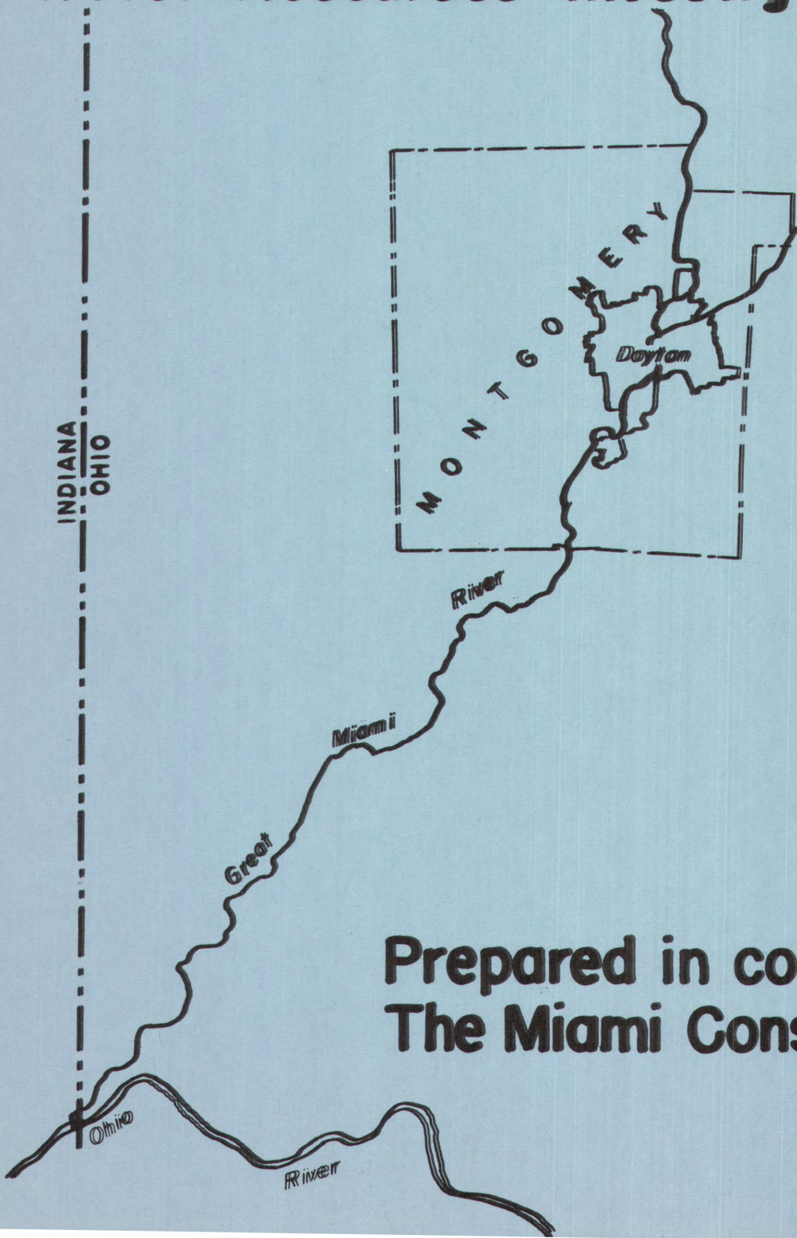
DIGITAL MODEL SIMULATION OF THE GLACIAL-OUTWASH AQUIFER AT DAYTON, OHIO

U.S. Geological Survey

Water Resources Investigations 18-75



Prepared in cooperation with
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A digital model which uses a finite-difference approximation technique to solve partial differential equations of flow through a porous medium was used to evaluate the effects of pumping stresses on water levels. The simulated head values presented in map form generally are in good agreement with potentiometric-surface maps prepared from field measurements.

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DIGITAL MODEL SIMULATION OF THE GLACIAL-OUTWASH AQUIFER AT DAYTON, OHIO

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ABSTRACT

Dayton, Ohio and its environs obtain most of their water from wells which penetrate the highly productive glacial-outwash deposits underlying the Great Miami River and its tributaries. Combined municipal and industrial use of ground water in this 90-square-mile (233-square-kilometre) area has increased from about 180 cubic feet per second (5 cubic metres per second) in 1960 to nearly 250 cubic feet per second (7 cubic metres per second) in 1972. The increased pumpage has resulted in continuing water-level declines in some parts of the area.

A digital model which uses a finite-difference approximation technique to solve partial differential equations of flow through a porous medium was used to evaluate the effects of pumping stresses on water levels. The model was made to simulate the aquifer as part artesian and part water table. The principal source of recharge to the unconfined part of the aquifer is from induced streambed leakage. Average head values for 1960 were calculated for each node in the digital model in response to ground-water pumping during the five pumping periods. The simulated head values presented in map form generally are in good agreement with potentiometric-surface maps prepared from field measurements for April 1959 and October 1960.

INTRODUCTION

Glacial-outwash deposits partly filling the valleys of the Great Miami River and its tributaries are the principal source of water for the Dayton, Ohio area. Ground-water use has increased steadily over the years in response to increased industrial activity and population growth. In 1960 average ground-water pumpage was about 180 ft^3/s (5 m^3/s). In 1972 the average pumpage was estimated at nearly 250 ft^3/s (7 m^3/s). The result in areas of concentrated withdrawal has been a lowering of water levels and a reduction in well yields.

In March 1973 the Miami Conservancy District asked the U.S. Geological Survey to develop a hydrologic model of the principal aquifers in the Dayton area. The model would be used to evaluate the ground-water situation as it presently exists and to predict the effect of future development. Two modeling techniques were considered -- a digital computer model and an electric-analog model.

Modeling techniques provide efficient means for calculating the response of an aquifer to real or hypothetical hydrologic stresses. Each technique, digital and electric-analog, offers advantages in solving fluid-flow equations for specific hydrologic conditions. The objective is to simulate the hydraulic head in an aquifer as a function of time and space. The decision was made to develop a digital model because it could be used more conveniently by the Miami Conservancy District.

The specific purpose of the digital model study was to develop a better understanding of the aquifer system and to show how it responds to various pumping stresses. The scope of the study was to develop a single-transmissive-layer model that could be used to evaluate the effects proposed ground-water development schemes would have on water levels. A simulation model of the aquifer system is particularly useful in determining areas where various types of hydrologic data would be needed. The model could aid the Miami Conservancy District and other local interests in designing future data collection programs and to help evaluate the need for additional ground-water management.

Most of the data used as input to the digital model were taken from previous investigations, particularly the reports by Norris and Spieker (1966) and Black, Crow, and Eidsness, Inc. (1970). These reports provide basic hydrologic data and a detailed description of the geology, hydrology, and the history of ground-water development in the Dayton area.

The author expresses his appreciation to Paul Plummer of the Miami Conservancy District for providing basic data and to P. C. Trescott of the U.S. Geological Survey, for making program modifications.

For use of those readers who may prefer to use metric units rather than English units, the conversion factors for the terms used in this report are listed below:

Multiply English unit	By	To obtain metric unit
feet (ft)	0.3048	metres (m)
feet per second (ft/s)	0.3048	metres per second (m/s)
cubic feet per second (ft^3/s)	0.02832	cubic metres per second (m^3/s)
miles (mi)	1.609	kilometres (km)
square miles (mi^2)	2.590	square kilometres (km^2)
inches (in)	25.4	millimetres (mm)

Location of Area

The study area (fig. 1) which includes the city of Dayton and environs lies in Montgomery County in southwestern Ohio. The rectangular area covers approximately 90 mi^2 (233 km^2), extending in a northeast-southwest direction along the Great Miami River from Miami Villa south through the greater Dayton metropolitan area, Moraine, and West Carrollton to the northern boundary of Miamisburg. That part of the valley of the Great Miami River that lies in the modeled area is about 16 mi (26 km) long and averages about 2.5 mi (4 km) wide. The principal tributaries of the Great Miami River in the modeled area are the Mad River, Stillwater River, Wolf Creek, Holes Creek, and Bear Creek.

HYDROGEOLOGY

Aquifer System

The principal source of ground water in the Dayton area is glacial-outwash deposits in the valleys of the Great Miami River and its tributaries. The outwash deposits in the study area range in thickness from a few feet to as much as 275 ft (84 m). In some places buried till separates the permeable deposits into more than one unit.

