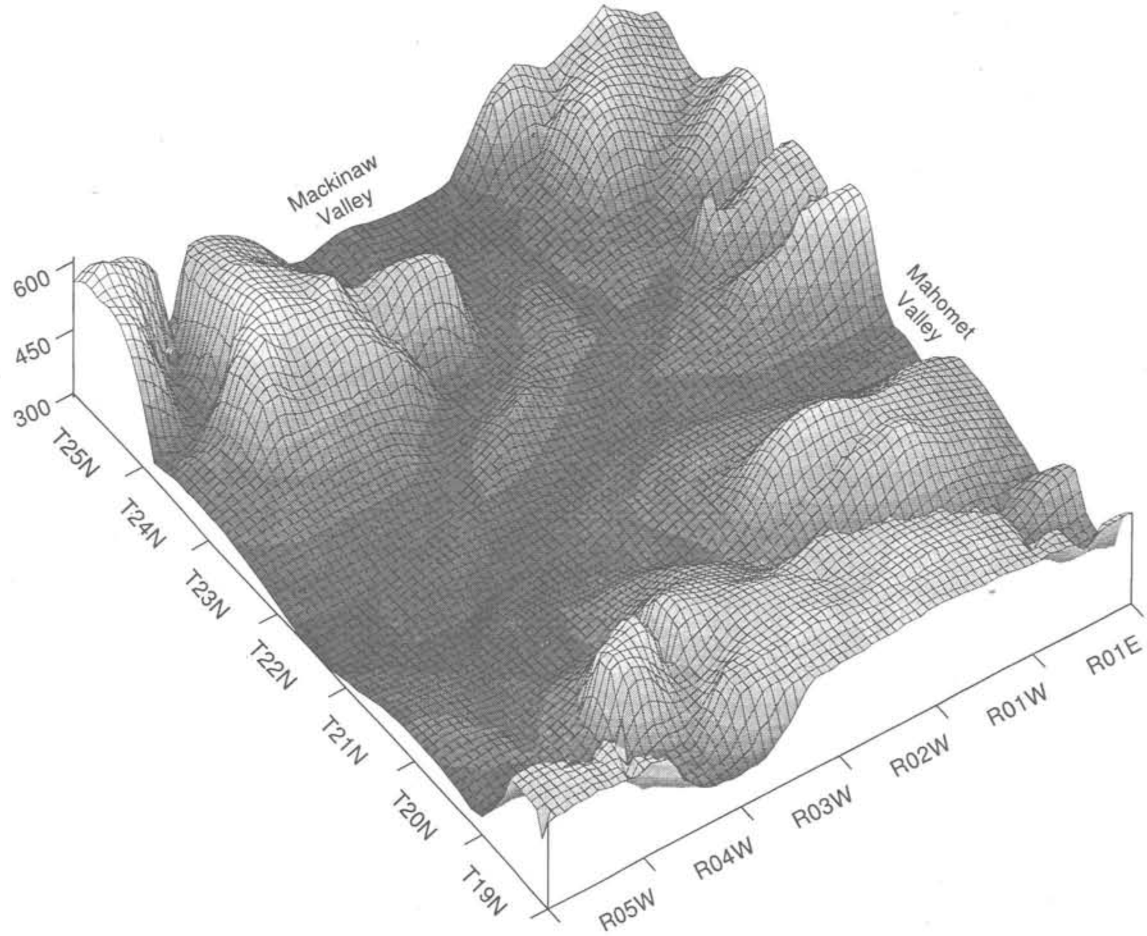
The Sankoty-Mahomet Aquifer in the Confluence Area of the Mackinaw and Mahomet Bedrock Valleys, Central Illinois

A Reassessment of Aquifer Characteristics

Steven D. Wilson Illinois State Water Survey
John P. Kempton Illinois State Geological Survey
R. Brandon Lott Illinois State Water Survey



1994 Cooperative Ground-Water Report 16
Department of Energy and Natural Resources

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Cover illustration depicts bedrock topography of the Sankoty-Mahomet aquifer

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by Steven D. Wilson, John P. Kempton, and R. Brandon Lott

ABSTRACT

A study was undertaken to assess the characteristics of the Sankoty-Mahomet aquifer, one of Illinois' largest buried sand-and-gravel aquifers, which formed in the confluence area of the Mahomet and Mackinaw Bedrock Valleys. A test drilling program was completed in 1992 to provide a better understanding of the aquifer system. Sixteen wells were installed to monitor water-level fluctuations in this lowermost basal sand-and-gravel aquifer composed predominately of Sankoty-Mahomet Sand. The borehole for each monitoring well was drilled into bedrock to assure penetrating the entire thickness of the aquifer. Water-level elevations were measured in more than 80 private wells during spring 1993, and a seven-day aquifer test was conducted on a municipal well at Hopedale, IL.

The drilling results indicate the very complex geologic framework of the system. Data from the latest drilling program indicated differences in bedrock elevations of as much as 75 feet from elevations in earlier bedrock surface maps of the area. Several locations, previously thought to have substantially thick aquifer materials, had thick lacustrine deposits below a relatively thin aquifer. The Sankoty-Mahomet aquifer appears to have been deposited on the often deeply eroded surface of these finer materials. One or more less extensive sand-and-gravel units, interbedded with tills above the aquifer, may be intermittently connected with it.

The conceptual model of the confluence area is notably different from earlier interpretations. The potentiometric surface map generated from measurements of private wells indicates that ground water flows both westward toward the Illinois River and northward into the Mackinaw Valley from the confluence area. Water levels rose throughout 1993 due to locally heavy rainfall. Vertical recharge is significant in the study area, as indicated by the aquifer's quick response to the rainfall as well as other research completed recently. In addition, hydrographs from the observation wells were studied to identify regions with similar water-level fluctuations, and locate at least two independent flow paths in the Mahomet portion of the aquifer: one from the Kenney Valley, the other from the main Mahomet Valley in Dewitt County. The flow from the Kenney Valley may be a significant portion of the horizontal flow entering the aquifer just west of the main part of the confluence area.

These data indicate that the geology and characteristics of the aquifer system are still not well understood. Additional drilling and aquifer tests are essential to developing a more detailed understanding of the aquifer system and its segments.

A separate, ongoing study of a portion of the confluence area is due to be completed in 1996 and should provide more insight about the amount of additional development that the system can maintain. Both the agricultural community and municipal planners view the Sankoty-Mahomet aquifer as the area's most promising water resource for future development.

ACKNOWLEDGEMENTS

This project was conducted as part of the program of research at the Illinois State Water Survey (ISWS). The Illinois State Geological Survey (ISGS) provided input and guidance in selection of drilling sites and in analysis of geologic information. The authors thank the east-central Illinois residents who permitted access to their wells and records. Several individuals went out of their way to help make our field efforts a success. Area residents Owen Romine, Harry Rogers, Robert Cremeens, and Sonny Litwiller were essential to the drilling program. The village of Hopedale and Water Superintendent Cal Willard provided use of the municipal well, as well as assistance in conducting the aquifer test.

Steve Burch (ISWS) served as site geologist for the well drilling program, and he provided considerable guidance in completing the field portion of the study. Sean Sinclair (ISWS) prepared computer-generated maps and figures in this report using ArcINFO, AutoCAD, and CANVAS software. Linda Hascall and Dave Cox prepared hand-drawn maps and figures, and drafted final maps and figures. Dorothy Woller and John Blomberg (ISWS) compiled the data in table 2 and table 5, respectively. Loretta Skowron and Brian Kaiser (ISWS Chemistry Division) reviewed table 5. Patti Hill, Pam Lovett, and Lori Woller typed the tables and text. Several people reviewed the manuscript: Dave Larson and Beverly Herzog (ISGS); Adrian Visocky and Steve Burch (ISWS); and Gary Clark of the Illinois Department of Transportation-Division of Water Resources (IDOT-DWR). Eva Kingston (ISWS) and Ellen Wolf (ISGS) edited the manuscript, and then Eva Kingston typeset the final report.

Several ISWS staff assisted in measuring water levels in the monitoring and private wells, and in performing the pump test: Scott Meyer, Bryan Coulson, Anthony Romanelli, and Jay Sheley. Adrian Visocky (ISWS) analyzed the aquifer test data. Phil Reed (ISGS) reviewed site selection criteria and performed downhole geophysical logging. Albrecht Well Drilling, Ohio, IL, was the contractor for the drilling program. This report also reaped the benefits of a concurrent cooperative mapping study of the 1:100,000 Champaign Quadrangle by the ISGS and the U.S. Geological Survey (Kempton et al., 1993).

Project funding originated through the Illinois Water Inventory and Aquifer Assessment Program administered by the Illinois Department of Energy and Natural Resources. The Illinois Department of Transportation - Division of Water Resources also provided substantial funds. The authors also want to thank Gary Clark, IDOT-DWR, for recognizing the importance of the study by supporting the drilling program because the unexpected findings from the drilling comprise the majority of this report. IDOT-DWR also provided a surveying crew to determine the elevations of the monitoring wells.

INTRODUCTION

Description and Background of the Study Area

This study encompasses an area of 720 square miles in Tazewell, McLean, and Logan Counties in Illinois (figure 1). Land use in the area is mainly agricultural (95.6 percent), with forest (2.0 percent), urban (2.2 percent), and gravel pits and lakes (0.2 percent) (USGS, 1983).

The population of the study area is 30.9 percent rural. In 1990, 12,705 of the 41,092 residents lived outside incorporated areas. Lincoln, the largest community in the study area, accounts for 15,418 residents, or 37.5 percent of the total population (U.S. Dept. of Commerce, 1991). The other incorporated municipalities are: Atlanta, Emden, Hartsburg, New Holland, and San Jose (Logan County); Danvers, McLean, and Stanford (McLean County); and Armington, Delavan, Hopedale, Mackinaw, Minier, Pekin, and Tremont (Tazewell County).

The surface topography reflects the landforms remaining after the last continental glaciers covering the area had melted, and also after subsequent weathering and erosion. These features include glacial end moraines forming relatively high ridges generally from northwest to southeast, and modern stream valleys, the lowest features of the landscape, from northeast to southwest.

Prior to and during glaciation of the area, large rivers and their tributaries carved deep valleys into the bedrock, forming two major bedrock valley systems in central Illinois: the Mahomet system from the east, and the Ancient Mississippi system from the north (figure 2). The two systems converged in Tazewell, McLean, and Logan Counties. Deposits left by glacial advances and retreats filled these channels with unlithified sands, silts, and clays (water and ice-laid deposits). Extensive sand-and-gravel deposits at the convergence of these channels are separated by layers of glacial till and/or lacustrine sediments. These sand-and-gravel deposits represent important aquifers or saturated bodies of earth materials that will yield sufficient ground water to wells. Previous studies (Kempton and Visocky, 1992) indicate that the Sankoty-Mahomet aquifer's potential yield significantly exceeds its current use. Consequently, irrigators and city planners frequently mention the area as a potential source for future ground-water development.

Ground-water issues in the confluence area of the Mackinaw and Mahomet Bedrock Valleys in central Illinois have become the focus of attention for several reasons. The need for additional water was recognized in the 1960s by the town of Normal, and more recently by the city of Bloomington. The town of Normal installed a well field in the Sankoty-Mahomet aquifer in the late 1970s. Because population growth and the drought in 1988 threatened Bloomington's surface water sources, there has been a heightened interest in developing a separate long-term regional water supply to provide potable water to most municipalities in and around the study area. There is also a growing agricultural interest in use of ground water for irrigation purposes.

Several municipalities and water authorities, headed by Bloomington, Normal, and McLean County, formed the Long Range Water Plan Steering Committee (LRWPSC) after the drought of 1988 to assure the responsible development of adequate water supplies for the future. The committee is examining three possibilities for future water-supply expansion: 1) a well field in the confluence area of the major bedrock valleys, 2) a pipeline from the Illinois River, and 3) a reservoir on the Mackinaw River. A feasibility study was undertaken to determine the most cost-effective choice to serve 15 to 20 communities in eastern Tazewell and McLean Counties, and concluded that a ground-water supply would cost considerably less than the other two alternatives. The committee has funded a research investigation, conducted by the ISWS and ISGS, to determine the feasibility of withdrawing 10 to 15 million gallons a day (mgd) from the aquifers in western McLean County, eastern Tazewell County, or both.

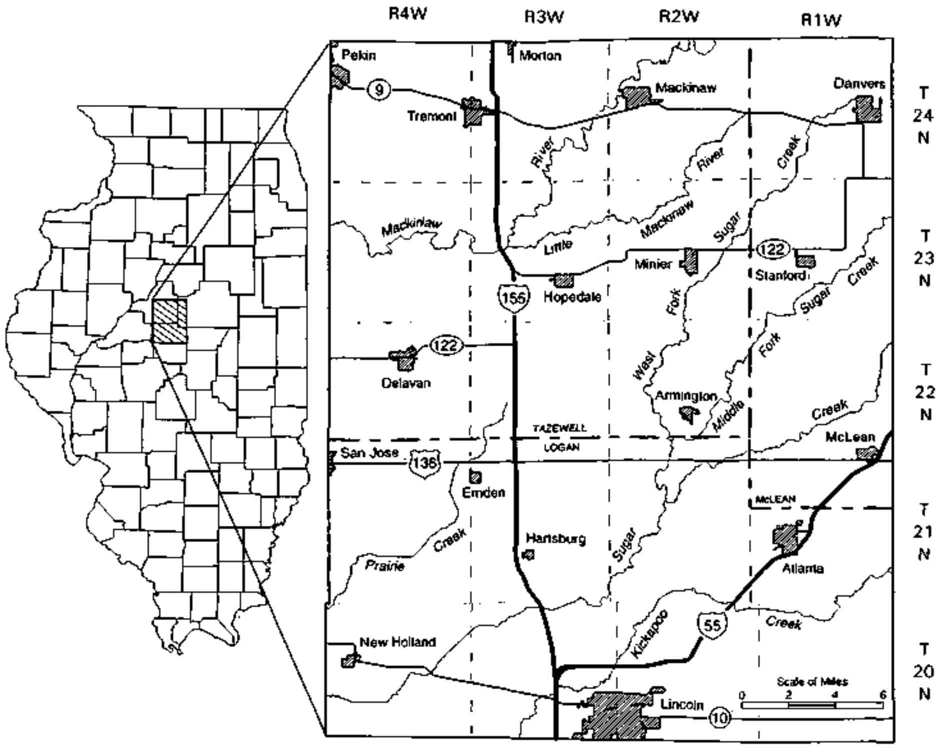

Figure 1. Location of study area

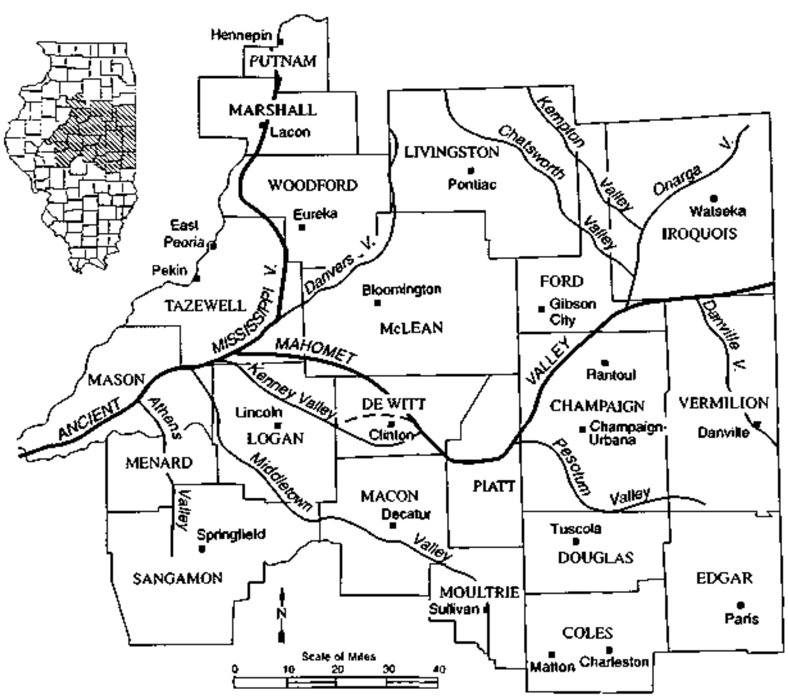

Figure 2. Thalweg of principal bedrock valleys and important tributaries in east-central Illinois (modified from Burch and Kelly, 1993)

Ground-Water Use

Total ground-water use in the study area includes municipal, private, livestock, and irrigation pumpage. Ground-water use from the Sankoty-Mahomet aquifer system is estimated to be 9.2 mgd for these four groups.

In the study area, 15 municipalities use ground water and withdrew approximately 12 mgd in 1992. Wells in several of these communities are finished in aquifers other than the Sankoty-Mahomet aquifer, and their pumpage accounts for 4 of the 12 mgd. Lincoln, Normal, and Mackinaw were the largest municipal users. Based on 1990 township populations, rural, private water use is estimated to be 0.7 mgd.

The study area includes only about 240 acres under irrigation, but irrigation is expanding rapidly near the study area and throughout Illinois. From 1977 to 1987, Illinois' irrigation increased from 110,000 to 200,000 acres (Bowman and Collins, 1987). Nearly 1,000 acres were irrigated between 1989 and 1992 within the region just two miles west of the study area (Rockford Map Publishers, 1993), and additional development in the study area is likely. Livestock are widespread in the study area, but do not account for much of the ground-water pumpage. Withdrawal for irrigation and livestock combined was estimated to be 0.5 mgd in the study area for 1992.

Purpose and Scope

The current study includes new field data collected from 1991-1994 but builds on an earlier study that resulted in an ISGS and ISWS cooperative report (Kempton and Visocky, 1992). That report summarized all existing geologic and hydraulic information in the most complete report on the geology and hydrology of the confluence area to date.

This report describes the study area and its geologic history; summarizes current research about the area; details field work completed during the study; presents updated maps of bedrock topography, aquifer thickness, and aquifer hydrology; summarizes study findings; and discusses implications of further development on this ground-water resource.

Previous Reports

There have been many papers, reports, and research studies about the Mahomet and Mackinaw Valleys. Horberg (1945, 1950) was one of the first to map the general shape of the two bedrock valleys in this region. His benchmark bedrock topography map led to additional research in which he defined the Mahomet Sand in the Mahomet Valley and the Sankoty Sand in the Mackinaw Valley (Horberg, 1953). The ISWS and ISGS later released a cooperative report of their investigation of the Havana Lowlands region, which included the western townships in the present study area (Walker et al., 1965). That report described ground-water flow characteristics of the region and compiled all data about aquifer properties to date, including the results of several aquifer tests from the current study area.

Between the late 1960s and 1985, the town of Normal supported several studies to investigate additional ground-water resources, which resulted in the development of the Normal West Well Field (Richards and Visocky, 1982). Before the well field was installed, a study was conducted to determine the aquifer properties for the region. Using the results of two 24-hour aquifer tests, researchers concluded that long-term interference effects from the planned well field would be minimal. When the well field became operational, however, the actual drawdowns were much greater than anticipated, causing interference with nearby private wells. The town of Normal eventually replaced or deepened several private wells that had been affected by the well field pumpage. Richards and Visocky (1982) subsequently re-evaluated the data and determined that the properties calculated from the tests did not represent regional values and that the tests should have been of longer duration. A direct connection between the shallow and deeper aquifers was found to occur locally. The revised conceptual model predicted much larger impacts due to pumpage from the well field and lowered the initial estimates of the safe yield at the site.

Two recent studies have been completed about the confluence area and the Mahomet Valley. One study updated the geology and hydrology of the confluence area based on well records added over the last 10-15 years (Kempton and Visocky, 1992). The second study, a paper by Kempton et al. (1991), was part of a collection of symposium papers about the history and development of the entire Mahomet-Teays aquifer system. Both studies drew several important conclusions about the confluence area: 1) there is no evidence to support Horberg's conclusion of a bedrock channel below 300 feet mean sea level (ft-msl); 2) there are few, if any, shallow Glasford Formation aquifers greater than 20 feet thick in the confluence area; 3) the Sankoty-Mahomet Sand is very complex and its distribution is not as extensive as was previously believed; and 4) based on estimates of ground-water discharge to streams, an estimated 75 mgd of water might be developed from the aquifers of the confluence area, mainly the Sankoty-Mahomet Sand.

REGIONAL GEOLOGIC SETTING

The bedrock formations in central Illinois consist of a succession of sedimentary rocks several thousand feet thick, including sandstone, limestone, dolomite, shale, and coal. These rocks were warped and tilted over many millions of years to form the Illinois Basin centered in southeastern Illinois and the north-south La Salle Anticlinal Belt east of the study area (Kempton and Visocky, 1992).

The older, generally deeper rocks of central Illinois are mainly limestone, dolomite, and sandstone, which frequently yield water. The younger rocks at or within a few hundred feet of the bedrock surface are composed largely of shale, which does not yield water, interbedded with a few, relatively thin layers of sandstone, limestone, and coal. Water is generally highly mineralized both in the younger and older rocks below depths of 200 to 400 feet. Consequently, ground-water resources are extremely limited in the shallow bedrock and are normally available only in small quantities when permeable sandstone or fractured limestone has been found at or near the top of the bedrock.

