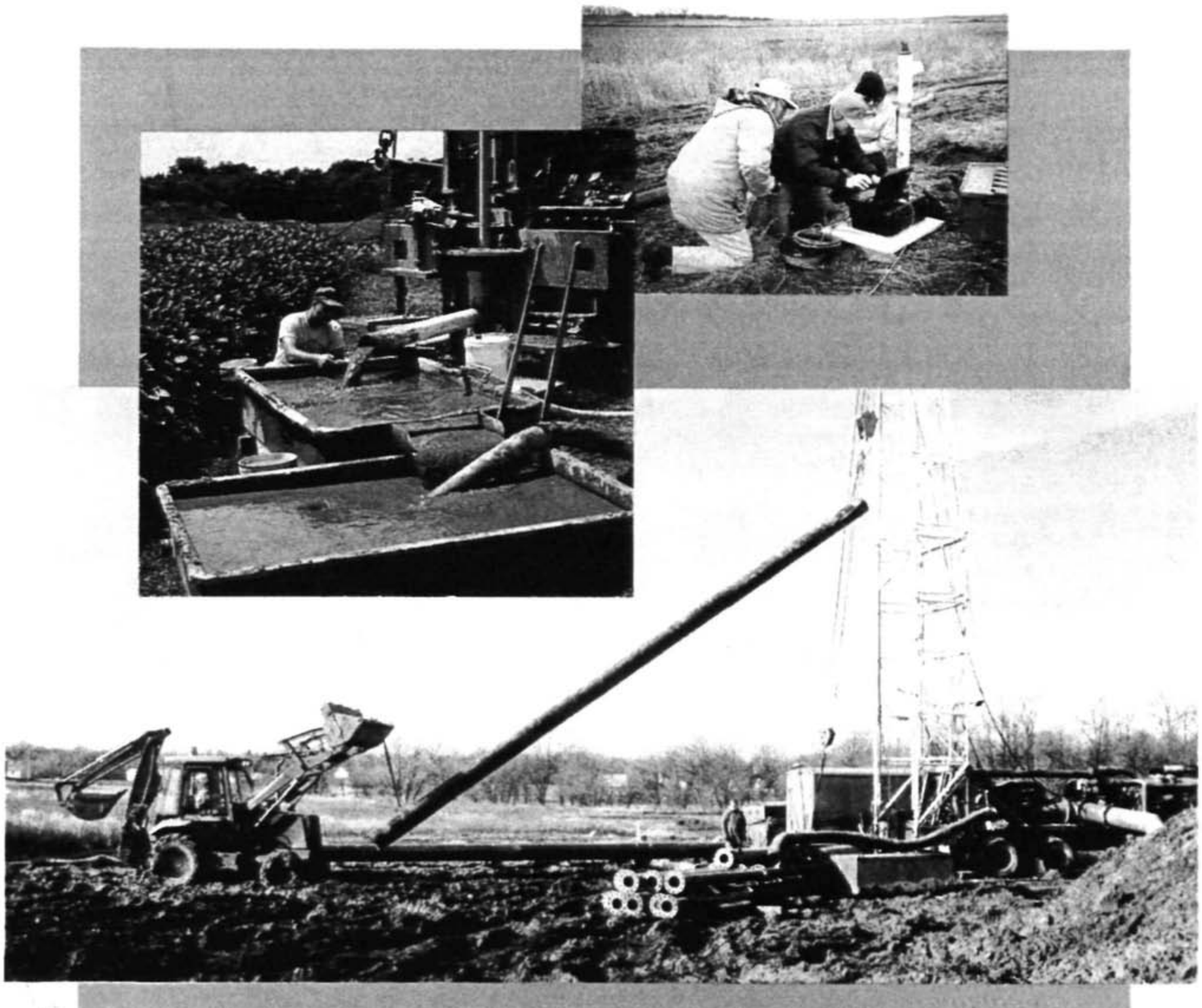
Hydrogeology and Groundwater Availability in Southwest McLean and Southeast Tazewell Counties

Part 1: Aquifer Characterization



1995
Cooperative Groundwater Report 17

Department of Natural Resources
ILLINOIS STATE GEOLOGICAL SURVEY
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Hydrogeology and Groundwater Availability in Southwest McLean and Southeast Tazewell Counties

Part 1: Aquifer Characterization

Beverly L. Herzog
Steven D. Wilson
David R. Larson
Edward C. Smith
Timothy H. Larson
Michael L. Greenslate

With contributions by

John P. Kempton
Samuel V. Panno
Keith C. Hackley
Philip C. Reed
Timothy C. Young
Stephen L. Burch
Robert D. Olson
Thomas R. Holm

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ILLINOIS STATE GEOLOGICAL SURVEY
615 East Peabody Drive
Champaign, IL 61820-6964

ILLINOIS STATE WATER SURVEY
2204 Griffith Drive
Champaign, IL 61820

Cover photos Clockwise from upper left. Collecting geologic samples during test drilling. Monitoring water levels during aquifer test. Installing well for aquifer test near Mackinaw. Well discharge at Mackinaw aquifer test: gas in the water causes the turbulence.

The appendixes cited in this volume are published under separate cover as *Hydrogeology and Groundwater Availability in Southwest McLean and Southeast Tazewell Counties—Part 1: Aquifer Characterization (Appendixes),* ISGS/ISWS Cooperative Groundwater Report 17A.

ILLINOIS DEPARTMENT OF NATURAL RESOURCES

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ABSTRACT

The confluence of the buried Mahomet and Mackinaw Bedrock Valleys, underlying southwest McLean and southeast Tazewell Counties, contains part of one of the largest sand and gravel aquifers in Illinois—the Sankoty-Mahomet Sand aquifer. This groundwater resources study was undertaken to estimate the quantity of water the Sankoty-Mahomet Sand aquifer can safely yield and to determine potential locations from which a supply of 10-15 million gallons of groundwater per day (mgd) could be developed.

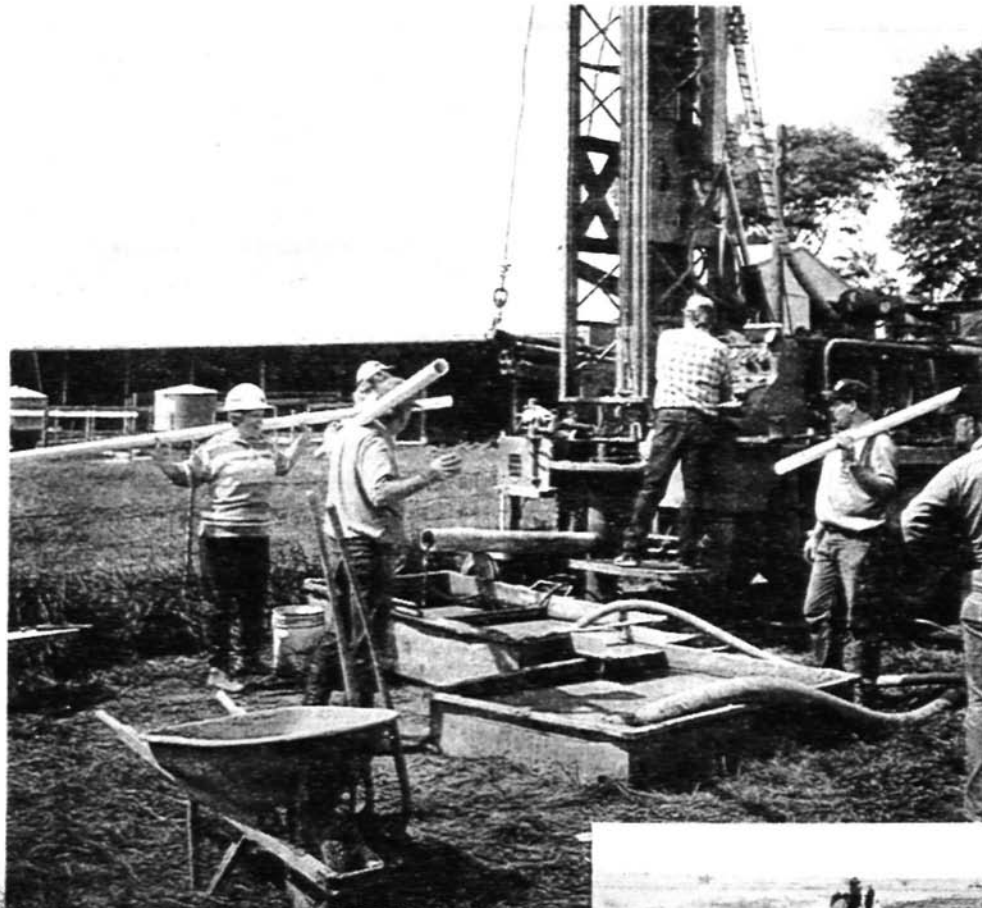
New data were generated from a surface geophysical survey, test drilling, mass measurements of water levels, aquifer tests, water quality samples, rain gauges, and stream measurements. These data show that, although the Sankoty-Mahomet Sand aquifer is more complex than previously thought, it is very prolific in this area. Shallow aquifers supplement the Sankoty-Mahomet Sand aquifer through leakage. The groundwater recharge rate to the Sankoty-Mahomet Sand aquifer is estimated at 200,000 gallons/day/square mile, or more than 50 mgd for the study area. Approximately 60% (30 mgd) of this volume is estimated to be available to wells. Thus, a new, large groundwater supply appears possible. An upcoming groundwater model presented in a subsequent report will address the area's potential for well-field development and suggest locations for a large well field.

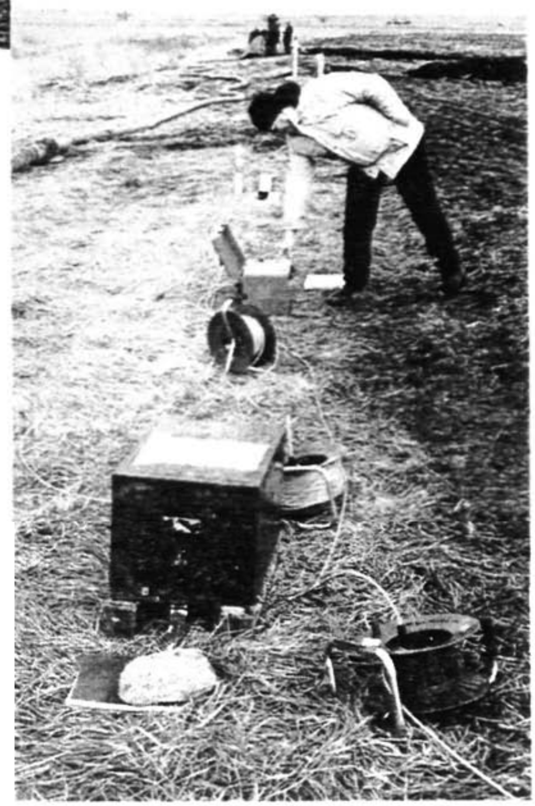
ACKNOWLEDGMENTS

This study was funded primarily by McLean County, the City of Bloomington, and the Town of Normal through the Long Range Water Plan Steering Committee, chaired by Bloomington mayor Jesse Smart. The engineering firm of Farnsworth and Wylie provided logistical assistance and surveying. Robert Kohlhase supervised these activities and critically reviewed the manuscript. Melvin Pleines, Mackinaw Valley Water Authority trustee, provided logistical help with many of the local field activities. Test drilling was done by Albrecht Drilling of Ohio, Illinois; Jet Hall was the driller. Layne Western drilled the production wells, furnished and installed test pumping equipment, and maintained steady pumping rates for the aquifer tests. Mark Graber provided access and assisted in preparing aquifer test site TST1.

Special thanks are offered to the county officials, township officials, and many citizens who generously allowed access for test drilling and for measuring water levels in private wells. Without their cooperation and interest, this field project would not have been possible.

Several staff members from the Illinois State Geological Survey and the Illinois State Water Survey participated in this project. Philip Orozco, Robert Vaiden, William Dey, and Donald Keefer helped on the seismic crew. Stephen Burch supervised test drilling. Philip Reed and Timothy Young ran geophysical logs on the test holes, measured sample moisture contents and density in the laboratory, and wrote the geophysical logging portions of this report. John Kempton contributed his expertise on the geologic framework, checked the stratigraphic interpretations on the well logs and cross sections, and reviewed the bedrock topography and Sankoty-Mahomet Sand aquifer maps. Adrian Visocky analyzed the aquifer test data and critically reviewed this manuscript. Daniel Adomatis ran the sieve analyses. Jay Sheley measured groundwater levels and river stage data, as well as collected rain gauge data. Samuel Panno, Keith Hackley, and Jack Liu collected and analyzed groundwater and gas samples, described the water chemistry of the study area, and calculated recharge from chemical data. Samuel Panno and Keith Hackley also wrote the sections of this report describing their work. Thomas Holm collected and analyzed water quality samples and wrote the corresponding parts of this report. Joseph Kamy assisted in collection of groundwater samples. Melisa McLean and Richard Rice helped with computer mapping. Robert Olson and Stephen Burch ran the aquifer tests. Mark Anliker, Brandon Lott, Anthony Romanelli, and Sean Sinclair assisted in the mass groundwater level measurements and conducted the well inventory. Amy Beckberger entered the groundwater level and river stage data into a computerized data base. Kris Klindworth set up the water well database used in the well inventory. Sean Sinclair also assisted in developing the hydrographs and aquifer test graphs. Wayne Wendland provided the precipitation data. William Roy, Richard Berg, and Ellis Sanderson served as peer reviewers for this manuscript.





(top) Installation of a 2-inch observation well. (left) Development of an observation well. (right) Set-up of southwest aqufer test. Test well is in the background.

EXECUTIVE SUMMARY

In 1993, the Illinois State Water Survey (ISWS) and the Illinois State Geological Survey (ISGS) began a 3-year study of the sand and gravel aquifers in southwest McLean and southeast Tazewell Counties to estimate the availability of groundwater. This area includes the confluence of the buried Mahomet and Mackinaw Bedrock Valleys and contains part of one of the largest sand and gravel aquifers in Illinois—the Sankoty-Mahomet Sand aquifer. The study has two goals: (1) to estimate the quantity of water the Sankoty-Mahomet Sand aquifer can safely yield, and (2) to determine locations, if any, within the study area from which a large supply of groundwater could be developed. The ideal location would provide 10 to 15 million gallons of water a day, while minimizing the impacts to existing wells.

Two major tasks must be completed to meet the study goals. The first task, characterization of the aquifer system within the confluence area, is presented in this report. The aquifer system includes the Sankoty-Mahomet Sand aquifer as well as shallower aquifers of the Banner and Glasford Formations. The second task, a computer model of the aquifers within study area, will use data generated by the first task to project future water levels and to predict the effects of a hypothetical well field at various locations within the study area. Results of the modeling effort will be published in a separate report that will provide an estimate of the safe yield of the Sankoty-Mahomet Sand aquifer and suggest potential locations for a well field that could provide 10 to 15 million gallons of water per day.

For the aquifer characterization, data from Survey files were supplemented with new data from a surface geophysical survey, test drilling, mass measurements of water levels, aquifer tests, water quality samples, rain gauges, and stream measurements. The surface geophysical survey was conducted along 45 miles of highway rights-of-way using the seismic refraction method. Data from this survey were used to update the bedrock topography map and to guide selection of sites for test drilling. Test holes were drilled into bedrock at 25 locations; 28 observation wells, including 6 in shallower aquifers, were installed. Due to insufficient aquifer thickness, no wells were installed at three sites. Water levels in the observation wells were measured biweekly to provide information on the long-term and seasonal fluctuations in water level elevations. These data were supplemented by data from mass water level measurements, which improve the accuracy of water level maps and will aid in calibrating the computer model. Water samples were collected both to provide regional water quality information and for isotope analyses, which were used to estimate aquifer recharge. Rain gauges were installed at 11 locations to assist in determining groundwater-surface water interactions and to address groundwater recharge.

Until recently, the confluence area was interpreted to be a wide, buried bedrock valley, the bottom of which is covered with sand and gravel deposits more than 100 feet thick. These deposits, called the Sankoty-Mahomet Sand aquifer, are overlain by finer grained deposits that contain interspersed sand and gravel aquifers. Geologic data from this study indicate that the Sankoty-Mahomet Sand aquifer includes two geologic units: the Sankoty-Mahomet Sand Member and a unit informally called the sub-Sankoty-Mahomet sand. Previous studies noted that the Sankoty-Mahomet Sand aquifer becomes coarser with depth; in this report, these deeper deposits are referred to as the sub-Sankoty-Mahomet sand. This distinction is irrelevant over most of the study area because the two sand units are merged. In places, however, the two sand units are separated by fine grained deposits, and elsewhere only one of the two sand units is present.

Seismic data from this project provided evidence of hills and depressions in the bedrock surface, greater in size and number than previously mapped. Instead of the sand and gravel aquifer, thick lacustrine deposits were found between three bedrock hills in the center of the study area. The most significant of the bedrock hills, located south of Hopedale (Tazewell County), interrupts the Sankoty-Mahomet Sand aquifer over several square miles. Several locations in the south-central and eastern part of the study area also have thick, lacustrine deposits that restrict the aquifer thickness. Such is the case near McLean, where the Sankoty-Mahomet Sand is absent and nearby wells are completed in the sub-Sankoty-Mahomet sand. The bedrock hills and lacustrine deposits effectively divide the Sankoty-Mahomet Sand aquifer into four regions, each containing significant areas where the aquifer is more than 100 feet thick. The aquifer is greater than 150 feet thick in two of the areas, and it is directly overlain by shallower sand and gravel aquifers for a combined thickness of more than 150 feet in another. Thus, although the Sankoty-Mahomet Sand aquifer is thinner than previously thought in parts of the study area, it is much thicker in others. Although no significant difference was discovered in the estimated volume of the Sankoty-Mahomet Sand aquifer in the confluence area, its geometry and geology are more complex than previous work indicated. This complexity limits the areas where a high capacity well field could be installed.

Local precipitation is the main source of recharge to the aquifer system. Groundwater recharge is very difficult to quantify. Calculations of groundwater recharge based on water chemistry suggest a rate of recharge to the aquifer system of 213,000 gallons/day/square mile (gpd/mi^2), which is equivalent to 4.4 inches/year. This estimate is close to the estimate of 194,000 gpd/mi^2 (4.0 inches/year), which was determined using groundwater runoff techniques. Thus, total groundwater recharge to the 264-square-mile study area is estimated at more than 50 millions gallons of water per day (mgd). Approximately 60% (30 mgd) of this volume is estimated to be available to wells. Groundwater recharge will be further evaluated in the forthcoming modeling study.

Fluctuations of water levels measured in observation wells completed in the Sankoty-Mahomet Sand aquifer mimic changes in the stage of the Mackinaw River over much of the river's reach through the study area, indicating hydraulic connection between the aquifer system and the river. A groundwater divide trends roughly southeast to northwest across the study area. North of the divide, groundwater flows north out of the confluence area into the Mackinaw Valley portion of the aquifer. South of the divide, groundwater flows toward the west.

Results from the two aquifer tests confirm the new geological data. In the southwest comer of the study area, the data indicate a prolific aquifer, but the aquifer test also confirmed that the boundary created by the bedrock high south of Hopedale will influence the long-term potential yield of the aquifer in that area. At the aquifer test site near Mackinaw, the aquifer is also substantial. Aquifer test data from this site show that the overlying sand and gravel aquifers appear to be a source of leakage into the Sankoty-Mahomet Sand.

Although the Sankoty-Mahomet Sand aquifer is more complex than previously thought, the estimated groundwater recharge rate remains essentially unchanged. In addition, the shallow aquifers supplement the Sankoty-Mahomet Sand aquifer in some parts of the study by leakance through the confining unit that separates the aquifers. The hydraulic properties of the Sankoty-Mahomet Sand aquifer show an excellent ability to transmit water and a water supply of 10 to 15 million gallons per day is still thought possible. The potential yield of the Sankoty-Mahomet Sand and locations where a high capacity well field could be developed will be evaluated in the upcoming modeling phase of the project and will be addressed in the second report.

INTRODUCTION

Background

The 1988-1989 drought in central Illinois focused attention on the need for a reliable, long-term water supply. Prior to the drought, little thought was given to a regional water supply for long-term future demands. During the drought, however, water levels in the City of Bloomington's reservoirs, Lake Bloomington and Evergreen Lake, dropped to alarmingly low levels, prompting Bloomington to institute severe water-use restrictions and to purchase supplemental water from the Town of Normal. To solve its short-term problem, Bloomington investigated two additional water sources: a supplemental groundwater supply from the vicinity of the two reservoirs and water from the Mackinaw River (Farnsworth and Wylie and Hanson Engineers 1989). A surface geophysical study indicated that insufficient groundwater was available to provide a supplementary municipal supply (Larson and Poole 1989). A side-channel pumping pool was built along the Mackinaw River in 1989. Shortly thereafter, it started discharging water into Evergreen Lake. This pumping pool remains in place for use only during emergency conditions.

After the drought, officials from Bloomington, Normal, and McLean County formed a Joint Steering Committee to investigate the feasibility of developing a regional water system to serve west-central McLean and eastern Tazewell Counties. A resulting study (Farnsworth and Wylie 1990) conducted for the Joint Steering Committee concluded that a regional supply would be both feasible and potentially beneficial to many communities. Four potential sources were identified for the regional water supply: constructing a reservoir on Panther Creek, building a pipeline to the Illinois River, building a water recycling facility, and pumping groundwater from the Sankoty-Mahomet Sand aquifer in southwestern McLean and southeastern Tazewell Counties. The first three options were not as appealing because of the costs of construction and obtaining regulatory approval. Pumping groundwater from the Sankoty-Mahomet Sand aquifer was determined to be the most viable source for a regional water supply (Farnsworth and Wylie 1990).

Purpose and Scope

In 1992, the Long Range Water Plan Steering Committee was formed by intergovernmental agreement between 18 public water suppliers and 4 water authorities. In 1993, the committee commissioned two additional, more detailed studies. The studies commissioned included a review of each community's water supply needs and a hydrogeologic investigation of the water supply potential from the Sankoty-Mahomet Sand aquifer. A report evaluating the existing and future water needs for 18 community water supplies in McLean and Tazewell Counties was completed in 1994 (Farnsworth and Wylie 1994). This study concluded that most of the rural communities in the study area have sufficient resources now and for the foreseeable future. Potential shortages were identified for Bloomington and Normal.

The Illinois State Geological Survey (ISGS) and the Illinois State Water Survey (ISWS) jointly undertook this study to evaluate aquifer characteristics and groundwater availability. The study has two goals: (1) to determine the quantity of water the Sankoty-Mahomet Sand aquifer can safely yield, and (2) to determine potential locations, if any, within the study area where a large supply of groundwater could be developed. The ideal location(s) would provide from 10 to 15 million gallons of water a day (mgd), while minimizing impacts to existing wells. The study area includes about 264 square miles of central Illinois (fig. 1). Most of the study area lies within McLean and Tazewell Counties, although it also includes a small part of northern Logan County.

Two major tasks must be completed to meet the study goals: (1) characterization of the aquifer system and (2) development of a computer model. Characterization of the aquifer system is the subject of this report.

The aquifer characterization study was primarily a field study. Data from ISGS and ISWS files were supplemented with new data collected from a surface geophysical survey, test drilling, mass measurements of water levels, aquifer tests, water quality samples, rain gauges, and stream measurements. The surface geophysical survey was conducted along 45 miles of highway rights-of-way using the seismic refraction method. Data from this survey were used to update the bedrock topography map and to guide selection of sites for test drilling. Test holes were drilled into bedrock at 25 locations. Wells were installed at 22 locations; the aquifer was essentially absent at the other three locations. Nests of two wells finished in different aquifers were installed at 6 locations, for a total of 28 observation wells.

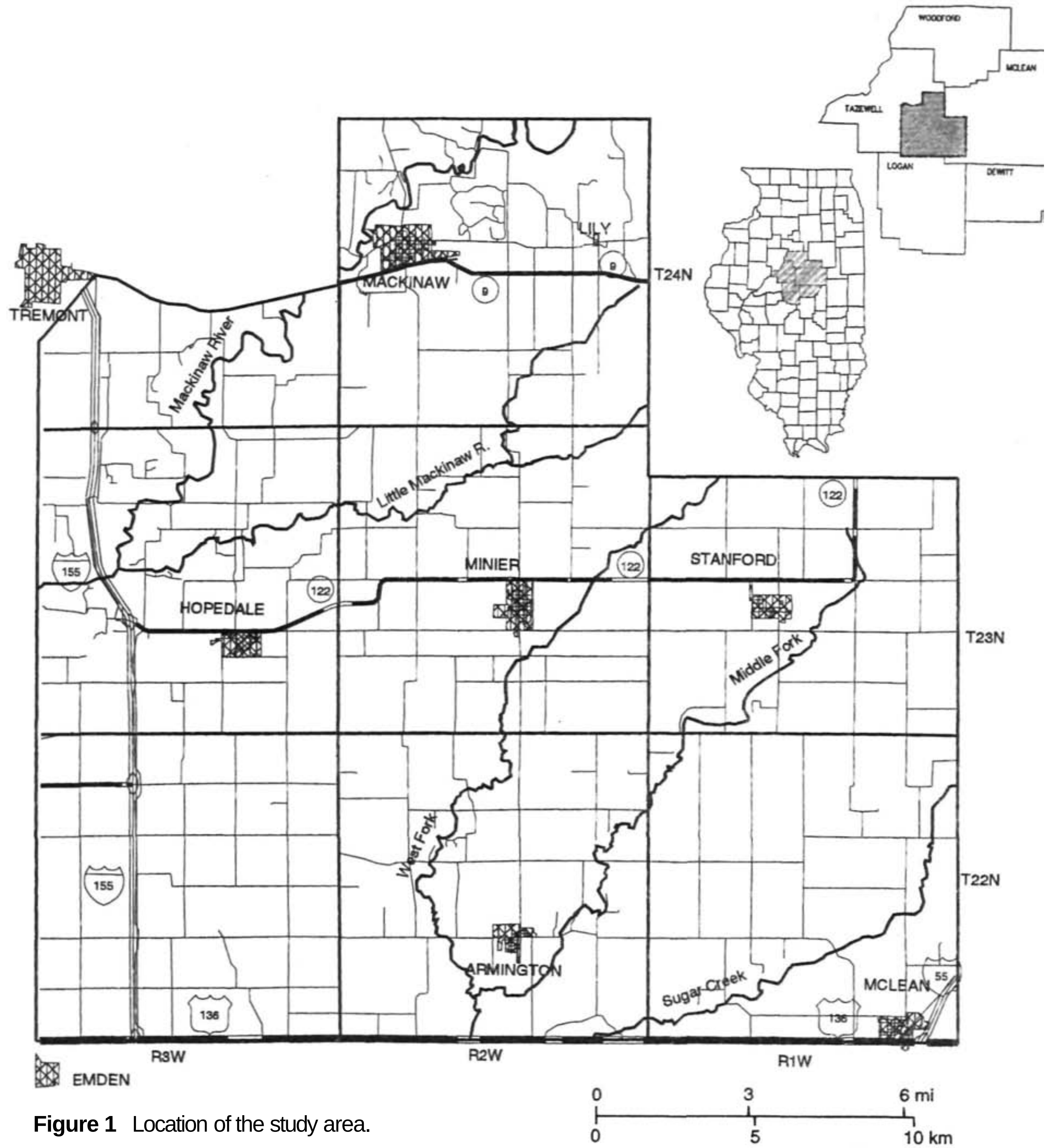


Figure 1 Location of the study area.

Water levels in the observation wells were measured biweekly to provide baseline information on the long-term and seasonal fluctuations in potentiometric surfaces of the aquifers. These data were supplemented by data from mass water level measurements. The resulting water level maps were used to determine groundwater flow directions and gradients as well to delineate recharge and discharge areas.

Water samples were collected both to provide water quality information for the study area and to allow recharge calculations from concentration gradients. Rain gauges were installed at 11 locations to assist in determining groundwater-surface water interactions and to address recharge to the aquifer system.

The data gathered for this project fostered a dramatically improved understanding of the hydrogeology and groundwater flow system in the study area. The data will be used in a computer model of the aquifer system that will simulate groundwater flow. The model will be used to identify the effects of various pumping scenarios at several different locations within the study area to estimate the magnitude of drawdowns caused by additional pumpage. It will also be used to evaluate what areas, if any, could support a withdrawal of 10 to 15 mgd without causing adverse impacts on the aquifer. Results of the modeling effort will be published in an upcoming final report.

This report characterizes the nature and occurrence of the sand and gravel aquifers within southwest McLean and southeast Tazewell Counties. It is a comprehensive source of information on the aquifer characteristics and groundwater resources in the study area.

OCCURRENCE AND QUALITY OF GROUNDWATER

General Concepts

Groundwater occupies the pore spaces found between particles of earth materials that comprise glacial sediments, as well as the fractures found in bedrock and in some glacial clays. Groundwater moves through these openings. Nearly all geologic materials will transmit water, but at different rates; the rate of movement depends on the size, homogeneity, and interconnection of the pore spaces (permeability) of the material. Earth materials are classified as aquifers or confining units (aquitards) on the basis of water transmission. An aquifer is a body of saturated earth materials that yields sufficient quantities of groundwater to a well for its intended use. Examples of aquifers include saturated sand and gravel, fractured and jointed carbonate bedrock, or sandstone. Deposits of diamicton (an unsorted and unstratified mixture of clay, silt, sand, gravel, and boulders), clay, shale, or other fine grained sediments form confining units, which restrict the flow of groundwater into or out of adjacent aquifers because water moves much more slowly through these materials.

Aquifer Characteristics

Aquifers are identified as artesian (confined), leaky artesian (semi-confined), or water table (unconfined) (fig. 2). Artesian conditions exist where the groundwater in the aquifer is under pressure greater than atmospheric pressure. This pressure buildup occurs where aquitards overlie and underlie an aquifer and impede the vertical movement of water into and out of it. The aquitards thus "confine" the aquifer. The water level in a well that is screened in an artesian aquifer will rise to a level (called the potentiometric surface) above the top of the aquifer. "Leakance" refers to the degree to which an aquitard allows the flow of water into or out of the artesian aquifer. Where there is flow of water vertically across an aquitard into or out of the aquifer, an artesian aquifer is described as leaky artesian. Water table conditions exist where an aquifer overlies an aquitard and the top of the

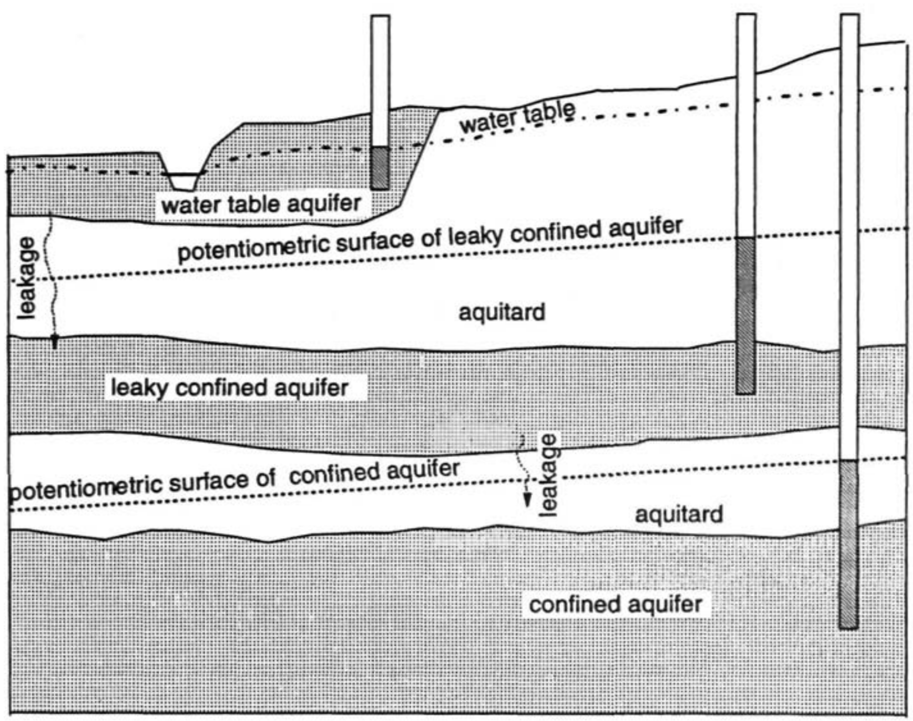


Figure 2 Types of aquifers.

aquifer is the top of the saturated zone or water table (i.e., the potentiometric surface is at atmospheric pressure). These water table conditions typically occur in surficial deposits near the land surface. The water table can be directly connected to rivers, lakes, streams, and other surface water bodies. Unconfined conditions can also occur if the water level in a confined aquifer drops below the top of the aquifer. In general, the water level in a well screened in an unconfined aquifer closely approximates the water table.

Hydraulic Properties of Aquifers and Aquitards

The hydraulic properties of an aquifer that describe its capability to transmit groundwater and take it into or release it from storage are hydraulic conductivity (K), transmissivity (7), and storage coefficient (S). These properties, and aquifer geometry, determine the response of water levels in an aquifer to pumpage. This response is determined by measuring water levels in wells. Vertical hydraulic conductivity *(K')* and thickness *(m')* are the significant hydraulic properties that characterize the rate of leakage through aquitards. All of these properties affect the ability of a confined aquifer to yield water. The sustained yield (amount of groundwater continuously available for use) of a confined aquifer depends not only on the hydraulic properties of the aquifer and aquitards surrounding it, but also on the leakage of water through the aquitards.

Hydraulic conductivity (K) is a coefficient of proportionality that describes the ease with which groundwater can move through permeable earth materials. The magnitude of hydraulic conductivity depends on the size and connection of pore openings in the sediment as well as the nature and extent of secondary permeability features (e.g., fractures). Aquifers have relatively high hydraulic conductivities while aquitards have relatively low hydraulic conductivities. Values of hydraulic conductivity for sand and gravel (aquifer materials) range from about 10 to 10^5 gallons per day per square foot (gpd/ft^2, 5 x 10^{-4} to 5 cm/s); values for diamicton (aquitard material) range from about 5 x 10^{-6} to 10 gpd/ft^2 (10^{-10} to 5 x 10^{4} cm/s; Heath 1989).

Transmissivity (7) is a measure of the capacity of an aquifer to transmit water. It equals the hydraulic conductivity (K) of the aquifer multiplied by the aquifer thickness (m).

The storage coefficient (S) describes the capacity of an aquifer to take into or release groundwater from storage. It is defined as the volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in the elevation of the potentiometric surface (or "head") (Heath 1989). For an artesian aquifer, this volume of water is derived from the slight expansion of the water resulting from the release of artesian pressure at a pumping well plus the compression of the aquifer due to the weight of the overlying materials. Typical values of the storage coefficient for confined aquifers range from 10^{-5} to 10^{-3} (Heath 1989). For a water table aquifer, the volume of water stems mostly from the gravity drainage or the refilling of the pore spaces in the sediments through which the water level change occurs. A relatively insignificant amount of the water volume comes from the expansion of the water and the compression of the aquifer. The storage coefficient for an unconfined aquifer is also referred to as the specific yield and typically ranges between 0.1 and 0.3 (Heath 1989).

Vertical hydraulic conductivity *(K')* is the rate of flow of water vertically through a horizontal unit area of aquitard under a unit vertical hydraulic gradient. Vertical hydraulic conductivity may vary locally by several orders of magnitude due to the variability of glacial deposits. Leakage through an aquitard is dependent upon the vertical hydraulic conductivity and thickness of the confining bed (m') plus the hydraulic gradient across the confining unit. The leakage coefficient, also called leakance, is the ratio *K/m'.* It describes the quantity of water moving through a confining bed into or out of an aquifer. Vertical leakage through a confining bed can be significant in determining the long term yield of a confined aquifer.

Groundwater Recharge and Discharge

Groundwater recharge is the addition of water to the zone of saturation. It is one of the most important factors to be considered when determining aquifer yield, but it is also one of the most difficult components to quantify. Several factors control the rate of groundwater recharge. Among these are the hydraulic characteristics, geologic characteristics, thickness, and distribution of the subsurface materials both above and below the water table; topography; land use; vegetation; soil moisture content; depth to the water table; intensity, duration, areal extent, and seasonal distribution of precipitation; type of precipitation (rain, snow, etc.); and air temperature (Walton 1962).

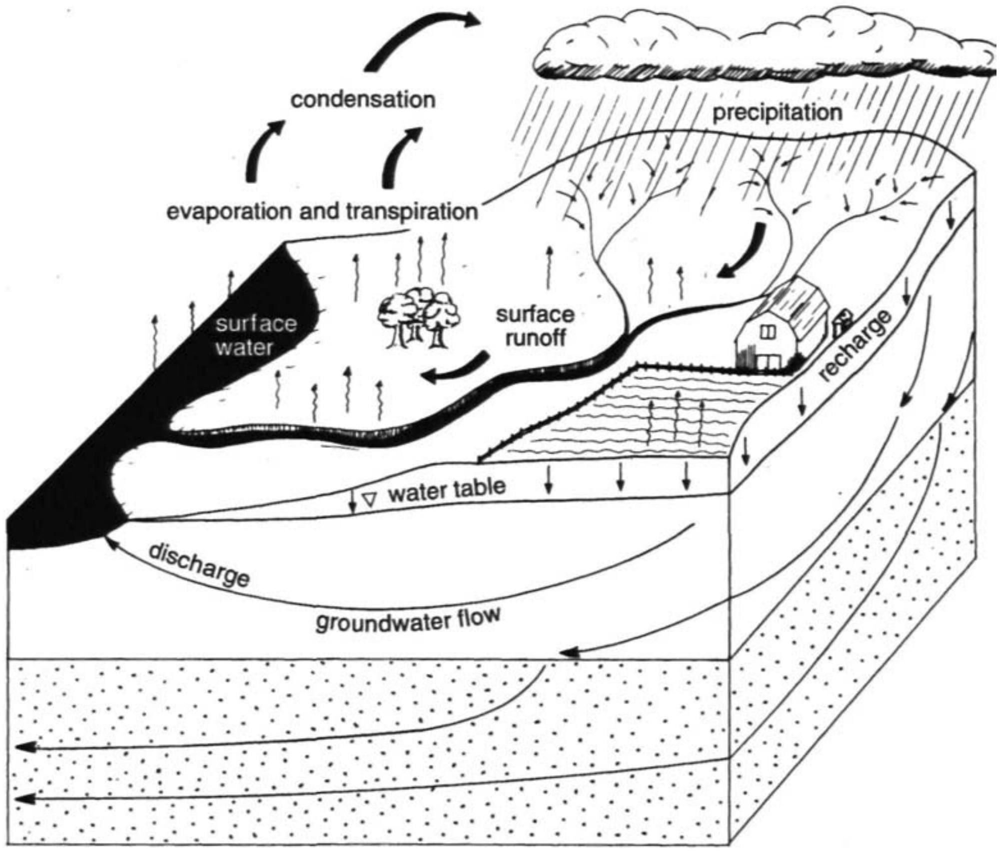


Figure 3 The hydrologic cycle.

