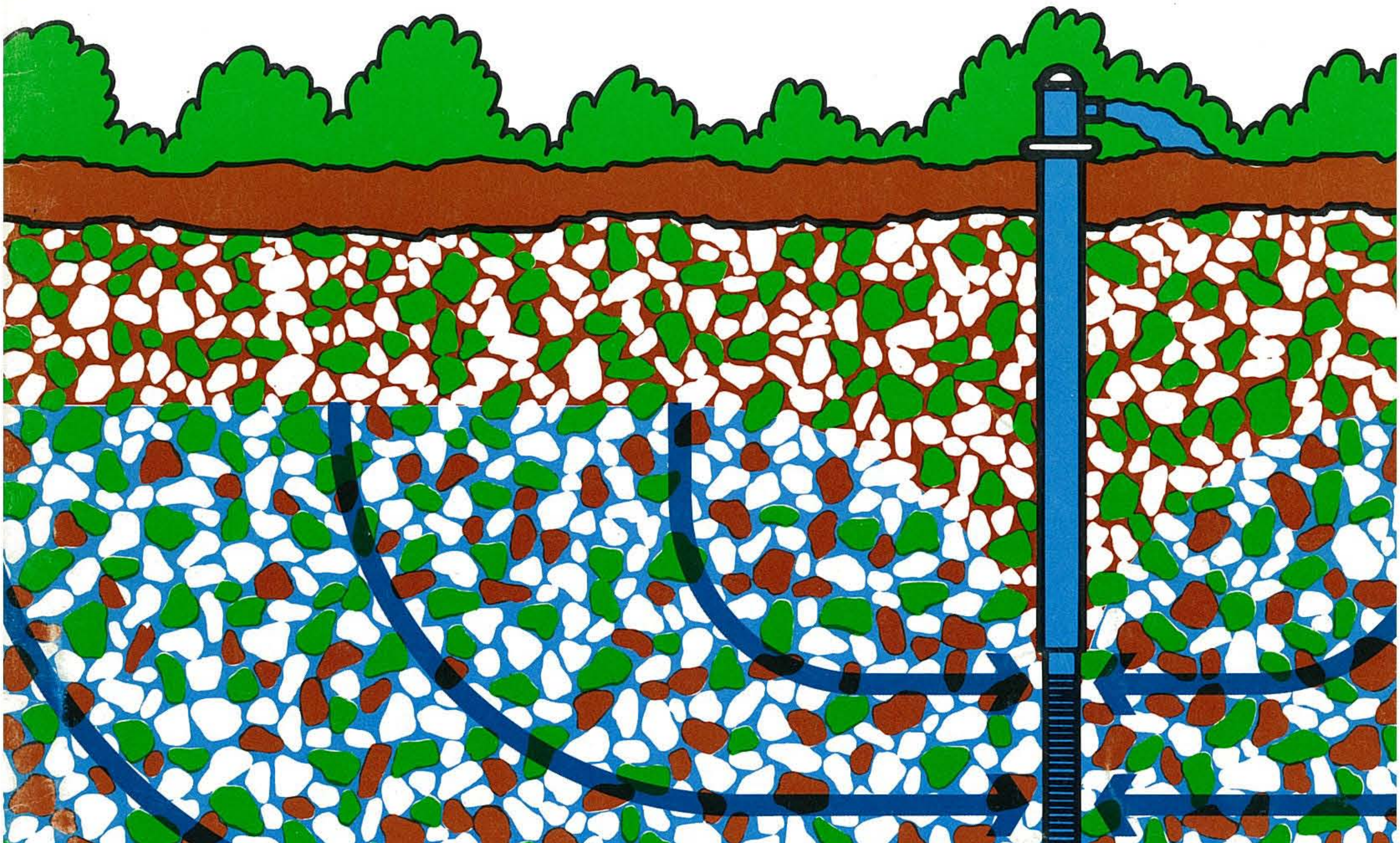
A model analysis of ground-water availability near Carmel, Indiana

U.S. Geological Survey
Water-Resources Investigations
76-46



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16. Abstracts

Flow in the unconsolidated glacial deposits near the city of Carmel in central Indiana was simulated by a digital-computer model in a study of hydraulic characteristics of the deposits. The study shows that 21.3 million gallons per day (933 litres per second) of additional water could be withdrawn from the aquifer for an indefinite period of time. This pumpage is approximately 5 million gallons per day (219 litres per second) above the projected water needs of Carmel for 1990. Saturated thickness, transmissivity, and storage coefficient of the outwash aquifer along the White River east of Carmel were determined, using available data supplemented by test drilling. The saturated thickness of the aquifer ranges from 10 to 110 feet (3 to 34 metres); transmissivity ranges from 1,000 feet squared per day (93 metres squared per day) to 24,000 feet squared per day (2,230 metres squared per day); and the average storage coefficient is 0.11.

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AVAILABILITY OF GROUND WATER

NEAR CARMEL, HAMILTON

COUNTY, INDIANA

By D. C. Gillies

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations 76-46

Prepared in cooperation with

Indiana Department of Natural Resources

Division of Water, and

the city of Carmel

May 1976

UNITED STATES DEPARTMENT OF THE INTERIOR

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GEOLOGICAL SURVEY

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SYSTEM (SI) UNITS

English units	Multiplied by	To obtain SI units
	Length	
inches (in)	25.4	millimetres (mm)
feet (ft)	.3048	metres (m)
miles (mi)	1.609	kilometres (km)
	Area	
square miles (mi^2)	2.590	square kilometres (km^2)
	Flow	
cubic feet per second (ft^3/s)	28.32	litres per second (l/s)
gallons per minute (gal/min)	.06309	litres per second (l/s)
million gallons per day (Mgal/d)	43.81	litres per second (l/s)
gallons per day per foot [(gal/d)/ft]	.00014	litres per second per metre [(l/s)/m)]
	Hydraulic Units	
transmissivity, feet squared per day (ft^2/d)	.0929	metres squared per day (m^2/d)
hydraulic conductivity, feet per day (ft/d)	.3048	metres per day (m/d)
feet per mile (ft/mi)	.1894	metres per kilometre (m/km)

AVAILABILITY OF GROUND WATER NEAR CARMEL,

HAMILTON COUNTY, INDIANA

by D. C. Gillies

ABSTRACT

A study of the hydraulic characteristics of the unconsolidated glacial deposits near the city of Carmel in central Indiana shows that 21.3 million gallons per day (933 litres per second) of additional water could be withdrawn from the aquifer for an indefinite period of time. This pumpage is approximately 5 million gallons per day (219 litres per second) above the projected water needs of Carmel for 1990. Saturated thickness, transmissivity, and storage coefficient of the outwash aquifer along the White River east of Carmel were determined, using available data supplemented by test drilling. The saturated thickness of the aquifer ranges from 10 to 110 feet (3 to 34 metres); transmissivity ranges from 1,000 feet squared per day (93 metres squared per day) to 24,000 feet squared per day (2,230 metres squared per day); and the average storage coefficient is 0.11. Seepage from the aquifer into the White River was estimated in November 1974, using data from U.S. Geological Survey gaging stations. Water-level information was obtained from a network of observation wells at that same time.

Flow in the aquifer was simulated with a digital-computer model. The model was used to estimate the rate of withdrawal that might be sustained from the aquifer and the effect of that withdrawal. The predicted average reduction of the flow of the White River is 21.3 million gallons per day (933 litres per second) in the study area. To develop the fullest potential of the aquifer and to maintain reasonable pumping rates, it would be advisable to locate pumping wells as near to the river as possible and where the aquifer saturated thickness equals or exceeds 50 feet (15 metres).

INTRODUCTION

The city of Carmel, Ind., is a rapidly growing suburban community in Hamilton County, about 10 mi (16 km) north of Indianapolis. An expanding population has accelerated demand on the city's public water supply. The projected population growth of Carmel, from 13,500 in 1975 to 90,000 by 1990, is expected to increase the average daily demand for water from 2 Mgal/d (90 l/s) in 1975 to 16 Mgal/d (700 l/s) in 1990. The purpose of this study was to investigate the availability of recoverable ground water near Carmel and the effects of increased ground-water pumpage on the ground-water system and on streamflow. The study was made by the U.S. Geological Survey in cooperation with the Indiana Department of Natural Resources and the city of Carmel for use in water management and planning.

Preliminary data indicate that the unconsolidated sand and gravel deposits along the White River, approximately 3 mi (5 km) east of Carmel, are potentially capable of meeting the city's projected water needs. These deposits occur along virtually the entire length of the White River in Indiana, although their extent, thickness, and lithologic consistency are variable. The part of these deposits initially chosen for study is a 2-mi (3-km)-wide band along the White River between the Marion-Hamilton County line (96th Street) and 146th Street to the north, an area of about 10 mi^2 (26 km^2). Location of the study area is indicated in figure 1. Prior to the study, only the approximate areal extent of the sand and gravel deposits was known. Consequently, the study involved determining both the horizontal and vertical extent of the deposits and their ability to transmit and store water. Other hydrologic factors of concern included: present water levels in the aquifer and their seasonal fluctuation; discharge from the aquifer, including pumping from wells; and sources of potential induced recharge to the aquifer resulting from increased pumping.

After the appropriate hydrologic data had been collected, flow in the sand and gravel deposits was simulated by a digital model. The model was used to evaluate selected pumping programs. This report describes the acquisition and interpretation of hydrologic data used to design the model and presents results of model analysis.

