Electrical-Analog-Model Study of Water Resources of the Columbus Area, Bartholomew County, Indiana

GEOLOGICAL SURVEY WATER-SUPPLY PAPER 1981

Prepared in cooperation with the State of Indiana Department of Natural Resources

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By F. A. WATKINS, JR., *and* J. E. HEISEL

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CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

ELECTRICAL-ANALOG-MODEL STUDY OF WATER RESOURCES OF THE COLUMBUS AREA, BARTHOLOMEW COUNTY, INDIANA

By F. A. WATKINS, JR., and J. E. HEISEL

ABSTRACT

The Columbus study area is in part of a glacial outwash sand and gravel aquifer that was deposited in a preglacial bedrock valley. The study area extends from the north line of Bartholomew County to the south county line and includes a small part of Jackson County south of Sand Creek and east of the East Fork White River. This report area includes about 100 square miles of the aquifer.

In the Columbus area, ground water in the outwash aquifer is unconfined. Results of pumping tests and estimates derived from specific-capacity data indicate that the average horizontal permeability for this aquifer is about 3,500 gallons per day per square foot. An average coefficient of storage of about 0.2 was determined from pumping tests. Transmissibilities range from near zero in some places along the boundary to about 500,000 gallons per day per foot in the thicker parts of the aquifer. About 800,000 acre-feet of water is in storage in the aquifer. This storage is equivalent to an average yield of 34 million gallons per day for about 21 years without recharge.

An electrical-analog model was built to analyze the aquifer system and determine the effects of development. Analysis of the model indicates that there is more than enough water to meet the estimated needs of the city of Columbus without seriously depleting the aquifer. Additional withdrawals will affect the flow in the Flatrock River, but if the withdrawals are made south of the city, they will not affect the river any more than present pumping. Future pumping should be confined to the deepest part of the outwash aquifer and (or) to the area adjacent to the streams.

On the basis of an hypothesized amount and distribution of pumping, the decline in water levels in the Columbus area as predicted by the model for the period 1970–2015 ranged from about 20 feet in the center of the areas of pumping to 3 feet or less in the areas upstream and downstream from these areas of pumping.

INTRODUCTION

An investigation of the outwash sand and gravel aquifer in the Columbus area was begun in April 1966. The available hydrologic data consisted of drillers' records, periodic measurements of water levels in about 80 wells or other points on the ground-water surface, water levels from six wells equipped with water-stage recorders, and measurements of the discharge of wells and streams. These data were collected as a part of the cooperative water resources program of the U.S. Geological Survey and the Indiana Department of Natural Resources.

PURPOSE AND SCOPE

The city of Columbus, Ind., obtains its water supply of about 6 mgd (million gallons per day) from a well field in the city. Growth predictions indicate the need for 34 mgd by the year 2015 (Hendricks and others, 1965). The purposes of this study are (1) to determine whether that amount of water can be obtained on a sustained basis from ground water and (2) to predict what long-term effect such withdrawals will have on ground-water levels and streamflow within the study area by imposing several hypothetical combinations of well-field locations and pumping rates on the aquifer. The purposes were accomplished through the use of an electrical-analog model.

LOCATION AND EXTENT OF THE AREA

Bartholomew County is in the northwest part of southeast Indiana. This report is concerned with an area of about 100 square miles that ranges in width from 2 miles to more than 7 miles and extends from the north county line to the south county line (fig. 1). It also includes that small area of Jackson County south of Sand Creek which would be enclosed by projecting the south Bartholomew County line eastward about 3 miles from the East Fork White River and the east county line southward about 2 miles from Sand Creek.

PREVIOUS INVESTIGATIONS

The geology and ground-water resources of this area have been discussed in the reports listed chronologically and summarized below:

1881. Elrod, M. N., Geology of Bartholomew County: Indiana Dept. Geology and Nat. History, 11th ann. rept., p. 150–213, 1 map. Describes the geology of the area with a short section on water supply from ground water.

1951. Klaer, F. H., Jr., Davis, G. E., and Kingsbury, T. M., Ground-water resources of the Columbus area, Bartholomew County, Indiana: Indiana Dept. Conserv., Div. Water Resources, 37 p., 2 pls. Summarizes and analyzes the available hydrologic data and geologic data by standard methods and makes a quantitative analysis of these data. The report concludes that ground-water conditions in the Columbus area are favorable for the development of 15 mgd.

1964. Watkins, F. A., Jr., Ground-water appraisal of the Clifty Creek basin and Clifty Creek Reservoir site, Indiana: U.S. Geol. Survey open-file report, 11 p., 1 table, 4 figs. Summary of hydrologic conditions with emphasis on the availability, adequacy, and usability of the ground water at the reservoir site.

1966. Schneider, A. F., and Gray, H. H., Geology of the upper East Fork drainage basin, Indiana: Indiana Geol. Survey Spec. Rept. 3, 55 p., 4 tables, 12 figs. Describes the physiography and geology of the basin as they affect the availability and suitability of reservoir sites.

