Kentucky Society of Professional Geologists
and the Kentucky Section of the
American Institute of Professional Geologists
2004 Joint Field Trip
October 21–23

The Bighill Exposure

And a Little Beyond

Compiled by Richard A. Smath

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On the Cover

Photographs of the new U.S. 421 highway under construction (September 1999) by Steven Henderson, Department of Highways

Section of the topographic map with new U.S. 421 by Willard Jackson, Department of Highways, Division of Planning

Photographs along the bottom:
Overlook at Windswept retreat by Richard A. Smath, Kentucky Geological Survey
Cloverbottom Limestone Mine and horses going through creek by Patrick Gooding, Kentucky Geological Survey

Contents

Thanks to:

- Belle Jackson, T.M.P., Executive Director of the Berea Tourism and Convention Commission
- Cheryl Chasteen, Administrative Assistant for the City of Berea Officials
- Vicki Spurlock, caterer
- Cane Break, musicians: Deborah Payne, Ryan Blevins, John Caudil, and Hollee Bragg

Field Trip Route and Physiography

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Deparment of Geological Sciences
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From the Kentucky Artisan Center at Berea (exit 77 off of I-75), we proceed east on Ky. 595. Proceed through Berea, and then turn left (east) onto Ky. 21 until it intersects U.S. 421 in Bighill. Turn right (south) onto U.S. 421 and proceed for 2 miles to the beginning of the exposure.

The area north of Berea is present wholly in the Lexington Plain or Bluegrass section of the Interior Low Plateaus physiographic province (Fenneman, 1938) (Fig. 1). The Lexington Plain largely coincides with apical parts of the *Jessamine Dome*, a structural culmination on the Cincinnati Arch. Because the dome is nearly symmetrical and truncated, lithologic units and resulting landforms form roughly concentric belts around the Lexington area, which is located near the center of the dome. Consequently, the Lexington Plain is divided into approximately concentric Inner and Outer Bluegrass belts. The Inner Bluegrass is relatively flat-lying to gently rolling and is the agriculturally richest part of the Lexington Plain. The area is largely underlain by soluble, phosphatic, Late Ordovician (Mohawkian) Lexington Limestone (Fig. 2), which produces very fertile soils. The flat terrain and fertile soils make this an agriculturally prosperous region with large-scale tobacco farming, grazing, as well as the breeding and rearing of horses. Karstic features are locally common, and the entire area may be a karstic solution plain.

The Inner Bluegrass typically gives way subtly to the Outer Bluegrass, an area of hummocky, irregularly rolling hills and low ridges underlain by Upper Ordovician shales and shaly limestones. Some workers divide the Outer Bluegrass into the Eden Belt and Outer Bluegrass proper. The Eden Belt is typically a highly dissected area with sharp, narrow ridge tops and steep-sided valleys developed on the

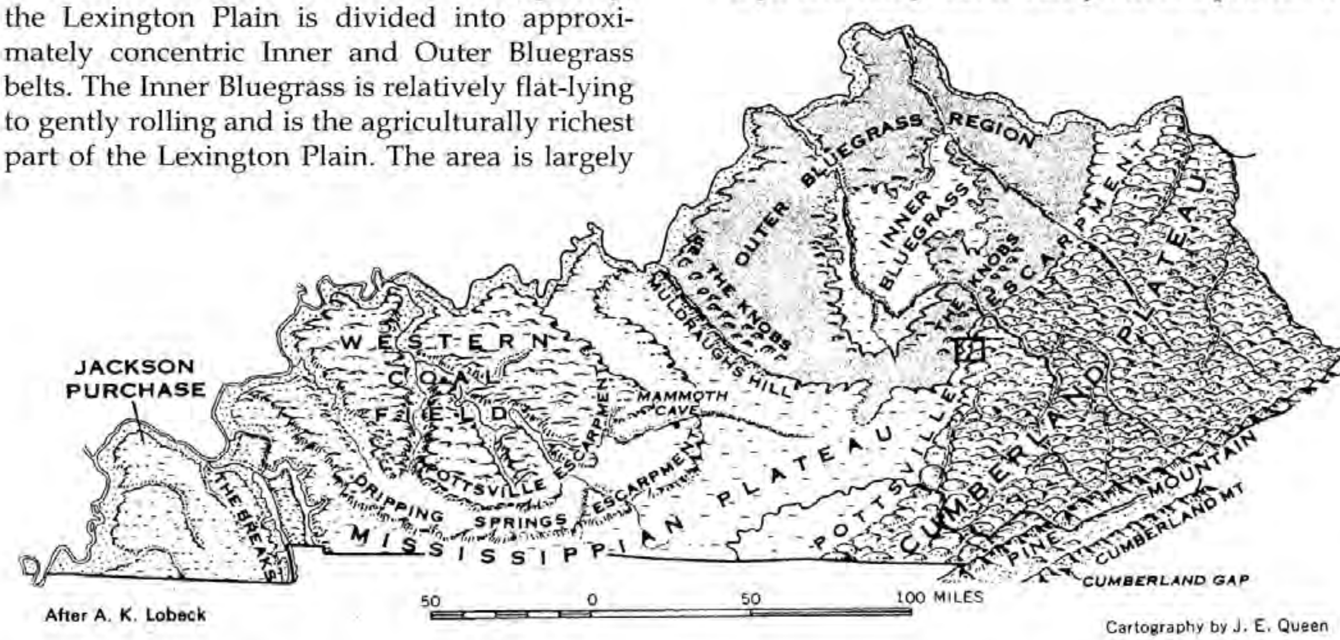
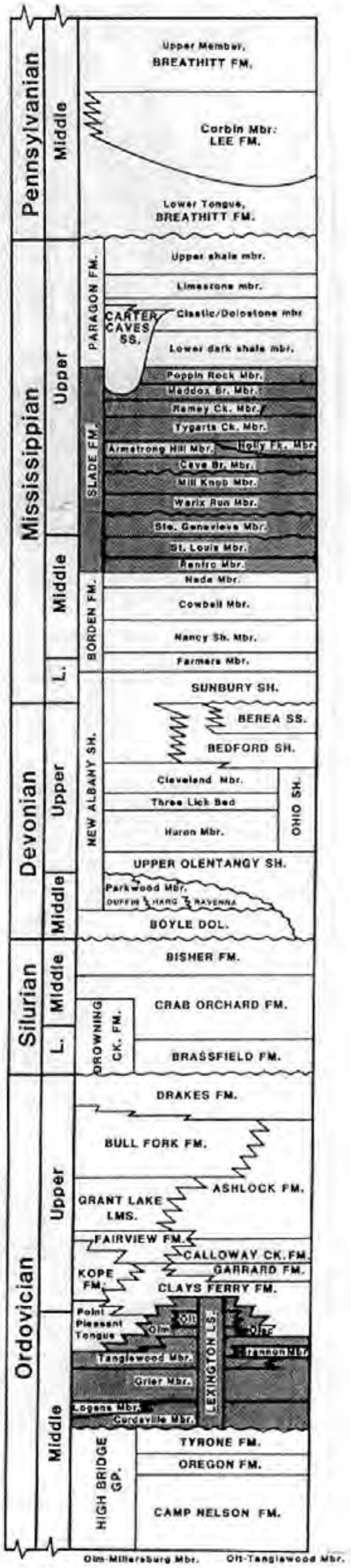


Figure 1. Physiographic map of Kentucky showing the location of the field trip area (box on map).



nonresistant shales and interbedded limestones of the Clays Ferry Formation. On the route to Berea, however, this belt is absent, because the Clays Ferry Formation is largely faulted out on the southern, downthrown side of the Kentucky River Fault Zone (Black, 1968). We cross the fault zone on the southern side of the Clays Ferry Bridge, just south of exit 99, and move from the Inner Bluegrass directly into the Outer Bluegrass proper. The Kentucky River in this area is part of 20-mile segment that roughly follows the Kentucky River Fault Zone.

The Outer Bluegrass proper is developed on Upper Ordovician limestones, siltstones, and interbedded shales of the Garrard Siltstone, Calloway Creek, Ashlock, and Drakes formations, and on its extreme western margins may include Lower and Middle Silurian carbonates and shales, as well as Middle Devonian carbonates (Fig. 2). Because carbonates are more abundant in this part of the section, the topography is not as steep as in the Eden Belt and the soils are more fertile. Hence, farms are larger, and both cattle grazing and burley tobacco are major sources of income.

The gently rolling hills and low ridges of the Outer Bluegrass proper continue up to the city of Berea, where the low-lying topography in and near the city abruptly gives way to a hilly and "mountainous" topography of conical hills and detached ridges known as the Knobs region of the Outer Bluegrass (Fig. 1). The Knobs form a horseshoe-shaped belt that surrounds the Bluegrass (Fig. 1). Near Berea, the Knobs are generally 300- to 600-feet (91- to 183-m) high and are erosional remnants formed as streams cut into the Highland Rim, which defines the outer margin of the Bluegrass in east-central Kentucky (Fig. 1). South and east of Berea, the broad bases of the knobs are generally developed in nonresistant shales of the Upper Devonian–Lower Mississippian New Albany Shale or the Lower and Middle Mississippian Nancy Shale Member of the Borden Formation, whereas the flat, resistant caps are in upper Borden siltstones (Cowbell

Figure 2. Generalized stratigraphic column for central and east-central Kentucky showing units traversed during the course of the trip. The lower darkened section is the Lexington Limestone, in which the trip will begin, and the upper darkened section is the Slade Formation, which comprises most of the section at Bighill.

Member), or more commonly, in the Middle and Upper Mississippian Slade Limestone or Lower Pennsylvanian Lee Sandstone (Weir, 1967; Weir and others, 1971) (Fig. 2). In fact, Indian Fort Mountain, a detached ridge north of Ky. 21 on the way to Bighill, is capped with Slade carbonates and Lee sandstones (Weir and others, 1971). Some of the intervening valley bottoms and glades are floored with Silurian or Devonian shales with veneers of Pleistocene or Holocene alluvial deposits and are broad and flat enough to support farming.

About a mile south of Bighill on U.S. 421, we begin to ascend the Highland Rim. Because the Mississippian outcrop belt is relatively narrow in this area and does not exhibit topography greatly different than that of Pennsylvanian rocks to the east, the Highland Rim here is effectively merged with the Pottsville Escarpment (Fig. 2). The cuts themselves are on the escarpment, so that climbing to the top of the cuts and looking to the north provide an excellent view of the Bluegrass, Knobs, and the escarpment.

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Stratigraphy and Depositional Setting of the Borden Formation (Lower Mississippian) at Bighill, Kentucky

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Stop 1A

Location. Our first stop is made up of two parts. The first part, stop 1A, begins at the very base of the Bighill section on the west side of U.S. 421, approximately 0.9 mile (1.5 km) south of the intersection of U.S. 421 and Ky. 21 at Bighill. The section is located in the southwestern corner (rectangle 7) of the Bighill 7.5-minute quadrangle in Madison County, east-central Kentucky.

Purpose. At this stop we will examine (**1**) the uppermost New Albany Shale and (**2**) the Jacobs Chapel Bed, Rockford Limestone, and lowermost foot of the Borden Formation (see Figure 1). Some of the more interesting features to observe and discuss include: (**1**) both anaerobic and dysaerobic conditions in the marine environment at the time of sediment deposition, and (**2**) basin floor/prodelta deposits.

Stratigraphy, Description, Interpretation. Stop 1A is a rather unimpressive exposure that shows, in ascending order, the New Albany Shale, Jacobs Chapel Bed, Rockford Limestone, and lowermost foot or two of the Nancy Member of the Borden Formation (see Fig. 3). In order to examine these contacts, we exposed approximately 2 feet of section along the base of the slope.

At the base of the section we can see the uppermost foot (30 cm) of the New Albany Shale. The New Albany is a carbonaceous, black to brownish black mud-shale that is noncalcareous and displays a platy to fissile parting property. Close examination of the New Albany shows a sparse scattering of finely crystalline pyrite. Fossils are rare in the unit, but include small inarticulate brachiopods, scattered conodont remains, spores, and fish teeth (Weir and others, 1971). The presence of the conodonts *Siphonodella* sp. and *Spathognathodus* sp. (Weir and others, 1971) points to an Early Mississippian (Kinderhookian) age for the upper part of the New Albany Shale. We interpreted this unit to be an anoxic basin-floor deposit. It covers a wide geographic area and provides a platform or base for the subsequent deposition of the greenish-gray shales of the Nancy Member's Borden Delta Complex (see Figs. 4 and 5). Evidence for this interpretation includes (**1**) the widespread nature of the New Albany Shale, (**2**) the high organic content of the unit along with its fine-grained nature, (**3**) a lack of any evidence of bioturbation (this is the

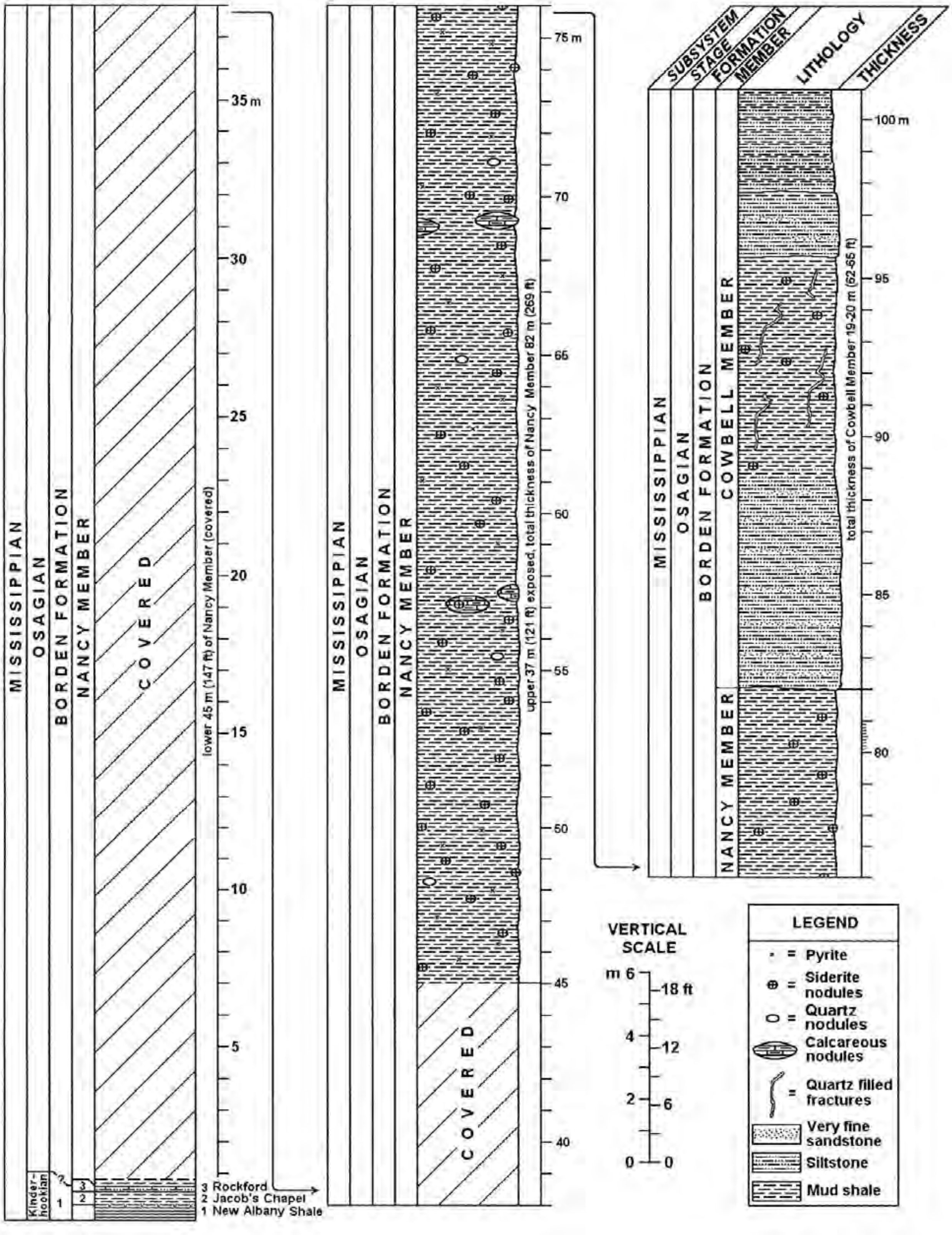

Figure 3. Composite stratigraphic section for the Nancy and Cowbell Members of the Borden Formation, along Bighill roadcut on U.S. 421, Madison County.

