Anthony Robinson, who helped with many aspects of the work presented herein.

GEOHYDROLOGY

The area of principal hydrologic interest of the Midwestern Basins and Arches RASA project encompasses approximately 44,000 mi^2, most of which is in the Midwestern Basins and Arches Region as defined in Shaver (1985). Boundaries of this study area (fig. 1) are coincident with the contact between Devonian limestones and younger Devonian shales (fig. 2) or surface-water bodies.

GEOLOGIC SETTING

The Midwestern Basins and Arches aquifer system generally lies between the Appalachian, the Illinois, and the Michigan (structural) Basins and is located along the axes of the Cincinnati, the Findlay, and the Kankakee Arches in parts of Indiana, Ohio, Michigan, and Illinois (fig. 2). The sedimentary rocks within the area range in age from Precambrian through Mississippian; however, bedrock units of primary interest range in age from Ordovician (Cincinnatian) through Lower Mississippian (table 1). The oldest bedrock units exposed at the bedrock surface are generally found along the axis of the Cincinnati Arch in the south-central part of the study area, owing to several periods of erosion (figs. 2 and 3). In general, units exposed at the bedrock surface are progressively younger with distance from the axes of the arches. Four faults or fault zones partially dissect these sedimentary rocks within the region (fig. 2).

The bedrock units of Ordovician age (Cincinnatian) consist of interbedded shales and limestones. Shales predominate in these units; less than one-quarter of the sequence is made up of limestones (Gray, 1972). This sequence of interbedded shales and limestones thickens eastward from the western border of Indiana toward Ohio and is overlain by carbonate rocks (limestones and dolomites) of Silurian and Devonian age. These carbonate rocks locally contain some evaporite deposits in northwestern Ohio and northern Indiana (French and Rooney, 1969; Janssens, 1977); they contain sulfide minerals in an area associated with the Findlay Arch (Botoman and Stieglitz, 1978). The carbonate rocks of Silurian and Devonian age range in thickness from 0 ft at the contact with the rocks of Ordovician age to 2,500 ft in southeastern Michigan (Casey, 1994) (fig. 4). Erosion has resulted in the loss of hundreds of feet of carbonate rock from across the central part of the study area. The carbonate-rock sequence has been completely eroded in places by the ancient Teays-Mahomet River system, described in Melhorn and Kempton (1991). As a result of this erosion, the older shales and limestones of Ordovician age are present at bedrock surface in sinuous exposures north of their principal area of exposure (fig. 2). The carbonate rocks are overlain by shales of Devonian and Mississippian age along the margins of the structural basins. Erosion has resulted in the loss of the shale sequence throughout the central part of the study area except for an area approximately 50 mi northwest of Columbus, Ohio (fig. 2). This shale outlier is referred to herein as the "Bellefontaine Outlier."

The bedrock is overlain by Quaternary glacial deposits throughout most of the study area (fig. 5 and table 1). These deposits directly overlie the carbonate rocks in the central part of the area (subcrop area of the carbonate rocks) and overlie the younger shales along the margins of the structural basins. Glacial deposits mask the ancient bedrock topography and bury numerous valleys in the bedrock surface.

The Quaternary glacial deposits—the result of multiple glacial advances—range in age from Kansan (oldest) to Wisconsinan (youngest) (Bennison, 1978). The deposits of Kansan and Illinoian age are not widespread within the study area and typically are present beyond the limit of the Wisconsinan ice sheet (fig. 5). The Kansan and Illinoian deposits are also thinner than the more widespread deposits of Wisconsinan age (Goldthwait and others, 1965; Geosciences Research Associates, 1982; Soller, 1986). The Wisconsinan ice sheet eroded much of these earlier glacial deposits; this resulted in landforms that contain material from multiple glacial advances. The resultant geomorphology is illustrated in figure 6; a photograph of a shaded relief generated from digital topographic data for every 30 seconds of latitude and longitude (U.S. Geological Survey, 1987).

The glacial deposits include ground- and end-moraine deposits, glaciolacustrine deposits, and outwash deposits (fig. 5); ice-contact stratified drift is present within the moraine deposits. The glacial deposits range in thickness from 0 to approximately 400 ft (fig. 7) (Mozola, 1969, 1970; Fleck, 1980; Gray, 1983; Soller, 1986). The areas dominated by ground- and end-moraine deposits are characterized by broad, low ridges with smooth, gentle slopes separated by flat, gently undulating plains (Mickelson and others, 1983). End moraines are close together where they abut highlands, such as the Bellefontaine Outlier (Young and others, 1985). The mineral composition of the moraines reflects local bedrock; about 4 percent of the material in Ohio was transported from the Canadian Shield north of the study area (Strobel and Faure, 1987).

Surficial glaciolacustrine deposits are present in the lowlands adjacent to Lake Michigan and Lake Erie and are the result of glacial lakes that formed along the margins of the retreating Wisconsinan ice (Young and others, 1985) (figs. 5 and 6). These glaciolacustrine deposits are dominated by lake bottom silts and clays. Minor sands and gravels mark the beaches of ancient shorelines (Goldthwait and others, 1965); some lakebed sands in Michigan just west of Lake Erie have been mapped (Western Michigan University, Department of Geology, 1981).

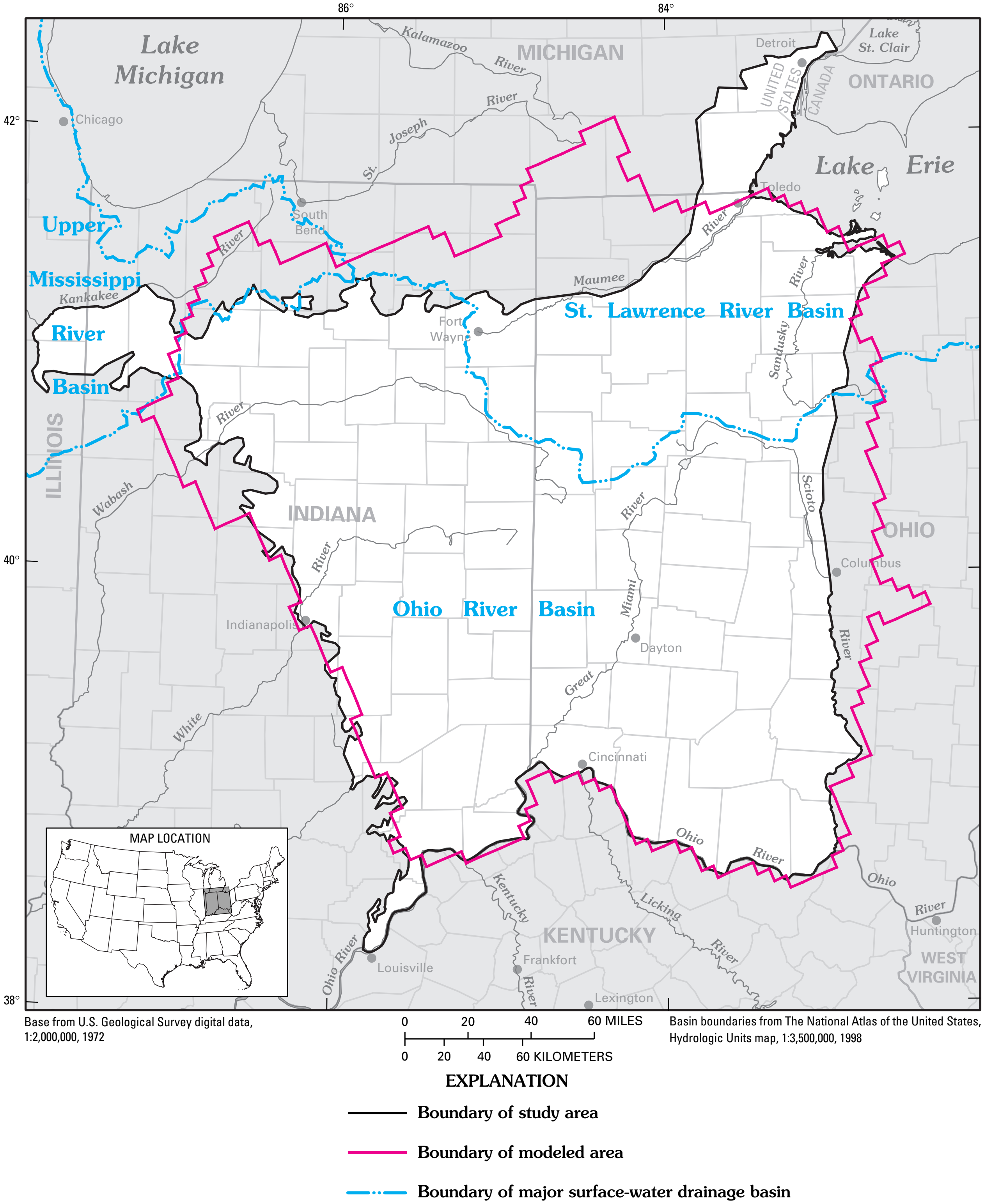


FIGURE 1.—Location of study and modeled areas.

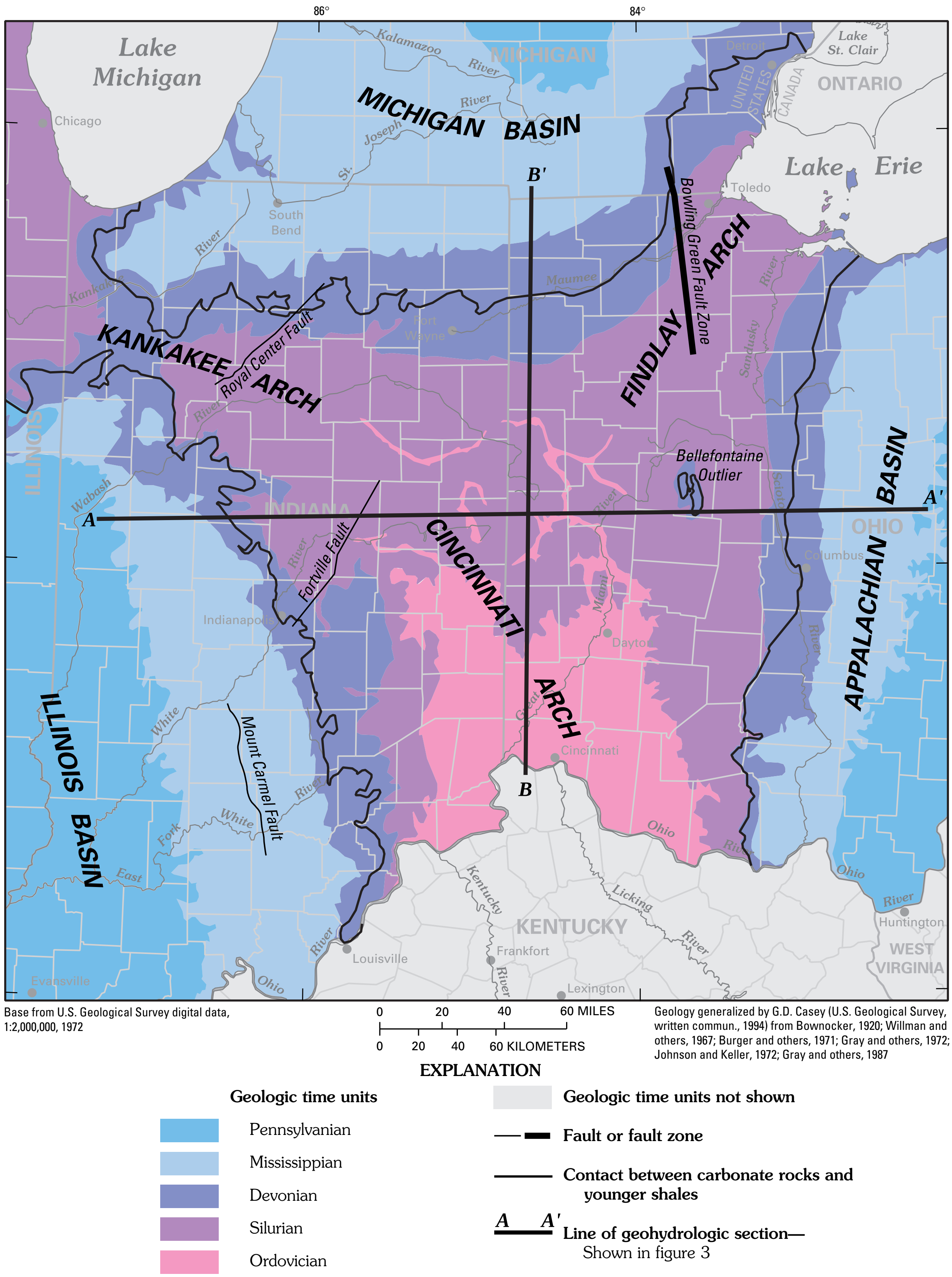


FIGURE 2.—Generalized bedrock geology and location of geohydrologic sections *A–A′* and *B–B′*.

