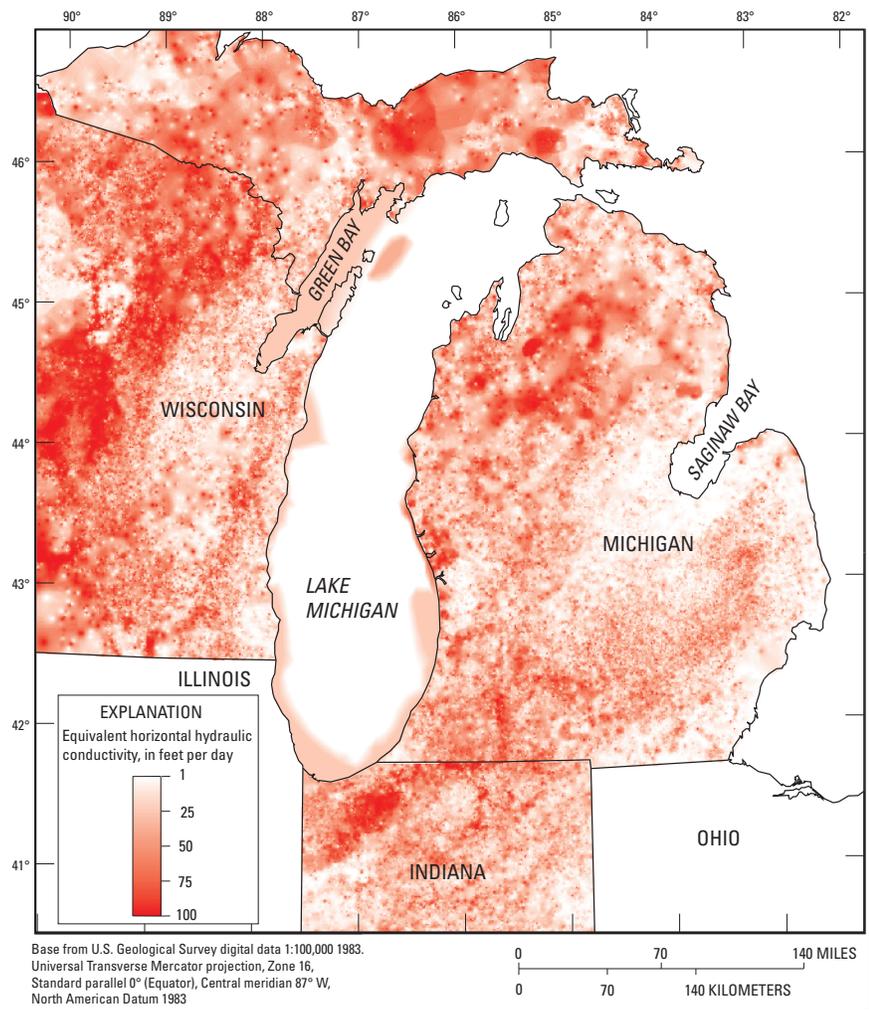


National Water Availability and Use Pilot Program

Processing, Analysis, and General Evaluation of Well-Driller Logs for Estimating Hydrogeologic Parameters of the Glacial Sediments in a Ground-Water Flow Model of the Lake Michigan Basin



Scientific Investigations Report 2008–5184

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By Leslie D. Arihood

National Water Availability and Use Pilot Program

Scientific Investigations Report 2008–5184

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Conversion Factors

Inch/Pound to SI		
Multiply	By	To obtain
Length		
inch (in.)	2.54	centimeter (cm)
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
Area		
square mile (mi ²)	259.0	hectare (ha)
square mile (mi ²)	2.590	square kilometer (km ²)
Flow rate		
gallon per minute (gal/min)	0.06309	liter per second (L/s)
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day (m/d)
Transmissivity*		
foot squared per day (ft ² /d)	0.09290	meter squared per day (m ² /d)
SI to Inch/Pound		
Multiply	By	To obtain
Length		
centimeter (cm)	0.3937	inch (in.)
millimeter (mm)	0.03937	inch (in.)
meter (m)	3.281	foot (ft)
kilometer (km)	0.6214	mile (mi)
Area		
hectare (ha)	0.003861	square mile (mi ²)
square kilometer (km ²)	0.3861	square mile (mi ²)
Flow rate		
liter per second (L/s)	15.85	gallon per minute (gal/min)
Hydraulic conductivity		
meter per day (m/d)	3.281	foot per day (ft/d)
Transmissivity*		
meter squared per day (m ² /d)	10.76	foot squared per day (ft ² /d)

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83)

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft³/d)/ft²]. In this report, the mathematically reduced form, foot squared per day (ft²/d), is used for convenience.

Abbreviations

AML	ARC Macro Language
ASCII	American Standard Code for Information Exchange
DEM	Digital elevation model
GWSI	Ground Water Site Inventory
K	Equivalent horizontal hydraulic conductivity
KZ	Equivalent vertical hydraulic conductivity
KSC	Horizontal hydraulic conductivity derived from water-well log specific-capacity data
T	Transmissivity, in foot squared per day

Processing, Analysis, and General Evaluation of Well-Driller Logs for Estimating Hydrogeologic Parameters of the Glacial Sediments in a Ground-Water Flow Model of the Lake Michigan Basin

By Leslie D. Arihood

Abstract

In 2005, the U.S. Geological Survey began a pilot study for the National Water Availability and Use Program to assess the availability of water and water use in the Great Lakes Basin. Part of the study involves constructing a ground-water flow model for the Lake Michigan part of the Basin. Most ground-water flow occurs in the glacial sediments above the bedrock formations; therefore, adequate representation by the model of the horizontal and vertical hydraulic conductivity of the glacial sediments is important to the accuracy of model simulations. This work processed and analyzed well records to provide the hydrogeologic parameters of horizontal and vertical hydraulic conductivity and ground-water levels for the model layers used to simulated ground-water flow in the glacial sediments. The methods used to convert (1) lithology descriptions into assumed values of horizontal and vertical hydraulic conductivity for entire model layers, (2) aquifer-test data into point values of horizontal hydraulic conductivity, and (3) static water levels into water-level calibration data are presented.

A large data set of about 458,000 well driller well logs for monitoring, observation, and water wells was available from three statewide electronic data bases to characterize hydrogeologic parameters. More than 1.8 million records of lithology from the well logs were used to create a lithologic-based representation of horizontal and vertical hydraulic conductivity of the glacial sediments. Specific-capacity data from about 292,000 well logs were converted into horizontal hydraulic conductivity values to determine specific values of horizontal hydraulic conductivity and its aerial variation. About 396,000 well logs contained data on ground-water levels that were assembled into a water-level calibration data set.

A lithology-based distribution of hydraulic conductivity was created by use of a computer program to convert well-log lithology descriptions into aquifer or nonaquifer categories and to calculate equivalent horizontal and vertical hydraulic conductivities (K and K_z , respectively) for each of the glacial layers of the model. The K was based on an assumed value

of 100 ft/d (feet per day) for aquifer materials and 1 ft/d for nonaquifer materials, whereas the equivalent K_z was based on an assumed value of 10 ft/d for aquifer materials and 0.001 ft/d for nonaquifer materials. These values were assumed for convenience to determine a relative contrast between aquifer and nonaquifer materials. The point values of K and K_z from wells that penetrate at least 50 percent of a model layer were interpolated into a grid of values. The K distribution was based on an inverse distance weighting equation that used an exponent of 2. The K_z distribution used inverse distance weighting with an exponent of 4 to represent the abrupt change in K_z that commonly occurs between aquifer and nonaquifer materials.

The values of equivalent hydraulic conductivity for aquifer sediments needed to be adjusted to actual values in the study area for the ground-water flow modeling. The specific-capacity data (discharge, drawdown, and time data) from the well logs were input to a modified version of the Theis equation to calculate specific capacity based horizontal hydraulic conductivity values (K_{sc}). The K_{sc} values were used as a guide for adjusting the assumed value of 100 ft/d for aquifer deposits to actual values used in the model.

Water levels from well logs were processed to improve reliability of water levels for comparison to simulated water levels in a model layer during model calibration. Water levels were interpolated by kriging to determine a composite water-level surface. The difference between the kriged surface and individual water levels was used to identify outlier water levels.

Examination of the well-log lithology data in map form revealed that the data were not only useful for model input, but also were useful for understanding the glacial hydrogeology of a multistate area. The distribution of K and K_z provided a three-dimensional view of aquifer and confining systems. The distribution of K_{sc} revealed an aerial difference from state to state and a relation to glacial sediment texture. Median K_{sc} was larger for Indiana (264 ft/d) than for Michigan (89 ft/d) or Wisconsin (48 ft/d). The difference could be related to past glacial processes. A pattern in K_{sc} was observed with sediment texture. Aquifers within sediment textures deposited

under a higher energy environment, for example, outwash textures, had higher values of K_{sc} than those deposited under a lower energy environment.

Introduction

At the request of Congress, the U.S. Geological Survey is assessing the availability and use of the Nation's water resources. In 2005, the U.S. Geological Survey established a program called the National Water Availability and Use (U.S. Geological Survey, 2002). The program is designed to characterize how much surface and ground water is currently available, how water availability is changing, and how much water will be available in the future. The Great Lakes Basin was chosen as a pilot study area. Methods developed during the pilot study are intended to best evaluate the resource and deliver accurate and timely information to planners at local, regional, and national levels.

Part of the study involves constructing a ground-water model for the Lake Michigan part of the Basin. Mandle and Kontis (1992, p. 92) found that most ground-water flow for the northern Midwest United States, which includes the Lake Michigan Basin west of Lake Michigan, occurs in the glacial sediments as opposed to the underlying bedrock. Therefore, adequate representation by the model of the horizontal and vertical hydraulic conductivity of the glacial sediments is important to the accuracy of model flow simulations. Adequate representation of the glacial sediments is difficult because of their extreme heterogeneity. The occurrence of interbedded sand and gravel deposits within the fine grained glacial sediments is difficult to predict. Many observation points are required to define the extent of intermittent zones of aquifer and nonaquifer sediments. Neff and others (2005, p. 7) explained the improvement in using the lithology information from well logs to better represent geology beneath the surface. Abundant data that can provide information on hydraulic conductivity, as well as ground-water levels, are available from well driller well logs that were provided to state regulatory agencies. The monitoring-, observation-, and water-well logs have been computerized by the states, stored in statewide data bases, and are available for processing and analysis by computer programs to provide data for model construction.

Purpose and Scope

This report documents the process of converting information from about 458,000 electronic well logs from the state data bases of Michigan, Wisconsin, and Indiana into data useful for construction of a regional ground-water model for the Lake Michigan Basin. The report first establishes the use and limitations of well logs in hydrogeologic studies, and then explains how lithologic descriptions recorded by drillers onto well logs were converted into hydraulic conductivities. The procedures are explained for converting specific-capacity data

(well pumpage rate, time, and water-level drawdown data) into values of horizontal hydraulic conductivity and for characterizing the horizontal hydraulic conductivities of different hydrogeologic units. A quality-control procedure to eliminate outlier water-level data from a well-log data set to be used as model-calibration data also is presented. No well-log data were available for the glacial sediments beneath Lake Michigan, therefore, the procedure for estimating glacial thickness and the horizontal and vertical hydraulic conductivities of the sediments beneath the lake is explained.

Results of the well-log processing are illustrated, analyzed, and compared to similar hydrogeologic products as a qualitative method to evaluate the accuracy of well-log data. Distributions of the well-log derived values of horizontal and vertical hydraulic conductivity are illustrated across three states for the top layer (layer 1) of the ground-water model. Patterns in the distributions are described and analyzed. The distributions are compared to surface-geology and ground-water availability maps and a type of flow-duration map as a way to qualitatively evaluate the accuracy of the hydraulic conductivity distributions. This comparison process revealed that well-log based products have potential advantages over conventional products like surface geology maps to determine hydrologic parameters, and those advantages are presented.

Geologic Setting

The 181,000 mi² study area covers nearly all of Michigan and parts of Wisconsin and Indiana (fig. 1). The study area corresponds to the area of the Lake Michigan ground-water model currently (2008) being constructed as another part of the pilot study on water availability and use. Although northeastern Illinois is included in the modeled area, hydraulic conductivity data for that area were not derived from well logs. Data for Illinois were derived from another model being constructed by the Illinois State Water Survey for northeastern Illinois (Lin and others, 2008). No well logs were analyzed for Ohio or the Canadian parts of the ground-water model. Hydraulic conductivity, to be discussed in the modeling report, was instead based on texture classes defined by Fullerton and others (2003) or Soller and Packard (1998).

