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Hydrogeology, Water Use, and Simulated Ground-Water Flow and Availability in Campton Township, Kane County, Illinois

Scientific Investigations Report 2006–5076

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

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By Robert T. Kay, Leslie D. Arihood, Terri L. Arnold, and Kathleen K. Fowler

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Conversion Factors

Inch/Pound to SI

Multiply	By	To obtain
Length		
inch (in.)	2.54	centimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
Area		
acre	0.4047	hectare
section (640 acres or 1 square mile)	259.0	square hectometer
Volume		
gallon (gal)	3.785	liter
Flow rate[1]		
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
gallon per minute (gal/min)	0.06309	liter per second
gallon per day (gal/d)	0.003785	cubic meter per day
million gallons per year (Mgal/yr)	0.003785	Million cubic meter per year
inch per year (in/yr)	25.4	millimeter per year
Specific capacity		
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter
Hydraulic conductivity[2]		
foot per day (ft/d)	0.3048	meter per day
Transmissivity[3]		
foot squared per day (ft^2/d)	0.09290	meter squared per day

[1] Flow Rate: In this report, recharge rate is presented in inch per year (in/yr). Elsewhere, recharge rate may be presented in the form gallon per day per square mile ($gal/d/mi^2$). In this report, leakage coefficient is presented in foot per day per foot (ft/d/ft). Elsewhere, leakage coefficient may be presented in the form gallon per day per square foot ($gal/d/ft^2$).

[2] Hydraulic conductivity: The standard unit for hydraulic conductivity is cubic foot per day per square foot of aquifer cross-sectional area $(ft^3/d)/ft^2$. In this report, the mathematically reduced form, foot per day (ft/d), is used for convenience. Elsewhere, hydraulic conductivity may be presented in the form gallon per day per square foot ($gal/d/ft^2$).

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

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:
°F=(1.8×°C)+32
Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:
°C=(°F-32)/1.8

Datums

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD29). The NGVD29 is a geodetic datum derived from a general adjustment of first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

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

HYDROGEOLOGY, WATER USE, AND SIMULATED GROUND-WATER FLOW AND AVAILABILITY IN CAMPTON TOWNSHIP, KANE COUNTY, ILLINOIS

By Robert T. Kay, Leslie D. Arihood, Terri L. Arnold, and Kathleen K. Fowler

ABSTRACT

Several aquifers underlying Campton Township in Kane County, Illinois provide virtually all of the water supply to the residents of the township. These aquifers consist of layers of unconsolidated sand and gravel in the glacial drift; dolomite and shale of the Alexandrian Series and the Maquoketa Group (the Silurian-Maquoketa aquifer); dolomite of the Platteville and Galena Groups (the Galena-Platteville aquifer); and sandstones of the Glenwood Formation and the St. Peter Sandstone (the Ancell aquifer). In 2002, total withdrawals from these aquifers underlying Campton Township exceeded 1.36 million gallons day.

Water-level altitudes in the shallow and deep glacial drift aquifers generally follow surface topography. Comparison of water levels measured in 1995 and 2002 does not indicate large (15 feet or more) water-level declines in these aquifers beneath most of the township.

Water-level altitudes in the Silurian-Maquoketa aquifer generally decrease from west to east. The potentiometric surface of the aquifer follows the bedrock-surface topography in parts of the township, but local low water-level altitudes and large declines in water levels between 1995 and 2002 indicate that withdrawals from the Silurian-Maquoketa aquifer may exceed recharge in some areas.

Water-level altitudes in wells completed in the Galena-Platteville aquifer vary by more than 300 ft. Large water-level declines in wells completed in the Galena-Platteville aquifer from 1995 to 2002 indicate that withdrawals from the Galena-Platteville and Silurian-Maquoketa aquifers exceed recharge in the northern part of the township.

Water-level altitudes in wells completed in the Ancell aquifer are also highly variable. Although there is no indication of large water-level declines in Ancell aquifer between 1995 and 2002, historical data for one well completed in the aquifer indicate large water-level declines over a period of decades.

Computer simulation of flow in the ground-water system indicates that most of the shallow ground water underlying the township is derived from precipitation near the ground-water divide in the western part of the township. Shallow recharge moves primarily through the glacial drift aquifers and the upper part of the Silurian-Maquoketa aquifer, with minimal flow into the Galena-Platteville and Ancell aquifers. Most of the water in the Ancell aquifer beneath the township originates as surface recharge in the area west of the township. Vertical recharge to the Ancell aquifer from the Galena-Platteville aquifer beneath the township is not substantial. The source of the water withdrawn from the Ancell is inflow through the aquifer from areas west of the township. About 10 percent of ground water flowing through the township in 2002 was withdrawn by wells, with 80 percent flowing through the township and discharging to surface water bodies, including the Fox River. Simulation of additional withdrawals from the Ancell aquifer to supply an additional 605 proposed homes indicates about 17 ft of drawdown in the aquifer in the vicinity of a production well, but virtually no drawdown in any of the overlying aquifers.

INTRODUCTION

The population of Campton Township (*township** 40 north, *range* 7 east) in the central part of Kane County in northeastern Illinois (fig. 1), has more than doubled since 1975 as suburban Chicago expands westward (Julia Glass, Campton Township, oral commun., 2003) . This growth is expected to

*First use of words defined in the Glossary of this report is italicized.

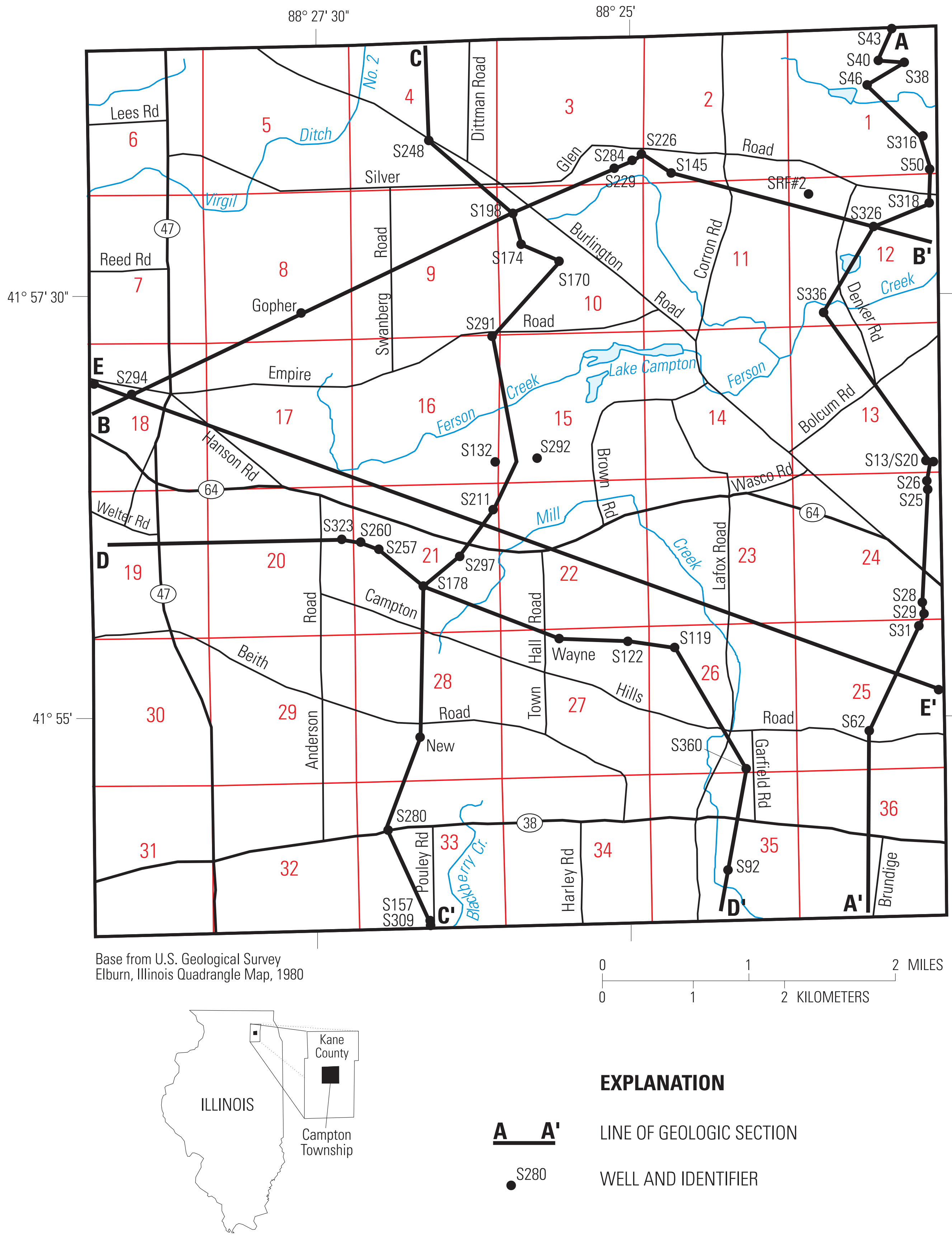


Figure 1. Location of selected wells and lines of geologic section, Campton Township, Kane County, Illinois.

continue. Because township residents rely exclusively on *ground water* for water supply, and most residents rely on water from residential-supply *wells*, there is concern about the effect of population growth on the availability of ground-water supplies. To gain a better understanding of the ground-water system, the U.S. Geological Survey (USGS), in cooperation with the Campton Township Board of Trustees, conducted a study of regional ground-water flow and water availability in Campton Township and the surrounding area.

This study was designed to determine the direction of regional ground-water flow and availability of water in the *aquifers* currently used for water supply (hereafter referred to as the aquifers used for residential-water supply) in the township. The study was divided into seven components: (1) compilation and analysis of available geologic and hydrologic data, (2) measurement of water levels in more than 200 residential-supply wells located throughout the township, (3) measurement of flow in 3 streams at a total of 12 locations, (4) assessment of water use in each of the aquifers used for residential supply within the township, (5) *geophysical logging* in 6 wells, (6) *aquifer testing* in 11 wells, (7) development of a computer *model* of the ground-water system underlying the township and the surrounding area.

Water-level declines due (at least in part) to pumping from residential-supply wells have resulted in many of the wells in the township not being capable of providing sufficient water to meet the demands of residential users. These declines have forced the affected homeowners to lower their pumps or drill new wells to meet their water needs. The information obtained from this investigation can be used by local residents to determine the potential for interruptions in their water supply and the most viable alternative source of water. Study results may also be used by governmental officials to help manage the effects of population growth on water supplies. Methods used in this study (for example, computer model simulation) can be used to assess similar hydrologic conditions in northern Illinois.

Purpose and Scope

This report describes the results of a study designed to determine the regional ground-water flow and water availability in the aquifers used for residential supply in Campton Township, Kane County, Illinois. In addition to a description of the geology, hydrology, and water use in the township, results of static water-level measurements, geophysical logging, aquifer testing, and streamflow measurements are presented. A computer model simulating regional ground-water flow and the response of water levels to current (2003) hydrologic and water-use conditions in the aquifers used for residential-water supply also is presented. In addition, results of model simulation of selected possible future water-use conditions are presented.

Geologic data were compiled and analyzed to identify the distribution of the geologic units that affect the availability of water supply. Compilation and analysis of the hydrologic data, including data from geophysical logging, aquifer testing, and water-level measurements, identified areas where the aquifers used for residential supply may have large (15 ft or more) declines in water level due to ground-water withdrawals exceeding recharge and storage. The hydrogeologic data also were analyzed to determine the effects of additional water withdrawals on water levels. Water use in the township was assessed to estimate the current demands on the aquifers and to correlate water use with identified areas of large declines in water levels. A ground-water-flow model was developed to further characterize the hydrogeology of the township and to simulate the effects of future water use or other changes on hydraulic conditions (drought, changes in *recharge* because of urbanization) on water supply in the aquifers.

Methods

Paper and electronic copies of all of the well logs available for the township were obtained from the Illinois State Water Survey (ISWS). Well logs contain driller-provided information on the well location by township, range, and *section* to 1/16th of the section, but also may include the well address or subdivision lot number. Driller's logs also contain information on the depth, thickness, and lithology of the geologic units encountered during drilling, the type and depth of well casing, the type and depth of grouting material used to seal the well casing, and the geologic unit(s) that yield water to the well. Driller-provided locations were cross-checked for accuracy to at least the quarter section by comparing driller-provided well address or subdivision lot number to the section and quarter-section location. Of the approximately 1,500 logs on record, locations of 74 wells were thought to be incorrect and were relocated to the true locations based on the address, owner name, or the lot number. The location of every well in the township was then plotted and the altitude of the land surface at that point was determined by use of a 1:24,000 scale topographic map. For the majority of properties, land-surface altitude varies by less than 10 ft and the altitude of the well should be accurate to within about 5 ft. Greater uncertainty is associated with properties with larger altitude variations across the property. Lithologic descriptions from the logs were then compiled and interpreted by use of geographic information system (GIS) software to determine the geometry and extent of the geologic units in the township and the aquifers used for water supply.

Lithologic interpretations were based primarily on analysis of interpretations provided by previous investigators, geophysical logs collected as part of this investigation, and the lithologic logs provided by the drillers. Lithologic logs were analyzed by use of various techniques, including a geostatistical method called kriging (Isaaks and Srivastava, 1989). Previous investigations in and near the township have focused on the glacial unit as well as the four *bedrock* units: the *Silurian*-aged units (Silurian System), the Maquoketa *Group*, the Galena and Platteville *dolomites* and the Ancell Group. Lithologic (and geophysical) logs were used to define the altitude of the top and bottom of each unit at each well where appropriate data were available. Ordinary kriging was used to fit a continuous surface through the defined points. Next, the extent of aquifer material in the unit was determined. Aquifers known to underlie the entire the township, such as the Ancell aquifer, were not investigated further for extent. For discontinuous aquifers, such as those within the glacial units, the extent of high and low *permeability* material was estimated by indicator kriging. A value of 1 was assigned if *sand*-and-*gravel* units indicative of aquifer material were described as being present at the well. If the well log indicates aquifer material was absent, a value of 0 was assigned. Indicator kriging then was used with the aquifer presence indicator to calculate an areal distribution of values between 1 and 0 throughout the aquifer. The distribution represents the probability of the aquifer being present at any random point. The cut-off point of 0.5 (50-percent probability) has been used as a reliable indicator of the boundary between aquifer and non-aquifer material (Johnson and Dreiss, 1989). In this study, a cut-off of point 0.7 was used to increase the probability of aquifer material being present at a given predicted location. However, interpretations based on the kriging results have some degree of uncertainty, as do all hydrogeologic interpretations.

Field work for this investigation consisted of four activities: streamflow measurements, ground-water-level measurements, geophysical logging, and aquifer testing. Streamflow measurements were made during summer and fall periods when flow was likely to have been sustained primarily by recharge from ground water (base flow). The size of the drainage basins above each of the 12 measurement stations were obtained by GIS analysis of surface topography.

Water levels were measured in about 220 wells open to each of the aquifers used for residential-water supply in Campton Township during June and July 2002 (appendix A). Water levels had also been measured in 140 of these wells during a previous investigation in May and June 1995 (Kay and Kraske, 1996). Water levels also were measured periodically in wells S354 and S360 from March 2003 through August 2005. Water levels typically were measured by use of a steel or electric tape calibrated to 0.01 ft. These wells typically had well logs with an owner name, address, and (or) a subdivision and lot number so that well construction and location information could be identified. Information on residential-well construction was obtained from the well logs and verified wherever possible by the homeowner. Efforts were made to ensure that the wells had not been pumped immediately prior to measurement and replicate water levels were collected where possible to ensure accurate data and (to the extent practical) *steady-state* ground-water-flow conditions. Measuring point altitude of the wells was estimated by use of topographic maps with a 2-ft contour interval provided by the Kane County Development Department. Potentially inaccurate measurements are denoted by a * (asterisk) in appendix A, and the results from these wells are not included in the discussion unless noted.

In June 2003 and August 2005 geophysical logs, including *caliper, natural gamma*, *normal resistivity*, fluid temperature, *fluid resistivity*, *flowmeter*, and *spontaneous potential* were recorded in five residential-supply wells (S302, S355, S356, S357 and S358) and one former production well (S354) in the township. These logs were used to identify the geologic properties of the units penetrated by the wells, to characterize the amount of vertical flow within the wells under ambient (non-pumping) conditions, and to calculate the hydraulic properties of the units intercepted by the wells (table 1). Techniques described by Paillet and others (2000) were used to determine the hydraulic properties.

Aquifer tests were performed in wells S354, S355, and S356 as they were being geophysically logged by recording water-level changes by use of a pressure transducer and datalogger during injection of between 2 and 4 gal/min of water (depending on the well) at a constant rate into the wells (table 1). The length of these tests varied from 24 to 80

Table 1. Summary of information obtained during geophysical logging of wells in Campton Township, Illinois June 2003 and August 2005.

[GPAn, Galena-Platteville and Ancell; <, less than; na, not available]

Well name (fig. 22)	Open interval (feet below measuring point)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Water-level altitude (feet above NGVD29)	Date of logging	Amount of vertical flow in borehole under ambient conditions (gallons per minute)	Altitude, top of Galena Group (feet above National Geodetic Vertical Datum of 1929)	Altitude, top of Ancell Group (feet above National Geodetic Vertical Datum of 1929)	Transmissivity of Ancell aquifer (feet squared per day)	Transmissivity of Galena-Platteville aquifer (feet squared per day)
S302	400-740	GPAn	954	541.4	6/16/03	10	na	na	13	25
S354	400-1020	GPAn	954	501.3	6/20/03	<0.1	559	224	<18	<18
S355	425-780	GPAn	932	500.0	6/17/03	0.1	557	222	<22	<22
S356	220-700	GPAn	930	512.8	6/18/03	<0.1	579	250	6	24
S357	199-720	GPAn	892	623.2	6/19/03	0.3	572	242	na	na
S358	445-745	GPAn	938	495.4	8/16/05	0.14	548	218	na	na

minutes. Aquifer tests also were performed in five wells open to bedrock aquifers and six wells open to the glacial *drift* aquifers by pumping between 4 and 15 gal/min of water (depending on the well) at a constant rate from the wells and recording water-level changes through time (table 2). The length of these tests typically was 100 minutes. The discharge or injection rate during the tests was held constant and verified by determining the length of time required to fill a 5-gal bucket. Water-level data were collected on a logarithmic time sequence by use of either a calibrated pressure transducer and datalogger or an electric water-level indicator. Aquifer test data were analyzed using the straight-line method of Cooper and Jacob (1946).

Description of the Township

Physiography and Climate

The northern two-thirds of Campton Township is in the Wheaton Morainal Plain of the Great Lakes section of the Central Lowland Physiographic Province (Leighton and others, 1948). The southern one-third of the township is within the Bloomington Ridged Plain Subsection of the *Till* Plains Section of the Central Lowland Physiographic Province. Glacial and *lacustrine* processes have produced most of the physiographic features in this area, including the *moraines* and *kames* that correspond to the

Table 2. Summary of aquifer-test information from residential-supply wells in Campton Township, Illinois.

[SMGP, Silurian-Maquoketa and Galena-Platteville; SM, Silurian-Maquoketa; DD, deep glacial drift; ? Denotes unknown information]

Well name (figs. 17, 20, 21)	Open interval (feet below land surface)	Initial depth to water (feet below top of casing)	Measuring point altitude (feet above NGVD29)	Aquifer	Transmissivity (feet squared per day)
S24	156-450	74.20	804	SMGP	4.9
S331	?-280	158.44	940	SM	1,410
S360	90-180	38.30	820	SM	8,370
S301	175-?	49.65	826	SM	12.9
S358	?-325	88.03	889	SM	1,680
S209	220-225	90.52	886	DD	2,350
S359	260-265	121.28	903	DD	150
S187	190-200	137.65	918	DD	490
S218	210-220	89.90	933	DD	49
S201	220-230	75.62	905	DD	4.7
S290	?-271	105.77	840	DD	190

topographic uplands present in most of the township (Willman, 1971). A kame is defined by the topographic high near Town Hall Road and Campton Hills Road (fig. 2). The area approximately between the 900- and 950-ft contours in the northwestern part of the township near Swanberg and Dittman Roads marks the location of part of glacial Lake Pingree. Topographic highs associated with the 950-ft contour in the northwestern part of the township near Route 47 define part of the location of the Marengo Moraine. Most of the landforms in the remainder of the township were formed as part of the Elburn Morainic Complex.

Land-surface altitude in Campton Township ranges from more than 1,000 ft near Town Hall Road and Campton Hills Drive to about 750 ft in the southeastern corner of the township (fig. 2). Land-surface altitude typically is above 850 ft in the northern and western parts of the township and between about 850 and 750 ft in the eastern part of the township. A topographic ridge in the northwestern part of the township (fig. 2) defines part of the surface-water drainage divide between the Rock River and Illinois River Basins. South of the ridge, surface water flows to Ferson, Mill, and Blackberry Creeks, which drain to the Fox River and eventually the to Illinois River. North of the ridge, surface water flows to Virgil Ditch, which drains to the Kishwaukee River and eventually to the Rock River.

The climate in this area is classified as temperate continental, with a mean annual temperature of about 10^{o} C and a mean annual precipitation of about 36.5 in. (National Oceanic and Atmospheric Administration, 2002). More than half the average annual precipitation falls from April 1 through August 31.

An estimated 70 percent of the average annual precipitation in Illinois is returned to the atmosphere by *evapotranspiration* and about 19 percent flows to streams as overland runoff (Mades, 1987; Illinois State Water Survey, 2005). On the basis of these percentages, average annual precipitation available for recharge to ground water is about 4 in.. Because of annual patterns in precipitation and evapotranspiration, recharge of precipitation to ground water occurs primarily during the spring and early summer.

Land Use

Prior to about 1963, land use in the township was primarily agricultural, except for small residential areas around Lake Campton and the towns of Wasco and Lily Lake, with some forest and wetlands (fig. 3). Residential development in Campton Township has increased continuously since 1963 and current land use is now more than 50-percent residential. Land used for agricultural activities is located primarily in the southern and western parts of the township. Wetland areas are located along much of Ferson Creek. Residential developments are primarily in the eastern and central parts of the township. Residential land use in the township is expected to increase (Julia Glass, oral commun., 2003).

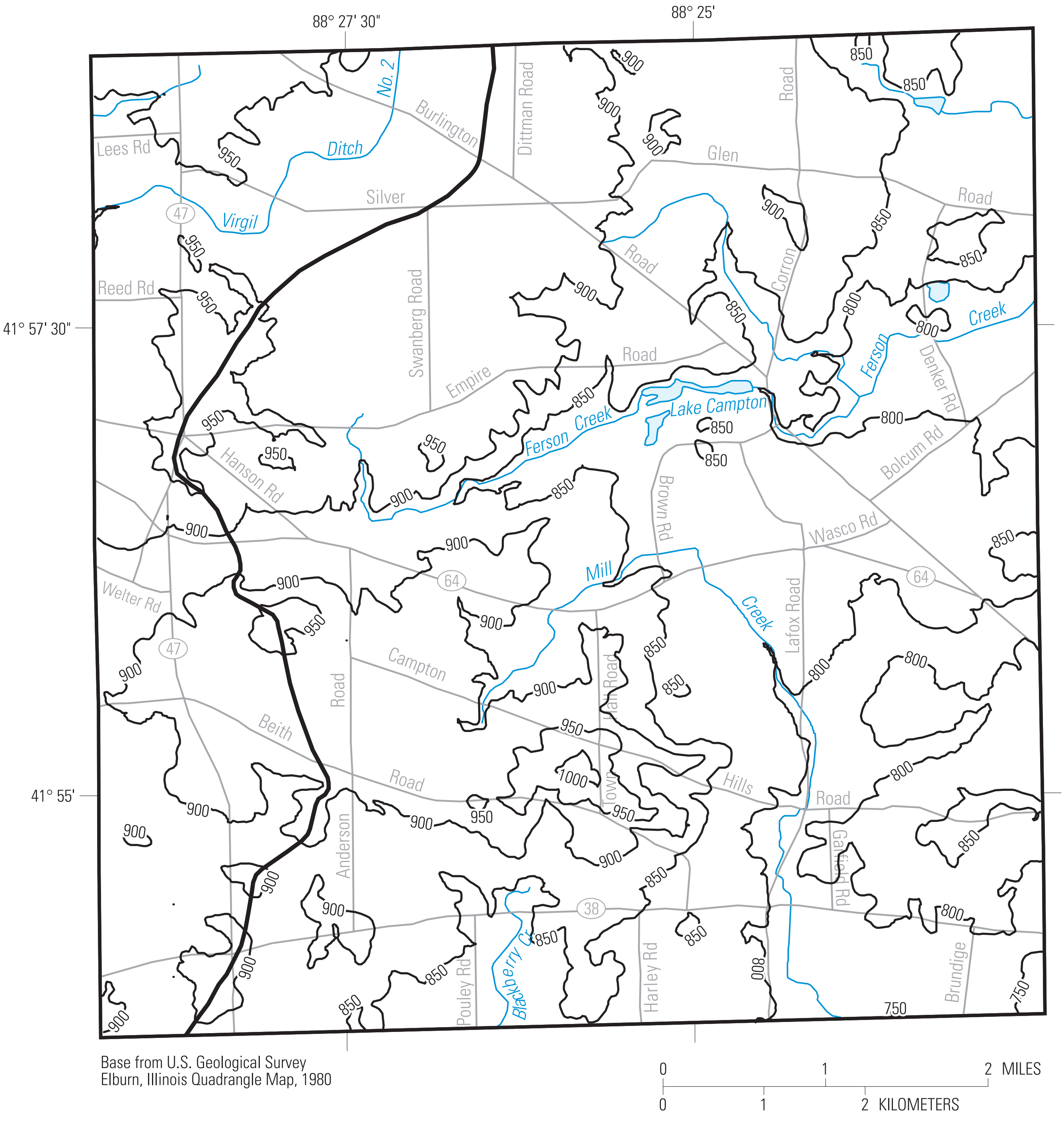
GEOLOGY

Assessment of the geology in the township is based on the description of previous investigators (Willman, 1971; Visocky and others, 1985; Graese and others, 1988; Schumaker, 1990; Kay and Kraske, 1996; Grimley and Curry, 2001), analysis of *lithologic logs* on file with the ISWS, and geophysical logging in six wells. Unless referenced as being used for this report, the *stratigraphic* nomenclature used in this report is that of the Illinois State Geological Survey (ISGS) (Willman and others, 1975; Hansel and Johnson, 1996) and does not necessarily follow the usage of the USGS (fig. 4).

Bedrock Units

The bedrock geologic units of concern to this study consist of *sandstones*, *shales* and dolomites of *Ordovician* and Silurian age (figs. 4, 5, and 6). From oldest to youngest, these units are the St. Peter Sandstone; the Glenwood *Formation*; the Platteville, Galena, and Maquoketa Groups; and the Egdewood and Kankakee Dolomites. Bedrock units are *unconformably* overlain by *unconsolidated* glacial and glacialfluvial units of *Quaternary* age.

The St. Peter Sandstone is a well-sorted, coarse- to medium-grained *quartz arenite*. The sandstone is well rounded, poorly cemented, and is typically between 150 and 250 ft thick beneath Campton Township (Graese and others, 1988). The Glenwood Formation overlies the St. Peter Sandstone and consists of interbedded coarse-grained dolomitic sandstone and sandy and *argillaceous* dolomite (Schumaker, 1990). The thickness of the Glenwood Formation is 25–50 ft in most of the township (Schumaker, 1990), but may be as much as 75 ft thick in some areas (Graese and others, 1988). The St. Peter Sandstone and the Glenwood Formation together constitute the Ancell Group. The St. Peter Sandstone and the sandstone units of the Glenwood Formation are referred to as the Ancell sandstone in this report. The altitude of the top of the Ancell



EXPLANATION

—900— LINE OF EQUAL ALTITUDE OF LAND SURFACE, IN FEET ABOVE NGVD29--Contour interval 50 feet

SURFACE-WATER DIVIDE

Figure 2. Surface topography of Campton Township, Illinois.

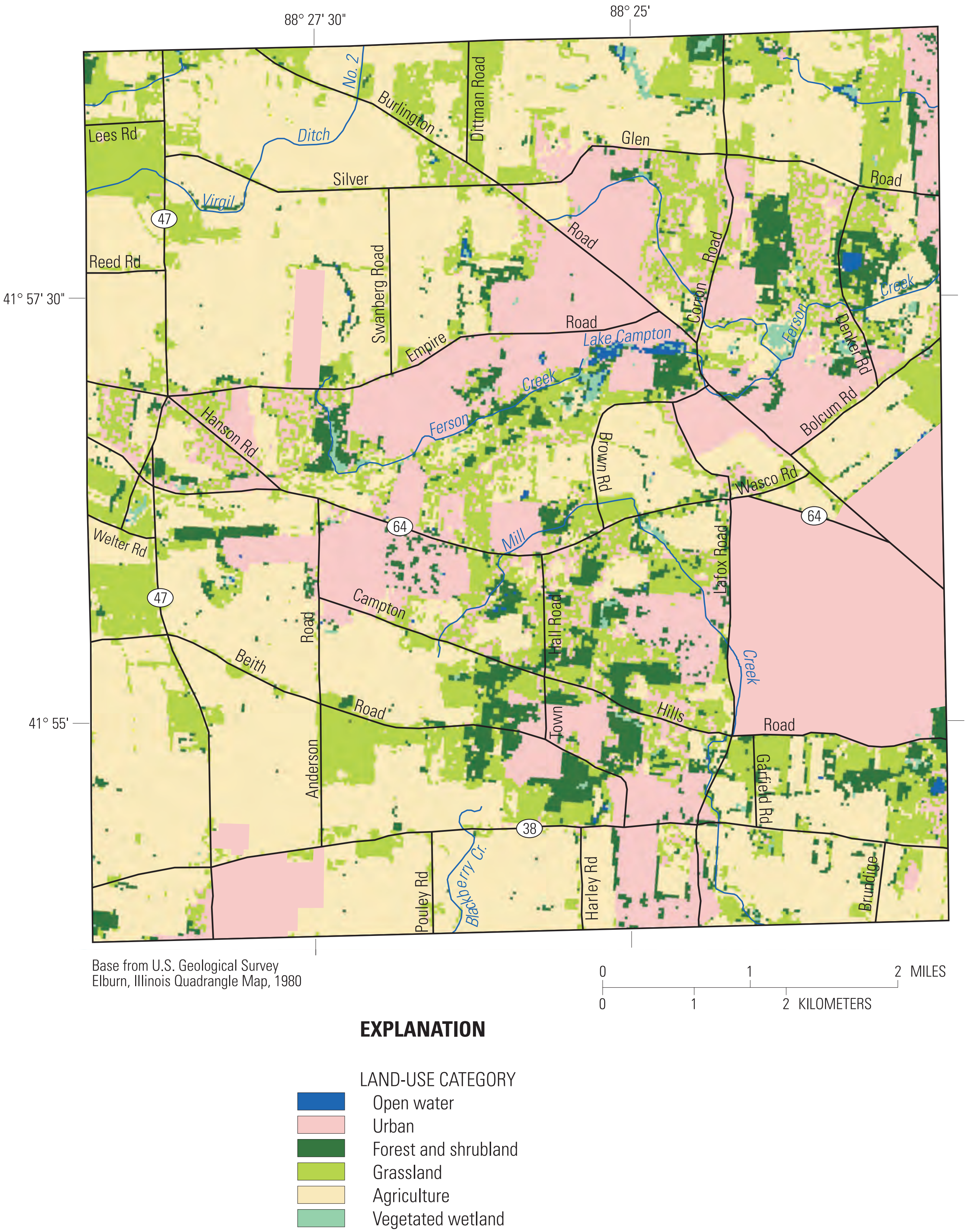


Figure 3. Land use in 2002, Campton Township, Illinois.

sandstone decreases from about 242-263 ft the northwestern part of the township to about 169-196 ft in the eastern part of the township (figs. 6 and 7). Natural-gamma logs indicate that the upper 50-80 ft of the Ancell sandstone, which typically is the part of the unit penetrated by residential-supply wells in the township, is more argillaceous than the rest of the sandstone (fig. 5)(appendix B).

The Platteville Group overlies the Glenwood Formation and consists of very fine- to medium-grained, pure to argillaceous dolomite (Graese and others, 1988). The Platteville Group can be divided into a lower, pure dolomite with a sandy base, and an upper *limestone* and dolomite characterized by green and red shale partings (Graese and others, 1988). The Platteville Group is about 145 ft thick beneath the township.

The Galena Group overlies the Platteville Group and can be divided into a basal unit of pure dolomite with brown shale partings, a middle unit composed of *cherty*, vuggy, medium-grained dolomite, and an upper unit of pure dolomite (Graese and others, 1988). The thickness of the Galena Group ranges from 160 to 200 ft beneath the township. The altitude of the top of the Galena Group decreases from about 590 ft in the northwestern part of the township to less than 460 ft in the southeastern part of the township (fig. 8). The altitude of the top of the Galena Group is between 560 and 520 ft beneath most of the township. The Galena Group is lithologically similar to the Platteville Group and these units are hereafter referred to as the Galena-Platteville dolomite.

The Maquoketa Group overlies the Galena Group and is the uppermost bedrock unit in most of the center and western parts of the township as well as its eastern edge (figs. 6 and 9). The total thickness of the Maquoketa Group ranges from 130 to 200 ft beneath the township (Graese and others, 1988). The *lithology* of the Maquoketa Group is highly variable laterally (Visocky and Schulmeister, 1988). The basal unit of the Maquoketa Group is predominately dolomitic shale with interbedded dolomite. This basal unit is overlain by a pure to

SYSTEM	GROUP	FORMATION	MEMBER	LITHOLOGY	THICKNESS (feet)	HYDROGEOLOGIC UNIT	
QUATERNARY		Greyslake Peat		Peat, silt, clay (where present)	0-10	Shallow Glacial Drift Aquifers	Semiconfining Unit
		Cahokia		Alluvial sand, silt, clay (where present)	0-20		
	MASON	Equality		Lacustrine silt, clay, sand (where present)	0-20		
		Henry		Glacial alluvial silt, sand, gravel, interspersed till, silt, clay	5-150		
		Lemont	Batestown	Till, some interspersed sand and gravel	0-35		
		Tiskilwa		Till, some interspersed sand and gravel	Up to 220	Deep Glacial Drift Aquifers	
			Robein Silt	Silt, peat, clay	0-10		
		Glasford		Till, sand, gravel	0-70		
SILURIAN		Kankakee		Dolomite	0-50	Silurian-Maquoketa aquifer	
		Edgewood					
ORDOVICIAN	MAQUOKETA			Shale, locally argillaceous dolomite and limestone	130-200		
	GALENA			Dolomite and limestone	300-350	Galena-Platteville aquifer	
	PLATTEVILLE						
	ANCELL	Glenwood		Sandstone, argillaceous dolomite at top	25-75	Ancell aquifer	
		St. Peter		Sandstone. Basal shale	150-250		

Figure 4. Generalized geohydrologic column showing stratigraphy, lithology, and hydrogeologic units, Campton Township, Illinois.

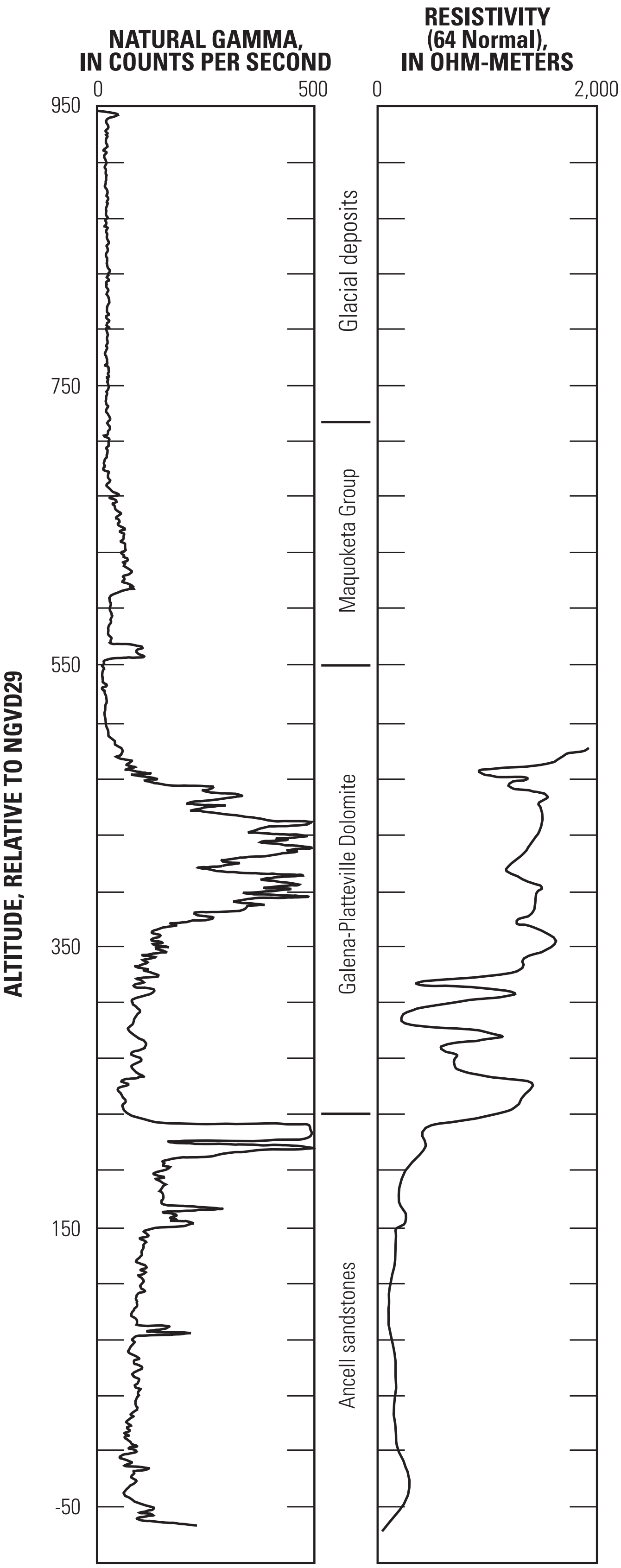


Figure 5. Select geophysical logs for well S354, Campton Township, Illinois.

argillaceous dolomite or limestone with some shale (Visocky and others, 1985). The argillaceous dolomite/limestone is overlain by dolomitic shale with interbedded dolomite and limestone (Schumaker, 1990). Each of the units within the Maquoketa Group can be observed in the natural-gamma logs (fig. 5). It should be noted that the high counts per second associated with the Galena-Platteville dolomite in well S354 (fig. 5) is not repeated in the other natural gamma logs (appendix B), which indicate the most argillaceous deposits penetrated by the wells are associated with the Maquoketa units. The anomalous response at well S354 is attributed to the larger diameter of this well at the interval of Maquoketa units in comparison to the interval of the Galena-Platteville dolomite, which decreases the response of the gamma signal.

Analysis of the lithologic logs indicates that shale units in the Maquoketa Group are likely to be encountered between 690 and about 740 ft in the northwestern part of the township, much of the south-central part of the township, and some of the northeastern part of the township. Shale is likely to be encountered below 650 ft in the east-central, southeastern, and southern parts of the township. The shale unit is estimated to be from 31 to 90 ft thick beneath most of the township, and is generally thickest in the northeastern and northwestern parts of the township (fig. 10). The shale unit is less than 15 ft thick beneath the southeastern part of the township and at scattered locations in the central and south-central parts of the township.

The youngest bedrock units beneath the township, where present, are the argillaceous dolomites of the Elwood and Kankakee Formations of the Silurian System (figs. 6 and 9). These dolomites unconformably overlie the Maquoketa Group and range from 0 to over 50 ft thick beneath the township (Graese and others, 1988).

Fractures, joints, and *solution openings* are most abundant in the upper 50 ft of the bedrock, where the bedrock is most weathered. These features decrease in size and number with depth (Graese and others, 1988).

The altitude of the bedrock surface varies by more than 150 ft beneath the township (fig. 11). Bedrock highs, approximately noted by the 750 ft contours, are located beneath much of the northern and southern parts of the township and at isolated locations in the eastern part. Bedrock valleys, defined by the 600 and 650 ft contours, are present in the central and eastern parts of the township (the Elgin bedrock valley) and the southwestern part of the township (the Elburn bedrock valley). These bedrock valleys cut through the bedrock and mark the location of pre-Pleistocene erosional surfaces

(Visocky and Schulmeister, 1988). Dolomite of the Silurian System is the uppermost bedrock unit beneath the north-central, southeastern and south-western parts of the township and generally is associated with the bedrock highs. Shale of the Maquoketa Group typically is the uppermost bedrock unit in the vicinity of the bedrock valleys (fig. 8).

Unconsolidated Units

The bedrock units are unconformably overlain by unconsolidated till, sand and gravel, and *silt* and *clay* of Quaternary age throughout the township (figs. 4, 12a, b, c, d). The unconsolidated units are typically thickest, between 250 and 300 ft, in the western half of the township, and thinnest, about 50 ft, near the southeastern corner of the township (figs. 12b, 12d). The deposits that compose the unconsolidated units have a complex distribution, often being present only in parts of the township and interfingering within and between units.

The Glasford Formation is the oldest unconsolidated unit in the township (Graese and others, 1988; Grimley and Curry, 2001)(fig. 4). The Glasford Formation, where present, is composed of sandy loam till with extensive basal sand and gravel (Curry and Seaber, 1990). The till thickness ranges from 0 to about 70 ft thick beneath the township, whereas the sand-and-gravel units range from 0 to 45 ft thick. The sand-and-gravel units are thickest in the bedrock valleys and are thin or absent at the bedrock highs.

The Robein *Member* of the Roxana Silt (hereafter referred to as the Robein Silt), where present, overlies the Glasford Formation (fig. 4). The Robein Member is composed of silts, sand, and clay with abundant organic carbon that has been modified by the process of soil formation. The Robein Silt typically is 0-10 thick beneath the township.

The Tiskilwa Formation overlies the Glasford Formation where the Robein Silt is absent (fig. 4). The Tiskilwa Formation is composed of clay loam till with interspersed beds of sand and gravel (Grimley and Curry, 2001). The Tiskilwa Formation is as much as 220 ft thick in the northwestern part of the township and thins to the east and south. The Tiskilwa Formation is the surficial geologic unit beneath much of the northwestern part of the township.

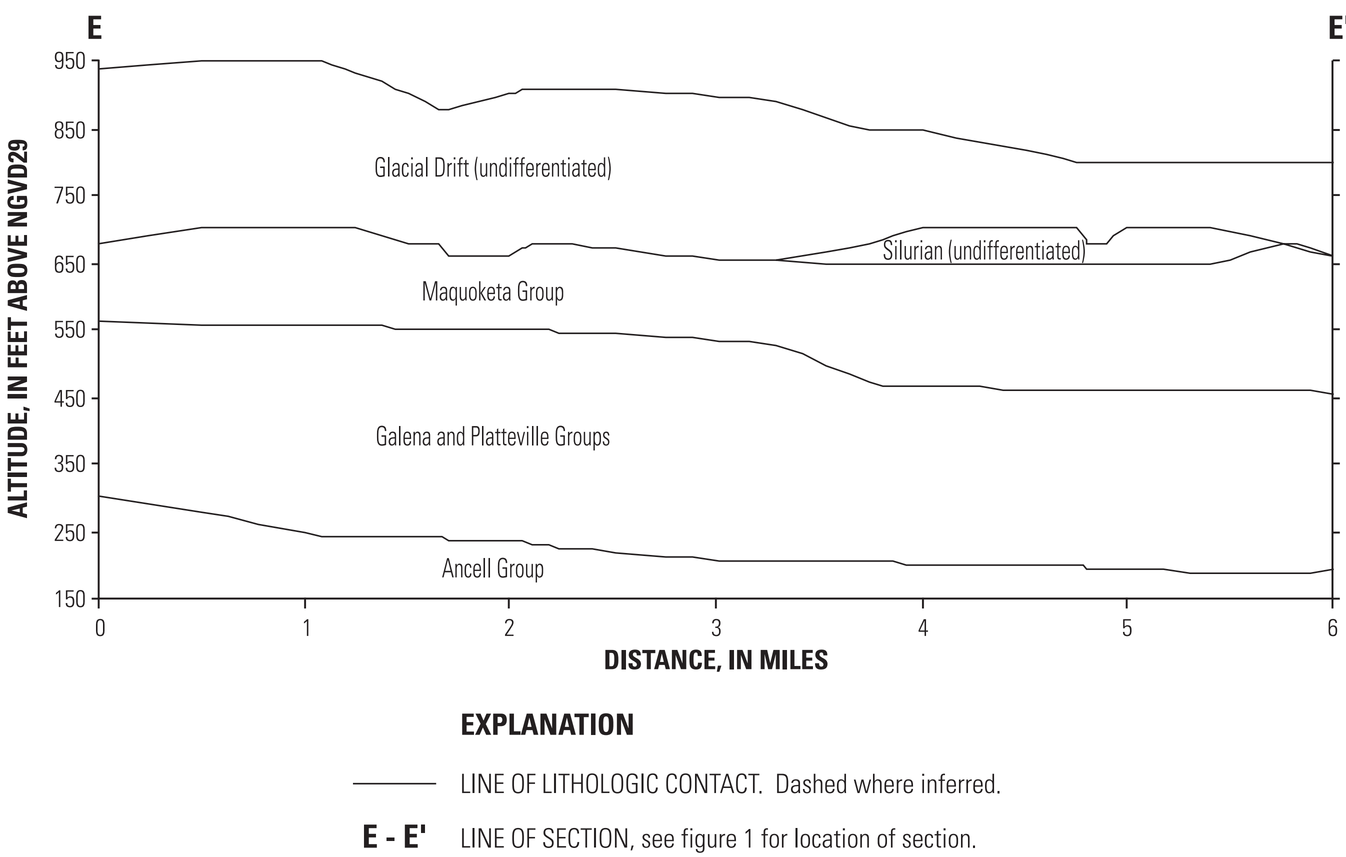


Figure 6. Geologic section E-E′ through the glacial drift and the bedrock, Campton Township, Illinois.