The sand and gravel aquifers in the confluence area are principally recharged by precipitation (fig. 3). Part of the precipitation that falls to the ground returns directly to the atmosphere through evaporation, part of it runs off into streams, and the rest percolates into the soil. Central Illinois receives approximately 38 inches of rain per year, with 32 inches percolating into the soil (Lamb 1994). Most of the soil water evaporates directly or is transpired by plants. Water in excess of the maximum retention capacity of the soil infiltrates downward through pore spaces to the zone of saturation where it enters the groundwater flow system. Shallow, unconfined aquifers are recharged by direct infiltration of precipitation, melting snow, or runoff in spring, early summer, and fall, when the volume of soil moisture is greater than evapotranspiration requirements. Typically, confined aquifers are recharged more slowly by leakage through the fine grained sediment of the confining beds. Hensel (1992) gives a more in-depth description of natural recharge in Illinois.

Groundwater moves from areas having higher hydraulic head (recharge areas) to areas having lower hydraulic head (discharge areas), as a result of differences in pressure and gravity. This movement is slow, typically on the order of tens to several hundreds of feet per year. The rate of groundwater movement is governed by the hydraulic conductivity of the material through which the water moves and by the prevailing hydraulic gradient. Groundwater eventually discharges to springs, seeps, and stream channels, or to wells. Changes in the volume of water stored in an aquifer are caused by changes in the rate of recharge to or discharge from the aquifer. The changes in volume induce water level fluctuations in wells completed in that aquifer. Earthquakes, changes in atmospheric pressure, and aquifer loading may generate short-term water level fluctuations, particularly in wells screened in a confined aquifer.

In areas where geologic data are limited, determining the size, extent, hydraulic properties, and water levels for shallow aquifers is difficult. However, it is essential to developing an accurate estimate of groundwater recharge. Historically, it has been assumed that water reaching an aquifer as recharge is balanced by groundwater discharge to streams (i.e., baseflow) and withdrawals from pumpage.

Aquifer Yield and Response to Pumping

Aquifer parameters (e.g., hydraulic conductivity, transmissivity, and storage coefficient), together with water level data, are used to estimate an aquifer's water-yielding capability and its response to groundwater pumpage. The sustainable yield of a well can be assessed from the hydraulic properties of the aquifer, measurements of water levels in wells, and well construction data.

Pumping a well completed in a water table aquifer lowers the water level in the well (drawdown) and causes the water table around the well to decline (fig. 4). The lowered water table indicates that the aquifer around the well is no longer saturated. Pumping a well completed in a confined aquifer produces a drawdown of the water level in the well and reduces the pressure on the groundwater around the well (fig. 4). The aquifer remains saturated so the water level in the well remains above the top of the aquifer. Water table conditions can be produced in an artesian aquifer if a well is

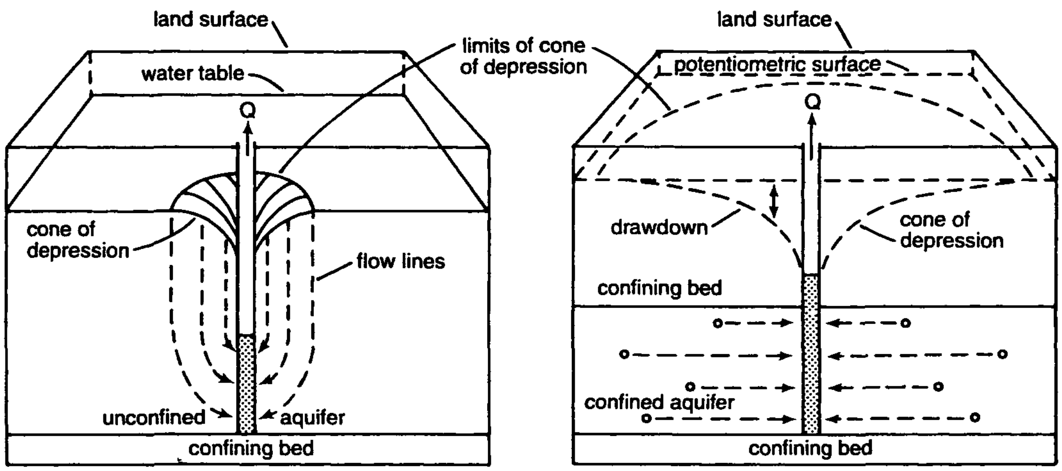


Figure 4 Flow to a well in an unconfined and a confined aquifer (adapted from Heath 1989).

pumped at a rate sufficient to lower the water level in the well to an elevation below the top of the aquifer. The aquifer around the well will dewater as in the case of a water table aquifer.

Groundwater flows to a pumping well, thereby creating an area of reduced hydrostatic pressure around the well (fig. 4). This reduction results in a decline in the potentiometric surface of the aquifer around the pumping well in the form of an inverted cone (called the cone of depression) with the well at the center (fig. 4). The amount of drawdown in a well pumping at a given rate plus the size and shape of the cone of depression depend on the amount of groundwater available to the well. This is determined by the hydraulic properties of the aquifer.

Water Quality

Groundwater contains dissolved mineral matter acquired initially as precipitation falls through the atmosphere and later as the water moves through the soil and sediments underlying the land surface. Dissolved minerals vary in type and concentration due primarily to the composition and solubility of the various deposits that the water encounters and the length of time that the water is in contact with these deposits. Temperature, pressure, and pH also affect dissolution rates and concentrations of the dissolved minerals. Longer residence time in the subsurface generally results in greater concentrations of dissolved minerals in groundwater. The suitability of groundwater for various uses depends mostly on the type and concentration of the various dissolved chemical constituents present and, to a lesser extent, on the physical properties of the water (e.g., temperature, color, and turbidity).

PREVIOUS STUDIES

The Mahomet and Mackinaw Bedrock Valleys (fig. 5) were first mapped, and their contents described, by Horberg (1945). Horberg refined the outline of the valleys as part of his statewide mapping of the bedrock topography (Horberg 1950) and studied the glacial deposits within the valleys (Horberg 1953). The Mahomet Valley was viewed as the major drainageway in east-central

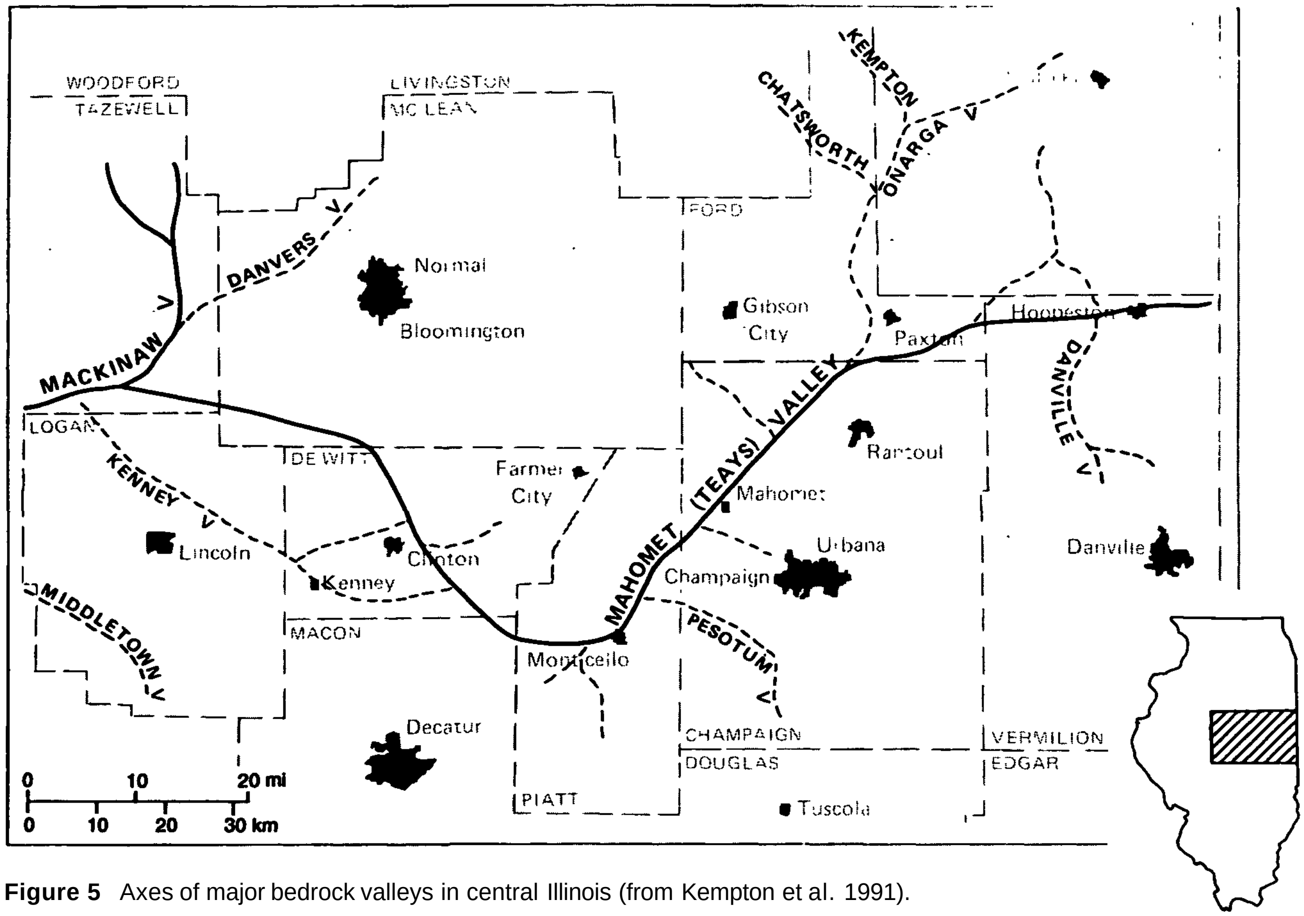


Figure 5 Axes of major bedrock valleys in central Illinois (from Kempton et al. 1991).

Illinois during preglacial times, carrying large volumes of glacial meltwater from the east. This meltwater filled the lower part of the valley with thick sand and gravel deposits, which he defined as the Mahomet Sand. The Mackinaw Bedrock Valley merges with the Mahomet Bedrock Valley in southeast Tazewell County. This area is referred to as the "confluence area" in this report. The Mackinaw Bedrock Valley is also filled with significant sand and gravel aquifers, which Horberg defined as the Sankoty Sand (Horberg 1953). Because of continued interest in the valley system, Horberg's regional framework and bedrock topography map were updated in two recent publications. Kempton et al. (1991) studied the geologic framework of the sediments within the Mahomet Bedrock Valley. Herzog et al. (1994) updated Horberg's bedrock topography map for the entire state.

Since their discovery, the Mahomet and Mackinaw Bedrock Valleys have generated great interest because of the significant sand and gravel aquifers they contain. Most of this work has concentrated on the Mahomet Bedrock Valley. Early work focused on regional groundwater geology (Foster 1953, Selkregg and Kempton 1958, Stephenson 1967) and groundwater hydrology (Visocky and Schicht 1969). Local groundwater availability studies were done for Champaign County (Smith 1950, Foster and Buhle 1951, Sanderson and Zewde 1976), Coles County (Foster 1952), and Piatt County (Sanderson 1971). Reports on public water supplies were published for the state (Hanson 1950, Visocky et al. 1978), Champaign County (Woller 1975), and several other municipalities (Visocky et al. 1978, Wehrmann et al. 1980). Included in the report by Visocky et al. (1978) were assessments of the adequacy of the aquifers used for 36 public water supplies in McLean and Tazewell Counties. All of these studies found that aquifers in the region were adequate to meet demands, except for a marginal rating for the town of Chenoa. Additionally, groundwater flow in and recharge to the Mahomet Sand was studied by Panno et al. (1994), who developed a conceptual model of flow within and recharge to the aquifer using major ion chemistry and isotope geochemistry.

In the 1970s, groundwater geology studies were expanded into "geology for planning" studies, which included evaluations of groundwater resources and other hydrogeologic issues of public interest.

Reports were published for the Springfield-Decatur area (Bergstrom et al. 1976) and DeWitt County (Hunt and Kempton 1977).

Local studies on the confluence area were done in the mid-1960s. The first of these focused on the Havana region (Walker et al. 1965). This study area borders the west side of the current study area. Several unpublished reports were written for the Town of Normal beginning in the 1960s. Unpublished open file reports on groundwater availability were prepared by Walker (1962) and Schicht (1966), and these were followed by unpublished correspondence and maps by John Kempton of the ISGS. The Town of Normal West Well Field was developed on the basis of these documents. Use of the well field lowered water levels in nearby domestic wells and caused Normal to replace or deepen the affected wells. In a reevaluation of the earlier data, Richards and Visocky (1982) learned that the aquifer that supplies Normal's wells was connected to a shallower aquifer that supplied the private wells. The duration of the aquifer tests performed in conjunction with the development of the West Well Field was too short to show this connection, resulting in the greater-than-expected drawdowns in the domestic wells.

After the Normal West Well Field was developed, several more studies were completed for this area. When Diamond Star Motors was considering building on their present location, Kempton and Poole (1985) updated work on aquifer distribution to provide more information on groundwater availability, and the drought of 1988-1989 prompted additional studies. A geophysical exploration north of Bloomington (Larson and Poole 1989) sought potential groundwater resources that could be quickly developed between Lakes Bloomington and Evergreen. The results were not promising enough for the city to pursue this option. Kempton and Visocky (1992) evaluated existing data for an area covering parts of DeWitt, Logan, McLean, and Tazewell Counties to make a preliminary assessment of the available groundwater resources for a regional groundwater supply. They concluded that the greatest potential for a regional supply was from the Sankoty-Mahomet Sand aquifer in southeast Tazewell and southwest McLean Counties (fig. 6). The current study area was selected on the basis of this recommendation.

A report on the confluence area of the Mahomet and Mackinaw Valleys was published within the past year (Wilson et al., 1994). The area covered in this report includes the current study area and extends 6 miles west and 12 miles south. Wilson et al. (1994) reported on the characteristics of the Sankoty-Mahomet Sand aquifer and listed several significant changes in the interpretation of its geologic configuration and groundwater flow system. They mapped a much wider bedrock channel between Mackinaw and Danvers, a bedrock surface elevation of 400 feet near Hopedale, and the presence of lacustrine material that limited the extent of the aquifer near Armington. They also mapped an east-west-trending groundwater flow divide in the Sankoty-Mahomet Sand aquifer, indicating flow both to the north and southwest out of the study area.

Two current studies that cover all or parts of the study area are nearing completion. Both focus more on areas east of the study area, but this project has benefitted from both. The ISGS is completing a study entitled "Geologic Assistance for Siting Solid Waste Disposal Facilities" for McLean County. The ISGS and the U.S. Geological Survey, under a federal-state cooperative geologic mapping program, are jointly conducting a three-dimensional mapping project of the 1:100,000-scale Champaign Quadrangle. Although the project does not directly include any of Tazewell County, it has a "buffer area" that extends beyond the edges of the quadrangle to improve the quality of the mapping along the edges of the study area. The buffer area includes the study area, thereby improving mapping accuracy along the western boundary of the Champaign Quadrangle and along the eastern boundary of the current study area. For the areas where these three study areas overlap in southwest McLean County, the three projects have been coordinated to provide a consistent geologic interpretation. In particular, the stratigraphic investigation performed for the Champaign Quadrangle improved the stratigraphic interpretations for this report.

As part of the current study of the confluence area, Holm (1995) sampled the observation wells and analyzed groundwater quality, focusing on naturally occurring arsenic. This water quality study concluded that the groundwater quality is generally good. No evidence of industrial or agricultural contamination was detected, and levels for hardness, iron, sulfate, and chloride were typical for groundwater in Illinois. Although arsenic was found in the groundwater over most of the study area, levels were less than the current standard of 50 µg/L (micrograms per liter) for public drinking water. If the U.S. Environmental Protection Agency (U.S. EPA) standard for arsenic is lowered to proposed levels of between 8 and 20 µg/L (B. Kaiser, U.S. EPA, pers. comm.,1995), much of the groundwater in the area will have arsenic concentrations greater than the new standard. Holm (1995) suggested

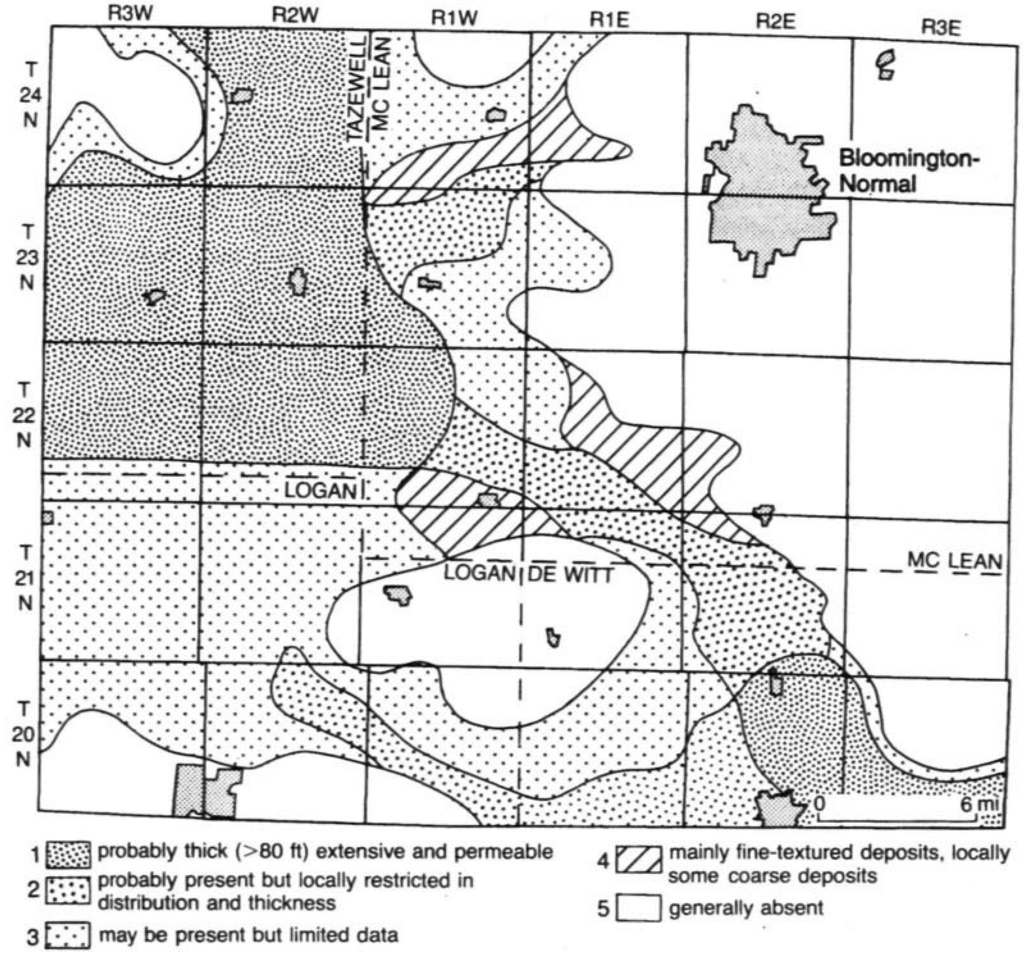


Figure 6 Occurrence of the Sankoty-Mahomet Sand in western McLean and eastern Tazewell Counties (from Kempton and Visocky 1992).

that any development of the aquifer for public water supplies be directed toward areas of low arsenic concentration.

REGIONAL SETTING

Location

The study area includes about 264 square miles of central Illinois (fig. 1). Most of the study area lies within McLean and Tazewell Counties, although it also includes a small part of northern Logan County. Six communities are located in the largely rural study area. These are McLean and Armington in the southern part of the study area; Stanford, Minier, and Hopedale in the central part; and Mackinaw in the north. Tremont is just west of the northwest comer and Emden is just south of the southwest comer of the study area.

Climate

The climate of central Illinois is humid continental (Koeppe 1935), characterized by warm and wet summers, cold and relatively dry winters, and wide fluctuations in both temperature and precipitation. The inconstancy of the weather patterns arises from the interactions of air masses moving across Illinois from the polar region, the Pacific Ocean, and the Gulf of Mexico.

As is characteristic of a continental climate, the occurrence and distribution of precipitation varies seasonally. Spring precipitation typically is more widespread than summer precipitation, which is mostly in the form of thunderstorms or larger storm cells associated with frontal systems. Most of the annual precipitation in the study area falls during the spring and summer months (fig. 7).

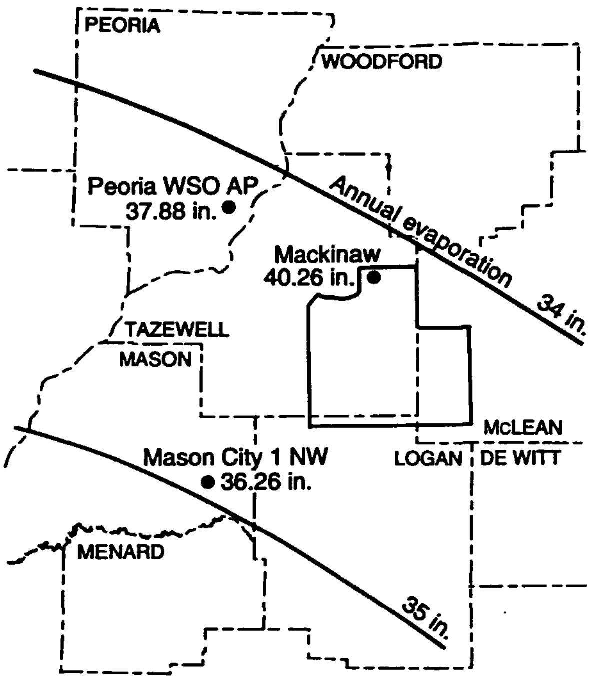


Figure 7 Average annual precipitation and evaporation for the 1970-1994 period of record (data from 1989 are excluded).

Precipitation data from 1970 to 1994 are available for three weather service stations located near the perimeter of the study area (table 1). Only the Mackinaw station is within the study area. These data suggest the wide variability of precipitation over time. Because the monthly precipitation total for April 1989 was missing from the Mackinaw record, values for all of 1989 were not included in the calculated averages for Mackinaw. For this report, the 1989 data were excluded from the Peoria WSO AP and Mason City 1 NW records to provide averages comparable to those for Mackinaw. Because 1989 was a dry year, excluding the 1989 data increases the 1970–1994 average annual precipitation for Peoria WSO AP by 0.62 inch and for Mason City 1 NW by 0.35 inch.

Average annual precipitation for 1970-1994 varied from 40.26 inches at Mackinaw to 36.26 inches at Mason City 1 NW (fig. 7), a difference of 4.00 inches. About two-thirds of the 1970–1994 annual precipitation arrived between April and September. These are the months of plant germination and growth as wel! as the time of peak evapotranspiration. The difference in average 1970–1994 April-September precipitation was 1.99 inches (table 1). The range in April-September precipitation for the three stations extended from 44.14 inches in 1993 to 7.15 inches in 1988; both of the values for these extremes were from the Mackinaw station. Rain or snow falling during the other 6 months of the year can have a greater effect on groundwater recharge than does April-September precipitation. From October through March, water losses due to evapotranspiration are at a minimum. The average 1970–1994 January-March and October-December precipitation (noted as Oct-March precipitation in table 1) varied as much as 2.01 inches (table 1). The range in January-March and October-December precipitation for the three stations extended from 27.27 inches in 1985 at Mackinaw to 8.06 inches in 1988 at Peoria WSO AP.

The greatest and least total annual precipitation in the 1970-1994 period of record indicate the range of variability in precipitation that needs to be included in any hydrological analysis of the study area. The greatest amount of yearly precipitation during the period was 59.70 inches at Mackinaw in 1993, or 19.44 inches above the 1970-1994 average annual precipitation of 40.26 inches. The least amount of yearly precipitation during the same time period was 21.56 inches in 1988 at Mason City 1 NW, or 14.70 inches less than the 1970-1994 average annual precipitation of 36.26 inches. These two readings represent a variation of 34.14 inches in total annual precipitation.

The amount of precipitation entering the groundwater flow system as recharge is strongly influenced by evapotranspiration. Water moves from the earth to the atmosphere through the processes of evapotranspiration, the combined effects of plant transpiration and direct evaporation of water from

Table 1 Precipitation values (inches) for three weather service stations near the study area.

Station	Average annual	Average April-Sept.	Average Oct-March	Highest annual	Lowest annual
Peoria WSO AP	37.88	23.11	14.77	55.35 (1990)	22.17 (1988)
Mackinaw	40.26	23.86	16.40	59.70 (1993)	22.16 (1988)
Mason City 1NW	36.26	21.87	14.39	56.56 (1993)	21.56 (1988)

Precipitation values are for the 1970-1994 period of record, excluding 1989.

the land surface and bodies of surface water. The rate of evapotranspiration responds to a variety of factors related to climate, vegetative cover, the degree of saturation of the soil profile, and the longevity of the water supply available for transpiration and evaporation. The amount of water moving into the atmosphere varies both seasonally, peaking during the summer, and from one year to the next.

Because the rate of evapotranspiration is difficult to measure, evaporation from natural water bodies can be used to indicate seasonal trends in the rate of evapotranspiration. Annual evaporation ranges from 34 to about 35 inches across the study area, increasing from northeast to southwest (fig. 7). The April-September evaporation ranges from 26 to about 26.5 inches, increasing from northeast to southwest across the study area (Neely and Heister 1987). Comparing precipitation with evaporation indicated that average annual precipitation exceeds annual evaporation, but that April-September evaporation exceeds normal April-September precipitation. This comparison suggests that little of the precipitation that falls during the growing season is typically available for groundwater recharge. Rainy periods during the summer may reduce the difference between evaporation and precipitation.

Regional Geologic Setting

The bedrock formations in central Illinois consist of a succession of sedimentary rocks several thousand feet thick. Lithologies include sandstone, limestone, dolomite, shale, and coal. These rocks were warped and tilted over many millions of years to form the Illinois Basin, which is centered in southeastern Illinois.

The older and generally deeper-buried rocks of central Illinois are mainly limestone, dolomite, and sandstone. These rocks commonly yield water from fractures (especially in limestone and dolomite) or permeable units (sandstone). Younger rocks, which are about 300 million years old, are found at or within a few hundred feet of the bedrock surface. These rocks are predominantly shale interbedded with occasional, relatively thin layers of sandstone, limestone, and coal. The bedrock may yield small quantities of groundwater from fractures in the shale, limestone, and coal, or from thin permeable sandstone beds. At depths of more than 200 to 400 feet below land surface, water in the bedrock of central Illinois is generally very mineralized and undrinkable.

After deposition of the sediments that now form the bedrock of the region, erosional topography developed on the bedrock surface during a long period of uplift. This erosional surface included major river valleys with numerous large and small tributary valleys. The major bedrock valleys in central Illinois are the ancient Mahomet and Mackinaw Bedrock Valleys (fig. 5), which merge in the confluence area.

The onset of continental glaciation some two million years ago profoundly changed the original bedrock landscape of central Illinois by disrupting drainage patterns and by deepening and ultimately burying the bedrock valleys. Repeated pulses of debris-carrying ice covered much of Illinois. The older pre-Illinois Episode and Illinois Episode glaciers covered larger areas of the state than the younger Wisconsin Episode glaciers. Pre-Illinois Episode ice sheets significantly modified the preglacial bedrock surface, initially by deepening the existing bedrock valleys through erosion and later by partially filling them mostly with sand and gravel. These sediments blocked and diverted several major channels away from the glacial margins. As each glacier melted, it left a layer of debris commonly called glacial drift. These deposits can be classified and mapped from subsurface information, such as logs and samples from test holes, as well as from information obtained from exposures of these deposits at the land surface. The complex series of events that ended about 14,000 years ago modified the landscape of the region to the extent that the bedrock is now covered with as much as 400 feet of glacial drift, including more than 150 feet of sand and gravel directly overlying the floors of the bedrock valleys in many locations. Within the study area, this sand and gravel is called the Sankoty-Mahomet Sand aquifer.

METHODS OF INVESTIGATION

Data Sources

Like all scientific studies, this one builds on previous research. Much of the previous work in this area has been of a regional scale and based primarily on analysis of data already on file at the ISGS and the ISWS. The recent study by Wilson et al. (1994), which included test drilling paid for by outside funding, is a notable exception.

The primary source of information for studies of groundwater geology is data from wells and borings drilled for resources (water, coal, oil, and gas) or engineering information (bridge and foundation borings, etc.) Most of these data consist of logs submitted by drillers. Both the ISGS and ISWS maintain files of water well logs. In addition, the ISGS maintains copies of logs from engineering borings and coal, oil, gas, and other exploratory test holes. More than 500 such logs are on file for this area (fig. 8). This number increases to more than 2,000 when logs for the surrounding townships, which are in the buffer area, are included. Although the volume of these data is large, the quality is generally inadequate for a detailed water resources evaluation. Coal, oil, and gas logs usually provide little information about the material above bedrock, which is the focus of this study, but they may have good data on bedrock elevation. Water well logs may lack the detailed information necessary to trace the continuity of aquifers, and they seldom provide information down to bedrock. Sometimes geophysical logs are available in addition to the drillers' logs, greatly improving interpretations of what material was encountered during drilling. Engineering logs are generally of very high quality, but the borings typically do not penetrate to the depth necessary to provide information on most aquifers. The ISGS also maintains a library of geologic samples from holes drilled throughout the state. For example, samples and geophysical logs are available for many of the test holes drilled

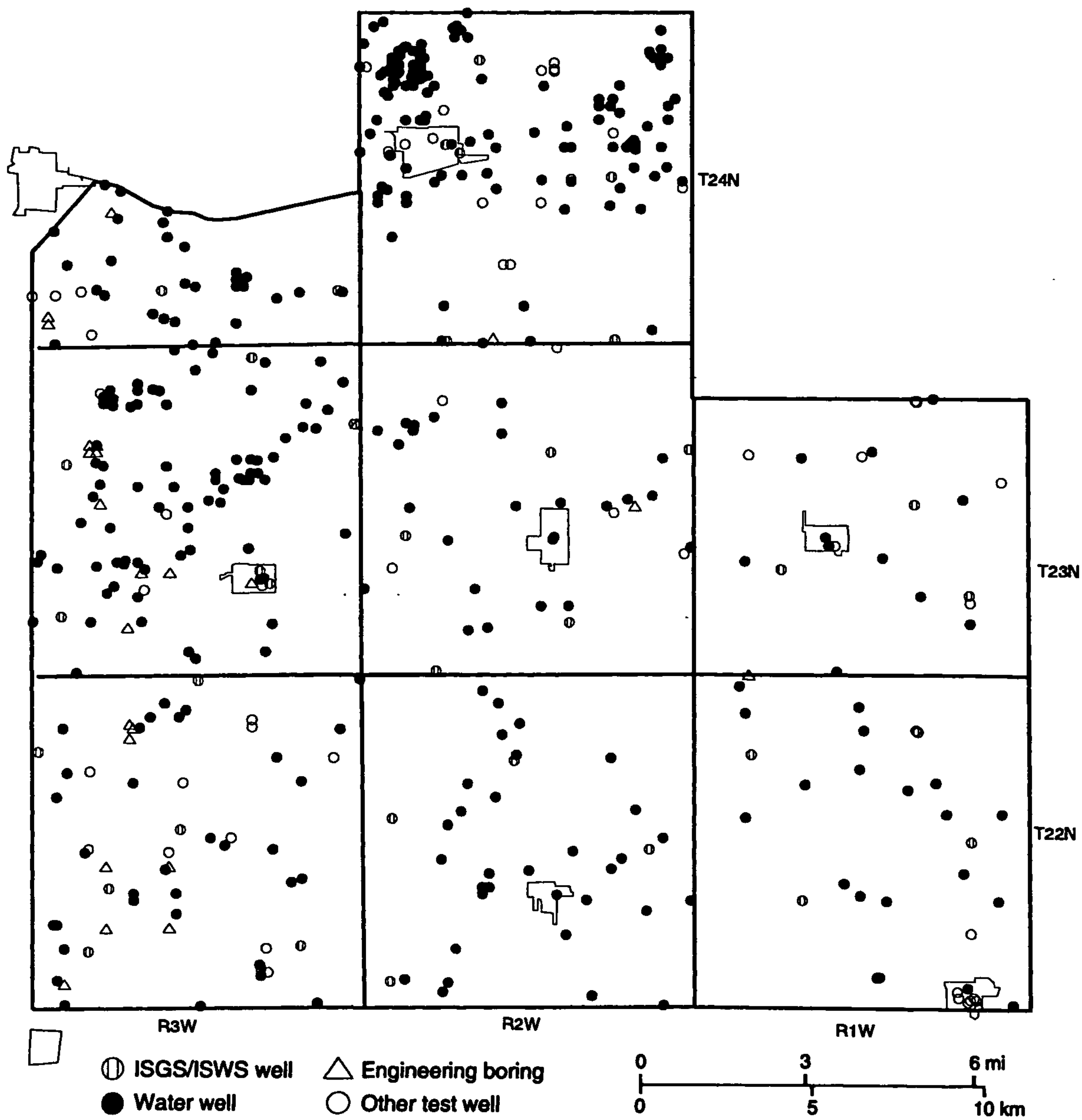


Figure 8 Locations for available well log data.

for the Normal West Well Field. Sample sets and geophysical logs provide the best information in the ISGS files.

Additional geologic data may come from previous geophysical studies of an area. The ISGS routinely performs electrical earth resistivity surveys for municipalities, industries, other public water supplies, and individuals who have been unsuccessful in finding water. Because this study area was selected for its vast potential resource, resistivity surveys have not been needed here. Data from large-scale surface geophysical studies are also available for some areas. An example of this is the work by Larson and Poole (1989) for the area near Bloomington's reservoirs. However, no previous large-scale geophysical studies have included the present study area.

Data from the logs on file at the ISGS were entered into a computerized database, along with calculations of bedrock elevations from the seismic refraction survey and the logs of the test holes drilled for this project. Locations of many of the wells had been previously verified (Kempton and Visocky 1992). All wells were plotted by computer and overlaid on 7.5-minute topographic quadrangle maps to check whether the locations plotted by the computer were accurate. Where the plotted locations were questionable, attempts were made to verify the correct location using plat books. For well logs that did not include elevation data, land surface elevations were estimated from the 7.5-minute topographic quadrangle maps.

The ISWS houses the primary hydrologic database for the state. It includes water quality analysis records, well construction reports, well sealing affidavits, well inventory records, water levels, data from pump tests, and other water well related records dating back to the 1890s. These data are filed according to township, range, section, and 10-acre plot using the numbering system shown in Appendix A. This numbering system is also used throughout this report. Selected data from these documents have been used to create five ISWS computer databases: the Private Well database, the Public-Industrial-Commercial Supplies database (PICS), the Ground-Water Quality database, the Illinois Water Inventory Program database (IWIP), and the Aquifer Properties database.

The Private Well database is a summary of well information from ISWS water records. It includes information such as the well's location, depth, owner, driller, Standard Industrial Classification code, and date drilled. The data come mostly from well construction reports, water quality analysis reports, and well inventory forms. The PICS database is a similar database of more detailed information about public water supply wells and large capacity commercial and industrial wells. The IWIP database contains data from annual surveys of facilities in the PICS database for water withdrawal and water use information. As the name implies, the Ground-Water Quality database contains water sample test results from samples analyzed at the ISWS laboratory and at the IEPA public water supply testing laboratory. Finally, the Aquifer Properties database summarizes the analysis of pumping tests conducted by the ISWS. Information includes owner, test duration, depth, aquifer being tested, location, descriptions of the site and well, and estimations of the hydraulic conductivity, transmissivity, and storage coefficient for the aquifer at that location. These databases were used to provide regional background information, indicate where new data were needed, and identify in which aquifer the wells used for the mass water level measurement were completed.

Seismic Refraction Surveys

Seismic refraction surveys record the seismic energy (from a small charge of explosive or "shot") that returns to the land surface after being refracted by an underground interface such as occurs at the bedrock surface (fig. 9). The recorded information is used to calculate the depth to the bedrock surface beneath the shot point and sensors.

For this study, 24 14-Hz geophone sensors were laid in 2,300-foot-long lines ("spreads"), with individual geophones spaced at 100-foot intervals along the spread. At each spread, energy was recorded from four shots, one at each end and one 500 feet inside each end. The maximum energy source was a 1 -pound charge of explosive buried in a 5-foot-deep hole. Energy from 180 shots along 45 seismic spreads was recorded in this investigation. The spreads were spaced at one-half-mile intervals along nine lines (fig. 10).

Refraction data were interpreted using the modified delay time method and the ray tracing computer program SIPT2 (Scott et al. 1972, Rimrock Geophysics 1992). The program calculates the elevation of the bedrock surface beneath each geophone. It automatically compensates for variations in land surface elevation and changes in the thickness of the near-surface, low-velocity zone. Geologic control in the form of logs from wells or test holes was available near several seismic lines. Refraction

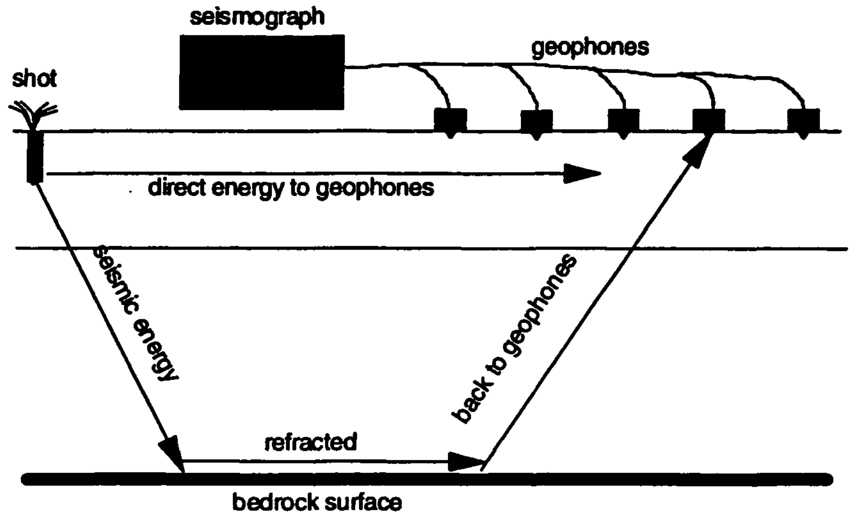


Figure 9 Schematic of the seismic refraction method. Seismic refraction surveys require a seismograph, geophones, and an energy source (or "shot"). A ray representing seismic energy is shown being refracted from the bedrock surface.

velocities were varied in the calculations until the calculated bedrock surface elevations gave a close match to the well data. A range of velocities was tried. The set most closely matching the well data was kept.