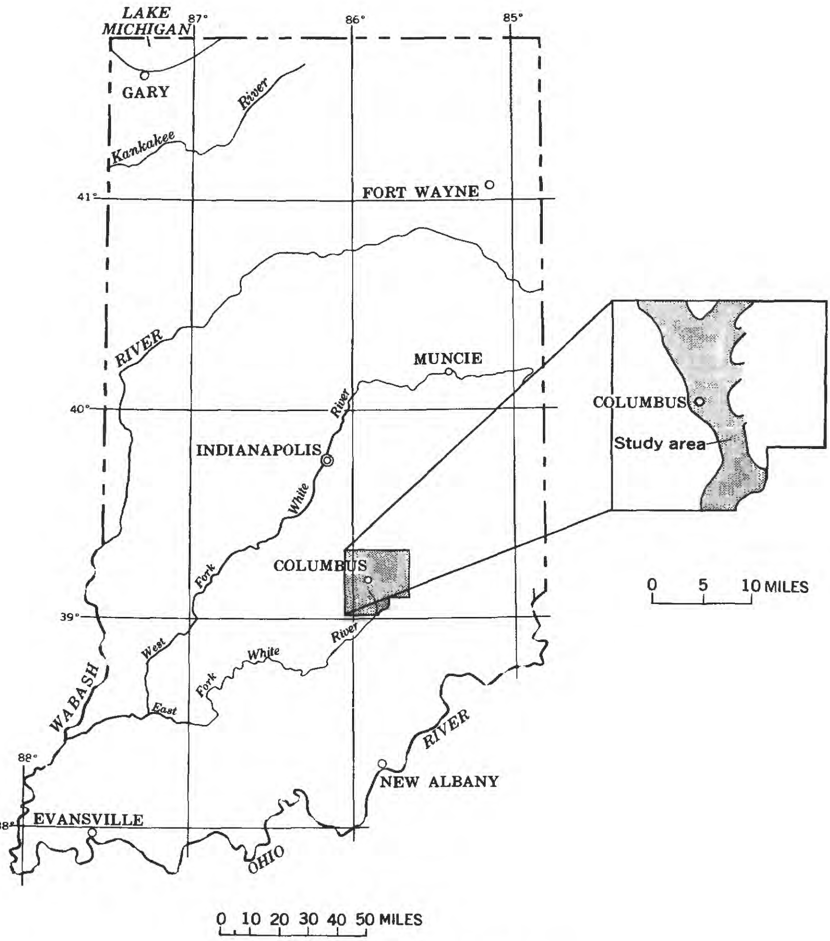

FIGURE 1.—Location of study area.

ACKNOWLEDGMENTS

The authors wish to thank all persons who contributed information and assisted in the preparation of this report. Mr. W. J. Steen of the Indiana Department of Natural Resources was particularly helpful in setting up the observation-well network and collecting and making a preliminary analysis of the data in the State files. We also wish to thank Mr. G. F. Hendricks and Mr. K. L. DeLap of Sieco, Inc., the consulting engineers for the city, for supplying pumping-test and other data and for running levels to many of the observation wells.

METHODS OF ANALYSIS

Standard U.S. Geological Survey techniques for the collection of basic data were utilized for the Columbus project. The extent of the aquifer was determined by surface and subsurface mapping. Water-level contour maps were prepared for periods of high and low water level. The data used on individual maps were collected on the same day in order to minimize any changes due to rising or falling water levels. A contour map of the bedrock surface was prepared for the area using data from wells and geophysical logs. A saturated-thickness map (pl. 1*A*) of the glacial outwash aquifer was prepared using the differen e between the water-level and bedrock altitudes from the water-level and bedrock contour maps. Hydraulic characteristics of the aquifer were estimated from well performance data and computed from pumping-test data. All these data were used to acquire an understanding of the hydrologic system so that an electrical-analog model of the glacial outwash aquifer could be constructed.

HYDROLOGIC SYSTEM OF THE COLUMBUS AREA

The subsurface material in the Columbus study area is predominantly outwash sand and gravel in a buried bedrock valley overlain at the surface by a sandy clay. There are some interfingering clay beds along the edges and a few clay lenses in the sand and gravel body. Figure 2 is a generalized geohydrologic section showing these features.

Water-table conditions prevail in the area, and the streams are in hydraulic connection with the aquifer (fig. 2).

There is a narrowing of the bedrock valley downstream (south) of Columbus (see pl. 1*A*). This constriction acts as a partial ground-water

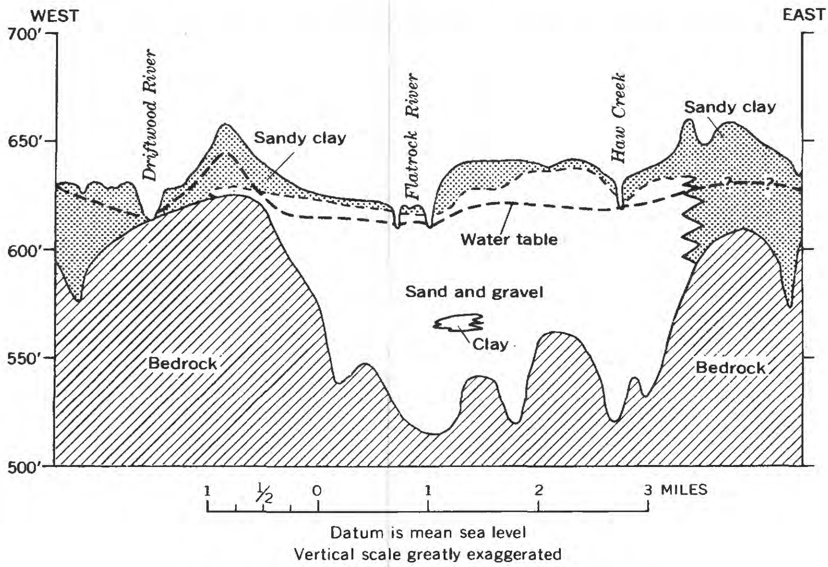

FIGURE 2.—Generalized geohydrologic section.

dam and causes an increase in ground-water discharge to the stream. The area upstream of this constriction is therefore especially promising for the development of ground-water supplies.

HYDROLOGIC CHARACTERISTICS

In areas where further development of ground water is contemplated, it is important to ascertain the hydraulic characteristics of the aquifer. The volume of water that the aquifer releases from or takes into storage per unit surface area per unit change in head normal to that surface is defined as the coefficient of storage. For water-table conditions it is equivalent to the specific yield of the aquifer. The rate at which an aquifer will yield water to wells is a function of the permeability or transmissibility of the aquifer. The coefficient of permeability can be defined as the rate of flow of water, in gallons per day, through a cross-sectional area of 1 square foot under a hydraulic gradient of 1 foot per foot. Transmissibility can be defined as the rate of flow of water, in gallons per day, through a vertical strip of the aquifer 1 foot wide extending the full saturated thickness of the aquifer under a hydraulic gradient of 1 foot per foot.

Many methods have been devised for determining the values of these aquifer characteristics. The methods used in this study and the results obtained are described below.

WELL-DATA ANALYSIS

The determination of hydraulic characteristics from well data was based on the data from wells in the study area, in addition to some wells in the same aquifer just out of the area to the north. The methods used were based primarily on specific capacity of the wells; checks were made from the analyses of four pumping tests. Specific capacities were used to estimate the transmissibility and permeability of the aquifer. The pumping tests were used to determine transmissibility, permeability, and also the storage coefficient of the aquifer. The specific capacity of a well is the relation of yield to drawdown, that is, its yield in gallons per minute per foot of drawdown caused by pumping, for a selected period of continuous pumping.

