In Cooperation with the City of Kalamazoo, City of Portage, Kalamazoo County Human Services Department, and Michigan Department of Environmental Quality

Simulation of the Ground-Water-Flow System in the Kalamazoo County Area, Michigan

Scientific Investigations Report 2004-5054

U.S. Department of the Interior
U.S. Geological Survey

Cover Photograph. Gourdneck Creek, Kalamazoo County, Michigan. (Photograph by Julie Jean, U.S. Geological Survey, 2001.)

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U.S. Department of the Interior
Gale A. Norton, Secretary

U.S. Geological Survey
Charles G. Groat, Director

U.S. Geological Survey, Reston, Virginia: 2004

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Conversion Factors, Abbreviations, and Vertical Datum

Inch/Pound to SI

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
acre	4,047	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
cubic foot (ft^3)	0.02832	cubic meter (m^3)
gallon (gal)	3.785	liter (L)
million gallons (Mgal)	3,785	cubic meter (m^3)
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
foot per day (ft/d)	0.3048	meter per day (m/d)
inch per hour (in/h)	0 .0254	meter per hour (m/h)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

$$°C=(°F-32)/1.8$$

Vertical coordinate information is referenced to the North Geodetic Vertical Datum of 1929 (NGVD 29).

Horizontal coordinate information is referenced to the Michigan State Plane Coordinate System, south zone, North American Datum of 1927 (MSP 27).

Altitude, as used in this report, refers to distance above the vertical datum.

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft^3/d)/ft^2]ft. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

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By Carol L. Luukkonen, Stephen P. Blumer, T.L. Weaver, and Julie Jean

Abstract

A ground-water-flow model was developed to investigate the ground-water resources of Kalamazoo County. Ground water is widely used as a source of water for drinking and industry in Kalamazoo County and the surrounding area. Additionally, lakes and streams are valued for their recreational and aesthetic uses. Stresses on the ground-water system, both natural and human-induced, have raised concerns about the long-term availability of ground water for people to use and for replenishment of lakes and streams. Potential changes in these stresses, including withdrawals and recharge, were simulated using a ground-water-flow model.

Simulations included steady-state conditions (in which stresses remained constant and changes in storage were not included) and transient conditions (in which stresses changed in seasonal and monthly time scales and storage within the system was included). Steady-state simulations were used to investigate the long-term effects on water levels and streamflow of a reduction in recharge or an increase in pumping to projected 2010 withdrawal rates, withdrawal and application of water for irrigation, and a reduction in recharge in urban areas caused by impervious surfaces. Transient simulations were used to investigate changes in withdrawals to match seasonal and monthly patterns under various recharge conditions, and the potential effects of the use of water for irrigation over the summer months.

With a reduction in recharge, simulated water levels declined over most of the model area in Kalamazoo County; with an increase in pumping, water levels declined primarily near pumping centers. Because withdrawals by wells intercept water that would have discharged possibly to a stream or lake, model simulations indicated that streamflow was reduced with increased withdrawals. With withdrawal and consumption of water for irrigation, simulated water levels declined. Assuming a reduction in recharge due to urbanization, water levels declined and flow to streams was reduced based on steady-state simulation results. Transient results indicated a reduction of water levels with the simulated use of water for irrigation over the summer months. Generally the transient simulation with recharge only in the winter provided the best fit to observed water levels collected during synoptic water-level measurements in some wells and to the trends observed in water levels for other wells.

Analysis of the regional hydrologic budgets provides an increased understanding of water movement within the ground-water-flow system in Kalamazoo County. Budgets for the steady-state simulations indicated that with reduced recharge, less water was available for streamflow and less water left the model area through the model boundaries. Similarly, with an increase in pumping rates, less water was available to enter streams and become streamflow. When recharge was assumed to remain constant and when it was allowed to vary throughout the year, the amount of water that entered storage was greater than that which left storage. However, when recharge was distributed through October–May only or when recharge rates were reduced from October to May, the amount of water that entered storage was less than that which left storage. Thus, on the basis of model simulations, with reduced recharge or increased withdrawals, water must come from storage, rivers, or from ground-flow-system boundaries to meet withdrawal demands.

Introduction

In Kalamazoo County, Michigan, and the surrounding area (fig. 1), ground water is widely used as a source of water for drinking and industry. Additionally, lakes and streams are valued for their recreational and aesthetic uses. Stresses on the ground-water system, both natural (changes in climate, for instance) and human-induced, raise concerns about the long-term availability of ground water for people to use and for replenishment of lakes and streams. Potential changes in these stresses were simulated using a ground-water-flow model.

Simulations with the ground-water-flow model included determination of the potential effects of changes in recharge and withdrawal rates on water levels and streamflow in the Kalamazoo County area and determination of regional ground-water budgets. The ground-water budget includes an evaluation of the amount of water entering the ground-water system (recharge), the amount of water in the system (storage), and the amount of water leaving the system (discharge). Sources of recharge include: areal recharge from precipitation that reaches the water table; recharge from losing streams, lakes, and wetlands; and injection wells. Sources of discharge include: discharge to streams, lakes, wetlands, and springs; evapotranspiration; and supply wells.

Changes in budget terms, such as reduced recharge during drought or ground-water pumpage, also need to be evaluated as they cause a temporary disequilibrium from long-term conditions when recharge equals discharge. For

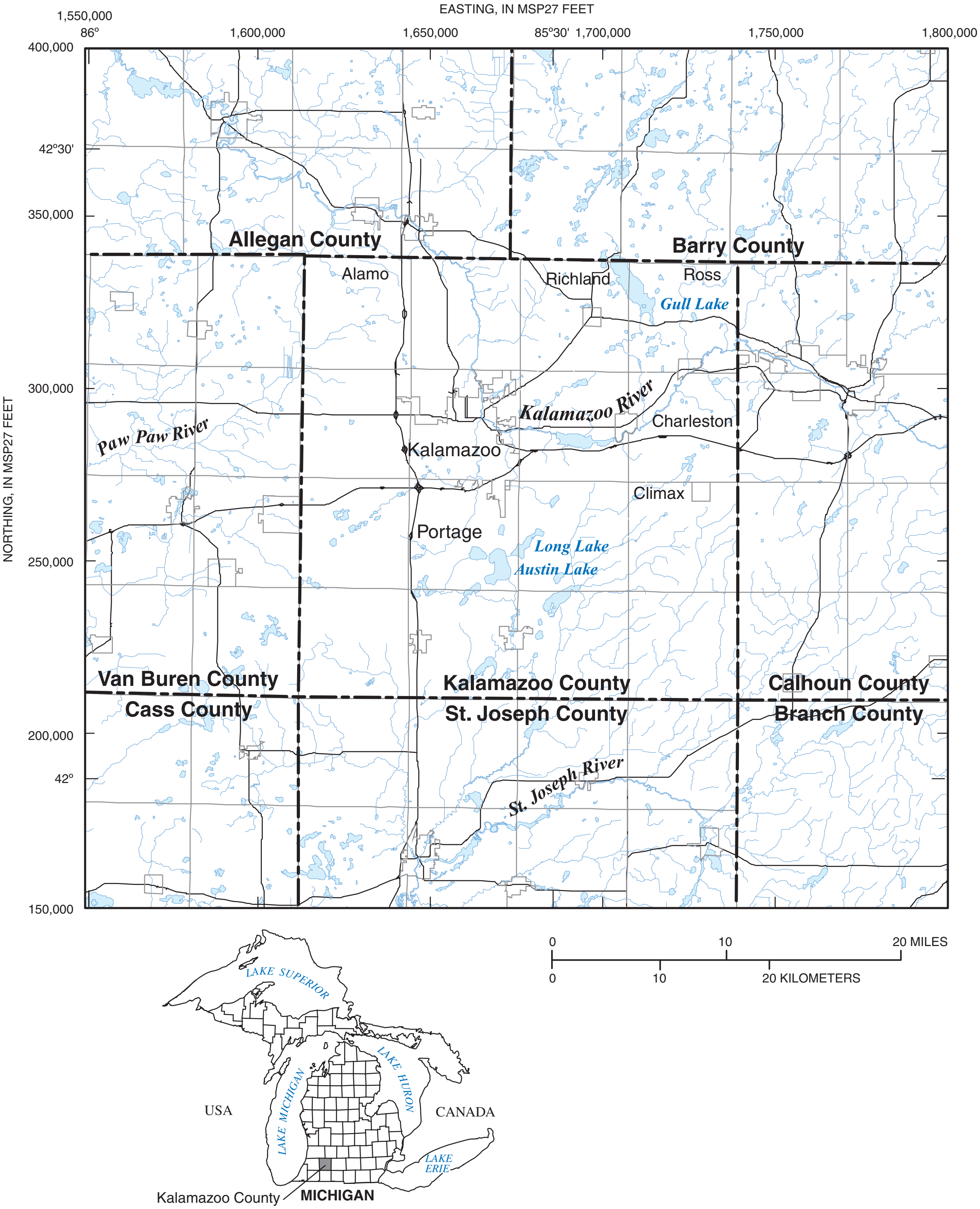


Figure 1. Location of Kalamazoo County, Michigan.

instance, pumpage can be supplied from storage in an aquifer, an increase in recharge, and a decrease in discharge. Any water withdrawn from the system will affect the system, perhaps in lower streamflow, loss of wetland areas, or reduced discharge to lakes. For most ground-water-flow systems, the change in storage in response to pumping is a transient phenomenon that occurs as the system readjusts to the pumping stress. The relative contributions of changes in storage, changes in recharge, and changes in discharge vary over time.

In 1997, the city of Kalamazoo, the city of Portage, and the Pharmacia and Upjohn Company began a study regarding the possible effects of ground-water withdrawals on surface-water and long-term ground-water resources. As a result of the 1997 study, the ground-water and surface-water monitoring networks in the county were expanded. In continuation of this effort, the U. S. Geological Survey, in cooperation with the city of Kalamazoo, the city of Portage, the Kalamazoo County Human Services Department, and the Michigan Department of Environmental Quality, investigated the effects of changes in stresses on the ground-water and surface-water systems in the Kalamazoo area. This study investigated the sources of water entering or leaving the ground-water-flow system and the potential effects of a reduction in recharge or an increase in withdrawals on ground-water levels and streamflow in the Kalamazoo County area.

Purpose and Scope

This report describes the results of numerical simulation of the ground-water-flow system in the Kalamazoo County area. The geologic and hydrologic settings are described and a conceptual model of the flow system is presented. Measurements of continuous and miscellaneous ground-water levels from wells, continuous and miscellaneous streamflow from streamflow-gaging stations and lake levels from lake-level gaging stations are presented. The investigation of the interaction between ground-water and surface water using seepage meters, piezometers, and temperature measurements is discussed. Numerical model construction, calibration, and simulation results are described. Effects of changes in stresses, such as recharge and pumpage, are evaluated. Simulations included investigation of changes in water levels and streamflows with different pumping conditions and different recharge rates under steady-state and transient conditions. Some additional simulations investigated potential effects of urbanization and irrigation on water levels. Hydrologic budgets for each simulation are compared to determine changes in the amounts of water within the various components of the system, including rivers and storage within aquifers; and amounts of water entering and leaving the ground-water-flow system.

Previous Studies

Previous studies have contributed to the understanding of the ground-water resources in the Kalamazoo County area. Deutsch and others (1960) described the ground-water hydrology and glacial geology of the Kalamazoo area. Allen and others (1972) studied the availability of water in Kalamazoo County. Rheaume (1990) investigated the geohydrology and water quality of Kalamazoo County. Numerous wellhead protection area and hydrogeologic survey reports were prepared by the city of Kalamazoo and Peerless Midwest (1996, 1999a, 1999b, 2002a, and 2002b), and the Ohio Drilling Company (1974, 1983a, 1983b, 1984, and 1987). Local hydrology and land use in Van Buren County are described by Cummings and others (1984). Geologic maps prepared by Monaghan and others (1983) and Puzio and others (1983) included information on surficial geology, thickness of glacial sediments, and bedrock geology.

Two reports describe the computer programs that the ground-water-flow model used in this study is based on. The computer code, MODFLOW-2000, along with the accompanying observation, sensitivity, and parameter estimation processes, is described by Harbaugh and others (2000) and Hill and others (2000).

Description of Study Area

Kalamazoo County is in southwestern Michigan and has an area of 576 mi^2 (fig. 1). Agriculture is the primary land-use category. The land surface is flat to rolling and ranges in elevation from 740 ft where the Kalamazoo River leaves the county, to 1,040 ft in the west-central part of the county. Mean precipitation in the study area averages 39 in./yr and is fairly evenly distributed throughout the year. September is the month of highest average precipitation (4.2 in.) and February is the month of lowest average precipitation (1.7 in.) (Midwestern Regional Climate Center, 2004a). Precipitation is slightly greater in the western upland areas than in the central and eastern parts of the county (Rheaume, 1990). This area averages about 58 in./yr of snowfall (Midwestern Regional Climate Center, 2004b). Mean daily average temperatures ranged from a low of 23.9°F in January to a high of 73.2°F in July (Midwestern Regional Climate Center, 2004c).

The three major drainage basins within the county include the Kalamazoo River Basin, which drains the northern two-thirds; the Paw Paw River Basin, which drains a small part of the western section; and the St. Joseph River Basin, which drains the remaining southern part of the county (fig. 2). These basins form the regional hydrologic boundaries of the flow system in the Kalamazoo County area. The county has 356 lakes and ponds that range in size from less than 1 acre to 2,050 acres. The largest lake is Gull Lake in the northeastern part of the county.

Methods of Investigation

Data were collected during this study to improve understanding of the water resources of the area and of the interaction between the ground-water and surface-water systems.

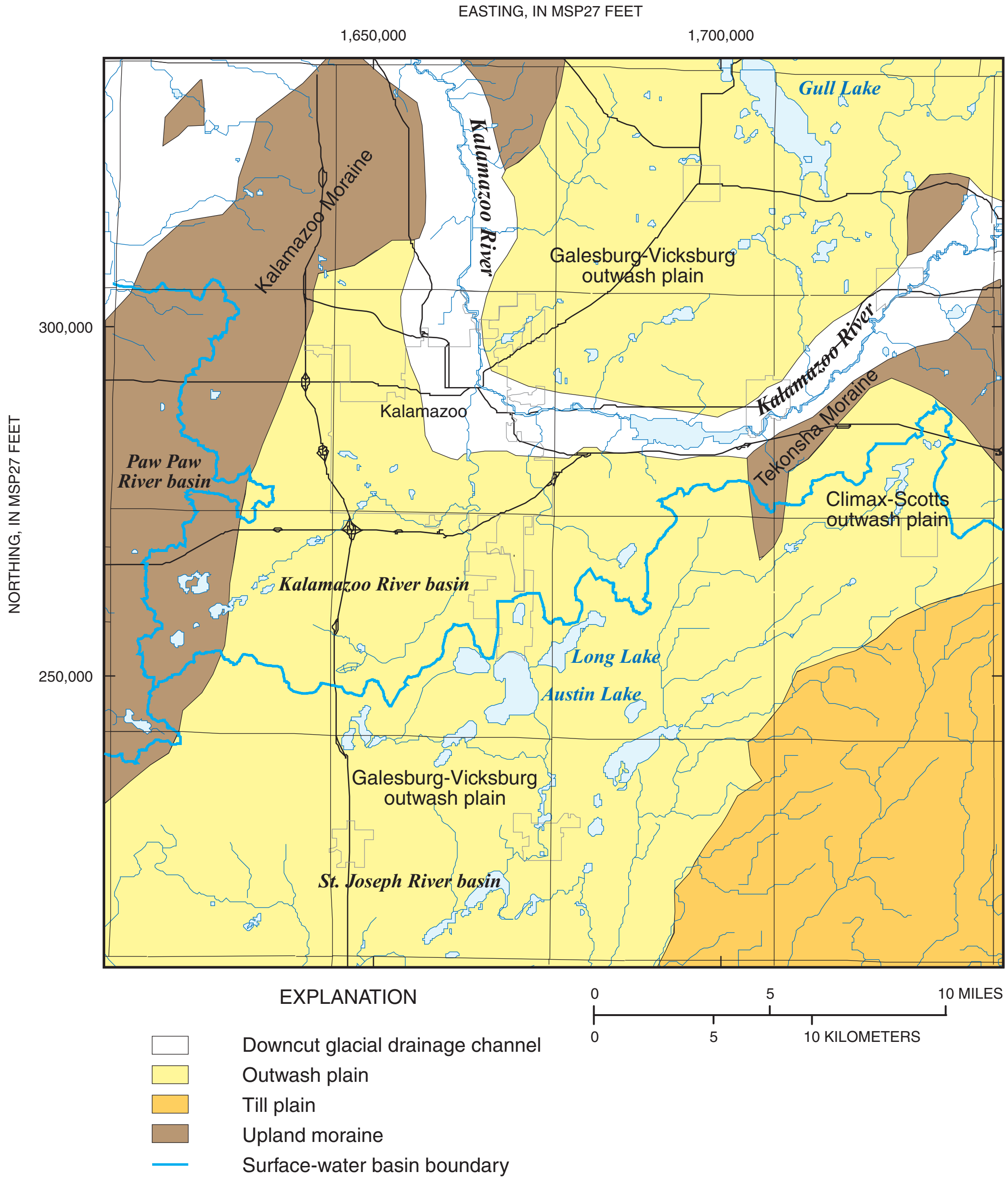


Figure 2. Surface-water basins and generalized spatial distribution of surficial glacial materials, Kalamazoo County, Michigan (Modifed from Rheaume, 1990).

Data collection included miscellaneous and continuous ground-water levels, miscellaneous and continuous stream discharge measurements, lake levels, and compilation of historic data. Data also were collected and compiled to provide measurements of water levels and streamflow to compare with model-simulated values. Comparison of observed and simulated values helps to ensure that the numerical model simulates the actual ground-water-flow system as closely as possible.

Ground-water/surface-water interactions were investigated using seepage meters, piezometers, and temperature sensors. Observations using the seepage meters yield measurements of flux between the stream and the ground-water-flow system. These flux measurements can be compared to model-simulated values as an additional model-calibration check. Sites where seepage meters were paired with piezometers yield estimates of the hydraulic conductivity of the lake or streambed materials; these estimates of riverbed hydraulic conductivities also are needed for model simulation. Data from piezometers placed in areas of different vegetative characteristics help to determine whether evapotranspiration is a important variable that should be included in model development. The following sections describe data collection methods.

Water Levels and Stream Discharge

Data collection during this study included ground-water levels at 42 miscellaneous and 18 continuous monitoring wells, stream discharge at 22 miscellaneous and 8 continuous streamflow- gaging stations, and water levels at 12 miscellaneous lake sites and at 4 continuous lake-level gaging stations. Ground-water and surface-water sites included sites being monitored at the start of the study (2000) or those monitored during previous studies, to provide consistency, to allow examination of any differences over time, and to indicate the possible range of variation at each site. Miscellaneous data collection was done at approximately quarterly intervals during synoptic surveys from August 2001 to September 2002. These measurements help to determine whether there are any seasonal or long-term differences in water levels and to provide calibration points for the numerical model. The date for each synoptic survey was chosen such that no appreciable precipitation occurred within the study area for 3–5 days prior to the synoptic date and there was a prediction of no appreciable precipitation on the synoptic dates. All continuous and miscellaneous sites were measured or flow determined from rating curves in August 2001. Ground-water levels and streamflow at a reduced number of sites were measured during the remaining synoptic surveys. Fewer sites were measured after the August synoptic survey because it was preferable to have all sites measured during the same day and relations between flows at some sites had been determined during previous studies, making additional measurements unnecessary.

Seepage Measurements

Seepage measurements were conducted in Long and Austin Lakes and selected streams in Kalamazoo County and surrounding areas at seven different times between July 2000 and September 2002, although not all sites were measured each time. Long Lake is in south-central Kalamazoo County and is approximately 2-mi long and from 0.25- to 0.75-mi wide. Long Lake covers 575 acres and has a maximum depth of 60 ft. Austin Lake is west of Long Lake and is approximately 2-mi long and 1-mi wide. Austin Lake covers 1,090 acres and has a maximum depth of 11 ft, although the average depth is only 4 ft. Sediment ranges in size from mud and silt to coarse-grained sand in Long Lake and, generally, from mud and silt to fine-grained sand in Austin Lake. Seepage measurements were made in August 2001 at selected streams within Kalamazoo and Van Buren Counties. Stream sediments ranged in size from mud and silt to large gravel.

Seepage, and thus ground-water/surface-water interactions, was monitored using mini-piezometers and seepage meters modeled after instruments described by Rosenberry (1990) and Winter and others (1988). The mini-piezometers measure hydraulic potential, or hydraulic-head gradient, between a lake or stream and the underlying material. Using Darcy's Law, the seepage rate can be calculated from hydraulic potential if the other factors are known. Seepage meters directly measure the volume and direction of seepage between the lake and lake bottom.

The piezometer used in this study was constructed from a 5-ft long, 0.5-in. galvanized steel pipe. A dowel rod with electrical tape wound around the end was inserted into each pipe during installation to prevent sediment from entering the pipe. The pipe was tapped gently into the lake or streambed near a seepage meter until approximately 6 in. of pipe was inserted. The dowel was removed and the pipe was allowed to equilibrate overnight. In cases where the pipe opening was beneath the water level, a 0.5-in. plastic tube was connected to the opening and tied above water. At each piezometer, distances from the top of pipe to the water surface both inside and outside the pipe were collected in addition to measurements from the top of the pipe to the lakebed and the total length of pipe.

Both small and large seepage meters were used in this study. The smaller seepage meter was constructed from a 2.5 L dishpan (fig. 3). The edges of the tub were cut off to minimize sediment disturbance. A 1-in diameter hole was cut in one corner of the tub for a standard 1-in brass fitting secured with a 1-in steel lock nut. The larger seepage meter, or barrel meter, was modeled after the half-barrel seepage meter described by Rosenberry (1990). The barrel meter used in this study was constructed from a 55-gallon drum that was cut cross sectionally into three sections of equal height. A 1-in brass fitting

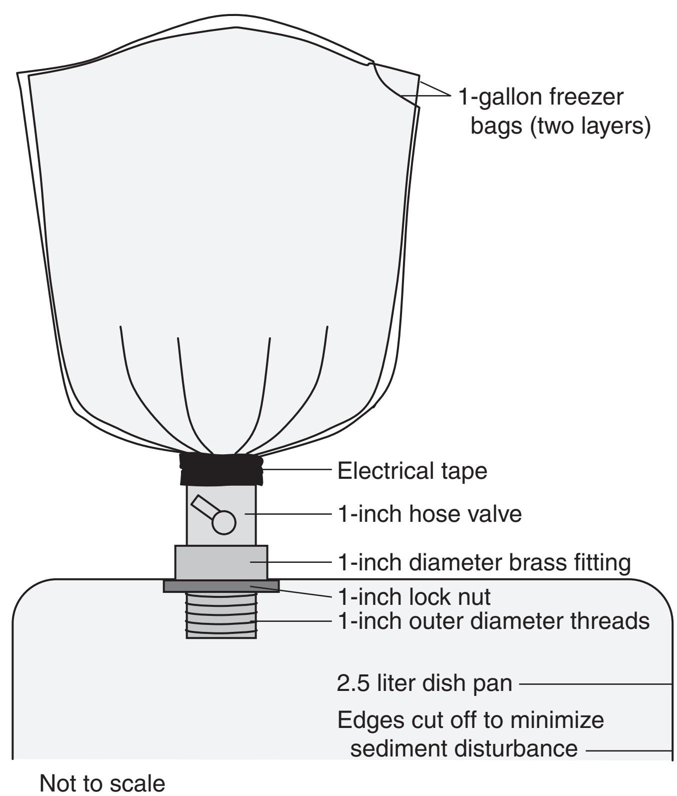


Figure 3. Example of a small seepage meter.

was installed into the closed end of each of the two drum end sections and near the edge of the remaining open ring section. All meters were worked gently into the stream or lakebed sediments trying to minimize sediment disturbance and so that a seal was formed. The ring barrel meter was placed in shallow water (less than 1 ft deep) with the open top of the meter exposed to air. The tub and closed barrel meter were installed at depths of 1 to 4 ft or greater so that the meters were completely submerged. The lake meters were placed in lines perpendicular to the shoreline with meters at varying distances from shore to determine whether there was any relation between seepage rate and distance from shore. The meters were allowed to equilibrate overnight. After equilibration, plastic bags with a known initial volume of water and all possible air removed were attached to the outside end of the brass fitting. These bags were constructed of two layers of 1 gal freezer bags attached to a 1-in. garden hose valve with electrical tape so that an airtight seal was formed. The corners of the outside bag were clipped to minimize trapped air and permit the bag to remain completely submerged when attached to the meter. The initial time when the bag was attached to the meter and the valve opened was recorded. The meter and bag were left undisturbed approximately 1 hour, after which the valve was closed and the time recorded. The final volume of water in the bag also was recorded. From one to three sets of seepage measurements were collected from each meter. At some sites, meters were paired, or placed next to each other at the same distance from shore, in order to check meter accuracy and to determine whether meter size had any effect on seepage rate.

Seepage rate was determined using the following equation:

$$\frac{\text{Final volume of water} - \text{initial volume of water}}{\text{Final time} - \text{initial time}} \tag{1}$$

The seepage rate was divided by the area of the stream or lakebed isolated by the meter to determine specific flux. Negative values of flux indicate water movement out of the stream or lake into the ground-water system; whereas positive values of flux indicate water movement into the stream or lake from the ground-water system. For those meters in which mini-piezometers were placed nearby, the hydraulic conductivity of the stream or lakebed sediments can be calculated using Darcy's Law as

$$Q = KA\ (dh/dl) \tag{2}$$

where

Q = volume of ground-water flow or seepage,

K = hydraulic conductivity of the material through which flow occurs,

A = cross-sectional area through which flow occurs,

and

dh/dl = hydraulic-head gradient.

The distribution of seepage rates can be highly dynamic, temporally and spatially (Rosenberry, 1990). Transient changes in the direction of seepage between ground water and an adjacent lake have been observed (Meyboom, 1966; Williams, 1968; Freeze and Banner, 1970; Jaquet, 1976; Zager, 1981; Price and Hendrie, 1983; and Rosenberry, 1985) and simulated by Winter (1983). Winter (1983) simulated a hypothetical cross-sectional setting common to many lakes. The model-simulation results show that after infiltration of snowmelt in the spring, the seepage direction reversed, and rather than lake water seeping into the ground-water-flow system, ground water began to seep into the lake. Transient flow reversals also could occur in response to large rainfalls as observed by Rosenberry (1985). Some studies have shown that most seepage occurs near the shoreline of a lake. McBride and Pfannkuch (1975) used digital models to illustrate that seepage will decrease exponentially with distance from shore. Field evidence of this seepage distribution was provided by Lee (1977), Erickson (1981), and Rosenberry (1985). However, these observations of an exponential decrease of seepage with distance from shore assume a uniform, homogeneously porous medium, whereas in many areas the geologic conditions are not homogenous (Rosenberry, 1990).

Well Measurements

Two sites were selected for monitoring and comparison of water levels; one site was in an area of sand and gravel near Long Lake and the other site was in an area surrounded by vegetation near Austin Lake. Site selection was based on the difference in the amount of vegetation between the two sites. With the presence of vegetation, water levels are expected to decrease during the day while the plants are consuming water and increase during the night while the plants are inactive (diurnal). In the absence of vegetation, water levels are not expected to vary diurnally. Continuous water-level measurements were collected to determine whether evapotranspiration was observed to be important and would need to be included in the ground-water-flow model.

Three wells initially were installed near Long Lake and two wells were installed near Austin Lake. Wells were 2-in. diameter, constructed of PVC, with 1-ft long, 10-slot screens. The wells were installed in hand-augered holes below the level of the water table and backfilled with native material. The well installed next to the shoreline at Long Lake was backfilled with bentonite in case the lake level rose to inundate the well and to prevent leakage down the well casing. Water levels were measured by transducers connected to a data logger. Water levels were monitored in wells at Long Lake from August 22, 2000, to January 11, 2002, and at Austin Lake from April 10, 2001, to February 1, 2002, although because of recording equipment problems, this record is not continuous at both sites. Also, over the data-collection period, water levels in Long Lake increased and eventually the two wells were submerged. These wells were no longer used for data collection.

Temperature Monitoring

From July 24, 2001, to November 24, 2001, temperature sensors were installed in the five wells located near Long and Austin Lakes. The temperature loggers consisted of a thermistor wired to a single channel data logger in a waterproof enclosure. Temperature measurements were recorded every 15 minutes. Additionally at each site, one temperature logger was placed within the lake near the piezometers to measure the surface-water temperature. The logger in Long Lake was at an initial approximate depth of 3 ft, whereas the logger in Austin Lake was at an approximate depth of 0.5 ft. Water levels increased in both lakes during the data-collection period (July 24, 2001, to November 24, 2001), although less in Austin Lake than in Long Lake. Although not performed during this study, with further analysis, these temperature data could be used to obtain an estimate of water movement between the surface water and ground-water-flow systems and be used to provide a further check on model calibration.

Hydrogeologic Setting

In the Kalamazoo County area, unconsolidated glacial deposits overlie bedrock units. The primary geologic and hydrologic characteristics of these units that affect regional ground-water flow are described. In addition, the water-withdrawal characteristics of production wells in the study area and the conceptual model of the regional ground-water-flow system are described below.

Geologic Setting

Kalamazoo County is on the southwestern part of the Michigan Basin, which is a gently sloping structural depression that extends beyond the political boundaries of the state into surrounding States and Ontario. Consolidated rock units underlying Kalamazoo County are overlain by a combination of Pleistocene and Holocene glacial, glaciofluvial, fluvial, and eolian deposits. An unconformity that represents several hundred million years of deposition and erosion separates Mississippian bedrock units from Pleistocene and Holocene deposits. Buried valleys eroded into the bedrock units indicate that drainage for various preglacial river valleys in Kalamazoo County was northwest trending (Deutsch and others, 1960). There are no known bedrock outcrops in the county (Straw, 1978). The landscape of Kalamazoo County includes broad, gently sloping outwash plains and sandurs, terraced valley-train outwash systems, a drumlinized till plain, a glacial lake plain, and smaller features such as kettles, kames, and till deposits (Monaghan and Larson, 1984). Glacial deposits range in thickness from less than 100 ft in the southeastern part of the county to more than 600 ft in the western part of the county (Forstat, 1983). Two major morainic systems are the prominent geomorphologic features of the county. The Kalamazoo and Tekonsha Moraines (fig. 2) and their respective features and landforms trend northwest across the county and represent a complex relation between the Lake Michigan and Saginaw glacial lobes during the late Pleistocene.

Description of Bedrock Units

Two Mississippian bedrock formations underlie Pleistocene and Holocene deposits throughout Kalamazoo County. The Coldwater Shale Formation subcrops under glacial and glaciofluvial deposits throughout the county, except the northeast corner, which is underlain by the Marshall Sandstone formation (Milstein, 1987). The Coldwater Shale thickens from west to east in the State (Cohee, 1979), and is laterally extensive, extending west under Lake Michigan and east under Lake Huron. The Coldwater Shale consists primarily of shale, although siltstone, sandstone, limestone, and dolomite comprise part of the unit in some areas of the Michigan Basin (Westjohn and Weaver, 1996; Cohee, 1979; Hale, 1941; Monnett, 1948). Numerous wells have been drilled in Kalamazoo County for purposes of obtaining water, oil, and gas, some with depths greater than 4,000 ft. Driller's logs show the thickness of the Coldwater Shale in Kalamazoo County ranges from 522 ft in Alamo Township (Martin, 1958) to 626 ft in Charlestown Township, and up to 862 ft in Richland Township. Because of its thickness and areal extent, the Coldwater Shale effectively forms an impermeable barrier between overlying unconsolidated deposits and underlying coarser-grained bedrock units, limiting available freshwater to unconsolidated deposits in the county.

The Marshall Sandstone underlies unconsolidated deposits in part of Charleston and Ross Townships in the northeast part of the county and overlies the Coldwater Shale in these locations. The Marshall Sandstone has been eroded away above the Coldwater Shale in the remainder of the county. The Marshall Sandstone unit consists mostly of sandstone, but is interbedded with limestone, dolomite, siltstone, and shale in some parts of the Michigan Basin. The Marshall Sandstone forms an important aquifer in parts of the Michigan Basin, including Calhoun and Jackson Counties and the northeastern corner of Kalamazoo County, where it is overlain by permeable glacial deposits (Westjohn and Weaver, 1996).

Description of Glacial Deposits

Glacial deposition in Kalamazoo County is complex. Many, if not most, of the past studies were based on the geomorphologic and stratigraphic work of Leverett and Taylor (1915). It is now recognized by most researchers that individual depositional units within broader sediment packages largely are discontinuous. No single study of the magnitude necessary to map the glacial sediments in detail in the county is known. The purpose of the following discussion is to summarize relevant results of some of the studies.

Present-day Kalamazoo County was at the intersection of two of the Great Lakes glacial lobes during late-Wisconsinan glaciation, with the Lake Michigan lobe from the west converging with the Saginaw lobe from the north (Martin, 1958). The late-Wisconsinan ice sheet reached its southernmost position in southern Illinois, Indiana, and Ohio around 20,000 years B.P. During retreat of the ice sheet over the period from about 20,000 to 11,000 years B.P., it became lobate. It was during this period that present-day southwestern Michigan, including Kalamazoo County, landforms were formed. Dominant glacial landforms in the county are areally extensive and are typically either outwash or morainal. Pleistocene depositional history as it is currently understood will be presented in this section along with a lithologic description of significant units within the depositional package.