During a long period of uplift, the bedrock surface developed an erosional topography that included major river valleys and their drainage system of numerous large and small tributary valleys (figure 2). The principal bedrock valleys in Illinois are the Ancient Mississippi Valley and its major tributary, the Mahomet Valley. The confluence of the Mahomet with the Mackinaw segment of the Ancient Mississippi lies in the southeastern portion of the study area.

The onset of continental glaciation about two million years ago began the process of stream or drainage diversion, then deepening, and ultimately burying the bedrock valleys. Repeated pulses of debris-carrying ice invaded much of Illinois, and older glaciers of pre-Illinoian and Illinoian age covered larger areas than the younger glaciers of Wisconsinan age. Each glacier left layers of debris as it melted, eventually accumulating from 400 to 600 feet of glacial drift on the bedrock surface. Glacial deposits can be classified and mapped from information collected at the surface (exposures) and from the subsurface (well logs and samples). The pre-Illinoian erosion from surges of meltwater significantly modified the preglacial bedrock surface: glaciers initially deepened the bedrock valleys and later filled them with coarse sand and gravel, which blocked and diverted several major channels away from the glacial margins. This complex series of events for at least 1.5 million years, has modified the depth and shape of the Mahomet Bedrock Valley and the Mackinaw segment of the Ancient Mississippi Valley (figure 2). The bedrock valley floor is now covered with up to 150 feet of sand and gravel, which forms the most important aquifer of the area, the Sankoty-Mahomet aquifer.

METHODS OF INVESTIGATION

Data Compilation and Sources of Data

The ISWS and ISGS maintain records of well logs, foundation or engineering borings, oil and gas logs, water quality samples, water use, and aquifer test results. These records along with pertinent published reports were used to provide information about the study area and to develop a project database for the Sankoty-Mahomet aquifer. A plot of key database components revealed large areas in which little was known about the sand thickness and bedrock configuration, so these areas were targeted as drilling locations. As the well verification progressed, the database was modified to reflect changes in existing log locations or other relevant data, to add wells not previously recorded, and to note wells no longer existing.

Field Methods and Procedures

Field efforts for this project included test drilling, monitoring well installation, geophysical logging, inventorying existing wells, measuring water levels, and aquifer testing. The work was conducted during 1992,1993, and early 1994.

The test drilling program consisted of 15 test holes designated as Sites 1-8, 10, A-E, and G (figure 3). Installation of monitoring wells began in June 1992. Site choices were based on a sensitivity plot of compiled data that pointed out locations where little or no information was available about the bedrock elevation, aquifer thickness, or both. A test hole was drilled through the drift and 4 to 8 feet into bedrock at each site, drill cuttings were collected at 10-foot intervals, and then a monitoring well was installed. Because an aquifer of sufficient thickness was not found at Site 4 or Site D, a well was not installed at either site. Consequently, these test holes were geophysically logged, properly plugged, and then abandoned. A monitoring well was installed in the Sankoty-Mahomet Sand at 13 sites (figure 3), and water levels were monitored regularly.

Monitoring wells were constructed of 2-inch diameter PVC pipe with either 5 or 10 feet of 0.020-inch (20 slot) screen. The PVC pipe extended below the screen into the bedrock to facilitate conducting the geophysical logging through the casing at a later date. When a significant upper aquifer was found at three of the 13 sites (Site 2, Site 3, and Site G), a second well was installed to monitor evidence of any hydraulic connection between the two aquifers at these sites.

The ISGS conducted the downhole geophysical logging. Each deep monitoring well was logged using a natural gamma radiation probe through the PVC pipe, except at Site 4 and Site D where logging was completed in the open hole.

A door-to-door search for private wells was also conducted to secure 80 to 100 private wells for periodic measurement. Once the depth and location of these wells were verified, an initial water-level measurement was taken. Then in May 1993, a "mass measurement" was conducted over a two-day period using both the private wells and the monitoring wells (figure 4).

An aquifer test was conducted at Hopedale using a municipal well, which was pumped at a continuous rate of 240 gpm for seven days. Two observation wells were monitored during the test. The time-drawdown data were analyzed to determine the aquifer properties under leaky artesian conditions.

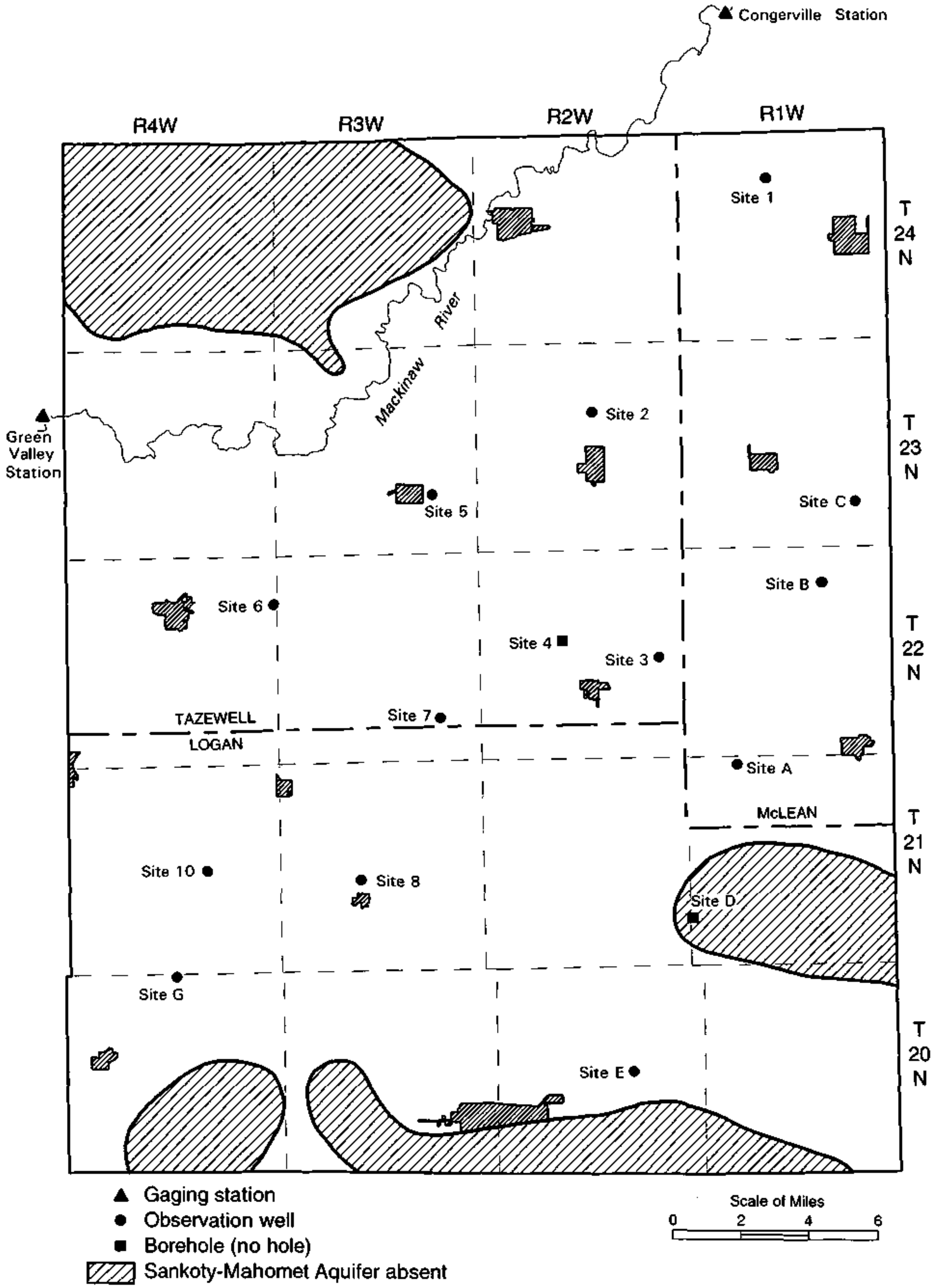

Figure 3. Location of drill sites, observation wells, and gaging stations

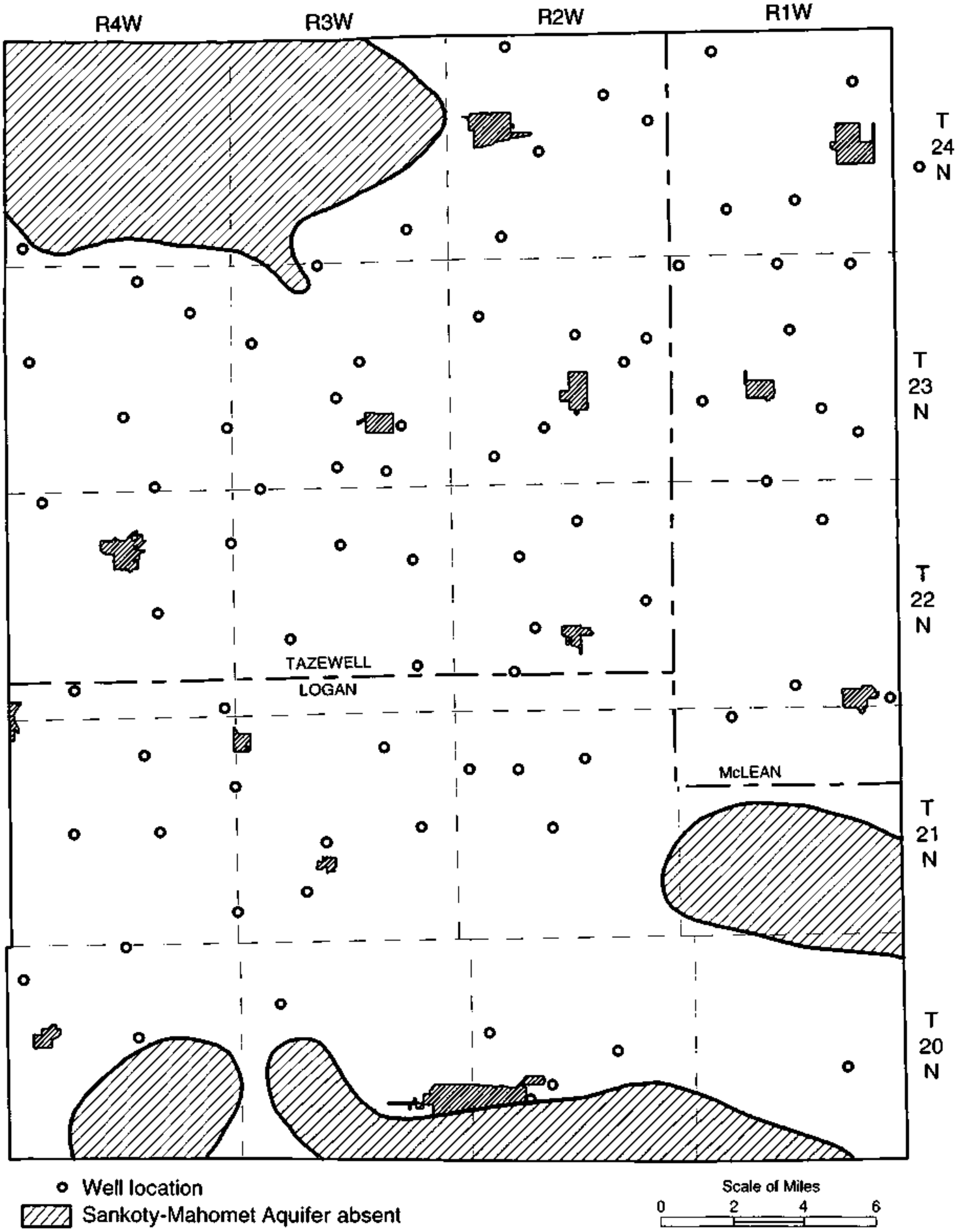

Figure 4. Location of the wells used for mass measurement

Laboratory/Office Methods and Procedures

The principal sources of geologic information were samples and field logs of materials encountered during test drilling along with the geophysical logs (mainly natural gamma). Sample descriptions and logs were compiled of previous test drilling programs in Normal and other communities. Drilling contractors routinely submit drillers logs to the State, and these logs were used to produce both geologic cross sections of materials encountered and maps of bedrock surface and aquifer thickness. Samples were taken at 10-foot intervals from each test hole drilled for the project, laid out in sequence, and described and classified (type and unit) based on the sequence and character of the units identified in the region. These descriptions were then matched with the field log and the natural gamma log to provide a more accurate determination of the top and base of each unit, and to note any units indicated by the gamma log that had not been observed in the samples.

Cross sections were constructed to include the maximum possible drill sites for which samples and geophysical logs were available as well as hydrogeologically significant areas. Within each cross section, geologic units were identified at each well or testhole site, and the geophysical log, where available, was reduced to match the vertical scale and placed next to the descriptive log. The representative nature and extent of the geologic materials was obtained by correlating units of similar characteristics and positions from well to well across the cross section. Thus the cross sections represent simplified vertical slices of the earth along the lines selected.

Elevations were determined for the top of bedrock and the top of the Sankoty-Mahomet Sand and its thickness for each well of record. These values were then plotted and contoured on separate maps.

Ground-water levels measured during the May 1993 mass measurement were contoured to produce a potentiometric surface of the aquifer. Measurements made at the monitoring wells over time were plotted to create ground-water hydrographs. The "fixed base-length" method of baseflow separation was used to estimate ground-water discharge to streams. The baseflow analysis used USGS streamflow data for water years 1990,1991, and 1992 from the Congerville and Green Valley gaging stations on the Mackinaw River (figure 3). Aquifer properties for the Hopedale test well were calculated using the Hantush-Jacob type curve and the Hantush semi-log methods (Hantush and Jacob, 1955; Hantush, 1956). Drawdown data collected during the test were corrected for effects caused by atmospheric pressure fluctuations by calculating a correlation coefficient for the changes in water levels in the aquifer relative to changes in atmospheric pressure.

GEOLOGY

Bedrock Topography

A significant amount of data about the bedrock topography below the Sankoty-Mahomet aquifer have been acquired since the last map of the area was completed in 1991. Findings from the 1992 drilling program are the most significant. Consequently, the map incorporates some locally significant changes that differ from previous maps, although use of previous regional maps (Kempton et al., 1991) ensured integration with regional patterns. Local maps prepared for recent studies of Normal (Kempton and Visocky, 1992) were also used, but they were modified as warranted by new data and a regional approach.

Figure 5 shows the current interpretation of the bedrock topography of the confluence area of the Mahomet and Mackinaw Bedrock Valleys. At Clinton in De Witt County, bedrock ridges to the southwest separate the main channel of the Mahomet Valley from a narrower channel, the Kenney Valley (Kempton and Visocky, 1992; figures 2 and 5). The main channel of the Mahomet is oriented nearly east-west in southwestern McLean County before it opens into the wide lowland of the Mackinaw Valley segment of the Ancient Mississippi River Bedrock Valley at the southeastern corner of Tazewell County (figure 5). Beginning near the village of Kenney, the narrower channel trends northwest across northeastern Logan County and joins the Mackinaw Valley a few miles west of the confluence of the main Mahomet channel with the Mackinaw.

Two major changes in the interpretation of the bedrock topography within the Mackinaw Valley significantly change the interpretation of the thickness and/or extent of the Sankoty Sand. Site 1, shown in figures 3 and 5, demonstrated that the bedrock valley wall was at least three miles farther east than previously mapped, thus potentially expanding the areal extent of the Sankoty-Mahomet Sand. A low bedrock ridge up to an elevation of 400 ft-msl is now mapped beneath the aquifer, northeast and southwest of Hopedale (figure 5), as indicated by information from Site 5.

The drilling program for this study provided a significant number of bedrock surface elevations for the area. Because the program was designed in part to provide information where little or none was previously available, it has resulted in new insights into the shape of the bedrock surface, particularly that of the Mackinaw and Mahomet confluence area. More recent maps of the bedrock topography of all or portions of the study area (Richards and Visocky, 1982; Kempton and Poole, 1985; Kempton et. al., 1991; Kempton and Visocky, 1992) depict the regional configuration and principal elements of the bedrock surface that have either been verified or modified by this study.

The thalweg (deepest part) of the Mahomet Bedrock Valley appears to be at the same elevation as that of the Mackinaw Bedrock Valley (slightly below an elevation of 350 feet). The confluence area forms a wide bedrock lowland emphasized by the Danvers Bedrock Valley, which opens into the lowland from the northeast (figure 2) just west of Bloomington and Normal. On previous maps the east valley wall of the Mackinaw Bedrock Valley north of the Danvers Valley extends westward, constricting the width of the deepest part of the Mackinaw Valley (below 400 feet) to about six miles in that area. A wider valley is now shown (figure 5), an interpretation based in part on data from Site 1 northeast of Danvers (figure 3).

Whereas previous maps have shown the thalweg of the Mackinaw Valley (below an elevation of 350 feet) to extend and widen directly south of these "narrows", the current map (figure 5) suggests that the thalweg follows along the west valley wall and is separated from the Mahomet/Danvers thalwag by a two- to four-mile-wide, low bedrock ridge that trends southwest-northeast from just south of Delavan to northeast of Hopedale. The highest elevation, slightly above 400 feet, is directly below Hopedale.

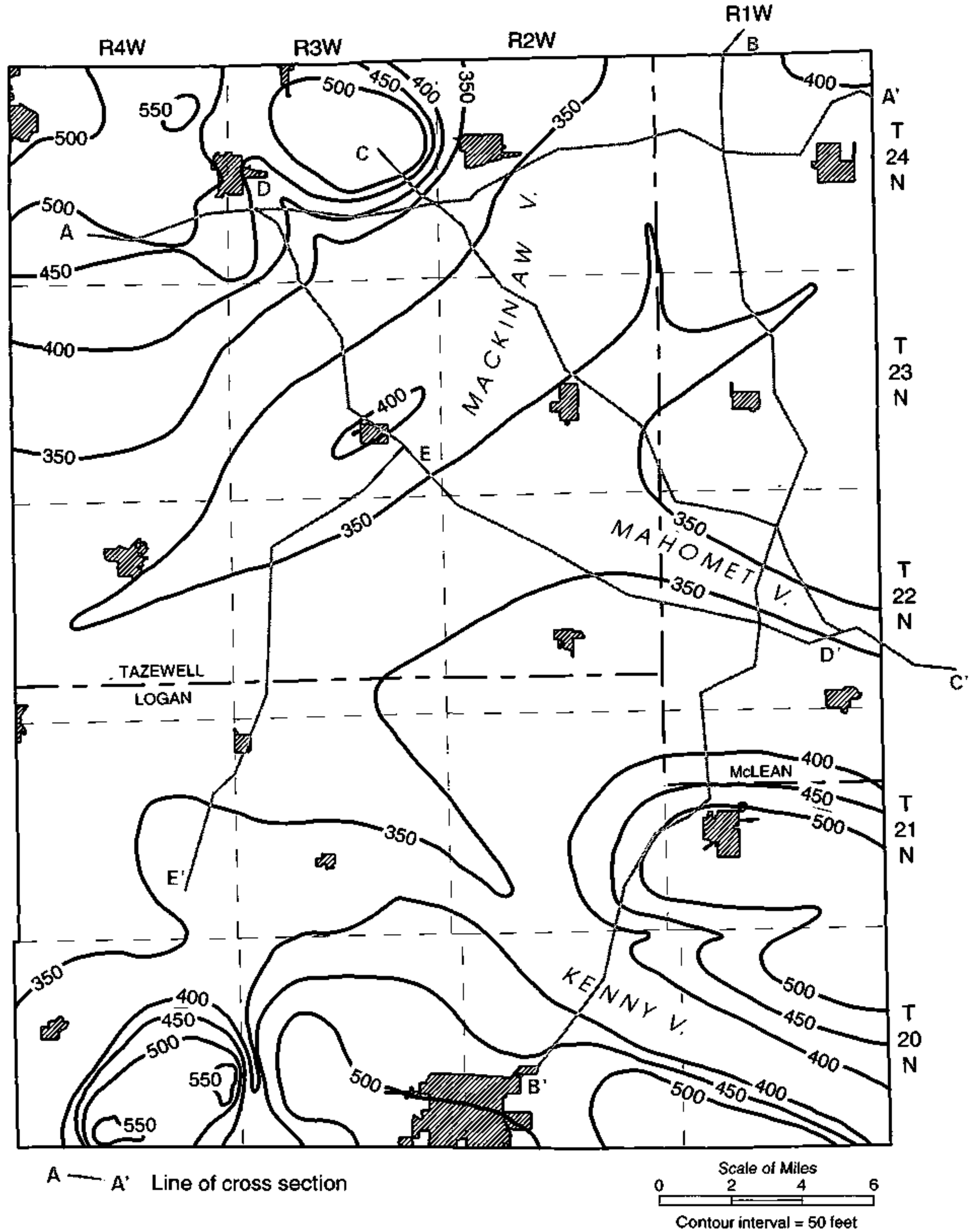

Figure 5. Bedrock topography and locations of cross sections

These two important changes in the interpretation of the bedrock configuration of the Mackinaw Bedrock Valley also have an impact on the interpretation of the thickness and extent of the Sankoty-Mahomet aquifer. Data from Site 1 indicate that the aquifer is potentially wider between the villages of Mackinaw and Danvers. The presence of the bedrock ridge between Delavan and Hopedale indicates that in some places the aquifer is up to 50 feet thinner than was previously assumed.

