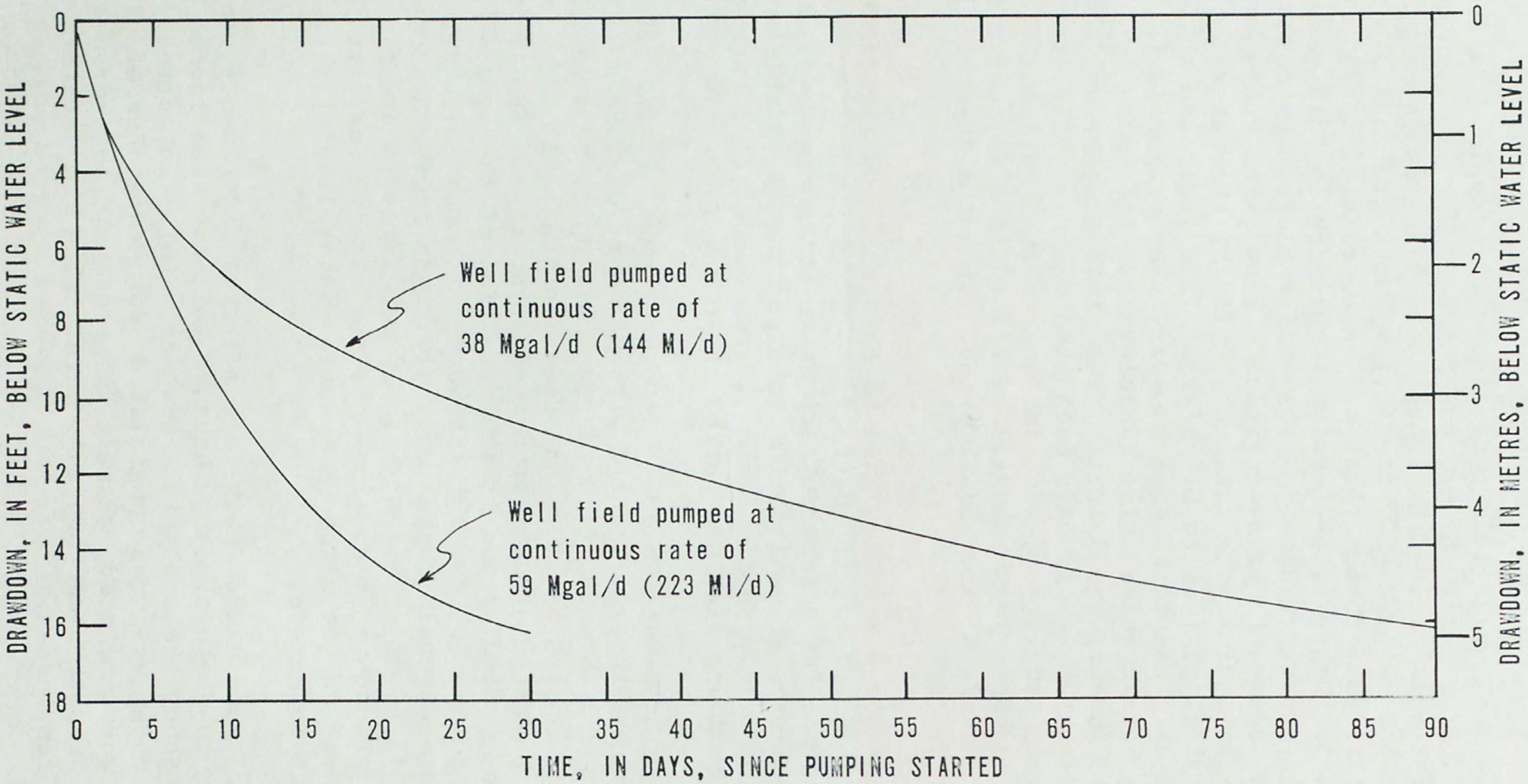

Figure 35.-- Water level declines and rates of continuous pumpage for 30 and 90 days in area 1

WATER QUALITY

by William J. Shampine

Water-quality samples were obtained from 32 small-diameter observation wells distributed fairly evenly throughout the glacial-outwash aquifer. These data (table 7) indicate that, in general, the ground water in the glacial-outwash aquifer is of acceptable quality for most uses although it is quite hard. Locally, however, the water quality appears to have been affected by the presence of man. The extent of this effect is indeterminate with the data available.