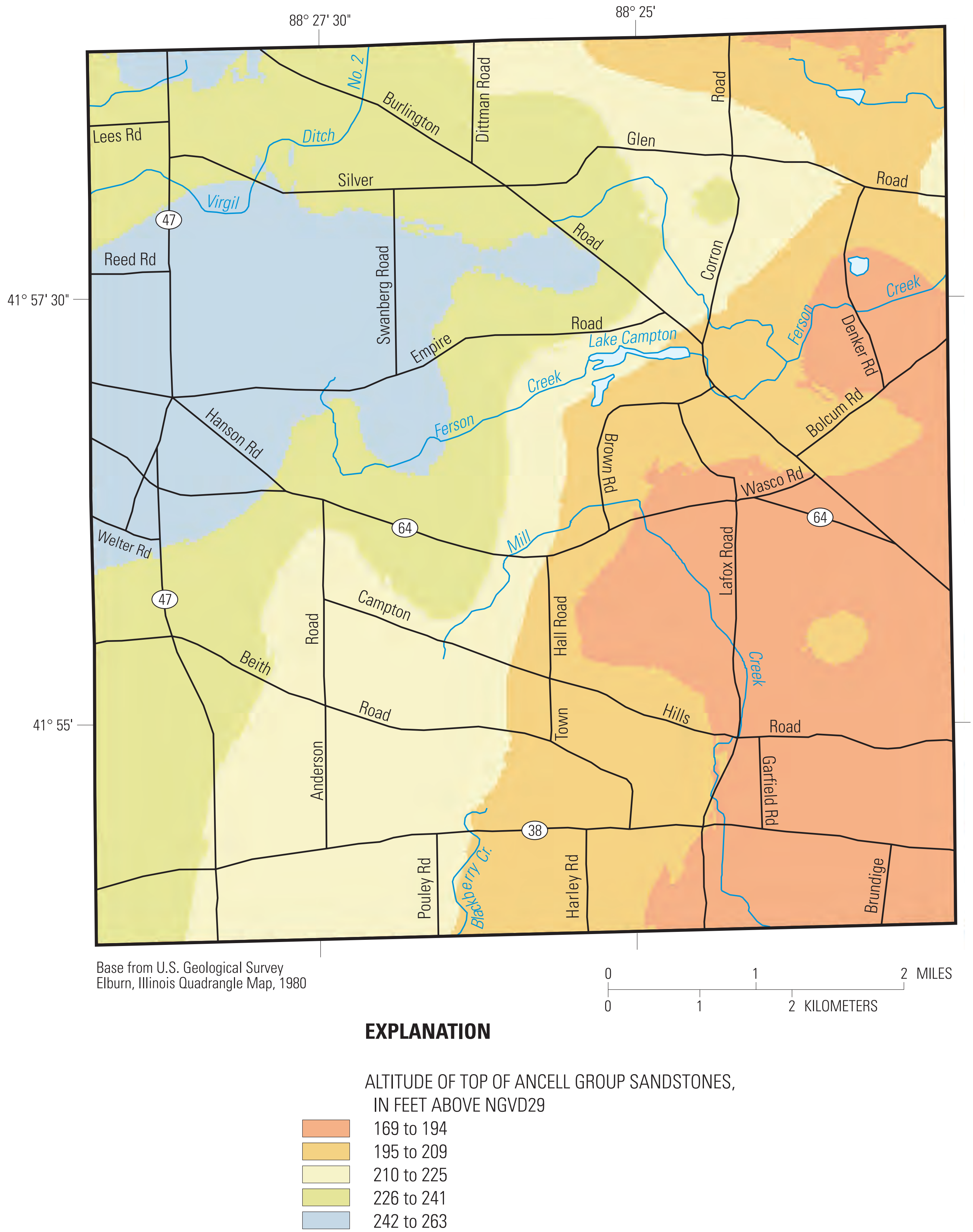
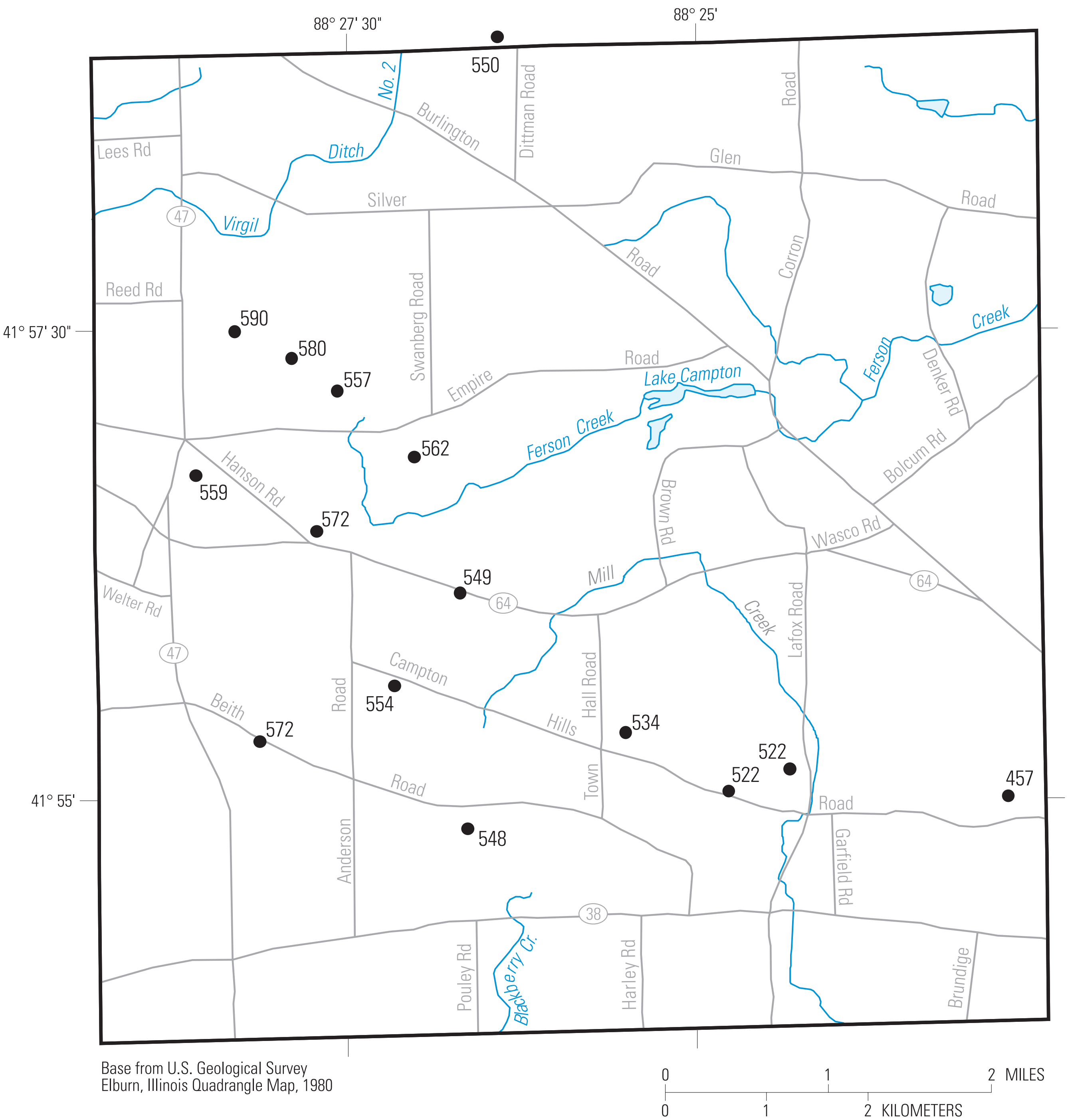


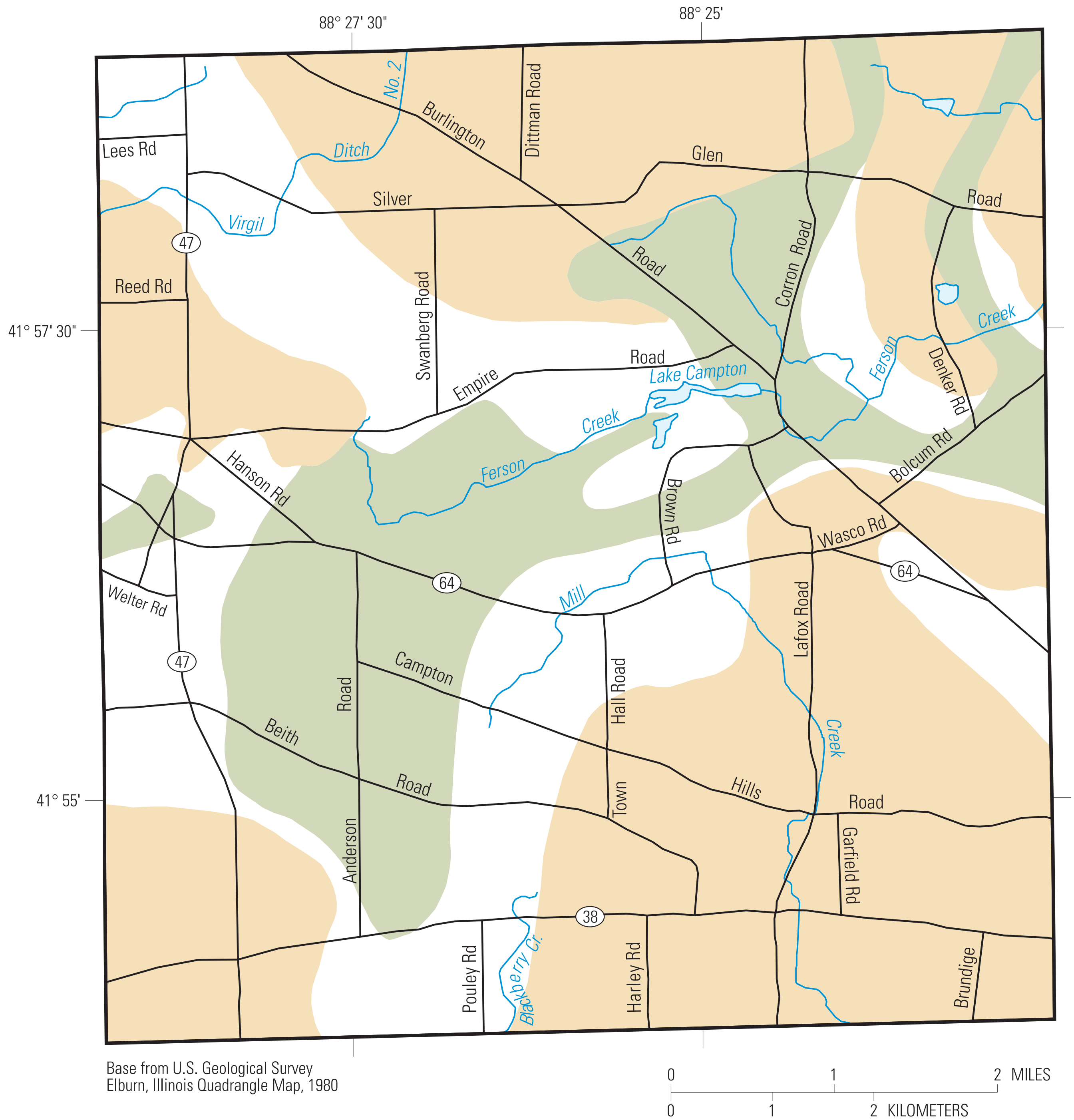
Figure 7. Altitude of top of Ancell Group sandstones, Campton Township, Illinois.



EXPLANATION

572 WELL--Number is altitude of top of Galena-Platteville dolomite, in feet above NGVD29.

Figure 8. Altitude of the top of the Galena-Platteville dolomite, Campton Township, Illinois.

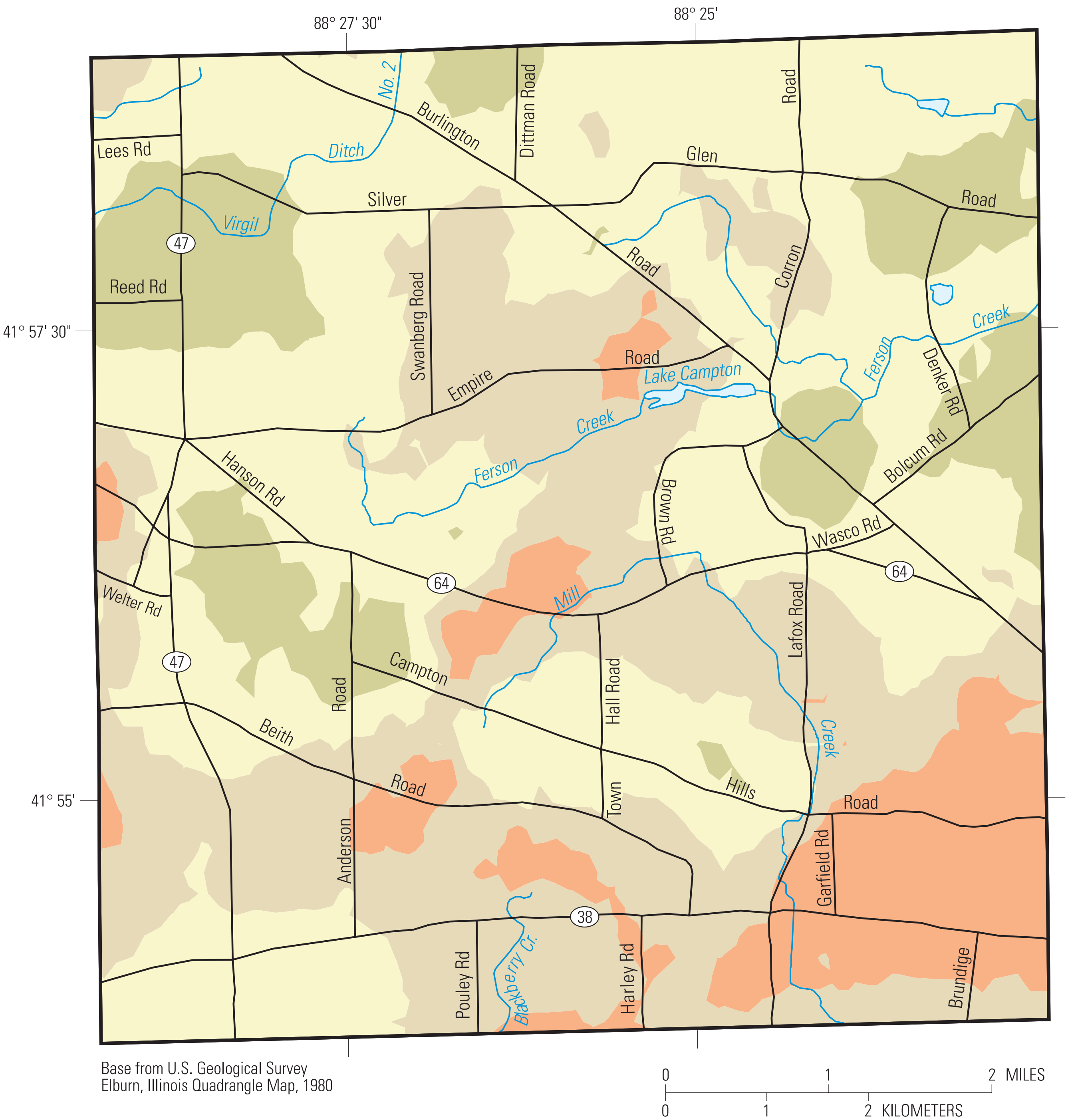


EXPLANATION

LITHOLOGY OF BEDROCK SURFACE

- Maquoketa Shale
- Maquoketa Dolomite
- Silurian Dolomite

Figure 9. Bedrock-surface lithology, Campton Township, Illinois.

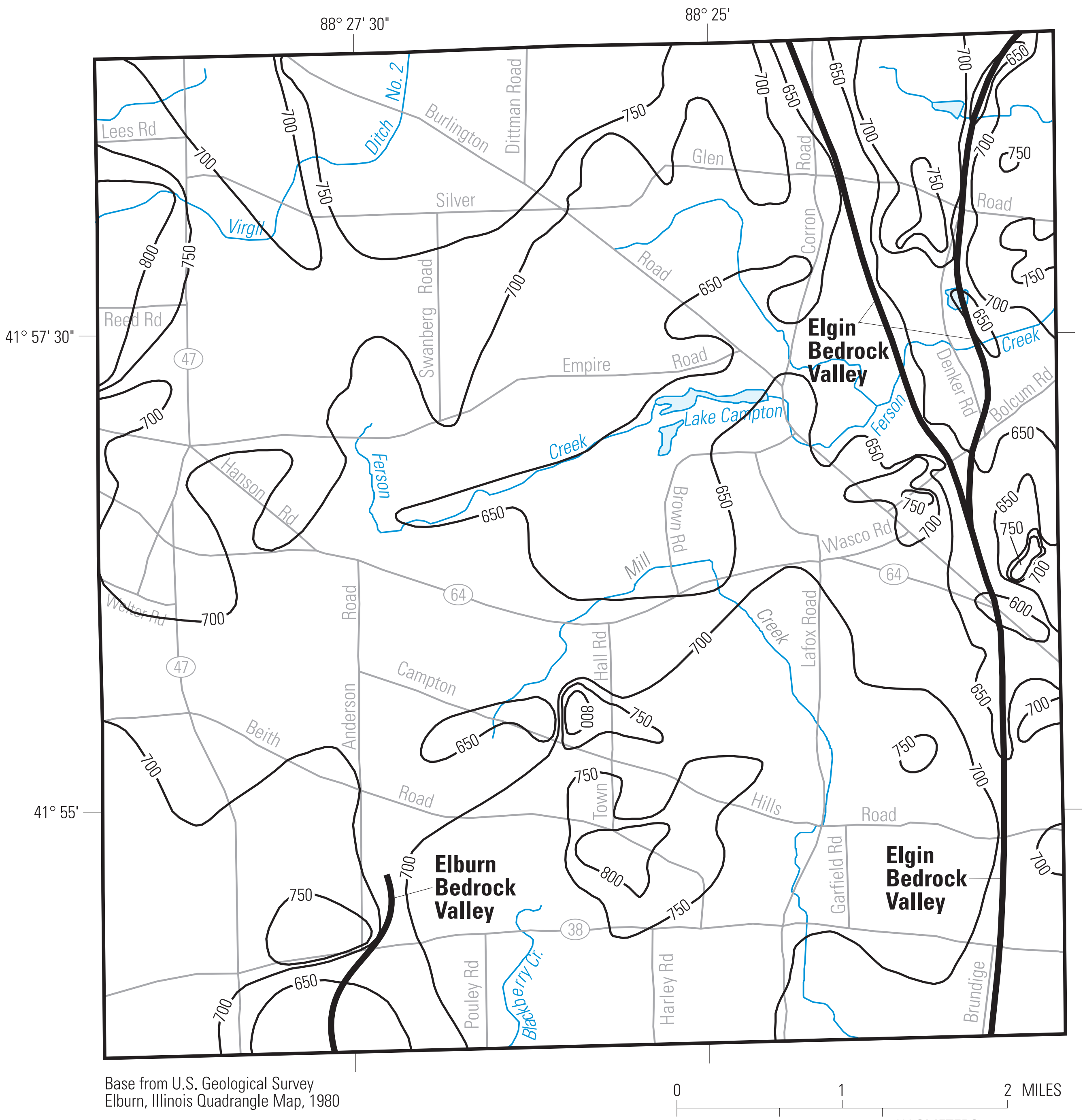


EXPLANATION

THICKNESS OF MAQUOKETA SHALE, IN FEET

- Less than 15
- 15 to 30
- 31 to 60
- 61 to 90

Figure 10. Thickness of Maquoketa Shale unit, Campton Township, Illinois.



EXPLANATION

750 LINE OF EQUAL ALTITUDE OF BEDROCK SURFACE, IN FEET ABOVE NGVD29--Contour interval 50 feet

BEDROCK VALLEY

Figure 11. Bedrock-surface topography, Campton Township, Illinois.

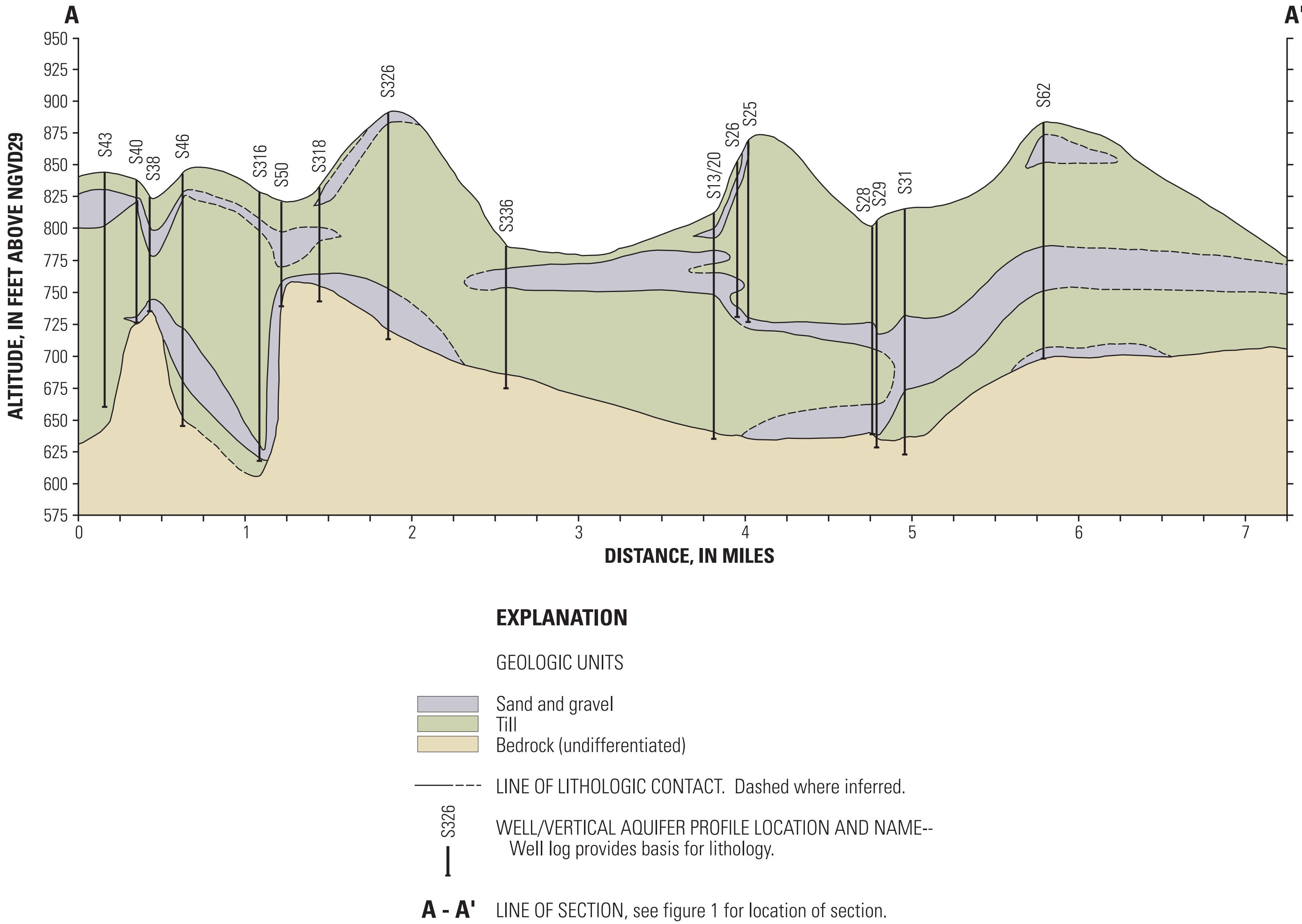


Figure 12A. Geologic section A-A′ through the glacial drift, Campton Township, Illinois.

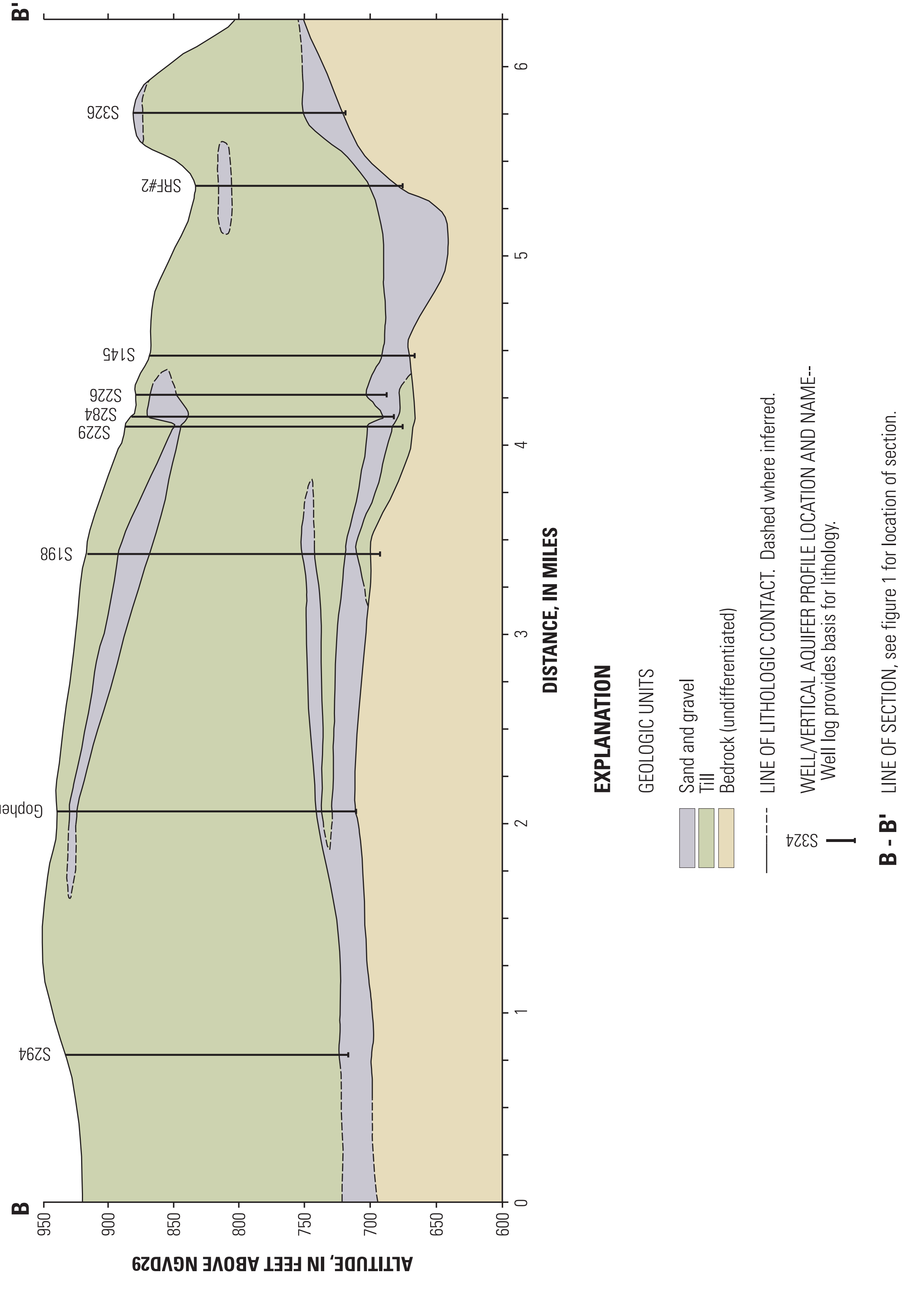


Figure 12B. Geologic section B-B′ through the glacial drift, Campton Township, Illinois.

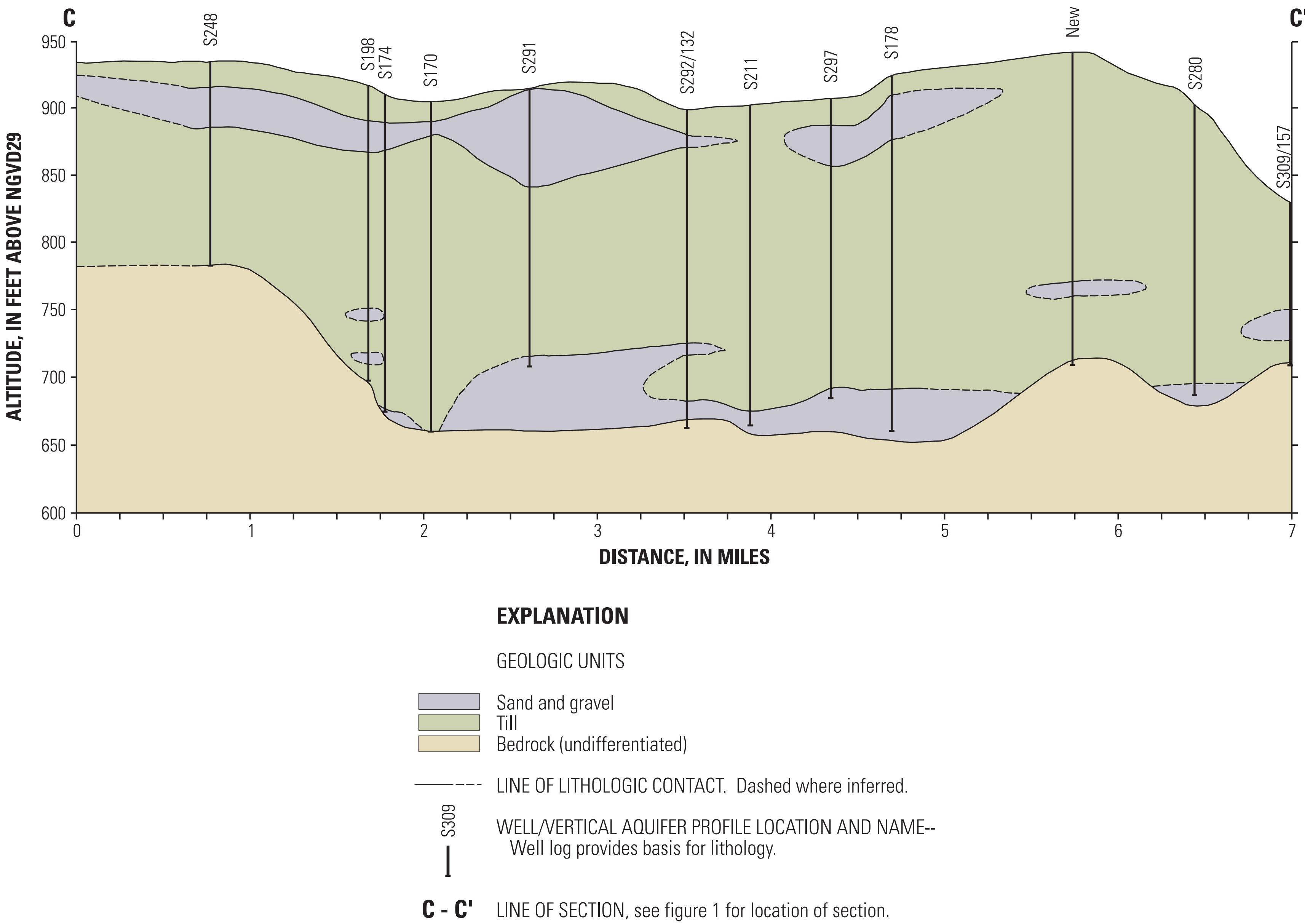


Figure 12C. Geologic section C-C′ through the glacial drift, Campton Township, Illinois.

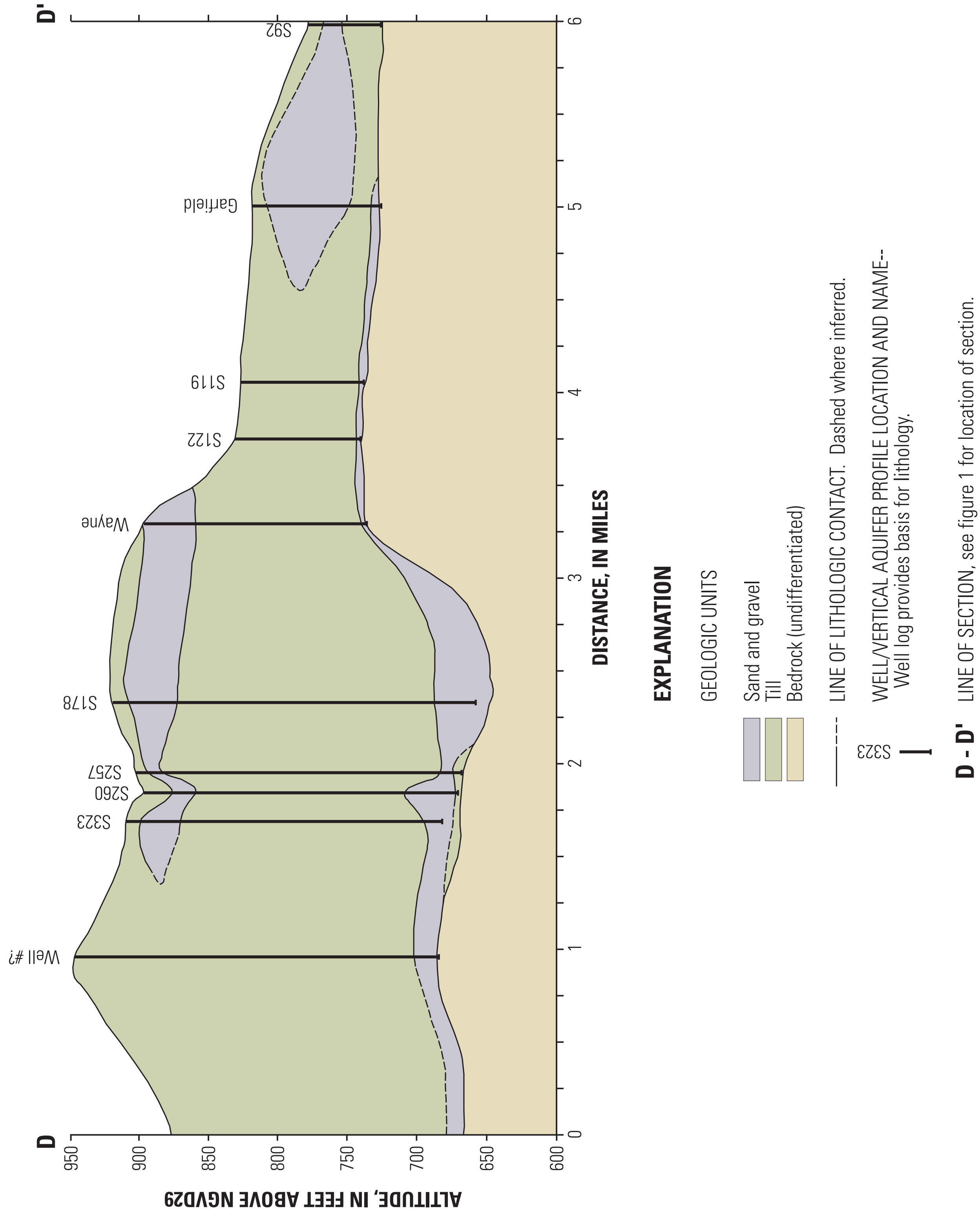


Figure 12D. Geologic section D-D' through the glacial drift, Campton Township, Illinois.

The Batestown Member of the Lemont Formation overlies the Tiskilwa Formation beneath much of the township and is composed of silt loam till with interspersed beds of sand (Grimley and Curry, 2001)(fig. 4). The Batestown Member is 0-35 ft thick and is the surficial geologic unit beneath much of the northeastern, central, and southern parts of the township.

The Henry Formation is composed predominately of sand and gravel with lenses of silt, clay, and till (fig. 4). The Henry Formation intertongues with the Batestown Member of the Lemont Formation, overlying the Tiskilwa Formation in parts of the township, and underlying the Batestown Member in other parts. The Henry Formation is as much as 150 ft thick beneath the township, and is thickest in the center of the township near Town Hall and Campton Hills Roads. The Henry Formation is the surficial geologic unit at scattered locations in the central, eastern, and southwestern parts of the township (Grimley and Curry, 2001). Areas where the Henry Formation is present at the land surface tend to be areas of locally elevated surface topography.

The Equality Formation is composed of lacustrine silt and clay with some beds of fine-to-medium sand (Graese and others, 1988; Grimley and Curry, 2001)(fig. 4). These units are commonly interbedded with the units of the Henry Formation. Where present, the Equality Formation is typically less than 20 ft thick and is the surficial geologic unit in much of the north-central, eastern, and southwestern parts of the township.

The Grayslake *Peat* is composed of peat, silt, and clay with some sand interbeds (fig. 4). Where present, these units are less than 10 ft thick (Graese and others, 1988). The Grayslake Peat is present at the land surface at a few scattered locations.

The Cahokia *Alluvium* (fig. 4) is composed of sand, silt, and clay deposited predominately along Ferson Creek and Mill Creek (Grimley and Curry, 2001). Where present, these units are 5-20 ft thick.

For the purposes of this report, unconsolidated geologic units were subdivided into two general categories, fine-grained units and coarse-grained units. Fine-grained units are composed predominately of till, with smaller amounts of silt and clay. Fine-grained units do not transmit water readily. Coarse-grained units are composed predominately of sand and gravel and transmit water readily. Interpretation of the well logs and data from previous investigations indicates that the sand-and-gravel units in the township can be generally subdivided into the shallow sand and gravel and the deep sand and gravel (figs. 12a, 12b, 12c, 12d, 13, and 14). The shallow sand-and-gravel units typically are overlain and underlain by fine-grained units and are composed primarily of the Cahokia, Equality, and Henry Formations, but likely include some discrete units in the upper part of the Tiskilwa Formation. The deep sand-and-gravel units typically directly overlie or are within 20 ft of the bedrock and are overlain by fine-grained deposits (figs. 12a, 12b, 12c, 12d). The deep sand-and-gravel units are composed primarily of deposits in the Glasford Formation and the lower part of the Tiskilwa Formation. The shallow and deep sand-and-gravel units typically are physically separated, except in the east-central part of the township where they are continuous in some areas (fig. 12a).

The shallow sand-and-gravel units are present in most of the eastern two-thirds of the township (fig. 13). These units typically are less than 25 ft thick, but may exceed 40 ft in thickness in some small areas. The deep sand-and-gravel units are present predominately in the center of the township (fig. 14). The deep sand-and-gravel units typically are less than 25 ft thick, but may exceed 40 ft in thickness north of Lake Campton. The deep sand-and-gravel units tend to be present and thickest in the bedrock valleys, particularly the area defined by the 650 ft contours on figure 11. Sand-and-gravel units typically are within about 25 ft of the land surface beneath much of the township, but are substantially deeper in the western part where the shallow sand-and-gravel units are absent (figs. 12b and 12d).

HYDROLOGY

Five aquifers are used for residential-water supply in Campton Township: the shallow glacial drift aquifers, the deep glacial drift aquifers, the Silurian-Maquoketa aquifer, the Galena-Platteville aquifer, and the Ancell aquifer (fig. 4). The various aquifers in the glacial drift commonly are separated by a discontinuous *confining unit*, which restricts flow among these aquifers. Though not used for water supply, the surface-water-bodies also are important hydrologic features in the township.

Surface Water

The primary surface-water bodies in the township are Virgil Ditch (gaging stations on Virgil Ditch have a VD prefix)(fig. 15), which drains the northwestern part of the township, Ferson Creek (gaging stations on Ferson Creek have a FC prefix), which drains the north-central part of the township, and Mill Creek (gaging stations on Mill Creek have a MC prefix), which drains the south-central part of the township. Mill and Ferson Creeks are peren-

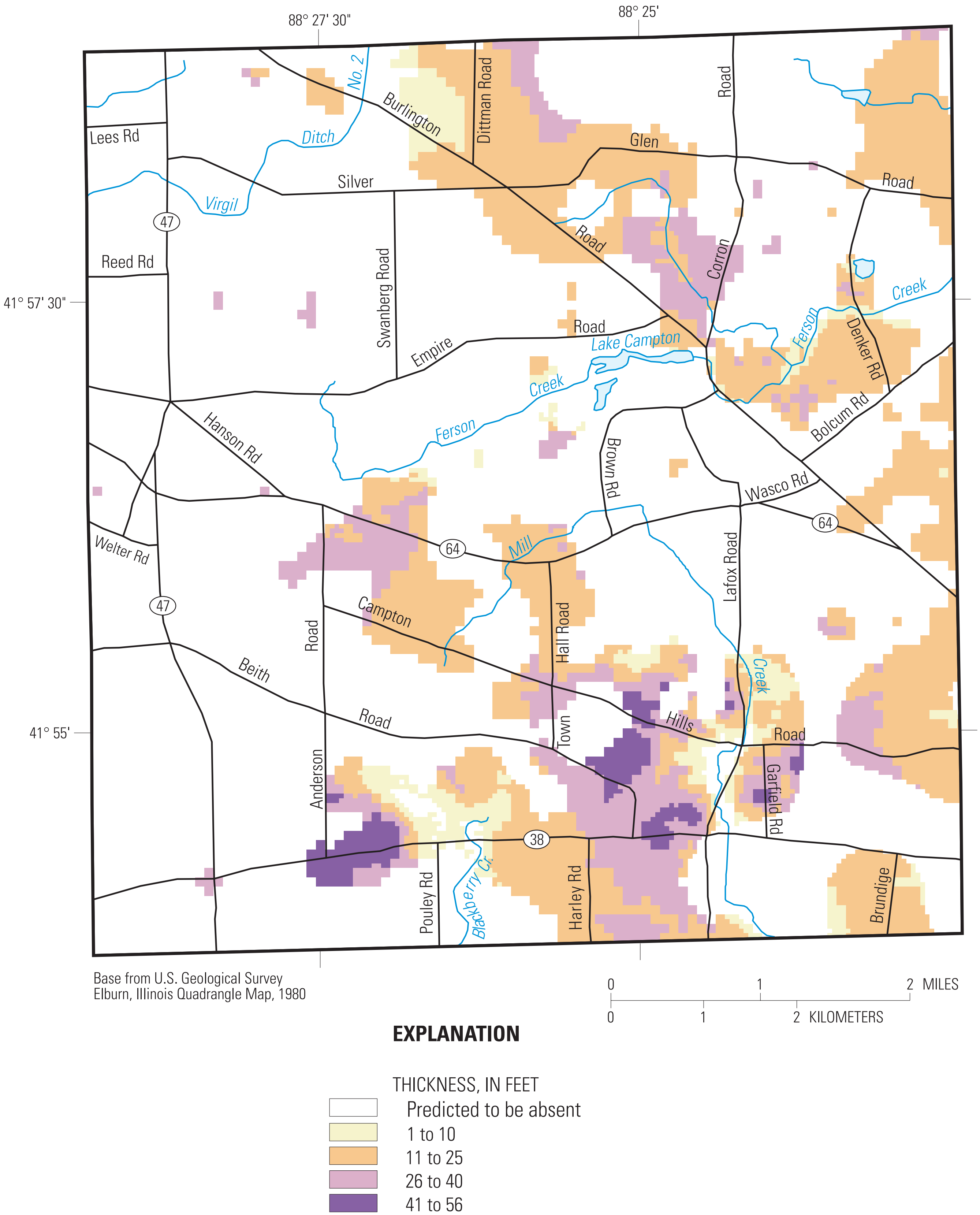
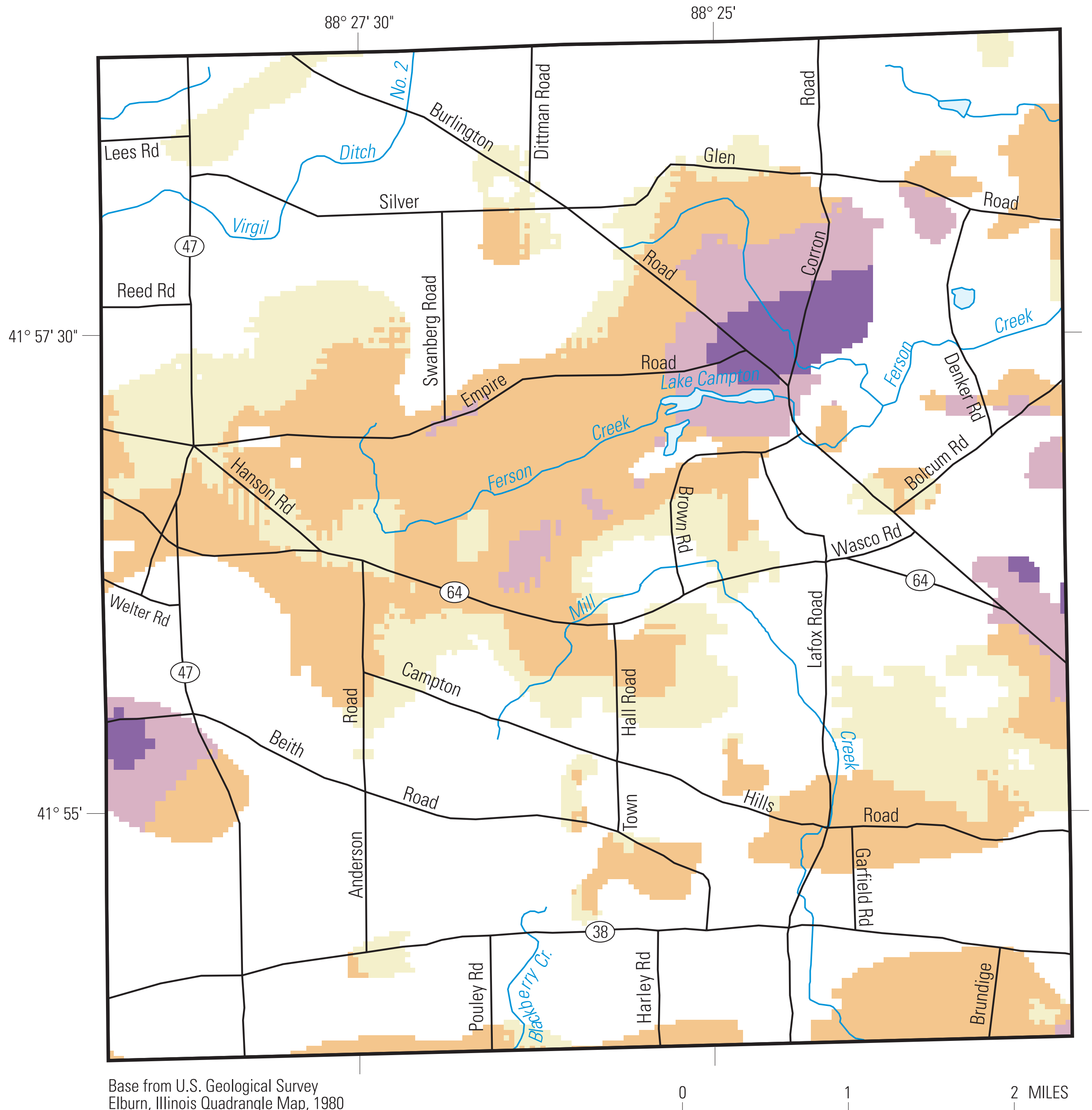


Figure 13. Location and thickness of the shallow glacial drift aquifers, Campton Township, Illinois.