GEOLOGY

Glacial Drift

The area of interest has undergone continental glaciation several times since the beginning of the Pleistocene Epoch, as well as subsequent erosion and reworking of glacial deposits by melt-water streams as the glaciers retreated. As a result, a mantle of glacial drift, ranging from 20 to 120 ft (6 to 37 m) thick, covers the bedrock in the area. The glacial drift

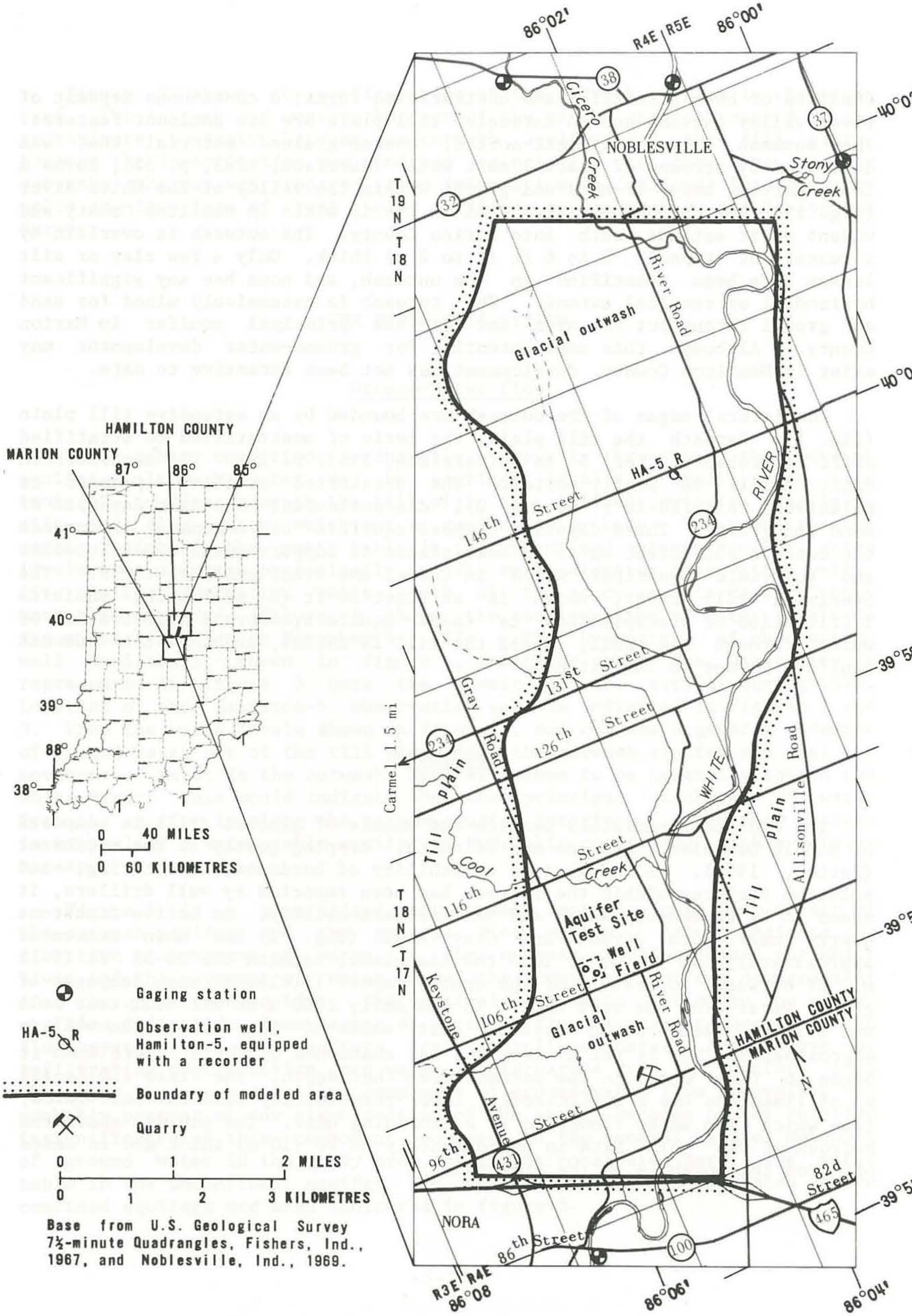


Figure 1.-- Location of study area.

consists of both stratified and unstratified forms; a continuous deposit of river-valley outwash and an extensive till plain are its dominant features. The outwash, a generally well-sorted, coarse-grained material that was deposited by streams of glacial melt water (Harrison, 1963, p. 52), forms a long, narrow body of sand and gravel within the valley of the White River (fig. 1). The outwash averages 2 mi (3 km) in width in Hamilton County and widens as it extends south into Marion County. The outwash is overlain by a veneer of topsoil 3 to 6 ft (1 to 2 m) thick. Only a few clay or silt lenses have been identified in the outwash, and none has any significant horizontal or vertical extent. The outwash is extensively mined for sand and gravel throughout the area and is the principal aquifer in Marion County. Although this same potential for ground-water development may exist in Hamilton County, development has not been extensive to date.

The lateral edges of the outwash are bounded by an extensive till plain (fig. 1). Beneath the till plain, the ratio of unstratified to stratified drift is probably 4 or 5 to 1 (Harrison, 1963, p. 19). The unstratified drift (till) is poorly sorted. The stratified drift is present as relatively thin (10-20 ft or 3-6 m), discontinuous, sheetlike deposits of sand and gravel. These deposits compose aquifers of secondary importance but contain sufficient water in most places to support small domestic wells and low-yield municipal wells in Carmel and other smaller towns. The overlying till layer, which is at least 20 ft (6 m) thick, inhibits infiltration of precipitation to these aquifers, whereas infiltration of water through the soil, where the till is absent, and into the outwash aquifer, is not so inhibited.

Bedrock

The bedrock immediately beneath the mantle of glacial drift is composed of Middle Devonian limestone and dolomite, dipping gently to the southwest (Patton, 1956). Although some variability of hardness, fracturing, and solution features within the bedrock has been reported by well drillers, it seems to be reasonably solid and has low permeability. An active limestone quarry near 96th Street and Gray Road (fig. 1) has been excavated approximately 140 ft (43 m) into the limestone, beneath the 30-35 ft (9-11 m) of outwash. Operators of the quarry report little, if any, seepage of ground water from the rock wall. In the early 1900's an oil well test hole was drilled along 116th Street, just east of Gray Road. The well is approximately 200 ft (61 m) deep and was abandoned by the drillers when it began to flow water to the surface from that depth. The first 170 ft (52 m) of limestone has a sufficiently lower permeability than the unit below, from which the water flows, to be a confining unit. The outwash above the bedrock at the well site is approximately 30 ft (10 m) thick and is cased off from the borehole in the limestone.

Figure 2 shows the configuration of the bedrock surface in the study area. The eastern and western borders of the mapped area correspond to the approximate boundary between the outwash aquifer and the till plain, because the investigation was confined principally to the area of the outwash. The map was prepared from information in drillers' logs of water wells and from logs of test holes drilled to bedrock by the Geological Survey.

THE HYDROLOGIC SYSTEM

Ground-Water Flow

Water-table conditions generally prevail throughout the outwash aquifer, although water may be confined locally in small areas. In contrast, aquifers beneath the till plain are confined, owing to overlying silt and clay layers, which act as semiconfining beds. To establish the configuration of the water table in the unconfined outwash aquifer, water levels were measured periodically in 21 small-diameter observation wells. Five similar observation wells south of the study area in Marion County were also measured. Figure 3 shows the configuration of water levels in the outwash aquifer on November 6, 1974. The hydrograph of observation well Hamilton-5, shown in figure 4, indicates that the water levels represented in figure 3 were the lowest in the aquifer during 1974. Location of the Hamilton-5 observation well is indicated in figures 1 and 3. From the water levels shown in figure 3, one can see a general movement of ground water out of the till plain into the outwash aquifer and that the movement of water in the outwash aquifer seems to be generally toward the White River. This would indicate that the principal discharge of water from the aquifer is along the stream-aquifer interface. This flow pattern is typical of a water-table aquifer in hydraulic connection with a gaining stream.

Figure 5 is a generalized west-to-east section along 126th Street of the outwash aquifer and the general flow system. Arrows indicate the direction of movement of water in the outwash aquifer toward the White River and the movement of water from the confined aquifers downgradient into the unconfined aquifer. The vertical components of flow in the system are also indicated. Some degree of vertical flow is present throughout the flow system, as the aquifers are naturally recharged from above by infiltrating precipitation, and water is discharged from the aquifers into the major streams. A smaller component of ground-water flow downvalley is probably present at any given section of the area. Because of the relative insignificance of this component compared to the general lateral movement of ground water in the study area, it is not considered here. The water table in the unconfined aquifer and the potentiometric surfaces of the confined aquifers are also indicated in figure 5.