In late-Wisconsinan times, following retreat of the ice sheet from its position in southern Illinois, Indiana, and Ohio, the Sturgis Moraine was formed during a re-advance. Only a small piece of the moraine is present in the southwest part of the county. The till sheet in the southeastern part of the county and the drumlin fields south of Climax may have formed during this time. The ice margin retreated for a period of time prior to either stabilizing or readvancing to the position of the Tekonsha Moraine, which is just south of the Kalamazoo River Valley from Galesburg east toward Calhoun County. The till sheet deposited behind the moraine is known as the Ganges Till and the Climax-Scotts outwash plain formed south of the moraine. Ice again retreated, and then the Lake Michigan lobe readvanced to form the Outer Kalamazoo Moraine. The till sheet deposited behind the moraine is known as the Saugatuck Till. The Galesburg-Vicksburg outwash plain formed east of the moraine, sloping gently eastward and burying part of the Tekonsha Moraine and the Climax-Scotts outwash plain. The ice retreated and the Lake Michigan Lobe readvanced a final time, forming the Inner Kalamazoo Moraine. Lacustrine sand between the Ganges and Saugatuck Tills in the Glenn Shores section indicates that the ice front of the Lake Michigan Lobe withdrew from the Tekonsha Moraine into the Lake Michigan Basin prior to readvancing to the Sturgis and Kalamazoo morainal positions (Monaghan and others, 1986). The distance of the withdrawal and readvance is at least 37 mi. Relatively thick layers of sand-and-gravel deposits associated with the Lake Michigan Lobe are much greater in volume than those associated with the Saginaw Lobe. It seems likely that this results because of the availability of greater quantities of coarser-grained material to the west and northwest of Kalamazoo County rather than north (Larson, G.J., Michigan State University, written commun., 2000).

Tunnel valleys are formed when subglacial water, which is often under immense hydrostatic pressure, seeks an outlet (Sugden and John, 1976). The results are channels that are frequently linear with steep sides often lacking downwasting features typical of channels formed subaerially. Some tunnel valleys preserve evidence that water flowed upgradient as a result of the hydrostatic pressure that might be found under the lobe of a continental glacier. The distinguishing feature of ice-directed tunnel valleys is their tendency to be aligned in a direction parallel to ice movement (Sugden and John, 1976). Kehew and others (1999) studied tunnel valleys in southwestern Michigan and northeastern Indiana and indicate that the tunnel valleys provide evidence about the relative timing of glacial advance, stagnation, and retreat in the area from around 20,000 years B.P. until the glaciers retreated from the area for the last time from around 11,000 to 12,000 years B.P. They believe that tunnel valleys formed beneath the Saginaw Lobe were filled with ice and debris when overridden by the Lake Michigan Lobe, which was either advancing while the Saginaw Lobe was stagnant or was advancing preferentially to the Saginaw Lobe. The result was tunnel valleys cut at high angles to the initial valleys. Kehew and others (1999) indicate that the Lake Michigan Lobe advanced southeastward over terrain previously altered by the Saginaw Lobe when the Lake Michigan and Lake Erie Lobes expanded much more strongly than the Saginaw Lobe or when the Lake Michigan Lobe expanded while the Saginaw Lobe was retreating. These Saginaw Lobe tunnel valleys may be more important relative to aquifer occurrence than previously thought (Kehew, Western Michigan University, written commun., 2003). Monaghan and others (1983) produced a series of geologic maps of Kalamazoo County that includes a surficial geology map that identifies the major geomorphologic features and includes drift thickness.

For this study, present-day landforms are termed till plain, upland moraine, outwash plains, and downcut glacial drainage channel as discussed in Rheaume (1990) (fig. 2). Monaghan and others (1983) describe the till as varying from mostly clay to primarily sand. The Tekonsha Moraine consists of massive to poorly bedded, coarse sand to sandy-clay till, and at places, as massive to poorly bedded sand and gravel containing boulders and cobbles; the Kalamazoo Moraine consists of sandy to very sandy till and massive to poorly bedded cobbly sand (Monaghan and others, 1983). The outwash plains are referred to as the Climax-Scotts outwash plain and the Galesburg-Vicksburg outwash plain and consist of medium- to very-coarse sand and gravel. The downcut glacial drainage channel was formed in the area of the Kalamazoo River and consists of medium- to very-coarse sand and gravel with some layers of clayey silt (Monaghan and others, 1983).

Hydrologic Setting

The hydrologic setting can be characterized in terms of aquifers and confining units, ground-water levels, streamflow, surface-water/ground-water interaction, and recharge. These elements of the hydrologic conditions are discussed in the following sections.

Aquifers and Confining Units

Glacial deposits, consisting largely of sands and gravels, are the source of water for most residents and businesses in Kalamazoo County. These deposits vary regionally in thickness and permeability. Aquifers underlying the outwash plains and the downcut glacial drainage channels, which together cover about two-thirds of the county, are the most productive (Rheaume, 1990). Allen and others (1972) identified an upper unconfined aquifer throughout most of the county and one to two lower semiconfined aquifers over about one-third of the county. At many locations, the hydraulic connection between aquifers is good enough that, under pumping conditions, water will move readily between aquifers.

For this study, the glacial deposits have been divided into one to three aquifers separated by confining units (fig. 4). Aquifer thicknesses and locations are based on the description and locations in Allen and others (1972) and the descriptions in local water, oil, and gas well logs. The upper unconfined aquifer ranges in thickness from 0 to 120 ft, the intermediate aquifer ranges in thickness from 0 to 100 ft; and the lower aquifer ranges in thickness from 0 to 120 ft. The upper unconfined aquifer is thickest along the western part of the study area. The intermediate aquifer is thickest in the central part of the study area south of the Kalamazoo River. The lower aquifer is thickest to the south and west of the city of Kalamazoo. The upper confining unit ranges in thickness from 0 to 120 ft and the lower confining unit ranges in thickness from 0 to 170 ft. The upper confining unit is thickest in the north central and west central parts of the study area. The lower confining unit is thickest in the west central and northeastern part of the study area. Thus, the glacial materials in the Kalamazoo County area can be combined into five units that control ground-water flow, although not all units are present throughout the study area.

Initial estimates of the hydraulic properties of the glacial aquifers and confining units are available from previous studies (Deutsch and others, 1960; Allen and others, 1972; The Ohio Drilling Company, 1974, 1983a, 1983b, 1984, and 1987;

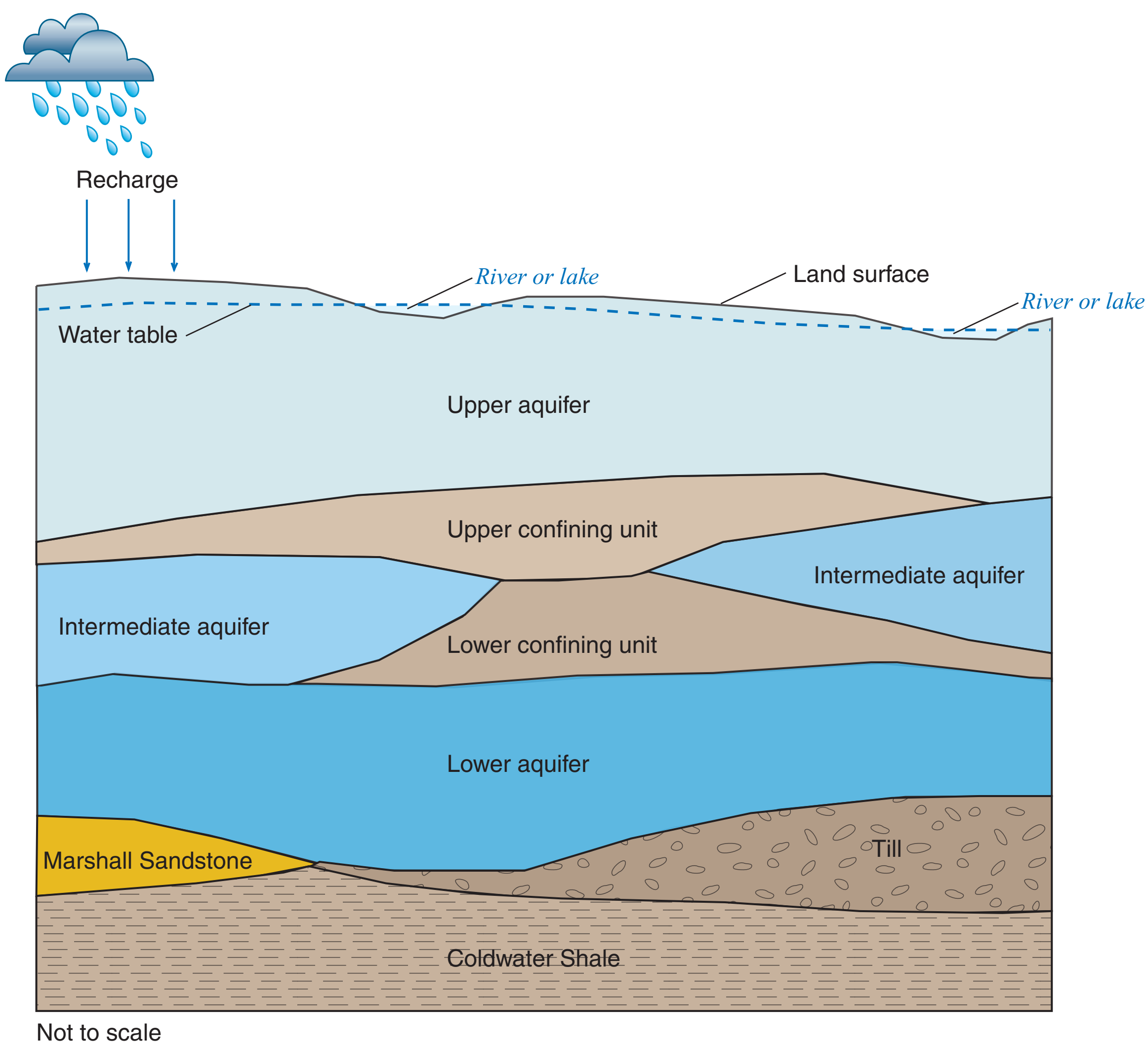


Figure 4. Generalized geologic section depicting hydrologic units in the Kalamazoo County area, Michigan.

Jones, Henry & Williams Consulting Engineers, 1961; Jones, 1992; City of Kalamazoo Department of Public Services, 1995; and City of Kalamazoo Department of Public Services and Peerless Midwest Company, Inc., 1999a and 1999b) (table 1, back of report). Aquifer test and recovery data collected during or prior to this study have been analyzed (Bob Snell, Prein and Newhof, written communication, 2001). Representative values estimated by Freeze and Cherry (1979) were used for the geologic materials in the area. In most cases estimates of hydraulic properties were reported for the upper unconfined aquifer or for the intermediate semiconfined aquifer; however, in some areas there are actually two lower aquifers. For the purposes of this study, the hydraulic properties of the intermediate and lower aquifers were initially assumed to be the same.

Over most of the county, the Coldwater Shale underlies the glacial deposits. Water from this shale unit is highly mineralized and yields are small; therefore, the Coldwater Shale is not generally used for water supply. The Marshall Formation underlies the glacial deposits in the northeastern part of the county. Sufficient quantities of water can be withdrawn from this formation for domestic supply. The lowermost unit that controls ground-water flow within the County consists of the Marshall Formation where it is present in the northeastern part of the study area (fig. 4). In the rest of the study area the lowermost unit consists of clay or other poorly permeable materials where they are present overlying the Coldwater Shale.

Ground-Water Levels

A preliminary understanding of ground-water flow directions can be based on surface-water elevations taken from topographic maps of the area, ground-water levels measured during this study, and previous reports covering the study area. Ground-water levels and locations are shown in table 2 (back of report) and figure 5. Ground-water levels during this study generally were highest in winter and late spring and lowest in late summer although some wells deviated from this pattern. Wells located within the downcut glacial drainage channel geologic setting generally had the shallowest depth to water, whereas wells located within the upland moraine geologic setting generally had the greatest depth to water. An initial estimate of water levels in the upper unconfined aquifer was based on water-level observations in area wells and points where contour lines crossed lakes and rivers on topographic maps (fig. 6). Regional ground-water flow is towards the major surface-water features in the area, including the Kalamazoo River, the St. Joseph River, and the Paw Paw River. Local ground-water flow is towards pumping centers.

During part or all of this study, continuous lake levels were recorded on four lakes, including Austin Lake, Long Lake, Hampton Lake, and Asylum Lake, in Kalamazoo County (table 3 and fig. 7). Austin, Long, and Hampton Lake levels generally were lowest in early 2000. Asylum Lake levels were low in late 2000 and mid 2001. All four lakes had the highest levels in early 2002. An additional 12 lake levels were referenced to local reference points during the first synoptic survey. However, because of the difficulty in obtaining nearby benchmarks, these miscellaneous lake levels could not be referenced to NGVD 29.

Water levels collected at Long Lake indicate water movement into the lake during the data-collection period in 2000 and early 2001; however, the rest of the data from 2001 indicates water movement out of the lake. Water levels collected at Austin Lake indicate water movement into the lake from mid-2001 to early 2002. Water levels in the wells near Long Lake generally showed no diurnal trends indicative of evapotranspiration. Water levels in the wells near Austin Lake generally indicated a diurnal trend; however, the magnitude of the change was less than 0.15 ft, which may be explained by problems with the data-collection equipment. Based on this information, as well as that this study is a regional evaluation of water resources and that recharge will be applied at the water table in model simulations, evapotranspiration is assumed to be negligible at the model scale.

Streamflow and Surface-Water Levels

Streamflow in Michigan generally is greatest in spring and early summer and lowest in late summer and winter. It increases slightly after the first autumn frosts when evaporation decreases and vegetation dies. Hydrographs for three streams within the county with the longest periods of record are shown in figure 8. Streamflow measured during this study is shown in table 4 and measurement locations are shown in figure 9. Flow at continuous sites was assigned to be the mean daily discharge for the dates of the synoptic surveys. For most stations, streamflow was greatest in December 2001 or March 2002; streamflow was lowest in August 2001 or September 2002. Three stations had the greatest streamflow in June 2002. In Kalamazoo County, the surface-water and ground-water systems are closely connected. During drier periods, flow of streams is almost entirely sustained by ground-water inflow.

Surface-Water/Ground-Water Interaction

Seepage measurements were collected six times at various sites around Long Lake from July 2000 until September 2002. Seepage rates, determined using equation (1), generally were lowest in October 2001 and highest in July 2000. Seepage rates ranged from -2.4 in./day to 1.8 in./day in October 2001 and from -10.8 in./day to 13.2 in./day in July 2000. However, the 0.5 in. of rain that fell in the morning just before data collection may have affected the July 2000 measurements. Comparing all sites on Long Lake, a greater range of seepage rates occurred during July and August. Generally, more negative seepage rates occurred in June 2002; thus, the seepage data indicate water movement primarily out of the lake into the ground-water system. In April 2002, higher positive and negative rates were observed close to shore; smaller rates were observed further from shore. In June 2002, negative rates were

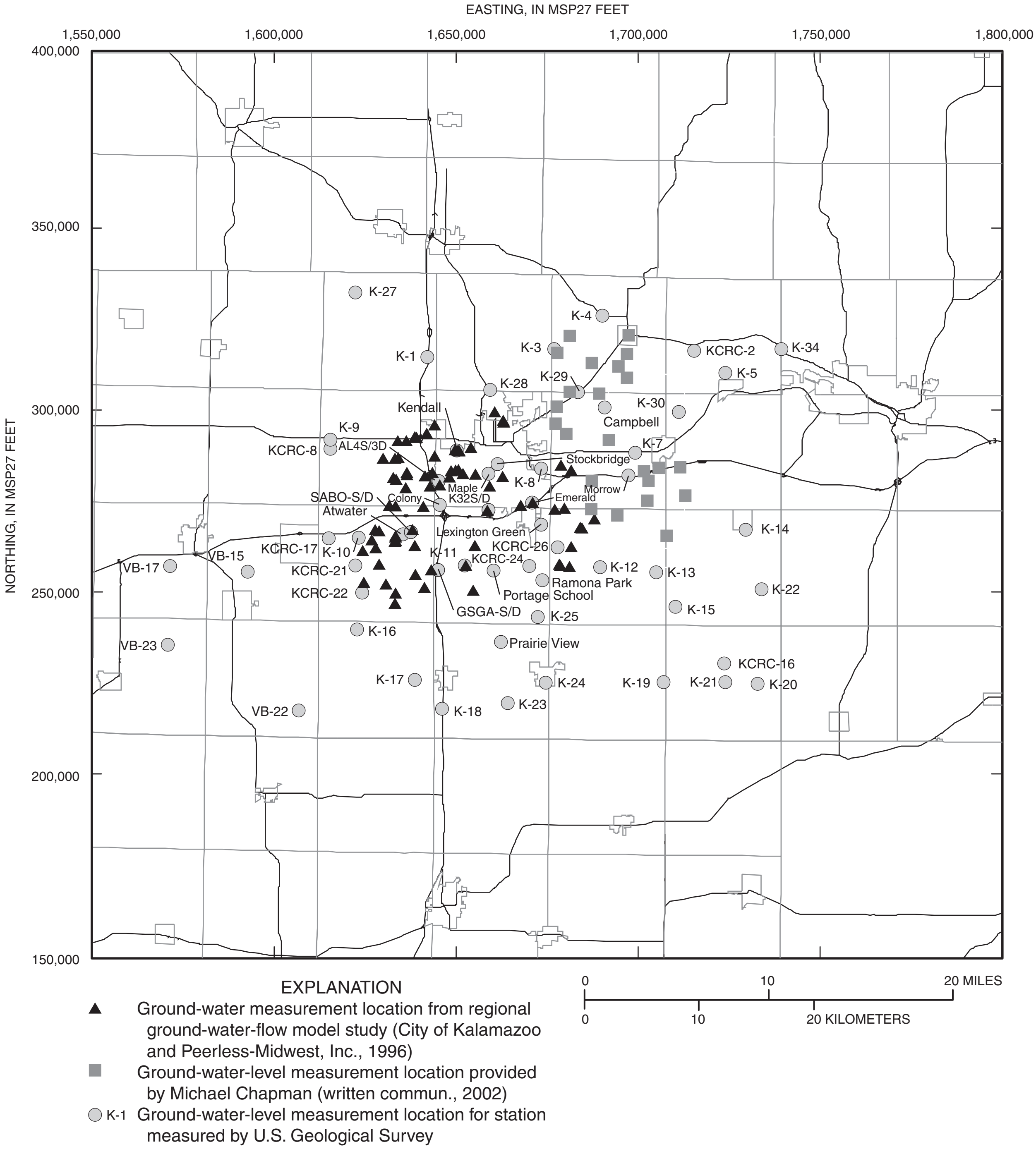
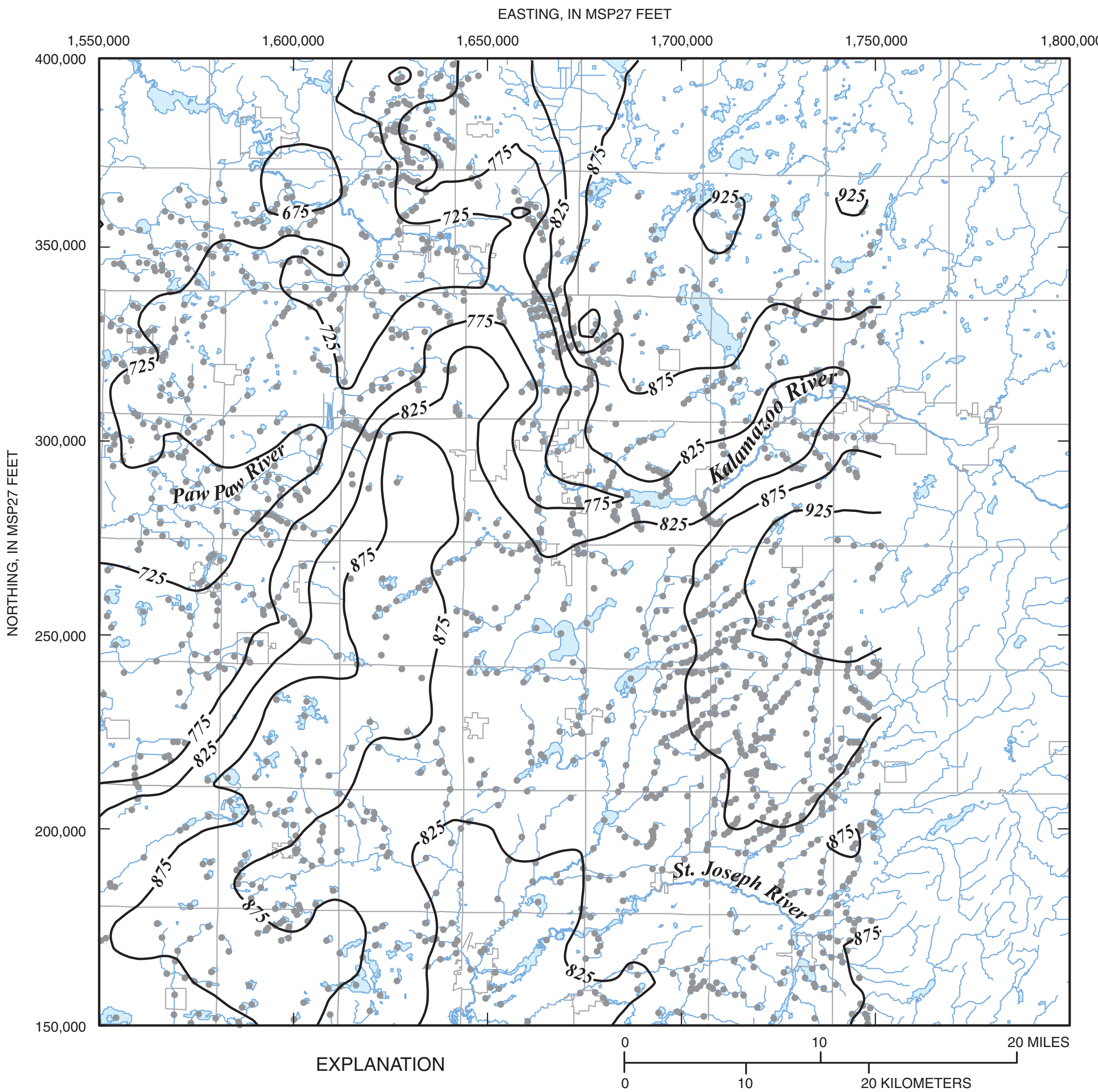


Figure 5. Ground-water-level measurement locations in the Kalamazoo County area, Michigan.



EXPLANATION

• Ground-water level observation location

— *825* — Water-level contour—Contour interval 50 feet. Datum is NGVD 29

Figure 6. Initial estimate of ground-water levels, Kalamazoo County area, Michigan.

Table 3. Lake water levels, during synoptic surveys, Kalamazoo, Michigan.

[Measurements referenced to feet above NGVD 29; --, measurement not available]

Local name	August 2001	December 2001	March 2002	June 2002	September 2002
Long Lake near Kalamazoo	853.94	854.46	856.15	856.1	855.39
Austin Lake near Kalamazoo	--	--	856.23	856.19	855.92
Hampton Lake near Portage	857.19	857.33	857.25	857.34	857.33
Asylum Lake near Kalamazoo	868.16	868.97	868.9	868.41	868.43

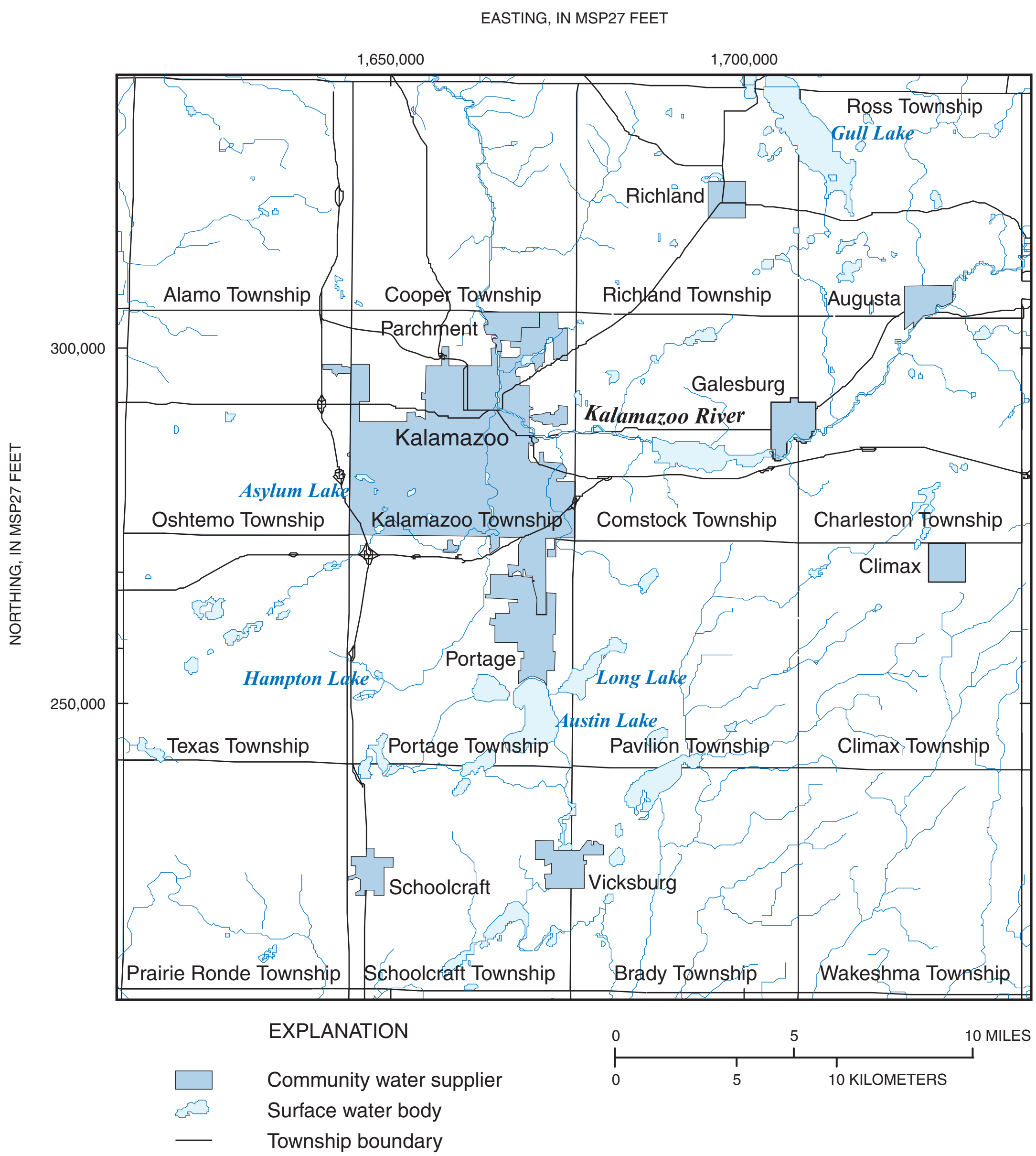


Figure 7. Location of community water suppliers and selected lakes in Kalamazoo County, Michigan.

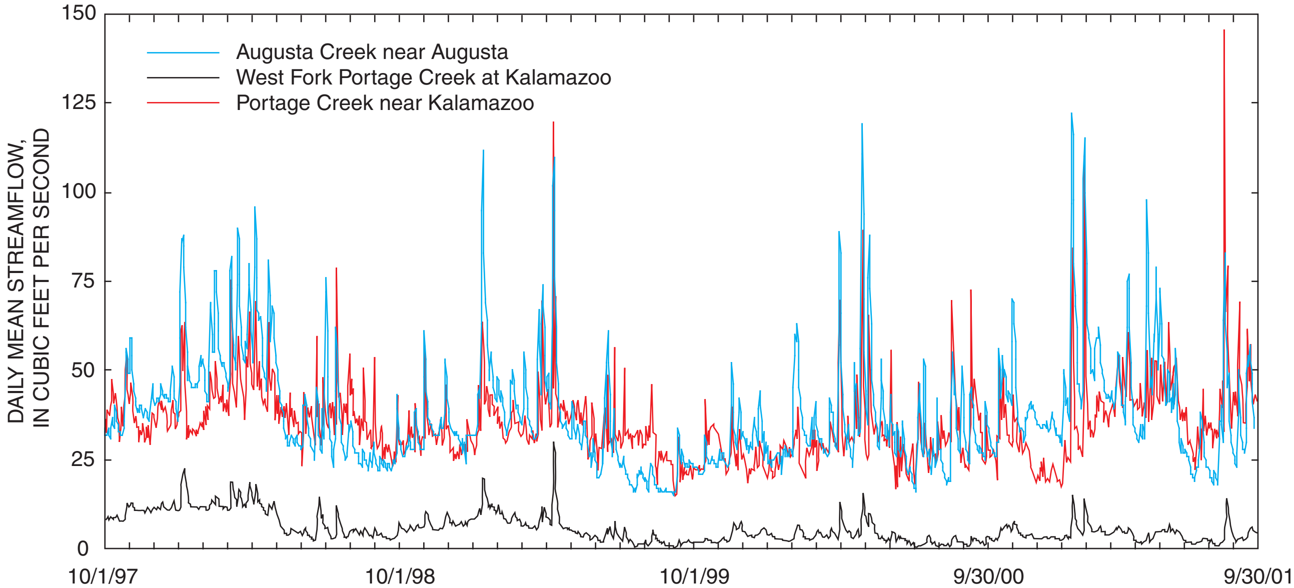


Figure 8. Daily mean streamflow for selected streams in Kalamazoo County, Michigan.

observed near shore and positive rates were observed further from shore. In September, higher negative rates were observed near shore, whereas small seepage rates were observed further from shore. Based on analysis of paired meters, the data indicate that there is not an appreciable difference between seepage rates measured with the tub and barrel meters.

Seepage measurements were collected at various sites around Austin Lake in July/August 2001. At Austin Lake, seepage rates varied over a much smaller range than at Long Lake, and no relation was indicated with distance from shore. Seepage rates ranged from -1.2 in./day to 0.6 in./day. These changes observed in seepage rates and directions over the data-collection period during this study likely indicate the presence of complex geologic or hydrologic controls in the Kalamazoo area.

Using Darcy's Law (2), hydraulic conductivities were determined for selected sites in Long Lake using the seepage and hydraulic-head gradient data. During this study, mini-piezometers were placed at 27 sites adjacent to seepage meters. Of these sites, measurements from 11 mini-piezometers were used for analysis. Data from the other mini-piezometers were discarded because the piezometers were not equilibrated at the time of measurement, visible plugging of the end of the piezometer at the time of removal from the lakebed, or to a different direction of seepage indicated by the piezometer than by the adjacent seepage meter. These different seepage directions likely are the result of difficulties incurred in measuring water levels when waves were present on the lake surface. Estimated hydraulic conductivities ranged from a high of 2,200 ft/d observed in June 2002 to a low of 3 ft/d observed in September 2002. The median lakebed hydraulic conductivity is 29 ft/d.

Seepage measurements were collected at 19 stream sites during the synoptic survey in August 2001 (table 5). All but two sites indicated ground-water discharge to the stream. Of the two that indicated water was leaving the stream, one contained shallow water barely above the level of the attached bag and the other was placed above a small rock dam. Thus, in both of these cases, the seepage measurements are questionable and only data from the other sites were included in analyses. Of the 17 remaining sites, seepage ranged from a low of 0.3 in./day (Unnamed tributary to the North Branch Paw Paw River, USGS streamflow-gaging station 04102217) to a high of 156.3 in./day (Augusta Creek, near Augusta, USGS streamflow-gaging station 04105700). The median streambed seepage rate is 7 in./day, whereas the average streambed seepage rate is 22.4 in./day. At one site, Augusta Creek near Augusta, three mini-piezometers were placed in an area near where the seepage meter was located because of the high seepage rates observed in this stream. Estimated hydraulic conductivity of the streambed materials averaged 12,000 ft/d.

Recharge

Determination of recharge is important because this water replenishes local aquifers and surface-water features. In the Kalamazoo County area, Allen and others (1972) reported that mean precipitation averages 35 in./yr. Allen and others (1972) estimated about 12 in. is discharged by streams, with about 3 in. originating as surface runoff and about 9 in. as ground-

Table 4. Stream discharge, measured during synoptic surveys, for Kalamazoo, Michigan, area streams.