EXPLANATION

THICKNESS, IN FEET

- Predicted to be absent
- 0 to 10
- 11 to 25
- 26 to 40
- 41 to 100

Figure 14. Location and thickness of the deep glacial drift aquifers, Campton Township, Illinois.

nial streams, and the part of Virgil Ditch within the township typically contains water. Those portions of Blackberry Creek and Otter Creek within the township typically do not contain water except during periods of high runoff such as during snowmelt and after storms. Streamflow measurements were collected monthly from July through October 2002 at three stations on Virgil Ditch, four stations on Ferson Creek, and five stations on Mill Creek (fig. 15). Ferson Creek receives wastewater effluent about 300 ft downstream of station FCRC and Mill Creek receives runoff from an area where wastewater is applied to the land surface by use of sprinklers about 0.5 mi east of station MCLFR.

Streamflow measurements indicate little or no flow in Virgil Ditch during most of the period of measurement, with July being the only period of flow above 1.0 ft^3/s. There was little measurable increase in the amount of flow downstream during August, September, and October (table 3). These results indicate there was little or no recharge from ground water to Virgil Ditch, except perhaps during July 2002.

Streamflow in Ferson and Mill Creek also was highest during the July measurement and was substantially lower in August through October. Flow in the creeks generally increased from upstream to downstream, indicating that the creeks are recharged by ground water in the township.

Confining Unit

The unconsolidated units in the township are composed primarily of fine-grained materials (primarily till, but some silt and clay) that form a confining unit that typically overlies, and commonly surrounds, the sand-and-gravel units that make up the glacial drift aquifers (figs. 4, 12a, 12b, 12c, 12d). The confining unit does not transmit water readily. For example, a long-term yield of 5 to 6 gal/min was reported for a 43-in.-diameter well drilled to a depth of 71 ft and drawing water from predominately silt and clay units north of Lake Campton (Benson, 1990). A horizontal hydraulic conductivity (Kh) value of 0.51 ft/d was calculated at a monitoring well open to silt and clay in the northeastern part of the township (William Morrow, U.S. Geological Survey, written commun., 2004). This value exceeds the upper range of the estimated Kh for glacial till in Kane County (0.000028-0.028 ft/d) (Graese and others, 1988). Because the confining unit does not transmit water readily, and is composed of materials that can adsorb some types of contamination, this unit can be effective in reducing the effects of contamination on the underlying aquifers.

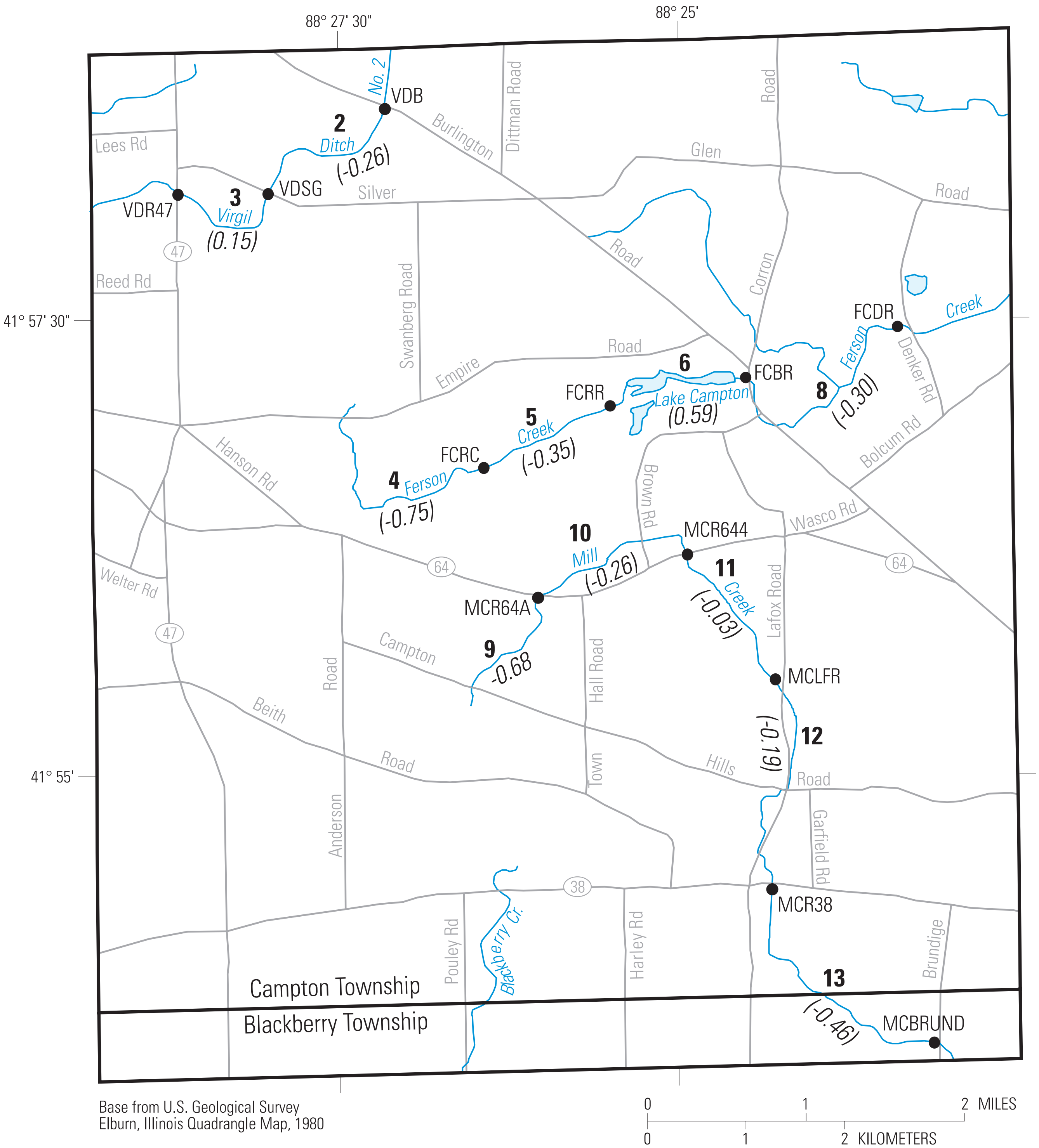
Shallow Glacial Drift Aquifers

Shallow, often discontinuous sand-and-gravel units in the Lemont, Cahokia, Equality, and Henry Formations compose the shallow glacial drift aquifers. The shallow glacial drift aquifers are not used extensively for water supply in Campton Township, but do supply water to some residences in the north-central, central, and east-central parts of the township. The top of these aquifers is often less than 20 ft below land surface (fig. 12a, 12b, 12c, 12d),

Table 3. Summary of steamflow data, Campton Township, Illinois, July-October 2002.

[Bold denotes impeded flow]

Location	Station name (fig. 15)	Drainage area (square miles)	Streamflow (cubic feet per second)			
			Date			
			7/12/02	8/9/02	9/12/02	10/15/02
Virgil Ditch at Burlington Road	VDB	1.48	0.00	0.04	0.02	0.00
Virgil Ditch at Silver Glen Road	VDSG	2.89	1.26	.009	.08	.08
Virgil Ditch at Route 47	VDR47	3.93	1.10	.02	.005	.01
Ferson Creek at Retreat Court	FCRC	3.43	1.54	.61	1.10	.81
Ferson Creek near Campton Ridge Road	FCRR	5.92	3.54	.71	1.04	.75
Ferson Creek at Burlington Road	FCBR	6.15	1.76	.87	1.30	1.04
Ferson Creek and Denker Road	FCDR	10.2	2.84	1.31	2.53	2.06
Mill Creek near Town Hall Road	MCR64A	2.27	.96	.74	.35	.69
Mill Creek at Highway 64	MCR64	3.04	.87	.67	.38	1.36
Mill Creek at LaFox Road	MCLFR	5.71	1.60	.96	1.19	1.08
Mill Creek at Route 38	MCR38	8.15	2.92	2.04	1.82	1.47
Mill Creek at Brundige Road	MCBRUND	11.86	5.48	2.87	2.44	2.65



EXPLANATION

● VDB STREAMFLOW-MEASUREMENT SITE AND IDENTIFIER

13 STREAMFLOW-MEASUREMENT SECTION NUMBER

(-0.46) STREAMFLOW RESIDUAL, IN FT^3/SEC

Figure 15. Location of streamflow-measurement sites, sections, and residuals, Campton Township, Illinois.

making the aquifers susceptible to drought and near-surface contamination.

Small, discontinuous units of sand or gravel in the Equality Formation, the Cahokia Alluvium, and the Greyslake Peat located primarily in the north-central and south-central parts of the township (fig. 13) are productive enough to be used for residential supply. Shallow sand-and-gravel units of the Henry Formation in the eastern part of the township may have the capacity to supply sufficient water for public supply (Curry and Seaber, 1990), with *yields* of 50-500 gal/min reported from wells utilizing the Henry Formation in Kane County (Graese and others, 1988).

The shallow glacial drift aquifers are recharged by precipitation at the land surface in and near the township. Water levels in the aquifers fluctuate by more than 10 ft annually, primarily in response to the amount and timing of precipitation (Graese and others, 1988). The direction of horizontal ground-water flow in these aquifers generally follows the land-surface gradient.

Data collected from seven wells open to shallow glacial drift aquifers in Campton Township during the summer of 2002 indicate the depth to water ranged from about 5 to 80 ft below the land surface (appendix A). These aquifers are not continuous throughout the township, so water levels in the aquifers were not contoured; however, water-level altitudes in shallow drift aquifers ranged from about 787 to 923 ft (fig. 16) in 2002 and generally showed a close relation to the altitude of the land surface. These data are consistent with the results of measurements from wells open to shallow drift aquifers during the 1995 survey of water levels in Campton Township (Kay and Kraske, 1996). Comparison of water levels measured from six wells open to shallow drift aquifers in 1995 with water levels measured from those wells in 2002 indicated less than 3 ft of variation between the two measurement periods. The small changes in water levels measured at these wells give no indication that volume of water being removed from the shallow glacial drift aquifers by pumping exceeds the volume being replaced by inflow from outside the township and recharge from precipitation. Additional withdrawals for residential and perhaps public water supply from shallow glacial drift aquifers may be feasible in the southeastern part of the township. However, the quality of water from these aquifers should be considered prior to their increased utilization.

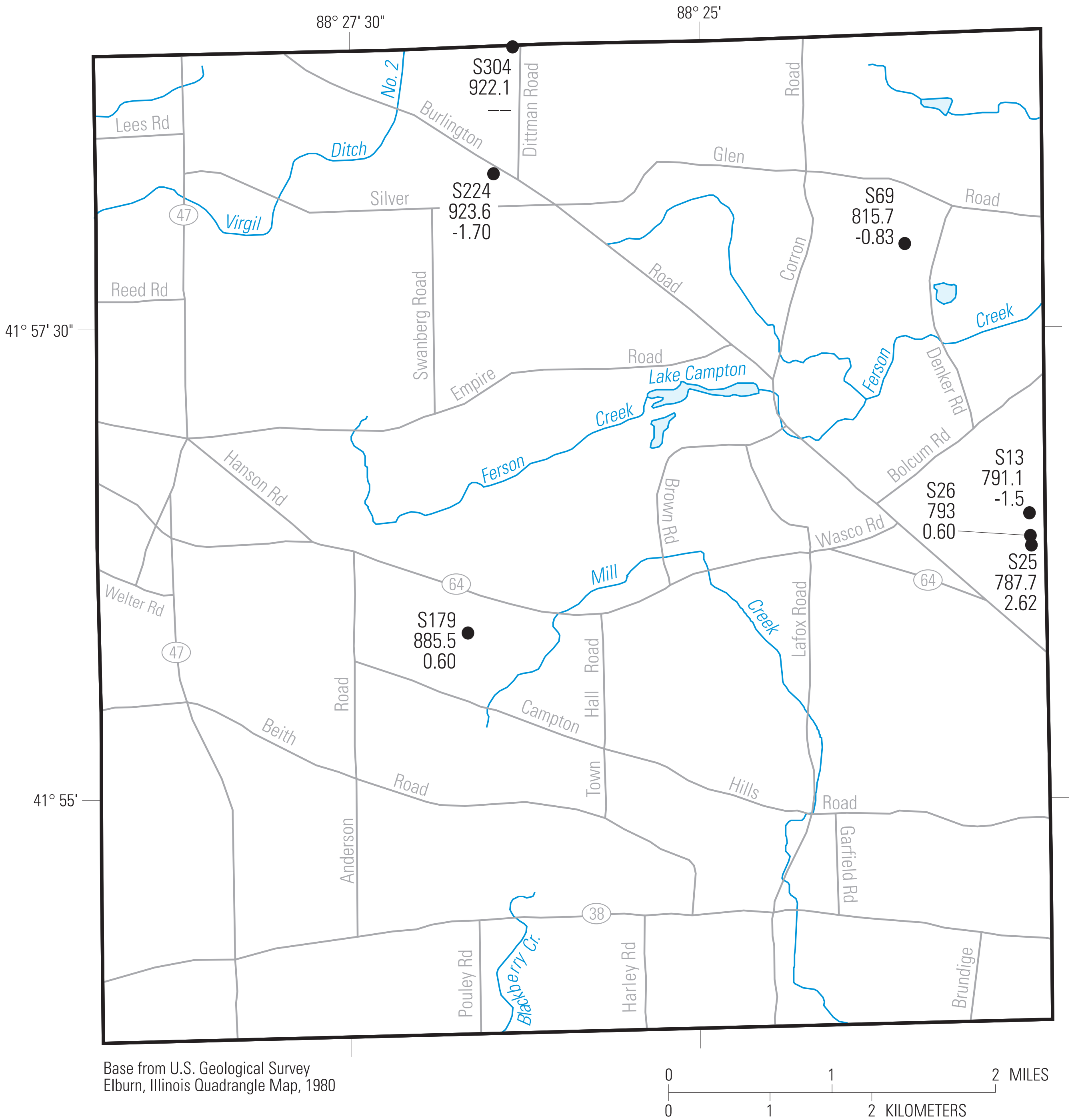
Deep Glacial Drift Aquifers

The deep, commonly discontinuous sand-and-gravel units embedded in the fine-grained glacial units of the Tiskilwa and Glasford Formations, and the massive, more continuous basal sands of the Glasford Formation compose the deep glacial drift aquifers underlying the township (fig. 4). Deep glacial drift aquifers are the primary drift aquifers used for residential supply in the township and are recharged by flow through the discontinuous confining unit and possibly the underlying bedrock units in some areas. Differentiation of shallow and deep glacial-drift aquifers can be difficult. However, data collected for this investigation indicate that wells open to the shallow glacial drift aquifers are drilled to an altitude above 725 ft NGVD29, whereas wells open to the deep glacial drift aquifers typically are drilled to an altitude below 725 ft NGVD29 (appendix A).

Many of the wells that draw water from the deep glacial drift aquifers are located within the bedrock valley in the eastern and central parts of the township (and beyond), where the aquifers can exceed 40 ft in thickness (figs. 4, 14). The deep glacial drift aquifers commonly directly overlie the bedrock where the aquifers are present. These aquifers are overlain by the discontinuous confining unit, which typically exceeds 75 ft in thickness. The presence of the overlying discontinuous confining unit makes the deep glacial drift aquifers less susceptible to contamination than the shallow glacial drift aquifers.

The hydraulic properties of the deep glacial drift aquifers vary with lithology. Yields of 50-500 gal/min have been reported from wells drawing from the deep glacial drift aquifers in Kane County (Graese and others, 1988) and some of the thicker, more continuous deep glacial drift aquifers within the bedrock valleys in the eastern part of the township may be capable of yielding sufficient water for public supply (Curry and Seaber, 1990). Thinner, discrete deep glacial drift aquifers embedded in the confining unit are capable of yielding water in sufficient quantities for residential use, typically between 10 and 20 gal/min. Constant-discharge aquifer tests performed as part of this investigation in six residential-supply wells drawing from deep glacial drift aquifers in the township yielded *transmissivity* estimates from 4.7 to 2,350 ft^2/d (table 2), with a geometric mean value of 140 ft^2/d. Horizontal-hydraulic conductivity estimates from these tests ranged from 0.47 to 94 ft/d, with a geometric mean value of about 13 ft/d.

Hydraulic properties of both shallow and deep glacial drift aquifers were calculated based on



EXPLANATION

S13
791.1
-1.5

WELL LOCATION--Top number is well name. Middle number is altitude of water level in well, in feet above NGVD29. Bottom number is change in water level (-, indicates decrease) since 1995 measurement, in feet.

——, indicates no 1995 measurement available.

Figure 16. Water levels and water-level change in wells completed in the shallow glacial drift aquifers, Campton Township, Illinois, June-July 2002.

analysis of data collected from constant-discharge aquifer tests conducted in 20 production wells drawing from glacial drift aquifers in the townships adjacent to Campton Township (Scott Meyer, Illinois State Water Survey, written commun., 2002). Transmissivities ranged from about 680 to 72,500 ft^2/d, with a geometric mean value of about 8,400 ft^2/d. Kh values ranged from about 32 to 970 ft/d with a geometric mean value of about 250 ft/d. Because most of these tests were performed in high-capacity production wells, it is probable that the most permeable parts of the aquifers were tested and these results represent the upper range of values for the glacial drift aquifers in the township.

Water-level data collected from 62 wells open to deep glacial drift aquifers in Campton Township during the summer of 2002 indicates water-level altitudes ranged from about 747 to 863 ft NGVD29 (appendix A, fig. 17). These aquifers are not continuous throughout the township, so water levels in these aquifers were not contoured; however, water-level altitudes in the wells open to deep glacial drift aquifers typically were higher than 800 ft in the western part of the township, between 775 and 800 ft in the central part of the township, and below 775 ft in much of the southern and eastern parts of the township (fig. 17). These patterns are consistent with those identified from analysis of water-level data collected from these aquifers in 1995 (Kay and Kraske, 1996) and are generally similar to patterns in land-surface topography (fig. 2).

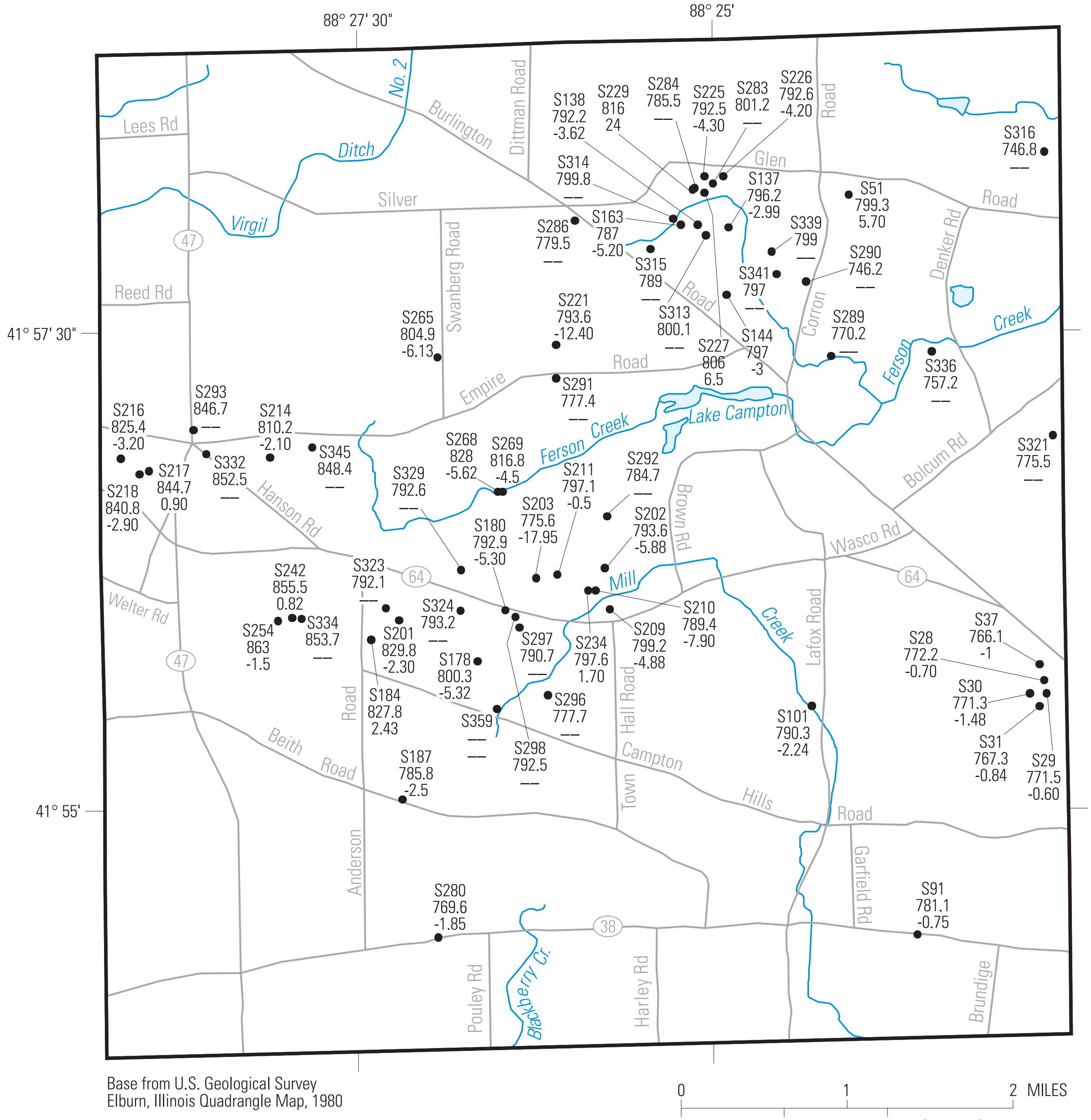
Comparison of water levels measured in 37 wells open to deep glacial drift aquifers in 1995 with water levels measured in those wells in 2002 indicated water levels declined by more than 7 ft in one well and rose by more than 7 ft in three wells. Of the remaining 33 wells, water levels in 6 declined by less than 7 ft and water levels in 26 increased by less than 7 ft. Water-level data do not indicate the widespread distribution of large declines in water level within the deep glacial drift aquifers. These aquifers appear to have the capacity to sustain additional withdrawals in most of the township. However, water-level declines of about 4 to 18 ft measured in the area southwest of Lake Campton (fig. 17) may warrant further monitoring. Water-level altitudes in the shallow glacial drift aquifers typically are above those in the deep glacial drift aquifers, indicating the potential for downward flow between these units.

Silurian-Maquoketa aquifer

The Silurian-Maquoketa aquifer underlies the entire township and comprises dolomite of the Alexandrian Series and dolomite and shales of the Maquoketa Group (fig. 4). This aquifer is hydraulically connected to the overlying deep glacial drift aquifers where the two are present (Visocky and Schulmeister, 1988), but it is typically *confined* by fine-grained glacial drift beneath the township.

The Silurian-Maquoketa aquifer can produce water in sufficient quantities to support residential use by individual households in Campton Township if the part of the aquifer intercepted by the well is sufficiently fractured. Ground-water flow though the Silurian-Maquoketa aquifer is predominately through the fractures concentrated in the upper 50 ft of the bedrock surface, where it is composed of weathered dolomite (Csallany and Walton, 1963), and potentially through fractures in the dolomite units deeper within the Maquoketa Group. Specific-capacity data from wells completed in shallow dolomite in northern Illinois (Csallany and Walton, 1963) indicate yields may be highest in wells completed in the weathered, upper part of the aquifer where it is composed of dolomite, in areas of bedrock highs, and where the bedrock is overlain by a deep glacial drift aquifer. Shale units in the Maquoketa Group do not typically transmit large amounts of water, even where the shale is at the bedrock surface (Schumaker, 1990).

The hydraulic properties of the Silurian-Maquoketa aquifer reported by previous investigators in and near Kane County range over four orders of magnitude and are affected primarily by the number, size, and degree of connection of the fractures and weathering features in the aquifer. The Kh of the Silurian-Maquoketa aquifer in Kane County ranges from 2.8×10^{-3} to 2.8×10^{1} ft/d for the Silurian dolomite (Graese and others, 1988). The Kh of the Silurian-Maquoketa aquifer in Kane County ranges from 2.8×10^{-3} to 2.8×10^{0} ft/d where the Maquoketa dolomite is present, and may be as high as 2.8×10^{1} ft/d where Maquoketa dolomite is near the bedrock surface (Graese and others, 1988). The Kh of fractured shale in the upper 10 ft of the Silurian-Maquoketa aquifer in Kane County is typically less than 2.8×10^{-2} ft/d, whereas the Kh of unfractured shale is typically less than 2.8×10^{-3} ft/d. The vertical hydraulic conductivity of the Silurian-Maquoketa aquifer has been estimated at about 2.8×10^{-6} ft/d in northern Illinois (Schumaker, 1990). Constant-discharge aquifer tests performed as part of this investigation in five residential-supply wells open to the Silurian-Maquoketa aquifer in the township yielded transmissivity estimates ranging from 4.9 to 8,370 ft^2/d (table 2), with a geometric mean value of 260 ft^2/d. Horizontal-hydraulic conductivity estimates from these tests ranged from 0.016 to



EXPLANATION

S280
769.6
-1.85

WELL LOCATION--Top number is well name. Middle number is altitude of water level in well, in feet above NGVD29. Bottom number is change in water level (-, indicates decrease) since 1995 measurement, in feet.
——, indicates no 1995 measurement available.

Figure 17. Water levels and water-level change in wells completed in the deep glacial drift aquifers, Campton Township, Illinois, June-July 2002.

93 ft/d, with a geometric mean value of about 2.3 ft/d.

Data collected by the ISWS during a survey of ground-water levels in north-central Illinois in 1987 indicated that water levels in wells completed in the Silurian-Maquoketa aquifer decreased from northwest to southeast in Kane County and were lower in parts of Campton Township than in the surrounding area (Visocky and Schulmeister, 1988). The lower water levels may indicate that more water is withdrawn from the aquifer in some parts of Campton Township than is replaced by recharge from overlying hydrologic units.

Water levels were measured periodically in monitoring wells open to the Silurian-Maquoketa aquifer in Kane County from December 1984 through May 1987 (Visocky and Schulmeister, 1988). Water levels in these wells typically varied by less than 10 ft. Water levels typically were highest in the spring, and declined through the remainder of the year, presumably in response to seasonal variations in pumping and recharge from overlying hydraulic units.

Water levels also were measured in well S360 completed in the Silurian-Maquoketa aquifer (table 2) as part of this investigation; the measurements were made approximately monthly from March 2003 through March 2004 and approximately quarterly through September 2004 (fig. 18). This well is in an agricultural area and it is assumed that water levels in this part of the aquifer are not substantially affected by nearby pumping. Precautions were taken to ensure that the well had not been pumped at least 15 minutes prior to measurement, and aquifer testing performed at the well indicates that drawdown in the well is less than 0.4 ft even while the well was being pumped. Water levels in well S360 fluctuated by about 4 ft during March 2003-September 2004. Water levels were generally related to the amount of precipitation, being higher following precipitation events and lower during periods of less precipitation.

Analysis of the long-term water-level data collected by Visocky and Schulmeister (1988) and in well S360 as part of this investigation indicates that water-level fluctuations of approximately 10 ft over a period of 5-10 years can likely be attributed to natural fluctuations in inflow and outflow of the Silurian-Maquoketa aquifer. Water-level declines in excess of 10 ft can potentially be attributed to the effects of pumping from the aquifer. Long-term water-level data also indicate that water levels in May through July of a year with no drought or flooding are probably representative of typical to higher than typical water levels for the year as a whole.

Data collected from 115 wells open to the Silurian-Maquoketa aquifer in Campton Township in June and July 2002 indicate that water-level altitudes ranged from 653 to 864 ft (figs. 19 and 20). Water-level altitudes were above 800 ft in the west-central part of the township and in a small area in the south-central part of the township; water-level altitudes were between 760 and 800 ft in most of the center of the township, were less than 760 ft in most of the eastern part of the township, and were below 720 ft in the northeastern part of the township. These patterns are consistent with those identified from water-level measurements in May and June 1995 (Kay and Kraske, 1996) and are generally consistent with water-level patterns in the deep glacial drift aquifers (fig. 17) and with surface topography (fig. 2). Ground water flows in the direction of declining water levels. Therefore, the general direction of ground-water flow in the Silurian-Maquoketa aquifer underlying the township is from west to east.

The *potentiometric surface* of the Silurian-Maquoketa aquifer appears to be affected by the topography of the bedrock surface in some areas. The high potentiometric surface defined by the 790- to 810-ft contours in the south-central part of the township (fig. 19) approximately

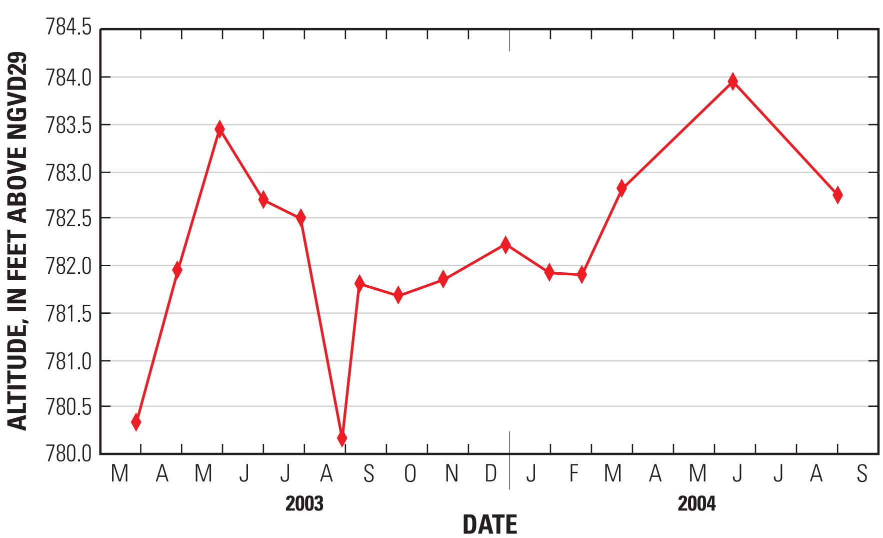
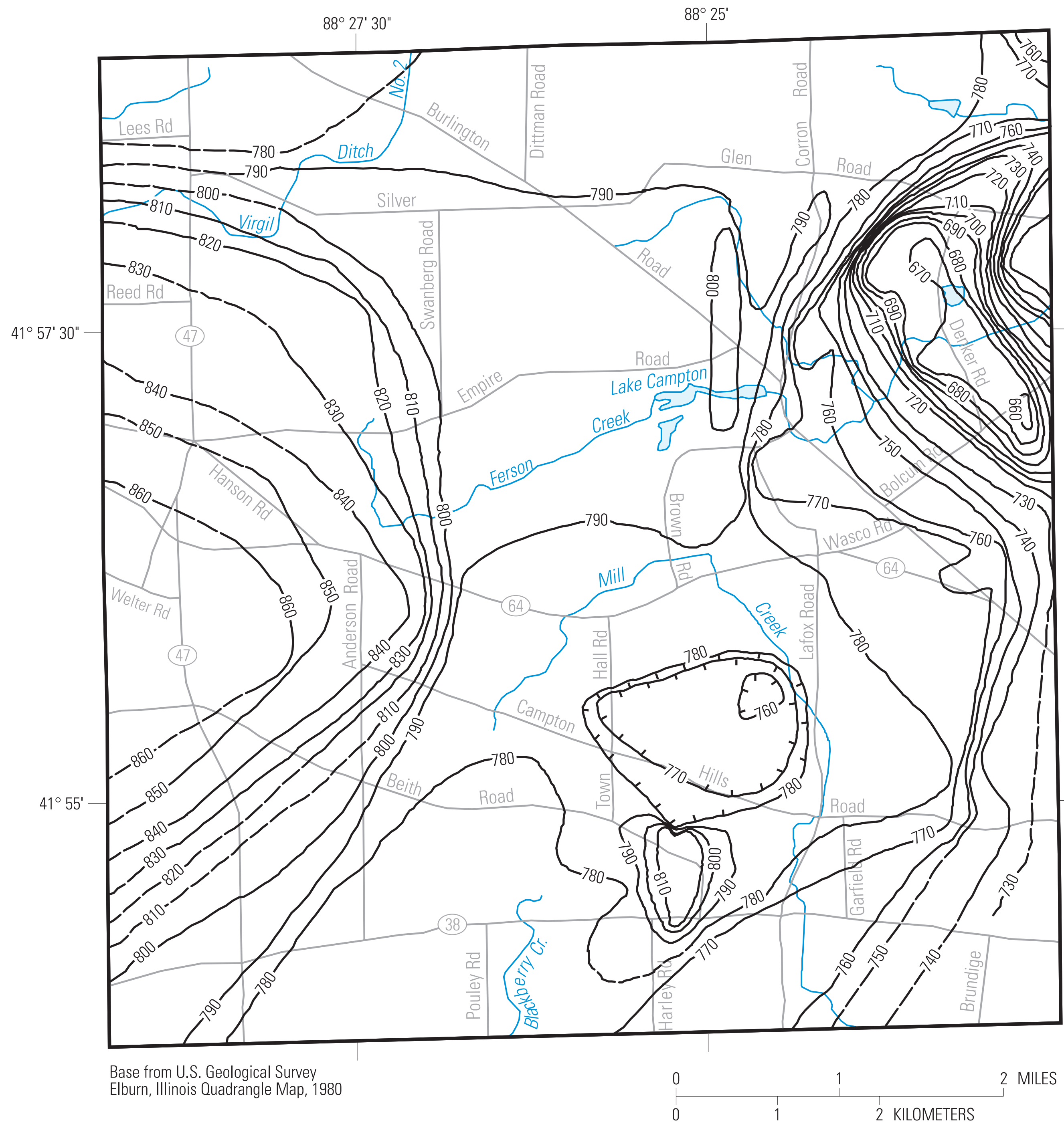


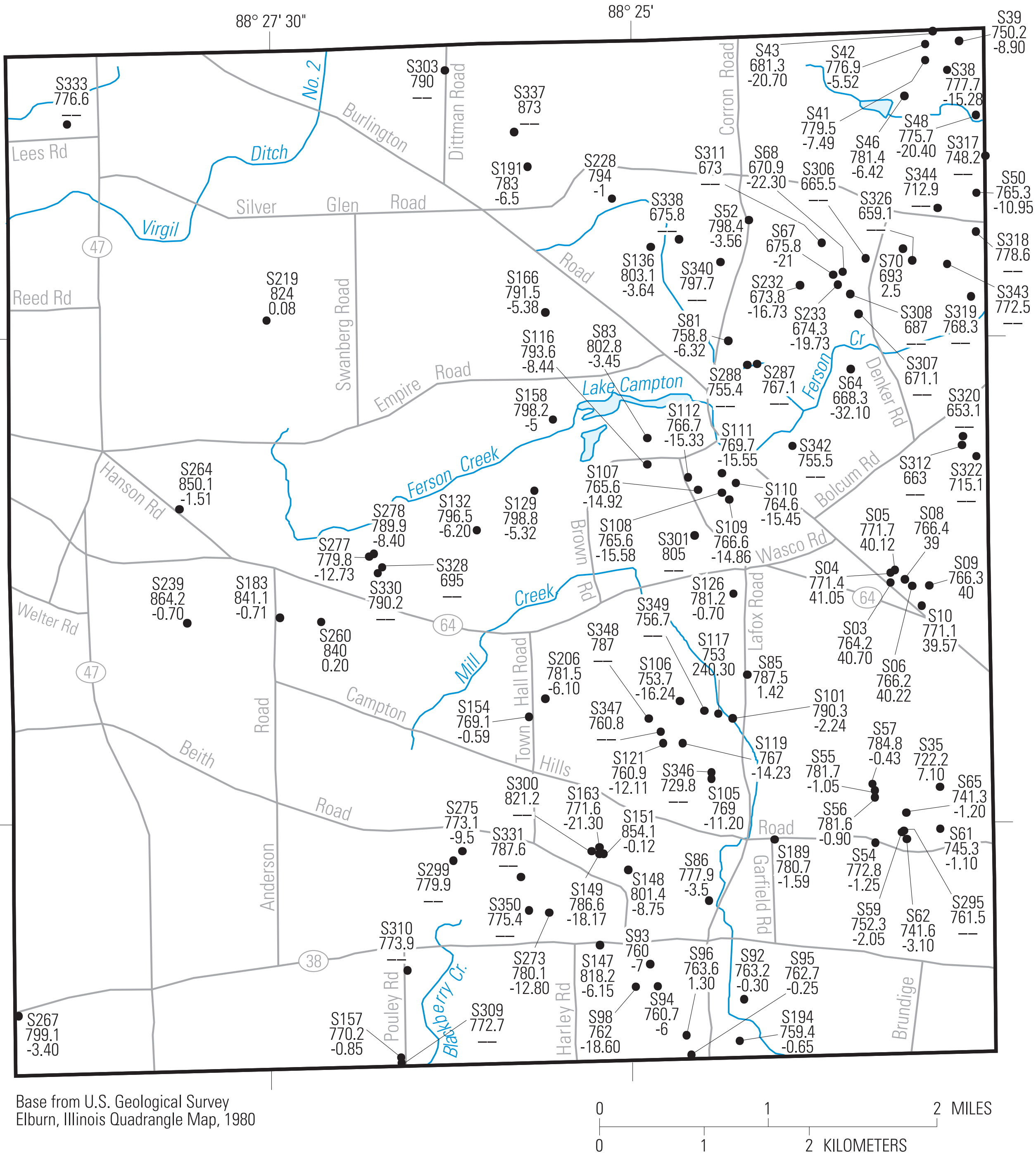
Figure 18. Water level in well S360 open to the Silurian-Maquoketa aquifer, Campton Township, Illinois, March 28, 2003-September 1, 2004.



EXPLANATION

—720— LINE OF EQUAL WATER-LEVEL ALTITUDE, IN FEET ABOVE NGVD29--
Contour interval 10 feet. Line dashed where approximate.

Figure 19. Potentiometric surface of the Silurian-Maquoketa aquifer, Campton Township, Illinois, June-July 2002.



EXPLANATION

S267
799.1
-3.40

WELL LOCATION--Top number is well name. Middle number is altitude of water level in well, in feet above NGVD29. Bottom number is change in water level (-, indicates decrease) since 1995 measurement, in feet.
—, indicates no 1995 measurement available.

Figure 20. Water levels and water-level change in wells completed in the Silurian-Maquoketa aquifer, Campton Township, Illinois, June-July 2002.

corresponds to the location of a bedrock mound defined by the 750 ft contours in figure 11. The low potentiometric surface defined by the 630- to 750-ft contours in the northeastern corner of the township approximately corresponds to an area where the bedrock surface is low because of the presence of the Elgin bedrock valley. However, some of the highest water levels in the Silurian-Maquoketa aquifer, for example, the area defined by the 790- to 860-ft contours in the western and central parts of the township, are associated with low areas in the bedrock surface (fig. 11). The high water levels in the Silurian-Maquoketa aquifer in the central part of the township may be due to inflow from deep glacial drift aquifers in this part of the township (fig. 14), as well as the effects of surface topography.

Water-level contours indicate the presence of a composite cone of depression defined approximately by the 750-ft altitude contour in the northeastern part of the township and the 780-ft contour in the south-central part of the township. The low water-level altitude in these areas cannot be attributed solely to variations in bedrock-surface topography. Land use in each of these areas is residential. Most of these residences were built prior to 1978 and the density of wells completed in the Silurian-Maquoketa aquifer exceeds about 0.66 well per acre in some of these areas (Kay and Kraske, 1996). Close examination of the water-level contours indicates that the water-level altitude tends to be lowest in the wells at the center of a residential area and highest in wells at the edges (compare figures 3 and 19). For example, the 770-ft contour shows a deflection near the edge of residential areas north of Wasco Road and east of LaFox Road, as well as in the area near the intersection of Burlington Road and Route 64.

Comparison of water-level altitudes measured in 75 wells open to the Silurian-Maquoketa aquifer in 1995 with altitudes in those wells in 2002 indicated that water levels declined by more than 10 ft in 24 wells, declined by less than 10 ft in 42 wells, rose by less than 2 ft in 2 wells, and rose by about 40 ft in 7 wells (fig. 20). The wells with the largest rise in water level are all located in the east-central part of the township near Burlington Road and Route 64. The cause of this large rise is unknown. Comparison of the 1995 and 2002 water-level data indicates that water levels in the Silurian-Maquoketa aquifer declined during this period in almost the entire township. Water-level declines in excess of 10 ft tended to be concentrated in the eastern part of the township (fig. 20), particularly in locations where the water-level contours indicated either a composite cone of depression or where deflections in the contours around subdivisions were observed. All of these data indicate that the amount of water being withdrawn by pumping from the Silurian-Maquoketa aquifer in much of Campton Township exceeds the amount of water being added by inflow from outside the township and recharge from other hydrologic units. Additional utilization of this aquifer should be considered with care.

Comparison of water-level altitudes in the deep glacial drift aquifers (fig. 17) with water levels in the Silurian-Maquoketa aquifer (fig. 20) at a given location indicates that water-level altitudes in these aquifers differ by less than 10 ft in most of the township. The small difference in water level between these aquifers indicates they may be hydraulically connected. Flow between the glacial drift and Silurian-Maquoketa aquifers appears to be complex within the township, with the movement of water between the aquifers being affected by bedrock topography and pumping from the Silurian-Maquoketa aquifer.

Galena-Platteville Aquifer

The Galena-Platteville aquifer underlies the entire township and is composed of the Galena-Platteville dolomite (fig. 4). This unit typically is considered to be a confining unit in northeastern Illinois. However, the Galena-Platteville dolomite yields enough water for residential supply within the township, typically in conjunction with the overlying Silurian-Maquoketa aquifer. Therefore, the Galena-Platteville dolomite is considered an aquifer for the purposes of this report. The Galena-Platteville aquifer is composed of massive dolomite with some fractures and vugs, which provide ground water to wells. The amount of water yielded to a well is affected by the number, size, and degree of connection of the fractures and vugs the well intercepts. The Galena-Platteville aquifer has low permeability beneath most of northeastern Illinois and most of the water withdrawn from residential wells completed in this aquifer is derived from well-bore storage that is slowly replaced from the aquifer when pumping has ceased.

The Kh of the Galena-Platteville aquifer in Kane County typically is less than 2.8×10^{-3} ft/d (Graese and others, 1988). An aquifer test on a 6-in.-diameter well open only to the Galena-Platteville aquifer north of Lake Campton resulted in a specific-capacity estimate of 0.057 (gal/min)/ft and a long-term yield of 10 gal/min (Benson, 1990). Geophysical logs recorded during this study indicated the presence of slightly permeable fractures and vugs in the Galena-Platteville aquifer at wells S356, S357, and S358, but not at wells S302, S354, and

S355 (Roger Morin, U.S. Geological Survey, written commun., 2003)(appendix 2). Aquifer testing conducted in conjunction with geophysical logging during this study indicated the transmissivity of the Galena-Platteville aquifer was less than 25 ft^2/d in wells S302, S354, and S356 (table 1) (Roger Morin, written commun., 2003). Assuming an aquifer thickness of about 330 ft, this transmissivity yields a Kh of about 7.6 X 10^{-2} ft/d.

Data collected during a survey of water-level altitudes in north-central Illinois in 1987 indicated that ground-water level altitudes in wells used for residential supply, which were open to both the Alexandrian-Maquoketa and Galena-Platteville aquifers, were lower in much of the northern part of Campton Township than in the surrounding area (Visocky and Schulmeister, 1988). The lower water-level altitudes may indicate that more water is withdrawn from the Alexandrian-Maquoketa and Galena-Platteville aquifers in this part of the township than is replaced by inflow.

Water-level altitudes measured in monitoring wells open to the Galena-Platteville aquifer in Kane County during 1984-87 indicated the overall direction of flow in the aquifer was from west to east in Kane County. Water-level fluctuations in these wells that could be attributed to natural variation were less than 10 ft during this period (Visocky and Schulmeister, 1988).

Water-level altitudes were measured in 25 wells completed in the Galena-Platteville aquifer as part of this investigation in 2002 (appendix A). All of these wells are open holes from the top of the bedrock, which is composed of Silurian or Maquoketa units, to the bottom of the well, which is composed of the Galena-Platteville dolomite. Water-level altitudes in the wells completed in the Galena-Platteville aquifer range from about 477 to 780 ft (fig. 21). The large range in water levels in these wells is probably the result of differences in the vertical distribution of water levels in the aquifers penetrated by the well and the amount of inflow to the well from the Silurian-Maquoketa aquifer, rather than natural variations in the water level of the Galena-Platteville aquifer. Where the water-level altitude is low, for example, in well S248, it is probable that the well receives only small amounts of water from the Silurian-Maquoketa aquifer and the water level in the well is approximately representative of the water level of the Galena-Platteville aquifer. Where the water-level altitude is high, for example, in well S123, it is probable that the well receives large amounts of water from the Silurian-Maquoketa aquifer and the water level in the well is more representative of the water level of the Silurian-Maquoketa aquifer. The water level measured in most of these wells is a composite of the water level of both the Galena-Platteville and Silurian-Maquoketa aquifers. Because water-level altitudes in most or all of the wells open to the Galena-Platteville aquifer do not accurately represent the water level of the Galena-Platteville aquifer, the potentiometric surface of the aquifer cannot be contoured with these data. However, the water levels do represent the likely maximum value of the water level at that point in the Galena-Platteville aquifer.