A method was used to estimate transmissibilities using specific capacities of wells in the area. First, permeability was estimated by multiplying the specific capacity of a given well by 1,000 and then dividing the result by the screen length. This figure was then multiplied by the saturated thickness of the aquifer to get an estimate of the transmissibility. The pumping test results are from Klaer, Davis, and Kingsbury (1951), some unpublished work by John Ferris in 1944, and computations from data from recent tests by Purdue University and the city of Columbus. The results of computations of these methods are shown in table 1.

TABLE 1.—*Hydrologic characteristics estimated from well data and computed from pumping tests*

[Discharge, in gallons per minute. Specific capacity, in gallons per minute per foot of drawdown. Transmissibility, in gallons per day per foot, and permeability, in gallons per day per square foot: A, estimated from specific capacity; B, from pumping tests]

Location	Well (pl. 1*A*)	Depth (feet)	Screen length (feet)	Saturated thickness (feet)	Pumping time (hours)	Discharge	Drawdown (feet)	Date measured	Specific capacity	Transmissibility		Permeability		Storage coefficient
										A	B	A	B	
T. 9 N., R. 6 E.	1	103	20	81	24	273	7.1	10–42	38	154,000		1,900		
	2	80	20	66	24	253	5.4	10–42	47	155,000		2,400		
	3	81	20	66	24	320		8–67			286,000		4,300	0.13
	4	124	25	106	8	1,486	20.6	11–47	72	305,000		2,900		
	5	98	30	81	24	1,000	11	9–50	91	245,000	300,000	3,000	3,700	.12
	6	114	25	90	24	2,010	27	8–64	74	265,000	380,000	3,000	4,200	.20
	7	113	25	100	27	1,160	10.1	10–64	115	460,000	380,000	4,600	3,800	
	8	93	15	76	5	1,068	28	8–50	38	190,000		2,500		
	9	94	20	74	5	681	11	9–66	62	230,000		3,100		
	10	85	10	81		325	11	4–46	30	245,000		3,000		
T. 10 N., R. 5 E.	11	55	12	80		550	30	54	18	120,000		1,500		
T. 11 N., R. 6 E.	12	75	15	60	24	880	22	8–57	40	160,000		2,700		
	13	80	15	65	24	754	18	9–57	42	180,000		2,800		
T. 11 N., R. 5 E [1]	14	82	20	68	24	2,000	19	10–43	105	355,000		5,200		
	[2] 14	82	20	68	72	2,000	19	10–43	105		146,000		1,900	.31
	15	46	10	31	5	550	13	6–62	42	130,000		4,200		
	16	46	10	32	2	618	16	10–55	39	125,000		3,900		
	16	46	10	32	6	617	15.5	11–55	40	130,000		4,000		
	17	110	20	92	24	1,093	30	2–56	36	165,000		1,800		
	18	81	30	66	8	1,767	34	4–67	52	110,000		1,700		
	19	110	20	90	5	979	14	5–48	70	315,000		3,500		
	20	112	30	95	8	800	4.6	10–38	174	550,000		5,800		
	20				10	1,001	22	7–53	46	146,000		1,500		
	21	101	20	84	8	710	6	2–45	118	495,000		5,900		
	21	101	20	84	6	1,470	14	11–53	105	442,000		5,300		

[1] Wells in T. 11 N., R. 5 E., are just north of the north county line in Johnson County.

[2] Unpublished information for Camp Atterbury well field.

It should be pointed out that some of the drawdowns listed in table 1 are large percentages of the saturated thickness of the aquifer. This may represent an actual drawdown in the aquifer, but it may also be due to well-entrance losses. Because the drawdowns at short distances from the pumping wells are very small, as stated in the following section entitled "Ground-water withdrawal," it is felt that the large drawdowns in some of the wells are due to turbulent flow.

Where possible, estimated transmissibilities were compared with those obtained from pumping tests. Generally those values from the analyses of pumping tests were higher than the estimated ones. For this reason it is felt that the estimated values are on the conservative side, and this would offset any decrease in transmissibility due to dewatering the aquifer.

The values for permeability shown in table 1 were averaged, and this average is 3,500 gpd per sq ft (gallons per day per square foot). The saturated thickness as shown on plate 1*A* and the average permeability were used to construct the transmissibility map (pl. 1*B*), on which transmissibility values are in thousands of gallons per day per foot.

GROUND-WATER WITHDRAWAL

When water is withdrawn from an aquifer by pumping a well, a cone of influence is created around the well. The volume of the cone is directly proportional to the amount of water withdrawn and inversely proportional to the storage coefficient of the aquifer. In water-table aquifers, the cone of influence spreads slowly. At the end of a 24-hour pumping test (Klaer and others, 1951) the drawdown was only 0.22 foot at a distance of 548 feet from a well pumping 1,000 gpm (gallons per minute). The water level in observation well BA–3, which is in the present city well field, fluctuates less than half a foot a day in response to changes in pumping.

The water-level map (pl. 2*A*) for February 16, 1967, does not show a cone of influence. This is due in part to the 10-foot contour interval of the map and in part to the ratio of the small quantity of water pumped to the quantity that could be pumped from the aquifer.

FLUCTUATION OF WATER LEVELS

Water levels fluctuate in response to changes in precipitation in the Columbus area. The hydrograph of well BA–2, which has the longest record of water levels, shows a decline from 1949 through 1954. This decline corresponds to a decline in the annual precipitation during the same period (fig. 3). In 1955 this trend was reversed and rose in response to increased precipitation, but well BA–2 was discontinued in 1956. This well was reestablished in 1965, and its response to changes in precipitation was again noted. The annual high water levels for the missing period (1956–65) were estimated (fig. 3), using the correlation between annual precipitation and annual high water level.