The data also indicate that there are some significant differences in water quality characteristics in the glacial-outwash aquifer. The specific conductance of water ranged from 508 to 1,760 micromhos (fig. 36). Limited data and complex flow patterns prevent accurate definition of areal patterns; however, the specific conductance in wells seems to decrease with increasing distances from urbanized areas.

The dissolved-solids concentration of the waters may be estimated by multiplying the specific conductance by a factor of 0.6. For example, water in the study area with a specific conductance of 1,000 micromhos will have a dissolved-solids concentration of about 600 mg/l (milligrams per litre).

Total organic carbon (TOC) is another general measure similar to specific conductance for interpreting water quality. It is a gross measure of the organic carbon in the water, with the limitation that it gives no indication of specific organic compounds. On a national basis, background levels of TOC in ground water generally are less than about 5 mg/l. TOC concentrations in the glacial-outwash aquifer ranged from 0.5 to 410 mg/l (fig. 37), with 9 of the 32 values greater than 20 mg/l and 4 values greater than 100 mg/l. Although it is conceivable that these high values are the result of decay of organic humus buried in the ground, it is more reasonable to assume they result from urban wastes.

Stiff (1951) proposed a method of plotting chemical analyses that would yield distinctive patterns for waters of different chemical compositions. Figure 38 contains a series of Stiff patterns and shows several distinct water types. Ground water in the glacial outwash is basically a calcium magnesium bicarbonate type water, as illustrated by the Stiff patterns for wells 1, 15, 20, and 67. The analyses represented by these four patterns probably approximate the "natural" quality expected in this area. More highly mineralized water or other water types could result from some activity of man. For example, the addition of sodium chloride (possibly from road salt) is clearly shown in the patterns for wells 23, 39, and 45. Gypsum (calcium sulfate) has been added to the water from well 47 and sodium bicarbonate has been added to the water from well 82. Data are too limited to define the sources or movement of any of these constituents, but the fact that there are considerable differences throughout the area is apparent.

In environments where salts do not occur in the rocks, the chloride concentration is a good general indicator of the influence of man on

Table 7.--Chemical analyses of water from selected wells in the glacial-outwash aquifer in the Indianapolis area, April 1975

(Chemical analyses, in milligrams per litre)

Well no.	Location	Alkalinity as $CaCO_3$	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Dissolved solids (Calc.)	Hardness as $CaCO_3$ Calcium, magnesium	Hardness as $CaCO_3$ Noncarbonate	Specific conductance (micromhos at 25°C)
1	Lat 393830 Long 0861201	244	83	25	7.2	1.4	298	36	19	0.1	328	310	66	570
2	Lat 393829 Long 0861201	331	91	26	28	2.0	403	33	30	.0	421	330	4	710
6	Lat 393954 Long 0861147	281	99	36	3.4	1.1	343	68	16	.2	405	400	110	673
8	Lat 393946 Long 0861359	192	63	24	16	1.5	234	49	32	.2	308	260	64	538
13	Lat 394129 Long 0861113	430	150	44	29	2.5	524	120	49	.1	663	560	130	1,080
15	Lat 394129 Long 0861308	310	120	30	26	2.1	378	69	42	.1	484	420	110	768
17	Lat 394225 Long 0861040	250	100	38	27	2.6	305	130	43	.0	499	410	160	810
20	Lat 395534 Long 0860530	268	100	31	7.1	1.5	327	65	14	.1	389	380	110	654
22	Lat 395443 Long 0860726	303	110	37	44	2.5	370	42	77	.2	508	430	120	887
23	Lat 395353 Long 0860555	350	130	53	84	2.8	427	67	210	.2	771	540	190	1,440
25	Lat 395255 Long 0860628	314	110	38	44	2.6	383	64	100	.0	559	430	120	993
26	Lat 395300 Long 0860751	391	130	45	21	1.7	477	35	51	.1	532	510	120	900
28	Lat 395156 Long 0860814	255	100	35	17	1.6	311	76	38	.1	140	390	140	756
30	Lat 395105 Long 0860952	328	150	48	25	2.0	400	130	84	.2	650	570	240	1,030
36	Lat 395027 Long 0860606	94	46	19	26	2.4	115	62	61	.1	274	190	99	508
37	Lat 395031 Long 0860724	282	90	30	15	2.3	344	45	45	.0	407	350	66	670
39	Lat 395013 Long 0860923	278	140	42	82	7.1	339	130	190	.0	764	520	240	1,350
45	Lat 394830 Long 0861036	310	140	48	36	2.7	378	130	100	.1	657	550	240	1,120
47	Lat 394744 Long 0860815	369	220	55	48	5.3	450	400	41	.1	998	780	410	1,400
51	Lat 394652 Long 0861031	343	160	50	39	8.4	418	180	98	.1	754	610	260	1,210
54	Lat 394551 Long 0861140	296	160	47	39	2.5	361	220	65	.0	723	590	300	1,090
63	Lat 394656 Long 0860852	399	130	45	16	4.0	487	120	15	.2	581	510	110	844
67	Lat 394828 Long 0861252	290	93	34	9.7	1.1	353	54	19	.1	400	370	83	685
69	Lat 394651 Long 0861231	308	120	37	18	2.4	375	97	32	.1	501	450	140	812
70	Lat 394647 Long 0861335	372	110	43	16	4.6	454	65	23	.1	495	450	79	775
75	Lat 394408 Long 0861258	371	130	38	35	3.5	452	95	32	.0	568	480	110	893
78	Lat 394321 Long 0861114	323	150	38	19	4.3	394	150	38	.1	606	530	210	961
82	Lat 394447 Long 0861251	861	4.8	4.6	190	9.5	1,050	3.0	83	.4	815	31	0	1,760
84	Lat 394436 Long 0861213	244	86	24	21	2.6	297	51	41	.1	381	310	70	664
94	Lat 394440 Long 0861043	271	140	36	13	2.7	331	130	11	.0	508	500	230	331
97	Lat 394837 Long 0860918	225	130	40	31	1.7	274	120	80	.2	549	490	260	274
100	Lat 394410 Long 0861214	346	95	31	12	1.8	422	.2	21	.1	385	360	19	422