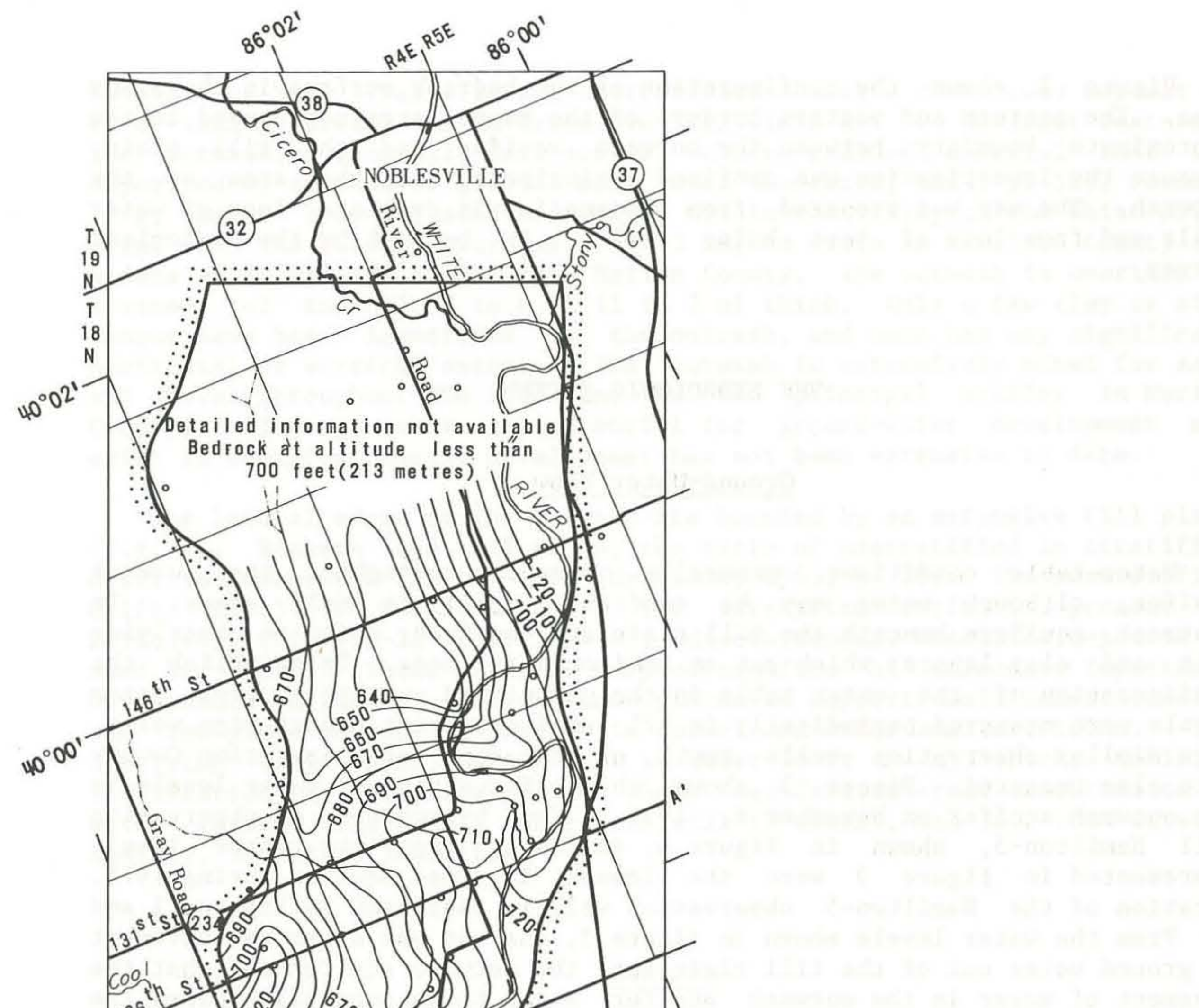
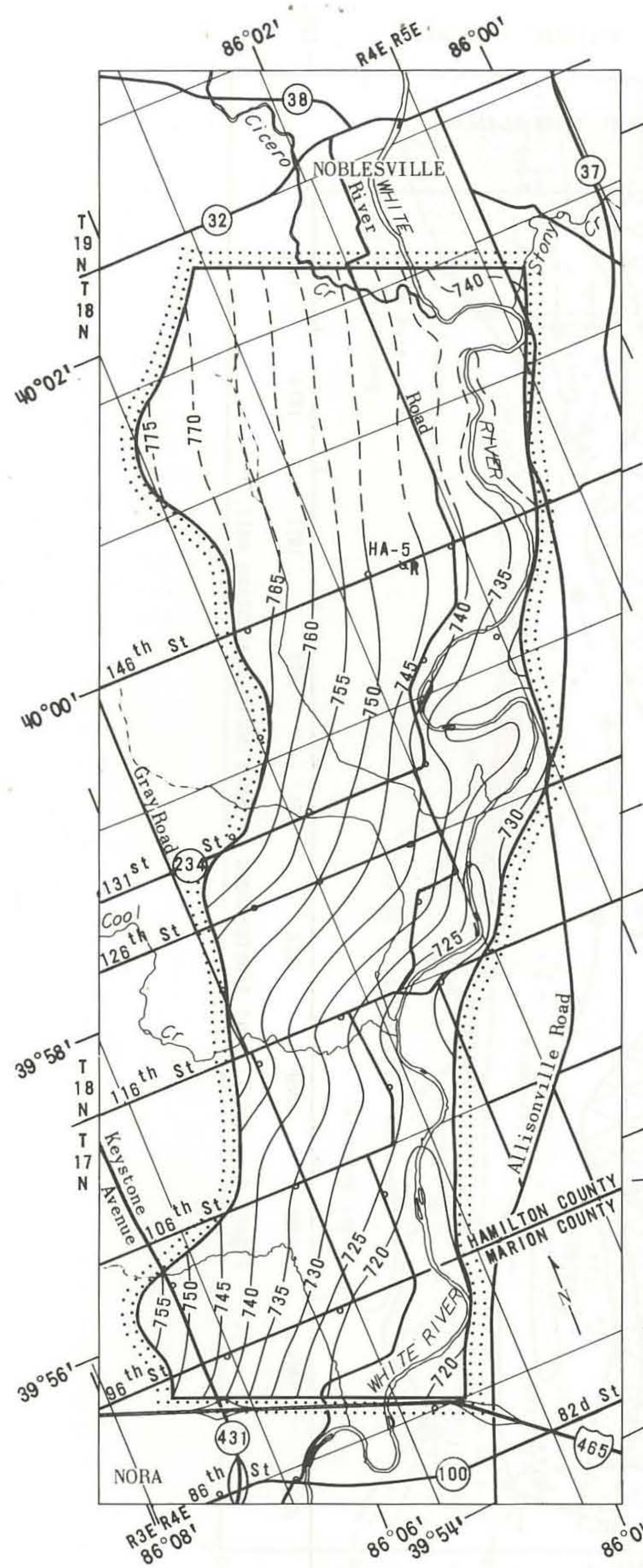


Figure 2.-- Bedrock configuration near Carmel.



EXPLANATION

—750--- Water-table contour. Dashed where approximately located. Contour interval 5 feet (1.5 metres). Datum is mean sea level

o Points of known water-level altitude

HA-5 R Observation well, Hamilton-5, equipped with a recorder

Boundary of modeled area

0 1 2 MILES

0 1 2 3 KILOMETRES

Base from U.S. Geological Survey 7½-minute Quadrangles, Fishers, Ind., 1967, and Noblesville, Ind., 1969.

Figure 3.-- Water-level contours for the outwash aquifer near Carmel, November 6, 1974.

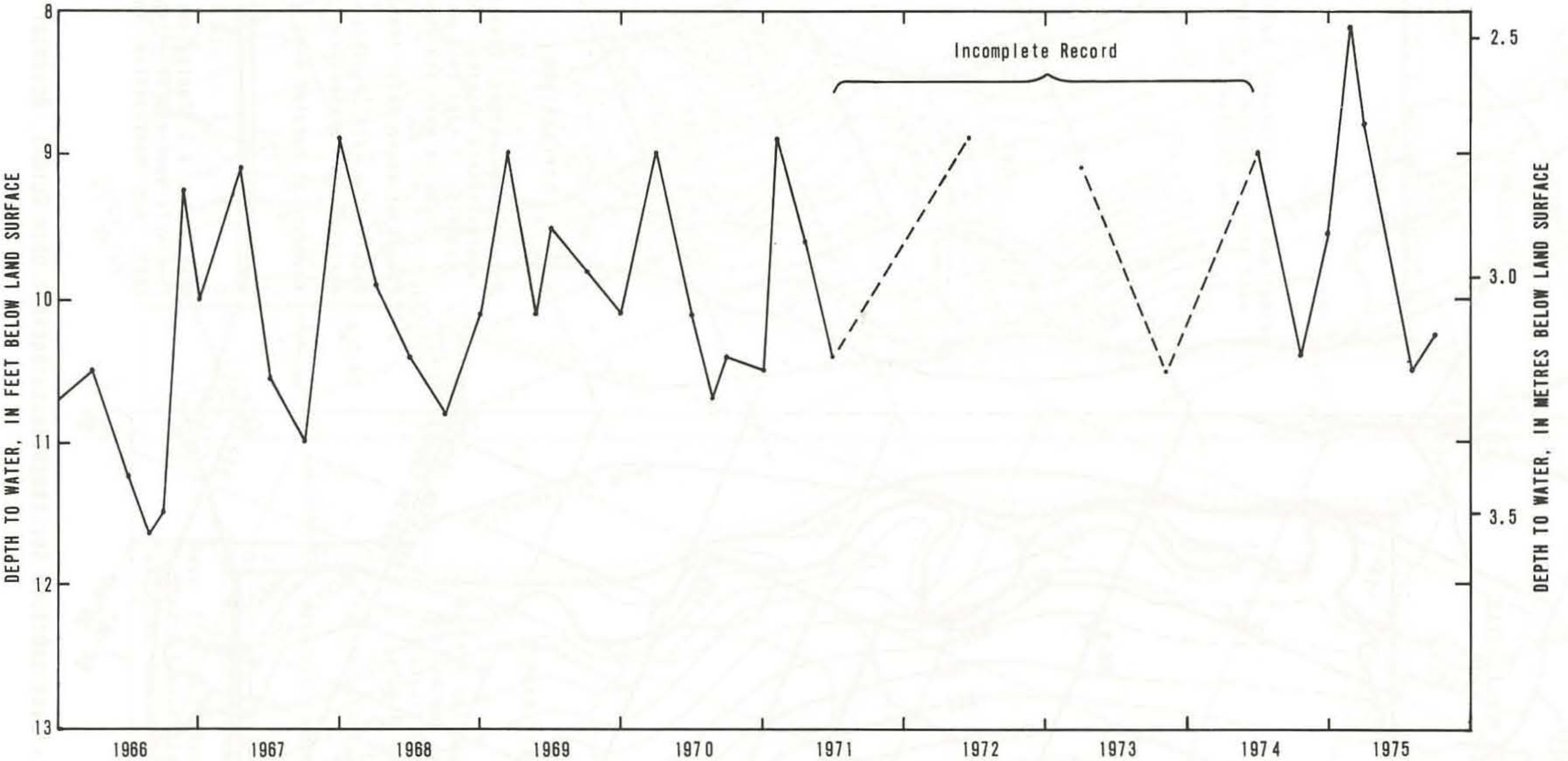


Figure 4.-- Water-level fluctuations in Hamilton-5 observation well.

