HYDROGEOLOGY AND SIMULATION OF GROUND-WATER FLOW AT THE SOUTH WELL FIELD, COLUMBUS, OHIO

by William L. Cunningham, E. Scott Bair, and William P. Yost

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
inch (in)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
foot per mile (ft/mi)	0.1894	meter per kilometer
square mile (mi^2)	2.590	square kilometer
foot per day (ft/d)	0.3048	meter per day
gallon per day (gal/d)	3.785	liter per day
million gallons per day (Mgal/d)	3,785	cubic meter per day

Water and air temperatures in degrees Celsius (^0C) can be converted to degrees Fahrenheit (^0F) by the following equation: $^0F = 1.8\ (^0C) + 32$

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)—a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

Hydrogeology and Simulation of Ground-Water Flow at the South Well Field, Columbus, Ohio

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Abstract

The City of Columbus, Ohio, operates four radial collector wells in southern Franklin County. The "South Well Field" is completed in permeable outwash and ice-contact deposits, upon which flow the Scioto River and Big Walnut Creek. The wells are designed to yield approximately 42 million gallons per day; part of that yield results from induced infiltration of surface water from the Scioto River and Big Walnut Creek. The well field supplied up to 30 percent of the water supply of southern Columbus and its suburbs in 1991. This report describes the hydrogeology of southern Franklin County and a transient three-dimensional, numerical ground-water-flow model of the South Well Field.

The primary source of ground water in the study area is the glacial drift aquifer. The glacial drift is composed of sand, gravel, and clay deposited during the Illinoian and Wisconsinan glaciations. In general, thick deposits of till containing lenses of sand and gravel dominate the drift in the area west of the Scioto River. The thickest and most productive parts of the glacial drift aquifer are in the buried valleys in the central and eastern parts of the study area underlying the Scioto River and Big Walnut Creek. Horizontal hydraulic conductivity of the glacial drift aquifer differs spatially and ranges from 30 to 375 feet per day. The specific yield ranges from 0.12 to 0.30.

The secondary source of ground water within the study area is the underlying carbonate bedrock aquifer, which consists of Silurian and Devonian limestones, dolomites, and shales. The horizontal hydraulic conductivity of the carbonate bedrock aquifer ranges from 10 to 15 feet per day. The storage coefficient is about 0.0002.

The ground-water-flow system in the South Well Field area is recharged by precipitation, regional ground-water flow, and induced stream infiltration. Yearly recharge rates varied spatially and ranged from 4.0 to 12.0 inches.

The three-dimensional, ground-water-flow model was constructed by use of the U.S. Geological Survey three-dimensional finite-difference ground-water-flow code. Recharge, boundary flux, and river leakage are the principal sources of water to the flow system. The study area is bounded on the north and south by streamlines, with flow entering the area from the east and west. Areal recharge is contributed throughout the study area, although a comparatively high percentage of precipitation reaches the water table in the area east of the Scioto River where little surface drainage exists. Ground-water flow is downward in the uplands of the Scioto River, and upward near the river in the glacial drift and carbonate bedrock aquifers.

The numerical model contains 53 rows, 45 columns, and 3 layers. The uppermost two layers represent the glacial drift. The bottom layer represents the carbonate bedrock. The horizontal model grid is variably spaced to account for differences in available data and to simulate heads accurately in specific areas of interest. The length and width of grid cells range from 200 to 2,000 feet; the finer spacings are designed to increase detail in the areas near the collector wells. The model uses 7,155 active nodes.

Measurements of water levels from October 1979 were used to represent steady-state

conditions before municipal pumping at the well field began. Measurements made during March 1986 were used to represent steady-state conditions after commencement of pumping at the well field. Water levels measured during March 1986 - June 1991 were used for calibration targets in the transient simulations.

The transient model was discretized into eight stress periods of 93 to 487 days on the basis of recharge, well-field pumpage, and available water-level data. Transient model calibration was based on seven sets of hydraulic-head measurements made during March 1986 - June 1991. This time period includes large-scale increases in well-field production associated with a drought in the summer of 1988, and a period of exceptionally high rainfall in 1990.

The ground-water-flow model was calibrated under steady-state and transient conditions by use of the hydraulic head and stream gain/loss data as calibration targets. The goodness-of-fit of a particular suite of input parameter values was evaluated visually by comparing contour maps of simulated versus measured hydraulic heads in each model layer, and quantitatively by computing summary statistics of the residuals between simulated and measured hydraulic heads. The mean absolute error and root mean square error were 2.3 and 3.8 feet for the October 1979 calibration and were 3.0 and 4.8 feet for the March 1986 calibration. For the transient simulation, the mean absolute error of the stress periods ranged from 3.1 to 5.5 feet. The root mean square error ranged from 3.6 to 7.0 feet. Maximum root mean square error was equal to about 10 percent of the saturated thickness of the glacial aquifer. As an additional calibration step, hydrographs were matched for wells in several areas of the model for which long-term water-level data were available.

The calibrated transient model is not highly sensitive to increases in hydraulic conductivity of layers 1 and 2 and changes in recharge, storage coefficient, or specific yield. The model is more sensitive to decreases in hydraulic conductivity, transmissivity, vertical conductance, and river stage. The model is most sensitive to changes in riverbed conductance.

The steady-state and transient models will be used in a subsequent study to determine the steady-state and transient contributing recharge areas to the collector wells.

INTRODUCTION

The City of Columbus, Ohio, operates four radial collector wells in southern Franklin County. The "South Well Field" is completed in permeable outwash and ice-contact deposits that are intersected by the Scioto River and Big Walnut Creek. The wells were designed to yield approximately 42 Mgal/d; part of that yield results from induced infiltration of surface water from the Scioto River and Big Walnut Creek. The well field supplied up to 30 percent of the water supply of southern Columbus and its suburbs in 1991.

Investigations in the South Well Field began in the late 1960's and 1970's to characterize the hydrogeology of the area. Ranney Water Systems (1970) and Stilson and Associates (1976) defined land needs and estimated production capabilities in the designated well field area bordered by State Route 104 to the west, Interstate-270 to the north, Lockbourne Road to the east, and State Route 665 to the south. Since that time, the U.S. Geological Survey (USGS) has performed hydrogeologic investigations in the vicinity of the well field, including modeling studies (Bloyd, 1974; Weiss and Razem, 1980; Razem, 1983; Bair and others, 1990; Eberts and Bair, 1990), geochemical studies (de Roche and Razem, 1981; de Roche and Razem, 1984; de Roche, 1985), and a combination thereof (Sedam and others, 1989; Childress and others, 1991).

These studies led to a characterization of steady-state flow in the glacial and bedrock aquifers and the interaction of these aquifers with the Scioto River and Big Walnut Creek. Past, current, and future ground-water fluctuations due to quarrying and other land-use changes, however, demonstrate that over long time periods the ground-water-flow system is transient in character. Razem (1983) simulated transient ground-water flow in the glacial drift aquifer by use of a two-dimensional finite-difference flow model. This study expands the previous investigation to three dimensions by including the flow contributed by the underlying bedrock aquifer, and examines the influence of transient-flow characteristics such as precipitation, streamflow, and pumping stress on drawdowns produced in the well field.

Purpose and Scope

The purpose of this report is to describe the hydrogeology and transient ground-water-flow system at southern Franklin County and the simulation of ground-water flow in the vicinity of Columbus' South Well Field. Specifically, the report describes the (1) hydrogeology of the study area, (2) assumptions made to represent the conceptual model of the flow system within the framework of the mathematical model, including justification of the assigned boundary conditions, (3) model-calibration process, and (4) results of simulations of steady-state and transient conditions. The scope of the study is limited to analysis of the shallow flow system in the glacial drift and carbonate bedrock aquifers.

Study Area

The 25.4-mi^2 study area is located in southern Franklin County, Ohio, in parts of Hamilton and Jackson Townships and the City of Columbus. The South Well Field is located south of Columbus between the Scioto River and Big Walnut Creek as shown on figure 1.

The area is characterized by generally flat topography with slopes of 40 to 70 ft/mi toward the major streams. The primary land use in the study area is agriculture (corn and soybeans). Other land uses include aggregate operations for sand, gravel, and limestone; transportation routes; residential housing; commercial areas; and light industry.

The climate of the area is moderate. Average annual temperature is 52° F and average annual precipitation is 37 in. There are approximately 171 days without killing frost; the growing season lasts from late April through mid-October (National Oceanic and Atmospheric Administration, 1989).

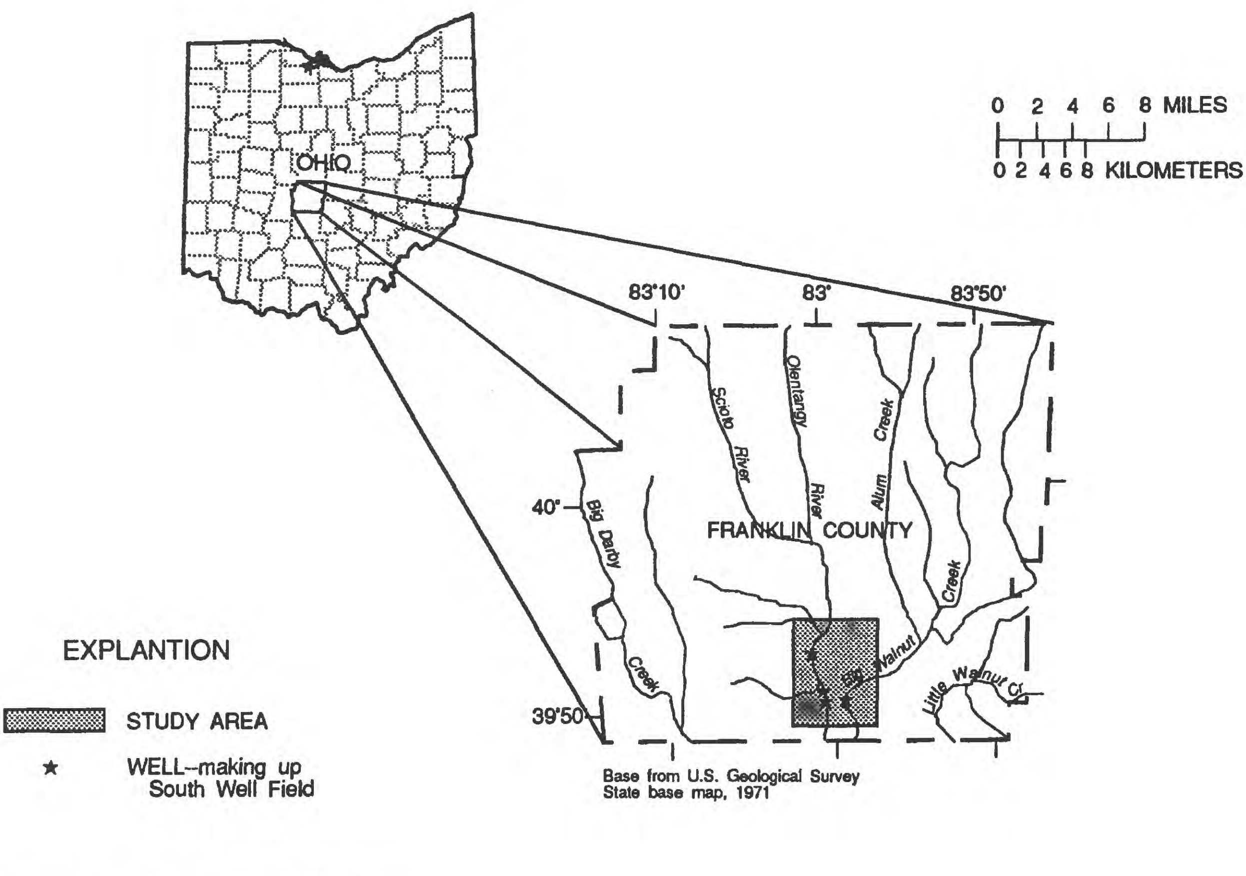

Figure 1. Location of study area.

Acknowledgments

The authors are grateful for the assistance from the City of Columbus personnel at Parsons Avenue Water Treatment Facility for access to the well field and associated data. David Parkinson of the Ohio Environmental Protection Agency provided outfall-volume data for flows into Big Walnut Creek and Scioto River. The authors also acknowledge the cooperation of area homeowners and businesses for access to their wells.

METHODS OF STUDY

The objectives of the study were met by collection of field data such as ground-water levels, precipitation, streamflow, and streambed permeability. These data were incorporated into a three-dimensional numerical ground-water-flow model.

Data Collection

A data-collection network was used to obtain information needed to construct a numerical ground-water-flow model of the glacial drift and carbonate bedrock aquifers and to assess their interaction with the surface-water-flow system. Results of previous aquifer testing, values of input parameters used in previous flow models, and water-level measurements from previous studies were evaluated for use in the numerical ground-water-flow model. Several sets of synoptic water levels were measured at times of differing stress on the flow system for use in the calibration of the transient flow model.

The locations of 74 wells used to make water-level measurements since 1979 are shown in figure 2. Some of these wells have been destroyed or abandoned. Various investigations with different purposes have produced data usable for several different time periods of interest. Water-level measurements during October 1979 (de Roche and Razem, 1984), were used to represent steady-state conditions before municipal pumping at the well field began. Water-level measurements during March 1986 (Sedam and others, 1989) were used to represent steady-state conditions after commencement of pumping at the well field. Water levels measured during March 1986 - June 1991 were used for calibration targets in the transient simulations. These water levels included USGS-measured water levels from previous studies and water-level records from a study conducted for the City of Columbus by Malcolm Pirnie, Inc. (1988). From December 1989 through June 1991, water levels were measured quarterly, when accessible, in 44 of these wells. Forty of the wells are completed in the glacial drift aquifer, whereas four of the wells are completed in the carbonate bedrock aquifer. Previous studies have shown that water levels in the carbonate bedrock and glacial drift aquifers are nearly the same in unstressed areas of the system (Bair and Norris, 1990). All these data were used to construct potentiometric-surface maps and hydrographs of the glacial drift and carbonate bedrock aquifers. Five wells were equipped with data loggers to record hourly water-level changes. Hourly data also were used to help calculate recharge rates to the aquifer and vertical hydraulic gradients within the glacial drift aquifer and between the glacial drift and carbonate bedrock aquifers.

Daily precipitation data were collected by City of Columbus personnel at the Parsons Avenue Water Treatment Facility throughout the study period. A U.S. Weather Bureau-type gage was read daily. In periods of missing data at the Parsons Avenue Water Treatment Plant, National Oceanic and Atmospheric Administration (1979-90) data from the Valley Crossing station, located 4 mi to the northeast, were substituted. These precipitation data were compared with data from previous studies (Sedam and others, 1989; Razem, 1983) to help determine recharge rates to the aquifer.

Pumpage records from all ground-water users in excess of 35,000 gal/d within the study area also were collected. Data were supplied by the City of Columbus (collector wells) and the Ohio Environmental Protection Agency (maximum discharges into the Scioto River and Big Walnut Creek). A computer-generated estimate of ground-water recharge was determined by use of the local-minimum method of streamflow hydrograph separation, as described by Pettyjohn and Henning (1979). These data were used to determine a hydrologic budget for the ground-water-flow model.

Previous assessments of hydraulic properties of the well-field area were reviewed in conjunction with previous numerical modeling studies to determine appropriate values of hydraulic parameters to be used in the ground-water-flow model. Parameters in areas where analytical work had not been performed were estimated from well logs, geologic sections, and trial-and-error adjustments during model calibration. Streambed permeabilities were estimated by use of seepage-meter and gain/loss studies on the Scioto River (de Roche, 1985; Childress and others, 1991) and Big Walnut Creek (Cunningham, 1992).

Flow Simulation

A transient ground-water-flow model was constructed by use of the USGS three-dimensional finite-difference ground-water-flow code (McDonald and Harbaugh, 1988), hereafter referred to as MODFLOW. MODFLOW is a block-centered finite-difference code written in a modular form that allows the user to incorporate various components of the conceptual flow system into different "packages" within the model. MODFLOW is a widely used and well-documented flow model.

The three-dimensional transient movement of ground water of constant density through porous earth material may be described by the partial-differential equation,

$$\frac{\partial}{\partial x}\left(-K_x\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(-K_y\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(-K_z\frac{\partial h}{\partial z}\right)\pm W = Ss\left(\frac{\partial h}{\partial t}\right),$$

where

K_x, K_y, and K_z are values of hydraulic conductivity along the x, y, and z coordinate axes, which are assumed to be parallel to the major axes of the hydraulic conductivity tensor;

h is the potentiometric head;

W is the volumetric flux per unit volume and represents sources and (or) sinks of water;

S_s is the specific storage of the porous material; and

t is time (McDonald and Harbaugh, 1988).

The partial-differential equation of ground-water flow is approximated numerically by use of finite-difference techniques. The continuous variables of the partial-differential equation are replaced with discrete variables that are defined at grid blocks (or nodes). Thus, the continuous differential equation that defines head everywhere in the aquifer is replaced by a finite number of linear algebraic equations that define head at specific points. The system of simultaneous algebraic equations then is solved to give an approximate solution to the time-varying head distribution that would be given by an analytical solution of the partial-differential equation of flow. The strongly implicit procedure was used to solve for head (McDonald and Harbaugh, 1988).

HYDROGEOLOGY

The hydrogeology of the study area consists of a heterogeneous glacial drift aquifer deposited on a carbonate bedrock aquifer.

Glacial Drift Aquifer

The primary source of ground water in the study area is the glacial drift aquifer. The glacial drift is composed of sand, gravel, and clay deposited during the Illinoian and Wisconsinan glaciations (Schmidt and Goldthwait, 1958). The heterogeneous and stratified nature of these deposits is shown in the geologic sections (figs. 3-5) through the study area. In general, thick deposits of till containing lenses of sand and gravel dominate the drift in the area west of the Scioto River. The thickest and most productive parts of the glacial drift aquifer are in the buried preglacial valleys in the central and eastern parts of the study area underlying the Scioto River and Big Walnut Creek.

There are various interpretations of the glacial depositional history of the area. Most authors present the interpretation discussed by Schmidt and Goldthwait (1958), which states that deposition occurred in two substages, the early Wisconsinan substage (about 50,000 years before present) and the late Wisconsinan substage (about 22,000 years before present). The early Wisconsinan substage is characterized by a widespread basal till overlain by outwash and valley-train deposits. The outwash is composed of very coarse sand and gravel ranging in thickness from 5 to 100 ft (Schmidt and Goldthwait, 1958). A thin, weathered and leached zone above the early Wisconsinan deposits represents the interval between Wisconsinan substages. Clayey till overlies this interglacial zone and represents the most recent, late Wisconsinan glacial substage. A kame and esker complex is present in the Spangler Hill area in the central part of the study area, although much of the surface expression of this complex has been removed by quarrying operations.

Prior to glaciation, Teays-stage river drainage carved much of the present bedrock topography in the study area. The main trunk of the ancient Teays drainage system entered southern Ohio from West Virginia, flowing north into Pickaway County and then northwest across Ohio and into Indiana. The ancient Cambridge River, a major Teays tributary flowing from the northeastern part of the State, probably joined the Teays River in central Pickaway County.

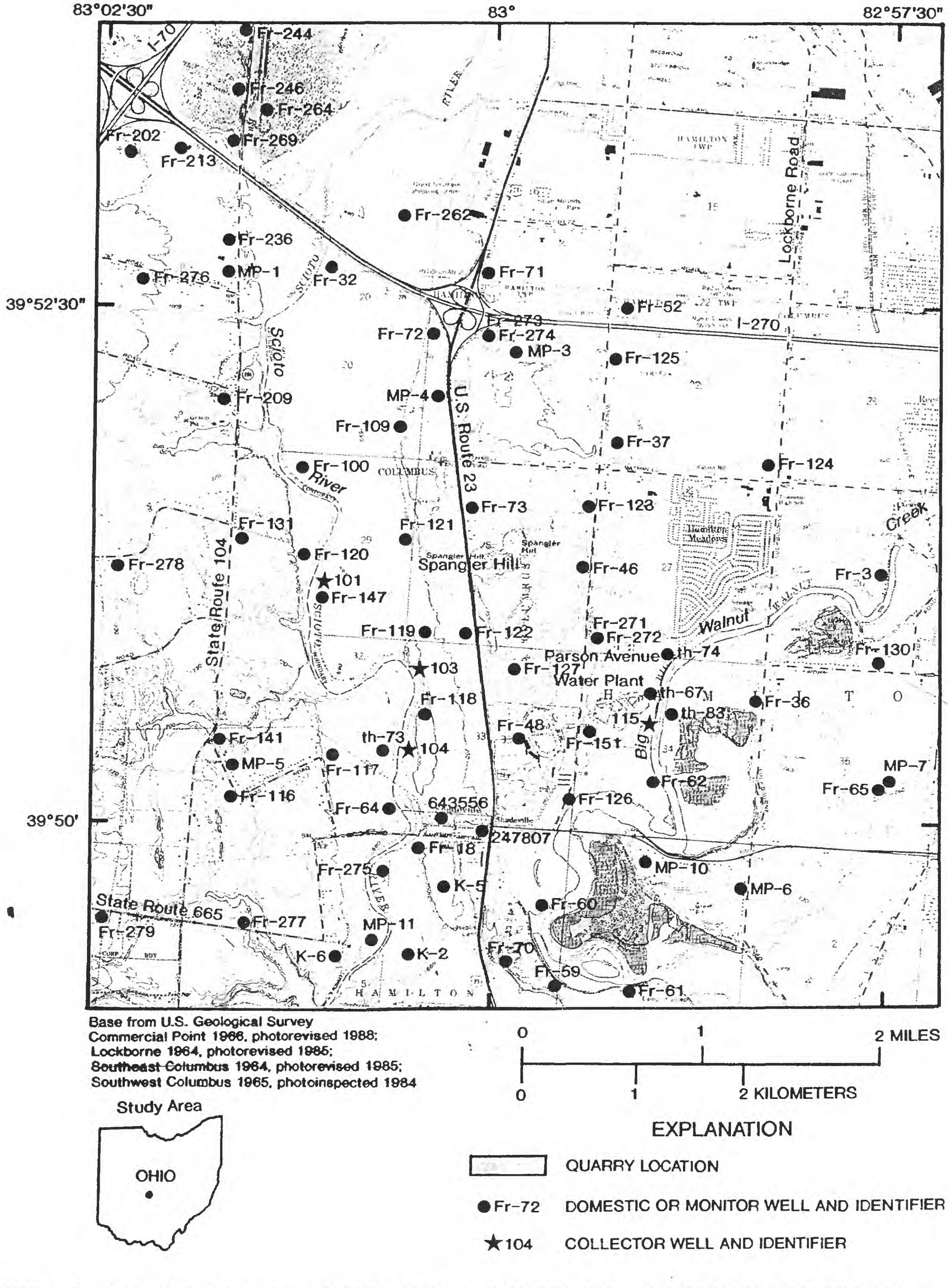

Figure 2. Location of ground-water-level-measurement sites. Water-level measurement tabled in Appendix 1.

seepage-meter and gain/loss studies on the Scioto River (de Roche, 1985; Childress and others, 1991) and Big Walnut Creek (Cunningham, 1992).