The study area is overlain by glacial sediments deposited by four glaciations, which created mostly flat to gently rolling topography (Weist, 1978, p. J3). Although glacial sediments cover the study area, in some areas those sediments are reworked into different deposits, such as alluvium. Glacial sediments can be as much as 1,100 ft thick in Michigan (Weist, 1978, p. J3) but usually are less than 200 ft thick (fig. 2). The glacial sediments at the surface consist of several textures, but fine grained sediments dominate (table 1). The distribution of the textures is shown in a surface geology map presented later in the report. A description of the regional surficial aquifer system within the glacial sediments is provided by Coon and Sheets (2006, p. 7–8). The glacial sediments overlie mostly consolidated sedimentary formations

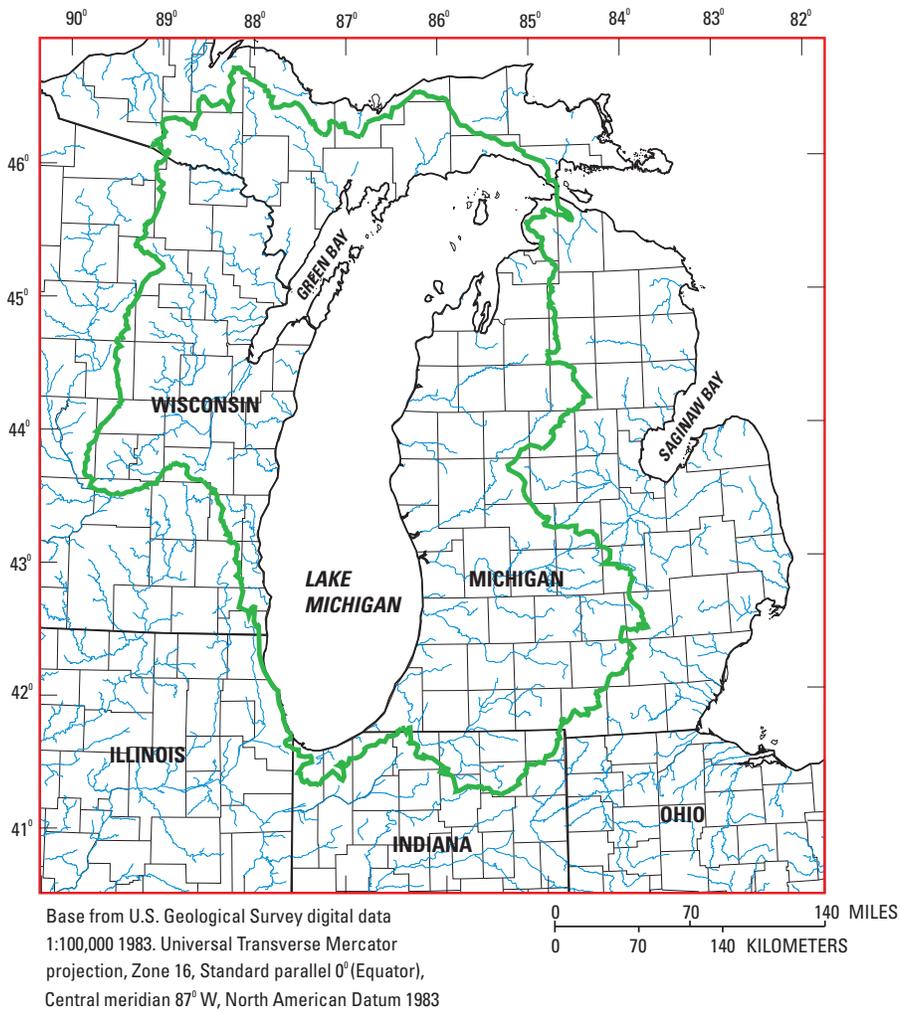
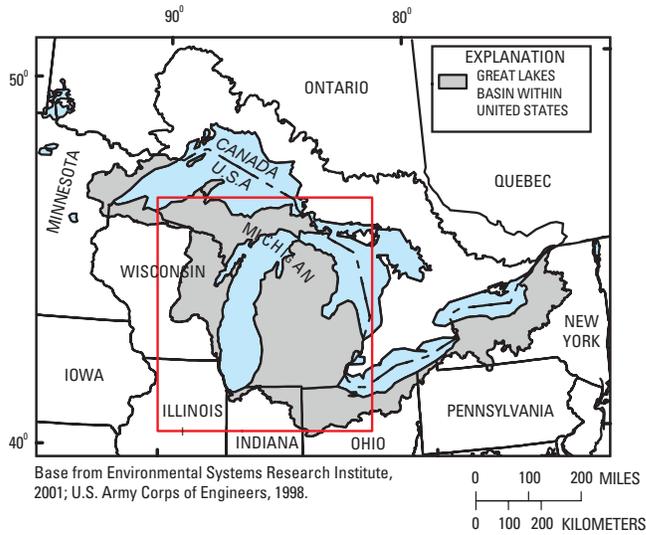


Figure 1. Study area of the Lake Michigan ground-water model and boundary of the Lake Michigan Basin.

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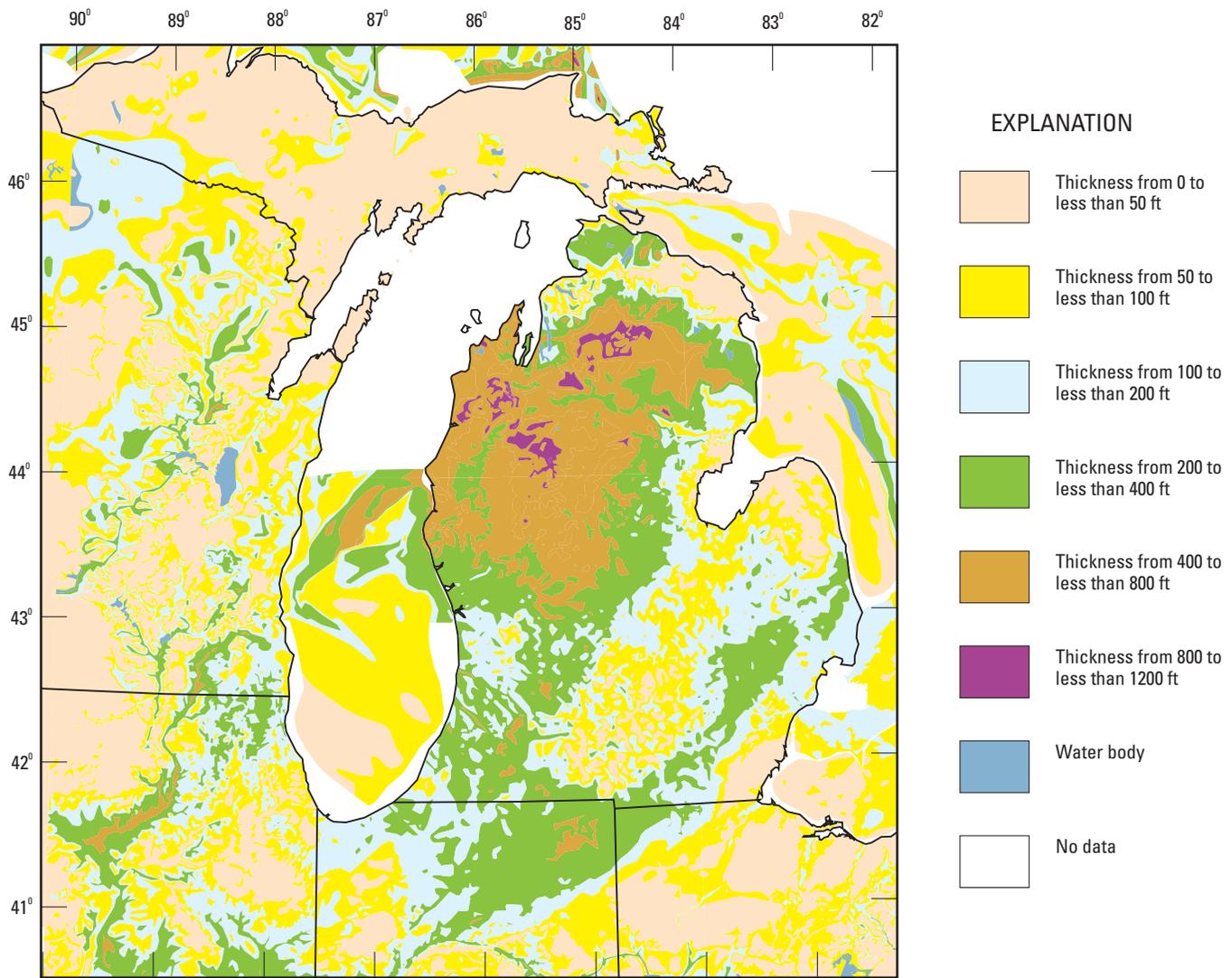
ranging in age from Jurassic to Precambrian Period (Sheets and Simonson, 2006, p. 4). The major bedrock feature is the deep sedimentary basin in Michigan (Weist, 1978, p. J3) that is reflected in the area of thick Quaternary Period sediments shown in figure 2. Additional information on the bedrock formations can be found in Sheets and Simonson (2006).

Table 1. Amount of surface area of the Lake Michigan ground-water model covered by individual sediment textures and sediment texture classifications.

[Data from Fullerton and others, 2003]

Sediment texture	Area covered by sediment texture (percent)	Sediment texture classification
Till	55.9	Fine grained
Coarse ¹	24.6	Coarse grained
Lake sediments	17.5	Fine grained
Peat and muck	1.5	Fine grained
Loess	0.5	Fine grained

¹Outwash, ice contact forms, sand, colluvium, alluvium



Base from U.S. Geological Survey digital data 1:100,000 1983. Universal Transverse Mercator projection, Zone 16, Standard parallel 0° (Equator), Central meridian 87° W, North American Datum 1983

Sediment thickness from Soller and Packard, 1998

0 70 140 MILES
0 70 140 KILOMETERS

Figure 2. Thickness of Quaternary Period sediments in the study area of the Lake Michigan ground-water model.

Processing of Well-Log Data

This section begins by providing a qualitative evaluation of data from all available well driller reports (well logs) for use in hydrogeologic interpretations, then describes the processing of data from well logs into hydrogeologic information useful for ground-water modeling. The methods used to convert (1) lithology descriptions into assumed values of horizontal and vertical hydraulic conductivity for entire model layers, (2) pump-test data into point values of horizontal hydraulic conductivity, and (3) static water levels into water-level calibration data are presented. Hydraulic conductivities for the glacial sediments beneath Lake Michigan also were required for the ground-water model, but well logs do not exist in that area. An alternate method for estimating horizontal and vertical hydraulic conductivity for Lake Michigan sediments is explained.

Evaluation of Well-Log Data

Data recorded on well logs from monitoring, observation, and water wells may not be thought as reliable as those collected by geologists or hydrologists who are typically guided by training programs or a quality-assurance plan during data collection. Still, the data can potentially be useful when other data are not available. For example, Fowler and Arihood (1998, sheet 1) used well-log water-level data to indicate general ground-water flow direction and water-level altitude for St. Joseph County in northern Indiana. By comparison, the mean absolute error between 46 water levels measured in 1990–1992 by U.S. Geological Survey staff (Bayless and Arihood, 1996) and a kriged water-level surface created from the well-log data was 4 ft. Horizontal hydraulic conductivity from a study area in northeastern Illinois calculated from driller-collected specific-capacity data (12 ft/d) compared favorably with values calculated from professionally supervised aquifer tests (8 ft/d) and ground-water model calibration (17 ft/d) (Kay and others, 2006). The following evaluation is a qualitative one that discusses the limitations of well-log data and the general approach of typical driller practice in well-log data collection.

Well logs can provide aerially extensive hydrogeologic information. However, as with all data sets, sources of errors need to be considered. Errors in location of the well log can result in misplaced lithologies and incorrect altitudes for those lithologies. Errors in location also result in incorrect altitudes for ground-water levels. The potential for sampling bias from water-well logs is present because wells are drilled only until they reach aquifer material that can yield sufficient water. Aquifers at higher or lower altitudes with possibly different characteristics are not sampled. Interpolation between points where aquifers yield sufficient water could result in the impression that all aquifers in the area are transmissive, yet the area may contain generally low-yielding aquifers. In addition, the low-yielding aquifers, if encountered, are not reported. These limitations to well-log information need to be consid-

ered during the interpretation of the data. The large quantity of well-log data may help to reduce the influence of occasional errors, but in areas with few well logs, the uncertainty in interpretation increases.

Well-log data are recorded by well drillers who generally are not required by an organization to follow specific procedures in data collection. Therefore, questions can arise as to the quality of data collected by drillers. During a series of ground-water availability studies by the U.S. Geological Survey during the 1970's and 1980's (for example, Lapham, 1981), project hydrologists decided to interview about 15 to 20 well drillers in four central Indiana counties about their data-collection practices. Information was acquired about how each driller recorded lithologies, static water levels, and specific-capacity data. Although some drillers used unacceptable practices, such as completing the well log long after drilling the well, the majority of drillers followed timely and reasonable procedures, such as waiting until a true static water level had developed before recording the static water level for the well. The impression from the interviews was that the typical driller exhibited a desire to provide accurate information and recorded data in a manner approaching that used by a geologist or hydrologist. The driller sometimes demonstrated the same desire as a scientist to accurately record his findings. In addition, the driller may be driven by economic interests in having to justify well placement on the basis of drilling records. That is, it may be difficult to sell a deep well when shallow wells have been successfully established.

Calculation of Equivalent Horizontal and Vertical Hydraulic Conductivities for the Glacial Layers of the Model

Well logs provided about 458,000 observation points with more than 1.8 million records of lithology (fig. 3). The well logs have become computerized and stored in statewide computerized data bases. The well logs from Wisconsin (Wisconsin Geological and Natural History Survey, 2004), Michigan (Michigan Department of Environmental Quality, 2003), and Indiana (Indiana Department of Natural Resources, 2002) were obtained for processing in 2006 and 2007 by computer programs into information useful for ground-water modeling.

