Michigan Department of Environmental Quality

Hydrogeology and Simulation of Regional Ground-Water-Level Declines in Monroe County, Michigan

Water-Resources Investigations Report 03-4312

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

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By Howard W. Reeves, Kirsten V. Wright, and J. R. Nicholas

In cooperation with the Michigan Department of Environmental Quality

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CONVERSION FACTORS, ABBREVIATIONS, AND DATUM

Multiply	By	To Obtain
foot (ft)	0.3048	meter
mile (mi)	1610	meters
square mile (mi^2)	2.590	square kilometer
acre	4046.9	square meter
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
million gallon per day (Mgal/d)	0.04382	cubic meter per second
foot per day (ft/d)	0.3048	meter per day
foot per second (ft/s)	0.3048	meter per second
square foot per day (ft^2/d)	0.09290	square meter per day
inch per year (in/yr)	0.0254	meter per year

Abbreviations

DNR	Department of Natural Resources
MDEQ	Michigan Department of Environmental Quality
NLCD	National Land Coverage Database
NPDES	National Pollutant Discharge Elimination System
RASA	Regional Aquifer-System Analysis
USGS	U.S. Geological Survey

Datum

Vertical coordinate information is referenced to the North Geodetic Vertical Datum of 1929 (NGVD 29). Horizontal coordinate information from various sources was projected into the Michigan Georef Coordinate system and referenced to the North American Datum of 1983 (NAD 83).

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ABSTRACT

Observed ground-water-level declines from 1991 to 2003 in northern Monroe County, Michigan, are consistent with increased ground-water demands in the region. In 1991, the estimated ground-water use in the county was 20 million gallons per day, and 80 percent of this total was from quarry dewatering. In 2001, the estimated ground-water use in the county was 30 million gallons per day, and 75 percent of this total was from quarry dewatering.

Prior to approximately 1990, the ground-water demands were met by capturing natural discharge from the area and by inducing leakage through glacial deposits that cover the bedrock aquifer. Increased ground-water demand after 1990 led to declines in ground-water level as the system moves toward a new steady-state. Much of the available natural discharge from the bedrock aquifer had been captured by the 1991 conditions, and the response to additional withdrawals resulted in the observed widespread decline in water levels.

The causes of the observed declines were explored through the use of a regional ground-water-flow model. The model area includes portions of Lenawee, Monroe, Washtenaw, and Wayne Counties in Michigan, and portions of Fulton, Henry, and Lucas Counties in Ohio. Factors, including lowered water-table elevations because of below average precipitation during the time period (1991 - 2001) and reduction in water supply to the bedrock aquifer because of land-use changes, were found to affect the regional system, but these factors did not explain the regional decline. Potential ground-water capture for the bedrock aquifer in Monroe County is limited by the low hydraulic conductivity of the overlying glacial deposits and shales and the presence of dense saline water within the bedrock as it dips into the Michigan Basin to the west and north of the county. Hydrogeologic features of the bedrock and the overlying glacial deposits were included in the model design. An important step of characterizing the bedrock aquifer was the determination of inputs and outputs of water—leakage from glacial deposits and flows across model boundaries. The imposed demands on the ground-water system create additional discharge from the bedrock aquifer, and this discharge is documented by records and estimates of water use including: residential and industrial use, irrigation, and quarry dewatering.

Hydrologic characterization of Monroe County and surrounding areas was used to determine the model boundaries and inputs within the ground-water model. MODFLOW-2000 was the computer model used to simulate ground-water flow. Predevelopment, 1991, and 2001 conditions were simulated with the model. The predevelopment model did not include modern water use and was compared to information from early settlement of the county. The 1991 steady-state model included modern demands on the ground-water system and was based on a significant amount of data collected for this and previous studies. The predevelopment and 1991 simulations were used to calibrate the numerical model. The simulation of 2001 conditions was based on recent data and explored the potential ground-water levels if the current conditions persist. Model results indicate that the ground-water level will stabilize in the county near current levels if the demands imposed during 2001 are held constant.

INTRODUCTION

In January 2002, the U.S. Geological Survey (USGS) published a report of hydrologic data documenting widespread ground-water-level declines in Monroe County, Michigan from 1991 to 2001 (Nicholas and others, 2002). Many of the 34 USGS monitoring wells in the county had declines ranging from 10 to 20 ft during this time period. Long-term ground-water-level declines result from a decrease in recharge, an increase in discharge by either natural or induced processes, or a combination of the two. Additional hydrologic data presented by Nicholas and others (2002), including precipitation, Lake Erie levels, and base flow to streams, did not show a systematic change consistent with a decrease in recharge. Ground-water use, however, increased appreciably from 1991 to 2001.

In March 2002, USGS and Michigan Department of Environmental Quality (MDEQ) entered into a cooperative agreement to study the hydrogeology and determine the cause or causes of the widespread ground-water-level declines in Monroe County. USGS and MDEQ agreed that the primary investigative tool should be a computer model—a numerical model of ground-water flow—suitable for evaluating changes in recharge and discharge and the effects of these changes on ground-water levels in Monroe County.

PURPOSE AND SCOPE

The purpose of this report is to investigate probable causes of ground-water-level declines in Monroe County, Michigan. The report focuses on the conceptual and numerical models of ground-water flow, including assumptions, model calibration, model simulations, and model limitations. Field data, both collected for this study and previously presented by Nicholas and others (2002), also are presented and analyzed to describe the hydrogeologic system. Potential reasons for the observed decline in ground-water levels are offered based on analysis of the field data and simulation results.

Description of the Study Area

The study area focuses on Monroe County in southeastern Michigan. The ground-water system, however, is regional, and the study area includes portions of Lake Erie to the east, Wayne and Washtenaw Counties to the north, Lenawee County to the west, and Lucas, Fulton, and Henry Counties, Ohio to the south (fig. 1).

Monroe County is relatively flat. The land-surface altitude ranges from 572 ft above NGVD 29 at Lake Erie to 740 ft in the northwestern corner of the county. Most of the study area is similar topographically to Monroe County; glacial moraines in the western part of the study area result in higher relief than the rest of the area. The land area of Monroe County is 550 mi^2; the study area is approximately 3,000 mi^2. The population of Monroe County in 2000 was 145,950 (U.S. Census Bureau, 2001).

Major rivers in the study area include the Detroit River, Huron River, River Raisin, Swan Creek, Stony Creek, and the Maumee River in Ohio. The Detroit and Maumee Rivers are considered the northeastern and southern borders of the study area, respectively. Historically, much of the study area was wetlands. These wetlands were drained through the installation of drainage tiles and tile drains in the 1800's to allow for agriculture, and there still are many drainage ditches and tile systems in the area.

Previous Studies

Historical studies that relate to this work were published by Sherzer (1900), Leverett (1915), Mazola (1970), Twenter (1975), and Allen (1977). Geologic maps showing glacial and bedrock stratigraphy in Michigan were developed by Farrand and Bell (1982) and Milstein (1987). These maps are used as base information for the Michigan portion of the study area. The Ohio Department of Natural Resources (2000a, 2000b) developed unconsolidated and bedrock aquifer maps that are used as base information for the Ohio portion of the study area. Breen (1989, 1991) provides recent ground-water information for northwestern Ohio.

Nicholas and others (1996) present a USGS study of the ground-water resources of Monroe County. Most of the monitoring wells used in the present study were installed as part of the previous USGS study. Nicholas and others (1996) provide the baseline ground-water-level data used for determining subsequent ground-water-level declines in Monroe County. They also provide water-quality information and a water-use summary for the county. Nicholas and others (2002) document the ground-water-level declines in the county from 1991 to 2001 using water levels measured in 34 monitoring wells by the U.S. Geological Survey (USGS) and the Monroe County Health Department (fig. 2).

Two USGS Regional Aquifer-System Analysis (RASA) projects provide the regional context for the simulation of ground-water flow described in this report. Westjohn and Weaver (1998) present the hydrogeologic framework for the Michigan Basin RASA. The Midwestern Basins and Arches RASA (Bugliosi, 1999) provides additional regional information especially relevant to the study area boundaries. The ground-water-flow model developed for the Midwestern Basins and Arches RASA overlaps slightly with the study area described here; however, the scale of that RASA study is too large to be used directly in the analysis of the observed ground-water-level declines in Monroe County.

Acknowledgments

The authors thank Elgar Brown and Brant Fisher from MDEQ for their technical support during this project. USGS Lansing Field Office staff, including Brian Heissenberger, Tom Morgan, and Daniel Obenauer, installed and maintained field equipment and monitored ground-water levels. USGS Network Operations staff, Rose McGowan and Suzanne Crowley, managed the databases used for water-level and water-quality data. Sharon Baltusis, Rose McGowan, and Michele Morenz provided important assistance in report preparation. Allen Shapiro, USGS, and Steven Wright, University of Michigan, helped with the instrumentation of wells near London Aggregates to monitor water levels after the quarry stopped pumping and provided additional assistance on this study aspect. The cooperation of Suzanne Hanf from SE Johnson and the personnel at London Aggregates for allowing instrument installation and equipment maintenance at their facility is appreciated. Jeff Stoll from Hanson Aggregates Midwest provided observation-well data. Water-level data collected by the Monroe County Health Department from 1993 to 2001 was invaluable for this study. The algebraic multigrid solver used in the

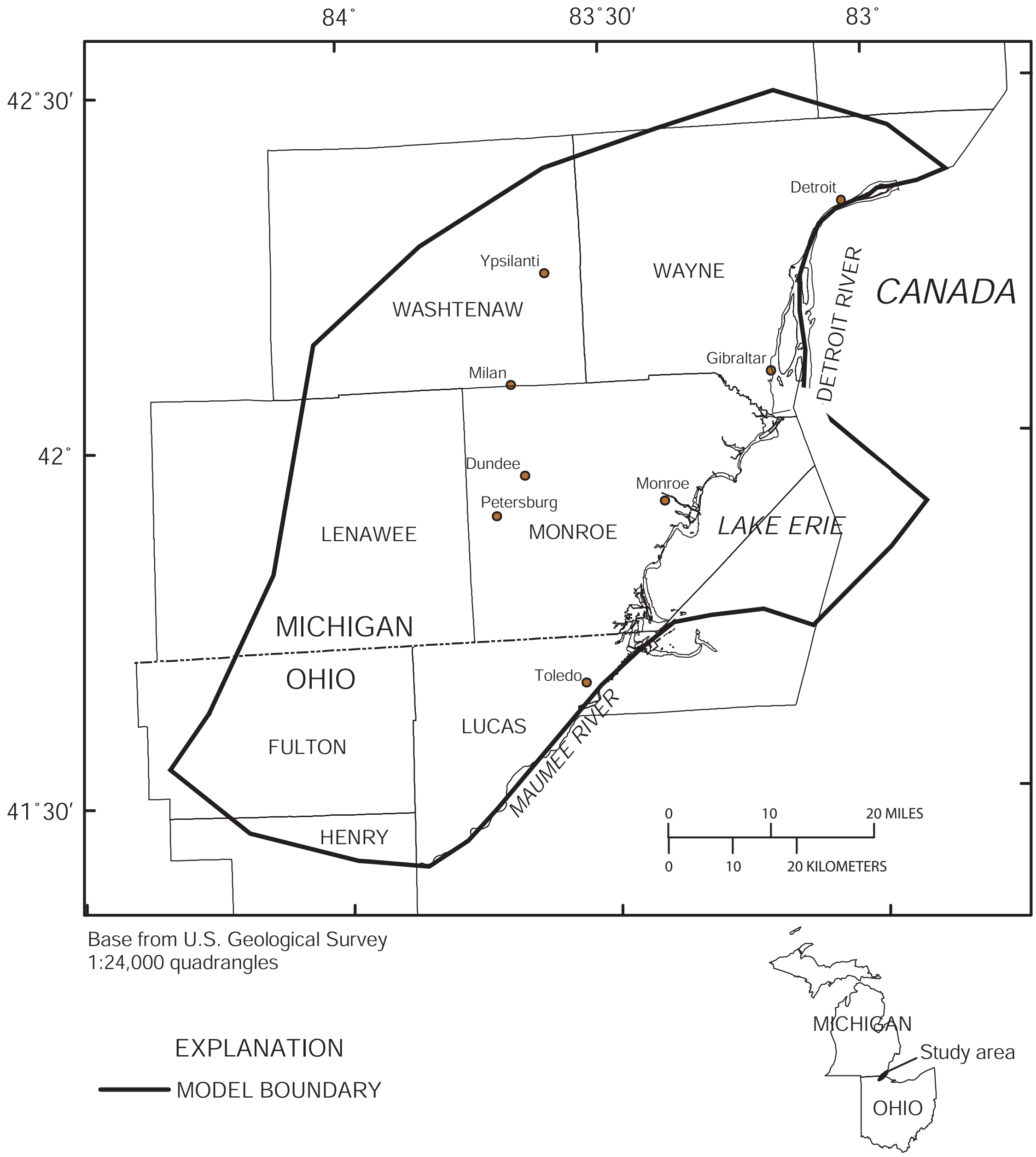


Figure 1. Study area in Monroe County and surrounding counties in Michigan and Ohio.

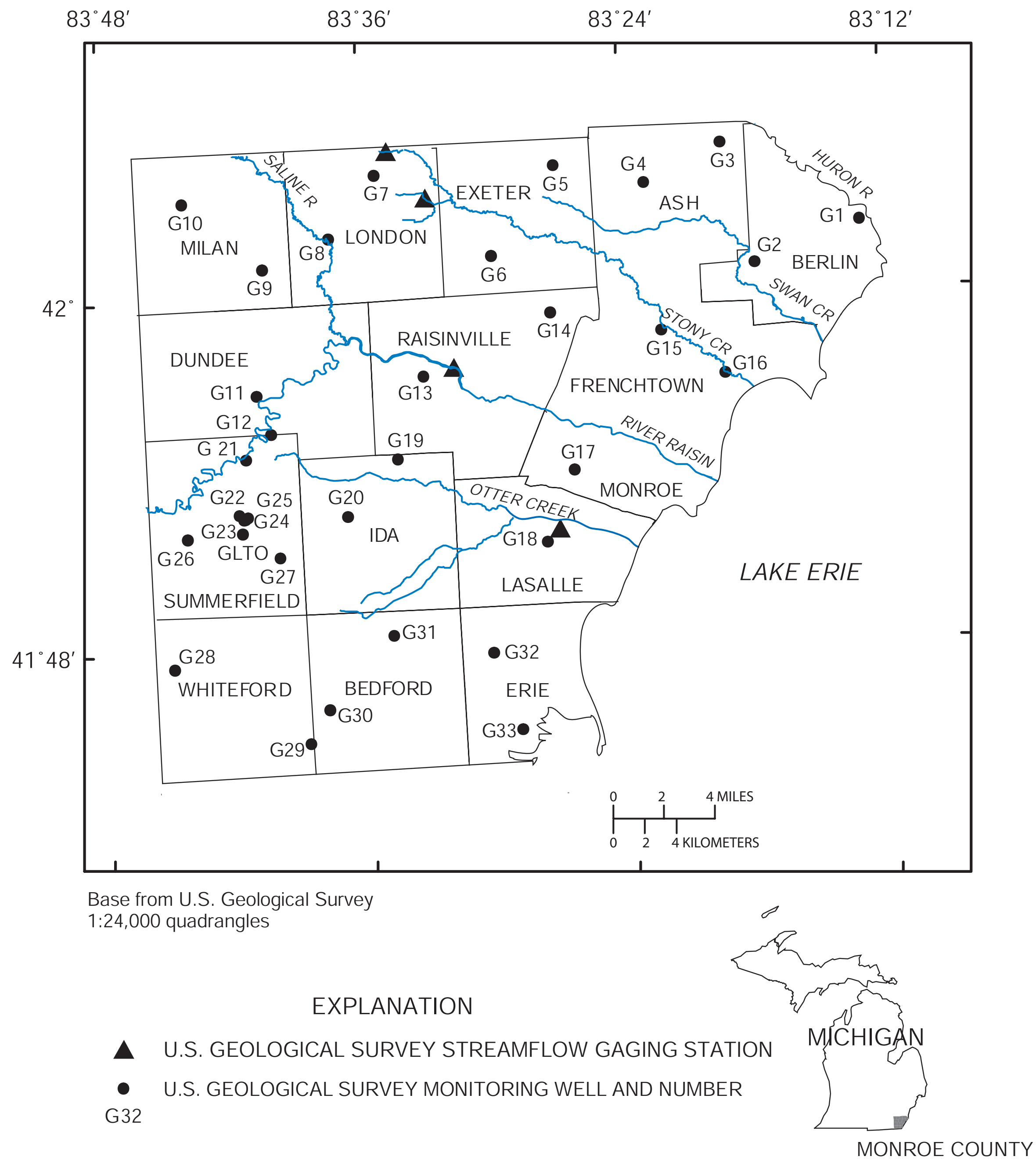


Figure 2. U.S. Geological Survey streamflow-gaging stations and monitoring wells in Monroe County, Michigan.

link-AMG package for MODFLOW-2000 (Mehl and Hill, 2001) was developed by the German National Research Center for Information Technology (GMD).

HYDROGEOLOGY

Understanding the hydrogeology of the bedrock aquifer is necessary for simulation of ground-water flow and investigation of water-level declines. The bedrock aquifer provides most of the ground-water resources in Monroe County, and the ground-water-level declines are in this aquifer. The bedrock aquifer is composed of several rock units dominated by dolomite and limestone with small amounts of sandstone. Glacial deposits overlie the bedrock aquifer in most of Monroe County and the study area. In Monroe County, the glacial deposits yield only small to moderate quantities of water to wells. In the northwestern part of the study area, the glacial deposits may yield more water than other parts of the area and provide drinking-water sources for most domestic wells. In this study, the glacial deposits are considered primarily with respect to how much water they transmit to or receive from the bedrock aquifer. The hydrogeologic features of the glacial deposits and bedrock aquifers are discussed below.

Glacial Deposits

Glacial deposits in the study area include unconsolidated silt, clay, and gravel deposited in landforms known as moraines; clay-rich glacial till; and glaciolacustrine sand and clay (fig. 3). The glacial deposits locally may act as aquifers where hydrogeologic properties allow them to store and transmit usable quantities of ground water. Clay-rich deposits often adequately store water, however, they typically limit the movement of ground water and are considered confining layers in the study area. In this study, individual glacial aquifers are not identified or simulated.

Only the upper portion of the glacial deposits are depicted in figure 3 (Farrand and Bell, 1982). These deposits, however, have considerable three-dimensional complexity. Typically, the bedrock surface is covered by clay-rich glacial till (Mozola, 1970), and the till is covered by the deposits shown in figure 3. Thus, where lacustrine sand deposits are shown, these deposits may be relatively thin compared to the underlying clay-rich till and separated from the bedrock. Glacial deposits are absent in the study area in small areas where rivers have exposed the bedrock near Lake Erie or where the glacial deposits have been removed to expose the bedrock for quarrying or construction. The detailed glacial geology at any location can be determined only by site-specific geologic or geotechnical investigation.

Hydraulic properties

Despite the heterogeneity of the glacial deposits, some general statements regarding expected hydraulic properties can be made. The capacity of a hydrogeologic unit to transmit water is described by the hydraulic conductivity. This hydraulic property may vary over many orders of magnitude, and the capacity of the unit to transmit water increases as the conductivity increases. The deposits in Monroe County are dominated by glaciolacustrine sand and clay (fig. 3). Both of these deposits are expected to have low hydraulic conductivity, although the glaciolacustrine sands may be able to transmit water better than the clays. The glacial deposits that should have the highest hydraulic conductivity in the study area are coarse moraine deposits in the northwestern and western portions of the study area. These deposits are predominately sand and may contain gravel layers. The glacial deposits are important to this study because they exchange water with the underlying bedrock aquifer.

The effective vertical hydraulic conductivity of glacial deposits was estimated using information from digital water-well logs compiled by the MDEQ in the electronic water-well-log database referred to as WelLogic (Michigan Department of Environmental Quality, 2003). Water-well drillers are required to log every well drilled in the State, and many drillers use the WelLogic system to enter well-log information. WelLogic defines a set of descriptions for the material types (lithologies) encountered during drilling. The lithologies reported on many older logs also have been converted to a standard set of descriptions by MDEQ and entered into the database. The water-well logs in the database contain information regarding the location and date drilled, construction details, static water level, and the lithologic description of the boring as reported by the driller. There were 8,759 water-well logs in the WelLogic database for Monroe County in May 2003. Note that these are not all of the well records for the county, some areas have not had many new wells drilled or have had only a few older well logs converted to the database, and, therefore, have few entries in the WelLogic system.

To estimate the effective vertical hydraulic conductivity, the thicknesses of sand, gravel, and clay were identified for each water-well log in Monroe, Lenawee, Washtenaw, and Wayne Counties where the well reached the bedrock aquifer. Only lithologies are identified on the well logs, and material tests were not performed to determine the vertical hydraulic conductivity of each lithology.

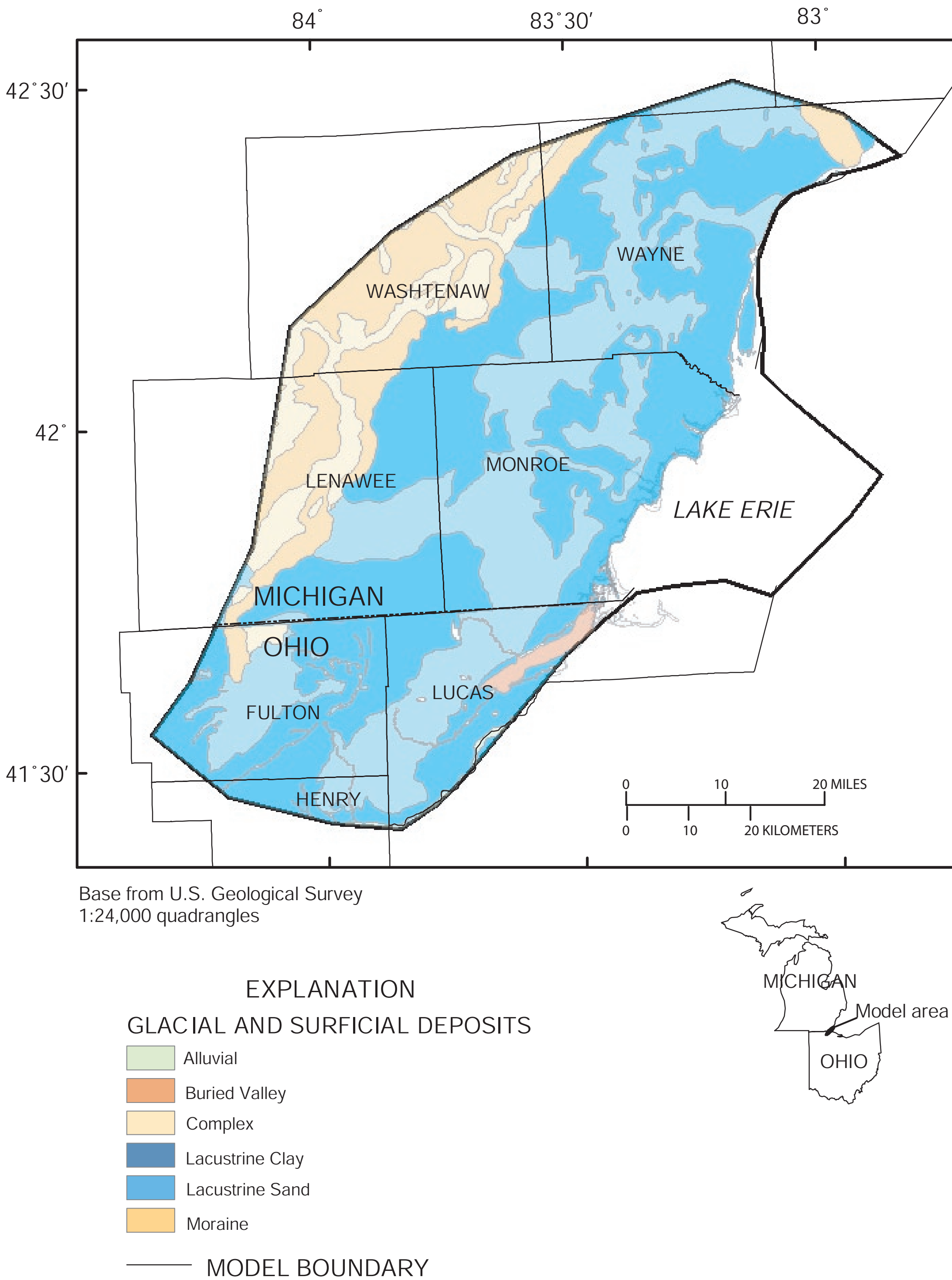


Figure 3. Generalized glacial deposit map for study area, modified from Ferrand and Bell (1982) and Ohio Department of Natural Resources (2000b).

Therefore, representative vertical hydraulic conductivities (Fetter, 2001) were assumed for the following lithologic classes: gravel = 3 x 10^{-3} ft/s, sand = 3 x 10^{-6} ft/s, and clay = 3 x 10^{-10} ft/s. The effective vertical hydraulic conductivity, K_v^{eff}, for the lithologic sequence described in each water-well log was calculated as

$$K_v^{eff} = \frac{b}{\sum_{m=1}^{n} \frac{b_m}{K_{v_m}}},$$

where b_m is the thickness of lithologic layer m, and n is the total number of lithologic layers for a given well (Fetter, 2001). Because the vertical hydraulic conductivity of each layer is assigned in this study, and not measured directly, the distribution of effective vertical hydraulic conductivities is indicative of relative values and not of absolute values in specific locations. The distribution of effective vertical hydraulic conductivity (fig. 4) indicates that the movement of water through glacial deposits is limited over most of Monroe County and that the area with greatest potential to transmit water is in the south-central part of the county. Comparison of figures 3 and 4 shows that even areas mapped as having glaciolacustrine sand at the surface may have a sufficient thickness of underlying clay to restrict vertical flow.

Water-level changes

Water levels in glacial deposits are important, as they affect the amount of water exchanged between the glacial deposits and the bedrock aquifer. If the water level in the glacial deposits is higher than that in the bedrock aquifer, then there is a hydraulic potential for water to flow from the glacial deposits to the bedrock. Most wells in the glacial deposits are not screened across the water table, and, therefore, because of the potential for vertical gradients within the glacial deposits, there are few available direct measurements of the level of the water table in the study area. The water-table surface, however, is commonly assumed to be shaped similar to the land surface (Fetter, 2001). In areas with tile-drain systems, the water table is near the land surface. The water table is at the same elevation as streams, lakes, and wetlands, and generally at a slightly higher elevation between these surface-water features. In the study area, perched water in the glacial deposits is common (Nicholas and others, 1996). Perched ground water is ground water that is above the water table and separated from the water table by an unsaturated zone. Perched ground water typically occurs in sand deposits that are underlain by clay, but they also may be present in any glacial deposit where the water level in the bedrock aquifer is below the glacial deposit/bedrock interface.

One USGS monitoring well, G-10, is completed in glacial deposits (fig. A10). This well is screened just above the contact between the glacial deposits and the bedrock aquifer at a depth of 107 ft. The depth to water in this well was 16 ft in 1991 and 20 ft in 2003. This decline in water level is less than the decline observed in nearby bedrock monitoring wells.

To assess water levels in the glacial deposits, information from the WelLogic database was used. The WelLogic database includes water levels in wells reported as “static water level” by drillers following completion of a new well. Driscoll (1986) defines “static water level” as “... the level at which water stands in a well or unconfined aquifer when no water is being removed from the aquifer [through the well in question] either by pumping or free flow.” The use of “static” may be misleading, because the water level in wells is rarely static. As discussed later in this report, the “static water level” changed over time in the study area. The term ground-water level is used herein to refer to the reported static water level in the database.

Statistical procedures used to evaluate the ground-water levels reported in the WelLogic database are described in appendix C. These procedures were applied to ground-water levels in wells screened in glacial deposits in Monroe, Washtenaw, and Wayne Counties. Results of the statistical analysis did not indicate any systematic declines in ground-water levels in wells screened to the glacial deposits. However, few well logs for glacial-deposit wells are available for Monroe or Wayne Counties. Therefore, little data are available to support the conclusion that ground-water levels in the glacial deposits underlying these counties have not declined.

Recharge

Water infiltrating into the ground-water system from surficial sources is referred to as recharge. In the study area, some recharge to the water table in the glacial deposits comes from losing reaches of streams, but most comes from precipitation that infiltrates below the root zone. Holtschlag (1996) estimated recharge to the water table for the Lower Peninsula of Michigan; estimated recharge values for the study area range from 4 to 6 in/yr. The procedure used to estimate recharge used statistical analysis of base-flow measurements from streams. Because the glacial deposits have low hydraulic conductivities in much of the study area, this base flow is assumed to primarily reflect the discharge of water from the glacial deposits to the streams. Examining the base-flow characteristics of streams in the study area may reveal changes in the recharge to the system.

The USGS program RORA implements the Rorabaugh model to estimate ground-water recharge from stream hydrographs (Rutledge, 1998; Rutledge, 2000). The yearly

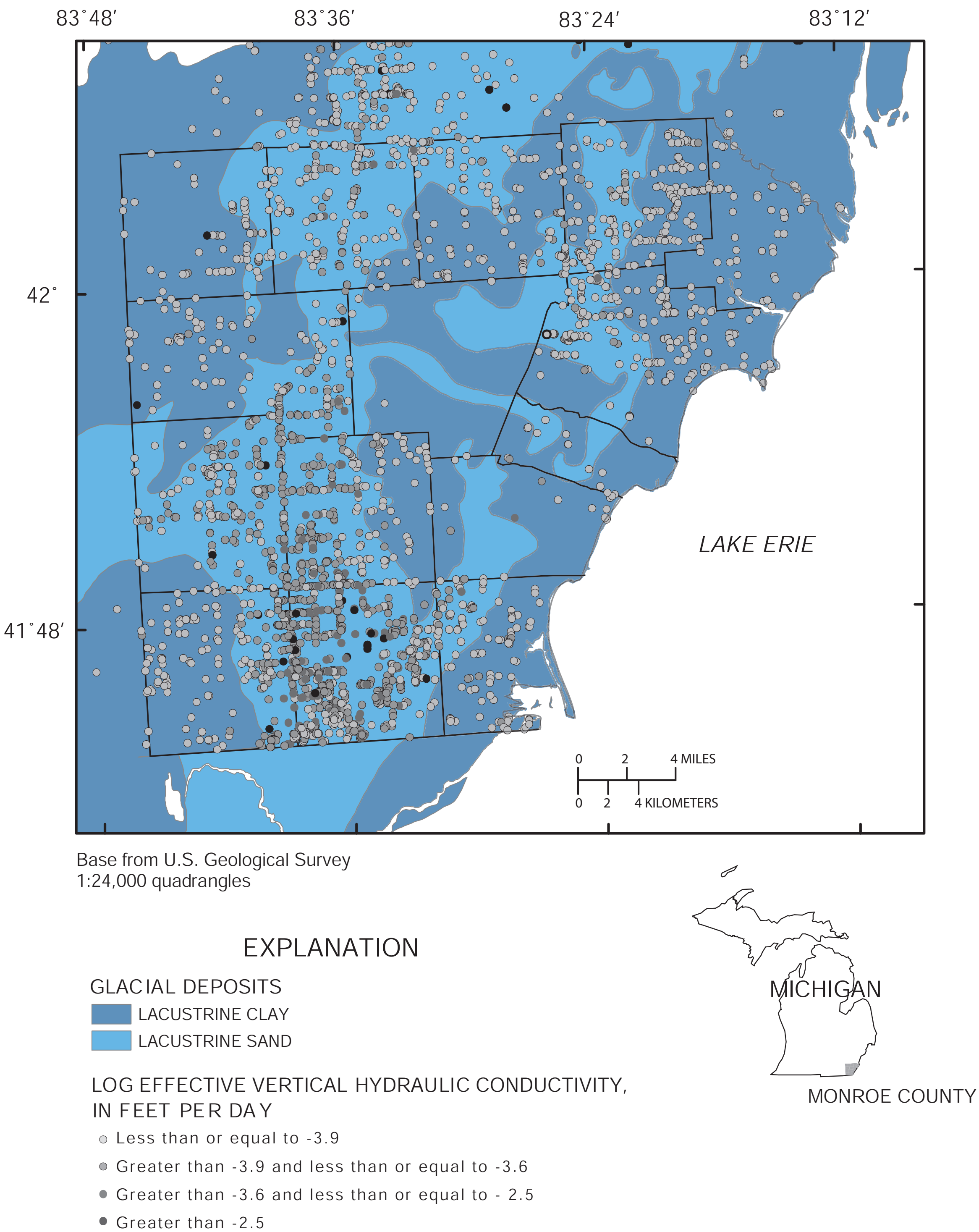


Figure 4. Estimated log(K_v^{eff}) for Monroe County, Michigan, based on thickness and lithology of units reported in the WelLogic database. Shaded areas are glacial deposits mapped by Ferrand and Bell (1982).

recharge in the River Raisin Basin and the Otter Creek Basin was examined using hydrographs from the USGS streamflow-gaging stations at the River Raisin near Monroe (U.S. Geological Survey Site 04176500) and Otter Creek near LaSalle (U.S. Geological Survey Site 04176605) (fig. 2). The estimated recharge to River Raisin and Otter Creek between 1990 and 2001 did not show a consistent downward trend. Base flow in Otter Creek was near its long-term average of 7.4 in/year for 8 years during the period. Base flow in River Raisin varied more than in Otter Creek. River Raisin base-flow values were below the basin long-term average of 7.5 in/year for 6 of the years during this period. The base flow was above average for the final 2 years of the period (2000-2001) and did not show a consistent downward trend. As another indication of streamflow, Nicholas and others (2002) showed that the 7-day low-flow values for the River Raisin station at Manchester, Michigan did not indicate a consistent downward trend.

Precipitation is the source of most of the recharge to the glacial deposits. The deviation from the annual average of 31.5 in/year for precipitation values from the Midwestern Regional Climate Center for the Monroe County Waterworks (station 205558) were used for this analysis (fig. 5). The annual differences and the difference for the period between October of one year and March of the next year are shown. Most recharge to the water table occurs during the period from October to March. From 1990 until 2002, the average precipitation from October to March tended to be below the long-term average. Average precipitation was above the long-term average during 1990-93, 1997, and 2001, and below the long-term average during 1994-96 and 1998-2000.

Discharge

Tile-drain systems, together with streams, lakes, and wetlands are areas of discharge from the glacial deposits. Tile-drain systems are common in current and former agricultural areas throughout the study area, because the water table is near the land surface. Historically, much of the study area had extensive wetlands including the Point Mouillee marshes and the Black Swamp (Robbins and others, 1994). Locally, domestic wells discharge small quantities of water from the glacial deposits. The base-flow estimates indicate that ground-water discharge to streams has not changed systematically over the past decade. None of the indicators of hydrologic changes in the glacial deposits — water levels, precipitation, or base flow — show systematic downward trends.

Bedrock Aquifer

The bedrock aquifer in the study area occurs in Silurian-Devonian carbonate rocks. In ascending order, and from oldest to youngest, the aquifer is composed of the Salina Group, the Bass Islands Dolomite, the Detroit River Group, which was subdivided into the Sylvania Sandstone and Detroit River Dolomite by Mozola (1970), and the Dundee Formation (fig. 6). The Traverse Group, a sequence of shale, sandstone, and dolomite, overlies the Dundee Formation and generally is not considered an aquifer. The characteristics of these units are given by Mozola (1970) and summarized by Nicholas and others (1996). Breen and Dumouchelle (1991) summarize the bedrock sequence in northwestern Ohio. They note that these formations grade from one lithology to the next and that contacts are difficult to identify. Intermediate weathering surfaces are not apparent in cores or in exposed faces such as quarries (D.B. Westjohn, U.S. Geological Survey, oral commun., 2003). Although the hydraulic properties of the individual bedrock units may be different, there is no evidence of extensive confining beds among the bedrock units composing the bedrock aquifer. Therefore, the entire bedrock sequence was considered to constitute a single heterogeneous bedrock aquifer.

The upper contact of the bedrock aquifer in the northwest portion of the study area consists of shale units, including the Antrim Shale and the extensive Coldwater Shale (fig. 6). These shales are important in that they have very low hydraulic conductivities that appear to prevent post-glacial recharge water from flowing to the Silurian-Devonian carbonates from above (Westjohn and Weaver, 1998; McIntosh and others, 2002; J.C. McIntosh, University of Michigan, oral commun., 2003). The Coldwater Shale was used as the lateral hydraulic boundary for the bedrock aquifers in the Michigan Basin RASA (Hoaglund and others, 2002). Glacial deposits, discussed previously, are the remainder of the upper boundary of the bedrock aquifer in this study. The lower boundary of the bedrock aquifer is assumed to be the Salina C-shale, which is part of the Salina Group (D.B. Westjohn, U.S. Geological Survey, written commun. 2002), and no evidence of upward flow from aquifers underlying the Salina Group has been presented in the literature.

The boundary to ground-water flow along the western border of the study area is saline water. Fresh ground water within the study area is bounded by brackish and saline water. The density difference between the fresh and saline water is assumed to limit migration of the freshwater. The Midwestern Basins and Arches RASA used the salinity boundary as the northern lateral boundary (Bugliosi, 1999). Monroe County is in a relatively limited area where the total dissolved solids concentration in the bedrock aquifer is low enough that the ground water is potable.

Lake Erie provides a hydraulic boundary to ground-water flow. Under natural conditions, the lake is a discharge area for ground-water-flow from the bedrock

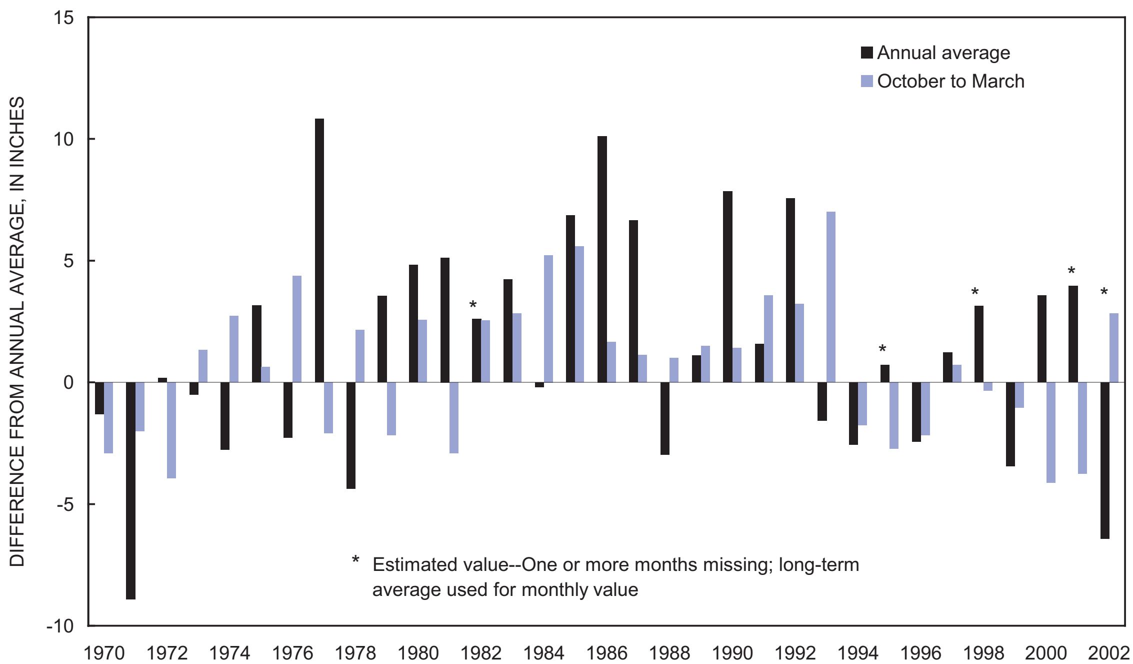


Figure 5. Graph showing yearly difference in annual precipitation from long-term average at the Monroe County Waterworks Monroe County, Michigan, 1970-2002.

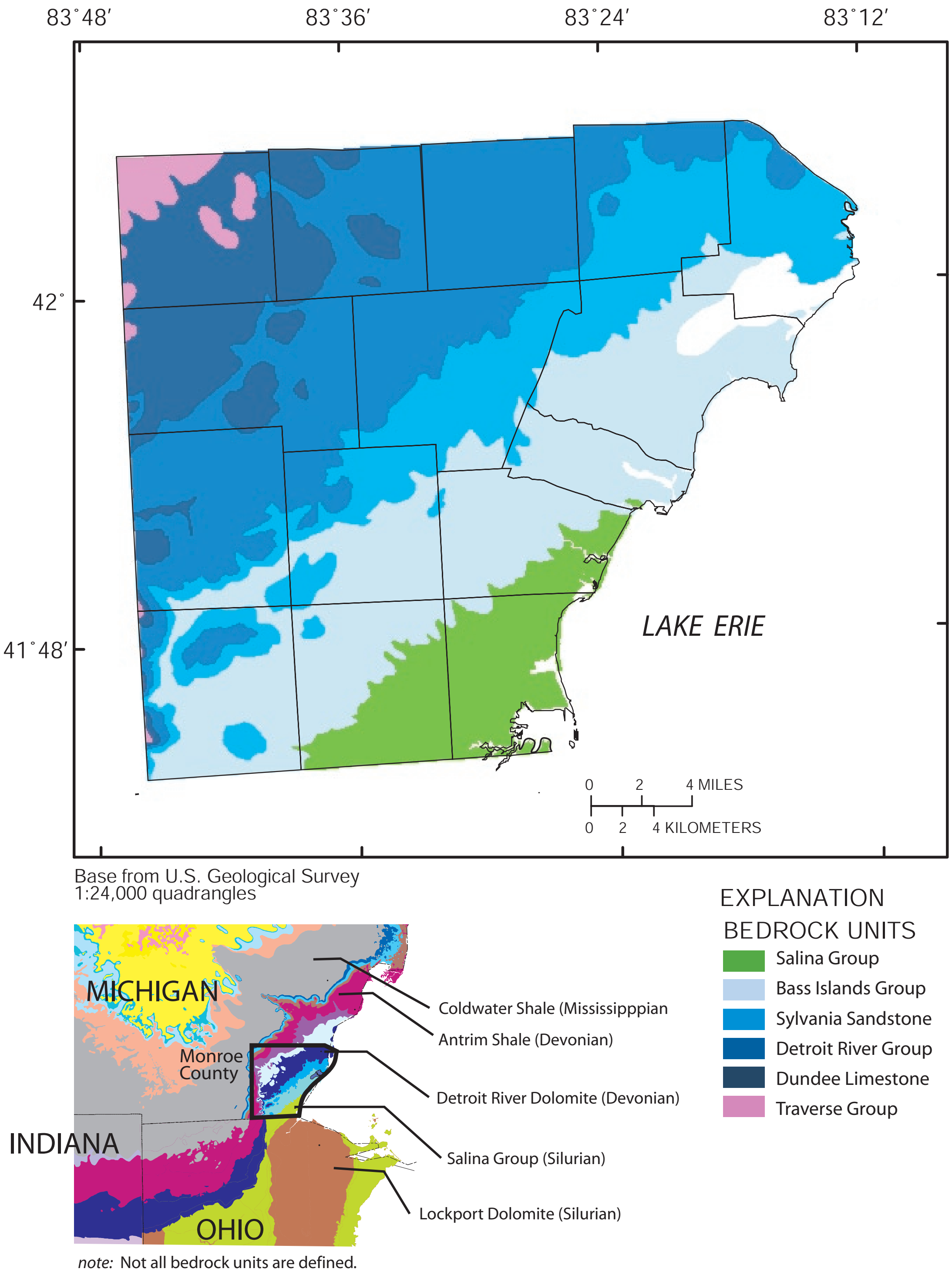


Figure 6. Bedrock units for Monroe County, Michigan and regional bedrock setting from Milstein (1987), Indiana Geological Survey (2002), Ohio Department of Natural Resources (2000a).

aquifer. Historically, water levels in the bedrock aquifer were above lake level along the Lake Erie shoreline in Monroe County. Bedrock water levels along most of the shoreline in the county ranged from slightly above lake level to tens of feet below lake level during the entire study period from 1991 through 2003. Water levels in the bedrock aquifer are expected to approach lake level at some distance eastward beneath Lake Erie.

Regionally, the bedrock aquifer is thought to be recharged predominantly in south-central Ohio and Indiana. From the recharge area, ground water moves northward until it meets saline water in the Michigan Basin and then discharges either to surface-water features in Indiana and northern Ohio or to Lake Erie (Bugliosi, 1999, fig. 15). Larger rivers, including the Maumee and Detroit Rivers, form internal boundaries to ground-water flow within the regional framework. Ground-water discharge in Lucas County, Ohio is towards the Maumee River and Lake Erie (Breen and Dumouchelle, 1991; R. Sheets, U.S. Geological Survey, written commun., 2002). There is no evidence for ground water in the bedrock aquifer crossing the Maumee River in Lucas County or the Detroit River. Based on this regional view, ground water in the bedrock aquifer is expected to move from west to east in Monroe County. A potentiometric surface for the bedrock aquifer for 1993 illustrates this movement (fig. 7).

Hydraulic properties

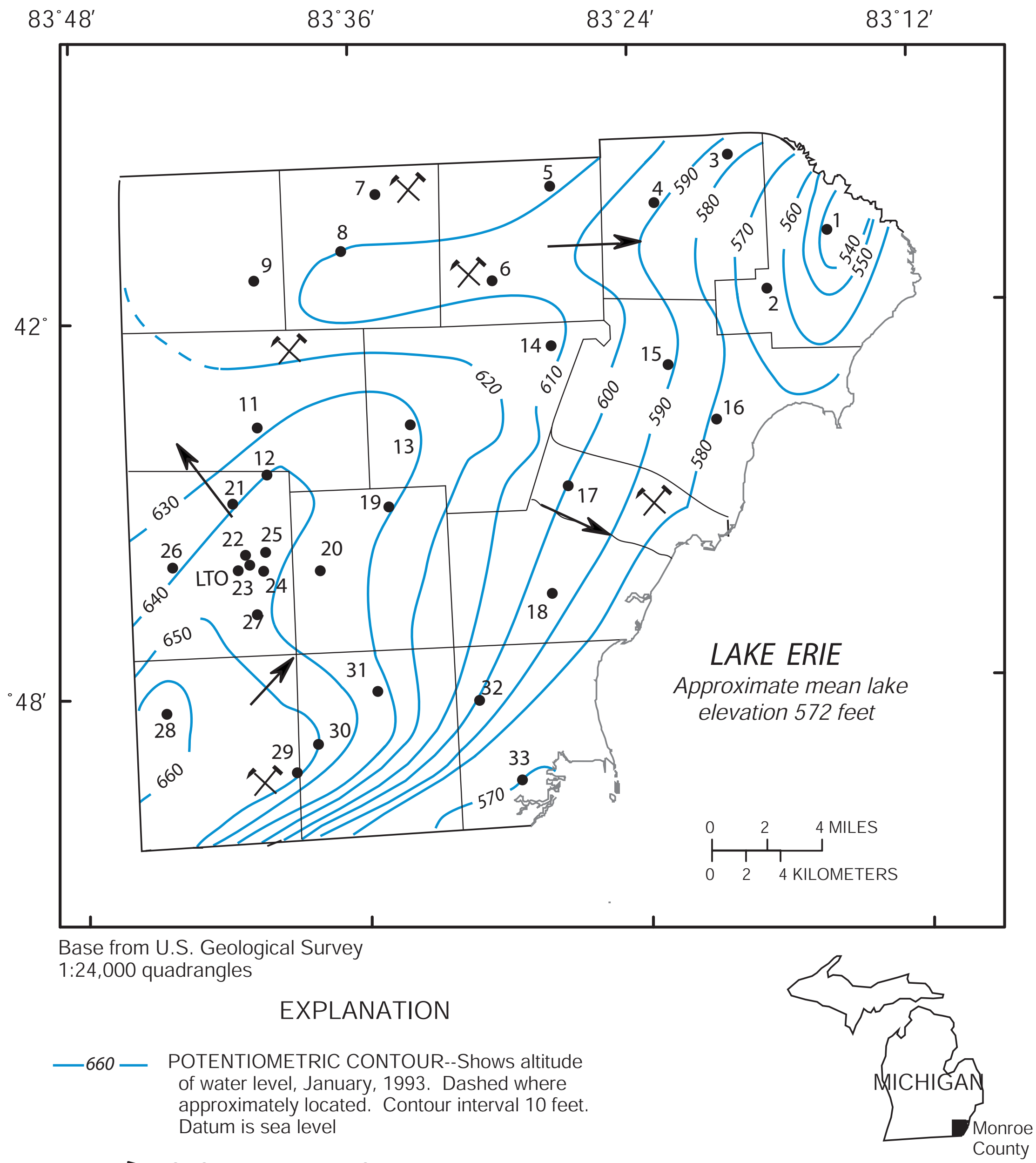
Ground-water flow in the bedrock aquifer is predominately through secondary openings consisting of fractures. Most of these fractures are along bedding-plane partings (Nicholas and others, 1996). Because of its location in the Michigan Basin, the bedding planes dip to the northwest in the study area. This dip coincides with a strike direction of southwest to northeast. Frequently, the fractures have been enlarged by dissolution of the bedrock. Some of the dissolution features are large enough to be considered karst (Sherzer, 1900) and may affect local flow patterns. This condition is especially true in southwestern Monroe County. Indications of the magnitude and distribution of hydraulic properties of the bedrock aquifer are given by aquifer tests, storage properties resulting from confined or unconfined conditions, and the barometric efficiency of monitoring wells.

Analysis of an aquifer test performed in the Petersburg game area using wells GLTO, G-22, G-24, and G-25, which are open to the Detroit River Dolomite, gave an estimated transmissivity of 1,000 - 1,300 ft^2/day and storativity of 0.00008 - 0.0001 (Nicholas and others, 1996). Test results also indicated a heterogeneous reponse at the monitoring wells. A more limited test performed near the London Aggregates quarry in London Township gave similar results (S.J. Wright, University of Michigan, written commun., 2003). Nicholas and others (1996) cite a test performed in Bedford Township that resulted in a slightly higher transmissivity than the Petersburg Game Area test (3,800 ft^2/day). These values were used as initial estimates in the calibration of the numerical model described in this report.

Slug-test results reported by Nicholas and others (1996) demonstrate the potential range and variability in hydraulic properties resulting from the fractured nature of the bedrock aquifer. The estimated transmissivity values determined from the slug-test analysis are questionable because the water level in the wells responded quickly to the applied slug and oscillations were recorded for many wells hampering the analysis. Importantly for this study, however, spatial patterns in transmissivity are not indicated in the reported results. For example, consider the reported transmissivities determined for monitoring wells G-29 and G-30 (fig. 2). These wells are 2 mi apart. Each well is completed in the Bass Islands Dolomite, and the top of both boreholes is approximately at the same elevation. Well G-30 has 119 ft of open borehole and well G-29 has 33 ft of open borehole. The reported transmissivities are 10 ft^2/day for G-30 and 3,400 ft^2/day for G-29. The longer borehole, G-30, evidently did not intersect as conductive of a fracture system as did the shorter borehole, G-29.