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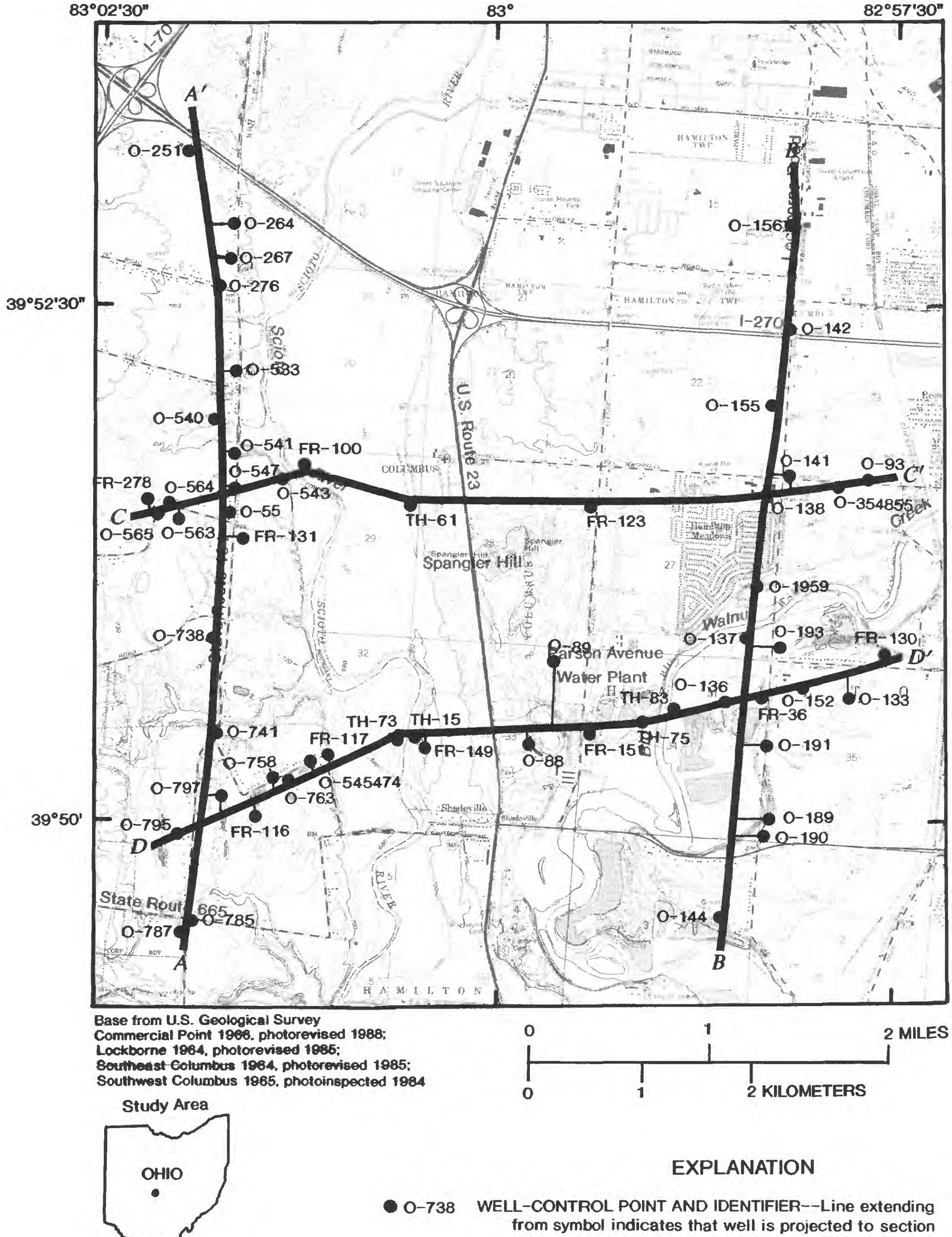

Figure 3. Location of geologic sections *A–A'*, *B–B'*, *C–C'*, and *D–D'*. (Sections shown in figure 4.)

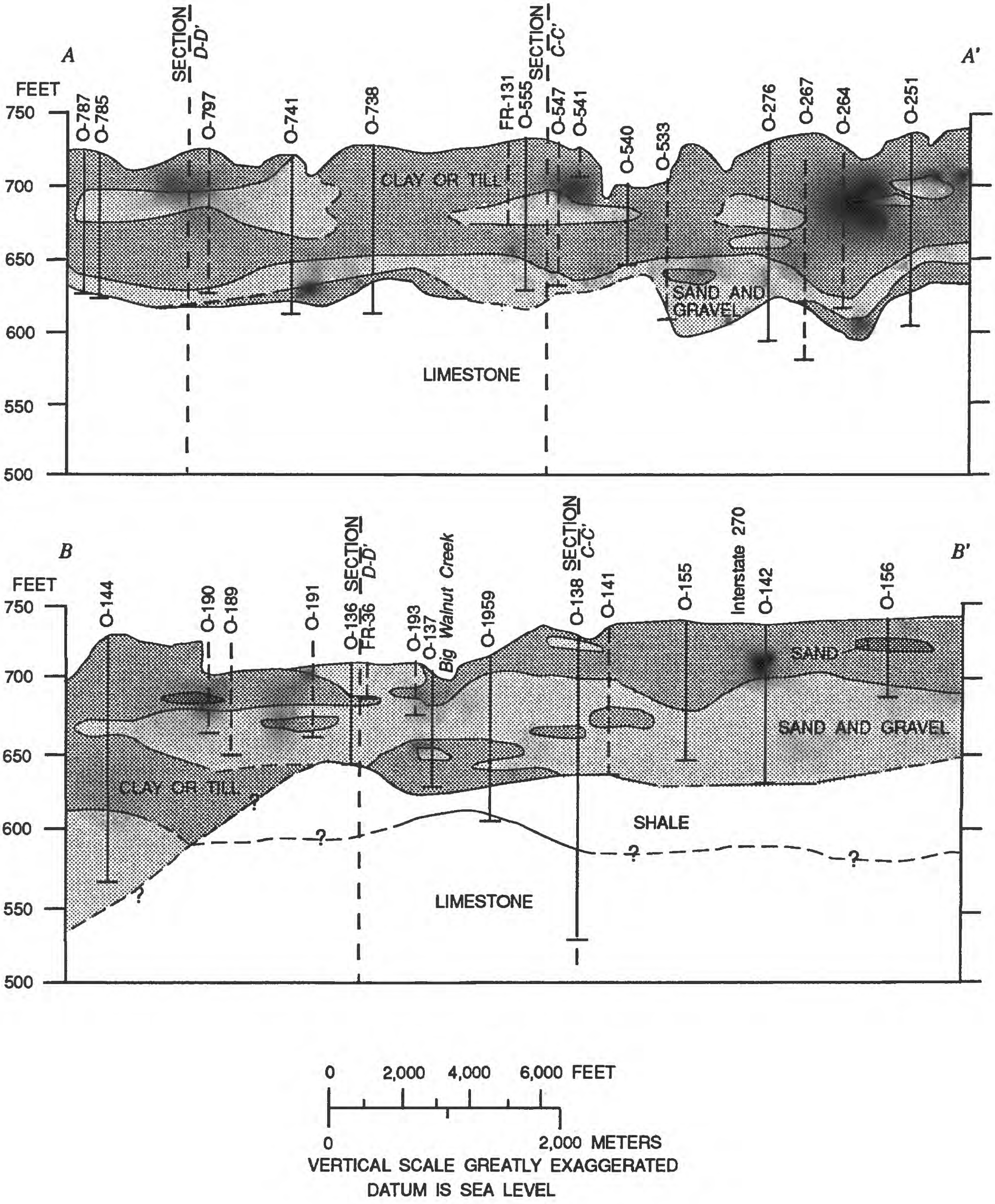

Figure 4. Geologic sections *A-A'* and *B-B'*. Reference well indicated by local identifiers. "Fr" prefix indicates U.S. Geological Survey identifiers; "O" indicates Ohio Department of Natural Resources identifier. (Section traces shown in figure 3.)

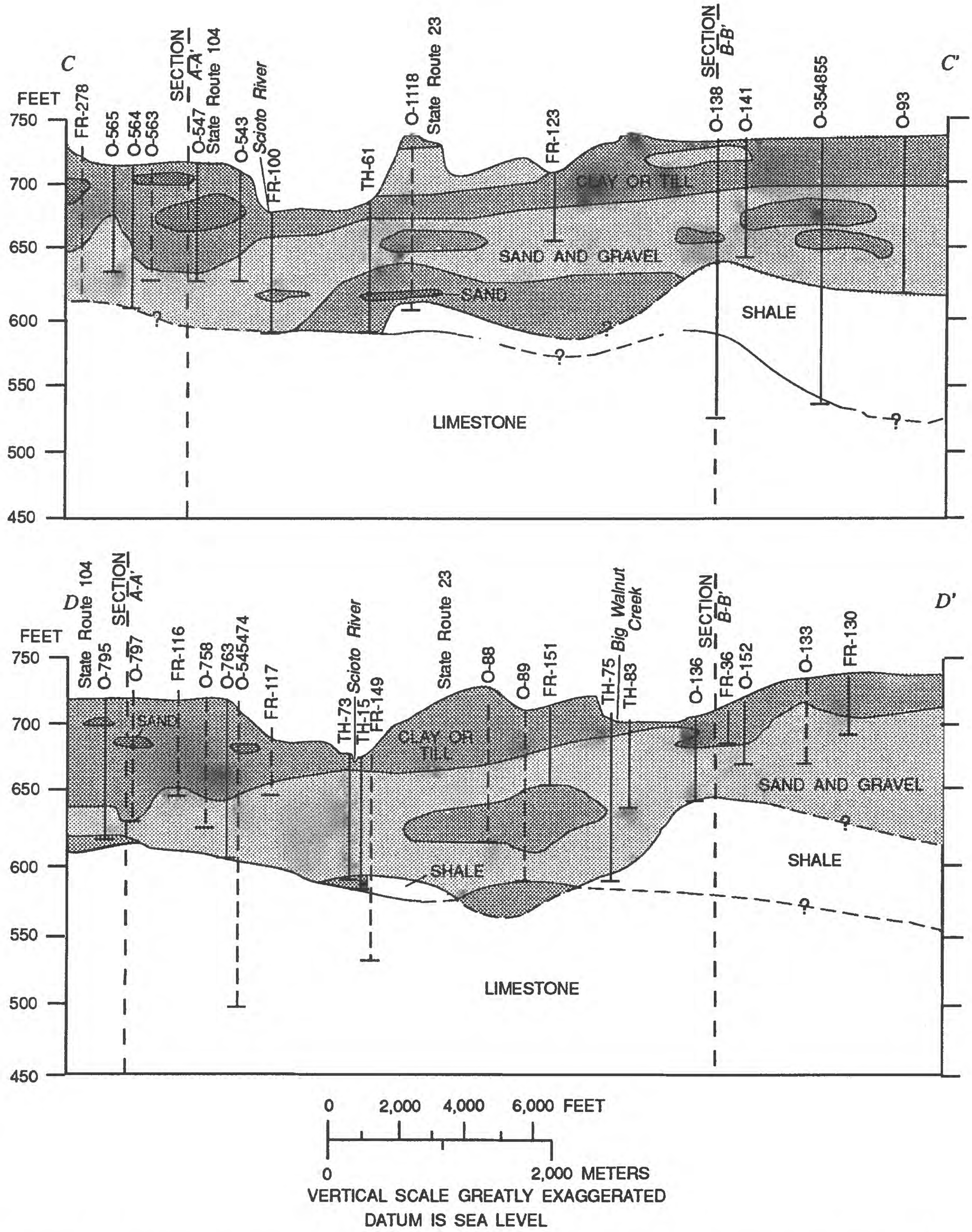

Figure 5. Geologic sections *C-C'* and *D-D'*. Reference well indicated by local identifiers. "Fr" and "TH" prefix indicates U.S. Geological Survey identifiers; "O" indicates Ohio Department of Natural Resources identifier. (Section traces shown in figure 3.)

Tributaries to the Cambridge River drained southeastern Franklin County and helped form the present bedrock topography in the study area (Dove, 1960).

Pleistocene glaciation blocked the northwestern course of the Teays system, and a new, southerly-flowing drainage system developed to carry water away from the melting glaciers. This new drainage system, known as the Deep-Stage drainage, generally followed the drainage path of the Cambridge River in Franklin County and further eroded the bedrock surface. When the glacial ice sheets melted, bedrock valleys became filled with unconsolidated glacial drift and outwash materials—in most areas, forming a single heterogeneous aquifer. The geologic sections (figs. 4-5) illustrate that the till deposits of the first substage locally have been removed by erosion, and outwash sand and gravel has been directly deposited over bedrock. The outwash deposits vary from 20 to 120 ft in thickness and are thickest near the center of the well field.

Recently, a more specific interpretation of the depositional history of the glacial drift emphasizes that the glacial drift aquifer is "not one great heterogeneous aquifer, but a series of smaller, more homogeneous aquifers that interact, yet also function independently of each other" (Williams and others, 1988). Unfortunately, even if these deposits do exist as hypothesized, given the available borehole information, the areal extent and geometry of these smaller, homogeneous aquifers cannot accurately be delineated. Therefore, the aquifer has been modeled by use of the traditional hydrogeologic interpretation of one major unconsolidated aquifer with spatial variations in hydraulic properties as determined from aquifer tests and well logs.

Carbonate Bedrock Aquifer

The secondary source of ground water within the study area is the carbonate bedrock aquifer, which consists of Silurian and Devonian limestones, dolomites, and shales (fig. 6). Geologic interpretations are based on logs from wells in other parts of the County, as no deep wells are completed within the study area. The lowermost unit represented in the stratigraphic section is the Middle Silurian Lockport Dolomite. The Lockport Dolomite consists of relatively pure, light-gray to white, fine to coarse crystalline dolomite and is approximately 65 ft thick (Norris and Fidler, 1973). The Upper Silurian Bass Island Group overlies the Lockport Dolomite. The Bass Island Group consists of the Greenfield Dolomite, the Tymochtee Formation, and undifferentiated dolomite. The Greenfield Dolomite is a light gray-brown, thinly bedded to massive dolomite that is about 50 to 60 ft thick in the study area (Norris and Fidler, 1973). The Tymochtee Formation is a medium to light gray, thinly bedded, argillaceous dolomite up to 100 ft thick. It is overlain by over 200 ft of undifferentiated brown to drab, fine-grained, argillaceous dolomite with some limestone layers (Schmidt and Goldthwait, 1958).

Devonian rocks in the study area consist of the Middle Devonian Columbus and Delaware Limestones and the Upper Devonian Olentangy and Ohio Shales (fig. 6). The Columbus Limestone is thickly bedded, varying from a limey dolomite in the lower portion to a low-magnesium limestone in the upper portion (Stout and others, 1943). The Columbus Limestone is approximately 70 ft thick in the study area. The Delaware Limestone and overlying shale units are not present across the entire study area due to pre-Pleistocene differential erosion. The Delaware Limestone is a dark brownish-gray to blue-gray limestone interbedded with calcareous brown shale and chert. Where present, the formation is up to 30 ft thick. The Olentangy Shale, also up to 30 ft thick, is a blue-black soft shale containing limestone concretions. The Ohio Shale consists of black carbonaceous shale and gray siliceous shale. The Ohio Shale is the youngest bedrock unit in the study area (fig. 6), occurring only in the eastern part and thickening eastward up to 100 ft on the eastern edge of the study area (Schmidt and Goldthwait, 1958).

The ground-water-flow system in the carbonate bedrock aquifer in central Ohio has been divided into two subsystems on the basis of differences in water quality between the Tymochtee Formation and the underlying Middle Silurian rocks (Norris and Fidler, 1973; Sedam and others, 1989; Childress and others, 1991). The subsystem that affects the study area is the upper part of the carbonate-bedrock flow system in the Tymochtee Formation and the Columbus and Delaware Limestones (fig. 7). Waters sampled below this subsystem have much higher dissolved solids concentration than waters sampled in the overlying Tymochtee Formation. This suggests that flow in the upper part of the carbonate bedrock aquifer is part of a local flow cell that discharges into the Scioto River, and flow in the deeper part of the flow system is part of a deeper, more regional flow cell. In general, the Ohio Shale thickens eastward from the Scioto River and

SYSTEM	SERIES	GROUP	FORMATION	LITHOLOGY	THICKNESS (feet)		DESCRIPTION
QUATERNARY	HOLOCENE		Alluvium		0-20		Silt, sand, and clay deposited on flood plains
QUATERNARY	PLEISTOCENE		Glacial drift		40-100		Silt, sand, gravel, and till, unconsolidated some lenses, some thick deposits
DEVONIAN	UPPER		Ohio Shale		0-100		Shale, black to dark-brown, carbonaceous sometimes arenaceous, grading from massive to thinly laminated beds
DEVONIAN	UPPER		Olentangy Shale		0-30		Shale, blue-black, soft, argillaceous with some argillaceous limestone concretions
DEVONIAN	MIDDLE		Delaware Limestone		30		Limestone, blue-gray, thin-bedded, with thin shaly layers, ovrite, and black chert
DEVONIAN	MIDDLE		Columbus Limestone		70		Limestone, brown, massive, pure, grading to porous, massive, and impure
SILURIAN	UPPER	SALINA GROUP	Un-differentiated dolomite		215-235	375	Dolomite, brown to drab, fine-grained compact, thin to massive, impure, argillaceous, with some limestone layers
SILURIAN	UPPER	SALINA GROUP	Tymochtee Dolomite		90-100		Dolomite, medium to light gray grading to light brown, thin-bedded, argillaceous, some thin shale beds
SILURIAN	UPPER	SALINA GROUP	Greenfield Dolomite		50-60		Dolomite, light gray-brown, some carbonaceous partings, beds vary in thickness from less than 6 inches to massive, thin beds are fine-grained and even textured, massive part are coarse grained and vesicular
SILURIAN	UPPER	SALINA GROUP	"Newburg zone"		15-20		Highly permeable zone of impure dolomite, stratigraphic position varies spatially above or within the upper portion of the Lockport Dolomite
SILURIAN	MIDDLE		Lockport Dolomite		65		Dolomite, pure, light gray to white, finely to coarsely crystalline

Figure 6. Summary of geologic characteristics of unconsolidated and consolidated rocks, south-central Franklin County, Ohio. (Modified from Schmidt and Goldthwait, 1958; Norris and Fidler, 1973; and U.S. Geological Survey well logs on file).

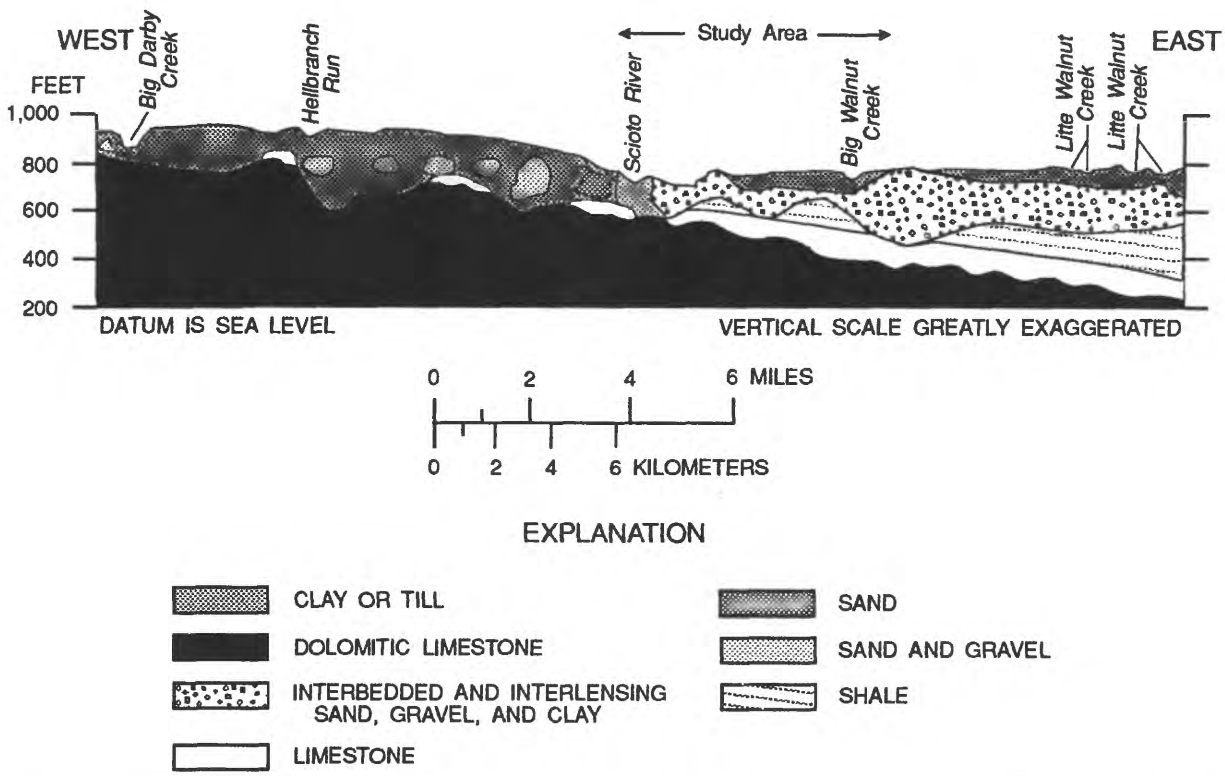

Figure 7. Generalized geologic section through southern Franklin County, Ohio. (Modified from Schmidt and Goldthwait, 1958, fig. 9).

restricts flow into and out of the underlying carbonate rocks. In the study area west of the Scioto River, the shale is absent.

Ground-Water Levels

Ground-water levels measured in previous studies (de Roche and Razem, 1984; Sedam and others, 1989) were used in concert with water levels measured for this study to calibrate the steady-state and transient models. Combining these data resulted in several synoptic sets of water levels from times of differing stress on the flow system. The locations of 74 different wells used to make water-level measurements since 1979 are shown in figure 2. Some of these wells have been destroyed or abandoned. Measurements from October 1979 (de Roche and Razem, 1984) were used as water levels representative of steady-state conditions before municipal pumping was initiated at the well field. A potentiometric surface of the glacial drift aquifer drawn from the October 1979 data is presented in figure 8. The potentiometric surface illustrates that flow is nearly parallel to the northern and southern boundaries of the study area, and, in most places, nearly perpendicular to the eastern and western boundaries of the study area. The Scioto River is losing flow in the northern part of the study area and gaining flow in the southern part of the study area. Big Walnut Creek is gaining in the upstream part of the study area and losing in the downstream part of the study area.

Measurements from March 1986 (Sedam and others, 1989) were used as water levels representative of steady-state conditions after commencement of pumping at the well field. The potentiometric surface of the glacial drift aquifer drawn from March 1986 water levels is presented in figure 8. The configuration of the potentiometric surface has not changed significantly at the boundaries of the study area; however, pumping in the collector wells along the Scioto River has lowered the water table enough to create additional losing reaches. The configuration of the potentiometric surface around Big Walnut Creek is largely unchanged.

