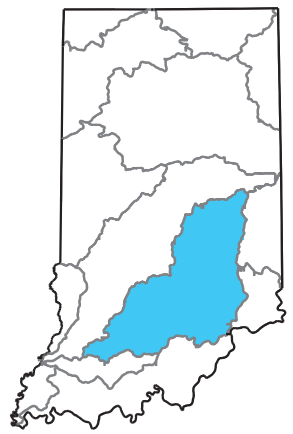


EAST FORK WHITE RIVER BASIN

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).

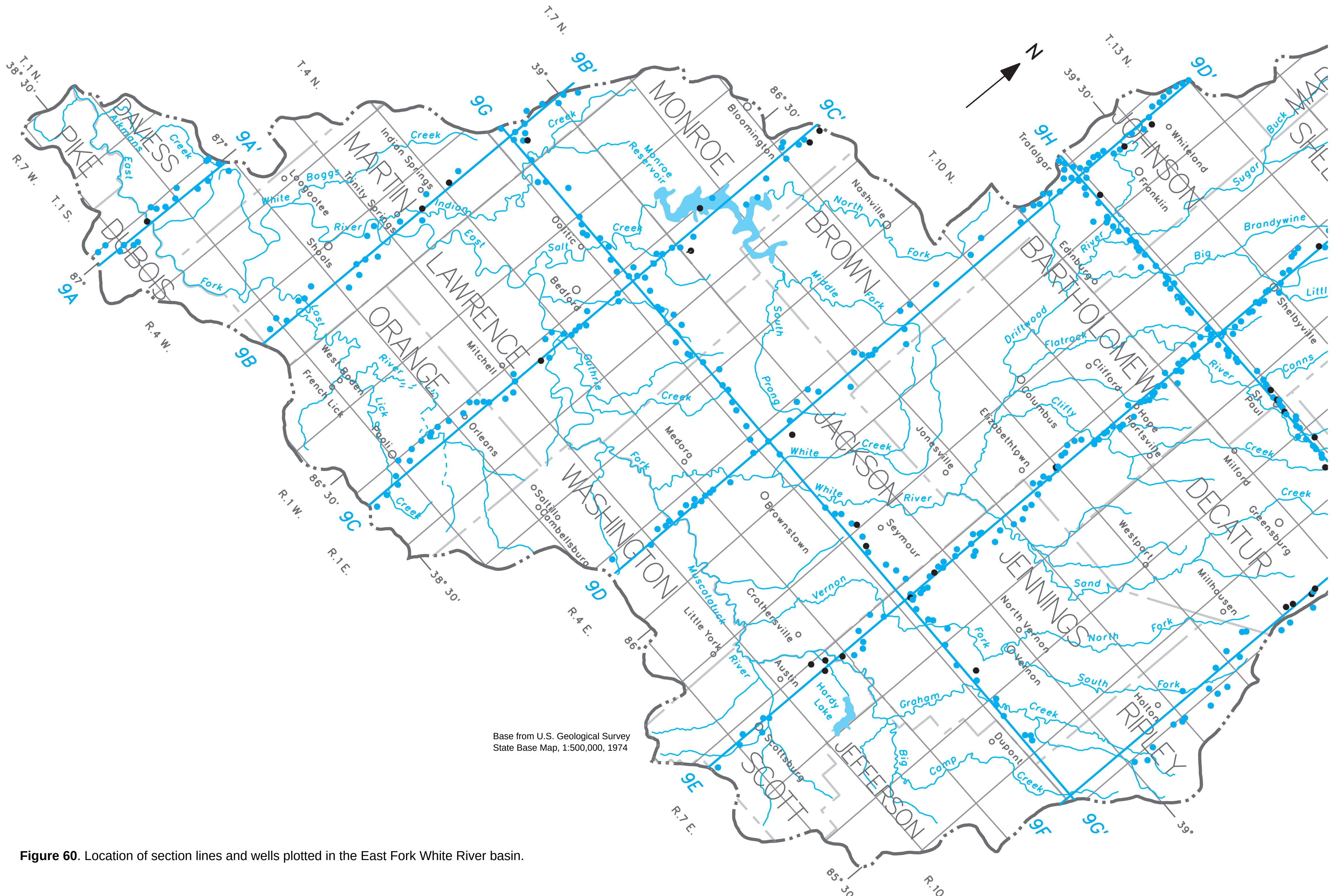
The easternmost physiographic unit in the East Fork White River basin is the Muscatatuck Regional Slope. The eastern boundary of the physiographic unit roughly coincides with the eastern boundary of the drainage basin. The Muscatatuck Regional Slope has a westerly dip of approximately 400 ft over 25 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.)

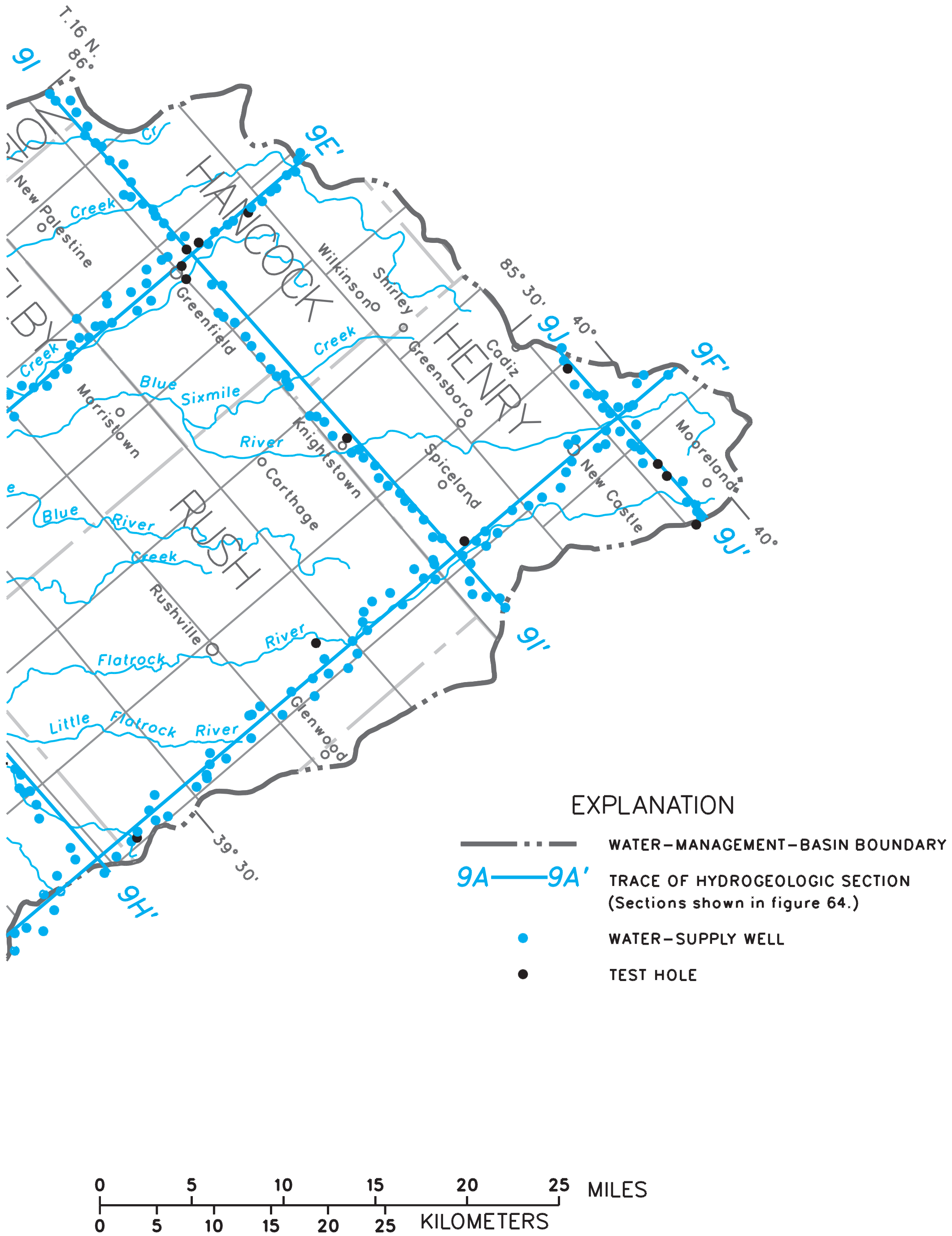
The Mitchell Plain, lying to the west of the Norman Upland, is underlain by Mississippian limestones. The area is a low-relief karst plain that is intensely pitted in some areas by thousands of sinkholes. Surface drainage is poorly developed because of the extensive internal drainage. Most of the precipitation 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).