Unconfined and confined aquifers respond differently to pumping or other imposed stresses. For confined aquifers, the response time to changes in an imposed stress is governed by the compressibility of water and the compressibility of the aquifer material (aquifer storativity). For unconfined aquifers, ground water drains from the pore space or fracture (specific yield). The storativity of a confined aquifer typically is orders of magnitude smaller than the specific yield of an unconfined aquifer, and aquifers under confined conditions respond much faster than unconfined aquifers to the same disturbance (Theis, 1940). As discussed previously, the bedrock aquifer in the study area is overlain by confining units composed of either shale or glacial deposits. Water levels in the bedrock aquifer may be below the contact between the glacial deposit and bedrock aquifer. In this situation, the aquifer is under unconfined conditions. A sequence of maps of Monroe County illustrating the development of unconfined areas in the bedrock aquifer is shown in figure 8. These areas were determined by comparing the ground-water levels reported in the WelLogic database to the elevation of the top of the bedrock in the same borehole. There is scatter in the points because of local heterogeneity and errors inherent in the database, but the unconfined areas have expanded over time as the water level in the bedrock aquifer has declined.

The water-level record for the Petersburg deep



Base from U.S. Geological Survey
1:24,000 quadrangles

EXPLANATION

660 — POTENTIOMETRIC CONTOUR--Shows altitude of water level, January, 1993. Dashed where approximately located. Contour interval 10 feet. Datum is sea level

GROUND-WATER FLOW--Arrow indicates direction of ground-water flow if the aqufier was isotropic and homogeneous with respect to transmissivity

28 WELL AND IDENTIFIER--Prefix the letter G to number

QUARRY--Active in 1992

Figure 7. Potentiometric surface for bedrock aquifer, 1993, for Monroe County, Michigan, (from Nicholas and others 1996).

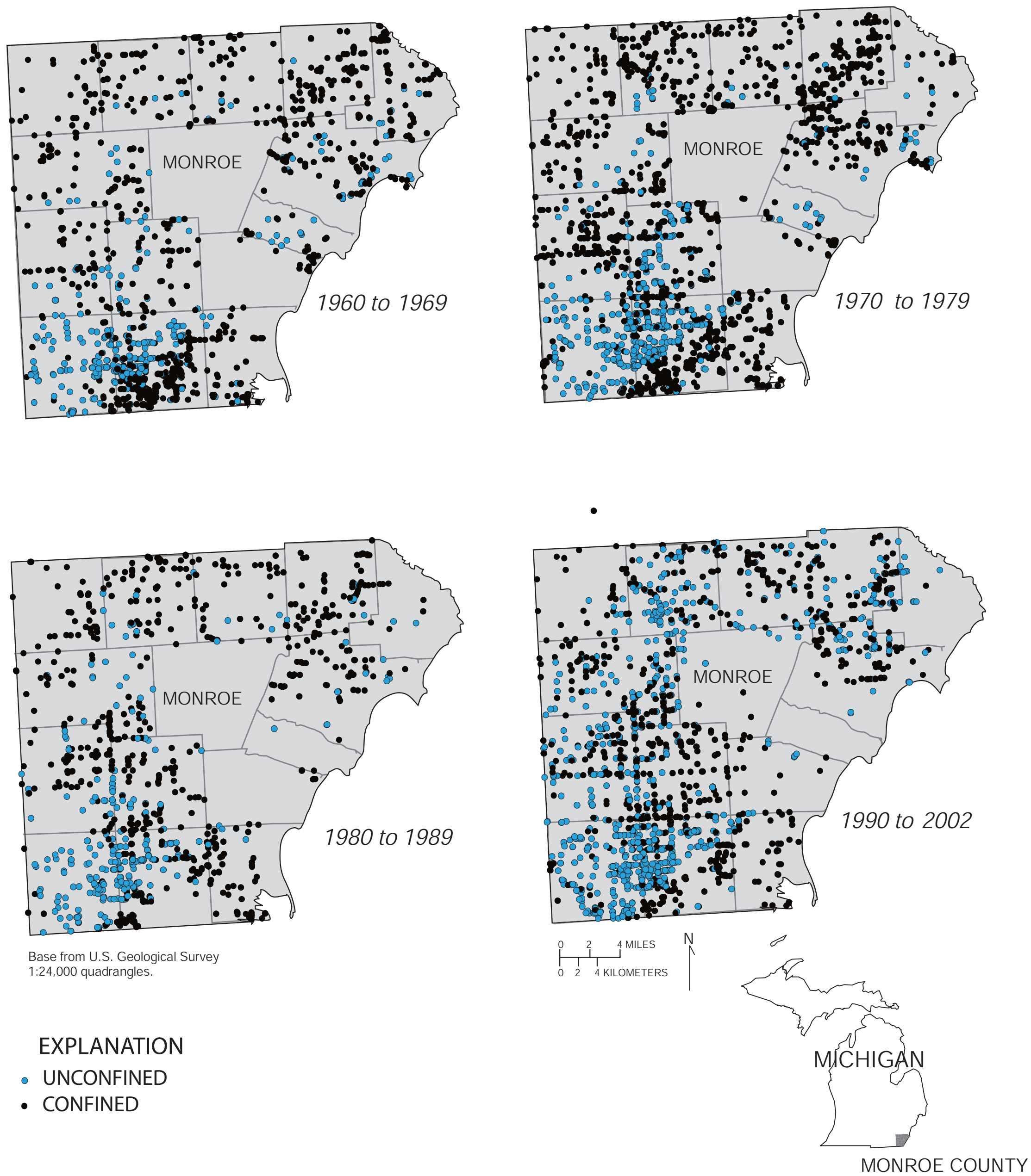


Figure 8. Time sequence of wells open to the bedrock aquifer showing confined (water level at least one foot above top of bedrock) or unconfined (water level at least one foot below top of bedrock) conditions, Monroe County, Michigan, 1960 to 2002 (data fromWelLogic database).

monitoring well (fig. 9), designated as GLTO (fig. 2), suggests that the aquifer in the area around this well has transitioned from confined to unconfined conditions. In the early part of the record, seasonal changes in the water level were on the order of 5 ft. The water level in the well declined about 12 ft from 1993 to 2001. Seasonal water-level changes since 1999 are less than 1 ft. Before 1999, aquifer yield resulted from the compressibility of the aquifer material and the compressibility of water (storativity). After 1999, aquifer yield results from dewatering fractures in the system (specific yield).

Barometric effects on water levels in wells in Monroe County were noticed by early settlers, and the response of water levels in monitoring wells to barometric changes may give insight concerning the hydraulic properties of the bedrock. Sherzer (1900) reported increased water levels and artesian flow described by residents. There were reports of faster flow or sediment in the water before storms, associated with a decrease in barometric pressure and a rise in water level. Precipitates in the wells may have been loosened by higher water velocities and introduced into the water in residents' wells. One well in Berlin Township that had been plugged by sand suddenly unclogged just before a storm. The well released water and sand "with such noise as to be heard some little distance" (Sherzer, 1900).

The values of barometric efficiency for 12 wells in Monroe County were calculated using Clark's method (Davis and Rasmussen, 1993) and seasonal variation and lithologic effects were evaluated. High and low values of barometric efficiency occurred in both areas mapped as having lacustrine sand and areas with lacustrine clay (fig. 3). Aquifers under unconfined conditions should have low or no barometric efficiency because air in the pore space should rapidly equilibrate with changes in atmospheric pressure. Barometric efficiencies, however, do not appear to correspond to confined and unconfined areas within the study area. This lack of correspondence indicates that the glacial deposits may have a low enough air diffusivity to allow wells open to both confined and unconfined areas of the bedrock aquifer to respond to barometric changes. Also, the bedrock surface topography may be locally variable, resulting in isolated unconfined areas that respond differently to barometric changes than does an extensive unconfined aquifer. The barometric efficiency correlated better with bedrock unit. Higher efficiency values were calculated for wells in the Detroit River Group and the Sylvania Sandstone, whereas lower values were calculated for monitoring wells in the Bass Islands Dolomite and Salina Group.

Water-level changes

Nicholas and others (2002) summarize the ground-water levels recorded from 1991 to fall 2001. The records for the monitoring wells through September 2003 are given in appendix A. Ground-water levels dropped in many of the monitored wells from 1991 to 2001. Water levels in 11 monitoring wells declined more than 10 ft, and the water level in 6 wells declined more than 20 ft. Measurements for the current study showed additional declines in water levels in most of the wells. Two wells, G-1 and G-2 (fig. 2), recently were dry indicating water levels below the bottom of the well. Well G-3, which is nearby and measured at 15-minute increments, exhibited a periodic pumping signature. Some of the wells (G-11, G-19, G-20, G-27 and G-28) had seasonal fluctuations on the order of 10 ft, with G-28 the most extreme. Some wells (G-4, G-5, G-6, G-7, G-8, G-10, G-17 and G-22 through G-26) showed little seasonal variation during the recent study period, which may be due to locally unconfined conditions in the bedrock aquifer. Ten wells (G-5, G-7, G-8, G-9, G-11, G-13, G-14, G-17, G-18, G-19) had water-level recoveries during 2002 and 2003. Some of these recoveries can be matched to seasonal variations. The general downward trend of water levels since 1991 can be seen in all the wells except for G-10 and G-18. The monitoring well G-10 is the only monitoring well screened to the glacial deposits. All of the other wells are open to the bedrock aquifer.

Ground-water levels in bedrock monitoring wells from four landfills were obtained from MDEQ and also are given in appendix A. Near the Monroe County line to the north, ground-water levels at Carleton Farms decreased approximately 15 ft between early 1995 and late 2001, and levels at Matlin Road Landfill decreased about 5 ft between 1996 and 2002. This decline in ground-water level is consistent with USGS well G-4, which declined about 10 ft over the same time period. The ground-water level at Jefferson Smurfit, near the city of Monroe, did not change from early 1999 to early 2002. This water-level record is consistent with the record from G-16, which is located northeast of the landfill nearly parallel with the Lake Erie shoreline. Rockwood Landfill is located within a mile of the Thompson and Rockwood Quarries, and ground-water levels are clearly affected by dewatering at these quarries (fig. A37). Landfill records from 1980 precede quarry operations and water levels were similar across the landfill. More recent measurements show a decrease in water levels on the south side of the landfill, from 60 to 80 ft below the levels on the north side.

The statistical significance of trends in water levels from 1991 through 2001 for the USGS monitoring well data was determined by using the methods described by

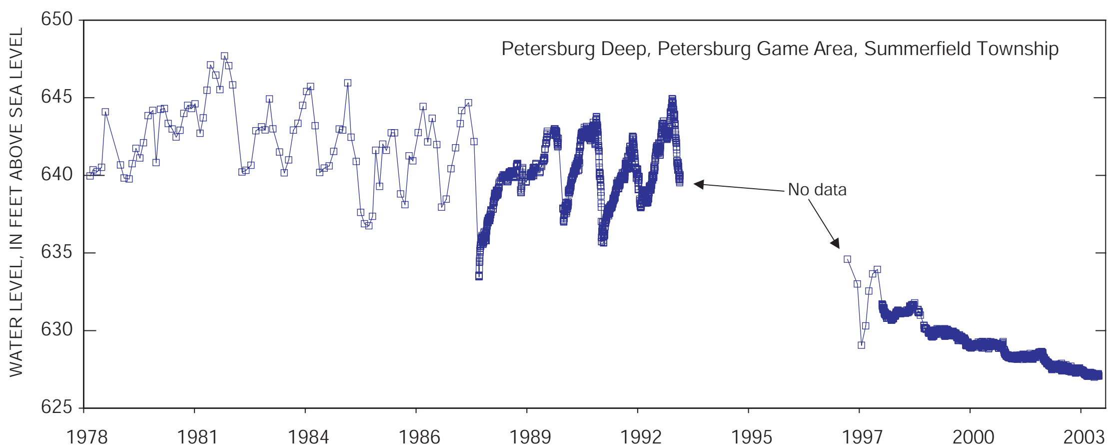


Figure 9. Hydrograph for U.S. Geological Survey observation well GLTO which is open to the bedrock aquifer, Monroe County, Michigan, 1978-2003. The change in amplitude of yearly response indicates a change from locally confined to locally unconfined conditions.

Kendall (1955) and by the Sen slope (Gilbert, 1987). Kendall's test was developed to determine if two series were correlated or not based on the ranking of the series. The test is summarized by the value referred to as the Kendall τ. The Kendall τ ranges from 1 for a strictly increasing series to -1 for a strictly decreasing series. A τ value of zero indicates no trend in the data. The test does not depend on the absolute value of the observations, and it is not affected by missing observations or changes in the frequency of observation of the time series. The analysis was performed using the statistical program S-Plus 2000 (MathSoft, 2000) and the S-Plus USGS library function "kensen" on both water-level time series and streamflow hydrographs. Details of the analysis are presented in appendix B, and a summary of the analysis follows.

The largest magnitude for the Sen slope for a water-level record was -8.7 ft/year at well G-17. Note that this well is less than 500 ft from a quarry that was deepened and increased ground-water withdrawals during the study period. At this distance, the water level is directly under the influence of the quarry depth, and this large decline is not indicative of declines throughout the county. Most wells had declines that can be grouped into three clusters. Wells in the northeastern part of the county (G-1 through 4), wells (G-12 through 15, G-19, G-20, G-29, G-31), and wells near Petersburg (G-21 through 28) had slopes between -1 and -2 ft/year. Wells in the northern part of the county (G-5 through 9, G-11) had slopes between -2 and -4.5 ft/year. The remaining wells (G-10, G-16, G-18 through 20, G-30, G-32, G33) showed little decline. Well G-10 does not follow the pattern in the northeastern portion of the county, but this is the only monitoring well completed in the glacial deposits overlying the bedrock aquifer. Well G-13 shows the only positive trend over the time period. The *p*-value for this well (0.16), however, indicates little statistical likelihood of a trend at this well. The only other well with a high *p*-value (0.13) is G-18. The remainder of the wells have *p*-values that are extremely small, and the statistical likelihood of no trend in these wells is less than 1 percent.

The water-well logs in the WelLogic database also were examined to determine if water-level trends in the bedrock aquifer are present. Despite the uneven coverage across the county, examination of the reported ground-water levels from the digitized logs may indicate trends and patterns of ground-water levels in the area. The ground-water levels from the water-well logs are subject to errors, including errors in determination of land-surface elevation at the well from topographic maps, water-level measurement error, and incomplete recovery of the water level in the well after disturbance during well drilling and development. These errors are assumed to be scattered around the true water level randomly, and, therefore, they are not assumed to impose a systematic bias to the analysis. An estimate of the mean ground-water level can be determined if a large number of well logs is included in the analysis. The water-well logs in the WelLogic data base were examined, and 7,229 logs for the bedrock aquifer had the date of drilling and other information to determine the ground-water level.

The analysis of WelLogic records described in appendix C was performed for the bedrock wells in Monroe County. This analysis grouped wells by township and date to discern changes in the ground-water level. The box-and-whisker diagrams were developed by grouping wells into 2-year groups by the drilling date. The mean, median, lower and upper quartiles, and outliers were determined for each group that had more than 10 wells. Mean values for all groups are connected by a dotted line on the plots. Groups with less than 10 wells drilled during a 2-year period are plotted using open circles for individual wells and solid circles at the means. Examination of the figures in appendix C reveals appreciable scatter for the depth-to-water for each 2-year group in each township. The boxes enclose data values between the lower quartile and upper quartile. These boxes tend to enclose a range up to 30 ft. The whiskers show the spread of the data by extending to either the maximum data value or the upper quartile + 1.5*(upper quartile-lower quartile) and to the minimum data value or the lower quartile - 1.5*(upper quartile and lower quartile). In the latter cases, the whiskers help highlight outliers in the data set. For most townships, the variation in the mean is less than the scatter of the groups. Despite this scatter, general trends are evident in the plots. Little change in the mean ground-water level is indicated for six townships. Townships where the depth to water increased (mean ground-water level declined) include Ash, Berlin, Dundee, Exeter, Frenchtown, London, Milan, Monroe, and Summerfield. For example, the mean ground-water levels decline in Exeter Township was approximately 25 ft (fig. 10). Estimated mean ground-water-level declines using the WelLogic database and the declines observed in the USGS monitoring wells generally agree (fig. 11). The townships that show the largest declines are consistent with the USGS monitoring wells. In the southern townships, the USGS wells have shown either no decline or smaller declines and the WelLogic interpretation agrees.

The spatial distribution of ground-water levels reported in the WelLogic database was examined further. Four-year groups of wells were plotted (figs. 12-13). A smooth surface was fit to the reported ground-water levels using the S-Plus 2000 loess function (Mathsoft, 2000). The loess surface fits a local trend surface to the data by using a weighted least-squares technique with greater weight given to data near the interpolation point (Venables and

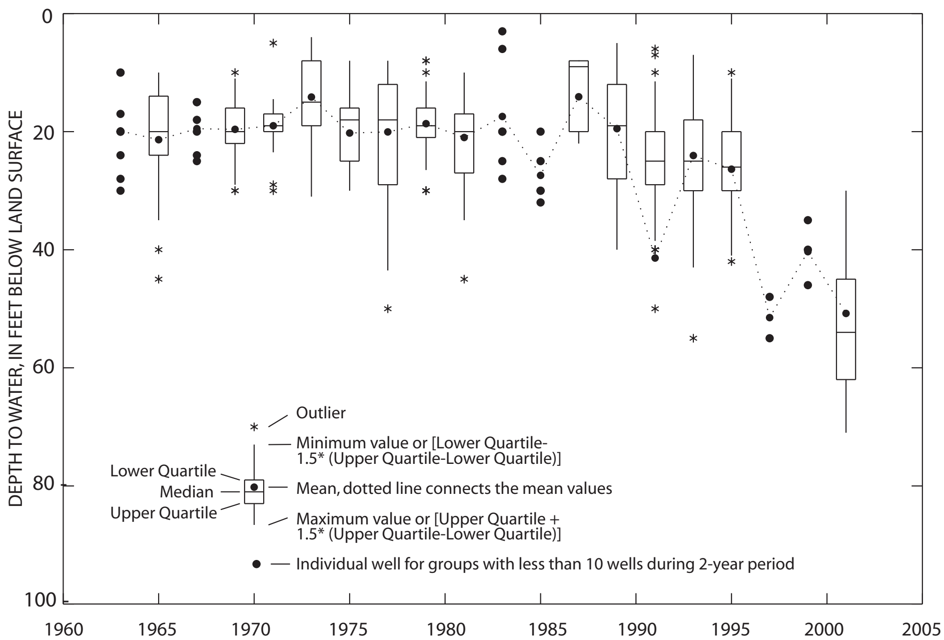


Figure 10. Trend in reported depth to ground water for Exeter Township, Monroe County, Michigan, data from the WelLogic database. Note the general decline in water level from 1990 to 2002.

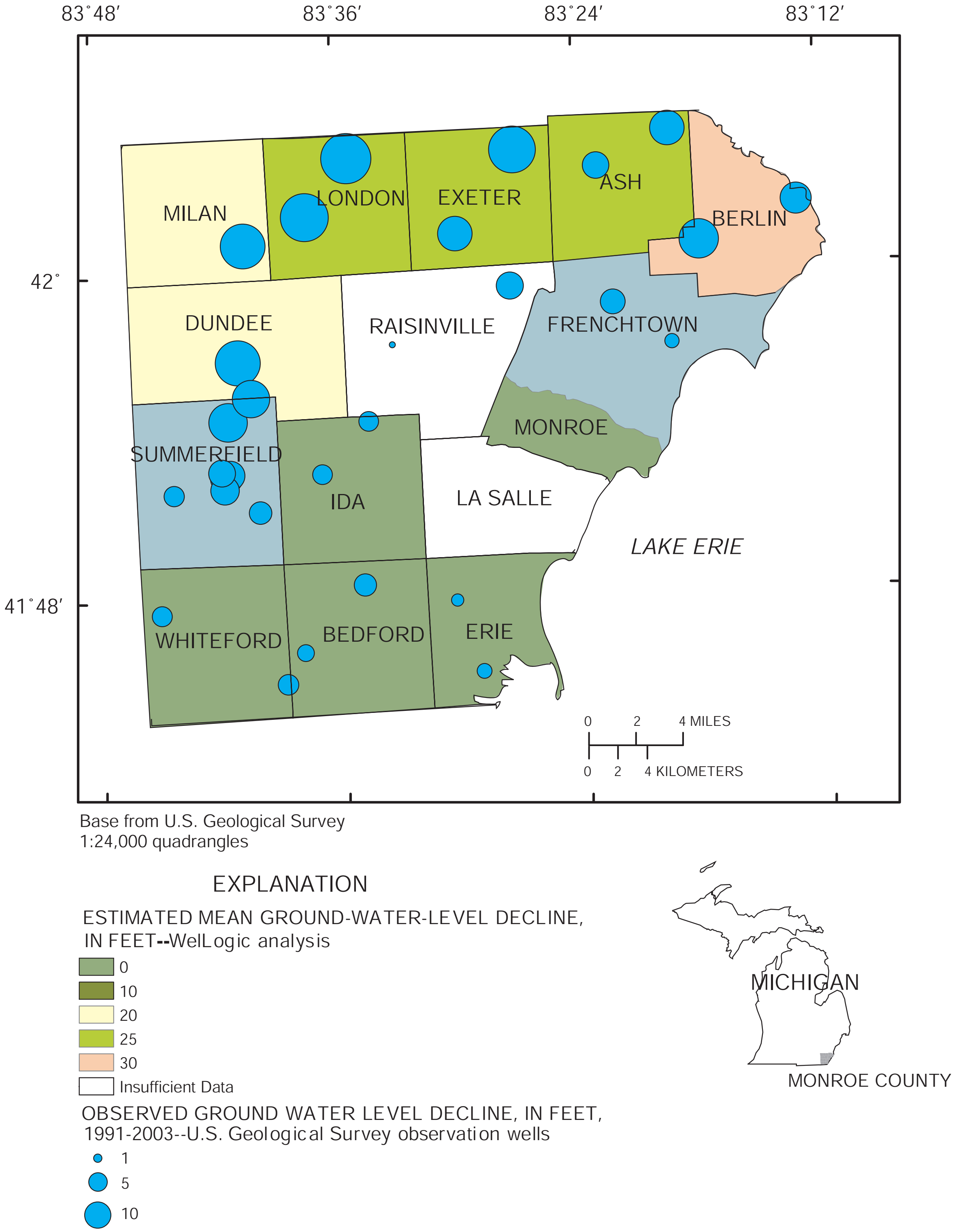


Figure 11. Estimated and observed ground-water-level declines, Monroe County, Michigan, 1991-2001.

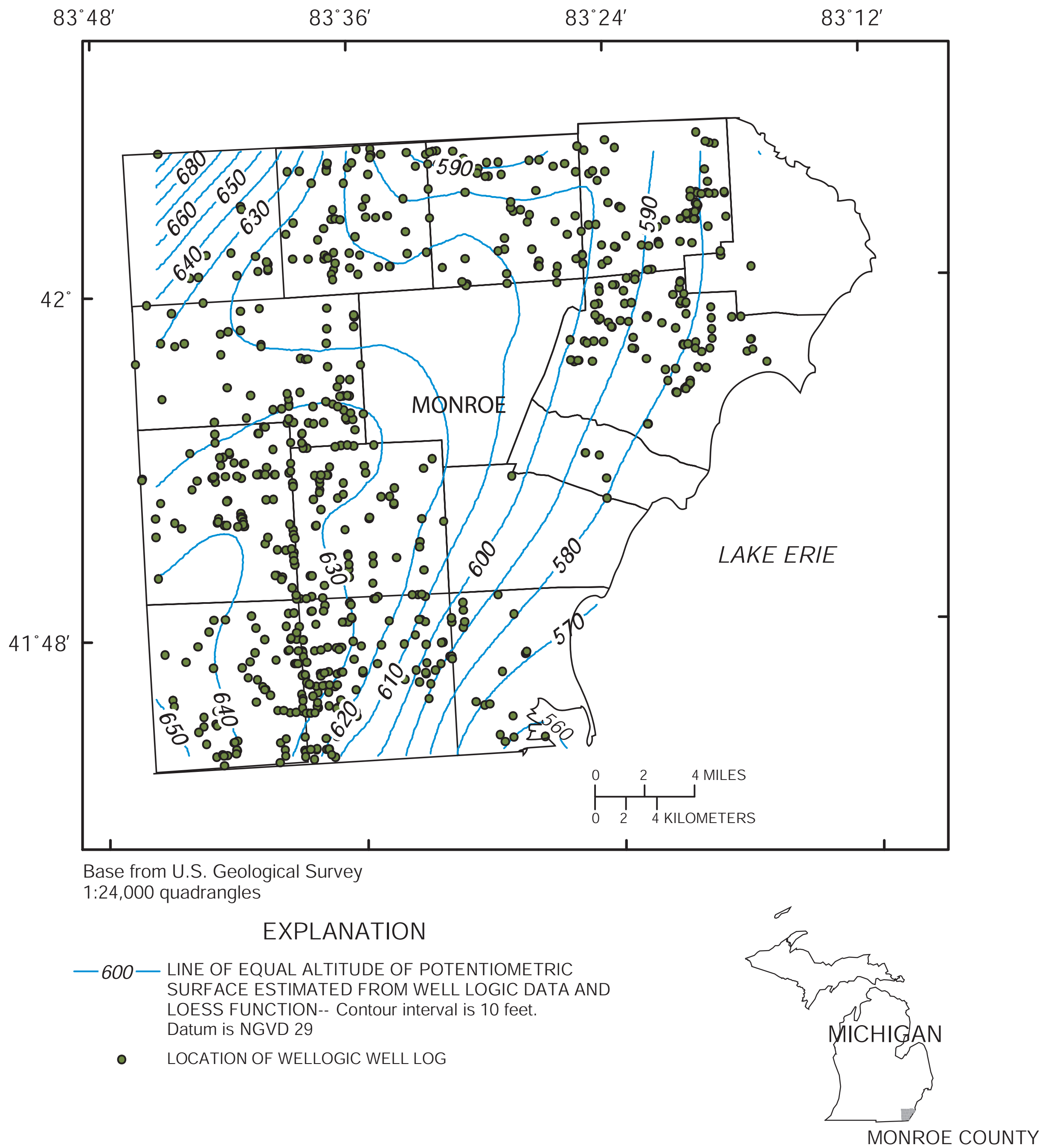


Figure 12. Potentiometric surface for bedrock aquifer, 1988-1991, developed from WelLogic ground-water-level records and loess function, Monroe County, Michigan.

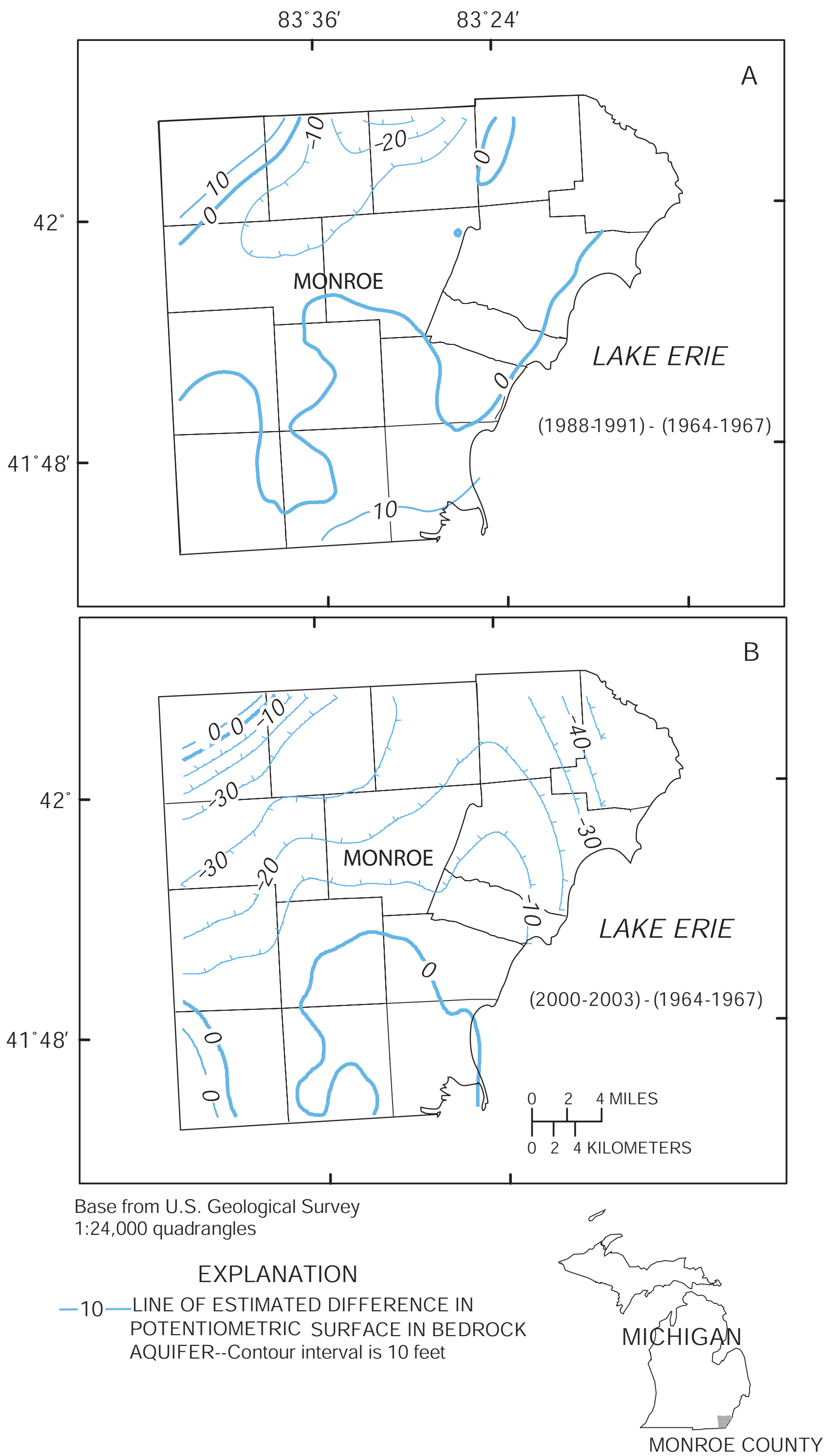


Figure 13. Estimated difference in potentiometric surface in bedrock aquifer in Monroe County, Michigan based on WelLogic ground-water levels and loess function. ((a) (1988-1991) - (1964-1967) surfaces and (b) (2000-2003) - (1964-1967) surfaces.)

Ripley, 1994). The distribution of water levels was fit using the well location. With these two independent variables, the loess surface essentially is a weighted-quadratic fit to the data. The fitted surface and observation points for the 1987-90 grouping is shown in figure 12. Note the general agreement between the loess surface and the interpreted potentiometric surface from Nicholas and others (1996) shown in figure 7.

The difference in the fitted surfaces between the 1988-91 grouping and the 1964-67 grouping is shown in figure 13(a). Little change in the mean ground-water level is shown. Zero changes are contoured over large parts of the county. Thus, the estimated change between these two time groupings is less than 10 ft. In contrast, large changes in ground-water levels are calculated using the 2000-03 grouping and the 1964-67 grouping (fig. 13(b)). In the northern portion of the county, 20-ft declines in water levels are contoured. The observed declines in the USGS monitoring wells also are plotted (fig. 13(b)). The contours indicate that the pattern of water-level changes produced with the WelLogic records are consistent with observed declines in water levels in the USGS monitoring wells. Exceptions to these results include G-17, which is close to an active quarry and not indicative of the regional decline, and G-1 which is now dry and, thus, not yielding a representative value for the local decline. The surface seems to overestimate the decline in the extreme northeast portion of the county and may overestimate the decline near G-16 where there are not many wells in the WelLogic database to control how the surface is contoured.

The ground-water-level patterns described above appear to relate to the expected pattern of vertical hydraulic conductivity of the glacial deposits. Superimposing the effective vertical hydraulic conductivity map and the apparent change in ground-water level shows that the area identified with the highest potential to allow vertical flow is in the region where ground-water levels have not declined substantially since 1991 (fig. 14). Regions with large change in ground-water level tend to have low potential for vertical flow based on the WelLogic lithologies. Note that the relation between the effective vertical hydraulic conductivity and the area of ground-water-level decline may be coincidental. The pattern of pumping in the region also is important, and the region with lower declines may simply correspond to a region where ground-water use has not increased as much as it has in the northern townships of the county.

USGS monitoring wells and the mean surface computed from the WelLogic records are consistent. Analysis of the time-series data from each set of wells indicates water-level declines up to 20 ft in the northern portion of Monroe County. Analysis of the WelLogic trends indicate fairly constant ground-water levels until the mid 1990s, when declines are evident in townships in northern Monroe County. The ground-water levels for monitoring wells in the bedrock aquifer reported from four landfills in the northeastern portion of the county also are consistent with levels observed in the USGS monitoring wells.

Leakage

Not all of the water that recharges the water table moves through the glacial deposits to the bedrock aquifer. The amount of water that reaches the bedrock aquifer depends on both the vertical hydraulic conductivity of the glacial deposits and the difference between the water level in the glacial deposits and the water level in the bedrock aquifer. Ground water only will flow from the glacial deposits to the bedrock aquifer if the water level in the glacial deposit is higher than that in the bedrock aquifer. For the bedrock, the term leakage is used to describe the supply of water flowing to the bedrock aquifer instead of recharge, because water from the surface must move through the overlying glacial deposits. Leakage also can be out of the bedrock aquifer to the overlying glacial deposits. Leakage to the bedrock aquifer has not been measured directly in the study area, but inferences regarding leakage can be deduced from water-level patterns and water quality.

Leakage to the bedrock aquifer is highest in areas where the water level in the glacial deposits is higher than that in the bedrock aquifer and in areas where the glacial deposits have higher vertical hydraulic conductivity than other areas. The moraine west of Monroe County appears to be a potential recharge area, however, much of the moraine lies on top of confining units composed of the Coldwater and Antrim Shales. Therefore, leakage to the bedrock aquifer in the study area is assumed to be through glacial deposits east of the Antrim Shale subcrop (fig. 6).

Ground-water chemistry can be used to identify leakage areas. Evaluation of the water-quality data summarized by Nicholas and others (1996) indicates that gypsum controls the major ion chemistry of the bedrock ground water in Monroe County. Geochemically, anhydrite ($CaSO_4$) acts like gypsum ($CaSO_4$*H_20), and these minerals only are absent in the formations below the Dundee Formation, indicating that gypsum saturation of the ground water occurs within these bedrock units. Spatial analysis of the data indicates higher saturation with respect to gypsum in water samples from wells under or near the lacustrine clay, and lower saturation with respect to gypsum in water samples from wells under or near the lacustrine sand areas (fig. 4). The pattern of lower saturation indicates that the sand units allow more leakage into the bedrock.

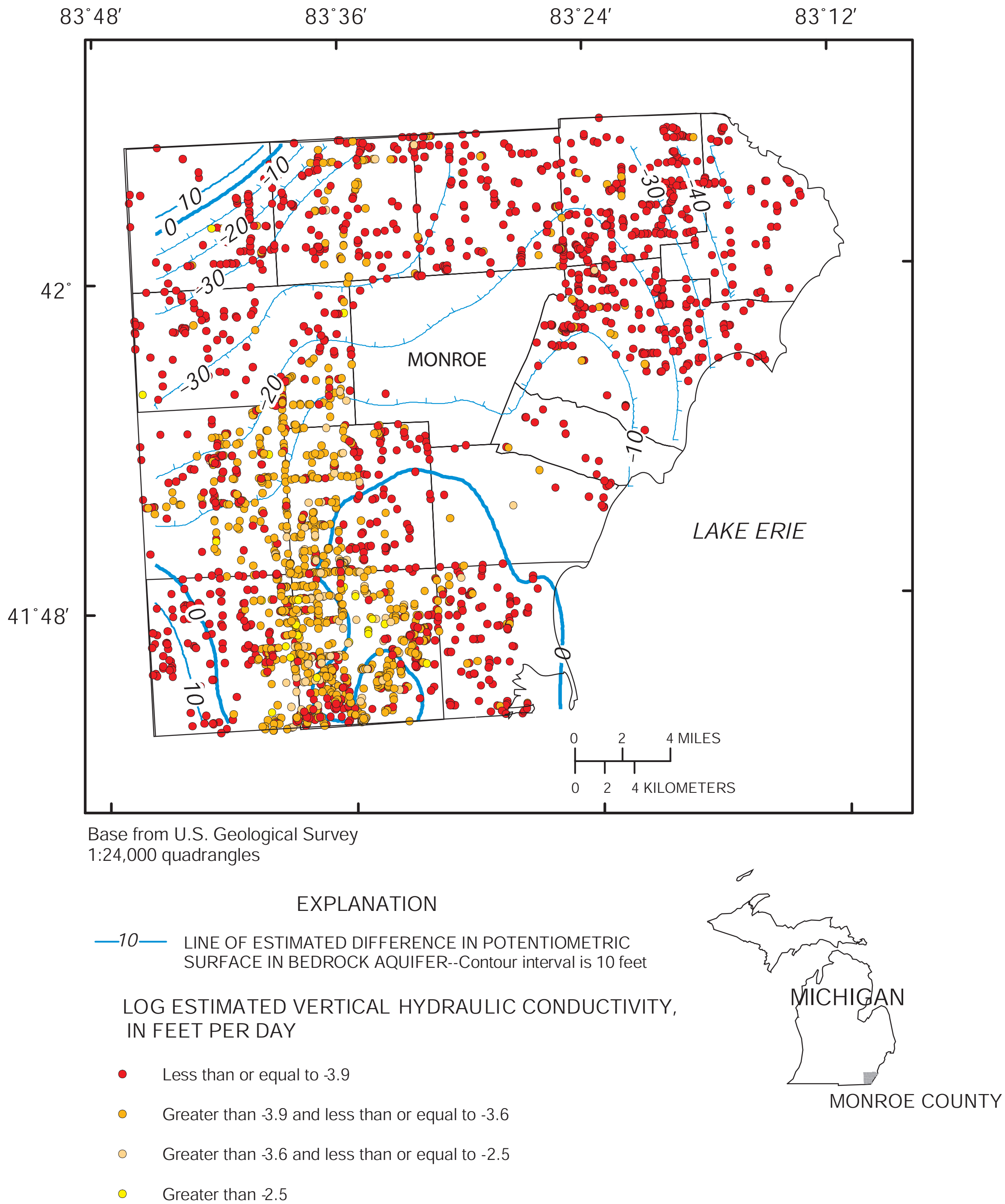


Figure 14. Estimated effective vertical hydraulic conductivity and estimated change in potentiometric surface from WelLogic data base and loess function (2000-2003) - (1967-1964) Monroe County, Michigan. Area of small change corresponds to area where the effective vertical hydraulic conductivity is greatest.

Another indicator of recharge is tritium content. Tritium concentrations above background in ground water indicates post-1953 recharge (Fetter, 2001). There is no clear relation between the geologic bedrock unit or the overlying glacial geology and the tritium values reported by Nicholas and others (1996). Water from 13 of 26 bedrock wells sampled for tritium indicated pre-bomb levels. The remaining 13 wells had elevated levels indicating some post-1953 recharge.

The water-chemistry results indicate only a small amount of leakage to the bedrock aquifer through the overlying glacial deposits. If extensive leakage were occurring, the ground water in the bedrock aquifer would be expected to have elevated tritium concentrations. The water also would be expected to be less saturated with respect to gypsum.

Discharge

Discharge from the bedrock aquifer is either natural or induced by human activities. Natural discharge occurs to Lake Erie, at springs, and in major streams. In areas where the potentiometric surface in the bedrock is higher than the water table in the glacial deposits, ground water will flow from the bedrock to the glacial deposits. Discharge caused by human activities is discussed in the "Water Use" section later in this report.

Historically, flowing wells were present in much of eastern Monroe County. In these areas, ground water would have flowed naturally from the bedrock to the glacial deposits. Earlier reports describing these wells were summarized by Allen (1977). Sherzer (1900) includes a photograph from 1899 of a flowing well near Otter Creek, 2.5 mi from LaSalle. The owner placed an 8-in pipe on the well, and it was reported to yield "... a very rapid stream two feet broad and four inches deep flowing from the well to Otter Creek" (Sherzer, 1900). The water level in flowing wells were reported to be from slightly above land surface to 20 ft above land surface near Lake Erie. Sherzer (1900) also reported that wells would interfere with each other and that, "wells which formerly flowed in the southern part of Erie township, back three miles from the lake, have now ceased although the water rises near the surface." He also noted that a deep well in the city of Monroe caused so much interference with other wells that it had to be plugged. The change from confined to unconfined conditions over portions of the county and the current absence of flowing wells indicates that natural discharge to the glacial deposits from the bedrock aquifer has decreased since 1900, and the decrease continues through 2001 as more of the bedrock aquifer becomes unconfined.

Lake Erie levels were examined to determine if there was a relation between the declining ground-water levels in the bedrock aquifer and the observed lake level. Lake Erie levels fluctuated from 1991 through 2002. A hydrograph of Lake Erie levels from the CO-OPS National Water Level Observation Network (NWLON) database at the Fermi Power Station (Station number 9063090) (National Oceanic and Atmospheric Administration, 2003) is shown in figure 15. The long-term mean Lake Erie levels for February and June also are shown (U.S. Army Corps of Engineers, 2003). Lake level was from 1 to 2 ft above the long-term average in 1997-98, and from 1 to 2 ft below the long-term average in 2001-02. The most important feature of this dataset for this study is that although the level of Lake Erie varied over the course of the study, the levels did not show a consistent downward trend. This lack of consistent trend indicates that the discharge to Lake Erie did not consistently increase for the time period leading to regional ground-water declines.

Great Sulphur Spring, in Erie Marsh near the Lake Erie shoreline in the southern part of Monroe County, is a discharge point for ground-water flow from the underlying bedrock aquifer. The spring was sampled in September 2003. The specific conductance was 2,590 microsiemens per centimeter at 25 degrees Celsius (μS/cm), and the flow rate from the spring during this period was approximately 5.2 ft^3/s. Any historical variation in its discharge would have been controlled by the water-level difference between the aquifer and the surrounding surface water that coincides with Lake Erie levels. Ground-water levels at G-33, which is nearby (see fig. 2), have been steady and Lake Erie levels have not dropped consistently. Therefore, ground-water discharge at Great Sulfur Spring should not have changed appreciably during the study period. Smaller springs may be present in the area, and their discharge would be controlled by this water-level difference.

WATER USE

Various data sources were used to estimate water use in the region. Agricultural irrigation and domestic self-supplied estimates were obtained through the USGS Water-Use Program that is performed in cooperation with MDEQ (U.S. Geological Survey, 2003; C.L. Luukkonen, U.S. Geological Survey, written commun., 2003), the U.S. Department of Agriculture (1999), and the U.S. Department of Commerce (1987). Industrial-use data were obtained through MDEQ following the MichiganWater Use Reporting Program that is mandated under the authority of Part 327, Great Lakes Preservation, Natural Resources and Environmental Protection Act, 1994 PA 451, as amended. Water users who have the capacity to withdraw over an average of 100,000 gallons (378,500 liters) per day over any 30-day period are required to register with the MDEQ and to report their water-use information on an annual basis. Quarry discharge was estimated from NPDES permits and associated reported discharges through MDEQ

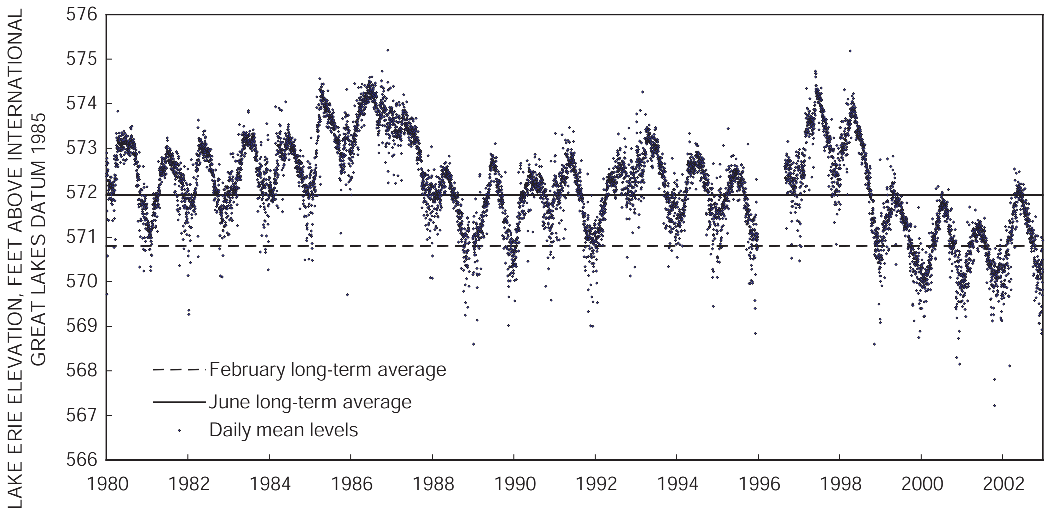


Figure 15. Hydrograph of Lake Erie from the CO-OPS National Water Level Observation Network (NWLON) database at Fermi Power Station (station number 9063090), Monroe County, Michigan. Lake Erie level was both above and below long-term average between 1991-2001 when ground-water levels were observed to decline.

and checked by field measurements. Comparing ground-water-use estimates from 1991 to 2000 shows an overall increase. The ground-water withdrawal for all recorded users for Monroe County in 1991 was approximately 20 Mgal/d, and was nearly 30 Mgal/d in 2001. The percentage of the total for public supply, self supplied, industrial, irrigation, and quarry dewatering remained about the same for each user.

Public, Self-Supplied, and Industrial Water Use

Individual domestic wells are important in Monroe County. The self-supplied water use estimates were made using census figures to estimate the number of people in the county who are not on public-water supply (table 1). The estimated number of self-supplied users was multiplied by 86 gallons per day, which is the average daily use per person (U.S. Geological Survey, 2003).

Most public supplies for communities in Monroe County rely on surface water. Historically, Carleton used approximately 0.1 Mgal/d of ground water. Petersburg and Flat Rock Village currently use approximately this amount (table 2). Use information for self-supplied industries and quarries in the study area from Ohio is summarized in table 3 (J. Remic, Ohio Department of Natural Resources, written communication, 2002).

Agricultural Irrigation

Agricultural irrigation is an important water use in the study area. Estimated irrigated farmland for Monroe County, Michigan, in 1982 was 1,952 acres; in 1987 it was 2,181 acres; in 1992 it was 2,385 acres; and in 1997 it was 5,047 acres (U.S. Department of Commerce, 1987; U.S. Department of Agriculture, 1999). Much of the irrigation, however, is attributed to surface-water sources (Nicholas and others, 1996; 2002). Ground-water irrigation estimates are given in table 4 (R. Van Til, Michigan Department of Environmental Quality, written commun., 2003). The values are given as an average daily withdrawal for each year. Because the irrigation season is much shorter than a calendar year, the actual pumping rates are higher than the values given in table 4 during the irrigation season and zero during the rest of the year.

Land-use maps were used to estimate the distribution of this irrigation across Monroe County. The distribution of row crops is widespread. Not all of the row crops farmed in the county are irrigated, but this distribution was used to simulate the general irrigation demand. The irrigation use from the bedrock aquifer estimated by MDEQ for Monroe County was divided by the acreage farmed to estimate an average demand that was applied to the entire agricultural area.

Golf-Course Irrigation

Golf-course irrigation is a visible water use in the study area. Some golf courses have registered with MDEQ as having the capacity to withdraw 100,000 gal/d for a 30-day period. Other courses in the area provided ground-water use estimates to USGS. The remaining courses are assumed to use limited amounts of ground water. To estimate an upper bound on golf course ground-water use, the remaining 12 courses were estimated to use between the median value of the other courses (0.012 Mgal/d) and the average value for the other courses (0.03 Mgal/d). With these values, the range for the maximum golf course ground-water use is 0.62 - 0.84 Mgal/d, and the annualized average reported ground-water use for golf courses is 0.48 Mgal/d.

Quarry Dewatering

Quarries have operated in the study area for decades. Aggregate (dolomite and limestone) quarries typically dewater the rock to provide access for the quarrying equipment and workers. The water pumped from the bedrock aquifer is either used in other processes at or near the quarry site, or it is discharged to surface water.

From 1985 to 2000, the estimated quarry dewatering has represented approximately 75 percent of the ground-water use in the study area. The amount of water pumped from the bedrock aquifer by quarries is estimated from reported discharges regulated by facility NPDES permits. Historically, some quarry operators accurately measured discharge, whereas others reported a discharge value that appears to be related to the permitted limit or pump capacity. Examination of the most recent records indicates that most quarries now measure discharge. The annualized average total reported discharge for quarries in Monroe County and nearby quarries in Wayne County in 1991 was approximately 17 Mgal/d. The reported discharge increased to approximately 23 Mgal/d in 2001 and ranged between 20 and 27 Mgal/d during 2001-02. In December 2002, one of the largest quarries stopped discharging and the total decreased by approximately 8 Mgal/d.

The largest ground-water discharge in 2001 in Monroe County, Michigan was reported for London Aggregates in London Township. To provide a confirmation of the reported values, a streamflow-gaging station was established on the North Branch (N.B.) of the Amos Palmer Drain downstream from the quarry (fig. 2). This drain received discharge from the quarry. Like many drains in the area, this drain dries out or has very little flow during the summer, except after large storms, without the quarry discharge. During 2002, the reported values from London Aggregates and discharges estimated from stage-discharge relations developed for the N.B. Amos Palmer Drain

Table 1. Self-supplied fresh ground-water use in Monroe County, Michigan.
[Mgal/d, million gallons per day]

Year	Population on self-supplied ground water (thousands)	Estimated per capita use (gallons per day)	Estimated use value	Per capita range (gallons per day) (Mgal/day)	Estimate use range (Mgal/d)
1985	38.66	75.0	2.9	60 - 120	2.3 - 4.6
1990	39.84	72.8	2.9	60 - 120	2.4 - 4.8
1995	67.26	72.6	4.9	60 - 120	4.0 - 8.0
2000	49.64	86.2	4.3	60 - 120	3.0 - 6.0

Table 2. Community public-water supplies in study area, January 2003.
[Mgal/d, million gallons per day, - not available]

Facility	Population served	Percapita use range (gallons per day)	Withdrawal range (Mgal/d)
Bennett Mobile Home Park	70	60-120	0.0042 - 0.0084
Flat Rock Village	830	60-120	.05- 0.1
Petersburg	1,201	60-120	.07 - 0.14
Stoney Trails Apartments	140	60-120	.0084-0.017
Whitehouse Municipal Water System	-	-	.4

Table 3. Reported self-supplied industrial and quarry ground-water use in the Ohio portion of the study area.
[Mgal/d, million gallons per day]

Facility	Use (Mgal/d)
BP Oil Toledo Refinery	0.8
Maumee Plant	.6
Seaway Sand and Gravel	3.8
Sylvania	2.4

Table 4. Estimated agricultural irrigation ground-water use in Monroe County, Michigan.
[Mgal/d, million gallons per day]

Year	Estimated ground-water use (Mgal/d)
1990	0.5
1995	.4
1998	.86
1999	.66
2000	.49
2001	.78

Station using methods described by Rantz and others (1982) were consistent (figs. 16 and 17). The reported quarry discharge generally is the stream discharge during storm events and generally higher than the stream discharge between storm events. Given the potential water loss because of evaporation and seepage, and the potential for additional water at the N.B. Amos Palmer Drain station from the surrounding fields, the reported discharge values appropriately represent the rate of water pumped by the quarry.