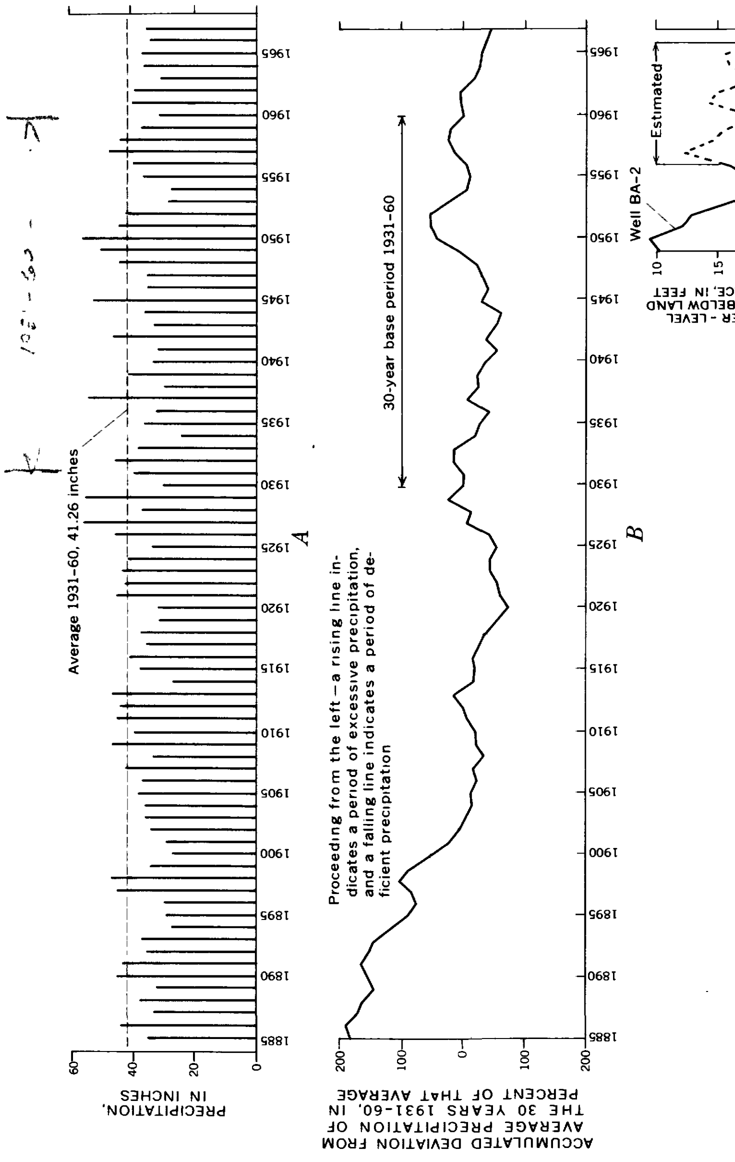

FIGURE 3.—*A*, Annual precipitation at Columbus, Ind., 1885–1967; *B*, Accumulated deviation from average precipitation of the 30 years 1931–1960; *C*, Annual high water levels in well BA–2 in sec. 18, T. 9 N., R. 6 E.

From the precipitation record and the hydrographs it is apparent that the long downward trend in water level is due to an extended period of deficient precipitation and not to pumping. Fluctuations due to pumping in this aquifer, under the present pumping conditions, are short term; water levels recover rapidly when pumping stops. The fluctuation for the period of record in well BA–2 is about 12 feet, and the maximum yearly fluctuation is about 6 feet.

VOLUME OF WATER AVAILABLE FROM THE SYSTEM

The amount of water in storage in an aquifer is determined by the volume of saturated sediments multiplied by their porosity. However, all this stored water is not available for use. A large amount of water will be held in the aquifer by molecular attraction or other forces of retention, so the amount that can be extracted is much less than the total amount in storage. In an artesian aquifer, the water released from or taken into storage is determined by the compressibility of the aquifer material and of the water. In a water-table aquifer, the volume of water released from or taken into storage is determined not only by the compressibility of the aquifer and the water but also by gravity drainage and refilling of the aquifer. In water-table aquifers, the amount of water obtained due to the compressibility factors is so small in comparison to the amount obtained from gravity drainage that it can be ignored.

In the Columbus area, calculated values of the storage coefficient range from 0.12 to 0.31, the average being about 0.2 (table 1). This is the figure used in subsequent analysis of the system.

The thickness and areal extent of the aquifer were determined from drillers' logs of wells, a seismic survey, and geologic reconnaissance of the area. The saturated thickness of the aquifer was found by subtracting the contours on the altitude of the bedrock surface from the contours on the water-level surface. The area between contours on the saturated thickness map was planimetered and multiplied by the average thickness between contours to determine the volume of the saturated sediments.

By using the average value of 0.2 for the coefficient of storage (specific yield), it was determined that there is about 800,000 acre-feet (260,000 million gallons) of water in the aquifer in the study area. Therefore, at the projected rate of withdrawal (34 mgd) the amount of water in storage in the aquifer is enough to supply Columbus for about 21 years assuming no recharge during this period. Although such an extreme period of no recharge is unlikely to occur, this illustrates the large amount of water available for short-term manipulation and management should the need arise.

ANALYSIS OF THE HYDROLOGIC SYSTEM BY ELECTRICAL-ANALOG MODEL

BASIS OF ELECTRICAL-ANALOG MODEL

Darcy's law for the rate of flow of water in an aquifer, wher expressed in the general form,

$$q=\frac{PAh}{L}, \tag{1}$$

is analogous to Ohm's expression for flow of electricity through a con-uctive medium,

$$I=\frac{C'A'V}{L'}. \tag{2}$$

The terms are used as in table 2.

TABLE 2.—*Hydrologic and analogous electrical characteristics*

Hydrologic concept			Electrical concept		
Symbol	Parameter	Units	Symbol	Parameter	Units
q	Discharge	Gallons per day	I	Current	Amperes.
h	Piezometric head difference over a distance L.	Feet	V	Voltage difference over L'	Volts.
P	Permeability	Gallons per day per square foot.	C'	Conductivity	Mhos per inch.
A	Cross sectional area	Square feet	A'	Cross sectional area	Square inches.
L	Distance between points from which h is determined.	Feet	L'	Length of conductive element.	Inches.
t	Time	Days	t'	Time	Seconds.
Q	Volume of water	Gallons	Q'	Electrical charge	Coulombs.

Both of these relations can be expressed in a simplified form. Consider m as saturated thickness of the aquifer and W as horizontal distance perpendicular to L; then $A=mW$, and we can substitute T, transmissibility, for mP. The result is another form of Darcy's law, $q \doteq T\frac{h}{L}W$. Since we use a square array, $W=L$, and so

$$q=Th. \tag{1a}$$

In Ohm's law,

$$\frac{C'A'}{L'}=\frac{1}{R},$$

where R is the resistance in a circuit element. This substitution into equation 2 results in

$$I=\frac{V}{R}. \tag{2a}$$

From the comparison of these two formulas the additional analogy of T to $\frac{1}{R}$ is evident.