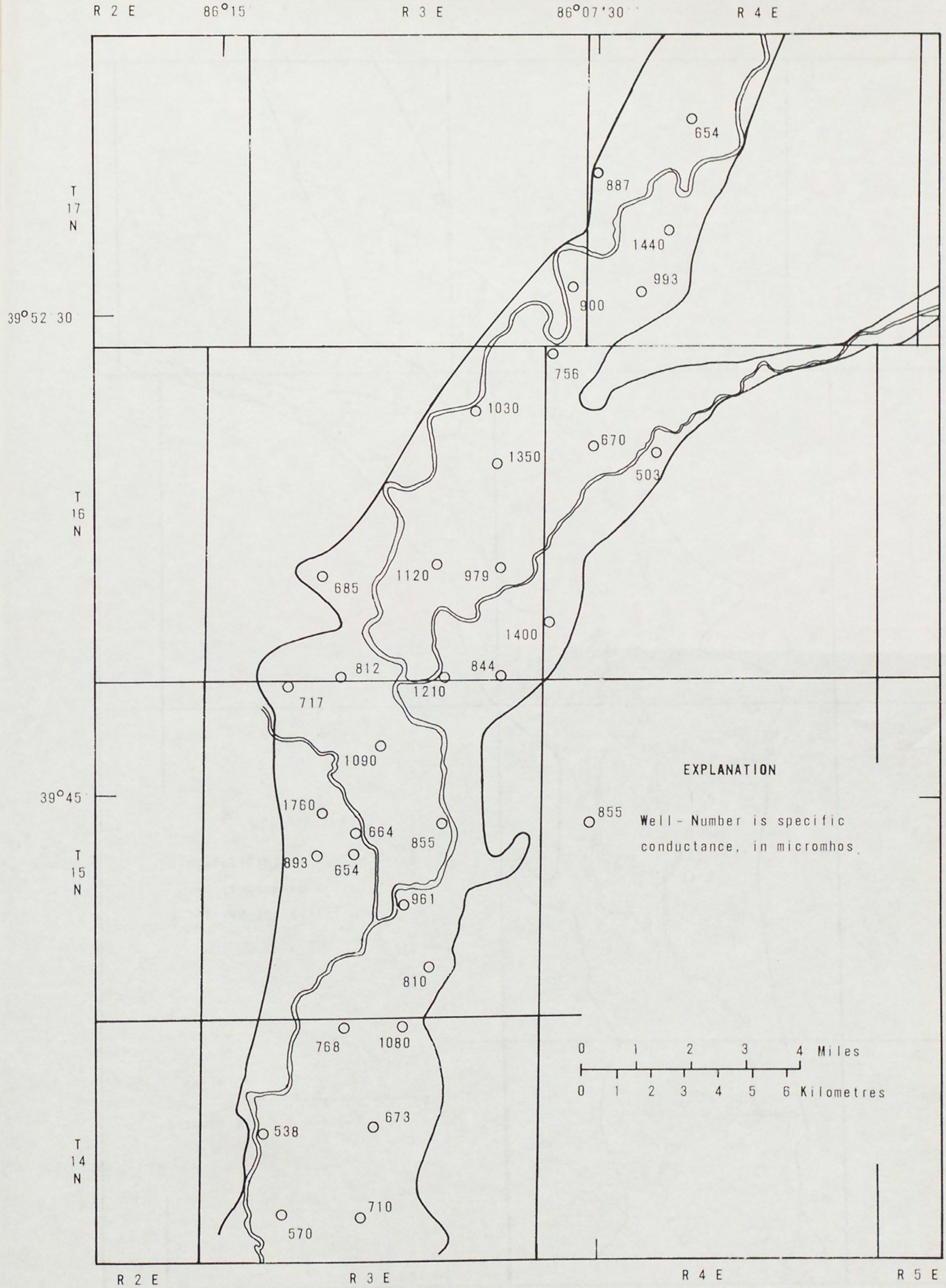

Figure 36.-- Specific conductance of water from selected wells in the Indianapolis area