Well-log processing used ARC Macro Language programming (AML) available in ArcInfo software (Environmental Systems Research Institute, 2003) and consisted of several steps that are illustrated in figure 4. The well-log data bases were obtained as two files formatted as American Standard Code for Information Interchange (ASCII) for each state. The first file contained basic well-site information, such as owner name, X-Y coordinates of the well, well depth, aquifer-test data, and water level. This file was converted into a point coverage. The second file was the record of lithologies recorded by the well drillers. The second file was converted into an INFO data base, which was then read by a second AML program that automatically converted driller lithology

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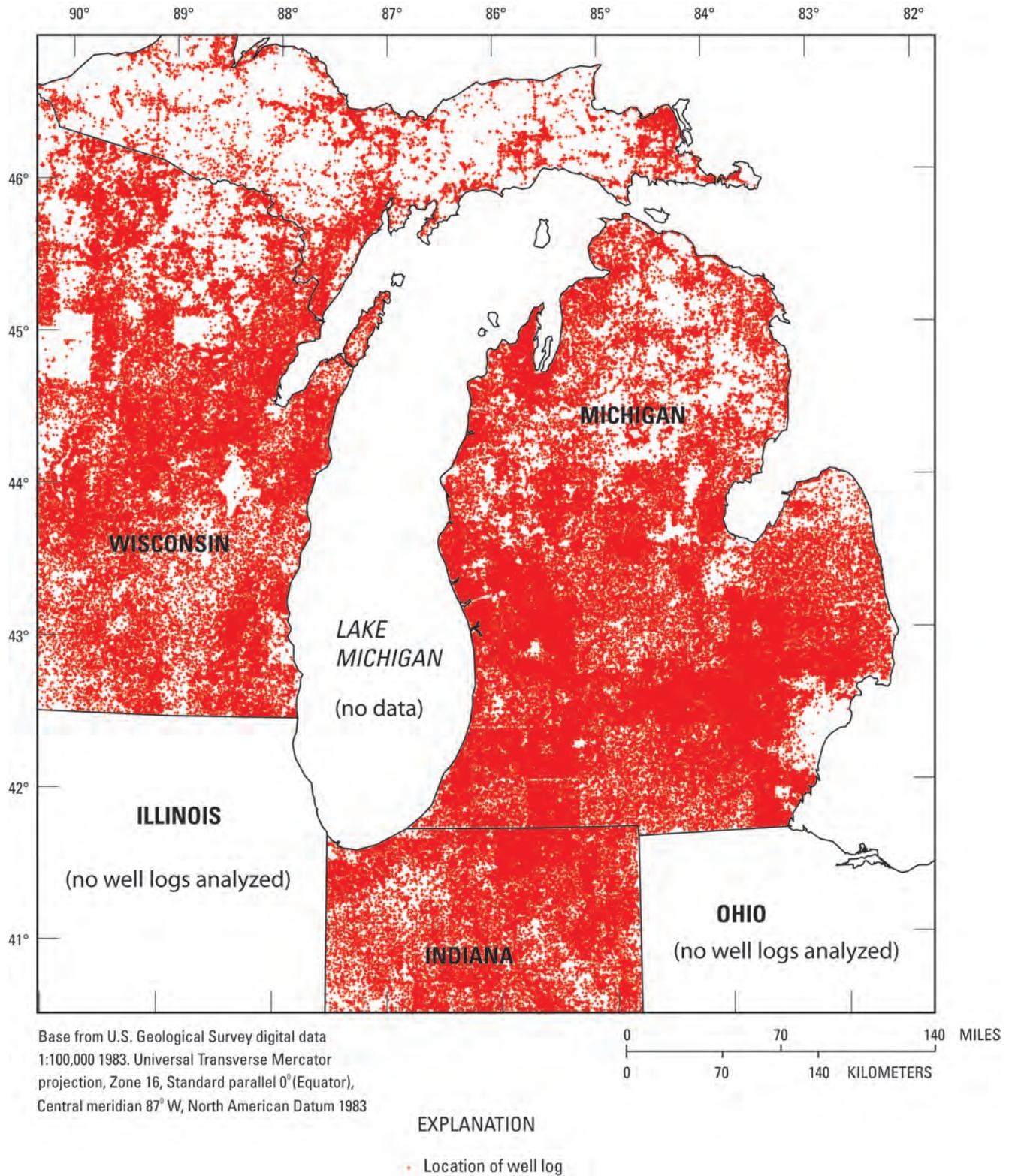


Figure 3. Distribution of logs from drilled wells used to generate hydraulic conductivity and water-level data for the glacial layers of the Lake Michigan Basin ground-water-flow model.

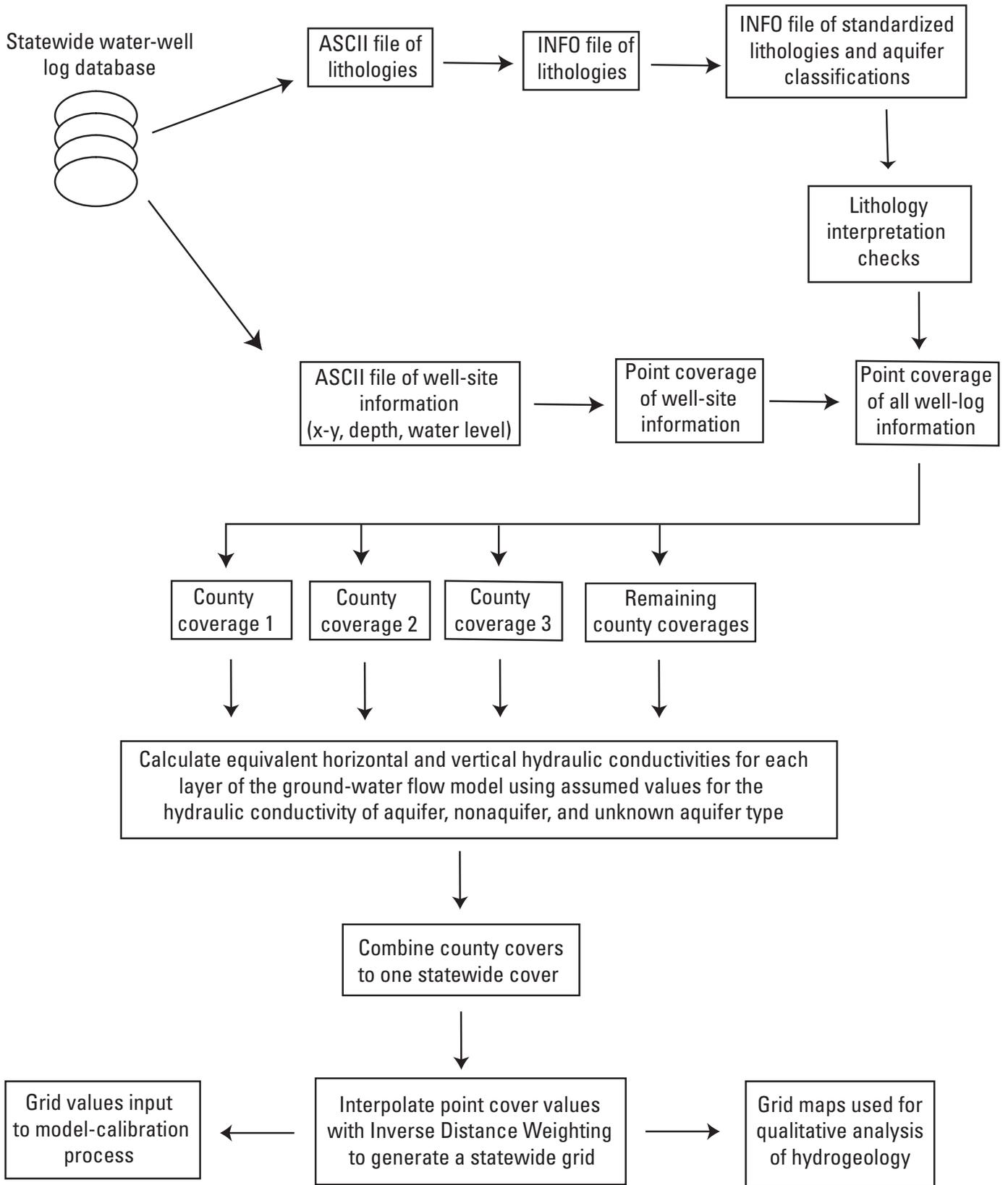


Figure 4. Flow chart for processing well-log lithologies into grids of horizontal and vertical hydraulic conductivity.

descriptions into standard U.S. Geological Survey Ground-Water Site Inventory System (GWSI) lithology codes (Mathey, 1989, p. 2–110). The program searches for key phrases that describe lithology, such as sand, clay, or silt, and compares the phrases to lithology text strings in an interpretation file that converts the driller text into a GWSI lithology. The second AML program searches the phrases for as many as three lithology descriptors to determine the best match to a GWSI code, which can have as many as three lithology descriptors. The AML also classified each lithology into either aquifer or nonaquifer material. After the interpretation, the lithology records were checked for lithology descriptions that could not be interpreted because of misspellings or new lithology phrases not referenced in the interpretation file. Additional lithology text strings were added to the interpretation file so that the undefined lithology records could be converted to GWSI lithologies. Errors in the lithology record, such as gaps in the lithology record or the appearance of glacial lithology in a bedrock section of the well log, resulted in eliminating the well record from the statewide coverage. On average, about 5 percent of the well logs were deleted from each original state well-log data base. The INFO lithology file was joined to the well-site information point coverage, and then subdivided into individual county coverages. Further processing is done on single county coverages, because data for an individual well can be accessed during subsequent programming more rapidly if the data are in a county coverage rather than in one set of the entire state.

The lithology records in the county coverages were converted by a third AML program into an equivalent, or effective horizontal and vertical hydraulic conductivity for each well site for specified thicknesses of the glacial sediments. The specified thicknesses correspond to the predefined thicknesses of glacial-sediment layers used in the ground-water model. The glacial sediments are represented in the model by three layers: a maximum 100-ft thick layer 1, a maximum 200-ft thick layer 2, and a layer 3 that accounts for all remaining glacial thickness. These layer thicknesses were used to provide a relatively even division of the glacial sediments in the majority of the modeled area and to provide an opportunity to vary the hydraulic conductivity with depth. Some areas have thin glacial sediments, therefore, a value of horizontal and vertical hydraulic conductivity is calculated at a well site for less than the 100- or 200-ft thickness and possibly for fewer than three layers.

The program calculates equivalent horizontal and vertical hydraulic conductivities (referred to subsequently as K and K_Z) for the predefined layers on the basis of percentage of aquifer and nonaquifer material in each layer. The equivalent K is based on an assumed value of 100 ft/day for aquifer material and 1 ft/day for nonaquifer material, whereas the equivalent K_Z is based on an assumed value of 10 ft/day for aquifer material and 0.001 ft/day for nonaquifer material. These values are assumed for convenience to determine a relative contrast between aquifer and nonaquifer material. The calculation of equivalent horizontal and vertical hydraulic conductivity for

each glacial layer of the model uses the following equations (DeWeist, 1965, p. 231–232):

$$K = \frac{\sum_{i=1}^n (t^i \times K^i)}{\sum_{i=1}^n t^i} \quad (1)$$

$$K_Z = \frac{\sum_{i=1}^n t^i}{\sum_{i=1}^n \frac{t^i}{K_z^i}} \quad (2)$$

where

K = equivalent horizontal hydraulic conductivity for a model layer

K_Z = equivalent vertical hydraulic conductivity for a model layer

K^i = assumed horizontal hydraulic conductivity of the i th lithology record within the layer

K_z^i = assumed vertical hydraulic conductivity of the i th lithology record within the model layer

t^i = thickness of the i th lithology record within the model layer

i = the i th lithology record,

and

n = total number of lithology records within the model layer.

After equivalent hydraulic conductivities have been calculated for well sites in each county, the county coverages are rejoined into a statewide coverage. Sites without lithology records were eliminated from the statewide coverage.

To further explain the relation between the equivalent hydraulic conductivities, assumed values of hydraulic conductivity, and quantity of aquifer / nonaquifer material, the relation is illustrated in figure 5. The K value is proportional to the thickness of aquifer material present within the model layer. For a 100-ft layer that is 50 percent aquifer material (50 ft of aquifer), K is 50 percent of the maximum value (100 ft/d), or 50 ft/d. The proportional relation can be applied to other assumed ranges for horizontal hydraulic conductivity of aquifer material, such as 1 to 1000 ft/d. That is, the distribution of horizontal hydraulic conductivity does not have to be recalculated from the well-log record; a simple multiplication can be applied to the distribution of values. The K_Z is not proportional to the percentage of aquifer material, but is strongly influenced by the value chosen for vertical hydraulic conductivity of the nonaquifer material. The lower the value for vertical hydraulic conductivity of nonaquifer material, the more quickly K_Z for the model layer decreases in value as the percentage of aquifer material decreases from 100 percent. If a different value for vertical hydraulic conductivity of nonaquifer material is chosen during model calibration, then K_Z has to be recalculated. According to equation 2, the recalculation requires values for

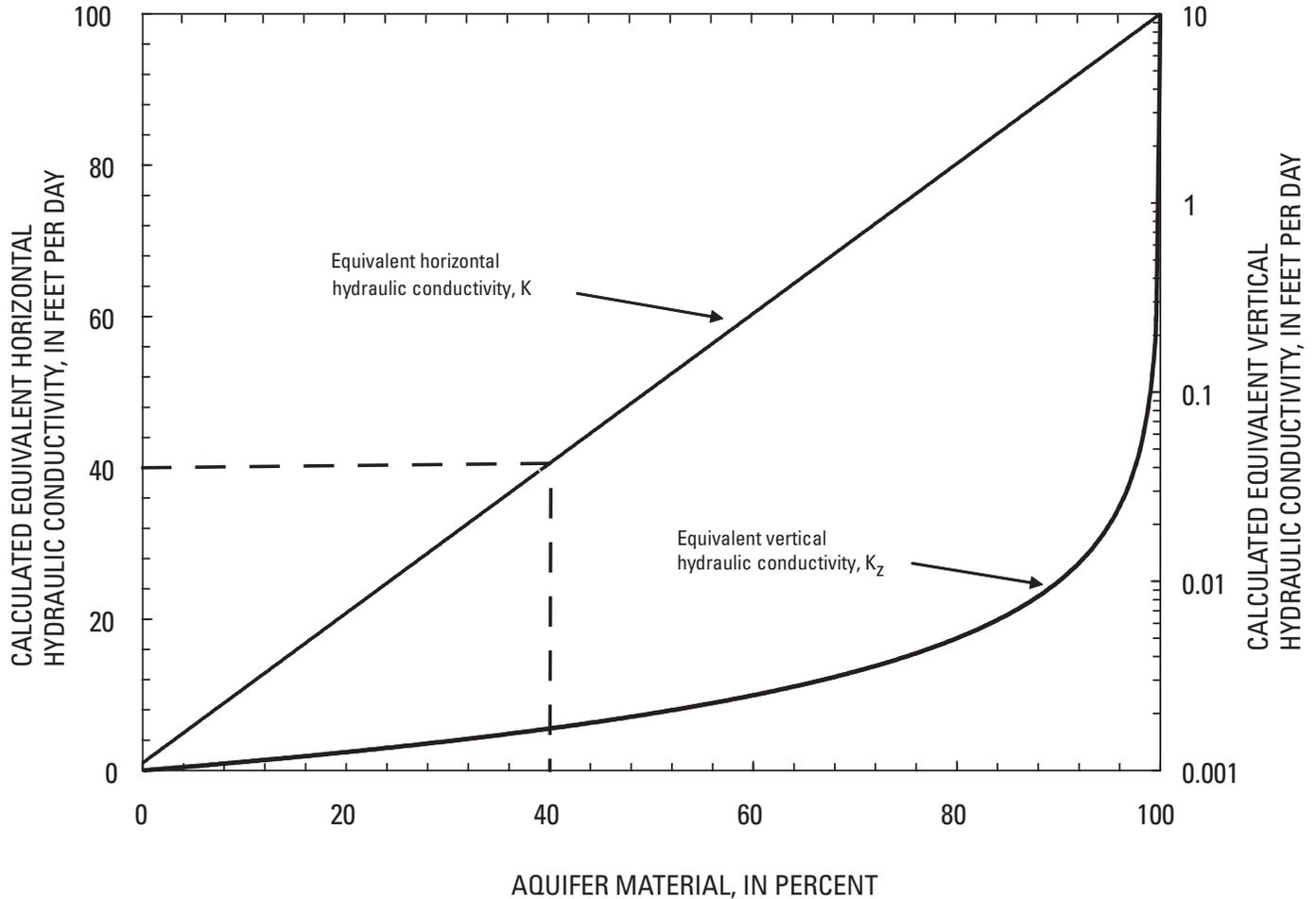


Figure 5. Relation between lithology-based equivalent hydraulic conductivity and the percentage of aquifer material in the model layer.