In July 2003, discharge measurements were made on streams receiving discharge from the Sylvania, Stone-Co Ottawa Lake, Stone-Co Maybee, and Hanson Quarries. Three of the four discharge measurements were consistent with reported discharge values (fig. 18), but the discharge measurement for the Sylvania quarry was lower than the typical reported discharge. Examination of the recent discharge records indicates that the reported values for 2002-2003 are reasonably accurate.

SIMULATION OF GROUND-WATER FLOW

The bedrock aquifer is modeled as a layered heterogeneous system. The complex heterogeneity implied by the slug-test results, however, is not considered in this regional model. Barometric efficiency analysis and preliminary modeling results indicate that the individual units of the bedrock aquifer should be modeled as separate units to allow for a variation in aquifer properties between units. The nature of flow in the fractured rock system suggests that anisotropy may be important in defining the flow regime.

As discussed previously, the ground water in the bedrock aquifer is under confined conditions for parts of the study area. In other areas, however, ground-water levels are below the glacial deposit/bedrock interface, and the bedrock aquifer is under unconfined conditions. The difference in aquifer response on a regional basis to changes in confinement of the aquifer is considered in the model.

The area of the model is approximately 3,000 mi^2 (fig. 19). Ground water does not flow into or out of the bedrock aquifer across the western salinity boundary, the Detroit River, and the Maumee River. The interaction between these rivers and the bedrock aquifer, however, is considered in the model. There is not a natural hydraulic boundary for the southwestern portion of the model area. This boundary was established to be far enough away from the main area of interest, Monroe County, to minimize the effect of the boundary on simulation results. The value assigned to the southwestern boundary was adjusted during model calibration to be consistent with the regional flow in the ground-water system described by Bugliosi (1999). A boundary in Lake Erie was established at a distance from Monroe County to lessen its direct impact on the simulation results, and the value of this boundary was set to the average lake level for each time period simulated. No flow is allowed through the bottom of the bedrock aquifer through the Salina C-Shale. Where the bedrock aquifer is confined by the Coldwater and Antrim Shales, no flow is allowed. Leakage between the bedrock aquifer and the glacial deposits is simulated.

The major ground-water withdrawals simulated are quarry dewatering; self-supplied domestic, municipal and industrial use; and agricultural and golf-course irrigation. Water may exit or enter the bedrock aquifer as leakage to or from the overlying glacial deposits. Water also may enter or exit the model area through the constant-head boundaries in Lake Erie and in the southwestern edge of the domain. Based on the regional flow analysis (Bugliosi, 1999), water enters the domain across the southwestern boundary and exits at Lake Erie.

Numerical Model

The equation describing ground-water flow cannot be solved analytically for the layered, irregular, and heterogeneous domain required for this regional study. Therefore, a numerical model that approximates this equation was used. MODFLOW-2000 (Harbaugh and others, 2000) was the computer code used to generate the numerical model of ground-water flow in the Monroe County region. Argus ONE and the Argus-ONE MODFLOW interface was used to generate the model grid and prepare the input files (Argus Holdings Limited, 2002; Winston, 2000). The ArcView and ArcMap Geographic Information Systems (ESRI, 2002a and ESRI, 2002b) also were used to generate information for the model input and to visualize model output. The numerical model requires specification of the geometry of the model domain, the boundary conditions, and parameters specifying the hydraulic conductivities, volumetric fluxes, and specific storage.

A finite-difference method is used in MODFLOW-2000 to approximate the equation that describes ground-water flow. The program allows the aquifer to convert between confined and unconfined conditions, and it has the capacity to simulate the other processes and boundary conditions required for the Monroe County regional model. The major limitation of the model for this application is that flow in the bedrock aquifer is dominated by flow through individual fractures, whereas flow through an equivalent porous media is assumed in the model simulation. This simulation of a fractured system as an equivalent porous media is typical of regional models (Anderson and Woessner, 1992; National Research Council, 1990), and at the scale of this model, this assumption should not hamper

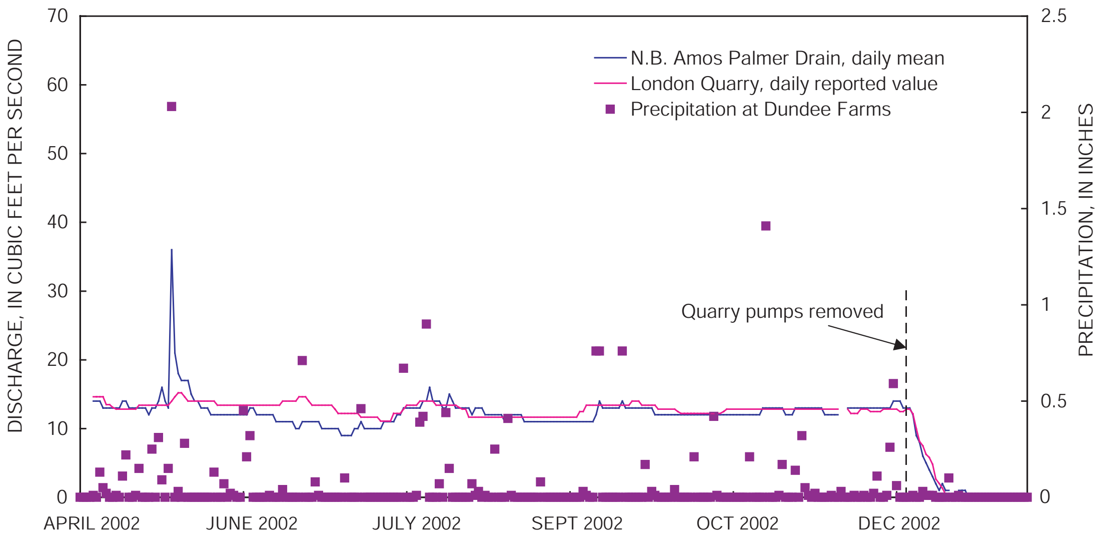


Figure 16. Reported London Quarry discharge and computed flow at the N.B. Amos Palmer Drain streamflow-gaging station, Monroe County, Michigan. Also shown are precipitation values recorded at the Dundee Farms weather station (Michigan Automated Weather Network, Michigan State University).

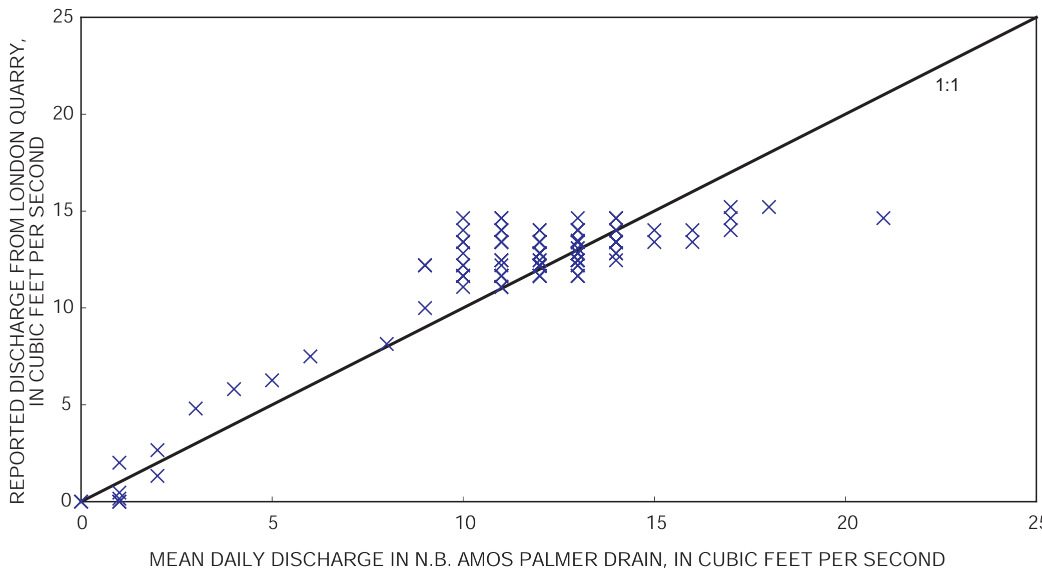


Figure 17. Reported discharge from London Quarry and computed discharge at the N.B. Amos Palmer Drain gage, Monroe County, Michigan.

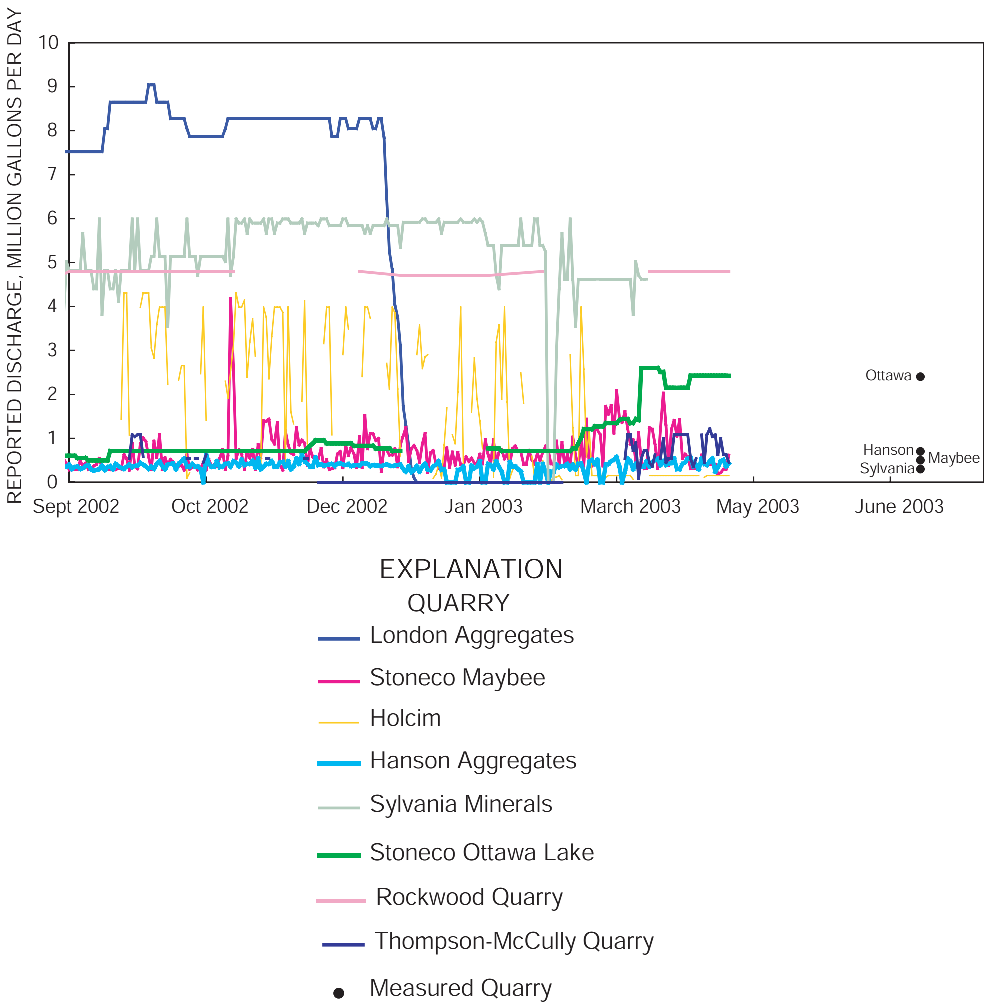


Figure 18. Reported quarry discharges, Monroe County, Michigan, 2002-2003. Values also are shown where gaged by U.S. Geological Survey at streams receiving discharge in July, 2003 .

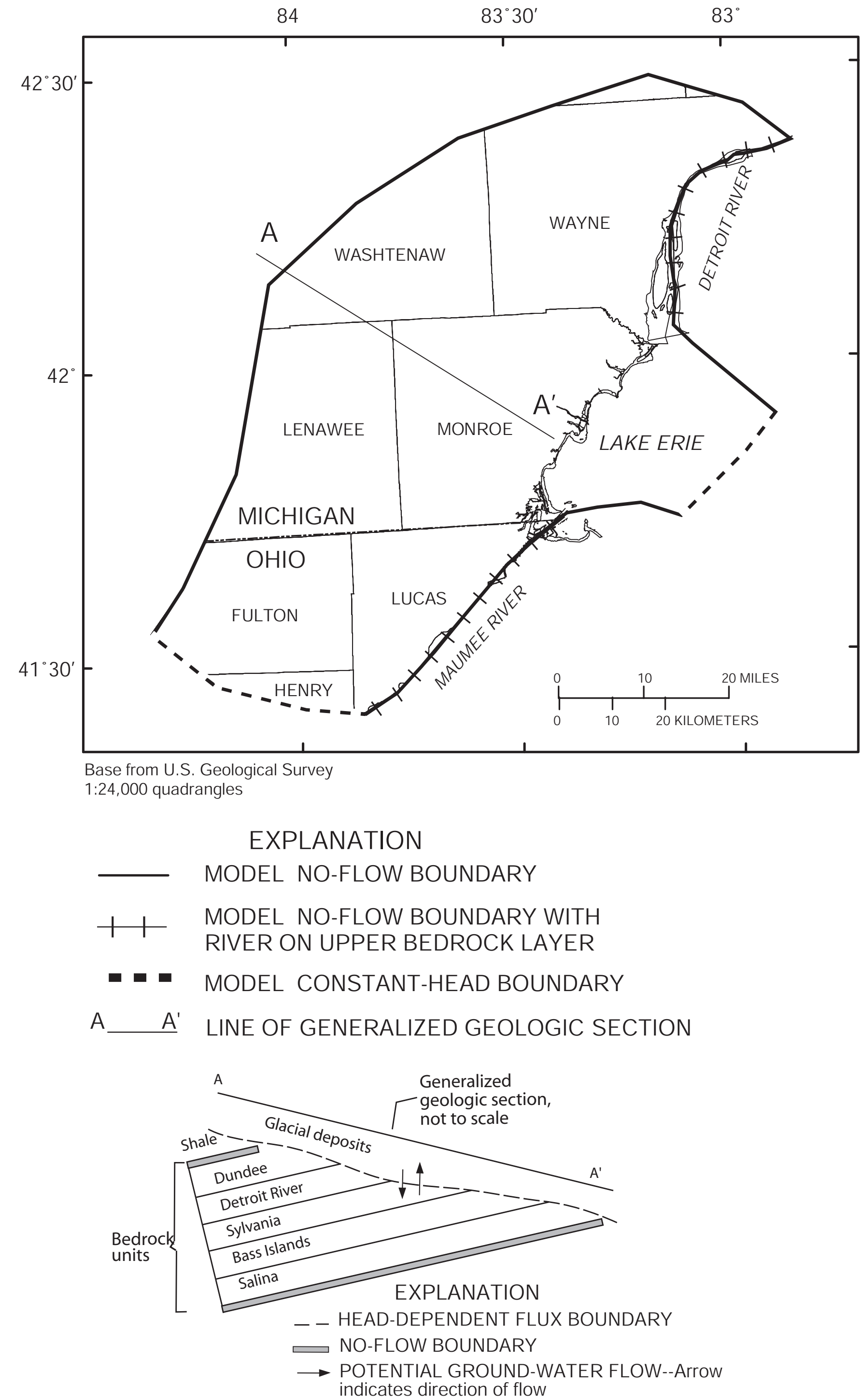


Figure 19. Model area and boundary conditions in the study area, Monroe County, Michigan and surrounding area.

the application of the model or the evaluation of the results. For site-specific evaluation, however, the fractured nature of the aquifer becomes important and field observations may deviate substantially from the simulation results.

Map Projection and Units Used

Michigan georef coordinates (Berry and Bormanis, 1970) were used as base coordinates for the model. Meters are used in this projection to define the map position, and Standard International units were used for the entire model. The simulation results have been converted to English units for this report to conform to standard usage in the area.

Relation Between Numerical and Conceptual Models

The conceptual model is the basis of the numerical model. The finite-difference approximation used in MODFLOW-2000 requires that the model area be subdivided into a regular grid of cells. For this application, the grid is composed of 10 layers, 297 rows, and 194 columns of finite difference cells. Cells outside the study area, and cells in layers eastward of the subcrop, are designated as inactive cells and are not used in the computation of water levels. There are approximately 400,000 active cells in the finite-difference grid. The grid is oriented with the strike of the bedrock units to anticipate anisotropy aligned with the strike (Nicholas and others, 1996). The grid is non-uniform. Because of the large withdrawals by quarries in the region, and the desire to evaluate the effects of all withdrawals on the ground-water system, the grid was refined to approximately 400 by 400 ft in quarries that were active in either 1991 or 2001. The largest model cells were approximately 6,500 by 6,500 ft in areas furthest away from Monroe County. The required input for the model are described in this section, but the specific values used in the model are difficult to present in the report. Model input data arrays are available by request from the Michigan District office.

Each bedrock unit is modeled as two numerical layers of equal thickness. The bedrock units dip from their subcrop northwest into the Michigan Basin. A generalized cross section through the model extending from Lake Erie through northwestern Monroe County into the Michigan Basin is shown on figure 19. The subcrop of each unit, and the entire Dundee formation, is bounded above by glacial material to the east and shales in the western portion of the study area. The lower boundary is a no-flow boundary. The northern, western, and southeastern boundaries of the model are also represented as no-flow boundaries. No flow was allowed across the outer boundary of the cells along the Detroit and Maumee Rivers. The upper cells for the numerical model along these rivers were modeled using the MODFLOW River Package, as discussed below. These rivers appear to act as regional hydraulic boundaries, but the vertical gradients near the rivers are not known. With this choice of boundary condition, the simulation considers the interaction between bedrock ground-water and the Maumee and Detroit Rivers. Constant-head boundaries were used for the remaining model boundaries (fig. 19).

Confined and unconfined conditions were considered in the designation of the MODFLOW layer types for each bedrock unit. The layer-property flow package (Harbaugh and others, 2000) was used to define the geometry and designations of the numerical layers. For each bedrock unit, both numerical layers were specified to be convertible layers. In convertible layers, if the water level for a cell falls below the top of the cell, then the saturated thickness of the cell is computed and used in the approximation. Under transient conditions, the storativity of the cell in this case is replaced by its specific yield. If the water level in the cell falls below the bottom of the cell, the cell is removed from the calculation for an iteration. These cells can be re-wet during the iterative solution process if surrounding cells have a water level greater than 3.28 ft above the bottom of the cell. The value of 3.28 ft was set during model calibration.

Leakage from the overlying glacial deposits is simulated using the general-head boundary package (Harbaugh and others, 2000). This package adds or removes water from the finite-difference cell based on the hydraulic conductance (hydraulic conductivity multiplied by the area available for flow, divided by boundary thickness) and the difference in water levels between the cell and the general-head boundary. As the water level decreases in the bedrock aquifer, the potential for leakage from the glacial deposits increases. In the field, once the bedrock aquifer becomes unconfined, the leakage from the glacial deposits will become independent of the water level in the bedrock aquifer. Downward leakage from the glacial deposits to the bedrock aquifer should become a function of the hydraulic properties of the glacial deposits and the perched water level within the glacial deposits. Complicating this process, capillary behavior also will change the response of the system as the bedrock aquifer becomes unconfined. The effects of capillarity, however, cannot be quantified without site-specific testing regarding the geometry, size, and distribution of fractures in the bedrock. Because of the uncertainties associated with quantifying flux across this boundary and the anticipated importance of this boundary to the regional model, a conservative choice that maximizes the potential leakage through the glacial deposits was made for this boundary.

The general-head boundary increases the flux from the glacial deposits to the bedrock aquifer as the water level in the bedrock aquifer declines. This choice may moderate the local effect of high-capacity pumping by allowing increased leakage compared to the field situation. This boundary condition was applied in the subcrop of each bedrock unit. Conductance values were defined by designating a MODFLOW parameter to the different glacial lithologies shown in figure 3. In Monroe County, where a sufficient number of WelLogic logs were available for hydrogeologic analysis, the glacial lithology parameter was scaled by the effective vertical hydraulic conductivity discussed earlier and shown in figure 4. The individual well-log estimates were averaged in the preprocessing software to generate a value for each finite-difference cell. This scaling allows the variability shown in figure 4 to be represented in the model. The Coldwater and Antrim Shales are expected to restrict leakage, and the hydraulic conductance through the Coldwater and Antrim Shales was set to zero in the model.

The general-head boundary was not assigned to a small zone around each quarry to prevent unrealistic values of leakage from being computed as the water level declined in cells designated as quarry cells. This zone may represent the disturbed area around the quarry where glacial materials are reworked after exposing the rock or the zone where the quarry operations essentially de-water adjacent glacial deposits.

The value of the boundary head was estimated by intersecting the elevation contours and stream reach information from the available 1:100,000 USGS digital line graphics (DLG) from the model area. The intersections approximate the average elevation of the water surface at streams and rivers. The water table is at this elevation at the streams and rivers. The water table is actually at a higher elevation between the rivers and streams, but for the regional model scale, this approximation is considered satisfactory. For cells under Lake Erie, the appropriate lake level from figure 15 was used. The sensitivity of the model to changes in water-table elevation is considered in this report, because it is important in the leakage to or from the bedrock aquifer.

Larger streams and rivers in the model area may affect ground-water levels and flow paths in the bedrock aquifer. The river package (Harbaugh and others, 2000) was used to simulate this effect. For the rivers, the exchange of water between the aquifer and river was estimated as the product of the river bed hydraulic conductance and the difference between the water level in the finite-difference cell below the river and the river stage. The river stage was estimated by intersecting the river with topographic contours as described previously. The conductance was set using a MODFLOW parameter and includes the thickness of the glacial deposits separating the river from the bedrock aquifer. In this way, the river has less effect on the bedrock aquifer when separated by thick glacial deposits and more effect where the glacial deposits are thinner.

Agricultural and domestic water-use demands were modeled as averaged values across the study area. Domestic use often is neglected for regional studies, because much of the water pumped by homeowner wells is returned to the ground-water system through septic tanks or the return flow associated with watering gardens and lawns. For the most conservative model simulation in terms of the effect on the bedrock aquifer system, the estimated domestic use was removed from the bedrock. Return flow was assumed into the overlying glacial material where it is available to recharge the bedrock through leakage as described previously. Average values were estimated using data from Monroe County and applied to the entire study area, in the case of domestic supply, and to the rowcrop land use from the National Land Coverage Database (NLCD), in the case of agricultural use. Because the row crop coverage is not uniform across the region, the 30 m NLCD data were aggregated into 600 m by 600 m cells. The percentage of 30 m pixels in the larger block that were designated as rowcrops was determined. The average irrigation rate for the rowcrops was multiplied by this percentage so that areas entirely covered by rowcrops were subject to the average demand, and those areas with less crops were subject to less demand. Model areas overlying shale are assumed to get water from glacial deposits and these irrigation demands were not imposed on the region of the model domain overlain by shale. Few wells are expected to be drilled through the shale into the carbonate aquifers. The average demands were modeled using negative values in the MODFLOW recharge package. The demands are input as length per time and multiplied by the area of the finite-difference cell to obtain the volume of water removed from the cell.

Pumpage from golf courses, municipal supply, and industrial wells were modeled as point stresses using the MODFLOW well package. The well package requires the pumping rate and location of the well. The Argus ONE MODFLOW preprocessor apportions pumping from different numerical layers based on the transmissivity of the layers, if the well spans multiple numerical layers. Three major wells were identified on the potentiometric-surface map of Lucas County, Ohio by Breen (1989). The pumping rates at these wells, however, were not provided. To simulate the effect of these wells on the regional ground-water system, the MODFLOW drain package was used. A drain was specified at the appropriate finite-difference cell and the mapped ground-water level in the well was used to specify the drain elevation.

Quarries were simulated using the MODFLOW drain package. The model requires the elevation of the drain and the hydraulic conductance between the drain and the aquifer. To simulate a quarry, the conductance was set to 328 ft/d so that the water level in the finite-difference cells adjacent to the quarry would be very close to the level of the quarry. The elevation of the drain was set to the elevation of the quarry sump. This information was available for some quarries and estimated for the others. The discharge rates reported by the quarry operators were for model calibration. All of these imposed withdrawals are illustrated in figure 20.

Time Periods Simulated

The model time periods simulated, the stresses on the system for each time period, and the calibration targets for each time period are summarized in table 5. Three time periods were considered: predevelopment (approximately 1900), developed (1991), and recent (2001). Each period was simulated under steady-state conditions. The purpose of the steady-state 2001 simulation was to examine the long-term ground-water levels in the region given 2001 conditions. Field observations imply that the ground-water system, although showing some seasonal variation, may have been at a cyclic steady state until the late 1980s or early 1990s.

Model Calibration

In model calibration, input values are sought that minimize the difference between the simulation and field observations. Two types of observations were used in the calibration: water levels and fluxes. Use of flux observations greatly improves model calibration compared to using water levels only. For this reason, the reported quarry fluxes were used in the calibration procedure and were not imposed on the model as other demands, such as wells. The model must be calibrated, because site-specific measurements of aquifer properties, such as hydraulic conductivity, may not be appropriate at the scale of the regional model. Other values, such as leakage between the glacial deposits and bedrock aquifer, have not been measured directly. In the MODFLOW-2000 calibration procedure, parameters are established to define various hydraulic properties of the model (Hill and others, 2000). For this application, parameters were set up for each geologic unit, for each glacial lithology with a general-head boundary conductance, for river bed conductance, and for the specified constant-head for the southwestern boundary.

For the southwestern boundary, the parameter for the constant-head boundary fixed the value near the Maumee River to the river elevation. For the remainder of the boundary, a factor multiplying the difference between the river elevation and the calculated water-table height was sought. The product of the parameter and the difference between the river elevation and the calculated water-table height was added to the river elevation for the entire boundary. Therefore, the constant-head boundary always specifies the river elevation on the eastern end of the boundary and allows the head specified for the rest of the boundary to increase more where the glacial deposits are thick and less where they are thin. A value of this parameter of one gives a constant-head boundary fixed to the estimated water-table height.

During calibration, many parameters were found to have low sensitivity or appeared to lack sufficient observations to yield reasonable results using inverse procedures. The values for these parameters were fixed at best estimates and the remaining parameters were used in the inverse procedure. The values of the fixed parameters were varied to try and yield the most reasonable and best calibrated model. As stated, the 1991 scenario was used for calibration. The set of calibrated parameter values also was checked using predevelopment simulations to provide an additional model constraint. The parameters used to define the model and the final calibrated values for each parameter are summarized in table 6.

To check the model at the calibration targets, simulated water levels in the uppermost active rock layer were compared against water levels from the USGS monitoring wells in Monroe County, WelLogic ground-water elevations, and water levels reported by Breen (1989) and Breen and Dumouchelle (1991). The observed values were treated as steady-state heads and values were estimated for the USGS monitoring wells to account for observed seasonal changes during the early portion of the time series. The observations were weighted differently in the inverse procedure. Water levels from the USGS monitoring wells (G wells) were assumed to be as accurate as the contour interval on the topographic maps used to estimate the elevation of the well head. Measurement or any other error for these wells is assumed to be small compared to the elevation error. The uncertainty in a USGS topographic map is quantified by the statement that 90 percent of the elevations are within one-half the contour spacing of the true value. The 90-percent confidence interval is 1.6, if the error follows a Gaussian distribution. The standard deviation, entered as the observation weight to the parameter-estimation process, is the value of the contour interval divided by twice the confidence interval. This computation yields a water-level observation weighting value of 3 ft for the simulation. Because the WelLogic data may have greater error in

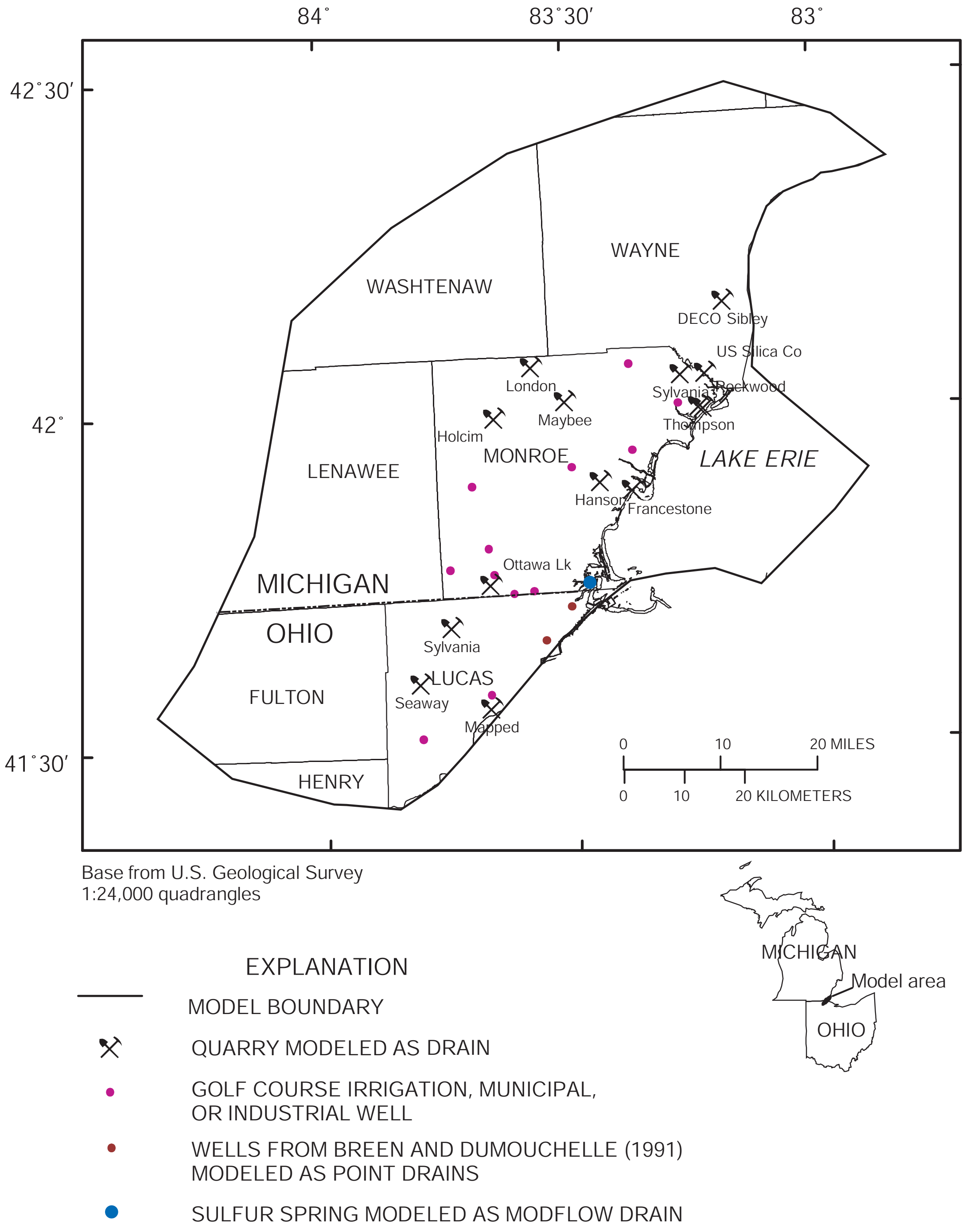


Figure 20. Map of major ground-water withdrawals within the study area, Monroe County, Michigan and surrounding area.

Table 5. Time periods simulated in the ground-water flow model of the Monroe County regional study

Model feature	Predevelopment	Developed	Recent
Time simulation	Steady-state	Steady-state	Steady-state
General-head boundary stress	Drained, crossing of surface water features for small drains and contours used to fix stress Lake Erie level = 571.7 feet above NGVD 29	Drained, crossing of surface water features for small drains and contours used to fix stress Lake Erie level = 571.7 feet above NGVD 29	Same as developed simulation Lake Erie level = 569.7 feet above NGVD 29
Lateral boundaries	As described in text	Same as predevelopment	Same as predevelopment
Domestic wells	None	Wells from Well Logic database with estimated demand	Wells from Well Logic data base includes Developed condition and additional wells
Irrigation wells	None	Estimated irrigation stress imposed for cells with rowcrop land-use coverage.	Estimated irrigation stress imposed for cells with rowcrop land-use coverage.
Industrial wells	None	As noted in table 3	As noted in table 3
Quarry dewatering	None	Reported values used in model calibration	Simulation results compared to reported values
Calibration Target head	Pattern of flowing wells and general magnitude of heads	WelLogic ground- water levels, U.S. Geological Survey monitoring wells	Not calibrated - values compared to U.S. Geological Survey monitoring wells and WelLogic values to test mechanisms for observed declines
flux	Checked flux from Sulfur Spring	Sulfur Spring and quarry dewatering estimates	

Table 6. Parameters used in numerical model with parameter type and final calibrated values. Parameters below the double line in the table are fixed during parameter estimation and were manually adjusted

Parameter	Type	Value (in feet per day)	Comment
Moraine	General-head boundary conductance factor	0.08	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance
Detroit River	Layer-property flow hydraulic conductivity	32	
Bass Islands	Layer-property flow hydraulic conductivity	5.0	
Sylvania	Layer-property flow hydraulic conductivity	54	
Salina	Layer-property flow hydraulic conductivity	5.1	
Alluvial	General-head boundary conductance	0.03	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance
Buried Valley	General-head boundary conductance	0.003	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance
Complex	General-head boundary conductance	0.03	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance
Erie lakebed	General-head boundary conductance	0.0026	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance
Lacustrine clay	General-head boundary conductance	0.0026	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance. Where WelLogic-based effective vertical conductances are available, this parameter also is scaled by an interpolated effective vertical conducrance divided by 8.64E-02 in areas mapped as lacustrine sand

Table 6. Parameters used in numerical model with parameter type and final calibrated values--Continued -

Parameter	Type	Value (in feet per day)	Comment
Lacustrine sand	General-head boundary conductance	0.0013 ft/d	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance. Where WelLogic-based effective vertical conductances are available, this parameter also is scaled by an interpolated effective vertical conducrance divided by 8.64E-02 in areas mapped as lacustrine sand
Not mapped	General-head boundary conductance	0.003 ft/d	This value is multiplied by area of cell and divided by thickness of glacial deposits to yield required conductance
Anisotropy	Horizontal anisotropy	1.0	Horizontal (x-y) anisotropy, aligned along rows that follow the general strike of bedrock units
Drain	Drain conductance	330. day-1	Multiplied by area of cell contributing to quarry flow
Southwest head	Constant head boundary head	0.9	Factor used to set constant-head boundary, a value of 1.0 sets the boundary to approximate water-table elevation
River bed	Riverbed conductance	0.03 ft/d	Multiplied by length of river in a cell to yield required river bed conductance
Detroit River	Riverbed conductance	3.3 ft/d	Multiplied by length of river in a cell to yield required river bed conductance
Maumee River	Riverbed conductance	3.3 ft/d	Multiplied by length of river in a cell to yield required river bed conductance
Dundee	Layer-property flow hydraulic conductivity	3.3 ft/d	

the estimated elevation or measurement of ground-water level, an 80-percent confidence interval was assumed. The standard deviation used for these observations was 4 ft. The water-level values tabulated by Breen and Dumouchelle (1991) were treated in the same manner as the G wells and assigned a weight of 3 ft.

The simulated hydraulic-head distribution (fig. 21) agrees with the expected distribution shown in figure 7. There is a component of flow from northwest of Monroe County and an area of elevated water levels because of leakage through the portion of the glacial moraine that lies east of the Coldwater and Antrim Shales. The regional flow is from the southwest, then east towards Lake Erie. This flow follows the regional pattern given by Bugliosi (1999). Areas of local drawdown are noted near the active quarries.

One check on the model calibration is the spatial distribution of residuals, the differences between the observed and simulated values. The unweighted water-level residuals for the calibrated model are shown in figure 22. Positive and negative residuals tend to be scattered across the model area. Points of closer agreement (+/- 5 ft) also are scattered across the model area. There is an overestimation in the simulation of the water levels in southeastern Monroe County. This overestimation and points with good agreement occur in the same geologic unit and this fit was accepted. There are less data from Lucas County, Ohio, and some pumping in this area may not be simulated as the hydraulic head in this area was consistently overestimated during the calibration process.

A direct comparison between the observed water level and simulated water level, which is the computed hydraulic head at the monitoring well, for the final calibrated trial is shown in figure 23. Note that the residuals follow the 1:1 correspondence line. The maximum underprediction of the model was 35 ft, and the maximum overprediction of the model was 29 ft. These largest two mismatches resulted from comparison of simulated water levels to values reported in the WelLogic database. Many of the residuals fall within 1 ft of the target water level. The water level was overpredicted at 78 observations and underpredicted at 96 observations. For this simulation, the mean weighted residual for water level was 0.74 ft, and the mean residual (unweighted) for water level was 0.88 ft. These values indicate that the model slightly underpredicts the water levels. The root-mean-square error of water-level residuals was approximately 10 ft. The observations from the USGS monitoring wells were given higher weight than the WelLogic values during the inversion procedure, and, in general, the simulation matches the G wells better. The most notable mismatches with the G well monitoring targets are large underestimations of head, greater than 15 ft, at G29 and G30. Both of these monitoring wells are in the karst area in southwestern Monroe County. The water levels in these monitoring wells tends to fluctuate more than other wells in the study area, making estimation of a steady-state observation difficult. In addition, the regional model generalizes the ground-water flow in karst areas.

The calibrated parameters were used to simulate the hydraulic-head distribution under predevelopment conditions. For this simulation, the quarries, and domestic and agricultural withdrawals were removed. Under these conditions, the model simulated water exchange with rivers, Lake Erie, and the glacial deposits. The simulated predevelopment water levels are shown in figure 24. This plot compares well to the regional patterns given by Bugliosi (1999) and the predevelopment conditions given by Nicholas and others (1996). Comparison of the water levels to ground elevation reveals that the hydraulic heads simulated are consistent with the flowing wells in the region summarized by Allen (1977). Heads in parts of Monroe County are as high as 20 ft above land surface, which is the value reported by Sherzer (1900).

In addition to calibrating to the observed hydraulic heads, reported quarry discharges were used in the parameter-estimation procedure. Discharges reported for the quarries were compared to simulated drain fluxes. These observations were weighted using an estimated scaled coefficient of variation for the observation. For flows where some values were reported and where the values did not vary widely, a scaled coefficient of variation of 10 percent was used. For other quarries, a scaled coefficient of variation of 20 percent was used. Various weighting values for the drain flows were used during the calibration process and had little effect on the final calibrated values for the parameters. The quarry discharges used in the 1991 calibration and the simulated flow from the model are summarized in table 7.

The model tends to underestimate most quarry and spring flows. This underestimation is especially true for the Sulfur Spring and Francestone calibration targets. The error at Sulfur Spring may be a result of the regional nature of the ground-water-flow model. Spring flow may result from local heterogeneities that are not captured in the simulation. The larger flow at Holcim quarry, however, is simulated reasonably well. The inherent uncertainty in quarry discharge and sump elevation used to set the drain elevations in the model, and the resolution possible with the scale of the regional model, make these calibration targets more uncertain. The reported discharges are in a reasonable range for the study area and accepted in the calibration.

Another flux constraint checked in the calibration was the discharge to the Detroit River. Gillespie and Dumouchelle (1989) used geophysical techniques to estimate the thickness of glacial deposits and sediments for Great Lakes connecting channels. Using estimated hydraulic

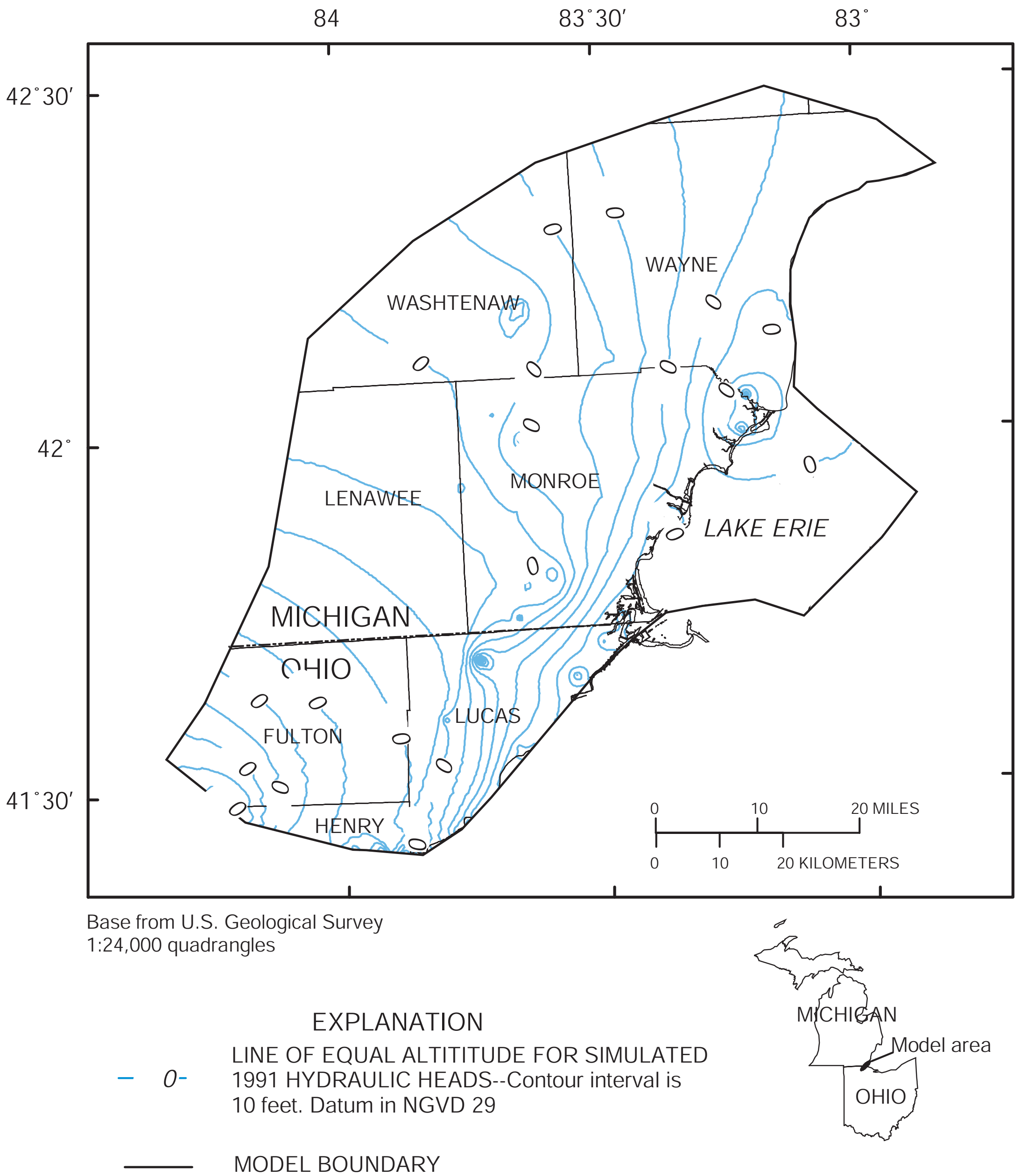


Figure 21. Simulated hydraulic heads for 1991 trial for study area (Monroe County, Michigan and surrounding area).

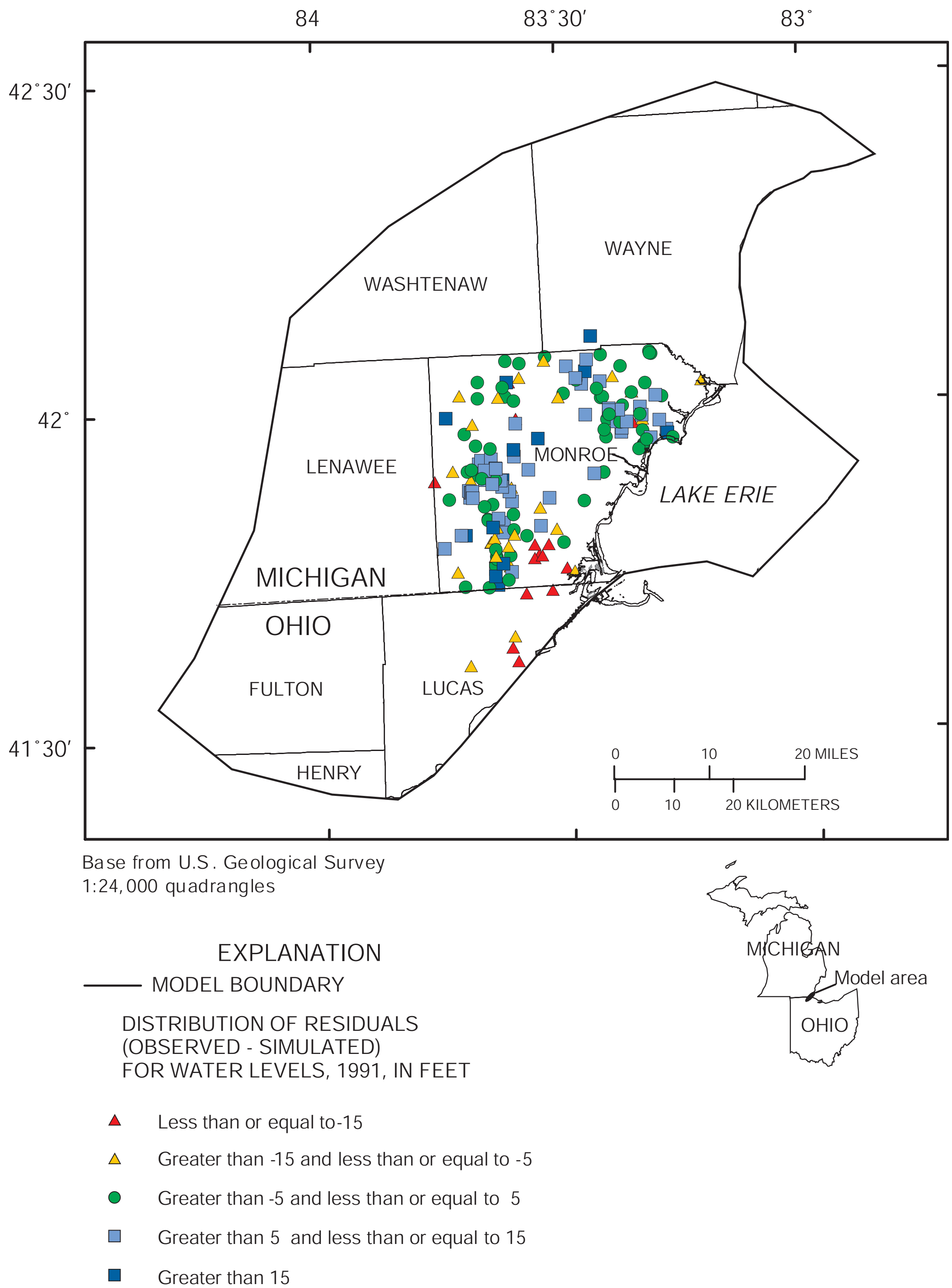


Figure 22. Distribution of residuals (observed - simulated) for water levels in the study area (Monroe County, Michigan and surrounding area) for 1991 final parameters.

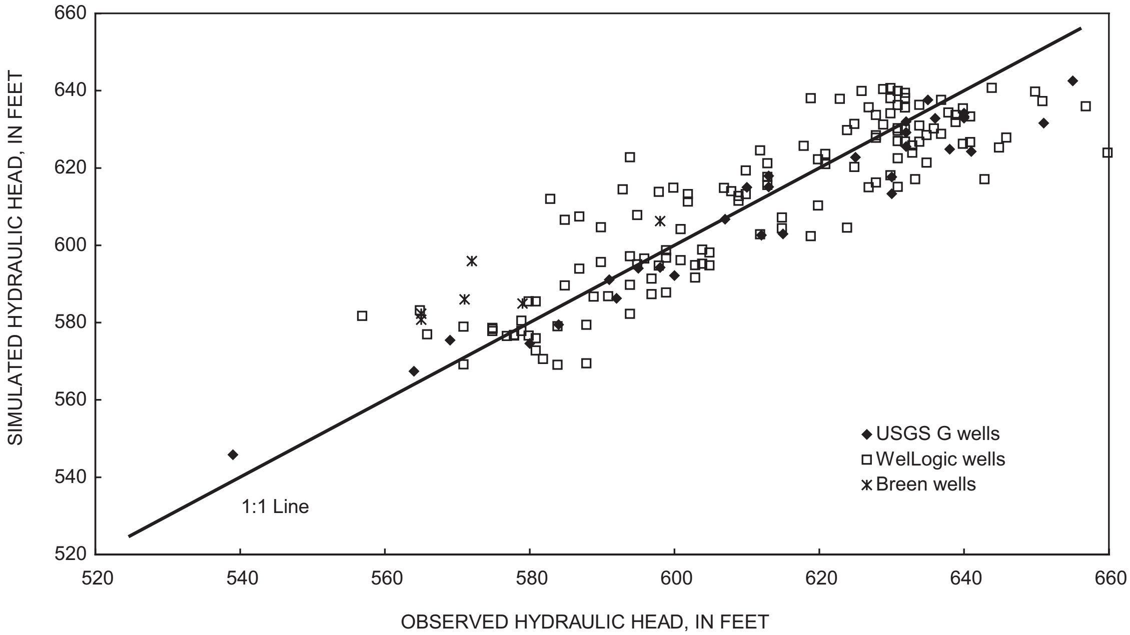


Figure 23. Simulated and observed hydraulic heads for bedrock aquifer wells in the study area (Monroe County, Michigan and surrounding area). Data from U.S. Geological Survey G wells and data from Breen and Dumouchelle (1990) are distinguished from WelLogic.