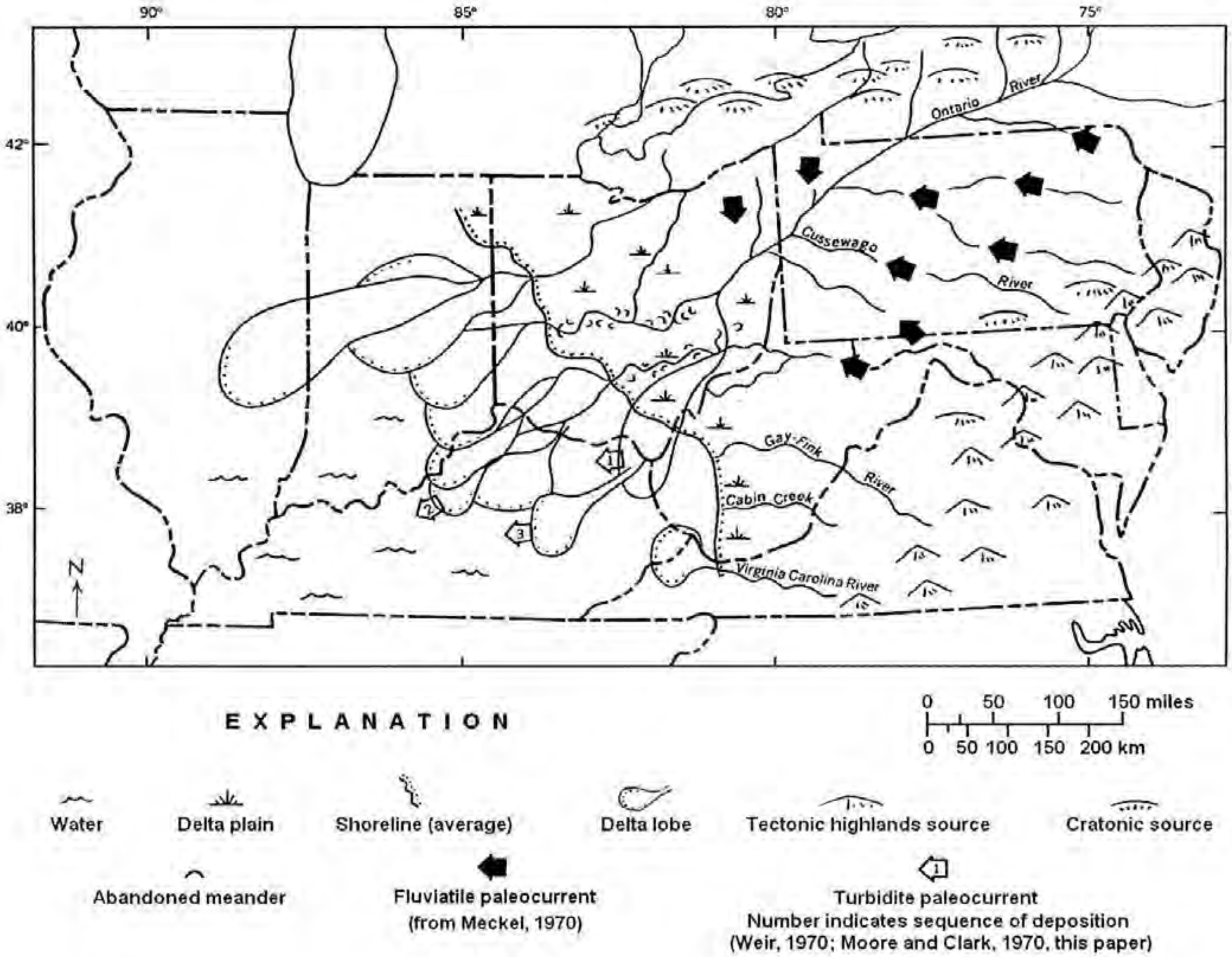

Figure 4. Paleogeographic map showing the development and progradation of the Borden Delta Complex in Kentucky, Indiana, Illinois, and Ohio, Early Mississippian times (adapted from Kepferle, 1977).

reason for the highly fissile nature of this unit), (4) presence of a low-diversity marine fauna, and (5) a scattering of pyrite crystals throughout the unit. Anoxic or anaerobic conditions occur whenever the dissolved oxygen content in a column of water is less than 0.1 mL of O_2 per liter of water.

Immediately above the New Albany Shale rests the Jacobs Chapel Bed. It is likewise of Early Mississippian (Kinderhookian) age, and at this locality is approximately 50 cm thick. The Jacobs Chapel is a mud-shale to clay-shale that is greenish gray, noncalcareous, and exhibits a platy to flaggy splitting property. The lower 10 to 12 cm contains a scattering of small (less than 3 cm), elliptical phosphate nodules, as well as phosphatic fossils such as conodonts and a scattering of other vertebrate remains. Also present along the very base of the Jacobs Chapel is a very thin (less than 2 cm thick) bed of argillaceous limestone, with abundant cone-in-cone structures. The term "Jacobs Chapel Shale" applies to a thin (less than 0.3 m) layer of greenish shale that overlies the New Albany (black) Shale and underlies the Rockford Limestone in southern Indiana. Conodonts collected from the Jacobs Chapel belong to the *Siphonodella lower crenulata* Assemblage Zone (Sandberg and others, 2002). This clearly places the Jacobs Chapel in the Kinderhookian Series. We think the Jacobs Chapel represents a basin-floor (hemipelagic) deposit

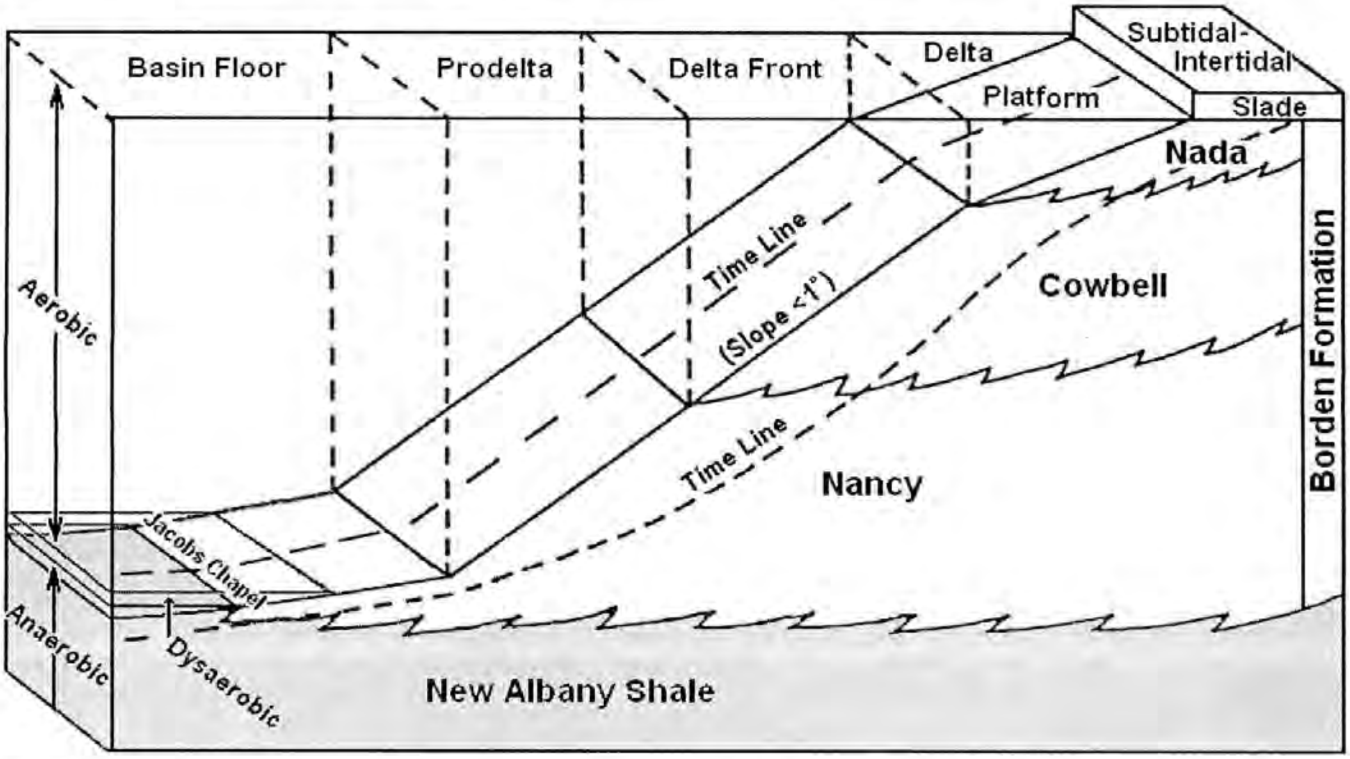

Figure 5. Block diagram showing the interpreted lithofacies of the New Albany Shale and Borden Formation in the Bighill area, east-central Kentucky (adapted from Kepferle, 1977).

that accumulated under dysaerobic conditions with very slow rates of sedimentation (see Fig. 4). In support of this interpretation is (**1**) the fine-grained (clay-shale to mud-shale) character of the Jacobs Chapel, (**2**) its greenish gray color, (**3**) the presence of small (less than 3 cm) phosphate nodules along the base of the unit, (**4**) abundance of conodonts in the samples collected, along with the presence of other phosphatic fossils. The overall color of the shale suggests that any iron making up the clay minerals in the shale was in a +2 oxidation state (ferrous iron). $Fe^{2+}O$ is stable whenever conditions are slightly reducing. The precipitation of apatite also appears to be favored by conditions that are slightly reducing. These types of environmental settings are called dysaerobic conditions, and occur whenever the dissolved oxygen content of water is between 1.0 and 0.1 mL of O_2 per liter of water (see Fig. 5) (Kammer and others, 1986). The Jacobs Chapel Bed also is substantially thicker (50 cm) than in the type section in southern Indiana (10 to 15 cm). We can account for this by suggesting that sedimentation rates were higher in the Appalachian Basin at this time than in the Illinois Basin where the type section is located.

At this locality the Rockford is a very thin-bedded (7 cm) siltstone that is greenish gray, lightly calcareous, and highly bioturbated (ichnofabric index 5 of Bottjer and Droser, 1994). It rests directly on the Jacobs Chapel and is immediately overlain by the Nancy Member of the Borden Formation. The Rockford is normally a limestone, but at this locality is more of a calcareous siltstone (about 10 percent carbonate). The base of the unit is sharp, almost erosional with the underlying Jacobs Chapel, and grades upward into the overlying Nancy Member of the Borden Formation. The unit is extensively bioturbated, almost to the point of completely obliterating any primary sedimentary structures. We believe the Rockford represents an isolated, distal-most part of a carbonate turbidity flow that came off a carbonate platform to the west. We suggest this because of (**1**) the sharp erosional base of the Rockford bed and gradational upper surface with the overlying shales of the Nancy Member, (**2**) the fine-grained nature of the deposit, along with its carbonate component, (3) the extensive bioturbation of the unit, (4) the stratigraphic position of the Rockford between two deeper-water marine shales (the Jacobs Chapel Bed and lower Nancy Member).

Only the lowermost part of the Nancy Member is exposed at this stop. It is a greenish gray, noncalcareous mud-shale that splits into platy-flaggy partings. We believe the lowermost Nancy Member along with the Jacobs Chapel represents a basin-floor deposit that accumulated under dysaerobic conditions. At the time of deposition this area was basinward of the Borden Delta Complex that was prograding out into east-central Kentucky from a northeastern direction (see Figs. 4–5). Only the finest grain fraction (clays and fine silts) was transported to this location at the time.

Stop (1B)

Location. This stop is some 0.4 mile (0.6 km) farther south along U.S. 421. We are now about 45 m (147 feet) up into the section (the lower half of the section is unfortunately covered by riprap or is heavily vegetated).

Purpose. Our second stop (1B) is a more extensive exposure of the Nancy Member of the Borden Formation. Some of the more interesting features to observe and discuss include (1) evidence of a prodelta deposit for the Nancy Member and (2) generally aerobic conditions in the marine environment at the time of sediment deposition.

Stratigraphy, Description, Interpretation. At this location you can observe a striking example of the Nancy Member of the Borden Formation (see Fig. 3). Here the Nancy Member is a dark gray to dark greenish gray mud-shale to silt-shale that is noncalcareous and exhibits platy to flaggy splitting properties. Sampling the unit demonstrates that these shales become increasingly more silty up-section, ranging from mud-shales lower in the section to silt-shales toward the top. These shales are extensively bioturbated with an ichnofabric index of 5 to 6 (Bottjer and Droser, 1994). Chaplin (1980, 1985) placed the trace fossils he found within the Nancy Member to be characteristic of the *Zoophycos* ichnofacies.

Scattered throughout the unit are nodules of siderite (ironstone). These are quite numerous and can vary from less than 1 cm (half an inch) across to more than 30 cm (1 foot). They are generally ellipsoidal to spherical in shape, though more complex shapes can be seen, and olive gray in color (when fresh). Internally, the nodules can be quite complex. Close inspection reveals the following types of nodules

1. Typic nodules, which have an undifferentiated internal fabric and generally sharp external boundaries.
2. Nucleic nodules, which have a foreign core inside them. In this case we can find a variety of fossils acting as an internal object or core in the nodules. The most common fossils found within the nodules are brachiopods, mollusks, especially straight-shelled, as well as coiled nautiloids (Plate 4), and goniatite ammonoids, and conularids. Less commonly encountered fossils include bryozoans (mainly fenestrate), trilobite fragments, and echinoderm debris.
3. Geodic nodules, with a hollow interior, often with a drusy lining of crystals.
4. Septaric nodules, which have an internal radiating crack pattern (Plate 5). These, along with the geodic nodules, are commonly infilled or replaced with a variety of secondary minerals, including barite, calcite, galena, pyrite, quartz, and sphalerite.
5. Concentric nodules or concretions, with concentric layers much like the layers of an onion. This feature is more commonly seen in nodules that have been exposed to weathering for a long period.
6. Pseudomorphic nodules, or nodules with a variety of internal fabrics. Most commonly fabrics seen here are those resulting from burrowing activity. Some of the nodules themselves are probably large burrows that have been replaced or permineralized by siderite. Some show smaller internal burrows within the nodules themselves, however. Often

these nodules have preserved within them three-dimensional burrow systems, which commonly show a horizontal attitude.

In addition to the siderite nodules, you can also find a scattering of both phosphate and quartz nodules (a few appear to be geodes) and large masses of pyrite. Also of some note is the occurrence of larger masses (nodules?) of a light-colored limestone. These limestone nodules are quite large, well over a foot across, and are very argillaceous (containing an insoluble residue content in excess of 50 percent). The limestone nodules occur along two distinct horizons, one at about 57 m and another at 69 m above the base of the Nancy Member. In viewing both these nodules, as well as the nodules of siderite, you can readily see that the shales are draped over and round the nodules. Both of these types of nodules can contain relict laminations and three-dimensional trace fossils within them. Where fossils are present in these nodules, in the majority of cases there are few signs of compaction other than a few minor fractures. This evidence suggests an early formation for the nodules within *the sediment before* extensive compaction of the shales had taken place.

The Nancy Member is interpreted to be a prodelta deposit, formed as the Borden Delta Complex prograded across east-central Kentucky. This unit is mainly deposited in a generally aerobic environmental setting. We are dealing with a stratified ocean basin in which bottom waters were anaerobic, lower waters were dysaerobic, and middle and upper waters were evidently aerobic. The evidence supporting this interpretation is (**1**) the fine-grained nature of the shales coupled with a gradual increase in grain size up-section, (**2**) the presence of an open marine fauna, (**3**) the high degree of bioturbation of these shales, resulting in a nearly complete homogenization of the sediment, (**4**) the occurrence of delta front sands and silts of the Cowbell Member that conformably overlie the Nancy Member (see Fig. 3).

The final unit present at this stop is the Cowbell Member of the Borden Formation. Unfortunately, we will not be able to examine this unit since it's about 37 m above us (beginning about the third bench up). Here the Cowbell is a series of siltstones and irregularly interstratified shales that are mostly greenish gray to yellowish gray. The siltstones are quite argillaceous, in places laminated, and locally bioturbated (ichnofabric index 3 to 5 of Bottjer and Droser, 1994). In places, the siltstone may grade into very fine sandstone. Fossils are scarce, but upon close inspection you can find small brachiopods, crinoid debris, and fenestrate bryozoans. Trace fossils are common, *Zoophycos* being the most conspicuous form seen. The more shaly intervals of the Cowbell also contain numerous siderite nodules and large and prominent joints and fractures that are infilled with very coarsely crystalline vein quartz.

One significant point to notice about the Bighill section is the thickness of the Cowbell: it is only some 19 to 20 m (62 to 65 feet) thick. This is quite thin for the Cowbell, which typically makes up a major part of the Borden Formation (often well over 200 feet thick). In contrast, the Nancy Member is unusually thick (some 82 m [269 feet]).

The Cowbell Member represents a series of delta front sands and silts. It is composed of a complex of subenvironments that are associated with the advancing Borden Delta Complex. Some of the subenvironments found within the Cowbell probably include silty/sandy distal-bar deposits and more shaly interdistributary bay deposits (see Chaplin 1980, 1985). At Bighill, considering the great thickness of the Nancy Member and corresponding thinness of the Cowbell Member, we are probably positioned between major delta lobes along the Borden Delta Complex.

Stop 2

Location. Stop 2 occurs higher up the Bighill section on either side of U.S. 421, about 2 miles (3.2 km) south of the intersection of U.S. 421 and Ky. 21, in Bighill. The section is located in the southwestern corner (rectangle 7) of the Bighill 7.5-minute quadrangle in Madison County, east-central Kentucky.