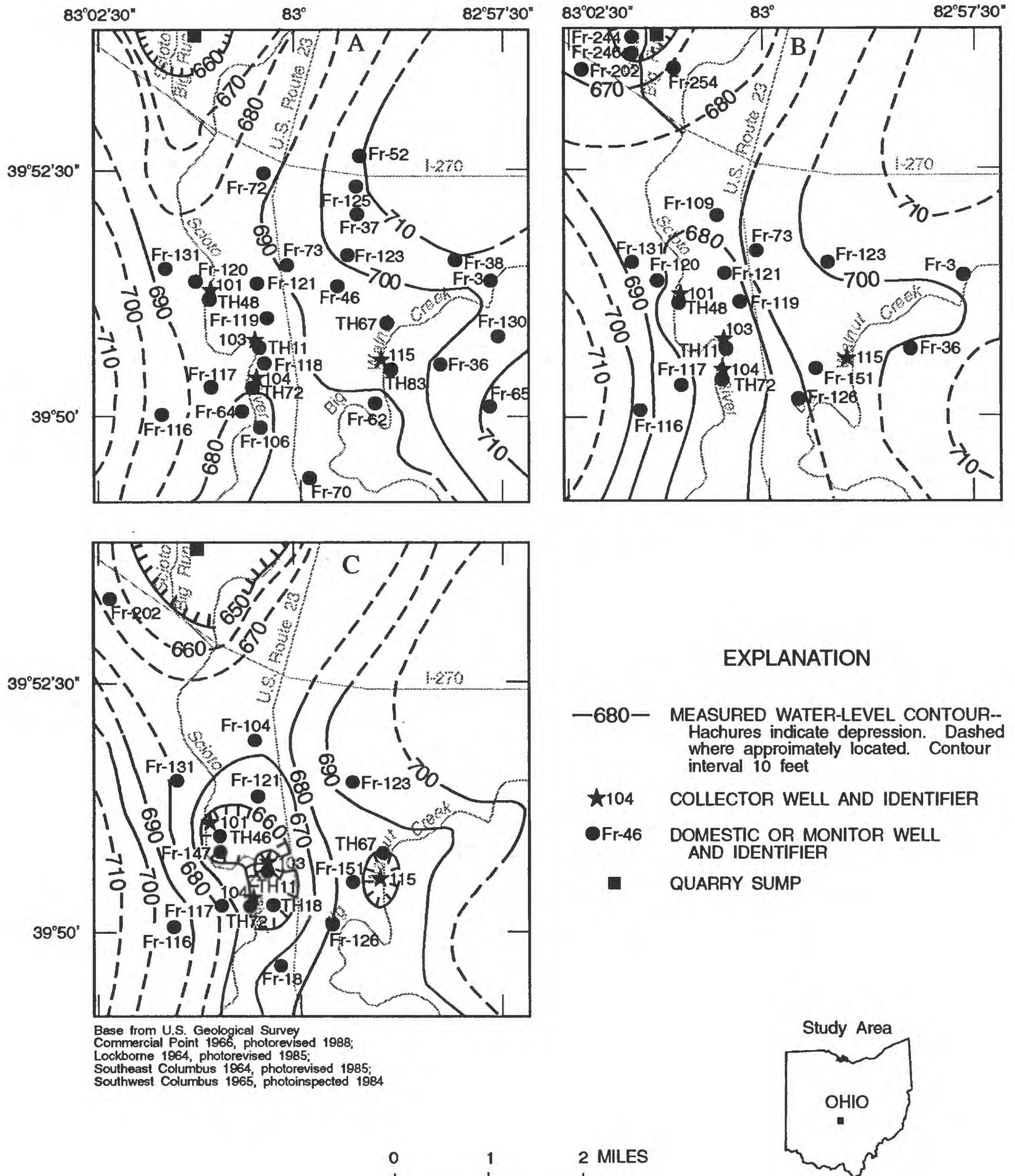

Figure 8. Measured potentiometric surface of glacial drift aquifer for (A) October 1979, (B) March 1986, and (C) August 1988. (From DeRoche and Razem, 1984, fig1).

Water levels measured from March 1986 through June 1991 were used for calibration targets in the transient simulations. The potentiometric surface of the glacial drift aquifer drawn from August 1988 water-level data is presented as an example of the transient nature of the flow system (fig. 8). Precipitation in 1987 and early 1988 was below normal. Well-field production also increased in response to increased demand. The combination of these two stresses resulted in the lowest water levels ever measured in many of the wells in the study area. The potentiometric surface illustrates overall declines in water levels, as well as an increase in the cone of depression of the Scioto River collector wells.

A table of selected records from wells in the study area is listed in the Appendix.

Ground-Water/Surface-Water Relations

Investigation of the ground-water/surface-water interactions in the study area focused on definition of a conceptual flow system based on potentiometric contours and stream gain/loss measurements, and measurement of streambed permeability by use of seepage meters.

Before any stresses such as municipal pumping and quarry dewatering were applied to the ground-water system in the study area, the Scioto River was a gaining stream, acting as a discharge area for the glacial drift aquifer and for the carbonate bedrock aquifer, which discharged water upward into the overlying glacial drift (Norris, 1959). Big Walnut Creek was a naturally losing stream throughout most of the study area. The elevation of the stream bottom is several feet higher than the nearby Scioto River, which acts as an underdrain and receives flow from Big Walnut Creek (Stowe, 1979; Sedam and others, 1989). With the advent of municipal pumping and quarry dewatering, the Scioto River has developed alternately gaining/losing reaches, depending on the proximity of the river to these stresses.

One of the most difficult parameters to estimate in a ground-water-flow model is streambed permeability. Previous studies indicated that streambed permeability is a sensitive parameter to consider in this flow system (Sedam and others, 1989; Childress and others, 1991). Previous modeling studies determined values of streambed permeability based on gain/loss studies and trial-and-error adjustment during model calibration until a good match between simulated and measured hydraulic heads and hydraulic fluxes (hereafter referred to as head and flux) was achieved.

Gain/loss studies on the Scioto River and Big Walnut Creek were uninformative. Measured gains or losses on both streams at low flow are within the measurement error of a discharge measurement rated good for most reaches of the streams. As a result, gaining and losing sections of the Scioto River and Big Walnut Creek were determined with additional seepage-meter and piezometer measurements. By 1991, the Scioto River was a losing stream in most of the study area. The river gains flow for about a mile in the area between the cone of depression created by the bedrock quarry and the cone of depression created by the well field. The river also gains flow south of collector well 104.

The gradient of Big Walnut Creek is less than that of the Scioto River, resulting in lower flow velocities. Most of the faster-moving part of the stream was mapped in the first 1.5 mi within the study area. The stream is gaining in this reach; however, the stream gradient decreases by eightfold near this point (at about Lockborne Road). The reduction in flow rate allows fine particles to settle out, reducing the streambed permeability (Stilson and Associates, 1977). The stream also begins to lose flow to the aquifer. The losing nature of the creek is natural because of the elevation difference between Big Walnut Creek and the Scioto River (as much as 15 ft), but it is enhanced by pumping at collector well 115 and the quarry operations on the eastern side of the creek (see fig. 2).

To more accurately estimate the streambed characteristics of the Scioto River in the study area, the river was mapped on the basis of three riverine settings: pools, riffles, and runs. Vertical hydraulic conductivity of the streambed materials (K_{vsb}) was measured directly at each of these settings by use of a streambed seepage meter. A uniform vertical hydraulic conductivity then was used throughout each of the mapped river settings. The results of these streambed-permeability studies are reported in Childress and others (1991) and Cunningham (1992).

Previous modeling efforts (Weiss and Razem, 1980; Razem, 1983; Sedam and others, 1989) used a uniform K_{vsb} value along the Scioto River. By use of the additional data from seepage measurements and river mapping, the transient ground-water-flow model incorporates field-measured variations in K_{vsb} along the course of the river.

Aquifer Recharge

The ground-water-flow system in the South Well Field area is recharged by precipitation, regional ground-water flow, and induced stream infiltration. Recharge from precipitation is controlled by the infiltration rate of soils, relief, temperature, and evapotranspiration. Recharge rates discussed below refer to the annual volume of water which reaches the water table.

Various recharge rates in the South Well Field area have been estimated. Stowe (1979, p. 70-78) estimated the recharge of four areas of differing surficial geology in southern Franklin County on the basis of separation of stream hydrographs. Stowe's estimates range from 4.2 in/yr in till-covered areas to 9.4 in/yr in areas covered with alluvium, outwash, and kames.

In the numerical model constructed by Weiss and Razem (1980, p. 9-11), annual recharge was estimated to be 12 in. This recharge rate was used throughout their study area and was based on hydrograph recessions from well FR-109. Although the method used to determine the aquifer recharge rate is valid, their analysis was probably affected by flooding along the Scioto River, which caused changes in bank storage and infiltration to the aquifer (Sedam and others, 1989). Analysis of subsequent recession data results in recharge rates as large as 30 in/yr based on water levels measured in wells in or near the flood plain. It also is likely that their specific yield estimate of 0.1 was incorrect. The result was an overestimate of the recharge rate for the entire study area but a rate which may be realistic for the flood plain during flood events.

Sedam and others (1989, p. 62) used trial-and-error adjustments of their steady-state numerical model to determine aquifer recharge rates. A recharge rate of 4.0 in/yr was used for the area west of the Scioto River. A recharge rate of 7.0 in/yr was assigned to the outwash and till deposits east of the Scioto River. The valley-train deposits, alluvium, kames, and eskers in the study area were assigned a recharge rate of 9.0 in/yr.

As one estimate of recharge and as a check of Stowe's (1979) estimate, recharge by hydrograph separation of more than 50 years of USGS stream discharge data was determined for Big Walnut Creek at Reese (1,000 ft downstream from the eastern boundary of the study area) and Scioto River at Columbus (about two river miles north of study area) by the local-minimum method described in Pettyjohn and Henning (1979) and adapted for the USGS Automatic Data Processing System (Dempster, 1990) data base by White and Sloto (1991). An example of the local-minimum method is shown in figure 9. Ground-water discharge to Big Walnut Creek over a drainage area of 544 mi^2 was about 4.0 in/yr, with a range of 2.1 in/yr at 9.8-percent base flow to 5.7 in/yr at 90-percent base flow (fig. 10). Ground-water discharge to the Scioto River over a drainage area of 1,629 mi^2 also was 4 in/yr, with a range of 2.2 in/yr at 10-percent base flow to 5.6 in/yr at 90-percent base flow (fig. 10).

These are conservative estimates of recharge in the study area because ground-water contribution to base flow is estimated over the entire drainage area, and the hydraulic properties of the glacial drift aquifer within the study area are generally more transmissive than those throughout the respective basins. Care should also be taken in using these recharge rates because, over most of the period of record, both streams were controlled by dams upstream from the respective gages. Hydrograph separation of USGS stream-discharge data from the Scioto River at Circleville also was performed because the geology of the study area more closely approximates the geology within the drainage area at Circleville than the drainage above the station at Columbus (fig. 10). On the basis of 6 years of discharge record (1974-79), ground-water discharge was 6.3 in/yr, with a range 3.5 in/yr at 14-percent base flow and 7.5 in/yr at 86-percent base flow.

Recharge rates determined for this study are based on a combination of the above methods and rates. The array of recharge rates used by Sedam and others (1989) was used as a starting point, with the higher recharge rate determined by Weiss and Razem (1980) used in the portion of the study area that is flooded frequently. These rates then were adjusted slightly during model calibration on the basis of measured precipitation and simulated water levels. Final yearly recharge rates ranged from 4.0 to 12.0 in. Areal and transient distribution of recharge are discussed in the "Model Parameters" section of this report. Actual rates input to the model varied with measured precipitation and hydrograph response during the simulation period.

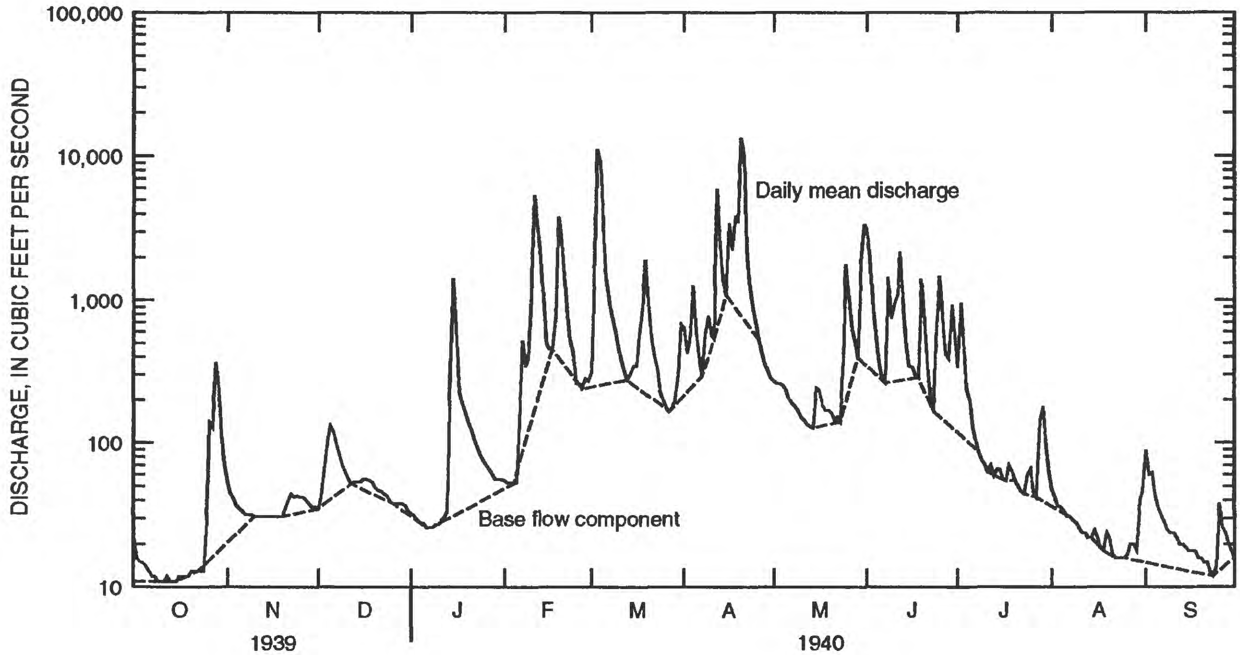

Figure 9. Hydrograph separation for Big Walnut Creek at Reese, Ohio, water year 1940.

SIMULATION OF GROUND-WATER FLOW

A transient ground-water-flow model was constructed by use of MODFLOW (McDonald and Harbaugh, 1988). The model was constructed to improve the understanding of the hydrogeology, ground-water/surface-water interactions, ground-water-flow directions, and hydrologic budget of the study area. The model will be used in future work to determine steady-state and transient contributing areas to the collector wells by use of particle tracking.

In a finite-difference model, space is discretized into a number of cells or blocks. A point, or node, is assigned to the center of each cell. The array of cells and nodes is called a grid. In the finite-difference method, the ground-water-flow equation is approximated for each cell in the system. The system of equations is then solved iteratively until there is sufficient agreement between two successive iterations. A user-defined convergence criterion determines the agreement between successive iterations. In a transient simulation, time is discretized into stress periods and time steps. Stress periods are periods of time when system stresses are constant. Time steps are increments of time within a stress period. Time must be discretized into smaller increments during a stress period because the finite-difference approximation assumes a constant gradient over the time step.

Description of Conceptual and Numerical Models

Construction of a ground-water-flow model begins with formulation of a conceptual flow model. A conceptual flow model is a simplified representation of the actual flow system. The conceptual flow model is transformed into a numericall model by use of equations that are assumed to incorporate the characteristics of the ground-water-flow system and the physics of ground-water flow.

The conceptual ground-water-flow system is illustrated in figure 11. Recharge, boundary flux, and river leakage are the principal sources of water to the flow system. The study area is bounded on the north and south by streamlines, with flow entering the area from the east and west. Areal recharge is contributed throughout the study area, although a higher percentage of precipitation reaches the water table in the area east of the Scioto River where little surface drainage exists. Vertical ground-water flow is downward in the uplands adjacent to the Scioto River, and upward near the river in both the glacial drift and the carbonate

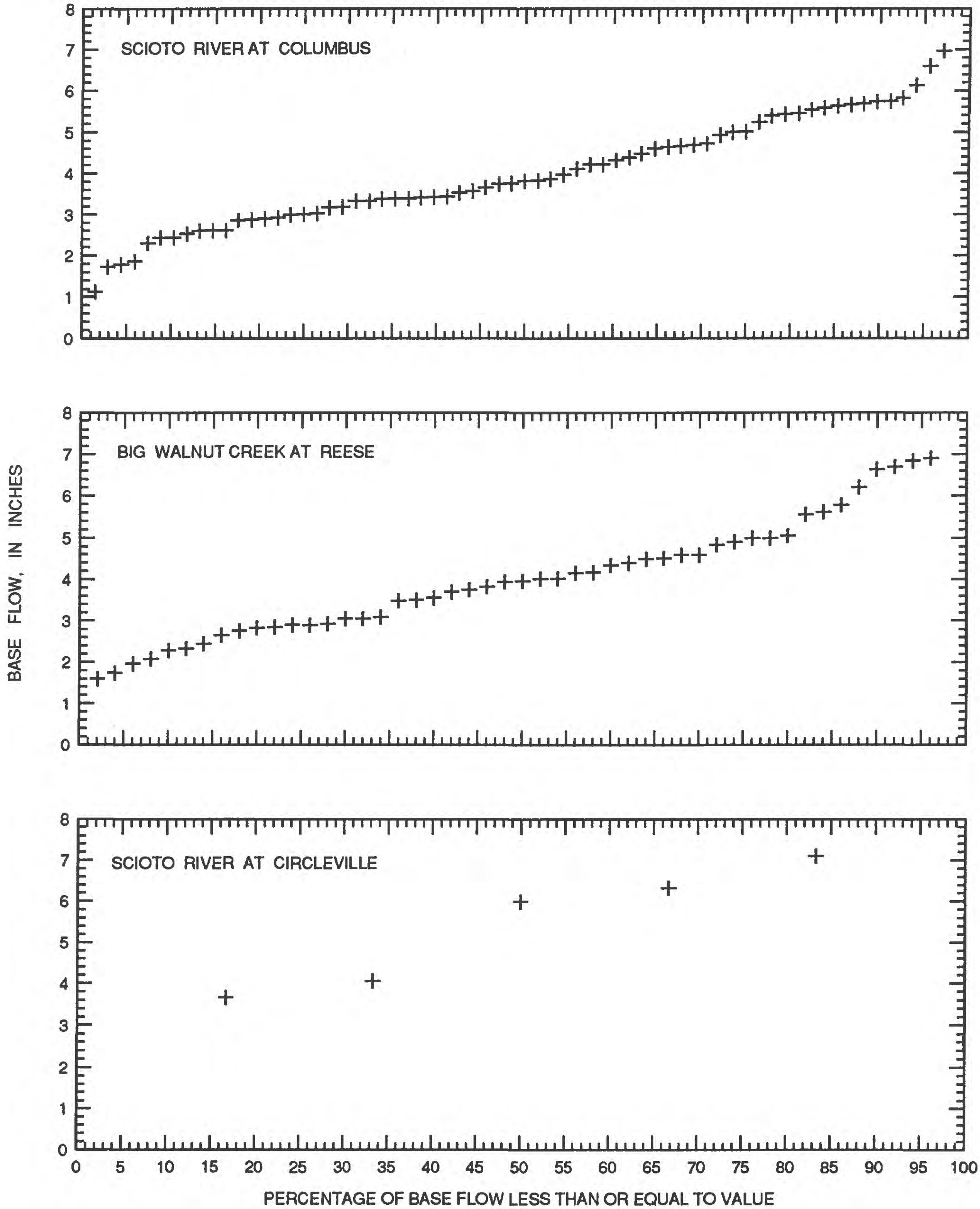

Figure 10. Frequency diagram of base flow for Scioto River at Columbus, Ohio, water year 1922-89, Big Walnut Creek at Reese, Ohio, water year 1940-88, and Scioto River at Circleville, Ohio, water year 1974-79.

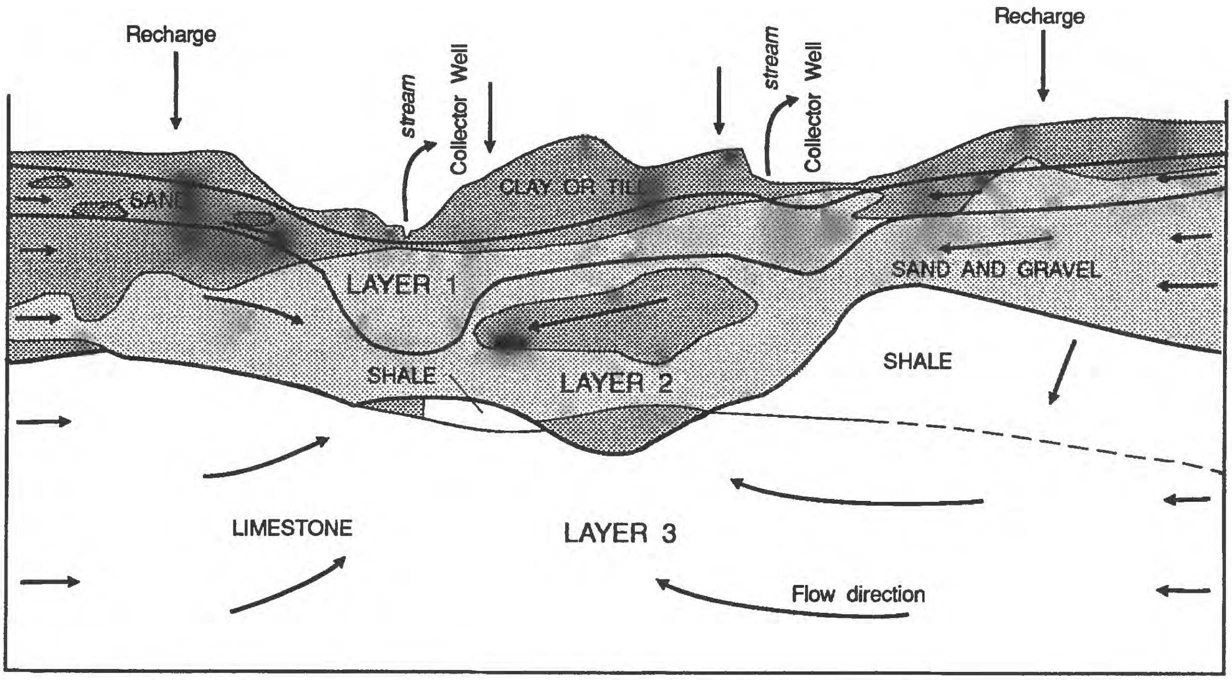

Figure 11. Conceptual ground-water-flow system.

bedrock aquifers.

The flow system is vertically discretized into three model layers. Layer 1 extends from the water table to a depth of approximately 15 ft, but is absent (mined out) at the bedrock quarry on the northern boundary of the study area. This layer separates the Scioto River and Big Walnut Creek from model layer 2 and allows the rivers and tributaries to partially penetrate a thin model layer, rather than a thick model layer. Vertically discretizing the glacial drift aquifer into two layers also increases the vertical resolution of the model and isolates the collector wells in a different model layer than the rivers. At the collector wells, the thickness of layer 2 is reduced, allowing the well to fully penetrate model layer 2, as opposed to partially penetrating a thicker layer. This vertical discretization is shown in figure 12.

