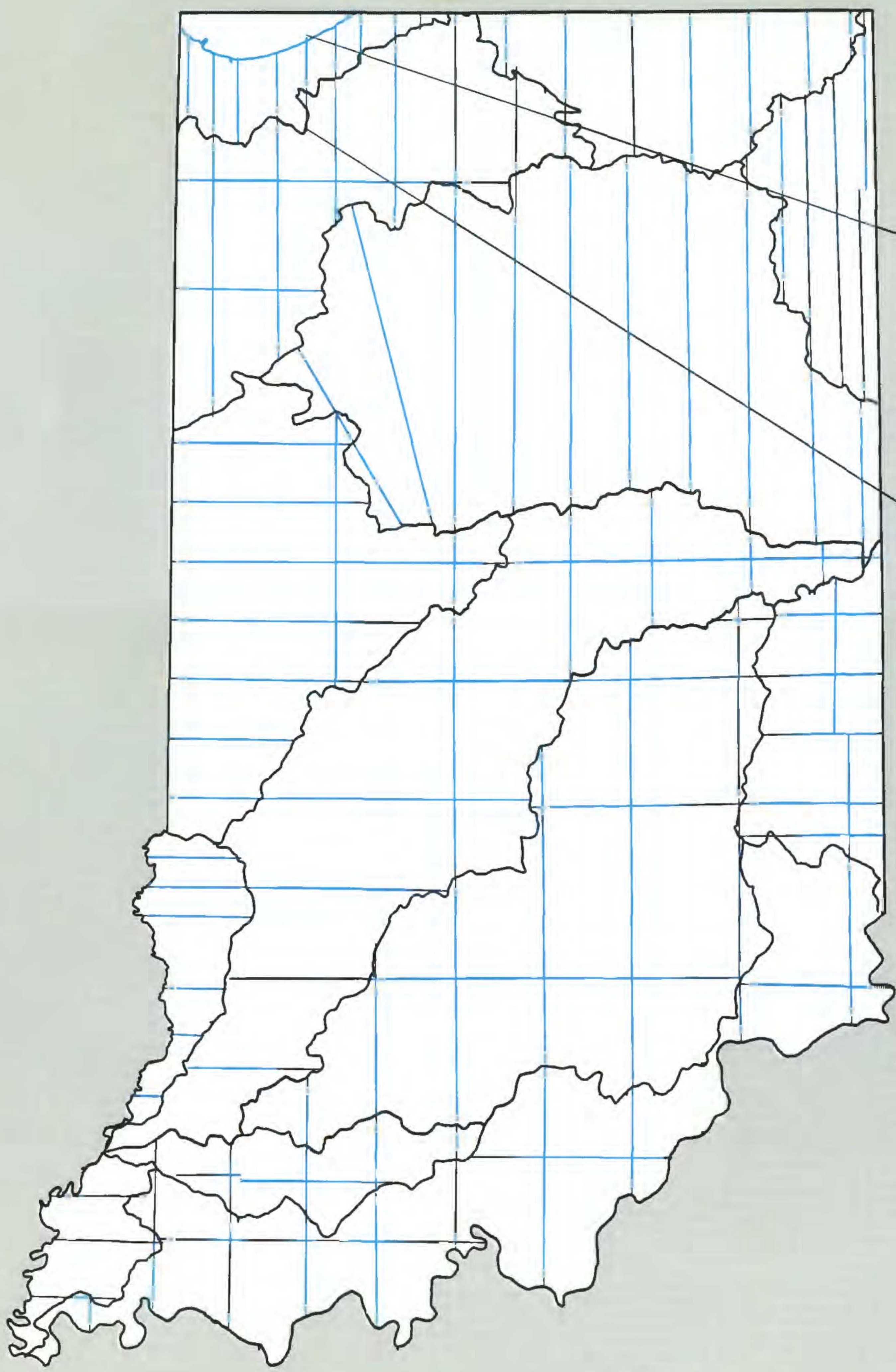
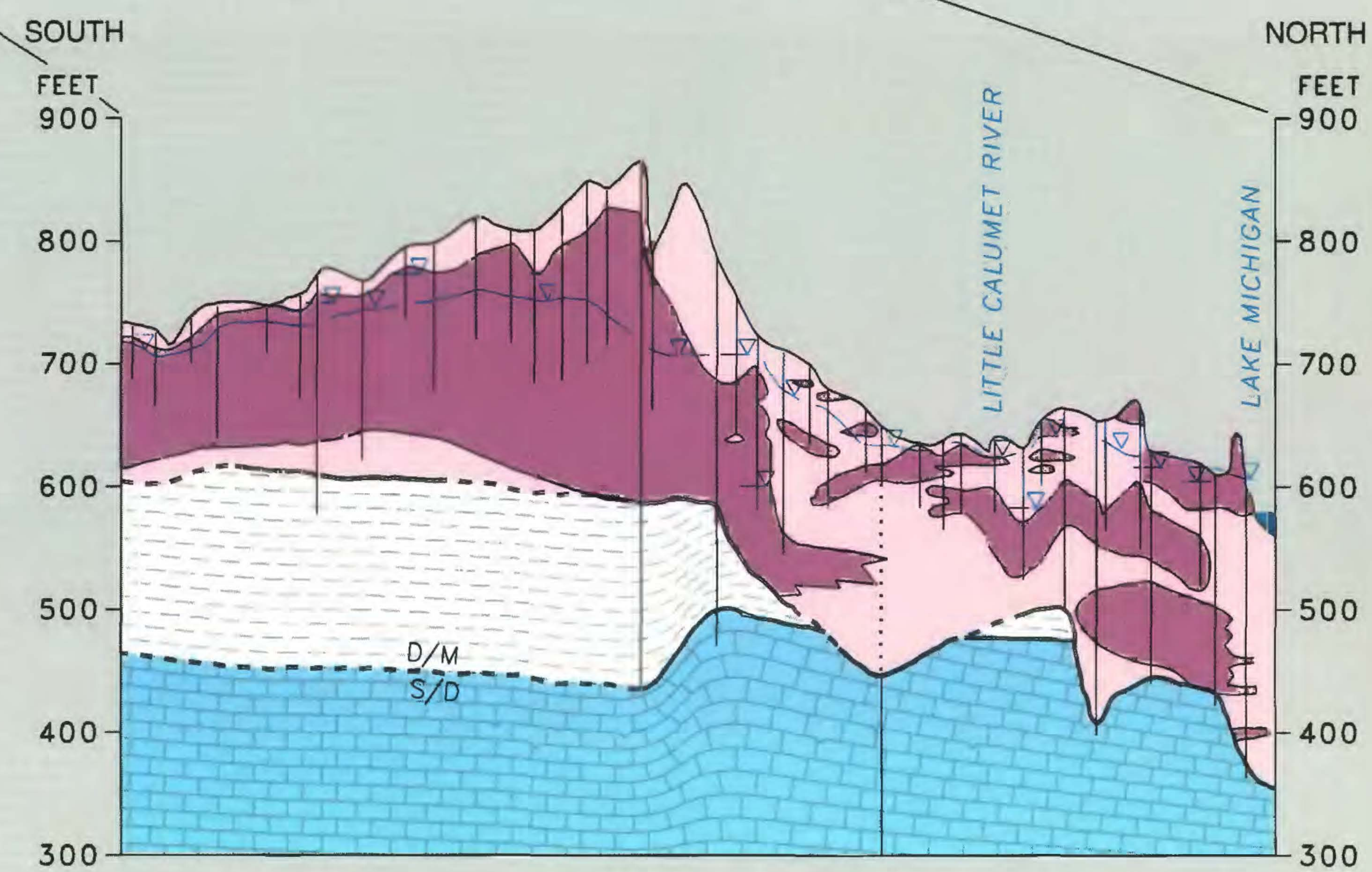
HYDROGEOLOGIC ATLAS OF AQUIFERS IN INDIANA



U.S. GEOLOGICAL SURVEY
Water-Resources Investigations Report 92-4142

Prepared in cooperation with the
INDIANA DEPARTMENT of NATURAL RESOURCES, DIVISION of WATER
INDIANA DEPARTMENT of ENVIRONMENTAL MANAGEMENT



COVER: A hydrogeologic section depicts sand and gravel aquifers (colored purple) and a carbonate bedrock aquifer (colored blue) adjacent to Lake Michigan in northern Indiana. The hydrogeologic section is one of 104 sections that were drawn for this atlas. Section lines (shown as blue lines on the State map) were drawn within each of Indiana's 12 water-management basins. The sections and maps included in this atlas provide a three-dimensional picture of the hydrogeology in Indiana.

Hydrogeologic Atlas of Aquifers in Indiana

By JOSEPH M. FENELON, KEITH E. BOBAY, and OTHERS

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To Obtain
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
acre	0.4047	hectare
square foot (ft^2)	0.09290	square meter
square mile (mi^2)	2.590	square kilometer
foot per mile (ft/mi)	0.1894	meter per kilometer
inch per year (in/yr)	2.54	centimeter per year
foot per day (ft/d)	0.3048	meter per day
mile per hour (mi/h)	1.609	kilometer per hour
foot squared per day (ft^2/d)	0.09290	meter squared per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
gallon per minute (gal/min)	3.785	liter per minute
gallon per day (gal/d)	3.785	liter per day
gallon per day per foot (gal/d/ft)	12.42	liter per day per meter
million gallons per day (Mgal/d)	0.04381	cubic meter per second
million gallons per year (Mgal/yr)	3,785	cubic meter per year
billion gallons per day (Bgal/d)	43.81	cubic meter per second
billion gallons per year (Bgal/yr)	3,785,000	cubic meter per year

Temperature, in degrees Fahrenheit (°F) can be converted to degrees Celsius (°C) as follows:

$$°C = 5/9\ (°F - 32)$$

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)—a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Hydrogeologic Atlas of Aquifers in Indiana

By Joseph M. Fenelon, Keith E. Bobay, *and* others

Abstract

Aquifers in 12 river basins (water-management basins) in Indiana are identified in a series of 104 hydrogeologic sections and 12 maps. Details of water-bearing units, including a generalized potentiometric surface, are derived from logs of more than 4,200 wells along 3,500 miles of section lines. Logs were obtained from water-well records, oil- and gas-well completion reports, coal-drilling records, and observation-well records. Well logs generally are plotted at 0.5- to 2-mile intervals. Hydrogeologic sections are spaced 6 to 20 miles apart. The horizontal scale of the sections is 1:250,000; vertical scale is greatly exaggerated. The scale of the maps depicting aquifers is 1:500,000. Aquifer maps are based on information from hydrogeologic sections and from previous studies. Where a type of aquifer was less than 15 square miles in areal extent, it was not mapped because of scale limitations. Types of aquifers depicted in the illustrations include unconsolidated and bedrock aquifers.

Unconsolidated aquifers are the most widely used aquifers in Indiana. Types of unconsolidated aquifers include surficial, buried, and discontinuous layers of sand and gravel. Most of the surficial sand and gravel is located in large outwash plains in northern Indiana and along the major rivers in the southern two-thirds of the State. Buried sand and gravel aquifers underlie much of the northern two-thirds of Indiana, where they are typically interbedded with till deposits and can be 10 to 400 ft deep. Discontinuous sand and gravel deposits are present as isolated lenses, primarily in glaciated areas.

Wells completed in the bedrock aquifers generally have lower yields than wells in most of the sand and gravel aquifers, but the bedrock aquifers are areally widespread and a major source of water for many domestic users and some large users of ground water. Carbonate rocks (limestone and dolostone); sandstones; complexly interbedded sandstone, siltstone, shale, limestone, and coal; and an upper weathered zone in low permeability rocks comprise the types of bedrock aquifers. Aquifers in carbonate rocks of Silurian, Devonian, and Mississippian age underlie about one-half of Indiana and are the most important of the bedrock aquifers in terms of yield and areal extent. The other principal type of bedrock aquifer is sandstone, which underlies large areas in the southwestern one-fifth of Indiana. The mapped sandstones are located within deposits of complexly interbedded sandstone, siltstone, shale, limestone, and coal of Mississippian and Pennsylvanian age. These complex deposits yield small quantities of water of variable quality, but they are important if they are the only available aquifer in a particular area. The remaining bedrock aquifer, which is used when it is the sole source of water for an area, is an upper weathered zone developed primarily in siltstone and shale of Mississippian and Devonian age and, to a lesser extent, in some of the shale and limestone of Ordovician age. No aquifer is mapped in the southeastern corner of Indiana, which is underlain by shale and limestone of Ordovician age.

INTRODUCTION

Ground water is the source of drinking water for nearly 60 percent of the residents of Indiana. Approximately 425 community water systems, 3,000 noncommunity water systems, 500 mobile-home parks, and 500,000 private homes are supplied by wells (Indiana Department of Environmental Management, 1990, p. 223; Indiana Department of Natural Resources, 1989, written commun.). In addition to drinking-water supplies, ground water is withdrawn for energy production, irrigation, and industrial, commercial, and agricultural uses. In 1991, about 204 Bgal (billion gallons) of ground water, or a daily average of 559 Mgal (million gallons), was withdrawn. The combined capability of registered ground-water withdrawal facilities in 1991 was 3,540 Mgal/d (million gallons per day) (Indiana Department of Natural Resources, 1993, written commun.).

Ground water is an important and abundant natural resource in Indiana; however, detailed maps and descriptions of the major **aquifers**[1] that pertain to the entire State have not been available. Published reports are currently limited to county-wide studies, a few basin studies, detailed site-specific investigations, and large-scale maps and assessments of the aquifers in the State.

The Indiana Ground-Water Protection and Management Strategy lists the delineation and mapping of aquifers as a primary need (Indiana Department of Environmental Management, 1987). The Strategy states that nearly all aspects of ground-water regulation, research, and utilization in the public and private sectors can benefit from maps and descriptions of the aquifers in the State. Therefore, one of the short-term goals of the strategy was the creation of a ground-water atlas of Indiana that would identify generic aquifers on a large scale (Indiana Department of Environmental Management, 1987, p. 5). The term "generic" was used to imply that aquifers do not necessarily conform to geologic age, group, or formation.

In response to the need for such an atlas, the U.S. Geological Survey (USGS), in cooperation with the Indiana Department of Natural Resources (IDNR) and the Indiana Department of Environmental Management (IDEM), prepared a series of hydrogeologic sections and maps that identify aquifers in 12 water-management basins of Indiana (fig. 1).

Purpose and Scope

This atlas describes and delineates aquifers in the Lake Michigan, St. Joseph River, Kankakee River, Maumee River, Upper Wabash River, Middle Wabash River, Lower Wabash River, White River, East Fork White River, Whitewater River, Patoka River, and Ohio River water-management basins in Indiana. The hydrogeologic sections were constructed at a horizontal scale of 1:250,000, whereas the maps were drawn at a scale of 1:500,000. The vertical scale of the sections is greatly exaggerated. Also included are maps that show the location of the hydrogeologic section lines, the thickness of unconsolidated deposits (from Gray, 1983), and the bedrock geology (from Gray and others, 1987) for each basin at a scale of 1:500,000.

[1] Terms in bold are defined in the "Definitions of Selected Terms" at the back of this report

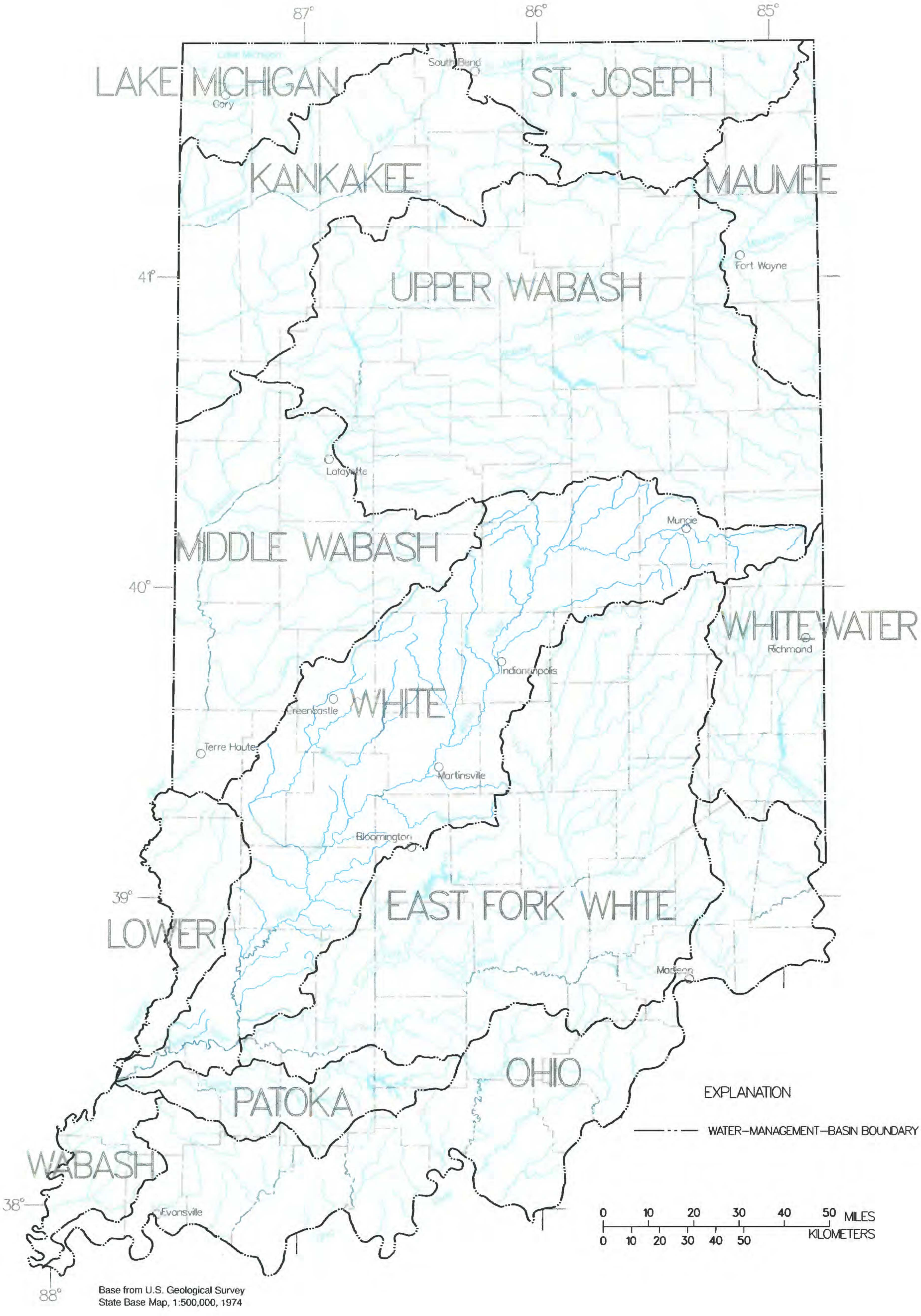

Figure 1. The 12 water-management basins of Indiana.

As defined in this atlas, an aquifer is a geologic formation, group of formations, or part of a formation that contains sufficient **saturated** permeable material to **yield** quantities of potable water adequate for domestic purposes (Lohman and others, 1972, p. 2). Location and delineation of aquifers throughout Indiana is the primary goal of this atlas. Types of aquifers are distinguished by lithology, thickness, depth, and continuity.

The hydrogeologic sections and areal maps can be used to understand and evaluate aquifer systems on a regional scale. This regional evaluation can then be used as a base for site-specific studies. **The sections and maps, however, do not replace site-specific hydrogeologic data**. The well logs used to plot sections were obtained from a 2 mile-wide path that bounded the traces of the sections. Thus, the data shown at a given location might not represent site-specific hydrogeology.

A glossary of hydrogeologic terms used herein is at the end of this report. Bedrock geologic names in this report follow the nomenclature of Shaver and others (1986).

Previous Studies

The ground-water resources of Indiana have been studied by many authors since the early 1900's. Harrell (1935) described the general physiographic features, hydrology, geology, and ground-water resources of each county in the State. Bechert and Heckard (1966) discussed the availability, flow, quality, and uses of ground water in Indiana. Bloyd (1974) summarized the ground-water resources of a region that includes greater than 80 percent of Indiana and provided regional estimates of hydraulic conductivity, specific yield, storage, **recharge**, and current and projected withdrawals. Clark (1980) characterized the availability, use, regulation, and future needs of ground water in Indiana. Geosciences Research Associates, Inc. (1982) summarized the "potential yield capability" and the water quality of the major bedrock hydrostratigraphic units in the State. The USGS (1988, p. 245-250) described the water quality of the principal aquifers of Indiana. The USGS is studying the flow and water quality in the regional carbonate bedrock and glacial aquifer system in Indiana (Bugliosi, 1990; Casey, 1992; Schnoebelen, 1992).

The location and extent of ground-water resources or aquifers in Indiana have been mapped by a few authors. Bechert and Heckard (1966, p. 109) mapped the availability of ground water on the basis of yields from "properly sized and developed" wells in eight ground-water provinces of Indiana. Gray (1973) mapped the general location and described "the principal resource units in ground-water production" for Indiana. Clark (1980, p. 33) updated the ground-water availability map of Bechert and Heckard using seven potential yield categories that were devised from a "range of probable maximum yields which can be expected from a properly constructed large-diameter well penetrating the full thickness of the aquifer." Geosciences Research Associates, Inc. (1982) mapped the structure and contour of the major bedrock hydrostratigraphic units in Indiana on the basis of geologic age and formation. The USGS (1985, p. 207) mapped the principal aquifers of Indiana as glaciofluvial, glacial **outwash**, sand and gravel lenses in **till** of Wisconsinan age, carbonate rocks of Mississippian age, and carbonate rocks of Silurian and Devonian age (1:5,000,000 scale); however, a large area of southwestern and south-central Indiana was mapped as being without a principal aquifer.

The Indiana Department of Natural Resources is preparing water-resource availability studies for the 12 water-management basins in Indiana. Published reports are currently (1990) available for the St. Joseph River, Whitewater River, and Kankakee River basins. These reports include maps that show the extent of aquifers, composite **potentiometric surface** maps for unconsolidated and **bedrock aquifer**s, a discussion of hydrogeologic characteristics, and information on ground-water quality and use. This atlas differs from these studies in the method used to delineate and name aquifers (generic aquifer types as compared to formal aquifer names for different geologic settings) and in the emphasis on the vertical distribution of the aquifers as shown in many detailed hydrogeologic sections.

Acknowledgments

The staff of the Indiana Department of Environmental Management and the Indiana Department of Natural Resources were helpful in all aspects of the atlas project including planning, format, and review of the Atlas. Special thanks are extended to the following people: Robert Hilton[2], Martin Risch[2], James Nowacki[2], Mike Yarling, Gregg Lemasters, and James Harris from the Indiana Department of Environmental Management; John Simpson, Thomas Bruns[2], William Steen, John Clark[2], John Barnhart[2], Michael Saul[2], and Sally Letsinger[2] from the Indiana Department of Natural Resources, Division of Water; and Norman Hester, Ned Bleuer, Anthony Fleming, Eric Kwale, John Rupp, and Henry Gray[2] from the Indiana Department of Natural Resources, Geological Survey.

The authors also wish to thank David Sperry and David Zetzl for providing the technical support to produce or help produce virtually every illustration in this atlas.

PHYSICAL SETTING OF INDIANA

The physical setting of Indiana, which includes the physiography, climate, and geology, controls the distribution, availability, and flow of ground water.

Physiography

Indiana has been divided into 13 major physiographic units (fig. 2) on the basis of similarities in topography and geology (Schneider, 1966, p. 41). The 13 units occupy 3 broad physiographic zones that trend in an east-west direction across the State. The zones are the Northern Lake and Moraine Region, the Central Drift Plain, and a southern zone dominated by bedrock landforms.

The Northern Lake and Moraine Region is subdivided into five lake (lacustrine) or morainal units: the Calumet Lacustrine Plain, the Valparaiso Morainal Area, the Kankakee Outwash and Lacustrine Plain, the Steuben Morainal Lake Area, and the Maumee Lacustrine Plain (fig. 2). A variety of glacial and postglacial landforms are in this 8,500 mi^2 area. Glacial depositional features include end moraines, till plains, **outwash plains** and **valley trains**, kames, and lake plains. These landforms have a diverse mix of sediments with highly variable hydrogeologic properties and numerous lithologic discontinuities. Related postglacial landforms include the many lakes of northeastern Indiana, the sand dunes along Lake Michigan, and peat bogs (Schneider, 1966, p. 40-42). Principal moraines and the extent of glaciation in Indiana are shown in figure 3.

The Tipton Till Plain, or Central Drift Plain, is a nearly flat glacial till plain covering central Indiana (fig. 2). This area of about 12,000 mi^2 is underlain by thick till and has been slightly eroded by postglacial streams. Most of the boundary between the Till Plain and the southern physiographic units coincides with the maximum extent of Wisconsinan glaciation, except in southeastern Indiana where the physiographic boundary is north of the Wisconsinan glacial boundary. The southeastern Indiana boundary was arbitrarily drawn along the edge of a broad transitional zone of thin glacial **drift** that does not obscure the bedrock physiography (Schneider, 1966, p. 40, 49).

Seven physiographic units composed of different bedrock types comprise 15,500 mi^2 in the southern one-third of the State (fig. 2). The bedrock is primarily sandstone, shale, siltstone, limestone, and dolomite. The physiographic units generally trend north-northwest following the strike of the bedrock. From east to west, the units are called the Dearborn Upland, the Muscatatuck Regional Slope, the Scottsburg Lowland, the Norman Upland, the Mitchell Plain, the Crawford Upland, and the Wabash Lowland (Schneider, 1966, p. 42-49). These southern units represent a sharply divided, alternating series of uplands and lowlands or plains. Large parts of the Wabash Lowland, Crawford Upland, Mitchell Plain, and Norman Upland were not glaciated during the Pleistocene Epoch. All seven bedrock physiographic units extend further north than shown in figure 2 but were buried by glacial deposits (Schneider, 1966, p. 54). Buried erosional surfaces of these units are evident in the hydrogeologic sections presented later in this report.

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[2]Person is no longer employed with the agency.

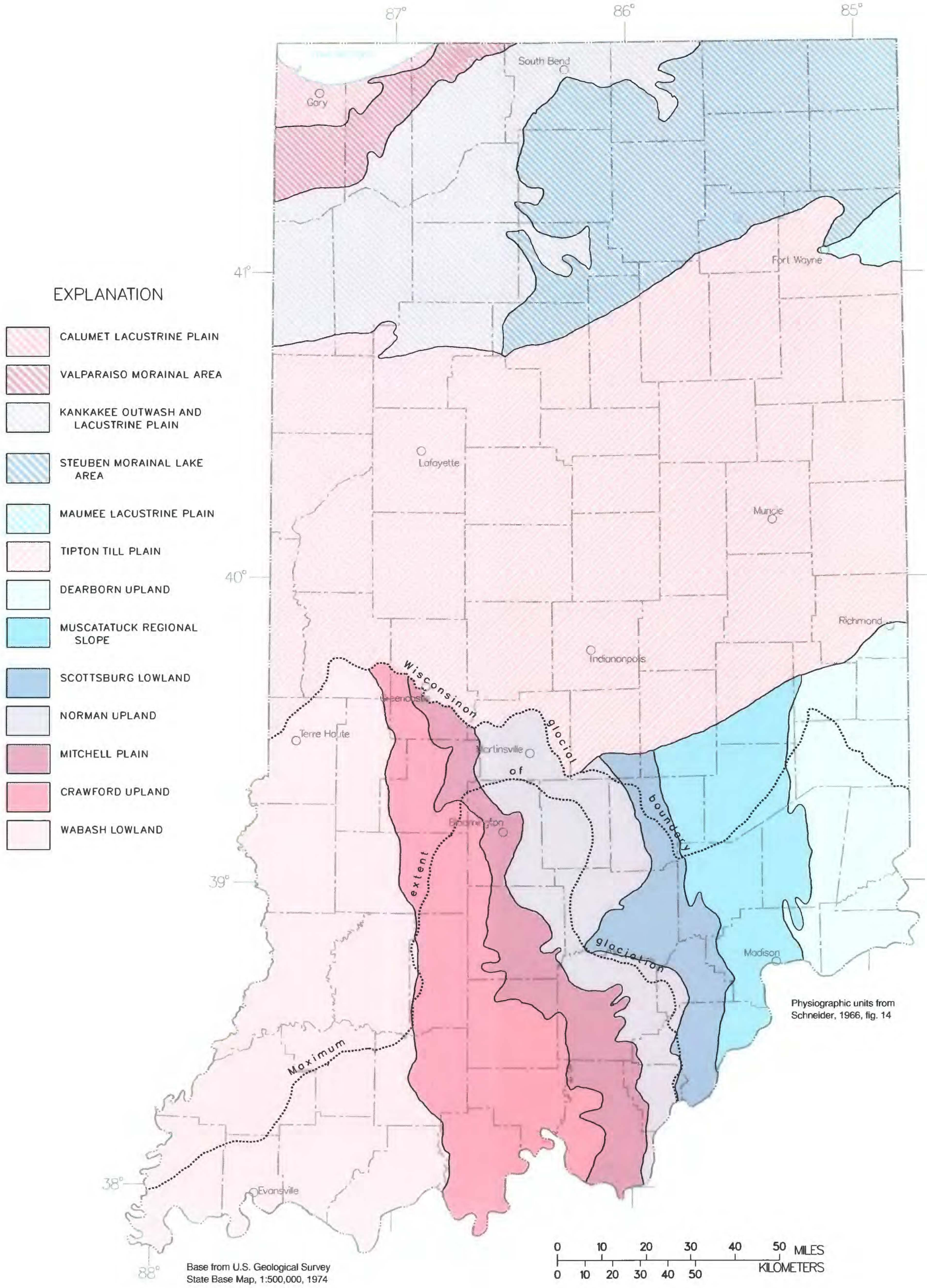


Figure 2. Physiographic units of Indiana.

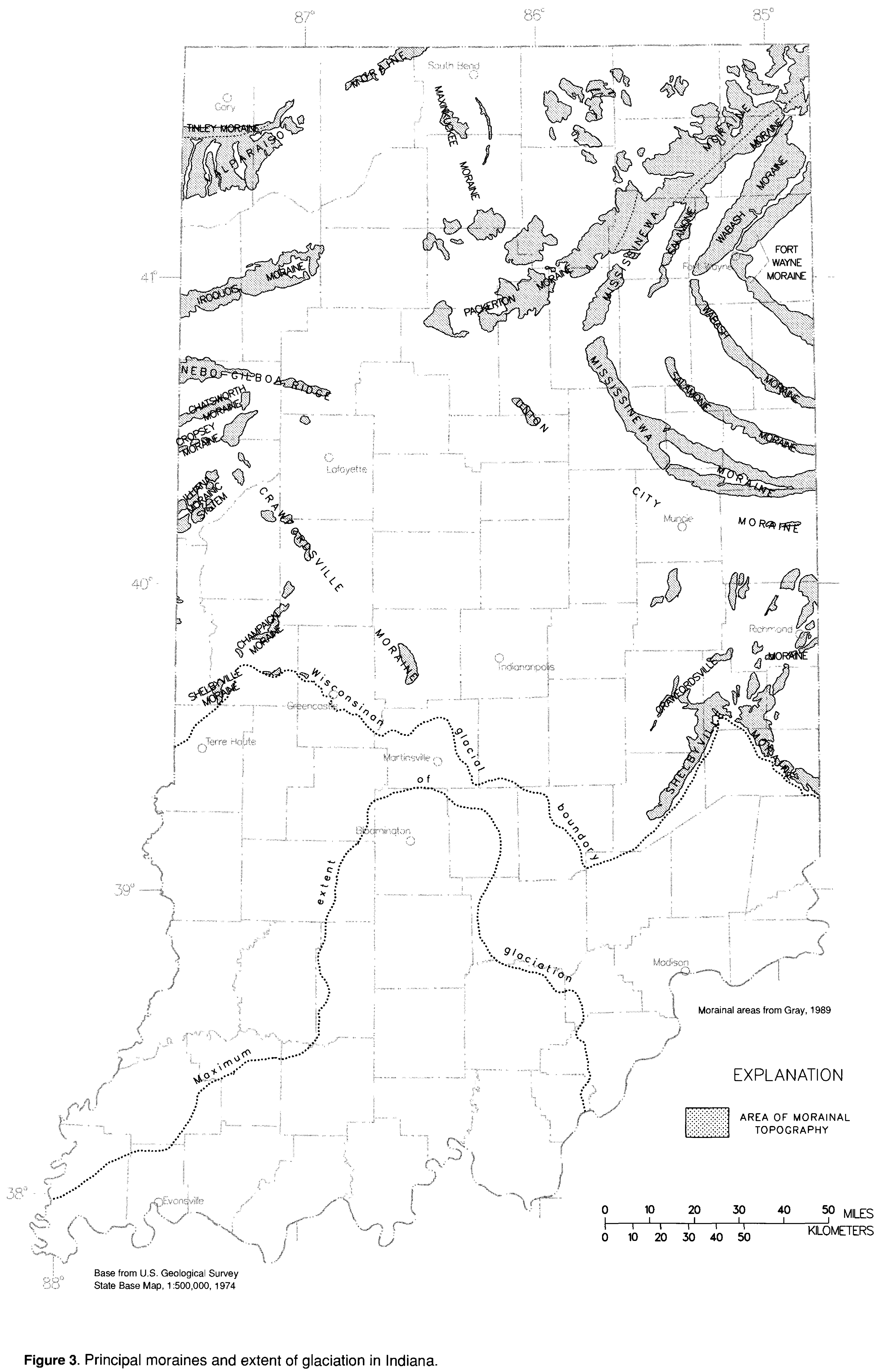


Figure 3. Principal moraines and extent of glaciation in Indiana.

The distribution of bedrock types and erosional characteristics determines the topography of the bedrock, which affects the thickness of unconsolidated deposits. Thick glacial sediments are present beneath moraines and in buried bedrock valleys. In the southern one-third of the State, unconsolidated deposits are thin and discontinuous, especially beyond the maximum extent of glaciation. The distribution, thickness, and hydraulic conductivity of these unconsolidated deposits control the near-surface occurrence and flow of ground water (Krothe and Kempton, 1988, p. 129).

Climate

Climate in Indiana is temperate with warm summers, cold winters, and no distinguishable wet or dry seasons. Precipitation is well-distributed throughout the year, although precipitation is somewhat greater during March through July because of increased frequency and intensity of showers and thunderstorms. The average length of the growing season, or freeze-free period, ranges from 173 days in northeastern Indiana to 199 days in southwestern Indiana (National Oceanic and Atmospheric Administration, 1988).

The interaction of tropical and polar air masses over Indiana normally results in abundant precipitation; however, temperature and precipitation vary considerably from year to year depending on the frequency of storms and frontal passages (National Oceanic and Atmospheric Administration, 1988). Average annual precipitation in Indiana ranges from 44 inches in the south to 36 inches in the northeast; average annual snowfall ranges from 10 inches in the south to 40 inches in northern Indiana. Average annual temperature ranges from 50° F in the north to 56° F in the southwest. Monthly evapotranspiration can be as much as 8 inches in southern Indiana in July (National Oceanic and Atmospheric Administration, 1988; Schaal, 1966; Visher, 1944, p. 450-461).

Geology

The regional structural features of Indiana bedrock include the Illinois Basin, the Michigan Basin, and the Cincinnati Arch (fig. 4). The two basins form the flanks of a saddle-like structure composed of the Cincinnati Arch and its branches, the Kankakee Arch and the Findlay Arch (in Ohio). The dip of the rocks into the two basins is about 10 to 30 ft/mi, although the dip can be less than 5 ft/mi at the top of the arches. Rocks of Ordovician, Silurian, Devonian, and Mississippian age crop out or are present as subcrops in both of the depositional basins, whereas rocks of Pennsylvanian age are present in Indiana only in the Illinois Basin. The older rocks are typically present on the crest of the arch; the progressively younger rocks are present in each of the basins. Individual beds in many formations are thin at the crest, but they thicken along the flanks and into the basins (Gutschick, 1966, p. 7-12). A geologic chart including age, group or stage, selected formations, and hydrogeologically important members or marker beds in Indiana is shown in figure 5.

Rocks of the Precambrian crystalline basement complex are found at estimated depths of 3,000 to 6,000 ft in the northeastern two-thirds of Indiana and 6,000 to 14,000 ft in the southwestern one-third (Rupp, 1991). Overlying the Precambrian bedrock is a Cambrian section of sandstone with lesser amounts of siltstone, and shale that is approximately 1,000 ft thick in eastern Indiana to 3,000 ft thick in northwestern Indiana. The Cambrian rock composes about one-third of the Paleozoic section in Indiana (Shaver and others, 1986, p. 119). Overlying the Cambrian rock is 20 to 4,500 ft of Lower Ordovician dolomite (Shaver and others, 1986, p. 70). The dolomite thickens toward the southwestern part of the State. The Lower Ordovician dolomite is unconformably overlain by 50 to more than 450 ft of dolomite, limestone, and sandstone of Middle Ordovician age (Shaver and others, 1986, p. 4).

Late Ordovician shale and limestone is exposed at the bedrock surface over large areas in southeastern Indiana (fig. 6). The shale and limestone range in thickness from approximately 500 ft in northwestern Indiana to 1,500 ft in southeastern Indiana (Shaver and others, 1986). The shale and limestone are unconformably overlain by Silurian limestone and dolomite. The Silurian rocks are present as subcrops or outcrops in east-central to northwestern Indiana (fig. 6), primarily along the axis of the Cincinnati Arch and the Kankakee Arch (fig. 4). The Silurian rocks generally range from 200 to 600 ft in thickness, except in the southeastern part of the State, where they were completely eroded. Large areas of carbonate platform and reef banks are present in these carbonate rocks along the flanks of the arches.

Figure 4. Regional structural features in Indiana.

GEOLOGIC AGE	Series or Group	Selected Formations	Selected Members
QUATERNARY	Holocene		
	Pleistocene: Wisconsinan	Wedron Formation: Lagro Formation, Trafalgar Formation	
	Pleistocene: Pre-Wisconsinan	Jessup Formation	
PENNSYLVANIAN	McLeansboro Group	Mattoon Formation	Merom Sandstone Member
		Bond Formation	St. Wendel Sandstone Member
		Patoka Formation	Inglefield Sandstone Member
		Shelburn Formation	West Franklin Limestone Member Busseron Sandstone Member
	Carbondale Group	Dugger Formation	Danville Coal Member (Coal VII) Anvil Rock Sandstone Member Hymera Coal Member (Coal VI)
		Petersburg Formation	Springfield Coal Member (Coal V)
		Linton Formation	Survant Coal Member (Coal IV) Coxville Sandstone Member
	Raccoon Creek Group	Staunton Formation	Seelyville Coal Member (Coal III)
		Brazil Formation	Upper Block Coal Member Lower Block Coal Member Buffaloville Coal Member
		Mansfield Formation	
MISSISSIPPIAN	Buffalo Wallow Group	Tar Springs Formation	
	Stephensport Group	Glen Dean Limestone Hardinsburg Formation Haney Limestone Big Clifty Formation Beech Creek Limestone	
	West Baden Group		
	Blue River Group	Paoli Limestone Ste. Genevieve Limestone St. Louis Limestone	
	Sanders Group		
	Borden Group	Edwardsville Formation Spickert Knob Formation New Providence Shale	
		Rockford Limestone	
DEVONIAN		Northern Indiana only: Coldwater Shale, Sunbury Shale, Ellsworth Shale, Antrim Shale	Southern Indiana: New Albany Shale
	Muscatatuck Group	Traverse Formation (Northern Indiana) Jeffersonville Formation (Southern Indiana)	Geneva Dolomite Member
SILURIAN	Salina Group; Bainbridge Group (Southwest Indiana)	Wabash Formation	Mississinewa Shale Member
		Northern Indiana: Louisville Limestone, Waldron Shale, Salamonie Dolomite, Cataract Formation	Southern Indiana: Louisville Limestone, Waldron Shale, Salamonie Dolomite
		Sexton Creek Limestone	Brassfield Limestone
ORDOVICIAN	Maquoketa Group	Southeastern Indiana: Whitewater Formation, Dillsboro Formation, Kope Formation	
		Trenton Limestone	Lexington Limestone (Southeastern Indiana)
	Black River Group		
	Ancell Group	Joachim Dolomite	St. Peter Sandstone
ORDOVICIAN / CAMBRIAN	Knox Supergroup		
CAMBRIAN	Munising Group	Franconia Formation Ironton Sandstone Galesville Sandstone Eau Claire Formation	
		Mt. Simon Sandstone	
PRE-CAMBRIAN	Basement Complex--Includes granite, basalt, and arkose of Middle Proterozoic age		

Figure 5. Geologic chart showing geologic age, group, and selected formations and members (geologic names from Shaver and others, 1986).

The Silurian carbonate rocks are unconformably overlain by Devonian dolomite and limestone (fig. 6), which attain thicknesses of 250 ft toward the centers of the Illinois Basin and the Michigan Basin (fig. 4). The carbonate rock sequence is overlain primarily by shales and siltstones of Late Devonian and Early Mississippian age. These shales and siltstones are present as subcrops in the Michigan Basin in the northeastern part of the State (fig. 6) and attain thicknesses of 800 ft in places. They also are present as subcrops or outcrops along the southwestern flank of the Cincinnati and Kankakee Arches in the south-central part of the State (fig. 6), where they range from 500 to 1,000 ft in thickness. Limestone of Middle Mississippian age overlies the shales and siltstones southwest of the arches. The area where the limestone is exposed at the bedrock surface trends northwest through the south-central part of the State. The Mississippian limestone ranges from 200 to more than 1,000 ft in thickness; the thicker deposits are in the southwestern corner of Indiana toward the center of the Illinois basin.

The limestone is overlain in the southwestern one-third of Indiana by rocks of Early Mississippian age and Middle and Late Pennsylvanian age (fig. 6). The rocks are composed of sandstone, shale, and thin but extensive beds of limestone, clay and coal. The beds of clay and coal are generally found above a major unconformity between the Mississippian and Pennsylvanian rocks. These rocks range from 1,000 to 2,000 ft in thickness.

There was little known deposition in Indiana between the end of the Pennsylvanian Period and the beginning of the Quaternary Period. During this time, the land surface was mostly an erosional surface that consisted of northwest-trending limestone plains, shale lowlands, and sandstone uplands (Wayne, 1966, p. 27). Rivers were entrenched in the bedrock; the main preglacial river valley in north-central Indiana was the Lafayette Bedrock Valley System (Bleuer, 1989), also known as the Teays valley, which drained most of the northern one-half of the State (fig. 7). The preglacial topography of most of the northern two-thirds of the State was buried beneath 50 to more than 400 ft of glacial debris during the Quaternary Period (Gray, 1983).

The earliest widespread evidence of continental glaciation in Indiana was from glaciers of pre-Illinoian and Illinoian Age. They extended through the northern three-quarters of Indiana (fig. 3; Wayne, 1966, p. 33). These pre-Wisconsinan glaciers deposited at least eight till units in Indiana that comprise approximately 75 percent of the glacial deposits in the Tipton Till Plain (A.J. Fleming, Indiana Geological Survey, 1990, written commun.) The only known pre-Wisconsinan deposits exposed at the surface are found over large areas south of the Wisconsinan glacial boundary (fig. 3). These deposits are composed of a sandy loam till of the Jessup Formation and deposits in proglacial lakes and outwash plains in southwestern Indiana (Gray, 1989). **Loess** was deposited downwind (east) of the valleys of the Wabash and Ohio Rivers.

There were several glacial advances in Indiana during Wisconsinan time by three different ice lobes (fig. 8). The furthest advance in the State was by the Erie Lobe, which covered the northern two-thirds of Indiana. The ice, which formed the Shelbyville Moraine in southeastern Indiana, was followed by another advance of the ice that formed the Crawfordsville Moraine (fig. 3). The two advances were from a northeastern source of ice and deposited till known as the Trafalgar Formation over large areas of central Indiana (Gray, 1989). As the ice receded, it left large amounts of sand and gravel in the form of valley train, kames, and eskers (Wayne, 1966, p. 36).

The next major Wisconsinan advance involved three ice lobes that competed for space in the northern one-third of Indiana. The Saginaw and Erie Lobes (fig. 8) advanced across north-central Indiana and formed the Packerton Moraine (fig. 3). The Saginaw Lobe left behind a complex suite of deposits of till, ice-contact stratified drift, and outwash (Gray, 1989) and formed most of the lakes in Indiana (Wayne, 1966, p. 36). The Lake Michigan Lobe (fig. 8) flowed out of the basin of Lake Michigan and formed the Maxinkuckee Moraine (fig. 3). The lobe receded and built the Valparaiso Moraine (fig. 3) and a large outwash fan south of the moraine. The Erie Lobe crossed into northeastern Indiana and formed the Union City Moraine and a series of concentric moraines to the northeast of the Union City Moraine (fig. 3). The ice lobe primarily deposited a clay-rich till of the Lagro Formation between the moraines (Gray, 1989).

Since the recession of the glaciers from Indiana about 8,000 years ago, deposition has been minor. The principal postglacial change was the redistribution of sand and silt of the glacial flood plains into windblown dune and loess deposits. Muck, peat, and marl formed in swampy areas and **alluvial deposits** formed along the modern rivers (Wayne, 1966, p. 37).

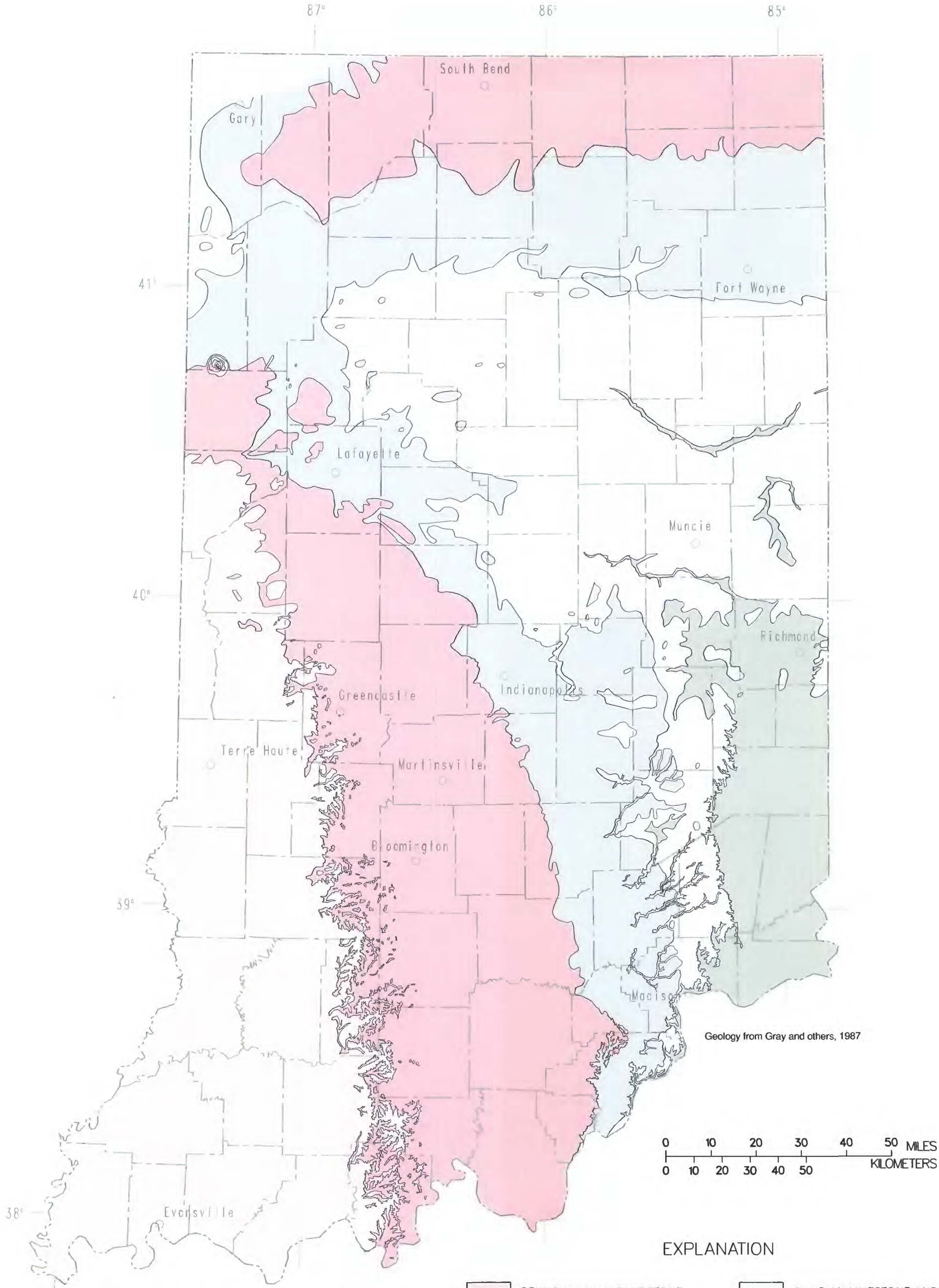
Hydrogeology and Ground-Water Flow

In northern Indiana, large areas of sand and gravel deposits in outwash plains and valley trains are capable of yielding as much as 2,000 gal/min of ground water. In addition, large yields (as much as several hundred gallons per minute) are available from productive Silurian and Devonian carbonate bedrock aquifers that underlie much of the area.

Significant ground-water resources are found in central Indiana along the valleys of the major rivers and streams (fig. 1). Intertill sand and gravel aquifers are present locally in the till plain throughout most of central Indiana. The Silurian and Devonian carbonate bedrock is a commonly used aquifer in central Indiana.

In the southern one-third of Indiana, major unconsolidated sources of ground water are limited to the valleys of the Wabash, White, Whitewater, and Ohio River systems (fig. 1). Mississippian, Devonian, and Silurian bedrock are sources of ground water in south-central and southeastern Indiana. Pennsylvanian sandstones are typically the most productive bedrock units in southwestern Indiana. Many areas in the southern one-third of the State lack adequate ground-water resources for purposes other than domestic (Clark, 1980, p. 34).

The productivity of different types of aquifers can differ greatly depending on certain fundamental characteristics. One fundamental characteristic of an aquifer is the ability to store water in pores. This **porosity** can be in the form of intergranular spaces as in sand and gravel; fractures and solution openings as in carbonate rocks; or intergranular spaces and fractures as in sandstones (Todd, 1980, p. 37-39). As unconsolidated sediment turns to stone, or becomes lithified, the original porosity of the sediment is reduced by cementation, compaction, and pressure solution (Davis, 1988, p. 325). Therefore, lithology is an important control on aquifer productivity, because it affects primary and secondary porosity and hydraulic conductivity. Different types of openings or pore spaces in geologic material are shown in figure 9. Openings that formed at the same time as the rock, such as pores in sedimentary deposits, are called primary openings (fig. 9a). Pores that formed after the rock is formed are called secondary openings (fig. 9b and 9c). The diameter of pores in sedimentary deposits can range from a few micrometers in clays to more than a centimeter in coarse gravel (Heath, 1988, p. 15) to the size of caves in carbonate rocks. An aquifer must be able to transmit water through such openings. This characteristic, called hydraulic conductivity, is dependent on the interconnected porosity of the material, the type of liquid, and the magnitude of the gravitational field (Lohman and others, 1972, p. 4). The hydraulic conductivity of an aquifer increases as grain size and the degree of sorting increase. Hydraulic conductivity is also usually greater in aquifers that have been enhanced by secondary porosity, such as fracturing. Finally, to be productive, an aquifer requires a source of water from precipitation or from adjacent geologic materials. Only 8 to 16 percent of the precipitation in Indiana, or about 3 to 8 inches per year, infiltrates into the ground-water system. Most of the precipitation is lost through evapotranspiration, and some runs off the land into surface waters (Bechert and Heckard, 1966, p. 100). Components of the hydrologic cycle are shown in figure 10.



Base from U.S. Geological Survey digital data, 1:500,000, 1983.
Lambert Conformal Conic Projection.
Standard parallels 33° and 45°, central meridian -86°

EXPLANATION

- PENNSYLVANIAN SANDSTONE, SHALE, LIMESTONE, AND COAL
- MISSISSIPPIAN SHALE, SILTSTONE, LIMESTONE, AND SANDSTONE
- DEVONIAN SHALE, LIMESTONE, AND DOLIMITE
- SILURIAN LIMESTONE AND DOLOMITE
- ORDIVICIAN SHALE AND LIMESTONE

Figure 6. Bedrock geology in Indiana.

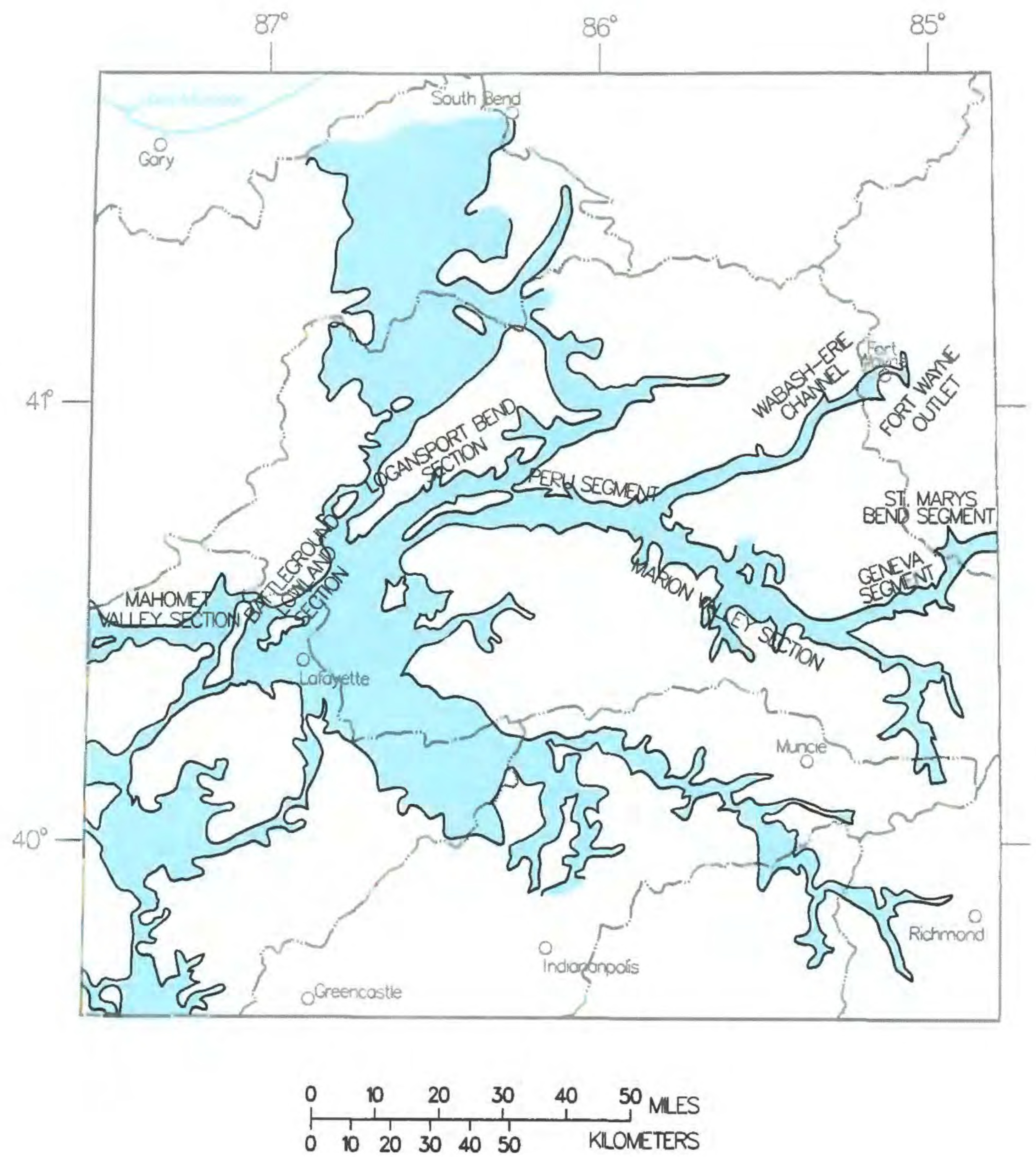


Figure 7. Location of buried bedrock valleys associated with the Lafayette Bedrock System in northern Indiana (modified from Bleuer, 1989).

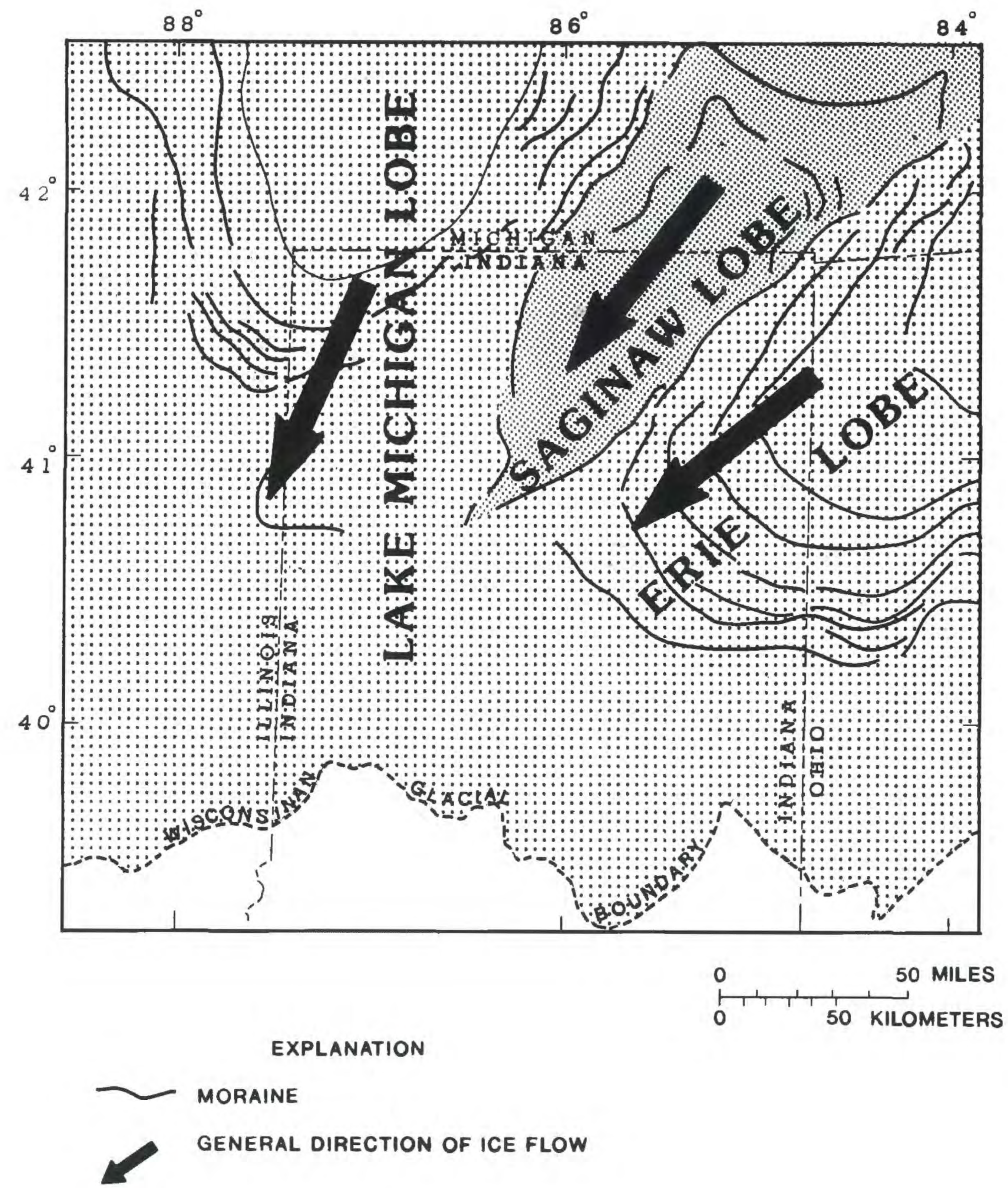
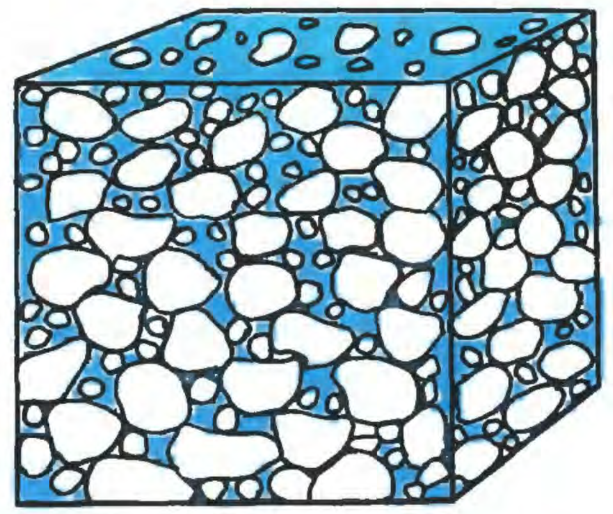


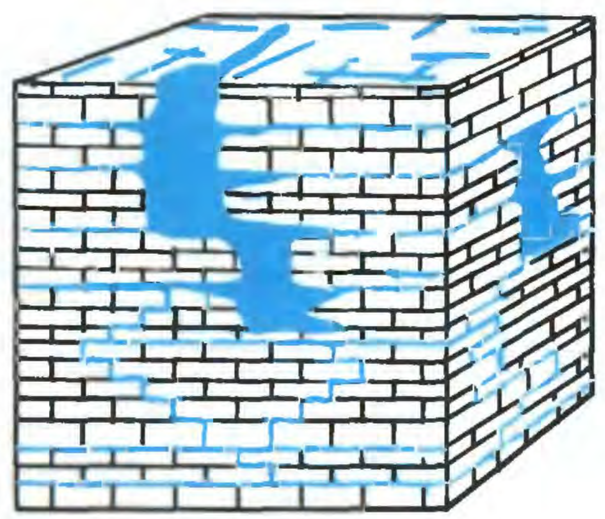
Figure 8. Primary glacial lobes and their principal directions of flow in Indiana during the Wisconsinan Age (modified from Wright and Frey, 1965).

Aquifers traversed by perennial streams commonly contain thick, extensive sand and gravel deposits that can produce large quantities of water. These aquifers are generally bounded by the flood-plain edge or by valley terraces. The hydraulic conductivity of sand and gravel buried in till or in buried bedrock valleys may be similar to these river-channel deposits, but recharge to the buried aquifers can be restricted by overlying sediments. In addition, buried sand and gravel deposits in till and preglacial valleys can be discontinuous and difficult to trace because of their complex depositional environments (Rosenshein, 1988, p. 167). Therefore, these buried deposits typically are less productive than the surficial deposits. The most productive carbonate rock aquifers contain large areas of solution openings along vertical joints and bedding planes as shown in figure 9. Similarly, sandstone aquifers are most productive where heavily jointed or fractured along bedding planes or where intergranular spaces have not been completely filled by cementing materials. Many sedimentary rocks—such as limestone, shale, and sandstone—have more vertical joints near the ground surface than at depth. This distribution of joints in the upper weathered zone tends to increase hydraulic conductivity locally by at least an order of magnitude. The hydraulic conductivity of most fractured rocks decreases rapidly with depth (Davis, 1988, p. 324). Coals can function as aquifers where fractures or cleats are well developed, resulting in increased hydraulic conductivity. Shales typically are not considered to be aquifers, because fractures in shale tend to be small and closely spaced, causing the hydraulic conductivity of the shale to be less than for most other rocks (Heath, 1984, p. 13; Todd, 1980, p. 38-41); however, in some cases, bedding-plane partings can create rather large horizontal hydraulic conductivities in shale.

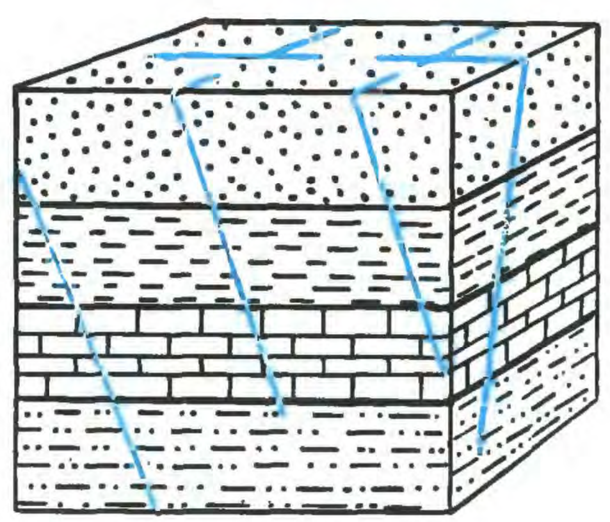
General rates of ground-water flow through geologic material can range from a few feet per second to less than a few feet per year (Todd, 1980, p. 92), although rates are typically on the order of a few feet to a few hundred feet per year. Ground-water flow can have three separate components: local, interbasin, and regional. A schematic diagram of local and regional ground-water discharge and recharge is shown in figure 10. Ground-water recharge is from precipitation, stream infiltration, infiltration from runoff and overland flow, and flow from adjoining aquifers. Ground-water discharge is by flow into surface water, evapotranspiration, and flow into adjoining aquifers. Local ground-water flow in shallow **unconsolidated aquifers** tends to be nearly horizontal, reflecting the surface topography, whereas regional flow patterns can range from simple (as in figure 10) to complex depending on the hydrologic properties of the aquifer. The direction of ground-water flow is determined by the potentiometric surface, an imaginary surface representing the **hydraulic head** distribution in an aquifer. The hydraulic head is the level that water will rise in a properly constructed well open to an aquifer. Ground water flows from areas of high head to areas of low head. Hydraulic head consists of two separate components: head due to elevation and head due to pressure. If elevation were the only component, then all ground water would simply flow from areas of high elevation to areas of low elevation (that is, downhill). Pressure head, however, is created by the weight of overlying water on a water particle. This pressure can create an upward component in the ground-water flow regime. Upward pressure is common near large rivers that function as regional ground-water **discharge areas**.



A. Pores in unconsolidated sand and gravel



B. Caverns and other solution-enlarged openings in limestone



C. Fractures in sedimentary rock

Figure 9. Types of openings in selected aquifers (modified from Heath, 1988).

Ground water in most surficial aquifers is at atmospheric pressure and is unconfined. Generally water only partly fills surficial aquifers, and water levels in the saturated zone are free to rise and decline (Heath, 1984, p. 6). All water pumped from an **unconfined aquifer** reaches the well through gravity draining, and the water is removed from storage.

Ground water in a **confined aquifer** fills the entire aquifer and is confined at higher than atmospheric pressure. The water is confined by overlying rocks and sediment that are substantially less permeable than the aquifer rocks. The pressure in a confined aquifer forces water in a well screened in the aquifer to rise above the top of the aquifer. The pressure gradient that moves water in a confined aquifer to the pumped well is created by the compression of the aquifer material and the expansion of the water as water is pumped from the aquifer. Some aquifers are under enough pressure to raise the potentiometric surface above the land surface; these wells are said to be "flowing" or "flowing artesian."

The direction of local ground-water flow can be opposite from the direction of regional flow, especially where aquifers are separated by geologic materials with low hydraulic conductivity. Generally, though, ground water tends to discharge to topographically low areas, such as rivers and wetlands. Discharge areas commonly cover 5 to 30 percent of the surface area of a watershed (Freeze and Cherry, 1979, p. 195-200).

Ground-Water Withdrawals

Most of Indiana has plentiful ground-water resources. The aquifers of Indiana contain approximately 100,000 Bgal of water (Bechert and Heckard, 1966, p. 106). Statewide ground-water withdrawals in 1991 were only about 0.20 percent of this estimated amount in storage (Indiana Department of Natural Resources, 1993, written commun.). At an average of 39 in/yr, total annual precipitation in Indiana is about 24,500 Bgal. Annual infiltration into Indiana aquifers, at 8 percent of precipitation, equals 1,960 Bgal/yr. Therefore, 1991 withdrawals (204 Bgal) were approximately 10 percent of annual recharge. Nearly 4,000 Bgal of water is stored in the outwash and alluvial aquifers of the White River and East Fork White River; estimated recharge to these aquifers is 2,500 Mgal/d (Bloyd, 1974, p. 12). In the two basins, withdrawals in 1991 were about 6 percent of the recharge and only about 1 percent of the ground water stored in the outwash and alluvial aquifers.

Total ground-water withdrawals from registered facilities in Indiana in 1991 was 204 Bgal, or an average of 559 Mgal/d. Withdrawals listed in table 1 are shown as totals and as daily averages (Indiana Department of Natural Resources, 1993, written commun.). Withdrawal-use categories in the table include energy, industrial, irrigation, public supply, rural supply, and miscellaneous. Categories not included in the table are domestic (self-supplied) and livestock. Also shown in table 1 is the capability of registered facilities to withdraw ground water from each basin. The withdrawal rates in table 1 reflect the demand for ground water, the availability of ground water, and the size and population of the basin.

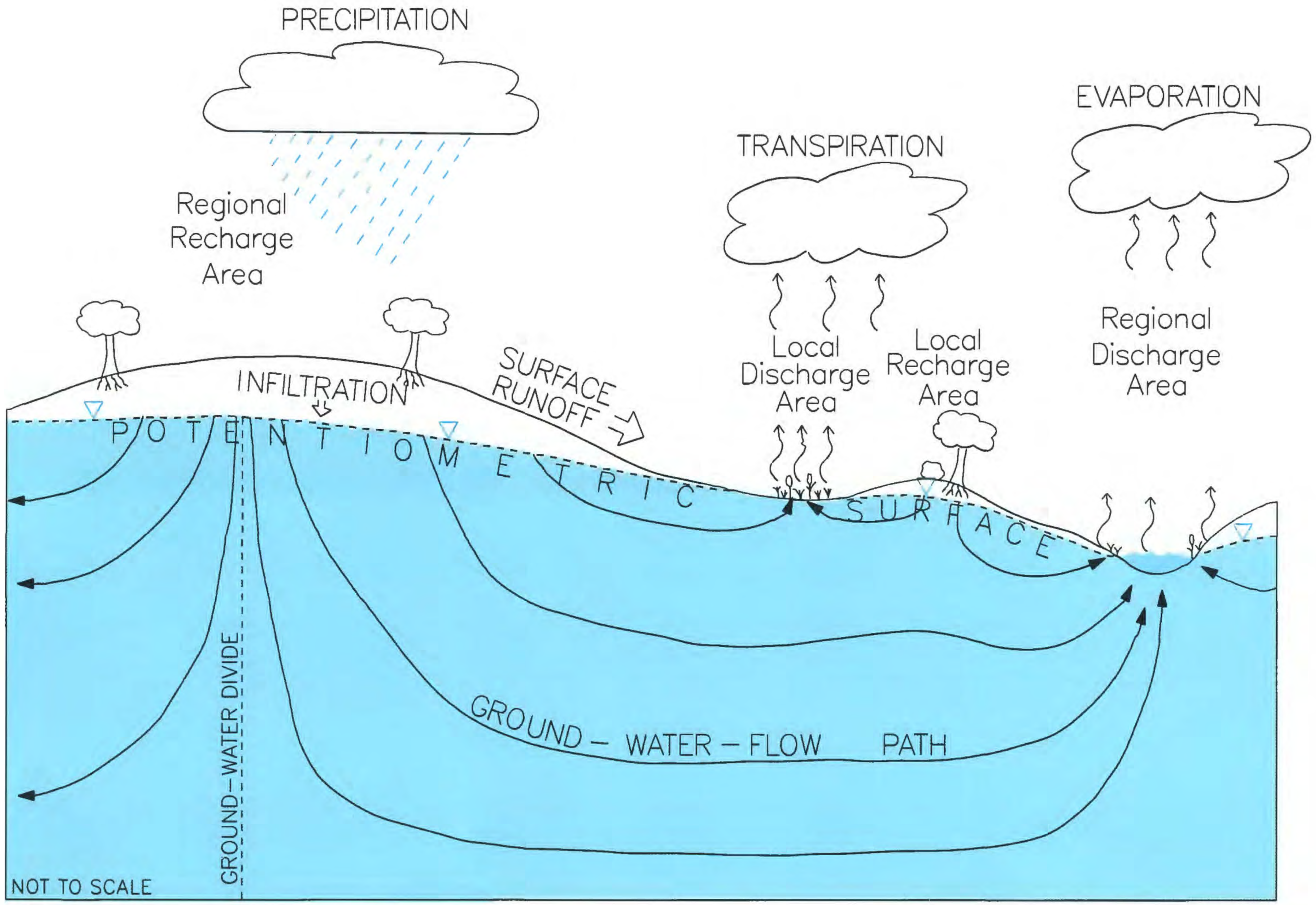


Figure 10. Generalized local and regional ground-water-flow paths and components of the hydrologic cycle.

More than 25,000 Mgal of ground water was withdrawn during 1991 in each of the White, St. Joseph, Middle Wabash, and Upper Wabash River basins. Amounts between 14,000 and 25,000 Mgal were pumped in the Ohio, Kankakee, and East Fork White River basins. Basins where between 3,000 to 7,500 Mgal were withdrawn in 1991 include Lake Michigan, Lower Wabash River, Whitewater River, and Maumee River. In the Patoka River basin, less than 32 Mgal of ground water were withdrawn in 1991.

METHODS OF STUDY

The Indiana Natural Resources Commission has divided the State into 12 water-management basins (fig. 1). The basin boundaries generally coincide with the surface-drainage divides of the major rivers in the State and with the State boundary. The management basins provide a hydrologic framework for surface-water and ground-water investigations in Indiana. The size of each basin (in square miles) is shown in table 2. Some river systems in Indiana drain into adjacent states. For these systems, the size of the water-management basin is not the same as the drainage area of the major river in the basin. These discrepancies are addressed in the individual discussions of each of the 12 basins.

Construction of Hydrogeologic Sections

Five to thirteen hydrogeologic sections were drawn for each basin to show the generalized hydro-stratigraphy. Stratigraphic details in hydrogeologic sections are from water-well records on file at the IDNR Division of Water; well-completion reports and lithologic logs on file at the IDNR Division of Oil and Gas; coal-test drilling records available from the IDNR Geological Survey; State and Federal highway drilling logs; USGS observation-well logs; and core samples collected by the IDNR. After drilling a hole, a driller is required by the State to file a form that lists information on lithology, water level, pumping rates, and other selected information on the well construction. All wells other than water-supply wells are labeled with a "t" on the hydrogeologic sections to indicate that they are test wells; dry water-supply wells or holes

Table 1. Ground-water withdrawals and pumping capability in Indiana, 1991
[Withdrawal and pumping-capability data are from Indiana Department of Natural Resources, 1993, written commun.; Mgal, million gallons; Mgal/d, million gallons per day]

Basin	Withdrawal		Pumping capability
	Daily (Mgal)	Annual (Mgal)	(Mgal/d)
White River	104.	37,800	524
St. Joseph River	87.1	31,800	607
Middle Wabash River	84.7	30,900	366
Upper Wabash River	74.8	27,300	518
Ohio River	55.9	20,400	263
Kankakee River	55.3	20,200	574
East Fork White River	39.2	14,300	254
Lake Michigan	19.4	7,090	96.4
Lower Wabash River	17.5	6,390	206
Whitewater River	11.5	4,200	53.6
Maumee River	9.34	3,410	81.6
Patoka River	.086	31.5	0.3
TOTAL	559	204,000	3,540

are labeled with a "d" on the sections. A few wells in each basin represent the combined lithologic data from shallow water-supply well logs and nearby deep test-hole logs. These combined wells are labeled as test wells on the hydrogeologic sections.

In the first five basins for which hydrogeologic sections were completed (Lake Michigan, St. Joseph, Maumee, Upper Wabash, and Lower Wabash), all sections were oriented roughly perpendicular to the major surface-water drainage in each basin to depict ground-water discharge to surface water. The remaining seven basins included one or more sections perpendicular to the other sections in the basin. For convenience, most hydrogeologic section lines run south to north or west to east. Logs of wells located within 1 mi of a hydrogeologic-section line were plotted at a density of one to three wells per mile.

Water levels shown on wells in the cross sections represent the hydraulic head in the aquifer in which the well is completed. Some of the water levels are connected to represent the generalized potentiometric surface in an aquifer that is being tapped by a group of wells along the section. Water levels were not indicated for all of the wells because they were not available from all drillers' logs. The locations of all hydrogeologic sections presented in the atlas are shown in figure 11. The number of section lines, length of section lines, and number of wells plotted for each section are listed in table 2.

The surface elevation shown on the hydrogeologic sections reflects the land surface at the well and does not necessarily portray the true topographic relief along the section line. For example, all the wells in a certain area might be located in a stream valley; no wells can be plotted in adjacent uplands even though the uplands are within the 2-mile width of the section line. Therefore, the surface topography depicts a valley and does not reflect the actual relief along the section line. The hydrogeologic sections are generally drawn to a depth of 300 ft below the land surface or at least 50 ft below the bedrock surface, whichever is greater. Only a small percentage of the water wells in the State are greater than 300 ft deep. Therefore, relatively little information at greater depths is available. Furthermore, use of water at depths greater than 300 ft is limited by low yields and salinity (W.J. Steen, Department of Natural Resources, 1990, written commun.; Indiana Department of Environmental Management, 1990, p. 223).

Aquifer types depicted in the hydrogeologic sections are sand and gravel, carbonate rock, sandstone, an upper weathered zone in low permeability rock, and interbedded bedrock material. Most bedrock was depicted as aquifer only where it is known to produce water. (This information is available from drillers' logs and previous studies.) If the bedrock formations are potentially water producing, then the material is mapped as "**aquifer—potential unknown**." Much of the complexly interbedded bedrock of Mississippian and Pennsylvanian age, the weathered zones, and some of the Silurian and Devonian carbonate rock was mapped as "aquifer—potential unknown" because of low yields, **dry holes**, or little knowledge about the productivity of the material. Almost all the Silurian and Devonian carbonate rock within 300 ft of land surface was depicted as aquifer even if no site-specific information was available to confirm its productivity. "Aquifer" and "aquifer—potential unknown" are colored on the hydrogeologic sections, whereas bedrock **nonaquifer material** and areas of unknown geologic material are not colored. In some sections, part of a formation is denoted as aquifer, but the rest is denoted as "aquifer—potential unknown" or nonaquifer—that is, a specific formation need not be hydrogeologically uniform throughout its extent. Unconsolidated material was grouped into two broad hydraulic categories: (1) sediments that have a relatively high hydraulic conductivity, such as sand and gravel; and (2) sediments that have a relatively low hydraulic conductivity, such as clay, silt, or mixed drift. Many of the sand and gravel deposits are aquifers, whereas the materials of low hydraulic conductivity were labeled as nonaquifer material. Areas where the geology is unknown are indicated by question marks.

Construction of Aquifer Maps

The maps showing lateral extent and continuity of aquifer types are based on interpretations of the hydrogeologic sections, on previously published surficial and bedrock geology maps, and on information available from previous studies of basin hydrogeology. The "Quaternary Geologic Map of Indiana" (Gray, 1989) was used to draw many of the surficial sand and gravel aquifers that were confirmed by the hydrogeologic sections. Bedrock geologic maps (Gray and others, 1987; Geosciences Research Associates, Inc., 1982; Gray, 1982) were used to indicate the extent of bedrock aquifers and the approximate boundaries where aquifers were buried by more than 300 ft of material.

The lithostratigraphic approach used in this study to define hydrogeologic settings resulted in seven aquifer types. Sand and gravel deposits are designated as surficial, buried, or discontinuous on the aquifer maps. **Surficial aquifer** indicates that the aquifer is covered by less than 10 ft of nonaquifer material. **Buried aquifer** indicates that the sand and gravel is covered by 10 ft or more of nonaquifer material and that the deposits are continuous in at least one direction for several miles or more. **Discontinuous aquifers** refer to lenses of sand and gravel that are not laterally extensive (discontinuous aquifers are typical of morainal areas). Buried bedrock valleys were mapped where productive aquifers are within them. The three unconsolidated aquifer types are shown as sand and gravel on the hydrogeologic sections. Carbonate rock (limestone and dolostone); sandstone; complexly interbedded sandstone, shale, siltstone, limestone, and coal; and an upper weathered zone in low-permeability rock are shown on the sections and on the aquifer maps. All of the complexly interbedded material shown on the sections is shown on the map as "aquifer—potential unknown" because of the uncertainty in mapping the water-producing zones on a regional scale. Areas typically devoid of formations capable of producing yields sufficient for domestic purposes are indicated as nonaquifer material on the maps.

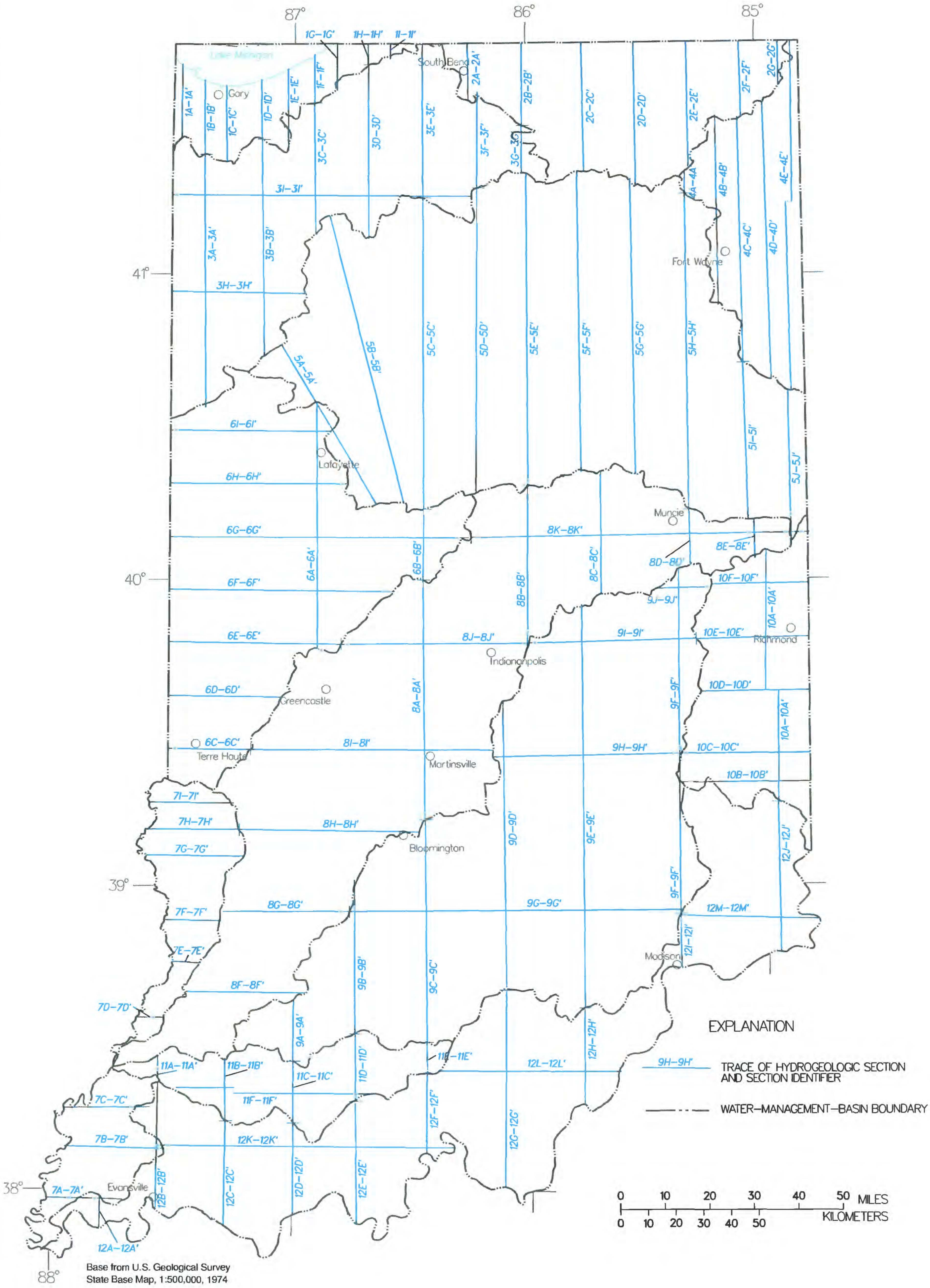


Figure 11. Location of hydrogeologic sections in the 12 water-management basins.

Table 2. Summary of basin areas and hydrogeologic section characteristics
[mi², square mile]

Basin	Area (mi^2)	Number of section lines	Length of section lines (in miles)	Total number of wells	Number of wells per mile
Lake Michigan	604	9	108	212	2.0
St. Joseph River	1,699	7	148	213	1.4
Kankakee River	2,989	9	339	490	1.4
Maumee River	1,283	5	244	305	1.3
Upper Wabash River	6,918	10	594	833	1.4
Middle Wabash River	3,453	9	347	470	1.4
Lower Wabash River	1,339	9	145	193	1.3
White River	5,603	11	409	354	0.9
East Fork White River	5,746	10	499	616	1.2
Whitewater River	1,425	6	189	226	1.2
Patoka River	862	6	99	98	1.0
Ohio River	4,224	13	375	214	0.6
TOTAL	36,145	104	3,496	4,225	[1]1.2

[1]Statewide average

Limitations of the Methods

In areas where large numbers of well logs were available for plotting, the logs with the location nearest the section line, the most complete well-record information, and the deepest hole were chosen for the hydrogeologic sections. This search for stratigraphic and hydrologic information resulted in a bias toward deep wells and deep aquifers in many locations.

The hydrogeologic sections represent the interpretation of the well-log data by the author(s) for each basin. The sections do not represent the only possible explanation or representation of the hydrogeology at that location. The reliability of the sections can vary within each basin depending on the quantity and quality of information on the drillers' logs. Logs were not always available at the desired density of two per mile. Some hydrogeologic sections include areas where only one well log is plotted in a 5-mile interval.

The interpretation of the well-log data on the hydrogeologic sections is a simplified picture of the geology on the section. Where lithologies change over short lateral distances, such as in the unconsolidated glacial deposits and the Pennsylvanian rocks, well logs spaced every 1/2 to 1 mi do not provide the needed information to depict detailed variations in geology. In the Pennsylvanian rocks, lithologies were commonly lumped together to avoid over-interpretation of well logs and, hence, a misleading and inaccurate representation of the system.

Many of the hydrogeologic sections contain logs of wells that were not drilled to the bedrock surface. In these areas, the topography of the bedrock surface was transferred from the "Map of Indiana Showing the Topography of the Bedrock Surface" (Gray, 1982). In these same areas, the bedrock geology and hydrostratigraphy was mapped with reference to the "Bedrock Geologic Map of Indiana" (Gray and others, 1987), the "Hydrogeologic Atlas of Indiana" (Geosciences Research Associates, Inc., 1982), structural maps of the base and top of the New Albany Shale in Indiana (Bassett and Hasenmueller, 1979a; 1979b), structural maps of the top of the Ordovician, Silurian, and sub-Pennsylvanian surfaces in Indiana (Bassett and Hasenmueller, 1980; Hasenmueller and Bassett, 1980; Keller, 1990), and other sources.

Water levels shown on the hydrogeologic sections were measured at different times by different drillers in aquifers that are not necessarily hydrologically connected. Water levels may also be a composite head from several aquifers or stratigraphic intervals, especially in uncased bedrock wells.

The aquifer map may not reflect the actual lateral extent and boundaries of the aquifer types. Because the section lines are 6 to 20 mi apart, the continuity of areas between the hydrogeologic section lines was extrapolated from the sections or inferred from published sources.

REFERENCES CITED

Back, William, Rosenshein, J.S., and Seaber, P.R., eds., 1988, Hydrogeology: Boulder, CO., Geological Society of America, The Geology of North America, v. O-2, 524 p.

Bassett, J.L., and Hasenmueller, N.R., 1979a, Map showing structure on base of New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources, Geological Survey, EGSP Series 800, scale 1:500,000.

_____1979b, Map showing structure on top of New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: State of Indiana Department of Natural Resources, Geological Survey, EGSP Series 801, scale 1:500,000.

_____1980, Map of Indiana showing structure on top of the Maquoketa Group (Ordovician): Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 812, scale 1:500,000.

Bechert, C.H., and Heckard, J.M., 1966, Ground water, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 100-115.

Bleuer, N.K., 1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley System (so-called Teays Valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

Bloyd, R.M., Jr., 1974, Summary appraisals of the Nation's ground-water resources—Ohio region: U.S. Geological Survey Professional Paper 813-A, 41 p.

Bugliosi, E.F., 1990, Plan of study for the Ohio-Indiana carbonate-bedrock and glacial aquifer system: U.S. Geological Survey Open-File Report 90-151, 25 p.

Casey, C.D., 1992, Hydrogeology of the basal confining unit of the carbonate aquifer system in the midwestern basins and arches region of Indiana, Ohio, Michigan, and Illinois: U.S. Geological Survey Open-File Report 92-489.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Davis, S.N., 1988, Sandstones and shales, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, Colo., Geological Society of America, The Geology of North America, v. O-2, p. 323-332.

Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, N.J., Prentice-Hall, Inc., 604 p.

Geosciences Research Associates, Inc., 1982, Hydrogeologic atlas of Indiana: Bloomington, Ind., 31 pl.

Gray, H.H., 1973, Properties and uses of geologic materials in Indiana: Indiana Department of Natural Resources, Geological Survey Regional Geologic Map Series, Supplementary Chart 1.

_____1982, Map of Indiana showing topography of the bedrock surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 35, scale 1:500,000.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Gutschick, R.C., 1966, Bedrock geology, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 1-20.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hasenmueller, N.R., and Bassett, J.L., 1980, Map of Indiana showing structure on top of Silurian rocks: Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 811, scale 1:500,000.

Heath, R.C., 1984, Ground-water regions of the United States: U.S. Geological Survey Water-Supply Paper 2242, 78 p.

_____1988, Hydrogeologic setting of regions, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, CO., Geological Society of America, The Geology of North America, v. O-2, p. 15-23.

Indiana Department of Environmental Management, 1987, Indiana ground water protection and management strategy: 127 p.

_____1990, Indiana water quality management 305(b) report for 1988-1989: Indiana Department of Environmental Management, Public Water Supply Section, 297 p.

Keller, S.J., 1990, Maps of southwestern Indiana showing geology and elevation of the sub-Pennsylvanian surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 51, scale 1:380,160.

Krothe, N.C., and Kempton, J.P., 1988, Central Glaciated Plains, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, CO., Geological Society of America, The Geology of North America, v. O-2, p. 129-132.

Lohman, S.W.; Bennett, R.R.; Brown, R.H.; Cooper, H.H., Jr.; Drescher, W.J.; Ferris, J.G.; Johnson, A.I.; McGuinness, C.L.; Piper, A.M.; Rorabough, M.I.; Stallman, R.W.; and Theis, C.V.; 1972, Definitions of selected ground-water terms—revisions and conceptual refinements: U.S. Geological Survey Water-Supply Paper 1988, 21 p.

National Oceanic and Atmospheric Administration, 1988, Climatological data—annual summary, Indiana: Asheville N.C., v. 93, no. 13.

Rosenshein, J.S., 1988, Region 18, alluvial valleys, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, CO., Geological Society of America, The Geology of North America, v. O-2, p. 165-175.

Rupp, J.A, 1991, Structure and isopach maps of the Paleozoic rocks of Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 48, 106 p.

Schaal, L.A., 1966, Climate, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 156-170.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Schnoebelen, D.J., 1992, Selected hydrogeologic data for the regional carbonate bedrock and glacial aquifers in eastern and central Indiana: U.S. Geological Survey Open-File Report 91-517, 102 p.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Todd, D.K., 1980, Groundwater hydrology: New York, John Wiley, 539 p.

U.S. Geological Survey, 1985, National water summary 1984—hydrologic events, related water-quality trends, and ground-water resources: U.S. Geological Survey Water-Supply Paper 2275, 467 p.

_____1988, National water summary 1986—hydrologic events and ground-water quality: U.S. Geological Survey Water-Supply Paper 2325, 560 p.

Visher, S.S., 1944, Climate of Indiana: Bloomington, Ind., Indiana University Publications, Science Series 13, 511 p.

Wayne, W.J., 1966, Ice and land—a review of the Tertiary and Pleistocene history of Indiana, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 21-39.

Wright, H.E., Jr., and Frey, D.G., eds., 1965, The Quaternary of the United States: Princeton, N.J., Princeton University Press, 922 p.

LAKE MICHIGAN BASIN

By Joseph M. Fenelon

General Description

The Lake Michigan basin, located in the far northwestern part of Indiana, encompasses a land area of 604 mi^2 (Hoggatt, 1975) within the northern halves of Lake and Porter Counties and the northern one-third of LaPorte County (fig. 12). In addition, the northern part of the basin includes a 241-mi^2 area beneath Lake Michigan. Within the basin is a major urban and industrial area that includes the cities of Gary, Hammond, East Chicago, and Merrillville.

Previous Studies

The first comprehensive evaluation of the ground-water resources in the Lake Michigan basin was done in a series of reports on the geohydrology and ground-water potential of Lake, Porter, and LaPorte Counties. Rosenshein (1961, 1962a) and Rosenshein and Hunn (1962) did preliminary evaluations of the ground-water resources in the three counties and tabulated well records for about 3,000 wells, including lithologic descriptions for about 1,200 of the wells and water-quality data from about 500 wells. The principal aquifers in Lake, Porter, and LaPorte Counties were described by Rosenshein (1963), Rosenshein and Hunn (1968a, 1968b), and Hunn and Reussow (1968). These authors determined and mapped the geometry and potentiometric surfaces of the aquifers, expected well yields, and general water quality. They also estimated hydraulic properties for the aquifers and associated confining units, and they determined sources and amounts of recharge to and discharge from the aquifers.

Hartke and others (1975) described the aquifers in Lake and Porter Counties, summarized ground-water usage in 1975 and the potential for future use, and qualitatively mapped the potential for aquifer contamination. General descriptions of the aquifers in the Lake Michigan basin have been written by Harrell (1935), the Great Lakes Basin Commission (1975), and Clark (1980). These authors described the major aquifers and the ground-water potential for the area.

A comprehensive study on the water resources of the Lake Michigan basin is being done by the Indiana Department of Natural Resources (J.E. Beaty, 1993, Indiana Department of Natural Resources, oral commun.). The Indiana Department of Natural Resources is characterizing the ground-water availability, use, flow, and quality, and mapping the bedrock elevation, and the geometry and areal extent of the primary aquifers.

Detailed studies have been done on the unconsolidated deposits of most of the Lake Michigan shoreline. Watson and others (1989) and Fenelon and Watson (1993) studied the surficial sand aquifer in most of Lake County north of the Little Calumet River. They mapped the bedrock surface and the geometry of the surficial aquifer, and they described the flow and water quality in the aquifer. Numerous studies have been done along Lake Michigan in the Indiana Dunes National Lakeshore, located between Gary and Michigan City (Meyer and Tucci, 1979; Shedlock and Harkness, 1984; Cohen and Shedlock, 1986; Wilcox and others, 1986; Shedlock and others, 1987; Thompson, 1987; Doss, 1991; Shedlock, Wilcox, Thompson, and Cohen, 1993; Shedlock, Cohen, Imbrigiotta, and Thompson, in press). Many of these studies are site specific, but together they constitute a detailed description of the unconsolidated aquifers and their associated flow paths and water quality within the Indiana Dunes National Lakeshore.

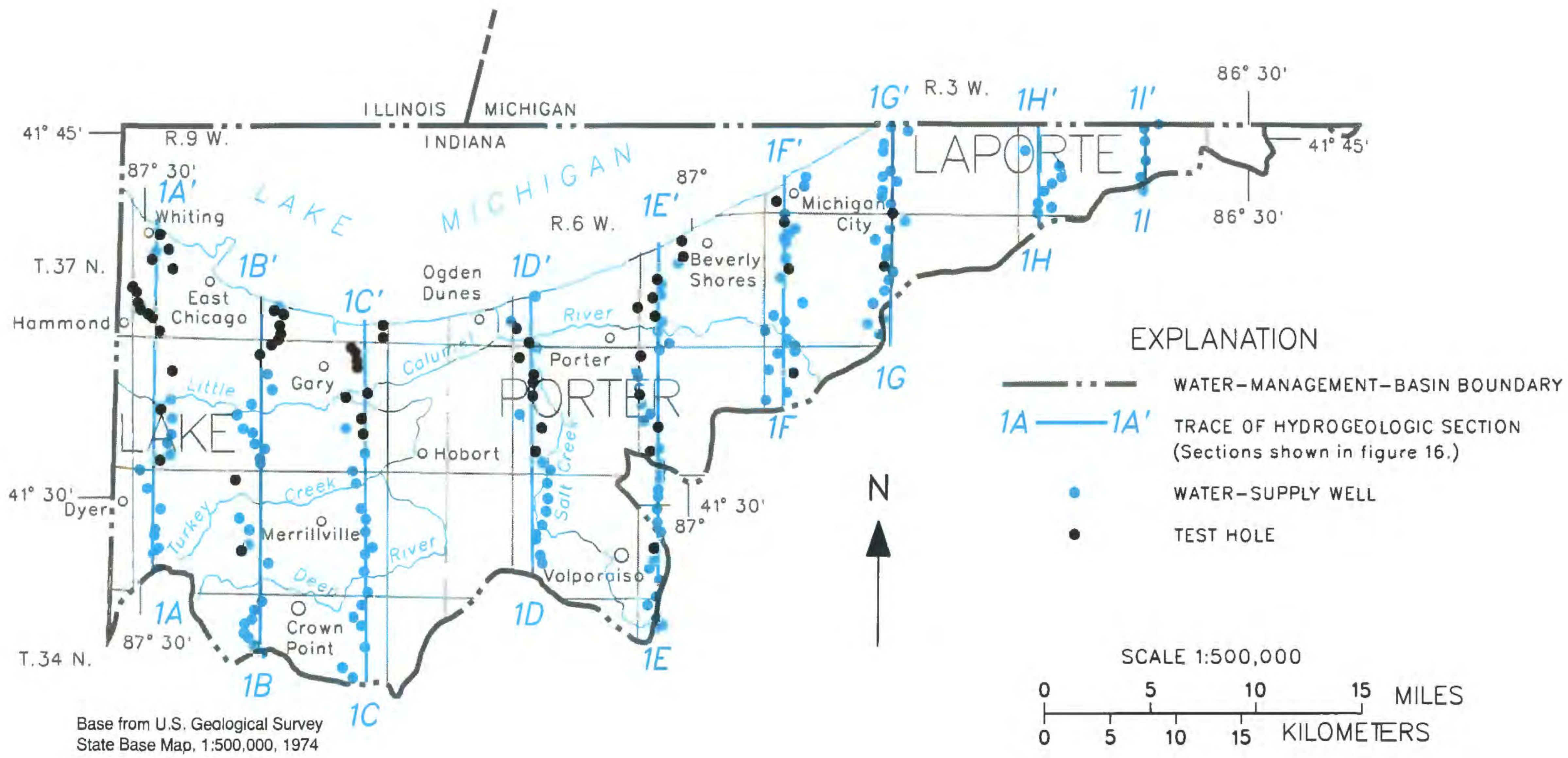


Figure 12. Location of section lines and wells plotted in the Lake Michigan basin.

Physiography

The Lake Michigan basin lies within the Calumet Lacustrine Plain, Valparaiso Morainal Area, and Kankakee Outwash and Lacustrine Plain, which are part of the Northern Moraine and Lake Region. The Kankakee Outwash and Lacustrine Plain, located in the extreme southeastern part of the basin, is discussed in the section on the Kankakee River basin in this report. The Calumet Lacustrine Plain (fig. 13), in the northern part of the Lake Michigan basin, occupies the lake bottom of the former glacial Lake Chicago—an extension of Lake Michigan in late Wisconsinan time (Bretz, 1955, p. 108). The lacustrine plain is not a completely flat area, but is a series of beach ridges, dunes, and interridge marshes. There are three dominant relict shorelines: the Glenwood, Calumet, and Toleston beach complexes, whose elevations are approximately 625, 607, and 600 ft above sea level, respectively (Thompson, 1987, p. 46-64). Relief in the Calumet Lacustrine Plain ranges from elevations greater than 650 ft above sea level in dunal areas associated with ancient beaches to approximately 580 ft above sea level on the present day Lake Michigan shoreline.

South of the Calumet Lacustrine Plain is the Valparaiso Morainal Area (fig. 13), composed of an arc-shaped end moraine complex that parallels the southern shore of Lake Michigan from Illinois, through northwestern Indiana, and into Michigan. The morainal complex is made up of several terminal moraines of Wisconsinan age including the Valparaiso and Tinley Moraines (fig. 13), which mark terminal positions of the Lake Michigan Lobe (Bretz, 1955, p. 106-108). The Valparaiso Morainal complex is about 150 ft higher than the Calumet Lacustrine Plain and forms a major divide that separates drainage to the Mississippi River from drainage to the Saint Lawrence River by way of Lake Michigan. Elevations in the complex generally range from 700 to 800 ft above sea level and are as high as 950 ft above sea level. The western end of the complex is wide and gently undulating, whereas the part of the complex east of Valparaiso, is more hilly and rugged (Schneider, 1966, p. 51-52).

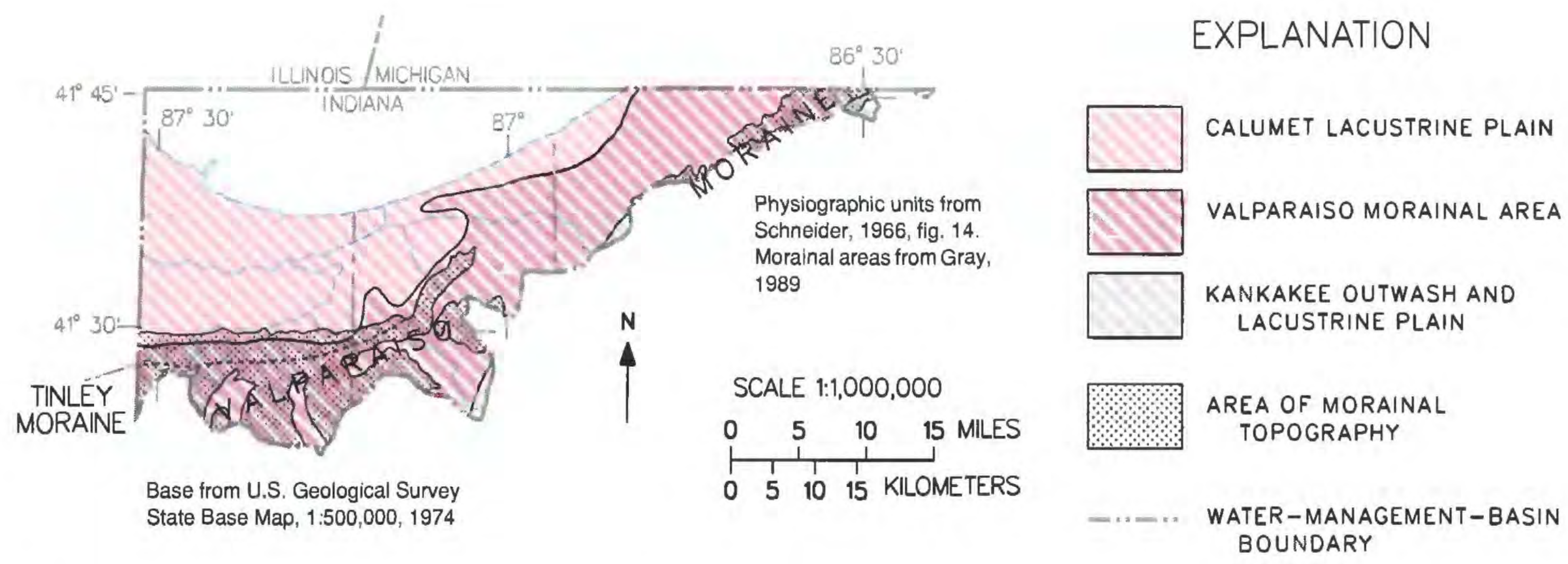


Figure 13. Physiographic units and moraines in the Lake Michigan basin.

Surface-Water Hydrology

The entire drainage area for Lake Michigan is approximately 67,900 mi^2 and includes 44,330 mi^2 of land in Indiana, Illinois, Wisconsin, and Michigan (Great Lakes Basin Commission, 1975, p. 21). Within Indiana, the Lake Michigan basin has an area of 845 mi^2, of which 604 mi^2 is land. The basin is drained in Indiana primarily by the Little Calumet River (fig. 12), which flows approximately parallel to the Lake Michigan shoreline and discharges to Lake Michigan through a ditch on the western side of Porter County. The major tributaries to the Little Calumet River are Turkey Creek, Deep River, and Salt Creek. Each tributary originates on the Valparaiso Moraine and flows north to the Little Calumet River. The eastern part of the Lake Michigan basin in LaPorte County is drained by smaller creeks that flow directly into Lake Michigan.

Geology

Bedrock Deposits

Overlying Precambrian bedrock in the Lake Michigan basin is more than 4,000 ft of sedimentary bedrock (Rosenshein and Hunn, 1968a, p. 7; Hartke and others, 1975, p. 4) that dips northeast at about 10 to 20 ft/mi. About 3,500 ft of the sedimentary bedrock is of Cambrian or Ordovician age. The Cambrian and Ordovician bedrock consists of about 2,000 ft of fine- to coarse-grained sandstone in the lower part and shale overlying dolomite and sandstone in the upper part (Rosenshein and Hunn, 1968a, p. 9; Hartke and others, 1975, p. 4). Overlying these rocks are Silurian rocks in the western part of the Lake Michigan basin and Silurian, Devonian, and Mississippian rocks further east (figs. 5 and 14).

The rocks of Silurian age, which consist of 400 to 600 ft of dolomite and some limestone (Great Lakes Basin Commission, 1975, p. 37), are divided into the Sexton Creek Limestone, the Salamonie Dolomite, and the Salina Group (Shaver and others, 1986). The Silurian rocks are composed of shaley to pure and fine- to coarse-grained carbonate rocks that include reef facies in the upper part.

The Devonian rocks consist of dolomite and limestone overlain by shale; these rocks contain the Muscatatuck Group and the Antrim and Ellsworth Shales. The Muscatatuck Group overlies the Silurian carbonate rocks; it is absent where Silurian rocks are exposed at the bedrock surface, and it is as much as 200 ft thick elsewhere (Shaver, 1974, p. 5). The Group is composed of a wide variety of impure to pure and dense to coarse-grained dolomite and limestone; in places, it contains anhydrite and gypsum in its lower part (Shaver and others, 1986, p. 99). The Antrim Shale, a brownish-black noncalcareous shale, overlies the Devonian carbonate rocks in the northeastern part of the basin (Shaver and others, 1986, p. 5). The Ellsworth Shale overlies the Antrim Shale and is of Devonian and Mississippian age. It is a grayish-green shale that contains limestone or dolomite lenses in its upper part (Shaver and others, 1986, p. 42).

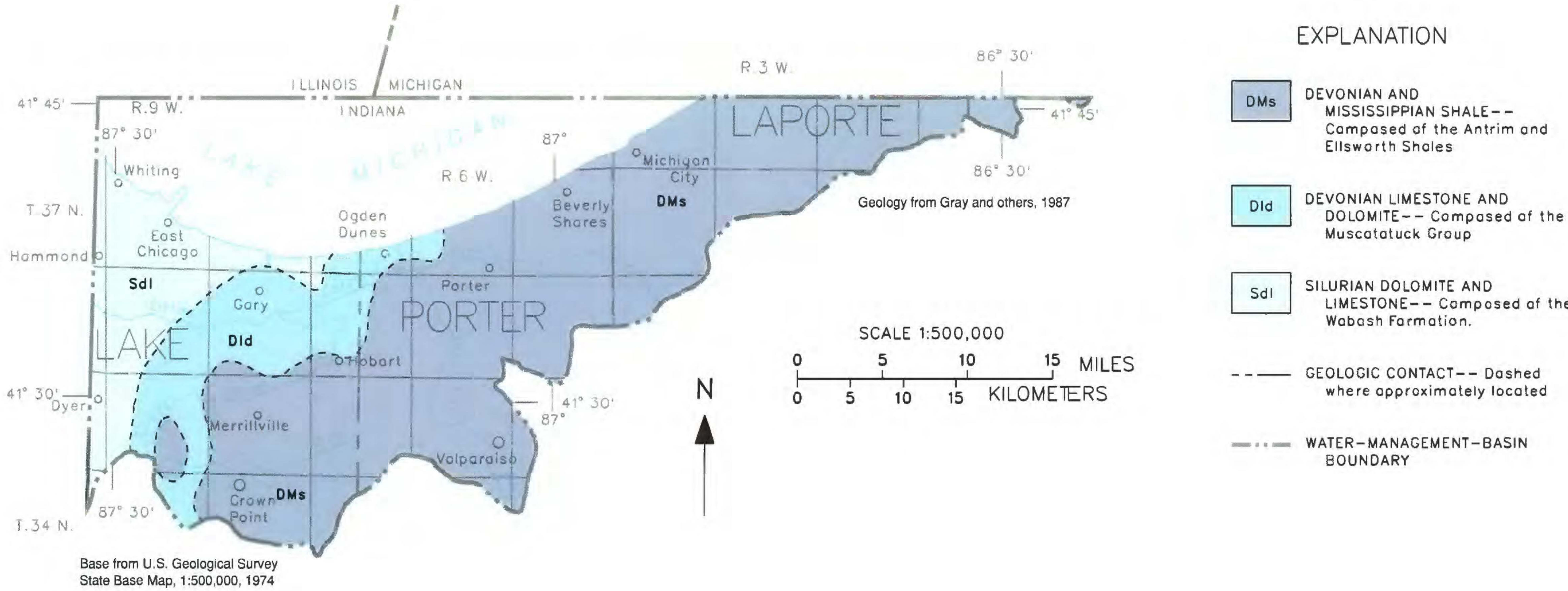
The bedrock surface is a preglacial erosional feature that has been further modified by glacial erosion. The Silurian and Devonian carbonate rocks exposed at the bedrock surface contain significant fractures and solution features in the upper 100 ft (Rosenshein and Hunn, 1968a, p. 10; Great Lakes Basin Commission, 1975, p. 24; Hartke and others, 1975, p. 4).

Unconsolidated Deposits

The unconsolidated deposits in the Lake Michigan basin are largely the result of glacial, glaciofluvial, shallow-water coastal and lake, wetland, and wind-blown sedimentation. They consist of clay-rich till, sand and gravel outwash, sand beaches and dunes, lake silt and clay, and peat. Thicknesses of unconsolidated deposits range from about 50 ft near the Indiana-Illinois State line to about 350 ft at the basin divide south of Michigan City (fig. 15). The Lake Michigan basin is overlain in most areas by two or more of four general unconsolidated units (Rosenshein, 1962b; Vig, 1962; Hunn and Reussow, 1968; Rosenshein and Hunn, 1968a, 1968b; Hartke and others, 1975).

The lowest unit overlies bedrock and is primarily a dense, clay-loam till that contains zones of intertill sand and gravel. This unit, which ranges in thickness from 0 to more than 100 ft, was formed by Wisconsinan and possibly pre-Wisconsinan glaciers that advanced through the basin. The basal part of the unit contains 0 to 15 ft of sand and gravel that fill the deepest parts of preglacial bedrock valleys.

Figure 14. Bedrock geology of the Lake Michigan basin.

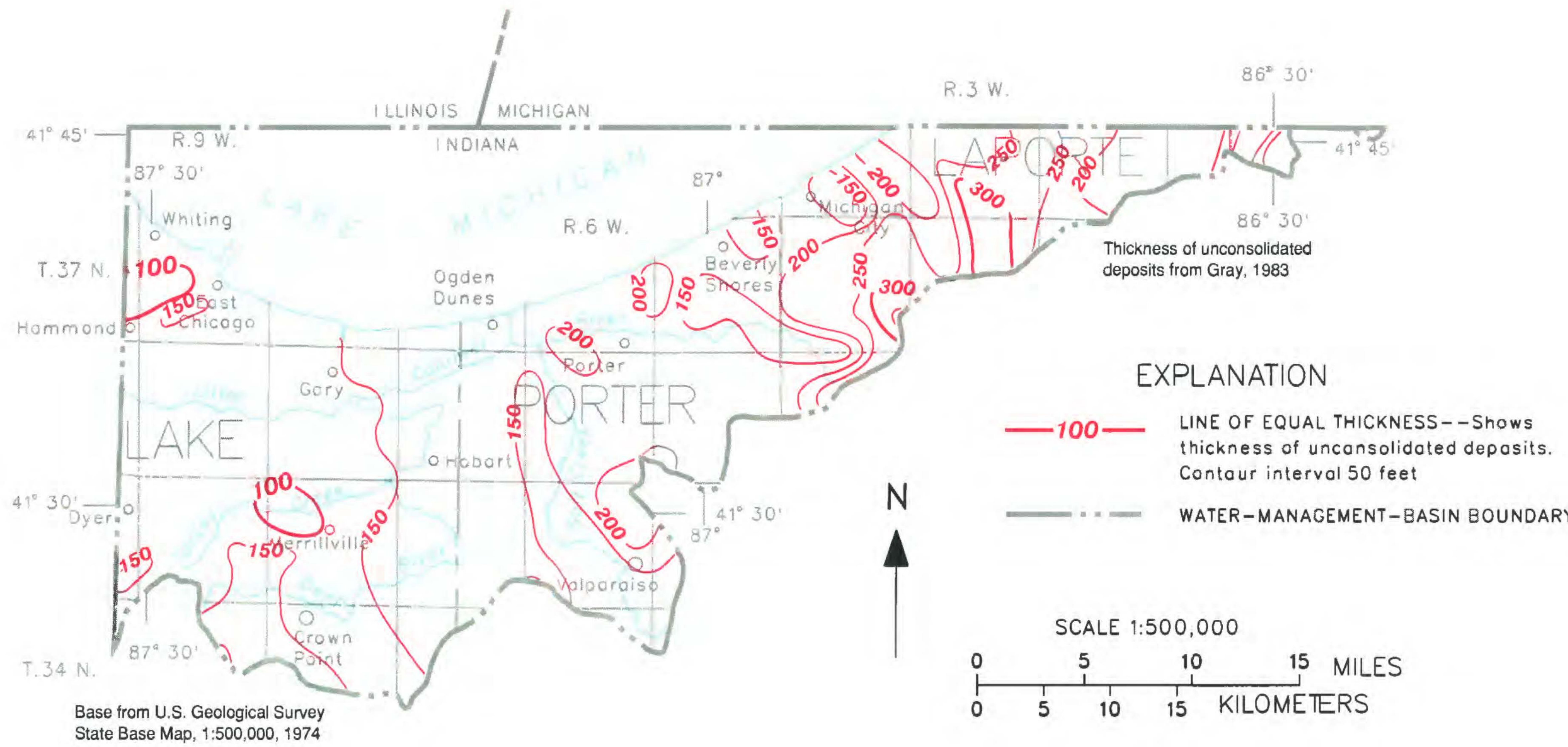


Glaciofluvial sand ranging from 0 to 100 ft in thickness overlies the basal till in the southern one-half of the basin. Locally, the sand contains sand and gravel interbeds and clay.

Overlying most of the glaciofluvial sand in the southern one-half of the Lake Michigan basin is a till that extends to the surface. The till ranges from 0 to more than 50 ft in thickness and contains an intertill sand and gravel layer (Rosenshein, 1962b, p. 129). The surficial till is similar lithologically but less dense than the basal till.

Overlying the basal till in the northern one-half of the basin is primarily fine to medium glaciolacustrine and wind-blown sand with some beach gravel, local peat, and lake silt and clay deposits. Most of these deposits were formed in association with glacial Lake Chicago, and they range in thickness from 0 to 70 ft (Rosenshein, 1962b, p. 129; Rosenshein and Hunn, 1968b, p. 17).

Figure 15. Thickness of unconsolidated deposits in the Lake Michigan basin.



Aquifer Types

Nine hydrogeologic sections (1A–1A′ to 1I–1I′) were produced for this atlas to show the general hydrostratigraphy of the Lake Michigan basin (fig. 16). All sections are oriented from south to north, approximately perpendicular to the Lake Michigan shoreline (fig. 12). Section lines were drawn at 5- to 7-mi intervals. A total of 212 well logs were used to produce the sections; 58 well logs are from test holes that are unrelated to water use. The average density of plotted wells along the section lines is 2.0 wells per mile.

In general, supplies of ground water throughout the Lake Michigan basin are adequate. Unconsolidated sands and gravels form the most productive aquifers. The primary unconsolidated aquifers are the buried sands and gravels overlying the basal till in the southern one-half of the basin and the surficial sands overlying the basal till in the northern one-half of the basin. Other less significant unconsolidated aquifers are discontinuous buried sands and gravels in the northeastern part of the basin, discontinuous surficial sands and gravels on top of the Valparaiso Morainal complex in the southeastern part of the basin, buried bedrock valley aquifers near Lake Michigan, and intertill sands and gravels throughout the basin. Carbonate bedrock underlies the basin but is used as an aquifer only in the far western part of the basin. The primary aquifers are shown in figure 17. Aquifers beneath Lake Michigan are excluded from the map and from most of the discussion because of insufficient data. Table 3 summarizes the four aquifer units mapped in figure 17. The table lists ranges of thickness and yield and names other authors have used to define the units.

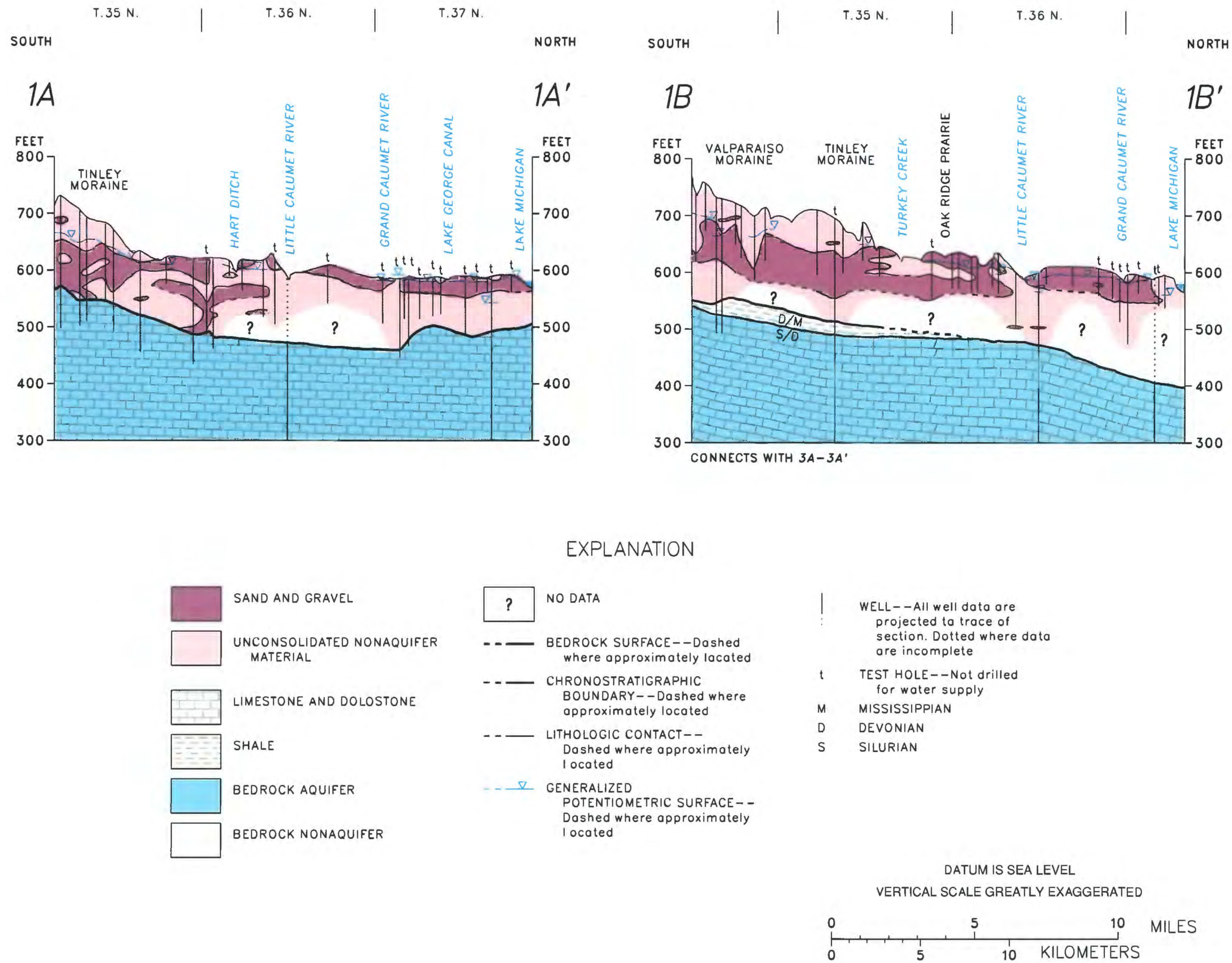
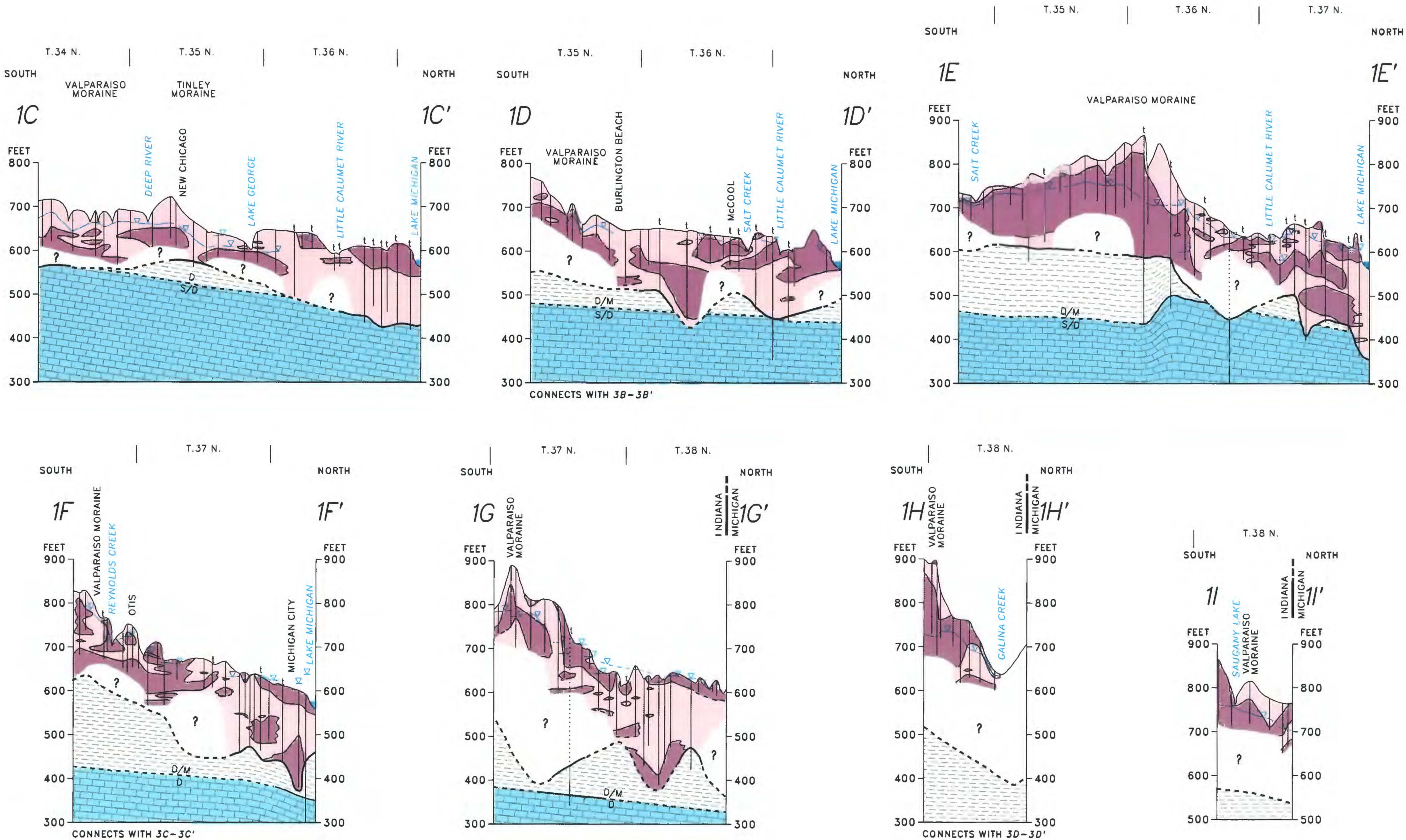


Figure 16. Hydrogeologic sections *1A–1A′* to *1I–1I′* of the Lake Michigan basin.


1C
1C'
SOUTH
NORTH
T.34 N.
T.35 N.
T.36 N.
VALPARAISO MORAINE
TINLEY MORAINE
DEEP RIVER
NEW CHICAGO
LAKE GEORGE
LITTLE CALUMET RIVER
LAKE MICHIGAN
FEET
D
S/D
1D
1D'
VALPARAISO MORAINE
BURLINGTON BEACH
McCOOL
SALT CREEK
LITTLE CALUMET RIVER
LAKE MICHIGAN
D/M
S/D
CONNECTS WITH 3B-3B'
1E
1E'
T.35 N.
T.36 N.
T.37 N.
VALPARAISO MORAINE
SALT CREEK
LITTLE CALUMET RIVER
LAKE MICHIGAN
D/M
S/D
1F
1F'
T.37 N.
VALPARAISO MORAINE
REYNOLDS CREEK
OTIS
MICHIGAN CITY
LAKE MICHIGAN
D/M
D
CONNECTS WITH 3C-3C'
1G
1G'
T.37 N.
T.38 N.
VALPARAISO MORAINE
INDIANA
MICHIGAN
D/M
D
1H
1H'
T.38 N.
VALPARAISO MORAINE
GALINA CREEK
INDIANA
MICHIGAN
CONNECTS WITH 3D-3D'
1I
1I'
T.38 N.
SAUGANY LAKE
VALPARAISO MORAINE
INDIANA
MICHIGAN

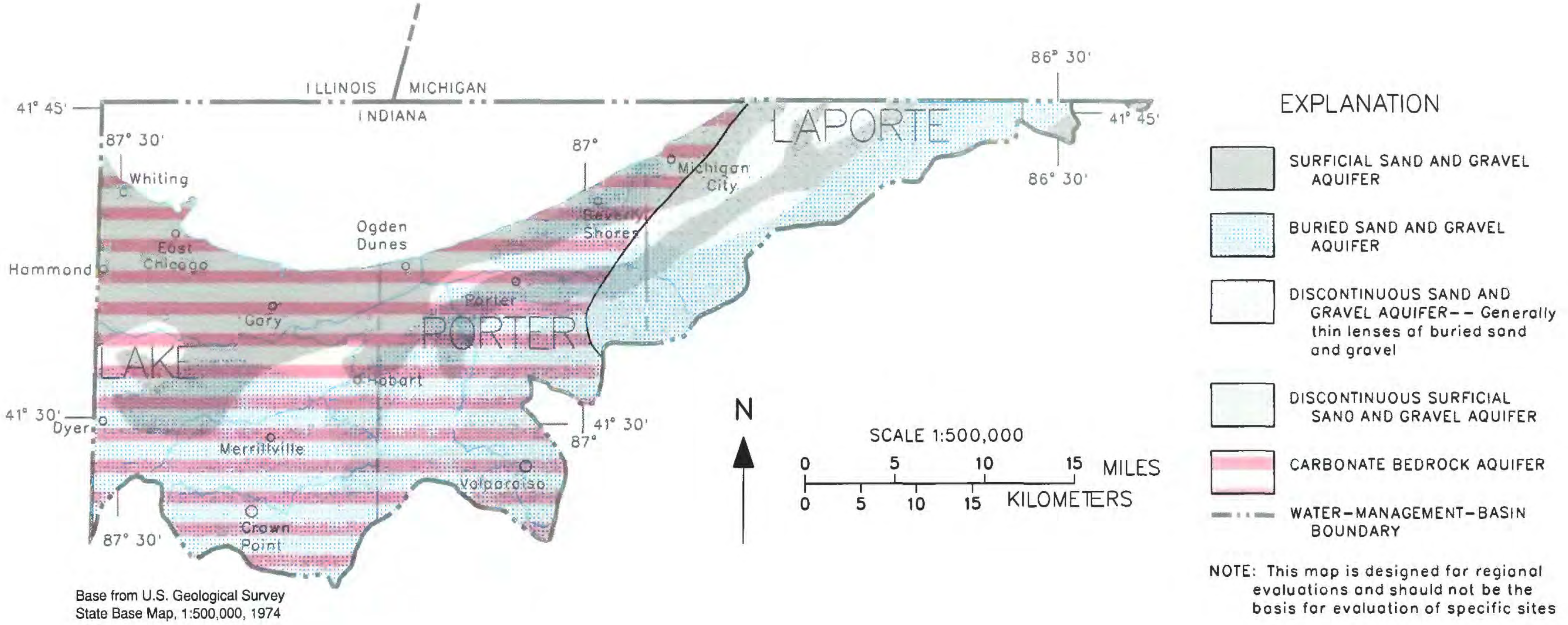


Figure 17. Extent of aquifer types in the Lake Michigan basin.

Unconsolidated Aquifers

Buried Sand and Gravel Aquifers

Buried sand and gravel aquifers are found in more than three-fourths of the basin (fig. 17). Most of the buried aquifers are part of a glaciofluvial sand aquifer, which is shown on the southern one-half to two-thirds of sections 1A–1A′ to 1F–1F′ (fig. 16). The aquifer as shown in the sections consists of sand bodies that are continuous for 3 to 6 mi in a north-south direction. These sand bodies are probably more continuous parallel to Lake Michigan and the Valparaiso Moraine. The buried glaciofluvial aquifer is as much as 200 ft thick (of which 150 ft is saturated) beneath the Valparaiso Moraine in section 1E–1E′ (fig. 16). The typical thickness penetrated by wells is about 50 ft; however, aquifer thicknesses may be greater because most of the wells do not penetrate the full thickness of the aquifer.

The glaciofluvial sand aquifer is overlain by a surficial till in most areas on the Valparaiso Morainal complex. Till thickness ranges from 0 to about 100 ft and is typically 20 to 50 ft. The aquifer is recharged primarily from the overlying till. The aquifer discharges to the land surface through the overlying till and to the bedrock through a basal till (Rosenshein and Hunn, 1968b). In sections 1F–1F′ and 1G–1G′ (fig. 16), north of the Valparaiso Moraine, hydraulic heads in the buried sand and gravel are above land surface; flowing wells can be found in these areas.

Several localized buried aquifers are between sections 1D–1D′ and 1F–1F′ in the northern part of the basin. These aquifers have been studied by a number of investigators (Wilcox and others, 1986; Shedlock and others, 1987; Shedlock, Wilcox, Thompson, and Cohen, 1993; Shedlock, Cohen, Imbrigiotta, and Thompson, in press) and have been named the "subtill" and "basal" aquifers. The "subtill" aquifer, shown on section 1E–1E′ in T. 37 N. (fig. 16), is buried beneath a surficial till and overlies another till. The aquifer extends almost 5 mi and is about 30 ft thick in section 1E–1E′. Beneath the underlying till in places is the "basal" aquifer, which extends about 2 mi in section 1E–1E′ and is about 50 ft thick.

Unmapped intertill sands and gravels in the basal and surficial tills in the southern one-half of the basin also contribute water to wells locally. Yields of some of these aquifers are high, but the aquifers are not extensive (Rosenshein and Hunn, 1968a; 1968b).

Surficial Sand and Gravel Aquifer

The unconsolidated surficial aquifer in the northern one-half of the basin is composed of glaciolacustrine and wind-blown sand. The aquifer extends south about 2 to 5 mi from the Lake Michigan shoreline in the eastern part of the basin and up to 10 mi from the shoreline in the western part of the basin. Thicknesses range from 0 to 70 ft and average about 30 ft. The surficial aquifer is recharged primarily from precipitation and from ground water flowing up from the basal till in the eastern part of the basin. Most discharge goes to streams, ditches, Lake Michigan, and to evapotranspiration (Rosenshein and Hunn, 1968b; Shedlock and others, 1987; Fenelon and Watson, 1993). The aquifer is used very little as a source of water in the western part of the basin because of its proximity to Lake Michigan, the major source of drinking water in the area. The aquifer is also rarely used because of its thin saturated zone (20 to 30 ft) and its susceptibility to contamination (Hartke and others, 1975, p. 25; Fenelon and Watson, 1993). The aquifer is tapped in the eastern part of the basin by households that do not have access to a public water supply or a better source of ground water.

The surficial aquifer extends for an undetermined distance beneath Lake Michigan. Sand or gravel has been found on the lake bottom at widely spaced sampling points within the State boundaries (Schneider and Keller, 1970). The distribution of sand and gravel indicates that the aquifer could extend more than 5 mi into the Lake, although insufficient data are available to map the areal extent or thickness.

Discontinuous Sand and Gravel Aquifers

Discontinuous buried sand and gravel aquifers are present in the northeastern part of the basin and are shown in the central part of section 1G–1G′ (fig. 16). Although the deposits are discontinuous, they are common enough that sources of domestic water supply are easy to find. Some of the discontinuous aquifers are in the deep bedrock valleys in the northeastern part of the basin. Section 1G–1G′ (fig. 16) crosses several of the valleys that are 200 to 300 ft below land surface and have about 100 ft of relief. The bottom of one of the valleys is filled with sand and gravel, which is tapped for a drinking-water supply; however, in general, these deep aquifers are not tapped unless they are the only unconsolidated aquifers in the area.

Table 3. Characteristics of aquifer types in the Lake Michigan basin
[>, greater than; <, less than; locations of aquifer types shown in fig. 17]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand	0- 70	[1,2]10- 500	Calumet aquifer[3]; Unit 1[1,2]
Buried sand and gravel	10- 200	[1,2,4]10- >500	Valparaiso aquifer[3]; Unit 3[1,2]; Subtill and basal aquifers[5]
Discontinuous sand and gravel	0- >100	0- 500	Unit 3[1,2]
Carbonate rock	400- 800	[1]<10- 400	Silurian-Devonian carbonate aquifer

[1]Rosenshein and Hunn, 1968a.
[2]Rosenshein and Hunn, 1968b.
[3]Hartke and others, 1975.
[4]Clark, 1980.
[5]Shedlock, Cohen, Imbrigiotta, and Thompson, in press.

Discontinuous surficial deposits are present throughout the southeastern part of the basin. Most of these deposits are in morainal areas and are a complex mix of ice-contact stratified drift and till. The deposits may contain surficial aquifers, which can be local to extensive and range from a few feet to more than 100 ft in thickness as in T. 37 N. of section 1G–1G′ (fig. 16). The discontinuous surficial aquifers are usually not used as a source of water because sources in the underlying buried aquifers are more reliable.

Bedrock Aquifers

Carbonate Bedrock Aquifer

Silurian and Devonian carbonate bedrock is the principal bedrock aquifer in the Lake Michigan basin. Although the carbonate bedrock aquifer is present throughout the basin (including the part covered by Lake Michigan), it is used mostly in the western part of the basin (sections 1A–1A′ and 1B–1B′ in fig. 16). Generally, wells penetrate only the upper 100 ft of the permeable carbonate bedrock. East of section 1B–1B′ (fig. 16), the aquifer is little used because of the availability of water in the unconsolidated aquifers, the increasing depth to the bedrock aquifer, and the potentially high concentrations of dissolved solids (Rosenshein and Hunn, 1968b, p. 5). The area where the carbonate bedrock aquifer is generally less than 300 ft below land surface is mapped in figure 17; however, the aquifer is not necessarily used in this area. For the same reasons as listed above, the aquifer is generally not used where more than 200 ft of unconsolidated material and(or) shale cover it, and is rarely used where there is more than 300 ft of cover. Depth to the carbonate bedrock ranges from about 100 ft below land surface in section 1A–1A′ to about 500 ft in section 1I–1I′ (fig. 16). Devonian and Mississippian shales, which overlie the carbonate bedrock aquifer in the eastern part of the basin (sections 1B–1B′ to 1I–1I′, fig. 16), are potential sources of small quantities of water (Rosenshein and Hunn, 1968a; 1968b).

Devonian and Mississippian shales and(or) the lower till unit confine the top surface of the carbonate bedrock aquifer throughout most of the basin. The bottom surface of the carbonate bedrock aquifer is underlain at depths of greater than 600 ft by 200 to 250 ft of shale interbedded with some limestone (Gray, 1972). The underlying shale functions as a confining unit for a deeper Ordovician and Cambrian aquifer system (not shown on fig. 17 or in hydrogeologic sections). Although the deeper aquifer is tapped extensively in northeastern Illinois (Schicht and others, 1976; Visocky and others, 1985), it is tapped very little in the Lake Michigan basin of Indiana because of small well yields and high construction and pumping costs (Rosenshein and Hunn, 1968a). An even deeper aquifer, consisting of about 2,000 ft of sandstone of early Cambrian age, is highly saline and is used only for deep waste injection (Hunn and Reussow, 1968; Hartke and others, 1975).

Summary

The Lake Michigan basin includes 604 mi^2 of land within the northern parts of Lake, Porter, and LaPorte Counties and a 241-mi^2 area beneath Lake Michigan in northwestern Indiana. The land is covered by a sandy lake plain overlying a basal till in the northern part of the basin and the till-capped Valparaiso Morainal complex to the south. The till covers a glaciofluvial sand that overlies a basal till. Silurian and Devonian carbonate bedrock is present as subcrop in the western part of the basin but is overlain by Devonian and Mississippian shales further east.

Four aquifer types were delineated in the Lake Michigan basin: (1) surficial sand and gravel, (2) buried sand and gravel, (3) discontinuous sand and gravel, and (4) carbonate bedrock. The primary aquifer, present in the southern one-half of the basin, is a buried glaciofluvial sand aquifer whose average thickness is 50 ft; it can yield more than 500 gal/min. The aquifer, which is covered by 0 to 100 ft of surficial till, overlies a basal till. Other buried sand and gravel aquifers are found in the northeastern part of the basin and in some of the buried bedrock valleys. A thin but extensive surficial sand aquifer in the northern one-half of the basin is little used because of its proximity to Lake Michigan, the major source of drinking water in the area. Discontinuous buried aquifers in the northeastern part of the basin are commonly used as a source of water. Discontinuous surficial aquifers are present throughout the southeastern part of the basin but are rarely tapped because of more reliable buried sources of ground water. The principal bedrock aquifer is a carbonate bedrock aquifer, used only in the far western part of the basin. The aquifer's yields are generally less than those of the unconsolidated aquifer material but can be substantial in some areas.

Regional ground-water discharge from all the aquifers is to Lake Michigan. Locally, however, ground water is commonly discharged through evapotranspiration or by flow to streams, ditches, and pumped wells in the area.

References Cited

Bretz, J.H., 1955, Geology of the Chicago region, part II—the Pleistocene: Illinois State Geological Survey Bulletin 65, 132 p.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Cohen, D.A., and Shedlock, R.J., 1986, Shallow ground-water flow, water levels, and quality of water, 1980-84, Cowles Unit, Indiana Dunes National Lakeshore: U.S. Geological Survey Water-Resources Investigations Report 85-4340, 25 p.

Doss, P.K., 1991, Physical and chemical dynamics of the hydrogeologic system in wetlands along the southern shore of Lake Michigan: Dekalb, Ill., Northern Illinois University, Doctoral dissertation, 267 p.

Fenelon, J.M., and Watson, L.R., 1993, Geohydrology and water quality of the Calumet aquifer in the vicinity of the Grand Calumet River/Indiana Harbor Canal, northwestern Indiana: U.S. Geological Survey Water-Resources Investigations Report 4115, 151 p.

Gray, H.H., 1972, Lithostratigraphy of the Maquoketa Group (Ordovician) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 7, 31 p.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Great Lakes Basin Commission, 1975, Great Lakes basin framework study—appendix 3, geology and ground-water: Ann Arbor, Mich., 152 p.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hartke, E.J., Hill, J.R., and Reshkin, Mark, 1975, Environmental geology of Lake and Porter Counties, Indiana—an aid to planning: Indiana Department of Natural Resources, Geological Survey Special Report 11, Environmental Study 8, 57 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hunn, J.D., and Reussow, J.P., 1968, Preliminary evaluation of the ground-water resources in the Calumet-Kankakee hydrologic area—appendix to the State Water Plan: Indiana Department of Natural Resources, Division of Water Resources, 39 p.

Meyer, William, and Tucci, Patrick, 1979, Effects of seepage from fly-ash settling ponds and construction dewatering on ground-water levels in the Cowles Unit, Indiana Dunes National Lakeshore, Indiana: U.S. Geological Survey Water-Resources Investigations 78-138, 95 p.

Rosenshein, J.S., 1961, Ground-water resources of northwestern Indiana—preliminary report, Lake County: Indiana Department of Conservation, Division of Water Resources, Bulletin 10, 229 p.

_____1962a, Ground-water resources of northwestern Indiana—preliminary report, Porter County: Indiana Department of Conservation, Division of Water Resources Bulletin 12, 131 p.

_____1962b, Geology of Pleistocene deposits of Lake County, Indiana: U.S. Geological Survey Professional Paper 450-D, Article 157, p. D127-D129.

_____1963, Recharge rates of principal aquifers in Lake County, Indiana: Ground Water, v. 1, no. 4, p. 13-20.

Rosenshein, J.S., and Hunn, J.D., 1962, Ground-water resources of northwestern Indiana—preliminary report, LaPorte County: Indiana Department of Conservation, Division of Water Resources Bulletin 13, 183 p.

_____1968a, Geohydrology and ground-water potential of Lake County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 31, 35 p.

_____1968b, Geohydrology and ground-water potential of Porter and LaPorte Counties, Indiana: Indiana Department of Conservation, Division of Water Resources, Bulletin 32, 22 p.

Schicht, R.J., Adams, J.R., and Stall, J.B., 1976, Water resources availability, quality, and cost in northeastern Illinois: Illinois State Water Survey Report of Investigation 83, 90 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Schneider, A.F. and Keller, S.J., 1970, Geologic map of the 1^{o} x 2^{o} Chicago Quadrangle, Indiana, Illinois, and Michigan, showing bedrock and unconsolidated deposits: Indiana Geological Survey Regional Geologic Map 4, Chicago Sheet, Part B, scale 1:250,000.

Shaver, R.H., 1974, The Muscatatuck Group (new Middle Devonian name) in Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 3, 7 p.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Shedlock, R.J., and Harkness, W.E., 1984, Shallow ground-water flow and drainage characteristics of the Brown ditch basin near the East Unit, Indiana Dunes National Lakeshore, Indiana, 1982: U.S. Geological Survey Water-Resources Investigations Report 83-4271, 37 p.

Shedlock, R.J., Loiacono, N.J., and Imbrigiotta, T.E., 1987, Effects of ground water on the hydrochemistry of wetlands at Indiana Dunes, northwest Indiana, *in* Interdisciplinary approaches to wetlands research: Michigan State University Press, Wilcox, D.A., and Herrman, R., eds., East Lansing, Mich., p. 37-55.

Shedlock, R.J., Wilcox, D.A., and Thompson, T.A., and Cohen, D.A., 1993, Interactions between ground water and wetlands, southern shore of Lake Michigan, USA: Journal of Hydrology, v. 141, p. 127-155.

Shedlock, R.J., Cohen, D.A., Imbrigiotta, T.E., and Thompson, T.A., in press, Hydrogeology and hydrochemistry of dunes and wetlands along the southern shore of Lake Michigan, Indiana: U.S. Geological Survey Open-File Report 92-139.