The near-surface geologic deposits were assumed to be in three seismic layers: a thin upper layer (about 10-25 feet thick) corresponding to soil, a relatively thick intermediate layer (about 200-250 feet thick) corresponding to glacial deposits, and a lower layer corresponding to bedrock. The seismic velocity of the upper layer was assumed to be 2,500 feet per second (ft/s). For each spread, the velocity of the other two layers was allowed to vary, and the depth of the two interfaces between the three layers was allowed to vary.

The seismic velocity of the glacial material ranged from 4,900 to 5,900 ft/s, averaging 5,590 ft/s for the 45 spreads. These velocities are somewhat less than typical glacial materials in Illinois (e.g., Larson and Orozco 1991). Velocities in this study are lower because of the presence of a thick sand and gravel layer within the glacial drift. Although the seismic data cannot distinguish the sand and gravel as a separate unit, its presence is noticeable by the decrease in velocity. Velocities for the bedrock are typical of shale, ranging from 10,000 to 10,700 ft/s and averaging 10,280 ft/s. The bedrock elevations were calculated using these velocity values. Results for each spread are reported as the average of the bedrock elevation calculated from the 24 geophones (fig. 10).

Data from the seismic work indicated that the bedrock surface is more irregular than shown on previous bedrock surface maps of the confluence area. South of Hopedale, five spreads produced average values for bedrock elevation of greater than 400 feet. One of these (Section 2, T22N, R3W, fig. 10) produced an elevation value of 529 feet, which was more than 100 feet greater than any previous data had shown. This value indicated that the bedrock high beneath Hopedale is a much taller and more extensive feature than had previously been recognized. Test drilling was planned near the location where the seismic work suggested bedrock was at 529 feet. Seismic data also produced bedrock elevation values of greater than 400 feet in Section 33, T24N, R2W, and Section 17, T23N, R2W (fig. 10). Two small depressions were seen from the seismic data, in Section 18, T23N, R1W, and Section 32, T23N, R2W, where elevations of approximately 315 feet were calculated. Both low values were detected in areas that had previously been mapped as less than 350 feet. Data from the seismic work was used to guide the selection of locations for test drilling.

Test Hole Drilling

Test drilling was undertaken at 25 sites across the study area (designated as MTH-1 to MTH-25 on fig. 11). Initial selection of locations for test drilling was made on the basis of an analysis of the availability and quality of existing data from water well records, logs of engineering and other borings, and oil and gas tests. This analysis identified areas with little or no subsurface information, or areas where more information would help clarify the subsurface geology or the hydrogeologic setting. The initial selection was modified because of the results of the seismic refraction survey. For example,

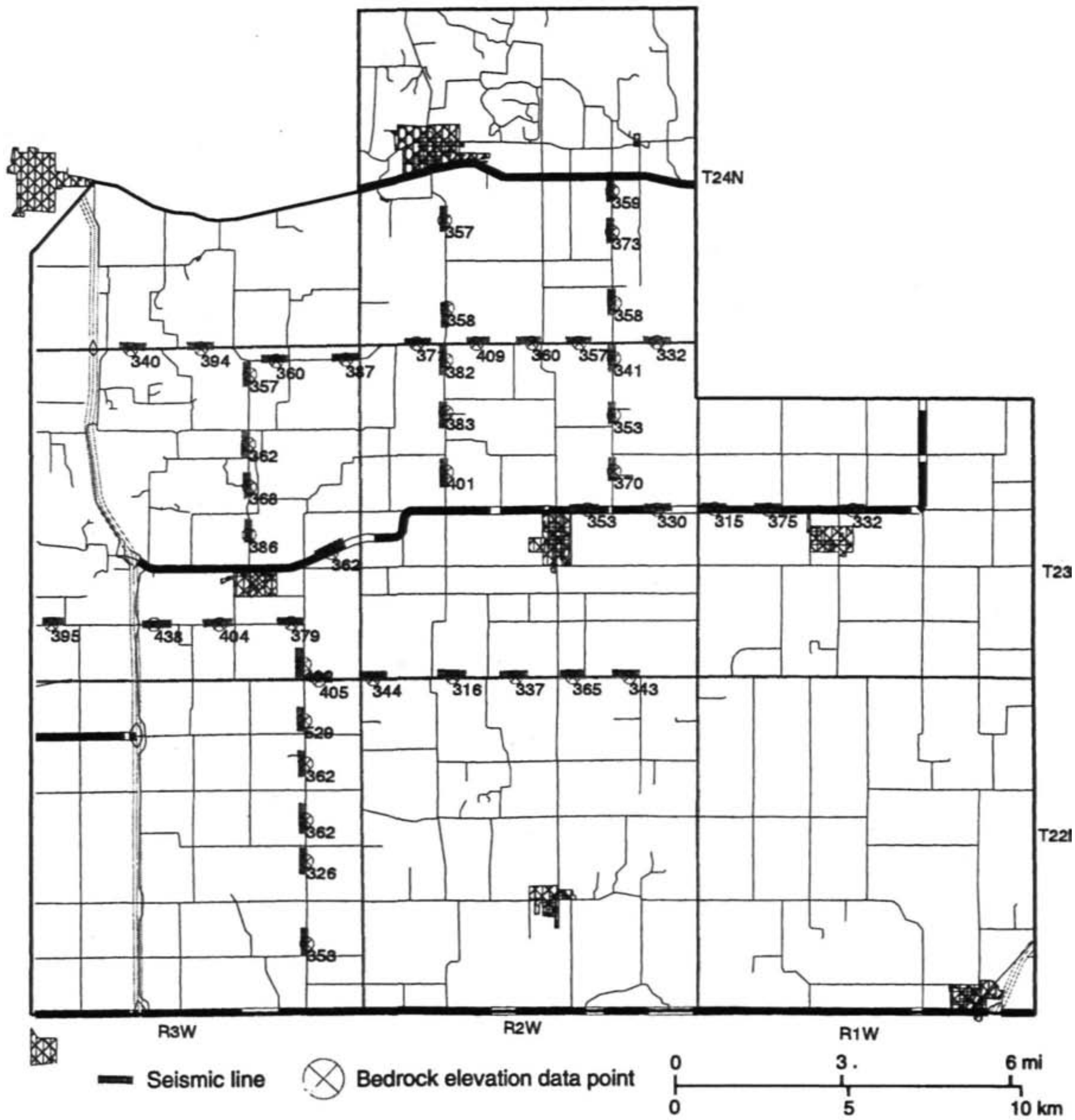


Figure 10 Location of seismic lines, showing the calculated bedrock elevation values.

seismic data revealed a steep bedrock hill south of Hopedale. Because of this unexpected feature, a test hole site was selected to confirm this buried hill and its effect on aquifer thickness.

Drilling method Albrecht Drilling used the forward rotary technique to drill a total of 8,010 feet of borehole. At each of the 25 sites, the initial test hole was drilled through the glacial drift and about 15 feet into the underlying bedrock. Coarse- and fine-mesh screens were used to collect samples of the drill cuttings from the drilling fluid as the fluid flowed into the portable mud pit from the discharge pipe connected to the surface casing. The drill cuttings were examined and described in the field by the driller and a geologist. Each kept a log noting depths, lithologies, and the physical characteristics of the sediments. The geologist also described the color and calcareousness of the samples. Colors were described using color names from the 1972 edition of the Munsell Color Charts (Munsell Color 1975). Calcareousness was characterized by noting reactions of the cuttings to a 10% hydrochloric acid solution. Representative samples for each 5 feet of drill cuttings were bagged for laboratory examination. In addition, cuttings were collected and placed on the ground in piles, each representing 10 feet of drilling. These were primarily used as an exhibit for visitors to the site, but they also helped field staff to visualize the sequence of materials.

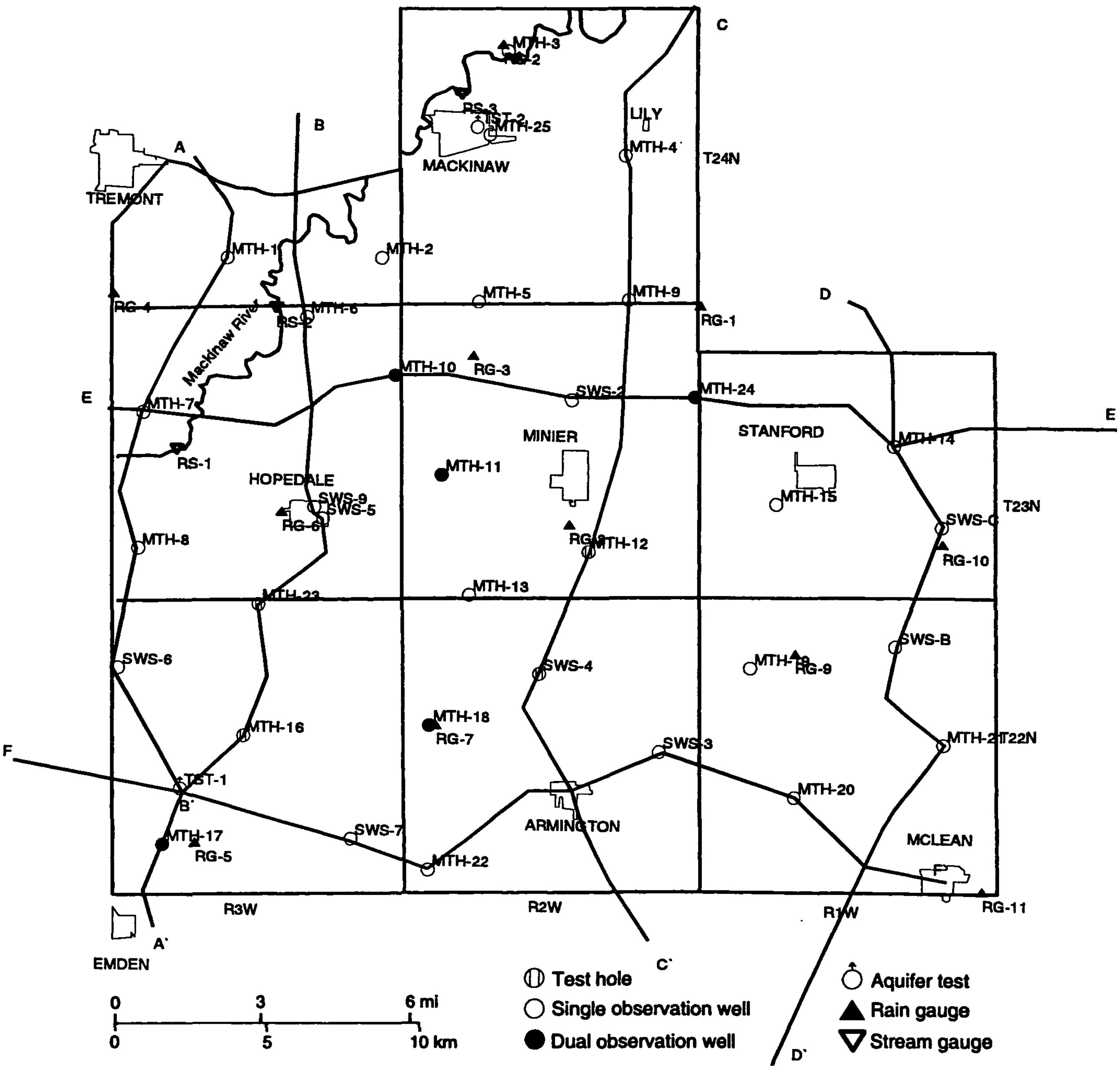


Figure 11 Locations of test holes, observation wells, rain gauges, river stage stations, and aquifer test sites.

Core collection Core samples were collected using a split spoon at sites MTH-6, 9,15, and 16 (fig. 11). The samples provided a comparison between the undisturbed sediments, the cuttings in the drilling fluid, and the geophysical logs. Coring provides the highest quality, but most expensive, samples. Therefore, coring was used only in a few locations to supplement cuttings.

Cores were collected with a split spoon sampler threaded onto a rod that was about 12 feet long and had a sliding weight attached to it. The sampler was lowered to the bottom of the test hole with a winch. A 140-pound weight was then raised 36 inches with a winch and released, driving the sampler into the sediments. The number of times the weight had to be dropped to advance the sampler each 6 inches was recorded. This number, called blow count, provides information on the type of material being penetrated, it is useful when no sample is recovered.

Samples were recovered at the following depths:

Site MTH-6:78 and 118 feet; the attempts at depths of 158 and 160 feet produced no recovery
Site MTH-9:48,80,140,200,240, and 280 feet
Site MTH-15: 54,100,120, 210, and 260 feet
Site MTH-16: 42,130, 206, 234,274, 284,296, 305, 315, 325, 335, and 347 feet

These samples were stored in glass jars for later analysis of moisture content and density in the laboratory.

Geophysical logging Downhole geophysical logs were run in the deep borehole at all test hole locations to supplement the field logs and samples, as well as to indicate the depth for placement of well screens. Such logs are critical for correlating geologic units encountered in different test holes, especially if continuous cores are not available. As discussed above, split spoon samples were collected only at selected locations because of their cost. Four types of logs were run in this project: natural gamma ray, gamma ray-neutron, acoustic, and caliper. Selection of the logging method for each site was dependent on site conditions and required data.

The ISGS downhole logging equipment is a Mineral Logging Systems Model 3500 truck-mounted system built by Gearhart-Owen Industries and modified by Mineral Logging Systems of Fort Worth, Texas. The equipment consists of several types of geologic sensing devices (called sondes), cables, and a recording device. Selected sondes are run sequentially in test holes that are cased or uncased and fluid or air filled. These sondes are lowered into a test hole attached to the end of a four-conductor armored cable. The cable is lowered and raised by a winch with a measuring device to show depth. The cable is connected to an electrical control panel and an analog chart recorder. Downhole signals are continuously recorded on as many as four channels on the analog chart as the sonde passes through the earth materials, providing a continuous in situ record.

Geophysical logging is ideally conducted in the open test hole after drilling is completed and the test hole has stabilized. At three of the sites, the logging was done through the casing because of concerns that the hole would collapse before logging could be completed. In these cases, the entire length of the test hole was cased, with the screen positioned at the desired depth, so that a log of the bedrock and overlying glacial deposits could be made. Because casing interferes with the other logging methods, only natural gamma ray and gamma ray-neutron logs were possible at these sites. Because gamma ray-neutron logging uses a radioactive source that requires permits for each site, its use is limited to holes where the owner will grant permission. It is also not used where caving material could trap the sonde and source in the hole. Table 2 lists the types of logs conducted at each test site. The five methods are discussed below.

Natural gamma ray log A natural gamma ray log is a graph of the gross gamma radiation (high-energy electromagnetic radiation) emitted by earth materials surrounding the sonde. In Illinois,

Table 2 Geophysical logs run on project test holes.

Site	Logs run	Site	Logs run
MTH-1	natural gamma ray	MTH-14	gamma ray-neutron, acoustic
MTH-2	natural gamma ray	MTH-15	gamma ray-neutron
MTH-3	natural gamma ray, gamma ray-neutron	MTH-16	gamma ray-neutron
MTH-4	gamma ray-neutron, acoustic	MTH-17	natural gamma ray
MTH-5	gamma ray-neutron, acoustic	MTH-18	natural gamma ray
MTH-6	gamma ray-neutron, acoustic	MTH-19	natural gamma ray
MTH-7	natural gamma ray	MTH-20	gamma ray-neutron
MTH-8	gamma ray-neutron, acoustic	MTH-21	gamma ray-neutron
MTH-9	gamma ray-neutron, acoustic, caliper	MTH-22	natural gamma ray
MTH-10	natural gamma ray	MTH-23	natural gamma ray
MTH-11	natural gamma ray	MTH-24	natural gamma ray
MTH-12	natural gamma ray	MTH-25	natural gamma ray
MTH-13	gamma ray-neutron, acoustic		

most natural earth radiation is generated from isotopes of potassium-40, thorium-232, and uranium-238. These elements are most abundant in clay minerals and less concentrated in clean quartz sand, gravel, and pure dolomite or limestone rock. Consequently, natural gamma ray logs often can distinguish aquifers from aquitards. Natural gamma ray logs are also useful for stratigraphic correlation and identification of lithology in fine grained sediments such as shale, buried soil zones, and silty lake clay, all of which show the greatest intensities. A strong linear correlation between the amount of clay and natural gamma radiation allows these uses. Fluid or air-filled, plastic- or steel-cased test holes generally do not affect levels of radiation required for a meaningful log.

For this project, natural gamma ray logs of open holes were used to guide well screen placement because these logs delineated aquifer units. Logs from both cased and uncased holes were used to aid stratigraphic correlation.

Gamma ray-neutron log The gamma ray-neutron log indicates hydrogen ion concentration in the area surrounding the test hole. Hydrogen is the most effective element in reflecting neutrons because its nucleus has nearly the same mass as a neutron. Because the flux of neutrons is proportional to the amount of hydrogen present in earth materials, and most of the hydrogen is present in water, the flux is proportional to the amount of water. Where water fills interstitial pore space, the neutron log records relative amounts of rock porosity. The ISGS gamma ray-neutron sonde uses a 3-curie americium-241/beryllium radioactive source that has a flux of 6.67×10^6 neutrons per second.

A gamma ray-neutron log cannot discriminate between hydrogen in pore water and hydrogen in hydroxyl ions (OH^-), which are principally found in clay minerals, the primary constituents of shales. As a result, shale and other lithologies containing minerals with hydrogen may appear similar to high porosity rocks on neutron logs. Gamma ray-neutron logs for sandstones, limestones, and dolomites with very low clay mineral content are good indicators of relative porosity.

Acoustic logs Acoustic or sonic logging employs the same principle used in the seismic refraction survey, providing data on the speed at which sound waves travel through earth materials. Lithology and porosity can be deduced from the graphic record.

Acoustic logs record the travel time of an acoustic wave from a transmitter in the sonde, through the borehole fluid and earth materials adjacent to the borehole, to two receivers in the sonde. A sonic wave created by the transmitter travels through the surrounding materials at a velocity related to the matrix minerals and porosity of the earth materials. The receivers relay refracted acoustic energy from the pulsed transducer to the recorder. Bow springs keep the acoustic sonde in the center of the borehole to ensure consistent spacing between the probe and the surrounding earth materials.

Acoustic logs were run on test holes located along the seismic lines to improve the interpretation of the seismic data. In most cases, the test hole data confirmed the bedrock elevation predictions from the seismic refraction survey; thus, refinements in the seismic interpretation were minor.

Caliper log The caliper sonde, which measures test hole diameter, has a set of three spring-loaded arms that extend radially, with a diameter of as much as 30 inches, to the test hole walls. The diameter of the hole may change because of material type (e.g., caving sand versus competent rock), fractures, or cavities. Changes in diameter, unusual obstructions, and damage to the test-hole walls also cause changes in diameter on the log. Because the hole diameter may change for many reasons other than the properties of the geologic material, caliper logs are not very useful for distinguishing types of glacial deposits. Caliper logs are, however, essential in deciding whether to run logs that require nuclear sources. They are also generally very helpful in the interpretation of other geophysical logs because most other logs are affected by changes in the diameter of the test hole.

Observation Well Construction

Observation wells provide access points for measuring groundwater levels and collecting ground-water samples for chemical analysis. Observation wells were installed in the Sankoty-Mahomet Sand aquifer at 22 of the 25 test hole sites (fig. 11); 16 of the 22 sites have one well and the other six sites have two, for a total of 28 wells. At most sites where a second, shallower, aquifer was found, a second, shallower, observation well was installed in a separate hole. Water levels measured in the shallower observation well provide a means for determining the degree of hydraulic connection between the two aquifers and identifying the vertical direction of groundwater flow.

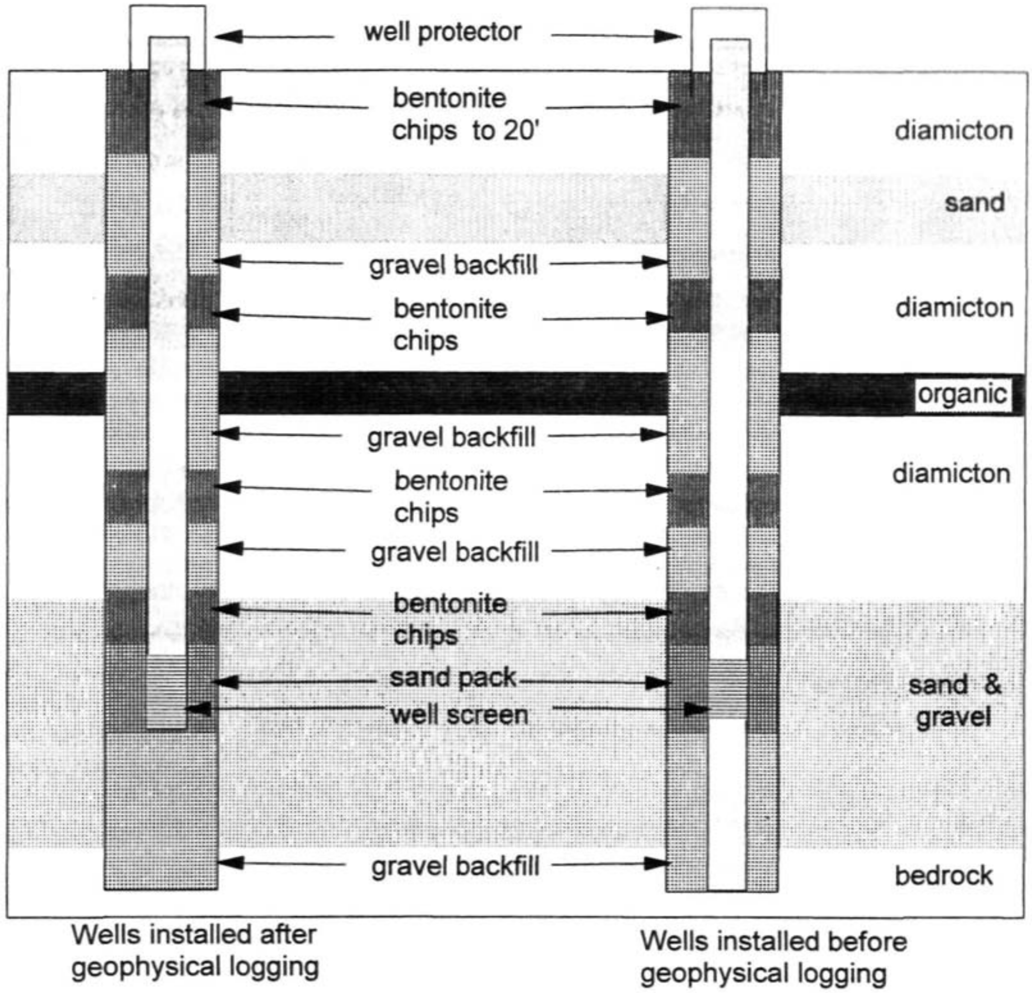


Figure 12 Schematic of observation wells.

The observation wells were constructed of 2-inch-diameter, schedule 40, flush-jointed PVC pipe with either 5 or 10 feet of screen (fig. 12). Either 0.020-inch (no. 20 slot) or 0.010-inch (no. 10 slot) screens were used, depending on the overall grain size of the interval to be screened. The screen was set at the desired depth, which was selected using information on the nature of the aquifer from the natural gamma-ray logs and sample descriptions. The test hole was backfilled with gravel until the screen rested on the gravel. Clear water was pumped down the casing and through the screen until clear fluid discharged from the annulus (the open space between the outside of casing and the sides of the borehole) at land surface. This cleared any remaining drilling fluid from the observation well and annulus. Sand was packed around the screen, and the annulus was filled with gravel to just above the top of the aquifer where the annulus was sealed with bentonite chips. The rest of the annulus was backfilled to a depth of 20 feet below land surface with gravel and several annular seals of bentonite chips. The remaining 20 feet of annular space was filled with bentonite chips to the land surface. A 4-inch square steel well protector was installed at land surface to secure the observation well. Each well was developed by pumping it with air until the well produced nonturbid water. This usually required several hours and sometimes needed to be repeated because the well was producing cloudy water. At three sites (MTH-3, 10, and 24), the observation wells were installed before the downhole logging because of the possibility that the sediments would cave into the hole. Thus, the well screen depth had to be selected without information from a geophysical log. Casing was set on bedrock at the bottom of these test holes with the screen placed at the desired depth so that the full extent of the test hole could be logged for geologic purposes. The annulus was backfilled with gravel up to the bottom of the screened interval, and the observation well was backflushed. The rest of the well construction procedure after this was the same as that given above.

Laboratory Study of Test Hole Samples

Sample studies Samples collected from the test holes drilled into bedrock were studied in the laboratory to provide a more detailed description than could be accomplished during drilling. Analysis of the samples from one test hole was completed before analysis of samples from another was begun. The sample bags were arranged according depth. Approximately 100 to 150 grams of material from each sample bag was placed in a watch glass. The samples were moistened with tap water using a spray bottle and analyzed for texture and Munsell color under consistent lighting conditions. The material in each watch glass was tested with several drops of a 10% solution of hydrochloric acid to determine the presence of carbonate minerals. The physical and lithologic characteristics of the material in each watch glass were described. The field logs were reviewed during the laboratory examination of the samples. A composite log of each test hole was developed by combining the field and laboratory descriptions. The geophysical logs, particularly the natural gamma and gamma ray-neutron logs, were crucial in determining the depths and helping to distinguish the sedimentologic character of the stratigraphic units. The composite logs are presented in Appendix B.

Grain-size analyses Grain-size data are used to characterize the distribution of grain sizes in sand and gravel aquifers, a characterization that is important in designing water wells. Grain-size data are also useful for in describing the sedimentological properties of diamicton units. The composite logs (Appendix B) were reviewed to select the test holes and intervals for grain-size analysis. The 10 test holes chosen were MTH-4, 6, 9, 10,15,16, 17, 18, 24, and 25. These were selected to best describe the character of the sand and gravel deposits present in the study area, particularly where multiple or thick aquifers were encountered. From each test hole, however, several intervals that represented all of the material types found at that location were selected for determination of grain-size distribution. These intervals are marked with asterisks in Appendix B.

As shown on the logs in Appendix B and the distribution curves in Appendix C, entire intervals were sampled for grain-size analyses. Representative samples were taken from each 5-foot increment within intervals greater than 5 feet thick. Thick sand and gravel units were divided into two or three intervals of equal size to adequately characterize grain-size changes within these units. Analysis of the sand and gravel units required more than 100 grams of material while that of the diamictons, silts, and clays required about 50 grams. All of the samples were analyzed in the ISGS Particle-Size Laboratory.

For particle-size analysis, the samples were air dried and then weighed. They were sieved through six nested screens the sizes of which were 2.0 mm, 1.0 mm, 0.5 mm, 0.25 mm, 0.125 mm, and 0.063 mm. These sizes separate the very coarse, coarse, medium, fine, and very fine sand components of sieved samples according to the Wentworth Classification system for grain size. The portion of the sieved samples retained on the 2.0-mm screen was considered to represent the gravel fraction of the sample. The portion that passed through all of the screens was considered to represent the 0.004-mm silt/clay fraction. The silt and clay fraction values were determined for selected samples using a standard hydrometer test. A 4.0-mm screen was added to the screen nest for sieving the samples from MTH-25. This size screen separates the granule component from the coarser pebble to boulder sizes in the Wentworth Classification system.

The amount of sample retained on each screen was weighed and added to the weight retained by the coarser screens. The total cumulative weight of a sample after sieving was compared with its total starting weight. The percent of the cumulative weight retained by each screen was calculated by dividing the cumulative weight retained by the total cumulative weight of the sample. The values for cumulative percent retained were graphed on semilog paper as grain-size distribution curves (Appendix C). These curves depict grain-size distribution and show the percentage of the sample that is above or below a certain grain size (Driscoll 1986).

Grain-size distribution curves indicate the degree of sorting and the overall coarseness of the sample within the range of grain sizes represented by the sieve sizes used. Well sorted sediments, which are mostly made up of a relatively narrow range of grain sizes, are depicted by relatively steep slopes in grain-size distribution curves. Poorly sorted sediments have a relatively wide range of grain sizes and are depicted by curves with flatter slopes. The 50% size (mean or average particle size) can be used to indicate the overall fineness or coarseness of a sample. Where grain-size curves have **a** relatively flat slope, which is generally true for this study, the 50% size is a poor indicator of fineness or coarseness. Other values, such as the 90% size, may be more useful in representing the fineness of the sample.

The effects of the forward rotary drilling method on the samples collected from the drilling fluid must be considered when interpreting the grain-size distribution curves presented in Appendix C. Grain-size distributions determined on samples collected during rotary drilling are less representative of the actual grain-size distribution than are similar analyses made on intact samples, such as those collected from split spoons. During rotary drilling, sediments that are at a shallower depth in the test hole may be incorporated into the drilling fluid as samples from a greater depth move up the annulus. Sediments of different sizes or densities may move up the annulus at different rates causing the finer grained sediments to arrive before the coarser grained sediments. The finer grained sediments may be recirculated through the test hole in the drilling fluid. The drilling action of the bit may break down larger particles into smaller ones. These effects are most noticeable in grain-size distribution curves obtained from the analysis of a single sediment type (e.g., the curve for MTH-16, 139-144 ft; Appendix C).

The grain-size distribution curves for most of the samples (Appendix C) show a wide degree of variation within the tested intervals. The curves exhibit a nonuniform grain-size distribution typical of poorly sorted sediments. Many of the distribution curves for the Banner Formation sands and gravels are S-shaped, suggesting that the distribution curve includes most of the grain sizes present at the site. Sediments with S-shaped grain-size distribution curves tend to be more porous than those having flatter curves. In many cases the curves seem to have a truncated S-shape, indicating the curve may extend into the pebble and cobble component of the grain-size range. This component may be under-represented due to the effects of the forward rotary drilling method. Grain-size distribution is also a rough predictor of hydraulic conductivity, so these data can supplement aquifer test data by showing the variability across the study area.

Moisture content and density Moisture content and density measurements were made on split-spoon samples collected from test holes MTH-9,15, and 16 and compared with values calculated from geophysical logs. Samples were removed from glass jars, weighed with their natural moisture content, placed in a water-filled graduated cylinder to determine volume, dried in a convection oven for 24 hours at 110° C, and then reweighed. Percent moisture content is equal to weight lost by drying (i.e., evaporated water) divided by the dry weight of the sample, expressed as a percent. Bulk density is equal to the weight of the dried sample divided by the sample volume. As is evident from table 3, these measurements are useful for distinguishing between different types of fine grained deposits, a distinction that can be very difficult to make from cuttings. Buried soils (the Robein Silt and Lierie Clay) had moisture contents of more than 15%. These units are significant because they serve as marker beds. The lacustrine deposits in the study area generally had a lower density and a higher moisture content than the diamictons. This should be expected because the lacustrine deposits were deposited by quiet water, causing them to be soft and wet. In contrast, the diamictons were deposited

Table 3 Density and moisture content measurements from spoon samples.

Geologic unit	Test hole	Depth (ft)	Density (g/cm^3)	Moisture content (%)	Gamma ray-neutron (cps)*
Wedron diamicton	MTH-9	48	1.88	16.6	33
Wedron diamicton	MTH-16	42	2.04	10.8	180
Robein Silt (Glasford Fm)	MTH-15	54	2.02	15.6	65
Robein Silt (Glasford Fm)	MTH-16	130	1.98	19.7	150
Radnor (Glasford Fm) diamicton	MTH-9	80	2.28	8.7	460
Vandalia (Glasford Fm) diamicton	MTH-15	100	—	6.4	260
Lierie Clay (Banner)	MTH-16	206	1.88	18.2	160
Hillery (Banner) diamicton	MTH-15	120	2.09	7.2	175
Banner lacustrine	MTH-16	234	1.97	12.1	175
sub-Sankoty diamicton	MTH-15	260	2.16	7.6	270

*cps = counts per second

by glacial ice, the weight of which squeezed out much of the sediments' original water content and left the resulting deposits very dense.

Moisture content was calculated from the gamma ray-neutron logs and compared with that determined from the samples for MTH-15 and 16. This comparison had do be done separately for each hole because the relationship differs as a result of different hole diameters. The correlation coefficients were 0.92 and 0.96, indicating close agreement between the two methods of calculating moisture content and density. Thus, the gamma ray-neutron logs could be used both to distinguish aquifers from fine grained sediments and to differentiate between different types of fine grained sediments.

Biweekly Groundwater Level Measurements

Water levels in 39 dedicated observation wells were monitored biweekly. Of these, 28 (prefix MTH on fig. 11) were installed for this project and 11 (prefix SWS on fig. 11) were installed for the recently completed study by Wilson et al. (1994). Thirty-one of the wells were finished in the Sankoty-Mahomet Sand aquifer and eight were finished in shallower sand and gravel aquifers. The observation well water levels were measured by lowering an electric tape into the well and directly reading the depth to water. The elevation of the measuring point location was determined and the depth to water was converted to an elevation in feet above mean sea level (msl). These data were collected beginning after well development through July 1995. .

The data are plotted as line graphs (hydrographs) of water level versus time and are presented in Appendix D. In the appendixes, sites with two wells use the suffix "d" for the deep and "s" for the shallow well. For a confined aquifer, changes in water level indicate fluctuations in the hydraulic head in the aquifer. For an unconfined aquifer, the changes in water level indicate when the water levels in the aquifer are rising (recharge greater than discharge) and lowering (recharge less than discharge). This information, when evaluated with rainfall and groundwater use data, provides a sense of how specific factors influence water levels of the aquifer. Evidence of aquifer interconnection can be determined by comparing hydrographs of the deep and shallow water levels at the same location. Where water level changes are similar or the water levels are at nearly the same elevation in different aquifers, the aquifers may be hydraulically connected.

River stage measurements were collected at three locations along the Mackinaw River (denoted with the prefix RVR on fig. 11) by lowering a tape from permanent locations on bridges over the river. These data were collected biweekly from September 1993 through July 1995. They are presented in Appendix E. Hydrographs of the river stage can be used to evaluate the recharge characteristics of an aquifer. When the river stage and groundwater level changes mimic each other, there may be a hydraulic connection between the two. If a connection is established, the elevations of each indicate whether the river is a discharge point or recharge point for the aquifer. Flow is from the unit of higher water level elevation toward the unit with lower water level elevation. At different times during the year, such as when the river is unusually high or low, the recharge/discharge roles of the two units may reverse.

Mass Groundwater Level Measurements

Mass groundwater level measurements consist of measuring water levels in a large number of wells over a short period of time over the entire aquifer. When plotted on a map and contoured, the data present an interpretation of the potentiometric surface for that point in time. From the groundwater surface map, groundwater flow direction, gradient, and general areas of recharge/discharge can be determined. For this study, wells in both the Sankoty-Mahomet Sand and shallower aquifers were monitored. Completing such a task required the use of private wells in addition to the dedicated observation wells installed for the project.

Several steps were taken to establish the network of measurable wells. Private wells were selected and well owners were contacted to obtain permission to use their wells. A field visit was conducted to verify the well location, accessibility, and information on the driller's log. For wells without a well log on file, as much information as possible was gathered from the well owner and, occasionally, the pump contractor. Information collected included owner, depth, year installed, screened interval, pump setting, driller, and pump contractor. This information was combined with the known regional geology to determine the aquifer being utilized. If it was not clear which aquifer was being used, such as in cases where the interval was close to the edge of a geologic unit or when accurate data were not available, the well was not used. Finally, the depth to water was measured if possible.

In a 2-week period in both August 1994 and July 1995, water levels in 200 wells throughout the study area were measured. Of these 200 wells, 126 were finished in the Sankoty-Mahomet Sand aquifer and 74 in the shallower Glasford and Upper Banner Formation aquifers.

Aquifer Tests

The hydraulic properties of an aquifer are tested by pumping water from a well and monitoring changes in water levels. Short duration tests, called specific capacity tests, are commonly performed by water-well drillers to provide an estimate of the ability of a well and pump to perform as desired. Tests of longer duration, called aquifer tests, give a better estimate of the hydraulic properties of an aquifer and the aquifer's ability to yield water to a well or group of wells.

An aquifer test is a controlled field experiment to determine the hydraulic properties of an aquifer using one pumped well operating at a constant flow rate and monitoring water level fluctuations in at least one observation well. These tests are conducted to provide the data necessary to quantify the aquifer's transmissivity and storage coefficient. Insight on leakage through confining units, hydraulic connection to rivers or other aquifers, and distance to aquifer boundaries may also be gained from aquifer tests. During a test, the water levels in the pumped well and in nearby observation wells are monitored frequently. These data are graphically analyzed using plots of drawdown versus time and drawdown versus distance. Several formulas that relate transmissivity and storage coefficient to the drawdown near a pumping well have been developed (Theis 1935, Cooper and Jacob 1946, Hantush and Jacob 1955, Hantush 1956, Ferris 1959, Walton 1962, Boulton 1963, Prickett 1965, Neuman 1975).

For this study, two methods of analysis were used for each test. The type-curve method (log-log) developed by Theis (1935) and the modified Jacob method (semilog) for artesian conditions with no leakage were used at one site (TST1 on fig. 11). The leaky artesian method (log-log) developed by Hantush and Jacob (1955) and the semilog leaky artesian method developed by Hantush (1956) were used to determine the aquifer properties at a second site (TST 2 on fig. 11). Appendix F describes these methods in detail.

The two aquifer tests were conducted to determine aquifer properties of the Sankoty-Mahomet Sand aquifer in the southwestern and northern parts of the study area (TST1 and TST2, respectively). These locations were chosen on the basis of aquifer thickness and site accessibility. At each site, a high capacity well and four observation wells were installed. Three of the observation wells were finished in the Sankoty-Mahomet Sand aquifer and one was finished in a Glasford sand aquifer. Water levels were measured during each test and for several days after pumping stopped (recovery) using an 8-channel HERMIT® data logger and accompanying pressure transducers. At TST2, a separate automated data collection system (the ISWS McDas system) was installed at OW-3 due to the distance from the pumped well. Periodically during the tests, measurements of groundwater levels were also taken using steel tape. A calibrated orifice tube was attached to the end of the pump discharge to monitor the flow rate during each test.

After the testing was completed, the production well and the observation wells at TST1 were abandoned and plugged according to Illinois Department of Health guidelines, except for OW3, which was added to the observation well network. The production well at TST2 was taken over by the Village of Mackinaw and was permitted as their Municipal Well no. 6. The observation wells installed for this aquifer test were abandoned and plugged according to Illinois Department of Health guidelines.

Rain Gauges

Rain gauges were installed at 11 sites throughout the study area (denoted by the prefix RG on fig. 11). Weighing bucket rain gauges, which use a battery driven paper chart to continuously record rainfall amount and intensity, were used. The paper charts were changed biweekly. Data from the charts were digitized in the office, allowing individual rainfall events to be listed separately.

Rain gauge data were plotted to correlate changes in aquifer water levels and river stages with rainfall patterns. The goal in analyzing this information was to determine if and how rainfall events affect the water levels in observation wells screened in the aquifers of the study area.