Layer 2 includes the remainder of the saturated glacial drift. Layer 2 ranges in thickness from 0 in the bedrock quarry at the northern boundary of the study area to 95 ft on the northeastern boundary of the study area. All observation wells in the glacial drift are completed in layer 2. The bottom of layer 2 coincides with the top of the Olentangy and Ohio Shales in the eastern part of the study area and the subcrop of the Delaware or Columbus Limestone in the western part of the study area (see fig. 7). Layer 3 includes all of the bedrock in the study area to an elevation of 250 ft above sea level, which approximates the elevation of the base of the Tymochtee Formation.

The rectangular finite-difference grid was superimposed on a topographic map of the study area (fig. 13) to simulate the hydrologic characteristics of the conceptual flow model. The numerical model contains 53 rows, 45 columns, and 3 layers. The grid is oriented nearly north-south so that most of the Scioto River flows parallel to grid lines. The model grid is variably spaced to account for differences in available data and to simulate heads and hydraulic gradients accurately in specific areas of interest. The length and width of grid cells range from 200 to 2,000 ft with finer spacings designed to increase detail in the areas near the collector wells. Finer grid spacing also

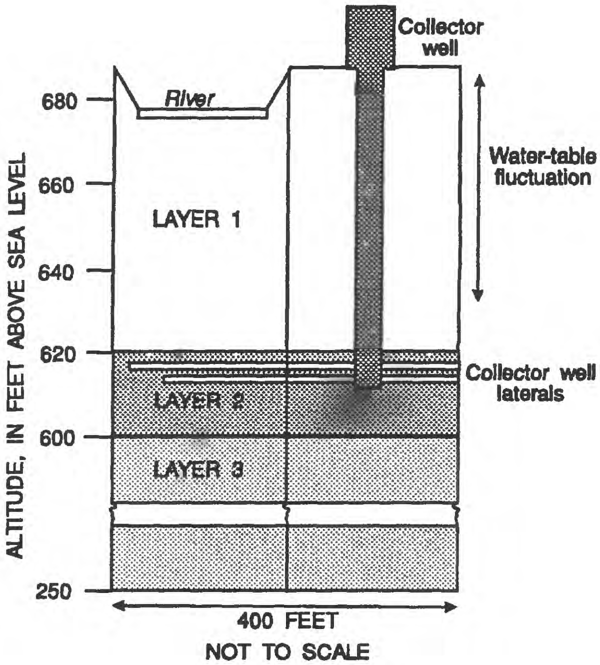

Figure 12. Vertical discretization of model in vicinity of collector well 101. Well completed in layer 2.

allows for segregation of cells containing rivers and wells. There are 7,155 active model nodes.

Rivers and their tributaries in the study area are simulated as head-dependent fluxes. The headwaters of small tributaries to the west of the Scioto River are simulated as drains in the till, and as river cells when the tributaries reach the buried valley. This simulates the conceptual flow system—streams drain the till to the west but can gain or lose flow as the stream crosses into the more permeable buried-valley deposits near the Scioto River. The Scioto River and Big Walnut Creek are simulated by use of the river package.

Streambed conductances differ along the course of the Scioto River and Big Walnut Creek as indicated by river mapping and seepage measurements discussed previously. Streambed-conductance values used in the model are defined as $K_{vsb}LW/T$ (McDonald and Harbaugh, 1988, p. 6-4), where

K_{vsb} is vertical hydraulic conductivity of the riverbed,

L is length of river reach within the grid cell,

W is width of river reach within the grid cell, and

T is thickness of the riverbed.

The length of river reach within a grid cell was measured from USGS 7-1/2-minute topographic maps. River width was estimated from the same topographic maps, although some adjustments were made on the basis of river cross-section measurements made at seepage-meter sites. Riverbed thickness was assumed to be 1 ft.

The bedrock quarry on the northern boundary of the study area has a pump that dewaters the area to an elevation of 562 ft above sea level. This is simulated by assigning specified heads of 562 ft to the cells in layer 3 that correspond to the sump pit in the quarry. The gravel quarries in the central and southeastern part of the study area are simulated by head-dependent fluxes. These are wet-mining quarries that pump small amounts of water solely for washing operations. Sedam and others (1989) showed that water levels in the southeastern quarries fluctuate with the water table.

Other pumpage within the study area is simulated as specified fluxes. Well-field pumpage was withdrawn from the model at various rates as recorded by Parsons Avenue Water Treatment personnel. Withdrawals in the steady-state simulations included pumpage by the Hamilton Meadows well field. This well field was not operating during the dates of the transient simulations. Smaller-scale pumpage within the study area includes Scioto Downs Raceway and the wet-mining quarries in the southeastern part of the study area.

Model Assumptions

The assumptions of the ground-water-flow model include (1) glacial drift deposits make up a single, heterogeneous but continuous unconfined aquifer, (2) no ground water flows into or out of the base of the Tymochtee Formation, (3) hydraulic conductivity within a model layer is uniform with depth but can differ laterally, (4) all streambeds have a thickness of 1 ft, (5) all wells fully penetrate the layer in which they are completed, (6) flow is parallel to the northern and southern external model boundaries, and (7) specified-flux boundaries on the east and west represent flux into the model and are not significantly affected by internal stresses. In MODFLOW, computed drawdown head in the grid cell, so actual drawdown

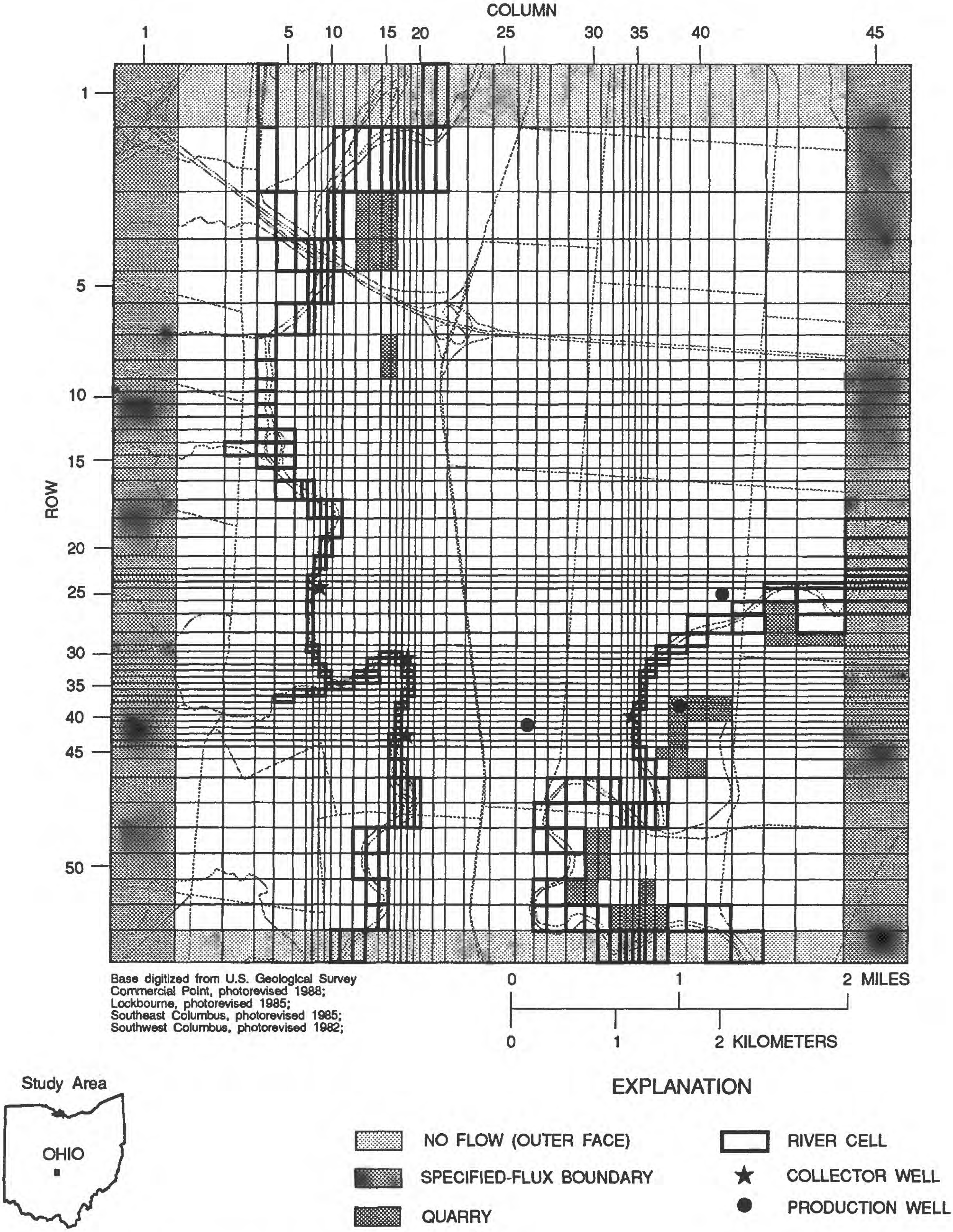

Figure 13. Finite-difference grid and model boundaries.

represents the average head in the grid cell, so actual drawdown at a pumped well within a grid cell can be much greater than that simulated by the model.

Model Boundaries

Quantitative modeling of a ground-water system entails solving the governing partial-differential equation in the flow domain while at the same time satisfying the specified boundary and initial conditions (Franke and others, 1987). The boundaries of this flow domain include external boundaries such as streamlines (no-flow) and specified flux, and internal boundaries such as specified flux (pumping, recharge) and head-dependent flux (stream and quarry seepage).

Boundary conditions are illustrated in figure 13. No-flow boundary conditions were set on the northern and southern borders of the flow model. The direction of ground-water flow at these boundaries is nearly east and west, parallel to the model boundary (see fig. 11). Thus, flow is nearly parallel to these boundaries, and flow rates across the boundaries are very small. One exception to the assumption of parallel flow lines at the southern boundary exists during the 1986 steady-state simulation. From the mid-1980's to 1988, a construction dewatering operation south of the study area had a slight impact on the flow field at the model boundary. Field data indicated that the cone of depression from the dewatering reached as far north as the town of Shadeville. To account for this effect, general head boundary conditions were assigned to the 10 model cells impacted by the dewatering operation during the 1986 steady-state simulation. Dewatering was reduced in 1986, and no longer impacted the model boundary. During the subsequent transient simulation, flow was assumed to be parallel to the boundary, and a no flow condition was assigned.

The eastern and western boundaries of the flow model were set as specified-flux boundaries. Hydrographs and periodic water-level measurements show that water levels at the boundaries change relatively little through time and are not greatly affected by seasonal stresses or pumping stresses within the study area (Shindel and others, 1980-91). Values of specified flux were assigned on the basis of K_h and interpolated heads on the eastern and western boundaries of the model. External boundaries also are at a relatively large distance from pumping centers to minimize any possible effects on model results.

Ground-water withdrawals in the model are based on rates for the collector wells which were obtained from records maintained at the Parsons Avenue Water Treatment Plant (Douglas Chambers, City of Columbus Division of Water, written commun., 1989-92). Other ground-water withdrawals in the study area were estimated from Ohio Environmental Protection Agency records of discharges to the Scioto River and Big Walnut Creek (David Parkinson, Ohio Environmental Protection Agency, written commun., 1989). Recharge is added to the model as a flux specified to the uppermost active model layer. Stream- and quarry-seepage rates were simulated as head-dependent flux boundaries.

Model Parameters

The data base for the numerical model was developed by reviewing previous analytical and numerical work performed in the area of the well field. Additional work was done to determine values of K_{vsb} because it had been determined to be a parameter critical to construction of a realistic model (Childress and others, 1991). Parameters in areas where aquifer-test data were not available were estimated from well logs, geologic sections, and trial-and-error adjustments during model calibration. Input parameters to the model include measured and estimated values of recharge, hydraulic head, hydraulic conductivity, bedrock altitude, streambed conductance, pumping rates, specific yield, and storage coefficient.

Because model layers 1 and 2 contain the same hydrogeologic unit, their horizontal hydraulic conductivities are identical in the vertical plane. Horizontal hydraulic conductivity (K_h) values of layer 1 varied spatially and ranged from 30 to 375 ft/d (fig. 14). (Figs 14-31 at back of report). The relatively low values of K_h to the west of the Scioto River represent the deposits of till with lenses of sand and gravel. K_h values in the northeastern part of the study area were set at 50 ft/d, on the basis of interpretation of an aquifer test in the USGS files and well logs. K_h in the flood plain of the Scioto River and Big Walnut Creek was 125 ft/d. K_h values near the collector wells were based on aquifer tests and ranged from 300 to 375 ft/d. The K_h values determined from analysis of aquifer tests were decreased somewhat for use in the numerical model. Estimates of saturated thickness for use in analytical solutions take into account only the more permeable parts of the saturated unconsolidated material. The saturated thickness in the numerical solution must account for the entire saturated thickness of the model layer. Because of the discontinuous nature of glacial deposits, this results in a lower mean K_h for the cell than commonly is determined from analysis of the pumping tests. Transmissivity of model layer 1 is shown in figure 15.

As previously stated, K_h values in model layer 2 are identical to those in layer 1. Two model layers are used to isolate the Scioto River and Big Walnut Creek, which are represented by head-dependent flux cells in model layer 1, from model layer 2. This enables more accurate simulation of stream/aquifer relations as well as better representation of vertical flow components. The thickness and transmissivity (fig. 16) of model layer 2 change due to the undulating bedrock surface (fig. 17).

The K_h values of the carbonate bedrock aquifer range from 10 to 15 ft/d (fig. 18). The higher K_h was assigned to the bedrock not capped by a shale unit and was based on aquifer tests performed to characterize the inflows to a sewerline constructed in the western part of the study area (Williams and others, 1988). The lower K_h was assigned to the eastern part of the study area where shale units overlie the carbonate bedrock. Transmissivity of layer 3 is shown in figure 19.

Vertical hydraulic conductance is a measure of the ability of the aquifer(s) to transmit water vertically. Vertical hydraulic conductance was calculated from horizontal hydraulic conductivity and the vertical distances between nodes in the three layers. Distributions of vertical hydraulic conductance are shown in figures 20 and 21. The storage coefficient of model layer 3 was set at a constant value of 0.0002 on the basis of published values from the carbonate bedrock aquifer in Franklin County (Schmidt and Goldthwait, 1958). The specific yield of layer 1 ranges from 0.12 to 0.30 as determined from aquifer tests, published values from unconsolidated materials (Johnson, 1967), and model calibration. Specific yield was distributed areally in correspondence to the final values of K_h used in the model. (See table at the top of the next column of text.) Values of storage coefficient assigned to model layer 2 differed according to its saturated thickness. A constant specific storage of 0.00006 was used to compute these values.

Horizontal hydraulic conductivity (feet per day)	Specific yield (dimensionles)
30	0.12
50	0.09
125	0.12
300	0.30
310	0.30
375	0.30

On the basis of precipitation and surficial geology, recharge was distributed to the uppermost model layer by areas shown in figure 22. The recharge factor is the percentage of precipitation that was added as a specified flux to the model. The recharge rate ranged from 4.0 to 12.0 in/yr for steady-state simulations. Recharge for transient stress periods was added as specified flux to the uppermost model layer as a percentage of the total precipitation during each stress period on the basis of figure 23 and hydrographs of wells measured hourly within the study area. Recharge rates for the steady-state simulations and for each stress period of the transient simulation are listed in Table 1.

Model Calibration

Model calibration is a process through which a mathematical representation of the physical system is developed to aid in accurate simulation of ground-water flow. The ground-water-flow model was calibrated under steady-state and transient conditions by using the hydraulic head and stream gain/loss data described previously as calibration targets. The goodness-of-fit of a particular suite of input parameter values was evaluated visually by comparing contour maps of simulated versus measured hydraulic heads in each model layer, as well as quantitatively by computing statistical measures of the residuals between simulated and measured hydraulic heads. The mean absolute error and the root mean squared error between simulated and measured heads were used as quantitative measures of the goodness-of-fit.

Steady-State Model

Simulated heads were compared with two sets of measured heads from two different steady-state periods to calibrate the steady-state model. Simulation of multiple sets of steady-state calibration targets increases the uniqueness of the flow solution. Measured heads from October 1979 were used as a calibration target for the time prior to construction of the collector wells. These head measurements are assumed to represent the steady-state configuration of the potentiometric surface because pumping stresses within the study area were confined to long-term, nearly constant pumpage associated with quarry operations and small domestic water supplies. Water levels measured in October 1979 were near mean water levels for the year because of high precipitation that fall. Measured heads from March 1986 were used as a calibration target for the period after commencement of pumping at the collector wells. These head measurements are believed to represent the near-steady-state configuration of the potentiometric surface in response to pumpage because well-field pumpage had been at nearly the same rate for several years.

For all model calibration runs, simulated heads were compared to measured heads by use of the mean absolute error (MAE) and root mean square error (RMSE) to determine the goodness-of-fit of the simulation based on a given number of observations (n). MAE is the mean of the sum of the absolute differences between the observed (H_{obs}) and simulated (H_{sim}) heads. RMSE is the square root of the mean squared difference between the observed and simulated heads.

$$MAE = \sum \frac{|Hsim - Hobs|}{n}$$

$$RMSE = \sqrt{\sum \frac{(Hsim - Hobs)^2}{n}}$$

Transient Model

A transient simulation adds complexity to the solution of the ground-water-flow equation in several ways: (1) time, in addition to space, must be discretized, (2) aquifer storage must be characterized, and (3) initial conditions must be specified.

Time was discretized into 8 stress periods differing in length from 93 to 487 days on the basis of recharge, well-field pumpage, and available water-level measurements for calibration targets. Stress periods were subdivided into a variable number of time steps on the basis of stress-period length. The length of the stress period and the significance of the changing stress determined the number of time steps in each stress period. Heads calculated from the previous stress period were used as the starting heads for the next stress period. Simulated heads from the March 1986 steady-state calibration were used as the

Table 1. Rates of recharge distributed to the uppermost active model layer, by area and time period

[A, B, and C refer to areas delineated in figure 22; SP, stress period]

Distribution area	Time period and recharge rate (inches per year)							
Steady-state simulation		1979			1986			
A		4.0			4.0			
B		9.0			9.0			
C		8.0			8.0			
D		12.0			12.0			
Transient simulation	SP1	SP2	SP3	SP4	SP5	SP6	SP7	SP8
A	0	3.5	0	3.5	0	6.4	6.1	4.4
B	0	7.9	0	7.9	0	14.4	13.8	9.9
C	0	7.0	0	7.0	0	12.8	12.2	8.8
D	0	10.6	0	10.6	0	19.2	18.3	13.2

starting heads for transient simulation. A comparison between precipitation and well-field pumpage by stress period is shown in figure 25. Table 3 summarizes the time discretization for the transient simulation.

Internal stresses to the model were computed for each stress period. These internal stresses include recharge, well-field and other pumpage, and stream stage. Recharge rates were varied on the basis of precipitation during the stress period and the response of hydrographs from wells in the study area. Well-pumpage rates were varied according to records maintained by personnel at the Parsons Avenue Water Treatment Plant and were determined by the median discharge in the final month of the stress period. This rate is nearly equivalent to the median discharge over the stress period. Stream-stage variation is based on the median gage height recorded over the period of uniform model stress. Stage is recorded by the USGS gages on Big Walnut Creek at Reese (eastern boundary of the study area), and Scioto River at Columbus (about 1 mi upstream from northern boundary of study area). Stage measurements from the Scioto River at Columbus gage were based on an adjustment between the gage at Scioto River at Columbus and the incomplete record from a City of Columbus river stage recorder on the Scioto River at collector well 104 (fig. 2). Well-field discharges (by individual collector well) and stream stage are presented in table 4 on the following page.

Transient model calibration was based on seven sets of hydraulic-head measurements made from March 1986 to June 1991. This time period includes large-scale increases in well-field production associated with a drought in the summer of 1988. The modeled time period also includes a period of exceptionally high rainfall in 1990.

Table 2. Steady-state water budget, March 1986

[Flow rates in million gallons per day, well-field pumpage of 8.4 Mgal/d]

Flow to aquifer from:	
Wells	0.0
Recharge	11.0
River leakage	9.4
Specified flux:	
Glacial drift	6.5
Bedrock	9.9
Total in	36.8
Flow from aquifer to:	
Quarry	20.5
Wells	8.4
Recharge	0.0
River leakage	7.9
Specified flux:	
Glacial drift	0.0
Bedrock	0.0
Total out	36.8

Table 3. Stress-period duration and observed water levels available for transient simulation

Stress-period number	Stress-period dates	Stress-period length (days)	Number of observed water levels
1	Mar. 1986 - Oct. 1986	214	35
2	Nov. 1986 - Mar. 1987	151	35
3	Apr. 1987 - Sep. 1987	183	21
4	Oct. 1987 - Apr. 1988	24	30
5	May. 1988 - Aug. 1988	93	14
6	Sep. 1988 - Dec. 1989	487	23
7	Jan. 1990 - Sep. 1990	274	29
8	Oct.1990 - June 1991	275	30

Table 4. Well-field discharge and stream-stage variation in transient simulation
[CW, collector well; SP, stress period;]

Collector well	Approximate pumpage, in million gallons per day SP1	SP2	SP3	SP4	SP5	SP6	SP7	SP8
CW-101	2.6	4.2	6.8	6.8	11.2	9.8	7.2	7.3
CW-103	2.8	2.3	4.6	4.6	5.7	2.7	5.7	5.0
CW-104	2.3	1.5	2.7	2.7	3.6	5.2	4.3	4.5
CW-115	1.6	2.4	1.8	1.8	3.7	5.0	3.8	4.9
Total	9.3	10.4	15.9	15.9	24.2	22.7	21.0	21.8

Stream	Stream stage[1], in feet above sea level SP1	SP2	SP3	SP4	SP5	SP6	SP7	SP8
Scioto River	684.8	684.5	684.5	684.5	684.8	684.8	686.3	686.3
Big Walnut Creek	699.9	699.5	699.5	699.5	700.0	700.0	700.6	700.6

[1]Stream stage in the model cell containing the uppermost stream reach in the study area.

Transient model calibration was based on seven sets of hydraulic-head measurements made from March 1986 to June 1991. This time period includes large-scale increases in well-field production associated with a drought in the summer of 1988. The modeled time period also includes a period of exceptionally high rainfall in 1990.

The transient calibration was evaluated in the same manner as the steady-state simulations—by visually comparing contour maps of simulated versus measured heads in each model layer and by minimizing values of MAE and RMSE. In addition, measured hydrographs from wells with water-level data available over the entire simulation period were compared visually with simulated hydrographs over the same period. In this manner, one can check that the measured and simulated water levels maintain the proper relation throughout the transient simulation.

Variations in the potentiometric surface of the glacial drift aquifer are shown in figures 26 and 27. The composite cone of depression around the Scioto River wells increased with each stress period through stress period 5 (August 1988). A water-level difference map, showing the water-level difference between simulation with and without well-field production, illustrates the minus-5-ft area of influence of the well field at maximum drawdown (fig. 28). These figures illustrate the increase in the area of influence of the collector well through time, as well as the changes in gaining and losing reaches of the Scioto River and Big Walnut Creek. The MAE and RMSE at the end of each stress period in the transient simulation are listed on each potentiometric surface. The values of MAE ranged from 3.1 to 5.5 ft. The values of RMSE ranged from 3.6 to 7.0 ft.

