recharge stable ground-water flow systems in these areas may be forced to discharge locally by means of drainage tile or shallow, transient ground-water flow systems. The Maumee River is also incised only a few feet, which may prevent it from intercepting flow from some stable ground-water flow systems. Poorly permeable glaciolacustrine sediments may also impede discharge from the carbonate-rock aquifer to the Maumee River. In general, glacial deposits in the Maumee River Basin are thin, absent, or poorly permeable. Toth (1963) notes that low ground-water discharge to streams within a drainage basin can be due to other areas of ground-water discharge within the basin. Before ditching in the early 1900's, much of the Maumee River Basin was swampland. Norris (1974) notes that the historic Black Swamp in this area resulted from a combination of poor drainage and ground-water discharge from regional ground-water flow into what was a relatively stagnant area of surface water and ground water.

The Sandusky River Basin is also associated with a fairly low percentage of sustained ground-water discharge to streams. Much of the ground water that flows through this drainage basin is likely to discharge to Lake Erie rather than to the streams within the basin.

REGIONAL GROUND-WATER FLOW

General concepts regarding flow within an aquifer system are reviewed herein to facilitate discussions of the conceptual and numerical models of the Midwestern Basins and Arches aquifer system. An aquifer system can comprise local, intermediate, and regional ground-water flow systems (fig. 18). In a local system of ground-water flow, recharge and discharge areas are adjacent to each other. In an intermediate ground-water flow system, recharge and discharge areas are separated by one or more topographic highs and lows. In a regional ground-water flow system, recharge areas are along ground-water divides, and discharge areas lie at the bottom of major drainage basins. Not all types of ground-water flow are present in every aquifer system (Toth, 1963).

The greatest amount of ground-water flow in an aquifer system is commonly in local flow systems. Ground-water levels and flow in local flow systems are the most affected by seasonal variations in recharge because recharge areas of these relatively shallow, transient ground-water flow systems make up the greatest part of the surface of a drainage basin (Toth, 1963). Regional flow systems are less transient than local and intermediate flow systems. For the remainder of this report, the term "regional flow systems" is used to describe flow systems that are minimally affected by seasonal variations in ground-water recharge and are capable of providing a fairly constant source of discharge to streams (sustained ground-water discharge). Although this use of the term "regional flow systems" refers, in large part, to intermediate and regional flow systems as defined by Toth (1963), some local-scale flow also may be included.

CONCEPTUAL MODEL

A conceptual model of an aquifer system is a simplified, qualitative description of the physical system. A conceptual model may include a description of the aquifers and confining units that make up the aquifer system, boundary conditions, flow regimes, sources and sinks of water, and general directions of ground-water flow. The conceptual model of the Midwestern Basins and Arches aquifer system presented herein is based on information presented in the "Geohydrology" section of this report.

The Midwestern Basins and Arches aquifer system is in a state of dynamic equilibrium with respect to hydrologic variations over the long-term period. As a result, the aquifer system may be adequately described on the basis of long-term average water levels and ground-water discharges. In addition, annual ground-water-level fluctuations are quite small (less than 10 ft) compared to the thickness of the aquifer system (hundreds of feet).

The water table within the aquifer system generally is within alluvium or glacial deposits; glacial aquifers can supply large yields of ground water in only a limited number of places. The glacial deposits are underlain by an areally extensive carbonate-rock aquifer, which is semiconfined or locally confined by the glacial deposits across most of the study area. The carbonate-rock aquifer is confined by shale along the margins of the aquifer system. Very little water is produced from the carbonate-rock aquifer under the shales because shallower freshwater sources are generally available.

Spatial patterns in hydraulic characteristics of the glacial aquifers or the carbonate-rock aquifer are not readily apparent from the available transmissivity data (figs. 9 and 10); however, some of the highest transmissivities in the glacial aquifers are associated with outwash deposits along the principal streams (figs. 5 and 9). Despite the spatial variability of hydraulic characteristics within the carbonate-rock aquifer, the aquifer functions as a single hydrologic unit at a regional scale (Arihood, 1994).

The upper boundary of the aquifer system coincides with the water table. The lower boundary generally coincides with the contact between the carbonate-rock aquifer and interbedded shales and limestones of Ordovician age where they underlie the aquifer. Where the carbonate-rock aquifer is hundreds of feet thick, the lower boundary of the aquifer system may be within the carbonate rocks. Lateral boundaries of the carbonate-rock aquifer include the limit of potable water (waters that contain dissolved-solids concentrations less than 10,000 mg/L; U.S. Environmental Protection Agency, 1984)) to the north, east, and west (fig. 34), Lake Erie to the north-

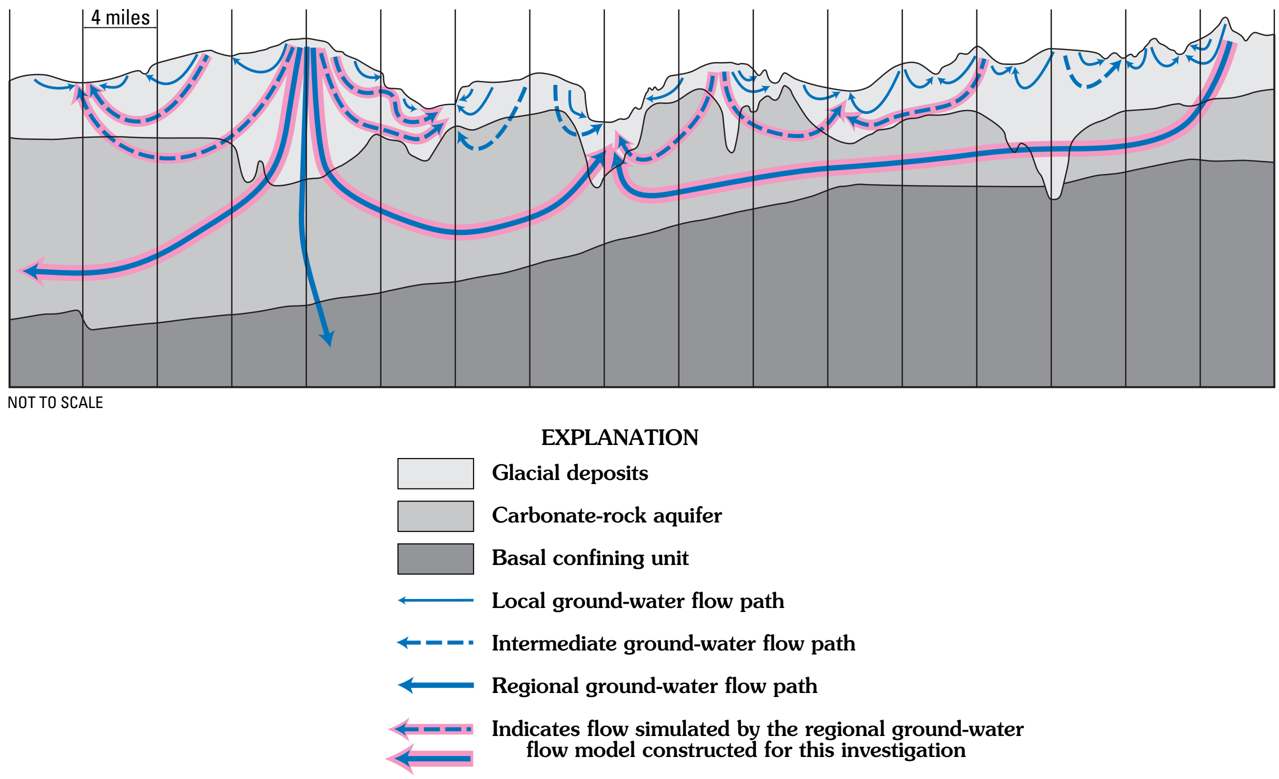


FIGURE 18.—Diagrammatic conceptual model of the Midwestern Basins and Arches aquifer system showing flow paths associated with local, intermediate, and regional flow systems (modified from Toth, 1963) and flow systems simulated by the regional ground-water flow model.

east, and the Ohio River and the upper weathered zone water-bearing unit to the south.

Several types of ground-water flow systems are present within the Midwestern Basins and Arches aquifer system, as evidenced by base-flow duration curves constructed for selected streamflow-gaging stations within the study area (discussed previously in the “Discharge” section). The amount of ground-water discharge to streams from fairly stable ground-water flow systems (regional, as defined in this report) relative to the amount of discharge from all scales of ground-water flow systems (local, intermediate, and regional) within the aquifer system (fig. 17) indicate that local flow systems dominate ground-water discharge to streams within the Midwestern Basins and Arches Region. Unless a large amount of ground water flows across lateral boundaries of the aquifer system or large volumes of ground water discharges to places other than streams, figures 14 and 17 can be used to infer that local flow systems dominate ground-water flow in the aquifer system.

The amount of recharge to regional flow systems (as defined in this report) within the Midwestern Basins and Arches aquifer system is approximately equal to mean sustained ground-water discharge to streams, ditches, lakes, and wetlands and losses from the relatively stable parts of the aquifer system by means of evapotranspiration and pumping. Recharge to the deepest parts of the aquifer system occurs predominantly in the upland areas.

The amount of ground water available to sustain streams during the driest periods (discharge from regional flow systems) within the study area is related to a number of factors. These factors include the availability of recharge, the geology and hydraulic gradients within the aquifer system, the position of the streams within the drainage basins and relative to the aquifer system in general, the relative incisement of the streams, and the presence of other places of discharge within the drainage basins.

Ground water flows from recharge areas, which are associated with high ground-water levels, to discharge areas, which are associated with low ground-water levels. General directions of regional flow in the aquifer system are away from several potentiometric highs toward the principal streams and Lake Erie. Most active flow of freshwater (less than 10,000 mg/L dissolved solids) in the carbonate-rock aquifer within the aquifer system is confined to the subcrop area of the aquifer.

NUMERICAL MODEL

A numerical model was constructed to test and improve upon the conceptual model of regional ground-water flow within the Midwestern Basins and Arches aquifer system. Concepts that were tested include the assumption that the carbonate rocks are (are not) productive throughout their entire thickness and hypotheses about what contributes to the differences in the percentages of mean sustained ground-water discharge to streams across the study area. The numerical model was also used to investigate the absence of systematic increases in ground-water ages along general directions of regional flow throughout most of the study area and the presence of isotopically distinct ground water beneath the Maumee River Basin (see “Geochemistry” section). Various aspects of the qualitative conceptual model were also quantified by use of the numerical model. Specifically, a regional ground-water budget was computed; rates and patterns of recharge and discharge to and from regional flow systems were mapped; and natural regional ground-water flow patterns and relative magnitudes of regional ground-water flow were determined.

In a numerical model, aquifers and confining units within an aquifer system are represented by cells organized into layers. Hydraulic heads and flow in each layer and the exchange of water between adjacent layers and across boundaries are computed simultaneously. These calculations are most commonly accomplished by use of a computer code that solves finite-difference or finite-element approximations of the partial differential equations (three-dimensional ground-water flow equation, boundary conditions, and initial conditions) that form the numerical model (Anderson and Woessner, 1992, p. 20).

The specific computer code used in this investigation is a three-dimensional modular model that solves a finite-difference approximation of the partial differential equations that describe ground-water flow (MODFLOW) (McDonald and Harbaugh, 1988). In the governing ground-water flow equation represented by this model, the density of water is assumed to be constant. Although the density of the water in the aquifer system may change within the carbonate-rock aquifer along the margins of the Michigan and Appalachian (structural) Basins, the effects of the density variations on ground-water flow within the modeled part of the aquifer system were assumed to be small enough that a variable-density flow model was considered unnecessary. This is a reasonable assumption because most of the aquifer system that was modeled is miles from these margins, stresses on the aquifer system do not affect the lateral limits of freshwater, and any affect density variations may have on model estimates of hydraulic conductivity or simulated hydraulic heads are likely to be within the confidence limits of the estimated hydraulic conductivity or the error associated with the hydraulic-head observations.

MODEL DESIGN

The numerical model built as part of this investigation was designed to simulate steady-state regional flow systems within the aquifer system. These are flow systems that are minimally affected by seasonal variations in ground-water recharge from precipitation and are capable of sustaining discharge to the streams during the driest periods. On the basis of stream base-flow estimates, much less than 50 percent of ground-water flow in the aquifer system is associated with these regional flow systems (figs. 14 and 17).

The model was not designed to simulate the preponderant local flow systems that are juxtaposed on the regional flow systems (figs. 15 and 18). Such local flow systems are too small and numerous to be adequately represented with a regional-scale model. Specifically, the model cell spacing chosen for this investigation (4 mi on a side) is not fine enough to capture the curvature of the water table associated with such local flow systems. This effect of scale results in the simulation of less cross-sectional area and hydraulic gradient than actually exists in the aquifer system; thus, the model cannot simulate the corresponding flow (fig. 19). Because local flow systems cannot be explicitly simulated in a model with a coarse regional-scale cell spacing, a boundary condition (discussed later in this section) was used to simulate the influence of the local flow systems on the deeper regional flow systems. The simulated regional ground-water budget, therefore, represents the budget of just the regional flow systems and not all flow systems within the aquifer system. Because water in local flow systems moves from recharge areas to the nearest stream valley, a map of the density of perennial streams that drain the modeled area is included to help illustrate the relative number of local flow systems that may be present within the aquifer system but are not included in the model (fig. 20).

Some flow systems may cross the basal confining unit of the Midwestern Basins and Arches aquifer system and become part of an even larger aquifer system. These flow systems also are not simulated by the regional ground-water flow model (fig. 18).

Unlike some areally extensive aquifer systems elsewhere in the Nation, the Midwestern Basins and Arches aquifer system is not subject to regional-scale pumping stresses. It was therefore unnecessary to construct a transient model that is capable of simulating changes in ground-water levels, discharge, and storage with time to represent regional ground-water flow (see “Conceptual Model” section). As a result, once a steady-state calibration was achieved, the numerical model was not calibrated to transient conditions, nor was it used to make predictions about the effects of future pumpage on regional ground-water flow. Known volumes of pumpage were also not included in the model because reported pumpage from the aquifer system is approximately 3 percent of total flow in the aquifer system. Although it is not known how

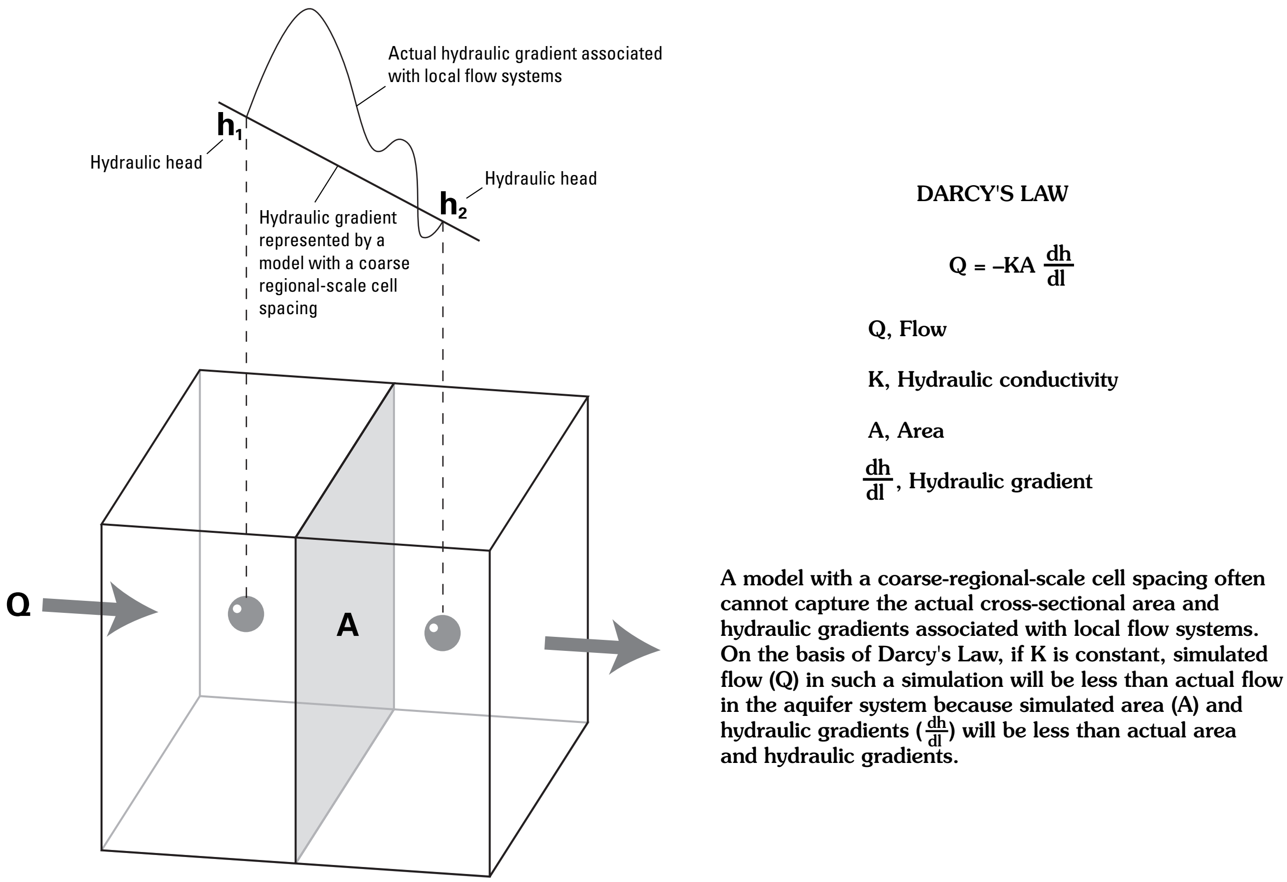


FIGURE 19.—Effect of model-cell spacing (model scale) on the amount of flow in an aquifer system that can be simulated with a numerical model.

much of this reported pumpage is associated with the predominant local flow systems as opposed to the simulated regional flow systems, most pumpage is assumed to be from the local flow systems. Even if all reported pumpage from the carbonate-rock aquifer is associated with regional flow systems, pumpage from the carbonate-rock aquifer (67 Mgal/d) would be less than 5 percent of flow in the regional flow systems based on estimates of mean sustained ground-water discharge to streams (fig. 17). Although some pumpage from glacial aquifers is also likely to be from regional flow systems, many of the largest water users produce water from outwash deposits along principal streams that are commonly associated with local flow systems, which are not simulated in the model.