Groundwater Chemistry

Groundwater chemistry data are used to calculate, on the basis of the spatial distribution of chloride concentrations, the volume of recharge to an aquifer. Chemical data are also used to help indicate the source of water in an aquifer and explain natural spatial variations in groundwater quality. Most of the groundwater samples were collected from wells installed and developed by the ISWS and ISGS for this study. Additional samples were obtained from private wells directly from the well head, or by using outside faucets that bypassed home water treatment units and holding tanks.

Ten wells were sampled during three separate sampling events; six wells were completed in the Sankoty-Mahomet Sand aquifer and four in the shallower sand and gravel aquifers of the upper Banner and Glasford Formations. For the first sampling event, groundwater and gas samples were collected from six wells, three in shallow Glasford aquifers and three in the deeper Sankoty-Mahomet Sand aquifer. The first set of samples was collected from wells along the groundwater flow divide, as mapped by Wilson et al. (1994, p. 36), where chloride (Cl") concentrations in the groundwater were greater than those in the Sankoty Sand aquifer in the Mackinaw Bedrock Valley to the north and south. The selection of subsequent sampling locations was dependent on the results obtained from the first sampling event; the second and third locations were north and south of the groundwater divide, respectively.

The groundwater chemistry of the Mahomet Sand aquifer, the Sankoty Sand aquifer, their union as the Sankoty-Mahomet Sand aquifer, and the overlying sand and gravel aquifers in the upper Banner and Glasford Formations was evaluated using groundwater chemical data from this study, supplemented by data selected from the ISWS groundwater quality database and published analyses (Hanson 1950,1958,1961, Pannoetal. 1994, Holm 1995).

Selected water samples were analyzed for major and minor cations and anions. All groundwater and gas samples were analyzed for ^{13}C (carbon-13) from CH4 (methane) and CO2 (carbon dioxide), ^{18}0 (oxygen-18) and D (deuterium) from H_2O (water), D from CH4, ^{14}C (carbon-14) from HCO3 (bicarbonate), and ^{3}H (tritium) from water. Techniques used for isotopic analysis are presented in Hackley et al. (1992) and the International Atomic Energy Agency (1983).

Temperature, Eh (platinum electrode potential), pH, and specific conductance of the water samples were measured in the field using the appropriate electrodes, meters, and standards (Barcelona et al. 1985, Garske and Schock 1986). These parameters, which are indicators of water quality, were monitored continually as water was pumped from each well. Water must be pumped from a well before a sample is collected because the chemical characteristics of the water in the well may be altered by isolation from the aquifer, reduced pressure on the water, and contact with air and the well casing. Stabilization of the indicator parameters signals that water being removed from the well is representative of water in the aquifer. Samples were collected after the rates of change of pH and Eh were less than 0.01 pH units per minute and 0.5 mV per minute.

Background and threshold concentrations for Cl^- were determined using cumulative probability graphs developed for geochemical exploration by Sinclair (1974). The values calculated were used to statistically separate different groundwater types on the basis of CI" concentration.

Groundwater Recharge

Because of the difficulties and uncertainties in determining groundwater recharge, two methods were selected to estimate this parameter (1) Walton's (1965) groundwater runoff estimates and (2) mass balance calculations. Both methods and the results obtained are presented in more detail in the Groundwater Recharge subsection of the Hydrology section.

Groundwater Quality

Development of a groundwater resource needs to consider not only the availability of groundwater, but also the quality of the groundwater. The ISWS maintains a groundwater quality database of private and municipal water sample analyses. These data were evaluated along with new data collected for this study.

During the course of the current study, the ISWS secured additional funding through the Environmental Protection Trust Fund to conduct a detailed investigation of the arsenic concentrations in the confluence area (Holm 1995). Holm collected samples from the observation wells in this study to

Table 4 Groundwater quality analyses.

alkalinWes (Alk)	cobalt (Co)	nitrite (NC_2)	silicon (Si)
aluminum (Al)	conductivity	nickel (NO	silver (Ag)
alachlor	copper (Cu)	nonvolatile organic	sodium (Na)
ammonia (NH3)	fluoride ($F^{-)}$	carbon (NVOC)	strontium (Sr)
antimony (Sb)	hydrogen sulfide (H_2S)	odor*	sulfate ($S0_4^{2-}$)
arsenic (As)	iron (Fe)	pH	sulfur (S)
atrazine	lead (Pb)	phosphorus (P)	tallium (Ti)
barium (Ba)	manganese (Mn)	platinum-electrode	tin (Sn)
beryllium (Be)	magnesium (Mg)	potential (Eh)	titanium (Ti)
cadmium (Cd)	mercury (Hg)	potassium (K)	vanadium (V)
calcium (Ca)	molybdenum (Mo)	selenium (Se)	zinc (Zn)
chloride (Cl^-)	nitrate ($N0_3$)	simazine	

*Odor was evaluated by sniffing samples, not by an analytical method.

evaluate arsenic concentration and additional parameters and to provide the general chemistry data required for this project. The samples were analyzed for an extensive group of analytes (table 4).

Groundwater was pumped from the wells using a submersible pump (Grundfos, Clovis, CA). The flow rate was estimated using a stopwatch and a calibrated bucket. The flow rate was approximately 10 liters per minute in most wells. Most of the water was pumped into the calibrated bucket and then discarded. The temperature was measured in the bucket using an alcohol-in-glass thermometer. A small fraction of the water was directed through a flow-through cell (Garske and Schock 1986) for measuring pH, platinum-electrode potential (Eh), and specific conductance. Pumping the well continued until the pH and specific conductance readings stabilized. These readings were recorded. The Eh never attained a steady value although the rate of change continually decreased. At least three well volumes were pumped from the well by the time the readings had stabilized.

Groundwater samples were collected after the pH and specific conductance readings had stabilized. The flow bypassed the flow-through cell during sample collection. Samples were collected without filtration for determination of nonvolatile organic carbon (NVOC), volatile organic carbon (VOC), herbicides, and ammonia. The sample tube was then connected to an in-line tangential-flow filter holder (Nuclepore) that held a polycarbonate filter (90 millimeter diameter, 0.1 micrometer pore size, Nuclepore). Samples were collected for determination of alkalinity, anions, metals, and arsenic. The samples were stored in an insulated cooler for transport to the laboratory and then stored at 4°C until analysis. Duplicate samples were collected from one well per sampling trip. One set of sample bottles filled with deionized water was taken along with the other sample bottles to serve as a trip blank, accounting for any contamination during transportation and storage. A filtration blank was collected on one sampling trip.

Alkalinity values were determined by titration with standardized acid immediately after sample collection. The endpoint was located by Gran's method (Butler 1982). At least two alkalinity titrations were performed for each water sample. The average value for each sample is given in this report. The relative average deviation (difference of duplicates or range of triplicates divided by the average) was less than 5% for all samples. All other analyses were performed at ISWS laboratories.

Arsenic was determined by graphite furnace atomic absorption spectrophotometry (GFAAS) using a palladium-magnesium matrix modifier (Welz et al. 1988). The detection limit of 3 ug/ L was estimated as three times the standard deviation of the blank (Keith et al. 1983). For each set of samples analyzed, at least 20% of the samples were run in duplicate, at least 10% were spiked, and a quality assessment standard (WS378 #4, U.S. EPA) was run at the beginning and at the end of the analysis. Ammonia concentrations were determined using an ammonia gas-sensitive electrode (Orion) (Clesceri et al. 1992). It is impossible to estimate a meaningful detection limit for a gas-sensitive electrode. However, the ammonia concentrations of all groundwater samples were within the linear range of the electrode. The ammonia concentrations in the trip blanks were all well below the linear range (i.e., less than 1% of sample concentrations). Therefore, the ammonia determinations were reliable.

Metals, anions, NVOC, herbicides, and VOCs were determined by a group of chemists certified by the Illinois Environmental Protection Agency. Metals were determined by inductively coupled plasma-atomic emission spectrometry (ICP-AES). Anions were determined by ion chromatography. Nonvolatile organic carbon was determined by irradiation with ultraviolet light and persulfate oxidation and measurement of CO_2 by nondispersive infrared absorption (Peyton 1993). The herbicides atrazine, alachlor, and simazine were determined by liquid-liquid extraction and gas chromatography using a thermionic specific detector (specific for nitrogen and phosphorus). The VOCs were determined by purge-and-trap gas chromatography with photoionization and Hall electrolytic conductivity detectors.

Groundwater Use

The ISWS collects annual groundwater use information from public water supplies and self supplied industries. For purposes of quantifying total groundwater withdrawal within the study area, pumpage was subdivided into five water-use categories: municipal, rural-residential, industrial, livestock, and irrigation. Use by rural residents, livestock, and for irrigation can be estimated using the method developed by Kirk (1987) as part of the Illinois Water Inventory Program at the ISWS. Kirk's method uses the per capita use by small communities to extrapolate rural water use. Data needed to compute these estimates include population census data. Illinois Agricultural Statistics Service data on livestock populations are used for livestock estimates. The irrigation estimates are calculated on the basis of inches applied per acre. The acreage was determined for this study by noting irrigation in the field during the collection of water level data and while compiling the well inventory information. The high capacity wells in the PICS database were reviewed and investigated to verify the irrigation acreage.

GEOLOGY

Surface Geologic Features

The study area is located within the Bloomington Ridged Plain, which is within the Till Plains Section of the Central Lowland Province (Leighton et al. 1948). Drift from the last glaciation, the Wisconsin Episode, forms the land surface everywhere in this area, except for the southwest comer (fig. 13). Erosion and deposition by the last continental ice sheet gave shape to the current landforms. Subsequent weathering and erosional processes have altered the landforms to their present shape. The land surface topography is marked by the relatively high, arcuate ridges of glacial end moraines (ice-marginal positions). These features trend northwest to southeast across the northern and southwestern parts of the study area. The highest elevations in the study area occur on the moraines. These elevations reach about 820 feet in the north, on the Bloomington and King's Mill Moraines, and about 730 feet in the southwest comer, on the LeRoy Moraine (fig. 13). A broad, flat area between these end moraines lies at an elevation of about 650 feet.

Surface drainage of the study area occurs within two watersheds. The modem streams have a general northeast to southwest orientation. A low drainage divide, trending northeast to southwest, separates the two watersheds. The divide extends from north of Stanford, passes between Minier and Hopedale, extends to the southwest of Hopedale, and exits the study area north of Emden. The principal stream in the north half of the study area is the Mackinaw River (fig. 1). It enters the study area north of Mackinaw at an elevation of about 600 feet and exits the study area west of Hopedale at an elevation of about 520 feet. The Mackinaw then joins the Illinois River south of Pekin. Its major tributaries are the Little Mackinaw River, Sargent Slough, and Indian Creek. Sugar Creek, a tributary of Salt Creek, is the primary drainage in the south half of the area. It enters the study area north of McLean at an elevation of about 655 feet and exits southeast of Armington at an elevation of about 595 feet. The West Fork and Middle Fork are its major tributaries. The West Fork enters the study area at an elevation of about 660 feet and exits at an elevation of 585 feet. Its course is diverted to a north-south orientation west of Armington by the LeRoy Moraine. The Middle Fork enters the study area at an elevation of about 678 feet. Its confluence with the West Fork south of Armington is at an elevation of about 585 feet. Tributaries to Prairie Creek drain the southwest side of the LeRoy/ Shelbyville Moraines in the southwest comer of the study area.

Geologic Framework

Bedrock geology Shallow bedrock in the study area is composed mostly of shales and limited thin sandstones, limestones, and coals that make up the upper part of the Pennsylvanian-age Carbondale Formation. Younger rocks are absent. The shales do not yield water. Limited quantities

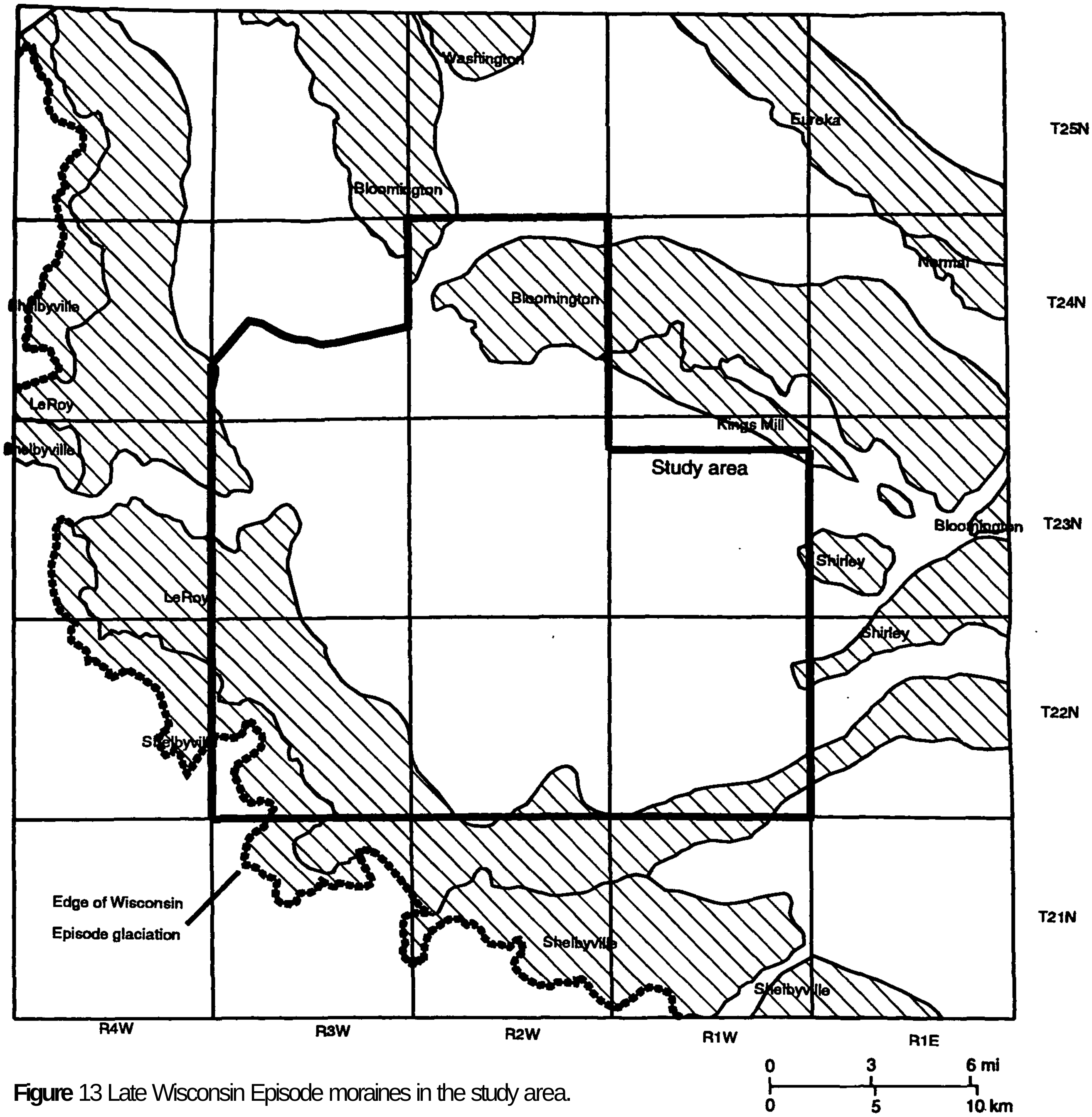


Figure 13 Late Wisconsin Episode moraines in the study area.

of water may be available from the relatively fine grained sandstones and from small, widely spaced fractures in the limestones and coals. These rocks may produce enough water for domestic use, but they cannot provide enough water for a municipal water supply. Mineralization of groundwater increases with depth, so that water found 50 to 100 feet below the bedrock surface may be too highly mineralized for most uses. Because the bedrock does not contain significant aquifers, no further discussion of it is included here.

Bedrock Topography Cutting into the bedrock surface are the Mackinaw and Mahomet Bedrock Valleys, which come together in the confluence area that underlies most of the study area. The configuration of the bedrock topography is of interest because it confines the dominant aquifer in this area. For this reason, many studies have produced maps of bedrock topography. Using well log data available at the time, Kempton and Visocky (1992, fig. 14) showed a broad valley covering most

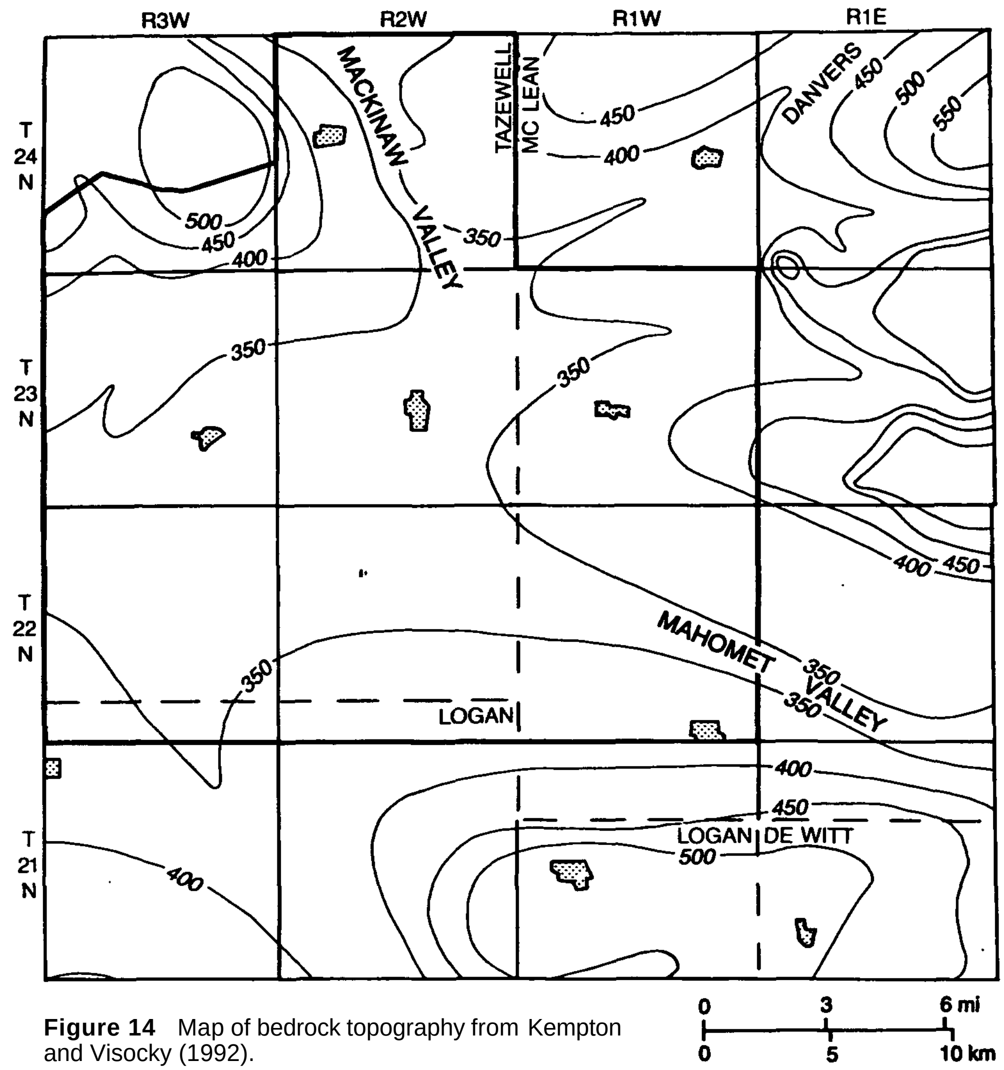


Figure 14 Map of bedrock topography from Kempton and Visocky (1992).

of the current study area. Few of the available logs were for holes that extended to bedrock, and bedrock elevations of less than 350 feet were predicted over much of the confluence area.

Wilson et al. (1994) drilled eight test holes into the bedrock of the current study area, adding some complexity to the interpretation of the bedrock surface (fig. 15). Three of their holes encountered bedrock at elevations of greater than 350 feet in the center of the confluence area; the elevation at Hopedale was above 400 feet. This resulted in a map showing the confluence area as two southwest-trending channels of similar size, divided by a ridge that outlined a feature informally called the "Hopedale high." Each channel had a bottom elevation of less than 350 feet. The center of the Mahomet Bedrock Valley was depicted as a narrow channel with an elevation of less than 350 feet.

The copious seismic and borehole data generated by this project, as well as by concurrent mapping projects in McLean and Champaign Counties and the statewide bedrock topography map of Herzog et al. (1994), have led to a more accurate interpretation of the bedrock surface and a reduction of the contour interval from 50 to 25 feet so more detail can be seen. Figure 16 presents the bedrock topography map developed for this study. The areas east and south of the study area are part of the buffer zone, which is included in work maps to reduce errors and to tie into other regions along the edge of a study area. These areas are included because they help explain the distribution of the Sankoty-Mahomet Sand aquifer within the confluence area. The eastern buffer zone for this project is in the study area or buffer zone for two concurrent projects, so the mapping along the east edge should be more accurate than is usually the case in buffer zones.

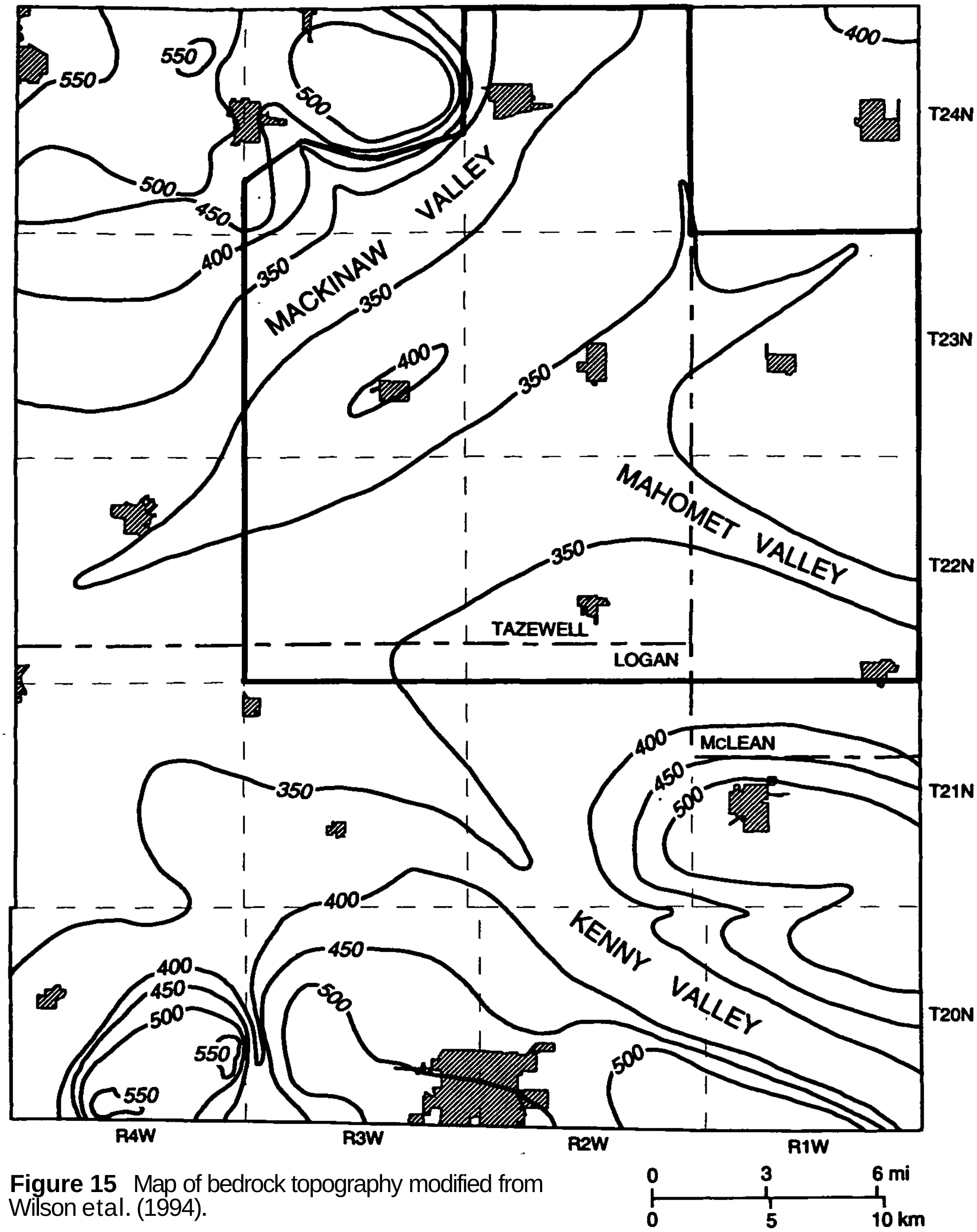


Figure 15 Map of bedrock topography modified from Wilson et al. (1994).

The most significant difference between the current map (fig. 16) and the two previous maps (figs. 14, 15) is the reduction of the area where the bedrock surface elevation is below 350 feet. The remaining areas include a depression southeast of Tremont and a band that roughly coincides with the southern channel shown in figure 15 but that now is considered to be less than 1 mile wide. A thin Mackinaw channel enters the study area from the center of the north boundary and joins the Mahomet channel, which enters from the southeast corner, southwest of Stanford. The composite channel heads west out of the study area north of Emden. Although the deepest part of the channel is narrower than previously thought, the 375-foot contour encloses an area very similar to that of the 350-foot contour of Kempton and Visocky (1992, fig. 14).

Although several hills and depressions occur within the area outlined by the 375-foot contour, the "Hopedale high" is the most significant feature. Test drilling confirmed that the top of this feature has an elevation of greater than 500 feet. The Hopedale high is an elongated hill with a westward-dipping

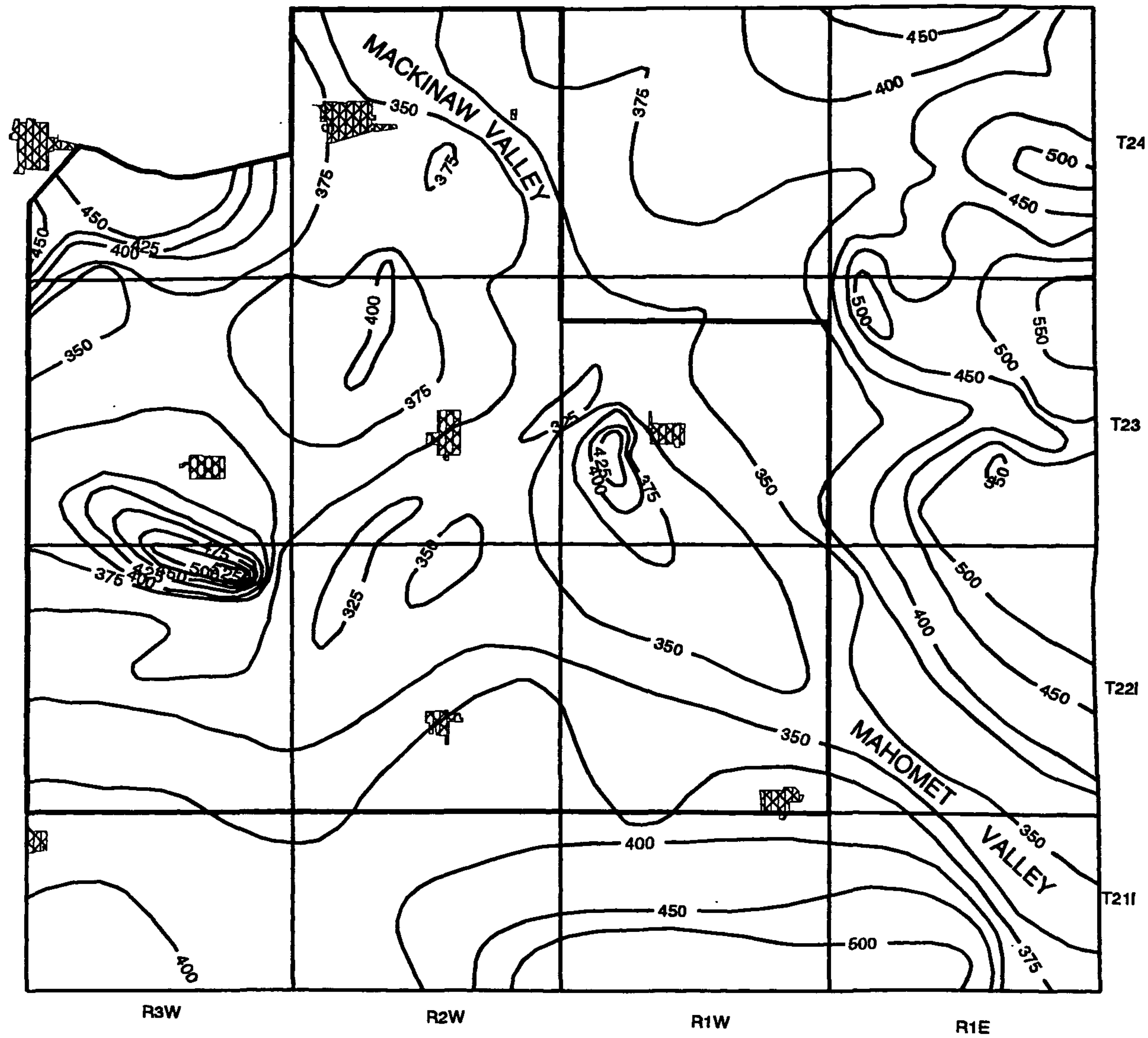


Figure 16 Map of bedrock topography generated from this study (includes buffer zone).

0 3 6 mi
0 5 10 km

crest and very steep slopes on its east and south sides. Two additional hills, originally identified from seismic data (fig. 10), extend to an elevation of greater than 400 feet. One is northwest of Minier and the other is southwest of Stanford. The log of a well recently drilled in Stanford revealed that the Stanford hill extends further east than previous data suggested. The Stanford bedrock high lies in the middle of the channel that joins the Mackinaw and Mahomet Bedrock Valleys. Two small depressions, each with bottom elevations below 325 feet, have also been identified southwest of Minier and west of Stanford.

Glacial Geology The glacial drift that rests on the bedrock ranges from approximately 100 feet to more than 300 feet thick (Piskin and Bergstrom 1975). The drift is thinnest along the sides of the bedrock valley, and over bedrock highs within the valleys, and thickest over the deepest sections of the bedrock valleys. Glacial drift includes diamicton, outwash sand and gravel, and lacustrine sediments. Outwash consists mainly of bedded sand and gravel deposited by glacial meltwater streams. It may occur either in the form of valley train outwash confined by valley walls into long, narrow deposits, or spread over large areas as a flat or gently sloping deposit called an outwash plain. Lacustrine sediments are fine grained clays and silts deposited in relatively quiet water, such as in a lake or slowly moving stream.

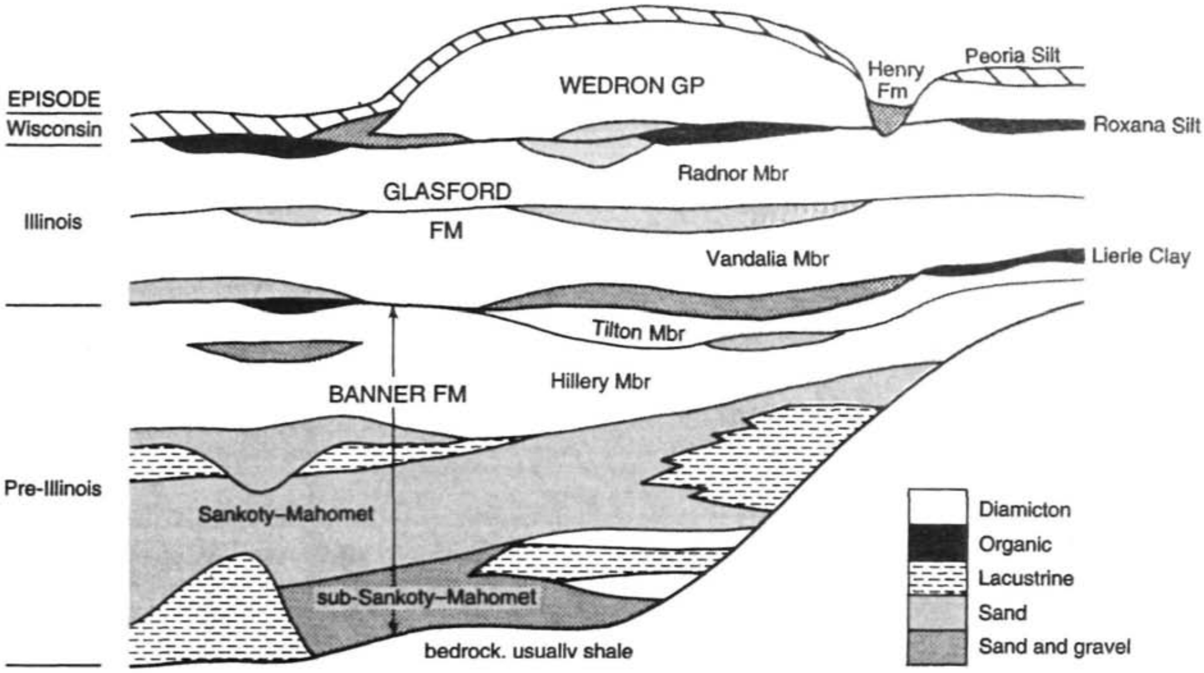


Figure 17 Sequence of geologic material in the study area (modified from Wilson et al. 1994).

As shown in figure 17, the glacial and related deposits were identified, distinguished, and classified in three principal ways: (1) by the physical characteristics of the deposits (lithostratigraphy) with names assigned to the most significant, extensive, and recognizable units (e.g., Sankoty-Mahomet Sand); (2) by the periods of time during which the materials were deposited (e.g., Illinois Episode); and, where present, (3) by buried soils or weathered surfaces (pedostratigraphy; e.g., Lierle Clay). Major buried soils, which indicate long times between glaciations, separate the deposits into bundles. Deposits within these bundles are identified in more detail. Cross sections were prepared to show continuity of units and provide the stratigraphic framework upon which the other geologic maps are based. Six cross sections were selected—two trending west to east and four trending north to south. These cross sections and their locations are shown in figure 18. Cross-sectional lines were selected to make maximum use of information on the geologists' and geophysical logs from this study and the previous study by Wilson et al. (1994). Supplementing these data were the best drillers' logs for wells located along the selected lines.

In the confluence area, the glacial deposits can be grouped into three bundles by age: (1) Banner Formation deposits of the pre-lllinois Episode, (2) Glasford Formation deposits of the Illinois Episode, and (3) Wedron Group deposits of the Wisconsin Episode. These are separated locally by well developed soils and organic horizons. The Lierle Clay separates the Banner and Glasford Formations, and the Robein Silt and Berry Clay separate the Glasford Formation and the Wedron Group.

Distribution of Sand and Gravel Aquifers

Banner Formation The lowermost bundle of units, deposited more than 300,000 years ago, is the Banner Formation. It contains several distinct lithologies and is bounded at the bottom by the bedrock surface and at the top by organic-rich deposits (Lierle Clay, where present) or by younger glacial deposits (Glasford Formation). The Banner Formation includes four geologic units (fig. 17): (1) sub-Sankoty-Mahomet sediments, (2) the Sankoty-Mahomet Sand Member, (3) the Tilton Member, and (4) the Hillery Member. The sub-Sankoty-Mahomet sediments, which are the deepest material and cover most of the valley floor, were recognized and described as a separate unit for the first time in this study. This unit ranges in thickness from 0 feet, where the bedrock is high south of Hopedale and within the valley in the east-central part of the study area, to nearly 150 feet at McLean, where much of it is diamicton. It averages approximately 50 feet in thickness. The sub-Sankoty-Mahomet sediments primarily consist of sand and gravel, referred to in this report as the sub-Sankoty-Mahomet Sand aquifer, overlain and sometimes interbedded with gray diamicton.