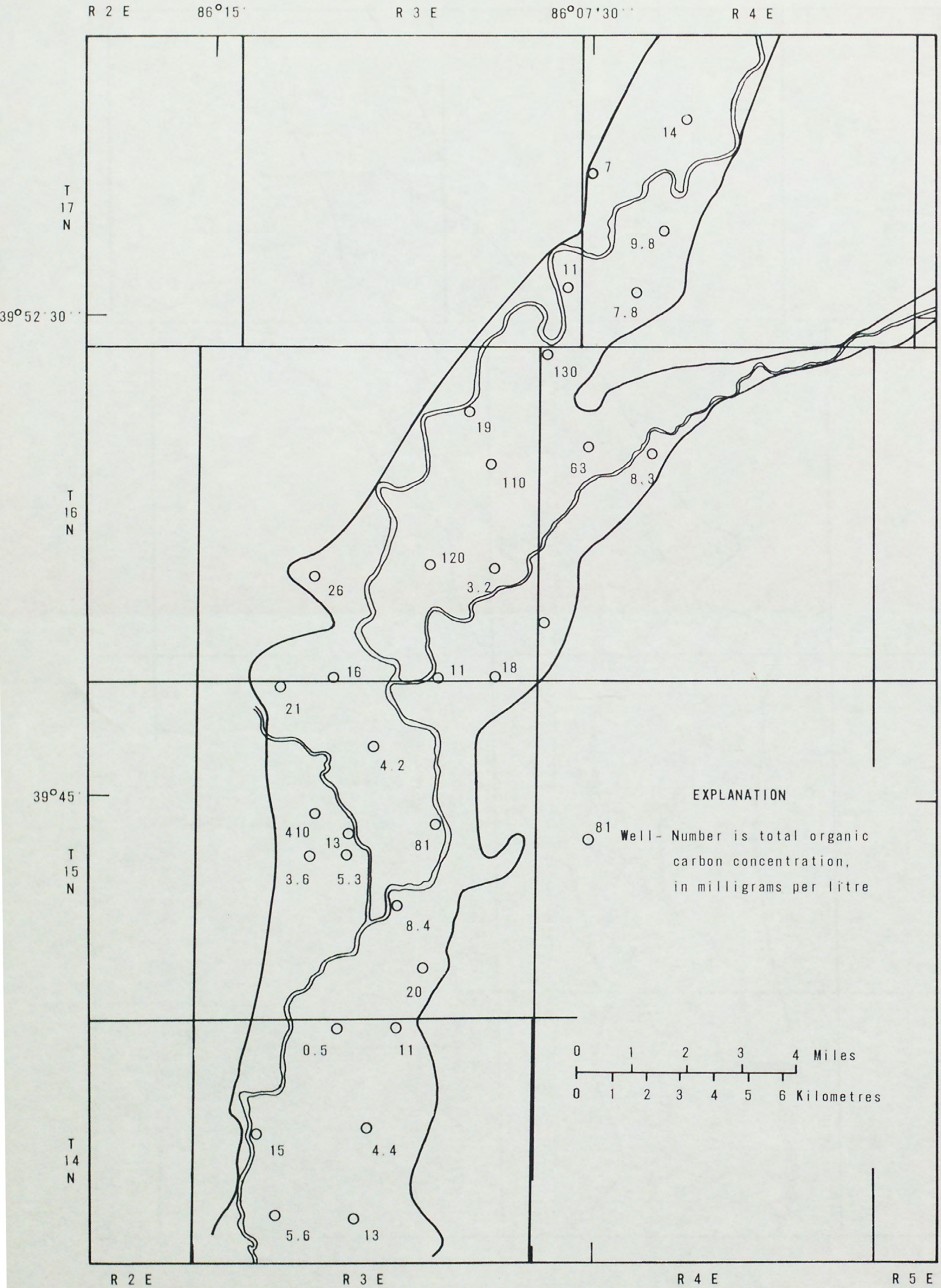

Figure 37.-- Total organic carbon concentration in water from selected wells in the Indianapolis area