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

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. Landsurface 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

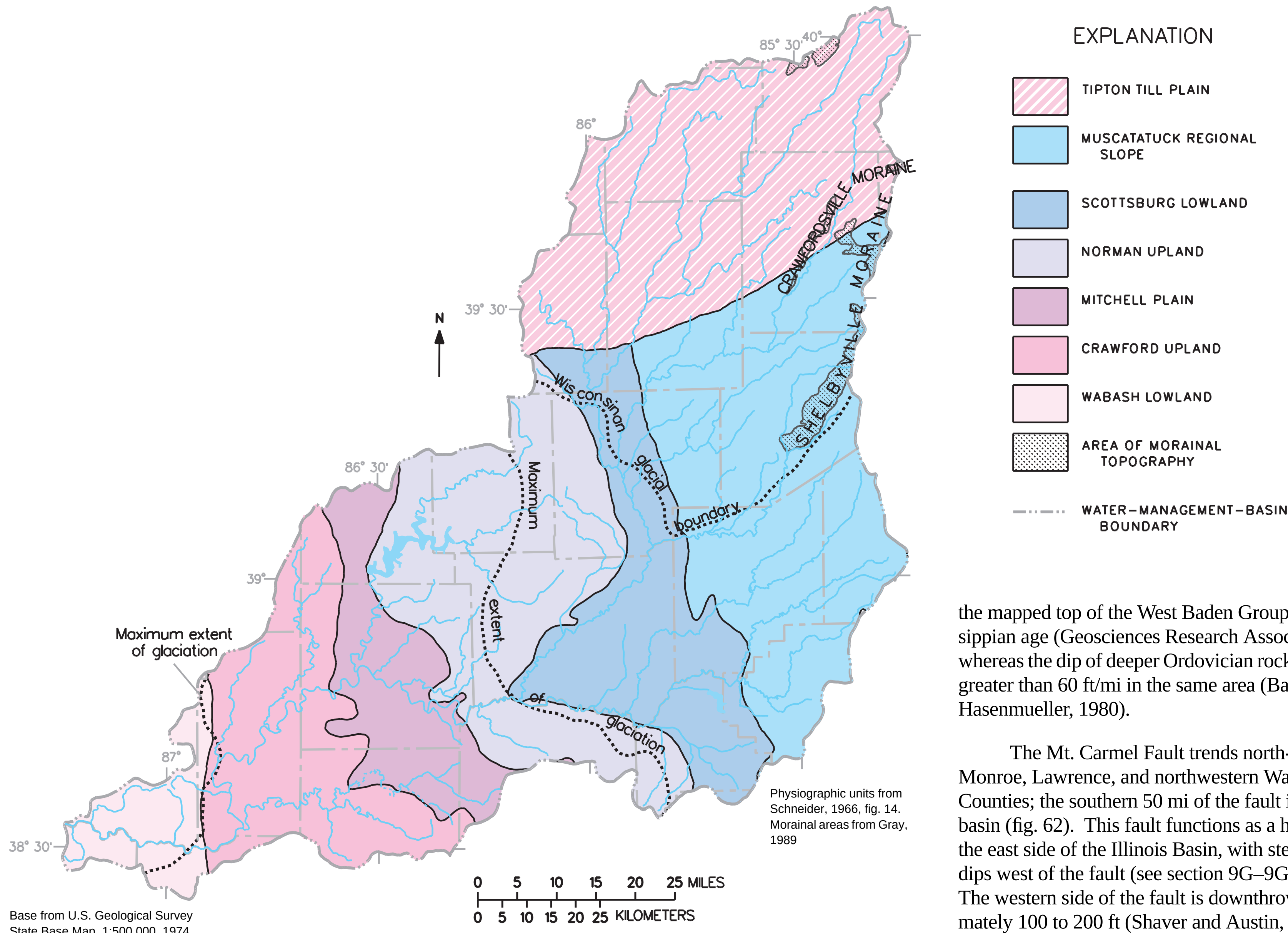


Figure 61. Physiographic units, moraines, and extent of glaciation in the East Fork White River basin.

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

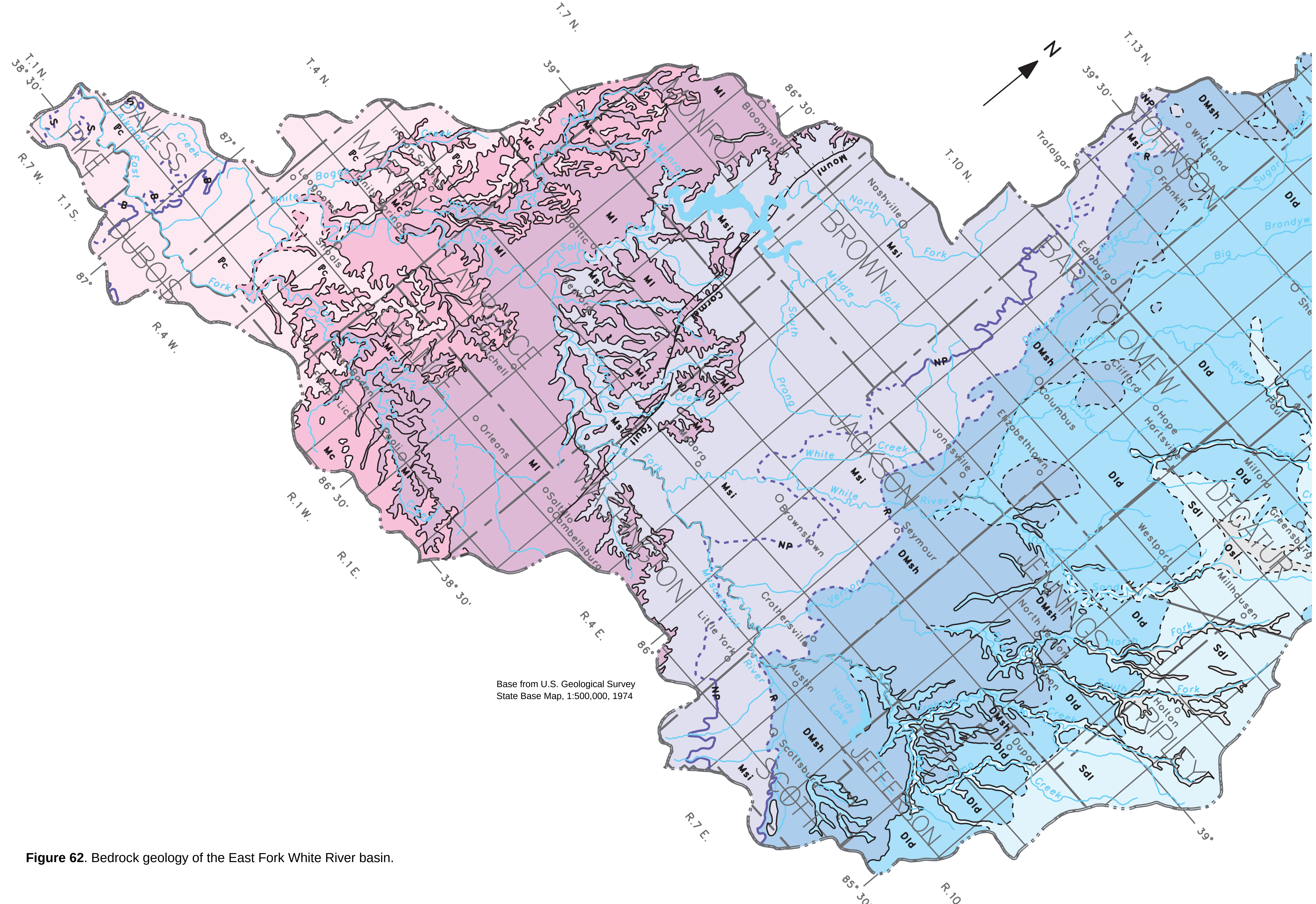
More than two-thirds of the East Fork White River basin was glaciated during the Pleistocene Period. Pre-Wisconsinan glaciers covered the northeastern two-thirds and the extreme downstream end of the basin (fig. 61). Wisconsinan ice overrode the earlier glacial deposits in the northeastern one-third 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).

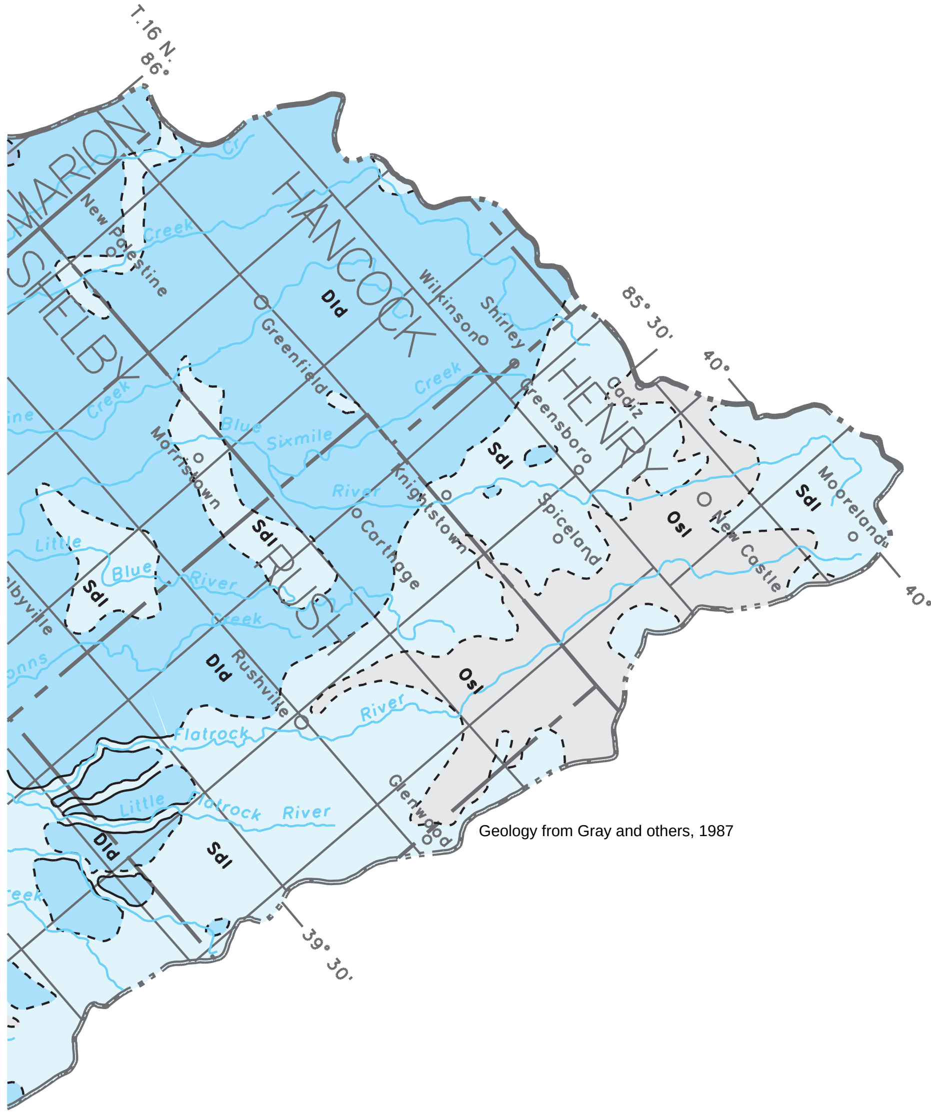
In general, the unconsolidated deposits in the central one-third of the basin are less than 50 ft thick (fig. 63). Exceptions can be found along the East Fork White River and part of the Muscatatuck River, where thicknesses range from 50 to more than 100 ft. The East Fork White River flows in a 3-mi-wide 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.