Thompson, T.A., 1987, Sedimentology, internal architecture and depositional history of the Indiana Dunes National Lakeshore and State Park: Bloomington, Ind., Indiana University, Doctoral dissertation, 129 p.

Vig, R.J., 1962, Geology of the unconsolidated deposits of Lake County, Indiana: Grand Forks, N.D., University of North Dakota, unpublished Masters thesis, 56 p.

Visocky, A.P., Sherrill, M.G., and Cartwright, Keros, 1985, Geology, hydrology, and water quality of the Cambrian and Ordovician systems in northern Illinois: Illinois State Geological Survey and Illinois State Water Survey Cooperative Groundwater Report 10, 136 p.

Watson, L.R., Shedlock, R.J., Banaszak, K.J., Arihood, L.D., and Doss, P.K., 1989, Preliminary analysis of the shallow ground-water system in the vicinity of the Grand Calumet River/Indiana Harbor Canal, northwestern Indiana: U.S. Geological Survey Open-File Report 88-492, 45 p.

Wilcox, D.A., Shedlock, R.J., and Hendrickson, W.H., 1986, Hydrology, water chemistry, and ecological relations in the raised mound of Cowles Bog: Journal of Ecology, v. 74, p. 1103-1117.

ST. JOSEPH RIVER BASIN

By Kathleen K. Fowler

General Description

The St. Joseph River basin, which encompasses an area of 1,699 mi^2 in northeastern Indiana, is part of the St. Lawrence drainage system. The basin includes all of Lagrange County, most of Elkhart, Steuben, and Noble Counties, and parts of St. Joseph, Kosciusko, and Dekalb Counties. The St. Joseph River flows into Indiana in Elkhart County and flows out of the State in St. Joseph County. Major cities within the basin are South Bend, Mishawaka, Elkhart, Goshen, Kendallville, and Angola (fig. 18).

Previous Studies

Studies of the ground-water resources of the St. Joseph River basin have been published since the late 1890's. Leverett (1899) compiled a report on the wells of northern Indiana that was used as a reference for many years. Capps (1910) produced one of the first investigations of the quantity, quality, and distribution of ground water in north-central Indiana. A summary of ground-water sources and occurrence for each county of the State was compiled by Harrell (1935). Klaer and Stallman (1948), Stallman and Klaer (1950), and Rosenshein and Hunn (1962) described ground-water resources of the South Bend area, Noble County, and St. Joseph County. The geohydrology and ground-water potential of St. Joseph County was also described by Hunn and Rosenshein (1969). Pettijohn (1968) provided data on the occurrence, availability, and chemical quality of ground water in the St. Joseph River basin as part of a comprehensive report of the ground-water resources of Indiana. Marie (1975) studied the water-supply potential of the aquifers in the South Bend area. Reussow and Rohne (1975) compiled three plates describing the availability, distribution, quality, and use of water in the St. Joseph River basin. The ground-water resources and quality of northwestern Elkhart County were evaluated by Imbrigiotta and Martin (1981). Bailey and others (1985) and Lindgren and others (1985) described local hydrologic systems in the St. Joseph River basin and modeled the effects of agricultural irrigation on those local systems. Crompton and others (1986) reviewed the hydrologic data-collection network in the St. Joseph River basin. Peters (1987) and Peters and Renn (1988) described the effects of agricultural irrigation on the water resources of the St. Joseph River basin. Ground-water and surface-water availability, distribution,

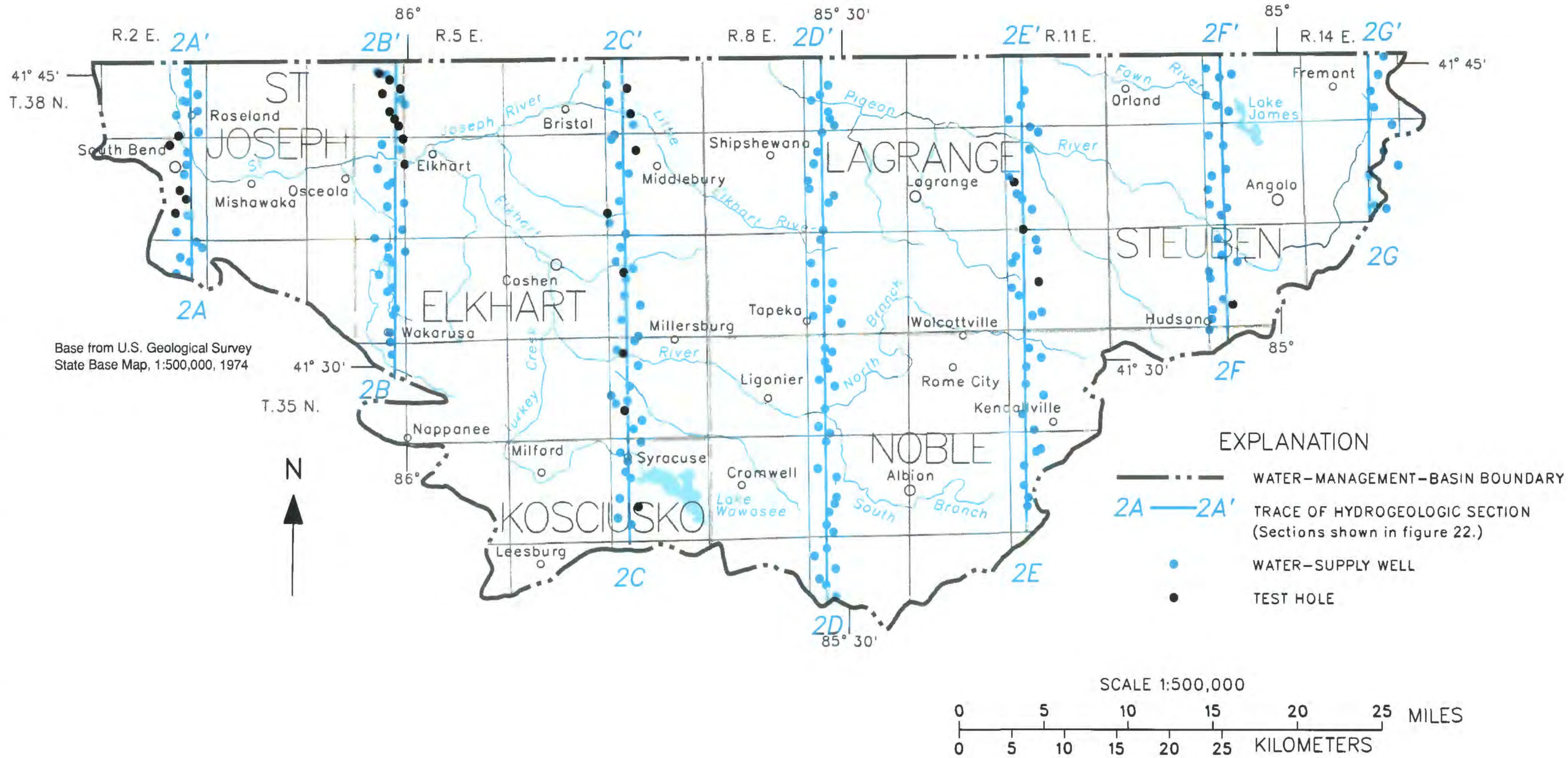


Figure 18. Location of section lines and wells plotted in the St. Joseph River basin.

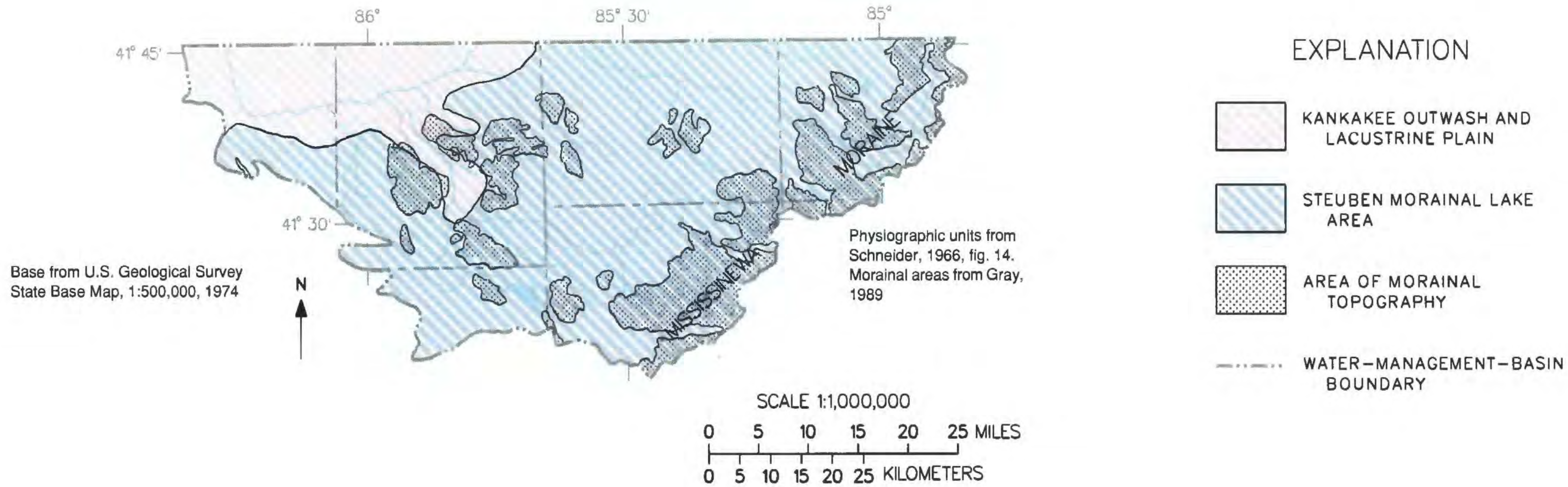


Figure 19. Physiographic units and moraines in the St. Joseph River basin.

quality, and use in the St. Joseph River basin were described by the Indiana Department of Natural Resources (1987).

Physiography

The St. Joseph River basin is part of the Northern Moraine and Lake Region physiographic area as described by Malott (1922, p. 112) and Schneider (1966, p. 42). This region has been divided into five physiographic units. Two units, the Kankakee Outwash and Lacustrine Plain and the Steuben Morainal Lake Area, compose the St. Joseph River basin (fig. 19).

The topography of the basin is variable; land-surface altitudes range from 700 ft near South Bend to 1,100 ft north of Angola. The Kankakee Outwash and Lacustrine Plain, recently described as the Kankakee Lowlands (Indiana Department of Natural Resources, 1987, p. 12), lies in the northwestern part of the basin and extends into eastern Illinois and southwestern Michigan. Prior to the glacial retreat from the Great Lakes drainage basin, the St. Joseph River drained southwest to the Kankakee River. After the St. Lawrence River became free of ice, the St. Joseph River drainage was captured and drainage direction reversed downstream of South Bend.

Much of this area is a poorly drained, level plain covered by fine-grained alluvium and underlain by thick outwash sand and gravel. Much of the sand and gravel was deposited in the form of broad valley trains and outwash plains by glacial meltwaters at several different times during the late Wisconsin glaciation (Schneider, 1966, p. 52). Sand transported by the wind, and formed into dunes, overlies the outwash in parts of this area.

The Morainal Lake Area, located south and east of the Kankakee Outwash and Lacustrine Plain, has a more complex physiography. Glacial and post-glacial activity have produced the present landforms. Some of the glacial landforms include: knob-and-kettle topography (which forms basins for the many lakes and peat bogs characteristic of northeastern Indiana), kames (composed of ice-contact sand and gravel deposits), meltwater channels, small lake plains, and dunes (Schneider, 1966, p. 52-53). The Maxinkuckee Moraine and the Packerton Moraine (fig. 3) near the southwestern and southern edges of the basin have traditionally been recognized as the major uplands formed by the recession of the Saginaw Lobe (fig. 8). Recent evidence indicates that the Maxinkuckee Moraine was formed by the recession of the Lake Michigan Lobe (fig. 8) and that the Packerton Moraine is a recessional feature of the Erie Lobe (Bleuer and Melhorn, 1989, p. 44). The Mississinewa Moraine (fig. 19), located along the southeastern border of the basin, was formed by the recession of the Erie Lobe (fig. 8). These and other morainic uplands (fig. 19) are composed of unsorted material in a clay matrix. The lowlands between moraines are remnants of meltwater channels and are composed of thick sand and gravel deposits (Crompton and others, 1986, p. 7). Local relief of 100 to 150 ft is common; relief in areas of kame deposits can be as much as 200 ft.

Surface-Water Hydrology

The St. Joseph River drains 1,699 mi^2 in Indiana and 2,586 mi^2 in southern Michigan. The river begins in a morainal area near Hillsdale, Mich., flows generally to the southwest, then to the north through South Bend, Ind., and empties into Lake Michigan near Benton Harbor, Mich. Forty-one miles of the St. Joseph River are in Indiana (fig. 18). The average channel gradient is approximately 2.5 ft/mi (Indiana Department of Natural Resources, 1987, p. 1 and 21). The major tributaries of the St. Joseph River are the Elkhart, Pigeon, and Fawn Rivers. Minor tributaries with greater than 100 mi^2 of drainage include North Branch of the Elkhart River, Turkey Creek, Little Elkhart River, and South Branch of the Elkhart River.

Geology

Bedrock Deposits

Four thousand feet of sandstones, siltstones, shales, limestones, and dolomites of Cambrian, Ordovician, Silurian and Devonian ages overlie Precambrian igneous and metamorphic basement rocks in the St. Joseph River basin. Paleozoic shale and limestone are present at the bedrock surface throughout the St. Joseph River basin in Indiana (fig. 20). A gently rolling bedrock surface is interspersed with a few entrenched, preglacial valleys. Bedrock is overlain by thick glacial drift throughout the basin. Bedrock formations dip northeast into the Michigan Basin (fig. 4) at approximately 30 ft/mi (Indiana Department of Natural Resources, 1987, p. 15).

Devonian carbonate rocks of the Muscatatuck Group are present at the bedrock surface in relatively small areas in Kosciusko and Noble Counties in the extreme southern part of the basin (fig. 20). The Devonian carbonate rocks and underlying Silurian carbonate rocks form a 400 to 900 ft thick carbonate rock sequence (Bassett and Hasenmueller, 1978 and 1979). Overlying the carbonate rock is the Antrim Shale of Devonian age. The Antrim Shale is present

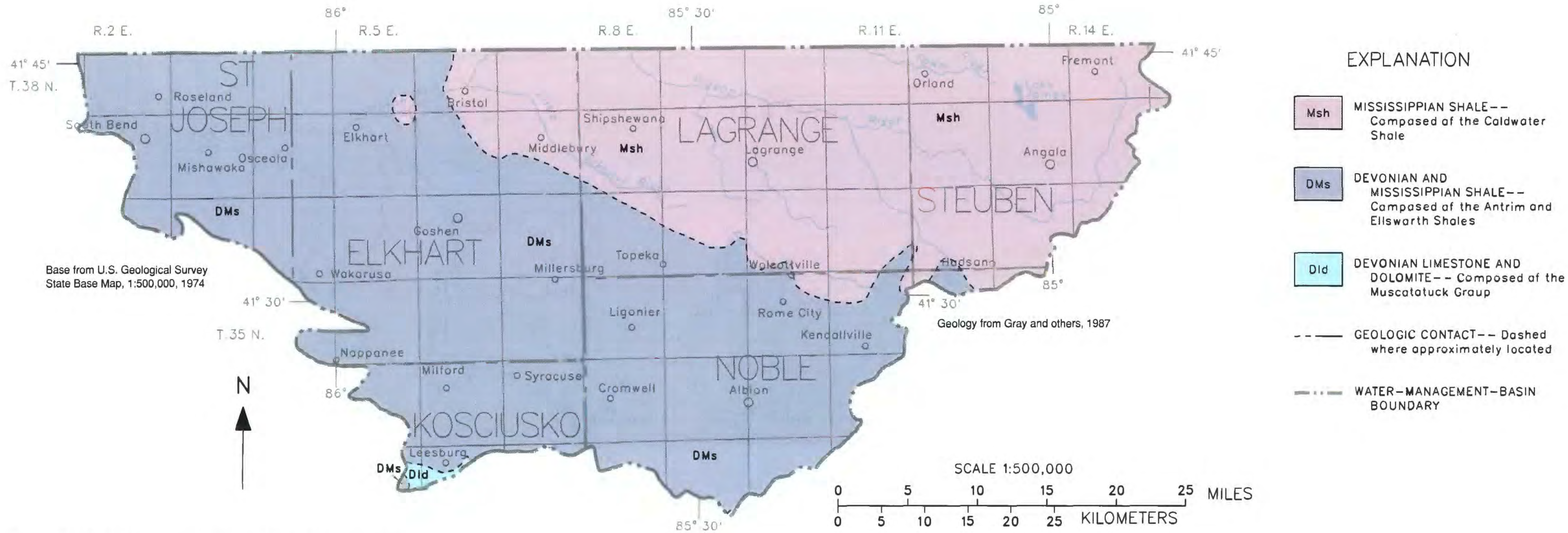


Figure 20. Bedrock geology of the St. Joseph River basin.

at the bedrock surface in the southern part of the basin. This shale is typically brownish-black and noncalcareous. Thickness of the Antrim Shale ranges from 0 to greater than 220 ft in the St. Joseph River basin. The Ellsworth Shale of Late Devonian and Early Mississippian age overlies the Antrim Shale, and is present at the bedrock surface in the western part of the basin. The lower part of the formation consists of alternating layers of gray-green shale and brownish-black shale. The upper part is a grayish-green shale that contains limestone and dolomite lenses. The Ellsworth Shale ranges in thickness from less than 40 ft to greater than 200 ft. The Coldwater Shale of Early Mississippian age is present in the northeastern part of the basin. It is predominantly a gray to greenish-gray, slightly silty shale with red shale stringers in the lower part. Thickness ranges up to 500 ft. (See Shaver and others, 1986.)

Unconsolidated Deposits

The St. Joseph River basin is covered by thick, unconsolidated glacial deposits (fig. 21). These deposits are the result of several glacial periods, but predominately the Wisconsinan glaciation (fig. 8) and the subsequent recessions of the Saginaw, Erie, and Lake Michigan Lobes (Schneider, 1966, p. 53; Bleuer, 1989, p. 44). Most of the sediments deposited in the St. Joseph River basin are from the ice advances of the Saginaw and the Erie Lobes of about 15,000 years ago (Wayne, 1966, p. 35).

Although, the thickness of the drift in this basin ranges from 100 ft to 500 ft, thicknesses of 200 ft to 400 ft are typical (Reussow and Rohne, 1975) (fig. 21). Drift thickness generally increases from west to east. The sand and gravel units within the drift were deposited as broad outwash plains or channels beyond the melting ice front, as kame and esker deposits within the ice sheet, and as lenticular masses within the morainal deposits of glacial till (Pettijohn, 1968, p. 7). These sand and gravel units within the drift are the major aquifers of the basin.

Aquifer Types

Seven hydrogeologic sections 2A–2A′ to 2G–2G′ (fig. 22), were produced for this atlas to show the general hydrostratigraphy of the St. Joseph River basin. Locations of the sections are shown in figure 18. All hydrogeologic sections are oriented from south to north and were drawn at intervals of 8 to 12 mi. The seven hydrogeologic sections of the St. Joseph River basin have a combined length of 148 mi and were produced from the logs of 213 water-supply and test wells. The average density of logged wells plotted along the section lines is 1.4 wells per mile.

The St. Joseph River basin is an area of highly variable and complex glacial deposits. The sand and gravel sequences of the unconsolidated deposits, which form the major aquifer systems in the basin, were laid down by the advances of as many as three ice sheets. These deposits are variable in extent and thickness and are widespread over the entire area. The major types of unconsolidated aquifers are the surficial sands and gravels and the buried sands and gravels of the northern and central parts of the basin. Other unconsolidated aquifers are the discontinuous lenses of sand and gravel distributed across the central and southern parts of the basin. Buried preglacial bedrock valleys, where they are filled with sand and gravel, are small but significant aquifers along the Indiana-Michigan State line. In the St. Joseph River

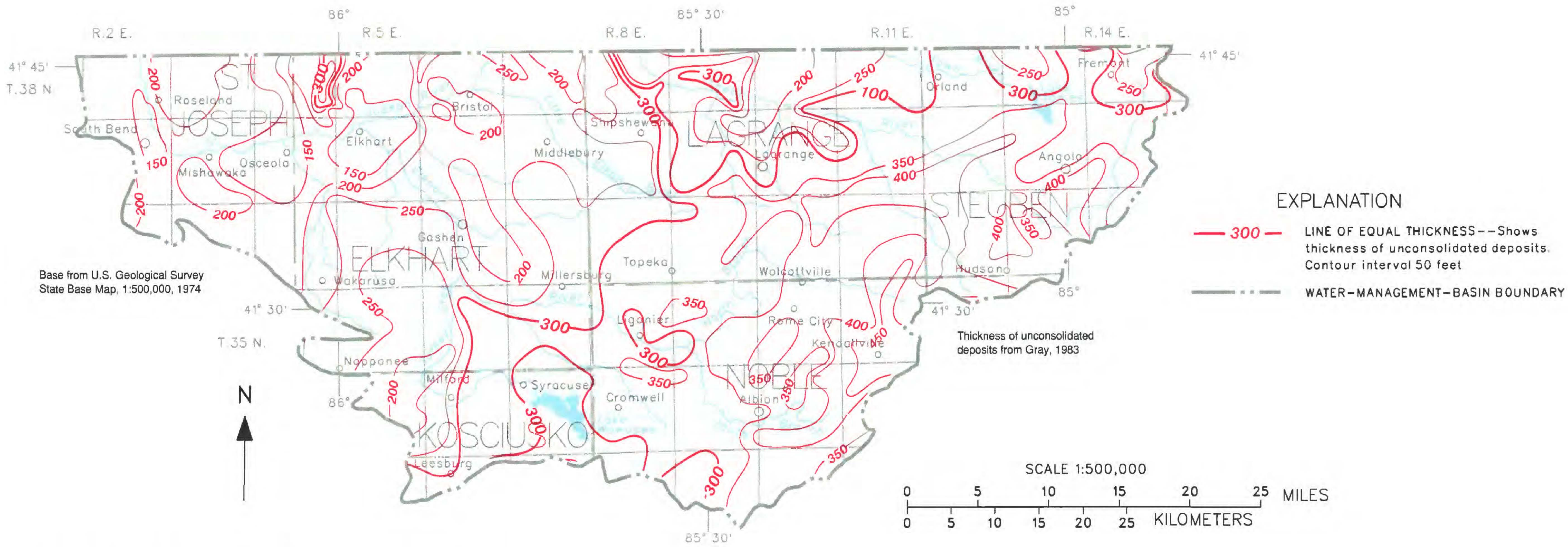


Figure 21. Thickness of unconsolidated deposits in the St. Joseph River basin.

basin, ground-water flow in the unconsolidated aquifers is generally toward the St. Joseph River and its tributaries. The four aquifer types mapped in the St. Joseph River basin (fig. 23) are summarized in table 4. The table includes range of thickness, range of yields, and aquifer names commonly used by other authors.

Unconsolidated Aquifers

Surficial Sand and Gravel Aquifers

Surficial sand and gravel aquifers are present in approximately half of the basin (fig. 23) and are shown in all seven hydrogeologic sections (fig. 22). The surficial sand and gravel aquifers consist mostly of outwash, outwash-fan deposits, isolated hills and ridges of ice-contact stratified drift, and Holocene alluvium (Gray, 1989). Thicknesses are generally greatest in the north and east and in the vicinity of buried bedrock valleys. Thicknesses of surficial sand and gravel aquifers range from a few feet in section 2C–2C′, T. 37 N. (fig. 22), to 160 ft in section 2B–2B′, T. 37 N. (fig. 22). Precipitation is the principal source of recharge to the surficial sand and gravel aquifers.

Buried Sand and Gravel Aquifers

Most of the buried sand and gravel aquifers are composed of outwash-related material. They are found in more than one-third of the basin, but they are predominantly in the northeastern and central parts of the basin, as seen in section 2C–2C′, Tps. 35 and 37 N., eastward to section 2G–2G′ (fig. 22). Small areas of buried sand and gravel are shown at the southern end of section 2A–2A′ (fig. 22). Much of the buried sand and gravel aquifer lies beneath the surficial sand and gravel aquifer. A clay layer of variable thickness separates these two aquifers.

Sand and Gravel Within Buried Preglacial Bedrock Valleys

Buried preglacial bedrock valleys are located in two areas of the basin. The far northern parts of sections 2B–2B′ and 2D–2D′, T. 38 N., (fig. 22) show the depths and thicknesses of parts of these valleys. This aquifer type does not cover a large area of the basin, but it is a locally significant aquifer. The bedrock valley is filled with sand and gravel in some areas, and with nonaquifer material in other areas (section 2B–2B′, fig. 22). The buried valleys filled by thick sequences of sand and gravel can produce as much as 2,000 gal/min (Pettijohn, 1968, p. 8). The buried bedrock valley beneath the Pigeon River in section 2D–2D′ (fig. 22) is filled mostly with nonaquifer material.

Discontinuous Sand and Gravel Aquifers

Discontinuous sand and gravel aquifers underlie approximately half of the basin. This aquifer type, which is present across the central one-third and

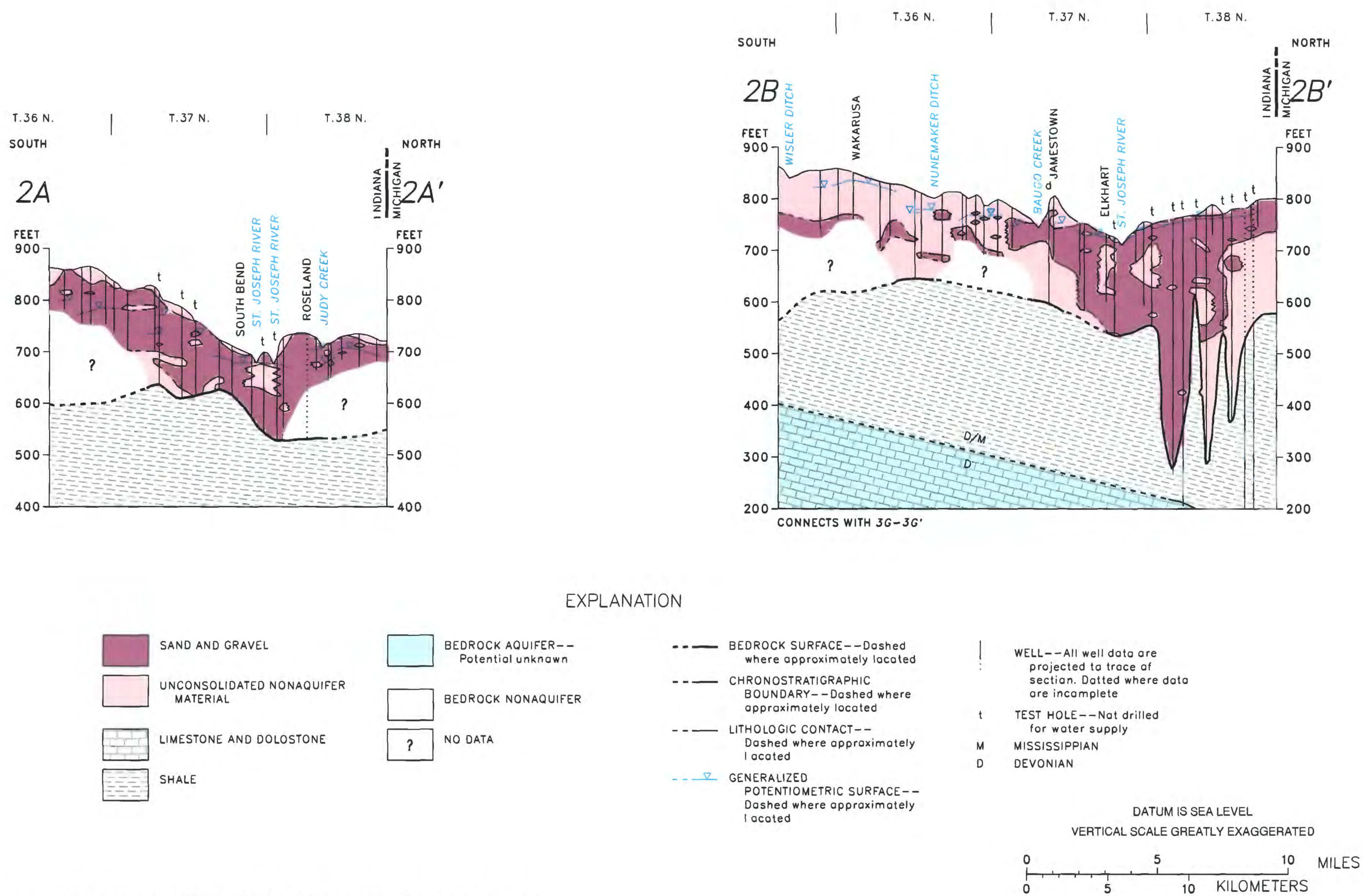
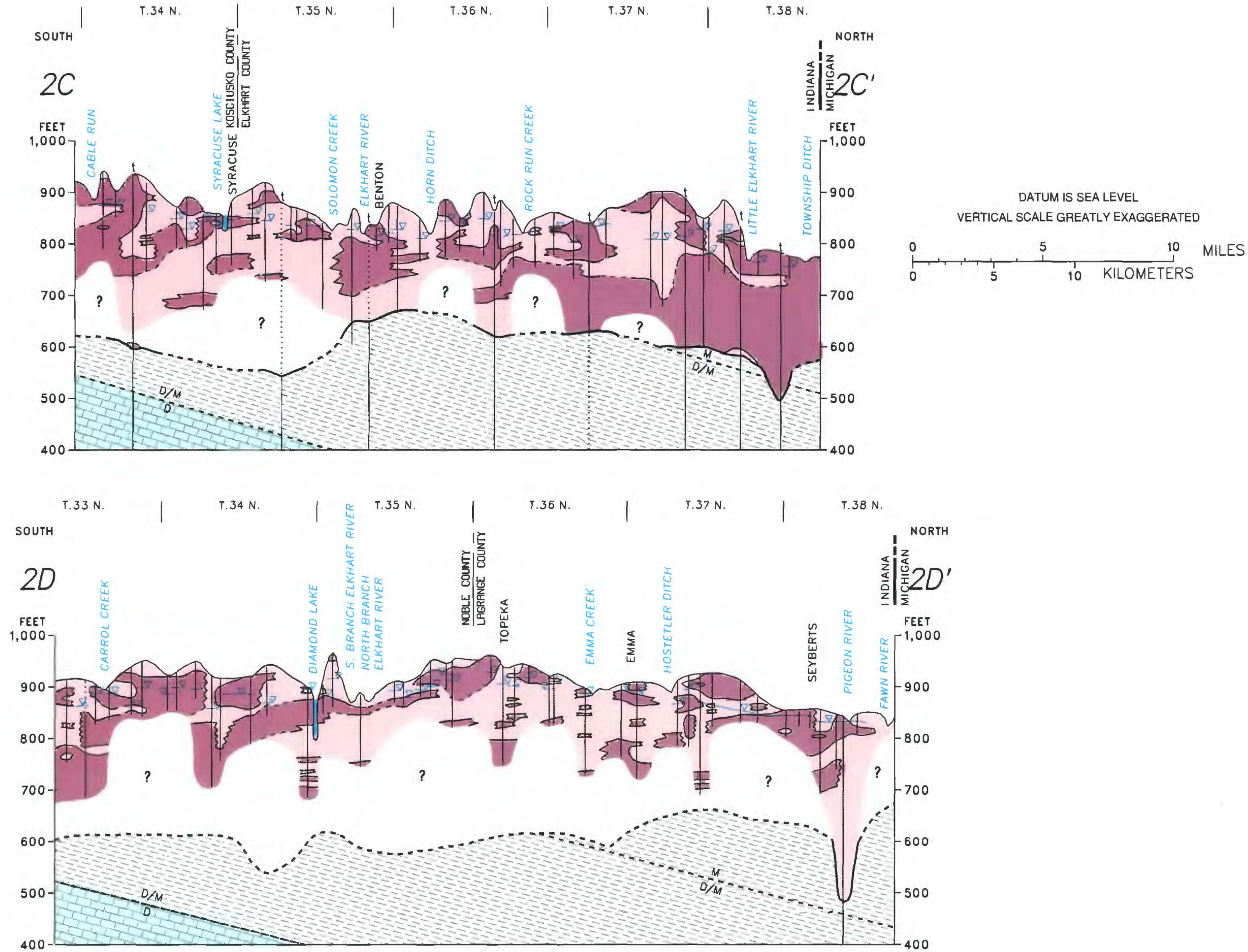
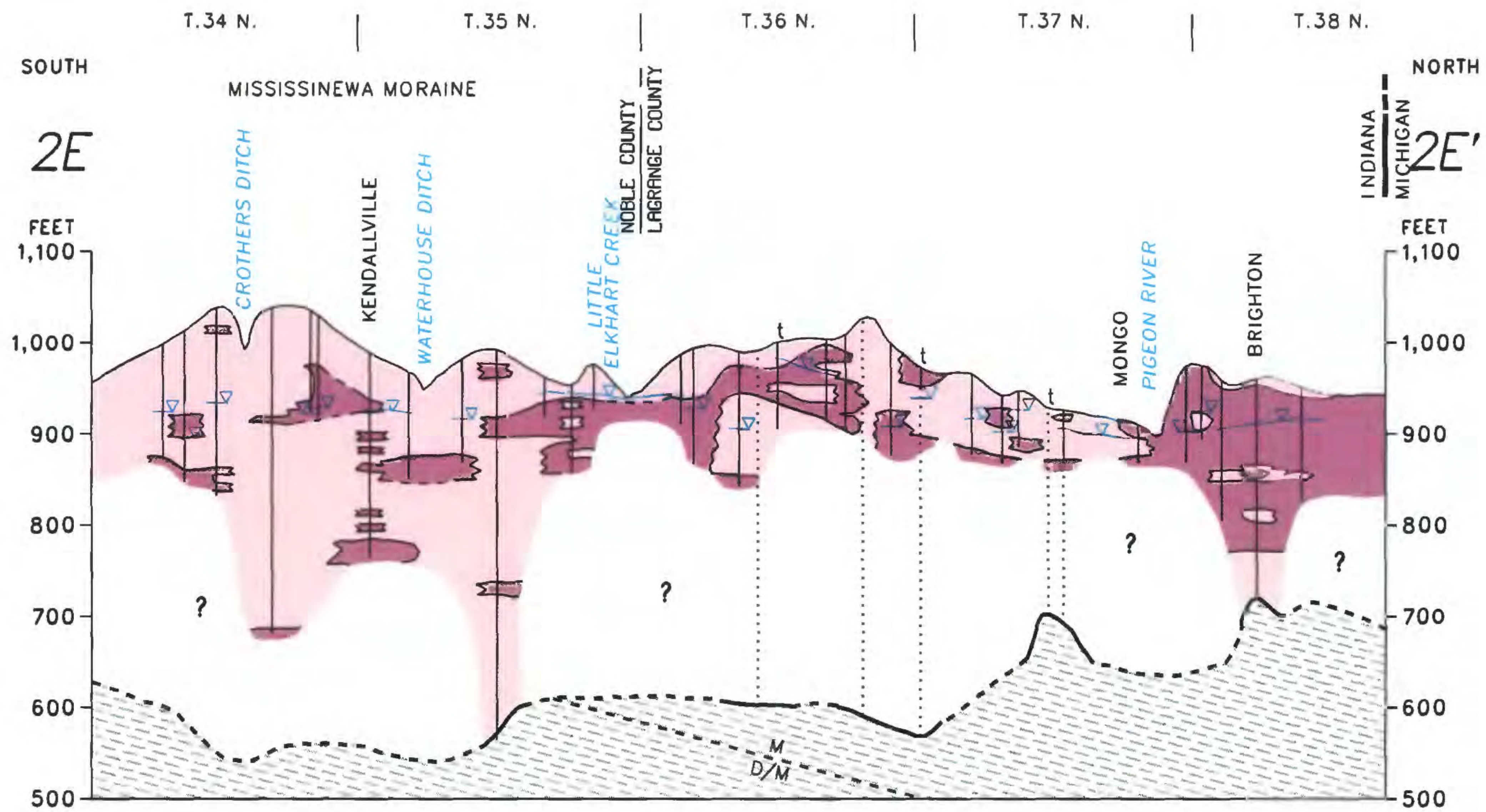


Figure 22. Hydrogeologic sections *2A–2A'* to *2G–2G'* of the St. Joseph River basin.

Figure 22. Hydrogeologic sections *2A–2A′* to *2G–2G′* of the St. Joseph River basin—Continued.




T.34 N.
T.35 N.
T.36 N.
T.37 N.
T.38 N.
SOUTH
NORTH
MISSISSINEWA MORAINE
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2E'
INDIANA
MICHIGAN
CROTHERS DITCH
KENDALLVILLE
WATERHOUSE DITCH
LITTLE ELKHART CREEK
NOBLE COUNTY
LAGRANGE COUNTY
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PIGEON RIVER
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FEET
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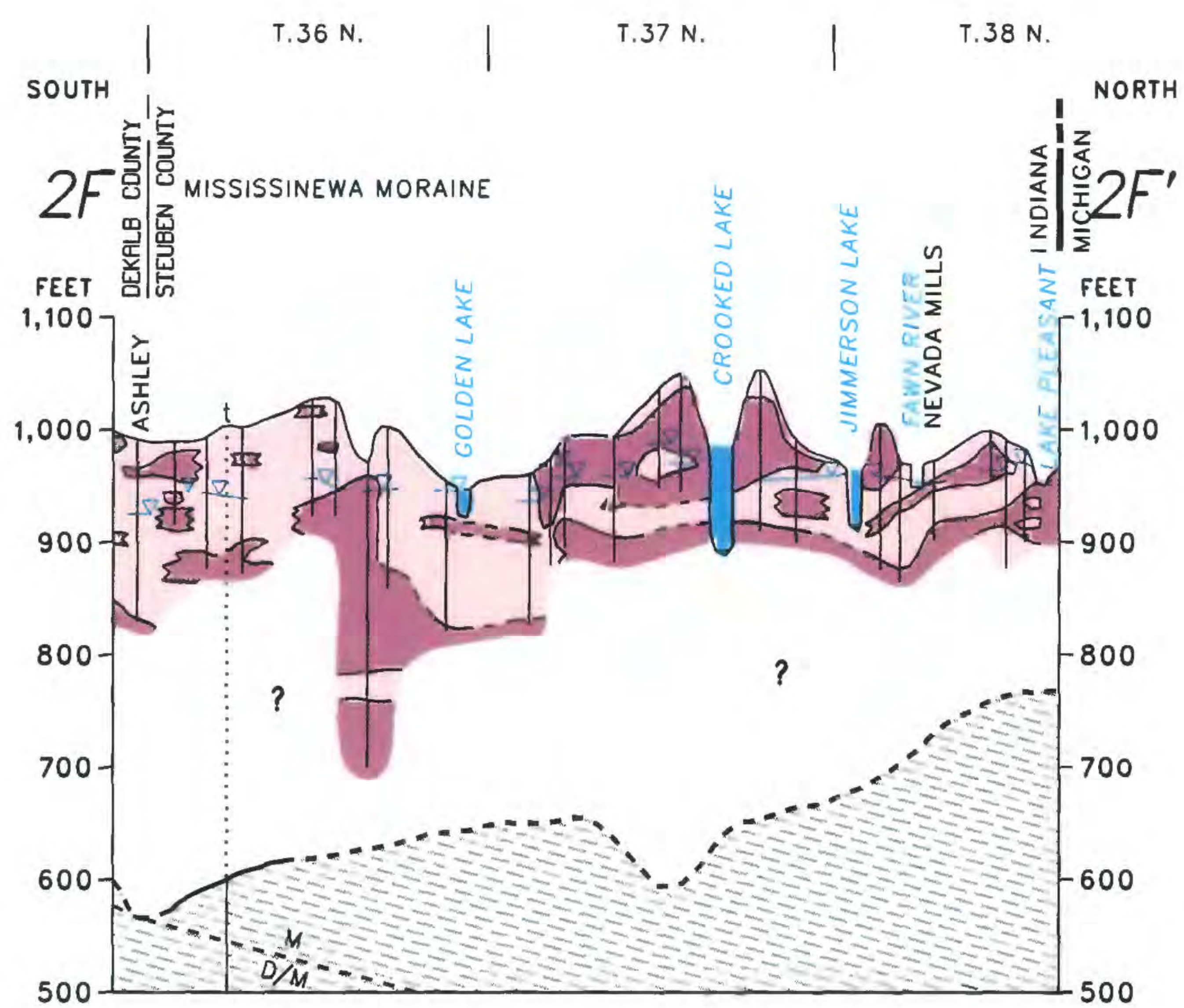
T.36 N.
T.37 N.
T.38 N.
SOUTH
NORTH
2F
2F'
DEKALB COUNTY
STEUBEN COUNTY
MISSISSINEWA MORAINE
INDIANA
MICHIGAN
ASHLEY
GOLDEN LAKE
CROOKED LAKE
JIMMERSON LAKE
FAWN RIVER
NEVADA MILLS
LAKE PLEASANT
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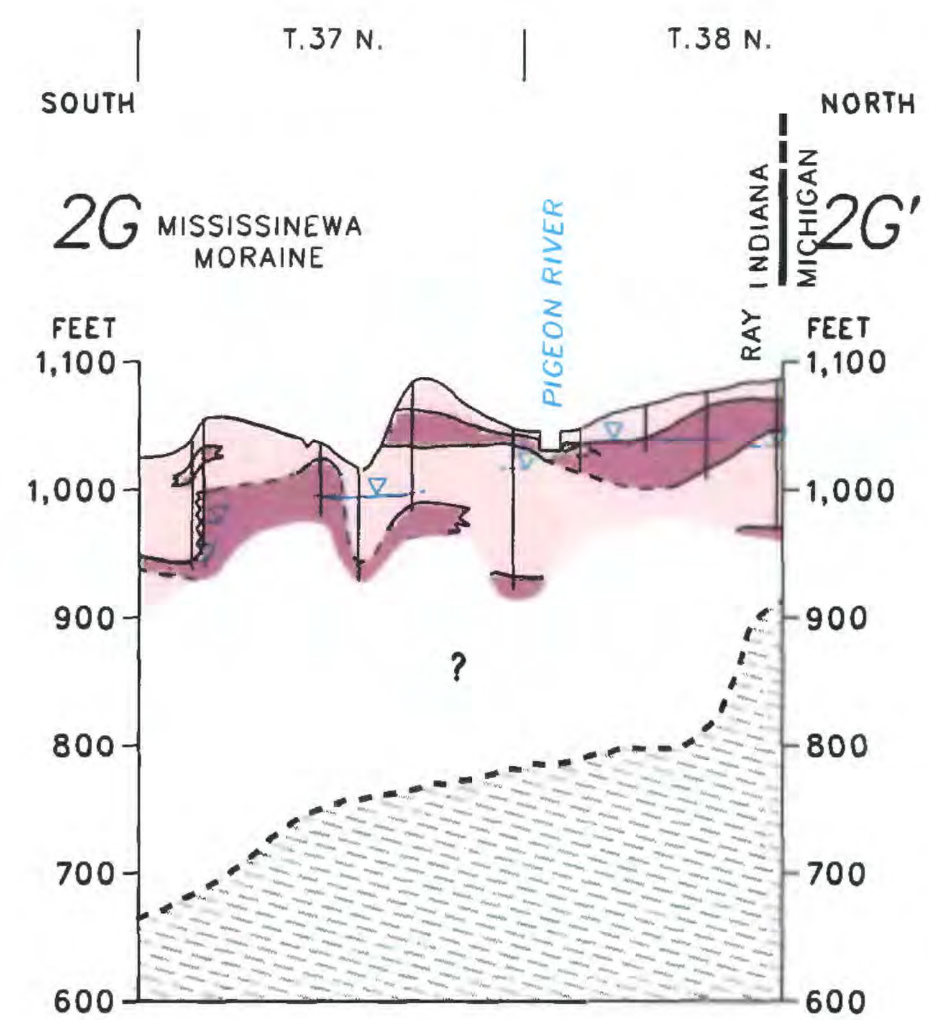
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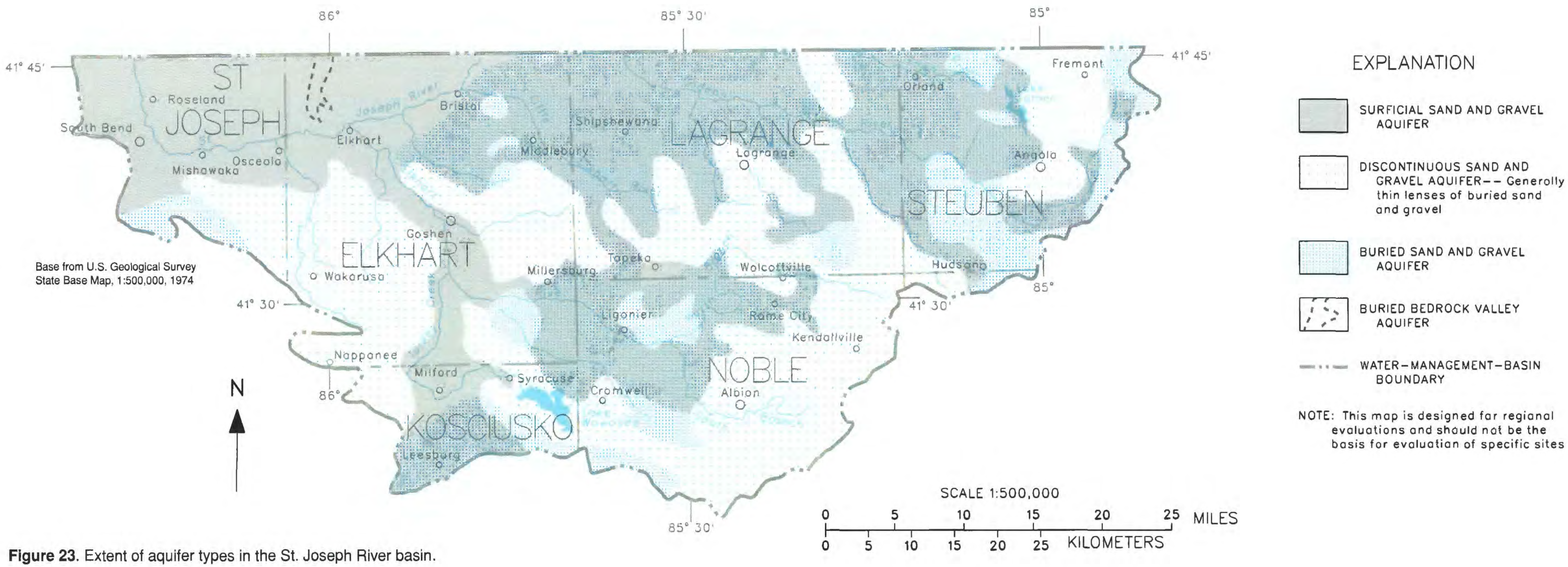


Figure 23. Extent of aquifer types in the St. Joseph River basin.

southern boundaries of the basin, is shown in hydrogeologic sections 2B–2B′ to 2G–2G′ (fig. 22). In general, this is an area of loam till and morainal topography (Gray, 1989). The discontinuous lenses of sand and gravel are variable in size but are typically thin and used locally for domestic and agricultural wells. In some areas, however, individual zones can be as thick as 60 ft, as shown in section 2E–2E′, T. 34 N. (fig. 22).

Bedrock Aquifers

There are several potential bedrock aquifers in the St. Joseph River basin. However, in over half of the area the potential bedrock aquifers are more than 300 ft below the land surface. Therefore, aquifers in the unconsolidated drift are more accessible, as well as adequate for all uses. Bedrock aquifers are not shown on the aquifer map (fig. 23) for the St. Joseph River basin.

The Silurian and Devonian carbonate rock sequence has the greatest potential as a bedrock source of ground water (Gray, 1973). At present, the carbonate bedrock is not used as a source of ground water because of the relative abundance of aquifers in the glacial drift. The other bedrock unit with aquifer potential, the Coldwater Shale of Mississippian age, is at the bedrock surface in the northeastern part of the basin (fig. 20). Isolated pockets of sandstone within this unit have a potential for domestic or light-industrial water supply (Gray, 1973).

Summary

The St. Joseph River basin encompasses 1,699 mi^2 in northeastern Indiana. The basin is composed of two physiographic units: the Kankakee Outwash and Lacustrine Plain and the Steuben Morainal Lake Area. The gently rolling bedrock surface of the basin is composed of shale and limestone. The entire basin is overlain by drift. This unconsolidated drift was deposited during the advances and retreats of Wisconsinan and older glaciations. The glacial drift generally thickens from west to east.

Four unconsolidated aquifer types are present in the St. Joseph River basin. The primary aquifer types are the surficial sands and gravels of the northern one-third and central parts of the basin and the buried sands and gravels of the northern and central

Table 4. Characteristics of aquifer types in the St. Joseph River basin
[<, less than; locations of aquifer types shown in fig. 23]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common names
Surficial sand and gravel	0-160	[1,2]25-2,000	St. Joseph and Tributary Valley, Howe Outwash, and Hilltop Aquifer Systems[2]; Valley train and outwash plain deposits[1]; Unit 3[1]
Buried sand and gravel	5- 90	25-1,000	Natural Lakes and Moraine, Howe Outwash, St. Joseph, Topeka, and Kendallville Aquifer Systems[2]
Sand and gravel in buried bedrock valley	20-480	[1,2]100-2,000	St. Joseph Aquifer System[2]; Preglacial valley deposits[3]
Discontinuous sand and gravel	<5- 80	[1,2]20- 600	Nappanee, Kendallville, and Natural Lakes and Moraine Aquifer Systems[2]; Unit 2[1]

[1]Hunn and Rosenshein, 1969.
[2]Indiana Department of Natural Resources, 1987.
[3]Pettijohn, 1968.

parts. Minor aquifer types include the buried preglacial bedrock valleys along the Indiana-Michigan State line and the discontinuous sand and gravel aquifer material across the central one-third and southern boundary of the basin.

Potential bedrock aquifers are not mapped in the St. Joseph River basin because of their great depth and infrequent use. The Silurian and Devonian carbonate rocks have the greatest potential as a bedrock ground-water source. Sandstone deposits within the Coldwater shale in the northeast part of the basin are also potential bedrock aquifers.

References Cited

Bailey, Z.C., Greeman, T.K., and Crompton, E.J., 1985, Hydrologic effects of ground- and surface-water withdrawals in the Howe area, Lagrange County, Indiana: U.S. Geological Survey Water-Resources Investigations Report 85-4163, 130 p.

Bassett, J.L., and Hasenmueller, N.R., 1979, Map showing structure on the base of New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources, Geological Survey, EGSP Series 800, scale 1:500,000.

Bleuer, N.K., and Melhorn, W.N., 1989, Glacial terrain models, north-central Indiana—the application of downhole logging to analysis of glacial vertical sequences: Field trip 2, Geological Society of America, North-Central Section, Notre Dame University, South Bend, Ind., April 19, 1989, p. 41-93.

Capps, S.R., 1910, The underground waters of north-central Indiana: U.S. Geological Survey Water-Supply Paper 254, 279 p.

Crompton, E.J.; Peters, J.G.; Miller, R.L.; Stewart, J.A.; Banaszak, K.J., and Shedlock, R.J., 1986, Review of the hydrologic data-collection network in the St. Joseph River basin, Indiana: U.S. Geological Survey Water-Resources Investigations Report 86-4157, 51 p.

Gray, H.H., 1973, Regional geologic map series: Indiana Department of Natural Resources, Geological Survey, Supplementary Chart 1.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Harrell, M.A., 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hunn, J.D., and Rosenshein, J.S., 1969, Geohydrology and ground-water potential of St. Joseph County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 33, 20 p.

Imbrigiotta, T.E., and Martin, Angel, Jr., 1981, Hydrologic and chemical evaluation of the ground-water resources of northwest Elkhart County, Indiana: U.S. Geological Survey Water-Resources Investigations Report 81-53, 149 p.

Indiana Department of Natural Resources, 1987, Water resource availability in the St. Joseph River basin, Indiana: Indiana Department of Natural Resources, Division of Water, Water Resource Assessment 87-1, 139 p.

Klaer, F.H., Jr., and Stallman, R.W., 1948, Ground-water resources of St. Joseph County, Indiana—part 1, South Bend area: Indiana Department of Conservation, Division of Water Resources Bulletin 3, 177 p.

Leverett, Frank, 1899, Wells of northern Indiana: U.S. Geological Survey Water-Supply Paper 21, 82 p.

Lindgren, H.A., Peters, J.G., Cohen, D.A., and Crompton, E.J., 1985, Hydrologic effects of ground- and surface-water withdrawals in the Milford area, Elkhart and Kosciusko Counties, Indiana: U.S. Geological Survey Water-Resources Investigations Report 85-4166, 75 p.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan W.N. and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 21, p. 112-124.

Marie, J.R., 1975, Hydraulic characteristics and water-supply potential of the aquifers in the vicinity of the wastewater treatment plant, South Bend, Indiana: U.S. Geological Survey Water-Resources Investigations 49-74, 26 p.

Peters, J.G., 1987, Description and comparison of selected models for hydraulic analysis of ground-water flow, St. Joseph River Basin, Indiana: U.S. Geological Survey Water-Resources Investigations Report 86-4199, 125 p.

Peters, J.G., and Renn, D.E., 1988, Effects of agricultural irrigation on water resources in the St. Joseph River basin, Indiana, and implications for aquifer yield: U.S. Geological Survey Water-Resources Investigations Report 87-4273, 35 p.

Pettijohn, R.A., 1968, Reconnaissance of the ground-water resources of the St. Joseph River basin in Indiana—appendix to the State water plan: Indiana Department of Natural Resources, Division of Water, 32 p.

Reussow, J.P., and Rohne, P.B., Jr., 1975, Water resources of the St. Joseph River basin in Indiana: U.S. Geological Survey Hydrologic Investigations Atlas HA-537, scales 1:500,000 and 1:250,000, 3 sheets.

Rosenshein, J.S., and Hunn, J.D., 1962, Ground-water resources of northwestern Indiana—preliminary report, St. Joseph County: Indiana Department of Conservation, Division of Water Resources Bulletin 15, 318 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Ind., Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E., 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Stallman, R.W., and Klaer, F.H., Jr., 1950, Ground-water resources of Noble County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 5, 103 p.

Wayne, W.J., 1966, Ice and land, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Ind., Indiana Academy of Science, p. 21-39.

KANKAKEE RIVER BASIN

By Joseph M. Fenelon

General Description

The Kankakee River basin, located in northwestern Indiana, is the sixth largest (2,989 mi^2) of the 12 water-management basins in the State. The basin includes most of Newton, Jasper and Starke Counties and one-half to two-thirds of Lake, Porter, LaPorte, St. Joseph, Marshall and Benton Counties (fig. 24). Most of the towns in the basin are farming communities; the largest cities are LaPorte, Plymouth, Knox, and Rensselaer.

Previous Studies

Preliminary reports of the ground-water resources have been published for all of the counties in the Kankakee River basin except Benton County (Rosenshein, 1961; 1962; Rosenshein and Hunn, 1962a; 1962b; 1964a; 1964b; 1964c; 1964d). The reports include preliminary evaluations of the ground-water resources, and tabulated well records for about 6,500 wells, including lithologic descriptions for about 2,500 wells and water-quality data from about 2,000 wells. Approximately 40 percent of the well records are from wells in the Kankakee River basin. The principal aquifers in Lake, Porter, LaPorte, and St. Joseph Counties were described by Rosenshein (1963), Rosenshein and Hunn (1968a; 1968b), Hunn and Reussow (1968), and Hunn and Rosenshein (1969). These authors described and mapped the geometry and potentiometric surfaces of the major aquifers, expected well yields, and general water quality. They also estimated hydraulic properties for the aquifers and associated confining units, and they determined sources and amounts of recharge to and discharge from the aquifers.

The geologic framework of the aquifers in the Valparaiso Moraine and the Kankakee River Lowland is discussed in Fraser and Bleuer (1991a; 1991b). The reports discuss the geology of the unconsolidated deposits and the principal aquifers throughout much of the northwestern part of the Kankakee River basin.

The western one-half of the basin was further studied by other authors. Hartke and others (1975) described the aquifers in Lake and Porter Counties. They summarized ground-water use and the potential for future use. They also qualitatively mapped the potential for aquifer contamination. The effects of ground-water withdrawals for irrigation on the ground-water system in Newton and Jasper Counties were described in several reports (Bergeron, 1981; Basch and Funkhouser, 1985; Arihood, in press).

A report on the water and land resources of the Kankakee River basin includes information on the general ground-water availability, ground-water flow, bedrock elevation, and the geometry and areal extent of the primary aquifers (State of Indiana and others, 1976). A comprehensive report on the water resources of the Kankakee River basin prepared by the Indiana Department of Natural Resources (1990) includes much of the same type of information as the 1976 report, but in more detail. In addition, information on ground-water use and ground-water quality is presented.

General descriptions of the aquifers in the Kankakee River basin have been reported by Harrell (1935) and Clark (1980) who described the major aquifers and the ground-water availability in the area.

Physiography

The Kankakee River basin lies primarily within the Northern Moraine and Lake Region, which includes the Valparaiso Morainal Area, the Kankakee Outwash and Lacustrine Plain, and the Steuben Morainal Lake Area; the southwestern part of the basin lies within the Tipton Till Plain (figs. 2 and 25).

The Valparaiso Morainal Area, in the northwestern part of the basin, is composed of an arc-shaped end moraine complex that parallels the southern shore of Lake Michigan from Illinois through northwestern Indiana into Michigan. The morainal complex marks a terminal position of the Lake Michigan ice lobe (Bretz, 1955, p. 106-108) and separates the Kankakee River basin from the Lake Michigan basin to the north. Elevations in the morainal complex generally range from 700 to 800 ft above sea level west of Valparaiso, Ind. (fig. 12) and 800 to 950 ft above sea level along the crest of the moraine east of Valparaiso. The western end of the complex is wide and gently undulating. It contains till ridges on the top of the complex and outwash sands on the southern flank that extend northward beneath the moraine (see Fraser and Bleuer, 1991b). East of Valparaiso, only a thin part of the Valparaiso Morainal Area near the crest of the morainal complex lies within the basin.

The Kankakee Outwash and Lacustrine Plain lies south and southeast of the Valparaiso Morainal Area and covers about two-thirds of the basin. It is a broad, flat, and poorly drained area that is primarily covered by glacial outwash, dune sand, alluvial deposits, and lake sand. The southwestern boundary encompasses the Iroquois Moraine of Wisconsinan age (fig. 25).

The Steuben Morainal Lake Area occupies southern St. Joseph County and most of Marshall County in the far eastern part of the basin. The part of the Steuben Morainal Lake area within the basin consists of gently undulating till plains created by the western advance of the Huron-Erie and Saginaw Lobes of the Wisconsinan ice sheet and the eastern advance of the Lake Michigan Lobe (Gray, 1989) (fig. 8).

The Tipton Till Plain extends throughout central Indiana and occupies a small part of the Kankakee River basin in Benton County and extreme southern Newton and Jasper Counties. It is bounded by the Kankakee Outwash Plain on the north side and is a nearly flat to gently undulating Wisconsinan till plain.

Surface-Water Hydrology

The Kankakee River drains 5,165 mi^2 in northeastern Illinois and northwestern Indiana (State of Indiana and others, 1976, p. III-1). Within Indiana, the Kankakee River basin has an area of 2,989 mi^2 (Hoggatt, 1975). The Kankakee River begins in northwestern St. Joseph County and flows southwest for about 80 mi before reaching Illinois (fig. 24). Before development of the area, the Kankakee River was a large, meandering river surrounded by marshes. Now the river in Indiana is ditched, has a gradient of about 1 ft/mi, and has been shortened to about one-third of its natural stream length (State of Indiana and others, 1976, p. III-24).

Most of the northern part of the basin is bounded by the Valparaiso Moraine (fig. 25), which forms a major divide separating drainage to the Mississippi River from drainage to the St. Lawrence River. The major northern tributaries of the Kankakee River, which flow from the Valparaiso Moraine, are the Little Kankakee River, Crooked Creek, and Singleton Ditch. Major tributaries to the south are Pine Creek, the Yellow River, and the Iroquois River (fig. 24). The Iroquois River drains about one-fourth of the basin (843 mi^2 in Indiana) and joins the Kankakee River in Illinois. The Yellow River has a drainage area of 439 mi^2, all within Indiana (State of Indiana and others, 1976, p. III-24). In all, there are more than 30 tributaries to the Kankakee River, many of them ditch systems.

Lakes occupy about 16 mi^2 of the basin, less than 1 percent of the surface area. Three lakes have areas of about 2 mi^2 and five lakes have areas of 0.5 to 1.5 mi^2 (State of Indiana and others, 1976, p. III-26).

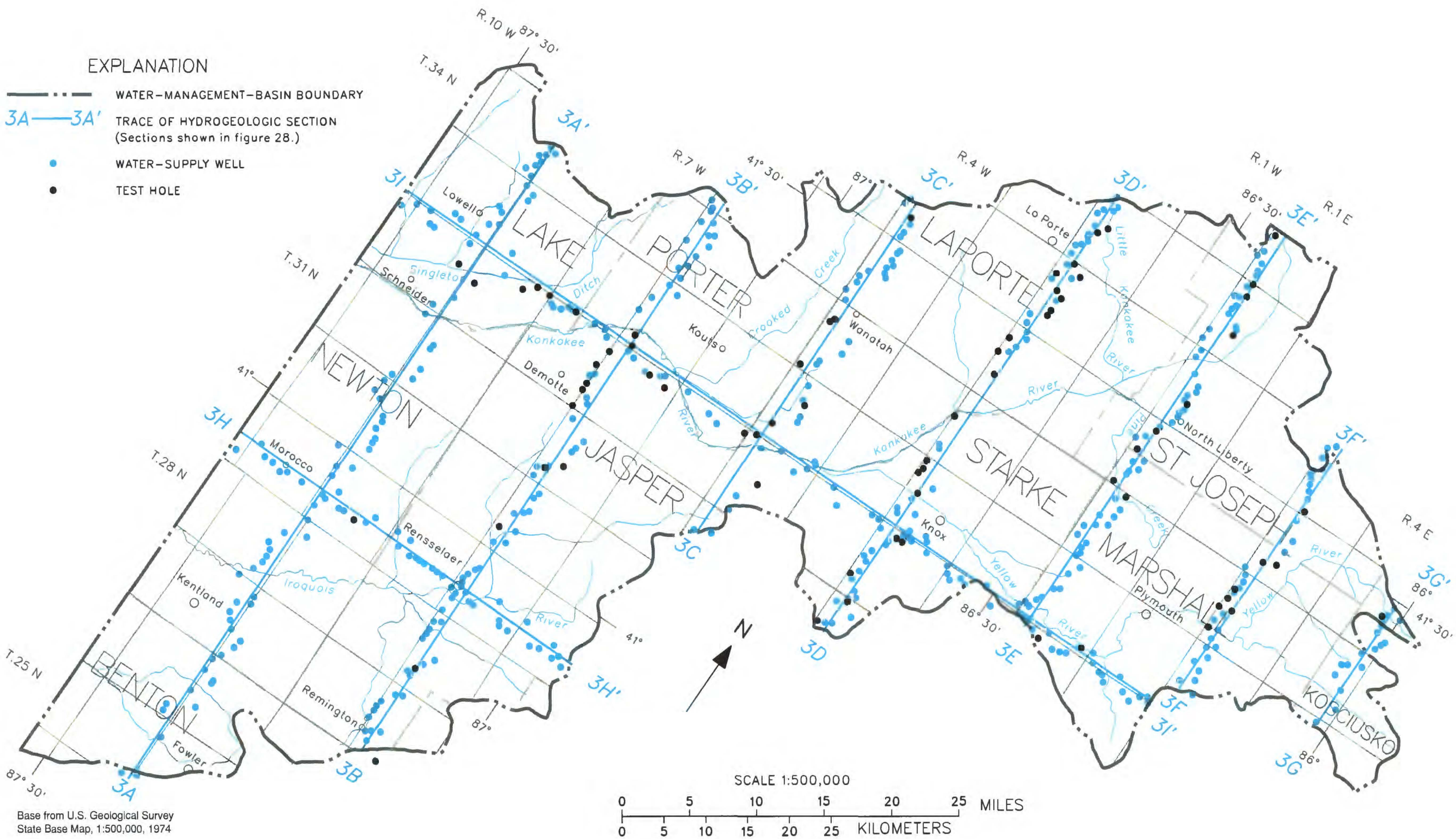


Figure 24. Location of section lines and wells plotted in the Kankakee River basin.

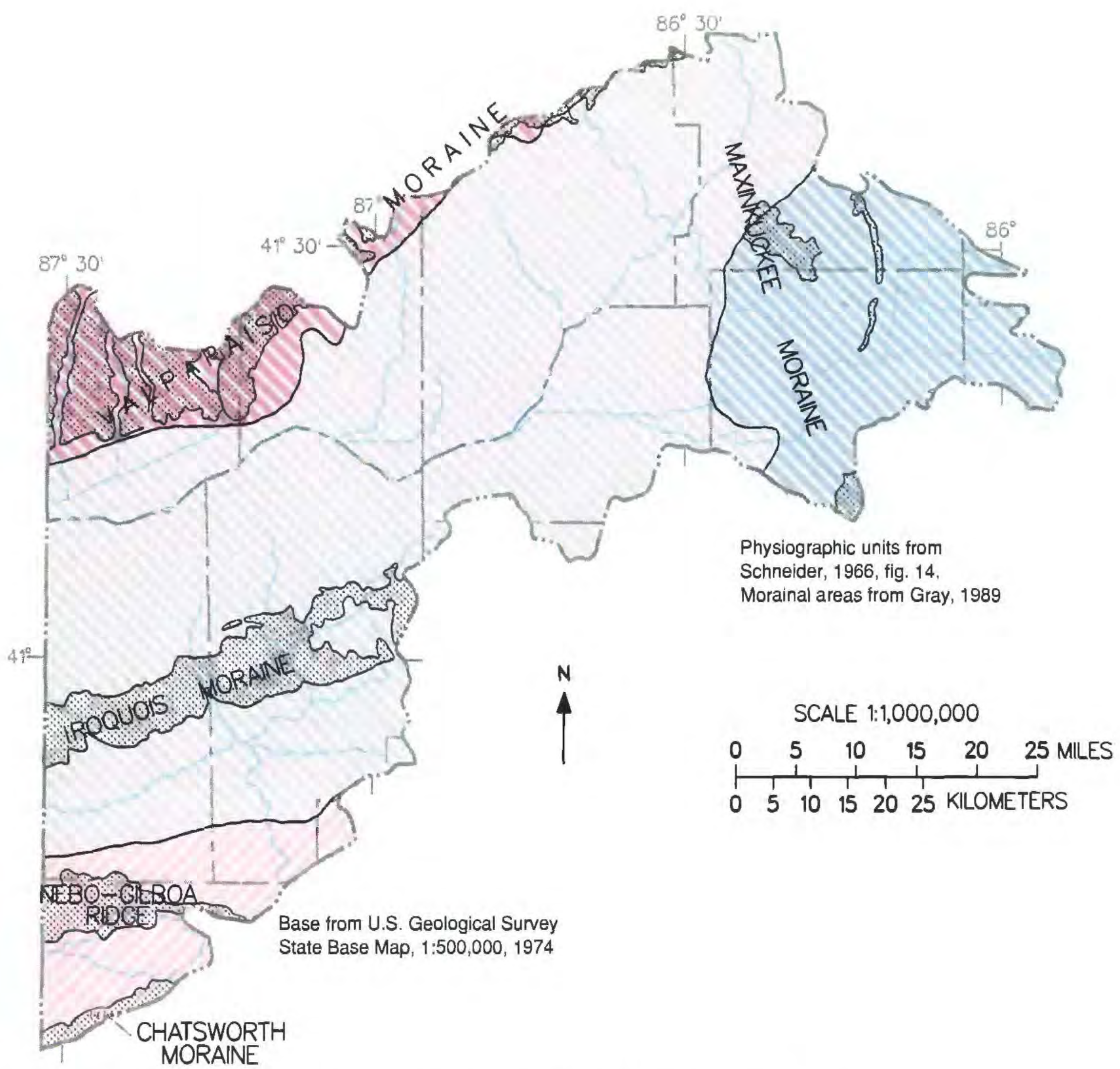


Figure 25. Physiographic units and moraines in the Kankakee River basin.

Geology

Bedrock Deposits

The major structural feature in the Kankakee River basin is the Kankakee Arch, a northern extension of the Cincinnati Arch, which trends northwest to southeast (fig. 4). The rocks on the north side of the Kankakee Arch dip northeast toward the Michigan Basin, whereas the rocks on the south side dip southwest toward the Illinois Basin. Within the Kankakee River basin, the rocks on both sides of the arch dip approximately 10 to 20 ft/mi. Rocks beneath large areas of Lake, Jasper, and Pulaski Counties, however, are on the crest of the arch and are nearly flat lying.

More than 4,000 ft of gently dipping sedimentary bedrock overlies Precambrian granitic bedrock (Rosenshein and Hunn, 1968a, p. 7; Hartke and others, 1975, p. 4). Approximately 3,500 ft of the sedimentary bedrock is of Cambrian and Ordovician age. The uppermost Ordovician rocks, collectively called the Maquoketa Group, consist of 200 to 300 ft of shale and minor limestone (Gray, 1972, p. 4-6) at depths of 600 to 1,000 ft below land surface.

Overlying the Maquoketa Group are Silurian rocks exposed at the bedrock surface in the northwestern part of the Kankakee River basin and Silurian, Devonian, and Mississippian rocks exposed at the bedrock surface in much of the rest of the basin (fig. 26). The Silurian rocks are composed of 400 to 600 ft of dolomite and some limestone (Hartke and others, 1975, p. 4) and consist of a wide range of carbonate rocks ranging from shaley to pure and fine to coarse-grained carbonate rocks; the lower 40 to 50 feet is very cherty (Shaver and others, 1986). The Silurian rocks compose the Sexton Creek Limestone, the Salamonie Dolomite, and the Salina Group. The Salina Group includes units consisting of reef facies and attains thicknesses of 400 ft in Lake and Newton Counties (Shaver and others, 1986, p. 134).

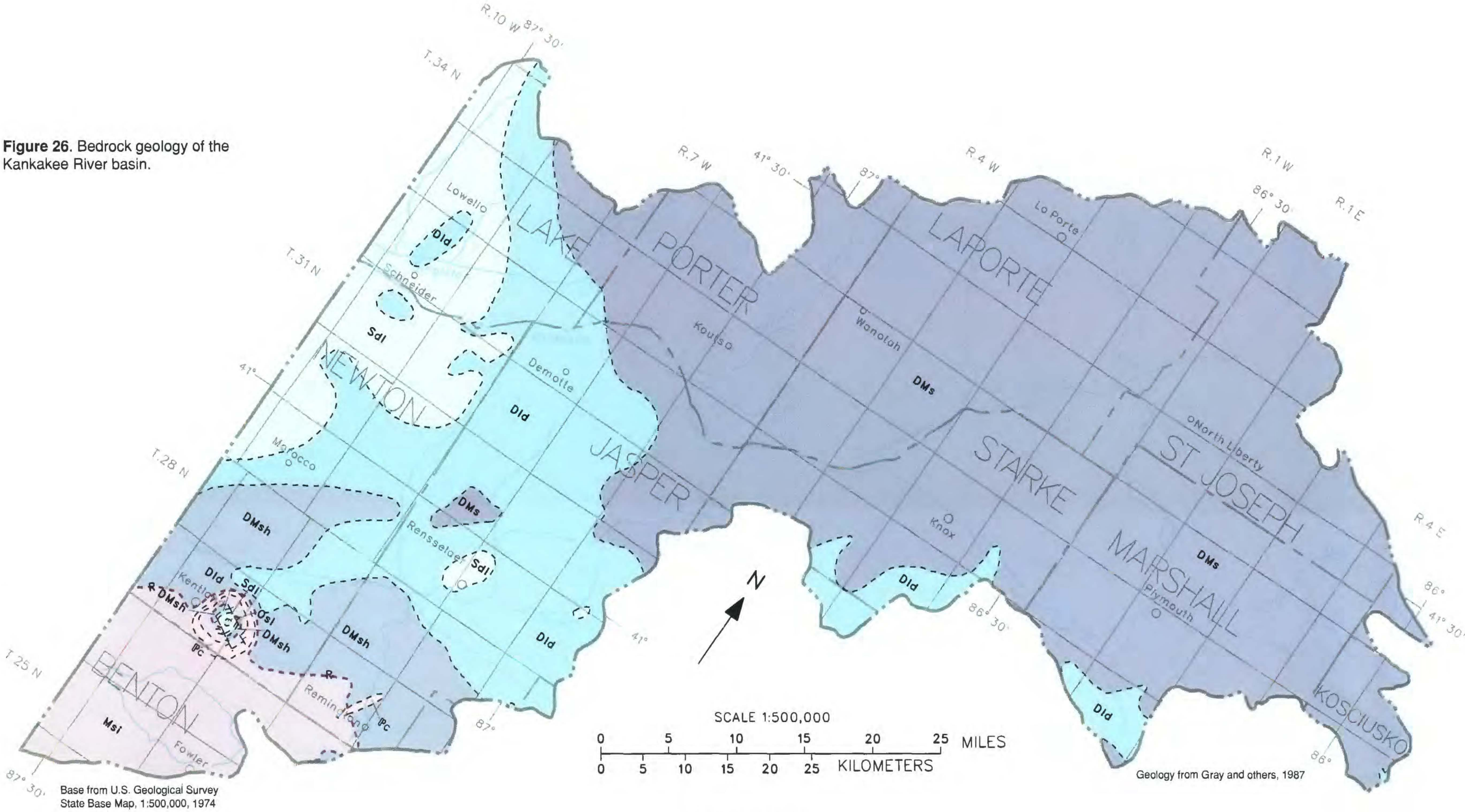
The Devonian and Mississippian rocks consist of several hundred feet of dolomite and limestone overlain by shale; these rocks compose the Muscatatuck Group and the New Albany Shale or the Antrim and Ellsworth Shales. The Muscatatuck Group overlies the Silurian rocks and attains thicknesses of as much as 200 ft (Shaver, 1974, p. 5). The Group is composed of a wide variety of impure to pure and fine to coarse-grained dolomite and limestone, and, in places, it contains anhydrite and gypsum in its lower part (Shaver and others, 1986, p. 99). Overlying the Devonian carbonate rocks in the northeastern part of the basin is the Antrim Shale, a brownish-black, noncalcareous shale (Shaver and others, 1986, p. 5). The Ellsworth Shale overlies the Antrim Shale and is of Devonian and Mississippian age. The Ellsworth Shale is a grayish-green shale that contains limestone or dolomite lenses in its upper part and alternating beds of grayish-green and brownish-black shale in its lower part (Shaver and others, 1986, p. 42). In the southwestern part of the basin, the New Albany Shale, which correlates with and is similar in lithology to the Antrim and Ellsworth Shales, overlies the Devonian carbonate rocks (Shaver and others, 1986, p. 101).

In the southwestern part of the basin, the Devonian and Mississippian shales are overlain by the Rockford Limestone and Borden Group of Mississippian age (Gray and others, 1987). The Rockford Limestone consists of 2 to 20 ft of gray clayey limestone and underlies 485 to 800 ft of gray clayey siltstone and shale known as the Borden Group (Shaver and others, 1986, p. 17-18).

A small (5 mi^2) complexly deformed structure known as the Kentland anomaly or dome is near Kentland, Ind. in southern Newton County (fig. 26). This structure consists of steeply dipping Ordovician and Silurian rocks that have been uplifted approximately 2,000 ft. Adjacent to the area are relatively flat-lying Mississippian and Pennsylvanian rocks. Meteorite impact, volcanism, and faulting have been proposed as explanations of the anomaly (Gutschick, 1976).

The bedrock surface is a preglacial erosional feature that has been further scoured by glacial erosion. Several preglacial bedrock-valley systems are buried beneath glacial deposits (Bleuer, 1989) (fig. 7). The Silurian and Devonian carbonate rocks exposed at the bedrock surface contain significant fractures and solution features in the upper 100 to 200 ft of the bedrock (Rosenshein and Hunn, 1968a, p. 10; Hartke and others, 1975, p. 4; Bergeron, 1981, p. 15; Basch and Funkhouser, 1985, p. 33).

Figure 26. Bedrock geology of the Kankakee River basin.



EXPLANATION

- **Pc** — PENNSYLVANIAN COMPLEXLY INTERBEDDED SHALE AND SANDSTONE, WITH THIN BEDS OF LIMESTONE AND COAL-- Composed of the Racoon Creek Group
- **Msi** — MISSISSIPPIAN SILTSTONE AND SHALE WITH MINOR SANDSTONE AND DISCONTINUOUS LIMESTONE-- Composed of the Borden Group
- **- -R- -** — ROCKFORD LIMESTONE
- **DMsh** — DEVONIAN AND MISSISSIPPIAN SHALE-- Composed of the New Albany Shale
- **DMs** — DEVONIAN AND MISSISSIPPIAN SHALE-- Composed of the Antrim and Ellsworth Shales
- **Dld** — DEVONIAN LIMESTONE AND DOLOMITE-- Composed of the Muscatatuck Group
- **Sdl** — SILURIAN DOLOMITE AND LIMESTONE-- Composed of the Wabash Formation. Louisville Limestone through Sexton Creek Limestone in Kentland area
- **Osl** — ORDOVICIAN SHALE AND LIMESTONE-- Composed of the Maquoketa Group to the upper part of the Knox Supergroup
- NORMAL FAULT-- Hachures on downthrown side. Dashed where approximately located
- GEOLOGIC CONTACT-- Dashed where approximately located
- WATER-MANAGEMENT-BASIN BOUNDARY

Unconsolidated Deposits

The unconsolidated deposits in the Kankakee River basin are largely the result of glacial, glaciofluvial, lake, wetland, and wind-blown sedimentation. They consist of clay, silt, and sandy loam tills; sand and gravel outwash; sand dunes; sand, silt and clay lake deposits; and peat. Thicknesses of unconsolidated deposits range from less than 50 ft in the western one-half of the basin to about 350 ft at the northern basin divide near Laporte, Ind. (fig. 27).

Glaciers advanced through the Kankakee River basin a number of times before Wisconsinan time. Most of what is preserved, however, is a result of the last several advances. During the Wisconsinan Age, ice actively advanced through the basin and eroded or overrode most of the older glacial deposits (fig. 8).

The Lake Michigan ice lobe overrode most of the Kankakee River basin as it advanced south of the basin. Where the eastern edge of the ice stalled at a till upland on the east side of the basin in central Marshall and St. Joseph Counties, it formed the Maxinkuckee Moraine (fig. 25), which has been interpreted as an outwash fan head (Bleuer and Melhorn, 1989, p. 44). Two other ice lobes were also active in northwestern Indiana at this time. The Saginaw Lobe advanced southwest into eastern Marshall and St. Joseph Counties. It sat in a lowland east of the Maxinkuckee Moraine and deposited till in the eastern part of the Kankakee River basin (Gray, 1989). The third lobe, the Huron-Erie Lobe, advanced west into Indiana, south of the Saginaw Lobe and east of the Lake Michigan Lobe.

As the Lake Michigan Lobe receded, it formed the till-cored Iroquois Moraine (fig. 25) and a proglacial lake that covered a large area of the Kankakee River basin between the ice to the north and the Iroquois Moraine to the south (Fraser and Bleuer, 1991b, p. 5). Lake muds, in excess of 30 ft, were deposited in the lake, and basal and ablation tills were later deposited over the muds (Fraser and Bleuer, 1991, p.5; Gray, 1989).

After the Lake Michigan Lobe receded, the Huron-Erie Lobe advanced west across the southern half of the basin, overriding the Iroquois Moraine and depositing thin (10 to 30 ft) tills in Benton and southern Newton and Jasper Counties (Gray, 1989).

The Lake Michigan Lobe retreated to a bedrock high on the northern edge of the Kankakee River basin, where it formed the Valparaiso Moraine (fig. 25). The upper part of the moraine is dominantly till on the western side, but it contains progressively more sand to the east until it becomes primarily a giant outwash fan head in LaPorte County.

While the ice was forming the Valparaiso Moraine, large volumes of outwash were deposited in a 5- to 15-mile wide band from the southern edge of the moraine to south of the Kankakee River. South of the river, large areas of the outwash were reworked by wind to form extensive dune sand deposits. Also deposited in the lowland between the Valparaiso, Iroquois, and Maxinkuckee Moraines were small areas of lake sand and clay. The sandy sediments overlie till, lake muds, or bedrock throughout much of the basin. The sand extends north into Lake and Porter Counties beneath the till and ice-contact stratified deposits of the Valparaiso Moraine. In central Newton and Jasper Counties, the sand pinches out on the till of the Iroquois Moraine. In western Marshall County and southern St. Joseph County, it terminates in the till and ice-contact stratified deposits of the Maxinkuckee Moraine.

Aquifer Types

Nine hydrogeologic sections (3A–3A′ to 3I–3I′) were produced for this atlas to show the general hydrostratigraphy of the Kankakee River basin (fig. 28). Seven of the sections are oriented from south to north; two are oriented from west to east (fig. 24). A total of 490 well logs were used to produce the sections; 72 well logs are from test holes that are unrelated to water use. The average density of logged wells plotted along the section line is 1.4 wells per mile. Several maps of Indiana were used to produce the sections including bedrock topography (Gray, 1982), surficial geology (Gray, 1989), bedrock geology (Gray and others, 1987), and structure on the base of the New Albany Shale (Bassett and Hasenmueller, 1979).

Adequate supplies of ground-water for domestic use can be found throughout the Kankakee River basin. Unconsolidated sands and gravels and Silurian and Devonian carbonate bedrock are the most productive aquifers. The primary unconsolidated aquifers are surficial sands in the central part of the basin and buried sands and gravels in the northern and eastern parts of the basin. The carbonate bedrock aquifer underlies the entire basin, but is only important as a source of water in the western one-half of the basin. Other locally important unconsolidated aquifers are discontinuous surficial sands and gravels in the three morainal areas and discontinuous buried sands in the southwestern part of the basin. An upper weathered-bedrock aquifer composed of limestone, shale, and minor amounts of sandstone is sometimes used in the southwestern part of the basin. Table 5 summarizes characteristics of the six aquifer types mapped in figure 29.

Unconsolidated Aquifers

Surficial Sand and Gravel Aquifer

A surficial sand and gravel aquifer covers about one-half of the basin (fig. 29) and is shown in the central part of sections 3A–3A′, 3B–3B′, and 3I–3I′, most of sections 3C–3C′ and 3D–3D′, and the northern part of section 3E–3E′ (fig. 28). It extends along a 15- to 25-mile wide band that trends east-northeast along the Kankakee River from Illinois to Michigan. It is bounded by the Valparaiso Moraine on the north, the Iroquois Moraine on the south, and the Maxinkuckee Moraine on the east (fig. 25). The aquifer, which is underlain by clay that can be more than 100 ft thick in some places, locally overlies bedrock, as shown in the central part of section 3C–3C′ and small areas of sections 3B–3B′ and 3I–3I′ (fig. 28).

The northern two-thirds of the surficial sand and gravel aquifer is composed of outwash that lies adjacent to and south of the Valparaiso Moraine. The southern one-third of the aquifer is composed of a mantle of dune sand over outwash. The dune sand is mostly reworked outwash sand at higher elevations than the adjacent outwash. The surficial sand and gravel aquifer also includes small areas of lake sand and alluvial material along the central parts of the Kankakee and Yellow Rivers. The aquifer material is mostly sand to the southwest and sand and gravel to the northeast (State of Indiana and others, 1976, p. III-28).

The surficial aquifer is unconfined and is recharged primarily from direct precipitation. Some recharge also comes from ground-water flow from the bedrock (Hartke and others, 1975, p. 30) and from the buried sand and gravel beneath the Valparaiso Moraine. The hydraulic connection between the surficial and buried sand and gravel aquifers can be seen in T. 33 N. of section 3B–3B′ and T. 36 N. of section 3C–3C′ (fig. 28). Most of the ground water flows from topographically high areas and discharges to the rivers and ditches at lower elevations (Bergeron, 1981, p. 21). The Kankakee, Iroquois, and Yellow Rivers function as the major regional discharge areas for the basin (State of Indiana and others, 1976, plate 10).

Depths to the water table range from 0 to more than 50 ft below land surface, but are generally 10 to 20 ft. Water levels in the aquifer fluctuate about 5 ft/yr because of variations in natural recharge and discharge (Arihood, in press); levels are highest in the early spring and the lowest in the summer (Bergeron, 1981, p. 23; Arihood, in press). Pumping of the bedrock aquifer for irrigation causes little noticeable effect on water levels in the surficial aquifer (Basch and Funkhouser, 1985, p. 31-33).

The saturated thickness of the surficial aquifer typically ranges from 20 to 50 ft in the southwestern part of the basin and from 50 to 100 ft in the northeastern part (State of Indiana, 1976, plate 11). Locally, the saturated thickness exceeds 150 ft, as shown in T. 37 N. of section 3D–3D′ (fig. 28). Ground-water yields from the surficial aquifer can be as great as 2,000 gal/min but are commonly much less. Generally, a properly constructed well can be expected to produce between 200 and 600 gal/min (Hartke and others, 1975, p. 30; Clark, 1980).

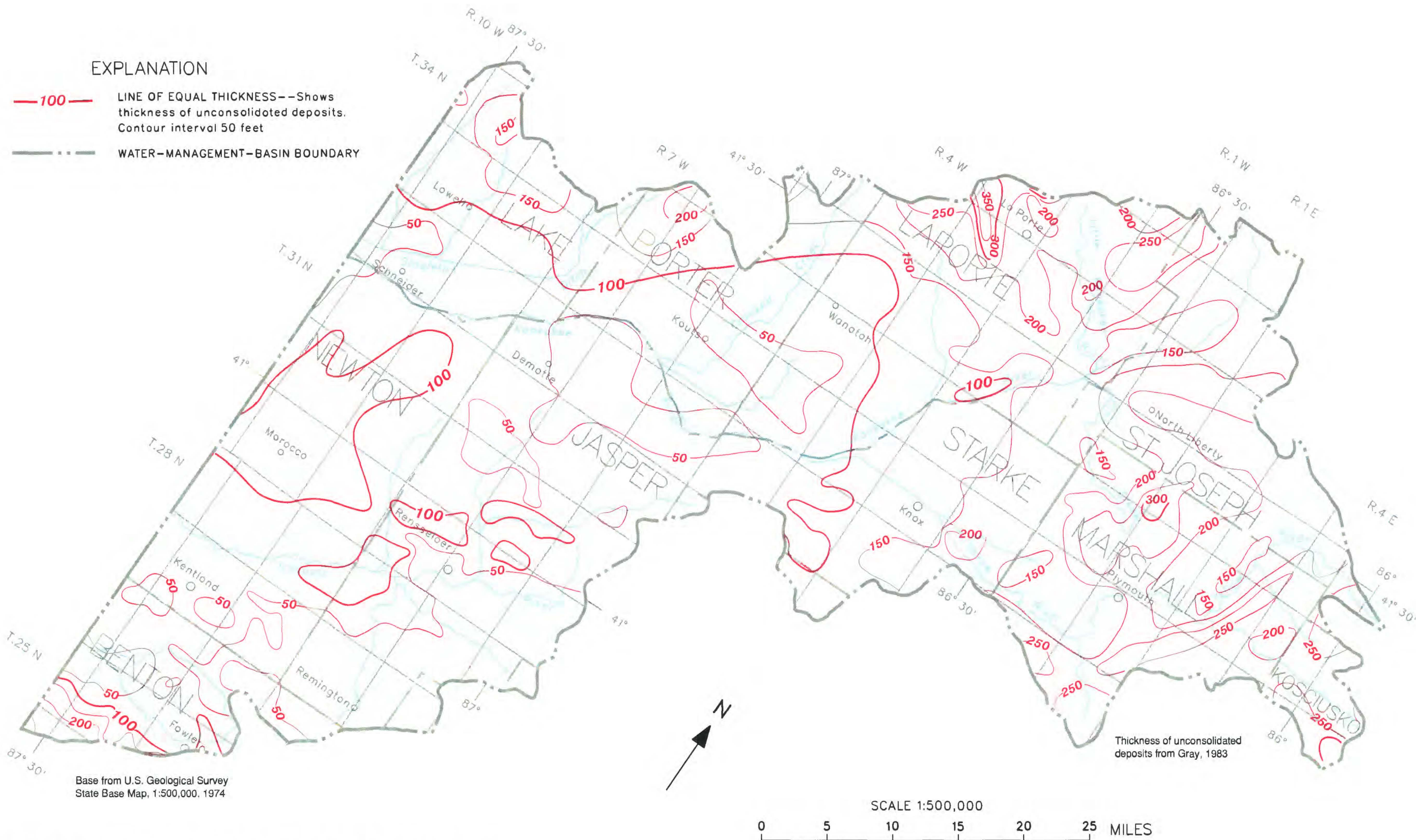


Figure 27. Thickness of unconsolidated deposits in the Kankakee River basin.

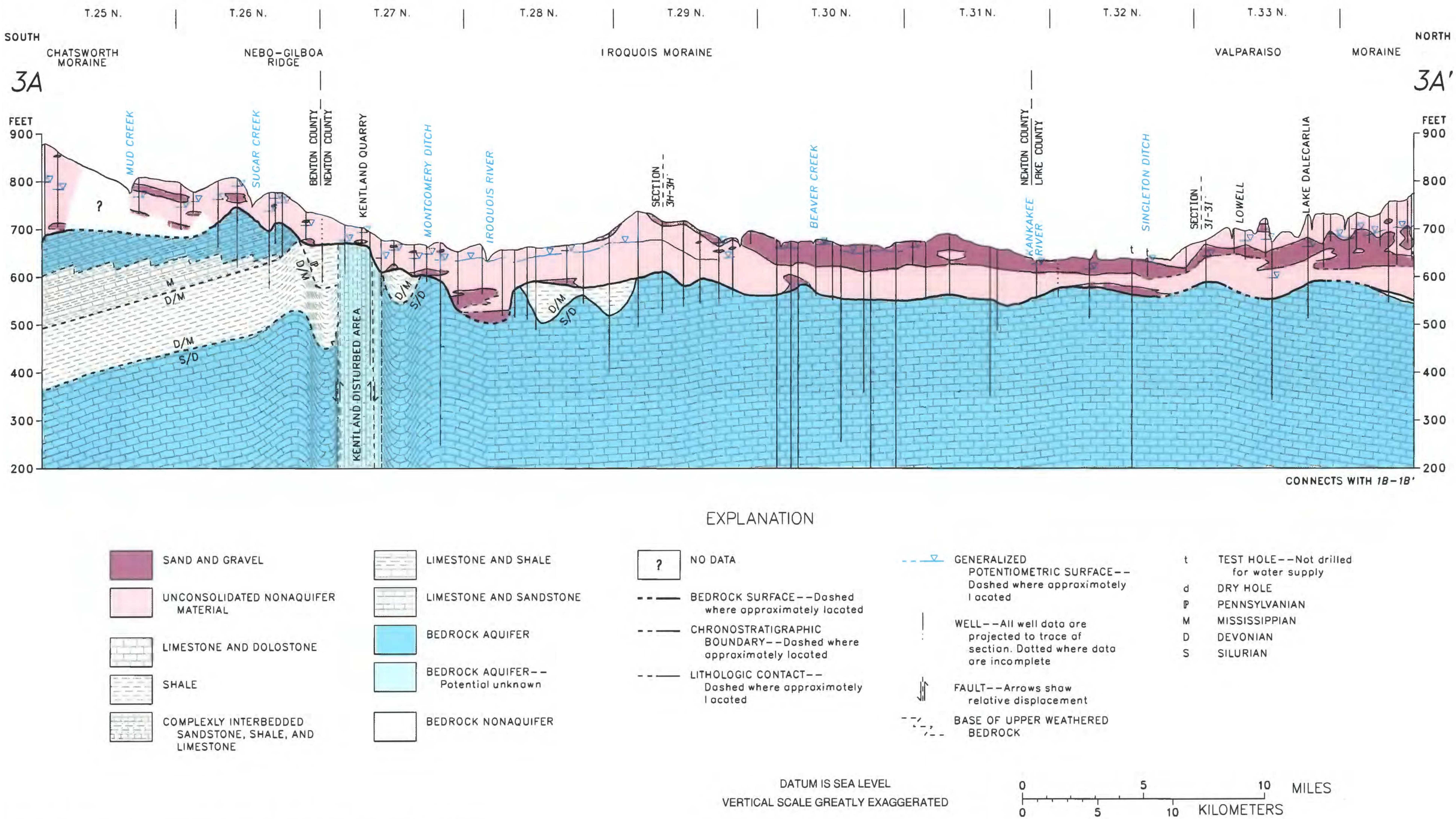


Figure 28. Hydrogeologic sections *3A–3A'* to *3I–3I'* of the Kankakee River basin.

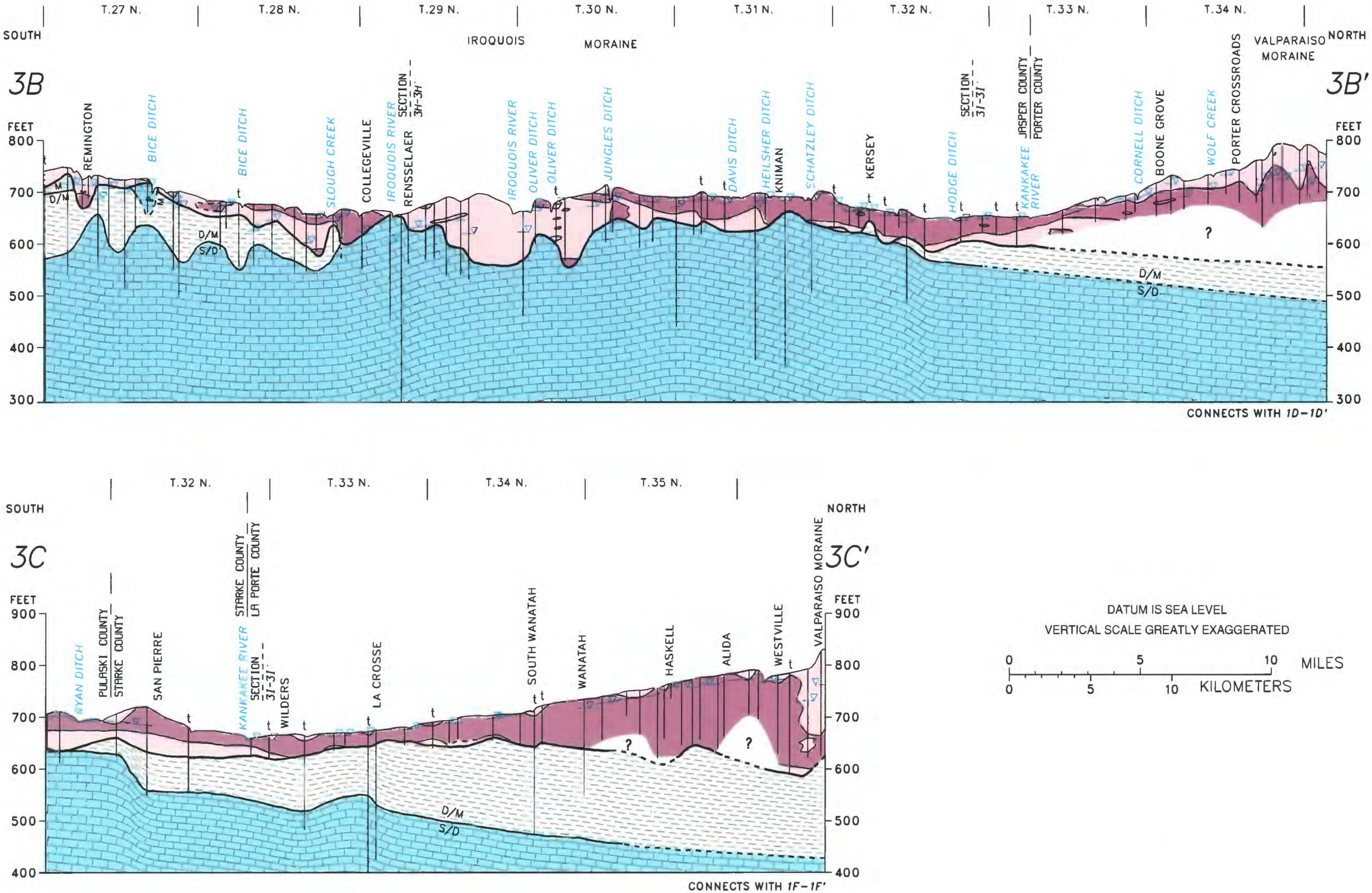
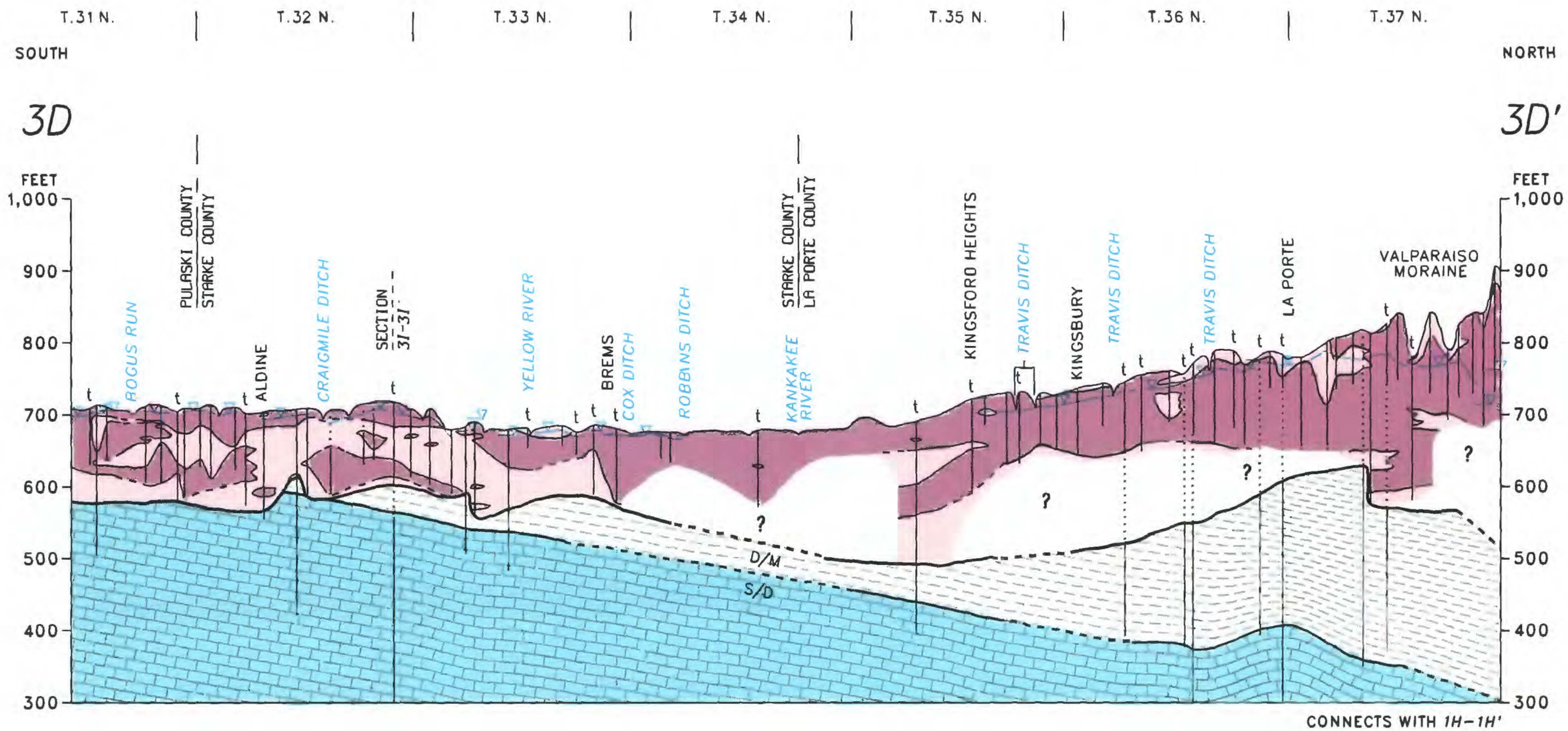


Figure 28. Hydrogeologic sections *3A–3A′* to *3I–3I′* of the Kankakee River basin—Continued.


T.31 N.
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SOUTH
NORTH
3D
3D'
FEET
1,000
900
800
700
600
500
400
300
ROGUS RUN
PULASKI COUNTY
STARKE COUNTY
ALDINE
CRAIGMILE DITCH
SECTION 31-31'
YELLOW RIVER
BREMS
COX DITCH
ROBBINS DITCH
KANKAKEE RIVER
STARKE COUNTY
LA PORTE COUNTY
KINGSFORD HEIGHTS
TRAVIS DITCH
KINGSBURY
TRAVIS DITCH
TRAVIS DITCH
LA PORTE
VALPARAISO MORAINE
D/M
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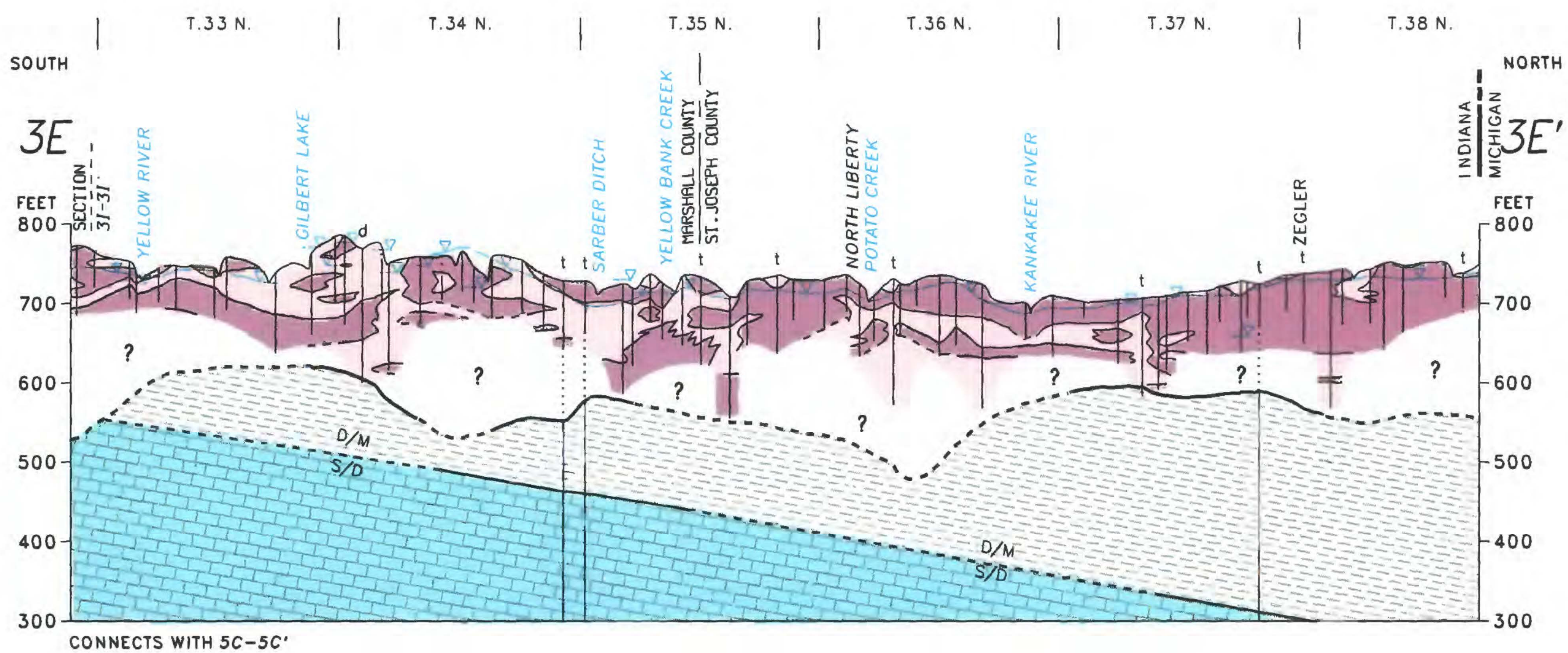
CONNECTS WITH 1H–1H'


T.33 N.
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T.38 N.
SOUTH
NORTH
3E
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600
500
400
300
SECTION 31-31'
YELLOW RIVER
GILBERT LAKE
SARBER DITCH
YELLOW BANK CREEK
MARSHALL COUNTY
ST. JOSEPH COUNTY
NORTH LIBERTY
POTATO CREEK
KANKAKEE RIVER
ZEGLER
INDIANA
MICHIGAN
D/M
S/D
CONNECTS WITH 5C–5C'

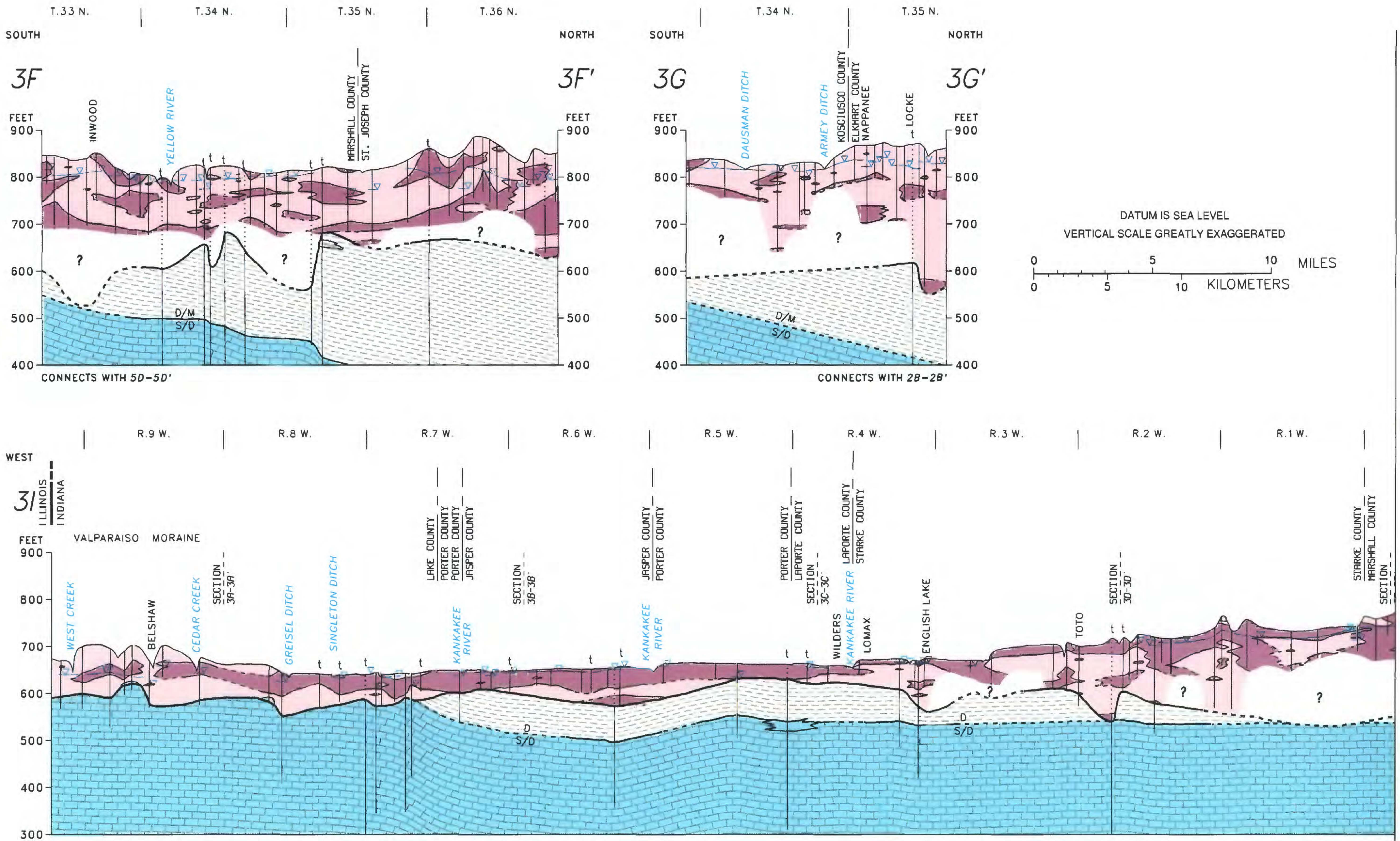
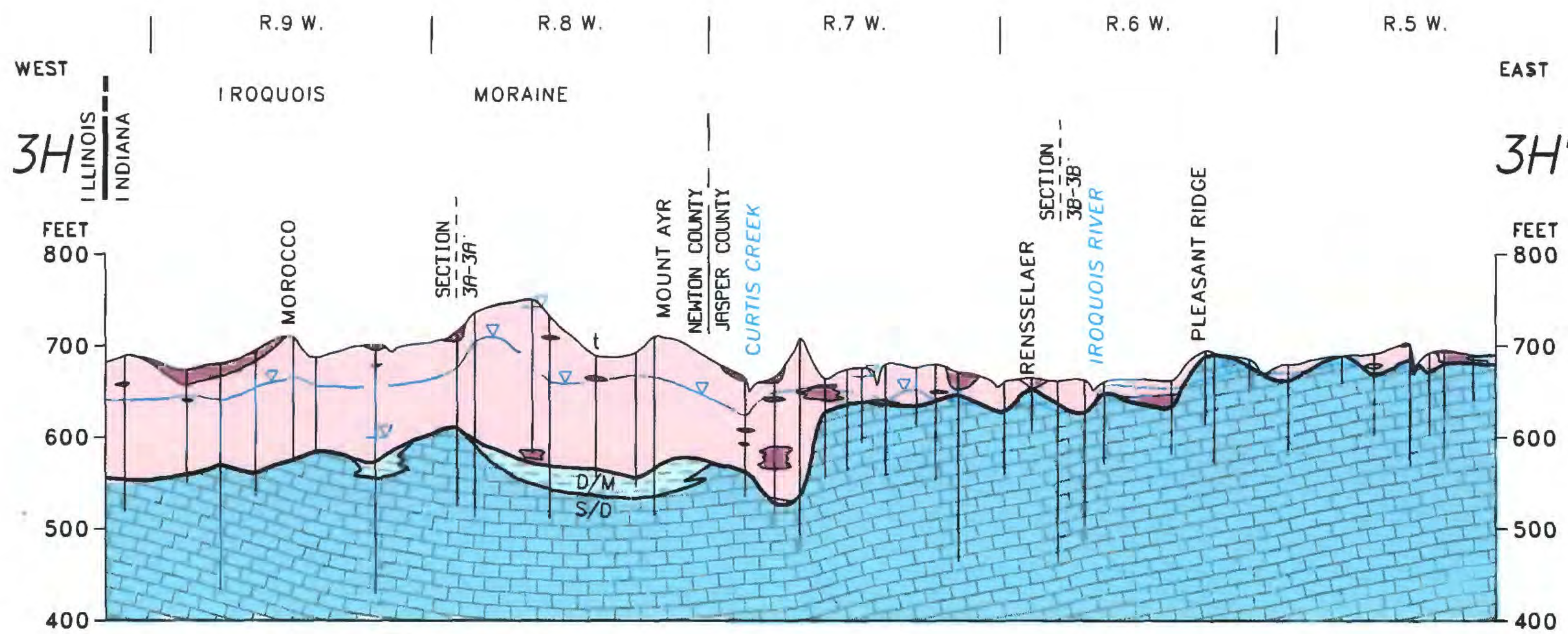
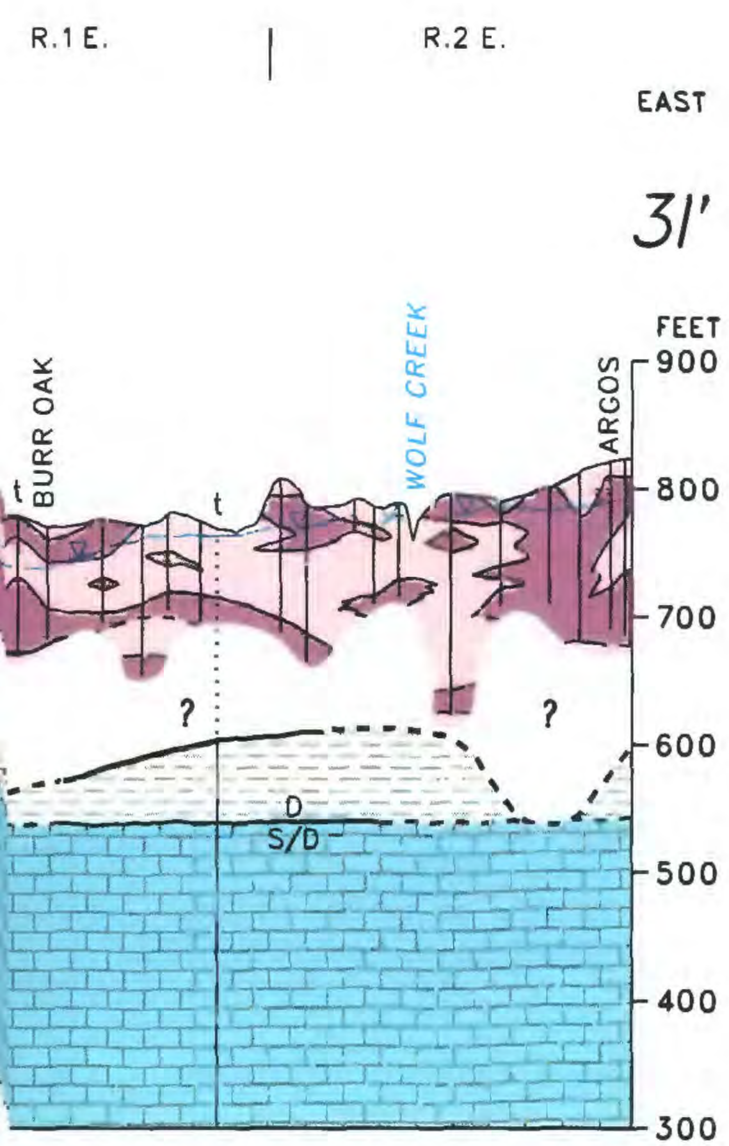


Figure 28. Hydrogeologic sections *3A–3A′* to *3I–3I′* of the Kankakee River basin—Continued.





Discontinuous Surficial Sand and Gravel Aquifers

Discontinuous surficial sand and gravel aquifers are found in the eastern part of the basin (fig. 29) and are shown in the southern part of section 3E–3E′, most of section 3F–3F′, and the eastern part of section 3I–3I′ (fig. 28). Most of the aquifers are located in the Maxinkuckee and Valparaiso morainal areas and are a part of a complex mixture of ice-contact stratified drift, outwash, and till. Individual aquifers range from a few feet to 100 ft in thickness. The discontinuous surficial sand and gravel is usually not used as a water supply, because it is commonly unsaturated or located above buried sand and gravel aquifers.

Buried Sand and Gravel Aquifers

Buried sand and gravel aquifers are present in more than one-quarter of the basin and are shown in three areas of figure 29. A major aquifer is in the northern part of the basin, a second smaller area of buried aquifers is in the southwestern part of the basin, and a third area of buried aquifers, discussed in the following section entitled "Discontinuous Buried Sand and Gravel Aquifers" is in the eastern part of the basin.

The northern buried sand and gravel aquifer is part of the Valparaiso Moraine and can best be seen in the northern parts of sections 3A–3A′ and 3B–3B′ (fig. 28). The sand and gravel form a single, partially confined aquifer (25 to 100 ft thick) that is buried beneath about 20 to 50 ft of clay till. The aquifer probably overlies clay throughout much of its extent in the Kankakee River basin, although in some areas, such as section 3B–3B′ (fig. 28), no information is available as to what underlies the sand and gravel.

The northern buried aquifer is recharged through the overlying till. The buried aquifer discharges to the surficial aquifer where they are hydraulically connected, as well as to the land surface through the overlying till or to the bedrock through the underlying till (Rosenshein and Hunn, 1968b). Water depths are generally 20 to 50 ft below land surface; the deeper water levels are to the north, near the crest of the moraine. Yields from the aquifer are generally 10 to 100 gal/min, but yields exceed 1,000 gal/min in places.

A second area of buried sand and gravel aquifer is the southwestern part of the basin beneath the Iroquois River in a buried bedrock valley. The aquifer, present beneath the river in section 3A–3A′ (fig. 28), has a thickness that ranges from several feet to about 70 ft and averages 25 ft. The aquifer directly overlies bedrock and is buried beneath 20 to 100 ft of clay. Recharge to the aquifer is primarily from the underlying bedrock, and discharge is through the overlying clay to the Iroquois River (Bergeron, 1981, p. 15-22).

Discontinuous Buried Sand and Gravel Aquifers

Most of the eastern part of the basin contains discontinuous buried sand and gravel aquifers. Buried aquifers that are continuous for 5 to 10 mi along a section line, such as those shown in the southern part of section 3E–3E′ and in section 3F–3F′ (fig. 28), were mapped as "buried sand and gravel aquifer" (fig. 29). Discontinuous buried sand and gravel aquifer was mapped to the east and west of this area where the aquifers are laterally continuous for only 1 to 5 mi. These discontinuous aquifers are a significant source of ground water. In general, the areas of "buried" and "discontinuous buried" aquifers in the eastern part of the basin differ little in their water-bearing capacity.

Throughout the eastern part of the basin, the "buried" and "discontinuous buried" aquifers are common; a domestic water supply generally can be found within 150 ft of the land surface (see southern part of sections 3D–3D′ and 3E–3E′, sections 3F–3F′ and 3G–3G′, and eastern part of section 3I–3I′; fig. 28). Multiple buried sand and gravel aquifers are common; individual aquifer thicknesses range from 5 to 50 ft. Well depths are generally 50 to 150 ft; a few wells in the far eastern part of the basin are deeper than 150 ft. Ground-water yields from these aquifers are generally 10 to 500 gal/min but are greater than 1,000 gal/min in some areas.

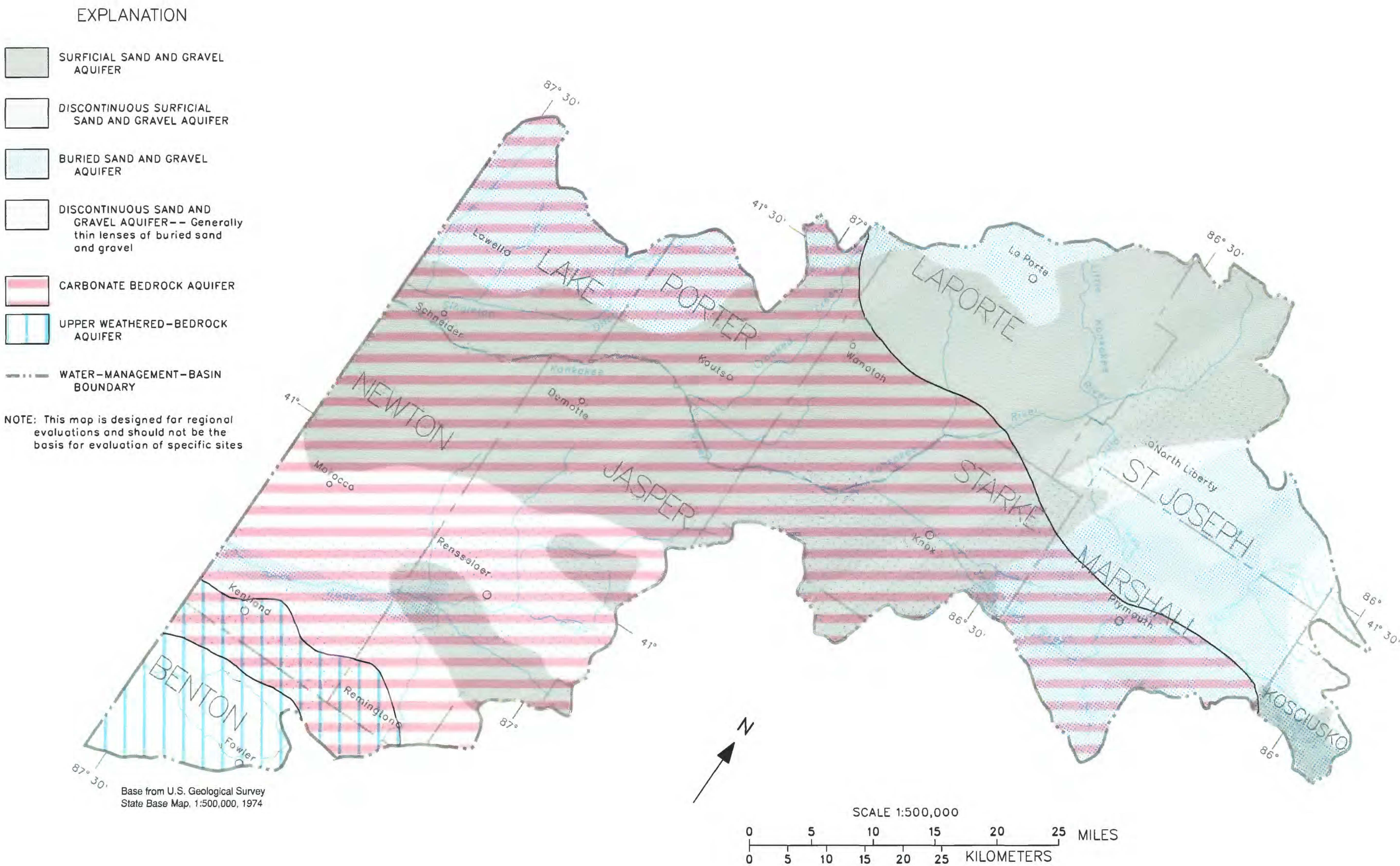


Figure 29. Extent of aquifer types in the Kankakee River basin.

Table 5. Characteristics of aquifer types in the Kankakee River basin
[>, greater than; <, less than; locations of aquifer types shown in fig. 29]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and gravel	0- 175	[1,2,3,4]10- 2,000	Kankakee aquifer[4,5]; Unit 3[1,2,3]; Kankakee and Valparaiso Outwash Apron Aquifer Systems[6]
Buried sand and gravel	5- 100	[1,3,4]10->1,000	Valparaiso aquifer[5]; Unit 3 [1,2,3]; Valparaiso Moraine, Maxinkuckee Moraine, and Nappanee Aquifer Systems[6]
Discontinuous surficial sand and gravel	10- 100	No data	Maxinkuckee Moraine Aquifer System[6]
Discontinuous buried sand and gravel	2- 50	5- 1,000	Unit 3[3]; Iroquois Basin and Eolian Sands Aquifer Systems[6]
Carbonate bedrock	500- 800	5- 2,000	Silurian-Devonian carbonate aquifer; Sexton Creek Limestone, Salamonie Dolomite, Salina Group, and Muscatatuck Group[7]
Upper weathered bedrock	50- 125	<1- 20	Mississippian Borden Group[7]

[1]Rosenshein and Hunn, 1968a.
[2]Rosenshein and Hunn, 1968b.
[3]Hunn and Rosenshein, 1969.
[4]State of Indiana and others, 1976.
[5]Hartke and others, 1975.
[6]Indiana Department of Natural Resources, 1990.
[7]Shaver and others, 1986.

An area of discontinuous buried aquifers in the southwestern part of the basin is within, and south of, the Iroquois Moraine (fig. 29). These aquifers can be seen in the southern parts of sections 3A–3A′ and 3B–3B′ and in the eastern part of section 3H–3H′ (fig. 28). The aquifers are smaller and less abundant than the buried discontinuous sand and gravel aquifers in the eastern part of the basin. Individual aquifers typically are thin (2 to 15 ft) and cover less than 1 mi^2. They are recharged from the surrounding clay material and are tapped for water supplies if no better aquifers are available (such as in section 3A–3A′, fig. 28). Yields from these intratill aquifers range from 5 to 50 gal/min but are commonly 10 to 20 gal/min. Depths to the aquifers can exceed 150 ft but are generally 50 to 100 ft.

Bedrock Aquifers

Carbonata Bedrock Aquifer

Silurian and Devonian carbonate bedrock, consisting mostly of dolomite and limestone, forms the principal bedrock aquifer in the Kankakee River basin. Although it is present throughout the basin, it is mostly used as a water supply in the western part of the basin (fig. 29; sections 3A–3A′, 3B–3B′, 3H–3H′, and 3I–3I′, and the southern parts of 3C–3C′ and 3D–3D′, fig. 28). The carbonate bedrock is found at the bedrock surface in most of Lake, Newton, and Jasper Counties. In the remaining parts of the basin, it is covered by as much as 300 ft of shale in addition to 100 to 300 ft of overlying unconsolidated deposits (fig. 27). Carbonate bedrock is shown at land surface in parts of section 3H–3H′ (fig. 28) but it is more than 500 ft below land surface in the northern parts of sections 3D–3D′ and 3F–3F′ (fig. 28).

The carbonate bedrock aquifer is generally 500 to 600 ft thick except in the northeastern part of the basin, where it attains a thickness of 800 ft. The upper part of the aquifer is highly permeable because of the enlargement of fractures, joints, and bedding planes by pre-Pleistocene weathering. The density of the fractures and joints decreases with depth (Bergeron, 1981, p. 15). The carbonate bedrock aquifer also includes reef structures within the Silurian rocks that can be highly fractured (Ault and others, 1976). These fractures are excellent conduits for ground water.

Most domestic wells penetrate only the upper 10 to 150 ft of the carbonate bedrock, as shown in section 3H–3H′ and the southern part of section 3B–3B′ (fig. 28). Many high-capacity irrigation, industrial, and municipal wells have been completed in the carbonate bedrock in Newton and Jasper Counties between the Kankakee and Iroquois Rivers. Most of the high-capacity wells are used to irrigate crops grown in the Kankakee outwash and surficial dune sands, such as those shown in the central part of section 3A–3A′ (fig. 28). The high-capacity wells generally penetrate more than 200 ft of carbonate bedrock; a few penetrate more than 500 ft.

In the northeastern and extreme southwestern parts of the basin, the carbonate bedrock aquifer is not used because of the availability of water in the unconsolidated aquifers, the greater depth to the carbonate bedrock aquifer, and potentially high concentration of dissolved solids in the water (Rosenshein and Hunn, 1968b, p. 5). The boundary of the carbonate bedrock aquifer is mapped where the top of the carbonate bedrock aquifer is generally greater than 300 ft below land surface (fig. 29), but this boundary does not necessarily separate the usable from the unusable parts of the aquifer. The aquifer is generally not used where more than 200 ft of unconsolidated material and(or) shale cover it, and it is rarely used where more than 300 ft of material cover it, because of the reasons given above.

Devonian and Mississippian shales and the lower till unit confine the upper surface of the carbonate bedrock aquifer throughout most of the basin. The shales are potential sources of small quantities of water (Rosenshein and Hunn, 1968a; 1968b; Arihood, in press). At depths of greater than 600 ft, the lower carbonate bedrock surface is underlain by 200 to 250 ft of shale interbedded with some limestone (Gray, 1972). Recharge to the carbonate bedrock aquifer is mostly from the till and upper shale, and discharge is to the major rivers in the basin (Arihood, in press). Well yields are generally 5 to 50 gal/min, although high-capacity wells can produce as much as 1,500 to 2,000 gal/min (Basch and Funkhouser, 1985, p. 34). During the irrigation season, water levels in the bedrock are lowered 5 to 80 ft by pumping (Arihood, in press), but most wells are able to recover fully before the beginning of the next season.

Upper Weathered-Bedrock Aquifer

Mississippian limestone, siltstone, and shale are found at the bedrock surface in the southwestern part of the basin in Benton County (fig. 29). As much as 200 ft of the Mississippian bedrock is shown in the southern part of section 3A–3A′ (fig. 28). Water wells in the section penetrate 10 to 125 ft of the upper part of the bedrock. The upper weathered bedrock, where permeability has probably been enhanced due to preglacial weathering, is shown as "aquifer." In general, the Mississippian bedrock is a poor water producer. Yields from wells into the Mississippian bedrock shown in section 3A–3A′ (fig. 28) range from less than 1 to 20 gal/min.

Summary

The Kankakee River basin includes 2,989 mi^2 in northwestern Indiana. The basin is bounded by morainal areas and till on the north, southwest, and east; a wide band of surficial outwash and dune sands crosses the central part of the basin.

Six aquifer types were delineated in the Kankakee River basin. Unconsolidated sands and gravels and Silurian and Devonian carbonate bedrock are the most productive aquifers in the Kankakee River basin. One of the primary unconsolidated aquifers is composed of surficial sands in the central part of the basin. The saturated thickness of the sands ranges from 20 to 100 ft, and well yields range from 10 to 2,000 gal/min. Two additional areas of productive unconsolidated aquifers are in the northern and eastern parts of the basin where buried continuous and discontinuous sands and gravels are present. The northern buried aquifer, located beneath the Valparaiso Moraine, forms a single, partially confined aquifer that is 25 to 100 ft thick. It is buried beneath 20 to 50 ft of clay till and overlies a basal till. The eastern buried continuous and discontinuous aquifers are the main source of ground water for the eastern part of the basin. Sufficient supplies of ground water for domestic use are commonly found within 150 ft of land surface. The buried aquifers can supply from 10 to more than 1,000 gal/min of ground water. Other locally important unconsolidated aquifers are discontinuous surficial sands and gravels in morainal areas and discontinuous buried sands in the southwestern part of the basin.

The carbonate bedrock aquifer underlies the entire basin but is only important as a source of water in the western one-half of the basin, where it is near or at the bedrock surface. Most domestic wells penetrate only the upper 10 to 150 ft of the carbonate rocks, whereas high-capacity wells penetrate from 200 to more than 500 ft of the carbonate bedrock aquifer. Yields are usually sufficient for domestic use, and they can be as great as 2,000 gal/min. Another bedrock aquifer composed of an upper weathered zone of various lithologies is used for a water supply in some areas of the southwestern part of the basin.

References Cited

Arihood, L.D., in press, Geohydrology and simulated ground-water flow in an irrigated area of northwestern Indiana: U.S. Geological Survey Water-Resources Investigations Report 92-4046.

Ault, C.H., Becker, L.E., Droste, J.D., Keller, S.J., and Shaver, R.H., 1976, Map of Indiana showing thickness of Silurian rocks and location of reefs and reef-induced structures: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 22, scale 1:500,000.

Basch, M.E., and Funkhouser, R.V., 1985, Irrigation impacts on ground-water levels in Jasper and Newton Counties, Indiana, 1981-1984: Indiana Department of Natural Resources, Division of Water, Water Resource Assessment 85-1, 109 p.

Bassett, J.L., and Hasenmueller, N.R., 1979, Map showing structure on base of New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources, Geological Survey, EGSP Series 800, scale 1:500,000.

Bergeron, M.P., 1981, Effect of irrigation pumping on the ground-water system in Newton and Jasper Counties, Indiana: U.S. Geological Survey Water-Resources Investigations 81-38, 73 p.

Bleuer, N.K., 1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley System (so-called Teays Valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

Bleuer, N.K., and Melhorn, W.N., 1989, Glacial terrain models, north-central Indiana—the application of downhole logging to analysis of glacial vertical sequences: Field Trip 2, Geological Society of America, North-Central Section, Notre Dame University, South Bend, Ind., April 19, 1989, p. 41-93.

Bretz, J.H., 1955, Geology of the Chicago region, part II—the Pleistocene: Illinois State Geological Survey Bulletin 65, 132 p.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Fraser, G.S., and Bleuer, N.K., 1991a, Geologic framework of the aquifers in the Kankakee River Lowland: Indiana Department of Natural Resources, Geological Survey Occasional Paper 60, 10 p.

_____1991b, Geologic framework of the aquifers of the Valparaiso Moraine: Indiana Department of Natural Resources, Geological Survey Occasional Paper 59, 8 p.

Gray, H.H., 1972, Lithostratigraphy of the Maquoketa Group (Ordovician) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 7, 31 p.

_____1982, Map of Indiana showing topography of the bedrock surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 36, scale 1:500,000.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Gutschick, R.C., 1976, Geology of the Kentland structural anomaly, northwestern Indiana: Kalamazoo, Mich., Western Michigan University, Field Guide of the North Central Section of the Geological Society of America, 60 p.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hartke, E.J., Hill, J.R., and Reshkin, Mark, 1975, Environmental geology of Lake and Porter Counties, Indiana—an aid to planning: Indiana Department of Natural Resources, Geological Survey Special Report 11, Environmental Study 8, 57 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hunn, J.D., and Reussow, J.P., 1968, Preliminary evaluation of the ground-water resources in the Calumet-Kankakee hydrologic area—appendix to the State Water Plan: Indiana Department of Natural Resources, Division of Water, 39 p.

Hunn, J.D., and Rosenshein, J.S., 1969, Geohydrology and ground water potential of St. Joseph County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 33, 20 p.

Indiana Department of Natural Resources, 1990, Water resource availability in the Kankakee River basin, Indiana: Indiana Department of Natural Resources, Division of Water, 247 p.

Indiana [State of], U.S. Department of Agriculture, and U.S. Department of Interior, 1976, Report on the water and related land resources, Kankakee River basin: Lincoln, Nebr., U.S. Department of Agriculture, Soil Conservation Service, 269 p.

Rosenshein, J.S., 1961, Ground-water resources of northwestern Indiana—preliminary report, Lake County: Indiana Department of Conservation, Division of Water Resources Bulletin 10, 229 p.

_____1962, Ground-water resources of northwestern Indiana—preliminary report, Porter County: Indiana Department of Conservation, Division of Water Resources Bulletin 12, 131 p.

_____1963, Recharge rates of principal aquifers in Lake County, Indiana: Ground Water, v. 1, no. 4, p. 13-20.

Rosenshein, J.S., and Hunn, J.D., 1962a, Ground-water resources of northwestern Indiana—preliminary report, LaPorte County: Indiana Department of Conservation, Division of Water Resources Bulletin 13, 183 p.

_____1962b, Ground-water resources of northwestern Indiana—preliminary report, St. Joseph County: Indiana Department of Conservation, Division of Water Resources Bulletin 15, 318 p.

_____1964a, Ground-water resources of northwestern Indiana—preliminary report, Jasper County: Indiana Department of Conservation, Division of Water Resources Bulletin 25, 83 p.

_____1964b, Ground-water resources of northwestern Indiana—preliminary report, Marshall County: Indiana Department of Conservation, Division of Water Resources Bulletin 19, 157 p.

_____1964c, Ground-water resources of northwestern Indiana—preliminary report, Newton County:

Indiana Department of Conservation, Division of Water Resources Bulletin 26, 69 p.

_____1964d, Ground-water resources of northwestern Indiana—preliminary report, Starke County: Indiana Department of Conservation, Division of Water Resources Bulletin 22, 87 p.

_____1968a, Geohydrology and ground-water potential of Lake County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 31, 35 p.

_____1968b, Geohydrology and ground-water potential of Porter and LaPorte Counties, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 32, 22 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H., 1974, The Muscatatuck Group (new Middle Devonian name) in Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 3, 7 p.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

MAUMEE RIVER BASIN

By Theodore K. Greeman

General Description

The Maumee River basin in northeastern Indiana is 1,283 mi^2 and includes parts of Adams, Allen, Dekalb, Noble, and Steuben Counties (fig. 30). Principal cities within the Maumee River basin include Auburn, Decatur, Fort Wayne, Garrett and New Haven. The Maumee River begins in Fort Wayne, Ind. at the confluence of the St. Marys and St. Joseph Rivers. Most of the Maumee River basin in Indiana is drained by these two tributaries. From the confluence, the Maumee River flows 28 mi east-northeast to the Indiana-Ohio State line. The mouth of the Maumee River is in northwestern Ohio, at the southwestern end of Lake Erie. In Ohio, the Maumee River flows 108 mi to Lake Erie; thus, the total length of the Maumee River is 136 mi.

Previous Studies

Before 1960, several authors studied the hydrogeology of the Maumee River basin; however, available data were limited. Leverett (1897) reported scattered observations on ground-water availability in Indiana. Harrell, who wrote the first comprehensive report on the ground-water resources in the Maumee River basin (1935), inventoried the ground-water resources of the counties and determined the "general principles of the occurrence of ground waters in Indiana" (1935, p. 1). Stallman and Klaer (1950) described ground-water availability in Noble County and presented lithologic logs of numerous wells, as well as a potentiometric-surface map.

Enactment of laws requiring drillers to report lithologic and hydrologic properties of water wells drilled in Indiana after 1958 quickly created a new data base. Watkins and Ward (1962) reported on the ground-water resources of Adams County and included ground-water-quality information. Herring (1969) described the ground-water resources of the Indiana part of the Maumee River basin and identified principal aquifers, potential yields, and ground-water quality. Pettijohn and Davis (1973) prepared a hydrologic atlas of the water resources of the Maumee River basin in Indiana; their report includes selected information on ground-water quality and surface-water quality, water budget, surface-water flow duration, surface-water stage and discharge, and potentiometric surface. Bleuer and Moore (1972) described and correlated glacial stratigraphy in the Fort Wayne area. Bleuer and Moore (1978) continued this work by identifying and mapping unconsolidated stratigraphic units throughout Allen County. The latter report describes the stratigraphic framework of glacial deposits, ground-water availability, ground-water quality, potential for deep-well disposal, and other hydrologic factors relating to environmental issues in Allen County. Planert (1980) modeled a 700-mi^2 area in northeastern Indiana. He studied several ground-water sources and simulated several ground-water-withdrawal scenarios to evaluate ground-water-level declines and streamflow losses in northwestern Allen County.

Physiography

The Maumee River drainage basin includes three distinct physiographic units in Indiana (Malott, 1922, p. 66): the Steuben Morainal Lake Area, the

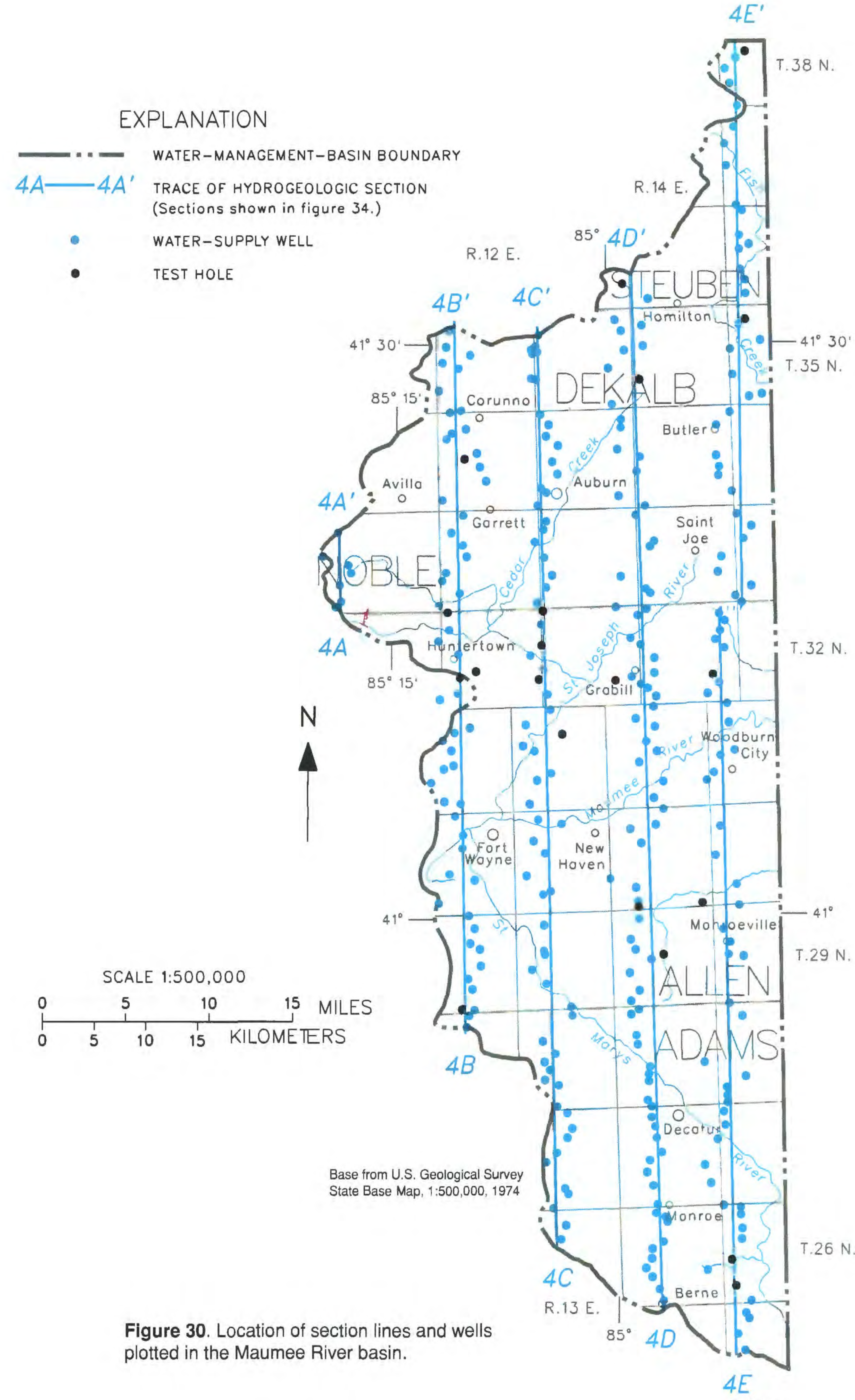


Figure 30. Location of section lines and wells plotted in the Maumee River basin.

Maumee Lacustrine Plain, and the Tipton Till Plain (fig. 2). Surface drift in the Maumee River basin was deposited between 21,000 and 14,000 years ago (Gooding, 1973, p. 24) during the latest Wisconsinan glaciation, although some alluvial deposits are post glacial.

The Steuben Morainal Lake Area (fig. 31) is an undulating region of moraines and kettle lakes, characterized by "knob and kettle" topography. "Knob" refers to the morainal hills whose local relief ranges from 100 to 200 ft. These knobs are composed of till or ice-contact sand and gravel (Schneider, 1966, p. 53). Kettle refers to the numerous surface depressions in the drift, many of which are water filled. When deposited, the moraines were composed of till mixed with sand, gravel and detached blocks of glacial ice. Kettles are depressions formed as buried ice melted and overlying sediments collapsed.

The morainal landforms (fig. 31) in the Steuben Morainal Lake Area were formed during Late Wisconsinan time, by at least two surges of the Huron-Erie Lobe and an intermediate surge of the Saginaw Lobe (fig. 8). Ice of the Huron-Erie Lobe deposited tills when it advanced into northern and central Indiana between 23,000 and 17,000 years ago (Shaver and others, 1970, p. 177). Following withdrawal of the Huron-Erie Lobe, ice of the Saginaw Lobe advanced into northeastern Indiana. A resurgence in the Huron-Erie Lobe then separated the Saginaw Lobe from its source. The resurgent Huron-Erie Lobe overran the older tills and deposited a new till on top. Deposition of this younger till was completed about 13,000 to 14,000 years ago (Wayne, 1963, p. 44). This surficial till commonly is less than 10 ft thick on northern parts of the Fort Wayne, Wabash and Salamonie Moraines (A.J. Fleming, Indiana Geological Survey, written commun., 1990).