As an additional calibration step, hydrographs were matched for wells in several areas of the model that had long-term water-level data. These wells include FR-116 from the till plain to the west, FR-120 and FR-121 from the Scioto River valley, and FR-123 from the northeastern part of the study area (fig. 2). The observed and simulated hydrographs for these wells are presented in figure 29. These hydrographs, from different hydrogeologic settings within the study area, show that the transient model realistically simulates measured water-level changes.

The transient water balance is summarized in table 5. The water balance indicates that the average well-field production for the transient simulation was 18.6 Mgal/d. Because the simulation began and ended under similar hydrologic conditions, the balance for the entire simulation indicates little change in storage. The major inflows to the aquifer were boundary flow, river leakage, and areal recharge. Major outflows were the collector wells and the bedrock quarry.

Table 5. Water budget for transient simulation

[Flow rates in million gallons per day]

Flow to aquifer from:	
Storage	2.6
Recharge	8.9
Surface-water leakage	22.3
Specified flux:	
Glacial drift	6.5
Bedrock	9.9
Total in	50.2
Flow from aquifer to:	
Storage	3.1
Quarry	20.6
Wells	18.6
Recharge	0.0
Surface-water gains	7.9
Specified flux:	
Glacial drift	0.0
Bedrock	0.0
Total out	50.2

Sensitivity Analysis of Transient Model

Sensitivity analysis is the aspect of model construction used to assess the confidence associated with the values of input parameters used in the calibrated model. It is a useful way to identify model inputs that have the most influence on model predictions. Systematic changes in selected hydraulic characteristics and boundary conditions allow for evaluation of model sensitivity and potential simulation error. The analysis is used to determine whether the differences between simulated and observed data can be accounted for by the range of uncertainty in the input parameters and boundary conditions.

Individual input parameters are increased and decreased by a constant factor while all other parameters were unchanged. The difference between simulated and observed values of head ("residual head") are used to evaluate model sensitivity. The factor used to change input parameters differed for each parameter based on a hydrologically reasonable range of that parameter.

For each sensitivity simulation, the residual head between simulated and observed head at each observation point was computed and graphically displayed in a boxplot.

Parameters that were changed in the sensitivity analysis included horizontal hydraulic conductivity of layers 1 and 2 ($K_{1,2}$), transmissivity of layer 3 (T_3), vertical conductance between layers 2 and 3 (*Vcont*), specific yield of layer 1 (*Sy*), storage coefficient of layers 1 and 2 ($S_{1,2}$), storage coefficient of layer 3 (S_3), river stage, riverbed conductance, and recharge. Riverbed conductance and *Vcont* were multiplied by 2, 5 and 10. Parameters $K_{1,2}$, T_3, *Sy*, $S_{1,2}$, S_3, and recharge were changed by ±25 and ±50 percent. The transient sensitivity analysis required 42 model runs.

For each sensitivity simulation, the residual head between simulated and observed heads at each observation point was computed as previously described. The mean absolute error and the root mean squared error of the residual heads are plotted in line graphs in figures 30 and 31. This method was used to standardize the presentation of the model sensitivity to changes in input parameters. In these figures, the minimum residual head is used as the measure of model sensitivity. These figures illustrate that the model is not highly sensitive to increases in $K_{1,2}$, *Sy*, $S_{1,2}$, S_3, and recharge, (positive changes in parameters result in little change in RMSE and MAE). The model is more sensitive to decreases in parameter values, particularly $K_{1,2}$, T_3, *Vcont*, and river stage. The model is most sensitive to changes in the streambed conductance

SUMMARY

The South Well Field consists of four radial collector wells completed in permeable outwash and ice-contact deposits that are intersected by the Scioto River and Big Walnut Creek. The wells are designed to produce approximately 42 Mgal/d, a portion of that yield resulting from induced infiltration of surface water from the Scioto River and Big Walnut Creek. The well field supplied up to 30 percent of the city's water supply in 1991, serving southern Columbus and its suburbs.

A transient ground-water-flow model was constructed by use of the USGS three-dimensional finite-difference ground-water-flow code MODFLOW. The model was constructed to improve the understanding of the hydrogeology, ground-water-flow directions, ground-water/surface-water interactions, and hydrologic budget of the study area. The steady-state and transient models will be used in a subsequent study to determine the steady-state and transient contributing recharge areas to the collector wells.

The primary source of ground water in the study area is the glacial drift aquifer. The aquifer consists of heterogeneous deposits of sand, gravel, and clay. The thickest, most productive parts of the glacial drift aquifer are in the buried valleys in the central and eastern parts of the study area underlying the Scioto River and Big Walnut Creek. A secondary source of ground water within the study area is the carbonate bedrock aquifer, which consists of Silurian and Devonian limestones, dolomites, and shales.

From December 1989 through June 1991, quarterly water levels were measured, when accessible, in 44 wells. These measurements were used to construct potentiometric-surface maps and hydrographs of the glacial drift and carbonate bedrock aquifers for calibration of the transient model. Five wells were equipped with data loggers to record hourly water-level changes. Hourly data also were used to help determine recharge rates to the aquifer and vertical hydraulic gradients within the glacial drift aquifer and between the glacial drift and carbonate bedrock aquifers.

The ground-water-flow system is conceptualized as follows. Recharge, boundary flow, and river leakage are the principal sources of water to the flow system. The study area is bounded on the north and south by streamlines, with flow entering the area from the east and west. Areal recharge is contributed throughout the study area, although a higher percentage of precipitation reaches the water table in the area east of the Scioto River where little surface drainage exists. Vertical ground-water flow is downward in the uplands of the Scioto River, and upward near the river in both the glacial drift and carbonate bedrock aquifers.

To more accurately estimate the streambed characteristics of the Scioto River in the study area, the river was mapped on the basis of three riverine settings: pools, riffles, and runs. Vertical hydraulic conductivity of the streambed materials was measured directly at each of these settings by use of a streambed seepage meter. A uniform vertical hydraulic conductivity was then used throughout each of the river settings. Because of the difference in elevation between Big Walnut Creek and the Scioto River, Big Walnut Creek is a naturally losing stream throughout most of the study area.

A variable-spaced finite-difference grid containing 53 rows, 45 columns, and 3 layers was used to simulate the flow system. The length and width of grid cells range from 200 to 2,000 ft with finer spacings designed to increase detail in the areas near the collector wells. Vertically discretizing the glacial drift aquifer into two layers increased the vertical resolution of the model and isolated the collector wells in a different model layer than the rivers and their tributaries. Layer 3 included all of the bedrock in the study area to an elevation of 250 ft above sea level.

K_h values of layer 1 varied spatially and ranged from 30 to 375 ft/d. The relatively low values of K_h to the west of the Scioto River represent the deposits of till with lenses of sand and gravel. K_h in the flood plain of the Scioto River and Big Walnut Creek was 125 ft/d. K_h values near the collector wells ranged from 300 to 375 ft/d. K_h values in the northeastern part of the study area were set at 50 ft/d. The K_h values of the carbonate bedrock aquifer range from 10 to 15 ft/d.

Simulated heads were compared with two sets of measured heads from two different steady-state periods to calibrate the steady-state model in order to increase the uniqueness of the flow solution. Measured heads from October 1979 were used as a calibration target for the time prior to construction of the collector wells. Measured heads from March 1986 were used as a calibration target for the period after commencement of pumping at the collector wells. In general, simulated heads matched observed heads to within ±5.0 ft in all model layers. The mean absolute and root mean square errors were 2.3 and 3.8 ft, respectively, for the October 1979 calibration, and were 3.0 and 4.8 ft, respectively, for the March 1986 calibration.

Simulated heads from the March 1986 steady-state simulation were used as the starting heads for the transient simulation. Specific yield (0.12 to 0.30) and storativity (0.0002) values were added to the model to account for transient changes in ground-water storage. Recharge, river stage, and well-field pumpage were varied by stress period.

The largest increase in drawdown for the transient simulation occurred in stress period 5 (August 1988). The mean absolute and root mean square errors for the transient simulation were computed for 7 of the 8 stress periods. Mean absolute error ranged from 3.1 to 5.5 ft, and root mean square error ranged from 3.6 to 7.0 ft.

This ground-water-flow model has several potential future uses. The model could be used to determine the effect of variations in pumping schedules and/or rates on aquifer water levels. The model also could be used to determine the effect of additional production wells placed within the study area. Delineation of recharge areas and discharge areas and estimates of advective times-of-travel can now be done in a subsequent study based on this model. With some modification, the model can be used to estimate the impact of increased mining activities in the area.

The ground-water-flow model is limited in several ways. Limited calibration targets in the carbonate-bedrock system reduce the confidence associated the simulation of those units. There is also a lack of data in the glacial drift aquifer in the northeast part of the modeled area. The eastern and western boundary conditions limit the amount of drawdown that can be simulated within the study area. Time and space discretization limit the effectiveness of the simulation in the proximity of the collector wells. The time discretization in the model is designed to simulate constant conditions over the stress period, so calibration targets adjacent to the wells may not be matched accurately due to short-term fluctuations in pumping rates. A periodic post-audit of the model would help to refine estimates of the hydraulic parameters and boundary conditions of the aquifer system as boundary conditions change and as additional hydraulic data are collected.

The current study is an improvement in several ways over previous investigations in and around the South Well Field. The ground-water-flow model was improved by increasing horizontal and vertical discretization in the vicinity of the wells. The refined finite-difference grid allowed for better simulation of stream/aquifer relations. The additional fieldwork and geologic investigation, in concert with the transient simulation, refined estimates of hydraulic parameters such as streambed permeability, horizontal and vertical hydraulic conductivity, and storativity.

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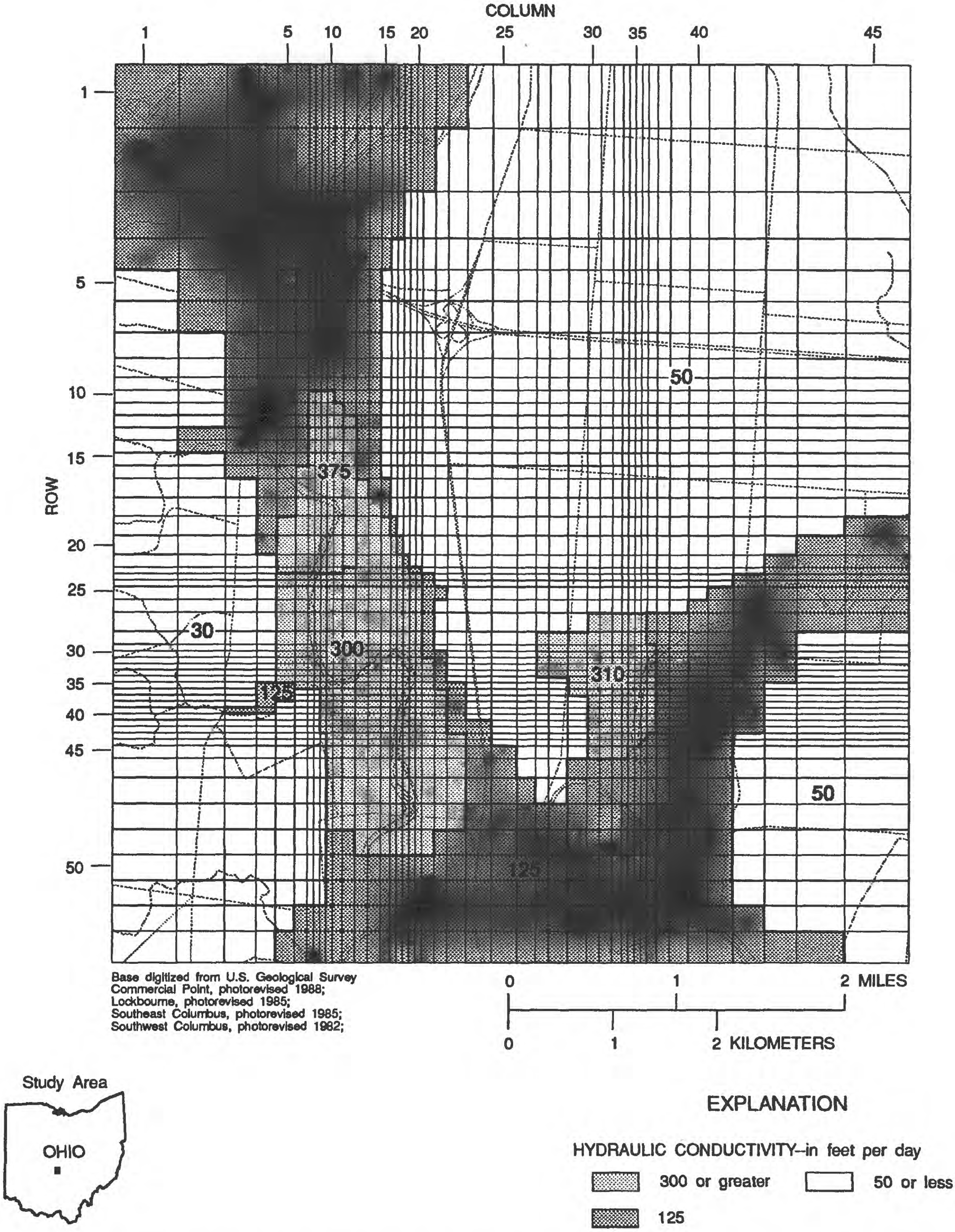


Figure 14. Areal distribution of horizontal hydraulic conductivity used in model layers 1 and 2 to simulate the glacial drift aquifer.

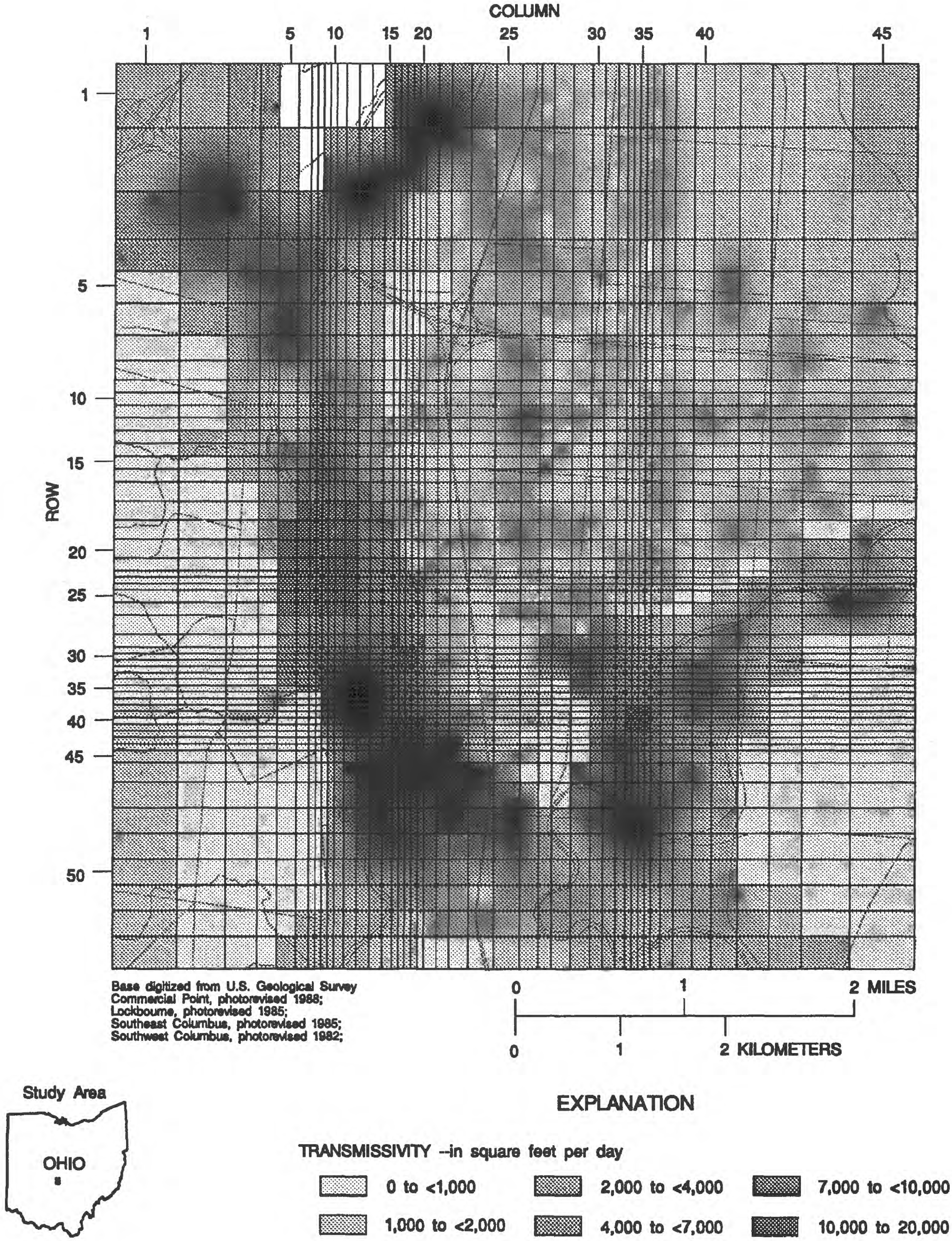


Figure 15. Areal distribution of transmissivity used in model layer 1.

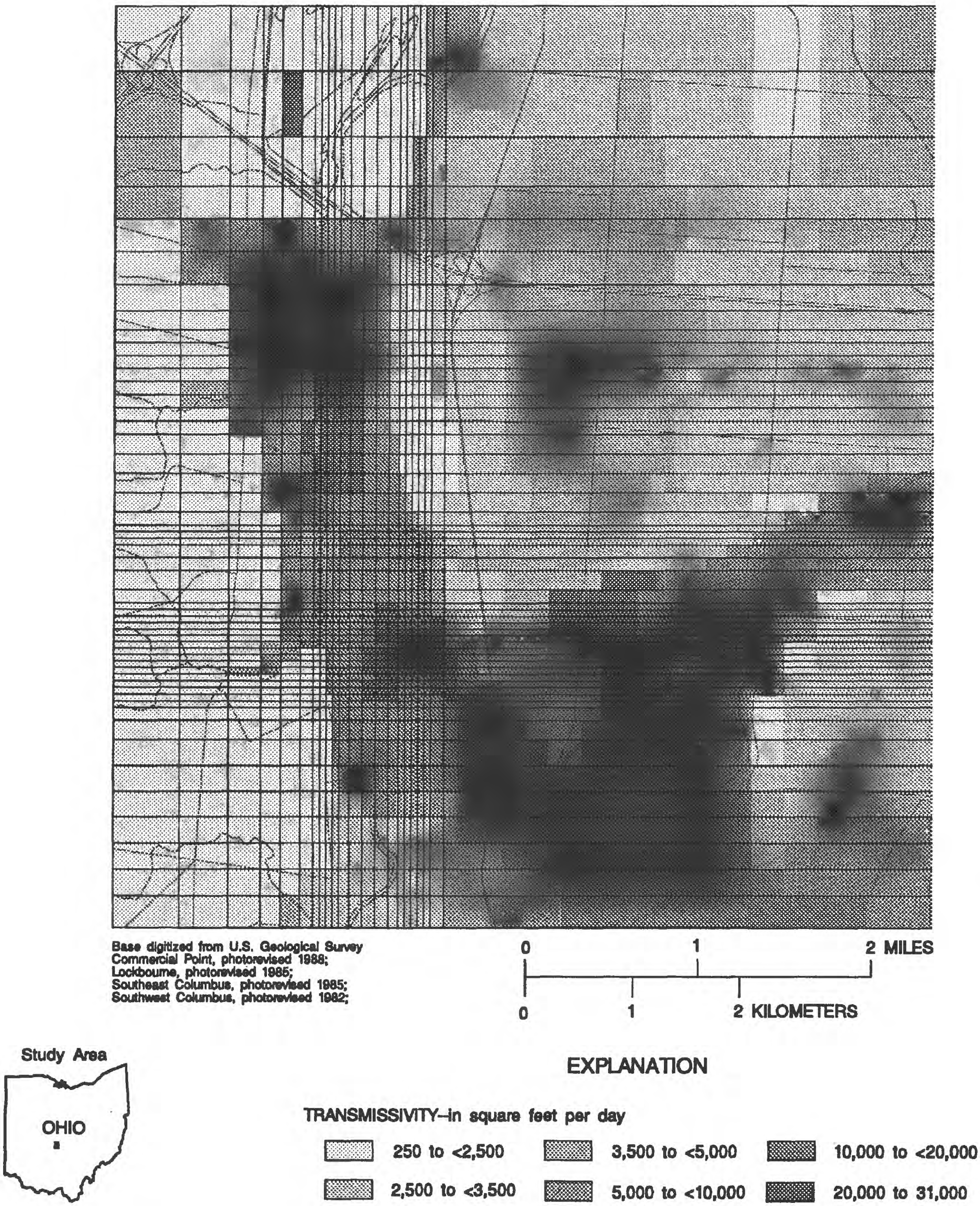


Figure 16. Areal distribution of transmissivity used in model layer 2.