Comparison of 1995 and 2002 water-level altitudes in the wells open to the Galena-Platteville aquifer indicates that water levels rose by more than 10 ft in 3 wells, rose less than 3 ft in 4 wells, declined less than 10 ft in 6 wells, and declined more than 10 ft in 3 wells (fig. 21). Two of the wells in which water-level altitudes rose more than 10 ft are located in the east-central part of the township, where water-level altitudes in the Silurian-Maquoketa aquifer also rose substantially since 1995, indicating that the water-level rise in both aquifers is related to changes in hydraulic conditions in the Silurian-Maquoketa aquifer. Areas of declining water-level altitudes had no consistent pattern, and the values likely reflect the difficulty of taking an accurate water-level measurement from these deep wells rather than a true indication of water-level trends.

The water-level altitude in the wells open to the Galena-Platteville aquifer typically is substantially less than that in the wells open to the Silurian-Maquoketa aquifer. This relation of water levels indicates the potential for downward flow of water from the Silurian-Maquoketa aquifer to the Galena-Platteville aquifer. The amount of water being exchanged between these aquifers, however, is likely to be small because of the low vertical hydraulic conductivity of the Galena-Platteville aquifer.

Ancell Aquifer

The Ancell aquifer underlies the entire township and consists of the St. Peter Sandstone and the sandstones of the Glenwood Formation (fig. 4). The Ancell aquifer is the uppermost aquifer in the *Cambrian*-Ordovician aquifer system, and is used for public water-supply in northern Illinois. The Ancell aquifer is the deepest aquifer penetrated by residential-supply wells in the township. Residential-supply wells typically penetrate only the upper 50 ft of the aquifer.

Many of the residential-supply wells open to the Ancell aquifer are cased through the unconsolidated units and are completed as open holes from

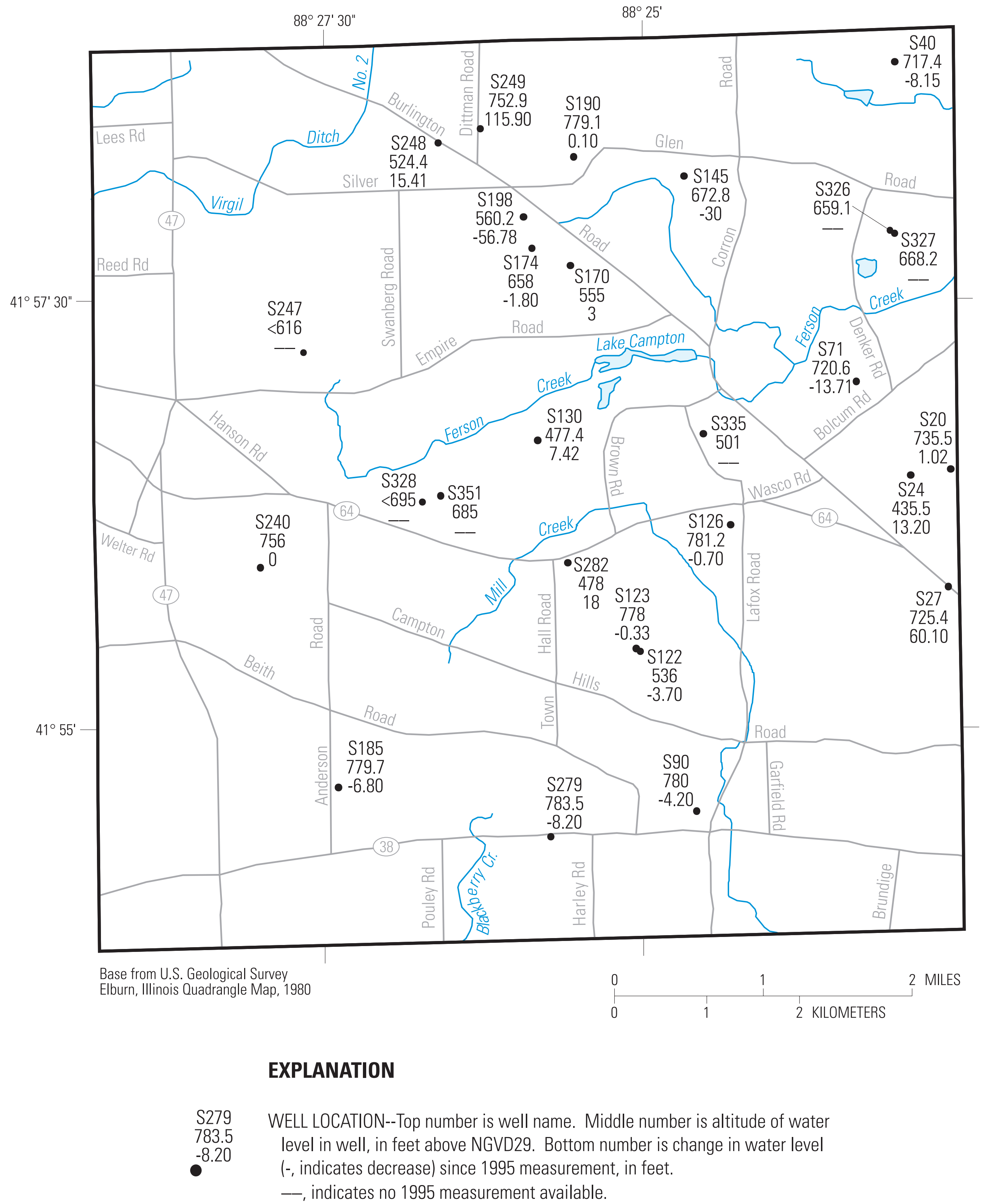


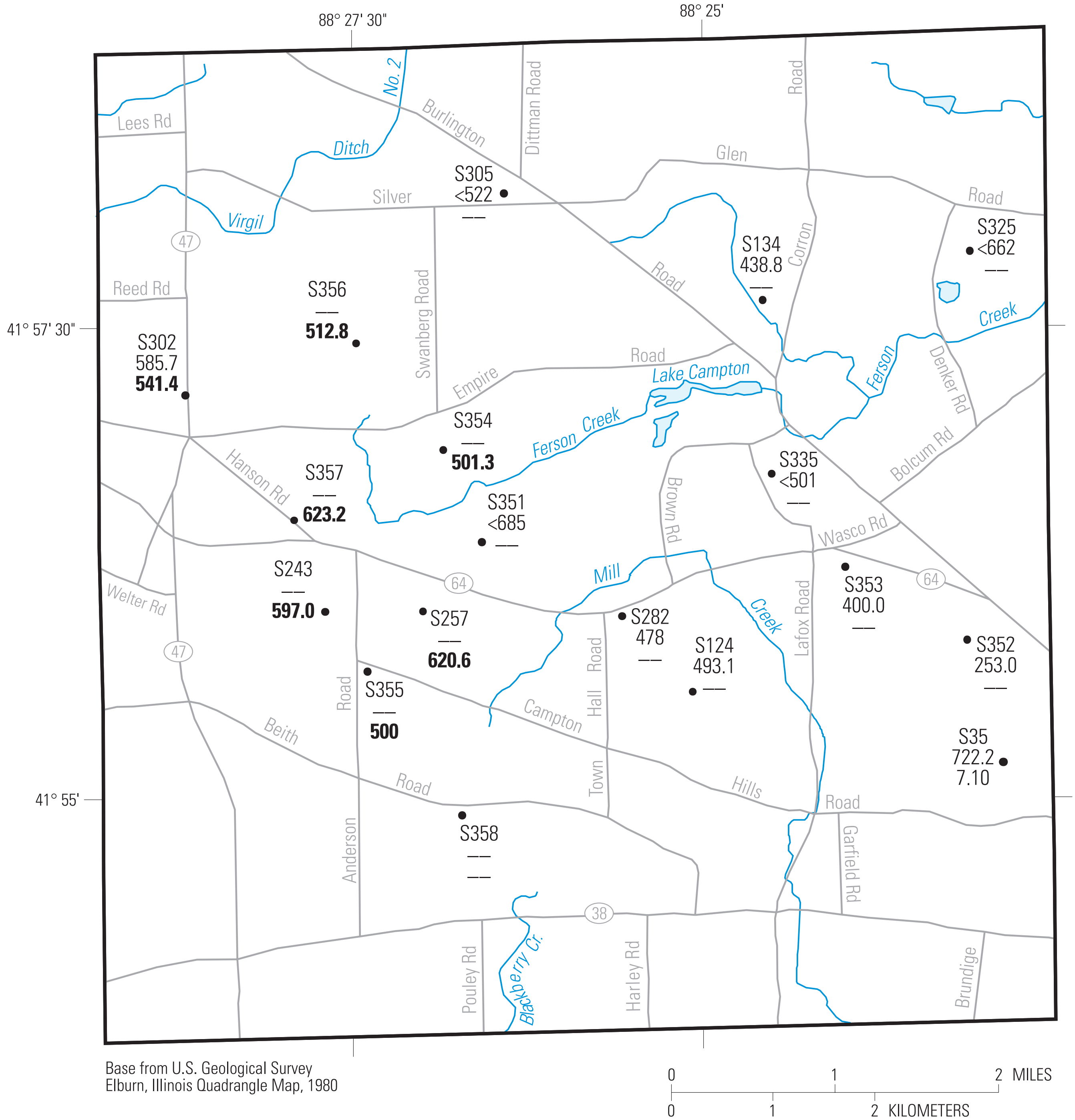
Figure 21. Water levels and water-level change in wells completed in the Galena-Platteville aquifer, Campton Township, Illinois, June-July 2002.

the Silurian-Maquoketa aquifer, through the Galena-Platteville aquifer, and into the top of the Ancell aquifer. Recent revisions to the well-construction codes in Kane County have required that new wells drilled into the Ancell aquifer be cased, though not necessarily sealed, through the upper part of the bedrock. Because wells open to the Ancell aquifer in the township are open over large intervals to multiple aquifers, vertical flow from the top of the open interval to the bottom of a well can be substantial. Flowmeter logging during this study identified about 10 gal/min of flow from the top of to the bottom of well S302 (table 1). Much smaller amounts of vertical flow were measured during flowmeter logging in wells S354 (less than 0.1 gal/min), S355 (about 0.1 gal/min), S356 (less than 0.1 gal/min), S357 (about 0.3 gal/min), and S358 (about 0.14 gal/min). All of these wells have been cased through the upper part of the bedrock and likely represent the minimum amount of vertical flow through a well that can be expected. Downward flow within a borehole has the effect of moving water out of the shallower aquifers, primarily the Alexandrian-Maquoketa, and into the deeper aquifers, primarily the Ancell but also the Galena-Platteville.

The Ancell aquifer yields between 50 and 200 gal/min to wells, making it the most consistently productive of the bedrock aquifers used for residential supply in Campton Township (Graese and others, 1988). A constant-discharge aquifer test conducted in a well open to the entire thickness of the Ancell aquifer about 3 mi southwest of the township yielded a transmissivity of about 2,800 ft^2/d, a Kh of 8.3 ft/d and a *storage coefficient* of 2.2×10^{-4} (Visocky and Schulmeister, 1988). Analysis of specific-capacity data obtained during production testing in wells S352 and S353 yielded an estimated transmissivity of about 260 ft^2/d and an estimated Kh of 1.0 ft/d. Transmissivity estimates for the Ancell aquifer obtained during aquifer testing and geophysical logging of wells S302, S354, S355, and S356 for this study were less than 30 ft^2/d (Roger Morin, written commun., 2003)(table 1). The smaller estimate of transmissivity (in comparison to the multiple-well test performed outside the township) obtained from the specific-capacity analysis of wells S352 and S353 is at least partly because of the effect of well loss on the analysis. The reason for the small transmissivity estimated from testing in wells S302, S3254, S355, and S356 is not clear. However, the large amount of argillaceous material in the upper part of the Ancell aquifer, which was the interval most of these wells were open to, may have reduced its permeability relative to the aquifer as a whole.

Water-level altitudes were measured in 14 wells open to the Ancell aquifer in June and July 2002 and in an additional 4 wells open to the Ancell aquifer in June 2003. Water-level altitudes in the wells completed in the Ancell aquifer vary from 253 to 840 ft (fig. 22). The measurement of 253 ft probably reflects the presence of pumping in well S352 at the time of measurement, but the large range in water-level altitudes in these wells is probably the result of differences in the amount of inflow to the well from the Silurian-Maquoketa aquifer, not natural variations in the water level in the Ancell aquifer. Where the water-level altitude is low, for example, in well S134, it is probable that the well receives little of no water from the Silurian-Maquoketa aquifer and the water level in the well approximates the water level of the Ancell aquifer. Where the water-level altitude is high, for example, in well S35, it is probable that the well receives large amounts of water from the Silurian-Maquoketa aquifer and the water level in the well partly reflects the water level in the Silurian-Maquoketa aquifer. The water levels in most of the wells are a composite of the water level of the Silurian-Maquoketa, Galena-Platteville, and Ancell aquifers and are not representative of any one aquifer. Because water-level altitudes in most or all of the wells open to the Ancell aquifer do not represent the water level in the aquifer, the potentiometric surface of the aquifer cannot be contoured with the available data and the direction of ground-water flow in the aquifer cannot be determined. These data can be used to determine the maximum water-level altitude in the Ancell aquifer. The data indicate that water levels tend to decrease from west to east in the township, indicating the potential for flow from west to east. This interpretation is consistent with interpretations made by previous investigators (Burch, 2002). Water-level altitudes in the wells completed in the Ancell aquifer are typically lower than the water-level altitudes in wells completed in the Galena-Platteville aquifer, indicating the potential for downward flow between these aquifers.

Periodic measurement of water levels in well S354 from June 2003 through August 2005 indicates that water levels decreased overall by about 5 ft during that time. However, this interpretation is based on a small number of measurements and additional data are required to determine if this variation can be attributed to natural causes or anthropogenic influences (pumping). Comparison of water-level measurements from well S354 collected by the driller in 1967 (Patrick Mills, U.S. Geological Survey, written commun., 2005) with the August 2005 measurement indicates that water levels have decreased about 80 ft since 1967.



EXPLANATION

S302
585.7
541.4

WELL LOCATION--Top number is well name. Middle number, where present, shows water level measured in June-July 2002. Bottom number, in bold where present, shows water level measured in June 2003. ——, indicates no measurement available.

Figure 22. Water levels in wells completed in the Ancell aquifer, Campton Township, Illinois, June-July 2002 and June 2003.

Water-level measurements collected by the USGS from residential-supply wells open to the Ancell aquifer in Campton Township in 2002 and 1995 give no clear indication that the volume of water being removed from the aquifer by pumping exceeds the amount of water being added by inflow from outside the township, recharge through the Galena-Platteville aquifer, and enhanced flow through open boreholes in the township. Long-term data collected from well S354, however, indicates that regional pumping from the Ancell aquifer over a period of decades may induce substantial declines in water level beneath the township

WATER USE

Information provided by the Kane County Development Department (written commun., 2002) indicates that there are approximately 5,025 residences and commercial establishments in Campton Township. Although exact numbers of residences and commercial establishments are not available, it was estimated that residences (including farms) constitute more than 95 percent of the water users in Campton Township. About 3,640 residences (72 percent) derive their water from private wells and about 1,390 (28 percent) derive their water from production wells. There are no known industrial, mining, or irrigation ground-water withdrawals in the township. All water in the township is supplied by ground water.

There are seven active production wells in Campton Township (table 4). Two wells are open to the deep glacial drift aquifer in section 16. Four production wells are open exclusively to the Ancell aquifer, one in section 3, one is section 23, one in section 24, and one in section 15 (fig. 1). The seventh active production well in Campton Township is open to the Ancell aquifer and the underlying Ironton-Galesville aquifer in section 32. Two unused wells open to the Ancell and Ironton-Galesville aquifers supplied water to a commercial facility just outside the southeastern edge of the township.

From 1999 through 2002, the two production wells in section 16 discharged a total of between about 35.6 and 38.7 Mgal/yr (average of about 103,000 gal/d) to 380 residences (Mark Menard, Utilities Inc., written commun., 2003), for an average water use of about 270 gal/d per household. The subdivision that receives water from the two production wells located in sections 23 and 24 (sections shown in figure 1) has been increasing in population since its inception in 1995, but these wells discharged about 137,000 gal/d to about 510 recipients (primarily residents but also some commercial establishments) in 2002 (Frank Gorham, Robert H. Anderson and Associates, written commun., 2003). The average water use for these wells has consistently been about 264 gal/d per residence.

Of the 3,638 households relying on residential wells, well-construction logs were available for 1,476 (about 40 percent) of the wells. The available well logs were grouped by section and divided into four categories; wells completed in the glacial drift, wells completed in the Silurian-Maquoketa aquifer, wells completed in the Galena-Platteville aquifer, and wells completed in the Ancell aquifer (table 5). Numerous bedrock wells are open to multiple aquifers, but this report considers withdrawals to be from the deepest aquifer penetrated by the well (the aquifer the well is completed in).

Analysis of the well-construction logs indicates that approximately 19 percent of the wells are completed in the glacial drift, about 19 percent are completed in the Silurian-Maquoketa aquifer, about 50 percent are completed in the Galena-Platteville aquifer, and about 13 percent are completed in the

Table 4. Estimated water withdrawals from production wells in Campton Township, Illinois, 2002.

[DD, deep glacial drift aquifer; An, Ancell aquifer; IG, Ironton-Galesville aquifer; *, total annual dischcharge from wells S352 and S353, discharge from indivual wells not known; **, assumes equal disharge from wells S352 and S353; ?, unknown]

Well name (fig. 32)	Open interval (feet below measuring point)	Aquifer open to	Total annual withdrawls in 2002 (gallons)	Average daily withdrawls in 2002 (gallons)
S268	?-186	DD	20,445,840	56,016
S269	?-186	DD	18,133,200	49,680
S352	613-875	An	57,086,000*	78,200**
S353	614-870	An		78,200**
S361	?-1393	An/IG	49,200,000	134,795
S362	?-790	An	2,417,760	6,624
S363	?	An	na	?
Total				**403,515**

Table 5. Well information and estimated residential water use, Campton Township, Illinois, 2002.

[-, not applicable]

Section (fig. 1)	Number of well records in the section	Number of residences on private water supply in the section	Number of wells finished in glacial drift aquifers in the well records for the section	Number of wells finished in bedrock in the well records for the section	Number of wells finished in the Silurian-Maquoketa aquifer in the well records for the section	Number of wells finished in the Galena-Platteville aquifer in the well records for the section	Number of wells finished in the Ancell aquifer in the well records for the section
1	100	186	1	99	22	74	3
2	25	34	15	10	4	6	0
3	41	94	7	34	12	18	4
4	18	51	2	16	2	11	3
5	4	14	0	4	0	4	0
6	5	20	1	4	2	2	0
7	4	5	0	4	0	2	2
8	16	26	0	16	0	5	11
9	4	17	2	2	0	1	1
10	101	323	15	86	14	54	18
11	74	164	25	49	25	18	6
12	91	191	2	89	9	67	13
13	107	181	5	102	12	80	10
14	28	158	0	28	6	15	7
15	61	149	14	47	9	34	4
16	24	81	6	18	2	14	2
17	89	187	18	71	8	40	23
18	34	106	7	27	3	17	7
19	3	10	1	2	1	0	1
20	43	131	3	40	1	24	15
21	119	278	69	50	7	33	10
22	21	88	8	13	6	5	2
23	19	72	1	18	6	10	2
24	80	186	35	45	17	21	7
25	95	299	32	63	18	29	16
26	84	121	1	83	16	55	12
27	52	180	0	52	20	28	4
28	23	45	2	21	15	4	2
29	7	14	2	5	1	3	1
30	3	10	0	3	0	3	0
31	2	10	0	2	1	1	0
32	3	3	0	3	0	3	0
33	5	9	3	2	0	2	0
34	44	115	1	43	21	22	0
35	45	76	1	44	15	27	2
36	2	4	0	2	1	1	0
Total	**1,476**	**3,638**	**279**	**1,197**	**276**	**733**	**188**

Ancell aquifer. These logs are representative of the wells drilled at the time of home construction but typically do not reflect the depths of wells that have been redrilled. Of the sections with 10 or more well records, glacial drift wells account for 20 percent or more of the wells drilled in sections 2, 11, 15, 16, 17, 18, 21, 22, 24 and 25 (sections shown in fig. 1). These sections are predominately along an east-west line in the central third of the township in areas of lower bedrock-surface altitude (fig. 11). Wells completed in the Silurian-Maquoketa aquifer account for 20 percent or more of the wells drilled in sections 1, 3, 11, 14, 22, 23, 24, 27, 28, 34, and 35. These sections are in the eastern half of the township and typically are areas of elevated bedrock-surface altitude. Wells completed in the Ancell

Table 5. Well information and estimated residential water use, Campton Township, Illinois, 2002—continued.

[-, not applicable]

Section (fig. 1)	Estimated number of wells finished in glacial drift aquifers in the section	Estimated number of wells finished in bedrock in the section	Estimated number of wells finished in the Silurian-Maquoketa aquifer in the section	Estimated number of wells finished in the Galena-Platteville aquifer in the section	Estimated number of wells finished in the Ancell aquifer in the section	Total daily withdrawals from the section (gallons)	Total daily withdrawals from glacial drift aquifers in the section (gallons)	Total daily withdrawals from the Silurian-Maquoketa aquifer in section (gallons)	Total daily withdrawals from the Galena-Platteville aquifer in the section (gallons)	Total daily withdrawals from the Ancell aquifer in the section (gallons)
1	2	184	41	138	6	49,290	493	10,844	36,475	1,479
2	20	14	5	8	0	9,010	5,406	1,442	2,162	-
3	16	78	28	41	9	24,910	4,253	7,291	10,936	2,430
4	6	45	6	31	9	13,515	1,502	1,502	8,259	2,253
5	0	14	0	14	0	3,710	-	-	3,710	-
6	4	16	8	8	0	5,300	1,060	2,120	2,120	-
7	0	5	0	3	3	1,325	-	-	663	663
8	0	26	0	8	18	6,890	-	-	2,153	4,737
9	9	9	0	4	4	4,505	2,253	-	1,126	1,126
10	48	275	45	173	58	85,595	12,712	11,865	45,764	15,255
11	55	109	55	40	13	43,460	14,682	14,682	10,571	3,524
12	4	187	19	141	27	50,615	1,112	5,006	37,266	7,231
13	8	173	20	135	17	47,965	2,241	5,379	35,862	4,483
14	0	158	34	85	40	41,870	-	8,972	22,430	10,468
15	34	115	22	83	10	39,485	9,062	5,826	22,008	2,589
16	20	61	7	47	7	21,465	5,366	1,789	12,521	1,789
17	38	149	17	84	48	49,555	10,022	4,454	22,272	12,806
18	22	84	9	53	22	28,090	5,783	2,479	14,045	5,783
19	3	7	3	0	3	2,650	883	883	-	883
20	9	122	3	73	46	34,715	2,422	807	19,376	12,110
21	161	117	16	77	23	73,670	42,716	4,334	20,429	6,191
22	34	54	25	21	8	23,320	8,884	6,663	5,552	2,221
23	4	68	23	38	8	19,080	1,004	6,025	10,042	2,008
24	81	105	40	49	16	49,290	21,564	10,474	12,939	4,313
25	101	198	57	91	50	79,235	26,690	15,013	24,188	13,345
26	1	120	23	79	17	32,065	382	6,108	20,995	4,581
27	0	180	69	97	14	47,700	-	18,346	25,685	3,669
28	4	41	29	8	4	11,925	1,037	7,777	2,074	1,037
29	4	10	2	6	2	3,710	1,060	530	1,590	530
30	0	10	0	10	0	2,650	-	-	2,650	-
31	0	10	5	5	0	2,650	-	1,325	1,325	-
32	0	3	0	3	0	795	-	-	795	-
33	5	4	0	4	0	2,385	1,431	-	954	-
34	3	112	55	58	0	30,475	693	14,545	15,238	-
35	2	74	25	46	3	20,140	448	6,713	12,084	895
36	0	4	2	2	0	1,060	-	530	530	-
Total	**688**	**2,950**	**680**	**1,807**	**463**	**964,070**	**185,162**	**183,723**	**466,788**	**128,397**

aquifer account for 20 percent or more of the wells drilled in sections 8, 14, 17, 18, and 20. These sections are predominately in the west-central part of the township.

Interviews with homeowners in Campton Township indicate that many of the wells originally completed in the Silurian-Maquoketa aquifer have been abandoned and new wells have been drilled into the Galena-Platteville or Ancell aquifers, particularly in the north-central part of the township. Therefore, the well breakdown shown in table 5 likely overestimates the number of wells currently completed in the Silurian-Maquoketa aquifer and underestimates the number of wells currently completed in the Galena-Platteville and Ancell aquifers.

To estimate the total number of residential-supply wells open to each aquifer in Campton Township, it was assumed that the percentage of wells penetrating each aquifer in the well log database for a section is representative of the percentages of the actual wells penetrating the various aquifers in that section. The total number of residences supplied by private wells in each section then was multiplied by the percentage of well logs penetrating each aquifer in that section to provide the final estimate of the total number of residential-supply wells using each aquifer in every section of the township (table 5).

To estimate the volume of water withdrawn by residential-supply wells, the number of wells in each section was multiplied by 265 gal/d, the approximate daily water use per residence based on data from production wells in the township. With this information, it is estimated that residential-supply wells withdrew more than 964,000 gal of water from the aquifers underlying Campton Township every day in 2002 (table 5). Based on the aquifer distribution in the well records, residential-supply wells withdrew more than 180,000 gal/d from both the glacial drift and Silurian-Maquoketa aquifers, nearly 500,000 gal/d from the Galena-Platteville aquifer (which draws some of its water from the Silurian-Maquoketa aquifer), and more than 125,000 gal/d from the Ancell aquifer. Production wells within the township withdrew about another 403,000 gal/d (table 4) during 2002, of which about 105,000 gal/d was from a deep glacial drift aquifer in section 16, about 157,000 gal/d was from the Ancell aquifer in sections 23 and 24, and about 6,600 gal/d was from the Ancell aquifer in section 3. If it is assumed that half of the water withdrawn by the production well in section 32 is taken from the Ancell aquifer and the rest from the Ironton-Galesville aquifer, about 67,400 gal/d was withdrawn from the Ancell aquifer by this well. No withdrawal data were reported for the production well completed in the Ancell aquifer in section 15, but withdrawals from this well are expected to be less than 2,000 gal/d. Total water withdrawals in the township, therefore, exceeded 1.36 Mgal/d in 2002, of which about 285,000 gal/d were withdrawn from the glacial drift aquifers, about 180,000 gal/d were withdrawn from the Alexandrian-Maquoketa aquifer, about 500,000 gal/d were withdrawn from the Galena-Platteville aquifer, about 356,000 gal/d were withdrawn from the Ancell aquifer, and about 67,000 gal/d were withdrawn from underlying aquifers.

SIMULATION OF GROUND-WATER FLOW AND AVAILABILITY

A computer ground-water model was developed to simulate the geohydrologic characteristics of the ground-water system and to estimate the source and availability of ground water in Campton Township. A computer model was selected for analysis because, generally, a computer model is superior to analytical solutions (a single equation). A fairly simplified ground-water-flow system would have to be assumed throughout Campton Township in order to apply an analytical solution, but a computer model can account for much more detail and variability in geology, hydrology, hydraulics, and stresses, and the interactions between these components of the ground-water system. The model can be used to: (1) quantify ground-water availability and flow through the system (determine the ground-water budget), (2) identify the recharge area to wells in Campton Township, (3) predict the effects of pumping on ground-water levels, and (4) simulate the effect of increased pumping on streams and other surface-water bodies, under current (2002) or a variety of hypothetical development conditions. This section describes the computer model chosen for the analysis, the conceptual hydrogeologic framework used to guide model construction, the calibration of the model to observations and simulated results, the sensitivity of simulation results to model input, and the model limitations and qualifications. The model used for this study is based on the three-dimensional, finite-difference computer code of Harbaugh and others (2000) called MODFLOW-2000. An iterative procedure is used in the model simulation to solve a finite-difference version of the continuity equation for steady flow in an anisotropic, heterogeneous, multi-layer, ground-water-flow system. The automated parameter-estimation process incorporated into MODFLOW-2000 is used to determine model-parameter values and estimate confidence limits of model predictions.

Simplifying Assumptions

A set of simplifying assumptions is used in the development of a ground-water model. The following assumptions were made for the geometry, hydraulic properties, and other characteristics of the ground-water-flow system underlying the area modeled in this study:

1. The multiple aquifers in both the shallow and deep parts of the unconsolidated unit within Campton Township are represented by two layers, one for all of the shallow glacial drift aquifers (layer 2), and a second for all of the deep glacial drift aquifers (layer 4). These aquifers are composed of sand and gravel separated by a confining unit composed of fine-grained materials (silt, clay, and till).
2. Horizontal and vertical hydraulic conductivity within the shallow and deep glacial drift aquifers are uniform thoughout the aquifers.
3. The bedrock is divided into four model layers representing the Silurian-Maquoketa aquifer, the Galena-Plattville aquifer, and the Ancell aquifer. The Silurian-Maquoketa aquifer is represented by two model layers so that the potential desaturation of this aquifer can be simulated.
4. The thickness of all simulated streambeds is 1 ft. The calibrated value of streambed vertical hydraulic conductivity is based on a 1-ft bed thickness.
5. Horizontal and vertical hydraulic conductivity of the confining units in the unconsolidated unit are uniform throughout their extent.
6. The ground-water-flow system is in dynamic equilibrium. Dynamic equilibrium is defined as a water-level fluctuation centered on a long-term average water level. The starting water levels are assumed to be at steady state (no change in water level with respect to time). Water levels for the Ancell aquifer are rebounding in parts of northern Illinois (Burch, 2002), and model simulation reflects water levels for the Ancell aquifer for 2002.

Model Design

The computer model is based on a rectangular, block-centered grid network that extends about 1.3 mi beyond Campton Township to the north, west, and south and extends to the Fox River in the east (fig. 23). The extension of the grid boundary beyond Campton Township was necessary so that simulations within the township were not affected appreciably by the imposed grid boundary conditions. The greater extension of the grid to the east as opposed to other directions allowed the model to include a natural hydrologic boundary (Fox River) and additional substantial pumpage between Campton Township and the river. The grid (12.2 mi by 8.9 mi) comprised 7,505 blocks that ranged in size from 500 ft by 500 ft in the central part of the modeled area to 1,000 ft by 1,000 ft at the corners. A node size of 500 ft by 500 ft was the most common in the model grid and provides sufficient water-level and flow detail within Campton Township without generating a computationally excessive number of model nodes that extend the time required for computer simulations.

Ground-water flow is simulated through nine model layers that represent the unconsolidated and bedrock aquifers and confining units (fig. 24). Model layers 1-5 represent the glacial drift aquifers and the confining unit in the unconsolidated materials. Layers 6-9 represent the bedrock aquifers. The Silurian-Maquoketa aquifer was represented by layers 6 and 7. Layers 8 and 9 represent the Galena-Platteville and Ancell aquifers, respectively. To improve the stability of model solutions, all layers were simulated as confined, including layer 1 (unconsolidated confining unit). To ensure that the confined thickness of layer 1 simulated the same conditions as the thickness of a water-table layer, an iterative approach was used. A confined simulation was run, and the simulated water levels for layer 1 were used in a second simulation as the new, more representative water levels for layer 1. The water levels in layer 1 obtained from the second simulation then were used to generate even more representative water levels for layer 1. This process is repeated until the latest water-level surface for layer 1 and the previously simulated surface are virtually the same. At that point, the simulation results from a layer under confined conditions and a layer under unconfined conditions are identical because the transmissivities are the same. In some areas, adjustment of the water-level surface of layer 1 resulted in the layer thinning to zero. In this case, the model node was made inactive. Layers that thin to zero thickness were assigned a minimum thickness of 1 ft, and, to compensate for this artificial thickness,

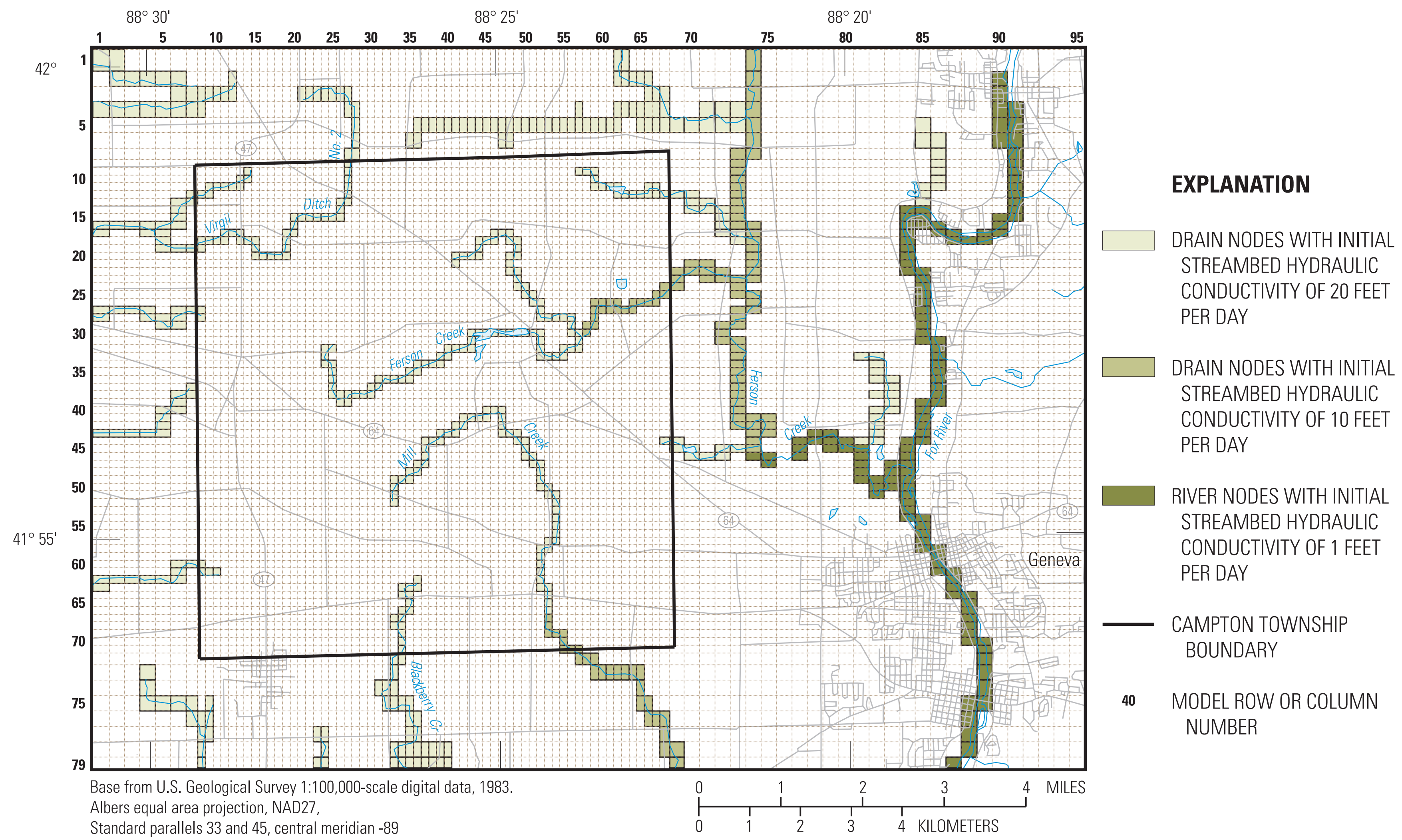


Figure 23. Model grid and river and drain nodes used in the simulation of ground-water flow in Campton Township and surrounding area, Illinois.

1 ft of thickness was deducted from the underlying model layer. In turn, the Kh of the aquifer layer below was assigned to the nodes representing the 1-ft layer.

River and drain nodes (McDonald and Harbaugh, 1988) were implemented in the uppermost active layer of the model to represent the streams shown in figure 23. A total of 132 river nodes were used to simulate the Fox River and the downstream end of Ferson Creek. River nodes represent large streams that can supply substantial amounts of water to the ground-water-flow system when the water table declines below the bottom of the stream. A total of 644 drain nodes were used to simulate the smaller streams, such as Mill Creek (fig. 15). Drain nodes receive ground water but do not allow inflow to the ground-water system. Drain nodes represent small streams that cease to flow when the water table declines below the bottom of the stream. Initial values of streambed vertical hydraulic con-

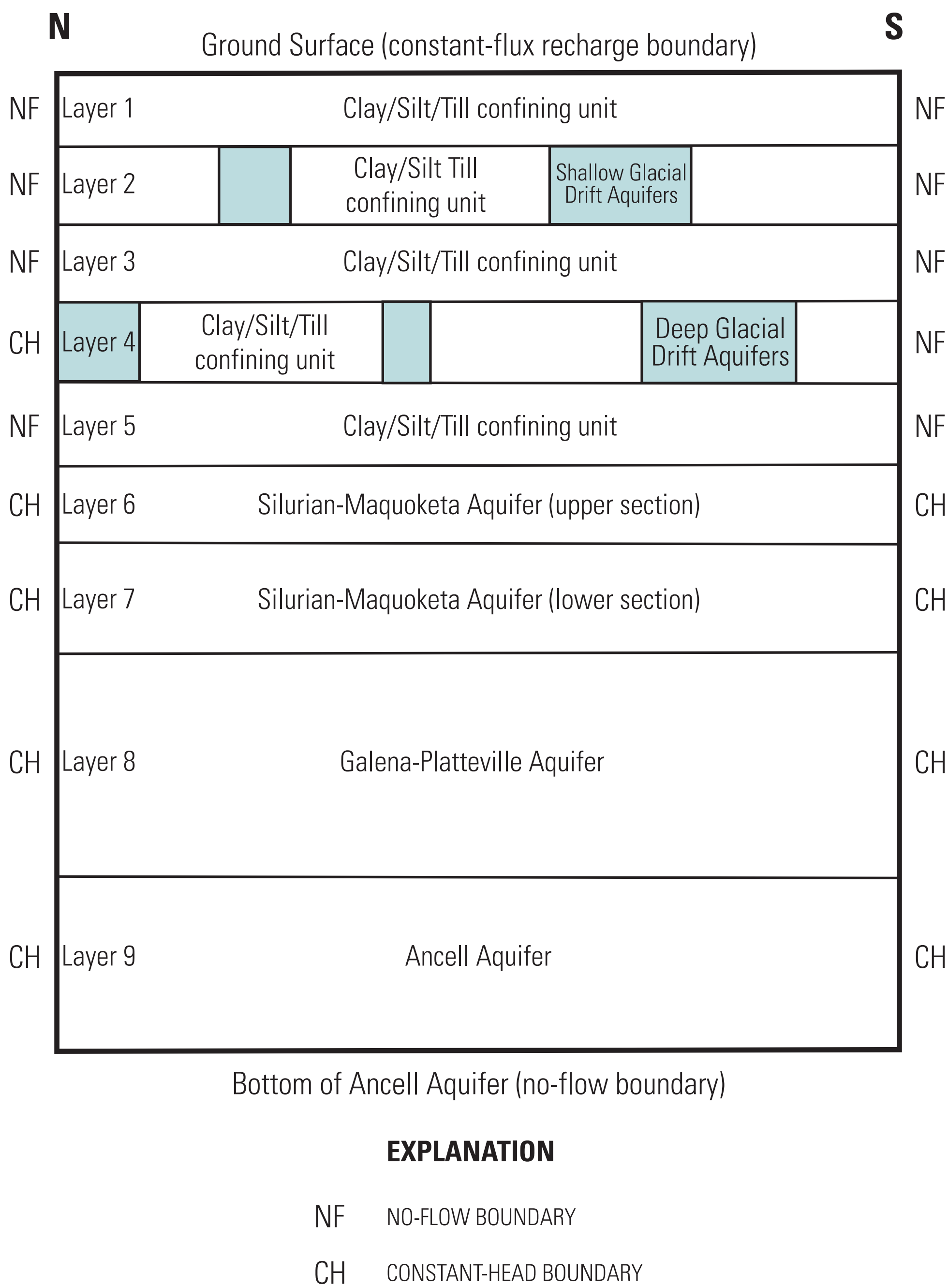


Figure 24. Idealized cross-sectional view of the ground-water model for Campton Township and surrounding area, Illinois.

ductivity were based on stream size. As stream size becomes larger, the grain size of streambed sediment becomes smaller (Leopold and others, 1964), which results in lower values of streambed vertical hydraulic conductivity. Based on this relation, three stream sizes were identified (small, intermediate, large) and three values of streambed hydraulic conductivity were defined (see table 6).

Boundary conditions in the ground-water model were selected so that the type and location of the boundary would have a minimal effect on the results of simulated pumping. Three boundary conditions were used in the simulation (fig. 24). A constant-flux boundary was used at the top of the model to represent recharge from precipitation. A no-flow boundary was used at the bottom of the model, which is the bottom of the Ancell aquifer. The no-flow boundary was chosen in this case because withdrawals are from the Ancell aquifer and shallower aquifers, and because the Ancell overlies a confining unit that is more than 300 ft thick (Visocky and Schulmeister, 1988). No-flow boundaries also were used on the sides of the model wherever fine-grained units are present. A constant-head boundary was used around the model layers representing the Silurian-Maquoketa, Galena-Platteville, and Ancell aquifers because lateral flow from outside the model area may contribute substantial amounts of water. The constant-head boundary for the Ancell aquifer also served to impose the general water-level decline measured in the Cambrian-Ordovician aquifer system (Visocky and Schulmeister, 1988). A small number of constant-head nodes also are used wherever the shallow and deep glacial drift aquifers are present in limited extent at the sides of the model.

Initial water levels for model simulations were estimated from water-level maps and computerized well-log records of water levels. The calculation of reasonable initial water levels was important for the assignment of constant-head values around the model boundaries. The water levels are assumed to represent conditions for 2002. Starting water levels for layer 2, representing the shallow glacial drift aquifers, were based on well-log records of water levels from wells in the shallow part of the unconsolidated unit. Water levels were kriged to create an estimated water-level surface, and surface values within each model node were averaged for model input. Initial water levels for the layer 1, representing the uppermost confining unit, were an average of the elevation surfaces representing land surface and water levels in the shallow glacial drift aquifers (layer 2). Starting water levels for layer 4, representing the deep glacial drift aquifers, were based on well-log records of water levels from wells in the deeper part of the glacial materials. Starting water levels for layers 3 and 5, representing the confining units above and below the deep glacial drift aquifers, are averages of the surfaces above and below the two layers. Starting water levels for layers 6 and 7, representing the Silurian-Maquoketa aquifer, were estimated from a water-level map, as well as from point values. The Silurian-Maquoketa water-level map from Kay and Kraske (1996) was scanned and converted to water-level contours. These contours provided data for Campton Township. For the rest of the model area, point values of water levels from well logs penetrating the Silurian-Maquoketa aquifer were used to estimate water-level surfaces in layers 6 and 7. These two data sources were used with an ARC/INFO command called TOPOGRID to estimate the water-level surface. Initial water levels for layer 9, representing the Ancell aquifer, were derived from a combination of water-level maps by Visocky and Schulmeister (1988) and Burch (2002). The maps show water levels for the Cambrian-Ordovician aquifer system, which includes, but is not exclusive to, the Ancell aquifer. Therefore, water levels from the map provided only an estimate of water levels in the Ancell aquifer. It was anticipated that simulated water level for the Ancell aquifer (layer 9) may require adjustment during calibration because of the uncertainty of using water levels for the Cambrian-Ordovician aquifer system to represent water levels in the Ancell aquifer.

The location and volume of water withdrawn from production wells in the township in 2002 was obtained from the ISWS Public-Industrial-Commercial Database (Timothy Bryant, Illinois State Water Survey, written commun., 2004). Volumes of water withdrawn from residential supply wells were estimated from analysis of the well-log database and records of home ownership in the township and described in the section on water use in this report. The pumpage rate for individual home wells, 0.00047 ft^3/s (304.5 gal/d), was based on the maximum water use recorded by the Wasco Sanitary District (Frank Gorham, written commun., 2003). This rate applies to the same period as for other calibration data. Pumpages from wells that penetrate multiple bedrock aquifers were divided among the model layers on the basis of aquifer transmissivity.

Initial values for the Kh of aquifers in each model layer were based on aquifer-test information collected as part of this investigation (tables 1 and 2), from aquifer tests on file with the ISWS, and analysis of data reported on driller's logs. Aquifer-test data obtained from driller's logs was analyzed by use of techniques developed by Prudic (1991) and Butler (1957). The Kh values calculated from

Table 6. Summary of aquifer-test results obtained from driller's logs in and around Campton Township, Illinois.

[ft/d, feet per day]

Aquifer and model layer	Range of horizontal hydraulic conductivity (ft/d)	Median horizontal hydraulic conductivity (ft/d)
Shallow glacial drift aquifers (model layer 2)	1.7 - 590	16
Deep glacial drift aquifers (model layer 4)	0.34 - 900	7.6
Silurian-Maquoketa aquifer (model layers 6 and 7)	0.27 - 68	4.4
All bedrock aquifers (model layers 6-9)	0.01 - 820	.37

aquifer testing provides a check against the model-calibrated values.

Information on recharge rates in the township was unavailable. Therefore, initial estimates of recharge were based on previous experience with recharge rates of glacial deposits in the Illinois and Indiana region (Eberts and George, 2000; Mills and others, 2002; Meyer and others, 1975). Fine-grained and sand-and-gravel units required appreciable different initial estimates of recharge. Recharge rates in areas of fine-grained materials were estimated to be 4 in/yr. Recharge rates in areas of sand and gravel were estimated to be 16 in/yr. Calibration of recharge rates was constrained by results of 12 streamflow measurements (table 3).

Model Calibration

Model calibration is the process of adjusting the model input variables, also called parameters, to produce the closest match between simulated and measured streamflow and water levels in the aquifers. During calibration, parameters representing aquifer hydraulic properties over specified areas were adjusted first manually by trial-and-error and then by automatic parameter-estimation techniques to match measured water levels in wells and measured streamflow. MODFLOW-2000 provides a parameter-estimation feature (Hill and others, 2000) that uses a nonlinear least-squares regression method to aid in estimating hydraulic properties and to evaluate model fit. The parameters estimated in the calibration process represent the hydraulic properties distributed as constant values over model areas or over extended linear features, such as rivers, and therefore are not intended to represent specific values of field tests at individual points.