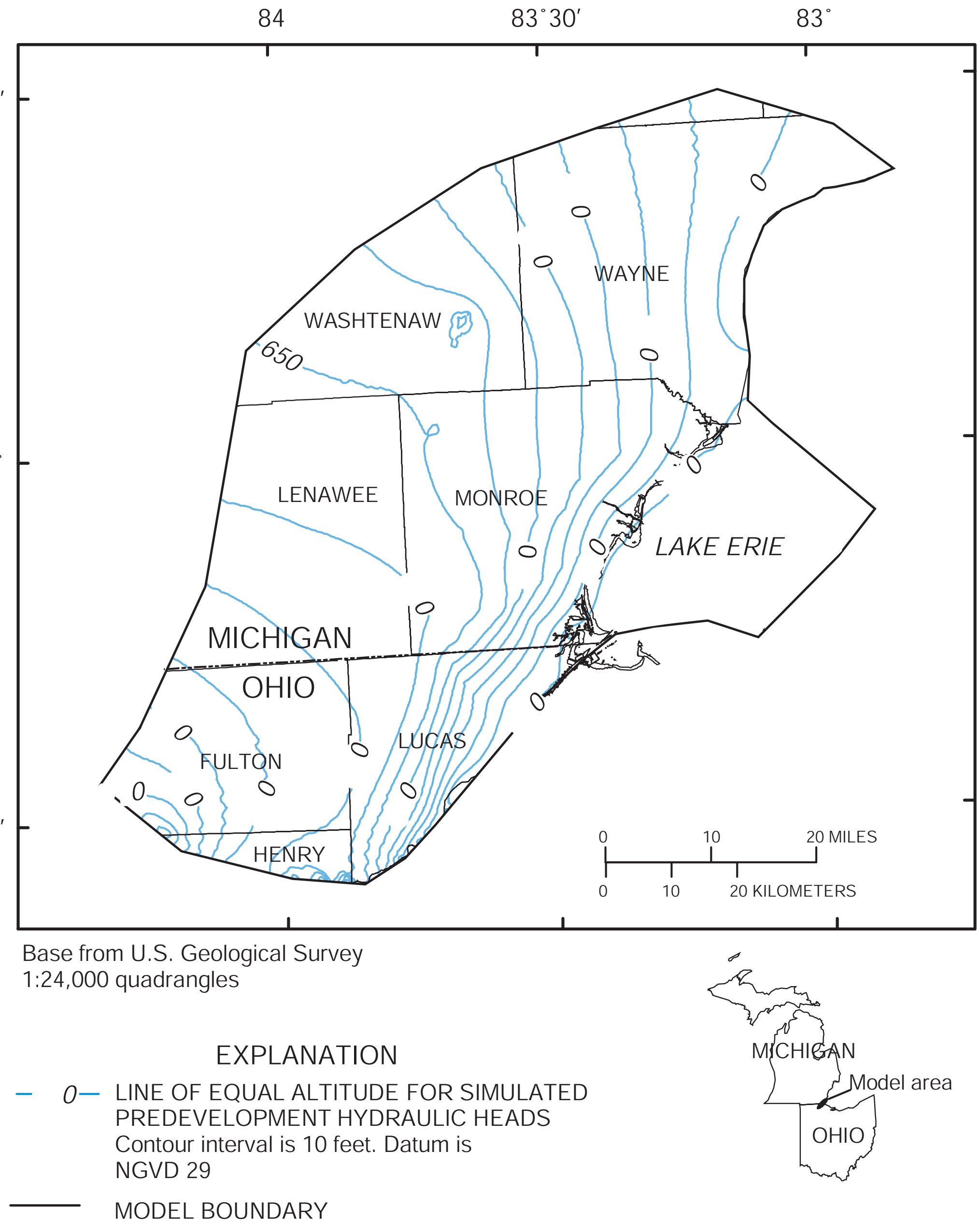


Figure 24. Simulated predevelopment hydraulic heads for the study area (Monroe County, Michigan and surrounding area) using calibrated values determined using 1991 simulation and constrained by predevelopment conditions.

Table 7. Quarry and spring discharges used in calibration of 1991 regional model of the study area and final simulated flows.

[Mgal/d, million gallons per day, - not available, negative values indicate water removed from aquifer]

Quarry or spring	Estimated observed discharge (Mgal/d)	Simulated discharge (Mgal/d)	Residual (observed-simulated) (Mgal/d)
Holcim	-4.8	-4.5	-0.3
Rockwood	-4.1	-3.4	-.7
Stoneco-Ottawa Lake	-.2	-.3	.1
Francestone	-1.5	-.7	-.8
Sulfur Spring	-3.8	-.2	-3.6
Seaway	-	- 3.9	-
Sylvania (Lucas Co.)	-1.6	-1.1	-.5
Quarry on Breen Map	-	-0.1	-

conductivities and an estimated upward gradient of 0.01, discharge values were estimated. For the Detroit River, these values ranged from 0.6 Mgal/d near Lake St. Clair to 20 Mgal/d near Lake Erie where the river and shipping channel reach the bedrock aquifer. Simulated discharge to the Detroit River is 1 Mgal/d. This value is in the reasonable range estimated by Gillespie and Dumouchelle (1989). The thickness of the glacial deposits in the model cells at the Detroit River may be too large leading to an underestimate of flux. On the scale of the model cells used, however, the values used were appropriate. More accurate discharge values would require additional grid refinement near the river and better input data regarding the thickness of the glacial deposits near and under the river, the hydraulic conductivity of these materials, and the actual vertical gradient at the river.

The model residuals were checked for correlation and trends to guard against a biased result despite adequate mean-error statistics. Plotting the weighted residual against the weighted simulation value tests for bias in the simulation (fig. 25). The weighted residuals for the WelLogic observations are distributed with both positive and negative values and show no apparent trend as the weighted simulation value is changed. The G wells and observations from Breen and Dumouchelle (1991) have slightly more positive residuals indicating that the model overestimates the head, but there is not a trend with the weighted simulation value. Simulated quarry fluxes, as discussed, tend to underestimate the observed discharges. In general, simulation results are not biased toward high or low observed values.

The composite-scaled sensitivity for all of the parameters, computed using the calibrated values for the parameters, are shown in figure 26. Note that the five parameters used in the final parameter-estimation simulation have among the highest composite-scaled sensitivities. The lacustrine clay, lacustrine sand, horizontal anisotropy, and southwest constant-head boundary parameters have similar magnitudes for the composite-scaled sensitivity, but during the calibration process these parameters tended to be unstable and prevented convergence of the parameter estimation. The instability may be caused by a lack of observations for some parameters, change in the sensitivity as the parameter values change, or correlation with other parameters in the ground-water system when too many parameters are chosen for parameter estimation. Effects on the simulation results of varying the parameters with larger composite-scaled sensitivity values were tested in a sensitivity analysis performed with the model. Changes in the low-sensitivity parameters should not appreciably affect the simulation results and sensitivity analysis was not performed on these parameters.

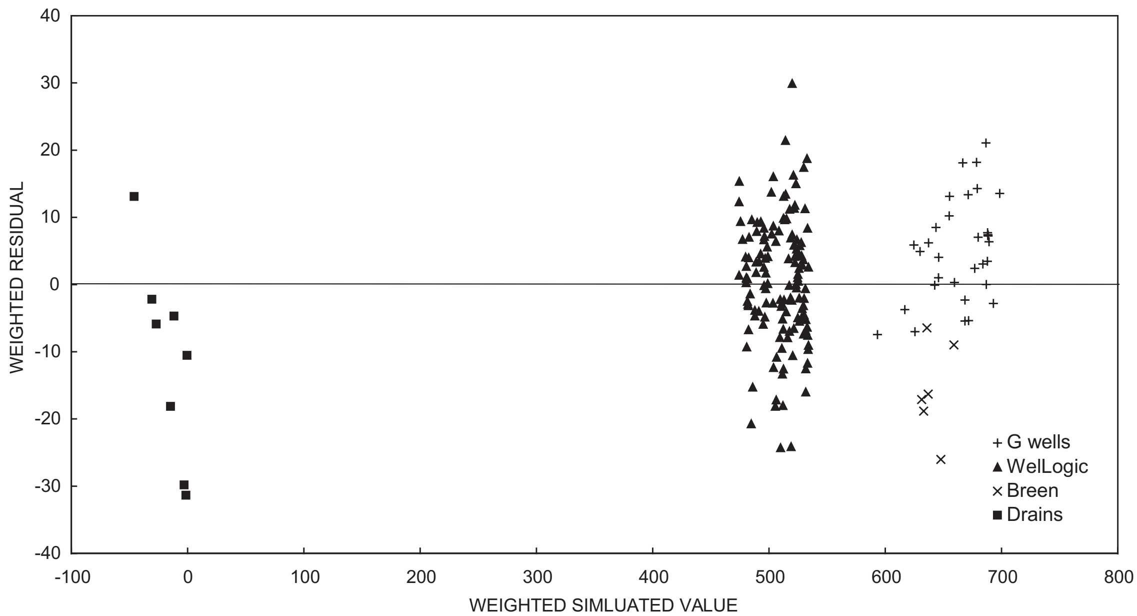


Figure 25. Weighted residuals and weighted simulated values for the model of the study area (Monroe County, Michigan and surrounding area). Residuals are reasonably scattered positive and negative, and they do not show definite trends as the weighted simulated value changes.

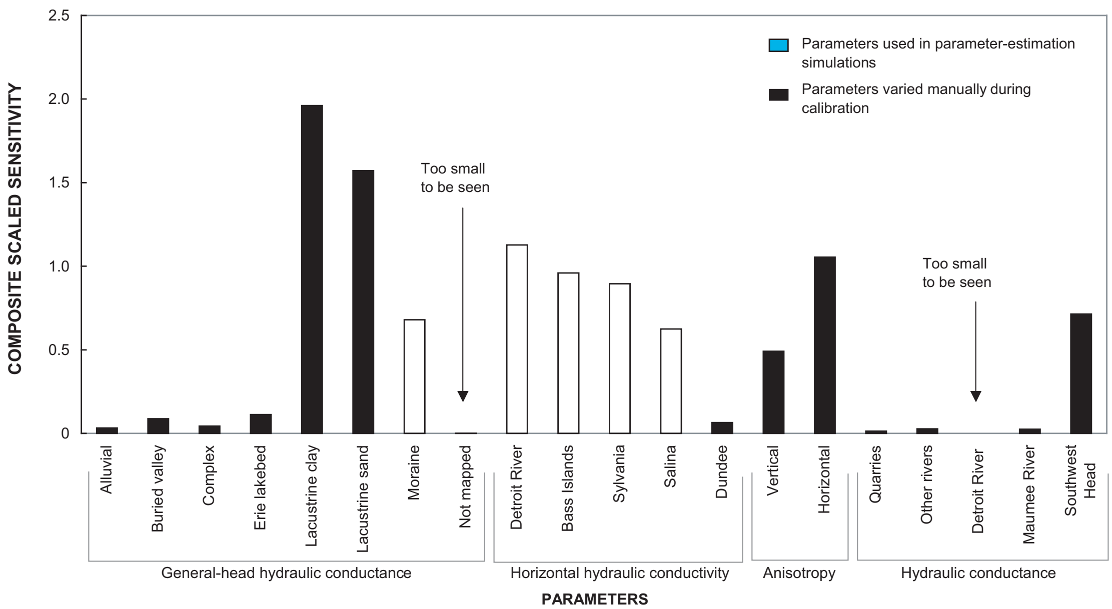


Figure 26. Composite-scaled sensitivities for all parameters used to define the model of study area (Monroe County, Michigan and surrounding area).

The correlation matrix estimated for the five parameters used in the parameter-estimation procedures is

	Moraine	Detroit River	Bass Islands	Sylvania	Salina
Moraine	1	0.6371	0.04088	0.5429	0.2475
Detroit River	.6371	1	-.0082	-.02048	-.09613
Bass Islands	.04088	-.0082	1	-.0878	-.2378
Sylvania	.5429	-.02048	-.0878	1	.05233
Salina	.2475	-.09613	-.2378	.05233	1

This matrix indicates that the parameters used in the estimation process are not excessively correlated, because the off-diagonal values in the matrix are not close to unity. If parameters are correlated, parameters may have to be combined, or more constraints added to the model to help provide a unique set of calibrated values.

A water budget for each simulation is reported in MODFLOW-2000. This budget will be used to investigate causes of ground-water-level declines. For the 1991 and predevelopment simulations, the error in the water budget was 0.5 percent or less. Large errors in the water budget may indicate inconsistencies in the ground-water model, too few iterations in the iterative solver used by the numerical model, or potential problems with simulated water flux across model boundaries. The lack of error in the simulations does not guarantee a consistent model, but it is another indicator of satisfactory model performance. The water-budget values will not be presented for the predictive simulations or sensitivity simulations presented in the next sections, but the mass-balance error consistently was 1.0 percent or less for all simulations.

Exploration of Ground-Water-Level Declines

Ground-water capture

The numerical model was used to investigate the observed ground-water-level declines in the Monroe County area. The evaluation is discussed following the classic series of papers by Theis (1940), Bredehoeft and others (1982), and Bredehoeft (2002) that explores the source of water to wells and the “water budget myth.” This series of papers explains that there is a concomitant decline in water level as withdrawal from an aquifer is increased by pumping — any pumping of water causes a removal of water from storage. Achievement of a new steady-state water level only is possible when the hydraulic gradients caused by the pumping either induce increased recharge to the ground-water system, decreased natural discharge from the ground-water system, or produce a combination of increased recharge and decreased discharge. It is the change in recharge, discharge, or both, which may be termed “capture”, that allows the system to reach a new steady-state, not the virgin recharge or discharge rates for the system. Bredehoeft and others (1982) refer to a focus on the relation between imposed pumping and the virgin recharge rate as the “water budget myth.” In this section, the induced changes in the capture of the bedrock aquifer in response to the change in withdrawals from predevelopment to 1991 and then to 2001 will be explored to help with the analysis of the observed water-level declines. Pertinent fluxes for the calibration simulations are presented in table 8. Information from the exploratory 2001 simulations will be presented in this same form.

The fluxes summarized demonstrate the factors required to achieve steady-state water levels with the 1991 demands imposed on the ground-water system. The flux from the ground-water use simulated for the entire model area is approximately 10 Mgal/d greater than the use estimated for Monroe County. The difference is due to quarries and pumping in Ohio and Wayne County, Michigan, and to the additional domestic and agricultural use across the study area outside of Monroe County. Capture is the total change in the flux across head-dependent boundaries, general-head boundaries, and the flux to rivers. Because these are steady-state simulations, this capture must equal the flux from the demands on the system. The increase in flux across constant head boundaries is a combination of an increase in virgin recharge of 1.1 Mgal/d and a decrease in natural discharge across the boundary in Lake Erie of 1.0 Mgal/d. In the same manner, water exchange with the glacial deposits and Lake Erie changes by both an increase in leakage to the bedrock aquifer (in) and a decrease in leakage from the bedrock aquifer (out). The flux to rivers also is important in this simulation. The change in discharge from the bedrock aquifer to the rivers is approximately 8.5 Mgal/d. This decrease might be questioned if the earlier discussion of base flow and the discussion that base flow may not reflect the condition of the bedrock aquifer is recalled. The major fluxes from the bedrock aquifer in the simulation, however, are to the Maumee and Detroit Rivers, and this small change in discharge may not be evident compared to the river discharge. The discharge in the Maumee River at Waterville, Ohio (USGS Station 04198500) ranged from 300 to 13,000 Mgal/d in 2000, and the average discharge of the Detroit River is approximately 120,000 Mgal/d (Holtschlag and Koschik, 2003).

The spatial distribution of fluxes from the general-head boundaries for the predevelopment and 1991 calibration simulations are shown in figures 27 and 28. In these figures, the change in leakage between the glacial deposits and the bedrock aquifer in Monroe County is clear. Under pre-

Table 8. Fluxes from calibrated models and assessment of changes from virgin rates of recharge and discharge for the study area
[All fluxes in million gallons per day, negative values indicate water removed from bedrock aquifer]

		Flux across constant head boundaries	Flux across general-head boundaries (leakage to/ from glacial deposits and Lake Erie)	Flux to rivers	Flux to quarries, Sulfur Spring, industrial and municipal wells, irrigation, and comestic use
Predevelopment	In	12.4	13.5	0.1	0
	Out	9	6.2	10.6	0.2
	Net	3.4	7.3	-10.5	-.2
1991 Calibrated	In	13.5	29.5	2.3	0
	Out	8	.7	4.1	32.5
	Net	5.5	28.8	-1.8	-32.5
Change from virgin rate		2.1	21.5	8.7	-32.3

development conditions (fig. 27), the bedrock is losing water to the glacial deposits over much of the study area and nearly half of Monroe County. In the 1991 simulation (fig. 28), the bedrock is gaining water over most of the study area and almost all of Monroe County. The localized leakage from the moraine northwest of Monroe County is evident in both simulations. The pumping demands imposed by the 1991 simulation are balanced by capture resulting from a virtual elimination of the natural discharge from the bedrock to the glacial deposits under predevelopment conditions and an increase in the recharge to the bedrock by leakage from the glacial deposits in the western portion of the study area.

2001 simulations

Simulations were done to determine the causes of ground-water-level declines by imposing the estimated 2001 demands described in this report on the model with the 1991 calibrated parameters. The transient stresses from 1991 to 2001 were not used in model calibration, because the conditions used for such calibration simulations would predetermine the causes of the simulated ground-water-declines. For example, if the model were calibrated using these transient stresses, and the general-head boundary were assumed not to change, then a calibrated 2001 model would indicate that the observed declines could be attributed only to increased demands. The effect of below-average precipitation on the water level in the glacial deposits is difficult to assess. Therefore, the best option for exploring water-level declines was to run simulations using the calibrated model based on 1991 and predevelopment conditions, and to perform sensitivity analysis to illustrate the range of reasonable solutions.

To determine if increased demands, both quarry dewatering and other uses, could produce the observed declines, the calibrated model was run using 2001 conditions as described in table 5. Note that the elevation of the water table in the glacial deposits, the general-head boundary value, was not changed. The simulated steady-state water levels under 2001 conditions are given in figure 29. A small area of elevated water levels to the northwest of Monroe County is evident in the figure, and there is additional drawdown across the modeled area compared to the 1991 simulation. The simulated decline in water level is more easily recognized when the 1991 simulated water levels are subtracted from the 2001 simulated heads (fig. 30). Negative values in figure 30 indicate a decline in water level. Note the similarity between the simulated change map and the change map estimated from WelLogic ground-water levels

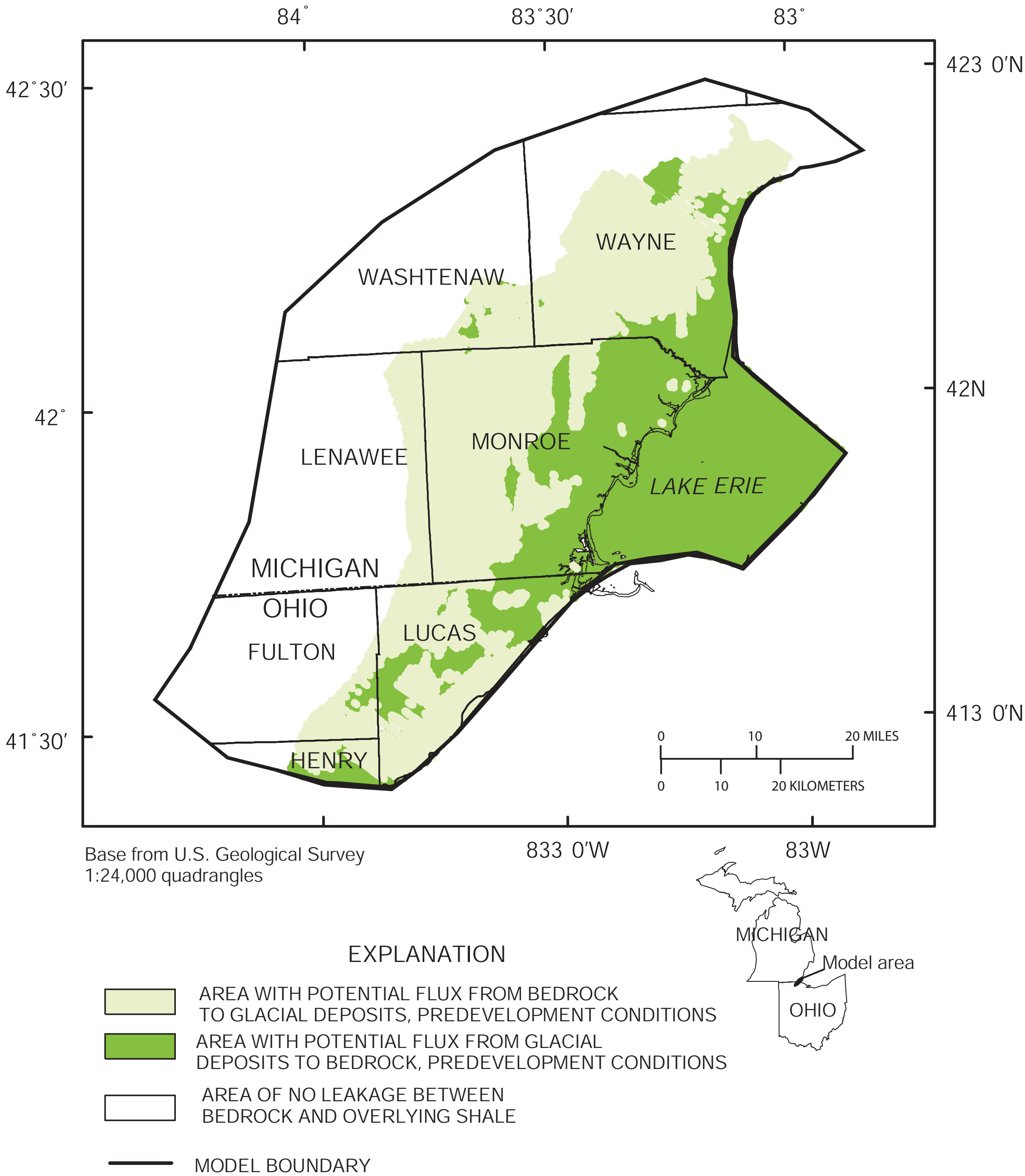


Figure 27. Simulated potential flux direction between glacial deposits and bedrock aquifer for the study area (Monroe County, Michigan and surrounding area) under predevelopment conditions.

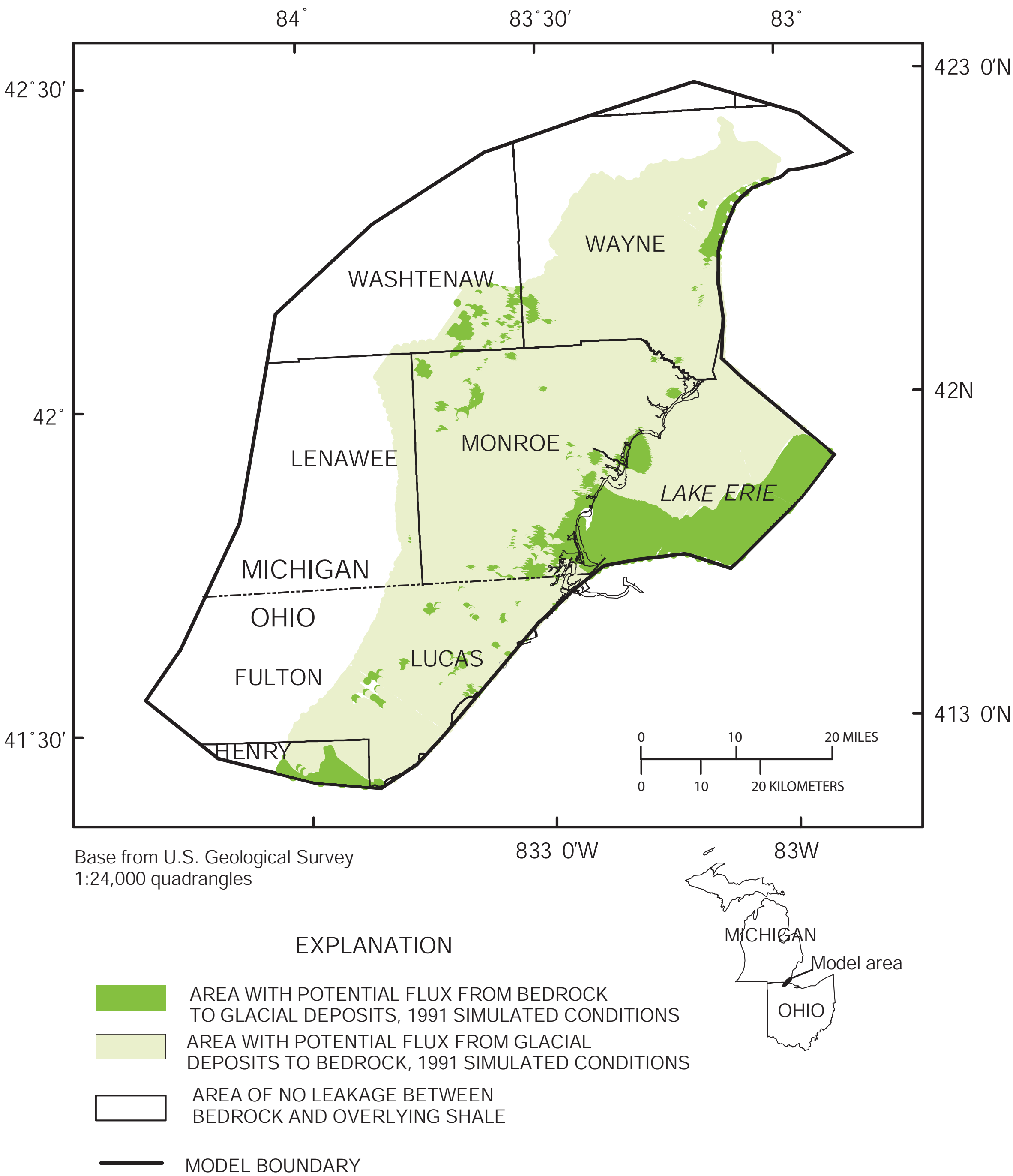


Figure 28. Simulated potential flux direction between glacial deposits and bedrock aquifer for the study area (Monroe County, Michigan and surrounding area) under 1991 conditions.

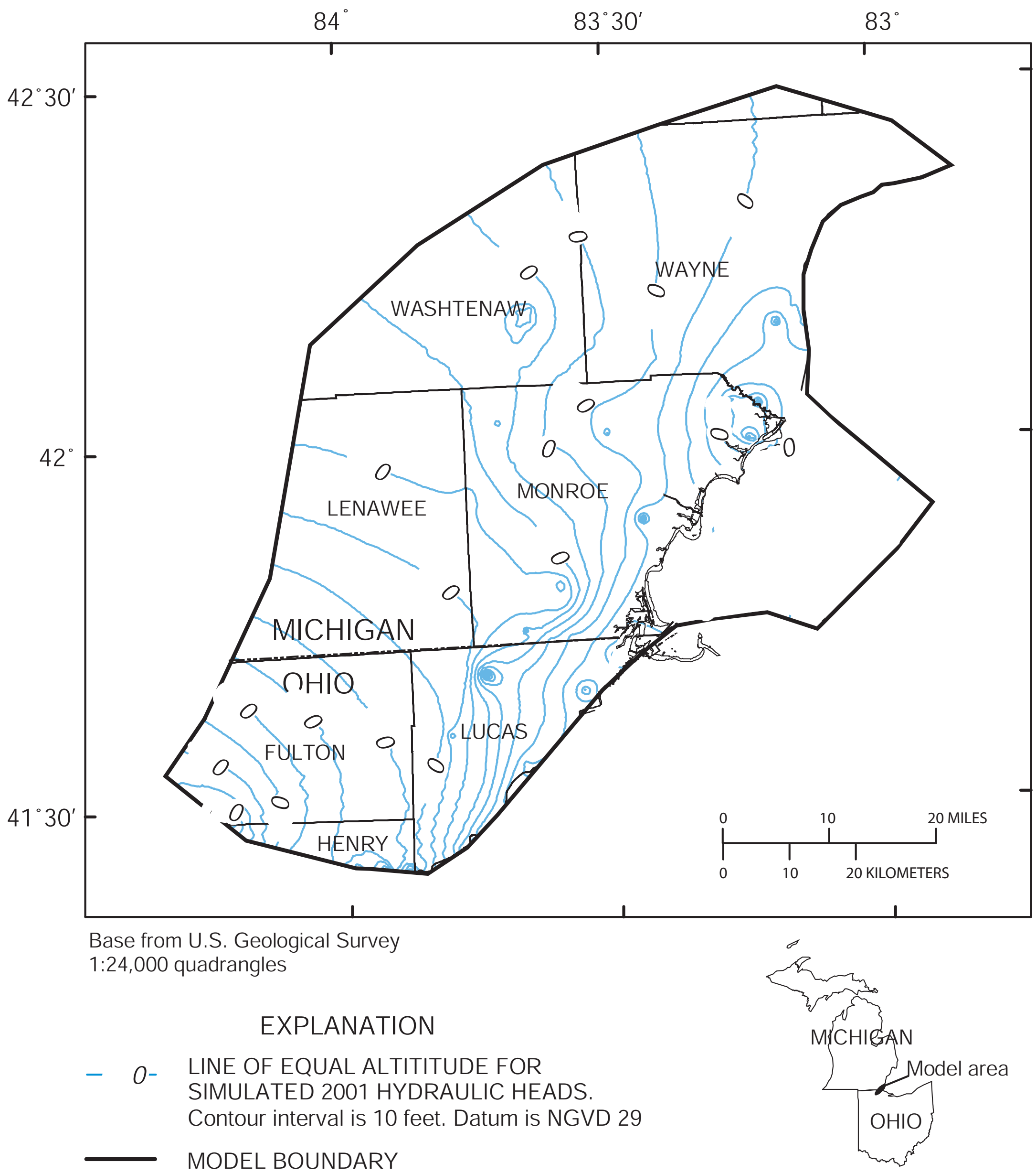


Figure 29. 2001 steady-state simulated hydraulic heads using 1991 calibrated parameters of the study area (Monroe County, Michigan and surrounding area).

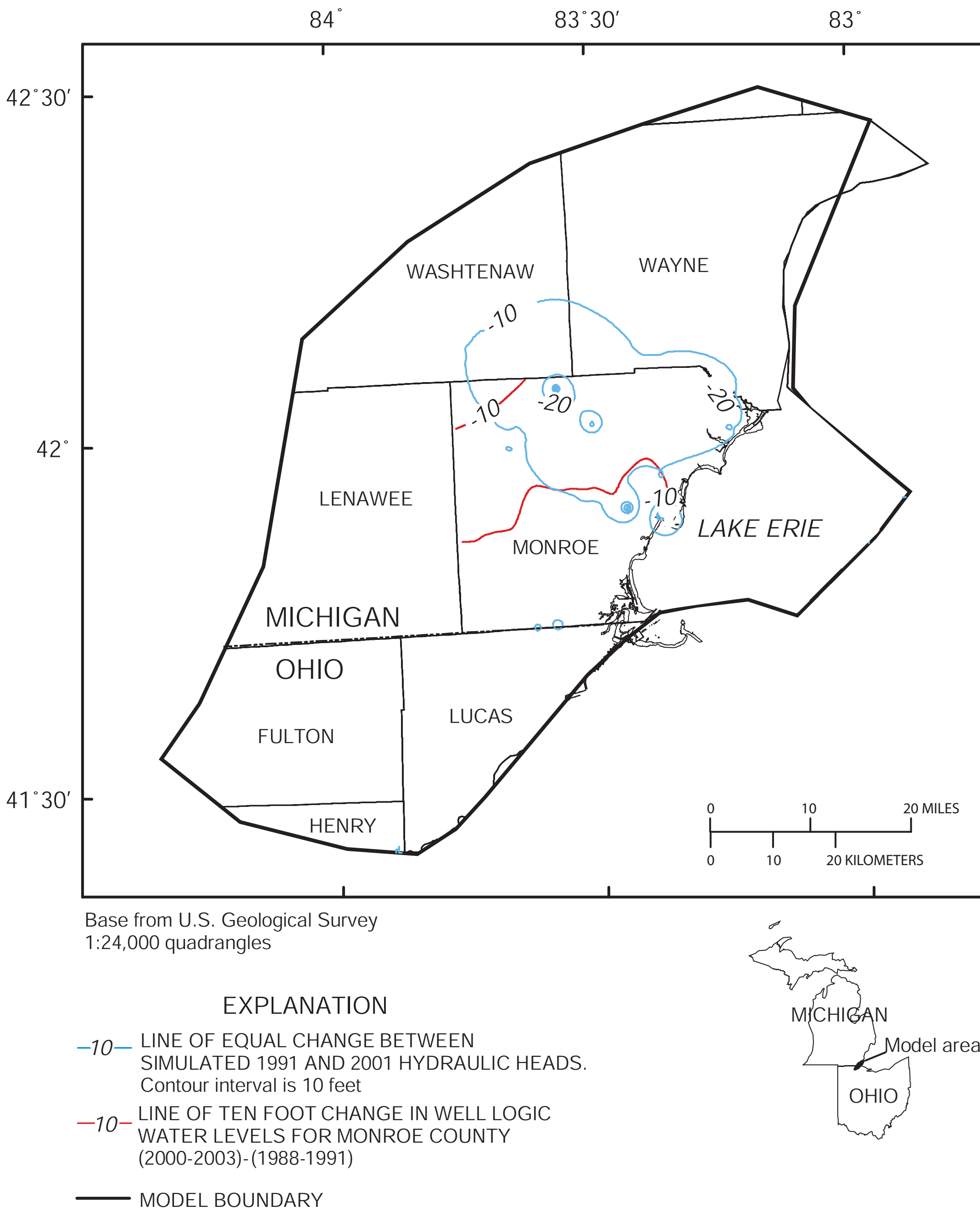


Figure 30. Difference between simulated hydraulic heads, 1991-2001 for the study area (Monroe County, Michigan and surrounding area).

Table 9. Simulated and reported quarry discharges for selected quarries from the 2001 steady-state simulation for the study area
[Mgal/d, million gallons per day]

Quarry	Reported discharge (Mgal/d)	Simulated value (Mgal/d)	Difference (Mgal/d)
Holcim	-3.2	-3.6	0.4
Hanson	-.5	-.9	.4
London	-8	-3.6	-4.4
Sylvania	-4	-3.2	-.8

(fig. 13b). The -10-ft contour from the WelLogic analysis also is shown on figure 30. Both the simulation and the WelLogic analysis show approximately 10 ft of decline across the middle of Monroe County, small declines in the southern townships in Monroe County, and larger declines in the northern townships. The area of 20 ft or more of decline in the north-central portion of the county is smaller in the simulation than in the WelLogic analysis (fig. 13b). The simulation results appear to slightly underestimate the observed regional water-level decline.

Selected quarry discharges estimated in the 2001 steady-state simulation are summarized in table 9. The simulation slightly overestimates discharge at Hanson and Holcim quarries and underestimates discharges at London Aggregates and Sylvania quarries. Some of the difference may be attributed to heterogeneities in the bedrock aquifer that are not simulated in the regional model. Changes in the unit hydraulic conductivity affect discharge at all of these quarries and improving the fit to discharge from one quarry would be expected to degrade the fit at the others. Some of the differences also may be attributed to inaccuracies in model estimates for the area or depth of these quarries.

A similar analysis for other water withdrawals is not necessary because they are input directly as demands into the ground-water model. The simulated domestic, municipal, and irrigation demands match those outlined in the “Water use” section of this report. Any uncertainty related to the ground-water system response to these flows would be related to the uncertainty in the water-use estimates. As in the 1991 simulations, the total flux in the 2001 simulations is approximately 10 Mgal/d greater than that estimated for Monroe County. The total increment in withdrawal from the system is consistent with the estimated increase in withdrawal based on available water-use information.

A more detailed examination of the results reveals that the estimated steady-state heads tend to be higher than the observed values in 2001, especially in the northwest and north-central portion of the county. This information for the USGS monitoring wells is summarized in table 10. Note that simulated water levels in monitoring wells G-22, G-23, G-24, G-25, and GLTO match, but some wells, including G-7, G-8, and G-9, overestimate the water level. Sensitivity analysis can help explore to what extent other factors affect the water levels and can reveal the reasonable range of declines expected because of the inherent uncertainty in the model parameter values.

The 2001 simulation is for steady-state conditions, therefore, the simulation provides an estimate of the ultimate capture, if conditions were held to 2001 values. The increased demands are balanced by increased flux across the general-head boundary (table 11). In the 2001 simulation, nearly all of the additional water withdrawn from the system is provided by leakage through general-head boundaries and interaction with rivers. Recall that the use of the general-head-boundary package to simulate the leakage

Table 10. Simulation results from 2001 steady-state simulation with observed 2001 values and residuals for U.S. Geological Survey monitoring wells in the study area.

Monitoring well	Observed water level (feet above NGVD 29)	Simulated water level (feet above NGVD 29)	Residual (Observed-Simulated)) (feet)
G-1	525	536.5	-11.5
G-2	545	551.4	-6.4
G-3	569	566.8	2.2
G-4	583	577.7	5.3
G-5	587	588.9	-1.9
G-6	593	583.3	9.7
G-7	582	592.7	-10.7
G-8	580	600.6	-20.6
G-9	585	607.7	-22.7
G-11	600	614.4	-14.4
G-12	622	616.8	5.2
G-13	627	601.1	25.9
G-14	603	587.4	15.6
G-15	582	573.7	8.3
G-16	576	566	10
G-18	594	587	7
G-19	626	608.3	17.7
G-20	628	618	9.9
G-21	615	621.9	-6.9
G-22	628	626.9	1.1
G-23	629	626.9	2.1
G-24	629	626.8	2.2
G-25	629	626.5	2.5
G-26	622	631.8	-9.8
G-29	635	620.1	14.9
G-30	649	628	21
GLTO	629	628	1

Table 11. Fluxes from calibrated model and 2001 simulation summarizing ground-water capture. -
[All fluxes in million gallons per day, negative values indicate water removed from bedrock aquifer] -

		Flux across constant head boundaries	Flux across general-head boundaries (leakage to glacial deposits and Lake Erie)	Flux to rivers	Flux across quarries, sulfur spring, industrial and municipal wells, irrigation and domestic use
2001 calibrated	In	14	36.5	3.1	0.0
	Out	7.8	0.9	3.1	41.8
	Net	6.2	35.6	0	-41.8
Change from virgin rate		2.8	28.3	10.5	-41.6
Change from 1991 rate		0.7	6.8	1.8	-9.3

from the glacial deposits to the bedrock aquifer maximizes this flux as the bedrock aquifer model layer becomes unconfined. Because this general-head-boundary flux is the key factor balancing the increase in demand in 2001, this choice of numerical representation of the boundary is a potential cause of the overestimation of head in the simulation. In the 2001 simulation, the flux out to Lake Erie through the general-head boundaries representing the lakebed is reduced to slightly less than 1 Mgal/d. Because the hydraulic-head gradients do not change appreciably at the Lake Erie or southwest constant-head boundaries between the 1991 and 2001 simulations, the flux across the model boundaries only changes by 0.7 Mgal/d.

Sensitivity to imposed boundaries

Sensitivity analysis examining the uncertainty inherent in the model and the effect of changes in different model inputs to the simulated hydraulic head was used to ascertain whether the observed heads are consistent with the 2001 steady-state model results or if other processes affect water levels. Three model boundaries were examined to determine how each can contribute to the observed ground-water declines, and to illustrate how each affects ground-water capture in the study area. The boundaries examined were the level of Lake Erie on the eastern boundary, the constant-head boundary on the southwest border of the model area, and the general-head boundary used to simulate the glacial deposits.

As discussed previously, the level of Lake Erie changed between 1991 and 2001. To examine how much of the observed ground-water decline could be attributed to a change in lake level, the 1991 calibration simulation was run again using the 2001 level of Lake Erie. The change in lake level was -1.7 ft. The steady-state simulation maximizes the effect of changing the lake level by neglecting any time lag in the response of the bedrock aquifer to changes in lake level. The maximum simulated change at a USGS monitoring well location was 0.8 ft at G16. The change in the fluxes across the boundaries because of this change were less than 0.5 Mgal/d. More water exited the system through leakage to Lake Erie, but, at the same time, slightly less water entered the system in areas where the ground-water level is below lake level. The simulation results indicate that changes in lake level minimally effect the observed ground-water declines in Monroe County.

Changes in the constant-head on the southwestern boundary of the modeled area affect the computed flux across the boundary, but if the remaining model parameters of the simulation are held constant, these changes did not greatly affect the simulated heads at the USGS monitoring wells in Monroe County. The boundary head was kept to a small range consistent with the regional pattern suggested by Bugliosi (1999), and extreme changes would be expected to have a greater effect on fluxes.

The potential effect of decreased precipitation or infiltration leading to a decrease in the ground-water level in the glacial deposits was tested through a simulation using the 2001 conditions, but with the water level in the glacial deposits simulated by the general-head boundary reduced by 3.3 ft. This simulation examines the combined effect of changing water-table elevation and imposed demands. Changing leakage through the glacial deposits by changing the water-table elevation, as might be expected during drought periods, affects areas where there is leakage to the bedrock (fig. 31). The difference between the two 2001 steady-state solutions is shown. Negative values result from the lowered water-table elevation in the general-head boundary package. Mass balances reported by the simulations yield a loss of approximately 1.6 Mgal/d in leakage to the bedrock because of the lowering of the water table. The results indicate that ground-water-level declines would be more evident in southern Monroe County if a decrease in leakage, because of low precipitation, caused the regional decline. This simulated distribution of declines does not match the WelLogic analysis or the water levels measured in the USGS monitoring wells.

The hydraulic conductance of the glacial deposits also controls the leakage between the glacial deposits and the bedrock aquifer. The calibrated model parameters prescribing the hydraulic conductance for the general-head boundary condition for the lacustrine clay, lacustrine sand, and moraine units were varied for the 2001 simulation. For each sensitivity simulation, all of the remaining parameters were fixed at the calibrated 1991 values, and the 2001 Lake Erie level was used. In general, increasing the conductance increases the heads for the modeled area. The response at monitoring wells G-7 and G-2 is shown in figure 32. Water levels in both of these monitoring wells were more sensitive to changes in the lacustrine clay parameter. The simulated heads did not respond as much when the conductances were lowered compared to when the conductance was increased. Increasing the conductance by an order of magnitude for the moraine parameter increased the head at G-7 by 12 ft, and decreasing the conductance by an order of magnitude caused a decrease in water level at G-7 of 6 ft. As the amount of water entering the system through the general-head boundary was decreased, the hydraulic heads in the bedrock aquifer lowered, and the flux through the drain cells representing the quarries decreased.

Ground-water capture in the system responds as anticipated (fig. 33). As the hydraulic conductance of the glacial deposits is increased, the water level in the bedrock aquifer tends to increase. This water-level increase causes an increase in quarry flux and a rise in the overall imposed demand of the ground-water system. The increased flux across the general-head boundary balances this increase in demand. The change in flux across the constant-head boundaries is small for these sensitivity simulations. The fluxes from the rivers and imposed demands are able to readjust to changes in the general-head boundary flux. As the flux through the general-head boundary is more restricted, quarry flows decrease and the system reaches steady state without as large of an adjustment to the overall fluxes or heads. The changes shown in figure 33 are with respect to the results of the 2001 steady-state simulation using the calibrated parameters.

Sensitivity to bedrock hydrogeologic parameters

The remaining model parameters examined were the hydraulic conductivities of the Detroit River, Sylvania, Bass Islands, and Salina units and the horizontal anisotropy ratio used for the simulations. The sensitivity to these parameters was assessed by examining the change in the 1991 steady-state simulation compared to the calibrated results. Similar analysis on the 2001 simulation shows similar trends. The most important feature of this portion of the sensitivity analysis is the range of head simulated by the model as the parameters are varied. The sensitivity to hydraulic conductivity parameter changes at monitoring wells G-2, G-7, G-13, G-33, and GLTO are shown in figure 34 and 34a. Note that the model responds differently in different locations to changes in hydraulic conductivity of the bedrock units. For example, a change in the hydraulic conductivity of the Salina unit causes an increase in hydraulic head at G-33 of 6 ft and a decrease at G-7 of 10 ft. For a factor-of-five change in all of these parameters, the resulting change in simulated hydraulic head is on the order of 10 ft.

Water-level declines

The WelLogic analysis and USGS monitoring-well data reveal regional ground-water-level declines, and model results and sensitivity analysis can be used to identify potential causes of these observed declines. In terms of the ground-water capture discussion, the declines may be caused by increased pumping or natural discharge, by decreased recharge to the ground-water system or by a combination of these factors. Effects of changes in natural discharge and recharge due to leakage changes were explored previously in the discussion of sensitivity analysis.

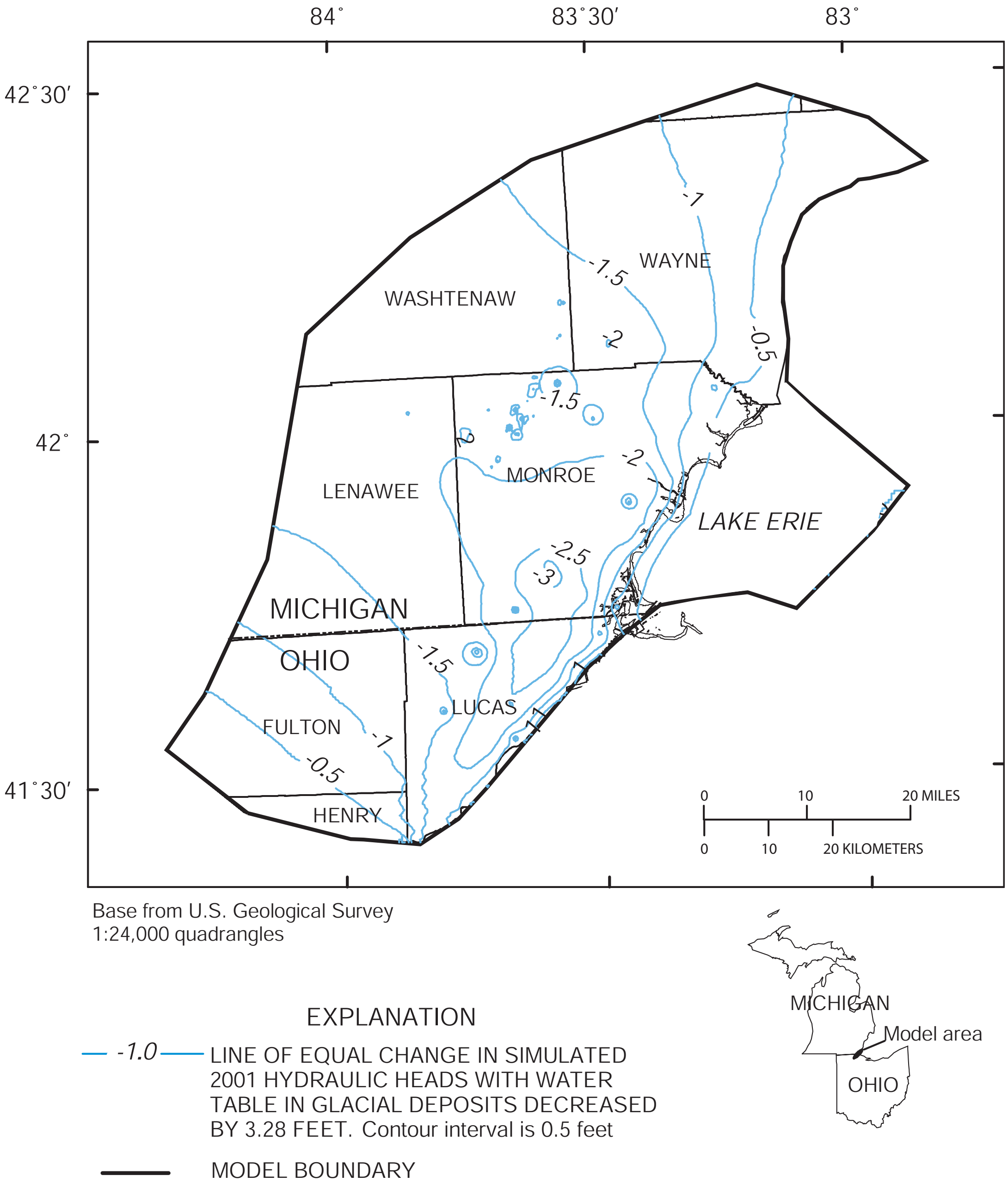


Figure 31. Simulated effect on 2001 hydraulic heads because of a 3.28-foot lowering of the water table in the glacial deposits across the study area (Monroe County, Michigan and surrounding area).

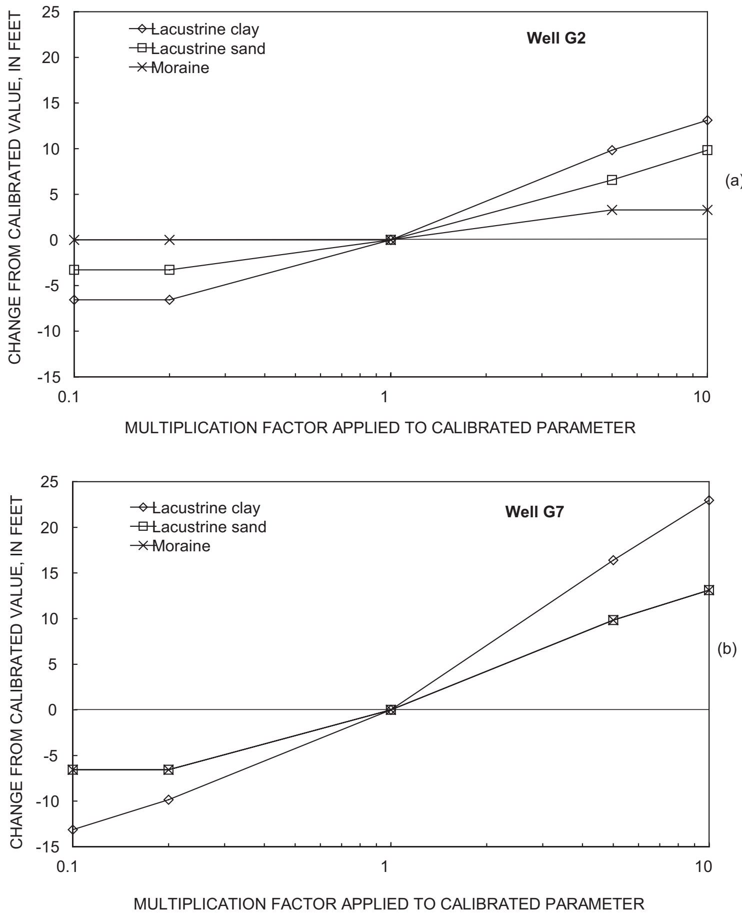


Figure 32. Change in hydraulic head for 2001 simulation at (a) G2, (b) G7, from varying model parameters.

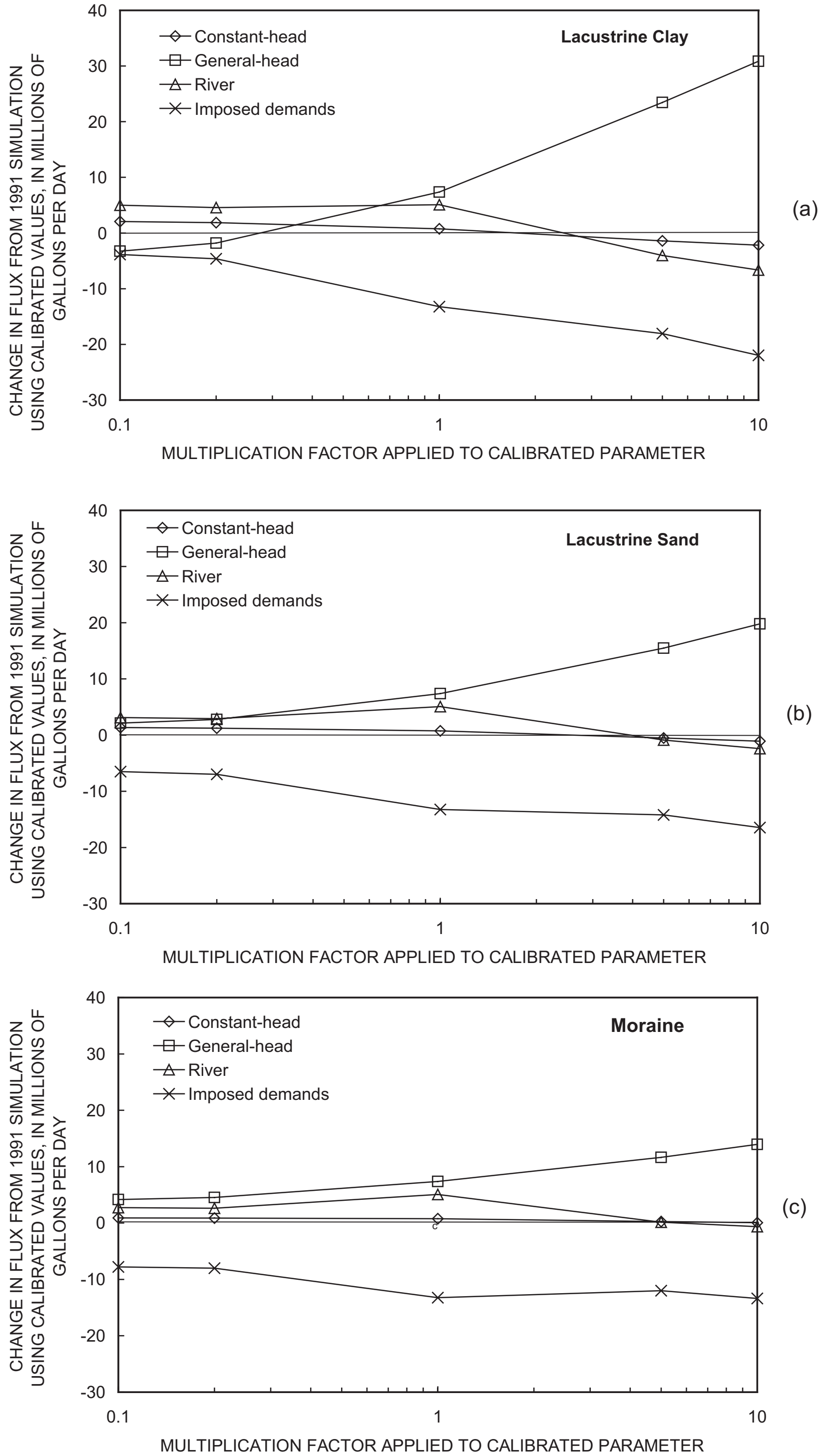


Figure 33. Change in flux from 2001 calibrated simulations because of changes in leakage from the glacial deposits in the study area for the (a) Lacustrine Clay, (b) Lacustrine Sand and (c) Moraine.