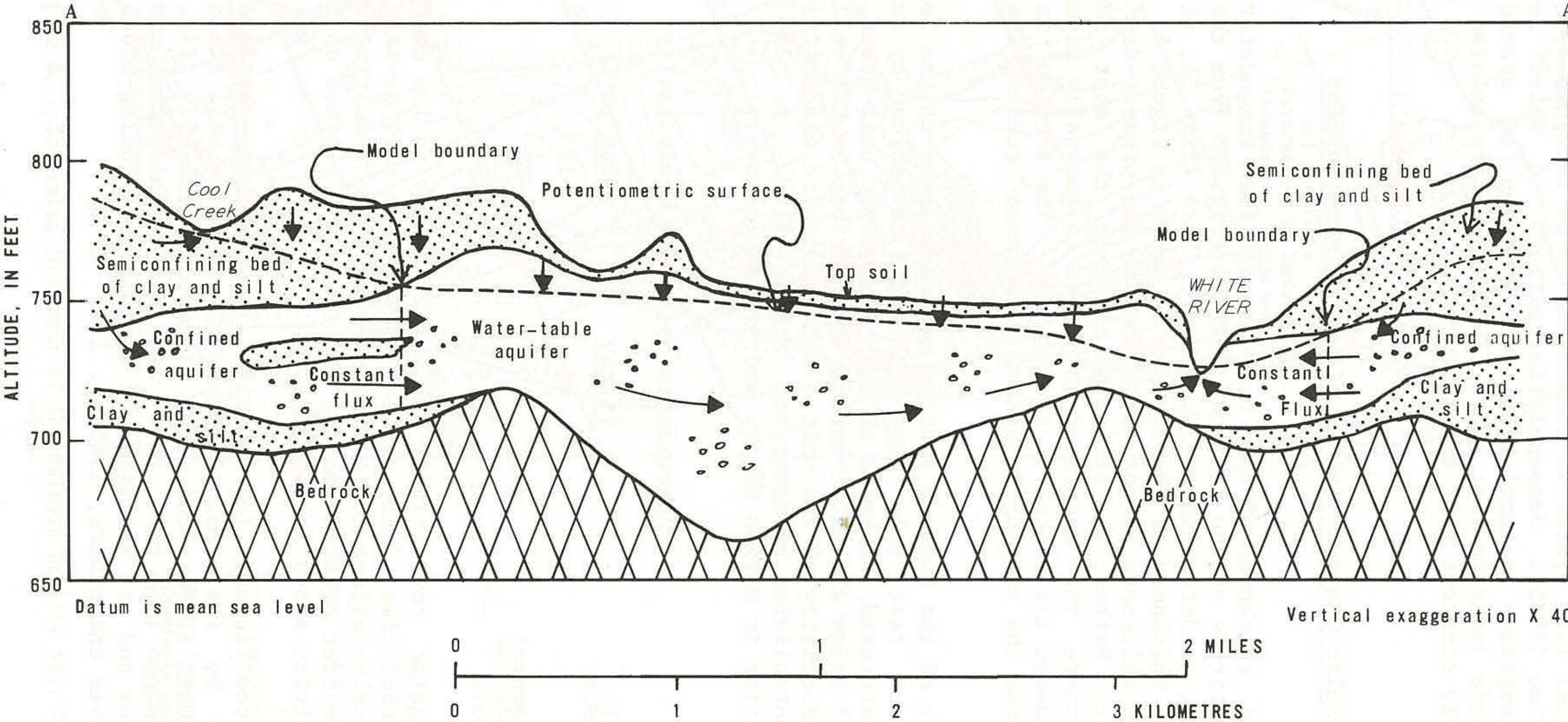


Figure 5.-- Generalized west-to-east section along 126th Street, with arrows indicating direction of ground-water flow.

The ground-water flow system is considered to be in dynamic equilibrium at the present time (1975). Seasonal fluctuations in water levels are accompanied by temporary accumulation or depletion of ground water in storage, but records indicate that these fluctuations are related to a mean that remains nearly constant year after year.

Hydraulic Characteristics of the Outwash Aquifer

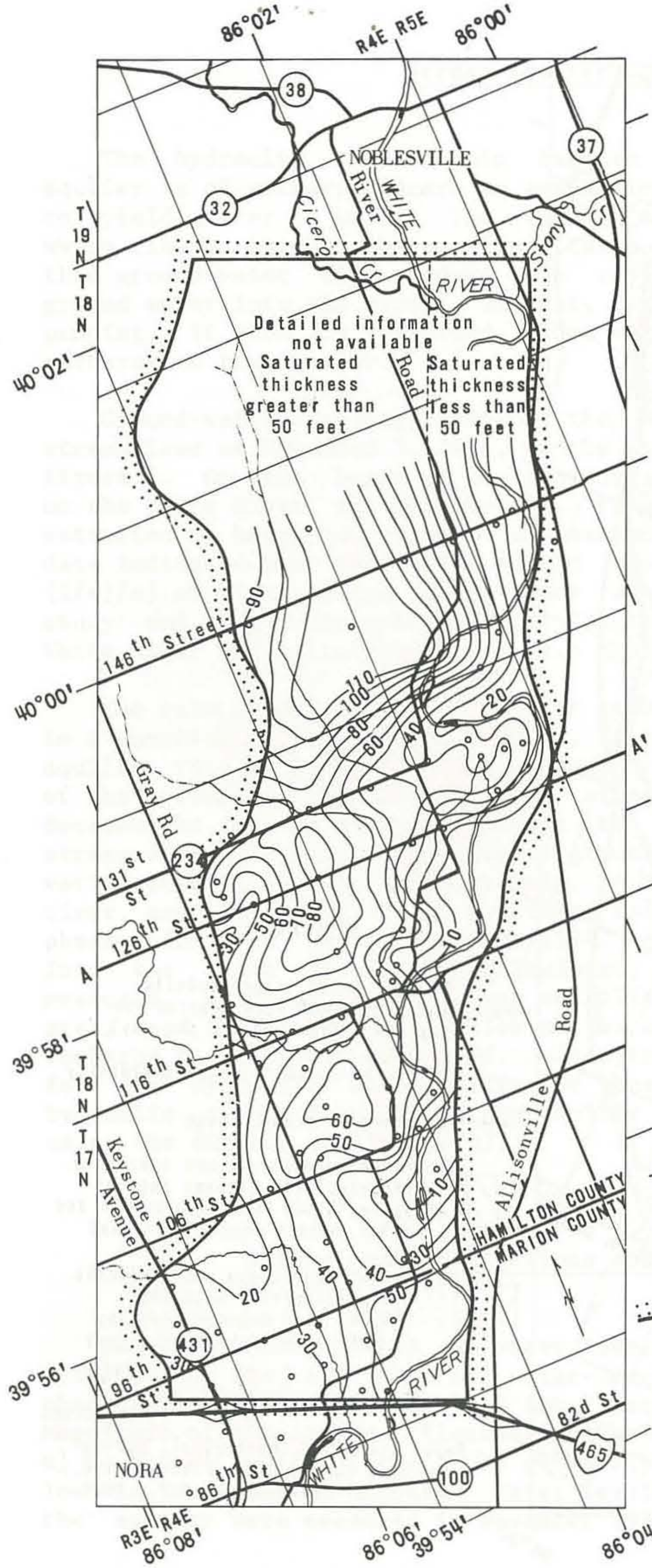
The saturated thickness of the outwash aquifer was determined by subtracting the altitude of the bedrock surface (fig. 2) from the altitude of the water table, as determined on November 6, 1974 (fig. 3), at selected points. Saturated thickness of the aquifer is shown in figure 6. Areas of greatest saturated thickness do not correspond to the present-day location of the White River. Rather, the stream is close to the east edge of the outwash aquifer, where the saturated thickness is generally significantly less than it is toward the center of the aquifer. The section in figure 5 further illustrates the varying saturated thickness across the outwash aquifer.

Transmissivity of the outwash aquifer was calculated at the location of each water well and test hole that penetrated to bedrock by summing the products of the saturated thickness of each lithologic unit (sand, sand and gravel, or gravel) below the water table, as described in the well log, and the hydraulic conductivity of the unit. The hydraulic conductivity values used were those established by Meyer and others (1975, p. 18) during a study of this aquifer in Marion County and are as follows:

	Approximate hydraulic conductivity	
Material	ft/day	m/day
Sand	40	12
Sand and gravel	240	73
Gravel	415	126

Because the hydraulic conductivity of silt or clay is low compared with that of other lithologies, silt or clay logged in a well or test hole was neglected in the calculations of transmissivity. The resulting values for transmissivity, in feet squared per day (ft^2/d), were plotted on the map of aquifer transmissivity shown in figure 7.

The storage coefficient (or specific yield) was obtained from an aquifer test made by the Geological Survey in the well field near the intersection of 106th Street and Gray Road in March 1974 (fig. 1). The average value obtained from this test was 0.11. This value was used for the outwash aquifer and was verified by the electric-analog-model study of Marion County (Meyer and others, 1975, p. 19).



EXPLANATION

—10— Line of equal thickness of saturated material, Nov. 6, 1974 Interval 10 feet(3 metres)

○ Points of known saturated thickness

Boundary of modeled area

0 1 2 MILES

0 1 2 3 KILOMETRES

Base from U.S. Geological Survey 7½-minute Quadrangles, Fishers, Ind., 1967, and Noblesville, Ind., 1969.

Figure 6.-- Saturated thickness of the outwash aquifer near Carmel.