[Discharge in cubic feet per second; --, discharge not available]

Station number	Station name	August 2001	December 2001	March 2002	June 2002	September 2002
04096950	Bear Creek at Z Avenue near Fulton	4.04	--	--	--	--
04097040	Little Portage Creek at TS Avenue, near Climax	1.81	--	--	--	--
04097060	Little Portage Creek at 38th Street, near Fulton	7.75	35.6	28.2	18	7.68
04097120	Portage River at S Avenue, near Pavilion	15.8	--	--	--	--
04097170	Portage River at W Avenue near Vicksburg	27.1	94.5	99.4	51.4	21.5
04097205	Gourdneck Creek at 23rd Street, near Vicksburg	8.36	24.6	26.7	18.3	11.1
04097207	Austin Lake Outlet at TU Avenue, near Vicksburg	0.46	1.52	5.23	3.61	0.7
04097210	Portage Creek at W Avenue, at Vicksburg	9.04	--	--	--	--
04097240	Portage Creek at Z Avenue near Mendon	29.0	73.1	86.2	65.9	35
04097340	Portage River at Parkville River, at Parkville	63.8	224	228	149	63.5
04097370	Flowerfield Creek at Flowerfield Road, at Flowerfield	9.86	35	39.1	20.9	12.7
04105500	Kalamazoo River near Battle Creek	380	1,200	1,060	602	353
04105671	Eagle Lake Drain, at Fort Custer Drive near Augusta	5.02	--	--	--	--
04105700	Augusta Creek at EF Road, near Augusta	20.2	50	48	25	25
04105800	Gull Creek at 37th Street, near Galesburg	9.91	41.3	38.9	41	19.3
04105990	Comstock Creek, at East Main Street, near Kalamazoo	3.42	9.38	8.63	7.91	6.64
04106000	Kalamazoo River at River Street, at Comstock	504	1,690	1,270	964	547
04106050	Davis Creek at Olmstead Road, at Kalamazoo	4.64	--	--	--	--
04106180	Portage Creek at Portage	12.3	19	22	16	13
04106300	Portage Creek at Lovers Lane, near Kalamazoo	35.8	42	32	46	33
04106320	West Fork Portage Creek at 12th Street, near Oshtemo	1.36	5.1	4.9	3.6	2.1
04106400	West Fork Portage Creek at Oakland Drive, at Kalamazoo	1.45	7.5	7.2	6.9	3.3
04106501	Portage Creek at Stockbridge Avenue, at Kalamazoo	41.9	47.6	42.6	54.7	38
04106513	Arcadia Creek at West Main Street, at Kalamazoo	1.99	--	--	--	--
04106750	Spring Brook at Riverview Road, near East Cooper	14.6	20.1	18.8	19.2	15.9
04106906	Kalamazoo River at Plainwell	710	1,820	1,520	1,120	669
04107750	Rupert Lake Outlet at AB Avenue near Plainwell	6.60	--	--	--	--
04107710	Sand Creek at 2nd Street, near Alamo	7.21	15.6	14.9	11.9	7.96
04102217	Unnamed tributary to North Branch Paw Paw River at 32nd Street, near Paw Paw	8.12	14.9	13.3	14.6	10.2
04102178	East Branch Paw Paw River at 30th Street, near Lawton	12.9	19.3	15.9	19.5	16.2

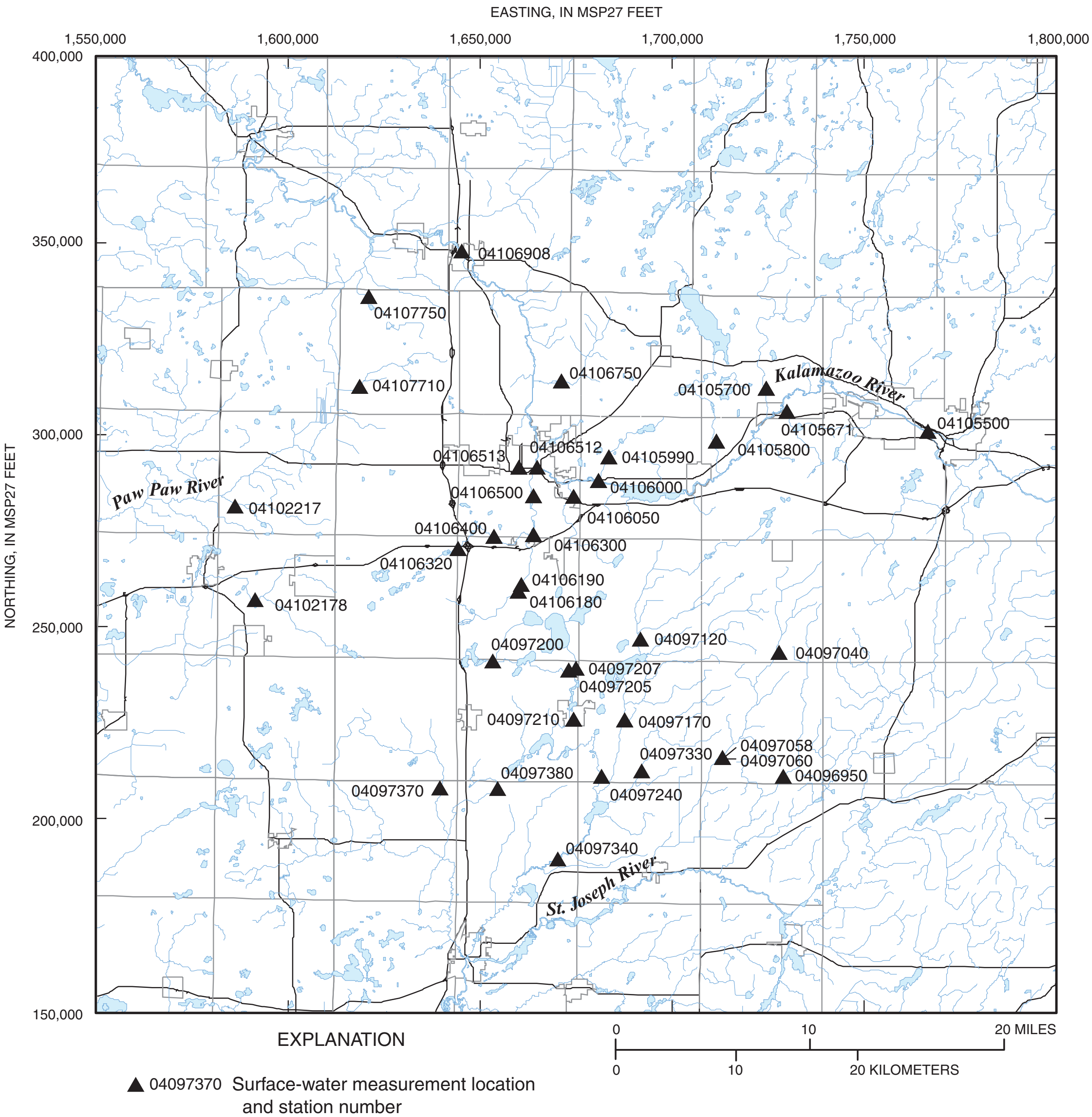


Figure 9. Surface-water measurement locations in the Kalamazoo County area, Michigan.

Table 5. Stream seepage, measured during August 2001 synoptic survey, for Kalamazoo, Michigan, area streams.

[Seepage measurements in inches per day]

Station number	Station name	August 2001 seepage
04096950	Bear Creek at Z Avenue, near Fulton	7.0
04097060	Little Portage Creek at 38th Street, near Fulton	13.7
04097120	Portage River at S Avenue, near Pavilion	7.0
04097170	Portage River at W Avenue, near Vicksburg	33.9
04097205	Gourdneck Creek at 23rd Street, near Vicksburg	2.4
04097210	Portage Creek at W Avenue, at Vicksburg	12.6
04097240	Portage Creek at Z Avenue, near Mendon	5.3
04097340	Portage River at Parkville Road, at Parkville	9.3
04097370	Flowerfield Creek at Flowerfield Road, at Flowerfield	127.6
04105671	Eagle Lake Drain, at Fort Custer Drive, near Augusta	.4
04105700	Augusta Creek at EF Road, near Augusta	156.3
04105800	Gull Creek at 37th Street, near Galesburg	16.6
04105990	Comstock Creek, at East Main Street, near Kalamazoo	3.3
04106750	Spring Brook at Riverview Road, near East Cooper	-25.9
04106906	Kalamazoo River at Plainwell	8.0
04107750	Rupert Lake Outlet at AB Avenue, near Plainwell	-.1
04107710	Sand Creek at 2nd Street, near Alamo	28.7
04102217	Unnamed tributary to North Branch Paw Paw River at 32nd Street, near Paw Paw	.3
04102178	East Branch Paw Paw River at 30th Street, near Lawton	19.2

water runoff. About 23 in. is evapotranspired or leaves the county as regional ground-water flow. Average yearly precipitation ranged from a low of 23.9 in. in 1999 to a high of 46 in. in 1975 based on data obtained from the National Weather Service Stations at Kalamazoo Hospital and Gull Lake Biological Station, and the Long-Term Ecological Research Station at Kellogg Biological Station (fig. 10). Precipitation was generally below average from 1960 to 1965 and 1992 to 1998 and at or above average from 1965 to 1990. The amount of water from precipitation that reaches the water table will vary depending on changes in precipitation and geologic factors, the extent of impervious surfaces, and the extent of municipal

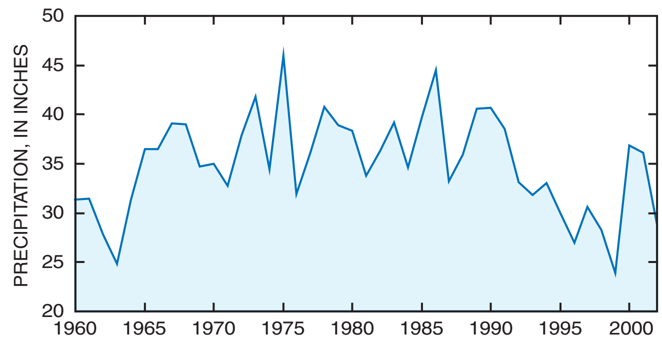


Figure 10. Average yearly precipitation from 1960 to 2001 for Kalamazoo County, Michigan [data from National Weather Stations at Kalamazoo Hospital and Gull Lake Biological Station and from the Long-Term Ecological Research Station at Kellogg Biological Station].

sewers that transfer water from one area to another. To estimate regional recharge rates, surface water and saturated-zone techniques are often used. Surface-water techniques include seepage meters and determination of base-flow discharge. Saturated zone techniques include numerical modeling (Scanlon and others, 2002).

Historical Estimates of Recharge

Initial estimates of recharge are available from previous studies. Holtschlag (1994) determined the spatial variation of average ground-water recharge rates for 1951–1980 from an analysis relating base-flow characteristics of streams to land use and basin characteristics in the Lower Peninsula of Michigan. On the basis of this study, the minimum average annual recharge in the Kalamazoo County area is 7.1 in./yr and the maximum is 17.7 in./yr; the average ground-water recharge rate is 11.3 in./yr. Allen and others (1972) prepared ground-water budgets for the Kalamazoo and St. Joseph River Basins to obtain estimates of ground-water recharge. Allen and others (1972) determined the ground-water-runoff part of the budget using streamflow records for the Kalamazoo River at Comstock and the Portage River near Vicksburg. A streamflow hydrograph, which is a graph of discharge over time, can be divided into a direct-runoff component and a ground-water-runoff component. The direct-runoff component is associated with precipitation that enters the stream as overland runoff, and the ground-water component, also called the base-flow component, is associated with ground-water flow into a stream. The long-term average recharge rate was determined to be 9 in./yr for each basin. Estimated ground-water recharge varied from 4 in./yr in 1964 to 13 in./yr in 1943 and 1950 in the Kalamazoo River Basin.

Rheaume (1990) estimated ground-water recharge rates in Kalamazoo County for different geologic settings using ground-water runoff to streams. In the outwash plain area, estimates of ground-water recharge rates varied from 9.5 to 13.3 in./yr and averaged 10.9 in./yr. In the upland moraine area, estimates of ground-water recharge rates varied from 4.7 to 6.7 in./yr and averaged 5.9 in./yr. In the downcut glacial drainage channel, estimates of ground-water recharge rates varied from 7.2 to 11.8 in./yr and averaged 8.8 in./yr. In the till plain area, estimates of ground-water recharge rates varied from 5.0 to 9.5 in./yr and averaged 6.9 in./yr. Rheaume (1990) estimated the countywide weighted recharge rate to be 9.3 in./yr, which is similar to the value estimated by Allen and others (1972).

Estimates Determined During this Study

During this study ground-water recharge rates were investigated using base-flow separation to determine the ground-water contribution to streamflow. Base-flow separation using the program PART, developed by Rutledge (1993 and 1998), was performed on 11 streams in the study area with streamflow data within 1965–2001. The period of record for these streams ranged from 2 to 36 years. Estimated ground-water-recharge rates ranged from a low of 2.9 in./yr in 2000 for West Fork Portage Creek near Oshtemo (USGS streamflow-gaging station 04106320), to highs of 19.2 in./yr in 1986 for Augusta Creek near Augusta (USGS streamflow-gaging station 04105700), and 29.1 in./yr in 1991 for Portage River near Kalamazoo (USGS streamflow-gaging station 04106300). An industry discharges some water to Portage River, which increases the estimated base flow for this stream. Estimates of average ground-water-recharge rates over the period of record for each station ranged from 5.9 in./yr to 22.9 in./yr.

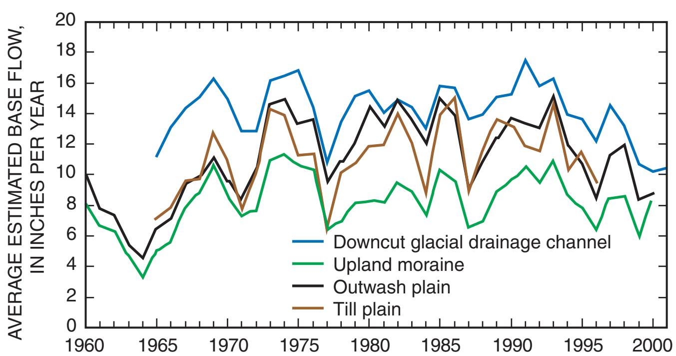


Figure 11. Average estimated base flow by major surficial material type for streams in the Kalamazoo area, Michigan.

Because of the differences in estimated ground-water-recharge rates in different geologic settings observed by Rheaume (1990), base flow of streams was compared for each geologic setting during this study (fig. 11). Average annual base-flow values were determined for the streams within each of the four geologic settings. Average annual estimated base flow was lowest in the upland moraine setting and highest in the outwash plains areas.

Surface-water techniques, such as the determination of base-flow discharge, typically provide an estimate of the potential recharge (Scanlon and others, 2002). Rushton (1997) distinguished potential from actual recharge that reaches the water table. Infiltrated recharge may or may not reach the water table because of unsaturated zone processes, bank storage and subsequent evapotranspiration, development of perched aquifers, and the inability of an aquifer to accept recharge because of a shallow water table or a low transmissivity (Lerner and others, 1990). Thus, actual ground-water-recharge rates likely will differ from those determined from stream base flow. Several control structures are present on the Kalamazoo River, which also complicates the estimation of ground-water-recharge rates.

Two additional factors affecting recharge, or the amount of water reaching the water table, are irrigation and urbanization. The withdrawal and subsequent addition of ground water to irrigate local crops may provide, depending on irrigation efficiency, for a decrease or increase in the amount of water reaching the water table. Impervious surfaces limit recharge if these areas drain to storm sewers that transport the water to another area or to a stream for release. However, if the impervious areas or rooftops drain to lawns or other permeable areas, recharge is not reduced; thus, recharge may or may not be reduced in urban areas. For areas served by individual residential wells and septic tanks, most of the water withdrawn for use is assumed to be returned to the ground-water-flow system. For areas served by individual residential wells and municipal sewers, the water that is withdrawn for use is transported to another area for discharge; therefore, these areas have reduced recharge.

History of Municipal Pumpage

Pumpage by municipal suppliers was obtained or estimated for various years from 1966 to 2001 (table 6 and fig. 7). Ground-water withdrawals generally follow county population totals. County population generally increased from 1960 to 1980, slightly declined in 1990, and increased again in 2000, while overall withdrawals increased from 1960 to 2000. Population within the city of Kalamazoo generally declined over this same time period while city withdrawals generally increased. These withdrawals include those for residential and business use. Population and ground-water withdrawals within the city of Portage increased over this time period. Population and ground-water withdrawals remained about the same for the communities of Augusta, Galesburg, Parchment, and Vicksburg over this time period.

Table 6. Estimated ground-water withdrawals for Kalamazoo, Michigan area communities, 1966-2001.

[In million gallons per year]

Community water supply	1966	1987	1994	2001
City of Galesburg	50.4	75.9	90.3	105.1
City of Kalamazoo	5,584.5	7,193.1	7,930.0	7,385.8
City of Parchment	105.9	167.1	145.0	128.8
City of Portage	629.5	1,353.4	1,513.6	1,860.1
Village of Augusta	32.6	30.4	34.3	31.2
Village of Climax	10.0	12.4	20.6	23.9
Village of Mattawan	102.8	110.6	118.9	127.9
Village of Schoolcraft	66.4	55.3	69.8	68.2
Village of Vicksburg	131.1	112.7	103.7	93.0
Charleston Township	10.7	11.5	12.4	13.3

Conceptual Model of Ground-Water Flow

The conceptual model describes ground-water flow within Kalamazoo County and surrounding areas and includes delineation of aquifers, confining units, and ground-water-flow-system boundaries. Geologic units within the Kalamazoo County area can be combined into units that control ground-water flow. Six units were used to conceptualize the ground-water-flow system, although not all units are present throughout the study area. These units consist of three permeable sand-and-gravel units separated by confining units consisting primarily of clay and silt, and a lowermost unit consisting of the Marshall Sandstone where it is present in the northeastern part of the study area and clay or other poorly permeable materials where they are present overlying the Coldwater Shale in the rest of the area. Geologic information obtained from about 3,200 logs of production, domestic, and oil and gas wells and from previous reports was used to delineate the extent of the ground-water-flow-system units.

Ground-water-flow systems have physical boundaries, which are formed by an impermeable body of rock or a large body of water, or hydrologic boundaries, which include ground-water divides and streamlines. Regional hydrologic boundaries to the flow system in the Kalamazoo County area are formed by the boundaries of the Kalamazoo River, St. Joseph River, and Paw Paw River Basins that divide the county and cover a large portion of southwestern Michigan. Boundaries for the local flow system in the Kalamazoo area consist primarily of surface-water features. Ground-water flow is assumed to be bounded by the St. Joseph River to the south; Pine Creek, North Branch Paw Paw River, and South Branch Paw Paw River to the west; numerous lakes, Gun River, and Kalamazoo River to the north; and Nottawa Creek, Pine Creek, and Kalamazoo River to the east. The upper boundary of the ground-water flow system is formed by the top of the unconfined aquifer and is equal to the water-table altitude. The lower boundary is formed by the upper surface of the Coldwater Shale, which is considered to be impermeable.

Simulation of Ground-Water Flow

Simulation of ground-water flow is made possible by first developing a conceptual model of the flow system and then developing a numerical model that is consistent with the conceptual model. The model area consists of Kalamazoo County and parts of the surrounding counties, including Allegan, Barry, Calhoun, St. Joseph, and Van Buren Counties. This larger area (larger than Kalamazoo County) is simulated to minimize boundary effects on the ground-water-flow solution in the interior part of the model by allowing natural physical and hydrologic boundaries to be used. The model design also allows for development of predictive simulations within Kalamazoo County to investigate water availability and the effects of water withdrawals.

The U.S. Geological Survey Modular Three-Dimensional Finite-Difference Ground-Water Flow model, MODFLOW-2000 (Harbaugh and others, 2000), was used to simulate ground-water flow in the Kalamazoo County area. This code allows the simulation of steady-state or transient ground-water flow in three dimensions with leakage between model layers. In addition, this code contains processes that allow inclusion of observations and comparison of measured and calculated values, development of parameter sensitivities, and estimation of parameter values (Hill and others, 2000).

Model development includes the determination of grid characteristics, boundary conditions, and input variables, such as layer hydraulic properties and stresses. Model calibration then proceeds with the comparison of model-simulated heads (water levels) and streamflows with measured values for steady-state or transient conditions. Under the steady-state assumption, all water entering the model area through the boundaries or as recharge is assumed to leave the model area through the boundaries, rivers, or wells. No ground-water storage or temporal discretization terms are required. Ground-water withdrawals and recharge remain constant during the simulation. During a transient simulation, water is released from or taken into storage from the porous material. Heads change with time as a result of this transfer of water. When the transfer to and from storage stops, the ground-water-flow system reaches steady state and heads stabilize (Anderson and Woessner, 1992). Ground-water withdrawals and recharge can change during the transient simulation. Both steady-state and transient models are useful for simulating the ground-water-flow system. Average conditions are used in steady-state models to predict long-term average heads and flows. Transient models, which consider water movement into and out of storage, can predict responses to short-term changes in stresses.

For this study the ground-water-flow model was calibrated to both steady-state and transient conditions to completely simulate the ground-water-flow system.

Steady-state models were developed to represent conditions in the Kalamazoo County area in the years for which both water-level and streamflow data were available for calibration. These years include 2001, 1994, 1987, and 1966. Available data on ground-water withdrawals and observations for each time period were averaged to be representative of average recharge years. Ground-water withdrawals and recharge rates were kept constant for each simulation.

As mentioned previously, recharge rates are known to vary annually and seasonally; therefore, transient models also were developed to investigate the ground-water-flow-system response to seasonal and monthly changes in recharge and withdrawal rates from October 2001 to September 2002. The following sections describe details of model development, calibration, and results of simulations with the steady-state and transient models.

Spatial Discretization

The model area is rectangular and covers about 1,023 mi^2. The area is approximately 38-mi long (north-south) and 40-mi wide (east-west). For model simulation, this area is horizontally discretized into a variably spaced grid of cells in 154 rows and 162 columns. In the central portion of the model area, each cell is approximately 660 by 660 ft. Cell spacing increases by a factor of 1.2 to a maximum grid spacing of about 2,730 ft. This maximum cell size was selected to not exceed a maximum cell length to width ratio of 10:1 as recommended by Anderson and Woessner (1992). Grids that do not exceed these recommended ratios between adjacent row and column widths and lengths for individual cells have a reduced likelihood of numerical difficulties when the model equations are solved. Each grid cell is assigned the average aquifer properties for the volume of aquifer represented by the cell; variations in properties that are within a grid cell cannot be represented. Glacial deposits are known to vary considerably in lithology and thickness over short distances (tens to hundreds of feet). This variability makes exact representation of the detailed hydrogeology impossible in a numerical model. Thus, hydraulic properties of the units within glacial deposits are generalized to represent the regional ground-water-flow system.

The model area is discretized vertically into six model layers (table 7). Layer 1 represents the upper aquifer in the glacial deposits. This aquifer is unconfined with water levels representing the water table. Layer 2 represents the underlying confining unit. Layer 3 represents the intermediate aquifer. Layer 4 represents a confining unit. Layer 5 represents the lower aquifer. Layer 6 represents the Marshall Sandstone in the northeastern part of the model area and clay where present over the rest of the model area.

Table 7. Relation between geologic units and model layers, Kalamazoo County area, Michigan.

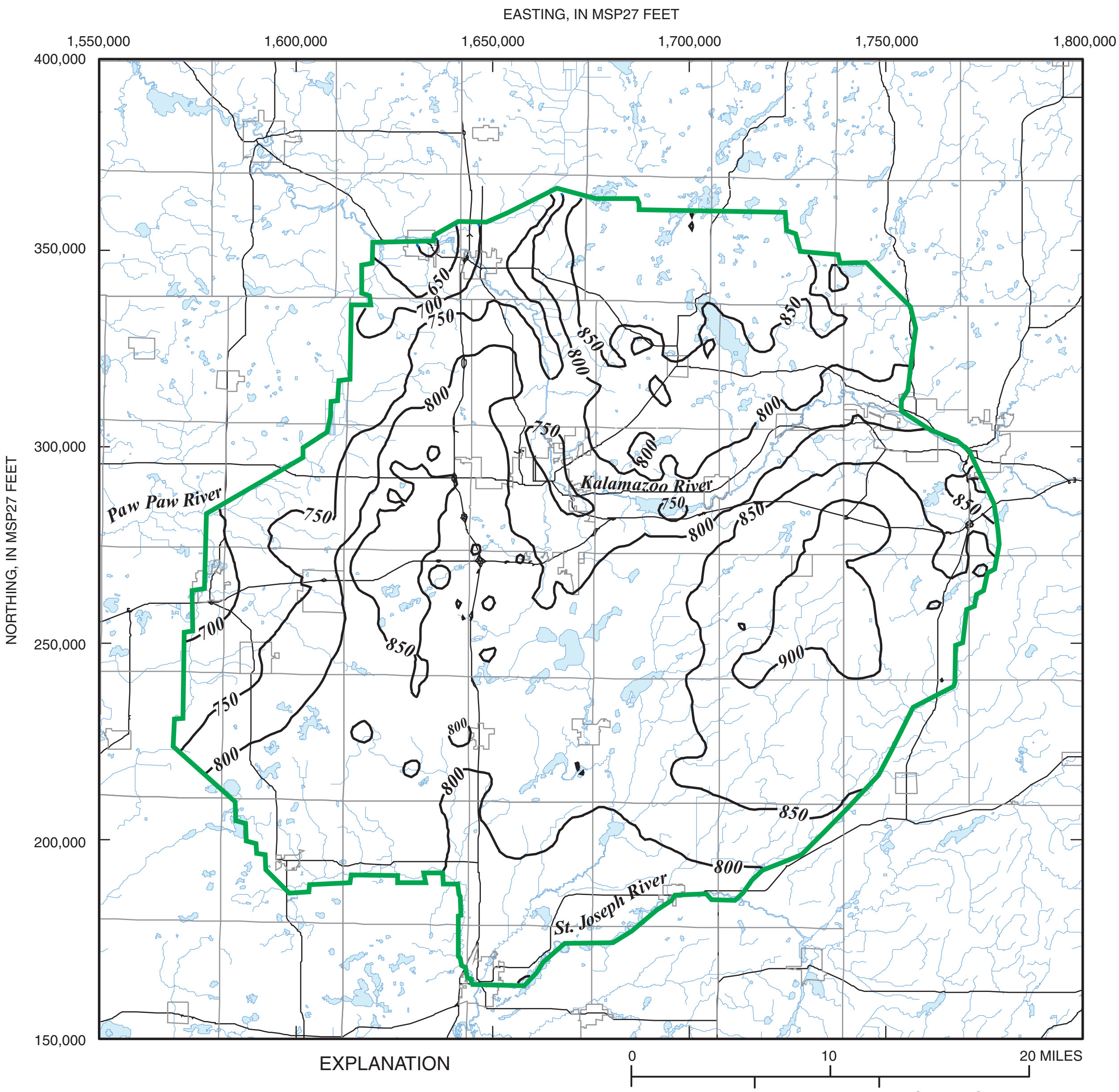
Geologic unit	Model layer
Upper aquifer	1
Upper confining unit	2
Intermediate aquifer	3
Lower confining unit	4
Lower aquifer	5
Till / Marshall Sandstone	6

Land-surface information for most of the model area is available from USGS 30 m elevation data with a vertical resolution of 5–10 ft. This information was used to construct model layers by subtracting depths of estimated layers determined using well log data from the land-surface elevation. The initial estimates of water levels in wells also were developed by subtracting the depth to water recorded on well logs from the land-surface elevations.

Layer surfaces were determined based on about 3,200 logs of public-supply and domestic wells obtained from the Geological Survey Division, Michigan Department of Environmental Quality (MDEQ) and the Statewide Ground-Water Database (now WELLOGIC), MIRIS-Geologic Resource Mapping Unit, MDEQ (figs. 12–16). Additional information was obtained from 145 logs of oil and gas wells provided by John Paquin, Water Resources Manager, city of Kalamazoo and from previous reports (Allen and others, 1972; city of Kalamazoo and Peerless-Midwest, 1999a, 1999b, 2002a, and 2002b). Information on the bedrock surface from the well logs was used to refine the estimated bedrock surface in the Regional Aquifer Systems Analysis (RASA) study (Hoaglund and others, 2000). In areas where the geologic materials comprising a layer are absent, the layer is assigned a minimal thickness of 1 ft and the hydraulic properties of the underlying layer. Additional delineation of layer surfaces in areas of sparse data was accomplished by interpolating from known nearby points.

Boundary Conditions

Model boundaries extend from less than a mile to about 9 mi beyond the boundaries of Kalamazoo County and consist primarily of local surface-water features. Boundaries for the model are based on surface-water elevations and topographic information, and are represented by specified-head and specified-flux conditions. Specified-head boundaries (also known as constant-head boundaries) are simulated by specifying head values that do not change during numeric simulation. Specified-flux boundaries can be simulated by



EXPLANATION

— *850* — Layer contour—Shows altitude of the bottom of the upper aquifer. Contour interval 50 feet. Datum is NGVD 29

Model area boundary

Figure 12. Estimated bottom of the upper aquifer, Kalamazoo County area, Michigan.

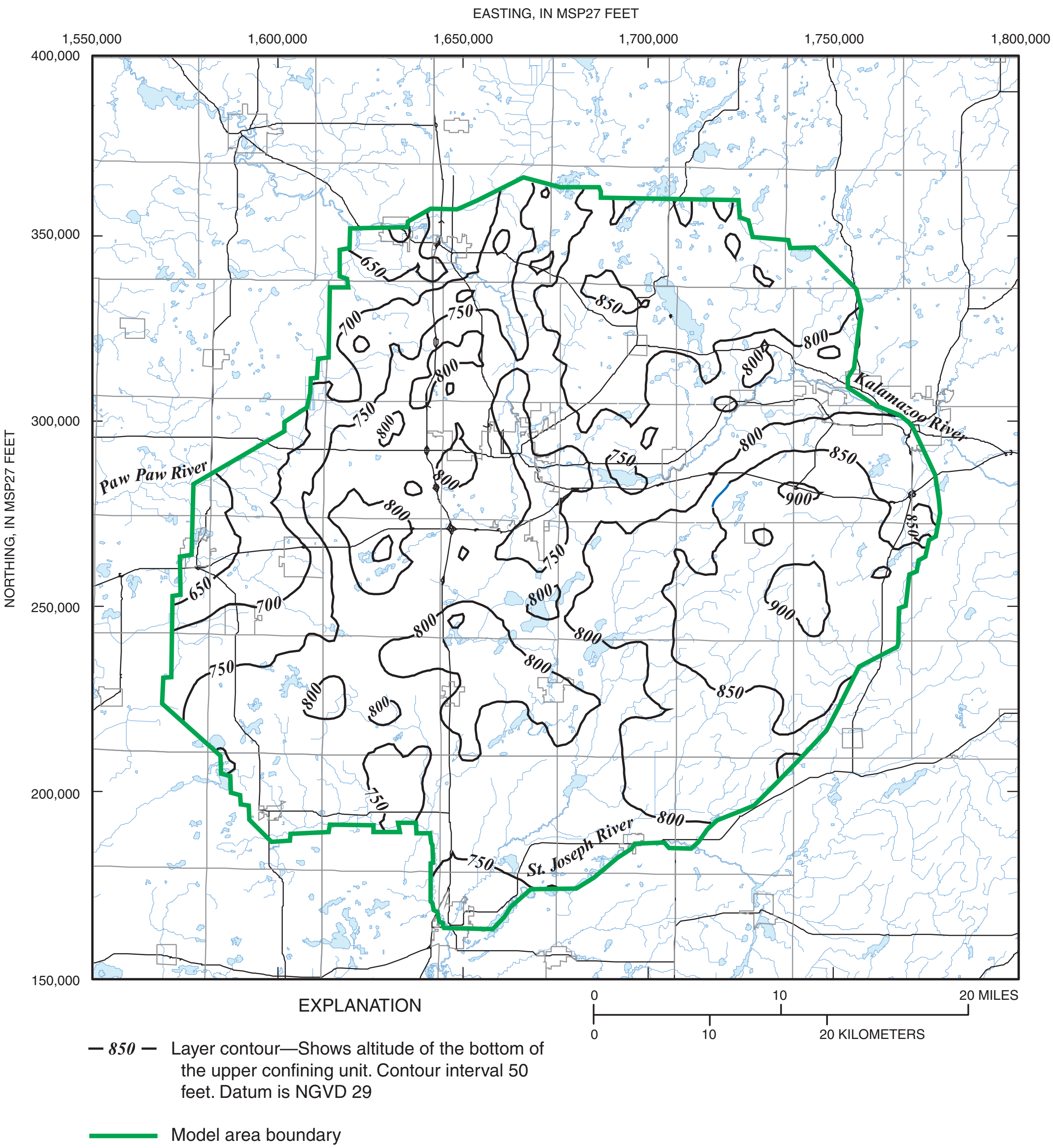


Figure 13. Estimated bottom of the upper confining unit, Kalamazoo County area, Michigan.

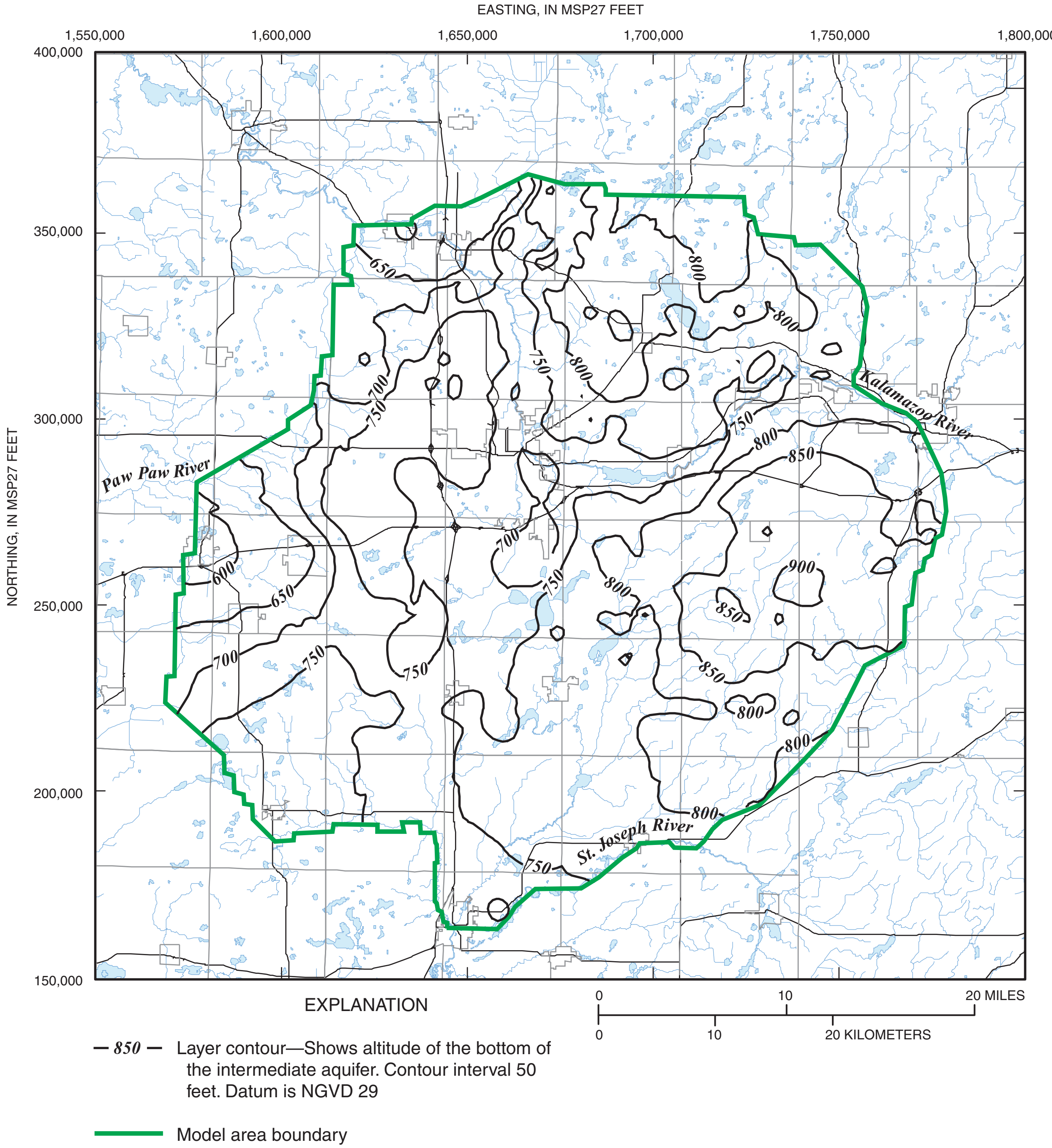


Figure 14. Estimated bottom of the intermediate aquifer, Kalamazoo County area, Michigan.