In the Rohrers Island area (fig. 1) an upper water-table aquifer and a lower artesian aquifer occur. The upper aquifer is hydraulically connected to the Mad River and is recharged artificially through a network of regulated lagoons. The lower aquifer is separated from the upper aquifer by semi-confining till beds and is recharged by leakage from the upper aquifer. The thickness of the lower aquifer ranges from a few feet near the model boundaries to as much as 100 ft (30.5 m) in those areas where the altitude of the bedrock surface is lowest (fig. 2). Water is pumped from both the upper and lower aquifers with the greatest amount from the upper aquifer. Water-level changes caused by pumping are usually insignificant in the upper aquifer because of the extremely efficient artificial recharging system. The aquifer system outside the Rohrers Island area is considered as water table, even though in many areas till layers are known to be interbedded with the outwash. It is assumed that both aquifers in these areas respond as a unit, hydraulically connected to the streams.

Bedrock and Upland Glacial Deposits

The bedrock bounding the glacial-outwash deposits consists of shale interbedded with thin layers of limestone. The bedrock in the study area is assumed to be everywhere an impermeable, no-flow boundary, and the altitude of the bedrock surface represents that of the bottom and sides of the aquifer.

Upland glacial deposits, consisting mostly of till and clay and minor amounts of sand and gravel, overlie the bedrock along the aquifer boundaries or valley walls and provide some recharge to the outwash aquifer. For the

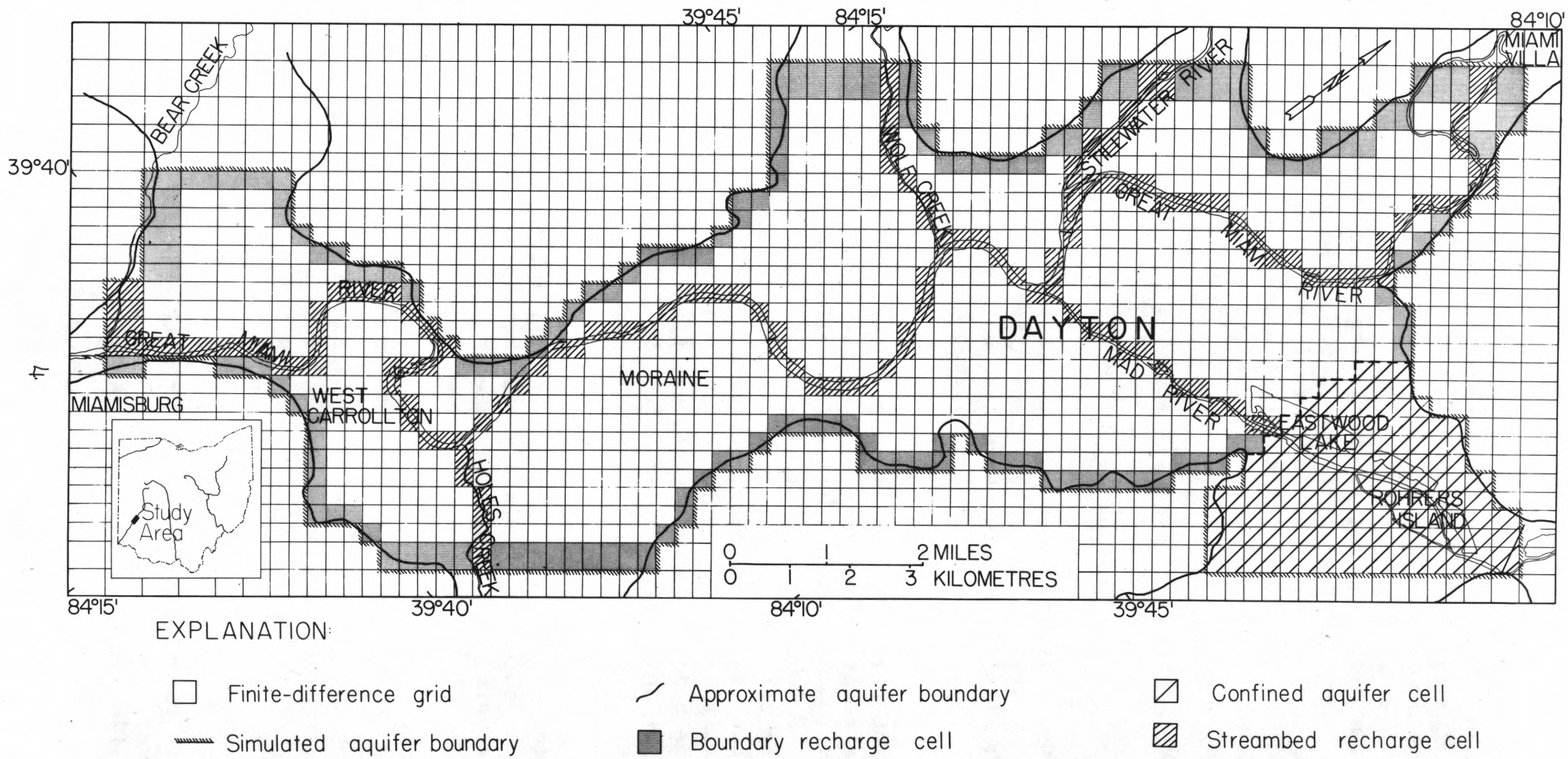
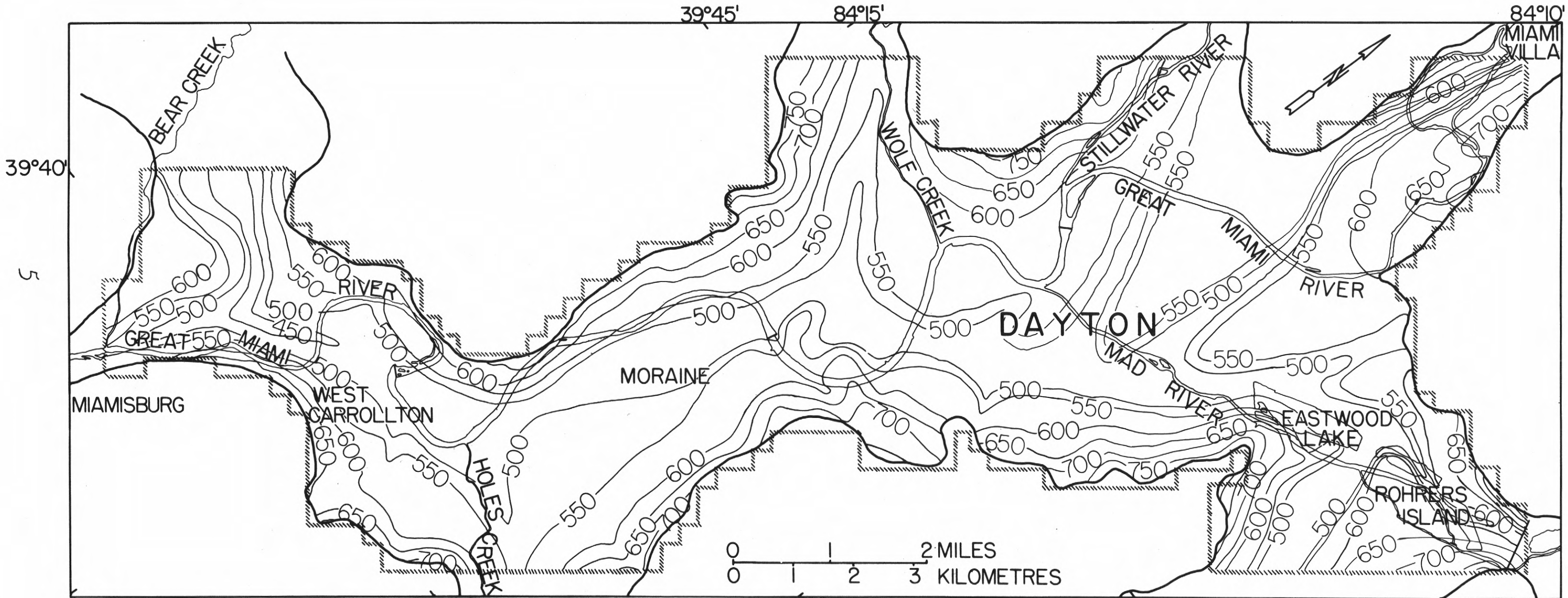


Figure 1.--Location map and map of modeled area



EXPLANATION:

Approximate aquifer boundary

Simulated aquifer boundary

600 Bedrock contour. Shows altitude of bedrock surface. Contour interval 50 feet (15 metres). Datum is mean sea level.

Figure 2.--Bedrock surface underlying the glacial-outwash aquifer (Modified from Norris and Spieker, 1966).

most part the upland deposits and the bedrock are relatively unimportant as sources of water except for farm and domestic supplies.

The configuration of the aquifer boundaries (fig. 2) is based on a bedrock map by Norris and Spieker (1966). A few changes in the bedrock contours were made for the Moraine and West Carrollton areas based on additional drillers' logs.