TABLE 1.—*Relation between geologic and hydrologic units of the Midwestern Basins and Arches aquifer system and model layers used to simulate regional flow within the aquifer system*

[Fm, Formation; Ls, Limestone; equiv, equivalent; modified from Casey, 1992, fig. 3]

Generalized Geologic Units								Generalized Hydrologic Units (North / South)	Model Layers
SYSTEM	**SERIES**	**Southeastern Indiana**	**Northwestern to east-central Indiana**	**Northern Indiana**	**Northwestern Ohio**	**Central-western Ohio**	**Southwestern Ohio**		
QUATERNARY	PLEISTOCENE	GLACIAL DEPOSITS						GLACIAL AQUIFERS AND CONFINING UNITS	LAYER 1
MISSISSIPPIAN	LOWER	Borden Group 1	Borden Group 1	Coldwater Shale 2				UPPER CONFINING UNIT	
DEVONIAN	UPPER	New Albany Shale	New Albany Shale Ellsworth Shale Antrim Shale	Ellsworth Shale Antrim Shale	Bedford Shale Antrim Shale	Ohio Shale Olentangy Shale	Bedford Shale Ohio Shale Olentangy Shale		
DEVONIAN	MIDDLE	Muscatatuck Group	Muscatatuck Group	Traverse Formation Detroit River Formation Muscatatuck Group	Traverse Formation Dundee Ls. 3 4 Detroit River Formation	Delaware Limestone Columbus Limestone Detroit River Group		CARBONATE-ROCK AQUIFER	LAYER 2
DEVONIAN	LOWER						Hillsboro Sandstone		
SILURIAN	CAYUGAN	Salina[5] Group	Salina[5] Group	Salina[5] Group	Salina Group	Salina Group	Salina Group		
SILURIAN	NIAGARAN	Salamonie[5] Dolomite	Salamonie[5] Dolomite Cataract Fm.	Salamonie Dolomite Cataract Fm.	Lockport Dolomite 7	Lockport Dolomite Guelph[6] Dolomite Goat Island[6] Dolomite Gasport[6] Dolomite 7	Lockport Dolomite Rochester Shale equiv. Dayton Limestone		
SILURIAN	ALEXANDRIAN	Brassfield Limestone	Sexton Creek[5] Brassfield Limestone Cataract Fm.	Brassfield Limestone Cataract Fm.	8 Brassfield Limestone Cataract Formation	8 Brassfield Limestone Cataract Formation	Brassfield Limestone	9	
ORDOVICIAN	CINCINNATIAN	Maquoketa Group Kope Formation	Maquoketa Group	Maquoketa Group	Undifferentiated Cincinnatian rocks	Undifferentiated Cincinnatian rocks	Undifferentiated Cincinnatian rocks	BASAL CONFINING UNIT	
ORDOVICIAN	MIDDLE	Trenton Limestone Kope Formation	Trenton Limestone	Trenton Limestone	Trenton Limestone	Trenton Limestone	Trenton Limestone		

[1] Rockford Shale
[2] Sunbury Shale
[3] Delaware Limestone
[4] Columbus Limestone
[5] Follows usage of the Indiana Geological Survey (Gray and others, 1985)

Nondeposition or erosion
Intervals not included in investigation

[6] Follows usage of the Ohio Geological Survey (Hull, 1990; Larsen, 1991)
[7] Rochester Shale equivalent
[8] Dayton Limestone
[9] Upper weathered zone water-bearing unit (not an aquifer)

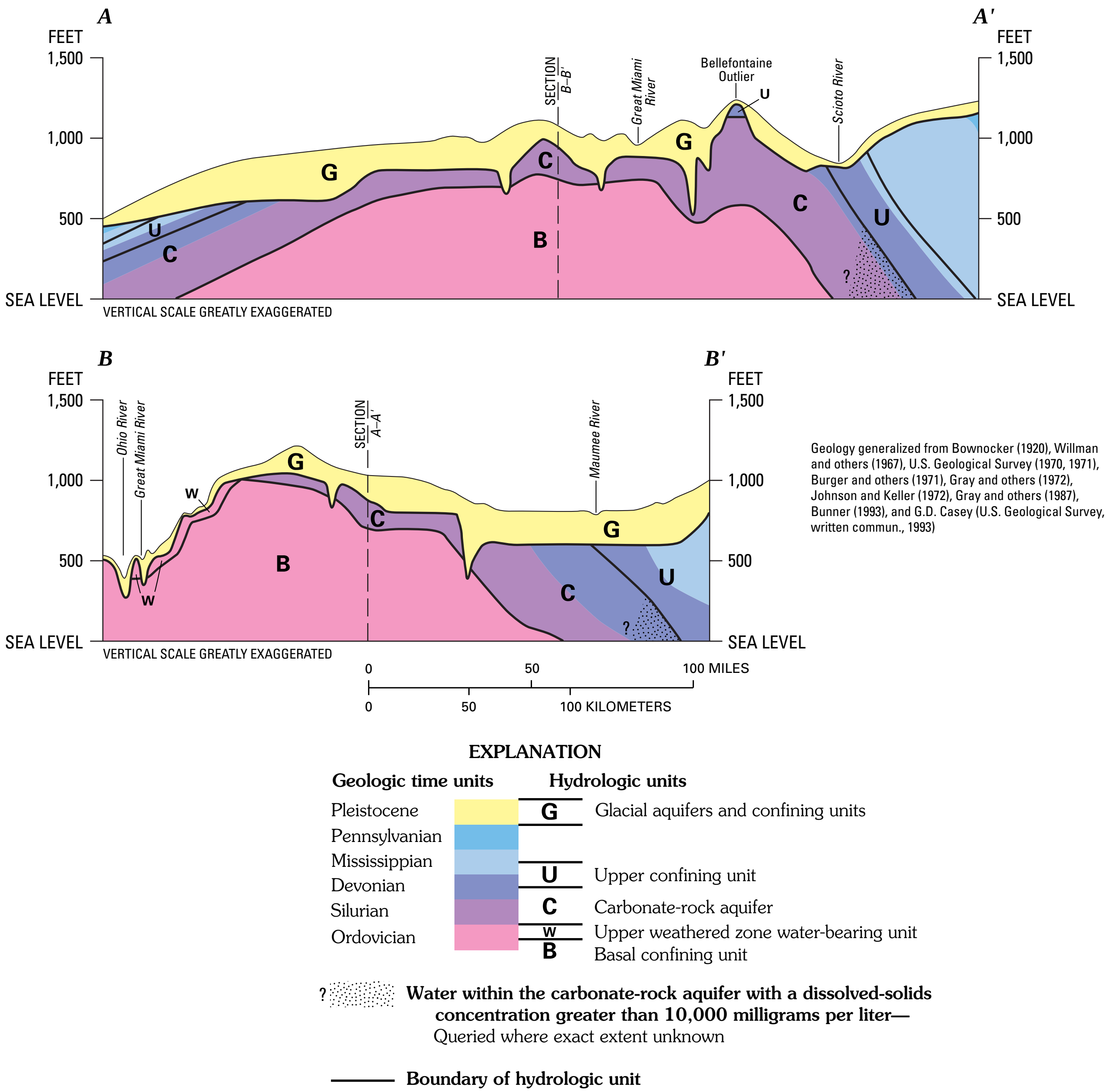
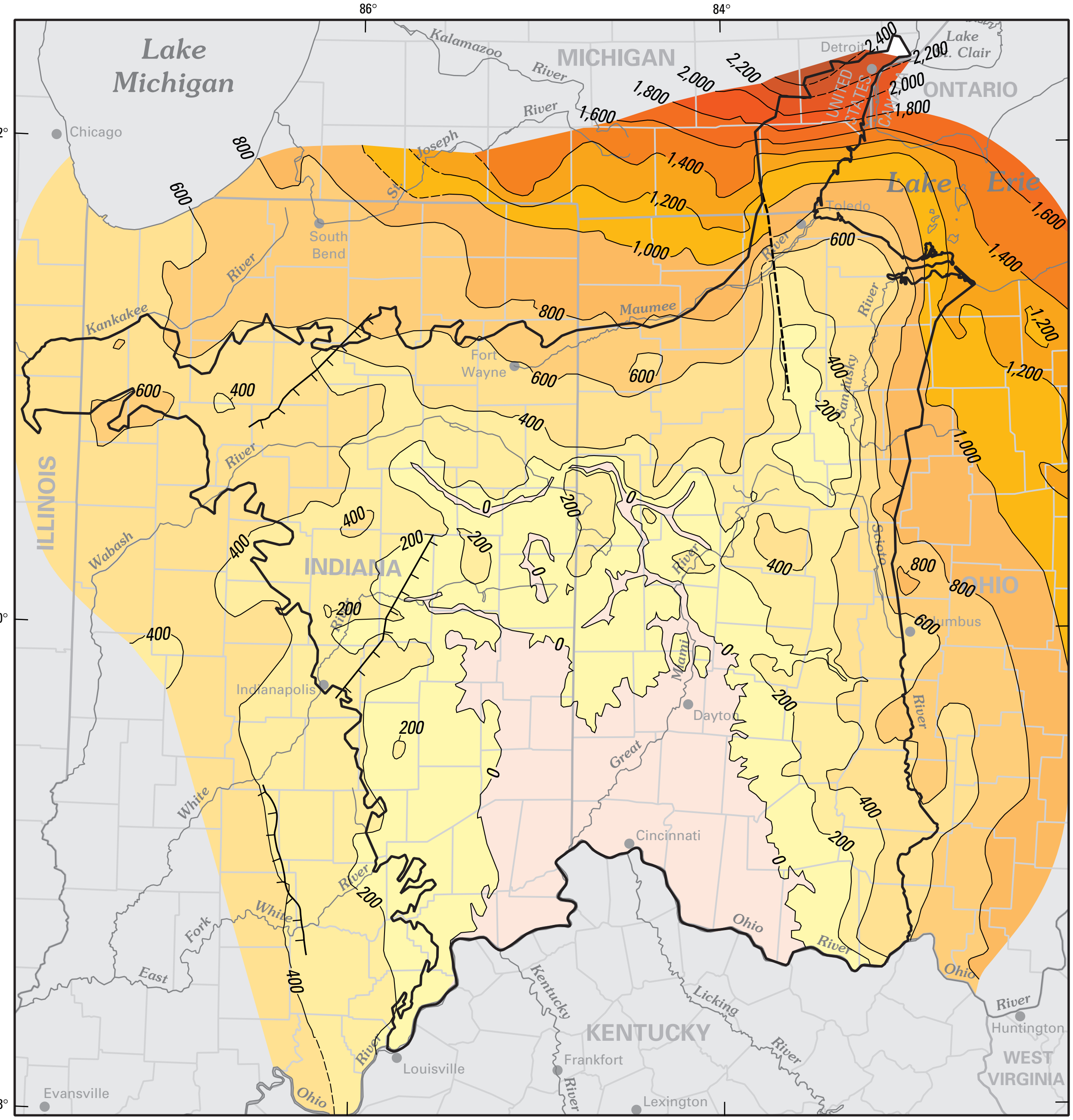


FIGURE 3.—Generalized sections *A–A′* and *B–B′* showing geologic and hydrologic units that compose the Midwestern Basins and Arches aquifer system (faults not shown; lines of section shown in fig. 2).