Non-linear least-squares regression is an automated parameter-estimation technique that is more efficient and objective compared to trial-and-error calibration because all parameter values are adjusted automatically, without bias, and concurrently to obtain the best possible fit between simulated and measured water levels and streamflows. The numerical difference between measured and simulated values is called a residual. In the regression method, parameter values are estimated by minimizing the squared weighted residuals, called the objective function (Hill, 1998).

Representation of Aquifer Properties and Conditions by Model Parameters

In the model, grid cells assumed to have similar hydrologic properties are grouped together as a parameter zone and assigned a parameter value that can be adjusted during the calibration process. The model contained 26 different parameters. Seven of these parameters were estimated automatically. The names of the parameters and the model component that the parameter represents at the beginning of model calibration are shown in table 7. In addition, the parameters that were estimated and information about parameters that were tested, but eventually dropped from model simulation, is shown in table 7. The final parameter data set used in model simulation is provided in table 8.

Although all parts of the ground-water-flow system are parameterized, the degree of detail used to represent the flow system by parameterization is limited by the availability and distribution of observation data (water levels and streamflow). For example, only one parameter for Kh of the fine-grained units was created (**kclay**) and used in all the layers containing fine-grained materials (layers 1 - 5). Calibration data were not available to estimate the Kh of individual clay/silt/till units; therefore, the Kh of fine-grained units for the individual model layers were lumped together, and represented by one parameter. However, parameters for Kh were assigned to individual bedrock layers because observation water levels were available for all these layers.

The decision to estimate a specific parameter was based on the sensitivity of simulated water levels to changes in model parameters. The sensitivity of simulated water levels with respect to various parameters was calculated using the sensitivity

Table 7. Initial and final model parameters and indication of parameter being estimated in simulation of the ground-water-flow system, Campton Township, Illinois.

[-, not applicable].

Parameter name	Model component represented by the model	Parameter included in original model design	Parameter included in final model design	Parameter estimation attempted?	Parameter successfully estimated?
kclay	Horizontal hydraulic conductivity of clay, silt, and till units representing the confining unit (layers 1-4)	Yes	Yes	Yes	No
kvclay	Vertical hydraulic conductivity of the clay, silt, and till units representing the confining unit (layers 1-4)	Yes	Yes	Yes	Yes
k2	Horizontal hydraulic conductivity of shallow glacial drift aquifers (layer 2)	Yes	No	-	-
k4	Horizontal hydraulic conductivity of the deep glacial drift aquifers (layer 4)	Yes	No	-	-
ksg	Horizontal hydraulic conductivity of all glacial drift aquifers	No	Yes	Yes	Yes
kvsg	Vertical hydraulic conductivity of all glacial drift aquifers	Yes	Yes	No	-
k1o	Horizontal hydraulic conductivity of the surficial outwash	Yes	Yes	No	-
k1a	Horizontal hydraulic conductivity of the alluvium	Yes	Yes	Yes	No
k6	Horizontal hydraulic conductivity of layer 6 (upper section of Silurian-Maquoketa aquifer)	Yes	No	-	-
k7	Horizontal hydraulic conductivity of layer 7 (lower section of Silurian-Maquoketa aquifer)	Yes	No	Yes	
kam	Horizontal hydraulic conductivity of the Silurian-Maquoketa aquifer	No	Yes	Yes	Yes
kv6	Vertical hydraulic conductivity of layer 6 (upper section of the Silurian-Maquoketa aquifer)	Yes	Yes	Yes	No
kshale	Hydraulic conductivity of layers 6 and 7 in areas where shale is at bedrock surface	No	Yes	Yes	Yes
kv7	Vertical hydraulic con¬ductivity for layer 7 (lower section of Silurian-Maquoketa aquifer)	Yes	Yes	Yes	Yes
k8	Horizontal hydraulic conductivity for layer 8 (Galena-Platteville aquifer)	Yes	Yes	Yes	No
kv8	Vertical hydraulic conductivity for layer 8 (Galena-Platteville aquifer)	Yes	Yes	Yes	Yes
k9	Horizontal hydraulic conductivity for layer 9 (Ancell aquifer)	Yes	Yes	Yes	No
kv9	Vertical hydraulic conductivity for layer 9 (Ancell aquifer)	Yes	Yes	No	-
d10	Hydraulic conductivity of intermediate-sized drain streambeds	Yes	Yes	Yes	No
d20	Hydraulic conductivity of small-sized drain streambeds	Yes	Yes	Yes	No
r1	Hydraulic conductivity of the Fox River bed	Yes	Yes	No	-
r10	Hydraulic conductivity of Ferson Creek near Fox River	Yes	Yes	No	-
rcht	Recharge to the till (layer 1)	Yes	Yes	Yes	Yes
rchls	Recharge to the loamy sand (layer 1)	Yes	No	-	-
rchcse	Recharge to the outwash (layer 1)	Yes	Yes	Yes	No
rchaluv	Recharge to the alluvial deposits around the streams (layer 1)	No	Yes	Yes	No

equation method (Hill and others, 2000). Composite scaled sensitivities (CSS) were calculated for each parameter. CSS aid in determining if there is adequate information in the calibration data to estimate a particular parameter. CSS less than about 0.01 times the largest CSS of all the parameters indicate that the non-linear regression method may not be able to estimate the parameter (Hill, 1998).

Observations and Observation Weights

The observations used for model calibration consisted of 217 water-level measurements and 12 streamflow measurements made during the summer of 2002. Most observation water levels were from the aquifers most often used for water supply, represented in layers 4, 6, and 7; however, observations were available for all the aquifers. Water levels were measured throughout Campton Township, but measurements were more concentrated in areas of greater ground-water use. The streamflow measurement locations were along Mill and Ferson Creeks, and Virgil Ditch (fig. 15).

The purpose of weighting the observations used in model calibration is two-fold. First, weighting reduces the effect of observations that are known to be less accurate and increases the effect of observations that are known to be more accurate. Second, weighting produces weighted residuals (a measure of the difference between the observation and its simulated equivalent) that have the same units, whether the residual is for water-level or streamflow observations. Water-level and streamflow residuals in the same units allow both residuals to be included in the sum of squared errors to be minimized. Weights on observation data account for measurement error associated with the accuracy of the sampling device, method of determining land surface, effects of recent pumping, unknown screened intervals of wells, and other sources of uncertainty. For the regression method to produce parameter estimates with the smallest possible variance, the weights should be proportional to 1 divided by the variance of the measurement errors for the observation (Hill, 1998). To estimate these variances, MODFLOW-2000 reads statistics on measurement error from which the variances of the observation errors, and the weights are then calculated. The standard deviation of the measurement error was used to estimate the weights for water- level observations and the coefficient of variation was used for the streamflow measurements. The calculations of the statistics are described in Hill (1998). For the water-level statistic, it was assumed that water levels could be estimated with a 95-percent confidence to be within 1ft of the true value because land surface was estimated with a 2-ft land-surface contour map. If the measurement errors are assumed to be normally distributed, then a table of the cumulative distribution of a standardized normal distribution (Cooley and Naff, 1990) can be used to determine the desired statistic (standard deviation) for water-level measurement error as follows:

$$1.96 * \text{standard deviation} = 1 \text{ ft}$$
$$\text{standard deviation} = 0.5$$

The value 1.96 was obtained from a normal probability table. For the flow statistic, a commonly calculated value of 0.20 was used for all flow measurements. The approximation was acceptable because the weighted residuals for flow-measurements did not affect the regression calculations.

Changes in the Design of the Model During Calibration

The calibration process identified which changes to the model design were required to produce more reasonable model results. The major changes consisted of (1) combining the Kh parameters for the Silurian-Maquoketa aquifer (model layers 6 and 7) into one parameter, (2) redistributing the amount of pumpage from residential wells penetrating multiple layers, (3) adding a Kh parameter for an area of consistently large negative (measured minus simulated value) residuals, (4) assuming a Kh value of 5 ft/d for fine-grained material, and (5) adjusting the water level for specified heads in the Ancell aquifer (layer 9).

Individual parameters for Kh of layers 6 and 7 (the Silurian-Maquoketa aquifer) were lumped into a single parameter because the parameter estimation process did not yield reasonable values. In some parameter-estimation runs, the estimated value for hydraulic conductivity of the lower section of the aquifer was higher than that for the upper section. The reason for this difference is that almost all the observation water levels used to evaluate both the upper and lower sections of the aquifer are from wells that are open to the entire aquifer. Therefore, a distinction in the hydraulic conductivities of the two model layers (6 and 7) can not be made in the parameter-estimation process.

The majority of pumping wells in the model area are open to more than one model layer. As a result, the water pumped from the well is derived from multiple model layers, and the proportion that is obtained from each layer must be determined. Typically, the amount of water derived from each layer is assumed to be directly proportional to the transmissivity of individual layers penetrated by the pumping well. Accordingly, the initial model design divided layer contributions to residential pumpage on the basis of transmissivity alone. However, the large differences in water level between model layers representing the bedrock create an additional effect on the proportion of pumpage being derived from the layers. Field measurements of downhole flow in residential wells ranged from 0.1 to 10 gal/min, with an average value of about 0.1 gal/min. The probable average amount of vertical flow through wells is sufficiently similar to estimated residential pumpage (about 0.2 gal/min) that simulated residential pumpage at the end of model calibration was mostly derived from the Silurian-Maquoketa aquifer (layers 6 and 7). As a result of this pumping change during calibration, the rate of well discharge assigned to the Silurian-Maquoketa aquifer is greater than the proportion dictated by relative transmissivities.

An analysis of intermediate model-calibration results revealed that water-level residuals (measured minus simulated water level) were consistently large in one area. To reduce the residuals, two approaches were tested. The first approach was to add a parameter called **kshale**, which represented the Kh of the Silurian-Maquoketa aquifer (layers 6 and 7) where shale is present at the bedrock surface. The area of the model represented by **kshale** was determined from the lithology descriptions on the well logs used to construct the model. In this approach, the Silurian-Maquoketa aquifer was represented by two parameters, **kam**, and **kshale**. The calibrated value of **kshale** became one order of magnitude lower than that for **kam**, which reduced, but did not eliminate, the associated large positive residuals (measured water level higher than simulated). The second approach added the same parameter **kshale**, but the parameter extended only over the area of the large negative residuals (measured water levels lower than simulated). As a result, the calibrated value of **kshale** became more than two orders of magnitude lower than that for **kam.** The second approach produced the least overall model error and best calibration statistics.

At the beginning of parameter estimation, an attempt was made to estimate the Kh of the fine-grained clay/silt/till confining units (**kclay**). Commonly, the estimated value was unusually large. Perhaps, the lack of water-level data from the fine-grained units prevents an accurate estimation of the parameter. Eventually, **kclay** was set at 5 ft/d to facilitate parameter estimation. A value of 5.0 ft/d was chosen because it is less than the Kh value determined for units comprised predominately of till in other studies. Eberts and George (2000) determined a calibrated Kh value of 21.3 ft/d for units composed predominately of till at or near the surface over two-thirds of Indiana. Mills and others (2002) determined a calibrated Kh value of 10 ft/d for units composed predominately of till in and around the City of Belvidere in northern Illinois. A value lower than those used in previous studies was chosen for this investigation because the sand and gravels within the predominately fine-grained confining unit underlying Campton Township are partly accounted for by the simulation of separate model layers for the confining unit and the shallow and deep glacial drift aquifers, whereas models developed in other studies simulated an average Kh of the aquifers and till confining units together. However, 5 ft/d is a high value for a fine-grained confining unit. This value probably reflects (1) fracturing in the fine-grained materials, (2) mixtures of coarse- and fine-grained materials, such as silt and gravel mapped as till, and (3) thin deposits of sand and gravel that were not recorded by drillers.

The initial water-level surface for the Ancell aquifer (layer 9) was estimated from a map of the potentiometric surface of the Cambrian-Ordovician aquifer system (Visocky and Schulmeister, 1988) and then adjusted for recent water-level recovery in the aquifer system mapped by Burch (2002). The Ancell aquifer composes the upper part of the Cambrian-Ordovician aquifer system and the potentiometric surface of the Ancell aquifer should be similar to that for the Cambrian-Ordovician aquifer system. However, it is likely that water levels in the Ancell aquifer are higher than in the aquifer system as a whole. The water-level values (specifically the constant-head water levels around the boundary of layer 9) were adjusted upward from the level of the 1988 water-level map (Visocky and Schulmeister, 1988) during model calibration to improve the match between measured and simulated water levels in layer 9, as well as in the units represented by the layers above layer 9. The adjustment was made iteratively along with parameter estimation so that simulated water levels in all model layers were improved. Eventually, water levels for the Ancell aquifer (layer 9) were increased by 150 ft from those representing the Cambrian-Ordovician aquifer system. The increase includes the effect of a rebound in water levels from the Cambrian-Ordovician aquifer system of 40 ft in the Campton Township area measured from 1988 to 2000 (Burch, 2002).

Two alternative model designs were tested during the study, but not adopted in the final model: (1) providing a downward leakage from the Ancell aquifer (layer 9), and (2) adding an alluvial zone around streams. Neither design change appreciably affected ground-water levels and streamflow.

Calibration Results

This section provides the final values for model parameters, indications of how well the model simulates measured water levels and streamflow with the calibrated parameters, and a representation of the ground-water-flow system as simulated with the model. The calibrated parameter values and associated confidence intervals (if the parameters are estimated) are listed in table 8.

The model-estimated values for Kh of the glacial drift aquifers (layers 1-5) and the Silurian-Maquoketa aquifer (layers 6 and 7) are similar to field-measured values. The median value of transmissivity determined from six aquifer tests performed in wells completed in glacial drift aquifers as part of the study is 168 ft^2/d. The geometric mean Kh of these tests was 13 ft/d, which is in good agreement with the model-calibrated value of 17.5 ft/d. The Kh value for glacial drift aquifers calculated from analysis of the aquifer-test data reported in the driller's logs (table 6) averages about 12 ft/d. Besides closely matching measured values of Kh for the glacial drift aquifers, parameter estimation also approximately matched measured values of Kh for the Silurian-Maquoketa aquifer (layers 6 and 7). The geometric mean transmissivity of six aquifer tests in the Silurian-Maquoketa aquifer is 262 $ft^{2/}/d$, and the geometric mean Kh was calculated to be 2.3 ft/d, which is comparable to the model-estimated value of about 6 ft/d.

Because the parameter was not sufficiently sensitive with respect to simulated water levels, parameter estimation failed to determine a value for vertical hydraulic conductivity of the Galena-Platteville aquifer (layer 8); therefore, the value was determined manually. The value presented in table 8 was chosen so that the simulated differences in water levels between the Silurian-Maquoketa aquifer (layers 6 and 7) and the Ancell aquifer (layer 9) were comparable to the measured differences.

Automated parameter estimation may be restricted by parameters that are correlated (changes in one parameter can be compensated by proportional changes in another). Parameter estimation of correlated parameters may result in non-unique values for the parameters. The test for parameter correlation is the correlation coefficient, and values close to 1 or -1 indicate the possibility that the two parameters are not uniquely estimated. Only one parameter correlation (-0.94) was observed, and that is between **kshale** (Kh of the Silurian-Maquoketa aquifer in the area of large residuals) and **kvam** (vertical hydraulic conductivity of the Silurian-Maquoketa aquifer). This result is an unexpected correlation and probably is not indicating non-unique solutions of the two parameters. When the area represented by **kshale** was changed from the area of shale outcrop to only the area of shale with large water-level residuals, the parameter **kvam** did not change substantially, but **kshale** did change appreciably. The substantial change in **kshale** is an indication that only changes in **kshale** could reduce the large residuals, not changes in **kvam**. Therefore, the values for **kshale** and **kvam** are likely unique. The absence of other correlations is probably because of the availability of flow observations to the calibration. Although different combinations of Kh and recharge can produce suitable water levels, only one unique set of values can produce satisfactory ground-water levels and flows to measured stream sections (fig. 25). The 12 flow observations available for calibration were apparently adequate for assisting the automated calibration process to determine unique and uncorrelated parameter values.

Analysis of Residuals

The degree of fit between field measured and model-simulated values is an indication of how well the model represents the actual ground-water-flow system. Model fit can be measured in multiple ways, including correlation coefficients, plots of measured water levels in relation to simulated water levels, and water-level residuals. Ideally, simulated values should be close to measured values such that when weighted observations are plotted against weighted simulated values, the residual values should fall close to a line with slope equal to 1 and intercept of 0. The correlation coefficient between weighted observations and weighted simulated equivalents reflects how well the values follow the 1:1 line. A correlation coefficient greater than 0.90 is desirable and the calibration resulted in a value of 0.99. The plot of weighted measured water levels in relations to weighted simulated water levels and two plots showing weighted residuals is shown in figure 25.

Two types of weighted residual plots are shown in figure 25. The weighted residuals are plotted according to their position in an assumed normal distribution (fig. 25b). If the residuals are truly normally distributed (a requirement for valid parameter estimation), then the residuals should plot along a straight line. The statistic that measures the linearity of the plot, as well as the independence of one residual to another, is called the correlation

Table 8. Initial and final parameter values used in the simulation of the ground-water-flow system, Campton Township, Illinois.

[ft/d, feet per day; -, not estimated; *, no initial value; in/yr, inches per year]

Parameter	Initial parameter value	Final parameter value	Approximate 95-percent Linear confidence interval
Horizontal hydraulic conductivity of clay, silt, and till units	0.05 ft/d	5.0 ft/d	-
Vertical hydraulic conductivity of clay, silt, and till units	$5 * 10^{-4}$ ft/d	$2.46 * 10^{-4}$ ft/d	$1.68 * 10^{-4} – 3.63 * 10^{-4}$ ft/d
Horizontal hydraulic conductivity of all glacial drift aquifers	16 ft/d (shallow glacial drift aquifers) 8 ft/d (deep glacial drift aquifers)	17.5 ft/d (both aquifers)	7.14 – 42.8 ft/d
Horizontal hydraulic conductivity of the Silurian-Maquokekta aquifer where dolomite is at the bedrock surface	4 ft/d	5.99 ft/d	4.55 – 7.88 ft/d
Vertical hydraulic conductivity of the Silurian-Maquoketa aquifer	0.04 ft/d (Silurian) $6 * 10^{-5}$ ft/d (Maquoketa)	$1.18 * 10^{-4}$ ft/d (both aquifers)	$6.03 * 10^{-5} – 2.29 * 10^{-4}$ ft/d
Horizontal hydraulic conductivity of the Silurian-Maquokekta aquifer in area of shale with large residuals	*	0.0137 ft/d	0.00582 – 0.0329 ft/d
Horizontal hydraulic conductivity of the Galena-Platteville aquifer (layer 8)	0.076 ft/d	0.076 ft/d	-
Vertical hydraulic conductivity of the Galena-Platteville aquifer (layer 8)	$1.6 * 10^{-4}$ ft/d	$2.24 * 10^{-7}$ ft/d	-
Horizontal hydraulic conductivity of the Ancell aquifer (layer 9)	8.3 ft/d	5 ft/d	-
Recharge rate to the fine-grained glacial units	4 in/yr	1.82 in/yr	0.95 – 3.53 in/yr
Vertical hydraulic conductivity of all glacial drift aquifers	3.0 ft/d (shallow glacial drift aquifers) 1.6 ft/d (deep glacial drift aquifers)	2.62 ft/d (both aquifers)	-
Vertical hydraulic conductivity of the Ancell aquifer (layer 9)	0.083 ft/d	0.05 ft/d	-
Hydraulic conductivity of small- sized drain streambeds	20 ft/d	20ft/d	-
Hydraulic conductivity of interme¬diate-sized drain streambeds	10 ft/d	10 ft/d	-
Hydraulic conductivity of the Fox River streambed	1 ft/d	1 ft/d	-
Hydraulic conductivity of Ferson Creek near Fox River	10 ft/d	10 ft/d	-
Recharge to the surficial outwash	16 in/yr	16 in/yr	-

between ordered weighted residuals and normal order statistics. This correlation coefficient also should be near 1, and the value associated with the model calibration is 0.99 for this study, indicating the residuals are normally distributed. Weighted residuals and their weighted simulated values are shown in figure 25c. Ideally, the weighted residuals should be evenly distributed around 0, and the size of the weighted residuals should not be related to the magnitude of the weighted simulated values (for example, large residuals associated with lower simulated values). These requirements were generally satisfied in model calibration.

The residual plots do not show the residuals associated with measured streamflow. The plotting position for the flow residuals are near the origin of the axis, whereas the water-level residuals are far from the origin. The relative plotting positions reflect the difference in magnitude of the sum of squared residuals for each type of observation: 397 for flow residuals and 763,040 for water-level residuals. Plotting all residuals on the same graph would result in the water-level residuals plotting so close together that details of the distribution of the water-level residuals would be obscured. To display the more important error associated with the water-

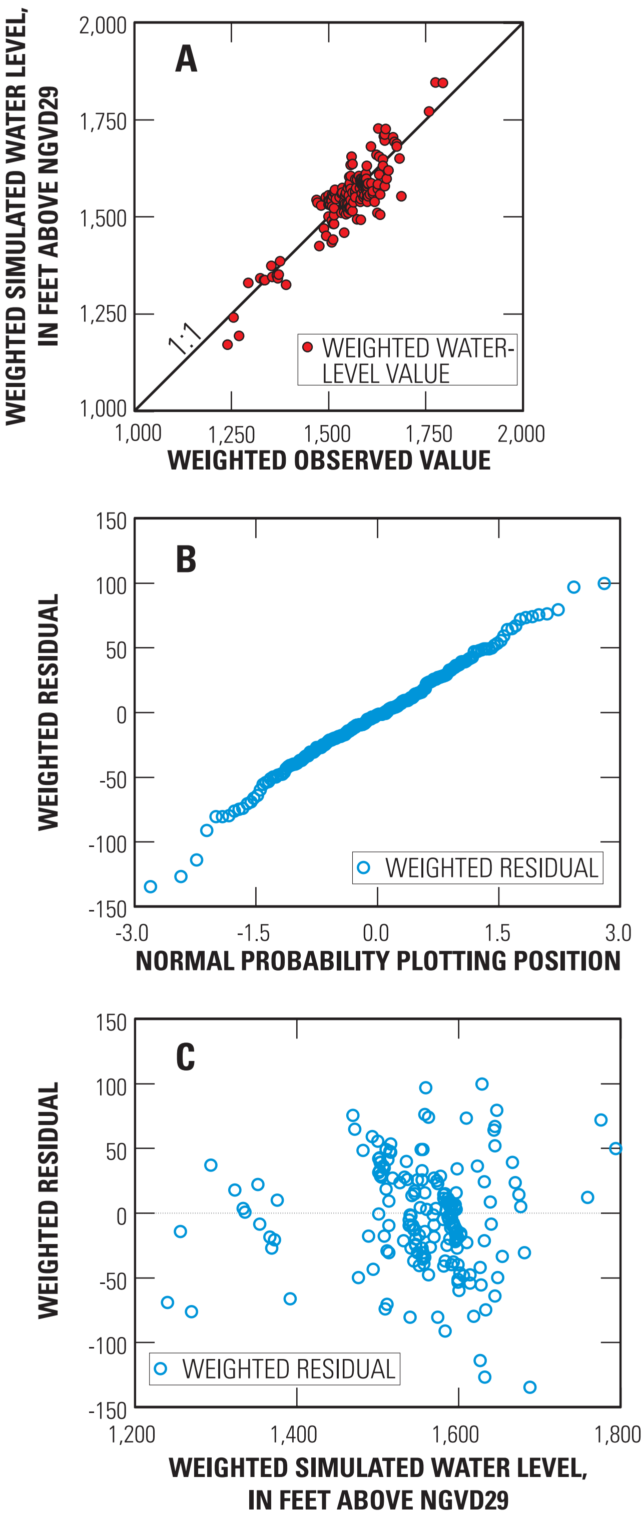


Figure 25. Graphical analysis of model fit for Campton Township and surrounding area, Illinois, A) weighted simulated and weighted observed water levels; B) normal probability plot of weighted residuals; C) weighted residuals and weighted simulated values.

level simulations, the axes on the residual plots were chosen to show only the water-level residuals. The flow residual data are presented in table 9 and referenced areally in figure 25. The tendency for error in simulated flows is to be greatest for the smaller streams (table 9 and fig. 25). During all of the process of calibration, matching measured flows was difficult, possibly indicating that the model does not include all the glacial drift aquifers connected to the streams. The flow residuals do not generate substantial model error because they plotted near the 1:1 line in figure 25a and near the 0 line in figure 25c.

Residuals also can be analyzed by their areal distribution and magnitude. The map of water-level residuals for all layers shows a mix of positive and negative residuals over the model area (fig. 26), which is characteristic of an adequately calibrated model. The magnitude of unweighted water-level residuals for different layer combinations is presented in terms of their median value and their standard deviation (table 10). In a set of normally distributed residuals, about two-thirds of the residuals are within one standard deviation of the mean residual, and about 95 percent are within two standard deviations. The spread of residuals can be used as a guide to the accuracy of the simulated water levels. The spread of residuals becomes greater for progressively deeper model layers, which represent the deeper bedrock aquifers, and indicates that model-derived estimates are less accurate for the deeper layers than for the shallower layers.

The magnitude of the weighted residuals provide a measure of model fit, and the standard error of the regression measures that magnitude The standard error of the regression for the model is dimensionless, so it is multiplied by the standard deviation of water-level measurement error to obtain a measure of overall model fit for water levels. The standard error for the model is 40.5 ft and the standard deviation of measurement error is 0.5; therefore, the overall model fit is + (-) 20.2 ft. To evaluate the error term relative to the degree that measured water levels can fluctuate for all layers in the model area (337 ft), 20.2 ft is a 6-percent model error.

The model fit of + (-) 20.2 ft was achieved by creating a parameter zone in the Silurian-Maquoketa aquifer with lower-than typical Kh corresponding to the area of large residuals in northeastern Campton Township. Other areas of substantially lower Kh may be present in Campton Township, but water-level measurements have not been made at every possible point in the township to help detect the possible areas of low hydraulic conductivity. If unknown areas of low Kh are present, then the

Table 9. Unweighted flow residuals for streamflow-measurement sections used for model simulation of the ground-water-flow system, Campton Township, Illinois.

[ft^3/s , cubic feet per second]

Streamflow-measurement section number (fig. 15)	Measured flow (ft^3/s)	Simulated flow (ft^3/s)	Unweighted flow residual (simulated – measured) (ft^3/s)
2	0.35	0.09	-0.26
3	-.08	.07	.15
4	1.02	.27	-.75
5	.49	.14	-.35
6	-.27	.32	.59
8	.95	.65	-.30
9	.68	.00	-.68
10	.14	.40	.26
11	.39	.33	-.03
12	.85	.66	-.19
13	1.30	.84	-.46

model fit calculated with available water-level data overestimates model accuracy.

Residuals can be associated with composite water levels of multiple layers. To obtain such residuals, the simulated water levels that are compared to measured levels are a weighted average of simulated water levels from individual layers penetrated by the well. The weighting is based on transmissivity of the individual layers.

Model Budget

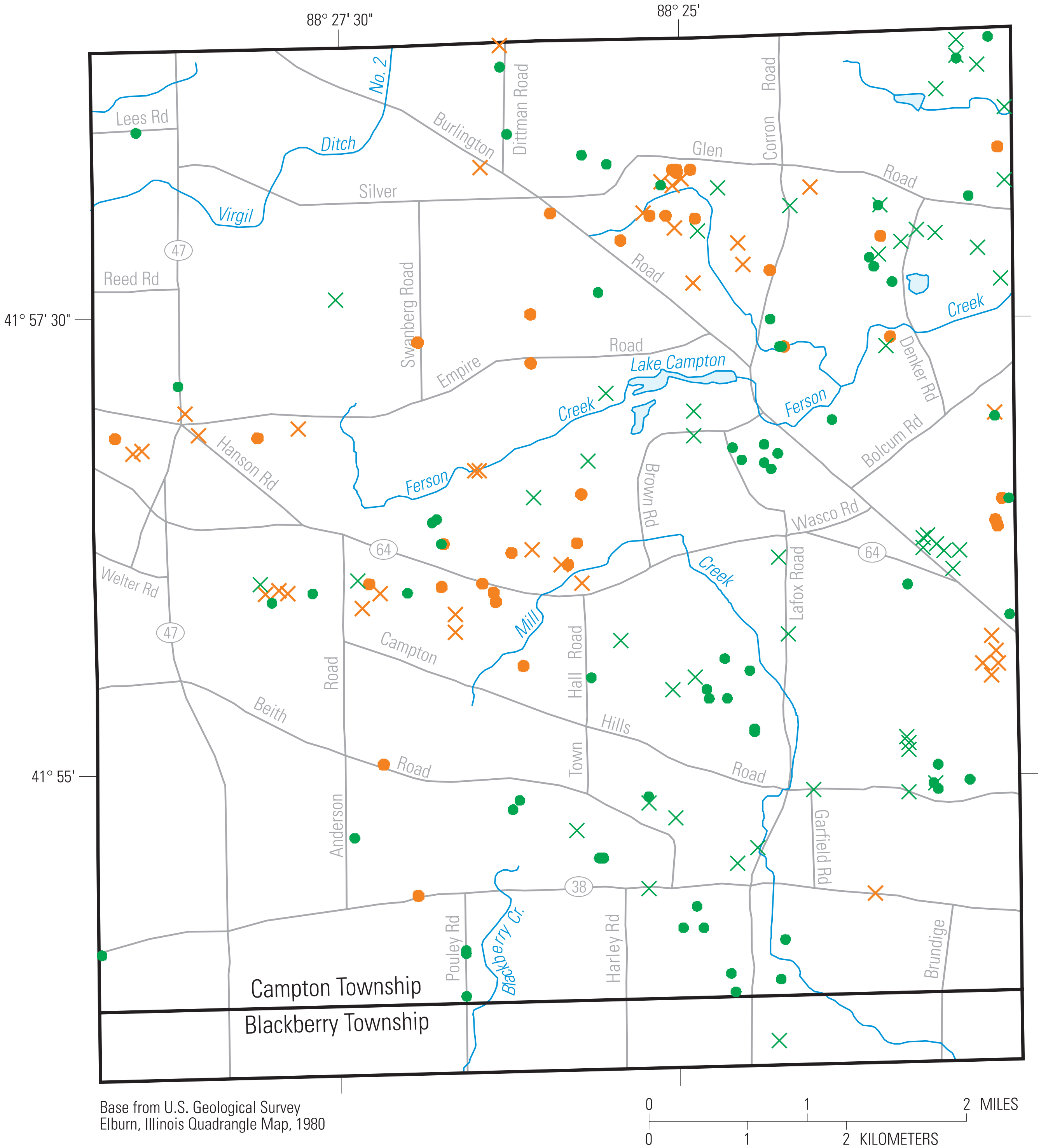
The calibrated model can be quantified and analyzed by estimating an overall flow budget, a budget for individual model layers, and a budget for the Campton Township portion of the model only. The overall model flow budget is shown in table 11. Two features of the budget stand out. First, about 75 percent of inflow to the local ground-water-flow system (the glacial drift aquifers and the Silurian-Maquoketa aquifer) is from local precipitation, and not as constant-head boundary flow from the sides of the model. Therefore, most of the recharge area to the local ground-water supply is within the township, and local land-use management decisions can have a substantial effect on local ground-water quantity and quality. Second, as of 2002, 14.8 percent of ground water is being discharged to wells, 13.8 percent to model boundaries, and 71.3 percent flows to streams, including the Fox River. These percentages may be used as a benchmark against future ground-water development within the township.

The average flow budget can be analyzed in further detail by examining the budgets for individual layers, as shown in figure 27. Note that in figure 27 the schematic representing the model has changed from the schematic for the conceptual model in figure 24. Model layers 1 through 5 are discontinuous in figure 27 because the calibrated water levels fell below the bottom of these layers in various places. One pattern indicated in figure 27 is that vertical flow interaction between layers decreases with depth. A practical implication of this result is that pumpage from the deeper aquifers, such as the Ancell (layer 9), would have the least

Table 10. Magnitude of unweighted water-level residuals for model layers in model simulation of the ground-water-flow system, Campton Township, Illinois.

[ft, feet]

Layer or layer combination	Hydrogeologic unit	Number of observations	Median residual (ft)	Bias	Standard deviation of residuals (ft)
2	Shallow glacial drift aquifers	7	0.57	-8.51	15.7
4	Deep glacial drift aquifers	63	-1.45	.01	19.7
6,7	Silurian-Maquoketa aquifer (upper and lower sections)	97	1.95	.12	19.7
6,7,8,9	Silurian-Maquoketa, Galena-Platteville, and Ancell aquifers	17	22.2	18.1	24.3
All layers	All aquifers	209	1.37	1.05	20.7



EXPLANATION

- ● LOCATION OF A POSITIVE RESIDUAL IN THE MODEL LAYERS REPRESENTING BEDROCK AQUIFERS (LAYERS 6-9)
- ● LOCATION OF A POSITIVE RESIDUAL IN THE MODEL LAYERS REPRESENTING GLACIAL DRIFT AQUIFERS (LAYERS 2 AND 4)
- × LOCATION OF A NEGATIVE RESIDUAL IN THE MODEL LAYERS REPRESENTING BEDROCK AQUIFERS (LAYERS 6-9)
- × LOCATION OF A NEGATIVE RESIDUAL IN THE MODEL LAYERS REPRESENTING GLACIAL DRIFT AQUIFERS (LAYERS 2 AND 4)

Figure 26. Distribution of positive and negative water-level residuals resulting from model simulation of ground-water flow, Campton Township, Illinois. A positive residual indicates that the measured value is greater than the simulated; a negative residual indicates that the simulated value is greater than the measured.

Table 11. Water budget associated with model calibration of the ground-water-flow system, Campton Township, Illinois.

[ft^3/s, cubic feet per second]

Source of flow into the model	Inflow rate (ft^3/s)	Source of outflow from the model	Outflow rate (ft^3/s)
Precipitation	18.6	Ground-water pumpage	3.65
Boundaries	6.10	Boundaries	3.41
Recharge from streams	0	Discharge to streams	17.6
Total inflow	24.7	Total outflow	24.7

effect on streamflow in comparison to pumpage from shallower aquifers. Also, the Ancell aquifer has the most boundary inflow, which provides a source of water for future pumpage development. The most heavily used and the most affected part of the ground-water-flow system is the Silurian-Maquoketa aquifer (layers 6 and 7). Pumping from this aquifer is five times greater than pumping from the Ancell aquifer.

The flow budget for the Campton Township portion of the model (table 12) indicates that recharge from precipitation to the shallow parts of the ground-water-flow system is greater within the township than in the rest of the model area. Campton Township represents 33 percent of the overall modeled area, and, similarly, accounts for 35 percent of the recharge from precipitation to the model. However, only 69.3 percent of the recharge entering Campton Township is discharged to local streams, as compared to 94.6 percent of recharge for the entire model. These differences may be because of the proximity of the township to major surface-water drainage divides, which can be areas of increased recharge to ground water. Within the township, 20.5 percent of the ground water goes to pumpage, 43.4 percent is discharged to streams, and 36.1 percent crosses the township boundaries and flows to the rest of the model area. These percentages can be used as a baseline for future ground-water development. For example, the effects of projected future pumpage on the ground-water-flow system can be determined and potentially minimized by considering the results of simulations of various well locations and pumpage volumes.

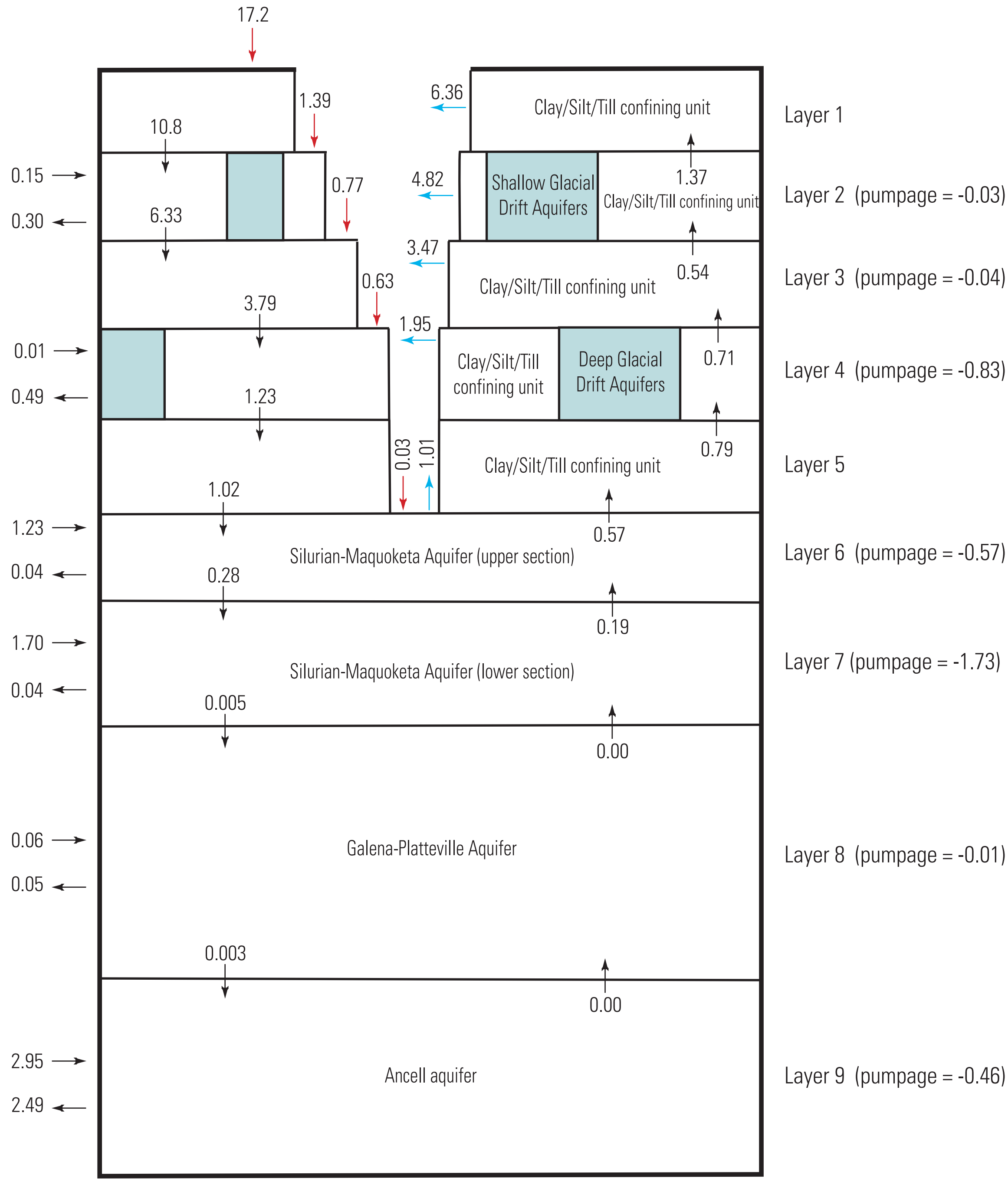
Simulated Water Levels

Water levels were simulated in the model layers representing the aquifers and confining units making up the ground-water-flow system. Simulated water levels for the shallow glacial drift aquifers (layer 2) and the deep glacial drift aquifers (layer 4) are presented in figure 28. Simulated water levels for the lower part of the Silurian-Maquoketa aquifer (layer 7) and the Ancell aquifer (layer 9) are presented in figures 29a and 29b, respectively. The contour lines indicate general ground-water-flow direction (flow is from higher to lower water levels) and the effect of pumping on water levels in each of the aquifers represented by model layers. Residential well locations are included to show the correlation between density of residential wells and the cumulative water-level drawdowns. Closed contours appear along streams (fig. 28a) because of changes in the slope of the stream channel. The deep glacial drift aquifers (layer 4) are affected by domestic pumpage (fig. 28b), primarily in the north-central part of the township. The largest effect of residential-well pumpage is simulated in the lower section of the Silurian-Maquoketa aquifer (layer 7) in the northeastern part of the township (fig. 29a). Model-simulated water-level altitudes are determined for individual layers, but not layer combinations, such as layers 6 and 7 representing the Silurian-Maquoketa aquifer. Therefore, model-simulated water levels cannot be directly compared to the measured

Table 12. Water budget associated with the model calibration of the ground-water-flow sytem for the Campton Township portion of the model, Illinois.

[ft^3/s, cubic feet per second]

Source of flow into the model	Inflow rate (ft^3/s)	Source of outflow from the model	Outflow rate (ft^3/s)
Precipitation	6.45	Ground-water pumpage	2.11
Boundaries	3.87	Boundaries	3.72
Recharge from streams	0	Discharge to streams	4.47
Total inflow	10.3	Total outflow	10.3



EXPLANATION

FT³/S CUBIC FEET PER SECOND

17.2 RECHARGE RATE, IN FT³/S

ARROW REPRESENTING RECHARGE

6.36 STREAM DISCHARGE, IN FT³/S

ARROW REPRESENTING STREAM DISCHARGE

3.79 DOWNWARD FLOW, IN FT³/S

0.71 UPWARD FLOW, IN FT³/S

1.23 LATERAL FLOW INTO MODEL, IN FT³/S

0.04 LATERAL FLOW OUT F MODEL, IN FT³/S

Figure 27. Water budget for the calibrated ground-water model, Campton Township and surrounding area, Illinois.

potentiometric surface of the Silurian-Maquoketa aquifer shown in figure 19. Nevertheless, the drawdown cone in northeastern Campton Township shown in figure 19 can be seen in figure 29a. Pumping from production wells in the township has some effect on the water-level altitude in the Ancell aquifer (layer 9)(fig. 29b).

Simulated Flow Paths

Ground-water-flow paths, which are perpendicular to the simulated water-level contours shown in figures 30 and 31, provide useful information about the source, distribution, and discharge of ground water in the model area. The flow paths are shown areally in figure 30 and in cross section through the model along row 40 (see fig. 23) in figure 31. Areal recharge in the form of precipitation enters the ground-water-flow system throughout the model area and usually discharges to streams. Some flow paths end before they enter a stream, indicating the flow is diverted to a pumping well. The flow lines are very densely drawn in figure 31 because they are from the entire modeled area and are projected onto row 40. Also, most of the flow lines shown in cross section (fig. 31) do not begin at the top of layer 1 because most of the lines begin along rows other than row 40, where water-surface altitudes differ from those at row 40. Most of the areal recharge in the model area flows through layers 1 through 6 (unconsolidated units and the upper part of the Silurian-Maquoketa aquifer) as indicated in figure 31. Additional flow-path analysis indicated that recharge in the model area also reaches the lower part of the Silurian-Maquoketa aquifer (layer 7), but the limited number of flow lines shown in figure 31 (a flow line every fifth row and column) do not indicate the movement of substantial volumes of water deeper into the ground-water-flow system. Simulated recharge does not reach the Galena-Platteville (layer 8) and Ancell aquifers (layer 9); ground-water flows into these aquifers from areas west of the model. The flow path analysis (figs. 30 and 31) and flows described in the water budget (fig. 27) indicate the unconsolidated materials and the upper section of the Silurian-Maquoketa aquifer do not appreciably interact with that part of the flow system in the Galena-Platteville and Ancell aquifers.

Ground-water-flow paths also can indicate the probable source of water to wells. Locations of residential and production wells within Campton Township are shown in figure 32. The points at which water flowing to the wells enters the ground-water system (recharge points) are shown in figure 33. The only areally extensive recharge area within the township is in the area of a ground-water divide in the west-central part. Many of the recharge locations plot as a straight line along column 1, row 1, and row 79 (figs. 23 and 33). This plotting pattern indicates that the actual recharge locations are outside the model, primarily to the north and west. Most of the water that is pumped from the lower section of the Silurian-Maquoketa aquifer, and all of the water pumped from the Galena-Platteville and Ancell aquifers, originates as recharge to ground water beyond the boundaries (west) of the model.

Sensitivity Analysis

The purpose of a sensitivity analysis of a model is to determine the input parameters that most affect simulated water levels. If certain parameters substantially affect simulated water levels, then accurate estimates of the value of these parameters is important to accurate model predictions. The process of automated parameter estimation provides three types of sensitivity-related information: dimensionless-scaled sensitivity, 1-percent scaled sensitivities, and composite-scaled sensitivities (Hill, 1998). The three types of sensitivity information reflect the degree of change in simulated water levels for a given change in an input parameter. The dimensionless-scaled sensitivities can be used to evaluate the importance of observations to the estimation of each of the parameters. This type of information is useful to guide model calibration. One-percent scaled sensitivities measure the variation in sensitivity of a parameter throughout the model. One-percent scaled sensitivities are calculated for each node of the model, and the sensitivities can be mapped. The areas of larger sensitivity for a parameter are good locations for obtaining additional observations to improve the parameter estimate. The composite-scaled sensitivities measure the sensitivity of simulated water levels to perturbations in each model parameter on the basis of all available observations. A large value of composite-scaled sensitivity associated with a specific parameter indicates that the available set of observations is more useful in estimating those parameters. The composite-scaled sensitivities for the calibrated model parameters and other parameters are listed in table 13.