Sequence and Distribution of Glacial Deposits

The glacial deposits covering the bedrock surface in east-central Illinois range in thickness from about 75 feet to more than 400 feet, and bedrock in the study area is not exposed (see figure 6 in Kempton and Visocky, 1992). The thickest glacial drift, generally found over bedrock valleys, averages more than 200 feet thick throughout the region. Deposited during multiple fluctuations of continental glaciers, drift is composed principally of glacial till.

Glacial tills are unsorted mixtures of clay, silt, sand, gravel, and scattered boulders—the widespread blanket deposits left by individual glaciers. Although many tills have similar properties, each till sheet can usually be distinguished by its specific properties or its relation to other identifiable underlying or overlying materials; therefore, its distribution can be mapped.

Glacial outwash is sand and gravel or other debris deposited by meltwater in front of the ice, in crevasses within the ice, in channels below the ice, and along or beyond the margin of the ice. Outwash deposited in sheets along the margin is called an outwash plain. Valley trains result from outwash deposited down valley from the ice, and ice contact deposits are associated directly with melting ice. Debris that collects on the surface of a melting glacier is also called ablation drift and may consist of both till and water-laid materials. Such deposits have been recognized from studying present glaciers and from mapping the surficial deposits of glaciated regions such as east-central Illinois.

Glacial till is common over much of the area; however, buried sand and gravel is usually less widespread and its presence is less predictable. Outwash plains are generally extensive. Other types of outwash deposits formed as long, narrow channels or ridges, and as mounds of variable size and distribution.

High-energy meltwater streams that flow from a glacier transport relatively coarse- to medium-textured materials (e.g., sand and gravel) and some fine material (silt and clay), much of which is deposited as outwash. Glacial meltwater normally carries the fine silt and clay great distances. Occasionally, these sediments are trapped behind rapidly accumulating outwash, especially in tributary valleys, and are deposited as fine-textured lake sediments (silt and clay).

All of these conditions occur in central Illinois throughout the succession of glacial deposits. Although the thicker, more extensive subsurface outwash materials have been identified and mapped (especially the Mahomet and Sankoty Sand Members of the Banner Formation), many thinner and less extensive sand-and-gravel deposits have been identified only at specific locations or from individual well records. Lineback (1979) mapped the distribution of the surficial materials.

Figure 6 shows a generalized sequence of glacial materials in the study area. Sankoty Sand is the sand and gravel at or near the base of the glacial deposits in the Mackinaw Bedrock Valley, and the Mahomet Sand in the Mahomet Bedrock Valley (Horberg, 1950); however, the deposits are similar in stratigraphic position and depositional environment. Collectively, these deposits are geologically referred to as the Sankoty-Mahomet Sand.

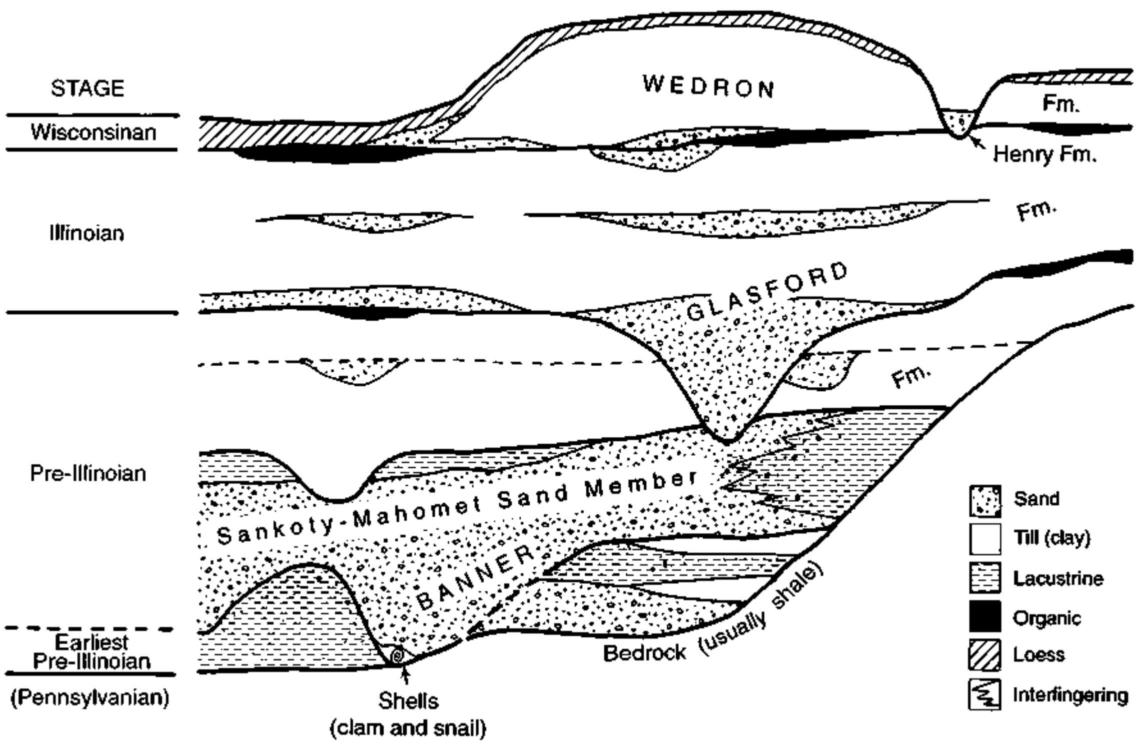

Figure 6. Sequence of geologic materials (stratigraphic column)

The pre-Illinoian drift, often called the Banner Formation, was deposited throughout the area about two million to 500, 000 years ago. Comprising the oldest glacial deposits identified in the area, the Banner includes the principal sand-and-gravel aquifer of the region, the Sankoty-Mahomet Sand. These sand-and-gravel deposits, where present, are found mostly in the bedrock valleys. Along the bedrock valley walls, these deposits may grade laterally into fine-textured silts and sands, overlie older fine-textured materials (figure 7), or interbed with glacial tills (figure 6). The top of the Sankoty-Mahomet Sand lies generally between an elevation of 425 and 500 ft-msl (figure 8).

Identification of Banner Formation tills is possible because their physical and mineralogical characteristics differ from those of the overlying tills. Occasionally, buried soils or organic deposits are found at the top of the formation, a feature suggesting a significant period of time between deposition of the Banner Formation and the overlying Glasford Formation. Banner Formation deposits have been identified locally near the base of the glacial deposits on the bedrock uplands away from the major bedrock valleys. In these cases, they are generally thinner and contain only scattered, thin, less extensive sand-and-gravel layers.

The overlying Glasford Formation, deposited 500,000 to 150,000 years ago during the Illinoian stage of glaciation, is present throughout the study area and contains several sand-and-gravel layers, which are interbedded with tills and used as aquifers throughout the area. As indicated in figure 6 and the cross sections (figure 7), one locally significant sand-and-gravel aquifer lies at or near the base of the Glasford Formation. East of the study area, it is the

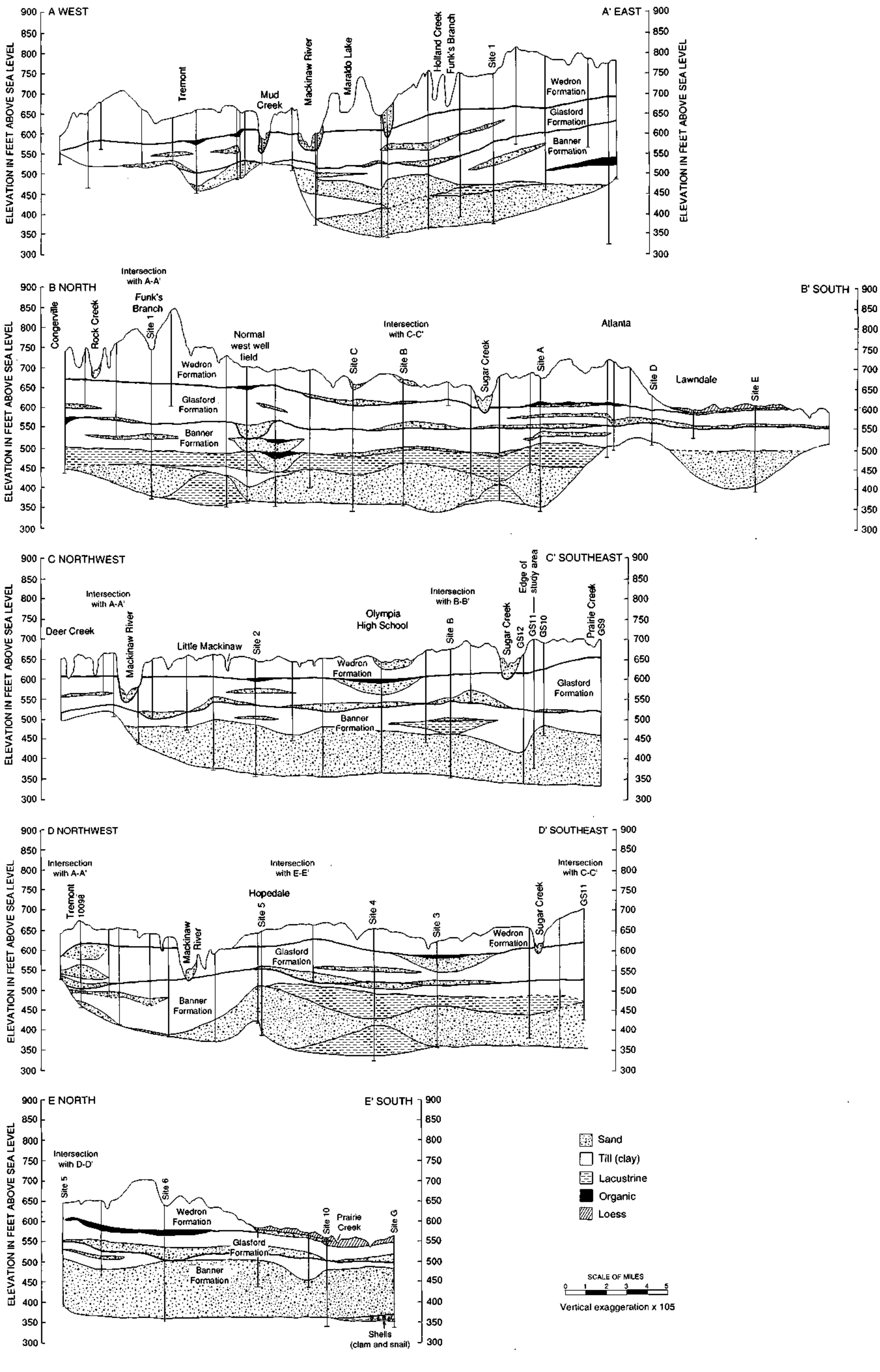

Figure 7. a) Northern east-west cross section A-A', b) Eastern north-south cross section B-B', c) Northern northwest-southeast cross section C-C, d) Southern northwest-southeast cross section D-D', and e) Western north-south cross section E-E' (see figure 5 for locations)

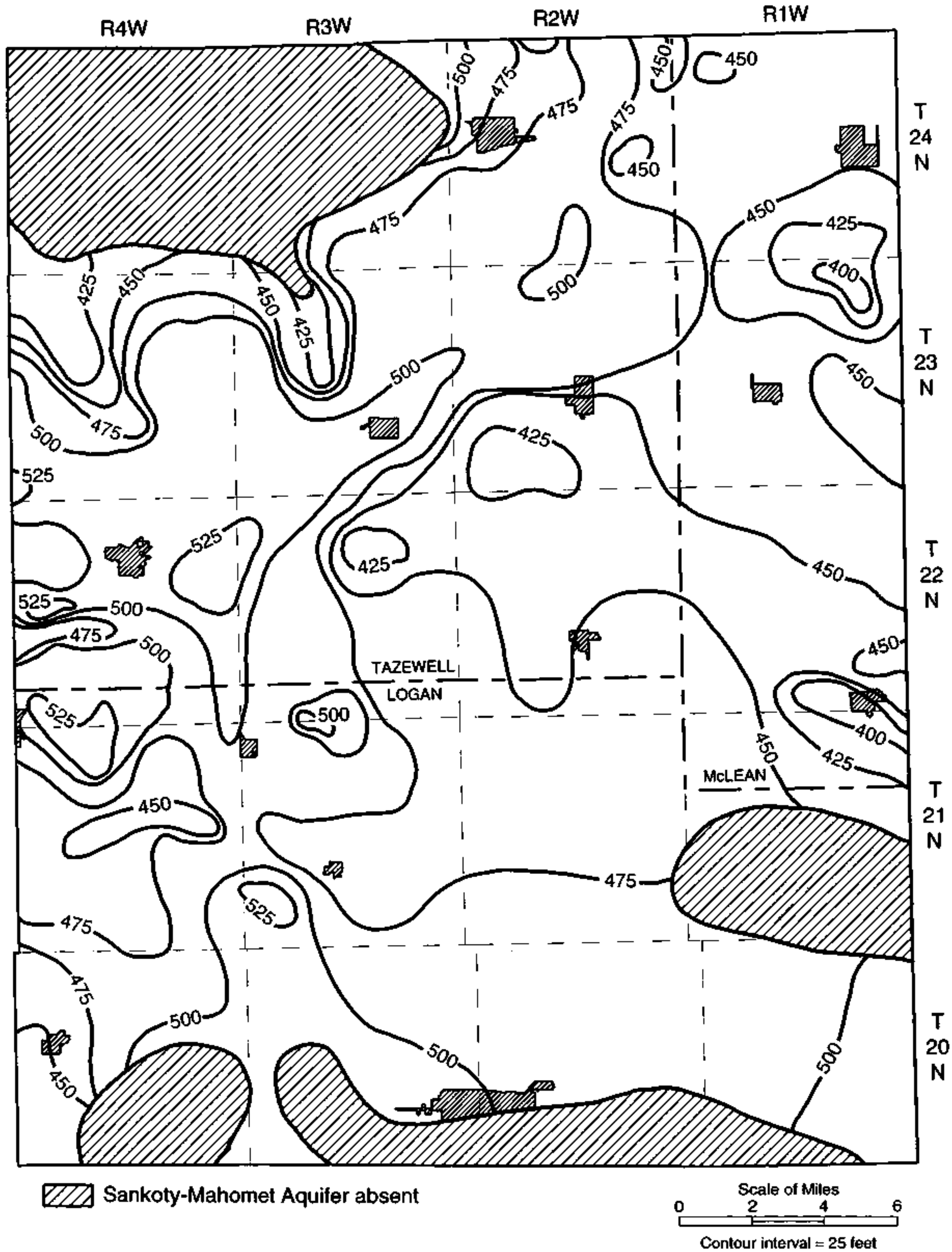

Figure 8. Elevation of the top of the Sankoty-Mahomet aquifer system

deepest aquifer used by the town of Normal, and it also directly overlies the Sankoty-Mahomet Sand locally within Normal's West Well Field and elsewhere in the study area. Where directly overlying the Sankoty-Mahomet aquifer, the two aquifers are hydraulically connected (Richards and Visocky, 1982).

An upper Glasford Formation sand and gravel may be regionally extensive but thin. It is used as an aquifer, however, supplying ground water to farms, residential subdivisions, and small businesses. The top of the Glasford Formation is capped by a widespread buried soil, silt beds, and discontinuous organic deposits (Robein Silt) that provide a regional marker horizon; some local sand-and-gravel channels may also be present at the top of the Glasford (figure 7c).

The Wedron Formation, principally till, was deposited 75,000 to 12,000 years ago during the Wisconsinan stage of glaciation and is the uppermost glacial deposit in Illinois. Only scattered sand-and-gravel lenses are present within or at the base of the Wedron Formation. Ground water from these lenses is used only locally for small households and farms.

The Henry Formation directly underlies the land surface along the principal streams of the area (Sugar Creek, Kickapoo Creek, and the Mackinaw River). It is composed of sand and gravel deposited by outwash draining from the melting Wisconsinan glacier (75,000 to 12,000 years ago). Locally, it may be 50 to 70 feet thick along these streams, although it averages 25 feet. Prior to Henry Formation deposition, these valleys were initially eroded into the Wedron Formation and, in some places, into the Glasford Formation. Therefore, in places the sand and gravel of the Henry Formation may connect directly with the sand and gravel in the upper part of the Glasford Formation. In such situations, there is the potential for developing moderate ground-water supplies (e.g., Kickapoo Creek near Hey worth, Sugar Creek in and just south of Bloomington-Normal, and the Mackinaw River at Mackinaw).

Position and Distribution of Sand-and-Gravel Aquifers

Banner Formation (pre-Illinoian)

Because of their generally deep burial beneath younger deposits, deposits currently assigned to the pre-Illinoian Banner Formation (figure 6) in central and east-central Illinois have not been well understood. Until recently, much of our knowledge of these deposits and the aquifers in them has been based on the work of Horberg (1950, 1953). Few wells penetrated to bedrock, and it was generally assumed that the Sankoty-Mahomet Sand was the basal unit, consisting almost entirely of a relatively uniform body of sand and gravel up to 150 feet thick and filling the lower part of the two bedrock valleys. Local studies (Hunt and Kempton, 1977; Kempton et al., 1982; Kempton and Visocky, 1991) mapped the surface and/or thickness of the Sankoty-Mahomet Sand in some detail, although much of these data were based on numerous downhole geophysical log interpretations. Geophysical logs (primarily natural gamma logs) were also used to determine depth to bedrock. Because samples from only a few wells were available, they provided little information. Initially it was assumed that the Sankoty-Mahomet Sand in these areas was continuous from its top to bedrock, although evidence from the geophysical logs indicated finer textured zones.

The first indication that the thickness might be less in some areas came when the local bedrock topography was remapped and revealed important changes from Horberg's (1950) mapping. These changes were summarized and shown by Kempton et al. (1991). In addition, work in the Mahomet Valley and the Mackinaw Valley (Richards and Visocky, 1982; Kempton and Visocky, 1992) had also suggested that the lithic characteristics of the Sankoty-Mahomet Sand sequence showed significant complexities, somewhat similar to those suggested in the Mahomet by Kempton et al. (1991). Kempton and Visocky (1992) suggested that the main body of coarse-textured sand and gravel was locally a fill into a partially eroded, finer textured deposit, particularly in the Mahomet Valley near the confluence with the Mackinaw Valley. Our study adds significant new data to support and expand on these conditions in the Mackinaw Valley and the confluence area in the Mahomet Valley.

The test drilling program documented the local occurrence of finer textured water-laid (lacustrine) deposits below the Sankoty-Mahomet Sand (Site 4; figure 3) and supported the interpretation of similar deposits elsewhere (figures 7b, 7d). Lacustrine deposits directly overlie the Sankoty-Mahomet Sand in some portions of the area, particularly in the northern and eastern portions. The recognition of additional areas of fine-textured, water-laid deposits indicates that the Sankoty-Mahomet Sand is not as thick and extensive as was originally mapped in 1953, at least locally, and as suggested by Kempton and Visocky (1992).

The main body of the Sankoty-Mahomet Sand is still recognized as the principal regional aquifer for the development of large ground-water supplies. The aquifer's highest surface in the area is about 525 ft-msl but may be as low as 400 ft-msl (figures 7 and 8). Previous mapping (figure 9) suggested thick deposits over much of the area. In most of the area, the aquifer averages 100 feet thick and may be as much as 150 feet thick locally, particularly in the western part (figure 10). The aquifer is usually overlain by till or lacustrine materials about 30 to 50 feet thick. Locally, however, post-Banner Formation erosion has cut channels through these finer materials into the Sankoty-Mahomet Sand. In some areas, later (Glasford Formation) sand and gravel was deposited directly on the Sankoty-Mahomet Sand (figure 6) as in the Normal West Well Field (Richards and Visocky, 1982; Kempton and Visocky, 1992), and this may occur throughout the present study area. Another layer of sand and gravel may be present in some areas between the upper till or lacustrine units of the Banner Formation but appears to be significant in terms of aquifer potential only locally in the vicinity of the Normal West Well Field (figure 7b).

Of interest from a geologic perspective is the presence of about 20 feet of dark gray gravelly, silty, sand-containing shells of clams and snails directly above bedrock at the base of the Sankoty-Mahomet Sand (figures 6 and 7e) in Site G (figure 3). A similar deposit has been reported for the Mahomet Valley at the base of the Mahomet Sand in southeastern De Witt County (Miller et al., 1992). Amino-acid dating of these shells supports the conclusion of other studies (Kempton et al., 1991) that some of the older deposits are more than 730,000 years old. These shell-bearing materials are probably related to the thick pre-Illinoian lacustrine materials found below the Sankoty-Mahomet Sand elsewhere in the area (figures 6 and 7).