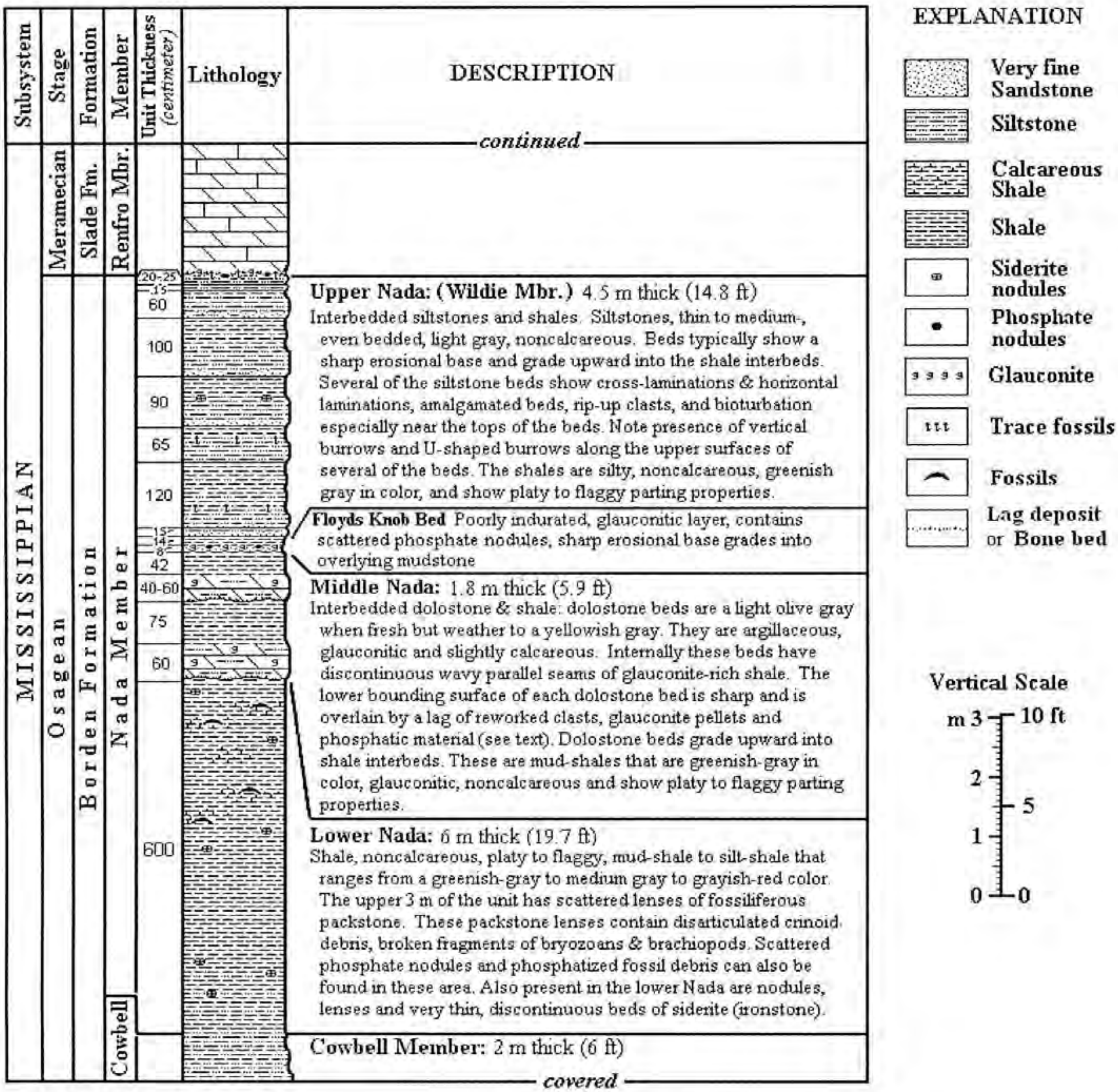

Figure 6. Block diagram showing the interpreted lithofacies of the New Albany Shale and Borden Formation in the Bighill area, east-central Kentucky (adapted from Kepferle, 1977).

Purpose. At this stop we will examine the upper Borden Formation including the Nada and Wildie Members (Plate 1). Some of the more interesting features to observe and discuss include (1) delta platform lithofacies, characterized by phases of marine transgression and delta destruction, and (2) evidence of pauses in the marine sedimentation, resulting in the development of what could be described as a transgressive surface.

Stratigraphy, Description, Interpretation. Our stop begins at the contact between the Nada and Cowbell Members of the Borden Formation and ends at the base of the Renfro Member of the Slade Formation (see Fig. 6, Plates 1–2). The contact between the Nada and Cowbell Members is at the base of the roadcut on the east side. The remaining Borden Formation is approximately 11.9 m (39 feet) thick at this locality, and includes the Nada Member and, tentatively, the Wildie Member (see Fig. 6). For descriptive purposes, we have divided the Nada into a lower, middle, and upper part. We think the upper part of the Nada probably corresponds to the Wildie Member as originally described by Weir and others (1966).

The lower part of the Nada Member is about 6 m thick (19.7 feet). It can be described as a noncalcareous, platy to flaggy mud-shale to silt-shale that ranges from a greenish gray to medium gray to grayish red color. The upper 3 m of the unit has scattered lenses of fossiliferous packstone (1 to 7 cm thick and anywhere from 30 to 100 cm wide). These packstone lenses contain disarticulated crinoid debris, broken fragments of bryozoans (mainly fenestrate and rhomboporid bryozoans), and brachiopods (mainly spiriferid brachiopods). Scattered phosphate nodules (Plate 3) and phosphatized fossil debris (Plates 4–6) can also be found in this unit. Also present in the lower Nada are nodules, lenses, and very thin, discontinuous beds of siderite (ironstone). The packstone lenses are probably formed by local currents that scoured out shallow depressions in the soft underlying shale. These scoured-out areas were filled with broken and disarticulated fossil debris, presumably by the same marine currents.

The middle part of the Nada Member is around 1.8 m (5.9 feet) thick and consists of greenish gray mud-shale, along with several beds of dolostone. The dolostone beds are a light olive gray when fresh, but weather to a yellowish gray color. They are argillaceous, glauconitic, and slightly calcareous. Internally, these beds have discontinuous, wavy, parallel seams of glauconite-rich shale (hummocky cross-beds?). Close inspection may also reveal rip-up clasts, some scouring, and bioturbation. The lower bounding surface of each dolostone bed is sharp and overlain by a lag of reworked clasts of dolostone, glauconite pellets, and phosphatic debris. The phosphatic material includes small phosphate nodules, fish teeth, scales, small fragmented bones, and conodonts. Phosphatized invertebrate remains, including, gastropods, brachiopods, and crinoid debris are also present.

The second dolostone bed has prominent quartz nodules in the lower two-thirds of the bed. In general, these nodules are elliptical to highly irregular in shape, with the long axis of the nodule lying parallel to bedding. The outer surface of each nodule appears rather bumpy and irregular, with the larger specimens taking on the appearance of a flattened head of cauliflower. The outer edge of the nodules consists of microquartz (equidimensional crystals less than 20 μm) and chalcedonic quartz (microcrystalline quartz crystals with a fibrous crystal habit). Toward the interior of the nodule the quartz changes to a megaquartz (equidimensional crystals greater than 20 μm) and becomes optically clear or lipid. Some of the nodules have the interior filled with coarse quartz crystals; others maintain a hollow interior and can therefore be described as geodes. A third variation is a final infilling of many of these nodules with iron-rich, saddle-shaped dolomite crystals. Saddle dolomite is a variety of dolomite with a warped crystal lattice. It is characterized by curved, nonplanar, crystal faces and cleavage, and shows a sweeping extinction pattern when viewed in thin section under cross-polarized light. These are clearly a type of void-filling dolomite. They are quite stunning to see in the outcrop, with the curved, light brown dolomite crystals standing out against the white exterior of the quartz nodule.

Each dolostone bed grades upward into the mud-shale. The upper contact is gradational with the overlying shale, which gradually changes from a dolostone to a dolomitic shale and finally into a mud-shale over a vertical distance of 1 to 2 cm. The mud-shales are greenish gray, glauconitic, noncalcareous, and show platy to flaggy parting properties.

Separating the middle part of the Nada Member from the upper part (i.e., the Wildie Member) is a very thin bed of dark green glauconite containing abundant phosphate nodules; it is locally known as the Floyds Knob Bed. This poorly indurated, glauconitic horizon is around 8 cm thick, has a sharp erosional base, and grades into the overlying mudstone. Upon close inspection you can just barely make out small-scale cross-laminations in this bed, along with small burrow casts. This is a very interesting unit, and has been locally used as a key bed or marker bed to facilitate correlation in field mapping.

The remainder of the upper part of the Nada Member consists of a series of interbedded siltstones and silty shales. The thin- to medium-bedded, even-bedded siltstones are light gray, noncalcareous, and grade from coarse silt to very fine sandstone in places. These beds typically show a sharp erosional base and grade upward into the shale interbeds. Several of the siltstone beds show interesting internal sedimentary structures, including cross-laminations and horizontal laminations, amalgamated beds, rip-up clasts, and some bioturbation, especially near the tops of the beds. Of particular interest are cylindrical, U-shaped burrows *(Arenicolites?)* (Plate 8) and vertical burrows *(Skolithos?)* (Plate 9) along the upper surfaces of several of the siltstone beds. This suggests a *Skolithos* ichnofacies, a trace fossil assemblage indicating a lower littoral to infralittoral environment with moderate to relatively high energy conditions.

The shales are typically silty, noncalcareous, greenish gray, and show platy to flaggy parting properties. The lowest of these shale interbeds rests directly on the Floyds Knob Bed and is itself a greenish gray, noncalcareous, glauconitic mudstone.

We believe that the upper part of the Nada Member as seen here corresponds to the Wildie Member of the Borden Formation as defined by Weir and others (1966). We are basically looking at a series of interbedded siltstones and silty-shales that are underlain by a layer of glauconitic mudstone and overlain by dolostones of the Renfro Member. To quote Weir and others (1966, p. F18–F19,

> Near Wildie the member is made up almost entirely of thick beds of resistant brownish-weathering, greenish-gray siltstone and very fine grained sandstone with shale in thin seams and partings. ... The shale is nonresistant, clayey to silty, in places glauconitic, and mostly greenish gray but locally grayish red and grayish purple. A persistent seam of glauconitic siltstone is present at the base of the unit and another glauconitic seam is at the top. ... The Wildie Member is overlain by the Renfro Member, whose yellowish-gray silty limestone contrasts strongly with the drab shale and siltstone of the Wildie Member.

The very top surface of the upper part of the Nada is represented by yet another pause in sedimentation (a bypass surface?) or possibly a transgressive lag deposit. The surface is highly bioturbated, with horizontal traces being most common. Note the high concentration of *Zoophycus* and other trace fossils along the upper surface. It is also highly glauconitic, and contains a dense scattering of phosphate nodules and phosphatic fossil remains. This concentration of phosphatic material (fluoroapatite) and glauconite extends down into the underlying bed some 20 cm or so. Phosphatic fossils found along this surface include fish teeth, scales, small fragmented bones, phosphatized internal molds of gastropods and brachiopods, and crinoidal debris. The phosphate nodules are noteworthy because they are highly bored and show evidence of organisms colonizing or attaching to the surface of the nodule. In addition, you can find current ripples along the upper surface of this layer. Note the ripple crests have a N25°W orientation with an apparent current direction to the northeast. The average wavelength of these ripples is 6.5 cm and wave height is 0.3 cm.

Discussion. The Nada Member of the Borden Formation has generally been interpreted as subtidal, open marine, shelf-margin environments that developed on top of the delta front silts and sands of the Cowbell Member (Chaplin and Mason 1979; Chaplin, 1980, 1985). These open-marine sediments interfinger and are interstratified with the underlying Cowbell. We feel that the upper parts of the Borden Formation, specifically the Nada Member, represent phases of delta destruction that occurred as the Borden

Delta Complex prograded. These phases of delta destruction are caused by the complete abandonment of a prograding delta lobe or a switching or diversion of a lobe to another deposition site. Whenever a delta lobe is abandoned, the sediments are naturally subjected to winnowing and reworking by ocean waves and currents. Occasional shifts in deposition of some of these delta lobes have introduced fine-grained clastic debris onto abandoned lobes.

The Nada Member, as seen here, is different from the Nada so carefully examined in northeastern Kentucky by Chaplin and Mason (1979) and Chaplin (1980, 1985). To the northeast, the Nada typically has more carbonate beds and contains a more diverse, open-marine invertebrate fauna. Chaplin (1980) reported the presence of abundant brachiopods, bryozoans, cephalopods, gastropods, pelecypods, pelmatozoan detritus, rugose corals, etc. He interpreted these carbonate buildups as series of crinoid-bryozoan shoals that grew on top of the abandoned lobes of the "Borden Delta Complex." The exposures seen here do not have the same degree of carbonate deposition as those in the Morehead area, nor does it have the diverse, open marine shelly faunas so typical of the Nada to the northeast.

The Nada Member (at least the lower part at Bighill), is muddier than that in northeastern Kentucky and deposition seems to have occurred in a quieter water setting. Most of the time sedimentation must have been slow enough to allow for the development and growth of quiet water, soft bottom communities of crinoids, bryozoans, and brachiopods. These communities were, however, periodically disrupted by large storm events. This is evident because of the presence of debris-filled scours that are infilled with the broken and disarticulated remains of these fossil communities.

The presence of dolostone beds in the middle part of the Nada, suggests a shallowing of this sequence. These dolostone beds began as carbonate muds that were washed into the area and were later modified by waves and currents (possible storm currents). Evidence of higher energy, storm related conditions includes: (1) the sharp/erosional lower contacts of the dolostone beds that are in turn overlain by a basal lag deposit; (2) the presence of rip-up clasts in the dolostone beds, occurrence of scouring along with hummocky cross-stratification (hummocky cross-stratification is believed to be formed by storm surge); (3) the gradational contact between the dolostone bed and the overlying shales. The shales that lie above each dolostone bed are very similar to shales of the lower part of the Nada Member. It would suggest a slow return to normal marine conditions following each storm event.

We think the Floyds Knob Bed, the 8 cm layer of glauconite and phosphate nodules, represents a marine transgressive lag deposit which resulted from a reworking of the underlying sediments along one of these abandoned delta lobes. The crossbedding and sharp lower contact suggests fairly strong current activity which winnowed away any fine-grained mud and concentrated the heavier qlauconitic sands and phosphate pebbles.

The upper Nada or Wildie Member represents a separate lithofacies that is found within the Nada Member and interfingers with it. We think this unit is more than likely a series of shallow (infralittoral to sublittoral) marine sands which developed under moderate to relatively high wave or current activity. Evidence for this environmental setting includes: (1) the presence of cross-laminations, amalgamated beds, and rip-up clasts in a number of the siltstone beds; (2) the sharp/erosional base of the siltstone beds and their fining upwards character from a coarse siltstone to a silty shale, this suggests a waning current; and (3) the present of a restricted, low diversity ichnofauna consisting of cylindrical, U-shaped and vertical burrows indicating a Skolithos ichnofacies (a trace fossils assemblage characteristic of low intertidal to subtidal marine environment, moderate to high-energy conditions, and typically found in clean well sorted sands). These characteristics could suggest that the siltstone beds could have been deposited as a series of storm related deposits.The shaly interbeds would therefore represent a return to normal marine sedimentation between storm events.

The last event must surely represent a pause in the deposition of the upper part of the Nada Member. This layer could easily be interpreted as a "bypass surface" or a transgressive lag deposit that occurred just prior to the full development of the carbonates of the Slade Formation. The intensive burrowing and concentration of authigenic sediments (glauconite and phosphate nodules) in this layer suggest that a long period of time had elapsed during which terrigenous deposition was slow to negligible. This was long enough to allow for the chemical/biochemical interaction of the existing sediments with overlying ocean waters. We lack the necessary paleontological evidence to call this contact a paraconformity. But it certainly represents a period of nondeposition, with a corresponding break in time (a hiatus). The concentration of glauconite and phosphate in these sediments and the presence of an abundant trace fossil biota, also suggests that the area was experiencing an influx of nutrient-rich waters. This was likely due to ocean upwelling onto this shallow-marine, delta platform. Any reworking of this sediment would occur in response to changes in relative sea level or to variations in local current patterns. In either case, the coarser patches of phosphatized and glauconized sediment would remain as a lag deposit while the finer grained matrix were winnowed away by marine currents.

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Horizontal Drains

From the Jensen Drilling Company Web Site

Why Horizontal Drains?

Subsurface water may act in any number of ways to reduce the stability of cuts and embankments (Fig. 7). Among these are decrease in cohesion, subsurface erosion, lateral pressure in fractures and joints, and excess pore water pressure. One way to stabilize a slope is to place riprap onto it (Fig. 8). Another way is to remove subsurface water through the use of horizontal drains, which are holes drilled into an embankment or cut slope and cased with a slotted PVC pipe.

How Are Horizontal Drains Installed?

The dozer-mounted drill is positioned perpendicular to the face of the embankment, and the angle above the horizontal is set (Fig. 9). A 4⅛-inch hole is advanced to the design depth (300 feet at the Bighill project) using rotary drilling. The drill bit, which is expendable, is attached to 3½-inch drill rods (each drill rod is 10 feet long). After completion of the drilling, 1½-inch slotted schedule #80 PVC is installed through the drill rods. The drill bit is then detached (and left behind in the borehole), and the drill rods are withdrawn while keeping the PVC in place. This completes the installation. In order to move the dozer-mounted drill as little as possible, multiple horizontal drains are drilled in the embankment in a close spacing. The outermost horizontal drains are drilled so as to fan away from the center horizontal drains in opposite directions (Fig. 10).

Figure 7. Failed slope during construction of Ky. 421.