The Maumee Lacustrine Plain, which covers more than 120 mi^2 of Indiana (fig. 31) (Malott, 1922, p. 151), is an area once occupied by glacial Lake Maumee. In this nearly level plain, lake sediments (lacustrine deposits) are incorporated within the youngest Wisconsinan till deposits that cover the lake-bottom plain (A.J. Fleming, Indiana Geological Survey, oral commun., 1990). Topographic relief on the lake plain is low with slopes commonly less than 5 ft/mi. Surficial sand deposits at the perimeter of the plain are Lake Maumee strand or beach deposits. The geomorphology of the lake plain can also be interpreted as a glacially scoured lowland. The lake is possibly best characterized as a subglacial, water-filled basin that catastrophically drained its water when the ice of the Huron-Erie Lobe melted (A.J. Fleming, Indiana Geological Survey, oral commun., 1990).

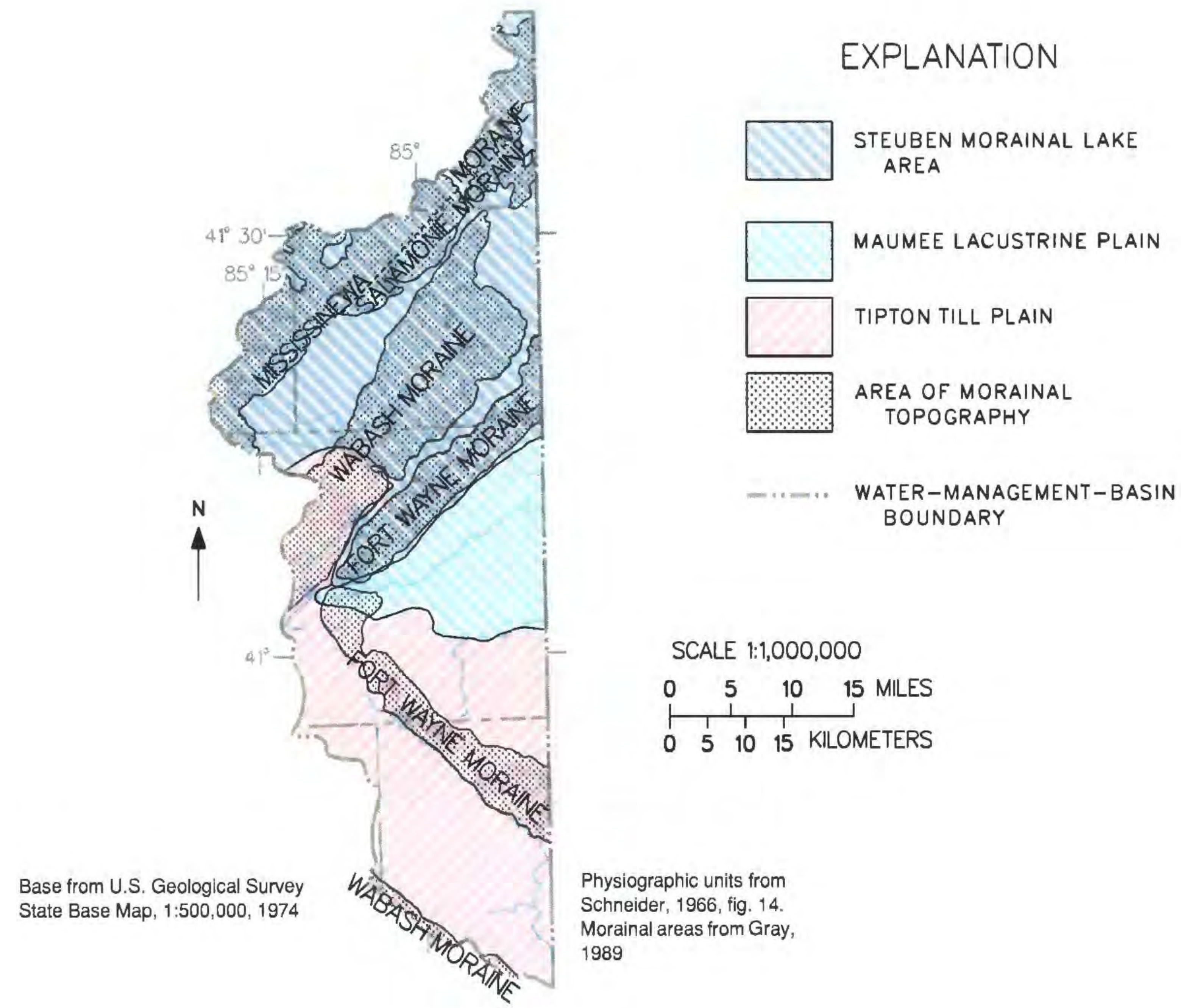


Figure 31. Physiographic units and moraines in the Maumee River basin.

Water levels in glacial Lake Maumee reached a high-stage altitude of 800 ft above sea level (Bleuer and Moore, 1978, p. 59) sometime between 14,000 and 11,800 years ago. At this stage, glacial Lake Maumee drained over the Wabash Moraine at Fort Wayne, continued to the southwest by way of the Little Wabash River (Wabash-Erie channel) (fig. 7) and joined the Wabash River at Huntington. Outflow from Lake Maumee scoured the Fort Wayne outlet to an altitude of about 750 ft.

After the glaciers retreated from the Lake Erie basin, water levels in Lake Maumee declined. The Fort Wayne outlet was abandoned as other drainage outlets opened. The Wabash Moraine became the major divide between drainage to the St. Lawrence River and the Mississippi River (Wright and Frey, 1965, p. 90). Lake Erie, the low-stage equivalent of ancient Lake Maumee, stabilized near its present pool altitude of approximately 570 ft (Department of the Army, 1989) at least 11,800 years ago (Wright and Frey, 1965, p. 90).

After Lake Erie stabilized, the lake bed was exposed to erosion. In Indiana, the Maumee River has downcut 25 to 40 ft into the glaciolacustrine sediments (Pettijohn and Davis, 1973, pl. 1) and, thus, has failed to develop a flood plain.

The Tipton Till Plain (fig. 31) is a nearly level to gently rolling, poorly drained glacial plain (Schneider, 1966, p. 49-50). The till plain is underlain by Huron-Erie Lobe tills and some stratified sediments. Resurgent periods during retreat of the last glacial ice produced the Fort Wayne and Wabash Moraines. Relief across the moraines is generally less than 50 ft, although the relief is slightly greater in several areas. These moraines range from 2 to 6 mi in width.

Surficial till composition is similar throughout the Tipton Till Plain of the Maumee River basin. Surficial tills are characterized by high clay content (50-55 percent) and low sand content (15-20 percent) (Gooding, 1973, p. 9). Glacial stratigraphy in the Tipton Till Plain is horizontally continuous and less complex than the stratigraphy of the Steuben Morainal Lake Area. Slopes are generally less than 1 ft per 100 ft.

Surface-Water Hydrology

The Maumee River and its principal tributaries drain 6,608 mi^2 of northeastern Indiana, southern Michigan, and northwestern Ohio. In Indiana, the Maumee River drains 1,283 mi^2 (Ohio Department of Natural Resources, 1985). The altitude of the Maumee River channel bottom at Fort Wayne is 728 ft above sea level, whereas at the Indiana-Ohio State line it is about 700 ft (Simpson, 1988). The average gradient of the meandering channel is 1.0 ft/mi in Indiana. This low gradient continues downstream 87 mi to Waterville, Ohio, where the Maumee River altitude is 610 ft. The gradient of the Maumee River in Indiana and northwestern Ohio is low because near-surface bedrock impedes downcutting of the channel. Between Waterville, Ohio, and Lake Erie, the Maumee River channel bottom drops 38 ft in 21 mi (a downstream gradient of 1.8 ft/mi).

Moraines divide the Maumee River basin into four subbasins that are drained by the Maumee River, the St. Marys River, the St. Joseph River, and Cedar Creek (fig. 30). The Maumee River drains a total of 2,129 mi^2 at the streamflow-gaging station at Antwerp, Ohio (7 mi downstream of the Indiana-Ohio State line). At Antwerp, the median discharge of the Maumee River (1939-85) is 631 ft^3/s, and the average runoff is 10.80 in/yr (D. V. Arvin, U.S. Geological Survey, written commun., 1991). Flatrock Creek drains 99.8 mi^2 of the Maumee River subbasin in Indiana before entering Ohio and joining the Maumee River downstream. Flatrock Creek has little or no flow at times during most dry years.

The St. Marys River begins in Ohio and flows northwest to Fort Wayne, draining a total of 840 mi^2, of which 401 mi^2 are in Indiana (Hoggatt, 1975, p. 56). The St. Marys River has a median discharge of 134 ft^3/s (as measured at a streamflow-gaging station 10.8 mi above the mouth, 1930-85), and has an average runoff of 10.32 in/yr (Arvin, 1989, p. 850).

The St. Joseph River flows southwest and drains the northern half of the Maumee River basin. Cedar Creek is the main tributary of the St. Joseph River. (Cedar Creek is 7 to 10 mi northwest of the St. Joseph River.) Cedar Creek flows parallel to the St. Joseph River through Dekalb County before turning southeast and joining the St. Joseph River in Allen County. The St. Joseph River and Cedar Creek drain a total of 1,086 mi^2, of which 843 mi^2 are in Indiana (Hoggatt, 1975, p. 56). At the streamflow-gaging station near Fort Wayne, the median discharge of the St. Joseph River (1942-55, 1984-85) is 391 ft^3/s (Arvin, 1989, p. 843); the average runoff for this period of record is 12.09 in/yr.

Several streams draining the Maumee River basin were tributary to adjacent basins during the last glacial period. Cedar Creek formerly flowed into the Eel River of the Wabash River basin near Huntertown in Allen County. As stream discharge decreased and glacial ice retreated from the area, flow over the Wabash Moraine stopped and the Maumee River basin began draining to the St. Lawrence River (Bleuer and Moore, 1978, p. 63). None of the rivers in the Maumee River basin of Indiana flow on bedrock, although alluvium thickness is less than 20 ft in several areas.

Geology

Bedrock Deposits

In northeastern Indiana, the bedrock is composed of Paleozoic limestone, dolomite, and shale that overlie Precambrian granite, basalt, arkose and other rocks (Shaver and others, 1986, pl. 2). These bedrock units have been moderately deformed by several structural elements (fig. 4).

The Cincinnati Arch is a broad bedrock anticline along the Indiana-Ohio State line. The axis of the anticline trends north-northwest from Cincinnati. In Randolph County, about 30 mi south of the Maumee River basin, the Cincinnati Arch splits. North of the split, the axis of the Cincinnati Arch trends northwest across Indiana. The axis of the other branch, the Findlay Arch, trends northeast across Ohio to Lake Erie. The Maumee River channel parallels the axis of the Findlay Arch. The entire Maumee River basin is located on the north-dipping flanks of the Cincinnati and Findlay Arches.

Bedrock in the Maumee River basin dips north into southern Michigan (Michigan Basin) at 15 to 22 ft/mi (sections 4A–4A′ to 4E–4E′, fig. 34). As the sequence of Paleozoic bedrock dips northward into the Michigan Basin, the entire sequence thickens and increasingly younger rocks are at the bedrock surface (fig. 32).

In the Maumee River basin, Precambrian rocks are buried under more than 3,000 ft of lithified sediments. About 1,000 ft of Cambrian sediments overlie the Precambrian basement in the Maumee River basin. About 2,000 ft of Ordovician sediments overlie the Cambrian sediments.

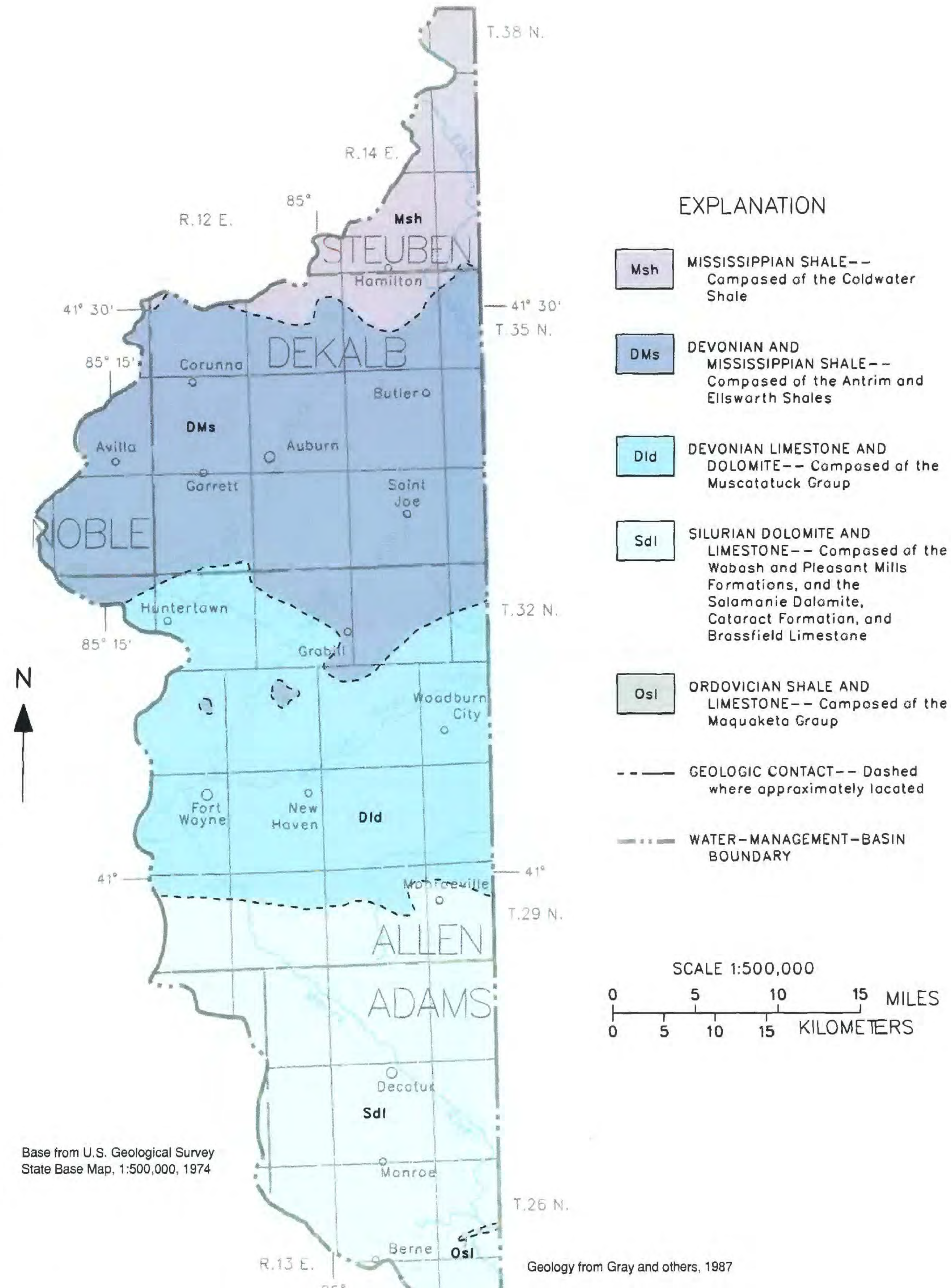


Figure 32. Bedrock geology of the Maumee River basin.

Although there are several potential aquifers in the Cambrian and Ordovician rocks, their potential as a source of potable water is doubtful. Water in these deep aquifers is highly mineralized, making the high transmissivity of the rock more desirable for waste disposal than for withdrawal of water (Bleuer and Moore, 1978, p. 48-49). The most notable of these potential disposal aquifers is the Cambrian-age Mt. Simon Sandstone. Two potential disposal aquifers within the Ordovician rocks include the Knox Dolomite and the Trenton Limestone.

Subcrops of Paleozoic bedrock underlie the drift throughout the Maumee River basin. Depth to bedrock in the Maumee River basin ranges from near zero in several areas to about 450 ft (fig. 33). The buried bedrock surface ranges in altitude from about 500 ft above sea level at the base of a deep preglacial stream channel to about 900 ft above sea level near the Indiana-Michigan State line. Rocks exposed at the bedrock surface range in age from 450 million to 355 million years (Palmer, 1983).

The oldest rocks that directly underlie the drift in northeastern Indiana are Upper Ordovician interbedded shales and limestones (fig. 32). Less than 50 ft of Upper Ordovician rock is exposed at the base of the preglacial St. Marys Bend Segment of the Lafayette Bedrock Valley (fig. 7) (Bleuer, 1989, table 1). This is the only area in the Maumee River basin where Ordovician rocks were exposed by preglacial erosion.

Overlying the Ordovician rocks throughout the Maumee River basin is a thick sequence of Paleozoic carbonate rocks. This sequence of carbonate rocks is composed of the Silurian Salina Group and the Devonian Muscatatuck Group (fig. 32). The carbonate rocks are composed of layered limestone, dolomite, and some thin shale beds. Their combined thickness attains 700 ft. The carbonate rocks are present as a subcrop below the drift south of the Maumee River. At the southern tip of the Maumee River basin, in the St. Marys Bend Segment of the Lafayette Bedrock Valley, the entire carbonate rock sequence has been eroded. The carbonate rock sequence transmits water through fractures and solution openings, and it is used as a source of ground water. A paleokarst topography preserved beneath the drift indicates that these carbonate rocks were drained by subterranean drainage. North of the Maumee River, the entire carbonate rock sequence dips below younger shales.

In Indiana and Ohio, the preglacial Lafayette Bedrock Valley (formerly Teays River Valley) drained a 35,000-mi^2 carbonate bedrock plain (L.D. Arihood, U.S. Geological Survey, oral commun., 1989). In the Maumee River basin, broad upland areas between deeply entrenched streams were underlain by 400 ft or more of carbonate rock. The entrenchment of Tertiary streams and the presence of buried karst topography indicate deep preglacial ground-water levels in the carbonate rocks. Locally, preglacial ground-water levels were near the base of the carbonate rocks.

The youngest rocks found in the Maumee River basin are Paleozoic shales (fig. 32). The Devonian Antrim Shale directly overlies the Devonian carbonate rocks. Three younger shale units overlie the Antrim Shale in the Maumee River basin: the Ellsworth Shale of Devonian and Mississippian age, and the Sunbury and Coldwater Shales of Mississippian age. North of the Maumee River a subcrop of these shales is commonly overlain by more than 200 ft of unconsolidated drift (fig. 33). Although these shales have been eroded south of the Maumee River (fig. 32), they attain a thickness of about 850 ft at the Indiana-Michigan State line. In Branch County, Mich., about 15 mi north of the Indiana-Michigan State line, shale is exposed at the land surface. These shales restrict the circulation of ground water.

Unconsolidated Deposits

During the Pleistocene Epoch, ice, thousands of feet thick, flowed repeatedly into Indiana. The Erie Lobe advanced west-southwest across northern Ohio following a carbonate-bedrock dip slope. The Huron Lobe advanced south out of the Lake Huron basin. The two lobes of glacial ice coalesced in northwestern Ohio to form the Huron-Erie Lobe. The Huron-Erie Lobe advanced up onto the Findlay and Cincinnati Arches. Some deposits from previous periods of glaciation were removed by further advances.

As the Huron-Erie Lobe moved into northeastern Indiana, it was forced to change direction by another lobe of glacial ice (fig. 8). The other lobe, known as the Saginaw Lobe, advanced into north-central Indiana from the Saginaw Bay area of Michigan. The Saginaw Lobe and the Huron-Erie Lobe abutted each other along the northwest boundary of the Maumee River basin in Indiana. This blockage forced the Huron-Erie Lobe to advance southward across eastern Indiana. The area where the two lobes of glacial ice abutted each other is underlain by thick drift. This thick drift deposit is herein called the "moraine complex." The moraine complex has the surface characteristics of end moraines.

The St. Marys River flows near the bedrock surface along much of its course through Indiana; however, bedrock is 90 ft below the streambed 3.5 mi upstream from the confluence with the St. Joseph River (section 4B–4B′, fig. 34). Drift thickness south of the Maumee River (fig. 33) is generally from 50 to 100 ft, except where as much as 300 ft of glacial deposits fill preglacial valleys. All areas where drift thickness is less than 50 ft are south of the Maumee River. In some areas south of the Maumee River (sections 4A–4A′ to 4E–4E′, fig. 34), a rubble layer composed of silt, fine sand, and broken dolomite overlies the bedrock. Where present, this rubble layer is generally less than 5 ft thick.

Tills are exposed at land surface over most of the Maumee River basin in Indiana. Nearly all of the till is from the Lagro Formation of Wisconsinan age (Gray, 1989). The Lagro Formation was deposited by the Huron-Erie Lobe more than 13,000 years ago (Wayne, 1963, p. 44). Along the northwestern boundary of the Maumee River basin, surface deposits are a mix of Huron-Erie Lobe and Saginaw Lobe associations, especially near Huntertown in Noble County (A.J. Fleming, Indiana Geological Survey, written commun., 1990).

Glacial drift south of the Maumee River contains at least two identifiable till units, the Lagro and the Trafalgar Formations (Bleuer and Moore, 1978, p. 11). The upper unit, the Lagro Formation was deposited over lake sediments during the last glacial advance into the area. The Lagro Formation contains 10 to 20 percent sand and 40 to 50 percent clay. The base of the Lagro Formation is characteristically rich in clay because of the incorporation of the lake sediments.

The lower till unit, known as the Trafalgar Formation, is also of Late Wisconsinan age. The Trafalgar Formation contains 35 to 45 percent sand, and 15 to 20 percent clay (Bleuer and Moore, 1978, p. 11). Typically the Trafalgar Formation is extremely hard. Dense, silty sand and gravel is locally present between these till units. In most areas, this sand and gravel is too silty to be used as an aquifer; locally however, it forms small channels of coarse, shaly gravel that are tapped by domestic wells. Where coarse, this intertill sand and gravel unit is composed of 50 to 90 percent shale clasts. (A.J. Fleming, Indiana Geological Survey, written commun., 1990).

North of the Maumee River, intertill sand and gravel deposits are abundant throughout the drift. These sand and gravel deposits were concentrated into layers by meltwater that transported the silt and clay away. Some intertill deposits form laterally continuous horizons, whereas others are discontinuous. Layered intertill deposits were formed by repeated advances and retreats of glacial ice during the Pleistocene. These horizontally continuous sand and gravel deposits are disrupted in many areas. Collapse of sediments into voids left by buried ice, and lateral movement of sediments resulting from postdepositional ice advances, have produced complex deposits.

Surficial sand and gravel deposits are uncommon in the Maumee River basin of Indiana. They can be found as valley-train deposits along the St. Joseph and St. Marys Rivers, Cedar Creek, and the Fort Wayne outlet of ancient Lake Maumee. A

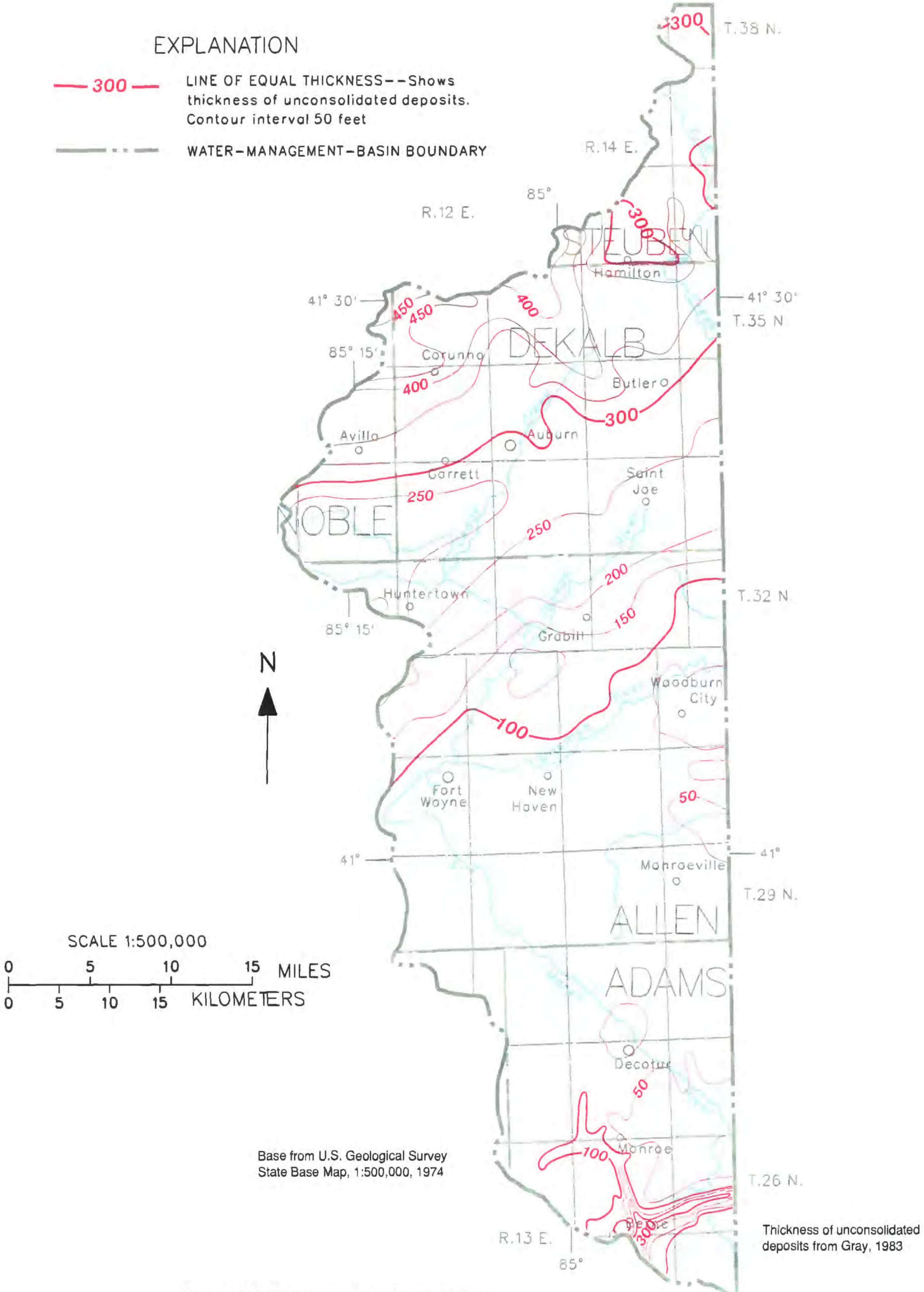


Figure 33. Thickness of unconsolidated deposits in the Maumee River basin.

large buried outwash-fan deposit (section 4C–4C′, fig. 34) is partially exhumed by Cedar Creek.

Few post-Wisconsinan deposits are found in the Maumee River basin. Some minor alluvial deposits, primarily composed of reworked outwash, are found in association with the Wisconsinan outwash deposits. Because the Maumee River formed after glaciation ended, none of the alluvial deposits along the Maumee River are reworked outwash. Minor deposits of muck, peat and marl, all of Holocene age, also are present in the basin.

Aquifer Types

Five hydrogeologic sections (sections 4A–4A′ to 4E–4E′, fig. 34) were produced for this atlas to depict aquifer types in the Maumee River basin. The hydrogeologic sections are oriented south-north, approximately perpendicular to the Maumee River, and are spaced 6 mi apart (fig. 30). The section lines produced from logs of 305 water-supply and test wells have a total length of 244 mi. The average density of logged wells plotted along the sections is 1.3 wells per mile, (approximately one well every 4,000 ft).

The map showing the extent of aquifers in the Maumee River basin (fig. 35) was constructed by use of the hydrogeologic sections. Additional information for the aquifer map was from the Quaternary geologic map of Indiana (Gray, 1989) and other publications referenced in this chapter.

The aquifer map shows the five aquifer types that are commonly used for water supply in the Maumee River basin. Of these aquifer types, four are restricted to the unconsolidated deposits. These include surficial sand and gravel aquifers, buried sand and gravel aquifers, discontinuous sand and gravel aquifers in isolated deposits, and sand and gravel in buried bedrock valleys. The only bedrock aquifer type used in this basin is carbonate bedrock of Silurian and Devonian age. Characteristics of the five aquifer types mapped in the Maumee River basin are summarized in table 6.

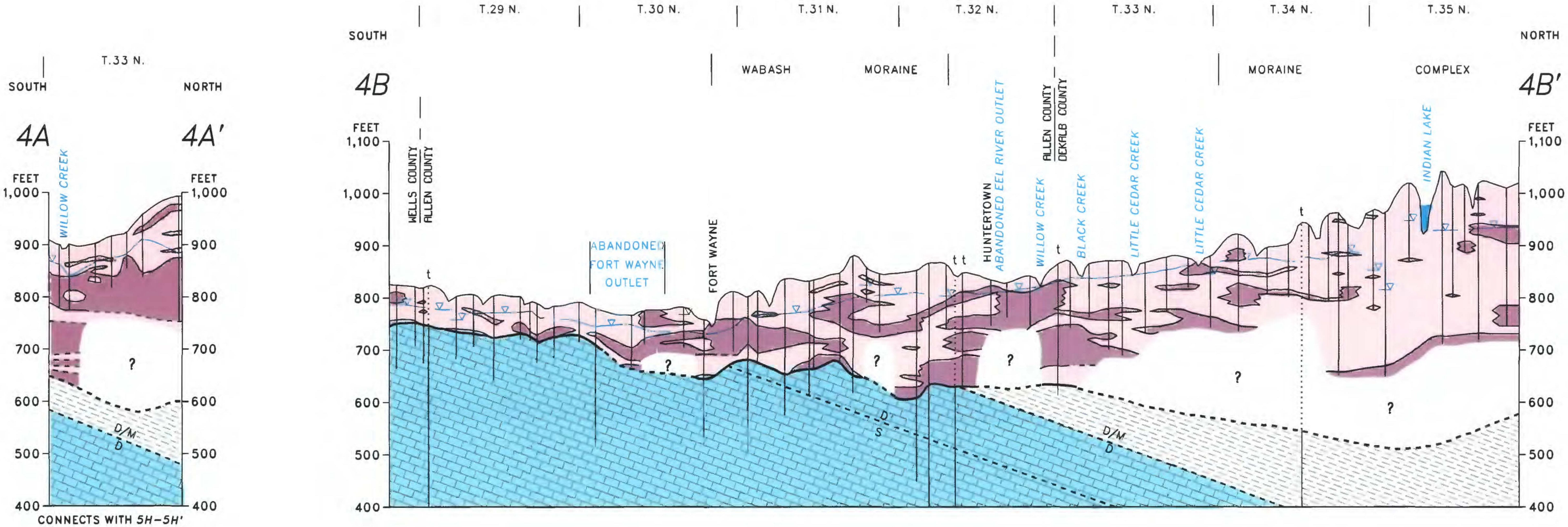
Unconsolidated Aquifers

Surficial Sand and Gravel Aquifers

Surficial sand and gravel aquifers have been deposited by present-day streams (alluvial) and by glacial meltwater (outwash). Alluvial deposits adjacent to the St. Marys and St. Joseph Rivers, and Cedar Creek (fig. 35) are generally thin and are not significant sources of ground water. Most present-day streams follow former glacial drainage channels and have reconfigured the glacial deposits. Because it does not follow a glacial drainage channel, the Maumee River has few alluvial deposits associated with it.

The bulk of the surficial sand and gravel aquifers in the Maumee River basin are composed of outwash. Most of the major streams draining the basin have outwash deposits adjacent to their channel (fig. 35). The outwash deposits in the basin are commonly 1/2 to 1 mi wide and as much as 40 ft thick (section 4B–4B′ and 4C–4C′, fig. 34). The Fort Wayne outlet to the Wabash-Erie Channel in southwestern Fort Wayne (fig. 7) and the Cedar Creek outlet to the Eel River near Huntertown have little or no outwash along their channels. These former outlet channels are located on divides, and streams no longer occupy them. Although the Fort Wayne outlet is up to 3 mi wide, only minor outwash deposits are present (section 4B–4B′, fig 34).

One additional type of surficial sand deposit in the Maumee River basin is beach sand. East of Fort Wayne, deposits of beach sand follow the former shoreline of ancient Lake Maumee. These sand deposits are at an altitude of about 760 to 775 ft, which corresponds to the altitude of the Fort Wayne outlet (sections 4D–4D′ and 4E–4E′, fig. 34). Beach ridge deposits are not on the aquifer map, because they generally are not productive aquifers in the Maumee River basin. Most unconsolidated sand and gravel aquifers were buried under clay-loam till during glacial disintegration.



EXPLANATION

SAND AND GRAVEL

UNCONSOLIDATED NONAQUIFER MATERIAL

LIMESTONE AND DOLOSTONE

SHALE

LIMESTONE AND SHALE

BEDROCK AQUIFER

BEDROCK NONAQUIFER

? NO DATA

BEDROCK SURFACE--Dashed where approximately located

CHRONOSTRATIGRAPHIC BOUNDARY--Dashed where approximately located

LITHOLOGIC CONTACT--Dashed where approximately located

GENERALIZED POTENTIOMETRIC SURFACE--Dashed where approximately located

WELL--All well data are projected to trace of section. Dotted where data are incomplete

t TEST HOLE--Not drilled for water supply

M MISSISSIPPIAN

D DEVONIAN

S SILURIAN

O ORDOVICIAN

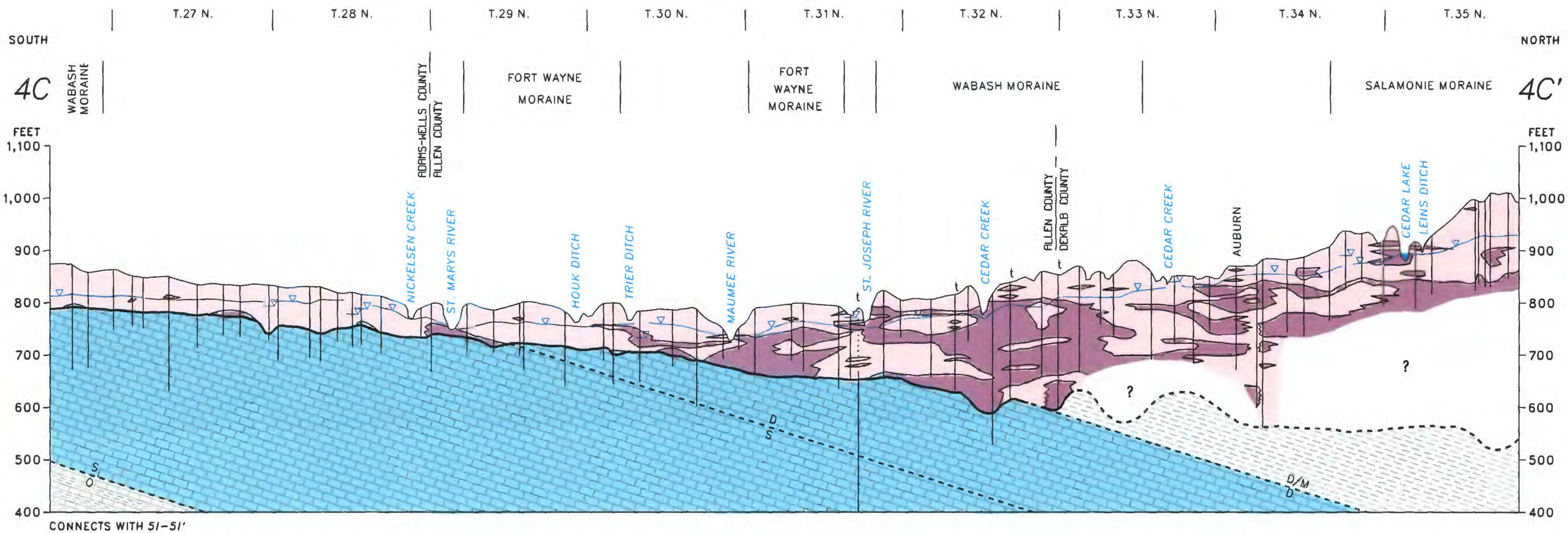
DATUM IS SEA LEVEL

VERTICAL SCALE GREATLY EXAGGERATED

0 5 10 MILES

0 5 10 KILOMETERS

Figure 34. Hydrogeologic sections *4A–4A′* to *4E–4E′* of the Maumee River basin.



Buried Sand and Gravel Aquifers

Buried sand and gravel aquifers (sections 4A–4A′ to 4E–4E′, fig. 34) are laterally continuous deposits that were formerly coalescing outwash fans, outwash plains, kame terraces and other ice-contact stratified deposits. Because these aquifers were not deposited uniformly, aquifer characteristics such as thickness, texture, and distribution are highly variable. Post depositional slumping of super saturated deposits, ice collapse, and erosion have caused additional disruption of these aquifer materials. The aquifer map (fig. 35) indicates areas where one or more of these buried aquifers is present.

Although most buried sand and gravel aquifers are not exposed by stream erosion, buried sand and gravel aquifers are exposed by Little Cedar Creek in T. 33 N. and by ditching on the north side of the abandoned Fort Wayne outlet in T. 30 N. (section 4B–4B′, fig. 34); by Cedar Creek in T. 32 N. and the St. Joseph River in T. 31 N. (section 4C–4C′, fig 34); by West Branch and a tributary to Fish Creek in T. 37 N. and by the Maumee River in T. 31 N. (section 4E–4E′, fig. 34). From the northeastern corner of Indiana south along section 4E–4E′ (fig. 34) to the Maumee River, the potentiometric surface of the buried sand and gravel aquifer slopes south at 7.8 ft/mi. This closely approximates the slope of the land surface.

Nearly all ground-water production north of the Maumee River is from buried sand and gravel aquifers (fig. 34). Although highly variable, the average thickness of buried aquifers in Allen County is 25 ft (Bleuer and Moore, 1978, p. 46). The median well yield from large diameter wells (10 in. or larger) that are finished in buried sand and gravel deposits in Allen County is 250 gal/min; well yields in this area range from 20 to 500 gal/min (Bleuer and Moore, 1978, p. 46). Water in buried sand and gravel aquifers is of suitable quality for drinking.

Clark (1980, p. 211) indicates two areas north of the St. Joseph River in Dekalb County where more than 1,000 gal/min of ground water is available from buried sand and gravel aquifers. Equally large yields are probably available throughout this part of the basin, as large yields have been found wherever ground-water exploration has been done. The highest yield noted from a well plotted on the Maumee River basin hydrogeologic sections is a well in T. 34 N., R. 14 E. (section 4E–4E′, fig. 34). This well is 24 in. in diameter and is 180 ft deep, and it was test pumped at 2,250 gal/min.

An extensive sand and gravel aquifer, a buried outwash fan of Saginaw Lobe association (A.J. Fleming, Indiana Geological Survey, oral commun., 1991), underlies a 110-mi^2 area in northwestern Allen County and southwest Dekalb County. Centered on Huntertown, this buried outwash fan slopes southeast into the Maumee River basin from the crest of a buried moraine. Portions of this buried fan are mapped in Tps. 32 and 33 N. (sections 4A–4A′, 4B–4B′, and 4C–4C′, fig. 34). Planert (1980, p. 52) found that 30 to 39 Mgal/d could be developed from the unconsolidated sand and gravel aquifers in northwestern Allen County; however, he also found that these ground-water withdrawals would decrease streamflow in the St. Joseph River at Fort Wayne by 32 to 38 ft^3/s.

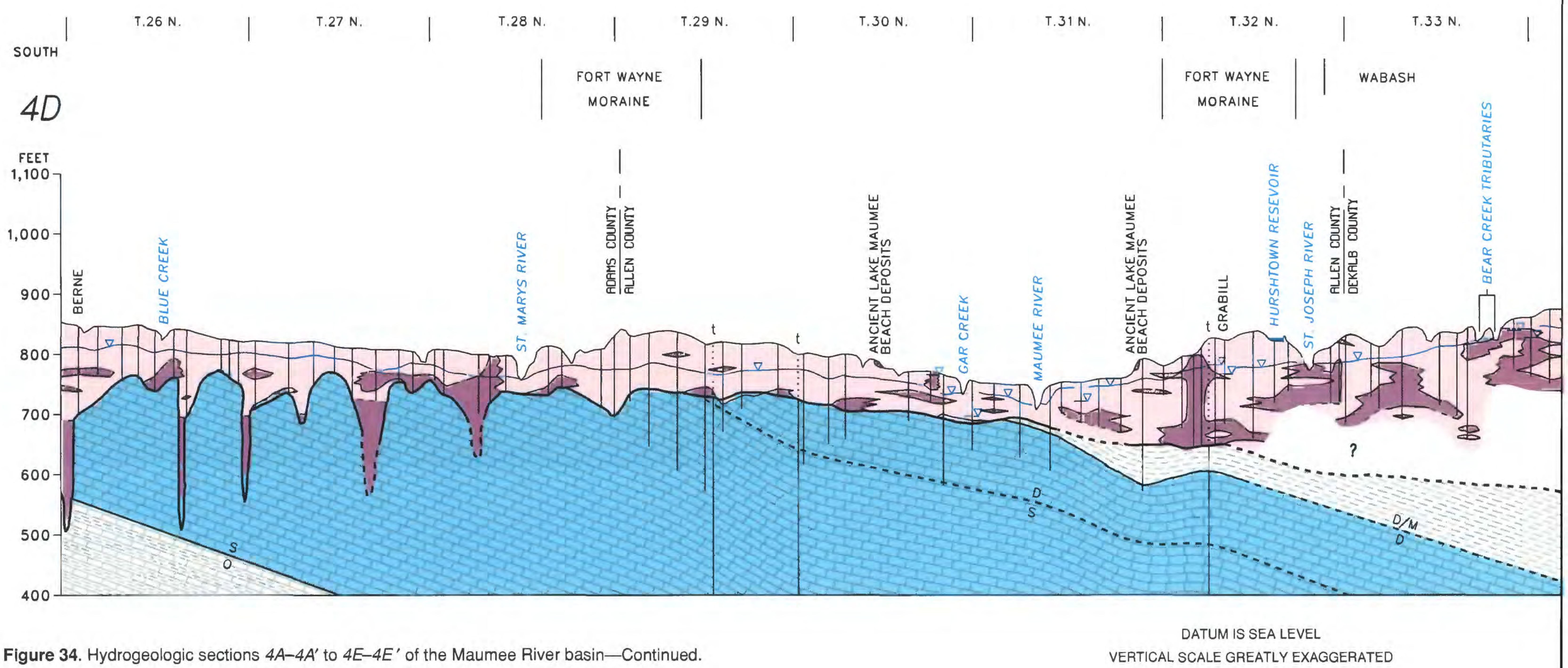


Figure 34. Hydrogeologic sections *4A–4A'* to *4E–4E'* of the Maumee River basin—Continued.

Discontinuous Sand and Gravel Aquifers

Discontinuous sand and gravel aquifers are similar to buried sand and gravel aquifers, except that they occur intermittently at one horizon or randomly dispersed throughout the drift. Some discontinuous buried sand and gravel was deposited in meandering channels. In the Maumee River basin, almost all discontinuous sand and gravel aquifers are south of the Maumee River. Drift in this area thins to less than 100 ft.

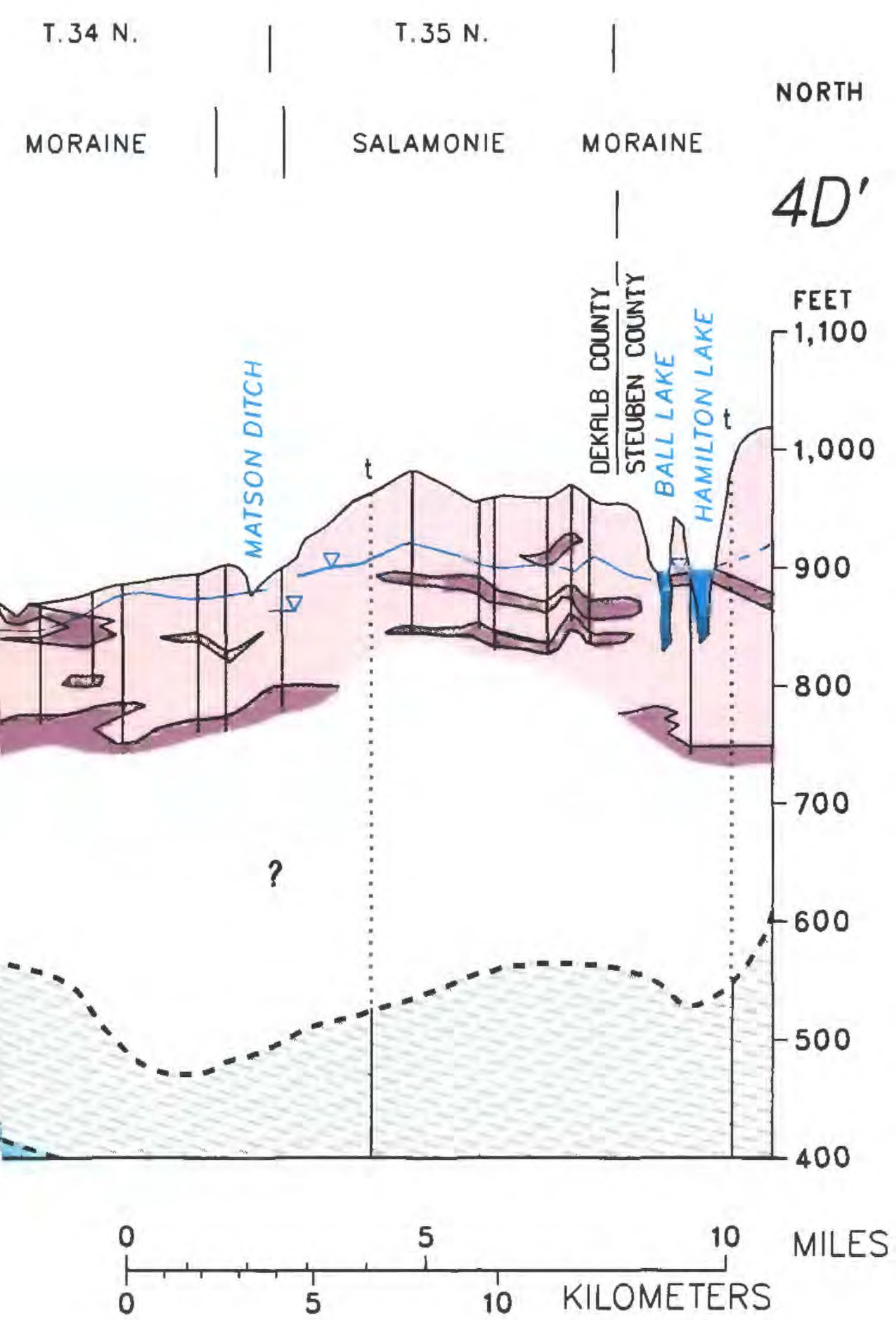
Discontinuous sand and gravel aquifers are rarely used for ground-water supply because of the abundance of fine sand, silt, and clay in the discontinuous aquifer and the availability of reliable ground-water supplies in the bedrock. Drillers commonly identify the discontinuous sand and gravel aquifer as "dirty gravel" in well logs. Yield from the discontinuous area south of the Maumee River is generally less than 20 gal/min. Few logs of wells drilled in this area report yields from the discontinuous aquifer, because drillers commonly finish the wells in the bedrock aquifer.

Sand and Gravel Aquifer in Buried Bedrock Valleys

A deep, preglacial bedrock valley underlies the drift in the southern tip of the Maumee River basin. This buried valley is known as the St. Marys Bend Segment of the Marion Valley Section of the Lafayette Bedrock Valley System (fig. 7) (Bleuer, 1989, p. 5), and was formerly known as the Teays Valley. Before burial, the Marion Valley Section had limestone cliffs that were 300 ft high. During the Pleistocene glacial period, this valley system filled with glacial and lake sediments. Most valley fill deposits in the basin are clay, although several sand and gravel zones are present. Some of the gravels are quite coarse (Bleuer and others, 1991, p. 86). High yields from these deposits are due to the coarseness of the aquifer, depth of burial, and the correspondingly large drawdowns available.

Bleuer (1991, fig. 6D, p. 63) indicates that there are three aquifer horizons in this part of the St. Marys Bend Segment. The upper aquifer is heavily used; the middle aquifer is fine grained sand and does not produce large volumes of water; the lowest aquifer is a coarse sand and gravel, which has the potential for the largest ground-water production.

The city of Decatur, Adams County, constructed its well field in the St. Marys Bend Segment of the Lafayette Bedrock Valley. This well field is near Berne, 12 mi south of Decatur, near the divide between the Maumee River basin and the Upper Wabash River basin. The well field pumps water from the upper aquifer. This aquifer is a 40-foot-thick, coarse-grained, cobbly sand and gravel deposit confined under 110 ft of clay. Well yields from this well field range from 1,000 to 1,400 gal/min with approximately 40 ft of drawdown (Bleuer and others, 1991, p 86).



Bedrock Aquifers

Carbonate Bedrock Aquifer

The carbonate bedrock aquifer underlies the entire Maumee River basin (sections 4A–4A′ to 4E–4E′, fig. 34), with the exception of the small area underlain by the Lafayette Bedrock Valley (section 4E–4E′, fig. 34) where carbonate rock has been eroded. In northern Allen, Dekalb, Noble and Steuben Counties, the bedrock dip and drift thickness restrict the use of the carbonate bedrock aquifer (fig. 35). The Silurian-Devonian carbonate bedrock aquifer is used predominately south of the Maumee River. The carbonate bedrock aquifer is below the water table throughout the Maumee River basin and is confined except where stone quarries dewater it for mining.

The median yield from large-diameter wells completed in the carbonate bedrock aquifer in Allen County is 175 gal/min; well yields range from 35 to 500 gal/min (Bleuer and Moore, 1978, p. 46). South of the Maumee River, the thickness of the drift overlying the carbonate bedrock aquifer averages about 100 ft, and most wells are completed in the bedrock. South of the Maumee River, the potentiometric slope in the carbonate bedrock aquifer slopes north at 3.7 ft/mi. The slope of the potentiometric surface closely approximates the land surface.

Karst development in this carbonate bedrock was extensive at the beginning of the Quaternary period. This secondary enhancement of bedrock permeability is responsible for the large well yields available from the carbonate bedrock aquifer. Planert (1980, p. 15) found the permeability of the carbonate bedrock aquifer in Allen County is greatest near the preglacial erosion surface and decreases with depth. The full carbonate bedrock aquifer sequence is about 700 ft thick. The smallest yields from the carbonate bedrock aquifer are found in the shallow bedrock areas south of the Maumee River and north of the St. Marys River. Northeast of Monroeville, in T. 30 N., R. 15 E., the maximum potential ground-water yield from the bedrock is 50 gal/min (Clark, 1980, p. 211).

The Maumee River is the lowest surface-water outlet draining the north-dipping part of the carbonate bedrock aquifer sequence in Indiana. Some of the ground-water flow in the carbonate bedrock aquifer sequence enters the Maumee River basin along the southwest boundary of the Maumee River basin. Greeman (1991) found that both the White River basin, and the Wabash River basin (fig. 1) contribute water to the ground-water flow in the carbonate bedrock aquifer.

The decision to complete a well in the carbonate bedrock aquifer can be based on absence of an adequate unconsolidated aquifer, problems associated with having to set a well screen in an unconsolidated aquifer, or water-quality considerations. The quality of water differs between the bedrock and unconsolidated aquifers. Locally, the carbonate rocks contain abundant gypsum ($CaSO4$), which is the major source of sulfate. Sulfate is a common ground water constituent that can occur at undesirable high concentrations. Bleuer and Moore (1978, p. 43) report that, in Allen County, water from the carbonate bedrock aquifer has significantly higher concentrations of dissolved strontium and sodium and significantly lower concentrations of dissolved bicarbonate, iron, and zinc than water from unconsolidated aquifers. None of these constituents present a health hazard in the concentrations reported.

Ordovician rocks form a confining unit underlying the Silurian-Devonian carbonate bedrock aquifer throughout the Maumee River basin. Similarly, till functions as a confining unit overlying the Silurian-Devonian carbonate bedrock aquifer. Because of these confining units, ground-water flow in the carbonate bedrock aquifer is isolated; long flow paths and a slow transmission rate cause ground water to remain in the carbonate bedrock aquifer for years. During the long transit, ground water can become heavily mineralized. Within the Maumee River basin, heavily mineralized water from the carbonate bedrock has been found south of the Maumee River. However, water in the Silurian-Devonian carbonate bedrock aquifer is substantially less mineralized than water in the more deeply buried Cambrian and Ordovician bedrock aquifers.

Summary

The Maumee River basin encompasses 1,283 mi^2 of northeastern Indiana and includes large parts of Adams, Allen, and Dekalb Counties and parts of Noble and Steuben Counties. The physiography of the area can be divided into three major regions: the Steuben Morainal Lake Area, the Lake Maumee Lacustrine Plain, and the Tipton Till Plain. Topography ranges from nearly level in the Maumee Lacustrine Plain to hilly (more than 200 ft of relief) in the Steuben Morainal Lake Area.

Moraines divide the Maumee River basin into four subbasins that are drained by the Maumee River, the St. Marys River, the St. Joseph River, and Cedar Creek. The St. Marys River and the St. Joseph River converge at Fort Wayne to form the Maumee River. Surface-water drainage in the Maumee River basin is part of the St. Lawrence River system.

Bedrock underlying the glacial drift ranges from Late Ordovician to Early Mississippian in age. Silurian and Devonian carbonate bedrock, less than 300 ft deep, underlies the southern half of the basin. The Silurian-Devonian carbonate sequence ranges from 0 to 700 ft in thickness. The carbonate bedrock sequence is absent at the base of the St. Marys Bend segment of the preglacial Lafayette Bedrock Valley. Younger shales overlie the carbonate bedrock sequence in the northern half of the basin. All surface sediments are glacial or fluvial in origin. Glacial drift covers most of the bedrock, and depths range from 0 to about 450 ft. All areas where drift thickness is less than 50 ft are south of the Maumee River.

Five aquifer types are commonly used for water supply in the Maumee River basin. Surficial sand and gravel aquifers are found along most of the major streams in the basin. Both buried and discontinuous sand and gravel aquifers are present in the Maumee River basin. Buried aquifers are used north of the Maumee River whereas discontinuous aquifers may be used south of the Maumee River. Adequate domestic supplies are available from both types of aquifers, although large yields are available only from buried aquifers. Large yields are available from sand and gravel aquifers in the St. Marys Bend Segment of the Lafayette Bedrock Valley in the southern tip of the Maumee River basin. The carbonate bedrock aquifer underlies the southern half of the basin, and is a productive source of ground water.

Ground water is readily available throughout the basin. Large yields of ground water are available from the unconsolidated aquifers in northern Allen County and southern Dekalb County. Yields of 200 gal/min or more are available from unconsolidated aquifers north of the Maumee River. Yields of 35 gal/min or more are available from the carbonate bedrock aquifer south of the Maumee River.

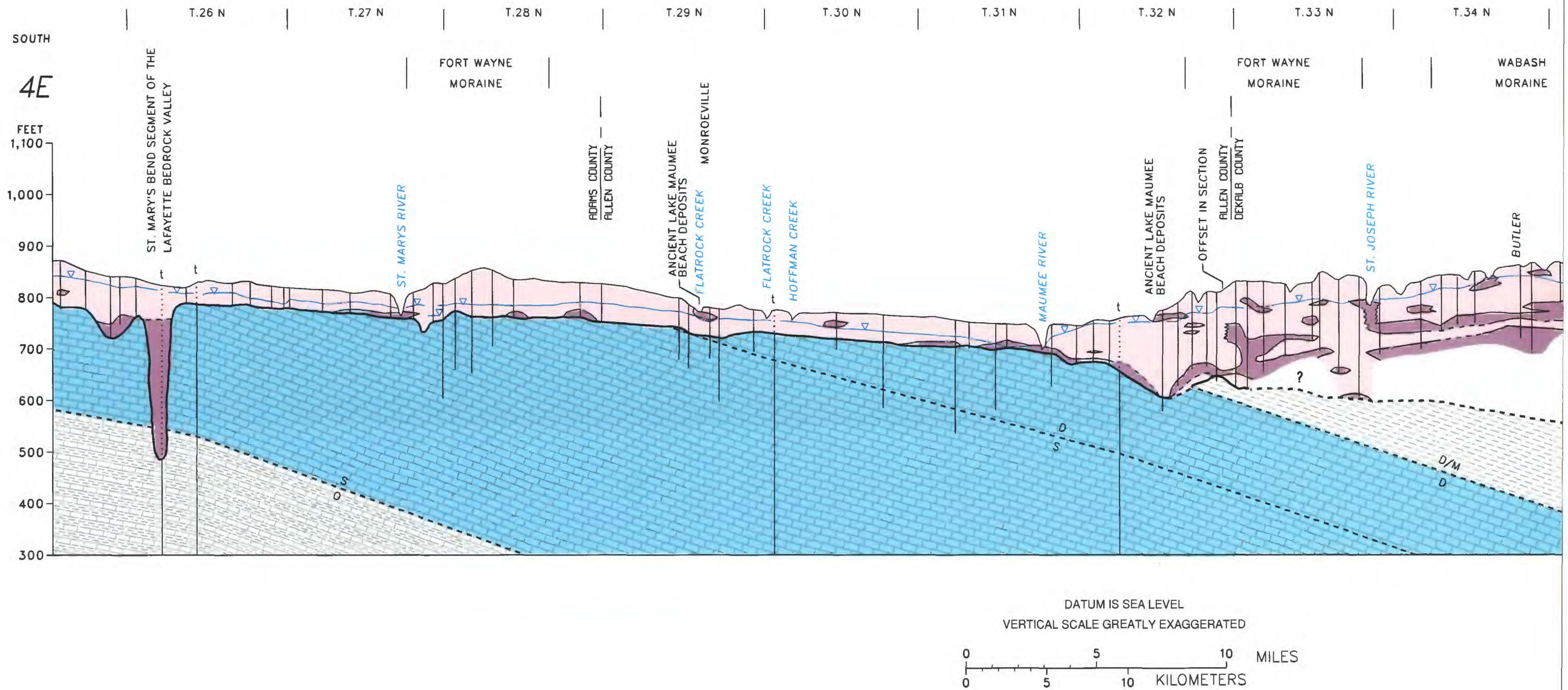


Figure 34. Hydrogeologic sections *4A–4A′* to *4E–4E′* of the Maumee River basin—Continued.

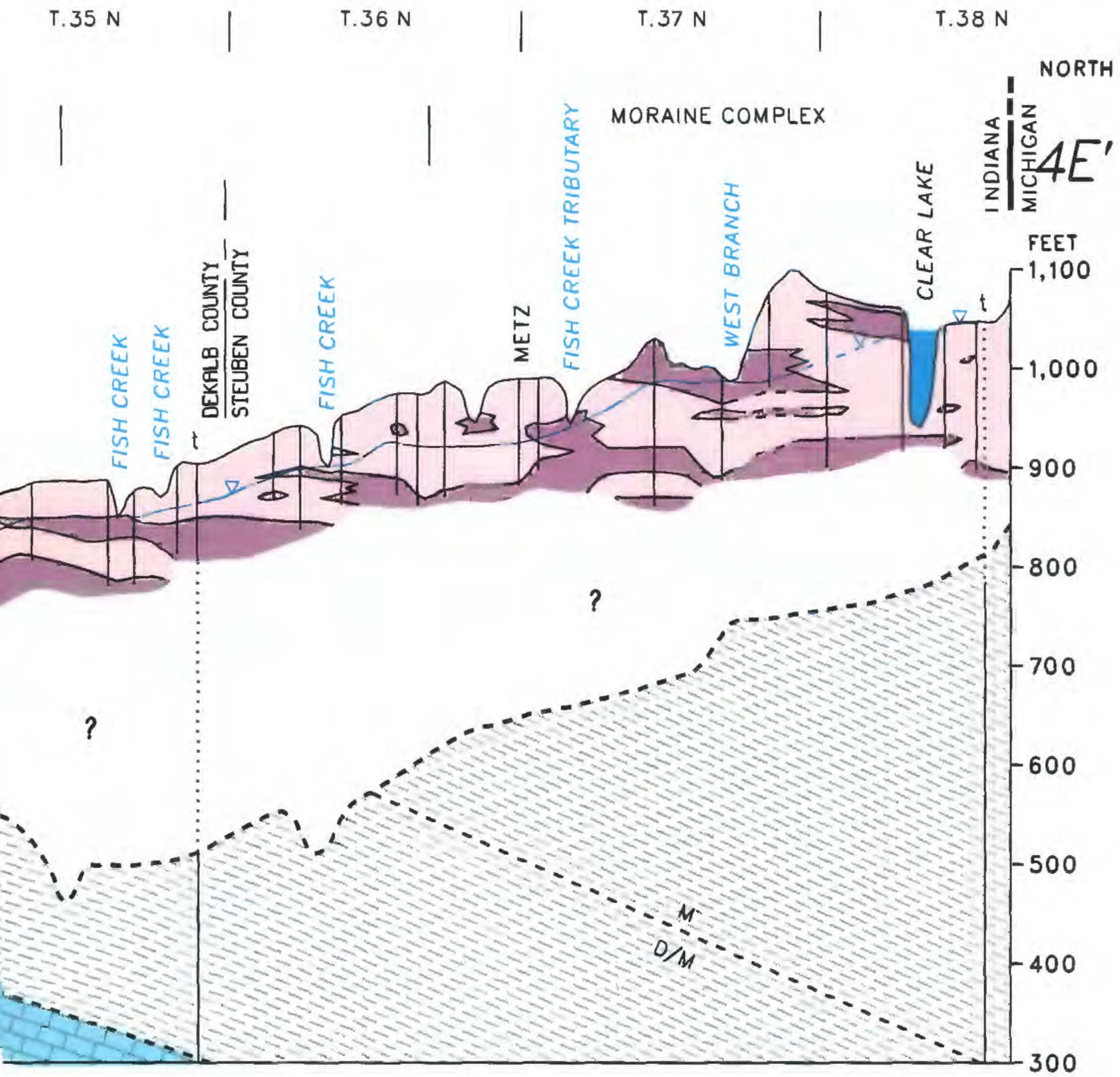


Table 6. Characteristics of aquifer types in the Maumee River basin

[≥ , greater than or equal; locations of aquifer types shown in fig. 35]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and gravel	0- 40	No data	
Buried sand and gravel	0- 150	20- 2,250	Intertill sand and gravel
Discontinuous sand and gravel	0- 80	≥20	
Sand and gravel in buried bedrock valley	0- 50	No data	Lafayette Bedrock Valley aquifer[1]
Carbonate bedrock	0- 700	35- 500	Silurian-Devonian carbonate aquifer

[1]Bleuer and others, 1991.

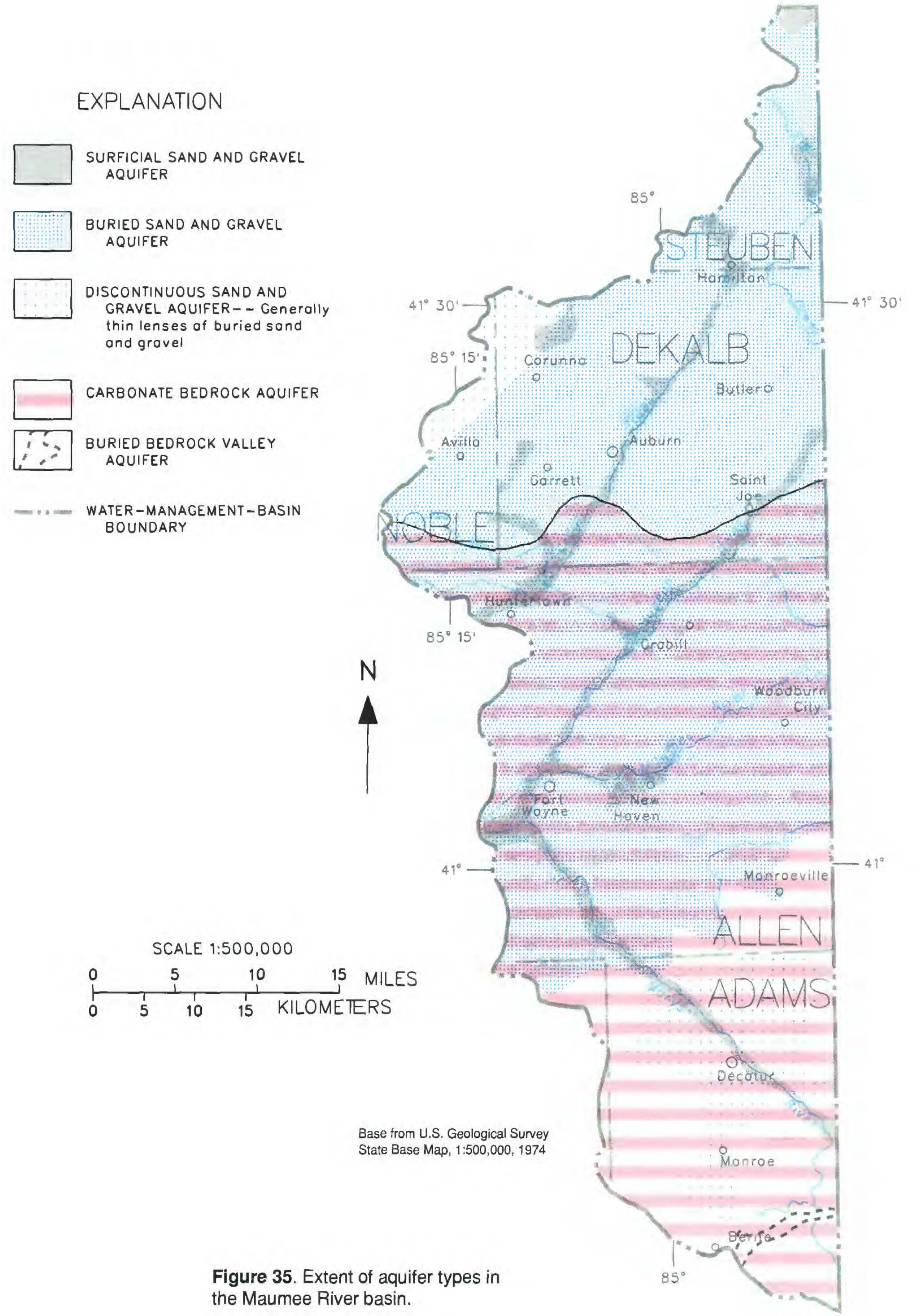


Figure 35. Extent of aquifer types in the Maumee River basin.

References Cited

Arvin, D.V., 1989, Statistical summary of streamflow data for Indiana: U.S. Geological Survey Open-File Report 89-62, 964 p.

Bleuer, N.K., 1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley System (so-called Teays Valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

_____1991, The Lafayette Bedrock Valley System of Indiana; concept, form, and fill stratigraphy, *in* Melhorn, W.N. and Kempton, J.P., eds., Geology and hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 51-77.

Bleuer, N.K., Melhorn, W.N., Steen, W.J. and Bruns, T.M., 1991, Aquifer systems of the buried Marion-Mahomet trunk valley (Lafayette Bedrock Valley System) of Indiana, *in* Melhorn, W.N., and Kempton, J.P., eds., Geology and hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 79-89.

Bleuer, N.K. and Moore, M.C., 1972, Glacial stratigraphy of the Fort Wayne area and the draining of glacial Lake Maumee, *in* Moulton, Benjamin, ed., Proceedings of the Indiana Academy of Science, v. 81, p. 105-209.

_____1978, Environmental geology of Allen County Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 13, 72 p.

Clark, G.D. ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Ind., Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Gooding, A.M., 1973, Characteristics of Late Wisconsin tills in eastern Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 49, 28 p.

Gray, H.H., 1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

Gray, H.H., 1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geology map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Greeman, T.K., 1991, Areal differences between groundwater and surface-water drainage basins in Indiana [abs.]: Midwest Ground Water Conference, 36th, Indianapolis, 1991, program with abstracts, p. 106-107.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Herring, W.C., 1969, Reconnaissance of the ground-water resources of the Maumee River basin, Indiana: Indiana Department of Natural Resources, Division of Water Report 17, 30 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Leverett, Frank, 1897, The water resources of Indiana and Ohio, *in* U.S. Geological Survey: Eighteenth Annual Report, 1896-97, Part IV, Hydrography: p. 419-560.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 21, part II, p. 59-256.

Ohio Department of Natural Resources, 1985, Principal streams and their drainage areas: Ohio Division of Natural Resources, Division of Water.

Palmer, A.R., 1983, Decade of North American geology—1983, geologic time scale: Geology, v. 11, p. 503-504.

Pettijohn, R.A. and Davis, L.G., 1973, Water resources of the Maumee River basin, northeastern Indiana: U.S. Geological Survey, Hydrologic Investigations Atlas HA-493, 3 sheets.

Planert, Michael, 1980, Ground-water availability near Fort Wayne, Allen County, Indiana: U.S. Geological Survey, Water-Resources Investigations 80-34, 54 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A. A., ed., Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Berger, A.M.; Gates, G.R.; Gray, H.H.; Hutchison, H.C.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; Smith, N.M.; Wayne, W.J.; and Wier, C.E., 1970, Compendium of rock-unit stratigraphy in Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 43, 229 p.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Simpson, J.N., 1988, Profile—Maumee River, tributary to Lake Erie: Indiana Department of Natural Resources, Division of Water, 3 sheets.

Stallmam, R.W. and Klaer, F.H., Jr., 1950, Ground-water resources of Noble County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 5, 103 p.

U.S. Army Corps of Engineers, 1989, Water levels of the great lakes—weekly data summary: 1 p.

Watkins, F.A., Jr., and Ward, P.E., 1962, Ground-water resources of Adams County, Indiana: Indiana Department Conservation, Division of Water Resources Bulletin 9, 67 p.

Wayne, W.J., 1963, Pleistocene formations of Indiana: Indiana Department of Conservation, Geological Survey Bulletin 25, 85 p.

Wright, H.E., Jr. and Frey, D.G., eds., 1965, The Quaternary of the United States: Princeton, N.J., Princeton University Press, 922 p.

UPPER WABASH RIVER BASIN

By Theodore K. Greeman

General Description

For management purposes, the Indiana Department of Natural Resources has divided the Wabash River basin into three subbasins: an upper basin, a middle basin, and a lower basin. The Upper Wabash River basin extends from the Indiana-Ohio State line downstream to include Wildcat Creek (fig. 36) near Lafayette, Tippecanoe County (fig. 1). This area is approximately 110 mi east-west by 70 mi north-south.

The Upper Wabash River basin is 6,918 mi^2 and includes all or most of Blackford, Carroll, Cass, Clinton, Fulton, Grant, Howard, Huntington, Jay, Miami, Pulaski, Wabash, White, Whitley, and Wells Counties, and parts of 13 other counties. Principal cities in the basin include Bluffton, Columbia City, Frankfort, Hartford City, Huntington, Kokomo, Logansport, Marion, Monticello, North Manchester, Peru, Portland, Rochester, Wabash, and Warsaw (fig. 36).

Previous Studies

Numerous reports have been written on the hydrogeology of areas within the Upper Wabash River basin. One of the first was by Capps (1910), who described ground-water sources and artesian-well areas in the central part of this basin. Harrell (1935), in his comprehensive hydrologic report on Indiana, described the general features, geology, drainage, ground-water resources and sources of water for each county in Indiana. Although some information in Harrell's report is dated, the report is useful as a historical documentation of early water use.

In the 1950's and early 1960's, the Indiana Department of Conservation, Division of Water Resources (now called Department of Natural Resources, Division of Water) published a partial series of county ground-water-resource reports. Nine of these reports, discussed below, describe areas within the Upper Wabash River basin. Of the counties described in these reports, only Fulton County is wholly within the Upper Wabash River basin.

Stallman and Klaer (1950) described ground-water availability in Noble County. Lithologic logs of numerous wells are included, as well as a potentiometric-surface map. Rosenshein and Cosner (1956) and Rosenshein (1958) reported on the ground-water resources of Tippecanoe County. The 1956 report presents well-log data and water-level measurements, whereas the 1958 report describes the glacial and bedrock geology, bedrock topography, hydrologic cycle, ground-water quality and ground-water use. Rosenshein's 1958 report also includes a potentiometric-surface map and nine geologic-sections; it identifies five confined aquifers near Lafayette. Watkins and Ward (1962) reported on the ground-water resources of Adams County. Their report includes ground-water-quality information and lithologic logs of numerous wells.

Four more county reports were published in 1964. These reports, authored by Rosenshein and Hunn (1964a, 1964b, 1964c, 1964d), address the four adjoining counties of Marshall, Fulton, Starke, and Pulaski . All four reports describe the general geology, general sources of ground water, and general water-quality information. Tables include lithologic logs of numerous wells and data on ground-water quality. The last in this group of nine county reports describing parts of the Upper Wabash River basin is Hoggatt and others (1968). This report disscusses the quantity, distribution, and quality of the water resources in Delaware County.

Watkins and Rosenshein (1963) were the first to report the hydrologic properties of the bedrock in the Upper Wabash River basin. They described the transmissivity, storage and recharge rates for the Silurian bedrock near Bunker Hill, Miami County. Although the values are site specific, they were derived from aquifer tests and are representative of the Silurian bedrock.

In 1971, the Wabash River Coordinating Committee released an eight-volume "plan for the conservation, management, development and proper preservation of the water and related land resources of the Wabash River Basin" (appendix B, p. ii). These reports describe the same management basins as are described in this report. The section about ground water by Nyman and Pettijohn (1971) contains a summary of nondomestic well data for the Upper Wabash River basin, specific-capacity and well-yield data for unconsolidated aquifer sources, and estimates of potential yields to large-diameter wells from different aquifer types.

Two Hydrologic Investigations Atlases published by the U.S. Geological Survey present information on the Upper Wabash River basin. The first report (Tate and others, 1973) contains information on the water resources in the eastern half of the basin. Map data on ground-water quality, surface-water quality, hydrologic balance, surface-water flow duration, surface-water stage and discharge, unconsolidated-aquifer transmissivity, and the potentiometric surface are presented. The study area includes all of the Wabash River drainage basin upstream from Logansport, east to the Indiana-Ohio State line (fig. 36).

The second Hydrologic Investigations Atlas (Marie and Davis, 1974) pertains to the western half of the Upper Wabash River basin. This map report presents much of the same type of information presented in the earlier report by Tate and others, but at a different map scale. The report by Marie and Davis pertains to the Wabash River basin downstream from Logansport to Lafayette, slightly beyond the western basin boundary used for this report (fig. 36).

Bleuer and Moore (1972) were the first to describe and correlate glacial stratigraphy in the Allen County area. Bleuer and Moore (1978) subsequently described the glacial geology, ground-water availability, ground-water quality, and the potential for deep-well waste disposal in Allen County. These two reports represent a continued effort to identify and map the stratigraphic units within the unconsolidated drift.

A recent trend in ground-water investigations is toward model studies. Planert (1980), in his report on the hydrogeology of parts of Allen, Noble, and Whitley Counties within the Upper Wabash River basin, modeled several ground-water sources and simulated ground-water withdrawal to evaluate streamflow losses and ground-water-level declines. Gillies (1981) examined the ground-water resource potential near Logansport, Cass County. Maps of the geology, a buried-valley aquifer, and the potentiometric surface are presented. A computer model was used to simulate the effects of pumping on ground-water levels and streamflow at two locations. Smith and others (1985) reported on the Wildcat Creek and Deer Creek drainage basins in Howard County and parts of adjacent counties. This report describes streamflow, hydrologic characteristics of aquifers and confining beds, aquifer-stream interaction, and ground-water and surface-water quality. The report also presents model simulations that indicate the effects of pumping on ground-water levels and streamflow.

Bleuer (1989) reported on the buried Lafayette Bedrock Valley System (formerly the Teays-Mahomet Valley System) that converges on Lafayette, Indiana. Bleuer (1991) interpreted the sequence of valley-segment formation and the sedimentary facies and stratigraphy of the valley fill in the buried-valley system. Bleuer and others (1991) described the availability of ground-water resources from various segments of the Lafayette Bedrock Valley System within the Upper Wabash River basin. They also described the geology of several aquifers and confining beds.

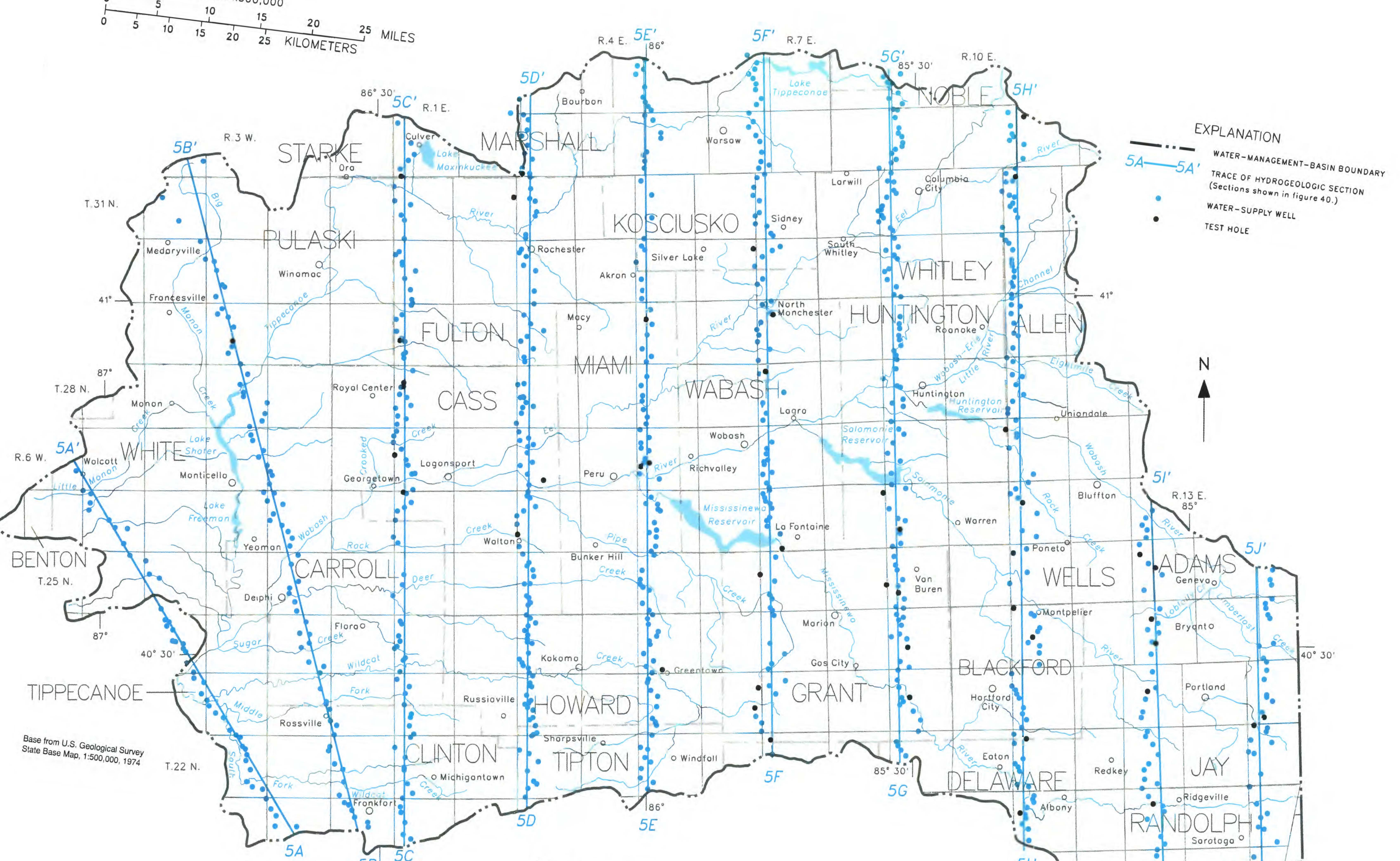


Figure 36. Location of section lines and wells plotted in the Upper Wabash

Physiography

The Upper Wabash River basin lies within two distinct physiographic areas (Schneider, 1966, p. 41): the Northern Moraine and Lake Region, located north of the Eel River, and the Tipton Till Plain, located predominately south of the Eel River. Except for occasional bedrock exposures and recent stream deposits, most physiographic features in the Upper Wabash River basin were formed by glaciers during Pleistocene time.

The Northern Moraine and Lake Region in the Upper Wabash River basin is characterized by moraines, outwash, and lake (lacustrine) plains. Depositional environments allow this northern area to be subdivided into two units: the Steuben Morainal Lake Area on the east and the Kankakee Outwash and Lacustrine Plain on the west (fig. 37).

The Steuben Morainal Lake Area is characterized by interlobate moraine topography. "Interlobate moraine" is a term coined by T.C. Chamberlin (1883) to describe the assemblage of moraines between two lobes of glacial ice. The moraine topography of the Steuben Morainal Lake Area is characterized by hummocky terrain and numerous kettle lakes that lack surficial drainage.

Glacial stratigraphy is very complex within the interlobate morainal deposits because slumping and ice thrusting obscured much of the original structure. Slumping occurred when entrapped ice melted and created surface depressions. Some depressions are filled with water, whereas others are drained. Hills composed of large blocks of reworked till, as well as ice-contact stratified sand and gravel (kames), are common. Local relief commonly ranges from 100 to 150 ft (Schneider, 1966, p. 52). Meltwater channels, outwash plains, and sand dunes also are common.

The Kankakee Outwash and Lacustrine Plain is characteristically flat and poorly drained. Sand, deposited as outwash by glacial meltwaters, lies at or near the surface throughout much of the area. Most deposits of sand are in valley trains, outwash plains, and lake-sand deposits associated with the Tippecanoe River. Prevailing westerly winds have rearranged the sand into dunes in White and Pulaski Counties.

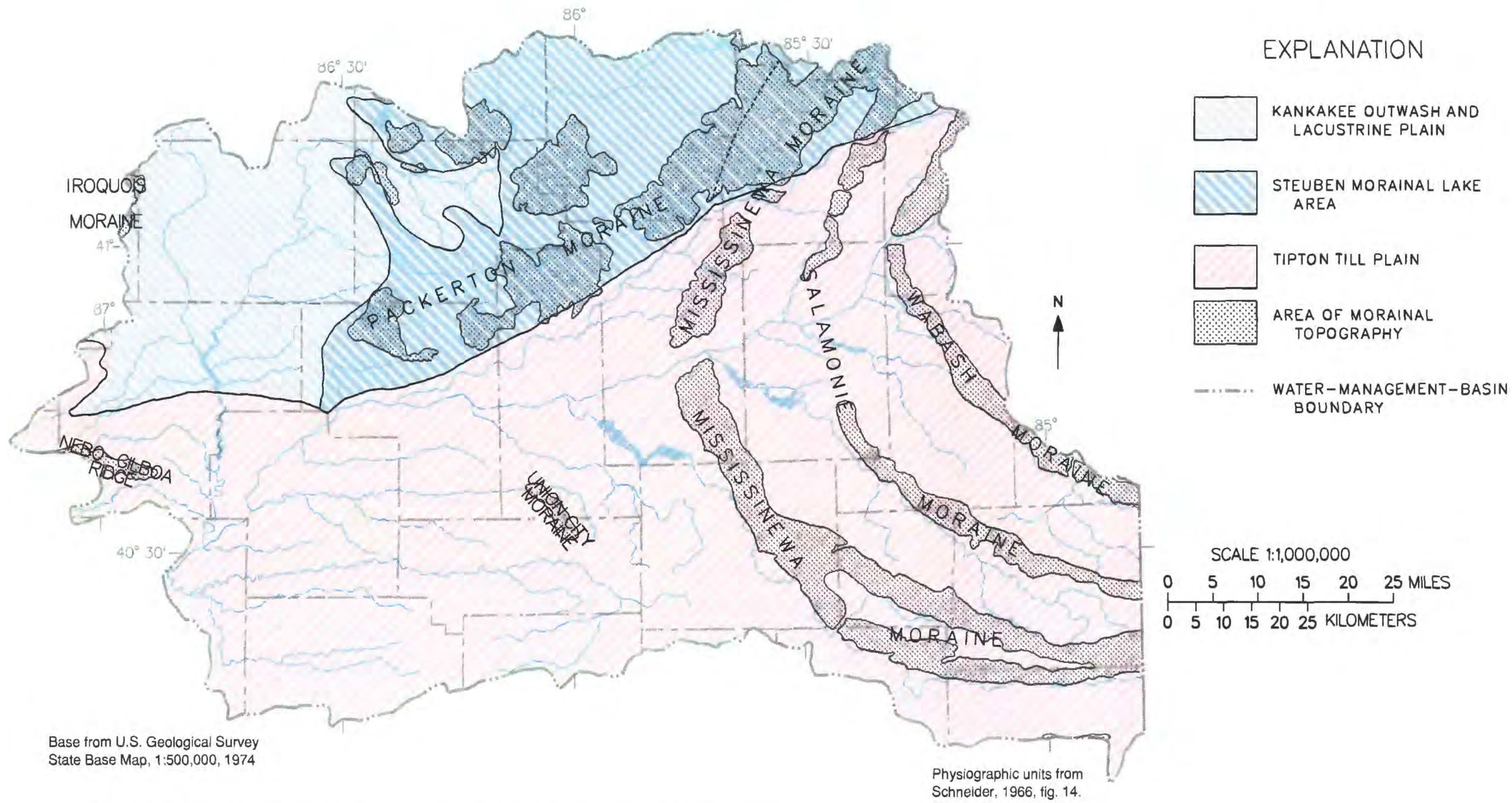


Figure 37. Physiographic units and moraines in the Upper Wabash River basin.

South of the Eel River is the Tipton Till Plain. The till plain surface is nearly flat to gently undulating, poorly drained, and featureless. Surface relief is generally less than 10 ft per 1,000 ft. The till plain is underlain by ground moraine and ablation tills. Resurgent periods during the retreat of the last glacial ice produced large concentric recessional moraine ridges (fig. 37). Relief across the moraines is low, generally less than 50 ft, although relief is slightly greater in several areas. The recessional moraines are 1 to 6 mi wide. Stratigraphy in the till plain area tends to be more horizontally continuous and less complex than in morainal deposits, except where thin till covers morainal deposits.

The Wabash Moraine (fig. 37), which forms the northeastern boundary of the Upper Wabash River basin, also forms the major divide between drainage to the St. Lawrence River and the Mississippi River. Lake Maumee, the high stage glacial-ancestor of Lake Erie, once overflowed the Wabash Moraine into the Wabash River basin. Shifting of the divide came about when the glacial ice retreated from the Lake Erie Basin, initiating drainage to the St. Lawrence River. Lake Maumee stopped draining across the Wabash Moraine about 12,000 years ago (Fullerton, 1980, pl. 1).

Surface-Water Hydrology

The Wabash River drains 32,910 mi^2 of Indiana, Illinois, and Ohio. Of the 23,921 mi^2 drained by the Wabash River within Indiana, 11,710 mi^2 are included in the three Wabash River management subbasins. Tributaries that are covered separately in this report drain the other 12,211 mi^2. The upper part of the Wabash River drains a total of 7,203 mi^2 of which 6,918 are in the Upper Wabash River management basin and 285 are in Ohio (Hoggatt, 1975, p. 2, 148, and 202).

The Wabash River begins in Mercer County, Ohio. From Mercer County, the Wabash River flows into Jay County, Ind. (fig. 36). Upstream from the Indiana-Ohio State line, the Wabash River drains 251 mi^2. Downstream from the State line, another 34 mi^2 of Ohio drainage eventually reaches the Wabash River by way of Limberlost Creek and the Mississinewa River (fig. 36).

This eastern part of the Wabash River basin is shaped like a shallow bowl as a result of continental glaciation. The glacially scoured landscape slopes northward from southern Randolph County to the Maumee River (sections 4E–4E′, fig. 34 and 5J–5J′, fig. 40). Moraines function as drainage-collection barriers on the slopes of the bowl. The moraines prevent tributaries draining the slope from reaching the center of the bowl. In the eastern part of the basin, the Mississinewa, Salamonie and Wabash Rivers drain these collection barriers, whereas the Loblolly-Limberlost (Adams and Jay Counties), Eightmile, and Rock Creeks (Wells County), plus many other small creeks, drain the slope. The streams follow the southern edge of the moraines in the southeast part of the basin, as shown in figure 37. Loblolly Creek is partially located in a surface trough that overlies the deeply buried Lafayette Bedrock Valley.

Upstream from the Indiana-Ohio State line, the Wabash River flows along the southern edge of the St. Johns Moraine (Ohio name for Salamonie Moraine of Indiana). At Fort Recovery, Ohio, 1 mile east of the State line, the Wabash River has cut across the St. Johns moraine and relocated against the Wabash Moraine. At the State line, the Wabash River is at an elevation of 835 ft above sea level.

Downstream from Bluffton, the Wabash River diverges from the Wabash Moraine and continues flowing northwest toward the Little River (Wabash-Erie Channel). The Wabash River flows on or near bedrock in this part of the basin. Huntington Reservoir, a flood control structure, regulates flow in the Wabash River upstream from the confluence of the Little River. The elevation of the Wabash River is 700 ft above sea level just downstream from Huntington Reservoir.

At Huntington, the Wabash River changes orientation, flowing west-southwest, and follows the Little River. The Little River (Wabash-Erie Channel) was the principal outflow channel for glacial Lake Maumee. Bedrock is exposed in the downstream end of the Little River channel. Downstream from Huntington, the Wabash River flows on bedrock. Terraces flanking the main channel are composed of stream-deposited sand and gravel.

Downstream from Huntington, the Salamonie River (fig. 36) is the next major tributary. The Salamonie River joins the Wabash River from the south bank at Lagro. A flood-control reservoir was constructed on the Salamonie River just upstream of the mouth. Downstream from Lagro to Richvalley (fig 36), the Wabash River flows on bedrock, commonly without terraces. Valley width in this stretch is slightly narrower than upstream from Lagro. This constriction may be because of the Mississinewa Moraine, through which the Wabash River cuts. At Richvalley, the Wabash River valley widens as it crosses the preglacial Lafayette Bedrock Valley.

Just downstream from Richvalley, the bottom of the Wabash River channel is at 635 ft above sea level, whereas the base of the Lafayette Bedrock Valley is at about 460 ft. At this location (section 5E–5E′, fig. 40), the buried drainage channel crosses under the Wabash River.

The Mississinewa River is the next principal tributary to the Wabash River, entering from the south bank (fig. 36). The city of Peru is 2 mi downstream from the mouth of the Mississinewa River. From Peru to Georgetown, the Wabash River flows west. Downstream from Peru, the bedrock surface rises, and the Wabash River again flows on rock with unconsolidated terraces. Pipe Creek is the next major tributary to join the Wabash River, also from the south bank.

At Logansport, the Eel River joins the Wabash River. The Eel River, which also began as a high-stage overflow drain for Lake Maumee, flows on unconsolidated materials along its entire course. The Eel River is the first major tributary entering the Wabash River from the north bank downstream of Huntington (fig. 36).

The Wabash River continues to flow on bedrock downstream from the mouth of the Eel River. At Logansport, terraces along the north side of the Wabash River are composed of unconsolidated material. Bedrock terraces, capped by as much as 60 ft of unconsolidated drift, form the south valley wall. At Georgetown (fig. 36), the orientation of the Wabash River changes again. Downstream of Georgetown, the Wabash River is oriented southwest.

The Wabash River flows on bedrock through most of Cass County and part of Carroll County. Just downstream from Georgetown, the Wabash River flows on unconsolidated deposits for several miles. At Delphi, the Wabash River exposes the top of a biohermal reef deposit, and the river temporarily flows on bedrock. Numerous bedrock exposures along the Wabash River are klintars, the exhumed tops of biohermal reef masses. Klintars form bedrock knobs that protrude above the surrounding land surface. In northeastern Tippecanoe County, the Wabash River flows near the bedrock surface with alluvial sand deposits in the channel.

In the area where Crooked Creek enters the Wabash River, the slope of the Wabash River channel decreases from greater than 2.5 ft/mi to less than 1.5 ft/mi. This change in slope is because of the large volumes of sediment carried into the Wabash River by the Eel and Tippecanoe Rivers. The Eel and Tippecanoe Rivers are the principal tributaries draining areas north of the Wabash River. Outwash-fan deposits adjacent to the Tippecanoe River (Tippecanoe Fan) have contributed large volumes of sediment to the Wabash River. This sediment loading begins choking the Wabash River channel more than 15 mi upstream of the Tippecanoe River.

The largest tributary to enter the Wabash River in the Upper Wabash River basin is the Tippecanoe River (fig. 36). Although the Tippecanoe River has no bedrock exposures in its channel, several tributaries, including Little Monon and Big Monon Creeks, flow on bedrock. Two reservoirs on the Tippecanoe River are located several miles upstream from the mouth. These reservoirs are for hydroelectric power and are not flood-control structures.

Continuing downstream, Wildcat Creek is the last tributary to enter the Wabash River in the Upper Wabash River basin. Wildcat Creek (fig. 36) enters the Wabash River from the southeast bank. From the Tippecanoe River downstream to Wildcat Creek, the Wabash River flows directly over the Battle Ground Lowland Section of the Lafayette Bedrock Valley (fig. 7), and the bedrock surface lies about 100 ft below the Wabash River in this area.

Geology

Bedrock deposits

Bedrock in the Upper Wabash River basin is composed of Paleozoic limestones, dolomites, sandstones and shales. Bedrock structure is dominated by the Cincinnati Arch (fig. 4), which plunges northwest across this basin. Along the axis of the Cincinnati Arch the rocks plunge from 4 to 13 ft/mi (0.04 to 0.14 degree) as indicated by a mapped Ordovician marker bed (Hasenmueller and Bassett, 1980c). The plunge of the axis is steepest in the northwestern part of the Upper Wabash River basin.

From the axis, bedrock dips both northeast into the Michigan Basin and southwest into the Illinois Basin (fig. 4). The dip of the bedrock away from the axis is slightly steeper than the plunge of the axis. Bedrock along the northeast-dipping flank has a maximum dip of 20 ft/mi (0.22 degree) in Kosciusko County, as indicated by a mapped Ordovician marker bed (Hasenmueller and Bassett, 1980c).

Most of the Upper Wabash River basin is on the northeast-dipping flank of the Cincinnati Arch. The Wabash River rises on the northeast flank of the arch and follows the plunge axis of the arch from the headwaters to Logansport. At Logansport, a structural bedrock sag in the crest of the arch, known as the Logansport Sag, allows the Wabash River to cross the axis of the Cincinnati Arch and drain toward the Illinois Basin (fig. 4).

Bedrock dips southwest in the part of the Upper Wabash River basin that includes Benton, Carroll, Clinton, Tippecanoe, and White Counties. The southwest-dipping flank of the Cincinnati Arch has a maximum dip of 17 ft/mi (0.18 degree) in southeast Tippecanoe County, as indicated by a mapped Ordovician marker bed (Hasenmueller and Bassett, 1980c). During part of the Paleozoic Era, the arch supported coral reef communities that are now deposits of carbonate rock. Throughout most of the Paleozoic Era, this anticline separated open seas to the northeast and southwest.

Although Paleozoic bedrock crops out at numerous locations in the Upper Wabash River basin, it is covered by drift in most places. Most bedrock exposures are near the Wabash River or its southern tributaries. The elevation of the buried bedrock surface ranges from just below 400 ft above sea level (Bruns and others, 1985a), at the base of a deep preglacial valley in Tippecanoe county, to about 1,050 ft above sea level (section 5J–5J′, fig. 40) in Randolph County near the Indiana-Ohio State line.

The age of bedrock exposed by preglacial erosion ranges from 315 to 440 million years (Palmer, 1983). Older Paleozoic rocks are present in the basin, but they were not exposed to post-Paleozoic erosion. A thick sequence of Cambrian sandstones, siltstones, shales, limestones, and dolomites overlies Precambrian igneous and metamorphic basement rocks. In the Upper Wabash River basin, the Cambrian rocks range from 2,000 to 3,500 ft in thickness. The Cambrian rocks are overlain by younger Ordovician sandstones, shales, and carbonate rocks. Ordovician rocks in the Upper Wabash River basin area are 1,000 to 1,400 ft thick.

The oldest rock to subcrop the drift in northeastern Indiana is the Maquoketa Group, an interbedded shale and limestone of Ordovician age. The subcrop of Upper Ordovician shales and limestones is present at the base of the Lafayette Bedrock Valley in the eastern one-third of the Upper Wabash River basin (fig. 38). The Geneva Segment of the Lafayette Bedrock Valley (fig. 7) is entrenched about 200 ft into the Ordovician rocks (section 5I–5I′, fig. 40).

Overlying the thinly bedded Upper Ordovician shale and limestone is a thick sequence of carbonate rocks, composed of limestones and dolomites. The carbonate rocks were deposited by marine organisms that lived in the seas during the Silurian and Devonian Periods. Silurian formations of the carbonate rock sequence are the Brassfield Limestone, Cataract Formation, Salamonie Dolomite, Pleasant Mills Formation, and the Wabash Formation. Overlying the Silurian formations is the Devonian Muscatatuck Group (fig. 38). Thick platform-reef deposits and associated debris are included within this carbonate rock sequence along the crest and flanks of the Cincinnati Arch.

The lithology of the Silurian-Devonian carbonate rock sequence is variable. Clayey carbonate rocks are more common than pure carbonate deposits. Dolomite has replaced limestone in some areas, and interbedded chert deposits are common in others. Bedded gypsum and anhydrite are found in the Devonian part of the carbonate rock sequence. The color of carbonate bedrock ranges from white to brown, and the texture ranges from micrite (a fine-grained crystalline precipitate) to bioclastic (a fossiliferous reef-framework detritus). Bedding ranges from thin to massive. Bedrock is nearly horizontal throughout most of this basin; however, steeply dipping beds (up to 45 degree dip) are observed locally in reef-flank deposits. Depositional environments ranged from deep-water marine to shallow-water marine and isolated tidal pools where evaporite deposits accumulated. Disconformities are present within Silurian deposits and at the Silurian-Devonian contact. These unconformities are nondepositional and erosional.

The Silurian and Devonian carbonate rock sequence attains a maximum thickness of 700 ft in northeastern Whitley County. Postdepositional erosion has thinned the carbonate rock sequence significantly in most of the basin. Because carbonate rocks are relatively resistant to weathering, they form the bedrock surface across the top of the Cincinnati Arch. Carbonate rocks are at the bedrock surface in about 75 percent of the Upper Wabash River basin (fig. 38). Some carbonate rocks are at the bedrock surface in every county within the Upper Wabash River basin.

After the deposition of carbonate rocks, a major change in sea level occurred. During the Devonian Period, sea level dropped relative to the arch, and the connection between the Michigan and Illinois Basins was interrupted. The Illinois and Michigan Basins both contain younger Devonian rocks, which display facies changes indicating shallowing seas and exposed land areas (Droste and Shaver, 1983, p. 24). The presence of sand indicates sandy coastal facies nearest to the arch edges. Kerogen-rich black shales were deposited in these anoxic basins. These carbonaceous shales are known as the New Albany Shale south of the Cincinnati Arch and as the Ellsworth and Antrim Shales north of the arch.

The New Albany Shale and equivalent shales were deposited during a 15 million year period (Palmer, 1983) extending from the Devonian Period into the early Mississippian Period. Although these fine-grained oil shales are as thick as 150 ft in the basin, the entire shale thickness remains in only a few subcrop locations in the Upper Wabash River basin. These formations have been thinned by erosion except where they are overlain by younger rocks.

Overlying the New Albany Shale is a thin limestone, known as the Rockford Limestone. Where present, the Rockford Limestone is less than 22 ft thick (Shaver and others, 1986, p. 124). In this basin, the top and bottom contacts of the Rockford Limestone are conformable and sharp. Both overlain and underlain by shales, this thin limestone is an excellent Lower Mississippian marker bed. This limestone has a uniform lithologic texture over a large area of Indiana and Illinois.

Where present, the Rockford Limestone is overlain by silts and shales of the Borden Group. The Borden Group shales are the youngest Mississippian subcrops in the Upper Wabash River basin. When deposited, the Borden Group probably was more than 500 ft thick in this area. Now, because of postdepositional erosion, the Borden Group is generally less than 50 ft thick in Benton, Clinton, Tippecanoe, and White Counties. In the Upper Wabash River basin, the Borden Group subcrop area covers less than 100 mi^2 and is buried under 15 ft or more of drift. In this basin, the New Providence Shale appears to be the only formation within the Borden Group that has not been eroded. The New Providence Shale is primarily a blue-gray shale. In this basin, drift unconformably overlies the Borden Group; the contact is characterized by moderate relief.

The youngest rocks in the Upper Wabash River basin were deposited during the early Pennsylvanian Period. In Benton and White Counties, a shaley sandstone of the Raccoon Creek Group (Shaver and others, 1986, p. 120 and 121) forms remnant caps overlying older rocks. The rocks are medium- to coarse-grained sandstones in alluvial-channel deposits containing carbonaceous shale lenses (S.J. Keller, Indiana Geological Survey, oral commun., 1990). These isolated subcrops of Pennsylvanian rock are generally less than 30 ft thick.

The Pennsylvanian rocks in the Upper Wabash River basin unconformably overlie Mississippian and Devonian rocks. After the deposition of the Mississippian rocks, a major erosional event removed most Mississippian rocks from this basin. At the northern end of hydrogeologic section 5A–5A′ (fig. 40), more than 1,300 ft of Upper Mississippian and Lower Pennsylvanian stratigraphic section is missing (Gray, 1979, p. K12). Hydrogeologic section 5A–5A′ (fig. 40) indicates erosion of the New Albany Shale in the Wolcott area prior to the deposition of the Raccoon Creek Group. The hiatus (period of missing record) between these rocks represents more than 35 million years. There are no younger Pennsylvanian bedrock units in this basin, although younger Paleozoic deposits have been found in Indiana and elsewhere.

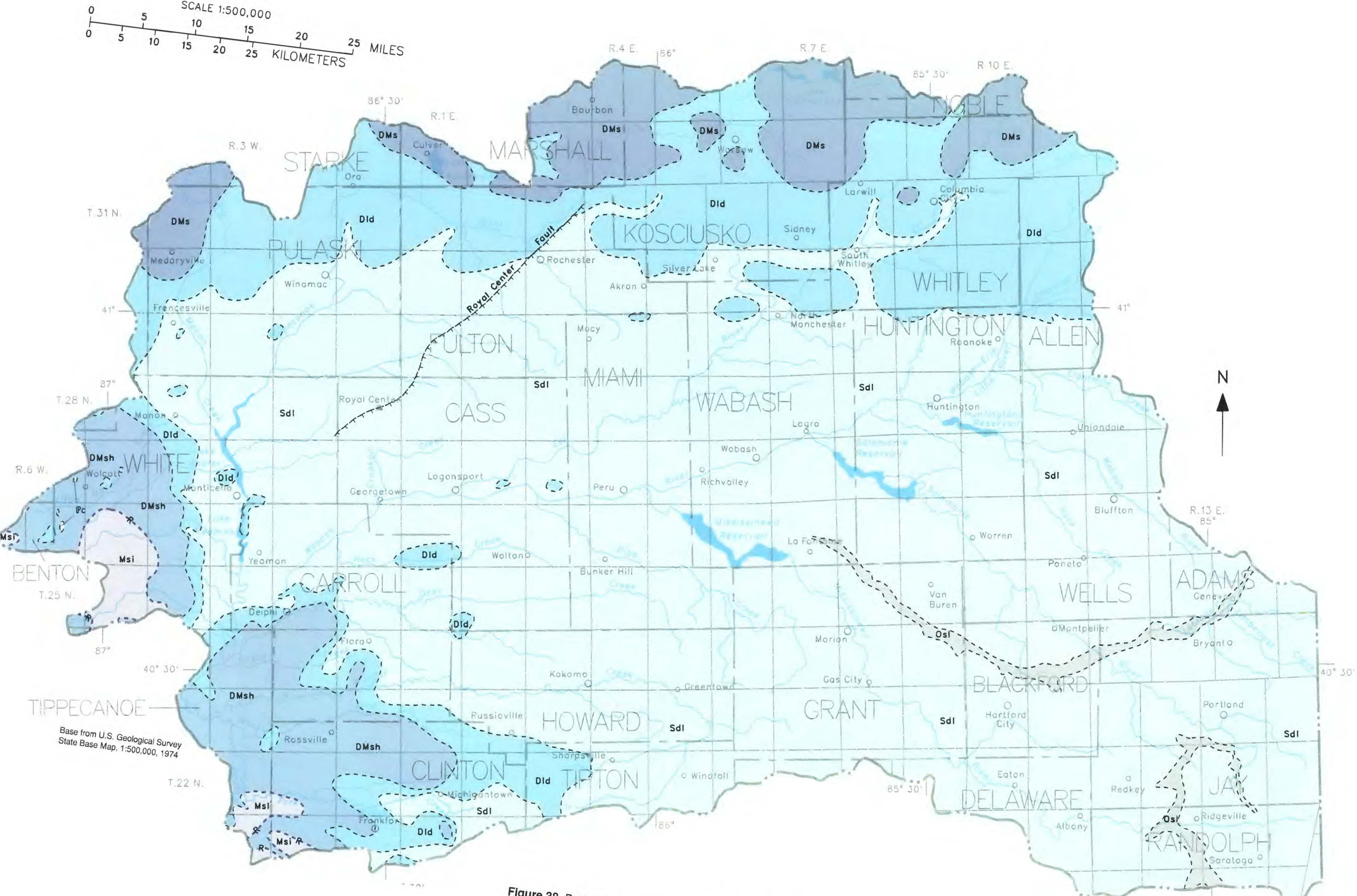


Figure 38. Bedrock geology of the Upper Wabash River ba

EXPLANATION

Pc — PENNSYLVANIAN COMPLEXLY INTERBEDDED SHALE AND SANDSTONE, WITH THIN BEDS OF LIMESTONE AND COAL-- Composed of the Racoon Creek Group

Msi — MISSISSIPPIAN SILTSTONE AND SHALE WITH MINOR SANDSTONE AND DISCONTINUOUS LIMESTONE-- Composed of the Borden Group

--R-- — ROCKFORD LIMESTONE

DMsh — DEVONIAN AND MISSISSIPPIAN SHALE-- Composed of the New Albany Shale

DMs — DEVONIAN AND MISSISSIPPIAN SHALE-- Composed of the Antrim and Ellsworth Shales

Dld — DEVONIAN LIMESTONE AND DOLOMITE-- Composed of the Muscatatuck Group

Sdl — SILURIAN DOLOMITE AND LIMESTONE-- Composed of the Wabash and Pleasant Mills Formations, and the Salamonie Dolomite, Cataract Formation, and Brassfield Limestone

Osl — ORDOVICIAN SHALE AND LIMESTONE-- Composed of the Maquoketa Group

NORMAL FAULT-- Hachures on downthrown side. Dashed where approximately located

GEOLOGIC CONTACT-- Dashed where approximately located

WATER-MANAGEMENT-BASIN BOUNDARY

Unconsolidated deposits

Glacial ice invaded Indiana several times during the Pleistocene Epoch (1 million to 10,000 years ago). Moving from several spreading centers in Canada, glaciers followed physiographic lowlands and major valleys to reach Indiana (fig. 8). The Huron-Erie Lobe moved into northeastern Indiana from the Lake Erie and Lake Huron basins several times. The Huron-Erie Lobe followed a bedrock lowland developed on Devonian and Mississippian shales. Synchronously, glacial ice moved south out of the Lake Michigan basin several times. The Lake Michigan Lobe also followed a physiographic lowland developed on the same Devonian and Mississippian shale bedrock.

A third lobe of ice, the Saginaw Lobe, filled the area between the other two lobes. This lobe moved through the Saginaw Bay area of Michigan. Glaciers competed for space in the Upper Wabash River basin several times. The Saginaw Lobe moved southward until its path was blocked by the Lake Michigan Lobe on the west and the Huron-Erie Lobe on the south and east. With its westward path blocked, the Huron-Erie Lobe advanced southward onto the carbonate bedrock platform in central and eastern Indiana.

Glaciers have followed these glacial flow paths into Indiana several times. More is known about the most recent glacial advance, the Wisconsinan glacial stage, than about earlier advances. This is primarily because of the presence of the Wisconsinan drift at the land surface. Many drift deposits from earlier glacial periods have been scoured and incorporated into younger sediments. Others are buried under younger deposits.

During each advance, thick glacial ice covered the land. Glacial ice in the Upper Wabash River basin is estimated to have been several thousand feet thick (Harrison, 1958, p. 84). In addition to tremendous volumes of water, glaciers also transported huge volumes of sediment into Indiana. Glacially transported sediment choked drainages and aggraded channels.

When the first pre-Illinoian glacial lobe reached Indiana from the north and northeast and spread out onto the carbonate bedrock platform, it encroached upon a major valley system known as the Lafayette Bedrock Valley (fig. 7). This preglacial channel was entrenched more than 350 ft into the nearly horizontal Paleozoic sediments. Sediment from the advancing glacier blocked this valley in at least two places. In eastern Indiana, a gray loam till and other sediments filled the valley from "rock rim to rock rim," whereas a claystone-bearing red till partially filled the valley from east-central to west-central Indiana (Bleuer, 1989, p. 6). The red till has been dated at greater than 730,000 years old (Bleuer, 1989, p. 7) and is among the oldest tills in the valley. Currently, the entire Lafayette Bedrock Valley is filled with glacial sediments.

Within a physiographic area, the surficial unconsolidated deposits generally have the same depositional history. Because of this regional similarity, the unconsolidated geology will be discussed by physiographic regions (fig. 37). Exceptions to this general rule are common, however, especially in areas underlain by the Lafayette Bedrock Valley.

The first physiographic area, the Steuben Morainal Lake Area (fig. 37), is located north of the Eel River and east of Logansport, and it includes the Tippecanoe River basin upstream of Monterey. During the Late Wisconsinan, the Huron-Erie Lobe of glacial ice moved west-southwest into Indiana from the Lake Erie basin as the Saginaw Lobe moved south-southwest into Indiana from the Saginaw Bay area. When the two lobes met, they blocked each other and stagnated. After stagnation, the Saginaw Lobe was partially overrun by the Huron-Erie Lobe.

A thick deposit of glacial drift, herein called the "moraine complex," was built up by repeated glacial advances of the two lobes. The moraine complex is composed of unsorted clay, silt, sand, gravel, and boulders in irregular deposits; drift is more than 250 ft thick in this area (fig. 39). The moraine complex includes areas depicted as Packerton Moraine and Mississinewa Moraine in figure 37 and in hydrogeologic sections 5C–5C′ to 5H–5H′ (fig. 40).

As the Saginaw and Huron-Erie Lobes competed for space, they formed very complicated geologic deposits. Any original depositional structure within the drift of the moraine complex was obliterated by glacial thrusting. Within the moraine complex, blocks of till were shoved together by glacial ice. Sand and gravel were carried into available spaces between till blocks by meltwater. Numerous kettle lakes are present in this area; the depressions for these lakes resulted from the burial of glacial ice.

The second physiographic area exhibiting distinct unconsolidated deposits corresponds with the Kankakee Outwash and Lacustrine Plain (fig. 37). This area is northwest of the Packerton Moraine and north of the Wabash River. The Tippecanoe River currently drains this area, but this was not always the case.

During the retreat of the Saginaw Lobe, meltwater drained from Indiana by way of the Kankakee River drainage basin. This meltwater carried large volumes of sediment, which choked the streams draining the area. Surficial sand deposits, some more than 50 ft thick, were laid down. As the ice terminus shifted and meltwater volume varied, however, the streams draining this area changed routes several times.

Stream piracy has been active in the development of the Tippecanoe River and Yellow River drainage basins. (The Yellow River is not in the Upper Wabash River basin; it is a tributary of the Kankakee River.) Several glacial sluiceways connected the Yellow River with the Tippecanoe River basin. At Ora, on the Pulaski County-Starke County line (fig. 36), the basin divide between the Tippecanoe River and Yellow River is only 15 ft above the normal flow elevation of the Tippecanoe River. Topographic relief and potentiometric relief are very low in this part of the basin.

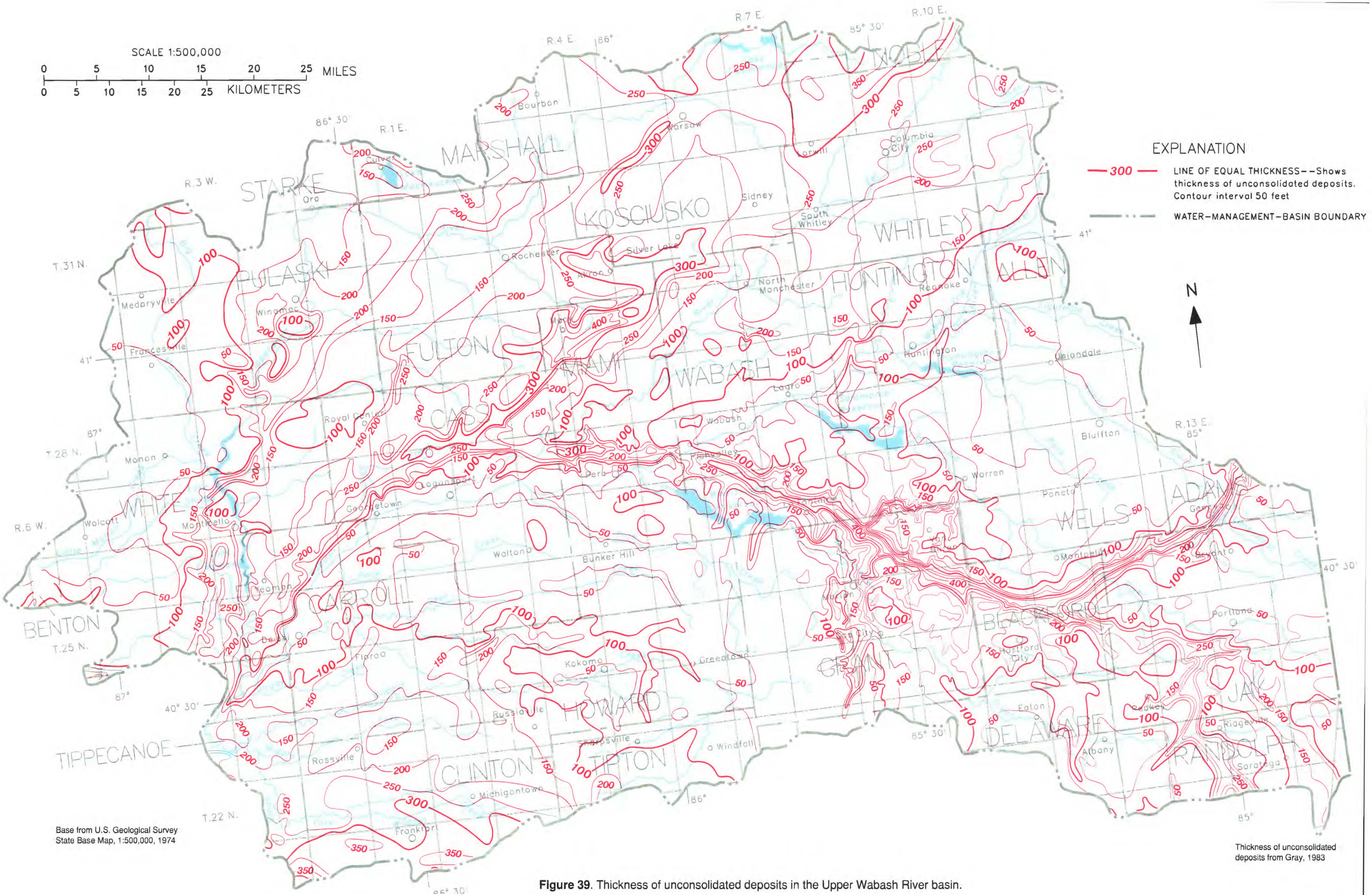


Figure 39. Thickness of unconsolidated deposits in the Upper Wabash River basin.

The third physiographic area of similar unconsolidated geology is south of the Eel River and east of Peru. This is the northeastern part of the Tipton Till Plain, which is covered with clay-loam till. These till deposits have been radiocarbon dated as more than 14,000 years old (Wayne, 1963, p. 44). In the Upper Wabash River basin, nearly all surface sediments south of the Eel River are of glacial origin and are considered to be part of the Lagro Formation of Late Wisconsinan age (Gray, 1989). The Mississinewa, Salamonie, and Wabash Moraines form arc-shaped ridges across this area (fig. 37).

South of the Eel River, till covers the broad Silurian carbonate bedrock platform. In the area between the Eel River and the Wabash-Little River Channel, drift thickness is greater than 150 ft (fig. 39). Lagro till composition ranges from 35 to 55 percent clay, 35 to 50 percent silt, and 10 to 30 percent sand by weight (Gooding, 1973, p. 8). Clay content in this till sheet is inversely proportional to sand content, with the combined weight of clay and sand equaling about 65 percent of the total weight.

The area south of Huntington and east of Peru was scraped clear of most pre-Wisconsinan drift, and the Lagro drift sheet is commonly less than 50 ft thick (fig. 39). Exceptions to the thin drift are areas underlain by the Lafayette Bedrock Valley (fig. 7). Drift greater than 400 ft thick is common within the buried preglacial valley (fig. 39). A basal sand and gravel overlies much of the bedrock south of the Wabash River (sections 5H–5H′ to 5J–5J′, fig. 40). This basal deposit is generally less than 10 ft thick and hydrologically connects the carbonate bedrock and the drift.

In the area of thin drift, erosion has removed the thin drift from the valleys of the large streams. The Wabash, Mississinewa and Salamonie Rivers flow on, or near, the bedrock. Exceptions are where the Wabash River crosses buried drainage channels. Two examples of this are (1) more than 300 ft of unconsolidated deposits underlie the Wabash River at Geneva and (2) more than 175 ft of unconsolidated deposits underlie the Wabash River just upstream from the mouth of the Mississinewa River (section 5E–5E′, fig. 40).

The fourth and final area of distinct unconsolidated deposits in the Upper Wabash River basin is south of the Wabash River and west of the Mississinewa River. This area corresponds to the part of the Tipton Till Plain that is covered by the Trafalgar Formation of Late Wisconsinan age (Gray, 1989). This area is underlain by buried moraines and thick deposits of drift. Drift thickness increases southward from the Wabash River and generally ranges from 0 to 200 ft in this area, although more than 350 ft of drift is found south of Frankfort, Clinton County (fig. 39). (See hydrogeologic section 5C–5C′, fig. 40.)

In this part of the Tipton Till Plain, the till is composed of an uncemented silty, sandy, calcareous till containing abundant pebbles and cobbles with scattered beds and lenses of silt, sand, and gravel (Shaver and others, 1970, p. 176-178). Drift originated from a northeastern (Huron-Erie Lobe) source. East of Pipe Creek, the Trafalgar Formation is covered by the younger Lagro Formation, both of Late Wisconsinan age (Wayne, 1963, p. 48).

Aquifer Types

Ten hydrogeologic sections in the Upper Wabash River basin (fig. 40) were constructed for this atlas. All are oriented perpendicular to the Wabash River. Eight of these hydrogeologic sections (5C–5C′ to 5J–5J′, fig. 40) are oriented south-north parallel to township boundaries and are 12 mi apart. The two westernmost hydrogeologic sections (5A–5A′ and 5B–5B′, fig. 40) are oriented south-southeast to north-northwest (fig. 36). Hydrogeologic sections 5A–5A′, 5B–5B′, and 5C–5C′ (fig. 40) are approximately 20 mi apart on their northern ends and 6 mi apart on their southern ends. The average density of logged wells plotted along the sections is 1.4 wells per mile. The 10 hydrogeologic sections (5A–5A′ to 5J–5J′) have a combined total length of 594 miles and a total of 833 plotted wells.

Additional bedrock-altitude information used in the hydrogeologic sections was from Bruns and others (1985a, 1985b, 1985c). Bedrock-stratigraphy information was supplemented by Bassett and Hasenmueller (1979, 1980) and Hasenmueller and Bassett (1980a, 1980b, 1980c). Other supplemental bedrock information was obtained from Bassett and Keith (1984) and Cazee (1988).

A map showing the extent of aquifers in the Upper Wabash River basin (fig. 41) was constructed by use of the hydrogeologic sections. Additional information for the aquifer map was obtained from the "Quaternary geologic map of Indiana" (Gray, 1989) as well as other studies referenced elsewhere in this section.

The aquifer map shows six types of aquifers used in the Upper Wabash River basin. Of these six aquifer types, four are restricted to the extensive unconsolidated deposits. These include surficial sand and gravel aquifers, buried sand and gravel aquifers, discontinuous sand and gravel aquifers in isolated deposits, and deep bedrock valleys, which may contain sand and gravel aquifers. A fifth aquifer type is carbonate bedrock of Silurian and Devonian age or of Early Mississippian age. The Silurian and Devonian carbonate rock is the primary aquifer used in about 50 percent of the basin. In another 25 to 35 percent of the basin, the carbonate bedrock aquifer is mapped where it is within 300 ft of the land surface but is not commonly used for water supply. The sixth aquifer type shown in figure 41, an upper weathered-bedrock aquifer in shale, is present in the extreme western part of the basin. Because the aquifer covers such a small part of the basin, it is not discussed in this section. Information on the upper weathered-bedrock aquifer is provided in the sections on the Kankakee River and Middle Wabash River basins in this report.

Characteristics of the five primary aquifer types in the Upper Wabash River basin are listed in table 7. Differences between the types of sand and gravel aquifers are not indicated on the hydrogeologic sections.

Unconsolidated Aquifers

Surficial Sand and Gravel Aquifers

Surficial sand and gravel has been deposited in three environments: (1) outwash plains and valley trains where sediment was deposited by glacial meltwaters; (2) alluvial environments where sediment is being deposited by present-day streams; and (3) sand dunes where sediment was deposited by the wind. Most of the surficial sand and gravel deposits in the Upper Wabash River basin are glacial in origin. Glacial meltwater carried sand and gravel away from the margin of active glaciers. Outwash-plain and valley-train deposits formed along glacial drainage routes. Large outwash deposits of unconsolidated sand and gravel are present along the Eel River, Tippecanoe River and along the Wabash River downstream from Delphi. Tremendous volumes of sediment were brought into the Wabash River by the Eel and Tippecanoe Rivers.

Outwash-plain deposits blanket more than 600 mi^2 of the Upper Wabash River basin with sand and gravel. Most surficial outwash-plain deposits are near the Tippecanoe River. These outwash-plain deposits form a blanket of sand from 20 to 80 ft thick. Outwash-plain deposits are mapped north of Lake Shafer at Buffalo (section 5B–5B′, fig. 40). Here, the surficial deposits are from 20 to 50 ft thick and as much as 720 ft above sea level. Surficial deposits of sand and gravel, as much as 80 ft thick, are mapped (section 5C–5C′, fig. 40) from the Tippecanoe River north to Lake Houghton and the basin boundary.

Beginning at Delphi, terraced valley-train sands and gravels are present along the Wabash River. These valley-train deposits form two terraces above the current flood plain. (See McBeth, 1902, p. 237.) The terraces formed at the time the glaciers were melting. The width of the outwash deposits range from 1.5 to 2.5 mi upstream from Logansport, and from 2 to 3 mi downstream from Delphi. Large flow volumes during glacial time scoured the shallow bedrock and prevented the accumulation of major unconsolidated deposits along the Wabash River upstream from Logansport.

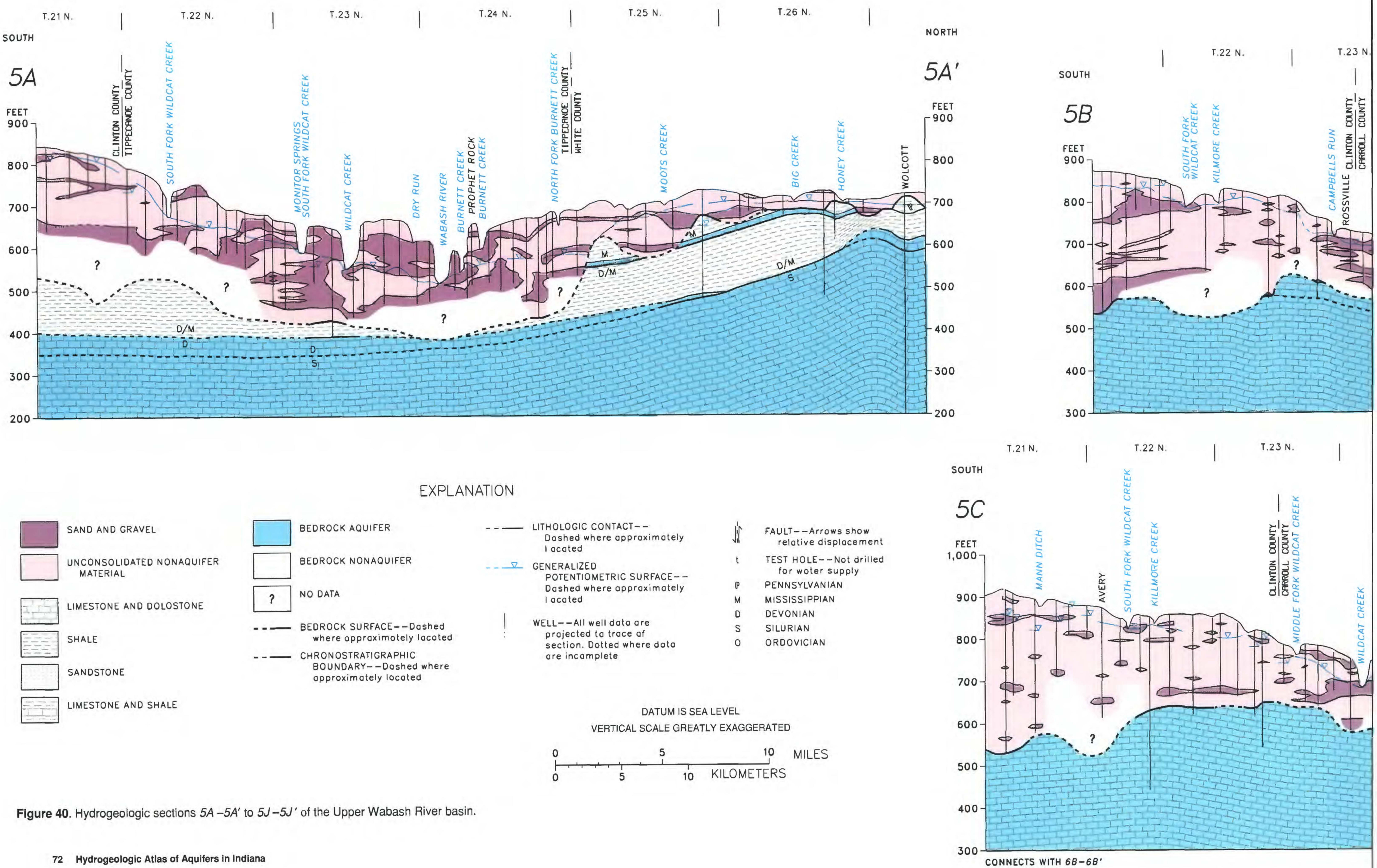
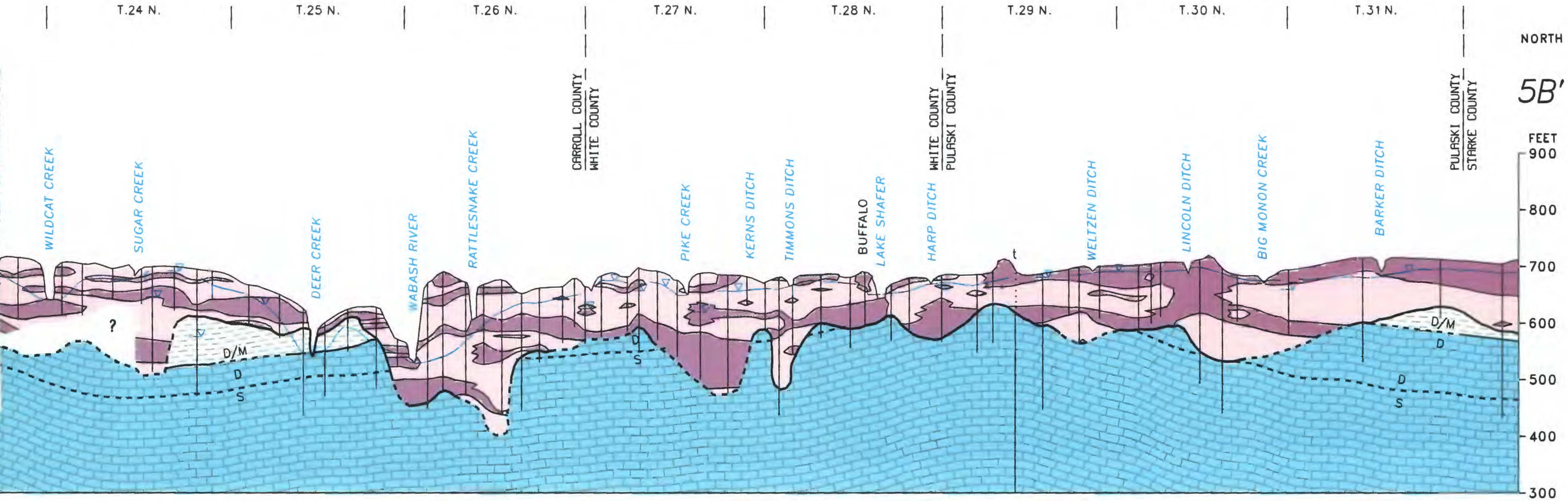
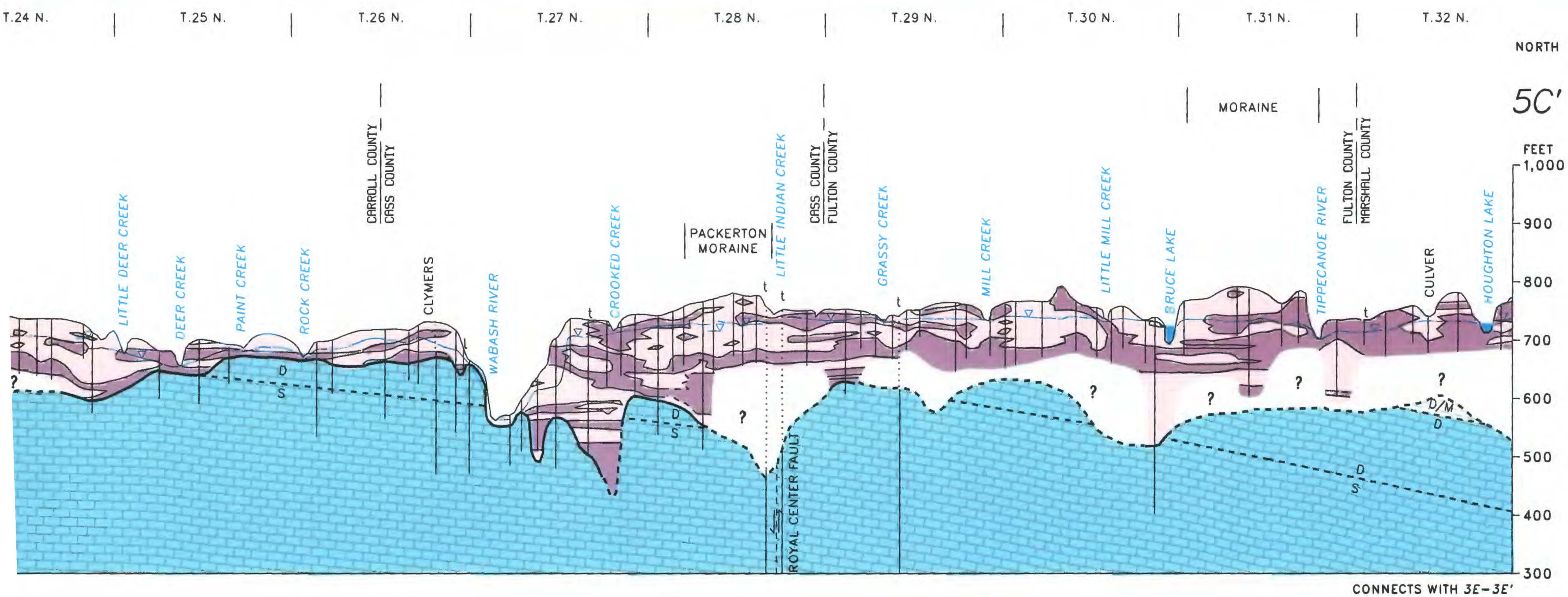


Figure 40. Hydrogeologic sections *5A–5A′* to *5J–5J′* of the Upper Wabash River basin.


T.24 N.
T.25 N.
T.26 N.
T.27 N.
T.28 N.
T.29 N.
T.30 N.
T.31 N.
NORTH
5B′
CARROLL COUNTY
WHITE COUNTY
WHITE COUNTY
PULASKI COUNTY
PULASKI COUNTY
STARKE COUNTY
FEET
900
800
700
600
500
400
300
WILDCAT CREEK
SUGAR CREEK
DEER CREEK
WABASH RIVER
RATTLESNAKE CREEK
PIKE CREEK
KERNS DITCH
TIMMONS DITCH
BUFFALO
LAKE SHAFER
HARP DITCH
WELTZEN DITCH
LINCOLN DITCH
BIG MONON CREEK
BARKER DITCH
D/M
D
S


T.24 N.
T.25 N.
T.26 N.
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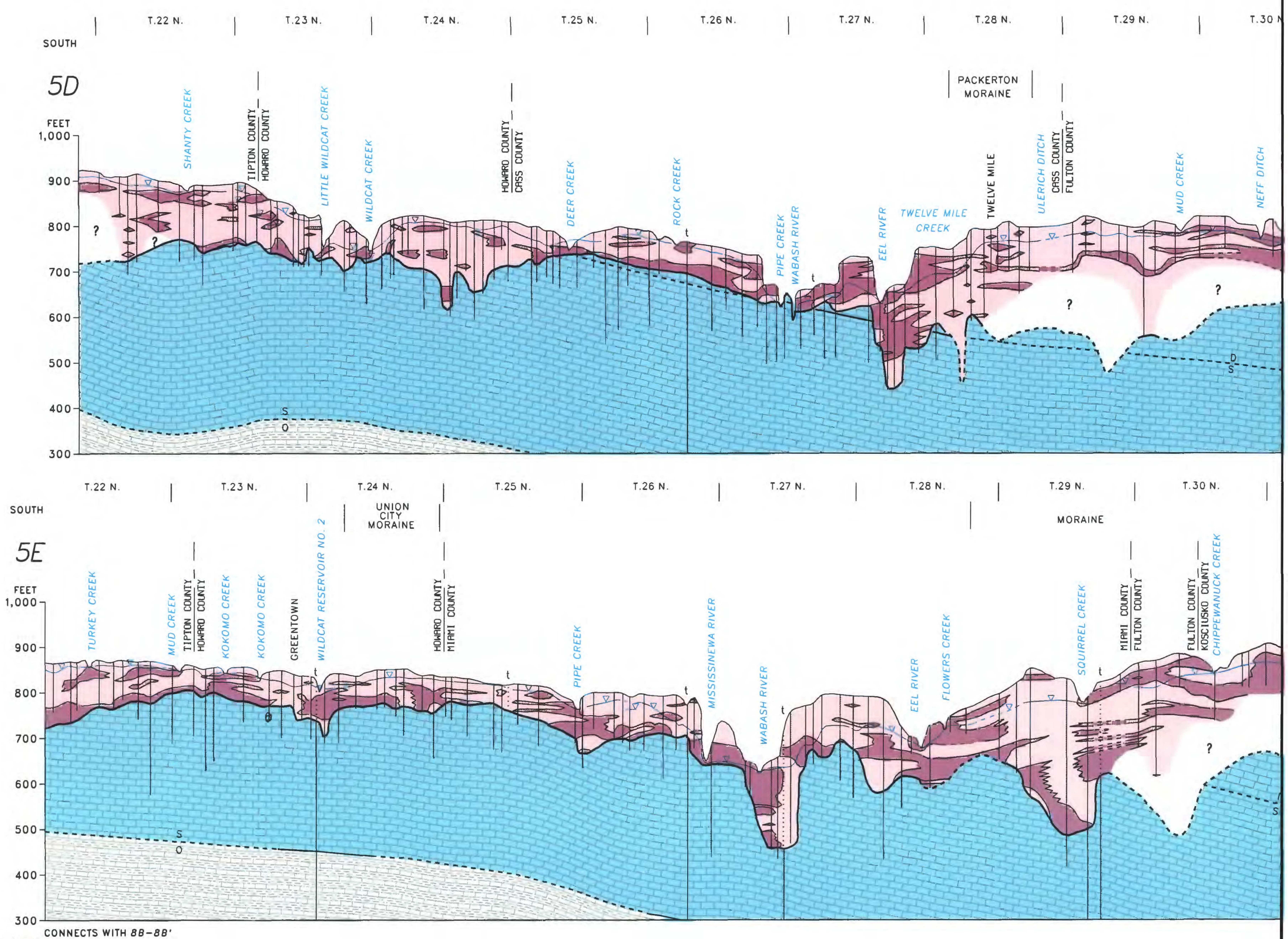
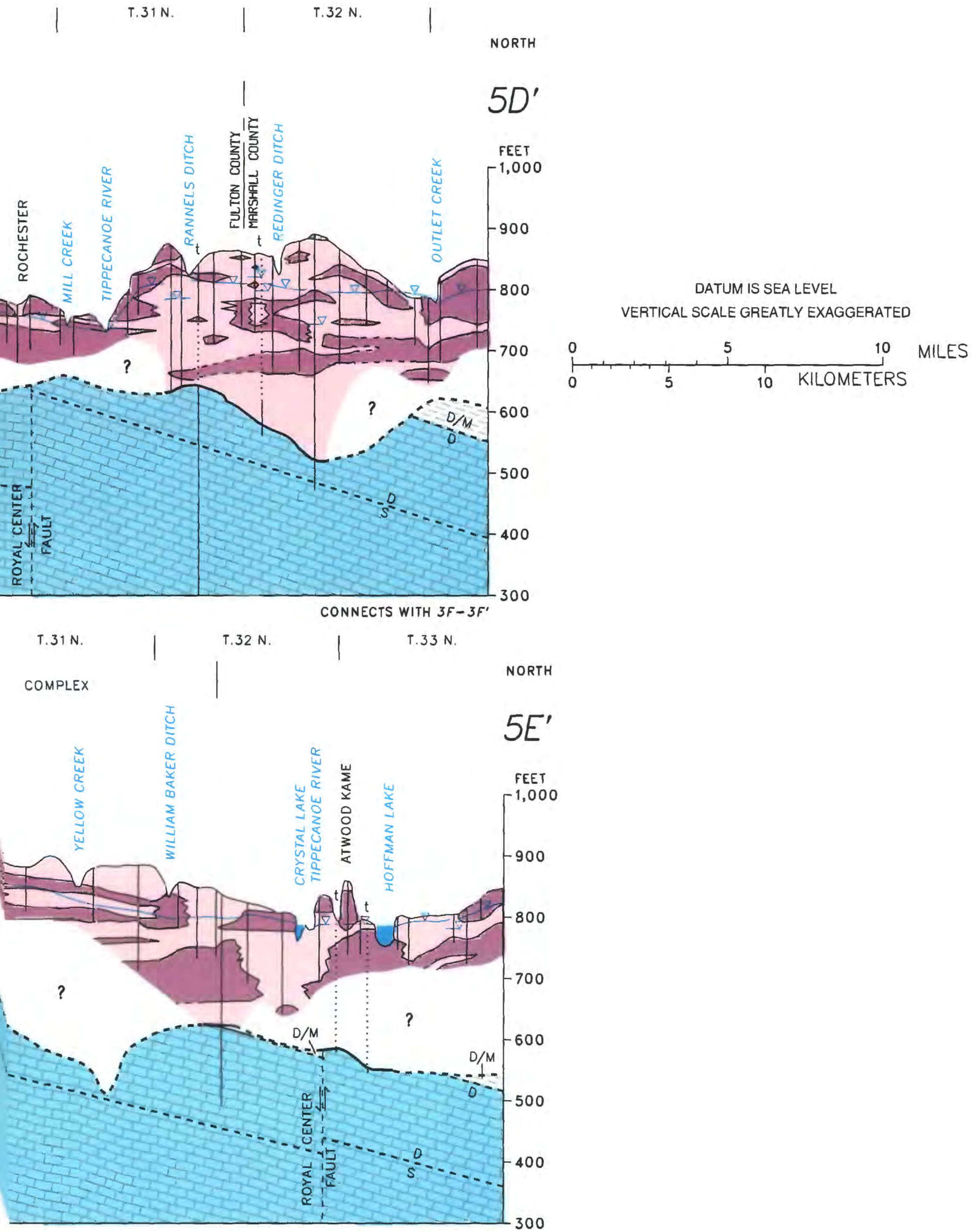
T.28 N.
T.29 N.
T.30 N.
T.31 N.
T.32 N.
NORTH
5C′
MORAINE
CARROLL COUNTY
CASS COUNTY
CASS COUNTY
FULTON COUNTY
FULTON COUNTY
MARSHALL COUNTY
PACKERTON MORAINE
FEET
1,000
900
800
700
600
500
400
300
LITTLE DEER CREEK
DEER CREEK
PAINT CREEK
ROCK CREEK
CLYMERS
WABASH RIVER
CROOKED CREEK
LITTLE INDIAN CREEK
GRASSY CREEK
MILL CREEK
LITTLE MILL CREEK
BRUCE LAKE
TIPPECANOE RIVER
CULVER
HOUGHTON LAKE
ROYAL CENTER FAULT
D/M
D
S
CONNECTS WITH 3E–3E′

Figure 40. Hydrogeologic sections *5A–5A'* to *5J–5J'* of the Upper Wabash River basin—Continued.





In Indiana, surficial sand and gravel deposits commonly contain large volumes of recoverable ground water, especially in outwash materials. Water levels in surficial aquifers are generally within 30 ft of the land surface, but are deeper adjacent to the Wabash River downstream from Delphi (section 5A–5A′, fig. 40). Terrace deposits in this area have deep, unconfined water tables. Because surficial aquifers are easily contaminated, some surface deposits are no longer suitable for potable ground-water production. In Kosciusko County, 30 of 83 private supply wells with nitrate contamination tap surficial, unconfined aquifers (State of Indiana, 1989, p. 9-11).

Alluvial deposits are another type of surficial aquifer that is common in the Upper Wabash River basin. Composed of reworked outwash sand and gravel and deposited adjacent to river channels, alluvial deposits commonly form a veneer on bedrock or drift. Deposits of alluvium are generally less than 30 ft thick and are covered by less than 10 ft of nonaquifer material.

The largest alluvial aquifers in the basin are found within the outwash deposits along the Eel River and along the Wabash River downstream from Delphi. Downstream from Delphi, the width of the alluvial deposits is from 0.5 to 1 mi along the Wabash River. Although bedrock reappears adjacent to the river in some areas, the alluvial deposits are continuous from Delphi downstream 330 mi to the mouth.

Sand dunes are another common type of surficial unconsolidated deposit in the Upper Wabash River basin. Wind has reworked both alluvial and outwash deposits into sand dunes. Two of the largest dunes are shown (T. 29 N. and 30 N.) in hydrogeologic section 5B–5B′ (fig. 40). These dunes stand 20 to 30 ft above the outwash plain. Dune deposits are very fine grained and may be above the water table, which makes dune-sand deposits a poor source of ground-water.