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

Contours from Casey, 1994

0 20 40 60 MILES

0 20 40 60 KILOMETERS

EXPLANATION

0, 200, 400, 600, 800, 1,000, 1,200, 1,400, 1,600, 1,800, 2,000, 2,200, 2,400 **Thickness of carbonate rocks**—Lines of equal thickness dashed where inferred. Interval 200 feet. Darker colors indicate greater thickness

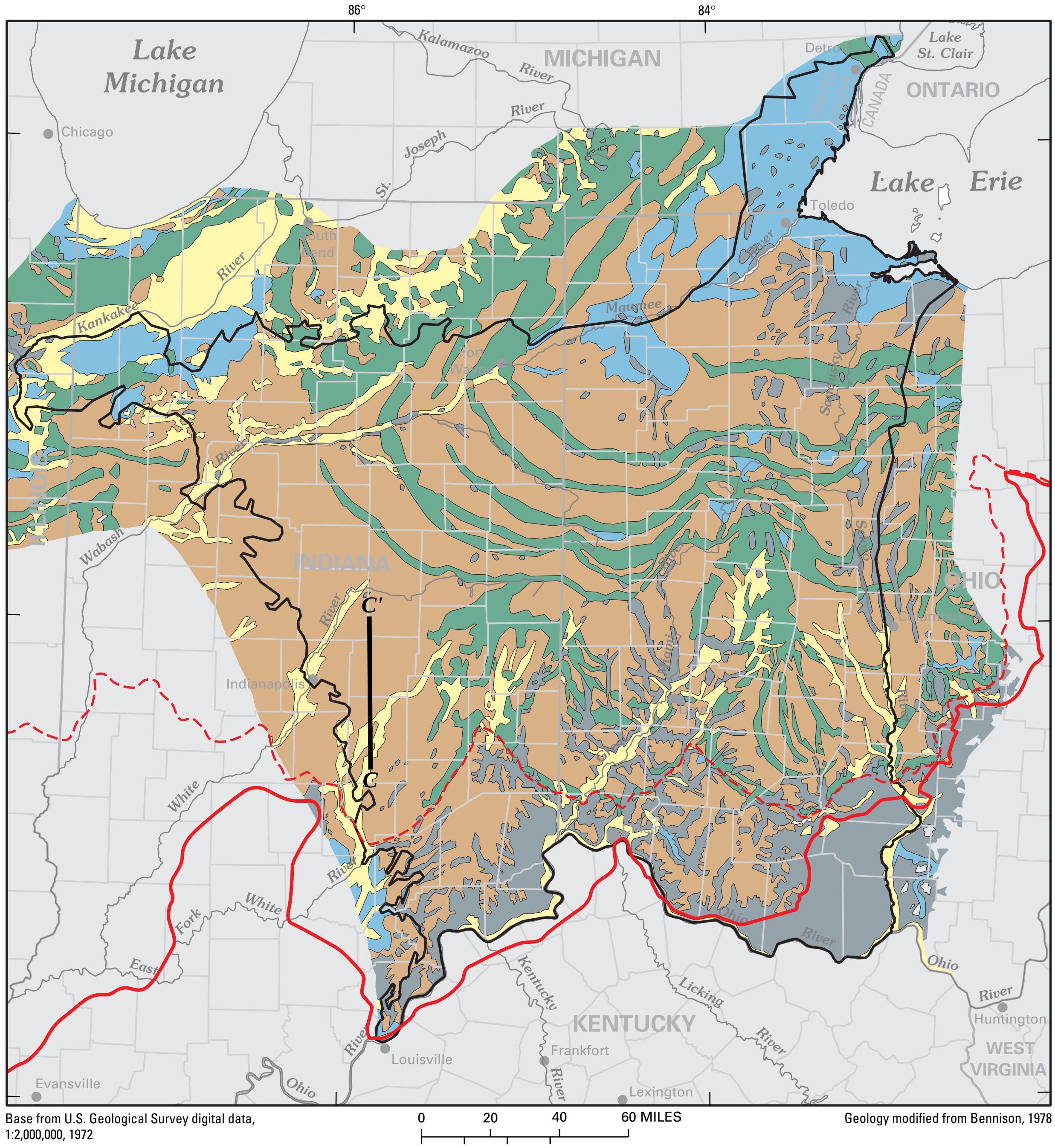
Rocks of Ordovician age—Upper weathered zone water-bearing unit. Not an aquifer. Silurian and Devonian carbonate rocks absent

Fault zone

Fault—Hachures on downthrown side

Boundary of study area

FIGURE 4.—Thickness of carbonate rocks of Silurian and Devonian age in the Midwestern Basins and Arches Region.



EXPLANATION

Glacial deposits

- Ground-moraine deposits
- End-moraine deposits
- Glaciolacustrine deposits
- Outwash deposits
- Bedrock

Limit of Wisconsinan ice sheet (Flint, 1959)

Limit of all glacial advances (Flint, 1959)

Boundary of study area

C *C′* **Line of geologic section**—Shown in figure 8

FIGURE 5.—Generalized surficial geology and location of geologic section *C*—*C′*.

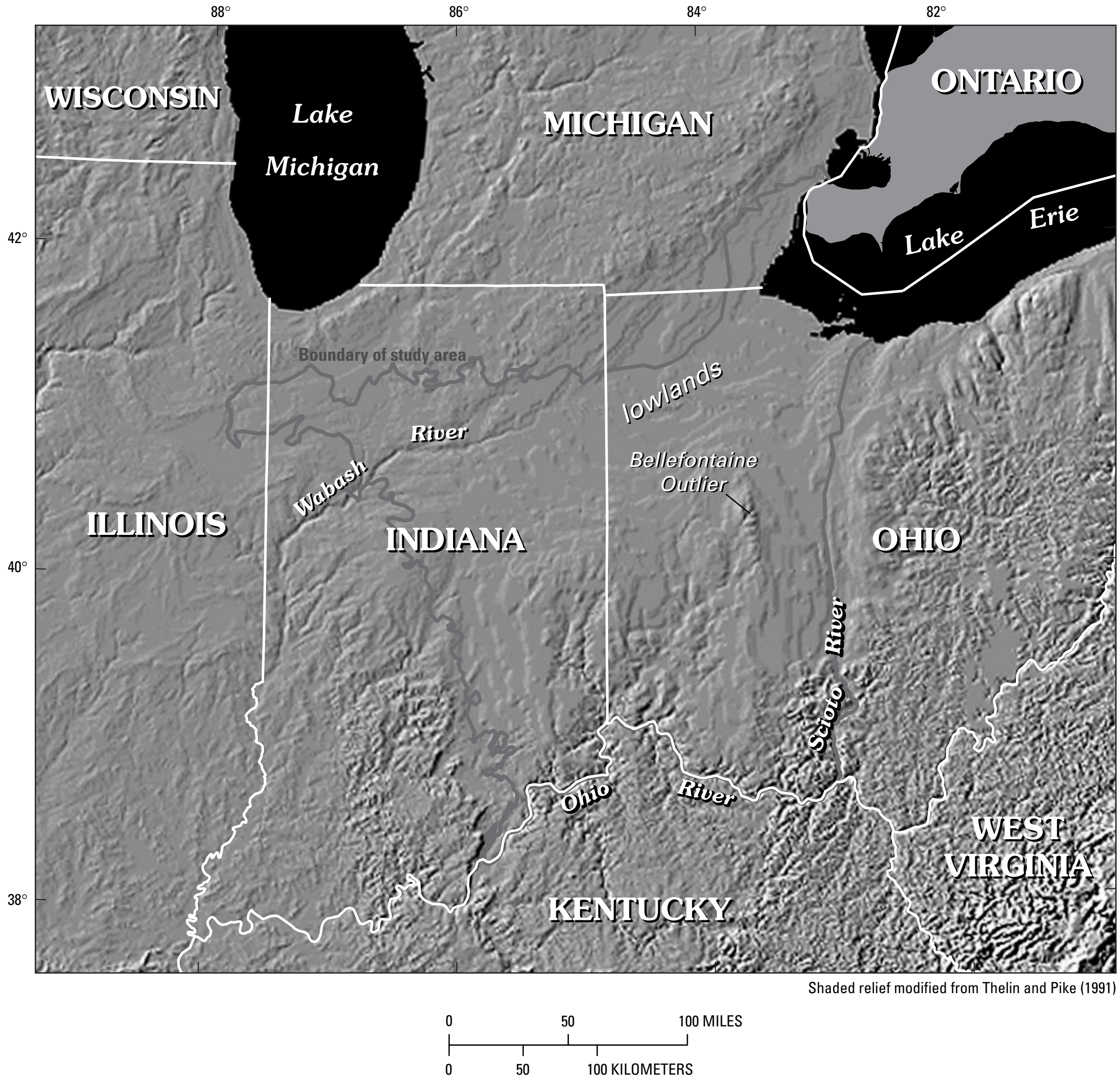


FIGURE 6.—Shaded relief of land-surface topography (geomorphology) in the Midwestern Basins and Arches Region, from digital data.

Outwash deposits commonly fill the ancient drainage systems, which served as channels for the deposition of such washed and sorted material. In many places, these outwash deposits underlie principal streams that currently drain the area.

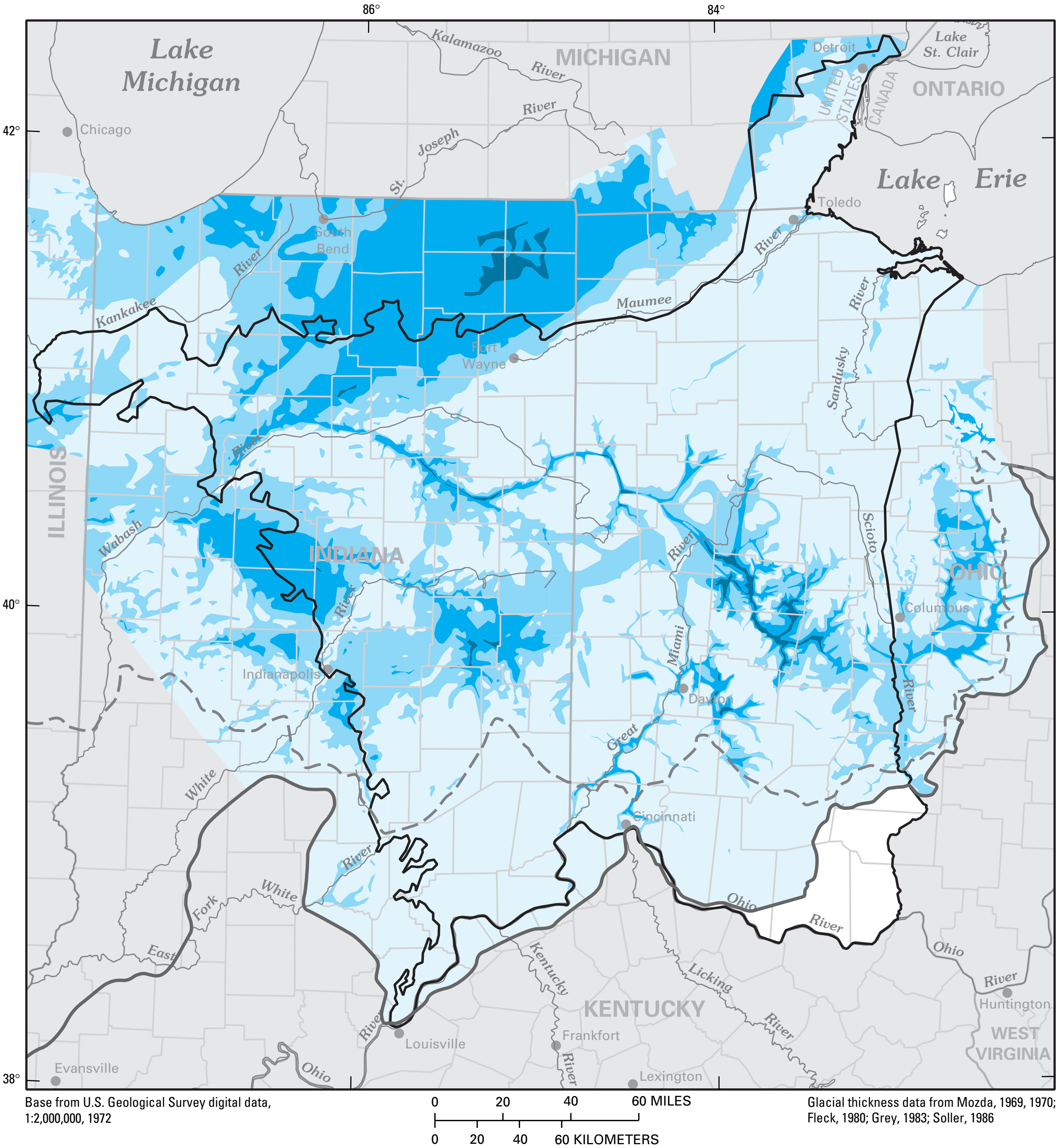
HYDROLOGIC SETTING

The study area has a distinctly seasonal humid temperate climate. Precipitation in Indiana is greatest from March through July (Glatfelter and others, 1991). The wettest months in Ohio tend to be April through August, whereas February and October tend to be the driest (Sherwood and others, 1991). Mean annual precipitation computed from stations with at least 50 years of data ranges from 33 to 43 in. across the study area (E.F. Bugliosi, U.S. Geological Survey, written commun., 1993). Approximately 26 in/yr are consumed by evapotranspiration in Indiana (Clark, 1980). Todd (1969) notes that potential evapotranspiration exceeded precipitation from mid-May through mid-September over a 30-year period in southwestern Ohio.

Parts of three major river systems—the Ohio, the St. Lawrence, and the Upper Mississippi—drain the study area (fig. 1).