thicknesses of aquifer and nonaquifer material (other values in equation 2 are known or given), which can be obtained from the equivalent horizontal hydraulic conductivity. For example, if the model layer is 100 ft thick at a model cell and aquifer material has an assumed horizontal hydraulic conductivity of 100 ft/d, and if K at a model node is 75 ft/d, then the thickness of nonaquifer material is 25 ft and aquifer material thickness is 75 ft. A value for vertical hydraulic conductivity is assumed, and the remaining unknown, K_z , can be calculated.

Several additional details associated with the calculation of hydraulic conductivities are provided to help understand the processing of well logs:

1. The value of assumed hydraulic conductivity for a lithology record is determined by a file that defines all GWSI lithology codes to be aquifer, nonaquifer, or undefined lithology, such as drift. For undefined lithologies, the geometric mean of the equivalent horizontal hydraulic conductivity of aquifer and nonaquifer material is assigned to the lithology record. Because aquifer material has an assumed horizontal hydraulic conductivity of 100 ft/d and nonaquifer material has an assumed horizontal hydraulic conductivity 1 ft/d, then the assigned horizontal hydraulic conductivity for the undefined lithology is approximately 10 ft/d.
2. The model design uses three layers (1, 2, and 3) to represent the glacial sediments. Where glacial sediments range from 1 to 100 ft thick, layer 1 has a thickness equal to the glacial thickness. Where glacial sediments are from 101 to 300 ft thick, layer 1 has a thickness of 100 ft and layer 2 has a thickness equal to the remainder of the glacial thickness. Where glacial sediments are in excess of 300 ft thick, layer 1 extends from the land surface to 100 ft; layer 2 extends from 100 ft to 300 ft below land surface; and layer 3 extends from 300 ft below land surface to the depth where the glacial sediments contact the top of bedrock.
3. A lithology record (for example, clay 85 to 120 ft) may begin in one layer and end in the layer below. For that case, the lithology record, along with the associated hydraulic conductivity, is split proportionally between the two layers.

4. A penetration depth into each model layer, expressed as a percentage, was calculated for each well. Wells that penetrate to a depth equal to or greater than 50 percent of a layer were selected from the coverage of equivalent hydraulic conductivity and used to create the continuous distributions of hydraulic conductivity.
5. Hydraulic conductivities are calculated only for that part of the lithology record that is below the driller-measured water level. In a few wells, the water level is greater than 100 ft deep, meaning that layer 1 is dry and calculated hydraulic conductivities are 0 ft/d. If no water-level value is available for the well, then the equivalent hydraulic conductivities are based on all lithology records within a layer. This decision should not noticeably alter the distributions because 85 percent of the wells have water-level data and the water table is usually close to the surface in fine grained sediments, which comprises most of the area. The near-surface water table is evidenced by the general need for tile drainage and the high surface-water levels in borrow pits along interstate highways.
6. In areas where the glacial sediments are thicker than 300 ft and a third glacial layer is present, the thickness of the third glacial layer is defined as the thickness from 300 ft below ground to the depth of bedrock as defined by the well log. If the well is not drilled to bedrock, then an estimated thickness of model layer 3 is determined by use of an interpolated value of bedrock surface altitude from a bedrock-surface map constructed in another one of the water-availability set of studies (David Lampe, U.S. Geological Survey, written commun., 2006). The percent penetration of the well into the third layer is based on the actual or estimated layer thickness.
7. If the well extends through a layer bottom, this process also computes an altitude of the layer bottom at each well. If a layer has zero thickness, then the altitude of that layer is assigned the same value as the bottom altitude of the first layer above with a finite thickness. For example, if layer 3 is zero thickness, and layer 2 is 50 ft thick with a bottom altitude of 650 ft, then the bottom altitude of layer 3 is also 650 ft. The altitude of the bedrock surface would also be 650 ft. During model construction, layer 3 would be given a small thickness of 0.2 ft so that the layer is continuous.

Calculation of Aquifer Hydraulic Conductivity from Well-Log Specific-Capacity Data

Values of horizontal hydraulic conductivity for aquifer material also were calculated directly from about 292,000 well logs by use of the specific-capacity data (well discharge, duration of pumping, and water-level drawdown) and are referred to as K_{sc} . The calculated K_{sc} provided an estimate of horizontal hydraulic conductivity. The modeling process used the estimate as a guide for adjusting the assumed K value of 100 ft/d to actual values used in calculating a lithology-based K for a model layer. The process of converting pump-test, or specific-capacity data, into K_{sc} is presented.

Well discharge, duration of pumping, and water-level drawdown for each well were input to a modified form of the Theis equation shown below (Prudic, 1991, p. 11) to determine an aquifer transmissivity, T . Because T appears on both sides of equation 3, an iterative process was used to calculate T by use of an initial value of 500 ft²/d. The process stops after several iterations when the difference between the old estimate and new estimate for T becomes less than 5 ft²/d, and the last new estimate is used. The value of T was adjusted for the effect of partial penetration by the well screen into the aquifer by use of a method described by Butler (1957, p. 160). Values of K_{sc} were calculated by dividing T by the thickness of saturated aquifer material penetrated by the well, which provides a conservative estimate for K_{sc} . Well loss was not accounted for because there were no available data to analyze well loss, and the well was assumed 100 percent efficient. It was assumed that, in whole, the gravel pack around the well screen compensated for the additional drawdown caused by turbulent flow through the well screen and in the well casing. The calculation of T from specific-capacity data is often done, and two alternative, computerized methods are referenced for convenience: (Bradbury and Rothschild, 1985; McLin, 2005).

$$T = 15.32 * \left(\frac{Q}{s}\right) * \left(-0.577 - \log_e \left[\frac{r^2 S}{4Tt}\right]\right) \quad (3)$$

where

- T = transmissivity, in feet squared per day,
- Q = well discharge, in gallons per minute,
- s = drawdown, in feet,
- r = well radius, in feet,
- S = storage coefficient (dimensionless),

and

- t = time, in days.

Equation 3 requires values for storage coefficient, well diameter, and screen length. Well logs often do not have all of these data (and almost never have storage coefficient), therefore estimates need to be determined. If screen length data was missing, then a screen length of 5 ft was assumed when pumping rate was less than 200 gal/min, and a screen length of 20 ft was assumed if the pumping rate was greater than or

equal to 200 gal/min. If well diameter data was missing, then a well diameter of 4 inches was assumed when pumping rate was less than 200 gal/min, and a well diameter of 8 inches was assumed if pumping rate was greater than or equal to 200 gal/min. If the well was confined, then the storage coefficient was assumed to be 0.0001; if the well was unconfined, then the storage coefficient was assumed to be 0.15. These values are well within ranges suggested by Freeze and Cherry (1979, p. 60–61), and their difference from actual values should not significantly affect the calculation of T (Prudic, 1991, p. 12). Confined or unconfined conditions were determined by observing if the pumping level for the water surface was above or below the top of the aquifer, respectively.

Calculation of horizontal hydraulic conductivity from transmissivity requires a value for well discharge. Well discharge could be generated by a submersible pump, an air line (air flow from a compressor), or other methods, like a bailer. Because different pumping methods might result in different values for K_{sc} , a comparison was made between K_{sc} values calculated using data from different methods of pumping. Table 2 presents the results of the effect from different pumping methods on median values of K_{sc} for wells in Michigan (Michigan Department of Environmental Quality, 2005) and Indiana (Indiana Department of Natural Resources, 2002). Sample size for each glacial texture was large, always greater than 1,000 samples. The values are for sites where the pumping method was known and do not include all of the sites with specific-capacity data. The table shows that median K_{sc} for the air flow method is greater than that for the submersible pump method in Indiana, however, the opposite is true in Michigan. A rank-sum test (Helsel and Hirsch, 1995) was done on the individual state data sets and confirmed that the air flow method of pumping produces larger values of K_{sc} in the Indiana data and smaller values in Michigan ($p > 0.999$). Because it is not known which method of pumping produces a more representative value of K_{sc} , and because the median K_{sc} using all methods of pumping is between the median K_{sc} for air flow and submersible pumping, the data from all pumping methods were used in the analysis of patterns in K_{sc} for the three states.

Table 2. Comparison of median horizontal hydraulic conductivity calculated using specific capacity data from Indiana and Michigan.

[ft/d, feet per day; (1,625), number of samples; data from Michigan Department of Environmental Quality (2005) and Indiana Department of Natural Resources (2002)]

Method of pumping	Median horizontal hydraulic conductivity in Indiana (ft/d)	Median horizontal hydraulic conductivity in Michigan (ft/d)
All	177 (6,425)	18.7 (162,496)
Air flow	226 (2,076)	14.2 (50,455)
Submersible pump	143 (976)	37.0 (12,471)

Compilation of Water-Level Calibration Data from Well Logs

About 396,000 water-level values (one value per well log collected at the time of well construction) are available from the state well-log data bases for the Lake Michigan ground-water model area. In some areas, the well-log water-level data are the only source of information on water surfaces of the various aquifers. Such a large data set should provide many reliable water levels to aid in model calibration. In fact, Fowler and Arihood (1998, sheet 1) used well-log water-level data to indicate general ground-water flow direction and water-level altitude for St. Joseph County in northern Indiana. Water levels measured by U.S. Geological Survey (Bayless and Arihood, 1996) were on the average (mean absolute error) 4 ft from a kriged water-level surface created from the driller data, which is sufficient accuracy for model calibration data. However, appropriate screening techniques were used to eliminate inaccurate and inconsistent data.

All well logs that contained a depth to static water level from land surface were processed into a coverage and evaluated for accuracy and consistency. The process of selecting water-level data for model calibration is illustrated in figure 6. A static water-level altitude was calculated by subtracting the depth to the static water level from the land-surface altitude. Land-surface altitude typically is recorded on the well log along with the depth to static water level. If land-surface altitude was not recorded, then a value interpolated from a 30-meter digital elevation model (DEM) was substituted for the missing altitude data. Substituted values of altitude represented less than 1 percent of the data set, except for the data from Wisconsin. Land-surface altitude data were not available from the well logs; therefore, all altitude data from Wisconsin (86,000 values) were derived from the DEM. The error in land-surface altitude will vary depending on the method used to determine altitude. Altitudes commonly are estimated from topographic maps. The author examined the potential error in altitudes estimated from topographic maps by comparing estimated altitudes with surveyed values for 35 well sites. The mean absolute error for altitudes estimated from a topographic map was 1.68 ft. The root mean square error for altitudes based on DEM data was estimated to be 8.01 ft (American Society for Photogrammetry and Remote Sensing, 2007, p. 106).

Although well-log water levels can be helpful, many water levels in the data set are expected to contain significant measurement and location error and should be eliminated from the set of calibration water levels. A method was developed for quality assurance of well-log water levels. The method used kriging to determine a water-level surface, and then used the difference between the kriged surface and water levels from a well log to identify and eliminate outlier water levels.

12 Processing, Analysis, and General Evaluation of Well-Driller Logs for Estimating Hydrogeologic Parameters

A kriging interpolator available in ArcInfo, version 8.3 (Environmental Systems Research Institute, 2003) was used to estimate a water-level surface grid that had grid cells every 2,000 ft. Kriging was used to improve the estimation of water levels by taking advantage of the spatial correlation existing

between water-level measurements. Ordinary kriging (McBratney and Webster, 1986) was applied to the individual water-level data sets from each state by use of the kriging parameters shown in table 3. Included in the table is the mathematical function used to fit the semi-variance data. Kriging was applied to the individual state data sets, but trends in population distributions between each state were not considered.