DIGITAL MODEL

Definition

A digital model of an aquifer system is a mathematical model which utilizes a digital computer for numerically solving the partial differential equations of flow through a porous medium. The numerical method used in this study for approximately solving the differential equations of flow is the finite-difference method. In this method the aquifer is subdivided into a system of rectangular cells in which the aquifer properties are assumed to be uniform. The differential equation of flow is approximated by a finite-difference equation which relates heads at the "centers" of the cells at selected "steps" in time.

Pinder and Bredehoeft (1968) reviewed the development of analytical methods and electric analog and digital models used in ground-water hydrology to evaluate the response of aquifer systems to various hydrologic stresses. They described the mathematical equations used, development of a finite-difference model, methods of solving the resulting system of simultaneous linear equations, and a field application. Trescott, Pinder, and Jones (1970) described a digital model study of an alluvial aquifer. Modeling techniques presented in these reports were used as a guide for the Dayton aquifer study.

Pinder (1970) prepared a report that lists and describes a digital model computer program that can be used to simulate two-dimensional flow in an aquifer system. One purpose of Pinder's work was to develop a computer program flexible enough that it could be easily used by U.S. Geological Survey hydrologists, without modification, for relatively simple problems and adapted to more complex problems with only minor reprogramming.

The basic digital model program developed by Pinder has since been modified. In December 1972, P. C. Trescott revised Pinder's program and made this new program available for testing and experimentation. This revised program (Trescott, 1973) was used for the digital model study of the aquifer system at Dayton, Ohio. The program was written in Fortran IV computer language and designed specifically for an IBM 360 computer.

The digital model has several capabilities. It can simulate the response of stresses in artesian aquifers, water-table aquifers, or a combination of the two. The aquifers can be irregular in shape and nonhomogeneous in character. The flow system in the aquifer is considered as two-dimensional. Source water is derived from aquifer storage, precipitation, flow across aquifer boundaries, or leakage through streambeds and confining beds. Discharge from the aquifer is by pumpage from one or more wells, evapotranspiration, and leakage to the stream.

Model Development

A map of the area was prepared showing the aquifer boundaries and the principal stream locations (fig. 1). The modeled area, 15.3 mi (24.6 km) in a north-south direction and 5.9 mi (9.5 km) in an east-west direction, covers approximately 90 mi^2 (233 km^2). A rectangular grid network, superimposed over the map consists of 27 rows and 74 columns, making a total of 1,998 square and rectangular cells. Most cells on the map are square and represent a distance of 1,000 ft (305 m) on a side. Each cell contains a node at its center. These nodes are points at which flow equations are evaluated even though the cell represents a volume of the aquifer through which flow is occurring.

An important consideration in developing a digital model is the decision on the size of the grid, as the total number of nodes is directly related to computer-storage and computation-time requirements. Cells located outside areas of significant pumping were made larger in order to limit the total number of nodes. This model with 1,998 nodes required approximately 200,000 bytes of computer storage, and the computation time for one computer run was between 10 and 14 minutes on the IBM 360 computer.

Conceptual Model

A conceptual model of the aquifer system was developed based on existing hydrologic information. The conceptual model consists of a number of simplifying assumptions which makes it possible to describe the aquifer mathematically. The assumptions do not exactly represent the real conditions; however, they represent, in concept, the hydrologic process being described.

The basic assumptions in the conceptual model of the Dayton aquifer are as follows:

(1) The glacial-outwash aquifer is a single unconfined aquifer except in the Rohrers Island area, where a till bed separates the aquifer into an upper unconfined unit and a lower semiconfined unit. These units coalesce downstream into the single unconfined aquifer as the till bed fingers out. In the Rohrers Island area, only the lower semiconfined aquifer was modeled because the digital model employed in this study can simulate only one transmissive layer. Pumpage from the upper aquifer in the Rohrers Island area was deleted from the total pumpage for the Dayton area. Thus, the full model simulated a single continuous aquifer which was represented as semiconfined in the Rohrers Island area and unconfined elsewhere.

(2) The aquifer is hydraulically connected to the Great Miami River and other principal streams except in the Rohrers Island area, where a constant hydraulic head is maintained above the semiconfining bed.

(3) The aquifer system is isotropic, and the flow in the aquifer is horizontal and two-dimensional.

(4) The bedrock forms an impermeable boundary.

(5) Recharge to the aquifer is from streambed leakage, boundary leakage, and precipitation that is uniformly distributed areally.

(6) Ground-water is discharged by pumping from wells and leakage to streams.

(7) Evapotranspiration from the ground-water reservoir is considered negligible since almost everywhere the potentiometric surface is below an effective depth at which evapotranspiration could occur.

(8) The average stream stage remains constant in all streams throughout the simulated period.

(9) In the confined section of the aquifer, water is derived from storage in the confining bed, as well as from storage in the aquifer itself, and from recharge from overlying sources. The release of water from storage in the confining bed is simulated using an approximation described by Bredehoeft and Pinder (1970).

Data Requirements

All parameters used to define the aquifer system are evaluated for each node and are considered representative of the whole cell. These various model parameters were then made into data sets and read into the computer as input data. Following is a list of the model parameters and their associated units. Only English units are shown here because the computer program, as it is written, will not accept metric units.

(1) Dimensions of the rectangular grid network, feet.

(2) Hydraulic conductivity of aquifer, feet per second.

(3) Altitude of top of aquifer, feet above mean sea level.

(4) Altitude of bottom of aquifer, feet above mean sea level.

(5) Storage coefficient, dimensionless.

(6) Specific storage, feet.

(7) Stage in streams and head in overlying aquifer, feet above mean sea level.

(8) Thickness of confining bed, feet.

(9) Hydraulic conductivity of confining bed, feet per second.

(10) Initial potentiometric surface in aquifer, feet above mean sea level.

(11) Recharge from precipitation, feet per second.

(12) Pumping rate, cubic feet per second.

Hydraulic Properties of the Aquifers

Transmissivity and Hydraulic Conductivity

The transmissivity, expressed in feet squared per day, is the product of the hydraulic conductivity, expressed in feet per day, and the saturated thickness, in feet. Stated in other terms, the transmissivity is the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient (Lohman, 1972.)

For artesian, or confined, aquifers the transmissivity remains a constant as long as the hydraulic head is not lowered beneath the top of the aquifer. In this instance, the saturated thickness remains the same and is calculated in the program as the difference between aquifer top and aquifer bottom. In water-table, or unconfined, aquifers the transmissivity changes in direct proportion to changes in the saturated thickness of the aquifer. Here, the saturated thickness is calculated by the computer as the difference in altitude between the water table and the aquifer bottom. The computer program checks the changes in saturated thickness at every node for each iteration and recalculates the transmissivity. Thus, in a dynamic water-table aquifer with variable stresses imposed on it, the values of transmissivity are continuously changing. Calculation of changes in transmissivity by conventional analytical methods would be virtually impossible.

An option available in the computer program permits simulation of aquifer systems which are confined in some areas and unconfined in others and which may change from confined to unconfined conditions within a given area as the water level is drawn below the top of the aquifer. If the aquifer is under water table conditions, data sets are required for the aquifer hydraulic conductivity, the altitude of the bottom of the aquifer, and the hydraulic head in the aquifer. For that part of the aquifer system that is artesian, data sets are required for the aquifer hydraulic conductivity and the altitude of the bottom and of the top of the aquifer.

The hydraulic conductivity used in this study was estimated from a few aquifer tests. Controlled aquifer tests are difficult to make in the Dayton area because of interference caused by variable pumping. Norris and Spieker (1966) report hydraulic conductivity values ranging from 0.15×10^{-2} to 0.46×10^{-2} ft/s (0.05×10^{-2} to 0.14×10^{-2} m/s). Other hydraulic conductivity values were estimated based on lithologic characteristics of the glacial deposits reported on drillers' logs.

The estimated hydraulic conductivity value used in the model for the water-table part of the aquifer was 0.116×10^{-2} ft/s (0.035×10^{-2} m/s) which is adjusted to compensate for till lenses in some parts of the aquifer. An average hydraulic conductivity value for the confined part of the aquifer in the Rohrers Island area was 0.193×10^{-2} ft/s (0.059×10^{-2} m/s). Values of hydraulic conductivity were recorded for each node and entered into the computer as a data set. Zero values around the perimeter of the grid network defined the model limits and were used to represent an impermeable, no-flow boundary.

Storage Properties

The storage coefficient is the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head (Lohman, 1972).