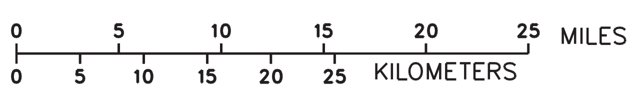
The northern one-third of the basin was initially glaciated by pre-Wisconsinan glaciers that deposited thick tills and some outwash of the Jessup Formation. During the Wisconsinan Age, loam till of the Trafalgar Formation was deposited in the basin. The Wisconsinan tills were deposited by the Huron-Erie 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



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

Figure 62. Bedrock geology of the East Fork White River basin.





EXPLANATION

- ₽c — PENNSYLVANIAN COMPLEXLY INTERBEDDED SHALE AND SANDSTONE, WITH THIN BEDS OF LIMESTONE AND COAL-- Composed of the Racoon Creek and Carbondale Groups
- 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 Wabash and Pleasant Mills Formations, and the Salamonie Dolomite, Louisville Limestone, Cataract Formation, and the Brassfield Limestone
- Osl — ORDOVICIAN SHALE AND LIMESTONE-- Composed of the 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 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).

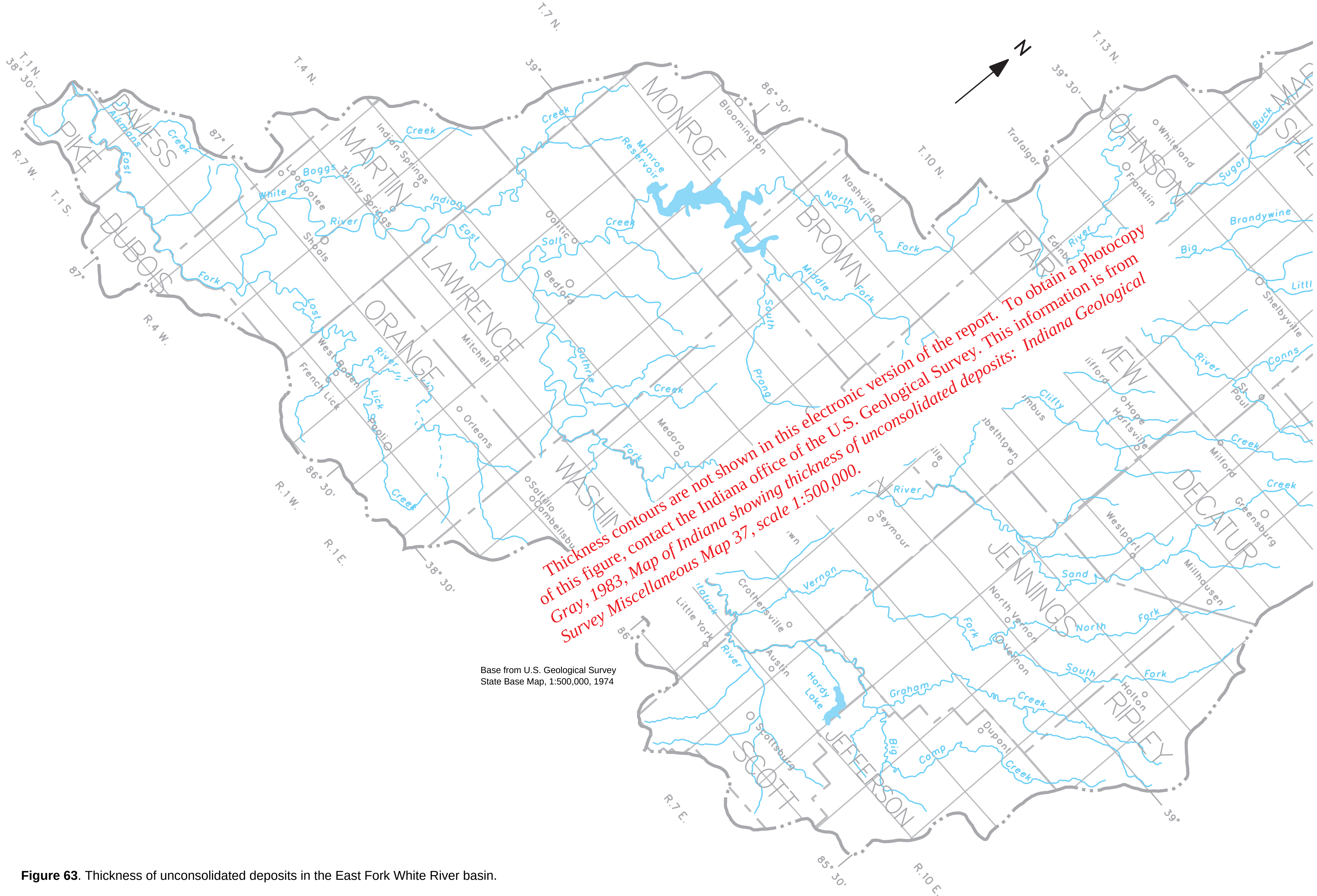
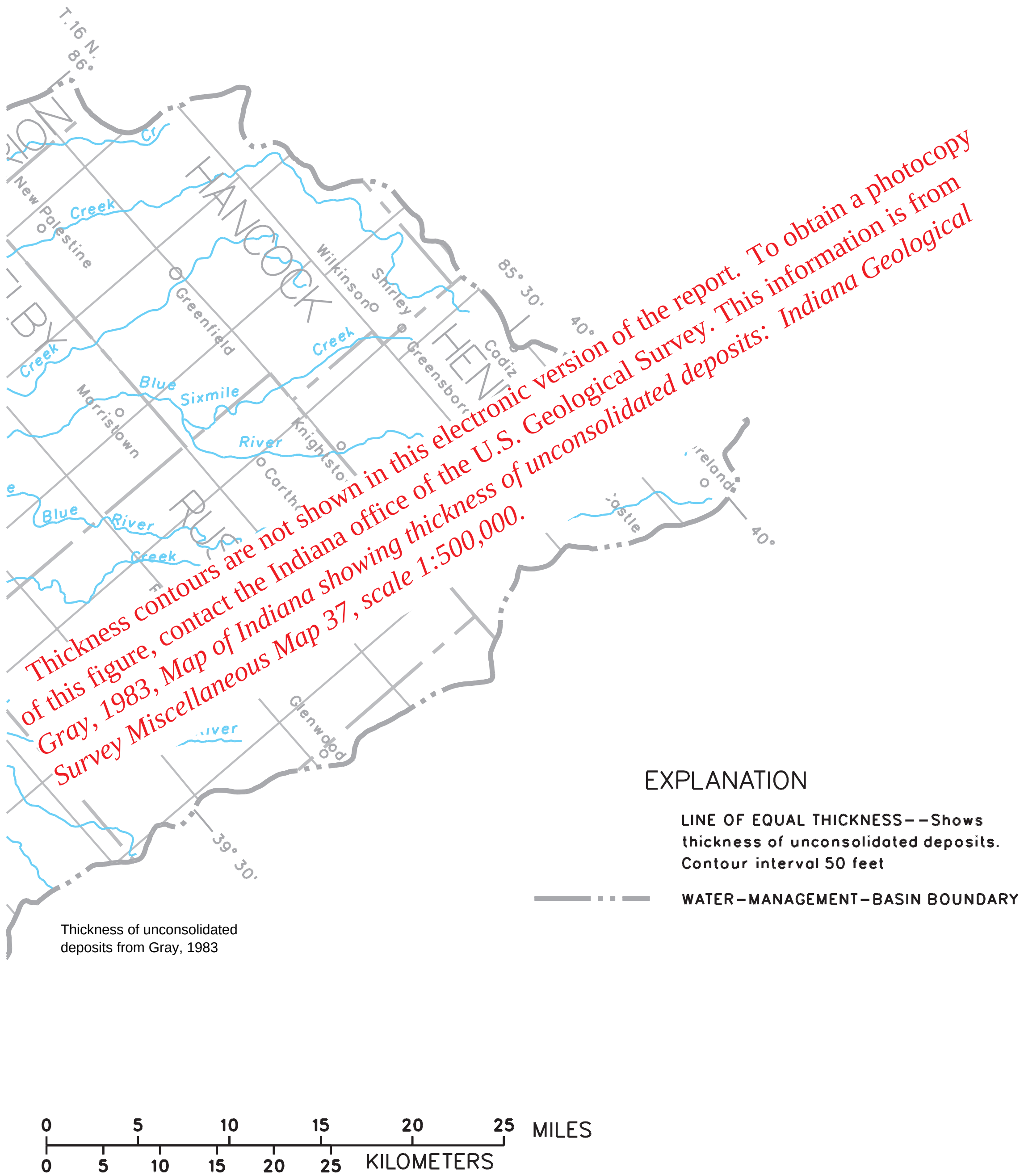


Figure 63. Thickness of unconsolidated deposits in the East Fork White River basin.