Buried Sand and Gravel Aquifers

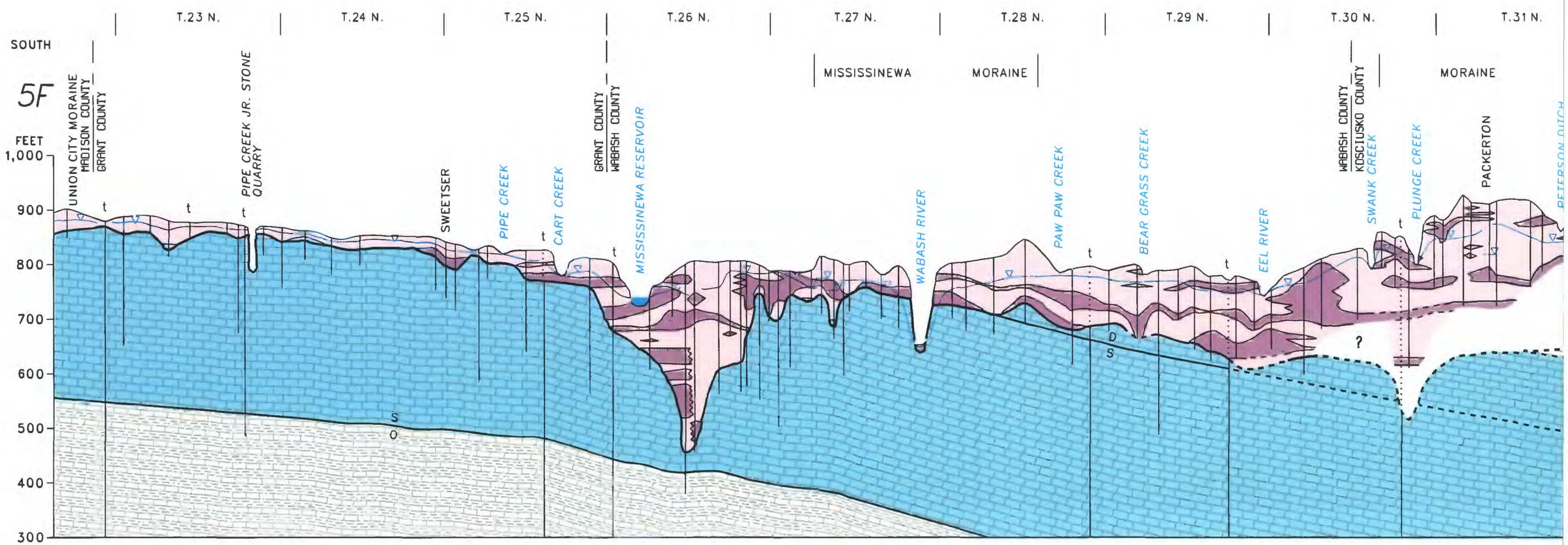
Most of the sand and gravel deposits in the Upper Wabash River basin are buried (fig. 41). Forming general horizons within the drift, buried sand and gravel aquifers formerly were isolated- to reticulating-fan, sub-ice channel, outwash-plain, and valley-train deposits. Buried aquifers are mapped as continuous although they were not deposited uniformly. Variations in the aggradational environment resulted in large variations of intertill aquifer thickness and distribution. Additional disruption of deposits resulted from glacial scour and shoving during burial. Most buried aquifers were originally surficial deposits, which have now been enclosed within the drift. Buried aquifers are covered by silty, clay-loam to loam tills and loess in the Upper Wabash River basin. The exact location and elevation of buried aquifers is unpredictable, as is their degree of interconnection.

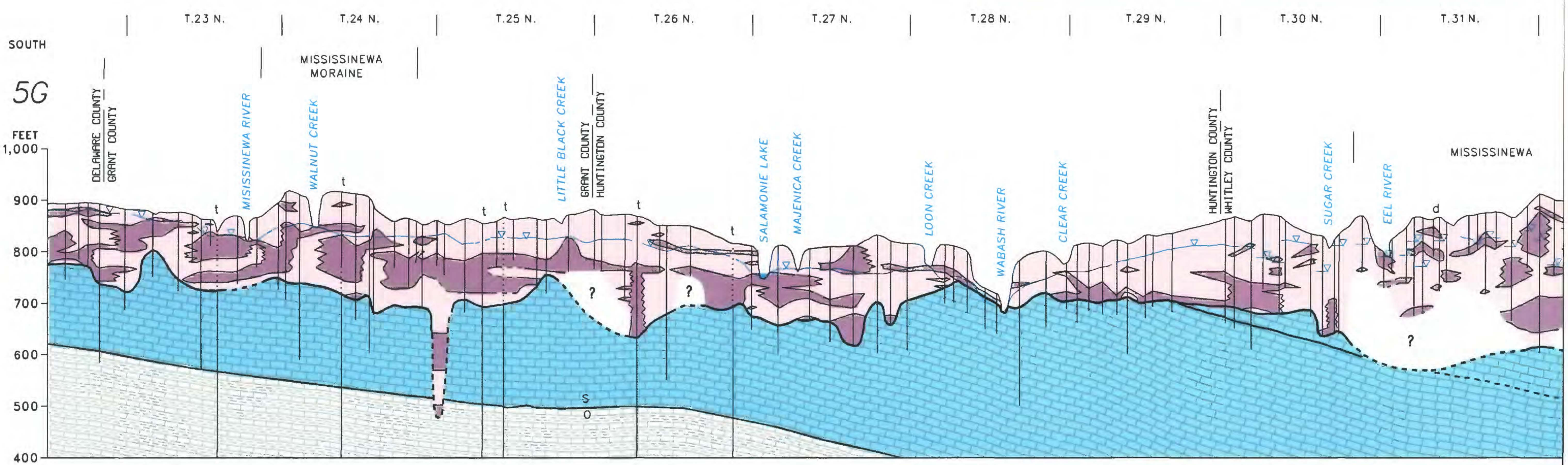
In the Upper Wabash River basin, numerous intertill sand and gravel deposits have been exposed by stream erosion. Examples are shown along the Wabash River (sections 5A–5A′ to 5F–5F′, fig. 40); the Salamonie River (sections 5G–5G′ to 5J–5J′, fig. 40); the Mississinewa River (sections 5E–5E′ to 5J–5J′, fig. 40); the Eel River (sections 5D–5D′ and 5E–5E′, fig. 40); the Tippecanoe River (sections 5C–5C′ to 5E–5E′, fig. 40); Pipe Creek (sections 5D–5D′ and 5E–5E′, fig. 40); and Wildcat Creek (sections 5A–5A′ to 5C–5C′, fig. 40).

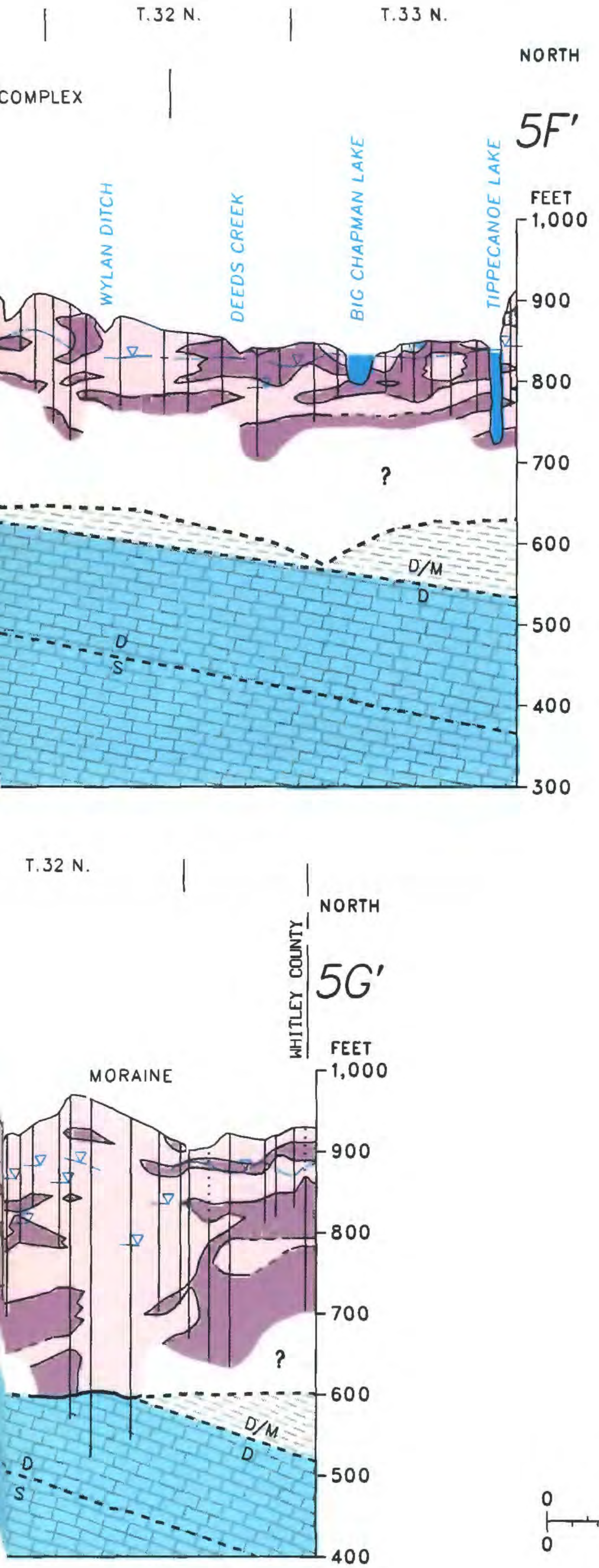
In addition to the buried sand and gravel deposits exposed along the major streams, numerous buried sand and gravel aquifers are exposed by tributary streams throughout the area. The hydrogeologic sections indicate the connection between surface drainage and the buried sand and gravel aquifers. Except during high stream stage, ground water commonly discharges to the streams.

South of the Wabash River and northwest of Kokomo, buried aquifers are common and fairly extensive. In many areas, a basal aquifer is present at the interface of bedrock and drift. North of the Wabash River in the Upper Wabash River basin most ground-water production comes from buried sand and gravel aquifers. Ground water in buried aquifers is generally confined. Large yields of water, ranging from 20 to 1,350 gal/min are available from buried aquifers in the Upper Wabash River basin. The largest yields available are in Kosciusko, Noble, and Whitley Counties (Clark, 1980, p. 33).

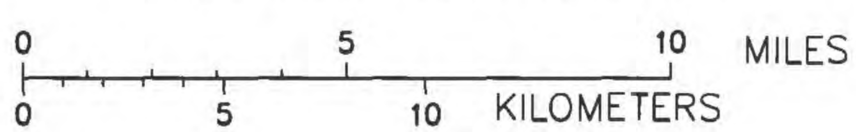
Figure 40. Hydrogeologic sections *5A–5A'* to *5J–5J'* of the Upper Wabash River basin—Continued.







DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED



Bleuer and Moore (1978, p. 43) report that, in a study for Allen County, water from the glacial aquifers has higher dissolved concentrations of bicarbonate, iron, and zinc than water from the carbonate bedrock aquifer. Water in wells sampled from glacial aquifers contained four times as much dissolved iron as water in wells sampled from the carbonate bedrock aquifer. None of these constituents present a health hazard in concentrations reported.

Numerous deposits of cemented sand and gravel have been found in the Upper Wabash River basin. Lithified Holocene deposits have formed from the chemical deposition of calcium carbonate and hydrous iron oxides dissolved in ground water. Deposition of the cementing agents occurs where confined aquifers are exposed to the atmosphere, as in aquifer outcrops along streams. One notable occurrence of a cemented sand and gravel deposit is Prophet's Rock, at the Tippecanoe Battleground site, Battle Ground, Ind. Prophet's Rock (T. 24 N.), shown on hydrogeologic section 5A–5A′ (fig. 40), is an outcrop of sand, gravel, and cobbles cemented by calcium carbonate.

Discontinuous Sand and Gravel Aquifers

Two main areas of discontinuous buried sand and gravel aquifers are shown on the aquifer map (fig. 41). Within these areas, aquifers are small, discontinuous deposits (lenses) of sand and gravel at scattered elevations. Braided streams and sub-ice channels formed sinuous sand and gravel deposits, which are not continuous over broad areas. Further disruption of deposits resulted from glacial scour and shoving. Discontinuous aquifers supply adequate water for domestic needs; however, the larger yields necessary for agricultural, industrial and municipal needs may be unavailable. Most wells penetrate several lenses of sand and gravel.

One area of discontinuous sand and gravel, located between the Tippecanoe and Eel Rivers in Kosciusko, Wabash, and Whitley Counties (fig. 41), extends northeast into the St. Joseph River basin. This area is within the Steuben Morainal Lake physiographic area and is mapped as morainal topography by Gray (1989). The area is labeled "moraine complex" on hydrogeologic sections 5F–5F′ (fig. 40) and as "Mississinewa Moraine" in Tps. 31 and 32 N. in hydrogeologic section 5G–5G′ (fig. 40).

The second area of discontinuous sand and gravel is in eastern Clinton and western Howard Counties, within the Tipton Till Plain (fig. 41). This area does not coincide with mapped surficial geologic features (Gray, 1989), although thick morainal deposits are present in the subsurface. The lack of aquifer continuity in this buried moraine area is shown in Tps. 22 and 23 N.of hydrogeologic section 5C–5C′ (fig. 40). Parts of another discontinuous aquifer deposit are present in Jasper, Pulaski, and Starke Counties (fig. 41). Most of this discontinuous aquifer area is outside this basin and within the Kankakee River basin.

In areas with discontinuous sand and gravel aquifers, the interconnection between ground water and surface water is not evident due to the low permeability of the enclosing materials. Water in discontinuous sand and gravel aquifers is generally confined. When several discontinuous aquifers are penetrated, the potentiometric head in each successive aquifer is generally lower than the potentiometric head of the aquifer above it. This is indicative of ground-water recharge through low-permeability materials. Well yields (exclusive of domestic wells) from discontinuous sand and gravel deposits range from 20 to 300 gal/min, with a median of 90 gal/min (Nyman and Pettijohn, 1971, p. 47). Median depth of wells in discontinuous sand and gravel aquifers is slightly greater than that of wells in continuous sand and gravel (buried and surficial) to accommodate larger drawdowns.

Sand and Gravel Within Buried Preglacial Bedrock Valleys

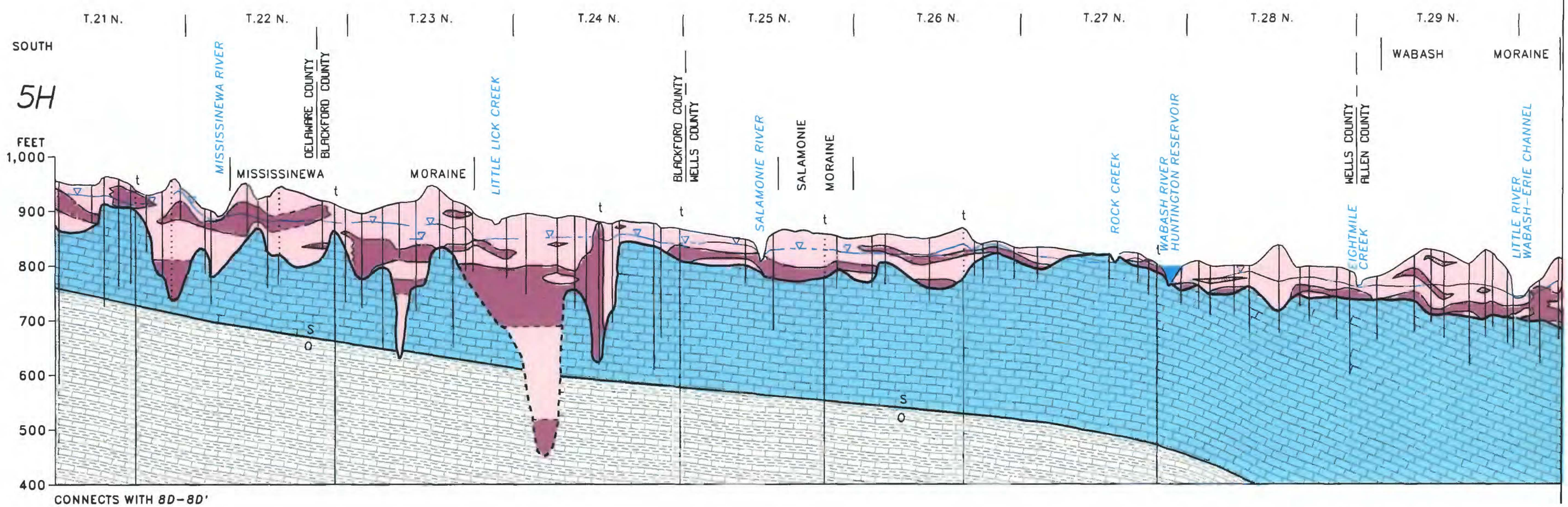
During this century, information obtained by exploratory drilling for oil and water has been used for delineation of a deeply buried preglacial valley system (fig. 7). The preglacial Lafayette Bedrock Valley (formerly known as the Teays Valley) crosses Indiana through the Upper Wabash River basin. Repeated glaciation filled the Lafayette Bedrock Valley with glacial and lake sediments. The valley was filled by different geologic events at different times. The Lafayette Bedrock Valley underlies 384 mi^2 (Nyman and Pettijohn, 1971, p. 55) of this basin.

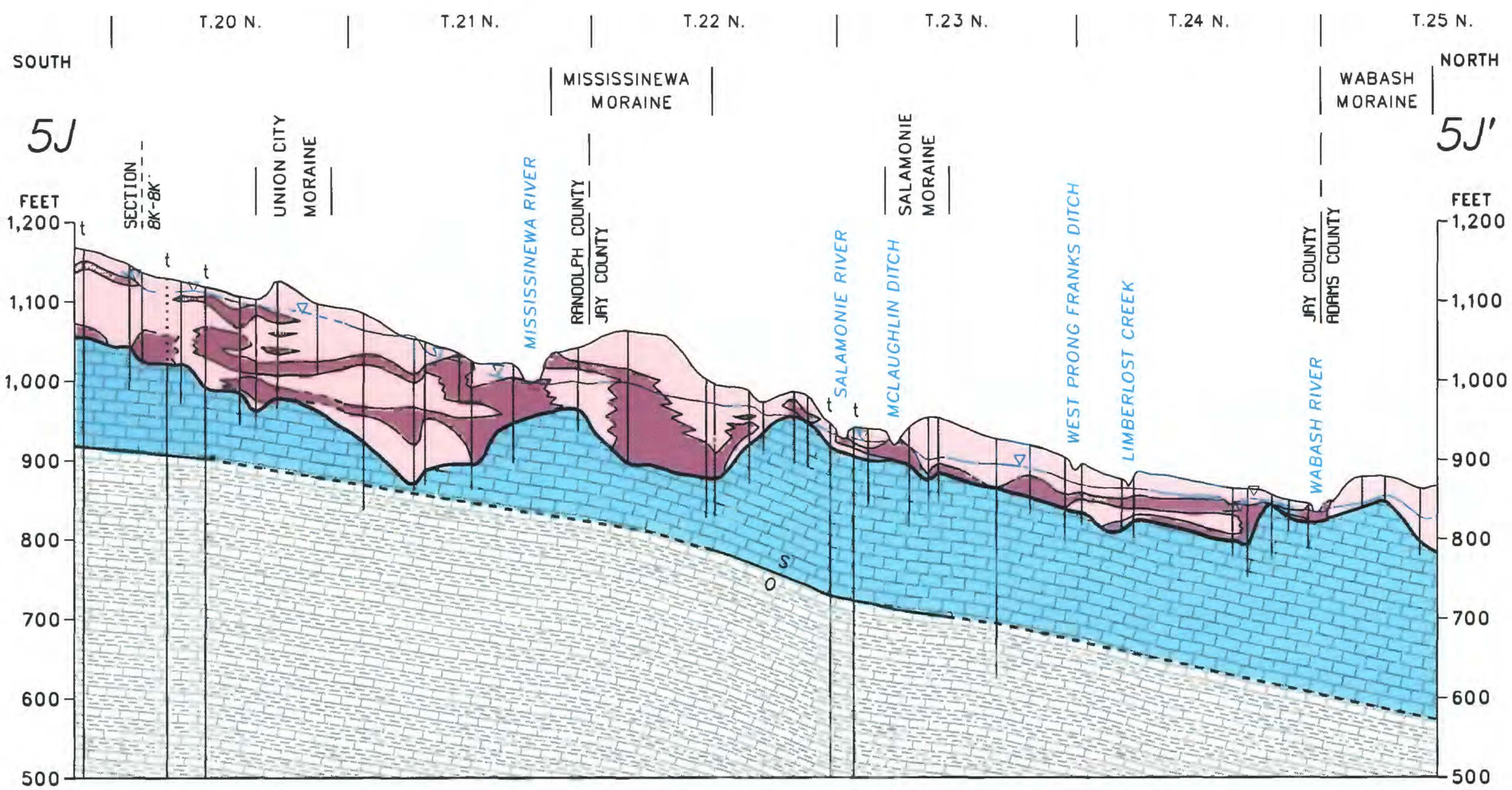
The only surface indications of the buried bedrock valley are in the area of Loblolly and Limberlost Creeks near Geneva in Adams and Jay Counties, and near Richvalley in Wabash and Miami Counties. Near Geneva, a swamp formed in the surface depression directly above the buried valley. This swamp is drained by Loblolly Creek, which follows the buried valley. The drape of the stratigraphy above this deep valley is shown in T. 24 N. of hydrogeologic section 5I–5I′ (fig. 40).

At Richvalley, the Wabash River's bedrock channel ends and the valley widens as the Lafayette Bedrock Valley tangentially crosses 170 ft below the river. Hydrogeologic section 5E–5E′ (fig. 40) transects the valley (T. 27 N.) between Richvalley and Peru. Although the Wabash River has a rock channel upstream and downstream, it has an alluvial channel in this area. Channel width and slump compaction of thick valley fill has facilitated the deposition of thin alluvial sediments over thick glacial outwash in the Richvalley area.

The stratigraphy of the valley fill has been recently studied. Bleuer (1991) and Bleuer and others (1991) report on the stratigraphy of unconsolidated deposits within the buried valley. Detailed stratigraphic information on the location of thick, highly transmissive unconsolidated deposits within the buried valley are intended to allow planners to locate high-yield wells. In summary, Bleuer and others (1991, p. 88) state the general rule "...west is best, east is least...," meaning that the buried aquifer has high yields from about Peru westward to Illinois, whereas productivity is less dependable eastward toward Ohio.

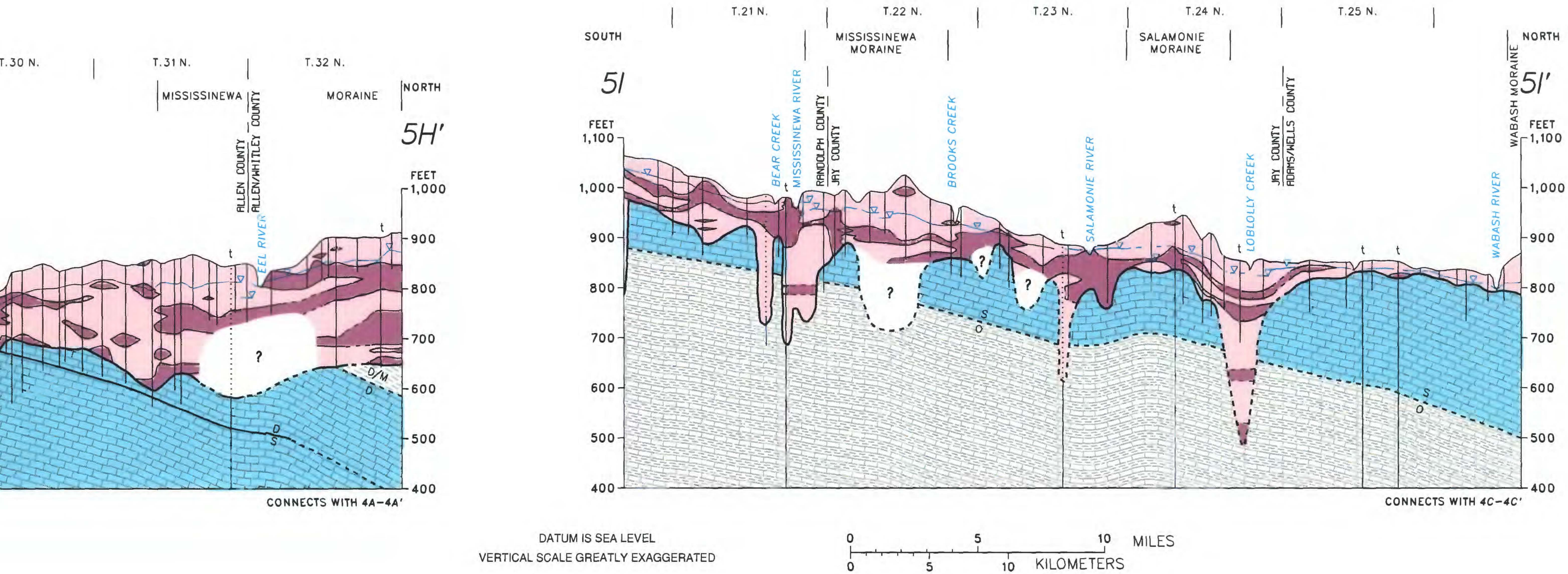
Figure 40. Hydrogeologic sections *5A–5A'* to *5J–5J'* of the Upper Wabash River basin—Continued.





The largest well yields in the Upper Wabash River basin are from unconsolidated aquifers in the buried bedrock valley. Yields from nondomestic wells ranged from 25 to 2,000 gal/min; a median yield of 700 gal/min is reported by Nyman and Pettijohn (1971, p. 47). The valley fill is very productive in several areas. Several municipal users withdraw ground water from aquifers in or just above the Lafayette Bedrock Valley. Cities in the Upper Wabash River basin that use the buried valley aquifer are Berne, Decatur, Geneva, Marion, La Fontaine, and Peru. Some valley fill aquifers are so thick and areally extensive that they can easily supply sustainable yields of tens of millions of gallons per day (Bleuer and others, 1991, p. 79).

Three areas of the Upper Wabash River basin are underlain by few unconsolidated aquifers. These three areas are (1) Adams and Wells County, (2) western Grant County, and (3) parts of White and Pulaski Counties. These three areas are similar in that thin drift deposits overlie scoured bedrock. Although yield from the drift-bedrock interface is small, a thin sand and gravel unit is generally present.



Bedrock Aquifers

Carbonate Bedrock Aquifers

A Silurian-Devonian carbonate bedrock aquifer underlies nearly all of the Upper Wabash River basin (fig. 41). Carbonate bedrock is absent from the base of the Lafayette Bedrock Valley in the eastern part of the basin, whereas it is 700 ft thick near the tricorner area of Allen, Whitley, and Noble Counties. Along the far northern side of the basin, increasing drift thickness and an overlying shale unit restrict access to the carbonate bedrock aquifer. The carbonate bedrock aquifer is not commonly used in areas where the overlying drift thickness exceeds 150 ft.

Ground-water flow in the carbonate bedrock aquifer is through vertical fractures, horizontal bedding planes, and solution openings. Karstification of the carbonate bedrock aquifer by surface water entering the ground-water system has further enhanced the secondary permeability of the carbonate bedrock aquifer throughout the basin. This secondary enhancement of permeability is responsible for the largest well yields available from the carbonate bedrock aquifer. Planert (1980, p. 15) found that the hydraulic conductivity of the carbonate bedrock aquifer in Allen County is greatest near the preglacial erosion surface and decreases with depth.

A ground-water divide separates regional ground-water flow in the northeastern part of the carbonate bedrock aquifer in the Upper Wabash River basin (Greeman, 1991). Northeast of this divide, more than 1,000 mi^2 of the Upper Wabash River basin, and part of the White River basin (fig. 1) drain toward the Maumee River basin. The ground-water flow in the regional carbonate bedrock aquifer is toward the Maumee River basin from as far west as Huntington, Ind. and as far south as Randolph County (fig. 36) (Greeman, 1991). Southwest of the regional ground-water flow divide in the Silurian-Devonian carbonate bedrock aquifer, ground-water flow is toward the Wabash River. Hydrogeologic sections 5G–5G′ to 5J–5J′ (fig. 40) illustrate these conclusions.

In the Upper Wabash River basin, yields from the Silurian-Devonian carbonate bedrock aquifer are suitable for domestic and stock uses. Yields suitable for small industries are common; however, large yields are less common. Well yields from the Silurian-Devonian carbonate bedrock aquifer system range from 15 to 1,250 gal/min. Median yields for nondomestic wells are 200 and 360 gal/min for the Silurian and Devonian carbonate rocks, respectively (Nyman and Pettijohn, 1971, p. 47).

The Silurian-Devonian carbonate bedrock aquifer system is the alternative ground-water source for the Upper Wabash River basin. If sufficient water is not found in the unconsolidated deposits, the carbonate bedrock is a dependable source in most of the area. North of the Eel and Tippecanoe Rivers, the carbonate bedrock aquifer is seldom used because water supply is available from shallower sand and gravel aquifers. South of the Eel and Tippecanoe Rivers, the carbonate bedrock aquifer is within 100 ft of the land surface in many areas; however, not all wells tap the carbonate rock, as domestic supplies are commonly available from sand and gravel deposits.

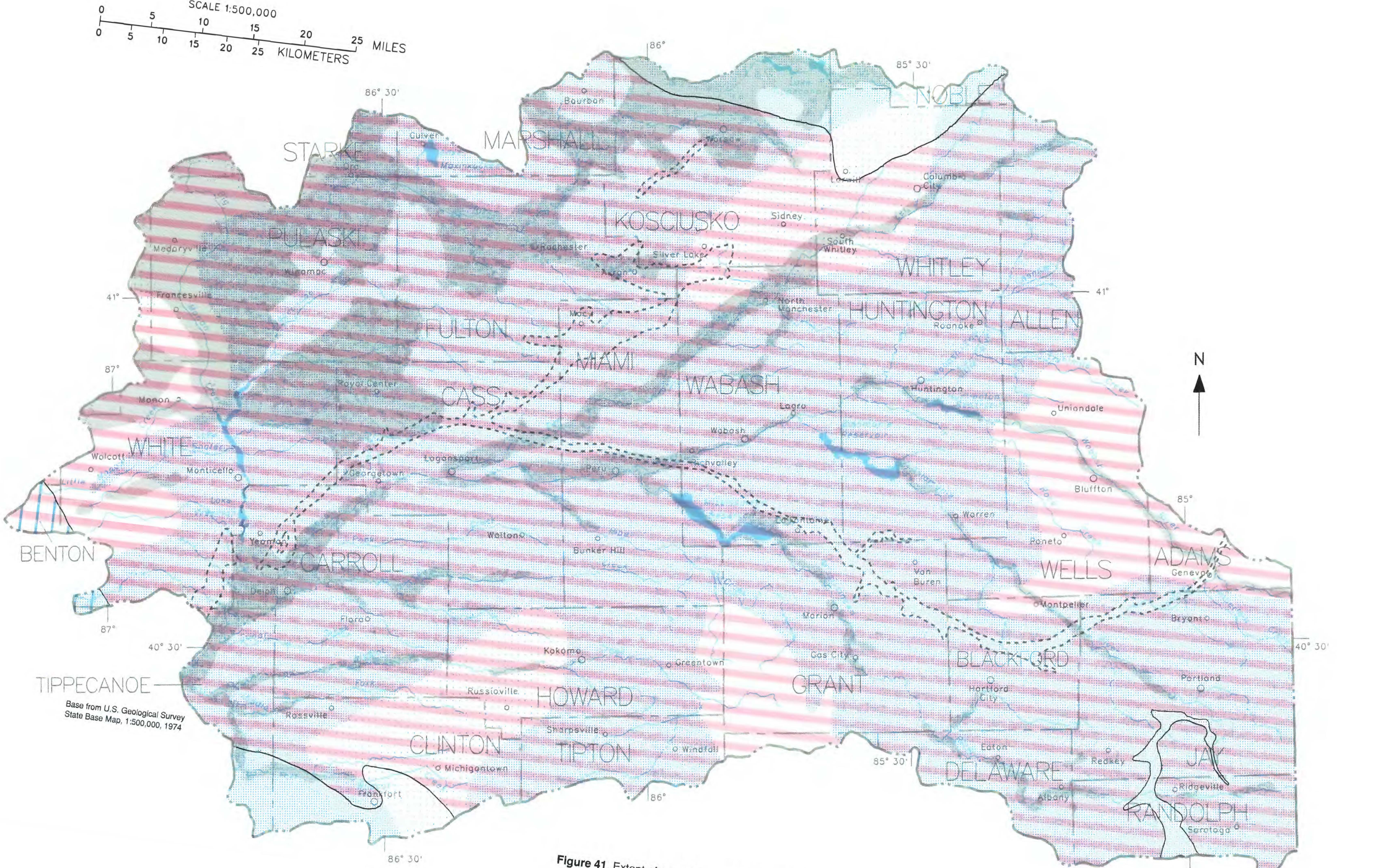


Figure 41. Extent of aquifer types in the Upper Wabash River basin.

EXPLANATION

SURFICIAL SAND AND GRAVEL AQUIFER

BURIED SAND AND GRAVEL AQUIFER

DISCONTINUOUS SAND AND GRAVEL AQUIFER-- Generally thin lenses of buried sand and gravel

CARBONATE BEDROCK AQUIFER

UPPER WEATHERED-BEDROCK AQUIFER

BURIED BEDROCK VALLEY AQUIFER

WATER-MANAGEMENT-BASIN BOUNDARY

NOTE: This map is designed for regional evaluations and should not be the basis for evaluation of specific sites

Table 7. Characteristics of aquifer types in the Upper Wabash River basin
[Locations of aquifer types shown in fig. 41]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and gravel	0- 80	20- 1,350	
Buried sand and gravel	0-120	20- 1,000	
Discontinuous sand and gravel	0- 80	20- 300	
Sand and gravel in buried bedrock valley	0-200	25- 2,000	Lafayette Bedrock Valley aquifer[1]
Carbonate bedrock	0-700	15- 1,250	Silurian-Devonian carbonate bedrock aquifer
	0- 22	0- 200	Rockford Limestone

[1]Bleuer and others, 1991.

Where the carbonate rock is overlain by shale, ground-water circulation is restricted. This restriction allows time for accessory minerals to dissolve and degrade the quality of the ground water in the carbonate bedrock aquifer. Concentrations of dissolved minerals become objectionable in areas where circulation is restricted. A common objectionable constituent is sulfate, which dissolves from gypsum. Elevated concentrations of sulfate are primarily reported from carbonate rocks of Devonian age. Bleuer and Moore (1978, p. 43) report that ground water from the carbonate bedrock aquifer has significantly higher concentrations of dissolved strontium and sodium than ground water from the glaciofluvial aquifers.

A Lower Mississippian carbonate bedrock aquifer (Rockford Limestone) underlies the 100 mi^2 Borden Group subcrop in the far western part of the Upper Wabash River basin. Hydrogeologic section 5A–5A′ (fig. 40) indicates the location of the Rockford Limestone and its distribution. The Rockford Limestone is not a productive aquifer. It is, however, approximately 130 ft shallower than the Silurian-Devonian carbonate bedrock aquifer in this area. The Rockford Limestone attains a thickness of as much as 22 ft in this basin. Although well yields are small, they may be suitable for domestic needs.

The deep Ordovician and Cambrian rocks contain several potential aquifers. In the Upper Wabash River basin, however, these Ordovician and Cambrian rocks are not used as aquifers, because they are deeply buried and the water is saline.

Summary

The Upper Wabash River basin, located in north-central Indiana, is the largest water management basin (6,918 mi^2) in Indiana. It includes the cities of Bluffton, Columbia City, Frankfort, Hartford City, Huntington, Kokomo, Logansport, Marion, Monticello, North Manchester, Peru, Portland, Rochester, Wabash, and Warsaw.

Six different aquifer types were mapped within the basin (fig. 41): (1) surficial sand and gravel, (2) buried sand and gravel, (3) discontinuous buried sand and gravel, (4) sand and gravel within buried preglacial bedrock valleys, (5) carbonate bedrock of Silurian and Devonian ages and Lower Mississippian age, and (6) an upper weathered-bedrock aquifer in Mississippian shale.

The Upper Wabash River basin contains large volumes of usable ground water. The principal source of ground water in this basin is the unconsolidated deposits. More than 600 mi^2 of the basin is covered by surficial sand and gravel deposits, which average 30 ft in thickness (Nyman and Pettijohn, 1971, p. 55). Most surficial deposits are in the Tippecanoe River drainage. Domestic water supplies from buried sand and gravel aquifers are available throughout most of the basin. Yields from the sand and gravel aquifers range from 20 to 1,350 gal/min; the largest yields are in Kosciusko, Noble, and Whitley Counties.

The most productive aquifer system in the basin is within the buried Lafayette Bedrock Valley deposits. Some segments of the Lafayette Bedrock Valley contain unconsolidated deposits that are capable of producing as much as 2,000 gal/min. The cities of Decatur, Geneva, Marion, La Fontaine, and Peru derive all or part of their water supply from the buried-valley segments. Some valley-fill aquifers are so thick and areally extensive that they can supply sustainable yields of tens of millions of gallons per day.

The Silurian-Devonian carbonate bedrock aquifer is the primary bedrock ground-water resource. The carbonate bedrock aquifer is within 300 ft of the land surface in more than 80 percent of the Upper Wabash River basin. Yields of wells that tap carbonate bedrock aquifers in this basin range from 0 to 1,250 gal/min; however, the bedrock aquifer is seldom used in most areas because of the abundance of ground water in unconsolidated aquifers.

References Cited

Bassett, J.L., and Hasenmueller, N.R., 1979, Map showing structure on base of the New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources, Geological Survey, EGSP Series 800, scale 1:500,000.

Bassett, J.L., and Hasenmueller, N.R., 1980, Map of Indiana showing structure on top of the Maquoketa Group (Ordovician): Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 812, scale 1:500,000.

Bassett, P.A. and Keith, B.D., 1984, Data base for deep wells in Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 46, 29 p.

Bleuer, N.K., 1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley system (so-called Teays valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

_______1991, The Lafayette Bedrock Valley System of Indiana—concept, form, and fill stratigraphy, *in* Melhorn, W.N. and Kempton, J.P., eds., Geology and hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 51-77.

Bleuer, N.K., Melhorn, W.N., Steen, W.J. and Bruns, T.M., 1991, Aquifer systems of the buried Marion-Mahomet trunk valley (Lafayette Bedrock Valley System) of Indiana, *in* Melhorn, W.N. and Kempton, J.P., eds., Geology and Hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 79-89.

Bleuer, N.K., and Moore, M.C., 1972, Glacial stratigraphy of the Fort Wayne area and the draining of glacial Lake Maumee, *in* Proceedings of the Indiana Academy of Science for 1971: v. 81, p. 105-209.

_______1978, Environmental geology of Allen County, Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 13, 72 p.

Bruns, T.M., Logan, S.M., and Steen, W.J., 1985a, Map showing bedrock topography of the Teays Valley, western part, north-central Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 42, scale 1:100,000.

_____1985b, Map showing bedrock topography of the Teays Valley, central part, north-central Indiana: Indiana Department Natural Resources, Geological Survey Miscellaneous Map 43, scale 1:100,000.

_____1985c, Map showing bedrock topography of the Teays Valley, eastern part, north-central Indiana: Indiana Department Natural Resources, Geological Survey Miscellaneous Map 44, scale 1:100,000.

Capps, S.R., 1910, The underground waters of north-central Indiana: U.S. Geological Survey Water-Supply Paper 254, 279 p.

Cazee, Sherry, 1988, Catalog of core samples of the Indiana Geological Survey: Indiana Department of Natural Resources, Geological Survey Directory 9, 83 p.

Chamberlin, T.C., 1883, Preliminary paper on the terminal moraine of the second glacial epoch: U.S. Geological Survey, Third Annual Report, 1881-1882, p. 291-402.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Droste, J.D., and Shaver, R.H., 1983, Atlas of early and middle Paleozoic paleogeography of the southern Great Lakes area: Indiana Department of Natural Resources, Geological Survey Special Report 32, 32 p.

Fullerton, D.S., 1980, Preliminary correlation of Post-Erie interstadial events (16,000-10,000 radiocarbon years before present), central and eastern Great Lakes region, and Hudson, Champlain, and St. Lawrence lowlands, United States and Canada: U.S. Geological Survey Professional Paper 1089, 52 p.

Gillies, D.G., 1981, Ground-water potential of the glacial deposits near Logansport, Cass County, Indiana: U.S. Geological Survey, Water-Resources Investigations 81-7, 98 p.

Gooding, A.M., 1973, Characteristics of Late Wisconsin tills in eastern Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 49, 28 p.

Gray, H.H., 1979, The Mississippian and Pennsylvanian (Carboniferous) systems in the United States—Indiana: U.S. Geological Survey Professional Paper 1110-K, 20 p.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Greeman, T.K., 1991, Areal differences between ground-water and surface-water drainage basins in Indiana [abs.]: Midwest Ground Water Conference, 36th, Indianapolis, 1991, program with abstracts, p. 106-107.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Public Works, Conservation Department, Division of Geology Publication 133, 504 p.

Harrison, Wyman, 1958, Marginal zones of vanished glaciers reconstructed from the preconsolidation-pressure values of overridden silts: Journal of Geology, v. 66, n. 1, January 1958, p. 72-95.

Hasenmueller, N.R. and Bassett, J.L., 1980a, Map of Indiana showing thickness of New Albany Shale (Devonian and Mississippian) and equivalent strata: Indiana Department of Natural Resources, Geological Survey, EGSP Series 805, scale 1:500,000.

_____1980b, Map of Indiana showing structure on top of Silurian rocks: Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 811, scale 1:500,000.

_____1980c, Map of Indiana showing structure on top of the Trenton Limestone (Ordovician): Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 813.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hoggatt, R.E., Hunn, J.K., and Steen, W.J., 1968, Water resources of Delaware County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 37, 57 p.

Indiana, State of, 1989, Nonpoint source assessment report, part II: Indiana Department of Environmental Management, 28 p.

Marie, J.R. and Davis, L.G., 1974, Water resources of the middle Wabash River basin, Indiana: U.S. Geological Survey Hydrologic Investigations Atlas HA-508, 3 sheets.

McBeth, W.A., 1902, Wabash River terraces in Tippecanoe County, Indiana, *in* Proceedings of the Indiana Academy of Science for 1901: Indianapolis, Ind., p. 237-243.

Nyman, D.J., and Pettijohn, R.A., 1971, Ground water, *in* Wabash River Coordinating Committee, Wabash River basin comprehensive study: U.S. Army Corps of Engineers, Louisville Ky., volume VIII, appendix G, 100 p.

Palmer, A.R., 1983, Decade of North American geology—1983 geologic time scale: Geology, v. 11, p. 503-504.

Planert, Michael, 1980, Ground-water availability near Fort Wayne, Allen County, Indiana: U.S. Geological Survey Water-Resources Investigations 80-34, 54 p.

Rosenshein, J.S., 1958, Ground-water resources of Tippecanoe County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 8, 37 p.

Rosenshein, J.S. and Cosner, O.J., 1956, Ground-water resources of Tippecanoe County, Indiana—appendix, basic data: Indiana Department of Conservation, Division of Water Resources Bulletin 8, 65 p.

Rosenshein, J.S. and Hunn, J.D., 1964a, Ground-water resources of northwestern Indiana—preliminary report, Marshall County: Indiana Department of Conservation, Division of Water Resources Bulletin 19, 157 p.

_____1964b, Ground-water resources of northwestern Indiana—preliminary report, Fulton County: Indiana Department of Conservation, Division of Water Resources Bulletin 20, 83 p.

_____1964c, Ground-water resources of northwestern Indiana—preliminary report, Starke County: Indiana Department of Conservation, Division of Water Resources Bulletin 22, 87 p.

_____1964d, Ground-water resources of northwestern Indiana—preliminary report, Pulaski County: Indiana Department of Conservation, Division of Water Resources Bulletin 24, 71 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Shaver, R.H.; Burger, A.M.; Gates, G.R.; Gray, H.H.; Hutchison, H.C.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; Smith, N.M.; Wayne, W. J.; and Wier, C.E., 1970, Compendium of rock-unit stratigraphy in Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 43, 229 p.

Smith, B.S., Hardy, M.A., and Crompton, E.J., 1985, Water resources of Wildcat Creek and Deer Creek basins, Howard and parts of adjacent Counties, Indiana, 1979-82: U.S. Geological Survey Water-Resources Investigations Report 85-4076, 92 p.

Stallmam, R.W. and Klaer, F.H., Jr., 1950, Ground-water resources of Noble County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 5, 103 p.

Tate, C.H.; Davis, L.G.; Johnson, L.E.; and Pettijohn, R.A., 1973, Water resources of the Upper Wabash River Basin, northeastern Indiana: U.S. Geological Survey Hydrologic Investigations Atlas HA-433, 3 sheets.

Wabash River Coordinating Committee, 1971, Wabash River Basin comprehensive study: Louisville, Ky., U.S. Army Corps of Engineers, 14 v., variable pagination.

Watkins, F.A., Jr., and Rosenshein, J.S., 1963, Ground-water geology and hydrology of Bunker Hill Air Force Base and vicinity, Peru, Indiana: U.S. Geological Survey Water-Supply Paper 1619-B, 32 p.

Watkins, F.A., Jr., and Ward, P.E., 1962, Ground-water resources of Adams County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 9, 67 p.

Wayne, W.J., 1963, Pleistocene formations of Indiana: Indiana Department of Conservation, Geological Survey Bulletin 25, 85 p.

MIDDLE WABASH RIVER BASIN

By Paul K. Doss

General Description

The Middle Wabash River basin, as defined in this report, encompasses 3,453 mi^2 of west-central Indiana (Hoggatt, 1975). The basin is bounded on the west by Illinois, extends eastward to approximately 12 mi east of Lebanon, and extends north-south from approximately 10 mi south of Terre Haute to approximately 18 mi north of Lafayette (fig. 42). The Middle Wabash River basin includes all of Fountain, Montgomery, Vermillion, and Warren Counties, significant parts of Benton, Boone, Parke, Tippecanoe, and Vigo Counties, and small parts of six other counties. The largest population centers in the Middle Wabash River basin, listed in order of relative size, are Terre Haute, Lafayette, West Lafayette, Crawfordsville, and Lebanon.

Previous Studies

Several facets of the water resources of a large part of the Middle Wabash River basin were examined by Wangsness and others (1983), who summarized regional ground-water quantity and quality, surface-water quantity and quality, and precipitation characteristics. Another large-scale study was done by the Wabash River Coordinating Committee (1971) on the entire Wabash River basin. The Middle Wabash River basin, as defined in that report, included the Middle Wabash River basin as defined here, along with drainage areas in Illinois. Nyman and Pettijohn (1971) concentrated on the ground-water resources component of that study and reported on aquifer definition, base flow to streams, chemical quality, and management considerations. A report by Marie and Davis (1974) refers to a Middle Wabash River basin that is geographically different from the Middle Wabash River basin defined in this report. The downstream limit of the basin studied by Marie and Davis is virtually the upstream limit of this report; however, the water budget, ground-water availability, surface-water data, and other aspects of the report by Marie and Davis are pertinent because aquifers common to both areas may be hydraulically connected and of common origin. In addition, climatic and surface-water characteristics are similar for both basins.

The cities of Lafayette and West Lafayette, combined, form the largest population center within the Middle Wabash River basin. Several studies have examined the hydrogeology of glacial deposits underlying the Lafayette area. Maarouf and Melhorn (1975) discussed shallow-bedrock and unconsolidated aquifers and produced lithofacies maps of the unconsolidated materials to a depth of 400 ft. The ground-water resources of buried bedrock-valley deposits west of the Lafayette area in Illinois, were described by Visocky and Schicht (1969). The potentiometric surface in, and hydraulic conductivity of, outwash deposits near Lafayette were determined by Pohlmann (1986), who also suggested values for recharge rates through overlying tills. Some management considerations for the potential exploitation of ground-water resources in the Lafayette area were suggested by Loganathan and others (1980).

Several reports have been prepared on a county scale that describe the ground-water resources of counties in the Middle Wabash River basin. Reports by Watkins and Jordan list well-records and give preliminary information on geology and ground-water resources in Montgomery, Fountain and Vermillion Counties (1965a, 1965b, 1965c), and Putnam and Parke Counties (1964a, 1964b). Other county reports not only describe these aspects of ground-water resources but also include discussions of geology and water quality for Boone County (Brown, 1949), Tippecanoe County (Rosenshein and Cosner, 1956; Rosenshein, 1958), and Montgomery County (Cable and Robison, 1974). Several reports discuss the ground-water resources in Vigo County (Steen and Uhl, 1959; Watkins and Jordan, 1963). Cable and others (1971) delineate and discuss the principal bedrock and unconsolidated aquifers in Vigo and Clay Counties. The flow system and ground-water quality in the Pennsylvanian Mansfield Formation in Clay County are described by Thomas (1980).

Physiography

Topographic relief within the Middle Wabash River basin is approximately 530 ft. Altitudes range from a low of nearly 440 ft above sea level to a high of 970 ft above sea level. The lowest point is in the valley of the Wabash River at its exit from the basin in Vigo County and the highest point is in the uplands of Boone County in the eastern part of the basin.

The valley of the Wabash River is a dominant physiographic feature within the basin. In places the valley is 2 to 3 mi wide. Large expanses of flood-plain lowlands are present at the confluences of major tributaries such as the Vermillion River and Big Raccoon Creek. Prominent terrace surfaces can also be seen in many areas along the course of the river valley.

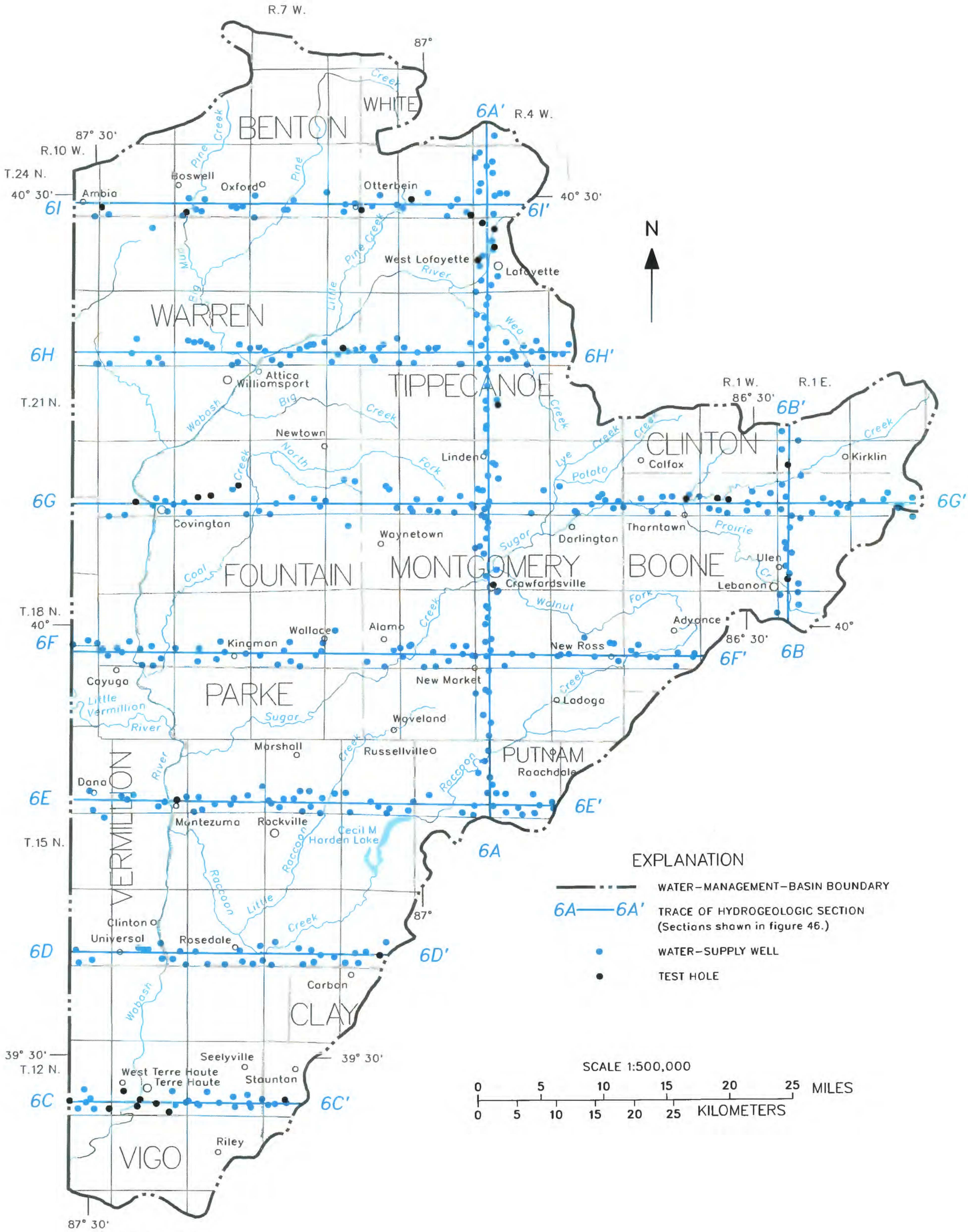
The surface physiography of the Middle Wabash River basin is dominated by the Tipton Till Plain (fig. 43). The southern extent of the Tipton Till Plain is marked by the Shelbyville Moraine and coincides with the southern extent of Wisconsinan glaciation in the basin. The Tipton Till Plain is a generally featureless, flat to gently-rolling plain, which is interrupted in places by very low-relief end moraines, including the Chatsworth, Crawfordsville, and Shelbyville Moraines (Malott, 1922, pl. III, p. 107; Schneider, 1966, p. 50). Some ice-disintegration features, including disintegration ridges and prairie mounds, also can be seen within the Tipton Till Plain in the Middle Wabash River basin (Bleuer, 1974).

South of the Wisconsinan glacial boundary, physiography is controlled by bedrock, although pre-Wisconsinan glacial deposits and Wisconsinan loess are present at land surface. The Wabash Lowland makes up most of the area south of the Wisconsinan glacial boundary within the Middle Wabash River basin (fig. 43). The Wabash Lowland is characterized by broad, terraced valley bottoms and undulating uplands (Schneider, 1966, p. 48). The overall subdued topography is controlled dominantly by the underlying fine-grained, clastic Pennsylvanian bedrock. Immediately east of the Wabash Lowland is the Crawford Upland (fig. 43), a dissected, westward-sloping upland (Schneider, 1966, p. 47). The diverse topography of the Crawford Upland reflects the underlying bedrock, an alternating sequence of resistant and nonresistant strata of Upper Mississippian Chester rocks and the basal-Pennsylvanian Mansfield Formation.

There are several large areas of human-disturbed land in the basin, particularly in the southern part in Vermillion, Vigo, Clay, and Parke Counties. Most of the disturbed land in those counties results from surface mining of coal and tailings disposal.

Surface-Water Hydrology

The Wabash River is the main surface drainage channel within the Middle Wabash River basin. Mean discharges at gaging stations which monitor the Wabash River range from 6,484 ft^3/s at Lafayette to 10,870 ft^3/s at Terre Haute (Arvin, 1989). Daily mean discharges for the 1988 calendar year at Lafayette range from a minimum of 702 ft^3/s to a maximum of 32,600 ft^3/s (Thompson and Nell, 1990, p. 99). The Wabash River flows from east to west in the northern part of the basin and from north to south in the central and southern parts; the major change in



Base from U.S. Geological Survey State Base Map, 1:500,000, 1974

Figure 42. Location of section lines and wells plotted in the Middle Wabash River basin.

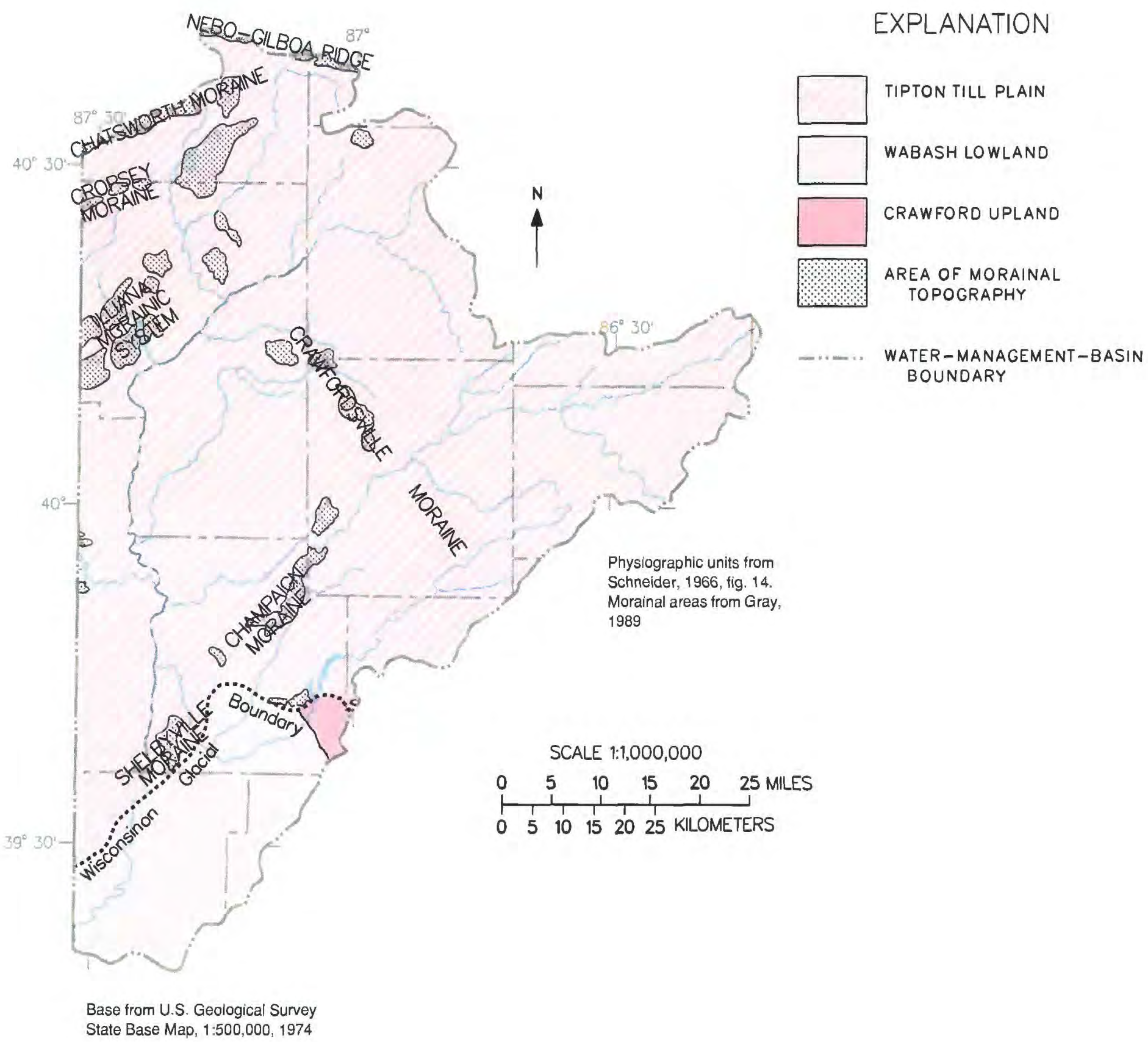


Figure 43. Physiographic units, moraines, and extent of glaciation in the Middle Wabash River basin.

direction of flow is near Covington, Ind. (fig. 42). The Wabash River enters the basin near Lafayette in Tippecanoe County, flows through Tippecanoe and Warren Counties, continues along the county line between Vermillion County and Fountain and Parke Counties, and then exits the basin through Vigo County at the Illinois-Indiana State line.

Little Pine, Big Pine, and Mud Pine Creeks drain the northern part of the Middle Wabash River basin (fig. 42). The confluence of Big Pine Creek with the Wabash River is at Attica, Indiana. The Coal Creek system drains the north-central and central part of the basin. The confluence of Coal Creek with the Wabash River is just east of Cayuga. Sugar Creek, a major tributary in the basin, flows southwest out of Crawfordsville and drains the entire eastern and south-central part of the basin. The confluence of Sugar Creek with the Wabash River is north of Montezuma. Raccoon Creek drains the southeastern and southern part of the basin. A reach of Raccoon Creek has been dammed to create Cecil M. Hardin Lake (Mansfield Reservoir), east of Rockville in Eastern Parke County. The confluence of Raccoon Creek with the Wabash River is south of Montezuma. The Vermillion River drains the western part of the basin and joins the Wabash River southeast of Cayuga. Much of the Vermillion River basin is in Illinois and hence, is not covered in this report. Most of the major tributaries to the Wabash River in western Indiana have, in places, cut through the cover of unconsolidated deposits and occupy valleys with exposed bedrock valley walls.

Geology

Bedrock Deposits

The Middle Wabash River basin lies on the eastern and northeastern margin of the structural Illinois Basin and on the southwestern limb of the Kankakee Arch (fig. 4). Bedrock units strike generally northwest, dipping gently into the interior of the Illinois Basin. Subcrops of rock units at the bedrock surface are progressively younger westward (fig. 44). The oldest rocks exposed at the bedrock surface are carbonate rocks of the Wabash Formation of Silurian age in Boone, Clinton, and Tipton Counties (fig. 44; Gray and others, 1987). The youngest bedrock subcrop includes clastic rocks of the Pennsylvanian McLeansboro Group in Vigo and Vermillion Counties along the Illinois-Indiana State line.

At least three major unconformities are present in the stratigraphic sequence, including the Silurian-Devonian disconformity, the Mississippian-Pennsylvanian angular unconformity, and the pre-Pleistocene angular unconformity (the Silurian-Devonian disconformity is not shown on hydrogeologic sections of the Middle Wabash River basin; Silurian and Devonian rocks are represented as a continuous sequence of parallel carbonate rocks). The erosional surface at the Mississippian-Pennsylvanian boundary, in some places, is characterized by topography with significant relief, including valleys as much as 115 ft deep (Gray, 1979, p. 12). A comprehensive interpretation of the configuration of the pre-Pennsylvanian surface by Keller (1990) was used as an aid in verification of the position of the Mississippian-Pennsylvanian boundary in the subsurface of the Middle Wabash River basin. An example of the Mississippian-Pennsylvanian angular unconformity can be seen in hydrogeologic section 6G–6G′ (fig. 46).

The bedrock surface underlying unconsolidated materials has preserved a topography that was at least partially acquired before Pleistocene glacial deposition. A regional, east-west paleodrainage system with many tributary valleys converge into a trunk valley in the vicinity of Lafayette, Ind. This bedrock valley system has historically been referred to as the Teays-Mahomet bedrock valley (Wayne, 1956, p. 36) and, more recently, as the Lafayette Bedrock Valley System (Bleuer, 1989; 1991; Bleuer and others, 1991). Excellent examples of this buried drainage system can be seen in hydrogeologic sections 6A–6A′ and 6I–6I′ (fig. 46). Detailed information on the altitude of the bedrock surface in the vicinity of Lafayette, Ind. near the complex coalescence of bedrock valleys was verified with information from Bruns and others (1985). Other bedrock elevations not specifically in the area of bedrock-valley convergence were verified with information from Gray (1982).

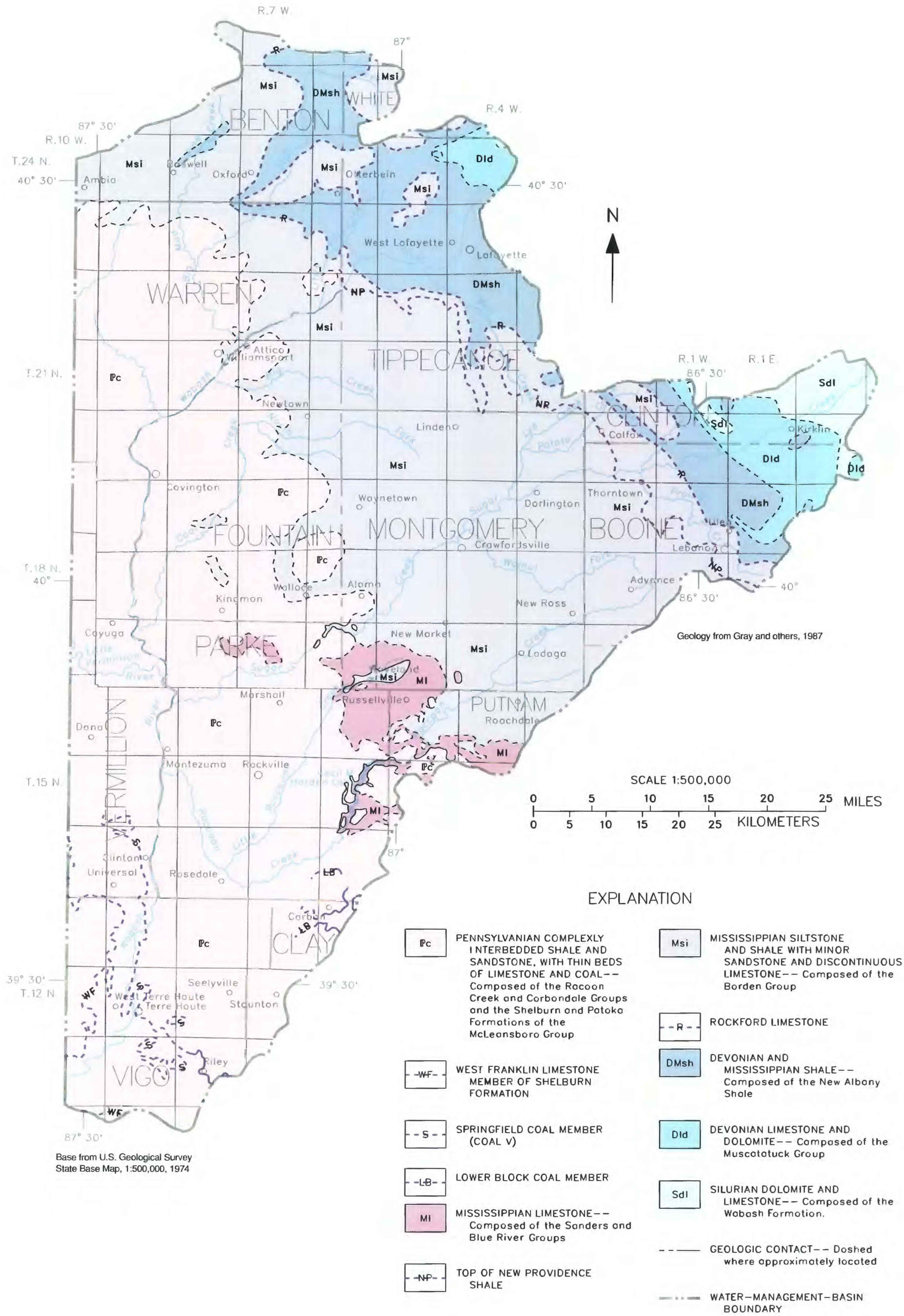


Figure 44. Bedrock geology of the Middle Wabash River basin.

The oldest rocks shown in the subsurface in this report are shales of the Ordovician Maquoketa Group. The Maquoketa Group is principally a shale unit that is slightly greater than 200 ft thick in northwestern Indiana (Shaver and others, 1986, p. 88). Only the upper 150 ft of the Maquoketa is shown in the subsurface of section 6G–6G′ (fig. 46).

Overlying Ordovician rocks is a sequence of Silurian and Devonian carbonate rocks. The entire thickness of these carbonate rocks is shown in the subsurface only in section 6G–6G′ (fig. 46) in which the thickness is 375 ft. The dominant unit in the Silurian and Devonian sequence is the Wabash Formation, which, in the vicinity of the Middle Wabash River basin, is greater than 250 ft thick (Shaver and others, 1986, p. 164). Other carbonate rocks that make up the Silurian and Devonian sequence include the Sexton Creek Limestone, the Salamonie Dolomite, and dolomites and limestones of the Muscatatuck Group (Shaver and others, 1986). Although not shown on the hydrogeologic sections, two test wells and one industrial water-supply well along sections 6B–6B′ and 6I–6I′ (fig. 46) penetrated the entire thickness of Silurian and Devonian carbonate rocks. Overlying the carbonate rock sequence is the New Albany Shale (Devonian and Mississippian). The New Albany Shale is a dark, carbon-rich shale that ranges in thickness from 100 to 120 ft in the Middle Wabash River basin (fig. 46).

Mississippian rocks in the Middle Wabash River basin are dominated by the Borden Group. Most of the Borden Group consists of siltstones, shales, fine sandstones, and discontinuous limestones (Shaver and others, 1986, p. 18). Where the entire thickness of the Borden Group has been interpreted in the subsurface of the Middle Wabash River basin, it is approximately 660 ft thick (section 6A–6A′, fig. 46).

In the northern part of the basin, the Borden Group is overlain unconformably by clastic rocks of the Pennsylvanian Mansfield Formation (fig. 44) (Gray and others, 1987). In the southeastern part of the basin, however, the Borden Group is overlain by truncated Mississippian carbonate rocks of the Sanders and Blue River Groups. The Sanders Group is made up of a variety of carbonate rocks, including fine-grained dolomites and fossiliferous limestones (Shaver and others, 1986, p. 137). The Blue River Group is composed largely of carbonate rocks (Shaver and others, 1986, p. 16). The maximum combined thickness of the Sanders and Blue River Groups in the Middle Wabash River basin is approximately 250 ft (section 6D–6D′, fig. 46).

Clastic rocks of the basal Pennsylvanian Raccoon Creek Group overlie Mississippian carbonate rocks in the south-central part of the basin and overlie Borden Group rocks in the northern part of the basin (fig. 44) (Gray and others, 1987). The Raccoon Creek Group is dominated by shales and sandstones, but it includes coals and minor limestones (Shaver and others, 1986, pp. 120-121). Much variability in the thickness of basal Pennsylvanian rocks is due to thick rock sequences in pre-Pennsylvanian valleys and thin rock sequences on top of pre-Pennsylvanian highlands (Gray, 1979, p. 13-14). The basal Mansfield Formation, ranging from 50 to 300 ft thick, is dominated by sandstones and contains coarse sands and conglomerates at its base (Shaver and others, 1986, p. 86). Thick Mansfield sandstones are interpreted along section 6D–6D′ (fig. 46) in R. 6 and 7 W. Similar thicknesses of the Mansfield Formation were documented for the same area by Hutchison (1976, pl. 4). The Mansfield Formation is overlain by the Brazil and Staunton Formations of the Raccoon Creek Group.

Clastic rocks of the Carbondale and McLeansboro Groups overlie the Raccoon Creek Group (fig. 44). The Carbondale Group, which is dominated by shales and sandstones and which contains four economically significant coals, consists of the Linton, Petersburg, and Dugger Formations (Shaver and others, 1986, p. 27). The only formation of the McLeansboro Group in the subsurface in the Middle Wabash River basin is the Shelburn Formation, which consists mainly of shale, siltstone, and sandstone and is less than 175 ft thick.

The maximum thickness of Pennsylvanian rocks interpreted in the Middle Wabash River basin is approximately 750 ft (sections 6C–6C′ and 6D–6D′, fig. 46).

Unconsolidated Deposits

The entire Middle Wabash River basin was glaciated during the Pleistocene, and unconsolidated materials at the surface are dominated by deposits of the Wisconsinan glaciation. The southern extent of Wisconsinan glaciation in the basin is in Vigo and Parke Counties (fig. 43). Total thickness of unconsolidated deposits ranges from zero at bedrock exposures along tributary valleys to greater than 350 ft in buried valleys near Lafayette (fig. 45). Bleuer (1989, 1991) examined the stratigraphy of bedrock-valley fill and geomorphic characteristics of the buried valleys to define the developmental history of the valley system.

North of the Wisconsinan glacial boundary, surficial deposits are dominated by loamy tills of the Trafalgar and Wedron Formations (Gray, 1989). At least three main Wisconsinan till deposits have been documented in a single exposure north of Williamsport, Ind. (fig. 42) (Bleuer, 1975). Sands and gravels deposited during Wisconsinan glaciation as outwash and stratified drift also are common. Sands and gravels, both at the surface and in the subsurface, are most extensive in the northern part of the basin and in bedrock valleys. Many large bodies of sand and gravel in the subsurface that are confined by tills and other nonaquifer material are correlative for lengths as much as 18 mi. Some basal sands and gravels at the bedrock surface and separated from overlying sands, as well as some basal sands in bedrock valleys are certainly of a pre-Wisconsinan origin. Surficial sand and gravel in recent river valleys may extend throughout the length of the basin.

South of the Wisconsinan glacial boundary, pre-Wisconsinan loamy tills of the Jessup and Glasford Formations are present at the land surface (Gray, 1989). Some pre-Wisconsinan lake silts and clays are at the surface southwest of Rosedale (fig. 42).

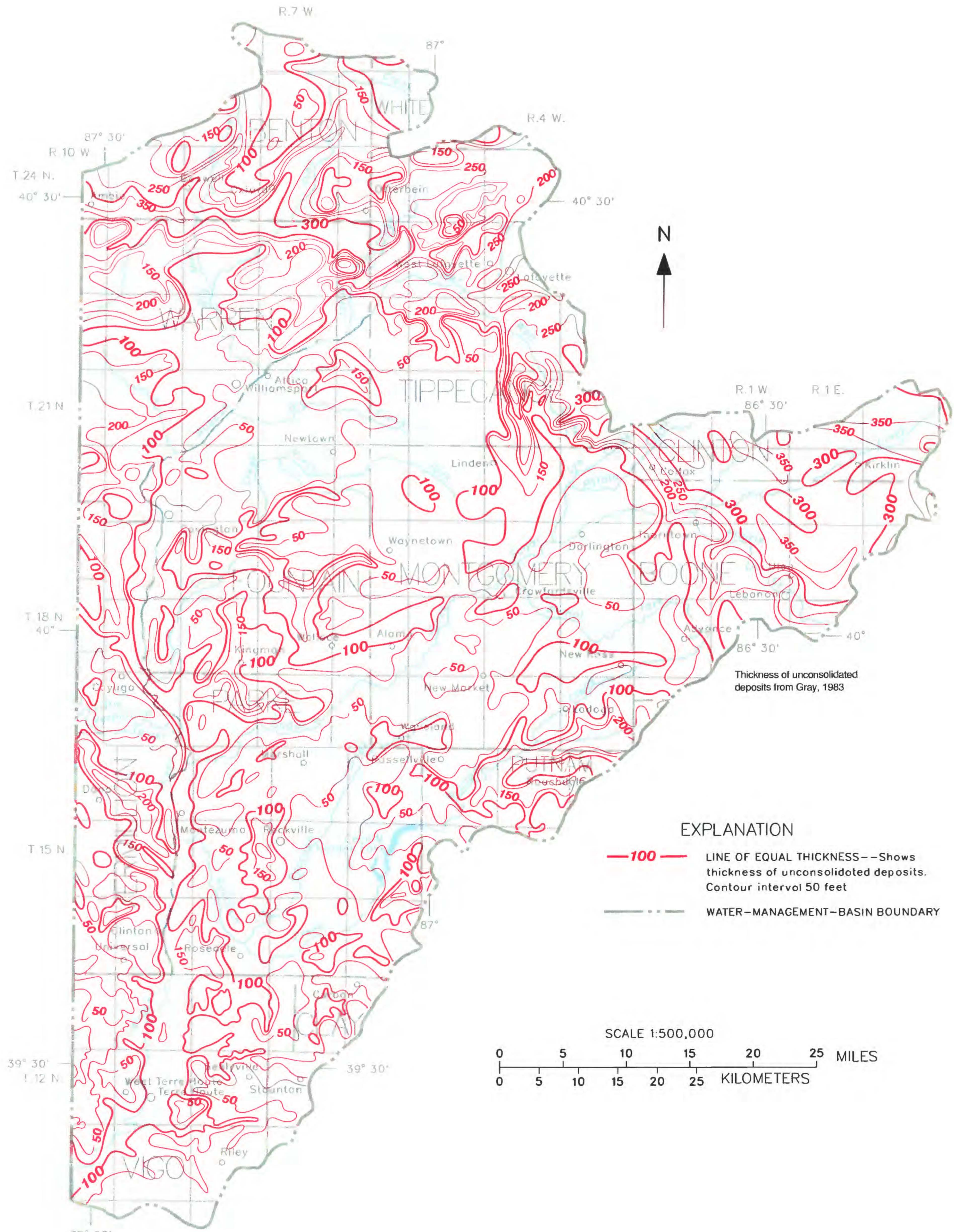
Holocene alluvium and colluvium are found in valleys of the Wabash River and its tributaries throughout the basin. Scattered loess (wind-blown silt) and dune-sand deposits are present in the central and southern parts of the basin. Most of these materials were derived from outwash in the Wabash Valley and are concentrated along the eastern margin and to the east of the valley.

Some Pleistocene sands and gravels have been cemented by calcium carbonate to form lithified sandstones and conglomerates interbedded with unconsolidated deposits (Rosenshein, 1955). These deposits are associated with the Wabash River and Wildcat Creek drainage basins and can be seen in outcrop overlying gray clays.

Aquifer Types

Nine hydrogeologic sections with a total length of 347 mi were constructed to describe aquifer types in the Middle Wabash River basin. Individual sections range in length from 16 to 67 mi. In total, 470 well records were used and the average density of plotted wells is 1.4 wells per mile. Sections 6A–6A′ and 6B–6B′ (fig. 46) are oriented south-north with a spacing of approximately 24 mi, and sections 6C–6C′ to 6I–6I′ (fig. 46) trend west-east with a spacing of 12 mi (fig. 42). The location and orientation of all the section lines provide coverage approximately perpendicular to each of the two major drainage directions of the Wabash and tributary rivers.

At least seven aquifer types are present in the Middle Wabash River basin: surficial sand and gravel; buried sand and gravel; discontinuous buried sand and gravel; complexly interbedded sandstone, shale, limestone, and coal; sandstone; Mississippian and Silurian-Devonian carbonate rocks; and an upper weathered-bedrock zone. Physical characteristics and some common or stratigraphic names for each aquifer type are listed in table 8. The areal distribution of these aquifers is shown in figure 47.



Base from U.S. Geological Survey
State Base Map, 1:500,000, 1974

Figure 45. Thickness of unconsolidated deposits in the Middle Wabash River basin.

Unconsolidated Aquifers

The most significant aquifer systems in the Middle Wabash River basin consist of unconsolidated surficial and buried sand and gravel that originated as outwash and alluvial valley fill and typically are found in both recent and relict river valleys. Some large areas of buried sand and gravel are not associated with bedrock valleys and most likely originated as ice-marginal stratified drift. Discontinuous buried sand and gravel is present in large areas throughout the basin. The surficial and buried sands and gravels are extensive in their distributions. Surficial sand and gravel is found within 10 ft of land surface, whereas buried sand and gravel is covered by more than 10 ft of nonaquifer material.

Surficial sand and gravel aquifers can be seen in sections 6A–6A′ and 6C–6C′ to 6I–6I′ (fig. 46). In all cases, the surficial aquifer is within and along the valley of the Wabash River and its tributaries. Surficial sand and gravel aquifers range from 10 to 150 ft in thickness, but, are commonly 80 to 120 ft thick. In places where an entire valley is filled from bedrock to land surface with sand and gravel (sections 6C–6C′ to 6F–6F′, fig. 46), that system is called a surficial aquifer. That entire thickness of permeable materials may not represent a single depositional unit, but because of the absence of interlayered, nonaquifer material, it functions as a single hydrogeologic unit.

Buried sand and gravel is found in various places throughout the basin, but it is most common in bedrock valleys (fig. 46). The most extensive buried sand and gravel is in the buried Lafayette Bedrock Valley System. The valley is mapped in figure 47 as "buried bedrock valley aquifer". Section 6I–6I′ (fig. 46) roughly coincides with the main axis of the bedrock valley where it exits the basin and enters Illinois. The bedrock knobs shown in the bedrock valley are a function of the 2-mile width of the hydrogeologic section and are actually bedrock highs on the interfluves between adjacent tributary bedrock valleys. More than 200 ft of buried sand and gravel are found in this part of the Lafayette Bedrock Valley System.

Some buried sand and gravel that does not appear to be basal fill in bedrock valleys can be seen along section 6B–6B′ (fig. 46), in R. 4 W. of section 6F–6F′ (fig. 46), and in several areas along section 6G–6G′ (fig. 46). In some places, multiple zones of buried sand and gravel are separated by nonaquifer material and are therefore stratigraphically and hydraulically distinct (sections 6A–6A′, 6H–6H′, and 6I–6I′, fig. 46). Buried sands and gravels that are not necessarily valley-fill deposits are large-scale intertill sands and gravels, outwash-fan deposits and ice-contact stratified drift within Wisconsinan and pre-Wisconsinan sediment sequences (Gray, 1989).

Potential well yields from the surficial sand and gravel range from 300 to 2,700 gal/min (Bechert and Heckard, 1966; Cable and others, 1971), whereas pumping rates noted on well logs that were used to construct sections range from 10 to 781 gal/min (table 8). Buried sand and gravel has a slightly lower expected range of yields, from 25 to 1,500 gal/min (Cable and others, 1971; Nyman and Pettijohn, 1971); yet, some high pumping rates from buried sand and gravels that are noted on well logs exceed pumping rates from the surficial sands. It is likely that the full production potential of the surficial and buried sand and gravel aquifers have not been realized in many cases and that each aquifer type is capable of significant yields.

The natural discharge points for the surficial sands and gravels are adjoining streams, such as the Wabash River (Pohlmann, 1986). Much ground-water flow in the thick sand and gravel sequences is lateral flow (Maarouf and Melhorn, 1975, p. 74; Pohlmann, 1986, p. 138) that follows regional flow paths. In some cases, natural discharge to buried sand and gravel and well pumpage may locally induce flow in the surficial sand and gravel away from the Wabash River and tributaries.

Recharge to surficial sands and gravels may range from 6.4 in/yr (Maarouf and Melhorn, 1975, p. 73) to 10.0 in/yr (Pohlmann, 1986, p. 138), whereas recharge to sands and gravels buried under low-permeability nonaquifer material may be less than 2 in/yr (Pohlmann, 1986, p. 138).

Discontinuous buried sand and gravel is another ground-water resource within the unconsolidated deposits. Examples of this aquifer type can be seen in virtually every hydrogeologic section; however, it is a significant source of water only in areas shown on sections 6A–6A′, 6B–6B′, 6F–6F′, 6G–6G′, and 6H–6H′ (fig. 46). The occurrence of discontinuous buried sand and gravel is similar to that of buried sand and gravel that is not related to bedrock valleys; however, it tends to be thinner and of smaller areal extent. Discontinuous stringers of sand and gravel are ubiquitous, in as much as almost 80 percent of the examined well logs note some discontinuous sand and gravel in the unconsolidated sediments. The sand and gravel deposits are 5 to 55 ft thick and are an important source of water for domestic needs. Expected yields from discontinuous buried sand and gravel range from 5 to 300 gal/min (Bechert and Heckard, 1966; Nyman and Pettijohn, 1971), whereas reported pumping rates noted on well logs range from 2 to 320 gal/min (table 8). The actual productivity of an individual well screened in a discontinuous buried sand and gravel aquifer is a function of the thickness of that unit, the storage characteristics (storativity), the hydraulic conductivity of the confining nonaquifer material, and the interconnection of the aquifer with other water-bearing sand and gravel bodies.

Most buried sand and gravel aquifers, both continuous and discontinuous, are artesian; that is, water levels in wells that are screened in buried sand and gravel are typically higher than the top of the permeable unit. One large buried sand and gravel unit that underlies the eastern part of R. 2 W. in section 6G–6G′(fig. 46) is penetrated by four wells, three of which are flowing wells. Buried sand and gravel deposits that are under artesian pressures will, in some cases, serve as sources of recharge for nearby unconsolidated aquifers, confining units, and bedrock units.

In general, discontinuous and continuous buried sands and gravels that are confined above and below by tills and other nonaquifer material have higher hydraulic heads than basal sands and shallow bedrock. These observations suggest that over most of the basin, some vertical ground-water flow is downward through the glacial cover, into shallow bedrock. In some areas however, most commonly near the Wabash River and major tributaries such as Sugar Creek, water levels in bedrock wells are higher than those in the overlying drift, and suggest vertically upward ground-water flow.

Other generalizations can be made regarding the overall distribution of sand and gravel aquifer material in unconsolidated sediments in the Middle Wabash River basin. Sands and gravels, whether they are surficial or buried, are most extensive in thick drift sequences (fig. 45). Thick drift is most common along buried bedrock valleys, the present valley of the Wabash River, and north of the Wisconsinan glacial boundary in the Tipton Till Plain (fig. 43). The overall lack of sand and gravel aquifers in the southern part of the basin and away from bedrock and river valleys can best be seen along sections 6A–6A′ and 6E–6E′ (fig. 46).

Bedrock Aquifers

At least four distinct aquifer types have been defined in different bedrock units within the Middle Wabash River basin. These bedrock aquifers include complexly interbedded sandstones, shales, limestones, and coals; sandstone; Mississippian and Silurian-Devonian carbonate rocks; and an upper weathered zone in low permeability rocks (table 8). Because the bedrock surface and the Mississippian-Pennsylvanian unconformity truncate bedrock units of different ages (fig. 44), not all bedrock aquifers are laterally continuous. Moreover, aquifers in the upper weathered bedrock and, in some cases, in the complexly interbedded Pennsylvanian bedrock, are not present throughout the stratigraphic unit but simply represent water-bearing horizons within these units (table 8). Where an aquifer does not include the entire stratigraphic unit, the interpretations of the basal or lateral boundaries of the aquifer are based on well-completion depths and locations of lithologic change.

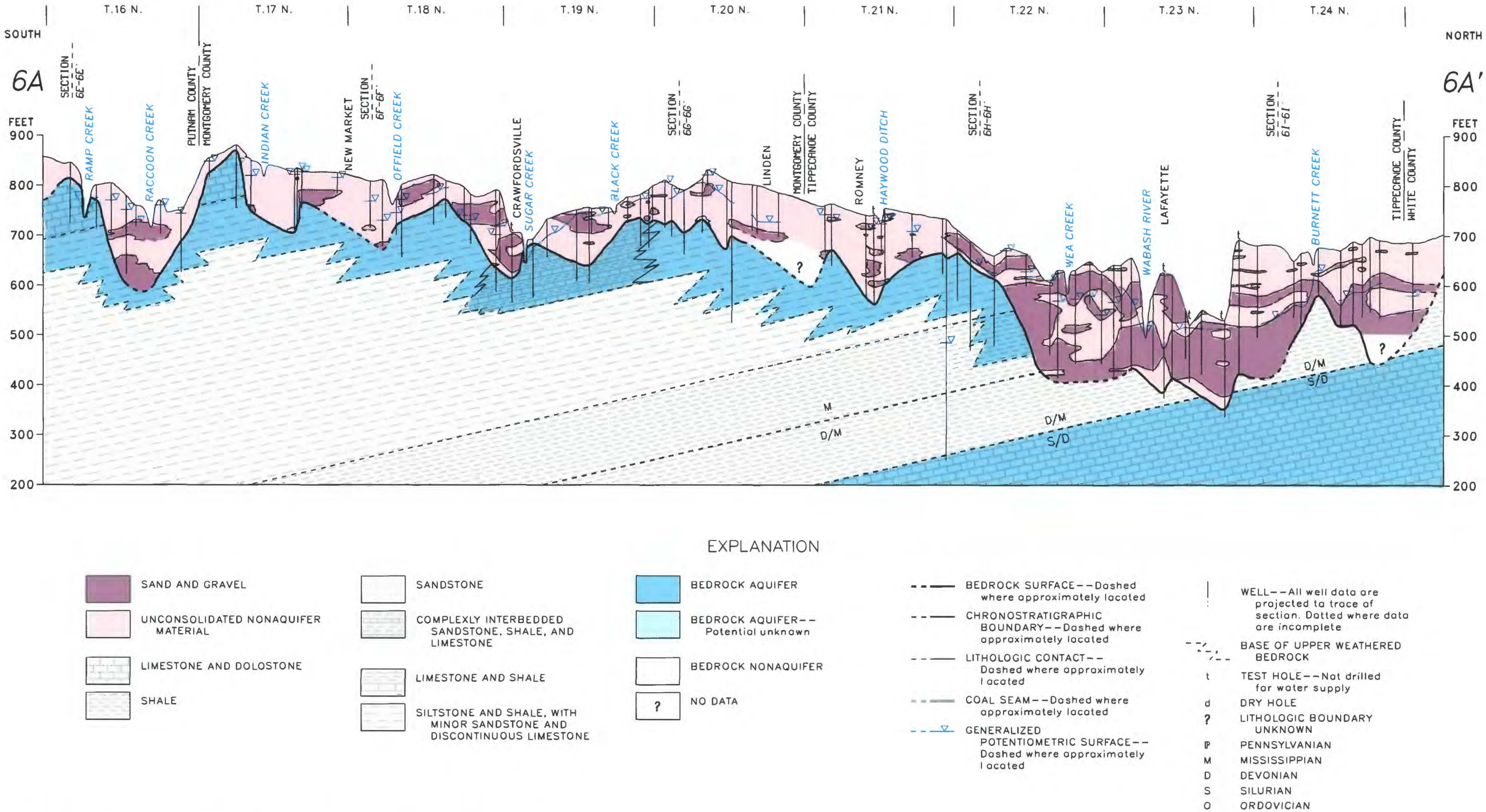
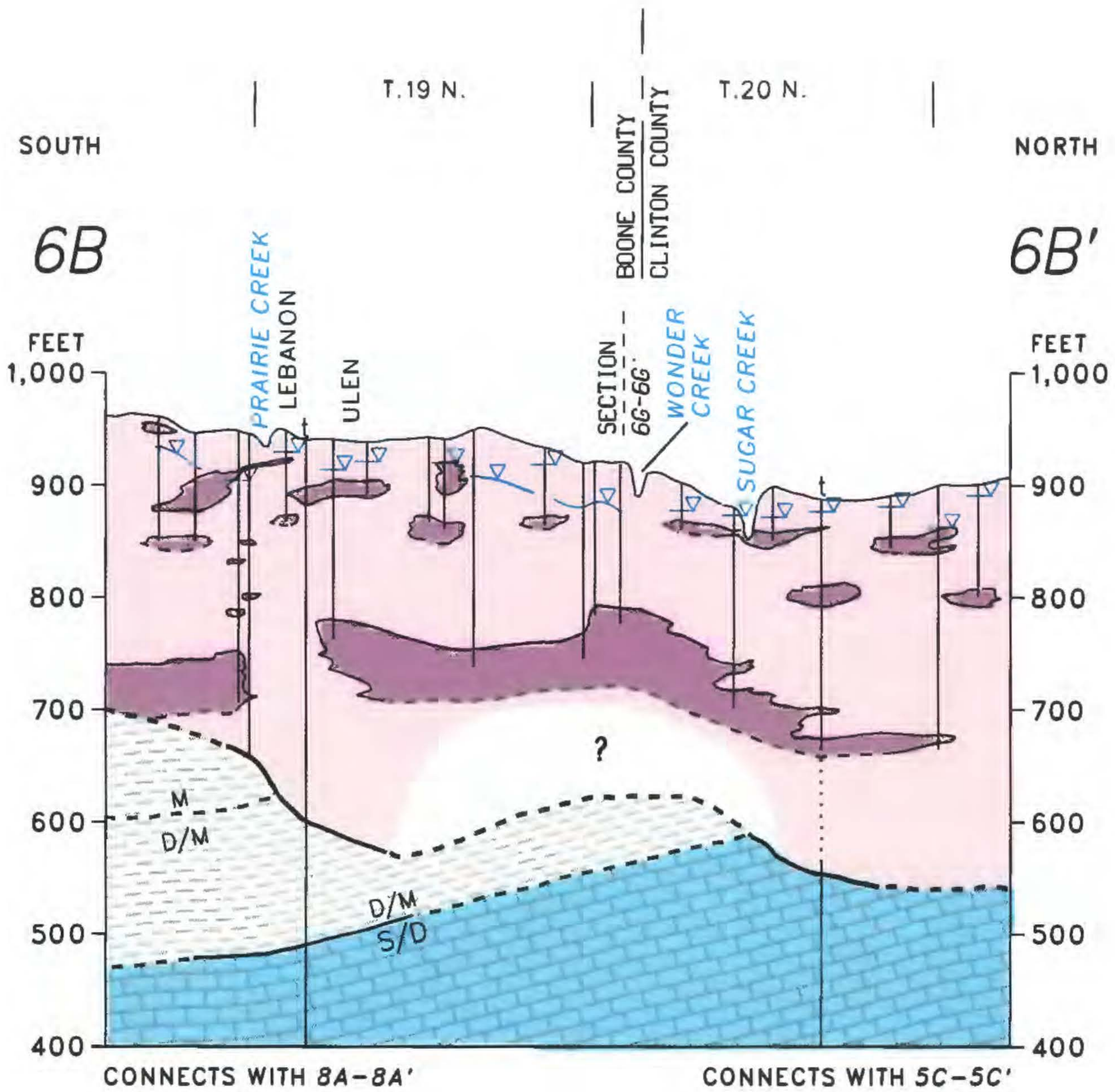
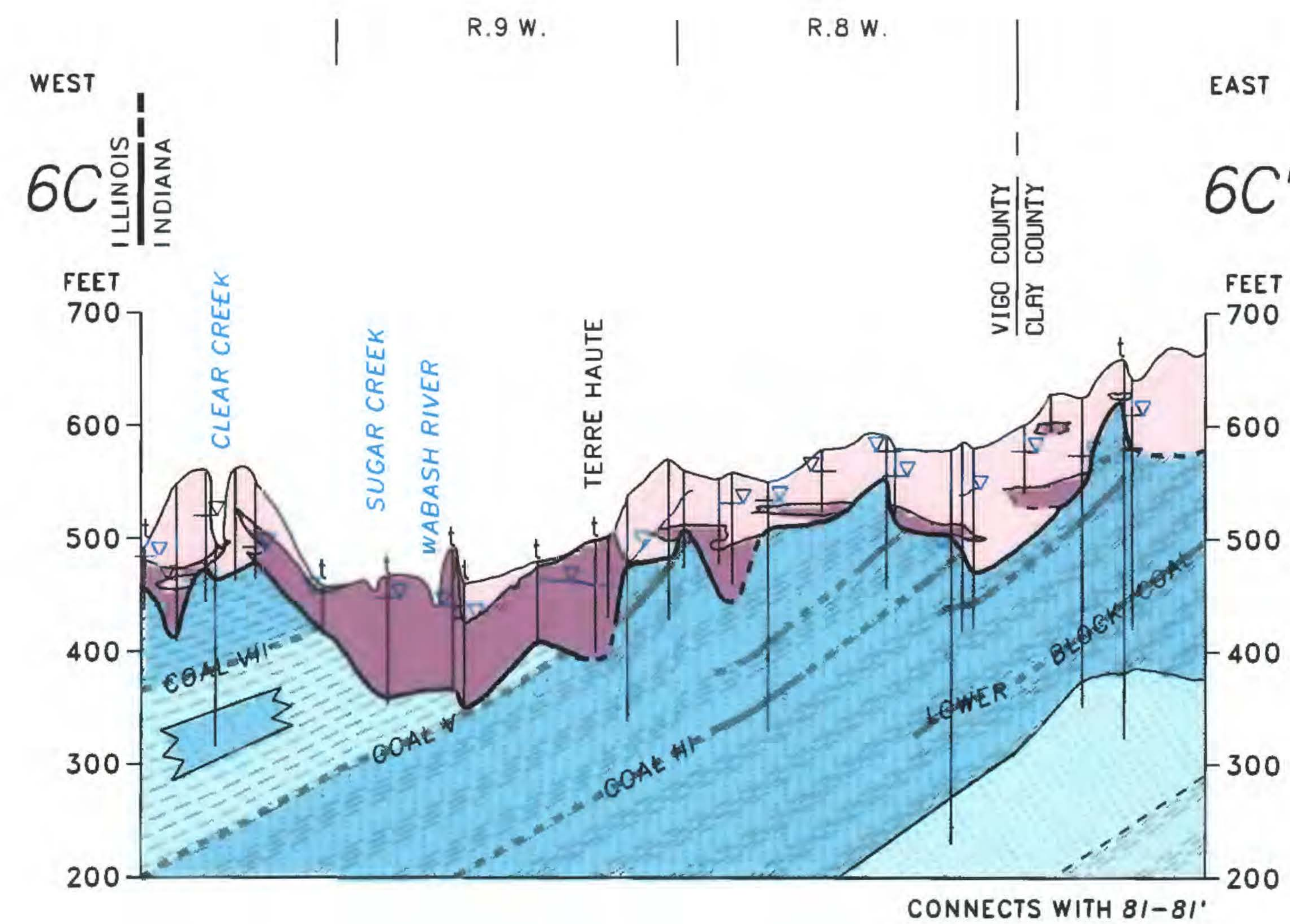
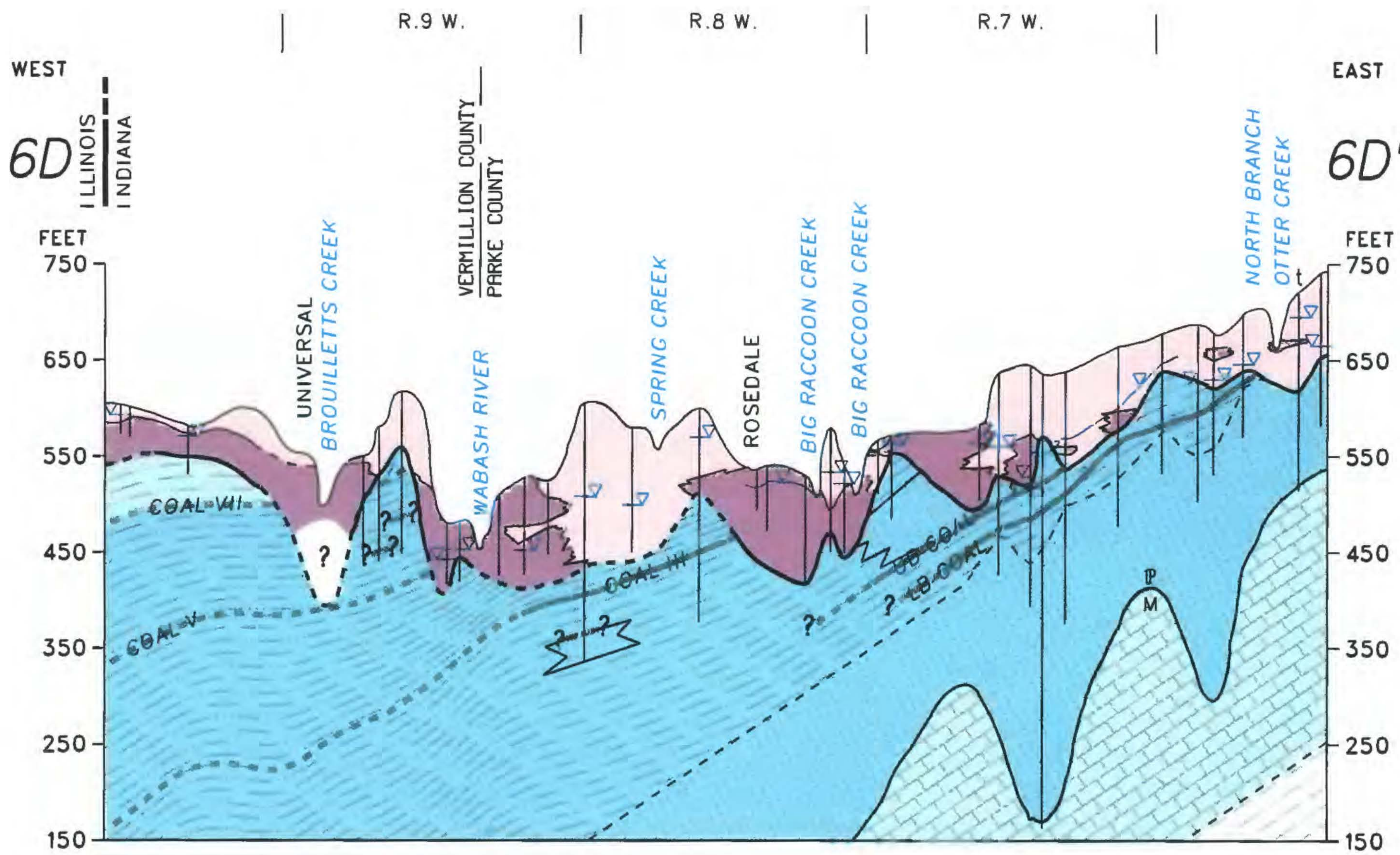


Figure 46. Hydrogeologic sections *6A–6A'* to *6I–6I'* of the Middle Wabash River basin.







DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED

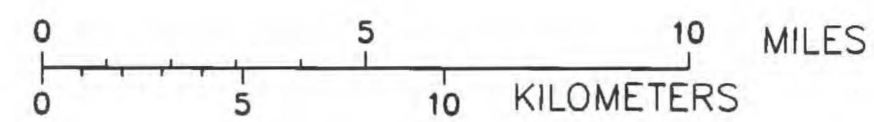
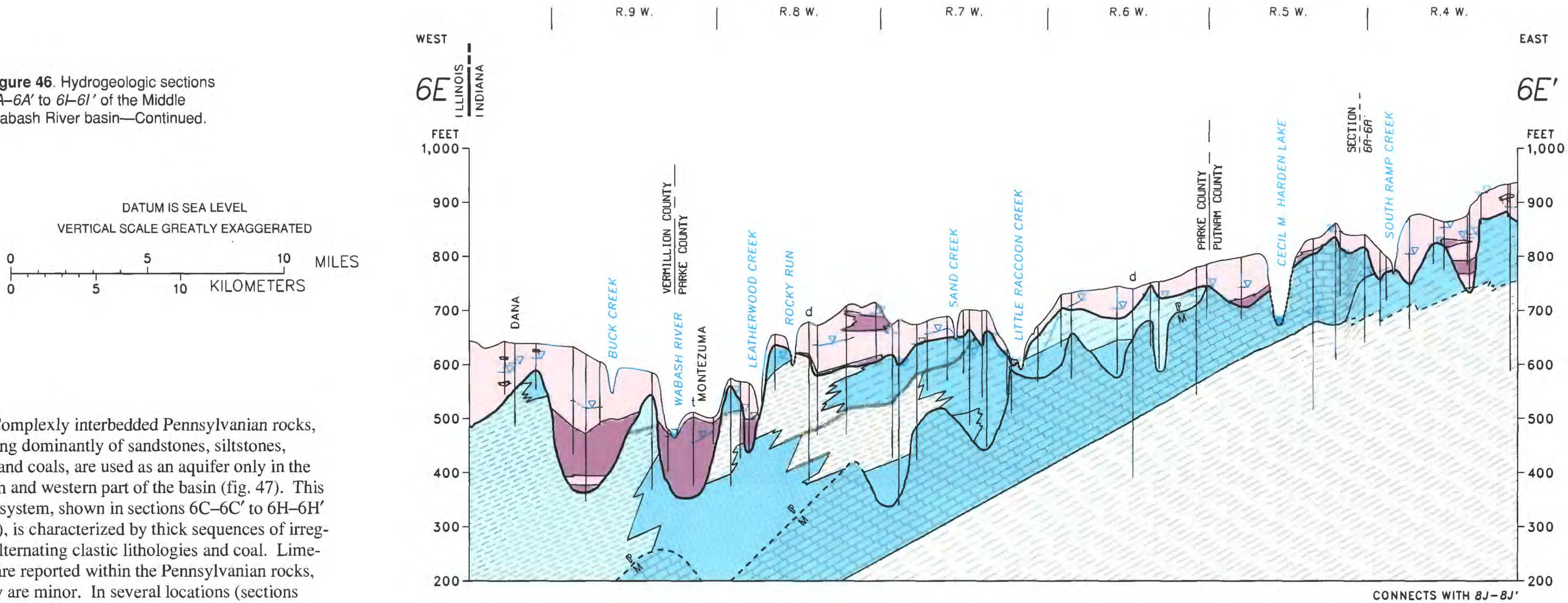


Figure 46. Hydrogeologic sections *6A–6A'* to *6I–6I'* of the Middle Wabash River basin—Continued.



Complexly interbedded Pennsylvanian rocks, consisting dominantly of sandstones, siltstones, shales, and coals, are used as an aquifer only in the southern and western part of the basin (fig. 47). This aquifer system, shown in sections 6C–6C′ to 6H–6H′ (fig. 46), is characterized by thick sequences of irregularly alternating clastic lithologies and coal. Limestones are reported within the Pennsylvanian rocks, but they are minor. In several locations (sections 6C–6C′ and 6D–6D′, fig. 46), major coals were identifiable in the subsurface with the aid of published coal maps (Powell, 1968; Hutchison, 1961; Wier, 1952). The coals are important as stratigraphic markers; in addition, they may influence water chemistry (Smith and Krothe, 1983, p. 24), and they can be the dominant source of water for some uncased wells that are open to the entire complexly interbedded sequence (Banaszak, 1980).

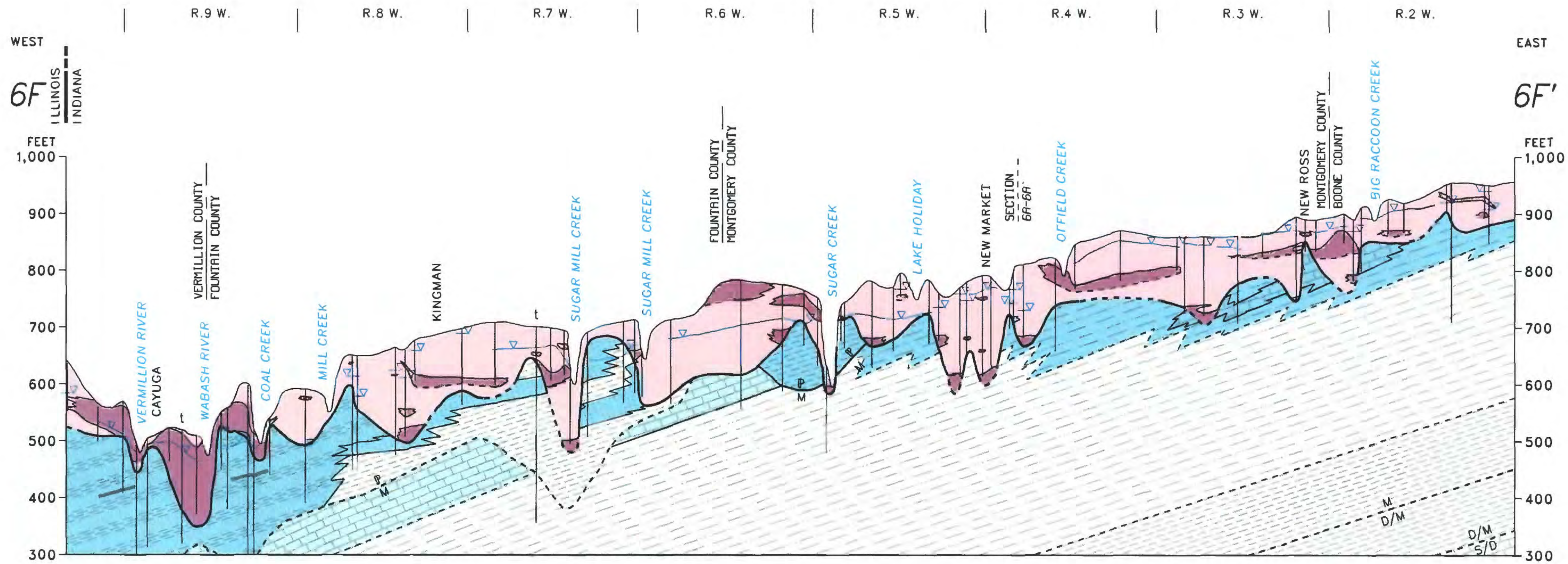
In almost all cases where wells tap the complexly interbedded aquifer, it is impossible to determine the exact source of water. Wells are commonly drilled several hundred feet into the bedrock and left uncased for the entire thickness of rocks penetrated. The interpreted thickness of the complexly interbedded aquifer ranges from 30 to 375 ft; however, those estimates are based on well penetrations and could be modified with additional data (table 8).

Expected yields from these Pennsylvanian interbedded bedrock units range from 3 to 70 gal/min (Cable and others, 1971; Nyman and Pettijohn, 1971); however reported pumping rates range from 0.1 to 35 gal/min (table 8). Smith and Krothe (1983) suggest that coal units in Vigo and Clay Counties have higher transmissivities and specific capacities than other water-bearing units in the same lithostratigraphic sequence. Drawdown during pumping at most wells open through the complexly interbedded bedrock is large (as much as a few hundred feet).

In several locations, individual lithologies were interpreted within the Pennsylvanian rocks. Shale and sandstone units are defined within the interbedded sequence, and are best represented in R. 7 W. and R. 8 W. along sections 6E–6E′ and 6F–6F′ (fig. 46). One well that penetrated a thick shale unit was noted as a dry hole. Some sandstone units can provide a supply of water and are discussed as a separate aquifer.

Many surface and underground coal mines have been developed in the coal-bearing, complexly interbedded strata of southwestern Indiana. Mining-related activities have affected the hydrogeology of the complexly interbedded aquifer and adjacent aquifers. Banaszak (1980, p. 240) suggests that cast overburden piles have significant water-storage capability and can increase the amount of infiltration and recharge to underlying coal aquifers. However, overburden piles that provide recharge to underlying aquifers can be detrimental to water quality. There also are abandoned coal-mine shafts and rooms in the subsurface of the coal-mining region. Many water wells drilled along sections 6C–6C′, 6D–6D′, and 6E–6E′ intercepted open mine rooms at various depths. In at least one location, water from saturated rocks above an open mine was discharged into the open room as soon as it was encountered.

The sandstones defined as a distinct aquifer in the Middle Wabash River basin are generally basal Pennsylvanian rocks of the Mansfield Formation. Most of these sandstones immediately overlie the Mississippian-Pennsylvanian unconformity and are depicted in sections 6D–6D′, 6E–6E′, and 6G–6G′ (fig. 46). Smaller sandstone units are noted above the unconformity in sections 6C–6C′, 6D–6D′, 6E–6E′, and 6F–6F′ (fig. 46). Basal sandstones, many of which contain conglomerates near their bases, typically fill "valleys" in the Mississippian-Pennsylvanian unconformity and originated as channel-fill and fluvial sands (section 6E–6E′, fig. 46).



The thickness of the sandstone aquifer was interpreted from the depth of well penetrations and ranges from 30 to 250 ft (table 8). The sandstones are laterally discontinuous and bounded on all sides by rocks of different lithology and(or) the underlying unconformity.

Expected well yields for Pennsylvanian sandstones range from 8 to 75 gal/min (Cable and others, 1971; Arihood and Mackenzie, 1983) and reported pumping rates range from 1 to 40 gal/min (table 8).

Recharge to the sandstone and the complexly interbedded aquifers is from precipitation on outcrop areas and downward percolation through overlying unconsolidated material (Cable and others, 1971, p. 5; Arihood and Mackenzie, 1983, p. 54). Recharge to the Mansfield Formation in the vicinity of Terre Haute was estimated on the basis of several assumptions and a simple water balance equation to be 1.64 in/yr (Thomas, 1980, p. 29).

Most of the sandstones rest on an erosional surface formed prior to and during the deposition of Mansfield sands; therefore, it is likely that the basal sandstones and conglomerates have a good hydraulic connection with the underlying Mississippian limestones. The erosional surface represents a period of weathering that most likely developed a zone of enhanced permeability at the surface of underlying carbonate rocks that is now in contact with the overlying clastic rocks. Sufficient data are not available to determine the degree or nature of the hydraulic connection between the sandstones and the limestones, and whether water is discharged from the limestones into the sandstones or vice versa.

The Mississippian carbonate bedrock aquifer is composed of rocks of the Mississippian Blue River and Sanders Groups. These carbonate rocks are truncated by the Mississippian-Pennsylvanian unconformity and are found only in the southeastern and south-central part of the basin (fig. 47). Rocks of the Blue River and Sanders Groups are seen on sections 6A–6A′ and 6D–6D′ to 6F–6F′ (fig. 46). The only known usage of these carbonate rocks as an aquifer, however, is in areas shown on sections 6A–6A′ and 6E–6E′. The carbonate rocks in these areas are 25 to 225 ft thick.

Expected yields for the Mississippian carbonate bedrock aquifer range from 5 to 60 gal/min (Bechert and Heckard, 1966; Cable and Robison, 1974); reported pumping rates range from 1 to 30 gal/min (table 8). The water-producing characteristics of these carbonate rocks are highly variable, with some dry holes occurring in the Mississippian carbonate rocks (see section 6E–6E′, fig. 46).

In general, well yields and permeability in carbonate bedrock aquifers can be highly variable (Siddiqui and Parizek, 1971; Lattman and Parizek, 1964). Variable yields result from an uneven distribution of permeability that develops from preferential dissolution of carbonate along fractures, joints, and bedding planes (Legrand and Stringfield, 1971; White, 1969). Statistical analyses have indicated a direct relation between well locations on fractures and fracture traces and a higher productivity of the well (Siddiqui and Parizek, 1971). Clearly, wells need to be in connection with water-bearing zones to maximize potential yields. In contrast, wells sited in nonfracture locations may not provide sufficient ground water for their intended use.

Another carbonate bedrock aquifer, the Silurian-Devonian aquifer, is present in the Middle Wabash River basin and shown in sections 6A–6A′, 6B–6B′, and 6F–6F′ to 6I–6I′ (fig. 46). The Silurian-Devonian carbonate bedrock aquifer is an important aquifer in the State; however, only three wells used to construct the sections in the Middle Wabash River basin were identified as producing water from the

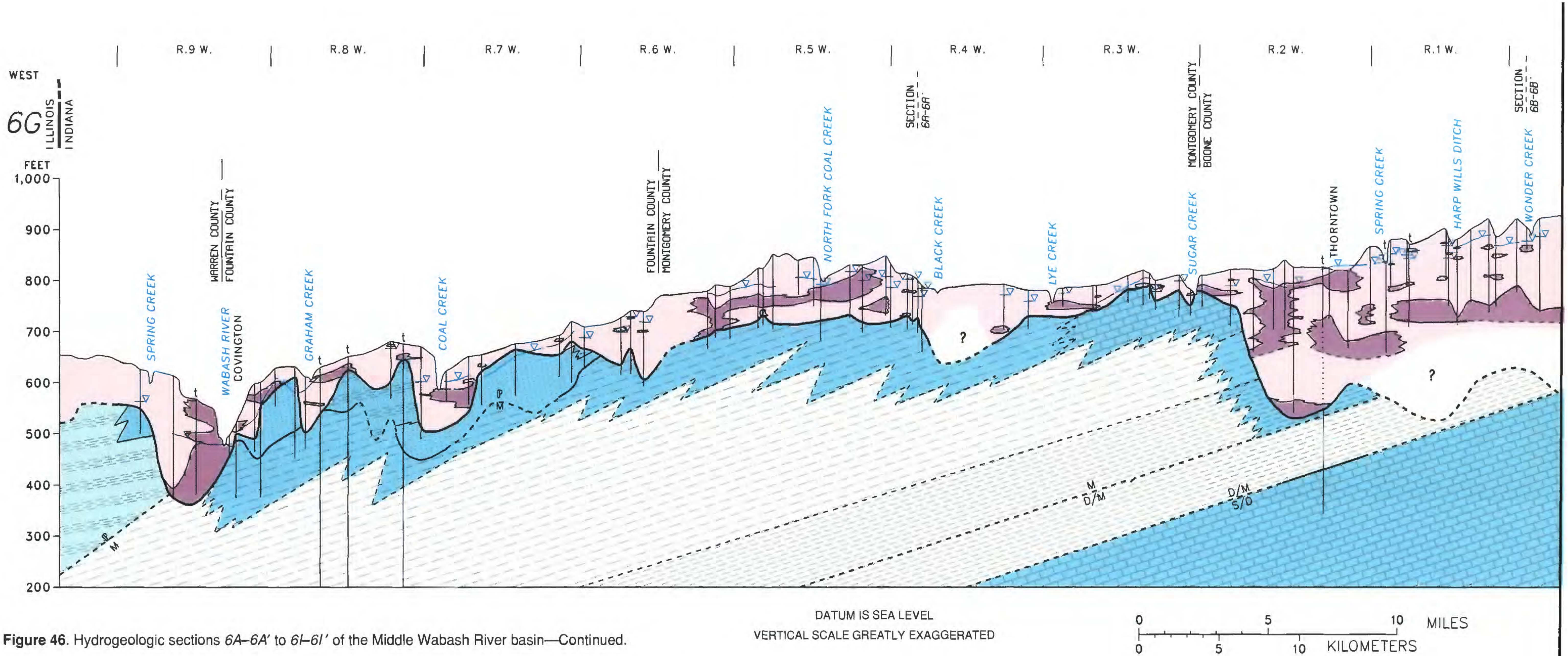


Figure 46. Hydrogeologic sections *6A–6A′* to *6I–6I′* of the Middle Wabash River basin—Continued.

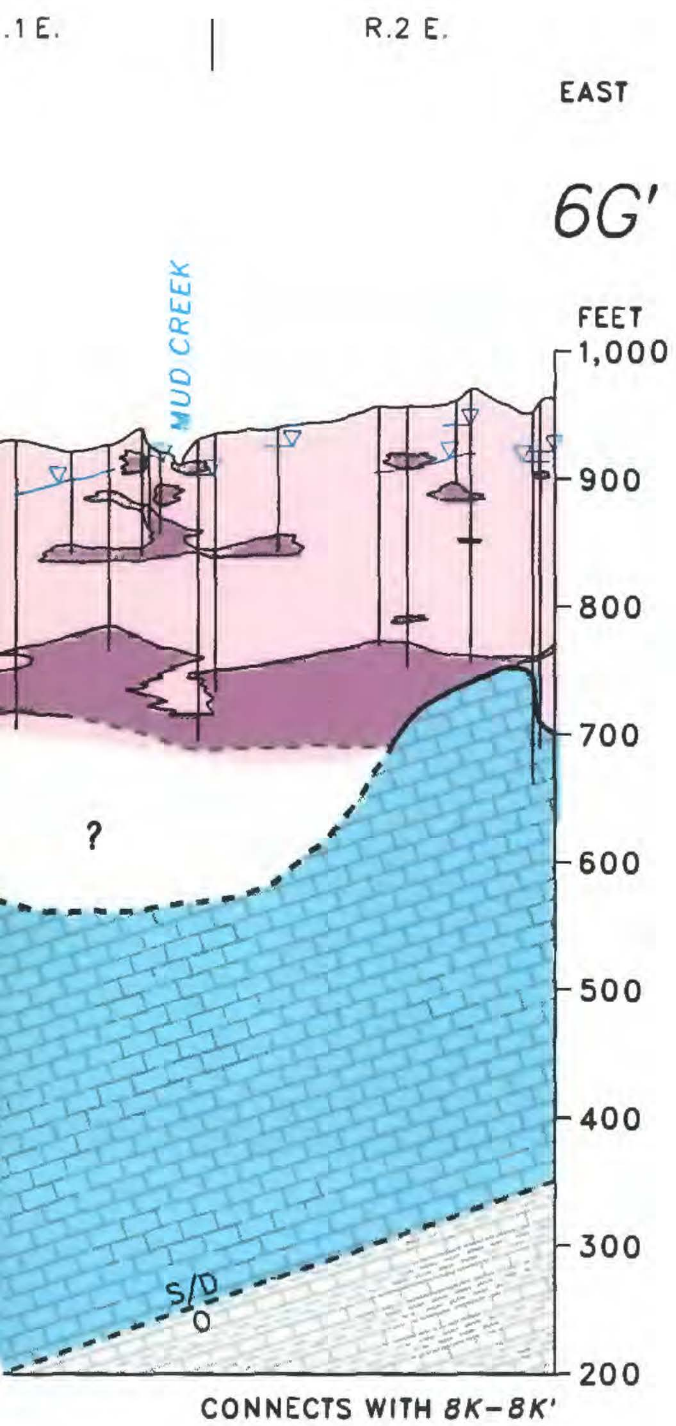
aquifer (two domestic wells in section 6G–6G′ and one industrial well in section 6I–6I′). The aquifer is not significant in the basin because of its considerable depth, the presence of highly mineralized water at depth, and the availability of water from shallower bedrock aquifers and overlying sand and gravel. Characteristics of the aquifer are listed in table 8. Detailed information on the Silurian-Devonian carbonate bedrock aquifer is provided in the sections on the Lake Michigan and Kankakee River basins in this report.

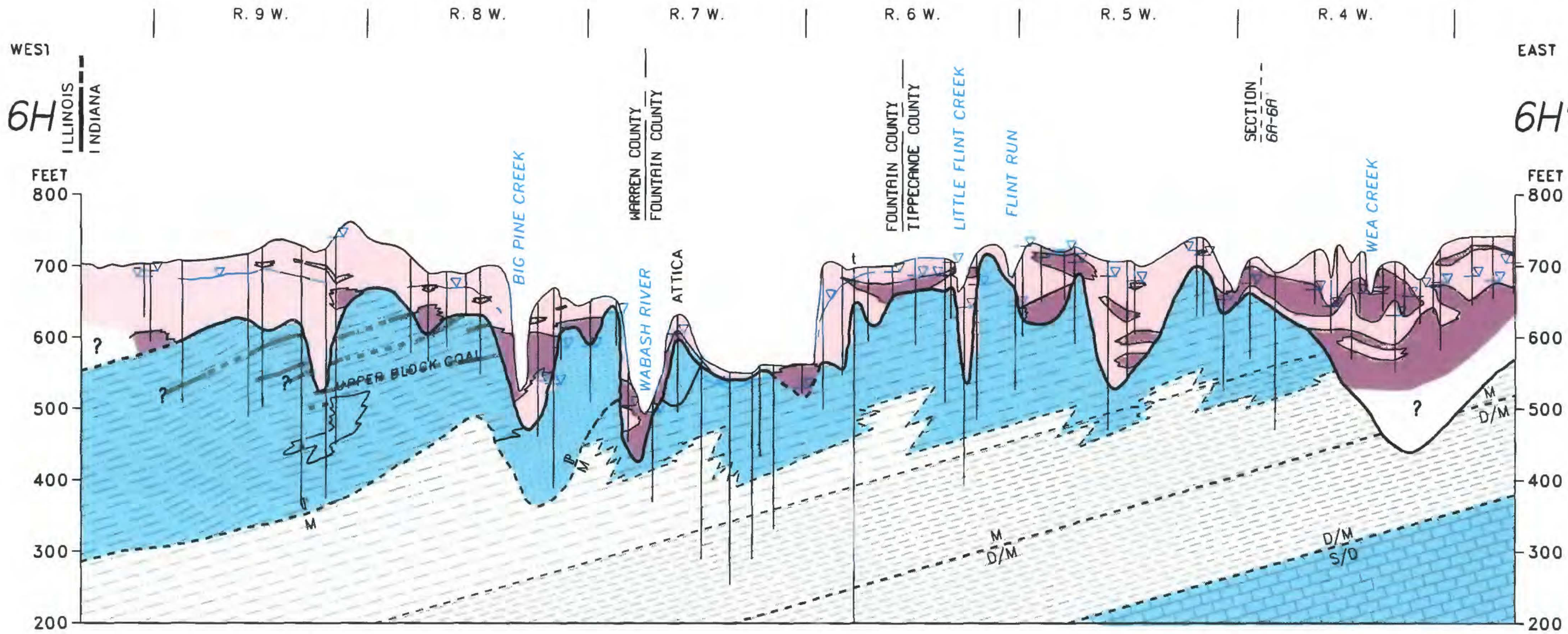
An upper weathered zone in low-permeability bedrock is another significant bedrock aquifer in the Middle Wabash River basin. The aquifer is primarily at and immediately below the bedrock surface in Mississippian Borden Group rocks. The aquifer is significant only in areas underlain by typically nonaquifer bedrock (Borden Group), however, it is likely that some of the characteristics of the upper weathered-bedrock aquifer are present in all units at the bedrock surface. The upper weathered zone formed in the bedrock during various weathering processes before and during deposition of the overlying drift. The aquifer characteristics result from the enhanced permeability in the weathered zone.

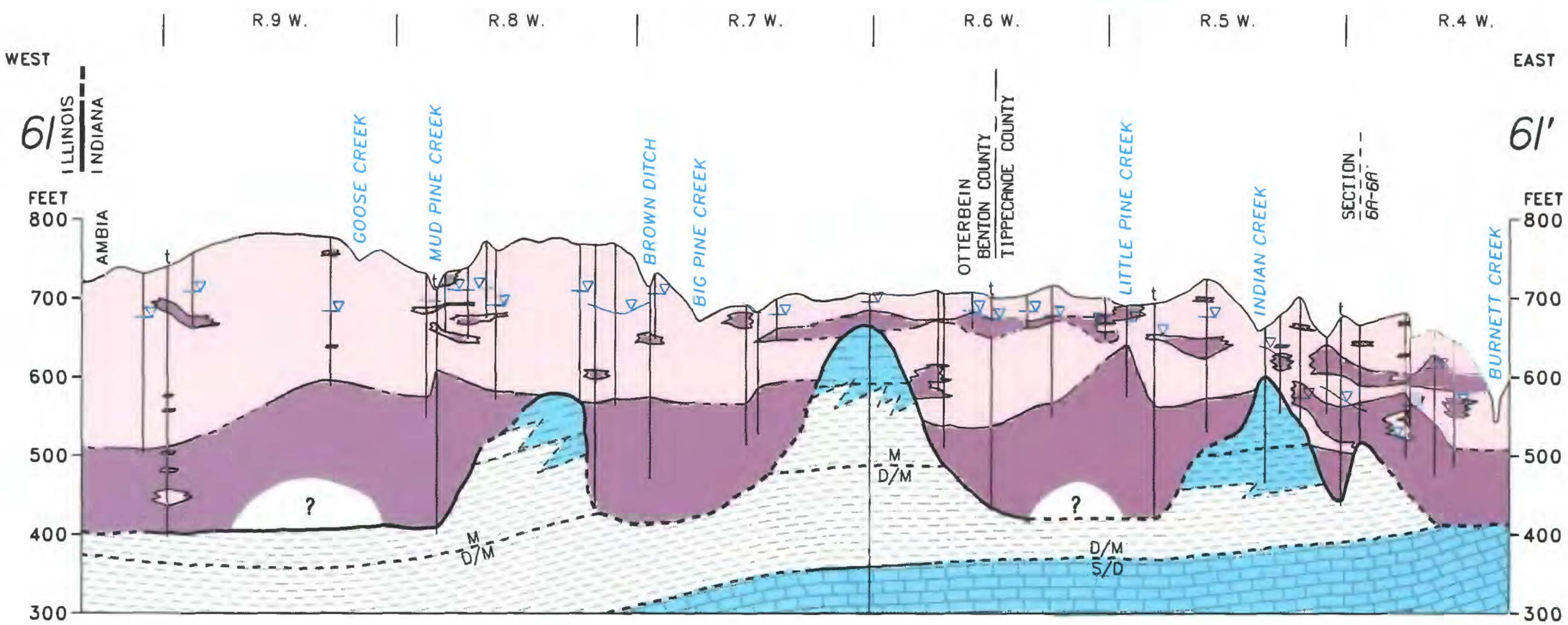
The upper weathered-bedrock aquifer is areally extensive (fig. 47), and is most significant in the thick Borden Group rocks shown in sections 6A–6A′ and sections 6F–6F′ to 6H–6H′ (fig. 46). Small areas have also been inferred in section 6I–6I′ (fig. 46). This aquifer, by its nature, is not restricted by lithologic boundaries. Rather its existence is entirely a function of weathering at and below the bedrock surface. Lithologic variations are shown within the upper weathered-bedrock aquifer along sections 6A–6A′, 6F–6F′ and 6G–6G′ (fig. 46) and produce spatial variability in hydraulic properties of the aquifer.

The true thickness of the upper weathered bedrock is unknown, but it has been inferred from the total penetration of wells that use it as an aquifer. Suggested thicknesses of the upper weathered-bedrock aquifer range from 25 to 175 ft (table 8), but the true enhanced-permeability zone is most likely limited to approximately the upper 50 ft.

Expected yields for Borden Group rocks in the upper weathered zone range from 1 to 270 gal/min (Cable and Robison, 1974; Wangsness and others, 1983) and up to 590 gal/min from siltstones and shales (Nyman and Pettijohn, 1971, p. 61); however, these yields are far greater than those noted on logs used to construct the sections. Reported pumping rates for the upper weathered-bedrock aquifer range from 2 to 150 gal/min (table 8). Some of the variability in yields from this aquifer are attributable to the observed lithologic variations in the Borden Group, the variability in







degree of weathering and fracturing, and the hydrogeologic character of the overlying drift.

Recharge to the upper weathered-bedrock aquifer is through the overlying drift. There is most likely a good hydraulic connection between rocks of the Borden Group and basal sands and gravels in buried bedrock valleys and buried sand and gravels on some areas of bedrock uplands (sections 6A–6A′ and 6G–6G′ to 6I–6I′, fig. 46). Most of the Borden Group rocks underlying the Lafayette Bedrock Valley along section 6I–6I′ are not interpreted as being an aquifer because no wells are completed in it and because sufficient ground-water resources are available in sand and gravel aquifers that overlie the bedrock.

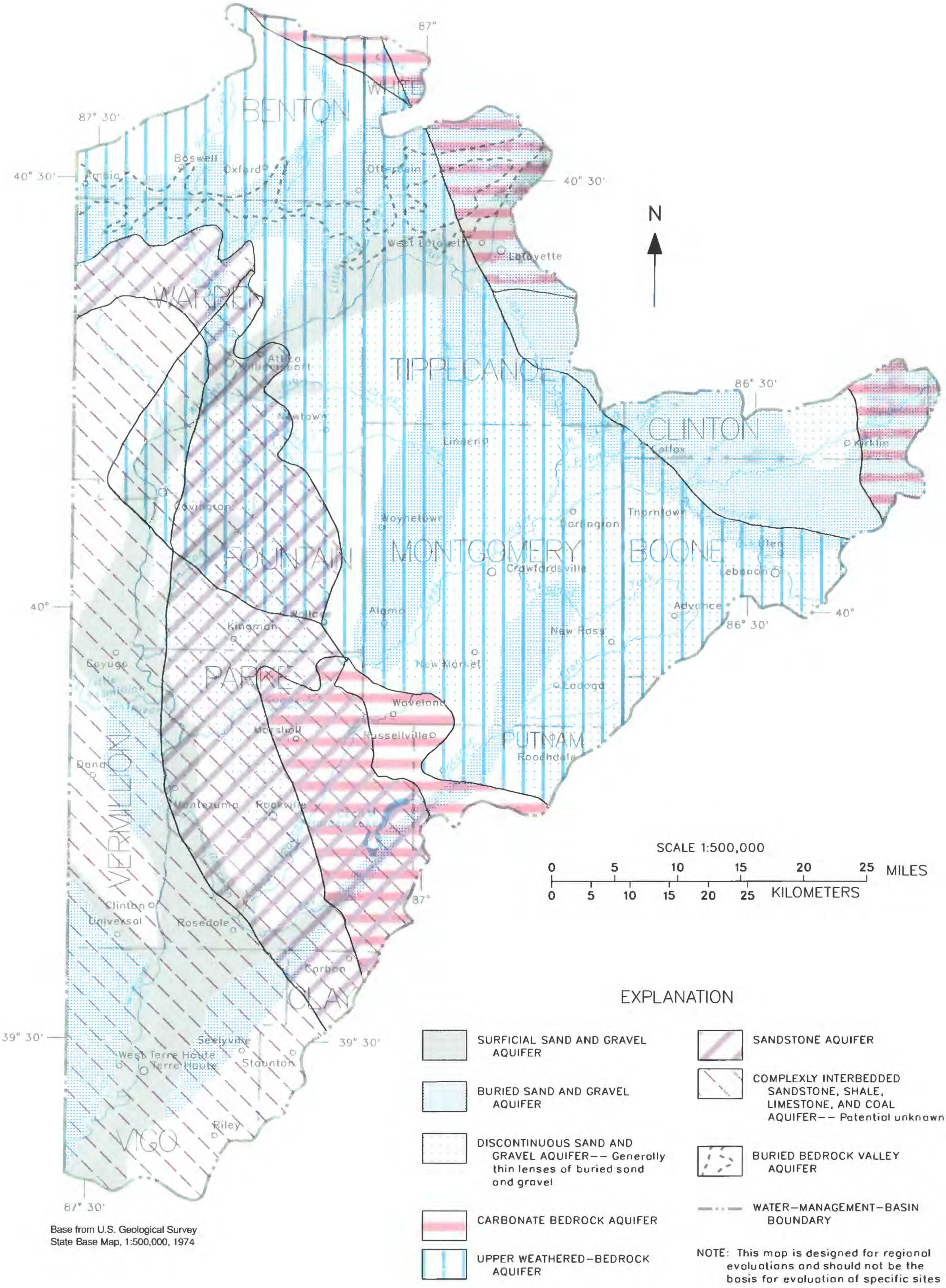


Figure 47. Extent of aquifer types in the Middle Wabash River basin.

Table 8. Characteristics of aquifer types in the Middle Wabash River basin
[<, less than; locations of aquifer types shown in fig. 47]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Reported[1] pump rates (gallons per minute)	Common name(s)
Surficial sand and gravel	10-150	[2,3]300-2,700	10- 781	Outwash, valley train
Buried sand and gravel	10-225	[3,4]25-1,500	5-1,012	Valley fill
Discontinuous sand and gravel	5- 55	[2,4]5- 300	2- 320	Intertill sands
Complexly interbedded sandstone, shale, limestone, and coal	[5]30-375	[3,4]3- 70	0.1- 35	Raccoon Creek Group, Carbondale Group
Sandstone	[5]30-250	[3,6]8- 75	1- 40	Mansfield Formation
Carbonate bedrock				
Mississippian	25-225	[2,7]5- 60	10- 30	Sanders Group Blue River Group
Silurian-Devonian	[5]20- 90	[2]<500	50- 800	Wabash Formation
Upper weathered bedrock	[5]25-175	[6,7]1- 270	2- 150	Borden Group

[1]Reported pump rates refer to noted pump rates at the time of well installation and may not represent a "true" well-yield.
[2]Bechert and Heckard, 1966.
[3]Cable and others, 1971.
[4]Nyman and Pettijohn, 1971.
[5]Reported thicknesses refer to observed and inferred permeable zones and may not correlate with specific formation thicknesses.
[6]Arihood and Mackenzie, 1983.
[7]Cable and Robison, 1974.

Summary

Ground water is available throughout the Middle Wabash River basin from unconsolidated and bedrock deposits; however, yields that tap these deposits differ greatly among aquifer types. The most productive aquifers and among the most extensive are the thick sands and gravels within buried bedrock valleys and along present river courses. These deposits currently yield 1,000 gal/min or more to individual wells and are capable of yielding greater than 2,500 gal/min. Discontinuous sands and gravels are also important and yield greater than 300 gal/min. Perhaps the next most important aquifer, simply because of its areal extent, is the upper weathered-bedrock aquifer. This aquifer supplies a large part of the domestic water requirement for the northern and eastern two-thirds of the basin. Pumping rates for wells in the upper weathered bedrock are as high as 150 gal/min. Sandstones and complexly interbedded sandstones, shales, limestones, and coals, of Pennsylvanian age, provide for the domestic water needs in the southwestern one-third of the basin and yield as much as 75 gal/min. The Mississippian carbonate bedrock aquifer is unpredictable because of permeability differences common to carbonate rocks. Dry holes are noted in these rocks, whereas some wells yield as much as 30 gal/min. Although the Silurian-Devonian carbonate bedrock aquifer is an important aquifer in the State, it is rarely used in the Middle Wabash River basin, generally because of its depth.

References Cited

Arihood, L.D., and Mackenzie, A.L., 1983, Ground water quantity, *in* Wangsness, D.J., and others, Hydrology of Area 30, eastern region, Interior coal province, Illinois and Indiana: U.S. Geological Survey, Water-Resources Investigations Open-File Report 82-1005, p. 52-57.

Arvin, D.V., 1989, Statistical summary of streamflow data for Indiana: U.S. Geological Survey Open-File Report 89-62, 964 p.

Banaszak, K.J., 1980, Coal as aquifers in the Eastern United States, *in* Proceedings of the Symposium on Surface Mining Hydrology, Sedimentology, and Reclamation: University of Kentucky, Lexington, Kentucky, December 1-5, 1980, p. 235-241.

Bechert, C.H., and Heckard, J.M., 1966, Ground water, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 100-115.

Bleuer, N.K., 1974, Distribution and significance of some ice-disintegration features in west-central Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 8, 11 p.

_____1975, The Stone Creek Section—a historical key to the glacial stratigraphy of west-central Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 11, 9 p.

_____1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley System (so-called Teays Valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

_____1991, The Lafayette Bedrock Valley System of Indiana—concept, form, and fill stratigraphy, *in* Melhorn, W.N. and Kempton, J.P., eds., Geology and hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 51-77.

Bleuer, N.K., Melhorn, W.N., Steen, W.J. and Bruns, T.M., 1991, Aquifer systems of the buried Marion-Mahomet trunk valley (Lafayette Bedrock Valley System) of Indiana, *in* Melhorn, W.N. and Kempton, J.P., eds., Geology and Hydrogeology of the Teays-Mahomet Bedrock Valley System: Geological Society of America Special Paper 258, p. 79-89.

Brown, E.A., 1949, Ground-water resources of Boone County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 4, 152 p.

Bruns, T.M., Logan, S.M., and Steen, W.J., 1985, Map showing bedrock topography of the Teays Valley, western part, north-central Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 42, scale 1:100,000.

Cable, L.W. and Robison, T.M., 1974, Ground-water resources of Montgomery County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 36, 20 p.

Cable, L.W., Watkins, F.A., Jr., and Robison, T.M., 1971, Hydrogeology of the principal aquifers in Vigo and Clay Counties, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 34, 34 p.

Gray, H.H., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Indiana: U.S. Geological Survey Professional Paper 1110-K, 20 p.

_____1982, Map of Indiana showing topography of the bedrock surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 36, scale 1:500,000.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geology map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hutchison, H.C., 1961, Distribution, structure, and mined areas of coals in Fountain and Warren Counties and the northernmost part of Vermillion County, Indiana: Indiana Department of Conservation, Preliminary Coal Map 9, scale 1 inch equals approximately 2 miles.

_____1976, Geology of the Catlin-Mansfield area, Parke and Putnam Counties, Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 54, 57 p.

Keller, S.J., 1990, Maps of southwestern Indiana showing geology and elevation of the sub-Pennsylvanian surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 51, scale 1:380,160.

Lattman, L.H., and Parizek, R.R., 1964, Relationship between fracture traces and the occurrence of ground water in carbonate rocks: Journal of Hydrology, v. 2, p. 73-91.

Legrand, H.E., and Stringfield, V.T., 1971, Development and distribution of permeability in carbonate aquifers: Water Resources Research, v. 7, no. 5, p. 1284-1294.

Loganathan, G.V., Delleur, J.W., and Talavage, J.J., 1980, Planning groundwater supply systems for urban growth—application to West Lafayette, Indiana: Purdue University Water Resources Research Center Technical Report 131, 75 p.

Maarouf, A.M.S., and Melhorn, W.N., 1975, Hydrogeology of glacial deposits in Tippecanoe County, Indiana: Purdue University, Water Resources Research Center Technical Report 61, 107 p.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 20, part 2, p. 59-256.

Marie, J.R., and Davis, L.G., 1974, Water resources of the Middle Wabash River basin, Indiana: U.S. Geological Survey Hydrologic Investigations Atlas HA-508, 3 sheets.

Nyman, D.J., and Pettijohn, R.A., 1971, Ground water, *in* Wabash River Coordinating Committee, Wabash River basin comprehensive study: U.S. Army Corps of Engineers, Louisville Ky., volume VIII, appendix G, 100 p.

Pohlmann, K.F., 1986, Hydrogeology of Quaternary deposits in the West Lafayette, Indiana area—preliminary assessment, *in* Lardner, J.P., ed, Seventh Annual Water Resources Symposium, June 4-6, 1986, Angola, Ind., Proceedings: Indiana Water Resources Association, p. 137-138.

Powell, R.L., 1968, Distribution, structure and mined areas of coals in Parke County and southern Vermillion County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 13, scale 1 inch equals approximately 2 miles.

Rosenshein, J.S., 1955, Interesting aspects of the conglomerate-sandstone phase of the glacial deposits of Tippecanoe County, Indiana, *in* Moulton, B., ed, Proceedings of the Indiana Academy of Science, 1954: v. 64, p. 176-177.

_____1958, Ground-water resources of Tippecanoe County, Indiana: Indiana Department of Conservation, Division of Water Resources Bulletin 8, 37 p.

Rosenshein, J.S. and Cosner, O.J., 1956, Ground-water resources of Tippecanoe County, Indiana—appendix, basic data: Indiana Department of Conservation, Division of Water Resources Bulletin 8, 65 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Siddiqui, S.H., and Parizek, R.R., 1971, Hydrogeologic factors influencing well yields in folded and faulted carbonate rocks in central Pennsylvania: Water Resources Research, v. 7, no. 5, p. 1295-1312.

Smith, C.R., and Krothe, N.C., 1983, Hydrogeology of the Carbondale and Raccoon Creek Groups, Pennsylvanian System, Vigo, Clay, and Sullivan Counties, Indiana: Indiana University Water Resources Research Center, 93 p.

Steen, W.J., and Uhl, J.E., 1959, Preliminary report on ground water conditions in the vicinity of New Goshen, Vigo County: Indiana Department of Conservation, Division of Water Resources, 29 p.

Stockdale, P.B., 1931, The Borden (Knobstone) rocks of southern Indiana: Indiana Department of Conservation Publication 98, 330 p.

Thomas, J.M., 1980, Water chemistry and flow system of the early Pennsylvanian Mansfield Formation Aquifer System in Clay County, Indiana: Bloomington, Ind., Indiana University, Master's thesis, 76 p.

Thompson, R.E., Jr., and Nell, G.E., 1990, Water resources data, Indiana, water year 1989: U.S. Geological Survey Water-Data Report IN-89-1, 336 p.

Visocky, A.P., and Schicht, R.J., 1969, Groundwater resources of the buried Mahomet bedrock valley: Illinois State Water Survey Report of Investigation 62, 52 p.

Wabash River Coordinating Committee, 1971, Wabash River basin comprehensive study: Louisville, Ky., U.S. Army Corps of Engineers, 14 v., variable pagination.

Wangsness, D.J., and others, 1983, Hydrology of area 30, eastern region, interior coal province, Illinois and Indiana: U.S. Geological Survey, Water-Resources Investigations Open-File Report 82-1005, 82 p.

Watkins, F.A., Jr., and Jordan, D.G., 1963, Ground-water resources of west-central Indiana—preliminary report, Vigo County: Indiana Department of Conservation, Division of Water Bulletin 17, 358 p.

_____1964a, Ground-water resources of west-central Indiana—preliminary report, Putnam County: Indiana Department of Conservation, Division of Water Bulletin 21, 83 p.

_____1964b, Ground-water resources of west-central Indiana—preliminary report, Parke County: Indiana Department of Conservation, Division of Water Bulletin 23, 125 p.

_____1965a, Ground-water resources of west-central Indiana—preliminary report, Montgomery County: Indiana Department of Natural Resources, Division of Water Bulletin 27, 107 p.

_____1965b, Ground-water resources of west-central Indiana—preliminary report, Fountain County: Indiana Department of Conservation, Division of Water, Bulletin 28, 91 p.

_____1965c, Ground-water resources of west-central Indiana—preliminary report, Vermillion County: Indiana Department of Conservation, Division of Water Bulletin 29, 89 p.

Wayne, W.J., 1956, Thickness of drift and bedrock physiography of Indiana north of the Wisconsin glacial boundary: Indiana Department of Conservation, Geological Survey Report of Progress 7, 70 p.

White, W.B., 1969, Conceptual models for carbonate aquifers: Ground Water, v. 7, p. 15-21.

Wier, C.E., 1952, Distribution, structure and mined areas of coals in Vigo County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 1, scale 1 inch equals approximately 2 miles.

LOWER WABASH RIVER BASIN

By Keith E. Bobay

General Description

The Lower Wabash River basin incorporates the drainage basin of the Wabash River between Honey Creek in Vigo County and the mouth of the Wabash River at the Ohio River in Posey County (fig. 48). The basin has an area of 1,339 mi^2 (Hoggatt, 1975) and includes most of Sullivan and Posey Counties, plus parts of Vigo, Greene, Knox, Gibson, and Vanderburgh Counties in southwestern Indiana. The major cities and towns in the basin are Vincennes, Sullivan, and Princeton.