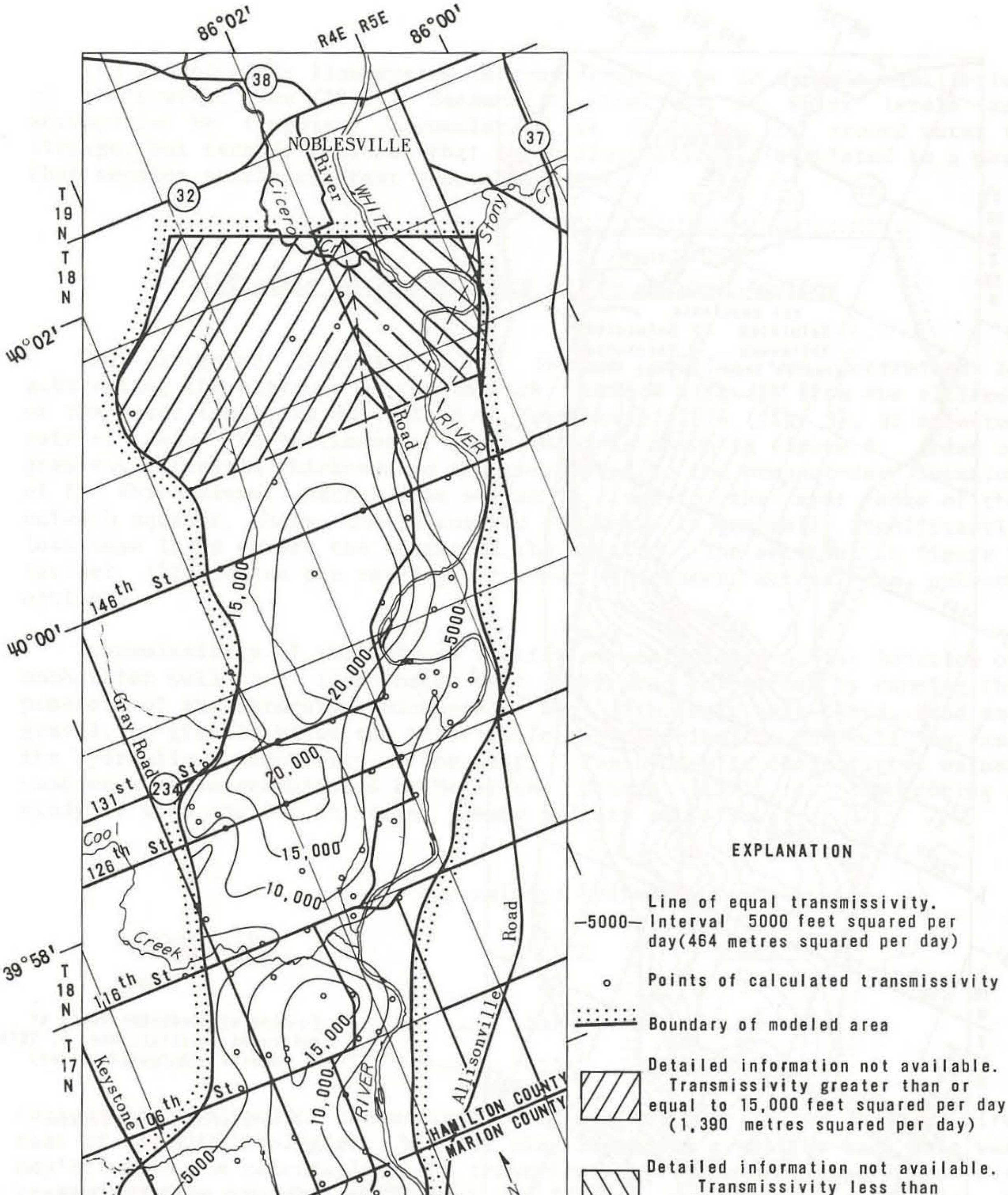


Figure 7.-- Transmissivity of the outwash aquifer near Carmel.

Stream-Aquifer Connection

The hydraulic relationship between the White River and the outwash aquifer is of primary concern in estimating the capability of the aquifer to yield water to wells. The river is the principal discharge area for water flowing through the aquifer. Large-scale pumping will divert part of this ground-water flow toward the river to the pumping wells. Flow of ground water into the stream, in fact, can probably be reversed entirely by pumping. If flow is reversed, the stream will become a major source of recharge to the aquifer.

Ground-water seepage into the White River was estimated from streamflows on November 3, 1974, at the stream-gaging stations indicated in figure 1. On the basis of the flow-duration curve for the gaging station on the White River at Noblesville, flow in the stream at that time was estimated to have been equaled or exceeded 72 percent of the time. These data indicated that the stream gained about 303 (gal/d)/ft [4.36×10^{-2} (l/s)/m] of river length in the study area. For the area modeled in this study and the corresponding river length, ground-water seepage into the White River was calculated to be 19.6 Mgal/d (859 l/s).

The rate at which the White River gains ground water in any given reach is a function of several variables, including hydraulic properties of the aquifer, rate of recharge to the aquifer, vertical hydraulic conductivity of the streambed, thickness of the streambed, and stage of the stream. Because the rate of natural recharge to the aquifer and the stage of the stream vary seasonally, the rate of ground-water seepage to the river also varies seasonally. In this analysis, however, the relationship between the river and the aquifer is described and is simulated on the basis of observations in early November 1974. At that time, ground-water seepage into the White River was calculated, water levels in the aquifer were measured, stage of the river was established along the reach in the study area, and hydraulic properties of the aquifer as well as rate of natural recharge to it were estimated. From these parameters, a range of values for the hydraulic conductivity of the streambed that expressed the hydraulic connection between the aquifer and the stream was synthesized using the digital aquifer model.

Water-Level Fluctuations and Evapotranspiration

Seasonal fluctuations in water levels in the outwash aquifer are evident in the 10 years of water-level records from the Hamilton-5 observation well. (See figure 1 for location and figure 4 for hydrograph.) Magnitude of the seasonal fluctuations at the well ranges from 2.0 ft (0.61 m) to 2.5 ft (0.76 m); the highs generally occur in March or April and the lows in September or October. Water levels in the 26 observation wells in the aquifer were measured in November 1974 and again in April 1975. All these wells had a higher water level in April than in November; average difference between high and low was approximately 2.5 ft (0.76 m). These annual fluctuations in water levels in the aquifer are due to seasonal fluctuations in recharge to the aquifer. However, precipitation is rather evenly distributed through the year because monthly averages do not vary significantly. Average annual precipitation is 37.4 in (942 mm) at Noblesville, approximately 2 mi (3 km) north of the study area. Lowest average monthly precipitation is 2.2 in (56 mm) in February, and the highest average is 4.2 in (110 mm) in May (National Oceanic and Atmospheric Administration, 1974). Fluctuations in recharge to the aquifer are probably not the result of the distribution of precipitation through the year, but, rather, the result of increased evapotranspiration from the unsaturated zone during the warmer months of the growing season and decreased evapotranspiration during the cooler months. Evapotranspiration from the unsaturated zone reduces recharge to the aquifer during the warmer months because it intercepts infiltrating precipitation before it can reach the water table. Most of the evapotranspiration is probably from the unsaturated zone above the water table because the plant types involved use soil moisture almost entirely, and the depths to the water table are generally too great (more than 10 ft or 3 m) for significant evaporation directly from the water table (Meyer and others, 1975, p. 38). The effect of evapotranspiration on the flow system, therefore, is indicated by a reduction of recharge to the aquifer. Lowering of water levels by pumping will probably not recover evapotranspiration losses above the water table (Meyer and others, 1975, p. 38).

AQUIFER SIMULATION AND ANALYSIS BY DIGITAL COMPUTER MODEL

Finite-Difference Models

The finite-difference aquifer model of Trescott and Pinder (1975) was used to simulate movement of water within the outwash aquifer. Because flow in the aquifer is unconfined, the basic flow equation, whose solution was approximated in the model, is given by:

$$\frac{\partial}{\partial x}\left(K_{xx}\, b \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\, b \frac{\partial h}{\partial y}\right) = S_y \frac{\partial h}{\partial t} \pm W_{(x,y,t)}$$

In this expression, K_{xx} and K_{yy} are the principal components of hydraulic conductivity; h is the height of the ground-water level above some arbitrary datum; b is the saturated thickness of the aquifer; S_y is the specific yield of the aquifer; t is time; and x and y are rectangular coordinates. The term, W, includes well discharge, transient leakage from a confining bed, direct recharge from precipitation, and evapotranspiration, and its sign depends on whether water is being added to or withdrawn from the system. This equation can be derived by combining Darcy's law and the principle of conservation of mass.

The finite-difference model program of Trescott and Pinder (1975) is a two-dimensional flow model that simulates ground-water flow in a horizontal plane. Such a simplification may result in an erroneous prediction of water levels in the aquifer simulated by the model if vertical head gradients or flow are significant in the system simulated. The nodal spacing used in the model (667 ft or 203 m) combined with the ratio of horizontal to vertical permeability of 10 to 1, established for the outwash aquifer during the aquifer test at 106th Street and Gray Road (fig. 1), and the saturated thickness of the aquifer, suggest that no serious errors should be introduced in the model from this simplification.

Other simplifications and assumptions that were made during simulation of the aquifer and its flow system are as follows:

1. The specific yield (or storage coefficient) of the unconfined aquifer is uniform throughout. The value 0.11 was used in all simulations.

2. The rate of recharge to the aquifer from precipitation is assumed to be uniform in distribution over the modeled aquifer and remains constant through time.

3. The rate of flow into the unconfined aquifer from the confined aquifers flanking it does not change in time or in response to any stress placed upon the model.

4. All pumping wells simulated are screened through the entire saturated thickness of the aquifer and incur no well loss when pumped.

The steady-state and transient analyses described were obtained by applying the strongly implicit procedure, which is one of the three equation-solving schemes available as options in the Trescott and Pinder model program.