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).

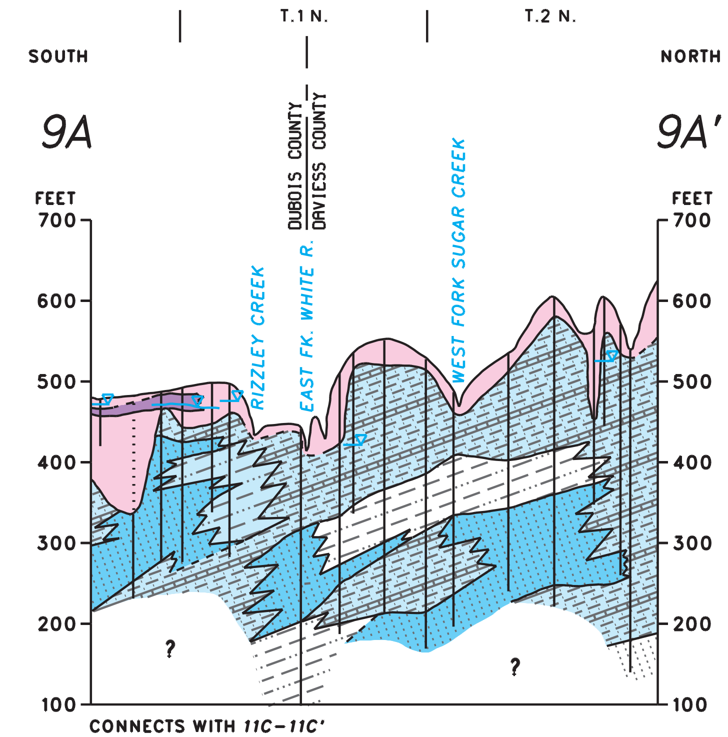
Buried Sand and Gravel Aquifers

Forming general horizons in the drift, buried sand and gravel aquifers are found in laterally continuous deposits covered by more than 10 ft of nonaquifer 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).



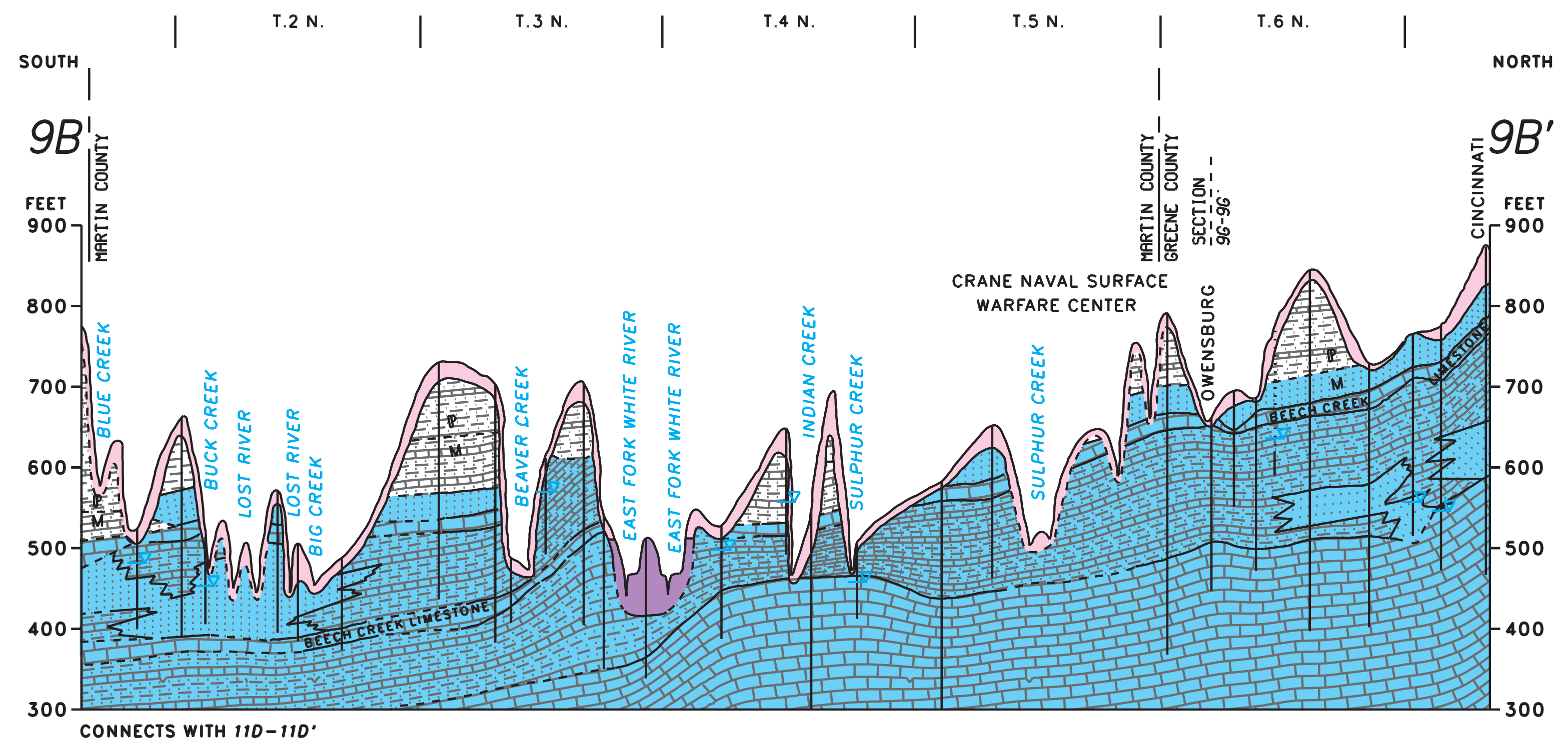
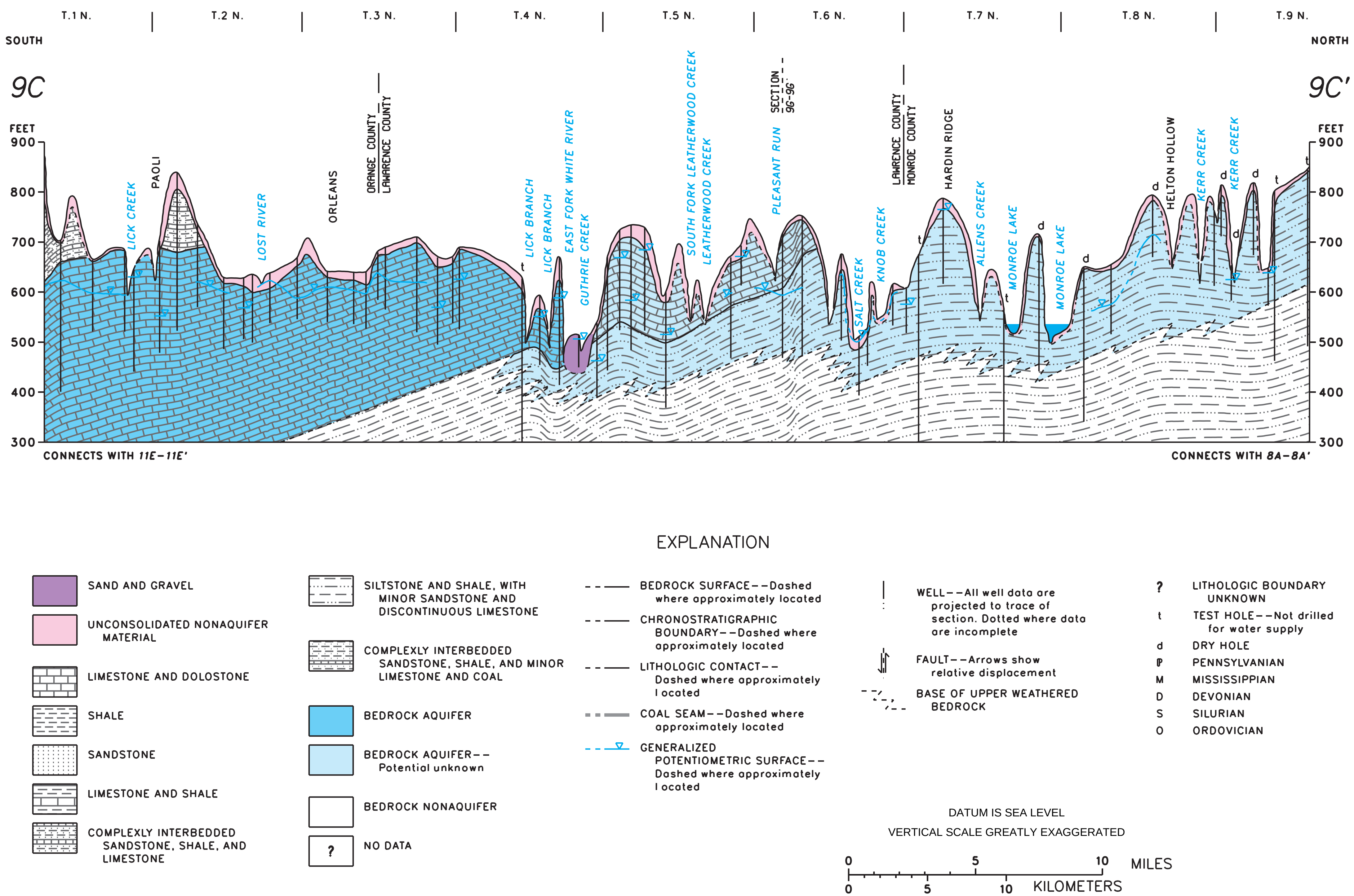


Figure 64. Hydrogeologic sections *9A–9A′* to *9J–9J′* of the East Fork White River basin.