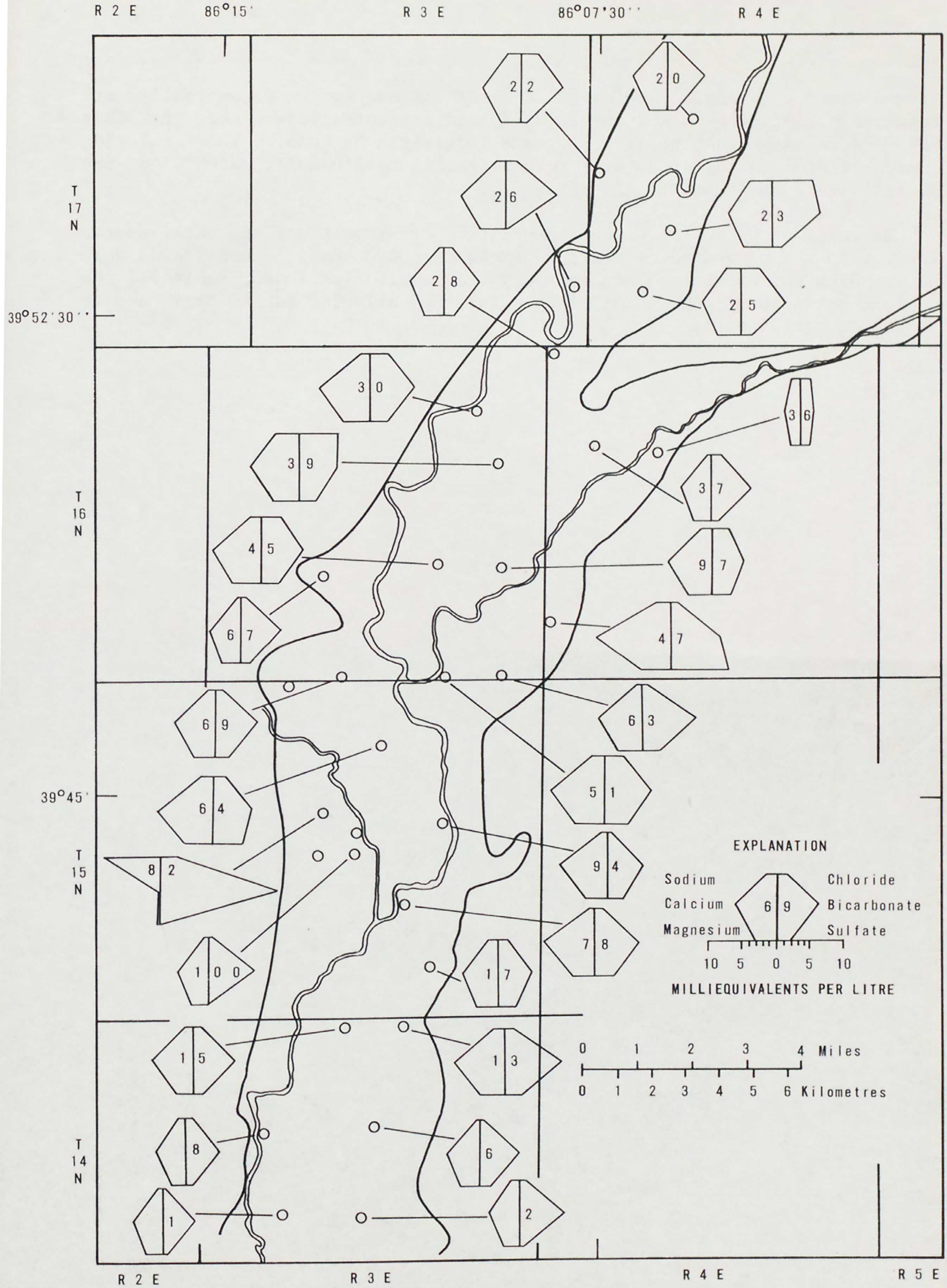

Figure 38.-- Stiff patterns for waters from selected wells in the Indianapolis area