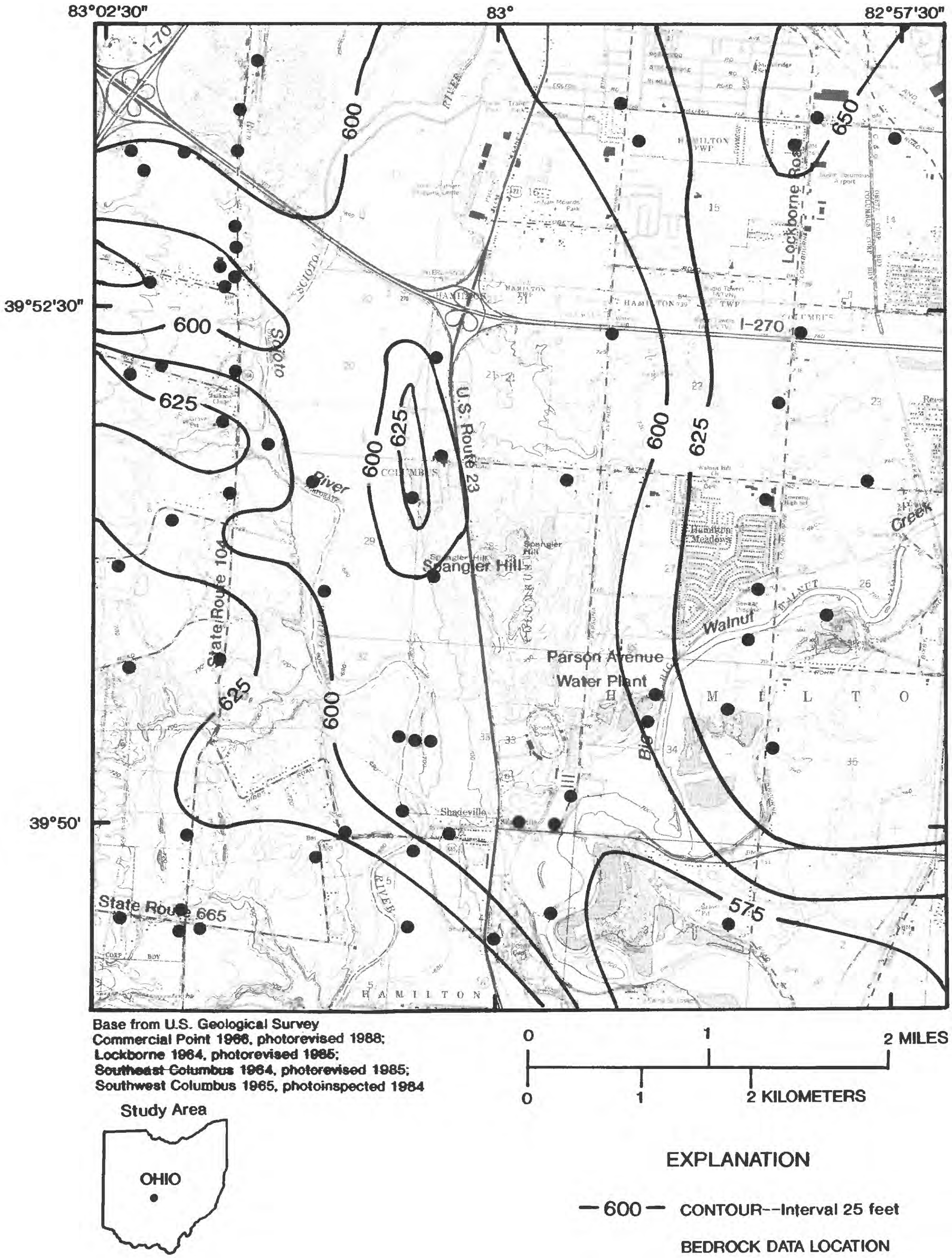
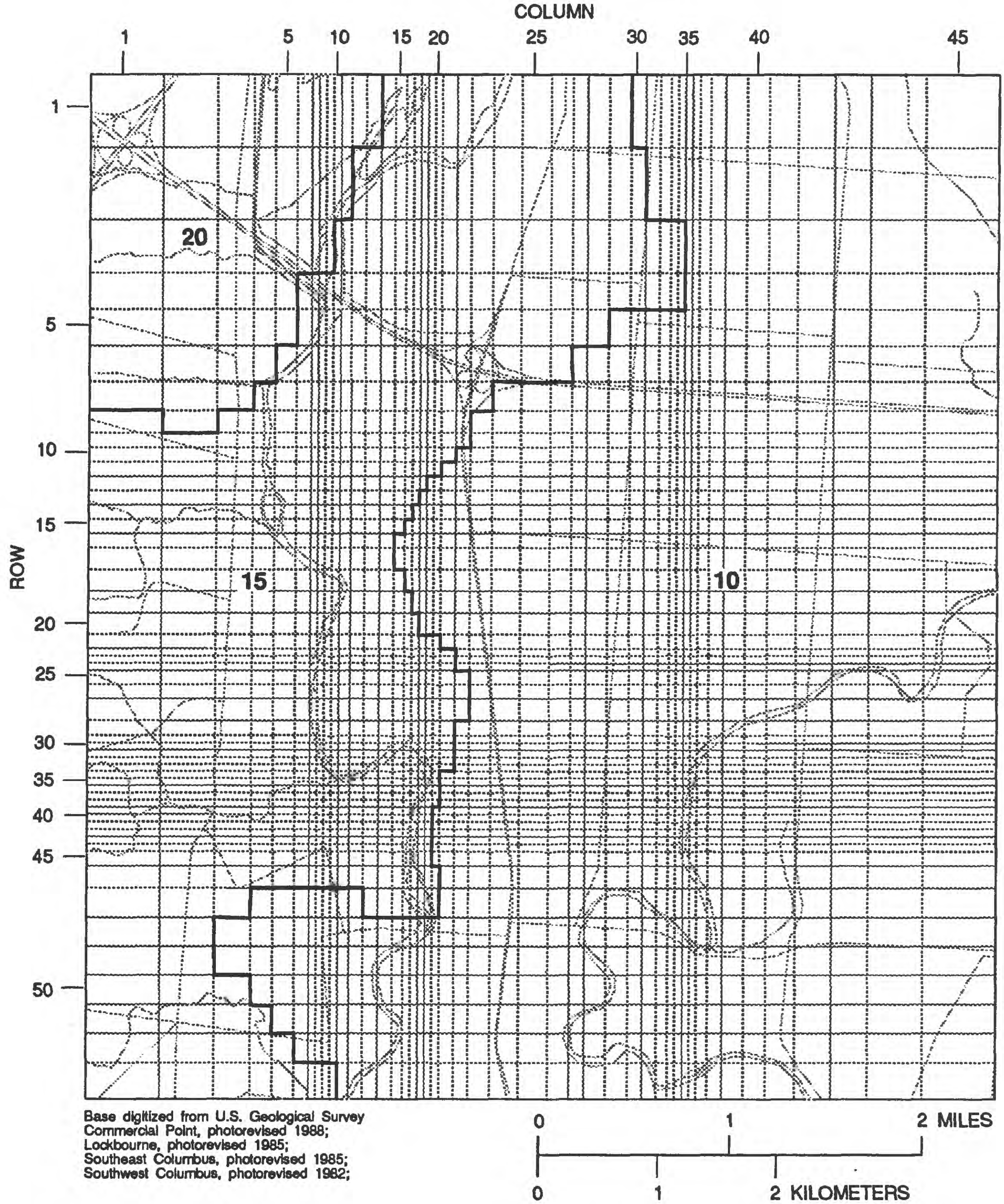


Figure 17. Configuration of bedrock surface.



EXPLANATION

15 HYDRAULIC CONDUCTIVITY–in feet per day

Figure 18. Areal distribution of horizontal hydraulic conductivity used to calculate transmissivity of the bedrock aquifer, model layer 3.

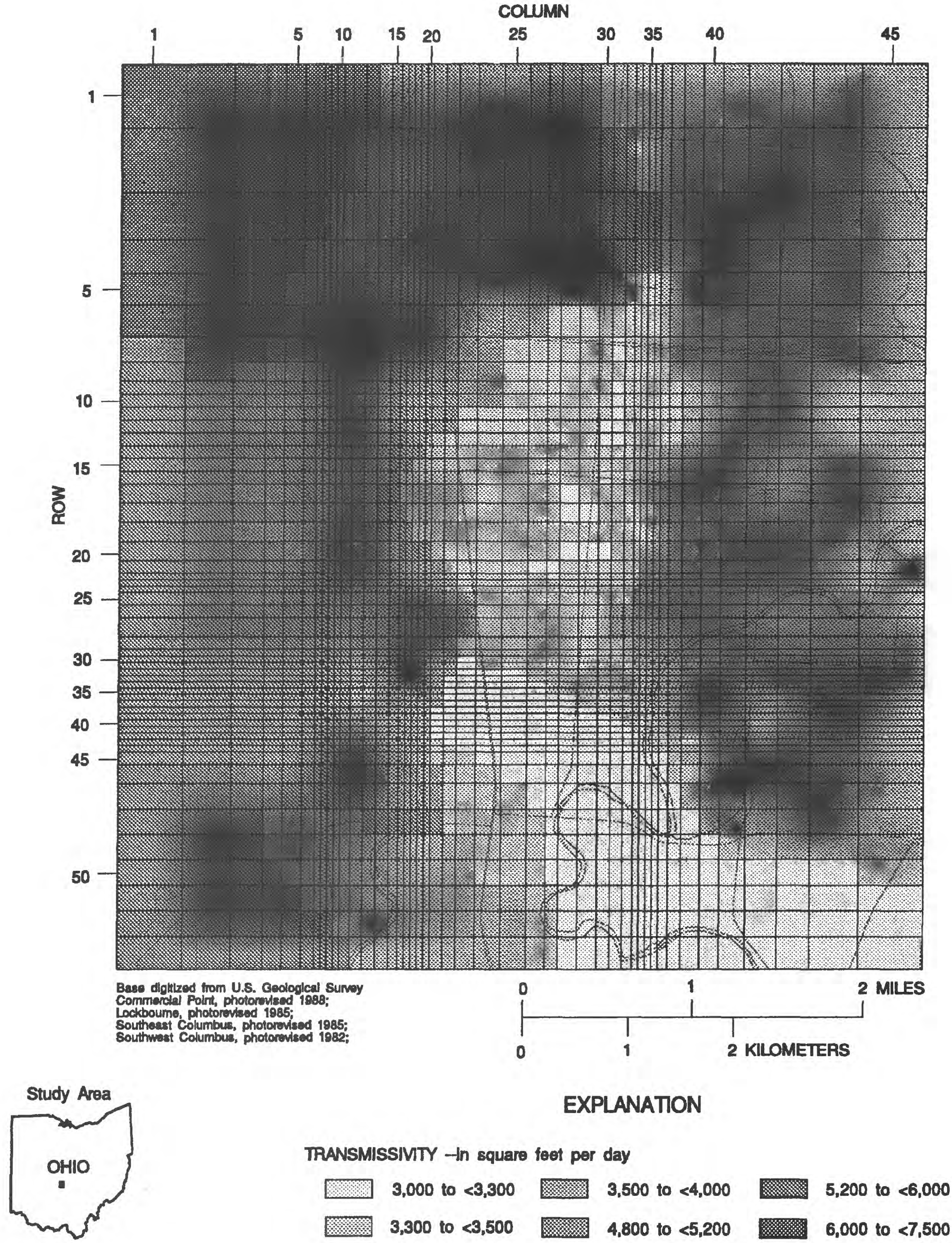


Figure 19. Areal distribution of transmissivity used in model layer 3.

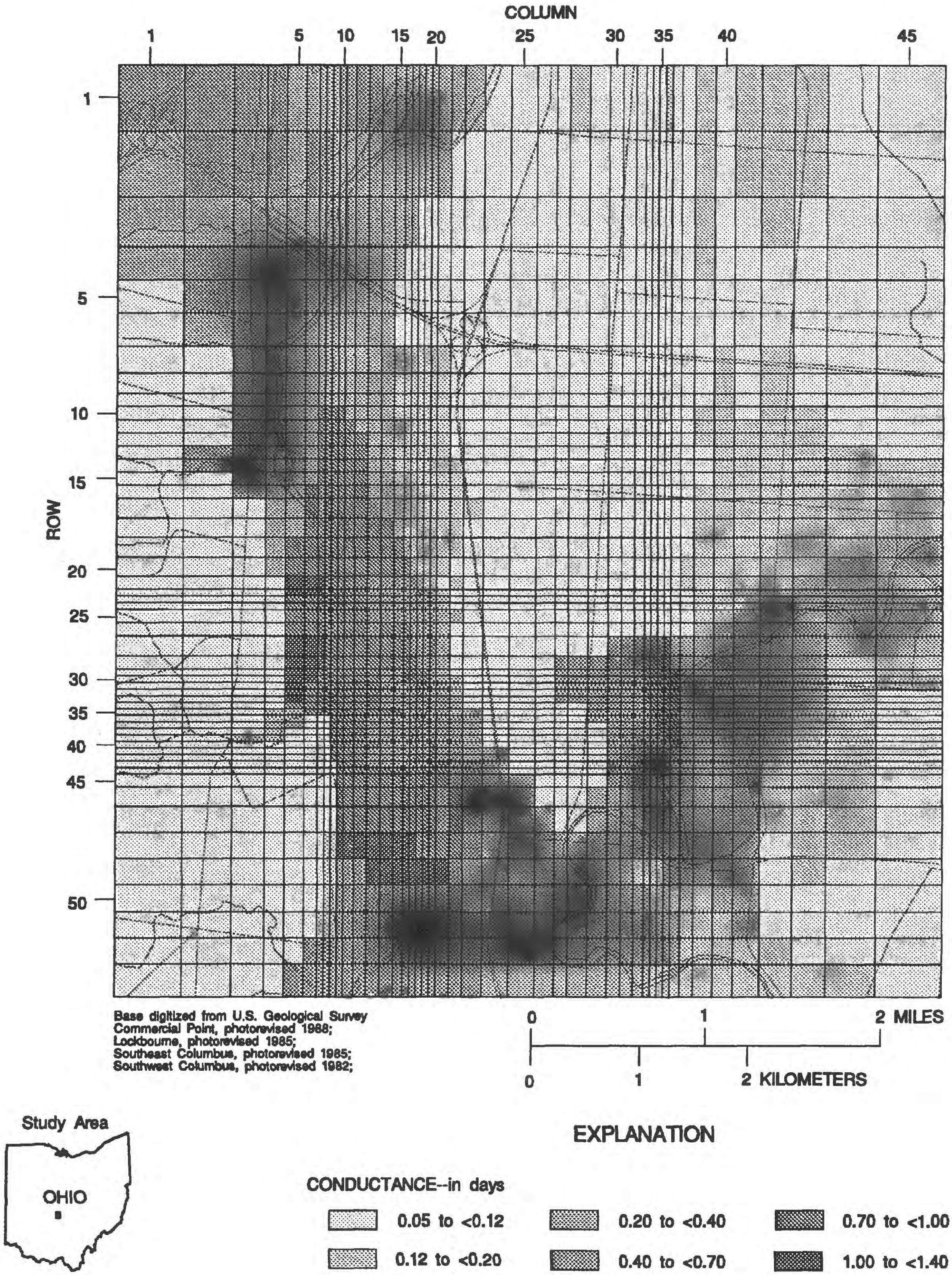


Figure 20. Areal distribution of vertical conductance between model layers 1 and 2.

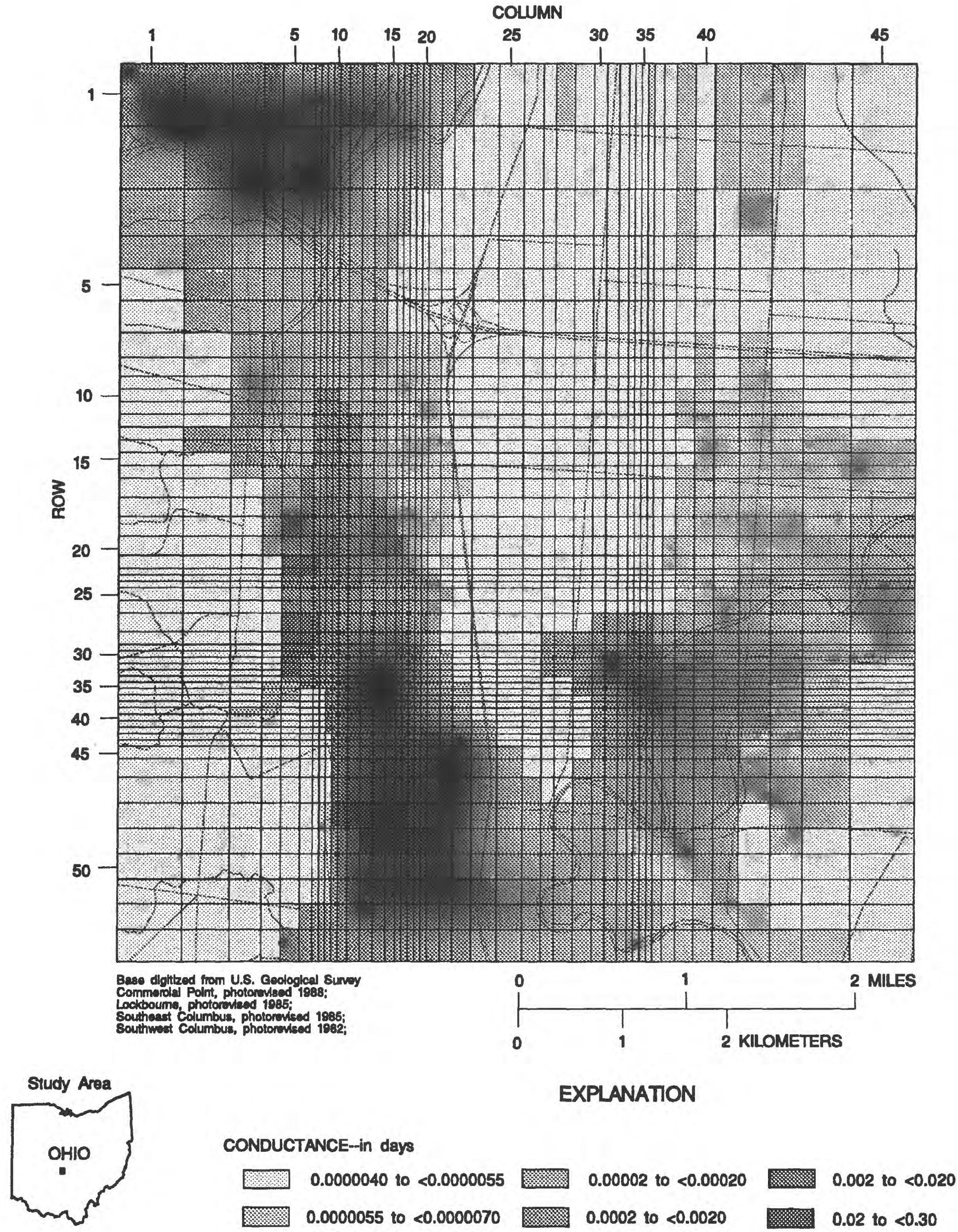
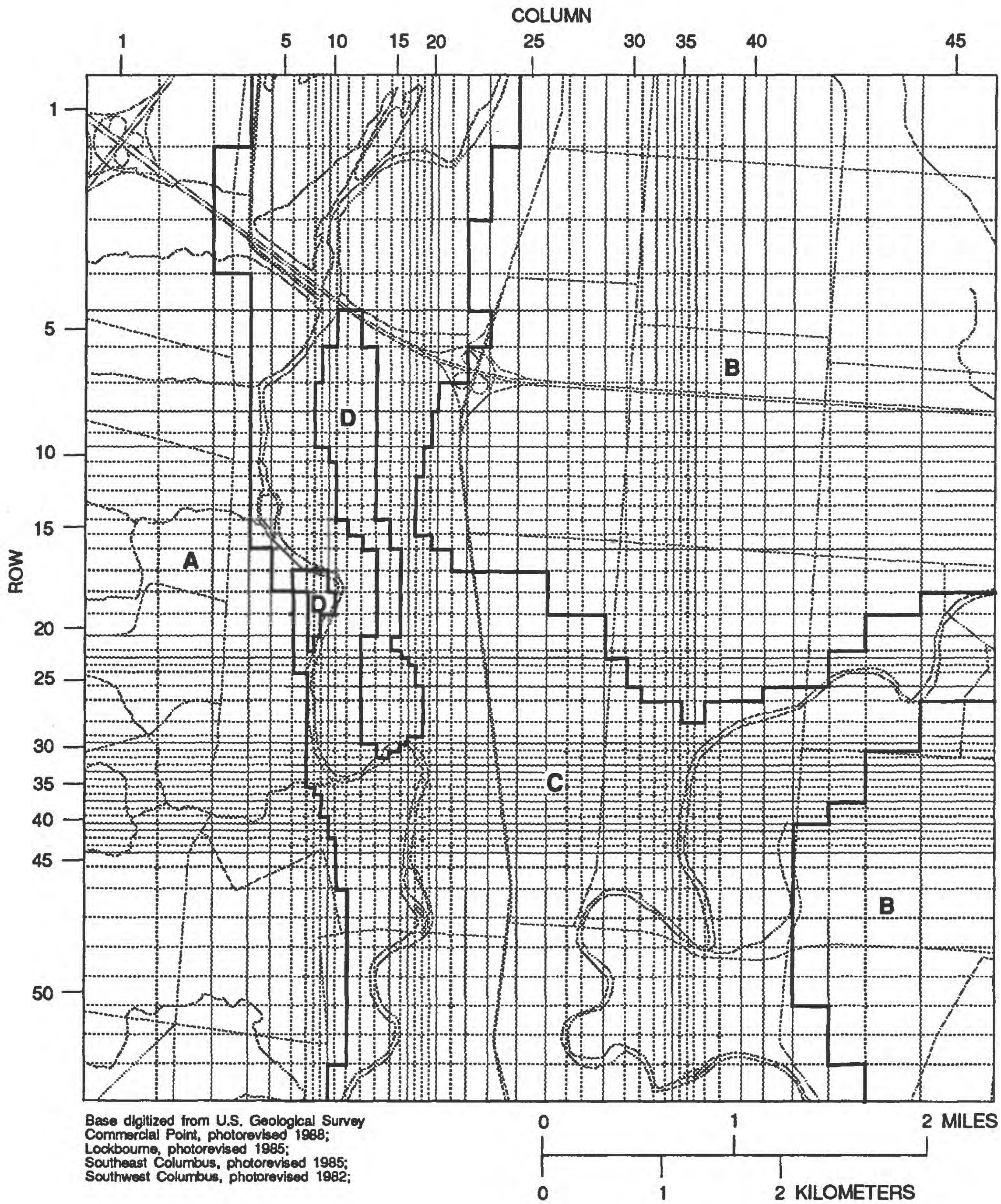
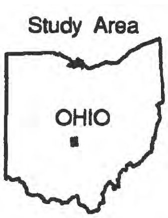


Figure 21. Areal distribution of vertical conductance between model layers 2 and 3.





EXPLANATION

A REFERENCE LETTER FOR RECHARGE RATES--listed in table 1

Figure 22. Areal distribution of recharge used to simulate the glacial-bedrock aquifer system.

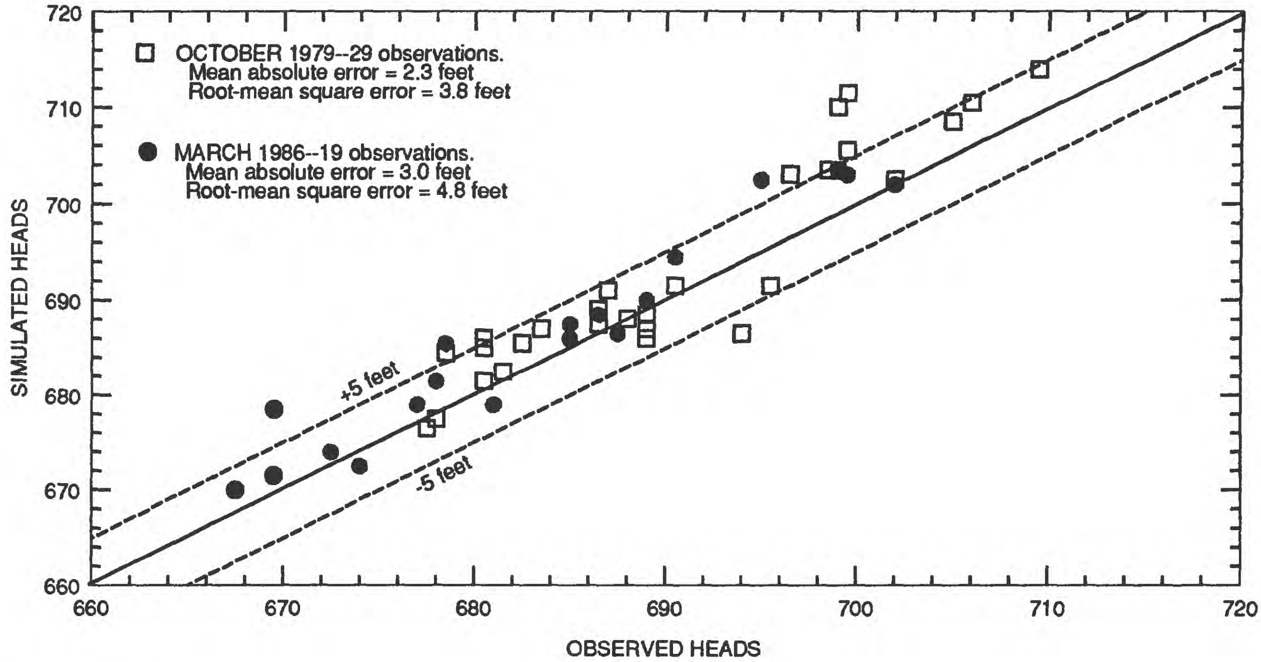


Figure 23. Comparison of the observed and simulated heads in all wells for all model layers, steady-state conditions, October 1979 and March 1986.