A uniform value for the storage coefficient of 2×10^{-1} was selected for the water-table part of the aquifer. This value is typical for glacial-outwash aquifers in Ohio. The storage coefficient for the artesian part of the aquifer at Rohrers Island was assumed to be 1×10^{-4} based on an aquifer test (Norris, 1959).

Different storage coefficient values were tested in the model to determine their effects on model output or results. The results showed that because pumpage was simulated for time periods of several years, the solution (head distribution) was relatively insensitive to the tested variations in the storage coefficient.

Specific storage is the volume of water released from or taken into storage per unit volume of the porous medium per unit change in head (Lohman, 1972). Specific storage of the confining bed is required for the calculation of rate of release of water from storage in the confining bed in the artesian part of the aquifer. An estimated specific storage of 2×10^{-5} /ft (0.61×10^{-5} /m) was used for all confining beds in the model.

Sources of Aquifer Recharge

Recharge from four sources was simulated in the digital model; (1) precipitation, (2) induced streambed leakage, (3) leakage through aquifer confining beds, and (4) leakage across aquifer boundaries.

Recharge From Precipitation

Precipitation varies from year to year and from season to season but in general is evenly distributed throughout the year. Even with fairly uniform precipitation the amount of recharge will vary seasonally as a result of other factors such as temperature, evapotranspiration, and rainfall intensity. Average precipitation at Dayton is 36.75 in (933 mm) per year.

The model did not simulate seasonal variation in precipitation. Recharge from precipitation was programmed into the model as a constant rate of 2×10^{-8} ft/s (0.6×10^{-8} m/s), equivalent to 7.5 in (191 mm) per year. This recharge rate was computed from estimates of recharge made by Norris and Spieker, 1966.

Induced Streambed Leakage

The principal source of recharge to the aquifer is from induced leakage through the streambeds of the Great Miami River and its tributaries. This leakage is induced by pumping ground water. The quantity of water leaking through the streambed depends on the hydraulic conductivity and thickness of the bed material, the area of the streambed, the temperature of the induced water, the stream stage, and the head in the aquifer. All of these factors except hydraulic conductivity will vary seasonally, and thus the leakage rate also will vary seasonally.

Hydraulic conductivity values of the streambed materials that were used in the digital model were based on results of direct measurements of seepage losses made in the Great Miami River and the Mad River during relatively low-flow conditions (Norris and Spieker, 1966). Values of seepage loss derived from this procedure are low values; under average conditions they would be higher. The seepage loss measurements showed a hydraulic conductivity rate of 0.072×10^{-5} to 6.0×10^{-5} ft/s (0.022×10^{-5} to 1.8×10^{-5} m/s).

A streambed thickness of 1 foot (0.3 m) was used at all stream nodes. The effective thickness of the streambed is unknown and could range from less than 1 in (25 mm) to a few feet, depending on local conditions.

The temperature of the water induced through the streambed was not considered in estimating the average leakage rate.

Throughout the simulation period the stream stage was modeled as a constant representing median flow conditions. The average stream stage at each node was estimated from altitudes on current topographic maps and measured cross sections at a stream discharge of about 1,000 ft^3/s (28 m^3/s). The mean discharge of the Great Miami River near the center of the model area is 2,039 ft^3/s (58 m^3/s). For the initial conditions three low-head dams on the Great Miami River were assumed to be in existence even though this was not true prior to significant ground-water development. The dams cause the stream slope to be relatively flat and to change abruptly at the dams.

Another factor is the area of the streambed relative to the size of the nodal cell. Most cells in the digital model are 1,000 ft (305 m) square, whereas the maximum stream width is about 400 ft (122 m). Leakage through the streambed is calculated in the model over the entire area of the cell through which the stream passes. Rather than rewrite the model program to handle streambed areas different from nodal areas, the hydraulic conductivity values for the streambed were adjusted in proportion to the effective area.

The top of the aquifer at each stream node was taken as 5 ft (1.5 m) below the stream stage. At all other nodes in the water-table part of the aquifer, the top of the aquifer was taken as the top of the saturated section and was recomputed at each step of calculation. During simulation, the aquifer top remained fixed at the stream nodes until the head in the aquifer dropped below the bottom of the streambed, at which time the stream nodes converted to conventional water-table nodes, and the rate of leakage from the stream became constant.

Initial values of head in the aquifer were computed in an equilibrium analysis, to be described later, and were generally above the local values of stream stage. In the pumping simulations, also to be described subsequently, leakage into the aquifer could occur only when the head in the aquifer fell below the stream stage. When a head change of 5 ft (1.5 m) or more occurred in the aquifer at the stream nodes, a maximum leakage rate was attained. Leakage was allowed to increase only up to this predetermined level.

Leakage Through Aquifer Confining Bed

In the Rohrers Island area two aquifers separated by a semiconfining bed exist. The lower aquifer is artesian and is recharged by leakage from the upper aquifer. Water leaks through the confining bed whenever the head in the upper aquifer is higher than the head in the lower aquifer. The amount of leakage to the artesian aquifer depends on the difference in hydraulic head between the two aquifers and on the hydraulic conductivity and thickness of the confining bed. Only the lower aquifer was actually simulated in the model in this area. The upper aquifer was considered only as a source of leakage to the lower flow system; for this purpose, constant values of head for the upper aquifer were specified at each model node in the Rohrers Island area.

Accurate values of hydraulic conductivity for a confining bed such as till or clay are difficult to determine for a large area. Most data are collected from small areas such as well fields. Norris (1959) describes methods used for determining hydraulic conductivity of the till beds at Rohrers Island; determinations by these methods range from 0.46×10^{-7} to 2.0×10^{-7} ft/s (0.14×10^{-7} to 0.61×10^{-7} m/s). Based on these values a uniform hydraulic conductivity rate of 0.775×10^{-7} ft/s (0.236×10^{-7} m/s) was selected to represent average conditions for the confining bed at Rohrers Island.

The confining bed was considered as being continuous; however, the thickness is known to vary. Norris (1959) reported that the thickness ranges from 11 to 58 feet (3 to 18 m). For the digital-model simulation, thickness values were selected from 25 to 50 ft (8 to 15 m), with an average of about 40 ft (12 m). The thinner sections normally occur near the bedrock boundaries, where the upper aquifer is thin or missing.

In the field situation, if the hydraulic head in the lower artesian aquifer is lowered below the bottom of the confining bed the aquifer will convert to water-table conditions and the amount of leakage will not further increase. This conversion is simulated in the digital model; it can occur at one or more artesian aquifer nodes and the aquifer may change back and forth from artesian to water table in the course of the simulation.

Leakage Across Aquifer Boundaries

The aquifer is bounded on the bottom and partly on the sides by impermeable limestone and shale bedrock. Upland glacial deposits, consisting mostly of till and clay, overlie the bedrock along the perimeter or valley walls and provide recharge by lateral movement of water from the upland to the valleys. Recharge rates were estimated based on a study by Walton and Scudder (1960) in the Fairborn, Ohio, area. Fairborn lies northeast of the modeled area along the Mad River and recharge conditions are somewhat similar. However, the author believes that recharge rates at Dayton are higher than at Fairborn. The maximum recharge rate was estimated to range from 0.82 to 2.0 ft^3/s (0.02 to 0.06 m^3/s) per valley-wall mile, depending on the character of the upland materials.

The lateral recharge into the aquifer was simulated by treating the boundary (first node inside model boundary) as a stream node. That is, each node along the boundary was assigned hypothetical values of stream head, streambed hydraulic conductivity, streambed thickness, and limiting seepage. These parameters were selected in such a way as to generate leakage into each boundary node in amounts roughly equal to the estimated lateral recharge through the valley walls, in the area represented by the node. This technique was used for convenience, to avoid programming a separate subroutine for computation of lateral inflow.

The hypothetical value of "stream head" used at these boundary nodes was taken as 5 ft (1.5 m) above the top of the aquifer. The position of the aquifer top, in turn, was selected on the basis of topographic altitude at the node, and was held fixed at these boundary nodes as it was at the actual stream nodes. The hypothetical "streambed hydraulic conductivity" was given values between the limits of 0.31×10^{-7} to 0.775×10^{-7} ft/s (0.094×10^{-7} to 0.236×10^{-7} m/s), according to the estimated lateral inflow at the node. The hypothetical "streambed thickness" was 1 ft (0.3 m), with the maximum or limiting seepage occurring with 5 ft (1.5 m) of head difference. That is, when the head at any boundary node is lowered by 5 ft (1.5 m), maximum recharge at that node is attained and further lowering of the head produces no additional increase in recharge. Once this condition is reached, the aquifer top becomes the same as the top of the saturated section at that node, and thus it no longer remains fixed.