Previous Studies

The ground-water resources of the counties in the Lower Wabash River basin were described by Harrell (1935). Watkins and Jordan (1962, 1963) made a preliminary evaluation of ground water in Sullivan County and Vigo County. They reported lithologic descriptions and published drillers' logs, well-construction and aquifer information, and water-quality analyses. Cable and others (1971) further described the hydrogeology of the principal aquifers in Vigo and Clay Counties. They mapped the bedrock geology and topography, the surficial geology, and the location and potentiometric surface of the major unconsolidated aquifers in the two counties. Cable and Robison (1973) described the hydrogeology of the principal aquifers of Sullivan County. Robison (1977) reported on the ground-water resources of Posey County. He described the sources and potential yield of the principal aquifers in the county and summarized the chemical quality of the ground water. In addition, he mapped the surficial geology and the bedrock topography of the county, as well as the topography of the Inglefield Sandstone. Cable and Wolf (1977) reported on the ground-water resources of Vanderburgh County. They mapped the areal extent, thickness, and surface topography of the Inglefield Sandstone aquifer in the Big Creek watershed (fig. 48). They also reported production potential and salinities of ground water from the major bedrock aquifers in the county.

Smith and Krothe (1983) evaluated the hydrogeology of the Carbondale and Raccoon Creek Groups of the Pennsylvanian System in the northern part of the Lower Wabash River basin. They reported transmissivities and the chemical quality of the sandstone and coal aquifers, and they mapped the potentiometric surface of the sandstone aquifers in the Brazil, Staunton, Linton, and Petersburg Formations. Glore (1970) and Wier and others (1973) studied the aquifer potential of the Middle Pennsylvanian sandstones of the Busseron Creek watershed in Sullivan County. Barnhart and Middleman (1990) studied the hydrogeology of Gibson County. They reported well yields and thicknesses of the bedrock aquifer formations and of the unconsolidated aquifers. They also mapped the topography of certain sandstone formations and reported the results of water-quality analyses. Clark (1980) mapped the potential yield of ground water from properly constructed large-diameter wells.

Shedlock (1980) mapped the thickness, distribution, and potentiometric surface of the glacial outwash aquifer in a 56-mi^2 area near Vincennes, Ind. Many authors (Wier, 1952a, 1952b, 1954; Friedman, 1954; Waddell, 1954; Wier and Powell, 1967; Tanner and others, 1981a, 1981b, 1981c; Smith and Krothe, 1983) have mapped the thickness and structure of the major Pennsylvanian coal seams in southwestern Indiana.

Physiography

The Lower Wabash River basin is part of the Wabash Lowland physiographic unit (Malott, 1922, p. 102; Schneider, 1966, p. 48) which is shown in figures 2 and 49. The Lower Wabash River valley is a broad, flat glacial drainage channel that includes winding channels, a wide flood plain, and adjacent terrace levels. The valley floor ranges from 3 to 10 mi in width. Local relief on the valley floor is typically less than 50 ft except for isolated hills (Fidlar, 1948, p. 10).

Undulating, rolling plains with a thin cover of till, loess, and silt characterize the area east of the Wabash terraces. Local relief is greater in the uplands of southern Posey County beyond the maximum extent of glaciation (fig. 49). Broad, flat lake plains that form present day bottomlands east of the terraces were created during Wisconsinan time when tributary valleys became ponded by the rapid aggradation of the valley floor (Fidlar, 1948, p. 102). In the surrounding uplands, bedrock terraces were eroded on resistant limestones and shales.

Steep bluffs rise from the Wabash Valley flood plain near the towns of Merom, Vincennes, and New Harmony, where the flood plain narrows. Relief is 100 to 150 ft in these areas. Springs are common along the bluffs. Isolated rock islands are exposed in the Wabash River flood plain. These erosional remnants, or braided-valley cores, are loess-covered bedrock that has withstood erosion. These hills rise nearly 100 ft above the flood plain (Fidlar, 1948; Thornbury, 1950).

Surface-Water Hydrology

The Wabash River has been divided into three separate water-management basins encompassing the upper, middle, and lower reaches (fig. 1). The Wabash River drains an area of 32,910 mi^2, including parts of Illinois and Ohio (Hoggatt, 1975, p. 174). The Lower Wabash River basin includes about 10 percent of the drainage area of the Wabash River basin in Indiana (Hoggatt, 1975). A streamflow gaging station on the Wabash River, located below the mouth of the Patoka River at drainage area 28,635 mi^2, has been in operation since 1927. During the period of record, discharges at this station ranged from a daily mean discharge of 1,650 ft^3/s to an instantaneous maximum of 305,000 ft^3/s, with an average flow of 27,569 ft^3/s (Glatfelter and others, 1989, p. 183). The average gradient of the Wabash River in the Lower Wabash River basin is approximately 0.7 ft/mi (Fidlar, 1948, p. 7).

The major tributaries to the Wabash River in the Lower Wabash River basin with drainage areas greater than 50 mi^2 include Prairie Creek, Turman Creek, Busseron Creek, Maria Creek, River Deshee, Black River, and Big Creek (fig. 48). The White River and the Patoka River also drain into the Wabash River in the Lower Wabash River basin; however, these two rivers are in separate water-management basins and are discussed in other sections of this report. The Patoka River and the White River are major tributaries that contribute more than 40 percent of the average flow in the Wabash River.

Geology

Bedrock Deposits

Pennsylvanian rocks are at the bedrock surface throughout the Lower Wabash River basin (fig. 50) and are more than 1,000 ft thick. A lithologic sequence of sandstone, shaly sandstone, shale, thin limestone, coal, and underclay comprise the Raccoon Creek, Carbondale, and McLeansboro Groups of

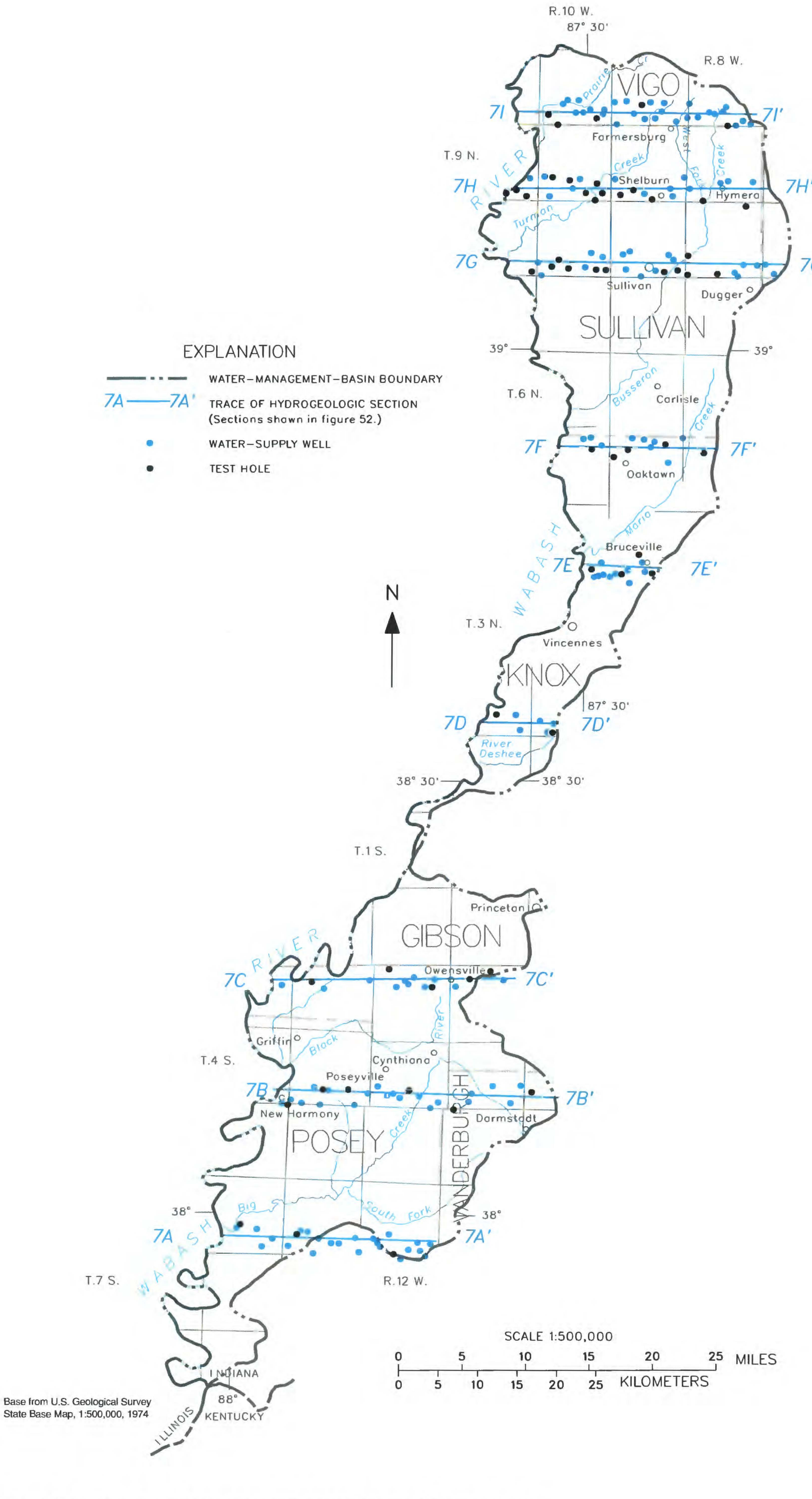


Figure 48. Location of section lines and wells plotted in the Lower Wabash River basin.

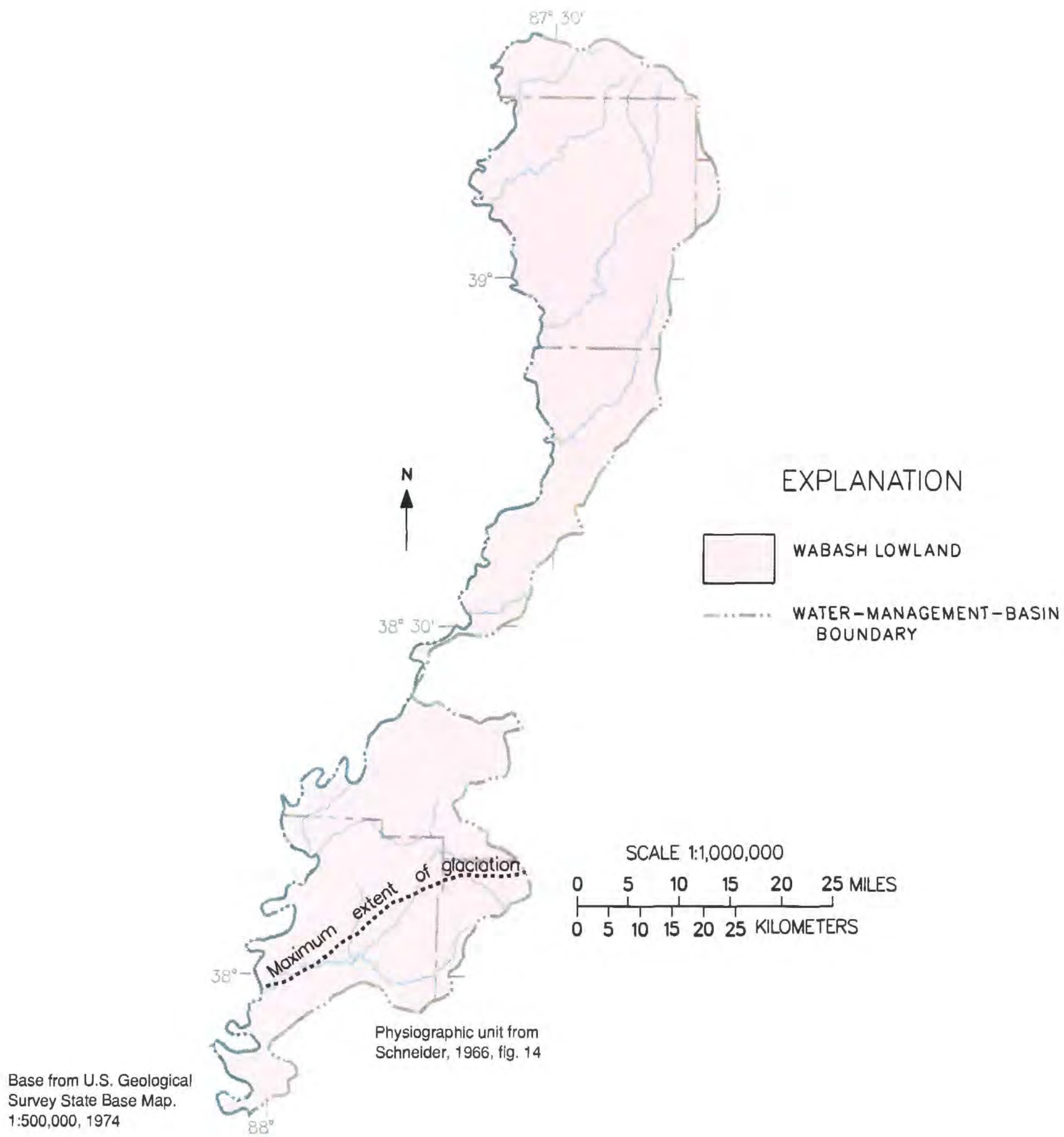


Figure 49. Physiographic unit and extent of glaciation in the Lower Wabash River basin.

Pennsylvanian age (fig. 5; Cable and others, 1971). Stratigraphic units are extremely difficult to trace in the subsurface because distinct lithologies of the clastic rock types are few, and lateral facies changes can be abrupt. Moreover, certain beds are discontinuous or absent locally, sharp variations in thickness occur over relatively short distances, and beds may have been eroded or never deposited. Formations of Pennsylvanian age are typically referenced with respect to the thin beds of the major coal seams and limestones that are the only consistent, areally extensive layers in the lithologic sequence (Harrell, 1935; Gray, 1979, p. 13).

Pennsylvanian bedrock exposed at the land surface or bedrock surface in the basin belong to the Linton, Petersburg, and Dugger Formations of the Carbondale Group and to the Shelburn, Patoka, Bond, and Mattoon Formations of the McLeansboro Group. These rocks overlie the Mansfield, Brazil, and Staunton Formations of the Early Pennsylvanian Raccoon Creek Group (fig. 5). These Pennsylvanian rocks unconformably overlie older Mississippian rocks (Shaver and others, 1986).

The Raccoon Creek Group ranges in thickness from 100 to 1,000 ft and is composed of 95 percent shale and sandstone, plus minor amounts of clay, coal, and limestone. Shale is more common than the massive, fine-grained, crossbedded sandstones. The Mansfield Formation is 50 to 300 ft thick and is divided into two general lithologies: a lower sandstone, and an upper division of mostly shale and mudstone. In the Lower Wabash River basin, the Mansfield Formation is typically overlain by more than 500 ft of rock. The Brazil Formation ranges in thickness from 40 to 90 ft and includes the Lower Block and Upper Block Coal Members (fig. 5). The Staunton Formation is 75 to 150 ft of sandstone, shale, and as many as eight coal beds, including the Seelyville Coal Member (Coal III). The Seelyville Coal averages 6 ft in thickness in the northern coal fields and has a 4- to 6-ft-thick plastic underclay with some shale (Shaver and others, 1986; Murray, 1957, p. 26).

The Carbondale Group ranges in thickness from 260 to 470 ft and averages 300 ft. The group, which is thickest in central Posey County, is composed primarily of variable shales and sandstones and includes some laterally extensive limestone and coal beds. The Linton Formation ranges in thickness from 43 to 162 ft and averages 80 ft. Lateral lithologic variations are common. The Petersburg Formation is 70 to 190 ft thick and includes the Springfield Coal Member (Coal V) (fig. 50). The Springfield Coal Member attains a maximum thickness of 13 ft and is split in places by as much as 65 ft of shale. This coal seam underlies as much as 90 ft of gray, silty shale in the Lower Wabash River basin. The clay layer beneath the Springfield Coal Member averages 2 to 5 ft in thickness. The Dugger Formation ranges in thickness from 73 to 185 ft, and includes the Danville Coal Member (Coal VII) and the Anvil Rock Sandstone Member. The Danville Coal seam attains thicknesses of 6.5 ft in Vigo County, then thins to 0.2 ft toward the south. The underclay associated with the Danville Coal Member averages 4 to 6 ft in thickness (Shaver and others, 1986; Murray, 1957, p. 26).

The McLeansboro Group attains its maximum thickness of 770 ft in northern Posey County. The group is made up of 90 percent shale and sandstone plus minor amounts of siltstone, limestone, clay, and coal. The Shelburn Formation ranges in thickness from 50 to 250 ft and is composed primarily of shale, siltstone, and sandstone. This formation crops out in the Lower Wabash River basin from the Ohio River in Vanderburgh and Posey Counties north through Vigo County. Included in the Shelburn Formation are the Busseron Sandstone Member and the West Franklin Limestone Member (fig. 50). The Busseron Sandstone Member rests unconformably on the Danville Coal Member. This sandstone is gray to tan, fine to medium grained, massive, and interbedded in places with a gray shale. Thickness ranges from 48 to 77 ft in Sullivan County; the member appears to be absent in many places within Posey County. The West Franklin Limestone Member consists of one to three limestone beds as much as 10 ft thick separated by as much as 25 ft of shale.

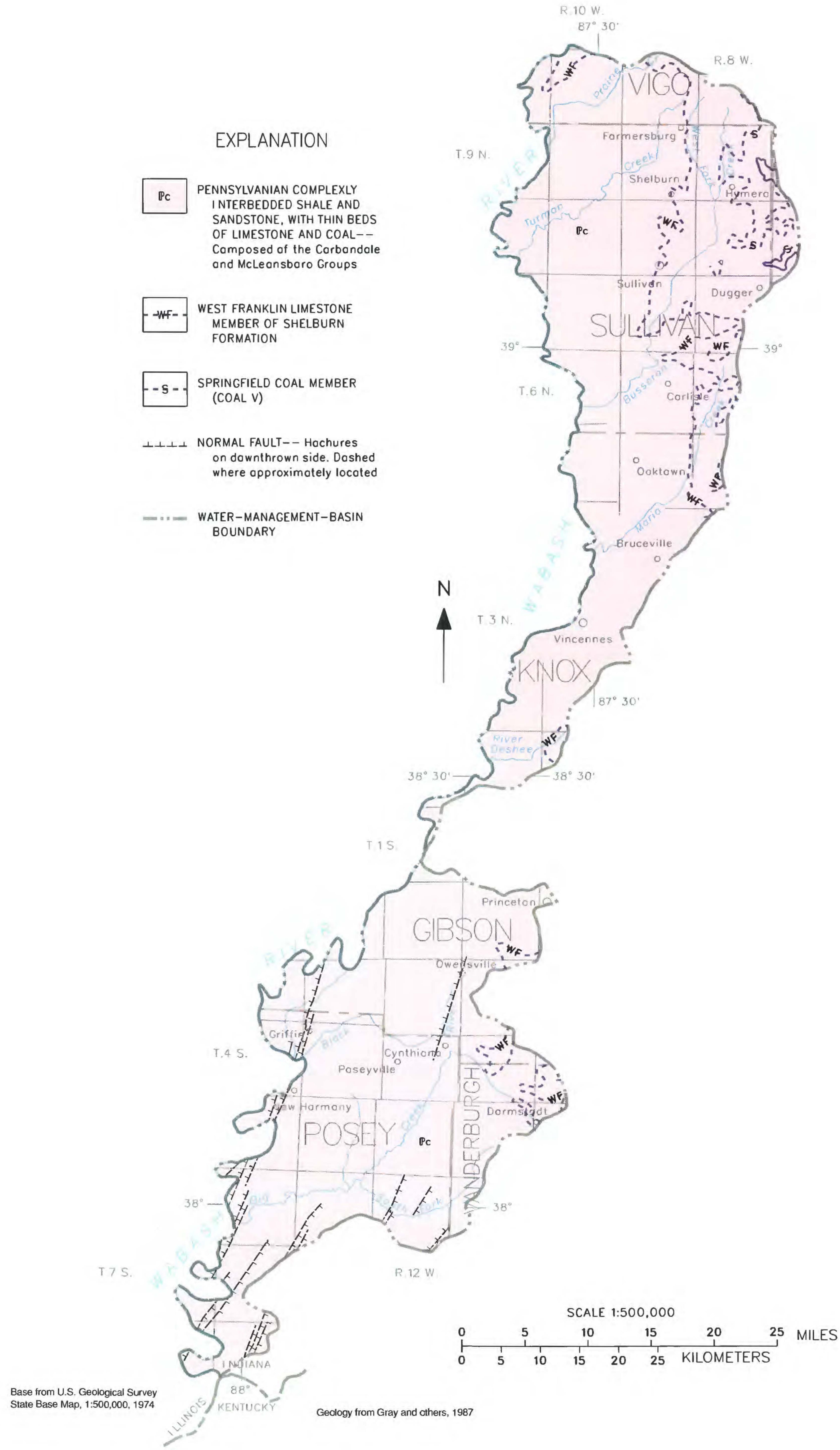


Figure 50. Bedrock geology of the Lower Wabash River basin.

One or more of the beds may not be present in parts of Posey and Gibson Counties (Shaver and others, 1986).

The Bond and Mattoon Formations are present only in the far western part of the Lower Wabash River basin (fig. 50). The formations are primarily shale, siltstone, and sandstone. The Bond Formation attains thicknesses of 250 ft in northwestern Posey County. Only the upper 40 ft of the Mattoon Formation is present in Western Sullivan County, whereas the lower 150 ft, including the Merom Sandstone Member is present in northwestern Posey County (Shaver and others, 1986).

A pronounced structural feature of the region is the Illinois Basin (fig. 4). The major stratigraphic units in the Lower Wabash River basin become thinner with distance from the center of the basin, which is in southeastern Illinois. In the Lower Wabash River basin, the Pennsylvanian rocks generally dip to the west-southwest at approximately 25 ft/mi (Gray, 1979, p. 3). In contrast to the topography of the formations that are exposed at the land surface, the bedrock surface beneath the glacial drift is generally a gently rolling plain. Valleys of the unglaciated areas are filled with valley-train deposits or lake sediments (Cable and Robison, 1973; Cable and Wolf, 1977, p. 6).

Faults in the Wabash Valley are another pronounced structural feature of the basin (fig. 50). Faults as much as 30 mi long, with displacements greater than 400 ft, have been identified (Ault and Sullivan, 1982). These post-Pennsylvanian faults are present as single planes and in zones of multiple planes. The faults in the Lower Wabash River basin are confined to Posey and Gibson Counties and trend north-northeastward, parallel to faults in Kentucky and Illinois (Ault and Sullivan, 1982, p. 7). Eleven distinct faults or fault systems, with displacement between 20 and 450 ft, have been mapped and named by Ault and Sullivan (1982, p. 12). Faults in the Inglefield Sandstone Member of Posey County have been mapped by Robison (1977, pl. 3) (see section 7A–7A′, fig. 52). Faulting in the Springfield Coal Member has been mapped in northern Posey County by Tanner and others (1981b).

Unconsolidated Deposits

In the aggraded valleys of the Wabash River and major tributaries, the primary unconsolidated deposits consist of alluvium that overlies thick Pleistocene valley-train sand and gravel deposits. The thickness of unconsolidated deposits in the Lower Wabash River basin is shown in figure 51. Thicknesses of sand and gravel as great as 150 ft have been measured adjacent to the Wabash River (Fidlar, 1948, pl. 3) and along Busseron Creek. In general, the sand and gravel deposits lie directly on the bedrock (Shedlock, 1980). Thickness of unconsolidated deposits decreases to 50 ft in minor tributary valleys and to less than 50 ft in the uplands (Gray, 1983). Many oxbow lakes and abandoned meanders are present in the modern Wabash River flood plain. Some of these depressions are filled with gravel and silt carried by floodwaters. Pre-Wisconsinan glacial drift of variable thickness covers the Pennsylvanian bedrock in the upland areas. Loess and sand dunes as much as 50 ft thick are scattered throughout the basin and cover the drift locally. Sand dunes have developed on terraces and lake plains along the edge of the Wabash River valley and on some nearby uplands. The dunes are primarily in Sullivan and Vigo Counties (Fidlar, 1948, p. 91-95).

Clay and silt beds were deposited in the lake plains along many of the tributary valleys. During Wisconsinan time, lakes once covered much of central and north-central Posey County and areas of Sullivan, Gibson, and Vanderburgh Counties (Fidlar, 1948, pl. 1). Lake-silt deposits up to 150 ft have been measured in northern Gibson County. Many of the lake plains have been buried beneath younger wind-blown sediments (Fidlar, 1948, p. 48).

Aquifer Types

Nine hydrogeologic sections (7A–7A′ to 7I–7I′) were produced for this atlas to depict aquifer types in the Lower Wabash River basin (fig. 52). All sections are oriented west to east, approximately perpendicular to the Wabash River (fig. 48). Section lines were drawn at about 12-mi intervals, except in northern Sullivan County where section lines were drawn at 6-mi intervals and in northern Gibson County where there is a 20-mi interval between two section lines. The average density of logged wells plotted along the section lines is 1.3 wells per mile.

The major aquifer type in the Lower Wabash River basin is the outwash and alluvial sand and gravel in the Wabash River valley. These thick sand and gravel deposits are relatively clean, well sorted, and coarse grained. A secondary unconsolidated source of ground water in the basin is the buried sand and gravel in tributary valleys. These deposits are covered by more than 10 ft of silt and clay. Additional unconsolidated ground-water resources include sand and gravel lenses interbedded with lake sediments and glacial till, and dune sands (Cable and others, 1971; Cable and Robison, 1973; Robison, 1977; Clark, 1980).

Wells drilled into the Pennsylvanian bedrock are commonly finished in the intervals that contain sandstones at the base of individual formations. Although these sandstones typically are the major sources of ground water in the bedrock, wells commonly are open to layers of shale, limestone, sandstone, and coal. The presence of laterally thin and discontinuous deposits results in extremely complex lithology and difficult aquifer definition in the basin. Therefore, it is impossible to delineate the exact source of the water to most wells. Yields of ground water from the bedrock are rarely enough for any use other than domestic. Thus, the importance of these bedrock aquifers lies not in their productivity but in their wide areal distribution. They are commonly the sole source of freshwater in the interior of the Lower Wabash River basin.

The four aquifer types of the Lower Wabash River basin are summarized in table 9. The table includes range of thickness, range of reported yields, and common or geologic names that previous authors have used to define the aquifers.

Unconsolidated Aquifers

A surficial sand and gravel aquifer is present along the Wabash River flood plain in the northern two-thirds of the basin (fig. 53). Although the surficial sand and gravel is depicted on the northern seven hydrogeologic sections, this aquifer is shown most extensively in sections 7D–7D′ and 7F–7F′ (fig. 52). Yields from wells located outside the Wabash River valley are typically much less than yields from the Wabash valley (table 9).

Sand dunes were mapped as surficial aquifer only where they are known to be producing water, such as in section 7F–7F′, R. 9 W. (fig. 52). Dune sands typically are thin deposits with a highly fluctuating water table; thus, dune sands are relatively undependable as a source of water. Instead of being used as aquifers, the dune sands in the Lower Wabash River basin commonly function as recharge areas for underlying aquifer material (Gray, 1973).

Buried sand and gravel aquifers generally are present in the southern one-third of the Wabash River flood plain and in the valleys of the major tributaries to the Wabash River. The buried sand and gravel in the Wabash River valley (sections 7A–7A′ to 7C–7C′, fig. 52) is connected to and of the same origin as the surficial sand and gravel aquifer to the north. The gradational boundary between the surficial and buried sand and gravel aquifers in the Wabash River valley is the area where the outwash is buried by more than 10 ft of nonaquifer alluvial material. In general, the aquifer is buried by about 10 to 20 ft of alluvial sand, silt and clay material. Buried sand and gravel aquifers also are present along Busseron, Prairie, Turman, Maria, and Big Creeks (fig. 53). The Busseron Creek and Turman Creek sand and gravel deposits are shown in section 7H–7H′ (fig. 52). Another buried sand and gravel aquifer underlies low-permeability clay and silt in the uplands near Poseyville (Robison, 1977, pl. 1,2).

Recharge to the unconsolidated aquifers typically occurs where precipitation infiltrates directly into the aquifer or through the overlying glacial or fine-grained alluvial material. Shedlock (1980) estimated recharge rates to the outwash aquifer near Vincennes to be 12 in/yr.

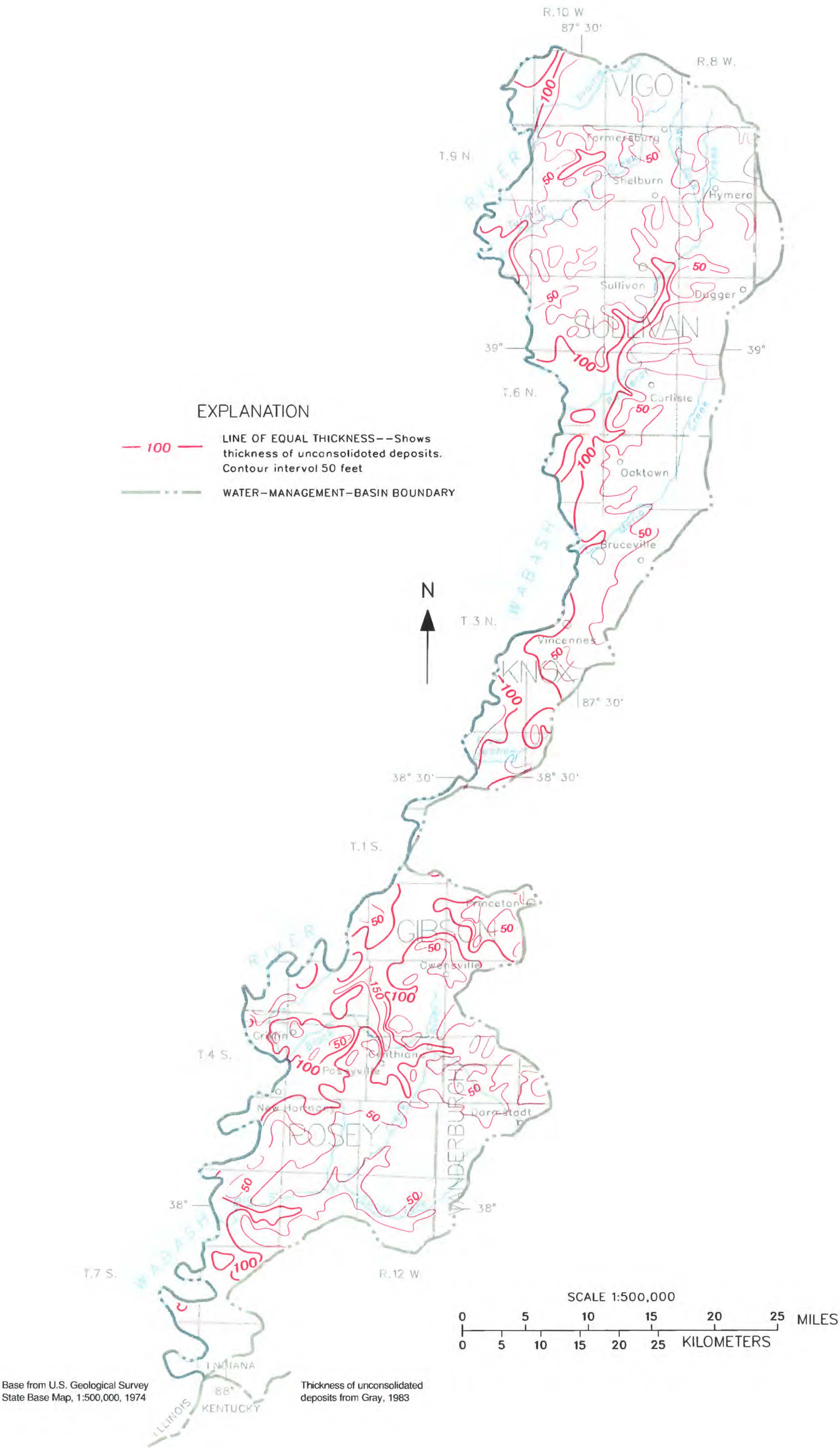


Figure 51. Thickness of unconsolidated deposits in the Lower Wabash River basin.

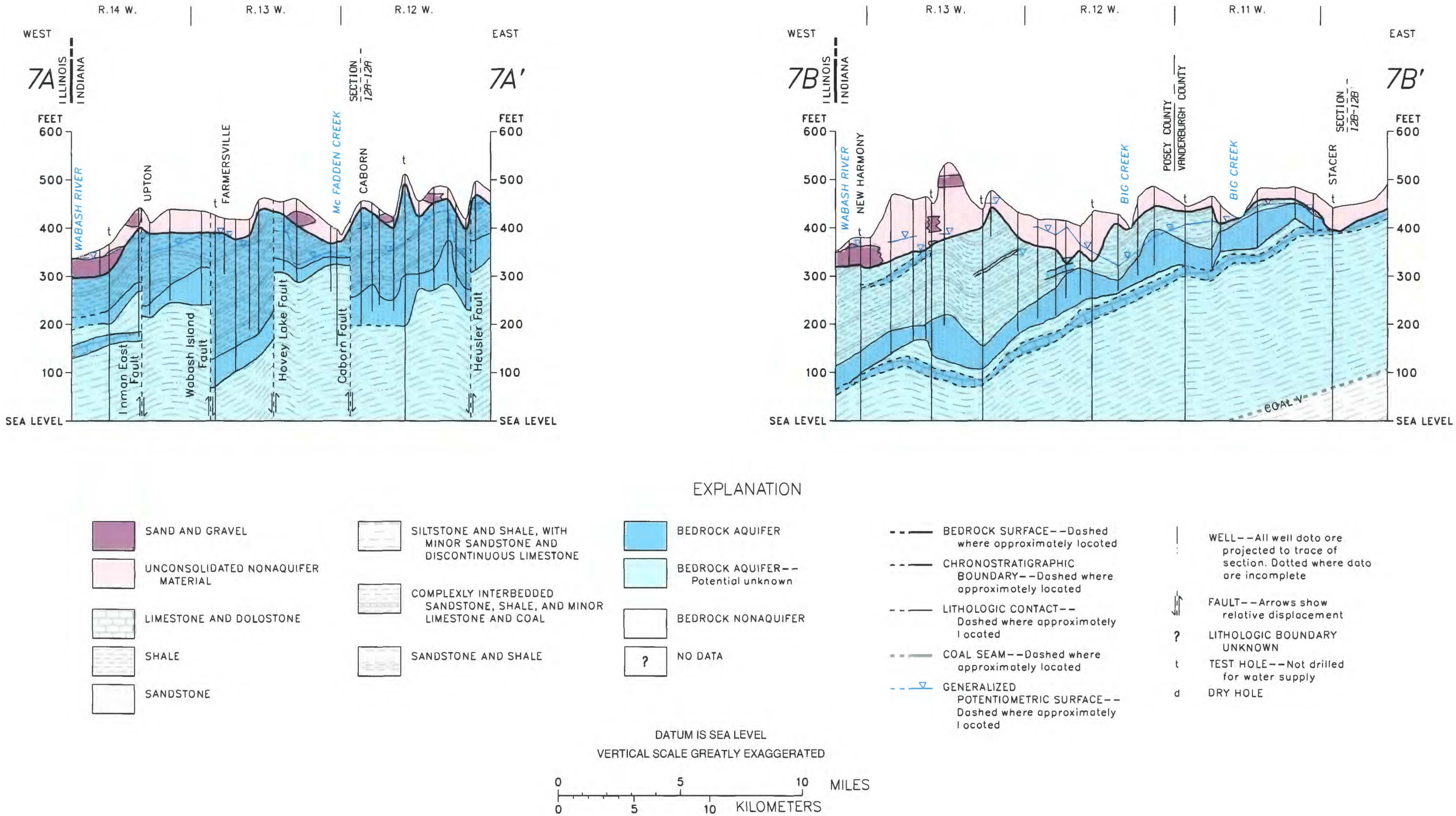


Figure 52. Hydrogeologic sections *7A–7A′* to *7I–7I′* of the Lower Wabash River basin.

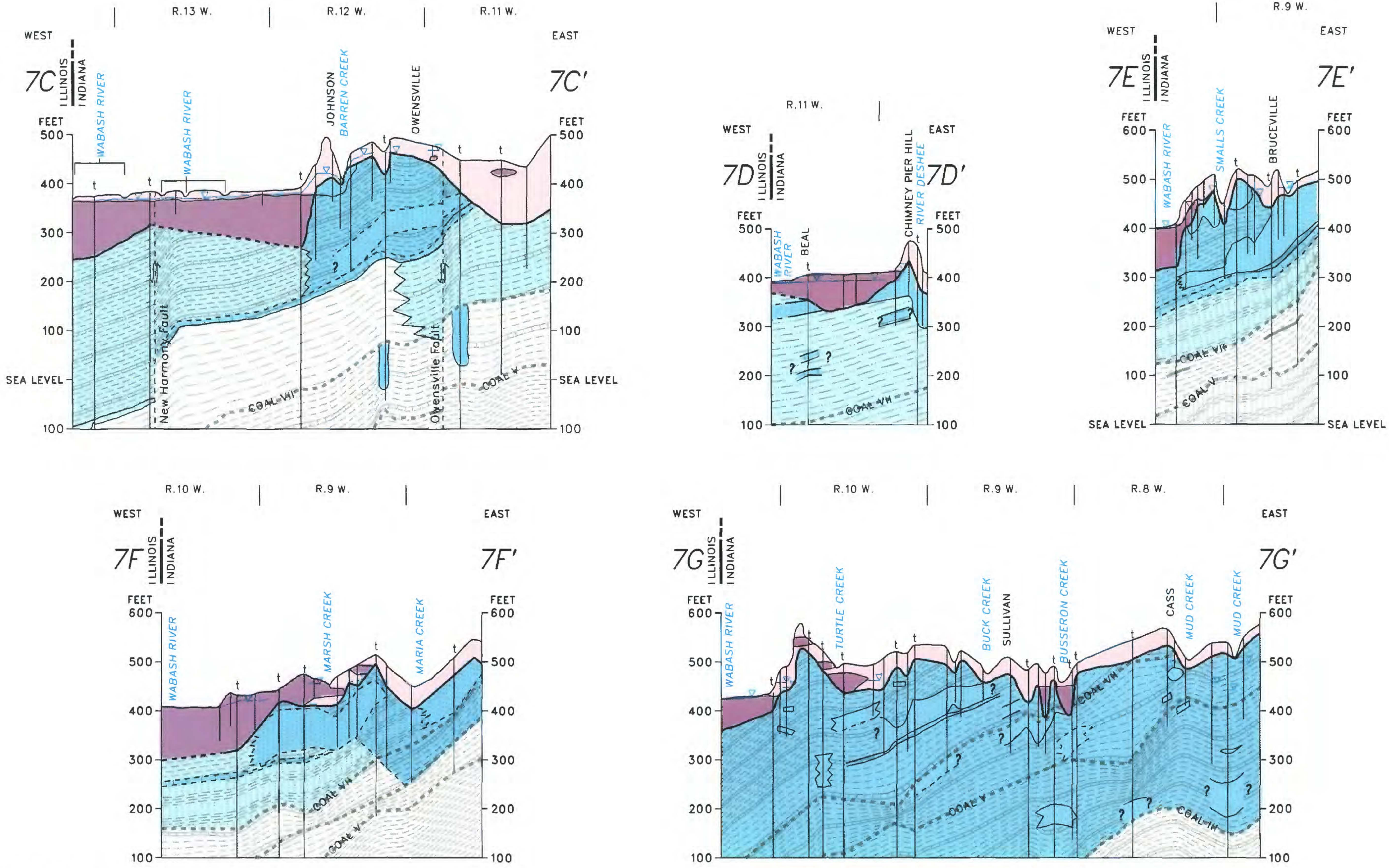
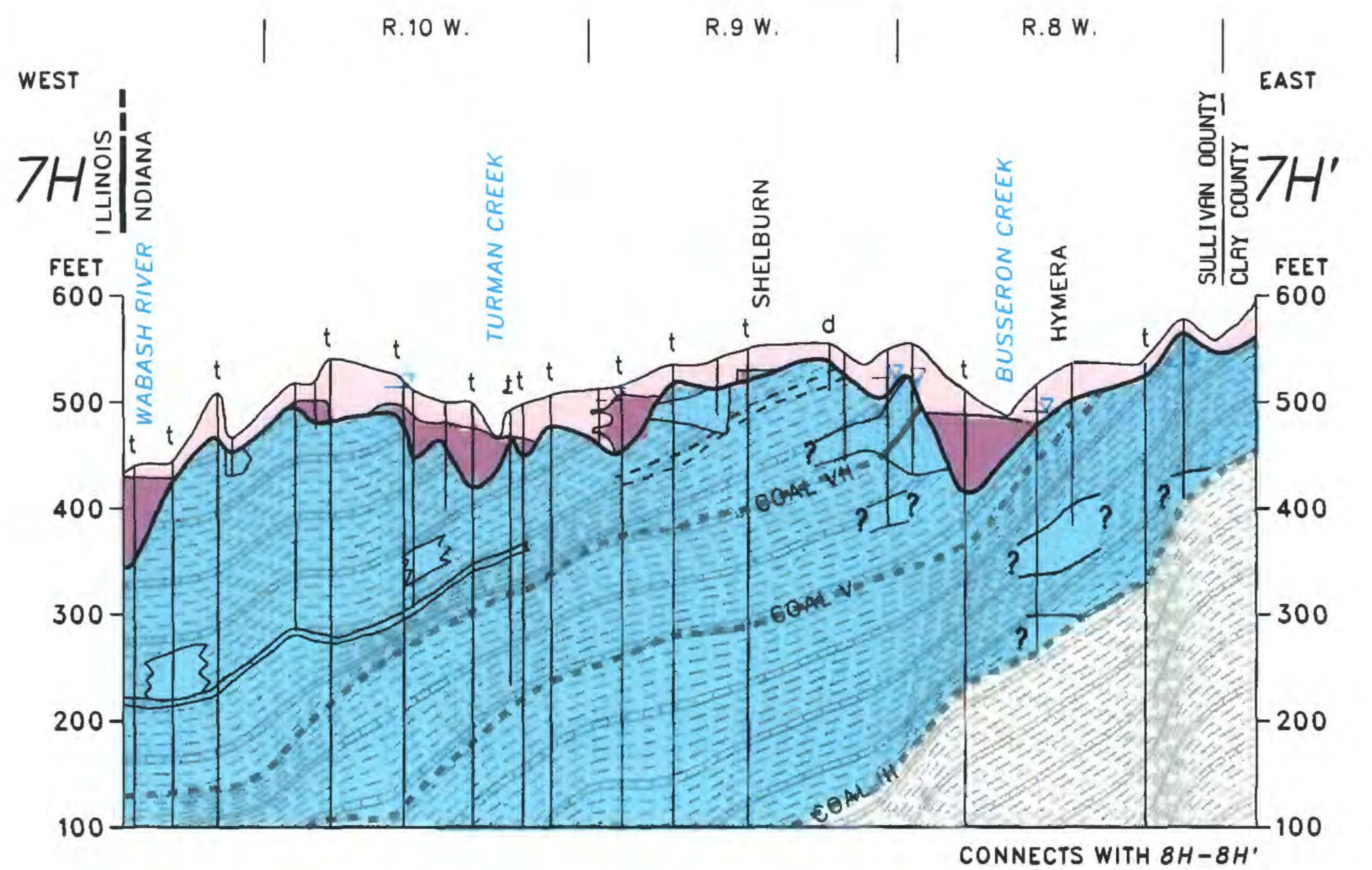
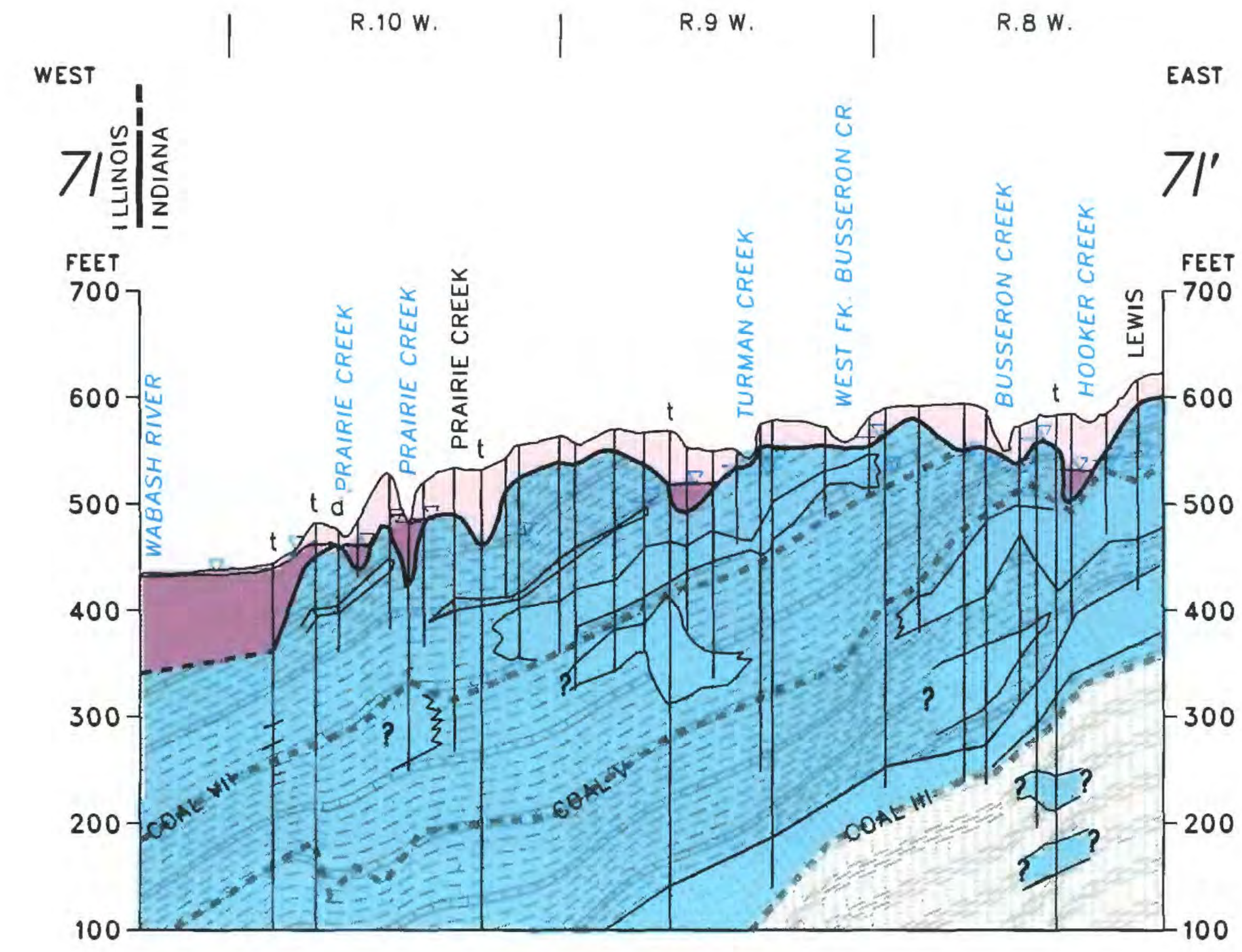
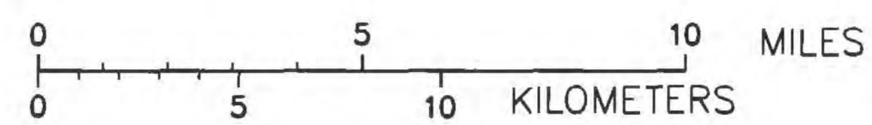


Figure 52. Hydrogeologic sections *7A–7A′* to *7I–7I′* of the Lower Wabash River basin.





DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED



Recharge from adjacent uplands and underlying bedrock is approximately 3 in/yr or 25 percent of the total recharge. Ground water in the unconsolidated aquifers naturally flows toward the Wabash River or its tributaries, where it discharges (Cable and others, 1971; Robison, 1977).

Bedrock Aquifers

The Inglefield Sandstone Member is the thickest and most laterally extensive bedrock aquifer in Vanderburgh, Posey, and Gibson Counties, as shown in sections 7A–7A′, 7B–7B′, and 7C–7C′ (fig. 52). The continuity of the sandstone aquifer is disrupted in section 7A–7A′ (fig. 52) by the many faults in Posey County. The areal extent of the sandstone shown on the aquifer map (fig. 53) generally follows the distribution of the Inglefield and Busseron Sandstone Members (Wier and Girdley, 1963; Barnhart and Middleman, 1990). The West Franklin Limestone Member typically underlies the Inglefield Sandstone Member and is used as a source of fresh water in places (see section 7B–7B′ in fig. 52). Normally, though, drillers complete wells in the more productive sandstone and do not penetrate the limestone below. The West Franklin Limestone Member is included with the complexly interbedded aquifer shown in figure 53. Other potential aquifers in Vanderburgh County include a sandstone in the Upper Dugger Formation which grades into shale, and the Coxville Sandstone Member, which is too deep and saline for water supply throughout much of the basin (Cable and Wolf, 1977, p. 12-14). The thick, narrow channel sandstones (for example, the Anvil Rock Sandstone Member) typically produce more water than the thin, broader sheet sandstones associated with the complexly interbedded aquifer (Cable and others, 1971). Two channel sandstones are depicted in section 7C–7C′ in R. 12 W. and R. 11 W. (fig. 52). Other less significant sandstones have been mapped on most of the sections. A few are laterally continuous, whereas many sandstones abruptly grade into shales, as shown in section 7I–7I′ (fig. 52).

Aquifers associated with complexly interbedded sandstone, shale, limestone, and coal are mapped together where most wells are open to the entire bedrock section below the unconsolidated material. Generally, the sandstones and coals are the primary water-producing units. Coal seams can be locally significant water-producing zones only if an underclay is present beneath the coal and the coal is fractured (Harrell, 1935, p. 76; Banaszak, 1980; Smith and Krothe, 1983). The underclays are important hydrologically as low-permeability, semiconfining beds capable of perching ground water in the coal seams and associated strata above. The approximate location of three major coal seams has been shown on the hydrogeologic sections. These coal seams are most likely to function as aquifers, owing to the presence of thick underclays. The complexly interbedded aquifer type is present throughout the entire basin (fig. 53); however, along the Wabash River valley in Gibson County, it is buried too deeply to be of use. The complexly interbedded aquifer material is best shown in the northern sections (7G–7G′, 7H–7H′, and 7I–7I′ in fig. 52).

Table 9. Characteristics of aquifer types in the Lower Wabash River basin
[Locations of aquifer types shown in fig. 53]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and ravel	10-150	[1]50-2,700	Wabash Valley outwash, valley train, and dune sand
Buried sand and gravel	10- 50	[2,3]5- 300	Interbedded sand and gravel lenses
Sandstone	20- 65	[4]0.5- 20	Inglefield Sandstone Member[2,3] or Patoka aquifer[4] or white water sand
Complexly interbedded sandstone, shale, limestone, and coal	[3,5]20- 50	[6]0.1- 15	Busseron Sandstone Member[1,3,5,6] or Unit 6[1]
	[5]1- 63	[6]8- 56	Coxville [4,6] or Linton aquifer[4]
			Anvil Rock [7], Merom[2], and St. Wendel[2] Sandstone Members
	0- 10	[1,3]0.5- 20	Danville (Coal VII), Springfield (Coal V), and Seelyville (Coal III) Members
			West Franklin Limestone Member[2] ; Units 2 and 3[8]

[1]Cable and others, 1971.
[2]Robison, 1977.
[3]Barnhart and Middleman, 1990.
[4]Cable and Wolf, 1977.
[5]Glore, 1970.
[6]Wier and others, 1973.
[7]Hopkins, 1958.
[8]Cable and Robison, 1973.

It is commonly impossible to laterally correlate a sandstone unit across a section or to know whether a specific sandstone or other stratigraphic unit is saturated or water-producing unless noted by the well drillers. Therefore, bedrock not indicated as "aquifer material" on the hydrogeologic sections implies that an abundance of low-permeability shale or sandy shale is present. If no information was available to indicate whether the rocks are aquifer or nonaquifer material, then the area is shown as "aquifer—potential unknown" on the section. An example is the material depicted beneath the sandstone aquifer in section 7A–7A′ (fig. 52). All of the complexly interbedded sandstone, shale, limestone, and coal was mapped as "aquifer—potential unknown" on figure 53 because of the difficulty in mapping aquifer zones in the complexly interbedded bedrock where hydrogeologic properties can abruptly change.

The cementation of the sandstones limits their hydraulic conductivity; the confining nature of the overlying shales and the underclays limits the recharge to the sandstones. These two factors contribute to the low yield from most sandstone aquifers in the basin (Glore, 1970, p. 34-35; Wier and others, 1973, p. 301). On the other hand, the sandstones of the Mansfield Formation at the Mississippian-Pennsylvanian contact (fig. 5) are reported to be a source of water, but they are too deep to be an economically desirable source of freshwater in the Lower Wabash River basin (Harrell, 1935; Watkins and Jordan, 1963; Cable and others, 1971; Cable and Robison, 1973; Robison, 1977; Banaszak, 1980; Clark, 1980; Smith and Krothe, 1983; Barnhart and Middleman, 1990).

The approximate locations of the Danville (Coal VII), Springfield (Coal V), and Seelyville (Coal III) Coal Members have been shown on the hydrogeologic sections (fig. 52) on the basis of data from the drilling records and previous studies (Wier, 1952a, 1952b; Friedman, 1954; Waddell, 1954; Wier and Powell, 1967; Tanner and others, 1981a, 1981b, 1981c). The coals are closer to the surface and are more commonly used as sources of water in the northern sections (7G–7G′, 7H–7H′, and 7I–7I′ in fig. 52), where sandstones are absent or discontinuous, than in other parts of the basin. Smith and Krothe (1983) found that many of the coal seams have higher transmissivities and specific capacities than the sandstones.

Recharge to bedrock aquifers occurs primarily near outcrop areas or indirectly through thin overlying unconsolidated deposits at a rate of a few inches per year (R.J. Shedlock, U.S. Geological Survey, written comm., 1982). The outcrop of the Inglefield Sandstone Member near the eastern boundary of the Lower Wabash River basin in Vanderburgh and Gibson Counties (Cable and Wolf, 1977, p. 16; Barnhart and Middleman, 1990, p. 6) could be a conduit for recharge to some of the coal seams (Banaszak, 1980). Ground water discharges from the Pennsylvanian sandstones near Vincennes by upward flow into the glacial outwash; ground water in the outwash discharges to the Wabash River and its tributaries (Shedlock, 1980).

Summary

The Lower Wabash River basin encompasses 1,339 mi^2 of Sullivan, Posey, Vigo, Greene, Knox, Gibson, and Vanderburgh Counties in southwestern Indiana. The basin includes all west-flowing drainage into the Wabash River from Honey Creek in Vigo County to the mouth of the Wabash River at the Ohio River in Posey County, nearly 304 river miles.

Four aquifer types were delineated in the Lower Wabash River basin. The principal aquifer in the basin is the thick sand and gravel deposits in the Wabash River valley. Yields of as much as 2,700 gal/min have been obtained from this aquifer, which is 150 ft thick in places. The aquifer boundary generally follows the flood plain along the river and was mapped as surficial sand and gravel in the northern part of the basin and buried sand and gravel in the southern part of the basin. Less productive surficial aquifers are present in the valleys of Busseron and Big Creeks, the major tributaries to the Wabash River within the basin.

Secondary aquifers are present in the buried sand and gravel deposits along minor tributary valleys and within upland lake deposits. In the uplands beyond the Wabash valley, sandstone aquifers provide water for domestic purposes where sand and gravel sources are absent. Thick sandstone aquifers are common in Posey and Gibson Counties, but the presence of faults in this area makes determination of the exact depth of the aquifer difficult. In many other areas within the interior of the basin, aquifers are limited to the complexly interbedded layers of sandstone, shale, limestone, and coal. Production is slight and generally limited to areas of fractures and joints, primarily in the sandstone and coal.

Discharge from surficial and buried sand and gravel aquifers in the basin is typically toward the Wabash River and its tributaries. Regional discharge from the bedrock aquifers appears to be upward to the Wabash River.

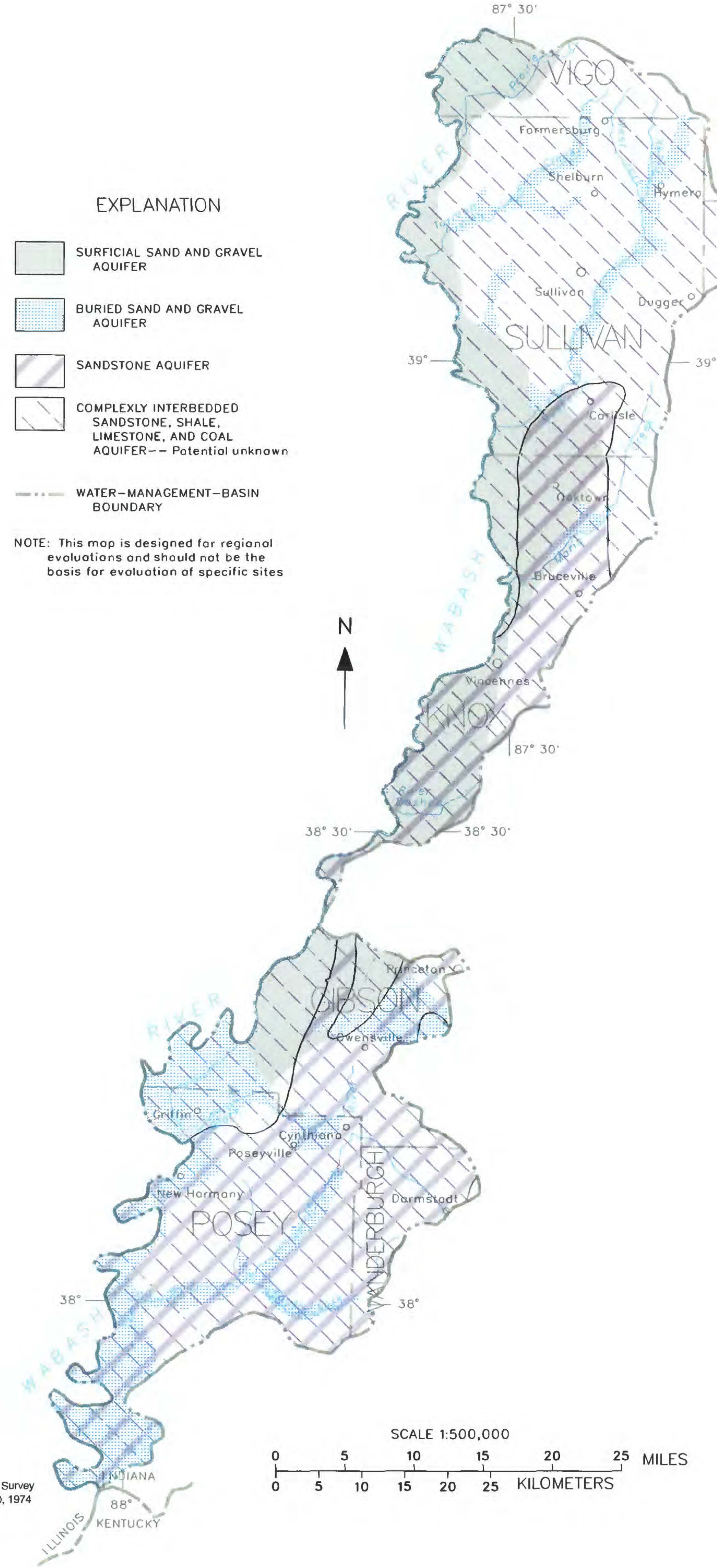


Figure 53. Extent of aquifer types in the Lower Wabash River basin.

References Cited

Ault, C.M., and Sullivan, D.M., 1982, Faulting in southwest Indiana: Washington, D.C., U.S. Nuclear Regulatory Commission, NUREG/CR-2908, 50 p.

Banaszak, K.J., 1980, Coal as aquifers in the Eastern United States, *in* Proceedings of the Symposium on Surface Mining Hydrology, Sedimentology, and Reclamation: University of Kentucky, Lexington, Kentucky, December 1-5, 1980, p. 235-241.

Barnhart, J.R., and Middleman, B.H., 1990, Hydrogeology of Gibson County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 41, 18 p.

Cable, L.W., and Robison, T.M., 1973, Hydrogeology of the principal aquifers in Sullivan and Greene Counties, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 35, 26 p.

Cable, L.W., Watkins, F.A., Jr., and Robison, T.M., 1971, Hydrogeology of the principal aquifers in Vigo and Clay Counties, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 34, 34 p.

Cable, L.W., and Wolf, R.J., 1977, Ground-water resources of Vanderburgh County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 38, 37 p.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Fidlar, M.M., 1948, Physiography of the lower Wabash valley: Indiana Department of Conservation, Geology Division Bulletin 2, 112 p.

Friedman, S.A., 1954, Distribution, structure, and mined areas of coals in Gibson County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 4, scale 1 inch equals approximately 2 miles.

Glatfelter, D.R., Thompson, R.E., Jr., and Nell, G.E., 1989, Water resources data, Indiana, water year 1988: U.S. Geological Survey Water-Data Report IN-88-1, 331 p.

Glore, C.R., 1970, A preliminary aquifer study of Pennsylvanian age sandstones, Busseron Creek Watershed, Sullivan County, Indiana: Bloomington, Ind., Indiana University, unpublished Master's Thesis, 40 p.

Gray, H.H., 1973, Properties and uses of geologic materials in Indiana: Indiana Department of Natural Resources, Geological Survey, Regional Geologic Map Series, Supplementary Chart 1.

_____1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Indiana: U.S. Geological Survey Professional Paper 1110-K, 20 p.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hopkins, M.E., 1958, Geology and petrology of the Anvil Rock Sandstone of Southern Illinois: Illinois Geological Survey Circular 256, 49 p.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 20, part 2, p. 59-256.

Murray, H.H., 1957, Pennsylvanian underclays—potential bonding clays for use in foundries: Indiana Department of Conservation, Geological Survey Report of Progress 11, 27 p.

Robison, T.M., 1977, Ground-water resources of Posey County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 39, 27 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Shedlock, R.J., 1980, Saline water at the base of the glacial-outwash aquifer near Vincennes, Knox County, Indiana: U.S. Geological Survey Water-Resources Investigations 80-65, 61 p.

Smith, C.R., and Krothe, N.C., 1983, Hydrogeology of the Carbondale and Raccoon Creek Groups, Pennsylvanian System, Vigo, Clay, and Sullivan Counties, Indiana: Indiana University Water Resources Research Center, 93 p.

Tanner, G.F., Stellavato, J.N., and Mackey, J.C., 1981a, Map of southwestern Gibson County, Indiana, showing structure on Springfield Coal Member (V) of the Petersburg Formation (Pennsylvanian): Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 32, scale 1:31,680.

_____1981b, Map of northern Posey County, Indiana, showing structure on Springfield Coal Member (V) of the Petersburg Formation (Pennsylvanian): Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 33, scale 1:31,680.

_____1981c, Map of southwestern Posey County, Indiana, showing structure on Springfield Coal Member (V) of the Petersburg Formation (Pennsylvanian): Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 34, scale 1:31,680.

Thornbury, W.D., 1950, Glacial sluiceways and lacustrine plains of southern Indiana: Indiana Department of Conservation, Division of Geology Bulletin 4, 21 p.

Waddell, Courtney, 1954, Geology and coal deposits of the Shelburn Quadrangle, Sullivan County, Indiana: U.S. Geological Survey Coal Investigations Map C17, scale 1:24,000.

Watkins, F.A., Jr., and Jordan, D.G., 1962, Ground-water resources of west-central Indiana—Preliminary report—Sullivan County: Indiana Department of Conservation, Division of Water Bulletin 14, 345 p.

_____1963, Ground-water resources of west-central Indiana—preliminary report, Vigo County: Indiana Department of Conservation, Division of Water Bulletin 17, 358 p.

Wier, C.E., 1952a, Distribution, structure, and mined areas of coals in Vigo County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 1, scale 1 inch equals approximately 2 miles.

_____1952b, Distribution, structure, and mined areas of coals in Sullivan County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 2, scale 1 inch equals approximately 2 miles.

_____1954, Geology and coal deposits of the Hymera Quadrangle, Sullivan County, Indiana: U.S. Geological Survey Coal Investigations Map C16, scale 1:24,000.

Wier, C.E., and Girdley, W.A., 1963, Distribution of the Inglefield and Dicksburg Hills Sandstone Members in Posey and Vanderburgh Counties, Indiana, *in* Proceedings of the Indiana Academy of Science, 1962: v. 72, p. 212-218.

Wier, C.E., and Powell, R.L., 1967, Distribution, structure, and mined areas of coals in Knox County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 12, scale 1 inch equals approximately 2 miles.

Wier, C.E., Glore, C.R., and Agnew, A.F., 1973, Sandstone aquifers in eastern Sullivan County, Indiana, *in* Proceedings of the Indiana Academy of Science, 1972: v. 82, p. 297-302.

WHITE RIVER BASIN

By Mary E. Hoover and James M. Durbin

General Description

The White River basin spans nearly the entire width of south-central Indiana. The basin, as defined in this report, includes the areas from the headwaters of the White River in Randolph County to the confluence with the Wabash River in Knox County, but does not include the basin of the East Fork White River (fig. 1). The White River basin encompasses 5,603 mi^2 in 27 counties and includes all or large parts of the following counties: Boone, Clay, Davies, Delaware, Greene, Hamilton, Hendricks, Knox, Madison, Marion, Monroe, Morgan, Owen, Putnam, Randolph, and Tipton. Principal cities within the basin are Anderson, Greencastle, Indianapolis, Linton, Martinsville, Muncie, Spencer, Washington, and Winchester (fig. 54).

Previous Studies

Because a large proportion of Indiana's population resides within the White River basin, many studies have been completed on ground water and characteristics of the aquifers that control ground-water availability. A series of reports by the U.S. Geological Survey describes the ground-water resources of five counties within the northern part of the basin: Madison (Lapham, 1981), Delaware (Arihood and Lapham, 1982), Hamilton and Tipton (Arihood, 1982), and Randolph (Lapham and Arihood, 1984). The authors of these studies examined the hydrogeology of the White River basin within each respective county and modeled expected yields given a variety of pumping schemes, geohydrologic characteristics of the aquifers, and locations of induced recharge.

Other studies that focused on northern counties in the basin include reports on the hydrogeology of Delaware County (Hoggatt and others, 1968), Madison County (Wayne, 1975), Marion County (Herring, 1976), and Hamilton County (Gillies, 1976). The study by Gillies (1976) included modeling of an aquifer system adjacent to the White River near Carmel, Ind., and evaluation of the effects of continued and increased production from the aquifer. Studies of the outwash aquifer along the White River in Marion County (Meyer and others, 1975; Smith, 1983) focused on the characteristics of the aquifer and modeling of the hydrology and water availability for Indianapolis. The outwash aquifer along the White River in Johnson and Morgan Counties was studied by Bailey and Imbrigiotta (1982) to estimate the geometry and hydraulic characteristics of the aquifer and to establish the nature and extent of the hydraulic connection between surface and subsurface hydrology. Watkins (1965) appraised the ground-water resources and effects of a proposed reservoir on the hydrology of the Big Walnut Creek watershed in parts of Putnam, Hendricks, and Boone Counties.

Another series of reports published by the Indiana Department of Conservation, Division of Water, in cooperation with the U.S. Geological Survey, describes the ground-water resources of a number of southwestern Indiana counties within the White River basin. Studies were done in Greene County (Watkins and Jordan, 1961), Clay County (Watkins and Jordan, 1962), and Owen County (Watkins and Jordan, 1963); the authors published well logs, delineated which lithologies were aquifers, and evaluated ground-water availability. Other reports published by the Indiana Department of Natural Resources, Division of Water, for Clay and Vigo Counties (Cable and others, 1971) and Greene and Sullivan Counties (Cable and Robison, 1973) refined the work done previously in those counties and expanded the research to include data on water quality. A report by Barnhart and Middleman (1990) detailed the hydrogeology and ground-water quality of Gibson County. A report by Wangsness and others (1981) summarized available hydrologic data for an area that includes the lower half of the White River basin downstream from Gosport, Ind. (fig. 54). The report includes surface-water, ground-water, and water-quality information. A Master's thesis by Thomas (1980) detailed the aquifer potential and characteristics of the Mansfield Formation within Clay County.

A ground-water study that describes the hydrogeology of the entire White River basin was done by Nyman and Pettijohn (1971). The report is a brief description of the important aquifers in the basin, and includes information on well yields and potential yields, ground-water quality, and ground-water discharge to the major streams in the basin. A major study by the U.S. Geological Survey is currently (1991-97) being done for the White and East Fork White River basins as part of the National Water-Quality Assessment Program. The study will assess the water quality of the surface- and ground-water resources of the White and East Fork White River basins (Jacques and Crawford, 1991).

In addition to written reports, various ground-water-availability maps have been published. The Indiana Department of Natural Resources, Division of Water, has published maps that delineate major aquifers along with recorded and potential well yields in the following counties: Morgan (Heckard, 1965), Johnson (Uhl, 1966), Madison (Steen, 1970), Hamilton (Herring, 1971), Marion (Herring, 1974), and Boone (Steen and others, 1977). Ground water availability maps have been completed for the entire state of Indiana by Bechert and Heckard (1966) and Clark (1980).

Physiography

The topographic relief across the White River basin is about 750 ft. The highest point, about 1,200 ft above sea level, is in Randolph County in the eastern part of the basin. The lowest point, about 450 ft above sea level, is in Gibson County in the southernmost part of the basin.

The basin lies within five physiographic units as defined by Malott (1922) and later refined by Schneider (1966) (fig. 55). The northern half of the basin is in the Tipton Till Plain. This plain of low relief is composed of thick glacial deposits that obscure the underlying bedrock topography. The Norman Upland, of which only a small part of the northernmost extent is within the basin, is characterized by narrow, flat-topped divides and deep V-shaped valleys; local relief is typically 125 to 250 ft. The Norman Upland is well drained by a strongly developed dendritic stream pattern. The Mitchell Plain in the White River basin, which in most places is less than 7 mi wide, occupies a narrow strip in the central part of the basin. The Mitchell Plain is a westward-sloping plain composed of limestones. The limestones are subject to karst development and they form numerous sinkholes into which some streams "disappear". The karst development in the White River basin is not as extensive as karst development further south in the State. The Crawford Upland is a westward-sloping plateau developed in interbedded sandstones, shales, and limestones capped by resistant sandstones. Differential erosion in this region has created a deeply dissected upland in which local relief is as much as several hundred feet. The Crawford Upland is about 25 mi wide and is adjacent to, and west of the Mitchell Plain. The Wabash Lowland is the southernmost physiographic unit in the basin. This unit is a broad lowland underlain by nonresistant siltstones and shales, which have been eroded by repeated glaciations into a subdued landscape.

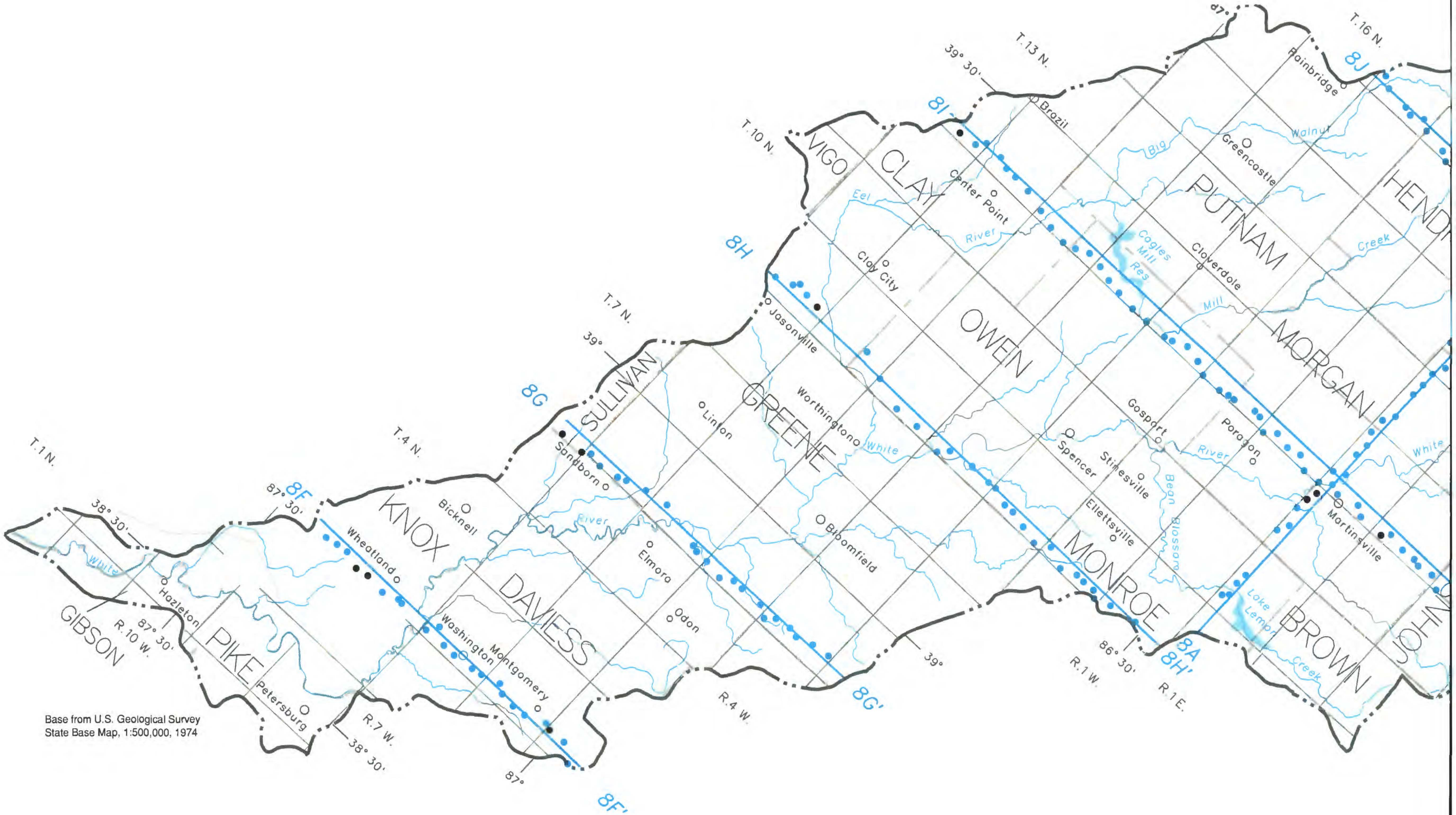
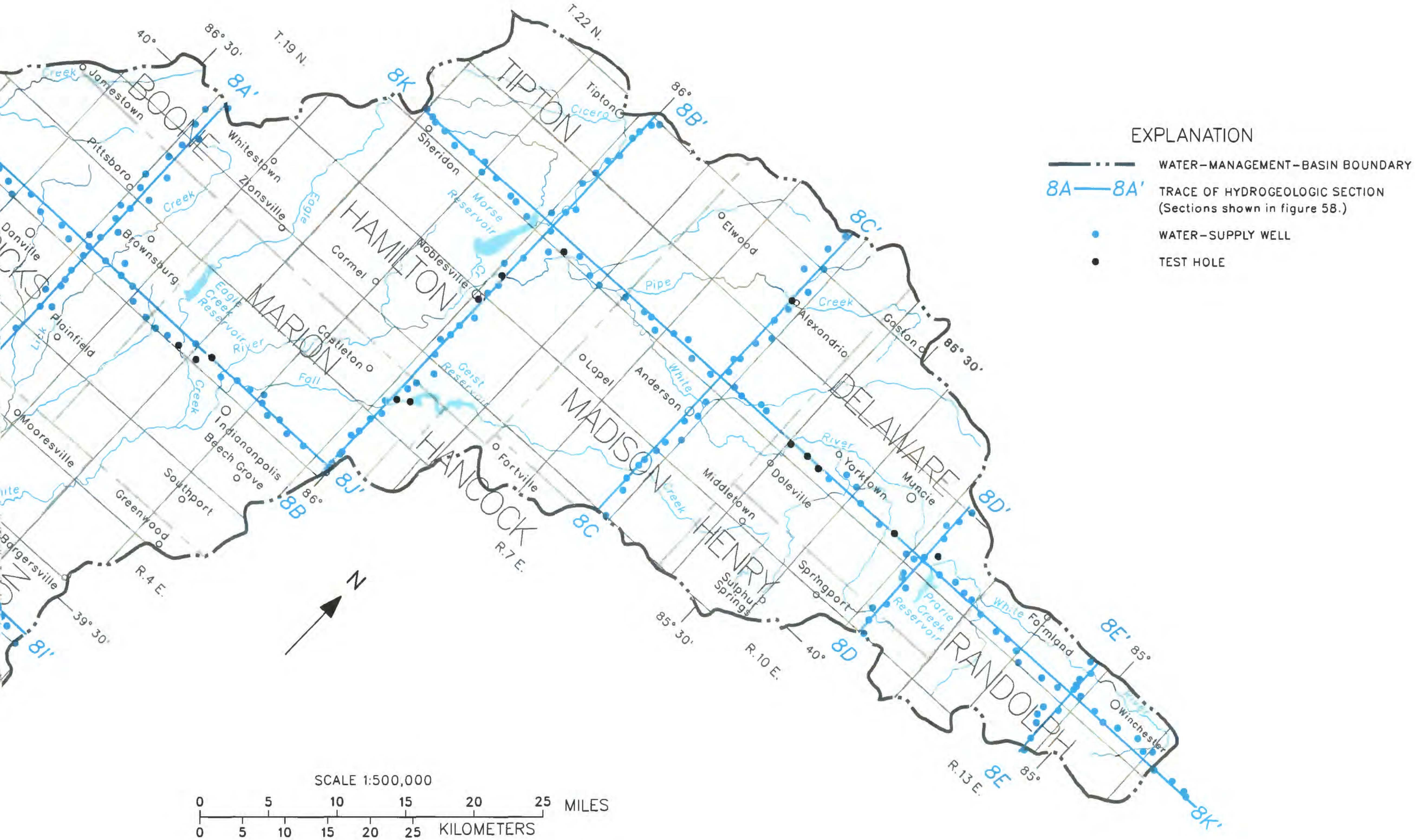


Figure 54. Location of section lines and wells plotted in the White River basin.


EXPLANATION
WATER–MANAGEMENT–BASIN BOUNDARY
8A——8A' TRACE OF HYDROGEOLOGIC SECTION (Sections shown in figure 58.)
WATER–SUPPLY WELL
TEST HOLE
SCALE 1:500,000
0 5 10 15 20 25 MILES
0 5 10 15 20 25 KILOMETERS
N
BOONE
HAMILTON
TIPTON
MARION
HANCOCK
MADISON
DELAWARE
HENRY
RANDOLPH
8A'
8K
8B'
8C'
8D'
8E'
8K'
8B
8C
8D
8E
8J'
8I'
40°
86° 30'
T.19 N.
T.22 N.
86°
85° 30'
39° 30'
85°
R.4 E.
R.7 E.
R.10 E.
R.13 E.
Jamestown
Pittsboro
Brownsburg
Danville
Plainfield
Mooresville
Whitestown
Zionsville
Carmel
Sheridon
Noblesville
Castleton
Indianapolis
Beech Grove
Southport
Greenwood
Bargersville
Tipton
Elwood
Alexandria
Gaston
Lapel
Anderson
Fortville
Middletown
Daleville
Yorktown
Muncie
Springport
Sulphur Springs
Farmland
Winchester
Cicero
Morse Reservoir
Geist Reservoir
Eagle Creek Reservoir
Prairie Creek Reservoir
Eagle
Creek
Pipe
White
River
Fall
Creek

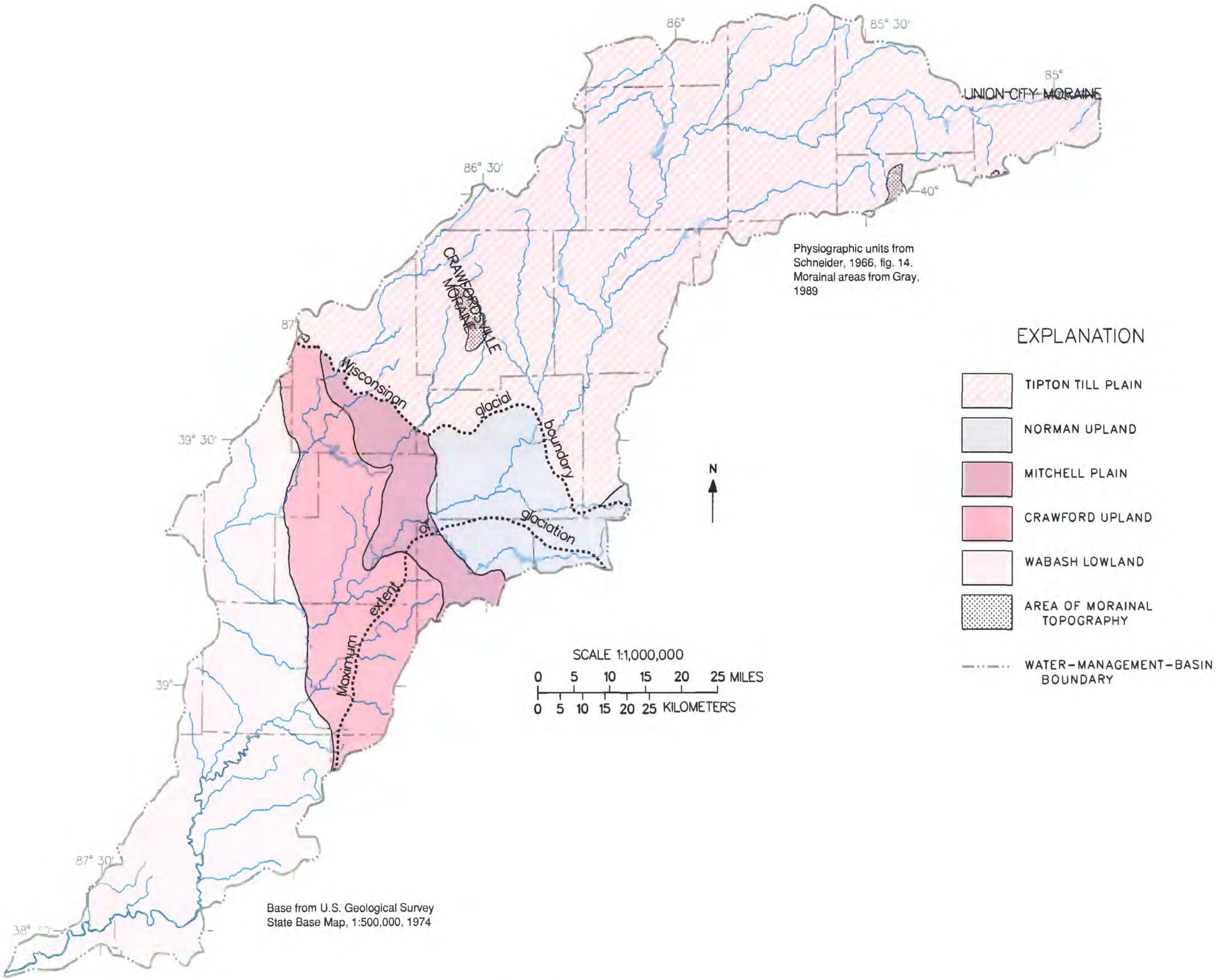


Figure 55. Physiographic units, moraines, and extent of glaciation in the White River basin.

Surface-Water Hydrology

The White River provides the major drainage within the basin; average discharges of the river are 208 ft^3/s near Muncie in Delaware County and 11,850 ft^3/s near Petersburg in Pike County (Arvin, 1989). Several large tributary drainage basins are within the White River basin (fig. 54). The Eel River tributary, in the southwestern part of the basin, has the largest drainage area (830 mi^2) of any tributary to the White River in the White River basin. Other tributaries whose drainage areas are greater than 100 mi^2 include Fall Creek, Eagle Creek, Big Walnut Creek, White Lick Creek, Mill Creek, Pipe Creek, and Cicero Creek. These tributaries are perennial streams and, depending upon climatic and aquifer conditions, are either recharge sources or discharge outlets for ground water.

A number of streams have been artificially dammed to form water-supply reservoirs. Principal reservoirs include Morse, Geist, Eagle Creek, Cagles Mill, and Prairie Creek Reservoirs.

Geology

Bedrock Deposits

The White River basin overlies two major structural features known as the Illinois Basin and the Cincinnati Arch (fig. 4). Bedrock strikes north-northwest, generally dipping gently to the southwest into the Illinois Basin; however, in the northeastern part of the basin where the Cincinnati Arch is present, bedrock dips northward toward the Michigan Basin, as shown in sections 8C–8C′, 8D–8D′ and 8E–8E′ (fig. 58). Successively younger rocks are exposed in the basin from east to west (fig. 56). Rocks of Ordovician age are exposed on top of the Cincinnati Arch in the northeastern part of the basin (fig. 56). To the west, rocks of Silurian, Devonian, Mississippian, and Pennsylvanian age are present at the bedrock surface either as subcrops where covered by unconsolidated materials or as outcrops where exposed in unglaciated areas or along some of the large streams (fig. 56). Erosional unconformities between the Silurian and Devonian contact and the Mississippian and Pennsylvanian contact are significant. Preglacial stream systems have eroded and dissected the entire bedrock surface, removing large amounts of Paleozoic rocks from the crest of the Cincinnati Arch and creating deep bedrock valleys (fig. 7). Examples of these valleys can be seen in most of the hydrogeologic sections (fig. 58).

The Fortville Fault and the Mount Carmel Fault, each about 50 mi long, transect the basin. The Fortville Fault strikes north-northeast from Marion County through Hancock and Madison Counties (fig. 56). The southeastern block of the fault is downthrown. The Mount Carmel Fault strikes north-northwest from Washington County through Lawrence and Monroe Counties (fig. 56). Only the northernmost 10 mi of the Mount Carmel Fault is within the basin.

Ordovician rocks of major lithostratigraphic significance in the White River basin are part of the Maquoketa Group. The Maquoketa Group is as much as 80 percent shale that is interbedded with limestone. The proportion of limestone increases toward the east in the White River basin (Shaver and others, 1986, p. 88).

Silurian rocks within the basin include the Brassfield Limestone, the Cataract Formation, the Salamonie Dolomite, and the Salina Group. The Brassfield Limestone, which is less than 10 ft thick in most places, interfingers with shales and dolostones of the Cataract Formation (Shaver and others, 1986, p. 20). The Salamonie Dolomite is a fairly pure dolostone that is about 50 ft thick in the central part of the State (Shaver and others, 1986, p. 180-132). The Salina Group contains the Pleasant Mills Formation and the Wabash Formation, both of which are composed of limestone and dolostone interbedded with shale members (Gray and others, 1987). Both the carbonate rocks and the shales are of variable thickness (Shaver and others, 1986, p. 114-116, 163-165).

Devonian bedrock consists primarily of dolomitic carbonate rocks (Muscatatuck Group) or shale (New Albany Shale). The Muscatatuck Group can be as much as 250 ft thick, but it is probably no thicker than 50 to 60 ft in the White River basin. The New Albany Shale, which is Devonian and Mississippian in age, is composed of dark carbonaceous shales (Shaver and others, 1986, p. 101) and is 85 to 150 ft thick within the White River basin.

Rocks of Mississippian age include the Borden, Sanders, Blue River, West Baden, and Stephensport Groups. The Borden Group ranges in thickness from 485 to 800 ft and consists of the New Providence Shale, the Spickert Knob Formation, and the Edwardsville Formation. The New Providence Shale, overlying the New Albany Shale, is composed predominantly of shale. The Spickert Knob Formation grades upward from a silty shale to a massive siltstone but includes some sandstone and limestone. The Edwardsville Formation consists of siltstone, sandy shale, and sandstone interbedded with minor limestones (Shaver and others, 1986, p. 18-19).

The Sanders and Blue River Groups consist of well-bedded and dense limestones that contain thin shale beds. Where limestone crops out or is covered by thin unconsolidated materials, it commonly is highly karstic and contains numerous sinkholes and caves. The thickness of the Blue River Group in outcrop within the basin ranges from 150 to 240 ft; in the subsurface, thickness may exceed 350 ft (Shaver and others, 1986, p. 16-17). Thickness of the Sanders Group is variable, ranging from 120 to 150 ft (Shaver and others, 1986, p. 136).

The West Baden Group is a mixture of sandstones, siltstones, shales, and mudstones, interbedded with thin limestone lenses; outcrop thickness is 100 to 140 ft and subsurface thickness is as much as 260 ft in Gibson County (Shaver and others, 1986, p. 167). The Stephensport Group is composed of equal parts of shales, sandstones, and limestones. Because of the erosional unconformity between the Mississippian and Pennsylvanian rocks, outcrops of the Stephensport Group are generally less than 50 ft thick and are absent in many places throughout the White River basin (Gray and others, 1987). The subsurface thickness of the Stephensport Group ranges from 130 to 230 ft (Shaver and others, 1986, p. 151).

Rocks of Pennsylvanian age within the White River basin include the Raccoon Creek, Carbondale, and McLeansboro Groups. These three groups are dominated by shales, but sandstones, siltstones, limestones, clays, and coal also are major components. Within the Raccoon Creek Group are the Mansfield, Brazil, and Staunton Formations. The Mansfield Formation, which can be as much as 300 ft thick, is mostly sandstone in the lower part of unit but contains increasingly more shale upward in the unit (Shaver and others, 1986, p. 87). The Brazil Formation is characterized by the lack of traceable beds; it is composed primarily of shale, sandstone, underclay, and coal, which have a combined thickness of 40 to 90 ft (Shaver and others, 1986, p. 21). The Staunton Formation consists of 75 to 150 ft of sandstones, shales, thin, areally limited coal beds, and minor limestone lenses (Shaver and others, 1986, p. 149-150). The Carbondale Group includes the Linton, Petersburg, and Dugger Formation. The Linton Formation consists of sandstones, shales, limestones and coal; it is typically about 80 ft thick but ranges from 60 to 162 ft in thickness (Shaver and others, 1986, p. 80). The Petersburg Formation consists of 40 to 120 ft of shale, fine-grained sandstone, and coal, including the Springfield Coal Member (Coal V) (Shaver and others, 1986, p. 112). The Dugger Formation contains several coal members and beds of limestone, shale, and clay, and ranges in thickness from 73 to 185 ft (Shaver and others, 1986, p. 39). The Shelburn, Patoka, and Bond Formations of the McLeansboro Group are present in the far southwestern corner of the basin. The McLeansboro Group is more than 90 percent shale and sandstone, but has small amounts of siltstone, limestone, coal, and clay (Shaver and others, 1986, p. 86). The West Franklin Limestone, a thin but persistent marker bed, is present within the Shelburn Formation (fig. 56).

Unconsolidated Deposits

Nearly all of the White River basin is covered by unconsolidated deposits, most of which were deposited by continental ice sheets. During the Pleistocene, continental ice sheets consisting of numerous lobes advanced into Indiana at least three times and deposited glacial sediments (Wayne, 1966, p. 21). These three glacial advances occurred during the Wisconsinan, Illinoian and pre-Illinoian glacial stages (in order from most to least recent). Thicknesses of deposits range from less than 25 ft in the southern part of the basin to as much as 400 ft in the northern part of the basin, although most of the unconsolidated deposits in the basin are from 50 to 150 ft thick (fig. 57). Glacial sediments, including outwash sand and gravel, from all three glacial stages filled preglacial stream valleys and created buried bedrock valleys (Bleuer, 1989). The location of these buried bedrock valleys is shown in figure 7.

Exposures of pre-Illinoian deposits are rare in the White River basin, and little information on the nature and extent of these deposits is available. During the Illinoian Age, ice covered as much as 80 percent of Indiana. Illinoian deposits are exposed throughout the southern half of the basin. These Illinoian deposits are predominantly loam tills that are heavily dissected; few morainal systems have been delineated. Pre-Illinoian and Illinoian glacial sediments are included in the Jessup Formation (Gray, 1989).

Overlying Illinoian and pre-Illinoian deposits are Wisconsinan glacial materials. During Wisconsinan glaciation, the Lake Michigan Lobe and Erie Lobe covered the upper one-third of the White River basin (fig. 8) and deposited extensive terminal and recessional morainal systems. Only small segments of these systems, the Union City and Crawfordsville Moraines (figs. 3 and 55), are within the boundary of the basin. The northern one-half of the basin is covered by thick ground moraine, which is composed of loamy tills interbedded with thin, discontinuous to continuous layers of stratified sand and gravel. Outwash that was transported south from the Wisconsinan glaciers filled in many of the large stream valleys beyond the glacial boundary, as well as valleys within the Wisconsinan glacial limits. During all of the glacial stages, the landscape was covered by windblown deposits to some degree; these deposits consisted chiefly of loess (windblown silt) and localized dune sand.

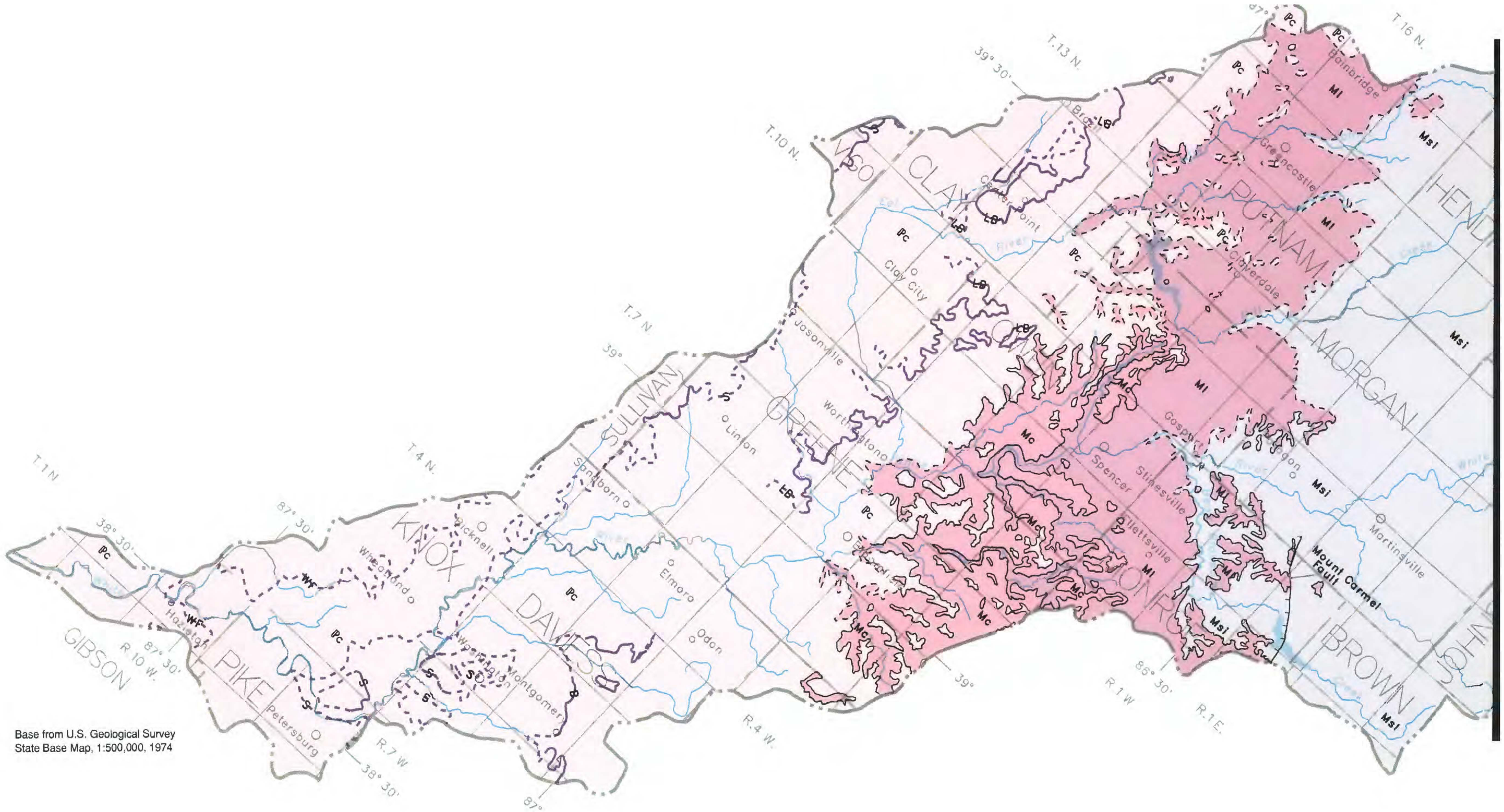
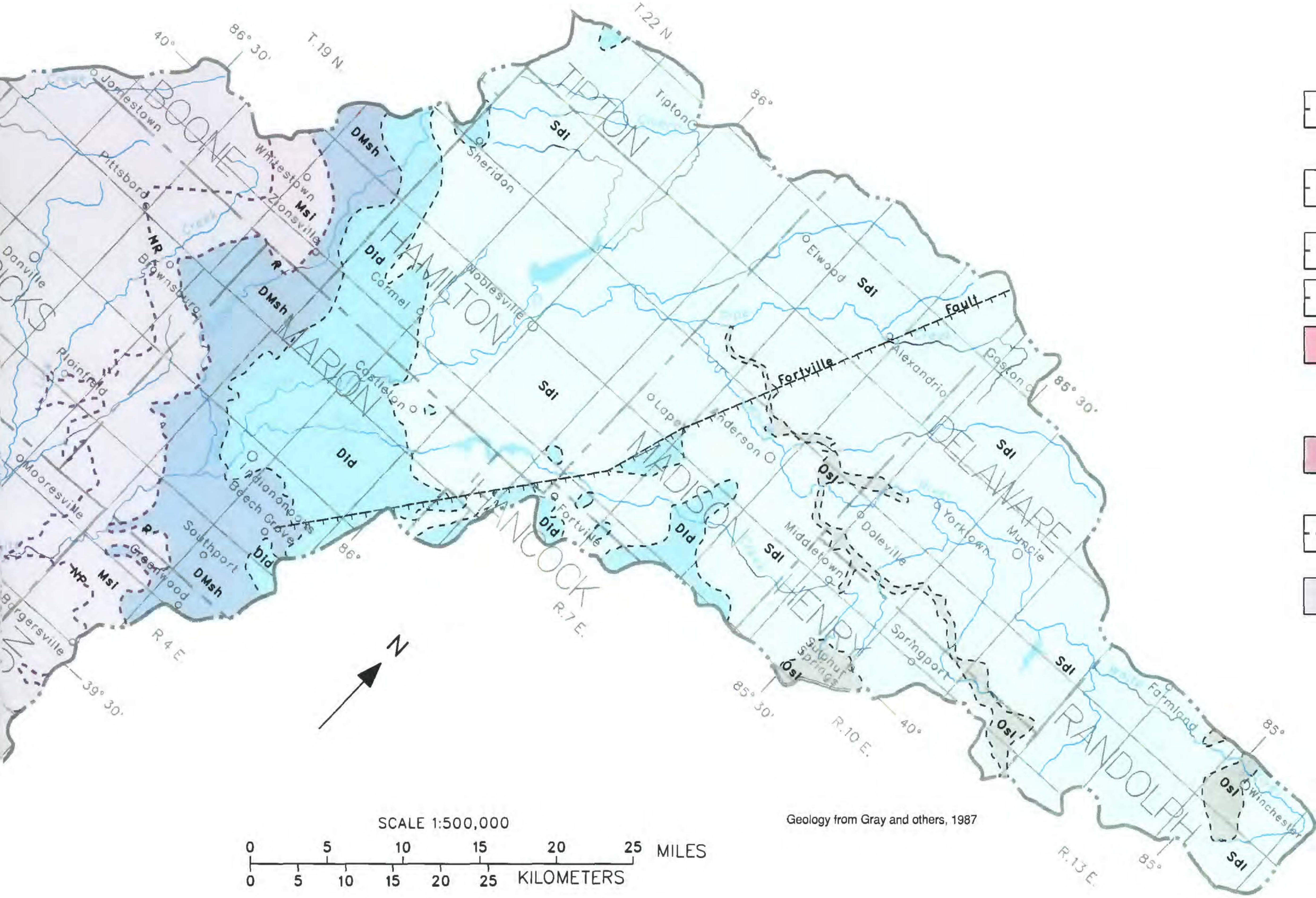


Figure 56. Bedrock geology of the White River basin.



Geology from Gray and others, 1987

EXPLANATION

- **Pc** — PENNSYLVANIAN COMPLEXLY INTERBEDDED SHALE AND SANDSTONE, WITH THIN BEDS OF LIMESTONE AND COAL-- Composed of the Racoon Creek and Carbondale Groups and the Shelburn, Patoka, and Bond Formations of the McLeansboro Group
- **-WF-** — WEST FRANKLIN LIMESTONE MEMBER OF SHELBURN FORMATION
- **-S-** — SPRINGFIELD COAL MEMBER (COAL V)
- **-B-** — BUFFALOVILLE COAL MEMBER
- **-LB-** — LOWER BLOCK COAL MEMBER
- **Mc** — MISSISSIPPIAN COMPLEXLY INTERBEDDED SHALE, SANDSTONE AND LIMESTONE-- Composed of the West Baden and Stephensport Groups
- **Ml** — MISSISSIPPIAN LIMESTONE-- Composed of the Sanders and Blue River Groups
- **-NP-** — TOP OF NEW PROVIDENCE SHALE
- **Msi** — MISSISSIPPIAN SILTSTONE AND SHALE WITH MINOR SANDSTONE AND DISCONTINUOUS LIMESTONE-- Composed of the Borden Group
- **-R-** — ROCKFORD LIMESTONE
- **DMsh** — DEVONIAN AND MISSISSIPPIAN SHALE-- Composed of the New Albany Shale
- **Dld** — DEVONIAN LIMESTONE AND DOLOMITE-- Composed of the Muscatatuck Group
- **Sdl** — SILURIAN DOLOMITE AND LIMESTONE-- Composed of the Wabash and Pleasant Mills Formations, and the Salamonie Dolomite, Cataract Formation, and Brassfield Limestone
- **Osl** — ORDOVICIAN SHALE AND LIMESTONE-- Composed of the Maquoketa Group
- NORMAL FAULT-- Hachures on downthrown side. Dashed where approximately located
- GEOLOGIC CONTACT-- Dashed where approximately located
- WATER-MANAGEMENT-BASIN BOUNDARY

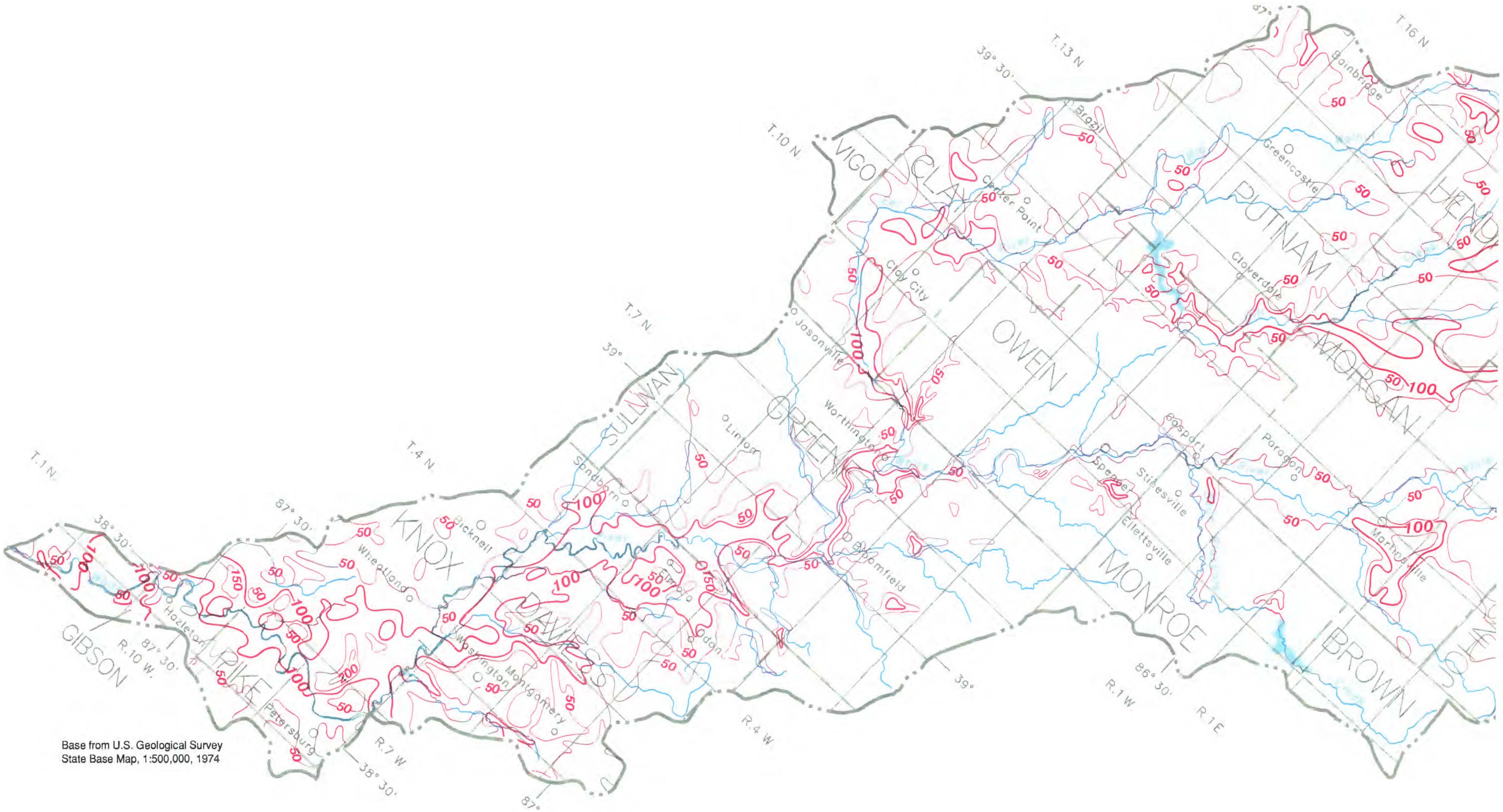
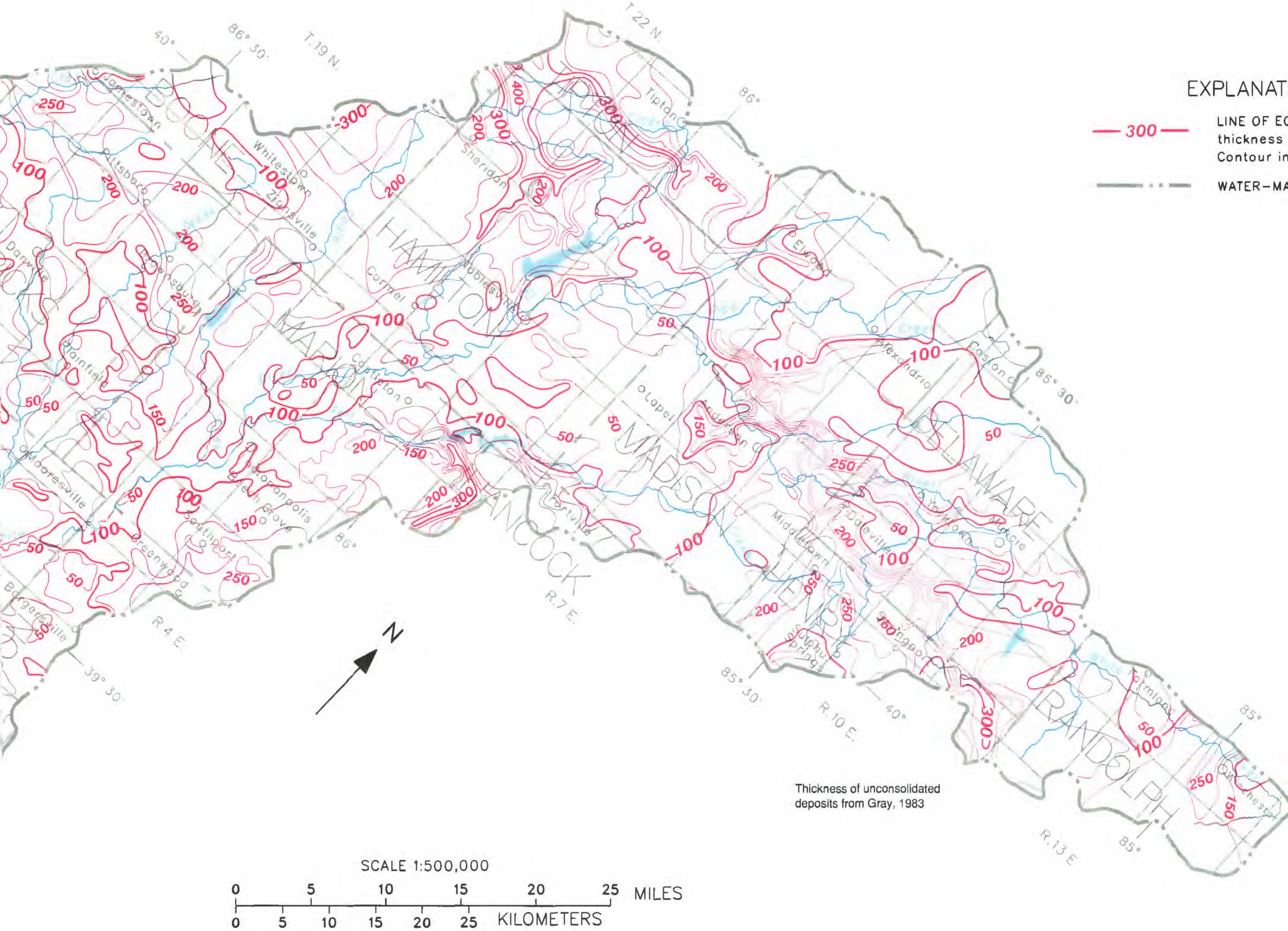


Figure 57. Thickness of unconsolidated deposits in the White River basin.



Thickness of unconsolidated deposits from Gray, 1983

EXPLANATION

— 300 — LINE OF EQUAL THICKNESS--Shows thickness of unconsolidated deposits. Contour intervol 50 feet

— · · — WATER-MANAGEMENT-BASIN BOUNDARY

Aquifer Types

The hydrostratigraphy of the White River basin is shown in 11 hydrogeologic sections (fig. 58). Hydrogeologic sections 8A–8A′ to 8E–8E′ are oriented south-north, and hydrogeologic sections 8F–8F′ to 8K–8K′ are oriented west-east (fig. 54). The typical spacing between hydrogeologic sections is about 18 mi, the exception being the spacing between 8D–8D′ and 8E–8E′, which is only 12 mi. The total length of the 11 hydrogeologic sections is about 410 mi. In all, 354 well logs were used to draw the sections. These well logs were plotted at an average density of one well log every 1.2 miles (fig. 54).

Throughout the northeastern one-third of the basin, the principal aquifers are buried continuous sand and gravel where the drift is greater than 25 ft thick, carbonate rock (limestone and dolostone) where drift is thin, and surficial sand and gravel near major streams (fig. 59). Where the glacial deposits are thick, the depth of wells ranges from 50 to 400 ft and averages 150 ft (Bechert and Heckard, 1966, p. 108-109). The carbonate bedrock aquifer in the northeastern one-third of the basin is Late Ordovician, Silurian, and Devonian in age. Wells in these rocks are as deep as 150 ft (Lapham, 1981, p. 16), but only the upper 100 ft is generally considered to be permeable (Cable and others, 1971).

Throughout the central one-third of the basin, principal aquifers include surficial, buried, and discontinuous sand and gravel; an upper weathered zone in siltstone and shale, and a carbonate bedrock aquifer (fig. 59). The characteristics of the sand and gravel aquifers are the same as those in the northeastern one-third of the basin. The siltstone-shale aquifer is used only where no other aquifer type is available. Water production from these normally low-yield rock types is from a zone of enhanced permeability created by weathering and fracturing of the shale and siltstone. Water produced from the carbonate bedrock aquifer is from Mississippian limestones.

The principal aquifers in the southwestern one-third of the basin are surficial sand and gravel; sandstone; complexly interbedded sandstone, shale, limestone, and coal; and carbonate rock (fig. 59). Surficial sand and gravel along large streams is the only productive sand and gravel aquifer in the southwestern one-third of the basin. Yields from all bedrock aquifers in the area are low (less than 20 gal/min). Sandstone aquifers are present in Pennsylvanian rocks, aquifers in the complexly interbedded materials are present in Late Mississippian and Pennsylvanian rocks, and carbonate bedrock aquifers are present in Mississippian rocks. Physical characteristics and some common or stratigraphic names for aquifer types within the basin are summarized in table 10.

Unconsolidated Aquifers

Surficial Sand and Gravel Aquifers

Surficial sand and gravel aquifers are restricted to the major river valleys throughout the basin (fig. 59) and can be seen in sections 8A–8A′, 8B–8B′, and 8F–8F′ to 8I–8I′ (fig. 58). In instances where an entire valley is filled from bedrock to land surface with sand and gravel (as shown in section 8A–8A′ (fig. 58) along the White River near Martinsville), the valley was mapped as surficial sand and gravel aquifer. The entire thickness of sand and gravel may not represent a single, continuous deposit but rather is an area of stratigraphic and hydraulic connection between the surficial and buried sand and gravel. The surficial sand and gravel consists of Wisconsinan and older glaciofluvial or fluvial sand and gravel and minor windblown deposits in the form of dune sands (Thornbury, 1950; Barnhart and Middleman, 1990; Gray, 1989). The dune sands, found in the southern part of the basin, may be a local source of water for shallow domestic wells, but these sands are generally considered insignificant as aquifers (Watkins and Jordan, 1961; 1962).

The areal extent of the surficial sand and gravel aquifer in the southern part of the basin is greater than that in the northern part; however, the demands on the aquifer in the north are much greater than in the south because of its use by the municipalities of Muncie, Anderson, and Indianapolis, and by nearby industries. Authors of previous studies have agreed that the "outwash" aquifers that underlie the major streams are the most productive aquifers in the basin (Watkins and Jordan, 1961, 1962; Lapham, 1981; Arihood, 1982; Arihood and Lapham, 1982; Lapham and Arihood, 1984). In the southern part of the basin, where the surficial sand and gravel aquifer is used for small-town and domestic supplies, it has not been developed to its full water-producing potential.

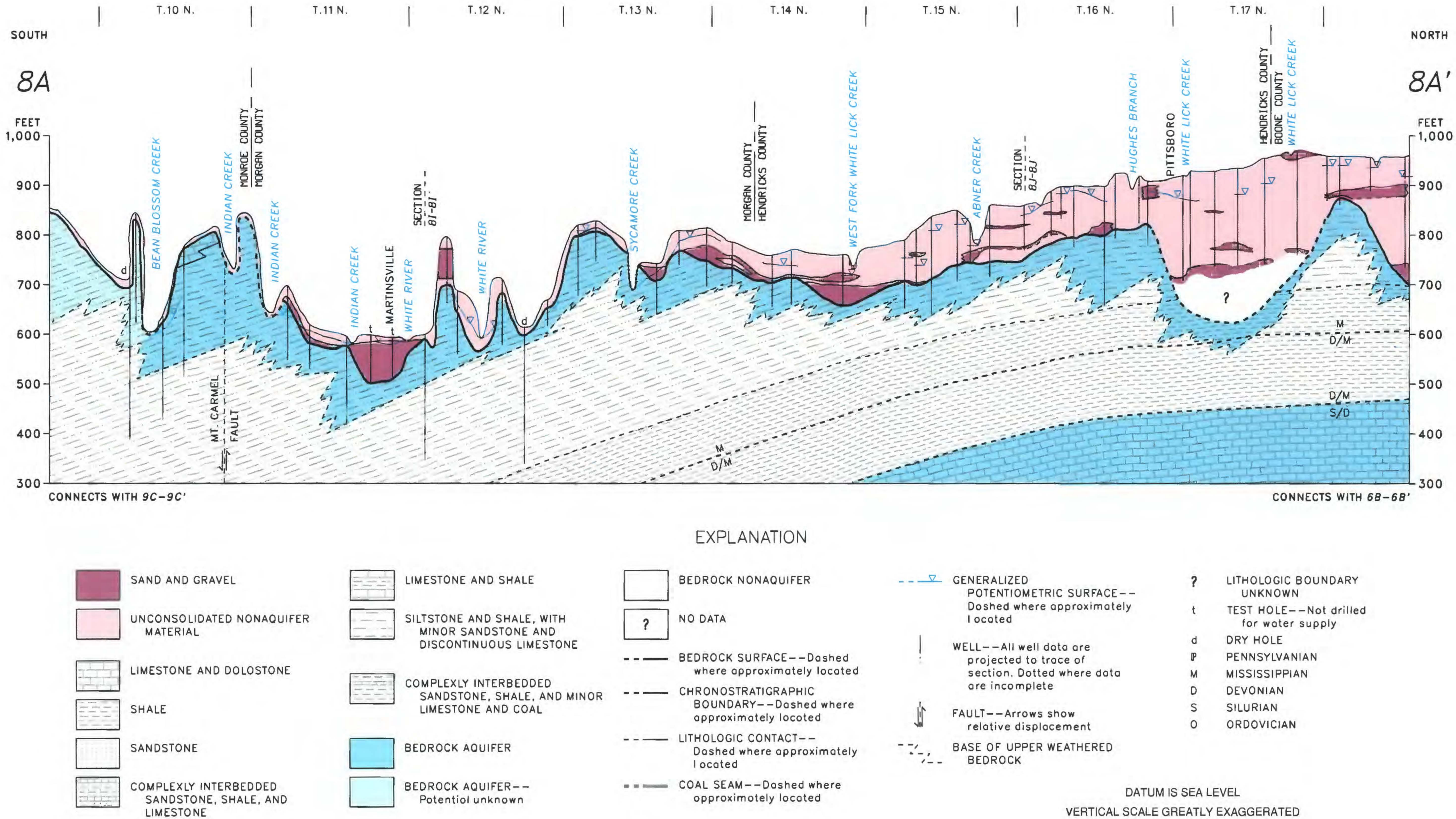
The surficial aquifer is generally unconfined along rivers (see section 8B–8B′ along the White River and Fall Creek, fig. 58). In places, the aquifer is hydraulically connected to buried sand and gravel aquifers that extend beneath the river, as shown in section 8D–8D′ (fig. 58) (Gillies, 1976, Smith, 1983). Recharge to the aquifer is from direct infiltration of precipitation and, at times, from the streams. The streams are connected hydraulically to the aquifer, usually gaining water from it; however, during drought or heavy pumping nearby, the streams can function as recharge sources for the aquifer (Gillies, 1976).

The thickness of the surficial sand and gravel aquifer ranges from 10 to more than 150 ft. Within the northern one-half of the basin, where the outwash aquifers have been studied extensively, the water table is generally within 10 ft of the surface. Saturated thickness, which ranges from 10 to 110 ft, depends on bedrock relief and thickness of the aquifer (Meyer and others, 1975; Smith, 1983). Hydraulic conductivities for the surficial sand and gravel aquifer range from 24 to greater than 1,500 ft/d (Arihood and Lapham, 1982; Smith, 1983). Well yields range from 10 to 2,000 gal/min (Meyer and others, 1975; Gillies, 1976; Smith, 1983; Barnhart and Middleman, 1990).

Buried and Discontinuous Sand and Gravel Aquifers

Buried and discontinuous sand and gravel aquifers have similar origins and exhibit similar characteristics, and therefore, are discussed here together. The major difference between the two aquifer types is that buried sand and gravel aquifers are thicker and areally more extensive than the discontinuous sand and gravel aquifers. Buried sand and gravel aquifers, used in the northern one-half of the White River basin (fig. 59), can be seen in sections 8A–8A′ to 8E–8E′ and in section 8K–8K′ (fig. 58). Discontinuous sand and gravel aquifers, used in the middle one-third of the White River basin (fig. 59), can be seen in section 8A–8A′ (northern one-third), 8I–8I′ (eastern one-quarter), and 8J–8J′ (western one-half) (fig. 58). The two aquifer types were deposited as outwash-plain deposits, valley fill in pre-Illinoian valleys, thin sheets of stratified drift, and small pockets of coarse-grained glaciolacustrine sediment (Watkins and Jordan, 1961, p. 6; Watkins and Jordan, 1962, p. 6; Watkins and Jordan, 1963, p. 6; Meyer and others, 1975, p. 7-9; Gillies, 1976, p. 4; Lapham, 1981, p. 10-31; Arihood, 1982, p. 8-23; Barnhart and Middleman, 1990, p. 9). Where buried sand and gravel deposits are continuous, they can be sources of large amounts of water. Discontinuous sand and gravel deposits tend to have low water yields; well contractors commonly drill through these deposits to obtain higher yields from the bedrock sources below (Watkins and Jordan, 1962, p. 6; Barnhart and Middleman, 1990, p. 9).

The buried and discontinuous sand and gravel aquifers are usually confined by layers of low-permeability till (see section 8A–8A′, fig. 58) (Watkins and Jordan, 1962, p. 6-7; Arihood and Lapham, 1982, p. 10-25). In some locations, the buried or discontinuous sand and gravel aquifers are contiguous with surficial sand and gravel aquifers along the major streams; together, the aquifers form a complex hydrogeologic system as shown in section 8C–8C′ (fig. 58) (Gillies, 1976, p. 9; Meyer and others, 1975, p. 9-16).



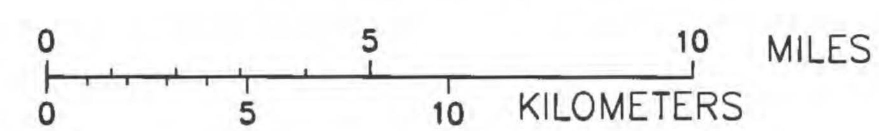


Figure 58. Hydrogeologic sections *8A–8A′* to *8K–8K′* of the White River basin.

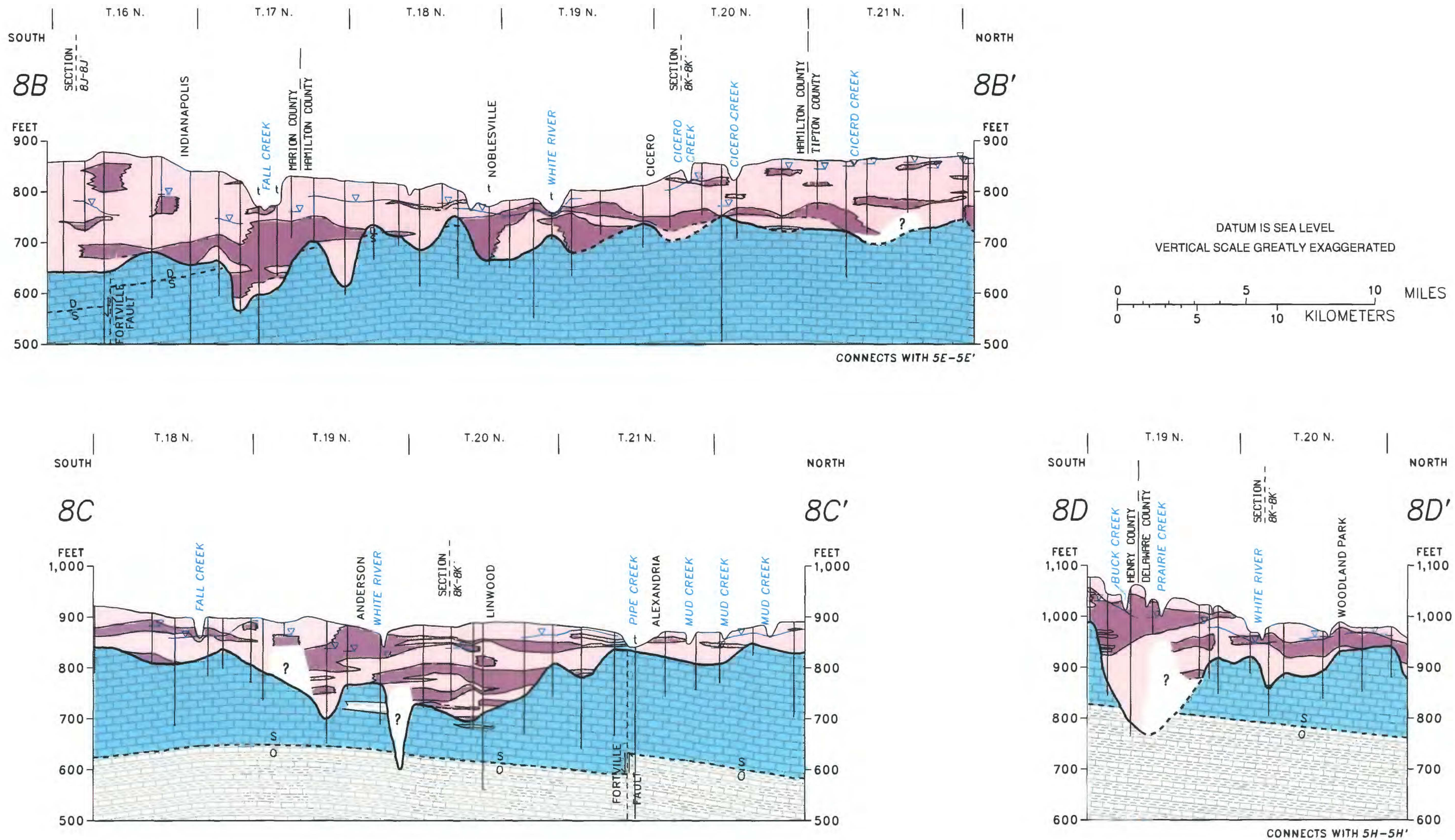
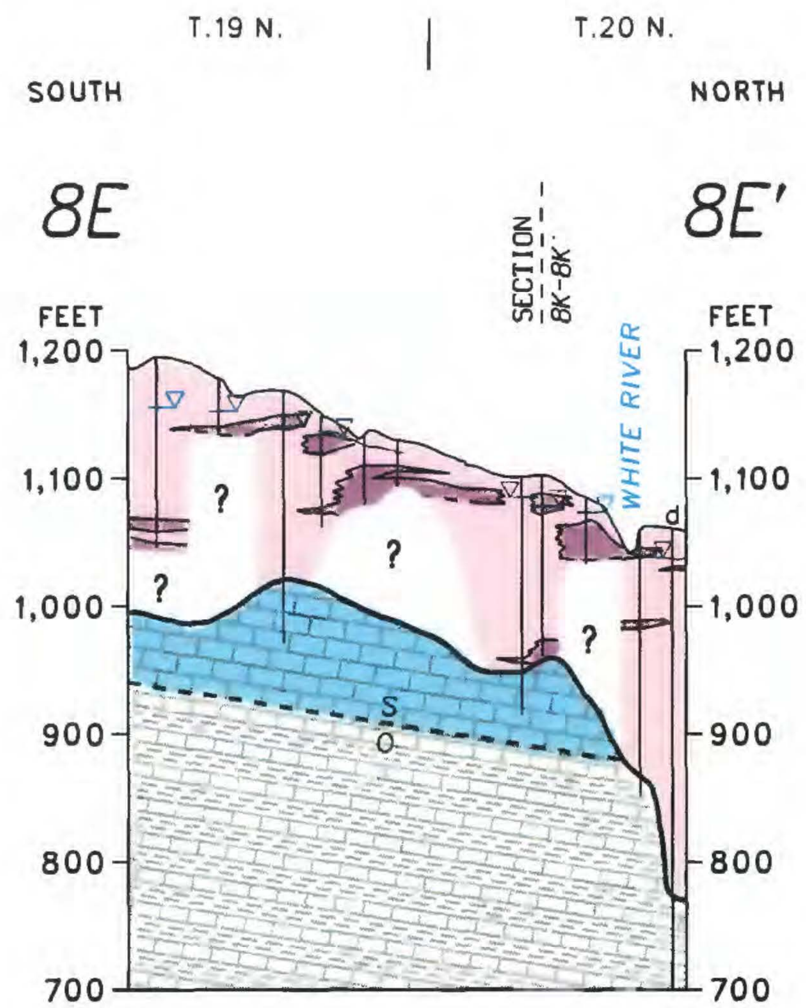
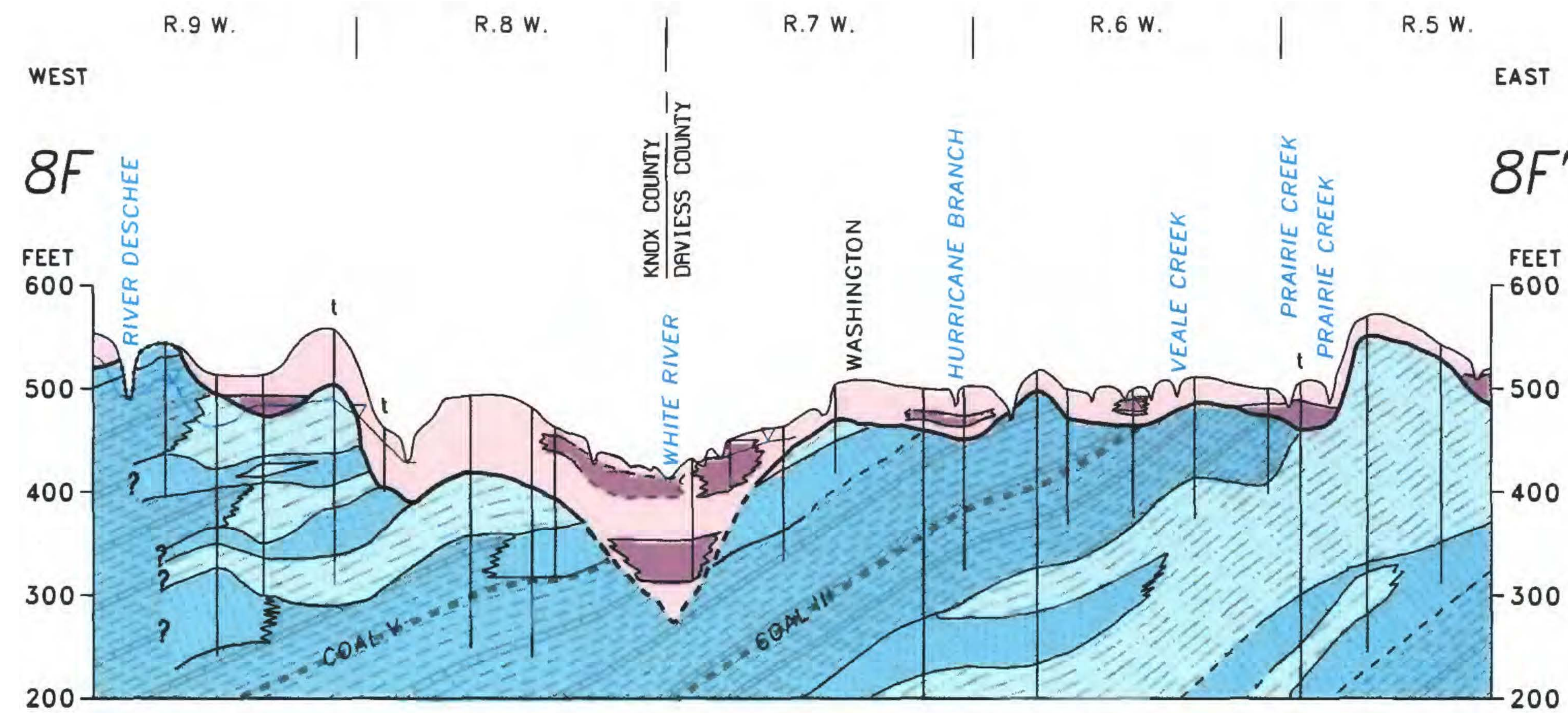
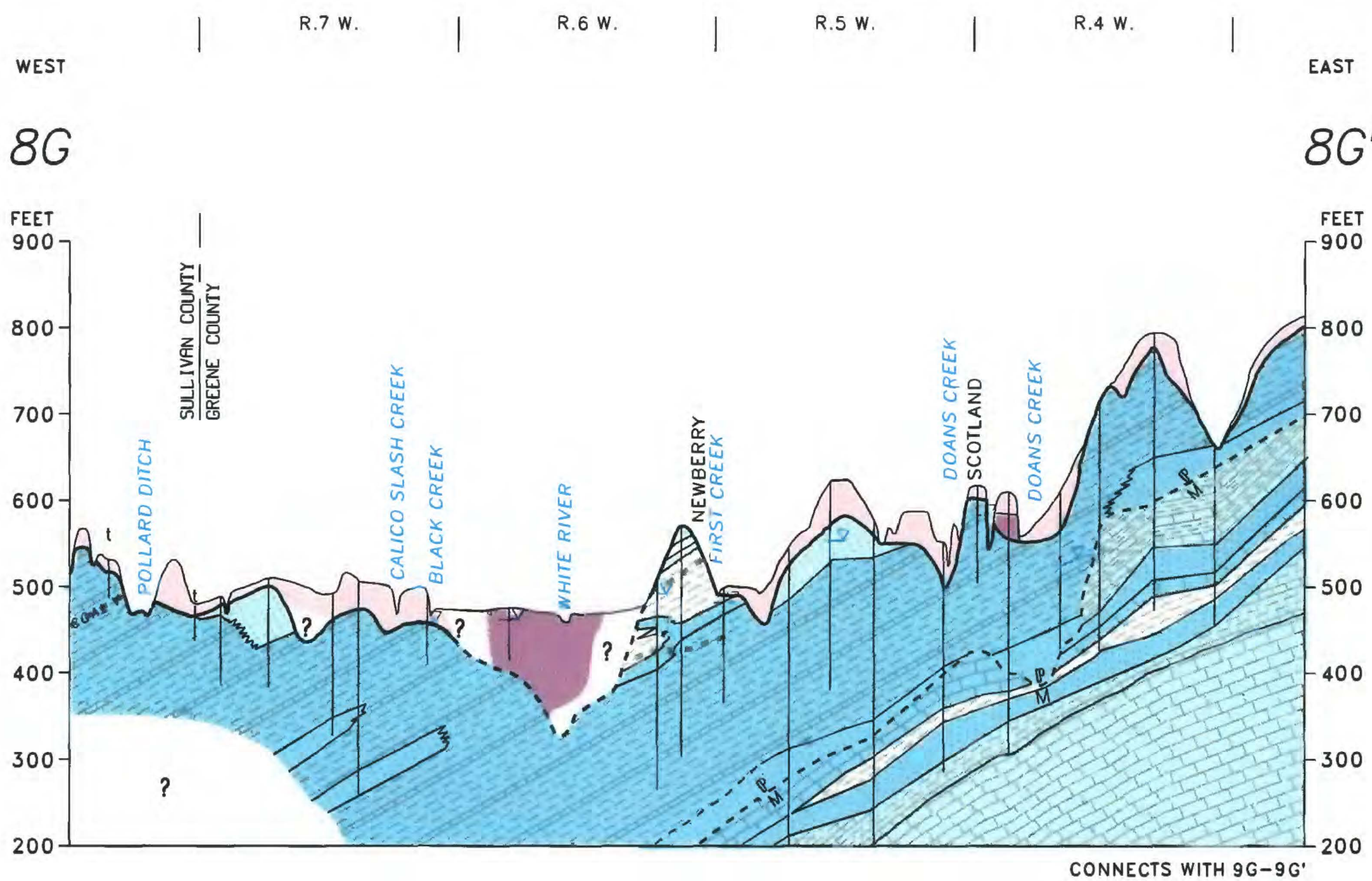


Figure 58. Hydrogeologic sections *8A–8A'* to *8K–8K'* of the White River basin—Continued.


T.19 N.
T.20 N.
SOUTH
NORTH
8E
8E′
SECTION 8K-8K′
WHITE RIVER
FEET
1,200
1,100
1,000
900
800
700
S
O


R.9 W.
R.8 W.
R.7 W.
R.6 W.
R.5 W.
WEST
EAST
8F
8F′
RIVER DESCHEE
KNOX COUNTY
DAVIESS COUNTY
WHITE RIVER
WASHINGTON
HURRICANE BRANCH
VEALE CREEK
PRAIRIE CREEK
PRAIRIE CREEK
FEET
600
500
400
300
200
COAL V
COAL III


R.7 W.
R.6 W.
R.5 W.
R.4 W.
WEST
EAST
8G
8G′
FEET
900
800
700
600
500
400
300
200
SULLIVAN COUNTY
GREENE COUNTY
POLLARD DITCH
CALICO SLASH CREEK
BLACK CREEK
WHITE RIVER
NEWBERRY
FIRST CREEK
DOANS CREEK
SCOTLAND
DOANS CREEK
CONNECTS WITH 9G–9G′

Recharge of buried and discontinuous aquifers is probably by infiltration of precipitation through the confining layers. Recharge rates reported by most of the studies of the White River basin were calculated from baseflow and static-water-level data (Lapham, 1981; Arihood and Lapham, 1982; Arihood, 1982; Lapham and Arihood, 1984). Such data can be used to determine recharge rates for buried or discontinuous aquifers if the aquifers are hydraulically linked to the stream. However, this linkage is not the case for many of the buried aquifers some distance from the surficial "outwash" aquifers, such as those shown in section 8K–8K′ (fig. 58). Because most of the buried and discontinuous aquifers are not regionally extensive, they have not been studied in detail; no information regarding recharge rates is available, other than average areal recharge rates for a particular modeled region. Arihood and Lapham (1982) calculated average areal recharge rates to the buried and discontinuous aquifers of 2 in/yr, or approximately 5 percent of the total precipitation, for a modeled region in the upper part of the White River basin.

The thickness of the buried and discontinuous sand and gravel aquifers ranges from 5 to 50 ft in most of the counties in the northern part of the basin (Lapham and Arihood, 1984, p. 11). Reported hydraulic conductivities of the confined buried and discontinuous aquifers range from 200 to 390 ft/d (Cable and others, 1971; Meyer and others, 1975). Many hydrologic studies in the northern part of the basin were based on the assumption that the average hydraulic conductivities of the buried and discontinuous sands and gravels were similar to those of the surficial sands and gravels, namely 433 ft/d (Arihood and Lapham, 1982; Lapham and Arihood, 1984; Lapham, 1981). Well yields of buried and discontinuous aquifers typically range from 10 to 250 gal/min (Herring, 1971, 1974).

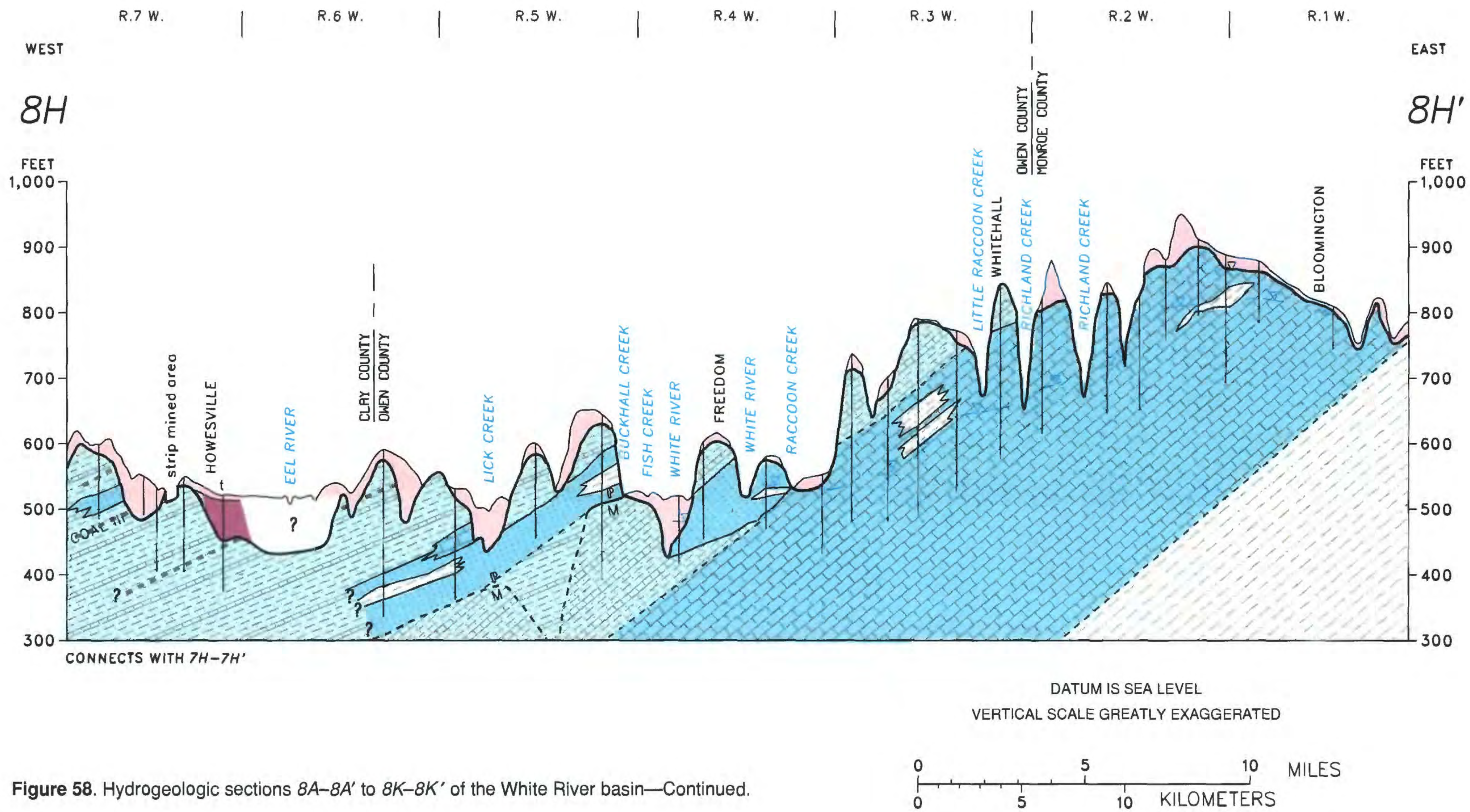


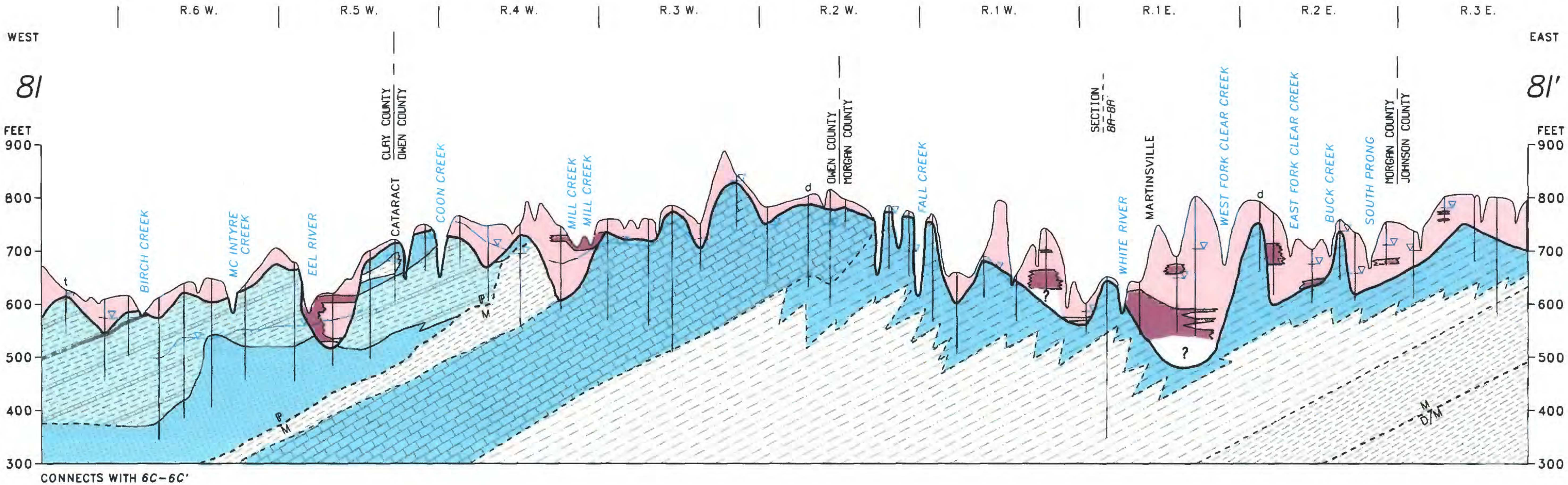
Figure 58. Hydrogeologic sections *8A–8A′* to *8K–8K′* of the White River basin—Continued.