DISCRETE GEOHYDROLOGIC FRAMEWORK

The numerical regional ground-water flow model is a quasi-three-dimensional two-layer model (table 1) structured within a 65-row by 61-column finite-difference grid (fig. 21). The upper model layer (layer 1) is used to simulate hydraulic heads and flow through glacial and other surficial deposits (fig. 22). Effective hydraulic conductivities were used to account for the heterogeneities with the glacial deposits. (An effective hydraulic conductivity is a hydraulic conductivity of an equivalent homogeneous formation for which the mean flux is equal to that prevailing in the heterogeneous formation (Indelman and Dagan, 1993).) The lower model layer (layer 2) is used to simulate hydraulic heads and flow through the bedrock. A single layer was considered sufficient to simulate flow in the carbonate-rock aquifer because vertical hydraulic-head gradients within the aquifer are small and the carbonate-rock aquifer functions as a single hydrologic unit at a regional scale. Where the carbonate-rock aquifer is absent in the south-central part of the study area, parameter values in model layer 2 were chosen to simulate hydraulic heads and flow in the upper weathered zone water-bearing unit.

Because relatively little horizontal flow occurs within the shale that separates the glacial deposits and the carbonate-rock aquifer along the margins of the modeled area, this upper confining unit is not represented as a separate layer in the model. Rather, a quasi-three-dimensional approach is used. With such an approach, only the resistance of the upper confining unit to vertical flow between the glacial deposits and the carbonate-rock aquifer is simulated (figs. 21*A* and 22).

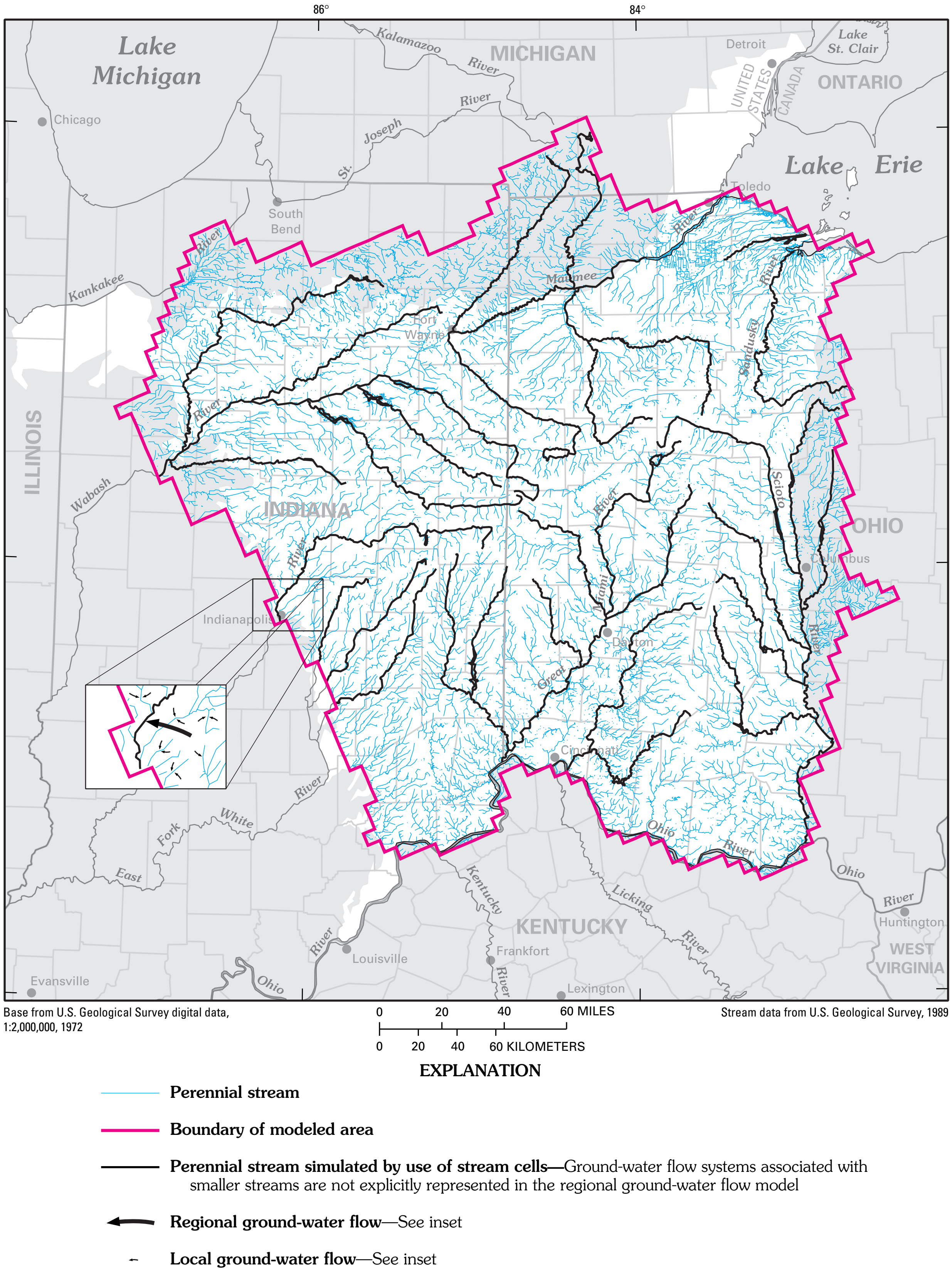
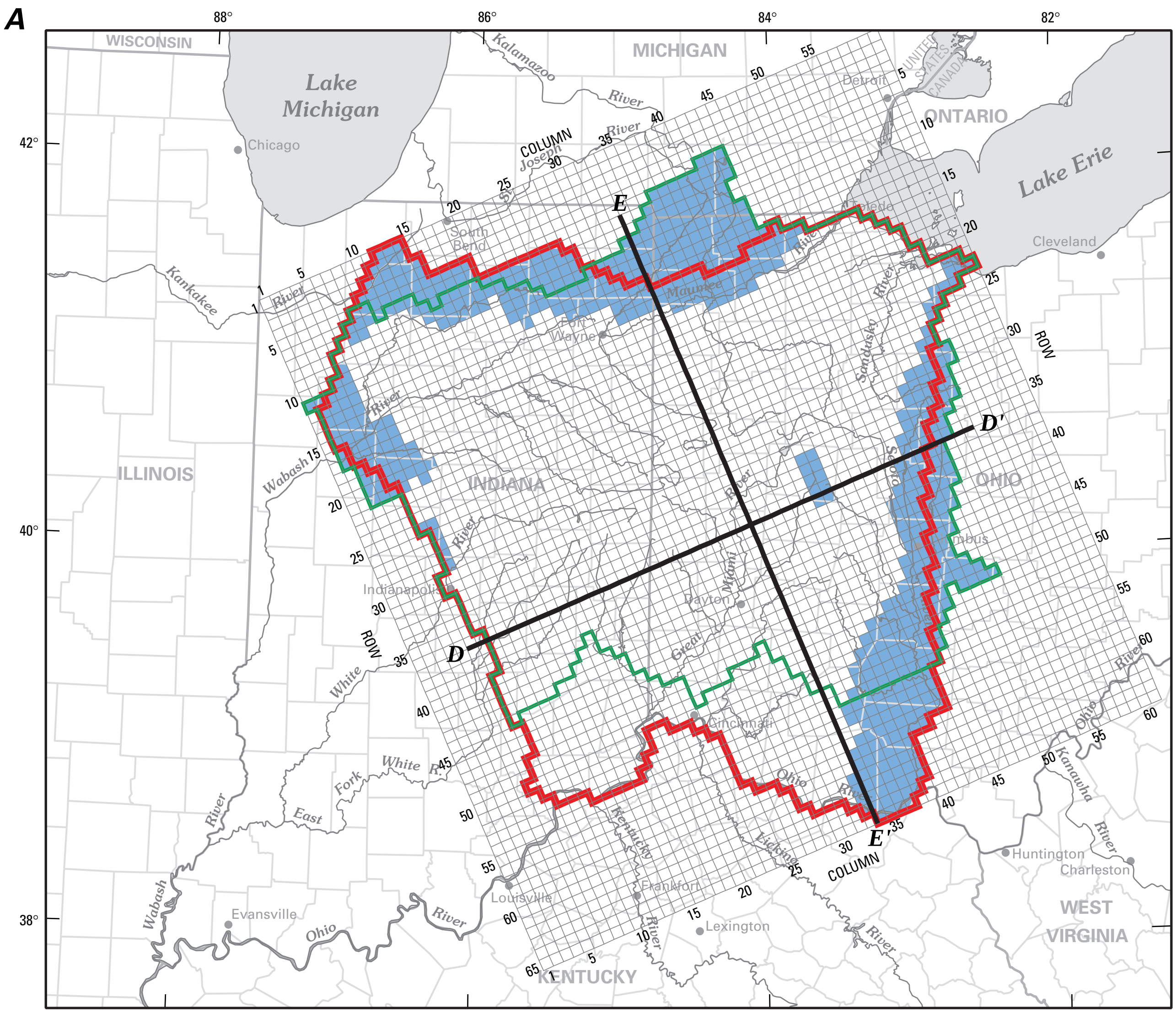


FIGURE 20.—Density of perennial streams that drain the modeled area at the scale of 1:100,000 and streams that are explicity represented in the regional ground-water flow model by use of stream cells.



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972

0 20 40 60 MILES

0 20 40 60 KILOMETERS

EXPLANATION

□ **Model cell**

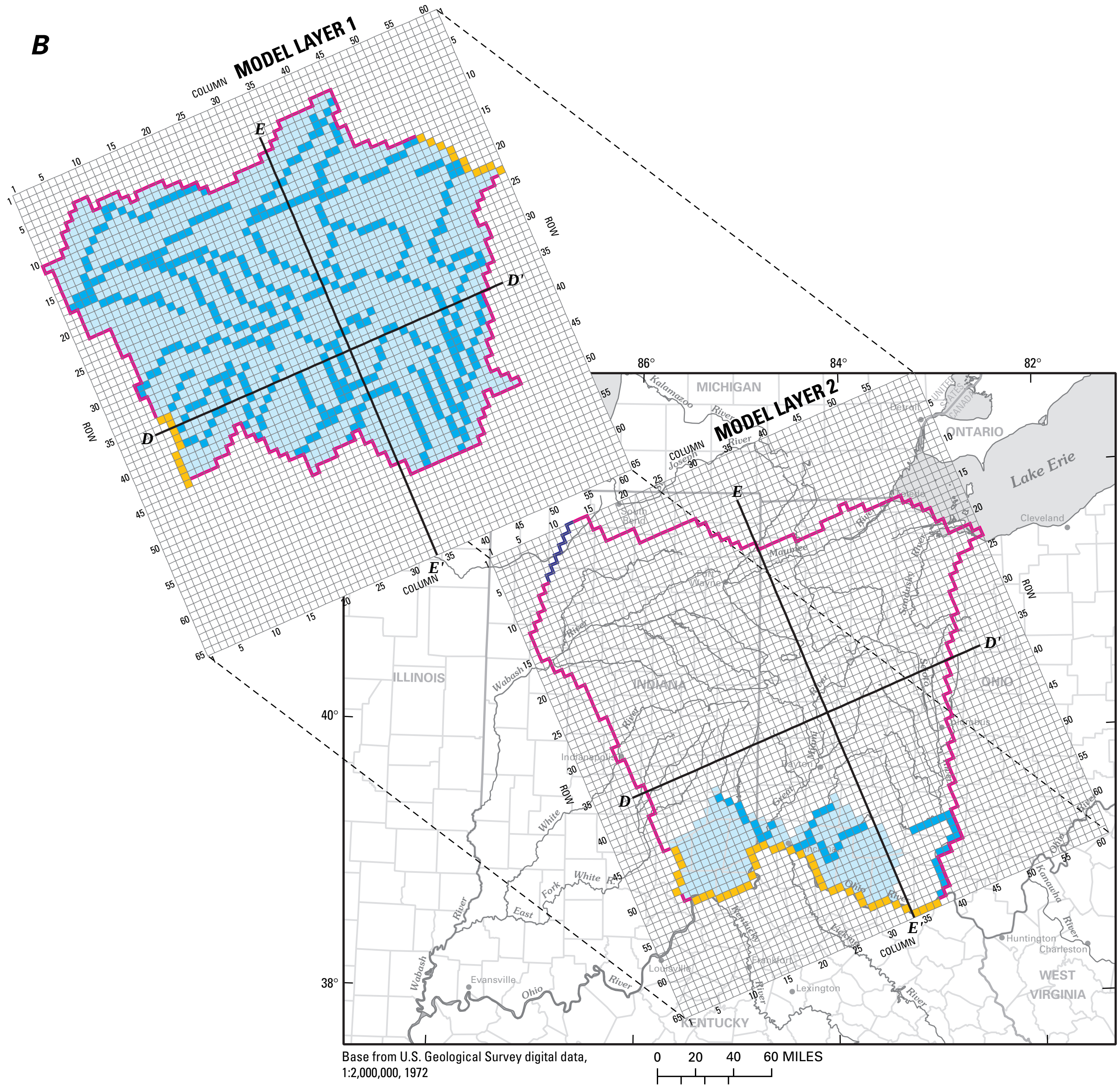
Area of upper confining unit—Not a separate model layer

Boundary of model layer 1

Boundary of model layer 2

D *D′* **Line of model section**

FIGURE 21.—Model layers used to simulate the Midwestern Basins and Arches aquifer system, and the location of model sections *D–D′* and *E–E′*: (*A*) areal extent.



EXPLANATION

- **Model cell**
- **Specified-head boundary**
- **Head-dependent flux boundary (stream cell)**—Used to simulate principal streams
- **General head-dependent flux boundary**—Used to simulate exchange of water at the regional water table. Recharge applied to model cell.
- **Specified-flux boundary**
- **No-flow boundary**
- D——D′ **Line of model section**

FIGURE 21. CONTINUED—Model layers used to simulate the Midwestern Basins and Arches aquifer system, and the location of model sections *D–D′* and *E–E′*: (*B*) boundaries.

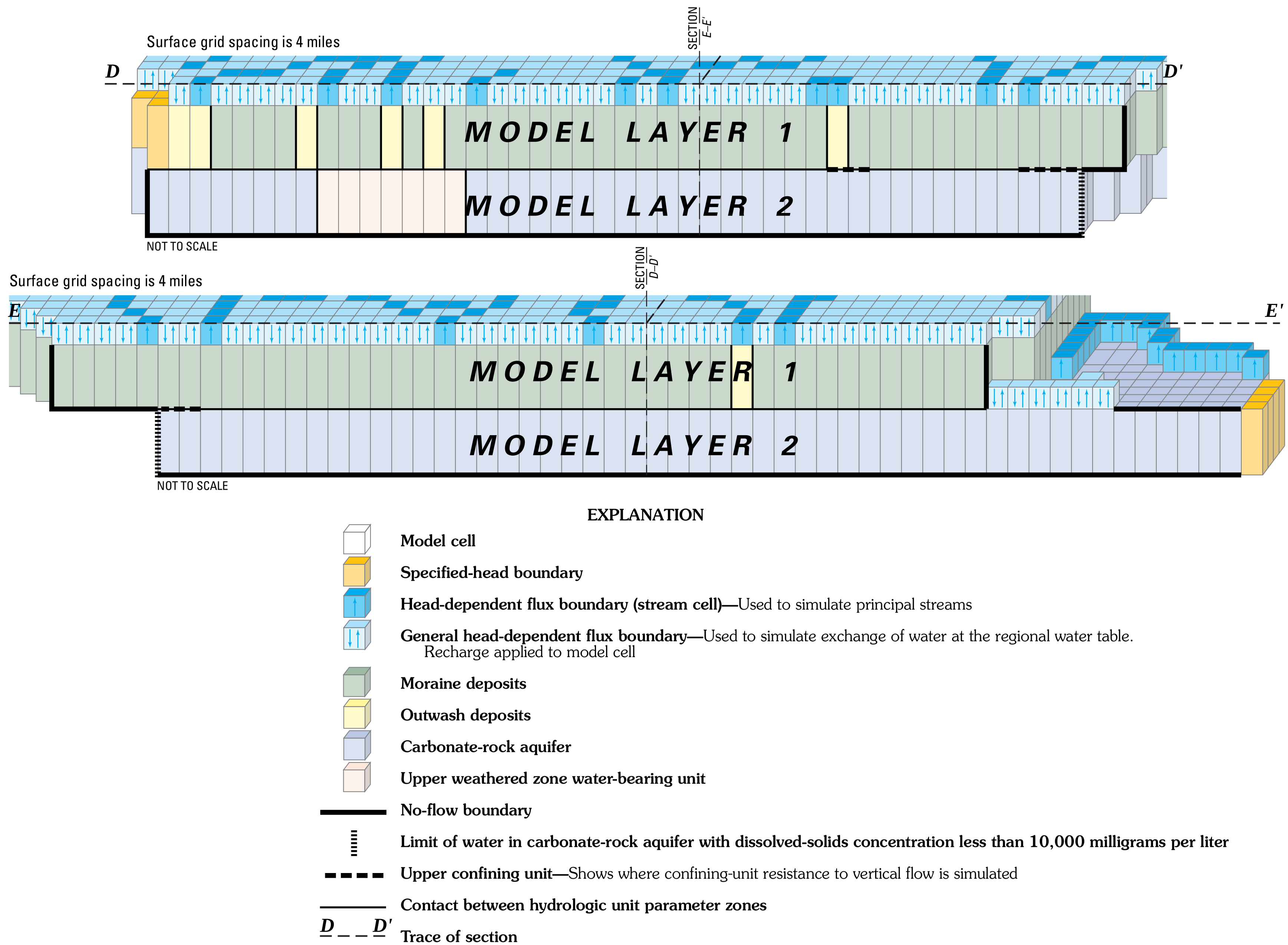


FIGURE 22.—Sectional diagrams showing the areal extent and boundaries of the model layers used to simulate the Midwestern Basins and Arches aquifer system (lines of section on figure 21).