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
PAOLI
LICK CREEK
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
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

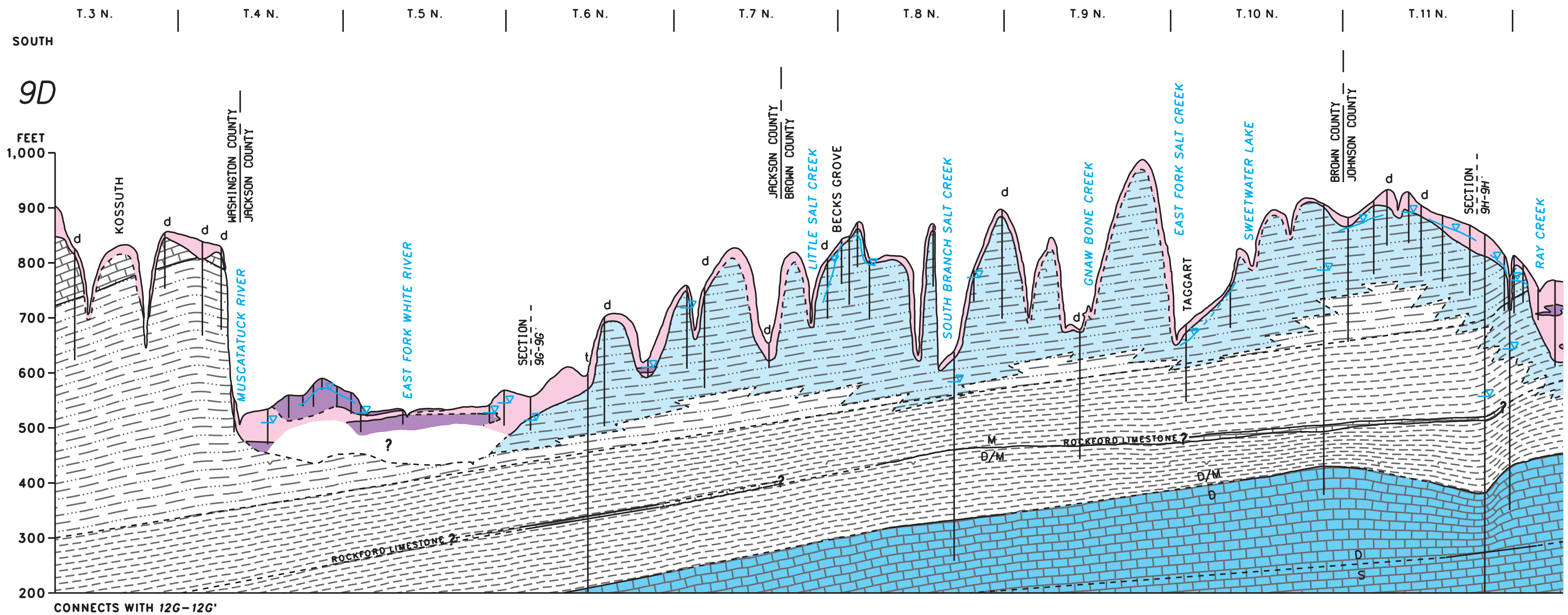


Figure 64. Hydrogeologic sections *9A–9A′* to *9J–9J′* of the East Fork White River basin—Continued.

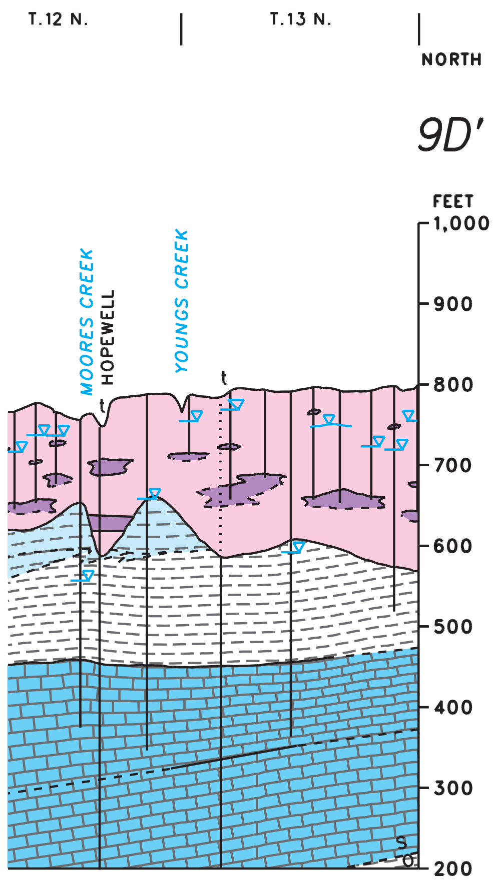
0 5 10 MILES
0 5 10 KILOMETERS

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).

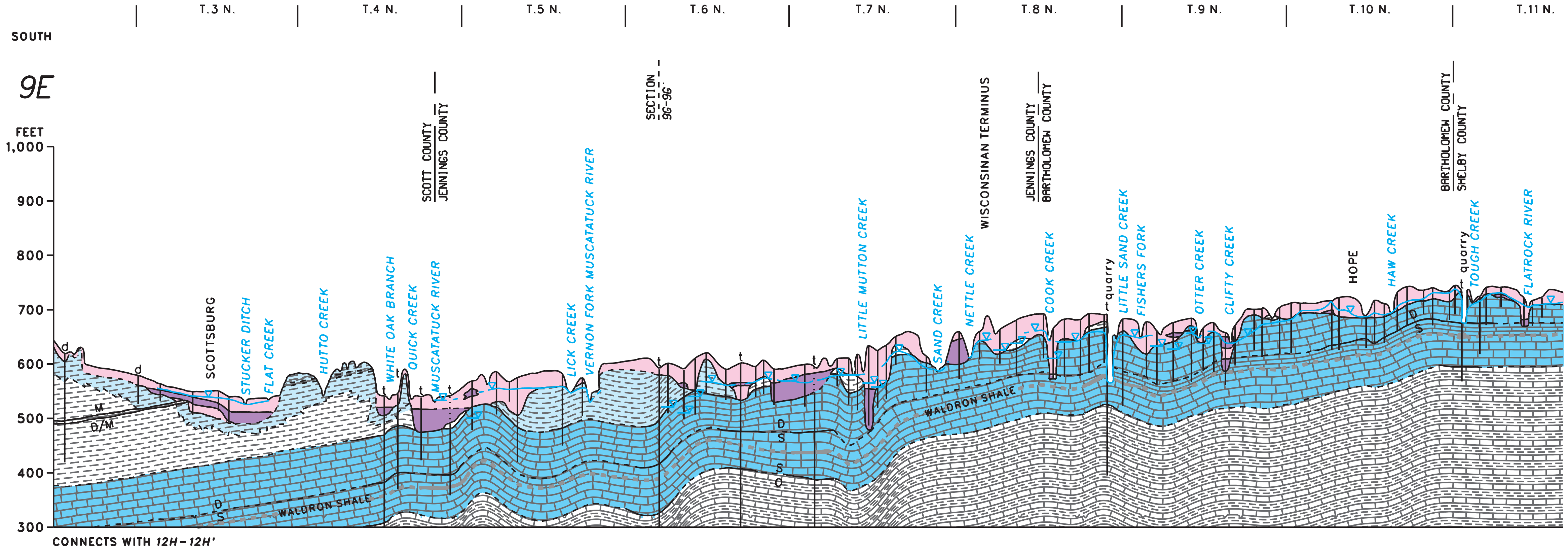
The thick Mississippian carbonate bedrock aquifer is composed primarily of relatively pure limestone, which is soluble in infiltrating precipitation. Carbonate dissolution has enlarged openings, forming underground channels within the aquifer. Typical well yields are 1 to 50 gal/min but can be as large 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.