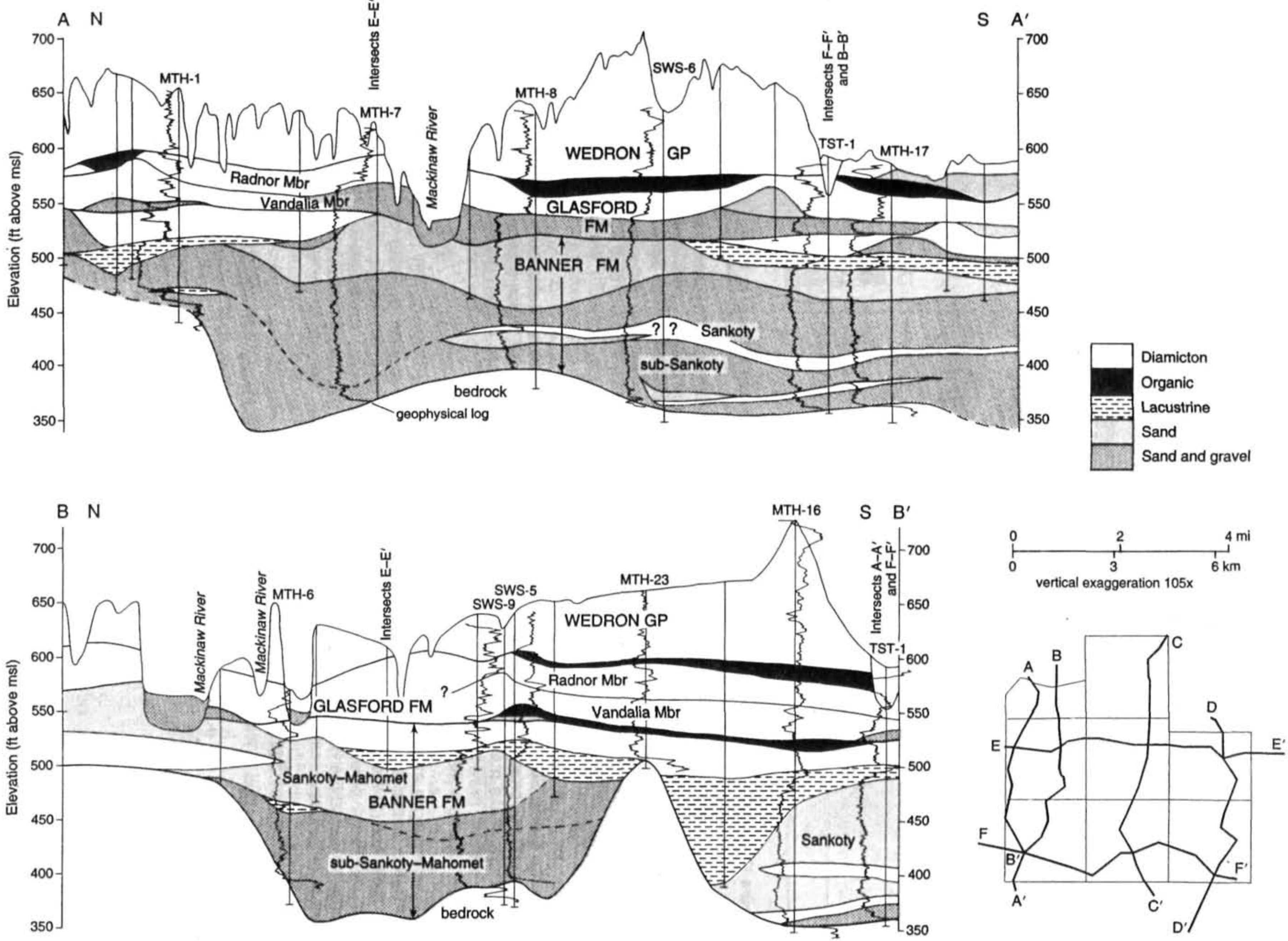
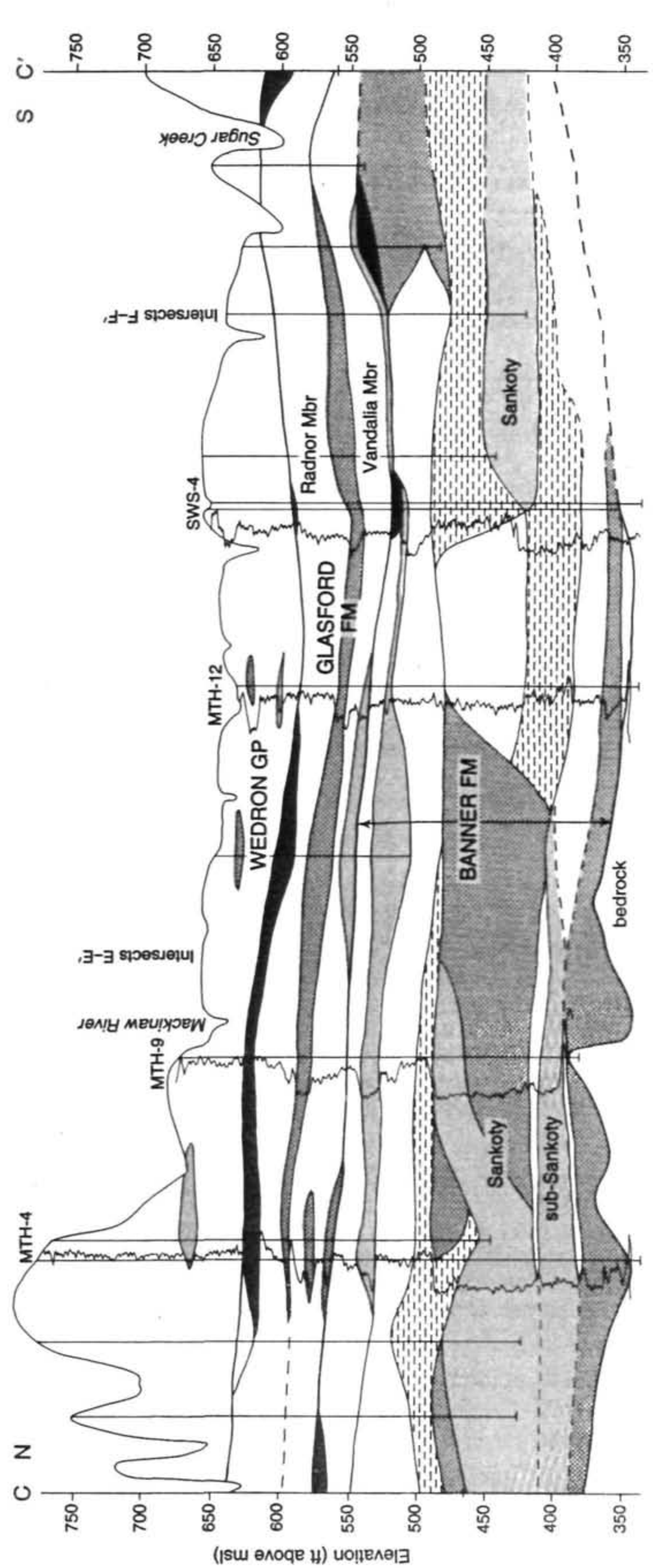
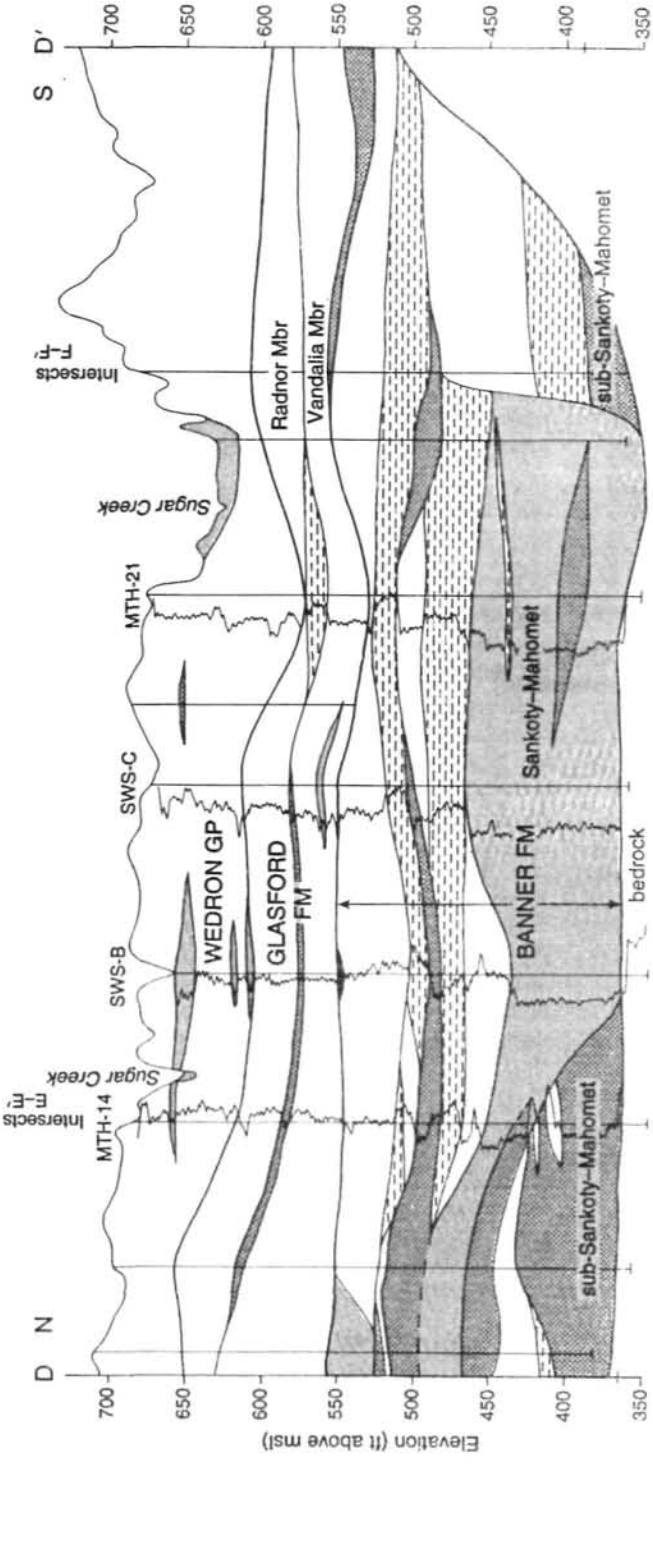
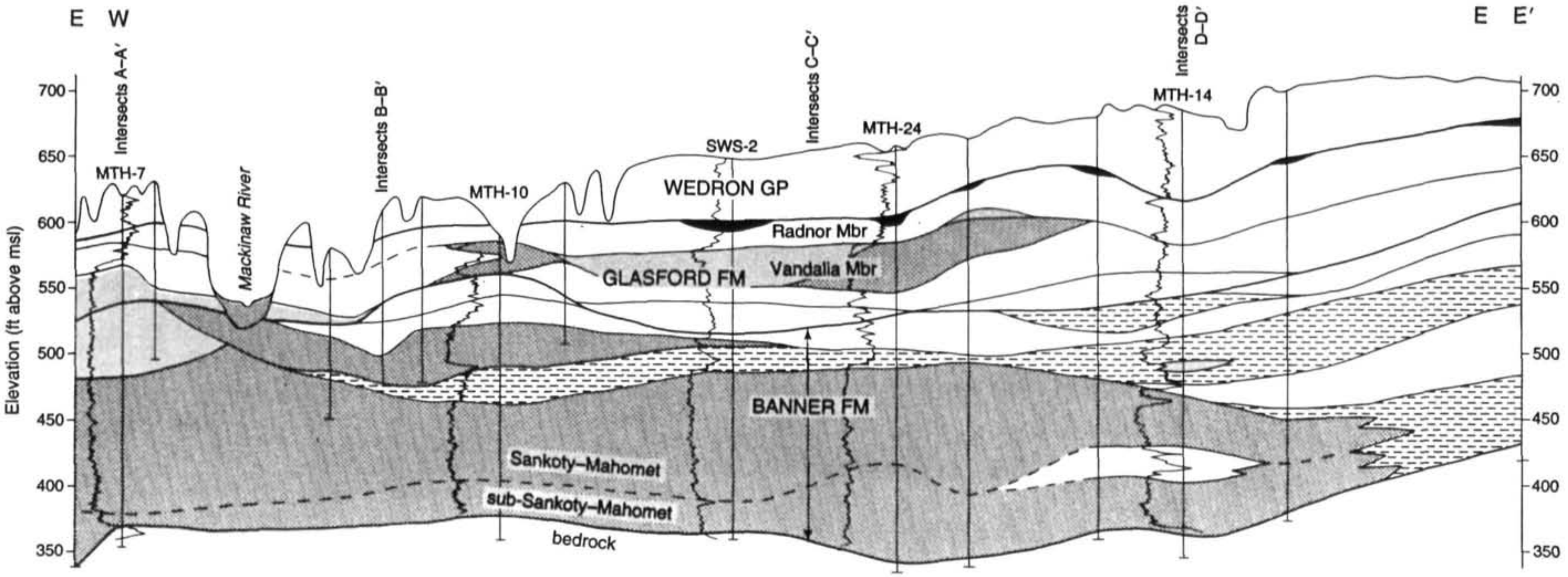


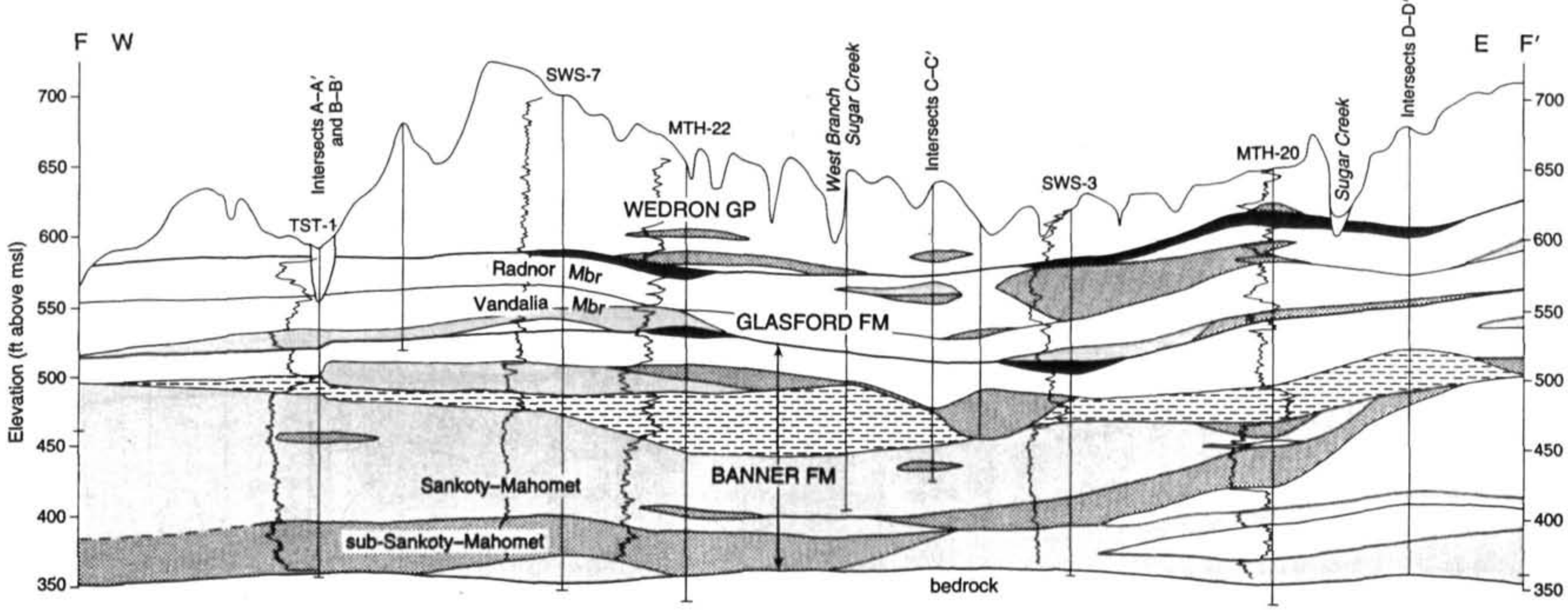
Figure 18 Cross sections through the study area: (A) north-south cross section A-A', through R3W, along west side of study area; (B) north-south cross section B-B', through R3W; (C) north-south cross section C-C', through R2W; (D) north-south cross section D-D', through R1W, along the east side of the study area; (E) northern east-west cross section E-E', through T23N; and (F) southern east-west cross section F-F', through T24N.


C
N
S
C'
Elevation (ft above msl)
750
700
650
600
550
500
450
400
350
MTH-4
MTH-9
Mackinaw River
Intersects E–E'
WEDRON GP
MTH-12
GLASFORD FM
SWS-4
Radnor Mbr
Vandalia Mbr
Intersects F–F'
Sugar Creek
BANNER FM
Sankoty
sub-Sankoty
bedrock


D
N
S
D'
Elevation (ft above msl)
700
650
600
550
500
450
400
350
MTH-14
Intersects E–E'
Sugar Creek
SWS-B
WEDRON GP
GLASFORD FM
SWS-C
MTH-21
Intersects F–F'
Radnor Mbr
Vandalia Mbr
BANNER FM
Sankoty–Mahomet
sub-Sankoty–Mahomet
bedrock



E W
E E′
Intersects A–A′
Intersects B–B′
Intersects C–C′
Intersects D–D′
MTH-7
MTH-10
SWS-2
MTH-24
MTH-14
Mackinaw River
WEDRON GP
Radnor Mbr
Vandalia Mbr
GLASFORD FM
BANNER FM
Sankoty–Mahomet
sub-Sankoty–Mahomet
bedrock
Elevation (ft above msl)
700
650
600
550
500
450
400
350



F W
E F′
TST-1
Intersects A–A′ and B–B′
SWS-7
MTH-22
West Branch Sugar Creek
Intersects C–C′
SWS-3
MTH-20
Sugar Creek
Intersects D–D′
WEDRON GP
Radnor Mbr
Vandalia Mbr
GLASFORD FM
BANNER FM
Sankoty–Mahomet
sub-Sankoty–Mahomet
bedrock
Elevation (ft above msl)
700
650
600
550
500
450
400
350

Where the diamicton is absent, it is difficult to distinguish these sand and gravel deposits from the overlying Sankoty-Mahomet Sand, although the lower unit is generally coarser grained. In southwest McLean County, approximately 30 feet of lacustrine sediment is also present (cross section D-D', fig. 18d). McLean obtains its water from the sub-Sankoty-Mahomet sand aquifer beneath these lacustrine deposits. Fine grained sub-Sankoty-Mahomet material is also present in the center of the study area, in the south half of T23N, R2W and the north half of T22N, R3W, between the largest three bedrock hills (figs. 16,18e).

Above the sub-Sankoty-Mahomet sediments lies the Sankoty-Mahomet Sand Member, a thick sand or sand and gravel outwash deposit. East of the study area, in the Mahomet Bedrock Valley, this unit is called the Mahomet Sand Member. In the Mackinaw Bedrock Valley, both in the northern and western parts of the study area, it is called the Sankoty Sand Member. Where these two sand units join in the confluence area, they are referred to by their combined names, as they are throughout this report. Where the sub-Sankoty-Mahomet material is absent, the Sankoty-Mahomet Sand Member fills the lowermost part of the bedrock valley. A layer of clayey silt to silty sand (lacustrine deposit) typically caps the Sankoty-Mahomet Sand. Along the valley walls and in tributary valleys, where stream velocities were slower, these lacustrine deposits are commonly the predominant lithology.

The Sankoty-Mahomet Sand Member and the sub-Sankoty-Mahomet sand lithologic units together make up the Sankoty-Mahomet Sand aquifer. Because a separating unit is thin or absent in much of the area, these two Banner Formation sands are hydraulically connected and function as a single aquifer. The elevation of the top of this unit is shown in figure 19 and its thickness in figure 20. The Sankoty-Mahomet Sand aquifer is greater than 150 feet thick between Tremont and Hopedale, southeast of Hopedale, and east of Mackinaw, but it thins toward the bedrock valley walls (fig. 20).

The elevation of the top of the Sankoty-Mahomet Sand aquifer ranges from more than 525 feet in the southeast corner of T22N, R2W to less than 400 feet south of Hopedale, where it is interrupted by **a** bedrock hill. The Sankoty-Mahomet Sand appears to be draped over the Mahomet Bedrock Valley; lower top elevations occur directly over the deepest part of the channel. A similar pattern is not obvious over the Mackinaw Bedrock Valley. Where the top of the Sankoty-Mahomet Sand is less than 425 feet in elevation, the sand is usually thin. In these areas, the Sankoty-Mahomet is usually limited by an overlying lacustrine facies. Principal among these areas is a location in the middle of the study area between the largest of the three bedrock hills. The bedrock topography suggests that a lake existed here during glacial times, implying that lacustrine sediments dominated the pre-lllinois Episode.

The Sankoty-Mahomet Sand Member is absent in southwest McLean County, where the sub-Sankoty-Mahomet deposits are thickest, and in the center of the study area, where a thick lens of lacustrine sediments occurs. These areas are shown in cross sections C-C' and F-F' (fig. 18c, 18f) and as depressions in the maps of the elevation of the top of the aquifer (fig. 19) and on the aquifer thickness map (fig. 20). Because the Sankoty-Mahomet Sand Member is absent, the aquifer is less than 25 feet thick in these two areas. Throughout most of the rest of the area, the Sankoty-Mahomet Sand aquifer is 75 to 125 feet thick (fig. 20).

In some areas, to the east and also to the southwest, the Sankoty-Mahomet Sand may grade into relatively fine grained, backwater and lacustrine deposits. Here, fine grained material is prominent in areas where the bedrock elevation is greater than 500 feet. At any given location, the Sankoty-Mahomet Sand exhibits a relatively uniform grain-size horizontally at a given depth, but it may show changes in texture in the vertical direction. It is generally more fine grained than the underlying sub-Sankoty-Mahomet sand. Because the texture of the Sankoty-Mahomet Sand generally coarsens downward, the two sand units can be difficult to distinguish, especially where no intervening diamicton is present.

Two diamictons, the Hillery and Tilton Members, mark the top of the Banner Formation. A locally significant sand and gravel deposit lies between two of these diamictons and at the top of the Banner Formation (fig. 18c-f). Within the confluence area, the Banner Formation reaches a maximum thickness of about 200 feet.

Glasford Formation The middle bundle of deposits is the Glasford Formation (fig. 17), which lies above the Banner Formation and below the Wedron Group. These materials were deposited during the Illinois Episode, 300,000 to 130,000 years ago, and average almost 100 feet thick. The cross

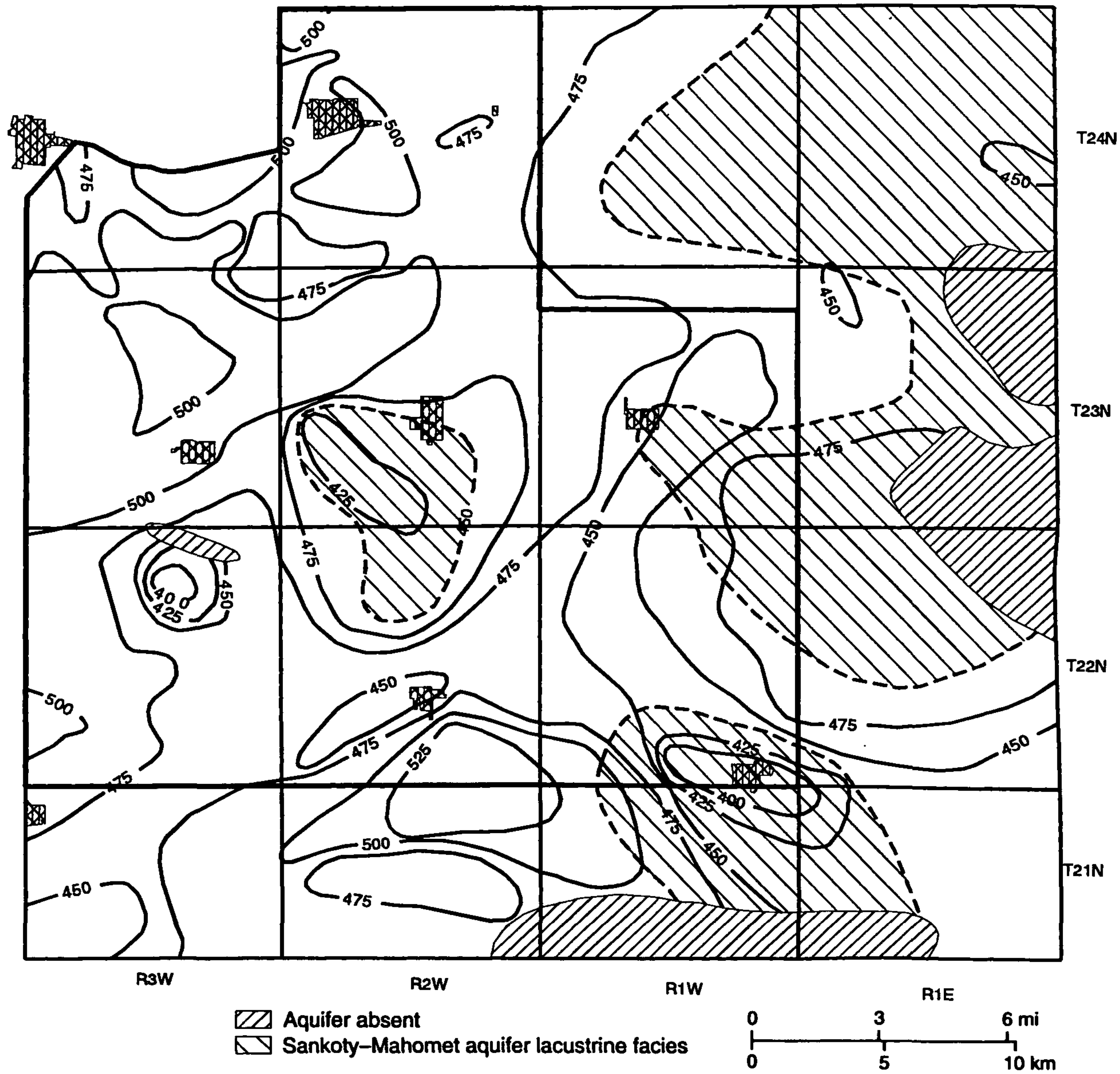


Figure 19 Map of the elevation of the top of the Sankoty-Mahomet Sand aquifer. Contours in feet above msl.

sections (fig. 18) show the succession of units. Although composed predominantly of two diamictons, the Radnor Member and Vandalia Member, the Glasford Formation does contain some locally significant deposits of sand and gravel (fig. 18), usually at the base of the diamicton units. The top of the Glasford Formation is marked by a buried soil (Robein Member), a remarkably persistent black, organic-rich silt (fig. 21), which is recorded in many of the well logs and sample sets available throughout the area.

Sand and gravel outwash is found primarily at two positions within the Glasford Formation: between the diamictons of the Radnor and Vandalia Members and at the base of the Vandalia Member. Figure 22 shows the highest elevation at which Glasford outwash was encountered, according to well records. Sand at the base of the Vandalia Member may directly overlie the Sankoty-Mahomet Sand, forming one thick aquifer. Such is the case in the northwestern part of the study area (fig. 19), where nearly 200 feet of continuous sand and gravel were encountered in MTH-7 (fig.11). Locally, the sand and gravel may separate two parts of the Radnor Member (fig. 18f). Some local channels containing sand and gravel deposits occur at the top of the Radnor Member. Outwash deposits in the Glasford Formation tend to be somewhat thicker and more widespread than the outwash within the overlying

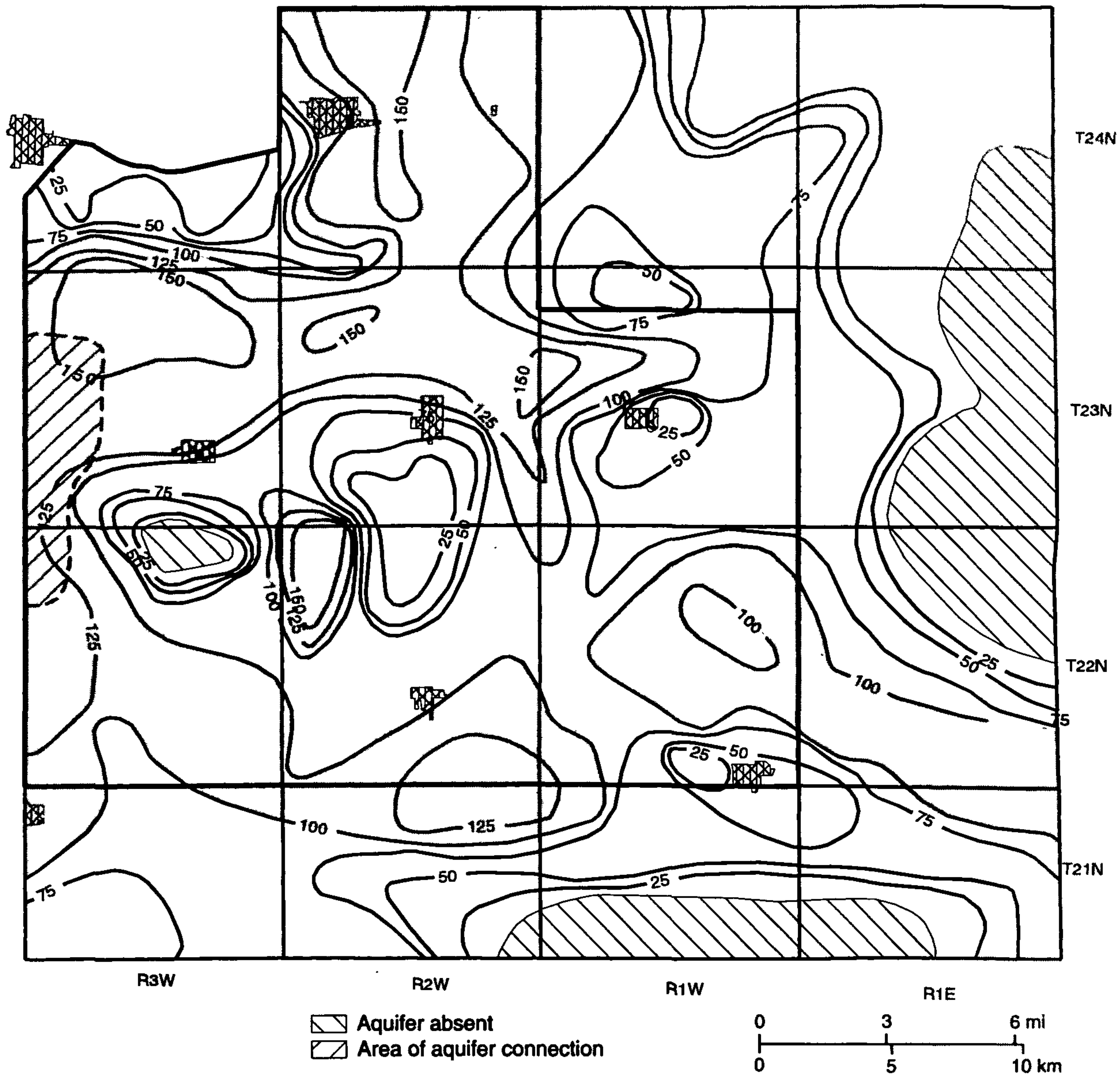


Figure 20 Map of the thickness of the Sankoty-Mahomet Sand aquifer. Contours are in feet.

Wedron Group. Kempton et al. (1991) noted that the upper Banner and Glasford Formation sands and gravels in the confluence area generally have a total thickness of less than 20 feet, thus lacking the adequate thickness and coarse grained texture needed for a large public water supply. Because upper Banner and Glasford Formation outwash sand and gravels tend to be thin and discontinuous in the confluence area, it was not possible to map them individually from the information contained on available logs. Instead, a composite map of the thickness and extent of these sands and gravels was produced (fig. 23). These combined units are more than 20 feet thick in a few locations, generally less than 2 miles wide, but they are scattered throughout the study area.

Wedron Group The Wedron Group, the principal surficial unit throughout most of the study area, directly overlies the Glasford Formation or the Robein Member. The Wedron Group was deposited 25,000 to 12,000 years ago. The southwestern margin of the Wedron Group, consisting of the Shelbyville and LeRoy Moraines (fig. 13), angles across the southwest corner of the study area. Consisting principally of diamicton, the Wedron Group contains only limited sand and gravel deposits (figs. 17,18). The thickness of the Wedron Group is quite variable (fig. 18), ranging from just a few feet to about 100 feet and averaging close to 50 feet.

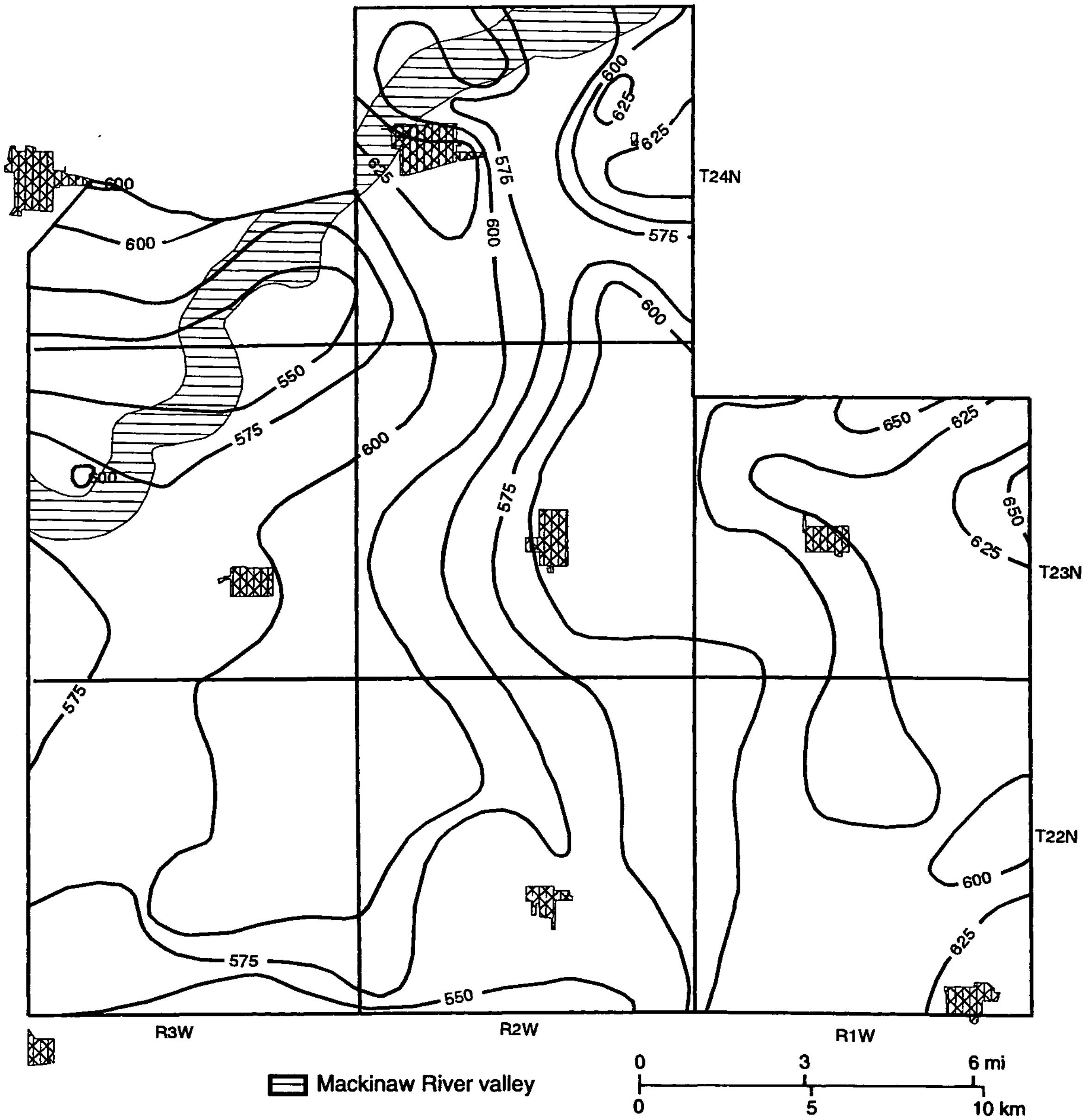


Figure 21 Map of the top of the Robein Silt (top of the Glasford Formation). Contours are in feet above msl.

The lower diamicton in the Wedron Group is the Tiskilwa Formation, a pinkish gray, pebbly clay that extends downward from the land surface to an approximate elevation of about 600 feet. Locally, there may be a thin, younger Wedron diamicton unit above the Tiskilwa Formation. Thin lenses of sand and gravel, the Ashmore Tongue of the Henry Formation, are locally present between the two diamictons. In a few places, a thin sand and gravel is also found at the base of the Wedron Group (fig. 18). Because these deposits are thin and discontinuous, they are difficult to show on a map. These outwash deposits offer no potential for development of a municipal water supply, but they may offer limited potential for supply from a large-diameter bored well. The surficial Peoria Silt covers the entire landscape with as much as 10 feet of clayey silt.

Sand and gravel of the Henry Formation (fig. 17) is locally present along the principal streams of the area (Mackinaw River, Little Mackinaw River, West Fork, and Middle Fork) and along the outer margin of the Wedron Group. While generally thin and restricted in distribution, these sand and gravel deposits may reach 60 feet thick is some areas. Data from this study and from drillers' logs provided insufficient data to map the Henry Formation.

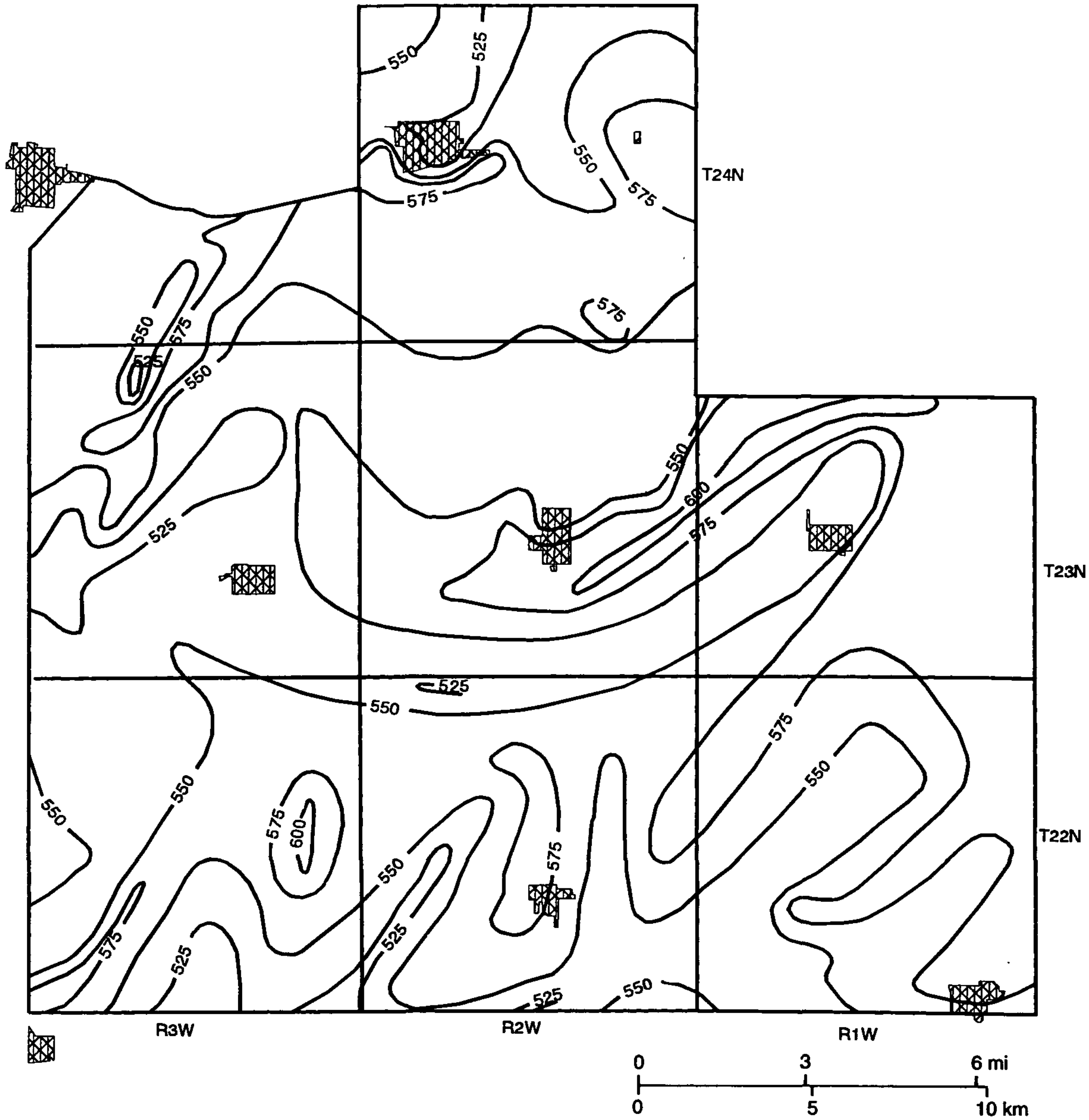


Figure 22 Elevation of the top of the Glasford Formation sands. Contours are in feet above msl.