Description of the Model

Figure 1 indicates that the area of aquifer modeled is somewhat larger than the study area, as originally defined. The modeled area was extended north of 146th Street and south of 96th Street so that the effects of pumping along the northern and southern boundaries of the model would be minimal. However, the eastern and western boundaries of the model remained those that define the study area and coincide with the lateral edges of the unconfined outwash aquifer. The total area modeled is 19.4 mi^2 (50.3 km^2). In several of the figures depicting bounds and characteristics of the water-table aquifer (figs. 2, 3, 6, and 7), aquifer properties had to be estimated, as the modeled area was extended north of 146th Street, because only a small amount of data is available in this area. Information used to extend the modeled area south of 96th Street was obtained from the ground-water availability study recently completed in Marion County (Meyer and others, 1975).

To place the data in a form compatible with the model, the modeled aquifer was divided into small rectangles in map view, or into small blocks in three dimensions through use of a finite difference grid. Each block has the volume $b\Delta x\Delta y$, where b is saturated thickness and Δx and Δy are the grid spacings in the x and y directions, respectively, on a coordinate axis. The center of each block is called a node. At these nodes, data representing applicable aquifer properties for that block are fed into the model. Also at these nodes, water levels are calculated by the model program and are then printed out for interpretation. Because of conditions in the aquifer and for simplicity, the modeled outwash aquifer was divided by a square-grid network into 1,215 equally sized nodal areas. Each side of the square was approximately equal to 667 ft (203 m) on the ground. The model contains 70 nodes north-south and 28 nodes east-west; its shape approximates that of the aquifer.

The square-grid network was positioned over the plots of bedrock altitude (fig. 2), water-level altitude (fig. 3), saturated thickness (fig. 6), and transmissivity (fig. 7), and an appropriate value for each property was assigned to each node in the model. In the simulation of a water-table aquifer, transmissivity at each node is recomputed as the water level changes to account for dewatering of the aquifer as a result of pumping. The model computes transmissivity as the product of saturated thickness and aquifer hydraulic conductivity, and, thus, the model must be fed these two parameters initially. The values of hydraulic conductivity for each node that were fed into the model were obtained by dividing the transmissivity (fig. 7) by the saturated thickness (fig. 6) at each node. As a result, hydraulic conductivity varies from node to node throughout the model, a condition of nonhomogeneity.

The White River was simulated in the model by identifying grid nodes that most nearly represented the course of the river and assigning appropriate values to appropriate parameters. The river was modeled by assigning a uniform gradient of 1.8 ft/mi (0.3 m/km), a uniform depth of 2 ft or 0.6 m (representing annual low-flow in November 1974), and a uniform streambed thickness of 1 ft (0.3 m). The hydraulic connection of the river with the aquifer was represented by assigning a hydraulic conductivity to the streambed, as discussed in the section, "Model Calibration and Steady-State Verification."

Boundary Conditions

In any aquifer model, boundaries must be accurately simulated or must be far enough away from simulated pumping stresses so that they will not be significantly affected by those stresses. In this particular study, available data did not permit accurate simulation of the confined aquifers flanking the outwash aquifer. An effort was made, however, to simulate minimum flow between the confined aquifers and the unconfined aquifer. This minimum flow across the east and west edges of and into the outwash aquifer was simulated as a constant (in time) flux at each node along the boundary of the model. Assignment of flux values along the constant-flux boundaries is discussed in the section, "Model Calibration and Steady-State Verification." Locations of the constant-flux boundaries are indicated on figures 8, 9, and 10. The relationship of the constant flux to the general flow system is also illustrated in the section in figure 5.

Northern and southern boundaries of the model were treated differently than the eastern and western boundaries. An important characteristic of the water-level map (fig. 3) is that the water-level contours meet the northern and southern boundaries of the modeled area at nearly right angles, which indicates that little ground water is flowing across these boundaries. Therefore, during both the steady-state and the transient analyses, all the northern boundary of the model and a part of the southern boundary were treated as impermeable. The remainder of the southern boundary was treated as a constant head. The constant-head boundary was aligned with the 718-ft (219-m) contour line in the outwash aquifer. (Location of the constant-head boundary is shown in figures 8, 9, and 10.)

Treatment of the model's boundaries in the preceding manner during the transient part of the model analysis resulted in a conservative estimate of ground-water availability. In the real system, the spreading cone of depression resulting from ground-water pumpage equal to that simulated would induce the movement of ground water toward the pumping centers from beyond the area of the aquifer where a given model boundary was imposed once the cone of depression had reached that boundary or area. Modeling the confined-unconfined boundaries as a constant flux and the northern and southern boundaries of the outwash as impermeable during the transient analysis did not permit such movement. The part of the southern boundary of the model simulated as a constant-head could allow an unlimited flow across this boundary and into the modeled aquifer. However, examination of the constant-head boundary after the transient simulation showed that the net effect was only a small reduction of ground-water outflow from the boundary.

The final boundary condition that had to be considered in the model was the hydraulic relationship between the sand and gravel outwash aquifer and the bedrock limestone beneath it. Because of its low permeability and its probable minor contribution to the total flow in the real ground-water system, the bedrock was not considered in the model analysis. Instead, the bedrock was represented in the model as the impermeable base of the unconfined outwash aquifer.

Model Calibration and Steady-State Verification

Before a model can be used to determine the availability of water from an aquifer, it must be capable of simulating flow in the aquifer to an acceptable degree. Acceptability of this model was established by comparing real water levels in the outwash aquifer, as shown in figure 3, and the real rate of ground-water seepage into the White River in November 1974 with the same parameters determined by the model.

In the Marion County study, Meyer and others (1975, p. 45) closely approximated April water levels in the outwash aquifer with an electric-analog model; simulated recharge to the aquifer was 13.5 in (343 mm) per year. Because November (minimum) water levels were being approximated in the Carmel study and because seasonal water-level fluctuations in the aquifer were about 2.5 ft (0.75 m), recharge of less than 13.5 in (343 mm) per year was simulated. In fact, several lesser recharge rates were simulated, and flow across the constant-flux boundaries was adjusted appropriately in each case. A constant rate of ground-water seepage into the White River was simulated. Simulated recharge that acceptably approximated November 1974 water levels in the aquifer and that permitted calculation of a reasonable rate of recharge to the confined aquifers underlying the till plain (simulated in the model as constant flux) was 11.9 in (302 mm) per year. To approximate real water levels in the outwash aquifer for a given rate of simulated flow across the till-outwash boundary, distribution of simulated flow was varied along the boundary. The hydraulic connection between the outwash aquifer and the White River was initially set equal to the vertical hydraulic conductivity of the outwash aquifer (31.4 ft/d or 9.6 m/d), as established by the aquifer test at 106th Street and Gray Road. Altitudes at stream nodes were set equal to values measured during November 1974. With these constraints, real water levels in all but one small area of the outwash aquifer (cross-hatched area in fig. 8) were closely approximated, as was ground-water seepage into the White River.

Water levels in the cross-hatched area of the outwash aquifer, shown in figure 8, were as much as 8 ft (2.4 m) too low (below measured levels) at the end of calibration. Successive adjustments of the hydraulic connection between the aquifer and the stream in this area permitted water levels to be sufficiently approximated by assigning a reduced hydraulic conductivity of 0.07 ft/d, (0.02 m/d) to the modeled streambed.

After water levels in the aquifer and ground-water seepage into the White River had been approximated, a series of simulations was made to calibrate the model with respect to hydraulic conductivity of the streambed. Use of a wide range of values of streambed hydraulic conductivity (0.67-67 ft/d or 0.2-20 m/d) had little effect on either the simulated rate of ground-water seepage into the stream or simulated water levels in the outwash aquifer near the river. Therefore, the lowest overall value of streambed hydraulic conductivity that would yield an acceptable approximation of water levels in the aquifer and that would yield the correct rate of ground-water seepage was used in the model.

Excluding that part of the river in the cross-hatched area of figure 8, the entire reach of the White River through the modeled area was represented with a streambed hydraulic conductivity of 0.67 ft/d (0.20 m/d); values smaller than this produced modeled water levels near the river that were too high to be acceptable. The simulated hydraulic conductivity of the streambed through the cross-hatched area in figure 8 remained 0.07 ft/d (0.02 m/d). Overall, this approach assured that the model would be conservative and that subsequent estimates of pumpage possible in the real aquifer system would be conservative as well because of the rather restrictive stream-aquifer connection simulated in the model.

Fifty-two percent of recharge to the outwash aquifer in the steady-state solution was represented by the constant flux across the confined-unconfined boundaries. This flux is 11.7 Mgal/d (513 l/s) and can be considered to be the recharge from precipitation to the confined aquifers under the till plain. This recharge subsequently moves downgradient and into the unconfined aquifer. Recharge per unit area to the confined aquifers beneath the till plain was estimated to be 3.7 in (94 mm) per year if ground-water divides correspond to surface-water divides. This rate of recharge was within the range of values found to be most satisfactory in the electric-analog-model study of Marion County, where the confined aquifers flanking the unconfined aquifer were modeled in detail (Meyer and others, 1975, p. 48). The remaining 48 percent of recharge to the outwash aquifer represented recharge from precipitation to the unconfined aquifer. This amounted to 10.87 Mgal/d (476.2 l/s) or 11.9 in (302 mm) per year.

Average ground-water runoff (or seepage) for subbasins within the upper White River basin has been previously estimated by a technique of stream-hydrograph separation (Cable and others, 1971, p. 18). This previous study also assumed that surface-drainage divides and ground-water divides correspond. In the subbasin that includes the area modeled in this study, analysis of hydrographs of the gaging stations (locations indicated in fig. 1) indicated a value for recharge per unit area equal to 5.64 in (143 mm) per year. This rate can be considered the integrated recharge to the aquifers underlying the subbasin, both confined and unconfined. By combining the two rates of recharge used in the model and applying the appropriate areas involved, integrated recharge for the area of the subbasin represented by the model was calculated to be 4.76 in (121 mm) per year. This integrated recharge rate represents conditions in the subbasin in early November 1974, when water levels in the aquifer and ground-water seepage into the White River were at their annual lows. The recharge rate is approximately 84 percent of the rate reported by Cable and others (1971, p. 18) as representing "average" conditions in the subbasin.