Each model cell is 4 mi on a side. This cell spacing was chosen to allow the curvature in the regional potentiometric-surface map of the carbonate-rock aquifer in figure 12 to be represented by the model. It was assumed that flow in the fractured carbonate-rock aquifer behaves as flow in a porous medium at this simulation scale. It is noted in the “Hydraulic Characteristics” section of this report that this is assumption is probably valid.

The model grid is oriented so that it parallels the edge of the carbonate-rock subcrop along the margin of the Illinois (structural) Basin. This is because it was outside the scope of the investigation to study flow in the carbonate-rock aquifer within the Illinois Basin; a hydraulic boundary—which is easiest to simulate parallel to a finite-difference grid—rather than a physical boundary is simulated along this edge of the model. Anisotropy was not a consideration for model-grid orientation; it is assumed that no principal direction of horizontal anisotropy dominates regional flow within the aquifer system.

BOUNDARIES, SOURCES, AND SINKS

The following types of boundaries are included in the numerical model and are represented on figures 21*B* and 22. Model boundaries were set for a given model construction and were not automatically adjusted by the nonlinear regression.

No-flow boundaries

Part of the eastern and all of the northern and northwestern boundaries of model layer 1, which represents glacial aquifers and confining units, coincide with principal surface-water drainage divides that are assumed to coincide with ground-water divides in the glacial deposits. The northern part of the eastern boundary and most of the western boundary of model layer 1 coincide with regional flowlines in the glacial deposits (fig. 11); ground water flows parallel to and not across flowlines. The southern boundary of model layer 1 coincides with the limit of the Wisconsinan ice sheet (fig. 5). Glacial deposits are thin or absent, and few glacial aquifers are present south of this limit. Horizontal flow in glacial deposits beyond the limit of the Wisconsinan ice sheet is assumed to be negligible at the regional scale. In addition, the limit of the Wisconsinan ice sheet coincides with regional flowlines in the glacial deposits throughout much of the modeled area.

The eastern, northern, and part of the western boundaries of model layer 2, which represents the carbonate-rock aquifer or the upper weathered zone water-bearing unit, coincide with flowlines or the position of water in the carbonate-rock aquifer with a dissolved-solids concentration of 10,000 mg/L or greater (fig. 12; see also fig. 34). Water in the carbonate-rock aquifer is assumed to move slowly where it becomes saline (greater than 10,000 mg/L dissolved solids) and, for the purposes of this investigation, no water is assumed to flow across these boundaries. In addition, the position of the saline waters is assumed to have remained constant over the short period of time (tens of years) represented by the model calibration targets. Part of the northwestern boundary of model layer 2 coincides with a ground-water divide in the carbonate-rock aquifer.

Most of the water in the carbonate-rock aquifer along the northeastern model boundary is likely to flow upward into overlying glacial deposits and ultimately into Lake Erie. It is assumed that no water in the carbonate-rock aquifer flows laterally beyond the shore of the lake. Hanover (1994) notes that discharge from the carbonate-rock aquifer to Lake Erie is concentrated near the lakeshore. The assumption of no flow along this boundary was tested by constructing an alternative model with a specified-head boundary in model layer 2. The results of this alternative model are nearly identical to the results of the calibrated final model in which the no-flow boundary forces simulated ground water in the carbonate-rock aquifer to discharge through the overlying glacial deposits. This finding indicates that the no-flow boundary does not limit the amount of simulated regional ground-water flow that can leave the aquifer system at this point.

Ground water flows toward part of the western boundary of model layer 2 (fig. 12); however, some of this western boundary is simulated as a no-flow boundary (fig. 21*B*). This decision was made because a specified-head boundary in model layer 1 (fig. 21*B*) was considered sufficient to simulate the outward flux from the glacial deposits and the carbonate-rock aquifer; at the regional scale, the hydraulic heads in these hydrologic units are very similar along this boundary, owing to the absence of the upper confining unit. In addition, available ground-water-chemistry data (see “Tritium and Carbon Isotopes” section) indicate a notable increase in the age of water in the carbonate-rock aquifer just west of the model boundary where the aquifer dips beneath the upper confining unit. This combination of factors may indicate that much of the water in the carbonate-rock aquifer discharges through the overlying glacial deposits rather than moving downdip into the Illinois (structural) Basin.

The carbonate-rock aquifer is directly underlain by poorly permeable interbedded shales and limestones. As a result, a no-flow boundary condition is assigned to the lower model boundary where it coincides with the bottom of the carbonate-rock aquifer. Simulation of an alternative conceptual model, however (see “Model Discrimination” section), indicates that the lower limit of active freshwater flow may actually be within the carbonate rocks of Silurian and Devonian age where these rocks are hundreds of feet thick. The lower model boundary coincides with the top of the upper confining unit where water in the carbonate rock beneath the upper confining unit is saline.

It is assumed that flow in the upper weathered zone water-bearing unit is restricted to shallow depths (tens of feet) and that any exchange of water between this zone and deeper bedrock units is negligible. Therefore, the lower model boundary beneath the upper weathered zone water-bearing unit is simulated as a no-flow boundary.

Specified-head boundaries

The northeastern model boundary in model layer 1 coincides with the shore of Lake Erie. It is assumed that ground-water levels in the glacial deposits along this boundary are near lake level and that they can be represented by the average lake level for the long-term period. A specified-head boundary condition is also imposed on the southern part of the western boundary of model layer 1 where the boundary is coincident with the 700-ft equipotential line on the ground-water-level map of the glacial deposits (fig. 11).

Water-level data reported by Eberts (1999) indicate that ground-water flow in the bedrock on both sides of the Ohio River is toward the river. As a result, the southern boundary of model layer 2 coincides with the position of the Ohio River. Specified hydraulic heads used to simulate this boundary were derived from 1:250,000 topographic maps. A specified-head boundary condition also is imposed along a small section of the western boundary of model layer 2 where notable glacial aquifers, and thus model layer 1, are absent. This boundary coincides with the 700-ft equipotential line on the potentiometric-surface map of the carbonate-rock aquifer in figure 12.

Specified-flux boundary

A specified-flux boundary condition is imposed on the most northwestern boundary of model layer 2. Ground-water flow across this boundary is approximated by use of Darcy's Law,

$$Q = -KA\frac{dh}{dl} \quad , (1)$$

where K is horizontal hydraulic conductivity,
A is the cross-sectional area through which flow occurs, and
$\frac{dh}{dl}$ is the hydraulic gradient approximated from the hydraulic-head contours in figure 12.

This boundary flow was manually recalculated and a new amount of flow was specified in the model after transmissivity estimates were updated by the model during calibration.

Head-dependent flux boundaries (sources and sinks)

Principal streams that drain the modeled area are explicitly simulated by use of head-dependent-flux boundary conditions (figs. 21*B* and 22). Although these streams only partially penetrate the glacial deposits or carbonate-rock aquifer, analysis of streamflow data indicates that they are discharge points for regional flow within the aquifer system. Hydraulic heads along these stream cells are set equal to the stage of the streams.

The upper boundary of the aquifer system coincides with the water table, which is generally present in glacial deposits but is locally present in bedrock where glacial deposits are thin or absent. A discussion of the treatment of flow across the water-table boundary in the model can be found in the following section.

APPROACH TO MAPPING REGIONAL RECHARGE AND DISCHARGE

Anderson and Woessner (1992, p. 152) note that no universally applicable method has been developed for estimating ground-water recharge across a water table and that most proposed methods have been used with limited success. Although recent investigations have demonstrated that spatial variation in the rate of recharge across the water table of an aquifer system can be significant (Stoertz and Bradbury, 1989), modelers have traditionally assumed a spatially uniform recharge rate to simulate the water-table flux across areas of similar surficial geology. Such an approach prohibits adequate representation of flow across the water table because ground-water basins often include areas where the net flux is upward (Anderson and Woessner, 1992, p. 152).

A few recently published concepts, which have been used by other researchers to simulate a water-table flux, are summarized below. These ideas were considered during construction of the numerical model of regional flow in the Midwestern Basins and Arches aquifer system.

Jorgensen and others (1989a, b) and Stoertz (1989) demonstrate that the water-table flux, which is appropriate for simulation of an aquifer system, is scale dependent. If the size of a model cell is larger than the length of some flow paths within the aquifer system, some ground water recharges and discharges within the area represented by a single model cell. The result is a need to reduce the amount of net recharge applied at the water-table boundary of the model to simulate the aquifer system correctly at the desired scale. Buxton and Modica (1992) show that despite uniformity of surficial geology (and thereby recharge rates) in the physical aquifer system across a modeled area, net recharge may vary across the modeled area because the water-table boundary combines the effects of recharge from precipitation and ground-water discharge to streams. Stoertz (1989) also notes that a model-cell spacing that captures the general water-table curvature is necessary in order to equate simulated recharge with basin yield. An additional observation by Stoertz (1989) is that simulated patterns of recharge and discharge are not affected if the permeability of the entire basin is changed; however, simulated recharge and discharge rates are affected. To map recharge and discharge areas and to simultaneously estimate appropriate rates, the modeler must constrain the model solution with some measurements of flow such as streamflow or pumpage.

In the current investigation, the assumption was made that the amount of net recharge appropriate for simulation of regional ground-water flow equals the amount of water necessary to maintain the regional trend of the water table and to simultaneously supply the principal streams with a base flow equal to long-term average ground-water discharge from fairly stable flow systems within the aquifer system (mean sustained ground-water discharge). This net recharge excludes recharge across the water table that discharges near the point of recharge by means of evapotranspiration or by means of local-flow-system discharge to small tributary streams.

Because net regional recharge results from the combined effects of recharge from precipitation and local ground-water discharge, net regional recharge is simulated in the numerical model by applying a uniform rate of recharge to areas of similar surficial geology and allowing recharge in excess of the appropriate net regional recharge to discharge by application of a general head-dependent-flux boundary condition above the uppermost active model layer (figs. 21 and 22). This general head-dependent-flux boundary condition is not coincident with the head-dependent-flux boundary condition used to simulated the principal streams (stream cells); no recharge is applied to stream cells because the principal streams are areas of known regional discharge, and estimates of discharge from regional flow systems to these streams are used to constrain the model solution.

Hydraulic heads specified for the general head-dependent-flux boundary condition used to help simulate the exchange of water at the regional water table are equal to the altitude of the regional water table. These altitudes were estimated by a method in which digital topographic data and empirical equations relate water-table altitudes and land-surface topography (Williams and Williamson, 1989). The conductance term for the general head-dependent-flux boundary condition is defined to be proportional to the total length of small tributary streams in each model cell (fig. 20) because such streams are assumed to dominate the exchange of water at this boundary.

The inclusion of the general head-dependent-flux boundary condition in the numerical model allows for simulation of some discharge from regional flow systems to areas that are not coincident with the principal streams. Such upward net flux across the regional water table may include water that flows from the point of recharge by way of regional flow systems and subsequently leaves the aquifer system through evapotranspiration or discharge to springs, seeps, ditches, and streams smaller than those represented by the stream cells in the model. This approach to simulation of the water-table boundary also allows for simulation of horizontal flow in the water-table aquifer.

Regional recharge and discharge areas were mapped on a cell-by-cell basis by computing the difference between the amount of recharge applied to the uppermost active model layer and the amount of water lost by means of the head-dependent-flux boundary conditions. In localized areas near regional potentiometric highs, recharge to the deep regional flow systems may be higher than the amount of recharge applied to the model in areas with similar surficial geology. In these places, additional water may enter the simulated regional flow systems by means of the general-head-dependent-flux boundary condition. No net regional recharge or discharge was simulated or mapped where layer 2 is the uppermost active model layer and the carbonate-rock aquifer is isolated from the water table by the upper confining unit (fig. 21).

Head-dependent-flux boundary conditions have been used by other modelers to simulate the flux across a regional water-table boundary (Williamson and others, 1990; Leahy and Martin, 1993). Because the Midwestern Basins and Arches aquifer system is a relatively unstressed steady-state system at the regional scale, the application of a head-dependent-flux boundary condition in this investigation had to differ slightly from previous applications. Specifically, the approach taken in this investigation, as described above, allows net regional recharge to be computed by a steady-state model on a cell-by-cell basis while horizontal flow in the water-table aquifer is simulated. This is possible because, in addition to observations of hydraulic head, base-flow observations along the stream cells are included in the model. The combination of hydraulic-head and base-flow observations was necessary to prevent the general head-dependent-flux boundary condition from overly constraining the model solution.

PARAMETERIZATION

To simulate steady-state regional ground-water flow in the aquifer system, the modeler specified the following system characteristics: (1) horizontal hydraulic conductivity or transmissivity, (2) vertical hydraulic conductivity, (3) streambed hydraulic conductivity, (4) recharge, and (5) a conductance term for the general head-dependent-flux boundary condition used to help simulate flux at the regional water table. These quantities were calculated by means of 16 parameters (a quantity that is estimated by use of trial and error or nonlinear regression) because it was found that regional ground-water flow in the aquifer system could be reasonably simulated with this few parameters. In addition, for reliable estimation of parameter values, the number of parameters must be a fraction of the number of observations of ground-water levels and flows used to estimate them (Hill, 1992, p.15).

Horizontal hydraulic conductivity in layer 1, used to simulate glacial deposits, is simulated with three parameters. The corresponding parameter zones (areas over which a parameter value is applied uniformly) are shown in figure 23*A* and represent areas of moraine deposits, outwash deposits, and glaciolacustrine deposits. The horizontal hydraulic conductivities are effective values that represent the combined effects of sands and gravels (glacial aquifers) and clayey till (glacial confining units) on regional ground-water flow. (These horizontal hydraulic conductivities are multiplied within the computer program by specified saturated thicknesses to compute transmissivity.) Transmissivity in layer 2 is simulated with two parameter zones representing the carbonate-rock aquifer and the upper weathered zone water-bearing unit (fig. 23*B*). (In the "Model Discrimination" section of this report, results are presented for an alternative model in which the parameter value for the carbonate-rock aquifer zone is horizontal hydraulic conductivity rather than transmissivity. In this alternative model, the carbonate-rock aquifer's transmissivity varies systematically with aquifer thickness.)

The vertical hydraulic conductivity between the glacial deposits (layer 1) and the bedrock (layer 2) (fig. 23) is simulated with four parameters. One parameter is used to represent the vertical hydraulic conductivity of the upper confining unit. The other three represent the effective vertical hydraulic conductivities of the glacial deposits and the underlying bedrock where the shale is absent. The associated parameter zones coincide with areas of moraine deposits underlain by bedrock, outwash deposits underlain by bedrock, and glaciolacustrine deposits underlain by bedrock.

Streambed hydraulic conductivity is simulated by use of two parameters. One parameter is used to simulate most streams within the modeled area, and the other is used to simulate the effect of the upper confining unit where it separates the streams and the carbonate-rock aquifer in the southeastern part of the modeled area (fig. 21). Streambed thickness and area for each stream cell are specified.

Recharge from precipitation is simulated with four parameters. The principal recharge zone represents recharge to moraine deposits (ground- and end-moraine deposits) or locally to the carbonate-rock aquifer. The other smaller zones represent recharge to outwash deposits, glaciolacustrine deposits, or the upper weathered zone water-bearing unit directly (fig. 23).