Revision of Conceptual Model

The data gathered for this study revealed that the configuration of the bedrock surface, which is a major control on the thickness and distribution of the Sankoty-Mahomet Sand aquifer, is more complex than previously thought. The current interpretation combines those given in the two most recent publications on the area (Kempton and Visocky 1992, Wilson et al. 1994). It follows the model of a single broad valley, similar to that depicted by Kempton and Visocky (1992), but their outline of the thalweg has been replaced with a contour line at 375 feet elevation. The remaining area at less than 350 feet defines a much narrower thalweg for the valley (fig. 16). Wilson et al. (1994) noted the nonuniformity of the valley bottom by mapping a ridge between two main channels. This ridge has been replaced with two large hills, one trending northwest to southeast and one trending north to south. Additional hills and depressions have been added. Most notable among these bedrock influences are the bedrock hills south and northeast of Hopedale and the bedrock hill southwest of Stanford. These hills slowed movement of glacial meltwaters entering from the Mackinaw and Mahomet Bedrock Valleys, forming a lake and leaving thick lacustrine deposits in the center of the study area (fig. 20) where sand and gravel previously had been predicted. Similarly, lacustrine

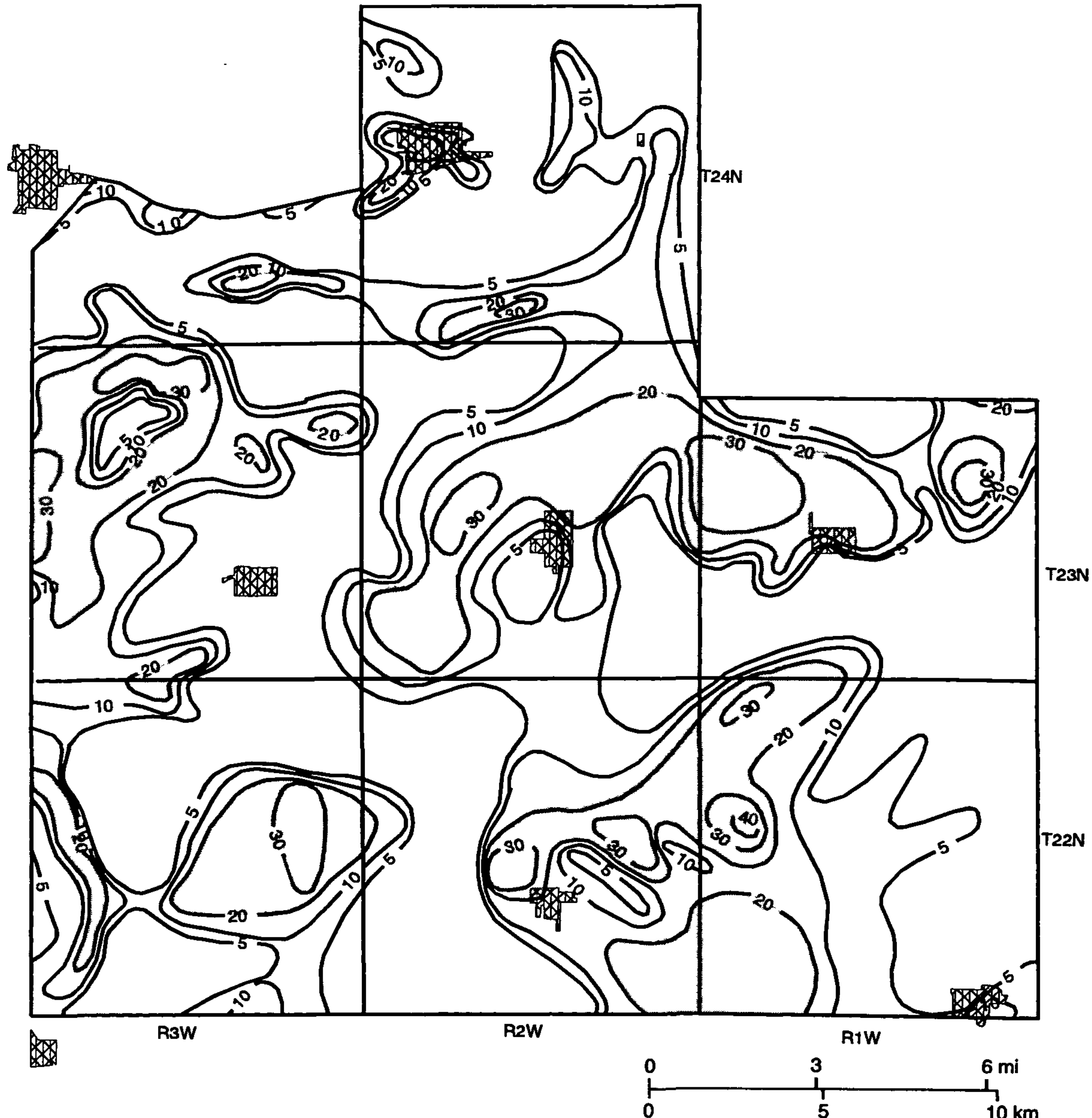


Figure 23 Map of the combined thickness and extent of the Glasford and upper Banner Formation aquifers. Contours are in feet.

sediments were deposited along the edges of the bedrock valley on the east side of the study area and beneath McLean.

As a result of new data gathered for this study, sub-Sankoty-Mahomet sediments were mapped in the confluence area and their significance was noted. Previous interpretations suggested that if a significant thickness of Sankoty-Mahomet Sand Member was present, it extended to the bedrock surface. Sub-Sankoty-Mahomet sand and gravel and finer grained deposits (fig. 18) are present throughout much of the study area. The significant thickness of sub-Sankoty-Mahomet diamicton and lacustrine deposits near McLean (fig. 18d) explains the village's difficulty in finding a municipal water supply. Throughout most of the study area, however, the sub-Sankoty-Mahomet material is mostly sand and gravel that is separated from the Sankoty-Mahomet Sand by less than 20 feet of diamicton.

Although this interpretation changes the current understanding of the shape of the bedrock topography and the distribution of the Banner Formation, the new data help to explain the variability seen in the Sankoty-Mahomet Sand aquifer and make the variations more predictable. The new interpretation does not rule out the Sankoty-Mahomet Sand aquifer as a potential municipal water supply. Areas where the Sankoty-Mahomet Sand aquifer is greater than 100 feet thick occur throughout the study area and are similar to the areas shown by Wilson et al. (1994). Five large areas where this aquifer is greater than 150 feet thick were identified; these areas had not been mapped previously. A sixth area, where the Sankoty-Mahomet Sand aquifer is directly overlain by shallower sand and gravel aquifers, corresponds with the only area identified as having a thickness of greater than 150 feet by Wilson et al. (1994). Thus, although the Sankoty-Mahomet Sand aquifer is thinner in parts of the study area than previously thought, it is much thicker in others. These findings mean that predictions of the size of the Sankoty-Mahomet Sand aquifer remain essentially unchanged, but the variability of its shape and extent will limit where high capacity well fields can be located.

HYDROLOGY

Biweekly Groundwater Level Measurements

Table 5 lists the data for the 39 observation wells (fig. 11) used for this study. Appendix D contains the well hydrographs. Included in the list are the 11 ISWS observation wells (prefix SWS on fig. 11) installed prior to this study.

The purpose of long-term monitoring of groundwater levels is to measure changes over time. This helps to identify connections between aquifers and documents the response of water levels to major climatic events, such as the precipitation amounts causing the floods of 1993. Groundwater levels are typically cyclic over the course of a year. Groundwater levels tend to be highest in the spring when recharge is greatest and consumption is low. During summer, when discharge typically exceeds recharge, groundwater levels tend to decline. The excessive rainfall that generated the flood of 1993 caused water levels to rise throughout 1993.

Well SWS-C (Appendix D) illustrates the cyclic nature of the water levels and also demonstrates the effect that the increased rainfall in 1993 had on the Sankoty-Mahomet Sand aquifer. Initially, groundwater levels dropped through the summer and fall of 1992, in response to increased consumption and decreased recharge. Then, as expected, water levels began to rise late in 1992 and early in 1993 as consumption decreased. In 1993, however, there was a significant period of increased rainfall. This caused water levels in the near surface aquifers (the source of leakage to the Sankoty-Mahomet Sand aquifer) to rise. Instead of declining through the summer of 1993 as they had in 1992, groundwater levels in the Sankoty-Mahomet Sand aquifer kept rising. By the fall of 1993, groundwater levels were nearly 4 feet higher than they were the previous year. As flooding subsided and precipitation returned to normal levels, groundwater levels began to decline. Groundwater levels dramatically declined during the 1994 growing season. By the fall of 1994, groundwater levels were again near the 1992 levels. The rainfall and groundwater level data suggest that the aquifer will respond to varying precipitation conditions. Groundwater levels measured in 1992 and 1994 probably represent the "normal" or typical condition of the aquifer, while 1993 groundwater represent an atypically wet year.

The groundwater level hydrographs support the conclusion that increased rainfall in 1993 affected groundwater levels in the Sankoty-Mahomet Sand aquifer. This was to be expected because the aquifer is under artesian pressure conditions and the pressure response from higher water levels in the near surface materials should transfer to the deep aquifer quickly. As water levels in the shallow aquifers rise, leakage into the deep aquifer will increase.

Water levels measured in the shallow observation wells also suggest a hydraulic connection. A comparison of plots of the shallow and deep water levels at sites with two observation wells generally shows similar fluctuations in water levels between the deep and shallow wells, an indication of hydraulic connection. For instance, at MTH-22, the hydrographs are identical in shape for the deep and shallow observation wells, but their water levels are about 4 feet different in elevation. This situation suggests a strong connection, with the head loss through the aquitard accounting for the difference in elevation. Similarly, at MTH-17, the hydrographs are similar in shape, but the elevation difference is about 10 feet. The difference in elevation is a measure of the head loss through the aquitard; therefore, the connection between aquifers is not as strong at MTH-17. At MTH-24 and SWS-2, the aquifers appear to be poorly connected. The hydrographs at these locations are not alike in shape or in elevation. Part of the reason the shallow aquifer in this part of the study area has a different hydrograph pattern may be that there are many private wells that utilize the shallow aquifer. Over most of the rest of the study area, the shapes of the shallow observation well hydrographs are fairly similar to the deep observation well hydrographs.

Table 5 Location, depth, screened formation, and measuring point elevations for observation wells.

Observation well	Location*	Depth of bottom of screen (ft below land surface)	Formation screened**	Measuring point elevation (ft above msl)
SWS-A	T21NR01W05.4g	317	S-M	678
SWS-B	T22NR01W02.8a	297	S-M	675
SWS-C	T23NR01W26.1C	289	S-M	667
SWS-01	T24NR01W09.4f	352	S-M	751
SWS-2d (east)	T23NR02W10.5a	267	S-M	645
SWS-2s (west)	T23NR02W10.5a	93	G	645
SWS-3d (east)	T22NR02W24.7g	252	S-M	619
SWS-3s (west)	T22NR02W24.7g	57	G	619
SWS-05	T23NR03W26.6f	242	S-M	642
SWS-06	T22NR04W12.1f	237	S-M	638
SWS-07	T22NR03W26.2a	339	S-M	700
SWS-09	T23NR03W26.6f	234	S-M	633
MTH-01	T24NR03W28.4a	172	S-M	653
MTH-02	T24NR03W36.4h	237	S-M	642
MTH-03	T24NR02W04.8C	237	S-M	200
MTH-04	T24NR02W14.4a	387	S-M	776
MTH-05	T24NR02W32.5a	242	S-M	663
MTH-06	T23NR03W02.8g	137	S-M	567
MTH-07	T23NR03W18.3h	230	S-M	626
MTH-08	T23NR03W30.3a	217	S-M	640
MTH-09	T24NR02W35.4a	222	S-M	674
MTH-10d (west)	T23NR03W12.1d	192	S-M	586
MTH-IOs (east)	T23NR03W12.1d	95	UB	586
MTH-11d (south)	T23NR02W19.2e	237	S-M	643
MTH-11s (north)	T23NR02W19.2e	112	UB	643
MTH-13	T23NR02W32.6a	272	S-M	653
MTH-14	T23NR01W15.1a	307	S-M	687
MTH-15	T23NR01W29.4h	237	S-M	665
MTH-17d (north)	T22NR03W29.8a	152	S-M	589
MTH-17s (south)	T22NR03W29.8a	72	UB	589
MTH-18d (west)	T22NR02W18.4d	222	S-M	650
MTH-18s (east)	T22NR02W18.4d	157	UB	650
MTH-19	T22NR01W08.8e	232	S-M	654
MTH-20	T22NR01W29.1h	222	S-M	649
MTH-21	T22NR01W23.1h	297	S-M	675
MTH-22d (west)	T22NR02W31.4d	237	S-M	662
MTH-22S (east)	T22NR02W31.4d	157	UB	662
MTH-24d (north)	T23NR02W12.1a	307	S-M	657
MTH-24S (south)	T23NR02W12.1a	122	G	657
MTH-25	T24NR02W17.1d	307	S-M	684

* Location is identified using the ISGS numbering scheme explained in Appendix A.
**S-M denotes Sankoty-Mahomet Sand aquifer, UB denotes upper Banner sand and gravel aquifer, and G denotes Glasford sand and gravel aquifer.

The observation well hydrographs point out several interesting aspects of the aquifer system. Recharge and discharge in the center of the study area are less variable than they are along the boundaries. The groundwater levels along the Mackinaw River are much more responsive to precipitation than are groundwater levels farther from the river.

Hydrographs of MTH-3 and MTH-4 reflect the pumping test at TST2, conducted from April 5 through May 5, 1995. Although these wells are approximately 3 miles from the pumped well, they clearly show the effects of the pumpage. Groundwater levels in MTH-3 and MTH-4 began to rise in early spring, but then they fell until after the test was completed. In contrast, groundwater levels approximately 4 miles from the pumping test, in wells MTH-5 and MTH-9, began rising in late February or early March and continued to rise throughout the spring.

The data collected at several wells are suspect. A truck hit the well at MTH-14 3 days after installation. The well was repaired and redeveloped twice, but the water level response indicates it may still be partially plugged. Three other wells (MTH-2, MTH-15, and SWS-7) appear to be plugged and not permitting free flow of groundwater into the well. All three wells were developed a second time with only limited success. MTH-20 has had to be developed several additional times due to the presence of precipitated iron in the well. It was such a problem that water level measurements were affected. When the well was developed, the initial surges of groundwater were very discolored with iron oxide because of the relatively high concentration of iron in the groundwater at this location.

Plots of the biweekly measurements of river stage made at three locations along the Mackinaw River (fig. 11) are given in Appendix E. Parallel trends of groundwater levels and river-stage data indicate that the river and groundwater levels fluctuate similarly and are close in elevation on the west side of the study area. The elevation of the river is always higher than the groundwater level in the aquifer, indicating that the hydraulic gradient is downward (toward the aquifer). Wilson et al. (1994) found that the gradient reverses and the river receives groundwater discharge from the aquifer in the area just west of this study area (R4W). As shown in figure 24, water levels in the river are related to rainfall, which is expected because rainfall runs off to the river. During the summer of 1994, river stage declined despite average rainfall. Declines were similar to the groundwater-level declines in the aquifer as shown by the hydrographs of MTH-6 and RVR-1. The decline in the aquifer was most likely a result of increased consumption and decreased recharge, which is the typical seasonal condition. In the river, the decline was most likely a result of increased evapotranspiration and the summer precipitation pattern of thunderstorms.

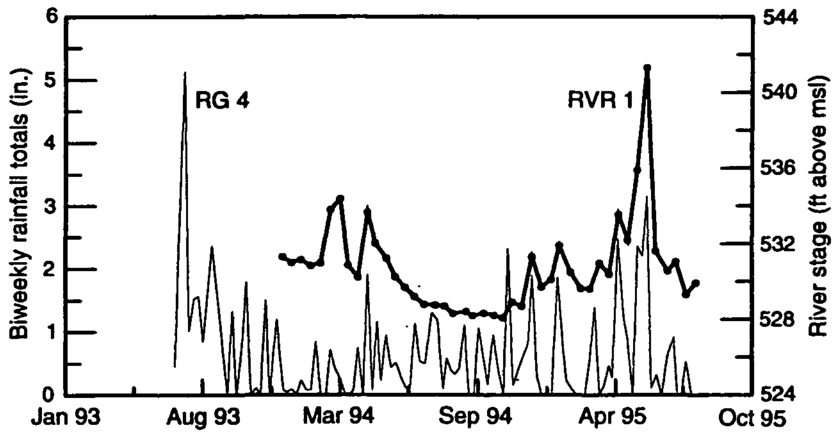


Figure 24 Data from Rain Gauge 4 superimposed on data from river measuring point RVR 1.

Mass Groundwater Level Measurements

The water levels measured in 200 wells in the study area over a 2-week period in August 1994 and July 1995 were plotted and contoured on maps to show the potentiometric surface maps of the Sankoty-Mahomet Sand aquifer for 1994 and 1995 (figs. 25 and 26, respectively). Figures 27 and 28 show the potentiometric surface maps of the Glasford/Upper Banner Formation aquifers for 1994 and 1995, respectively.

The `aSankoty-Mahomet Sand aquifer has a groundwater divide (figs. 25, 26) that trends south-east-northwest across the study area. North of the divide, the groundwater flow direction is north into the Mackinaw Bedrock Valley. South of the divide, groundwater flows west toward the Havana lowlands. The highest groundwater levels were measured in the southeast part of the study area where the potentiometric surface approaches 590 feet. Groundwater levels decrease to about 520 feet in the western portion of the study area near the Mackinaw River. The average groundwater gradient of the potentiometric surface is about 3.5 feet per mile, east to west.

Because the Glasford Formation and the upper part of the Banner Formation include several thin sand and gravel aquifers that are not very extensive, their thicknesses and lateral extent are very difficult to map. As can be seen from the cross sections (fig. 18), two or more of these sand and gravel aquifers may be present at some locations. Consequently, the decision was made to create

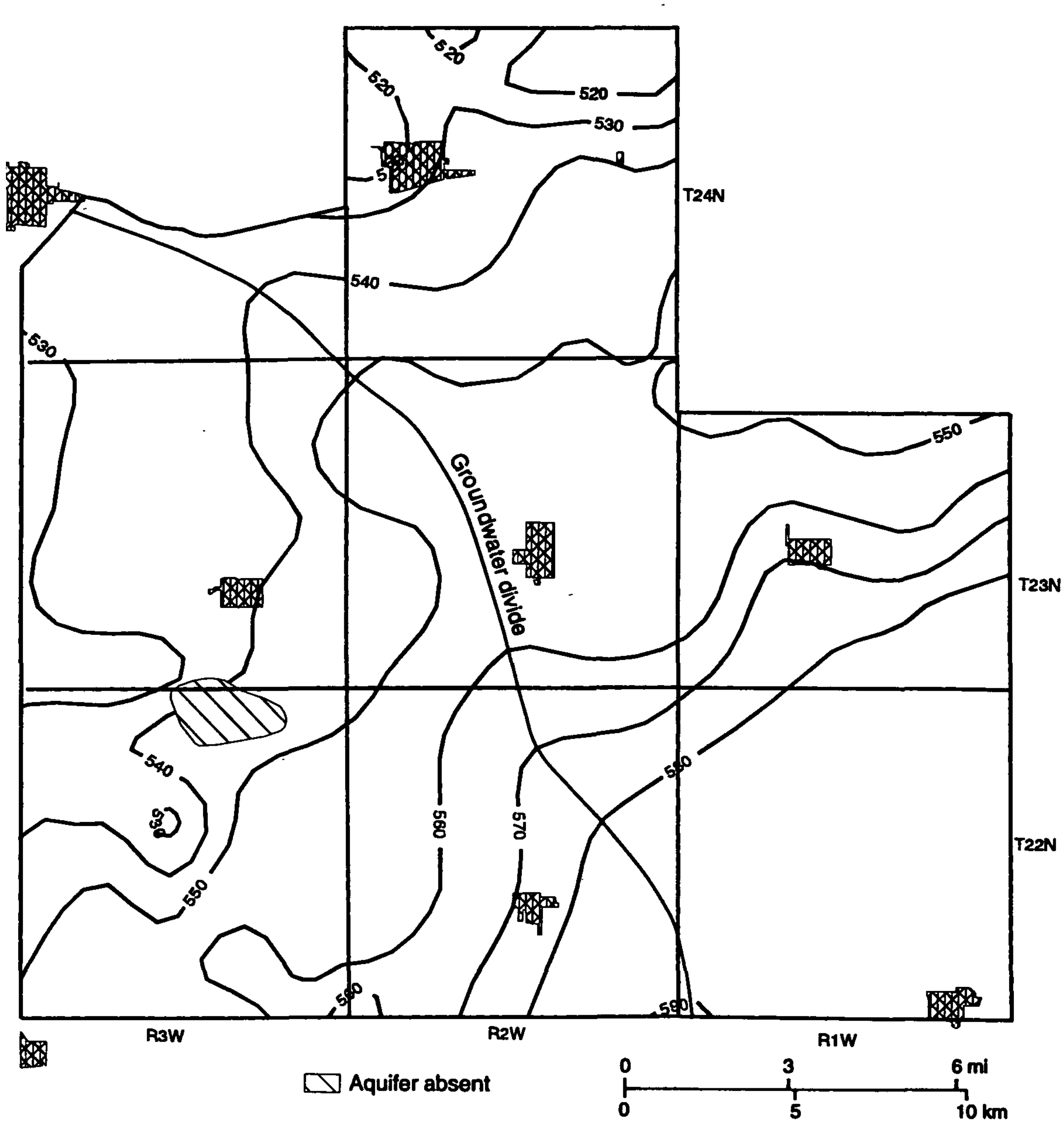


Figure 25 Map of the potentiometric surface of the Sankoty-Mahomet Sand aquifer in August 1994. Contours are in feet above msl.

one potentiometric surface map from the water level data for those wells finished in the Glasford and upper Banner Formations. Anliker et al. (in press) chose a similar approach in their evaluation of data from a mass measurement of water levels in the vicinity of DeWitt and Piatt Counties. Although the resulting map suggests a continuous sand and gravel aquifer within the Glasford and upper Banner Formations, the continuity of the these aquifers has yet to be defined. Presenting the water level information in this manner allows analysis and interpretation of the groundwater resources in a regional context.

The few data points that were uncharacteristically high or low and that could not be explained by physical phenomena were removed. The data removed were assumed to be in discontinuous and/or limited aquifers and not part of the general upper aquifer flow system. Potentiometric surface maps for the Glasford and upper Banner sand and gravel aquifers in the confluence area were constructed by employing the methodology used for DeWitt and Piatt Counties,.

Both sets of potentiometric surface maps (figs. 25-28) show areas where groundwater levels indicate localized flow patterns. These areas coincide with geologic features of the aquifer systems.

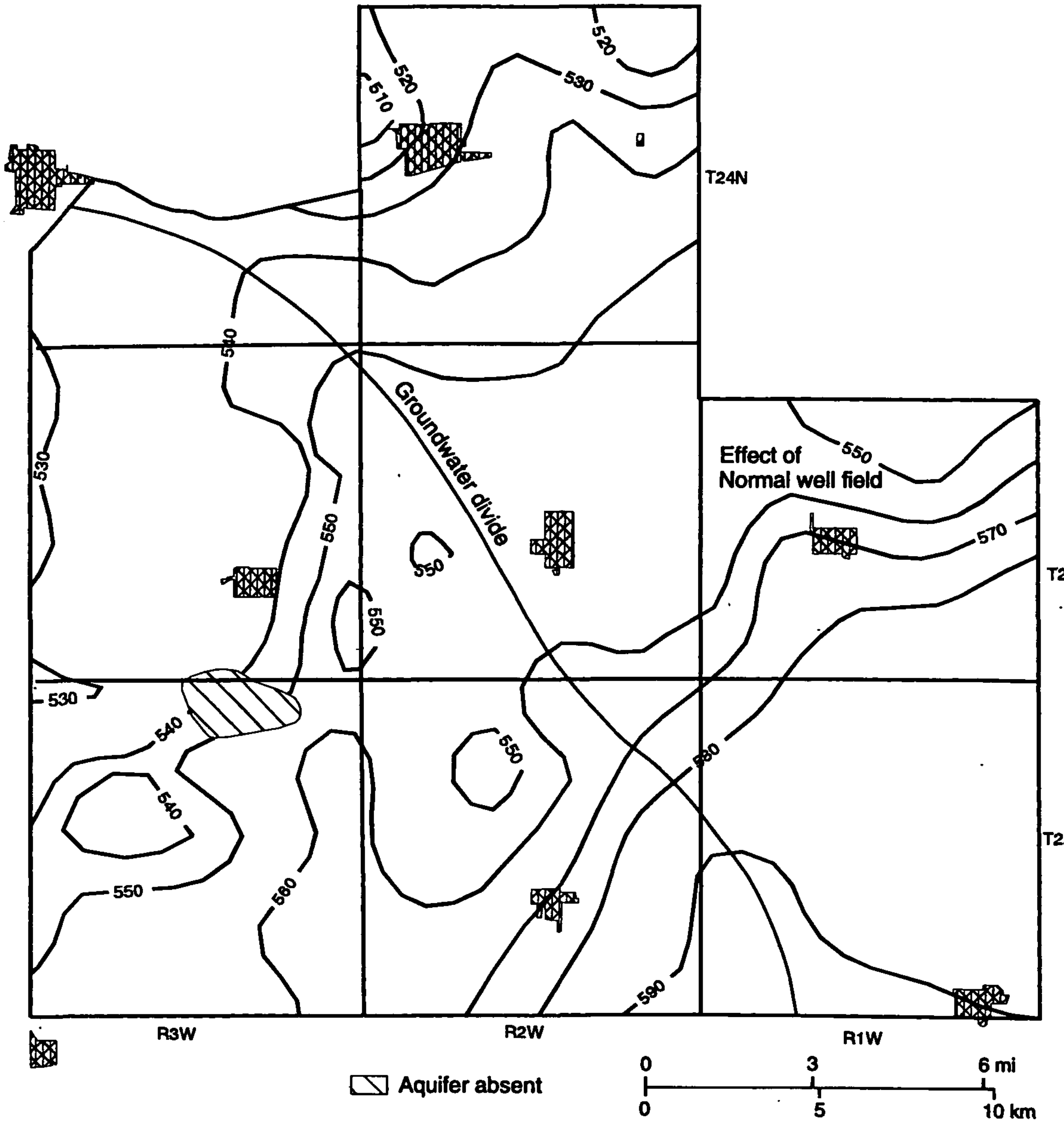


Figure 26 Map of the potentiometric surface of the Sankoty-Mahomet Sand aquifer in July 1995. Contours are in feet above msl.

For example, the wide, flat area within the 560-foot contour in the center of figure 26 coincides with an area where the Sankoty-Mahomet Sand Member is absent and the aquifer is composed only of the sub-Sankoty-Mahomet sand. Thus, the aquifer is thin (fig. 19) and the elevation of its top is lower than it is in the surrounding area (fig. 20).

Although the sand and gravel aquifers of the Glasford and upper Banner Formation are thin in the study area, the depths of shallow wells in the well inventory revealed that many rural wells use these sand and gravel aquifers for their water supply. Water levels in this aquifer ranged from 650 feet to 530 feet above msl. Although this large range suggests a steep hydraulic gradient, it is unlikely that one continuous Glasford and/or upper Banner sand and gravel aquifer is present across the entire study area. The hydraulic gradient across any single sand and gravel unit may be considerably less.

Figures 29 and 30 represent the water level difference between the Glasford/upper Banner Formation sand and gravel aquifers and the Sankoty-Mahomet Sand aquifer for 1994 and 1995, respectively. Comparing water level differences between the aquifers can indicate areas of direct connection between the aquifers and/or the land surface. The greatest differences between water

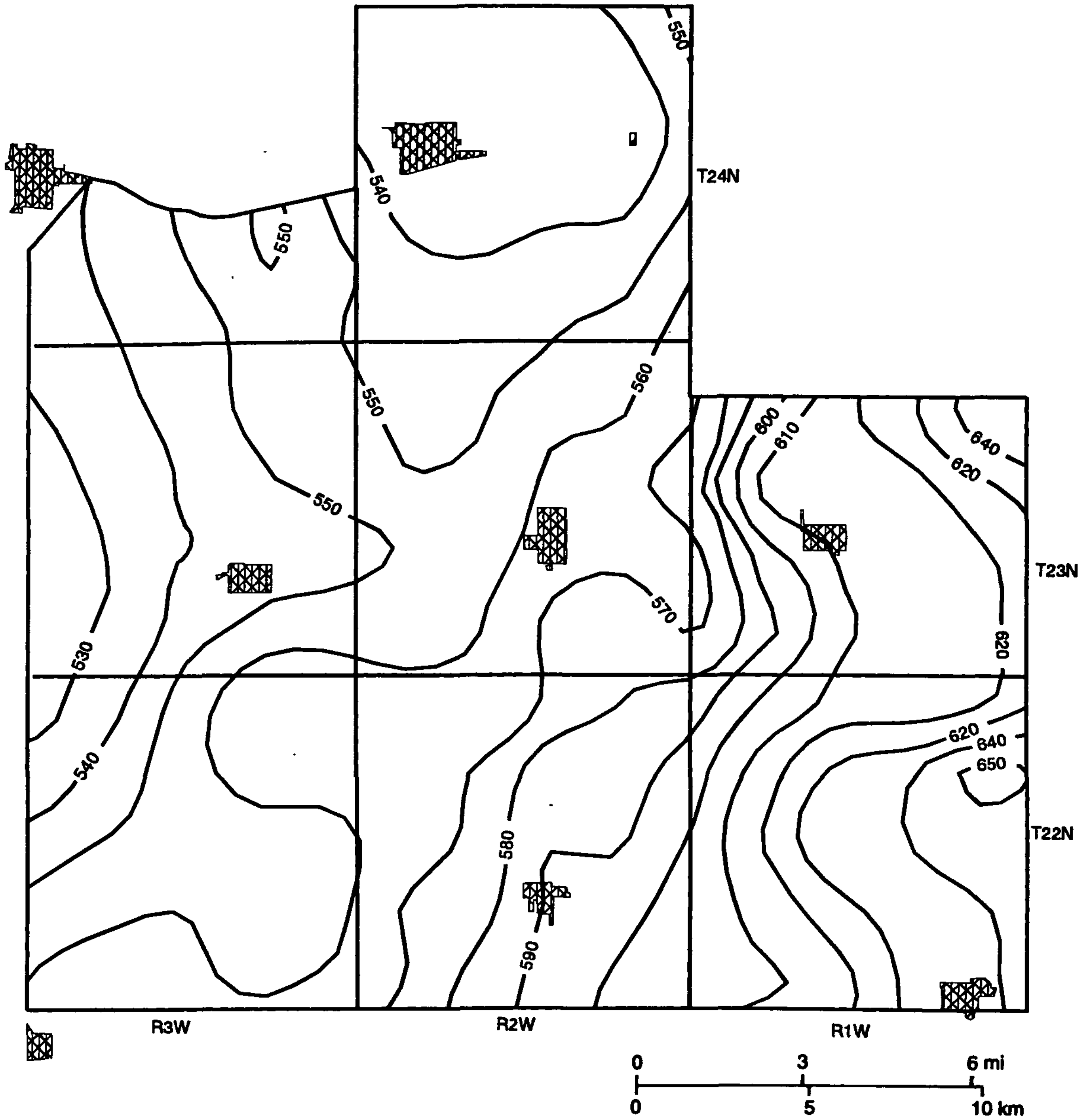


Figure 27 Map of the potentiometric surface of Glasford and upper Banner aquifers in August 1994. Contours are in feet above msl.

levels occurs in the northeast part of the study area, north of Stanford. This difference may be accentuated by the Town of Normal West Well Field, which is located along Highway 9 just outside of the study area. Areas of near-zero water level difference may indicate a hydraulic connection between the aquifers (i.e., that the aquifers function as a single hydrologic unit). The geologic log for MTH-7 (fig. 11) confirms the lithologic connection in that area. There are two areas of greatest hydraulic connection: along the west edge of the study area, where geologic logs also indicate a lithologic connection between the two aquifers, and along the Mackinaw River. The connected areas shown in figures 29 and 30 appear somewhat different because different wells were available for measurement on the two dates.

Aquifer Tests

The historical pump test data for the study area and other nearby sites within the Sankoty-Mahomet Sand aquifer listed in Appendix G include eight tests to determine specific capacity. Specific capacity, the yield of a well in gallons per minute per foot of drawdown, varies with the radius of the well and the duration of pumpage. Walton (1962) developed a modified non-equilibrium formula that

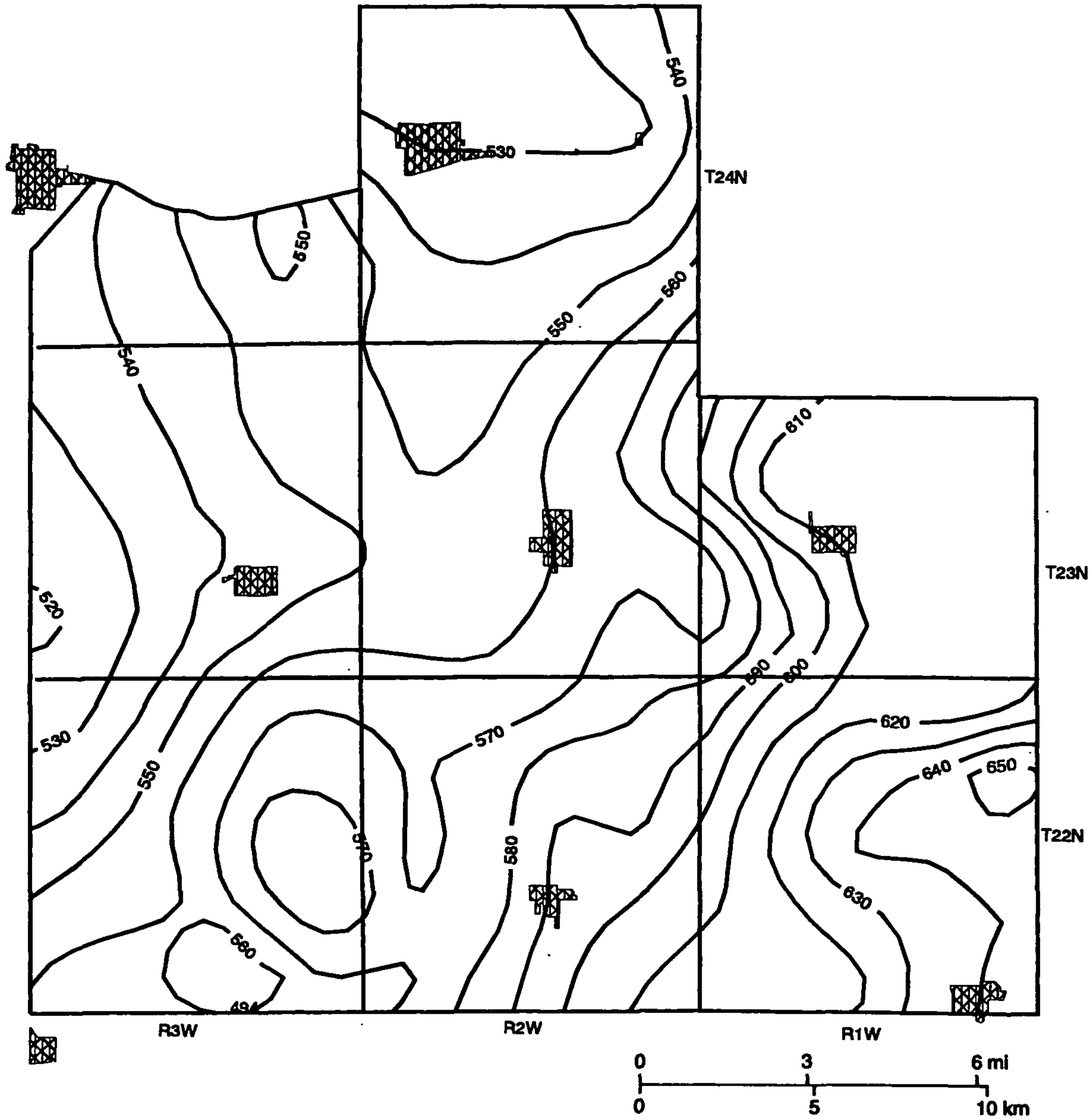


Figure 28 Map of the potentiometric surface of Gtasford and upper Banner aquifers in July 1995. Contours are in feet above msl.

relates specific capacity to transmissivity (7). This formula assumes that the well fully penetrates the aquifer, there is no influence on the effective radius of the well from the drilling, and well loss is negligible. A storage coefficient (S) is assumed. There are several limitations to this method, most notably that the effective well radius is usually not known. Water table conditions also create significant overestimates of transmissivity for short duration tests. Because of the imprecise nature of these analyses, the values for transmissivity in Appendix G determined by the specific capacity method should be used with caution. Typical values of the hydraulic properties for the study area are presented in table 6. The values in table 6 were calculated from the parameters presented in Appendix G. Hydrologic data typically do not conform to a normal statistical distribution. Where the data were log normally distributed, the geometric mean was calculated, and where the data were neither normally nor log normally distributed, the median was determined instead.

Two aquifer tests were completed during this study (TST1 and TST2 on fig. 11). For both tests, production wells were installed in the Sankoty-Mahomet Sand aquifer for the purpose of evaluating

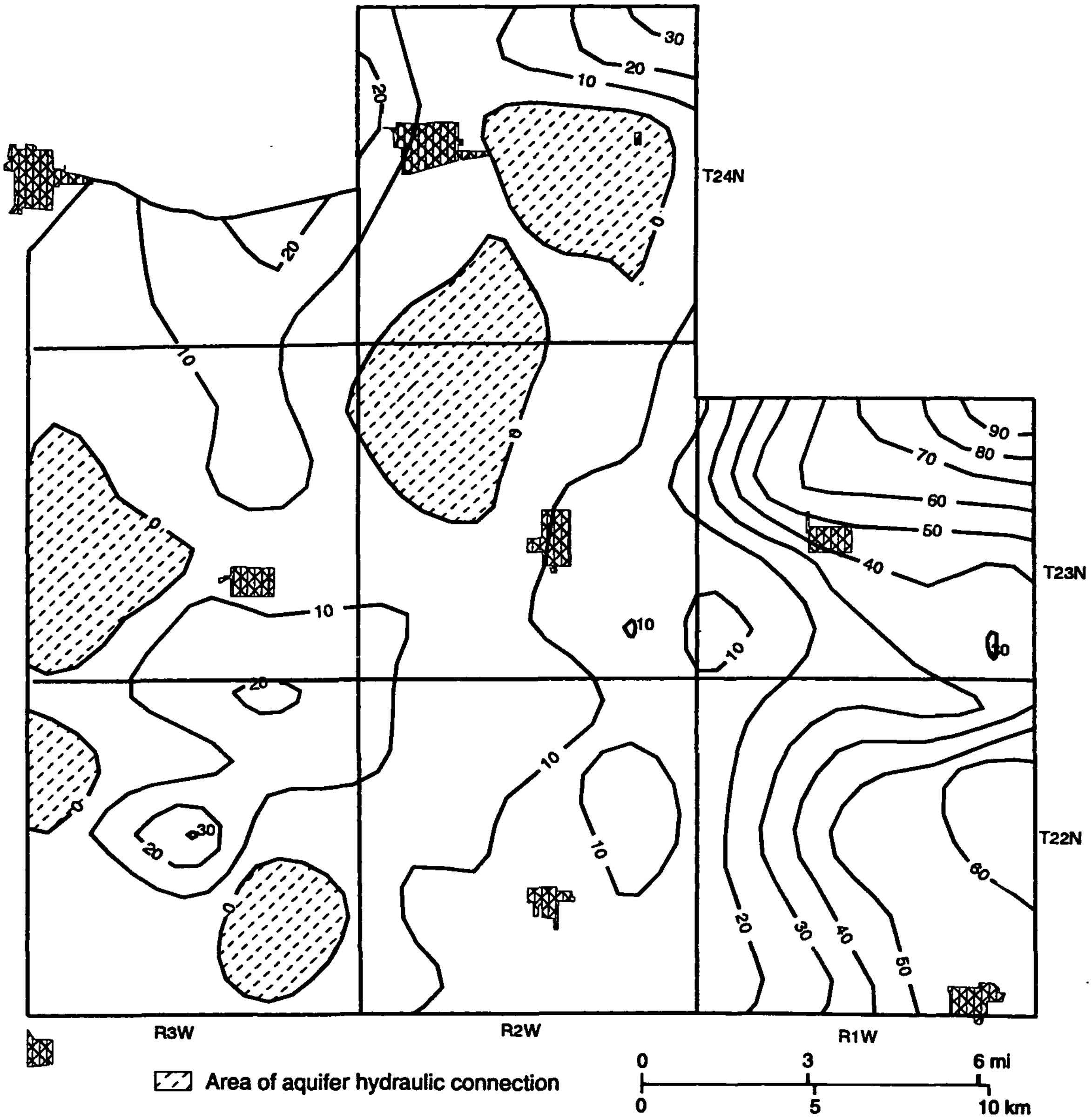


Figure 29 Map showing the difference between the potentiometric surfaces of the combined Glasford aquifers and the Sankoty-Mahomet Sand aquifer in August 1994. Contours are in feet.

the aquifer hydraulic properties. Three tests were originally proposed, but budget constraints forced cancellation of the test planned for the eastern part of the study area.