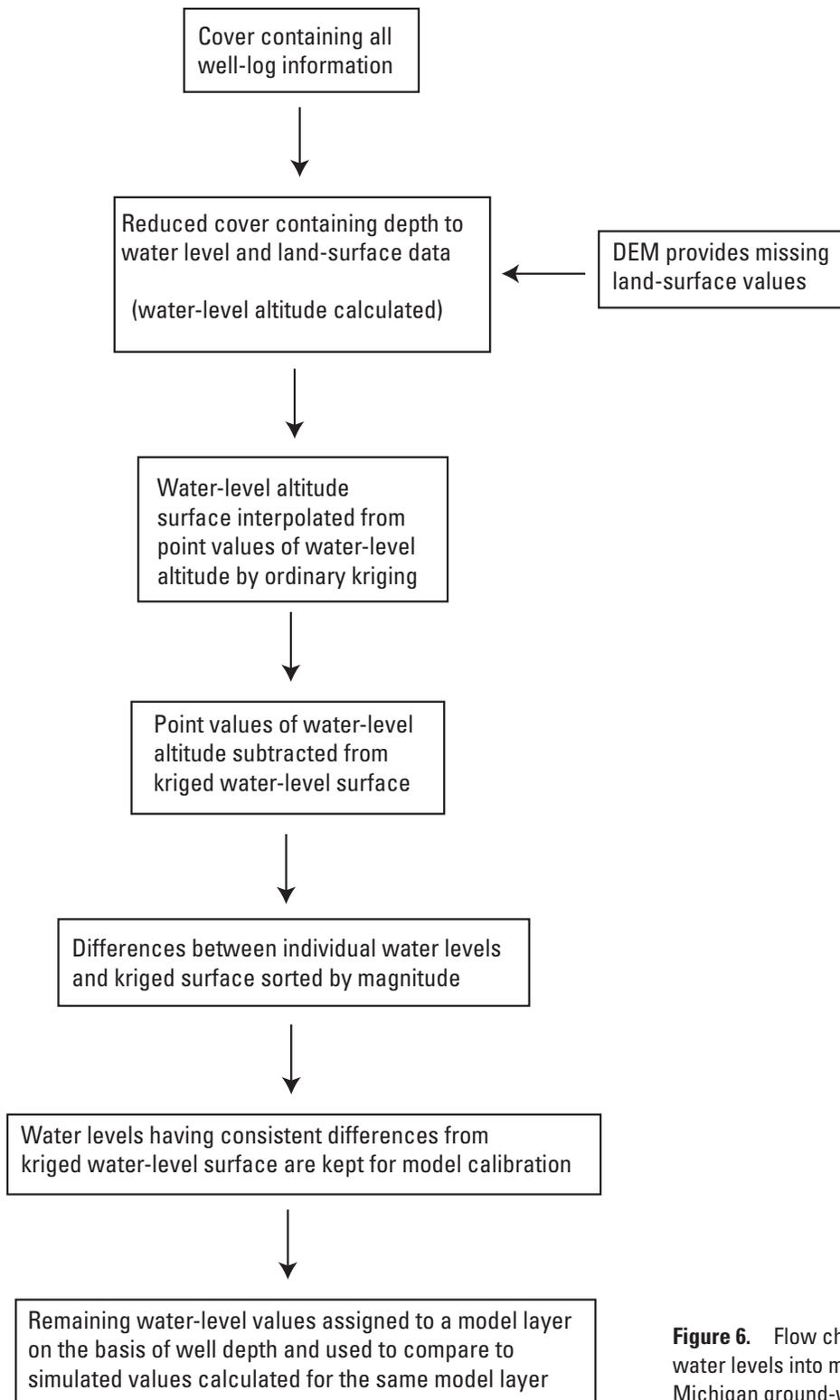


Figure 6. Flow chart for the processing of well-log water levels into model-calibration data for the Lake Michigan ground-water-flow model.

Lack of consideration of the trends probably introduced error in the estimates, but the errors were not considered significant relative to the use of the water-level surface, which was to eliminate outlier water levels. As explained in the next paragraph, only large differences between individual measurements and the kriged water-level surface were used to identify outliers.

Table 3. Kriging parameters used to interpolate ground-water-level surfaces in Indiana, Michigan, and Wisconsin for the Lake Michigan ground-water-flow model.

State	Number of data points	Mathematical function	Sill	Range
Indiana	41,407	Gaussian	866	25,754
Michigan	265,399	Spherical	3,057	36,000
Wisconsin	113,886	Gaussian	3,060	36,000

After the water-level surface was generated, the difference between the surface and water level for each well was calculated and stored in the coverage. The difference was used to identify and eliminate outlier water-level values from each state data set. The difference values were sorted from smallest to largest, and they were then examined to determine a break between large differences that were unique in value and differences that were repetitive. For example, the sorted differences might change from -450 ft to -400 ft, then -380 ft. Eventually, the sorted differences change by less than 1 ft between two values, for example from -39 ft to -38 ft, to -37.5 ft. At that point, water levels with similar values for difference exist within the data set. Similar difference values may indicate that the measured water-level values are confirming each other and should remain in the data set. The example difference value of -39 ft may seem so large that it should not be included, but legitimate reasons may exist for the large difference. For example, the water levels might be from a deep or shallow layer or near a pumping center. For each state, only the water levels with similar values for the differences were kept in the coverage. The break in value for acceptable differences varied from state to state, but was somewhat above 100 ft.

All water levels in the data set were used together to create a single water-level surface rather than create a different water-level surface for each of the three glacial layers in the ground-water model. Past experience with dividing water-level data into layers and then interpolating several individual surfaces has resulted in some areas having an unreasonable water-level surface because of insufficient data. The use of all water levels together typically results in a reasonable surface, but it may create the concern that the resulting water-level

surface does not represent any specific model layer. If the surface does not represent a specific layer, then the surface may not be thought an appropriate indicator of outlier water levels for any individual layer. However, in Indiana the water-level difference between aquifers is often only 1 to 2 feet (Arihood and Cohen, 1998). Water-level differences between layers may be greater than a few feet in many parts of the study area (for example, see Bayless and Arihood, 1996, p. 17). In those areas, the goal is to avoid eliminating water levels that have large differences between their values and the interpolated surface, but that still may be representative of the model layer to which the water levels belong. The process of retaining water-level values that have consistent differences from the average surface, as described previously, will help avoid eliminating water levels different from the average surface, and that still may be useful for calibration.

Although a single water-level surface was generated, the individual water-level values were grouped by model layer. The grouping was done by assigning a layer to the water-level value on the basis of well depth. During model calibration, the water-level values assigned to a specific layer were compared to model-simulated values for the same layer.

The interpolated water-level surface also was created using water levels from all periods of the year and from all available years, which covers decades of water-level data. The vast majority of the data typically span about a 40-year period from the 1960's to about 2005. Water-level data from a specific year and time of the year could be reselected to create a time-specific surface, but the smaller number of data points would lead to areas with few data points and uncertain water-level surface. The use of data from all times of the year and all years results in a data set that contains the entire range of water levels. From such a data set, the kriging interpolator would determine the most likely, or average, water-level surface, which may avoid representations of extreme conditions.

Another way in which the single interpolated water-level surface can be effectively used to detect outlier water levels that should be eliminated from model-calibration data is to define areas where wells are located mostly within a specific aquifer. In those areas, the interpolated water-level surface reflects mostly the water-level surface of that specific aquifer. Water levels that vary by several feet from the interpolated surface can be eliminated by future users of the data set.

The number of locations where well-log water-level data could be used to calibrate the ground-water model for changes in water level from one layer to the next is probably limited. The sites with water-level data representing both model layers must be essentially at the same location. Also, multiple water levels that generally confirm each other must be available for both layers.

Estimation of Thickness and Hydraulic Conductivity for Glacial Sediments beneath Lake Michigan

Few data sources (and no well logs) were available to help provide information on thickness and horizontal and vertical hydraulic conductivity of the sediments beneath Lake Michigan. Therefore, the values of thickness and hydraulic conductivity for sediments beneath Lake Michigan are inferred from existing map information and, to a large degree, based on what could be considered reasonable by study hydrologists. These values may require adjustment during model calibration.

Different data sources for thickness of sediments were used for different parts of the lake bottom. For the southern half of Lake Michigan (see fig. 2 for extent), work presented by Soller and Packard (1998) was used to estimate the thickness of lacustrine clay and till. For the northern half of Lake Michigan, a shaded relief map (Colgan and Principato, 1998) and a discussion of the geomorphology of the lakebed (National Oceanic and Atmospheric Administration, 2005) were used to estimate the thickness and hydrologic characteristics of the glacial sediments. East of Green Bay, Wis. and for two isolated areas of alluvial fan/deltaic deposition, the glacial sediments were modeled as a sequence of sand over nonaquifer materials. In areas where Colgan and Principato (1998) interpreted smoothly varying lake bottom topography, it was assumed that the map indicated deposition of sand and silt. In areas where the map showed highly variable lake bottom topography, it was assumed that the map reflected bedrock surface control of the lake bottom. For areas of highly variable lake bottom topography, it was assumed that a 10 ft thick clay deposit was the only glacial sediment covering bedrock. For the entire near-lakeshore area, the underlying clay and tills were assumed to be capped by sand deposits derived from fluvial sources. The resulting thickness map is shown in figure 7.

Equivalent hydraulic conductivity values (K and K_z) were assigned based upon the composition of the materials within each mapping area. For clay and till, $K = 1$ ft/d and $K_z = 0.1$ ft/d; for sand, $K = 100$ ft/d and $K_z = 10$ ft/d. For areas mapped as being capped by sand, such as around the lake perimeter, the sand deposits were assumed to represent 25 percent of the total thickness of the glacial sediments. For example, where sediments were mapped as 100 ft thick, sand and clay thickness were assumed to be 25 ft and 75 ft, respectively. The assumption of 25 percent sand deposits ensured that there was always a sand and clay mixture, regardless of the total thickness of glacial sediments. The resulting horizontal hydraulic conductivity distribution is shown in the next section.

Analysis of Well-Log Data

The processing of well logs was originally intended to provide distributions of horizontal and vertical hydraulic conductivity for a ground-water model. The values associated with the distributions were manipulated by the calibration process to obtain final values while maintaining the relative differences in the distribution. During the processing, observations of the well-log data in map form indicated additional benefits that accrue from this analysis in the understanding of the hydrogeology of a multistate area. The distributions of K and K_z data provide a three-dimensional view of aquifer and confining systems and the distribution of specific-capacity-based hydraulic conductivity reveal aerial differences from state to state in horizontal hydraulic conductivity of sand and gravel deposits.

The point values of K and K_z calculated from the well-log lithology records for each layer were used as input to an interpolation program within ArcInfo to estimate a grid of values for hydraulic conductivity for each state data set. The K distribution was based on an inverse distance weighting equation shown below (Isaaks and Srivastava, 1989, p. 258–259). According to Isaaks and Srivastava (1989, p. 259), as the value of the exponent p becomes greater, the closest input value has greater influence on the estimated value. The K distribution was calculated with an exponent of 2, which is commonly used (Sheppard, 1968). The K_z distribution used an exponent of 4 on the log of vertical hydraulic conductivity to better represent the abrupt change in K_z that occurs between aquifer and nonaquifer materials. The distributions were calculated for each state, and then the individual state distributions and the distribution beneath Lake Michigan were merged together to construct a single distribution of K and K_z for each model layer. The K and K_z distributions for layer 1, the most continuous layer, are shown in figures 8 and 9, respectively. The figures show the K and K_z distributions based on assumed values for aquifer and nonaquifer materials, but the distributions also reflect the total amount of aquifer and nonaquifer material in the first layer of the model. The grids are based on well logs that penetrate at least 50 percent of the layer thickness.

$$\hat{v} = \frac{\sum_{i=1}^n \frac{1}{d_i^p} v_i}{\sum_{i=1}^n \frac{1}{d_i^p}} \quad (4)$$

where

- \hat{v} = estimated value,
- v_i = input value,
- i = ith input value,
- n = last input value,
- d = distance between the input value and the point of estimation,

and

- p = exponent.

The distributions shown in the previous two figures provided the pattern of change in hydraulic conductivity for the ground-water model. Final values of hydraulic conductivity associated with the distribution were determined during model calibration. For calibration, the original values of hydraulic conductivity were to be expanded from those in the grids (1–100 ft/d) to a wider range that was more effective in

matching observed values of ground-water levels and flows. In addition, the modeled area was divided into zones representing different glacial environments. The addition of zones provided the flexibility of using different multipliers for hydraulic conductivity in areas that may have a consistently higher or lower value because of glacial setting. Final details about the expansion and zonation will be discussed in the future modeling report.

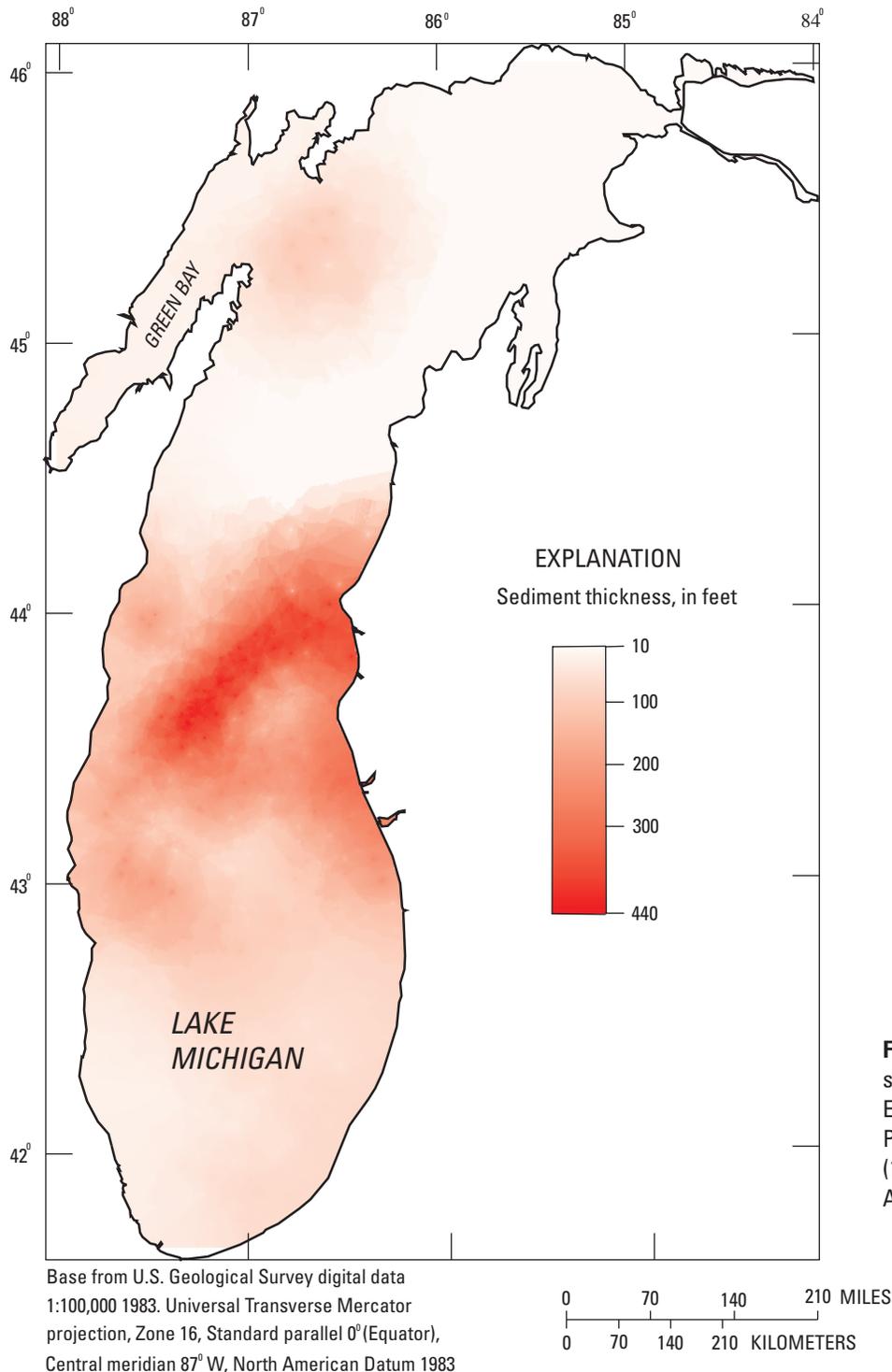


Figure 7. Estimated thickness of glacial sediments beneath Lake Michigan. Estimates based on data from Soller and Packard (1998), Colgan and Principato (1998), and National Oceanic and Atmospheric Administration (2005).