Finally, the conductance term of the general head-dependent-flux boundary condition used to help simulate the exchange of water at the regional water table (fig. 21*B*) is simulated by use of one parameter. This conductance parameter is multiplied by the lengths of small streams present within each respective model cell to attain the conductance needed by the head-dependent boundary package of MODFLOWP.

MODEL CALIBRATION

Calibration of a numerical ground-water flow model is the process of finding a set of boundary conditions, parameter values, and stresses that produce simulated ground-water levels and flows that match field-based measurements or estimates within a preestablished range of error (Anderson and Woessner, 1992, p. 223). The difference between the observed and simulated ground-water levels and flows are hydraulic-head and flow residuals, respectively. The observed values used for the regional ground-water flow model include 389 synoptic measurements of ground-water levels in the carbonate-rock aquifer and the upper weathered zone water-bearing unit, and 43 estimates of mean sustained ground-water discharge to principal streams that represent long-term steady-state conditions in the aquifer system. (These data are discussed in the "Levels and Discharge" sections of this report.) No observed ground-water levels in the glacial deposits were included in the model.

An estimate of the standard deviations for the errors in these observations was made in advance of model simulations to calculate weights for the regression, discussed below in the "Procedure" section. The estimated standard deviations for the errors in the ground-water-level data include the error associated with determination of the measuring-point elevations from topographic maps, deviation of measured values from long-term average ground-water levels (Eberts, 1999), and vertical hydraulic gradients in the aquifer system due to measurement of open-hole wells that may not represent water levels strictly associated with the regional flow systems. These sources of error were evaluated for each measurement; standard deviations of the errors ranged from 6 to 12 ft.

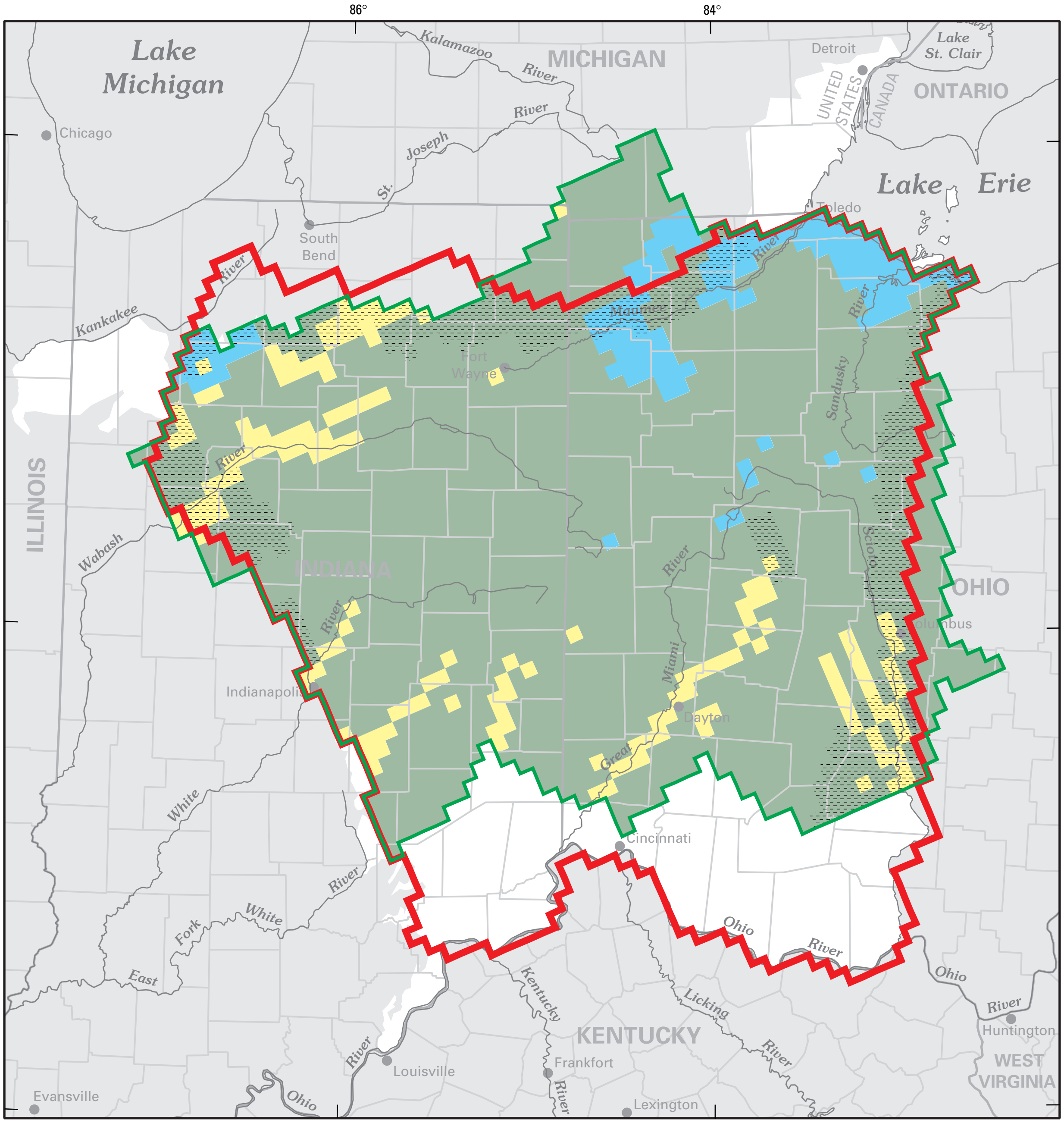
Estimates of mean sustained ground-water discharge to selected streams were assumed to be appropriate calibration values for simulation of steady-state regional flow in the aquifer system. These means range from 88 to 98 percent streamflow duration—streamflow that is equaled or exceeded 88 to 98 percent of the time—and all but four of the means fall between 88 and 94 percent streamflow duration. (Previous researchers (Cross, 1949; Schneider, 1957) have used streamflow that is exceeded 90 percent of the time as an approximate index of dry-weather flow in Ohio.)

For calculating the weights in the regression (see below), it is assumed that the error associated with the estimates of mean sustained ground-water discharge has a 90 percent chance of being 20 percent of the estimated discharge. Estimation of standard deviations associated with these values followed procedures described in Hill (1992, p. 49).

PROCEDURE

An automated nonlinear-regression approach to calibration developed by Cooley and Naff (1990) and extended for complicated three-dimensional problems by Hill (1992) was used in this investigation. Specifically, parameter values were automatically adjusted to achieve the smallest possible value of the objective function. The objective function in this method is the weighted sum of squared differences between observed and simulated hydraulic heads and flows:

A



Base from U.S. Geological Survey digital data,
1:2,000,000, 1972

0 20 40 60 MILES
0 20 40 60 KILOMETERS

EXPLANATION

Glacial deposits

Moraine deposits—Ground and end moraines

Glaciolacustine deposits

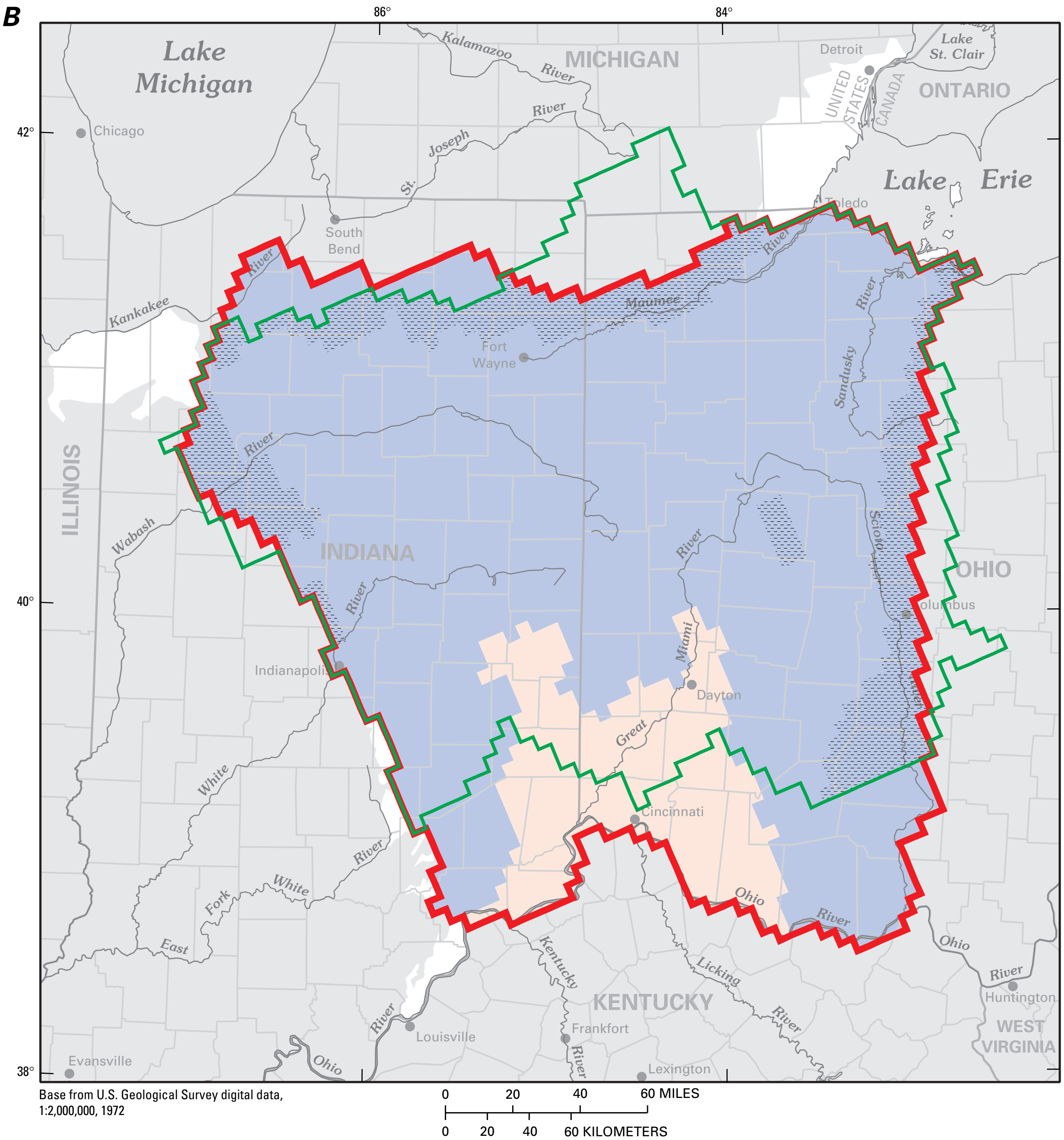
Outwash deposits

Upper confining unit—Shows where confining-unit resistance to vertical flow is simulated between model layers

Boundary of model layer 1

Boundary of model layer 2

FIGURE 23.—Zones used for model parameterization: (*A*) model layer 1



EXPLANATION

Carbonate-rock aquifer

Upper weathered zone water-bearing unit

Upper confining unit—Shows where confining-unit resistance to vertical flow is simulated between model layers

Boundary of model layer 1

Boundary of model layer 2

FIGURE 23. Continued—Zones used for model parameterization: (*B*) model Layer 2

$$SSE = \Sigma[w_i^{1/2}e_i]^2, i = 1, n, \quad (2)$$

where e_i is the difference between the observed and calculated values of measurement i,
$w_i^{1/2}$ is the square root of the weight assigned to the error in the observed value of measurement i;
$w_i^{1/2}e_i$ is the weighted residual corresponding to measurement i; and
n is the number of observations.

The weights in this equation reflect the assumed reliability (standard deviations) of the hydraulic-head measurements or flow estimates (observations) and account for the different units of measure associated with hydraulic heads (L) and flows (L^3/T). The weights equal 1 divided by the variance of the observation error. The parameter values that correspond to the smallest SSE possible for the parameterization and boundary conditions imposed on the model are called the optimal parameter values.

Scaled sensitivities for each of the model parameters also can be computed by use of the nonlinear regression method. Scaled sensitivities equal

$$\frac{\partial y_i}{\partial b_j} w_i^{1/2} b_j, i = 1, n; j = 1 \quad , \quad (3)$$

where b_j is one of the model parameters and
y_i is a calculated hydraulic head or flow.

A comparison of the scaled sensitivities for various parameters (hydraulic conductivity, transmissivity, recharge, conductance terms) in a specific model provides information on the relative effect of each parameter in the regression. Relatively small scaled sensitivities are associated with parameters that have little effect on simulated results and cannot be estimated by nonlinear regression.

Application of the nonlinear-regression procedure of model calibration in this investigation ensures that model error is due to model design rather than to suboptimal parameter values. This enables comparison of various model designs so that various aspects of the numerical and conceptual models can be tested. In general, the best models have (1) the smallest parameter coefficients of variation, (2) parameter correlations of less than 0.95, (3) the smallest calculated error variance (SSE divided by the difference between the number of observations and the number of estimated parameters (Draper and Smith, 1981)), and (4) weighted residuals that are normal, independent, and of equal variance. On the basis of these standards, the calibrated final model presented herein is the best representation of regional ground-water flow in the Midwestern Basins and Arches aquifer system among the alternatives tested. (A brief discussion of what was learned from two alternative models is found in the section "Model Discrimination.")

ESTIMATES OF PARAMETER VALUES

For 8 of the 16 model parameters described in the "Parameterization" section of this report, scaled sensitivities are large enough for the parameter values to be estimated by nonlinear regression. The number and location of observations used for the model calibration affect these scaled sensitivities and are directly responsible for which parameters can be estimated. Estimated parameters in the calibrated final model, ordered from highest to lowest in terms of sensitivity, include (1) transmissivity of the carbonate-rock aquifer, (2) horizontal hydraulic conductivity of the moraine deposits, (3) recharge applied to the moraine deposits, (4) effective vertical hydraulic conductivity of the combined moraine/bedrock areas, (5) the conductance term for the general head-dependent-flux boundary condition used to simulate the regional water table, (6) hydraulic conductivity of streambeds throughout most of the modeled area, (7) horizontal hydraulic conductivity of the outwash deposits, and (8) vertical hydraulic conductivity of the upper confining unit.

The other model parameters were assigned values from available data in the literature and were held constant during the regression. Any adjustments to these values were made by trial and error. These values include an effective vertical hydraulic conductivity of 0.1 x 10^{-2} ft/d for the combined glaciolacustrine/bedrock deposits and 0.1 ft/d for the combined outwash/bedrock deposits. The value for the hydraulic conductivity of the streambeds that are underlain by the upper confining unit is set at 0.1 x 10^{-2} ft/d. Horizontal hydraulic conductivities of the glaciolacustrine deposits and the upper weathered zone water-bearing unit are set at 0.05 and 0.06 ft/d, respectively. Values for recharge applied to the glaciolacustrine deposits, the upper weathered zone water-bearing unit, and the outwash deposits range from 0.1 x 10^{-2} to 11.8 in/yr.

Estimated values for the optimal parameter set from the calibrated final model are listed in table 3. Each of the parameter estimates falls within the range of published field values for the Midwestern Basins and Arches aquifer system where data are available (table 2). The estimated transmissivity for the carbonate-rock aquifer not only is within the range of field-determined estimates of transmissivity but also is within 16 percent of the geometric mean of these values. The estimated effective horizontal hydraulic conductivity for the moraine deposits falls within the range of textbook values for these materials (Freeze and Cherry, 1979). The estimated recharge value in table 3 does not represent net regional recharge to the regional flow systems nor does it represent all recharge to the entire aquifer system, which would include recharge to local, intermediate, and regional flow systems.

TABLE 3.—*Parameter estimates and reliability of the optimal parameter set from the calibrated final model of regional flow in the Midwestern Basins and Arches aquifer system*

[ft^2/d, feet squared per day; ft/d, feet per day; in/yr, inches per year]

Parameter	Parameter estimate	Approximate 95-percent linear confidence interval	Relative parameter reliability[a]
Transmissivity of the carbonate-rock aquifer	1,610 ft^2/d	1,030 – 2,500 ft^2/d	0.23
Horizontal hydraulic conductivity of the moraine deposits	21.3 ft/d	13.7 – 33.1 ft/d	.23
Recharge applied to the moraine deposits[b]	2.15 in/yr	1.41 – 2.88 in/yr	.18
Effective vertical hydraulic conductivity of the combined moraine/bedrock areas	0.375×10^{-2} ft/d	0.139×10^{-2} – 0.101×10^{-1} ft/d	.59
Conductance term for the general head-dependent flux boundary condition used to simulate the regional water table[b]	0.259 ft^2/d	0.161 – 0.418 ft^2/d	.25
Hydraulic conductivity of streambeds throughout most of the modeled area[c]	0.0149 ft/d	0.0045 – 0.05 ft/d	.77
Horizontal hydraulic conductivity of the outwash deposits	168 ft/d	46.1 – 620 ft/d	.87
Vertical hydraulic conductivity of the upper confining unit	0.466×10^{-3} ft/d	0.645×10^{-4} – 0.338×10^{-2} ft/d	1.80

[a] Coefficient of variation for the recharge parameter; comparable measure of reliability for the other parameters, which were log-transformed for the regression. Smallest values indicate greatest parameter reliability.

[b] The estimated recharge value is not net recharge to regional flow systems; net recharge to regional flow systems is computed by subtracting the flux associated with the head-dependent flux boundary conditions from this estimated value of the recharge parameter on a cell-by-cell basis (fig. 28).