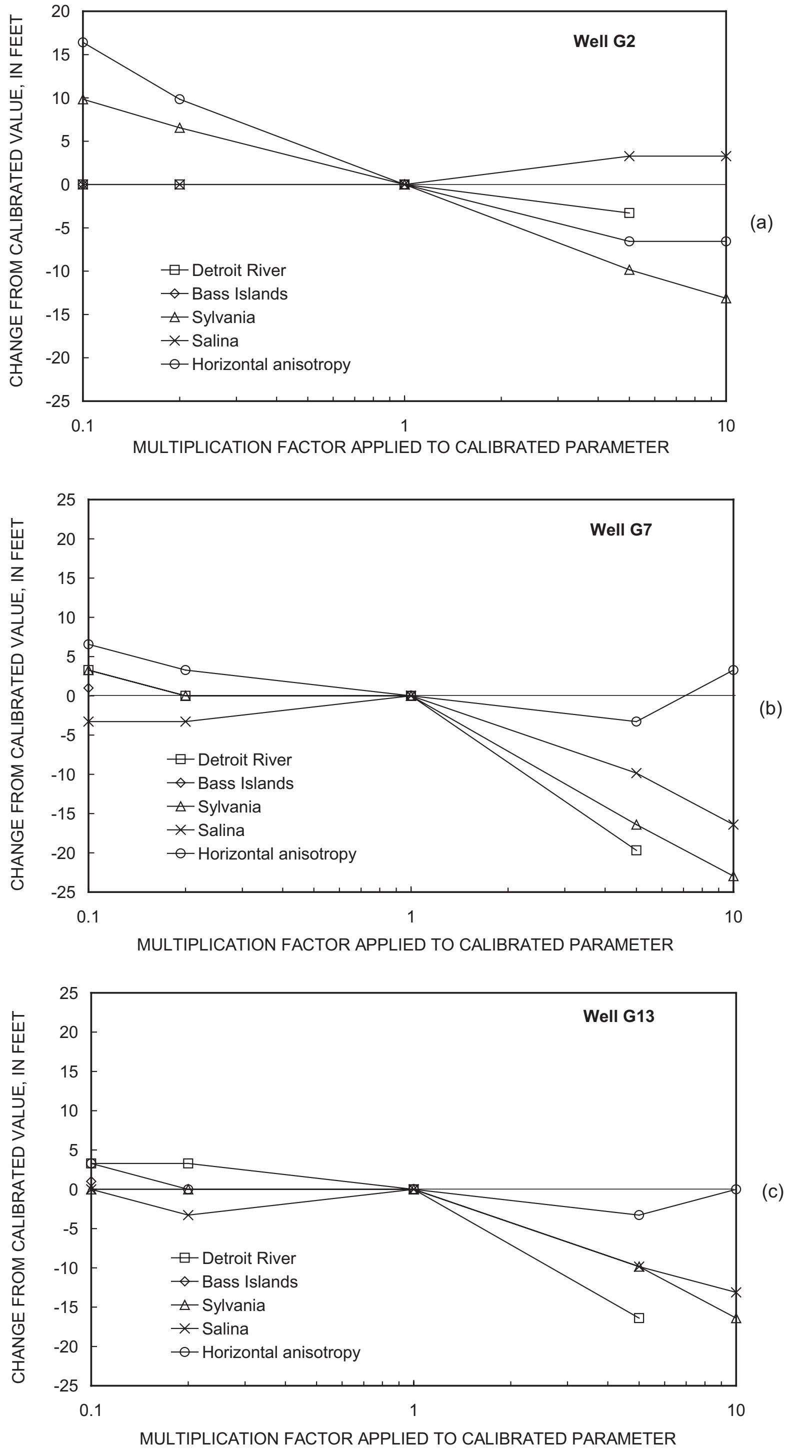
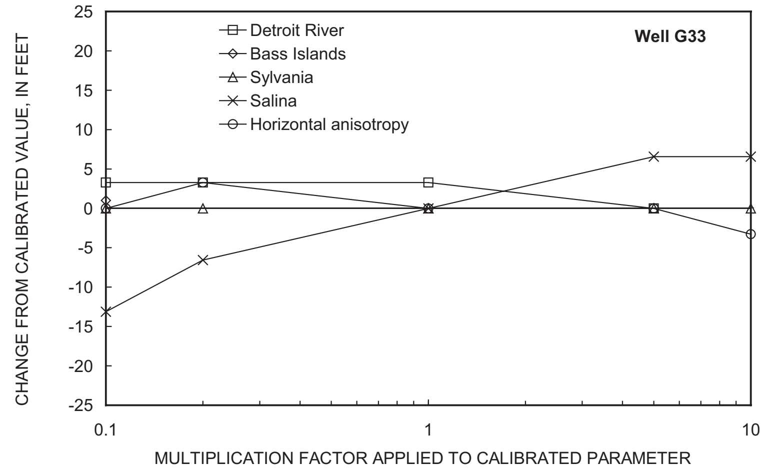
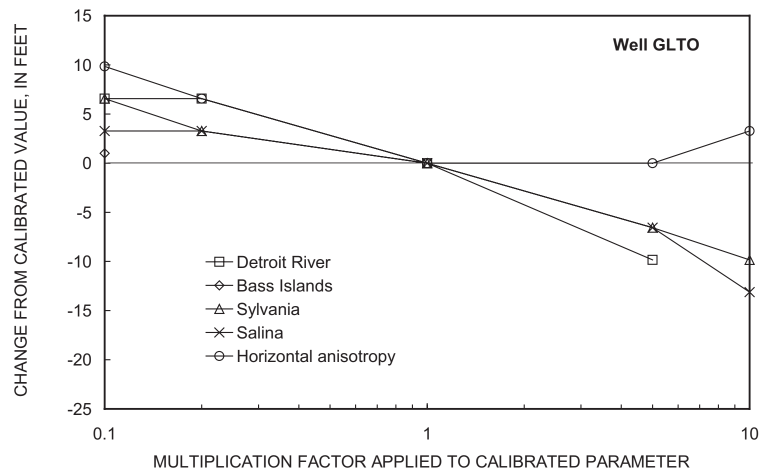


Figure 34. Change in hydraulic head for 1991 simulation at (a) G2, (b) G7, (c) G13 from varying model parameters.



(d)



(e)

Figure 34a. Change in hydraulic head for 1991 simulation at (d) G33, (e) GLTO, from varying model parameters.

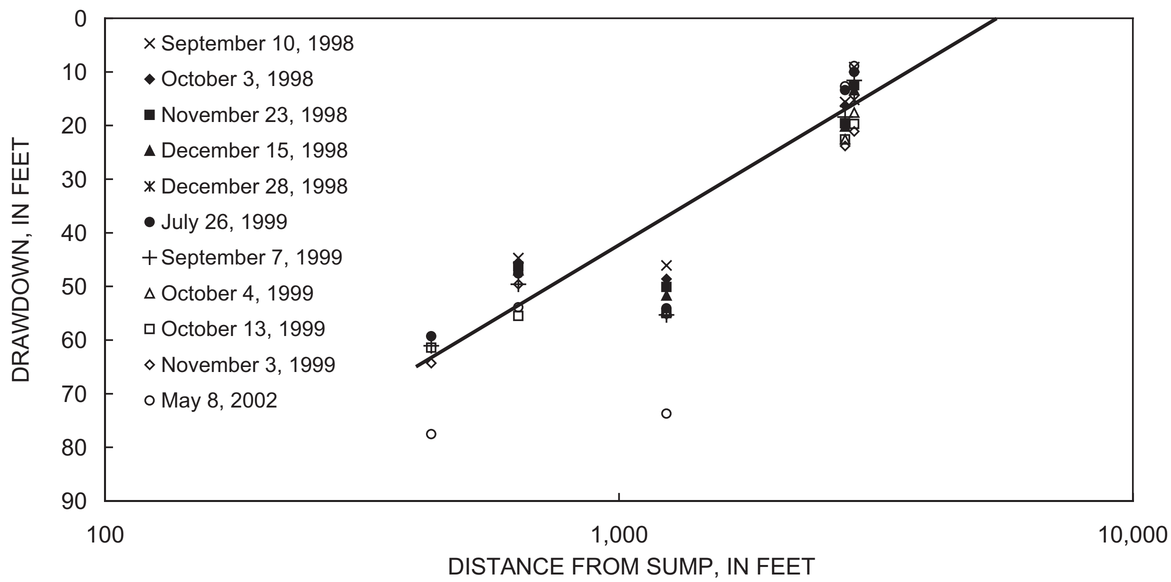


Figure 35. Drawdown near Hanson Quarry and radius of influence line.

The simulation results indicate that the effect of Lake Erie levels on a regional decline in ground-water levels appears to be small. Low rainfall or drought may affect the ground-water system by lowering the water-table elevation in the glacial deposits. The effect of this lowered water table, however, should result in greater changes in the water levels in south-central Monroe County. Neither the USGS monitoring well data or the analysis of the WelLogic ground-water-level information support a lowered water table in the glacial deposits as the mechanism leading to widespread bedrock aquifer water-level declines. In addition, geochemical analysis of the system indicates that the ground water in the bedrock aquifer is not recharged to a great extent by recent rainfall indicating that short-term changes in rainfall would not produce the observed declines.

Simulation results indicate that the effect of reducing the hydraulic conductance of glacial deposits and thereby reducing the leakage to the bedrock aquifer is greater than the effect of changing Lake Erie level or the decreased water-table elevation in the glacial deposits. A decrease in the hydraulic conductance of glacial deposits may be a result of land-use change. This effect, however, is smaller than the observed regional declines and may be ascertained through the sensitivity results that examined the conductances of the general-head boundary. An order of magnitude reduction in the capacity of the glacial deposits designated as moraines to transmit water to the bedrock aquifer decreased the general-head boundary flux by 5 Mgal/d or less and resulted in a change in water level of 6 ft at monitoring well G-7 and 3 ft at G-2 (fig. 32). The decline at GLTO also was 3 ft. Land-use change as a contributing factor to the observed declines cannot be discounted by the simulation results; however, the simulation results do not support this factor as the sole cause for the observed ground-water declines.

Simulation results indicate that the effect of increased ground-water withdrawals on a regional decline in ground-water levels is significant. The area of declines is the northern portion of Monroe County based on the WelLogic analysis and USGS monitoring wells (fig. 13b). Based on the simulation results, the increase in total ground-water use can explain most of the observed ground-water-level decline. As discussed previously, the majority of the ground-water use in the region is for quarry dewatering. The numerical model can be used to illustrate the relative effects of increased quarry dewatering, irrigation, and self-supplied domestic, municipal, and industrial use.

The relative effect of increased self-supplied domestic, municipal, and industrial use, and agricultural and golf-course irrigation can be illustrated by considering the results of a simulation where the simulated quarries from 1991 are used with the remaining conditions for the 2001 simulation. The maximum decline in simulated water level at a USGS monitoring well was 3.2 ft at G-29. The average simulated drawdown for all of the USGS monitoring wells from the 1991 simulation was 1.6 ft. The increased withdrawals for self-supplied domestic, municipal, and industrial use, agricultural irrigation, and golf-course irrigation contribute to the regional decline, but do not explain much of it. Note that increased pumping for any of these uses potentially can have a large site-specific effect. For example, the continuous record of water level for monitoring well G-3 (fig. A3) indicates local pumping. The measured water level decreases rapidly and then recovers more slowly to its previous elevation consistent with a nearby well intermittently pumping from the bedrock aquifer.

The effect of quarry dewatering can be considered through two analyses. The first is classic dewatering analysis. Simulation results (fig. 29, for example), clearly show localized drawdown in the vicinity of active quarries. These "cones of depression" are superimposed on the regional hydraulic surface. To illustrate the cone of depression from an active quarry in more detail, monitoring-well data from USGS well G-17 and observation wells operated by Hanson Aggregates (J. Stoll, Hanson Aggregates, written commun., 2002) are shown in a distance-drawdown curve (fig. 35). From this analysis, the hypothetical "radius of influence" for a well is estimated by extending a straight-line through the data until it intersects the axis of zero drawdown. Implicit in this analysis is that the drawdowns plotted on the figure all result only from the dewatering at the facility. This assumption may be violated for the wells furthest from the quarry as they may show regional declines. Violation of the assumption, however, will increase the radius of influence for the quarry and does not change the conclusion of this evaluation. For this quarry, the distance at which little impact on nearby water levels is expected is on the order of 1 mi. This analysis, however, does not account for regional effects or assess the capture of the ground-water system.

On the regional scale, the relation between the radius of influence and regional effect must be evaluated. Alley and Schefter (1987) and Alley and others (1999) discuss the "egg carton" and "bathtub" conceptual models to frame the discussion of water-level declines and ground-water management for the High Plains aquifer. These conceptual models are useful analogies for Monroe County. The "bathtub" model proposes that any withdrawal from the ground-water system lowers the water level for the entire system instantaneously. Thus, any pumping in the aquifer is immediately noticed throughout the ground-water system. Resistance to ground-water flow, however, makes this an inappropriate model for water management. Pumping ground water can induce local drawdowns such as the cone of depression shown in figure 29, and ground-water withdrawals do not instantaneously and uniformly lower regional water levels from the point of the withdrawal to the hydraulic boundaries of the system. The inadequacy of the "bathtub" model leads to the conceptual model of an "egg carton", where individuals are not seriously effected by nearby pumps because of the local nature of the cone of depression. Alley and Schefter (1987) demonstrate that the "egg carton" model also is not appropriate for water management for the High Plains aquifer. The cumulative effects of many wells interact to generate regional drawdowns. The actual situation lies between these two extreme analogies: pumping leads to both localized drawdown and contributes to regional effects (Alley and others, 1999). The situation in Monroe County is similar to the High Plains aquifer analysis in that large-scale withdrawals at quarries cause both local cones of depression and contribute to the regional decline.

Dewatering at individual quarries creates small cones of depression, approximately 1 mi in radius, but the withdrawals are part of the overall demand causing the observed regional decline. The depth of dewatering is controlled by the elevation of the quarry sumps in the vicinity of the quarries and by proximity to sources of ground-water capture in the remainder of the area. Simulation results and sensitivity analysis suggest that if regional demands were held to 2001 levels, water levels would stabilize near the observed elevation. Note that London Aggregates in London Township, Monroe County stopped pumping in December 2003 under a consent agreement with the U.S. Environmental Protection Agency (PIRGIM v. Wolcottville, 2002; United States v. Wolcottville, 2002). At the time of the shutdown, this quarry was the largest single ground-water user in Monroe County based on reported discharges from 8 to 10 Mgal/d. The observed water level at monitoring well G-7 recovered approximately 10 ft in September 2003. The model underestimated the withdrawal from this quarry, but simulation without pumping at London Aggregates shows some regional recovery under steady-state conditions (fig. 36). This simulation again illustrates both the local nature of the cone of depression caused by a large withdrawal and the regional effect of large withdrawals on the system. The greatest recovery shown in the figure is at the quarry as expected, and much of the recovery occurs within a mile of the quarry. This is the recovery of the short-range cone of depression. More regional recovery also is predicted, however, for steady-state conditions. Recovery between 1 and 6 ft is predicted outside the cone-of-depression, and this recovery reflects the regional effect of quarry dewatering.

The predicted recovery if quarry withdrawals are removed and the remaining 2001 stresses applied to the model is illustrated in figure 37. The recovery contours show more than 10 ft of recovery in water levels for most of the model area. Recoveries greater than 30 ft are simulated in parts of Monroe County. In the simulations of water-level recovery, the hydraulic properties of the bedrock aquifer are assumed to have not changed because of dewatering conditions. Because without quarries the withdrawals are less than the calibrated 1991 values, the model simulates water levels above 1991 levels in parts of the model area. As noted above, however, the ground-water-capture analysis must consider all ground-water use. If other uses increase, the ground-water level will decline again, until sufficient capture is produced by the ground-water system to balance the new demands.

Model Limitations

There are limitations to the model and analysis presented herein. As discussed previously, despite the widespread use of analysis techniques based on flow through porous media for fractured rock systems, local flow conditions in the ground-water system will be affected by fractures and conditions may be locally different than the averaged condition simulated with the regional model. In the same way, local heterogeneities in hydrogeologic properties of the bedrock deposits and in the bedrock/glacial interface could not be considered in the regional model. Finally, karst features in southwestern Monroe County are not simulated by the model. The effect of karst features on the regional system was not assessed other than to note that the model produces results consistent with observed values in the remainder of Monroe County.

The simulated ground-water capture depends on the ground-water demands placed on the system. Errors in water-use estimates will affect the results and major changes in water-use estimates would necessitate recalibration of the model.

The configuration of the moraine/bedrock aquifer boundary is not well known. Large-scale maps were used to create the basemap for the model, and the area where moraine or more conductive deposits overlie the Dundee Formation to the west and northwest of Monroe County could be different than used in the simulation. The effect of a smaller recharge area to the regional system was illustrated with the sensitivity analysis by decreasing the conductance of the moraine general-head boundary parameter. A larger recharge area could provide more water to help the system achieve steady-state. The larger recharge area may be less sensitive to land-use changes, because these changes affect the local scale and the influence would be buffered by a larger unchanged area.

Leakage between the glacial deposits and bedrock aquifer is an important factor in the regional water budget. The model simplifies the system by treating the glacial deposits as a boundary condition to the bedrock aquifer. The flux also is potentially maximized by the use of the general-head boundary that increases the flux as water levels decline in

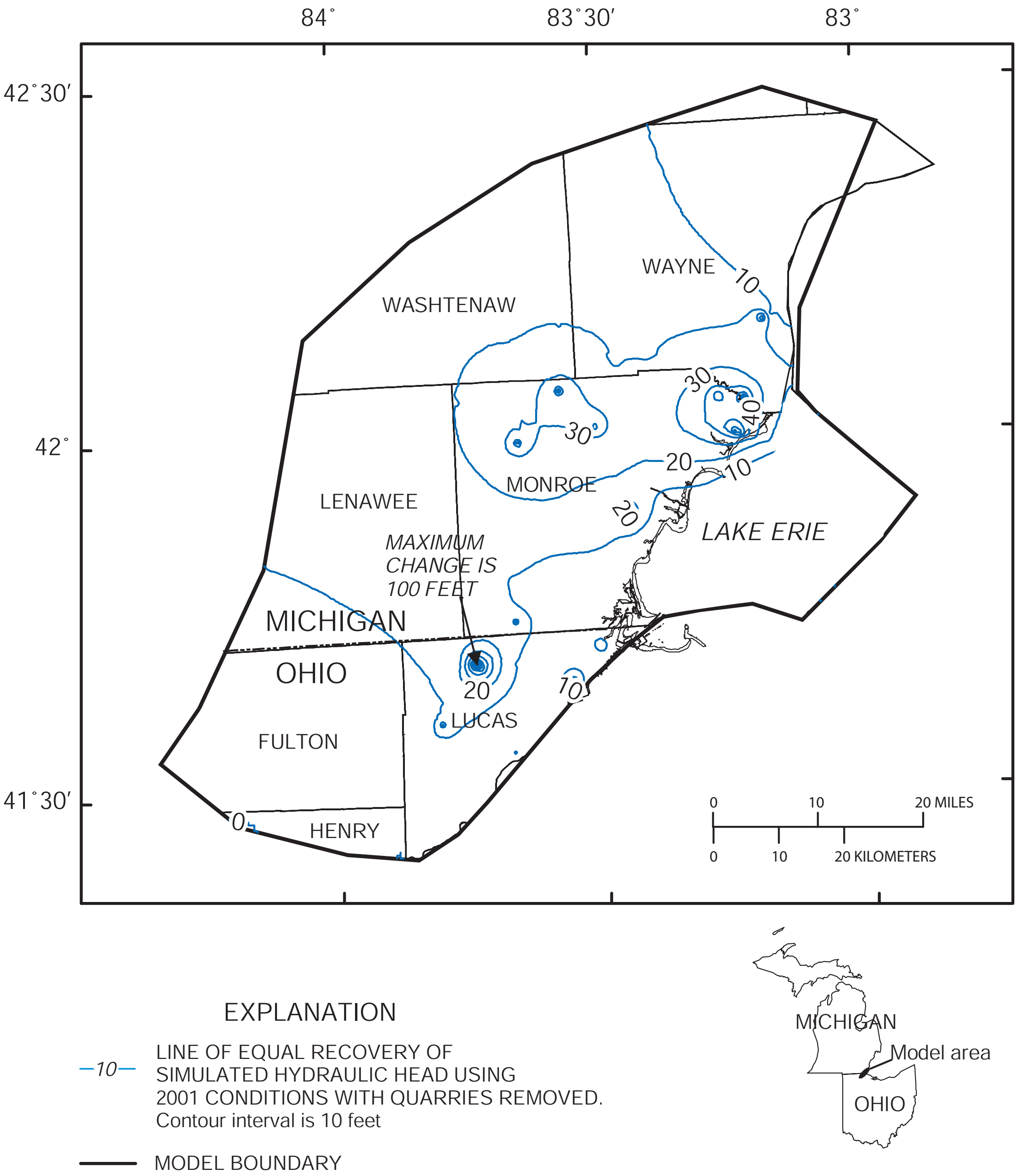


Figure 36. Simulated recovery from 2001 conditions if quarry discharges are removed for the study area (Monroe County, Michigan and surrounding area).

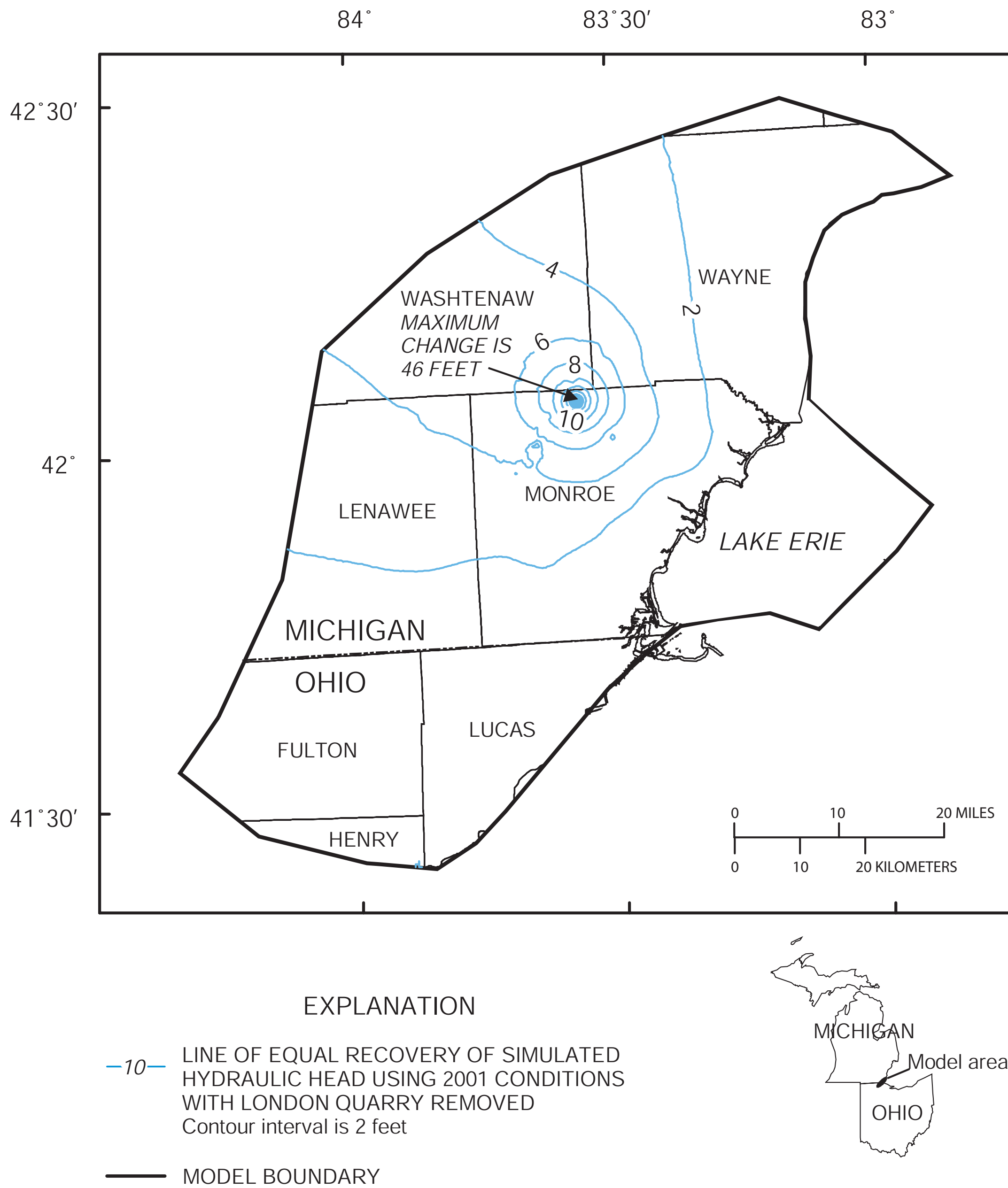


Figure 37. Simulated recovery for study area (Monroe County, Michigan and surrounding area) under 2001 steady-state conditions if London Quarry discharge is removed.

the bedrock aquifer regardless of whether this unit is locally confined or unconfined. A more rigorous approach would be to explicitly simulate water flow in the glacial deposits. This approach also could include a transient analysis so time-varying conditions in both the bedrock aquifer and glacial deposits could be considered.

A consistent model was sought by using predevelopment and 1991 conditions to constrain model performance. An improved simulation may result if 1991 to 2003 information is used to develop a transient model. This simulation could help estimate the time to reach a new steady-state condition given changes in the imposed demands. This simulation may not improve the ability to identify the causes of ground-water declines beyond the steady-state models, but it would provide additional insight into aquifer properties and could produce a more well-calibrated model.

SUMMARY AND CONCLUSIONS

The U.S. Geological Survey (USGS), in cooperation with the Michigan Department of Environmental Quality (MDEQ), began a study in 2002 to study the hydrogeology and identify the cause or causes of observed widespread ground-water-level declines in the bedrock aquifer in Monroe County, Michigan. The study area focuses on Monroe County, but includes portions of neighboring counties in Michigan and Ohio. Previous reports provide necessary background regarding water use, water chemistry, and hydrogeology. The present report analyzes the hydrologic data in more detail than previous work and uses a numerical ground-water-flow model to investigate the causes of ground-water-level declines.

The main focus of the study is the bedrock aquifer composed of five Silurian-Devonian bedrock units. These rocks are fractured, exhibit karst features in some areas, and are locally heterogeneous with respect to hydrogeologic properties. The overlying glacial deposits are important in that they control leakage to the bedrock aquifer, and leakage is important to the response of the bedrock ground-water system to changes in pumping. In addition to the 32 bedrock monitoring wells and 1 glacial monitoring well installed for an earlier study, the MDEQ WelLogic database of water-well logs was used extensively in this study to both estimate hydrogeologic properties of the glacial deposits and to document water-level changes. The USGS monitoring wells were monitored from 1991 through 2003 and provide documentation of the water-level declines in the bedrock aquifer across the county.

Ground-water levels declined in many of the monitoring wells from 1991 to 2001. Water levels in 11 monitoring wells declined more than 10 ft, and the water level in 6 wells declined more than 20 ft. The general downward trend of water levels since 1991 is evident for all but 2 wells, one of which is completed in glacial deposits. Water levels in 10 wells, all in northern Monroe County, increased in 2003, and some of this increase can be attributed to seasonal fluctuations in water levels.

Estimated ground-water use in Monroe County increased from approximately 20 Mgal/d in 1991 to nearly 30 Mgal/d in 2001. The percentage of the total used for public supply, self supplied, industrial, irrigation, and quarry dewatering in 1991 and 2001 is about the same. The major ground-water use is attributed to quarry dewatering. In 1991, 80 percent of the total was from quarry dewatering. In 2001, 75 percent of the total was from quarry dewatering.

The conceptual model of the study area uses an equivalent porous media approach to simulate ground-water flow in the bedrock aquifer. The bedrock aquifer is simulated as 10 layers—2 layers for each geologic unit. Glacial deposits are not simulated explicitly. Rather, these units are simulated as an upper boundary condition to the bedrock aquifer. The bedrock aquifer can both receive water from the overlying glacial deposits and transmit water to the glacial deposits. Shale units overlying the bedrock aquifer in the west and northwestern portions of the study area are considered to be no-flow upper boundaries for the model. The lower boundary condition for the bedrock aquifer also is a no-flow boundary. Lateral boundaries were no-flow across the Detroit and Maumee Rivers and the western boundary of the study area, where the ground water in the bedrock is saline. Flow to the Maumee and Detroit Rivers, however, is simulated in the model. The remaining lateral boundaries are constant-head boundaries in Lake Erie and on the southwestern model boundary. Discharge from the bedrock aquifer includes leakage to the overlying glacial deposits. In addition, quarry dewatering; self-supplied domestic, municipal, and industrial water use; and irrigation water use were simulated in the model.

Simulation results indicate that regional declines in ground-water levels are caused by increased ground-water demands. Results indicate that a simulated lower level of Lake Erie has little effect on regional ground-water levels. Similarly, results indicate that a simulated decreased leakage from the glacial deposits to the bedrock aquifer, consistent with land-use changes, does not explain the observed declines in regional ground-water levels in the bedrock aquifer.

The largest ground-water withdrawal in the region was quarry dewatering operations. Increases in other uses contribute to the observed decline, but cannot account for the total observed regional decline. Dewatering at individual quarries creates limited cones of depression, approximately 1 mi in radius, but the withdrawals are part of the overall regional demand causing the observed decline. The depth of dewatering is controlled by the elevation of the quarry sumps in the vicinity of the quarries and by proximity to sources of ground-water capture in the remainder of the area.

Application of model results is limited by the assumptions inherent in numerical models of ground-water flow. The assumption of porous media flow in the model limits its application in local areas where heterogeneities in hydrogeologic properties dominate the response of ground-water levels to stresses. Leakage between glacial deposits and the underlying bedrock aquifer is an important factor in the model simulations and is simplified in the numerical ground-water-flow model. Whereas regional patterns of leakage are sufficiently accurate for simulation purposes, direct measurements of leakage and glacial deposit conductances are necessary to calculate leakage in local areas.

Ground-water capture controls the magnitude of water-level declines when new demands are introduced to the ground-water system. Simulation results were used to illustrate ground-water capture for the regional model. Prior to 1991, ground-water demands appear to have been balanced by interception of natural discharge from the bedrock aquifer to the glacial deposits in eastern Monroe County. At some time, demands in the region exceeded the amount of water that could be readily captured from natural discharge, and ground-water levels declined. Observations from 1991 until 2003 document this regional decline. Simulation results indicate, if demands were held constant to 2001 levels, the decline would stop when the combination of decreased discharge and additional recharge balances the demand. Limits to future ground-water withdrawals in the region should be assessed on a regional basis because analysis relying on local drawdown does not consider regional effects. Decisions regarding future ground-water development also must be made with recognition that dispersed withdrawals, while not creating large local drawdowns, still affect the regional ground-water system.

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Appendix A

Water-Level Data

Ground-water wells originally drilled in 1991 for the study done by Nicholas and others (1996) were monitored by the U.S. Geological Survey (USGS) through August 1993. Measurements were taken by Monroe County from this time until August 2001 when USGS began monitoring for the present study until September 2003. Eleven of the 33 wells were electronically monitored at 15-minute increments during this study, and 21 of the wells were manually measured at approximately 6-week intervals. One well, G31, was unavailable for measurement during this study. Well GLTO (figure 13) located in Petersburg State Game Area has a record dating back to 1978, and water levels from this well were used in the study.

U.S. Geological Survey Observation Wells

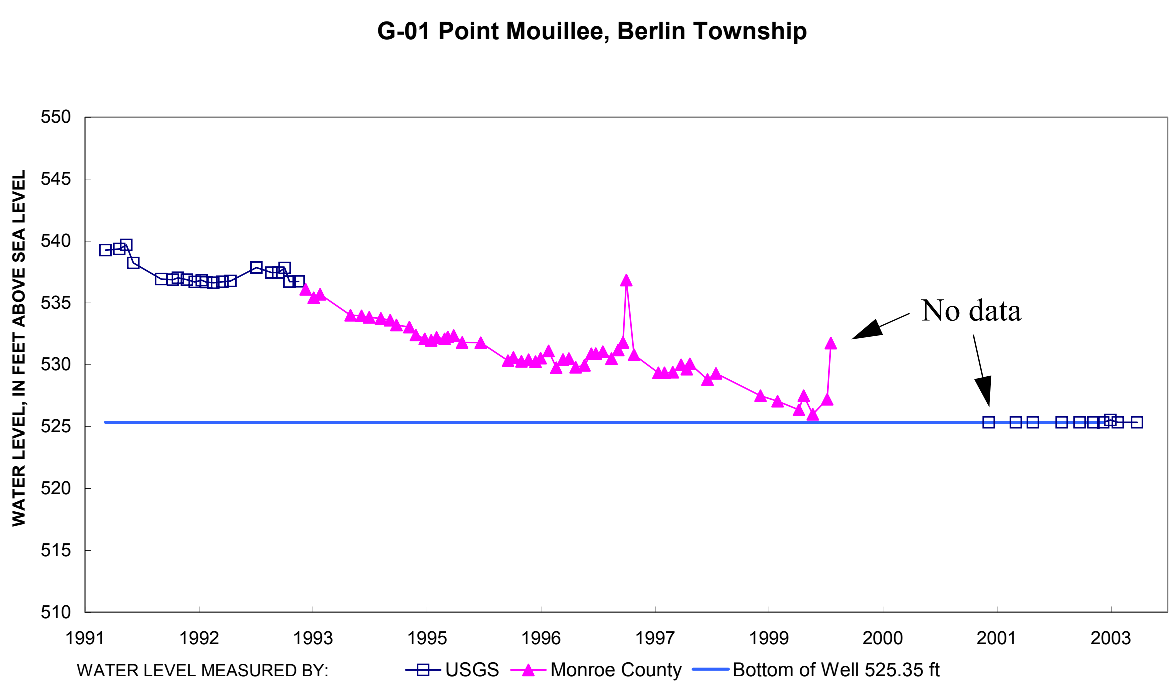

Figure A1. Ground-water levels for well G-1 Berlin Township, Monroe County, Michigan.

G-02 Berlin Township Park, Berlin Township

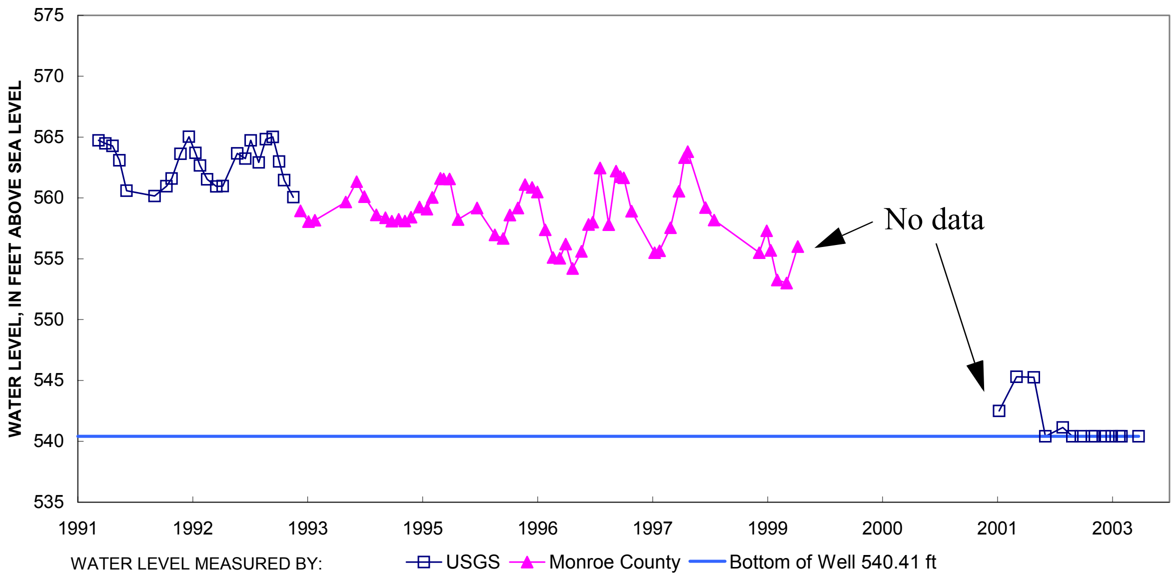

Figure A2. Ground-water levels for well G-2 Berlin Township, Monroe County, Michigan.

G-03 Evergreen Cemetery, Ash Township

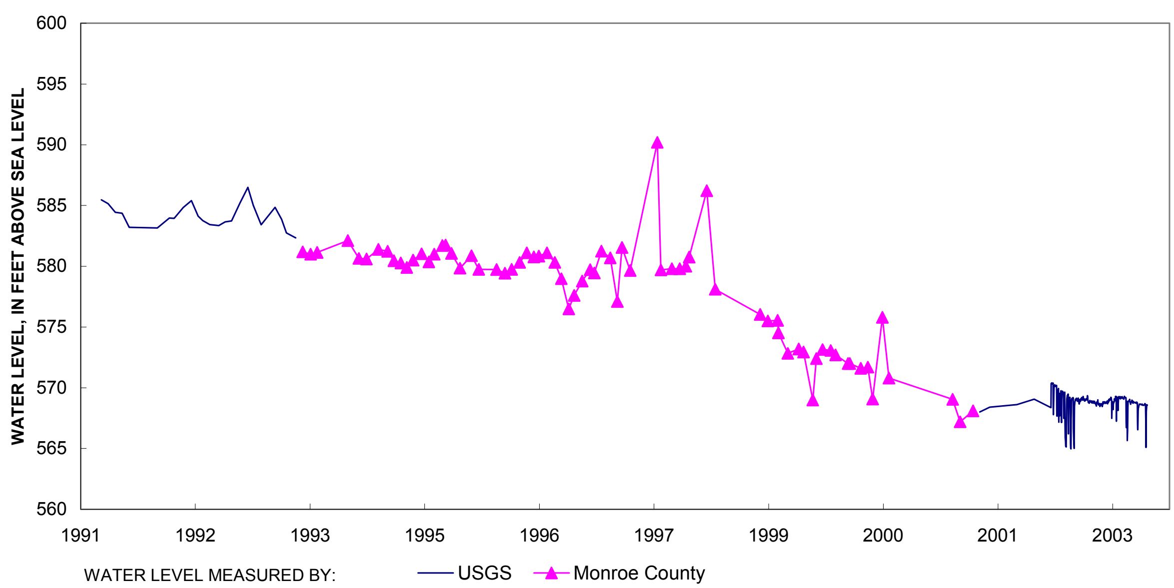

Figure A3. Ground-water levels for well G-3 Ash Township, Monroe County, Michigan.

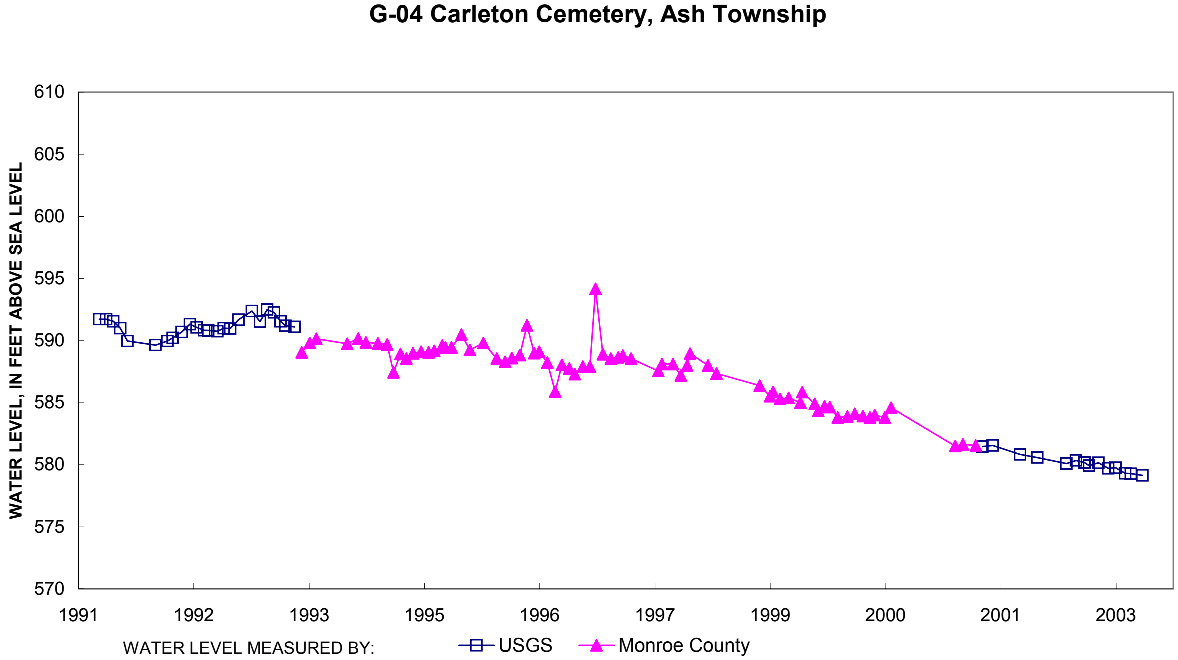

Figure A4. Ground-water levels for well G-4 Ash Township, Monroe County, Michigan.

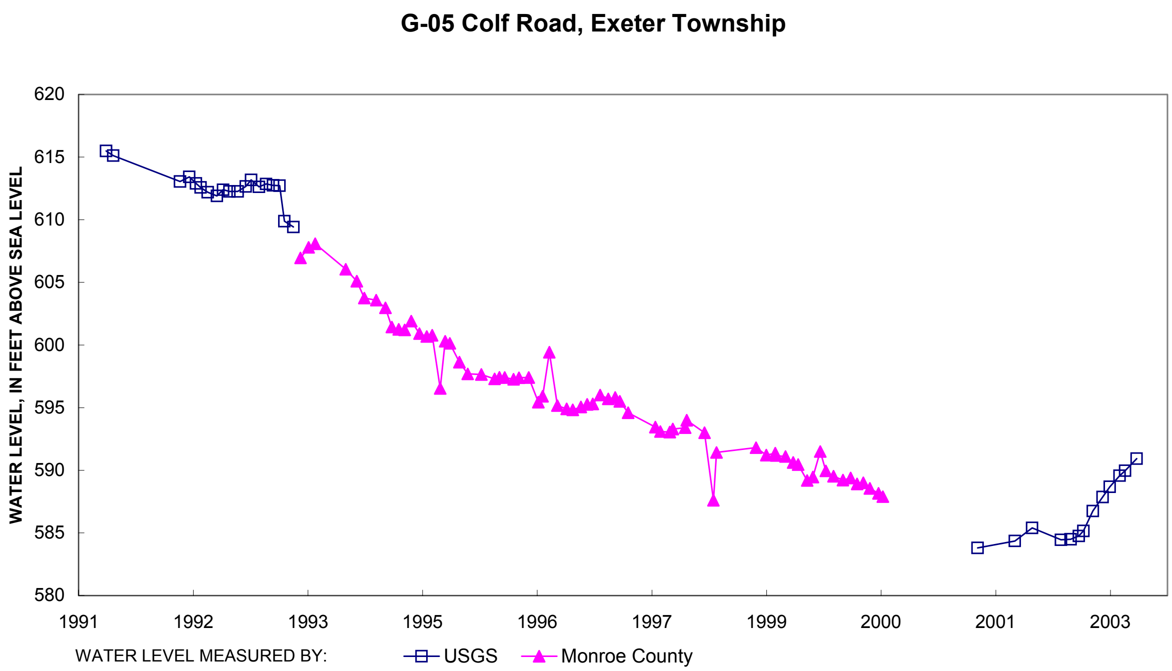

Figure A5. Ground-water levels for well G-5 Exeter Township, Monroe County, Michigan.

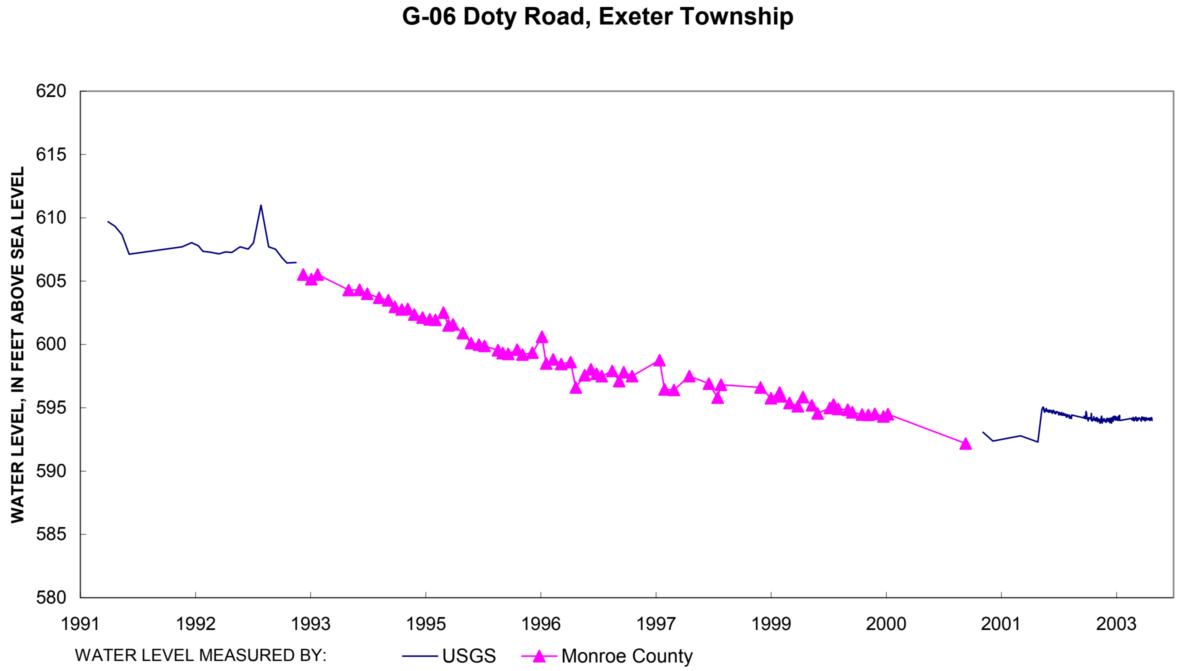

Figure A6. Ground-water levels for well G-6 Exeter Township, Monroe County, Michigan.

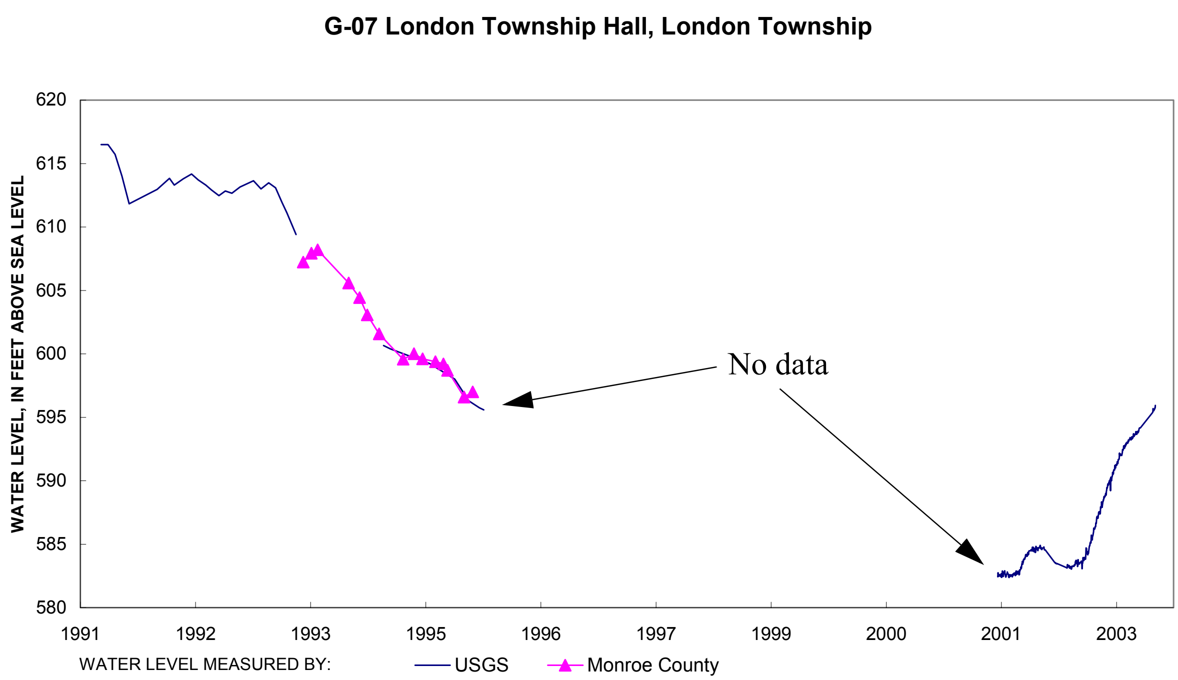

Figure A7. Ground-water levels for well G-7 London Township, Monroe County, Michigan.