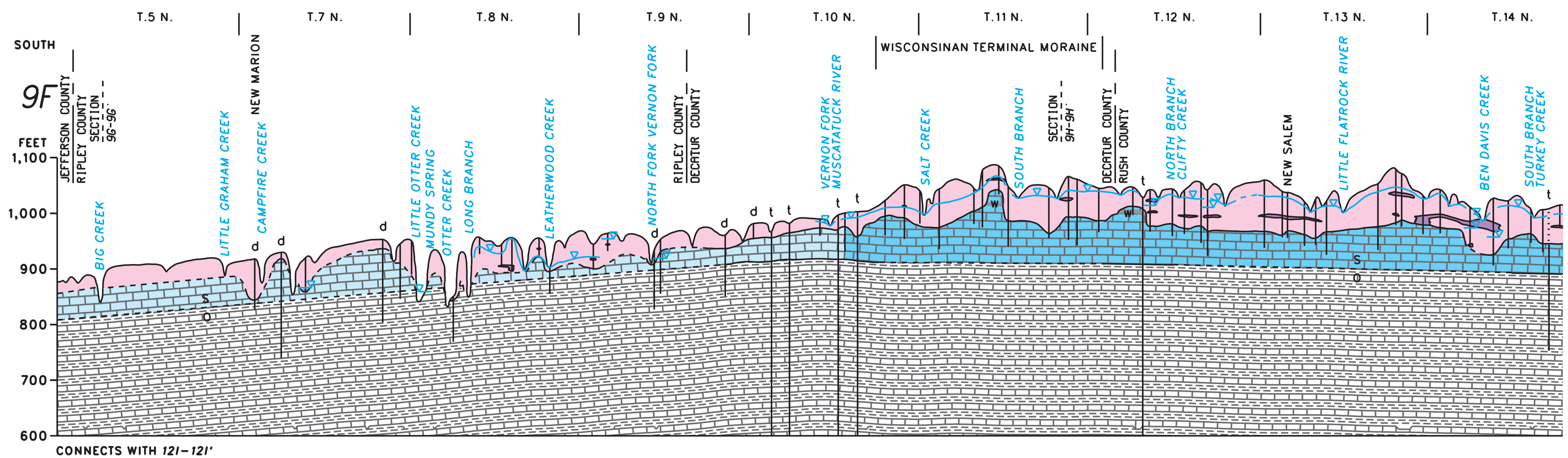
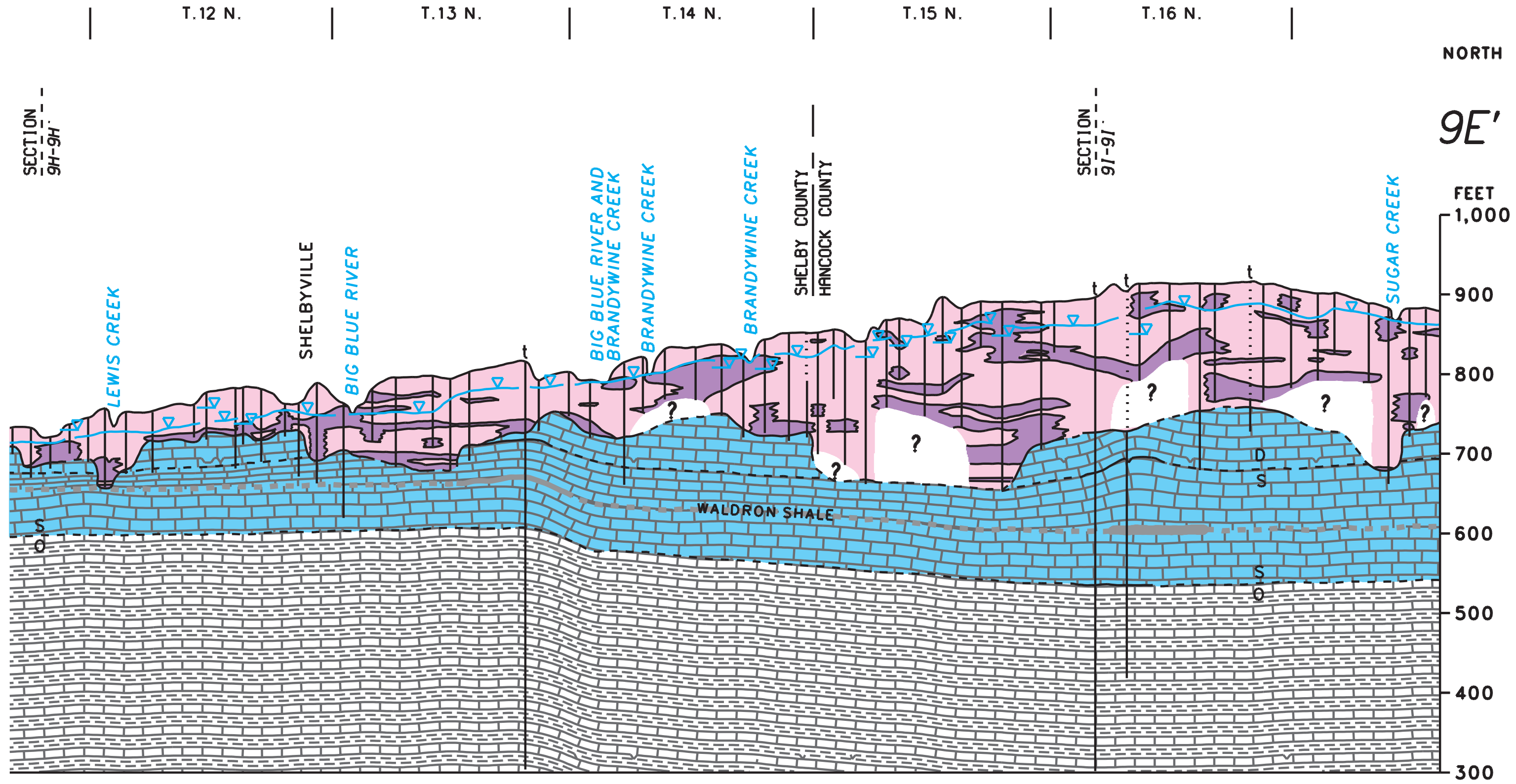
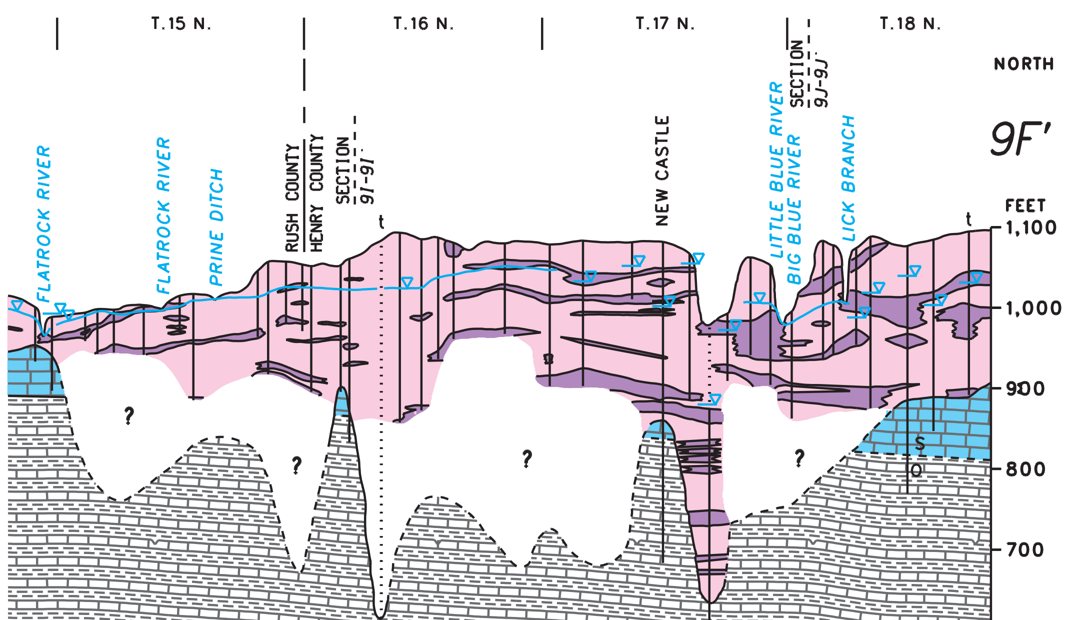


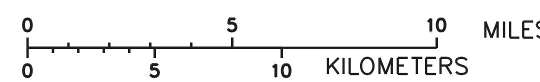
Figure 64. Hydrogeologic sections *9A–9A′* to *9J–9J′* of the East Fork White River basin—Continued.