ground-water quality. Chloride concentrations in water normally are relatively low, less than about 25 mg/l, and concentrations higher than this may represent the result of some activity by man. The chloride concentrations in the study area ranged from 11 to 210 mg/l, with 25 of the 32 values greater than 25 mg/l (fig. 39).

Hardness of water is a characteristic of importance to water users. Water with a hardness concentration exceeding 180 mg/l is considered very hard. With the exception of one atypical well (well 82, table 7), the minimum hardness of the water found in the alluvium is 260 mg/l, and it ranges up to 780 mg/l.

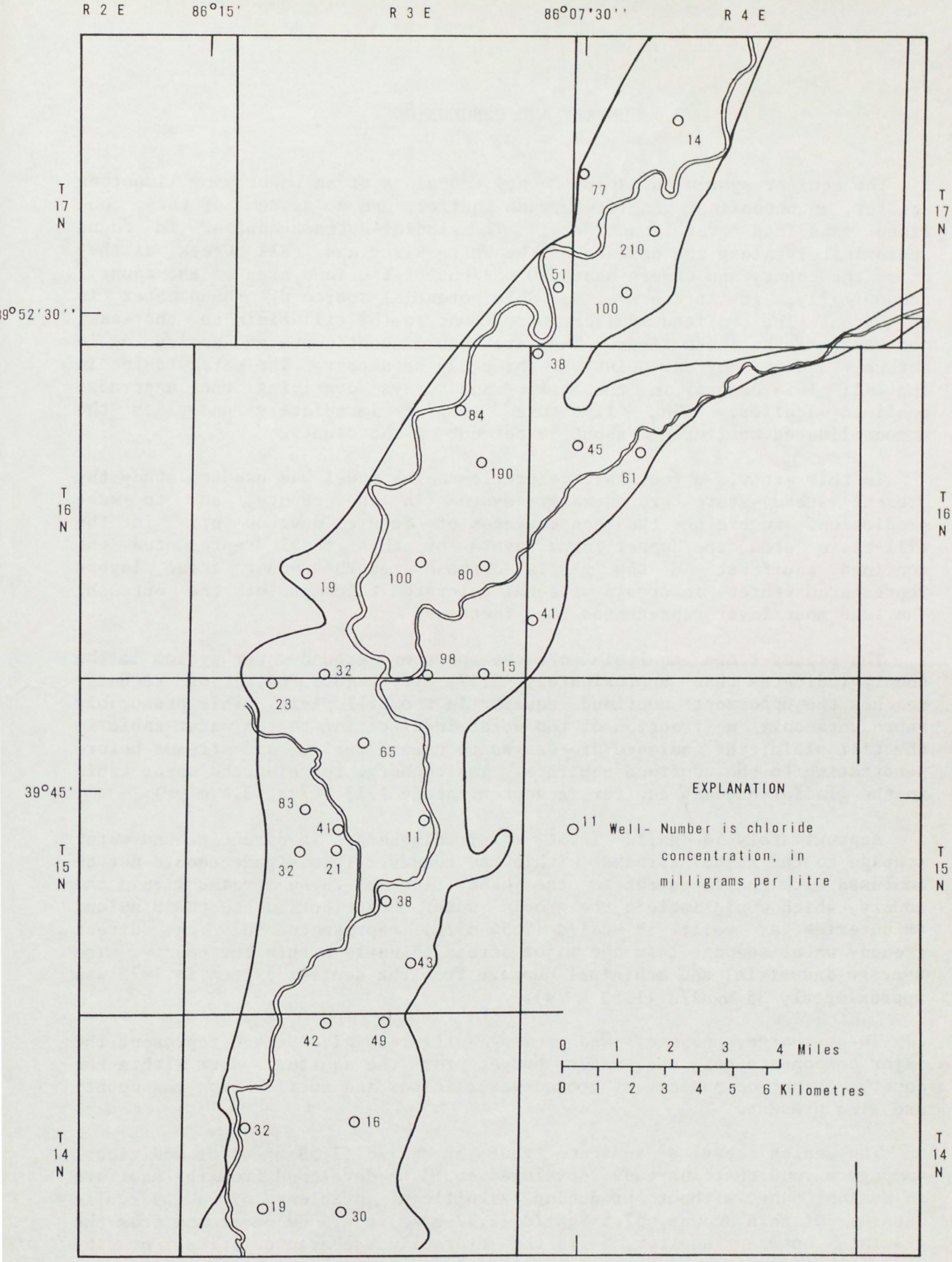

Figure 39.-- Chloride concentration in water from selected wells in the Indianapolis area

SUMMARY AND CONCLUSIONS

The aquifer system in Marion County consists of an underlying limestone aquifer, an unconfined glacial-outwash aquifer, and a system of three confined sand and gravel aquifers. The glacial-outwash aquifer is found approximately along the courses of the White River and Fall Creek as they cross the county and covers about 20 percent of the land area of the county. Economically, it is the most valuable potential source of ground water in the area. The confined aquifers were found in the till plain to the east and west of the White River. They are both discontinuous and coalescing in nature, and at any one point all three may be absent. The water table in the till plain occurs in the silts and clays overlying the uppermost confined aquifer. The limestone aquifer immediately underlies the unconsolidated aquifers in about 50 percent of the county.

In this study, a four-layer electric-analog model was used to study the present steady-state ground-water system in the county, and to make predictions regarding the consequences of future development. In the till-plain area the upper three layers of this model represented the confined aquifers; in the glacial-outwash area the upper three layers represented three intervals of equal saturated thickness of the outwash. The lowermost layer represented the limestone.

The steady-state analysis of the present ground-water system in the county indicates that approximately 0.17 ft/yr (0.5 m/yr) of recharge reaches the uppermost confined aquifer in the till plain. This presumably represents only a fraction of the water infiltrating to the water table in the till plain; the balance discharges to upland springs and streams before penetrating to the confined aquifers. The recharge reaching the water table in the glacial-outwash aquifer is approximately 1.13 ft/yr (3.4 m/yr).