Glasford Formation (Illinoian)

The Glasford Formation overlies the Banner Formation throughout the study area (figures 6 and 7). Although composed predominantly of glacial till, the Glasford has three zones that contain sand-and-gravel layers: at the base, between the two main tills, and at the top (figures 6 and 7). The basal layer may be the most hydrogeologically important where it directly overlies the Sankoty-Mahomet Sand (Site 6 in figures 3 and 7e). The basal Glasford sand and gravel appears consistently throughout the area and is probably the source of ground water for numerous domestic and farm wells. Its presence may also be the reason that so few wells of record were drilled into the Sankoty-Mahomet Sand. Although it is generally no more than 20 feet thick on the average and somewhat discontinuous locally, its greatest thickness appears to be where it fills channels cut into the Banner Formation. This is apparent in various cross sections (Richards and Visocky, 1982; Kempton and Visocky, 1992; figures 7a-e).

The middle sand-and-gravel unit is generally less continuous and thinner than the basal unit and is more limited in areal extent. Similarly, the upper sand and gravel occurs only locally as channel fill (figures 7c-e) deposited by the last Illinoian glacier.

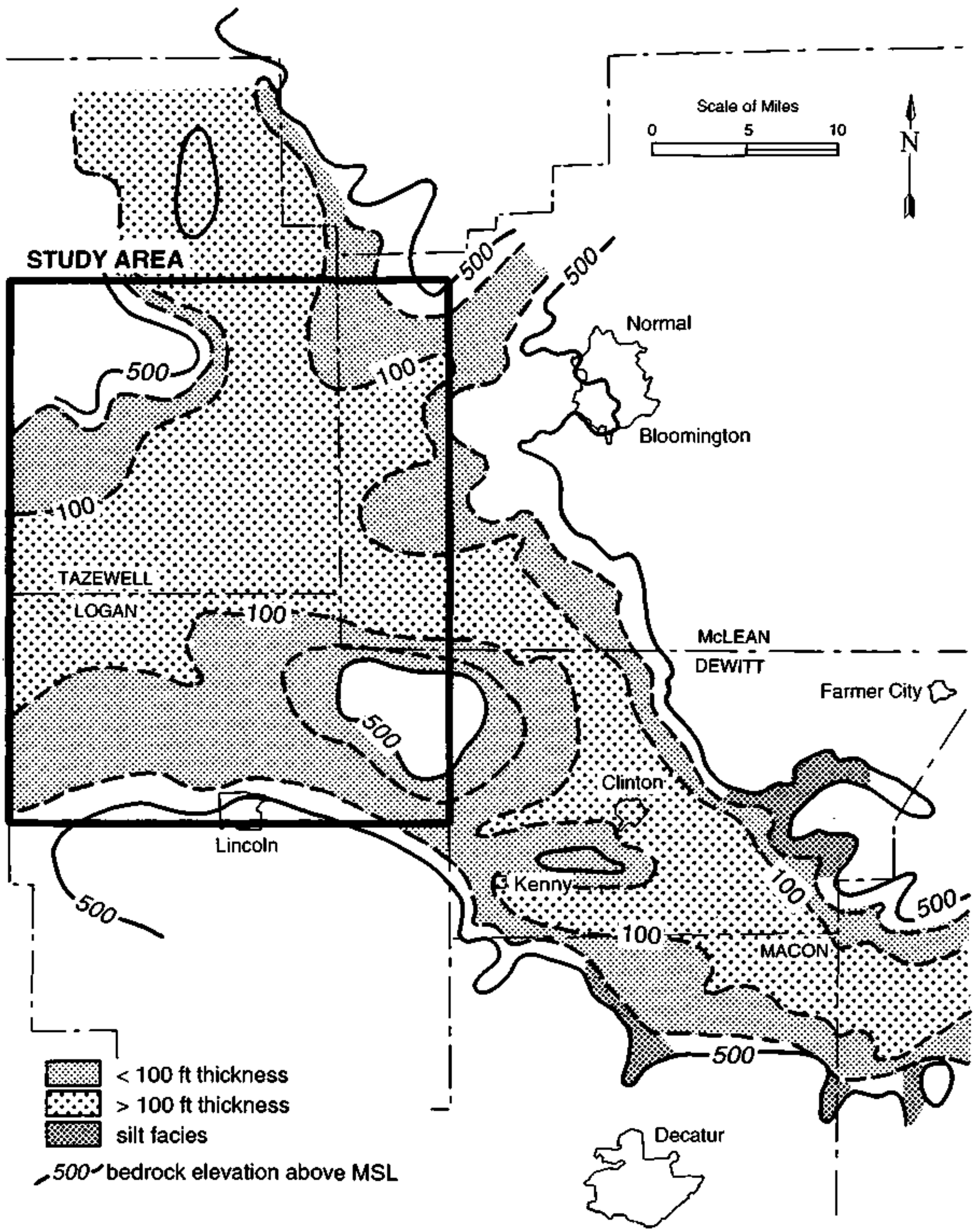

Figure 9. Regional thickness and distribution of the Sankoty-Mahomet Aquifer system (Kempton and Visocky, 1992)

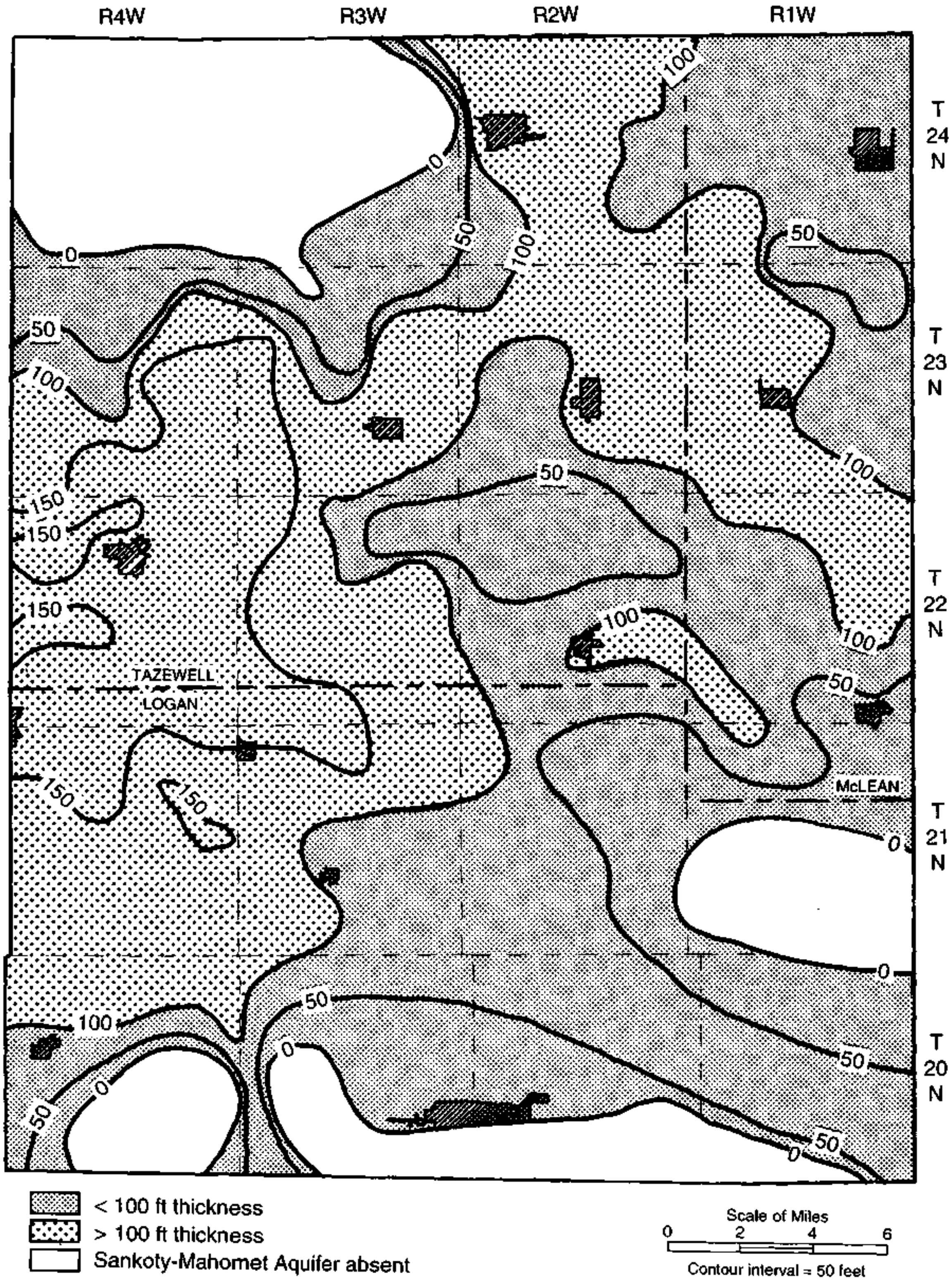

Figure 10. Thickness of the Sankoty-Mahomet aquifer system

Wedron Formation (Wisconsinan)

Sand and gravel associated with the Wedron Formation is found primarily as a discontinuous layer at the base of the till (figures 7b-d). No data suggest these deposits are extensive, but, where present, they may provide a local source of small supplies of ground water. Because the Wedron Formation is composed almost entirely of a single till in the study area, no other gravel layers of significance are likely to be found.

Henry Formation (Wisconsinan)

Surficial sand and gravel, related to melting of the glacier that produced the till of the Wedron Formation, is found along the larger streams and river valleys of the area, especially in Kickapoo Creek, Sugar Creek, and the Mackinaw River Valleys. These deposits may be up to 60 feet thick and occasionally lie against older sand-and-gravel layers (figure 7). They may yield small municipal ground-water supplies locally.

Relationship of Geology to Ground-Water Availability

Nearly all geologic materials will transmit water, but at different rates depending on the type of material; for example, water moves very slowly through clay but relatively rapidly through gravel. The amount of water available from an aquifer during a given period depends on (1) the rates at which the materials transmit water, (2) thedimensions of the body of materials yielding water, and (3) the amount and rate of recharge to or discharge from the body of materials. Thicker, more extensive sand-and-gravel deposits have a greater potential as a municipal ground-water supply source.

The availability of ground water is, therefore, controlled by the nature and arrangement of various earth materials beneath the surface. Because geologic conditions vary, ground water is readily available in some areas and difficult to obtain in others. Consequently, gathering information on the distribution, sequence, and character of geologic materials as aquifers assists community planners in developing ground-water resources.

The fundamental classification of geologic material is the lithostratigraphic unit (group, formation, member). As indicated previously, this classification is based on the lithologic characteristics of the material (i.e., type of material), controlled by the method of deposition and the environment into which the material was deposited during a given period of time. Therefore, the sequence of glacial sediments varies widely both vertically and horizontally, and depositionally related sediments can differ significantly in their lithic characteristics because of the highly variable nature of glacial melting and depositional processes. As a result, materials within the same depositional sequence, assigned to a geologic formation, may not have similar lithic characteristics (e.g., water-laid sand and gravel grading to silts and clays).

Because lithostratigraphic units and hydrologically similar units are not always coincident, a separate classification of geologic materials, referred to as hydrostratigraphic units, has been developed (Maxey, 1964; Seaber, 1988). Visocky et al. (1985) introduced a formal hydrostratigraphic classification in Illinois. They divided geologic materials into three major hydrostratigraphic units, the uppermost of which consists of local and intermediate flow systems in nonindurated geologic materials, including alluvium and glacial drift. They considered local precipitation to be the major component of recharge to this system. This unit is composed of confining units, or aquitards (primarily glacial tills and fine-textured lacustrine sediments that transmit water slowly), and aquifers (mainly sand and/or gravel that transmit water rapidly).

For hydrologic evaluations in this report, we have used the concept of hydrostratigraphic unit (Sankoty-Mahomet aquifer) as opposed to the geologic lithostratigraphic unit (Sankoty-Mahomet Sand Member, Banner

Formation). Consequently, the lacustrine units at the top of Sankoty-Mahomet Sand Member are not part of the aquifer, although they are part of the formation. When because of erosion, younger overlying sand-and-gravel units of different formations directly overlie the Sankoty-Mahomet aquifer, they become part of that aquifer system.

The basic hydrostratigraphic unit referred to in this report, therefore, consists of the Sankoty and Mahomet Sand Members and the confining units that separate it from other aquifers. An upper aquifer, such as a Glasford Sand, is considered part of the hydrostratigraphic unit when it is present (figures 6 and 7e) and is hydraulically connected to the Sankoty-Mahomet aquifer.

HYDROLOGY

The goal of any aquifer assessment is to determine the quantity and quality of ground water available for use. This study emphasizes ground-water availability but also includes an overview of available water chemistry data. A water quality assessment of an aquifer frequently warrants a completely separate, directed research effort.

This chapter is limited to a discussion of the Sankoty-Mahomet aquifer system, where "system" includes any hydraulically connected upper aquifers. Although the aquifer extends well beyond the study boundaries, the study area is unique because it is where the Ancient Mississippi and Mahomet Bedrock Valley systems converge. This region of the aquifer system has not previously been studied in any detail because of a lack of wells extending to bedrock. Few private wells penetrate to bedrock because the shallower glacial deposits have provided ample water supplies.

Aquifer Characteristics

Ground-water exists in the pore spaces between soil particles that compose glacial materials and in fractures in bedrock. Water moves through these openings within an aquifer either toward a pumping well or toward a discharge point. The size, homogeneity, and interconnection of the pore space openings in part determine the rate of water flow through them. Pumping a well induces flow toward the well, and the water-level decline due to pumpage is called drawdown. The areal extent of the drawdown is the radius of influence that a given pumping well has on the aquifer. A potentiometric surface around a pumping well forms an inverted cone, the cone of depression, with the well at the center. Drawdown associated with a well pumping at a given rate is dependent upon the areal extent, thickness, and hydraulic properties of the aquifer system.

An aquifer may exist under artesian, leaky artesian, or water-table conditions. Artesian refers to the aquifer being under pressure greater than atmospheric pressure. The result is that water levels in artesian wells are above the top of the aquifer. Artesian conditions exist where overlying fine-grained materials with low permeabilities (aquitards or confining beds) limit vertical movement of water into and out of the aquifer. This "confines" the aquifer under artesian pressures. The term "leaky" refers to the degree of limitation of vertical water movement into or out of an aquifer. A nonleaky condition exists if no vertical water movement occurs and a leaky condition exists when the confining unit does allow some vertical water movement into or out of the aquifer. The degree of leakage is determined by the rate of water movement into or out of the aquifer across the confining bed. A water-table (unconfined) system exists when the aquifer is directly connected to the surface. Under these conditions the potentiometric surface of the aquifer is at atmospheric pressure and is referred to as the water table. Under water-table conditions, drawdown due to pumping dewaters a portion of the aquifer, allowing air to fill that pore space. In the artesian case, pumping a well lowers the water pressure in and around the well. The aquifer itself does not dewater, however; it remains fully saturated unless water levels fall below the top of the aquifer. When this occurs, the aquifer changes to water-table conditions in the dewatered area.

The Sankoty-Mahomet aquifer is commonly overlain by glacial till or other fine-grained materials throughout the study area (figures 7a-e). In the confluence area, the aquifer generally exists under leaky artesian conditions, but the aquifer system becomes unconfined (exists as a water-table aquifer) west of the study area in Mason County. In the western part of the study area there are limited regions where water-table conditions exist in the aquifer. These conditions occur where the Sankoty-Mahomet aquifer is hydrologically connected to an upper aquifer (Glasford Formation or Henry Formation), that is itself under water-table conditions.

Aquifer Hydraulics

The hydraulic properties that determine aquifer yield and aquifer response to pumpage, or drawdown, include the aquifer's transmissivity and storage capacity, and vertical hydraulic conductivity and thickness of the overlying confining units. Transmissivity (T), the aquifer's capacity to transmit water, is a measure of the horizontal velocity of ground water through an aquifer of a specific thickness. Transmissivity is defined as the rate of flow of water through a vertical strip of aquifer of unit width under a unit hydraulic gradient. It is the product of the aquifer's hydraulic conductivity (K) and its saturated thickness (m).

The storage coefficient (S) represents the aquifer's ability to release water. It is the change in volume of water released or stored per unit surface area of aquifer due to a unit change in water level. A change in storage can be attributed to (1) compressibility of the water and the aquifer material or (2) change in pore space either dewatered from pumping or refilled from recharge of the aquifer. Under artesian conditions, the aquifer is completely saturated and the storage coefficient is based on the first condition. Typical values range from 10^{-5} to 10^{3} (tables 1 and 2). Water-table aquifers, however, are not completely saturated and a change in storage is attributed to both conditions. The change in storage is almost entirely due to the draining or refilling of the pore spaces, however, since the portion attributed to aquifer and/or water compressibility is negligible by comparison. With values generally between 0.05 and 0.3, water-table aquifers have a much higher storage coefficient than artesian aquifers (Kempton and Visocky, 1992).

Vertical leakage through a confining bed can be significant. Leakage depends upon vertical hydraulic conductivity (K') of the confining bed, thickness of the confining bed (m'), and the water-level difference between the aquifer and the source bed above the confining unit. Leakage is typically defined in terms of the ratio of vertical hydraulic conductivity and thickness of the confining bed (K'/m'), termed by Hantush (1956) as the leakage coefficient. This coefficient is the discharge of water through a unit area of interface between an aquifer and a confining bed for a unit head difference across the confining bed.

Aquifer Tests

A controlled pumping test is conducted to provide data necessary to determine specific aquifer properties. During the test, water levels are continually measured in the pumped well and at nearby observation wells. These data are analyzed using plots of time versus drawdown and distance versus drawdown. Several formulas have been developed that relate the aquifer properties (T and S) to the drawdown near a pumping well (Theis, 1935; Cooper and Jacob, 1946; Hantush and Jacob, 1955; Hantush, 1956; Ferris, 1959; Walton, 1962; Boulton, 1963; Prickett, 1965; Neumann, 1975). When only the pumped well data are gathered, transmissivity can accurately be determined, but there is the potential for considerable error in determining the storage coefficient from such data. Results of historical pump tests for the study area and other nearby tests in the Sankoty-Mahomet aquifer are listed in table 1 and summarized in table 2.

Table 1 includes data from eight tests to determine specific capacity or the yield of a well in gallons per minute per foot of drawdown. Specific capacity varies with the radius of the well and the length of pumpage. Walton (1962) developed a modified nonequilibrium formula that relates specific capacity to transmissivity by assuming a particular storage coefficient. This formula also makes several other assumptions: that the well fully penetrates the aquifer, that there is no influence from the drilling around the screen, that well loss is negligible, and that the aquifer is homogeneous, isotropic, nonleaky, and of infinite areal extent. There are several limitations to this method, most notably that the effective well radius usually exceeds the assumed nominal well radius. Water-table conditions also create excessively high estimates for short duration tests because time-drawdown relations are distorted. Due to the imprecise nature of these analyses, those values in table 1 determined by the specific capacity method are only estimates.

Table 1. Aquifer Tests and Specific Capacity Data in Basal Aquifer in or near Study Area

Well location	*Well owner*	*Depth (ft)*	*Pumping rate (gpm)*	*Specific capacity (gpm/ft)*	*Analysis method*	*Transmissivity (gpd/ft²)*	*Hydraulic conductivity (gpd/sqft)*	*Storage coefficient*
Dewitt								
20N1E-								
28.4al	County Nursing Home	336	9	5.9	S	12,047		
28.4a3	County Nursing Home	326	75	7.4	T	53,500	920	
21N1E-								
29.7b8	Waynesville (OW1)	217			T	6,200		0.0007
Logan								
20N4W-								
18.2d	New Holland	155	75	3.1	S	49,500	2,150	
21N3W-								
6.7c	Emden	124	205	14.6	S	32,000		
McLean								
22N1E-								
16.7dl	IDOT (Funk's Grove)	322	37	2.9	T*	5,900		
22N1W-								
6.1hl	Olympia High School	250	215	21.5	T	153,400	1,870	
35.1b1	McLean County	353			S*	1,400		
35.1b3	McLean County	340	203	16.4	R	75,800	2,920	
35.8c4	McLean County	332	203	11.8	T	133,700	2,060	
23N1E-								
6.8hl00	Normal	346	1416	128.3	T	340,400	3,660	

Note: Four different analysis methods were used: time-drawdown (T), time-recovery (R), specific capacity (S), or a combination (TR). An asterisk indicates that wells may be finished in finer Mahomet or pre-Mahomet deposits.