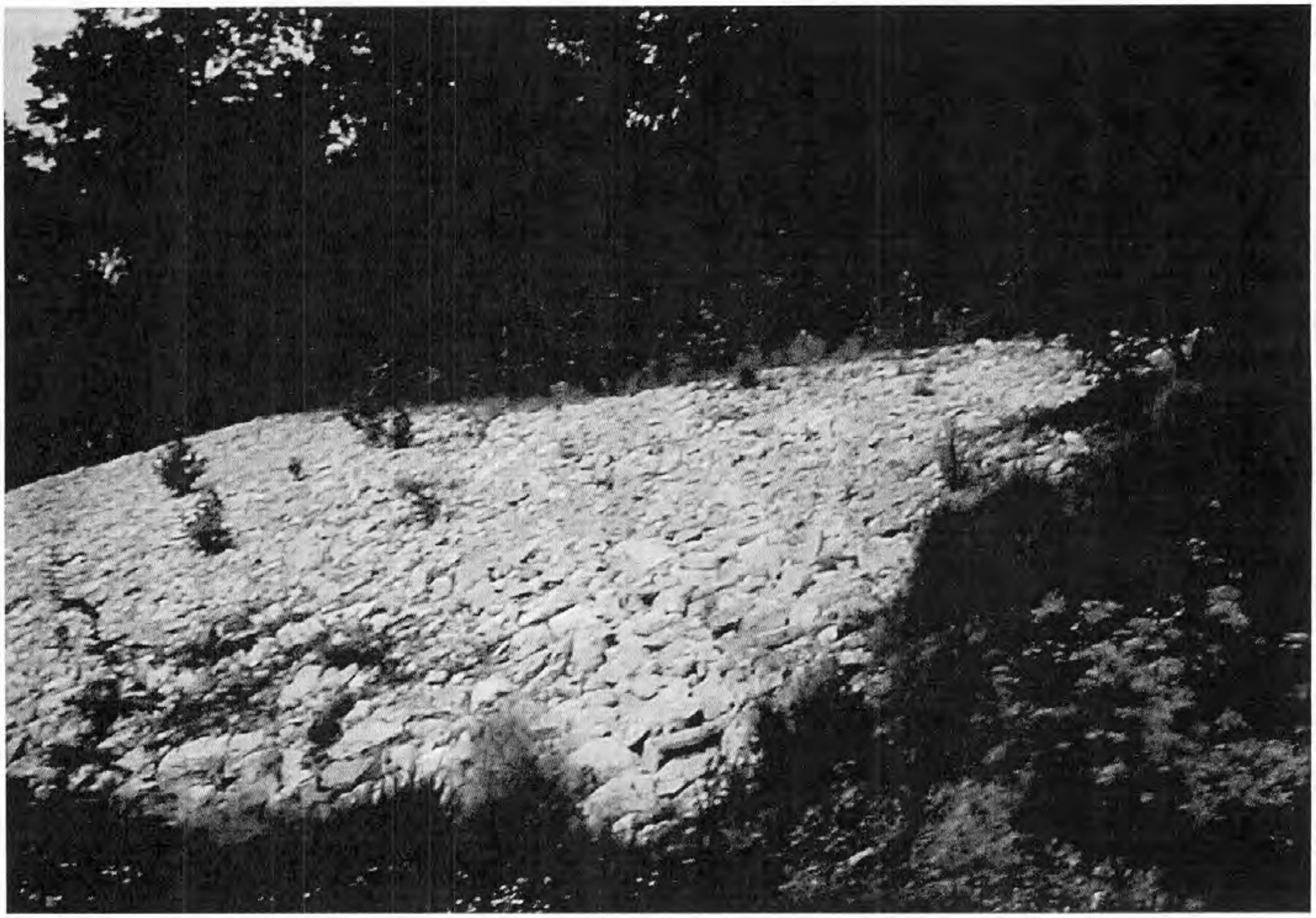

Figure 8. Rip-rap placed on failed slope to stabilize.

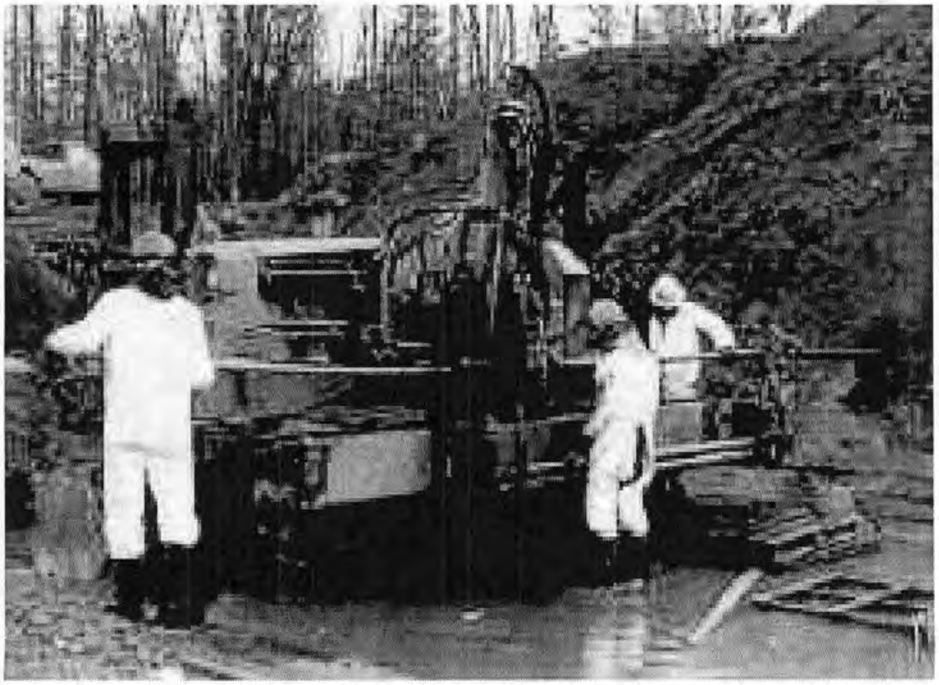

Figure 9. Dozer-mounted drill.

Figure 10. Completed project with three drainage pipes (right side in grass) in the Borden Formation.

Stratigraphy and Depositional Environments of the Middle and Upper Mississippian Slade and Paragon Formations, Bighill Exposure, East-Central Kentucky

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Introduction

Approximately 30 percent of Kentucky is underlain at the surface by Mississippian rocks, especially carbonates, and these rocks are responsible for much of Kentucky's physiography and karst landforms. These rocks are also economically important as sources of crushed stone, groundwater transit routes, and hydrocarbon reservoirs. On this trip to Bighill, we will examine a representative exposure of these rocks, which contains clastic and carbonate components and illustrates both physiographic and economic aspects of the section. Although the section is about 7 years old, its presence on a part of the Pottsville Escarpment that is narrow (Fig. 1) provides a fresh, nearly continuous, and easily accessible exposure of Middle and Upper Mississippian carbonate rocks from the Slade and lower Paragon formations over a short distance along a new stretch of highway with broad shoulders. Not only is the section easily accessible for study, its location also allows for ready explanation of relationships between lithology and physiography, and many features are available for interpreting depositional environments, paleoecology, possible hydrocarbon reservoirs, and karst hydrology. These features will be briefly described for each unit below.

Procedures. The Bighill section was measured and described using standard U.S. Geological Survey procedures by the 2001 Advanced Stratigraphy class at the University of Kentucky; colors were approximated using the Geological Society of America's Rock-Color Chart (Rock-Color Chart Committee, 1991). Representative samples were collected for thin sections, and the entire exposure was scintillated with a hand-held scintillometer according to procedures outlined in Ettensohn and others (1979) to generate the artificial gamma-ray log in Figure 11.

Mississippian Paleogeography, Paleoclimate, and Eustasy

Based on the latest geologic time scale (Gradstein and others, 2004; Gradstein and Ogg, 2004), the carbonates of the Slade Formation were deposited from about 345 to 322 Ma ago and are Visean through middle Serpukhovian (latest Osagean to late Chesterian) in age, whereas the preserved clastic and carbonate sediments of the overlying Paragon Formation probably represent only 1 or 2 million years of mid-Serpukhovian (late Chesterian) time (Fig. 12). During Late Devonian through Early Mississippian (Famennian–Tournaisian) time, the field trip area was situated at about 30° south latitude in a warm temperate climatic zone (Scotese, 2003) with sufficient moisture to ensure weathering and transport of resulting clastic materials from Acadian highlands to the nearby sea during the major episode of Price-Pocono-Grainger-Borden deltaic development. By the later part of Early Mississippian time, however, the continent had

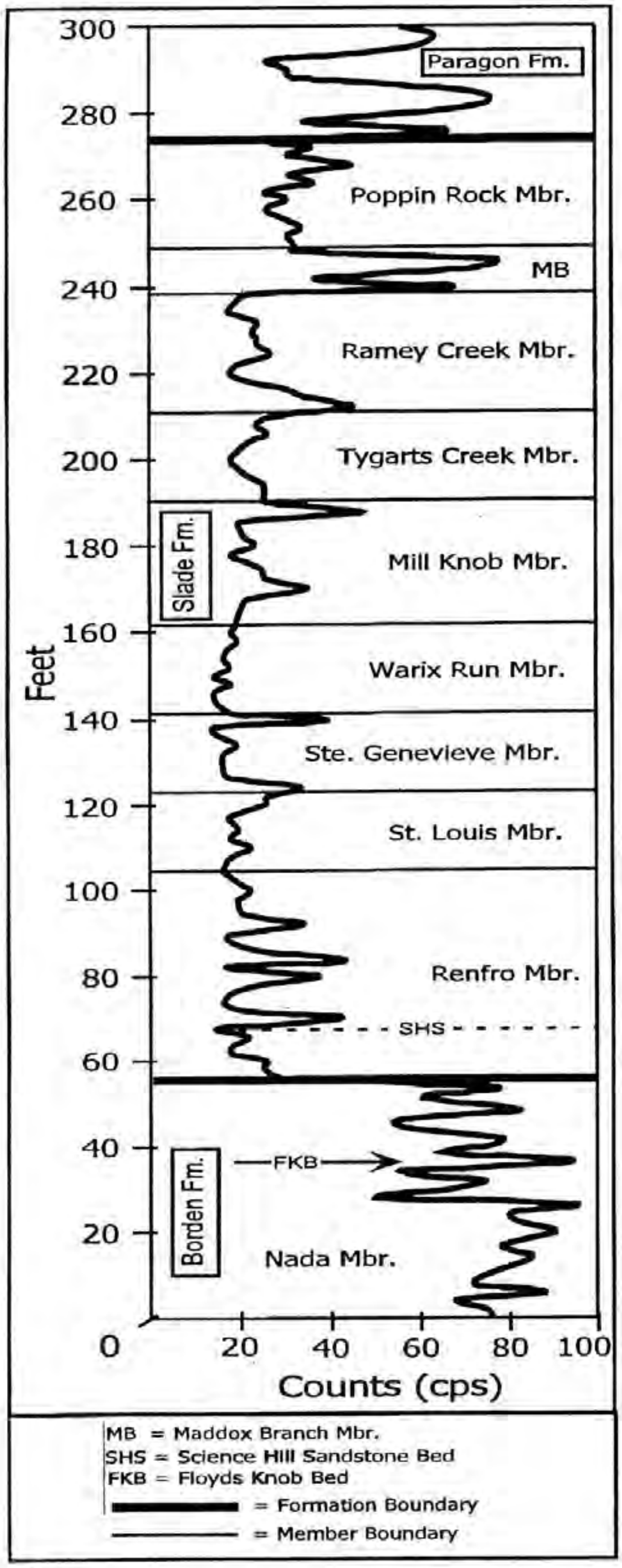

Figure 11. Artificial gamma-ray log of the Mississippian section at Bighill, showing the gamma-ray signature of each unit.

moved northward into the arid, subtropical, trade-wind belt, and the field trip area was located at about 15 to 20° south latitude (Plate 11) and remained there until latest Mississippian (latest Chesterian) time, when the area moved into the humid, tropical zone (Scotese, 2003). Because higher temperatures and aridity promote biological and physiochemical precipitation of carbonates (Lees, 1975), presence in a belt of subtropical aridity is typical of thick, pure carbonate units like the Slade Formation (Lees, 1975; Heckel and Witzke, 1979). Hence, paleoclimate, in combination with the paleogeographic and tectonic conditions discussed later, was an important factor in the deposition of the Slade carbonates.

The upper Borden-Slade units exposed at Bighill reflect the transition from a generally west-dipping, clinoformed, deltaic sequence to a subtly east-dipping, carbonate ramp. The transition is in the declining or shallowing part of the second-order, Variscan-Hercynian tectono-eustatic cycle (Ettensohn, in press) and exhibits parts of two third-order cycles that are broadly similar to global trends (see Ross and Ross, 1987, for example); in this area, however, both cycles probably reflect tectonic and eustatic components (Ettensohn, 1994; Ettensohn and others, 2002; Al-Tawil and Read, 2003). The first third-order, deepening cycle ends with the extreme shallowing indicated in the Renfro Member, while another broadly deepening cycle begins with the St. Louis, culminates in the Maddox Branch Member, and continues to decline and shallow through the Paragon Formation (Fig. 12). The many smaller fourth-order cycles (Fig. 12) are probably eustatic, reflecting glacio-eustatic transition into Carboniferous icehouse conditions (Al-Tawil et al., 2003; Al-Tawil and Read, 2003), although local tectonism has certainly controlled certain cycle durations (Dever 1980,1990; Dever and others, 1977, 1990; Ettensohn, 1977, 1979a, 1980, 1981; Ettensohn and Dever, 1979; Ettensohn and others, 1988) (Fig. 13).

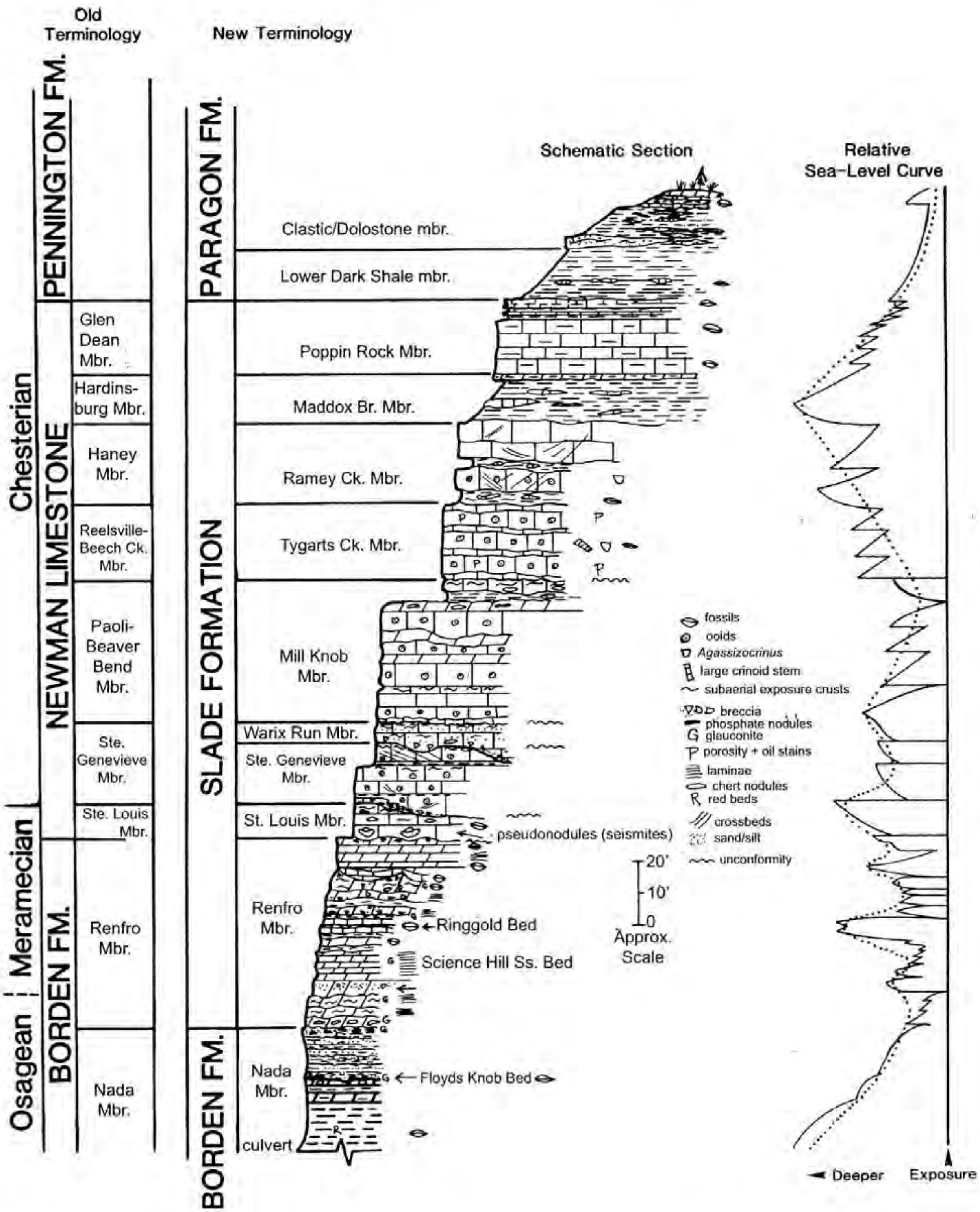

Figure 12. Schematic drawing of the section at Bighill, showing both old (McFarlan and Walker, 1956) and new (Ettensohn and others, 1984) stratigraphic terminologies and a generalized, relative sea-level curve for the section. Parts of two third-order tectono-eustatic cycles and many smaller fourth-order cycles, of probable glacio-eustatic origin, are apparent.