Figure 8 shows the steady-state solution that represents the best approximation achieved by the model of the water levels in the aquifer in November 1974. The map shown here was contoured directly from the plot of water levels generated by the model. The points are shown where water-level measurements were obtained in the aquifer and where direct comparison of observed and modeled water levels can be made. Locations of historical pumpage of ground water assumed to be in equilibrium with the

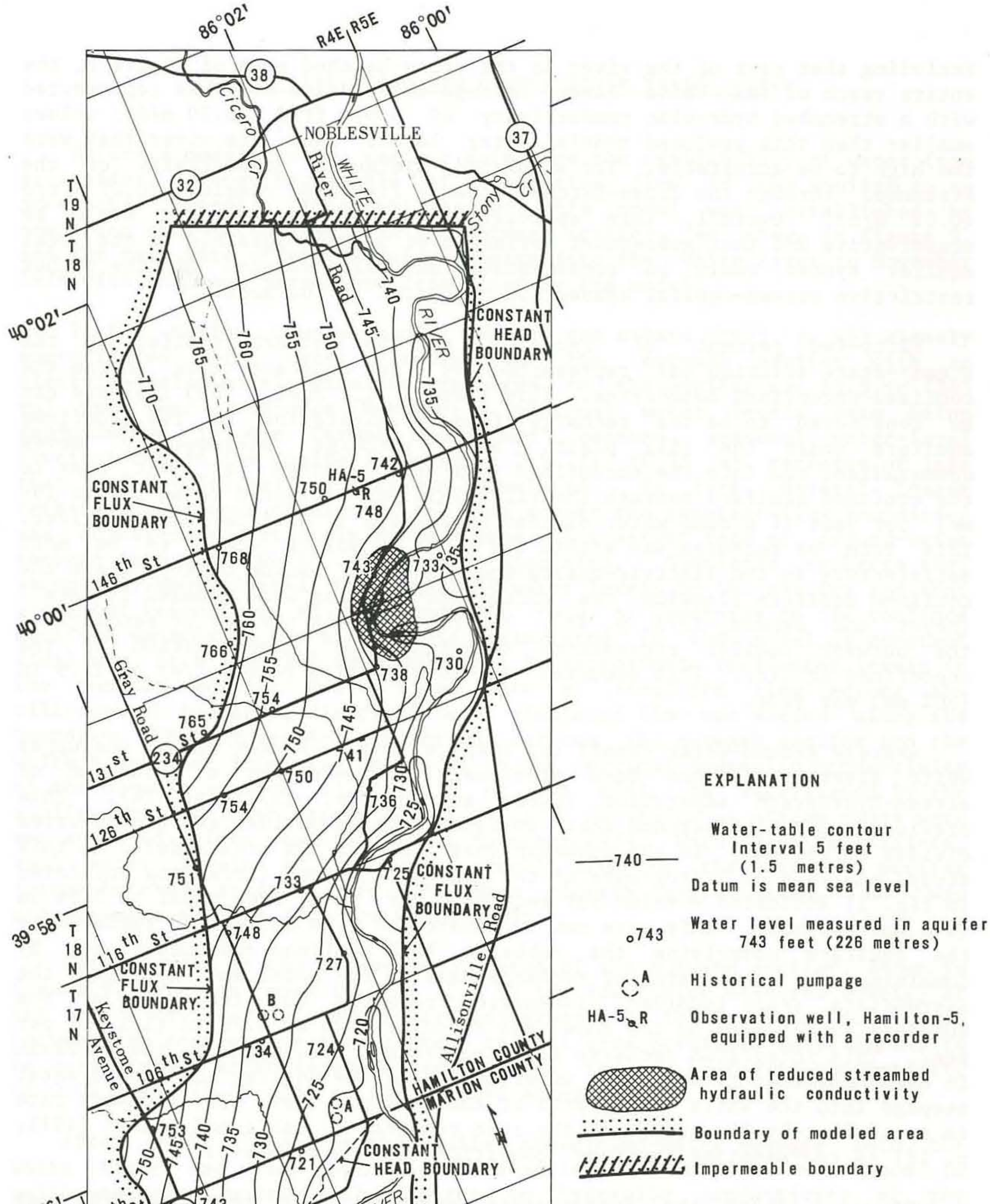


Figure 8.-- Simulated steady-state water levels in the outwash aquifer near Carmel, November 6, 1974.

flow system are also indicated: (A) stone quarry, and (B) well field. Pumping rates at these two sites are 1 Mgal/d (44 l/s) and 0.5 Mgal/d (22 l/s), respectively.

Table 1 summarizes the sources and discharges characterizing the steady-state solution described in the preceding paragraph. This tabulation indicates the magnitude of the various flow components in the water budget for the assumed set of steady-state conditions simulated in the model. However, steady-state conditions do not normally prevail because seasonal fluctuations in recharge as well as long-term cycles of wet and dry years will continuously modify this budget. The significance and utility of these approximations are that they document a conservative basal period with an annual frequency from which consistent hydrologic interpretations can be made. Subsequent model estimates of the long-term yield of the aquifer can then be weighed against the effects of withdrawal of this estimated yield on a set of less than average conditions of water availability.

Table 1.--Steady-state ground-water budget in the outwash aquifer for conditions modeled in November 1974

Sources	(Mgal/d)	(l/s)
Constant-flux boundaries	11.70	512.6
Direct recharge 11.9 in/yr (302 mm/yr)	10.87	476.2
Discharges		
Ground-water seepage	19.79	867.0
Historical pumpage	1.50	65.7
Constant-head boundary	1.28	56.1

Simulated Ground-Water Pumpage

The main objective of this study was to determine if the projected 1990 water needs of Carmel could be obtained from the part of the outwash aquifer studied and to determine the effects of such a large withdrawal on water levels in the aquifer and flow of the White River. In addition, pumpage that might be sustained from the aquifer in the immediate vicinity of the well field at 106th Street and Gray Road (fig. 1) was studied.

For the main objective, simulated well sites were selected on the basis of maximum transmissivity and saturated thickness and proximity to the White River. In addition, the simulated wells were distributed in single lines approximately parallel to the river to minimize mutual interference. Only one well was simulated per node, so that the minimum distance between pumping wells was approximately 667 ft (203 m). In practice, of course, pumpage simulated at a given node in the model could be distributed among more than one well in the nodal area.

Pumping rates in the individual wells were limited by the restriction that the steady-state drawdown reached in a given well did not exceed 50 percent of the prepumping saturated thickness of the aquifer at that site. This drawdown restriction should result in a fairly conservative estimate of the total amount of ground water available for the well distribution modeled. A well radius of 3 ft (0.9 m) was selected for the simulation because most high-yield production wells in the Indianapolis metropolitan area are gravel packed to this radius and are highly developed.

By use of the preceding criteria, pumpage of 21.3 Mgal/d (934 l/s) was obtained from the distribution of pumping in the outwash aquifer shown in figure 9. This simulated pumpage is in addition to the 1.5 Mgal/d (66 l/s) already being pumped (historical pumpage). Pumping rates in individual wells for this program varied from site to site and ranged from 350 to 800 gal/min (22 to 50 l/s), depending on hydrologic conditions, and averaged 587 gal/min (37 l/s). This distribution is the one accepted after a set of trial-pumping simulations, where distribution and rate of pumping were altered in the model. Figure 9 also shows the steady-state water levels that resulted from this simulated pumping; figure 10 shows a map of the steady-state drawdown for the same pumping program. For this simulation, the modeled flow system reached virtual equilibrium at the end of 14.6 years of pumping.

Table 2 lists sources and discharges characterizing the pumping simulation described above at the end of 14.6 years of continuous pumping.

Comparison of table 2 with table 1 (which documents the 1975 flow system) indicates that the additional 21.3 Mgal/d (933 l/s) pumpage is primarily supported by diverting the total amount of ground-water flow that had previously entered the White River within the modeled area. In addition, table 2 indicates that at this virtual equilibrium a net rate of 0.68 Mgal/d (30.0 l/s) is induced from the flow of the White River into the ground-water system.