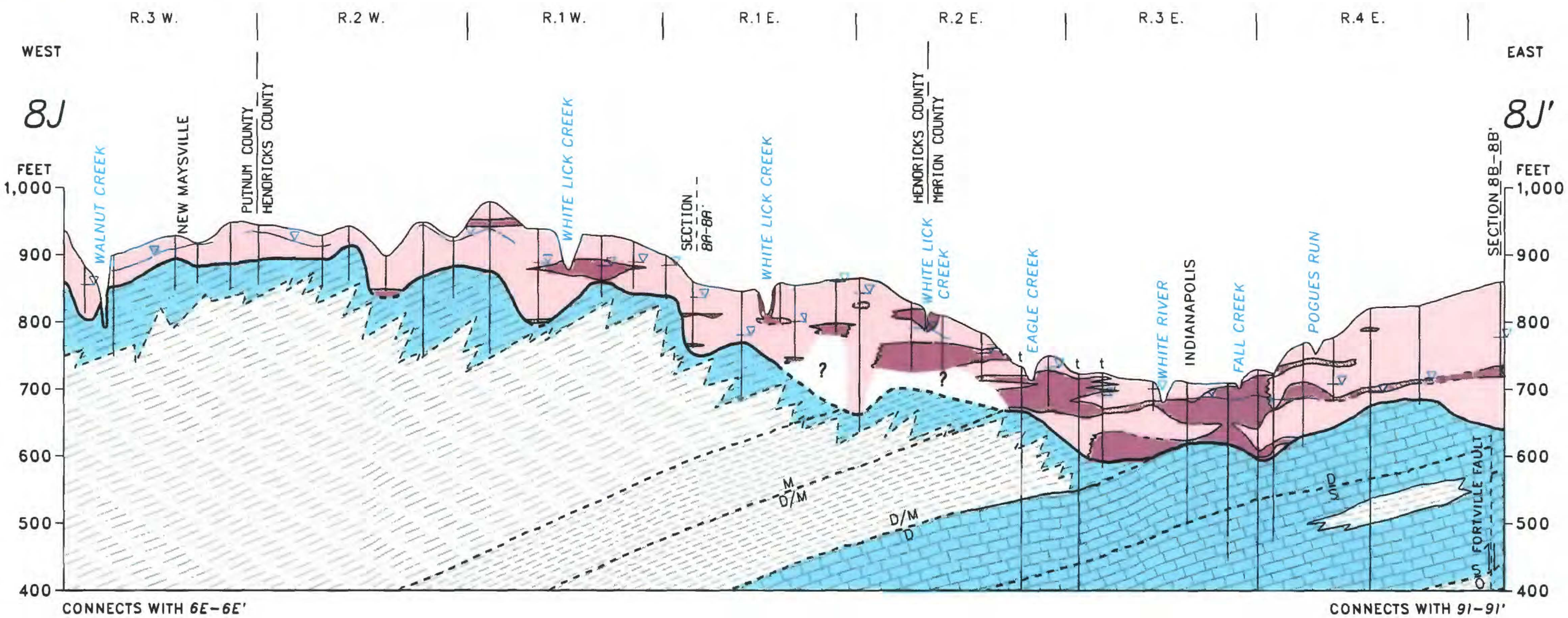
Bedrock Aquifers

Carbonate Bedrock Aquifers

Carbonate bedrock aquifers are present in the northern one-third of the White River basin and in a north-south band that is about 15 to 20 mi wide near the middle of the basin (fig. 59). Carbonate bedrock aquifers are shown in all hydrogeologic sections (fig. 58) except section 8F–8F′.

In the northern part of the basin, Ordovician shales and limestones of the Maquoketa Group are overlain by thick carbonate rock sequences with shale-dominant facies of Silurian and Devonian age (sections 8B–8B′ to 8E–8E′ and section 8K–8K′, fig. 58) (Wayne, 1975, p. 16-17; Lapham and Arihood, 1984). The upper Ordovician rocks of the Maquoketa Group consist of a large proportion of carbonate rock in the northeastern part of the basin (Gray, 1972) and are adequate for domestic water supplies in some places; however, Silurian and Devonian carbonate bedrock aquifers are preferred to Ordovician aquifers as water sources. The Silurian and Devonian carbonate rocks, which are now covered by glacial deposits, were once exposed and underwent some karst development (Wayne, 1966, p. 30). Because the primary permeabilities of the carbonate rocks tend to be low, it is this weathered zone within the carbonate rocks that is most likely to produce significant amounts of water, owing to solution-enhanced bedding planes, joints, and fractures (Lapham and Arihood, 1984, p. 10).


R.6 W.
R.5 W.
R.4 W.
R.3 W.
R.2 W.
R.1 W.
R.1 E.
R.2 E.
R.3 E.
WEST
EAST
8I
8I'
FEET
900
800
700
600
500
400
300
BIRCH CREEK
MC INTYRE CREEK
EEL RIVER
CATARACT
CLAY COUNTY
OWEN COUNTY
COON CREEK
MILL CREEK
MILL CREEK
OWEN COUNTY
MORGAN COUNTY
FALL CREEK
SECTION 8A-8A'
WHITE RIVER
MARTINSVILLE
WEST FORK CLEAR CREEK
EAST FORK CLEAR CREEK
BUCK CREEK
SOUTH PRONG
MORGAN COUNTY
JOHNSON COUNTY
CONNECTS WITH 6C-6C'


R.3 W.
R.2 W.
R.1 W.
R.1 E.
R.2 E.
R.3 E.
R.4 E.
WEST
EAST
8J
8J'
FEET
1,000
900
800
700
600
500
400
WALNUT CREEK
NEW MAYSVILLE
PUTNUM COUNTY
HENDRICKS COUNTY
WHITE LICK CREEK
SECTION 8A-8A'
WHITE LICK CREEK
HENDRICKS COUNTY
MARION COUNTY
WHITE LICK CREEK
EAGLE CREEK
WHITE RIVER
INDIANAPOLIS
FALL CREEK
POGUES RUN
SECTION 8B-8B'
FORTVILLE FAULT
CONNECTS WITH 6E-6E'
CONNECTS WITH 9I-9I'

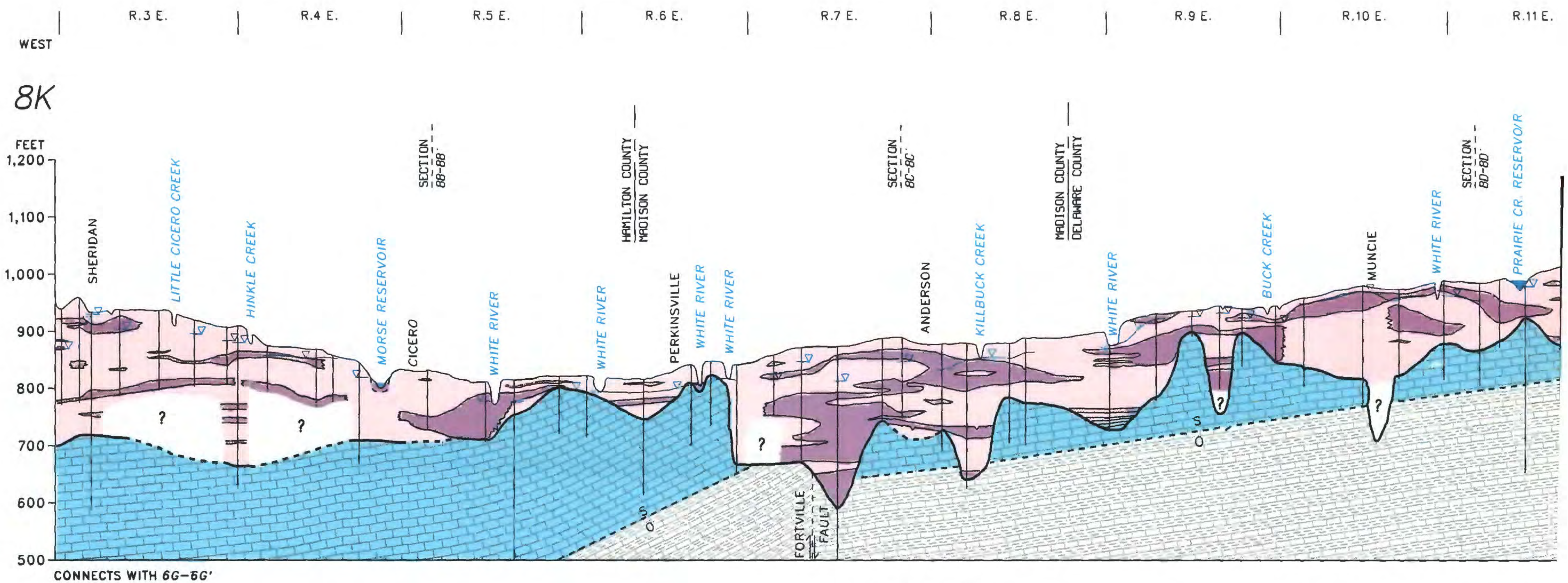


Figure 58. Hydrogeologic sections *8A–8A′* to *8K–8K′* of the White River basin—Continued

DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED

0 5 10 MILES
0 5 10 KILOMETERS

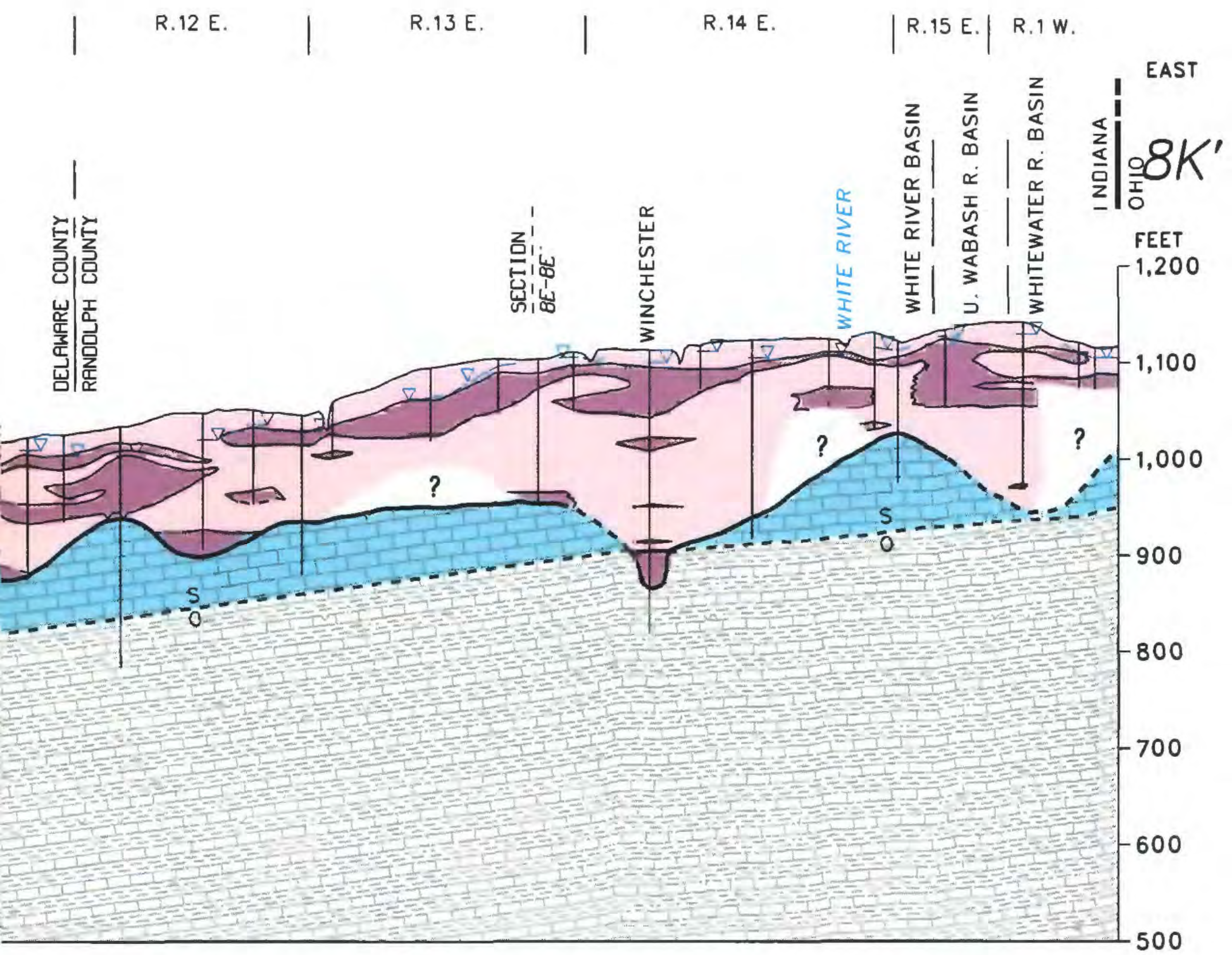
Wayne (1975), in a Madison County report, states that nearly all of the rocks within the Silurian and Devonian Systems will yield water. Specific rock units that are particularly good water producers include the Salamonie Dolomite and the Louisville Limestone (Wayne, 1975, p. 16).

Recharge to the carbonate bedrock aquifers is mostly by infiltration and percolation of rainwater through the overlying glacial deposits. Thicknesses of specific carbonate bedrock aquifers within the Ordovician, Silurian, and Devonian systems range from 40 to 300 ft, but only the upper 150 ft is generally tapped (Arihood, 1982, p. 8). The water-bearing capability of the Silurian and Devonian aquifers is chiefly dependent on the fracture density and degree of weathering. Because of this, the hydraulic conductivity of these aquifers is highly variable. Cable and others (1971) estimated the average hydraulic conductivity of the aquifers to be 13.4 ft/d. Well yields of more than 100 gal/min are possible from these aquifers (Steen, 1970; Wayne, 1975, p. 16).

Other carbonate bedrock aquifers within the basin include the Mississippian Blue River and Sanders Groups. The carbonate rocks are well-cemented, dense, medium-bedded limestones; ground water commonly flows along fractured and weathered surfaces. Intense karst development in the limestone of these groups is common where they are exposed at the surface. Flow of ground water through the fracture and joint systems enhances preexisting avenues of water flow. Recharge of these aquifers is by infiltration of precipitation through thin glacial deposits, exposed bedrock fracture surfaces, and karst terrain. Wells within these aquifers generally yield less than 30 gal/min, and dry holes are not uncommon (Bechert and Heckard, 1966, p. 108-109).

Upper Weathered-Bedrock Aquifer

In the central one-third of the basin, aquifers are developed in an upper weathered zone of the Devonian and Mississippian New Albany Shale and siltstones and shales of the Mississippian Borden Group. This aquifer type is shown in hydrogeologic sections 8A–8A′, 8I–8I′, and 8J–8J′ (fig. 58). The upper weathered zone is a zone of enhanced permeability produced by weathering before, during, and after glaciation. The availability of water in this weathered zone is highly variable and is dependent on the degree of enhanced permeability, the type and thickness of overlying deposits, and the bedrock topography. The dependence on type and thickness of overlying deposits is evident in hydrogeologic section 8A–8A′ (fig. 58) where, as glacial deposits thin toward the south, dry wells are increasingly common. Where the aquifer is unreliable, the weathered zone is mapped as "aquifer—potential unknown." This boundary is located near the maximum extent of glaciation (fig. 55 and 59).



The shale-siltstone upper weathered-bedrock aquifer is used primarily for domestic and stock water supplies in areas where no other aquifers are available. The exact thickness of the weathered-bedrock aquifer is unknown but is inferred by the depth of the wells that are completed in it. These depths range from 20 to greater than 200 ft, but the zone of enhanced permeability is generally limited to the upper 150 ft (table 10). Because shales and siltstones are generally considered to be confining units, hydraulic conductivities are thought to be low; owing to secondary permeability caused by weathering, however, the actual value is unknown. Well yields range from 0 to 10 gal/min (Bechert and Heckard, 1966; Clark, 1980).

Complexly Interbedded Sandstone, Shale, Limestone, and Coal Aquifer

In the southwestern part of the basin, laterally discontinuous basal sandstones and to a lesser degree, limestones and coals are the principal aquifers (Cable and Robison, 1973, p. 8-9). These aquifers are contained within complexly interbedded sandstones, siltstones, shales, limestones and coals of Mississippian and Pennsylvanian age. The complexly interbedded sequence is shown in hydrogeologic sections 8F–8F′ to 8I–8I′ (fig. 58). The sections may or may not show individual aquifer units within the complexly interbedded sequence depending on well density and(or) detail of the well logs. The coals and limestones are typically less than 10 ft thick and can serve as useful stratigraphic markers. Shales and siltstones are generally much thicker, but variable in thickness as well.

Because aquifers within the complexly interbedded sequences are discontinuous, water-bearing capabilities are variable and can be assessed only on a local basis. These complexly interbedded sequences are therefore mapped as "aquifer—potential unknown" in the hydrogeologic sections and on the aquifer map (fig. 59). Wells finished in the complexly interbedded bedrock are usually not screened but are open throughout the length of the well; it is not always possible, therefore, to identify the unit that is the source of water. Well yields from the complexly interbedded aquifers tend to average about 5 gal/min and rarely exceed 20 gal/min (Bechert and Heckard, 1966, p. 108-109; Cable and Robison, 1973, p. 23). Hydraulic conductivities are probably low also.

Sandstone Aquifers

Most of the sandstones are sheetlike deposits or sinuous channel sandstones (Cable and Robison, 1973) that range from less than 20 to 100 ft. Thin discontinuous sandstones are combined with other shales, siltstones, limestones, and coal deposits and mapped as "aquifer—potential unknown." The more extensive sandstones are shown in hydrogeologic sections 8F–8F′ to 8I–8I′ (fig. 58) where they are mapped as aquifers (fig. 59). These sandstones produce greater yields than do the thin, discontinuous sandstones within the complexly interbedded deposits. The most frequently used sandstone aquifer is the lower Pennsylvanian Mansfield Formation (Thomas, 1980). This sandstone, confined above and below by shales, ranges from 20 to 100 ft in thickness in Clay County (Thomas, 1980, p. 14). Other sandstones that are considered to be aquifers are in the Linton Formation and the Petersburg Formation of Middle Pennsylvanian age (Cable and others, 1971, p. 11). Recharge to these sandstone aquifers occurs where the formations crop out at the surface, primarily in the southern, unglaciated parts of the basin.

Permeability of most of the sandstones is low, and yields from wells that tap any of the relatively continuous sandstone aquifers are correspondingly low; maximum yield is 30 gal/min, and average yield is 10 gal/min (Cable and others, 1971; Cable and Robison, 1973).

Summary

Several large cities, including Indianapolis, and all or parts of 27 counties lie within the White River basin. The basin contains unconsolidated glacial deposits which overlie bedrock that ranges in age from Ordovician to Pennsylvanian. The unconsolidated deposits consist of clay-rich, loamy, tills interbedded with stratified sand and gravel, as well as sand and gravel deposited as outwash along the major streams. A variety of lithologies are present in the bedrock system. Limestones and shales dominate the rocks of Ordovician, Silurian, Devonian, and early Mississippian age. Almost all sedimentary lithologies are present in the Late Mississippian and Pennsylvanian Systems.

Seven different aquifer types have been identified within the basin: three unconsolidated aquifer types and four bedrock aquifer types. The most productive aquifers are the surficial sands and gravels. Wells completed in this type of aquifer can yield as much as 2,000 gal/min; such wells are major water sources for Indianapolis, Anderson, and Muncie. The surficial sand and gravel aquifers are generally unconfined, are variable in thickness, and have high hydraulic conductivities.

Buried and discontinuous sand and gravel aquifers are commonly used where the drift is thick. The hydrologic character of these aquifers is similar to surficial sand and gravel aquifers, but the aquifer is confined by relatively impermeable till layers.

Carbonate rocks form the primary bedrock aquifer in the northern one-third and the west-central part of the basin. Well yields are moderate to high in the northern part of the basin, ranging from 20 gal/min to greater than 600 gal/min, but recharge rates are probably low because recharge occurs by infiltration and percolation of rainwater through the overlying fine-grained glacial deposits. Yields from the carbonate bedrock aquifer in the west-central part of the basin are lower than in the north, ranging from 0 to 20 gal/min.

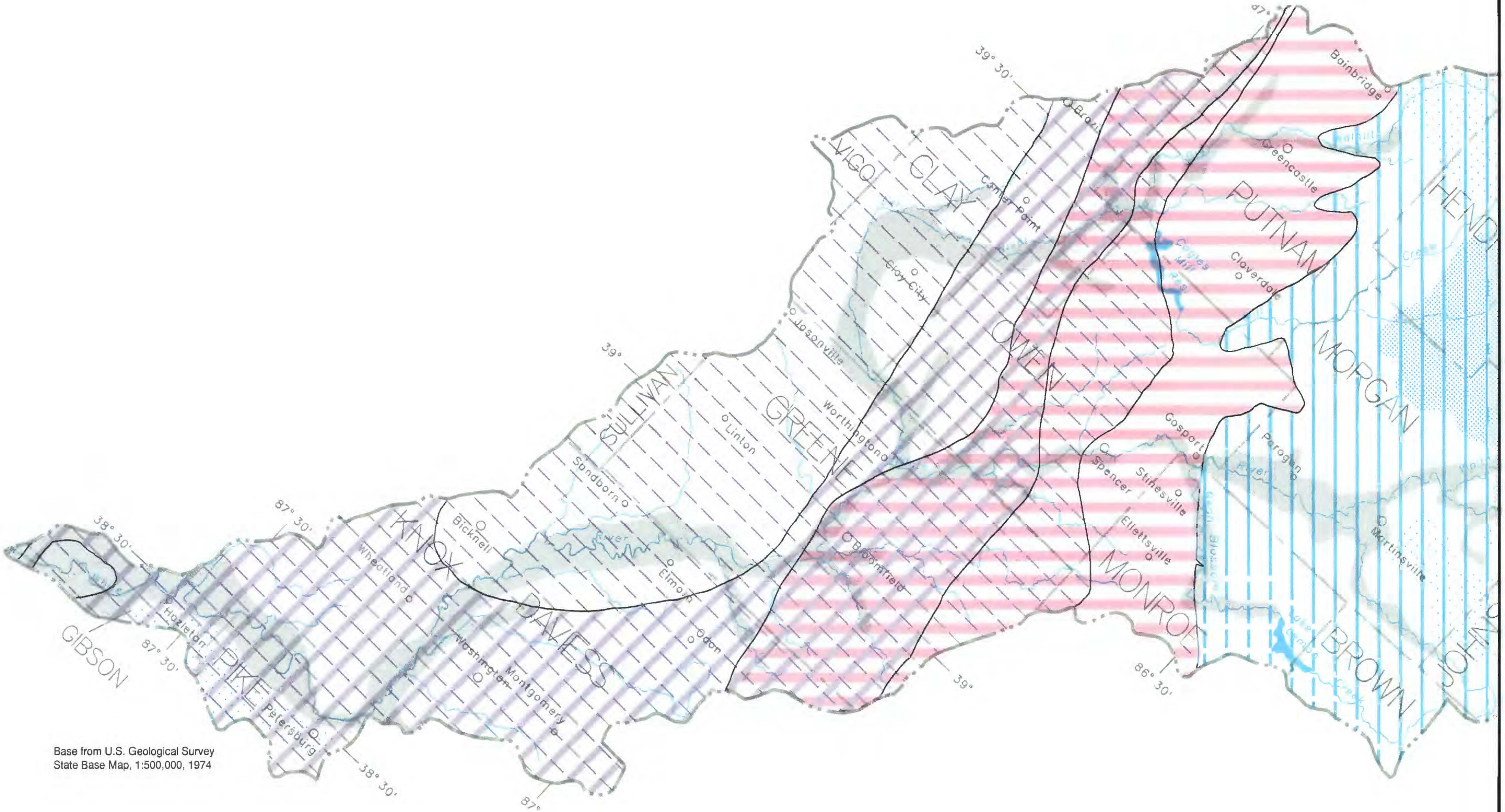


Figure 59. Extent of aquifer types in the White River basin.

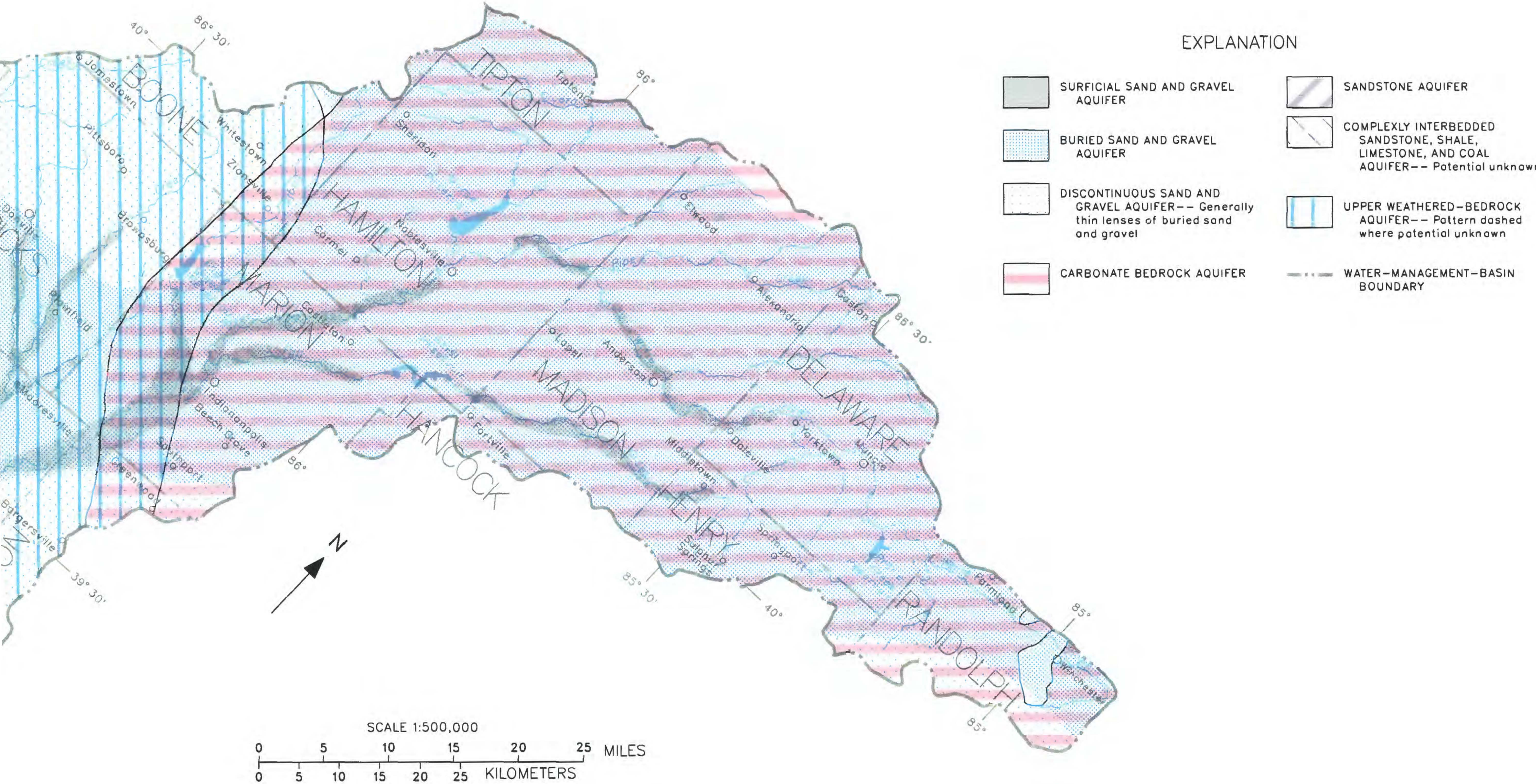

EXPLANATION
SURFICIAL SAND AND GRAVEL AQUIFER
BURIED SAND AND GRAVEL AQUIFER
DISCONTINUOUS SAND AND GRAVEL AQUIFER-- Generally thin lenses of buried sand and gravel
CARBONATE BEDROCK AQUIFER
SANDSTONE AQUIFER
COMPLEXLY INTERBEDDED SANDSTONE, SHALE, LIMESTONE, AND COAL AQUIFER-- Potential unknown
UPPER WEATHERED-BEDROCK AQUIFER-- Pattern dashed where potential unknown
WATER-MANAGEMENT-BASIN BOUNDARY
BOONE
TIPTON
HAMILTON
MARION
MADISON
HANCOCK
DELAWARE
HENRY
RANDOLPH
Jamestown
Pittsboro
Brownsburg
Whitestown
Zionsville
Sheridan
Tipton
Carmel
Noblesville
Elwood
Alexandria
Gaston
Castleton
Lapel
Anderson
Indianapolis
Beech Grove
Southport
Greenwood
Mooresville
Plainfield
Bargersville
Fortville
Middletown
Daleville
Yorktown
Muncie
Sulphur Springs
Spiceland
Farmland
Winchester
40°
86° 30'
86°
85° 30'
85°
39° 30'
N
SCALE 1:500,000
0 5 10 15 20 25 MILES
0 5 10 15 20 25 KILOMETERS

Table 10. Characteristics of aquifer types in the White River basin
[<, less than; locations of aquifer types shown in fig. 59]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and gravel	5- 150	[1,2,3]100- 2,000	Outwash, alluvium, valley train[4,5,6]
Buried sand and gravel	5- 90	[1,2,3]50- 300	Interbedded sand and gravel, outwash plain[4,5]
Discontinuous sand and gravel	5- 40	[1,2,5]10- 200	Interbedded sand and gravel, outlier[4,5]
Carbonate bedrock			
Mississippian	[7]150	[1,2]<20	Sanders and Blue River Groups[8]
Devonian	[7]150	[1,2]100- 600	Muscatatuck Group[8]
Silurian	[7]150	[1,2]200- 600	Salamonie Dolomite, Brassfield Limestone, Cataract Formation, and Salina Group[8]
Upper weathered bedrock	[7]150	[1,2]0- 10	Borden Group and New Albany Shale[8]
Complexly interbedded sandstone, shale, limestone, and coal	highly variable[9]	[1,2]5- 20	West Baden, Stephensport, Raccoon Creek, and Carbondale Groups, and Patoka Formation[8]
Sandstone	20- 100	[1,2,10]5- 20	Raccoon Creek Group[8]

[1]Bechert and Heckard, 1966.
[2]Clark, 1980.
[3]Herring, 1971; 1974.
[4]Arihood and Lapham, 1982.
[5]Barnhart and Middleman, 1990.
[6]Watkins and Jordan, 1961; 1962; 1963.
[7]Reported thickness is not total thickness of unit but thickness of unit considered permeable or water bearing.
[8]Shaver and others, 1986.
[9]Water is commonly found in thin beds within complexly interbedded sequence.
[10]Thomas, 1980.

Complexly interbedded rocks of different lithologies are used as aquifers in the southern one-third of the basin, but yields from these aquifers are generally low. The major water producers within the complexly interbedded sequence are thin sandstones but limestones and coals also can produce water.

Relatively continuous sandstone units, mostly within the Pennsylvanian System, such as the Mansfield Formation, are used as aquifers and are mapped as a separate aquifer type. Well yields from sandstone aquifers are slightly higher than from the complexly interbedded aquifers.

In the central part of the basin, the only source of usable quantities of water is a weathered zone within shale and siltstone. These rocks have sufficient secondary permeability to serve as a source of water, but only for small supplies. Well yields range from 0 to 10 gal/min within this aquifer type.

References Cited

Arihood, L.D., 1982, Ground-water resources of the White River basin, Hamilton and Tipton Counties, Indiana: U.S. Geological Survey Water-Resources Investigations 82-48, 69 p.

Arihood, L.D., and Lapham, W.W., 1982, Ground-water resources of the White River basin, Delaware County, Indiana: U.S. Geological Survey Water-Resources Investigations 82-47, 69 p.

Arvin, D.V., 1989, Statistical summary of streamflow data for Indiana: U.S. Geological Survey Open-File Report 89-62, 964 p.

Bailey, Z.C., and Imbrigiotta, T.E., 1982, Ground-water resources of the glacial outwash along the White River, Johnson and Morgan Counties, Indiana: U.S. Geological Survey Water-Resources Investigations 82-4016, 87 p.

Barnhart, J.R., and Middleman, B.H., 1990, Hydrogeology of Gibson County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 41, 17 p.

Bechert, C.H., and Heckard, J.M., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 100-115.

Bleuer, N.K., 1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley System (so-called Teays Valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

Cable, L.W. and Robison, T.M., 1973, Hydrogeology of the principal aquifers in Sullivan and Greene Counties, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 35, 26 p.

Cable, L.W., Watkins, F.A., Jr., and Robison, T.M., 1971, Hydrogeology of the principal aquifers in Vigo and Clay Counties, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 34, 34 p.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Gillies, D.C., 1976, Availability of ground water near Carmel, Hamilton County, Indiana: U.S. Geological Survey Water-Resources Investigations 76-46, 27 p.

Gray, H.H., 1972, Lithostratigraphy of the Maquoketa Group (Ordovician) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 7, 31 p.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Heckard, J. M., 1965, Water resources of Morgan County, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

Herring, W.C., 1971, Water resources of Hamilton County, with emphasis on ground-water availability: Indiana Department on Natural Resources, Division of Water, 1 sheet.

_____1974, Water resources of Marion County, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

_____1976, Technical atlas of the ground-water resources of Marion County, Indiana: Indiana Department of Natural Resources, Division of Water, 53 p.

Hoggatt, R.E., Hunn, J.K., and Steen, W.J., 1968, Water resources of Delaware County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 37, 57 p.

Jacques, D.V., and Crawford, C.G., 1991, National water-quality assessment program—White River basin: U.S. Geological Survey Open-File Report 91-169, 2 p.

Lapham, W.W., 1981, Ground-water resources of the White River basin, Madison County, Indiana: U.S. Geological Survey Water-Resources Investigations 81-35, 112 p.

Lapham, W.W., and Arihood, L.D., 1984, Ground-water resources of the White River basin, Randolph County, Indiana: U.S. Geological Survey Water-Resources Investigations Report 83-4267, 86 p.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 20, part 2, p. 59-256.

Meyer, William, Reussow, J.P., and Gillies, D.C., 1975, Availability of ground water in Marion County, Indiana: U.S. Geological Survey Open-File Report 75-312, 87 p.

Nyman, D.J., and Pettijohn, R.A., 1971, Ground water, *in* Wabash River Coordinating Committee, Wabash River basin comprehensive study: U.S. Geological Survey and U.S. Army Corps of Engineers, Louisville District, volume VIII, appendix G, 100 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Smith, B.S., 1983, Availability of water from the outwash aquifer, Marion County, Indiana: U.S. Geological Survey Water-Resources Investigations Report 83-4144, 70 p.

Steen, W.J., 1970, Water resources of Madison County, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

Steen, W.J., Bruns, T.M., and Perry, A.O., 1977, Water resources of Boone County, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

Thomas, J.M., 1980, Water chemistry and flow system of the early Pennsylvanian Mansfield Formation aquifer system in Clay County, Indiana: Bloomington, Ind., Indiana University, Master's thesis, 76 p.

Thornbury, W.D., 1950, Glacial sluiceways and lacustrine plains of southern Indiana: Indiana Department of Conservation, Division of Geology Bulletin 4, 21 p.

Uhl, J.E., 1966, Water resources of Johnson County, with emphasis on ground water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

Wangsness, D.J., Miller, R.L., Bailey, Z.C., and Crawford, C.G., 1981, Hydrology of area 32, eastern region, Interior Coal Province, Indiana: U.S. Geological Survey Water-Resources Investigations/ Open-File Report 81-498, 76 p.

Watkins, F.A., Jr., 1965, Wabash basin comprehensive study, ground-water appraisal of the Big Walnut Creek basin above Little Walnut Creek and Big Walnut Creek Reservoir Site, Indiana: U.S. Geological Survey Open-File Report, 11 p.

Watkins, F.A., Jr., and Jordan, D.G., 1961, Ground-water resources of west-central Indiana—preliminary report, Greene County: Indiana Department of Natural Resources, Division of Water Bulletin 11, 255 p.

_____1962, Ground-water resources of west-central Indiana—preliminary report, Clay County: Indiana Department of Natural Resources, Division of Water Bulletin 16, 309 p.

_____1963, Ground-water resources of west-central Indiana—preliminary report, Owen County: Indiana Department of Natural Resources, Division of Water Bulletin 18, 99 p.

Wayne, W.J., 1966, Ice and land, a review of Tertiary and Pleistocene history of Indiana, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 21-39.

_____1975, Urban geology of Madison County, Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 10, 24 p.

EAST FORK WHITE RIVER BASIN

By Joseph M. Fenelon and Theodore K. Greeman

General Description

The East Fork White River basin, located in south-central Indiana, extends from the southwestern to the east-central part of the State. The basin has an area of 5,746 mi^2, and its long axis trends northeast-southwest for a distance of approximately 150 mi. The East Fork White River basin includes all, or part of, the following counties: Bartholomew, Brown, Daviess, Decatur, Dubois, Hancock, Henry, Jackson, Jefferson, Jennings, Johnson, Lawrence, Marion, Martin, Monroe, Orange, Pike, Ripley, Rush, Scott, Shelby and Washington. Principal cities in the basin include Bedford, Bloomington, Columbus, Franklin, Greenfield, Greensburg, Loogootee, New Castle, North Vernon, Rushville, Seymour, and Shelbyville (fig. 60).

Previous Studies

The only ground-water study that describes the hydrogeology of the entire East Fork White River basin was done by Nyman and Pettijohn (1971). The report is a brief description of the important aquifers in the basin, and includes information on well yields and potential yields, ground-water quality, and ground-water discharge to the major streams in the basin. A major study by the U.S. Geological Survey is currently (1991-97) being done for the East Fork White and White River basins as part of a National Water-Quality Assessment Program. The study will assess the water quality of the surface- and ground-water resources of the East Fork White and White River basins (Jacques and Crawford, 1991). Generalized ground-water availability maps have been completed for the entire state of Indiana by Clark (1980) and Bechert and Heckard (1966).

A number of publications contain information on localized hydrogeology of the eastern half of the basin. These publications include a series of county ground-water-availability maps, which emphasize the reported and potential well yields from the major aquifers in the northeastern counties of the East Fork White River basin. These maps, published by the Indiana Department of Natural Resources, Division of Water, are for Shelby (Bruns and Uhl, 1976), Hancock (Uhl, 1975), Henry (Uhl, 1973), Johnson (Uhl, 1966), and Marion (Herring, 1974) Counties. Other publications describing the ground-water resources of Marion County are by Roberts and others (1955), Meyer and others (1975), and Herring (1976).

Hydrogeologic studies in or near Columbus, Bartholomew County, have defined the ground-water resources of that area (Klaer and Kingsbury, 1948; Klaer and others, 1951), mapped the glacial outwash aquifer along the Flatrock River and East Fork White River (Davis and others, 1969), and modeled ground-water availability (Watkins and Heisel, 1970). Ground-water models have also simulated water-level declines that might result from different arrangements of municipal water-supply wells for the cities of Columbus and Taylorsville (Planert, 1976; Planert and Tucci, 1979).

Ground-water resources in three watersheds in the northeastern one-third of the basin were evaluated to determine the effects of proposed reservoirs upon the hydrology of the Big Blue River (Nyman and Watkins, 1965a), the Flatrock River (Nyman and Watkins, 1965b), and Clifty Creek (Watkins, 1964). A complete hydrologic balance of Summit Lake, in the headwaters of the Big Blue River, was determined by Duwelius, (1993) for water years 1989 and 1990 (a water year begins October 1 and ends September 30, the following year).

Two publications on the southeastern part of the basin provide detailed maps of lineament and fracture-trace locations in Jennings County (Greeman, 1981) and Decatur County (Greeman, 1983). These studies describe the bedrock aquifers and explain the hydrologic significance of the mapped lineaments and fracture traces to ground-water well yield.

A brief description of the aquifers in the southwestern one-fifth of the basin is included in Wangsness and others (1981). Ruhe (1975) studied the Lost River watershed to investigate the connection between surface-water and ground-water flow in the karst terrain.

Physiography

Seven physiographic regions in the East Fork White River basin were defined by Malott (1922) and later refined by Wayne (1956) and Schneider (1966). The Tipton Till Plain (fig. 61), located in the northern one-fifth of the basin, is a nearly flat to gently undulating till plain. The southern boundary of the Tipton Till Plain is approximate and is located where drift thickness obscures the underlying bedrock physiography. The remainder of the basin is within six bedrock-dominated physiographic units that trend approximately north-south, paralleling the regional bedrock strike (fig. 61).

The easternmost physiographic unit in the East Fork White River basin is the Muscatatuck Regional Slope. The eastern boundary of the physiographic unit roughly coincides with the eastern boundary of the drainage basin. The Muscatatuck Regional Slope has a westerly dip of approximately 400 ft over 25 mi or 0.17 degree (Schneider, 1966, p. 43). The slope is controlled by the regional dip of the Silurian and Devonian carbonate bedrock. In general, river valleys are deeply entrenched along joints and fracture zones in the carbonate bedrock, and commonly make near-right-angle turns.

The Scottsburg Lowland is west of the Muscatatuck Regional Slope. The lowland is a 10- to 20-mi-wide trough with little relief and is underlain by Devonian and Mississippian shales. Pre-Wisconsinan glaciers followed the Scottsburg Lowland into southern Indiana and northern Kentucky. During the pre-Wisconsinan and later glacial advances and retreats, the Scottsburg Lowland became a principal discharge route for meltwater and outwash. North of Scottsburg, this lowland is now filled with outwash deposits ranging from 50 to more than 100 ft in thickness (fig. 63).

Further west, in the central part of the East Fork White River basin, is the Norman Upland. The Norman Upland is separated from the Scottsburg Lowland by the Knobstone Escarpment, which stands as much as 300 ft above the Scottsburg Lowland. (See hydrogeologic section 9G–9G′, R. 3 E., fig. 64.) This escarpment is capped by sandy siltstones that are more resistant to weathering than underlying Devonian and Mississippian shales. The upland is generally flat topped but thoroughly dissected by steep-sloped stream valleys (Schneider, 1966, p. 45). The escarpment also marks the location of a major change in bedrock dip, which becomes steeper to the west. (This change in bedrock dip is discussed in the "Bedrock Geology" section.)

The Mitchell Plain, lying to the west of the Norman Upland, is underlain by Mississippian limestones. The area is a low-relief karst plain that is intensely pitted in some areas by thousands of sinkholes. Surface drainage is poorly developed because of the extensive internal drainage. Most of the precipitation and some of the rivers drain underground through swallow holes.

The Crawford Upland is underlain by complexly interbedded Mississippian and Pennsylvanian sandstones, shales, and limestones, which cause the topography to be very diverse. The area is a westward-sloping, deeply dissected upland with local relief of as much as 350 ft (Schneider, 1966, p. 48).

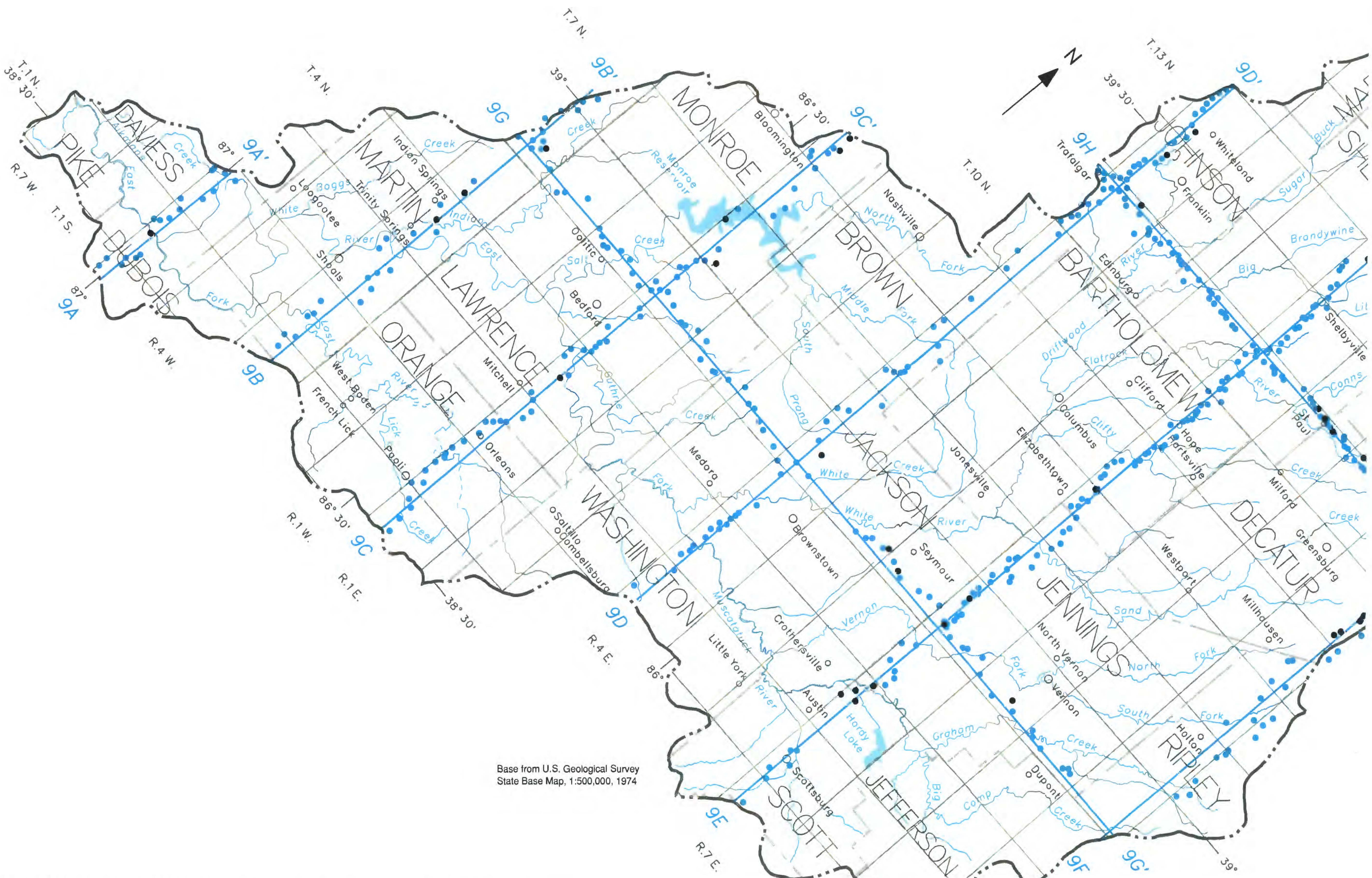
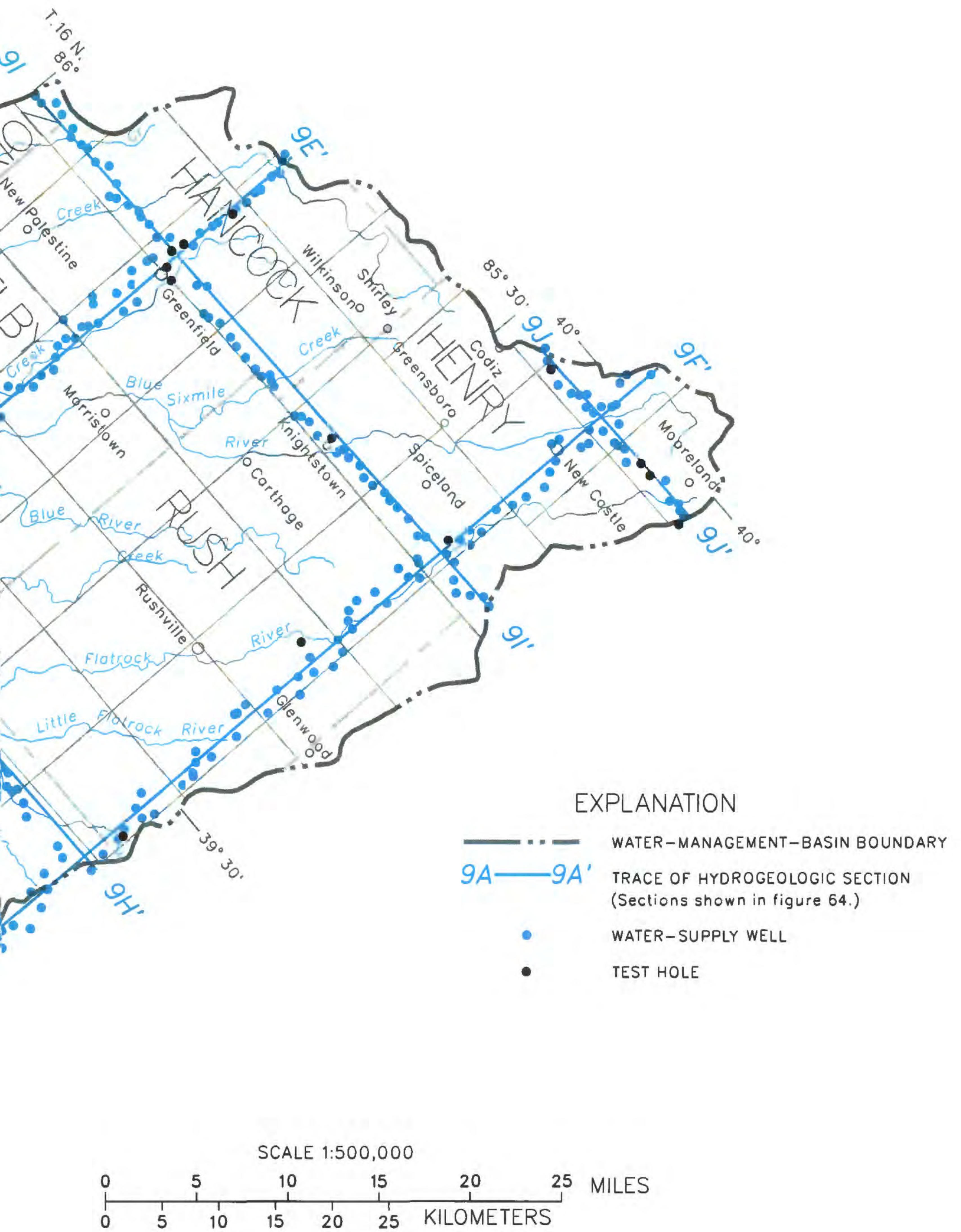


Figure 60. Location of section lines and wells plotted in the East Fork White River basin.



The westernmost physiographic region in the basin is the Wabash Lowland. Land surface elevations in the Wabash Lowland are 300 to 400 ft below the top of the Crawford Upland (Schneider, 1966, p. 48). The lowland is underlain by Pennsylvanian siltstones, sandstones, and shales and is covered by thin glacial drift within the basin. The Wabash Lowland is generally characterized by low relief and gentle slopes (Schneider, 1966, p. 49).

Surface-Water Hydrology

Most of the rivers in the East Fork White River basin drain to the southwest because of the regional slope of the bedrock. The East Fork White River, which begins at the confluence of the Driftwood and Flatrock Rivers, is the largest river in the basin (fig. 60). From its origin at Columbus to its mouth in the southwest corner of the basin, the East Fork White River flows 239 mi (Hoggatt, 1975, p. 58). The East Fork White River flows into the White River near Petersburg, Ind. (figs. 1 and 54).

Major tributaries to the East Fork White River with drainage areas greater than 500 mi^2 (fig. 60) include (1) the Muscatatuck River, which drains the southeastern part of the basin; (2) Salt Creek, which drains the west-central part of the basin; and (3) the Driftwood River, Flatrock River, and Big Blue River, which drain the northern part of the basin. Drainages in the basin that are from 100 to 500 mi^2 in drainage area include the Lost River, Sugar Creek, Graham Creek, Clifty Creek, Big Creek, Indian Creek, White Creek, Brandywine Creek, and the Little Blue River (fig. 60).

Rivers in the eastern half of the basin have a subparallel drainage pattern that reflects the regional dip of the bedrock. The rivers exhibiting subparallel drainage down the regional bedrock slope are Sugar Creek, the Big Blue River, the Little Blue River, the Flatrock River, Clifty Creek, Sand Creek, Vernon Fork (both North and South Forks), Graham Creek, and the East Fork White River from Medora to Jonesville. These rivers flow southwest, into the Scottsburg Lowland (fig. 61), which is bounded on the west by the Knobstone Escarpment.

Only two rivers in the East Fork White River basin breach this escarpment. The East Fork White and the Muscatatuck Rivers breach the escarpment about 15 mi apart in Jackson County. Flowing southwest and west, respectively, from their cuts through the escarpment, the East Fork White River is joined by the Muscatatuck River near Medora.

Drainage of the Mitchell Plain (fig. 61) in northeast Orange County, central Lawrence County, and Monroe County is considerably different from the rest of the basin; most runoff quickly leaves the surface by entering sinkholes and becoming part of the ground-water system. In the streams that do flow across the Mitchell Plain, some surface water is intercepted by swallow holes and diverted underground into either the ground-water system or subterranean channels. For example, in Orange County, the Lost River loses flow in a series of swallow holes between R. 1 W. and 1 E., T. 2 N. (fig. 60). The water then flows through underground channels and reemerges 7 mi to the west and 168 ft lower (Ruhe, 1975, p. 33).

Monroe Reservoir and Hardy Lake are the two principal lakes in the basin (fig. 60). They were formed from rivers that were dammed to provide flow regulation, water supply, and recreation. Monroe Reservoir is the largest maximum-capacity reservoir in Indiana (second largest normal capacity) with a surface area of 16.8 mi^2 (Ruddy and Hitt, 1990, p. 99-103).

Geology

Bedrock Deposits

The East Fork White River basin is southwest of the Cincinnati Arch (fig. 4). Bedrock dips to the southwest into the Illinois Basin at approximately 20 ft/mi in the northeastern part of the basin, as determined from the mapped top of the Ordovician rocks (Bassett and Hasenmueller, 1980). In the southwestern part of the basin, the dip of the shallow bedrock increases to about 43 ft/mi as measured from

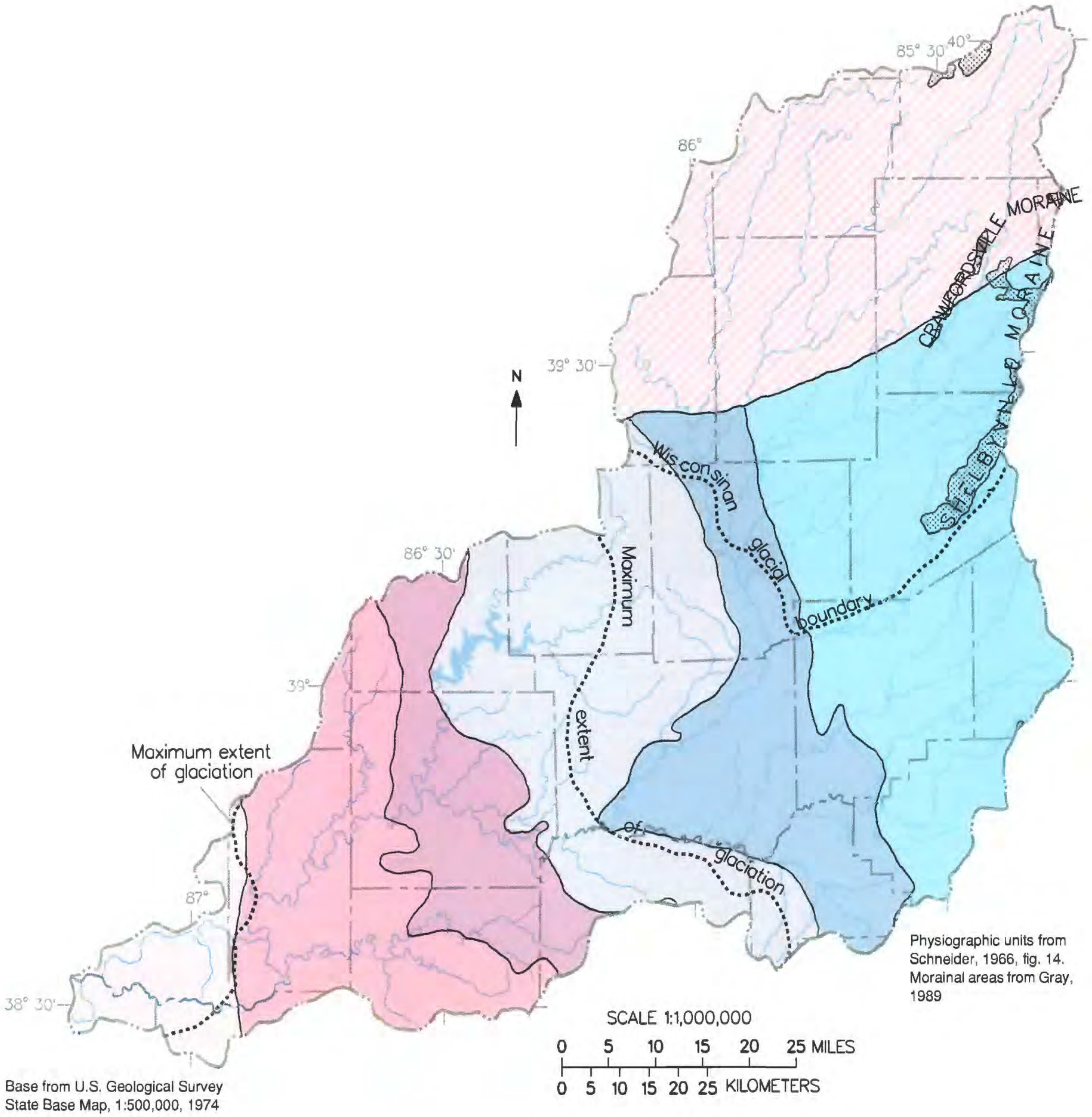


Figure 61. Physiographic units, moraines, and extent of glaciation in the East Fork White River basin.

the mapped top of the West Baden Group of Mississippian age (Geosciences Research Associates, 1982), whereas the dip of deeper Ordovician rocks increases to greater than 60 ft/mi in the same area (Bassett and Hasenmueller, 1980).

The Mt. Carmel Fault trends north-northwest in Monroe, Lawrence, and northwestern Washington Counties; the southern 50 mi of the fault is within the basin (fig. 62). This fault functions as a hinge line on the east side of the Illinois Basin, with steeper bedrock dips west of the fault (see section 9G–9G′, fig. 64). The western side of the fault is downthrown approximately 100 to 200 ft (Shaver and Austin, 1972, p. 11 and 20). Locally, shorter parallel faults (about 5 mi in length) are present (Shaver and Austin, 1972, p. 4).

Rocks of Ordovician through Pennsylvanian ages are present at the bedrock surface in the East Fork White River basin. The oldest rocks at the bedrock surface underlie thick drift in buried bedrock valleys in the far northeastern part of the basin and are exposed in streambeds in the southeastern part of the basin (fig. 62). These Ordovician rocks are the Dillsboro and Whitewater Formations of the Maquoketa Group (fig. 5). The Maquoketa Group consists of thin interbedded shale and limestone and is more that 400 ft thick in the basin.

Silurian rocks, which overlie the Ordovician rocks, are present along the eastern edge of the basin (fig. 62). Silurian formations in the basin are (from oldest to youngest) the Brassfield Limestone, the Salamonie Dolomite, the Waldron Shale, the Louisville Limestone, and the Wabash Formation (fig. 5). The Waldron Shale and Louisville Limestone form the Pleasant Mills Formation in the northern one-third of the basin. The Silurian rocks are composed primarily of limestone, dolomite, dolomitic limestone, and minor amounts of shale and chert. The Silurian rocks have a combined thickness of 90 to 500 ft within the East Fork White River basin (Hasenmueller and Bassett, 1980; and Bassett and Hasenmueller, 1980). The Waldron Shale is a thin (0 to 12 ft thick) shale that hydrologically separates the underlying Silurian carbonate rocks from the overlying Silurian and Devonian carbonates (Greeman, 1981, p. 6). Pre-Devonian erosion thinned the upper part of the Silurian rocks near the Cincinnati Arch. In the northern part of the basin, only the lower 50 ft of the Wabash Formation remains; further south, postdepositional erosion removed all of the Wabash Formation and the underlying Louisville Limestone and Waldron Shale (Schneider and Gray, 1966).

The Devonian Muscatatuck Group unconformably overlies the Silurian rocks. The Muscatatuck Group, in areas of outcrop, consists of 50 to 90 ft of dolomite and limestone and small amounts of anhydrite and gypsum (Shaver and others, 1986, p. 99; Gray and others, 1985). Devonian carbonate rocks are present at the bedrock surface in more than 1,000 mi^2 of the eastern part of the East Fork White River basin, although they have been eroded from the extreme eastern edge of the basin (fig. 62). The combined thickness of the Silurian and Devonian carbonate rocks range from 90 ft in the eastern part to about 1,000 ft in the southwestern part of the basin (Geosciences Research Associates, 1982, pl. 21).

The Silurian and Devonian carbonate rocks are overlain by the Devonian and Mississippian New Albany Shale. This greenish-gray to black, fissile shale crops out in a 5- to 20-mi-wide northwest-trending band in the Scottsburg Lowland (east-central part of the basin) (fig. 62). Eroded or not deposited across the Cincinnati Arch, the New Albany Shale ranges from 85 to 150 ft in thickness in the East Fork White River basin. The shale is considered a confining unit, greatly restricting the connection between surface water and ground water in the underlying carbonate bedrock aquifer.

Rocks of Mississippian age include the Rockford Limestone and the Borden, Sanders, Blue River, West Baden, and Stephensport Groups (fig. 5). The Rockford Limestone, averaging 3 ft in thickness, is a widespread marker bed that separates the New Albany Shale from the overlying Borden Group. The Borden Group is a thick (500 to 800 ft) unit with a north-northwest trending outcrop area of almost 1,000 mi^2 in the central part of the basin (fig. 62). The Borden Group is composed of siltstone and shale interbedded with some sandstone and minor limestone; the lower 200 ft is primarily shale (Shaver and others, 1986, p. 17-18). The Borden Group underlies the Norman Upland and crops out along the eastern edge of the Knobstone Escarpment.

Cropping out to the west and overlying the Borden Group are the Sanders and Blue River Groups (fig. 62). Both groups are primarily carbonate rocks that contain minor amounts of chert, shale, siltstone, anhydrite, gypsum, and calcareous sandstone (Shaver and others, 1986, p. 16 and 137). These Mississippian carbonate rocks range in thickness from about 350 ft in Monroe County to 550 ft in Orange County. In some areas, the thick-bedded carbonate rocks are quarried for fine building stone. Other horizons contain geodes, joints, and solution features that make them unsuitable for quarrying. Underlying the Mitchell Plain, the Sanders and Blue River Groups have well-developed karst solution features (sinkholes and caves) throughout much of their outcrop area.

The youngest Mississippian rocks in the basin are the West Baden and Stephensport Groups. Both groups are composed of shale, sandstone, and limestone; however, the West Baden Group is dominated by shale and sandstone. The West Baden Group is 100 to 120 ft thick in the East Fork White River basin, and the Stephensport Group is 130 to 150 ft thick (Gray and others, 1985). The West Baden and Stephensport Groups underlie the eastern half of the Crawford Upland. An erosional surface with as much as 150 ft of local relief (Shaver and others, 1986, p. 86) marks the Mississippian-Pennsylvanian boundary.

Pennsylvanian rocks above the erosional surface include the Raccoon Creek Group and the Carbondale Group. They are found in a small area in the far southwestern corner of the basin (fig. 62). The Raccoon Creek Group is 150 to 500 ft thick and is 95 percent shale and sandstone, the remainder consisting of clay, coal, and limestone (Shaver and others, 1986, p. 120-121). Shale is more common than sandstone in the Raccoon Creek Group, even though a 50- to 185-ft-thick sandstone, the Mansfield Formation, is at the base (Shaver and others, 1986, p. 87 and 121). The youngest rocks in the basin are in the Carbondale Group. The group is composed mostly of shale and sandstone, but it contains some thin, but laterally extensive, limestone beds and economically important coal beds (Shaver and others, 1986, p. 27). The Carbondale Group is typically less than 300 ft thick in the basin.

The geologic record from the end of the Pennsylvanian Period to the Quaternary Period is missing. This hiatus could represent either a nondepositional period or sediments that were deposited and later eroded. At the beginning of the Quaternary period, preglacial rivers draining the eastern half of the East Fork White River basin flowed southwest down the bedrock slope and into the lowland developed on the New Albany Shale (Scottsburg Lowland). These preglacial rivers drained into one river, which flowed south along the lowland. Near Seymour, this predecessor of the East Fork White River turned west through a low gap in the escarpment separating the Scottsburg Lowland from the Norman Upland. The main difference between preglacial and postglacial drainage in the eastern half of the basin, is the thick deposits of glacial drift now filling the Scottsburg Lowland. Most of the western half of the basin is unglaciated, and the present-day drainages are similar to those of preglacial time except for raised channel levels caused by the addition of valley fill.

The bedrock surface in the far northeastern part of the basin, in Henry County and northern Rush County, indicates a north-flowing preglacial stream. This buried bedrock valley is part of the Lafayette Bedrock Valley System (fig. 7). Hydrogeologic sections 9F–9F′ and 9J–9J′ (fig. 64) show relief on the buried bedrock surface exceeding 300 ft in northeastern Henry County.

Unconsolidated Deposits

More than two-thirds of the East Fork White River basin was glaciated during the Pleistocene Period. Pre-Wisconsinan glaciers covered the northeastern two-thirds and the extreme downstream end of the basin (fig. 61). Wisconsinan ice overrode the earlier glacial deposits in the northeastern one-third of the basin. Three general areas characterized by different types of surficial deposits are (1) the unglaciated part of the basin, (2) the glaciated area south of the Wisconsinan glacial boundary, and (3) the glaciated area north of the Wisconsinan glacial boundary (fig. 61).

The unglaciated part of the basin is in the western one-third but excludes the far downstream end. Unconsolidated deposits in the unglaciated area are mostly soils that have developed on the underlying bedrock. Exposed bedrock types include siltstone, shale, carbonate rock, and sandstone. Residual reddish-brown soils developed on the carbonate rocks can be as thick as 50 ft (Gray, 1989). Most of the area however, is covered by thin deposits of soils and loess that are generally from 5 to 20 ft thick (fig. 63). Exceptions to this are the valley-train outwash deposits along the East Fork White River and Salt Creek where thicknesses of silt, sand, and alluvium can be as much as 100 ft. (See hydrogeologic section 9B–9B′, fig. 64.)

The central one-third of the basin and the far western part were glaciated only by pre-Wisconsinan glaciers (fig. 61). The pre-Wisconsinan glaciated area is mantled by a complex mix of deposits. Streams have exposed bedrock in many places throughout the pre-Wisconsinan part of the East Fork White River basin (Gray, 1989). Drift overlying the bedrock throughout much of the area is a deeply weathered loam to sandy-loam till of the Jessup Formation, the oldest Pleistocene unit recognized in Indiana (Schneider and Gray, 1966, p. 23). Comprised of two till members of pre-Wisconsinan age, the Jessup Formation is typically only a few tens of feet thick and rests directly on bedrock (Schneider and Gray, 1966, p. 23). Overlying bedrock and older till in some areas is a poorly stratified combination of weathered bedrock, sand, silt, and loess that has accumulated by mass wasting, stream deposition, and windblown deposition (Gray, 1989).

In general, the unconsolidated deposits in the central one-third of the basin are less than 50 ft thick (fig. 63). Exceptions can be found along the East Fork White River and part of the Muscatatuck River, where thicknesses range from 50 to more than 100 ft. The East Fork White River flows in a 3-mi-wide glacial drainageway. This drainageway was filled when outwash from Wisconsinan and pre-Wisconsinan glaciers was deposited in the river valley. Recent alluvial deposits of silt, sand, and gravel overlie the outwash sand and gravel. Sand dunes and blanket sand deposits are present along the eastern side of the East Fork White River channel in Bartholomew and Jackson Counties.

The northern one-third of the basin was initially glaciated by pre-Wisconsinan glaciers that deposited thick tills and some outwash of the Jessup Formation. During the Wisconsinan Age, loam till of the Trafalgar Formation was deposited in the basin. The Wisconsinan tills were deposited by the Huron-Erie ice lobe (fig. 8), which advanced out of the Lake Huron and Lake Erie basins to the northeast. During the Wisconsinan Age, ice advanced and retreated from the basin on several occasions, forming moraines. The Shelbyville Moraine forms part of the eastern boundary of the basin (fig. 61) and represents

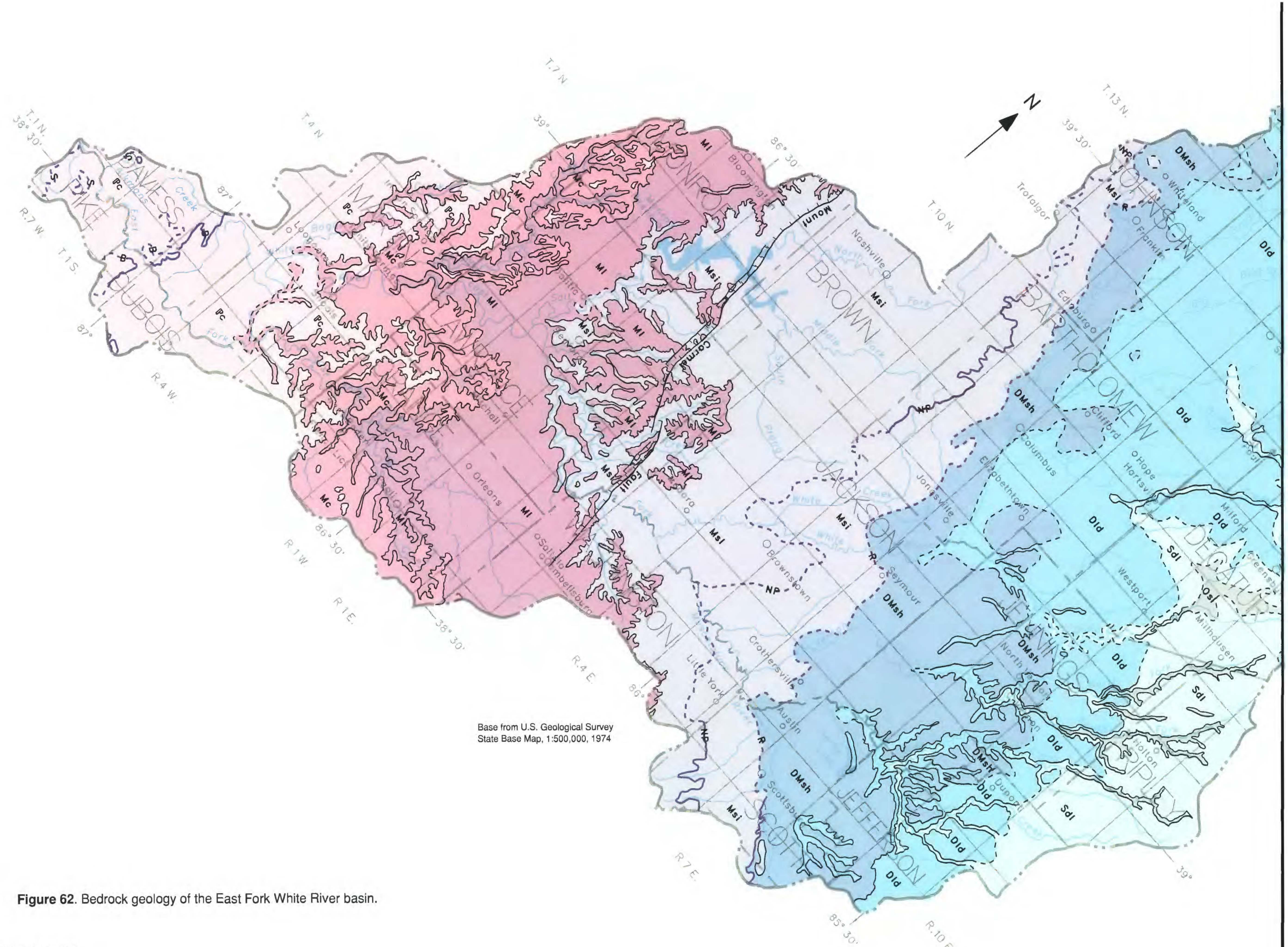
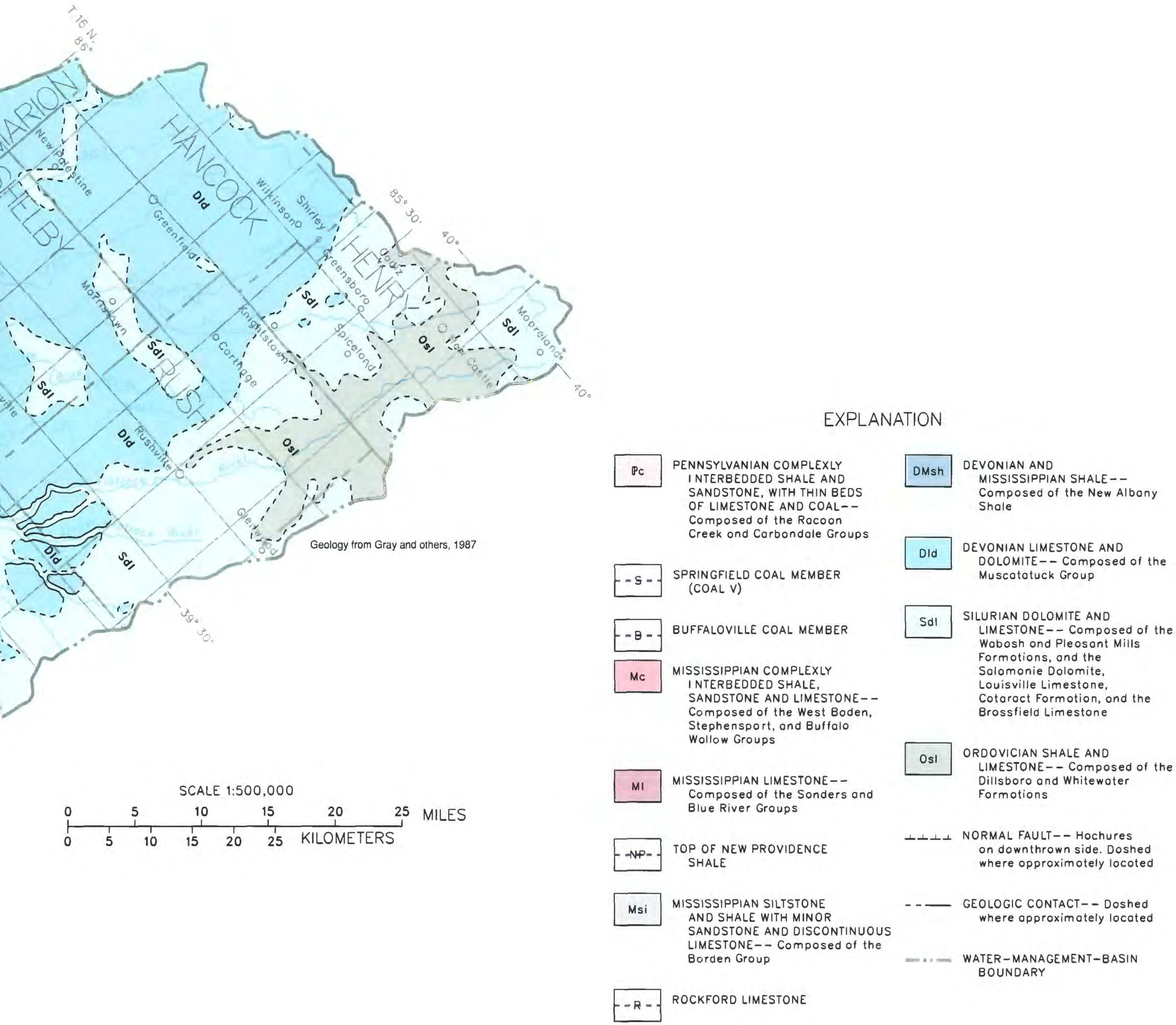


Figure 62. Bedrock geology of the East Fork White River basin.



the furthest Wisconsinan advance into the East Fork White River basin. The Shelbyville Moraine is well developed and low-lying, with only about 20 ft of relief (Schneider and Gray, 1966, p. 10). The Crawfordsville Moraine is also present in the basin, but it is not as well developed (fig. 61). The thickness of the Wisconsinan drift ranges from 0 to 150 ft and is typically 20 to 50 ft (Schneider and Gray, 1966, p. 20). Till generally thickens northward in the basin.

The total thickness of all unconsolidated deposits in the northeastern one-third of the basin ranges from 0 to more than 400 ft (fig. 63). Substantial deposits of sand and gravel are located along or near Sugar Creek, the Big Blue River, the Flatrock River, and in several abandoned channels. These deposits consist of glacial outwash and recent stream deposits (Nyman and Pettijohn, 1971, p. 24). The thickest unconsolidated deposits are in buried bedrock valleys and consist of till interbedded with sand and gravel. The deepest valley is in the northeastern part of the basin. (See hydrogeologic sections 9F–9F′ and 9J–9J′, fig. 64.)

Aquifer Types

Ten hydrogeologic sections (9A–9A′ to 9J–9J′, fig. 64) were constructed for this atlas to depict aquifer types in the East Fork White River basin. Hydrogeologic sections 9A–9A′ to 9F–9F′ are oriented south-north, whereas hydrogeologic sections 9G–9G′ to 9J–9J′ are oriented west-east (fig. 60). Almost 620 well logs were used to construct the sections; average density of logged wells plotted along the sections is 1.2 wells per mile. Information from the following authors aided in interpretation of well logs and construction of hydrogeologic sections: Pinsak (1957); Sullivan (1972); Bassett and Hasenmueller (1979, 1980); Bassett and Keith (1984); Hasenmueller and Bassett (1979, 1980); Gray (1982, 1983, 1989); Gray and others (1987); and Keller (1990).

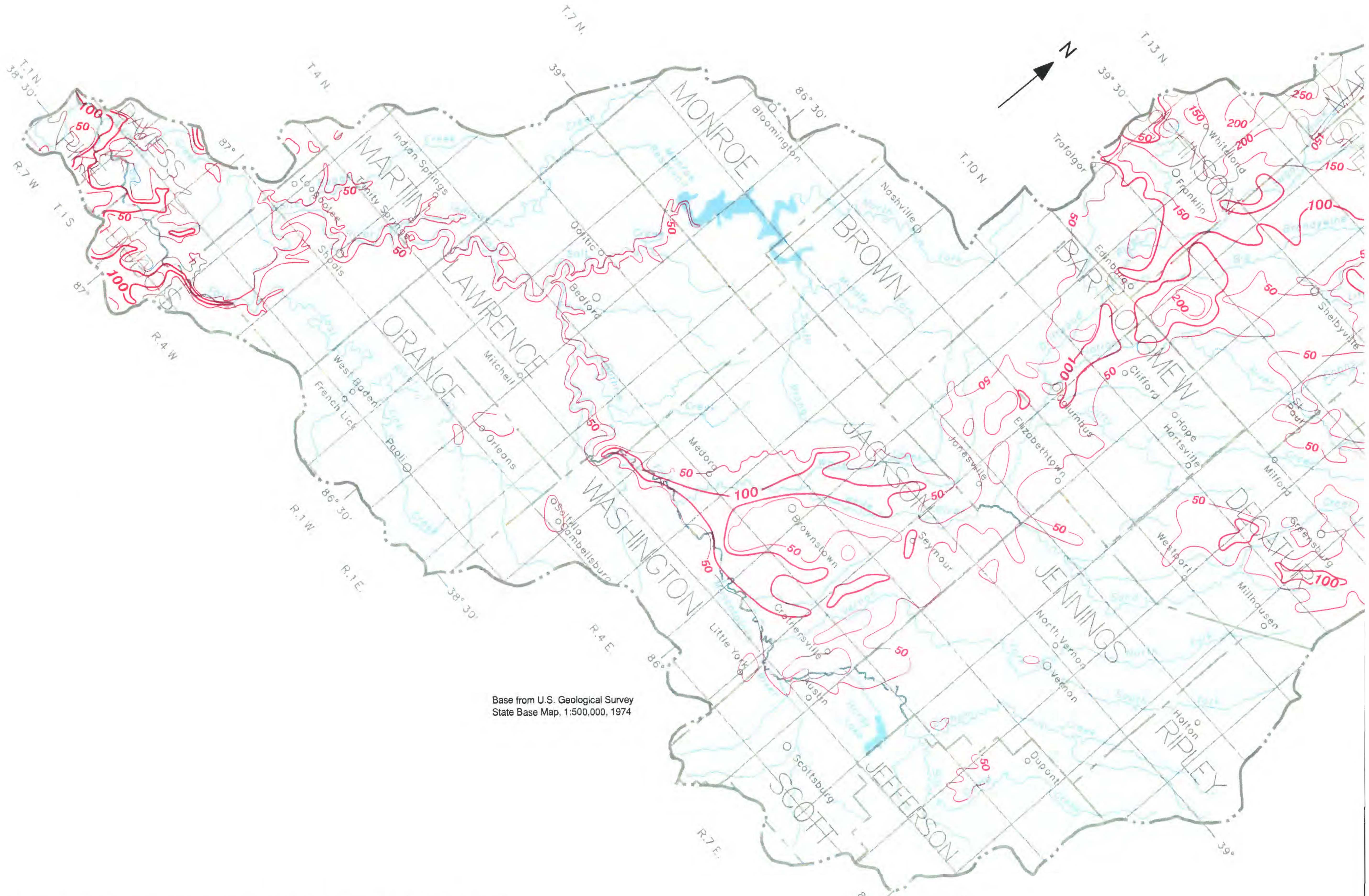
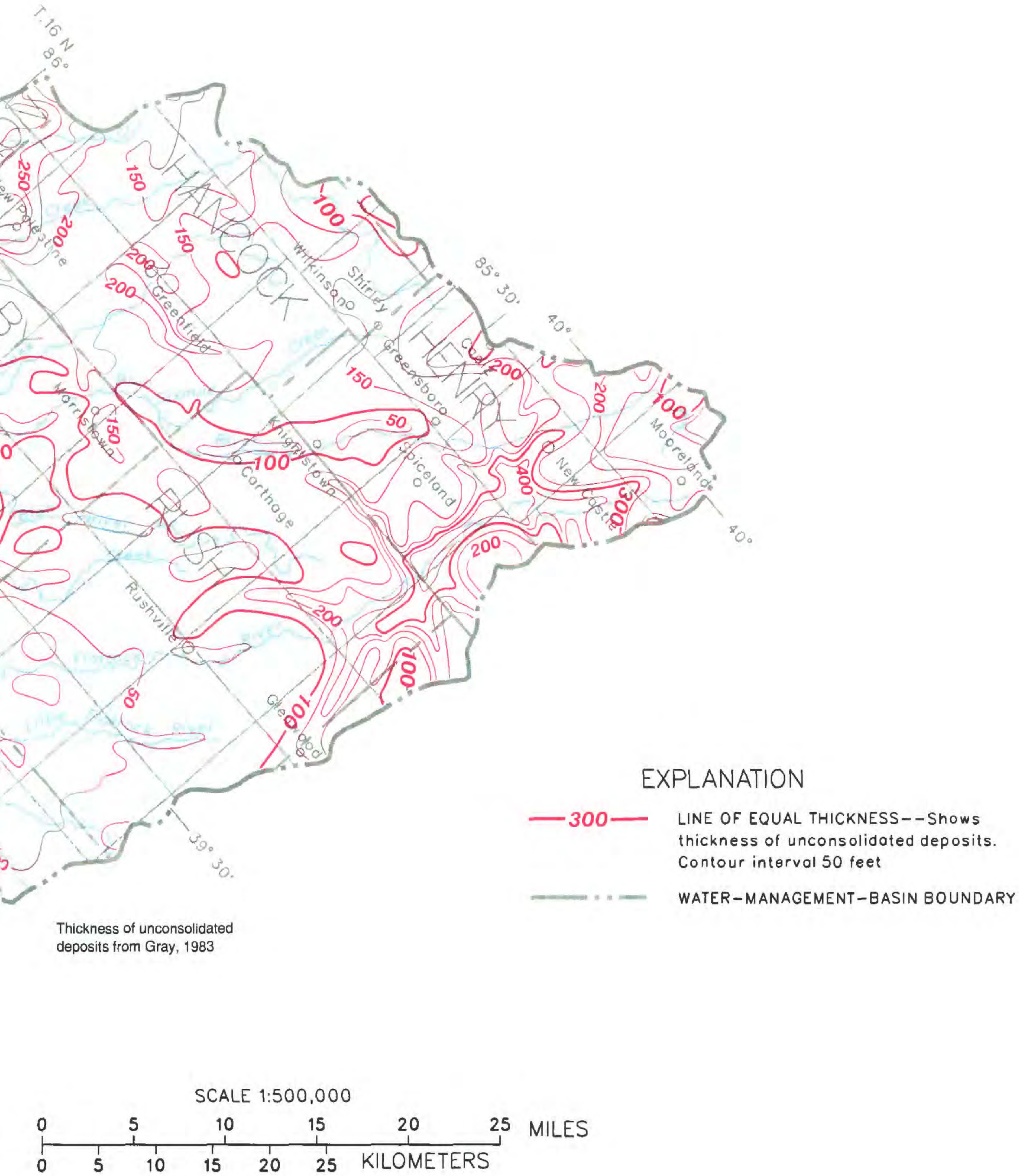


Figure 63. Thickness of unconsolidated deposits in the East Fork White River basin.



A map showing the extent of aquifers in the East Fork White River basin (fig. 65) was constructed. The aquifer map depicts seven different aquifer types in the basin: (1) surficial sand and gravel; (2) buried sand and gravel; (3) discontinuous sand and gravel; (4) carbonate rocks; (5) complexly interbedded sandstone, shale, limestone, and coal; (6) sandstone; and (7) an upper weathered zone in siltstone and shale. The aquifer map was constructed from the widely-spaced hydrogeologic sections and reports listed in the previous studies section.

The principal unconsolidated aquifers in the East Fork White River basin are associated with the glacial drift and outwash deposits along the major rivers. All buried sand and gravel deposits are in the northern one-third of the basin, whereas discontinuous sand and gravel aquifers are concentrated into four areas shown in figure 65. Surficial sand and gravel deposits are adjacent to the main rivers throughout the basin. The principal source of water in the bedrock is the carbonate rocks, which underlie about two-thirds of the basin. Except along a 1,000-mi^2 northwest-trending band of shales and siltstones in the central part of the basin, well yields are generally adequate for domestic and stock needs. Locally, especially in the outwash sand and gravel deposits, yields are sufficient for municipal and industrial needs. Details about each aquifer type are listed in table 11.

Unconsolidated Aquifers

Three types of unconsolidated sand and gravel aquifers are mapped in the East Fork White River basin: surficial outwash deposits located adjacent to some of the major rivers; buried sand and gravel; and discontinuous sand and gravel, including surficial and buried deposits (fig. 65). In general, where a surficial sand and gravel aquifer or buried sand and gravel aquifer is mapped, it is the primary aquifer in that area.

Surficial Sand and Gravel Aquifers

Major surficial sand and gravel aquifers are adjacent to the East Fork White River in Jackson and Bartholomew Counties and upstream from the confluence along Sugar Creek, the Big Blue River, and the Flatrock River (fig. 65). The surficial sand and gravel is primarily Quaternary outwash and smaller amounts of recent stream deposits, wind-blown sand, and Quaternary ice-contact stratified sand and gravel in isolated hills and ridges (Gray, 1989). In the northern one-third of the basin, surficial sand and gravel thickness along Sugar Creek, Brandywine Creek, the Big Blue River and the Flatrock River generally ranges from 10 to 40 ft (Nyman and Pettijohn, 1971, p. 24). Surficial deposits attain a maximum saturated thickness of 120 ft and a width of more than 5 mi just north of Columbus (Watkins and Heisel, 1970, pl. 1). South of Columbus, in the central one-third of the basin, surficial aquifers are 2 to 4 mi wide with 20 to 100 ft of saturated thickness (Planert and Tucci, 1979, p. 8). All surficial deposits in the western one-third of the East Fork White River basin are along the East Fork White River, Salt Creek, and the Lost River. The outwash deposits along the East Fork White River narrow to less than 1 mi and less than 100 ft in thickness. Southwest of Shoals, Ind., the outwash aquifer is usually less than 50 ft thick and contains more fine sand and less gravel (Nyman and Pettijohn, 1971, p. 24). Although, by definition, the mapped surficial sand and gravel aquifer is within 10 ft of the land surface, the aquifer is deeper in places.

Along the river valleys containing surficial sand and gravel, infiltration rates are high, and a large amount of the average annual 40 in/yr of precipitation reaches the water table. In Bartholomew County, recharge rates to the surficial sand and gravel are approximately 15 to 18 in/yr (Jack Whitman, Indiana University, oral commun., 1992). Horizontal hydraulic conductivity of the outwash aquifer in Bartholomew County ranges from 200 to 800 ft/d and averages 470 ft/d (Watkins and Heisel, 1970, p. 6-7). Transmissivity as high as 67,000 ft^2/d is reported by Watkins and Heisel (1970, p. 1). The outwash valleys are the major ground-water discharge areas for the basin. Most of the ground water in the surficial sand and gravel flows into, and contributes to, the base flow of the rivers within the valleys (Watkins, 1964, table 1; Nyman and Watkins, 1965a, table 1; 1965b, table 1). Ground water contributes about 0.75 ft^3/s per mile of river reach to the Big Blue River and Sugar Creek during base flow. Further downstream along the East Fork White River from Columbus to Seymour, base-flow discharge is about 1.0 ft^3/s per mile of reach (Nyman and Pettijohn, 1971, p. 28). According to Nyman and Pettijohn (1971, p. 27), this reach of river from Columbus to Seymour has the greatest potential for ground-water development in the basin. Ground-water yields in the surficial sand and gravel aquifers can be greater than 1,000 gal/min. In general, properly constructed wells within these aquifers are able to produce several hundred gallons per minute or more (Bechert and Heckard, 1966, p. 108-123; Nyman and Pettijohn, 1971, p. 24).

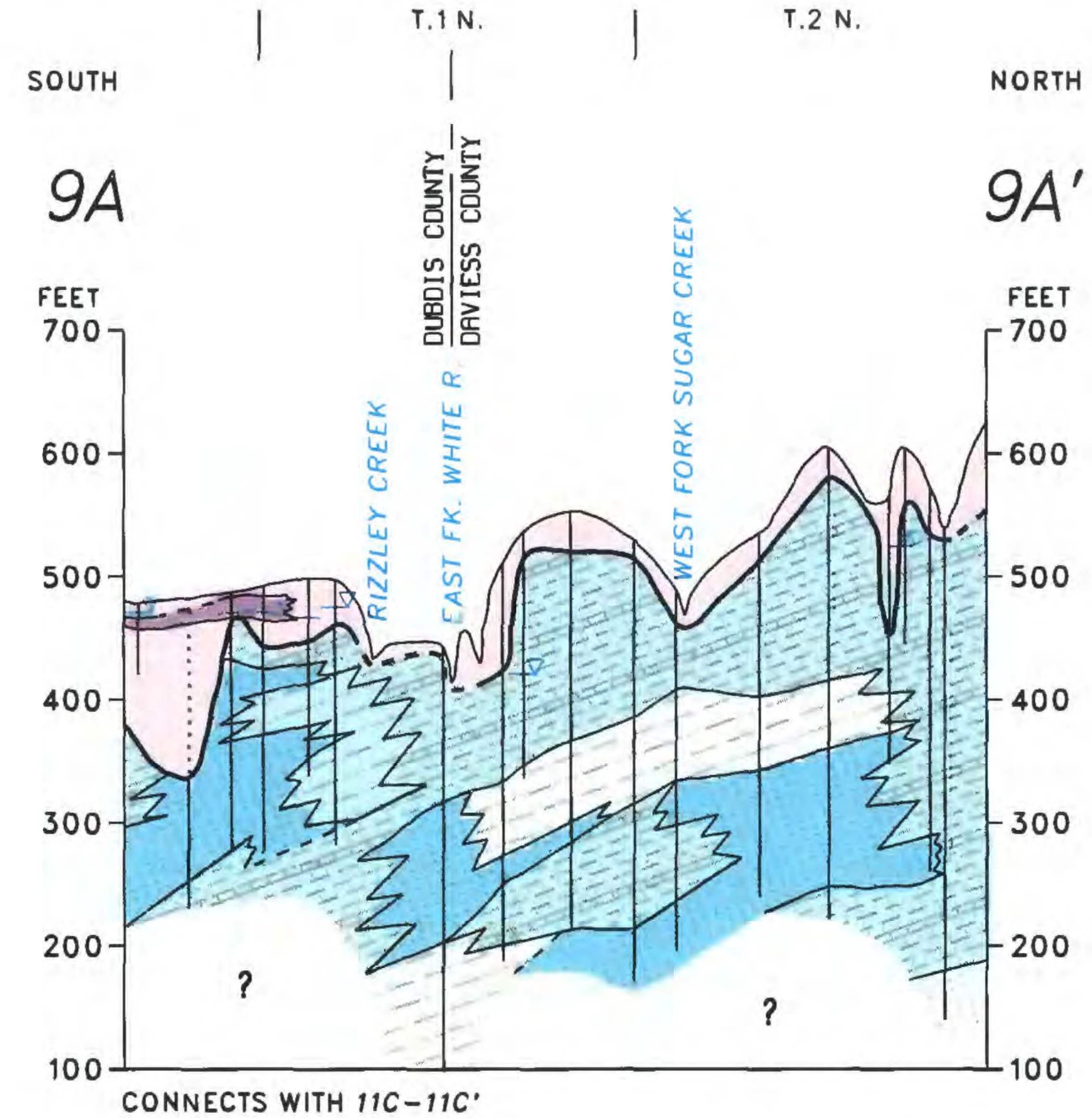
Buried Sand and Gravel Aquifers

Forming general horizons in the drift, buried sand and gravel aquifers are found in laterally continuous deposits covered by more than 10 ft of non-aquifer material. Buried aquifers underlie about one-sixth of the basin, primarily in Henry, Hancock, Shelby, and Johnson Counties (fig. 65). Buried aquifers are shown in the west end of hydrogeologic sections 9H–9H′ to 9J–9J′ and the north end of hydrogeologic sections 9E–9E′ and 9F–9F′ (fig. 64).

In the East Fork White River basin, multiple buried aquifers are commonly found at different horizons in the thick drift. Some of the buried aquifers in the East Fork White River basin correspond to buried aquifers reported in several previous studies of the adjacent White River basin (Lapham, 1981; Arihood, 1982; Arihood and Lapham, 1982). For example, Arihood and Lapham (1982) identified the tops of four buried aquifers in northern Henry County at altitudes of 900 ft, 960 ft, 1,000 ft, and 1,040 ft above sea level. These aquifers can be traced south into the East Fork White River basin in the northern part of section 9F–9F′ (fig. 64).

Just as several of the buried sand and gravel aquifers continue across the basin divides, ground-water flow also crosses the divides. In hydrogeologic section 9E–9E′ (fig. 64), the ground-water divide for the buried aquifers (T. 16 N.) is about 6 mi south of the surface-water divide between the White and East Fork White River basins. Ground water flows north under the surface-water divide toward Fall Creek (located in the White River basin, fig. 54), which is entrenched about 50 ft deeper than the headwaters of Sugar Creek.

In general, recharge to the buried sand and gravel is from ground-water flow through overlying tills and other confining units. In the upstream reaches of many of the rivers, some of the river water and shallow ground water probably flows downward to recharge the buried sand and gravel (Watkins, 1964, table 1; Nyman and Watkins, 1965a, table 1; 1965b, table 1). Much of the ground water probably flows toward major river valleys, where it discharges into the rivers. Ground-water yields to wells in buried sand and gravel aquifers generally range from about 10 to several hundred gallons per minute (Clark, 1980, p. 33).



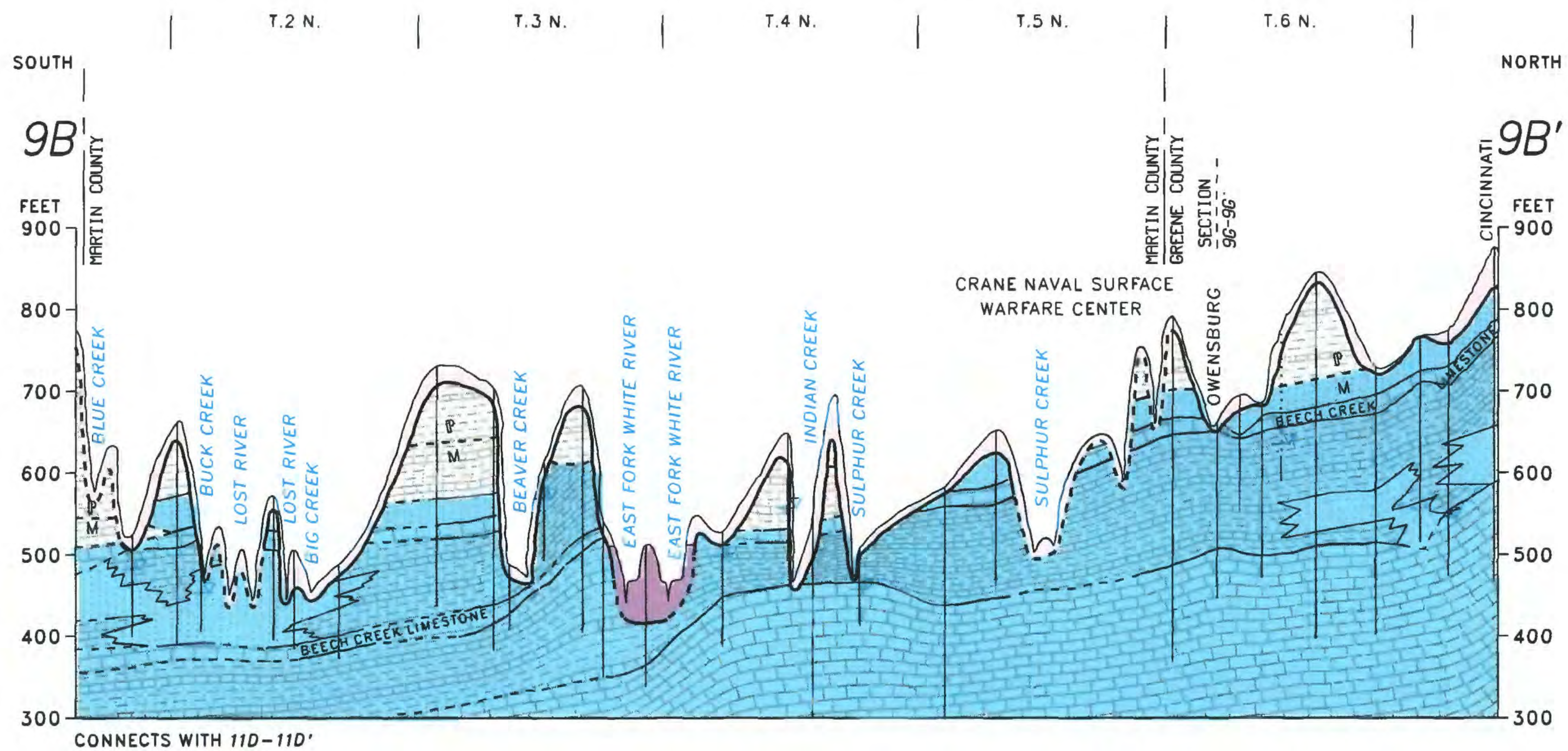
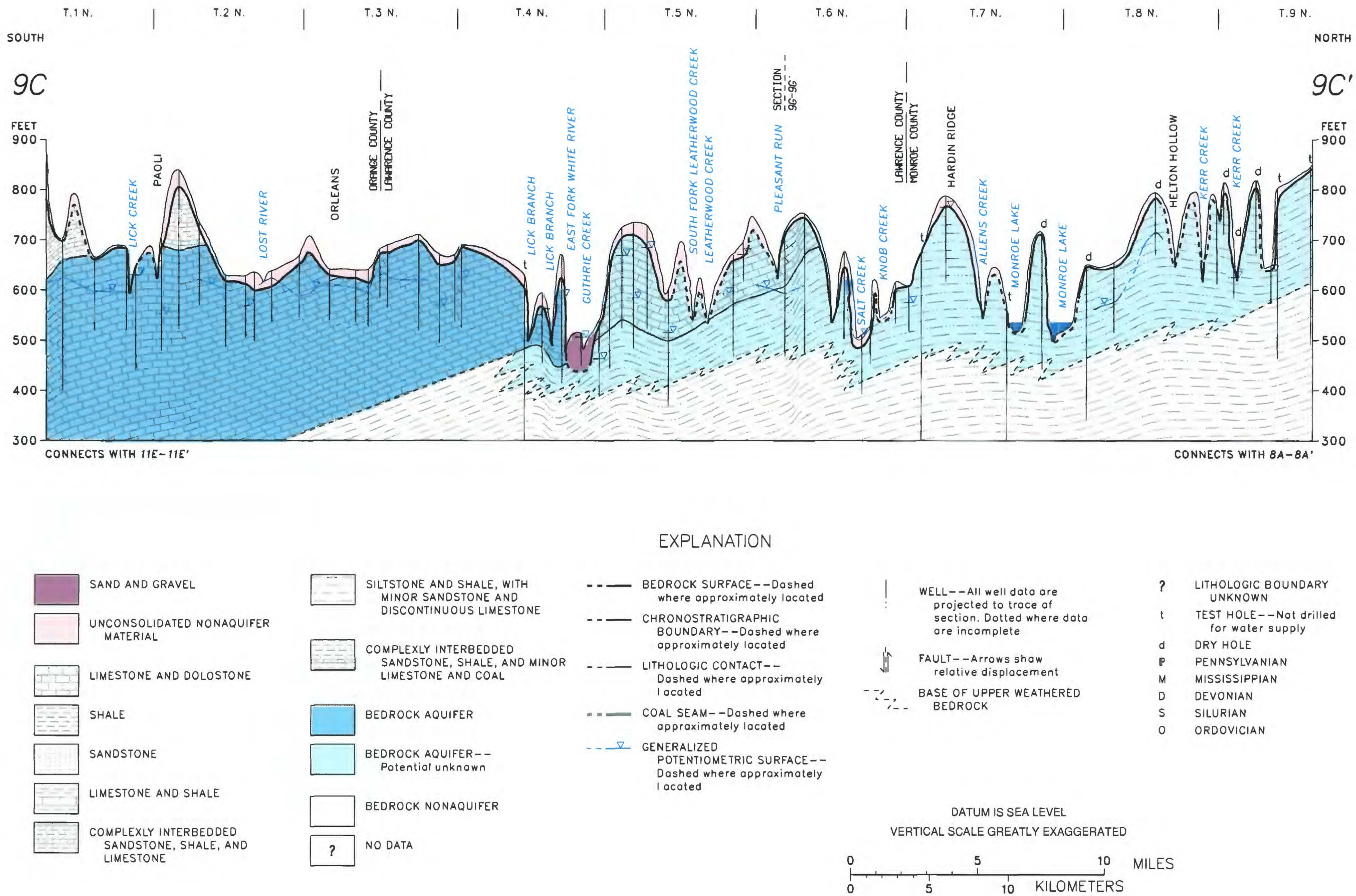


Figure 64. Hydrogeologic sections *9A–9A′* to *9J–9J′* of the East Fork White River basin.



T.1 N.
T.2 N.
T.3 N.
T.4 N.
T.5 N.
T.6 N.
T.7 N.
T.8 N.
T.9 N.
SOUTH
NORTH
9C
9C'
FEET
900
800
700
600
500
400
300
LICK CREEK
PAOLI
LOST RIVER
ORLEANS
ORANGE COUNTY
LAWRENCE COUNTY
LICK BRANCH
LICK BRANCH
EAST FORK WHITE RIVER
GUTHRIE CREEK
SOUTH FORK LEATHERWOOD CREEK
LEATHERWOOD CREEK
PLEASANT RUN
SECTION 9G-9G'
SALT CREEK
KNOB CREEK
LAWRENCE COUNTY
MONROE COUNTY
HARDIN RIDGE
ALLENS CREEK
MONROE LAKE
MONROE LAKE
HELTON HOLLOW
KERR CREEK
KERR CREEK
CONNECTS WITH 11E–11E'
CONNECTS WITH 8A–8A'
EXPLANATION
SAND AND GRAVEL
UNCONSOLIDATED NONAQUIFER MATERIAL
LIMESTONE AND DOLOSTONE
SHALE
SANDSTONE
LIMESTONE AND SHALE
COMPLEXLY INTERBEDDED SANDSTONE, SHALE, AND LIMESTONE
SILTSTONE AND SHALE, WITH MINOR SANDSTONE AND DISCONTINUOUS LIMESTONE
COMPLEXLY INTERBEDDED SANDSTONE, SHALE, AND MINOR LIMESTONE AND COAL
BEDROCK AQUIFER
BEDROCK AQUIFER--Potential unknown
BEDROCK NONAQUIFER
? NO DATA
BEDROCK SURFACE--Dashed where approximately located
CHRONOSTRATIGRAPHIC BOUNDARY--Dashed where approximately located
LITHOLOGIC CONTACT--Dashed where approximately located
COAL SEAM--Dashed where approximately located
GENERALIZED POTENTIOMETRIC SURFACE--Dashed where approximately located
WELL--All well data are projected to trace of section. Dotted where data are incomplete
FAULT--Arrows show relative displacement
BASE OF UPPER WEATHERED BEDROCK
? LITHOLOGIC BOUNDARY UNKNOWN
t TEST HOLE--Not drilled for water supply
d DRY HOLE
IP PENNSYLVANIAN
M MISSISSIPPIAN
D DEVONIAN
S SILURIAN
O ORDOVICIAN
DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED
0 5 10 MILES
0 5 10 KILOMETERS

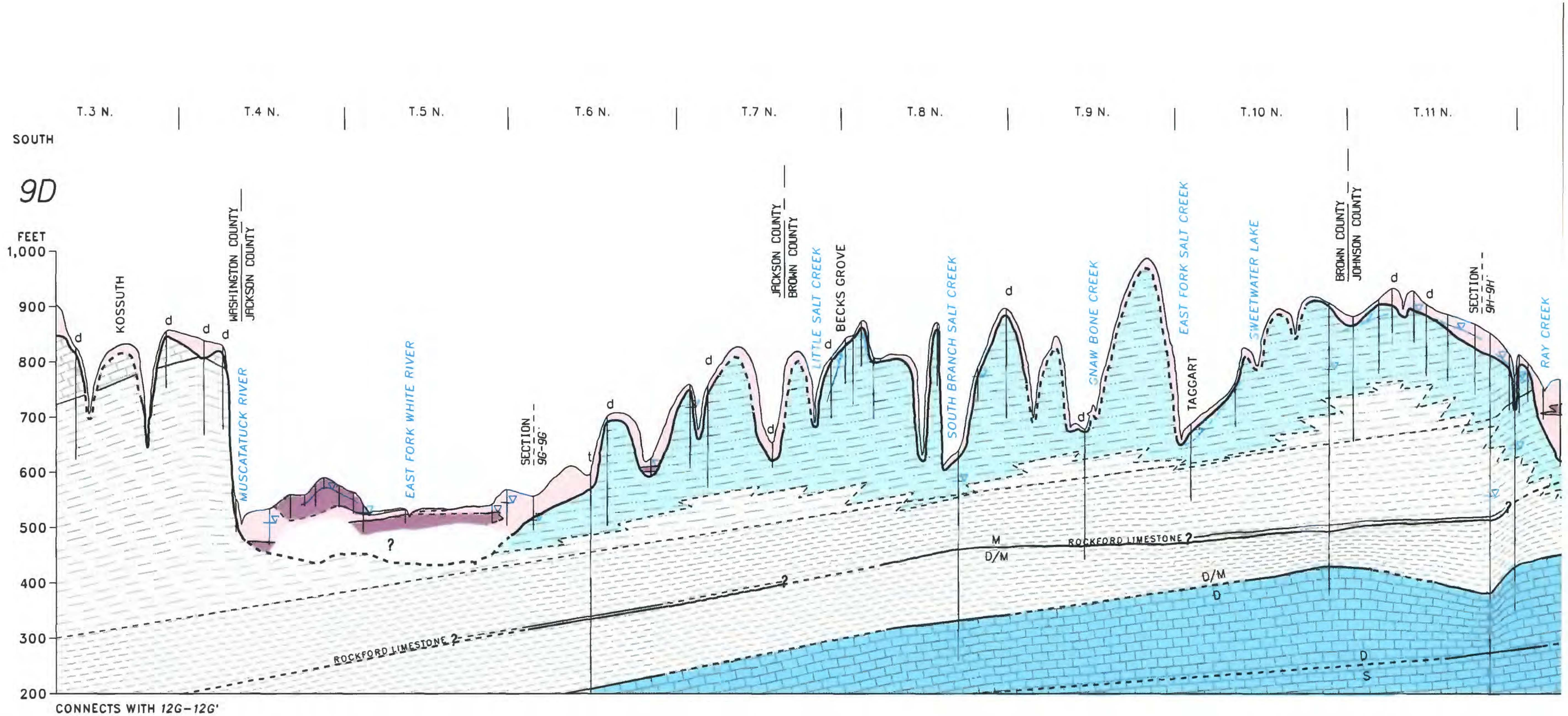


Figure 64. Hydrogeologic sections *9A–9A′* to *9J–9J′* of the East Fork White River basin—Continued.

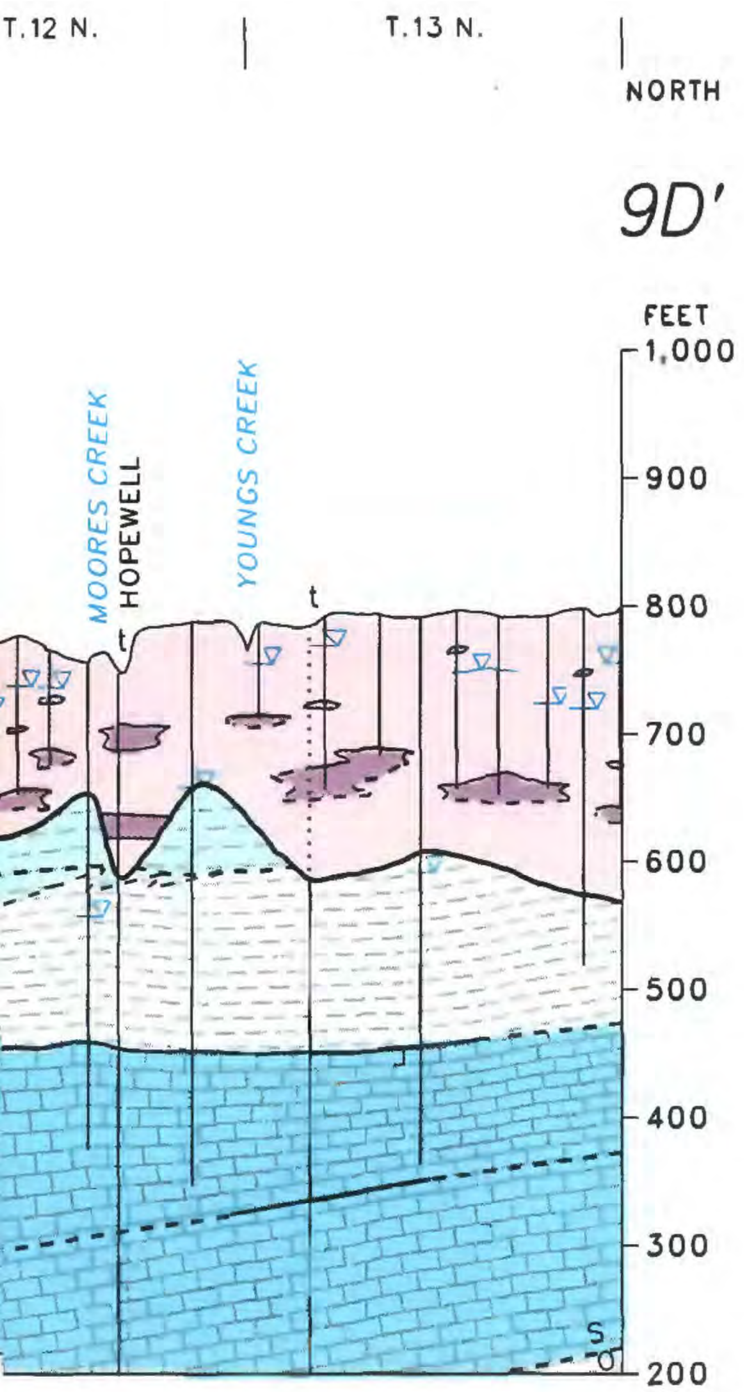
0 5 10 MILES
0 5 10 KILOMETERS

DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED

Several buried bedrock valleys in the northeastern part of the basin contain buried sand and gravel aquifers. Located in northeastern Rush County and southern Henry County (fig. 63), these buried valleys can be seen in hydrogeologic sections 9F–9F′ and 9J–9J′ (fig. 64). Nearly 500 ft of drift overlies the deepest parts of these valleys. Generally, adequate supplies of ground water can be found in the buried sands and gravels within the upper half of the valley fill. The lower half is predominately nonaquifer material.

Discontinuous Sand and Gravel Aquifers

Small, discontinuous lenses of sand and gravel, either buried in general stratigraphic horizons or forming a basal deposit on the bedrock surface form discontinuous sand and gravel aquifers. In the East Fork White River basin, the area underlain by discontinuous sand and gravel aquifers is generally south of the more continuous buried sand and gravel aquifers (fig. 65). The discontinuous sand and gravel aquifers in Rush County and parts of Henry, Marion, and Johnson Counties are typically present in multiple layers. South of southern Shelby County, unconsolidated deposits thin, and a discontinuous basal sand is generally the only unconsolidated aquifer present. Some discontinuous sand deposits can be found along the Muscatatuck River and its tributaries. These sand deposits are generally very fine grained and can pass through well screens. Another area of discontinuous sand and gravel is within a buried valley south of the East Fork White River in Pike and Dubois Counties (southern end of section 9A–9A′, fig. 64). This buried valley is more than 150 ft deep. In preglacial times, the valley drained an area now partially drained by the Patoka River. Although discontinuous sand and gravel aquifers can not supply large volumes of water, they can be an important resource where they are the only source. In the northern part of the basin, domestic yields are generally available, and yields as high as a 100 gal/min are reported.



Bedrock Aquifers

Carbonate Bedrock Aquifers

The most widespread aquifers in the East Fork White River basin are carbonate aquifers, which underlie about three-fourths of the basin. Although limestone and dolomite are not considered highly permeable, solution of the carbonate rock along joints and bedding planes by infiltrating precipitation can significantly increase the permeability of the formations and the availability of ground water. For purposes of discussion, the carbonate bedrock aquifers in the East Fork White River basin are divided into three groups: thin (5 to 30 ft) Mississippian limestone aquifers, a thick (350 to 500 ft) Mississippian carbonate bedrock aquifer, and a Silurian-Devonian carbonate bedrock aquifer.

The stratigraphic relation of the Mississippian carbonate bedrock aquifers to the Silurian-Devonian carbonate bedrock aquifer can be seen in hydrogeologic section 9G–9G′ (fig. 64). Thin Mississippian carbonate aquifers in the far western part of the hydrogeologic section are interbedded within sandstones, shales, and limestones. These are underlain by the thick Mississippian carbonate bedrock aquifer. Underlying this aquifer and overlying the Silurian-Devonian carbonate bedrock aquifer are approximately 800 ft of siltstone and shale. The Silurian-Devonian carbonate aquifer is confined at its lower boundary by nearly impermeable Ordovician shale and limestone.

The thin Mississippian limestones are the least important carbonate bedrock aquifers. They are also the youngest Mississippian limestones. They are exposed at the bedrock surface in a 15-mi-wide band along the western edge of the carbonate bedrock aquifers in Martin County (fig. 65). These thin aquifers are shown within 150 ft of the land surface in hydrogeologic section 9B–9B′, and in R. 3 W. of section 9G–9G′ (fig. 64). They are interbedded with sandstone aquifers and complexly interbedded sandstone, shale, and limestone deposits. The most important thin Mississippian limestone aquifers are within the Stephensport Group and include the Beech Creek Limestone, the Haney Limestone, and the Glen Dean Limestone. The Beech Creek Limestone is labeled on hydrogeologic section 9B–9B′ (fig. 64). Ground water moves along fractures, bedding planes, and solution openings within these limestone beds. Yields are highly variable and range from 0 to 15 gal/min.

The thick Mississippian carbonate bedrock aquifer underlies the complexly interbedded sandstone, shale, limestone, and coal deposits and the thin Mississippian limestone aquifers. The thick carbonate bedrock aquifer is 350 to 500 ft thick and consists of the Blue River Group and Sanders Group. The aquifer, found in the southwestern one-third of the basin, is used primarily in Orange, western Monroe, southern and western Lawrence, and western Washington Counties (fig. 65). The aquifer is shown in the subsurface in hydrogeologic section 9B–9B′, the southern half of section 9C–9C′, and the western end of section 9G–9G′ (fig. 64).

The thick Mississippian carbonate bedrock aquifer is composed primarily of relatively pure limestone, which is soluble in infiltrating precipitation. Carbonate dissolution has enlarged openings, forming underground channels within the aquifer. Typical well yields are 1 to 50 gal/min but can be as large as 100 gal/min. The lowermost 100 ft of the carbonate rocks, just above the Borden Group, produces very little water (0 to 1 gal/min). (See hydrogeologic section 9G–9G′, Rs. 1 W. and 1 E., fig. 64.)

The thick Mississippian carbonate bedrock aquifer is confined above by low permeability interbedded sandstone, shale, and limestone and below by nearly impermeable siltstones. Most recharge probably enters the aquifer from direct infiltration of precipitation. Because of the high permeability of the fractured limestones, ground-water flow can be rapid. Dye-trace measurements of ground-water flow velocity through karst terrain range from 0.03 to 0.21 mi/h, and ground-water gradients range from 13 to 37 ft/mi (Ruhe, 1975, p. 34-35). Ground-water levels in karst terrains may fluctuate rapidly because of high flow rates through the joint system and low storage capacities of the aquifers (Gray and others, 1960, p. 51). Ruhe (1975, p. 63) reported a water-level change of 24.5 ft in 36 hours in a sinkhole within the carbonate bedrock aquifer. Most of the ground water in the thick Mississippian carbonate bedrock aquifer probably flows to the major rivers in the area (East Fork White River, Lost River, and Indian Creek). Some of the ground-water flow discharges to underground rivers and springs.

Mineralized springs at French Lick and West Baden (fig. 60) have been used for health spas for nearly 150 years. Highly mineralized sulfur water or "Pluto Water" emanates through Mississippian sandstones from the thick Mississippian carbonate bedrock aquifer below. Much of the ground water comes from deep within the carbonate aquifer. Sulfate in the spring water at French Lick comes from gypsum beds in the St. Louis Limestone (base of the Blue River Group) that are 350 to 400 ft below land surface (Hill, 1986, p. 6). There are also mineral springs at Trinity Springs and Indian Springs, approximately 1 mi west of hydrogeologic section 9B–9B′, T. 4 N. (fig. 64). At Trinity and Indian Springs, ground water also flows up from deep within the thick carbonate bedrock aquifer (Hill, 1986, p. 7)

The Silurian-Devonian carbonate bedrock aquifer is the most widely used of the carbonate bedrock aquifers. It underlies the eastern half of the basin and is used extensively, especially where unconsolidated deposits are thin. It is the primary aquifer in an area that includes Jefferson, Jennings, Decatur, eastern Bartholomew, southern Shelby, and southern Rush Counties (fig. 65). The Silurian-Devonian aquifer is shown in hydrogeologic sections 9D–9D′ to 9J–9J′ (fig. 64). The permeability of the Silurian-Devonian carbonate rocks results from fracturing and subsequent solution activity along fractures and bedding planes (fig. 9).

The Silurian-Devonian carbonate bedrock aquifer is composed of limestone, dolostone, and some shale, and ranges from 50 to 250 ft in thickness in its principal area of use. (See the eastern carbonate bedrock aquifer area in fig. 65.) The Waldron Shale separates the Silurian-Devonian carbonate bedrock aquifer into an upper and a lower carbonate bedrock aquifer sequence. The upper sequence has a much higher permeability than the lower sequence (Greeman, 1981, p. 12). In particular, one unit in the upper sequence, the Geneva Dolomite Member of the Muscatatuck Group, is commonly tapped for water supply. Hydrogeologic section 9E–9E′ (fig. 64) shows how reliable this formation is for water supply; more than half of the plotted wells are completed near the base of the Devonian Muscatatuck Group rocks. The Geneva Dolomite Member is a vuggy, sugary-textured dolostone commonly logged as sandstone by drillers.

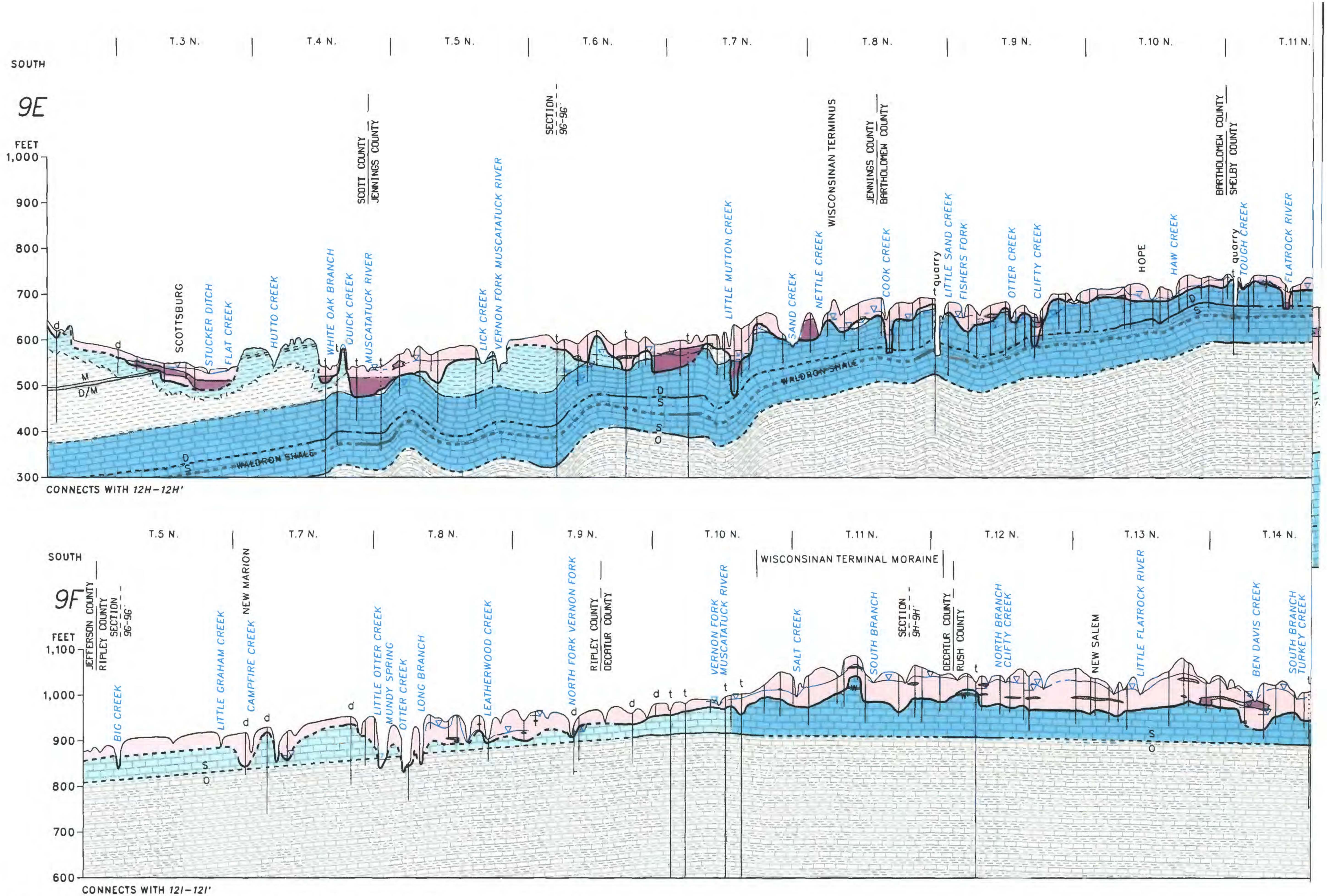
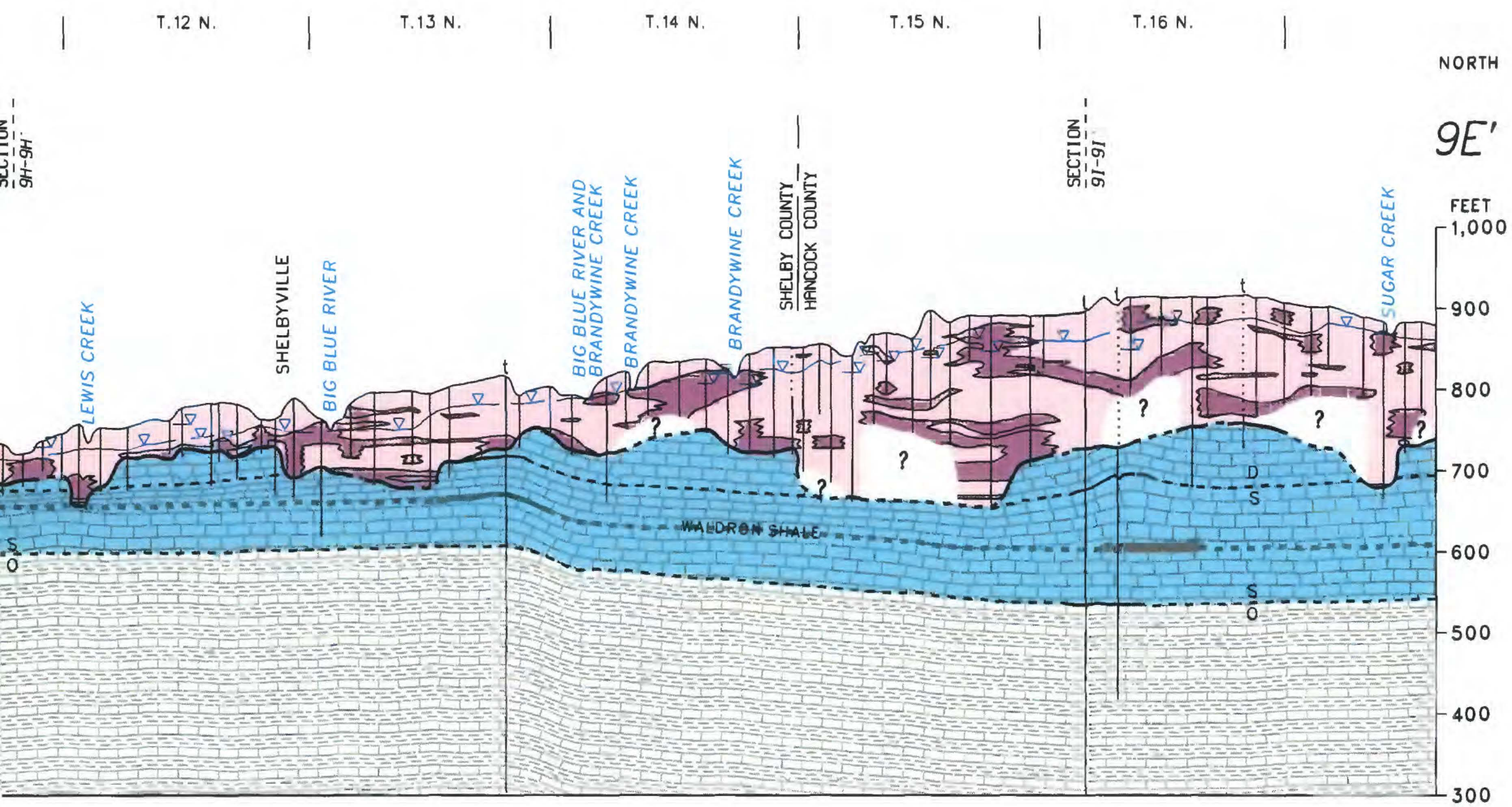
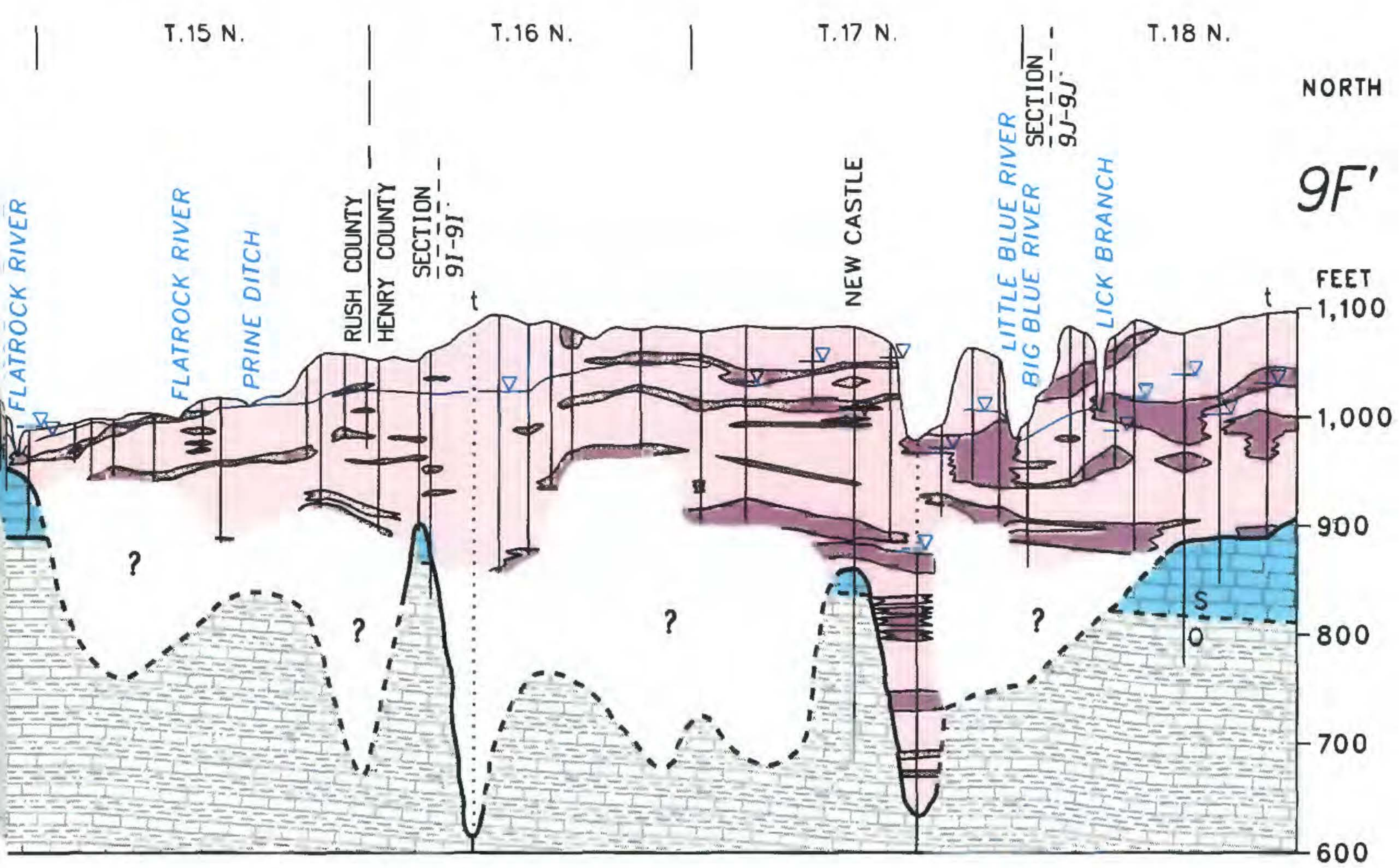


Figure 64. Hydrogeologic sections *9A–9A'* to *9J–9J'* of the East Fork White River basin—Continued.





DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED

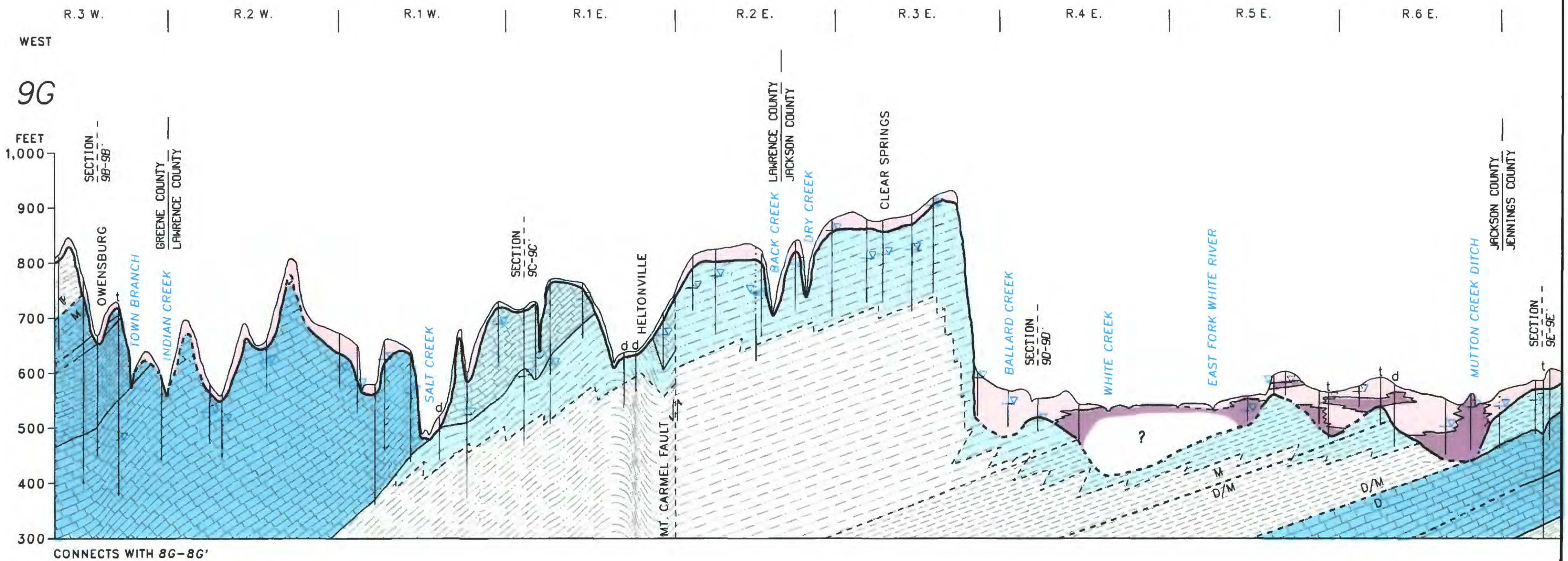
0 5 10 MILES
0 5 10 KILOMETERS

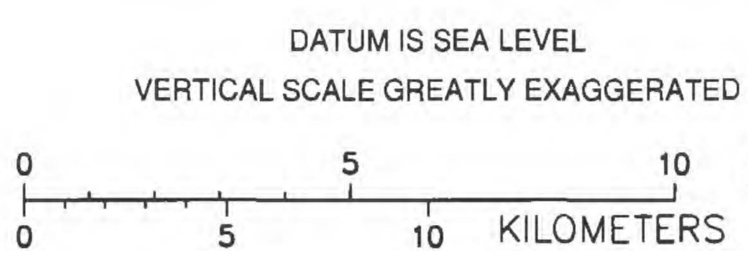
The area in the extreme southeastern part of the basin (Ripley County and eastern Decatur, Jennings, and Jefferson Counties) is mapped as "carbonate bedrock aquifer—potential unknown" because many drilled holes in the area are dry (fig. 65). This area is underlain by the lower carbonate bedrock aquifer sequence, which includes the Salamonie Dolomite and Brassfield Limestone (Greeman, 1981, p. 10). These rocks are unproductive aquifers because of a siliceous cap on the Salamonie Dolomite that is resistant to erosion and solution activity (Greeman, 1981, p. 10-11). Some drillers in the area locate ground-water drilling sites on lineaments and fracture traces that have been mapped from aerial photographs. Mapped lineaments and fracture traces indicate solution-enlarged fractures in the bedrock that transport ground water. These features are mappable through the 30 to 50 ft of till, which is common in this area. The intersection of two lineaments increases the chances that a well will produce sufficient water for domestic use (Greeman, 1981 and 1983).

The western boundary of the Silurian-Devonian carbonate bedrock aquifer (near hydrogeologic section 9D–9D′) was arbitrarily drawn where the top of the aquifer dips to more than 300 ft below the land surface. This boundary does not necessarily reflect the full extent of the aquifer as a water resource, but is a general boundary where the aquifer is not easily accessible because of depth. Several wells tap the aquifer at depths greater than 300 ft where there is no adequate supply of ground water above it. Shown in hydrogeologic section 9D–9D′ (fig. 64) are seven wells that penetrate 300 to 450 ft of rock above the aquifer. Reported pump rates from these seven wells range from 2 to 150 gal/min; all but one well yields greater than 5 gal/min. Within the boundary of the mapped Silurian-Devonian carbonate bedrock aquifer, reported pump rates rarely exceed 100 gal/min and are typically 5 to 25 gal/min.

Recharge to the Silurian-Devonian aquifer is principally from infiltration of precipitation, although some recharge from streams occurs when ground-water levels are lower than stream levels. An example of this downward infiltration can be seen in hydrogeologic section 9D–9D′, T. 12 N. (fig. 64), where water levels in the bedrock are 50 ft to more than 100 ft lower than water levels in the unconsolidated deposits. This area is near the basin divide, where the greatest downward gradients are expected.

Some ground water in the East Fork White River basin discharges to rivers that are not in this basin because the surface-water divide is not the same as the ground-water divide. Along the northwestern side of the basin, ground water flows northwest under the basin divide and discharges toward White River and Fall Creek (fig. 54). These two rivers in the White River basin are entrenched deeper than any of the rivers in the northern part of the East Fork White River basin and, therefore, are able to divert ground water





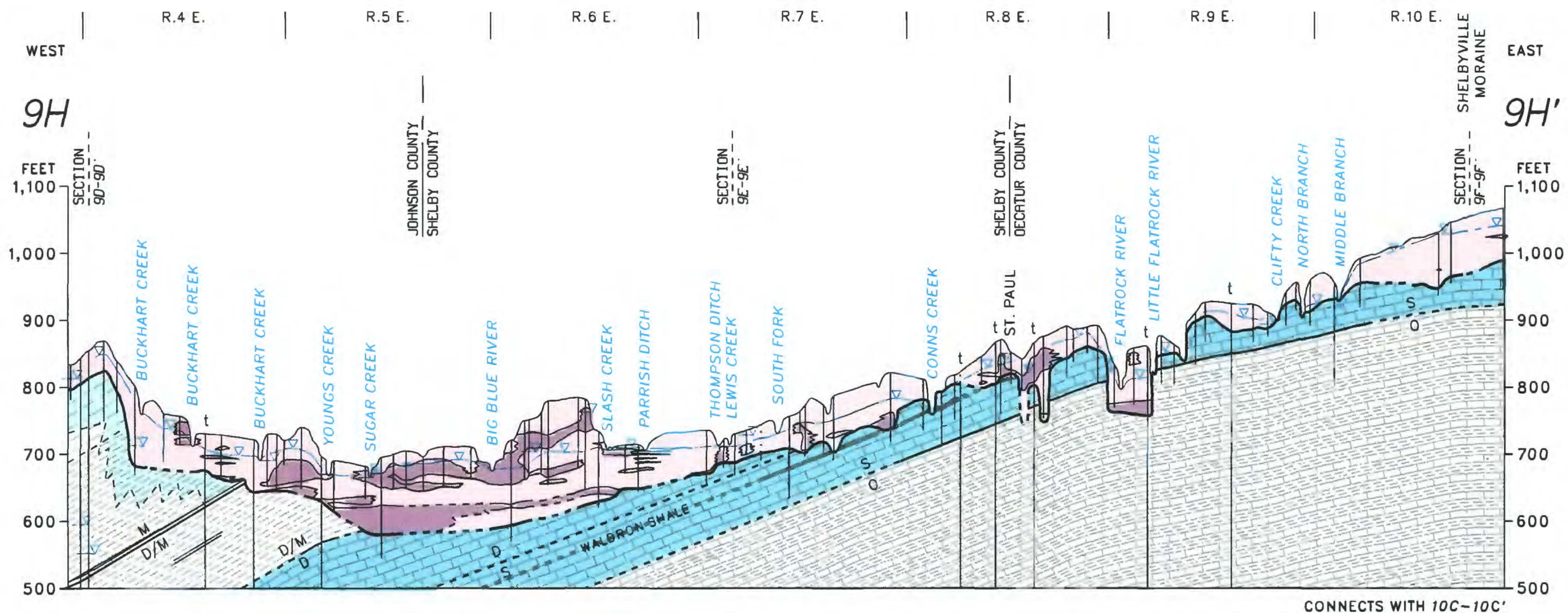
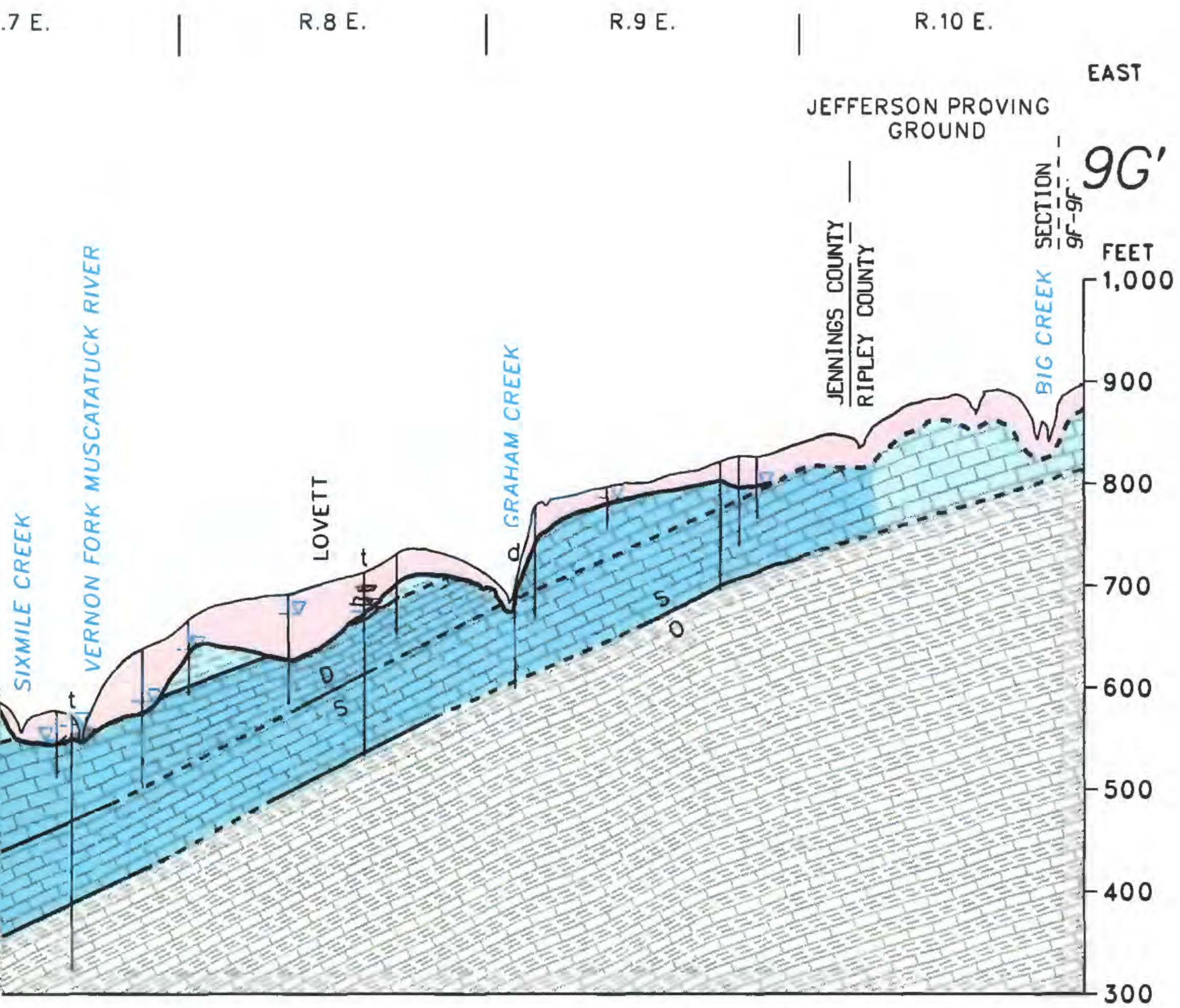


Figure 64. Hydrogeologic sections *9A–9A'* to *9J–9J'* of the East Fork White River basin—Continued.