GROUND-WATER USE

Ground water is plentiful throughout much of the study area and serves as an important resource. Approximately 433 Mgal/d of ground water was reported to have been withdrawn from the Midwestern Basins and Arches aquifer system in



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

Glacial thickness data from Mozda, 1969, 1970; Fleck, 1980; Grey, 1983; Soller, 1986

EXPLANATION

Thickness of glacial deposits—In feet

- 0 – 100
- 101 – 200
- 201 – 400
- Greater than 400

Limit of Wisconsinan ice sheet (Flint, 1959)

Limit of all glacial advances (Flint, 1959)

Boundary of study area

FIGURE 7.—Generalized thickness of glacial deposits in the Midwestern Basins and Arches Region.

Indiana and Ohio during the 1990 calendar year (Beary, 1993). Only 15 percent (67 Mgal/d) of this water was withdrawn from the carbonate rocks. Of the remaining 85 percent, much of the water was withdrawn from outwash deposits that underlie principal streams (E.A. Beary, U.S. Geological Survey, written commun., 1993). These pumpage figures reflect only ground-water withdrawals reported by users capable of pumping 100,000 gal/d or greater, and not all of this withdrawn water is consumed. Regardless of the actual amount of pumpage from the aquifer system, the system is not heavily stressed at the regional scale, as is apparent when the pumpage figures are compared to the amount of ground water that discharges to streams within the study area. Eberts (1999) estimates that, over a period of long-term steady-state conditions in the aquifer system, greater than 13,000 Mgal/d discharges from the aquifer system to streams within the study area.

No regional-scale cones of depression are present within the aquifer system. At the subregional scale, irrigation pumpage in northwestern Indiana results in seasonal water-level declines in the carbonate rocks. On an annual basis, however, the carbonate-rock aquifer appears to be able to support high-capacity irrigation pumpage without significant long-term depletion (Indiana Department of Natural Resources, 1990).

AQUIFERS AND CONFINING UNITS

The water table within the Midwestern Basins and Arches aquifer system generally is within Quaternary alluvium or glacial deposits. Glacial aquifers typically consist of sands and gravels that compose outwash deposits (fig. 5) or discontinuous lenses of ice-contact stratified drift within ground- and end-moraine deposits (fig. 8). These aquifers are most commonly unconfined where the outwash deposits are present along principal streams and are locally semiconfined or confined by clayey till elsewhere in the region. Because the glacial aquifers are not normally extensive, individual aquifers can supply large yields of ground water only locally (Ohio Department of Natural Resources, Division of Water, 1970).

The shale sequence of Mississippian and Devonian age functions as a confining unit. Specifically, the shale sequence restricts the flow of ground water between the glacial aquifers and the underlying carbonate-rock aquifer along the margins of the structural basins (fig. 3). In this report, these shales are referred to as the “upper confining unit.”

The carbonate-rock aquifer directly underlies the upper confining unit along the margins of the structural basins and underlies the glacial deposits, which collectively function as a semiconfining unit, within the central part of the study area.

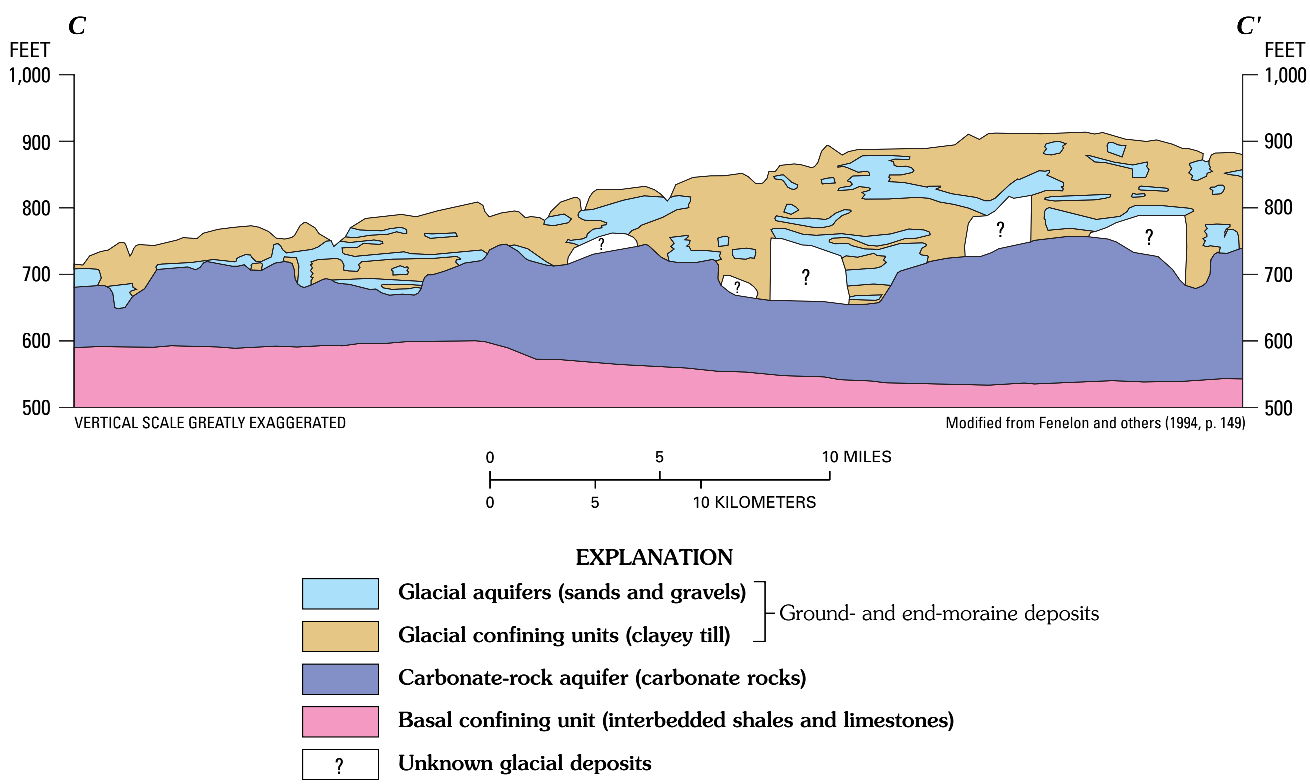


FIGURE 8.—Generalized hydrologic section *C–C′* showing typical relation between glacial aquifers and glacial confining units in the Midwestern Basins and Arches aquifer system (line of section shown in fig. 5).

The carbonate-rock aquifer is unconfined in areas where it is locally exposed at the land surface. Lateral boundaries of the carbonate-rock aquifer generally are coincident with the occurrence of waters that have a dissolved solids concentration of 10,000 mg/L or greater (see fig. 34, p. C65) or where the aquifer pinches out in the south-central part of the study area (fig. 4).

The carbonate-rock aquifer is confined below by a basal confining unit that is composed of interbedded shales and limestones of Ordovician age. Gupta (1993) demonstrates that these shales significantly limit the flow of ground water into or out of the bottom of the carbonate-rock aquifer but that some water moves across the basal confining unit of the Midwestern Basins and Arches aquifer system to become part of an even larger aquifer system. There is some evidence that the bottom of the carbonate-rock aquifer may actually be within the Silurian and Devonian carbonate rocks in some areas. Arihood (1994) notes that fractures in the carbonate rocks in a few fully penetrating wells in northwest Indiana are not productive in the bottom 60 to 400 ft of the wells.

Although the Ordovician rocks (basal confining unit) underlie the carbonate-rock aquifer throughout most of the study area, they are laterally contiguous with the aquifer along the axis of the Cincinnati Arch in the south-central part of the area. The contact between Silurian and Ordovician rocks, where it is exposed at the bedrock surface, has been described throughout the literature as a spring horizon. The role of the shales along this contact was summarized succinctly by Norris and others (1950, p. 23): “The chief importance of the impervious Ordovician shale with respect to ground water is that it deflects the water to the surface as springs.” These interbedded shales and limestones, however, are used locally as a source of water in the south-central part of the study area where they are exposed at the bedrock surface and other aquifers are absent. Weathering has increased secondary porosity and permeability within this area and has allowed water circulation to increase at shallow depths. Yields from wells completed in the interbedded shales and limestones within this area are typically less than 10 gal/min, drawdowns commonly are extreme, and dry holes are common (Indiana Department of Natural Resources, 1988). In this report, the interbedded shales and limestones of Ordovician age that are exposed at the bedrock surface are referred to as the “upper weathered zone water-bearing unit.” The upper weathered zone water-bearing unit is not considered to be an aquifer but may be hydraulically connected to the carbonate-rock aquifer.

GROUND WATER

The occurrence and flow of ground water in the Midwestern Basins and Arches aquifer system are controlled by the geohydrologic framework of the aquifer system and by the distribution and rate of recharge and discharge. Recharge and discharge also affect the long-term availability of ground water.

HYDRAULIC CHARACTERISITCS

The productivity of the glacial aquifers varies spatially within small distances because of variations in the composition, continuity, and structure of the deposits (Strobel, 1993). On the basis of data from 101 aquifer tests, transmissivities of the glacial aquifers within the study area range from 300 to 69,700 ft^2/d (fig. 9 and table 2). Storage coefficients for the same material range from 0.00002 to 0.38 (Joseph and Eberts, 1994). Transmissivities at two wells within the study area that are completed in clayey till (not considered to be aquifer material) are 1.5 and 2.1 ft^2/d (Strobel, 1993).

On the basis of available aquifer-test data, the vertical hydraulic conductivity at wells completed in glacial deposits within the study area ranges from 0.0001 to 0.77 ft/d (Norris, 1959, 1979, 1986; Fleming, 1989; Strobel, 1993). Strobel (1993) notes that clayey till within the study area may be fractured at shallow depths as a result of desiccation, biological action, oxidation of minerals, or isostatic rebound after the retreat of the last ice sheet. He observed fractures in till within the study area to depths of 15 ft, and he suggests that the intersection of such fractures with one another and with sand and gravel lenses within the clayey till can result in vertical hydraulic conductivities greater than those commonly considered restrictive to ground-water flow.

Vertical and horizontal hydraulic conductivities of the shales of Mississippian and Devonian age range from 10^{-7} to 10^{-5} ft/d, as determined from laboratory analysis of core samples (G.D. Casey, U.S. Geological Survey, written commun., 1993). Because these values do not account for fractures in the shales, effective hydraulic conductivities that represent field conditions may be orders of magnitude larger (Freeze and Cherry, 1979, p. 158).

Water in the carbonate-rock aquifer is primarily present in fractures, joints, bedding planes, and solution channels within the rock. These openings are due, in part, to the effects of weathering during the period of geologic history when the carbonate-rock aquifer was exposed at the land surface before glaciation (Ohio Department of Natural Resources, Division of Water, 1970). The productivity of the aquifer varies with the concentration of openings within the rock, which seldom approach conditions associated with karst terranes. These openings are interconnected on an areal basis. Previous researchers (Ohio Department of Natural Resources, Division of Water, 1970) noted that the hydraulic characteristics of the carbonate-rock aquifer approach those of a regionally homogeneous medium as the study area increases. Arihood (1994) also notes that aquifer tests that create drawdown cones over several miles depict the carbonate-rock aquifer as an equivalent porous medium. On the basis of data from 171 aquifer