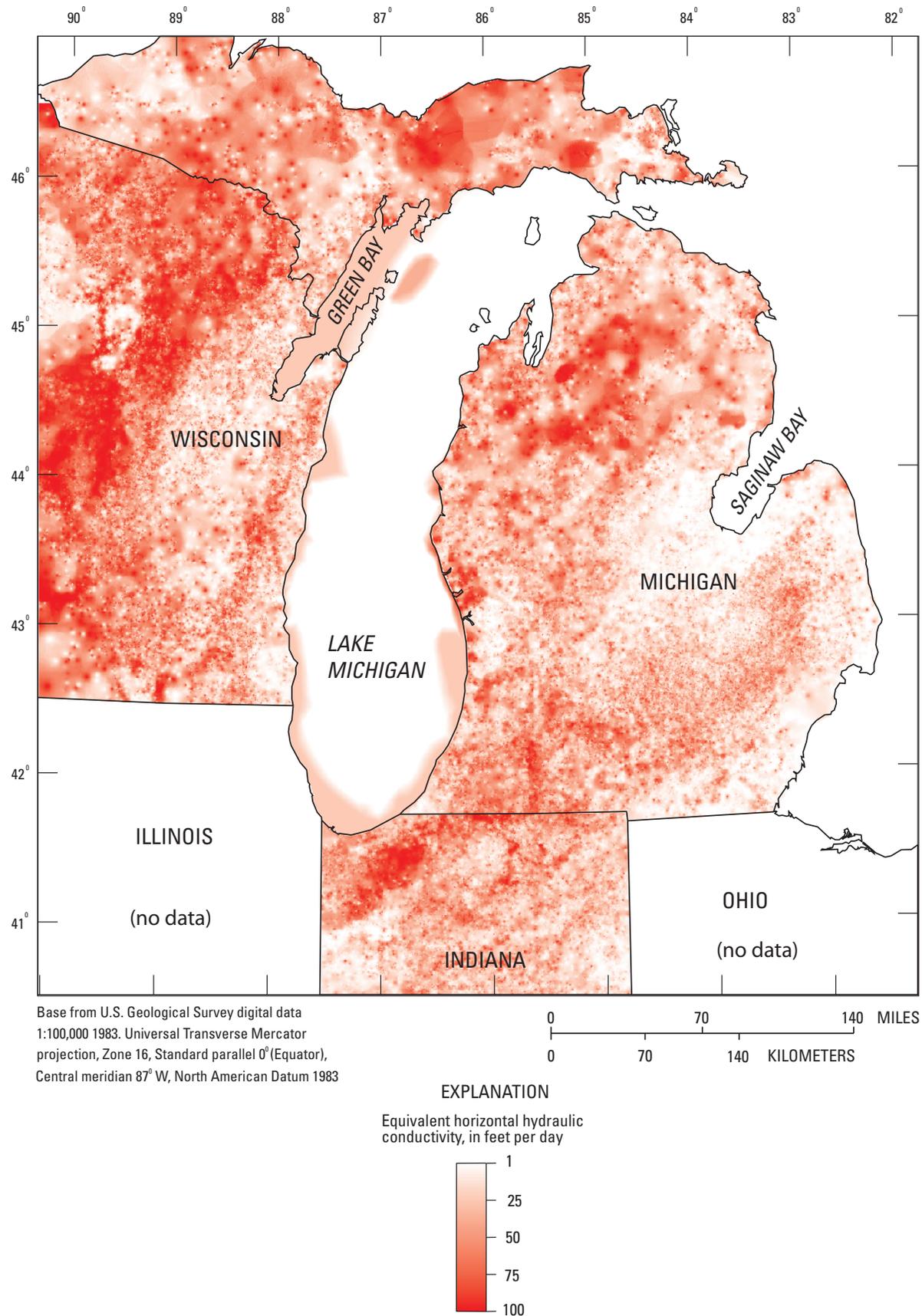
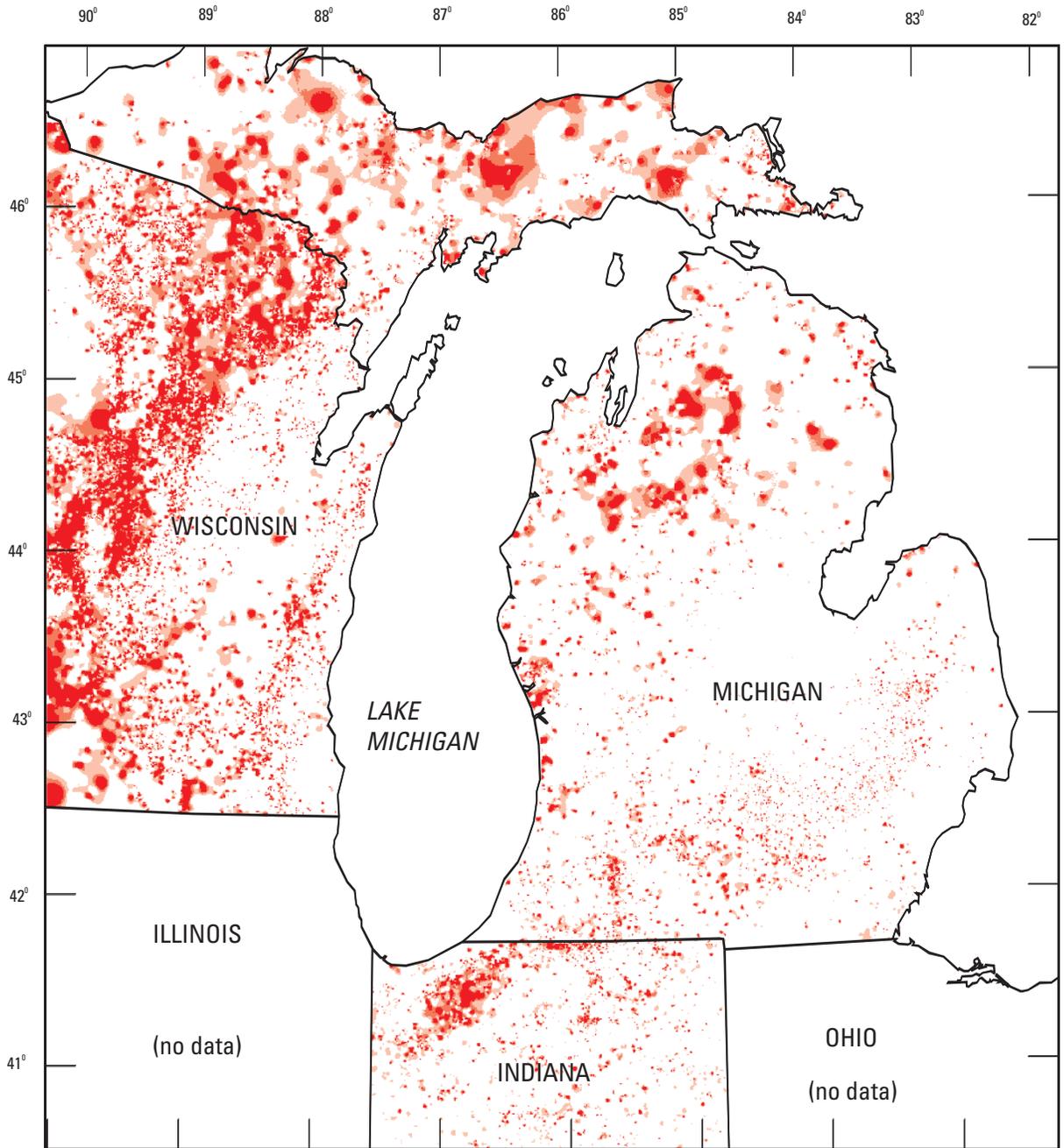
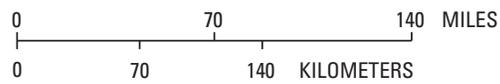


Figure 8. Distribution of equivalent horizontal hydraulic conductivity for layer 1 of the model based on assumed hydraulic conductivity of 1 foot per day for nonaquifer material and 100 feet per day for aquifer material.



Base from U.S. Geological Survey digital data
 1:100,000 1983. Universal Transverse Mercator
 projection, Zone 16, Standard parallel 0° (Equator),
 Central meridian 87° W, North American Datum 1983



EXPLANATION

Vertical hydraulic conductivity,
 in feet per day

	Less than or equal to 0.01		Greater than 0.1 and less than or equal to 1.0
	Greater than 0.01 and less than or equal to 0.1		Greater than 1.0 and less than or equal to 10.0

Figure 9. Distribution of equivalent vertical hydraulic conductivity for layer 1 of the model based on assumed vertical hydraulic conductivity of 0.001 foot per day for nonaquifer material and 10 feet per day for aquifer material.

Patterns can be seen in the distributions of hydraulic conductivity. Patterns of high and low K over large areas are evident within each state, such as the high values in northwestern Indiana and the low values southwest of Saginaw Bay in Michigan (fig. 8). Patterns also continue across state boundaries, such as the southwest-to-northeast set of linear features in northern Indiana and southeastern Michigan (fig. 8). High values of K_z appear in the same areas in figure 9 as seen in figure 8, but the extent of high values is more limited in figure 9 than in figure 8. The limited extent of high values is related to the calculation of K_z . An area can have mostly aquifer material, but a thin layer of nonaquifer material within the aquifer can significantly lower the overall value of K_z calculated for the model layer.

About 270,000 point values of K_{sc} also were mapped and observed for patterns. Using inverse distance weighting with an exponent of 2, distributions of K_{sc} were generated from the specific-capacity data for individual states, and the state distributions were merged into a single distribution for the three-state area (fig. 10). Figures 8 and 10 both show values for horizontal hydraulic conductivity, but they differ in the volume of material that they represent. Figure 8 is the equivalent hydraulic conductivity for model layer 1, which consists of aquifer and nonaquifer material. Figure 10 provides an estimate of the hydraulic conductivity of aquifer material within layer 1. In figure 8, horizontal hydraulic conductivity of aquifer material is assumed to be 100 ft/d, whereas in figure 10 the horizontal hydraulic conductivity of aquifer material is calculated from specific capacity data. Extremely high values of greater than 10,000 ft/d for K_{sc} were calculated for a few sites, but the distribution shown does not exceed 1,000 ft/d because that number is considered a realistic maximum value. Median K_{sc} values decrease from Indiana (264 ft/d) to Michigan (89 ft/d) to Wisconsin (48 ft/d). The decline could be related to past glacial processes. Kevin Kincare (U.S. Geological Survey, oral