Initial Potentiometric Surface

The initial potentiometric surface is the one that existed before groundwater development. Data on the initial potentiometric surface are not available in the Dayton area; however, by using a few basic assumptions an initial surface was simulated.

Before significant development, it was assumed that ground-water inflow to the aquifer was approximately equal to the outflow and that the system was under steady-state conditions. Steady-state conditions can be simulated in the model by setting the storage coefficient of the aquifers and the specific storage of the confining beds to zero, computing one time step of any length and iterating to a solution (Trescott, 1973). On applying this technique, the "stream heads" at boundary nodes were fixed, according to topography; and the stream stages at the actual stream nodes were also fixed, using stream cross section data. With these values as boundary conditions, starting values of head were inserted at all other nodes of the model, and the simulation was run at equilibrium. The result was a set of hypothetical heads within the aquifer which were compatible, in a steady-state flow pattern, with the heads along the boundaries and along the streams.

The hypothetical head distribution is shown in figure 3. As can be seen, it represents the normal configuration that would be expected for flow from an upland area to a draining stream and therefore was accepted as a reasonable representation of the initial potentiometric surface. For the above simulation technique to generate the initial potentiometric surface, the values selected for all parameters used in the model must be reasonably correct. A trial and error procedure is usually required until the system responds satisfactorily.

Aquifer Discharge

Pumpage

Complete ground-water pumpage records are usually unavailable for industrialized areas such as Dayton, Ohio. Most ground-water users do not meter their wells. Only the larger users such as the Dayton Water Department and other public supply and industrial users maintain reasonably good records.

Pumpage from the upper aquifer at Rohrers Island, which accounts for about 25 percent of the total ground-water pumpage for the study area, was not included because the model simulates only the lower aquifer in this area. Pumpage from the lower aquifer at Rohrers Island, however, was included. Estimates of pumpage from the West Carrollton area were added to the records compiled by Norris and Spieker (1966).

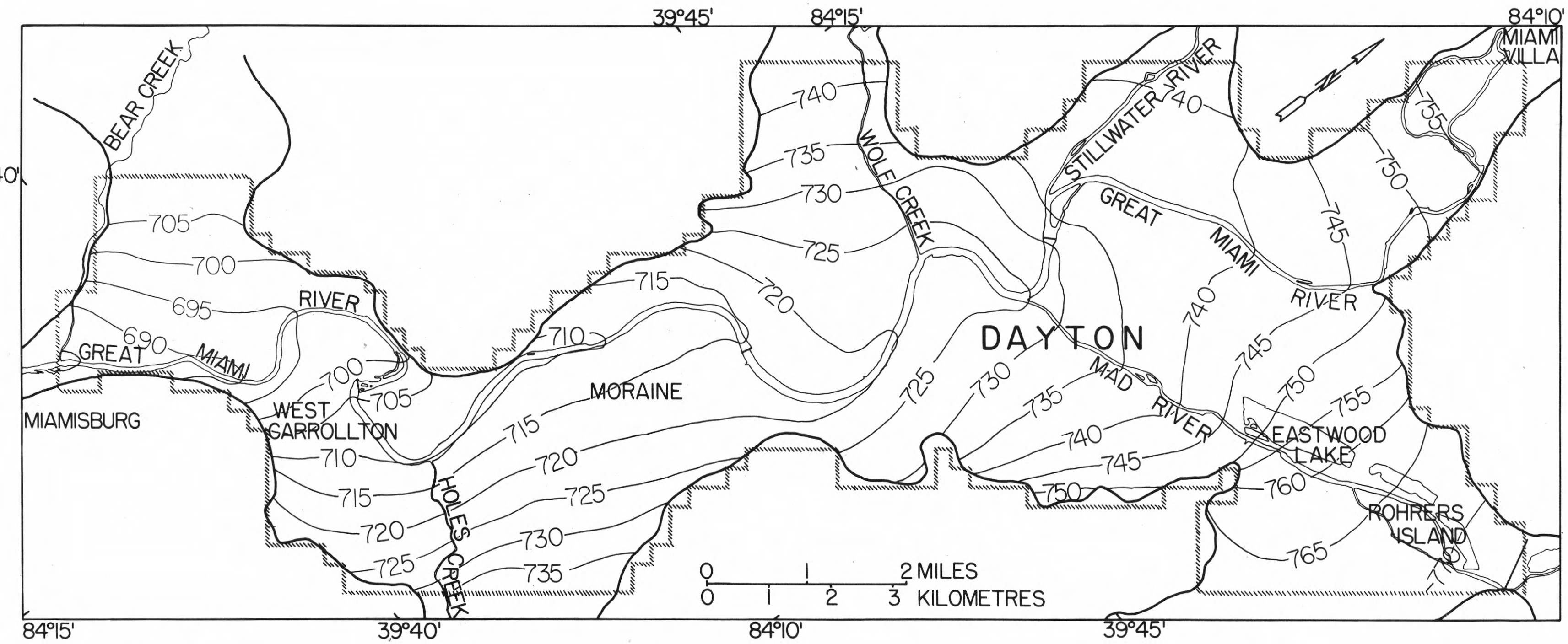
Norris and Spieker (1966) compiled estimates of pumpage for 10-year intervals covering the years 1900-50. A water-use inventory was made in 1946 (Norris and others, 1948). Relatively good records are available for 1958, which were based on a water-use inventory made in 1954-55 by the Miami Conservancy District.

In compiling pumpage records for the digital model it was necessary to combine yields from two or more wells in one cell area. Wells located within the same cell were represented in the model as one well pumping at a constant rate over the specified time interval. As noted by Prickett and Lonnquist (1971), a simulated well has an effective diameter equal approximately to $a/4.81$ where a is the finite-difference grid interval. Thus, the drawdown for a particular node would be less than that observed in a real well at that location.

Ground-water pumpage was separated into five pumping periods through 1960 (fig. 4). The time intervals used were based on the availability of pumpage records.

The first pumping period was made to correspond to ground-water pumpage before 1941. The time interval used for this pumping period was determined by allowing simulated pumping to continue at a constant rate until such time that steady-state conditions were reached -- that is, no further head changes occurred in the aquifer with time. The assumption is that steady-state conditions actually did exist in 1940 and that not until the mid-1940's did water levels generally decline in the Dayton area.

The other four pumping periods selected were 1941-46, 1947-50, 1951-54, and 1955-60.



EXPLANATION:

Approximate aquifer boundary

Simulated aquifer boundary

—715— Potentiometric contour, shows altitude of potentiometric surface. Contour level 5 feet (1.5 metres) datum is mean sea level.

Figure 3.--Hypothetical potentiometric surface of the glacial-outwash aquifer before ground-water development.