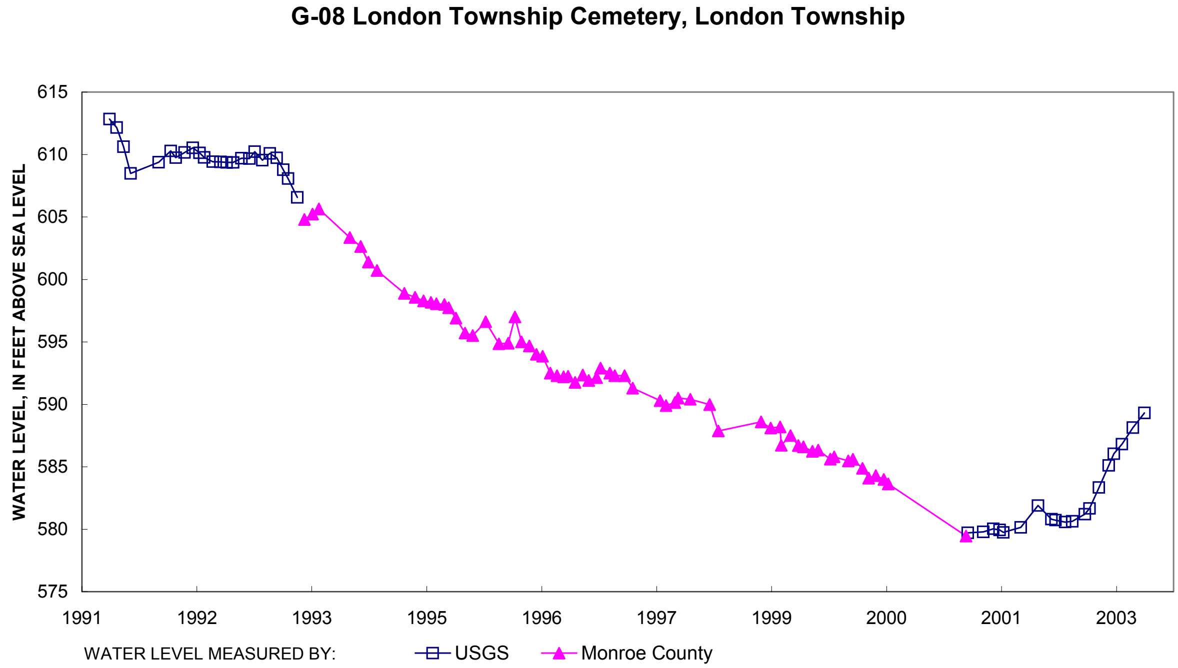

Figure A8. Ground-water levels for well G-8 London Township, Monroe County, Michigan.

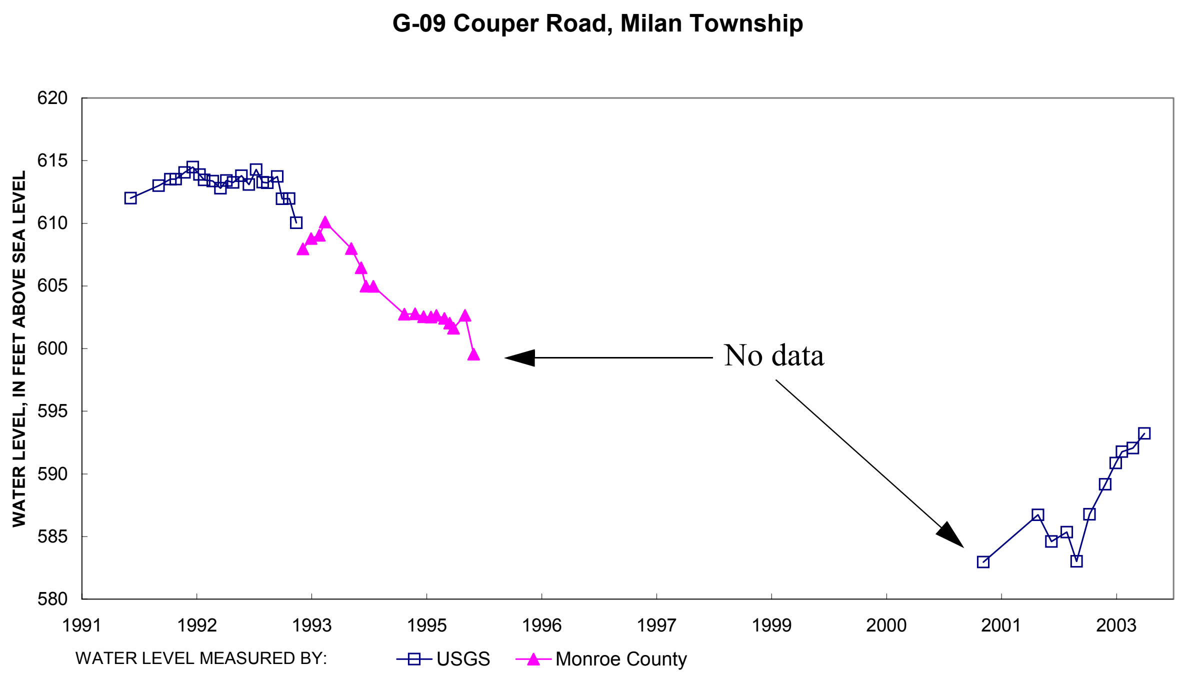

Figure A9. Ground-water levels for well G-9 Milan Township, Monroe County, Michigan.

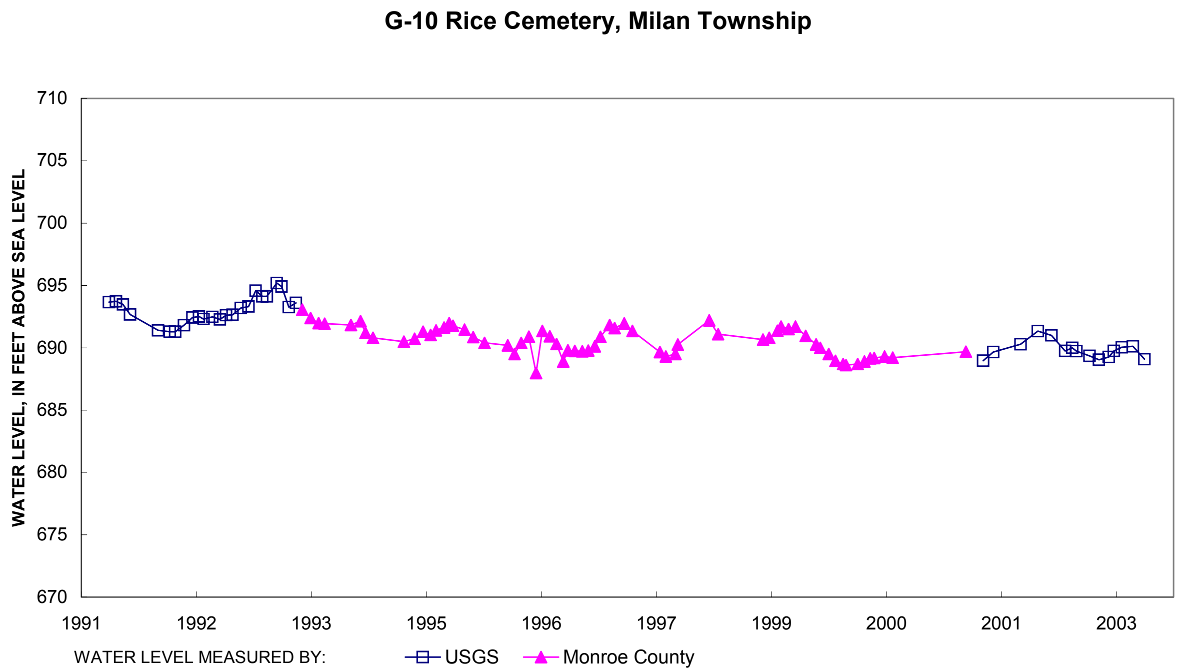

Figure A10. Ground-water levels for well G-10 Milan Township, Monroe County, Michigan.

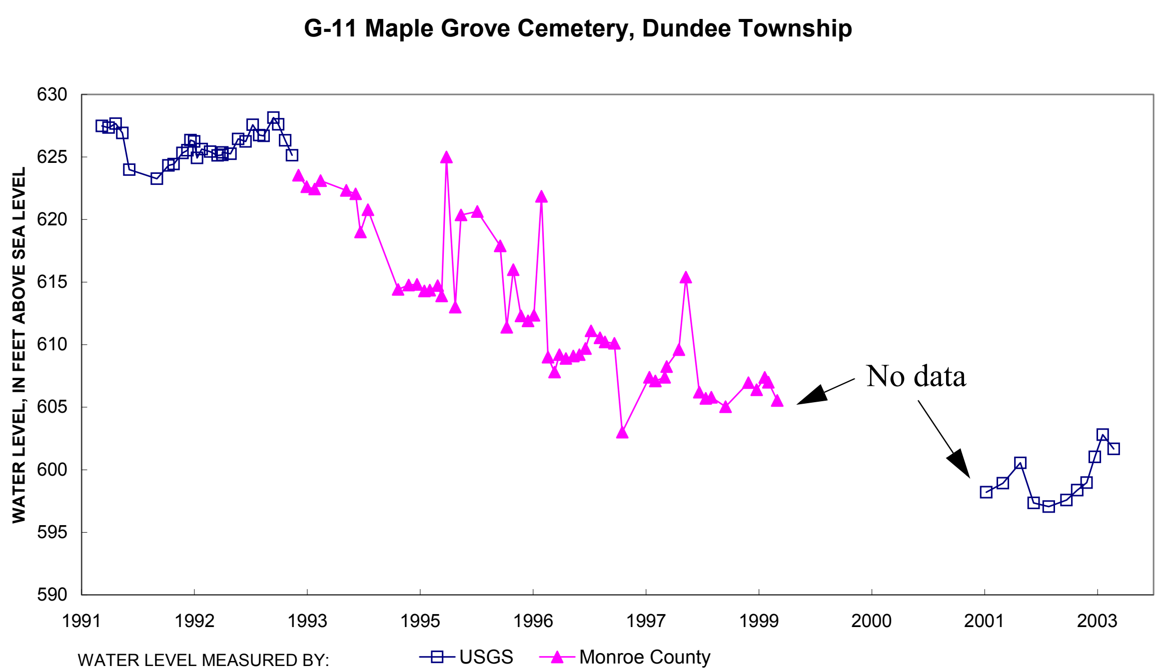

Figure A11. Ground-water levels for well G-11 Dundee Township, Monroe County, Michigan.

G-12 Oak Grove Cemetery, Dundee Township

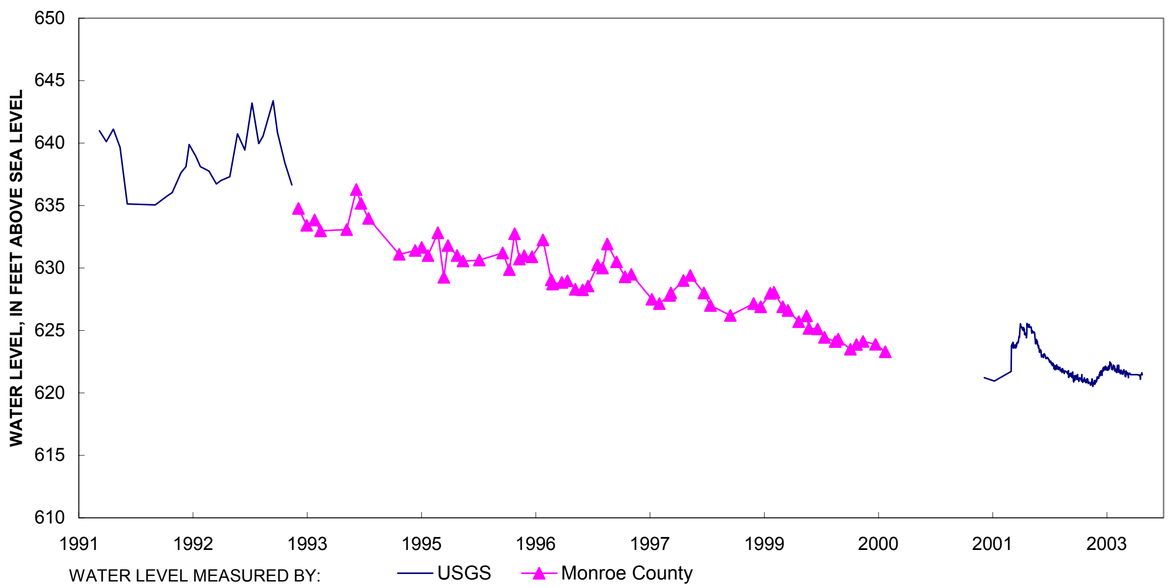

Figure A12. Ground-water levels for well G-12 Dundee Township, Monroe County, Michigan.

G-13 Rath Cemetery, Raisinville Township

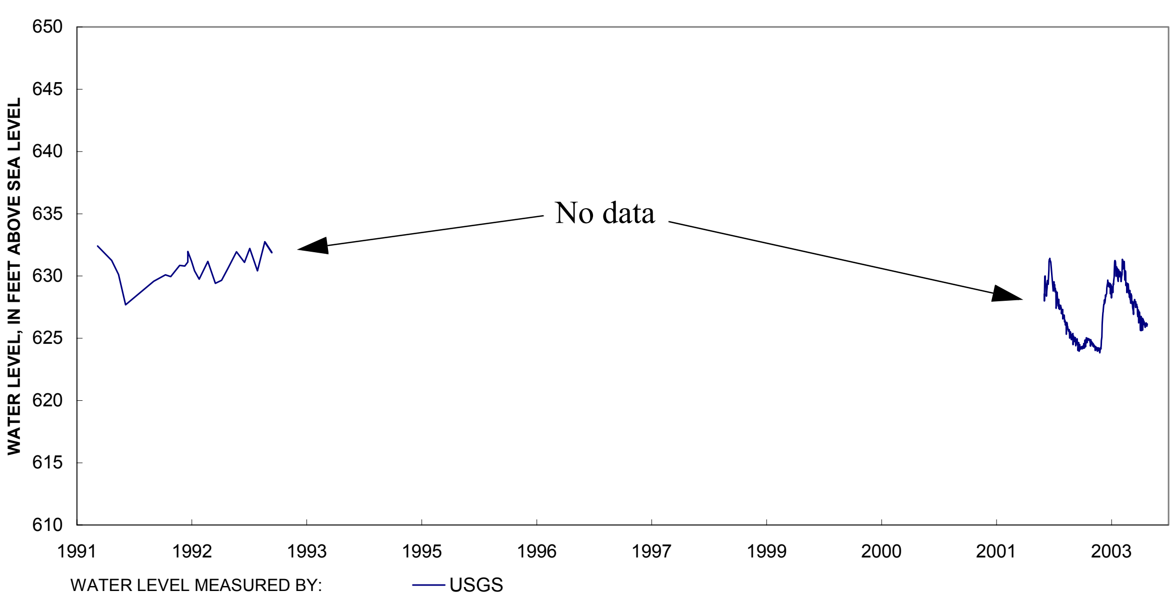

Figure A13. Ground-water levels for well G-13 Raisinville Township, Monroe County, Michigan.

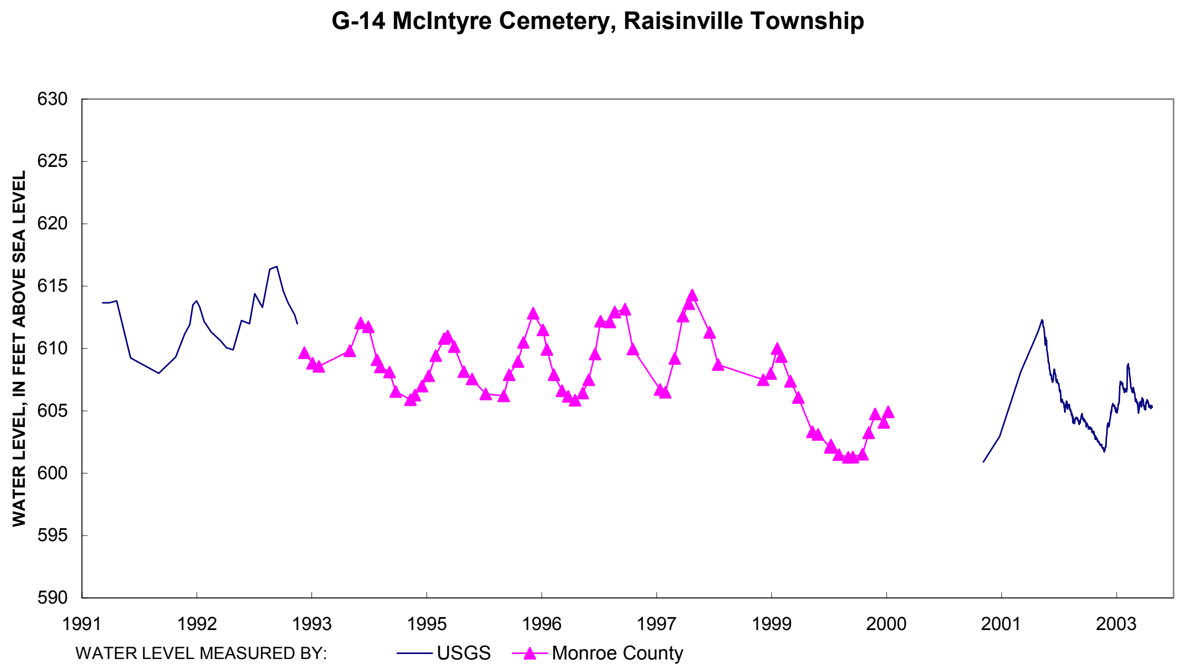

Figure A14. Ground-water levels for well G-14 Raisinville Township, Monroe County, Michigan.

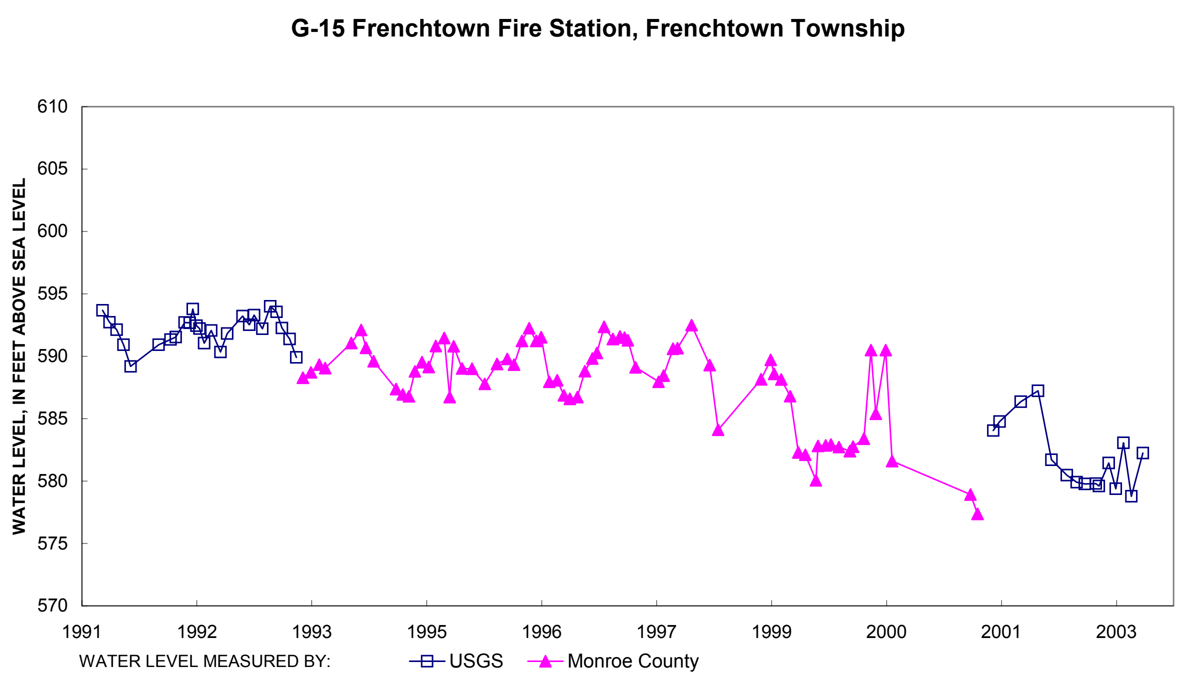

Figure A15. Ground-water levels for well G-15 Frenchtown Township, Monroe County, Michigan.

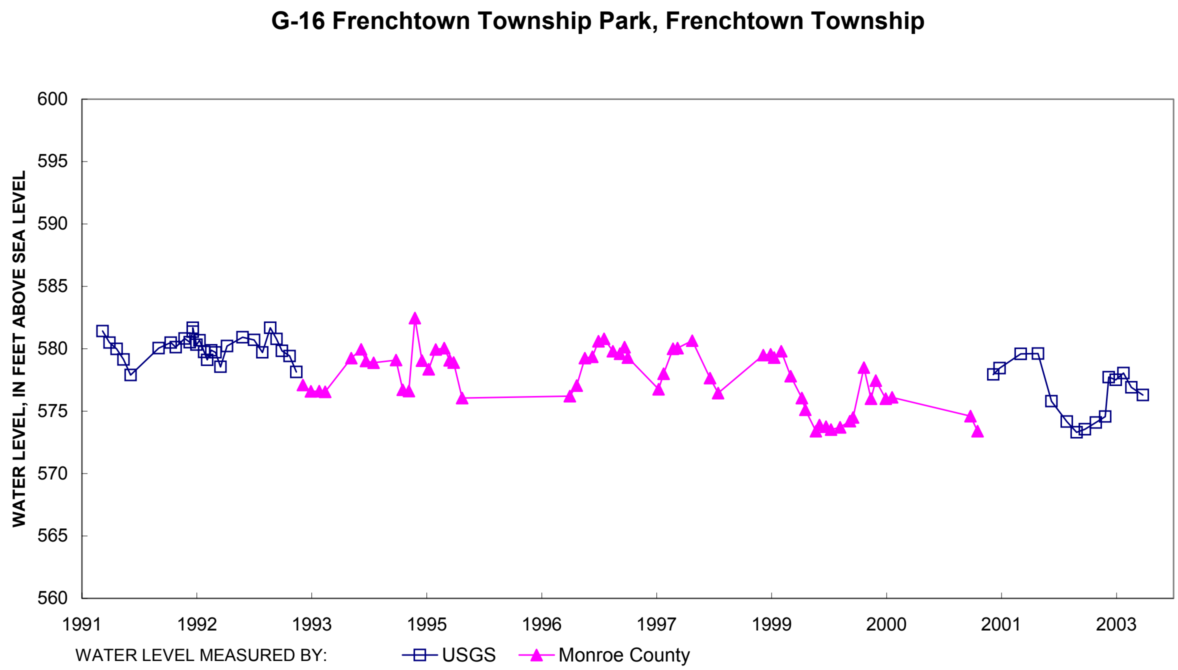

Figure A16. Ground-water levels for well G-16 Frenchtown Township, Monroe County, Michigan.

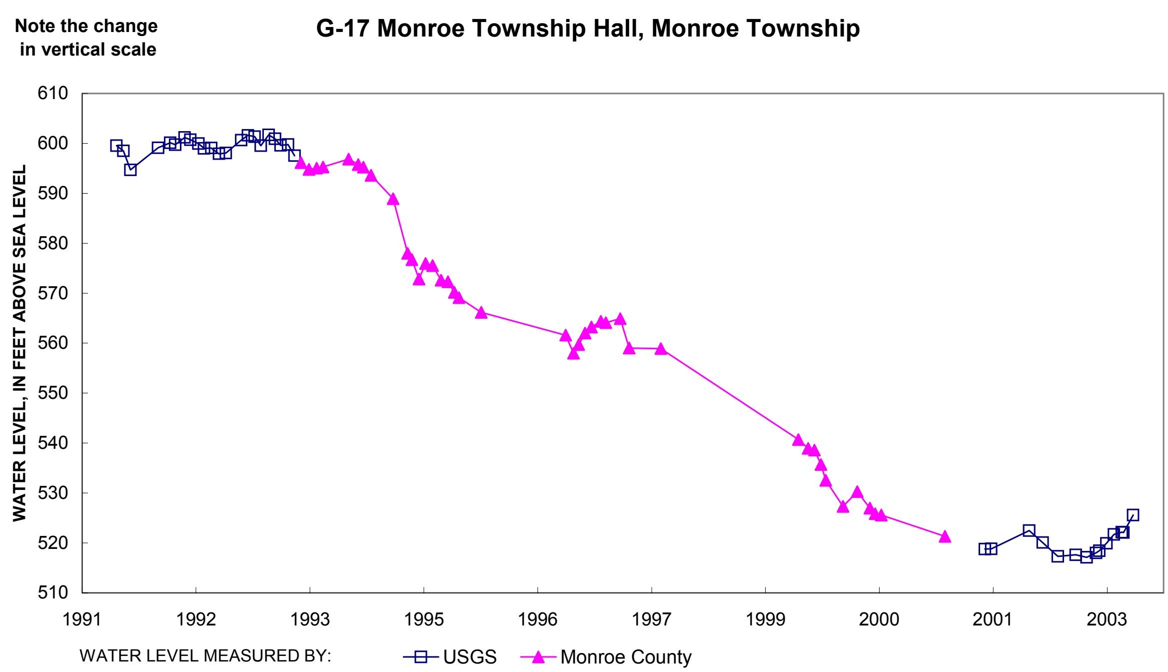

Figure A17. Ground-water levels for well G-17 Monroe Township, Monroe County, Michigan.

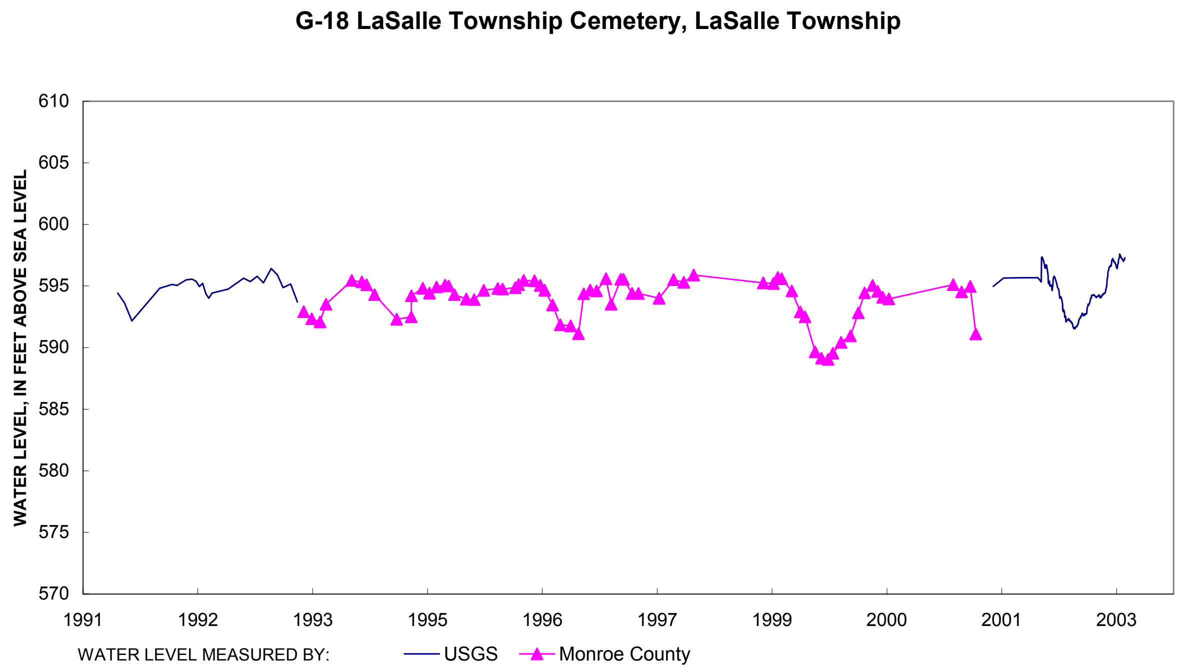

Figure A18. Ground-water levels for well G-18 LaSalle Township, Monroe County, Michigan.

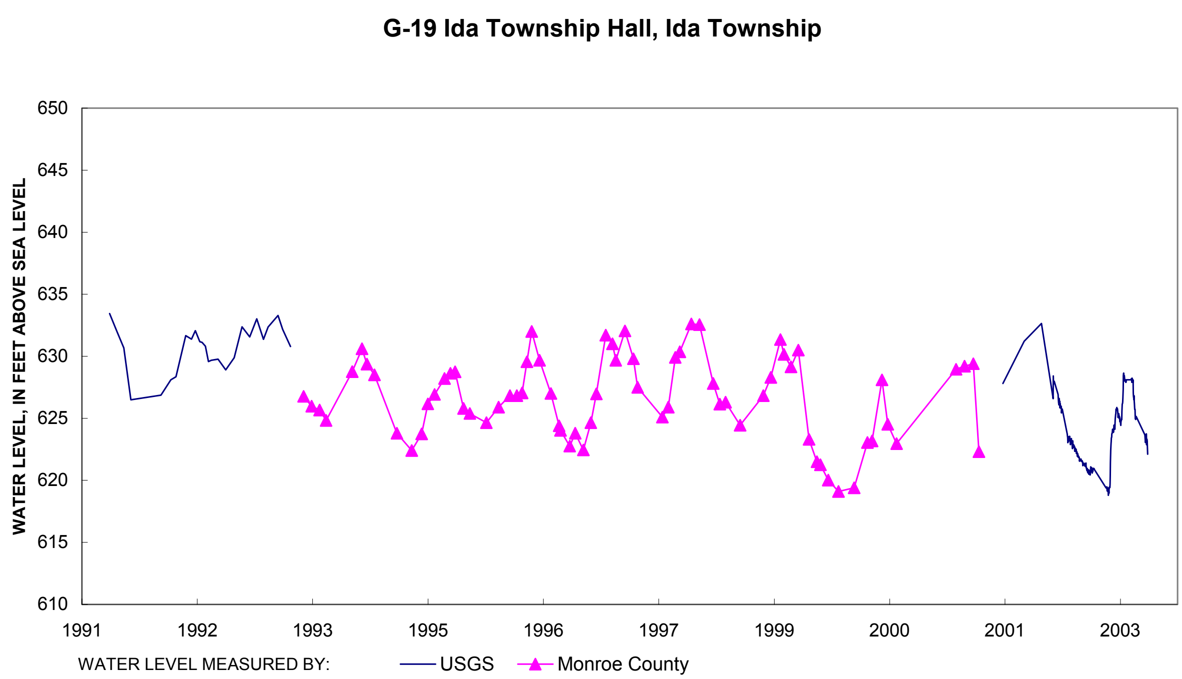

Figure A19. Ground-water levels for well G-19 Ida Township, Monroe County, Michigan.

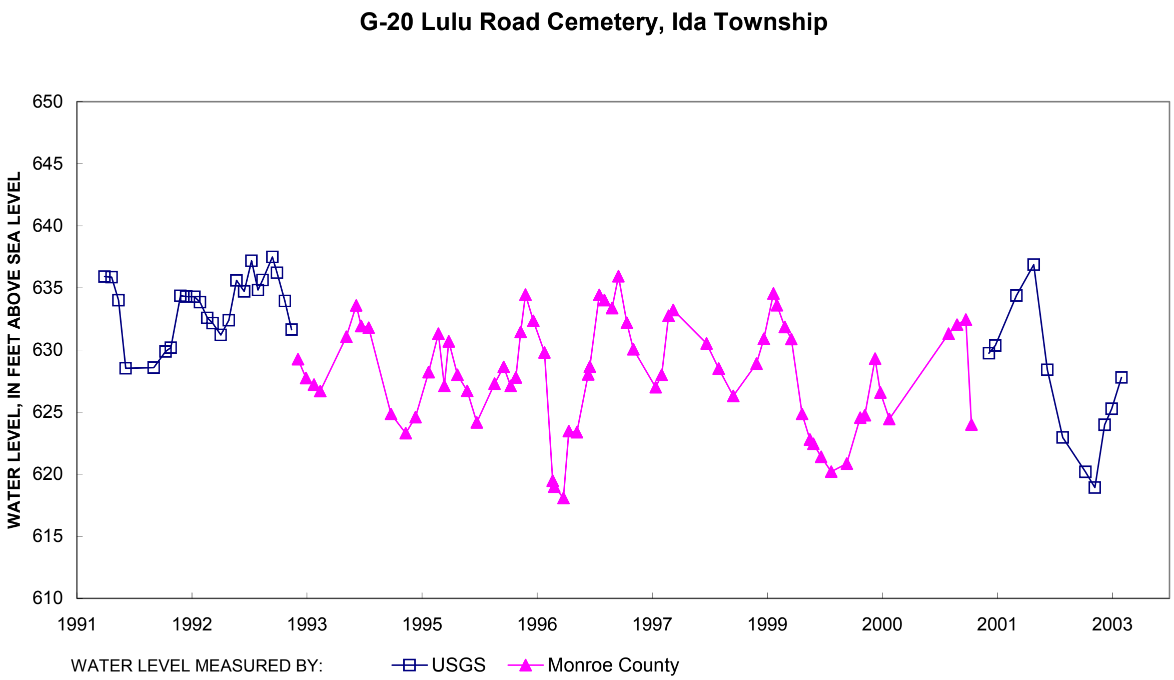

Figure A20. Ground-water levels for well G-20 Ida Township, Monroe County, Michigan.

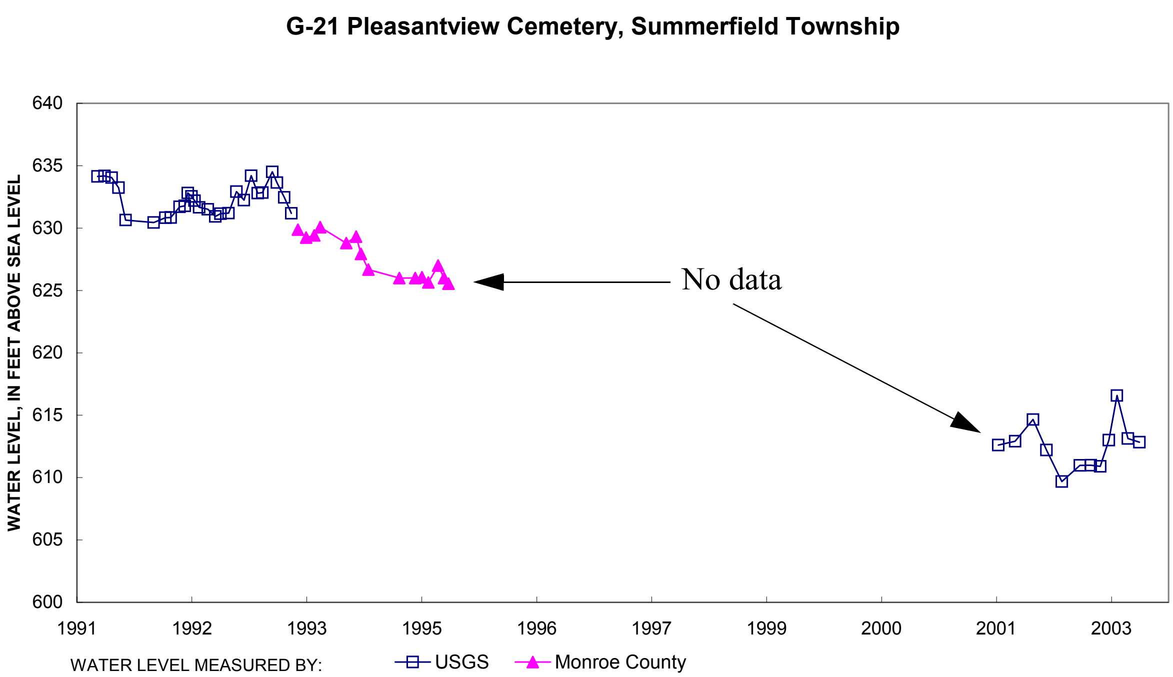

Figure A21. Ground-water levels for well G-21 Summerfield Township, Monroe County, Michigan.

G-22 Petersburg Game Area, Summerfield Township

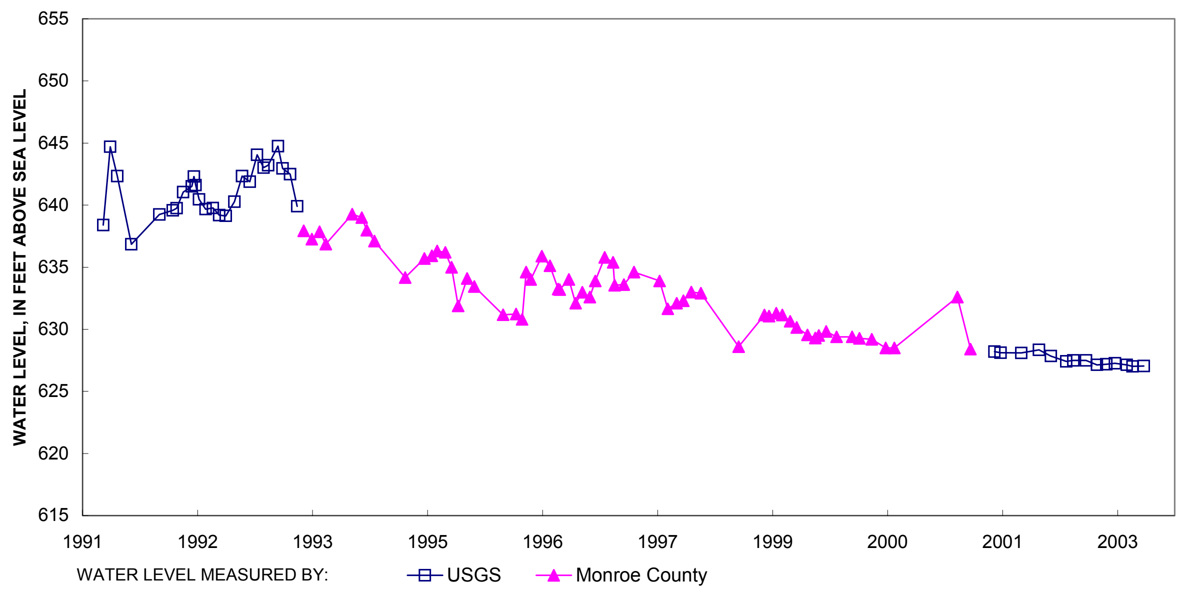

Figure A22. Ground-water levels for well G-22 Summerfield Township, Monroe County, Michigan.

G-23 Petersburg Game Area, Summerfield Township

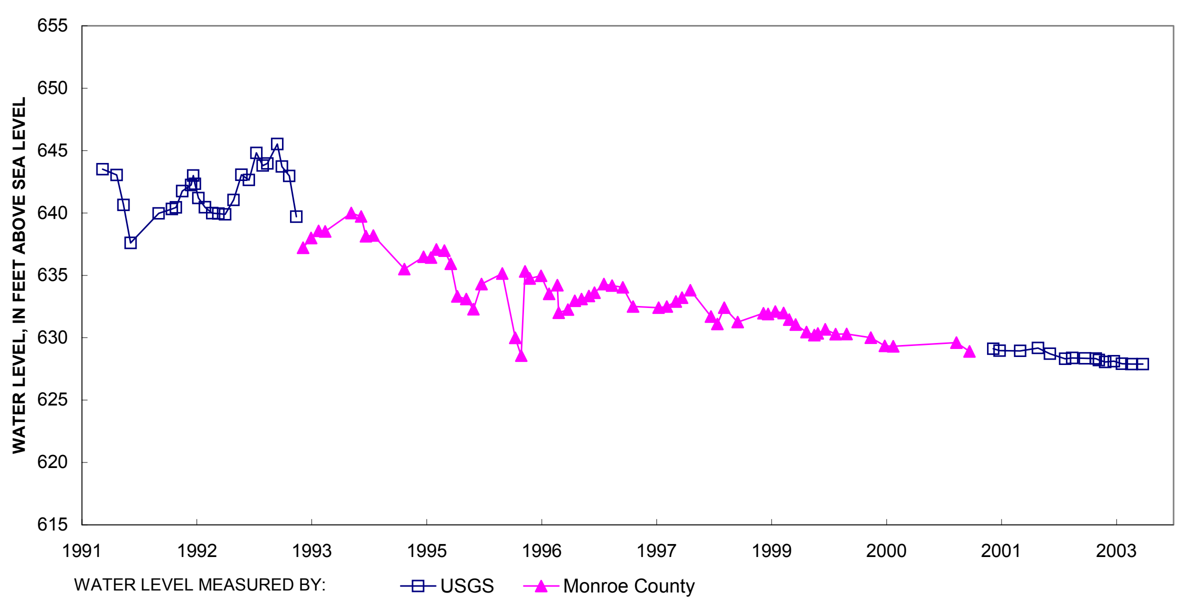

Figure A23. Ground-water levels for well G-23 Summerfield Township, Monroe County, Michigan.

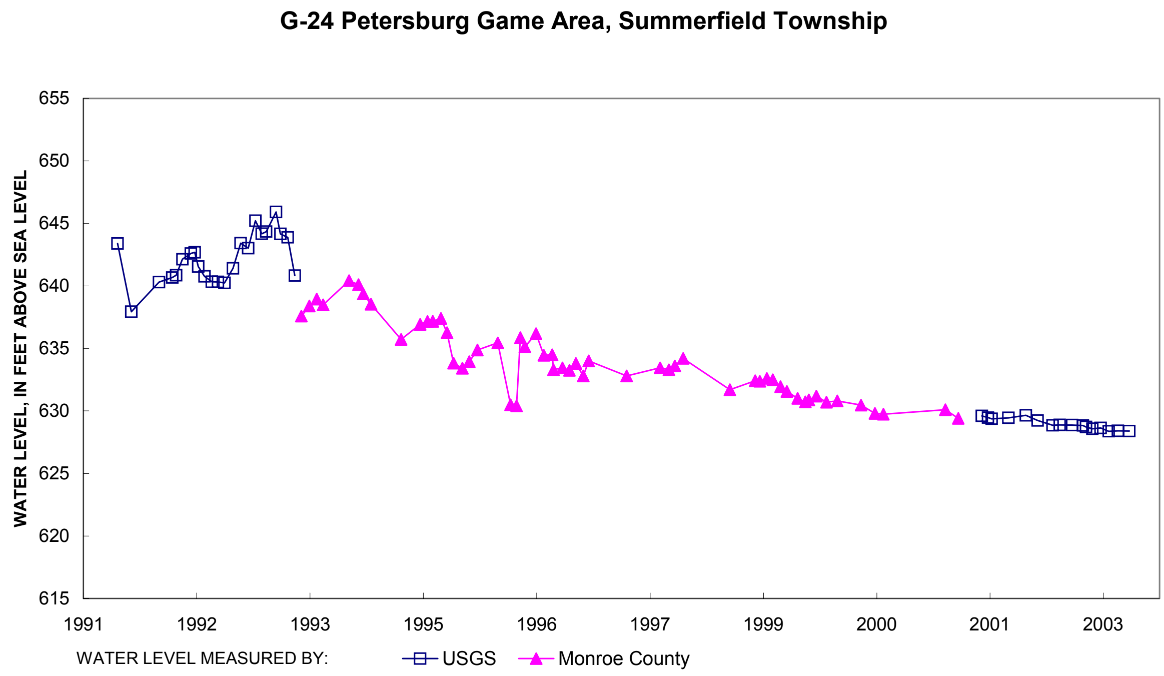

Figure A24. Ground-water levels for well G-24 Summerfield Township, Monroe County, Michigan.

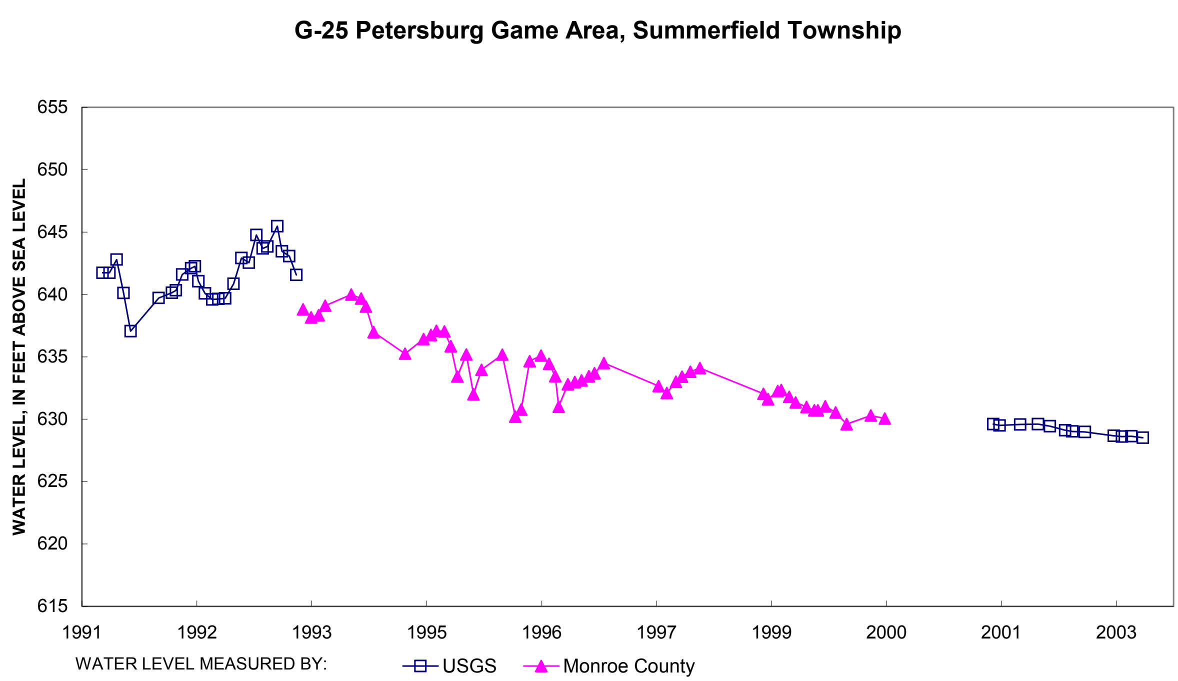

Figure A25. Ground-water levels for well G-25 Summerfield Township, Monroe County, Michigan.

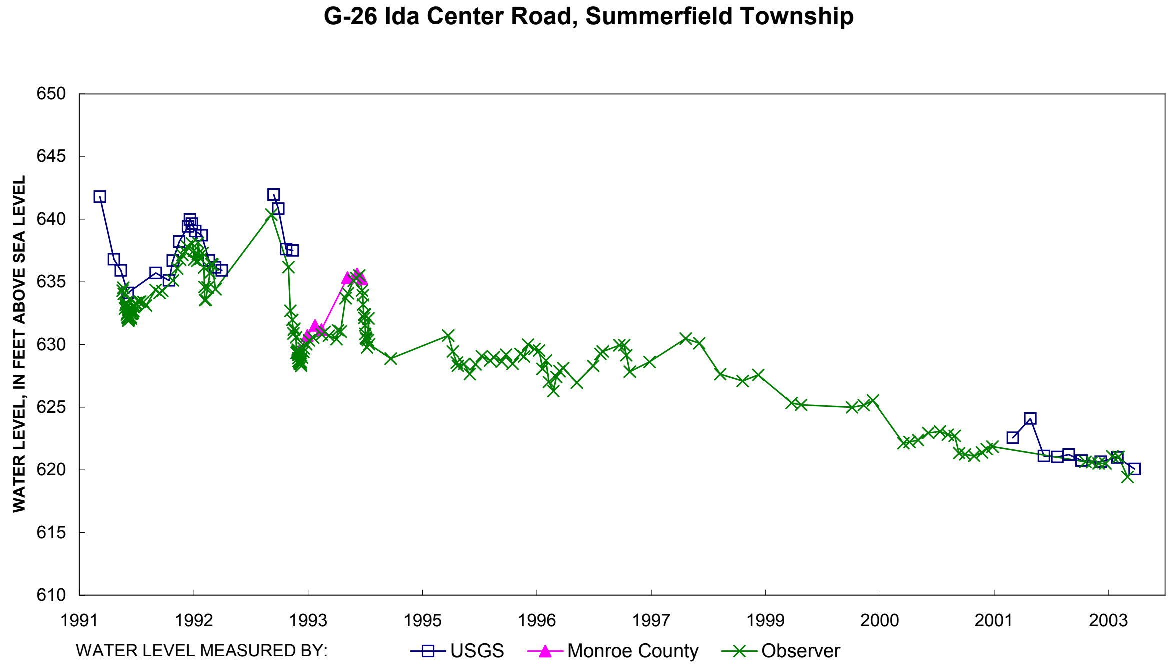

Figure A26. Ground-water levels for well G-26 Summerfield Township, Monroe County, Michigan.

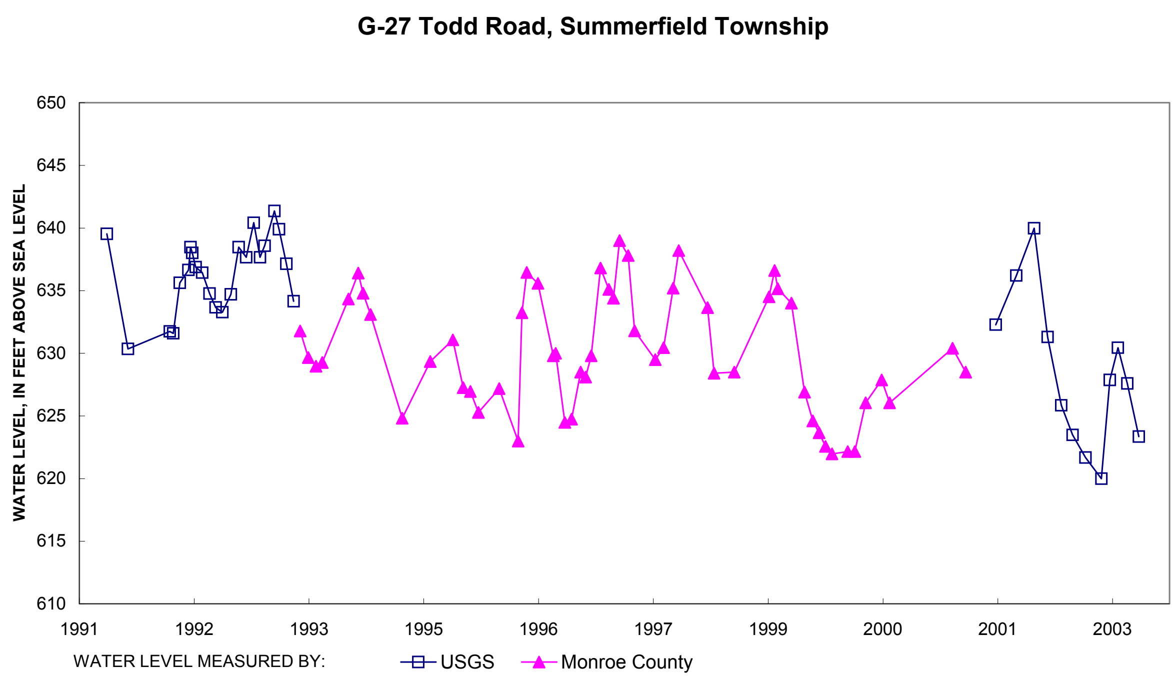

Figure A27. Ground-water levels for well G-27 Summerfield Township, Monroe County, Michigan.