Table 1. Continued

Well location	*Well owner*	*Depth (ft)*	*Pumping rate (gpm)*	*Specific capacity (gpm/ft)*	*Analysis method*	*Transmissivity (gpd/ft^2)*	*Hydraulic conductivity (gpd/sqft)*	*Storage coefficient*
6.8hl00	Normal (OW1)	268			T	201,800	2,170	0.0050
6.8hTH20	Normal	268	425	153.4	T	298,700	3,280	
6.8hTH20	Normal (OW1)	275			T	266,400	2,930	0.0002
6.8hTH20	Normal (OW2)	269			T	289,900	3,260	0.0020
6.8hTH20	Normal (OW3)	268			T	273,800	2,980	0.0100
6.8hTH20	Normal (OW4)	200			T	289,900		0.0008
3N1W-								
10.1hl03	Normal	328	1078	59.5	T	147,400		
10.1hl03	Normal (OW1)	253			TR	146,700		0.0004
10.1h103	Normal (OW2)	256			TR	136,200		0.0004
10.1h103	Normal (OW3)	250			TR	138,300		0.0003
10.1hTW21	Normal	324	567	31.2	T	126,700	1,690	
10.1hTW21	Normal (OW1)	325			T	117,900	1,340	0.0002
10.1hTW21	Normal (OW3)	323			T	117,900	1,370	0.0002
10.1hTW21	Normal (OW4)	324			T	127,700		0.0002
21.5c	Stanford (OW4)	235			T	76,700		0.0001
21.5c3	Stanford	247	81	21.8	T	89,100	2,480	
21.7d4	Stanford	246	150	13.0	T	77,600		
24N1E-								
9.7aTHl	D. Grieder Sod Farm	280	55	7.2	R	132,000	4,130	
24N1W-								
23.1g3	Danvers	417	195	9.1	S	16,500		

Note: Four different analysis methods were used: time-drawdown (T), time-recovery (R), specific capacity (S), or a combination (TR). An asterisk indicates that wells may be finished in finer Mahomet or pre-Mahomet deposits.

Table 1. Concluded

Well location	*Well owner*	*Depth (ft)*	*Pumping rate (gpm)*	*Specific capacity (gpm/ft)*	*Analysis method*	*Transmissivity (gpdlft²)*	*Hydraulic conductivity (gpd/sqft)*	*Storage coefficient*
23.1g4	Danvers	438	120	6.6	T	28,300	620	
35.2al02	Normal	364	1409	107.4	T	470,400	2,630	
35.2al02	Normal (supply)	239	488	22.2	S	45,166		
35.2al02	Normal (OW1)	239			T	173,600		0.0900
36.5al01	Normal	345	1409	143.3	T	516,600		
36.5al01	Normal (supply)	243	480	25.3	S	52,560		
36.5al01	Normal (OW1)	324			T	127,700		0.0002
TAZEWELL								
22N2W-								
22.5a	Armington	213			S	11,500		
22N4W-								
16.8b1	H. Walker Dist.	209	2248	97.7	T	280,200		
16.8d3	H. Walker Dist.	212	199	17.8	T	132,000	770	
24N2W-								
3.4aTWl	IDOC hatchery (OW2)	299			TR	309,700	2,230	0.0005
3.4aTWl	IDOC hatchery (OW3)	300			TR	337,000	2,530	0.0003
10.4hTWl	IDOC hatchery (OW1)	308			TR	249,100	1,850	0.0009
10.4hTH	IDOC hatchery (1-75)	295	1001	23.3	R	283,900		
18.4d5	Mackinaw	151	180	41.4	R	71,700		

Note: Four different analysis methods were used: time-drawdown (T), time-recovery (R), specific capacity (S), or a combination (TR). An asterisk indicates that wells may be finished in finer Mahomet or pre-Mahomet deposits.

Table 2. Summary of Aquifer Test and Specific Capacity Data from Table 2

Property	*Low*	*High*	*Average*
Transmissivities (gpd/ft)	6,200	516,600	163,200
Hydraulic conductivities (gpd/ft^2)	620	4,130	2,212
Storage coefficients	.0001	.09	
Specific capacities (gpm/ft)	3.1	153.4	38.4

A seven-day aquifer test was conducted on Hopedale Municipal Well 4. Two observation wells were monitored throughout the test: Observation Well 1 (OW1) 140 feet east of the pumped well and Observation Well 2 (OW2) 570 feet north of the pumped well.

To maintain an adequate water supply for the village of Hopedale during the test, water was pumped through the existing water system at an average rate of 240 gpm. Excess water was released through a fire hydrant.

There were several complications during the test. First, the test was run inline as a pressurized system with no open flow discharge, which caused the rate to fluctuate with changes in pressure when the water level in the tower changed or a hydrant was opened to discharge excess water. Second, without an open flow discharge, flow rates had to be measured using a totalizing meter which allowed calculation of average discharge rates but did not provide instantaneous discharge rates. Third, ground-water levels in both observation wells responded dramatically to changes in atmospheric pressure during the week of the test.

Analysis of data from OW1 indicates that the aquifer at Hopedale is very permeable. This was expected because geologic samples collected during drilling of observation wells contained more than 100 feet of sand and gravel at this location. Transmissivity was calculated as 343,800 gallons per day per ft (gpd/ft), and the hydraulic conductivity was 2,900 gpd/ft^2. The storage coefficient was 0.00057, in the typical range for artesian conditions. The analysis indicated a relatively small amount of leakage from the glacial till directly above the aquifer. The leakage coefficient was calculated to be 0.54 gpd/ft^3 and the vertical hydraulic conductivity was approximately 1.3 gpd/ft^2.

Data collected at OW2 appeared to be influenced either by barometric pressure changes or by mechanical difficulties with the pressure transmitter. Consequently, confidence in the results was compromised. Nonetheless, the results suggested that the transmissivity was lower at OW2 (estimated to be between 85,000 and 99,000 gpd/ft), the storage coefficient was in the artesian range (0.00068), and the leakage coefficient was an order of magnitude lower than at OW1 (0.13 gpd/ft^3).

Two analytical methods were used to determine the aquifer properties: log-log (type curve) and semilog (Hantush and Jacob, 1955; Hantush, 1956). Consistency and quality of the results at OW1 suggest that these properties are representative of actual conditions.

Observed Ground-Water Levels

Sixteen observation wells (figure 3) were installed at 13 sites in June and July of 1992 and serve several purposes. Periodic water-level measurements are taken to monitor fluctuations in the potentiometric surface over time, or determine the hydraulic properties of the aquifer during a controlled pumping test. Water samples from these wells can be collected to determine the chemical character of the ground water. Table 3 lists observation well data.

The wells have been measured regularly since July 1992, and the data gathered from each monitoring well are presented as hydrographs in appendix 2. The hydrographs provide a useful visual tool in interpreting various events that occurred over time during the study (differences in flow paths, connection between aquifers, and effects of the 1993 flood).

One might assume that the water levels in the aquifer would fluctuate as a single unit, i.e., when water levels increase in one part, there would be a corresponding increase in water levels in the rest of the aquifer. However, water-level fluctuations observed in separate parts of the study area have exhibited characteristics independent of each other. Because water levels along similar flow paths will react similarly, observed differences in the data suggest two distinct flow regions in the study area. Figure 11a presents hydrographs for Site B and Site 6, and figure 11b presents hydrographs for Site E and Site 10. Each set of wells has reacted consistently for the duration of the study. Notice the location of these wells in figure 3. Site B and Site 6 are 16 miles apart, and Site E and Site 10 are 13 miles apart. Site C, Site 2, and Site 5 have similar hydrographs to Site B and Site 6. Hydrographs for Site G and Site 8 are also very similar to those for Site E and Site 10.

Figures 11a and 11b imply that the southern wells react differently than the northern wells. An explanation may be that the wells exist in two distinct flow regimes: northerly wells follow a flow path from the main channel of the Mahomet Bedrock Valley, while the southerly wells follow the flow path of the Kenney Valley. Because geologic appraisals had indicated that the Kenney Valley probably had little significance to the regional flow system, it had not been hydraulically evaluated previously. If the two channels do have

Table 3. Observation Well Data

Observation Well	*Location*	*Depth of bottom of 5-foot Screen (ft below ls)*	*Aquifer screened*	*Measuring point elevation (ft above msl)*	*Bedrock elevation (ft above msl)*
Site 1	T24NR01W09.4f	352.0	B	750.7	384
Site A	T21NR01W05.4g	317.0	B	677.6	355
Site B	T22NR01W02.8a	297.0	B	675.9	362
Site C	T23NR01W26.1c	289.5	B	667.3	366
Site 2a (east)	T23NR02W10.5a	267.0	B	645.3	361
Site 2b (east)	T23NR02W10.5a	92.0	G	645.1	---
Site 3a (west)	T22NR02W24.7g	252.0	B	619.0	358
Site 3b (west)	T22NR02W24.7g	57.0	G	619.4	---
Site 5	T23NR03W26.6f	242.0	B	642.1	389
Site 6	T22NR04W12.1f	237.0	B	637.8	358
Site 7	T22NR03W26.2a	339.5	B	700.2	348
Site 8	T21NR03W21.4e	217.0	B	584.7	357
Site 10	T21NR04W14.8a	187.0	B	557.3	356
Site E	T20NR02W22.1h	177.0	B	608.1	410
Site Ga (north)	T20NR04W03.8f	172.0	B	560.8	347
Site Gb (south)	T20NR04W03.8f	57.0	G	561.1	---

Note: The aquifer was screened by Banner Formation (B) or Glasford Formation (G).

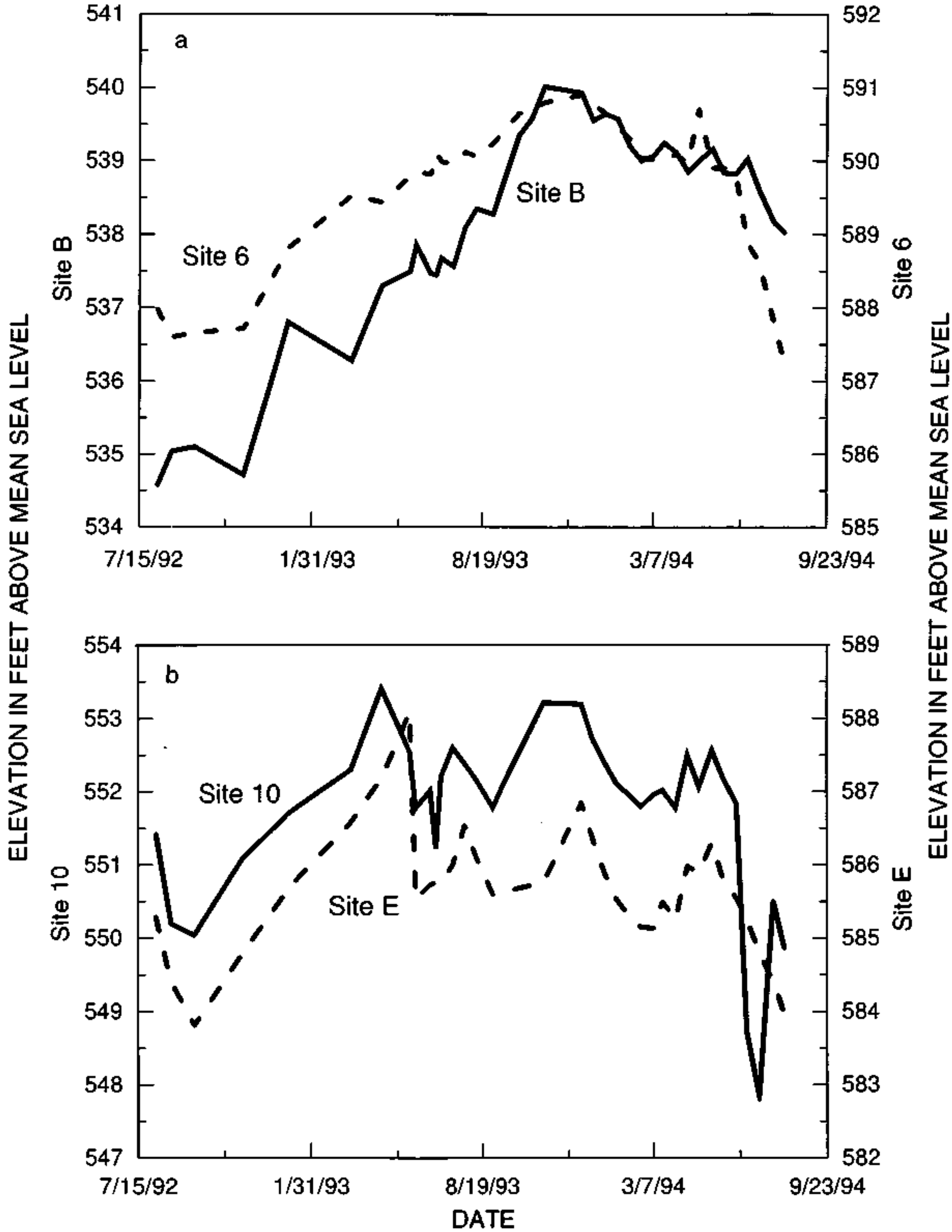

Figure 11. Observation well water-level hydrographs at a) Site B and Site 6 and b) Site 10 and Site E

separate flow regimes, ground-water flow from the Kenney Valley appears to be much more significant to the confluence area than was previously believed and it appears to influence water levels in the western part of the study area.

At three of the well sites, a second well was installed in an upper aquifer, when the upper aquifer was at least 20 feet thick. Figure 12 compares water levels for the wells at Site 2: the deep well is finished in the Sankoty-Mahomet aquifer, the shallow well in the upper Glasford aquifer. Hydrographs for the two aquifers indicate that they are not connected at this location since the water levels change independently. Site G also has two wells; however, the water-level trends parallel each other and remain at about the same elevation. Although the cross section (figure 7e) does not show a lithologic connection, the water-level data suggest that the two aquifers may be hydraulically connected.

In May 1993, water levels were measured in 80 private and ISWS wells throughout the study area. Such a "mass measurement" provides a picture of the potentiometric surface at a particular time (in this case, in the spring when water levels are typically at their highest). The results of this mass measurement are presented as a potentiometric surface map (figure 13).

Several important characteristics of the ground-water flow system are evident from figure 13, specifically the ground-water flow direction and the flow gradient. Ground water flows through the aquifer, toward lower water-level elevations to the west and north. The average gradient in these directions is approximately 2.5 feet per mile. Historically, it has been assumed that ground water flows into the confluence area from the north to the southwest in the Mackinaw Valley, and from east to west in the Mahomet Valley. From the confluence area, it then flows west where it discharges into the Illinois River or one of its tributaries. McComas (1969) was the

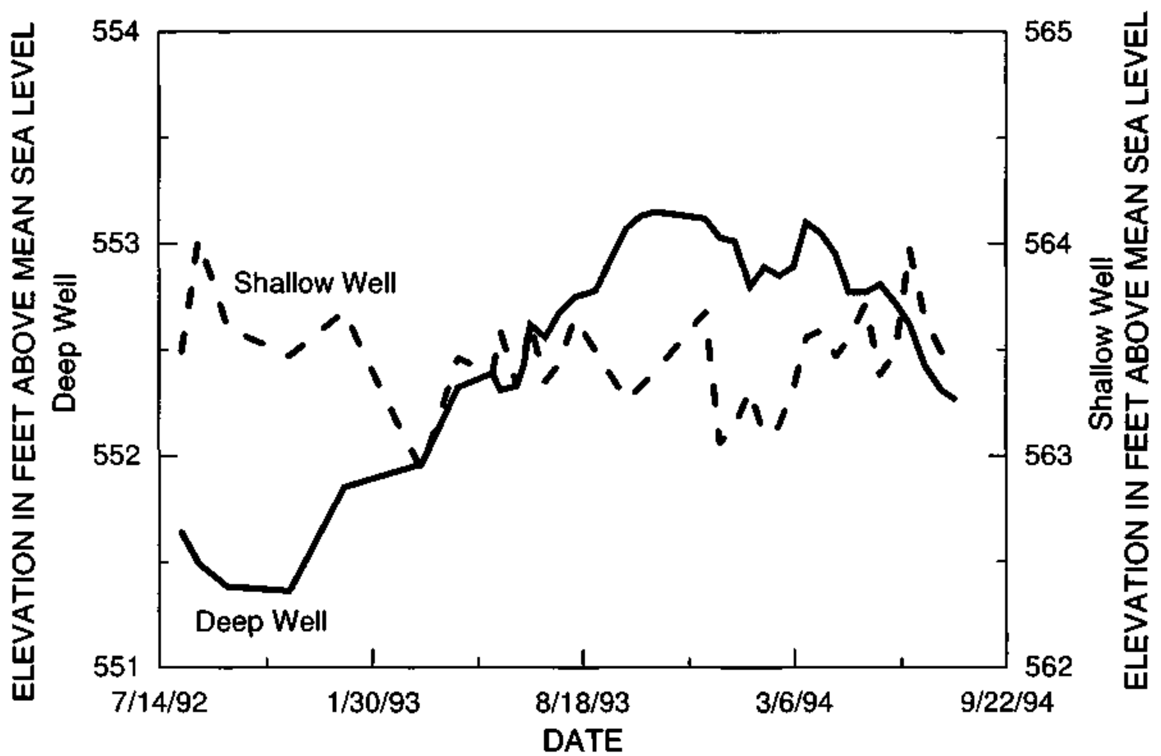

Figure 12. Observation well water-level hydrographs at Site 2

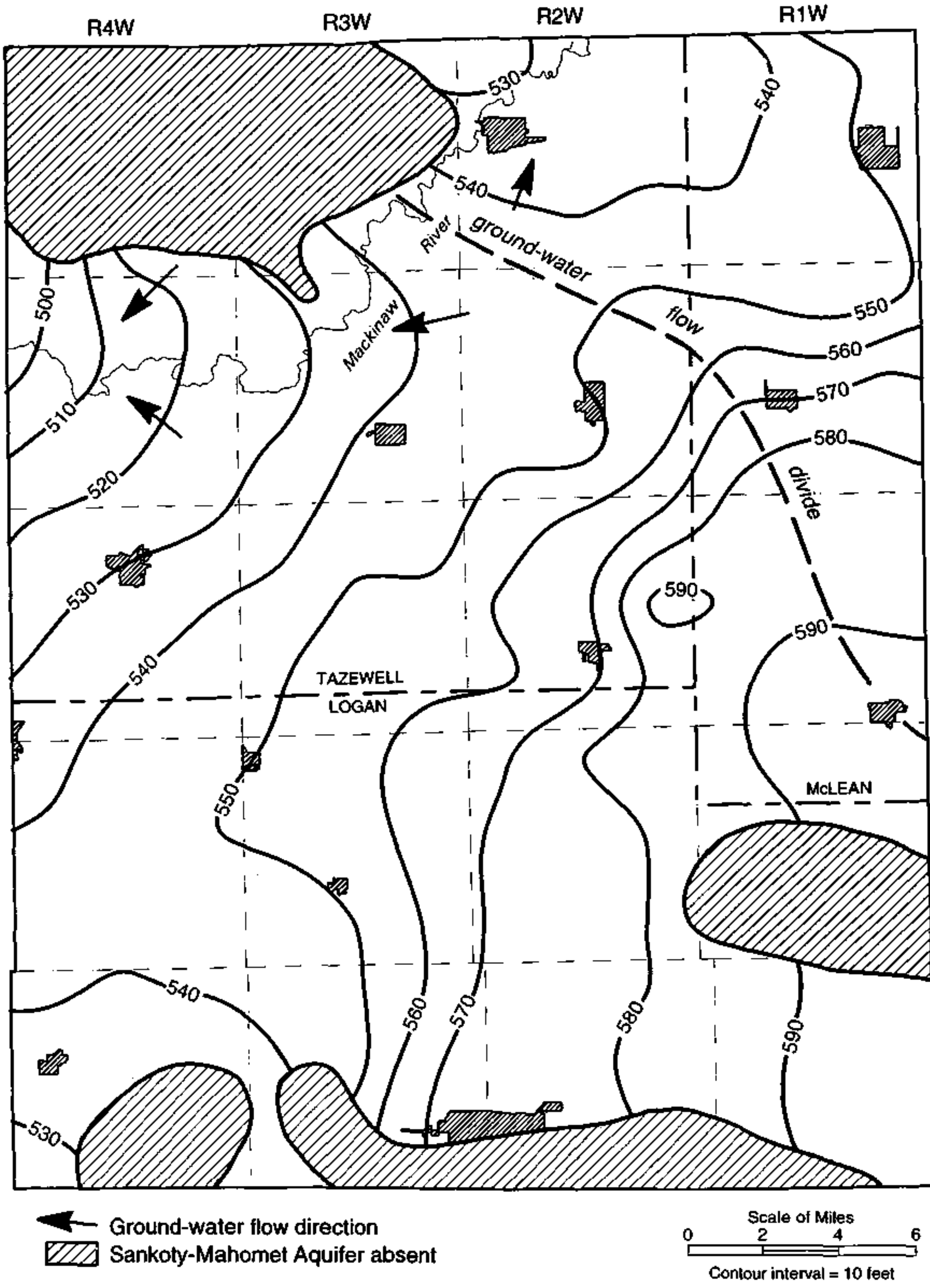

Figure 13. Potentiometric surface of the Sankoty-Mahomet aquifer system during spring 1993

first to suggest a ground-water divide in the Mackinaw Valley flow system, separating the Sankoty aquifer in the Mackinaw Valley into north and southwest components. Figure 13 indicates that the flow in the Mackinaw Valley is directed north and west out of the confluence area. This suggests the Mahomet portion actually provides the majority of the horizontal ground-water recharge for the region and contributes to a portion of the Mackinaw Valley north of the study area.