[c] Reported values range from 0.0007 – 18.7 ft/d (Meyer, 1978; Smith and others, 1985; Cunningham, 1992, Dumouchelle and others, 1993).

Rather, it represents the recharge rate applied to the moraine deposits that may be used in conjunction with the effects of the general head-dependent-flux boundary condition to estimate regional recharge rates. In a few areas near the regional potentiometric highs, the estimated recharge value was not great enough to balance observations of hydraulic heads and flow used to constrain the model solution, so water entered the model by means of the general head-dependent-flux boundary condition. This result was expected because the amount of recharge that reaches the deepest parts of an aquifer system is typically greatest near regional potentiometric highs. Net values of regional recharge or discharge are not apparent from table 3 but are presented in map form later in this report.

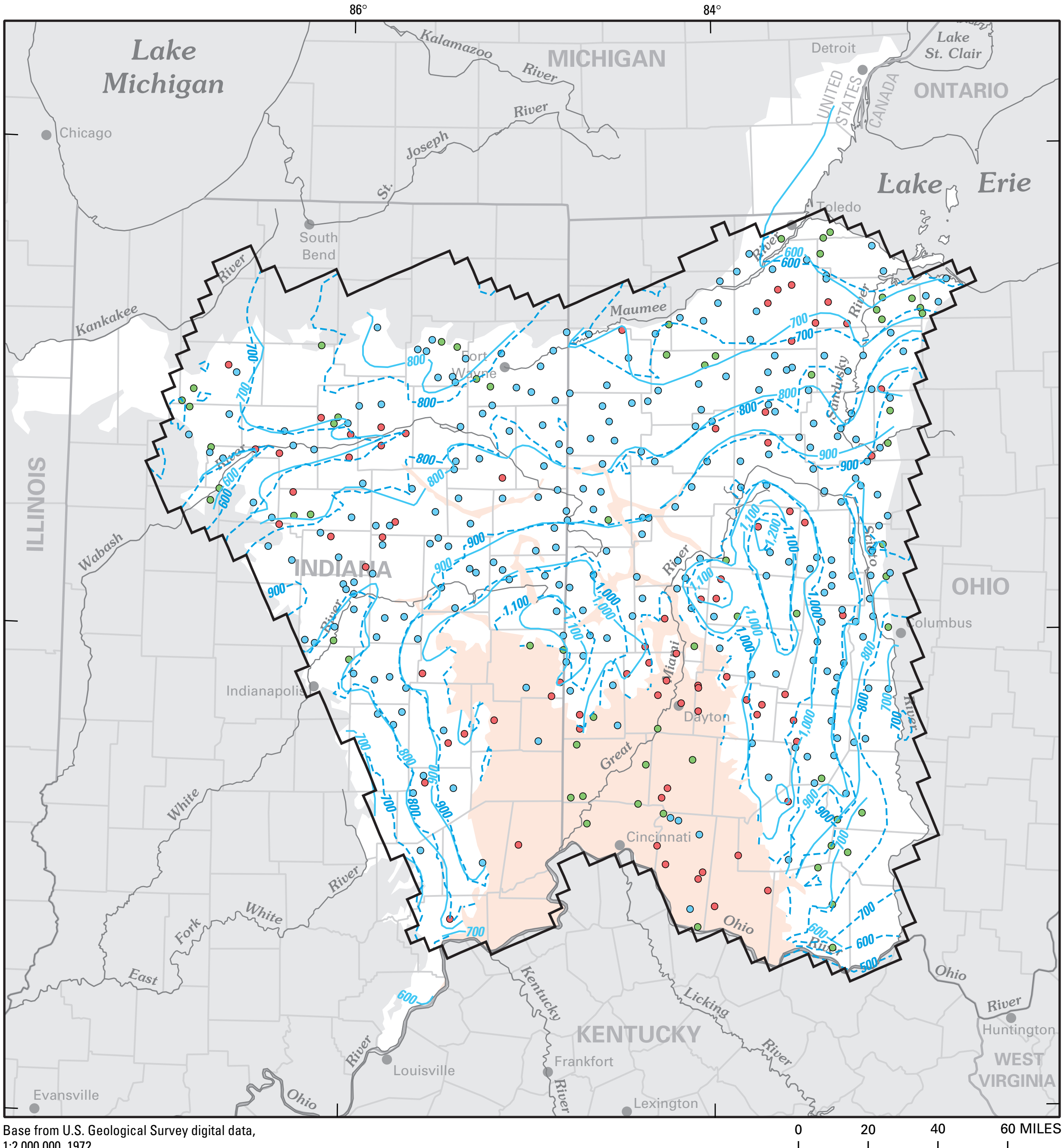
Model output indicates that no parameter correlations exceed 0.90. The greatest correlation (0.87) is between the recharge parameter and the conductance parameter associated with the general head-dependent-flux boundary condition.

SIMULATED HYDRAULIC HEADS

A total of 389 measured ground-water levels in the carbonate-rock aquifer and the upper weathered zone water-bearing unit (model layer 2) were used as observations in the regression. No water levels in the glacial deposits (model layer 1) were used because the available data, which are from drillers' logs, are likely to reflect a local water table associated with local flow systems not explicitly simulated in the model. In addition, estimates of the regional water table, which is typically present in glacial deposits, were included as part of the general head-dependent-flux boundary condition.

A comparison of simulated and measured potentiometric surfaces in the carbonate-rock aquifer is illustrated in figure 24. Simulated equipotential lines closely follow equipotential lines contoured from measured ground-water-level data. Observation locations are coded on the map to indicate locations where the simulated and measured (observed) hydraulic heads differ by less than three times the standard deviation of the errors associated with the observation. Locations where simulated hydraulic heads are above or below this range also are noted. Figure 24 indicates that the simulated hydraulic head most commonly differs from the observed hydraulic head by more than three times the standard deviation of the errors associated with the observation in the areas along the Great Miami River and along the northeastern and southeastern edges of the model. Although these patterns indicate some lack of model fit in these areas, the overall model fit is good.

A graph of weighted residuals plotted against weighted simulated values (Draper and Smith, 1981; Hill, 1994) (fig. 25) shows that the hydraulic-head residuals are indeed ran-



EXPLANATION

Upper weathered zone water-bearing unit— Not an aquifer. Carbonate-rock aquifer absent

Potentiometric contour—Carbonate-rock aquifer. Contour interval 100 feet. datum is sea level

700 Simulated

700 Measured

Boundary of model layer 2

Observation location

Reasonably predicted. Simulated and observed hydraulic heads differ by less than three times the standard deviation of the errors associated with the observation

Underpredicted. Simulated hydraulic head less than observed hydraulic head by more than three times the standard deviation of the errors associated with the observation

Overpredicted. Simulated hydraulic head greater than observed hydraulic head by more than three times the standard deviation of the errors associated with the observation

FIGURE 24.—Simulated and measured (observed) hydraulic heads in the carbonate-rock aquifer and the upper weathered zone water-bearing unit (model layer 2).

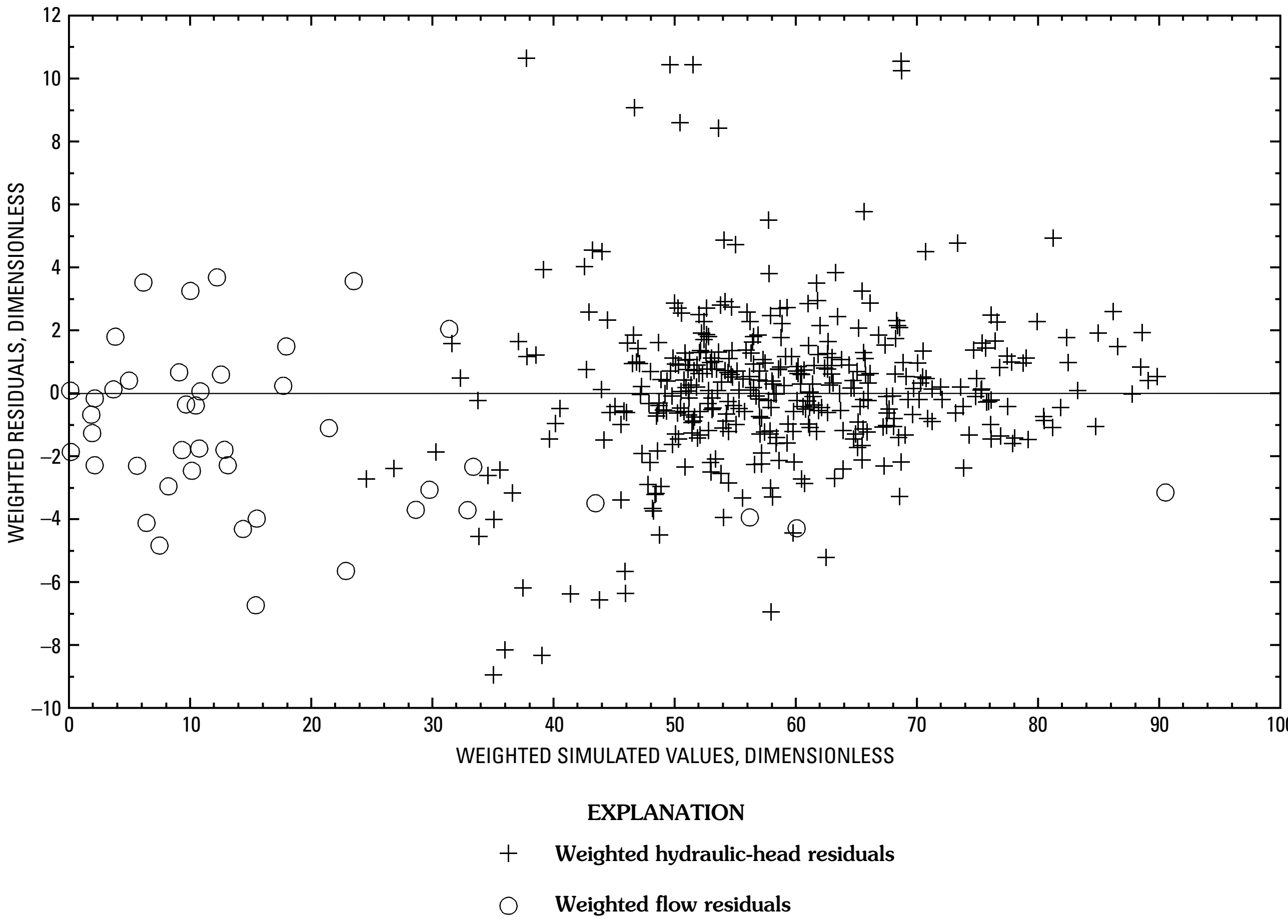


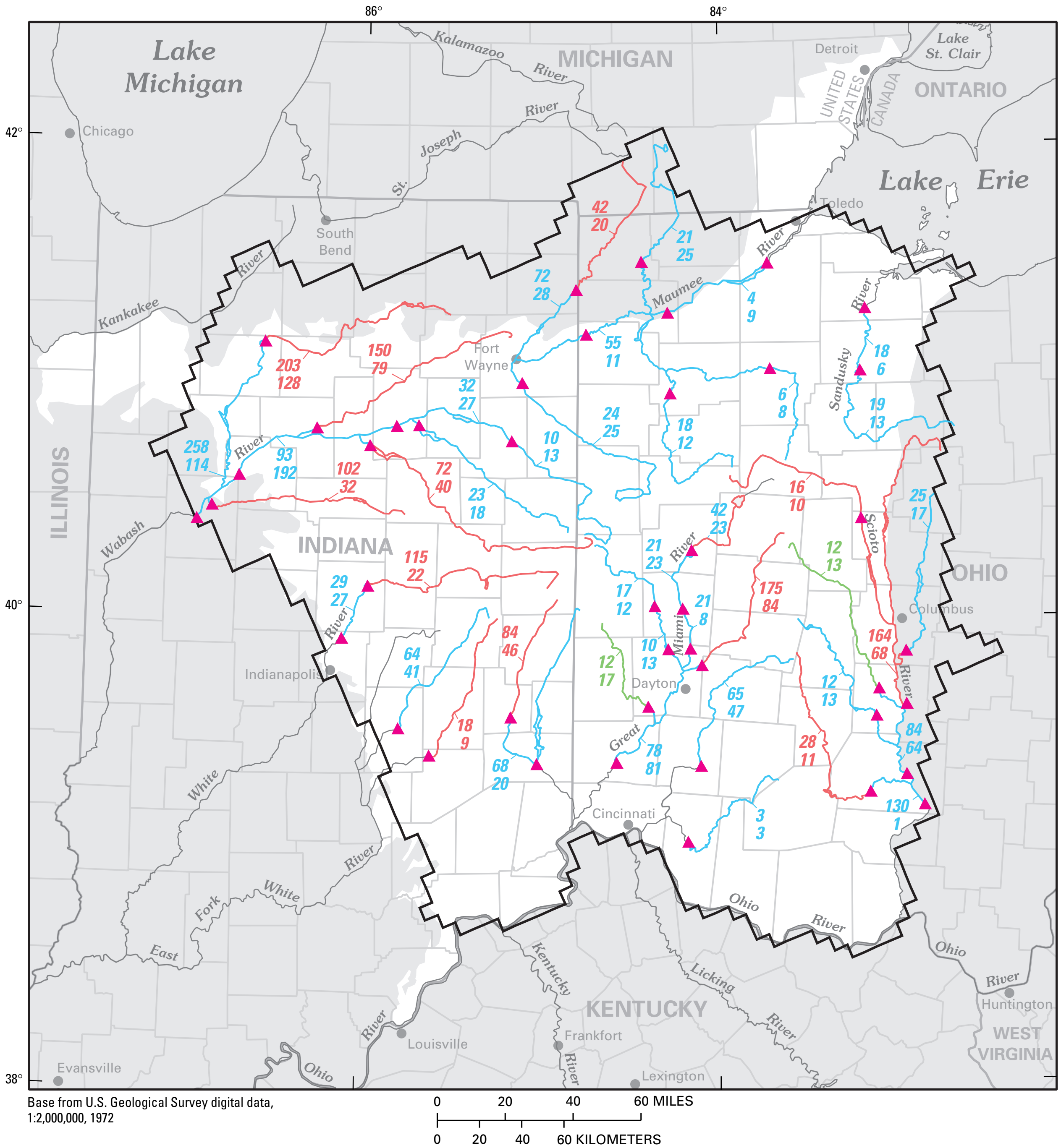
FIGURE 25.—Weighted residuals of hydraulic heads and flows plotted against weighted simulated values from the regional ground-water flow model for the Midwestern Basins and Arches aquifer system.

dom and have equal variance. Results of a runs test printed by MODFLOWP show that the hydraulic-head residuals are also independent. [The runs test takes into account the order of the residuals; too few runs commonly indicates positive serial correlation between residuals at individual locations (Hill, 1992).]

The root mean squared (RMS) error associated with hydraulic heads, which is the average of the squared differences in measured and simulated hydraulic heads, is another measure of model fit. Anderson and Woessner (1992, p. 241) note that if the ratio of the RMS error to the total head loss in the system is small, then the errors are only a small part of the overall model response. The RMS error computed from measured and simulated hydraulic heads in the regional ground-water flow model is 40 ft. The total head loss from the highest recharge area to the lowest discharge area in the model is 710 ft. The ratio of the RMS error to the total head loss in the system is 0.06. In summary, the model errors are only a small part of the overall model response; thus the model satisfactorily approximates ground-water-level observations.

SIMULATED FLOWS

Simulated and observed flows (estimates of mean sustained ground-water discharge) along 43 stream reaches were compared to help evaluate overall model response. Estimates of mean sustained ground-water discharges and simulated flows are listed by stream reach in figure 26. These values are difficult to compare without knowledge of the error associated with the observation for each stream reach. This is because each streamflow-gaging station that bounds a selected stream reach is assumed to contribute the same amount of error to the observation; some stream reaches are bounded by one streamflow-gaging station, whereas others are bounded by as many as five. Mean sustained ground-water discharges to stream reaches bounded by five streamflow-gaging stations are less well known than observations for other reaches bounded by fewer gaging stations. Stream reaches in figure 26 are coded to indicate locations where the simulated and observed flows differ by less than three times the standard deviation of the errors associated with the observation. Most simulated flows fall within this range. Reaches



EXPLANATION

Boundary of modeled area

68 / 20 **Ground-water discharge to selected stream reach**—Color corresponds to deviation of simulated from observed flow. Upper number is estimated mean sustained ground-water discharge (observed flow), in cubic feet per second. Lower number is simulated ground-water discharge, in cubic feet per second

Reasonably predicted—Simulated and observed flows differ by less than three times the standard deviation of the errors associated with the observation

Underpredicted—Simulated flow less than observed flow by more than three times the standard deviation of the errors associated with the observation

Overpredicted—Simulated flow greater than observed flow by more than three times the standard deviation of the errors associated with the observation

Streamflow-gaging station

FIGURE 26.—Simulated ground-water discharge to selected stream reaches from regional flow systems and estimated mean sustained ground-water discharge (observed flow) to the reaches.

where simulated flows are above or below this range are also noted.