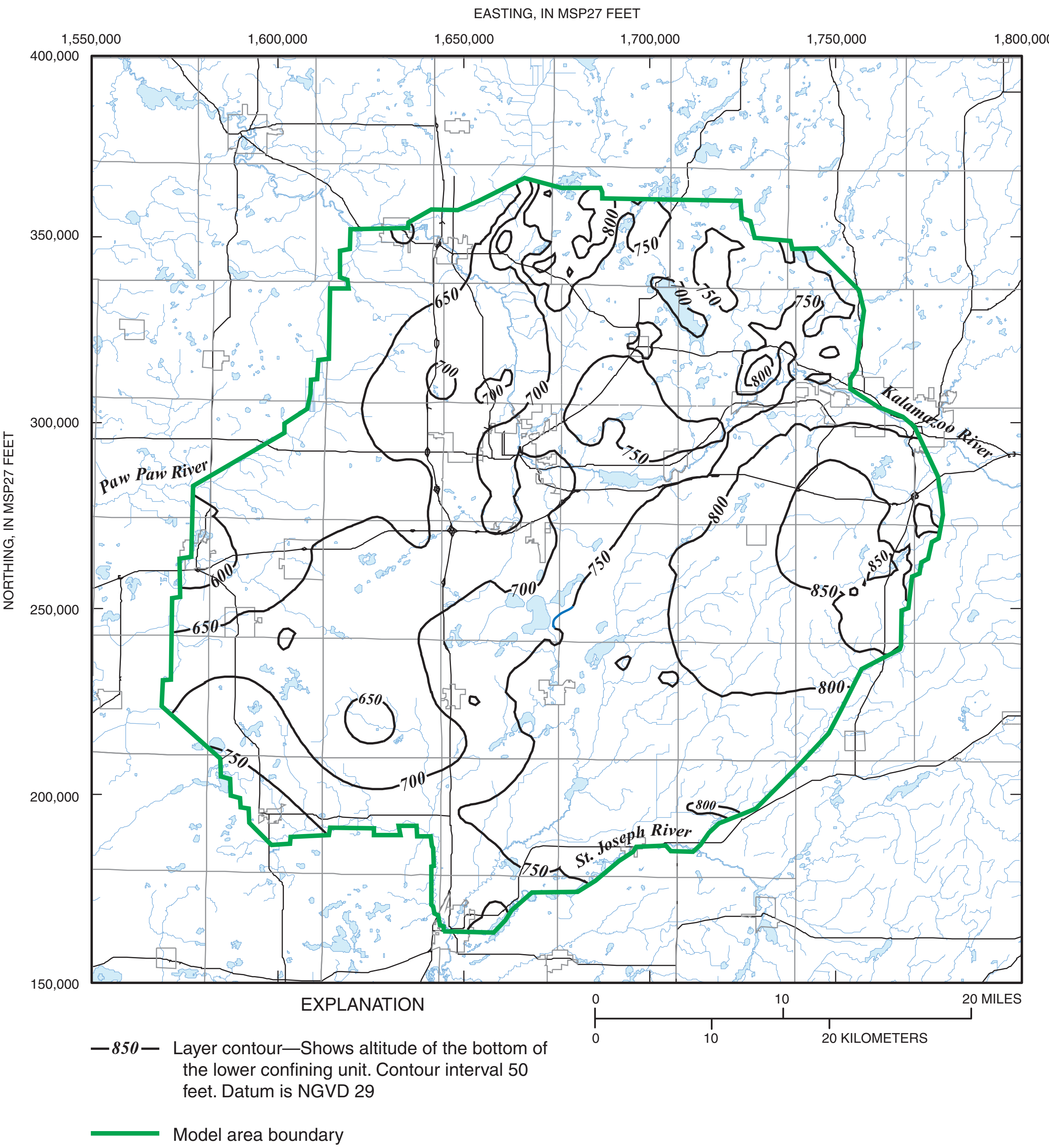


Figure 15. Estimated bottom of the lower confining unit, Kalamazoo County area, Michigan.

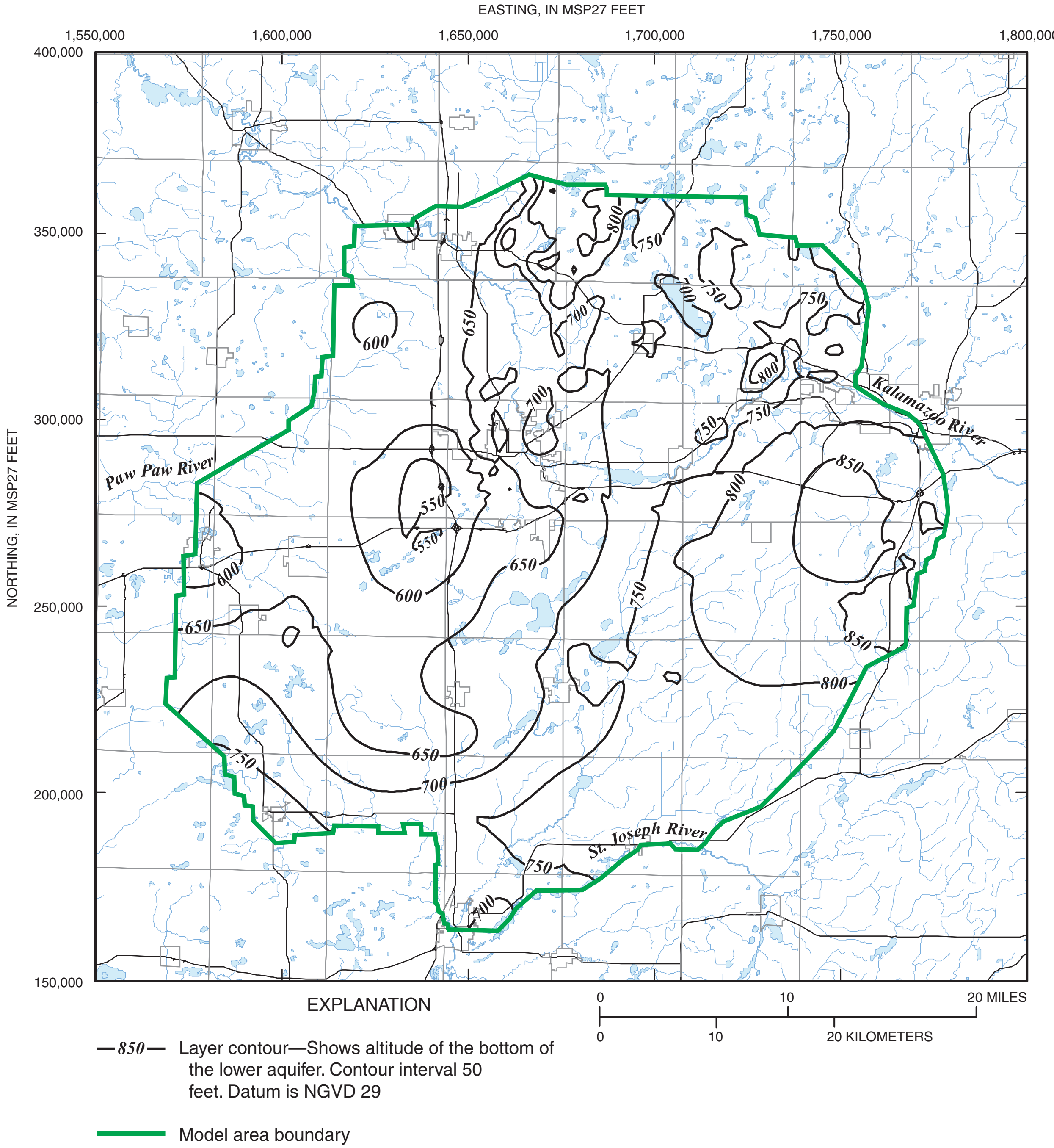


Figure 16. Estimated bottom of the lower aquifer, Kalamazoo County area, Michigan.

specifying the flux equal to zero to represent ground-water-flow divide boundaries and streamlines. Specified-head boundaries are used to represent hydraulic conditions along the model boundaries that coincide with surface-water features and are based on surface-water elevations (fig. 17). Specified-flux boundaries with flux equal to zero are used to represent natural physical boundaries along the remaining model boundaries. All boundaries of layers 3 and 5 are the same as those of layer 1. Layers 2 and 4 are bounded by no-flow boundaries. After model development, sensitivity of model results to boundary conditions was analyzed.

Hydraulic Properties and Stresses

Hydraulic properties used in model simulations include layer hydraulic conductivities and leakances, recharge rates, and streambed conductances. Aquifer hydraulic properties and leakances affect ground-water flow through and between model layers. Recharge rates indicate the amount of water movement through the upper boundary of the model. Streambed conductance affects vertical flow of ground water from an aquifer to a stream or from a stream to an aquifer. Stresses include withdrawal of water from the ground-water-flow system by wells.

As mentioned previously, information on hydraulic properties of the glacial aquifers and confining units is available from previous studies, analysis of aquifer test and recovery data collected during or prior to this study, and from representative values estimated by Freeze and Cherry (1979) (table 1). Initial estimates of hydraulic conductivity ranged from 0 to 375 ft/d for the upper aquifer (model layer 1) and from 0 to 433 ft/d for the intermediate and lower aquifers (model layers 3 and 5). Initial estimates of 0 indicate areas where the aquifer is absent; during model simulations, areas where an aquifer is absent from a layer are assigned the estimated hydraulic conductivity of the underlying layer. Initial estimates of hydraulic conductivity ranged from 0.02 to 1.3 ft/d for the upper confining unit (model layer 2) and were assumed equal to 0.05 ft/d for the lower confining unit (model layer 4). Vertical hydraulic conductivities were assumed to be 0.1 times the horizontal hydraulic conductivity for each model cell. Layer 6 comprises parts of the clay units overlying the Coldwater Shale and the Marshall Sandstone. The hydraulic conductivity of the clay units initially was estimated to be 0.01 ft/d. The hydraulic conductivity of the Marshall Sandstone initially was estimated to be 100 ft/d.

Initial estimated recharge rates were assumed to differ depending on the type of surficial geologic material (fig. 2). Initial estimates for each geologic setting are defined by Rheaume (1990) and equal 6.89 in/yr for the till plain, 10.86 in/yr for the outwash plains, 8.79 in/yr for the downcut glacial drainage channel, and 5.87 in/yr for the upland moraine areas. Additional areas of increased recharge were included in the model to account for water that is returned to the ground-water and surface water systems after use by a local industry for all 4 calibration years. These areas include Austin Lake, Upjohn Pond, and Portage Creek. In 2001, some additional recharge from Pfizer, Inc. was added to the east wetland area that is east of Upjohn Pond and additional recharge was added to Long Lake to account for a well that was installed by the local homeowners to supplement the lake. In 2002, an additional area, known as the East Road recharge area, was included in transient model simulations as ground-water recharge during the winter months.

Streambed conductance is calculated as the product of the hydraulic conductivity of the streambed materials, stream length, and stream width, divided by the streambed thickness. Stream lengths for cells representing rivers were equal to the length of the stream segment in each cell. Stream widths varied depending on the size of the stream and were assigned values ranging from 10 ft for most streams in the area to 400 ft for the Kalamazoo River. Streambed thicknesses were assumed to equal 1 ft. Streambed hydraulic conductivities initially were assigned a value of 25 ft/d. Horizontal hydraulic conductivity of the lakebed materials initially was assigned a value of 200 ft/d.

Ground-water withdrawals were simulated from the center of the cell containing a pumping well and were based on estimated values. For the purposes of this study and to obtain representative estimates for model simulation, ground-water withdrawals were averaged using 1 to 5 years of data. At times a well field or portions of a well field may be down because of well maintenance or other issues; therefore, 1 year may not be representative of the typical ground-water withdrawal amounts from the well field. For the cities of Kalamazoo and Portage with consecutive withdrawal data from 1963 to 2001, withdrawals were determined as follows. For 1966, ground-water-withdrawal data from the cities were averaged with data from Allen and others (1972). For 1987 and 1994, withdrawals for the 3 years centered on the calibration year were averaged. Thus, for the 1987 calibration period, data from 1986 to 1988 were used and, for the 1994 calibration period, data from 1993 to 1995 were used. For 2001, data were averaged using available data from 1997 to 2001. For other communities within the study area, withdrawals were estimated using data supplied by municipal suppliers and included in annual ground-water data reports prepared by the USGS from 1970 to 1990 (Huffman and Thompson, 1971, 1973; Huffman, 1974a-88; Huffman and Whited, 1988; 1989; 1991; 1993) or previous studies of this area (Allen and others, 1972 and Rheaume, 1990).

Data collected and estimated for Pfizer, Inc. included both withdrawals from the ground-water-flow system and releases back to ground water and surface water at Austin Lake, Upjohn Pond, the east wetland area adjacent to Upjohn Pond, and Portage Creek. These releases were simulated using areas of increased recharge in the model. For 1994, withdrawal data for Pfizer, Inc. were obtained from the report by City of Kalamazoo and Peerless-Midwest (1996), whereas withdrawal data for other years were reported by Pfizer, Inc.

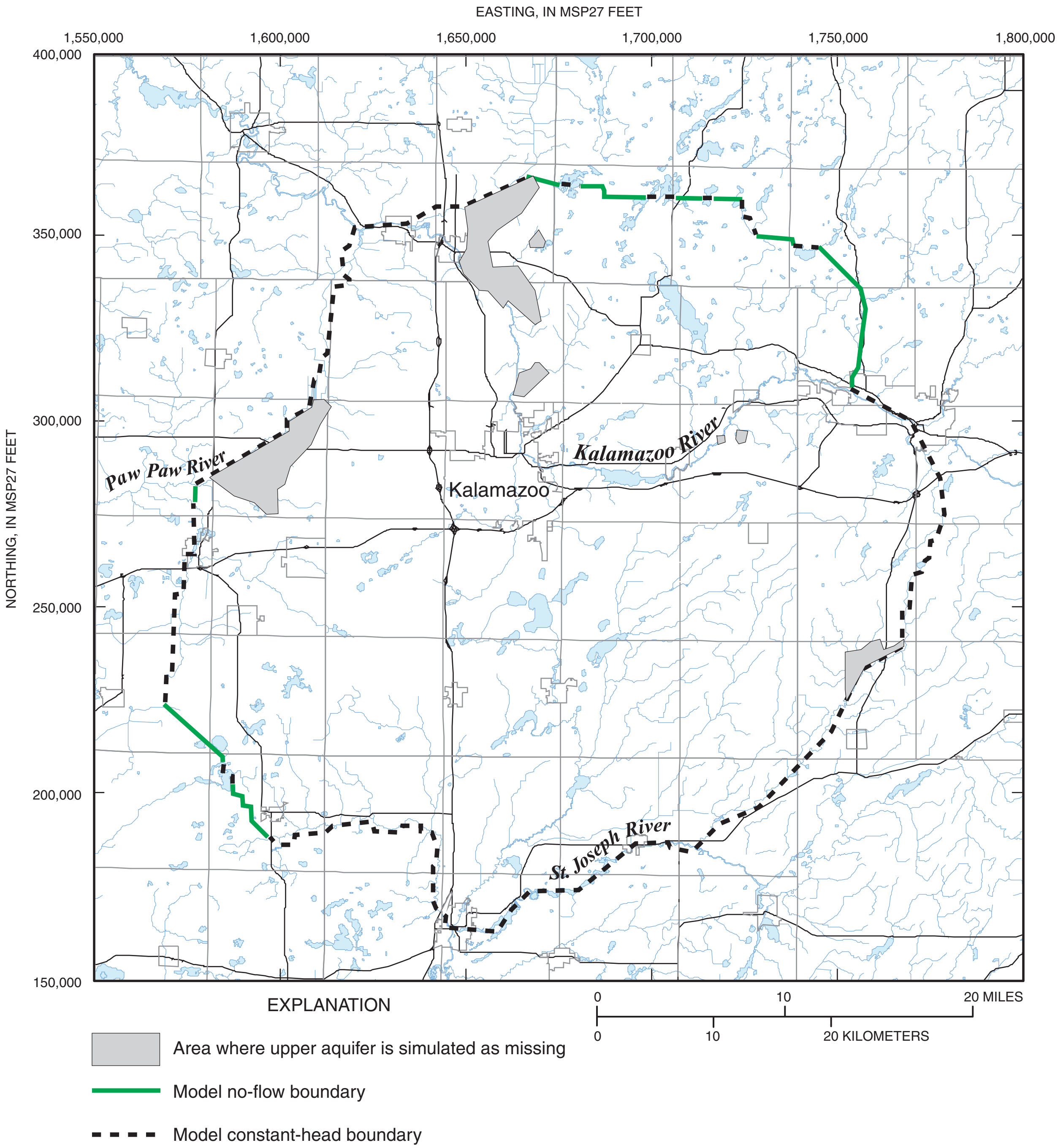


Figure 17. Ground-water-flow model boundaries, Kalamazoo County area, Michigan.

Steady-State Calibration

Model calibration is the process of reducing the difference between observed and simulated water levels and flows by adjusting model parameters. MODFLOW-2000 (Hill and others, 2000) contains observation, sensitivity, and parameter-estimation processes as part of the ground-water modeling computer program. The observation process allows comparison of observed and model-calculated values. The sensitivity process determines the sensitivity of hydraulic heads to specified model parameters, or in combination with the observation process, determines sensitivities for the simulated values associated with observations. A modified Gauss-Newton method is used in the parameter-estimation process to adjust the values of input parameters in an iterative procedure to minimize the difference between observed and simulated values. A weighted least-squares objective function, also known as the sum of squared, weighted residuals, is used to evaluate the fit between simulated and observed hydraulic heads and flows. Additional programs are included with MODFLOW-2000 to facilitate the evaluation of the accuracy of the model in simulating actual processes and the quantification of the uncertainty of model-simulated values (Hill and others, 2000).

Model fit is evaluated by comparing the magnitude and distribution of the residuals between simulated and observed water levels and flows. Simulated water levels and flows can be plotted against observed values, and the deviation from a straight line gives one indication of model fit. Plotting the residuals, which is the observed value minus the simulated value, shows the distribution and indicates possible biases in the model. A generally random distribution of positive and negative residuals can indicate that the model is not overpredicting or underpredicting water levels in parts of the model area. Water levels and streamflow from steady-state simulations were compared with observed and compiled data for 2001, 1994, 1987, and 1966. For this study, two additional comparisons were made to determine model fit to observed conditions. Fluxes through stream cells corresponding to sites where stream seepage was measured were compared to model-simulated seepage. Also, using data from an aquifer test in Ross Township (City of Kalamazoo, 1995), model-simulated drawdown was compared to observed drawdown values.

Parameter Estimation

For this study, model calibration was achieved by manual trial-and-error adjustment of parameters and by use of the automated parameter-estimation program in MODFLOW-2000. The following paragraphs describe the data used for calibration and the model parameters that were adjusted to achieve the best fit between observed and simulated water levels and flows.

For each steady-state model simulation, available miscellaneous and continuous water-level and flow observations were used for calibration. These observations were weighted for the parameter-estimation process to account for differences in measurement accuracy and differences in units of measurement. This weighting strategy allows those measurements with a higher degree of confidence to have more effect on parameter estimates and ensures that both water level and flow data, which are measured in different units, affect parameter estimates.

Water-level observations collected during this or previous USGS studies generally were assigned a higher degree of confidence because datums commonly are established by field surveys and measurements are obtained by specially trained and equipped observers. Water-level observations from other sources were assigned a lower degree of confidence because of the uncertainty associated with estimating a datum from a topographic map, measuring depth to water in a recently developed or currently used domestic well, and fluctuations in levels of surface-water bodies. Streamflow data included daily mean discharge data from streamflow-gaging stations and streamflow-measurement data from miscellaneously measured sites. Miscellaneous streamflow-measurement data collected during this study was during a time of little or no precipitation; therefore, the surface-runoff component of streamflow is assumed to be small and the measured streamflow was assumed to be composed primarily of ground-water discharge. These discharge observations likely are higher than actual base flow, so simulated flux to streams should be lower for these stations. For miscellaneous streamflow-measurement data collected prior to this study, no information is available on the hydrologic conditions immediately preceding and during the data-collection period; therefore, these observed streamflow data also are expected to be higher than model-simulated flows. Incremental ground-water discharge between gaging stations and miscellaneous sites was calculated as the difference between flows at each measurement location. For those wells and streams where various measurements were available over the course of the calibration period, the average and standard deviation were calculated from the observations and used as the calibration target and observation weight. For those streamflow-measurement locations where calculation of the incremental values left the calibration target much smaller than the weight, weights were assigned to be equal to half of the incremental value. For continuous-record streamflow-gaging stations and water-level sites, the calibration period consisted of the 5-year period centered on the calibration year. Thus, for 1966, available data from 1964 to 1968 were used; for 1987, available data from 1985 to 1989 were used; and for 1994, available data from 1992 to 1996 were used. For 2001, streamflow data collected from August 2001 to June 2002 and water-level data collected from October 2001 to September 2002 were used.

For the 2001 calibration period, observations from 82 wells, collected during this study and provided by Peerless-Midwest, Inc. (Mike Chapman, Peerless Midwest, Inc., written commun., 2002), and 30 streams were used to improve and evaluate model fit. Observation weight was determined to be the standard deviation of available measurements for the calibration period for wells and streams measured during this study. These weights ranged from 0.21 to 2.95 for wells and from 19,900 to 7,710,000 for streams. Water-level observations from Peerless-Midwest, Inc. were assigned a weight of 5. For the 1994 calibration period, observations from 75 wells, obtained from continuously monitored wells and the city of Kalamazoo and Peerless-Midwest, Inc. report (1996), and 6 continuous-record streams were used. Observation weights for measurements collected from continuous wells and streams ranged from 0.17 to 3.32 for wells and from 36,300 to 7,700,000 for streams. Observations from City of Kalamazoo and Peerless-Midwest, Inc. (1996) were assigned a weight of 5 for measurements collected in 1994 and a weight of 8 for measurements collected prior to 1994. For the 1987 calibration period, observations from 49 wells, obtained from continuously monitored wells and from Rheaume (1990), and 7 continuous-record streams were used. Observation weights for measurements collected from continuous wells and streams ranged from 0.006 to 6.74 for wells and from 65,900 to 5,930,000 for streams. If only one measurement was available for the calibration period, a weight of 5 was assigned. For the 1966 calibration period, observations from 10 continuous wells and 9 continuous-record streams were used. Observation weights for measurements collected from continuous wells and streams ranged from 0.23 to 3.11 for wells and from 150,700 to 4,850,000 for streams.

Parameterization is the process of identifying the aspects of the simulated ground-water-flow system that are to be represented by estimated parameters (Hill and others, 2000). Possible choices for parameters in a ground-water-flow model include the hydraulic characteristics of the aquifers, confining units, and streambed materials, and the ground-water recharge rates. Typically, it is impossible to estimate all parameters of interest because of limitations in the data sets available for calibration. Therefore, in this study, automated parameter estimation was combined with manual adjustments of some parameters to investigate their effects on model fit. Parameters were selected to scale the initial estimates of the layer hydraulic conductivities and leakances, streambed conductance, and recharge rates. The layer hydraulic-conductivity parameters were estimated only for the model layers representing aquifers (model layers 1, 3, and 5). The parameters associated with the model layers representing confining and lowermost units (model layers 2, 4, and 6) could not be estimated because there were no observations in these units. There were two leakance parameters (model layers 2 and 4; leakance parameters for the other layers also could not be estimated) defined to estimate the vertical conductance between layers and one streambed conductance parameter defined to estimate the vertical conductance of the streambed materials. The recharge parameters consisted of four zones that coincided with the different surficial materials present in the model area because recharge is believed to vary, at least in part, based on the type of surficial material and four zones for the areas of increased recharge from Pfizer, Inc. The four parameters defining areas of increased recharge by local industry were not allowed to vary during model calibration because the values were provided by the local industry. All parameters initially were assigned a value of 1.0 as a multiplier of the matrix of initial estimates; these parameter values were adjusted during model calibration.

Model calibration progressed by first determining the set of parameter values that produced the best match between observed and simulated water levels and flow under 2001 conditions. Model fit was determined first for 2001 conditions because this calibration period contained the most water-level and streamflow-observation data. Parameters describing the hydraulic conductivity of model layers 1, 3, and 5; leakances of layers 2 and 4, streambed conductance, and the recharge parameters defining the outwash zones had the highest sensitivities and could be estimated by the automated parameter estimation process in MODFLOW-2000. After some initial parameter estimation runs, the outwash recharge parameter was divided into three separate parameters to allow more flexibility in adjusting the recharge rates in different parts of the model area. Likewise, the parameters representing the hydraulic conductivity of the upper and lower aquifers (model layers 1 and 5) was divided into separate zones; however, after inspection, the estimated values did not correspond with the estimates of hydraulic conductivities available from aquifer tests and this division was eliminated. The other parameters were modified manually within reasonable ranges to determine the best fit between observed and simulated values.

Final estimated parameters and the resulting ranges of model hydraulic properties are shown in table 8 and the spatial distribution of layer hydraulic conductivities is shown in figures 18, 19, and 20. Horizontal hydraulic conductivity of the upper aquifer (model layer 1) is highest along the Kalamazoo River near the eastern edge of Kalamazoo County and within the city of Kalamazoo over to the western edge of Kalamazoo County. Horizontal hydraulic conductivities of the intermediate and lower aquifers (model layers 3 and 5, respectively) are highest in the central portion of Kalamazoo County. Recharge is highest in the southern part of the model in the area with outwash surficial materials and lowest in the moraine areas. The resulting set of optimal parameter values were used then to determine model fit under 1994, 1987, and 1966 pumping conditions.

Simulation Results

Simulation results for 2001, 1994, 1987, and 1966 conditions were compared to observations from each calibration period. Observed and simulated ground-water levels were similar for each calibration period (fig. 21), with the majority of differences less than 10 ft. Additionally, the spatial distribution of residuals, the difference between the observed and simulated values, did not indicate any major patterns for 2001 pumping conditions (fig. 22). Observed streamflow values also were within the same order of magnitude as the simulated values for most streams for each calibration period (fig. 23). As expected, observed streamflows were higher than simulated

Table 8. Final parameter estimates and resulting values for model hydraulic properties, Kalamazoo County area, Michigan.

[Average values for horizontal hydraulic conductivities are in parentheses]

Model parameter	Initial value	Final value	Value of hydraulic property
Horizontal hydraulic conductivity (feet/day)			
Layer 1 (upper aquifer)	1	0.4	0.2-149 (88)
Layer 3 (intermediate aquifer)	1	.8	.2-347 (187)
Layer 5 (lower aquifer)	1	.5	.2-217 (106)
Recharge (cubic feet/day)			
Drainage channel	1	0.65	6.0
Moraine	1	.9	5.5
Outwash area to north	1	1.2	14.2
Outwash area to south	1	1.4	16.6
Outwash area to west	1	.5	5.9
Till	1	1	7.0
Streambed conductivity (feet/day)	1	2	50
Vertical hydraulic conductivity (feet/day)			
Layer 2 (upper confining unit)	1	2	0.01-42
Layer 4 (lower confining unit)	1	1	.01-4.3

flows for most streams. Miscellaneous streamflow-measurement data collected for this study was during a time of little or no precipitation; therefore, the surface-runoff component of streamflow is assumed to be small. These measurements likely are higher than actual base flow, so model-simulation results should be lower for these stations, as well as for the stations with streamflow data collected prior to this study.

Additional checks on model calibration consisted of comparing model-simulated and observed seepage and comparison of model-simulated and observed drawdown because of a pumping well in Ross Township. Model-simulated seepage compared favorably with the observed seepage rates collected during the August 2001 synoptic survey (fig. 24). Seepage rates measured at two streams were large and are not shown in the figure. Model simulated seepage rates were lower at these locations. Possibly, the seepage meters were placed in locations of unusually high seepage that likely are not representative of the rest of the stream reach. Seepage rates are expected to be variable because the streambeds are not composed of uniform, homogeneously porous materials.

Simulated drawdown in the cell containing the pumping well in Ross Township reached 30.5 ft in the intermediate aquifer (model layer 3) where the pumping well was located. Simulated drawdown in this area reached 19.3 ft in the lower aquifer (model layer 5) and 5.1 ft in the upper aquifer (model layer 1). The model-simulated drawdown represents the average water level in the cell in each layer; that is, the average of the lowered water levels at the well location and the higher water levels away from the well and beyond the cone of depression formed as a result of ground-water withdrawals. This model-simulated value is not the actual water level that would be observed in a well pumping in an area with the layer and hydraulic characteristics that are used in the model and, thus, cannot be compared directly with the actual observed drawdown at the pumping well of 33.5 ft (City of Kalamazoo, 1995). Drawdown at the location of the well in the cell would be greater than the simulated value and, thus, greater than the observed value. That the model-simulated drawdown is greater than the observed value can be explained at least, in part, in that grid spacing is large in this part of the model area. The cell containing the test well is 2,471 ft by 2,081 ft. The test in Ross Township was conducted in an area that is away from municipal or other large ground-water withdrawals and, therefore, is not in the part of the model area with the finer grid spacing. As such, aquifer properties and layer surfaces are averaged based on all information from wells in this large area. Layer thickness for the intermediate aquifer in this cell is 12 ft, whereas actual thickness at the specific well site is 52 ft. This reduced cell thickness in the model means a lower simulated transmissivity and, thus, a greater area of influence and a deeper cone of depression from the pumping well. Reducing the grid spacing and incorporating more local information on aquifer and confining unit thicknesses would improve the simulation of withdrawals in this area in Ross Township.

Model Sensitivity Analysis

The calibrated model is influenced by uncertainty because model parameters are averaged and are based on limited data available from well drillers' logs or from previous studies. A sensitivity analysis can help to determine the effect of uncertainty on the calibrated model results because of estimated aquifer parameters, stresses, and boundary conditions. During a sensitivity analysis, calibrated values for model parameters are systematically changed within reasonable ranges to determine the magnitude of the changes in water levels from the calibrated solution (Anderson and Woessner, 1992). Model sensitivity to changes in the hydraulic parameters for the aquifers and streambed was determined and is described in this section. Changes in water levels and streamflow from the calibrated solution because of changes in recharge and ground-water withdrawal rates are described in subsequent sections. The sensitivity of boundary conditions also is investigated later in the report.

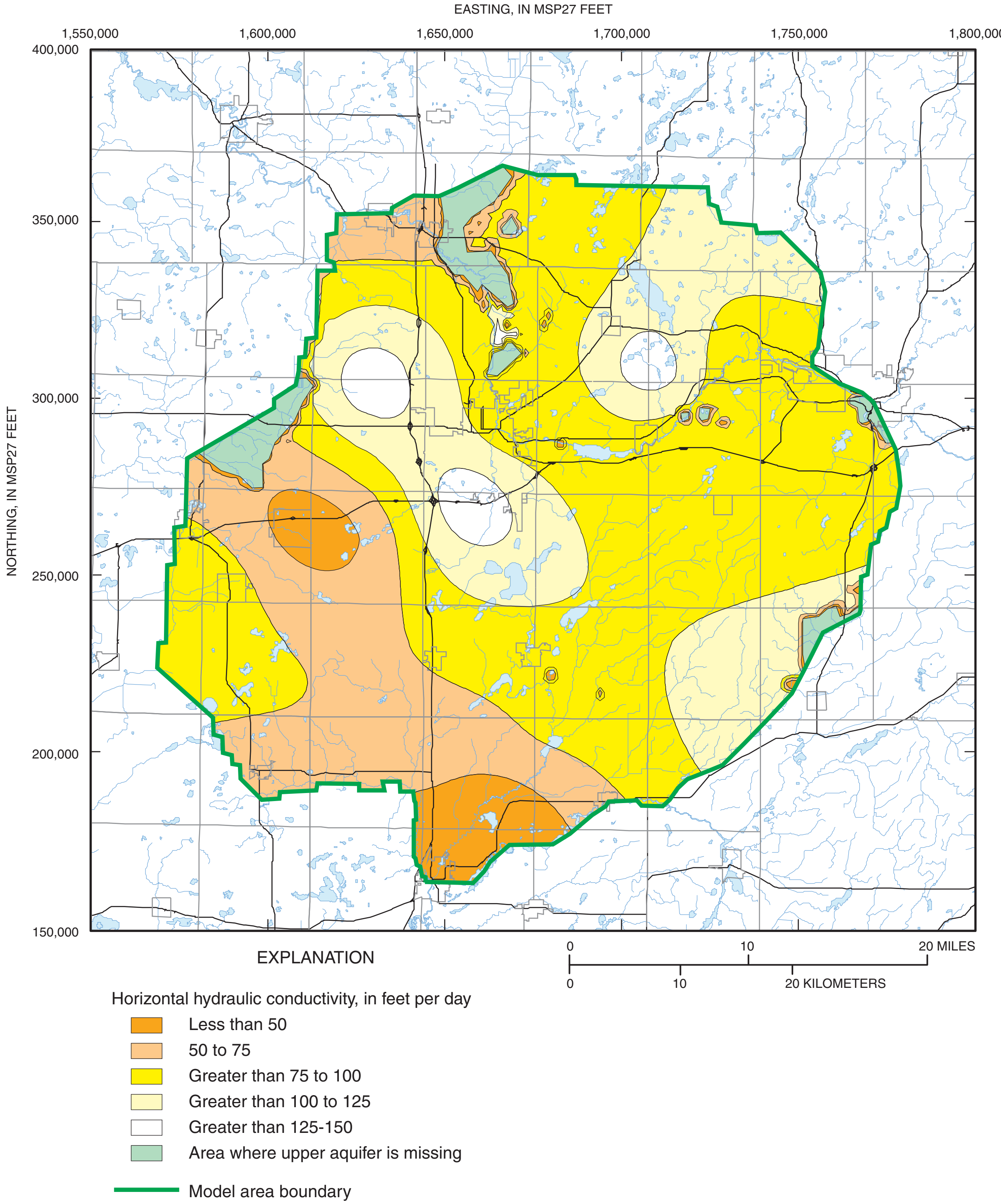


Figure 18. Estimated horizontal hydraulic conductivity of the upper aquifer (model layer 1) determined by parameter estimation, Kalamazoo County area, Michigan.