Aquifer test 1 The first aquifer test was conducted north of Emden at site TST1. Pumping continued for 7 days, between January 18 and January 25,1995. Recovery data were collected from January 25 to January 30, 1995. The pumping rate was held constant at 909 gallons per minute (gpm) for the duration of the test. Water level changes were monitored during the test in three observation wells installed in the Sankoty-Mahomet Sand aquifer and one in the overlying Glasford sand and gravel aquifer. The three Sankoty-Mahomet observation wells, OW1, OW2, and OW3, were located 83 feet, 165 feet, and 387 feet, respectively, in a line south-southwest from the pumped well. The Glasford observation well was about 120 feet south-southwest from the pumped well. A diagram of the TST1 site is included in Appendix H.

The pumped well was 12 inches in diameter and 233 feet deep. Twenty feet of 40-slot screen were installed 213 to 233 feet below land surface. The Glasford sand and gravel aquifer occurs in the interval between 60 and 70 feet below land surface at this site. The Sankoty-Mahomet Sand aquifer

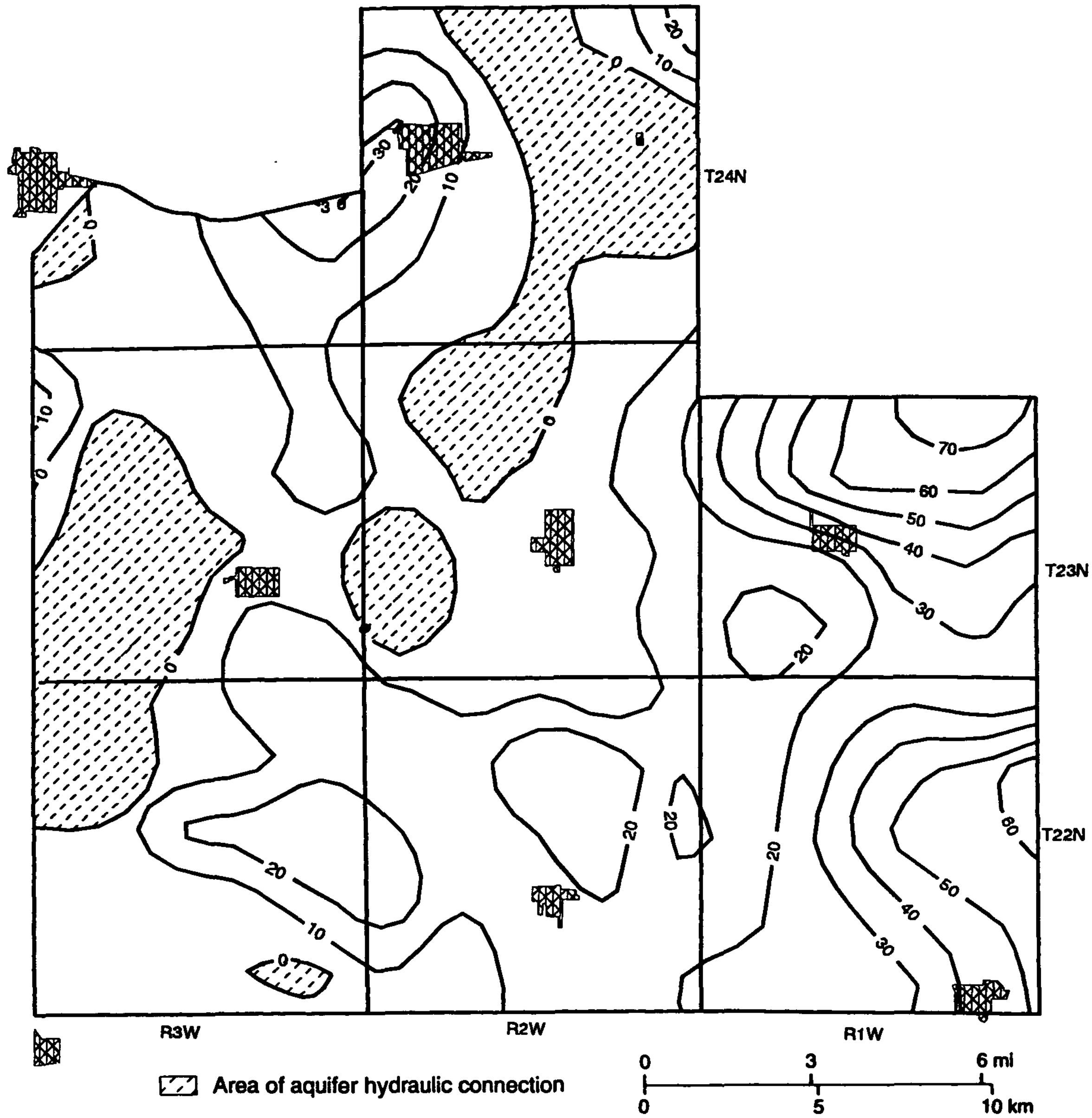


Figure 30 Map showing the difference between the potentiometric surfaces of the combined Glasford aquifers and the Sankoty-Mahomet Sand aquifer in July 1995. Contours are in feet.

was encountered at about 105 feet and extended to the bedrock at a depth of 233 feet below land surface.

No connection was found between the Glasford sand and gravel aquifer and the Sankoty-Mahomet Sand aquifer (i.e., there was no drawdown in the shallow observation well). A summary of results for the three deep observation wells are presented in table 7, and the aquifer test data are presented in Appendix H. The data for OW2 are included, but they should not be considered reliable. During the data analysis, it became apparent that the water level data for OW2 were Inconsistent. Three transducers used during the test were taken back to the site after the test and reinstalled. Data were collected for a week during static conditions to evaluate the transducers. The reovaluation of the pressure transducers indicated that the transducer in OW2 had a plugged vent tube that caused the transducer to produce erroneous data. The accepted results for the TST1 aquifer test (T = 294,000 gpd/ft and $S = 3.0 \times 10^{-4}$) are the average values from OW1 and OW3. Hydraulic conductivity (K) was calculated, using 126 feet as the aquifer thickness, as approximately 2,300 gpd/ft^2. Graphical

Table 6 Summary of aquifer test and specific capacity data. Data from individual tests are presented in Appendix G.

Property	Low	High	Median	Geometric mean
Sankoty-Mahomet Sand aquifer				
Transmissivity (gpd/ft)	6,200	516,600	146,700	136,500
Hydraulic conductivity (gpd/ft^2)	620	4,130	2,300	2,320
Storage coefficient	0.0001	0.09	0.000495	0.000613
Specific capacity (gpm/ft)	3.1	153.4	21.6	23.5
Glasford sand and gravel aquifers				
Transmissivities (gpd/ft)	4,000	121,900	16,800	18,800
Hydraulic conductivities (gpd/ft^2)	290	9,500	1,390	1,670
Storage coefficients	0.00002	0.08	0.00009	0.000216
Specific capacities (gpm/ft)	2.0	71.4	6.1	6.5

analysis indicated that the drawdown followed the Theis curve, indicating there was no measurable leakage.

In addition to these properties, the analysis revealed the presence of a single barrier boundary. A boundary is determined from analysis as an increase in the slope of drawdown curve with time on a semilog plot. Image well analysis suggested the boundary was approximately 2 miles from the site. Geological data from north of the site confirm that the aquifer is pinched off by fine grained material south of the Hopedale high, between 2 and 3 miles away. The boundary will increase the volume of the cone of depression around any high-capacity wells in this part of the study area, resulting in lower water levels than would be expected without such a boundary.

Aquifer test 2 The second aquifer test was conducted near Mackinaw at site TST2. Pumping continued for 30 days, between April 5 and May 5,1995. Recovery data were collected from May 5 to May 10, 1995. The pumping rate was held constant at approximately 1,050 gpm. Water level changes were monitored in three observation wells installed in the Sankoty-Mahomet Sand aquifer

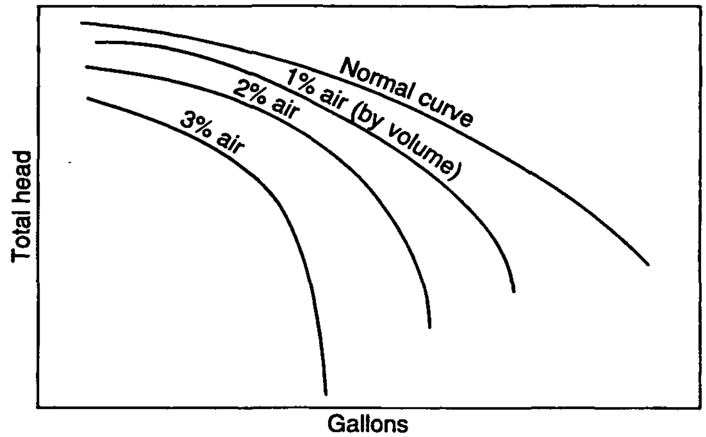
The Glasford sand and gravel well approximately 140 feet from the pumped well was dry and was not monitored during the test. Because the shallow sand and gravel aquifer was well below the elevation of the Mackinaw River, the aquifer was expected to be fully saturated. Therefore, drilling to the bottom of the aquifer was not thought be necessary. It appears that the shallow aquifer is only partially saturated (under water table conditions), and the installed shallow observation well was screened above the saturated portion. The shallow aquifer was encountered at about 121 feet below land surface. The screened interval for the shallow well was 135 to 140 feet below land surface. Logs for the deeper observation wells at the site suggest the bottom of the shallow aquifer is at about 155 feet.

Table 7 Summary of data for the southwest aquifer test (site TST1).

Well	Analysis type	T (gpd/ft)	S
OW-1	log-log	289,000	$1.0x10^{-4}$
	semi-log	293,000	$8.9x10^{-5}$
OW-2	tog-log	264,000	$7.5x10^{-4}$
	semi-log	264,000	$6.4x10^{-4}$
OW-3	log-log	298,000	$5.3x10^{-4}$
	semi-log	294,000	$4.6x10^{-4}$
Average*		294,000	$3.0x10^{-4}$

The values for OW-2 are excluded from the average (see text).

The three Sankoty-Mahomet observation wells, OW1 (MTH-25), OW2, and OW3, were 199 feet south, 317 feet west, and 602 feet west from the pumped well, respectively. The pumped well was 16 inches in diameter and 325 feet deep. Forty feet of 60-slot screen were installed between the depths of 285 and 325 feet. At this location, the Glasford sand and gravel aquifer was present between 120 and 150 feet below land surface. The top of the Sankoty-Mahomet Sand aquifer was at about 198 feet and extended to bedrock at 325 feet below land surface.



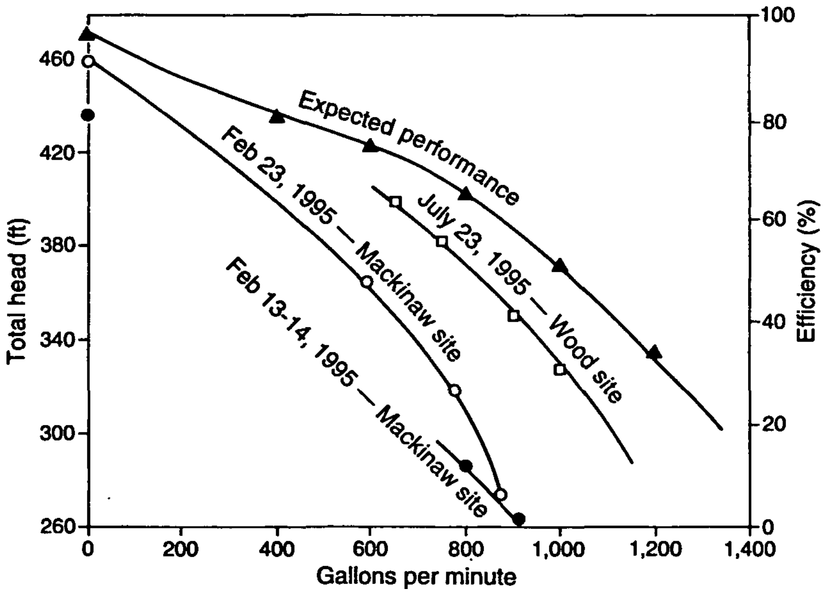


Figure 31 Effects of gas entrainment on pump efficiency in general (top) and for the pump used at aquifer test site TST2.

Several complications delayed the start of the test and made the analysis of the data difficult. The aquifer in this area has a significant amount of naturally occurring gas, mostly methane. This gas came out of solution when the water was pumped to the surface. The gas/water mixture caused a decrease in pump efficiency so that initially the desired pumping rate of 1,100 gpm could not be maintained. Figure 31 shows the effects of gas entrainment on pump efficiency in general and on the efficiency of the pump used for this test. A small amount of gas coming out of solution can create large decreases in pump efficiency. Several trial tests were conducted on the well to determine what operating conditions, if any, would allow an adequate pumping rate to be maintained. Testing revealed that using significant back pressure (i.e., pumping against a partially closed valve) caused the gas to remain in solution as a result of the increased water pressure in the column pipe. After nearly a month of pretesting, the aquifer test began on April 5. Although gas discharge still occurred, the pumping rate was maintained at approximately 1,050 gallons per minute for the duration of the test. Because of a concern that gas entrainment might cause significant changes in the flow rate, the rate was checked at regular intervals and hand measurements of water levels were taken at each observation well during the test.

The hydraulic parameters calculated for test TST2 are as follows: T = 346,000 gpd/ft, K = 2,700 gpd/ft^2, and $S = 6.7 \times 10^{-4}$ (table 8). Table 8 also lists the values calculated lor leakage and vertical hydraulic conductivity of the aquitard at each observation well. The analysis data are presented In Appendix H. These results suggest that leaky artesian conditions exist and that the shallow aquifer acts as the source of the leakage to the Sankoty-Mahomet Sand aquifer. The observation well logs

Table 8 Summary of data for the Mackinaw aquifer test (TST2).

Well	Analysis Type	T (gpd/ft)	K (gpd/ft^2)	S	K'/m' (gpd/rf^2/ft)	K' (gpd/ft^2)
OW-1	log-tog	354,000	2,700	$5.0x10^{-4}$	$2.0x10^{-3}$	0.480
	semi-tog	345,000	2,635	$4.9x10^{-4}$	$2.0x10^{-3}$	0.480
OW-2	log-log	344,000	2,690	$5.3x10^{-4}$	$1.23x10^{-3}$	0.180
	semi-log	337,000	2,640	5.6x10"	$1.23x10^{-3}$	0.210
OW-3	log-log	354,000	2,700	9.8x10"	1.56x10*	0.030
	semi-log	343,000	2,620	9.5x10"	2.27x10*	0.043
Average		346,000	2,660	6.7x10"		

and the production well log also suggest that the shallow aquifer may be discontinuous in the area. The water level data for the Mackinaw River indicate that the river stage is about 40 feet above the elevation of the shallow well screen, suggesting that, because this aquifer is not fully saturated, the shallow aquifer and Mackinaw River are not connected near the test site.

During the test, the cable between the sensor at OW1 and the automated data logger was severed twice. On both occasions, it was repaired, but some data were lost. In both cases, the cable appeared to have been chewed instead of cut, suggesting that an animal damaged the cable. The data for OW3 were collected using an electric water level monitor because the transducer data were corrupted in the field and were unusable. Nevertheless, aquifer properties were determined using data collected from all three observation wells.

Gas conditions A gas sample was collected at OW1 to determine the makeup and amount of gas in solution in the groundwater in this area. In addition, six gas samples were collected throughout the study area as part of the effort to estimate recharge using groundwater chemistry parameters. Figure 32 presents the locations and associated analytical data for the collected gas samples. These data, although not conclusive, warrant attention in any future effort to develop a water supply. This type of information may indicate a potential gas problem in high capacity wells. Both MTH-25 (OW1 at TST2) and MTH-17 (1 mile from TST1) were sampled. At MTH-25, the amount of gas was 106.3 cm^3/L, 66.0 cm^3/L of which was methane. At MTH-17, the corresponding values were 68.5 and 22.3 cm^3/L, respectively. The methane values may be the most important parameter because methane has a much lower solubility than the other constituents in the gas, meaning it will come out of solution before the other constituents. The lower gas values may explain why there was no gas entrainment problem at TST1. The data, if applicable to other locations, indicate there are several regions in the study area where the presence of gas would be a consideration in the design and development of high capacity wells.

Rain Gauge Measurements

The purpose of collecting the rain gauge data from the eleven rain gauges installed throughout the study area was to compare rainfall events to changes in water levels in the aquifers as another tool in determining the degree of connection between the aquifers and the land surface. Measuring river stage proved to be a necessary intermediate step toward comparing rainfall to water level changes. The variability in a rainfall hydrograph, spanning several years, was difficult to use for comparison. It is apparent that river stage is related to rainfall (fig. 24). In addition, the hydrographs in Appendix E indicate the groundwater levels in the Sankoty-Mahomet Sand aquifer are related to the Mackinaw River stage.

Groundwater Chemistry

Inorganic chemistry Groundwater within the Sankoty-Mahomet Sand aquifer in the study area has a chemical signature that is different from the groundwater within the Sankoty Sand aquifer in the Mackinaw Bedrock Valley (Panno et al. 1994). A plot of selected data on a trilinear diagram (fig. 33) indicates distinct water types within the study area. The presence of the distinct groundwater types is the result of rock-water interactions, the discharge of saline groundwater into the Mahomet Sand aquifer from the underlying bedrock, and the mixing of groundwater from the Mahomet Sand aquifer with that from the Sankoty Sand aquifer.

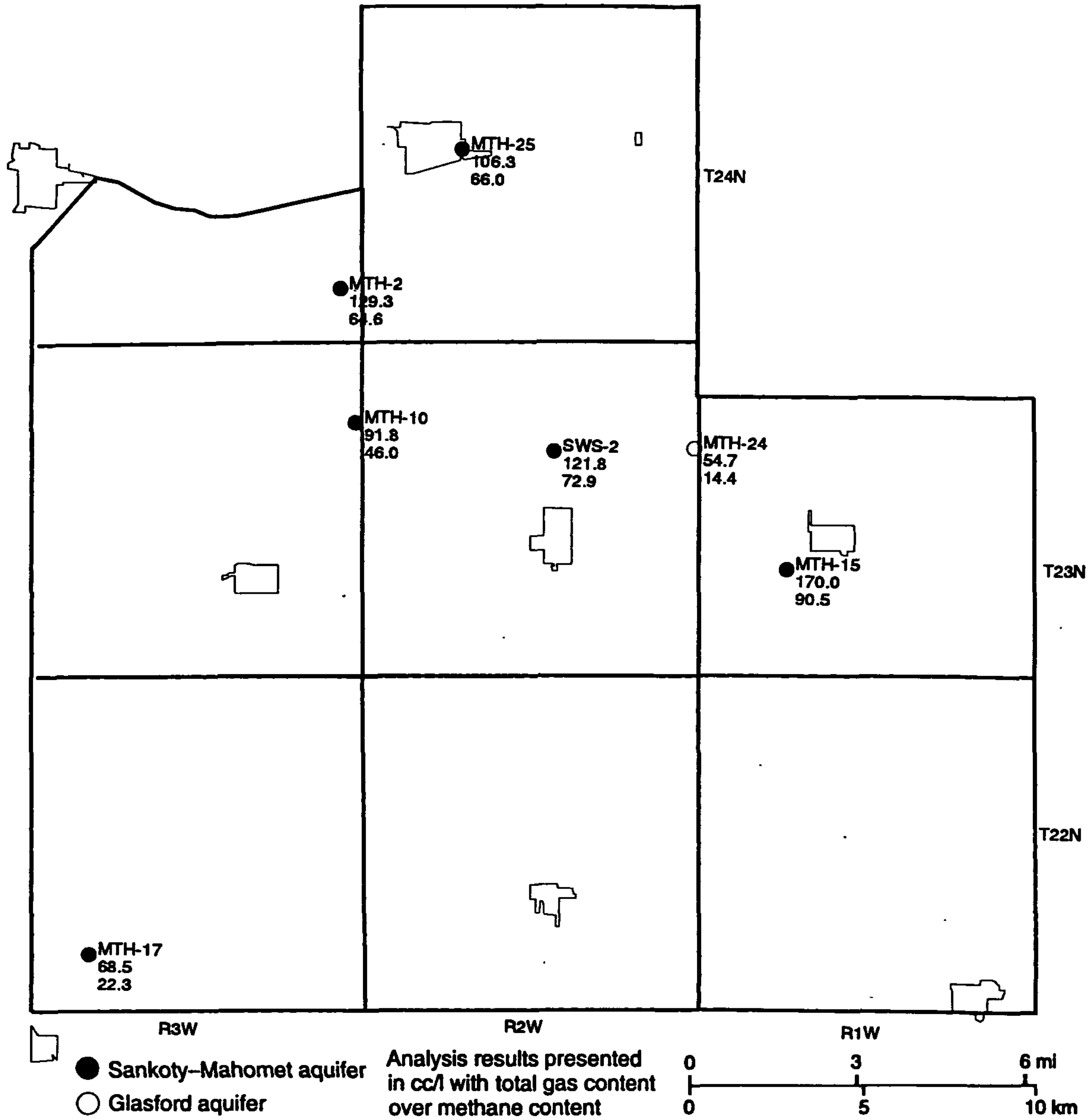


Figure 32 Location and results of gas sample analyses.

Groundwater in the Mahomet Sand aquifer is a mixed cation-bicarbonate ($HCO3^-$) type with characteristically high concentrations of chloride (Cl^-), sodium (Na^+), bicarbonate ($HCO3^-$), arsenic (As), and total dissolved solids (TDS). Chloride concentrations in the Mahomet Sand aquifer stabilize at approximately 70 mg/L. The chloride-enriched water extends into the eastern part of the study area. Chloride concentrations along the groundwater divide indicate the groundwater in the confluence area is the Mahomet Sand aquifer type. Mixing or dilution of the high-chloride groundwater apparently has been minimal along the divide (fig. 34).

Groundwater away from the chloride-enriched water along the groundwater divide is a relatively dilute calcium-bicarbonate (Ca^{2+}-$HC03^-$) type with relatively low concentrations of sodium (Na^+), chloride (Cl^-), sulfate ($SO4^{2-}$). bicarbonate ($HCO3^-$), and total dissolved solids (TDS). This water is quite different from the chloride-enriched, mixed cation-bicarbonate type groundwater of the Mahomet Sand aquifer, but it is similar to the shallower groundwater in the Glasford Formation (fig. 33). There is no indication of a bedrock component in the Sankoty Sand aquifer groundwater.

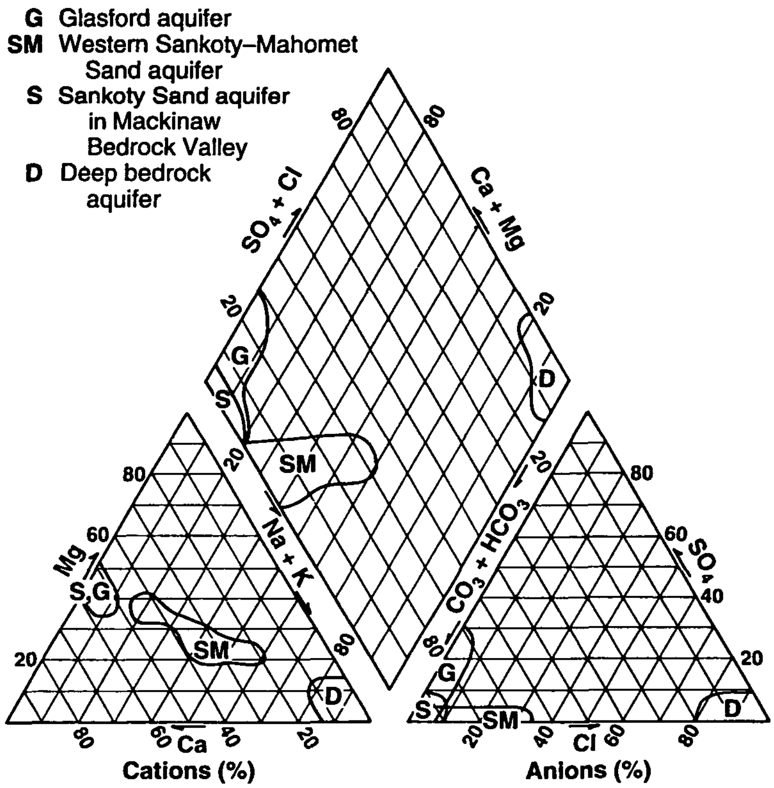


Figure 33 Trilinear diagram showing groundwater types in the Glasford Sand, Sankoty-Mahomet Sand, Sankoty Sand, and deeper bedrock aquifers (modified from Panno et al. 1994).

The distribution of chloride is a natural tracer of groundwater flow in the study area because of the relatively constant chloride concentration of the Mahomet Sand aquifer (71 mg/L) and the relatively small concentration of chloride in the Sankoty Sand aquifer (<3.5 mg/L) away from the mixing zone (Panno et al. 1994). Three distinct types of groundwater were found within the study area by using the graphical technique of Sinclair (1974). Groundwater containing <3.5 mg/L chloride is water that infiltrated through overlying sediments and recharged the Sankoty Sand aquifer. Groundwater with chloride concentrations of 3.5 to 45 mg/L is water of the Sankoty Sand aquifer that has mixed with groundwater of the Mahomet Sand aquifer in the study area. Groundwater with chloride concentrations of >45 mg/L is water of the Mahomet Sand aquifer or Mahomet Sand groundwater that has invaded the Sankoty Sand aquifer water with little or no mixing.

Contour lines of equal chloride concentration coincide with the location of a groundwater divide (dotted line) in the study area (fig. 34; Wilson et al. 1994). The chloride-rich groundwater follows the center of the Mahomet Sand aquifer where it enters the Sankoty Sand aquifer and extends the length of the groundwater divide. It is apparent from figure 34 that the chloride concentrations in the groundwater are diluted rapidly south of the divide.

Contours of chloride concentrations in the study area indicate groundwater flow directions, dilution effects of recharge through overlying sediments, and the effects of transmissivity on the movement of the chloride-rich groundwater. Dilution of the chloride-rich groundwater north of the divide occurs over a much larger area than south of the divide. This relationship appears to be related to transmissivity of the Sankoty Sand aquifer in the northern part of the study area. Groundwater south of the divide mixes with a larger volume of recharge than groundwater north of the divide. Flow lines, based on potentiometric surface maps of the study area (figs. 25-28), show greater divergence south of the groundwater divide and indicate a rapid dilution of solutes.

Isotope geochemistry The isotopic values for the 10 groundwater samples collected from in the study area are presented in table 9. The δD and $\delta^{18}0$ data plot close to the global meteoric line (fig. 35) and surround the average value for modem meteoric water of northern Illinois (5D ≈ -46%o, 6 0

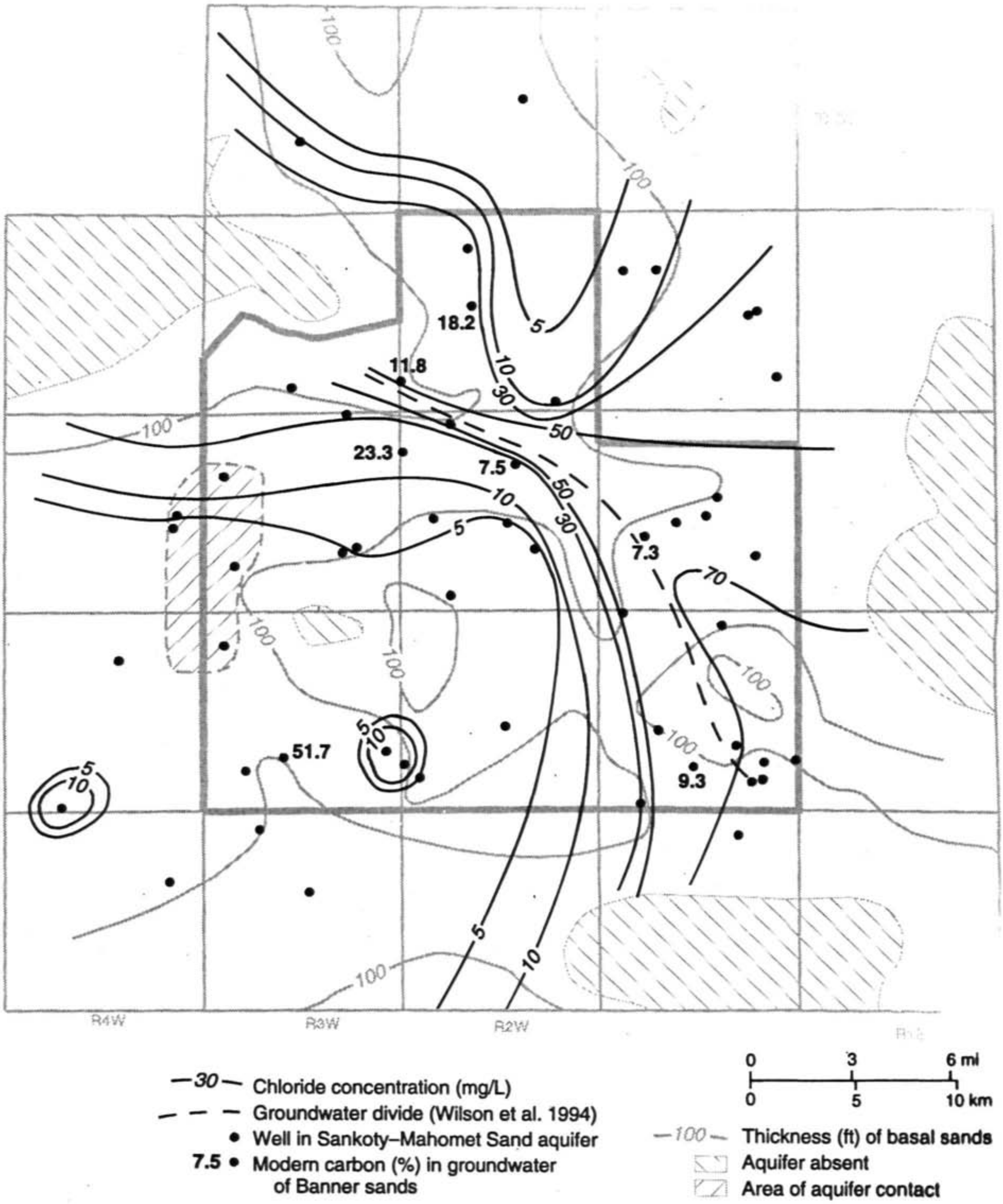


Figure 34 Contour map of chloride in the confluence area of the Mahomet Sand aquifer and the Mackinaw Bedrock Valley aquifer (modified from Panno et al., unpublished).

≈ -7‰; Graf et al. 1965). These results suggest that groundwater in the Sankoty Sand aquifer originated as precipitation under climatic conditions similar to those present today. The 6D and 6 0 results indicate an upper age limit of approximately 14,000 years for this groundwater. Water depleted in the heavy isotopes (D and ^{18}0) is assumed to be older than 14,000 years and associated with the colder glacial periods of the Pleistocene, which would result in more negative isotopic values relative to the average isotopic composition of recent precipitation in Illinois.

Table 9 Isotope concentrations from water well samples.

SITE	Isotopic Values: δD(‰)*	δ18 O (%o)	Dissolved Inorganic Carbon* δ13C(%o)	14 C (pMC)	Tritium (Miami) (TU)	Tritium (ISGS) (TU)
MTH-2	-45.0	-6.99	-2.4	11.8		0.21
MTH-10d	-43.3	-6.46	-4	23.3		
MTH-10S	-42.8	-6.82	-13.68	69.9		
MTH-15	-50.3	-7.98	-3.2	7.26		0.21
MTH-17d	-43.0	-6.78	-3.94	51.69		
MTH-24	-44.0	-7.17	-1.3	40.67		
MTH-25	-42.5	-7.84	-1.61	18.15		
SWS-2d	-46.9	-7.00	-3.8	7.45	0.11	0.38
SWS-2S	-43.7	-6.94	-1.3	38.20	0.013	
Sauder-17	-49.8	-7.41	-9.9	62.97	0.11	0.013

* %o denotes parts per thousand; pMC denotes percent modem carbon; TU denotes tritium units; location of lab is given in parentheses.

The carbon isotopic composition of the dissolved inorganic carbon (DIC) in the Sankoty Sand aquifer is surprisingly similar in all samples, ranging from -3.9 to - 1.6%o. These samples are enriched in the heavy isotope of carbon (^{13}C) and reflect the influence of microbial methane production in this area. Isotopic measurement of the methane in this area yielded $\delta^{13}C$ and 5D values that range from -77.9 to -80.5%o and -219.2 to -229.7%o, respectively. These values are typical of Illinois "drift gas" formed by microbial reduction of CO2 (Coleman et al. 1988). The amount of dissolved methane in the Sankoty-Mahomet Sand aquifer ranged from approximately 0.9 to 3.7 mmol/L. Aquifers in the shallower Glasford Formation contained approximately 0.1 to 0.7 mmol/L of dissolved methane.

During methane formation by microbial processes, isotopically lighter carbon is preferentially incorporated into the methane, resulting in an enrichment of ^{13}C in the remaining DIC. The carbon composition and the concentration of inorganic carbon can be strongly affected by microbial methane production within an aquifer (Barker et al. 1979, Grossman et al. 1989, Hackley et al. 1992). The carbon geochemistry in the study area appears to have been significantly influenced by the formation of microbial methane. Figure 36 shows a positive correlation between concentrations of bicarbonate and methane. Figure 37 shows a negative correlation between ^{14}C activity and methane concentration. Such influences must be considered when groundwater ages are calculated.

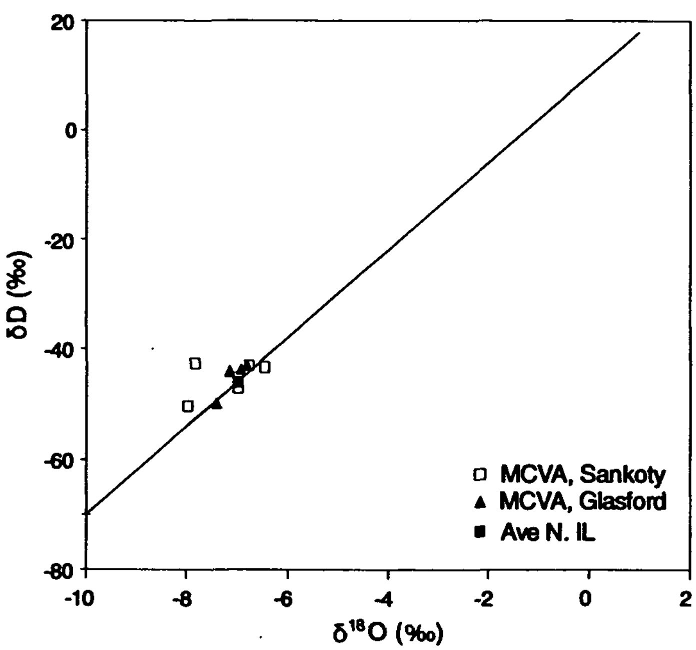


Figure 35 Plot of 5D versus δ^{18}O in groundwater samples from the aquifers in the confluence area compared with average meteoric water of northern Illinois and the global meteoric line (from Panno et al., unpublished).

The results of the ^{14}C analyses of the 10 groundwater samples range from approximately 7 to 70 pMC. With the exception of MTH-17d, shallower groundwater generally exhibits higher ^{14}C activities. The relatively high ^{14}C content at MTH-17d indicates that this location has probably received a significant component of younger water. MTH-17d is located in the southwestem portion of the study area, beyond the groundwater divide and zone of groundwater mixing between the Mahomet and Sankoty Sand aquifers. The Sankoty-Mahomet Sand aquifer in the vicinity of MTH-17d is 100 feet closer to the land surface than it is at sites to the north. This may help explain the relatively high ^{14}C activity in the area of MTH-17d. The Wedron Group thins significantly in this part of the study area and is absent just to the south and west of MTH-17d.

The ^{14}C data from groundwater of the Sankoty-Mahomet Sand aquifer are in good agreement with the contours of chloride concentrations (fig. 34). The lowest ^{14}C values are found in the Sankoty-Mahomet Sand aquifer and along the groundwater divide (fig. 34). The ^{14}C activity increases significantly north and south of the divide, indicating an influx of water with a higher ^{14}C content (i.e., younger water). These data suggest that the Sankoty Sand aquifer in the study area is not as isolated from the shallow aquifers and surface recharge as is the Mahomet Sand aquifer.

Preliminary tritium results indicate that no significant amounts of tritium have penetrated into aquifers in the Glasford Formation or into the Sankoty-Mahomet Sand aquifer within the study area. The tritium data indicate that groundwater in sediments of the Illinois and pre-Illinois Episodes is older than 50 years.