The graph of weighted residuals plotted against weighted simulated values in figure 25 indicates that the flow residuals are approximately random and have nearly equal variance except for weighted simulated values greater than about 40. These values are consistently less than observed values. (Negative weighted flow residuals calculated by MODFLOWP indicate underprediction of flow because the convention within the model is to represent ground-water losses to streams as negative values.) Simulated flows underpredict the flow observations slightly more often than they predict and overpredict them. All of the underpredicted stream reaches are in the upbasin areas; all of the streams at the bottom of the basins are well predicted considering the error on the observations. An alternative model that was constructed to test a hypothesis about why some upstream reaches are underpredicted in the regional ground-water flow model is discussed below. Results of a runs test for combined hydraulic-head and flow residuals, however, indicate randomness among weighted residuals.

MODEL DISCRIMINATION

Model discrimination is the process of comparing different hypotheses about an aquifer system by comparing results of models constructed using the different hypotheses (Hill, 1992). Two alternative models were developed to test two hypotheses used in the construction of the regional ground-water flow model. For the first alternative model, it was hypothesized that flow to streams in the upbasin areas may be underpredicted because a notable amount of ground water that sustains flow in the principal streams during the driest periods may be recharged at the water table within the area of the stream cells. No recharge is applied to the stream cells in the calibrated final model, therefore, this intracell flow is not represented in the model. Such a model design would have a greater affect on model calibration along upstream reaches as compared to downstream reaches because the area represented by the stream cells makes up a greater proportion of the drainage basins associated with upstream reaches.

To test whether some upstream reaches were underpredicted simply because no recharge is applied to the stream cells, an additional recharge parameter that represents recharge to stream cells was added to the model. The optimal recharge rate for this new recharge parameter is virtually zero, and the same stream reaches are underpredicted by this new model. In other words, the results of the new model are similar to the results of the model without the additional recharge parameter. It was concluded that lack of recharge to stream cells in the calibrated final model is not a factor that affects the overall model response. Locally, however, the underprediction of upstream reaches that flow along highly permeable outwash valleys, such as the Mad River (a tributary to the Great Miami River), may be related to this lack of simulated recharge.

Because simulation of an increased amount of curvature at the water table equates with simulation of an increased amount of flow within an aquifer system (fig. 19), the underprediction of some upstream reaches in the calibrated final model is possibly related to the cell spacing and the inability of the selected spacing to capture the curvature of the water table necessary to balance mean sustained ground-water discharge to the underpredicted stream reaches. Such a scale effect would be smallest in relation to the most downstream reaches because a greater proportion of sustained ground-water discharge to these streams is associated with the most regional trends of the water table, which are well represented by the coarse cell spacing of the calibrated final model. In other words, the mean sustained ground-water discharges used as observations in the regional ground-water flow model may be slightly high for some of the upstream reaches because of the cell spacing chosen for this investigation.

The calibrated final model presented in this report, however, is a reasonable representation of regional ground-water flow in the Midwestern Basins and Arches aquifer system. This is demonstrated, in part, by the estimated transmissivity for the areally extensive carbonate-rock aquifer, which is within 16 percent of the geometric mean of reported transmissivities. On the basis of Darcy’s Law, calculated flows vary in direct proportion to aquifer transmissivity and hydraulic gradient. If the mean sustained ground-water discharges to streams used to help calibrate the numerical model were not generally appropriate as calibration targets, transmissivities and hydraulic conductivities could not have been so reasonably estimated while hydraulic gradients were so well predicted. Stated another way, if flow observations and thereby simulated flows were not generally appropriate for the scale of the model, transmissivities or hydraulic gradients would have to have been inappropriately adjusted to accommodate the associated excess or missing flow.

A second alternative model was used to test whether the transmissivity of the carbonate-rock aquifer varies systematically with the thickness of the carbonate rocks. In this model, horizontal hydraulic conductivity rather than transmissivity of the carbonate-rock aquifer is estimated and multiplied by the thickness of the carbonate rocks to determine optimal transmissivities. This alternative model construction is graphically depicted in figure 27. It differs from the calibrated final model in that the lower part of the carbonate rocks is represented in this second alternative model, whereas the use of a single transmissivity in the calibrated final model would be similar to a model with a third layer of near-zero transmissivity used to represent the deepest part of the carbonate rocks.

The optimal parameter set for this second alternative model includes an estimated vertical hydraulic conductivity for the upper confining unit that is orders of magnitude higher than is considered reasonable. In addition, vertical hydraulic

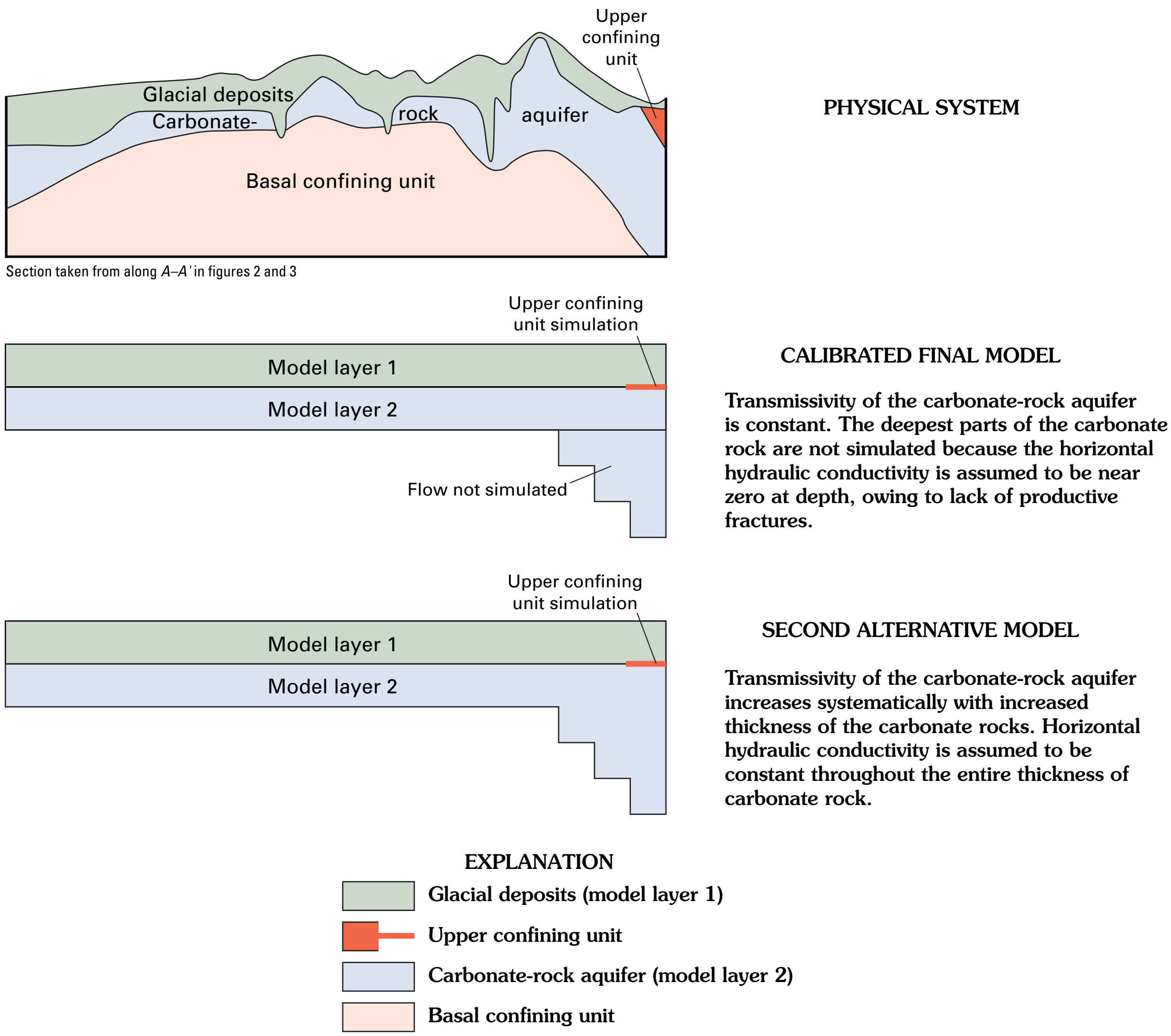


FIGURE 27.—Diagrams of two regional ground-water flow models used to test whether the carbonate rocks in the Midwestern Basins and Arches Region are transmissive throughout their entire thickness.

gradients between the carbonate-rock aquifer and the glacial deposits are tens of feet greater than gradients considered reasonable. Results of the runs test indicate that the residuals for this alternative model are not independent. In general, the largest head residuals are associated with the areas where the carbonate rock is very thick and computed transmissivities are thereby quite large.

To summarize, the second alternative model cannot match field conditions and simultaneously accommodate the ground-water flows associated with the large simulated transmissivities in areas where the carbonate rocks are very thick. It is concluded that the entire thickness of the carbonate rocks may not contribute substantially to the transmissivity of the carbonate-rock aquifer. This is consistent with findings that fractures at depth in the carbonate rocks may not be transmissive (Arihood, 1994). In addition, local anhydrite deposits are present at depths greater than 150 ft in the Sandusky Bay area, an area where the carbonate rock is very thick (Carlson, 1991). The presence of anhydrite is indicative of little active freshwater flow in the carbonate rock at these depths in at least one part of the aquifer system.

RELIABILITY OF PARAMETER ESTIMATES

Confidence intervals on estimated parameter values can help indicate the reliability of the estimates. Linear confidence intervals on the parameters can be computed if the model is correct and linear in the vicinity of the optimal set of values and if the parameters are normally distributed (Hill, 1994). Beale's measure and its critical values can be used to test model linearity (Cooley and Naff, 1990). The Beale's measure for the calibrated final model is 0.38. This value falls between the critical value of 0.046 (a value below which Beale's measure would indicate model linearity) and 0.52 (the value above which would indicate model nonlinearity).

If weighted residuals are independently distributed, the model is likely to be correct; if weighted residuals are normally distributed and the model is linear, the estimated parameter values are generally normally distributed (Hill, 1994; Seber and Wild, 1989). To test whether the weighted residuals are independent and normally distributed, they were compared with expected independent values from a standard normal distribution. The R_N^2 statistic (Hill, 1992) for the calibrated final model is 0.948, which is less than the critical value of 0.987 at the 95-percent confidence level. This statistic indicates that the weighted residuals may not be independent and normally distributed. However, this test is more restrictive than the less powerful Kolmogorov test (Hill, 1992, p 63). On the basis of the Kolmogorov test, Yager (1993) demonstrated that a R_N^2 statistic of 0.946 indicates that weighted residuals are independent and normally distributed at the 99-percent confidence interval for a model with a similar number of parameters and observations as for the calibrated final model. These results indicate that the weighted residuals from the calibrated final model are at least nearly independent and normally distributed. The distribution of the residuals was not investigated further.

Beale's measure indicates that the calibrated final model is at least slightly nonlinear; moreover, figure 24 indicates some spatial patterns in the weighted residuals for hydraulic head, and figure 25 indicates some nonrandomness of the weighted residuals for flow. As a result, the confidence intervals given in table 3 should be considered approximate.

The relative reliability of estimated parameter values were compared by use of coefficients of variation. Smaller coefficients of variation indicate greater reliability than do larger coefficients of variation. For parameters that are log-transformed during the regression, a substitute for the coefficient of variation (pseudo coefficient of variation) can be calculated by determining the difference between the upper and lower confidence limits divided by the exponential of the estimated parameter value and dividing this result by two times the critical value from the Students-t probability distribution used to compute the confidence limits. This measure is exactly the coefficient of variation for the parameters that were not log-transformed and is a comparable measure of reliability for those parameters that were log-transformed. All of the parameter values except the recharge parameter in the calibrated final model of this investigation were log-transformed during calibration. (Hydraulic-conductivity measurements in various geohydrologic situations are commonly lognormally distributed (Hill, 1992, p. 18)). Coefficients of variation and pseudo coefficients of variation for the optimal parameter set of the calibrated final model are listed in table 3. The parameter values for recharge applied to the moraine deposits, transmissivity of the carbonate-rock aquifer, and horizontal hydraulic conductivity of the moraine deposits are the most precisely known parameter values, whereas the vertical hydraulic conductivity of the upper confining unit is the least precisely known parameter value.

It should be noted once more that the recharge parameter estimate listed in table 3 does not represent net recharge to regional flow systems. Because net regional recharge (or net regional discharge) was computed by subtracting the simulated flux associated with the head-dependent-flux boundary conditions from the flux associated with this estimated value of recharge applied to the moraine deposits (see fig. 28), the reliability of net recharges or net discharges is not known.

SIMULATED REGIONAL GROUND-WATER FLOW

The calibrated final model can be used to quantify various aspects of the conceptual model and to draw conclusions about regional ground-water flow in the Midwestern Basins and Arches aquifer system. A ground-water budget that quantifies flow associated with the regional flow systems represented by the model is given in table 4; simulated net recharges or net discharges across the regional water table were used in the computations.

The calibrated final model represents the movement of 1,292 Mgal/d of water through the parts of the aquifer system not greatly affected by seasonal variations in ground-water recharge from precipitation; this is approximately 10 percent of the total flow in the aquifer system. Ninety-nine percent of this water is from recharge at the water table. Seventy-eight percent of the water (1,006 Mgal/d) leaves the system by means of the principal streams that were explicitly represented in the model by use of the stream cells. Nineteen percent of simulated regional ground-water flow discharges by means of seeps, springs, ditches, small streams, or evapotranspiration—discharge that was simulated by use of the general head-dependent-flux boundary condition. Two percent of the water (24 Mgal/d) leaves the system along the margin of the Illinois (structural) Basin. Much of this water probably discharges to streams just beyond the model boundary, but some of it may move downdip into the Illinois Basin. (Previous work (Cartwright, 1970) demonstrates, by means of temperature data, that some water enters the Illinois Basin and that water in the deep parts of the Illinois Basin ultimately discharges near the center of the basin through fracture zones associated with faults and anticlines in the basin.) Approximately 1 percent (18 Mgal/d) of the water associated with regional ground-water flow represented in the model discharges to the Ohio Valley aquifer that follows the Ohio River (local-scale alluvial aquifer not represented in the calibrated final model) and ultimately discharges to the Ohio River. Most of this discharge is from the carbonate-rock aquifer; less than 1 Mgal/d is from the upper weathered zone water-bearing unit (not an aquifer). Discharge attributed to regional ground-water flow in these bedrock units is a very small percentage of total ground-water discharge to the Ohio River; discharge directly to the Ohio River from the Ohio Valley aquifer along the boundary of the modeled area has been computed at approximately 1,400 Mgal/d (C.G. Norman, Ohio River Valley Water Sanitation Commission, written commun., 1989). These relative amounts of discharge to the Ohio River are consistent with a conceptual model in which local flow systems dominate flow in the Midwestern Basins and Arches aquifer system. Less than 1 percent of the regional ground-water flow represented in the model discharges to Lake Erie.

A ground-water budget that quantifies regional flow within the carbonate-rock aquifer, as represented by the model, was also computed (table 5). Of the 386 Mgal/d of water that moves along simulated regional flow paths in the carbonate-rock aquifer, 85 percent enters the aquifer by means of percolation through the overlying glacial deposits. Fifteen percent is associated with recharge by precipitation directly onto the carbonate-rock aquifer. Most of the water that enters the carbonate-rock aquifer flows back into the overlying glacial deposits. Eight percent of simulated regional flow in the carbonate-rock aquifer discharges to seeps, springs, ditches, and streams smaller than those represented in the model by use of the stream cells. Less than 1

TABLE 4.—*Simulated ground-water budget of regional flow systems in the Midwestern Basins and Arches aquifer system*

[Mgal/d, million gallons per day]

	Flow (Mgal/d)	Percent recharge or discharge
Recharge		
Across the regional trend of the water table	1,277	99
From losing stream reaches	15	1
Total	**1,292**	**100**
Discharge		
To principal streams	1,006	78
Across the regional trend of the water table to seeps, springs, ditches, small streams, or by means of evapotranspiration	242	19
Along the margin of the Illinois (structural) Basin	24	2
From the carbonate-rock aquifer to the Ohio River	18	1
To Lake Erie	1	<1
Across the northwestern boundary of the modeled area	1	<1
From the upper weathered zone water-bearing unit to the Ohio River	<1	<1
Total	**1,292**	**100**

TABLE 5.—*Simulated ground-water budget of regional flow systems in the carbonate-rock aquifer in the Midwestern Basins and Arches aquifer system*

[Mgal/d, million gallons per day]

	Flow (Mgal/d)	Percent recharge or discharge
Recharge		
From percolation of water through glacial deposits	327	85
Across the regional trend of the water table from precipitation directly onto the aquifer	59	15
From the upper weathered zone water-bearing unit	<1	<1
Total	**386**	**100**
Discharge		
By means of flow into overlying glacial deposits	322	84
Across the regional trend of the water table to seeps, springs, and small streams	32	8
To the Ohio River	18	5
To principal streams	9	2
Along the margin of the Illinois (structural) Basin	5	1
Into the upper weathered zone water-bearing unit	<1	<1
Across the northwestern boundary of the modeled area	<1	<1
Total	**386**	**100**

percent of the simulated ground-water flow in the carbonate-rock aquifer is into the upper weathered zone water-bearing unit.