Large composite-scaled sensitivities are desirable for parameters that greatly affect the estimation of water levels from pumpage, such as the Kh of an aquifer. Large sensitivities are associated with the Kh parameters **ksg, kam,** and **kshale** (layers 2, 4, 6, and 7) and with the vertical hydraulic conductivity parameters **kvclay** and **kvam**. These parameters affect the horizontal and vertical flow through the

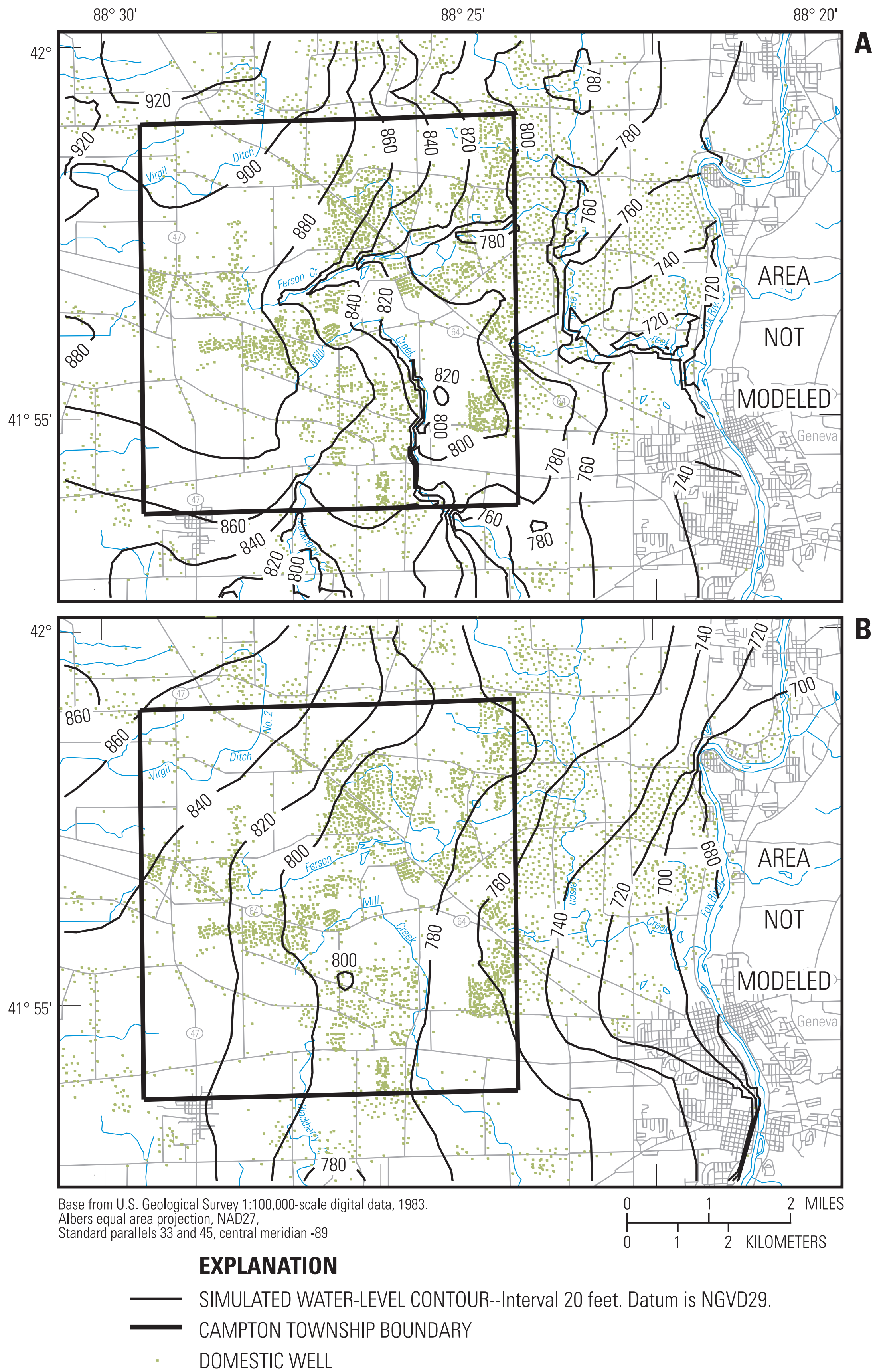


Figure 28. Model-simulated water levels for the A) shallow glacial drift aquifer (layer 2), and B) deep glacial drift aquifer (layer 4), Campton Township and surrounding area, Illinois.

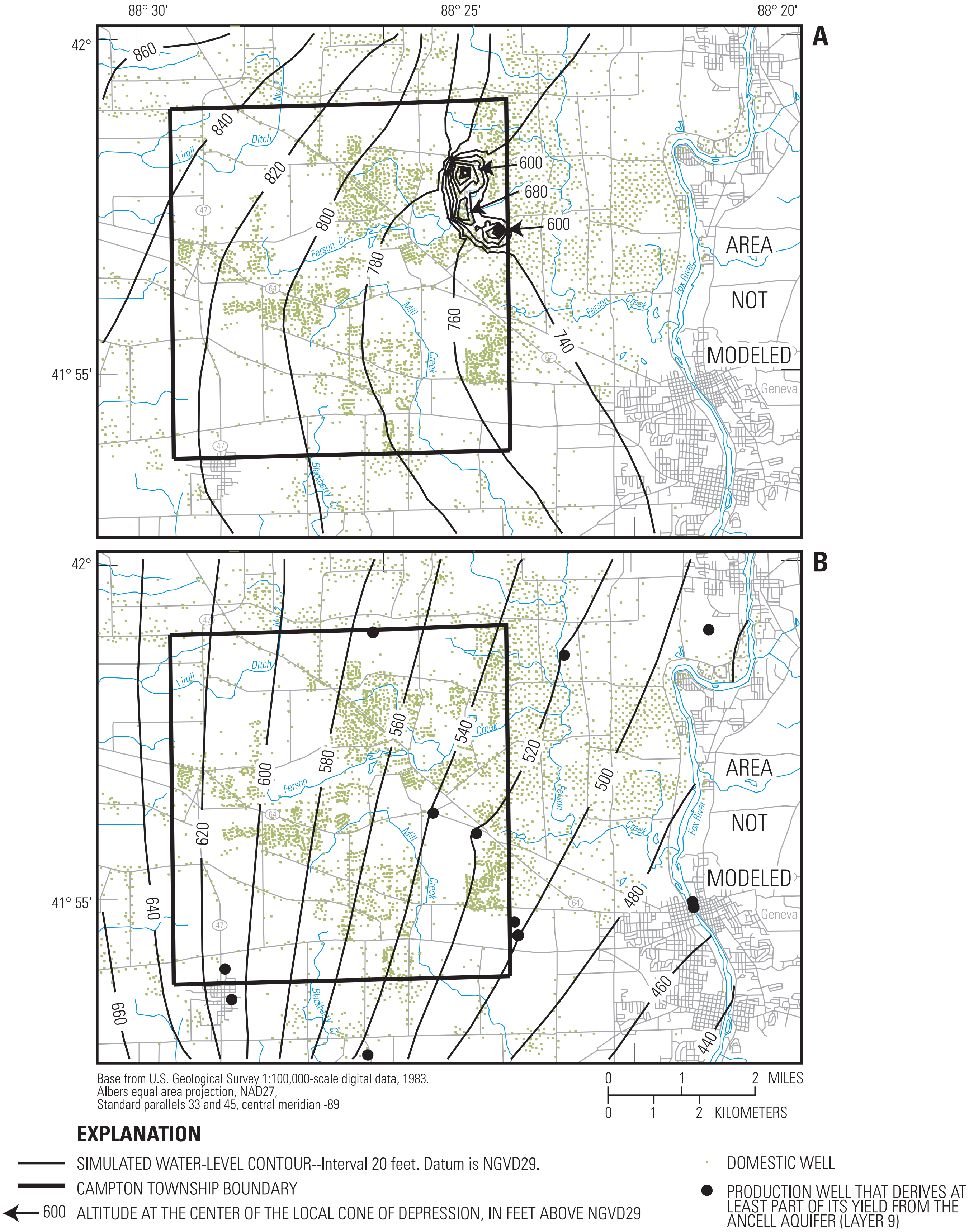


Figure 29. Model-simulated water levels for the A) lower section of the Silurian-Maquoketa aquifer (layer 7), and B) Ancell aquifer (layer 9), Campton Township and surrounding area, Illinois.

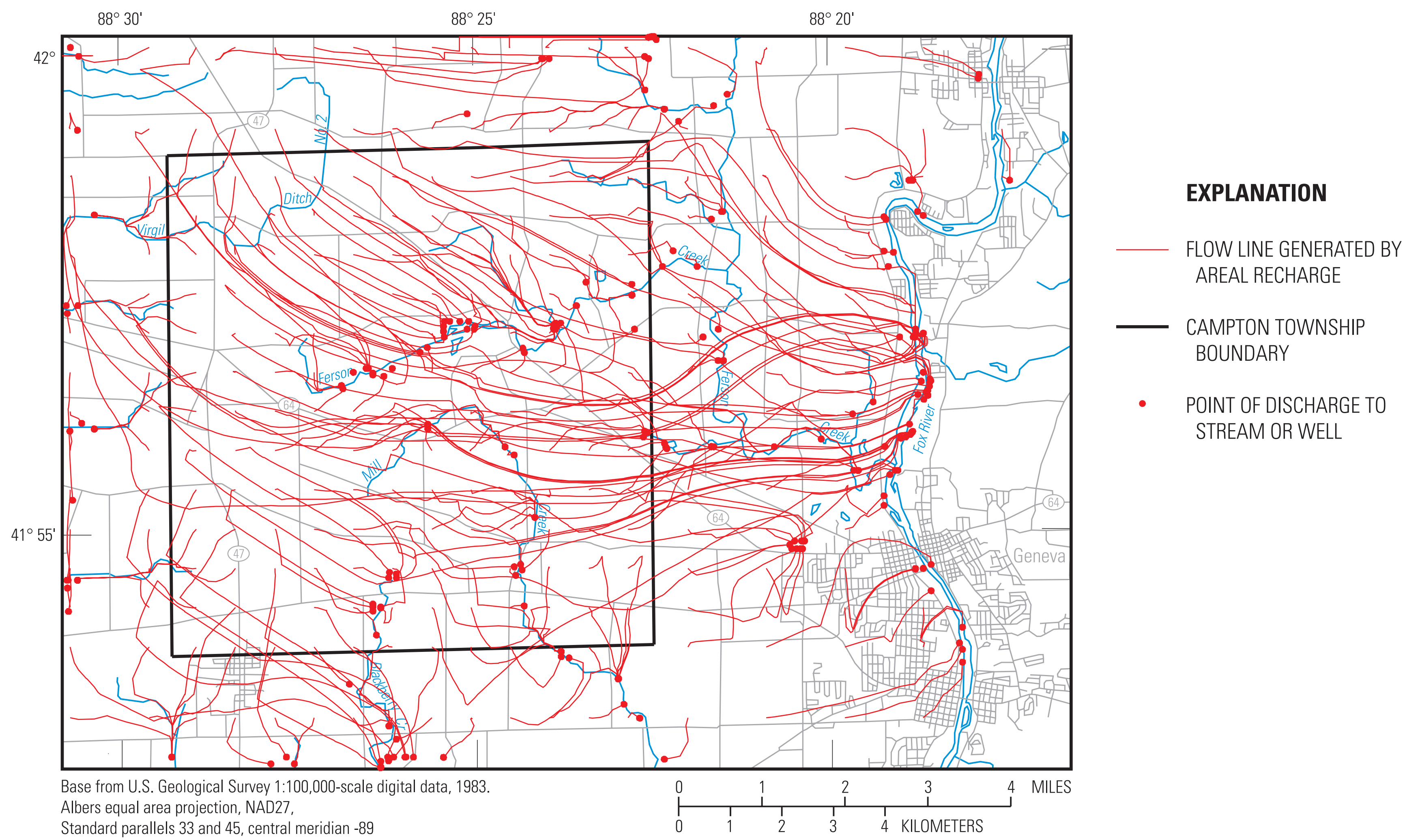


Figure 30. Simulated flow paths from points of recharge to points of discharge in Campton Township and surrounding area, Illinois.

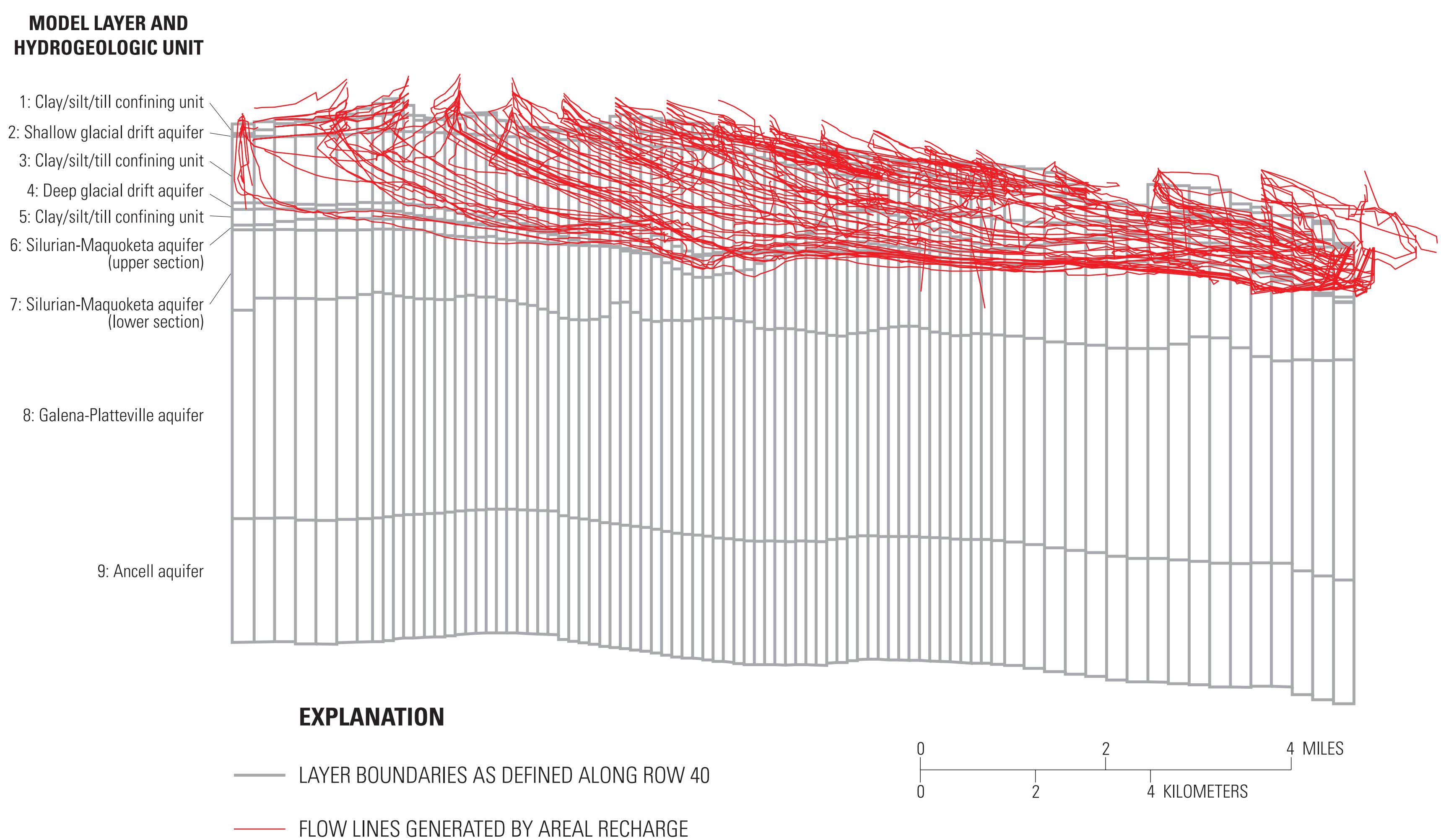


Figure 31. Flow paths from points of recharge to points of discharge shown in vertical section projected in the model grid along row 40 in the model of Campton Township and surrounding area, Illinois.

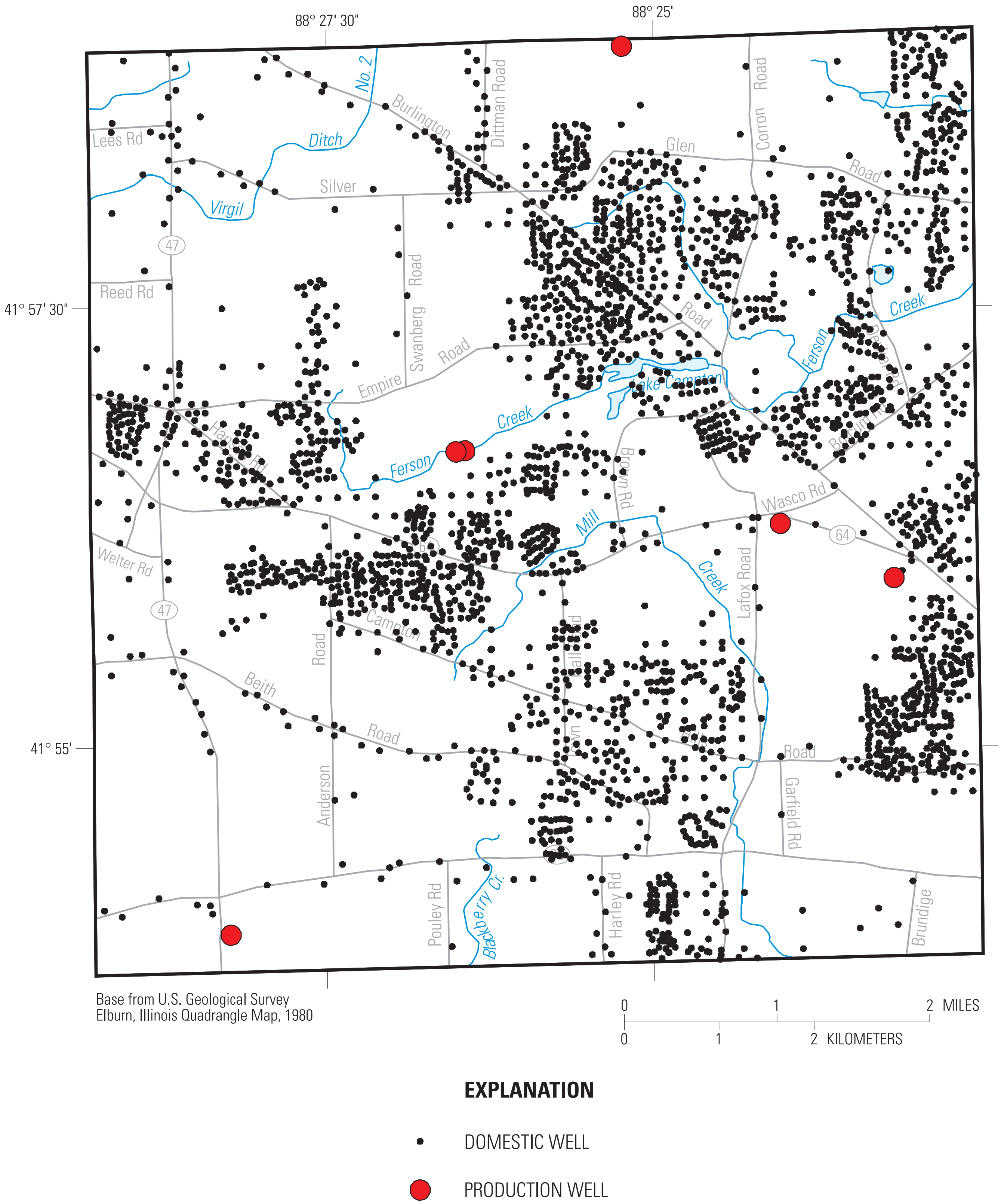


Figure 32. Locations of residential and production wells in Campton Township, Illinois.

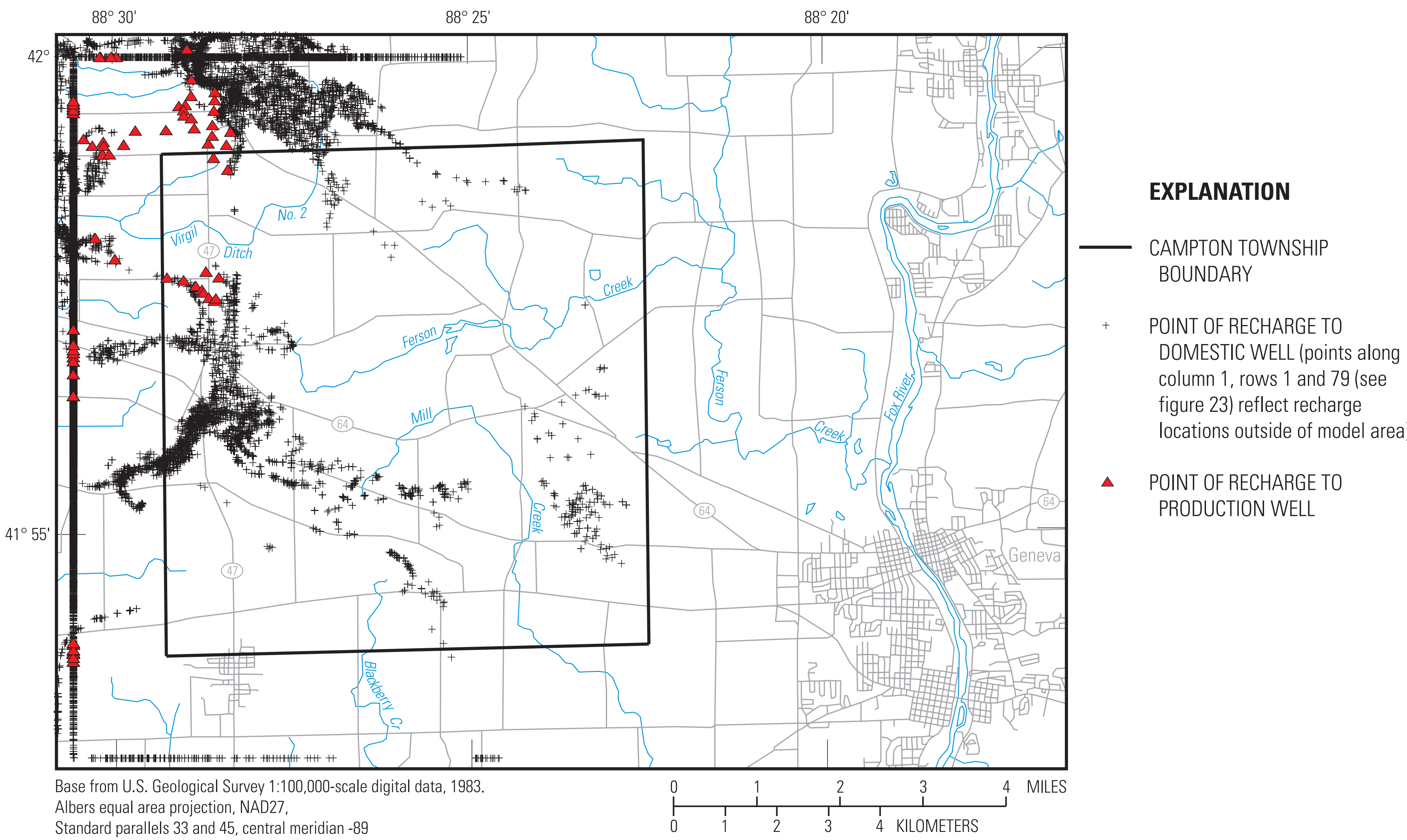


Figure 33. Source areas for residential and production wells, Campton Township and surrounding area, Illinois.

models layers representing the unconsolidated units and the upper section of the Silurian-Maquoketa aquifer and any predictive simulations involving pumping in these layers. The composite-scaled sensitivity for **k9**, the Kh of the Ancell aquifer (layer 9) is low, and, as a result, could not be determined by parameter estimation. More measured water levels unique to the Ancell aquifer would improve the sensitivity number for **k9**, and, subsequently, improve estimates of Kh of the aquifer and predicted water-level decline caused by new pumpage simulated in layer 9. The Kh value of the Ancell aquifer used in model layer 9 (5 ft/d) is an intermediate value of available data, and should provide reasonable estimates of drawdown caused by predictive pumping simulations. The sensitivities of the parameters for Kh of the glacial drift aquifers and the streambeds are low, indicating changes in their Kh values do not substantially change water levels throughout the model. The main recharge parameter, **rcht** (recharge to fine-grained material) is sensitive, indicating that a reliable estimate of flow into the model is possible. The uncalibrated values of Kh for streambeds are reasonable values based on values used in past modeling studies in similar areas (Meyer and others, 1975; Lapham, 1981: Arihood and Cohen, 1998).

Results of Predictive Simulations

Population growth in Campton Township has increased interest by the local government in understanding the effect of increased pumping on ground-water resources. Of specific interest to Campton Township is whether additional water supply can be obtained from the Ancell aquifer (model layer 9) and the effects of additional pumpage on water levels in the aquifer. Township records provided the number of planned new, unsold homes as of May 2004, and whether the homes will be supplied by individual residential wells or by available production wells. The records indicate that 140 of the new homes will be supplied by residential wells and 465 of the new homes will be supplied by available production wells. The Wasco Sanitary District estimated the per household water consumption in 2002 to be 304 gal/d (0.00047 ft^3/s). That pumpage rate was applied in the model node nearest the location of new homes supplied by their own well. The 304 gal/d was multiplied by 465 and applied to two available production wells. The additional withdrawals from the production wells was distributed such that both wells withdrew an equal amount of 0.216 ft^3/s, (original plus additional withdrawals). The location of the new homes supplying their own water, location of the two production wells, and simulated water-level contours for the Ancell aquifer, both before and after the additional with-

Table 13. Composite-scaled sensitivities of select model parameters for simulation of the ground-water-flow system, Campton Township, Illinois.

Parameter	Parameter name	Composite- scaled sensitivity
Horizontal hydraulic conductivity of clay, silt, and till units (layers 1-5)	kclay	11.7
Vertical hydraulic conductivity of clay, silt, and till units (layers 1-5)	kvclay	21.2
Horizontal hydraulic conductivity of glacial drift aquifers (layers 2 and 4)	ksg	7.62
Horizontal hydraulic conductivity of the Silurian-Maquokekta aquifer (layers 6 and 7)	kam	27.3
Vertical hydraulic conductivity of the Silurian-Maquoketa aquifer (layers 6 and 7)	kvam	25.9
Horizontal hydraulic conductivity of the Silurian-Maquoketa aquifer in the area of large water-level residuals (layers 6 and 7)	kshale	20.0
Horizontal hydraulic conductivity of the Galena-Platteville aquifer (layer 8)	k8	.001
Vertical hydraulic conductivity of the Galena-Platteville aquifer (layer 8)	kv8	.53
Horizontal hydraulic conductivity of the Ancell aquifer (layer 9)	k9	.41
Vertical hydraulic conductivity of the Ancell aquifer (layer 9)	kv9	1.33 * 10-9
Recharge rate to the fine grained glacial units (layers 1-5)	rcht	12.9
Vertical hydraulic conductivity of all glacial drift aquifers (layers 2 and 4)	kvsg	.005
Hydraulic conductivity of small- sized drain streambeds	d20	.016
Hydraulic conductivity of interme¬diate-sized drain sreambeds	d10	.001
Hydraulic conductivity of the Fox River streambed	r1	.002
Recharge to the surficial outwash (layer 1)	k1o	2.92

drawals, are shown in figure 34. A comparison of model-simulated water-level contours indicates the effect of the additional simulated withdrawals (fig. 35). The greatest drawdown (16.9 ft) is simulated at the northwest production well because the greatest increase in pumpage (0.20 ft^3/s) was simulated there. The additional pumpage in the Ancell aquifer (layer 9) did not cause any measurable predicted drawdown in the overlying aquifers. The lack of predicted additional drawdown in the overlying model layers indicates that the flow in the Ancell aquifer originates as recharge beyond the model boundaries and not as areal recharge from overlying units within the township. Average drawdown in the Ancell aquifer at the model boundaries is only 0.13 ft, indicating that the boundaries are not substantially affecting maximum drawdowns at pumping centers.

Confidence intervals on the predicted drawdowns were calculated, but uncertainty in some parameters result in a large range in the 95-percent confidence limit. In this case, the 95-percent confidence limit was not useful. Prediction scaled sensitivities indicate the parameters that are most important to these predictions. The prediction scaled sensitivities indicate that the Kh of the Ancell aquifer was about two orders of magnitude more important than any other model parameter in determining the drawdown predictions. Additional Kh data and reliable discrete water-level measurements for the Ancell aquifer would be the most beneficial data for increasing the accuracy of model predictions of withdrawals from the aquifer.

Model Limitations and Qualifications

The model has the greatest certainty in the accuracy of simulation in the model layers with the most water-level measurements because model parameters have been best estimated in those layers. The layers with the most observations are layer 4 (representing the deep glacial drift aquifers) and layers 6 and 7 (representing the Silurian-Maquoketa aquifer). The Kh of the Ancell aquifer (layer 9) could not be estimated with certainty because of the relatively small number of measurements. Even with the large number of wells with measured water levels, only a few multi-aquifer wells penetrated the Ancell aquifer. Additional water-level measurements from the Ancell aquifer, especially from single-aquifer wells and near production wells, would appreciably improve model calibration and predictive simulations. This and previous studies estimated the Kh of the Ancell aquifer from 1 to about 10 ft/d. Predicted drawdowns presented here are based on a Kh of 5 ft/d; therefore, if the Kh is actually 10 ft/d, the predicted drawdowns would roughly be reduced by half.

Records of public and private pumping withdrawals provided some measurement of ground-water flow in deeper parts of the model. However, the uncertainty in inter-aquifer flow within boreholes of residential wells limited the value of the pumpage records. Fortunately, tests of different assumptions of borehole flow did not substantially change model parameter estimates.

The method of calculating a composite simulated water level for comparison to measured levels from wells penetrating multiple aquifers represented by multiple model layers does not incorporate the effect of borehole flow on the simulated water levels. Only the transmissivity of the individual model layers is used to determine the effect of an individual layer on the composite simulated water level. Although the method used to compare simulated to measured water levels does not consider all important parameters, transmissivity is considered the most important parameter to accurate simulation of ground-water flow in Campton Township.

In small areas, the unweighted residuals were consistently high or low by tens of feet. The reason for the large water-level residuals is probably related to local variations in geology such as fracture density in the bedrock aquifers. Available information, such as geologic maps and well logs, do not provide sufficient detail for estimating these local variations. Therefore, available sources of information can not be used to predict where other local variations may occur, and simulated water levels could be high or low in such areas. The only way to reduce the large water-level residuals would be to collect additional field data (water levels and hydraulic parameters from aquifer tests) throughout the township. The average model uncertainty can be incorporated into the development of a "safety factor" for ground-water development. For example, if a predictive simulation resulted in 20 ft of drawdown in the lower section of the Silurian-Maquoketa aquifer (model layer 7), an additional 20.2 ft (overall model fit) of drawdown may result. In this example, the total drawdown (considering the "safety factor") would be 40.2 ft.

All aquifers represented by model layers have been simulated as confined layers, including layers under water-table (unconfined) conditions. For layers under water-table conditions, the thickness of the confined layers has been adjusted to be equivalent to the thickness of a layer under water-table conditions. Quality assurance of predictive simulations should include a check that aquifers simulated in model layers do not become substantially (greater

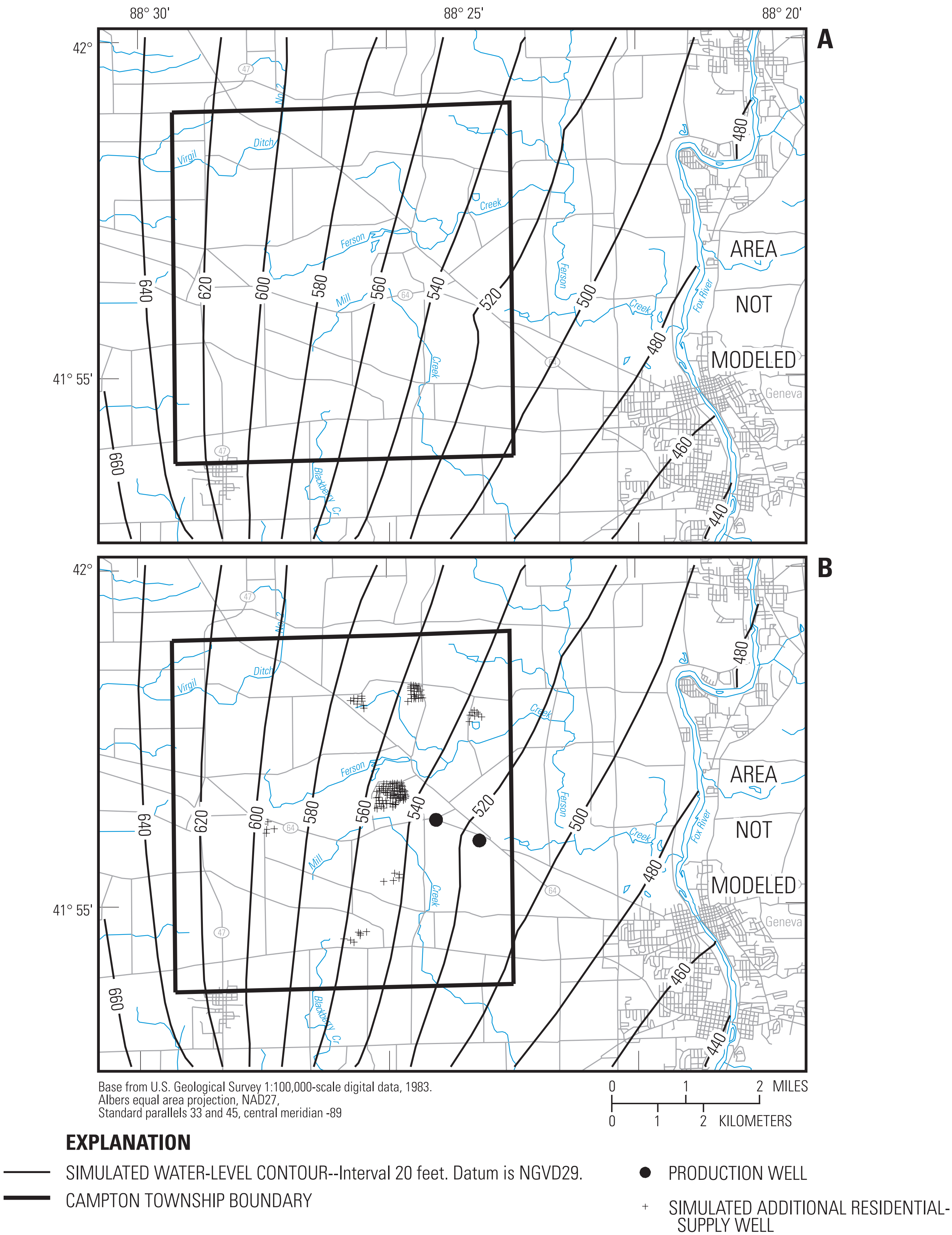
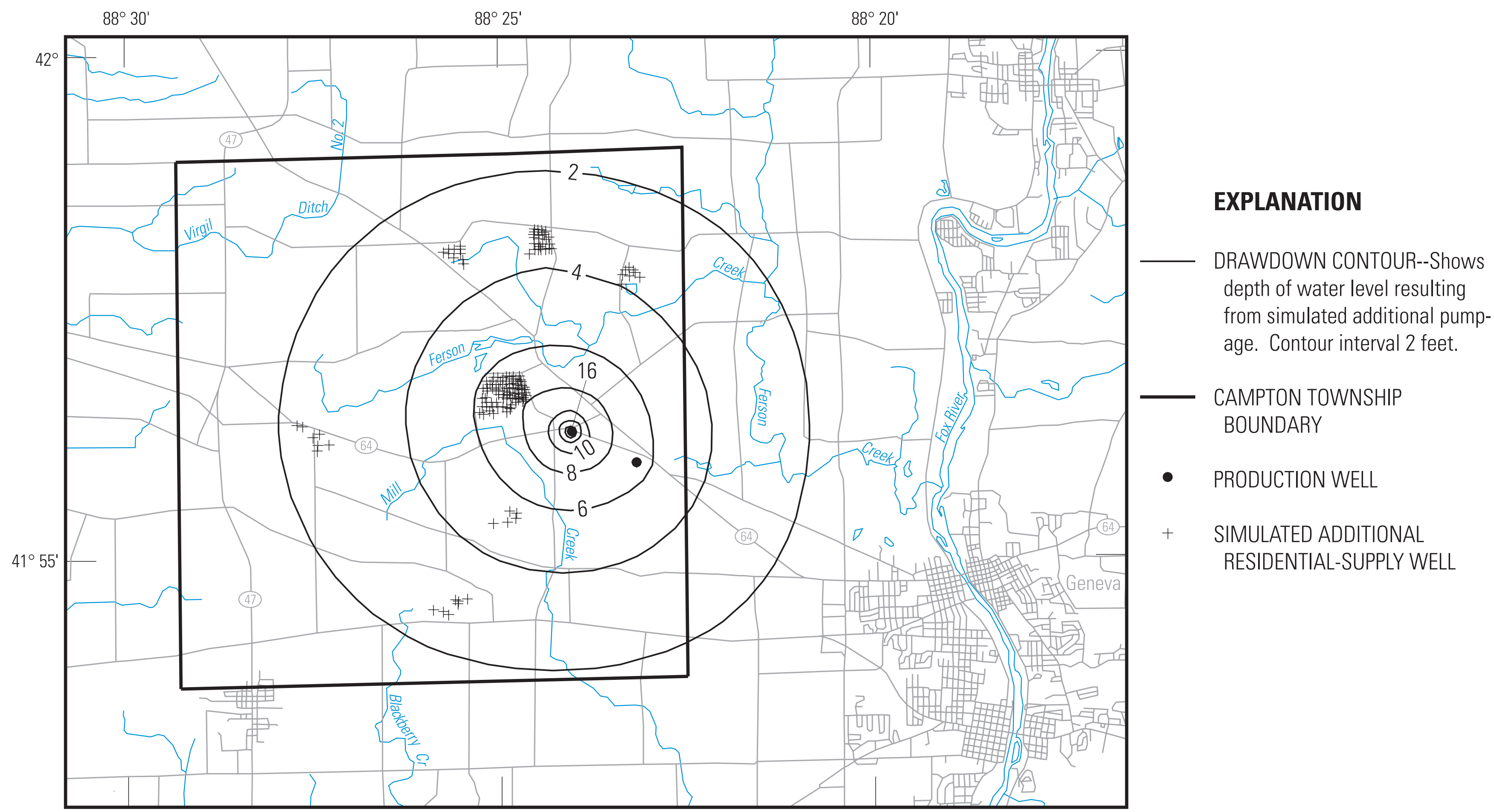


Figure 34. Model-simulated water-levels in the Ancell aquifer (layer 9), A) before and, B) after additional withdrawals, Campton Township and surrounding area, Illinois.



Base from U.S. Geological Survey 1:100,000-scale digital data, 1983.
Albers equal area projection, NAD27,
Standard parallels 33 and 45, central meridian -89

Figure 35. Drawdown in the Ancell aquifer caused by simulated additional withdrawals in Campton Township and surrounding area, Illinois.

than 10 percent) dewatered. If substantial dewatering results, then predicted drawdowns would be underestimated. In such cases, aquifer thicknesses represented by model layers should be adjusted.

SUMMARY AND CONCLUSIONS

Population growth in Campton Township, Kane County, Illinois has caused concern about the availability of water supplies from aquifers used for residential supply. In 1995, the U.S. Geological Survey, in cooperation with the Campton Township Board of Trustees, began a study of regional ground-water flow in Campton Township and the surrounding area. The study included examination of water levels in order to determine the current (2002) distribution of, and trends in, water levels in the shallow and deep glacial drift aquifers, the Silurian-Maquoketa aquifer, the Galena-Platteville aquifer, and the Ancell aquifer.

Water-level altitudes were measured in more than 225 residential-supply wells open to one or more of the aquifers in Campton Township during June and July 2002. Changes in water level between 1995 and 2002 were calculated by comparing measurements in a well taken in 1995 with measurements taken in the same well in 2002.

The water-level altitude in the wells open to the shallow glacial drift aquifers ranged from about 787 to 923 ft . Water-level altitudes in wells open to the deep glacial drift aquifers were typically above 800 ft in the northwestern part of the township, between 775 and 800 ft in the central part of the township, and less than 775 ft in much of the southern and eastern parts of the township, and approximately mirrored surface topography. Water-level changes in the shallow and deep glacial drift aquifers were less than 5 ft in most of Campton Township between 1995 and 2002. The deep glacial drift aquifers may warrant additional study in an area southwest of Lake Campton because of some potentially substantial water-level declines. The glacial-drift aquifers appear to be capable of sustaining additional withdrawals particularly beneath the eastern part of the township.

Water-level altitude in wells completed in the Silurian-Maquoketa aquifer generally decrease from west to east in the township. Local areas of low water-level altitudes near residential areas indicate that more water is removed from the Silurian-Maquoketa aquifer by pumping and vertical flow through wells open to the underlying aquifers than is recharged to the Silurian-Maquoketa aquifer, particularly in residential areas in the eastern part of the township. Additional withdrawals from the Silurian-Maquoketa aquifer for water supply should be considered with care.

Water-level altitudes in the wells open to the Ancell aquifer range from 253 to 722 ft. The large range in water-level altitudes in the wells open to the Ancell aquifer can be attributed, at least in part, to differences in the amount of inflow to the well from overlying aquifers, because these wells typically are open to multiple aquifers. Water-level declines in the Ancell aquifer over a period of decades have been indicated beneath the township. Determination of the cause (natural or anthropogenic) and magnitude of this decline cannot be accurately determined without additional investigation. The Ancell aquifer appears to have the capacity of supplying additional residential water use in the township.

A nine-layer ground-water-flow model was constructed to simulate ground-water flow in the unconsolidated materials and in the bedrock aquifers used for residential supply. Data sources for model simulation included driller's logs, geologic maps, aquifer test data, and 217 water-level and 12 streamflow measurements. Model calibration resulted in representing the horizontal hydraulic conductivity of the Silurian-Maquoketa aquifer with a single parameter. Model predictions should be most accurate, on average, in the model layers representing the aquifers with the most observations, which were layers 4 (representing the deep glacial drift aquifers), and 6 and 7 (representing the Silurian-Maquoketa aquifer).

Model simulations revealed information on flow patterns and drawdown response. For example, 75 percent of the inflow to the model is areal recharge from local precipitation. This recharge flows through the unconsolidated materials (model layers 1-5) and the upper section of the Silurian-Maquoketa aquifer (model layer 6). Little local recharge flows into the Galena-Platteville or Ancell aquifers. Simulation of additional pumpage from 605 new homes supplied primarily from two existing production wells completed in the Ancell aquifer is predicted to cause 16.9 ft of drawdown in the Ancell aquifer, but virtually no drawdown in the overlying aquifers. The source of the water pumped from the Ancell is from the aquifer in areas west of the modeled area. The source of water to wells pumping from the shallow and deep glacial drift aquifers and the Silurian-Maquoketa aquifer is mainly from the area along the western part of Campton Township and northwest of the township. Within the township boundaries, 20.5 percent of the ground water is discharged to pumpage, 43.4 percent is discharged to streams, and 36.1 percent flows across the township boundaries to the rest of the model area.

On the basis of the large differences in values for the hydraulic conductivity parameters of the Silurian-Maquoketa aquifer, the productivity of this aquifer can be considered highly variable. Water-level measurements and aquifer tests that can reveal the variability in the hydraulic conductivity in this aquifer are not available everywhere, and areas of low hydraulic conductivity beyond those identified are possible. A sufficient number of residential homes pumping from the Silurian-Maquoketa aquifer in new areas could cause large drawdowns, as seen in northeastern Campton Township. Additional aquifer tests and water-level measurements in new residential areas, particularly in wells completed in the Ancell aquifer, would improve the definition of the ground-water-flow system and the accuracy of the model simulation.

Acknowledgments

The authors thank the local residents who allowed the USGS access to their wells for data collection. The authors also thank John Pree, Denise Morgan, Ronald Petrucci, Daniel Korth, Fredrick Robinson, William Lovis, Gregory van Zandt, Susan Clarkson, Paul Pellico, James Horlock, Dudley Case, and Pamela Reed for their assistance with collection of water levels. Denise is especially thanked for her assistance with the data compilation.

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GLOSSARY

Alluvial—pertaining to or composed of alluvium, or deposited by a stream or running water.

Aquifer—a formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield appreciable quantities of water to wells and springs.

Aquifer test—a test to determine hydraulic properties of the aquifer involving the withdrawal of measured quantities of water from or addition of water to a well and the measurement of resulting changes in water level in the aquifer both during and after the period of discharge or additions.

Arenite—a "clean" sandstone that is well-sorted, contains little or no matrix material, and has a relatively simple mineralogic composition; specifically, a pure or nearly pure, chemically cemented sandstone containing less than 10 percent argillaceous matrix

Argillaceous—containing clay minerals.

Bedrock—the solid rock that lies below soil and other loose surface materials.

Caliper log—a continuous record of hole diameter.

Cambrian System—period of geologic time extending from about 570 to 500 million years before the present.

Chert—a rock type composed primarily of cryptocrystalline silica.

Clay—detrital particles less than 1/256 millimeter in size or an unconsolidated detrital sedimentary deposit composed primarily of particles less than 1/256 millimeter in size.

Confined—a modifier that describes a condition in which the potentiometric surface is above the top of the aquifer.

Confining unit—a hydrogeologic unit of impermeable or distinctly less permeable material stratigraphically adjacent to one or more aquifers.

Drawdown—the decline in water level at a point caused by the withdrawal of water from a hydrogeologic unit.

Drift—all glacial and fluvioglacial deposits left after the retreat of glaciers and ice sheets.

Dolomite—rock composed of calcium and magnesium carbonate.

Evapotranspiration—the combined loss of water from a given area by evaporation from the land and transpiration from plants.

Flowmeter log—a device used to measure the volume and direction of vertical flow at various depths within the borehole.

Fluid resistivity logs—logs that are designed to make quantitative measurements of the specific resistance of a fluid to the flow of electric current.

Formation—aggregation of related strata distinguishable from beds above and below and of mappable extent.

Fracture—breakage in the rock not related to the crystalline structure of the minerals that compose the rock, often having a preferred orientation.

Geophysical log—a record of the chemical or physical properties of fluids or rocks at a well site, such as thermal or electrical properties, density, resistivity, sonic potential and gamma-ray activity. The record is obtained by dropping a tool down the borehole and recording the property values as a function of depth.

Gravel—coarse-grained particles between 2 and 4 millimeters in size. Loosely used to denote unconsolidated detrital sedimentary deposits composed primarily of particles greater than 2 millimeters in size.

Ground water—subsurface water that fills available openings in rock or soil materials to the extent they are considered water-saturated.

Group—the lithostratigraphic unit next in rank above formation, consisting partly or entirely of two or more adjacent formations having prominent features in common.

Hydraulic conductivity—a proportionality constant relating hydraulic gradient to specific discharge, which for an isotropic medium and homogeneous fluid equals the volume of water at the prevailing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow.

Joint—a break in rock mass with no relative movement on opposite sides of the break.

Kame—a low mound, knob, hummock, or short irregular ridge, composed of stratified sand and gravel deposited by a subglacial stream as a fan or delta at the margin of a melting glacier; or as a ponded deposit on the surface or at the margin of stagnant ice; or by a supraglacial stream in a low place or hole on the surface of the glacier.