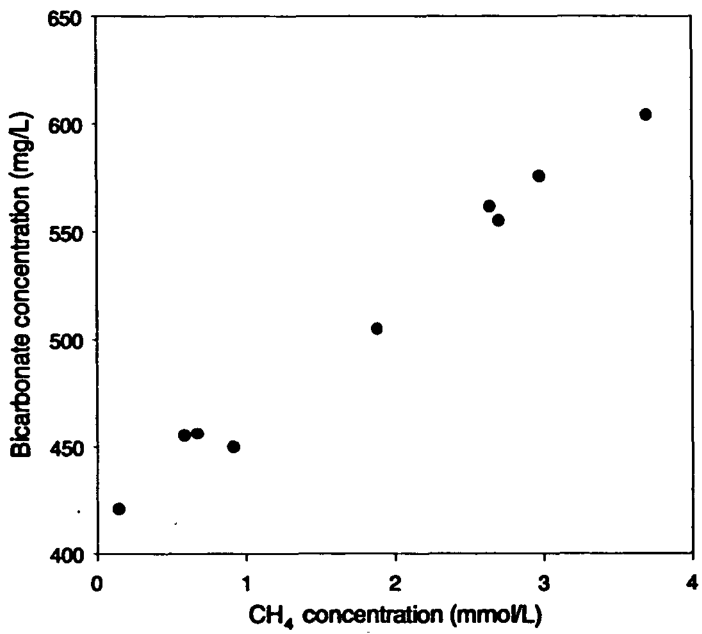


Figure 36 Plot of concentrations of bicarbonate versus methane in groundwater from the confluence area showing a positive correlation (from Panno et at., unpublished).

Groundwater Recharge

Groundwater runoff calculation method Walton (1965) used U.S. Geological Survey streamflow data for Illinois and flow duration studies reported by Mitchell (1957) to determine annual groundwater runoff in the state. He concluded that groundwater runoff in Illinois could be rapidly estimated without excessive error with standard period flow duration curves and an empirical relationship he observed between flow duration curves and the latitude of the basins. By using the flow duration curves and the empirical relationship, Walton estimated groundwater runoff for 109 basins in Illinois for years of near-, below-, and above-normal precipitation. The average ratio between recharge to cones of depression in deeply buried aquifers and groundwater runoff was found to be about 60%.

Walton's (1965) runoff estimates, when applied to the current study area, indicate that two runoff categories exist for years of normal pre-

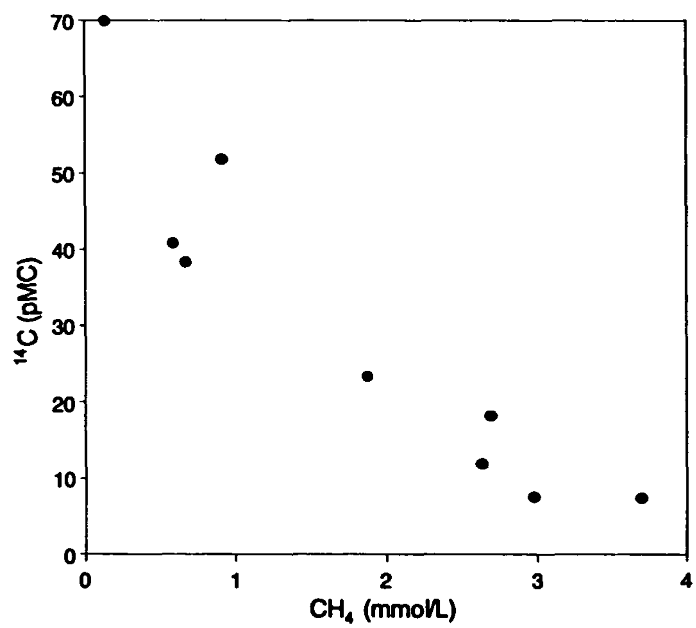


Figure 37 Plot of percent modem carbon versus methane concentrations in groundwater from the confluence area showing a negative correlation (from Panno et ai., unpublished).

cipitation: one with runoff between 129,000 and 194,000 gpd/mi.^{2} and the other with runoff between 194,000 and 259,000 gpd/mi.^{2}. If an average value of 194,000 gpd/mi.^{2} is assumed, groundwater recharge in the study area is estimated to be approximately 52 mgd. According to Walton, 60% (about 30 mgd) of that might be available for use to pumping centers.

Mass balance calculations The lines of equal chloride concentration for the Sankoty-Mahomet Sand aquifer within the study area (fig. 34) can be used to calculate recharge from land surface. The approximate volume of groundwater entering the study area from the Mahomet Sand aquifer to the east was calculated in the following manner. A cross section through the main channel of the Mahomet Sand aquifer (fig. 11 in Kempton and Visocky 1992) was used to calculate a cross sectional area of 4.8×10^6 ft^2 for the Mahomet Sand aquifer on the east side of the study area. A hydraulic gradient of 0.0002 (Kempton et al. 1991) and an average hydraulic conductivity of 1,940 gpd/ft^2 (Kempton and Visocky 1992, Wilson et al. 1994) were used in Darcy's law:

$$Q = KiA$$

where *K* is hydraulic conductivity, *i* is the hydraulic gradient, and *A* is area of the cross section. The quantity of groundwater flow (*Q*) from the Mahomet Sand aquifer into the study area was calculated to be approximately 2.6 mgd.

The area of the mixing zone bounded by the 70 mg/L and 5 mg/L concentration contours in figure 34 (approximately 264 square miles) was measured in order to determine the volume of surface recharge occurring in the study area. Because the chloride concentration of the Sankoty-Mahomet Sand aquifer is diluted in this area, as compared with concentrations in groundwater to the east, the volume of recharge can be calculated from the amount of dilution. An average Cl^- concentration of 2.45 mg/L from the upper Banner and Glasford sand and gravel aquifers was used in the mass balance equation to calculate the volume of water necessary to dilute the water in the Sankoty-Mahomet Sand aquifer from a Cl^- concentration of 70 mg/L to 5 mg/L. Recharge was calculated using the mass balance equation:

$$2.5 \text{ mg/L } Q_s + 70 \text{ mg/L } Q_{\text{Mahomet Sand aquifer}} = 5 \text{ mg/L}(Q_s + Q_{\text{Sankoty–Mahomet Sand aquifer}})$$

where Q_s is recharge from the surface and $Q_{\text{Mahomet Sand aquifer}}$ is discharge from the Mahomet Sand aquifer into the study area. The volume of water necessary to dilute the chloride-enriched ground-water from the east to the concentrations found in the study area is 213,000 gpd/mi^2. This volume is about 10% more than the volume determined using groundwater runoff techniques. Given the high level of uncertainty in both techniques for estimating recharge, this difference is considered insignificant.

Groundwater Quality

Groundwater quality in the study area was recently characterized by Holm (1995). The chemical data are presented in Appendix I. Tables 10 and 11 present summary statistics for several chemical parameters from historical data from samples collected in the study area. The data for tables 10 and 11 come from the ISWS Water Quality Database. Most of the data in Appendix I fall in the ranges presented in tables 10 and 11. The ammonia nitrogen concentrations in wells SWS-C and MTH-15 were greater than the highest value presented in table 10. Both wells are in the vicinity of Stanford. The potassium (K) concentration in well SWS-A was greater than the highest potassium value in table 10. The arsenic (As) concentrations in wells SWS-3s and MTH-24s were greater than the highest value listed in table 11. The As concentrations in wells MTH-18d and MTH-24d were greater than the highest value in table 10.

The major cations were calcium, magnesium, and sodium. Ammonium was a significant cation in samples from wells SWS-C and MTH-15. The main anion was bicarbonate, a component of most of the alkalinity (Stumm and Morgan 1981).

Panno et al. (1994) compiled chemical data for the Mahomet Sand aquifer and plotted contours of concentrations of several ions. They found that bicarbonate concentrations increased from less than 400 mg/L as $CaCO_3$ in the northwest comer of the study area to greater than 500 mg/L in the southeast comer. The alkalinities presented in Appendix I are consistent with this trend. For example, the values in wells MTH-1 and MTH-6 (northwest) were 342.5 and 331.8 mg/L. while the values in wells MTH-21 and SWS-A (southeast) were 485.3 and 509.3 mg/L.

Table 10 Summary of the general groundwater chemistry of the Sankoty-Mahomet Sand aquifer In and near the study area.

Constituent	Number of samples	Minimum	Maximum	Mean	Median	Standard deviation
Calcium	39	15.9	193	76.3	72.0	24.74
Magnesium	38	25.5	67.8	35.3	33.8	8.39
Sodium	36	0.0	160	55.5	27.0	50.23
Potassium	25	1.2	3.3	2.0	2.1	0.65
Chloride	61	1.0	100	25.2	8.8	27.93
Sulfate	34	0.0	122	16.1	2.5	26.85
Arsenic*	25	0.0	51.0	16.0	14.0	12.80
Iron	62	0.073	12.5	2.95	2.35	2.359
Lead*	19	0.0	40.0	8.5	6.0	10.55
Manganese*	44	0.0	235	53.4	40.0	57.9
pH	31	6.8	8.1	7.3	7.40	0.3
Alkalinity	64	300	644	416	426.0	78.3
Ammonia	20	0.0	9.2	3.3	2.2	2.96
Hardness	58	197	761	333	322.0	77.64
TDS	62	312	789	469	465.0	112

Constituent concentrations are in mg/L; *constituents are measured in μg/L; hardness and alkalinity are measured as $CaCO^3$. Results of 0.0 indicate that concentration was below method detection limit.

Table 11 Summary of the general groundwater chemistry of the Glasford sand and gravel aquifers in and near the study area.

Constituent	Number of samples	Minimum	Maximum	Mean	Median	Standard deviation
Calcium	29	35.0	202	98.03	92.0	36.59
Magnesium	30	25.0	224	53.83	42.1	36.67
Sodium	30	0.0	254	46.77	32.0	50.65
Potassium	13	1.0	2.2	1.7	1.6	0.38
Chloride	86	1.0	452	28.88	5.0	63.03
Sulfate	33	0.0	250	57.12	24.0	70.38
Arsenic*	12	0.0	20.0	5.17	4.0	6.59
Iron	86	0.0	28.5	3.00	2.50	4.19
Lead*	12	0.0	30.0	7.67	6.0	8.30
Manganese*	35	0.0	800	56.23	28.0	136.61
PH	20	6.4	8.0	7.21	7.20	0.45
Alkalinity	86	154	652	396.32	396.0	95.35
Ammonia	9	0.13	5.0	1.91	1.30	1.79
Hardness	86	216	1056	423.26	360.0	165.27
TDS	86	263	1556	534.56	476.0	227.46

Constituent concentrations are in mg/L; constituents are measured in μg/L; hardness and alkalinity are measured as $CaCO^3$. Results of 0.0 indicate that concentration was below method detection limit.

The sodium and chloride plots of Panno et al. (1994) showed steep concentration gradients at the east edge of the present study area. Sodium concentrations increased from less than 50 mg/L to greater than 110 mg/L over a distance of less than 6 miles. The sodium and chloride concentrations in Appendix I are mostly consistent with these trends. For example, for the transect of wells MTH-18, SWS-3, and MTH-19, the sodium concentrations were 10.2, 59.5, and 94.6 mg/L. However, well MTH-21, to the east of well MTH-19, had only 45.5 mg/L.

The concentrations of minor ions indicated sub-oxic (low dissolved oxygen, detectable iron, undetectable sulfide) to reducing (detectable sulfide) conditions in the aquifer system: Microorganisms use oxidants in the following decreasing order oxygen, manganese oxides, nitrate, iron oxides, and sulfate. Iron concentrations were greater than 1 mg/L in most samples. Manganese concentrations in many samples were greater than 0.2 mg/L. Dissolved oxygen readily oxidizes soluble ferrous iron to the relatively insoluble ferric form, so high dissolved iron concentrations indicate very low oxygen concentrations. The soluble manganous ion is not as rapidly oxidized to the insoluble MnC>2 as is ferrous iron to ferric, but detectable manganese concentrations usually confirm low dissolved oxygen concentrations. Nitrate was only detected in samples from wells MTH-1 and MTH-15, while ammonia was detected in all samples but MTH-1. Therefore, the microorganisms in the aquifer apparently depleted the dissolved oxygen, nitrate, and iron and manganese oxides. Although sulfide was not determined, the odor of hydrogen sulfide was detected in several groundwater samples during other determinations. Sulfate was undetectable in most of the samples with hydrogen sulfide odor. Therefore, in at least some parts of the aquifer, sulfate reduction was occurring.

The arsenic concentrations in wells MTH-24d and MTH-18d exceeded the maximum contaminant level of 0.05 mg/L. However, arsenic concentrations in surrounding wells were much lower, so the high arsenic concentrations were localized. Arsenic concentrations ranging from 0.01 to 0.03 mg/L were common in the study area. However, arsenic was undetectable (.003 mg/L) in many wells.

Copper, lead, nickel, cobalt, and selenium were detected in a few samples. (Detection limits are given in Appendix I.) The following elements, with detection limits given in parentheses, were undetectable in all samples: cadmium (0.003 mg/L), mercury (0.02 mg/L), silver (0.002 mg/L), molybdenum (0.012 mg/L), antimony (0.18 mg/L), tin (0.04 mg/L), thallium (0.12 mg/L), beryllium (0.001 mg/L), and vanadium (0.006 mg/L).

Aluminum and titanium were detected in some samples. In the pH range of the samples, concentrations of aluminum are expected to be from 0.002 to 0.02 mg/L (Weslowski 1992) and concentrations of titanium are expected to be less than 0.001 mg/L (Ziemniak et al. 1993). The reason for the anomalous aluminum and titanium concentrations may have been inadequate filtration or dust contamination during sampling. Arsenic concentrations appeared to be unrelated to either aluminum or titanium, so particulate contamination did not affect arsenic concentrations. However, three of the samples with detectable lead also had high aluminum, so the lead concentrations are also suspect.

The herbicides atrazine, alachlor, and simazine were undetectable in all samples. Similarly, the following organic compounds, which include many industrial solvents, were not detected in all samples: dichlorodifluoromethane; 1,1,2-trichloroethane; n-butyl benzene; chloromethane; 1,3-dichloropropane; 1,2-dichlorobenzene; vinyl chloride; tetrachloroethylene; 1,2-dibromo-3-chloropropane; bromomethane; dibromochloromethane; 1,2,4-trichlorobenzene; chloroethane; 1,2-dibromoethane; hexachlorobutadiene; trichlorofluoromethane; chlorobenzene; naphthalene; 1,1-dichloroethene; 1,1,1,2-tetrachloroethane; 1,2,3-trichlorobenzene; dichloromethane; ethyl benzene; t-1,3-dichloropropene; t-1,2-dichloroethene; m + p xylene; 1,4-dichlorobenzene; 1,1-dichloroethane; o-xylene; 2,2-dichloropropane; styrene; c-1,2-dichloroethene; isopropyl benzene; chloroform; bromoform; bromochloromethane; 1,1,2,2-tetrachloroethane; 1,1,1-trichloroethane; 1,2,3-trichloropropane; 1,1-dichloropropene; n-propyl benzene; carbon tetrachloride; bromobenzene; 1,2-dichloroethane; 1,3,5-trimethylbenzene; benzene; 2-chlorotoluene; trichloroethene; 4-chlorotoluene; 1,2-dichloropropane; t-butyl benzene; bromodichloromethane; 1,2,4-trimethylbenzene; dibromomethane; sec-butyl benzene; c-1,3-dichloropropene; para-isopropyltoluene; toluene; and 1,3-dichlorobenzene.

Groundwater Use

The total groundwater withdrawal in and near the study area in 1993 was approximately 5 mgd. Table 12 summarizes the groundwater usage for the study area. Appendix J presents a detailed listing of these data. Eleven municipalities located within or near the study area rely upon ground-

Table 12 Summary of total water withdrawals (mgd) by category and year within the study area.

	1988	1989	1990	1991	1992	1993
Municipal	3.472	3.084	3.168	3.098	2.984	2.812
Rural-Residential	1.146	1.436	1.450	1.423	1.418	1.435
Industrial	0.000	0.000	0.000	0.000	0.000	0.001
Livestock	0.512	0.496	0.500	0.499	0.467	0.463
Irrigation	0.200	0.200	0.200	0.200	0.200	0.200
Total	5.330	5.216	5.321	5.220	5.096	4.911

water as their source of water. Persons residing within the limits of incorporated municipalities account for 41% of the total population in the study area. Municipal water use also includes water from the municipal supply that is consumed by businesses. The Town of Normal is the largest single user of groundwater within the study area, pumping approximately 1.8 mgd in 1993. The towns of Delavan, Mackinaw, and Atlanta, the next three largest users of groundwater, pumped 0.18, 0.16, and 0.14 mgd, respectively. In 1993, total groundwater withdrawal by the eleven municipalities was approximately 2.8 mgd.

Rural-residential refers to consumption by that sector of the study area's population living outside of an incorporated municipality. This group includes both farmsteads with private well(s) used for household water supplies and those subdivisions that have their own private well(s) used to supply residents with water. A rural population was first calculated in order to arrive at an approximation of the total withdrawal attributed to this category of consumers. This estimate was done by subtracting the number of residents living in all cities located within a given township from that township's total population (U.S. Department of Commerce 1991). A percent rural population was then calculated. The rural population accounts for 59% of the total population of the study area. The resulting population value was then multiplied by a per capita water use figure that was calculated using a method published by Kirk (1987). The calculated per capita water use is approximately 98.25 gpd (gallons per day). Total groundwater withdrawal in 1993 for rural residents is estimated to be 1.5 mgd. Per capita use is greater for municipalities than for rural-residential use because the former includes some business consumption.

The industrial category includes only those industries within the study area that are self-supplying with respect to the water they use for manufacturing and/or processing. A search of the IWIP database indicated that there is one industry meeting this criterion. Groundwater withdrawal for this industry was approximately 520 gallons per day in 1993.

For purposes of this study, livestock includes all cattle, hogs, sheep, chickens, and turkeys. County livestock population information was taken from the Illinois Agricultural Statistics Annual Summaries (Illinois Agricultural Statistics Service 1988-1994). Total livestock usage for 1993 was 0.5 mgd, a figure derived from a calculation of the total livestock population in the study area multiplied by the daily water requirements for each particular type of animal (Kirk 1987).

Approximately 240 acres are irrigated via a center pivot irrigation system within the study area. The total groundwater withdrawal by this system is estimated to be 0.2 mgd.

Revision of Conceptual Model

The new data gathered for this study add significantly to the conceptual flow model outlined by Wilson et al. (1994). Most notable are data on the influences that the Mackinaw River and sand and gravel aquifers within the upper Banner and Glasford Formations have on the Sankoty-Mahomet Sand aquifer. Wilson et al. concluded that the shallow aquifers-were connected to the Sankoty-Mahomet Sand aquifer on the west side of their study area in R4W. It is now clear that the connection begins further east, especially in the Mackinaw River Valley. Information from logs is inadequate to identify a specific location or "boundary" where the connection begins.

The response of the Sankoty-Mahomet Sand aquifer to the excessive rainfall in 1993 and to changes in river stage indicate connection with the surface and with near-surface aquifers. This result confirms the conclusion that recharge to the Sankoty-Mahomet Sand aquifer is significant.

Using Walton's runoff method, recharge in the study area was estimated to be 52 mgd. Groundwater chemistry data indicated a recharge rate of 56 mgd.

The water level measurements suggest that recharge may increase westward through the study area. The water levels in the western observation wells have more significant responses to surface influences. Also, and the head difference between the two aquifers decreases westward through the study area, indicating better hydraulic connection (figs. 29,30)

CONCLUSIONS AND RECOMMENDATIONS

Conclusions

Given the amount of recharge available to the Sankoty-Mahomet Sand aquifer and its size and hydraulic properties, the prospects for additional development of high capacity wells in the study area appear favorable. Location of such developments will be governed by the thickness and boundaries of the aquifer.

Until recently, the confluence area of the Mahomet and Mackinaw Bedrock Valleys was interpreted to be a wide valley filled with a 100-foot-thick, sand and gravel aquifer (the Sankoty-Mahomet Sand aquifer) overlain by finer grained deposits with interspersed aquifers of the upper Banner and Glasford Formations. Seismic data from this study, however, have provided evidence of hills and depressions in the bedrock surface that are greater both in size and number than previously mapped. The most significant of the bedrock hills, located south of Hopedale, pinches off the aquifer.

The geometry and geology of the Sankoty-Mahomet Sand aquifer is more complex than previously thought. It includes two sand deposits, including the previously undifferentiated sub-Sankoty-Mahomet sand, and ranges in thickness from 0 to more than 150 feet. At several locations in the south-central and eastern part of the study area, thick, lacustrine deposits replace the aquifer material. The bedrock hills and the lacustrine deposits effectively divide the Sankoty-Mahomet Sand aquifer into four regions, each containing aquifer material in excess of 100 feet thick. In three of these areas, the aquifer is greater than 150 feet thick. In the fourth area, along the west edge of the study area, the Sankoty-Mahomet Sand aquifer is directly overlain by shallower sand and gravel aquifers, for a combined thickness of up to 200 feet in localized areas. These variations in aquifer thickness will affect the potential of various sites for the development of a potential high capacity well field.

The Sankoty-Mahomet Sand aquifer generally exhibits leaky artesian conditions and has high transmissivity values. Results from the two aquifer tests confirm the significance of the new geological data and their interpretation. In the southwest corner of the study area (TST1), the test on the Sankoty-Mahomet Sand aquifer suggested the presence of a significant aquifer. However, the analysis also revealed that the boundary created by the bedrock hill south of Hopedale would increase drawdown in high capacity wells in the southwest part of the study area, thereby increasing the volume of the cone of depression. Near Mackinaw (TST2), the aquifer also appeared to be capable of producing substantial quantities of water. At this site, the upper Glasford sand and gravel aquifer serves as the source bed for leakage into the Sankoty-Mahomet Sand aquifer.

Data on water levels verified the previous hydrologic interpretation presented by Wilson et al. (1994). A groundwater divide trends southeast-northwest across the area. Northeast of the divide, groundwater in the Sankoty-Mahomet Sand aquifer flows northward out of the study area and into the Mackinaw Valley. Southwest of the divide, groundwater in the Sankoty-Mahomet Sand aquifer flows westward. Groundwater levels also indicate that the hydraulic connection between the Glasford and the Sankoty-Mahomet Sand aquifers increases toward the west. Changes in stage of the Mackinaw River are mimicked by the water level fluctuations in the Sankoty-Mahomet Sand aquifer over much of the river's reach in the study area.

Recharge to the Sankoty-Mahomet Sand aquifer from the shallower upper Banner and Glasford Formation sand and gravel aquifers and the Mackinaw River is substantial. Local precipitation is the main source of recharge to the groundwater flow system. Calculations of groundwater recharge made on the basis of water chemistry indicate a recharge rate of 213,000 gpd/mi^2. This volume is similar to the average recharge value of 194,000 gpd/mi^2 determined using groundwater runoff techniques. Approximately 60% of recharge to the study area, more than 30 mgd, may be available to wells.

Future Work

The data acquired during this study have dramatically improved the conceptual understanding of the regional groundwater flow in the study area. These data will now be used to assemble a computer flow model to assess the impacts of various pumping scenarios at several locations within the study area. The potential yield of the Sankoty-Mahomet Sand aquifer, the feasibility of a high capacity well field, and the locations where such a well field might be developed will be evaluated in the upcoming modeling phase of the project and will be addressed in the final report.

This investigation also indicates, however, the need for some additional work. Four areas have been identified where the Sankoty-Mahomet Sand aquifer is more than 100 feet thick. Aquifer tests to determine hydraulic properties were performed only in two of these areas because of financial constraints. If the other two areas are to be considered for a municipal groundwater supply, aquifer tests should be conducted in these areas to verify current interpretations. Furthermore, the aquifer system apparently has a much more direct hydraulic-connection to the Mackinaw River than previously thought. Additional measurements to determine recharge and discharge parts of the river will aid in confirming and quantifying the river's role in recharge and/or discharge from the aquifer system. The modeling will assist in determining where further research is needed.

Development of the Sankoty-Mahomet Sand aquifer for irrigation, industry, and municipal use will undoubtedly increase over the next several decades. The best way to document the effects of additional pumpage on the available groundwater resources is to monitor water levels as these developments occur. Monitoring will also be essential for updating the groundwater flow model. Some regular monitoring of the observation well network and river stage, therefore, should be maintained after this study is completed. Such monitoring would have to include maintenance of the network and periodic redevelopment of the wells to assure useful data.

REFERENCES

Anliker, M.A., E.W. Sanderson, and A.P. Visocky, in press, Reconnaissance Study of Ground-Water Levels and Ground-Water Withdrawals in the Vicinity of DeWitt and Piatt Counties: Illinois State Water Survey Contract Report.

Barcelona, M.J., J.P. Gibb, J.A. Helfrich, and E.E. Garske, 1985, Practical Guide fro Ground-Water Sampling. Illinois State Water Survey Contract Report 374,94 p.

Barker J.F., P. Fritz, and R.M. Brown, 1979, Carbon-14 Measurements in Aquifers with Methane; Isotope Hydrology 1978, Proceedings of the International Symposium on Isotope Hydrology - Vol. 2, p. 661-678.

Bergstrom, R.E., K. Piskin, and LR. Follmer, 1976, Geology for Planning in the Springfield-Decatur Region: Illinois State Geological Survey Circular 497,76 p.

Boulton, N.S., 1963, Analysis of data from nonequilibrium pumping test allowing for delayed yield from storage: Proceedings of the Institution of Civil Engineers, v. 26, no. 6693, p. 469-482.

Butler, J.N., 1982, Carbon Dioxide Equilibria and Their Applications: Addison-Wesley, Reading, MA, 547 p.

Clesceri, L.S., A.D. Eaton, and A.E. Greenberg (eds.), 1992, Standard Methods for the Examination of Water and Wastewater, 18th edition: American Public Health Association, American Water Association, and Water Environment Federation.

Coleman, D.D., C.L. Liu, and K.M. Riley, 1988, Microbial Methane in the Shallow Paleozoic Sediments and Glacial Deposits of Illinois, U.S.A.: Chemical Geology, v. 71, p. 23-40.

Cooper, H.H., Jr., and C.E. Jacob, 1946, A generalized graphical method for evaluating formation constants and summarizing well-field history: Transactions of the American Geophysical Union, v. 27, no. 4, p. 526-534.

Driscoll, F.G., 1986, Groundwater and Wells (2nd ed.): Johnson Division, St. Paul, Minnesota, 1089 p.

Farnsworth and Wylie, P.C., and Hanson Engineers, 1989, Report of the Drought Emergency Water Sources and Options to Improve Existing Lake Supplies for the City of Bloomington, Illinois. 74 p.

Farnsworth and Wylie, P.C., 1990, Report on Feasibility of Establishing a Regional Water Supply for Central Illinois: submitted to McLean County Board, City of Bloomington, Town of Normal Joint Steering Committee, 44 p. plus figures.

Farnsworth and Wylie, P.C., 1994, Regional Water Needs Assessment Study: Summary of Findings. Prepared for the Long Range Water Plan Steering Committee.

Ferris, J.G., 1959, Groundwater, *in* CO. Wisler and E.F. Brater, editors, Hydrology: John Wiley and Sons, New York, p. 198-272.

Foster, J.W., and M.B. Buhle, 1951, An Integrated Geophysical and Geological Investigation of Aquifers in Glacial Drift NearChampaign-Urbana, Illinois: Illinois State Geological Survey Report of Investigations 155, 31 p.

Foster, J.W., 1952, Major Aquifers in Glacial Drift Near Matoon, Illinois: Illinois Academy of Science Transactions, v. 44, p. 85-94.

Foster, J.W., 1953, Significance of Pleistocene deposits in the groundwater resources of Illinois: Economic Geology, v. 48, no. 7, p. 568-573.

Garske, E.E., and M.R. Schock, 1986, An inexpensive flow-through cell and measurement system of monitoring selected chemical parameters in ground water: Ground Water Monitoring Review, v. 6, no. 6, p. 79-84.

Graf D.L., I. Friedman, and W.F. Meents, 1965, The Origin of Saline Formation Waters, II: Isotopic Fractionation by Shale Micropore Systems: Illinois State Geological Survey Circular 393,32 p.

Grossman E.L., B.K. Coffman, S.J. Fritz, and H. Wada, 1989, Bacterial production of methane and its influence on the groundwater chemistry in east-central Texas aquifers: Geology, v. 17, no. 6, p. 495-499.

Hackley K.C., C.L. Lui, and D.D. Coleman, 1992, ^{14}C dating of groundwater containing microbial CH4: Radiocarbon, v. 34, p. 686-695.

Hanson, R., 1950, Public Ground-Water Supplies in Illinois: Illinois State Water Survey Bulletin 40, 1036 p.

Hanson, R., 1958, New Public Ground-Water Supplies in Illinois—1950-1957: Illinois State Water Survey Supplement I to Bulletin 40.

Hanson, R., 1961, Additions to Public Ground-Water Supplies in Illinois: Illinois State Water Survey Supplement II to Bulletin 40.

Hantush, M.S., 1956, Analysis of data from pumping tests in leaky aquifers: Transactions of the American Geophysical Union, v. 37, no. 6, p. 702-714.

Hantush, M.S., and C.E. Jacob, 1955, Nonsteady radial flow in an infinite leaky aquifer: Transactions of the American Geophysical Union, v. 36, no. 1, p. 95–100.

Heath, R.C., 1989, Basic Ground-Water Hydrology: U.S. Geological Survey Water-Supply Paper 2220, 84 p.
Hensel, B.R., 1992, Natural Recharge of Groundwater in Illinois: Illinois State Geological Survey Environmental Geology 143,33 p.
Herzog, B.L, B.J. Stiff, C.A. Chenoweth, K.L. Wamer, J.B. Sieverling, and C. Avery, 1994, Buried Bedrock Surface of Illinois: Illinois State Geological Survey, Illinois Map 5, scale 1:500,000.
Holm, T.R., 1995, Ground-Water Quality in the Mahomet Aquifer, McLean, Logan, and Tazewell Counties. Illinois State Water Survey Contract Report 579,42 p.
Horberg, L, 1945, A Major Buried Valley in East-Central Illinois and its Regional Relationships: Illinois State Geological Survey Report of Investigations 106,11 p.
Horberg, L, 1950, Bedrock Topography of Illinois: Illinois State Geological Survey Bulletin 73,111 p.
Horberg, L., 1953, Pleistocene Deposits Below the Wisconsin Drift in Northeastern Illinois: Illinois State Geological Survey Report of Investigations 165, 61 p.
Hunt, C.S., and J.P. Kempton, 1977, Geology for Planning in DeWitt County, Illinois: Illinois State Geological Survey Environmental Notes 83,42 p.
International Atomic Energy Agency, 1983, Isotope Techniques in the Hydrogeological Assessment of Potential Sites for the Disposal of High-Level Radioactive Wastes: Technical Report Series no. 228, International Atomic Energy Agency, Vienna, 151 p.
Keith, L.H., W. Crummett, J. Deegan Jr., R.A., Libby, J.K., Taylor, and G. Wentler, 1983, Principles of environmental analysis: Analytical Chemistry, v. 55, p. 2210-2218.
Kempton, J.P., W.H. Johnson, K. Cartwright, and P.C. Heigold, 1991, Mahomet Bedrock Valley in east-central Illinois: topography, glacial drift stratigraphy, and hydrogeology, *in* W.N. Melhorn and J.P. Kempton, editors, Geology and Hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 91-124.
Kempton, J.P., and A.P. Visocky, 1992, Regional Groundwater Resources in Western McLean and Eastern Tazewell Counties: Illinois State Geological Survey and Illinois State Water Survey Cooperative Groundwater Report 13, 41 p.
Kempton, J.P., and V.L. Poole, 1985, Geology and distribution of the aquifers west of Normal, Illinois—an update: Illinois State Geological Survey unpublished report, 13 p.
Kirk, J.R., 1987, Water Withdrawals in Illinois in 1986: Illinois State Water Survey Circular 167,43 p.
Koeppe, C.E., 1935, Weather and Climate: McKnight and McKnight, Bloomington, Illinois, 135 p.
Lamb, M.K., 1994, Groundwater: Illinois' Buried Treasure—Education Activity Guide: Illinois Department of Energy and Natural Resources, 114 p.
Larson, T.H., and V.L. Poole, 1989, Geophysical Exploration for Potential Groundwater Resources Near Bloomington, Illinois: Illinois State Geological Survey Contract/Grant Report 1989-2,51 p.
Larson, T.H., and P.G. Orozco, 1991, Results of a Shallow Seismic Refraction Survey nearthe Village of North Aurora, Illinois: Illinois State Geological Survey Open File Series 1991-15,18 p.
Leighton, M.M., G.E. Ekblaw, and L Horberg, 1948, Physiographic Divisions of Illinois: Illinois State Geological Survey Report of investigations 129,19 p.
Lineback, J.A., 1979, Quaternary Deposits of Illinois: Illinois State Geological Survey Map, scale 1:500,000.
Lineback, J.A., 1981, Quaternary Deposits of Illinois: Illinois State Geological Survey Map, scale 1:2,500,000.
Mitchell, W.D., 1957, Flow Duration of Illinois Streams: Illinois Department of Public Works, Division of Waterways, Springfield, 189 p.
Munsell Color, 1975, Munsell Soil Color Charts, 1975 Edition: Munsell Color, Baltimore, Maryland.
Neely, R.D., and C.G. Heister (compilers), 1987, The Natural Resources of Illinois: Illinois Natural History Survey Special Publication 6, 224 p.
Neuman, S.P., 1975, Analysis of pumping test data from anisotropic unconfined aquifers considering delayed gravity response: Water Resources Research, v. 11, no. 2, p. 329-342.
Panno, S.V., Hackley, K.C., K. Cartwright, and C.L. Liu., 1994, Hydrogeochemistry of the Mahomet Bedrock Valley Aquifer, east-central Illinois: indicators of recharge and ground-water flow: Groundwater, v. 32, no. 4, p. 591-604.
Peyton, G.R., 1993, The free-radical chemistry of persulfate-based total organic carbon analyzers: Marine Chemistry, v. 41, p. 91–103.
Piskin, K., and R.E. Bergstrom, 1975, Glacial Drift in Illinois: Thickness and Character: Illinois State Geological Survey, Circular 490,35 p.
Prickett, T.A., 1965, Type-curve solution to aquifer tests underwater-table conditions: Ground Water, v. 3, no. 3, p. 5-14.
Richards, S.S., and A.P. Visocky, 1982, A Reassessment of Aquifer Conditions West of Normal, Illinois: Illinois State Water Survey Circular 153,33 p.

Rimrock Geophysics Inc., 1992, User's guide to SIPT2 V-3.2,12 p.
Sanderson, E.W., 1971, Groundwater Availability in Piatt County: Illinois State Water Survey Circular 107, 83 p.
Sanderson, E.W., and E. Zewde., 1976, Ground-Water Availability in Champaign County: Illinois State Water Survey Circular 124,139 p.
Schicht, R.J., 1966, Ground-Water Availability for the City of Normal: Illinois State Water Survey unpublished open-file report prepared for the Statewide Water Plan.
Scott, J.H., B.L. Tibbetts, and R.G. Burdek, 1972, Computer Analysis of Seismic Refraction Data: U.S. Bureau of Mines Report of Investigation 7595,95 p.
Selkregg, L.F., and J.P. Kempton, 1958, Groundwater Geology in East-Central Illinois: A Preliminary Geology Report: Illinois State Geological Survey Circular 248, 36 p.
Sinclair, A.J., 1974, Selection of threshold values in geochemical data using probability graphs: Journal of Geochemical Exploration, v. 3, no. 2, p. 129-149.
Smith, H.F., 1950, Groundwater Resources in Champaign County: Illinois State Water Survey Report of Investigation 6, 44 p.
Stephenson, D.A., 1967, Hydrogeology of Glacial Deposits of the Mahomet Bedrock Valley in East-Central Illinois: Illinois State Geological Survey Circular 409,51 p.
Stumm, W., and J.J. Morgan, 1981, Aquatic Chemistry, 2nd ed.: John Wiley, New York, 780 p.
Theis, C.V, 1935, The Relation Between the Lowering of the Piezometric Surface and the Rate and Duration of Discharge of a Well using Ground-water Storage: Transactions of the American Geophysical Union 16th Annual Meeting, part 2, p. 519-524.
U.S. Department of Commerce, 1991, 1990 Census of Population and Housing, Summary of Population and Housing Characteristics, Illinois: U.S. Department of Commerce, Washington, D.C.
Visocky, A.P., and R.J. Schicht, 1969, Groundwater Resources of the Buried Mahomet Bedrock Valley: Illinois State Water Survey Report of Investigation 62, 52 p.
Visocky. A.P., H.A. Wehrmann, K.L. Kim, and R.W. Ringler, 1978, Assessment of Public Groundwater Supplies in Illinois: Illinois State Water Survey Contract Publication 209,195 p.
Walker, W.H., 1962, Groundwater Availability in the Vicinity of Normal, McLean County, Illinois: Illinois State Water Survey unpublished open-file report.
Walker, W.H., R.E. Bergstrom, and W.C. Walton, 1965, Preliminary Report on the Ground-Water Resources of the Havana Region in West-Central Illinois: Illinois State Water Survey and Illinois State Geological Survey Cooperative Report 3, 61 p.
Walton, W.C, 1962, Selected Analytical Methods for Well and Aquifer Evaluation: Illinois State Water Survey Bulletin 49, 81 p.
Walton, W.C, 1965, Groundwater Recharge and Runoff in Illinois: Illinois State Water Survey Report of Investigation 48, 55 p
Wehrmann. H.A., A.P. Visocky, C.B. Burris, R.W. Ringler, and R.D. Brower, 1980, Assessment of Eighteen Public Groundwater Supplies in Illinois: Illinois State Water Survey Contract Report 237,185 p.
Welz, B., G. Schlemmer, and R. Mudadavi, 1988, Palladium nitrate-magnesium nitrate modifier for graphite furnace atomic absorption spectrometry—Part 2, determination of arsenic, cadmium, copper, manganese, lead, antimony, selenium and thallium in water: Journal of Analytical Atomic Spectrometry, v. 3, p. 695-701.
Weslowski, D.J., 1992, Aluminum sepciation and equilibria in aqueous solution: I. the solubility of gibbsite in the system Na-K-CI-Oh-AI$(OH)_4$ from 0 to 100°C: Geochimica et Cosmochimica Acta, v.56, p. 1065-1091.
Wilson, S.D., J.P. Kempton, and R.B. Lott, 1994, The Sankoty-Mahomet Aquifer in the Confluence Area of the Mackinaw and Mahomet Bedrock Valleys, Central Illinois: A Reassessment of Aquifer Characteristics: Illinois State Water Survey and Illinois State Geological Survey Cooperative Report 16, 64 p.
Woller, D.M., 1975, Public Ground-Water Supplies in Champaign County: Illinois State Water Survey Bulletin 60-15, 55 p.
Ziemniak, S.E., M.E. Jones, and K.E.S. Combs, 1993, Solubility behavior of titanium (IV) oxide in alkaline media at elevated temperatures: Journal of Solution Chemistry, v. 22, p. 601-623.