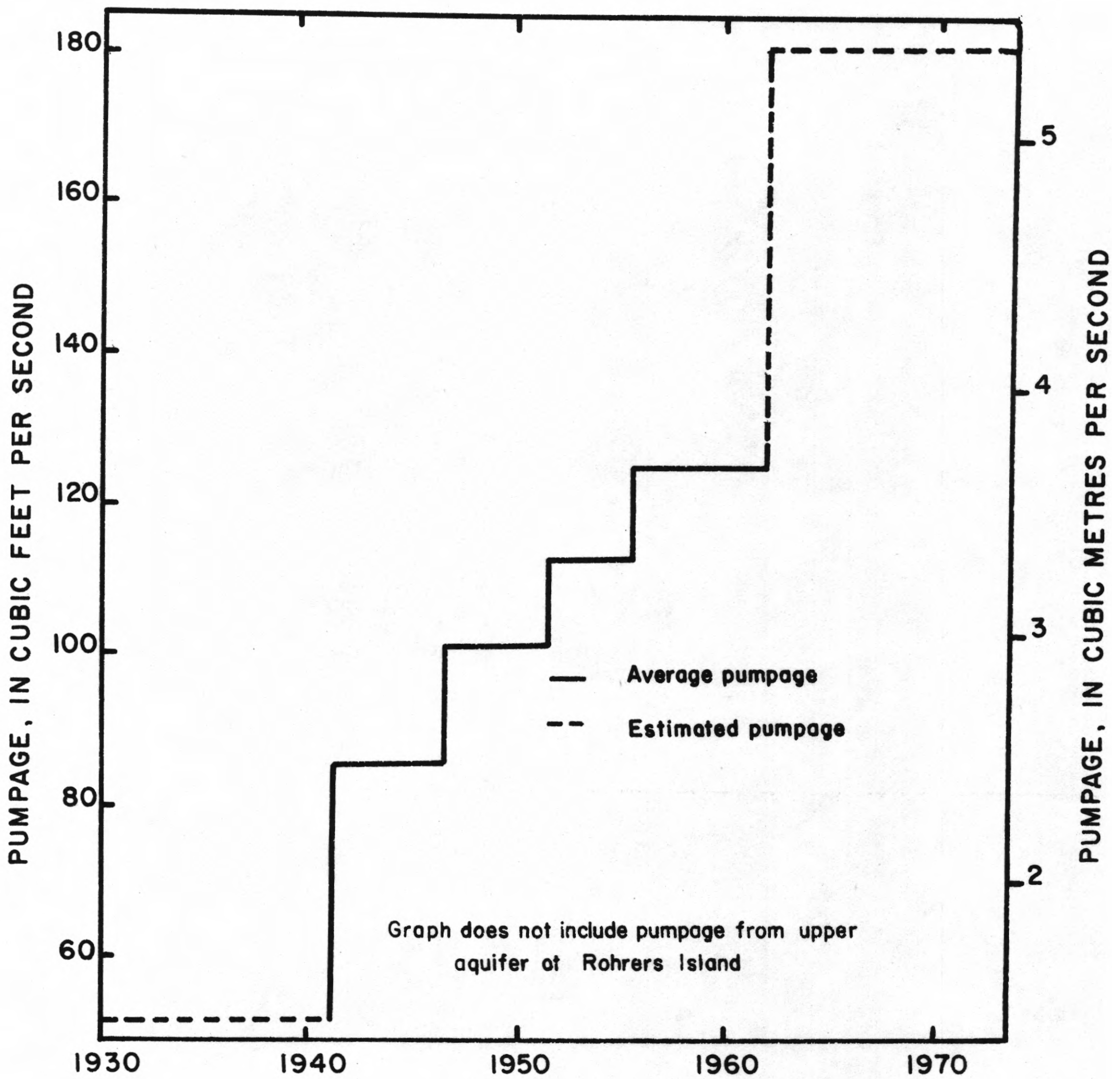


Figure 4.--Average ground-water pumpage used for digital-model simulation.

The following is a list of the total ground-water pumping rates for the area and the number of nodes used to simulate wells for each pumping period.

Pumping period	Number of nodes	Pumping rates (ft^3/s)	(m^3/s)
Before 1941	49	53.85	1.53
1941-46	62	89.03	2.52
1947-50	65	105.84	3.00
1951-54	75	117.62	3.33
1955-60	75	129.52	3.67
1961-72 (estimated)	85	187.49	5.31

Pumping at each node was entered into the model as data sets by designating the grid coordinates for the node and the pumping rate, in cubic feet per second. Locations of nodes used are shown in figures 5 and 7.

Leakage to streams

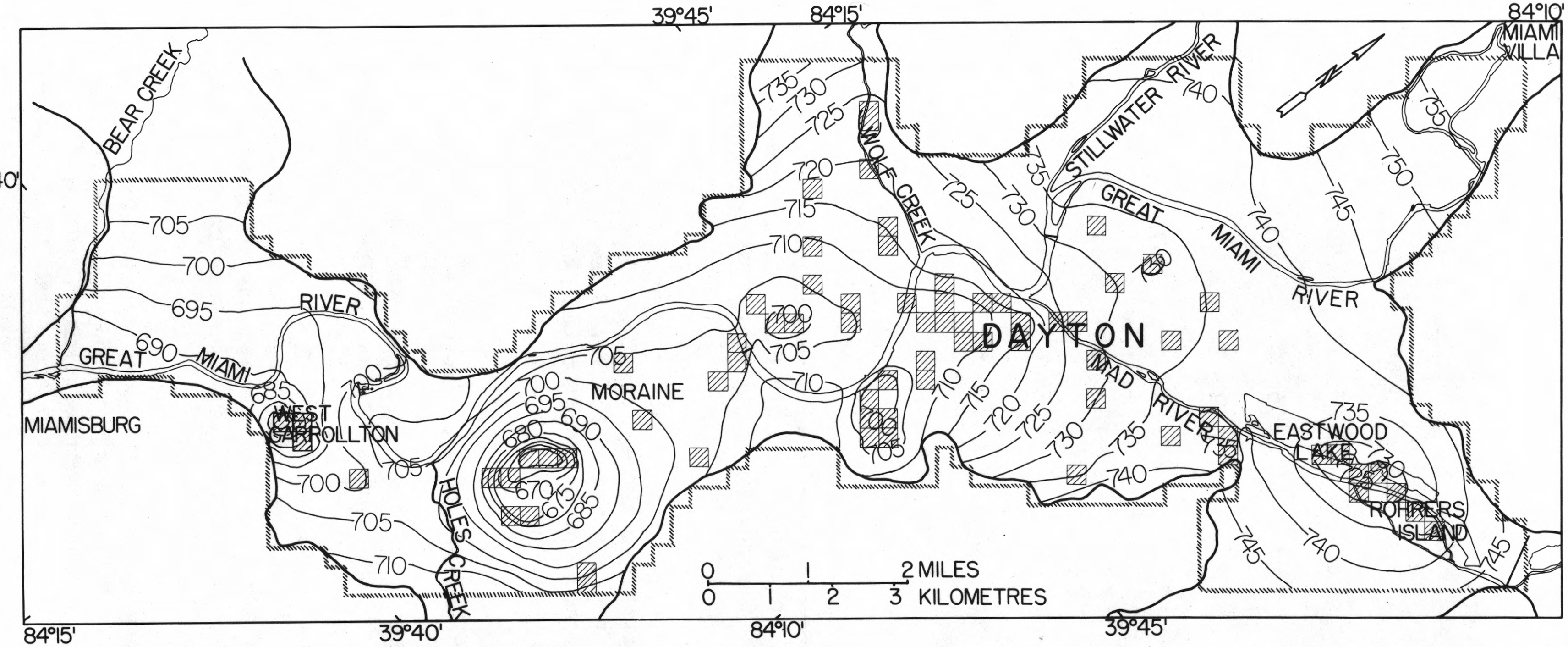
Under natural conditions, when the head in the aquifer is below the stream stage, water from the stream moves into the aquifer. When the head in the aquifer is above the stream stage, water from the aquifer discharges to the stream. Both recharging and discharging conditions may occur in a steady-state system.

When pumping occurs in an aquifer and significant quantities of water have been removed from aquifer storage, the gradient is usually away from the stream, and the aquifer is recharged by water leaking from the stream. However, in some areas, where pumpage is small or does not occur, conditions remain nearly natural and water continues to discharge from the aquifer to the stream.

CALIBRATION OF THE DIGITAL MODEL

After data sets were prepared for quantification of each parameter, a computer run was made to calculate the hydraulic head at each node in the model at the end of each pumping period. Water levels in the Dayton area fluctuate seasonally as a result of variable pumpage and other natural stresses, such as changes in stream discharge. If is not possible to simulate seasonal fluctuations when time increments of more than 1 year are used. Therefore, head values determined by the model are assumed to represent average water-level conditions for any selected time period.

Calculated head values were printed by the computer for the end of each simulated pumping period, then plotted on a map of the area and contoured to show the relative position of the potentiometric surface. Figure 5 is a contour map of the potentiometric surface, as calculated from the model. These contours



EXPLANATION:

Approximate aquifer boundary

Simulated aquifer boundary

Pumping cell, represents pumping from one or more wells.

-715- Potentiometric contour, contour interval 5 feet, datum is mean sea level.

Figure 5.--Potentiometric surface of the glacial-outwash aquifer, 1960, based on computer calculations.

represent average water-level conditions in 1960. Figure 6 shows the decline in water levels from the hypothetical initial potentiometric surface (fig. 3) to 1960.

Norris and Spieker (1966) collected data on water levels by measuring approximately 60 wells twice annually in 1958-60. They used these measurements, along with published water-level measurements for other wells, to prepare two potentiometric-contour maps which reflect the period of highest and lowest water levels measured. Contours from both maps are traced onto figure 7 and they represent conditions in April 1959 and October 1960. The October 1960 map generally shows water levels lower than the April 1959 map, but the differences are not uniform throughout the area. Because these are the only potentiometric-contour maps available, they were used to calibrate the digital model. An attempt was made to simulate hydraulic heads in the model that would equal or fall between the values shown on the two potentiometric-contour maps. When this condition was met the results were considered satisfactory.