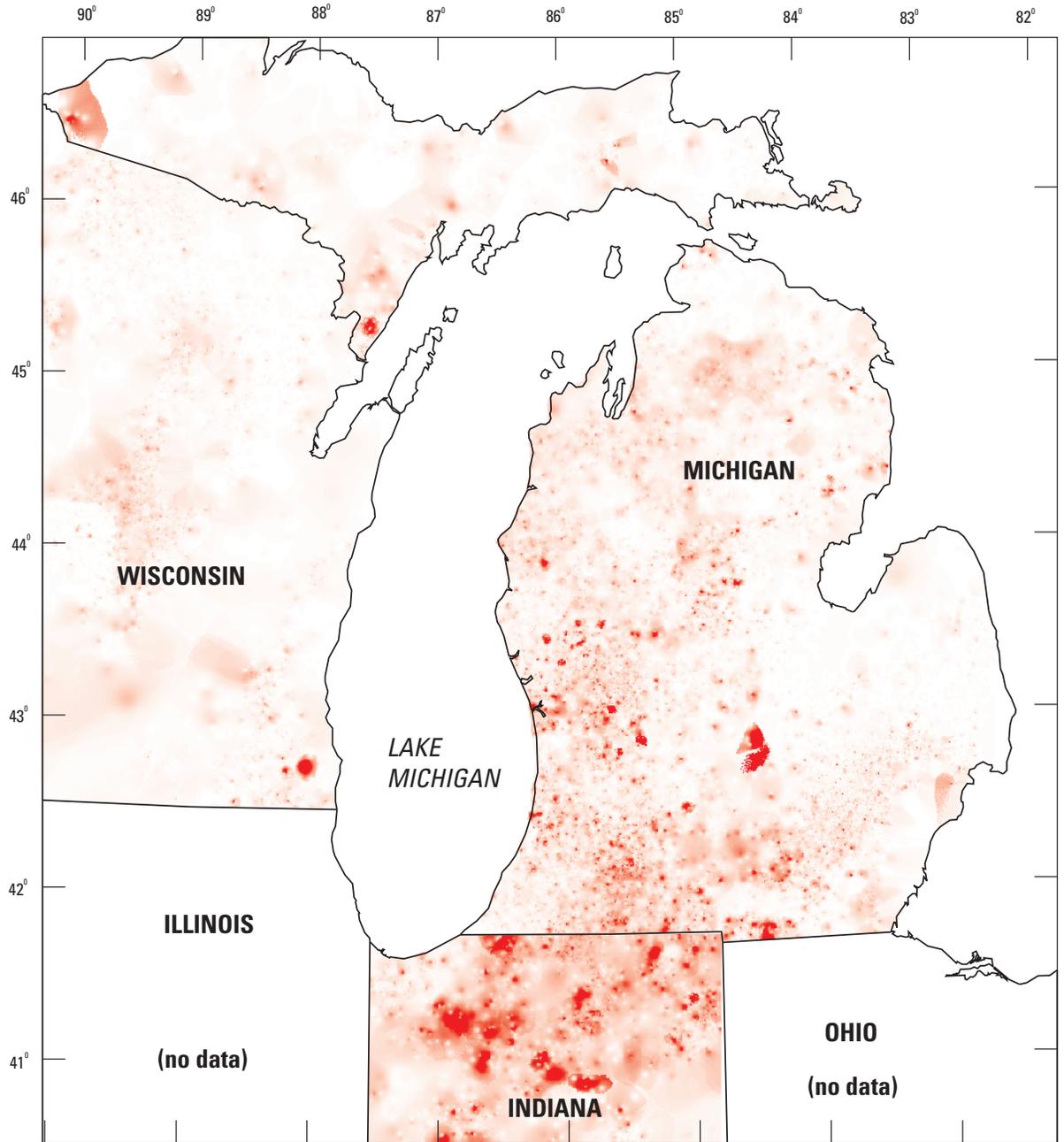
commun., 2006) described the glaciers as melting in place in Michigan. Such melting may have allowed nonaquifer material to settle in with the aquifer material and resulted in low horizontal hydraulic conductivities for aquifer material in Michigan. Meltwater may have been more active in Indiana, generating high energy flows and resulting in cleaner aquifer material in Indiana. Some correlation can be seen between areas of high horizontal hydraulic conductivity and current drainage ways in Indiana (fig. 10). More hydrogeologic investigation is required to confirm this general observation.

A relation was found between K_{sc} and sediment texture. The point data for K_{sc} were grouped by the surface geology texture class (defined by Soller and Packard, 1998) within which the well resides, and the median value was calculated for each texture class, layer, and state (table 4). Median values are used to avoid the influence of extremely high values that can noticeably influence the computation of average values. Median values also help to eliminate the influence of extremely high or low values of horizontal hydraulic conductivity caused by errors in data collection techniques. Again, K_{sc} values in Indiana are greater than those in the other two states. In addition, the percentage difference between the values associated with the different textures in Indiana and Michigan are greater than those in Wisconsin for layer 1. For example, in model layer 1, the difference between median K_{sc} from areas of till to areas of coarse grained sediments is 104 percent for Indiana, 25 percent for Michigan, and 14 percent for Wisconsin. Higher values are found in areas where surface geology textures consist of coarse grained sediments. Coarse grained sediments are associated with a high energy environment present during their deposition. Similar values were found for the three textures in Wisconsin. Except for Indiana, the values for the three textures decrease with depth, reflecting possibly the greater compression and decrease in pore size of the sediments with depth.

Table 4. Median horizontal hydraulic conductivity for different sediment texture classes calculated from specific-capacity data from Indiana, Michigan, and Wisconsin for model layers 1, 2, and 3 of a ground-water-flow model of the Lake Michigan Basin.

[ft/d, feet per day; —, no data]

Glacial texture at the surface	Horizontal hydraulic conductivity in Indiana (ft/d)	Horizontal hydraulic conductivity in Michigan (ft/d)	Horizontal hydraulic conductivity in Wisconsin (ft/d)
Model layer 1			
Coarse	200	35	25
Till	98	28	22
Fine grained stratified	74	12	29
Model layer 2			
Coarse	240	18	12
Till	130	16	16
Fine grained stratified	56	6.6	22
Model layer 3			
Coarse	—	7.7	0.73
Till	—	6.3	7.8
Fine grained stratified	—	6.3	—



Base from U.S. Geological Survey digital data
 1:100,000 1983. Universal Transverse Mercator
 projection, Zone 16, Standard parallel 0°(Equator),
 Central meridian 87° W, North American Datum 1983

0 70 140 MILES
 0 70 140 KILOMETERS

EXPLANATION

Horizontal hydraulic conductivity,
 in feet per day

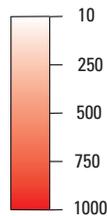


Figure 10. Distribution of horizontal hydraulic conductivity calculated from well-log derived specific-capacity data for aquifers in layer 1 of the ground-water-flow model.

Comparison of Hydrogeologic Products Derived from Well-Log Data with Other Hydrogeologic Products

The map of equivalent horizontal hydraulic conductivity of model layer 1, hereinafter termed K map, prepared by this study (fig. 8) is compared to three hydrogeologic products: a surface geology map of the study area, a ground-water availability map for Indiana, and a map that shows flow-duration ratios (the 20-percent flow duration divided by the 90-percent flow duration) for northern Indiana. The comparisons indicate similarities between the map of K and the three published maps, as well as potential advantages that accrue to the K map. The similarities indicate additional uses of the map of K in interpreting ground- and surface-water hydrology.

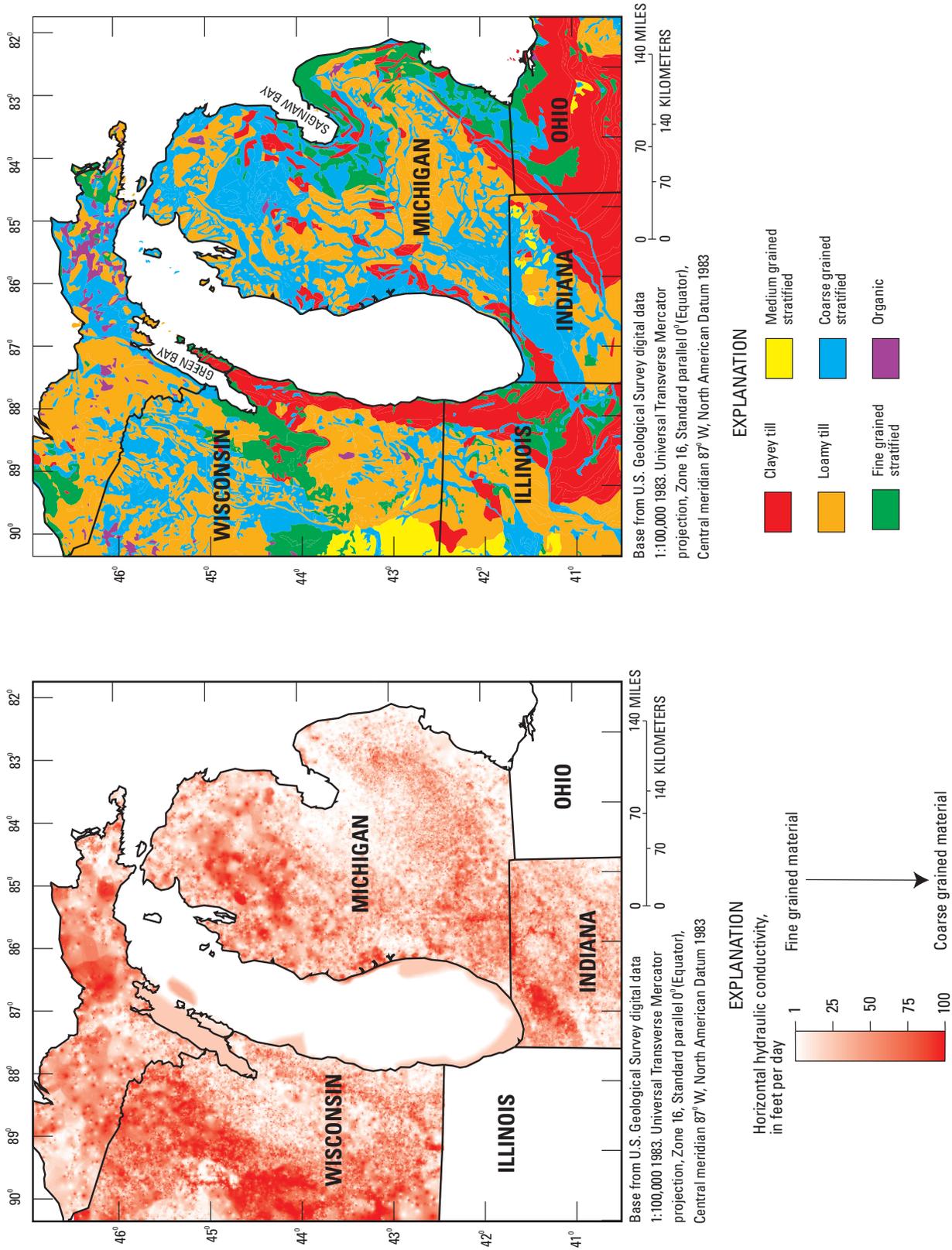
The patterns of fine and coarse grained sediments shown on the map of K (fig. 8) and the occurrences of those sediments on the surface geology map are similar (fig. 11). For example, a large area of coarse grained sediments is present in northwestern Indiana on both maps in fig. 11. In addition, a fine grained area is shown southwest of Green Bay in both parts of the figure. The K map shows additional, three-dimensional information that is not evident on the surface geology map. The surface geology map depicts a patchy deposit of coarse grained sediments in north-central Wisconsin, but the K map indicates that the coarse grained sediments are more aerially extensive and continue beneath the patchy loamy till. The surface geology map also shows surficial coarse and fine grained stratified sediments that rim and extend southwest of Saginaw Bay, but the K map suggests that fine grained sediments dominate the upper part of glacial sediments in the same area.

In figure 12, the K map for northern Indiana is presented alongside a ground-water availability map that shows potential yields to properly constructed large-diameter wells penetrating the full thickness of aquifers (Indiana Department of Natural Resources, Division of Water, 1980, p. 33). The ground-water availability map is based on maximum well yields reported in an area. The blue and green areas along the top of the ground-water availability map correspond to areas of high yield and also correspond favorably to the darkest area of red on the northern border of the K map (fig. 12). The dark red areas represent high K values, but they also represent a high percentage of sand and gravel (fig. 12), which could be the reason for the high yields. The correspondence suggests that the K map could possibly be used in combination with other indicators of ground-water availability. The correspondence between high ground-water availability and darker red areas is not exact, possibly because well-yield data may not have been available to define an area on the ground-water availability map as high yield. A correspondence between areas with lower values of K

and low well yields also can be seen in figure 12. The low-yielding brown and yellow areas in the southwestern corner of the ground-water availability map correspond to near-white areas of the K map.

The information in figure 12 or figure 8 (K distribution of the entire study area) could be combined with the specific capacity-based horizontal hydraulic conductivity information in figure 10 for additional ground-water availability information. That is, for two areas of equally high K values (equally high percentages of sand and gravel), the area indicated in figure 10 as having higher horizontal hydraulic conductivity would probably yield greater amounts of water to wells.

The last qualitative comparison between well-log based products and published maps will demonstrate a correspondence between the K distribution and low-flow statistics. Figure 13 shows the distribution of K overlain by areas of equal flow duration ratio for northern Indiana (Arihood and Glatfelter, 1991, Plate 1). Flow-duration ratio was defined as the 20-percent flow duration at a streamflow-gaging site divided by the 90-percent flow duration. Flow duration ratio was mapped for Indiana by calculating the ratio for all the qualifying streamflow gaging stations in Indiana, then transferring the ratios to the watershed areas associated with the gages. The flow-duration ratio data is used in a regression equation (Arihood and Glatfelter, 1991, p. 9) to estimate the 7-day, 10-year low flow at any point along a stream not affected by flow regulation. Low values of flow-duration ratio are related to areas predominately containing sand and gravel (Arihood and Glatfelter, 1991, p. 8). As can be seen in figure 13, the red areas of the K map along the northern boundary of the state are those with high percentages of sand and gravel, and they correspond to areas with low values of flow-duration ratio. The large red area in northwestern Indiana is associated with the lowest flow-duration ratio value of 3. In addition, the near-white areas in the southeastern and southwestern corners of the K map correspond to areas on the flow-duration map where base flow is so limited that the 7-day 10-year low flow is sometimes zero. The near-white areas not only show where ground water may not be sufficiently available, but, in contrast, show where surface-water supplies could be developed because they have a high percentage of surface runoff and low percentage of recharge to ground water.



Geology of surficial deposits (Fullerton and others, 2003)

Figure 11. Distribution of equivalent horizontal hydraulic conductivity for layer 1 from this study and surface geology from Fullerton and others (2003).