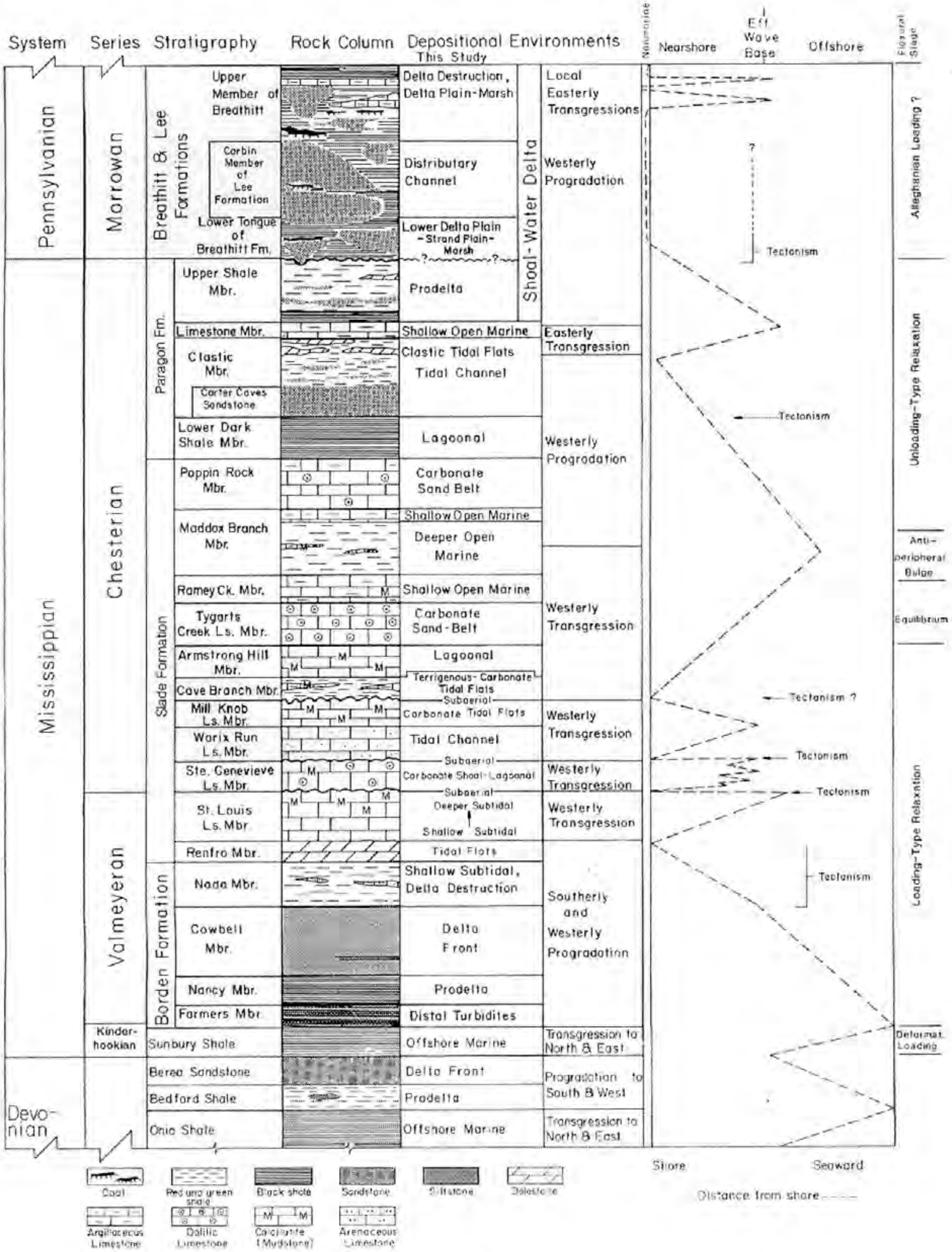

Figure 13. Generalized sequence of Upper Devonian and Carboniferous units in east-central Kentucky, showing inferred succession of environments, larger environmental continua, local tectonic events, and probable flexural stages (see Fig. 16) (from Ettensohn, 1992).

Transition to Carbonate Deposition: Upper Part of the Nada Member, Borden Formation

As already discussed, the Borden Delta Complex reflects the flooding of the Appalachian Basin with relaxational clastic deposits following the final Early Mississippian tectophase of the Acadian Orogeny (Ettensohn, 1994, 2001, 2004; Ettensohn and others, 2002). This deltaic wedge, known as the Price-Pocono, Grainger, or Borden Delta, immediately succeeded basinal, black-shale deposition in south-central parts of the basin. Price-Pocono clastics generally reflect subaerial parts of the delta complex, whereas the Grainger and Borden represent more distal, subaqueous parts of the complex.

The Nada is the upper member of the Borden Formation, and at least 37 feet (11.3 m) are present at the base of the middle exposure (Fig. 12, Plate 12). We will only discuss here the upper 23 feet (7.0 m) of the unit, however, since this is the most significant part for understanding the transition to carbonate deposition in the overlying Slade Formation.

In contrast to the greenish gray, dark gray, or grayish red silty mudstones and shales that comprise underlying parts of the unit, the upper part of the Nada is largely composed of pale blue-green (5BG 7/2), silty mudstones, siltstones, and shales, in what some workers have called a verdine facies (see for example Odin, 1985, 1988, 1990). Other parts of the unit are rich in glauconite, comprising a glaucony facies (see for example Odin and Matter, 1981; Odin, 1988).

Lithology. The upper part of the Nada begins with two rusty brown beds of argillaceous, silty, glauconitic dolostone, separated by greenish gray shale. The base of each dolostone is a glaucony- and phosphorite-rich pause or lag horizon, containing reworked fish bones and teeth, invertebrate fossil debris, and phosphorite nodules. Each bed is amalgamated, contains sparse fossil fragments, and was probably composed of calcareous-mud and glaucony pellets. Sedimentary structures are subtle, but hummocky cross-beds, scours, bioturbation, and rip-up clasts are present; the top of the lower dolostone may exhibit megaripples. The upper dolostone bed contains randomly oriented silica (chalcedony and quartz) nodules that appear to replace dolostone.

The upper dolostone is overlain by mudstone and shale that contain a prominent, moderate blue-green (5BG 4/6) layer of phosphorite and glaucony. This intensely glauconitic horizon is a typical glaucony facies and forms a prominent reentrant below the first siltstone. This horizon represents the widespread Floyds Knob Bed (Stockdale, 1939; Kepferle, 1971; Whitehead, 1978; Sable and Dever, 1990), which is 0.3 feet (0.1 m) thick, has a sharp base, and is gradational upward; it contains phosphorite nodules, fossil fragments, and burrows, which are infilled with glaucony and phosphorite.

Immediately above the Floyds Knob Bed, the true verdine facies begins, and it is composed of 15 feet (4.6 m) of thin, even-bedded, blue-green siltstones interbedded with silty, glauconitic mudstones and shales of the same color. At present the siltstones are not weathered enough for sedimentary structures to stand out prominently; however, close examination reveals the presence of amalgamated beds, crude grading succeeded by micro-cross-laminae, swaley cross-beds, scours, rip-up clasts, and bioturbation at or near the tops of each siltstone. The second siltstone bed has a horizon of phosphorite nodules at its top.

The top of the Nada is bounded by another horizon of glaucony facies, up to 0.7 feet (0.2 m) thick, composed of glauconitic siltstone with phosphorite nodules that have been bored. Phosphatized gastropods, brachiopods, and fish bones, as well as bioturbation, are present.

Interpretation. The Nada Member has been interpreted to represent delta destruction after abandonment or diversion of Borden deltas in eastern Kentucky (Ettensohn, 1979a, 1980, 1981) (Plate 11). In northeastern Kentucky, where much detailed work on the Borden Delta has been done (see for example Chaplin and Mason, 1979; Chaplin, 1980), the Nada is characterized by abundant limestones and diverse faunas interpreted to represent shallow-subtidal, open-marine environments. This contrasts with the Nada in this exposure, 70 miles (117 km) to the south, in which limestones and fauna are rare. The differences are probably best explained by position relative to the Kentucky River Fault Zone, which was periodically reactivated during Mississippian time (Dever and others, 1977; Ettensohn, 1979a, 1980, 1992). Northeastern Kentucky occupied the upthrown side of the fault near the Waverly Arch, a situation that generated a very shallow platform setting conducive to carbonate deposition and diverse faunas. The Bighill locality, however, is located about 50 miles (83 km) south of the fault zone on the downthrown side in conditions that were probably much deeper. South of the fault, the Nada is composed of mostly mudstones and shales, and appears to represent an area of deep-ramp shelf muds. The ramp, however, was not below storm wave base, because the presence of debris-filled scours, starved ripples, and laminae of fossil debris probably represents distal tempestites emplaced by storm backflow (see for example Aigner, 1985). At times, moreover, sedimentation must have been slow enough to allow quiet-bottom communities of delicate bryozoans, brachiopods, and crinoids to develop on the soft bottoms.

The upward change into thicker, coarser dolostones and siltstones in the upper Nada marks a transition into more proximal, mid-ramp, storm-shelf conditions with banks formed first of transported carbonate and glaucony pellets and then silt. The presence of thicker bedding, bed amalgamation, and swaley crossbeds all point to more proximal, mid-ramp conditions (Aigner, 1985; Pashin and Ettensohn, 1987).

Between the dolostones and siltstones, however, is the glaucony- and phosphorite-rich Floyds Knob Bed (Plate 12); a similar bed also occurs at the Nada-Renfro contact. Glaucony typically forms in upper-slope and outer-shelf environments deeper than 60 m (Odin and Matter, 1981). Glaucony and phosphorite commonly occur together in areas transitional between deep and shallow waters, slightly deficient in oxygen, with slightly lower-than-normal pH, and where clastic sedimentation is very slow to nonexistent (Carozzi, 1960; Hatch and Rastall, 1965). Initial precipitation of phosphorite may have been related to episodic upwelling into the area (see for example Carozzi, 1960; Prévôt and Lucas, 1990), but the fragmentation, boring, and probable rolling of many nodules suggest repeated reworking and concentration, perhaps by storm currents, during times of sediment starvation.

The verdine facies that characterizes the uppermost Nada reflects the in situ modification of clays by cation exchange in very iron-rich waters (Odin, 1988). Although the iron content of the waters had to be high, overall sediment influx was low, and in modern settings the verdine facies occurs in extensive, subtropical to tropical, continental-margin settings in waters 20 to 60 m deep, 30 to 60 km from the coast (Odin, 1988; Thamban and Rao, 2000). These conditions provide some idea of how the transitional Nada-to-Renfro shelf must have appeared.

Overall, the presence of carbonate- and clastic-rich storm deposits in the upper Nada reflects shallowing into upper-slope, storm-shelf conditions, whereas the co-occurrence of verdine, glaucony, and phosphoritic facies strongly supports sharply reduced clastic sedimentation on shallowing, subaqueous parts of an abandoned delta, subject to reworking by storms in an upper-slope to ramp setting. In final parts of this discussion, we will suggest that the shallowing and reduced clastic influx have both eustatic and tectonic causes (Ettensohn, 1994, 2004; Ettensohn and others, 2002). Nonetheless, the important point is that establishment of these conditions by the end of Nada deposition generated the shallow, subtropical, reduced-clastic platform-to-ramp setting necessary for the beginning of Slade carbonate deposition.

Slade Formation

The Slade Formation consists largely of shallow-water, Middle (late Osagean–Meramecian; Visean) and Late (Chesterian; Serpukhovian) carbonates (Figs. 12–13), formerly known as the Newman Limestone in east-central and northeastern Kentucky (see for example Weir and others, 1971). In 1984, the name "Newman" was restricted to the type outcrop belts on Pine and Cumberland Mountains in southeasternmost Kentucky and adjacent states, and the name "Slade Formation" was instated throughout east-central Kentucky (Ettensohn and others, 1984). The concept of the unit was enlarged, new units were designated, and others were renamed. A comparison of old and new unit nomenclature is provided in Figure 12

Renfro Member

The Renfro Member is unusually thick and well exposed in this cut, consisting of 56 feet (17 m) of grayish orange to pale orange dolostone with a few, thin, interbedded limestones and shales (Fig. 12, Plate 13). The contact with the underlying Nada is sharp and represents another glaucony- and phosphorite-rich lag horizon, but elsewhere in east-central Kentucky the contact may be gradational, with intertonguing between the two members. The Renfro in this exposure exhibits many complexities and, for ease of description, is subdivided into three parts by the two prominent limestone intervals.

Lower Part. The lower part, about 33 feet (10 m) thick, begins with glauconitic, hummocky crossbedded dolarenites that pass upward into ribbon-bedded dolosiltites, which contain breccias, reworked glaucony, and possible subaerial exposure crusts; prominent horizons of nodular, displacive chalcedony and quartz are present locally (Plate 14). This interval is interrupted by a 4-foot-thick (1.2-m-thick) bed of sandy dolostone, which was first recognized by Dever and Moody (1979a) as an equivalent to the Science Hill Sandstone of Lewis and Taylor (1979). Overlying the Science Hill equivalent is about 9 feet (2.7 m) of massive dolosiltite with four shoaling-upward cycles of glauconitic, bioturbated, ribbon-bedded dolostone passing upward into laminated dolostones; rare mudcracks are present. This interval is capped by about 5 feet (1.5 m) of massive dolosiltite to skeletal dolarenite with rare fragmented fossils and subtle hummocky crossbedding, which underlie the first limestone.

The lower part of the Renfro represents a continuation of the regressive, shallowing-upward regime begun in the underlying Nada. The lowermost dolarenites with hummocky crossbeds seem to represent continuation of storm-shelf conditions, only in a very shallow, upper-ramp setting. The presence of glaucony in these dolostones also confirms the storm-transport origin, because glaucony typically forms in deeper, less-oxygenated waters than those represented by the Renfro. The transport of glaucony into very shallow, tidal-flat or evaporitic settings is not unusual, however, and has been reported from similar settings (Ryan and Hillier, 2002).

Nodular chalcedony and quartz are locally common in the Renfro (Plate 14) and probably represent the replacement of evaporitic anhydrite with silica (see for example Chowns and Elkins, 1974). Although silica nodules are also present in the upper Nada dolostone, they are typically isolated and randomly oriented, suggesting a replacement origin. In the Renfro, however, the nodules occur in distinct horizons, commonly at breaks in sedimentation, coalesce with each other, and displace dolomitic and green clayey muds; they also occur with likely solution breccias and other intertidal-supratidal features. Nodular anhydrites are displacive and typically occur in shallow standing waters or more landward sabkha environments (Boggs, 2001; Tucker, 2001). Together all the information suggests that at least lower parts of the Renfro reflect the transport of nearshore carbonate muds and more distal, outer-shelf, verdine- and glaucony-facies muds, as well as open-marine waters, into very proximal, shallow-shelf lagoons and pans, where high temperatures and evaporation formed concentrated solutions that displacively precipitated anhydrite in the lagoonal sediments. The common occurrence of these nodules at breaks in sedimentation (Plate 14) suggests that evaporation and precipitation reflect the last stage in lagoon or sabkha development before marine inundation and the beginning of the next cycle. Replacement of anhydrite by silica is common (see for example Chowns and Elkins, 1974; Tucker, 2001), but the necessary thin-section work to verify its presence here has yet to be completed.

The regression and shallowing upward continue with several thin, cyclic sequences, composed of ribbon bedding, undulating laminae, and possible exposure crusts, and probably culminates with exposure just below the Science Hill equivalent. The Science Hill equivalent has been interpreted to represent distal parts of a highly constructive, shoal-water delta (Lewis and Taylor, 1979), but it may also be a lowstand fan on an underlying surface of exposure. The overlying interval of cyclic ribbon beds and laminites represents a return to alternating shelf-lagoon/intertidal-supratidal conditions, whereas the overlying crossbedded dolarenites represent inception of transgression and return to slightly deeper, storm-shelf conditions.

Middle part. The middle part of the Renfro begins with about 5 feet (1.5 m) of greenish gray to bluish gray, argillaceous limestone and interbedded shale with bioturbation and locally abundant corals and brachiopods. The top of the limestone is erosionally truncated and slightly melanized, suggesting ephemeral exposure. This limestone is probably equivalent to the widespread Ringgold bed of Dever (1990) in equivalents to the south (Fig. 12). The overlying 12 to 15 feet (3.7 to 4.6 m) includes massive, ribbon-bedded to laminated dolostones like those below, but they have been subsequently altered by exposure, solution, and pedogenic processes. Up to six paleosols, indicated by brecciation (Plate 15), microkarst, soil teepees, subtle exposure crusts, and truncation, are present, and each caps major dolostone unit. The last dolostone unit exhibits a prominent erosional knoll with relief up to 3.5 feet (1.1 m), around which the overlying limestone onlaps (Plate 16).

The Ringgold limestone equivalent that begins this part of the Renfro represents regional transgression and a return to shallow, open-marine, mid-ramp conditions; the muddy limestones and fossils suggest that the environment was situated below normal wave base. The paleosols capping each unit, however, indicate rapid sea-level lowering and exposure. Overlying dolostones and shales represent a return to cyclic shelf-lagoon/intertidal-supratidal environments with a few brief incursions of deeper-water or storm-shelf conditions. Each environmental interval, however, was abruptly ended by sea-level drop, exposure, and the development of a paleosol. The three massive dolostone units in the interval contain pockets of breccia, which are probably solution breccias, reflecting either karstic solution or dissolution of evaporites, both of which have been noted in equivalent rocks to the south (Dever and others, 1979a; Dever and Moody, 1979b; Dever, 1990).

Upper Part. The upper part of the Renfro begins with 1 to 3 feet (0.3mto 0.9 m) of a second limestone, which laps onto erosional highs in the underlying dolostone. The limestone is an argillaceous, fine-grained calcarenite with subtle cross laminae and fossil corals, bryozoans, brachiopods, and crinoid debris; individual parts of the unit seem to form sand waves. Two feet (0.6 m) of dolarenite with colonial rugose and tabulate corals and crinoid debris sharply overlies the limestone and passes upward into 5.5 feet (1.7 m) of ribbon-bedded to laminated dolosiltites with possible exposure crusts at the top. The overlying 8.5 feet (2.6 m) of dolostone consists of thin-bedded dolarenites to dolosiltites with bands of fenestrate bryozoan, brachiopod, and crinoid fossils. This part of the Renfro appears to have been altered throughout by subaerial exposure crusts, and the upper 2 feet exhibit breccias, erosional relief, and cavities filled with shale from the overlying St. Louis.

The upper 12 to 19 feet (3.7 to 5.8 m) of the Renfro represents two smaller transgressions (Fig. 12). The lower one begins with a shallow, open-marine, mid-ramp limestone succeeded by shelf-lagoonal and intertidal-supratidal dolostones, capped with a probable paleosol. The overlying sequence, however, reflects rapid transgression and establishment of shallow, open-marine, muddy-shelf environments, which at the top of the Renfro appear to have been abruptly exposed, for there is no intervening, shoaling-upward component. The top of the Renfro has been interpreted to represent a regional, mid-Valmeyeran unconformity (Ettensohn, 1994).