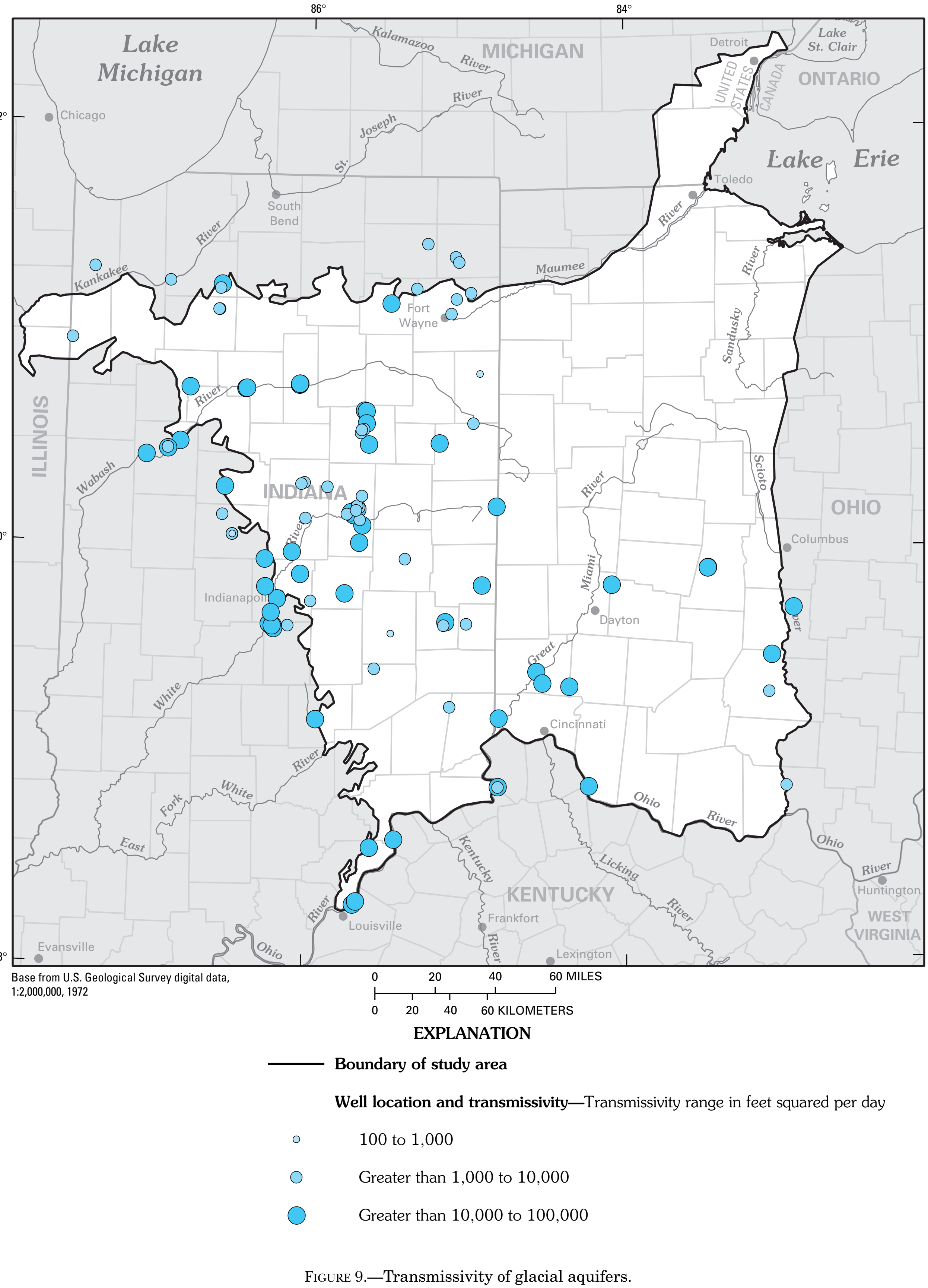


FIGURE 9.—Transmissivity of glacial aquifers.

TABLE 2.—*Summary of hydraulic characteristics of aquifers and confining units in the Midwestern Basins and Arches aquifer system*

[--, data not available; ft/d, feet per day; ft^2/d, feet squared per day]

Aquifer or confining unit	Range of horizontal hydraulic conductivities (ft/d)	Range of vertical hydraulic conductivities (ft/d)	Range of transmissivities (ft^2/d)	Range of storage coefficients
Glacial aquifers	--	[b]0.0001 – 0.77	[a]300 – 69,700	[a]0.00002 – 0.38
Glacial confining units	--		[c]1.5, 2.1	
Upper confining unit	[d]10^{-7} – 10^{-5}		--	--
Carbonate-rock aquifer	--	--	[a,e]70 – 52,000	[a]0.00001 – 0.05
Upper weathered zone water-bearing unit	[f]0.0016 – 12	--	--	--
Basal confining unit	[g]10^{-7} – 10^{-5}		--	--

[a]Joseph and Eberts (1994).
[b]Norris (1959, 1979, 1986), Fleming (1989), Strobel (1993).
[c]Strobel (1993).
[d]G.D. Casey (U.S. Geological Survey, written commun., 1993). From laboratory analyses of core samples.
[e]Geometric mean, 1,912 ft^2/d.
[f]Dumouchelle (1992).
[g]Lawrence Wickstrom (Ohio Geological Survey, written commun., 1991). From laboratory analyses of core samples.

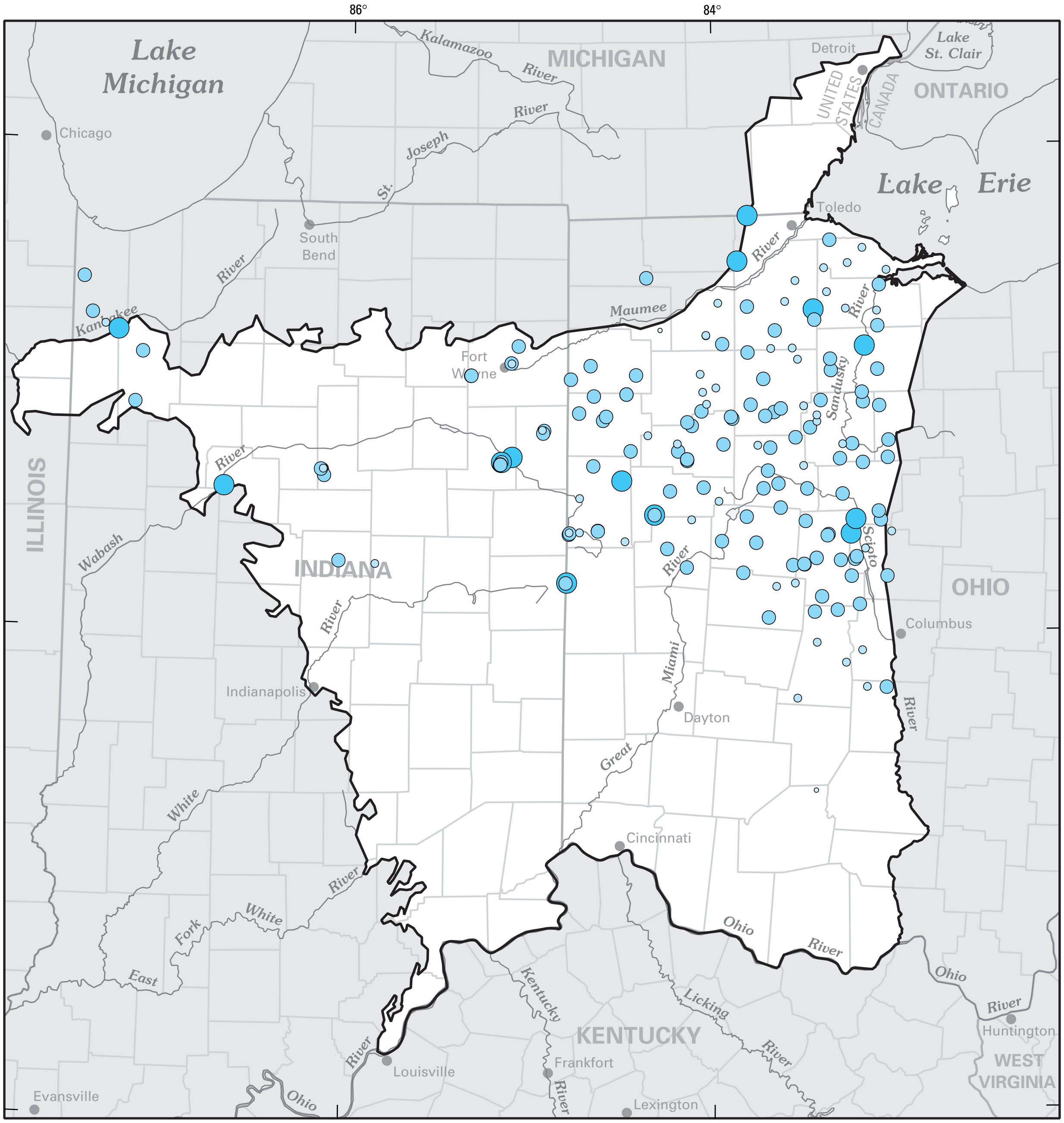
tests, transmissivities of the carbonate-rock aquifer range from 70 to 52,000 ft^2/d (fig. 10). These data were tested for normality by use of the Shapiro-Wilk test and were found to follow a lognormal distribution; they have a geometric mean of 1,912 ft^2/d. Storage coefficients range from 0.00001 to 0.05 (Joseph and Eberts, 1994). No pumped-well test data are available for estimation of vertical-hydraulic conductivities of the carbonate-rock aquifer.

Very little information is available to describe the hydraulic characteristics of the interbedded shales and limestones that function as a basal confining unit to the Midwestern Basins and Arches aquifer system. Analyses of core collected from the upper part of the interbedded shale and limestone sequence in southwestern Ohio provide estimates of vertical and horizontal hydraulic conductivities that range from 10^{-7} to 10^{-5} ft/d (Lawrence Wickstrom, Ohio Geological Survey, written commun., 1991). These values do not account for secondary porosity within the rocks. Shales of Ordovician age, however, are considered favorable for underground storage of liquefied natural gas in southern Indiana, where they underlie the carbonate-rock aquifer, because of their low hydraulic conductivities (Droste and Vitaliano, 1976); thus, these shales can be assumed to be very restrictive to ground-water flow in this area. Hydraulic conductivities are likely to be higher in the upper weathered zone water-bearing unit where the shales have been exposed at the land surface. On the basis of slug-test data at four wells completed in the upper part of the interbedded shale and limestone sequence near the upper weathered zone water-bearing unit, horizontal hydraulic conductivities range from 0.0016 to 12 ft/d (Dumouchelle, 1992).

LEVELS

Review of historical ground-water-level data indicates a long-term steady-state condition in the aquifer system. (Long-term steady-state conditions refer to a state of dynamic equilibrium in which no net change in storage in the aquifer system occurs over a long-term period. The long-term period referred to herein is a minimum of 10 years and includes wet and dry periods.) Extensive ditching to drain swampland in low-lying areas in northwestern Ohio (Kaatz, 1955) and northwestern Indiana in the late 1800's and early 1900's resulted in some dewatering of shallow glacial deposits (5 to 7 ft) in Indiana (Rosenshein, 1963; Indiana Department of Natural Resources, 1990) and possibly similar dewatering in Ohio; however, a new equilibrium has been established in these areas (Eberts, 1999). Annual ground-water-level fluctuations related to ground-water recharge range from 3 to 7 ft in the aquifer system (Clark, 1980; Shindel and others, 1991a, b). Annual high water levels are reached between March and June, and annual low water levels are reached near the end of the growing season.

The altitude of the water table, which typically is in glacial deposits, is a controlling factor for regional flow in the Midwestern Basins and Arches aquifer system. Most regional variation in water-table altitude is a consequence of the variation in land-surface altitude, and depth to the water table varies predictably at the regional scale. The specific relation between land-surface altitude and water levels in glacial deposits was determined by use of a least-squares method of linear regression (Eberts, 1999). Depth to water is greatest in topographically high areas and decreases in areas such as stream valleys. A composite regional potentiometric-surface map of the glacial deposits was constructed from water levels



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

0 20 40 60 MILES
0 20 40 60 KILOMETERS

EXPLANATION

Boundary of study area

Well location and transmissivity—Transmissivity range in feet squared per day

1 to 100

Greater than 100 to 1,000

Greater than 1,000 to 10,000

Greater than 10,000 to 100,000

FIGURE 10.—Transmissivity of the carbonate-rock aquifer.

reported on drillers' logs and is shown in figure 11. Also shown in figure 11 is the lack of available drillers' logs for most areas south of the limit of the Wisconsinan ice sheet. Perhaps so few domestic wells have been completed in glacial deposits within these areas because the units are not productive.

The regional potentiometric surface in the carbonate-rock aquifer is a subdued reflection of the land surface and further illustrates the effect of variations in land-surface altitude on the aquifer system. A regional potentiometric-surface map of the carbonate-rock aquifer (fig. 12) was constructed from water levels synoptically measured during July 1990 (Eberts, 1999). Potentiometric highs are in west-central Ohio near the Bellefontaine Outlier and near the southern limit of the carbonate-rock aquifer along the border between Indiana and Ohio. Potentiometric lows less than or equal to 600 ft are along the Wabash and the Ohio Rivers and Lake Erie.

RECHARGE

Ground-water recharge at the water table of the Midwestern Basins and Arches aquifer system is primarily from infiltration of precipitation and the associated flow away from the water table within the saturated zone. Recharge from precipitation at the water table varies seasonally because evapotranspiration, which can intercept infiltrating precipitation, varies seasonally. Recharge to the carbonate-rock aquifer is primarily from percolation of ground water through overlying units. The rate at which the carbonate-rock aquifer is recharged by percolation depends on the permeability and thickness of the overlying deposits and the difference between the water table in the overlying deposits and the potentiometric surface in the carbonate-rock aquifer.

Very few estimates of recharge to the Midwestern Basins and Arches aquifer system have been made. Daniels and others (1991) estimated recharge rates through unsaturated glacial till to be 1.4 and 1.8 in/yr from a tritium profile obtained from a core collected in Indiana, but they stated that such rates are more applicable to a local scale than a regional scale. Walton and Scudder (1960) report recharge rates of 12 in/yr through outwash deposits and 8 in/yr through glacial till on uplands within parts of the Great Miami River Basin. All other available recharge estimates, excluding those determined by use of previously constructed numerical ground-water flow models, are for recharge to the carbonate-rock aquifer. These estimated recharge rates, based on analyses of flow nets and on cones of influence of pumped wells, range from 0.14 to 6.3 in/yr (Rosenshein, 1963; Watkins and Rosenshein, 1963; Rowland and Kunkle, 1970; Cravens and others, 1990; Roadcap and others, 1993).