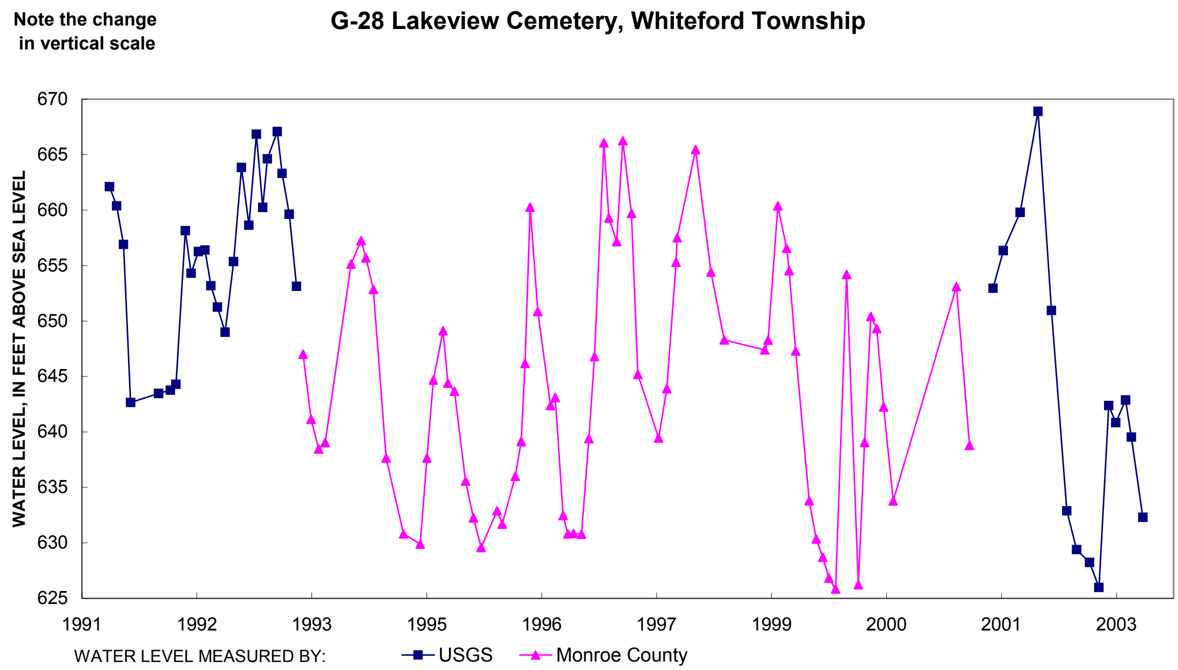

Figure A28. Ground-water levels for well G-28 Whiteford Township, Monroe County, Michigan.

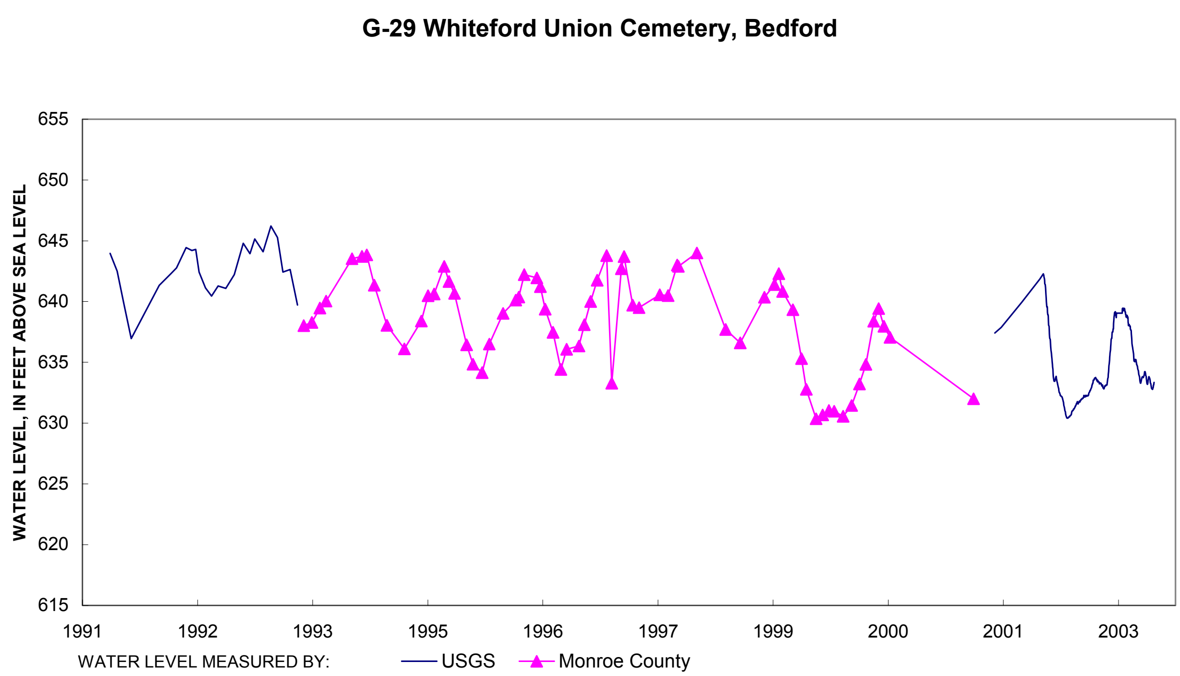

Figure A29. Ground-water levels for well G-29 Whiteford Township, Monroe County, Michigan.

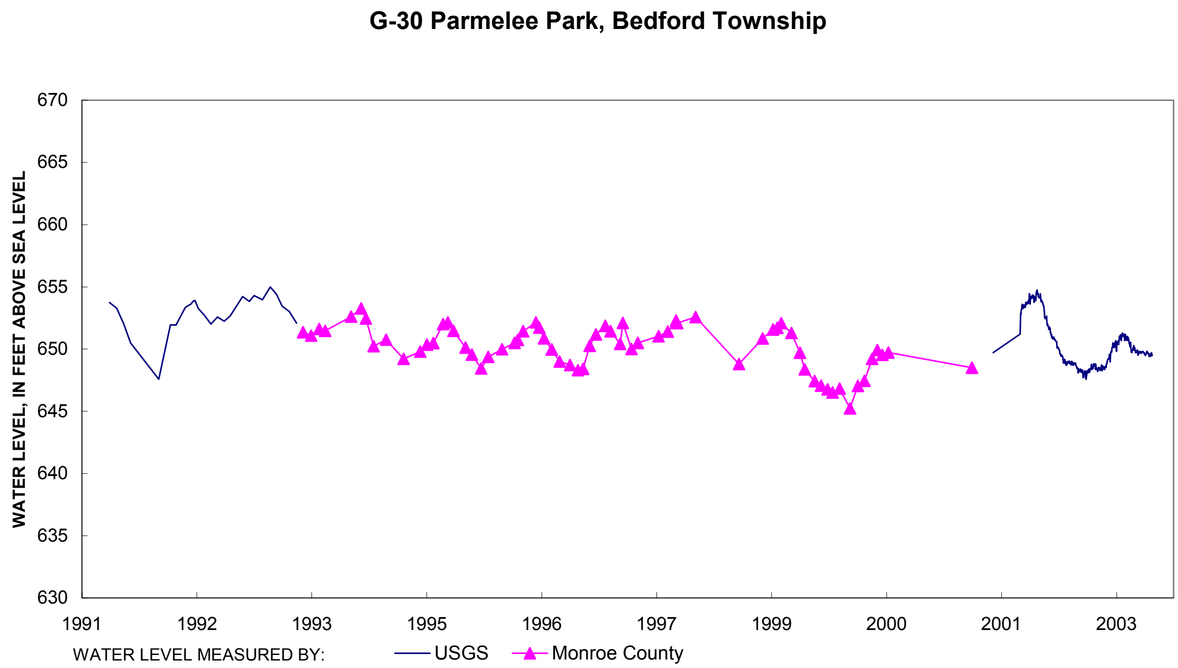

Figure A30. Ground-water levels for well G-30 Bedford Township, Monroe County, Michigan.

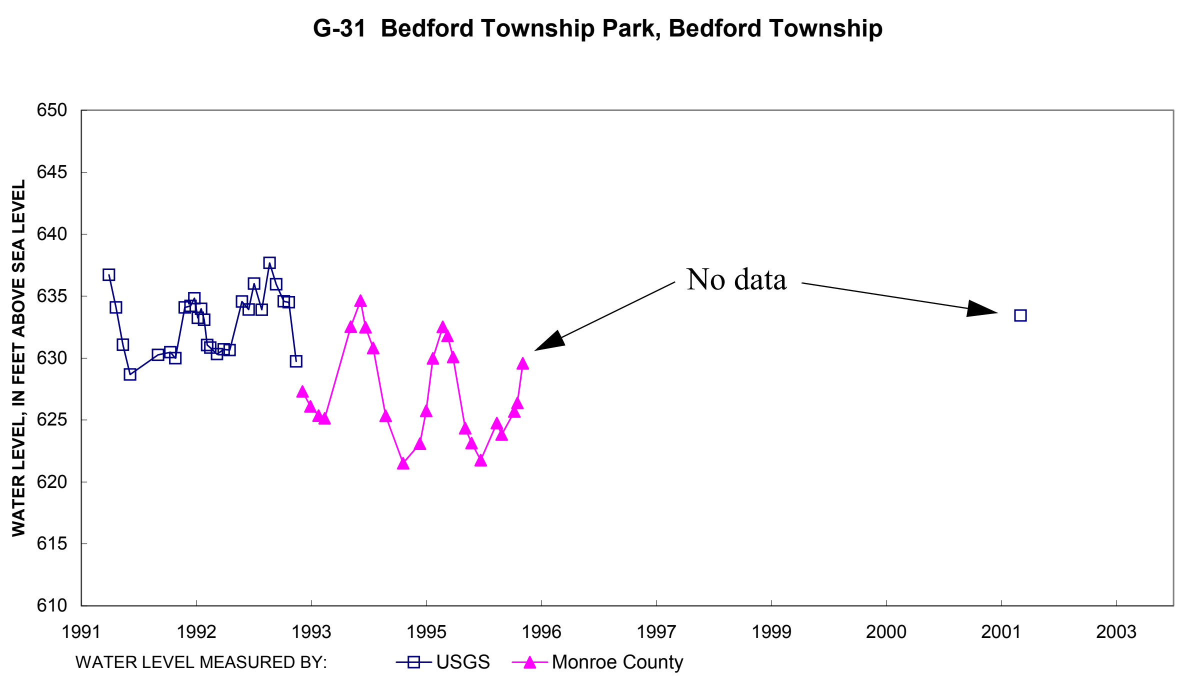

Figure A31. Ground-water levels for well G-31 Bedford Township, Monroe County, Michigan.

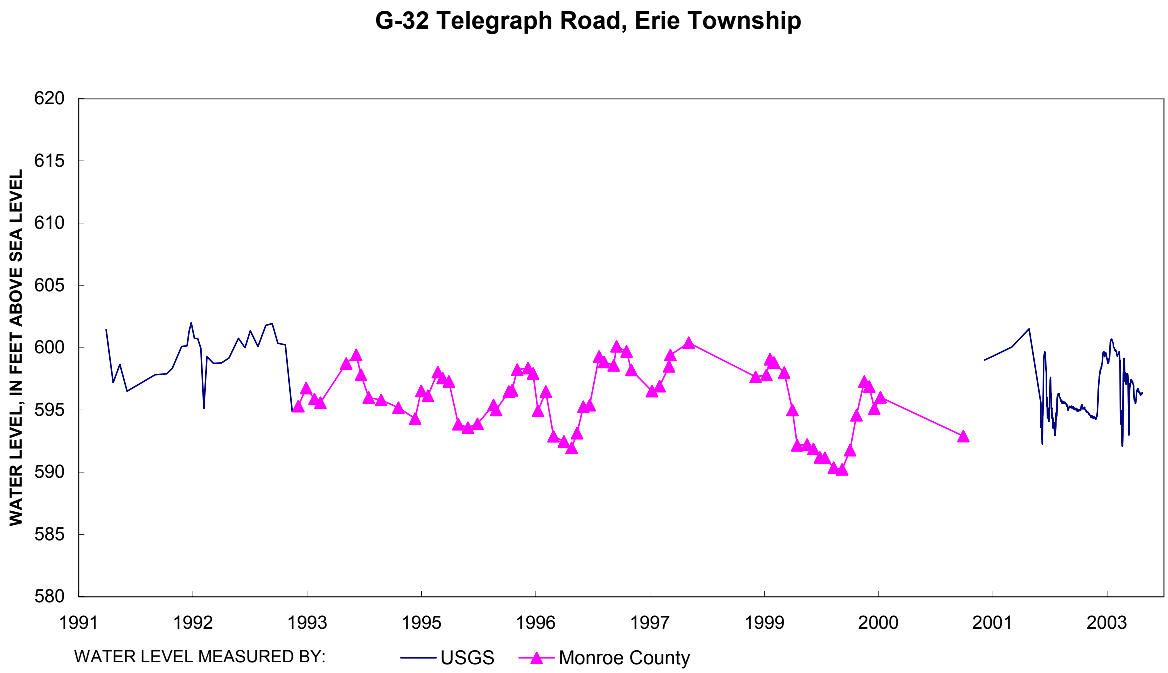

Figure A32. Ground-water levels for well G-32 Erie Township, Monroe County, Michigan.

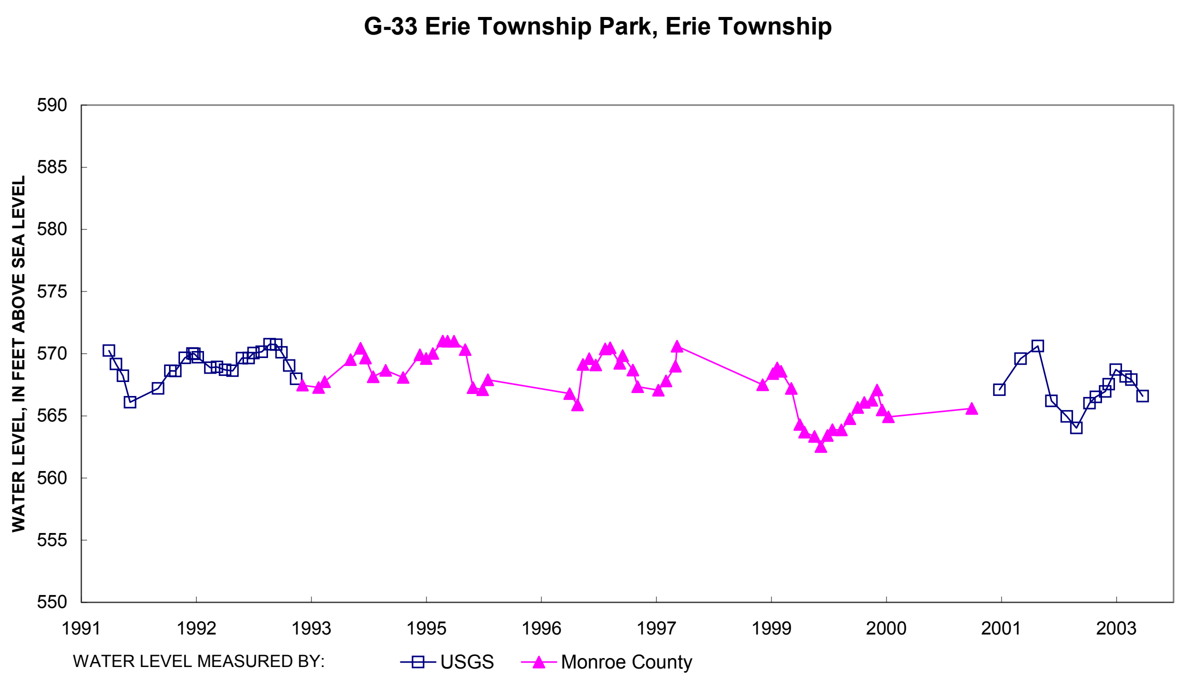

Figure A33. Ground-water levels for well G-33 Erie Township, Monroe County, Michigan.

Landfill records from MDEQ

Water-level measurements from landfill observation wells were obtained from the Michigan Department of Environmental Quality. Water-levels in bedrock wells were available from four landfills: Carlton Farms, Matlin Road, Rockwood Landfill, and Jefferson Smurfit. Rockwood Landfill records from 1980 predate the two nearby quarries, and the influence of the quarries can be seen in more recent measurements. Other landfills appear to match local trends observed in USGS observation wells (Carleton Farms and Matlin Road with G4 and Jefferson Smurfit with G16).

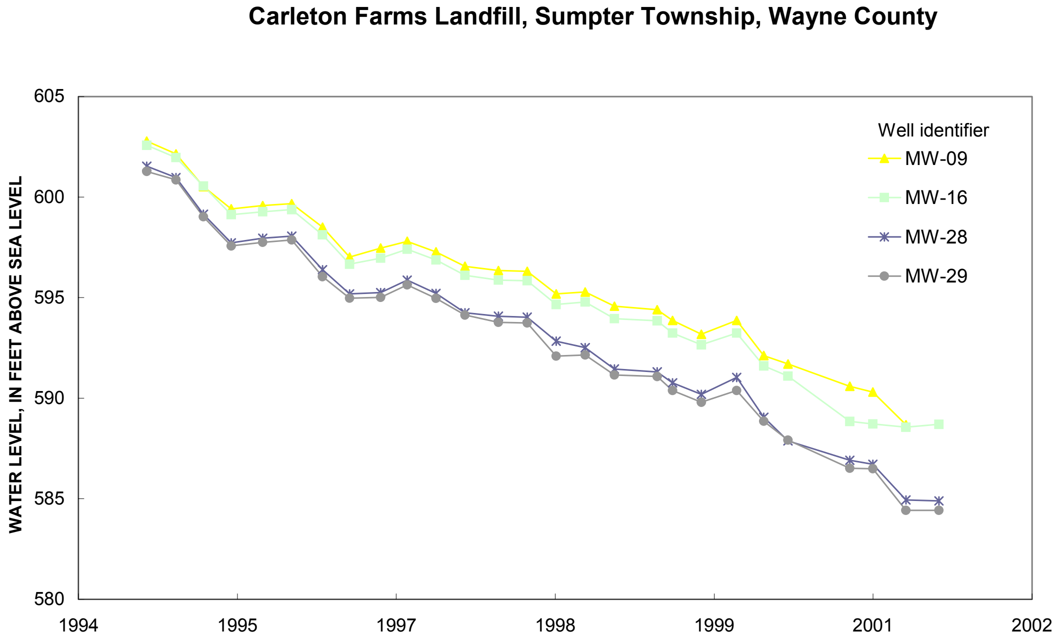

Figure A34. Ground-water levels for observation wells at Carleton Farms Landfill, Monroe County, Michigan.

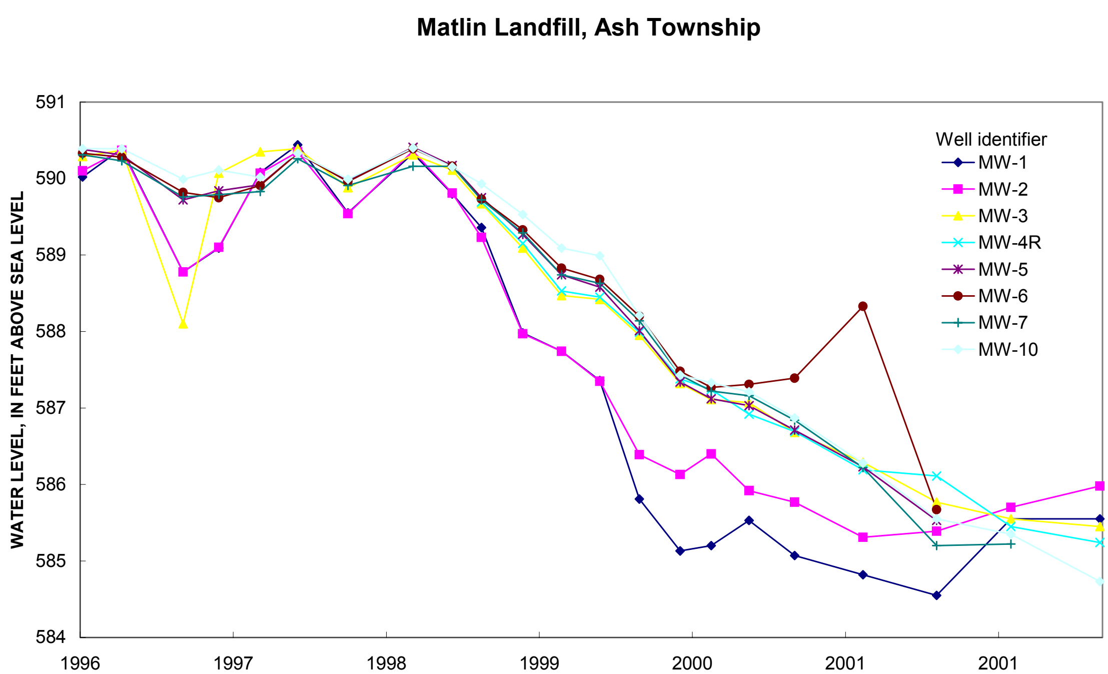

Figure A35. Ground-water levels for observation wells at Matlin Landfill, Monroe County, Michigan.

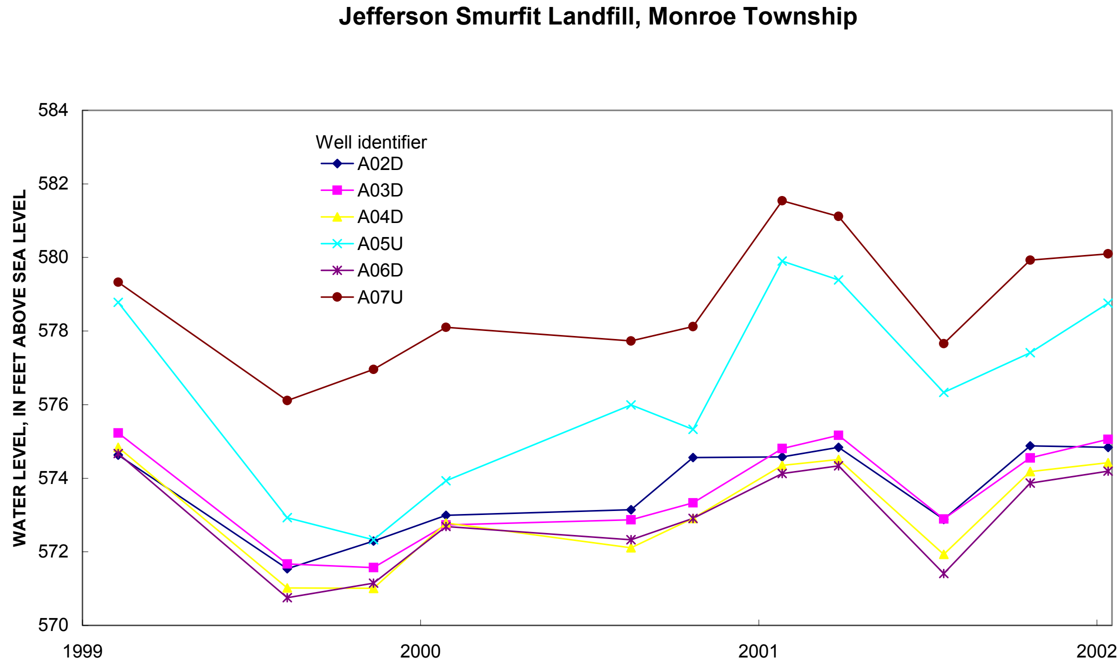

Figure A36. Ground-water levels for observation wells at Jefferson Smurfit Landfill, Monroe County, Michigan.

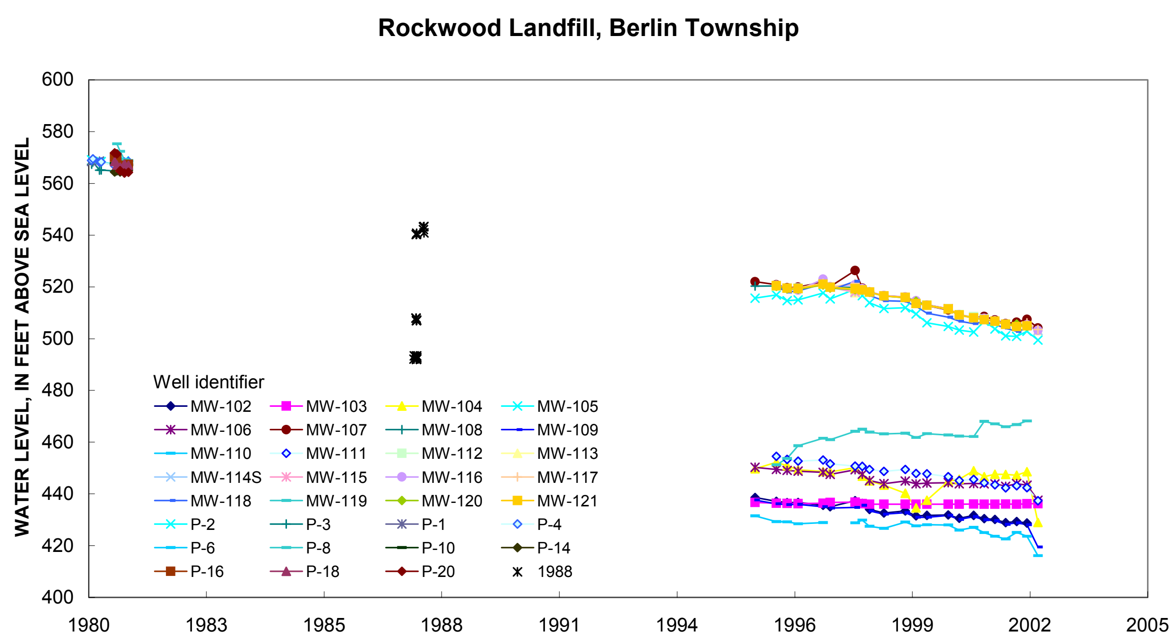

Figure A37. Ground-water levels for observation wells at Rockwood Landfill, Monroe County, Michigan.

Appendix B

Monroe County Statistics Summary

Data Used

Ground-water-level data summarized by Nicholas and others (2002) were used in this analysis. These data are presented in Appendix A of this report. All of the wells except G10 are open to the bedrock aquifer. G10 extends to 110 feet below the ground surface and is screened to glacial deposits approximately four feet above the glacial deposit/bedrock contact.

Daily mean discharge, monthly mean discharge, annual means, and seven-day average low flow data for the River Raisin at three gaging stations were used in an analysis of surface-water trends. The sites and periods of record are: near Monroe in Monroe County (1937-2001), near Adrian in Lenawee County (1953 - 2001), near Manchester in Washtenaw County (1970-2001), and Otter Creek near LaSalle in Monroe County (1987-2001).

Methods

The index developed by Kendall to describe correlation is now referred to as the Kendall τ and is computed by determining examining the differences between every possible pair of observations in chronological order. If the difference between a pair is positive, N_c is incremented by one. If the difference is negative N_d is incremented by one. The total number of pairs is *n(n-1)/2*. τ is defined as,

$$\tau = \frac{N_c - N_d}{n(n-1)/2}$$

If the observations are monotonically increasing, then N_c will be equal to the total number of pairs and τ will be 1. If the observations are monotonically decreasing, τ will be -1. The algorithm used in S-PLUS (Mathsoft, 2000) also allows for tied values. This rank test indicates where a positive (τ close to 1) or negative (τ close to -1) trend is present in the observations. A τ close to zero implies no trend is indicated by the data. In addition to the Kendall τ, the statistical significance of the trend is determined. For values of *n* larger than 40, τ has a normal distribution, and this property allows the critical level (*p*-value) to be determined for a statistical evaluation of the null hypothesis of the test that there is no trend supported. Large *p*-values imply that the null hypothesis cannot be rejected.

The magnitude of the trend was computed using the Sen slope procedure. In this method, the slope for every pair in the ranked series is obtained,

$$Q = \frac{o_{i'} - o_i}{t_{i'} - t_i} \forall i' > i$$

Where o_i and t_i are the values of the observation and the observation time at *i*. If there is only one observation per time period, which is the case for the data used in this study, then this evaluation yields *n(n-1)/2* values of *Q*. These values are then put in rank order, and the Sen estimator of the slope is the median value of the ranked list. The results of this analysis are summarized in Table B-1.

The largest magnitude for the Sen slope for a water-level record was -8.7 ft/year at well G-17. Note that this well is less than 500 ft from a quarry that was deepened and increased ground-water withdrawals during the study period. At this distance, the water level is directly under the influence of the quarry depth, and this large decline is not indicative of declines throughout the county. Most wells had declines that can be grouped into three clusters. Wells in the northeastern part of the county (G-1 through 4), wells (G-12 through 15, G-19, G-20, G-29, G-31), and wells near Petersburg (G-21 through 28) had slopes between -1 and -2 ft/year. Wells in the northern part of the county (G-5 through 9, G-11) had slopes between -2 and -4.5 ft/year. The remaining wells (G-10, G-16, G-18 through 20, G-30, G-32, G33) showed little decline. Well G-10 does not follow the pattern in the northeastern portion of the county, but this is the only monitoring well completed in the glacial deposits overlying the bedrock aquifer. Well G-13 shows the only positive trend over the time period. The *p*-value for this well (0.16), however, indicates little statistical likelihood of a trend at this well. The only other well with a high *p*-value (0.13) is G-18. The remainder of the wells have *p*-values that are extremely small, and the statistical likelihood of no trend in these wells is less than 1 percent.

The same statistical procedures were used to evaluate the potential for trends in the mean daily discharge values from 1991 through 2001 for the River Raisin near Monroe, Manchester, and Adrian, and Otter Creek near Lasalle. Two analyses were performed. The first used the mean daily discharge values for the gaging stations with time, and the second normalized the daily discharge values by the drainage area associated with each station. These analyses showed a statistically significant downward trends in the mean daily discharge for each station. The trend in mean daily discharge, when normalized by the annual mean discharge is approximately 3 percent decline per year. When normalized by the drainage area, the trends for the three stations on the River Raisin are fairly close in magnitude implying that any changes in River discharge cannot be solely attributed to activities within Monroe County downstream of the Adrian and Manchester gages. These results are summarized in Table B-2. The approach was repeated using the average monthly discharge values for the period of record for each station. The results of this analysis indicate a slightly positive, but statistically significant, trend at each station on the River Raisin. Changes in land use over the period of record may have caused these trends.

The time-series analysis for surface water support the conclusions made by Nicholas and others (2002) that stream discharge numbers do not show the same systematic and sigificant decline as noted in ground-water levels. Examination of the variation of stream discharge over the period of record reveals that stream discharge was much lower during the drought period in the 1960s than during for the past ten to fifteen years. These findings indicate that baseflow to streams is dominated by flow from the overlying glacial material and not from the bedrock aquifer.

Table B-1. Results of trend analysis for ground-water levels in U.S. Geological Survey monitoring wells in Monroe County, Michigan, 1991-2001 [ft/yr, feet per year]

Monitoring well	Kendall τ	*p*-value	Sen slope (ft/yr)
G-1	-0.786	9.90E-02	-1.38
G-2	- .454	2.50E-09	- .888
G-3	- .726	6.00E-26	-1.37
G-4	- .748	2.00E-27	- .833
G-5	- .91	2.00E-36	-3.
G-6	- .992	1.00E-38	-1.72
G-7	- .876	7.00E-19	-4.5
G-8	- .928	1.20E-39	-3.1
G-9	- .759	5.70E-12	-3.7
G-10	- .55	1.60E-14	- .4
G-11	- .74	1.54E-22	-2.95
G-12	- .806	1.00E-30	-1.7
G-13	.21	1.60E-01	.96
G-14	- .44	2.70E-10	- .86
G-15	- .49	7.30E-13	- .8
G-16	- .423	8.00E-09	- .4
G-17	- .82	1.00E-21	-8.7
G-18	- .106	1.26E-01	-0.05
G-19	- .29	2.00E-05	- .52
G-20	- .27	7.90E-05	-0.6
G-21	- .538	3.80E-07	-1.87
G-22	- .74	7.00E-25	-1.4
G-23	- .76	2.70E-26	-1.37
G-24	- .77	2.90E-24	-1.36
G-25	- .72	1.70E-21	-1.36
G-26	- .51	4.00E-27	-1.07
G-27	- .34	8.00E-06	-0.91
G-28	- .18	9.00E-03	-1.11
G-29	- .38	4.90E-08	-.7
G-30	- .46	6.00E-11	- .46
G-31	- .38	8.20E-05	-1.6
G-32	- .31	8.10E-06	-.48
G-33	- .35	5.00E-06	- .35
GLTO	- .54	4.20E-28	- .62

Table B-2. Surface-water flow statistical trends, 1991 - 2001
[cfs/year, cubic feet per second per year; ft/sec/year, feet per second per year]

Site	Kendall τ	Slope	Slope/Annual Mean
River Raisin near Monroe	- 0.086	-17. cfs/year	-0.023 1/year
River Raisin near Manchester	- .104	-3.3 cfs/year	- .031 1/year
River Raisin near Adrian	- .084	-8.6 cfs/year	- .025 1/year
Otter Creek	- .134	-1.3 cfs/year	- .028 1/year
River Raisin near Monroe normalized by drainage area	- .086	-6. E-10 ft/sec/year	
River Raisin near Manchester normalized by drainage area	- .104	-9. E-10 ft/sec/year	
River Raisin near Adrian normalized by drainage area	- .084	-7. e-10 ft/sec/year	

Appendix C

WelLogic Analysis

The box-and-whisker diagrams in this appendix were developed by grouping wells from the WelLogic database (Michigan Department of Natural Resources, 2003) into two-year groups by the date the well was drilled. The mean, median, lower and upper quartiles, and outliers were determined for each group with more than ten wells. The dotted line on the figure connects the mean values. Groups with less than ten wells drilled during a four-year period are plotted using open circles for individual wells and solid circles at the means. Examination of the figures reveals significant scatter for the depth to water for each group of four years in each township. The boxes enclose data values between the lower quartile and upper quartile. These boxes tend to enclose a range of several feet. The whiskers show the spread of the data by extending to either the maximum data value or the upper quartile + 1.5*(upper quartile-lower quartile) and to the minimum data value or the lower quartile - 1.5*(upper quartile and lower quartile). In the latter cases, the whiskers help highlight outliers in the data set. The outliers are shown on the plots using asterisks. For most of the townships, the variation in the mean is generally less than the scatter of the groups. Despite this scatter, general trends are evident in the plots. The appendix includes time series for wells open to the bedrock aquifer in Monroe County, Michigan and for wells screened in glacial deposits in Washtenaw County, Michigan.

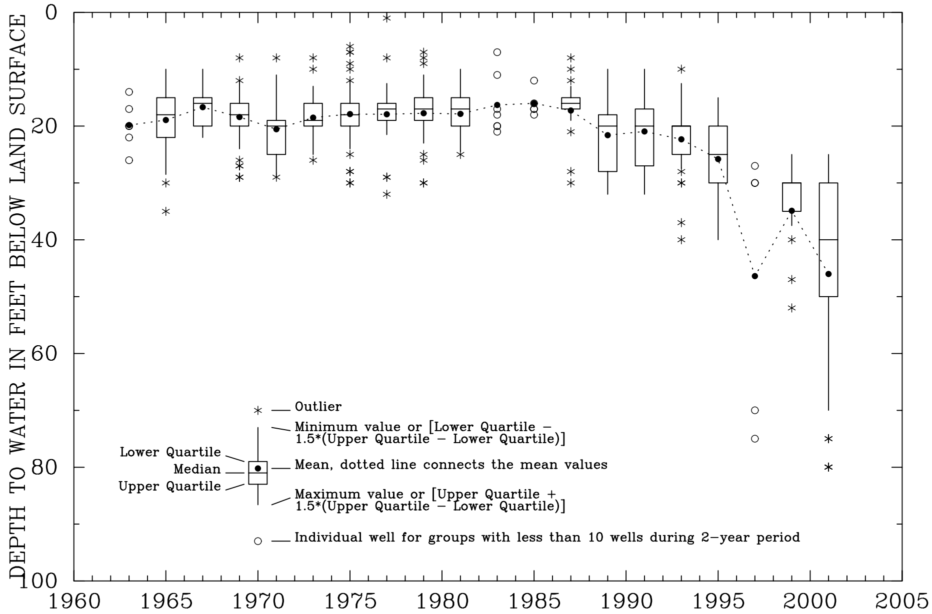

Figure C1. Bedrock aquifer ground-water levels reported in WelLogic database for Ash Township, Monroe County, Michigan.

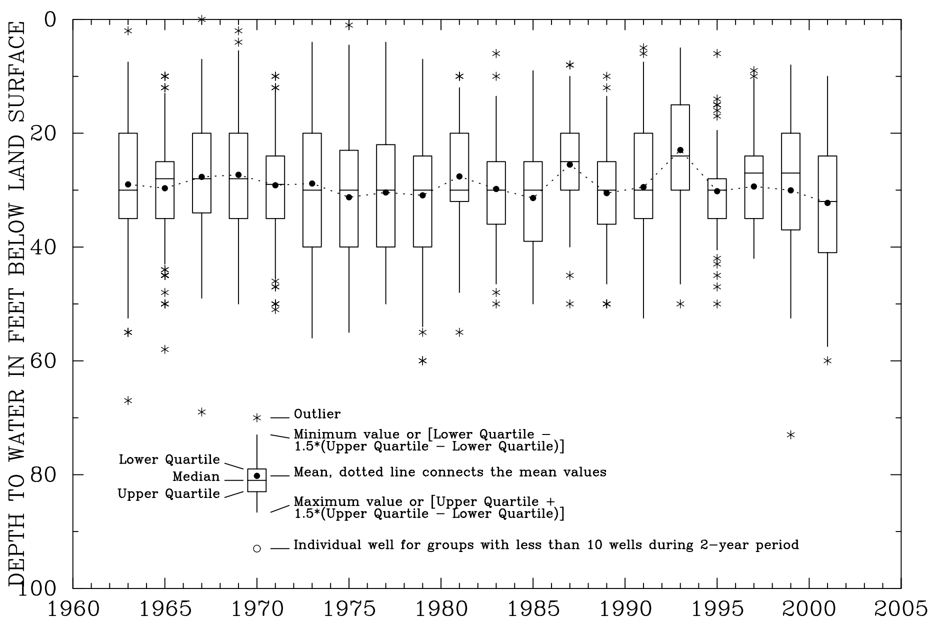

Figure C2. Bedrock aquifer ground-water levels reported in WelLogic database for Bedford Township, Monroe County, Michigan.

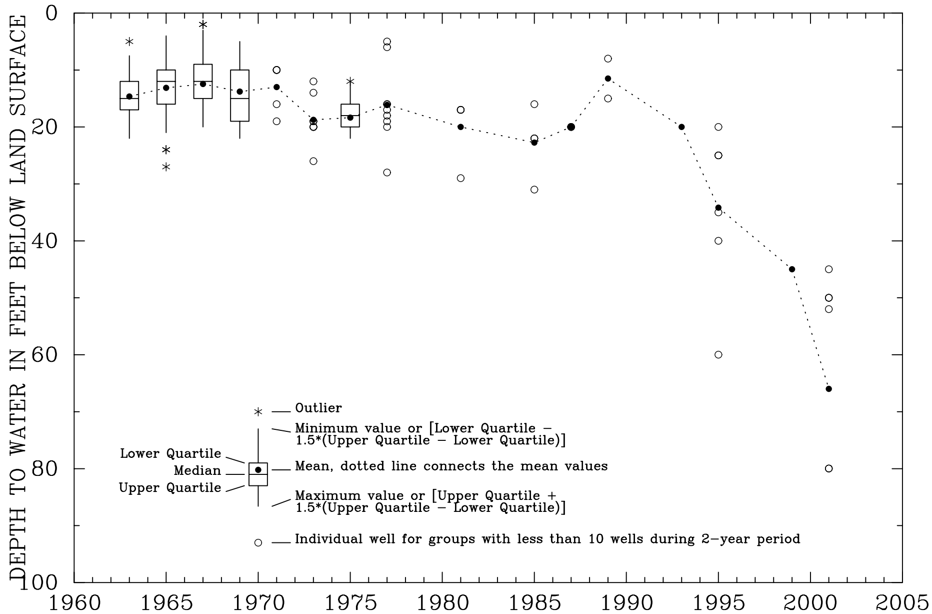

Figure C3. Bedrock aquifer ground-water levels reported in WelLogic database for Berlin Township, Monroe County, Michigan.

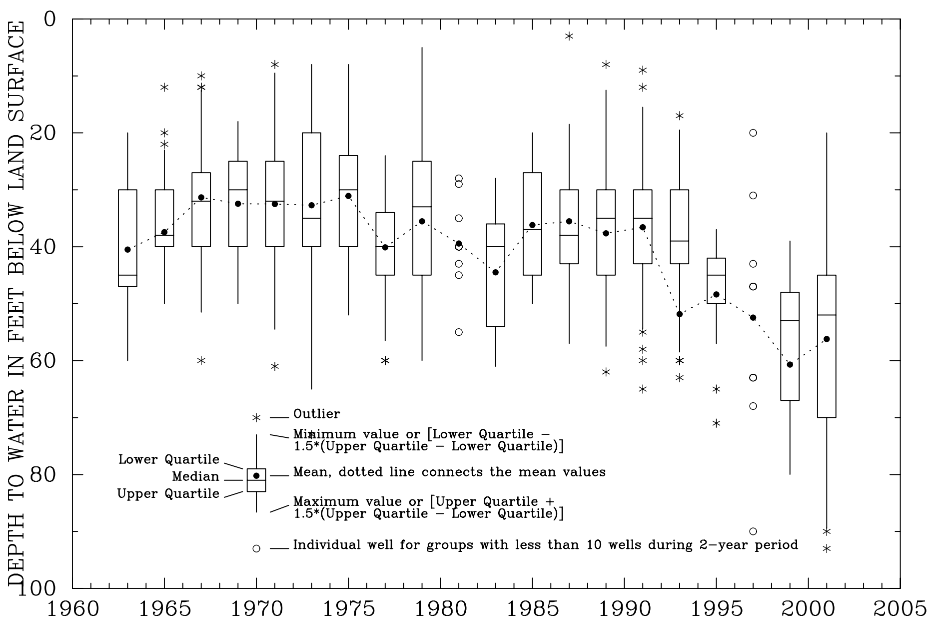

Figure C4. Bedrock aquifer ground-water levels reported in WelLogic database for Dundee Township, Monroe County, Michigan.

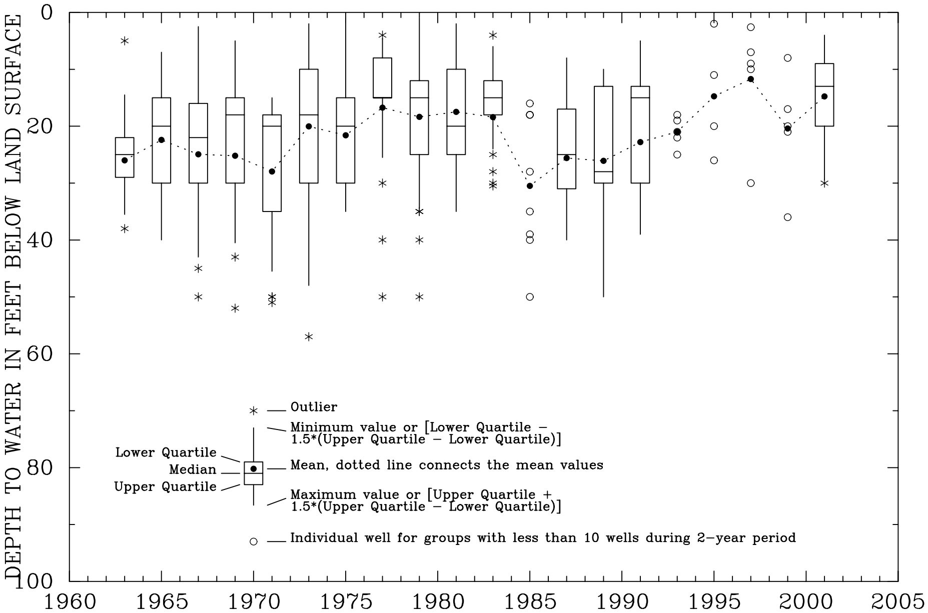

Figure C5. Bedrock aquifer ground-water levels reported in WelLogic database for Erie Township, Monroe County, Michigan.

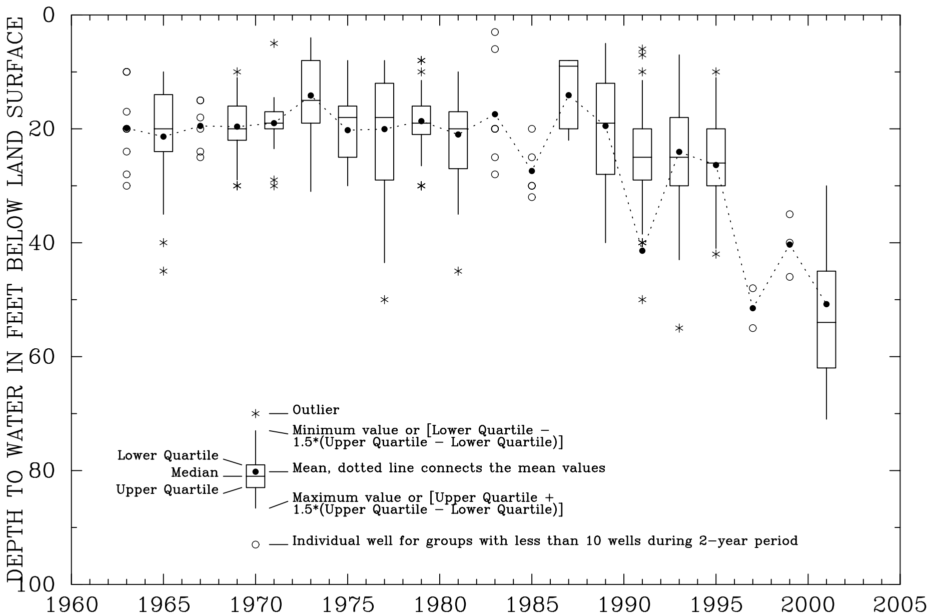

Figure C6. Bedrock aquifer ground-water levels reported in WelLogic database for Exeter Township, Monroe County, Michigan.

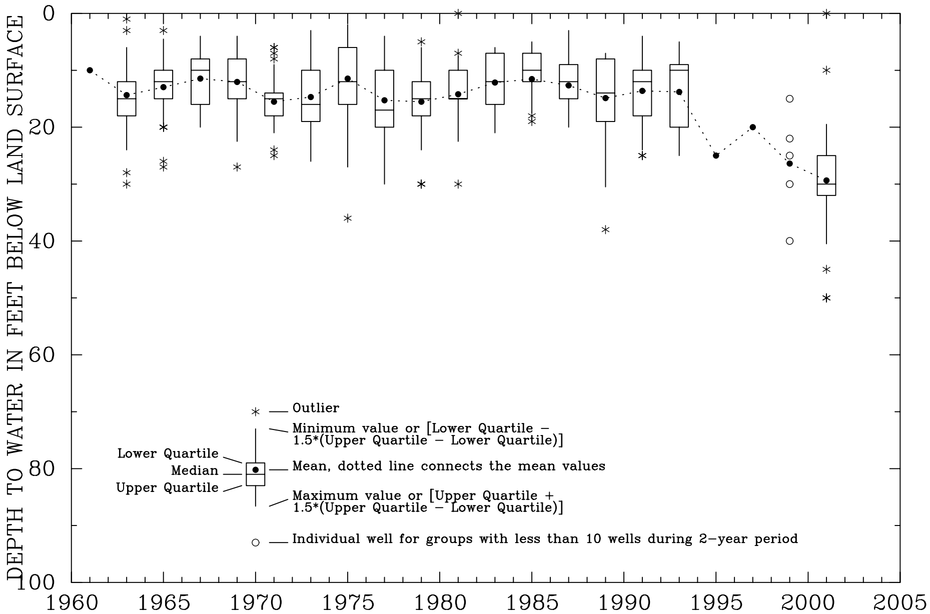

Figure C7. Bedrock aquifer ground-water levels reported in WelLogic database for Frenchtown Township, Monroe County, Michigan.

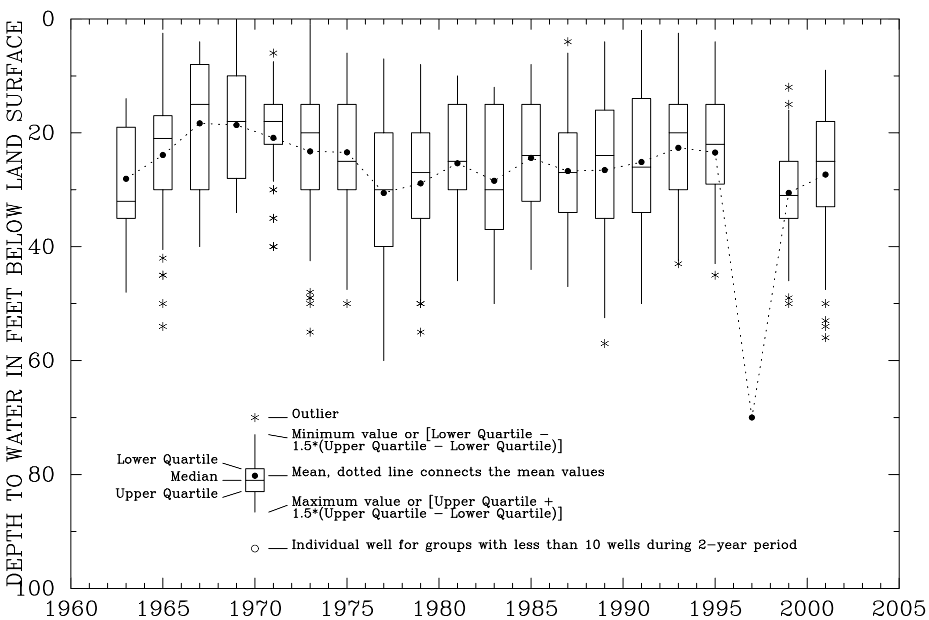

Figure C8. Bedrock aquifer ground-water levels reported in WelLogic database for Ida Township, Monroe County, Michigan.

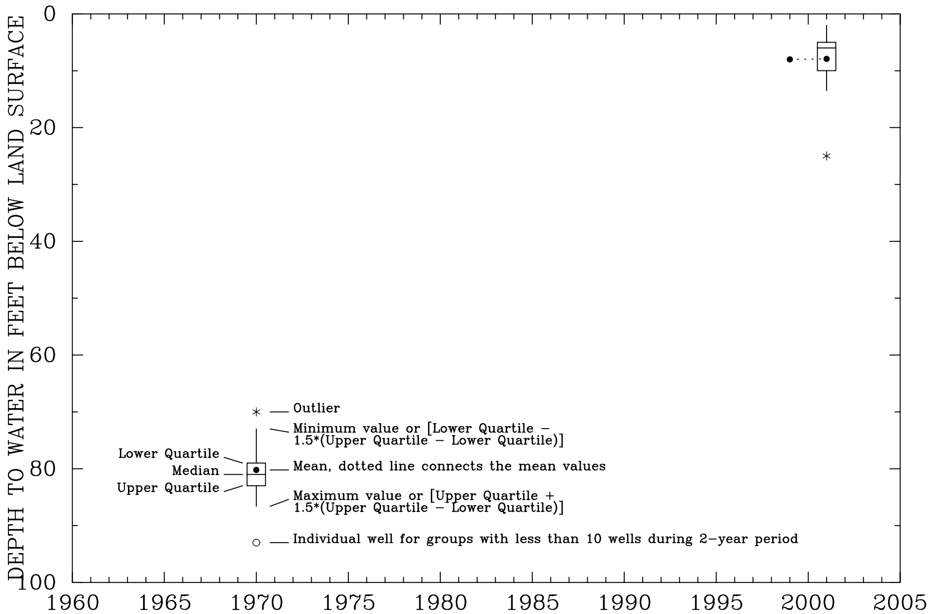

Figure C9. Bedrock aquifer ground-water levels reported in WelLogic database for Lasalle Township, Monroe County, Michigan.