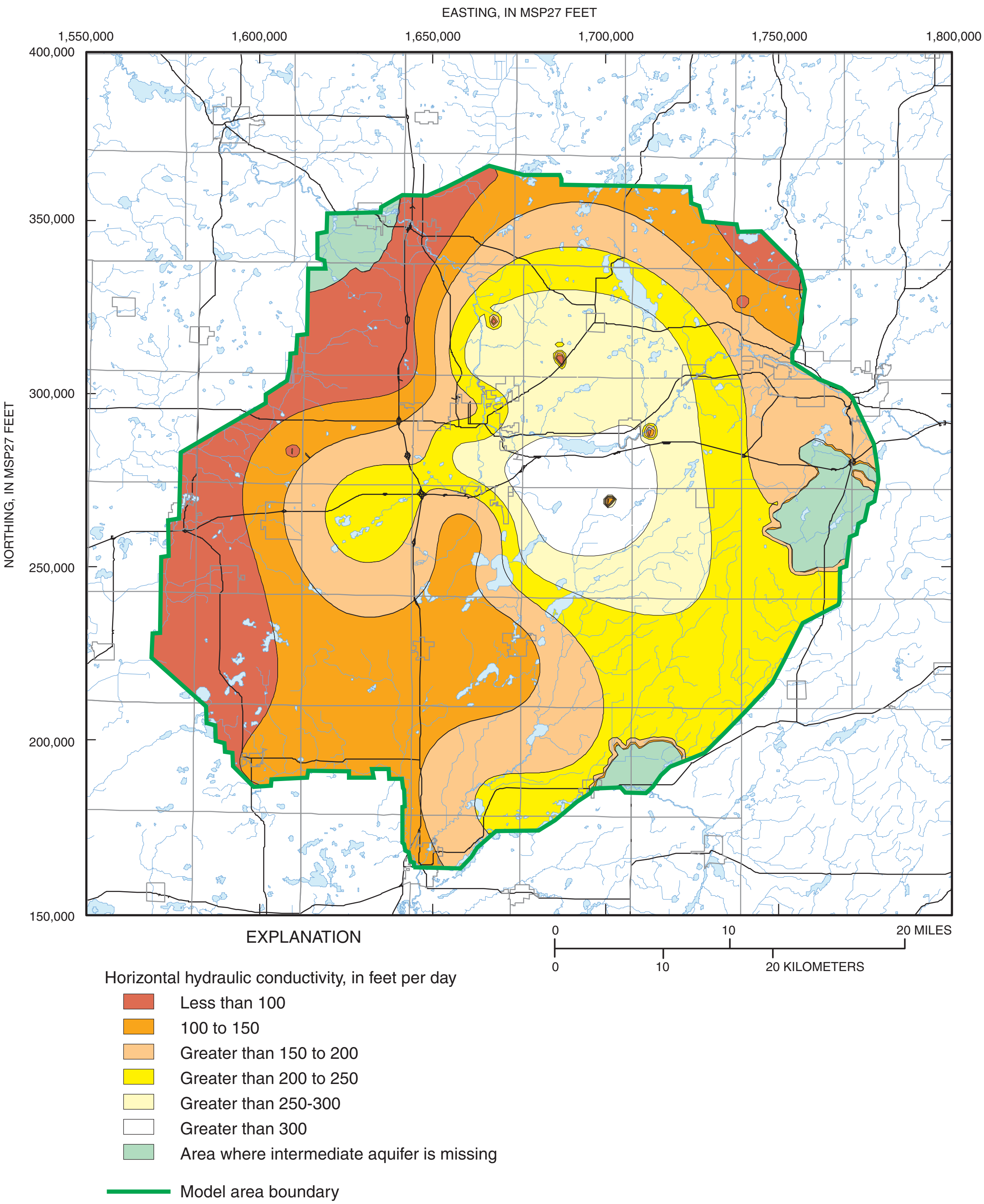


Figure 19. Estimated horizontal hydraulic conductivity of the intermediate aquifer (model layer 3) determined by parameter estimation, Kalamazoo County area, Michigan.

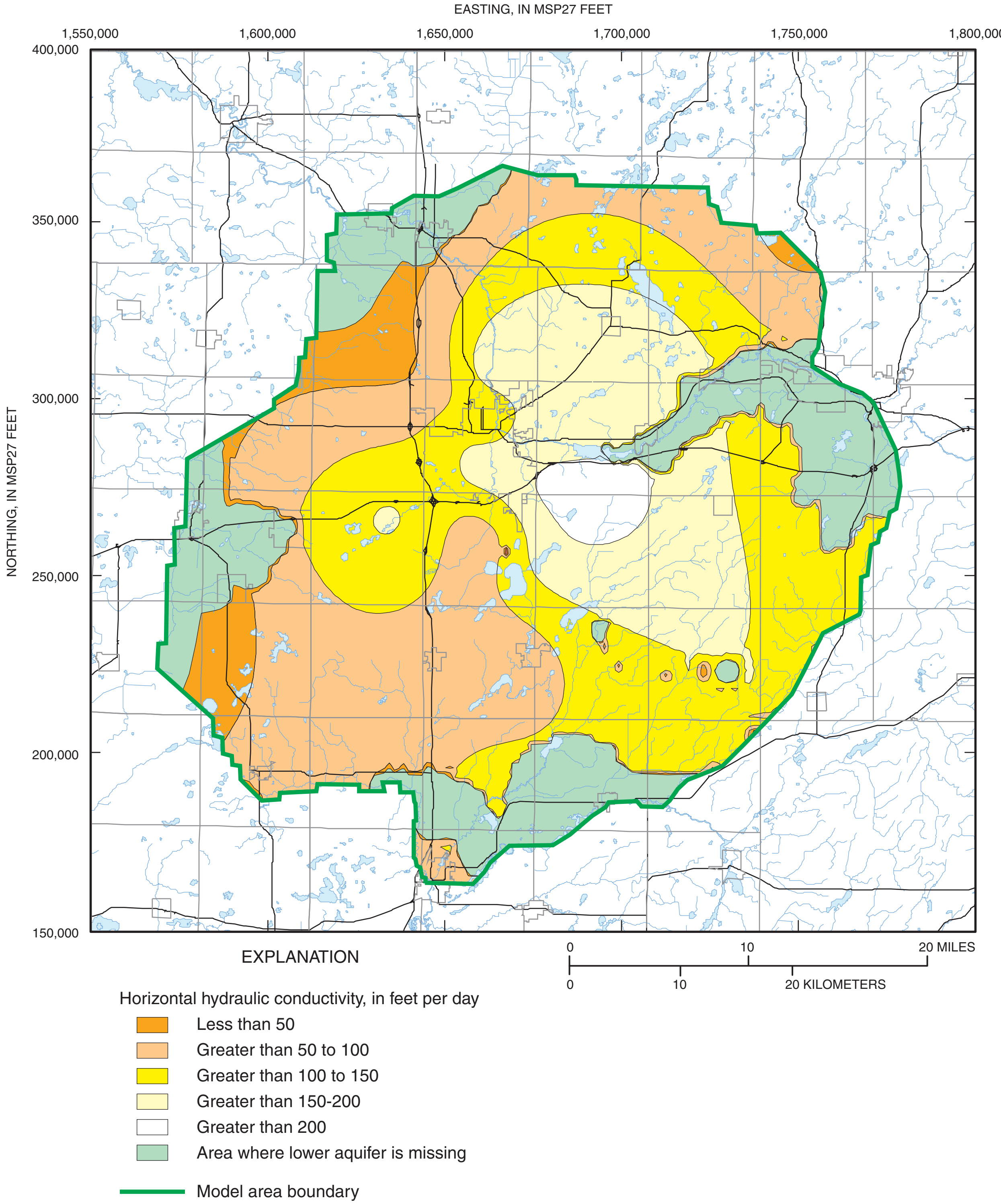


Figure 20. Estimated horizontal hydraulic conductivity of the lower aquifer (model layer 5) determined by parameter estimation, Kalamazoo County area, Michigan.

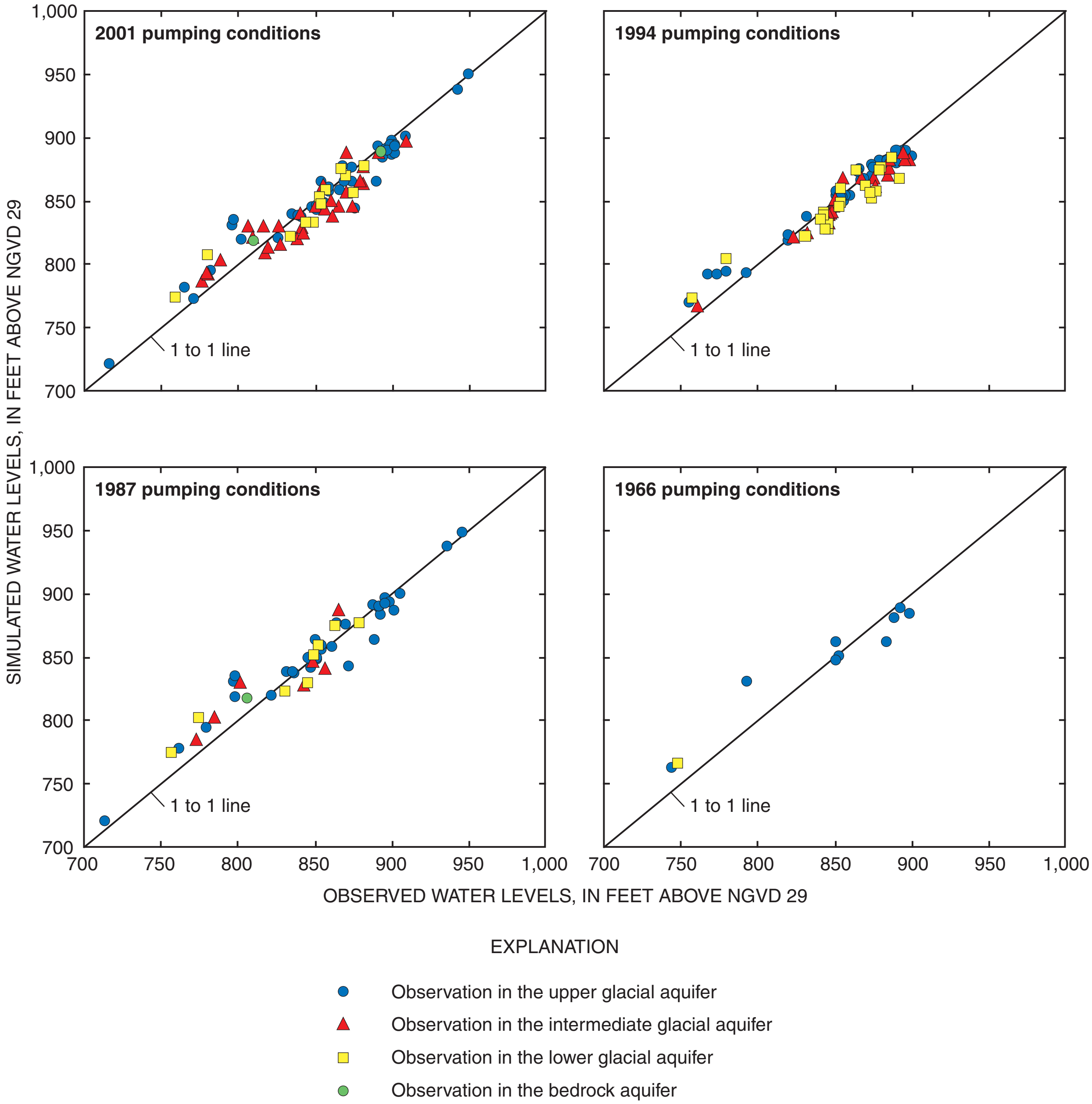


Figure 21. Observed and simulated water levels in wells for 2001, 1994, 1987, and 1966 pumping conditions, Kalamazoo County area, Michigan.

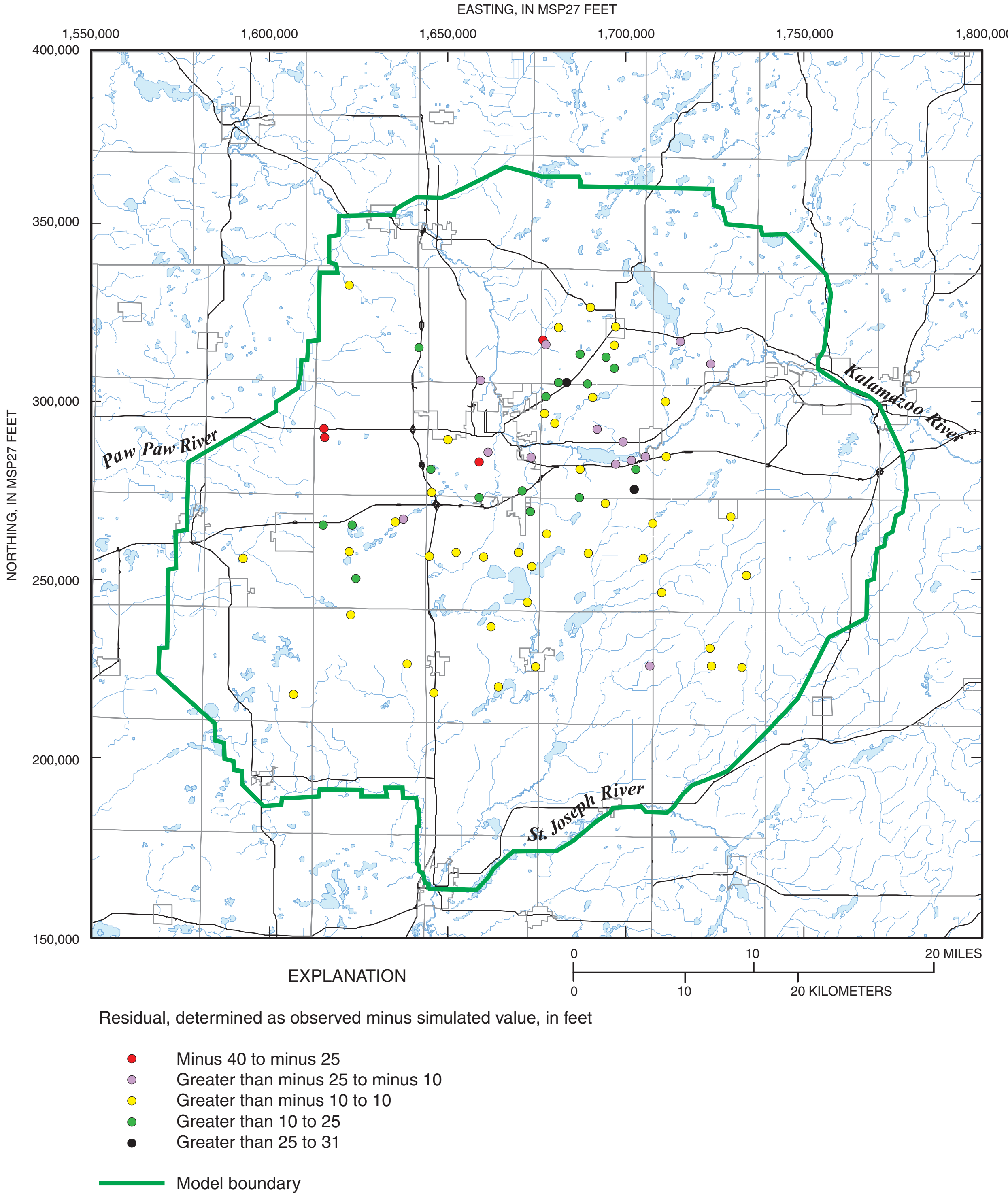


Figure 22. Distribution of water-level residuals for 2001 pumping conditions, Kalamazoo County area, Michigan.

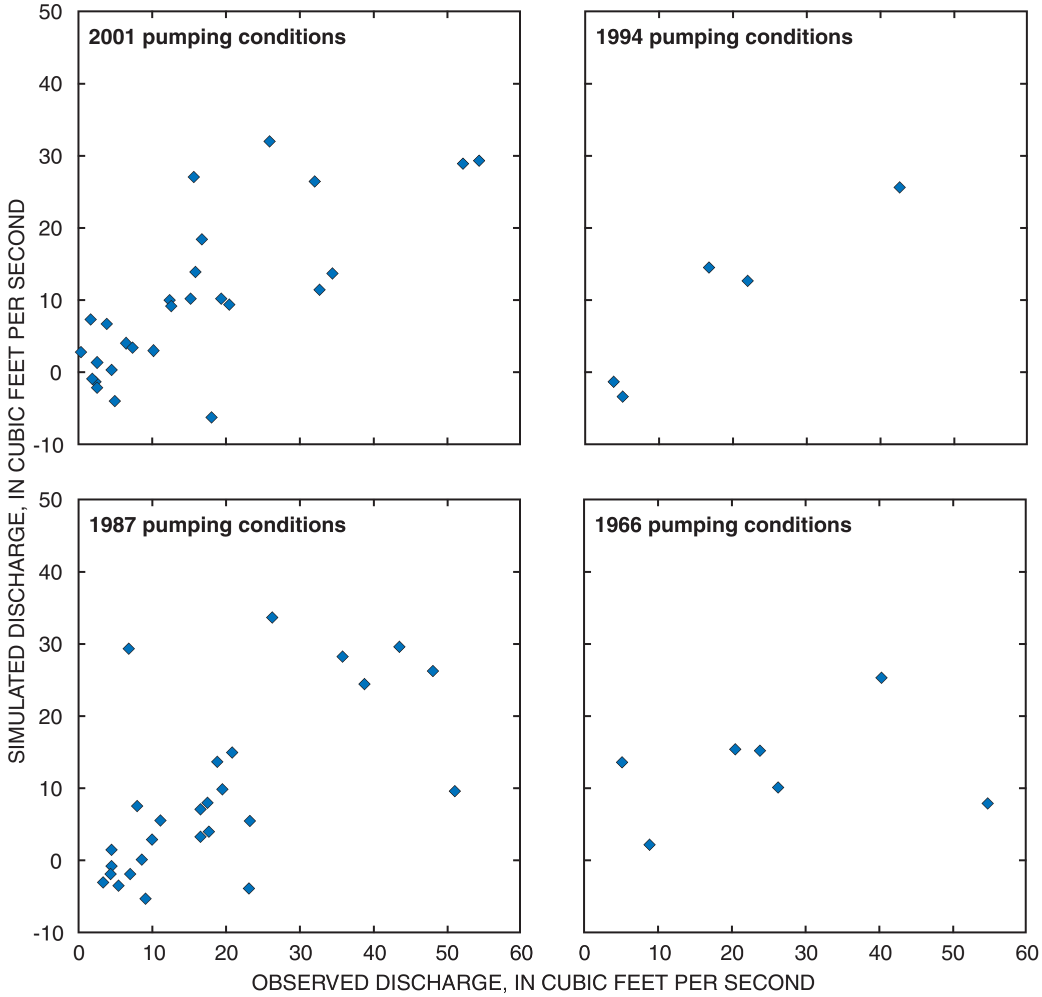


Figure 23. Observed and simulated streamflow for 2001, 1994, 1987, and 1966 pumping conditions, Kalamazoo County area, Michigan.

Hydraulic conductivity parameters for the aquifer layers (model layers 1, 3, and 5), leakance parameters for the confining units (model layers 2 and 4), and the streambed conductance parameter were varied in order to determine the resulting changes in the root mean square error (RMSE) (fig. 25). The RMSE is the square root of the average of the squared differences in simulated and observed water levels. The RMSE was determined for the model and for the individual aquifer layers because parameter changes could result in an overall lower RMSE; however, the resulting change in water levels in an individual layer may be unacceptable. The model was most sensitive to changes in the hydraulic conductivity parameters and least sensitive to changes in the streambed parameter. The RMSE generally is lower for lower values of the hydraulic conductivity parameters for the intermediate and the lower aquifers (model layers 3 and 5) and for higher values of the leakance for the upper confining unit (model layer 2). The RMSE generally is higher for lower values of the hydraulic conductivity parameter for the upper aquifer (model layer 1), for higher values of the hydraulic conductivity parameters for the intermediate and lower aquifers (model layers 3 and 5), and for lower values for the leakance parameter for the upper confining unit (model layer 2). Model fit was determined based on this calibration criteria and comparison of observed and simulated water levels and streamflow throughout the model area for each measurement location. Based on all calibration criteria and measures of model fit, the present regional model reasonably approximates the magnitude and trends observed in water levels in the modeled area.

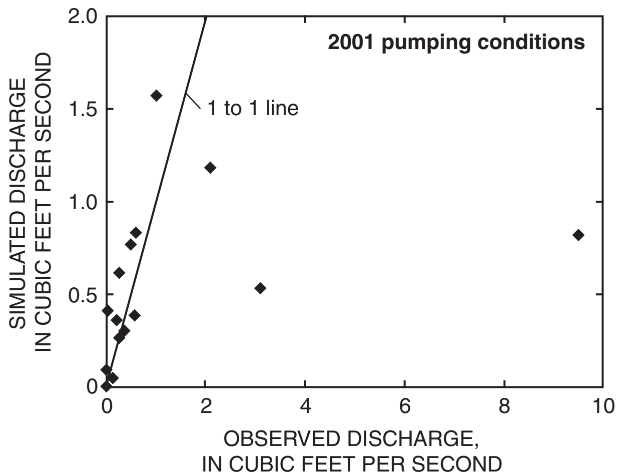


Figure 24. Observed and simulated flow at seepage measurement sites for 2001 pumping conditions, Kalamazoo County area, Michigan.

Transient Calibration

For the steady-state simulations, measured water levels and flows were assumed to represent long-term average conditions. However, true steady-state conditions do not occur; as mentioned previously, water levels, flows, and recharge rates were known to vary from 1966 to 2001. Model fit to each of the different time periods under steady-state conditions provides one check on the model capacity to represent changing ground-water withdrawals and stream discharges. With simulated water levels and flows matching observed values for 2001, 1994, 1987, and 1966 conditions, the steady-state model is shown to represent the long-term average values used for model calibration. However, recharge rates and ground-water withdrawals vary during the year and this variation is not represented in the steady-state calibrations. Therefore, the steady-state model was calibrated to transient conditions representing both seasonal and monthly changes to confirm that the model reasonably approximates the observed variations in the ground-water-flow system underlying the Kalamazoo area.

Two different transient simulations using varying pumping rates with three different recharge conditions each were performed to check model responses to changes in stresses. The first transient simulation involved investigating model fit to seasonal winter conditions from October 2001 to May 2002 and summer from June 2002 to September 2002. These season divisions were selected on the basis of ground-water-withdrawal data showing increased municipal withdrawals for June to September and lower withdrawals for the other months of the year. The time from June to September also is a time of expected increased potential evapotranspiration that could limit recharge to ground water. The time period from October 2001 to September 2002 was selected to coincide with data available from synoptic surveys conducted during this study. Average ground-water withdrawals for winter and summer were based on available municipal data. Water-level and flow observations were averaged from available data during each season.

The recharge conditions consisted of natural recharge remaining constant in the first simulation, distributed throughout the year in the second simulation, and distributed throughout the winter months only in the third simulation. Winter and summer recharge rates were based on average precipitation rates observed for 1997–2001 for each season. Water from precipitation may recharge surface-water bodies, may evaporate or be taken up by plants, or may recharge the ground-water system. In this study, ground-water recharge is assumed to be some percentage of precipitation and to vary along with monthly or seasonal precipitation amounts. Average recharge for the regional model area is less than precipitation; however, local recharge rates may approach, or exceed, or be appreciably less than the average precipitation rate. Because of a lack of site-specific information on local or regional ground-water recharge rates, recharge for model simulations is assumed to vary temporally based on the observed variations in precipitation rates. The average natural recharge rate determined during the steady-state calibration was distributed throughout the year using the percentage of observed rainfall during each month. An average recharge rate then was determined for winter and summer and used to modify the steady-state rates for the winter and summer stress periods. Some recharge occurred during the summer for the last simulation in local recharge areas. This recharge was assumed to remain at the simulated rate and be unaffected by evapotranspiration.

The second transient simulation involved investigating model fit to monthly changes from October 2001 to September 2002. Ground-water withdrawals were varied monthly based on available municipal data. Water-level and flow observations were specified for the stress periods coinciding with the dates of the synoptics. Recharge again was assumed to remain constant in the first simulation, to be distributed throughout the year in the second simulation, and to be distributed throughout the winter months only in the third simulation. Monthly recharge rates were based on average precipitation rates observed for 1997–2001 for each month. As for the seasonal simulation, some recharge is simulated in the model during the summer months in local recharge areas. This recharge was assumed to remain at the simulated rates and be unaffected by evapotranspiration.

For the transient model simulations, additional information must be specified, including the storage characteristics of the aquifers, the initial conditions, and the time dimension. Storativity describes the capacity of an aquifer to transfer water to and from storage. For a confined aquifer, the storage coefficient is equal to the volume of water released per unit area of aquifer per unit decline in head. For an unconfined

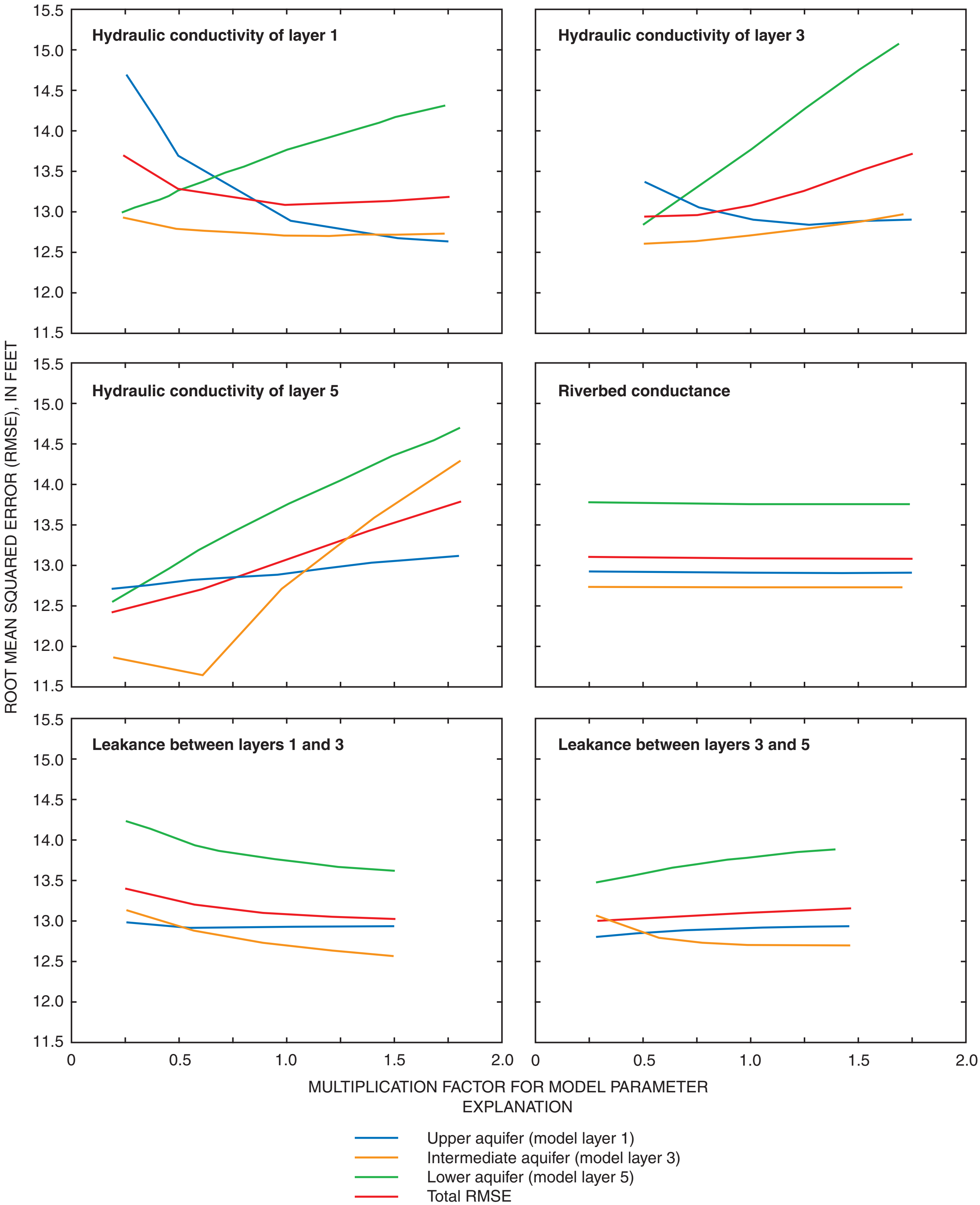


Figure 25. Sensitivity of water levels to changes in hydraulic conductivity and streambed parameters, Kalamazoo regional model.

aquifer, the equivalent term is specific yield that is defined as the volume of water released per unit surface area of aquifer per unit decline in the water table. The initial conditions provide the head distribution at the beginning of the simulation and provide a boundary condition in time. The stress periods and time steps specify the length of time used in determining the ground-water-flow solution and when selected stresses, such as ground-water withdrawals and recharge rates, can change during the simulation. Details of the development and calibration of the transient models are explained below.

Aquifer Storage Characteristics

Some initial estimates of the aquifer storage characteristics are available from the aquifer tests conducted in the county (table 1). Estimated values range from 0.02 to 0.2 for the upper unconfined aquifer (represented by layer 1). Estimated values range from 0.17 to 0.000054 for the intermediate aquifer (represented by layer 3). Typically, storativities range from 0.005 to 0.00005 for confined aquifers; thus, the higher values for the intermediate aquifer likely indicate areas where this aquifer is unconfined or possible problems with data analyses. The specific yields of unconfined aquifers typically are higher than the storativities of confined aquifers. This results because releases from storage in unconfined aquifers represent an actual dewatering of soil pores, whereas releases from storage in confined aquifers represent only the secondary effects of water expansion and aquifer compaction caused by changes in fluid pressure (Freeze and Cherry, 1979). In the model, the specific yield of the upper aquifer (model layer 1) was estimated to be 0.1 and the specific storage values of the intermediate and lower aquifers (model layers 3 and 5) were estimated to be 0.002 and 0.00001, respectively.

Initial Conditions

The initial conditions for the transient model were specified to be the water levels resulting from the steady-state model simulating 2001 conditions. This initial water-level distribution is known as dynamic average steady-state condition in which water levels vary spatially and flow into the ground-water system equals flow out of the ground-water system. Use of this water-level distribution ensures that the initial water-level data and the model hydraulic properties and parameters are consistent (Franke and others, 1987).

Temporal Discretization

Transient calibration requires specifying how the simulation period will be divided into time periods. This division involves specifying the number and length of stress periods within the simulation, and the number of time steps within each stress period. The use of stress periods offers the option of changing some of the parameters or stresses while the simulation is in progress (Anderson and Woessner, 1992). For the first transient simulation, three stress periods were specified to represent the initial steady-state conditions, winter conditions (October to May), and summer conditions (June to September). The effect of stress-period length on the solution and the movement of water into and out of storage was investigated for winter stress period lengths of 243 days, 1 year, 2 years, and 3 years and a summer stress period length of 122 days. The recharge and withdrawal conditions specified for the winter stress period were allowed to continue for the actual number of days from October to May or as long as 3 years; thus, allowing investigation of the amount of time needed for the movement of water into and out of storage to come to equilibrium. For the second transient simulation, 1 stress period was specified to represent the initial steady-state conditions, and the remaining 12 stress periods representing each month from October to September with stress-period length equaling the number of days in each month.

Selection of the time step is important because the values of time discretization strongly affect the numerical results. Ideally, it is desirable to use small nodal spacing and small time steps so that the numerical representation better approximates the ground-water-flow equation (Anderson and Woessner, 1992). Time steps can be increased as the simulation progresses. The ground-water-flow solution is sensitive to rapidly fluctuating water levels caused by introducing a stress; thus, small time steps allow the early response of the system to be captured. Even with a large time step the solution becomes more accurate as steady-state is reached. Instead of using small time steps when a stress is introduced, only solutions from later time steps could be used (Anderson and Woessner, 1992). As a rule of thumb, the solution should proceed through five time steps, during which there are no appreciable changes in values of sources and sinks or boundary conditions, before the solution is considered accurate (de Marsily, 1986).

To determine the effect of time-step length, results with a time-step length of 3.5 days was compared with those using a time-step length of 14 days for the transient calibration to seasonal conditions. Because little difference was seen in simulated water levels and storage, a time-step length of 14 days was used for the seasonal simulations. For each stress period in the simulation representing monthly conditions, changing time-step lengths were used with 15 time steps for each month except February for which 14 time steps were used. A multiplier of 2 allowed the initial time steps in each month to be small, whereas later time steps were longer. For comparison of model results to observations, the ground-water-flow solution for the last time step of each stress period was used.

Simulation Results

Simulated water levels under each recharge condition for the transient simulations of seasonal and monthly conditions were compared to observed water levels. Observed ground-water levels in the upper aquifer generally were higher in the winter than in the summer. These observations indicate that possibly more water is available to recharge the ground-water system during the fall, winter, and spring allowing water levels to rise; during the summer months as vegetation becomes active and intercepts potential recharge, water levels decrease as recharge is reduced. Wells in the intermediate and lower aquifers generally also displayed the trend of higher water levels in the winter and lower in the summer even though recharge from precipitation likely occurs slowly for these wells. The more immediate effect on water levels for wells in these aquifers is from nearby pumping; however, because pumping generally is higher in the summer and lower in the winter, water levels in the intermediate and lower aquifers follow the same trend with higher levels in the winter and lower levels in the summer.

Selected wells from each aquifer are shown to illustrate the fit between observed and simulated water levels for the transient simulations of seasonal and monthly conditions (figs. 26 and 27). Wells K-17 and K-12 are located in the upper aquifer, wells K-1 and Kendall are located in the intermediate aquifer, and wells Colony and Prairie are located in the lower aquifer. Wells Kendall, Colony, and Prairie likely are affected by local pumping, whereas K-17, K-12 and K-1 likely are unaffected by pumping. Well K-1 was not measured during the last synoptic survey, so the observed and simulated values are missing for the last time step.