DISCHARGE

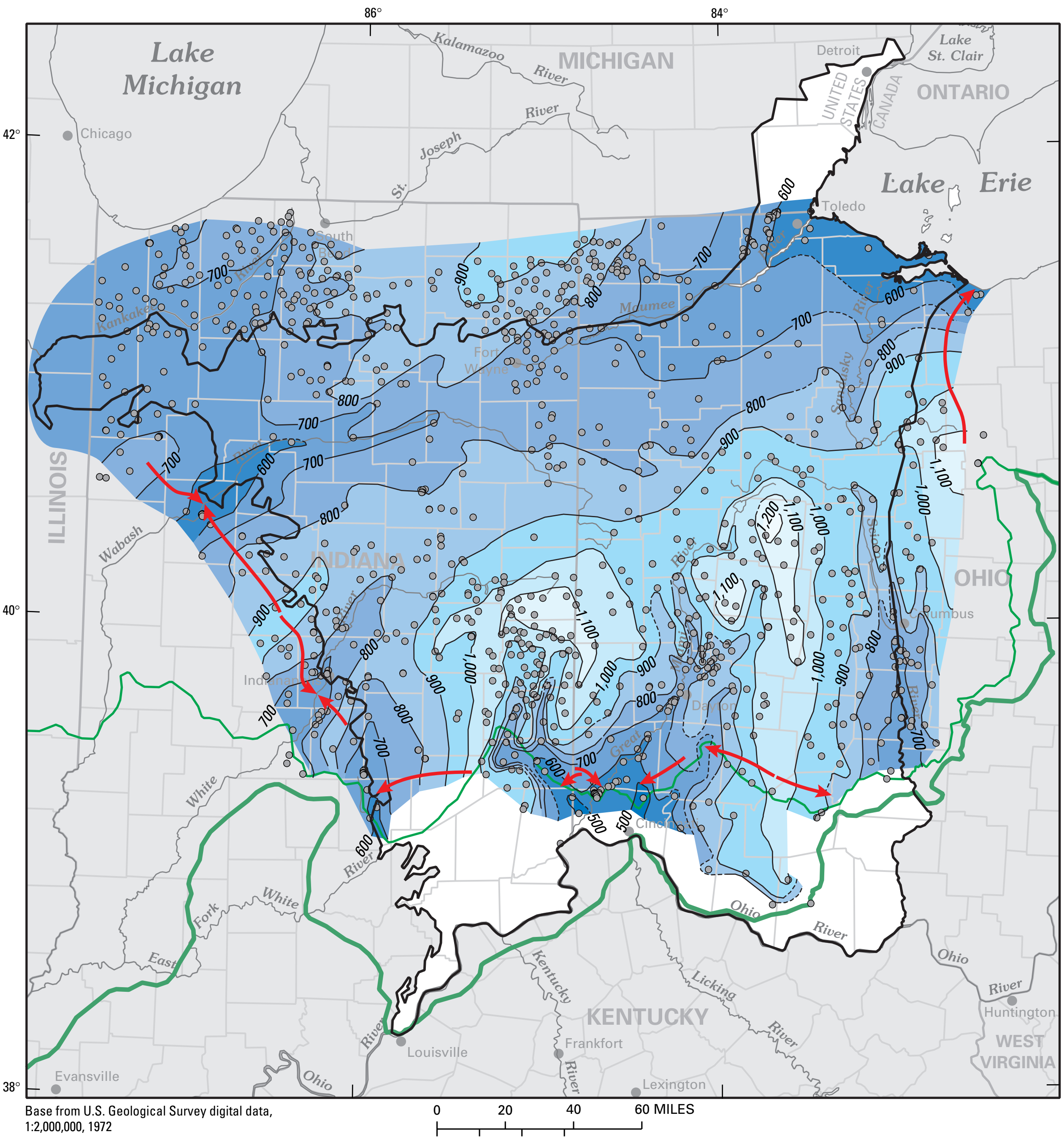
Ground-water discharge from the aquifer system includes discharge to streams, ditches, lakes, and wetlands and the removal of water from the saturated zone by evapotranspiration and pumping. Ground water discharges to a stream if the water table or potentiometric surface is above the stage of the stream, whereas the stream loses water to the aquifers if the water table is below the stream stage. Ground-water discharge to streams (base flow) can be estimated from streamflow data by separating streamflow hydrographs into direct-runoff and base-flow components.

Streamflow data were used to estimate ground-water discharge from the aquifer system to streams that drain the study area. Daily mean base flows for the period associated with unregulated or only minimally regulated low flow were computed for selected streamflow-gaging stations by means of the local-minimum method of hydrograph separation (Pettyjohn and Henning, 1979). A computer program (R.A. Sloto, U.S. Geological Survey, written commun., 1988) was used in this investigation to automate the local-minimum method of hydrograph separation; use of a computer program ensured that the separation technique was applied consistently.

Mean ground-water discharge to stream reaches above each selected streamflow-gaging station was estimated from the daily mean base flows. This was accomplished for each station by computing the average of all the daily mean base flows for the period selected for analysis (Eberts, 1999). Mean ground-water discharge to stream reaches between streamflow-gaging stations was estimated by computing the difference between mean ground-water discharge estimates for adjacent stations (fig. 13). These means describe the central tendency of ground-water discharge to the selected streams within the study area for long-term steady-state conditions in the aquifer system.

Mean ground-water discharge ranges from 17 to 80 percent of mean streamflow for the 43 selected stream reaches for which streamflow data were analyzed. (These values are the upper numbers in figure 14.) Mean ground-water discharge as a percentage of mean streamflow increases with distance downstream in about half of the principal surface-water drainage basins. Stated another way, ground water generally makes up a greater proportion of streamflow at the bottom of these drainage basins than in areas higher up in the basins. Unusually large percentages of mean ground-water discharge occur along stream reaches that drain areas underlain by large amounts of outwash deposits (figs. 5 and 14). In contrast, a notable decrease in mean ground-water discharge as a percentage of mean streamflow with distance downstream occurs where the Maumee River drains an area underlain by glaciolacustrine deposits (figs. 5 and 14). A decrease in mean ground-water discharge as a percentage of mean streamflow also occurs in the south-central part of the study area where the streams flow over areas where the carbonate-rock aquifer is absent (figs. 4 and 14). Cross (1949) also reported a relation between geology and base flow in Ohio.

Fluctuations of base flow in streams within the study area result from changes in hydraulic gradients in the Midwestern



EXPLANATION

1,200
1,100
1,000
900
800
700
600
500

Potentiometric contour—Line of equal water-level altitude. Dashed where glacial deposits are absent. Contour interval 100 feet. Datum is sea level. Darker blues indicate lower altitude

Boundary of study area

Limit of Wisconsinan ice sheet (Flint, 1959)

Limit of all glacial advances (Flint, 1959)

Flowline—Used to define boundaries of regional ground-water flow model

Observation well

FIGURE 11.—Composite regional potentiometric surface in glacial deposits constructed from water levels on drillers' logs.

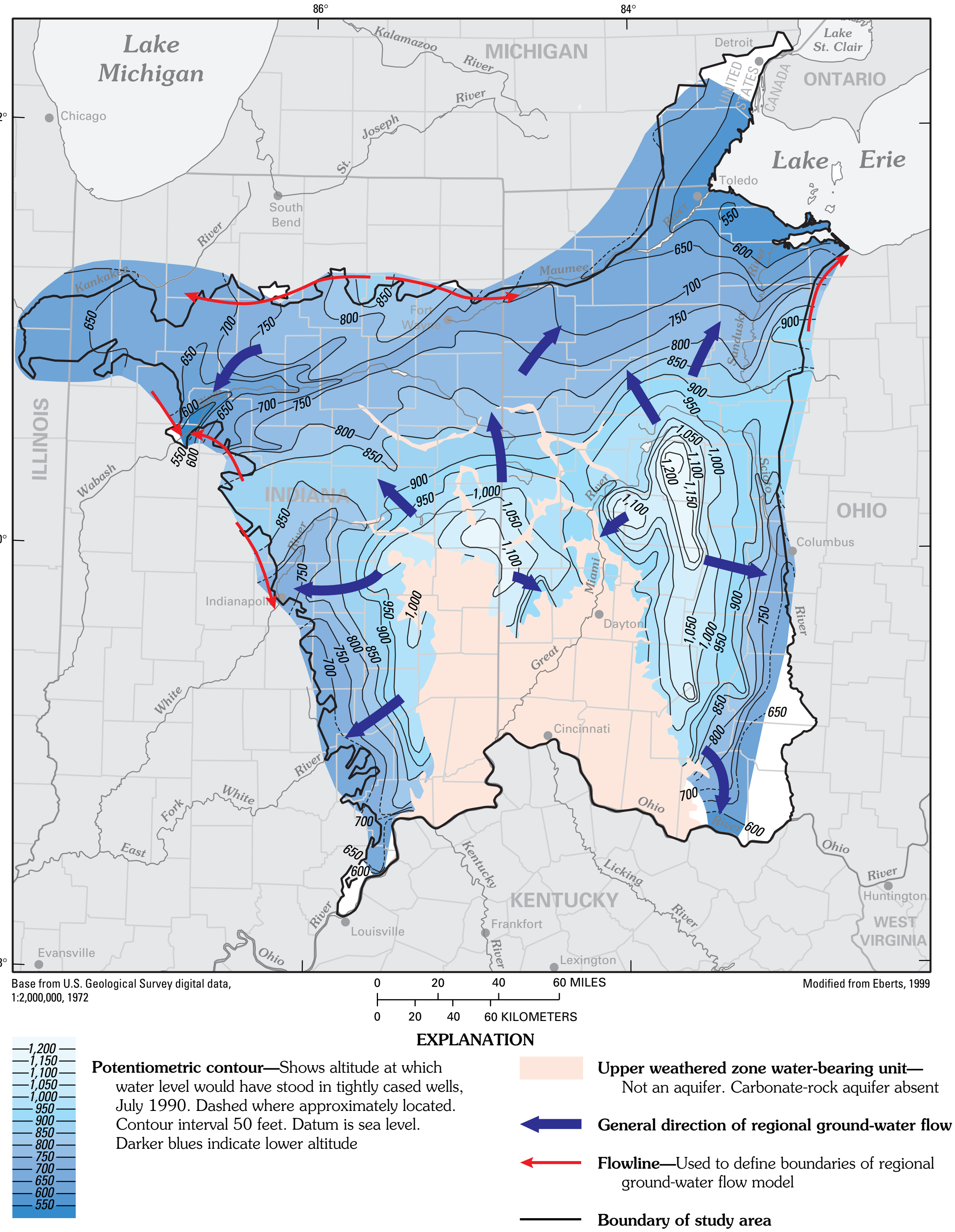
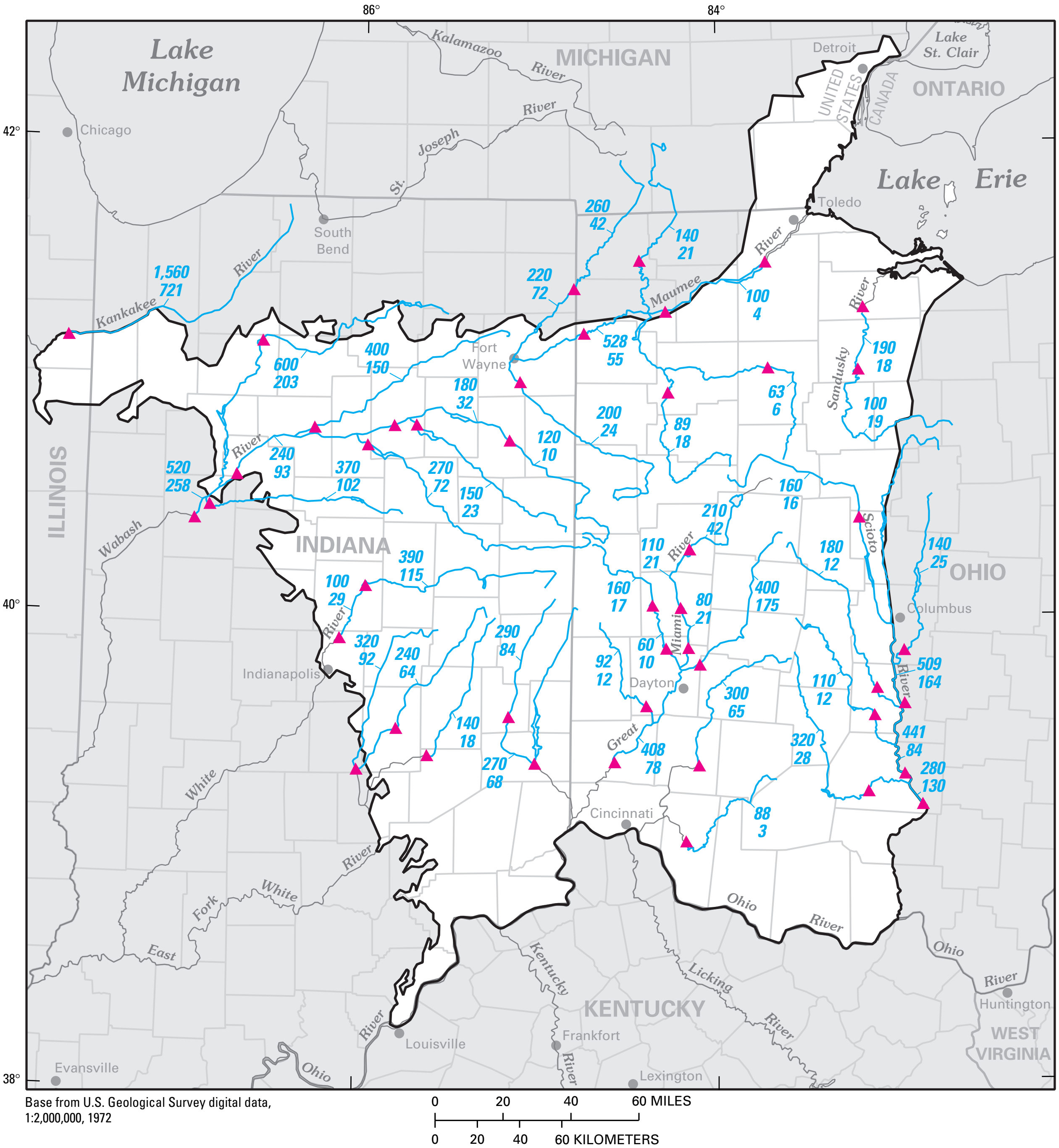