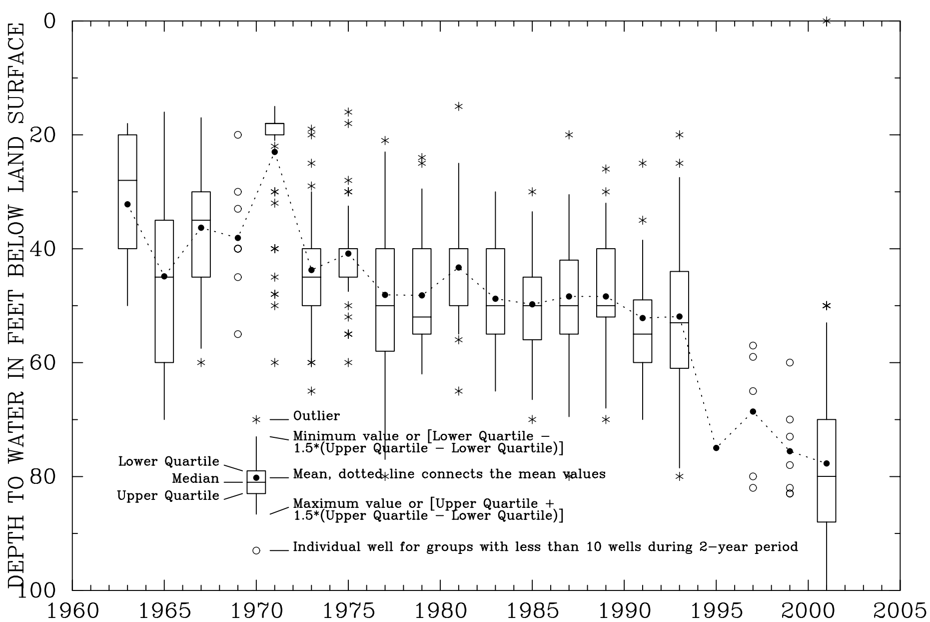

Figure C10. Bedrock aquifer ground-water levels reported in WelLogic database for London Township, Monroe County, Michigan.

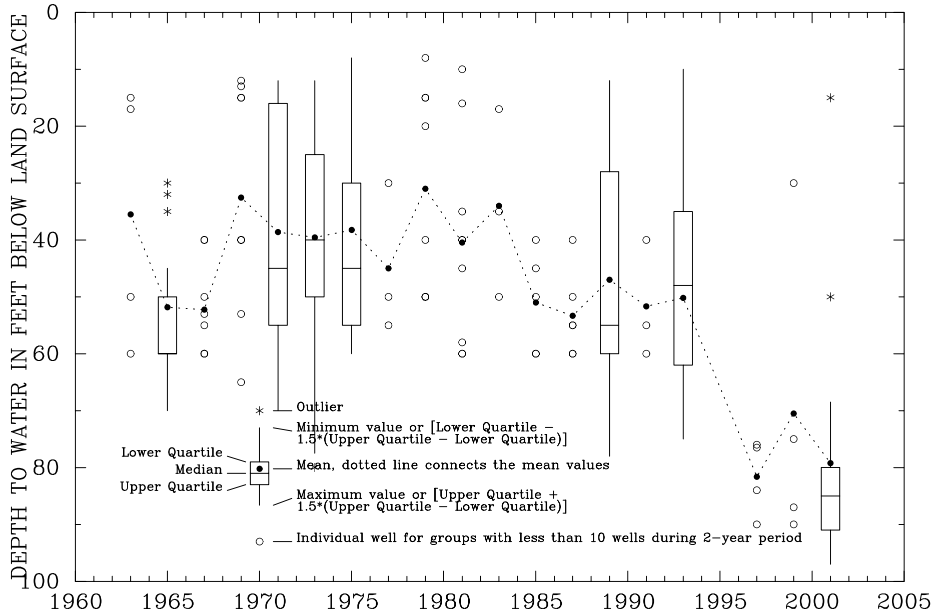

Figure C11. Bedrock aquifer ground-water levels reported in WelLogic database for Milan Township, Monroe County, Michigan.

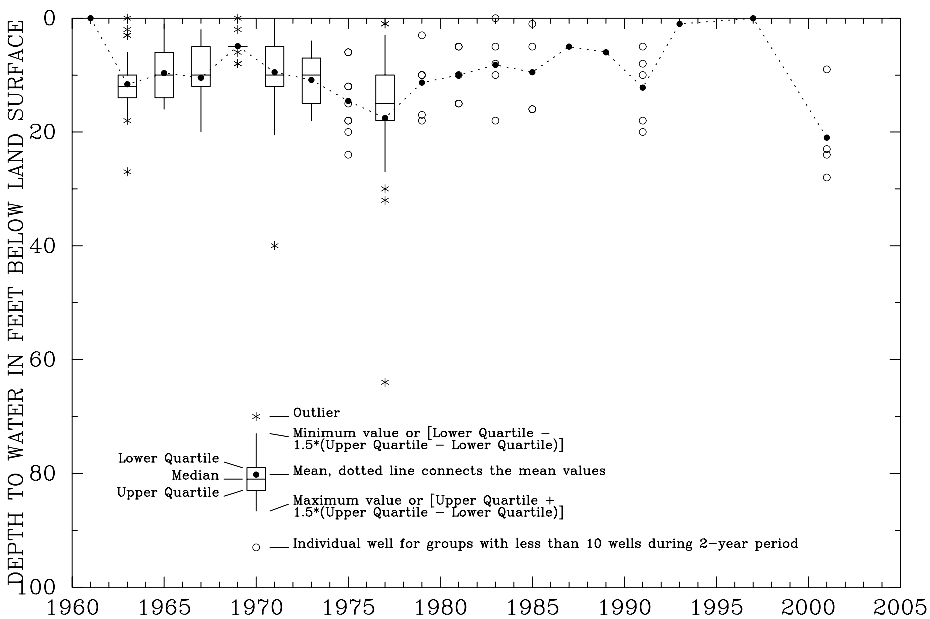

Figure C12. Bedrock aquifer ground-water levels reported in WelLogic database for Monroe Township, Monroe County, Michigan.

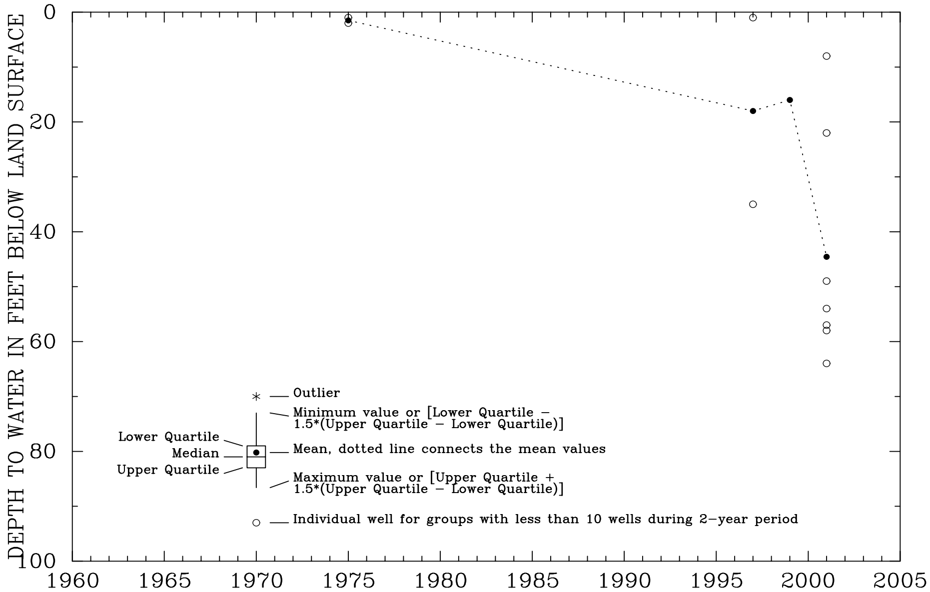

Figure C13. Bedrock aquifer ground-water levels reported in WelLogic database for Raisinville Township, Monroe County, Michigan.

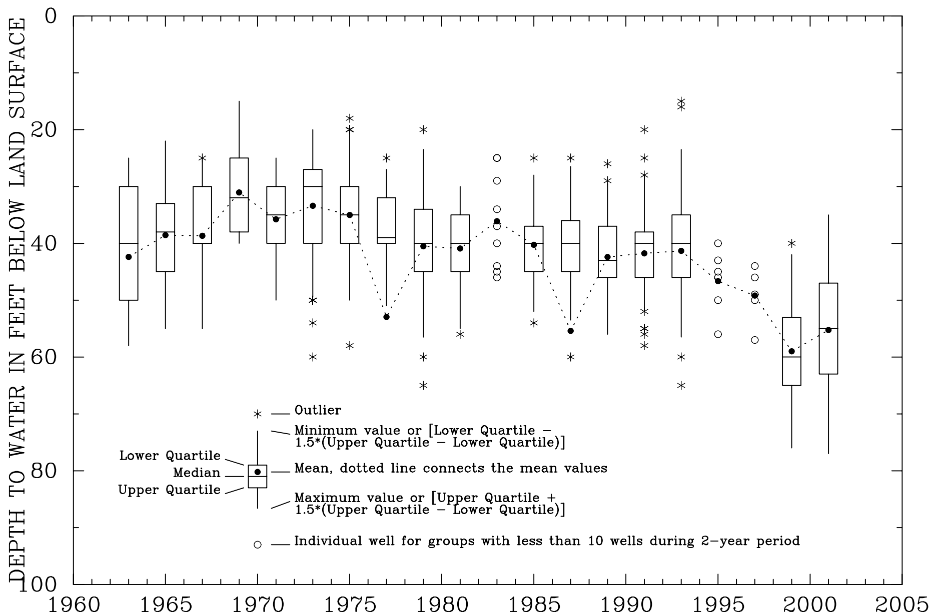

Figure C14. Bedrock aquifer ground-water levels reported in WelLogic database for Summerfield Township, Monroe County, Michigan.

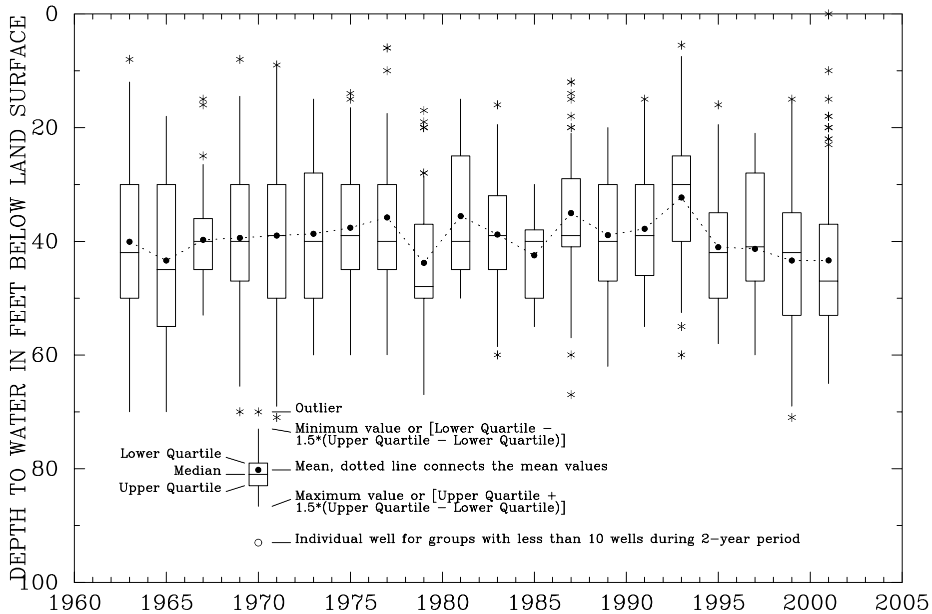

Figure C15. Bedrock aquifer ground-water levels reported in WelLogic database for Whiteford Township, Monroe County, Michigan.

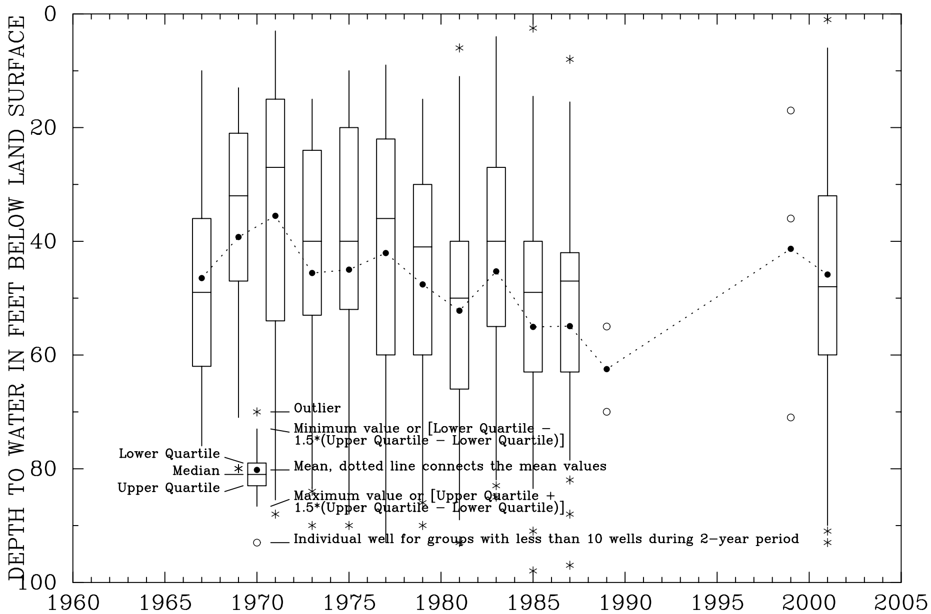

Figure C16. Glacial deposit ground-water levels reported in WelLogic database for Ann Arbor Township, Washtenaw County, Michigan.

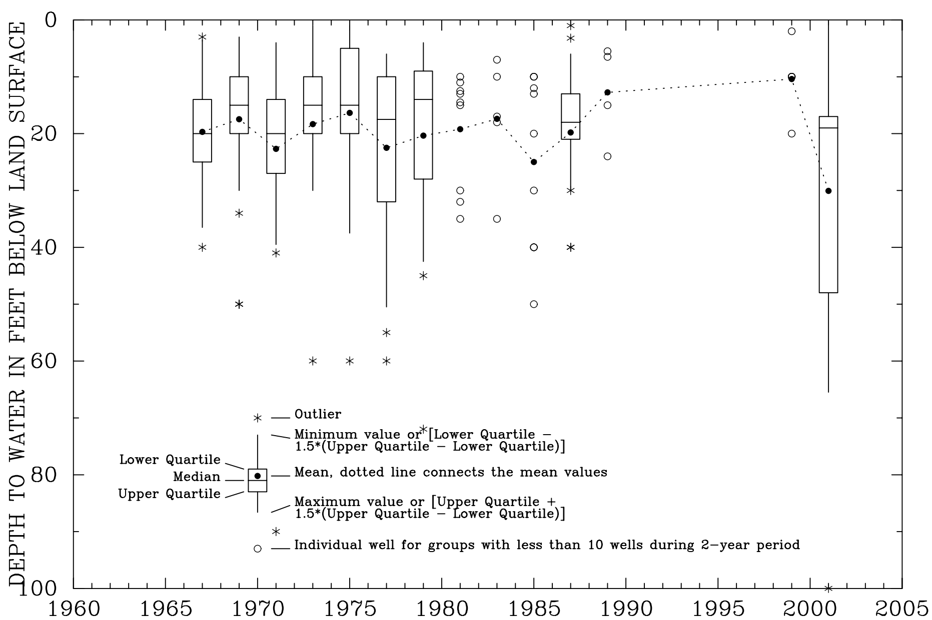

Figure C17. Glacial deposit ground-water levels reported in WelLogic database for Augusta Township, Washtenaw County, Michigan.

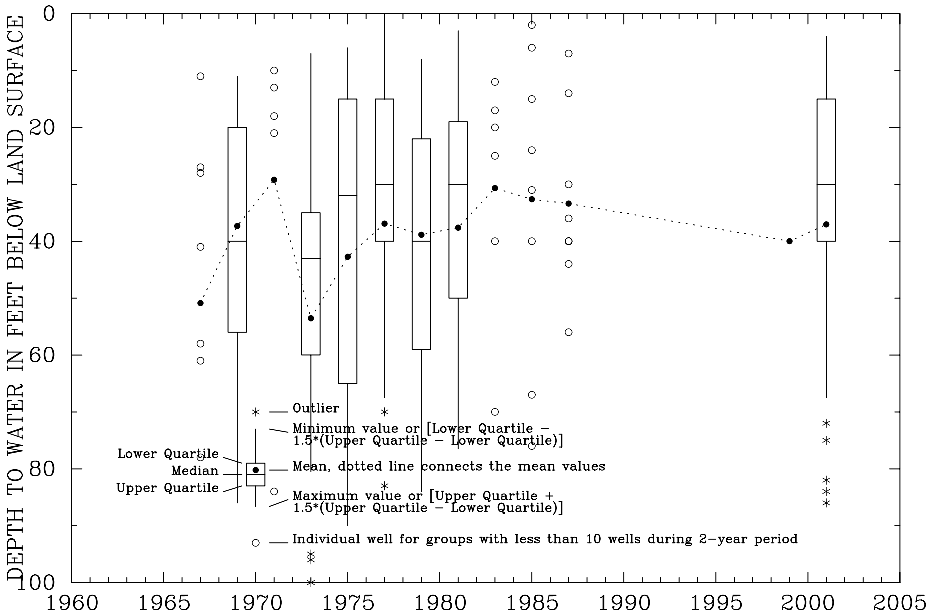

Figure C18. Glacial deposit ground-water levels reported in WelLogic database for Bridgewater Township, Washtenaw County, Michigan.

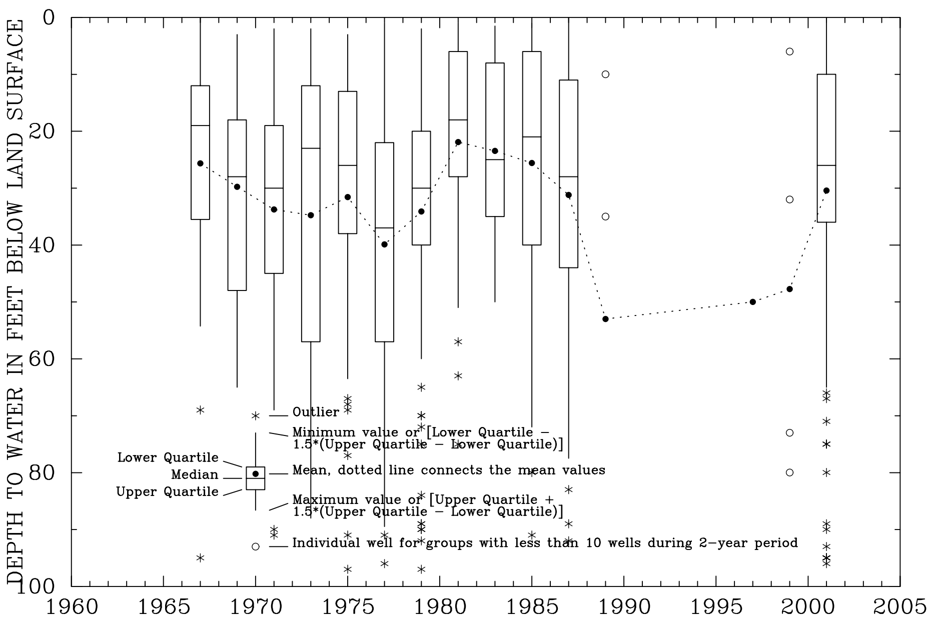

Figure C19. Glacial deposit ground-water levels reported in WelLogic database for Dexter Township, Washtenaw County, Michigan.

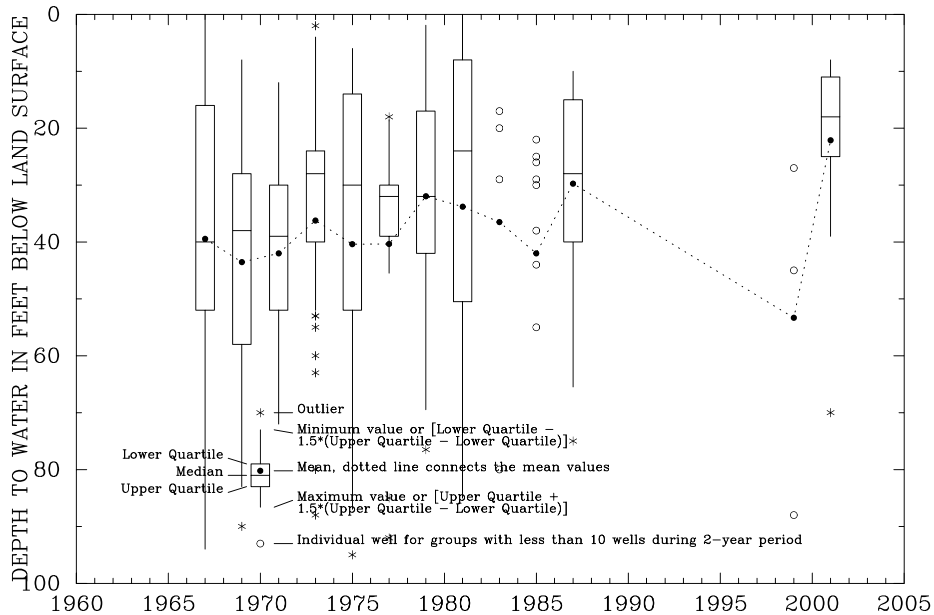

Figure C20. Glacial deposit ground-water levels reported in WelLogic database for Freedom Township, Washtenaw County, Michigan.

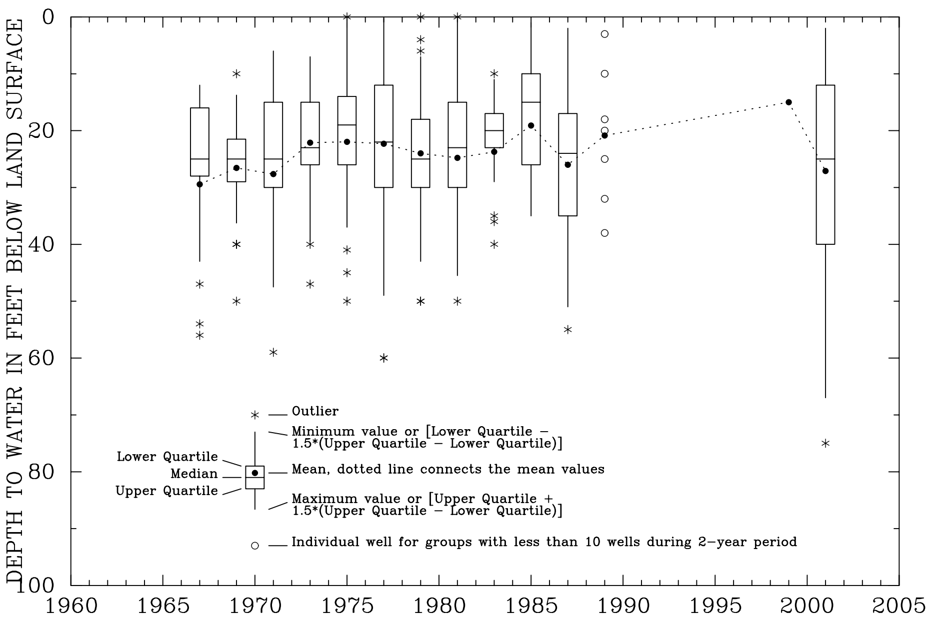

Figure C21. Glacial deposit ground-water levels reported in WelLogic database for Lima Township, Washtenaw County, Michigan.

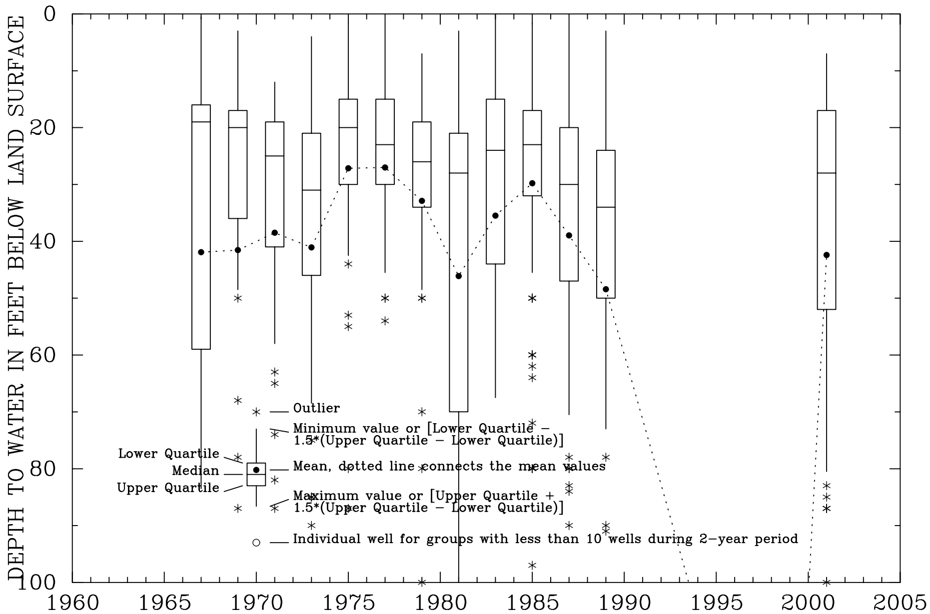

Figure C22. Glacial deposit ground-water levels reported in WelLogic database for Lodi Township, Washtenaw County, Michigan.

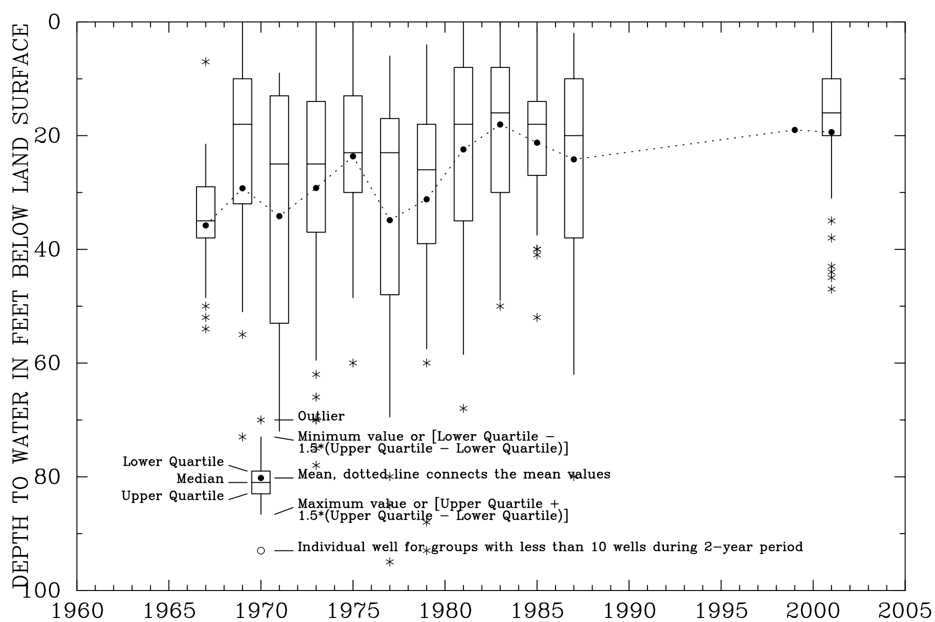

Figure C23. Glacial deposit ground-water levels reported in WelLogic database for Lyndon Township, Washtenaw County, Michigan.

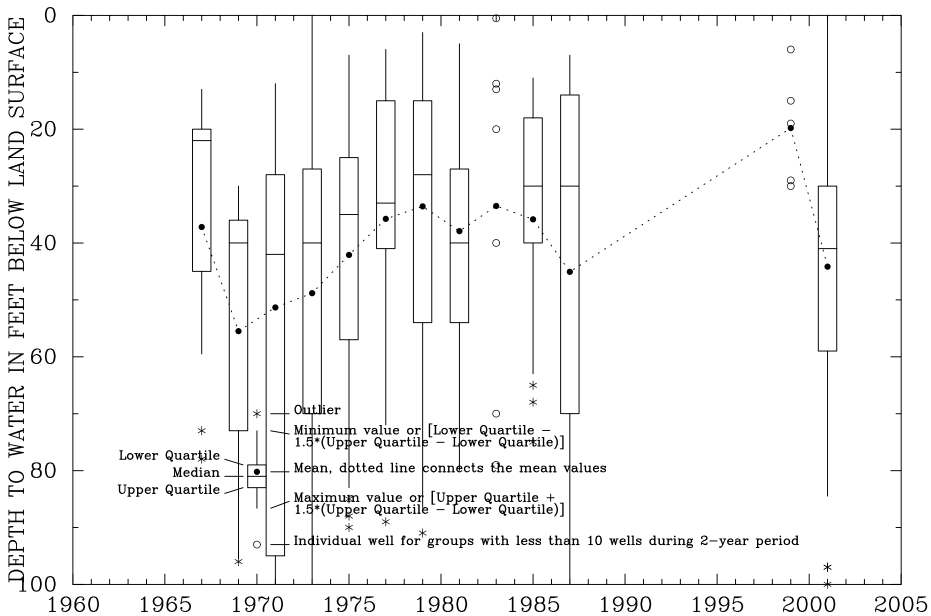

Figure C24. Glacial deposit ground-water levels reported in WelLogic database for Manchester Township, Washtenaw County, Michigan.

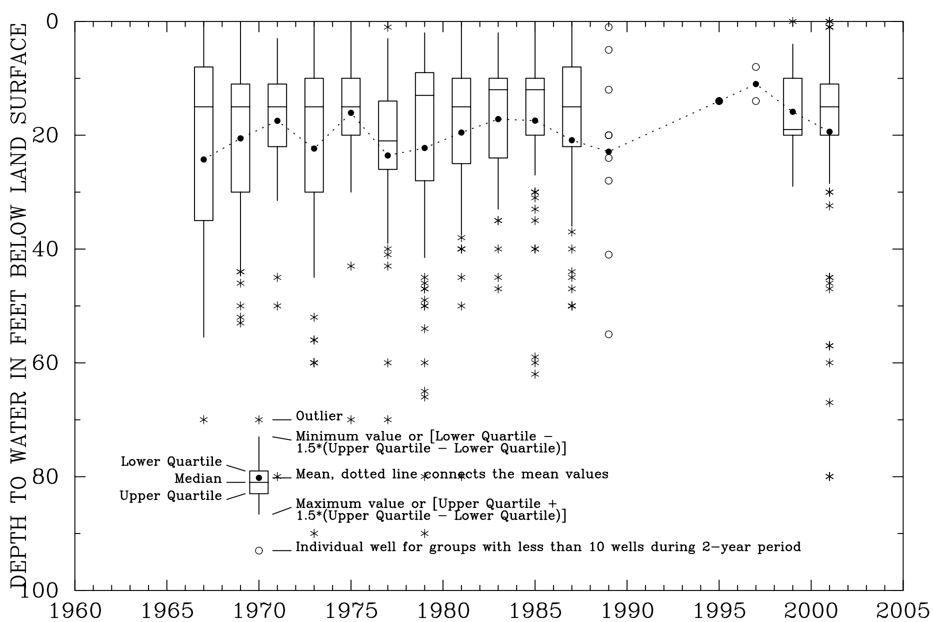

Figure C25. Glacial deposit ground-water levels reported in WelLogic database for Northfield Township, Washtenaw County, Michigan.

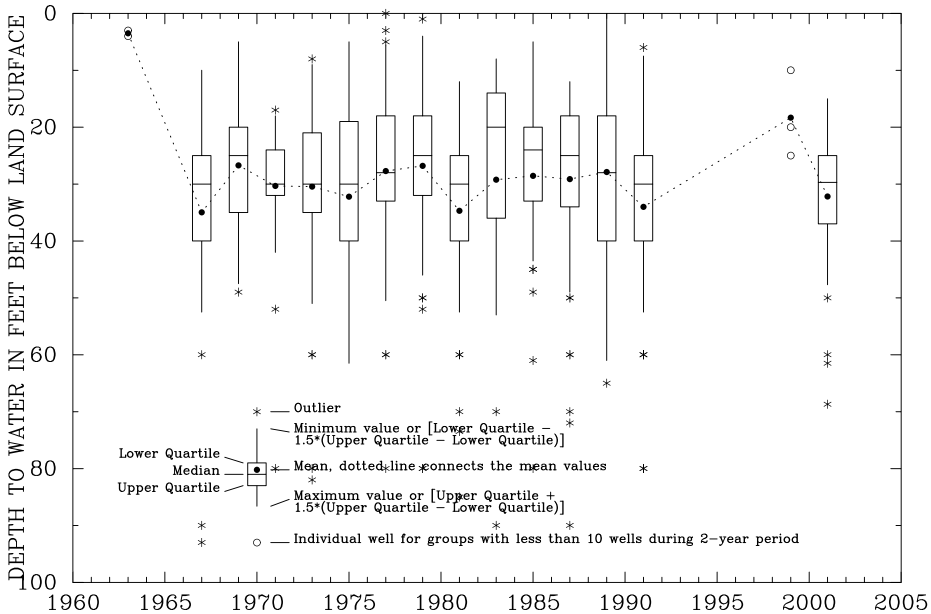

Figure C26. Glacial deposit ground-water levels reported in WelLogic database for Pittsfield Township, Washtenaw County, Michigan.

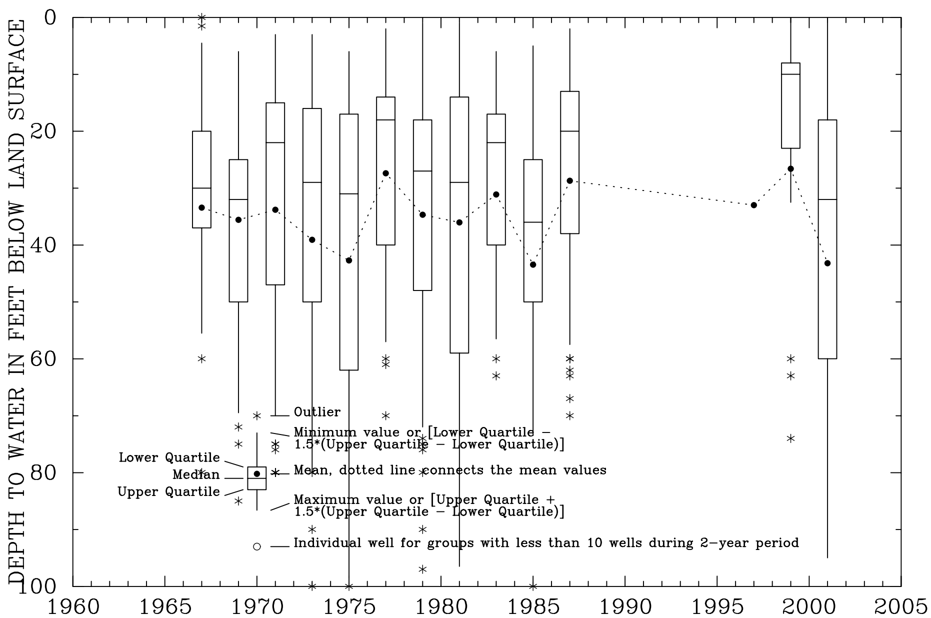

Figure C27. Glacial deposit ground-water levels reported in WelLogic database for Salem Township, Washtenaw County, Michigan.

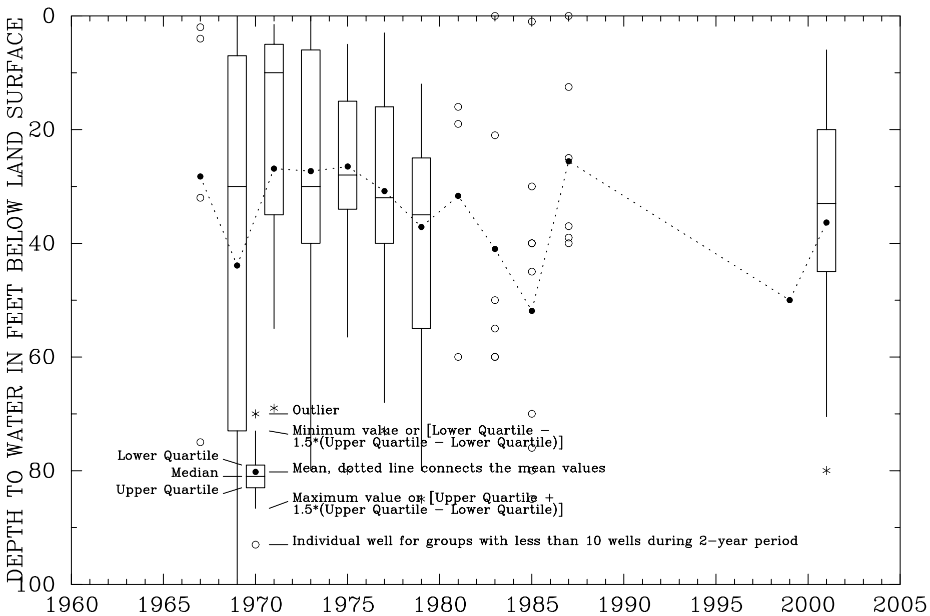

Figure C28. Glacial deposit ground-water levels reported in WelLogic database for Saline Township, Washtenaw County, Michigan.

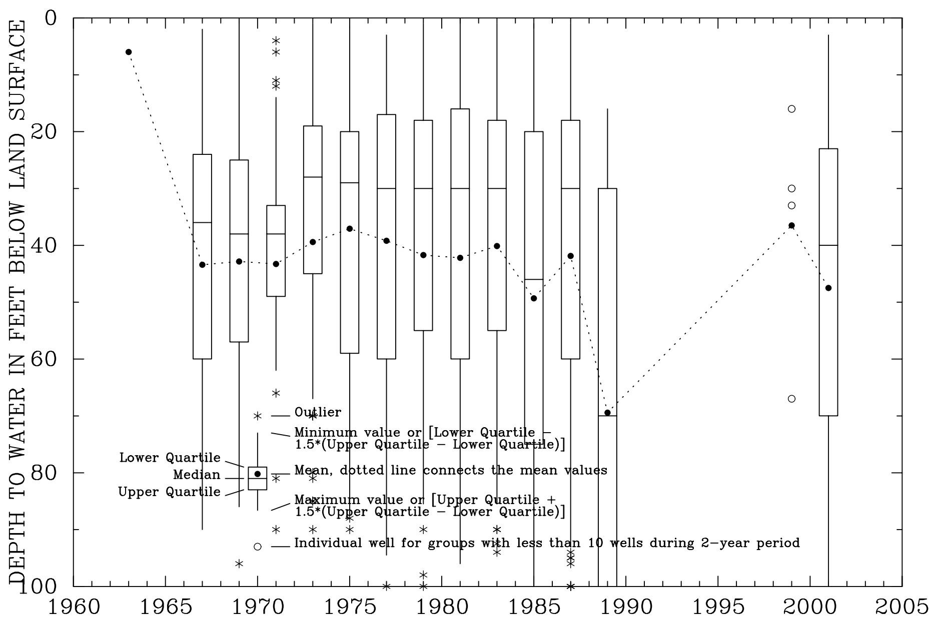

Figure C29. Glacial deposit ground-water levels reported in WelLogic database for Scio Township, Washtenaw County, Michigan.

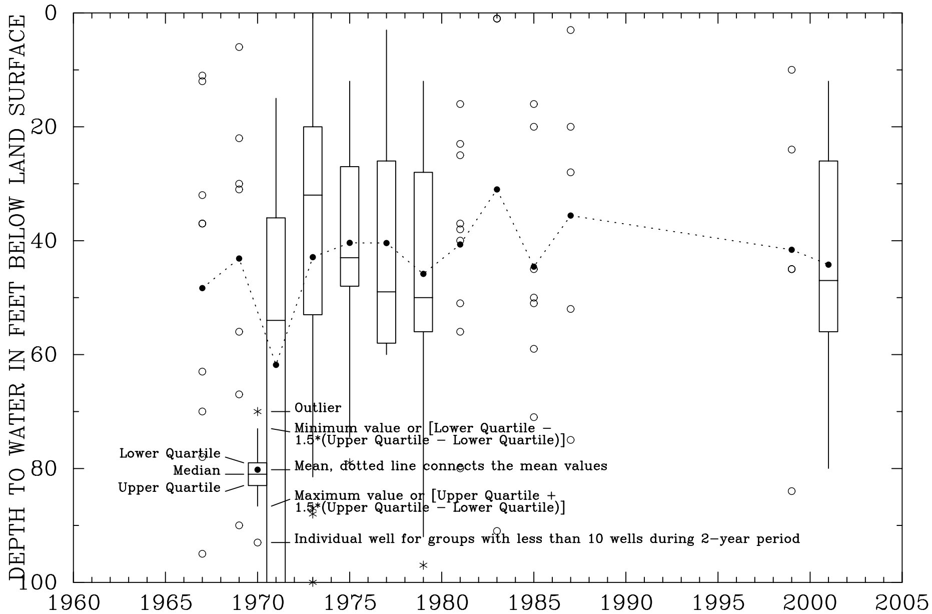

Figure C30. Glacial deposit ground-water levels reported in WelLogic database for Sharon Township, Washtenaw County, Michigan.

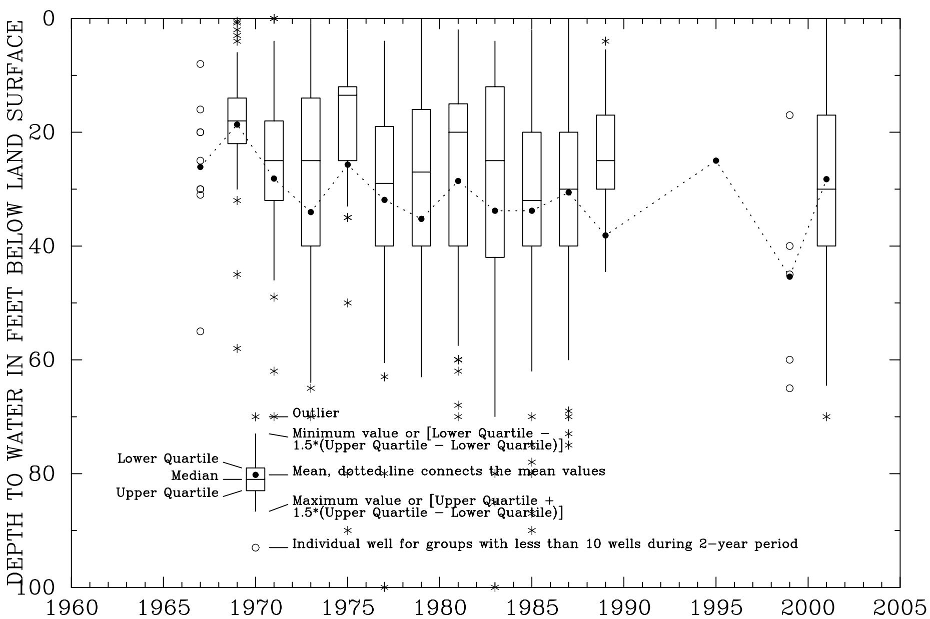

Figure C31. Glacial deposit ground-water levels reported in WelLogic database for Superior Township, Washtenaw County, Michigan.

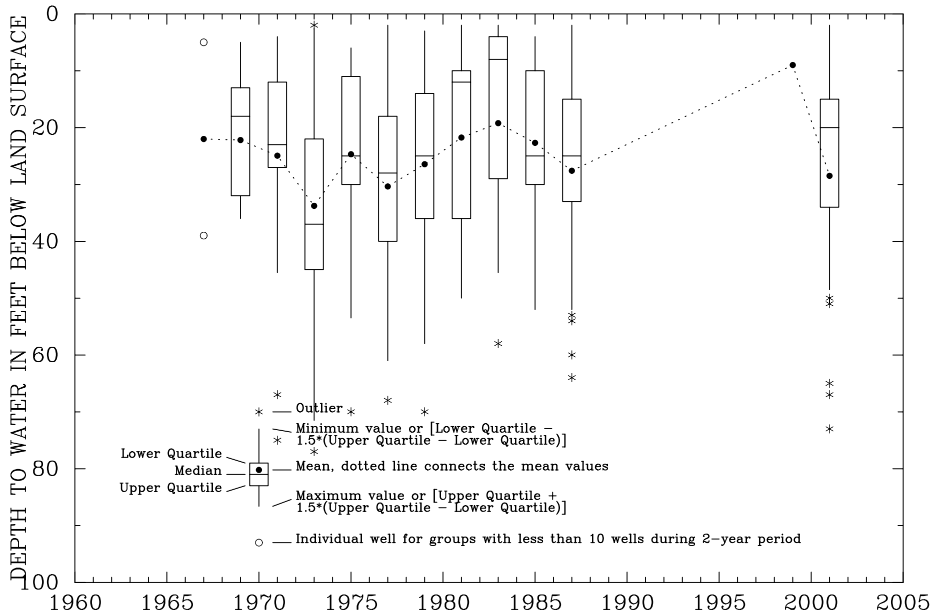

Figure C32. Glacial deposit ground-water levels reported in WelLogic database for Sylvan Township, Washtenaw County, Michigan.

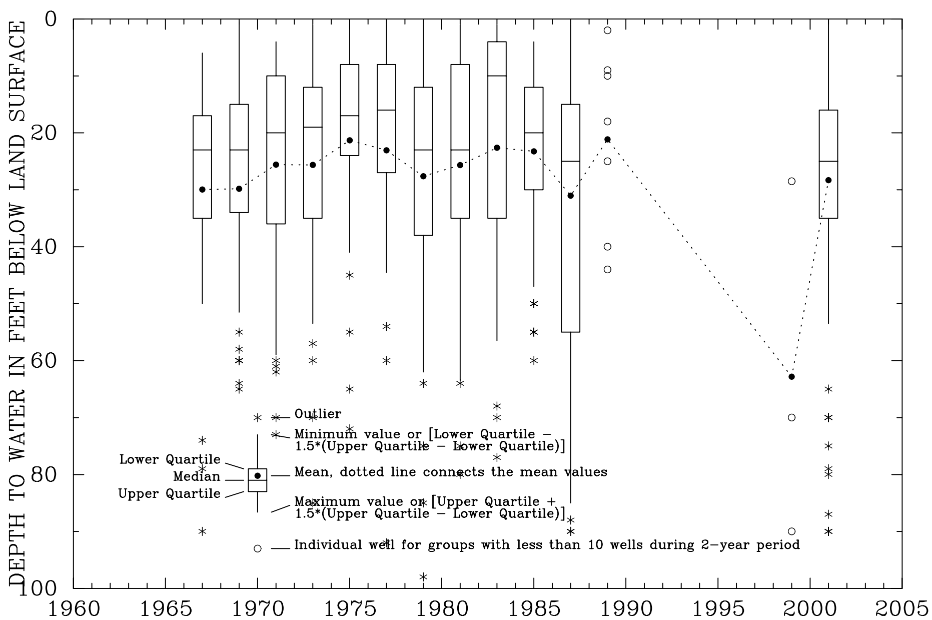

Figure C33. Glacial deposit ground-water levels reported in WelLogic database for Webster Township, Washtenaw County, Michigan.

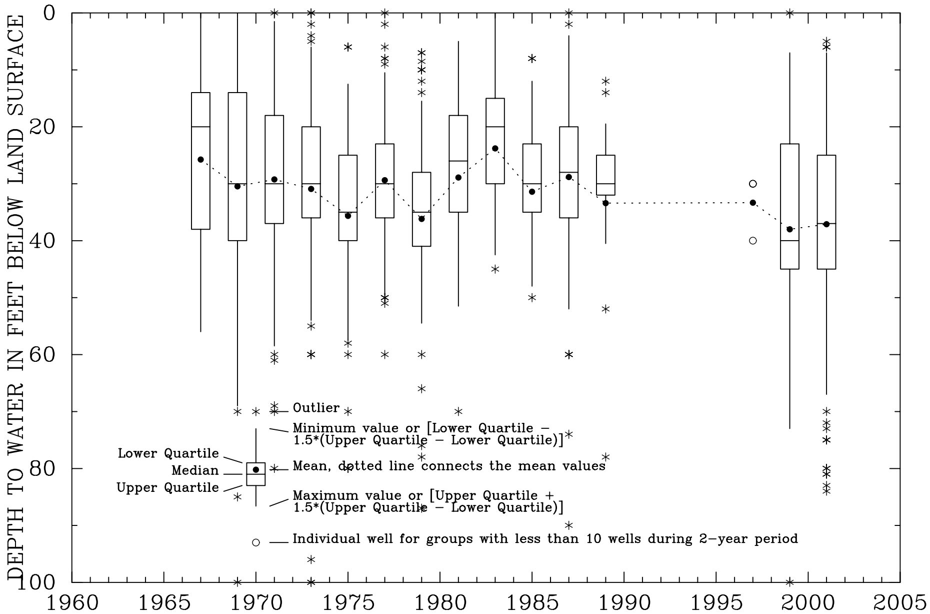

Figure C34. Glacial deposit ground-water levels reported in WelLogic database for York Township, Washtenaw County, Michigan.

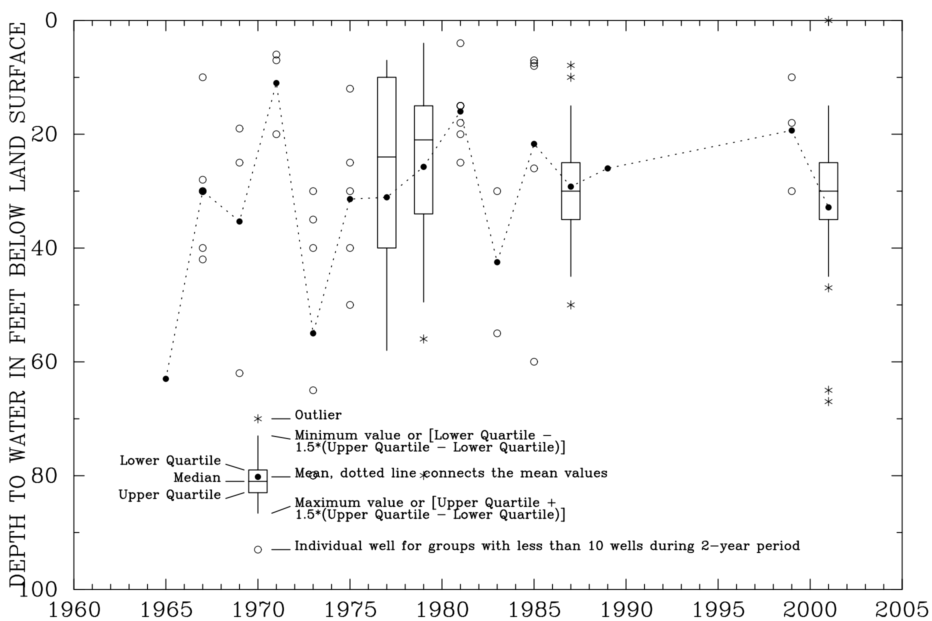

Figure C35. Glacial deposit ground-water levels reported in WelLogic database for Ypsilanti Township, Washtenaw County, Michigan.