Simulated water levels from each transient simulation compared favorably to observed water levels collected during synoptic surveys for some wells and to the trends observed in water levels for other wells. Transient simulations in which natural ground-water recharge occurred only during the winter season (which runs from October until May in the model simulation) provided the best fit to observed trends in water levels for most wells. However, for some wells, simulation results under the two other recharge conditions provided a closer match to observed water levels than the simulation in which natural ground-water recharge occurred only from October until May. In areas of municipal pumping, such as Kendall and Colony, simulated water levels do not necessarily match the trends in observed values under any recharge condition. Pumping rates were averaged for each season or month in model simulation and may not represent actual high and low pumping rates. Also, some wells show the trend displayed in K-1 with low levels in December, increasing levels in the spring, and low levels in September. In these cases, the simulation with recharge distributed throughout the year matches the trend of lower levels in December and the simulation with no natural recharge in the summer generally matches the increase and decrease in water levels observed in spring and summer.

Precipitation generally was lower in December 2001 than the overall average value and may help to explain the trend in measurements observed at some of these wells, such as K-1. Comparison of simulated water levels for well K-1 indicates little difference between the different recharge conditions (fig. 27). K-1 is located in a thick morainal sequence along the western edge of Kalamazoo County (fig. 2). Other wells within this area and within the till areas along the southeastern part of the model area also have simulated water levels that vary little with changing recharge condition. Water levels in wells within the areas of surficial outwash materials do vary during model simulation depending on recharge conditions. This variation results because the moraines and till features generally have lower recharge rates and tighter surficial materials than the outwash areas.

Results from the simulations in which recharge did not occur during the summer season (which runs from June until September in model simulation) provided a better fit to the trends in observed water levels in most cases seems to contradict results from Machavaram and Krishnamurthy (1995). Their research suggests that ground water in this area actually is a mixture of summer and winter recharge based on analysis of del O^{18} data. Based on this research, the model simulation with recharge occurring throughout the year should provide the best fit to observed water levels. It is likely that not all potential recharge in the summer is intercepted by vegetation as was simulated and that some water moves beyond the root zone, especially during high precipitation events, to recharge the ground-water system; however, it also is likely that actual recharge in the summer is reduced from the potential amount that would be available in the absence of vegetation. Additional information on local recharge areas and actual timing of recharge could improve model simulation and the representation of water levels in this area; however, these simulation results indicate that the present regional model reasonably approximates the trends observed in water levels in the modeled area.

Model Assumptions and Limitations

The ground-water-flow model was developed to simulate the regional ground-water-flow system in the Kalamazoo County area. Hydraulic properties in the aquifers were assumed to be isotropic. (Within a cell, hydraulic properties are the same in the north-south direction as in the east-west direction; hydraulic properties vary from location to location). Each grid cell represents the average hydrologic and hydraulic properties in the volume of aquifer represented by the cell. Vertical variations in aquifer properties within layers and any variations in head or flow within the aquifers are not simulated in the model. Local flows over distances smaller than the dimensions of the grid cell also cannot be accurately simulated. Additional geologic and hydrologic data, as well as finer discretization of the model, would be needed to simulate smaller-scale flow systems. The accuracy of layer surfaces and hydraulic conductivity estimates are limited by the available data at well and boring locations. Additional control and accuracy could be achieved by inclusion of more data points. Inclusion of available information on the location and extent of tunnel valleys may improve the representation of the ground-water-flow system in the model area. As indicated by the model simulation, some refinement of the model is needed to better simulate the effects of withdrawals in the area of the test well in Ross Township.

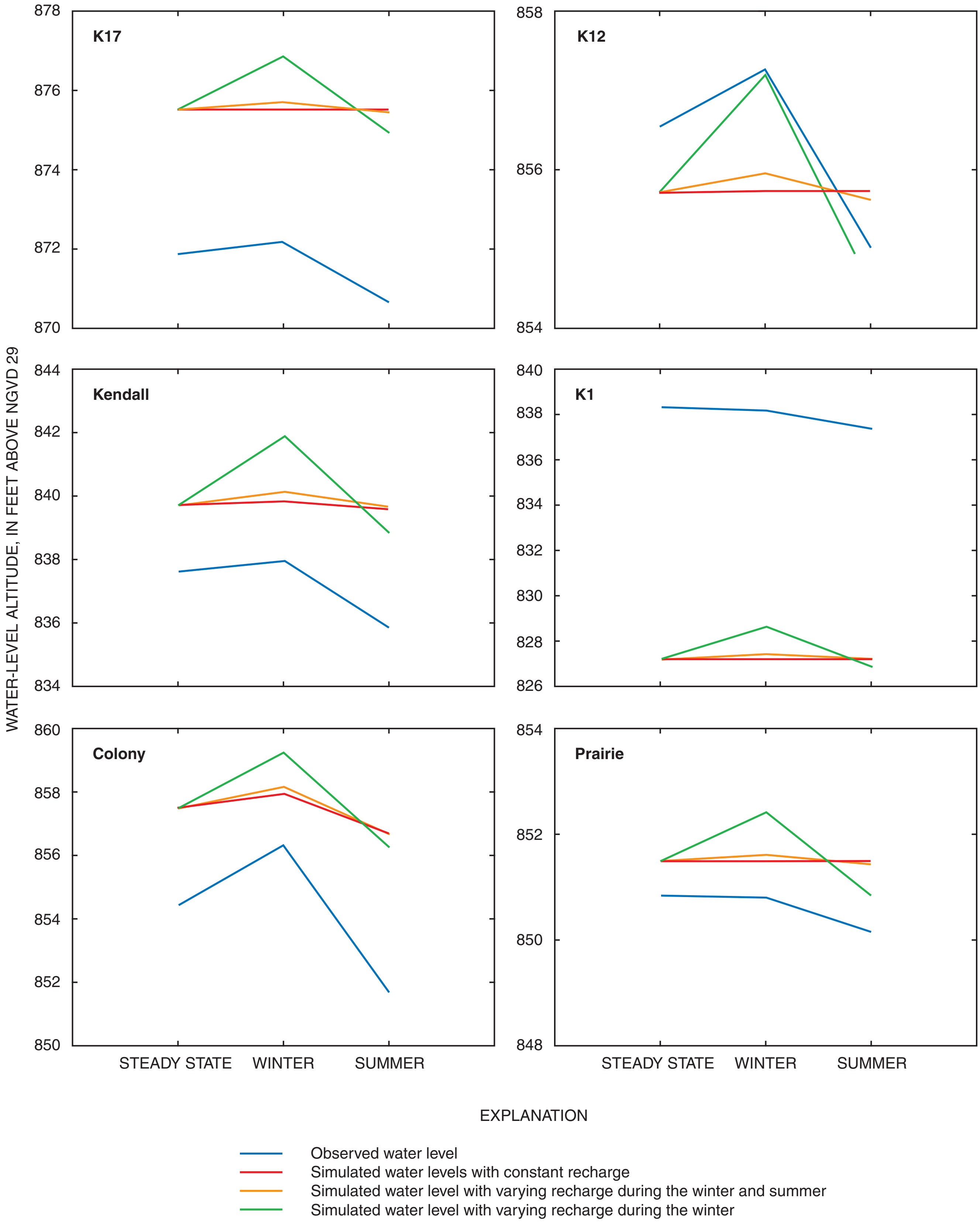


Figure 26. Water levels in selected wells of the transient model simulation representing seasonal conditions, Kalamazoo County area, Michigan. (See figure 5 for location of wells.)

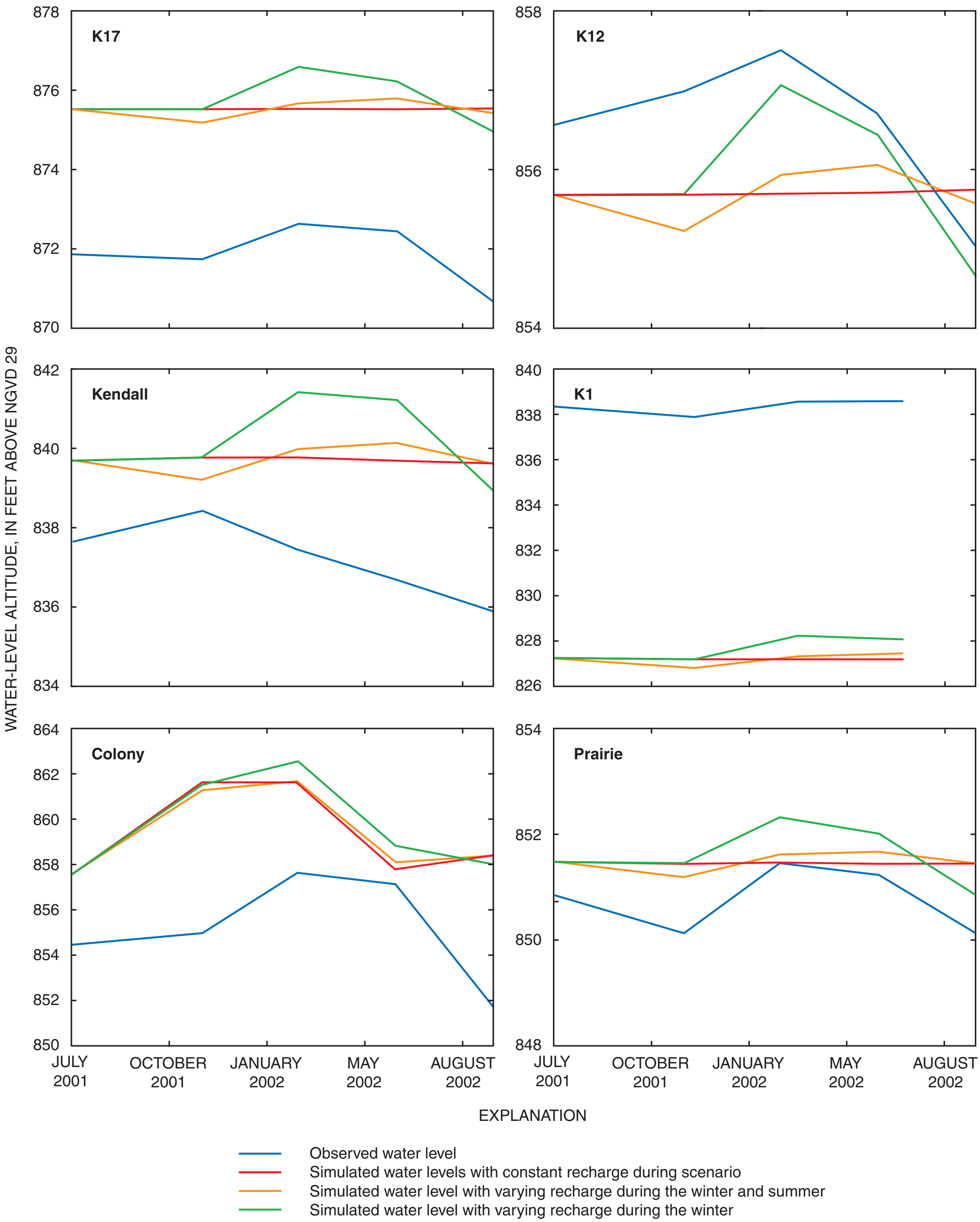


Figure 27. Water levels in selected wells of the transient model simulation representing monthly conditions, Kalamazoo County area, Michigan. (See figure 5 for location of wells.)

It is assumed in steady-state model simulations that all stresses within and inputs to the system, including well withdrawals and recharge rates, remain constant throughout the simulation. No net gain or loss of flow is simulated in the system and no changes in ground-water storage result. It is assumed in the transient simulations that the storage characteristics of the aquifers can be represented as specified by an average value that is constant for each layer. The actual storage properties of the aquifers likely vary from location to location. It is assumed in transient simulations that generalized values of monthly and seasonal recharge vary based on available precipitation data. Actual monthly or seasonal recharge rates were not measured and may vary both regionally and locally from those used in transient simulations. Estimates of withdrawals for irrigation (described later) also were generalized because of a lack of site-specific information; furthermore, areas representing as being irrigated in the model simulation were based on visible observation (actual irrigated areas may be larger or smaller than those observed). Therefore, represented changes in water levels because of irrigation represent only a general approximation. Likewise, effects because of urbanization (described later) would require more detailed study to determine which, and to what extent, urban areas actually reduce recharge.

Recharge to ground water was assumed to vary over the model area depending on the type of surficial material present; thus, local variations in recharge rates, such as those associated with impermeable surfaces or local differences in the surficial materials, are not simulated in the model. Simulated well withdrawals are assumed to come from the centers of grid cells. Small withdrawals from domestic wells were not included because of the difficulty in obtaining reliable data and the limitations in representing small-scale flow systems (systems considerably smaller than simulated as part of this study). However, domestic ground-water withdrawals probably are small at the scale of the model. Streams and lakes are represented in the model as river cells. This type of boundary condition provides for determining the amount of flow to and from the river cell; however, the amount of water flowing into the cell from an upstream cell is not accounted for. Therefore, a river cell could lose more water than actually is flowing in the stream. Thus, for detailed analysis of flow within particular streams, an accounting of actual flow within the stream needs to be part of the simulation.

The base of the model is assumed to be impermeable. This assumption is considered adequate for model development because of the limited flow available from the Coldwater Shale. External boundary conditions, based on natural hydrologic conditions and distant from Kalamazoo’s well fields, are assumed to have minimal effect on water levels and flow in the interior portion of the model. Enlargement of the model area to natural physical hydrologic boundaries might improve model simulations. The model may not accurately represent the ground-water-flow system for any predictive simulations involving ground-water withdrawals near the model boundaries.

Simulation of Potential Effects of Urbanization, Irrigation, and Changes in Withdrawals and Recharge

As mentioned previously, ground-water recharge may be affected by urbanization and irrigation. Three additional simulations were conducted to determine whether either of these factors affected model results for 2001 conditions. Urbanization could cause a possible reduction of recharge because of impervious surfaces or because of areas with residential wells and municipal sewers. Irrigation could cause a reduction in recharge because of withdrawals and consumption by crops or a local increase in recharge if excess water was added to crop areas.

Model simulations using both the steady-state and the transient models also were conducted to investigate the response of the ground-water-flow system in the Kalamazoo area to changes in stresses, including withdrawals and recharge. The steady-state model represents the long-term average response of the ground-water-flow system to the given set of input conditions. Thus, if withdrawals remained constant and recharge decreased, the steady-state model results would indicate the potential effects of reduced recharge on water levels and streamflow. The transient model was used to simulate water levels and flow conditions that result from temporary increases or decreases in withdrawals and recharge that occur over seasonal or monthly time scales.

Potential Effects of Urbanization

Urbanization could cause a possible reduction of recharge resulting from impervious surfaces or from areas with residential wells and municipal sewers. Urban areas that are connected to storm sewers that discharge the water to surface-water features likely cause a reduction in recharge; however, urban areas, such as rooftops and paved areas, that drain to lawns likely do not cause a reduction in recharge. Available data on urban areas for Kalamazoo, Portage, Parchment, Schoolcraft, and Vicksburg were used to define areas with potentially lower recharge than other areas. Within the city of Kalamazoo, approximately 75 percent of the urban areas are connected to storm sewers. These areas were assigned a recharge value of 25 percent of the recharge rate in the steady-state model simulating 2001 conditions. Detailed information on storm-sewer coverage or impervious areas was not available for Portage, Parchment, Schoolcraft, or Vicksburg. These areas were assigned a recharge value of 90 percent of the recharge rate in the steady-state model simulating 2001 conditions; a value that likely underestimates the actual reduction in recharge. Thus, in model simulation, it is assumed that recharge is reduced by 75 percent in the city of Kalamazoo urban areas and by 10 percent in the other urban areas. Because of limited data on areas with residential wells and municipal sewers, this potential factor that could limit recharge was not included in model simulation. With additional data, model simulations could predict potential effects on water levels of residential areas where water is withdrawn and released to municipal sewers.

Simulation results indicate that the assumed reduction in recharge because of urbanization caused a reduction of about 4.3 ft in water levels in the upper aquifer, about 4 ft in the intermediate aquifer, and about 2.1 ft in the lower aquifer in the city of Kalamazoo urban area. With the reduction in recharge and the steady-state assumption of no change in the amount of water in storage, the source of water to wells in the area must come from increased inflow from areas outside the urban areas or from decreased discharge to streams, lakes, or wetlands. In this simulation with reduced recharge and 2001 pumping rates, there is a reduction in the amount of water available for local streams and lakes (table 9). Thus, local streamflow, as well as local water levels, are affected by a reduction in recharge. A more detailed analysis of the nature of the urban areas in these cities as well as a simulation that accounts for changes in storage would provide more information on the potential effects of the reduction in recharge because of urbanization.

Potential Effects of Irrigation

Irrigation could cause a reduction in water levels because of withdrawals and consumption by crops or a local increase in recharge if excess water was added to crop areas. Because of the difficulty in obtaining estimates of withdrawals or application rates for individual farms, well logs and visible observation of irrigation equipment were used to determine irrigation areas. Approximately 200 irrigation well logs with

Table 9. Change in budget components for steady-state model simulation with reduced recharge in urban areas, Kalamazoo County area, Michigan.

Budget component	Steady-state simulation representing 2001 conditions (cubic feet per day)	Steady-state 2001 simulation with reduced recharge in urban areas (cubic feet per day)	Percent difference [(2001 conditions minus 2001 conditions with reduced recharge) divided by 2001 conditions times 100]
Kalamazoo urban area			
IN			
Boundaries	0	0	0.0
Rivers	3,713,900	4,384,500	-18.1
Recharge	2,030,800	507,710	75.0
Rest of model area	5,002,400	5,201,100	-4.0
OUT			
Boundaries	0	0	.0
Wells	1,729,400	1,729,400	.0
Rivers	6,055,200	5,537,300	8.6
Rest of model area	2,642,400	2,521,800	4.6
Rest of model area			
IN			
Boundaries	799,500	799,500	0.0
Rivers	36,943,000	37,139,000	-.5
Recharge	72,370,000	72,239,000	.2
Kalamazoo urban area	2,642,400	2,521,800	4.6
OUT			
Boundaries	8,058,400	8,058,400	.0
Wells	6,307,400	6,307,400	.0
Rivers	90,898,000	90,738,000	.2
Kalamazoo urban area	5,002,400	5,201,100	-4.0

installation dates ranging from 1962 to 2000, from a total of 523, were available within the parts of Kalamazoo, Van Buren, and St. Joseph Counties within the model area. For each county, the percentage of irrigation wells within the model area was determined. An amount of water withdrawn for irrigation for each well in each county was determined using the total amount of ground water withdrawn for irrigation within the county for 2001 and was applied to wells located within the model area. All irrigation wells likely are not included in model simulations and withdrawals likely are not equally distributed as is assumed in model simulations; however, this representation of irrigation withdrawals is assumed to be representative of the potential regional effects of irrigation on water levels. Estimated withdrawals equaled 5,997 ft^3/d for each irrigation well in Kalamazoo County, 13,184 ft^3/d for each irrigation well in St. Joseph County, and 8,958 ft^3/d for each irrigation well in Van Buren County. These withdrawals were assumed to come from the upper aquifer (model layer 1). Observed areas of irrigation were noted on county maps and used within the model to determine areas with adjusted recharge. Areas determined to be irrigated based on personal inspection totaled approximately 13,700 acres in Kalamazoo County and 8,700 acres in St. Joseph County. During trips by USGS personnel within the county during this study, no observations were made of visible irrigation in the part of Van Buren County in the model area; however, based on well logs, nine irrigation wells are located in Van Buren County in the model area. Large irrigated areas are located primarily within the model area in the southern part of Kalamazoo County and the northern part of St. Joseph County.

Two model simulations were developed to investigate potential effects of irrigation. Using the steady-state model simulating 2001 pumping conditions, estimated ground-water withdrawals were added for each irrigation well within the model area as specified above and, within observed irrigation areas, the estimated ground-water recharge rate was reduced to zero. This simulation presents a different recharge condition. The model developed and calibrated to 2001 conditions already included the effects of evapotranspiration (and consumption of water by crops) by having recharge equal to that portion that has bypassed the root zone and entered the ground-water system. It is assumed with this reduction in recharge rate for the irrigation simulation that all water available to recharge the ground-water system from precipitation and from irrigation equipment is intercepted and consumed by crops in the irrigated areas and no water is available to recharge the ground-water system in irrigated areas. Using the transient model simulating seasonal conditions with recharge during the winter, estimated ground-water withdrawals were added during the summer stress period for each irrigation well in the model area as specified above. No adjustment was made in the ground-water recharge rate in observed areas of irrigation because during the summer, no natural recharge is assumed in the transient model simulation. Thus, like the steady-state simulation, irrigation is assumed to be efficient with all water that is added consumed by the crops. In reality, this efficiency may not be the case; however, as stated above, detailed information for individual farms and crops was not available.

Steady-state model results indicate simulated differences in water levels of -0.5 to 11 ft in the upper aquifer (model layer 1), -0.4 to 10.7 ft in the intermediate aquifer (model layer 3), and -0.3 to 10.7 ft in the lower aquifer (model layer 5). Largest differences are in the southern part of the model area in St. Joseph County (fig. 28). Transient model results indicate differences in water levels of 0 to 2.5 ft in the upper aquifer (model layer 1), 0 to 1.7 ft in the intermediate aquifer (model layer 3), and 0 to 1.7 ft in the lower aquifer (model layer 5). Largest differences are in the southern part of the model area in St. Joseph County (fig. 29). No recharge is assumed in steady-state and transient simulations during the summer in the irrigated areas and withdrawals are increased for irrigation. The reductions in water levels are less in the transient simulation because the summer season runs for only 4 months, whereas the steady-state simulation represents the long-term effects of the increased withdrawals.

Potential Effects of Changes in Withdrawals and Recharge

Because recharge likely will not remain constant, water levels and streamflow resulting after a prolonged dry period were simulated using the steady-state model. For model simulations, determination of the change in recharge used in model simulation to represent dry conditions was based on the historical changes observed in precipitation (fig. 10). A low of about 24 in./yr was recorded in 1999, a value equal to approximately 70 percent of average precipitation. Therefore, recharge rates were reduced to 70 percent of simulated values in the steady-state model and resulting water levels and streamflows compared to those resulting from average recharge values. The change in water levels from average recharge conditions to drier conditions is shown in figure 30. Simulated water levels in the upper aquifer generally were up to 8 ft higher during the average recharge conditions than during the reduced recharge conditions. Model areas to the north and west show the greatest difference under the change in recharge conditions. Simulated streamflows are lower for the reduced recharge conditions than for the average recharge conditions (fig. 31). Under 2001 pumping conditions with average recharge, five streams are simulated as losing water in the model; whereas under reduced recharge conditions, seven streams are simulated as losing water in the model.

To simulate results of short-term reductions in recharge, the transient models representing seasonal and monthly conditions also were simulated with reduced recharge. Water levels with reduced recharge conditions in the winter were compared to results of the transient simulation using average recharge rates during the winter and no recharge during the summer (except for local industry recharge). Simulation results for the transient model representing seasonal conditions indicated

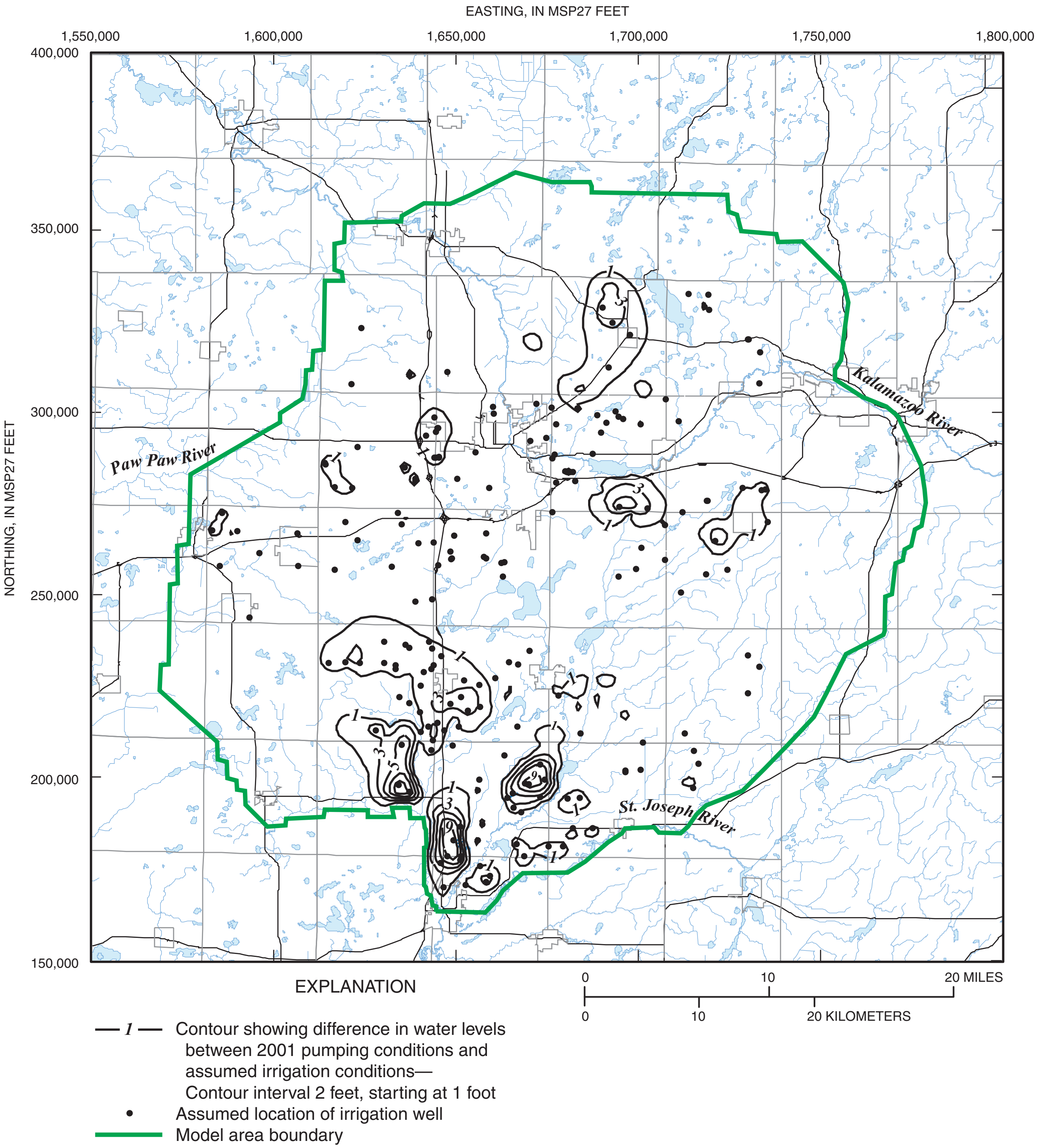


Figure 28. Difference in steady-state water levels in the upper aquifer (model layer 1) between 2001 pumping conditions and assumed irrigation conditions, Kalamazoo County area, Michigan.

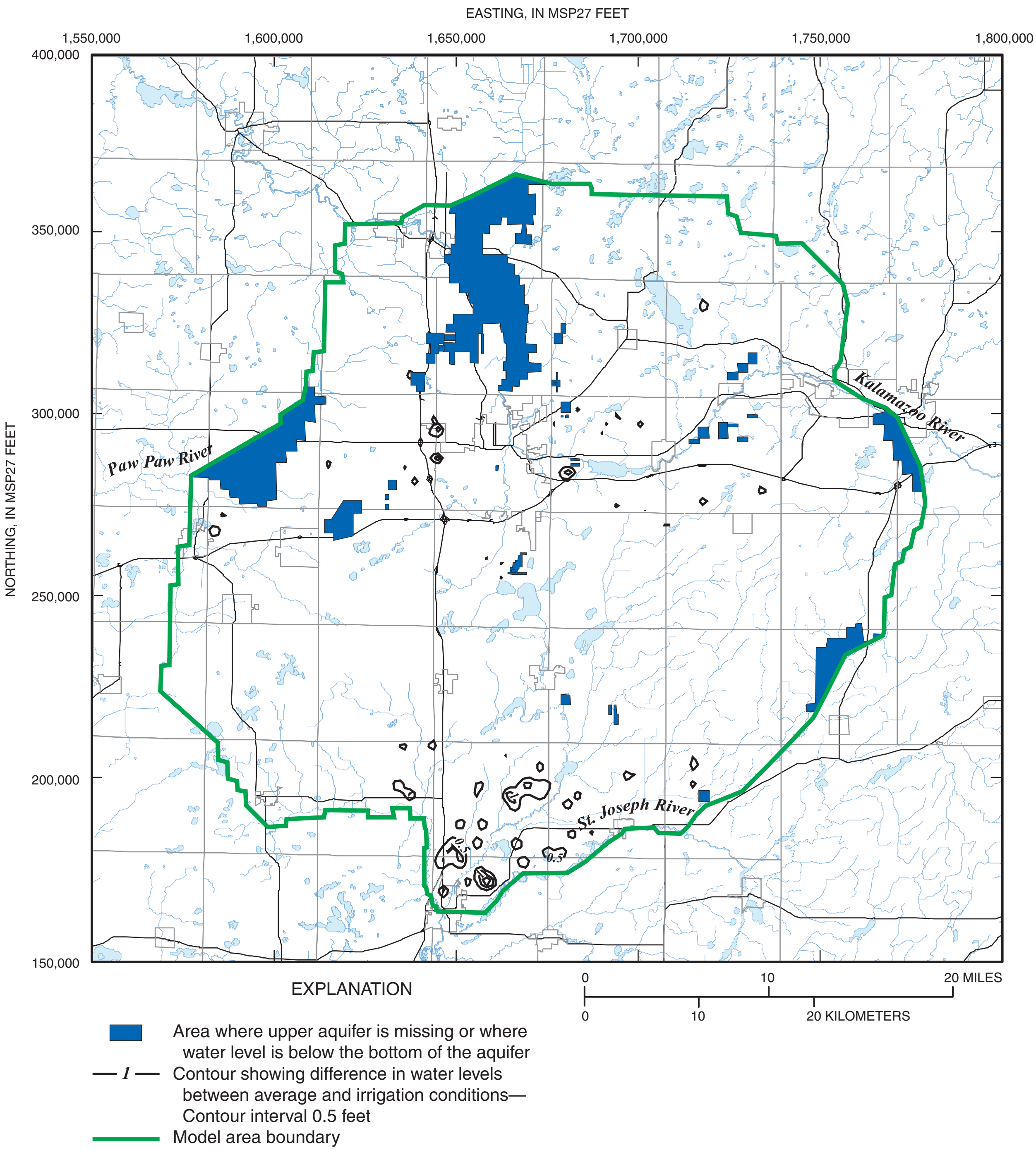


Figure 29. Difference in transient seasonal water levels in the upper aquifer (model layer 1) between 2001 pumping conditions and simulated irrigation conditions, Kalamazoo County area, Michigan.

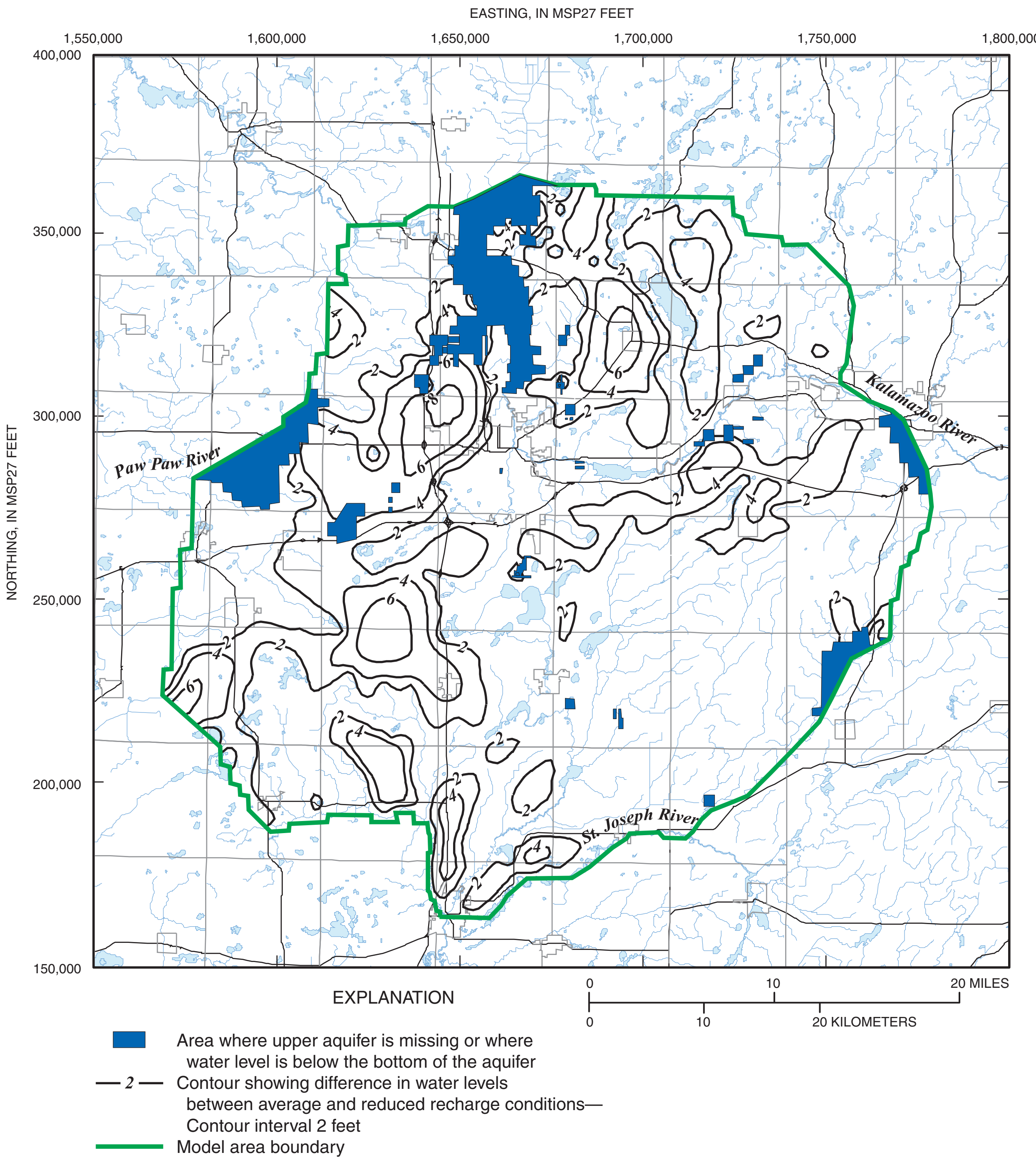


Figure 30. Difference in steady-state water levels in the upper aquifer (model layer 1) between average and reduced recharge, 2001 pumping conditions, Kalamazoo County area, Michigan.

changes in water levels of -0.3 to 3.6 ft in the upper aquifer (model layer 1), 0 to 3.1 ft in the intermediate aquifer (model layer 3), and 0 to 3.1 ft in the lower aquifer (model layer 5) when recharge rates are reduced. Simulation results for the transient model representing monthly conditions indicated changes in water levels of -0.6 to 3.7 ft in the upper aquifer (model layer 1), 0 to 3.2 ft in the intermediate aquifer (model layer 3), and 0 to 3.2 ft in the lower aquifer (model layer 5) when recharge rates are reduced. Areas of water-level changes for the upper aquifer for each transient simulation are shown in figures 32 and 33. For both simulations, declines in water levels are greatest in the north-central and southwestern parts of the model area. A detailed analysis of changes in streamflow was not made for each individual stream because the streams are represented in the model as river cells that do not account for the total amount of water flowing in the stream from neighboring upstream cells. Thus, a river cell could lose more water than actually is flowing in the stream. Overall, river cells tended to gain less water in both the seasonal and monthly simulations under reduced recharge conditions than under average recharge conditions. River cells also indicated that more water flowed out of these cells under reduced recharge conditions than under average recharge conditions for both transient simulations.