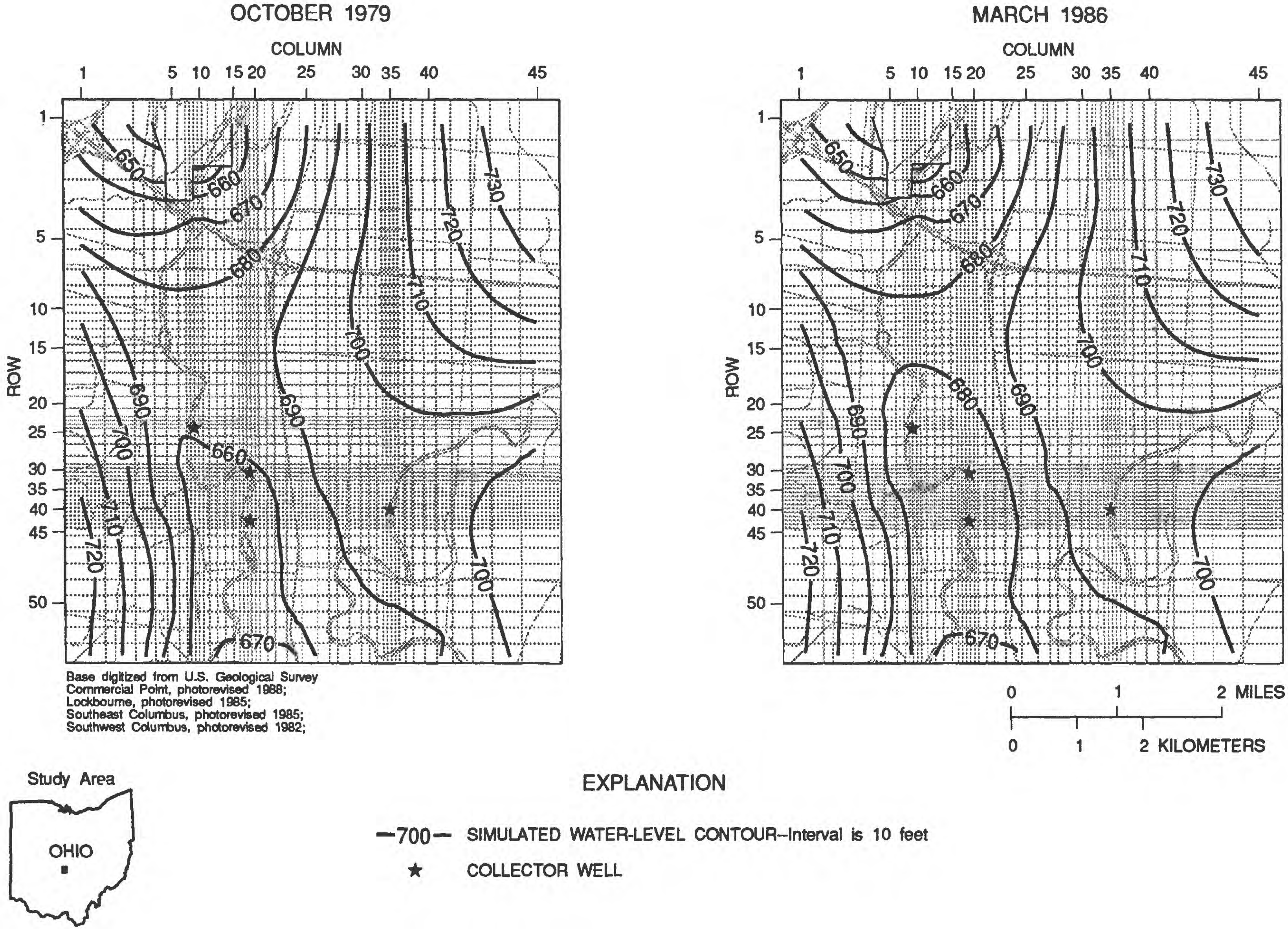


Figure 24. Simulated potentiometric surface of glacial-drift aquifer steady-state simulation, October 1979 and March 1986.

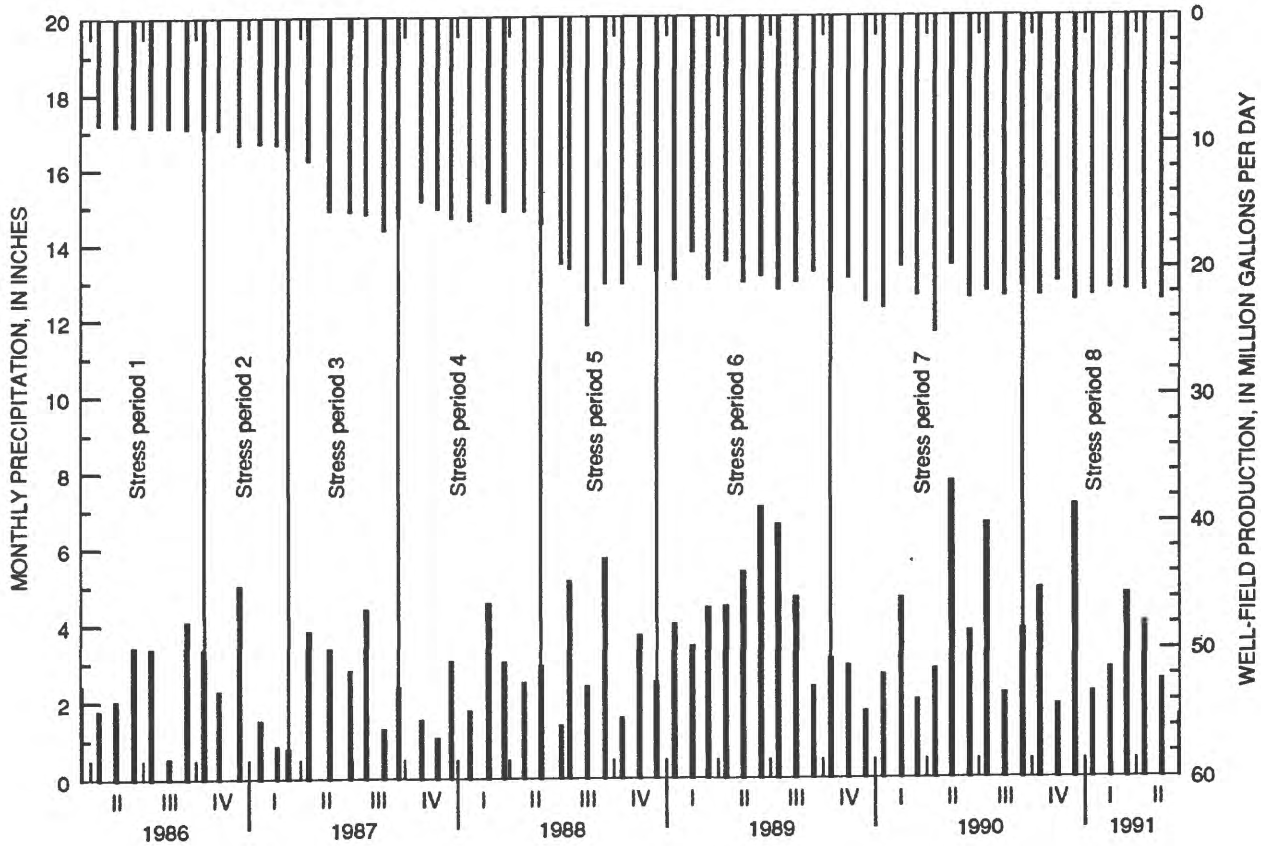


Figure 25. Total monthly precipitation and well-field production, by stress period.

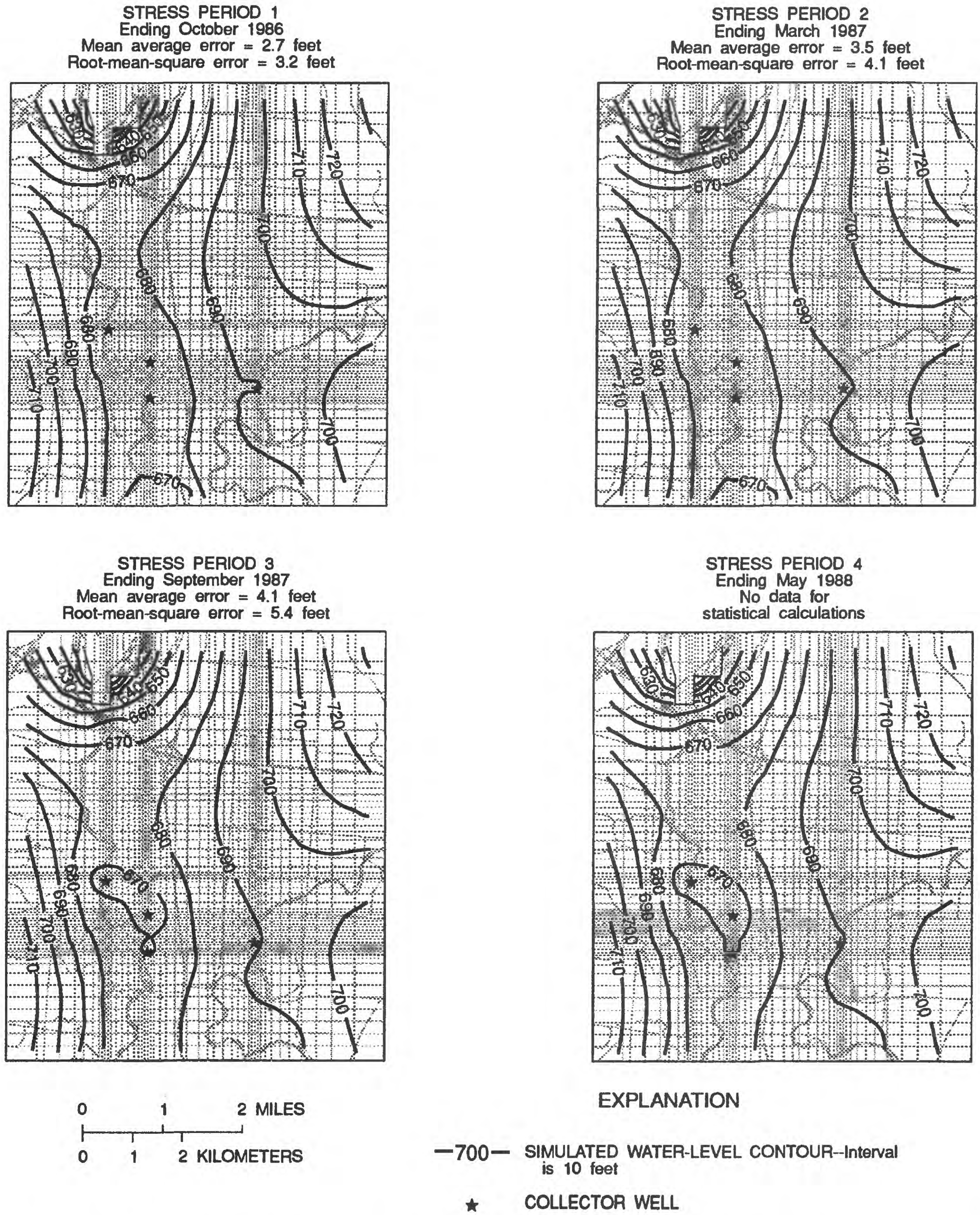
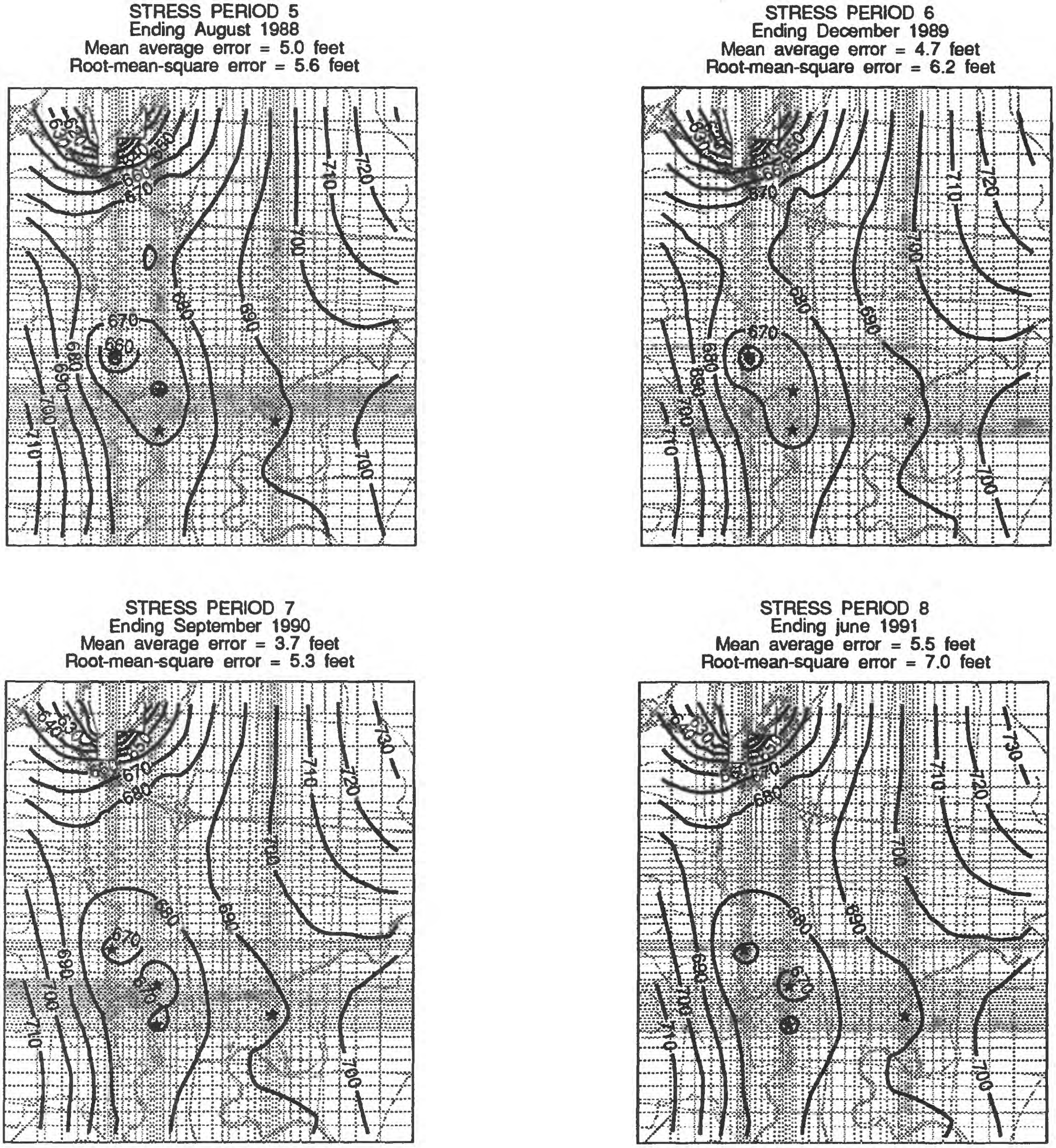


Figure 26. Simulated potentiometric surface of the glacial drift aquifer, model layer 2, for stress period 1-4.



0 1 2 MILES

0 1 2 KILOMETERS

EXPLANATION

—700— SIMULATED WATER-LEVEL CONTOUR--Interval is 10 feet

★ COLLECTOR WELL

Figure 27. Simulated potentiometric surface of the glacial drift aquifer, model layer 2, for stress period 5-8.

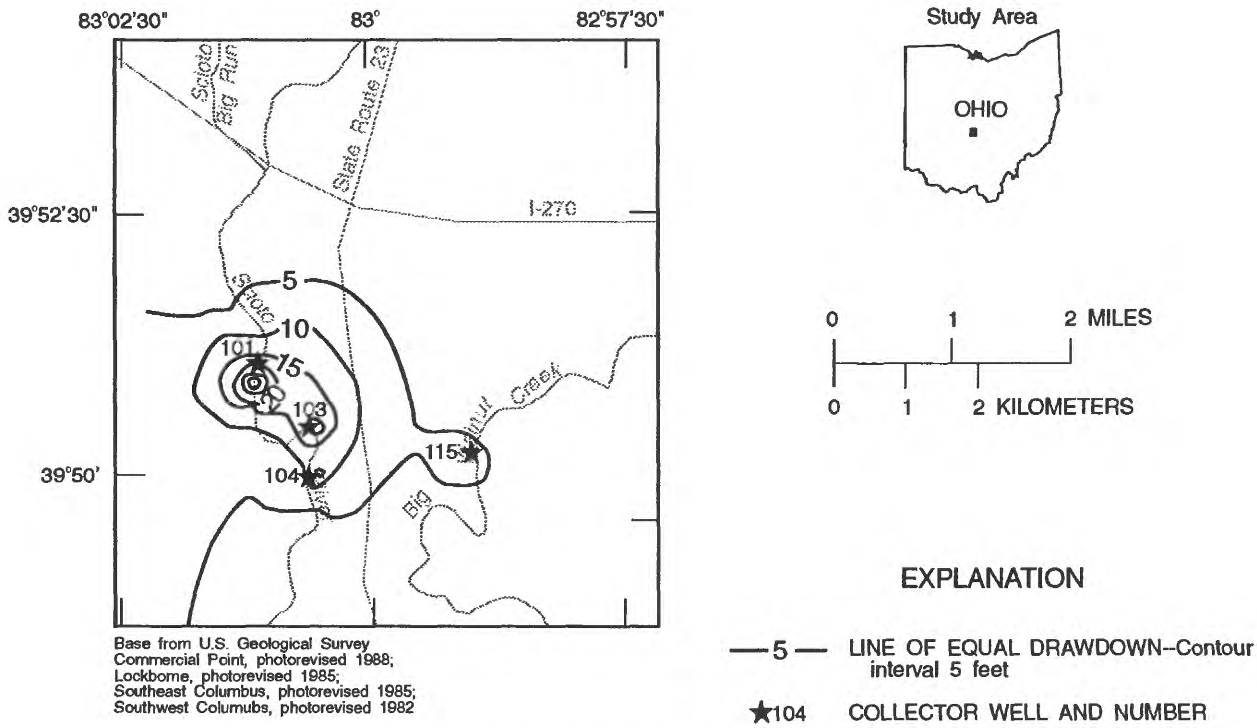


Figure 28. Head-differences, stress of August 1988, subtracted from simulation with no pumping.

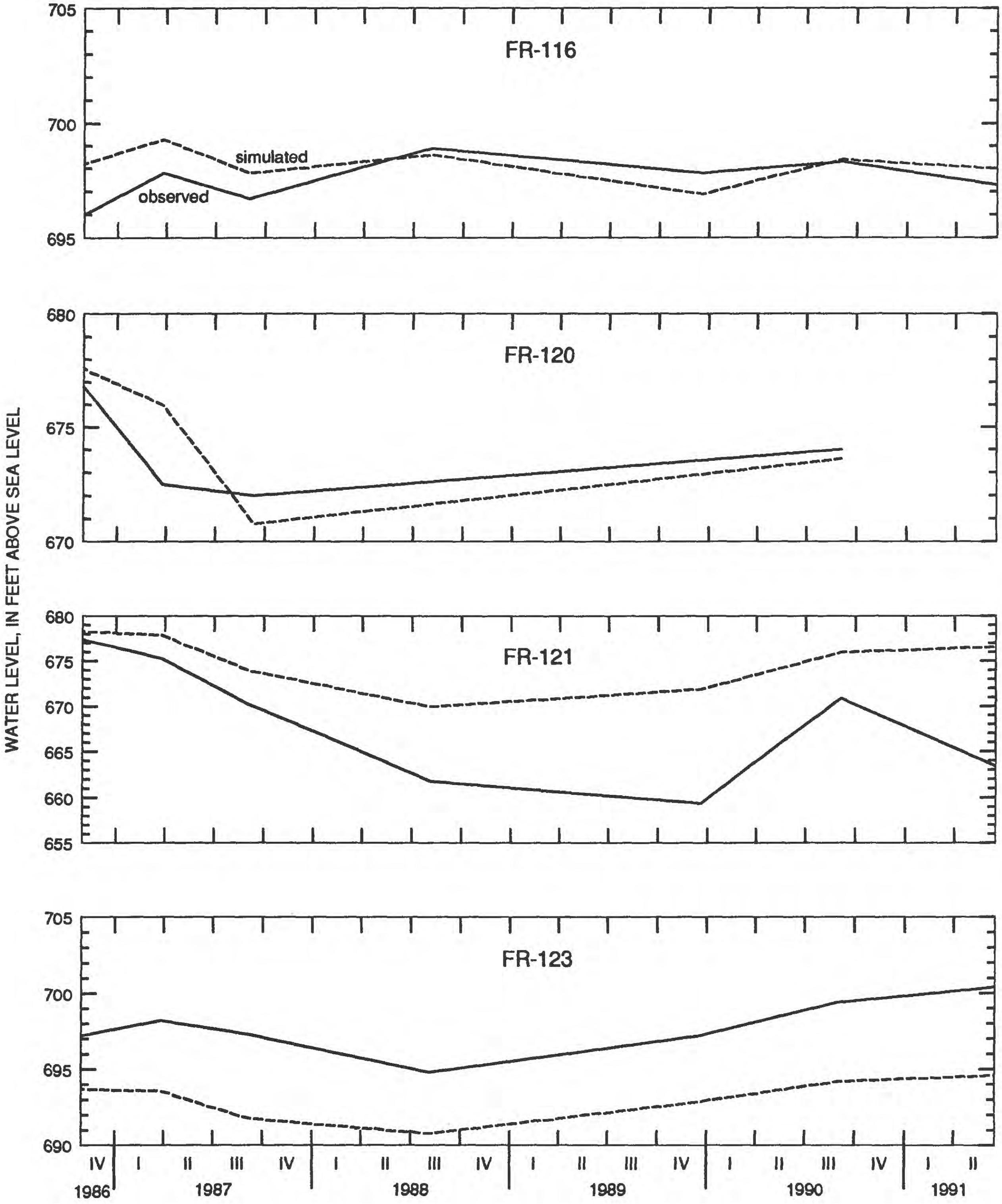


Figure 29. Hydrograph showing observed and simulated hydraulic heads at wells FR-116, FR-120, FR-121, and FR-123.