It was stated earlier in this report that water-table conditions may exist locally in the study area. Figure 13 suggests that such an area does exist near the Mackinaw River in Township 23 North, Range 04 West. As shown in the figure, the water-level contours bend around the river. Typically, this occurs when the flow paths in the aquifer are toward the river. Thus, the ground water is flowing toward and potentially discharging into the Mackinaw River. Observed water levels in the aquifer equal or exceed the bed elevation of the river, which indicates lateral movement of the ground water, with the river being the likely point of discharge.

Ground-water levels are typically cyclic (highest in the spring and lowest in the fall), but this cycle did not occur in 1993. Near-record rainfall provided additional recharge that allowed water levels in the aquifer to continue rising during the summer 1993. This is significant for two reasons. First, water levels were as much as 5 feet higher in fall 1993 than during fall 1992, representing a significant amount of additional water in storage over an area of several hundred square miles. Secondly, and more meaningful, the aquifer reacted to surface stresses quickly. Because the aquifer is overlain by as much as a hundred feet of till, a delay was expected in water-level response to the rainfall and flooding. Because aquifer responce was much more rapid than expected, it is obvious that the aquifer has a significant recharge component that has not been identified yet.

Ground-Water Discharge to Streams

Ground-water seepage into streams and rivers in the form of baseflow can account for large percentages of an aquifer's total discharge. Streamflow data can be used to estimate the amount of water discharged from an aquifer system into a stream system.

Streamflow consists of direct runoff and baseflow. Direct runoff is that portion of a rainfall event that enters a stream channel through surface runoff and interflow. Direct runoff is present in stream hydrographs for short periods of time following a runoff-producing rainfall event. Baseflow is that portion of streamflow produced by ground-water seepage into a stream. Baseflow is present in streamflow year-round, and during most low-flow events the entire streamflow consists of baseflow.

The Mackinaw River is a major surface water tributary in the study area. To determine the extent, if any, of the Mackinaw River's influence on the aquifer system in the study area, baseflow analyses were performed on streamflow data from two USGS gaging stations, one just upstream and one just downstream of the study area (figure 3). The upstream point was the USGS station at Congerville, IL, and the downstream point was the USGS station at Green Valley, IL. The Congerville station has a drainage area of 767 square miles, the Green Valley station, 1,073 square miles. Between the two stations, the watershed area is almost completely enclosed within the study area. Therefore, if the baseflow at the two stations is compared, significant changes in the amount of baseflow can be attributed to ground-water seepage within the study area.

The data used for the baseflow analysis came from USGS streamflow data for the two gaging stations as presented in the USGS publications for the Water Years 1990, 1991, and 1992. A water year begins on October 1 and ends on September 30 of the following calendar year.

The method of baseflow separation employed in this analysis was the fixed base-length method, which assumes that direct runoff ends at a fixed time (T) after the hydrograph peak for a rainfall event. A line is drawn on the hydrograph to project the baseflow decline of the previous peak to a point directly beneath the next hydrograph peak (line A-B in figure 14). Line B-C connects this point to a point on the hydrograph at distance T from the hydrograph peak. The T value is determined from the following formula (Cravens et al., 1990; after Linsley et al., 1958):

$$T = A^{0.2}$$

where A is the drainage basin area in square miles. The value of T was calculated to be 3.78 days for the Congerville station and 4.04 days at the Green Valley station.

Using the fixed base-length method of baseflow separation, daily baseflow values were obtained for the two gaging stations for a three-year period. In order to compare the values from the two gages, yearly baseflow averages were produced, as well as total baseflow averages for the three-year period. To analyze any influences occurring between the two stations, the amount of baseflow per square mile of drainage basin area was calculated by dividing the averaged baseflow values by the station's respective drainage area. The streamflow hydrograph and its baseflow component for the Congerville station and the Green Valley station are shown (figures 15a and 15b).

Consistent trends in both streamflow and corresponding baseflow between the two stations are expected since the two stations are on the same river and in such close proximity. Table 4 summarizes the findings of the analyses. The average yearly baseflows in cubic feet per second per square mile (cfs/mi^2) and the three-year average baseflows are presented. The Congerville station had an average baseflow of 0.42 cfs/mi^2, while the Green Valley station had an average baseflow of 0.50 cfs/mi^2.

These results suggest that the amount of baseflow from ground-water seepage increases 19 percent along the Mackinaw River within the study area. Exactly which aquifer or aquifers are contributing which amounts cannot be determined by this analysis. It does indicate, however, that some ground-water discharge is occurring.

It is important to note that this baseflow separation analysis is an approximation of ground-water discharge. This technique is a qualitative approach that indicates that ground-water discharge is occurring in the study area.

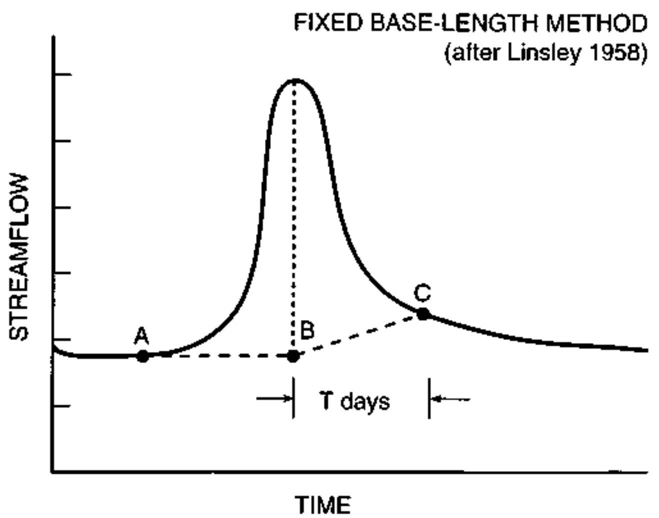

Figure 14. Fixed base-length method of baseflow separation

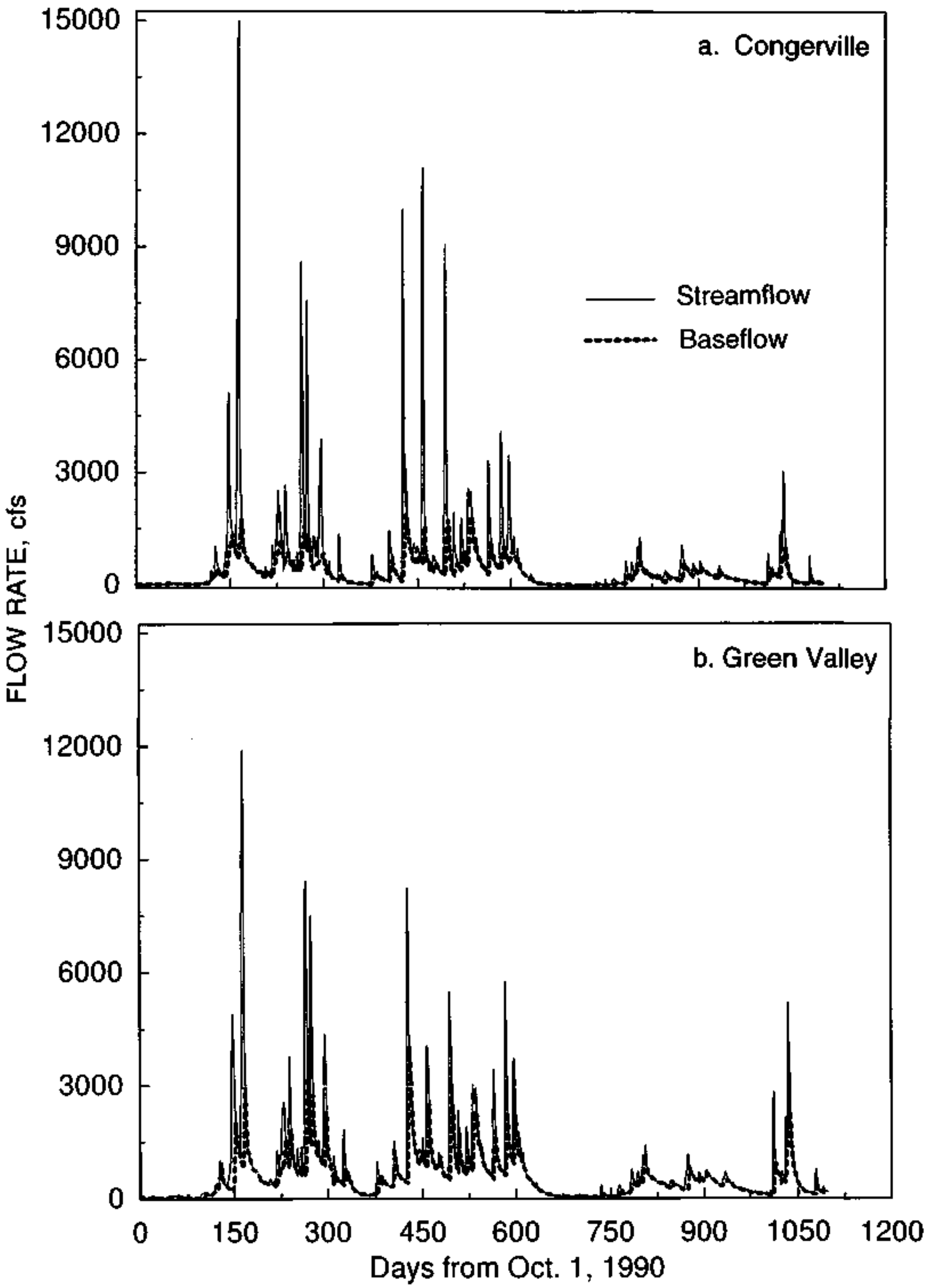

Figure 15. Streamflow and baseflow hydrographs for the a) Congerville and b) Green Valley gaging stations

Table 4. Estimated Mean Baseflow from Hydrograph Separation (cfs/mi^2)

	Water Years			
USGS gaging station	*1990*	*1991*	*1992*	*1990-92*
Congerville	0.45	0.55	0.25	0.42
Green Valley	0.50	0.66	0.32	0.50

Ground-Water Recharge

One of the most important factors in determining aquifer yield is recharge, which is also the most difficult component to determine. Recharge occurs from vertical movement of water through the upper geologic formations to the aquifer and from induced horizontal movement into the aquifer. Rainfall runs off into streams, it is discharged to the atmosphere by evapotranspiration, or it infiltrates to the water table (top of zone of saturation). The portion that infiltrates either runs off in a shallow flow system to nearby streams as interflow or moves downward into the deeper formations—in our case, to the Glasford aquifers, the Sankoty-Mahomet aquifer, or both. Recharge typically occurs due to infiltration, but it could also come from streams hydraulically connected to the aquifer. The rate of vertical movement depends on the head differential between the source bed and the deep aquifer, as well as the vertical hydraulic conductivity of the intervening aquitard. Locally, vertical conductivity could vary by orders of magnitude in the aquitards due to the complexity of the glacial materials. A good general understanding of the geologic framework of the area is available, but glacial geology is often so locally variable that the makeup of the materials at a particular point is never certain without good geologic control. Therefore, in some areas where geologic data are limited, knowing the size, extent, and water levels in the shallow aquifers is difficult but essential to developing an accurate estimate of areal recharge. The drilling program has shown that shallow aquifers exist in the study area, but not over large regions. Historically, it has been assumed that water reaching the aquifer as recharge is balanced by ground-water discharge to streams (i.e., baseflow) and withdrawals from pumpage.

Kempton and Visocky (1992) estimated recharge based on the work of Walton (1965), who concluded that recharge, representing runoff diverted to cones of depression in the Champaign-Urbana area, was equal to about 50 percent of ground-water runoff. For our study area, Walton (1965) classified two areas: one with an estimated runoff between 129,000 and 194,000 gallons per day per square mile (gpd/mi^2), another with an estimated runoff between 194,000 and 259,000 gpd/mi^2. Assuming an average of 194,000 gpd/mi^2, total ground-water runoff in the study area is about 140 mgd. If we assume that 50 percent of the runoff could be diverted into pumping centers, recharge is estimated to be 70 mgd.

The hydrochemistry of the ground water in the aquifer may provide clues to understanding the flow and recharge in the aquifer system. Panno et al. (1994) characterized the water chemistry in the Sankoty-Mahomet aquifer and the confluence area. They determined that the ground water in the Sankoty-Mahomet aquifer in the confluence area is most similar to the ground water in the Glasford Formation Sands. They concluded that 17 percent of the flow into the confluence area was from the Mahomet aquifer. The other 83 percent came from the Sankoty aquifer (north of the study area) and vertical recharge.

Ground-Water Quality

Water quality has become an important consideration when evaluating an existing or planned water system. Small communities are extremely vulnerable to changes in regulations dealing with ground-water quality, since the U.S. Environmental Protection Agency (USEPA) routinely reviews and modifies drinking water standards. Consequently, the ISWS maintains a water quality database forthe entire state. Table 5 summarizes information from

66 private and municipal wells finished in the Sankoty-Mahomet aquifer in the study area. Kaiser (1994) reviewed the chemical data and had these comments, "In general, the available water quality data indicate the water is hard and contains enough iron to stain porcelain and laundry. All of the sulfate values were below drinking water standards. However, many of the values for Total Dissolved Solids (TDS) exceeded the taste threshold of 500 milligrams per liter (mg/L). Some of the lead values exceeded the lead action level of 15 micrograms per liter (ug/L), but this is likely due to contact with metal plumbing. More than half of the water samples also contained arsenic in excess of the proposed standard set to go into effect in January 1996." The current standard for arsenic is 50 ug/L. The new standard will be 2-10 ug/L.

The study area has a history of high arsenic levels that are thought to be naturally occurring. Holm and Curtiss (1988) found arsenic levels near or above 50 ug/L at 8 of 20 private wells sampled in Tazewell County. The lowest value was 9 ug/L and the highest value was 226 ug/L, with 17 of the 20 samples having arsenic levels greater than 20 ug/L.

As part of the Safe Drinking Water Act, many other water quality regulations are set to be issued in June 1995 (Prendergast, 1993). Exactly what these new, stricter, regulations will mean is unclear, but many small communities may find that they are unable to provide acceptable drinking water. *Ultimately, ground-water quality regulations rather than ground-water availability* may determine the extent of ground-water development of the Sankoty-Mahomet aquifer in the confluence area of the Mahomet and Mackinaw Bedrock Valleys.

Table 5. Summary of the General Ground-Water Chemistry of the Sankoty-Mahomet Aquifer in the Study Area

Constituent	*Number of cases*	*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

Notes: Constituent concentrations are in mg/L; constituents marked with an "*" are measured in ug/L; and hardness and alkalinity are measured as $CaC0_3$. Results of 0.0 indicate that concentration was below method detection limit.

DISCUSSION

Prior to this research, the authors' conceptual model of the ground-water flow system was relatively simplistic. Horizontal movement of ground water was believed to be southwest through the Sankoty aquifer in the Mackinaw Valley and west through the Mahomet aquifer in the Mahomet Valley into the confluence area. From there, the flow was westerly into Mason County, eventually discharging into the Illinois River or one of its tributaries. The role of the Kenney Valley (figure 2) was presumed to be minimal. The Sankoty-Mahomet aquifer was thought to be situated directly atop the bedrock and bounded in extent by bedrock highs, with the top of the aquifer at an elevation near 500 ft-msl. Recharge was assumed to occur as vertical leakage at about 194,000 gpd/mi^2 (Walton, 1965; Kempton and Visocky, 1992).

Significance of Drilling Program

The importance of the test drilling phase of this project cannot be overemphasized because it was an essential first step toward accurately estimating the availability of the ground-water resource in the area. Drilling has revealed locally significant differences in the bedrock topography and consequently in thickness and extent of the Sankoty-Mahomet aquifer.

Information available prior to this study consisted principally of drillers' logs of water wells (primarily domestic and farm wells scattered throughout the area), engineering logs, and occasionally records from municipal wells and test holes. A few records for commercial structures and stratigraphic tests for gas storage or oil and coal exploration provided most of the information on the top of bedrock, largely because downhole geophysical logs were performed. Only a few water wells of record penetrated bedrock, and their distribution was poor, highly localized, or both.

The test drilling program, therefore, provided a significant amount of the information available in much of the area on the elevation of the top of bedrock and the characteristics of the lower part of the glacial sequence of deposits. Samples of rotary cutting were collected at 10-foot intervals. These data were supplemented by downhole natural gamma logs. Monitoring wells were established in the Sankoty-Mahomet aquifer and a significant shallower aquifer, providing regional controlled coverage for water-level measurements.

Revision of Bedrock Topography

While previous studies have provided maps of the elevation of the top of bedrock, they relied on few data points and were highly interpretive and generalized, except in a few local areas where wells had penetrated bedrock. It is not surprising therefore that locally significant changes in the interpretation of the bedrock surface configuration resulted from this study. The two most significant changes (figure 16) were the addition of a southwest-northeast elongated ridge with its highest area just above an elevation of 400 ft-msl at Hopedale, and the widening of the Mackinaw Valley northwest of Danvers by up to four miles eastward. The area of elevations of the bedrock surface below 350 feet-msl were modified and somewhat reduced, particularly near Hopedale in the areas of the bedrock ridge.

Subsequent drilling, currently being conducted for an ongoing ISWS/ISGS cooperative study mentioned earlier, and geophysical surveys in the area, will likely modify the current interpretation, just as the authors have modified previous interpretations. Each new map includes additional data and provides a more accurate depiction of the true shape of the bedrock surface. Previous studies (Kempton et al., 1982; Kempton et al., 1991; Kempton and Visocky, 1992) have also suggested a complicated history for the formation of the bedrock valley system and a variety of land forms within the valleys. Additional local features such as bedrock mounds, ridges, terraces, and overdeepened segments can be expected.

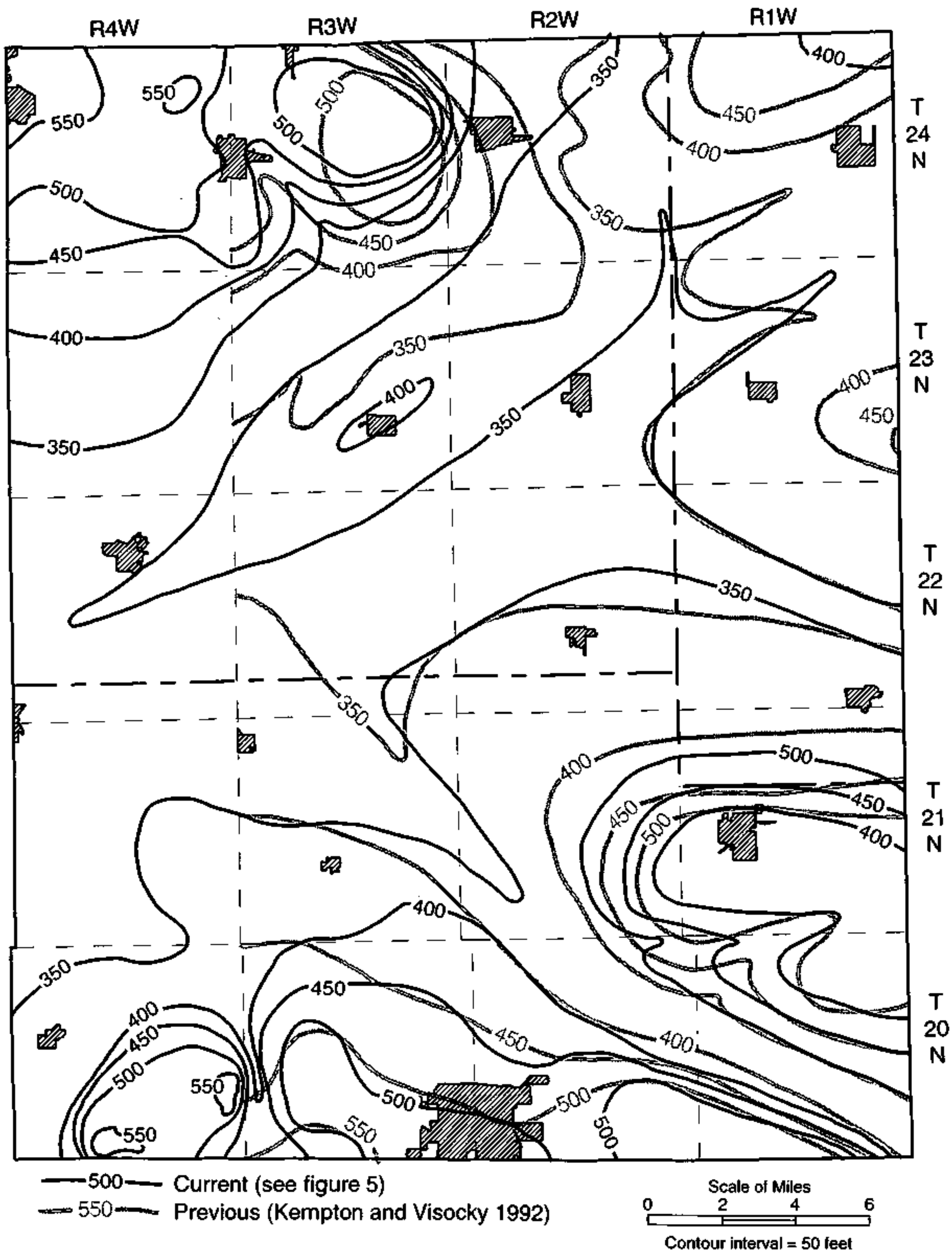

Figure 16. Combination of bedrock topography maps (from Kempton and Visocky, 1992, and figure 5)

Thickness and Extent of Aquifers

The revised bedrock topography map obviously changes the previously mapped extent and thickness of the Sankoty-Mahomet aquifer. New data and maps indicating a wider Mackinaw Valley northwest of Danvers provide the potential for the presence of several square miles of aquifer in areal extent not previously mapped (figures 8-10 and 17). On the other hand, the presence of a bedrock ridge in the vicinity of Hopedale restricts the interpreted thickness of the Sankoty-Mahomet aquifer (figures 9 and 10) in that area.