Ground-water-withdrawal rates have changed historically (table 6) and also are expected to change in the future. Estimated withdrawals for 2010 were simulated using the steady-state model with average and reduced recharge rates. Comparison of municipal withdrawals from 1966 to 2001 indicates an average increase of about 1 percent per year. Not all withdrawals increased from 1966 to 2001. Some communities decreased for parts of this time interval; therefore, the estimated increase represents only one possible withdrawal condition. Actual municipal withdrawals in 2010 are likely to differ from those simulated; however, these simulated withdrawals are assumed to be representative of actual withdrawals. To estimate future ground-water-withdrawal conditions for 2010, withdrawals from 2001 were increased by 9 percent. Simulated steady-state water levels using 2001 and 2010 pumping conditions were compared for both average and reduced recharge conditions.

Steady-state model results indicate simulated differences in water levels of -2.1 to 4.9 ft in the upper aquifer (model layer 1), -0.04 to 5.7 ft in the intermediate aquifer (model layer 3), and -0.04 to 6.6 ft in the lower aquifer (model layer 5) with the change in pumping conditions under average recharge rates. When pumping rates are increased from 2001 to projected 2010 rates, water levels decrease the most in the intermediate aquifer in an area west and north of Upjohn Pond. When recharge rates are reduced, steady-state model results from 2001 to 2010 pumping conditions indicate simulated differences in water levels of -0.45 to 5.6 ft in the upper aquifer (model layer 1), -0.02 to 5.6 ft in the intermediate aquifer (model layer 3), and -0.02 to 6.5 ft in the lower aquifer (model layer 5). Water-level declines are greatest in an area west and north of Upjohn Pond. Under reduced recharge rates, water-level declines in the intermediate and lower aquifers are similar to the declines in these aquifers resulting from the simulation with average recharge rates. However, declines in the upper aquifer are slightly greater from 2001 to 2010 pumping conditions with reduced recharge rates. Under both recharge conditions, recharge rates for areas receiving water from Pfizer, Inc. were not changed. As withdrawals increased, the rates of ground-water recharge also would likely increase and may offset the lowered water levels resulting from increased simulated withdrawals in this area.

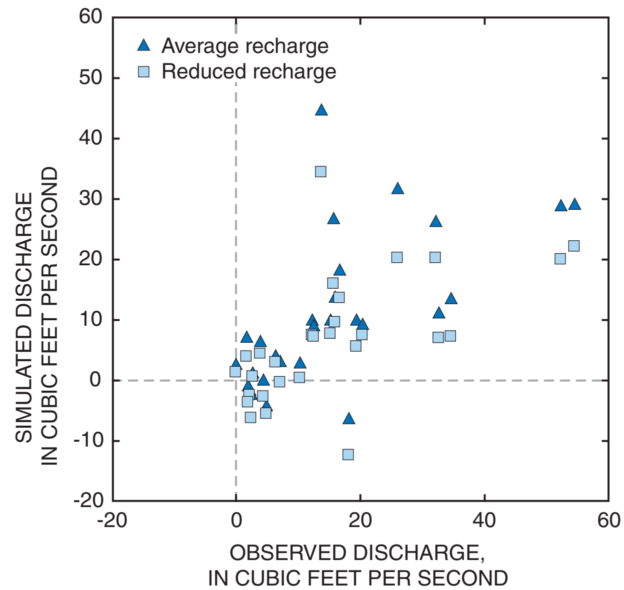


Figure 31. Change in streamflow from average to reduced recharge conditions, Kalamazoo County area, Michigan.

Comparison of streamflows from 2001 to 2010 under average recharge conditions indicates flow remains unchanged in streams that are not located near pumping centers. Streams located near pumping centers have lower flows with the increased ground-water withdrawals and include Davis Creek, West Fork Portage Creek at Oshtemo, Portage Creek at Stockbridge, Arcadia Creek, and Portage Creek at Lovers Lane. Comparison of streamflows from 2001 to 2010 under reduced recharge conditions also indicates that flow remains unchanged in streams that are not located near pumping centers. Streams located near pumping centers have lower flows with the increased withdrawals and include Davis Creek, Comstock Creek, West Fork Portage Creek at Oshtemo, Portage Creek at Stockbridge, and Portage Creek at Lovers Lane.

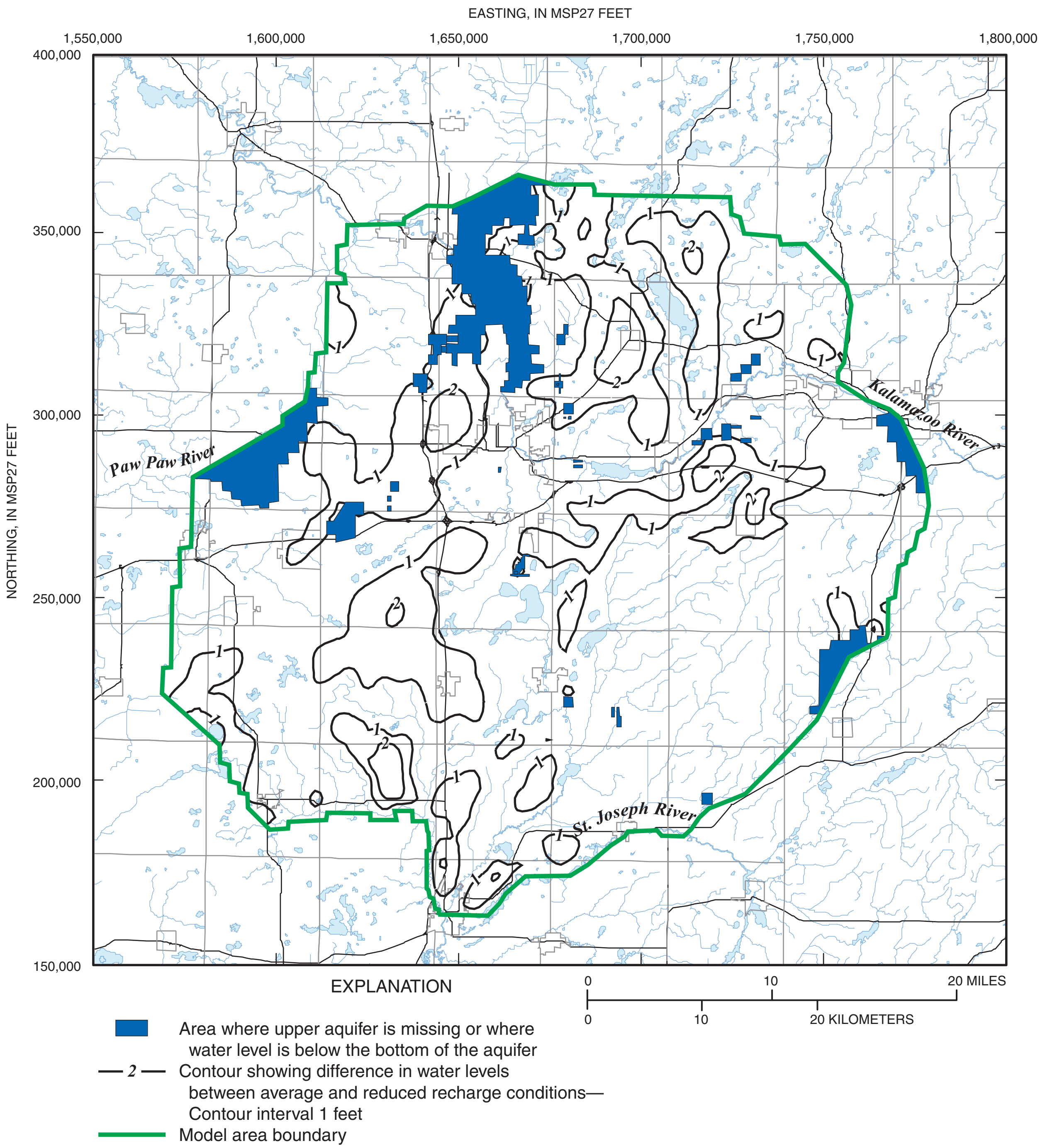


Figure 32. Difference in transient summer water levels in the upper aquifer (model layer 1) between average and reduced recharge, 2001 pumping conditions, Kalamazoo County area, Michigan.

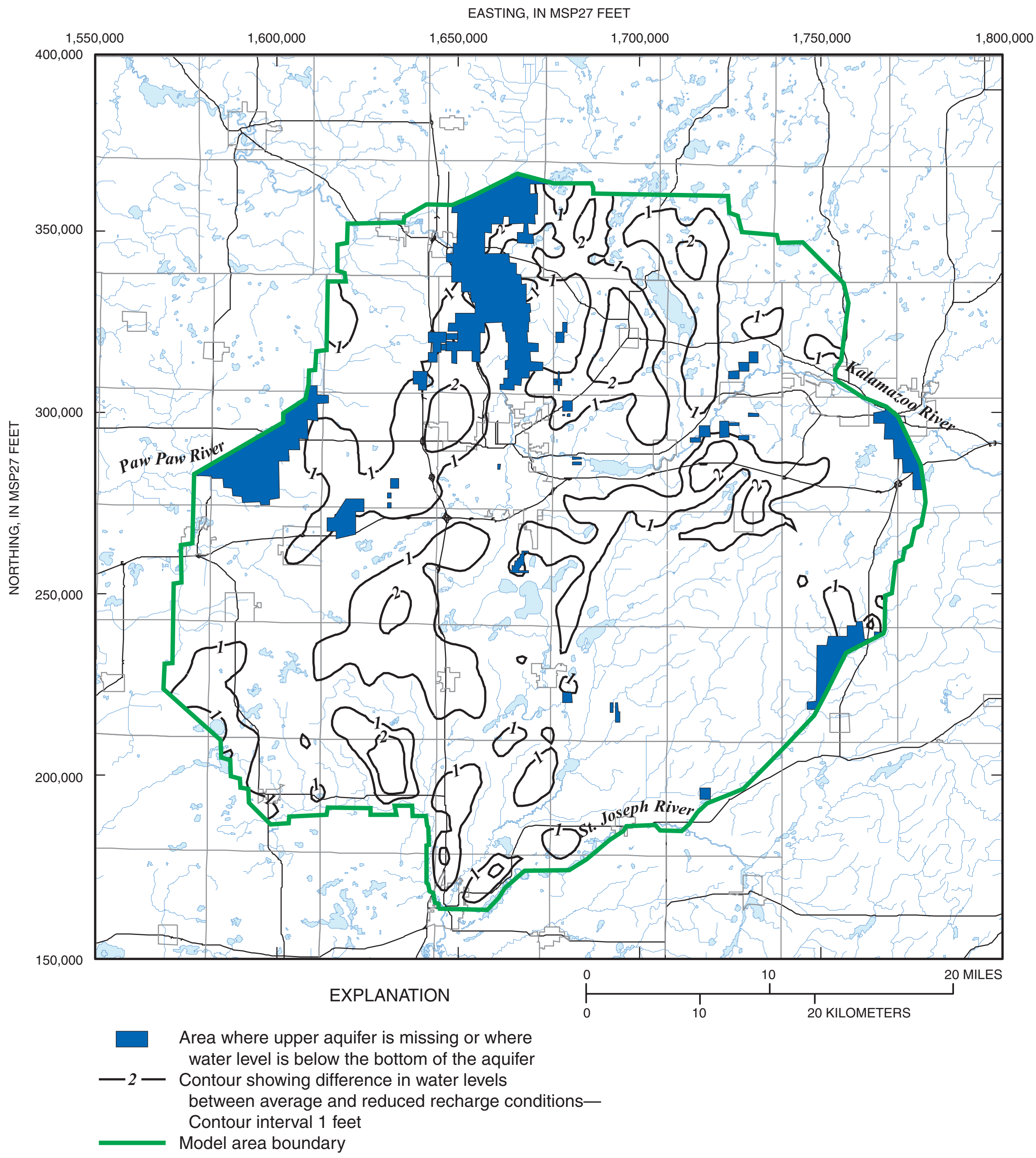


Figure 33. Difference in transient September water levels in the upper aquifer (model layer 1) between average and reduced recharge, 2001 pumping conditions, Kalamazoo County area, Michigan.

Regional Ground-Water Budgets for Kalamazoo County

For the purpose of developing a hydrologic budget, determination of the components of the budget is needed so that the amount and movement of water within the system is represented. Within the ground-water-flow system, some water is stored within the ground-water system, whereas some water recharges or discharges from the system through various pathways, which include system boundaries, rivers, infiltration from precipitation, and wells. Any water that is withdrawn by a well will have some immediate and/or long-term effect on the system, perhaps in lower streamflow, loss of wetland areas, or reduced discharge to lakes. With development and the addition of wells, a new discharge is superimposed upon a previously stable system that is at equilibrium, and it must be balanced by an increase in the recharge of an aquifer, or by a decrease in the old natural discharge, or by a loss of storage in the aquifer, or by a combination of changes in recharge, discharge, or storage (fig. 34).

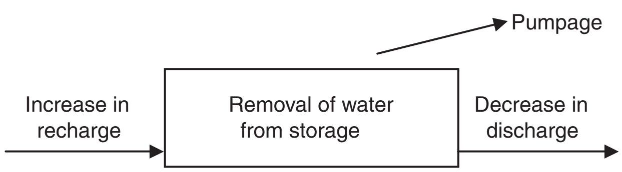


Figure 34. Hydrologic budget components.

For most ground-water systems, the change in storage in response to pumping is a transient condition that occurs as the system readjusts to the pumping stress. The relative contributions of changes in storage, changes in recharge, and changes in discharge evolve with time, potentially over a period of years, decades, or even centuries. The long-term source of water to discharging wells typically is a change in the amount of water entering or leaving the system. The numerical model can be used to aid in estimating water availability and the effects of extracting water from the ground-water and surface-water system. The model results can be used to evaluate ground water and surface water together on a systemwide basis and can be used to help predict the effects of alternative management simulations, so acceptable effects can be determined. The simulations conducted during this study were developed so that various stresses, such as pumping or recharge, were changed to allow comparison of the resulting changes in water levels or streamflow.

For the ground-water-flow system in the Kalamazoo County area, water may enter the system as recharge from precipitation, seepage from lakes and rivers, and as inflow from outside the study area. Water may leave the ground-water-flow system in the Kalamazoo County area as evapotranspiration, seepage into lakes and rivers, outflow from the study area, and as withdrawals by wells. Some changes in the amounts of water within parts of the system may occur in response to changes in stresses that bring the system out of equilibrium. These stresses could result in changes in the amount of water in storage within the aquifers. For the steady-state simulations, the amount of water recharging the ground-water system is assumed to equal the amount of water discharging from the ground-water system; changes in storage do not occur. Changes in the amounts of water in storage are represented in the transient simulations.

Simulation of the ground-water-flow system with MODFLOW-2000 allows determination of the hydrologic budget for each calibration period for each simulation. The hydrologic budget for each model simulation lists the amount of water entering and leaving the various components in budget calculations, including boundaries, rivers, recharge, and wells. Budgets for selected steady-state simulations are shown in tables 10 and 11. The budgets for the steady-state simulations investigating changes in recharge and withdrawals (table 11) indicate both changes in flow to and from rivers and model boundaries. With a reduction in recharge under either 2001 or 2010 pumping conditions, less water is available for streamflow and less water flows through the model boundaries. With a change in withdrawals, flow through boundaries does not change but less water is available for streamflow under both average and reduced recharge conditions.

The budgets for selected transient simulations are shown in tables 12 and 13. The budgets for the transient simulations with seasonal conditions indicate changes in flow to boundaries, rivers, and storage at the end of the summer. Thus, the simulations with no summer recharge and either average or

Table 10. Summary of hydrologic budgets for selected steady-state model scenarios representing current and historical conditions, Kalamazoo County area, Michigan.

Budget component	Volume of water (million gallons)			
	2001	1994	1987	1966
IN				
Boundaries	6.0	6.0	6.0	6.0
Rivers	304.1	299.4	281.7	275.5
Recharge	556.5	556.5	551.1	543.1
OUT				
Boundaries	60.3	60.3	60.2	60.2
Rivers	725.2	728.8	703.7	697.3
Wells	60.1	52.5	56.7	52.8
Percent discrepancy [(IN-OUT)/ IN *100]	2.45	2.38	2.2	1.75

Table 11. Summary of hydrologic budgets for selected steady-state model scenarios representing varying withdrawals and recharge, Kalamazoo County area, Michigan.

Budget component	Volume of water (million gallons)			
	2001 withdrawals with average recharge	2001 withdrawals with reduced recharge	Estimated 2010 withdrawals with average recharge	Estimated 2010 withdrawals with reduced recharge
IN				
Boundaries	6.0	6.8	6.0	6.8
Rivers	304.1	349.0	307.1	352.3
Recharge	556.5	396.7	556.5	396.7
OUT				
Boundaries	60.3	50.6	60.3	50.6
Rivers	725.2	625.7	723.1	623.9
Wells	60.1	60.1	65.2	65.2
Percent discrepancy [(IN-OUT)/IN *100]	2.45	2.18	2.45	2.17

reduced winter recharge indicate a large amount of water leaving storage and no water entering storage, less water available for streamflow, and reduced flow out of model boundaries at the end of the summer. The budgets for the transient simulations with monthly conditions indicate changes in flow to boundaries, rivers, and storage at the end of September that are similar to those observed with the seasonal simulation.

Changes in storage for the transient simulations with varying withdrawals and recharge for each month of the simulation are shown in figure 35. In the top graph, recharge remained constant during the entire simulation. Water enters storage during the winter months and is removed from storage during the summer months, as expected. With increased withdrawals to meet summer demands, water that may have entered storage is instead withdrawn by wells. In the middle graph, recharge was assumed to vary each month during the year using the estimated rate from the steady-state simulations proportioned to each month based on the average precipitation rate for that month. Water enters storage from mid-January to June and is removed from storage the other months of the year. In the bottom graph, recharge was assumed to vary during the winter months using the estimated rate from the steady-state simulations proportioned to each winter month based on average precipitation rates. No natural recharge occurred during the summer months. Water enters storage from mid-November to mid-June, although water movement into and out of storage was about equal at the end of December. Water is removed from storage during the first part of November and from mid-June until the end of the simulation in September. These graphs show the rate of water movement into and out of storage. The area under each line represents the actual volume of water moving into and out of storage. Determination of the volumes of water for each recharge condition indicates that when recharge is assumed to remain constant and when recharge is allowed to vary throughout the year, the amount of water that enters storage is greater than that which leaves storage. However, when recharge is distributed throughout the winter months only or when recharge rates are reduced during the winter months, the amount of water that enters storage is less than that which leaves storage.

A final transient simulation was conducted to investigate changes in budget components as withdrawals were allowed to increase from 2001 to 2010 conditions (fig. 36). As mentioned previously, withdrawals were estimated to increase about 1 percent per year based on historical data. Using the transient model simulating seasonal changes, summer and winter withdrawals were increased to projected 2010 pumping rates. Recharge remained at estimated 2001 rates for each winter and summer season. After the first year, changes in flow to and from boundaries, rivers, and storage remain relatively constant, although there was a slight increase in the amount of water discharged from rivers.

Table 12. Summary of hydrologic budgets at the end of selected transient model scenarios representing seasonal conditions, Kalamazoo County area, Michigan.

Budget component	Volume of water (billion gallons)			
	Constant recharge in winter and summer	Recharge occurs in winter and summer	Recharge occurs in the winter	Reduced recharge occurs in the winter
IN				
Storage	0.8	6.9	48.1	39.8
Boundaries	2.2	2.2	2.1	2.3
Rivers	109.9	108.2	105.9	114.7
Recharge	203.7	206.7	206.5	149.5
OUT				
Storage	.8	6.0	32.8	6.0
Boundaries	22.1	22.3	23.2	20.8
Rivers	275.1	278.1	290.3	263.8
Wells	21.5	21.5	21.5	21.5
Percent discrepancy [(IN-OUT)/IN *100]	-.89	-1.2	-1.38	-1.86

Table 13. Summary of hydrologic budgets at the end of selected transient model scenarios representing monthly conditions, Kalamazoo County area, Michigan.

Budget component	Volume of water (billion gallons)			
	Constant recharge in winter and summer	Recharge occurs in winter and summer	Recharge occurs in the winter	Reduced recharge occurs in the winter
IN				
Storage	1.6	19.3	53.7	50.9
Boundaries	2.2	2.2	2.1	2.3
Rivers	112.9	112.0	110.0	118.8
Recharge	203.7	204.2	204.3	146.4
OUT				
Storage	1.8	18.2	39.7	17.6
Boundaries	22.1	22.2	23.0	2.1
Rivers	273.6	278.2	289.0	263.0
Wells	21.2	21.2	21.2	21.2
Percent discrepancy [(IN-OUT)/ IN *100]	.53	-.63	-.74	-1.24

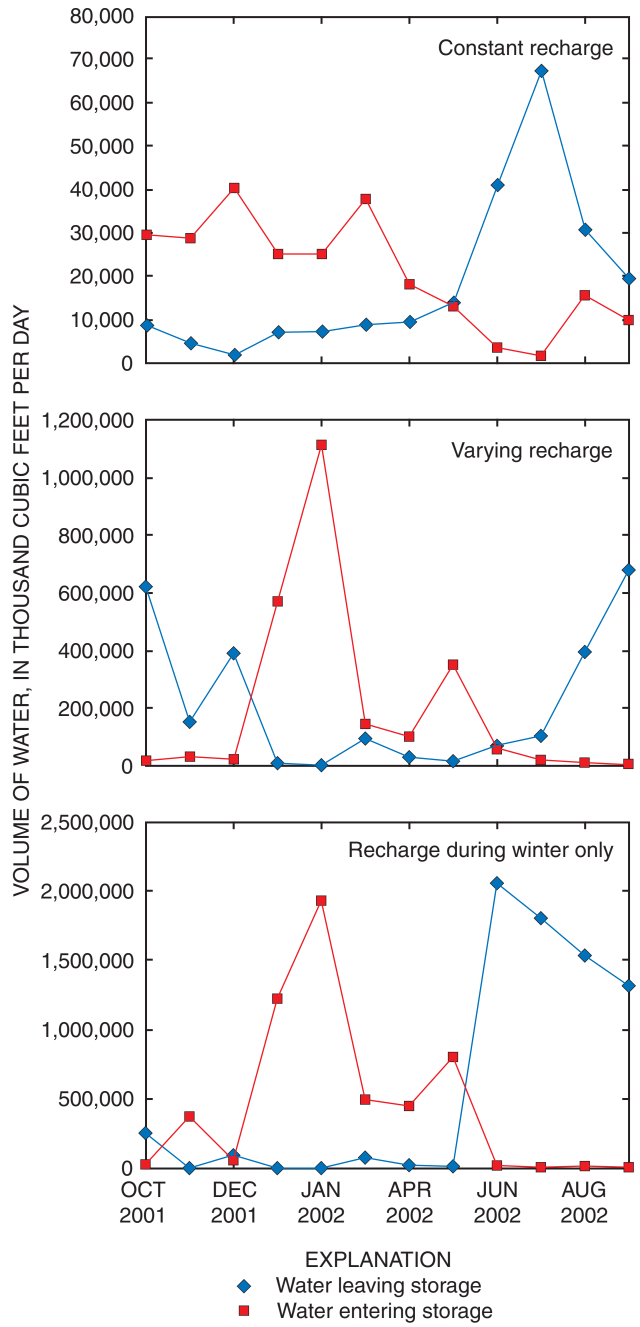


Figure 35. Changes in volume of water entering and leaving storage from selected transient simulations, Kalamazoo County area, Michigan.

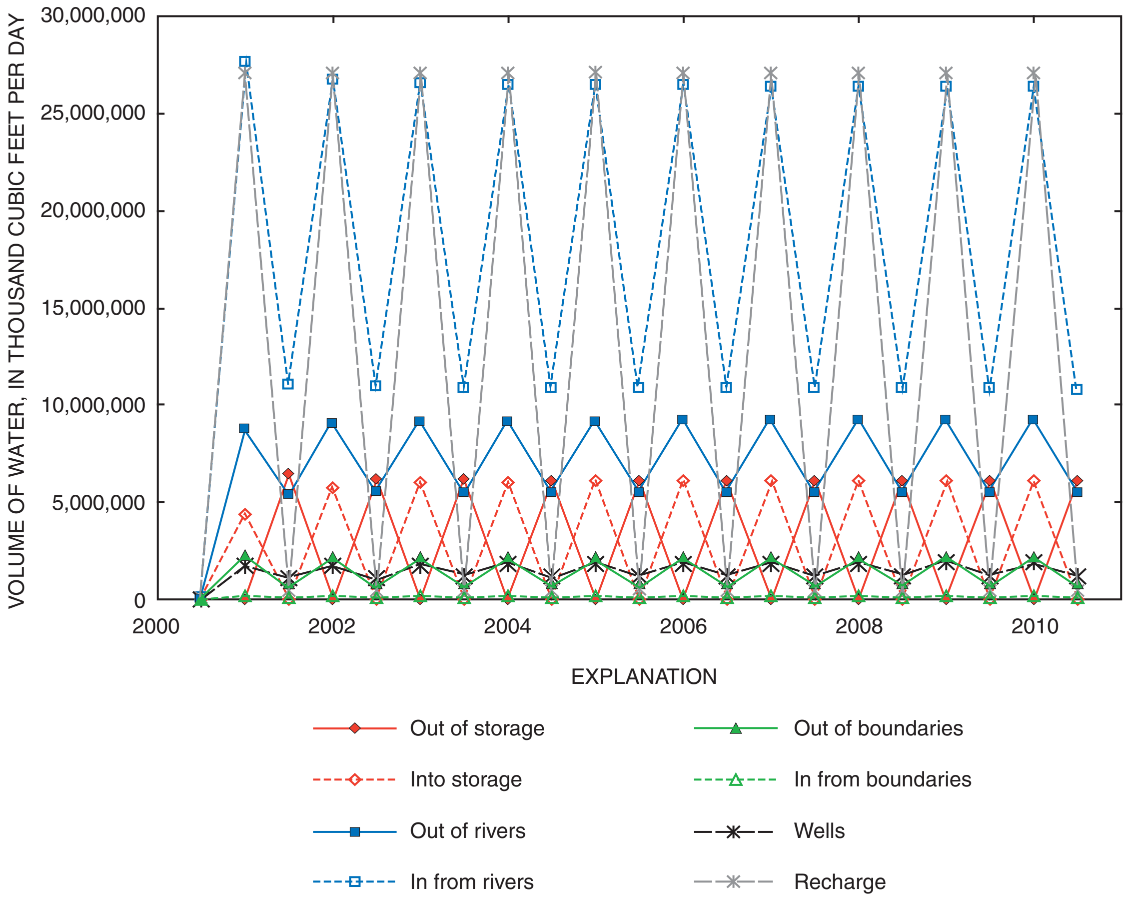


Figure 36. Changes in budget components for transient simulation representing increasing withdrawals from 2001 to 2010 pumping conditions, Kalamazoo County area, Michigan.

Conclusions

In Kalamazoo County and the surrounding area, ground water is widely used as a source of water for drinking and industry. Additionally, streams and lakes are valued for their recreational and aesthetic uses. As natural and human-induced stresses change, the long-term availability of ground water and surface water for people to use and enjoy may be questioned. To address concerns about the potential effects of changes in withdrawals and recharge, the U.S. Geological Survey, in cooperation with the city of Kalamazoo, city of Portage, the Kalamazoo County Human Services Department, and the Michigan Department of Environmental Quality, began a study in 2000 of the ground-water-flow system in the Kalamazoo County area. In this study, a ground-water-flow model was constructed to investigate possible effects of changes in withdrawals and recharge on water levels, streamflow, and the amount of water in the system.

Glacial deposits, consisting largely of sands and gravels, are the source of water for most residents and businesses in Kalamazoo County. The glacial deposits can be divided into an upper unconfined aquifer that occurs throughout most of the county and one to two lower semiconfined aquifers that occur in about one-third of the county. The upper unconfined aquifer ranges in thickness from 0 to 120 ft, the intermediate aquifer ranges in thickness from 0 to 100 ft; and the lower aquifer ranges in thickness from 0 to 120 ft. The upper confining unit ranges in thickness from 0 to 120 ft and the lower confining unit ranges in thickness from 0 to 170 ft. The Coldwater Shale, which is not generally used for water supply, underlies the glacial deposits in most of the county. The Marshall Formation underlies the glacial deposits in the northeastern part of the county. Sufficient quantities of water can be withdrawn from this formation for domestic supply.

Regional ground-water flow is towards the major surface-water features in the area, including the Kalamazoo River, the St. Joseph River, and the Paw Paw River. Local ground-water flow is towards pumping centers. Natural ground-water recharge, derived from precipitation, is assumed to vary depending on the surficial materials, which include till, moraines, a glacial drainage channel, and outwash plains.

Data collected during this study included ground-water-level measurements at 42 miscellaneous and 18 continuous wells, stream-discharge measurements at 22 miscellaneous and 8 continuous-recording stream gages, and water levels at 12 miscellaneous lake sites and at 4 continuous lake-level gaging stations. To provide consistency and to allow examination of any differences over time, ground-water and surface-water sites were selected because they are currently (2002) monitored sites or were monitored during previous studies. The interaction between surface water and ground water was investigated using seepage meters, piezometers, and temperature measurements. Seepage measurements were collected in Long and Austin Lakes and selected streams in Kalamazoo, St. Joseph, and Van Buren Counties at seven different times between July 2000 and September 2002, although not all sites were measured each time. Piezometer measurements were collected in conjunction with seepage measurements at selected sites. Water-level and temperature measurements were collected at two sites differing in the amount of vegetation near Long and Austin Lakes.

Data collected during this study and compiled from previous studies provided input values and calibration information for the model. The model area consists of Kalamazoo County and parts of the surrounding counties, including Allegan, Barry, Calhoun, St. Joseph, and Van Buren Counties. For model simulation, this area is horizontally discretized into a variably spaced grid of cells in 154 rows and 162 columns. The model area is vertically discretized into six model layers. Layers 1, 3, and 5 represent the upper, intermediate, and lower aquifers, whereas layers 2 and 4 represent the upper and lower confining units. Layer 6 represents the Marshall Sandstone in the northeastern part of the model area and clay where present over the rest of the model area.

Simulations consisted of steady-state model runs (in which stresses remained constant and changes in storage were not included) and transient model runs (in which stresses changed in seasonal and monthly time scales and storage within the system was included). Parameter estimation was used with the steady-state model representing 2001 pumping conditions to find the best fit between observed and simulated water levels and streamflow. Available information on water levels and streamflow from previous studies was used to check model fit to 1994, 1987, and 1966 conditions.

Steady-state simulations compared changes in water levels and streamflow under average and reduced recharge conditions. Simulated water levels in the upper aquifer generally were up to 8 ft higher during the average recharge conditions than during the reduced recharge conditions. Simulated streamflows were lower for the reduced recharge conditions than for the average recharge conditions. Steady-state simulations also were used to compare changes in water levels and streamflow from 2001 to projected 2010 pumping conditions for average and reduced recharge conditions. When pumping rates are increased, water-level declines are greatest in the intermediate aquifer in an area west and north of Upjohn Pond under either recharge condition; however, likely increases in local recharge were not accounted for in these simulations. In response to the increase in pumping under either recharge condition, streamflow generally remained about the same in streams away from pumping centers or decreased in streams near pumping centers.

Transient simulations were used to compare changes in water levels under four different recharge conditions for seasonal and monthly stress periods. Withdrawals were allowed to change in the transient simulations to better represent the actual pattern of lower withdrawals in the winter and higher withdrawals in the summer to meet increased demands. The recharge conditions consisted of constant recharge, recharge distributed throughout the year, recharge occurring in the winter only, and reduced recharge in the winter only. Generally the simulation with recharge occurring only in the winter provided the best fit to observed water levels in some wells and to the trends in observed water levels in other wells.

Some additional simulations were conducted to investigate potential effects of reduced recharge because of urban areas and potential effects of withdrawals and interception of water by crops because of irrigation. Assuming a reduction in recharge of 75 percent within the city of Kalamazoo because of urbanization, water levels declined up to 4.3 ft and flow to streams was reduced. These results were based on steady-state simulation. Using the steady-state model with additional irrigation withdrawals in the upper aquifer and no recharge in observed areas of irrigation, declines of up to 11 ft were simulated in the southern part of the model area, primarily in St. Joseph County. Using the transient model with additional irrigation withdrawals and no natural recharge in the summer (thus, as in the steady-state simulation, assuming completely efficient irrigation with all applied water used by crops), declines of up to 2.5 ft were simulated in the southern part of the model area, primarily also in St. Joseph County. The transient simulation indicates a smaller decline in water levels because the summer season only lasts 4 months, whereas the steady-state simulation represents the potential long-term effects of the increased withdrawals and reduced recharge.