Discharge q is the volume of water moving past a point in a certain time, $q=\frac{Q}{t}$. Similarly, I is the amount of electrical charge passing a point in a certain time, $I=\frac{Q'}{t'}$.

The analogy between these parameters is defined by four transformation constants. They are used in the following manner:

$$Q=K_1Q' \tag{3}$$

$$h=K_2V \tag{4}$$

$$q=K_3I \tag{5}$$

$$t=K_4t' \tag{6}$$

These constants were used to select components for the model. Their values are

$K_1=2,000,000,000,000,000$ (or 2×10^{15}) gallons per coulomb,
$K_2=10$ feet of head per volt,
$K_3=10,000,000,000$ (or 10^{10}) gallons per day per ampere,

and

$K_4=200,000$ (or 2×10^5) days of real time per second of computational time.

The performance of an analog model is dependent on the components used through the conversion factors. Therefore, it is necessary to select components that most nearly approximate the hydrologic parameters in the desired analogy. The model must be developed with understanding of the kind of results desired, and data must be chosen to answer and solve as well as possible the specific questions and problems. Because of time limitations, very little fieldwork went into this model, and it was conceived and built using existing well logs, pumping tests, and reports.

Results of the investigation of the model are termed solutions since the model is actually a computer designed to solve the partial differential equation for this particular aquifer with varying boundary and initial conditions. For a discussion of analog-model theory see Patten (1965).

CONSTRUCTION OF THE MODEL

COEFFICIENT OF TRANSMISSIBILITY

The aquifer material is distributed in space in a continuous fashion. This areal distribution cannot be duplicated in the model, but a part of this material must be represented by a single resistor. Because of the scale used, there are 32 resistors per square mile.

The resistors were selected by first making an areal map of the transmissibilities (pl. 1*B*) and then using the formula

$$R=\frac{1}{T}\frac{K_3}{K_2}\frac{D_{h1}}{D_{h2}}, \tag{7}$$

where R, T, K_3, and K_2 are as previously described and $D_{h1}=D_{h2}=1{,}320$ feet. D_{h1} is a horizontal distance on the model, and D_{h2} is a distance measured perpendicular to D_{h1} and represented by the same resistor. Numerically, equation 7 reduces to

$$R=\frac{10^9}{T}. \tag{8}$$

Resistors were chosen for increments of transmissibility of 25,000 gpd per ft and matched as closely as possible. The transmissibilities of 25,000 to 425,000 gpd per ft are represented by resistors of 39,000 to 2,200 ohms. There was a maximum matching error of 10 percent. This magnitude of error occurred only once, and the fact that a larger value of resistance was used tended to increase head drop for any flow condition and made results more conservative than if precision resistors had been used. Since the resistors react to the voltage difference the same regardless of the voltage level and the transmissibilities change with the saturated thickness, it was necessary to limit the drawdown to 10 percent of the saturated thickness in the simulation.

STORAGE

Capacitors were used to provide the required storage when the model was pulsed to simulate pumping. There was one capacitor for each node in the model. The capacitors used were 0.015 microfarad, representing a storage coefficient of approximately 0.2.

The capacitor value was chosen using the formula

$$C=7.48S\frac{K_2}{K_1}A,$$

where S is the storage coefficient, K_2 and K_1 are as previously described, and A is (1,320 ft)^2. The value chosen represented an increase of 15 percent over the average storage coefficient chosen. The capacitors had a tolerance of 5 percent.

ACCRETION

Accretion as used in this report is that part of rainfall which reaches the aquifer as recharge, going directly through the zone of aeration. The accretion circuit consisted of a number of resistors connected in parallel between a controllable bus and the modeled aquifer. These resistors were spaced so that there were four accretion resistors per square mile.

The values of the resistors were chosen so that the modeled accretion rate was constant over the area under the piezometric conditions of October 26, 1964: an accretion rate of 5.76 inches per year (Klaer and others, 1951), approximately 300,000 gpd per sq mi (gallons per day per square mile). This circuit had the feature of increasing the accretion for a corresponding drop in the piezometric-potential level.

The amount of increase in accretion is analogous to the amount of water that will be recovered because of reduction of evapotranspiration due to the greater depth to water. The amount recovered was simulated at from 10 percent to 23 percent of the 300,000 gpd per sq mi rate. This amount occurs only in the square mile of the greatest pumpage and decreases for greater distances from the pumping center. In the area of greatest simulated pumping, the natural water levels are less than 10 feet below the land surface.

STREAMFLOW

The streams are in hydraulic connection with the aquifer. Modeling of the streamflow was done by selecting resistors so that the correct simulated gradient was obtained for a known condition of flow. The model was then tested for the most likely and most conservative flow conditions, the median or 50-percent flow duration and 90-percent flow duration.

UNDERFLOW

In this report, the term underflow refers to the water flowing in the aquifer at the limits of the modeled area. This includes the water moving in the outwash deposits approximately parallel to the streams as they enter the modeled area. It also includes water moving out of the modeled area. The underflow moving into the modeled area was simulated by applying a voltage at the extremities of the modeled portion of the buried valleys. The voltage was applied through a meter and variable resistor so that the current could be measured and adjusted.

FLOW FROM LESS PERMEABLE FORMATIONS

Upon first testing the model it was found that the potentials along the east edge of the modeled aquifer were lower than the observed piezometric levels. This indicated that a significant amount of water was being supplied to the aquifer from adjacent formations. This was most noticeable in the northeast part of the area. Since the streams and rivers flowing in the center of the aquifer tend to drain the more permeable aquifer rapidly, whereas the less permeable material along the borders yields its supply of water to the aquifer slowly, the border material has a regulating effect on the piezometric levels.

The amount of water contributed by these adjacent formations varies with time. Table 3 shows the range in flow used for this contribution, the length of the border between the aquifer and the contributing formation, and the flow per mile across this border.

TABLE 3. — *Water moving into the aquifer from less permeable material*

Location	Range in flow (mgd)	Length of border (miles)	Flow across border (mgd per mi)
North end to Haw Creek (east border)	9–20	6.15	1.5–3.25
Haw Creek to Clifty Creek (east border)	4– 5	6.40	.6– .8
Clifty Creek to Little Sand Creek (east border)	2– 3	6.60	.3– .5
Little Sand Creek to Sand Creek (east border)	3– 4	4.70	.6– .9
North border, center	1	5.15	.2
Driftwood River above confluence with Flatrock River (west border)	.5	3.15	.16
Driftwood River below confluence with Flatrock River (west border)	2– 3	8.55	.2– .35

USING THE ANALOG MODEL

Potentials in the proper amount were applied to the input circuits as described in the previous section. Ammeters were used on all circuits to read the modeled flow.