FIGURE 12.—Regional potentiometric surface in the carbonate-rock aquifer, July 1990, and general directions of regional ground-water flow.



EXPLANATION

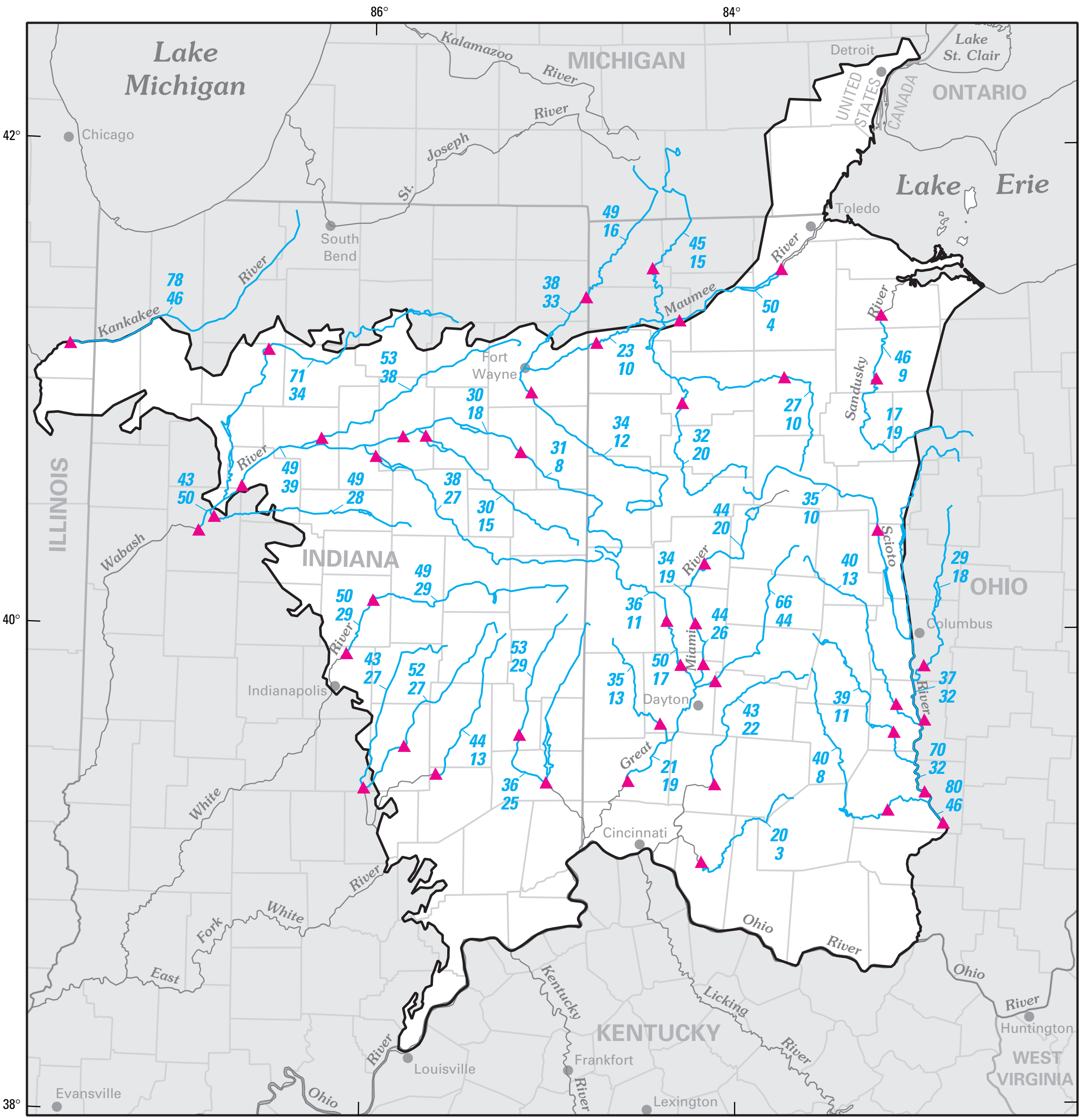
Boundary of study area

Ground-water discharge to selected stream reach—Upper number is mean ground-water discharge, in cubic feet per second. Lower number is mean sustained ground-water discharge, in cubic feet per second

270
68

Streamflow-gaging station

FIGURE 13.—Estimated ground-water discharge to selected stream reaches for long-term steady-state conditions in the Midwestern Basins and Arches aquifer system.



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

0 20 40 60 MILES

0 20 40 60 KILOMETERS

EXPLANATION

Boundary of study area

36
25

Discharge percentages for selected stream reach—Upper number is mean ground-water discharge as a percentage of mean streamflow. Lower number is mean sustained ground-water discharge as a percentage of mean ground-water discharge. Percentages represent long-term steady-state conditions in the Midwestern Basins and Arches aquifer system

Streamflow-gaging station

FIGURE 14.—Mean ground-water discharge as a percentage of mean streamflow, and mean sustained ground-water discharge as a percentage of mean ground-water discharge for selected stream reaches.

Basins and Arches aquifer system. Winter (1983) illustrates the effects of ground-water recharge from precipitation on hydraulic gradients in a water-table aquifer. He shows that hydraulic gradients in the areas of a water-table aquifer nearest a surface-water body are the first to respond to a recharge event because the water table is closest to land surface in these areas. Eventually, hydraulic gradients in the upland areas are affected by the recharge event (fig. 15). Seasonal fluctuations of base flow in streams are related to seasonal variations in ground-water recharge and the corresponding changes in hydraulic gradients at the water table. Some ground-water flow systems within an aquifer system are minimally affected by recharge events; these more stable ground-water flow systems provide a relatively constant source of ground-water discharge to streams over the course of a year and throughout long periods (fig. 15). The term "sustained ground-water discharge" is used herein to refer to this relatively constant source of base flow.

Base-flow duration curves constructed from daily mean base flows can be used to identify the component of base flow that is sustained during long periods, which include the driest periods (Eberts, 1999). (Base-flow duration curves are cumulative frequency curves that show the percentage of time during which specified base flows were equaled or exceeded in a given period; they are constructed by use of the method described by Searcy (1959), except that daily mean base flows are used instead of daily mean streamflows.)

Base-flow duration curves for streamflow-gaging stations along the principal streams within the study area are generally made up of two limbs when plotted on log-probability paper (fig. 16). The upper limb of each curve is commonly concave, whereas the lower limb of each curve is commonly a straight line, representing a flattening of the overall curve. This two-limb shape indicates the presence of at least two sources of ground-water discharge to these streams. Superimposed base-flow duration curves, constructed from periods of record that represent different ground-water recharge conditions, provide insight into the sources of ground-water discharge that result in the upper and lower limbs of the curves (fig. 16). Specifically, superimposed curves constructed from (1) the entire period of record, (2) from only summer months, when potential evapotranspiration exceeds precipitation (Todd, 1969), and (3) from a period of drought (U.S. Geological Survey, 1991) for single streamflow-gaging stations within the study area show that daily mean base flows that make up the upper limbs of the curves are from a ground-water source that readily responds to variations in ground-water recharge from precipitation. Conversely, the daily mean base flows that make up the lower limbs are from a ground-water source not greatly affected by variations in ground-water recharge (sustained ground-water discharge), as evident from the minimal differences in the lower limbs of the same curves. The base flows that make up the upper limb of each curve are likely to include a major component of transient, local-scale ground-water flow, because ground-water levels in local ground-water flow systems commonly decline in the summer and during droughts in response to a decrease in ground-water recharge from precipitation. The base flows that make up the lower limb of each curve are from a more stable ground-water flow, and likely represent a more dominant influence of intermediate- and regional-scale ground-water flow systems.

Lower limbs are absent or indistinct on base-flow duration curves constructed for streams within the study area that cease or nearly cease to flow during periods of dry weather. Such streams include small tributaries and streams that drain areas underlain by poorly permeable rocks. The absence of a lower limb on a base-flow duration curve cannot be used in itself to infer that stable flow systems are absent in the underlying aquifer system. Rather, relatively stable ground-water flow systems may be present in the underlying aquifer system but may simply discharge at some other point.

Mean sustained ground-water discharge to the stream reaches above selected streamflow-gaging stations was estimated by constructing base-flow duration curves and computing the average, for each curve, of all the daily mean base flows that make up the lower limb (Eberts, 1999). These base-flow duration curves were constructed, with the aid of a computer program developed by Lumb and others (1990), for the same period of record used to estimate mean ground-water discharge to the streams. Mean sustained ground-water discharge to stream reaches between streamflow-gaging stations was estimated by computing the difference between mean sustained ground-water discharge estimates for adjacent stations (fig. 13). These means describe the central tendency of sustained ground-water discharge to the streams for long-term steady-state conditions in the aquifer system.

Mean sustained ground-water discharge ranges from 3 to 50 percent of mean ground-water discharge for the 43 selected stream reaches. (These values are the lower numbers in figure 14.) Mean sustained ground-water discharge as a percentage of mean ground-water discharge increases with distance downstream along many stream reaches. Notable exceptions are stream reaches in the south-central part of the study area, where the carbonate-rock aquifer is absent, and stream reaches along the Maumee and Sandusky Rivers, which drain into Lake Erie.