Just as significant are the data from several test holes, especially Site 4 near Armington (figure 3), that suggest fine-textured lacustrine sediments nearly 80 feet thick occur locally below the Sankoty-Mahomet aquifer (figures 7b and 7d). In Site 4, not only is thick lacustrine material present below the Sankoty-Mahomet, but a thick lacustrine clayey silt and sand replaces most of the upper portion of the coarser textured deposit, effectively leaving only 17 feet of aquifer at this site (figure 10).

Kempton and Visocky (1992) reported a similar situation in the Mahomet Bedrock Valley near the confluence with the Mackinaw Bedrock Valley. Their cross section A-A' (figure 11, page 23), between Bloomington and Atlanta, shows the coarser Mahomet Sand filling a trench apparently eroded into fine-textured silts and sands, now thought to be older than the Sankoty-Mahomet Sand. The older sediments remain on both sides of the younger channel (see also their figure 18, unit 4, page 37). The absence of thick coarse-textured sediments at the village of McLean fits this interpretation. Therefore, the occurrences of pre-Sankoty-Mahomet fine-textured sediments appear to represent some recurring uneroded patches of these older sediments, deposited as a blanket over the entire valley floor prior to deposition of the Sankoty-Mahomet Sand.

Richards and Visocky (1982) and Kempton and Visocky (1992) have reported a similar but even more complex area of sediments in and around the existing town of Normal West Well Field (figure 7b). Logs and sample descriptions from the numerous test holes, observation wells, and production wells in that area (south half of T.24N., R.1W., and north half of T.23N., R.1W.) show considerable variation in the sequence and distribution of Banner Formation sediments, including interbedded tills or till-like materials and lacustrine sediments. In addition the town of Normal test hole 1-88, drilled in Section 28, T.24N., R. 1W., encountered only clayey silts and sands at the expected position of the Sankoty-Mahomet Sand although the natural gamma log had the typical curve of that sand unit. Figure 18 of Kempton and Visocky (1992) therefore also included that area in their category depicting mainly fine-textured deposits.

Hydrological Appraisal

This new conceptual model of the ground-water flow system reveals important changes from earlier theory. Most notable are the potential importance of the flow from the Kenney Valley, the ground-water divide separating the confluence area from the Sankoty aquifer, the large role of vertical recharge, and the implications of these changes on overall ground-water availability in the confluence area.

As mentioned earlier, figure 11 points out distinct differences in flow paths for different regions of the study area. The distribution of the southern wells indicates that the flow from the aquifer in the Kenney Valley is potentially as important a portion of the horizontal flow into the confluence area as the other valley segments. In addition, it broadens the confluence area to the west.

The potentiometric surface map shows clearly that there is little, if any, contribution of the flow into the confluence area from the Mackinaw Valley, north of the study area. A ground-water divide exists; north of the divide the flow direction in the aquifer is northwesterly toward the Illinois River Valley. If the Sankoty portion does not

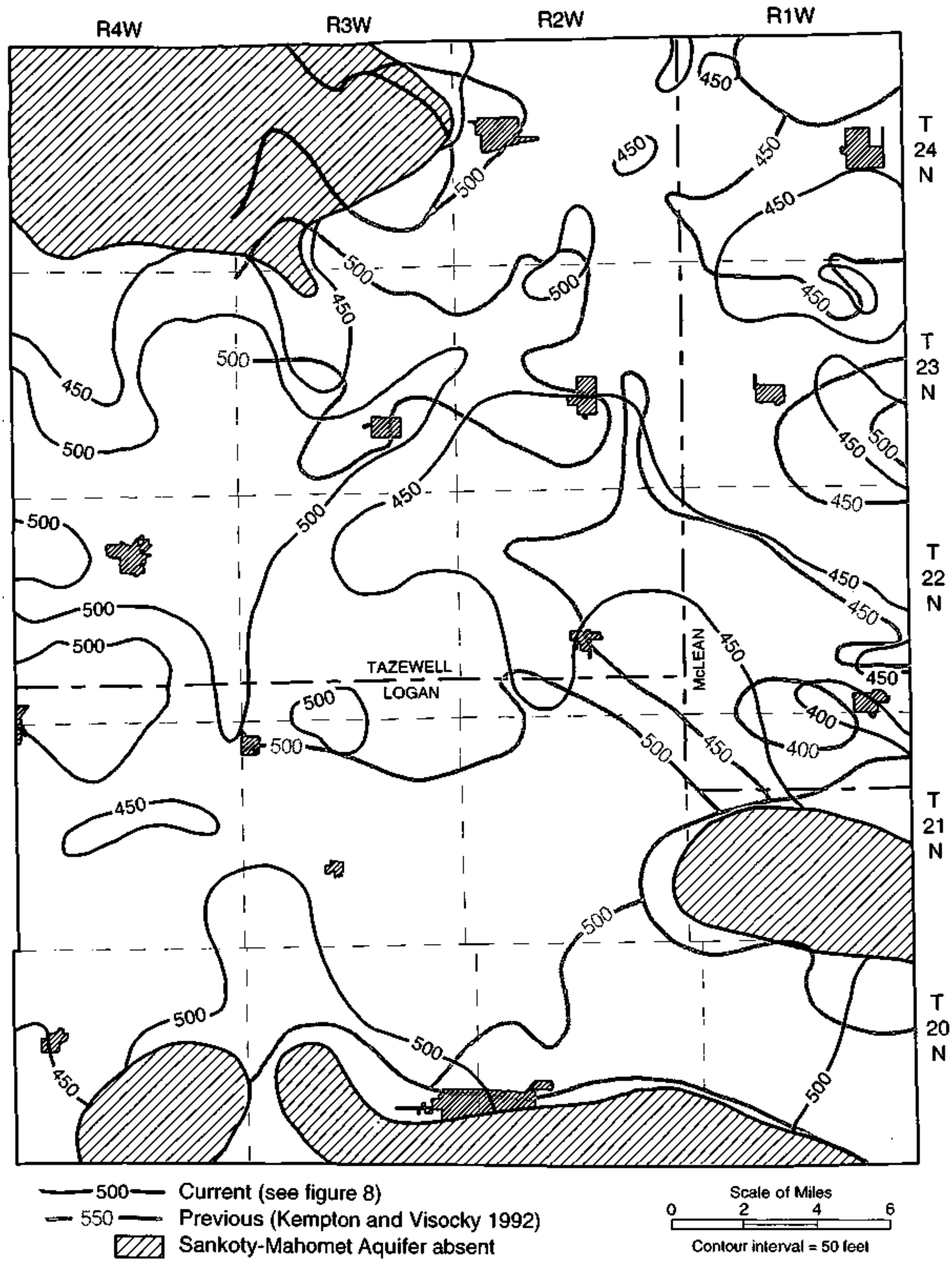

Figure 17. Combination of the elevation of the top of the Sankoty-Mahomet aquifer system (from Kempton and Visocky, 1992, and figure 8)

contribute to the flow in the confluence area, other sources must be significant: the Mahomet portion must contribute more flow, vertical recharge must be greater than was previously believed, or both.

There are several indications that vertical recharge in the study area is higher than previously believed, among them being the water-level response to rainfall during the flood of 1993. Observation well hydrographs in the Sankoty-Mahomet aquifer (appendix 2), showed a prompt response of water levels to the increased rainfall, with water levels rising throughout 1993. In summer 1994, water levels began to decline as expected. These results could signify increased recharge from the surface.

Work by Panno et al. (1994) tends to support this concept and found that the water chemistry of the confluence area was most similar to upper Glasford aquifers. Using water chemistry parameters, they concluded that 17 percent of the flow in the confluence area came from the Mahomet Valley and the rest came from the Mackinaw Valley and from vertical recharge. Since we have determined that the Mackinaw portion provides little, if any, flow to the confluence area, the statements from Panno et al. (1994) infer that vertical recharge provides the other 83 percent of flow to the confluence area. Panno et al. also list data from Hackley et al. (in press) whose data suggest that ground water in the confluence area is younger than any of the ground water found east of the study area in the Mahomet Valley. This observation implies significant vertical recharge.

Revised Conceptual Model

The new information gathered in this study led us to an updated conceptual model for the confluence area. Our new model has vertical leakage as the major flow component. Horizontal flow in the confluence area is entirely from the Kenney Valley and Mahomet Valley portions of the system. Horizontal flow in the Mackinaw portion is northerly away from the confluence area. The aquifer system is underlain by bedrock, lacustrine deposits, or fine-grained pre-Sankoty sediments. The unconsolidated fine-grained deposits between the aquifer and the bedrock restrict the size and extent of portions of the Sankoty-Mahomet aquifer, and also limit vertical movement of ground water from the bedrock to the aquifer. The system is hydraulically connected to the Glasford Sands in some western portions of the study area. Seldom does the top of the aquifer rise to an elevation of 500 ft-msl except in the Kenney Valley and some western parts of the study area where the Glasford Sands directly overlie the Sankoty-Mahomet Sand.

This new conceptual model changes the authors' presumptions about the confluence area and ground-water availability. The model is no longer an enormous bowl of aquifer material greater than 100 feet thick; its size and extent have been reduced in the eastern portions of the study area.

Kempton and Visocky (1992) estimated the aquifer's potential yield to be 70 to 75 mgd. Using the same method, the authors have estimated recharge to be 70 mgd for the study area; however, these estimates are likely too high. The map initially used to create the recharge areas (Walton, 1965) is now outdated and does not reflect any geologic changes in configuration that have been discovered since. The fundamental observation is that portions of the study area are significantly different. Consequently, in some areas, such as near Site 4, any large stress on the system could cause large drawdowns over a significant area, whereas near Site 6, the same stress would cause much less drawdown due to the increased ground-water availability in that portion of the study area.

Future Development

A separate, ongoing study of the confluence area (Wilson and Herzog, 1994) should provide additional insight into what areas, if any, may be suited for future development of large ground-water supplies. That study, sponsored by the LRWPSC and being conducted jointly by the ISWS and ISGS, is due to be completed in Summer

1996. One goal of that study is to suggest what locations, if any, could support a 15 mgd well field, while minimizing drawdowns and impacts to nearby existing wells.

The Sankoty-Mahomet aquifer system is extensive, but very complex. It remains the most promising source for ground-water development for central Illinois. This is true not only for municipal growth but also for agricultural growth. Irrigation use continues to grow rapidly throughout the state, making the Sankoty-Mahomet aquifer a very promising, available supply of ground water. Competition for ground water is an emerging issue that is already significant in other parts of Illinois where conflicting interests vie for a limited resource.

GLOSSARY OF TERMS

Ablation—All processes by which snow and ice are lost from a glacier. These processes include melting, evaporation (sublimation), wind erosion, and calving.

Aquifer—A saturated body of earth materials that will provide a generally sustainable yield of suitable quantities of ground water.

Aquifer Test—A controlled scientific field experiment to determine hydraulic properties of an aquifer using a single pumped well with a constant pumping rate and at least one observation well.

Aquitard—A saturated geologic unit that may transmit ground water, but not in sufficient quantities to permit economic development. A source for leakage to underlying or overlying aquifers. Aquitards are significant in the study of regional ground-water flow.

Artesian Aquifer—An aquifer in which the water level in a well is above the top of the aquifer. The terms *leaky* and *nonleaky* are often used to describe artesian conditions. Leaky refers to the situation where significant ground-water flow through the aquitard occurs. Nonleaky conditions refer to the circumstances where flow through the aquitard is insignificant.

Cone of Depression—A conical depression in a water table (unconfined aquifer) or a piezometric surface (artesian confined aquifers), created by pumping a well.

Confined Aquifer—An aquifer that is confined between two units of material with significantly lower hydraulic conductivity (i.e., aquitards or aquicludes).

Confining Bed—An aquitard or aquiclude that is contiguous to a confined aquifer.

Deposition—The laying or placing of any material; specifically, the constructive process of accumulation into beds, veins, or irregular masses.

Dewatering—Physical process of evacuating water from a water-table aquifer or the lowering of ground-water levels below the top of a confined aquifer being pumped.

Discharge Area—Point where water flowing through an aquifer leaves the aquifer (well, lake, river, etc.).

Drawdown—Difference between the nonpumping or static water level and the pumping or dynamic water level.

Geohydrologic Unit (Hydrostratigraphic Unit)—An aquifer, a confining unit (aquitard), or a combination of aquifers and confining units comprising "a framework for a reasonably distinct hydraulic system."

Geophysical Exploration—Use of geophysical techniques, e.g., electric, gravitational, magnetic, seismic, or radioactive, to gather information on the physical properties of the earth.

Ground Water—Water in the zone of saturation.

Ground-Water Divide—A ridge in the water table or other potentiometric surface from which the ground water represented by that surface moves away in both directions.

Homogeneity—An aquifer is *homogeneous* if the magnitude of all significant properties are independent of position in the aquifer; synonymous with uniformity.

Hydraulic Conductivity—Capacity of water-bearing material to transmit water, measured by the quantity of water passing through a unit cross section in a unit time under a unit hydraulic gradient (gpd/ft^2).

Hydrograph—A graph showing stage, flow, velocity, or other characteristics of water with respect to time. A stream hydrograph commonly shows rate of flow; a ground-water hydrograph, water level or head.

Hydrostratigraphic Unit—A body of earth materials having considerable lateral extent and composing "a geologic framework for a reasonably distinct hydrologic system." See Geohydrologic Unit.

Lacustrine—Pertaining to, produced by, or formed in a lake or lakes; e.g., "lacustrine silts" deposited on the bottom of a lake. Fine-textured water-laid deposits.

Lithification—Conversion of a newly deposited, unconsolidated sediment into a coherent, solid rock, involving processes such as cementation, compaction, desiccation, and crystalization.

Model—A representation of a physical system. Ground-water models may be physical, electric analog, or mathematical.

Observation Well—A general term indicating any well used to measure water levels or obtain water samples.

Permeability—Capacity of a porous rock, sediment, or soil for transmitting a fluid; a measure of the relative ease of fluid flow under unequal pressure.

Potable—Safe and drinkable for human use.

Potentiometric Surface—An imaginary surface representing the total head of ground water and defined by the level to which water will rise in a well.

Pumping Level—Water level in a well being pumped at which the observed water level has little or no change within a reasonable period of time.

Radius of Influence—Distance from the discharge well to the edge of the cone of depression.

Recharge Area—Localized or regional area where water enters an aquifer.

Specific Capacity—Rate of discharge of water from a well divided by the drawdown of water level within the well; varies with duration of discharge, which should be stated if known. If the specific capacity is constant except for the time variation, it is roughly proportional to the transmissivity of the aquifer. The relationship between discharge and drawdown is affected by the construction of the well, its development, the character of the screen or casing perforation, and the velocity and length of flow up the casing.

Specific Yield—Ratio of (1) the volume of water that the rock or soil, after being saturated, will yield by gravity to (2) the volume of rock or soil. The definition implies that gravity drainage is complete. In the natural environment, specific yield is generally observed as the change that occurs in the amount of water in storage per unit area of unconfined aquifer as the result of a unit change in head. Such a change in storage is produced by the draining or filling of pore space and is therefore dependent upon particle size, rate of change of the water table, time, and other variables. Hence, specific yield is only an approximate measure of the relationship between storage and head in unconfined aquifers. It is equal to porosity minus specific retention (water held by soil particles).

Static Water Level—Water level in a nonpumping well outside the area of influence of any pumping well. This level registers one point on the water table in a water table well or one point on the piezometric surface in a well intersecting an artesian aquifer.

Storage Coefficient—Volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. In a confined system, water derived from storage with decline in head comes from the expansion of the water and compression of the aquifer; similarly, water added to storage with a rise in head is accommodated partly by compression of the water and partly by expansion of the aquifer. In an unconfined system, the amount of water derived from or added to the aquifer by these processes generally is negligible compared to that involved in gravity drainage or filling or pores; hence, in an unconfined system the storage coefficient is virtually equal to the *Specific Yield.*

Stratigraphic Unit—A body of adjacent geologic materials recognized as a unit in the classification of rock sediments.

Thalweg—Line connecting the lowest or deepest points along a streambed or valley.

Till—Dominantly unsorted and unstratified drift deposited directly by and underneath a glacier without subsequent reworking by meltwater, and consisting of a heterogeneous mixture of clay, silt, sand, gravel, and boulders ranging widely in size and shape.

Transmissivity—Formerly called the "coefficient of transmissibility;" rate at which water flows through a unit width of the aquifer (perpendicular to flow) under a unit hydraulic gradient; product of hydraulic conductivity and saturated thickness.

Unconfined Aquifer—Also called a water-table aquifer; an aquifer in which the water table forms the upper boundary.

Water Table—That surface in an unconfined water body at which the pressure is atmospheric; defined by levels at which water stands in wells that penetrate the water body just far enough to hold standing water.

Well Loss—Component of drawdown in a well due to frictional losses from turbulent flow as water passes through the screen or well face and inside the casing to the pump intake.

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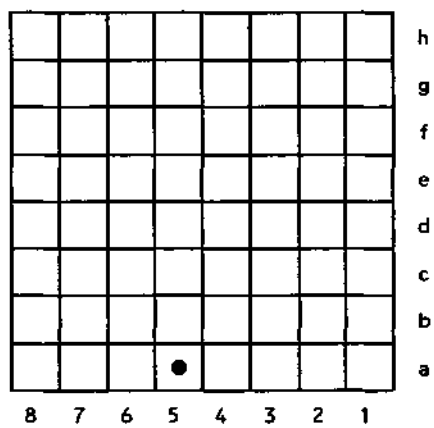
APPENDIX 1. WELL NUMBERING SYSTEM

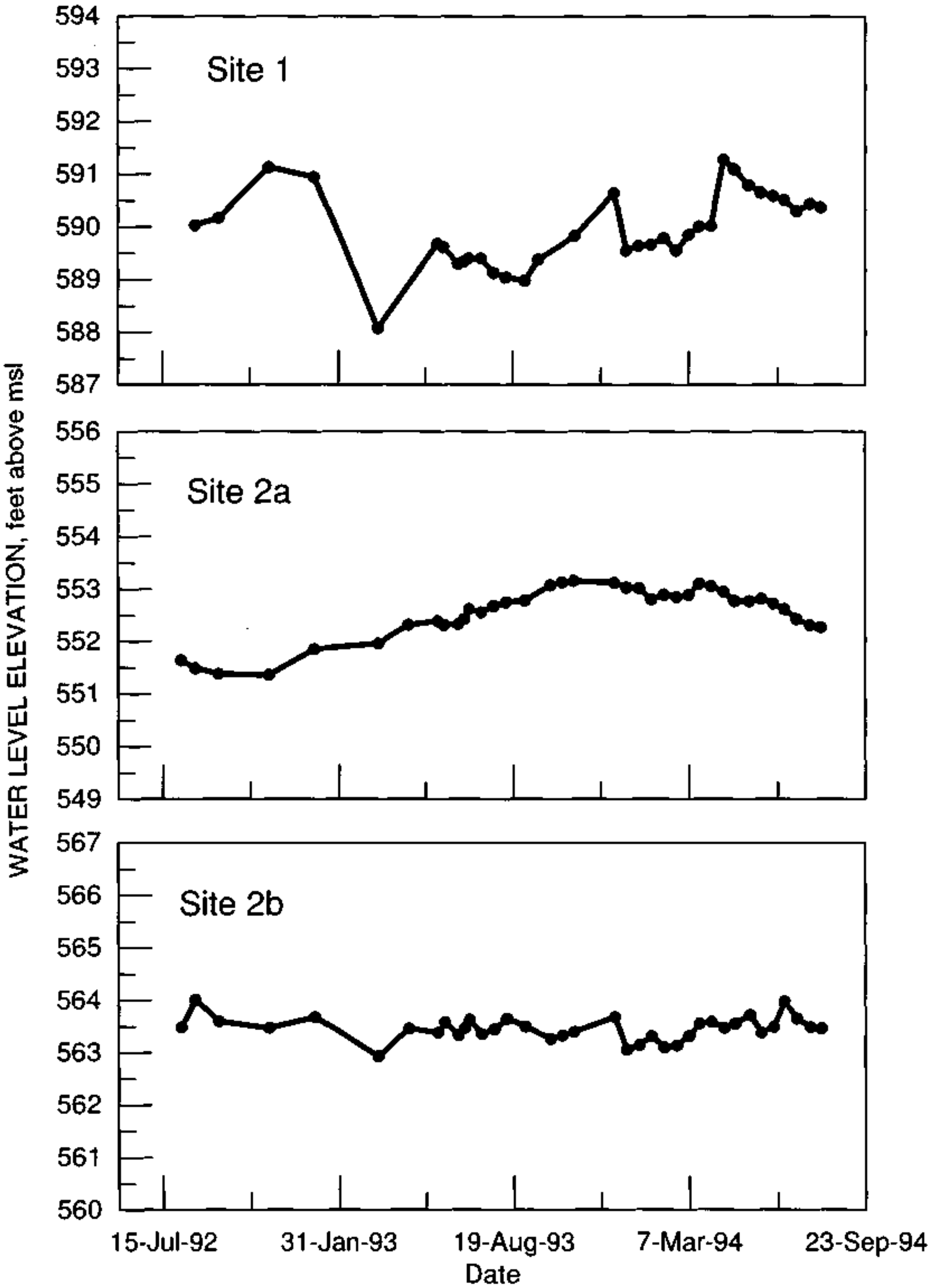
The well numbering system used in this report is based on the location of the well and uses the township, range, and section for identification. The well number may consist of up to five parts: county abbreviation, township, range, seciton, and coordinates within the section. Each 1/8-mile square contains ten acres and corresponds to a quarter of a quarter of a quarter section. A normal section of one square mile contains eight rows of 1/8-mile squares; an odd-sized section contains more or fewer rows. Rows are numbered from east to west and lettered from south to north as shown in the diagram.