The rows and columns of resistors were numbered so that each node or joining of four field resistors and a storage capacitor could be described by two numbers. A digital voltmeter was used to determine the piezometric-potential level of each of these nodes. Maps were made by drawing contours for voltage levels as determined by the above processes. All input and output current values were recorded.

VERIFYING THE MODEL

Verification implies that measured water level and known flows are simulated in the model for imposed boundary conditions. These boundary conditions must be known through measurement or determined using aquifer parameters.

Verification of the model was done by duplicating the piezometric surfaces of October 26, 1965, and February 16, 1967. On both of these days there had been a period of precipitation prior to the measuring of the wells, which resulted in streamflow conditions approximating the 50-percent flow duration. On February 16 the water-level recorders in the area exhibited no change in storage. Plate 2*A* is a map showing altitude of observed and simulated water levels for February 16, 1967.

Flow values from the verified model are listed in table 4.

TABLE 4.—*Simulated inflow and outflow values from the verified model (Feb. 16, 1967)*

	mgd
Inflows at the model boundaries:	
Streamflow:	
Flatrock River	100
Driftwood River	650
Clifty Creek	53
Haw Creek	14
Little Sand Creek	10
Sand Creek	105
Underflow:	
Driftwood Valley	36
Flatrock Valley	27
Clifty Valley	12
Little Sand Valley	3
Sand Valley	15
Contiguous formations:	
East	24
West	15
Accretion	78
Outflows at the model boundaries:	
East Fork White River	1140
Underflow	4
Pumping (city well field and industrial)	7

SIMULATED PUMPING PLANS

Following verification of the model, several different hydrologic conditions were applied to the model, and pumping was simulated at selected sites in the area to determine the effects on water levels and streamflow. Six different pumping plans were simulated and are referred to by number. These plans are defined in table 5. The results of these pumping situations are given in the figures and described in the following sections.

TABLE 5.—*Hydrologic and pumping simulation conditions for six pumping plans*

Plate	Pumping plan	Number of well-field locations	Total pumping (mgd)	Simulated accretion rate	Simulated streamflow duration (percent)	Type of model	Length of time of pumping
2*B*	1	6	36.7	Average	50	Steady state	Infinite.
2*C*	2	6	35.4	Average	50	Steady state	Infinite.
3*A*, *B*	3	4	35.5	72 percent of average	90	Nonsteady state	1,000 days.
3*C*, *D*	4	6	35.3	72 percent of average	90	Nonsteady state	1,000 days.
3*E*, *F*	5	6	34.9	Average	50	Nonsteady state	45 years, by 9-year increments.
3*G*, *H*	6	6	35.7	Average	50	Nonsteady state	45 years, by 9-year increments.

STEADY-STATE SOLUTION

The steady-state solution consists of "pumping" the model at various places and not using the storage features to obtain the ultimate long-term drawdown. This would amount to pumping in the aquifer over an extremely long period of time; all water that is withdrawn is recharge. No water is removed from storage. The piezometric surface will be lowered, but it is not possible to evaluate the rate of this lowering.

The steady-state pumping is simulated by drawing current from the model at the pumping center. The current is measured by an ammeter, and by using the K_3 relation the amount of water withdrawn can be computed. Simulated values of water-level altitude are read at the nodes using a voltmeter.

Pumping plan 1 (table 6) withdrew 36.7 mgd. The resulting piezometric surface is shown on plate *2B*. Since this surface was determined with simulated median-flow conditions (50-percent flow duration) and an average figure for accretion, the resulting water levels must be viewed as average water levels.

TABLE 6.—*Pumping-center locations and amounts withdrawn for pumping plan 1 shown on plate* 2B

Location	Amount (mgd)
1. NE¼NW¼NE¼ sec. 2, T. 8 N., R. 5 E.	5.8
2. SW¼NW¼SE¼ sec. 7, T. 8 N., R. 6 E.	5.6
3. NW¼SW¼NE¼ sec. 36, T. 9 N., R. 5 E.	5.8
4. SE¼NE¼SE¼ sec. 18, T. 9 N., R. 6 E.	7.8
5. SW¼NW¼NE¼ sec. 30, T. 9 N., R. 6 E.	4.9
6. NE¼NE¼NE¼ sec. 35, T. 10 N., R. 5 E.	6.8
Total	36.7

Only local lowering of the water levels is indicated by the solution shown on plate *2B*. Comparing this with plate *2A* shows that this is especially true in the area north of the city and west of the Flatrock River, where the maximum lowering of the water table is 15 feet and almost 7 mgd is being withdrawn from the aquifer. Less than 1 square mile would experience a level change of more than 10 feet in this area, and about 2 square miles of the aquifer would experience a water-level change of more than 8 feet.

Three well fields in the area south of the Driftwood River and west of the East Fork, pumping a total of 17.5 mgd, were simulated in this plan. Again, the depressions of more than 10 feet in the water level are local phenomena. These areas of 10-foot change total about 2 square miles but are not connected. The largest of these three areas of 10-foot changes is the one that is farthest from the East Fork, and it is about 1½ square miles in area. The entire area west of the river will experience a drop in water level of 6 feet under this plan.

The final area of interest is the city of Columbus itself. If two well fields pump 12.4 mgd, the decline of 10 feet in water level covers about

2¼ square miles. Most of the pumping, 7.8 mgd, is simulated in the city well field in the SE¼ sec. 18, T. 9 N., R. 6 E., and the other pumping, 4.6 mgd, is about 1¾ miles further south in the NW¼ sec. 30, T. 9 N., R. 6 E. The 10-foot change will take the shape of a narrow oval extending about one-half mile south of the southern field and five-eighths mile north of the city well field.

This plan will cause a water-level decline of 5 feet in the aquifer between a line drawn across the aquifer parallel to and coincident with Clifty Creek and a line drawn in an east-west direction 2 miles south of the north county line. There will be areas in this general region that will not experience this amount of water-level decline—for example, those adjacent to rivers and away from the pumping centers.

Approximately 4 mgd of the water pumped will come from an increase in accretion rates due to lowering of the water table. The balance of the water will come from interception of discharge to the East Fork.