The simulated amount of water that reaches the carbonate-rock aquifer by means of regional flow paths is only 30 percent of simulated regional ground-water flow in the aquifer system. The rest of the water remains within the glacial deposits. Because glacial deposits contribute so much to regional ground-water flow as defined in this report, it is not surprising that drainage basins where glacial deposits are thin, absent, or poorly permeable are associated with small amounts of sustained ground-water discharge to streams (see “Discharge” section). Such basins include those that drain to Lake Erie and those in the southeastern part of the study area (fig. 17).

One of the principal objectives of this investigation, which was met by use of the calibrated final model, was to map regional recharge and discharge areas. This was an important objective because water and contaminants that enter the aquifer system in regional recharge areas are likely to traverse a greater length of aquifer than water that enters the system at local recharge areas (Stoertz, 1989). Computed net amounts of recharge to and discharge from regional flow systems, as represented by the model, were used to construct the regional recharge and discharge map shown in figure 28. This map is regional in scale; therefore, the map is not meant to imply that recharge to flow systems too small to be represented in the model is not possible in areas designated as regional discharge areas and that discharge from local flow systems is not possible in areas designated as regional recharge areas. The map simply implies that more water recharges the fairly stable flow systems than discharges from such systems within areas mapped as regional recharge areas. The opposite holds true for regional discharge areas.

The Midwestern Basins and Arches aquifer system is characterized by alternating regional recharge and discharge areas, typically on a scale of less than 10 mi, except in the northeastern part of the modeled area (fig. 28). Ground water generally does not move from recharge areas associated with the very highest potentiometric levels (figs. 11 and 12) along long, continuous flow paths to areas associated with the very lowest potentiometric levels, such as the Wabash and Ohio Rivers and Lake Erie, while remaining isolated from additions of recharge. Rather, regional recharge and discharge areas are present all along the regional potentiometric gradient depicted by the potentiometric-surface maps in figures 11 and 12, except in the northeastern part of the study area. These patterns of regional recharge and discharge may explain the differences in ground-water ages between the northeastern part of the study area and the rest of the aquifer system (see “Geochemistry” section).

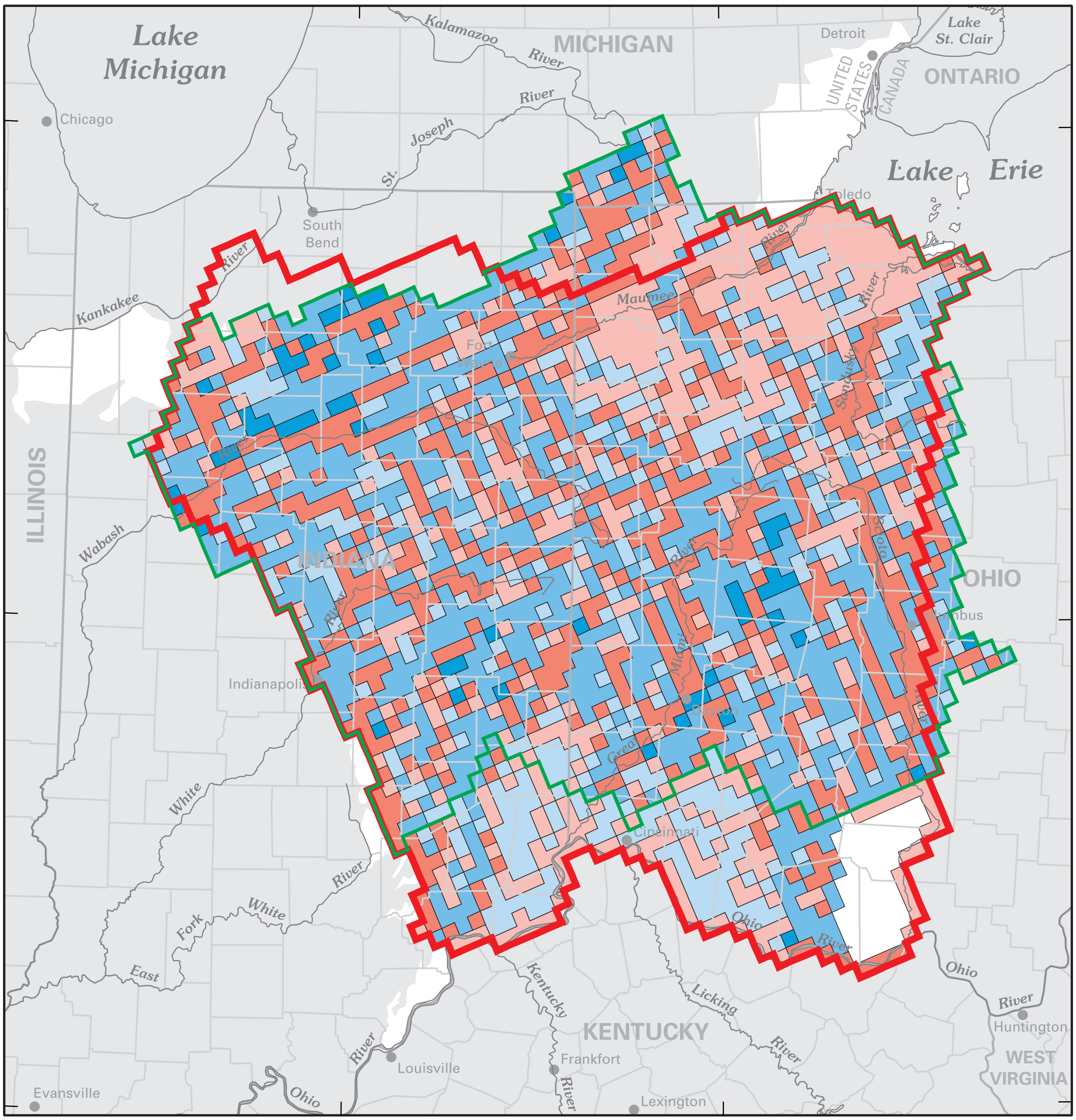
The regional potentiometric high near the Bellefontaine Outlier (fig. 6) is associated with some of the highest recharge rates. Another area associated with high recharge rates is the area of extensive outwash deposits north of the Wabash River in Indiana (fig. 5). The lowest regional recharge rates are associated with the area where the upper weathered zone water-bearing unit is exposed at the land surface or is overlain by thin glacial deposits.

High discharge rates are commonly associated with the principal streams within the modeled area. Mapped discharge areas, however, are not limited to the width of the stream cells. Specifically, a broad area (tens of miles) of regional discharge was simulated in the northeastern part of the modeled area. This area likely represents an area in which water that follows regional ground-water flow paths leaves the system by means of ditches, small streams, or evapotranspiration because ground-water levels are near land surface. The fine-grained glaciolacustrine deposits in the area may be associated with a thick capillary fringe that could help facilitate evapotranspiration. In addition, the hydraulic gradient towards the Maumee River is minimal. Toth (1963) notes that ground-water discharge in basins characterized by low relief, such as the Maumee River Basin in the lowlands near Lake Erie (fig. 6), takes place between the midline and the bottom of the drainage basin. In addition, he notes that only a small proportion of the ground water discharges as base flow in the principal streams in such basins. This pattern is similar to what is observed and simulated for the drainage basins in the northeastern part of the modeled area near Lake Erie. Because some regional ground-water flow discharges before it reaches the streams, estimates of base flow in the streams cannot be equated with recharge to the aquifer system in this area.

Rates of simulated discharge associated with this broad regional discharge area are extremely low. Typically, discharge rates are less than 0.5 in/yr in these areas. It is noteworthy that this broad area of weak discharge is largely coincident with an area characterized by the highest concentrations of dissolved sulfide within the modeled area, which also indicates that oxygenated recharge is not readily available to this part of the aquifer system (Ohio Department of Natural Resources, 1970). In addition, low recharge rates may also explain the presence of isotopically distinct ground water at depth beneath the Maumee River Basin (see “Geochemistry” section).

Areas that are not designated as regional recharge or discharge areas are evident on the regional recharge and discharge map (fig. 28). These are areas in which model layer 2 was the uppermost active layer of the model and the carbonate-rock aquifer is isolated from the water table by means of the upper confining unit. Neither recharge nor the general head-dependent-flux boundary condition was applied to these model cells.

On an areal map, discharge areas commonly constitute a smaller part of the surface area of a watershed than recharge areas do (Freeze and Cherry, 1979, p.197). Percentages of the surface area simulated as regional recharge and discharge



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972

0 20 40 60 MILES

0 20 40 60 KILOMETERS

EXPLANATION

Simulated regional recharge, in inches per year	**Simulated regional discharge, in inches per year**
0 to 0.5	0 to 0.5
Greater than 0.5 to 4.0	Greater than 0.5
Greater than 4	**Boundary of model layer 1**
No simulated regional recharge or discharge	**Boundary of model layer 2**

FIGURE 28.—Simulated regional recharge and discharge areas. (These areas do not necessarily coincide with recharge and discharge areas of local flow systems.)

areas were computed for eight selected surface-water drainage basins that drain the modeled area (fig. 29). Regional recharge areas are larger than regional discharge areas across most of the modeled area; however, regional discharge areas predominate in the northeastern part of the study area.

Regional discharge areas that are associated with the largest simulated upward hydraulic gradients are near the downstream end of the Wabash and Scioto Rivers, where model layers 1 and 2 are present, and just east of the mouth of the Sandusky River. Simulated hydraulic heads in the carbonate-rock aquifer in a few of these model cells were above land surface. The area east of the mouth of the Sandusky River was previously mapped as an area of flowing wells and large springs (Breen, 1989).

Patterns of advective regional ground-water flow that were computed from the model output are shown in figure 30. The flow paths represent flow in glacial deposits and bedrock units rather than flow associated with individual aquifers, although some individual flow paths may represent flow solely in either the glacial deposits or the bedrock units. Regional ground-water flow is generally from regional recharge areas to adjacent regional discharge areas. Some simulated regional flow paths on this figure bypass an adjacent regional discharge area and indicate discharge in an area further down the potentiometric gradient. Such areas worth noting include the downstream end of the Wabash River (A), the areas west of the Scioto River (B) and near Lake Erie (C), the area north of the Maumee River (D), and the areas near the highest regional potentiometric levels (E) (fig. 30). These regional flow paths that bypass adjacent regional discharge areas cannot be determined from two-dimensional potentiometric-surface maps of the aquifer system. Such flow paths develop because regional recharge is available across most of the aquifer system; the result is a three-dimensional flow field.

The flow paths can also be used to identify areas where regional ground-water flow does not discharge to principal streams. One example is in the northeastern part of the modeled area (F, fig. 30). Many of the flow paths that discharge in this area are relatively short and may be associated with discharge to ditches and streams that are too small to be represented in the model by use of the stream cells or by means of evapotranspiration. A second example is the south-central part of the modeled area (G, fig. 30). These simulated flow paths likely indicate discharge to springs and seeps, which are common in the interbedded shales and limestones of the upper weathered zone water-bearing unit. They could also indicate discharge to small streams.

A ground-water divide is noted in the southeastern part of the modeled area (H, fig. 30). Water south of a certain point is diverted away from the Scioto River and discharges to the Ohio River.

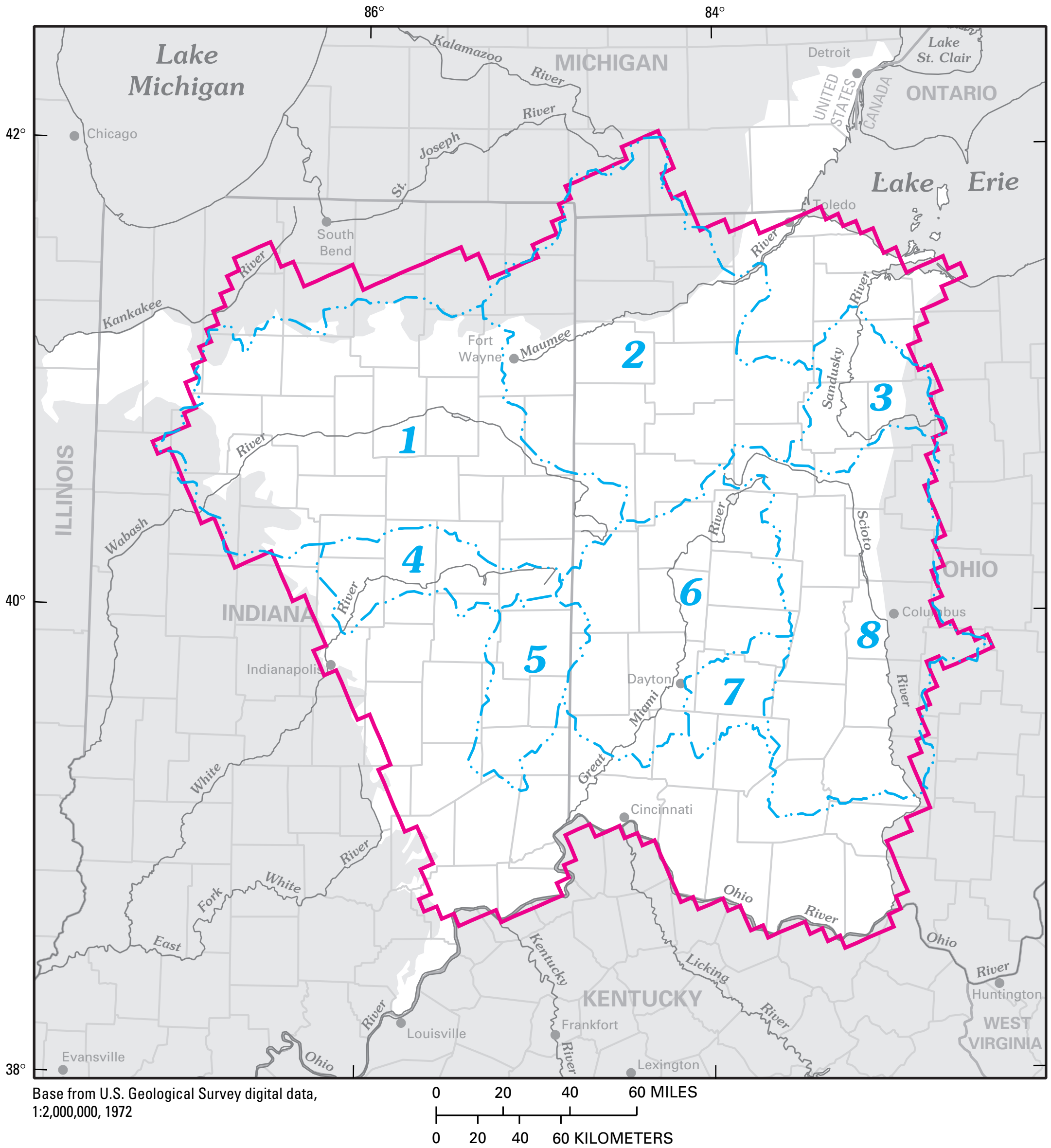
Simulated flow paths near Lake Erie indicate that the Lake diverts water away from some of the principal streams within the area (I, fig. 30). Flow paths near Lake Erie also indicate that recharge at the potentiometric high along the eastern boundary of the modeled area is likely to be the source of water that discharges in the area characterized by flowing wells and large springs east of the mouth of the Sandusky River (J, fig. 30; see also figs. 12 and 28).

Some of the simulated ground-water flow paths that terminate at Lake Erie are the longest within the modeled area (at nearly 50 mi). Simulated flow paths that begin just west of the Sandusky River and continue to Lake Erie are associated with a recharge area characterized by thin or absent glacial deposits. This recharge area has been previously referred to as the “limestone ridge area” and is the site of an exposed fossil coral reef (K, fig. 30). Other researchers have noted that this area of the carbonate-rock aquifer is vulnerable to contamination; nitrate contamination of ground water in the vicinity of this regional recharge area has been recognized since 1965 (Richards, 1990).

A comparison of regional ground-water flow paths and the position of the continental divide within the modeled area (the surface-water drainage divide that separates streams that flow toward the Atlantic Ocean from those that flow toward the Gulf of Mexico) illustrates the relation between surface-water and ground-water drainage basins (fig. 30). Ground-water divides associated with regional flow in the aquifer system are generally coincident with surface-water divides. Locally, however, deep regional ground-water flow paths can cross even major surface-water drainage divides (L, fig. 30). The amount of ground water that flows across the major surface-water drainage divides is likely to be a very small percentage of water that moves through the aquifer system because very few flow paths cross these divides.

Relative magnitudes (or volumes) of regional ground-water flow are not apparent from the map of regional ground-water flow patterns. Discharge vectors that illustrate the relative magnitude and resultant direction of horizontal regional flow within each cell of each model layer are shown in figure 31. The lengths of these discharge vectors are scaled linearly; units are feet cubed per day.