Relative amounts of mean ground-water discharge and mean sustained ground-water discharge to streams are illustrated by principal drainage basin in figure 17. These values are for the entire area above the most downstream streamflow-gaging station in each selected drainage basin. Circles are used to illustrate the relative volumes of ground-water discharge within the basins. The sizes of the circles were determined by use of an exponential-curve-scaling method because the range of volumes was too large to be represented effectively by linear scaling. Estimated mean ground-water discharges are noted in the figure for reference. Also repre-

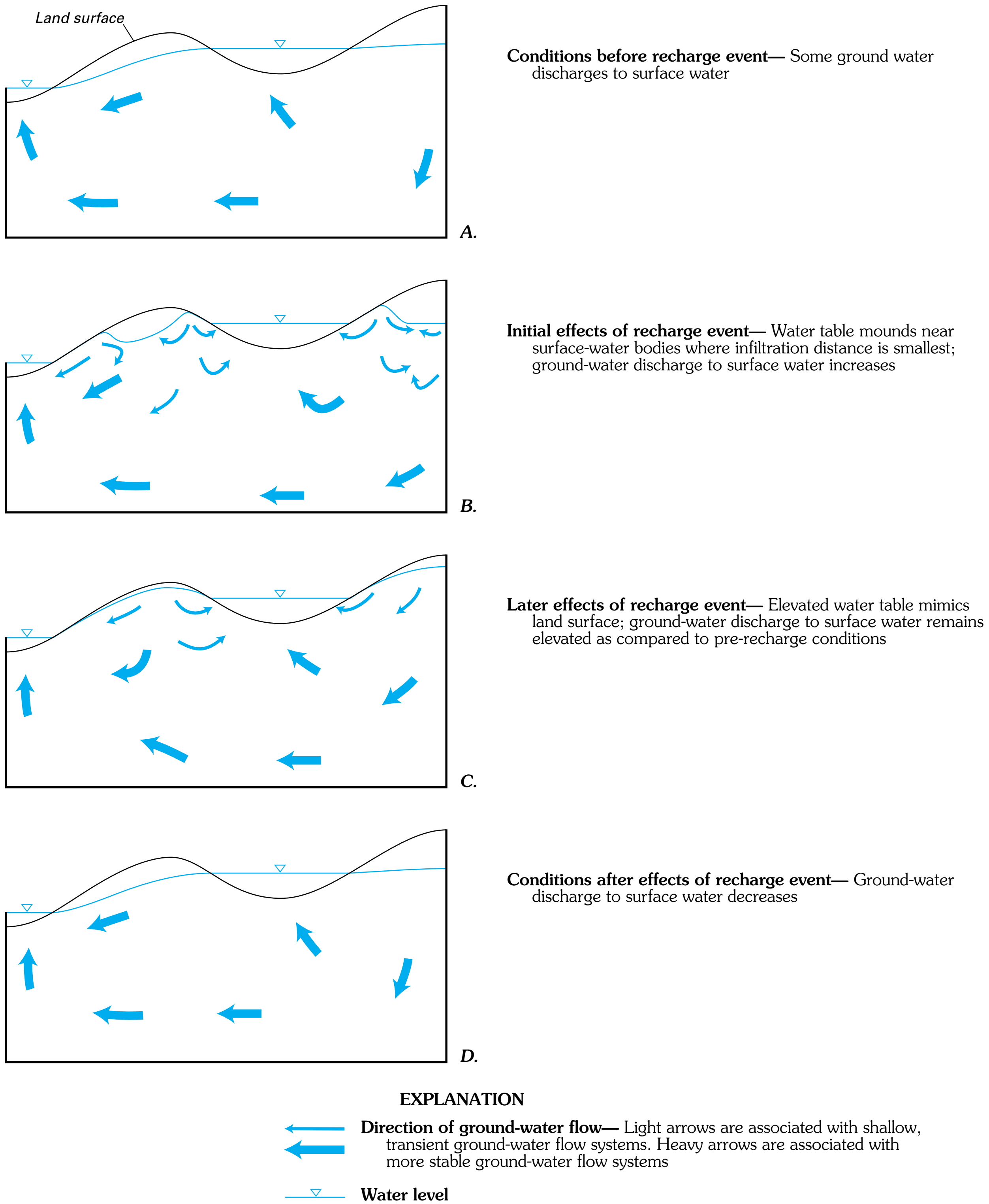


FIGURE 15.—Diagrams showing the effect of recharge from precipitation on the configuration of a water table and associated ground-water flow (modified from Winter, 1983).

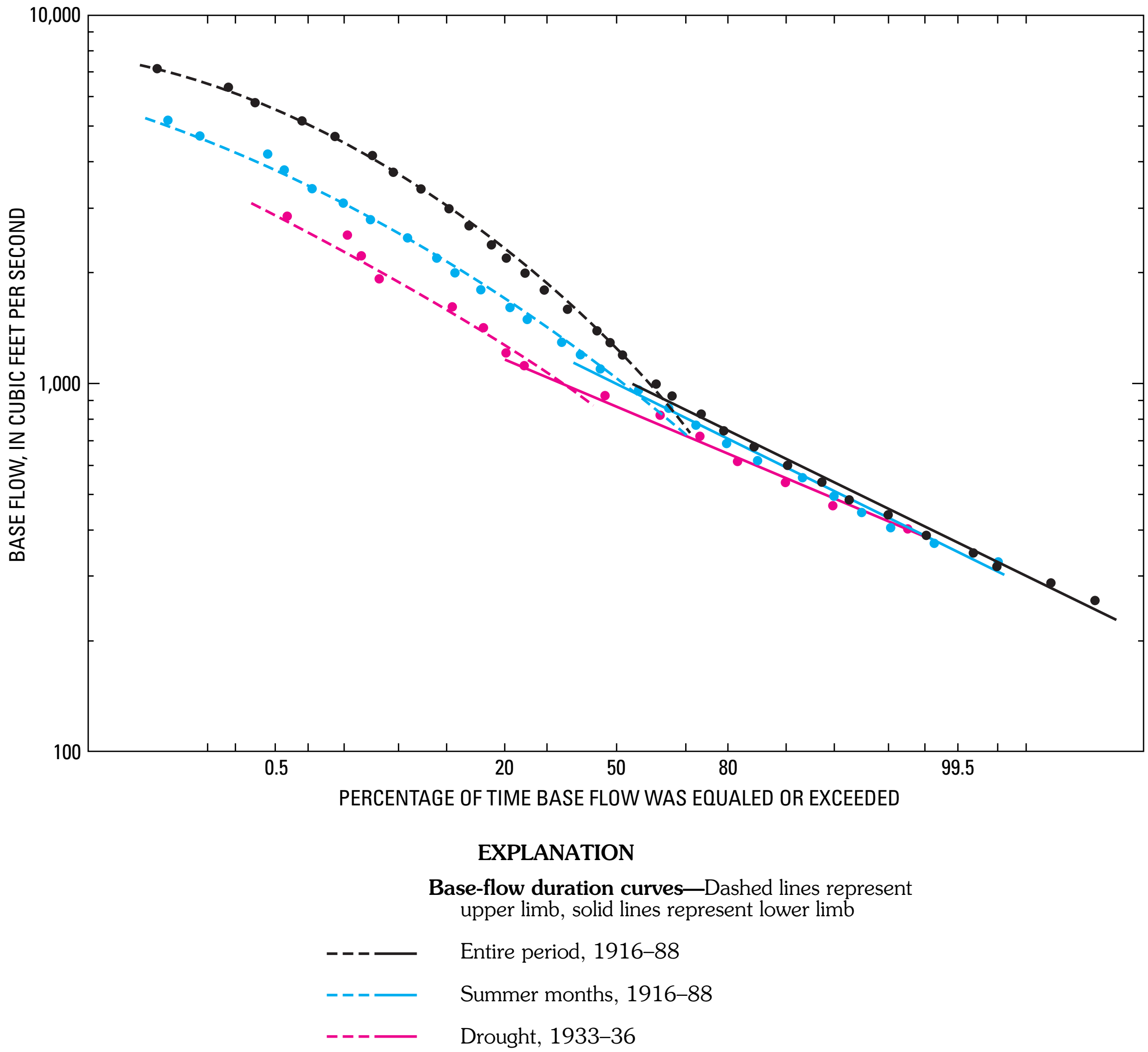


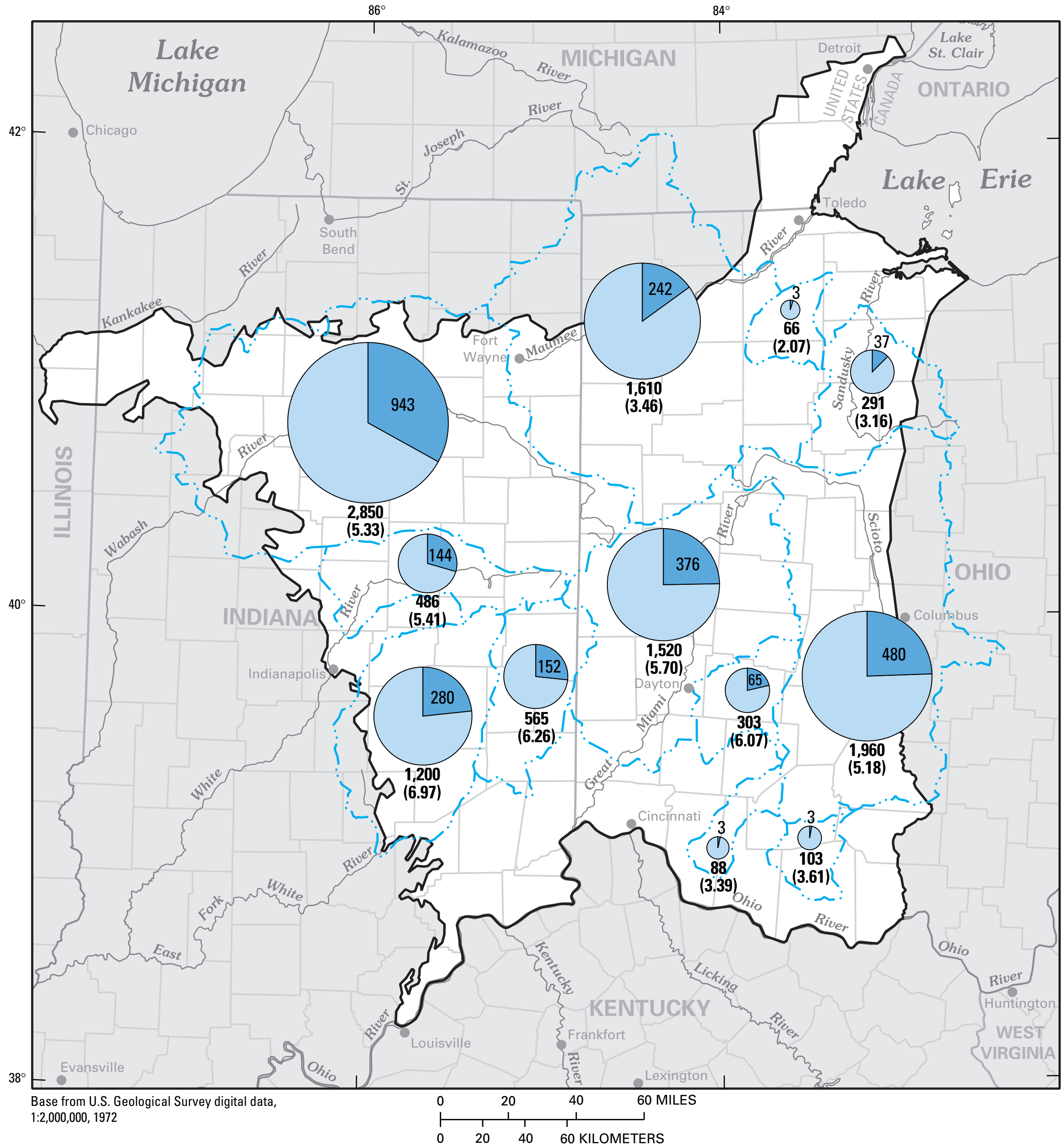
FIGURE 16.—Base-flow duration curves for various ground-water recharge conditions for a streamflow-gaging station on the Kankakee River.

sented in figure 17 are pie slices in each of the circles that shows the percentage of mean ground-water discharge to streams attributed to mean sustained ground-water discharge.

The greatest volume of ground-water discharge to streams is within the Wabash River Basin. This basin also has the highest mean sustained ground-water discharge as a percentage of mean ground-water discharge to the streams. Large amounts of outwash deposits are present within the basin. It is also the largest drainage basin within the study area. Bedrock crops out locally along the main stem of the Wabash River. In addition, the Wabash River is relatively deeply incised, and it has nearly the lowest base level within the study area. These factors likely contribute to the availability of ground-water recharge, resulting in a large volume of ground-water flow beneath the basin, and facilitate the interception of flow paths associated with the more stable ground-water flow systems.

The surface-water drainage basins with the smallest mean sustained ground-water discharge to streams as a percentage of mean ground-water discharge include the basins in the southeastern part of the study area. This may be due to the relatively small size of the basins, as well as their substantial local relief; transient, local ground-water flow systems generally dominate in such areas. These basins also differ from the Wabash River Basin in that glacial deposits are thin or absent within this area.

Flows of streams that drain into Lake Erie also consist of small percentages of mean sustained ground-water discharge. Hydraulic gradients within the aquifer system beneath the Maumee River Basin are fairly low. Because this basin is near Lake Erie and the base level of the aquifer system, the low hydraulic gradients may limit the capacity of the aquifer system to carry ground water away from recharge areas. As a result, much of the precipitation that potentially would



EXPLANATION

37 — **Mean sustained ground-water discharge to streams within selected surface-water drainage basins**—Represented by pie slice. Number is discharge, in cubic feet per second, for long-term steady-state conditions in the aquifer system

291 (3.16) — **Mean ground-water discharge to streams within selected surface-water drainage basins**—Represented by entire circle. Numbers are discharge, in cubic feet per second and inches per year (in parentheses), for long-term steady-state conditions in the aquifer system

Boundary of surface-water drainage basin

Boundary of study area

FIGURE 17.—Relative amounts of mean ground-water discharge and mean sustained ground-water discharge to streams in selected surface-water drainage basins.

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-