Pumping plan 2 (table 7) was similar to pumping plan 1, but all the pumping, except the city well field, is imposed south of the Driftwood River. The resulting piezometric surface is shown on plate 2*C*.

In this plan, the area north of the Driftwood River has less than 4 feet of depression of the water level. Owing to the heavy pumping, 20.3 mgd, water levels in the area of the aquifer west of the East Fork and between the south line at sec. 13 and sec. 14, T. 8 N., R. 5 E. and the Driftwood River are lowered 8 feet or more, and about 5½ square miles are lowered by 10 feet or more. One pumping site east of the White River and near the mouth of Clifty Creek has a small area of lowered water table around it. Pumping in this area amounts to 7.3 mgd and causes minimal lowering of present water levels.

The water levels in the city well field are such that without the pumping of the 7.8 mgd they would recover 10 feet in an area slightly larger than 1 square mile and in the shape of a 2:1 ellipse having its major axis parallel to Haw Creek. Since much of this pumping is already present, these changes may not be as severe as indicated by this analysis. About two-thirds of the area east of the Flatrock River and north of Columbus will have water levels 5 feet or more lower than if no pumping were present.

TABLE 7.—*Pumping-center locations and amounts withdrawn for pumping plan 2 shown on plate 2C*

Location	Amount (mgd)
1. NE¼NW¼NE¼ sec. 2, T. 8 N., R. 5 E.	4.1
2. NW¼NE¼NW¼ sec. 13, T. 8 N., R. 5 E.	4.3
3. SW¼NW¼SW¼ sec. 8, T. 8 N., R. 6 E.	7.3
4. SE¼SE¼NW¼ sec. 26, T. 9 N., R. 5 E.	4.7
5. NW¼SW¼NE¼ sec. 36, T. 9 N., R. 5 E.	7.2
6. SE¼NE¼SE¼ sec. 18, T. 9 N., R. 6 E.	7.8
Total	35.4

EFFECT OF PUMPING ON STREAMFLOW

The amount of water obtainable from the Flatrock River was evaluated for median flow with these pumping plans. The streamflow changed near the pumping centers but was relatively unaffected when the pumping was at a distance from the river.

Voltage measurements across resistors connecting the aquifer circuit to the stream circuits gave an indication of the exchange of water between the aquifer and the stream. Values of the resistors were selected on the basis of an estimated permeability of the stream bed. These measurements showed that under a no-pumping condition, the aquifer contributed about 20 mgd to the Flatrock River from the bridge at the corner of secs. 7, 8, 17, and 18, T. 10 N., R. 6 E. to the mouth, whereas under plan 1 about 11 mgd was discharged into this reach of stream, and about 18 mgd entered the stream under conditions of pumping plan 2. Historically, Flatrock River has maintained 29 mgd 90 percent of the time at Columbus based on estimates from East Fork White River at Columbus, the Driftwood River at Edinburg, and occasional measurements of the Flatrock River at Columbus. The lowest ever measured in Flatrock River at Columbus was 19 mgd. Under pumping plan 1 this flow would be approximately 10 mgd.

NONSTEADY-STATE SOLUTION

Unlike the steady-state solution, the nonsteady-state solution includes the effects of storage depletion in the aquifer and shows the change of water levels with time. The results of pumping are read on an oscilloscope screen; only changes in voltage level, corresponding to a change in water levels, are available. These voltage changes are measured and are superimposed (algebraically added) on the piezometric-potential map that was made assuming a no-pumping condition. Water-level altitudes are then determined by subtracting the changes from the no-pumping condition.

Pumping plans 3 and 4 were performed to show aquifer reaction under extreme low-flow conditions. The flow of all streams was simulated at 90-percent flow duration, and only 72 percent of the annual average accretion was used. Pumping continued for an equivalent of 1,000 days under these extreme conditions of low recharge.

Pumping plan 3 withdrew 35.5 mgd at four locations (table 8). Plate 3*A* shows the decline in water level due to this pumping. Plate 3*B* is a water-level-altitude map showing the levels which would occur after the indicated pumping.

Pumping plan 4 (table 9) withdrew 35.3 mgd at six locations. As shown on plate 3*C*, water levels declined at least 10 feet in the entire area bounded by the Driftwood River on the north, the south line of secs. 13

TABLE 8.—*Pumping-center locations and amounts withdrawn for pumping plan 3 shown on plate* 3A, B

	Location	*Amount (mgd)*
1.	SW¼NW¼SW¼ sec. 8, T. 8 N., R. 6 E.	8.7
2.	NW¼SW¼NE¼ sec. 36, T. 9 N., R. 5 E.	10.0
3.	SE¼NE¼SE¼ sec. 18, T. 9 N., R. 6 E.	6.3
4.	SE¼NE¼SE¼ sec. 23, T. 10 N., R. 5 E.	10.5
	Total	35.5

TABLE 9.—*Pumping-center locations and amounts withdrawn for pumping plan 4 shown on plate* 3C, D

	Location	*Amount (mgd)*
1.	NW¼NW¼NE¼ sec. 2, T. 8 N., R. 5 E.	4.2
2.	NW¼NE¼NW¼ sec. 13, T. 8 N., R. 5 E.	4.5
3.	SW¼NW¼SW¼ sec. 8, T. 8 N., R. 6 E.	7.9
4.	SE¼SE¼NW¼ sec. 26, T. 9 N., R. 5 E.	6.2
5.	NW¼SW¼NE¼ sec. 36, T. 9 N., R. 5 E.	6.4
6.	SE¼NE¼SE¼ sec. 18, T. 9 N., R. 6 E.	6.1
	Total	35.3

and 14, T. 8 N., R. 5 E., on the south, the East Fork of the White River on the east, and the valley wall on the west.

This area is the most heavily pumped and accounts for 21 mgd, or about 60 percent of the total. In the city of Columbus and near the mouth of Clifty Creek, the effects of the pumping are very much localized. Plate 3*D* shows the water-level contours determined for the changes shown on plate 3*C*.

Since no period as long as 1,000 days of 90-percent flow duration and 72-percent average accretion rate can reasonably be expected, the results of these two solutions point out that the aquifer will easily provide the indicated withdrawals.

Pumping plans 5 and 6 were made with the same median streamflow and average accretion conditions as plans 1 and 2 of the steady-state model. Well-field locations of plan 5 are the same as plan 1 and those of plan 6 are the same as plan 2.