Interpretation. The Renfro overall reflects the declining or shallowing parts of a major third-order eustatic cycle and the final part of a major regional regression that ended Middle Mississippian deltaic deposition throughout most of the Appalachian Basin and adjacent areas (Plate 14), although middle and upper parts of the Renfro also show brief episodes of minor transgression (Fig. 12). The resulting delta destruction, which is first apparent in the Nada at this exposure, transforms deeper-water, Borden delta-front slopes into a shallow-water, carbonate-ramp setting, on which overlying Middle and Upper Mississippian Slade carbonates were deposited. The transition from clastics to carbonates and from deeper water to very shallow water is clearly seen in the Nada-Renfro succession. On a regional scale, however, these changes necessitate the complete shut-off or diversion of clastic influx and an episode of eustatic lowering or regional uplift, or both (Ettensohn, 1994; Ettensohn and others, 2002). The likely tectonic and eustatic causes are discussed in a later section.

St. Louis Member

The St. Louis Member sits unconformably on top of the Renfro at the first major bench in the exposure (Fig. 12, Plate 12). The member is up to 12 feet (3.7 m) thick and consists of three widespread subunits, A, B, and C (Philley, 1971; Dever, 1980). The lowermost unit, A, is composed of thin- to medium-bedded, light gray, skeletal calcarenites containing clasts of reworked Renfro dolostone and sparse, colonial rugose corals. As is typical of the St. Louis in other places (Dever, 1980), the unit contains large elliptical nodules and ball-and-pillow structures of medium gray to pale green dololutite. Unit B consists of thin-bedded, light gray, fossiliferous calcarenites and calcilutites with interbedded greenish gray shales; chert nodules occur locally; brachiopods, bryozoans, and crinoid debris are common. Unit C, at the top, is composed of massive, medium- to thick-bedded, fine-grained calcarenite to calcilutites; irregular chert nodules, sometimes containing colonial rugose corals, are common. The dark gray or brownish gray color (Plate 12) is a product of melanization, and along with rare subaerial exposure crusts and erosional truncation, is a product of subaerial exposure and pedogenesis on an unconformity below the Ste. Genevieve Member (see for example Ettensohn and others, 1988). At the north end of the exposure, unit C is up to 5 feet (1.5 m) thick, but toward the south it is truncated below the Ste. Genevieve to less than 2 feet (0.6 m). Colonial rugose corals in the St. Louis have been identified as *Acrocyathus floriformis* and *A. proliferus,* which are common St. Louis guide fossils (see for example Butts, 1922).

Interpretation. The three-part, fining-upward St. Louis sequence has been interpreted to be a transgressive sequence (Ettensohn and Dever, 1979; Dever, 1980), with unit A representing mid-ramp skeletal shoals at wave base; unit B representing lower-ramp shallow, open-marine conditions, generally below wave base; and unit C representing deeper, open-marine conditions, well below wave base (Plate 11). The abrupt alteration of these subtidal carbonates by subaerial diagenesis, pedogenesis, and erosion along an unconformity, without an intervening regressive interval, suggests rapid drop in sea level and exposure. In northeastern Kentucky, this uplift has been clearly associated with basement structures (Ettensohn and Dever, 1979; Ettensohn and others, 1988), but the regional nature and distribution of the unconformity may suggest the importance of regional bulge uplift accompanying a tectophase of the Ouachita Orogeny (Ettensohn, 1993). The dolomitic nodules and ball-and-pillow structures in unit A are similar to soft-sediment deformation associated with seismicity and may be seismites (see for example Ettensohn and others, 2000; Greb and Dever, 2002).

Ste. Genevieve Member

The Ste. Genevieve Member unconformably overlies the St. Louis and begins with a transgressive lag of reworked, dark, upper St. Louis clasts

(Fig. 12, Plate 13). The unit is 28 to 36 feet (8.5 to 11.0 m) thick and thins to the northwest because of erosion from the overlying Warix Run Member. The unit can be separated into two parts by an erosional hiatus with a dark paleosol horizon. The lower part is an intertonguing and internally truncated facies complex of thin-bedded, light greenish gray calcilutites; massive, dark brownish gray, clotted, birdseye calcilutites; and skeletal-oolitic calcarenites (Plate 17). The calcarenites contain, high-angle, planar-tabular crossbeds, some of which show a herringbone pattern. The complex is 18 feet (5.5 m) thick, and facies are uniformly truncated along a melanized, paleosol horizon up to 3 feet (0.9 m) thick with breccias and thin, wispy, subaerial exposure crusts. The upper part of the member begins with breccias eroded from the lower unit and consists of 10 to 18 feet (3.0 to 5.5 m) of crossbedded, skeletal/oolitic calcarenite and interbedded dark calcilutites. *Platycrinites penicillus,* a Ste. Genevieve guide fossil (Butts, 1922), has also been found in the upper unit. The top of the unit is unconformable with the overlying Warix Run, and contains remnant soil teepees, root tubules, breccias, faint crusts, and melanization (Plate 18), all indications of a major paleosol. In northern parts of the cut, the Ste. Genevieve shows several relatively recent solution pits, some of which are lined with flowstone.

Interpretation. Both parts of the Ste. Genevieve are interpreted to represent very shallow, high-energy, tidal sandbar belts of migrating bars and shoals; the calcilutites are interpreted to represent lime-mud accumulation in protected lagoons behind the bars and shoals (Fig. 13, Plate 17). As the bars migrated, some of the lime muds were eroded, giving the complex truncation surfaces in the unit. The unusual, dark brownish gray, clotted calcilutites from the lower part are interpreted to represent migrating mud-mound facies that formed on erosional highs through the accretion of mud clasts ripped up from nearby lagoons by storms. Similar banks have been described from present-day Florida Bay (Ginsburg, 1972). The peculiar birds-eye texture reflects early sparry-calcite infilling of voids between mud clasts. The Ste. Genevieve contains prominent paleosol horizons at its middle and top (Plate 14), which represent episodes of subaerial exposure related to eustatic or tectonic causes, or both (Ettensohn and others, 1988; Ettensohn, 1993).

Warix Run Member

This member is very similar to the Ste. Genevieve from which it was formally separated in 1984 (see Dever, 1980; Ettensohn and others, 1984) (Figs. 12–13). The main difference is the presence of quartz sand and peloids. The Warix Run also typically occupies deep erosional lows cut into underlying units (Ettensohn and Dever, 1979; Dever, 1980; Ettensohn, 1980, 1981, 1992). At this exposure, the Warix Run is separated from the Ste. Genevieve by a green sandy breccia, up to 1 foot (0.1 m) thick, containing reworked clasts of the underlying Ste. Genevieve paleosol; most of the quartz sand is concentrated near the base of the unit. Overall, the unit is a dark brownish gray, peloidal calcarenite with large, planar-tabular crossbeds (Plate 18). The unit occupies a channel cut into the Ste. Genevieve here and thickens to the north as the channel deepens in that direction (Fig. 12). Accordingly, thickness varies from 3 to 13.5 feet (0.9 to 4.1 m) because of the erosional contact with the Ste. Genevieve (Plate 18). The dark color of the unit is the product of melanization, which together with sparse, wispy exposure crusts and root tubules indicates periodic exposure. Contact with the overlying Mill Knob member is gradational. Fossils of any kind are extremely rare.

Interpretation. The unconformity atop the Ste. Genevieve is a major regional unconformity, which probably reflects bulge-related uplift during a tectophase of the Ouachita Orogeny (Ettensohn, 1993, 1994). Unconformity formation was accompanied by deep erosion into underlying units, and the Warix Run represents initial lowstand flooding of these lows (Ettensohn and Dever, 1979; Dever, 1980; Ettensohn, 1980, 1981, 1992). The initial flooding is reflected in the basal, sandy breccia, whereas other parts of the unit represent migrating, tidal sand bars and dunes in the high-energy tidal channels (Plate 11). Combinations of eustatic drawdown and rapid upward accretion of the dunes into sea level apparently left the sands periodically exposed to pedogenic processes.

Mill Knob Member

The Mill Knob Member is composed of about 45 feet (13.7 m) of cyclically alternating shaly calcilutites, oolitic/skeletal calcarenites, and birdseye calcilutites or dolostones, locally capped with paleosols (Fig. 12, Plates 11, 19). Bedded or nodular cherts are common. Although every cycle is slightly different, most of the cycles are shoaling-upward, and five such cycles are present in this exposure (Plate 19), two of which are capped with paleosols. The upper 5 to 6 feet (1.5 to 1.8 m) of the member are unusual in the presence of a coarsening-upward cycle containing much shale (Fig. 12, Plate 20). This cycle begins with a basal zone that reflects reworking of the underlying paleosol and grades upward into greenish gray mudstones and shales interbedded with thin calcilutites, which are laminated and contain possible rare mudcracks. The top of this cycle is erosionally truncated and capped with a major paleosol, exhibiting melanization, exposure crusts, breccias, and soil teepees (Plate 20). Complete fossils are uncommon throughout the unit.

Interpretation. Unlike other underlying Slade units, the Mill Knob does not reflect a new transgression, but rather a continuation of flooding begun in the Warix Run and succeeding highstand conditions. The Warix Run and Mill Knob are parts of a single, large fining- and shoaling-upward regressive sequence and intertongue locally, although not at this exposure (Fig. 13). As the large tidal channels filled with Warix Run sands, shallow seas apparently expanded outward above and beyond the channels to form a shallow-water, tide-dominated shelf, subject to sea-level variations that produced the shoaling-upward cycles noted here and elsewhere. Rare cycles may begin with nodular, shaly, fossiliferous calcilutites that grade into bedded calcilutites (Plate 19), representing lower-ramp, shallow, open-marine environments, but only one such cycle is present here at the base of the unit. More commonly, a cycle begins with crossbedded skeletal/oolitic calcarenite, representing a tidal sand-belt or shoal, and grades upward into a birdseye calcilutite or dolostone (Plate 19), representing protected lagoonal and intertidal environments. Occasionally, the upper calcilutite or dolostone is altered by pedogenic (paleosol) features (melanization, breccias, exposure crusts, soil peds and teepees) (Plate 20), indicating complete exposure. The last cycle at the top of the Mill Knob is atypical in its high shale content and coarsening-upward nature (Plate 20), and appears to represent a short-lived, transgressive succession, reflecting the transition from paleosol to tidal flat and lagoon. Whether or not sediments representing deeper environments were ever present is uncertain, because the succession was truncated and subaerially exposed along the major regional unconformity that separates lower and upper parts of the Slade Formation. The unconformity is widespread in eastern Kentucky and is probably related to a combination of local tectonism and eustacy.

Tygarts Creek Member

This unit consists of 23 feet (7 m) of light gray to white, oolitic/skeletal calcarenite organized into five shoaling-upward cycles (Fig. 12, Plate 19). Each cycle begins with 0.4 to 0.5 feet (0.1 to 0.2 m) of thin-bedded, fine-grained, argillaceous calcarenite with thin shale partings and grades upward into 3 to 4 feet (0.9 to 1.2 m) of massive, crossbedded oolitic/skeletal calcarenite. Petrographically, the calcarenites are grainstones, and grainstones in the first two cycles, especially oolitic parts, exhibit local intergranular and moldic porosity containing dead oil (Plate 20); when the weather is warm, oil oozes out of the exposure from these areas. The upper 1.5 feet (0.5 m) of the unit is a dark, argillaceous, skeletal calcarenite. The Chesterian guide fossil *Agassizocrinus* and a large unidentified crinoid stem, which is indicative of the mid-Chester Gasper Stage (McFarlan and Walker, 1956), are both present in the unit. Complete fossils are rare. In northern parts of the exposure, recent karstic processes have affected the unit.

Interpretation. The unit has been interpreted to represent a mid-ramp, high-energy, carbonate sand belt (Ettensohn, 1977, 1980) (Fig. 13), formed at the point where wave base impinged on the bottom. In this setting, ooids were formed and skeletal debris was comminuted. The agitated, unstable bottoms were not conducive for most benthic fauna, and only a few heavily constructed, vagile gastropods and crinoids regularly inhabited the sands (Ettensohn, 1975). The cycles (Plate 19) apparently reflect episodes of abrupt deepening followed by upward aggradation into wave base, whereas the dark, muddy upper part of the unit is a transition into rocks representing deeper open-marine environments above.

The Tygarts Creek is part of a major regional transgression that inundated the once-exposed Mill Knob surface below, and it is part of a sequence that would normally include underlying units representing tidal-flat and lagoonal environments (Ettensohn, 1977, 1979a, 1980). Those units (Cave Branch Bed and Armstrong Hill Member) (Fig. 13) are absent here, perhaps because of erosion accompanying sand-belt formation.

Ramey Creek Member

At this exposure, the Ramey Creek Member is 28 feet (8.5 m) thick and consists of two shoaling-upward cycles (Fig. 12, Plate 21). The basal parts of each cycle contain bluish-gray, fossiliferous shales or mudstones that intertongue with crossbedded, argillaceous calcarenites, nodular calcilutites, or dolarenites. Overlying parts of each cycle consist of massive, medium- to thick-bedded skeletal/oolitic calcarenites, with high-angle crossbeds, some of which are herringbone (Plate 21). The lower body of calcarenite clearly intertongues with the underlying shaly section. The shaly parts of the unit are more fossiliferous than any other unit at Bighill, containing abundant ramose and fenestrate bryozoans, brachiopods, corals, gastropods, crinoids, and blastoids; bioturbation is ubiquitous. Among the most distinctive fossils are the bryozoan *Archimedes*, the brachiopods *Anthracospirifer*, *Composita*, and *Diaphragmus*, the coral *Zaphrentoides spinulosus*, the crinoid *Pterotocrinus*, and the blastoid *Pentremites*. *Agassizocrinus*, which occurs in the upper calcarenite, and *Pterotocrinus* are Chesterian guide fossils.

Interpretation. The Ramey Creek represents shallow, open-marine, outer ramp deposition just seaward of the carbonate sand belt (Ettensohn, 1977, 1980) (Plate 11). Deposition occurred in a lacework of tidally influenced shoals, where skeletal/oolitic sands were deposited, and in deeper, intervening basins, where calcareous mud and silt, as well as argillaceous muds, predominated. The argillaceous calcarenites, nodular calcilutites, and interbedded shales and mudstones represent basinal deposition below wave and tidal influence. The presence of crossbedded calcarenites among these lithologies apparently reflects storm-generated backflow currents from the sand belt and intervening shoals. The Ramey Creek contains the most diverse and populous fauna of any Slade member. The increased diversity and abundance of fossils, abundant burrowing, and the increased presence of muds (low sparite/micrite ratios) indicate the deeper (generally below wave base), more stable, offshore nature of Ramey Creek environments.

Maddox Branch Member

The Maddox Branch Member includes 12 to 14 feet (3.7 to 4.3 m) of predominantly dark gray to bluish gray clay shale and mudstone, which forms a prominent bench or reentrant above the upper Ramey Creek calcarenite (Fig. 12, Plate 21). Lenses and nodular layers of argillaceous calcarenite or calcilutite are commonly interbedded with the shales. Some of the calcilutite nodules are brecciated or contorted, and at this exposure, a crossbedded calcarenite reworked into megaripples is the most prominent interbed. The uppermost 1.5 to 2 feet (0.5 to 0.6 m) of the unit is an argillaceous dolosiltite/dolarenite, which appears to be a part of the overlying Poppin Rock Member (Plate 21); it is burrowed, fossiliferous, and locally absent because of post-depositional erosion. The Maddox Branch is commonly fossiliferous, but contains a fauna with less diversity and abundance than overlying or underlying members. The most common faunal elements include fenestrate and rhomboporid bryozoans, the productid brachiopod *Diaphragmus elegans*, and crinoid debris.

Interpretation. The predominance of argillaceous and calcareous muds in the Maddox Branch Member reflects deposition in a deeper, open-marine, outer-ramp environment in quiet conditions well below wave base (Fig. 13). The crossbedded and megarippled calcarenite probably represents a major episode of storm deposition. Brecciation and contortion in some calcilutites appear to have been caused by compaction. The low-diversity fauna dominated by productid brachiopods and delicate bryozoans may reflect restriction by soft muddy bottoms.

The Maddox Branch apparently represents the culmination of a major third-order deepening cycle and of the transgression begun at the base of the Tygarts Creek Member (Figs. 12, Plate 21), and as a consequence should contain a maximum flooding surface before the shallowing-upward, progradational (highstand systems tract) sequence begins. Although the time of maximum water depth in this sequence almost certainly occurred during Maddox Branch deposition, in this location this time is not represented by a surface of condensation or starvation. In some other exposures, however, this time may be represented by an interval of dark, organic-rich shale. Overlying parts of the unit then reflect the beginning of middle Chesterian shallowing and progradation. Accordingly, uppermost dolomitic parts of the unit (Plate 17), which are seldom well preserved, have been interpreted to represent a return to nearer-shore, shallow, open-marine conditions (Ettensohn, 1977, 1980, 1981).

Poppin Rock Member

The Poppin Rock Member includes 22 feet (6.7 m) of bluish gray, argillaceous, thin- to medium-bedded calcarenite (Fig. 12, Plate 21). Crossbedding and scours are common in the beds, and the tops of these beds typically exhibit slightly reworked bryozoan-brachiopod-crinoid communities. Fenestrate (e.g., *Archimedes* and *Lyropora*), ramose, and rhomboporid bryozoans; the brachiopods *Diaphragmus, Composita,* and *Anthracospirifer;* the coral *Zaphrentoides spinulosus;* and echinoderm debris are common on these bedding planes. The stemless crinoid *Agassizocrinus conicus,* a middle to late Chesterian guide fossil (Burdick and Strimple, 1982), occurs commonly within the calcarenite layers.