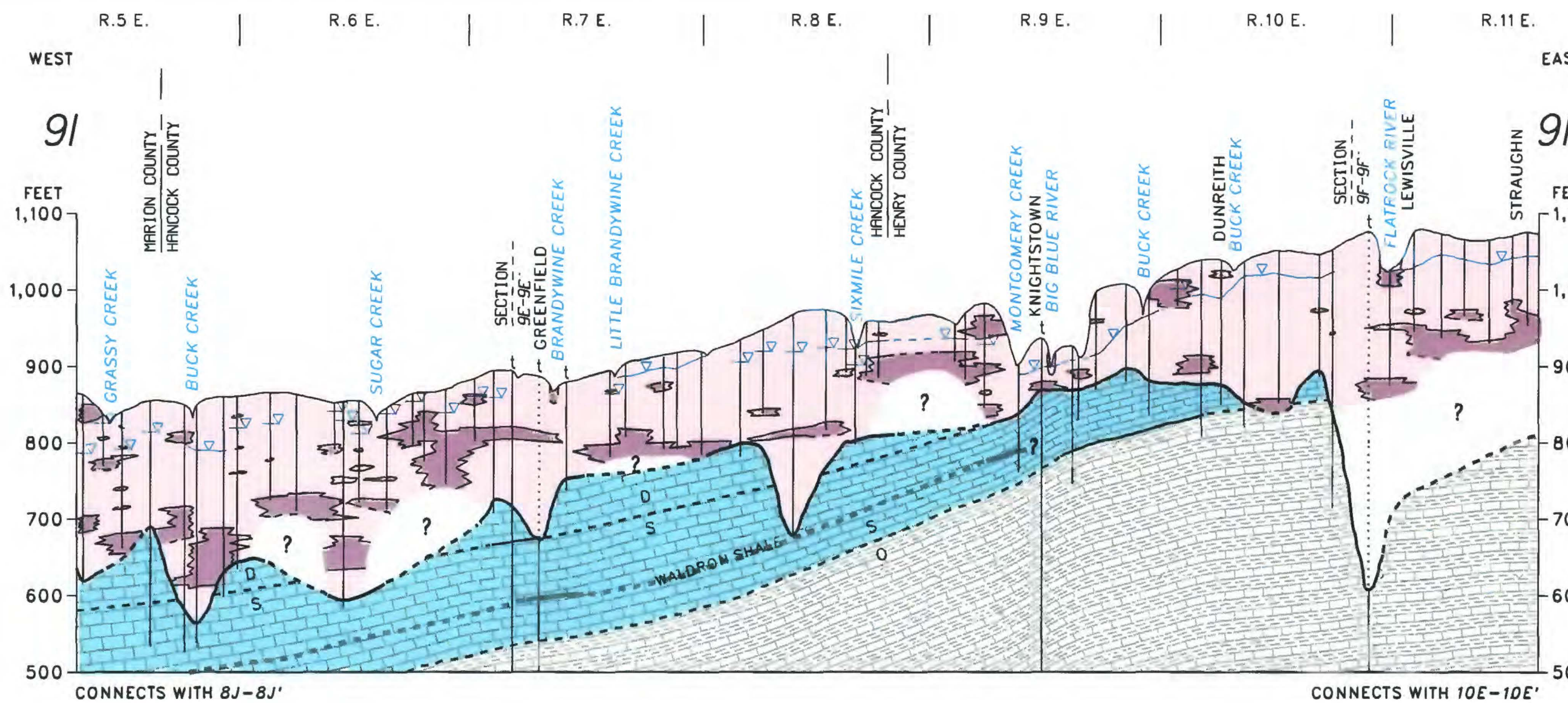


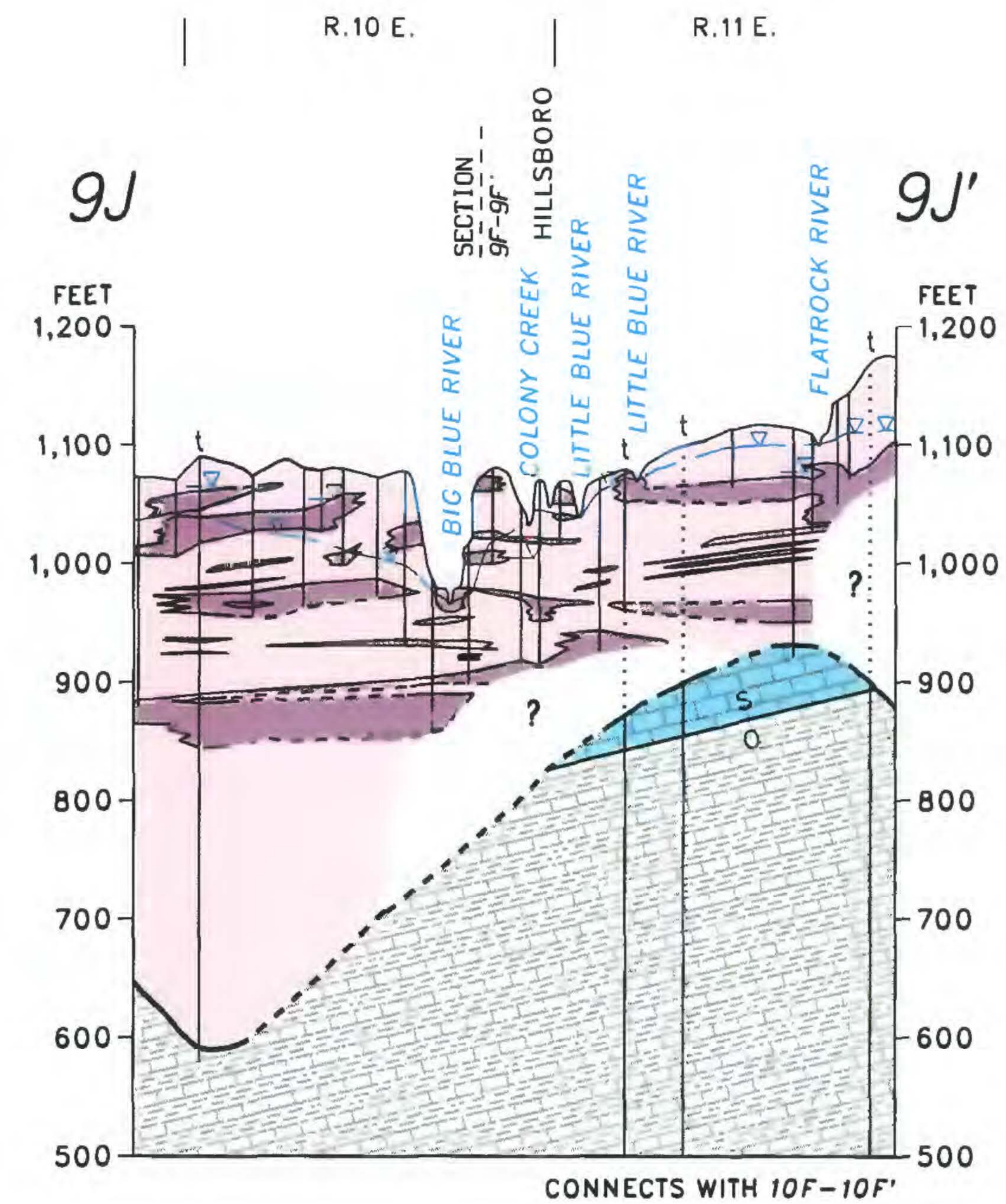
across the basin boundary. An example of ground water crossing the basin boundary can be seen on hydrogeologic section 9I–9I′ (fig. 64), where bedrock water-level elevations decline westward and are below most of the local streams. Ground water in the carbonate bedrock aquifer flows toward the White River, about 11 mi west of the transect shown in this figure. In hydrogeologic section 9E–9E′ (fig. 64), ground water in the carbonate bedrock aquifer flows north out of the basin toward Fall Creek and the White River. Regional flow from this aquifer may discharge into the Maumee River (fig. 1) (Greeman, 1991).

Complexly Interbedded Sandstone, Shale, Limestone, and Coal Aquifers

Aquifers within the complexly interbedded Mississippian and Pennsylvanian rocks are not mapped as individual lithologies because of the complex and discontinuous nature of the deposits. Deposited in marine and nonmarine environments, these complex deposits are composed of sandstone, shale, limestone, and coal. They are found in a 500-mi^2 area at the southwestern end of the basin (fig. 65), and they are shown in hydrogeologic sections 9A–9A′ and 9B–9B′ (fig. 64).

The complex material shown in hydrogeologic section 9A–9A′ (fig. 64) is Pennsylvanian bedrock and consists of the Raccoon Creek Group and the lower part of the Carbondale Group (Lloyd Furer, Indiana Geological Survey, written commun., 1990). The complex Pennsylvanian bedrock is composed primarily of shale and sandy shale that contains numerous coal beds and minor sandstone and limestone beds. The complex bedrock in hydrogeologic section 9A–9A′ is shown as "aquifer—potential unknown." Although most wells are completed in sandstone, the complex material is water bearing only in places. Mississippian rocks are less than 100 ft below the bottom of the hydrogeologic section 9A–9A′ (fig. 64).





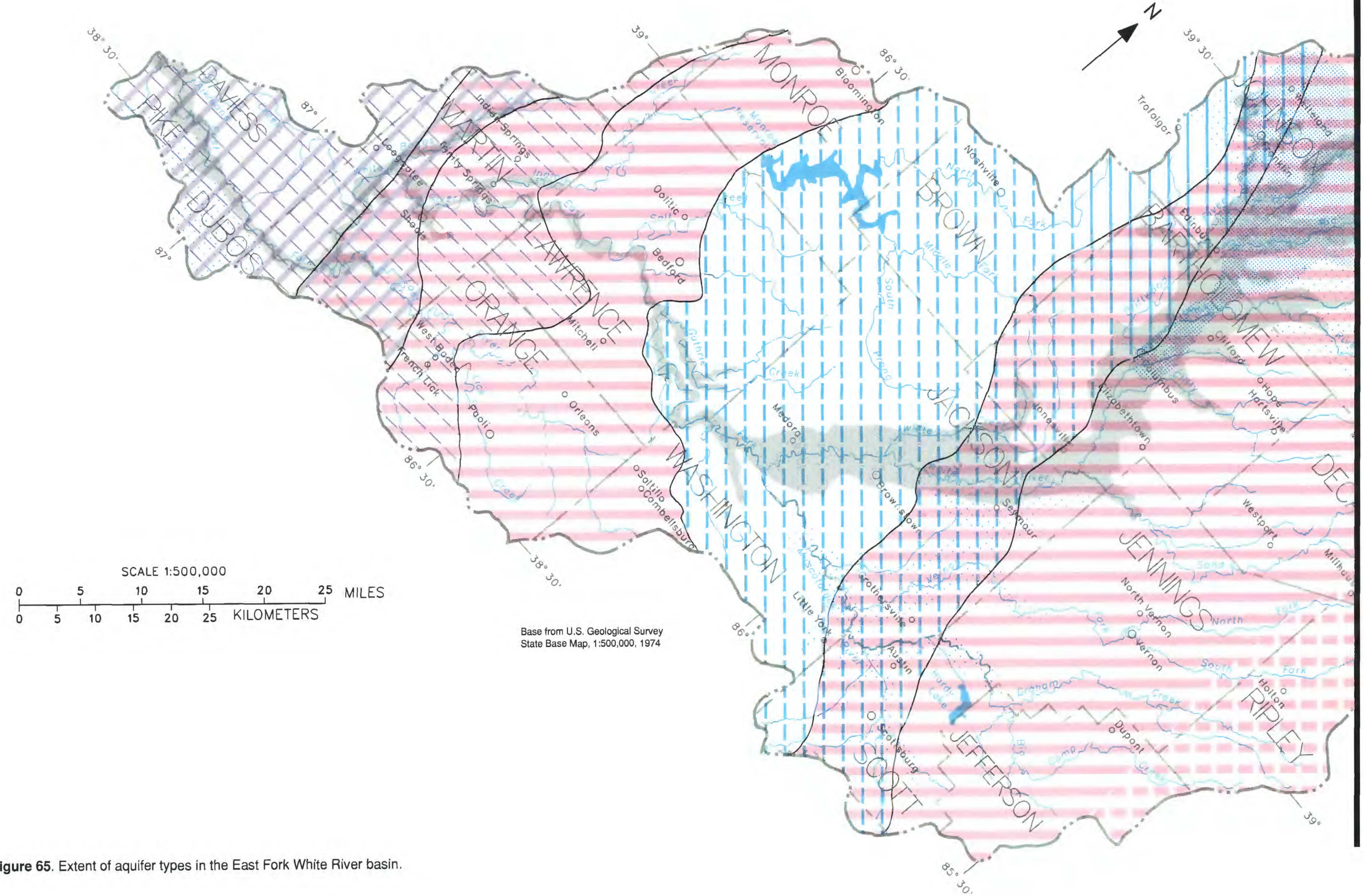


Figure 65. Extent of aquifer types in the East Fork White River basin.

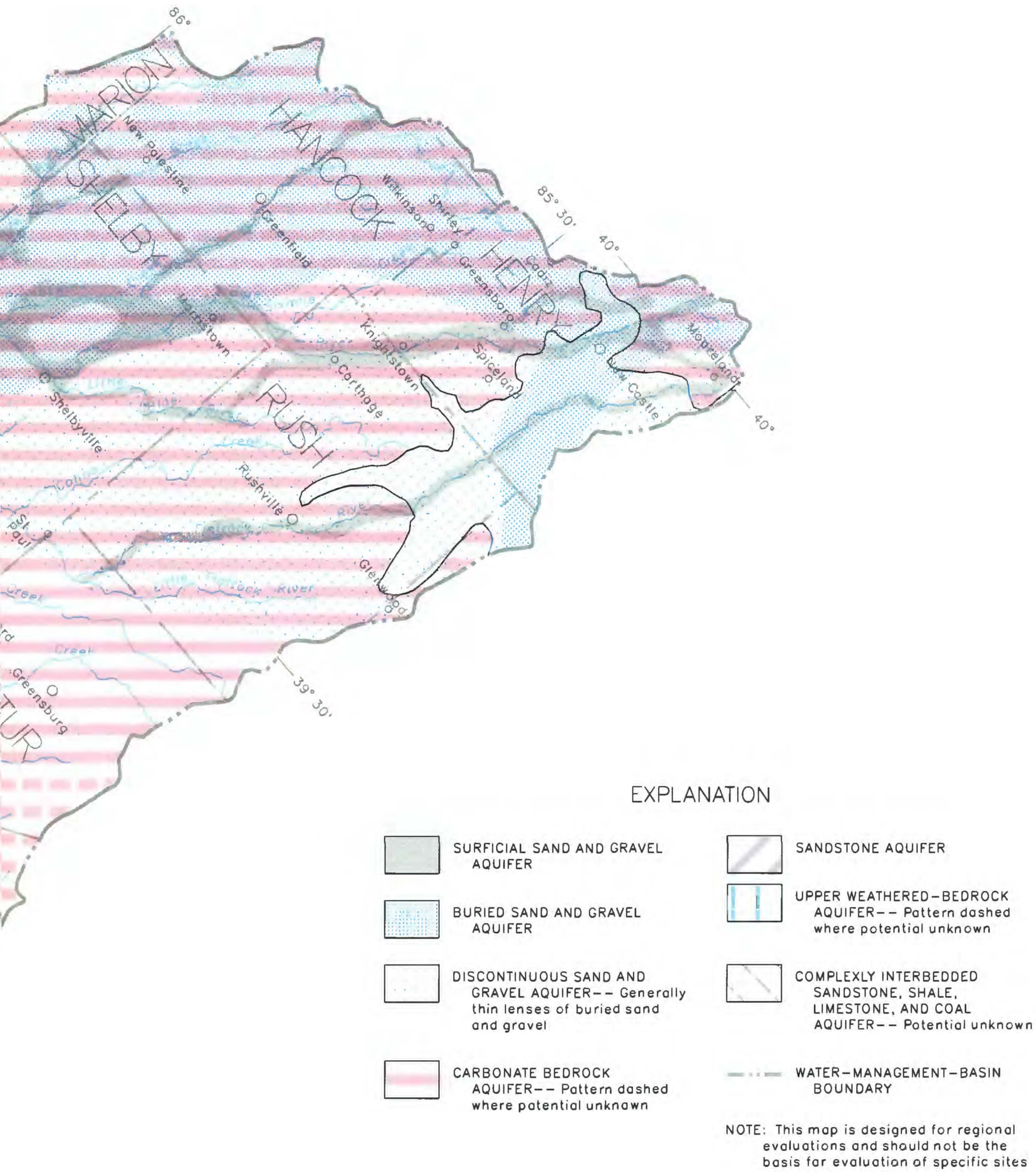


Table 11. Characteristics of aquifer types in the East Fork White River basin

[>, greater than; <, less than; locations of aquifer types shown in fig. 65]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and gravel	10- 100	[1,2,3]10- >1,000	Outwash, alluvium, valley train[3,4]
Buried sand and gravel	10- 50	[1,2]10- 200	
Discontinuous sand and gravel	5- 50	<20	
Carbonate bedrock			
Thin Mississippian limestone	5- 30	<15	Glen Dean, Haney, and Beech Creek Limestones[5]
Thick Mississippian carbonate	350- 550	1- 50	Blue River and Sanders Groups[5]
Upper Devonian and Silurian sequence	100- 225	5- 50	Muscatatuck Group, Wabash Formation, and Louisville Limestone[5]
Lower Silurian sequence	50- 60	0- 25	Salamonie Dolomite and Brassfield Limestone[5]
Complexly interbedded sandstone, shale, limestone, and coal	highly variable[6]	<15	Carbondale, Raccoon Creek, Stephensport, and West Baden Groups[5]
Sandstone	10- 150	1- 50	Mansfield and Big Clifty Formations[5]
Upper weathered bedrock	[7]<150	0- 5	Borden Group and New Albany Shale[5]

[1]Bechert and Heckard, 1966.
[2]Clark, 1980.
[3]Nyman and Pettijohn, 1971.
[4]Watkins and Heisel, 1970.
[5]Shaver and others, 1986.
[6]Water commonly found in thin beds within complexly interbedded unit.
[7]Thickness represents that which is considered permeable, not the thickness of the rock groups or formations.

The complexly interbedded material in hydrogeologic section 9B–9B′ (fig. 64) is primarily Mississippian sandstone, shale, and limestone. The thin Beech Creek Limestone (labeled on hydrogeologic section 9B–9B′, fig. 64) overlies the West Baden Group and is the lowest member of the Stephensport Group. Most of the complexly interbedded Mississippian bedrock is shown as aquifer in hydrogeologic section 9B–9B′ (fig. 64). Numerous wells are completed in the complexly interbedded Mississippian bedrock, and they produce adequate supplies of water for domestic needs.

The entire sequence of complexly interbedded Mississippian and Pennsylvanian bedrock is mapped as "aquifer—potential unknown" on the aquifer map (fig. 65). Even though the complexly interbedded Mississippian and Pennsylvanian bedrock may be the primary aquifer for many households, the location of productive zones cannot be mapped regionally. Because of low yields, the complexly interbedded bedrock aquifer is used only where other source aquifers are unavailable.

Most wells in the southwestern part of the basin are open (uncased) below the unconsolidated cover. Because many of the wells are more than 300 ft deep, it is difficult to determine which rock units supply ground water to the well. Where sandstones, limestones, or coal are mapped, they usually provide most of the water. Many of the wells, however, produce water from several low-productivity units rather than from one primary aquifer. The complexly interbedded bedrock, by itself, can provide domestic supplies of water (as much as 15 gal/min), but yields are variable.

Ground-water flow in the complexly interbedded bedrock is probably through thin limestone, coal, and sandstone beds. Ground water within these aquifers is confined by nearly impermeable shales that are common in the complex material. Deep circulation of ground water is limited in the complexly interbedded bedrock. In many places, ground water is saline at depths of 400 to 500 ft below the land surface. For example, in hydrogeologic section 9A–9A′ (fig. 64), the northernmost well produced saline water at about 400 ft below the land surface. Other well logs indicate that the drillers discontinued drilling to avoid encountering saline water and losing small but usable yields.

Sandstone Aquifers

Sandstones are commonly used as aquifers in the southwestern quarter of the East Fork White River basin (fig. 65). All of the sandstone aquifers are within the complexly interbedded bedrock. Formed as blanket sands, channel-fill deposits, and isolated lenses, these sandstone aquifers are 25 to 150 ft thick. Several sandstone units are used as aquifers. Sandstone aquifers are more common in the Pennsylvanian rocks (hydrogeologic section 9A–9A′, fig. 64) than in the Mississippian rocks (hydrogeologic section 9B–9B′, fig. 64). Most sandstone aquifers are within 300 ft of the land surface.

More than half of the wells in hydrogeologic sections 9A–9A′ and 9B–9B′ (fig. 64) penetrate sandstone aquifers. The sandstones shown in hydrogeologic section 9A–9A′ (fig. 64) are all from the Pennsylvanian Raccoon Creek Group; this group includes the Mansfield, Brazil, and Staunton Formations. Most of the sandstone in hydrogeologic section 9B–9B′ (fig. 64) is from the Mississippian Big Clifty Formation. A member of the Stephensport Group, the Big Clifty Formation contains a sandstone that ranges from 25 to 40 ft in thickness.

Yields from sandstone aquifers generally range from 1 to 50 gal/min. Average yields are higher in the Pennsylvanian sandstones than in the Mississippian sandstones (Wangsness and others, 1981, p. 34). For example, half of the wells in the Pennsylvanian sandstones in hydrogeologic section 9A–9A′ (fig. 64) yield greater than 10 gal/min, whereas most of the wells in the Mississippian sandstones in hydrogeologic section 9B–9B′ (fig. 64) yield less than 10 gal/min. Recharge to the sandstone aquifers is from infiltration of precipitation into the sandstones where they crop out, or by flow of ground water into the sandstones from another permeable unit. Recharge is limited where confining units within the sandstone restrict recharge.

Upper Weathered-Bedrock Aquifer

Although not a desirable source, an upper weathered zone in siltstone and shale is used as an aquifer in the central part of the East Fork White River basin. This aquifer is at, or near, the bedrock surface, where a weathered zone of siltstone and shale bedrock is present. The siltstone-shale bedrock consists of the Borden Group and the New Albany Shale. The upper weathered-bedrock aquifer can be seen on the following hydrogeologic sections: the northern half of section 9C–9C′; most of section 9D–9D′; the extreme southern part of section 9E–9E′; the central one-third of section 9G–9G′; and the far western part of section 9H–9H′ (fig. 64).

The upper weathered-bedrock aquifer is an unproductive source of water in the southern part of the East Fork White River basin but becomes more productive further north. On the aquifer map (fig. 65), only the far northeastern part of the upper weathered bedrock was mapped as aquifer. This area corresponds to thicker drift. Yields from this northeastern area can be as large as 10 gal/min, but dry holes occur. South of this area, the upper weathered bedrock was mapped as "aquifer—potential unknown" (fig. 65), rather than as nonaquifer, because the weathered siltstone-shale is the only source of ground water available in approximately 1,000 mi^2 of the basin. Reported pumpage rates in this area are generally less than 1 gal/min and rarely exceed 5 gal/min. Only the upper 50 to 150 ft of the siltstone and shale bedrock is considered to be potentially water bearing.

Summary

The East Fork White River basin, located in south-central Indiana, has an area of 5,746 mi^2 and includes the cities of Bedford, Bloomington, Columbus, Franklin, Greenfield, Greensburg, Loogootee, New Castle, North Vernon, Rushville, Seymour, and Shelbyville. Seven different types of aquifers were mapped in the basin: (1) surficial sand and gravel; (2) buried sand and gravel; (3) discontinuous sand and gravel; (4) carbonate rocks; (5) complexly interbedded sandstone, shale, limestone, and coal; (6) sandstone; and (7) an upper weathered zone in siltstone and shale.

The principal unconsolidated aquifers in the basin are the surficial and buried sand and gravel deposits located primarily within glacial drift in the northern one-third of the basin and in outwash deposits along some of the major rivers. Where these aquifers are present, they are generally the primary aquifers for their respective areas. Yields of wells that tap these aquifers are adequate for most uses and can exceed 1,000 gal/min. Discontinuous sand and gravel lenses are found primarily in the northern part of the basin where Wisconsinan glacial deposits are thin (less than 100 ft) and are locally an important source of water.

The principal bedrock aquifers in the basin are carbonate bedrock aquifers, which underlie about two-thirds of the basin. Yields of wells that tap these aquifers typically range from 1 to 25 gal/min but can exceed 100 gal/min. The carbonate rocks along the southeastern edge of the basin yield only small amounts of water, and dry holes are common. The southwestern part of the basin contains sandstone aquifers and complexly interbedded sandstone, shale, limestone, and coal aquifers. These aquifers are important, because they are the only aquifers in the area. The smallest yields in this basin are from wells that tap an upper weathered zone in siltstone and shale, which underlies about 1,000 mi^2 in the central part of the basin. Well yields in this area are generally less than 5 gal/min, and dry holes are common.

References Cited

Arihood, L.D., 1982, Ground-water resources of the White River basin, Hamilton and Tipton Counties, Indiana: U.S. Geological Survey Water-Resources Investigations 82-48, 69 p.

Arihood, L.D. and Lapham, W.W., 1982, Ground-water resources of the White River basin, Delaware County, Indiana: U.S. Geological Survey Water-Resources Investigations 82-47, 69 p.

Bassett, J.L., and Hasenmueller, N.R., 1979, Map showing structure on base of New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources, Geological Survey, EGSP Series 800, scale 1:500,000.

_____ 1980, Map of Indiana showing structure on top of the Maquoketa Group (Ordovician): Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 812, scale 1:500,000.

Bassett, P.A., and Keith, B.D., 1984, Data base for deep wells in Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 46, 35 p.

Bechert, C.H. and Heckard, J.M., 1966, Ground water, *in* Lindsay, A.A., ed., Natural features of Indiana: Indianapolis, Ind., Indiana Academy of Science, p. 100-115.

Bruns, T.M., and Uhl, J.E., 1976, Water resources of Shelby County, Indiana, with emphasis on ground-water availability: Indiana Division of Water, 1 sheet.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Ind., Governor's Water Resource Study Commission, Indiana Department of Natural Resources, 508 p.

Davis, L.G., Turner, A.K., and Melhorn, W.N., 1969, Analysis of ground-water use, replenishment, and aquifer characteristics in Bartholomew County, Indiana: Lafayette, Ind., Purdue University Water Resources Research Center Technical Report 3, 83 p.

Duwelius, R.F., 1993, Hydrologic data and hydrologic budget for Summit Lake Reservoir, Henry County, east-central Indiana, water years 1989 and 1990: U.S. Geological Survey, Water-Resources Investigations Report 92-4025, 47 p.

Geosciences Research Associates Inc., 1982, Hydrogeologic atlas of Indiana: Bloomington, Ind., 31 pl.

Gray, H.H., 1982, Map of Indiana showing topography of the bedrock surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 36, scale 1:500,000.

_____ 1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 38, scale 1:500,000.

_____ 1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Gray, H.H.; Droste, J.B.; Patton, J.B.; Rexroad, C.B., and Shaver, R.H., 1985, Correlation chart showing Paleozoic stratigraphic units of Indiana: Indiana Department of Natural Resources, Geological Survey, Supplement to Miscellaneous Map 48, 1 sheet.

Gray, H.H., Jenkins, R.D., and Weidman, R.M., 1960, Geology of the Huron area, south-central Indiana: Indiana Department of Conservation, Geological Survey Bulletin 20, 78 p.

Greeman, T.K., 1981, Lineaments and fracture traces, Jennings County and Jefferson Proving Ground, Indiana: U.S. Geological Survey Open-File Report 81-1120, 17 p.

_____ 1983, Lineaments and fracture traces, Decatur County, Indiana: U.S. Geological Survey Open-File Report 82-918, 17 p.

_____ 1991, Areal differences between ground-water and surface-water drainage basins in Indiana [abs.]: Midwest Ground Water Conference, 36th, Indianapolis, 1991, program with abstracts, p. 106-107.

Hasenmueller, N.R. and Bassett, J.L., 1979, Map showing structure on top of the New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources, Geological Survey, EGSP Series 801, scale 1:500,000.

_____ 1980, Map of Indiana showing structure on top of Silurian rocks: Indiana Department of Natural Resources, Geological Survey, METC/EGSP Series 811, scale 1:500,000.

Herring, W.C., 1974, Water resources of Marion County, Indiana, with emphasis on ground-water availability: Indiana Division of Water, 1 sheet.

_____ 1976, Technical atlas of the ground-water resources of Marion County, Indiana: Indiana Department of Natural Resources, Division of Water, 53 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hill, J.R., 1986, Springing to life—mineral wells attract Hoosiers for medicinal purposes: Outdoor Indiana, v. 51, no. 2, p. 4-9.

Jacques, D.V., and Crawford, C.G., 1991, National water-quality assessment program—White River basin: U.S. Geological Survey Open-File Report 91-169, 2 p.

Keller, S.J., 1990, Maps of southwestern Indiana showing geology and elevation of the sub-Pennsylvanian surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 51, scale 1:380,160.

Klaer, F.H., Jr., Davis, G.E., and Kingsbury, T.M., 1951, Ground-water resources of the Columbus area, Bartholomew County, Indiana: U.S. Geological Survey Open-File Report, 37 p.

Klaer, F.H., Jr., and Kingsbury, T.M., 1948, Provisional report on the ground-water resources of the Columbus area, Indiana: U.S. Geological Survey Open-File Report, 29 p.

Lapham, W.W., 1981, Ground-water resources of the White River basin, Madison County, Indiana: U.S. Geological Survey Water-Resources Investigations 81-35, 112 p.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 20, part 2, p. 59-256.

Meyer, William, Reussow, J.P., and Gillies, D.C., 1975, Availability of ground water in Marion County, Indiana: U.S. Geological Survey Open-File Report 75-312, 87 p.

Nyman, D.J., and Watkins, F.A., Jr., 1965a, Ground-water appraisal of the Big Blue River basin and Big Blue Reservoir site, Indiana: U.S. Geological Survey, Open-File Report, 12 p.

_____ 1965b, Ground-water appraisal of the Flatrock River basin and Downeyville Reservoir site, Indiana: U.S. Geological Survey, Open-File Report, 6 p.

Nyman, D.J., and Pettijohn, R.A., 1971, Ground water, *in* Wabash River Coordinating Committee, Wabash River basin comprehensive study: U.S. Geological Survey and U.S. Army Corps of Engineers, Louisville District, v. VIII, app. G, 100 p.

Pinsak, A.P., 1957, Subsurface stratigraphy of the Salem Limestone and associated formations in Indiana: Indiana Department of Conservation, Geological Survey Bulletin 11, 62 p.

Planert, Michael, 1976, Digital model analysis to predict water level in a well field near Columbus, Indiana: U.S. Geological Survey Water-Resources Investigations 76-63, 15 p.

Planert, Michael, and Tucci, Patrick, 1979, Effects of pumping on ground-water levels near Taylorsville, Bartholomew County, Indiana: U.S. Geological Survey Water-Resources Investigations 79-20, 22 p.

Roberts, C.M., Widman, L.E., and Brown, P.N., 1955, Water resources of the Indianapolis area, Indiana: U.S. Geological Survey Circular 366, 45 p.

Ruddy, B.C., and Hitt, K.J., 1990, Summary of selected characteristics of large reservoirs in the United States and Puerto Rico, 1988: U.S. Geological Survey Open-File Report 90-163, 295 p.

Ruhe, R.V., 1975, Geohydrology of karst terrain, Lost River watershed, southern Indiana: Bloomington, Ind., Indiana University, Water Resources Research Center Report of Investigation 9, 139 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Ind., Indiana Academy of Sciences, p. 40-56.

Schneider, A.F., and Gray, H.H., 1966, Geology of the upper East Fork drainage basin, Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 3, 55 p.

Shaver, R.H., and Austin, G.S., 1972, A field guide to the Mt. Carmel Fault of southern Indiana: Indiana Department of Natural Resources Geological Survey Guidebook 13, 25 p.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.D. Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B., and Weir, C.E., 1986; Compendium of Paleozoic rock-unit stratigraphy in Indiana—A revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Sullivan, D.M., 1972, Subsurface stratigraphy of the West Baden Group in Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 47, 31 p.

Uhl, J.E., 1966, Water resources of Johnson County, Indiana, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

_____ 1973, Water resources of Henry County, Indiana, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

_____ 1975, Water resources of Hancock County, Indiana, with emphasis on ground-water availability: Indiana Department of Natural Resources, Division of Water, 1 sheet.

Wangsness, D.J., Miller, R.L., Bailey, Z.C., and Crawford, C.G., 1981, Hydrology of area 32, eastern region, Interior Coal Province, Indiana: U.S. Geological Survey Water-Resources Investigations Open-File Report 82-566, 43 p.

Watkins, F.A., Jr., 1964, Ground-water appraisal of Clifty Creek basin and Clifty Creek Reservoir site, Indiana: U.S. Geological Survey, Open-File Report, 10 p.

Watkins, F.A., Jr., and Heisel, J.E., 1970, Electrical-analog-model study of water resources of the Columbus area, Bartholomew County Indiana: U.S. Geological Survey Water-Supply Paper 1981, 22 p.

Wayne, W.J., 1956, Thickness of drift and bedrock physiography of Indiana north of the Wisconsin glacial boundary: Indiana Geological Survey Report of Progress 7, 70 p.

WHITEWATER RIVER BASIN

By M. Catharine Woodfield

General Description

The Whitewater River water-management basin is located in southeastern Indiana. The basin extends approximately 75 mi along the Indiana-Ohio State line. Its maximum width is approximately 30 mi, south of the Brookville Reservoir (fig. 66). The basin encompasses an area of 1,425 mi^2 and includes all of Wayne and Union Counties, most of Fayette and Franklin Counties, and parts of Randolph, Henry, Decatur, and Dearborn Counties. The largest cities in the basin are Richmond and Connersville.

Previous Studies

The most comprehensive study of the Whitewater River basin's ground-water resources was done by the Indiana Department of Natural Resources (1988). Information on ground-water quality and availability, a potentiometric-surface map, and an aquifer-system map were included in this study. Wolf (1969) presented information on the hydrogeology of the unconsolidated water-bearing units and bedrock aquifers, including data on the thickness and permeability of the aquifers, well yields, ground-water discharge to streams, and water quality. Reports on ground-water availability, quality, bedrock topography, and piezometric surface for Union and Fayette Counties were prepared by Clark (1992) and Reynolds (1993), respectively.

Several authors have reported on the ground-water resources for the entire State of Indiana. Harrell (1935) wrote a comprehensive report on the ground-water resources for each county in Indiana. Bechert and Heckard (1966) delineated ground-water provinces on the basis of well yields and sources of ground water. Clark (1980) examined the different types of aquifers and their potential yields.

A detailed description of the unconsolidated materials for the entire basin was presented by Gooding (1963, 1966, 1973, and 1975). Gruver (1984) described the outwash deposits along the Whitewater River.

Physiography

Two physiographic units are included in the Whitewater River basin. The Tipton Till Plain, in the northern one-third of the basin, is characterized by gently rolling topography. In contrast, the southern two-thirds of the basin is characterized by bedrock-controlled rugged relief of the Dearborn Upland (fig. 67) (Schneider, 1966, p. 42).

The glaciated northern part of the basin (Tipton Till Plain) is underlain by a moderate thickness of glacial till (fig. 68), ranging from greater than 400 ft in a segment of the Lafayette Buried Bedrock Valley (fig. 7) to less than 50 ft near the southern boundary of the Tipton Till Plain. Headwater tributaries of the Whitewater River—namely, the West Fork, Greens Fork, Nolands Fork, and East Fork (fig. 66)—are incised into the Tipton Till Plain and have local relief of greater than 100 ft. Other geomorphic features of the Tipton Till Plain include moraines, kames, ice-channel fills (eskers), outwash plains, and valley trains. The predominant features are the moraines (fig. 67).

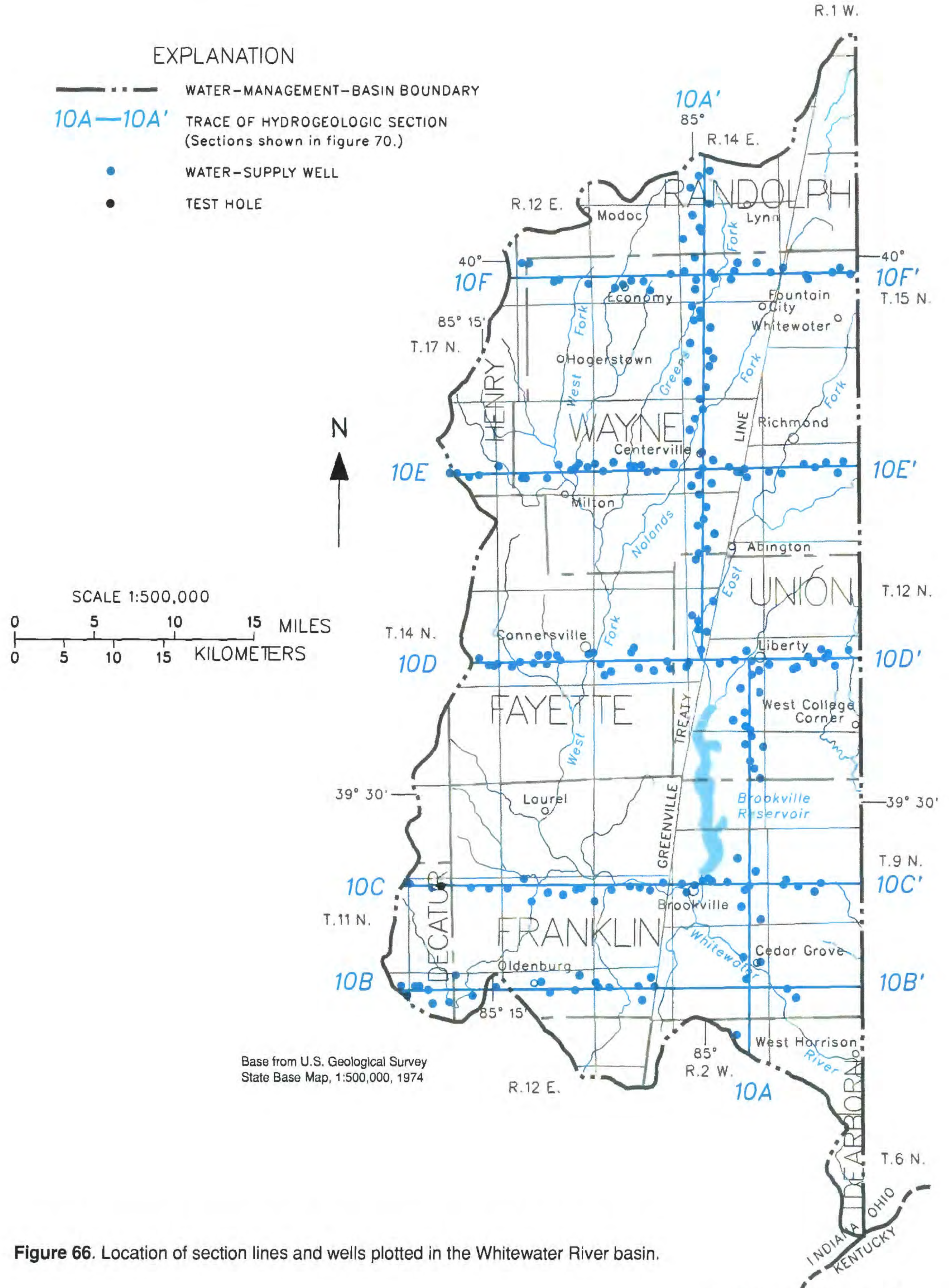


Figure 66. Location of section lines and wells plotted in the Whitewater River basin.

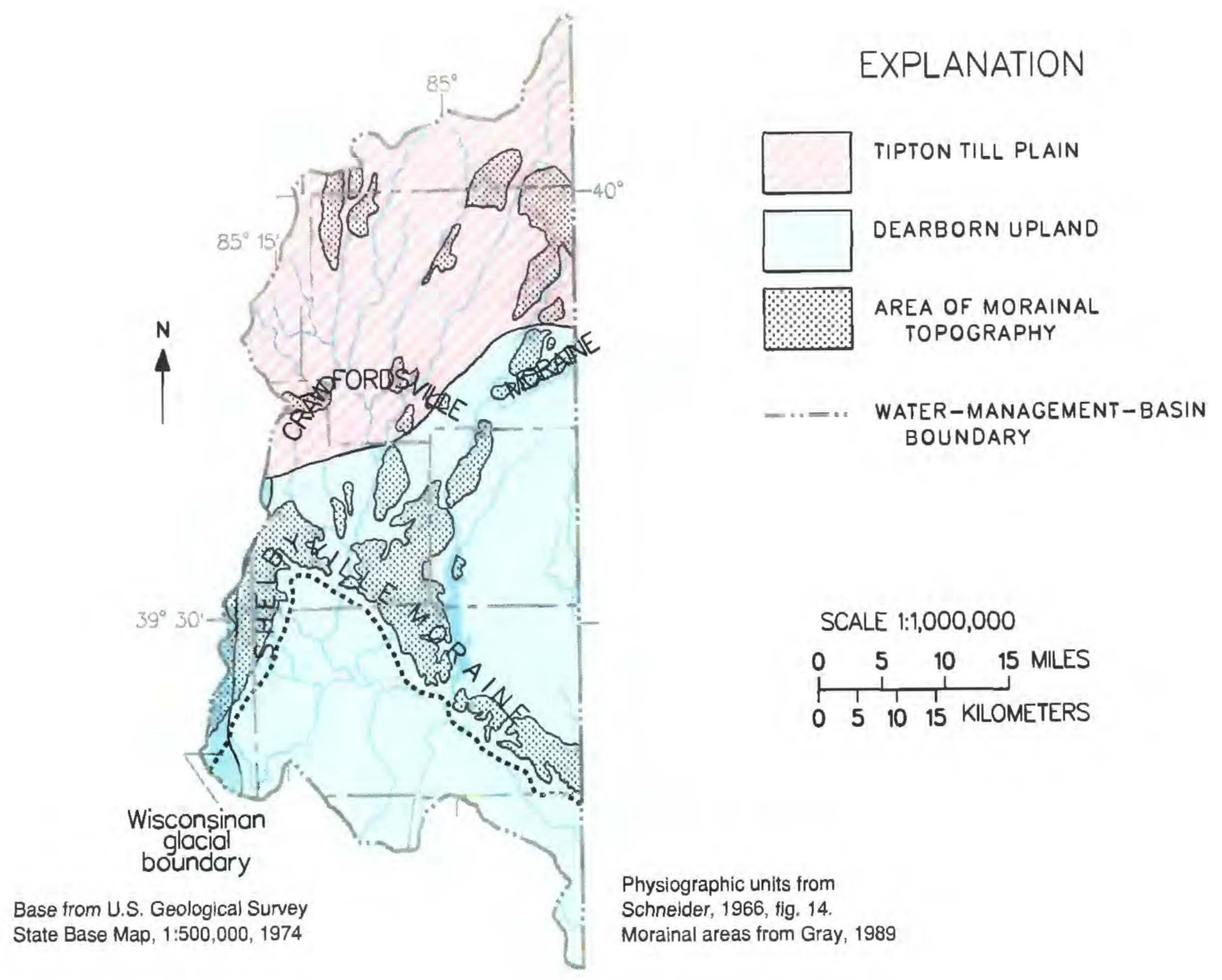


Figure 67. Physiographic units, moraines, and extent of glaciation in the Whitewater River basin.

The first and southernmost advance of the Erie Lobe (fig. 8) is marked by the segmented terminal Shelbyville Moraine (Wayne, 1965, p. 12). The Crawfordsville Moraine is a segmented recessional moraine marked by many kames and a large outwash plain (Wayne, 1965, p. 11). A segmented moraine deposited after the Crawfordsville Moraine (formerly called the Knightstown Moraine), covers most of the northern part of the basin and merges with the continuation of the Farmersville and Camden Moraines of Ohio (Wayne, 1965, p. 12). The boundary between the Tipton Till Plain and the Dearborn Upland is a broad arbitrary transitional zone where the glacial till is too thin to obscure the bedrock-surface relief of the Dearborn Upland (Schneider, 1966, p. 49).

The Dearborn Upland is a highly dissected bedrock plateau of rugged relief that is underlain by nearly flat-lying limestones and shales of Silurian and Ordovician age (Schneider, 1966, p. 42). The top of the plateau is covered by a cap of glacial till 15 to 50 ft thick (Schneider, 1966, p. 43). Large volumes of water from rapid ice melts have eroded deep valleys in this plateau, producing the major tributary valleys of the East and West Forks of the Whitewater River. These valleys have functioned as sluiceways for glacial meltwaters and extensive valley trains have developed within their courses. The valley-train deposits form a series of five to six outwash terraces that stand well above the present flood plains (Gooding, 1957, p. 1).

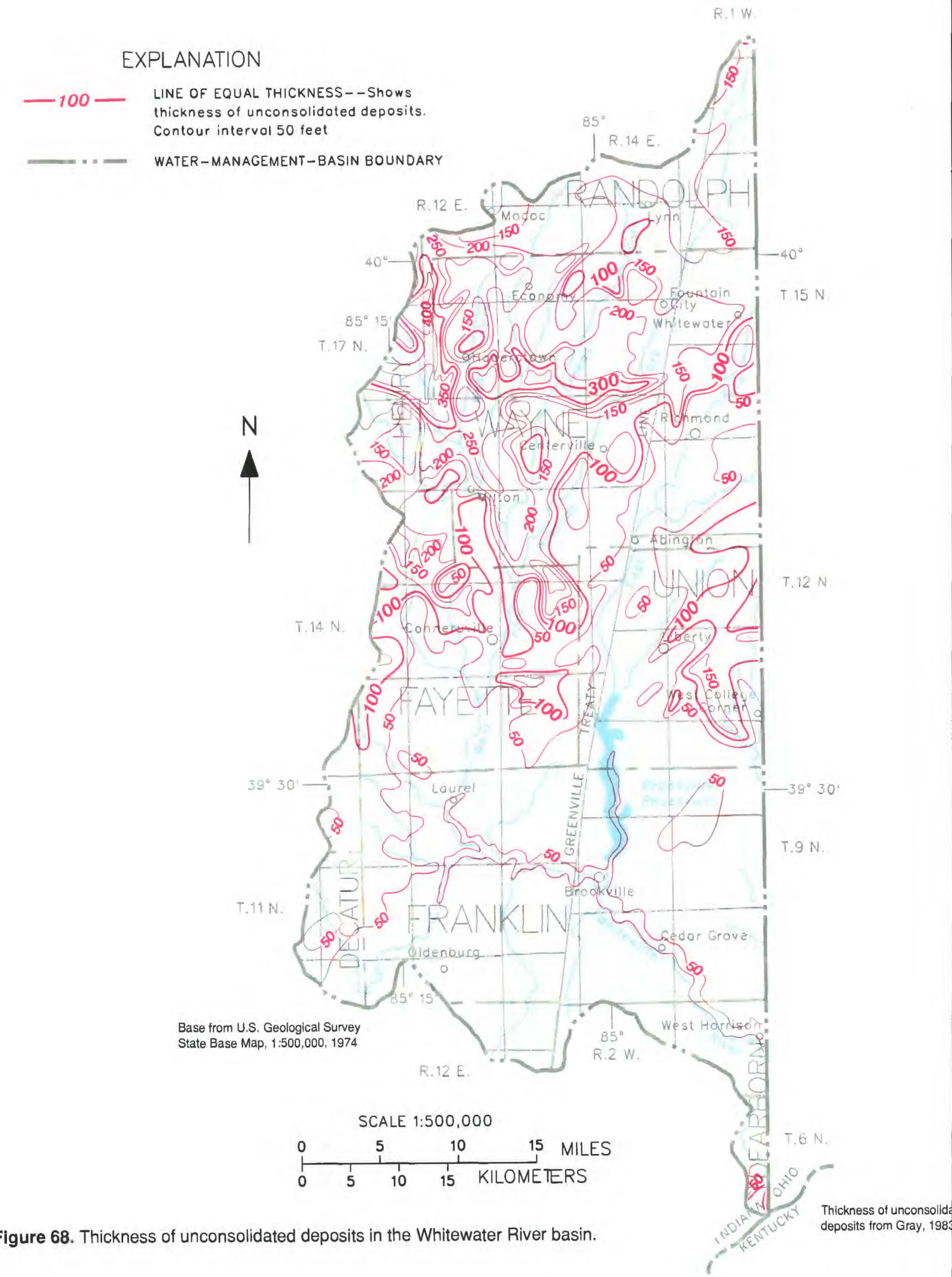


Figure 68. Thickness of unconsolidated deposits in the Whitewater River basin.

Surface-Water Hydrology

The Whitewater River, a major tributary of the Great Miami River in Ohio, drains 1,296 mi^2 in southeastern Indiana. Greens Fork, Nolands Fork, and the East and West Forks of the Whitewater River in southern Randolph County drain this basin to the Indiana-Ohio State line, near West Harrison, Ind. (fig. 66). The East Fork Whitewater River drains 352 mi^2 (312 mi^2 in Indiana), the West Fork drains 842 mi^2, and the main stem south of Brookville drains 145 mi^2 (142 mi^2 in Indiana) (Hoggatt, 1975, p. 78 and 94).

The Whitewater River water-management basin in Indiana also includes areas with headwater streams that drain into the Great Miami River in Ohio. These areas are in the northeastern, east-central, and southeastern parts of the basin and total more than 100 mi^2, accounting for the remainder of the basin's 1,425 mi^2.

Greens Fork and Nolands Fork, headwater streams of the West Fork Whitewater River, originate in the northern part of the basin and flow south-southwest (fig. 66). The East Fork Whitewater River originates in Ohio and drains the northeastern part of the basin as it flows south-southwest. The average discharge of the East Fork Whitewater River at Brookville at drainage area 380 mi^2 (1954-88) is 396 ft^3/s (Glatfelter and others, 1989, p. 49). The West Fork Whitewater River abruptly turns to the southeast in Franklin County where it joins the East Fork Whitewater River at Brookville. The average discharge of the Whitewater River at Brookville at drainage area 1,224 mi^2 (1916-17 and 1924-88) is 1,266 ft^3/s; flow ranged from a daily mean discharge of 60 ft^3/s to an instantaneous peak of 81,800 ft^3/s (Glatfelter and others, 1989, p. 50). The Whitewater River flows to the southeast out of Indiana through Dearborn County into the southwest-flowing Great Miami River in Ohio, which empties into the Ohio River at the intersection of Indiana, Ohio, and Kentucky.

The Whitewater River basin includes several manmade lakes that are used for reservoirs, especially in the southern part of the basin where ground-water supplies are limited. The East Fork Whitewater River flows into Brookville Reservoir (fig. 66), Indiana's second deepest (Indiana Department of Natural Resources, 1988, p. 1) and largest normal capacity reservoir (Ruddy and Hitt, 1990, p. 100).

Geology

Bedrock Deposits

The major geologic structure in the Whitewater River basin is the Cincinnati Arch (fig. 4) (Malott, 1922, p. 128). The crest of the arch trends north-northwest and bisects the basin. On the crest of the arch the rocks are nearly horizontal; the rocks on the western side dip west-southwest at an average of 25 ft/mi into the Illinois Basin (fig. 4) (Gutschick, 1966, p. 10), and rocks on the eastern side dip east to northeast at a similar rate into the Appalachian Basin. Because of this geologic structure and erosional processes, rocks of Ordovician age are exposed in the center of the basin, whereas rocks of Silurian age are exposed in the northeastern corner and southwestern edge of the basin (fig. 69).

The oldest rocks in the basin, at depths of 1,500 to 2,000 ft, are Precambrian granite, basalt, and arkose (fig. 5). These rocks are overlain by Cambrian rocks consisting of sandstone, shale, limestone, and dolomite (Gray, 1987). Unexposed Ordovician rocks that overlie the Cambrian rocks in the basin are composed primarily of dolomite, limestone, and sandstone. These Cambrian and Ordovician bedrock units are too deep to serve as a practical source for water supply, and the ground water is not potable in most places because of excessive concentrations of dissolved solids.

The oldest exposed bedrock in the Whitewater River basin is the Ordovician Maquoketa Group (figs. 5 and 69). This group is a westward-thinning wedge of rocks, approximately 700 to 1,000 ft thick in the basin, which consists primarily of shale in its lower part and limestone with smaller amounts of shale in its upper part (Gray, 1972, p. 5). The formations contained within this group include the Kope, Dillsboro, and Whitewater Formations.

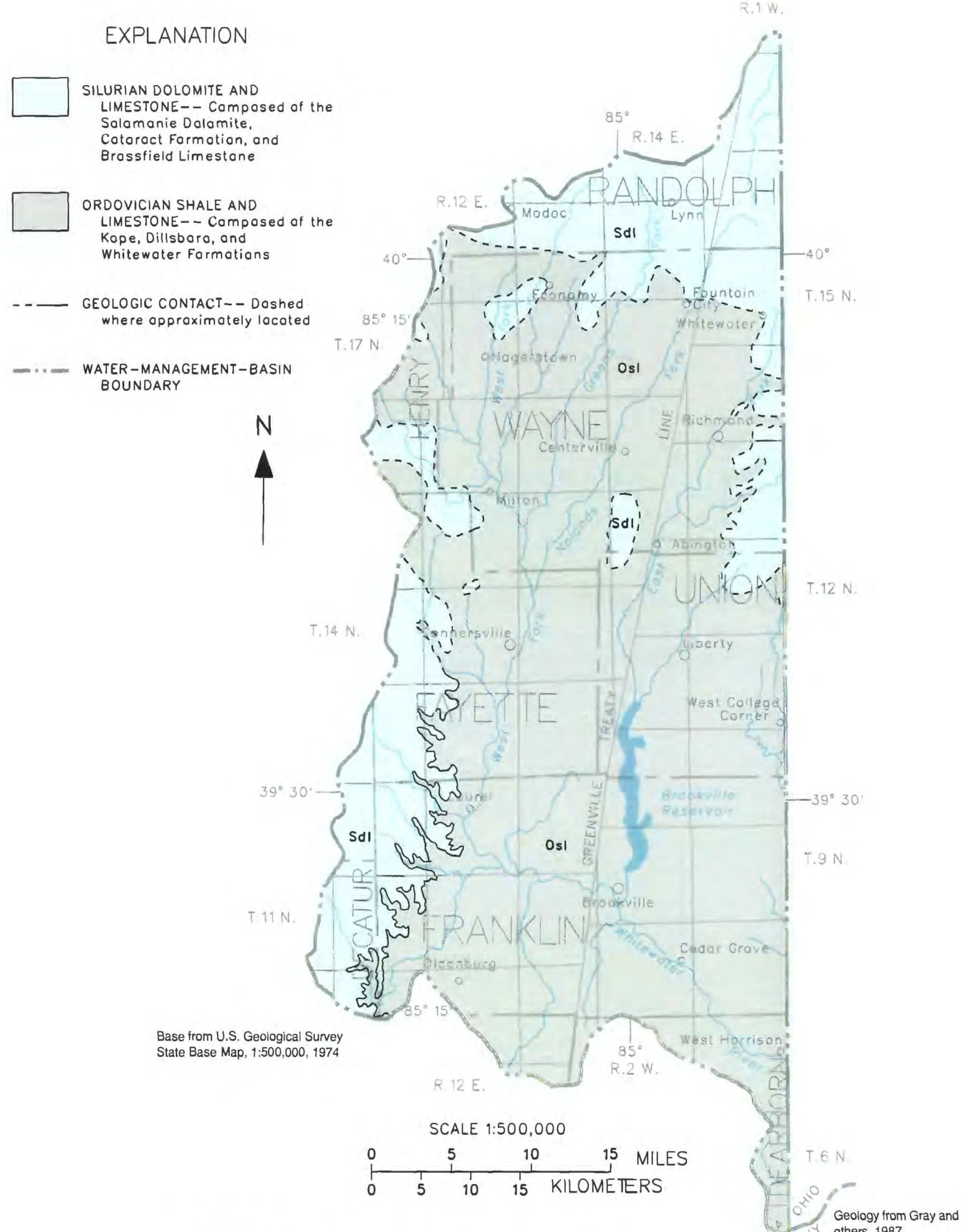


Figure 69. Bedrock geology of the Whitewater River basin.

The Kope Formation is a 300- to 400-ft-thick sequence, 95 percent of which consists of bluish- to brownish-gray shale (Shaver and others, 1986, p. 72). Exposed along the entrenched valleys of the lower reaches of the Whitewater River, this shale contains a thick basal dark-brown to nearly black shale. The upper contact between the Kope and the Dillsboro Formations is transitional; it is marked by a comparatively sharp upward increase in the proportion of limestone (Gray, 1972, p. 14).

In the Whitewater River basin, the Dillsboro Formation is an approximately 400-ft-thick sequence of shale with interbedded limestone (Shaver and others, 1986, p. 37). The Dillsboro is conformably overlain by the Whitewater Formation, which is a bluish-gray rubbly limestone interbedded with calcareous shale (Shaver and others, 1986, p. 168). The Whitewater Formation is disconformably overlain by the Silurian Brassfield Limestone.

Laferriere and others (1986, p. 1) noted that the disconformity between the Whitewater Formation and the Brassfield Limestone contains evidence of a complex erosional and depositional history. The disconformity is clearly defined by an abrupt vertical lithologic change (Laferriere and others, 1986, p. 4).

The Silurian Brassfield Limestone generally is a medium- to coarse-grained fossiliferous limestone that contains small amounts of shale and fine-grained dolomite (Shaver and others, 1986, p. 20). The Salamonie Dolomite, which unconformably overlies the Brassfield Limestone, is an impure dolomite that includes finer-grained clayey limestone, dolomitic limestone, and shale. Along the eroded edges of the unit in southeastern Indiana, the thickness ranges from 0 to 60 ft.

Unconsolidated Deposits

The Whitewater River basin is divided into three distinct areas of surficial unconsolidated deposits (Gray, 1989). The Jessup Formation of pre-Wisconsinan age is present in the southern one-third of the basin and is from an eastern source. The Trafalgar Formation of Wisconsinan age was deposited during several events by glaciers from a northeastern source and covers the central and northern parts of the basin. Buried pre-Wisconsin till is found beneath the Trafalgar Formation within the basin. Ice from two advances separated by a brief retreat deposited this pre-Wisconsinan till sequence (Gooding, 1966).

Pre-Wisconsinan loam to sandy-loam tills of the Jessup Formation are present in the southern part of the basin (Gray, 1989). Classically this formation has been referred to as the Illinoian drift. Gooding (1963) describes the drift in this area as consisting of calcareous oxidized till; noncalcareous, cherty, grayish-brown till; and calcareous sand and gravel. Minor deposits of outwash and alluvium are present within this area. The upland areas are capped by a silt complex composed of poorly stratified and poorly sorted sand and silt. These sediments are composed of weathering products and windblown silt (loess) (Gray, 1989).

The Trafalgar Formation in the central part of the basin is underlain by loam till and covered by a surficial layer of loess that ranges in thickness from 20 to 40 in. (Gray, 1989). The Wisconsinan glacial boundary and the Shelbyville Moraine (fig. 67) mark the southern edge of this unconsolidated material. Gooding (1975) described unnamed till units in the area as consisting of calcareous, brown to yellow-brown oxidized till along vertical joints, and gray unoxidized till. These units also contain some discontinuous and thin buff-colored calcareous sands and gravels. Extensive outwash material lies along the Whitewater River valley. Alluvium underlies the modern flood plain of the Whitewater River.

The northern one-third of the basin contains a thick sequence of older tills separated by buried soils and covered by Wisconsinan tills of the Trafalgar Formation (Gray, 1989). The northernmost part, an area of morainal topography, includes layers of ablation till, and dead-ice landforms. Within the headwater valleys of the Whitewater River and its major tributaries are linear, complex sequences of mixed drift, till, and stratified drift that appear to be collapse features. Further southward along the Whitewater River valley are a series of terraces composed of valley-train outwash (Gruver, 1984, p. 10). The outwash material contains large volumes of commercially extractable sand and gravel. The flood plain along the Whitewater River consists of Holocene alluvium of silt, sand, and gravel. The Crawfordsville Moraine marks the southernmost part of the northern region (fig. 67).

Within the northern part of the basin (western Wayne County and eastern Henry County) is a buried bedrock valley (fig. 7) that Bleuer (1989, p. 3) refers to as the New Castle Valley section of the Lafayette Bedrock Valley (formerly called the Teays Valley). This major preglacial drainageway originally flowed northwest, but it was blocked by ice and filled with glacial till. The valley is filled in places with more than 400 ft of unconsolidated deposits (fig. 68) consisting of clay interbedded with smaller amounts of sand and gravel. The valley is shown in Rs. 12 and 13 E. of section 10E–10E′, fig. 70).

Aquifer Types

Six hydrogeologic sections (10A–10A′ to 10F–10F′, fig. 70) were produced for this atlas to show the hydrostratigraphy of the Whitewater River basin. Hydrogeologic section 10A–10A′ bisects the basin and is oriented south to north (except for a 3 mi segment in the middle of the section that is oriented east-west and is also part of section 10B–10B′). The southern part of section 10A–10A′ crosses the lower reaches of the Whitewater River and roughly parallels the east side of the Brookville Reservoir, whereas the northern part of the section crosses Nolands Fork and Greens Fork. The remaining hydrogeologic sections (10B–10B′ to 10F–10F′) are oriented west to east and are approximately perpendicular to the surficial drainage of the basin (fig. 66). The west-east sections are spaced at intervals of approximately 8 to 12 mi. The average density of the 226 logged wells plotted along the hydrogeologic sections is 1.2 wells per mile.

Both the unconsolidated and bedrock deposits in the Whitewater River basin contain aquifers. The unconsolidated aquifers are composed of outwash in valley trains, intertill lenses of sand and gravel, and recently deposited alluvium. The bedrock aquifers are in the Silurian carbonate rocks and in an upper weathered zone in the Ordovician shale and limestone. Five aquifer types are mapped in the Whitewater River basin (fig. 71) and are summarized in table 12. The table includes information on the thickness, range of yields, and common aquifer names used by other authors.

Unconsolidated Aquifers

The unconsolidated aquifers, which contain the largest supply of ground water in the basin, consist of complexly interbedded glacial deposits. For simplicity, these sand and gravel aquifers are subdivided into three distinct aquifer types (fig. 71): discontinuous zones of sand and gravel, buried sand and gravel, and surficial sand and gravel.

Numerous layers of discontinuous sand and gravel aquifers underlie the northern two-thirds of the basin (fig. 71). Sand and gravel layers range from 2 to 50 ft in thickness and are not areally extensive. Hydrogeologic sections 10A–10A′ and 10D–10D′ to 10F–10F′ (fig. 70) show this type of aquifer material. Wells in areas of discontinuous sand and gravel are typically drilled and screened in the first significant water-bearing zone; however, some wells tap multiple zones.

An area with buried sand and gravel aquifers (fig. 71) that range in thickness from 2 to 50 feet is in the northwestern part of the basin. The layers of buried sand and gravel are more continuous (greater than 15 mi^2) than the discontinuous sand and gravel lenses. The buried sand and gravel is shown in hydrogeologic sections 10A–10A′, 10E–10E′, and 10F–10F′ (fig. 70). In many places, water-bearing zones of discontinuous sand and gravel are encountered above or below the more continuous aquifer. These multiple layers of sand and gravel deposits are separated by as little as 10 feet to as much as 50 feet of glacial till. Sand and gravel layers at different elevations represent deposition of outwash during various advances and retreats of the ice during Wisconsinan and pre-Wisconsinan glaciations.

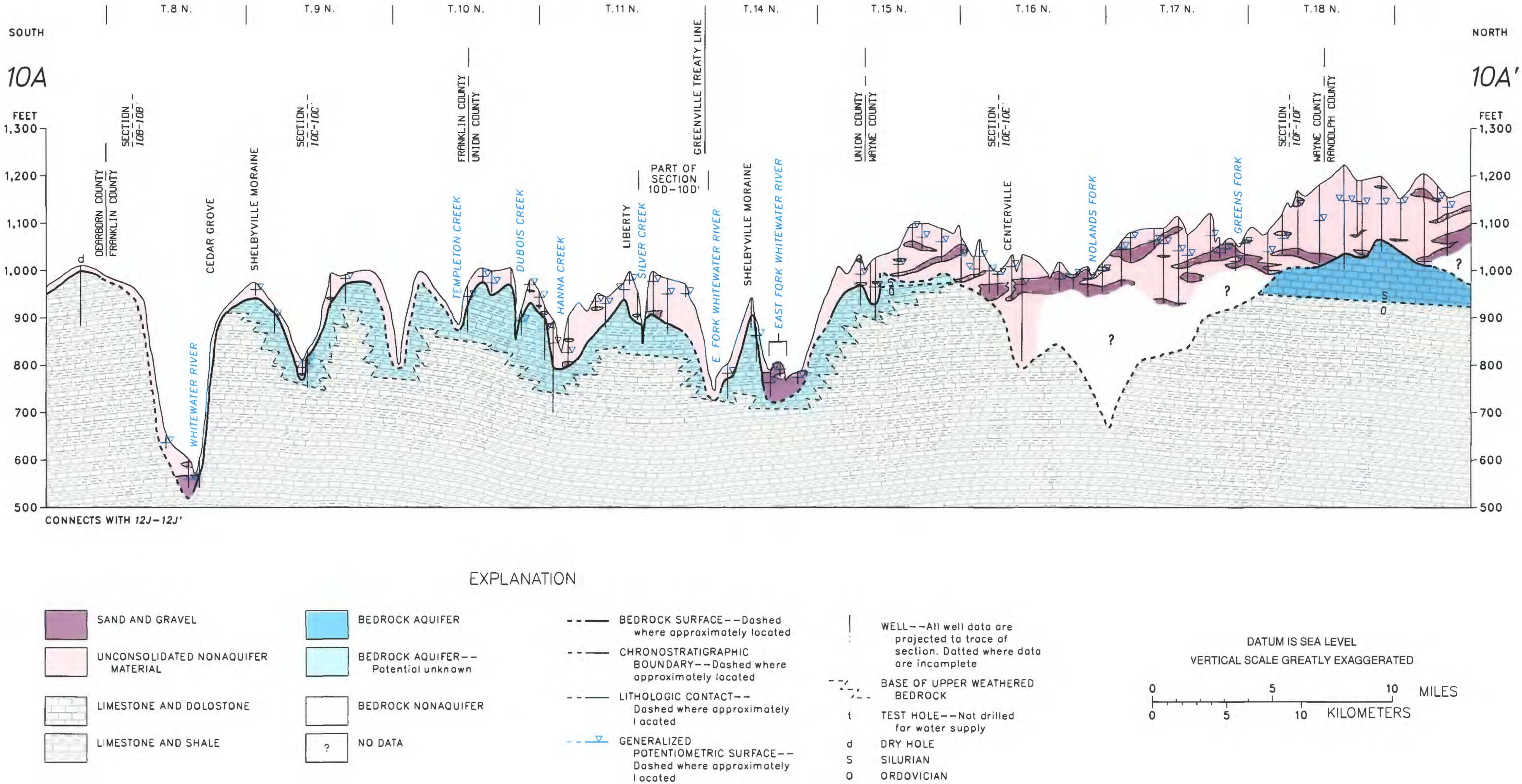


Figure 70. Hydrogeologic sections *10A–10A'* to *10F–10F'* of the Whitewater River basin.

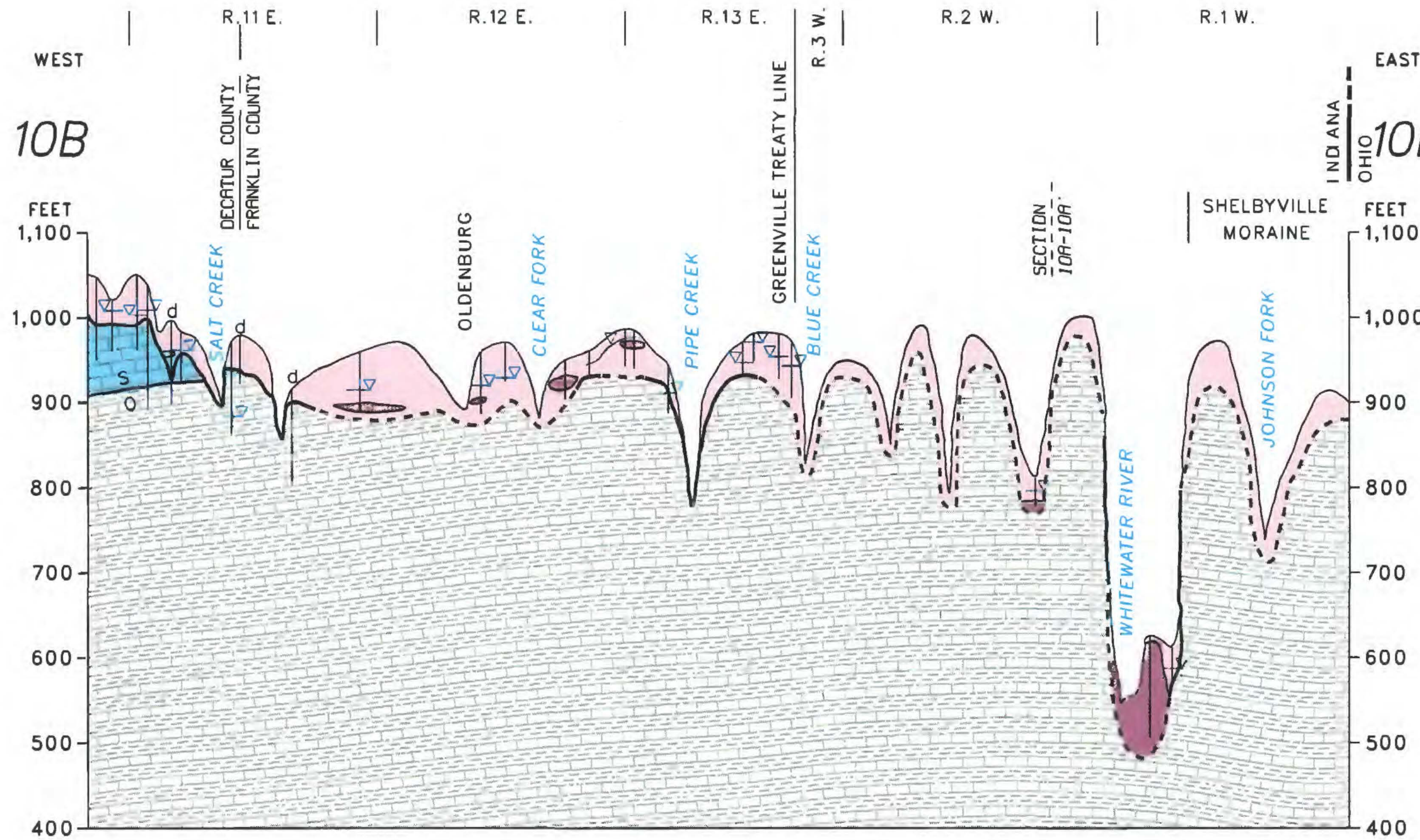
Figure 70. Hydrogeologic sections *10A–10A′* to *10F–10F′* of the Whitewater River basin—Continued.

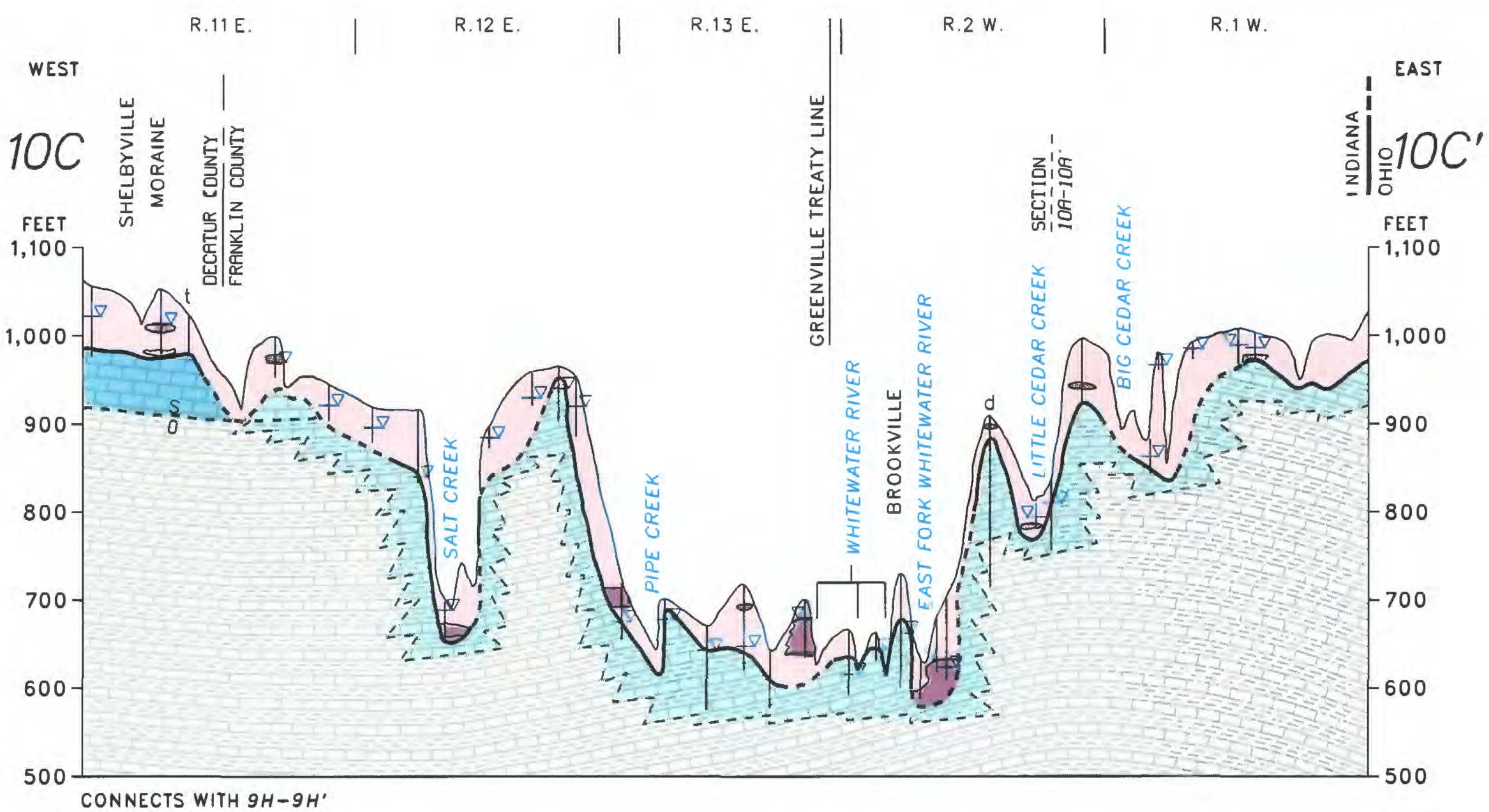
DATUM IS SEA LEVEL

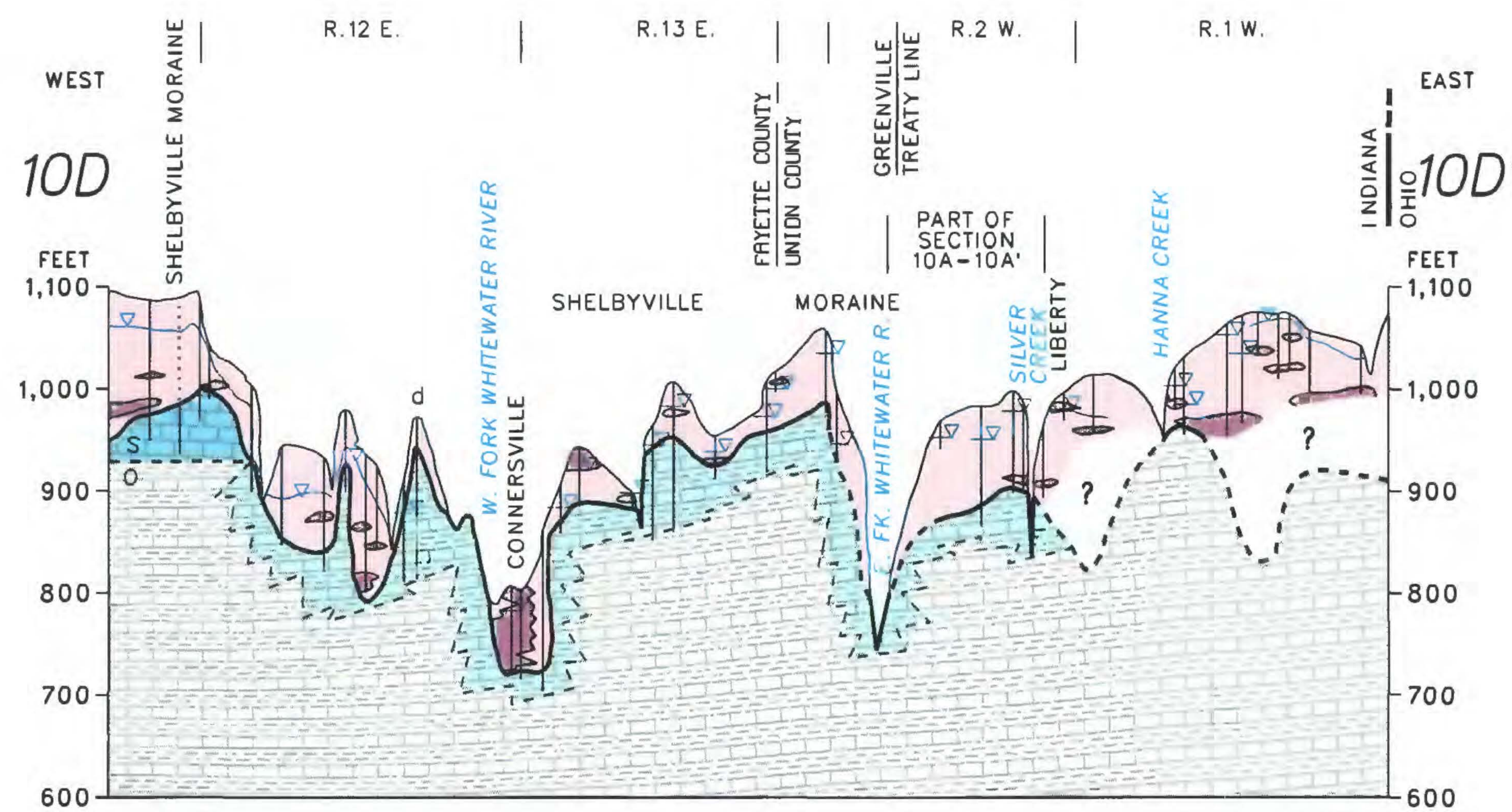
VERTICAL SCALE GREATLY EXAGGERATED

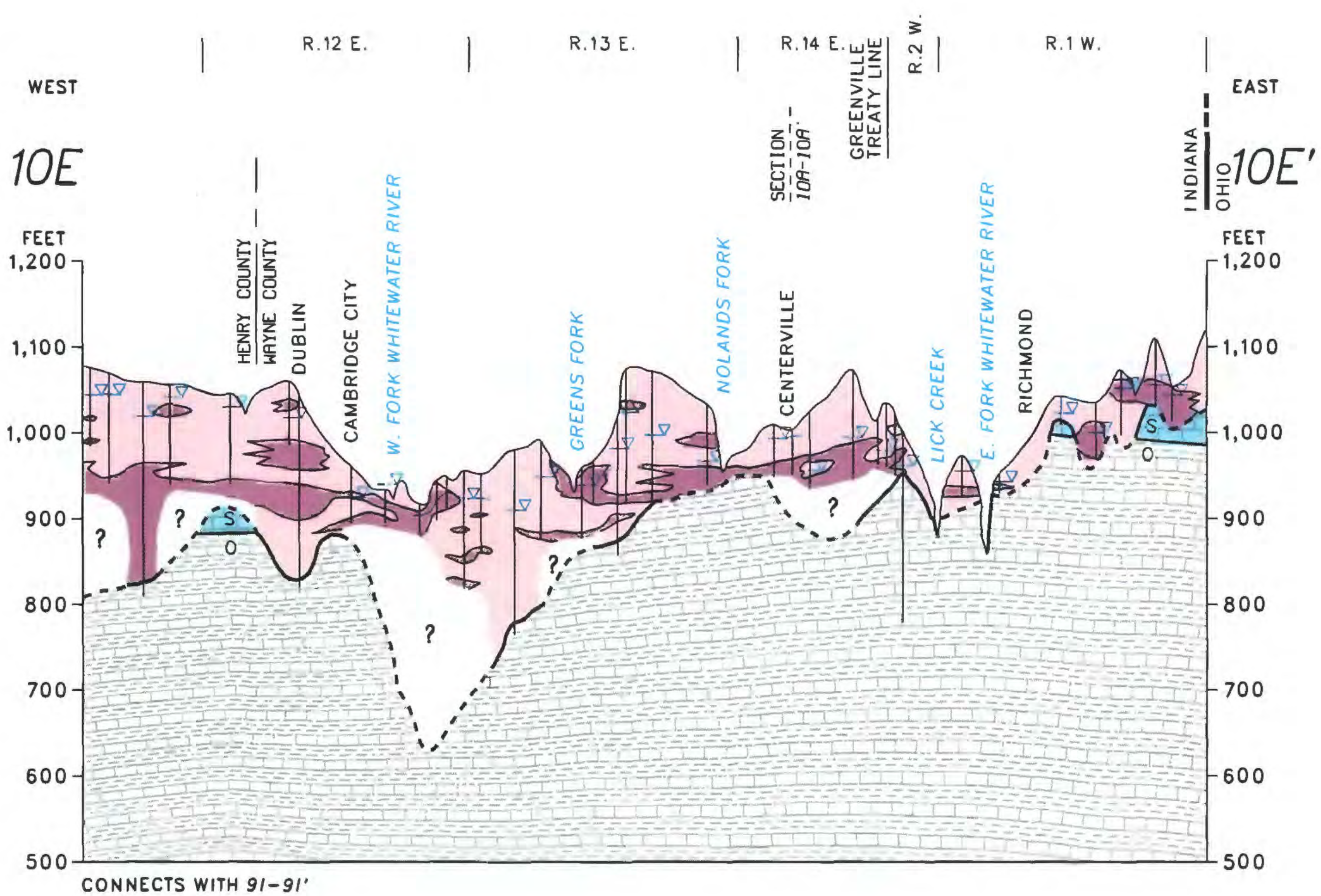
0 5 10 MILES

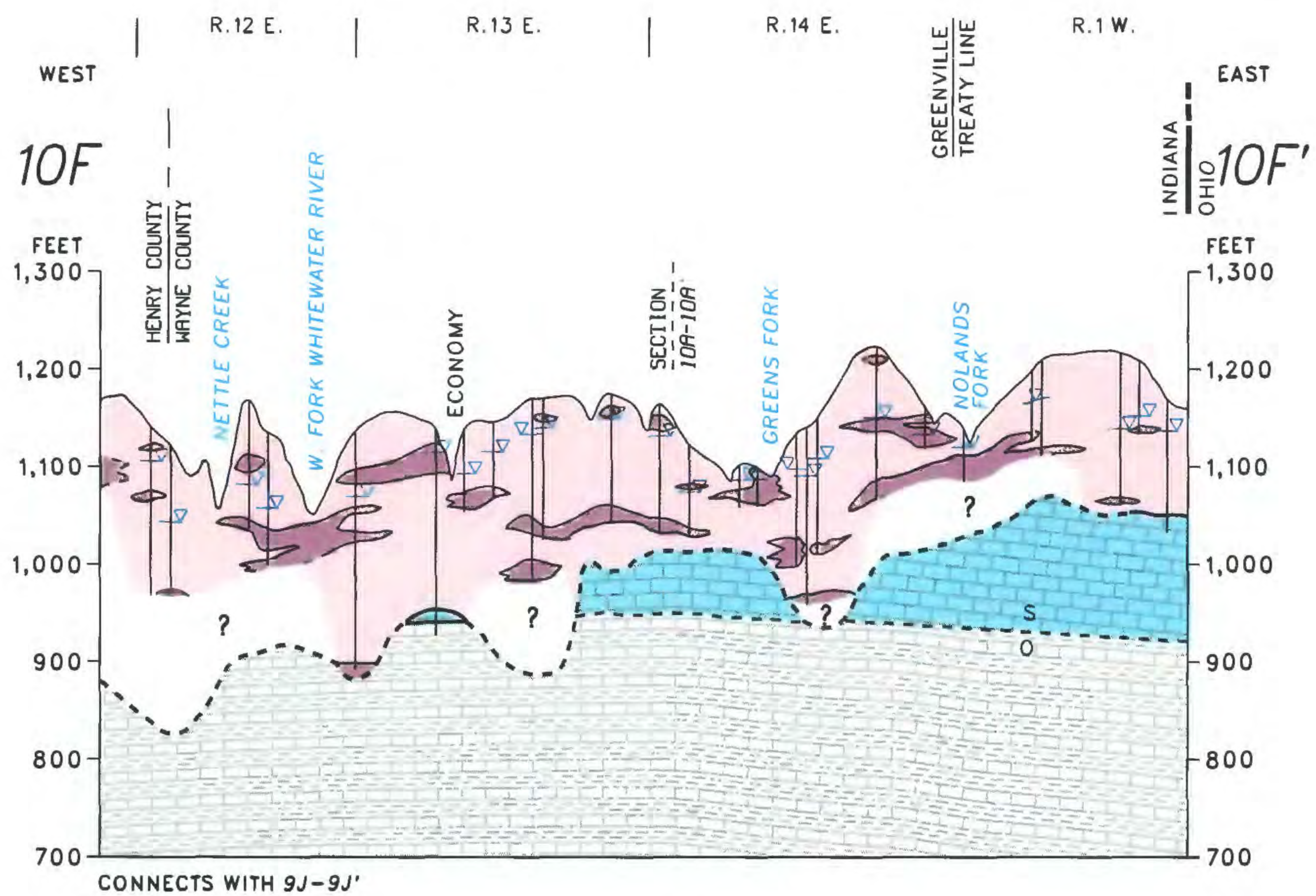
0 5 10 KILOMETERS











The New Castle Valley Section, a southeastern tributary of the Lafayette Bedrock Valley (Bleuer, 1989) (fig. 7) is located along the northwestern edge of the basin. The bedrock valley is shown beneath Nolands Fork in section 10A–10A′ (fig. 70) and beneath the West Fork Whitewater River in section 10E–10E′ (fig. 70). The buried bedrock valley is overlain by a confining layer of clay and contains discontinuous and buried sand and gravel layers and lenses. The extent of the aquifers within the buried bedrock valley is not known because wells within this area penetrate to a depth of only 200 ft or less, whereas the valley extends to a depth of 400 ft or more (fig. 68).

Surficial sand and gravel aquifers are present in the valleys of the Whitewater River and its major northern tributaries (fig. 71). These surficial aquifers are shown on all the hydrogeologic sections except 10F–10F′ (fig. 70). The principal water-bearing materials are extensive sand and gravel outwash deposits that were deposited as valley fill during the various periods of glacial advance and retreat, and more recently, as alluvium. Ground-water yields from these surficial sand and gravel aquifers are the highest measured for the entire basin (table 12).

Nonaquifer material, consisting mostly of sandy and pebbly clay, covers the southern part of the basin. Although no aquifer material is mapped in this area, large-diameter domestic wells are constructed in marginal water-bearing zones using an upper fractured clay zone or thin sand and gravel lenses within the clay as a source of water. Areas underlain by nonaquifer material can best be seen in hydrogeologic section 10B–10B′ (fig. 70).

Bedrock Aquifers

The Silurian carbonate bedrock aquifer forms an important aquifer in the northeastern and in the southwestern margins of the basin (fig. 71). This aquifer ranges in thickness from a few feet to more than 100 ft and is shown in all hydrogeologic sections (fig. 70). The carbonate bedrock aquifer is composed of thick-bedded limestone with thin interbedded shales. Most wells screened in these carbonate rocks receive water from the top few feet of the weathered bedrock and(or) intersect bedrock fractures or joints at depth. The fractures and joints were probably produced by the tectonic stresses that upwarped the Cincinnati Arch and by the overriding of glaciers at various times. Zones of moderate to high permeability along bedding planes and disconformities between successive rock strata, such as the Silurian-Ordovician unconformity, also increase well yields. Lateral changes in permeability can result from facies changes and fracture density. Ground-water yields from the carbonate bedrock aquifer are generally greater in the northern part of the basin where yields are usually more than 10 gal/min (Indiana Department of Natural Resources, 1988, p. 44); overall, well yields are generally sufficient for domestic use, although attempts to construct wells occasionally result in dry holes.

Shale and limestone of the Ordovician Maquoketa Group are present as subcrop or are exposed in the entrenched valleys in most of the basin from the northwest to the southeast (fig. 69). The Ordovician bedrock consists predominantly of nearly flat-lying interbedded layers of thick shale and thin limestone. These rocks form the bedrock surface across more than two-thirds of the basin and underlie the carbonate bedrock elsewhere. Because of the high shale content, permeability of these rocks is low. The small amount of water that does circulate through these rocks is within small joints and fractures at or near the bedrock surface. An upper weathered zone in the shale and limestone bedrock was mapped as "aquifer—potential unknown" in the central part of the basin (fig. 71). In this part of the basin, the shale and limestone are at the bedrock surface and are commonly used as a water supply. Areas where the shale and limestone bedrock is rarely used were not mapped as an aquifer even though bedrock could possibly provide ground water in these areas. Wells drilled into the shale and limestone can be seen in all six hydrogeologic sections, although most of the wells are shown in sections 10A–10A′, 10C–10C′, and 10D–10D′ (fig. 70). Most wells penetrate less than 100 feet into the upper weathered bedrock and yield less than 10 gal/min; dry holes are not uncommon. The upper weathered-bedrock aquifer is generally used if no other aquifer is available.

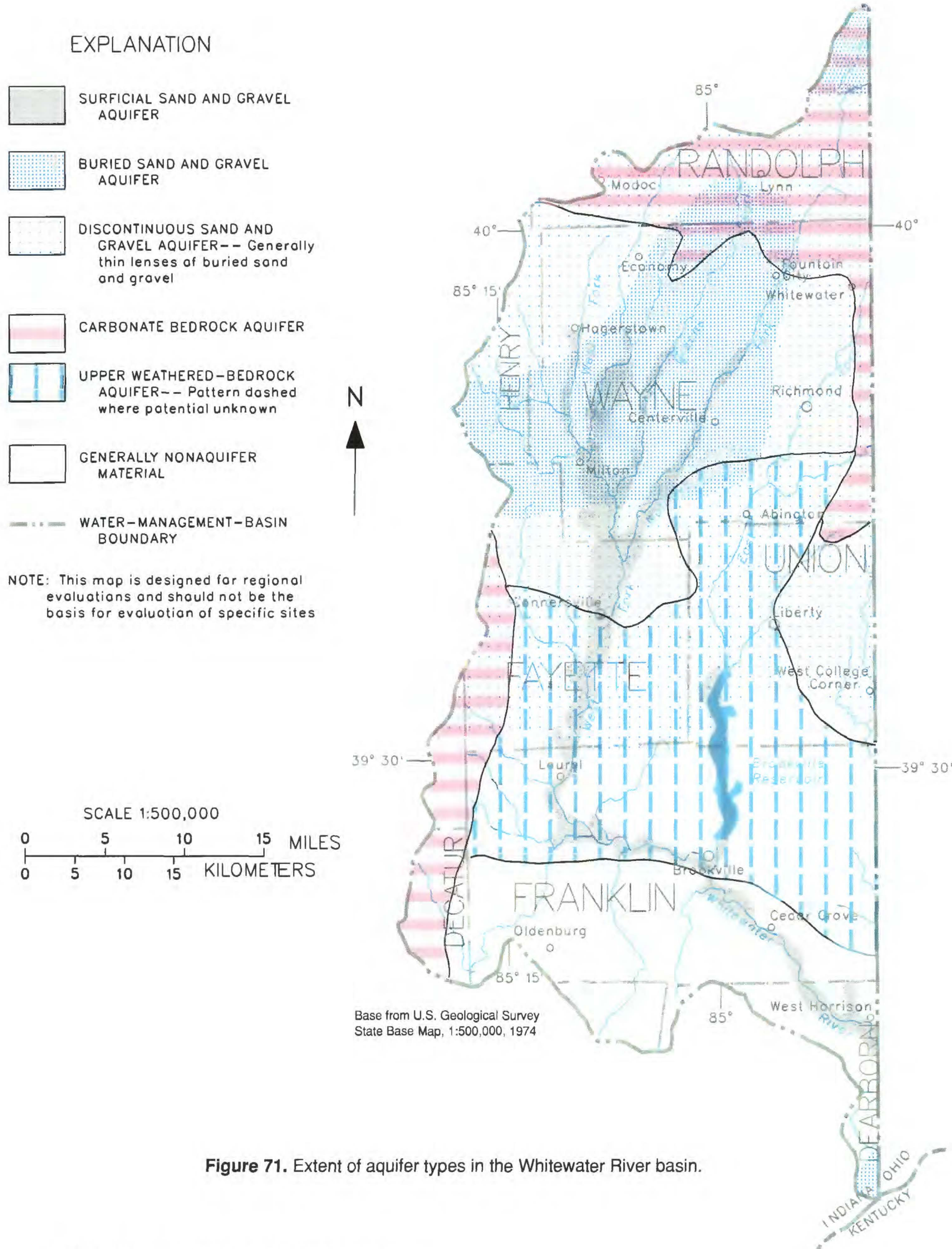


Figure 71. Extent of aquifer types in the Whitewater River basin.

Table 12. Characteristics of aquifer types in the Whitewater River basin
[<, less than; locations of aquifer types shown in fig. 71]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Discontinuous sand and gravel	2- 50	0- 150	Fayette-Union and Wayne-Henry Aquifer Systems[1]
Buried sand and gravel	2- 50	5- 150	Centerville Subsystem[1]
Surficial sand and gravel	30- 150	50- 1,200	Whitewater Valley Aquifer System[1]
Carbonate bedrock	0- 100	0- 60	Silurian Bedrock Aquifer System[1]
Upper weathered bedrock	<100	0- 10	Ordovician Bedrock Aquifer System[1]

[1]Indiana Department of Natural Resources, 1989.

Summary

The Whitewater River basin encompasses 1,425 mi^2 in southeastern Indiana. The basin is composed of two physiographic units; the undulating Tipton Till Plain in the north and the rugged Dearborn Upland in the south. Silurian carbonate rocks are at the bedrock surface on the western, northern, and northeastern edges of the basin, whereas Ordovician shale and limestone are at the bedrock surface in the remaining three-quarters of the basin. The bedrock is covered with as much as 400 ft of loam to sandy-loam till, loess, outwash, and alluvium.

Five different aquifer types are delineated in the Whitewater River basin: discontinuous sand and gravel; buried sand and gravel; surficial sand and gravel; carbonate bedrock; and an upper weathered zone in shale and limestone. The unconsolidated aquifers produce the largest supply of ground water for the basin; typical well yields range from approximately 10 gal/min within the discontinuous zones to several hundred gallons per minute in the surficial sand and gravel aquifers along the Whitewater River. The carbonate bedrock aquifer is a secondary supply of ground water, mostly for domestic use. Wells in the carbonate bedrock aquifer yield from 0 to 60 gal/min; the larger yields are obtained in the northern part of the basin where one or more fractures or joints are intersected. The upper weathered-bedrock aquifer is generally used if no other aquifer is available. Well penetration into the weathered zone is usually less than 100 ft, and most well yields are less than 10 gal/min. No aquifer is mapped in the southern part of the basin. Most ground water in this area is obtained from low-permeability unconsolidated deposits.

References Cited

Bechert, C.H., and Heckard, J.M., 1966, Ground water, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 100-115.

Bleuer, N.K., 1989, Historical and geomorphic concepts of the Lafayette Bedrock Valley System (so-called Teays Valley) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 46, 11 p.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Governor's Water Resource Study Commission, Indiana Department of Natural Resources, 508 p.

Clark, J.C., 1992, Hydrogeology of Union County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 42, 48 p.

Glatfelter, D.R., Thompson, R.E., Jr., and Nell, G.E., 1989, Water resources data, Indiana, water year 1988: U.S. Geological Survey Water Data Report IN-88-1, 331 p.

Gooding, A.M., 1957, Pleistocene terraces in the Upper Whitewater drainage basin, southeastern Indiana: Richmond, Ind., Earlham College Science Bulletin 2, 65 p.

_____1963, Illinoian and Wisconsin glaciations in the Whitewater Basin, southeastern Indiana, and adjacent areas: Journal of Geology, v. 71, no. 6, p. 665-682.

_____1966, The Kansan glaciation in southeastern Indiana: Ohio Journal of Science, v. 66, p. 426-433.

_____1973, Characteristics of Late Wisconsin till in eastern Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 49, 28 p.

_____1975, The Sidney Interstadial and late Wisconsin history in Indiana and Ohio: American Journal of Science, v. 275, p. 993-1101.

Gray, H.H., 1972, Lithostratigraphy of the Maquoketa Group (Ordovician) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 7, 31 p.

_____1983, Map of Indiana showing the thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

_____1989, Quarternary geologic map of Indiana: Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Gruver, B.L., 1984, Pleistocene valley-train deposition, Whitewater River, southeastern Indiana and southwestern Ohio: Bloomington, Ind., Indiana University, Master's thesis, 403 p.

Gutschick, R.C., 1966, Bedrock geology, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 1-20.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Indiana Department of Natural Resources, 1988, Water resource availability in the Whitewater River basin, Indiana: Indiana Department of Natural Resources, Division of Water, Water Resource Assessment 88-2, 126 p.

Laferrierre, A.P., Hattin, D.E., Foell, C.J., and Abdulkareem, T.F., 1986, The Ordovician-Silurian unconformity in southeastern Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 53, 12 p.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana geology: Indiana Department of Conservation, Division of Geology Publication 20, 1120 p.

Reynolds, R.J., 1993, Hydrogeology of Fayette County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 43, 29 p.

Ruddy, B.C., and Hitt, K.J., 1990, Summary of selected characteristics of large reservoirs in the United States and Puerto Rico, 1988: U.S. Geological Survey, Open-File Report 90-163, 295 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed., Natural features of Indiana: Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Wayne, W.J., 1965, The Crawfordsville and Knightstown Moraines in Indiana: Indiana Department of Conservation, Geological Survey Report of Progress 28, 15 p.

Wolf, R.J., 1969, Reconnaissance of the ground-water resources of the Whitewater River basin and selected adjacent areas, southeastern Indiana: Indiana Department of Natural Resources, Appendix to the State Water Plan of the Division of Water, 47 p.

PATOKA RIVER BASIN

By David A. Cohen

General Description

The Patoka River drains 862 mi^2 (Hoggatt, 1975) within a long, narrow basin in southwestern Indiana. The basin is approximately 12 to 16 mi wide throughout most of its 78-mi length. The Patoka River basin includes parts of northern Gibson County, the southern three-quarters of Pike and Dubois Counties, the southern one-third of Orange County, the northeastern corner of Crawford County, and smaller areas in three adjacent counties (fig. 72).

Previous Studies

The ground-water hydrology of the Patoka River basin has not been intensively studied. Watkins (1963) briefly outlined the water resources of the basin with an emphasis on the area surrounding Patoka Lake (fig. 72). The major aquifers in Gibson County were described by Barnhart and Middleman (1990). Harrell (1935) summarized the water resources of each county in Indiana, and Bechert and Heckard (1966) and Clark (1980) constructed ground-water-availability maps for the entire State.

The effects of surface coal mining on ground- and surface-water hydrology were investigated by Corbett (1965, 1968) and Banaszak (1985). Banaszak (1980) also investigated the use of coal beds as aquifers in southwestern Indiana. Wangsness and others (1981) summarized hydrologic data for ground water and surface water in an area of the Eastern Coal Region that includes the Patoka River basin.

Physiography

The Patoka River basin lies in two distinct physiographic units as defined by Schneider (1966) and Malott (1922). The eastern one-half of the basin is part of the Crawford Upland and the western one-half is part of the Wabash Lowland (figs. 2 and 73).

The Crawford Upland is underlain by Upper Mississippian and Lower Pennsylvanian sandstone, shale, and limestone. Differential erosion has produced a deeply dissected upland with abundant stream valleys and a well-integrated drainage system (Gray and others, 1957, p. 5; Schneider, 1966, p. 47-48). This part of the Patoka River basin is typified by generally flat-topped drainage divides and steep-walled valleys with as much as 350 ft of relief (Renn, 1989, p. 9; Schneider 1966, p. 48). Level tracts of land are generally found only along the flood plains of the Patoka River and some large tributaries, such as Straight River and Polson Creek.

The boundary between the Crawford Upland and the Wabash Lowland (fig. 73) is not well defined, but it is probably best placed at Jasper, Ind., (fig. 72) or just a little to the west of Jasper (Gray, 1963, p. 6). In this area, relief gradually decreases to the west over several miles as valleys widen and become increasingly filled with alluvium above their bedrock floors (Malott, 1922, p. 98-99).

The Wabash Lowland is characterized by extensively aggraded valleys and uplands that consist of rolling plains. Drainage divides in this part of the Patoka River basin are commonly 100 ft above the valley floors, and maximum relief is about 150 ft (Schneider, 1966, p. 49). Lake plains are found in the northern part of the basin in Gibson, Pike, and Dubois Counties. These plains mark the sites of glacial lakes that were formed when northward-flowing streams were ponded by advancing Illinoian ice or Wisconsinan outwash (Fuller and Ashley, 1902, p. 4; Fuller and Clapp, 1904, p. 4; Malott 1922 p. 144-146, 251-255; Thornbury, 1950, p. 5, 10). Isolated bedrock hills that rise above the alluvium or silt of the surrounding lake plain are common in the Wabash Lowland (Malott, 1922, p. 103).

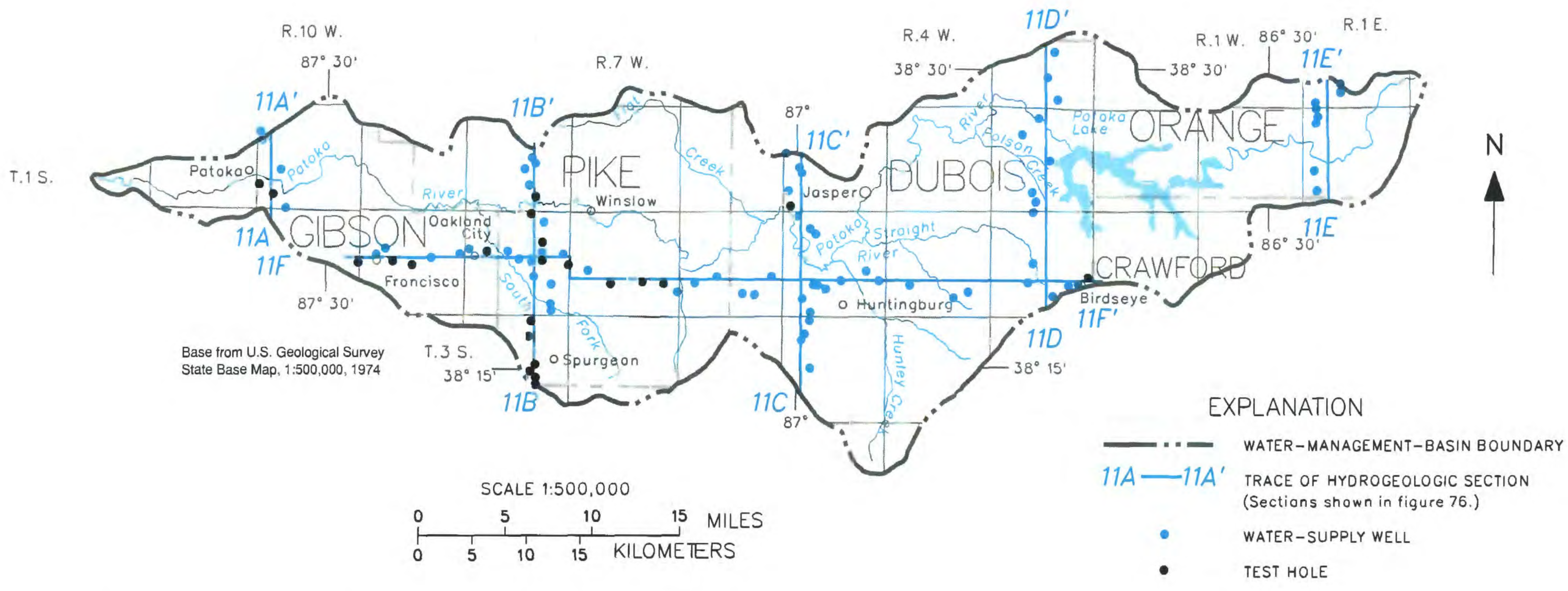


Figure 72. Location of section lines and wells plotted in the Patoka River basin.

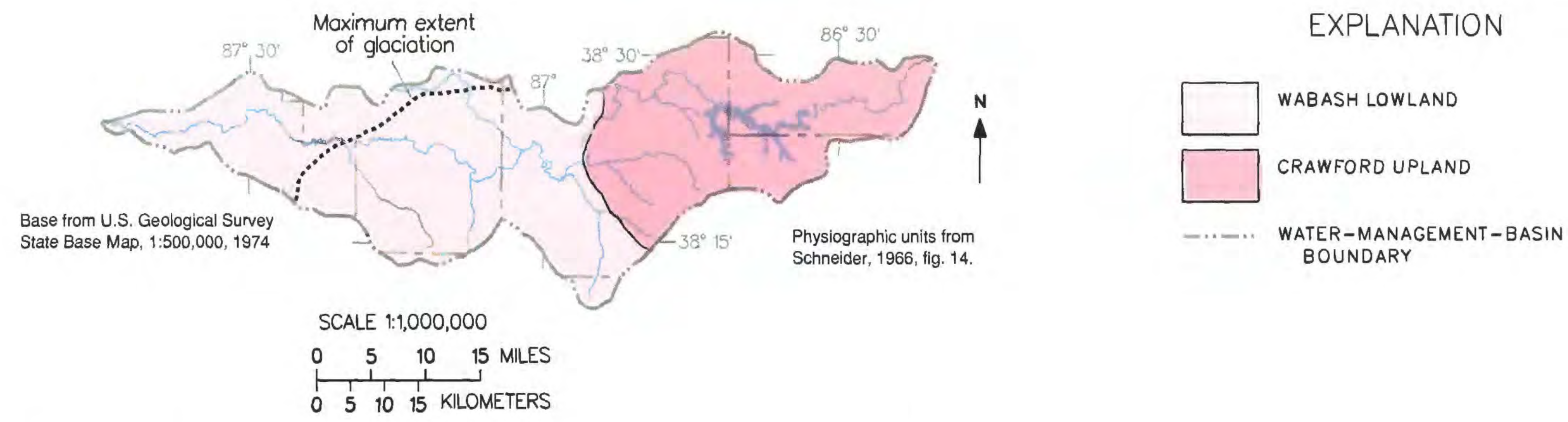


Figure 73. Physiographic units and extent of glaciation in the Patoka River basin.

Surface-Water Hydrology

The Patoka River originates in a group of hills in southeastern Orange County and flows westward to its mouth at the confluence with the Wabash River in extreme western Gibson County (fig. 72). The average gradient over this 161-river-mile course is approximately 1.74 ft/mi (Gray, 1963, p. 6; Hoggatt, 1975). The four largest contributing watersheds are Hunley Creek, South Fork Patoka River, Straight River, and Flat Creek (fig. 72) whose drainage areas are 82.0, 76.3, 67.6, and 58.9 mi^2 respectively (Hoggatt, 1975). These watersheds compose approximately one-third of the total area drained by the Patoka River.

For the period 1934 through 1985, the average discharge for the Patoka River approximately 1 mi upstream of hydrogeologic section 11A–11A′ (fig. 72) was 1,037 ft^3/s with a minimum daily mean discharge of zero during the period August 29 through September 12, 1936, and a maximum instantaneous discharge of 18,700 ft^3/s on January 26, 1937 (Arvin, 1989, p. 738).

Streamflow in the Patoka River has been regulated since 1978 by Patoka Lake (fig. 72). The reservoir, which is used for flood control, water supply, and recreation, is the third largest body of water in the State and has a capacity of 178,730 acre-ft (Renn, 1989, p. 11). Flow-duration analysis by Renn (1989, p. 27) indicates that regulation has generally increased low streamflows and decreased high streamflows downstream from the reservoir.

Streamflow in the Patoka River watershed can also be affected by spoil from surface coal mines (Martin and Crawford, 1987, p. 4). Corbett (1965, p. 2-3) concluded that cast overburden from surface coal mining is a significant source of streamflow in mined watersheds of the Patoka River during periods of extreme drought.

Geology

Bedrock Deposits

Bedrock underlying the Patoka River basin is part of the eastern limb of the Illinois Basin, a prominent regional downwarp centered in southeastern Illinois (fig. 4). Outcrops and subcrops in the Patoka River basin include, from east to west and oldest to youngest, the Blue River, West Baden, Stephensport, and Buffalo Wallow Groups of Mississippian age, and the Raccoon Creek, Carbondale, and McLeansboro Groups of Pennsylvanian age (Gray and others, 1987) (figs. 5 and 74). These rocks dip west-southwest at an average rate of approximately 25 to 30 ft/mi (Gray, 1979, p. K3).

The bedrock surface in the eastern quarter of the Patoka River basin is composed of Mississippian rocks (fig. 74). The middle Mississippian Blue River Group ranges in thickness from 150 to 650 ft and is composed predominantly of carbonate rocks with lesser amounts of evaporites, shale, chert, and calcareous sandstone. The three formations of the Blue River Group, in ascending order, are the St. Louis, Ste. Genevieve, and Paoli Limestones (Gray, 1979, p. K9; Shaver and others, 1986, p. 16).

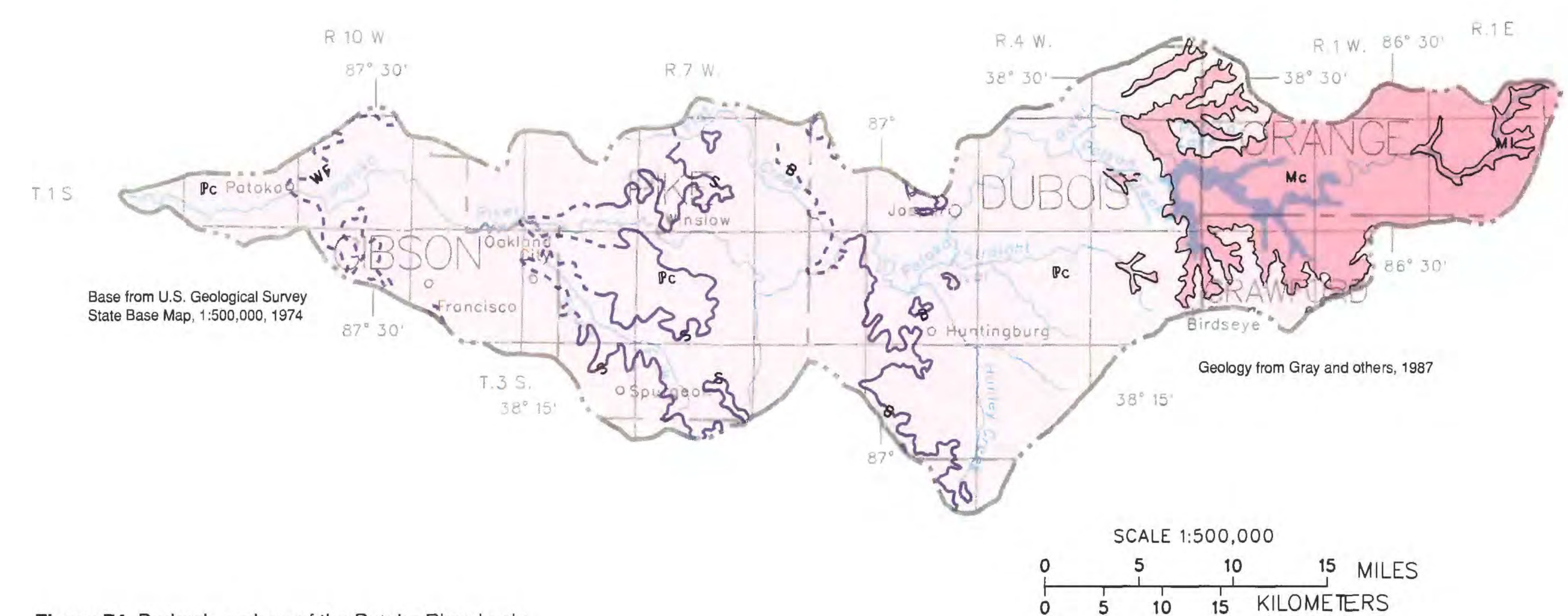
Overlying the limestone sequence of the Blue River Group are three Upper Mississippian groups that consist primarily of sequences of sandstone, shale, and limestone. The lowermost of these, the West Baden Group, ranges in thickness from 100 to 260 ft and is predominantly shale, mudstone, and sandstone, with lesser amounts of limestone. The West Baden Group is characterized by a predominantly clastic nature and by the irregularity of its limestone formations (Sullivan, 1972, p. 6; Shaver and others, 1986, p. 167).

Overlying the West Baden Group is the Stephensport Group, which consists of about equal parts of limestone, sandstone, and shale; it ranges in thickness from 130 to 230 ft. The formations of the Stephensport Group are, in ascending order, the Beech Creek Limestone, the Big Clifty Formation, the Haney Limestone, the Hardinsburg Formation, and the Glen Dean Limestone. The limestones in this group typically maintain their characteristic lithology throughout their outcrop areas, and they are more continuous and distinct than those in the underlying West Baden Group. The Beech Creek Limestone, at the base of the formation, is one of the best known, most widespread, and most reliable marker beds in rocks of Mississippian age in the Illinois Basin. In contrast, the clastic formations of the Stephensport Group commonly display abrupt lateral and vertical variations in lithology (Gray and others, 1957, p. 5-6; Gray and others, 1960, p. 41; Gray, 1979, p. K10; Shaver and others, 1986, p. 150-151).

The Buffalo Wallow Group overlies the Stephensport Group, and consists predominantly of shale, mudstone, and siltstone, with lesser amounts of sandstone and limestone. The Tar Springs Formation, at the base of the Buffalo Wallow Group, is probably the only formation in this group that is an outcrop or subcrop in the Patoka River basin. This formation is commonly about 65 ft thick and is primarily shale, although it contains local lenses of massive sandstone (Gray, 1978, p. 5; Shaver and others, 1986, p. 24).

Upper Mississippian rocks in the Patoka River basin are overlain unconformably by Lower Pennsylvanian rocks. The surface of this unconformity is a southwest-sloping plateau entrenched as much as 300 ft by southwest-trending relict stream valleys (Gray, 1979, p. K3).

Pennsylvanian rocks in Indiana consist of a dominantly clastic sequence of shale, siltstone, and sandstone, with intercalated thin but widespread beds of clay, coal, black shale, and limestone. Lateral facies changes are common and can be abrupt. Within short distances, individual beds or whole sequences can change in character so completely that they are scarcely traceable laterally (Gray and others, 1957, p. 5; Gray, 1979, p. K13). Pennsylvanian rocks compose the bedrock surface of the western three-quarters of the Patoka River basin (fig. 74), and include, in ascending order, the Raccoon Creek, Carbondale, and McLeansboro Groups.



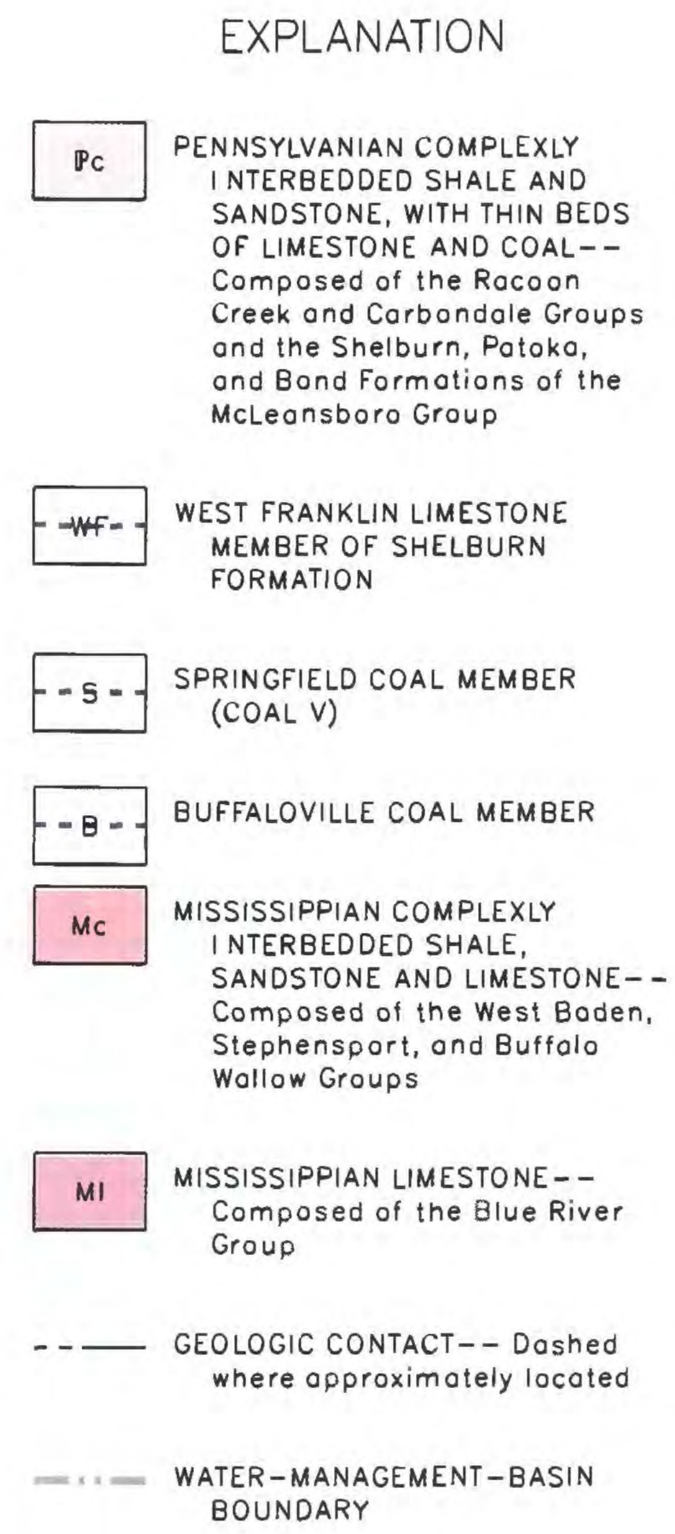


Figure 74. Bedrock geology of the Patoka River basin.

The Lower Pennsylvanian Raccoon Creek Group consists of more than 95 percent shale and sandstone and less than 5 percent clay, coal, and limestone. In ascending order, this group consists of the Mansfield, Brazil, and Staunton Formations. The Mansfield Formation is composed primarily of sandstone, shale, and mudstone, and it includes six named coal members. This formation is generally divided into an upper and lower unit; the upper unit is predominantly shale and mudstone, whereas the lower unit is mostly sandstone. The Mansfield Formation is roughly 200 to 400 ft thick in the Patoka River basin (Gray and others, 1960, p. 22-26; Gray, 1963, p. 11-12; Shaver and others, 1986, p. 86-87, 89, 120-121). The Brazil Formation, approximately 40 to 90 ft thick, consists of shale, sandstone, underclay, and coal. This formation is characterized by irregularities in both the thickness and lateral continuity of recognizable beds. The Brazil Formation contains the Buffaloville Coal Member, and the Upper Block Coal Member (Coal IV), a low-ash, low-sulfur coal. The Staunton Formation consists of 75 to 150 ft of sandstone and shale and as many as eight coal beds (Shaver and others, 1986, p. 21, 25, 150, 159; Weir, 1973, p. 18).

The Middle Pennsylvanian Carbondale Group includes, in ascending order, the Linton, Petersburg, and Dugger Formations. The Linton Formation consists of 40 to 160 ft of shale, sandstone, limestone, coal, and clay. At the base of the formation is the Coxville Sandstone Member, a 10- to 60 ft-thick fine- to coarse-grained, thickly bedded sandstone. The Petersburg Formation, 40 to 120 ft thick, includes three coal members, one limestone member, and unnamed beds of shale, sandstone, siltstone, and underclay. At the top of the formation is Indiana's most economically important coal, the Springfield Coal Member (Coal V) (Wier, 1973, p. 19, 27). The Dugger Formation includes 73 to 185 ft of sandstone, shale, limestone, coal, and clay. In the upper part of the Dugger Formation is the Anvil Rock Sandstone Member, a siltstone and sandstone that is present not only as a widespread sheet deposit but also as a channel-fill deposit. The Anvil Rock Sandstone Member can exceed 100 ft in thickness where the sheet phase overlies the channel-fill phase (Shaver and others, 1986, p. 6, 27, 32, 39, 80, 112, 149).

The Upper Pennsylvanian McLeansboro Group in the Patoka River basin consists of the Shelburn, Patoka, and Bond Formations, in ascending order. The Shelburn Formation ranges in thickness from 50 to 250 ft and is composed of shale, siltstone, and sandstone with lesser amounts of limestone, clay, and coal. At the base of this formation is the Busseron Sandstone Member, a fine- to medium-grained, massive sandstone, 48 to 77 ft thick, that fills erosional cutouts. More than 85 percent of the Patoka Formation is shale and sandstone; less than 15 percent is clay, limestone, and coal. Near the base of the formation is the fine-grained, 20-to 80-ft-thick Inglefield Sandstone Member (Shaver and others, 1986, p. 62, 85, 109, 142-43).

Unconsolidated Deposits

Unconsolidated deposits in the Patoka River basin are 0 to 150 ft thick (fig. 75). In the extreme western part of the basin, where the Patoka and Wabash River flood plains coincide, outwash sands and gravels overlain by alluvium commonly exceed 100 ft in thickness (Fidlar, 1948, p. 23-24, 43). Farther upstream, along the Patoka River, most unconsolidated deposits thin to 50 ft or less (fig. 75) and consist primarily of fine-grained alluvial materials of the Martinsville and Prospect Formations (Gray, 1963, p. 7-10).

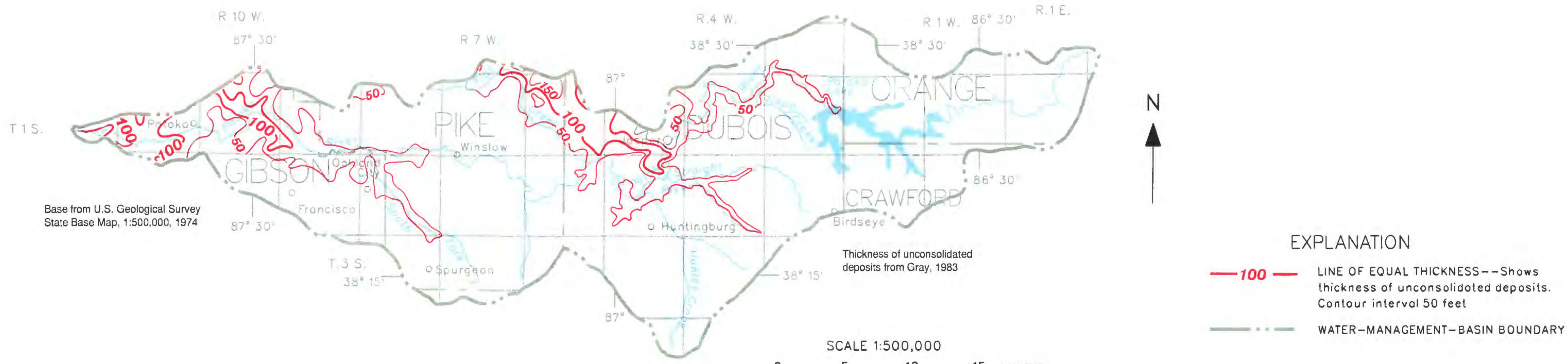


Figure 75. Thickness of unconsolidated deposits in the Patoka River basin.

In some northern parts of the basin in Gibson, Pike, and Dubois Counties, fine-grained lake (lacustrine) sediments, deposited in lakes of Illinoian and Wisconsinan ages, are 50 to more than 100 ft thick (Fidlar, 1948, p. 62-63; Gray and others, 1957, p. 7; Gray, 1963, p. 7-10). Illinoian till deposits are restricted to the northwestern part of the basin (fig. 73) and are generally less than 20 ft thick (Fuller and Clapp, 1904, p. 4).

Aquifer Types

Six hydrogeologic sections, totaling almost 100 mi in length, were constructed for the Patoka River basin (fig. 72). A total of 98 drillers' logs from water wells, coal-exploration test holes, and oil- and gas-exploration test holes, were used to construct the sections. Information from Wier and Stanley (1953), Friedman (1954), Hutchison (1964), Gray and others (1970), Gray and others (1987), and Keller (1990), aided interpretation of the drillers' logs. The average density of drillers' logs along the section lines is one per mile. Sections 11A–11A′ to 11E–11E′ (fig. 76) are oriented south-north, approximately perpendicular to the Patoka River, and are 14 to 16 mi apart. Section 11F–11F′ (fig. 76) is oriented west-east and is parallel to the long axis of the basin.

The different types of aquifers in the Patoka River basin include surficial sand and gravel deposits composed of outwash; discontinuous sand and gravel lenses interbedded with glacial, lake, and alluvial deposits; Upper Mississippian and Pennsylvanian sandstones; and Middle and Upper Mississippian carbonate rocks. Quantities of ground water sufficient to meet commercial and industrial needs can be obtained only from thick outwash sand and gravel deposits in the extreme western part of the basin. Yields from this aquifer can exceed 1,000 gal/min. The remaining aquifers in the Patoka River basin are generally capable of producing only enough water to meet domestic and livestock needs. Yields from these aquifers generally range from less than 1 to 25 gal/min and rarely exceed 100 gal/min. The four aquifer types in the Patoka River basin are summarized in table 13. The table includes range of aquifer thickness, range of reported yields, and common aquifer names used by other authors. The general areal extent of each aquifer type in the basin is shown in figure 77.

Unconsolidated and bedrock aquifers in the Patoka River basin are recharged primarily by direct infiltration and percolation of precipitation. Recharge may also occur as percolation from surface water in upland lakes and from mining impoundments. Flow in the shallow bedrock is probably complex and locally controlled; recharge occurs in upland areas, and ground water discharges to streams that flow on or near the bedrock surface (Martin and others, 1990, p. B22). Flow in deeper parts of the bedrock is primarily regional and probably follows the southwestward dip of the strata to points of discharge along the Wabash and Ohio Rivers.

Unconsolidated Aquifers

Unconsolidated aquifers in the Patoka River basin consist primarily of surficial sand and gravel, and discontinuous sand and gravel lenses. These aquifers are present in the western three-quarters of the basin (fig. 77). Unconsolidated deposits in the eastern one-quarter of the basin are relatively thin and impermeable and are not suitable sources of ground water (Gray and others, 1960, p. 22; Harrell, 1935, p. 162-163, 388).

Surficial Sand and Gravel Aquifers

Outwash sands and gravels are a major source of ground water along the lower reach of the Patoka River where the flood plains of the Wabash, White, and Patoka Rivers coalesce (Barnhart and Middleman, 1990, p. 9; Harrell, 1935, p. 215-218). The general areal extent of this aquifer type is shown as surficial sand and gravel aquifer in figure 77. Some wells completed in these deposits yield more than 1,000 gal/min (Barnhart and Middleman, 1990, p. 9).

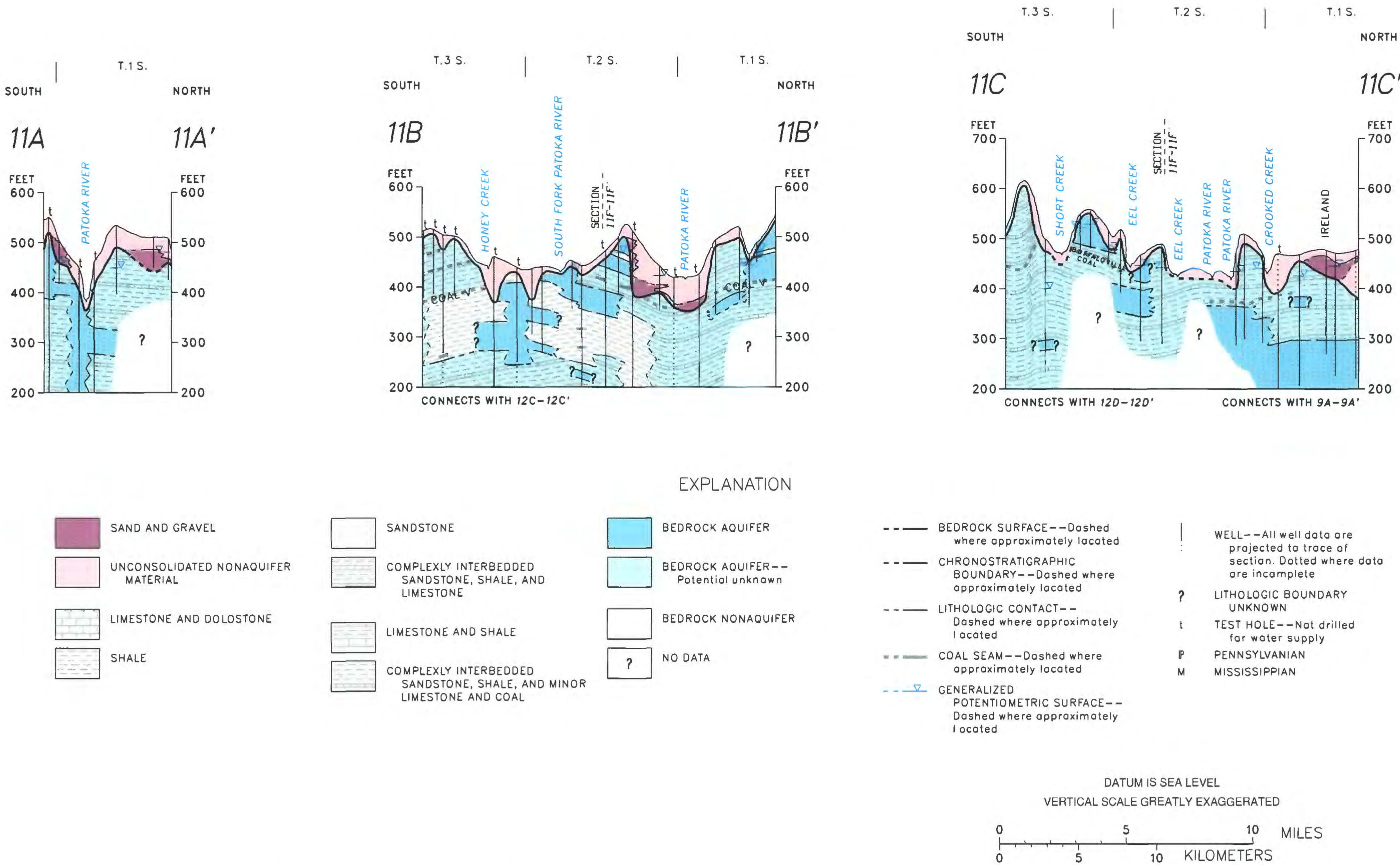


Figure 76. Hydrogeologic sections *11A–11A'* to *11F–11F'* of the Patoka River basin.

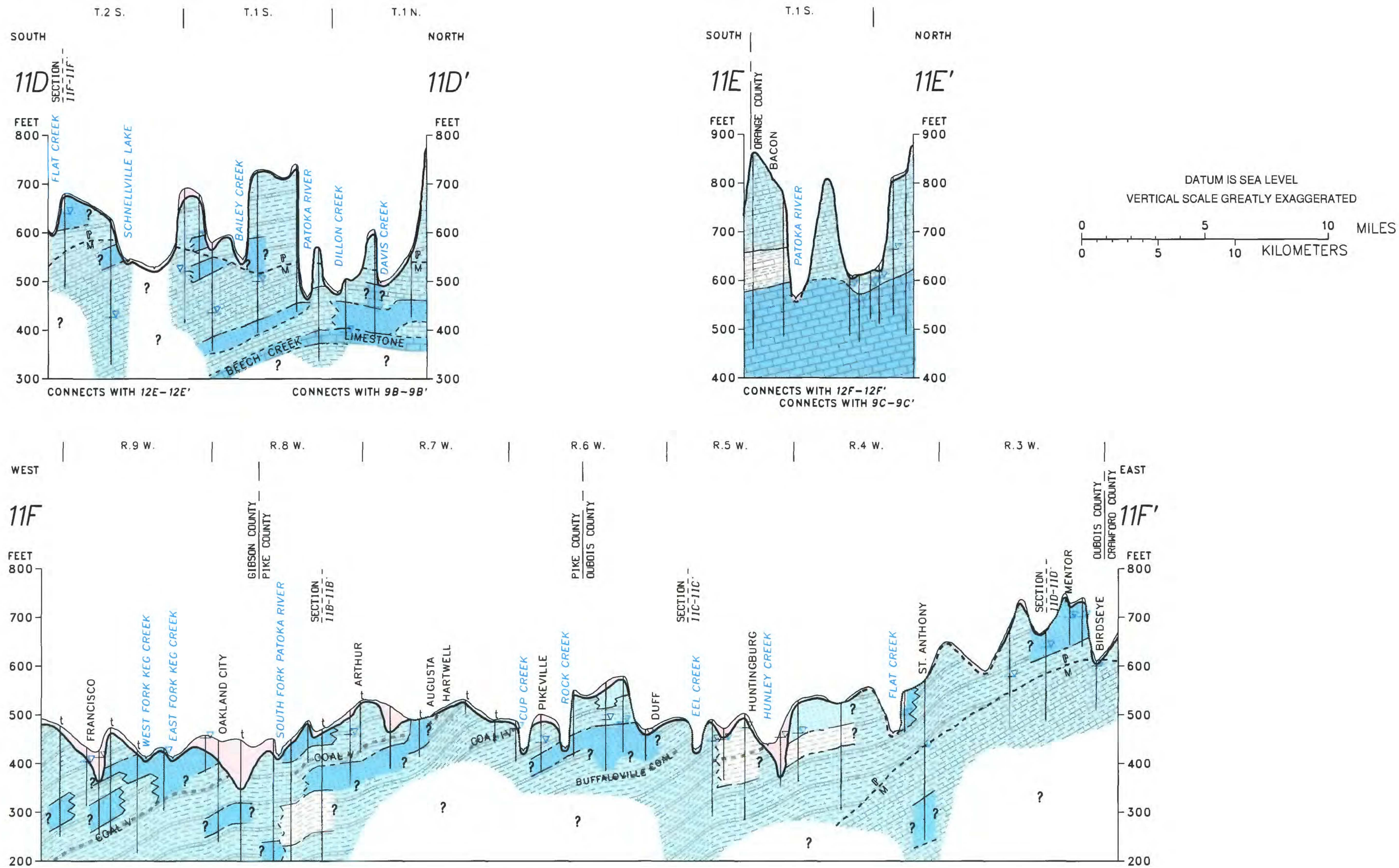


Figure 76. Hydrogeologic sections *11A–11A'* to *11F–11F'* of the Patoka River basin—Continued.

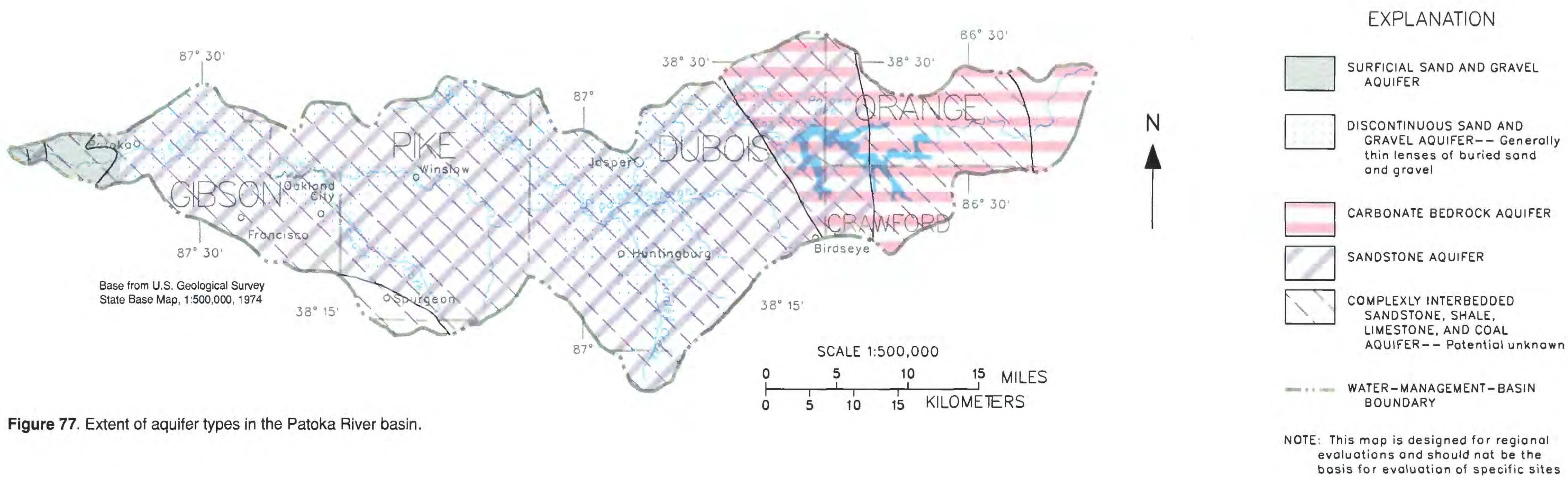


Figure 77. Extent of aquifer types in the Patoka River basin.

Discontinuous Sand and Gravel Aquifers

Another source of ground water from unconsolidated deposits is laterally discontinuous sand and gravel lenses interbedded with finer grained glacial, lake, and alluvial deposits (Barnhart and Middleman, 1990, p. 9). These deposits are found in the Patoka River and tributary valleys, other bedrock valleys, and in the northern part of the basin in Gibson, Pike, and Dubois Counties. The sand and gravel lenses are generally less than 30 ft thick but can be as much as 125 ft thick in places (Clark, 1980, p. 412; Barnhart and Middleman, 1990, p. 9). Wells in these deposits generally yield less than 100 gal/min; these deposits are mapped as discontinuous sand and gravel aquifer in figure 77. One well along the northern part of section 11A–11A′ (fig. 76) taps this type of aquifer. Other examples of this type of aquifer can be seen beneath the Patoka River valley in sections 11A–11A′ and 11B–11B′, and along the northern part of section 11C–11C′ (fig. 76).

Bedrock Aquifers

Carbonate Bedrock Aquifers

Middle and Upper Mississippian limestones are present in approximately the eastern one-third of the Patoka River basin and include limestone members of the Blue River, West Baden, and Stephensport Groups. These limestones are identified in figure 77 as the carbonate bedrock aquifer. Dissolution along joints, fractures, and bedding planes has generally increased the permeability of these rocks and is responsible for well yields of 5 to 10 gal/min (Harrel, 1935, p. 162-63; Jenkins, 1956, p. 13, 20; Gray and others, 1960, p. 43-44, 51; Gray and Powell, 1965, p.13-14; Bechert and Heckard, 1966, p. 108; Clark, 1980, p. 30). Mississippian carbonate bedrock aquifers are depicted in sections 11D–11D′ and 11E–11E′ (fig. 76).

The carbonate bedrock aquifers in section 11D–11D′ (fig. 76) are part of the Stephensport Group. Gray (1963, p. 13) and Gray and others (1960, p. 43-44, 48) indicate that the Beech Creek Limestone is the most extensive and highest yielding aquifer in the Stephensport Group. The wells along section 11E–11E′ (fig. 76) are completed in more than 50 ft of continuous limestone belonging to the Paoli Limestone and possibly the upper part of the Ste. Genevieve Limestone of the Blue River Group. Drillers' logs from four of the wells along this section indicate yields of 5 to 23 gal/min.

Sandstone Aquifers

Upper Mississippian and Pennsylvanian sandstone aquifers underlie virtually all but the eastern end of the Patoka River basin (fig. 77). With the exception of sand and gravel deposits, these sandstones are probably the primary source of ground water in the western two-thirds of the basin. In the Patoka River basin, yields of wells that tap these sandstones are generally low, ranging from less than 1 to 5 gal/min, with one notable exception; sandstones in the lower part of the Mansfield Formation, which are reputed to have the highest yields in the entire Pennsylvanian section (Gray and others, 1960, p. 29; Cable and Wolf, 1977, p. 9, 12-14), produce as much as 100 gal/min (Clark, 1980, p. 27-30). Wells along the northern one-half of section 11C–11C′ (fig. 76) tap thick sandstones that are probably the lower Mansfield Formation. Drillers' logs for three of these wells indicate yields that range from 7 to 20 gal/min—yields that are higher than those of most other wells that tap sandstones along other sections in the basin.