Under pumping plans 5 and 6 a total pumping time of 45 years was used. However, since it was felt that actual development of the Columbus water supply would be over a period of time and based on demand, a staggered system of well-field production was used to duplicate real conditions more nearly.

Instead of simulating pumping at all six locations for 45 years, a well field was added every 9 years until all six were in production. For example, in plan 5, the well field at location 4 was assumed to be producing from the start of the period—1970. The well field at location 5 was put into production 9 years later, that at location 3, 9 years after that, and so on, the last two fields going into production in 2006. All six fields were pumped for the last 9 years of the 45-year period. In each of the 9-year periods the aquifer system reached a condition of approximate equilibrium, as indicated by the decrease in the rate of decline of the water levels and spread of the cone of influence.

Pumping plan 5 simulated 34.9 mgd of pumpage at the end of 45 years (table 10). The decline in water level at the end of the 45-year period is shown on plate *3E*, and the final altitude of water levels is on plate *3F*.

TABLE 10.—*Pumping-center locations, amounts withdrawn, and timing of withdrawals for pumping plan 5 shown on plate* 3E, F

	Location	Pumping period	Amount (mgd)
1.	NE¼NW¼NE¼ sec. 2, T. 8 N., R. 5 E.	2006–2015	4.8
2.	SW¼NW¼SE¼ sec. 7, T. 8 N., R. 6 E.	2006–2015	4.8
3.	NW¼SW¼NE¼ sec. 36, T. 9 N., R. 5 E.	1988–2015	5.5
4.	SE¼NE¼SE¼ sec. 18, T. 9 N., R. 6 E.	1970–2015	7.8
5.	SW¼NW¼NE¼ sec. 30, T. 9 N., R. 6 E.	1979–2015	4.5
6.	NE¼NE¼NE¼ sec. 35, T. 10 N., R. 5 E.	1997–2015	7.5
	Total		34.9

Pumping plan 6 (table 11) is a plan in which 35.7 mgd was being pumped at the end of 45 years. The decline in water level after 45 years is shown on plate *3G*, and the resulting water levels are on plate *3H*.

Figure 4 is a representation of the water-level decline resulting from plan 6 at a point midway between two pumping areas 1.4 miles apart, both areas being south of the Driftwood River and west of the East Fork of the White River. The first stage of pumping is barely perceptible on this diagram, and it is due to pumping in the present city well field about 3 miles away. The second stage is due to pumping in one of the closer areas only 0.71 mile away. The third step is due to the other nearby pumping area the same distance away. The fourth step is from pumping 1.3 miles away. The last is from two pumping areas 3.5 and 3.2 miles away. Under both of these plans, changes in water level range

TABLE 11.—*Pumping-center locations, amounts withdrawn, and timing of withdrawals for pumping plan 6 shown on plate* 3G, H

	Location	Pumping period	Amount (mgd)
1.	NW¼NW¼NE¼ sec. 2, T. 8 N., R. 5 E.	1997–2015	7.9
2.	NW¼NE¼NW¼ sec. 13, T. 8 N., R. 5 E.	2006–2015	4.2
3.	SW¼NW¼SW¼ sec. 8, T. 8 N., R. 6 E.	2006–2015	8.3
4.	NE¼SE¼NW¼ sec. 26, T. 9 N., R. 5 E.	1979–2015	5.1
5.	NW¼SW¼NE¼ sec. 36, T. 9 N., R. 5 E.	1988–2015	3.9
6.	SE¼NE¼SE¼ sec. 18, T. 9 N., R. 6 E.	1970–2015	6.3
	Total		35.7

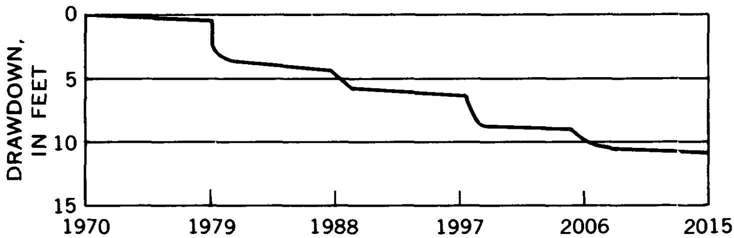

FIGURE 4.—Water-level decline due to pumping, 1970–2015, NW¼SW¼SW¼ sec. 25, T. 9 N., R. 5 E.

from more than 20 feet in the center of the pumping areas to less than 3 feet near the boundaries of the study area.

SUMMARY AND CONCLUSIONS

In the Columbus study area the subsurface material is predominantly outwash sand and gravel in a bedrock valley. The narrowing of the bedrock valley downstream from Columbus acts somewhat as a dam causing an increase in ground-water runoff to the stream, making the area upstream from the narrowest part of the bedrock valley especially promising for the development of ground-water supplies. Ground water is under water-table conditions in the outwash aquifer. The streams are partially incised into and are hydraulically connected with the aquifer.

In the Columbus area, analyses of well data gave values of transmissibility that range from near zero, along some parts of the boundary where the aquifer is thin, to approximately 500,000 gpd per ft, in the thickest part of the aquifer. The average coefficient of permeability was calculated to be about 3,500 gpd per sq ft. This value of permeability was used with the saturated thickness to construct the transmissibility map. The coefficient of storage of the aquifer was found to average about 0.2 (specific yield of aquifer is therefore 20 percent).

The total volume of water available from storage is about 800,000 acre-feet. This would supply the projected needs of Columbus (34 mgd) for about 21 years if there were no recharge during this period.

Analysis of the model indicates that the projected required yield of 34 mgd can be obtained with proper planning and spacing of wells. Future pumping will lower water levels for 2 or 3 years, but they will reach a level of approximate equilibrium by the end of 3 years. The natural yearly variations in water level are greater than those which will be caused by pumping except for changes in the immediate pumping area.

The flow in the Flatrock River will be affected by pumping, but if the withdrawals are made south of the city this depletion will be negligible. There will be less change in water levels if the pumping is done adjacent to streams and in the thickest part of the aquifer.

This study demonstrates the application of electrical-analog analysis to a water-supply problem. It also emphasizes the need for adequate data to evaluate the system realistically and the limits placed on interpretation due to the type of available data. Reliable water levels and pumping rates are especially important for verification of the model.

The model is to be maintained and modified in accordance with new data. More accurate predictions of effects of pumping will be made when these data are utilized in the model.

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