Tazewell County
Township 23N, Range 02W
Section 10



The location of the well above is 23NN02W10.5a. This is Site 2a (TAZ-92A).

APPENDIX 2. OBSERVATION WELL HYDROGRAPHS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593
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590
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588
587
Site 1
556
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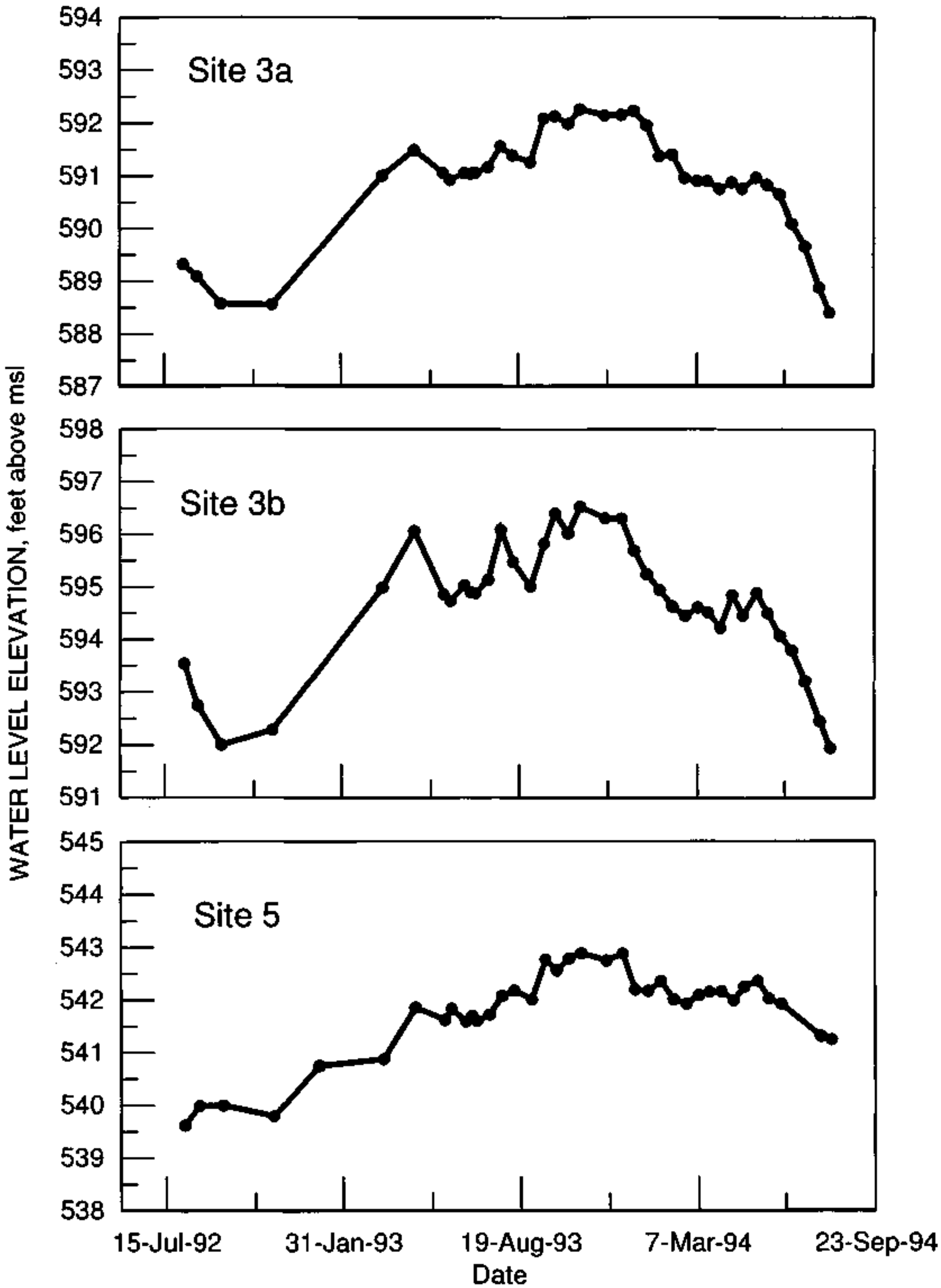
554
553
552
551
550
549
Site 2a
567
566
565
564
563
562
561
560
Site 2b
WATER LEVEL ELEVATION, feet above msl
15-Jul-92
31-Jan-93
19-Aug-93
7-Mar-94
23-Sep-94
Date


Site 3a
Site 3b
Site 5
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593
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541
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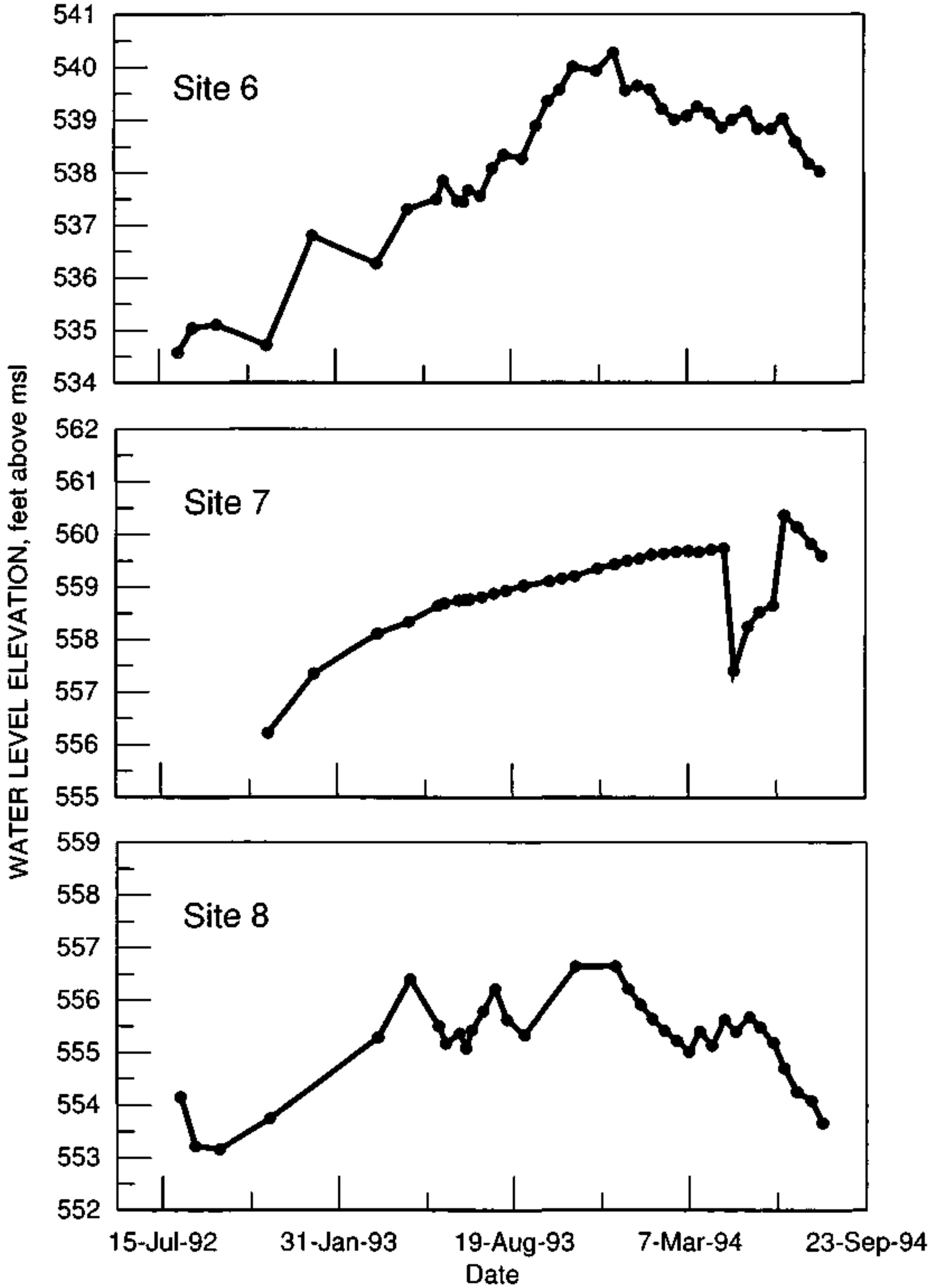
WATER LEVEL ELEVATION, feet above msl
15-Jul-92
31-Jan-93
19-Aug-93
7-Mar-94
23-Sep-94
Date


WATER LEVEL ELEVATION, feet above msl
Site 6
541
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539
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535
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Site 7
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558
557
556
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Site 8
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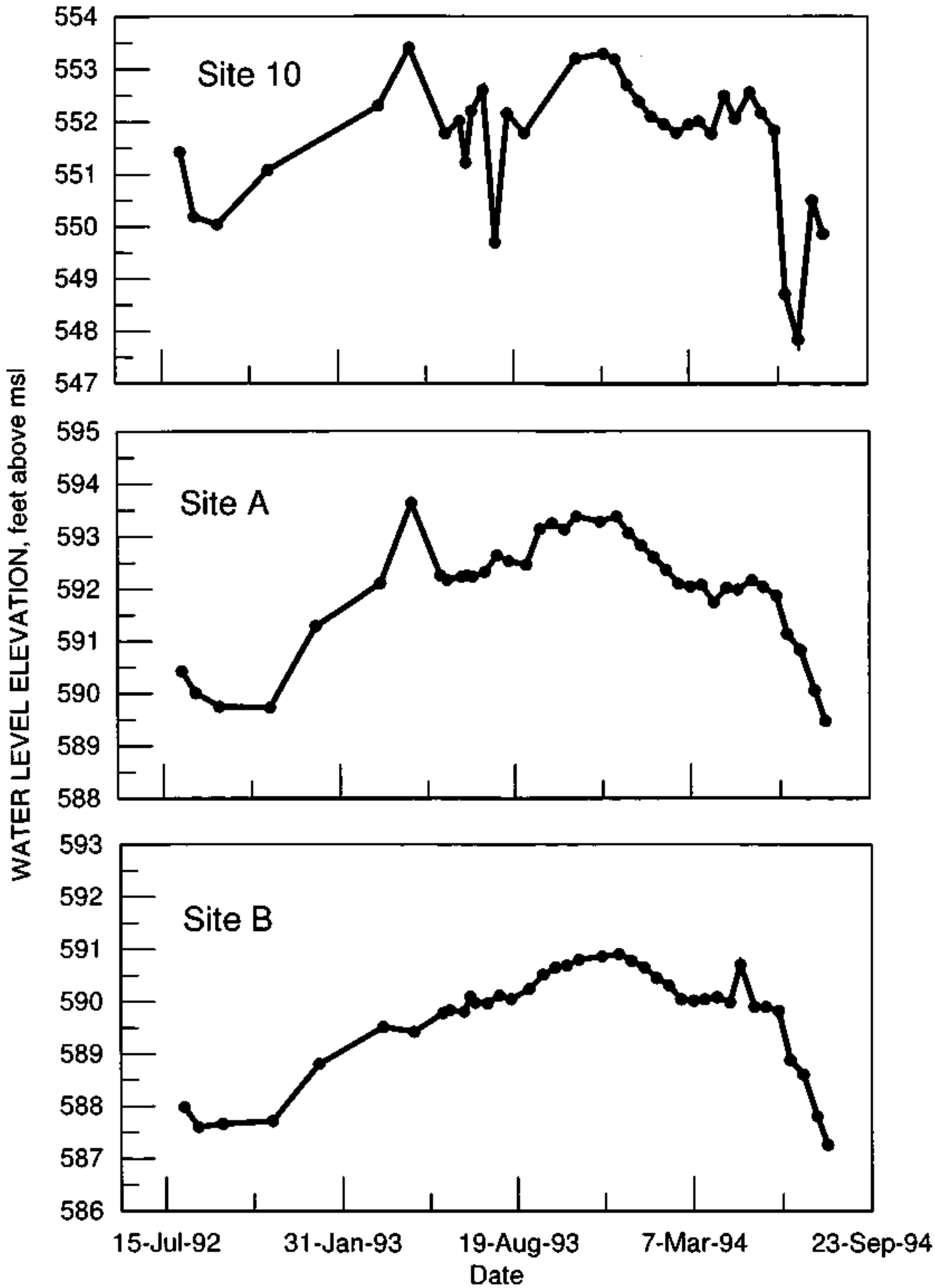
15-Jul-92
31-Jan-93
19-Aug-93
7-Mar-94
23-Sep-94
Date


WATER LEVEL ELEVATION, feet above msl
Site 10
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Site A
595
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592
591
590
589
588
Site B
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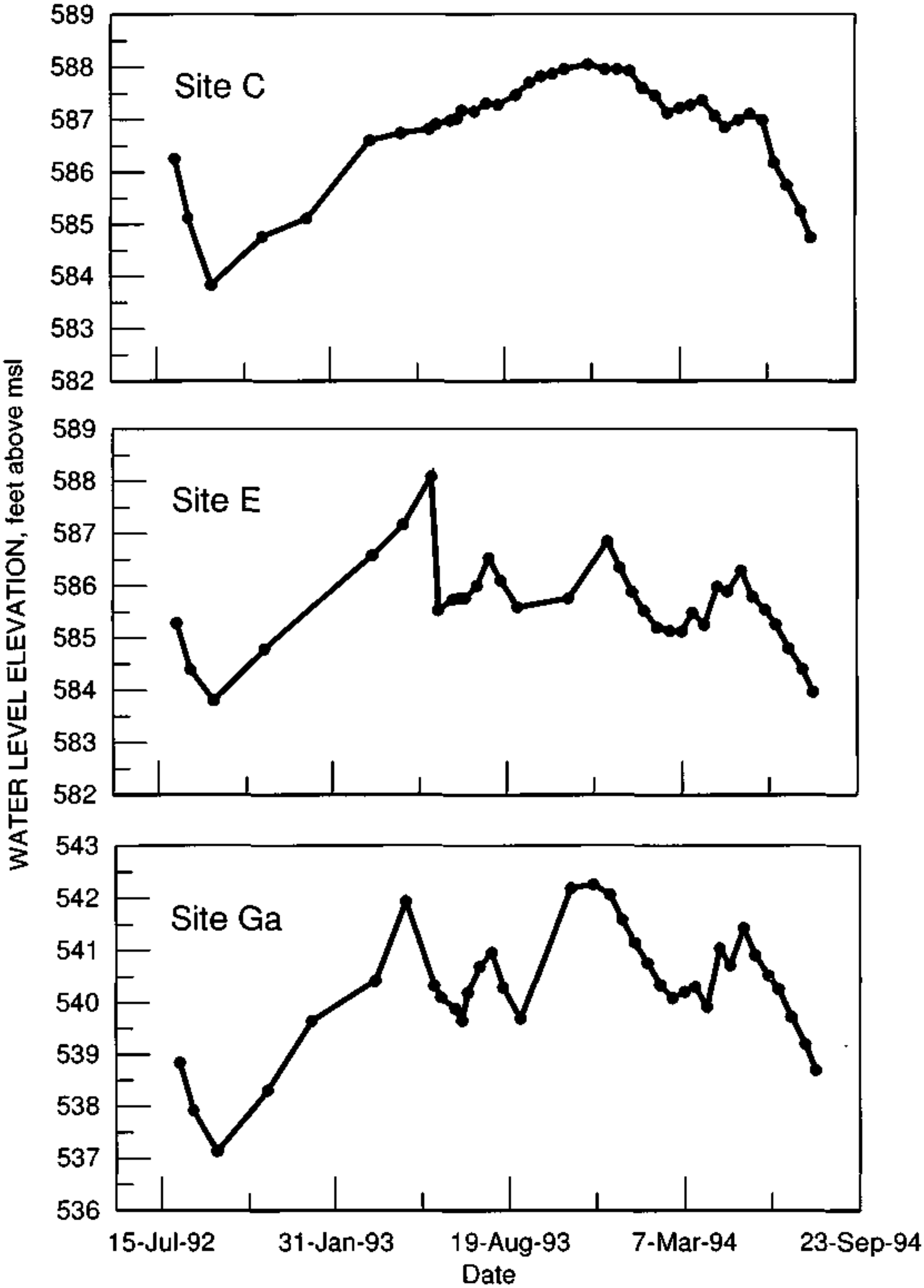
586
15-Jul-92
31-Jan-93
19-Aug-93
7-Mar-94
23-Sep-94
Date


WATER LEVEL ELEVATION, feet above msl
Site C
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588
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586
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584
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582
Site E
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588
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584
583
582
Site Ga
543
542
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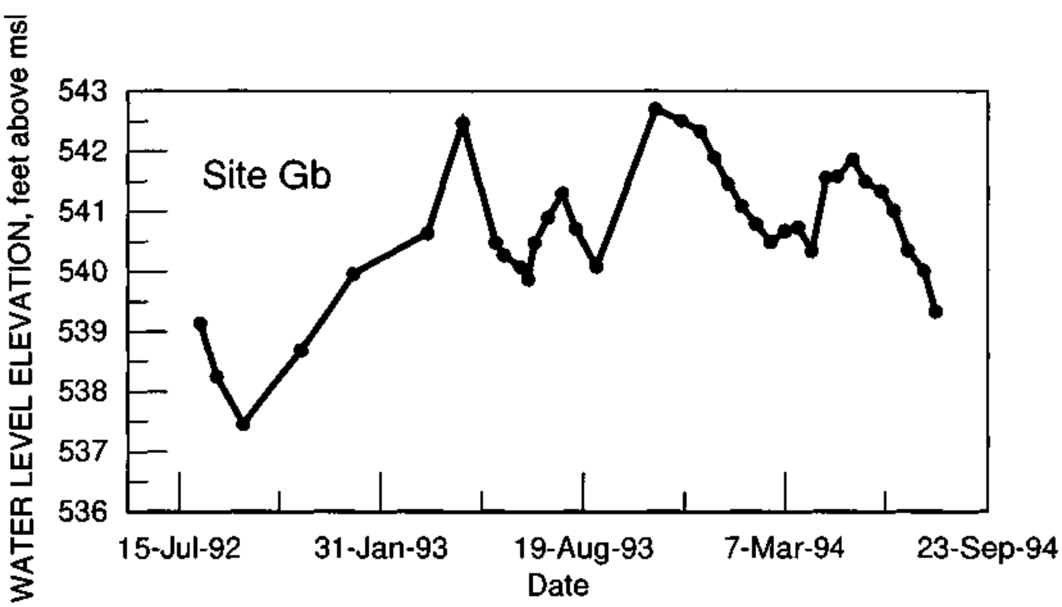
19-Aug-93
7-Mar-94
23-Sep-94
Date


WATER LEVEL ELEVATION, feet above msl
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541
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539
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537
536
Site Gb
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31-Jan-93
19-Aug-93
7-Mar-94
23-Sep-94
Date