Approximately 58 Mgal/d (2.54 m^3/s) discharges in direct ground-water seepage to the major streams within the county. This figure should not be confused with the increment to the base flow of these streams within the county, which would include the ground water contribution to their upland tributaries as well; 58 Mgal/d (2.54 m^3/s) represents only the direct ground- water seepage into the major stream channels within the county. The average industrial and municipal pumpage from the aquifer system in 1973 was approximately 35 Mgal/d (1.53 m^3/s).

The recharge, seepage, and pumpage figures given above represent the major components of the water budget for the aquifer system within the county. Minor components of ground-water inflow and outflow from the county are also present.

The analog results indicate that 59 Mgal/d (2.58 m^3/s) of additional pumpage beyond that already developed could be developed from the aquifers in Marion County without producing significant problems of a hydraulic nature. Of this amount, 57.5 Mgal/d (2.52 m^3/s) would be developed from the glacial- outwash aquifer. If the hydraulic characteristics of the stream-aquifer connection, as determined in the steady-state analysis, are substantially correct, this is close to the limit which could be developed using the well configuration and pumpage distribution employed in the experiments. Withdrawal of 72 Mgal/d (3.15 m^3/s), for example, could cause some problems of dewatering (excessive drawdown) in areas of present pumpage. However, assuming that the necessary water management practices could be implemented in the field, it is possible that through further model experimentation, more efficient well configurations and pumping distributions could be achieved that would permit greater withdrawals.

If the hydraulic nature of the stream-aquifer connection is actually better than determined in the steady-state analysis, withdrawals of as much as 103 Mgal/d (4.51 m^3/s) might be possible, again using the well configuration and pumpage distribution employed in the model experiments. Such discharges would induce heavy seepage from Fall Creek, and there is some doubt as to whether the flow of the creek could sustain the losses during the drought. More importantly, however, it is doubtful that the stream-aquifer connection throughout the county is actually good enough to permit such discharges. In any event, the results indicate that the glacial-outwash aquifer is a highly valuable source of water from which at least 57.5 Mgal/d (2.52 m^3/s) of additional pumpage could be developed.

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