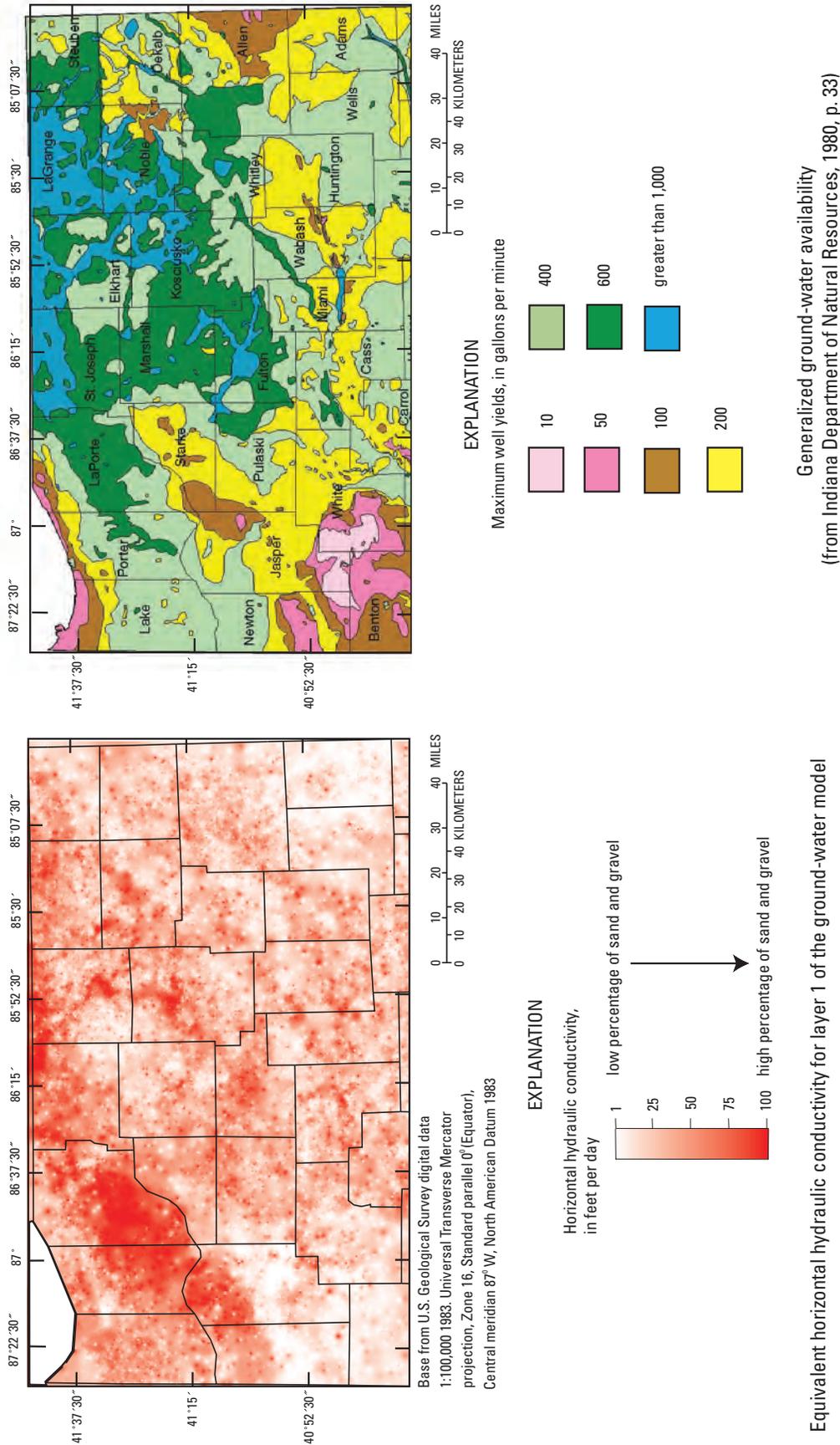


Figure 12. Distribution of equivalent horizontal hydraulic conductivity for layer 1 from this study and generalized ground-water availability for a part of northern Indiana (Indiana Department of Natural Resources, 1980).

Low values of flow-duration ratio also are related to watersheds with high base flows, which, in turn, are related to high rates of ground-water recharge. Because of its correspondence to flow-duration ratio, the K map could also possibly be used as an indicator of ground-water recharge. Ground-water recharge is often determined by a regression equation that uses surface geology as an independent variable because surface-

geologic information is readily available. However, Neff and others (2005, p. 7) explained the improvement in using the lithology information from well logs to better represent geology beneath the surface. The K distribution in figure 13 represents geologic composition beneath the surface, and its use as a basin characteristic may improve the prediction of recharge. Other possibilities may exist to improve the results

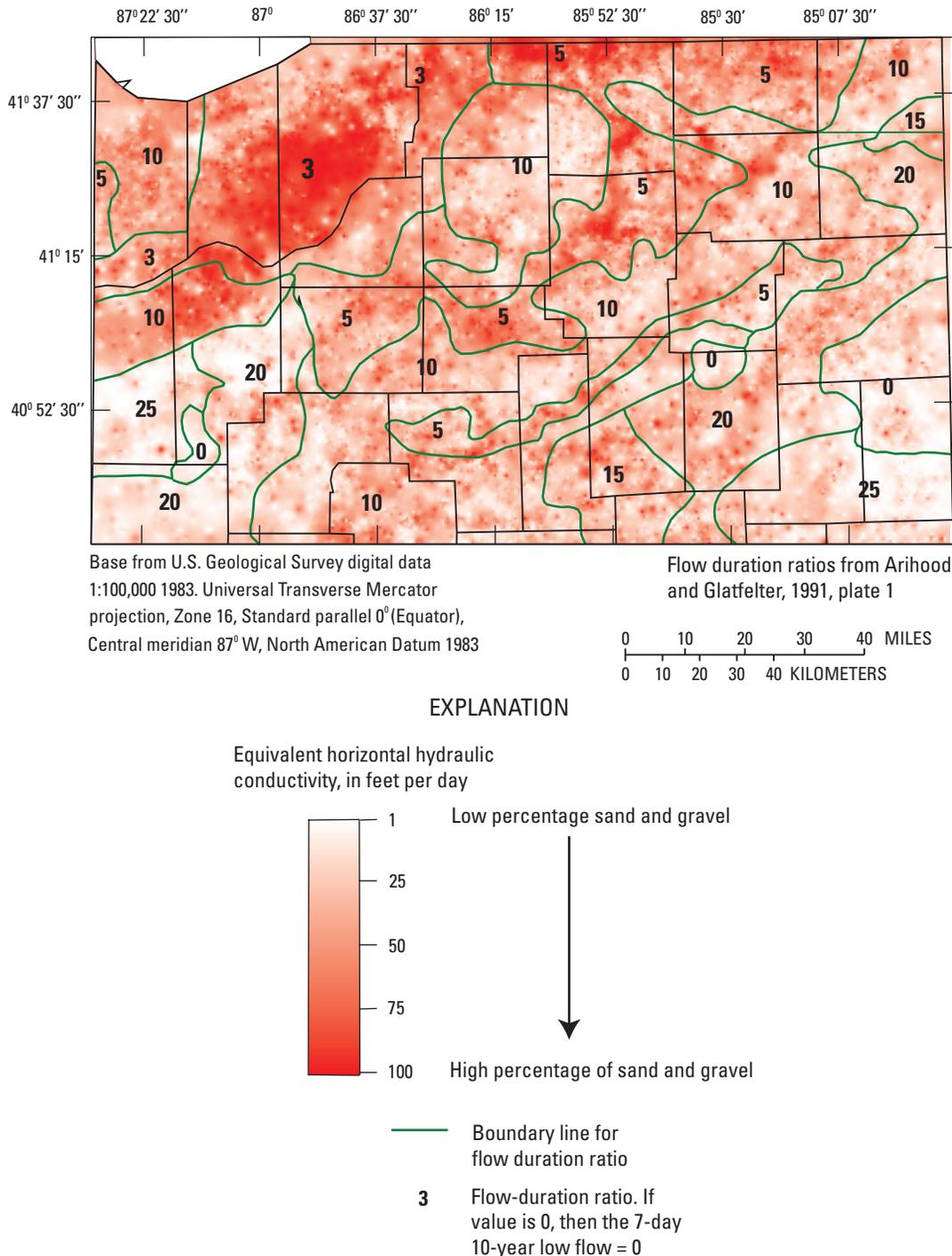


Figure 13. Distribution of equivalent horizontal hydraulic conductivity distribution for layer 1 of the model and flow-duration ratios for a part of northern Indiana.

of regression estimation of other hydrologic parameters whenever surface geology is used as an independent variable. For example, flow quantity, such as 10-year low flow, or water-quality characteristics could be estimated using K values.

Well logs may provide additional sources of information useful to water availability and use. Coon and Sheets (2006) estimated the volume of ground water available in the Great Lakes Basin. To estimate the volume of ground water in glacial sediments, the characteristics of surficial aquifers (thickness, extent, and composition) from the Great Lakes study area were used (Coon and Sheets, 2006, p. 12). If a well-log data base were available for a future study area, then the K distribution could be constructed for the area. The volume of aquifer material within the distribution could be calculated by available software, and then the volume of water available in the aquifer material could be calculated using an assumed storage coefficient. The volume of water calculation would be based on more detailed data than was done by Coon and Sheets (2006, p. 12). Water use is often recorded on well logs. Although water-use data from well logs were not used in this study, future studies might benefit from those data. Well-log records will not document all water use in a region, but they may provide a satisfactory sample of water use, or at least, an indication of water use that could supplement other data. The usefulness of water-use data from well logs would be based on the quality and completeness of data from any given area.

Summary and Conclusions

In 2005, the U.S. Geological Survey established a program called the National Water Availability and Use. The Great Lakes Basin was chosen as a pilot study area where methods were to be developed to best evaluate water resources. Part of the pilot study involves constructing a ground-water model for the Lake Michigan part of the Basin. Most ground-water flow occurs in the glacial sediments; therefore, adequate representation by the model of the horizontal and vertical hydraulic conductivity of the glacial sediments is important to the accuracy of model simulations. About 458,000 well driller logs were available in computerized form to help characterize hydraulic conductivity of the glacial sediments. Computer programs were written to process well-log information into representations of the equivalent horizontal and vertical hydraulic conductivity distribution for the glacial layers of the model. In addition, the specific-capacity data was converted into horizontal hydraulic conductivity data that could be used during ground-water modeling as a guide for adjustment of equivalent horizontal hydraulic conductivity distributions, and well-log water levels were processed to assemble a water-level calibration data set.

A lithology-based distribution of equivalent hydraulic conductivity was created by first automatically converting well-log lithology descriptions into aquifer or nonaquifer categories, and then calculating equivalent horizontal and vertical

hydraulic conductivities (K and K_z , respectively) for predetermined thicknesses of glacial sediments. The equivalent K was based on an assumed value of 100 ft/d for aquifer material and 1 ft/d for nonaquifer material, whereas the equivalent K_z was based on an assumed value of 10 ft/d for aquifer material and 0.001 ft/d for nonaquifer material. These values were assumed for convenience to determine a relative contrast between aquifer and nonaquifer sediments. The point values of K and K_z from wells that penetrate at least 50 percent of a model layer were interpolated into a continuous distribution. Each K distribution was based on an inverse distance weighting equation that used an exponent of 2, which is commonly used. The K_z distribution used an exponent of 4, which helped to represent the abrupt change in K_z that often occurs between aquifer and nonaquifer materials.

The assumed values of hydraulic conductivity for aquifer sediments needed to be adjusted to actual values that occur in the study area. The specific-capacity data (discharge, draw-down, and time data) from driller well logs were used with a modified version of the Theis equation that corrects for partial penetration of the aquifer to calculate horizontal hydraulic conductivities (K_{sc}). The K_{sc} values were used as a guide during ground-water modeling to adjust the assumed value of 100 ft/d for aquifer deposits to actual values used in the modeling.

A method was developed to obtain potentially more reliable water levels from well logs for comparison with simulated water levels in a model layer during model calibration. The method uses kriging of the water-level data to determine a water-level surface, and then uses the difference between the kriged surface and individual water levels to identify outlier water levels.

Examination of the well-log data in map form revealed that the data was not only useful for model input, but useful in understanding the hydrogeology of a multistate area. The distribution of K and K_z provide a three-dimensional view of aquifer and confining systems. The distribution of specific-capacity based hydraulic conductivity reveals an aerial difference from state to state and a relation to sediment texture. Patterns in high and low hydraulic conductivity over large areas are evident within a state, such as the high hydraulic conductivity in northwestern Indiana and the low hydraulic conductivity southwest of Saginaw Bay in Michigan. Patterns also continue across state boundaries, such as the southwest-to-northeast set of linear features in northern Indiana and southeastern Michigan. Median K_{sc} are larger in Indiana (264 ft/d) and relatively smaller in Michigan (89 ft/d) and Wisconsin (48 ft/d). This difference could be related to past glacial processes. Another pattern in K_{sc} was observed with sediment texture. Aquifers within sediment textures deposited under a higher energy environment, for example, outwash textures, had higher values of K_{sc} than those deposited under a lower energy environment.

The products derived from well-log data appear similar to published maps and possibly convey more information than available in the published maps. The patterns of fine and coarse grained sediments shown on the distribution of K for model layer 1 (K map) and the surface geology map are similar. In areas where the two maps differ, the difference may be related to the content of geologic deposits beneath the surface. In an area where the surface geology map shows a mixture of fine and coarse grained sediments, the K map shows that fine grained sediments dominate the glacial sediments. Areas of high and low well yield on a ground-water availability map for Indiana correspond to areas of high and low percentages of sand and gravel on the K map for the same area. The correspondence suggests that a K map could be a supporting indicator of ground-water availability. A flow-duration ratio map (20-percent flow duration divided by the 90-percent duration) indirectly maps the amount of sand and gravel in glacial deposits in Indiana. Areas that have low values of the flow-duration ratio and high percentages of sand and gravel correspond to areas of high sand and gravel percentage on the K map.

The similarities between the K map and hydrogeologic maps suggest uses of the map data beyond the generation of input data to a ground-water model that are beneficial to the understanding of ground-water availability. Areas of high K (high percentages of sand and gravel) are related to high rates of ground-water recharge and ground-water availability. Ground-water recharge is often determined by a regression equation that uses surface geology as an independent variable because surface-geology information is readily available. But the entire thickness of geologic materials beneath the surface determine recharge rates, and a K map based on well logs penetrating 100 ft or more of the glacial sediments better represents overall geologic composition. Data from a K map could be used as a basin characteristic to improve the prediction of recharge or any low or high flow statistic, such as the 7-day, 10-year low flow or the 10-year flood.

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