Once true equilibrium is reached, the small percentage (0.4) of pumpage still being derived from aquifer storage at the end of 14.6 years of pumping would eventually be derived from (1) induced flow from the White River, or (2) reduced underflow of ground water at the southern boundary of the modeled area, or (3) some combination of 1 and 2. Because data concerning the outwash aquifer downstream from the study area indicate that ground-water underflow from the southern boundary of the modeled area eventually enters the White River, one can assume that the total simulated

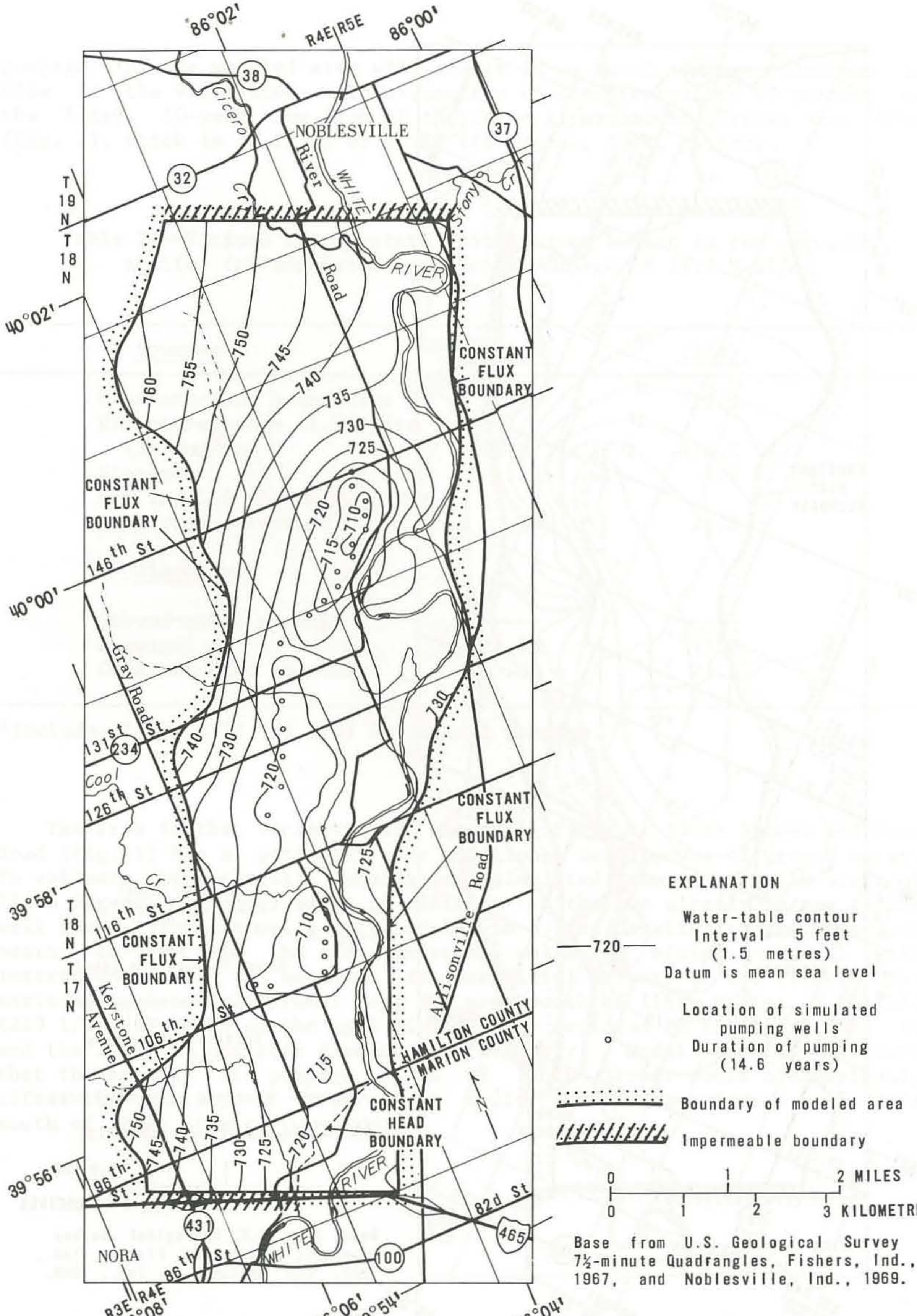


Figure 9.-- Computed steady-state water levels in the outwash aquifer near Carmel for simulated ground-water pumpage equal to 21.3 million gallons per day (934 litres per second).

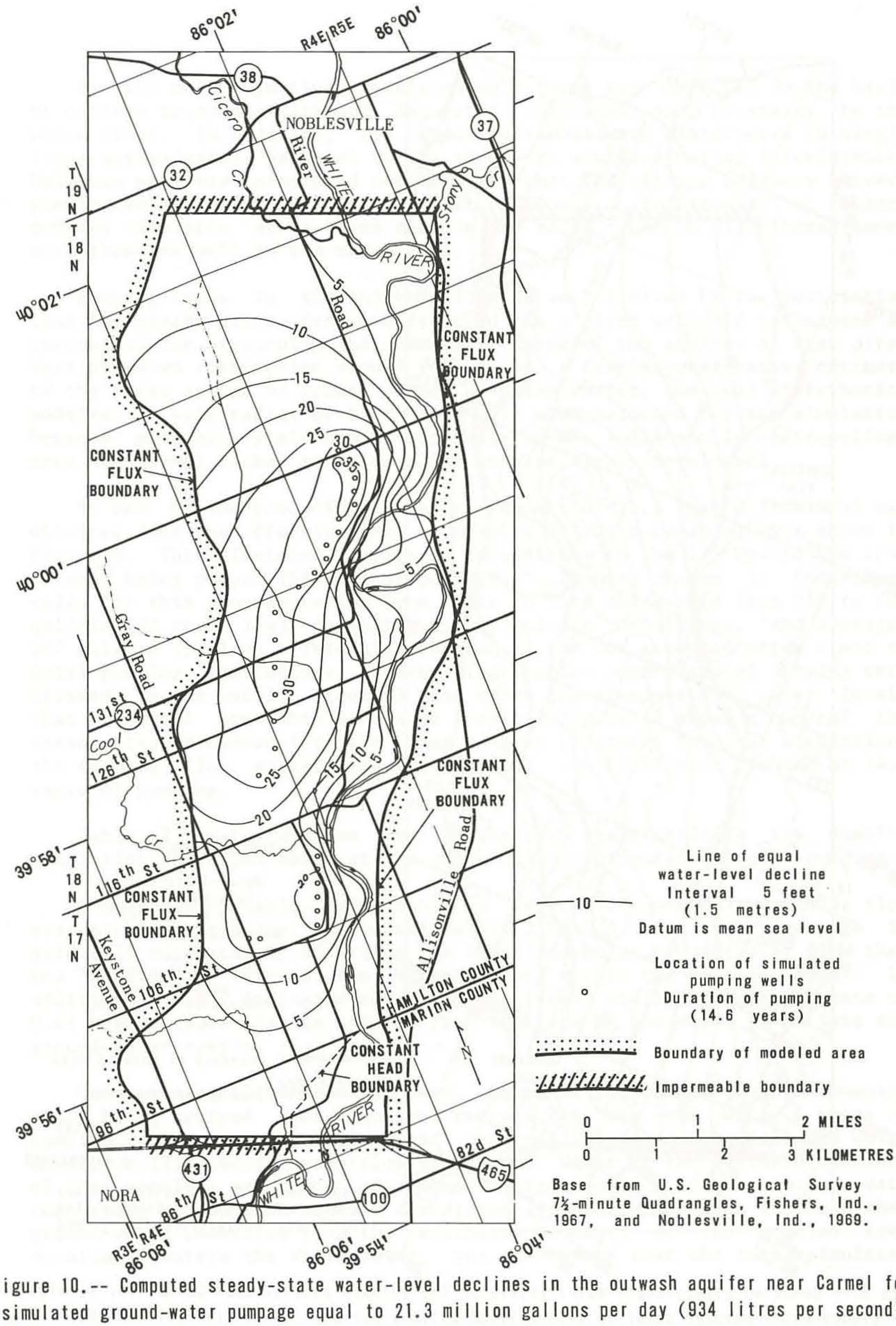


Figure 10.-- Computed steady-state water-level declines in the outwash aquifer near Carmel for simulated ground-water pumpage equal to 21.3 million gallons per day (934 litres per second).

pumpage in the modeled area will result in an equal average decrease in flow in the White River. This quantity of flow represents 47 percent of the 7-day, 10-year low flow of the White River at 86th Street near Nora (fig. 1), which is 70 ft^3/s or 1,980 l/s (Rohne, 1972, p. 165).

Table 2.--Virtual steady-state ground-water budget in the outwash aquifer for simulated additional pumpage of 21.3 Mgal/d

Sources	(Mgal/d)	(l/s)
Constant-flux boundaries	11.70	512.6
Direct recharge 11.9 in/yr (302 mm/yr)	10.87	476.2
Storage	.10	4.4
Net induced flow from river through streambed	.68	29.8
Discharges		
Ground-water seepage	-----	-----
Pumpage[1]	22.81	999.0
Constant-head boundary	.54	23.7

[1]Includes 1.5 Mgal/d (66 l/s) historical pumpage.

The area in the vicinity of the well field at 106th Street and Gray Road (fig. 1) has a potential for additional development of ground water. To estimate the potential, pumping was simulated from eight wells south of 116th Street, six model wells in addition to the two already in use in the well field. The six wells were modeled in a line parallel to the river and nearer to it than the two existing wells to minimize mutual well interference and to maximize the beneficial effects of the river. The maximum simulated withdrawal from the area south of 116th Street, 5 Mgal/d (219 l/s), is based on the well distribution depicted in figures 9 and 10 and the limiting criteria discussed previously. Model results indicate that the effects of pumpage north of 116th Street would not seriously affect the quantity of water that could be withdrawn from a well field south of 116th Street.

CONCLUSIONS

Results of modeling the outwash aquifer along the White River east of Carmel indicate that substantial quantities of water could be withdrawn from the aquifer. The amount is limited to the sum of (1) ground-water seepage to the river within the study area, (2) a small net amount of recharge induced from the river through its streambed by pumping, and (3) underflow diverted by pumping. When the model was stressed by additional pumpage of 21.3 Mgal/d (933 l/s), steady-state conditions were approached, an indication that this pumpage could be sustained indefinitely under the given set of limiting conditions. Pumping rates from individual wells could range from 350 to 800 gal/min (22.1 to 50.5 l/s), if mutual interference of pumping wells were minimized by adequate well spacing. To develop the fullest potential of the aquifer and to maintain reasonable pumping rates, it would be advisable to locate pumping wells as near to the river as possible and where the aquifer saturated thickness equals or exceeds 50 ft (15.2 m). Pumpage of 21.3 Mgal/d (933 l/s) in addition to the 1975 pumpage would reduce flow in the White River by this same amount, 21.3 Mgal/d (933 l/s), in the study area. This is about 47 percent of the 7-day, 10-year low flow of the river at the Survey's gage at 86th Street near Nora. Further verification of model results will be possible if ground water in the area is developed, as anticipated. A properly designed aquifer test between 131st and 146th Streets would also substantially contribute to verification and improvement of the model simulation.

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