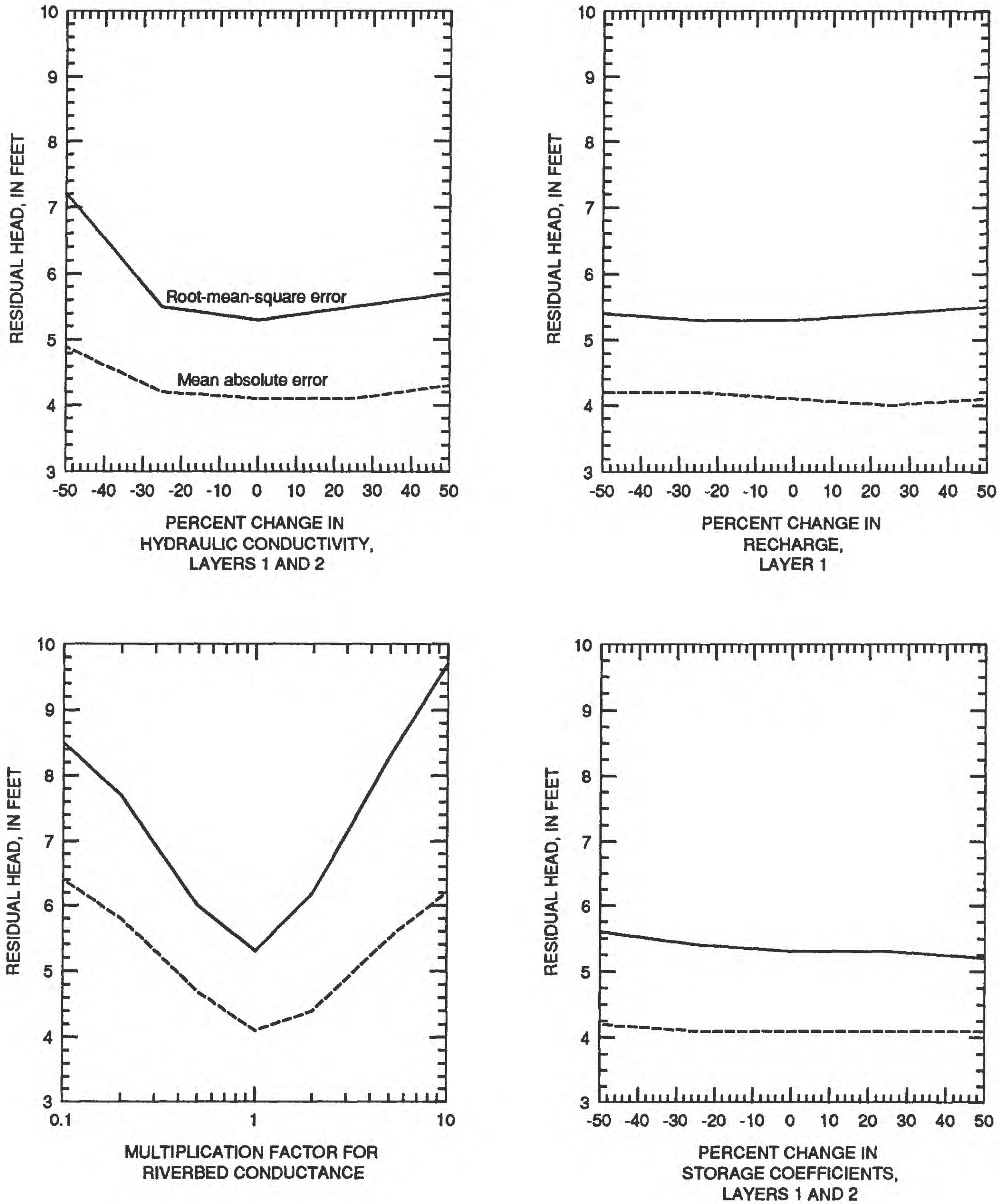


Figure 30. Changes in summary statistics in response to changes in hydraulic conductivity of layers 1 and 2, recharge, riverbed conductance, and storage coefficient of layers 1 and 2.

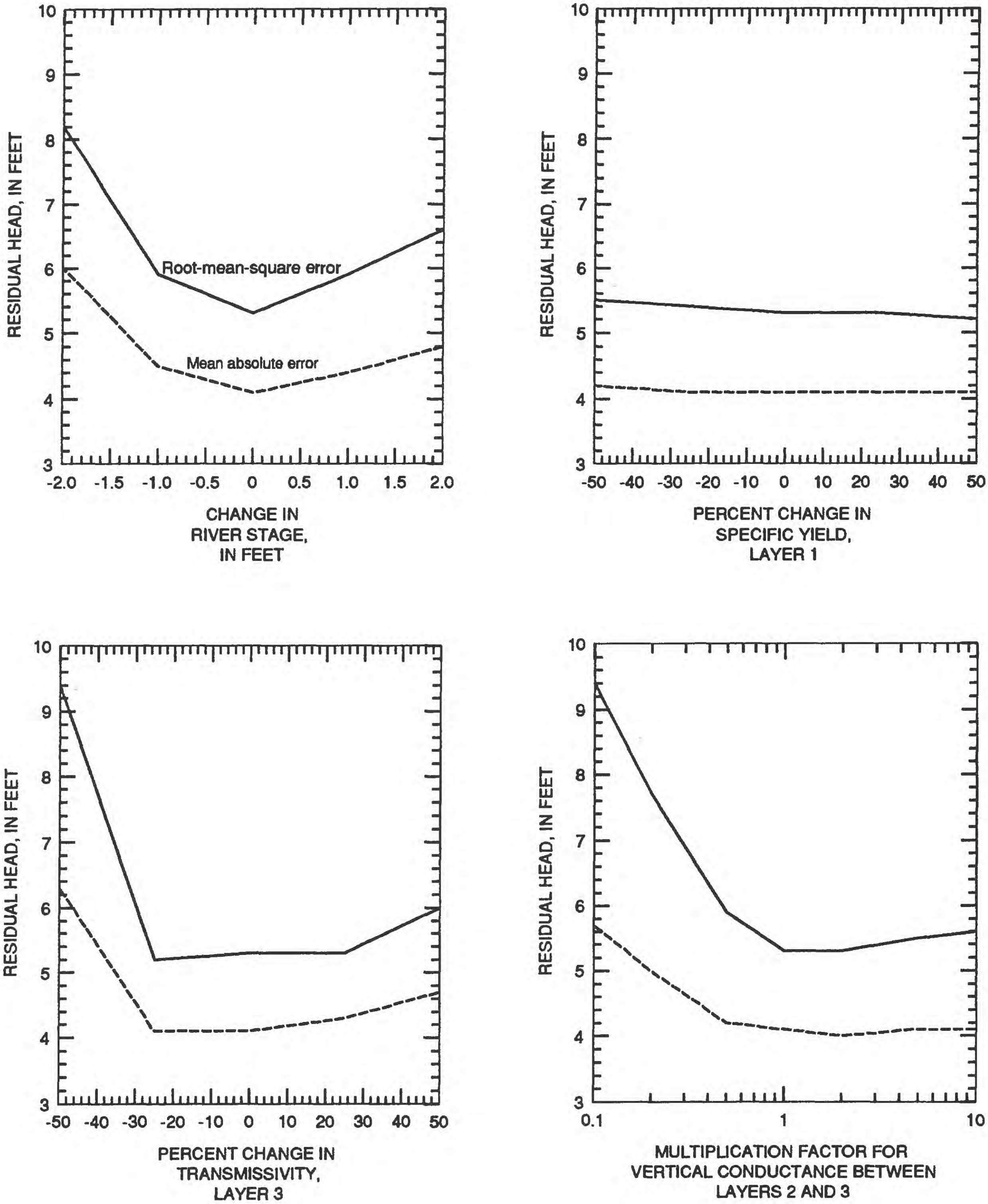


Figure 31. *Changes in summary statistics in response to changes in river stage, specific yield, transmissivity, and vertical conductance.*

APPENDIIX

Appendix 1. Records of selected wells in southern Franklin County, Ohio

[* indicates water level measurement from Malcolm-Pirnie, Inc. (1988), --- indicates missing data]

Local well number	Latitude (degrees)	Longitude (degrees)	Year of construction	Altitude of land surface (feet)	Depth of well (feet)	Water-level date	Depth below measuring point (feet)[a]	Water level altitude (feet)[b]
FR-100	395134	830100	1975	688.0	56.8	10-15-79	5.94	682.1
						10-31-86	13.50	674.5 *
						03-26-87	20.00	668.0 *
						12-20-89	23.55	664.4
						06-20-91	21.50	666.5
FR-101	395114	830104	1975	688.1	70.0	10-15-79	6.27	681.8
						10-31-86	18.50	669.6 *
						03-26-87	15.30	672.8 *
FR-101 TH-41	395114	830105	1975	685.6	82	12-20-89	35.80	649.8
						09-05-90	19.00	666.6
						06-20-91	26.52	659.1
FR-101 TH-42	395116	830104	1975	687.3	81	12-20-89	26.31	661.0
						09-05-90	20.89	666.4
						06-20-91	20.69	666.6
FR-101 TH-40	395115	830103	1974	685	80	09-05-90	18.36	666.6
						06-20-91	28.08	656.9
FR-101 TH 46	395114	830102	1975	687.5	80.2	09-03-87	24.42	663.1
						08-05-88	33.65	653.8
						12-20-89	34.30	653.2
						09-05-90	18.18	669.3
						06-20-91	28.80	658.7
FR-103	395046	830031	1974	699.0	101	10-15-79	22.90	676.1
FR-103 TH-11	395045	830025	1974	699	93	03-12-86	27.28	671.7
						10-31-86	31.20	667.8 *
						03-26-87	46.60	652.4 *
						09-02-87	43.13	655.9
						08-05-88	59.47	639.5
						12-20-89	52.28	646.7
						09-05-90	44.58	654.4
						06-20-91	62.50	636.5
FR-104	395020	830031	1975	691.0	80.0	10-15-79	6.45	684.5
						10-31-86	21.80	669.2 *
						03-26-87	21.20	669.9 *
FR-104 TH-18	395021	830029	1975	691.0	76	03-12-86	18.94	672.1
						10-31-86	19.60	671.4 *
						03-26-87	21.20	669.8 *
						09-02-87	27.67	663.3
						08-05-88	38.16	652.8
						12-20-89	54.45	636.5
						09-05-90	32.04	659.0
						06-20-91	43.41	647.6

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Local well number	Latitude (degrees)	Longitude (degrees)	Year of construction	Altitude of land surface (feet)	Depth of well (feet)	Water-level date	Depth below measuring point (feet)[a]	Water level altitude (feet)[b]
FR-104 TH-20	395020	830033	1975	690	82.0	09-03-87	22.30	667.7
						08-05-88	31.49	658.5
						12-20-89	45.66	644.3
						09-05-90	17.69	672.3
						06-20-91	26.39	663.6
FR-104 TH-72	395020	830034	1975	680	100	03-12-86	9.91	670.1
						09-02-87	18.94	661.1
						08-05-88	30.95	649.0
						12-20-89	55.37	624.6
						09-05-90	26.28	653.7
						06-20-91	36.98	643.0
FR-104 TH-73	395020	830037	1975	685	67.0	10-31-86	14.10	670.9 *
						03-26-87	16.30	668.7 *
						09-09-87	17.21	667.8
FR-109	395157	830035	1975	702.2	92.0	10-15-79	14.66	687.5
						03-12-86	16.64	686.6
						10-31-86	19.20	683.0 *
						03-26-87	24.90	677.3 *
						09-02-87	22.76	679.4
						08-05-88	30.56	671.6
FR-115	395042	825858	1975	710	116	10-17-79	14.45	695.5
						10-31-86	20.30	689.7 *
						03-26-87	25.20	684.8 *
FR-115 TH 67	395039	825858	1975	721	116	10-31-86	32.70	688.3 *
						03-26-87	36.10	684.9 *
						09-02-87	34.45	686.5
						08-05-88	40.50	680.5
						12-19-89	41.02	680.0
						08-30-90	36.99	684.0
						06-20-91	36.95	684.0
FR-116	395006	830136	1977	722	62.0	03-12-86	21.93	700.1
						10-31-86	26.00	696.0 *
						03-26-87	24.20	697.8 *
						09-02-87	25.29	696.7
						09-04-87	25.35	696.6
						08-05-88	26.05	695.9
						12-19-89	24.17	697.8
						09-05-90	23.72	698.3
						06-21-91	24.67	697.3
FR-117	395016	830103	1977	700	45.0	10-15-79	14.09	685.9
						03-12-86	14.53	685.5

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Local well number	Latitude (degrees)	Longitude (degrees)	Year of construction	Altitude of land surface (feet)	Depth of well (feet)	Water-level date	Depth below measuring point (feet)[a]	Water level altitude (feet)[b]
FR-117	395016	830103	1977	700	45.0	10-31-86	16.50	683.5 *
						03-26-87	18.30	681.7 *
						09-02-87	17.59	682.4
						08-05-88	19.66	680.3
						12-19-89	19.47	680.5
						09-05-90	17.37	682.6
						06-21-91	18.70	681.3
FR-118	395039	830026	1977	700	98.0	10-15-79	19.50	680.5
FR-119	395111	830026	1977	700	85.0	10-15-79	14.80	685.2
						03-12-86	21.48	678.5
						10-31-86	22.50	677.5 *
						03-26-87	29.20	670.8 *
						09-02-87	32.17	667.8
						09-05-90	30.72	669.3
						06-20-91	39.75	660.3
FR-120	395117	830116	1977	685	72.0	10-16-79	2.52	682.5
						10-31-86	8.20	676.8 *
						03-26-87	12.50	672.5 *
						09-11-87	13.02	672.0
						09-05-90	11.04	674.0
FR-121	395123	830033	1977	690	45.0	10-15-79	8.13	681.9
						03-12-86	11.19	678.8
						10-31-86	12.60	677.4 *
						03-26-87	14.70	675.3 *
						09-02-87	19.67	670.3
						08-05-88	28.20	661.8
						12-20-89	30.53	659.5
						09-05-90	19.07	670.9
						06-20-91	26.59	663.4
FR-123	395131	825924	1977	710	36.5	03-12-86	7.66	702.3
						10-31-86	12.80	697.2 *
						03-26-87	11.80	698.2 *
						09-02-87	12.63	697.4
						08-05-88	14.96	695.0
						12-19-89	12.72	697.3
						09-05-90	10.57	699.6
						06-20-91	9.56	700.4
FR-124	395141	825814	1977	740	44.5	10-17-79	29.88	710.1
FR-125	395213	825919	1977	712	51.0	10-17-79	4.74	707.3
FR-126	395008	825931	1977	703	122	10-17-79	4.97	698.0
						03-12-86	15.42	687.6

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Local well number	Latitude (degrees)	Longitude (degrees)	Year of construction	Altitude of land surface (feet)	Depth of well (feet)	Water-level date	Depth below measuring point (feet)[a]	Water level altitude (feet)[b]
FR-126	395008	825931	1977	703	122	10-31-86	18.10	684.9 *
						03-26-87	20.20	682.8 *
						09-02-87	19.31	683.7
						08-05-88	21.07	681.9
						12-19-89	19.58	683.4
						09-05-90	13.38	689.6
						06-20-91	14.24	688.8
FR-127	395048	825954	1977	730	54.0	10-15-79	26.68	703.3
FR-130	395046	825734	1977	740	48.0	10-17-79	31.72	708.3
						10-31-86	37.10	702.9 *
						03-26-87	37.50	702.5 *
						12-20-89	36.99	703.0
FR-131	395126	830140	1977	728	53.0	10-15-79	41.11	686.9
						03-12-86	41.45	686.5
						10-31-86	47.10	680.9 *
						03-26-87	46.10	681.9 *
						08-05-88	48.86	679.1
						12-19-89	49.05	678.9
						09-05-90	45.58	682.4
						06-21-91	46.39	681.6
FR-141	395020	830144	1976	720	64	09-02-87	28.66	691.3
						12-19-89	27.96	692.0
						09-05-90	27.43	692.6
						06-21-91	28.52	691.5
FR-147	395108	830106	1975	685	78.8	03-12-86	6.46	678.5
						10-31-86	14.60	670.4 *
						03-26-87	4.10	680.9 *
						09-03-87	20.13	664.9
						08-05-88	28.96	656.0
						12-20-89	29.84	655.2
						09-05-90	16.36	668.6
						06-20-91	26.06	658.9
FR-148	395114	830102	1981	687	140	09-03-87	21.93	665.1
						08-05-88	31.79	655.2
						12-20-89	33.69	653.3
						06-20-91	29.02	658.0
FR-149	395024	830030	1981	683	144	09-03-87	16.28	666.7
						08-05-88	22.33	660.7
						12-20-89	25.20	657.8
						09-05-90	16.21	666.8
						06-20-91	21.86	661.1

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Local well number	Latitude (degrees)	Longitude (degrees)	Year of construction	Altitude of land surface (feet)	Depth of well (feet)	Water-level date	Depth below measuring point (feet)[a]	Water level altitude (feet)[b]
FR 151	395027	825925	1983	718	60.3	03-12-86	23.59	694.4
						03-26-86	23.00	695.0
						10-31-86	29.40	688.6 *
						03-26-87	31.90	686.1 *
						09-02-87	29.12	688.9
						08-17-88	32.98	685.0
						12-19-89	31.90	686.1
						09-05-90	28.67	689.3
						06-20-91	28.46	689.5
FR-18	394956	830027	1975	695	86.4	03-27-86	14.66	680.3
						10-31-86	20.00	675.0 *
						03-26-87	22.40	672.6 *
						09-08-87	24.11	670.9
						08-05-88	23.44	671.6
						08-15-88	23.99	671.0
						09-05-90	18.52	676.5
						06-21-91	21.96	673.0
FR-202	395314	830219	1977	752	220	03-13-86	83.10	668.9
						10-31-86	89.90	662.1 *
						03-26-87	96.10	655.9 *
						09-08-87	94.70	657.3
						08-22-88	90.03	662.0
						12-19-89	91.27	660.7
						09-05-90	88.45	663.5
						06-21-91	89.19	662.8
FR-209	395206	830145	1977	704	—	03-12-86	12.80	691.2
						10-31-86	25.30	678.7 *
						03-26-87	31.30	672.7 *
						09-02-87	16.08	687.9
						12-19-89	15.86	688.1
						09-05-90	14.51	689.5
						06-21-91	14.87	689.1
FR-213	395315	830200	1981	730	97.0	03-13-86	77.17	652.8
						09-01-87	80.29	649.7
						12-19-89	80.61	649.4
FR-244	395335	830137	1979	710	75.0	10-18-79	41.89	668.1
						03-13-86	65.15	644.8
						09-01-87	68.77	641.2
						12-19-89	70.03	640.0
						09-05-90	67.88	642.1
						06-21-91	68.77	641.2

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FR-246	395331	830139	1972	722	142	10-18-79	104.40	617.6
						03-13-86	117.19	604.8
						09-01-87	122.34	599.7
						12-19-89	122.83	599.2
						09-05-90	121.20	600.8
						06-21-91	122.12	599.9
FR-262	395255	830030	1982	692	49.4	10-31-86	38.00	675.0 *
						03-26-87	44.90	668.1 *
FR-264	395329	830131	1982	659	140.5	09-01-87	61.49	597.5
						09-08-87	94.70	564.3
FR-268	395321	830057	1982	680	64.0	09-01-87	34.22	645.8
FR-269	395323	830140	1988	705	90	12-19-89	69.00	636.0
						09-05-90	70.67	634.3
						06-21-91	68.21	636.8
FR-271	395055	825924	1986	710	86.0	10-31-86	16.20	693.8 *
						03-26-87	17.20	692.8 *
						12-20-89	18.50	691.5
						09-05-90	15.82	694.2
						06-21-91	14.86	695.1
FR-272	395055	825924	1986	710	45	09-05-90	16.35	693.6
						06-21-91	15.39	694.6
FR-273	395224	830005	1986	710	91.5	03-26-87	26.10	683.9 *
						09-05-90	14.41	695.6
						06-20-91	15.51	694.5
FR-274	395224	830005	1986	710	25.0	09-05-90	13.67	696.3
						06-20-91	13.40	696.6
FR-275	394941	830044		680	25	10-31-86	8.60	671.4 *
						03-26-87	10.20	669.8 *
FR-276	395239	830214	1984	755	155	09-05-90	72.25	682.8
						06-21-91	72.56	682.4
FR-277	4394930	830131	1972	713	52.0	10-31-86	23.30	689.7 *
						03-26-87	21.70	691.3 *
						09-05-90	18.08	694.9
						06-21-91	18.52	694.5
FR-278	395115	830226	1986	735	114	09-05-90	31.54	703.5
						06-21-91	31.67	703.3
FR-279	394932	830227	1985	735	145	09-05-90	17.99	717.0
						06-21-91	15.62	719.4
FR-3	395114	825732	1946	713.0	60.0	10-17-79	10.67	702.3
						03-31-86	11.22	701.8
						09-30-87	13.28	699.7

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FR-3	395114	825732	1946	713.0	60.0	08-31-88	12.96	700.0
						12-20-89	16.48	699.8
						09-05-90	13.25	699.8
						06-21-91	12.77	700.2
FR-32	395234	830113	1968	700	80.0	10-15-79	14.42	685.6
FR-36	395037	825819	1970	715	31.0	10-17-79	10.03	705.0
						03-12-86	12.07	702.9
						10-31-86	15.90	699.1 *
						03-26-87	17.00	698.0 *
						12-19-89	16.38	698.6
						09-05-90	16.95	698.0
						06-21-91	14.98	700.0
FR-37	395153	825916	1950	726.0	38.0	10-17-79	22.36	703.6
FR-46	395114	825926	1960	720	37.5	10-17-79	21.61	698.4
FR-48						10-31-86	31.90	688.1 *
						03-26-87	33.50	686.5 *
FR-52	395230	825913	1940	735	84.4	10-17-79	23.81	711.2
FR-59	394912	825937	1965	730	63.0	10-16-79	41.47	688.5
FR-60						10-31-86	48.70	681.3 *
						03-26-87	51.20	678.8 *
FR-61	394911	825909	1950	735	73.0	10-16-79	44.86	690.1
FR-62	395012	825857	1972	705	---	10-17-79	18.69	686.3
FR-64	395008	830042	1969	680	94.0	10-15-79	1.83	678.2
FR-65	395008	825734	1950	742	57.0	10-17-79	30.66	714.3
FR-70	394927	825958	1950	705	59.0	10-16-79	15.67	689.3
						09-05-90	16.66	688.3
						06-21-91	16.39	688.6
FR-71	395238	830005	1950	700	40.0	10-15-79	11.57	688.4
FR-72	395217	830023	1950	715	47.6	09-05-90	26.13	688.9
						06-21-91	31.35	683.6
FR-73	395132	830012	1960	730	---	10-15-79	43.66	686.3
TH-83	395027	825856	1977	707	64.0	10-17-79	15.39	691.6
						03-26-87	22.60	684.4 *
						09-05-90	26.19	680.8
						06-21-91	26.82	680.2
K-2	394924	830034	---	682	---	10-31-86	13.10	668.9 *
						03-26-87	14.90	667.1 *
K-5	394941	830020	---	703	---	10-31-86	33.40	669.6 *
						03-26-87	35.30	667.7 *
MP-3	395218	825953	---	706	---	10-31-86	15.10	690.9 *
						03-26-87	15.50	690.5 *

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MP-4	395202	830021	---	722	---	10-31-86	35.60	686.4 *
						03-26-87	33.80	688.2 *
MP-5	395016	830140	---	723	---	10-31-86	30.20	692.8 *
						03-26-87	29.00	694.0 *
MP-6	394941	825825	---	741	---	10-31-86	53.80	687.2 *
						03-26-87	54.90	686.1 *
MP-7	395011	825730	---	740	---	10-31-86	34.60	705.4 *
						03-26-87	34.30	705.7 *
247807	394956	830003	---	708	---	10-31-86	33.30	674.7 *
						03-26-87	36.00	672.0 *
643566	395001	830020	---	694	---	10-31-86	17.30	676.7 *
						03-26-87	19.20	674.8 *

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