The greatest magnitudes of horizontal regional ground-water flow in the glacial deposits are associated with the most extensive outwash deposits. These outwash deposits tend to be concentrated along the principal streams within the modeled area. Notably large magnitudes of horizontal regional flow can be found in areas of outwash deposits within the Wabash River Basin. This drainage basin has the highest mean sustained ground-water discharge as percentage of mean ground-water discharge to the streams within the modeled area. In the northeastern part of the study area, magnitudes of horizontal regional flow in the glacial deposits are so small the vectors do not show up in figure 31*A*. Glacial deposits in this region are thin, locally absent, or poorly permeable. In addition, horizontal hydraulic gradients are fairly low. Such low magnitudes of horizontal regional flow are



Base from U.S. Geological Survey digital data, 1:2,000,000, 1972

EXPLANATION

8 — Boundary of surface-water drainage basin and basin number

Boundary of modeled area

Basin Number	Basin Name	Percent Regional Recharge Area	Percent Regional Discharge Area
1	Wabash River Basin	58	42
2	Maumee River Basin	44	56
3	Sandusky River Basin	48	52
4	White River Basin	59	41
5	Whitewater River Basin	63	37
6	Great Miami River Basin	62	38
7	Little Miami River Basin	67	33
8	Scioto River Basin	52	48

FIGURE 29.—Percentages of the surface area of selected surface-water drainage basins simulated as regional recharge and discharge areas.

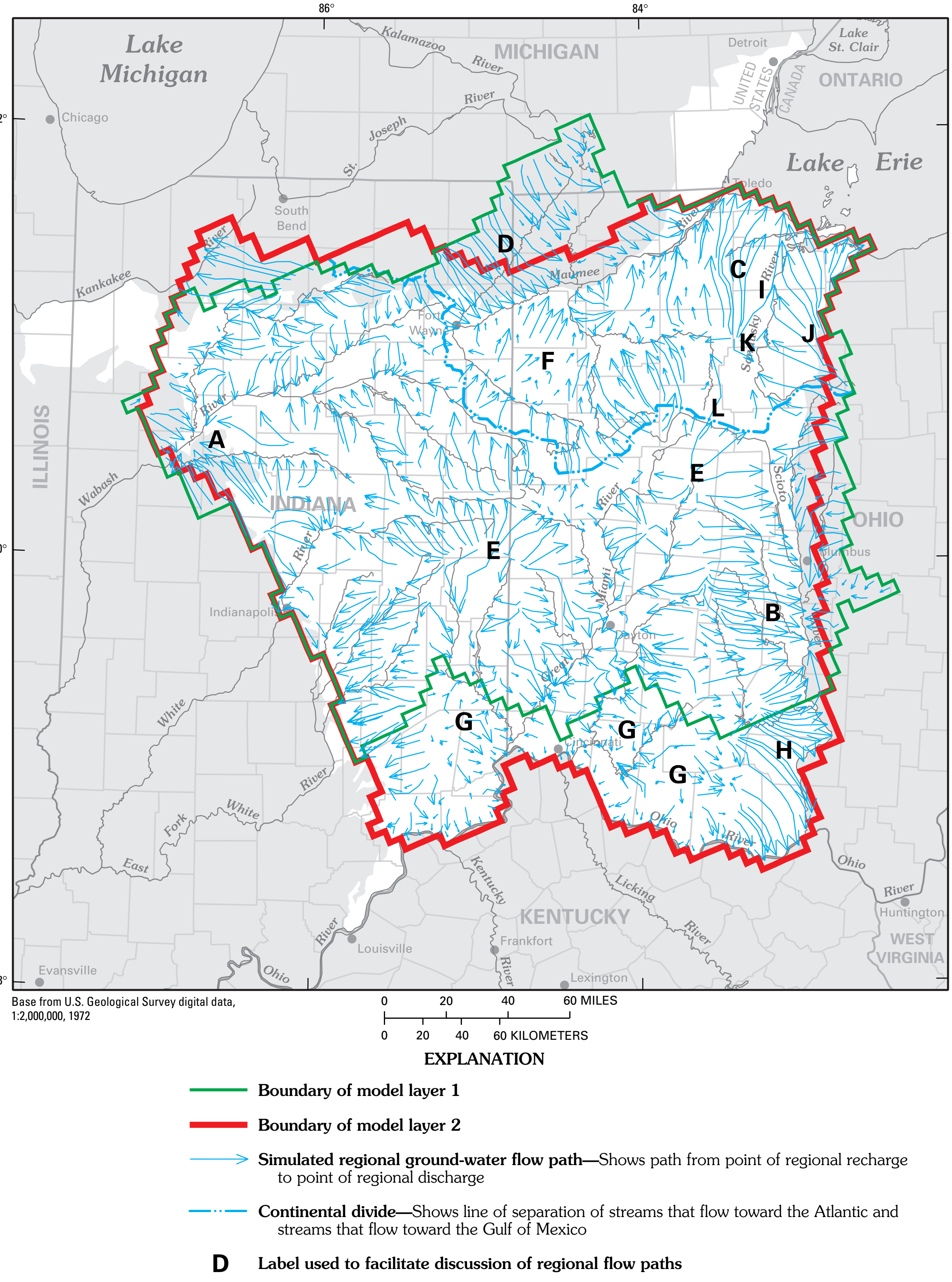


FIGURE 30.—Simulated patterns of advective regional flow in the Midwestern Basins and Arches aquifer system.

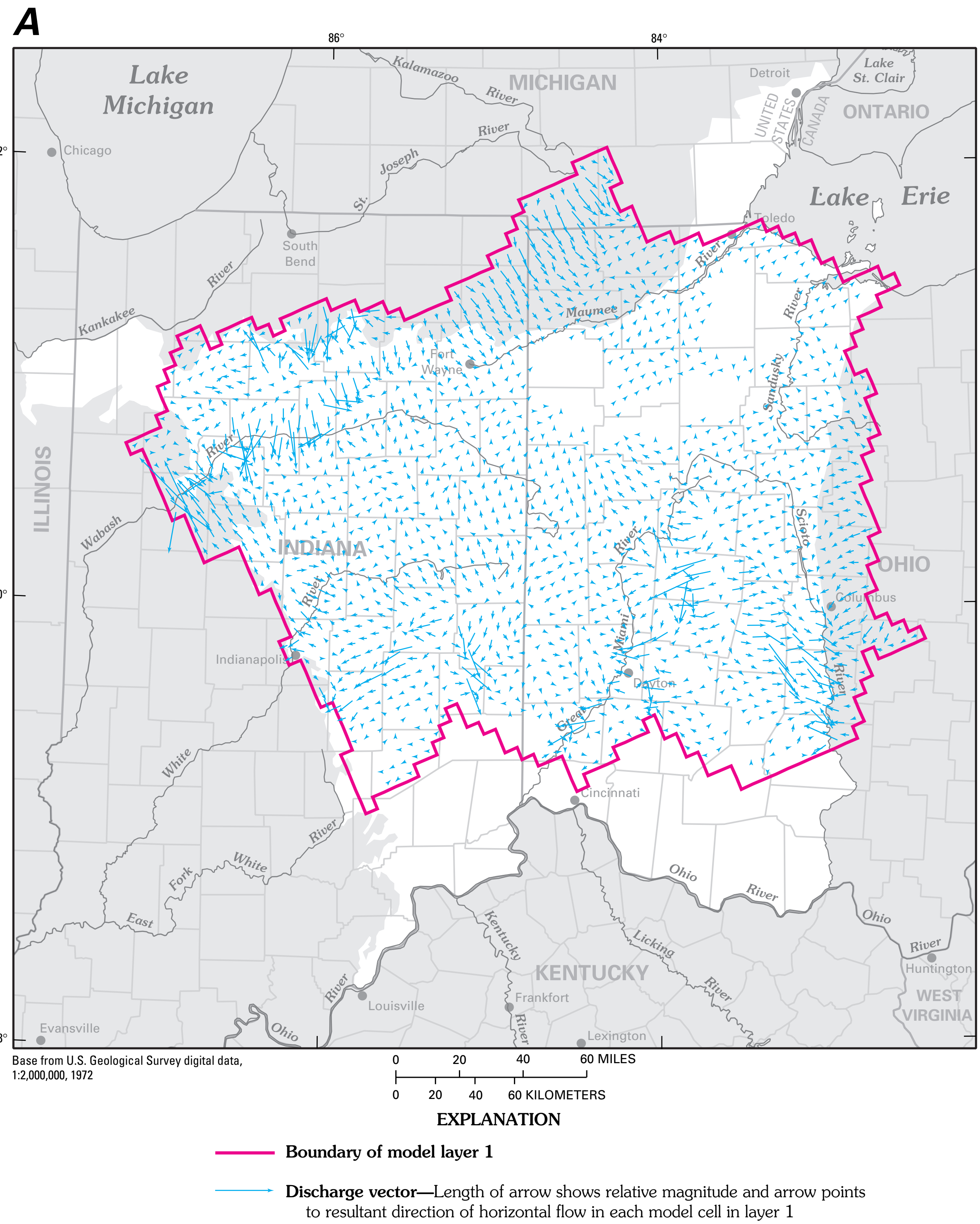


FIGURE 31.—Simulated relative magnitudes of horizontal regional flow in the Midwestern Basins and Arches aquifer system: (*A*) glacial deposits.

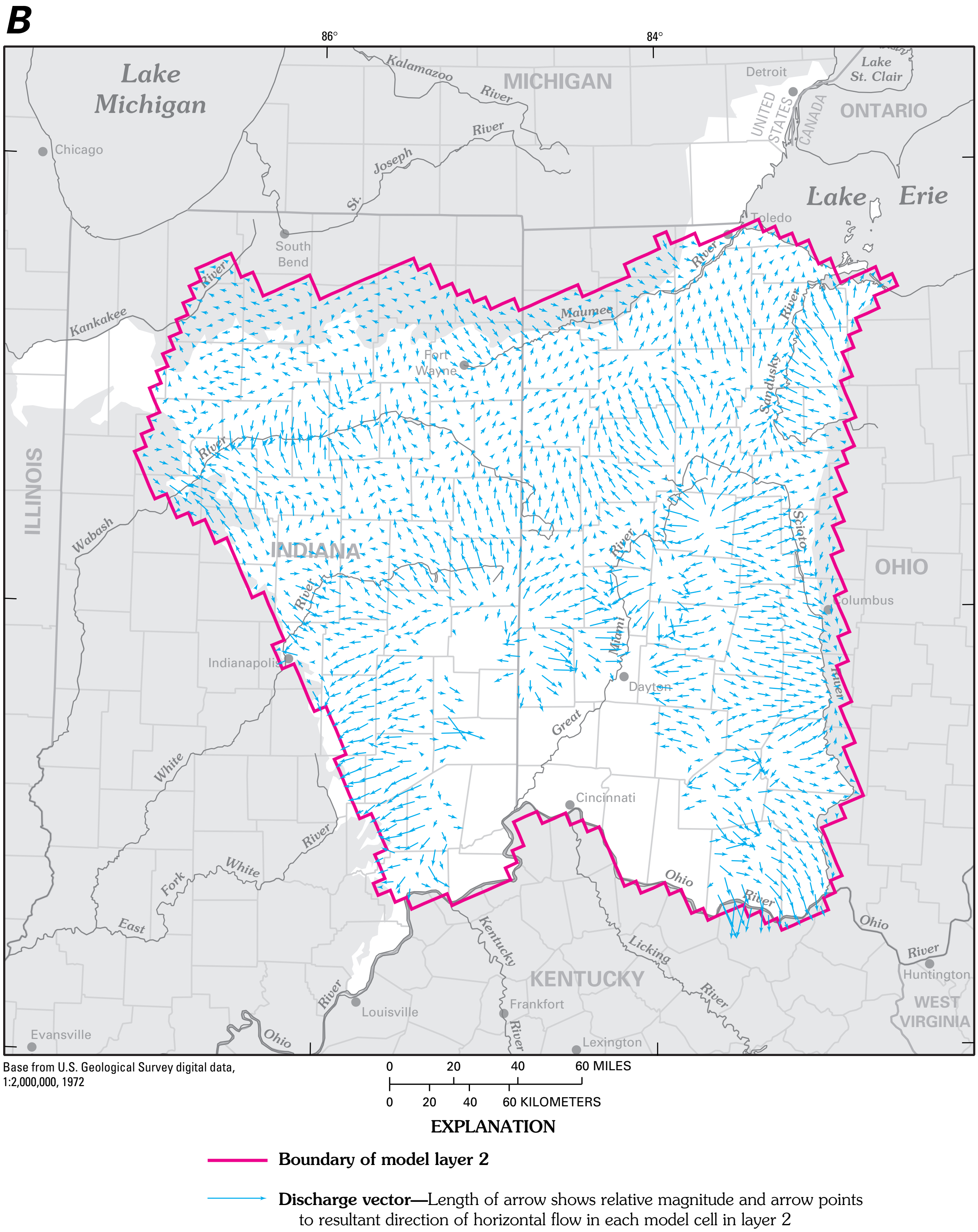


FIGURE 31. Continued—Simulated relative magnitudes of horizontal regional flow in the Midwestern Basins and Arches aquifer system: (*B*) bedrock.

consistent with the hypothesis that regional recharge to this part of the aquifer system could be limited by the inability of the aquifer system to carry ground water away from the area. This area is largely coincident with the area of weak regional discharge (fig. 28).

Simulated discharge vectors indicate high magnitudes of horizontal regional flow in the carbonate-rock aquifer in the areas around the regional potentiometric highs (fig. 31*B*). High magnitudes of horizontal regional flow are also associated with the downstream end of the Wabash and White Rivers, the margin of the Illinois (structural) Basin, the Ohio River, an area west of the Scioto River, and the area east of the Sandusky River. Discharge vectors along part of the Lake Erie shore indicate that the magnitude of horizontal regional flow in the carbonate-rock aquifer in this area is fairly small. Ground-water flow may be predominantly vertical in this area because it is an area of regional ground-water discharge. Simulated discharge vectors were computed for the upper weathered zone water-bearing unit, but the relative magnitudes of flow in this poorly permeable unit are so small that the vectors do not show up at the scale of figure 31*B*.

It should be noted that the discharge vectors show only relative magnitudes of horizontal regional ground-water flow and do not indicate flow velocities. Additional information on the effective porosity of the aquifers would be necessary to compute flow velocities. Appropriate effective-porosity data for fractured carbonate rock are difficult to obtain and were not available for this investigation. Ground-water ages presented in the following section, however, provide insight into ground-water residence times.

The calibrated final model was not used to simulate potential effects of future pumpage on regional ground-water flow in the aquifer system. Data on future pumpage needs at the regional scale are not available, and any simulations of future pumpage at this time would be contrived. It is noteworthy, however, that only a small percentage of current pumpage is associated with the regional flow systems explicitly simulated with this model. Therefore, more water associated with such regional flow systems almost certainly could be used. The quality of the ground water associated with some parts of the aquifer system, however, may limit its use.

GEOCHEMISTRY

Geochemical data were collected from the Midwestern Basins and Arches aquifer system to investigate the relations among ground-water chemistry, aquifer mineralogy, and present and past patterns of regional flow. The data include a synthesis of basic data from more than 1,300 ground-water analyses of water samples from the aquifer system, as well as detailed chemical and isotopic analyses of ground water and aquifer material along general directions of regional flow. The analyses represent two hydrologic units (table 1) within the Midwestern Basins and Arches aquifer system — aquifers within glacial deposits and the carbonate-rock aquifer — and were obtained from records in the U.S. Geological Survey's National Water Information System (NWIS) data base; files of the Indiana Department of Natural Resources, the Ohio Department of Natural Resources, and the Ohio Environmental Protection Agency; various published reports; and samples collected as part of this investigation. The data were compiled and analyzed to investigate the ground-water chemistry of the aquifer system on a regional scale. Ground-water chemistry of subregional areas of the Midwestern Basins and Arches aquifer system is described in the following reports: in Ohio, by Ohio Department of Natural Resources, Division of Water (1970), Norris and Fidler (1973), Norris (1974), Deering and others (1983), Breen and Dumouchelle (1991); and in Indiana, by Geosciences Research Associates, Inc. and Purdue University, Water Resources Research Center (1980) and Indiana Department of Natural Resources (1988, 1990). Analyses of brines from rocks of Silurian and Devonian age are found in Stout and others (1932), Lamborn (1952), Walker (1959), Stith (1979), Keller (1983) and Wilson and Long (1993a, b).

Data compiled from the literature and the available data bases were selected on the basis of the following criteria: (1) major-ion concentrations (Ca, Mg, Na, Cl, SO_4, and HCO_3) were determined, (2) the analyses balanced electrochemically within 10 percent and, (3) lithologies of the water-producing units were determined. In cases where multiple analyses were available for a well, the most recent analysis that met the above criteria was selected. The dissolved-solids data for most of the analyses that were used in this report were calculated by summing the concentrations of all major constituents according to the method described in Fishman and Friedman (1989). Dissolved-solids concentrations for waters in the Illinois and Michigan Basins were estimated from borehole geophysical data where available laboratory determinations were sparse (D.J. Schnoebelen, U.S. Geological Survey, written commun., 1993).

New data that were collected during this investigation include detailed chemical and isotopic analyses of ground water from the aquifer system along general directions of regional ground-water flow, as determined from the map of the potentiometric surface of the carbonate-rock aquifer (fig. 12), and isotopic analyses of aquifer material collected from cores of glacial deposits and carbonate rock. The locations of the ground-water and aquifer-material samples are shown in figure 32. At each sampling location along four transects across the aquifer system, ground-water samples were collected from the carbonate-rock aquifer, and, where possible, from a glacial aquifer. Sampling was restricted to existing domestic wells or test wells; wells with short open intervals in the deep parts of the aquifer were generally not available. At each sampling location, an attempt was made to sample the deepest available well in the carbonate-rock aquifer in