Lacustrine—pertaining to lakes.

Lithologic log—description of the geologic deposits encountered during drilling.

Lithology—the description of rocks on the basis of color, structure, mineral composition, and grain size; the physical character of a rock.

Limestone—sedimentary rock composed primarily of calcium carbonate.

Member—units of lesser rank in a heterogeneous formation that are lithologically distinct.

Model—a simplified representation of the appearance or operation of a real object or system. In a ground-water-flow model, the ground-water-flow system is represented by a computer model that mathematically approximates the flow system. Governing equations indirectly represent the physical processes that occur beneath the ground and between ground water and surface water.

Moraine—a mound, ridge, or other distinct accumulation of glacial drift, predominantly till, deposited chiefly by direct action of glacial ice in a variety of topographic landforms that are independent of control by the surface on which the drift lies.

Natural gamma log—a process whereby gamma rays naturally emitted by formations traversed by a borehole are measured by a tool containing a radiation detector that is lowered into the borehole and records the gamma rays detected in correlation with depth in the borehole.

Normal resistivity logs—any of a group of logs that are designed to make quantitative measurements of the specific resistance of a material to the flow of electric current.

Ordovician System—period of geologic time extending from about 500 to 435 million years before the present.

Outwash—material deposited by streams eminating from a melting glacier.

Peat—a partially decomposed mass of vegetation that has typically grown in a marsh or shallow lake.

Permeabilityva measure of the relative ease with which a porous medium can transmit a fluid under a potential gradient and is a property of the medium alone.

Physiography—the study and classification of the surface features of earth on the basis of similarities in geologic structure and the history of geologic changes.

Potentiometric surface—an imaginary surface representing the static head of ground water and defined by the level to which water will rise in a tightly cased well.

Quartz—mineral consisting of silicon dioxide.

Quaternary System—the latest period of geologic time represented in Illinois by accumulations of glacial (Pleistocene) and post-glacial (Holocene) deposits that were deposited above the Pliocene series.

Range—one of the north-south rows of townships in a U.S. public-land survey that are numbered east and west from the principal meridian of the survey

Recharge—the process of addition of water to the saturated zone or a well.

Sand—typically a silicate mineral between 1/16 and 2 millimeters in size or an unconsolidated detrital sedimentary deposit composed primarily of particles between 1/16 and 2 millimeters in size.

Sandstone—detrital sedimentary rocks composed primarily of silicate minerals that typically vary in size between 1/16 and 2 millimeters.

Section—a piece of land one square mile in area forming one of the 36 subdivisions of a township

Shale—detrital sedimentary rock composed primarily of particles less than 1/256 millimeter in size.

Silt—particle between 1/16 and 1/256 millimeter in size or an unconsolidated detrital sedimentary deposit composed primarily of particles between 1/16 and 1/256 millimeter in size.

Silurian System—period of geologic time extending from about 435 to 410 million years before the present.

Solution opening—a large cavity in a rock formed by chemical dissolution.

Spontaneous potential log—is a record of potentials or voltages that develop at the contacts between shale or clay beds and a sand aquifer, where they are penetrated by a drill hole.

Steady state—a condition in which flow velocity at any point in the flow field is a stable through time with respect to magnitude and direction.

Storage coefficient—the volume of water an aquifer releases from storage per unit surface area of the aquifer per unit change in head.

Stratigraphy—the study, definition, and description of major and minor natural divisions of rocks, particularly the study of their form, arrangement, geographic distribution, chronologic succession, classification, correlation, and mutual relations of rock strata.

Till—unsorted, unstratified material deposited beneath glacial ice or deposited from melting glacial ice.

Township—a division of territory in surveys of U.S. public land containing 36 sections or 36 square miles. As described in this report, township is also a unit of local government.

Transmissivity—the rate at which water of the prevailing kinematic viscosity is transmitted through the unit width of the aquifer under a unit hydraulic gradient. Transmissivity is equal to an integration of the hydraulic conductivities across the saturated part of the aquifer perpendicular to the flow paths.

Unconfined aquifer—an aquifer that has a water table.

Unconformable—feature of strata that do not succeed the underlying rocks in immediate order of age or in parallel position. A general term applied to any strata deposited directly upon older rocks after an interruption in sedimentation, with or without any deformation or erosion of the older rocks.

Unconsolidated deposit—geologic material that has not been lithified.

Water table—the upper surface of the zone of saturation on which the water pressure equals the atmospheric pressure.

Well—a bored, drilled, or driven shaft of a dug hole, whose depth is greater than the largest surface dimension.

Yield—volume of water supplied to a receptor, typically a well, per unit time.

APPENDIX A—Summary of well information and water-level measurements taken during the survey of residential-supply wells in Campton Township, Illinois, May-June 1995 and June-July 2002

Appendix A. Summary of well information and water-level measurements taken during the survey of residential-supply wells in Campton Township, Illinois, May-June 1995 and June-July 2002.

[SM, Silurian-Maquoketa aquifer; nm, not measured; DD, deep drift aquifer; SD, shallow drift aquifer; GP, Galena-Platteville aquifer; An, Ancell aquifer; * denotes uncertain data; ?, unknown; <, less than; >, greater than. Bold denotes uncertain data]

Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S01	163-235	SM	820	735.7	nm	nm
S02	147-220	SM	812	726.9	nm	nm
S03	167-220	SM	825	723.5	764.2	40.7
S04	149-205	SM	812	730.3	771.4	41.1
S05	149-220	SM	812	731.6	771.7	40.1
S06	186-250	SM	839	726.0	766.2	40.2
S07	125-147	DD	836	791.3	nm	nm
S08	175-225	SM	830	727.4	766.4	39.0
S09	196-250	SM	847	726.3	766.3	40.0
S10	171-220	SM	826	731.5	771.1	39.6
S11	186-250	SM	834	718.7	nm	nm
S12	44-78	SD	813	794.1	nm	nm
S13	57-63	SD	811	792.6	791.1	-1.5
S14	150-400	SMGP	804	650.0	nm	nm
S15	159-440	SMGP	820	729.0	nm	nm
S16	198-500	SMGP	827	639.4	nm	nm
S17	198-500	SMGP	831	657.3	nm	nm
S18	190-460	SMGP	820	572.0	nm	nm
S19	300-500	SMGP	802	706.2	nm	nm
S20	175-400	SMGP	812	734.5	735.5	1.0
S21	195-470	SMGP	846	640.0	nm	nm
S22	?-470	SMGP	836	595.0	nm	nm
S23	?-158	DD	863	790.1	nm	nm
S24	156-450	SMGP	804	722.3	735.5	13.2
S25	?-143	SD	868	785.1	787.7	2.6
S26	?-116	SD	851	792.4	793.0	.6
S27	165-400	SMGP	800	665.3	725.4	60.1
S28	150-165	DD	802	772.9	772.2	-.7
S29	155-175	DD	807	772.1	771.5	-.6
S30	?-175	DD	814	772.8	771.3	-1.5
S31	162-182	DD	816	768.2	767.3	-.8
S32	?-182	DD	816	770.1	nm	nm
S33	163-445	SMGP	850	674.0	nm	nm
S34	140-150	DD	830	762.6	nm	nm
S35	126-680	SMGPAn	818	715.1	722.2	7.1
S36	124-660	SMGPAn	818	716.0	nm	nm
S37	145-165	DD	792	767.1	766.1	-1.0
S38	88-240	SM	824	793.0	777.7	-15.3
S39	?-250	SM	825	759.1	750.2	-8.9
S40	110-280	SMGP*	834	725.5	717.4	-8.2
S41	120-175	SM	838	787.0	779.5	-7.5
S42	135-185	SM	841	782.4	776.9	-5.5
S43	20-320	SMGP	843	702.0	681.3	-20.7
S44	210-325	SM	850	672.0	nm	nm
S45	208-340	SM	856	684.9	nm	nm
S46	148-240	SM	846	787.8	781.4	-6.4
S47	91-145	SM	830	790.5	nm	nm
S48	76-180	SM	818	796.1	775.7	-20.4
S49	84-115	SM	819	797.3	nm	nm

Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S50	66-200	SM	821	776.2	765.3	-11.0
S51	?-251	DD	896	793.6	799.3	5.7
S52	264-295	SM	900	802.0	798.4	-3.6
S53	?-253	DD	907	816.9	nm	nm
S54	115-210	SM	845	774.0	772.8	-1.3
S55	88-160	SM	814	782.7	781.7	-1.1
S56	84-160	SM	808	782.5	781.6	-.9
S57	72-150	SM	803	785.2	784.8	-.4
S58	101-170	SM	804	802.0	nm	nm
S59	128-240	SM	840	754.3	752.3	-2.1
S60	160-210	SM	860	736.8	nm	nm
S61	>160	SM*	845	746.4	745.3	-1.1
S62	161-260	SM	880	744.7	741.6	-3.1
S63	104-160	SM	784	735.8	nm	nm
S64	133-260	SM	790	700.4	668.3	-32.1
S65	130-180	SM	836	742.5	741.3	-1.2
S66	78-160	SM	783	725.2	nm	nm
S67	165-300	SM	840	696.8	675.8	-21.0
S68	153-270	SM	826	693.2	670.9	-22.3
S69	?-50	SD	826	816.5	815.7	-.8
S70	218-350	SM	893	690.5	<693	?
S71	149-320	SMGP	804	734.3	720.6	-13.7
S72	146-308	SMGP	805	736.8	nm	nm
S73	316-420	GP	837	507.4	nm	nm
S74	166-455	SMGP	824	531.6	nm	nm
S75	165-445	SMGP	818	520.3	nm	nm
S76	163-440	SMGP	831	528.8	nm	nm
S77	172-210	SM	829	774.5	nm	nm
S78	163-200	SM	823	784.0	nm	nm
S79	196-335	SMGP	840	697.3	nm	nm
S80	165-200	SM	814	778.8	nm	nm
S81	203-260	SM	850	765.1	758.8	-6.3
S82	220-280	SM	863	763.8	nm	nm
S83	200-240	SM	861	806.2	802.8	-3.5
S84	?-400	SMGP	819	548.0	516.1	-32.0
S85	110-260	SM	854	786.1	787.5	1.4
S86	50-120	SM	793	781.4	777.9	-3.5
S87	87-200	SM	822	782.5	nm	nm
S88	95-140	SM	836	815.2	nm	nm
S89	85-150	SM	824	807.8	nm	nm
S90	93-375	SMGP	834	784.2	780.0	-4.2
S91	?	DD*	813	781.8	781.1	-.8
S92	53-170	SM	780	763.5	763.2	-.3
S93	94-140	SM	831	767.0	760.0	-7.0
S94	93-200	SM	821	766.7	760.7	-6.0
S95	88-107	SM	805	762.9	762.7	-.3
S96	91-200	SM	813	762.3	763.6	1.3
S97	100-115	SM	827	767.9	nm	nm
S98	?	SM*	840	780.6	762.0	-18.6
S99	110-205	SM	863	817.7	nm	nm
S100	180-195	SM	849	796.8	nm	nm
S101	45-120	DDSM	802	792.5	790.3	-2.2
S102	100-500	SMGP	843	673.0	nm	nm
S103	91-500	SMGP	830	420.0	nm	nm
S104	93-175	SM	833	769.8	nm	nm

Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S105	?	SM*	835	780.2	769.0	-11.2
S106	?-180	SM	842	769.9	753.7	-16.2
S107	172-225	SM	829	780.5	765.6	-14.9
S108	163-220	SM	825	781.2	765.6	-15.6
S109	163-220	SM	827	781.5	766.6	-14.9
S110	162-200	SM	826	780.0	764.6	-15.5
S111	172-260	SM	827	785.2	769.7	-15.6
S112	174-220	SM	826	782.0	766.7	-15.3
S113	169-200	SM	829	783.4	nm	nm
S114	162->342	SMGP*	825	621.0	nm	nm
S115	?	SM*	823	780.6	nm	nm
S116	?->160	SM*	857	802.0	793.6	-8.4
S117	?-665	SMGPAn	808	512.7	<753	nm
S118	79-200	SM	828	773.3	nm	nm
S119	86-295	SM	840	781.2	767.0	-14.2
S120	80-160	SM	821	781.9	nm	nm
S121	89-170	SM	830	773.0	760.9	-12.1
S122	95-515	SMGP	834	539.7	536.0	-3.7
S123	?-485	SMGP	834	778.3	778.0	-.3
S124	360-700	GPAn	862	492.3	493.1	.8
S125	90-380	SMGP	832	784.0	nm	nm
S126	147-350	SMGP	816	781.9	781.2	-.7
S127	206-450	SMGP	849	576.0	nm	nm
S128	205-460	SMGP	842	560.0	nm	nm
S129	248-290	SM	872	804.1	798.8	-5.3
S130	10*-500	SMGP	840	470.0	477.4	7.4
S131	177-490	SMGP	838	613.0	nm	nm
S132	234-238	SM	898	802.7	796.5	-6.2
S133	245-600	SMGP	920	485.0	nm	nm
S134	370-660	GPAn	830	434.0	438.8	4.8
S135	193-260	SM	840	671.0	nm	nm
S136	198-260	SM	865	806.7	803.1	-3.6
S137	?-200	DD	869	799.2	796.2	-3.0
S138	200-210	DD	885	795.8	792.2	-3.6
S139	328-700	GPAn	883	522.5	nm	nm
S140	?-560	SMGP	861	na	639.4	na
S141	200-240	SM	870	805.5	nm	nm
S142	?-178	DD	859	799.8	nm	nm
S143	?-170	DD	856	803.1	nm	nm
S144	139-145	DD	851	800.0	797.0	-3.0
S145	199-500	SMGP	868	702.8	672.8	-30.0
S146	253-498	SMGP	836	<536	nm	nm
S147	105-150	SM	840	824.3	818.2	-6.2
S148	120-230	SM	858	810.1	801.4	-8.8
S149	182-287	SM	930	804.8	786.6	-18.2
S150	166-240	SM	902	810.5	nm	nm
S151	175-240	SM	910	854.2	854.1	-.1
S152	?-320*	SM*	920	789.4	nm	nm
S153	204-295	SM	920	792.9	771.6	-21.3
S154	163-280	SMGP	902	769.7	769.1	-.6
S155	?->240*	SM*	910	795.7	nm	nm
S156	230-285	SM	960	792.5	nm	nm
S157	?-218	*SM	830	771.0	770.2	-.9
S158	202-320	SM*	858	803.2	798.2	-5.0
S159	?	SM*	850	797.3	nm	nm

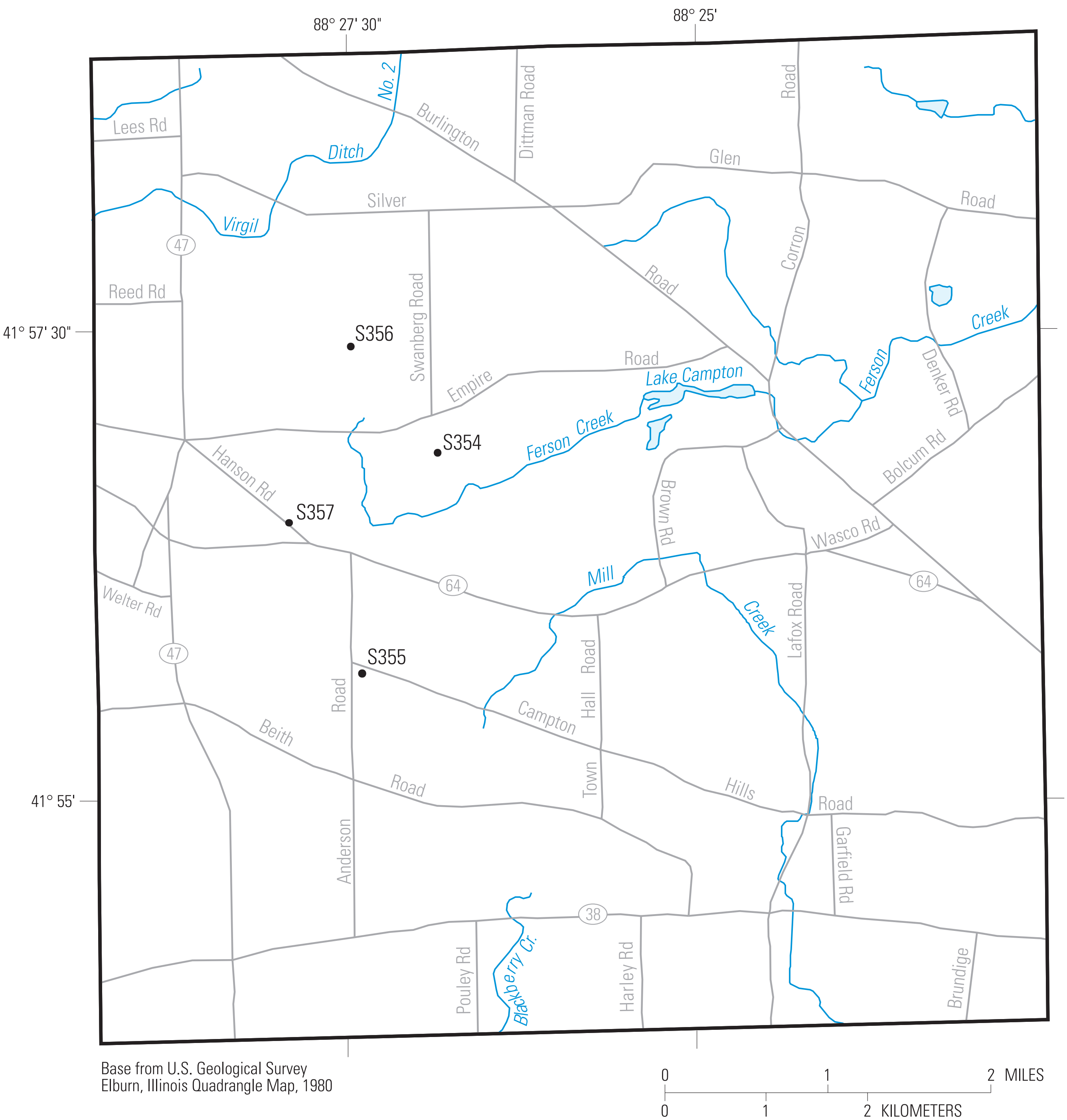
Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S160	202-360	SMGP	865	722.0	nm	nm
S161	270-510	SMGP	880	634.0	nm	nm
S162	?->400	SMGP	888	600.0	nm	nm
S163	?-200	DD	877	792.2	<787	>-5.2
S164	?-270	SM	830	692.1	nm	nm
S165	240-600	SMGP	899	492.5	nm	nm
S166	225-300	SM	901	796.9	791.5	-5.4
S167	233-595	SMGP	893	<513	nm	nm
S168	?-500	SMGP	887	<587	nm	nm
S169	236-500	SMGP	882	<620	nm	nm
S170	242-500	SMGP	905	552.0	555.0	3.0
S171	245-495	SMGP	902	560.0	nm	nm
S172	253-700	SMGPAn	906	462.0	nm	nm
S173	?-540	SMGP	912	650.0	nm	nm
S174	246-620	SMGP	910	659.8	658.0	-1.8
S175	176-186	DD	834	802.1	nm	nm
S176	171-296	SM	845	715.7	nm	nm
S177	?-225	DD	900	825.2	nm	nm
S178	252-260	DD	920	805.6	800.3	-5.3
S179	33-41	SD	900	884.9	885.5	.6
S180	?-212	DD	908	798.2	792.9	-5.3
S181	200-235	DD	911	798.5	nm	nm
S182	218-228	DD	901	800.0	nm	nm
S183	211-300	SM	893	841.8	841.1	-.7
S184	?-240	DD	906	825.4	827.8	2.4
S185	?-380	SMGP	892	786.5	779.7	-6.8
S186	?-50	SD	911	891.7	nm	nm
S187	?-200	DD	918	788.3	785.8	-2.5
S188	?-230	DD	920	817.0	nm	nm
S189	?-200	SM	794	782.3	780.7	-1.6
S190	219-380	SMGP	898	779.0	779.1	.1
S191	225-250	SM	909	789.5	783.0	-6.5
S192	221-500	SMGP	902	<667	nm	nm
S193	165-365	SMGP	908	834.4	nm	nm
S194	57-160	SM	781	760.0	759.4	-.6
S195	?-400	SMGP	822	<522	nm	nm
S196	?-440	SMGP	844	680.8	nm	nm
S197	226-325	SM	892	790.3	nm	nm
S198	222-600	SMGP	917	<617	560.2	nm
S199	?-205	DD	877	798.2	nm	nm
S200	220-600	SMGP	872	<572	nm	nm
S201	?-230	DD	905	832.1	829.8	-2.3
S202	?-210	DD	880	799.5	793.6	-5.9
S203	?	DD*	904	793.5	775.6	-18.0
S204	?-226	DD	920	789.2	nm	nm
S205	231-300	SM	904	803.2	nm	nm
S206	?-300	SM	886	787.6	781.5	-6.1
S207	255-360	SM	894	762.8	nm	nm
S208	?-234	DD	901	799.2	nm	nm
S209	?-215	DD	886	804.1	799.2	-4.9
S210	210-238	DD	883	797.3	789.4	-7.9
S211	?-235	DD	902	797.6	797.1	-.5
S212	?-210	DD	920	853.0	nm	nm
S213	?-210	DD	925	855.7	nm	nm
S214	?-242	DD	946	812.3	810.2	-2.1

Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S215	205-360	SMGP	902	630.2	nm	nm
S216	?-230	DD	932	828.6	825.4	-3.2
S217	?-213	DD	942	843.8	844.7	.9
S218	?-215	DD	933	843.7	840.8	-2.9
S219	242-365	SM	938	823.9	824.0	.1
S220	222-600	SMGP	910	744.3	nm	nm
S221	?-252	DD	925	806.0	793.6	-12.4
S222	303-440	SMGP	921	801.9	nm	nm
S223	162-400	SMGP	926	776.8	nm	nm
S224	42-46	SD	929	925.3	923.6	-1.7
S225	?-200	DD	891	796.8	792.5	-4.3
S226	170-185	DD	873	796.8	792.6	-4.2
S227	186-200	DD	878	799.5	806.0	6.5
S228	?	SM*	866	795.0	794.0	-1.0
S229	186-200	DD	888	792.0	816.0	24.0
S230	175-185	DD	893	799.1	nm	nm
S231	160-260	SM	908	790.9	nm	nm
S232	196-305	SM	855	690.5	673.8	-16.7
S233	179-296	SM	854	694.0	674.3	-19.7
S234	236-240	DD*	883	795.9	797.6	1.7
S235	235-520	SMGP	946	700.2	nm	nm
S236	?-235	DD	926	843.6	nm	nm
S237	?-248	DD	964	846.0	nm	nm
S238	236-520	SMGP	925	467.0	nm	nm
S239	250-300	SM	953	864.9	864.2	-.7
S240	265-544	SMGP	956	832.5	<756	nm
S241	260-560	SMGP	957	617.2	nm	nm
S242	?-276	DD	967	854.7	855.5	.8
S243	?-800	SMGPAn	966	<666	597.0	nm
S244	?-270	DD	960	850.9	nm	nm
S245	280-420	SMGP	966	850.0	nm	nm
S246	232-720	SMGPAn	939	463.0	nm	nm
S247	252-560	SMGP	944	<644	<616	nm
S248	153-580	SMGP	934	509.0	524.4	15.4
S249	152-505	SMGP	931	637.0	752.9*	115.9*
S250	168-590	SMGP	922	621.5	nm	nm
S251	231-600	SMGP	930	679.0	nm	nm
S252	248-600	SMGP	906	545.0	nm	nm
S253	243-540	SMGP	888	541.0	nm	nm
S254	?-245	DD	953	864.5	863.0	-1.5
S255	240-540	SMGP	906	571.0	nm	nm
S256	244-600	SMGP	909	<609	nm	nm
S257	240-710	SMGPAn	904	645.3	620.6	-24.7
S258	225-530	SMGP	897	595.0	nm	nm
S259	243-600	SMGP	914	612.0	nm	nm
S260	?	SM or DD	899	839.8	840.0	.2
S261	?-505	SMGP	887	<587	nm	nm
S262	375-800	SMGPAn	950	520.5	nm	nm
S263	226-540	SMGP	918	592.0	nm	nm
S264	?-240	SM*	912	851.6	850.1	-1.5
S265	?-226	DD	931	811.0	804.9	-6.1
S266	?-250	DD	960	837.9	nm	nm
S267	?-190	SM	895	802.5	799.1	-3.4
S268	?-186	DD	851	833.6	828.0	-5.6
S269	?-186	DD	846	821.3	816.8	-4.5

Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S270	?	SMGPAn*	910	475.0	nm	nm
S271	220-240	SM	922	818.8	nm	nm
S272	?-230	DD	898	793.5	nm	nm
S273	148-220	SM	895	792.9	780.1	-12.8
S274	119-190	SM	882	811.5	nm	nm
S275	207-280	SM	925	782.6	773.1	-9.5
S276	177-260	SM	900	789.3	nm	nm
S277	261-280	SM	869	792.5	779.8	-12.7
S278	194-280	SM	866	798.3	789.9	-8.4
S279	111-300	SMGP	850	791.7	783.5	-8.2
S280	?-212	DD	901	771.4	769.6	-1.8
S281	240-275	SM	910	<812	nm	nm
S282	233-580	SMGP	901	<460	478*	nm
S283	180-192	DD	870	nm	801.2	nm
S284	194-204	DD	887.5	nm	785.5	nm
S285	?-360	SM	883	nm	788.3	nm
S286	220-223	DD	915	nm	779.5	nm
S287	178-250	SM	836	nm	767.1	nm
S288	190-265	SM	840	nm	755.4	nm
S289	?-175	DD	836	nm	770.2	nm
S290	?-271	DD	840	nm	746.2	nm
S291	200-207	DD	915	nm	777.4	nm
S292	220-230	DD	890	nm	784.7	nm
S293	?-258	DD	962	nm	846.7	nm
S294	?	?	934	nm	<734	nm
S295	118-205	SM	839.8	nm	761.5	nm
S296	?-227	DD	897.1	nm	777.7	nm
S297	218-220	DD	906.3	nm	790.7	nm
S298	?-218	DD	907.1	nm	792.5	nm
S299	215-280	SM	942.2	nm	779.9	nm
S300	?	DD?SM?	928	nm	821.2	nm
S301	177-?	SM	826	nm	779.0	nm
S302	400-740	GPSP	954	nm	585.7	nm
S303	145-240*	SM*	930	nm	<790	nm
S304	?-40	SD	943	nm	922.1	nm
S305	363-720	SP	930	nm	<522	nm
S306	?	SM*	827	nm	665.5	nm
S307	132-235	SM	788	nm	671.1	nm
S308	137-220	SM	805	nm	687.0	nm
S309	118-183	SM	828	nm	772.7	nm
S310	?	SM	853	nm	773.9	nm
S311	177-295	SM	852	nm	673.0	nm
S312	?-260	SM	800	nm	663.0	nm
S313	?-169	DD	853	nm	800.1	nm
S314	?-228	DD	886	nm	799.8	nm
S315	?-210	DD	891	nm	789.0	nm
S316	214-224	DD	827	nm	746.8	nm
S317	?-260	SM	816	nm	748.2	nm
S318	81-205	SM	834	nm	778.6	nm
S319	?-260	SM	790	nm	768.3	nm
S320	?-300	SM	785	nm	653.1	nm
S321	?-101	DD	790	nm	775.5	nm
S322	143-180	SM	795	nm	715.1	nm
S323	220-227	DD	911	nm	792.1	nm
S324	?-214	DD	905	nm	793.2	nm

Well name	Open interval (feet below land surface)	Aquifer open to	Measuring point altitude (feet above NGVD29)	Altitude of water level in 1995 (feet above NGVD29)	Altitude of water level in 2002 (feet above NGVD29)	Change in water level 1995 to 2002 (feet)
S325	191-651	GP/SP	882	nm	<662	nm
S326	191-651	SMGP	882	nm	659.1	nm
S327	195-340	SM/GP	877	nm	668.2	nm
S328	244-310?	SMGPAn*	913	nm	<695	nm
	DEEPENED					
S329	?-231	DD	913	nm	792.6	nm
S330	234-310	SM	905	nm	790.2	nm
S331	?-280	SM	940	nm	787.6	nm
S332	?-194	DD	898	nm	852.5	nm
S333	280-355	SM	976	nm	776.6	nm
S334	?-279	DD	966	nm	853.7	nm
S335	167-500*	SMGP*	829	nm	<501	nm
S336	98-110	DD	783	nm	757.2	nm
S337	?-255	SM	903	nm	873*	nm
S338	?-300	SM	865	nm	675.78*	nm
S339	?-239	DD	888	nm	799.0	nm
S340	?-250	SM	887	nm	797.7	nm
S341	?-206	DD	853	nm	797.0	nm
S342	?-270	SM	800	nm	755.5	nm
S343	?-220	SM	837	nm	772.5	nm
S344	?-360	SM	843	nm	712.9	nm
S345	?-245	DD	950	nm	848.4	nm
S346	?-340	SMGP	835	nm	729.8*	nm
S347	?-205	SM	818	nm	760.8	nm
S348	?-210	SM	836	nm	787.0	nm
S349	?-255	SM	810	nm	756.7	nm
S350	?-140	SM	865	nm	775.4	nm
S351	216-600	SMGP	885	nm	<685	nm
S352	613-875	An	800	nm	253.0	nm
S353	614-870	An	810	nm	400.0	nm

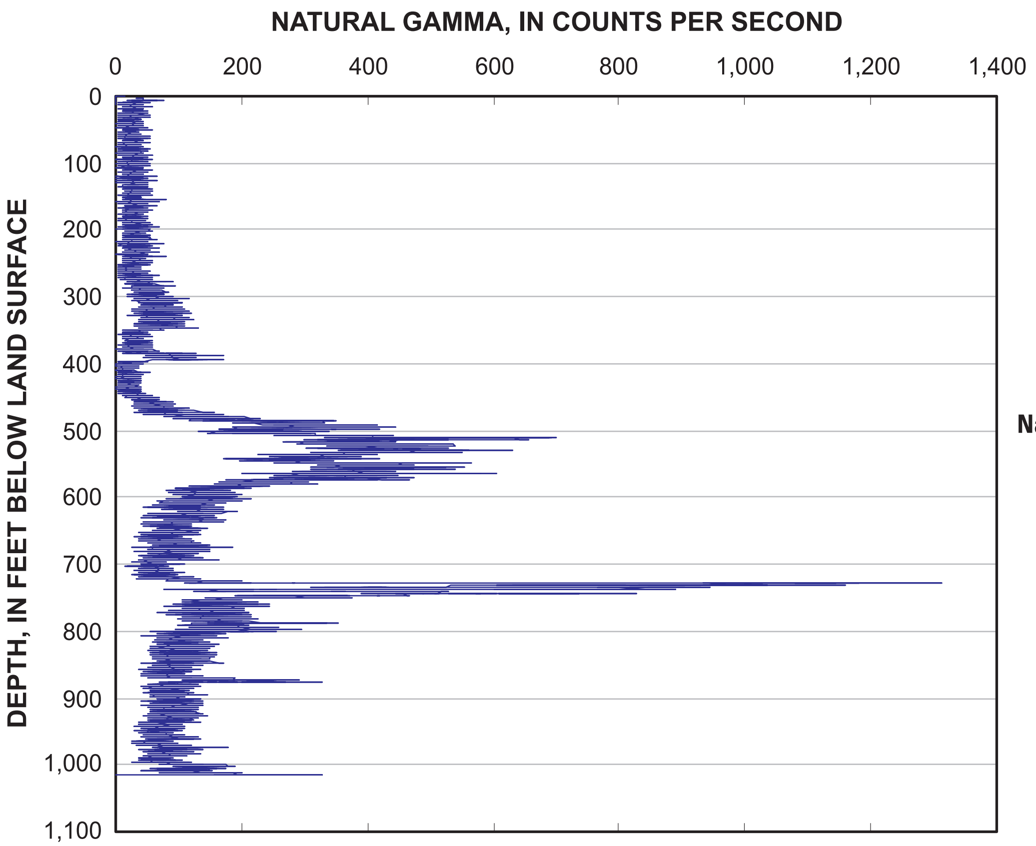
APPENDIX B—Geophysical logs collected from wells in Campton Township, Illinois, June 2003



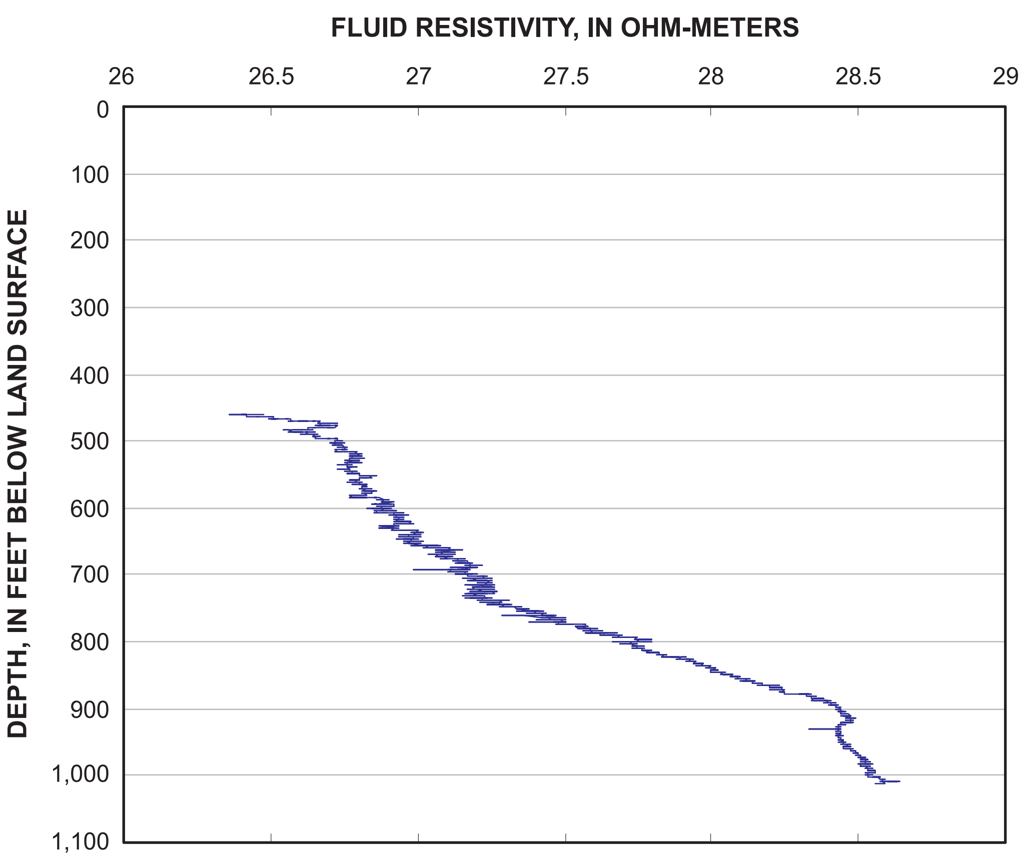
EXPLANATION

• S355 WELL LOCATION WHERE GEOPHYSICAL LOGS WERE COLLECTED, JUNE 2003

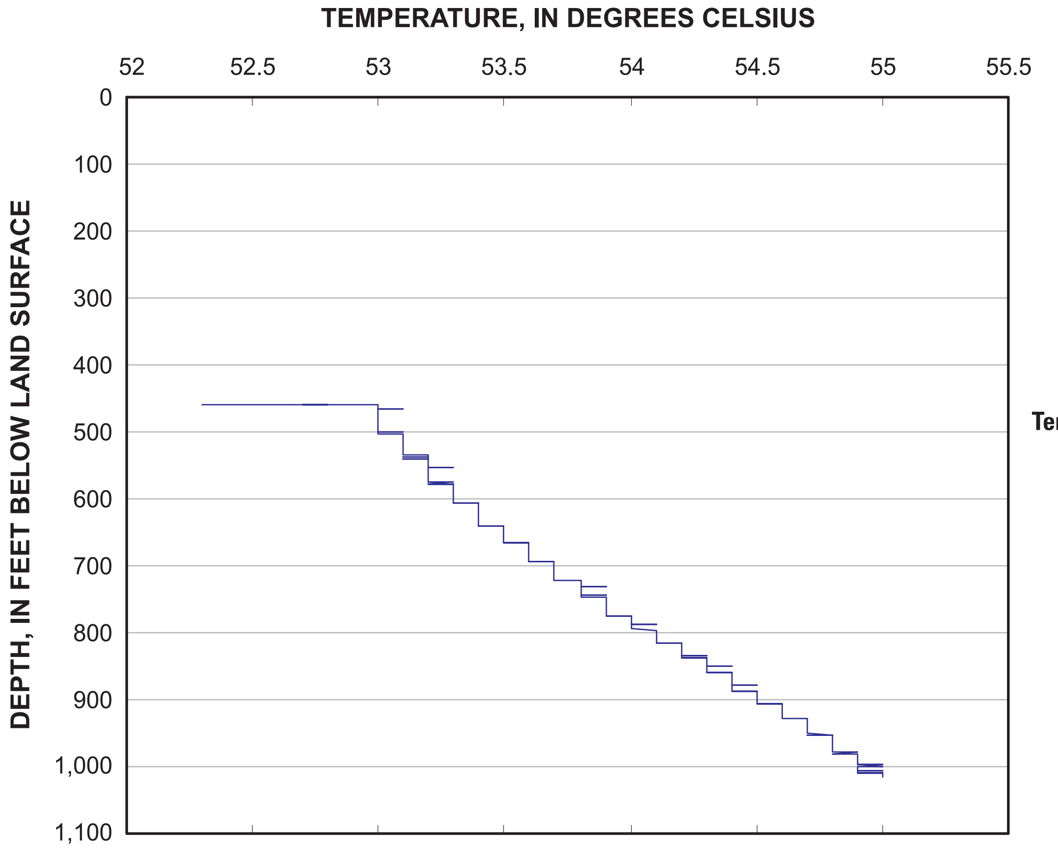
Figure B1. Wells where geophysical logs were collected in Campton Township, Illinois, June 2003.