To better understand the amount and movement of water within the ground-water-flow system in the Kalamazoo County area, hydrologic budgets for each simulation were determined. Withdrawals by wells intercept water that would have discharged at other locations, possibly to a stream or lake. With the development and addition of new wells, a new discharge is imposed upon a previously stable system, and it must be balanced by an increase in the recharge of an aquifer, a decrease in the old natural discharge, a loss of storage in the aquifer, or by a combination of these. The initial source of water to a well is from storage, but over the long term, the source typically is a change in the amount of water entering and leaving the system.

Budgets for the steady-state simulations indicate that with reduced recharge, less water is available for streamflow and less water leaves the model area through the boundaries. With an increase in pumping rates, flow through the boundaries does not change, but less water is available for streamflow. Budgets for the transient simulations with no natural summer recharge indicate that water enters storage during the winter months and leaves storage during the summer as withdrawals increase. During the summer, less water is available for streamflow and flow out of model boundaries is reduced. Determination of the volumes of water for each recharge condition indicates that when recharge is assumed to remain constant and when recharge is allowed to vary throughout the year, the amount of water that enters storage is greater than the amount that leaves storage. However, when recharge is distributed throughout the winter months only or when recharge rates are reduced during the winter months, the amount of water that enters storage is less than the amount that leaves storage.

An additional simulation investigated changes in budget components as withdrawals increased from 2001 to projected 2010 pumping rates. Recharge rates remained at estimated 2001 winter and summer rates for each year of the simulation. After the first year, changes in flow to and from boundaries, rivers, and storage remained relatively constant. For all transient simulations, flow to rivers is less with reduced recharge and with increased withdrawals. Flow out of rivers is greater with reduced recharge and with increased withdrawals. Flow out of model boundaries is less with reduced recharge and with increased withdrawals. Flow in through model boundaries did not change appreciably during the transient simulations. The 9-year simulation from 2001 to 2010 indicates water entering storage and rivers during the winter and leaving storage and rivers during the summer. With the trend of increasing withdrawals over this time period, the amount of water entering and leaving storage remains at about a constant rate, whereas the amount of water available for streamflow slightly decreases. This simulation depicts a possible effect on regional streamflow values. Determination of possible effects at specific stream sites would need some accounting of the amount of water within the stream channel to ensure that model simulation is not inducing more water out of the stream than is actually flowing in the stream. Likely changes in recharge during the time period of this simulation that would also affect water movement within the various components of the ground-water-flow system are not accounted for in model simulation. With reduced recharge, water must come from storage, rivers, or in from boundaries to meet withdrawal demands.

Acknowledgments

The authors gratefully acknowledge the assistance of Connie Cousins-Leatherman, Kalamazoo County Human Services Department; Richard Gendernalik, Pfizer, Inc.; William Hunsberger, Pfizer, Inc.; Robert Kimmer, city of Portage; Patrick Krause, Kalamazoo County Human Services Department; Peter Maas, Pfizer, Inc.; Bruce Minsley, city of Kalamazoo; and John Paquin, city of Kalamazoo. This advisory group provided much information, including hydrogeological and historical insights, about the Kalamazoo County area and provided guidance for this study to help make it a useful tool for water-resources managers to address future water-management issues. Pfizer, Inc., which was known as Pharmacia when this study began, provided information for this study. The Kalamazoo Community Foundation provided support for the 1997 study that involved expanding the ground-water and surface-water-monitoring network.

Numerous local homeowners provided access to area lakes for data collection and include Julie Bosma, William Channels, Julie Ellis, Seth Guim, David Kosacek, Robert Seely, David Shires, Alice Vemich, and Oliver Woods. Julie Ellis assisted with location of monitoring sites and provided valuable current and historic insights about the area. Oliver Woods permitted numerous visits to his property for lake-level gage installation and monitoring and for seepage data collection, provided tools, collected local precipitation data, and also provided numerous current and historical insights about the area. Robert Seely provided access to Austin Lake for seepage measurements and provided information on reports that described local conditions. Well logs and results of aquifer tests were provided by Robert Snell of Prein and Newhof. Christopher Everts of Mactec Engineering and Consulting, Inc. provided some local water-level information. Barbara Marczak of Prein and Newhof provided information on water levels and well elevations in the Schoolcraft area. Donald O. Rosenberry, USGS, assisted with collection and analysis of seepage data. Dr. Alan Kehew, Western Michigan University, and Dr. Grahame Larson, Michigan State University, provided valuable assistance in characterizing the geology and hydrology of the area. Additionally, thanks go to the many other people who assisted with data collection and many other aspects of the project, including Jaye Lunsford and Brian Neff of USGS, Dan Lunsford, Michele Morenz, and Jason Killian.

References Cited

Allen, William B., Miller, John B., and Wood, Warren W., 1972, Availability of water in Kalamazoo County, southwestern Michigan: U.S. Geological Survey Water-Supply Paper 1973, 129 p.

Anderson, Mary P., and Woessner, William W., 1992, Applied Ground-Water Modeling: Simulation of flow and advective transport: Academic Press, Inc., San Diego, California, 381 p.

City of Kalamazoo, Department of Public Services, 1995, Hydrogeologic investigation report regarding the proposed municipal wellfield site, 37th Street and Greer Drive vicinity, Ross Township, Kalamazoo County, Michigan: Kalamazoo, Michigan, variously paged.

City of Kalamazoo, Department of Public Services and Peerless-Midwest Company, Inc., 1996, Kalamazoo, Michigan regional groundwater flow model: Granger, Indiana, variously paged.

City of Kalamazoo, Department of Public Services and Peerless-Midwest Company, Inc., 1999a, Kalamazoo, Michigan, Water pumping stations 1, 2, 3, 4, & 7, Groundwater flow model and capture zone delineations: Kalamazoo, Michigan and Mishawaka, Indiana, variously paged.

City of Kalamazoo, Department of Public Services and Peerless-Midwest Company, Inc., 1999b, Kalamazoo, Michigan, Water pumping stations 11 & 12, Ground-water flow model and capture zone delineations: Kalamazoo, Michigan and Mishawaka, Indiana, variously paged.

City of Kalamazoo, Department of Public Services and Peerless-Midwest Company, Inc., 2002a, Kalamazoo, Michigan, Water pumping stations 5 and 14, Ground-water flow model and capture zone delineations: Kalamazoo, Michigan and Mishawaka, Indiana, variously paged.

City of Kalamazoo, Department of Public Services and Peerless-Midwest Company, Inc., 2002b, Kalamazoo, Michigan, Water pumping stations 25 and 39, Ground-water flow model and capture zone delineations: Kalamazoo, Michigan and Mishawaka, Indiana, variously paged.

Cohee, G.V., 1979, Michigan Basin region, in Craig, L.C., and Connor, C.W., eds., Paleotectonic investigations of the Mississippian System in the United States, Part I—Introduction and regional analyses of the Mississippian System: U.S. Geological Survey Professional Paper 1010, p. 49–57.

Cummings, T.R., Twenter, F.R., and Holtschlag, D.J., 1984, Hydrology and land use in Van Buren County, Michigan: U.S. Geological Survey Water-Resources Investigations Report 84-4112, 124 p.

de Marsily, G., 1986, Quantitative Hydrogeology: Academic Press, 440 p.

Deutsch, Morris, Vanlier, K.E., and Giroux, P.R., 1960, Ground-water hydrology and glacial geology of the Kalamazoo area, Michigan: Michigan Geological Survey Progress Report 23, 122 p.

Dorr, J.A., Jr., and Eschman, D.F., 1970, Geology of Michigan: Ann Arbor, Michigan, University of Michigan Press, 476 p.

Erickson, D.R., 1981, A study of littoral groundwater seepage at Williams Lake, Minnesota, M.S. Thesis, University of Minnesota.

Forstat, D.W., 1983, Drift thickness of Kalamazoo County, in Monaghan, G.W., Larson, G.J., Forstat, D.W., and Sorenson, H.O., Selected geologic maps of Kalamazoo County, Michigan: County geologic map series, Michigan Department of Natural Resources, Geological Survey Division, 5 pls., scale 1:100,000.

Franke, O.L., Reilly, T.E., and Bennett, G.D., 1987, Definition of boundary and initial conditions in the analysis of saturated ground-water flow systems; an introduction: U.S. Geological Survey Techniques of Water Resources Investigations 03-B5, 15 p.

Freeze, R. Allen, and Cherry, John A., 1979, Groundwater: Prentice-Hall, Inc., Englewood Cliffs, New Jersey, 604 p.

Freeze, R.A. and Banner, J.A., 1970, The mechanism of natural ground-water recharge and discharge 2. Laboratory column experiments and field experiments: Water Resources Research, v. 6, p. 138–155.

Hale, Lucille, 1941, Study of sedimentation and stratigraphy of Lower Mississippian in western Michigan: American Association of Petroleum Geologists Bulletin, v. 25, p. 713–723.

Harbaugh, Arlen W., Banta, Edward R., Hill, Mary C., and McDonald, Michael G., 2000, MODFLOW-2000, The U.S. Geological Survey modular ground-water model—User guide to modularization concepts and the ground-water flow process: U.S. Geological Survey Open-File Report 00-92, 121 p.

Hill, Mary C., 1998, Methods and guidelines for effective model calibration: U.S. Geological Survey Water-Resources Investigations Report 98-4005, 90 p.

Hill, Mary C., Banta, Edward R., Harbaugh, Arlen W., and Anderman, Evan R., 2000, MODFLOW-2000, The U.S. Geological Survey modular ground-water model—User guide to the observation, sensitivity, and parameter-estimation processes and three post-processing programs: U.S. Geological Survey Open-File Report 00-184, 209 p.

Hoaglund, J.R., Huffman, G.C., and Grannemann, N.G., 2000, Simulation of ground water flow in the Glaciofluvial, Saginaw, Parma-Bayport, and Marshall aquifers, central Lower Peninsula of Michigan: U.S. Geological Survey Open-File Report 00-504, 36 p.

Holtschlag, D.J., 1994, A generalized estimate of ground-water recharge rates in the lower peninsula of Michigan: U.S. Geological Survey Water-Supply Paper 2437, 81 p.

Huffman, G.C. and Thompson, T., 1971, Summary of ground-water hydrological data in Michigan in 1970: U.S. Geological Survey Report (unnumbered), 94 p.

Huffman, G.C. and Thompson, T., 1973, Summary of ground-water hydrological data in Michigan in 1971: U.S. Geological Survey Report (unnumbered), 88 p.

Huffman, G.C., 1974a, Summary of ground-water hydrological data in Michigan in 1972: U.S. Geological Survey Report (unnumbered), 90 p.

Huffman, G.C., 1974b-88, Ground-water data for Michigan—1973-86: U.S. Geological Survey Report (unnumbered, published annually).

Huffman, G.C. and Whited, C.R., 1988, Ground-water data for Michigan—1987: U.S. Geological Survey Open-File Report 88-704, 56 p.

Huffman, G.C. and Whited, C.R., 1989, Ground-water data for Michigan—1988: U.S. Geological Survey Open-File Report 89-597, 54 p.

Huffman, G.C. and Whited, C.R., 1991, Ground-water data for Michigan—1989: U.S. Geological Survey Open-File Report 91-55, 51 p.

Huffman, G.C. and Whited, C.R., 1993, Ground-water data for Michigan—1990: U.S. Geological Survey Open-File Report 92-114, 51 p.

Jaquet, N.G., 1976, Ground-water and surface water relationships in the glacial province of northern Wisconsin—Snake Lake: Groundwater, v. 14, p. 194–199.

Jones, Henry & Williams Consulting Engineers, 1961, Kalamazoo, Michigan, Groundwater development in the Atwater Valley, Texas Township: Toledo, Ohio.

Jones, Linda G., 1992, Hydrogeologic characterization of a multi-aquifer system at station 11 in Kalamazoo, Michigan: M.S. Thesis, Western Michigan University.

Kehew, A.E., Nicks, L.P., and Straw, W.T., 1999, Palimpsest tunnel valleys: evidence for relative timing of advances in an interlobate area of Laurentide ice sheet: Annals of Glaciology 28, p. 47–52.

Lee, D.R., 1977, A device for measuring seepage flux in lakes and estuaries: Limnology and Oceanography, v. 22, p. 155–163.

Lerner, D.N., Issar, A.S., and Simmers, I., 1990, Groundwater recharge, a guide to understanding and estimating natural recharge: International Association of Hydrogeologists, Kenilworth, Rep. 8, 345 p.

Leverett, F, and Taylor, F., 1915, The Pleistocene of Indiana and Michigan and history of the Great Lakes: U.S. Geological Survey Monograph 53, 527 p.

Machavaram, M. and Krishnamurthy, R.V., 1995, Earth surface evaporative processes: a case study from the Great Lakes region of the Unites States based on deuterium excess in precipitation: Geochimica Cosmochimica Acta, v. 59, p. 4279–4283.

Martin, H.M., 1958, Outline of the geologic history of Kalamazoo County: Michigan Department of Conservation, Geological Survey Division, 16 p.

McBride, M.S. and Pfannkuch, H.O., 1975, The distribution of seepage within lakes: U.S. Geological Survey Journal of Research, v. 3, p. 505–512.

Meyboom, P., 1966, Unsteady groundwater flow near a willow ring in hummocky moraine: Journal of Hydrology, v. 4, p. 38–62.

Midwestern Regional Climate Center, 2004a, Monthly precipitation data for selected year, previous year, and 30-year average, climate data available on the Web, accessed May 17, 2004, at URL *http://mcc.sws.uiuc.edu/cgi-bin/Data-Ret/item2-3_2.cgi*

Midwestern Regional Climate Center, 2004b, Monthly snowfall data for selected year, previous year, and 30-year average, climate data available on the Web, accessed May 17, 2004, at URL *http://mcc.sws.uiuc.edu/cgi-bin/Data-Ret/item2-3_2.cgi*

Midwestern Regional Climate Center, 2004c, Monthly temperature data for selected year, previous year, and 30-year average, climate data available on the Web, accessed May 17, 2004, at URL *http://mcc.sws.uiuc.edu/cgi-bin/Data-Ret/item2-3_2.cgi*

Milstein, R.L., comp., 1987, Bedrock geology of Southern Michigan: Michigan Geological Survey, scale 1:500,000.

Monnett, V.B., 1948, Mississippian Marshall Formation of Michigan: American Association of Petroleum Geologists Bulletin, v. 32, p. 629–688.

Monaghan, G.W., and Larson, G.J., 1983, Surficial geology of Kalamazoo County map, in Selected geologic maps of Kalamazoo County, Michigan: County geologic map series, Michigan Department of Natural Resources, Geological Survey Division, 5 pls., scale 1:100,000.

——-1984, Glacial geology of Kalamazoo County, Michigan Sate University: Department of Geology publication, 37 p.

Monaghan, G.W., Larson, G.J., Forstat, D.W., and Sorenson, H.O., 1983, Selected geologic maps of Kalamazoo County, Michigan: County geologic map series, Michigan Department of Natural Resources, Geological Survey Division, 5 pls., scale 1:100,000.

Monaghan, G.W., Larson, G.J., and Gephardt, G.D., 1986, Late Wisconsinan drift stratigraphy of the Lake Michigan lobe in southwestern Michigan: Geological Society of America Bulletin, v. 97, p. 329–334.

Ohio Well Drilling Company, 1974, Evaluation of long term ground water supply of the metropolitan Kalamazoo area: Massillon, Ohio, variously paged.

Ohio Well Drilling Company, 1983a, Hydrogeologic survey of the Morrow Lake area for the City of Kalamazoo, Michigan—Electric resistivity survey and aquifer test on south side of Morrow Lake: Massillon, Ohio, variously paged.

Ohio Well Drilling Company, 1983b, Hydrogeologic survey of the Morrow Lake area for the City of Kalamazoo, Michigan—Electric resistivity survey and aquifer test on north side of Morrow Lake: Massillon, Ohio, variously paged.

Ohio Well Drilling Company, 1984, Report on the hydrogeologic study of station 19, City of Kalamazoo, Michigan: Massillon, Ohio, variously paged.

Ohio Well Drilling Company, 1987, Report on station 4, City of Kalamazoo, Michigan: Massillon, Ohio, variously paged.

Price, A.G. and Hendrie, L.K., 1983, Water motion in a deciduous forest during snowmelt: Journal of Hydrology, v. 64, p. 339–356.

Puzio, M.C., Larson, G.J., Rieck, R.L., Forstat, D.W., and Sorenson, H.O., 1983, Selected geologic maps of Allegan County, Michigan: County Geologic Map Series, Michigan State Geological Survey.

Reed, J.E., Deutsch, Morris, and Wiitala, S.W., 1966, Induced recharge of an artesian glacial-drift aquifer at Kalamazoo, Michigan: U.S. Geological Survey Water-Supply Paper 1594-D, 62 p.

Rheaume, S.J., 1990, Geohydrology and water quality of Kalamazoo County, Michigan, 1986–1988: U.S. Geological Survey Water-Resources Investigations Report 90-4028, 102 p.

Rosenberry, D.O., 1985, Factors contributing to the formation of transient water-table mounds on the outflow side of a seepage lake, Williams Lake, Central Minnesota: M.S. Thesis, University of Minnesota.

Rosenberry, D.O., 1990, Inexpensive ground-water monitoring methods for determining hydrologic budgets of lakes and wetlands: in Enhancing States' Lake/Wetland Programs, p. 123–131.

Rushton, K., 1997, Recharge from permanent water bodies. In: Simmers, I. (ed) Recharge of phreatic aquifers in (semi)arid areas: AA Balkema, Rotterdam, p. 215–255.

Rutledge, A.T., 1993, Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records: U.S. Geological Survey Water-Resources Investigations Report 93-4121, 45 p.

Rutledge, A.T., 1998, Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records: U.S. Geological Survey Water-Resources Investigations Report 98-4148, 43 p.

Scanlon, Bridget R., Healy, Richard W., and Cook, Peter G., 2002, Choosing appropriate techniques for quantifying groundwater recharge: Hydrogeology Journal, v. 10, p. 18–39.

Straw, W.T., 1978, Bedrock Geology, in Passero, R.N., principal editor, Kalamazoo County, geology and the environment: Western Michigan University, p. 25–34.

Sugden, D.E., and John, B.S., 1976, Glaciers and Landscape: Great Britain, Edward Arnold, Division of Hodder and Stoughton Limited, 376 p.

Westjohn, D.B., and Weaver, T.L., 1996, Hydrogeologic framework of Mississippian rocks in the central Lower Peninsula of Michigan: U.S. Geological Survey Professional Paper 1418, 47 p.

Williams, R.E., 1968, Flow of groundwater adjacent to small, closed basins in glacial till: Water Resources Research, v. 4, p. 777–784.

Winter, T.C., 1983, The interaction of lakes with variably saturated porous media: Water Resources Research, v. 19, p. 1203–1218.

Winter, T.C., LaBaugh, J.W., and Rosenberry, D.O., 1988, The design and use of a hydraulic potentiomanometer for direct measurement of differences in hydraulic head between ground-water and surface water: Limnology and Oceanography, v. 33, n. 5, p. 1209–1214.

Zager, J.P., 1981, The interaction of surface and groundwater in lower Nashotah Lake, Waukesha County, Wisconsin: M.S. Thesis, University of Wisconsin.

Table 1. Properties of aquifers and confining units, Kalamazoo County, Michigan.

[--, estimate not available; >, greater than]

Well location	Estimated transmissivity (gallons/day/foot)	Estimated perme-ability (feet/day)	Storage coefficient/ specific yield	Geologic unit	Source
2S 11W 10-46	18,000-24,000	47	--	intermediate aquifer	Deutsch and others, 1960.
2S 11W 22-51	>100,000	84	--	intermediate aquifer	Deutsch and others, 1960.
2S 11W 22-91	300,000	200	0.17	intermediate aquifer	Deutsch and others, 1960.
3S 11W 14-1	65,000	97	--	intermediate aquifer	Deutsch and others, 1960.
3S 11W 3-3	42,000	112	.00024	intermediate aquifer	Deutsch and others, 1960.
3S 11W 4-3	110,000	147	--	intermediate aquifer	Deutsch and others, 1960.
1S 11W 34-7	230,000	1,230	.00027	intermediate aquifer	Deutsch and others, 1960.
3S 11W 9-2	130,000	122	--	intermediate aquifer	Deutsch and others, 1960.
1-9-29-5	150,000	214	.02	upper aquifer	Allen and others, 1972.
10-14-3	71,000	254	--	upper aquifer	Allen and others, 1972.
11-34-7	190,000	307	.00006	intermediate aquifer	Allen and others, 1972.
34-40	130,000	348	.02	intermediate aquifer	Allen and others, 1972.
2-11-6-1	35,000	227	.0051	intermediate aquifer	Allen and others, 1972.
11-14-5	80,000	348	.0024	intermediate aquifer	Allen and others, 1972.
20-12	60,000	321	.00057	intermediate aquifer	Allen and others, 1972.
22-4	79,000	107	.19	intermediate aquifer	Allen and others, 1972.
29-2	44,000	160	.00047	intermediate aquifer	Allen and others, 1972.
5	46,000	120	.0037	intermediate aquifer	Allen and others, 1972.
31-2	110,000	414	.0018	intermediate aquifer	Allen and others, 1972.
3-11-3-3	40,000	94	.0024	intermediate aquifer	Allen and others, 1972.
11-4-3	105,000	321	.0037	intermediate aquifer	Allen and others, 1972.
4-10	52,000	468	.0016	intermediate aquifer	Allen and others, 1972.
40	110,000	428	.03	upper aquifer	Allen and others, 1972.
42	100,000	374	.04	intermediate aquifer	Allen and others, 1972.
9-2	128,000	107	.0002	intermediate aquifer	Allen and others, 1972.
14-101	84,000	374	--	intermediate aquifer	Allen and others, 1972.
12-11-4	150,000	401	.0064	intermediate aquifer	Allen and others, 1972.
4-11-13-34	54,000	107	.0004	intermediate aquifer	Allen and others, 1972.
11-21-4	100,000	174	--	intermediate aquifer	Allen and others, 1972.
12-25-3	100,000	147	--	intermediate aquifer	Allen and others, 1972.
Kalamazoo River ground-water reservoir	20,000-120,000	19-401	.2	intermediate aquifer	Allen and others, 1972.
Schoolcraft ground-water reservoir	40,000-80,000	67-535	.2	upper aquifer	Allen and others, 1972.
Schoolcraft ground-water reservoir	10,000-100,000	17-167	.005	intermediate aquifer	Allen and others, 1972.
Kalamazoo-Portage ground-water reservoir	10,000-100,000	22-1337	.2	upper aquifer	Allen and others, 1972.

Table 1. Properties of aquifers and confining units, Kalamazoo County, Michigan.—Continued

[--, estimate not available; >, greater than]

Well location	Estimated transmissivity (gallons/day/ foot)	Estimated perme- ability (feet/day)	Storage coefficient/ specific yield	Geologic unit	Source
Kalamazoo-Portage ground-water reservoir	10,000-160,000	19-306	0.005	intermediate aquifer	Allen and others, 1972.
Cooper ground-water reservoir	20,000-80,000	45-1069	--	intermediate aquifer	Allen and others, 1972.
Morrow Lake ground-water reservoir	40,000-80,000	89-1069	--	intermediate aquifer	Allen and others, 1972.
Augusta-Galesburg ground-water reservoir	20,000-80,000	45-1069	--	intermediate aquifer	Allen and others, 1972.
Alamo-Ostemo ground-water reservoir	20,000-60,000	27-802	--	upper aquifer	Allen and others, 1972.
Sherman Lake ground-water reservoir	20,000-140,000	33-1872	--	upper aquifer	Allen and others, 1972.
Vicksburg ground-water reser- voir	20,000-40,000	200.5	.2	upper aquifer	Allen and others, 1972.
Vicksburg ground-water reser- voir	20,000-60,000	134	.005	intermediate aquifer	Allen and others, 1972.
Texas ground-water reservoir	20,000-80,000	100	.2	upper aquifer	Allen and others, 1972.
Texas ground-water reservoir	20,000-140,000	213.9	.005	intermediate aquifer	Allen and others, 1972.
Kalamazoo, Station 19	20,000	66.8	--	intermediate aquifer	Allen and others, 1972.
Kalamazoo, Station 4	367,000	748.7	.15	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 8	91,400	347.6	.00035	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 9	110,000	367.6	.00036	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 11	73,300	327.00	.0027	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 12	55,500-57,300	83.7	.000099	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 13	140,000	467.9	--	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 14	128,000	285	.043	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoon, Station 15	103,000	393	--	intermediate aquifer	Ohio Drilling Company, 1974.
Kalamazoo, Station 22	201,000	163	.000054	intermediate aquifer	Ohio Drilling Company, 1974.
Central Pumping Station	427,000	407.8	--	intermediate aquifer	Ohio Drilling Company, 1974.
Atwater Well Field	260,000	--	.0063	intermediate aquifer	Ohio Drilling Company, 1974.
Atwater Well Field	--	1.29	--	upper confining unit	Ohio Drilling Company, 1974.
Lake Campbell Well Field	168,950	161-301	.164	upper aquifer	Ohio Drilling Company, 1974.
Morrow Lake (South Side)	109,371	1,462	.134	intermediate aquifer	Ohio Drilling Company, 1983a.
Morrow Lake (North Side)	25,172	224	.00237	intermediate aquifer	Ohio Drilling Company, 1983b.
Atwater Valley, Texas Township	130,000	--	.0025	intermediate aquifer	Henry Jones and Williams Con- sulting Engineers, 1961.
Report on Station 4	83,000	170.7	.0003	intermediate aquifer	Ohio Drilling Company, 1987.
Report on study of Station 19	9,360	31	.00625	intermediate aquifer	Ohio Drilling Company, 1984.
Station 11 Report	73,630	328	.00012	intermediate aquifer	Linda G. Jones, 1992.
Station 11 Report	61,830	276	.000198	intermediate aquifer	Linda G. Jones, 1992.
Station 11 Report	--	.001	--	lower confining unit	Linda G. Jones, 1992.

Table 1. Properties of aquifers and confining units, Kalamazoo County, Michigan.—Continued

[--, estimate not available; >, greater than]

Well location	Estimated transmissivity (gallons/day/foot)	Estimated perme-ability (feet/day)	Storage coefficient/specific yield	Geologic unit	Source
Station 11 Report	55,900-140,300	249-625	.00012-.00145	intermediate aquifer	Linda G. Jones, 1992.
Station 11 Report	50,200-53,600	224-239	.000326-.00034	intermediate aquifer	Linda G. Jones, 1992.
Station 11 Report	--	0.0017-0.96	--	lower confining unit	Linda G. Jones, 1992.
Station 11 Report	--	.072-13.6	--	upper aquifer	Linda G. Jones, 1992.
Station 11 Report	--	.215-33	--	intermediate aquifer	Linda G. Jones, 1992.
Station 11 Report	--	.34-6.36	--	lower aquifer	Linda G. Jones, 1992.
Ross Township Site	154,800	151-689	0.1	upper aquifer	City of Kalamazoo Department of Public Services, 1995.
Ross Township Site	--	.022	--	upper confining unit	City of Kalamazoo Department of Public Services, 1995.
Ross Township Site	206,075	190-530	.00069	intermediate aquifer	City of Kalamazoo Department of Public Services, 1995.
Delineation Report - stations 1, 2, 3, 4, & 7	21,400-403,750	202.9	--	intermediate aquifer	City of Kalamazoo Department of Public Services and Peer-less-Midwest Company, Inc., 1999a.
Delineation Report - stations 11 & 12	21,400-403,750	32-600	--	intermediate aquifer	City of Kalamazoo Department of Public Services and Peer-less-Midwest Company, Inc., 1999b.
Long Lake Pumping Well Site	306,100	422	.00373	intermediate aquifer	Bob Snell, written commun, 2002.
Long Lake Pumping Well Site	332,600	458	.00054	intermediate aquifer	data collected during this study.
Long Lake Pumping Well Site	297,800	410	.00316	intermediate aquifer	data collected during this study.
Long Lake Pumping Well Site	--	284	--	upper aquifer	data collected during this study.
Long Lake Pumping Well Site	--	.468	--	upper confining unit	data collected during this study.
Long Lake Pumping Well Site	--	435	--	intermediate aquifer	data collected during this study.

Table 2. Observation well water levels, measured during synoptic surveys, Kalamazoo, Michigan.

[Measurements referenced to feet above NGVD 29; --, measurement not available]

Local name	August 6, 2001	December 4, 2001	March 25, 2002	June 12, 2002	September 3, 2002
K-23 at Barton Lake	837.09	838.17	838.63	838.6	837.35
K-20 Fulton & West Avenue	897.44	897.61	897.85	897.98	896.92
K-17 10th Street & West Avenue	870.46	871.78	872.6	872.47	870.65
K-19 36th & West Avenue	865.79	867.52	868	867.6	865.17
K-24 Vicksburg Waste Water Treatment Plant	835.83	836.78	836.88	836.56	835.88
K-21 Fulton Elementary School	898.6	899.11	900	900.25	898.15
Prairie View Park (recorder)	850.07	850.13	851.47	851.23	850.15
K-16 4th Street Gravel Pit	887.24	887.8	889.02	889.09	888.13
K-25 Austin Lake Airport	852.24	853.64	854.08	853.63	852.43
K-15 S Avenue & 37th Street	905.75	907.46	907.43	906.65	904.45
K-22 R. Avenue & 46th Street	938.51	941.35	941.44	940	936.07
Ramona Park at Long Lake	847.94	850.74	852.26	851.48	849.69
K-13 Q. Avenue & 35th Street	893.96	894.8	895.38	894.62	892.85
Gourdneck State Game Area (S) (recorder)	865.75	865.72	866.52	866.52	866.08
Gourdneck State Game Area (D) (recorder)	867.14	866.91	868.13	867.9	867.66
Portage Central High School (recorder)	848.39	847.2	--	849.57	848.78
K-11 off Centre near Oakland	862.44	863.26	863.39	863.77	862.92
K-12 Q. Avenue & 29th Street	855.7	856.99	857.5	856.7	855.02
K-10 Eagle Lake	896.1	896.68	896.89	897.43	896.63
Kalamazoo County (Atwater)	878.44	878.97	880.17	880.03	879.38
Kalamazoo County (SABO-D)	862.36	865.53	865.4	863.25	862.7
Kalamazoo County (SABO-S)	864.45	865.69	866.24	865.62	864.95
K-14 44th Street (American Legion)	946.98	947.78	948.04	947.56	946.27
Lexington Green (Sprinkle Road)	841.64	843.05	844.04	--	--
Kalamazoo County (K-32-S) Kilgore Road #9	844.36	844.28	850.51	850.24	846.54
Kalamazoo County (K-32-D) Kilgore Road #9	841.22	841.81	848.39	848.25	845.31
Kalamazoo County (Colony) (near US-131)	855.3	854.96	857.63	857.13	851.68
Kalamazoo County (Emerald) (recorder)	831.08	829.74	832.85	832.55	831.07
AL 3-D	872.39	871.47	873.77	874.27	873.94
AL 4-S	872.73	872.03	873.94	874.21	873.84
Kalamazoo County (Morrow)	778.21	781.93	779.39	778.37	780.85
Kalamazoo County (Maple)	771.25	781.21	769.66	769.32	769.25
K-8 Olmstead & Lake Street	773.33	774.64	774.59	--	773.42
Kazoo County (Crosstown & Stockbridge)	752.87	754.32	761.25	757.17	754.98
K-7 M-96 (River Oaks Park)	785	786.22	786.7	785.9	785
Kalamazoo County (Kendall) Deep (recorder)	834.6	838.4	837.46	836.71	835.85
K-9 1st Street & M-43	793.75	--	--	--	--
K-30 Crum Park (37th Street)	822.27	823.89	824	823.5	822.03
Kalamazoo County (Campbell)	830.95	830.72	833.54	833.32	832.45
K-29 G Avenue & 27th Street	870.1	872.19	873	873.95	874.08

Table 2. Observation well water levels, measured during synoptic surveys, Kalamazoo, Michigan.—Continued

[Measurements referenced to feet above NGVD 29; --, measurement not available]

Local name	August 6, 2001	December 4, 2001	March 25, 2002	June 12, 2002	September 3, 2002
K-28 Maple Glen Park (B Avenue & Westnedge)	762.1	763.3	764.3	763.1	762.26
K-5 42nd Street south of 27th Avenue	799.35	799.8	800.1	--	799.45
K-1 southbound US-131 Rest Area	837.34	837.88	838.5	838.6	--
K-3 DE Avenue & Old railroad	810.92	811.66	811.8	814.65	811.38
K-34 48th Street & M-89	806.87	807.76	808.2	808.75	806.65
K-4 C Avenue & M-89	898.09	898.72	899.2	900	899.18
K-27 Rupert Lake (B Avenue)	712.96	714.55	714.7	714.45	713.59
K-18 XY Avenue & US-131	855.5	856.5	857	--	--
KCRC #22 RS Avenue	799.32	799.48	800	--	--
KCRC #26 (P Avenue)	848.98	--	--	--	--
KCRC #21 Pretty Lake Boy's Camp	890.73	891.11	891.89	892.67	891.35
KCRC #2 38th Street near M-89	851.22	--	--	--	--
KCRC #8 1st Street north of KL Avenue	794.69	--	--	--	--
KCRC #17 1st Street north of P Avenue	885.73	885.95	886.65	887.9	887.4
KCRC #16 V Avenue at 42nd Street	892.87	893.48	893.9	893.61	892.18
KCRC #24 (Centre Street)	844.52	844.56	849.28	843.86	--
VB-22 24th Street (652) & 90th Avenue	894.65	894.92	895.68	896.85	896.34
VB-23 38th Street, north of 78th Avenue	750.93	--	--	--	--
VB-17 Paw Paw Road, north of 38th Street	733.47	--	--	--	--
VB-15	768.88	769.29	--	--	--

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Luukkonen, C.L., Blumer, S.P., Weaver, T.L., and Jean, J.—**Simulation of the Ground-Water-Flow System in the Kalamazoo County Area, Michigan**—Scientific Investigations Report 2004-5054

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