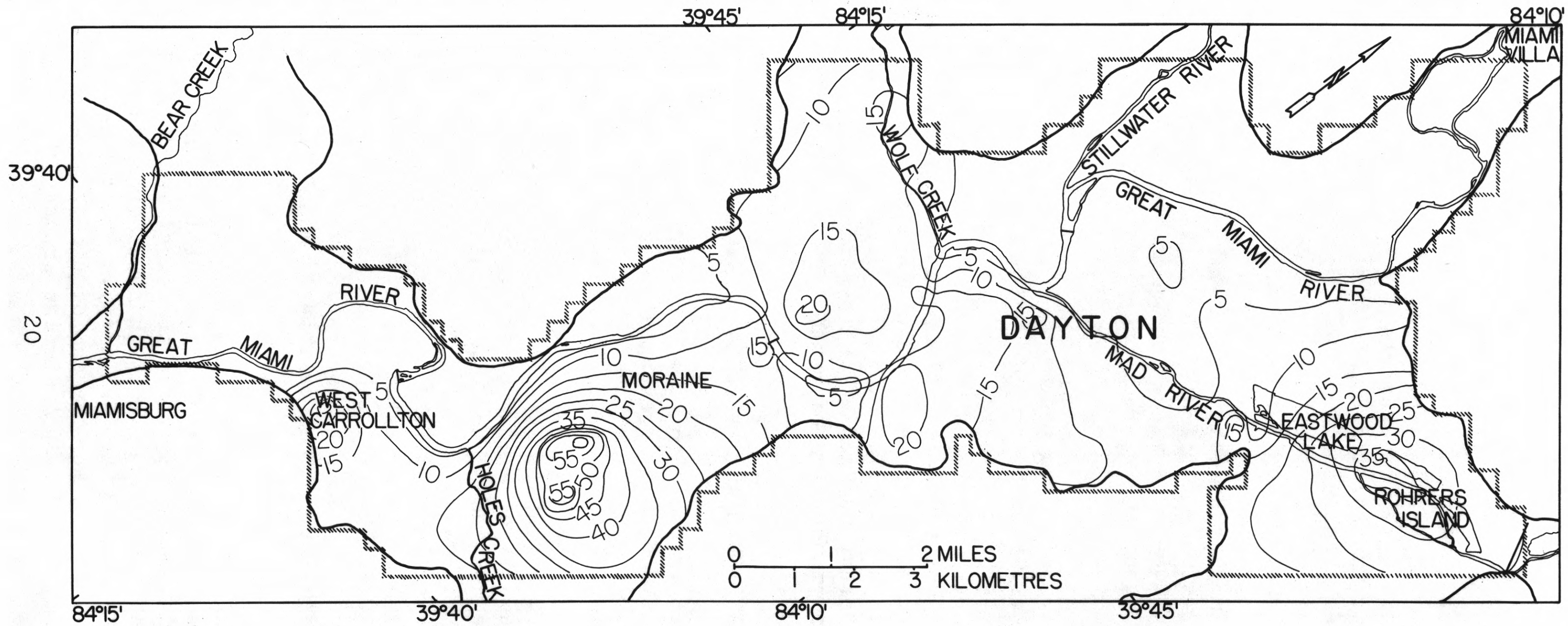
Several computer runs were required to test the validity of the hydrologic parameters. These tests were made by comparing the calculated hydraulic head values after each computer run with the potentiometric maps for April 1959 and October 1960. When head values did not meet the criterion, adjustments were made to the parameter values. The parameter requiring the most frequent adjustment in order to simulate the real condition was the hydraulic conductivity of the streambed. The hydraulic conductivity affects the leakage rate which is the poorest defined of the hydrologic properties because of seasonal variation and inaccuracy of parameter constants.

THE DIGITAL MODEL AS A PREDICTIVE TOOL

Once the model is calibrated it can be used as a management tool to predict the effects on water levels caused by changes in ground-water pumpage.

A test was made by estimating the ground-water pumpage for 1961-72 and making a computer run to calculate the hydraulic head at each node at the end of the period. This computer run was designated as pumping period 6, since it was a continuation of the previous five pumping periods. Only the pumping rates and the time interval were changed. From the calculated hydraulic head values a potentiometric-contour map was prepared (fig. 8) that represents the predicted average potentiometric surface in 1972. A contour map prepared for 1972 (fig. 9) shows the water-level decline relative to the pre-development potentiometric surface shown in fig. 3. The results compared favorably with the few available water-level measurements at observation wells and provide an independent check on the credibility of the calibrated model which was based on the record up to 1960.

Ground-water pumpage estimates for 1961-72 were made from water-use inventories prepared by the Miami Conservancy District for 1968 and 1971. The water-use inventory included 5 municipal water supplies, 10 of the largest industrial water users and an estimated combined withdrawal from about 85 ground-water users pumping less than 1 ft^3/s (0.3 m^3/s). The total average pumping rate (fig. 4) reflects the rate near the end of the 12-year period and probably is the best estimate available. A similar procedure could be



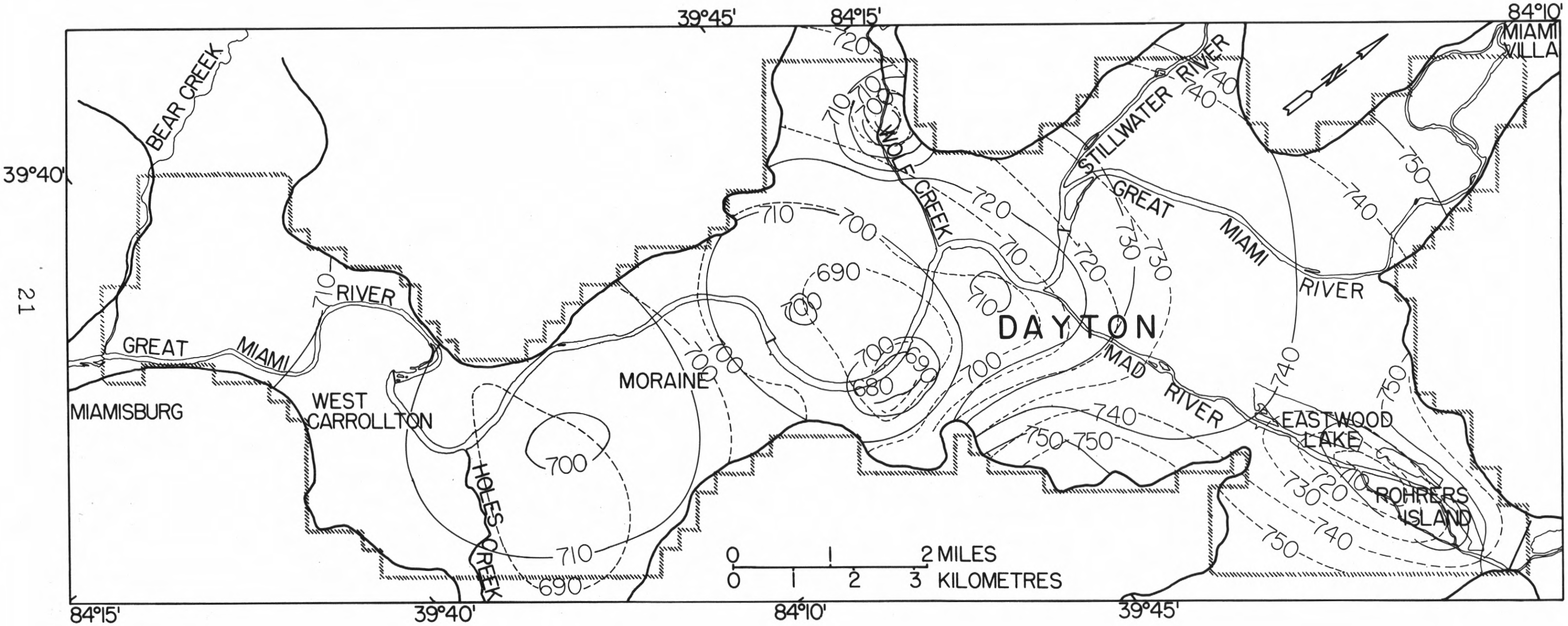
EXPLANATION:

Approximate aquifer boundary

Simulated aquifer boundary

—25— Line of equal water-level decline, interval 5 feet (1.5 metres)

Figure 6.--Decline in water level in the glacial-outwash aquifer, from beginning of ground-water development to 1960.