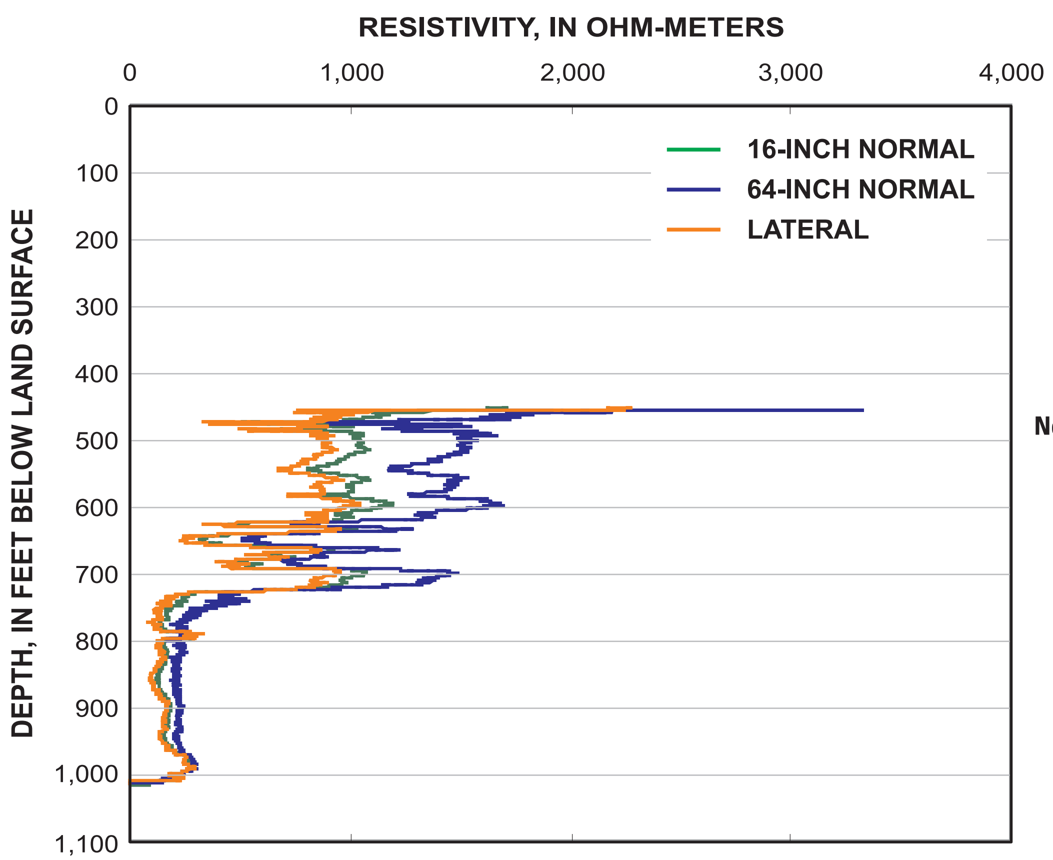
Natural-Gamma Log, Well S354



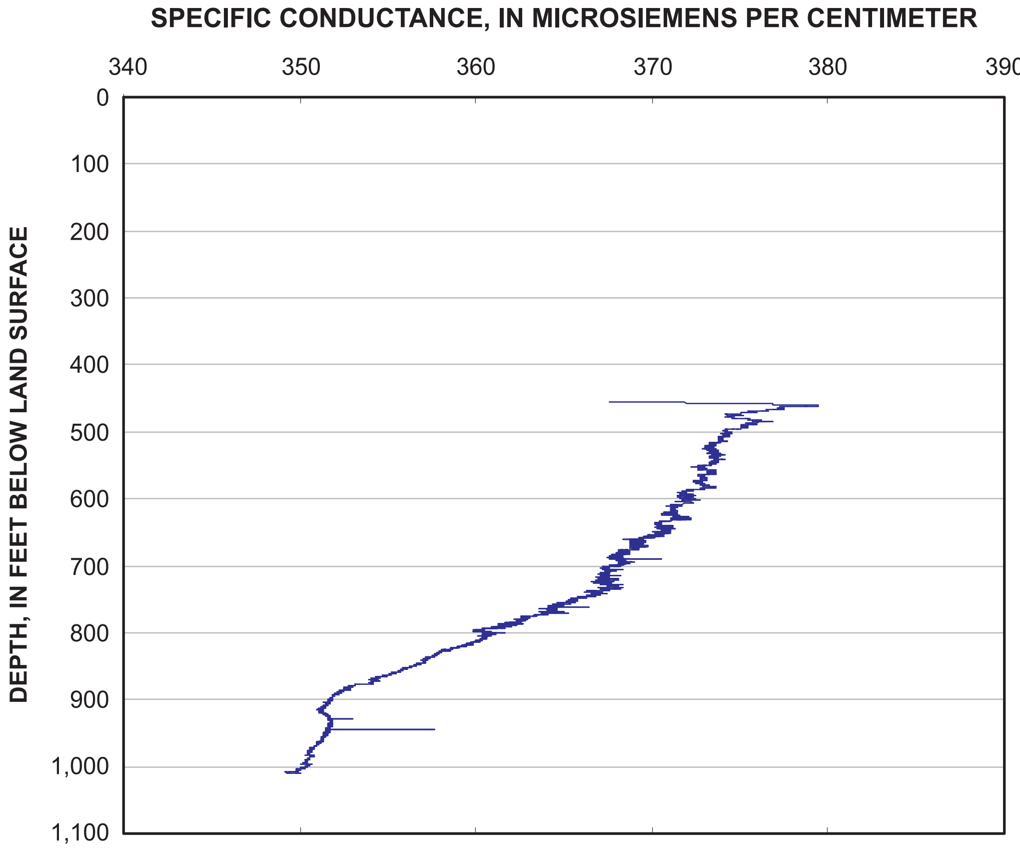
Fluid-Resistivity Log, Well S354



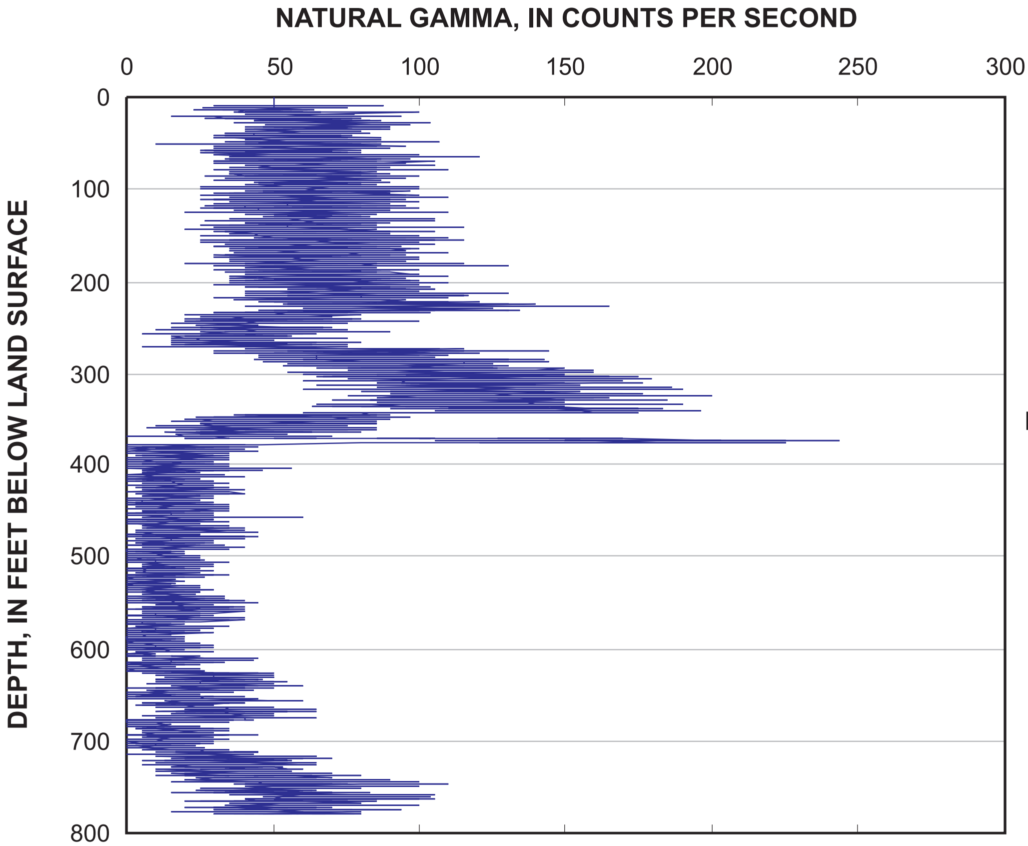
Temperature Log, Well S354



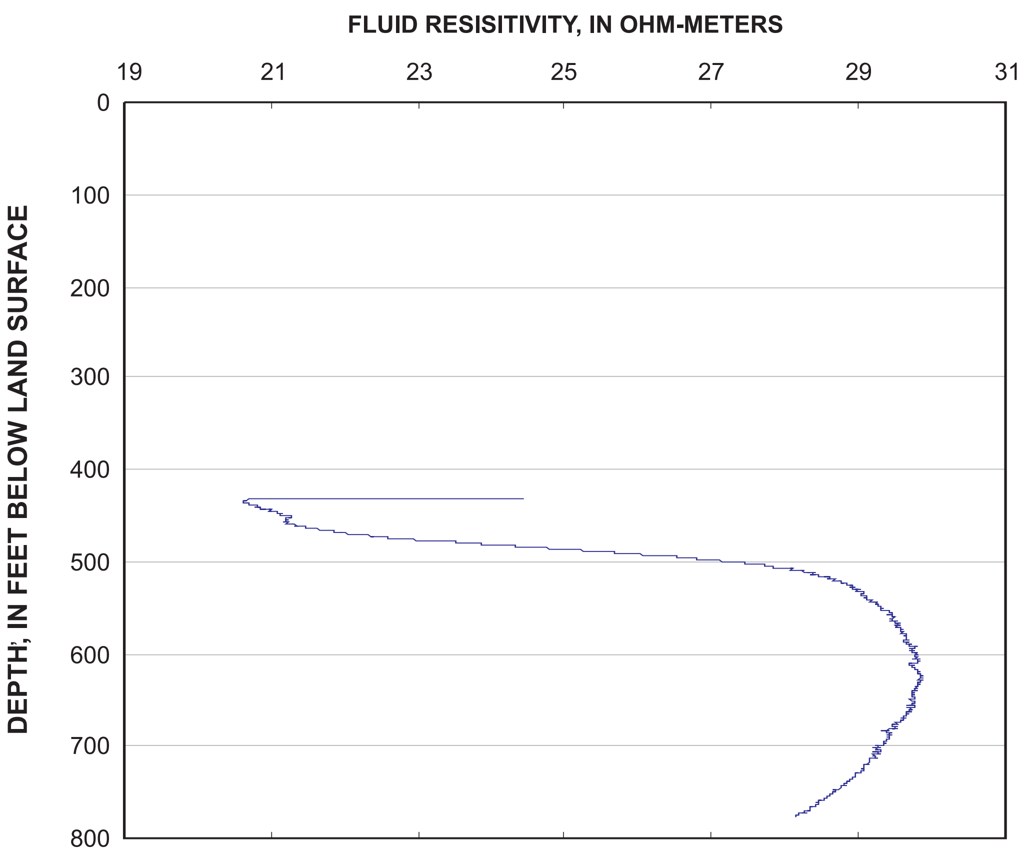
Normal-Resistivity Log, Well S354



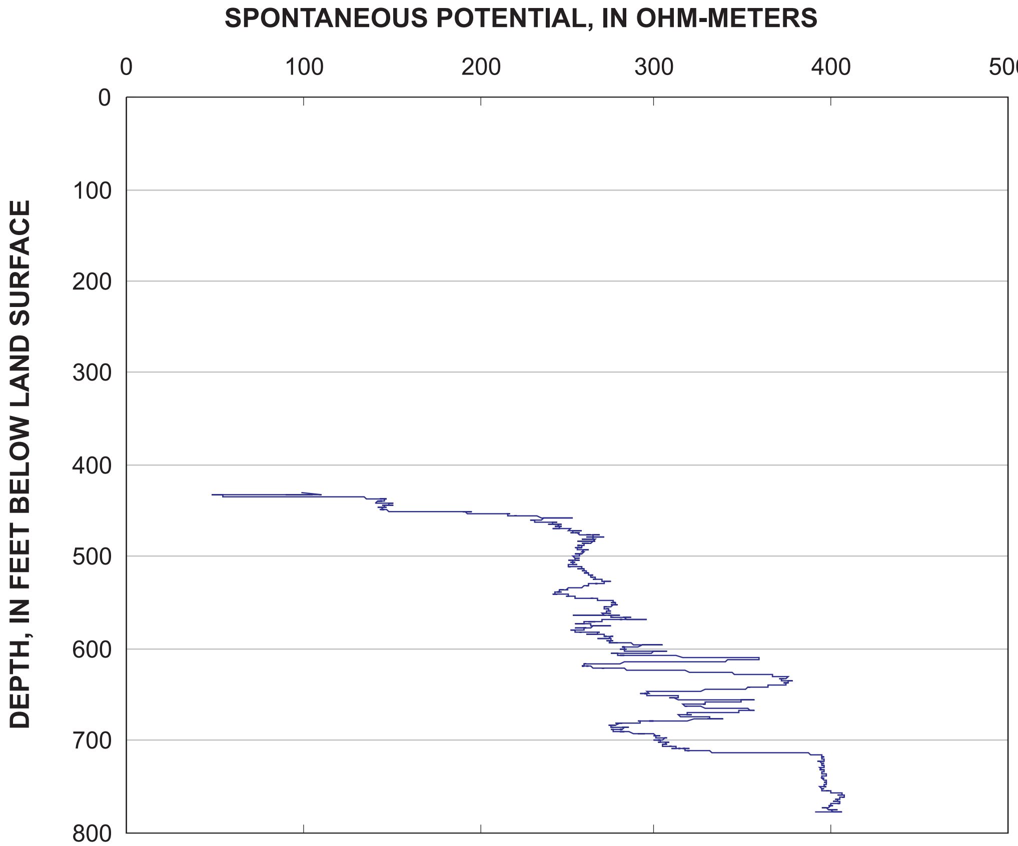
Specific-Conductance Log, Well S354



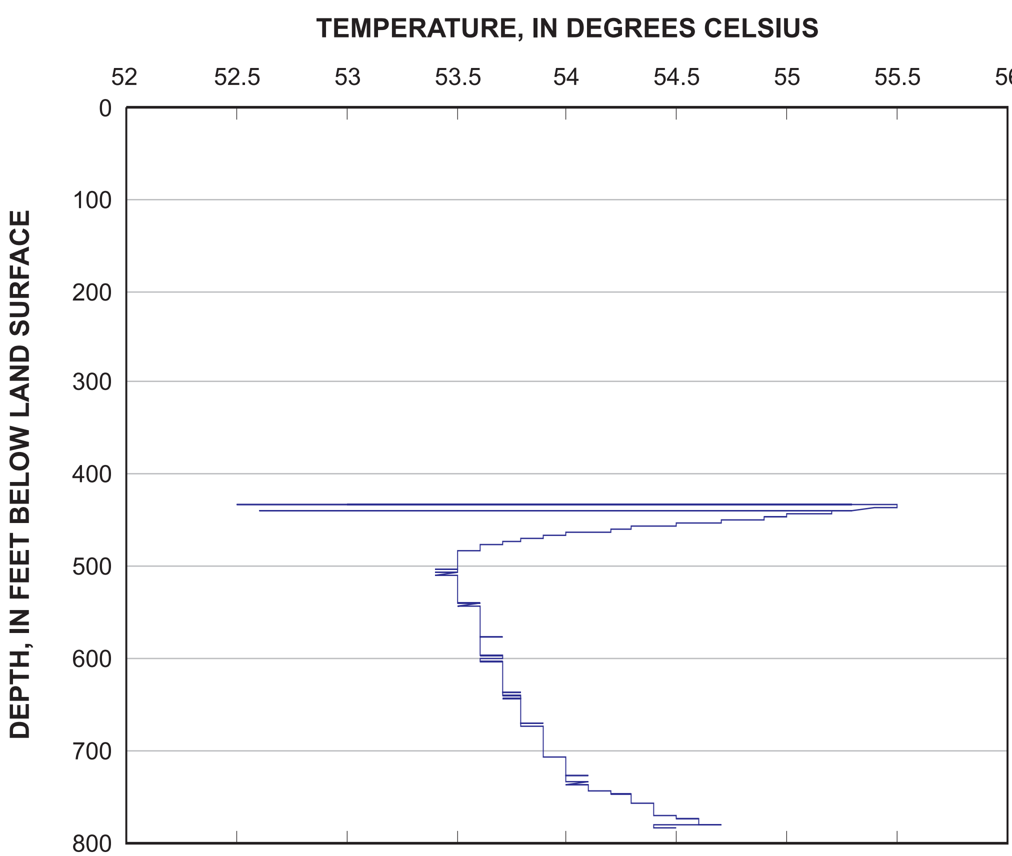
Natural-Gamma Log, Well S355



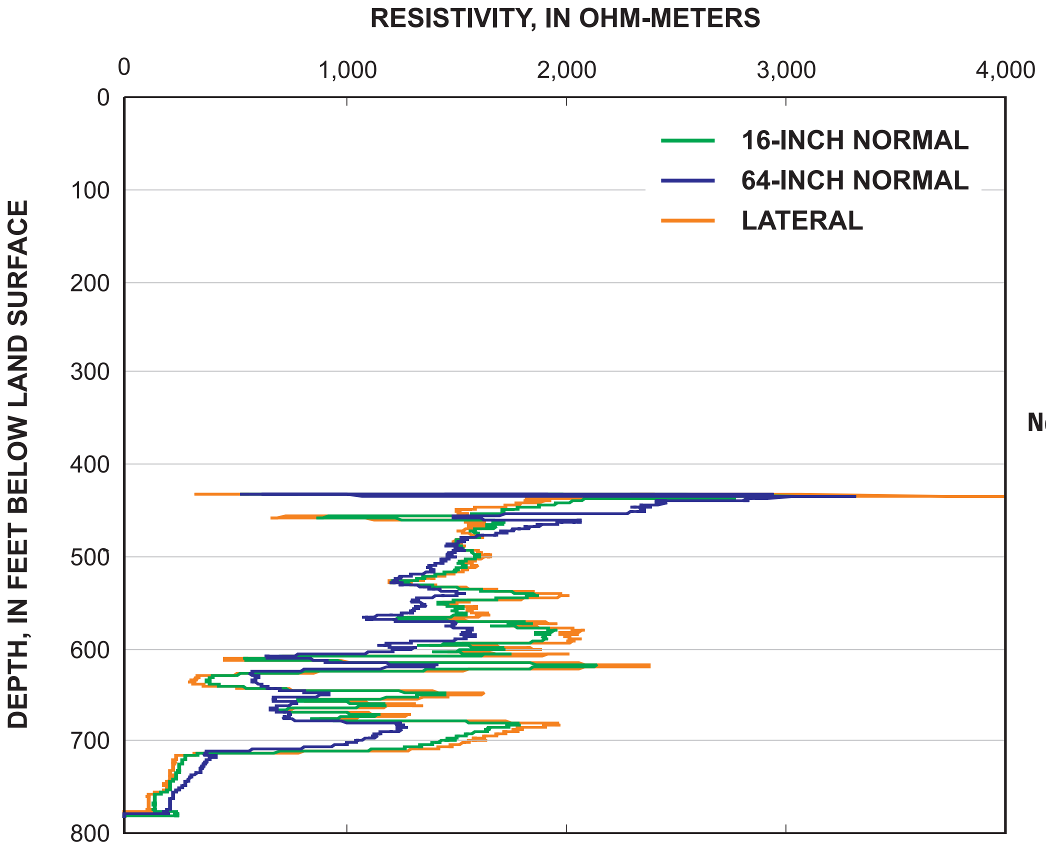
Fluid-Resistivity Log, Well S355



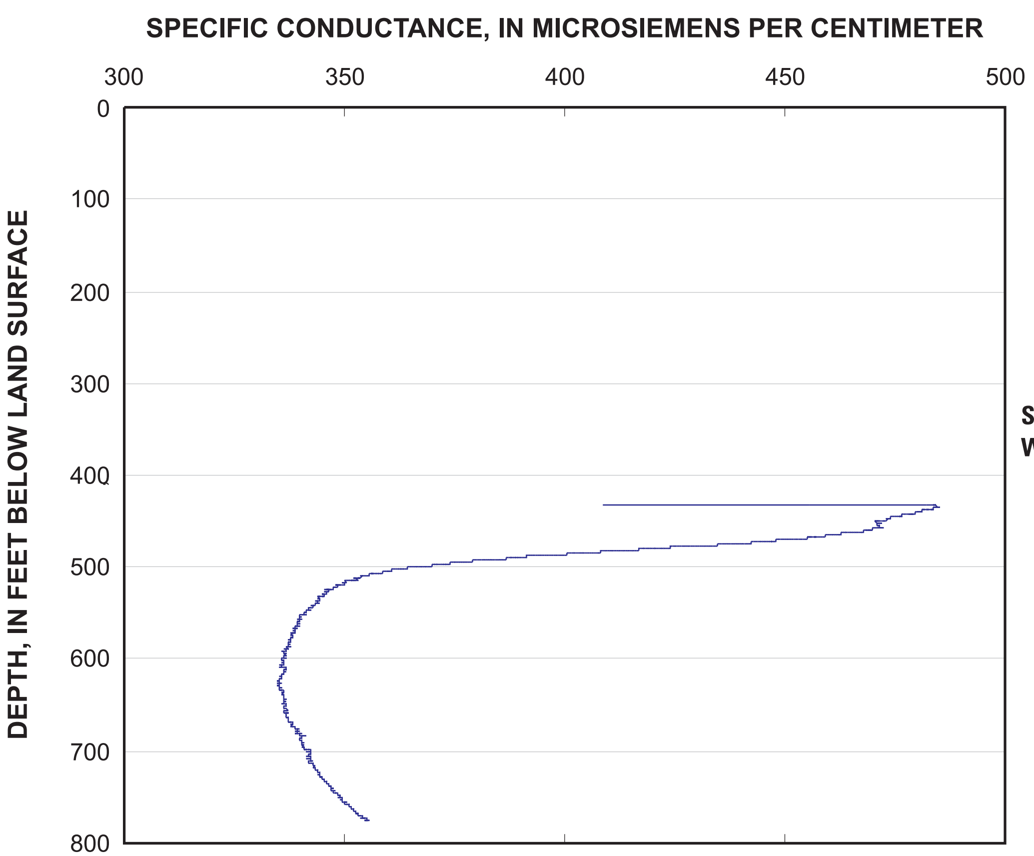
Spontaneous-Potential Log, Well S355



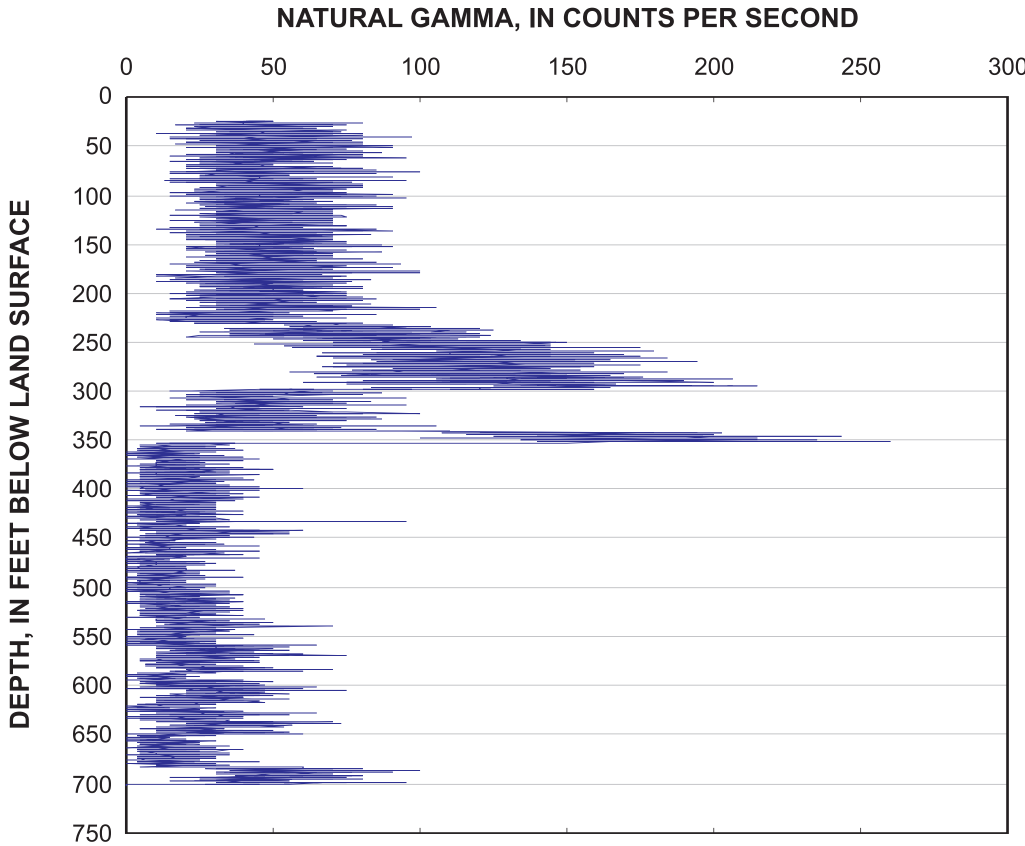
Temperature Log, Well S355



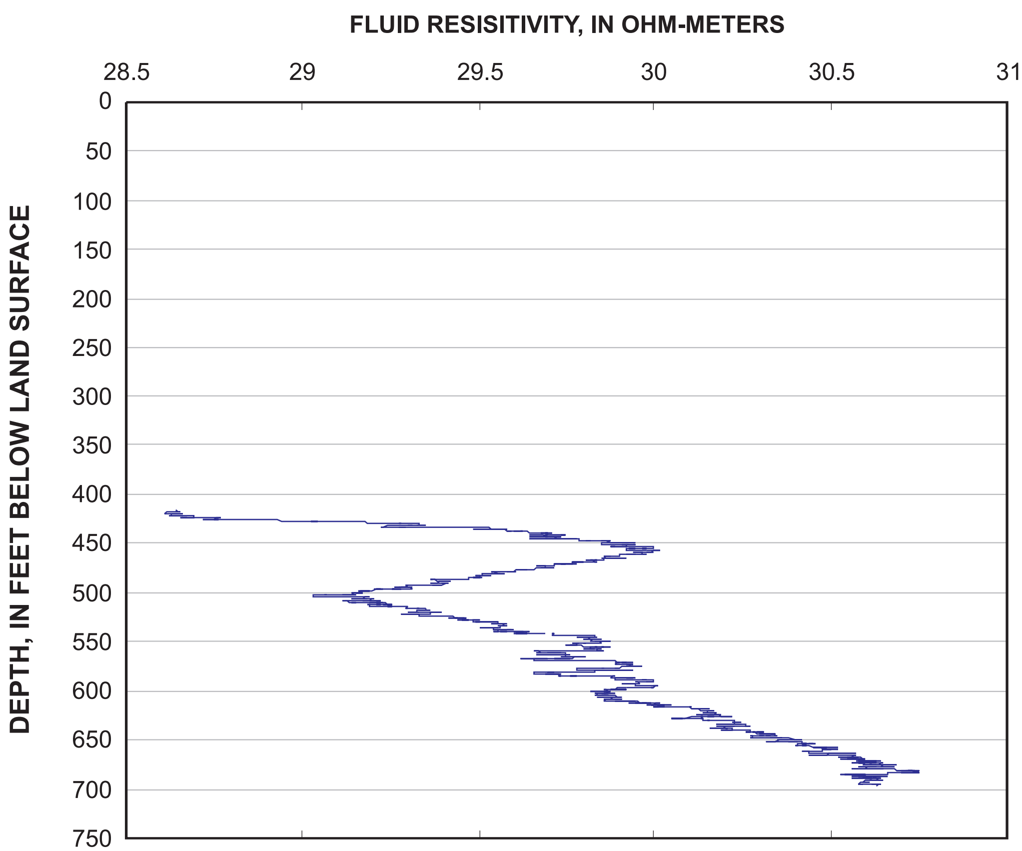
Normal-Resistivity Log, Well S355



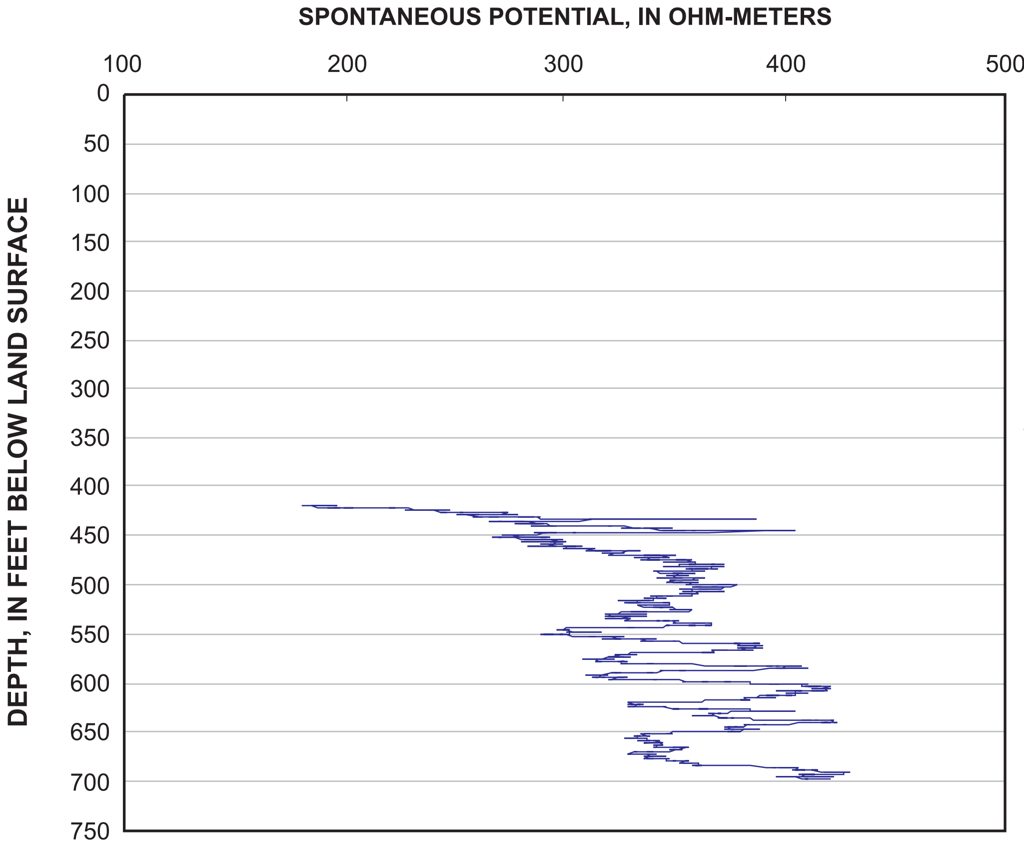
Specific-Conductance Log, Well S355



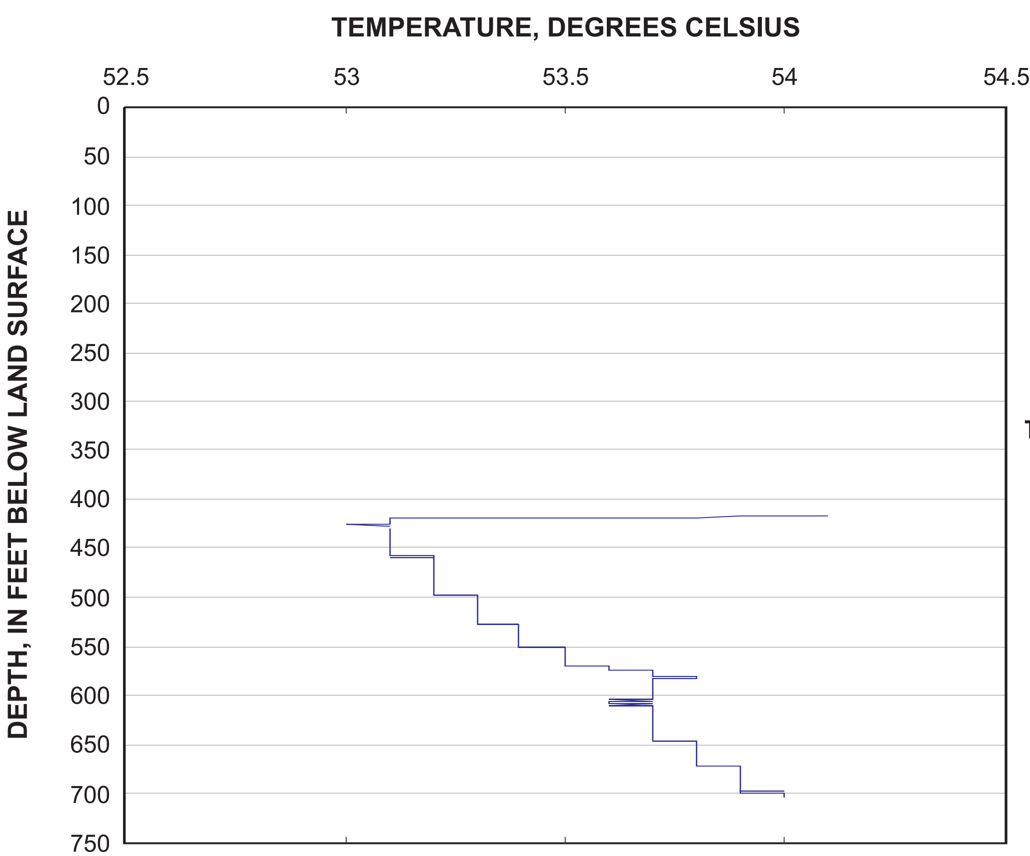
Natural-Gamma Log, Well S356



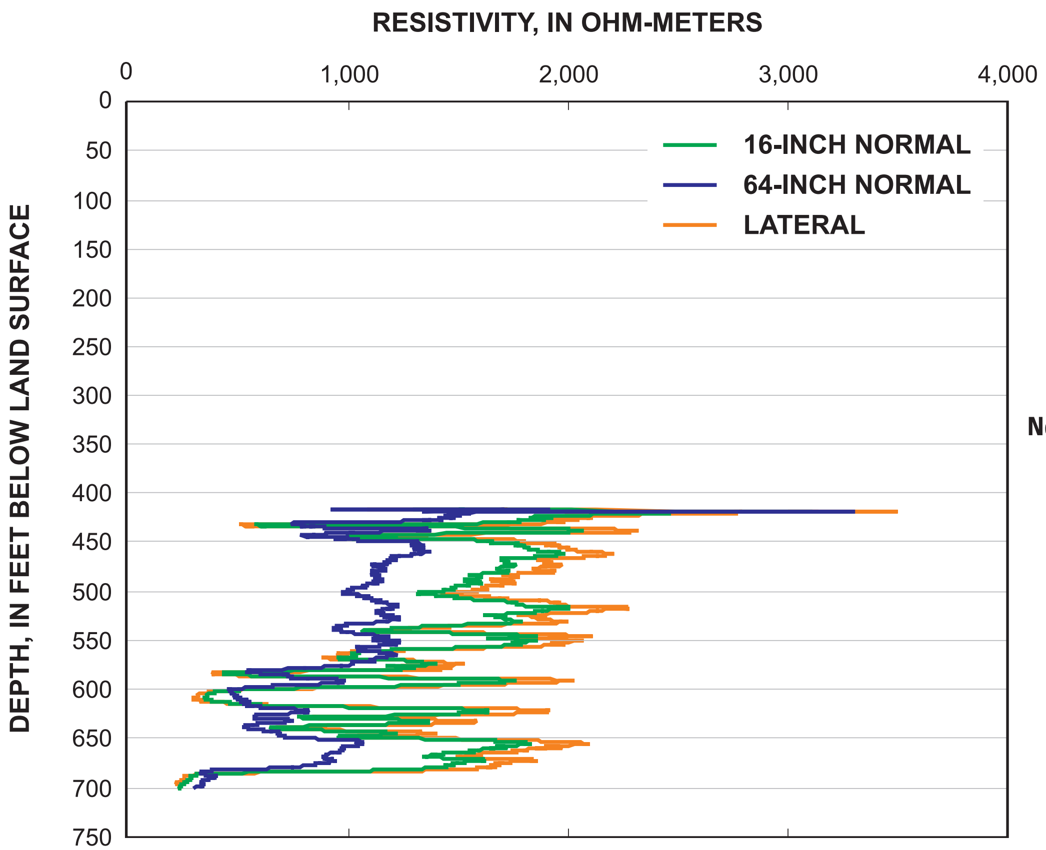
Fluid-Resistivity Log, Well S356



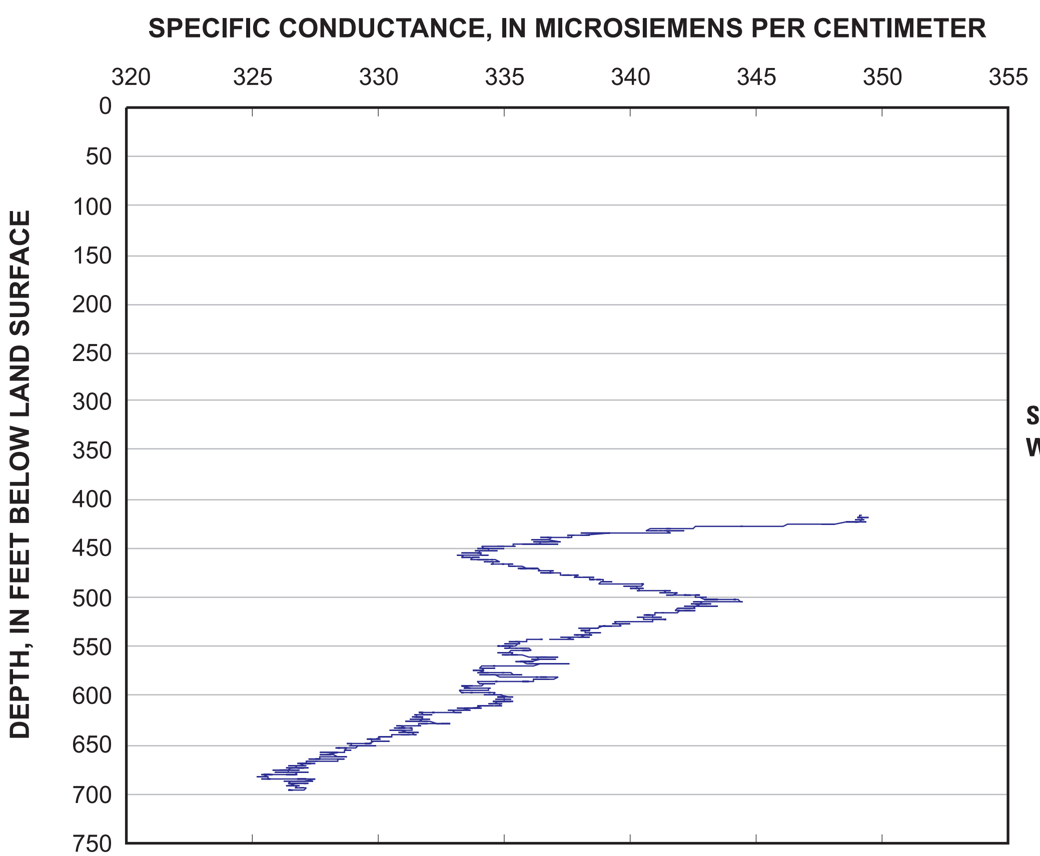
Spontaneous-Potential Log, Well S356



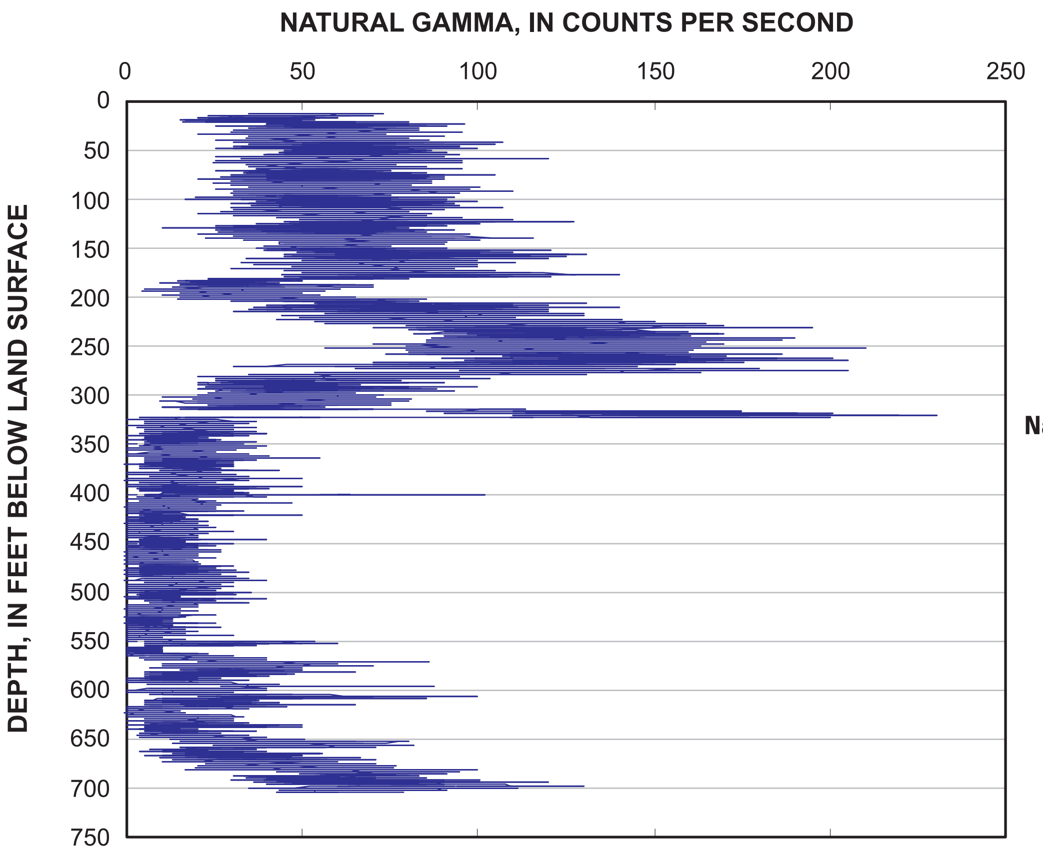
Temperature Log, Well S356



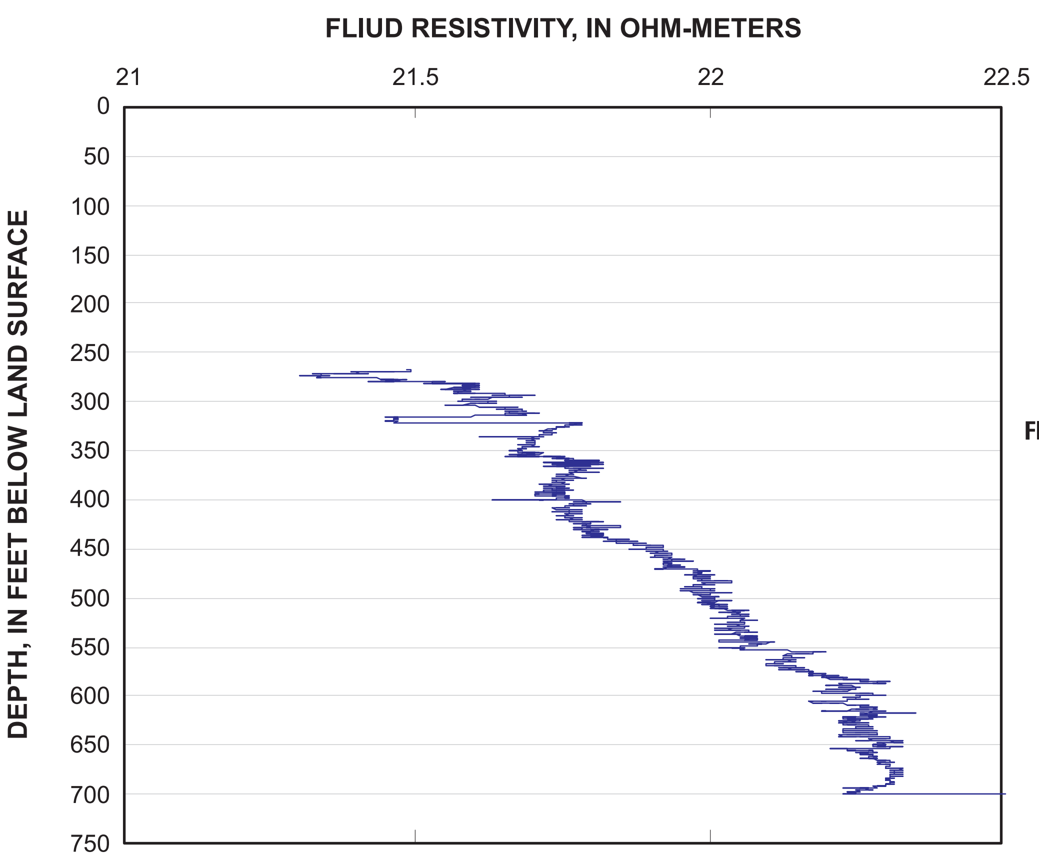
Normal-Resistivity Log, Well S356



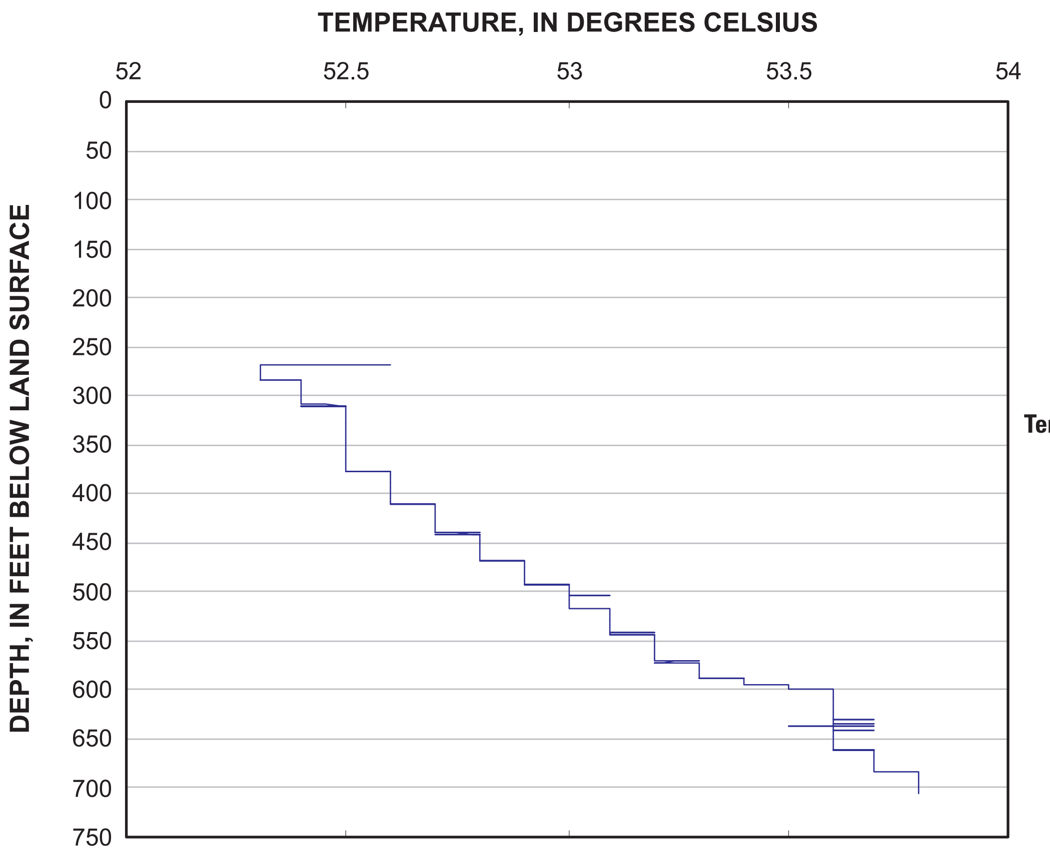
Specific-Conductance Log, Well S356



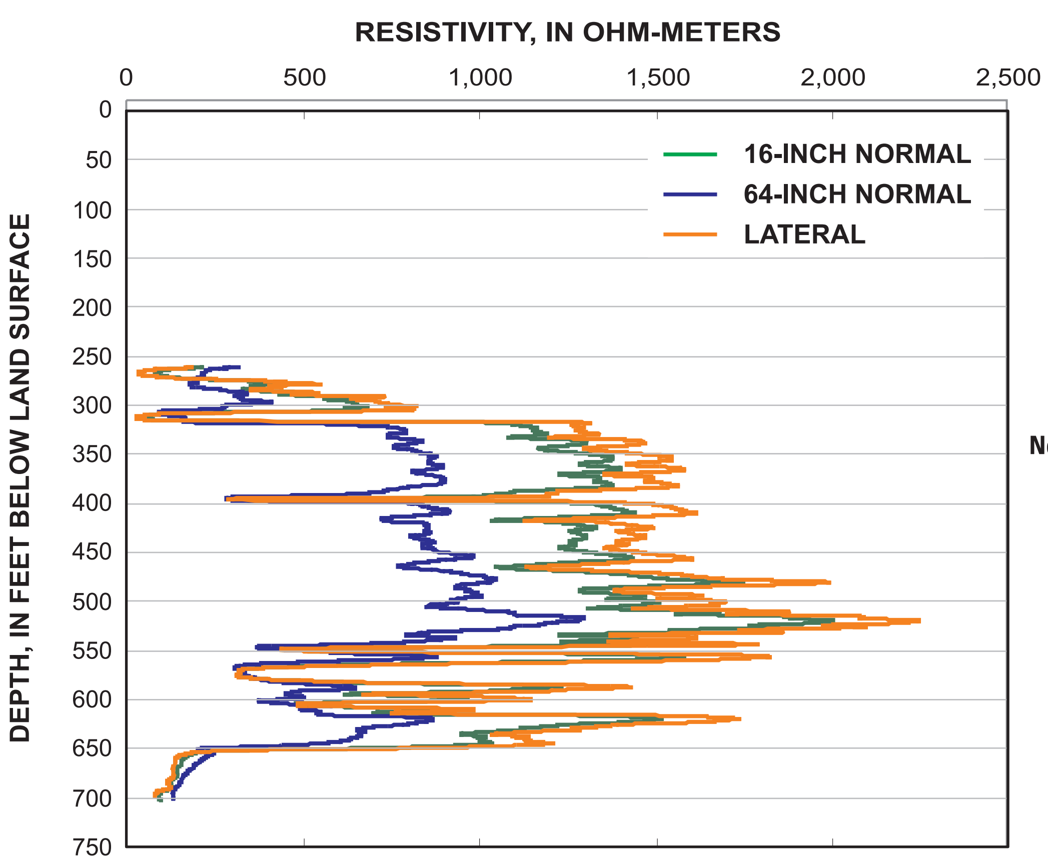
Natural-Gamma Log, Well S357



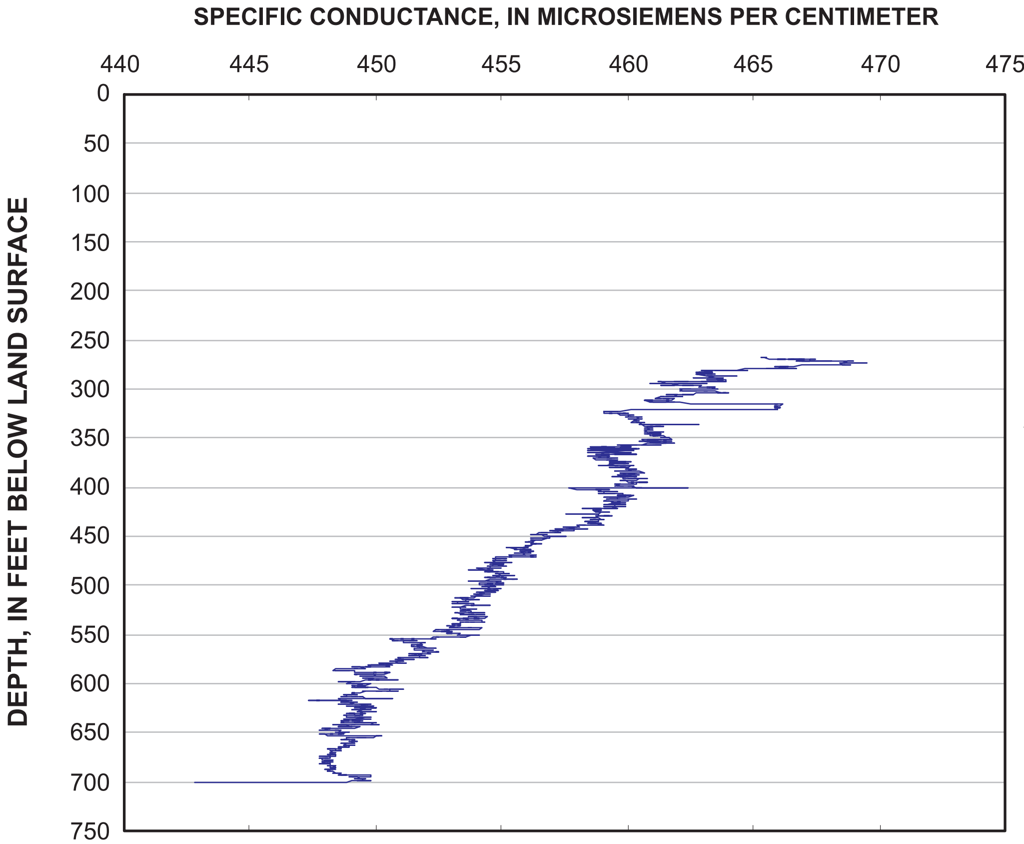
Fluid-Resistivity Log, Well S357



Temperature Log, Well S357



Normal-Resistivity Log, Well S357



Specific-Conductance Log, Well S357

Kay, R.T., and others—**Hydrogeology, Water Use, and Simulated Ground-Water Flow and Availability in Campton Township, Kane County, Illinois**—
Scientific Investigations Report 2006–5076