DATUM IS SEA LEVEL

VERTICAL SCALE GREATLY EXAGGERATED

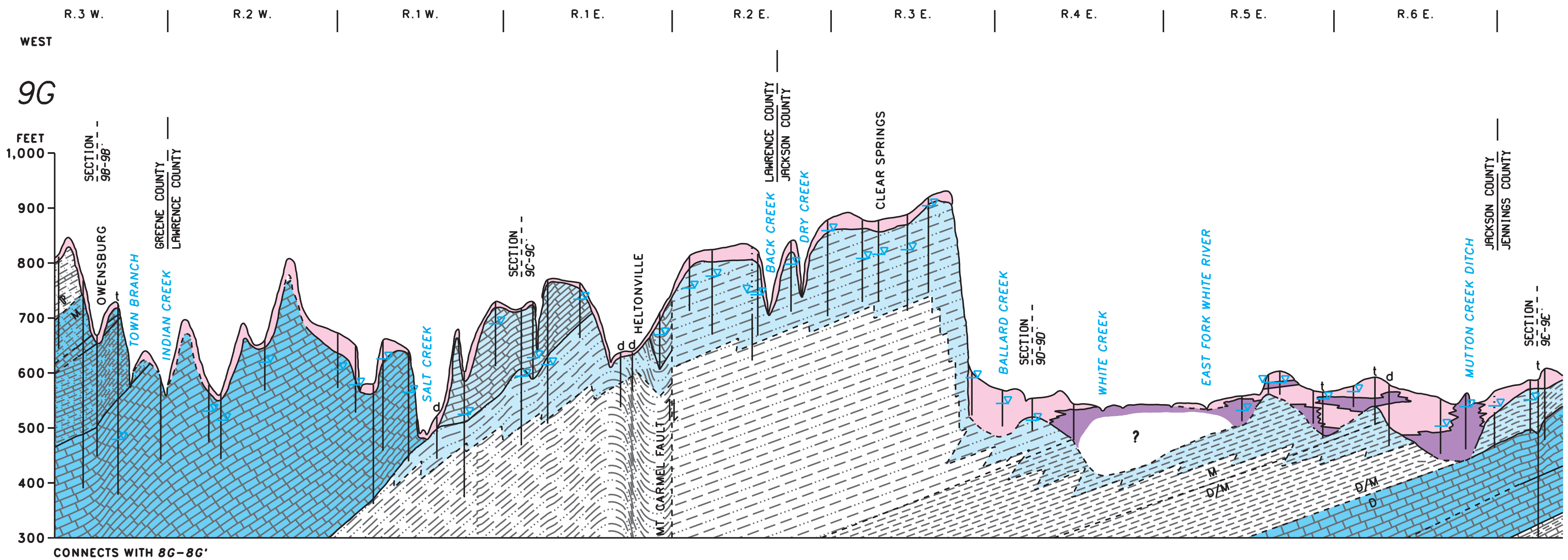


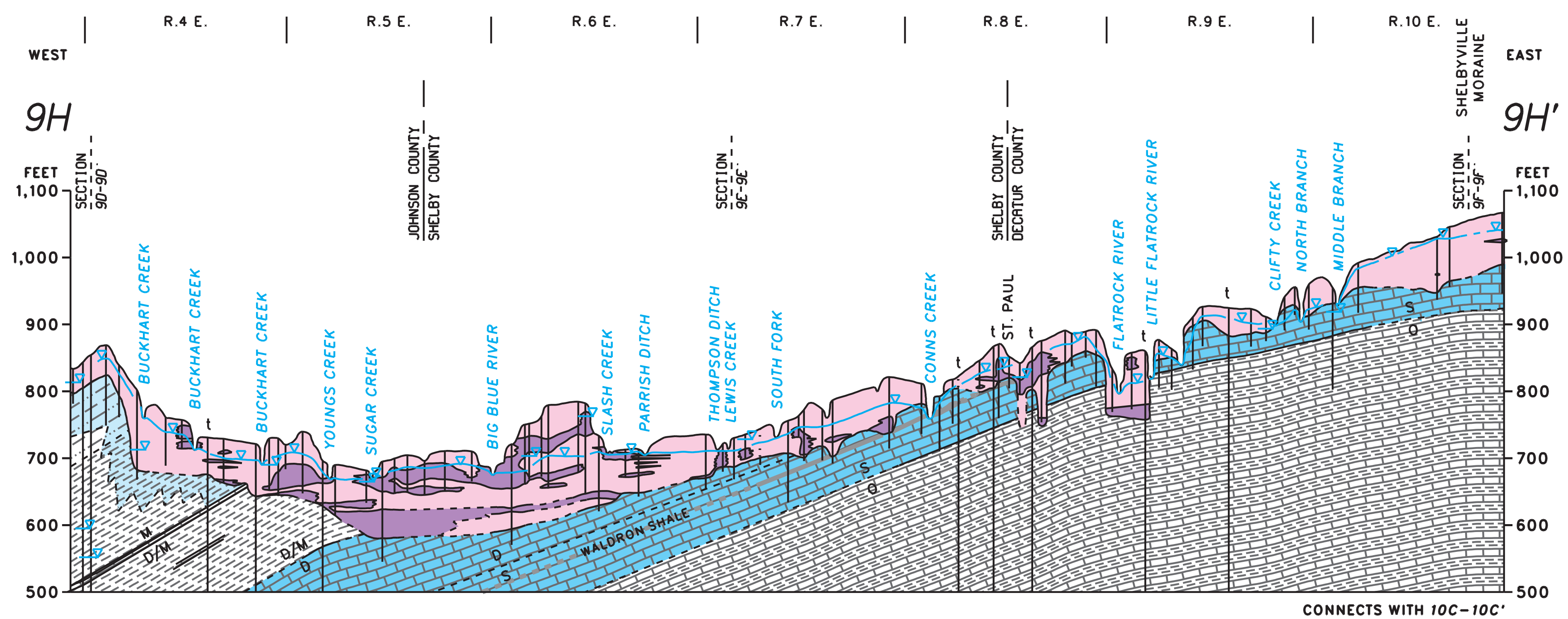
The area in the extreme southeastern part of the basin (Ripley County and eastern Decatur, Jennings, and Jefferson Counties) is mapped as "carbonate bedrock aquifer—potential unknown" because many drilled holes in the area are dry (fig. 65). This area is underlain by the lower carbonate bedrock aquifer 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





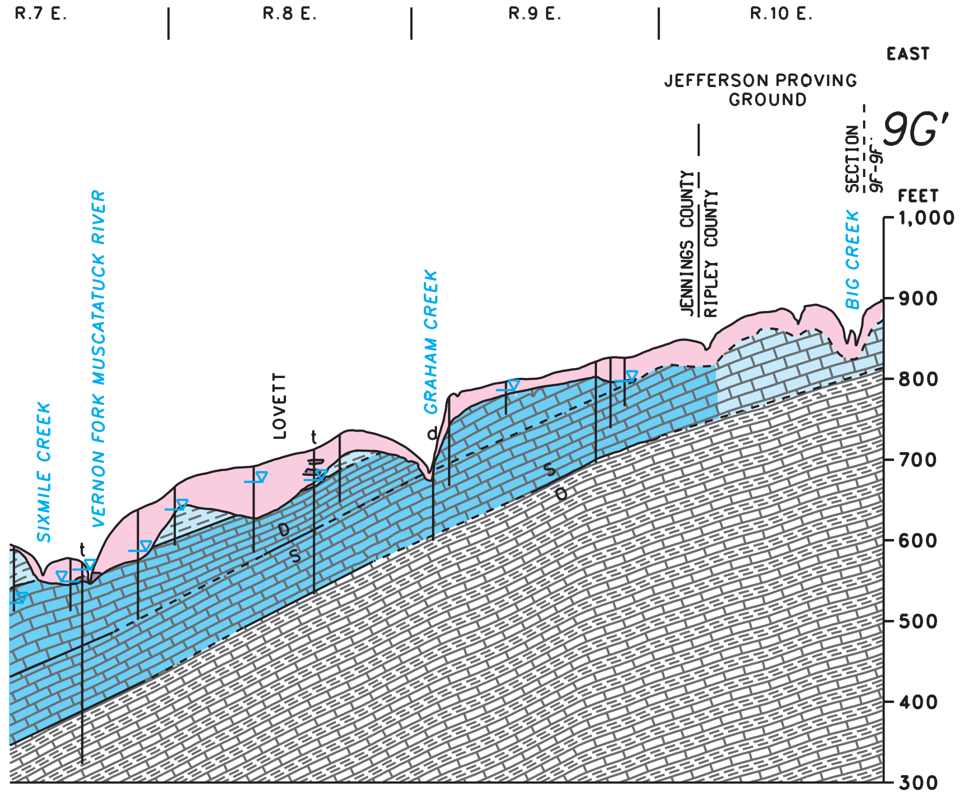
DATUM IS SEA LEVEL

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.