The upper 5.5 feet (1.7 m) of the unit becomes thinner bedded, finer grained, and more shaly; fine-grained calcarenites and calcisiltites predominate. Fossils are more abundant, and bioturbation on upper bedding surfaces can be intense.

Interpretation. The more massive, crossbedded skeletal calcarenites of the lower Poppin Rock Member represent deposition on a carbonate sand belt that was the progradational analogue of the underlying, transgressive Tygarts Creek Member (Fig. 13). The major difference between the two units is the presence of detrital muds and quartz in the Poppin Rock. Each bed in the Poppin Rock represents a migrating shoal, which became stabilized long enough to support a community on its top. The upper, more shaly part of the unit represents a somewhat deeper, more protected transition from the agitated sand belt to a quiet, deeper, back-sand-belt lagoon, represented by immediately overlying parts of the Paragon Formation (Ettensohn and Chesnut, 1979; Chesnut and Ettensohn, 1988).

Paragon Formation

The Paragon Formation is largely composed of Upper Mississippian, shallow-water, marginal-marine clastics, which were formerly included in the Pennington Formation (Figs. 142–13, Plate 21). Because these rocks are not closely related to or correlative with Pennington rocks in the type Pine and Cumberland Mountain outcrop belts in extreme southeastern Kentucky, all Upper Mississippian clastic rocks above the Slade carbonates on the east-central Kentucky outcrop belt were incorporated into the newly designated Paragon Formation (Ettensohn and others, 1984). The rocks are rarely exposed because of the predominance of shale and because large parts of the unit are commonly cut out by post-depositional Pennsylvanian erosion near the Mississippian-Pennsylvanian transition.

Lower Dark Shale Member. A complete exposure of this unit is present at the top of the southeastern cut, but it is being rapidly covered with slump debris and vegetation (Plate 21). The unit contains up to 16 feet (4.9 m) of greenish gray to dark gray, silty, fissile clay shale, typically with macerated plant debris (Fig. 12). The unit may be truncated at the top by the clastic/dolostone member and intertongues at its base with the Poppin Rock Member of the Slade Formation. Lower parts of the unit contain argillaceous calcarenite lenses and layers, which may be very fossiliferous and bioturbated. Fossils are very similar to those in the underlying Poppin Rock Member, but are generally more abundant. To the south, this unit has been informally called the Sloans Valley member (Ettensohn and Chesnut, 1979; Chesnut and Ettensohn, 1988).

Interpretation. The shales in this member have been interpreted to represent a protected shelf lagoon behind the Poppin Rock sand belt (Fig. 13). The fossils found in lower parts of the unit formed communities on small sandy shoals that developed in the lagoon as spillover lobes from the Poppin Rock sand belt (Ettensohn, 1977; Ettensohn and Chesnut, 1979; Chesnut and Ettensohn, 1988). Presence of plant debris suggests proximity of terrestrial source areas to the east.

Clastic/Dolostone Member. This unit is up to 17 feet (5.2 m) thick and begins with a fining-upward sequence of sandstone and shale. Basal parts of the unit consist of a thin- to medium-bedded, micaceous, medium-grained sandstone, which weathers to a yellowish orange or brown color; it typically contains ripples, crossbeds, and shale rip-up clasts, and at this exposure is up to 4 ft (1.2 m) thick. This sandstone grades laterally and vertically into 2.5 feet (0.8 m) of thin-, flaser-bedded, micaceous, fine-grained sandstones with ripples, which in turn grade upward into 6 feet (1.8 m) of dark gray shale and laminated siltstones and fine-grained sandstones with wavy and lenticular bedding; bioturbation is ubiquitous. Sharply overlying the sands is 3.5 feet (1.1 m) of interbedded dark shale and fossiliferous limestone. The limestones are succeeded by 1 foot (0.3 m) of greenish gray shale and a foot of yellowish orange dolostone (Fig. 12). The uppermost 13 feet (4.0 m) of the Paragon section just described is only visible at a grassy cut to the south of the main exposure on Burnt Ridge Road.

Interpretation. This succession of variable lithologies has been interpreted to represent a fining-upward, tidal-flat sequence on the shoreward side of the lagoon (Ettensohn, 1977; Ettensohn and Chesnut, 1979; Chesnut and Ettensohn, 1988) (Fig. 13). The presence of thin, fossiliferous limestone units apparently reflects brief transgressive episodes that inundated the tidal flats. North of this area, this member of the Paragon is composed largely of clastics, reflecting proximity to northern source areas; to the south, in contrast, the same interval is largely laminated dolostones, probably reflecting the absence of major clastic influx (Ettensohn and Chesnut, 1979; Ettensohn and others, 1984; Chesnut and Ettensohn, 1988). Apparently, this exposure occurs in a transition area between the two.

Tectono-Stratigraphic Implications

The Mississippian section at Big Hill is very typical of Mississippian rocks throughout much of the Appalachian Basin in that it consists of a three-part succession composed of Lower and Middle Mississippian clastics (Borden Formation), Middle and Upper Mississippian carbonates (Slade Formation), and uppermost Mississippian clastics (Paragon Formation). This succession has traditionally been interpreted to represent post-Acadian clastic influx, carbonate deposition accompanying tectonic quiescence, and renewed clastic influx marking inception of the Alleghanian Orogeny (see for example Perry, 1978; Chesnut, 1991). More recent interpretations based on flexural models suggest instead that these rocks are part of a third-order, Mississippian sequence that largely represents an Acadian relaxational response, and that the

Alleghanian Orogeny did not begin until Early Pennsylvanian time (Ettensohn and Chesnut, 1989; Ettensohn, 1994, 2001). In contrast, some of the smaller subsequences, such as the St. Louis, Ste. Genevieve, and Warix Run–Mill Knob in the lower Slade (Figs. 12–13), have been interpreted to represent eustatic and tectonic events superimposed on the larger relaxational response.

The flexural models predict a distinct sequence of lithologies and unconformities similar to those seen here and in other parts of the Appalachian Basin. Of the various flexural models, the viscoelastic models of Quinlan and Beaumont (1984) and Beaumont and others (1987, 1988) seem to most consistently and easily explain the observed sequence. Moreover, the relatively long time involved in developing critical parts of the Mississippian section (approximately 20 m.y.) is most typical of relaxation in viscoelastic regimes (Sinclair and others, 1991, p. 600). These models are briefly described below.

Models. Most flexural models predict that during orogeny surface and subsurface deformational loading by flakes, blocks, thrusts, nappes, and folds produces a downwarped flexural or retroarc foreland basin cratonward of the orogen and a peripheral bulge on the cratonward margin of the basin due to regional isostatic compensation by the lithosphere (Fig. 14A). As orogeny proceeds and thrust loads shift cratonward, the foreland basin and peripheral bulge continue to migrate cratonward away from the load. Once active orogeny and thrusting cease, however, lithospheric relaxation will cause the bulge to be uplifted and migrate back toward the orogen, while the adjacent foreland basin deepens and begins to receive major sediment influx (Fig. 15A); after the basin is largely infilled and adjacent highlands eroded low, a new phase of relaxation involving lithospheric rebound begins (Quinlan and Beaumont, 1984; Beaumont and others, 1987, 1988) (Fig. 15B). Hence, even during

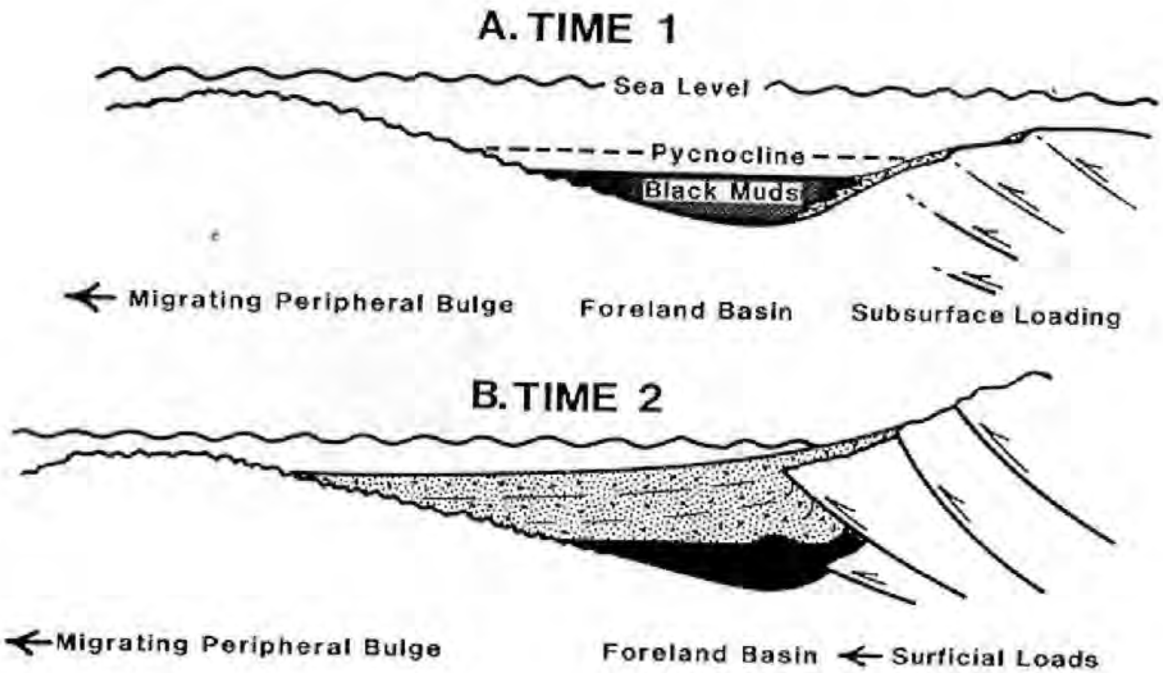

Figure 14. Schematic diagrams showing sedimentologic/stratigraphic responses of the foreland basin during two stages of loading. A. Rapidly subsiding foreland basin with little clastic influx during subsurface loading—corresponds to Sunbury/uppermost New Albany deposition in this area (see Plate 13). B. Subsiding foreland basin infiled with coarser clastics derived from surficial load—corresponds to Borden deposition in this area (see Fig. 16B). The pycnocline is a zone of thermohaline density stratification with decreasing oxygen content. Dark stipple: black shales. Large stipple: coarser clastic sediments. Wavy lines: unconformities.

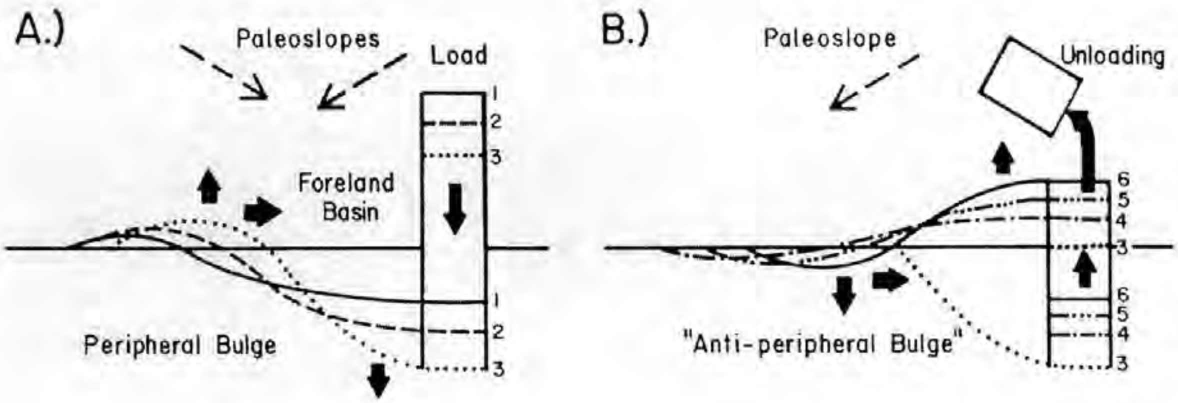

Figure 15. Schematic diagrams showing two types of flexural response to lithospheric-stress relaxation (redrawn from Beaumont and others, 1988). A. "Loading-type" relaxation when thrust migration ceases. B. "Unloading-type" relaxation when erosional unloading generates rebound near unloaded area (see Fig. 16).

times of orogenic quiescence after active thrusting, sedimentation and bulge-basin reorganization continue because of relaxation. Some of this lithospheric reorganization commonly involves reactivation of basement structures or surficial faulting (see for example Bradley and Kusky, 1986), resulting in local uplift and related facies changes in parts of the basin.

Assuming that deformational loading and relaxation go to completion, each flexural stage will generate a typical sedimentologic/stratigraphic response (Ettensohn, 1994). In following parts of the guide, we will briefly describe the nature and origin of these responses and compare them with the section at Bighill.

Unconformity Development. As convergence begins and a deformational load accumulates on the continental margin, the lithosphere responds by generating a compensating foreland basin and peripheral bulge (Fig. 14A). With cratonward movement of the thrust load in time, the subsiding basin and uplifted bulge migrate in the same direction, and erosion on the uplifted bulge generates a lower bounding unconformity throughout much of the foreland basin (Quinlan and Beaumont 1984). In proximal and central parts of a basin, closer to the locus of active deformation, however, subsidence may outstrip any effects of bulge uplift, resulting in a conformable contact at the base of the sequence. In parts of the Appalachian Basin to the north and south, this unconformity occurs at the Devonian-Mississippian boundary below the Sunbury Shale and its equivalents (Figs. 13, 16A). At our location in west-central parts of the Appalachian Basin, however, the Devonian-Mississippian boundary occurs in uppermost parts of the New Albany Shale, and is apparently conformable (Weir and others, 1971; Ettensohn, 1979b). Although we will not examine this part of the section on the field trip, this transition represents initiation of the fourth and final tectophase of Acadian Orogeny (Ettensohn, 1985).

Foreland-Basin Subsidence and Regional Transgression. As active tectonism and deformational loading ensue, rapid foreland-basin subsidence follows bulge uplift and moveout. Because initial subsidence is largely related to subsurface loading (Karner and Watts, 1983), or loading that never breaks the ocean/sea surface, no major source of orogenically derived sediment is available (Fig. 14A). In the absence of major clastic influx, suspended clay and abundant organic matter from the water column compose most sediment in the early foreland basin, and because subsidence exceeds sedimentation, the water column soon becomes stratified, and organic-rich sediments are preserved as black muds in resulting oxygen-poor environments (Fig. 14A). In the typical central Appalachian Mississippian section, this stage is represented by the Lower Mississippian black Sunbury Shale and its equivalents (Figs. 13, 16A), present here

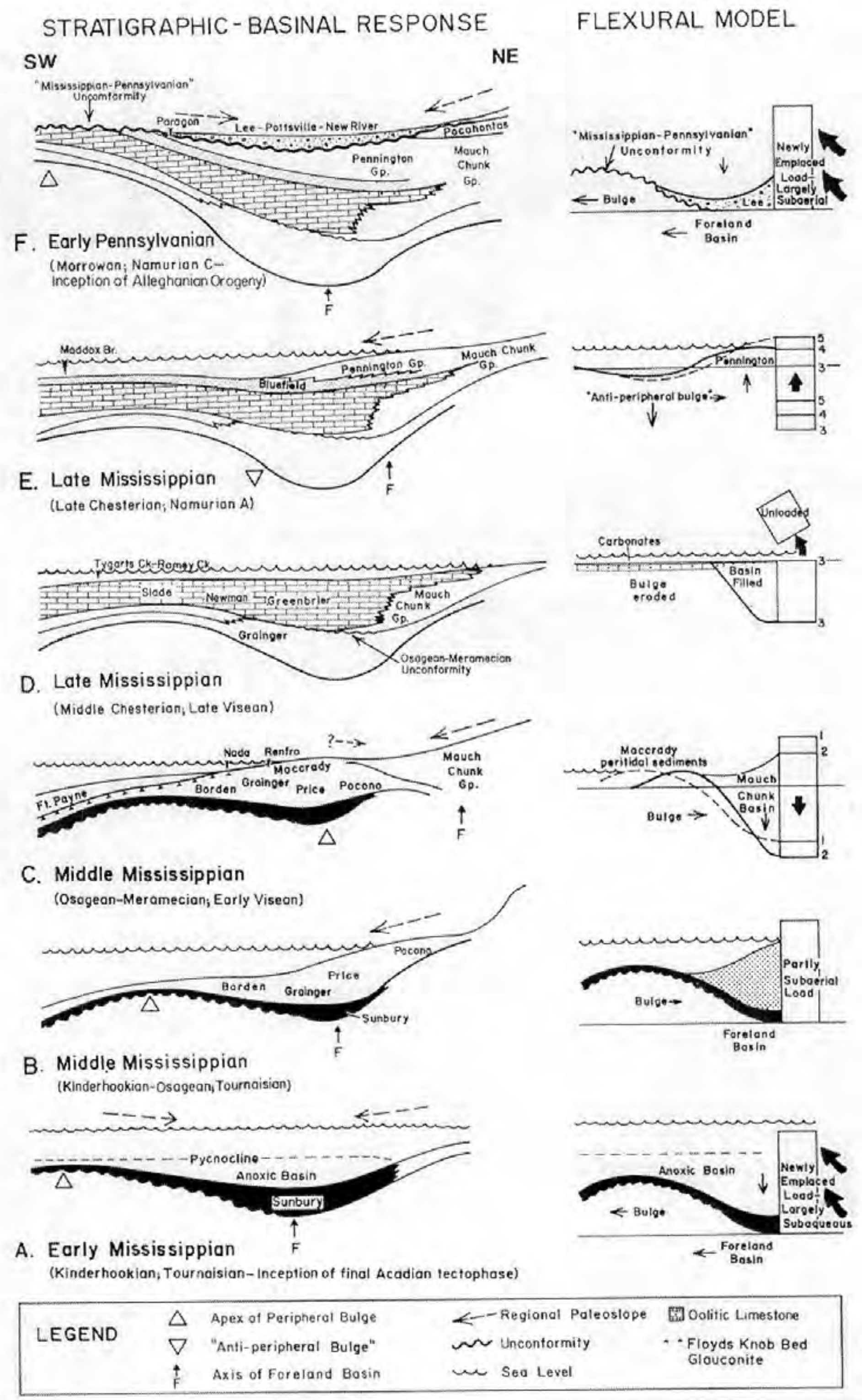

Figure 16. Schematic east–west section across central Appalachian Basin, showing the probable succession of flexural events between the last tectophase of the Acadian Orogeny (A) and the inception of the alleghanian Orogeny (F) and their sedimentary/stratigraphic responses relative to units in the central Appalachian Basin and at Bighill (adapted from Ettensohn, 1994, 2004).

in uppermost parts of the New Albany Shale (Weir and others, 1971; Ettensohn, 1979b).