EXPLANATION:

Approximate aquifer boundary

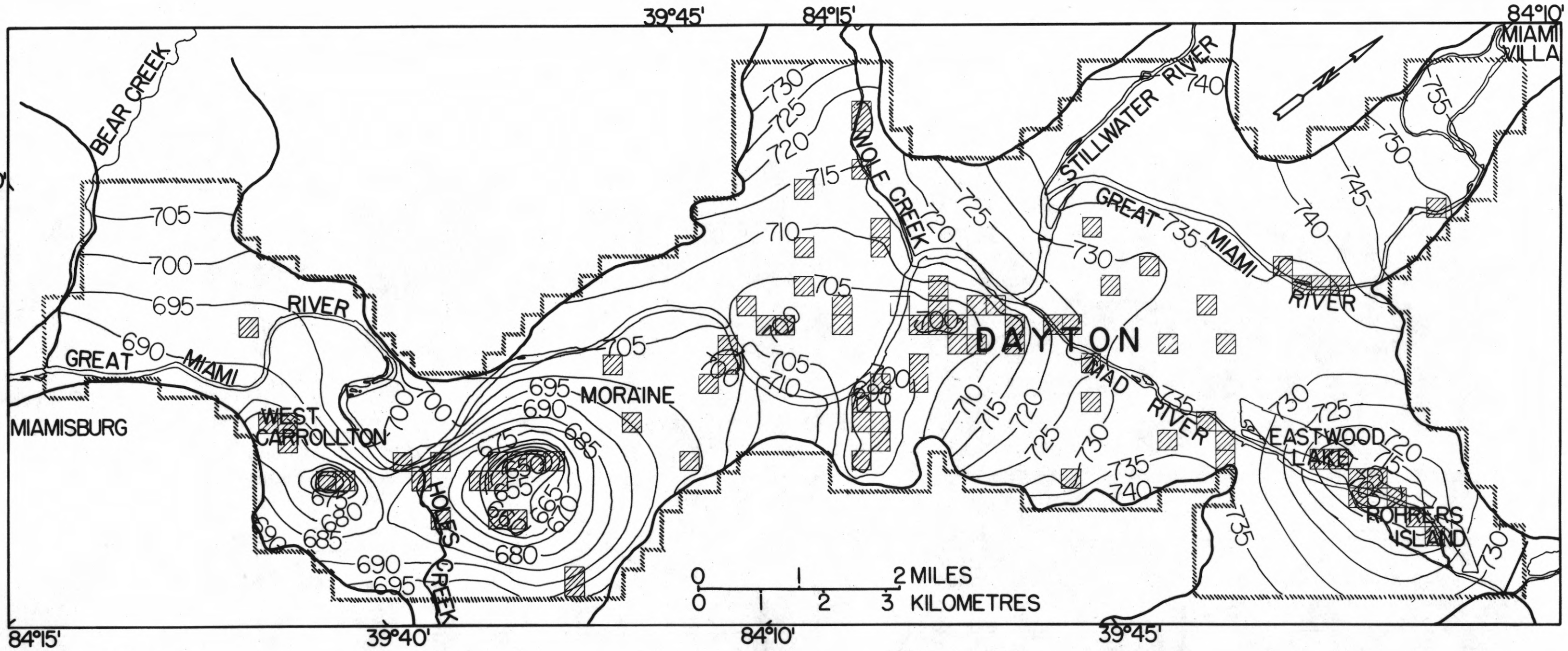
Simulated aquifer boundary

750 April 1959

750 October 1960

Potentiometric contours; contour interval 10 feet (3 metres), datum is mean sea level.

Figure 7.--Potentiometric surface of the glacial-outwash aquifer in April 1959 and October 1960, based on measurements in wells. (After Norris and Spieker, 1966.)



EXPLANATION:

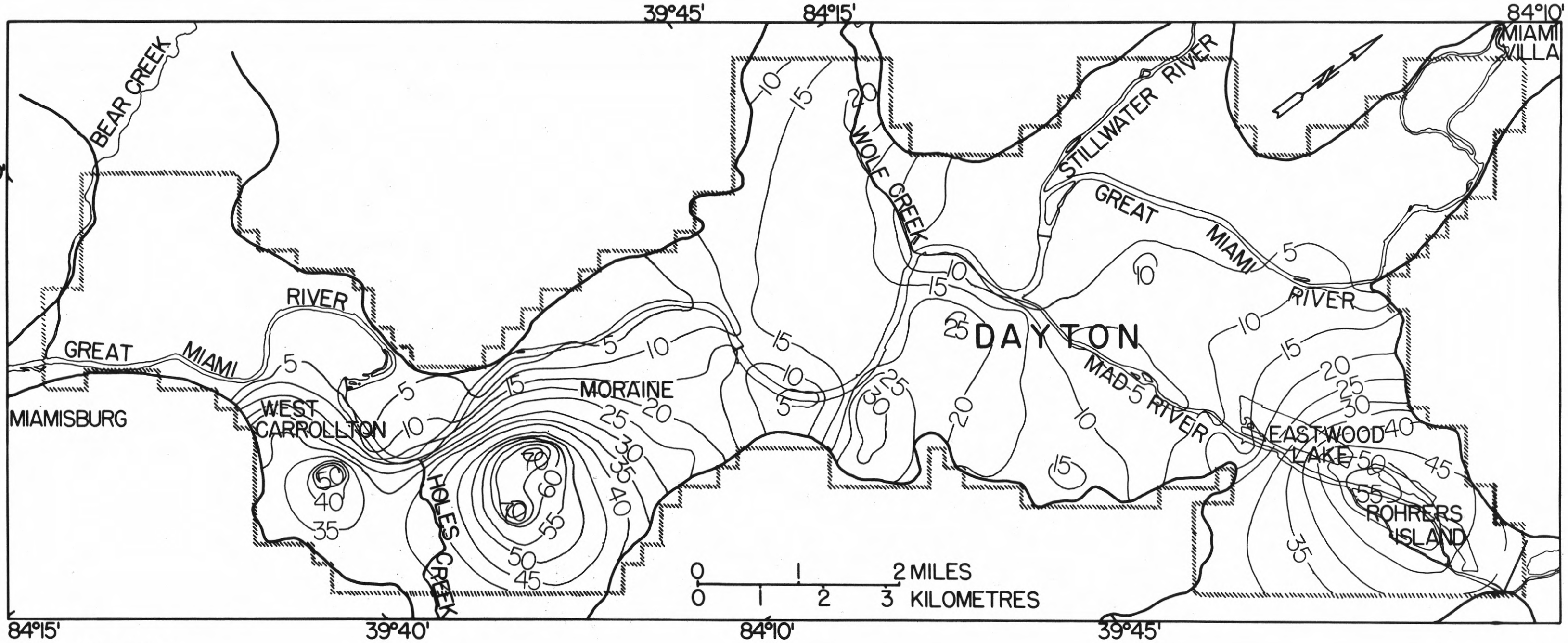
Approximate aquifer boundary

Simulated aquifer boundary

Pumping cell, represents pumping from one or more wells.

—750— Potentiometric contour, contour interval 5 feet (1.5 metres), datum is mean sea level

Figure 8.--Potentiometric surface of the glacial-outwash aquifer, 1972, based on computer calculations.



EXPLANATION:

Approximate aquifer boundary

Simulated aquifer boundary

—5— Line of equal water-level decline, interval 5 feet (1.5 metres)

Figure 9.--Decline in water level in the glacial-outwash aquifer from beginning of ground-water development to 1972.

used to predict pumping effects on water levels at any future time period desired. The only variables that would need to be changed in the model are the pumping rates and the projected time.

CONCLUSIONS

Average hydraulic head values for 1960 were calculated for each node in the digital model in response to ground-water pumpage over five pumping periods. The simulated head values generally are in good agreement with potentiometric-contour maps for April 1959 and October 1960 compiled by Norris and Spieker (1966).

The model as presently developed should be considered a first step in the modeling of the aquifer system at Dayton. A continued updating procedure would produce an improved model as more basic data are collected.

The types of data that are needed to develop this sort of model are known and therefore an increased effort could now be made to improve the collection and utilization of these data. Careful evaluation of drillers' logs would lead to improved values for the transmissivity and hydraulic conductivity. Pumpage records already collected could be used to test the model through the most recent data collection period. Additional seepage-loss measurements at various discharge levels in the Great Miami River and other streams would, no doubt, provide a better evaluation of recharge from induced streambed leakage.

Additional surveys of water levels in wells in the Dayton area, if made periodically, will enhance the observation-well program and provide data for up to date potentiometric contour maps. Without these contour maps for the whole area, the validity of the model calibration is questionable.

Mathematical modeling is becoming an accepted way to evaluate aquifer systems. Models can be used as management tools to show how the local hydrologic system functions and to forecast aquifer response to any anticipated future development. This is particularly advantageous for an area such as Dayton, which depends on ground water as its principal source of water.

Modeling techniques are being advanced rapidly, along with advances in computer technology. However, some present model limitations are computer-related -- limitations in storage capability and use of excessive computation time. Model simulation techniques are already being tested by the U.S. Geological Survey for multiple-aquifer studies and for ground-water-quality studies, which may become increasingly important for the Dayton area.

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