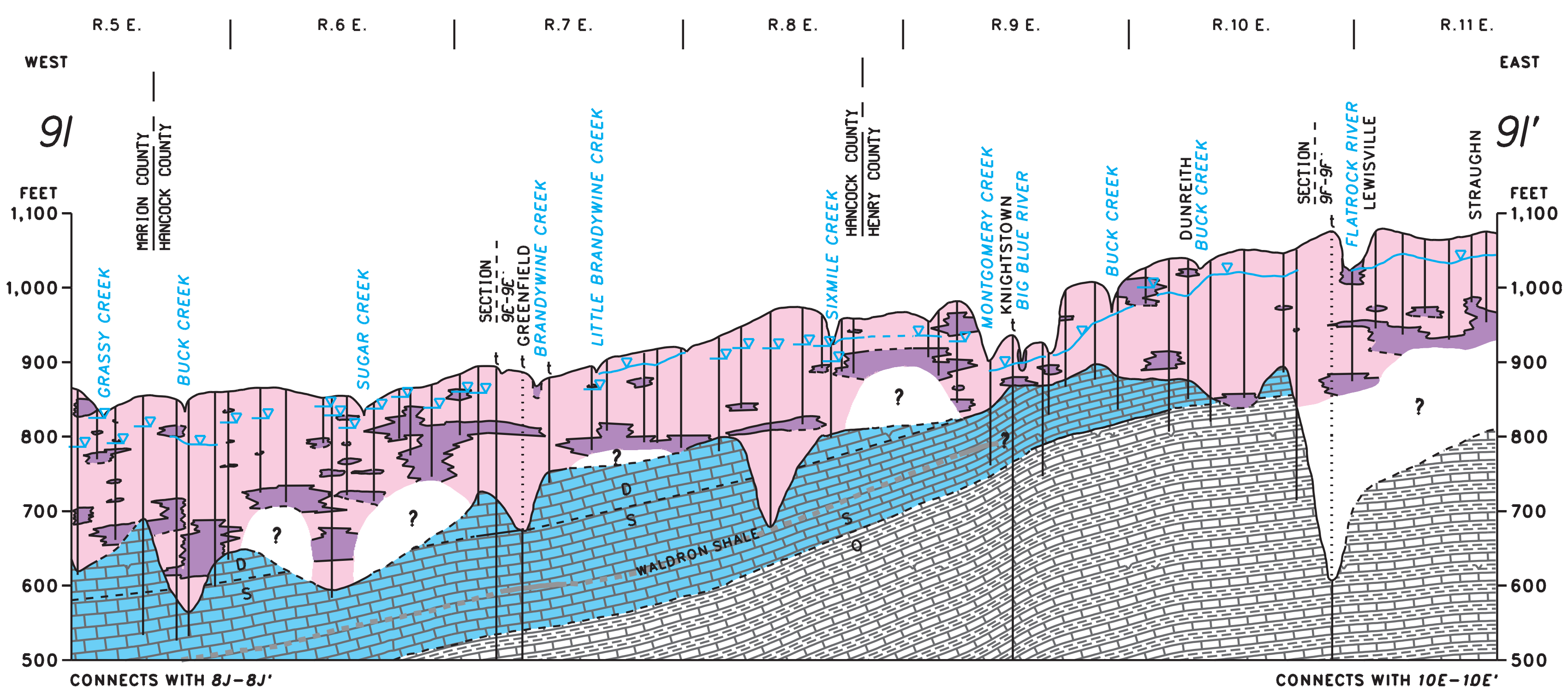


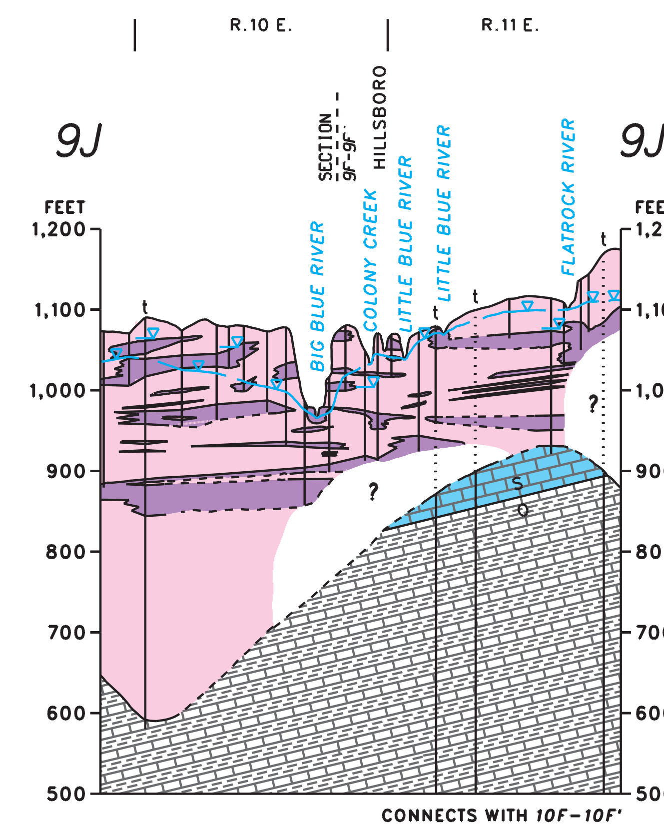
across the basin boundary. An example of ground water crossing the basin boundary can be seen on hydrogeologic section 9I–9I′ (fig. 64), where bedrock water-level elevations decline westward and are below most of the local streams. Ground water in the carbonate bedrock aquifer flows toward the White 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).





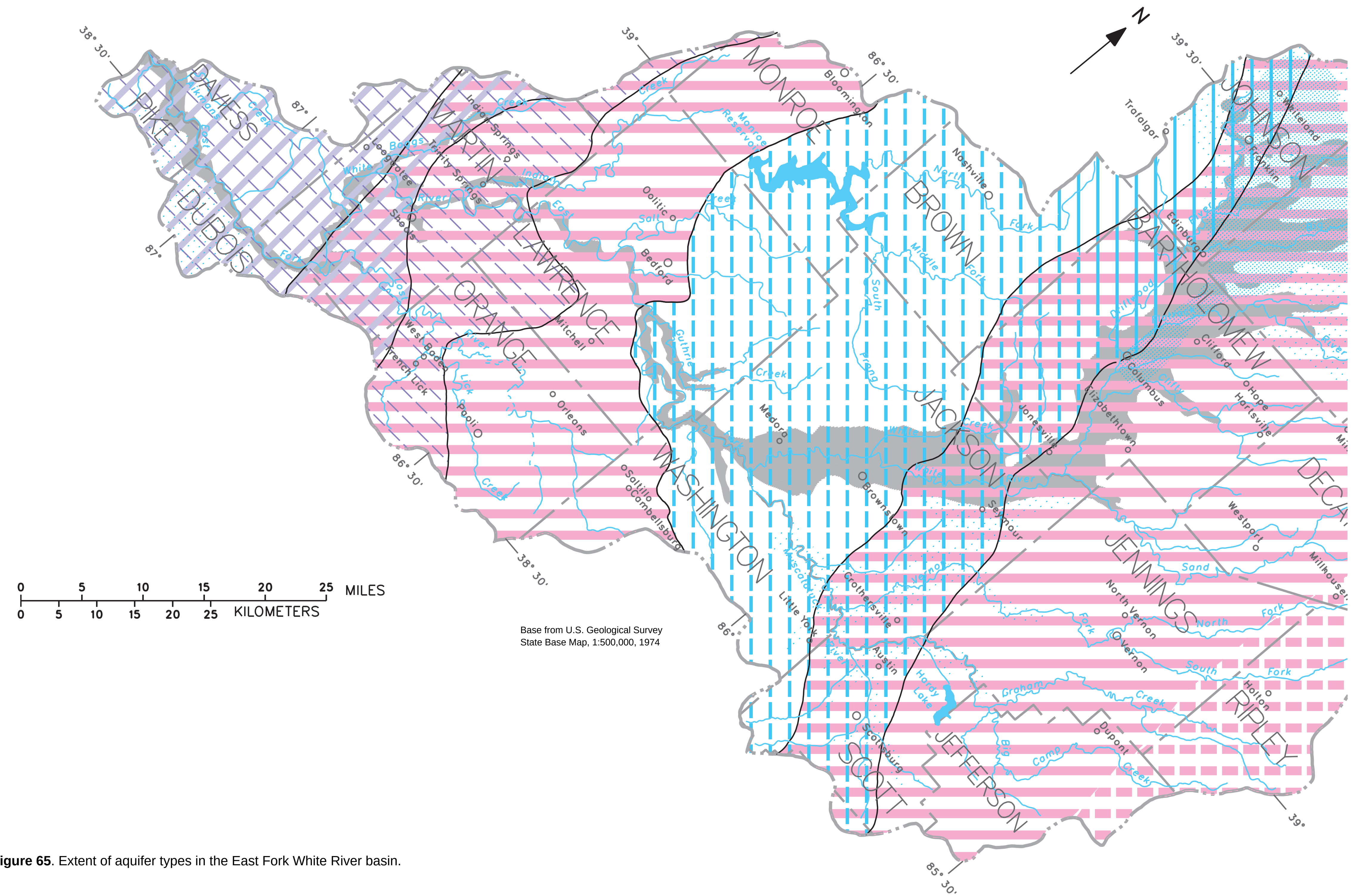


Figure 65. Extent of aquifer types in the East Fork White River basin.

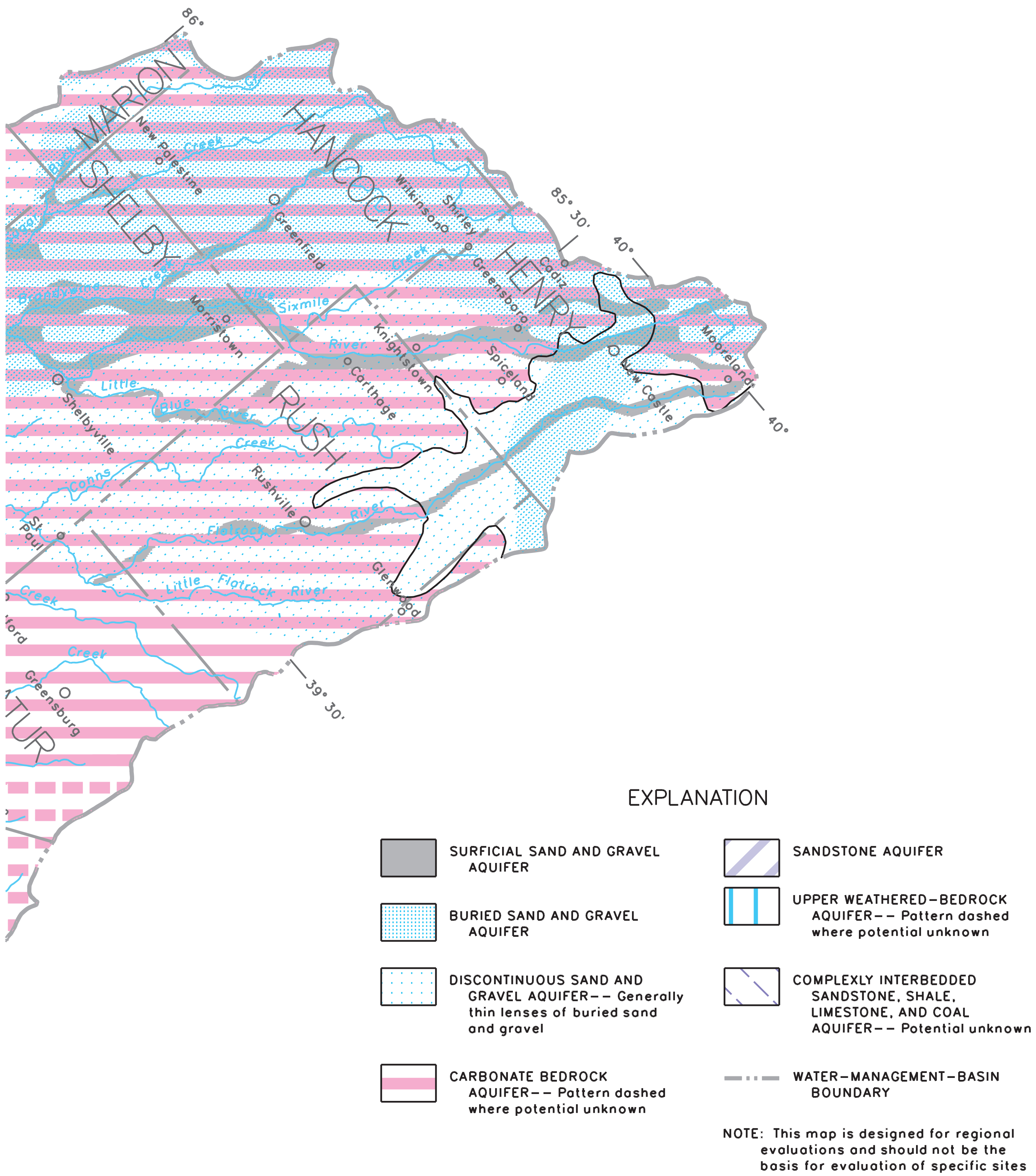


Table 11. Characteristics of aquifer types in the East Fork White River basin

[>, 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 non-aquifer, 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.

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