Most of the wells along sections 11B—11B′, 11C–11C′, and 11F–11F′ (fig. 76) tap Pennsylvanian sandstone aquifers. Some of these sandstones are fluvial (also known as channel-fill) deposits, which are relatively thick, narrow deposits that fill erosional

Table 13. Characteristics of aquifer types in the Patoka River basin
[<, less than; >, greater than; locations of aquifer types shown in fig. 77]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Surficial sand and gravel	30-<150	[1,2]25->1,000	Valley train[1]; flood plain deposits[3]; alluvium and outwash[4]; outwash sand and gravel[2]
Discontinuous sand and gravel	<30- 125	[1,2]5- 250	Sand and gravel lenses, outwash plain deposits[1]; sand and gravel deposits[2]
Carbonate bedrock			
Middle Mississippian	>100	[2,5]5- <100	Blue River Group[6]
Upper Mississippian	8- 40	[2]1- 10	Beech Creek Limestone[6]
	20- 40	No data	Haney Limestone[6]
	9- 30	No data	Glen Dean Limestone[6]
Sandstone			
Upper Mississippian	30- 70	[2]<1- 5	Big Clifty Formation[6]
	20- >60	[2]<1- 5	Hardinsburg Formation[6]
Pennsylvanian	10- 150	[2]<1- 100	Mansfield Formation[2,3,5,7]
	10- 60	[2]<1- 10	Coxville Sandstone Member[6] or Linton Aquifer[7]
	10- 160	[2]<1- 10	Anvil Rock Sandstone Member[6]
	20- 90	[1,2]<1- 10	Busseron Sandstone Member[1]
	20- 40	[1,2]3- 300	Inglefield Aquifer[1] or Inglefield Sandstone Member[2]

[1]Barnhart and Middleman, 1990.
[2]Clark, 1980.
[3]Harrell, 1935.
[4]Fidlar, 1948.
[5]Bechert and Heckard, 1966.
[6]Shaver and others, 1986.
[7]Cable and Wolf, 1977.

cutouts in underlying strata. Wells in channel sandstones may have higher yields than wells in sheet sandstones because of the greater thickness and higher permeability of channel sandstones. Their higher permeability is attributed to generally larger grain size, better sorting, and smaller amounts of clay compared to sheet sandstones (Hopkins, 1958, p. 12; Cable and Wolf, 1977, p. 9; Martin and others, 1990, p. B22). The thick sandstone sequence shown in section 11A–11A′ (fig. 76) probably consists of several stacked channel-fill deposits and includes, at least in part, the Anvil Rock Sandstone Member of the Dugger Formation. Another channel sandstone is evident on section 11B–11B′ (fig. 76) between Honey Creek and South Fork Patoka River.

Complexly Interbedded Sandstone, Shale, Limestone, and Coal

In areas where unconsolidated aquifers are absent, and where sandstone and limestone aquifers are either unproductive or too deep for the economical installation of wells, ground water can be obtained from coal seams and(or) complexly interbedded sequences of sandstone, shale, limestone, and coal. Banaszak (1980, p. 235-240) indicates that coal seams hydraulically connected to a source of recharge and underlain by a confining unit can yield as much as 10 gal/min. Only 8 of the well logs used to construct sections in the Patoka River basin indicate that the wells obtain water primarily from coal seams and(or) complexly interbedded lithologies; three of the logs indicate well yields of less than 1 gal/min. In contrast, a fourth well log, plotted on section 11B–11B′ (fig. 76) as the first water-supply well south of the Patoka River, indicated a yield of 12 gal/min. The source of ground water for this well is strata immediately downdip of a bedrock surface overlain by sand and gravel. Basal sands and gravels probably increase recharge to the underlying bedrock and increase net storage in the immediate area, resulting in anomalously high yields. With local exceptions such as this, coal seams and(or) complexly interbedded lithologies are not generally a primary source of ground water in the Patoka River basin, and these materials are shown as "aquifer—potential unknown" on all the hydrogeologic sections.

Summary

The Patoka River basin encompasses 862 mi^2 in southwestern Indiana. Major sources of ground water in the Patoka River basin include surficial sand and gravel composed of outwash, discontinuous sand and gravel, Mississippian carbonate rocks, and Mississippian and Pennsylvanian sandstones.

Outwash sand and gravel deposits can produce in excess of 1,000 gal/min but they are present only in the extreme western part of the basin. Discontinuous sand and gravel interbedded with fine-grained material are present in the western three-quarters of the basin along the Patoka River and tributary valleys and in areas once covered by lakes during Illinoian and Wisconsinan glaciations. Well yields as much as 250 gal/min are available from these deposits.

Sandstone and carbonate bedrock aquifers in the basin are generally capable of producing only enough water to meet domestic and livestock needs. Sandstone aquifers underlie most of basin, whereas carbonate bedrock aquifers, composed primarily of limestone, are present only in the eastern one-third. Well yields from the sandstone aquifers are generally less than 1 to 5 gal/min but seem to be somewhat higher in the lower Mansfield Formation. Wells in the carbonate bedrock aquifers typically yield 5 to 10 gal/min. Recharge to bedrock aquifers and net storage may be increased in areas where sand and gravel directly overly the bedrock surface. Bedrock wells downdip from these areas may have higher yields. Coal seams and complexly interbedded sequences of sandstone, shale, limestone, and coal are infrequently a primary source of ground water, and their potential as aquifers can be adequately evaluated only on a local basis.

References Cited

Arvin, D.V., 1989, Statistical summary of streamflow data for Indiana: U.S. Geological Survey Open-File Report 89-62, 964 p.

Banaszak, K.J., 1980, Coal as aquifers in the Eastern United States, *in* Proceedings of the Symposium on Surface Mining Hydrology, Sedimentology, and Reclamation: University of Kentucky, Lexington, Kentucky, December 1-5, 1980, p. 235-241.

_____1985, Potential effects on ground water of a hypothetical surface mine in Indiana: Ground Water Monitoring Review, v. 5, no. 1, p. 51-57.

Barnhart, J.R., and Middleman, B.H., 1990, Hydrogeology of Gibson County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 41, 17 p.

Bechert, C.H., and Heckard, J.M., 1966, Ground water, *in* Lindsey, A.A., ed. Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 100-115.

Cable, L.W., and Wolf, R.J., 1977, Ground-water resources of Vanderburgh County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 38, 37 p.

Clark, G.D. ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resources Study Commission, Indiana Department of Natural Resources, 508 p.

Corbett, D.M., 1965, Water supplied by coal surface mines, Pike County, Indiana: Indiana University, Water Resources Research Center Report of Investigations 1, 67 p.

_____1968, Ground-water hydrology pertaining to surface mining for coal, southwestern Indiana, *in* Coal Mining Drainage Research Symposium, Second Proceedings, p. 164-189.

Fidlar, M.M., 1948, Physiography of the lower Wabash valley: Indiana Department of Conservation, Division of Geology Bulletin 2, 112 p.

Friedman, S.A., 1954, Distribution, structure, and mined areas of coals in Gibson County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 4, scale 1 inch equals approximately 2 miles.

Fuller, M.I., and Ashley, A.H., 1902, Ditney folio of the Geologic Atlas of the United States: U.S. Geological Survey, 16 p.

Fuller, M.I., and Clapp, F.G., 1904, Patoka folio of the Geologic Atlas of the United States: U.S. Geological Survey, 17 p.

Geosciences Research Associates, Inc., 1982, Hydrogeologic atlas of Indiana: Bloomington, Ind., 31 pl.

Gray, H.H., 1963, Geology of the upper Patoka drainage basin: Indiana Department of Conservation, Geological Survey Special Report 2, 23 p.

_____1978, Buffalo Wallow Group, Upper Chesterian (Mississippian), of southern Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 25, 28 p.

_____1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Indiana: U.S. Geological Survey Professional Paper 1110-K, 20 p.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 37, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Gray, H.H., Dawson, T.A., McGregor, D.J., Perry, T.G., and Wayne, W.J., 1957, Rocks associated with the Mississippian-Pennsylvanian unconformity in southwestern Indiana: Indiana Department of Conservation, Geological Survey Field Conference Guidebook 9, 42 p.

Gray, H.H., Jenkins, R.D., and Weidman, R.M., 1960, Geology of the Huron area, south-central Indiana: Indiana Department of Conservation, Geological Survey Bulletin 20, 78 p.

Gray, H.H., and Powell, R.L., 1965, Geomorphology and groundwater hydrology of the Mitchell Plain and Crawford Upland in southern Indiana: Indiana Department of Natural Resources, Geological Survey Field Conference Guidebook 11, 26 p.

Gray, H.H., Wayne, W.J., and Wier, C.E., 1970, Geologic map of the 1^{o} x 2^{o} Vincennes quadrangle and parts of adjoining quadrangles, Indiana and Illinois, showing bedrock and unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Regional Geologic Map 3, scale 1:250,000.

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Hoggatt, R.E., 1975, Drainage areas of Indiana streams: U.S. Geological Survey, 231 p.

Hopkins, M.E., 1958, Geology and petrology of the Anvil Rock Sandtone of Southern Illinois: Illinois Geological Survey Circular 256, 49 p.

Hutchison, H.C., 1964, Distribution, structure, and mined areas of coals in Dubois County, Indiana: Indiana Department of Conservation Preliminary Coal Map 10, scale 1 inch equals approximately 2 miles.

Jenkins, R.D., 1956, The geology of the Hillham Quadrangle, Martin, Orange, and Dubois Counties: Bloomington, Ind., Indiana University, unpublished Master's thesis, 52 p.

Keller, S.J., 1990, Maps of southwestern Indiana showing geology and elevation of the sub-Pennsylvanian surface: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 51, scale 1:380,160.

Malott, C.A., 1922, The physiography of Indiana, *in* Logan, W.N., and others, Handbook of Indiana Geology: Indiana Department of Conservation Publication 20, part 2, p. 59-256.

Martin, J.D., and Crawford, C.G., 1987, Statistical analysis of surface-water quality in and near the coal-mining region of southwestern Indiana, 1957-80: U.S. Geological Survey Water-Supply Paper 2291, 92 p.

Martin, J.D., Duwelius, R.F., and Crawford, C.G., 1990, Effects of coal mining and reclamation on the geohydrology of six small watersheds in west-central Indiana: U.S. Geological Survey Water-Supply Paper 2368-B, 71 p.

Renn, D.E., 1989, Streamflow and stream quality in the coal-mining region, Patoka River basin, southwestern Indiana, 1983-85: U.S. Geological Survey Water-Resources Investigations Report 88-4150, 68 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed. Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.D.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

Sullivan, D.M., 1972, Subsurface stratigraphy of the West Baden Group in Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 47, 31 p.

Thornbury, W.D., 1950, Glacial sluiceways and lacustrine plains of southern Indiana: Indiana Department of Conservation, Division of Geology Bulletin 4, 21 p.

Wangsness, D.J., Miller, R.L., Bailey, Z.C., and Crawford, C.G., 1981, Hydrology of area 32, eastern region, Interior Coal Province, Indiana: U.S. Geological Survey Water-Resources Investigations/Open-File Report 81-498, 76 p.

Watkins, F.A., Jr., 1963, Wabash basin comprehensive study, ground water appraisal of the Patoka River basin and Patoka reservoir site: U.S. Geological Survey Open-File Report, 4 p.

Wier, C.E., 1973, Coal resources of Indiana: Indiana Department of Natural Resources, Geological Survey Bulletin 42-I, 40 p.

Wier, C.E., and Stanley, J.T., 1953, Distribution, structure and mined areas of coals in Pike County, Indiana: Indiana Department of Conservation, Geological Survey Preliminary Coal Map 3, scale 1 inch equals approximately 2 miles.

OHIO RIVER BASIN

By M. Catharine Woodfield and
Joseph M. Fenelon

General Description

The Ohio River basin is the southernmost water-management basin in Indiana. It extends approximately 200 mi across southern Indiana, from Lawrenceburg in eastern Indiana to about 10 mi southwest of Mt. Vernon in western Indiana (fig. 78). The Ohio River basin, the fourth largest basin in the State, encompasses 4,224 mi^2. The basin includes all of Ohio, Switzerland, Floyd, Harrison, and Perry Counties and large parts of Dearborn, Ripley, Jefferson, Clark, Washington, Crawford, Spencer, Warrick, and Vanderburgh Counties. Principal cities within the basin include Evansville, New Albany, Madison, Lawrenceburg, Jeffersonville, Mt. Vernon, Salem, Boonville, Tell City, and Charlestown.

Previous Studies

Pettijohn and Reussow (1969) authored a study of ground-water resources in the Ohio River basin. Their report defined the geohydrology of the principal water-bearing units, well yields to streams, and the ground-water quality within the water-bearing units.

Ground-water resources were assessed in comprehensive reports on Posey County (Robison, 1977), Vanderburgh County (Cable and Wolf, 1977), and Gibson County (Barnhart and Middleman, 1990). These county reports include well yields and thicknesses for bedrock and unconsolidated aquifers, and maps of the surfaces of some of the sandstone aquifers. Results of water-quality analyses are also reported and discussed. Gallaher and Price (1966) wrote a comprehensive report of the hydrogeology of the alluvial deposits within the Ohio River Valley in Kentucky and the availability and quality of the ground water within these deposits. Price (1964a, 1964b, 1964c) prepared hydrogeologic sections and mapped the hydrology and geology of the alluvial deposits along the Ohio River in Kentucky. Harvey (1956) described the geology and ground-water resources of the Henderson, Ky., area along the Ohio River. He included chemical composition, yield, and recharge and discharge areas for selected geologic formations.

Several general references on ground-water hydrology of Indiana include the Ohio River basin. Clark (1980) described the various types of aquifers and their potential yields. Bechert and Heckard (1966) delineated ground-water provinces on the basis of well yields and the sources of ground water. Ground-water resources of each county in Indiana were described by Harrell (1935).

Physiography

The physiography of the Ohio River basin is primarily bedrock controlled but was affected by the relocation of the pre-Pleistocene Teays River drainage into the Ohio River valley during the Pleistocene. The basin is characterized by considerable relief and a wide variety of topographic features that are the result of weathering, stream erosion, and mass movement. During Pleistocene time, the western two-thirds of the Ohio River basin was not glaciated; thus, the topography strongly reflects bedrock control. The eastern one-third of the basin was glaciated, but glacial deposits are not thick enough to obscure the bedrock relief.

The seven physiographic regions in the Ohio River basin trend north-northwest in conformity with the strike of the bedrock (fig. 79). East to west, these regions are the Dearborn Upland, Muscatatuck Regional Slope, Scottsburg Lowland, Norman Upland, Mitchell Plain, Crawford Upland, and Wabash Lowland (Schneider, 1966, p. 42).

The Dearborn Upland is a dissected plateau underlain by flat-lying limestones and shales. Streams that originate in this area are short; they have steep slopes and are entrenched as much as 450 ft in the upland plain (Schneider, 1966, p. 43). Parts of the plateau are thoroughly dissected; however, broad plains of virtually unmodified land remain well above the dissecting streams. The western boundary of the Dearborn Upland is defined by an escarpment that marks a major drainage divide between the generally west-southwest-flowing streams of the East Fork White River and Muscatatuck River drainage basins and the south and southeast-flowing streams of Indian-Kentuck Creek and Laughery Creek drainage basins.

A structurally controlled area, referred to as the Muscatatuck Regional Slope, lies west of the Dearborn Upland. Resistant, westward-dipping Silurian and Devonian carbonate rocks underlie this gently sloping plain. Near the Ohio River, the plain slopes westward from 875 ft to about 500 ft above sea level (Malott, 1922, p. 87).

The Muscatatuck Regional Slope subtly grades westward into the Scottsburg Lowland. The Scottsburg Lowland is an asymmetric trough underlain by New Albany Shale. The Lowland attains an elevation of 500 ft above sea level near the Ohio River and is primarily an expansive valley of little relief (less than 75 ft) (Schneider, 1966, p. 45). The western boundary of the Scottsburg Lowland lies at the base of a 400 to 600-ft high scarp, known as the Knobstone Escarpment (R. 6 E. in section 12L–12L′, fig. 82). The escarpment, one of the most prominent topographic features in Indiana, marks the boundary between the pre-Wisconsinan drift and the driftless area.

The Norman Upland, a dissected plateau above the Ohio River underlain by siltstones interbedded with soft shales, lies west of the escarpment (Schneider, 1966, p. 46). The area is characterized by flat-topped narrow divides and steep, V-shaped valleys. The western boundary of the Norman Upland extends along a line between deeply dissected clastic rocks and a limestone valley plain called the Mitchell Plain (Malott, 1922, p. 93-94).

The Mitchell Plain is formed on the dip slope of a thick sequence of limestones. The area has low relief and is marked by sinkholes and other solution features. In a few places, the Mitchell Plain is crossed by deeply entrenched streams, but much of the drainage is underground. Streams generally traverse the surface for only a short distance before disappearing into sinkholes.

The Crawford Upland, west of the Mitchell Plain, is characterized by a considerable increase in altitude and by a great diversity of relief and landforms. The relief is due to differential erosion of alternating beds of sandstone, shale, and limestone (Schneider, 1966, p. 48). Abundant stream valleys in the upland form a mature drainage system.

The Wabash Lowland is the farthest west of the physiographic regions. It is underlain primarily by sandstone and shale of Pennsylvanian age and is characterized by broad aggraded valleys and rounded hills. Thick lake (lacustrine) and river (alluvial) sediments and outwash deposits overlie the bedrock in places. Along the Ohio River are broad, terraced valley-fill sediments.

One of the most important hydrogeologic and physiographic features of the Ohio River basin is the Ohio River valley. It is widest in the Wabash Lowland and extends along the entire southern boundary of the basin. The width of the Ohio River valley in Indiana ranges from a few hundred feet near Leavenworth to about 6 mi near Evansville (Pettijohn and Reussow, 1969, p. 4). The Ohio River flows in only a small part of this old glacial drainageway. The drainageway is a deeply entrenched U-shaped valley that is very shallow in most places relative to its width. Tributary streams drain the valley floor, depositing fine-grained alluvium over coarse-grained outwash deposits (Pettijohn and Reussow, 1969, p. 10).

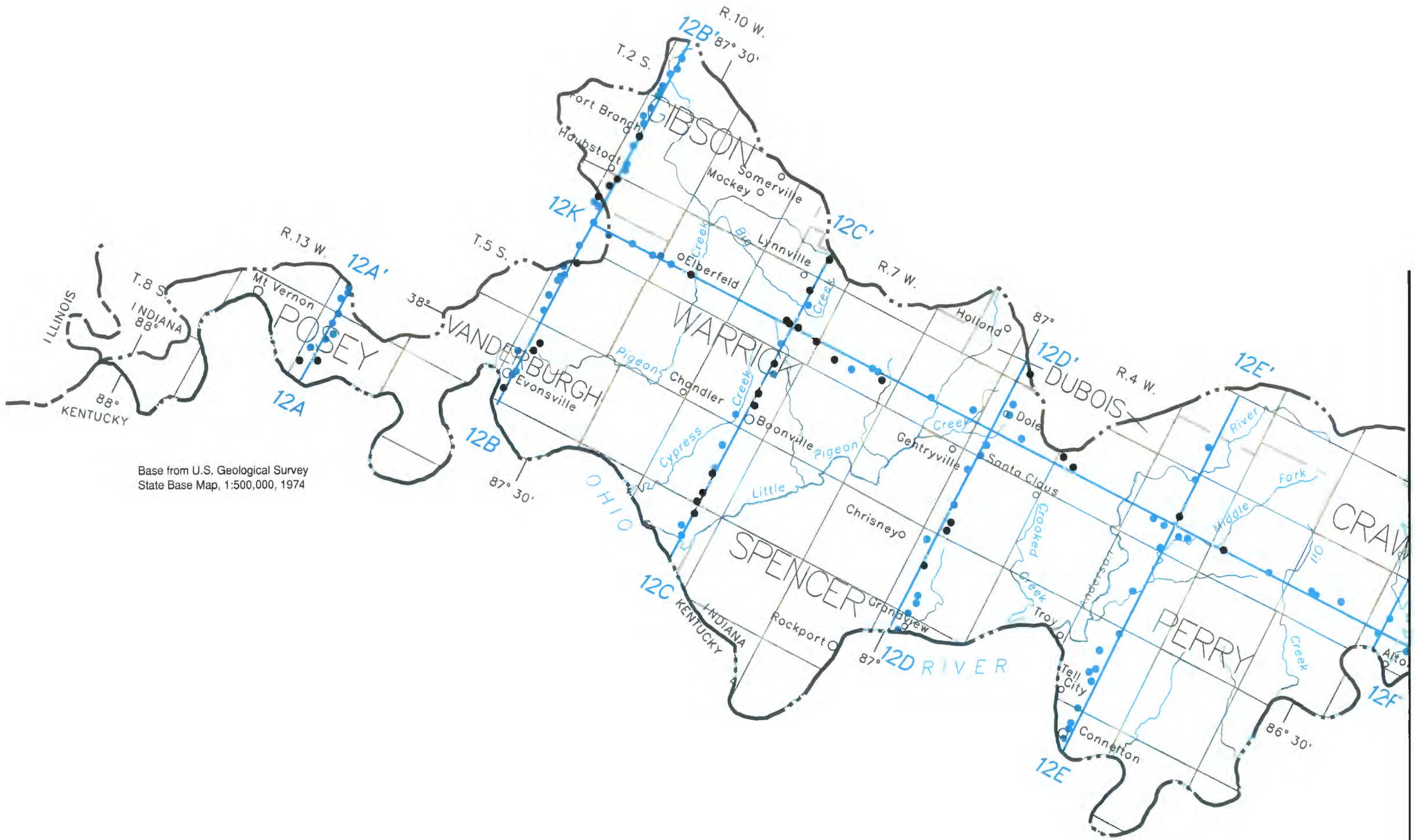
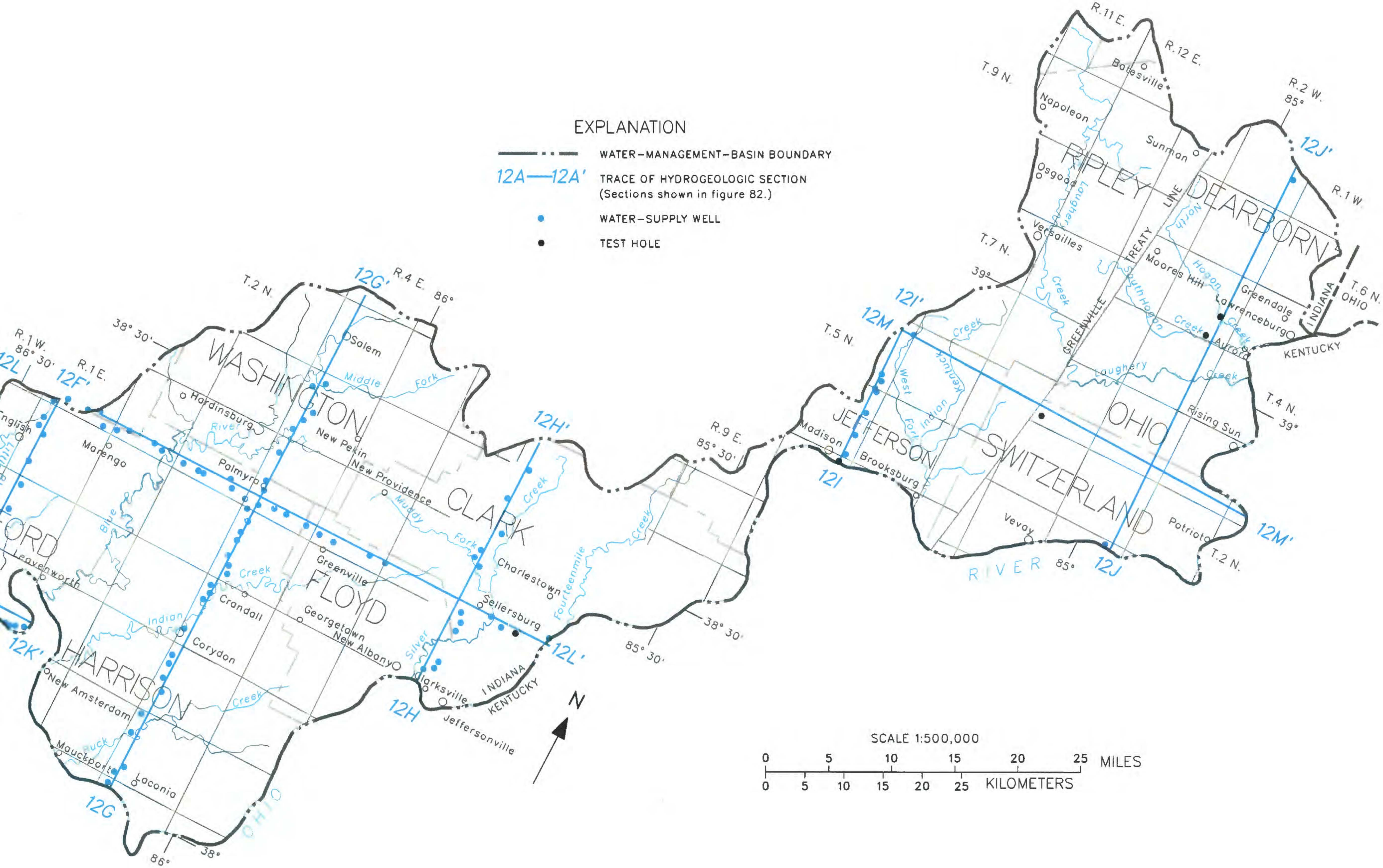


Figure 78. Location of section lines and wells plotted in the Ohio River basin.


EXPLANATION
WATER-MANAGEMENT-BASIN BOUNDARY
12A—12A'
TRACE OF HYDROGEOLOGIC SECTION
(Sections shown in figure 82.)
WATER-SUPPLY WELL
TEST HOLE
SCALE 1:500,000
0 5 10 15 20 25 MILES
0 5 10 15 20 25 KILOMETERS
N
WASHINGTON
CLARK
FLOYD
HARRISON
JEFFERSON
SWITZERLAND
OHIO
RIPLEY
DEARBORN
Salem
Hardinsburg
New Pekin
New Providence
Palmyra
Marengo
English
Leavenworth
Greenville
Crandall
Georgetown
New Albany
Corydon
New Amsterdam
Mauckport
Laconia
Charlestown
Sellersburg
Clarksville
Jeffersonville
Madison
Brooksburg
Vevay
Patriot
Rising Sun
Aurora
Lawrenceburg
Greendale
Moores Hill
Versailles
Osgood
Napoleon
Batesville
Sunman
INDIANA
KENTUCKY
OHIO
RIVER
GREENVILLE
TREATY
LINE
12F'
12G'
12G
12H'
12H
12I'
12I
12J'
12J
12K'
12L
12L'
12M
12M'
T.2 N.
T.5 N.
T.7 N.
T.9 N.
T.6 N.
T.4 N.
R.1 W.
R.1 E.
R.4 E.
R.9 E.
R.11 E.
R.12 E.
R.2 W.
38° 30'
86° 30'
86°
85° 30'
85°
39°
38°

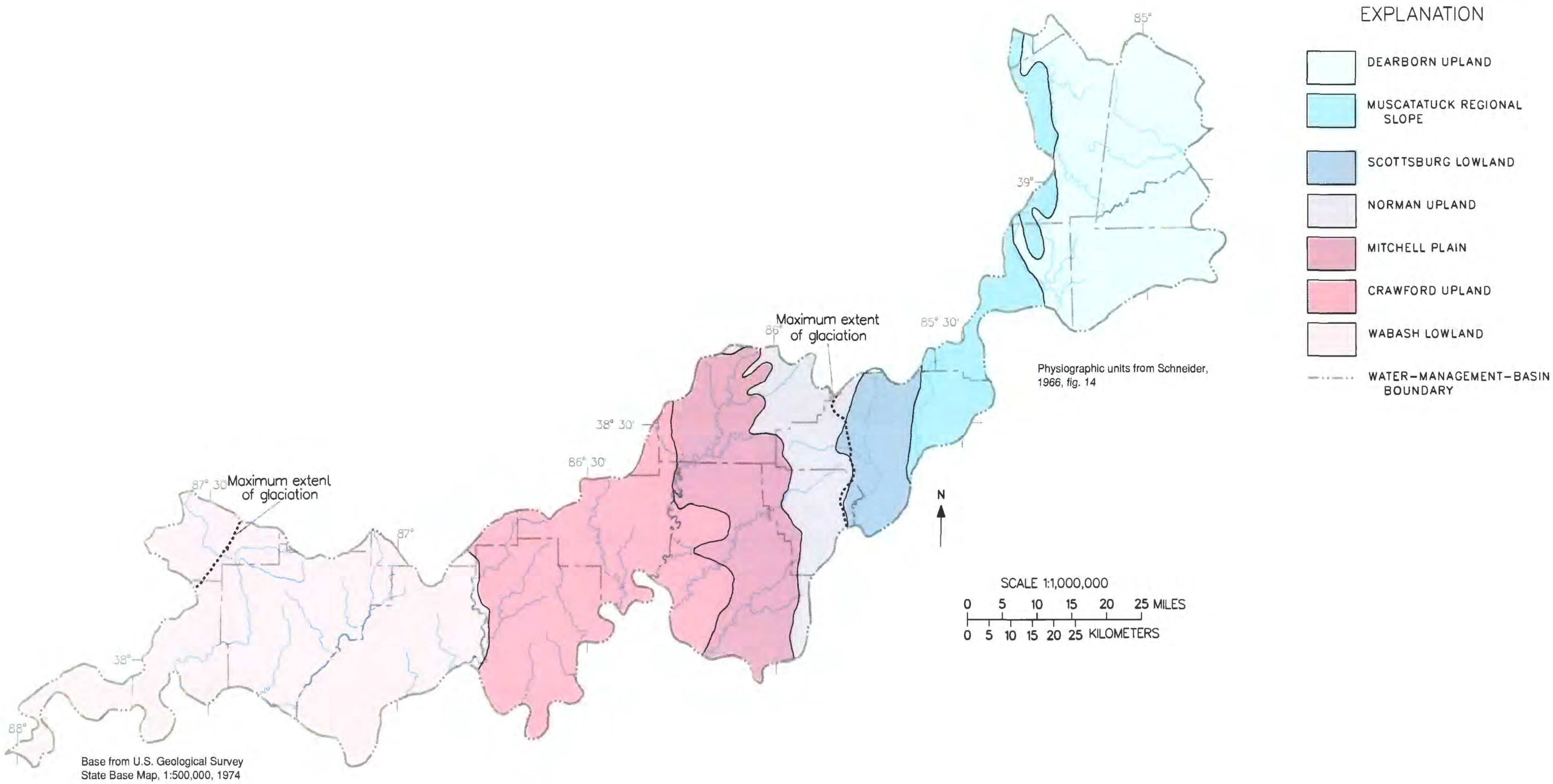


Figure 79. Physiographic units and extent of glaciation in the Ohio River basin.

Surface-Water Hydrology

The Ohio River is a major eastern tributary of the Mississippi River and forms the entire State boundary between Indiana and Kentucky. The total drainage area of Indiana streams that discharge into the Ohio River, excluding the Wabash and Whitewater River basins, is 4,224 mi^2. The Lower Ohio River was the name given to the Indiana part of the Ohio River by the State of Indiana in the State Water Plan (Pettijohn and Reussow, 1969, p. 4), but for the purpose of this report, it will be referred to as the Ohio River.

Meandering along Indiana's southern border for 357 mi, the Ohio River flows in a general southwesterly direction. River segments upstream of Cannelton follow bedrock fractures through many angled turns. The river descends 135 ft in elevation from the eastern side of the basin to the western side, resulting in an average gradient of 0.38 ft/mi. The flow of the river is affected by numerous dams operated by the U.S. Army Corps of Engineers.

Tributary streams with more than 100 mi^2 of drainage area that flow into the Ohio River are from east to west, Hogan Creek, which includes the North and South Forks (128 mi^2); Laughery Creek (343 mi^2); Indian-Kentuck Creek (153 mi^2); Fourteenmile Creek (101 mi^2); Silver Creek (219 mi^2); Indian Creek (257 mi^2); Blue River (524 mi^2); Little Blue River (172 mi^2); Anderson River (258 mi^2); Little Pigeon Creek (360 mi^2); and Pigeon Creek (368 mi^2). Some of the streams (including South Hogan Creek, Indian-Kentuck Creek, Silver Creek, and Middle Fork of Anderson River) have gone dry at times during most years of record (Arvin, 1989).

All of the tributary streams of the Ohio River in Indiana originate within 30 mi of the Ohio River, with the exception of the Blue River, which originates approximately 52 mi from the Ohio River (Malott, 1922, p. 75). Just west of Madison, the drainage divide between the Ohio River and the East Fork White River is within 2 mi of the Ohio River.

Geology

Bedrock Deposits

The major geologic structures within the Ohio River basin are the Cincinnati Arch and the Illinois Basin (fig. 4). The crest of the Cincinnati Arch trends northwest and crosses the eastern part of the Ohio River basin. The bedrock at the crest of the arch is nearly horizontal. West and southwest of the arch crest, the rocks dip at an average of 25 ft/mi into the Illinois Basin (Gutschick, 1966, p. 10).

The Paleozoic carbonate rocks that underlie the Ohio River basin range in age from Cambrian to Pennsylvanian (fig. 5). The deeply buried Cambrian rocks are composed of a sequence of sandstone, shale, limestone, and dolomite. Several thousand feet of buried Ordovician rocks overlie the Cambrian rocks (Gray and others, 1987). The Ordovician rocks are predominantly composed of dolomite, limestone, and some sandstone. Ground water is not withdrawn from these rocks because of their depth and because the water has high concentrations of dissolved solids.

The buried Ordovician rocks are overlain by the Ordovician Lexington Limestone, the oldest exposed rock unit in the basin. This fossiliferous limestone attains a maximum thickness of 256 ft in southeastern Indiana (Shaver and others, 1986, p. 78). The upper part of the Lexington Limestone is exposed at the bedrock surface in discontinuous stretches along the banks of the Ohio River in eastern Switzerland County (sections 12J–12J′ and 12M–12M′, fig. 82).

The Ordovician Maquoketa Group, consisting of the Kope, Dillsboro, and Whitewater Formations (fig. 80) overlies the Lexington Limestone. The Kope Formation is a 300- to 400-ft-thick sequence, 95 percent of which is shale (Shaver and others, 1986, p. 72). The formation is exposed mainly along the lower reaches of the deeply entrenched streams of North and South Hogan Creeks, Laughery Creek, Indian-Kentuck Creek, and along other smaller tributaries to the Ohio River. A transitional contact between the Kope Formation and the overlying Dillsboro Formation is marked by an upward increase in the proportion of limestone (Gray, 1972, p. 14). The Dillsboro Formation is a 400-ft-thick sequence consisting of about 30 percent fossiliferous and clayey limestones and 70 percent calcareous shale (Shaver and others, 1986, p. 37). It is conformably overlain by the Whitewater Formation, which is composed of less than 60 to 100 ft of bluish-gray limestone interbedded with calcareous shale (Shaver and others, 1986, p. 168). The Whitewater Formation is exposed in a north-northeastern band along the eastern margin of the basin and in the entrenched headwaters of Laughery and Indian Creeks.

The Whitewater Formation is overlain unconformably by the Silurian Brassfield Limestone, Salamonie Dolomite, Waldron Shale, and Louisville Limestone (fig. 80). The Brassfield Limestone consists of 4 to 20 ft of medium- to coarse-grained fossiliferous limestone containing minor amounts of shale (Shaver and others, 1986, p. 20). The Salamonie Dolomite overlies the Brassfield Limestone and consists predominantly of limestone and dolomite with minor amounts of shale and chert. The Salamonie Dolomite is 0 to 60 ft thick along the eroded edges of the outcrop area in southeastern Indiana. The Waldron Shale overlies the Salamonie Dolomite and is composed of about 5 ft of shale that contains siltstone and fossiliferous limestone beds, which are reeflike in many places. The Louisville Limestone, a fine-grained, thick-bedded, clayey and dolomitic limestone that is approximately 40 to 75 ft thick overlies the Waldron Shale (Shaver and others, 1986, p. 83). Outcrops of the Louisville Limestone in the southeastern part of the basin are overlain unconformably by the Devonian Muscatatuck Group (fig. 80).

The Muscatatuck Group is composed predominantly of fine-grained to granular dolomite and limestone. The carbonate rocks range from pure to sandy or shaly. The lower part of the Muscatatuck Group contains some anhydrite and gypsum (Shaver and others, 1986, p. 99). The Muscatatuck Group is exposed on the western side of the Cincinnati Arch and is zero to more than 250 ft thick in the basin.

Overlying the Muscatatuck Group is the Devonian and Mississippian New Albany Shale, composed primarily of brownish-black, carbon-rich shale and greenish-gray shale. The New Albany Shale is about 100 ft thick near its outcrop area in Clark County.

Rocks of Mississippian age in the Ohio River basin include the Borden, Sanders, Blue River, West Baden, Stephensport, and Buffalo Wallow Groups (fig. 80). The Borden Group overlies the New Albany Shale and is composed primarily of gray clayey siltstone and shale with fine sandstone. Sparse interbedded limestones form discontinuous lenses. This group, extending from the Ohio River in southern Floyd County northward, is about 500 ft thick in the basin, and has an outcrop width of about 6 to 12 mi within the east-central part of the basin.

Carbonate rocks of the Sanders and Blue River Groups overlie and crop out west of the Borden Group. A complex relation among a variety of carbonate rocks exists within the Sanders Group (Shaver and others, 1986, p. 136). At the base is a mixture of fine-grained dolomite and limestone with chert. Above these rocks lies a sequence of well-cemented bioclastic limestones and dolomites. The thickness of the Sanders Group is about 120 ft in the outcrop area. The Blue River Group is composed primarily of carbonate rocks but contains significant amounts of gypsum, anhydrite, shale, chert and calcareous sandstone (Shaver and others, 1986, p. 16). The Blue River Group is about 540 ft thick at outcrops in southern Crawford and Harrison Counties.

Complexly interbedded rocks of the West Baden, Stepensport, and Buffalo Wallow Groups overlie the Blue River Group. The West Baden Group is composed predominantly of shale, mudstone, thin-bedded to cross-bedded sandstone, and limestone (Shaver and others, 1986, p. 16). Along the outcrop area in the central part of the basin, the thickness ranges from 100 to 140 ft. The Stephensport Group, which is 130 to 230 ft thick, is composed of approximately equal parts of limestone, shale, and sandstone (Shaver and others, 1986, p. 150). The Buffalo Wallow Group consists of a shale, mudstone, and siltstone sequence with thin, laterally extensive beds of limestone and sandstone (Shaver and others, 1986, p. 24). Its maximum exposed thickness is 270 ft near the Ohio River. The Buffalo Wallow Group does not crop out north of southwestern Orange County because it was truncated by pre-Pennsylvanian erosion.

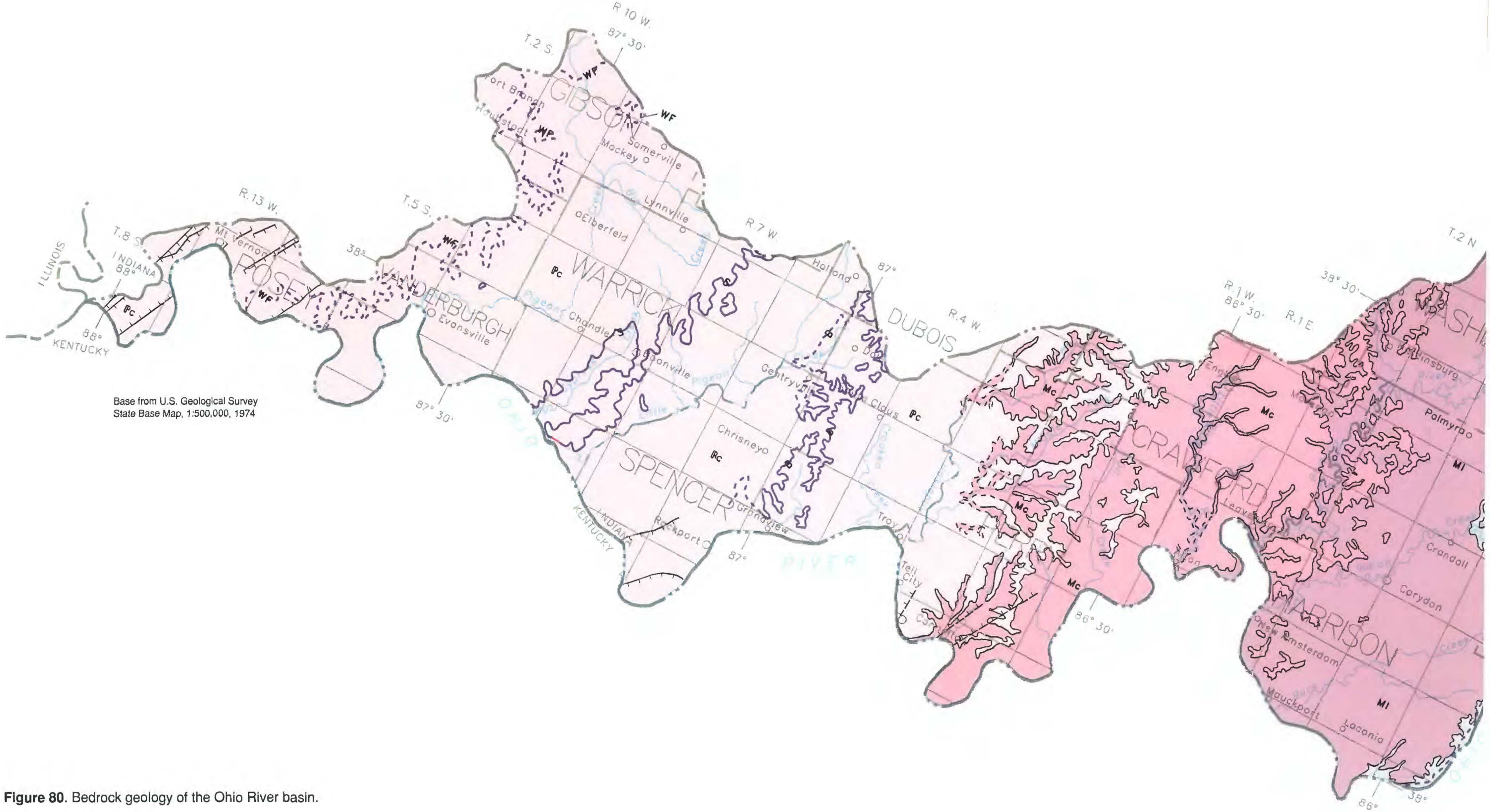
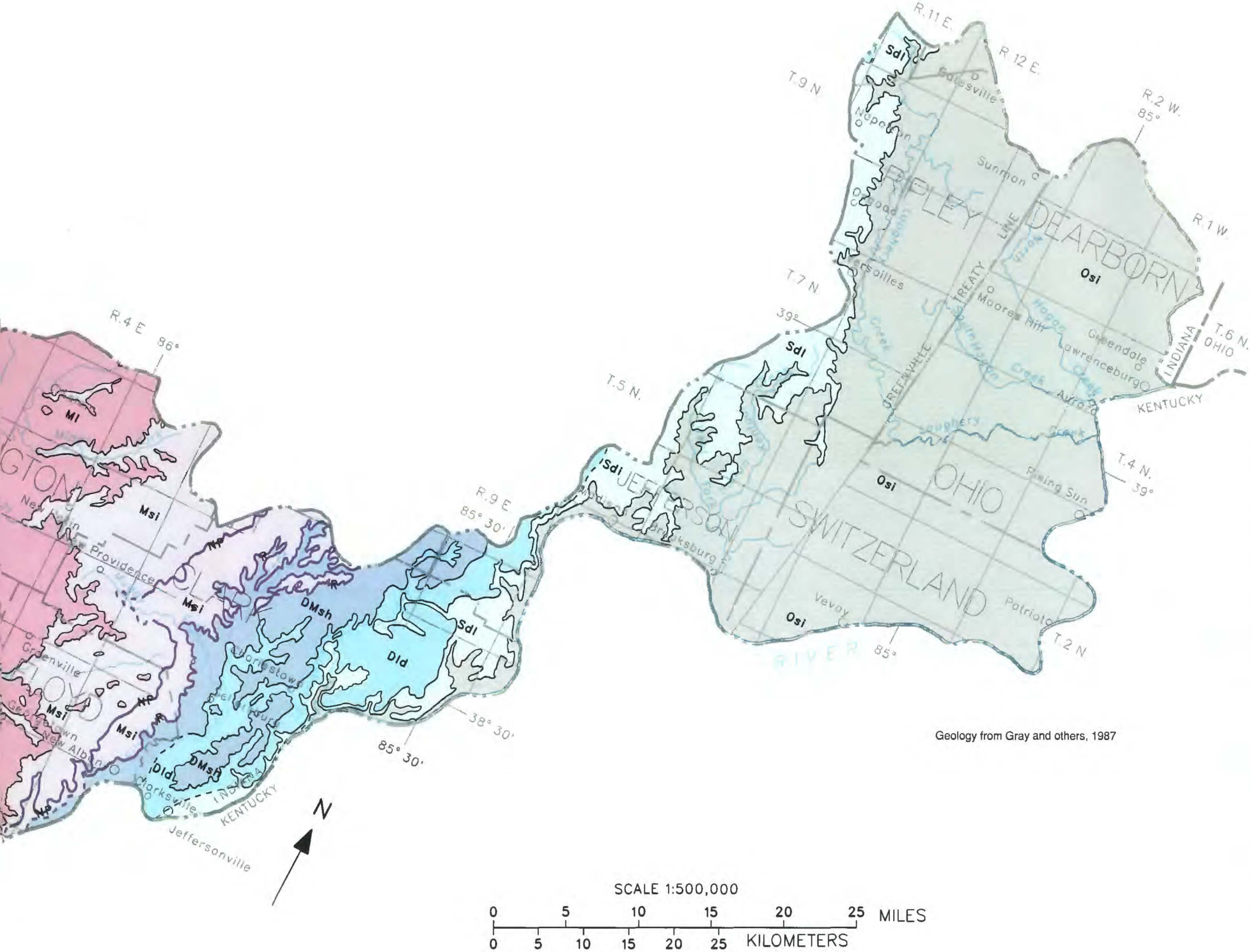


Figure 80. Bedrock geology of the Ohio River basin.



Geology from Gray and others, 1987

EXPLANATION

- **Pc** — PENNSYLVANIAN COMPLEXLY INTERBEDDED SHALE AND SANDSTONE, WITH THIN BEDS OF LIMESTONE AND COAL-- Composed of the Racoon Creek and Carbondale Groups and the Shelburn, Patoka, and Bond Formations of the McLeansboro Group
- **-WF-** — WEST FRANKLIN LIMESTONE MEMBER OF SHELBURN FORMATION
- **-S-** — SPRINGFIELD COAL MEMBER (COAL V)
- **-B-** — BUFFALOVILLE COAL MEMBER
- **Mc** — MISSISSIPPIAN COMPLEXLY INTERBEDDED SHALE, SANDSTONE AND LIMESTONE-- Composed of the West Baden, Stephensport, and Buffalo Wallow Groups
- **Ml** — MISSISSIPPIAN LIMESTONE-- Composed of the Sanders and Blue River Groups
- **-NP-** — TOP OF NEW PROVIDENCE SHALE
- **Msi** — MISSISSIPPIAN SILTSTONE AND SHALE WITH MINOR SANDSTONE AND DISCONTINUOUS LIMESTONE-- Composed of the Borden Group
- **-R-** — ROCKFORD LIMESTONE
- **DMsh** — DEVONIAN AND MISSISSIPPIAN SHALE-- Composed of the New Albany Shale
- **Dld** — DEVONIAN LIMESTONE AND DOLOMITE-- Composed of the Muscatatuck Group
- **Sdl** — SILURIAN DOLOMITE AND LIMESTONE-- Composed of the Louisville Limestone through Brassfield Limestone
- **Osl** — ORDOVICIAN SHALE AND LIMESTONE-- Composed of the Lexington Limestone, and the Kope, Dillsboro, and Whitewater Formations
- NORMAL FAULT-- Hachures on downthrown side. Dashed where approximately located
- GEOLOGIC CONTACT-- Dashed where approximately located
- WATER-MANAGEMENT-BASIN BOUNDARY

The youngest exposed bedrock in the Ohio River basin is of Pennsylvanian age (fig. 80). In ascending order, the groups that compose this rock assemblage are the Raccoon Creek, Carbondale, and the McLeansboro Groups. The Raccoon Creek Group is 95 percent shale and sandstone and 5 percent clay, coal, and limestone (Shaver and others, 1986, p. 120); small amounts of chert and sedimentary iron ore are found in the lower part of the group. The shale varies from soft and nonsilty to hard, silty, and sandy. The sandstone contains crossbedding and is mostly fine grained. The thickness of this group is variable because of the irregular unconformity between the Mississippian and Pennsylvanian rocks. The Raccoon Creek Group thickens southeastward from 100 ft in west-central Indiana to more than 1,000 ft in Vanderburgh County. The Carbondale Group consists primarily of about 300 ft of shale and sandstone (Shaver and others, 1986, p. 27). The Group contains four of Indiana's commercially important coals and some thin but laterally extensive limestones. The McLeansboro Group crops out in Gibson, Vanderburgh, Posey, and northwestern Warrick Counties. Shale and sandstone comprise more than 90 percent of the group (Shaver and others, 1986, p. 85-86). Minor amounts of siltstone, limestone, clay, and coal also are present. The McLeansboro Group is composed of the Shelburn, Patoka, and Bond Formations.

Unconsolidated Deposits

Unconsolidated deposits within the Ohio River basin are in three distinct regions: an eastern glaciated region, a western unglaciated region, and a southern river region.

The eastern glaciated region is primarily east of Floyd and Washington Counties. The thickness of the unconsolidated deposits in the upland area of the region is generally less than 50 ft (fig. 81). The material covering this area is predominantly composed of pre-Wisconsinan loam to sandy-loam till of the Jessup Formation and an upland silt complex of poorly stratified and poorly sorted sand and silt (Gray, 1989). The upland silt complex was derived from underlying weathered material and from windblown silt (loess). The composition of this material, and its topographic location, limit its potential as a source of ground water. A surficial deposit known as the lowland silt complex is present in narrow valleys of the glaciated region (Gray, 1989). The complex is composed of poorly stratified sand and silt from alluvial, colluvial, and wind-blown material. Even though this complex may contain some favorable material for aquifers, it is not considered to be a reliable source of ground water.

The unconsolidated surficial deposits are less than 50 ft thick in the uplands of the western unglaciated region of the basin (fig. 81). The unglaciated region contains terrace remnants of the lowland silt complex along many of its narrow major tributary valleys (Gray, 1989). What differentiates the unglaciated region from the glaciated region is the lack of till and the presence in the unglaciated region of large areas of loess in the upland areas. In many places, the loess is greater than 5 ft thick.

Along the Ohio River and in the middle to lower reaches of the major tributaries, the unconsolidated deposits are 50 ft to more than 100 ft thick (fig. 81). The sediment contained within the Ohio River Valley is mostly valley-train material deposited from melting glaciers. This material is undifferentiated outwash composed of sand and gravel, and it is the major source of ground water in the Ohio River basin. The outwash is commonly overlain by more than 10 ft of fine-grained sediments. In the lower to middle reaches of many of the tributaries are wetland or lake silts and clays. Covering the outwash and slack-water deposits in many areas is recently deposited alluvium composed of clay, silt, and sand. Colluvium is present along some of the tributary-valley margins.

Aquifer Types

Thirteen hydrogeologic sections (12A–12A′ to 12M–12M′, fig. 82) were drawn to show the hydrogeology of the Ohio River basin. Hydrogeologic sections 12A–12A′ to 12J–12J′ (fig. 82) are oriented south-north and are roughly perpendicular to the Ohio River. These hydrogeologic sections are spaced on average 18 mi apart; the average density of logged wells plotted along the sections is 0.64 wells per mile (fig. 78). Hydrogeologic sections 12K–12K′, 12L–12L′, and 12M–12M′ (fig. 82) are oriented west-east. These sections were spaced 21 mi apart on average; density of wells along the sections averages 0.46 wells per mile (fig. 78).

Because of the scarcity of water-well logs for the Ohio River basin, approximately 20 percent of the well logs used were from test holes. Many previous maps and reports were used in the production of these sections: (Gallaher, 1963a, 1963b, 1963c, 1964a, 1964b; Gallaher and Price, 1966; Price, 1964a, 1964b, 1964c; Bassett and Hasenmueller, 1979a, 1979b, 1980; Hasenmueller and Bassett, 1980; Geosciences Research Associates, Inc., 1982; Bassett and Keith, 1984; Gray, 1982, 1983, 1989; Gray and others, 1987). These maps and publications were especially helpful where well logs were scarce or wells were shallow.

Unconsolidated and bedrock aquifers are used as a water source in the Ohio River basin. The unconsolidated aquifers are mapped as buried sand and gravel aquifers (fig. 83). The buried sand and gravel aquifer is composed of outwash and alluvium that was deposited in the Ohio River valley, in lower reaches of the major tributary valleys, and in glacial lakes. Throughout the basin, the outwash is overlain by more than 10 ft of fine-grained sediments; therefore, the aquifer type is denoted as "buried" sand and gravel. The bedrock aquifers were subdivided into sandstone aquifers, carbonate bedrock aquifers, an upper weathered-bedrock aquifer, and a complexly interbedded sandstone, shale, limestone, and coal aquifer. The thicknesses, typical yields, and common names of these aquifer types are shown in table 14 at the back of this section.

Unconsolidated Aquifers

Buried Sand and Gravel Aquifer

An extensive zone of buried sand and gravel extends along nearly all of the Ohio River basin's southern border and along some of the lower reaches of the major tributaries (fig. 83). This zone of outwash and recently deposited alluvium is restricted to the Ohio River valley. The buried sand and gravel aquifer is shown in all 13 hydrogeologic sections (fig. 82). The aquifer is typically 35 ft to 150 ft thick, but it thins toward the valley margins. It is typically covered by 10 ft to 30 ft of clay, silt, and fine sand, although these fine-grained surface deposits can range from 0 to 100 ft in thickness. The coarseness of the deposits in the Ohio River valley generally increases with depth. Gravel is common near the bedrock surface, and boulders are sometimes found (Gallaher and Price, 1966, p. 44). The deposits also become finer with distance from the Ohio River. Sand deposits are thin or absent near the valley margins. Most of the valley-margin deposits consist of a mix of fine sand, silt, and clay and are commonly called "mud" or "quicksand" by drillers. The valley margin deposits are shown on the sections as nonaquifer material because they are not generally used for water supply; however, in some areas, low-yielding water-bearing units occur. Generally, where a tributary joins the Ohio River, the deposits in the tributary valley are fine-grained lake deposits and produce water only in some locations.

The buried sand and gravel deposits in the Ohio River valley are the most productive water-bearing units in the entire basin and yield the largest supplies of ground water. Properly constructed wells can yield as much as 2,000 gal/min, although typical yields are generally several hundred gallons per minute. This aquifer is extensively used by many of the major cities and industries along the Ohio River (Clark, 1980).

Horizontal hydraulic conductivities calculated from about 100 aquifer tests made in the buried sands and gravels along the Ohio River in Kentucky ranged from about 13 to 375 ft/d; the median was 61 ft/d (Gallaher and Price, 1966, p. 21). Specific capacities ranged from about 1 to 500 (gal/min)/ft with a median of 30 (gal/min)/ft (Gallaher and Price, 1966, p. 21). Water levels in the aquifer generally slope toward the Ohio River; ground water discharges to the river. During flooding, river water can flow into the aquifer near the stream (Gallaher and Price, 1966, p. 12). Infiltration of river water can be induced by heavy pumping in the outwash deposits; this infiltration reverses the natural pattern of ground-water discharge to the river and enables large sustained well yields.

A small buried sand and gravel aquifer in the northwestern part of the basin is about 50 ft thick and covers about 40 mi^2 (fig. 83; section 12B–12B′, fig. 82). Most of the aquifer material consists of fine sand; however, small deposits of gravel also are present. The aquifer is confined above and below primarily by deposits of fine sand, silt, and clay. Where present, the aquifer provides an adequate supply of ground water for domestic use.

Bedrock Aquifers

The bedrock aquifers in the Ohio River basin are less productive but more widespread than the unconsolidated aquifers. Several conditions limit the availability, quantity, and quality of the ground water contained within the bedrock: (1) the lack of thick, permeable, unconsolidated deposits above the bedrock throughout most of the basin; (2) low hydraulic conductivity of much of the bedrock; (3) deep water levels; and (4) highly mineralized ground water at depth.

Sandstone Aquifers

Sandstone aquifers underlie two large areas in the western one-half of the basin (fig. 83). Most of the sandstone aquifers are of Pennsylvanian age. These thick, laterally discontinuous sandstones are shown in hydrogeologic sections 12A–12A′, 12B–12B′, 12D–12D′, 12E–12E′, and 12K–12K′ (fig. 82). These sandstones are a dependable source of ground water, and they are very important aquifers in the western one-half of the basin. Wells in some of the thick sandstones can yield as much as 75 gal/min (Pettijohn and Reussow, 1969, p. 20); however, typical well yields from the sandstones are 1 to 20 gal/min.

Some of the sandstone aquifers in the far western part of the basin have been named and mapped in previous studies and are shown, but not labeled, on sections in figure 82. These sandstone aquifers include the Inglefield aquifer (Barnhart and Middleman, 1990, p. 4-7), also known as the Inglefield Sandstone aquifer (Robison, 1977, p. 9) or the Patoka aquifer (Cable and Wolf, 1977, p. 14), which is shown in the northern part of section 12A–12A′ and the central part of section 12B–12B′ (fig. 82). A second, less important aquifer, the Busseron Sandstone Member, generally lies at the base of the Danville Coal Member (Coal VII) (Shaver and others, 1986, p. 34; Barnhart and Middleman, 1990, p. 4). The Busseron Sandstone Member, (mapped by Barnhart and Middleman, 1990, pl. 1) is shown in the northern part (T. 2 S.) of section 12B–12B′ (fig. 82). The Dugger aquifer (Cable and Wolf, 1977, p. 13-15) is a discontinuous sandstone aquifer within the Dugger Formation that is present in parts of Vanderburgh, Gibson, Warrick, and Pike Counties. It is shown on the southern part of section 12K–12K′ just above the Springfield Coal Member (Coal V). Major sandstone aquifers are found in the Mansfield Formation at the base of the Raccoon Creek Group. In Vanderburgh County, the sandstone aquifers are 800 to 1,200 ft below land surface; the water in these deep sandstone aquifers is too mineralized to be used (Cable and Wolf, 1977, p. 9). Further east in Spencer County, these sandstone aquifers are only several hundred feet below land surface and can be used as a water source. These aquifers are shown at the base of section 12D–12D′ and near the land surface (above the Mississippian-Pennsylvanian unconformity) in Perry County in the northern part of section 12E–12E′ (fig. 82).

Water-bearing sandstones at the bedrock surface, such as those shown in the central part of 12B–12B′ or the northern part of 12E–12E′ (fig. 82), are recharged by precipitation infiltrating through the thin soil or glacial cover. Deeper sandstones are recharged much more slowly by overlying shales and siltstones or by limestones or coals, which can function as conduits of flow to the sandstones. Flow in the sandstone aquifers could be restricted because of their discontinuity and the abundance of surrounding shale and siltstone. The average hydraulic conductivity of the sandstones in the basin was estimated to be 1.6 ft/d (Pettijohn and Reussow (1969, p. 20).

In general, water quality diminishes with depth, but the depth to potable ground water is highly variable. In Vanderburgh County, freshwater was found in sandstone at depths of more than 500 ft; the ground water was highly mineralized at 800 ft below land surface (Cable and Wolf, 1977, p. 9, 22). Locally, ground water can be salty at depths of less than 300 ft. In section 12E–12E′ (fig. 82), a well near the intersection of T. 4 S. and T. 5 S. contained "saltwater" at a depth of about 280 ft.

Complexly Interbedded Sandstone, Shale, Limestone, and Coal Aquifer

A complexly interbedded sandstone, shale, limestone, and coal aquifer has been mapped over the western one-half of the basin (fig. 83). The complexly interbedded aquifer was mapped as "aquifer—potential unknown" because its aquifer characteristics are so widely variable throughout the basin. The aquifer is a poorly productive, secondary source of water, yet it is areally and vertically extensive; in general, it can supply enough water to support a household. Where sandstone aquifers have been mapped (fig. 83), the complexly interbedded aquifer is typically not used unless the sandstone aquifers are very deep (several hundred feet below land surface).

The complexly interbedded aquifer includes the West Baden, Stephensport, and Buffalo Wallow Groups of Mississippian age and the Raccoon Creek, Carbondale, and McCleansboro Groups of Pennsylvanian age. Where the sandstones are mapped as separate aquifers from the complex material (fig. 83), then the complexly interbedded sandstone, shale, limestone, and coal aquifer is dominated by shale and siltstone. Aside from the major sandstone bodies, shale and siltstone compose about 80 percent of the bedrock in well logs from sections 12A–12A′ to 12C–12C′ (fig. 82); however, large sandstone bodies may be present where complex material is shown with no well log information. Shale and siltstone, even though widespread, are not the primary water-bearing units in the complex aquifer. Although some well logs note shale and sandy shale as water bearing, about one-half of the well logs indicate that water is derived from thin (less than 10 ft), discontinuous sandstone bodies (not shown on the sections). The remaining wells get water from coal and limestone.

Most wells are uncased throughout the bedrock and, therefore, can produce water from several low-yielding units that are penetrated by the well. Yields of wells in the complex material range from 0 to about 20 gal/min. Commonly, yields are less than 2 gal/min.

Carbonate Bedrock Aquifers

Limestones and dolomites of Silurian, Devonian, and Mississippian age are another source of ground water in the Ohio River basin. The Mississippian carbonate rocks, consisting of the Blue River and Sanders Groups, are the most extensive and thickest of the carbonate bedrock aquifers in the Ohio River basin. They underlie most of Perry, Crawford, and Harrison Counties and the western one-half of Washington County (fig. 83). These thick, massive Mississippian carbonate bedrock aquifers are shown in hydrogeologic sections 12F–12F′, 12G–12G′, the eastern part of 12K–12K′, and the western part of 12L–12L′ (fig. 82). In western Perry and Crawford Counties, the carbonate rocks are overlain by younger complexly interbedded rocks.

Wells in the Mississippian carbonate rocks have yields of 1 to 100 gal/min; typical yields are 3 to 15 gal/min (Bechert and Heckard, 1966; Pettijohn and Reussow, 1969; Clark, 1980). These moderate yields are attributed to the presence of highly transmissive solution features, joints, cracks, and bedding planes (fig. 9b). Most of the solution features and greatest permeabilities are in the Blue River Group, which constitutes the upper two-thirds of the carbonate bedrock aquifer. Low well yields and dry holes are common where the aquifer thins toward the east, as shown along R. 5 E. of section 12L–12L′ (fig. 82). Several spring horizons have formed within the Mississippian limestones where permeable limestone overlies relatively impermeable limestone.

The Silurian-Devonian carbonate bedrock aquifer is in the east-central part of the Ohio River basin in Clark, Jefferson, and Ripley Counties (fig. 83). The aquifer can be seen in hydrogeologic sections 12H–12H′ and the far eastern part of 12L–12L′ (fig. 82). Most of the extractable ground water within the carbonate rocks is in joint planes, fractures, and solution features. The carbonate rocks identified as an aquifer are a reliable source of ground water, and well yields are generally less than 20 gal/min (Bechert, and Heckard, 1966; Pettijohn and Reussow, 1969; Clark, 1980).

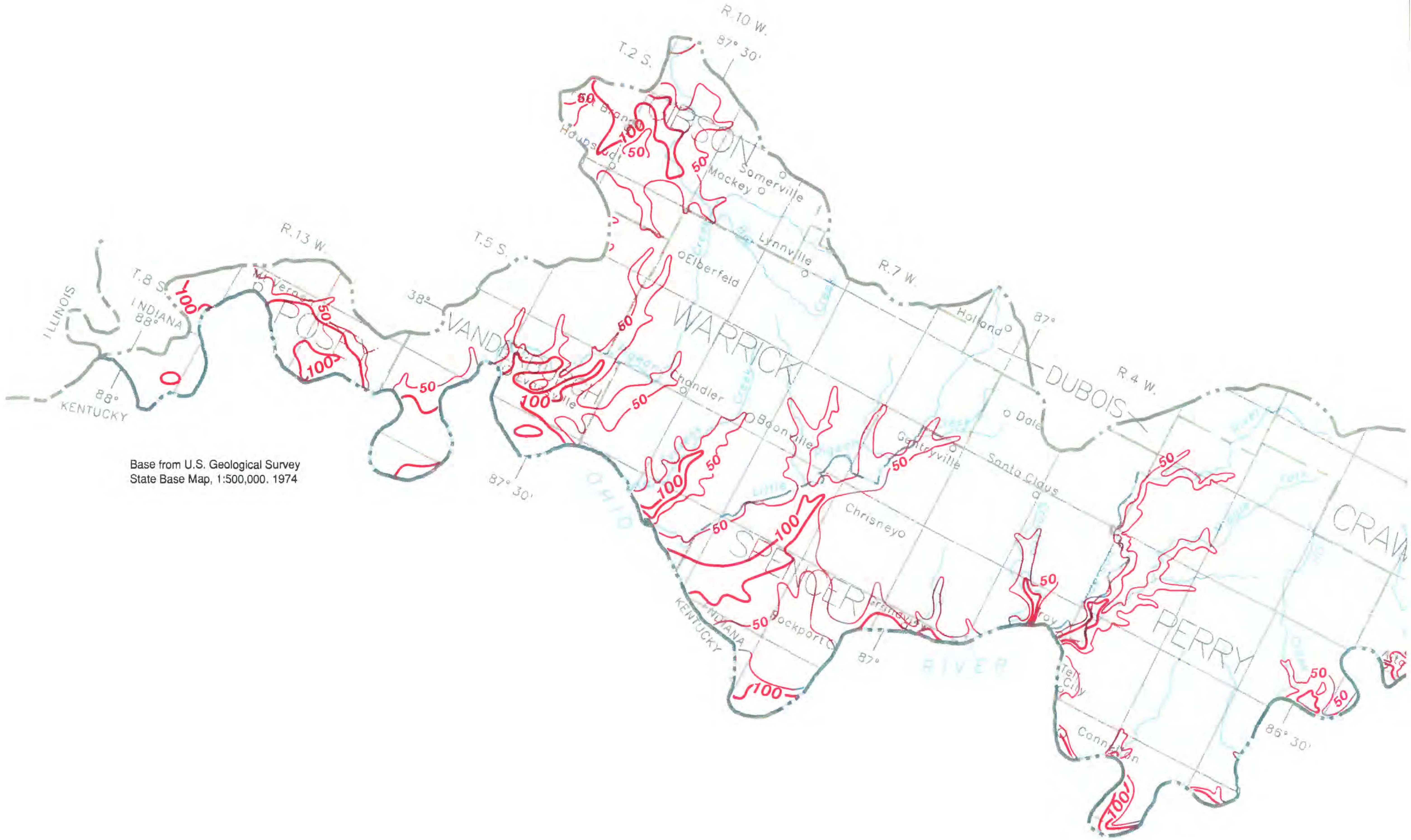
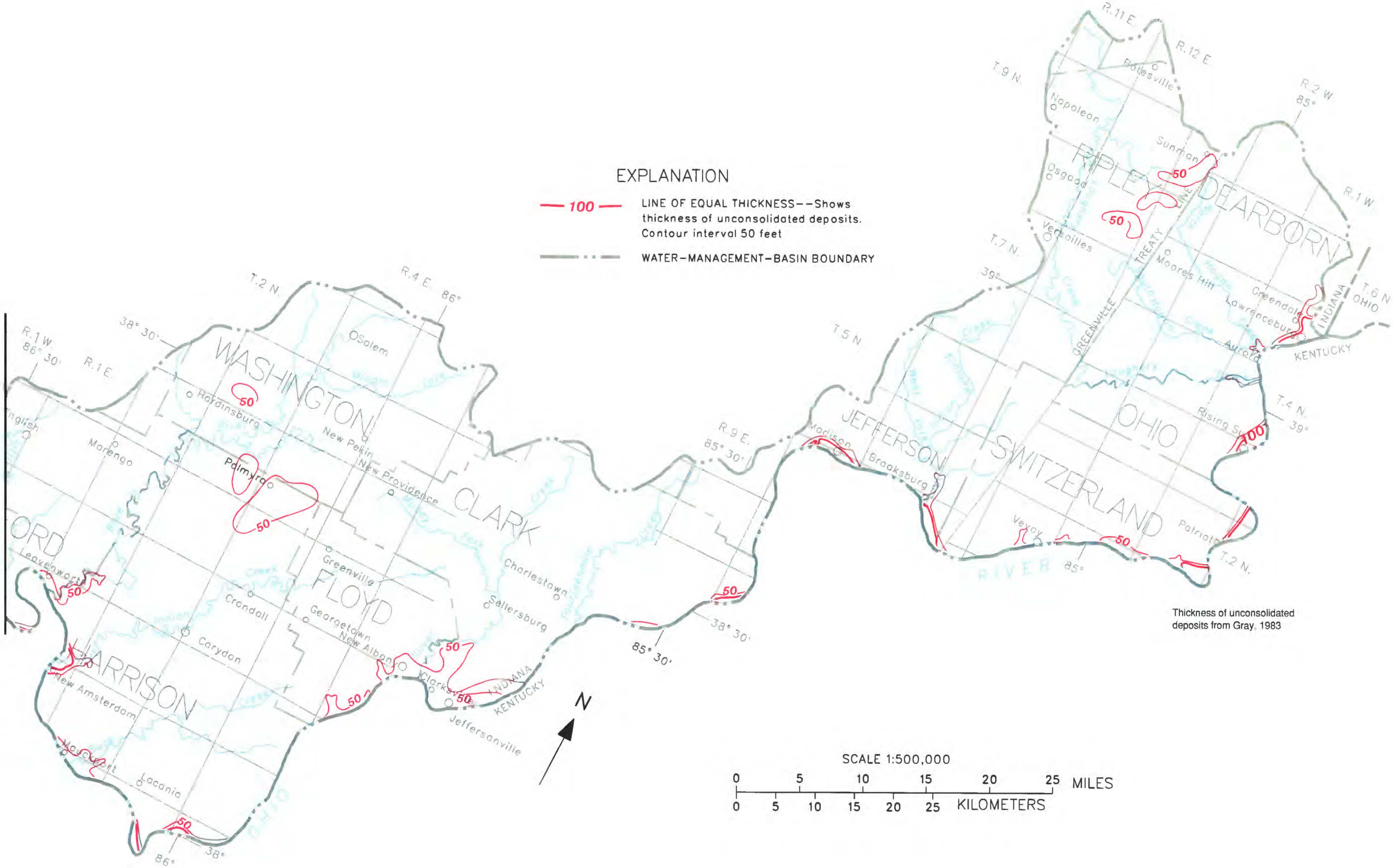
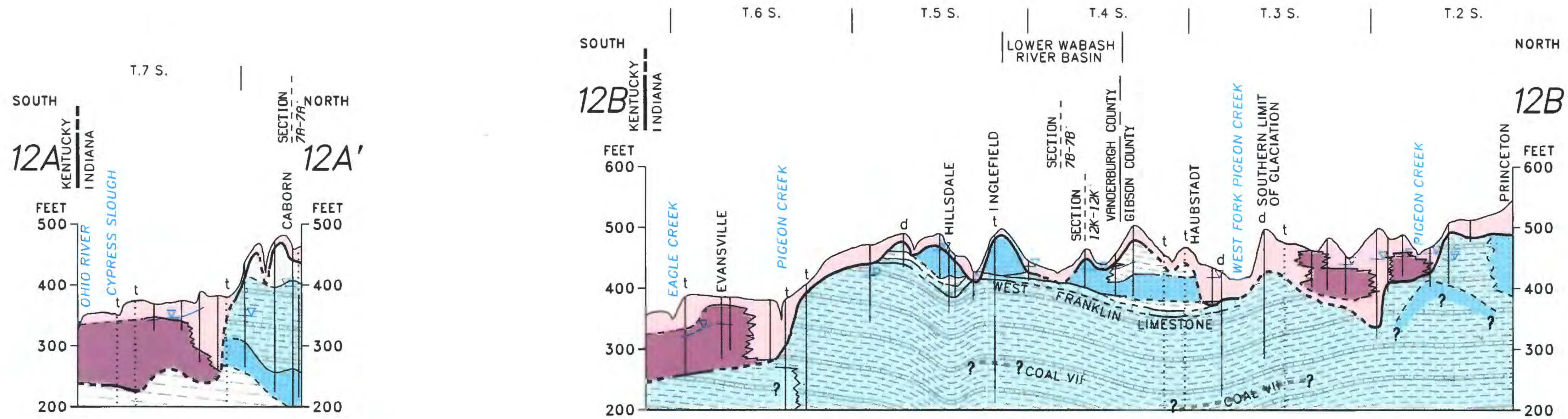


Figure 81. Thickness of unconsolidated deposits in the Ohio River basin.



Thickness of unconsolidated deposits from Gray, 1983



EXPLANATION

SAND AND GRAVEL

UNCONSOLIDATED NONAQUIFER MATERIAL

LIMESTONE AND DOLOSTONE

SHALE

SANDSTONE

COMPLEXLY INTERBEDDED SANDSTONE, SHALE, AND LIMESTONE

LIMESTONE AND SHALE

SILTSTONE AND SHALE, WITH MINOR SANDSTONE AND DISCONTINUOUS LIMESTONE

COMPLEXLY INTERBEDDED SANDSTONE, SHALE, AND MINOR LIMESTONE AND COAL

BEDROCK AQUIFER

BEDROCK AQUIFER--Potential unknown

BEDROCK NONAQUIFER

BEDROCK SURFACE--Dashed where approximately located

CHRONOSTRATIGRAPHIC BOUNDARY--Dashed where approximately located

LITHOLOGIC CONTACT--Dashed where approximately located

COAL SEAM--Dashed where approximately located

GENERALIZED POTENTIOMETRIC SURFACE--Dashed where approximately located

WELL--All well data are projected to trace of section. Dotted where data are incomplete

BASE OF UPPER WEATHERED BEDROCK

? LITHOLOGIC BOUNDARY UNKNOWN

t TEST HOLE--Not drilled for water supply

d DRY HOLE

P PENNSYLVANIAN

M MISSISSIPPIAN

D DEVONIAN

S SILURIAN

O ORDOVICIAN

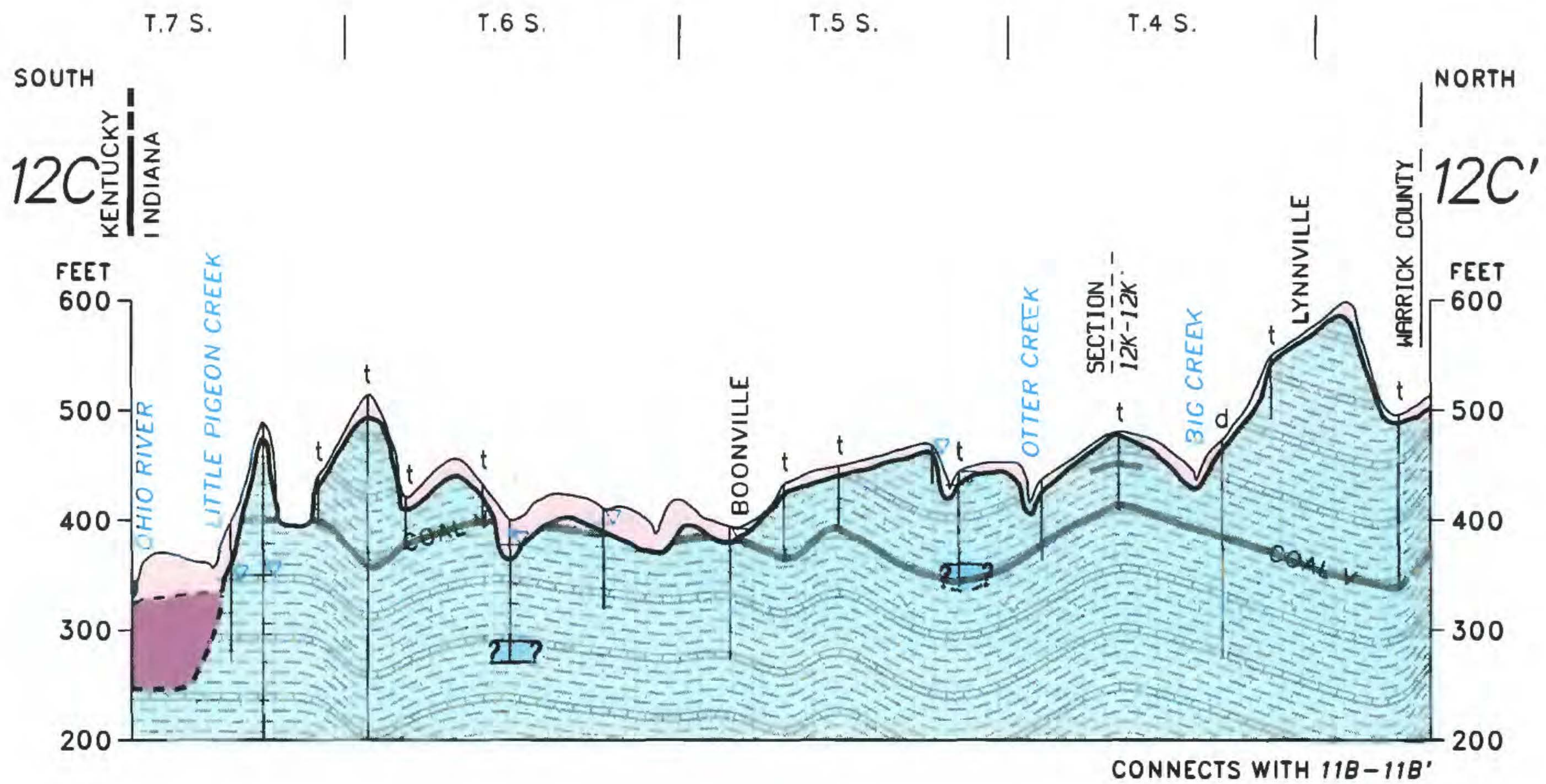
DATUM IS SEA LEVEL

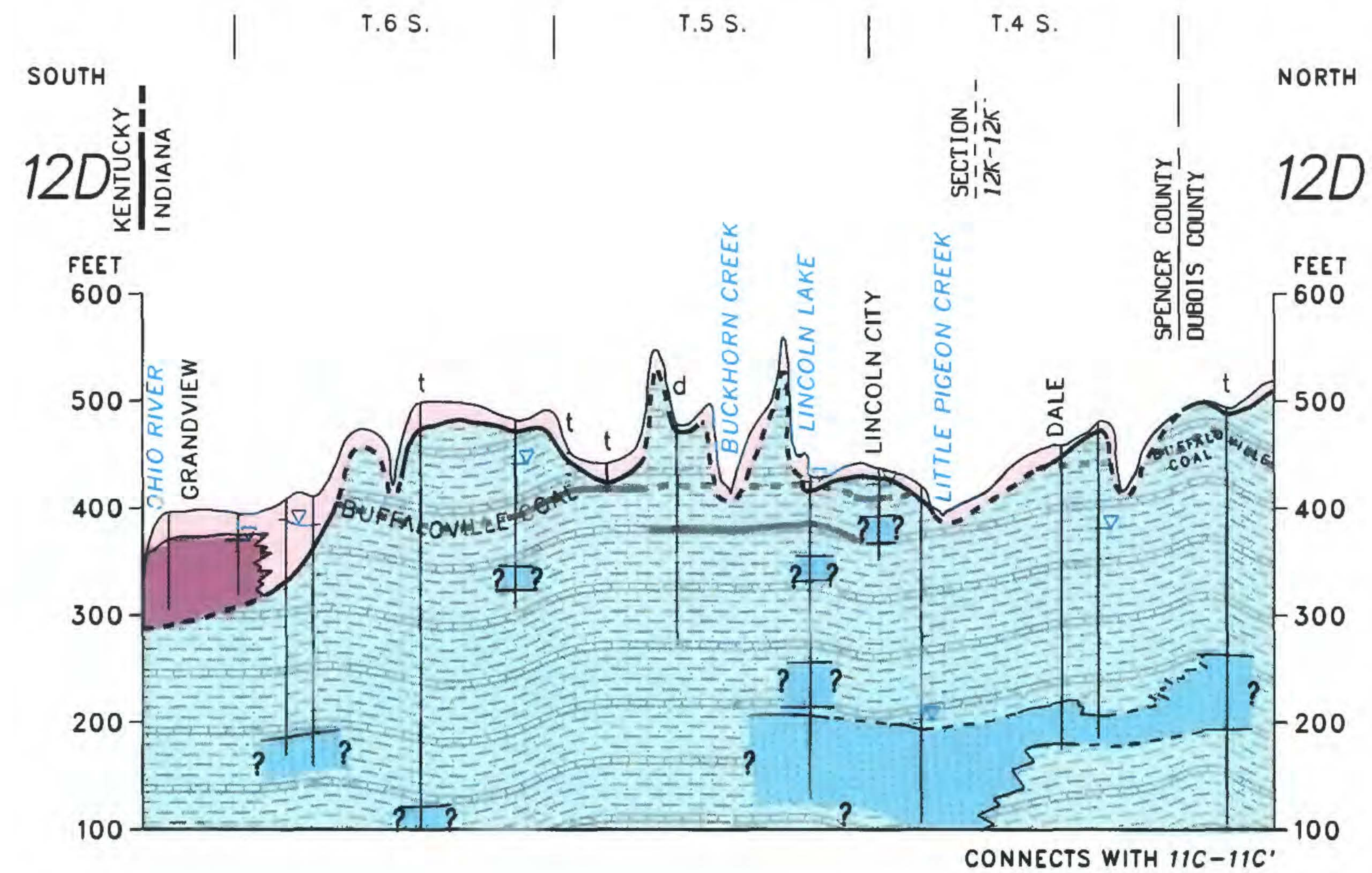
VERTICAL SCALE GREATLY EXAGGERATED

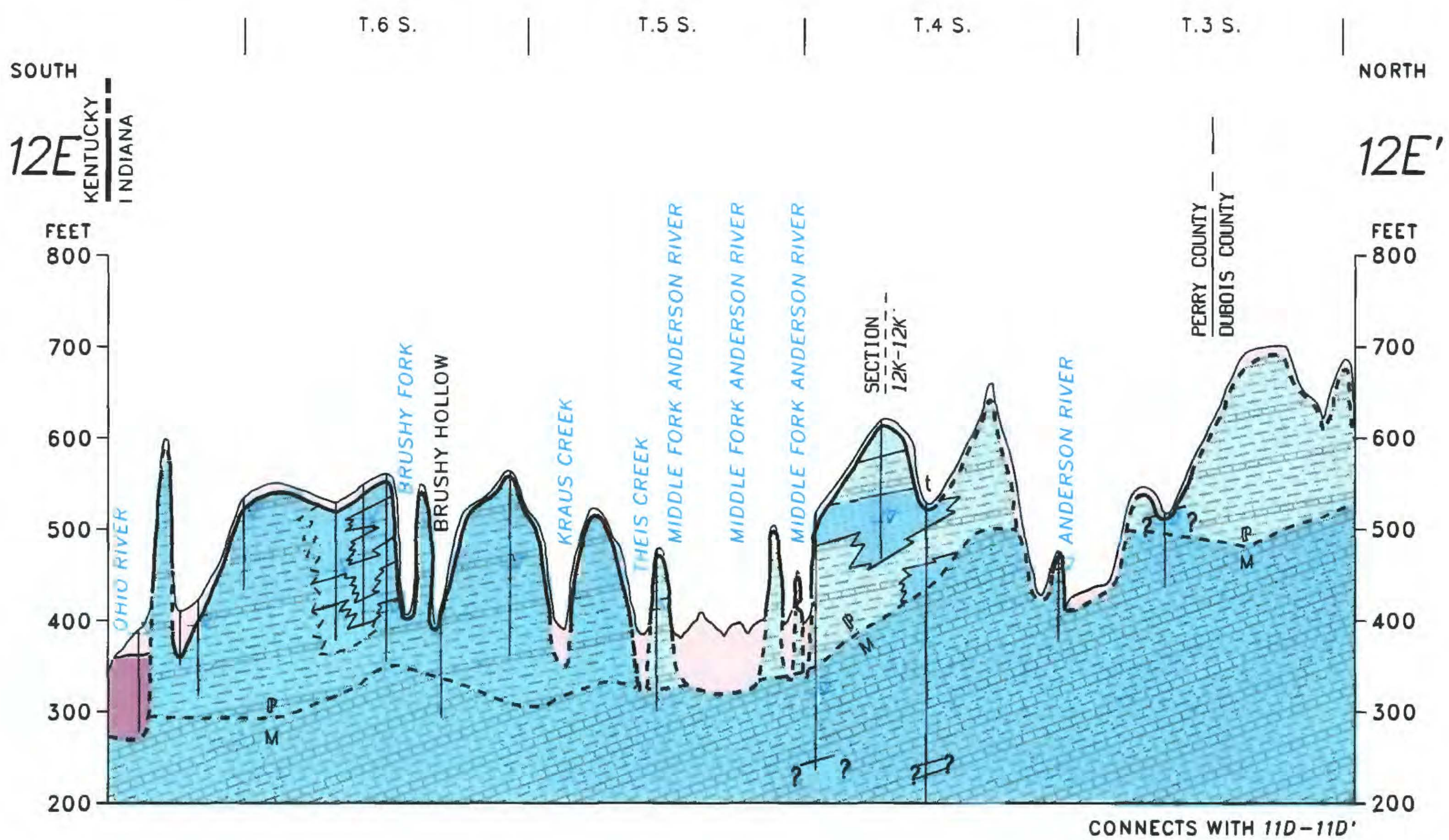
0 5 10 MILES

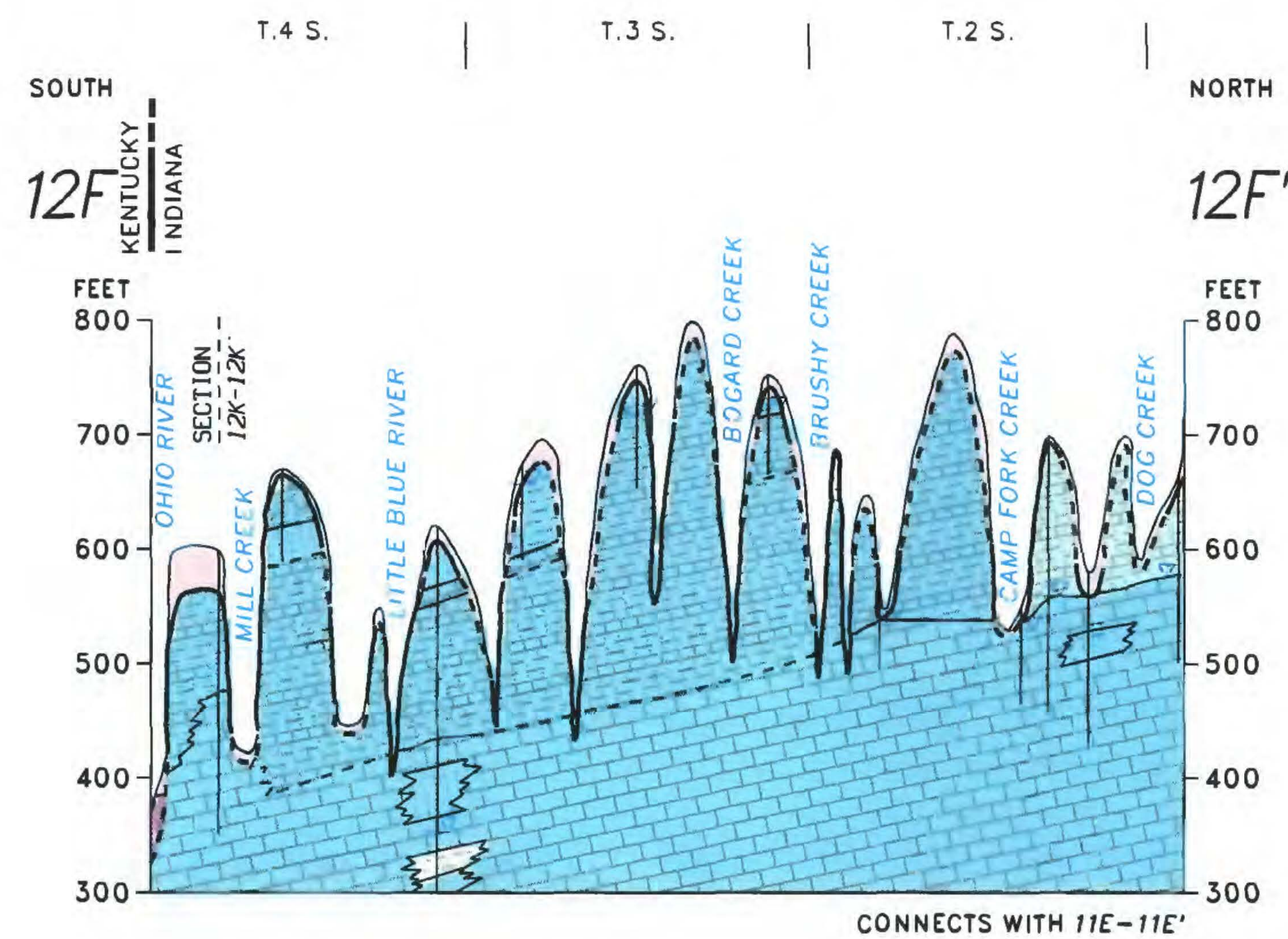
0 5 10 KILOMETERS

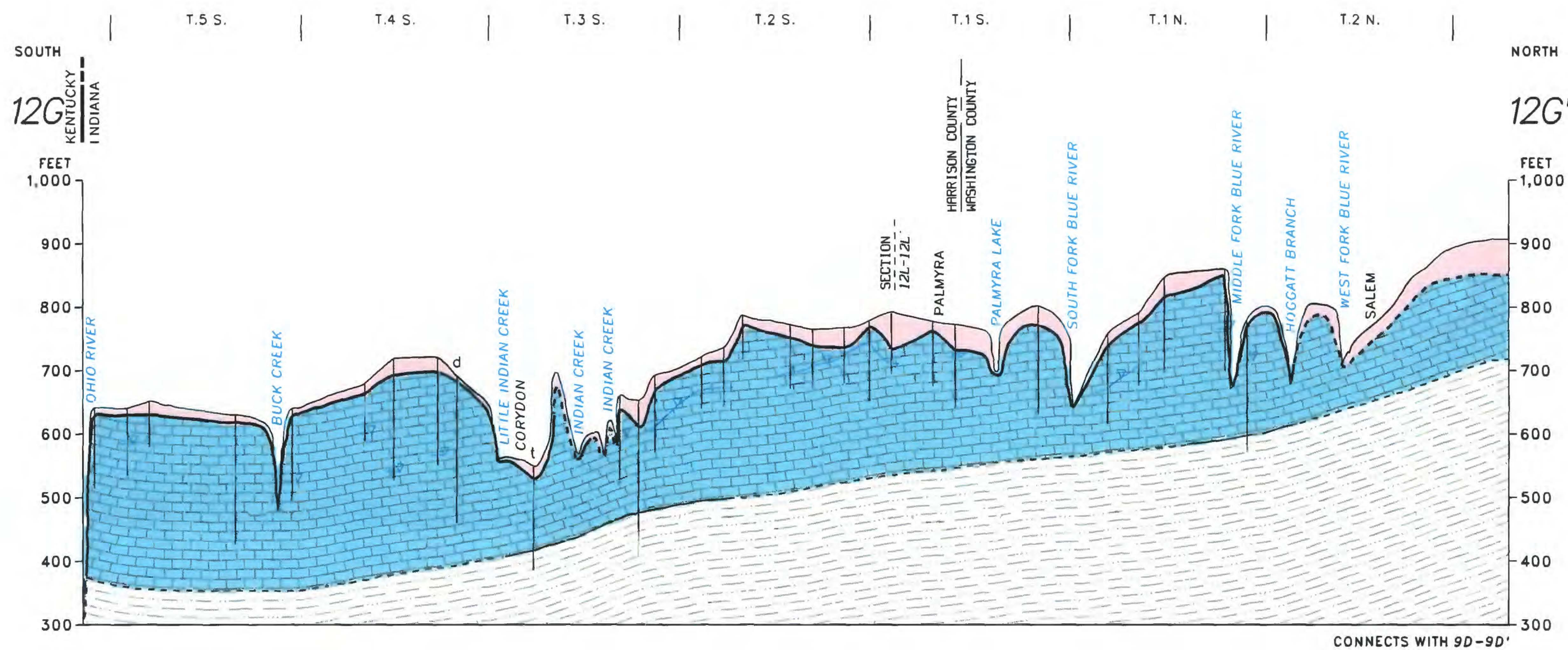
Figure 82. Hydrogeologic sections *12A–12A'* to *12M–12M'* of the Ohio River basin.

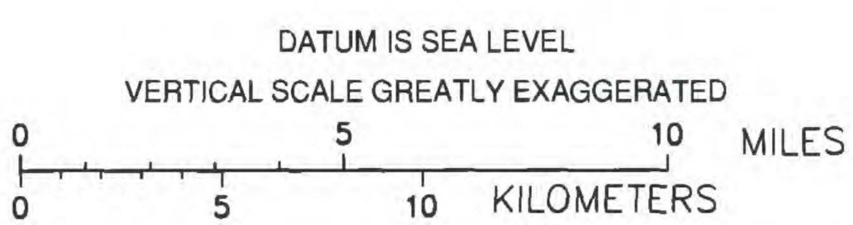

T.7 S.
T.6 S.
T.5 S.
T.4 S.
SOUTH
NORTH
12C
12C'
KENTUCKY
INDIANA
FEET
600
500
400
300
200
OHIO RIVER
LITTLE PIGEON CREEK
BOONVILLE
OTTER CREEK
SECTION 12K-12K'
BIG CREEK
LYNNVILLE
WARRICK COUNTY
COAL V
CONNECTS WITH 11B-11B'


T.6 S.
T.5 S.
T.4 S.
SOUTH
NORTH
12D
12D'
KENTUCKY
INDIANA
FEET
600
500
400
300
200
100
OHIO RIVER
GRANDVIEW
BUCKHORN CREEK
LINCOLN LAKE
LINCOLN CITY
LITTLE PIGEON CREEK
SECTION 12K-12K'
DALE
SPENCER COUNTY
DUBOIS COUNTY
BUFFALOVILLE COAL
CONNECTS WITH 11C-11C'


T.6 S.
T.5 S.
T.4 S.
T.3 S.
SOUTH
NORTH
12E
12E'
KENTUCKY
INDIANA
FEET
800
700
600
500
400
300
200
OHIO RIVER
BRUSHY FORK
BRUSHY HOLLOW
KRAUS CREEK
THEIS CREEK
MIDDLE FORK ANDERSON RIVER
MIDDLE FORK ANDERSON RIVER
MIDDLE FORK ANDERSON RIVER
SECTION 12K-12K'
ANDERSON RIVER
PERRY COUNTY
DUBOIS COUNTY
CONNECTS WITH 11D-11D'


T.4 S.
T.3 S.
T.2 S.
SOUTH
NORTH
12F
12F'
KENTUCKY
INDIANA
FEET
800
700
600
500
400
300
OHIO RIVER
SECTION 12K-12K'
MILL CREEK
LITTLE BLUE RIVER
BOGARD CREEK
BRUSHY CREEK
CAMP FORK CREEK
DOG CREEK
CONNECTS WITH 11E-11E'





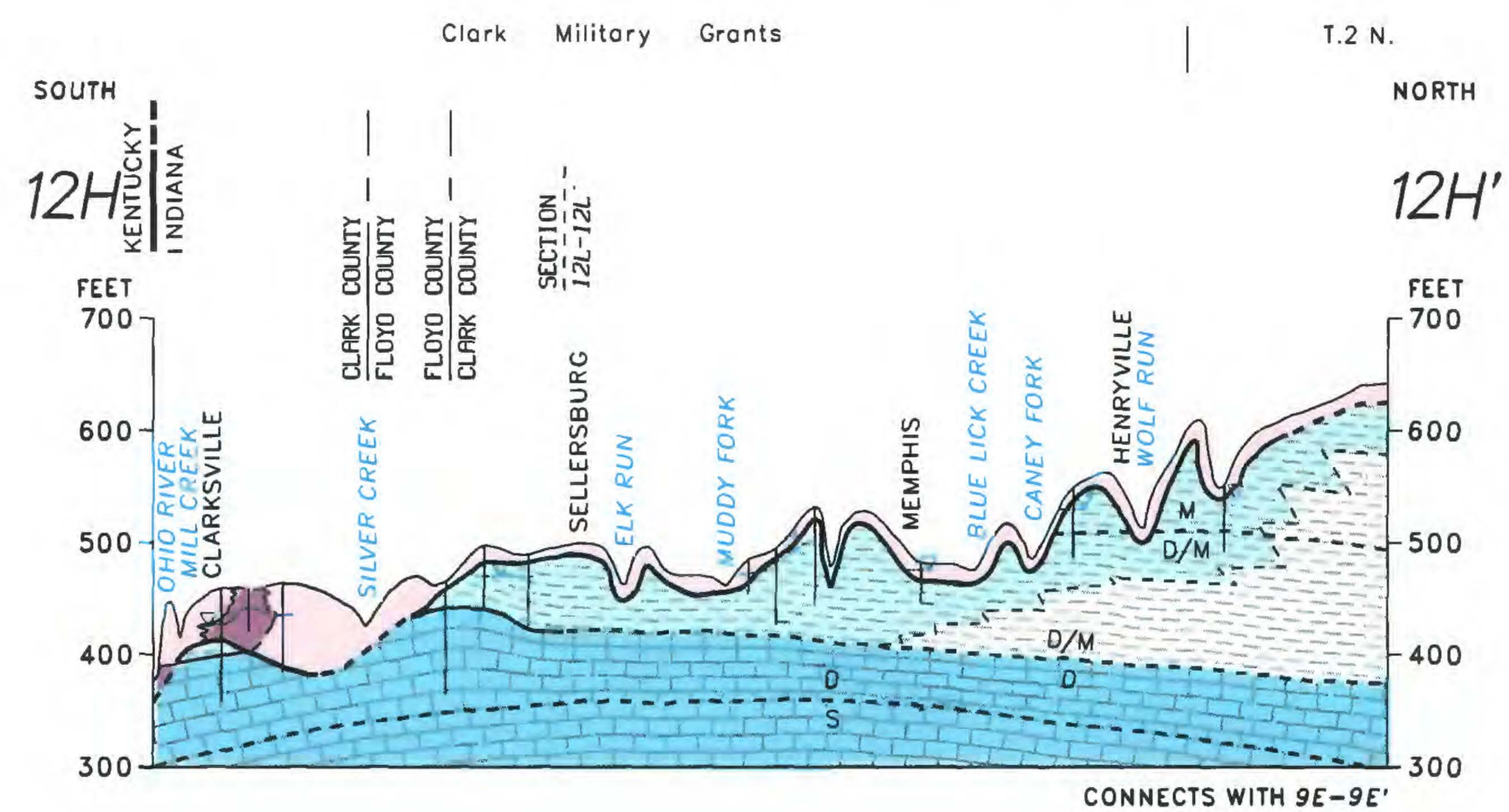
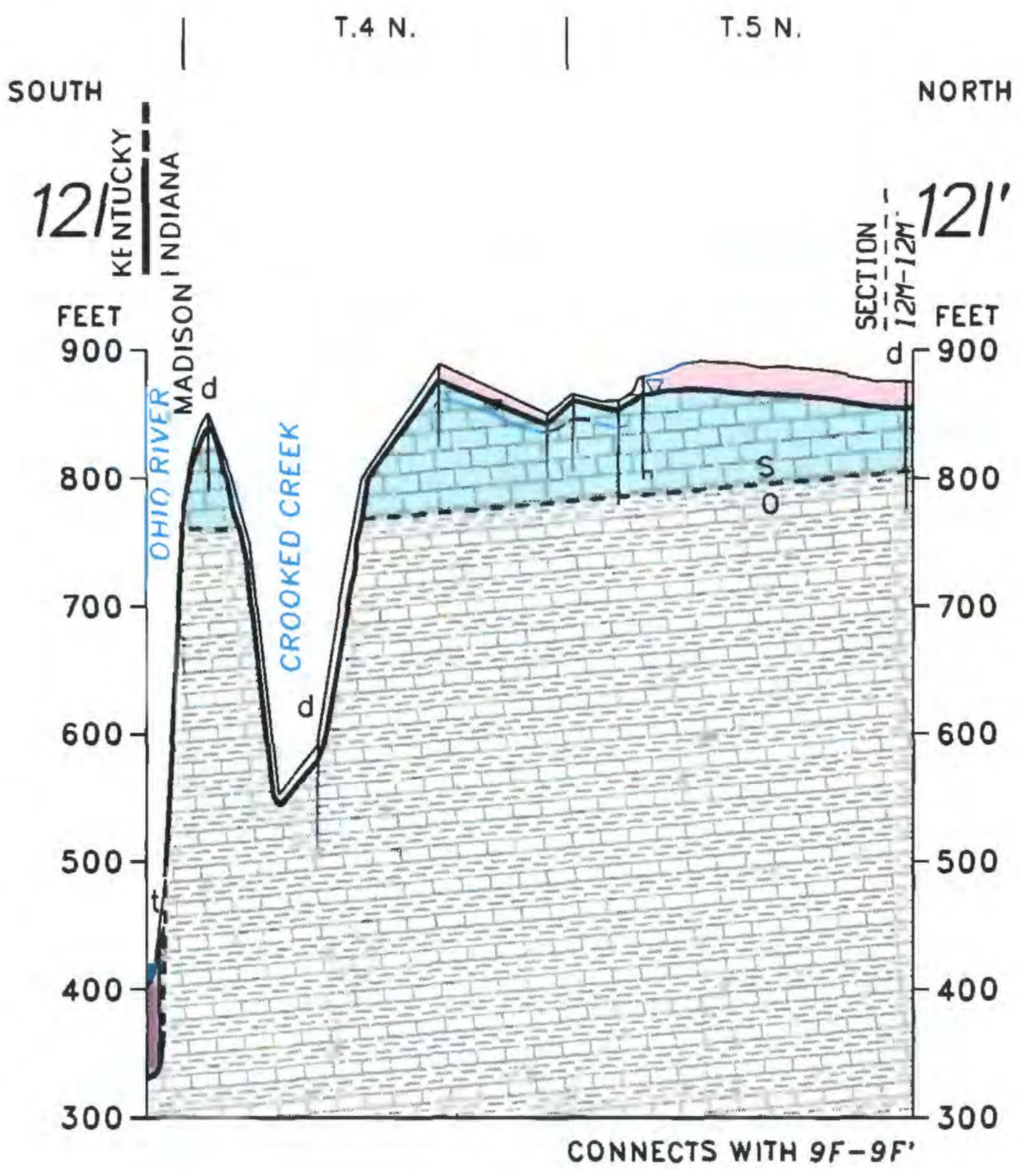
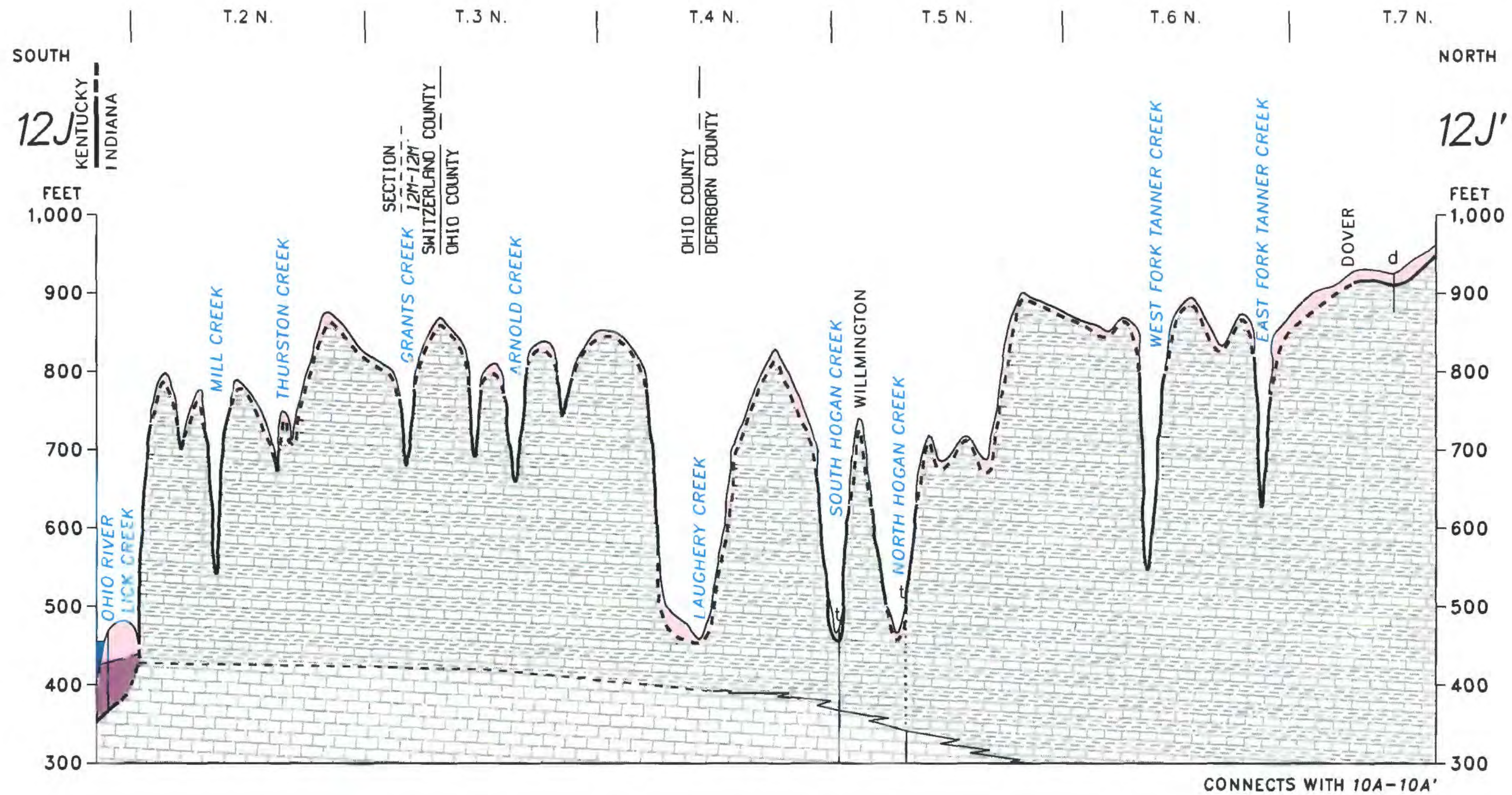


Figure 82. Hydrogeologic sections *12A–12A'* to *12M–12M'* of the Ohio River basin—Continued.





The eastern edge of the Silurian-Devonian carbonate bedrock aquifer, where only the lower part of the Silurian carbonate rocks remain, is mapped as "aquifer—potential unknown." (See the eastern part of section 12I–12I' and the western part of section 12M–12M'; fig. 82). Well yields are highly variable, and dry holes are common. Many wells that produce water in this area are on lineaments and fracture traces (Greeman, 1981; 1983). Lineaments and fracture traces are commonly surface expressions of subsurface fractures and solution features. Yields of wells that intersect these fractures are usually less than 20 gal/min; holes can be dry if they do not intersect a fracture. Another source of water within the Silurian carbonate bedrock aquifer is from minor solution features at the Ordovician-Silurian unconformity.

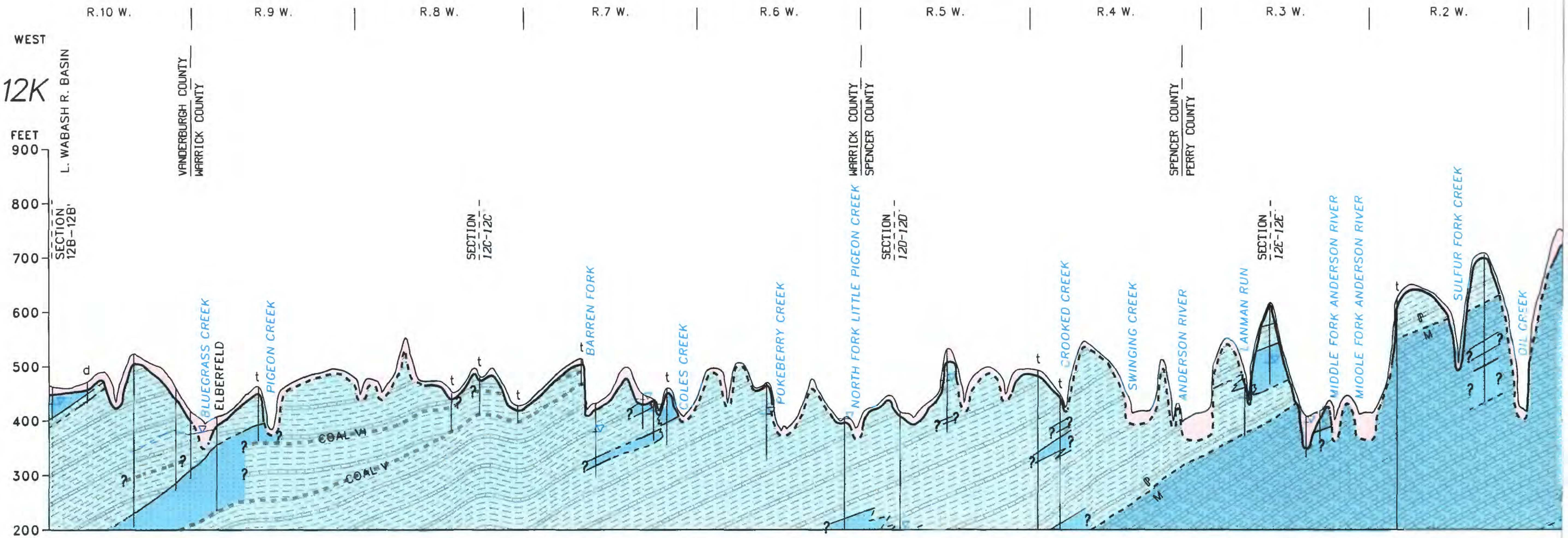
Upper Weathered-Bedrock Aquifer

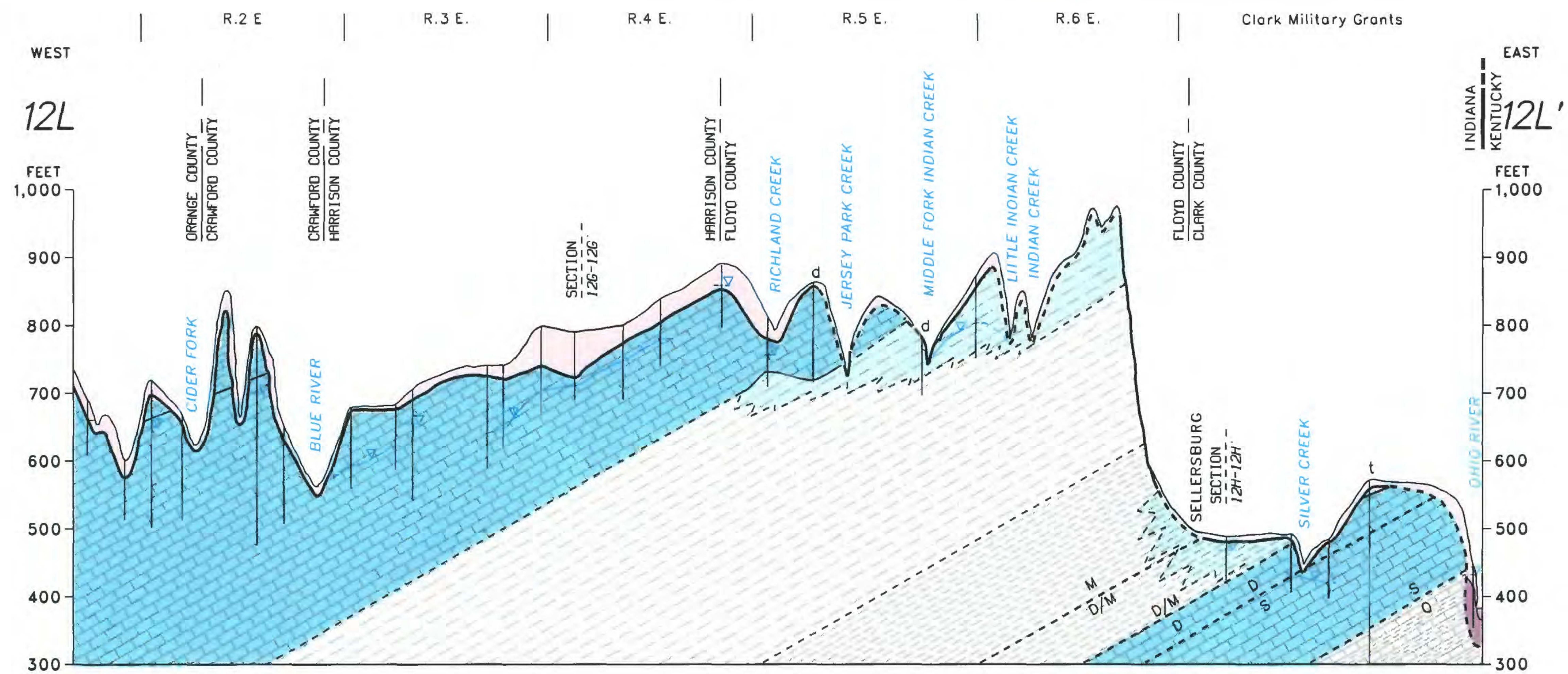
An upper weathered-bedrock aquifer in shales and siltstones underlies Washington, Clark, and Floyd Counties in the east-central part of the basin (fig. 83). The aquifer, composed of the Devonian and Mississippian New Albany Shale and Borden Group, lies between Silurian-Devonian carbonate rocks and Mississippian carbonate rocks (section 12L–12L', fig. 82). The weathered bedrock was mapped as "aquifer—potential unknown" because of the low water-yielding character of the rocks and the frequency of dry holes reported by well drillers. Ground water is obtained from joints and bedding planes that are common in the upper 50 to 100 ft where the permeability has been enhanced by weathering. The weathered shales (section 12H–12H', fig. 82) appear to be better water producers than the weathered siltstones, possibly because of better development of bedding planes in the shales.

The upper weathered-bedrock aquifer is a useful aquifer only where it is at or very near the bedrock surface. At depth, the shales and siltstones are not considered an aquifer because they yield very little or no water because of low permeabilities and limited recharge. In addition, the productive Mississippian carbonate bedrock aquifer is available at shallow depths.

Generally Nonaquifer Material

No bedrock aquifer is present in the far eastern part of the basin (fig. 83). Furthermore, few unconsolidated deposits overlie the bedrock, which consists predominantly of Ordovician shale and limestone. Most of the shale in the Ordovician bedrock is soft and does not enable joints to develop; therefore, permeabilities in the Ordovician rock are low (Gray, 1972, p. 22). Conventional water wells are rare in this area (sections 12J–12J' and 12M–12M', fig. 82); most people living there obtain drinking water from cisterns, small reservoirs, large-diameter bucket-rig wells, the Ohio River, or rural water utilities which obtain ground water from the alluvium along the Ohio River (Clark, 1980; W.J. Steen, Indiana Department of Natural Resources, written commun., 1991).

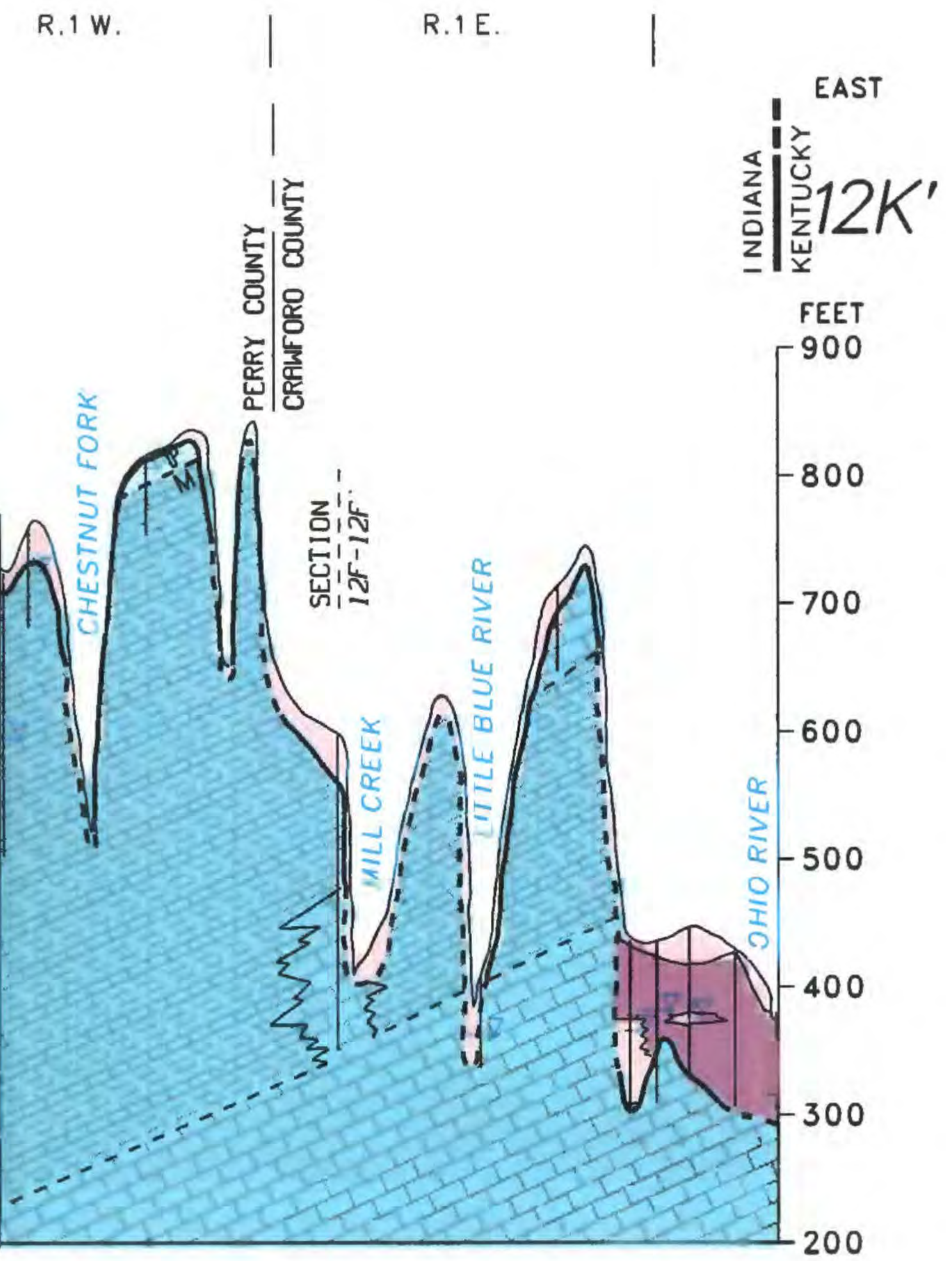


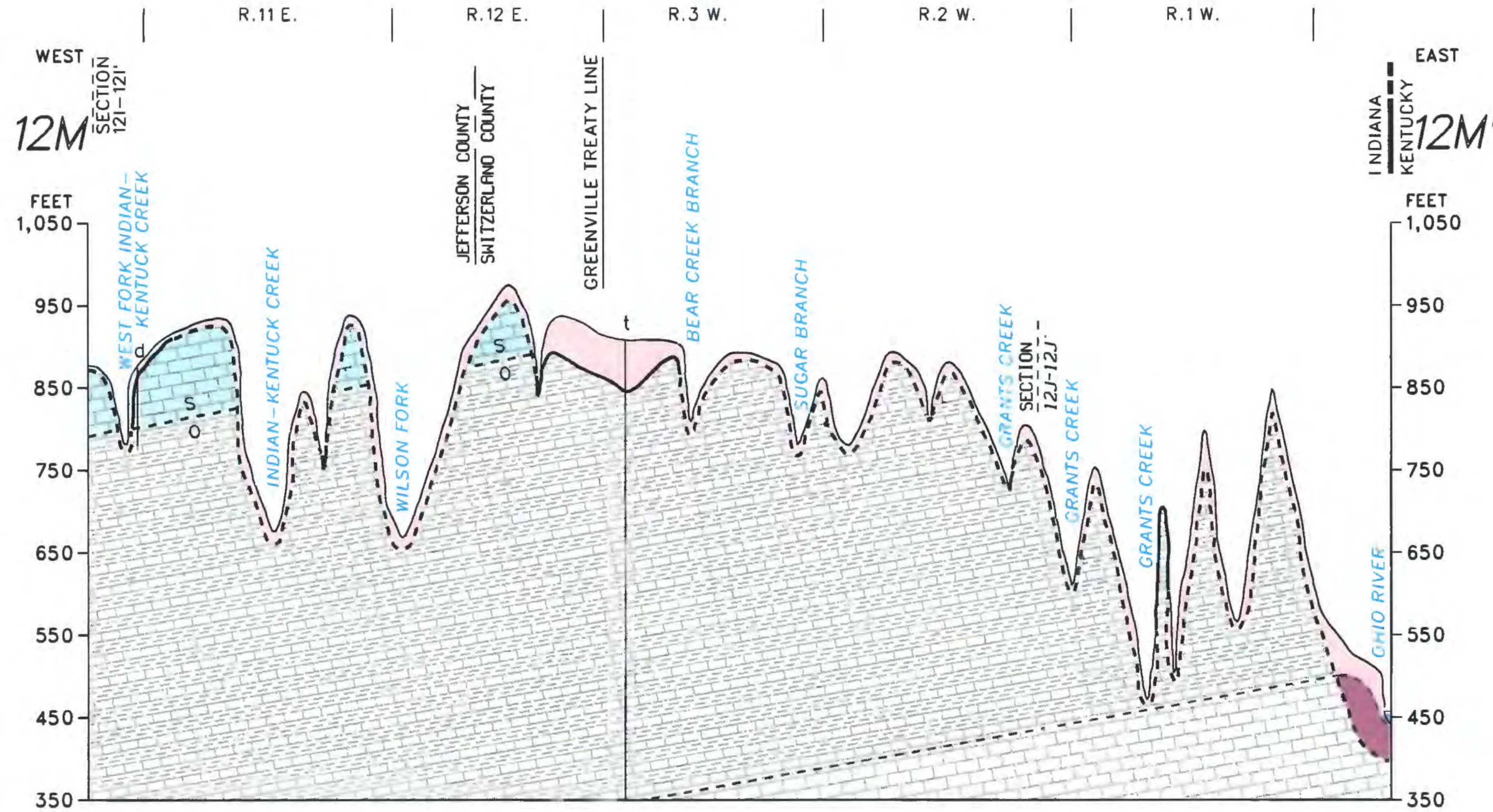


DATUM IS SEA LEVEL
VERTICAL SCALE GREATLY EXAGGERATED

0 5 10 MILES
0 5 10 KILOMETERS

Figure 82. Hydrogeologic sections *12A–12A′* to *12M–12M′* of the Ohio River basin—Continued.





Summary

The Ohio River basin encompasses 4,224 mi^2 of southern Indiana. The oldest exposed rocks in the basin (Ordovician in age) are in the eastern part of the basin and the youngest rocks (Pennsylvanian in age) are exposed in the western part of the basin. These Paleozoic rocks are covered by as much as 100 ft of loam to sandy-loam till, lake silt and clay, alluvium, loess, and outwash sand and gravel deposits.

Five aquifer types were delineated in the Ohio River basin: (1) buried sand and gravel, (2) sandstones of Pennsylvanian age, (3) complexly interbedded sandstone, shale, limestone, and coal, (4) limestone and dolomite of Silurian, Devonian, and Mississippian age, and (5) an upper weathered zone of siltstone and shale. Buried sand and gravel is found primarily in the Ohio River valley. The sand and gravel yields the largest supply of ground water for the basin; well yields typically range from 10 to 1,000 gal/min. The bedrock aquifers are a less productive supply of ground water for the basin and generally support only domestic use. Well yields from the bedrock aquifers normally range from 0 to 20 gal/min. The highest yields from bedrock come from the sandstone and carbonate bedrock aquifers. The complexly interbedded aquifer, upper weathered-bedrock aquifer, and the eastern edge of the Silurian-Devonian carbonate bedrock aquifer were all mapped as "aquifer—potential unknown." These areas are characterized by generally low yields (less than 5 gal/min) and dry holes. The eastern end of the basin contains no mappable aquifer.

Four factors greatly limit the distribution, quantity, and quality of the ground water contained within the Ohio River basin: (1) the lack of thick, permeable unconsolidated deposits over most of the basin, (2) low hydraulic conductivity of the bedrock, (3) deep water levels, and (4) highly mineralized ground water at depth.

References Cited

Arvin, D.V., 1989, Statistical summary of streamflow data for Indiana: U.S. Geological Survey Open-File Report 89-62, 964 p.

Barnhart, J.R., and Middleman, B.H., 1990, Hydrogeology of Gibson County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 41, 18 p.

Bassett, J.L., and Hasenmueller, N.R., 1979a, Map showing structure on top of the New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources Geological Survey, EGSP Series 801, scale 1:500,000.

Bassett, J.L., and Hasenmueller, N.R., 1979b, Map showing structure on base of New Albany Shale (Devonian and Mississippian) and equivalent strata in Indiana: Indiana Department of Natural Resources Geological Survey, EGSP Series 800, scale 1:500,000.

Bassett, J.L., and Hasenmueller, N.R., 1980, Map of Indiana showing structure on top of the Maquoketa Group (Ordovician): Indiana Department of Natural Resources Geological Survey, METC/EGSP Series 812, scale 1:500,000.

Bassett, P.A., and Keith, B.D., 1984, Data base for deep wells in Indiana: Indiana Department of Natural Resources, Geological Survey Occasional Paper 46, 35 p.

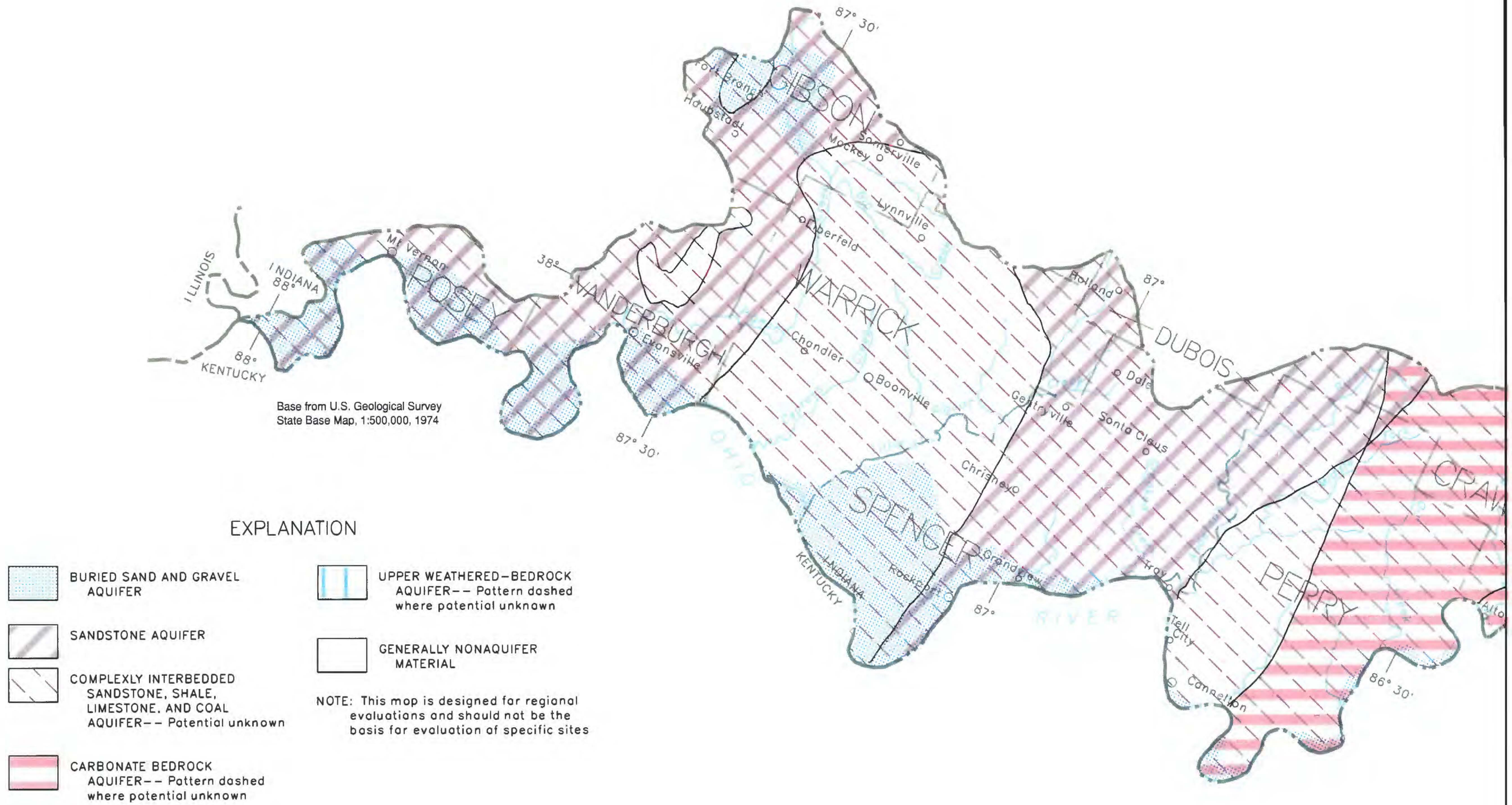


Figure 83. Extent of aquifer types in the Ohio River basin.

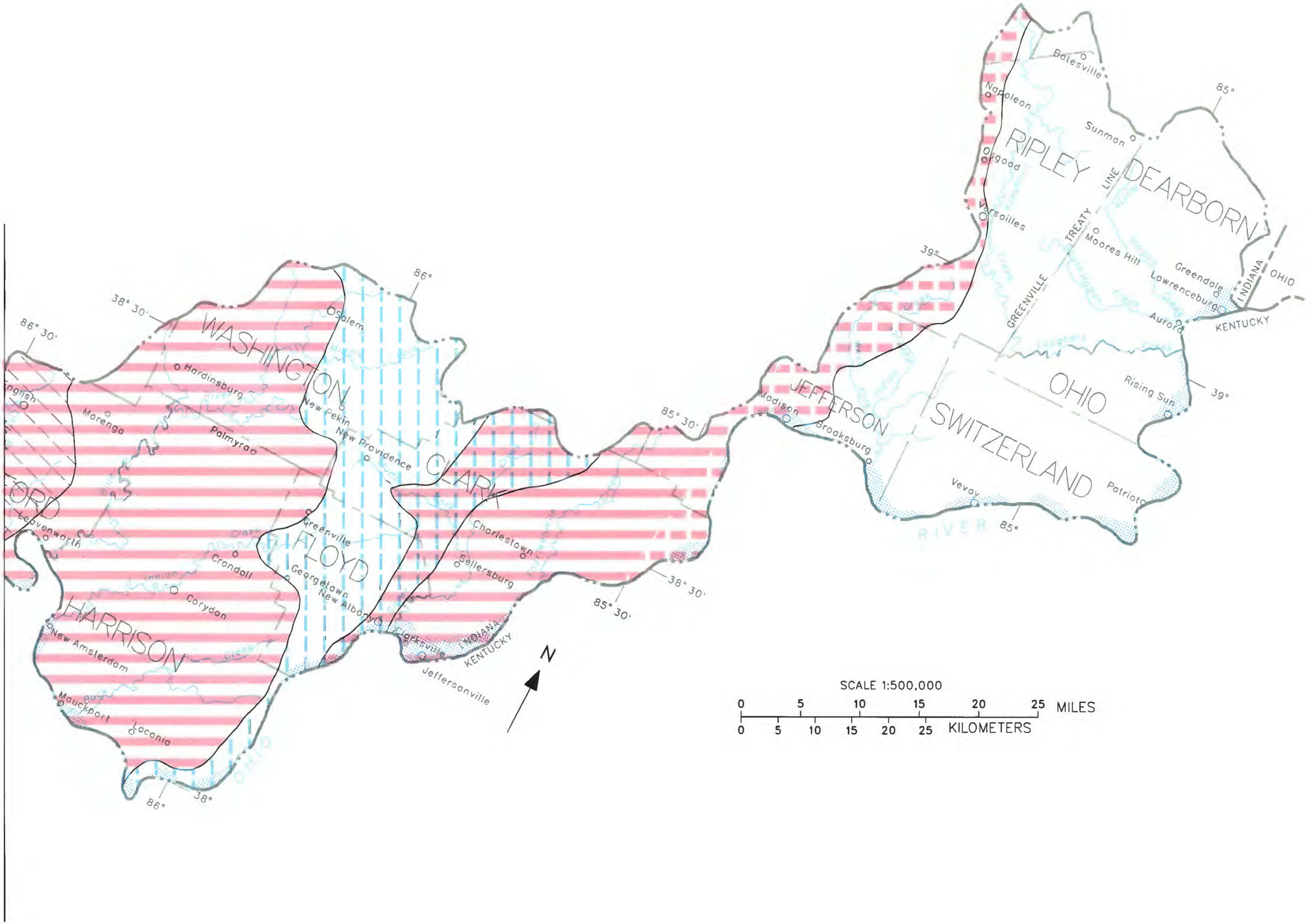

WASHINGTON
CLARK
FLOYD
HARRISON
JEFFERSON
SWITZERLAND
OHIO
RIPLEY
DEARBORN
Salem
Hardinsburg
New Pekin
New Providence
Morengo
Palmyra
Leavenworth
Crandall
Corydon
Greenville
Georgetown
New Albany
Clarksville
Jeffersonville
Charlestown
Sellersburg
New Amsterdam
Mauckport
Laconia
Madison
Brooksburg
Vevay
Patriot
Rising Sun
Aurora
Lawrenceburg
Greendale
Moores Hill
Versailles
Osgood
Napoleon
Batesville
Sunman
GREENVILLE
TREATY
LINE
INDIANA
OHIO
KENTUCKY
RIVER
OHIO
SCALE 1:500,000
0 5 10 15 20 25 MILES
0 5 10 15 20 25 KILOMETERS
N

Table 14. Characteristics of aquifer types in the Ohio River basin
[Locations of aquifer types shown in fig. 83]

Aquifer type	Thickness (feet)	Range of yield (gallons per minute)	Common name(s)
Buried sand and gravel	35-150	[1,2,3,4,5]10-1,000	Ohio River sand and gravel[1]; valley train deposits[3]; alluvial deposits[5]
Sandstone	10- 50	[1,2,3,4,6]1- 75	Inglefield Sandstone Member[1]; Linton, Dugger, and Patoka aquifers[2]
Complexly interbedded sandstone, shale limestone, and coal	highly variable[7]	0- 20	West Baden, Stephensport, Buffalo Wallow, Raccoon Creek, Carbondale, and McCleansboro Groups[8]
Carbonate bedrock			
Mississippian	100-600	[1,4,6]1- 100	Blue River and Sanders Groups
Silurian-Devonian	50-250	[1,4,6]0- 20	Muscatatuck Group through Louisville Limestone[8]
Upper weathered bedrock	50- 100	[1,4,6]0- 5	New Albany Shale and Borden Group[8]

[1]Clark, 1980.
[2]Cable and Wolf, 1977.
[3]Robison, 1977.
[4]Bechert and Heckard, 1966.
[5]Gallaher and Price, 1966.
[6]Pettijohn and Reussow, 1969.
[7]Water commonly found in thin beds within complexly interbedded sequence.
[8]Shaver and others, 1986.

Bechert, C.H., and Heckard, J.M., 1966, Ground water, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 100-115.

Cable, L.W., and Wolf, R.J., 1977, Ground-water resources of Vanderburgh County: Indiana Department of Natural Resources, Division of Water Bulletin 38, 37 p.

Clark, G.D., ed., 1980, The Indiana water resource—availability, uses, and needs: Indianapolis, Governor's Water Resource Study Commission, Indiana Department of Natural Resources, 508 p.

Gallaher, J.T., 1963a, Geology and hydrology of alluvial deposits along the Ohio River in the Hawesville and Cloverport areas, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-72.

_____1963b, Geology and hydrology of alluvial deposits along the Ohio River in Lewisport and Owensboro areas, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-74, 2 sheets.

_____1963c, Geology and hydrology of alluvial deposits along the Ohio River in the Spottsville and Reed areas, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-96, 2 sheets.

_____1964a, Geology and hydrology of alluvial deposits along the Ohio River in the Henderson area, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-91, 2 sheets.

_____1964b, Geology and hydrology of alluvial deposits along the Ohio River in the Stanley area, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-110, 2 sheets.

Gallaher, J.T., and Price, W.E., Jr., 1966, Hydrology of the alluvial deposits in the Ohio River valley in Kentucky: U.S. Geological Survey Water-Supply Paper 1818, 80 p.

Geosciences Research Associates, Inc., 1982, Hydrogeologic atlas of Indiana: Bloomington, Ind., 31 pl.

Gray, H.H., 1972, Lithostratigraphy of the Maquoketa Group (Ordovician) in Indiana: Indiana Department of Natural Resources, Geological Survey Special Report 7, 31 p.

_____1982, Map of Indiana showing topography of the bedrock surface: Indiana Department of Natural Resources, Geological Survey, Miscellaneous Map 36, scale 1:500,000.

_____1983, Map of Indiana showing thickness of unconsolidated deposits: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 38, scale 1:500,000.

_____1989, Quaternary geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 49, scale 1:500,000.

Gray, H.H., Ault, C.H., and Keller, S.J., 1987, Bedrock geologic map of Indiana: Indiana Department of Natural Resources, Geological Survey Miscellaneous Map 48, scale 1:500,000.

Greeman, T.K., 1981, Lineaments and fracture traces, Jennings County and Jefferson Proving Ground, Indiana: U.S. Geological Survey Open-File Report 81-1120, 17 p.

_____1983, Lineament and fracture traces, Decatur County, Indiana: U.S. Geological Survey Open-File Report 82-918, 17 p.

Gutschick, R.C., 1966, Bedrock geology, *in* Lindsey, A.A., ed, Natural features of Indiana: Indianapolis, Indiana Academy of Science, p. 1-20

Harrell, Marshall, 1935, Ground water in Indiana: Indiana Department of Conservation, Division of Geology Publication 133, 504 p.

Harvey, E.J., 1956, Geology and ground-water resources of the Henderson Area Kentucky: U.S. Geological Survey Water-Supply Paper 1356, 227 p.

Hasenmueller, N.R., and Bassett, J.L., 1980, Map of Indiana showing structure on top of Silurian rocks: Indiana Department of Natural Resources Geological Survey, METC/EGSP Series 811.

Malott, C.A., 1922, The physiography of Indiana, *in* Handbook of Indiana geology: Indiana Department of Conservation, Publication 20, part 2, p. 59-256.

Pettijohn, R.A., and Reussow, J.P., 1969, Reconnaissance of the ground-water resources of the lower Ohio River basin in Indiana: U.S. Geological Survey, Open-File Report, 47 p.

Price, William E., Jr., 1964a, Geology and hydrology of alluvial deposits along the Ohio River between Newport and Warsaw, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-98, 2 sheets.

_____1964b, Geology and hydrology of alluvial deposits along the Ohio River between southwestern Louisville and West Point, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-111, 2 sheets.

_____1964c, Geology and hydrology of alluvial deposits along the Ohio River between Prospect and southwestern Louisville, Kentucky: U.S. Geological Survey Hydrologic Investigations Atlas HA-130, 2 sheets.

Robison, Tully M., 1977, Ground-water resources of Posey County, Indiana: Indiana Department of Natural Resources, Division of Water Bulletin 39, 27 p.

Schneider, A.F., 1966, Physiography, *in* Lindsey, A.A., ed., Natural features of Indiana: Indianapolis, Indiana Academy of Science p.40-56.

Shaver, R.H.; Ault, C.H.; Burger, A.M.; Carr, D.D.; Droste, J.B.; Eggert, D.L.; Gray, H.H.; Harper, Denver; Hasenmueller, N.R.; Hasenmueller, W.A.; Horowitz, A.S.; Hutchison, H.C.; Keith, B.D.; Keller, S.J.; Patton, J.B.; Rexroad, C.B.; and Wier, C.E.; 1986, Compendium of Paleozoic rock-unit stratigraphy in Indiana—a revision: Indiana Department of Natural Resources, Geological Survey Bulletin 59, 203 p.

DEFINITIONS OF SELECTED TERMS

The following are definitions of selected terms as they are used in this report; they are not necessarily the only valid definitions of these terms.

Alluvial deposits (or alluvium). Unconsolidated sediment deposited in river channels or on flood plains by a stream.

Aquifer. A formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield quantities of potable water for domestic purposes. An aquifer may include unsaturated parts of the permeable material.

Aquifer—Potential unknown. An aquifer-type classification that implies a formation of unknown or poor production capabilities. The use of the formation as a water supply may result in low yields and(or) dry holes.

Bedrock aquifer (or consolidated aquifer). An aquifer composed of limestone, dolostone, sandstone, coal, siltstone, or shale bedrock.

Buried aquifer. A sand and gravel aquifer whose upper surface is greater than 10 feet beneath the land surface; a buried aquifer is not discontinuous.

Confined aquifer. An aquifer whose potentiometric surface is higher than the top of the aquifer.

Discharge area. An area where water is lost from an aquifer; commonly a surface-water body or an area of intensive ground-water pumping.

Discontinuous aquifer. A sand and gravel aquifer composed of small, detached sand and gravel deposits that are less than about 15 square miles in extent, and separated from other aquifers by non-aquifer material. Discontinuous aquifers can be either surficial or buried.

Drift. A general term for all material transported by glacial processes and deposited directly from melting ice or by meltwater streams.

Dry hole. A hole abandoned during drilling for lack of water.

Hydraulic head (or static head). The height of the surface of a column of water above a standard datum that can be supported by the static pressure at a given point; the sum of the elevation head and the pressure head; the level to which water will rise in a properly constructed well.

Loess. A blanket of fine-grained material, typically silt, deposited by the wind.

Nonaquifer material. Sediments with low hydraulic conductivity that normally will not transmit quantities of potable water adequate for domestic purposes.

Outwash. Stratified unconsolidated material, typically sand and gravel deposited by meltwater streams flowing beyond the glacial ice; proglacial stratified drift.

Outwash plain. A broad, gently sloping sheet of outwash.

Porosity. The ratio of the volume of the voids or openings in a rock to its total volume.

Potentiometric surface (generalized). An imaginary surface representing the total head of ground water in an aquifer and defined by the level to which water will rise in a properly constructed well.

The generalized potentiometric surface of unconsolidated aquifers usually represents the static head from a discrete screened interval and generally is a subdued reflection of the land surface. The generalized potentiometric surface of bedrock deposits is typically a composite water level from a hole open through many lithologies.

Recharge. Water that is gained by an aquifer.

Saturated. The condition in which the pores of a material are filled with water.

Surficial aquifer. A type of aquifer whose upper surface is within 10 feet of the land surface; a surficial aquifer is not discontinuous.

Till. An unsorted, unstratified sediment deposited directly by glaciers with little or no reworking by meltwater. Till is composed of clay, silt, sand, gravel and(or) boulders. The term till is used in place of diamicton in the text.

Unconfined aquifer (or water-table aquifer). An aquifer whose upper surface is the water table.

Unconsolidated aquifer. A type of aquifer composed of sand, gravel, or a mixture of sand and gravel.

Valley train. A long, narrow body of outwash deposited by meltwater streams far beyond the margin of active glaciation, and confined laterally within a valley.

Yield (or potential well yield). The maximum pumping rate that can be sustained in a well without lowering the water level below the water intake. The maximum potential well yield is supplied by a properly constructed, fully penetrating, large diameter well.

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