Loading-Type Relaxation and Regional Regression. Once active deformation and thrust migration cease, the deformational load becomes static. The lithosphere responds to the static load by relaxing stress so that the foreland basin deepens and narrows while the peripheral bulge is uplifted and shifts toward the load (Fig. 16A) (Beaumont and others, 1988). If sea level is low at this point, bulge uplift and migration may generate a regional unconformity. Moreover, by this time, emplacement of surface loads allows drainage nets to develop so that much of the static surface load can be eroded and transported to the foreland basin as turbidites, debris flows, deltas, and tempestites, burying basinal black shales below flysch-like sediments (Fig. 14B).

This relaxation in the central Appalachian Basin is represented by turbidites and deltaic deposits in the Borden Formation and equivalents in the Grainger, Price, and Pocono formations, as well as in the lower Mauch Chunk Group in eastern parts of the basin (Figs. 13, 16B). Eastward bulge migration apparently disrupted deltaic sedimentation (Fig. 16C), however, and in distal parts of the basin west of the bulge, sediment starvation and delta destruction ensued, as seen in the Nada and in the glaucony-rich rocks of the Floyds Knob Bed (Stockdale, 1939; Kepferle, 1971); in more proximal parts of the basin on and near the bulge, deltaic sedimentation gave way to shallow open-marine and peritidal sedimentation in the Maccrady Formation and in succeeding units such as the Renfro (Fig. 16C). In more distal areas to the west and southwest, declining clastic influx gave rise to the deeper-water, cherty Ft. Payne carbonates.

The Maccrady, Ft. Payne, and equivalents such as the Renfro, however, were truncated by a regional unconformity during the Osage-Meramec transition (Fig. 16C). Although uplift on structures has been implicated locally (Warne, 1990), the influence of bulge uplift in unconformity formation is indicated by timing and distribution of the unconformity (Ettensohn, 1994). Nonetheless, delta destruction and unconformity formation are anomalous compared to other flexural sequences at similar stages, requiring an additional means of sea-level lowering—a requirement supporting the mid-Mississippian period of eustatic lowstand noted by Vail and others (1977) and Harland and others (1989). In fact, the sea-level drop and exposure reflected in Maccrady and Renfro tidal flats apparently resulted from the unique coincidence of eustatic lowstand and bulge uplift and culminated in a regional unconformity during the Osage-Meramec transition, represented here by the Renfro–St. Louis unconformity. With the decline in clastic influx and advent of shallow waters in a subtropical setting, widespread carbonate deposition in largely coeval units such as the Slade, Newman, Greenbrier, and Monteagle limestones supplanted the clastic infilling of the basin that would have normally characterized remaining parts of loading-type relaxation (Fig. 16D).

Anomalies in this phase of the sequence include three unconformities in lower parts of the Slade. Of the unconformities, the early Chesterian Ste. Genevieve–Warix Run unconformity is the most widespread, and its regional distribution parallel to the Ouachita orogen suggests bulge uplift during a Ouachita tectophase (Ettensohn 1993, 1994; Ettensohn and Pashin, 1993, 1997). The St. Louis–Ste. Genevieve and Mill Knob–Tygarts Creek unconformities, on the other hand, seem to reflect some combination of local structural reactivation and eustatic lowstand (Dever and others, 1977; Ettensohn and others, 1988) (Figs. 12–13).

Immediately following early Chesterian uplift and erosion (Ste. Genevieve–Warix Run unconformity), quartzose sands flooded parts of the central Appalachian Basin, generating local sandstones and arenaceous calcarenites such as the Loyalhanna Member of the Greenbrier, the Trough Creek Member of the Mauch Chunk, the drillers' Greenbrier Big Injun and Keener sands, and the Warix Run Member of the Slade Formation. Although these sand-rich units are interpreted to represent tidally influenced, coastal sand-flat environments (Carney and Smosna, 1989; Vest, 2000), their provenance is unknown. Their association with the early Chesterian unconformity and overlying Mill Knob transgression, however, may indicate that they are lowstand deposits with sources to the north and east generated by Ouachita bulge uplift.

Equilibrium Phase. Eventually the foreland basin fills with sediments while adjacent orogenic highlands undergo extensive lowering caused by subsidence and erosion, leading to a brief period of near-elevational equilibrium between the filled basin and the eroded highlands (Fig. 16D). This phase is the culmination of loading-type relaxation, and in the resulting, shallow low-gradient seas, a blanket of shallow-water carbonates spreads rapidly throughout the basin. In the Appalachian Basin, this phase peaked in Late Mississippian (late middle Chesterian) time prior to deposition of the lower Bluefield, Maddox Branch, Pencil Cave, or Lillydale shales, when a sheet of very shallow-water skeletal and oolitic sands, represented at Bighill by the Tygarts Creek and Ramey Creek members (Figs. 12–13), spread across much of the basin and adjacent craton (Fig. 16D). At this time, Mississippian carbonate deposition attained its greatest extent in the foreland basin.

Unloading-Type Relaxation, Regression, and Cratonward Progradation. Deposition of the carbonate blanket (Fig. 16D) is the culmination of regression begun with loading-type relaxation and is relatively short-lived, because the area of the former orogen and foreland basin soon rebounds upward in isostatic response to "unloading." A compensating "anti-peripheral bulge," or peripheral sag, develops cratonward of the rebounding area, deepens, and migrates toward it in time (Fig. 15B) (Beaumont and others, 1988). As a result, shallow-water carbonates from the previous phase are abruptly overlain by transgressive, deeper-water shales or carbonates. This brief transgressive episode is followed by a regressive, cratonward-prograding wedge of marginal-marine and terrestrial, clastic sediments as the peripheral sag fills (Fig. 16E). Because the rebounded area includes parts of the former foreland basin and parts of the already beveled orogenic upland, most derived sediment is relatively fine grained. Moreover, because this phase of relaxation begins from a state of approximate elevational equilibrium at or near sea level, a single cratonward-dipping paleoslope develops (Fig. 15B), and sediment is deposited in mainly marginal-marine or terrestrial environments.

In the central Appalachian Basin, this phase of relaxation begins with abrupt deepening in a peripheral sag represented by the deeper-water Maddox Branch Member of the Slade, the Pencil Cave or Lillydale shales of the upper Newman Limestone, and the lower Bluefield Formation (Figs. 12–13, Plate 21). The overlying clastics of the Paragon Formation (Plate 21) and Pennington or upper Mauch Chunk groups represent the cratonward-prograding wedge of marginal-marine sediments (Ettensohn and Chesnut, 1985) (Fig. 16E). Based on the few places where the Mississippian-Pennsylvanian boundary is apparently gradational (Ettensohn and Chesnut, 1989; Ettensohn, 1994), this marginal-marine sedimentation probably continued into earliest Pennsylvanian time (Pocahontas Formation and equivalents), when it was abruptly ended by truncation along the sub-Absaroka or Mississippian-Pennsylvanian unconformity (Ettensohn and Chesnut, 1989) (Fig. 16F).

"Mississippian-Pennsylvanian" Unconformity

In eastern parts of the central Appalachian Basin, Mississippian and Pennsylvanian sections have been interpreted to be gradational, and the major erosive event reflected in the systemic unconformity apparently began later in Early Pennsylvanian time along a surface that first truncates Lower Pennsylvanian sediments (Englund, 1979; Englund and others, 1979) and cuts progressively deeper into Upper Mississippian rocks to the west and northwest. Although the systemic unconformity is not present in this exposure, on Indian Fort Mountain just northwest of Bighill, Lower Pennsylvanian conglomerates unconformably overlie lower Slade carbonates, and at least 215 feet (66 m) of Upper Mississippian rocks are missing on the unconformity (Weir and others, 1971). Hence, what appears to be a Mississippian-Pennsylvanian unconformity throughout most of its distribution is actually an Early Pennsylvanian unconformity that almost certainly reflects true inception of the Alleghanian Orogeny (Figs. 13, 16F).

Discussion and Conclusions

The Mississippian section at Bighill (Fig. 12) represents part of a largely Mississippian tectonic sequence defined at the base by Sunbury equivalents in the New Albany Shale and at the top by the Early Pennsylvanian unconformity present elsewhere in the area (Fig. 13). Based on the section and general knowledge of Appalachian Basin Mississippian stratigraphy, in this sequence Sunbury equivalents represent a transgressive systems tract at the base, whereas overlying parts of the Borden, Slade, and Paragon formations represent parts of a succeeding highstand systems tract, interrupted by smaller transgressive events (Ettensohn, 1998, 2004; Ettensohn and others, 2002). Parts of two third-order tectono-eustatic sequences and many smaller fourth-order glacio-eustatic cycles are also apparent (Fig. 12).

In another sense, however, flexural modeling suggests that the typical three-part clastic-carbonate-clastic Mississippian succession is mostly of tectonic origin related to the closing phase of the Acadian Orogeny, such that basal transgressive parts of the sequence reflect a final phase of active deformational loading, whereas overlying regressive parts represent succeeding phases of lithospheric relaxation (Figs. 13, 16). Although not visible at Bighill, Early Pennsylvanian truncation of the Mississippian section on the sub-Absaroka or "Mississippian-Pennsylvanian" systemic unconformity apparently marks inception of the Alleghanian Orogeny (Figs. 13, 16F).

Although the clastic-carbonate-clastic succession is typical for the Appalachian Basin, the middle carbonate section, so well developed at Bighill and elsewhere, is anomalous in a typical flexural sequence (Ettensohn, 1994, 2004). In fact, carbonate deposition like that observed in the Slade Formation apparently interrupted normal accumulation of deltaic clastics during an early phase of relaxation and effectively altered the course of Mississippian sedimentation throughout the Appalachian Basin. This "interruption" is most likely related to a unique coincidence of relaxational bulge uplift and a sea-level lowstand that resulted in cessation of major clastic influx and generated a rising, sediment-starved, upper-slope or outer-shelf setting (Nada, Fort Payne, Floyds Knob Bed) that resulted in very shallow-water conditions and exposure (Renfro, Maccrady) during the Osage-Meramec transition (see Warne, 1990) (Fig. 16C). When seas returned during St. Louis transgression, the necessary conditions for major carbonate deposition (absence of major clastic influx, very shallow waters, and presence in an arid subtropical belt) were in place, and major carbonate deposition expanded throughout the Appalachian Basin and adjacent regions for the next 20 Ma, even in the face of perturbations from the Ouachita Orogeny to the south.

By mid-Chesterian time, abrupt deepening and renewed clastic influx, apparent in the Maddox Branch Member and overlying Mississippian units at this exposure (Figs. 12–13, 16E–F), mark the end of major carbonate deposition throughout the basin and the beginning of Mississippian clastic deposition related to unloading-type (rebound) relaxation in eastern parts of the basin. This clastic deposition apparently continued into Early Pennsylvanian time, when uplift and erosion accompanying inception of the Alleghanian Orogeny truncated the underlying section on the "Mississippian-Pennsylvanian unconformity."

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Landscape Context of the Field Trip Area

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The field trip area is near the intersection of three major physiographic (landform) regions: the Bluegrass, Pennyroyal, and Cumberland Plateau (Fig. 17). Each region is separated from the others by significant escarpments developed on resistant bedrock units (Fig. 18). The Bluegrass Region in this area is characterized by gently rolling hills developed on Upper Ordovician to Devonian limestone, dolostone, and shale. The Pennyroyal is a dissected plateau developed on Mississippian limestones, and separated from the Bluegrass by the Knobs Escarpment. The Cumberland Plateau is a higher dissected plateau developed on Pennsylvanian sandstones and younger coal-bearing strata. The Cumberland Plateau is separated from the other two regions by the Cumberland Escarpment. The Cumberland and Knobs Escarpments merge between the Bluegrass and Cumberland Plateau Regions.

Bedrock geology is a major control on the development of landforms, on both local and regional scales. Locally, the field trip area is near the margin of the Livingston Conglomerate, a Lower Pennsylvanian channel cut into and removing the shales and limestones of the Upper Mississippian Paragon Formation. Where the Paragon is present, a broad bench is formed from the relatively easy erosion of the clay-rich shales (Fig. 19). Where the Livingston is present and the shales of the Paragon are absent, steep cliffs and hillsides are formed, lacking a prominent bench (Fig. 19).

The distribution of bedrock geology has led to distinctly different landscapes in these regions, which in turn has led to development of different ecosystems and land-use patterns in the regions. The ecological boundaries are not sharp and clear, however. Even where they are visually dramatic, landscape boundaries are generally diffuse in the ecological sense: plants and animals are notorious for ignoring lines on maps. Therefore, intersections of landscape regions tend to be areas of high ecological diversity, as the plants and animals characteristic of each region blend along the margins.

Dramatic physiographic boundaries, especially escarpments such as those developed in the field trip area, can serve as barriers to transportation and commerce. Early settlers found the few traversable passages through the escarpments, and developed the earliest roads along those routes. Many of these routes are still used as transportation corridors today. During the American Civil War (1861–1865), these routes became strategic military routes both for northern units moving troops and supplies southward, and for southern troops moving through the state on raids and invasions throughout the war.

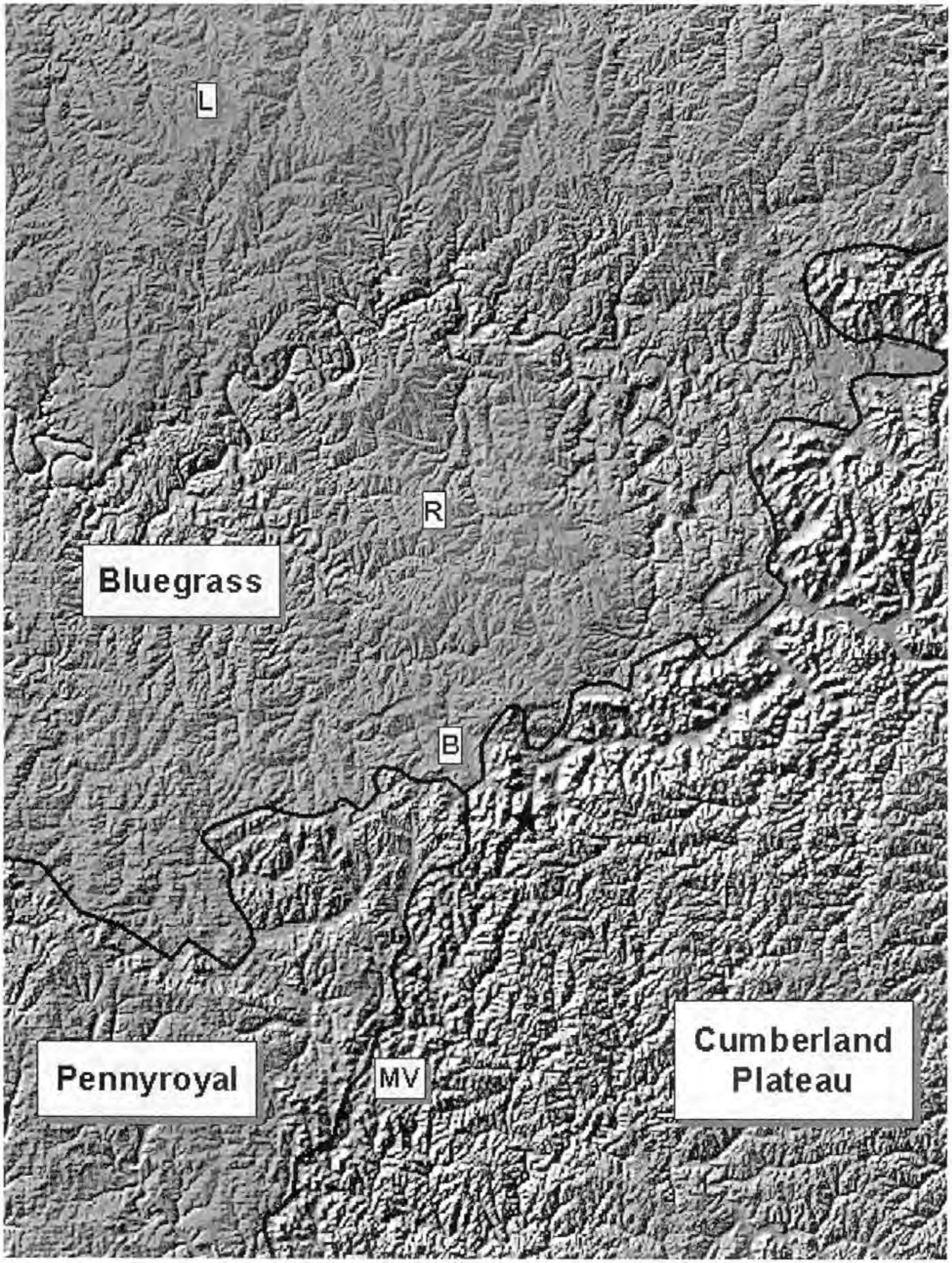

Figure 17. Physiographic regions of the field trip area.

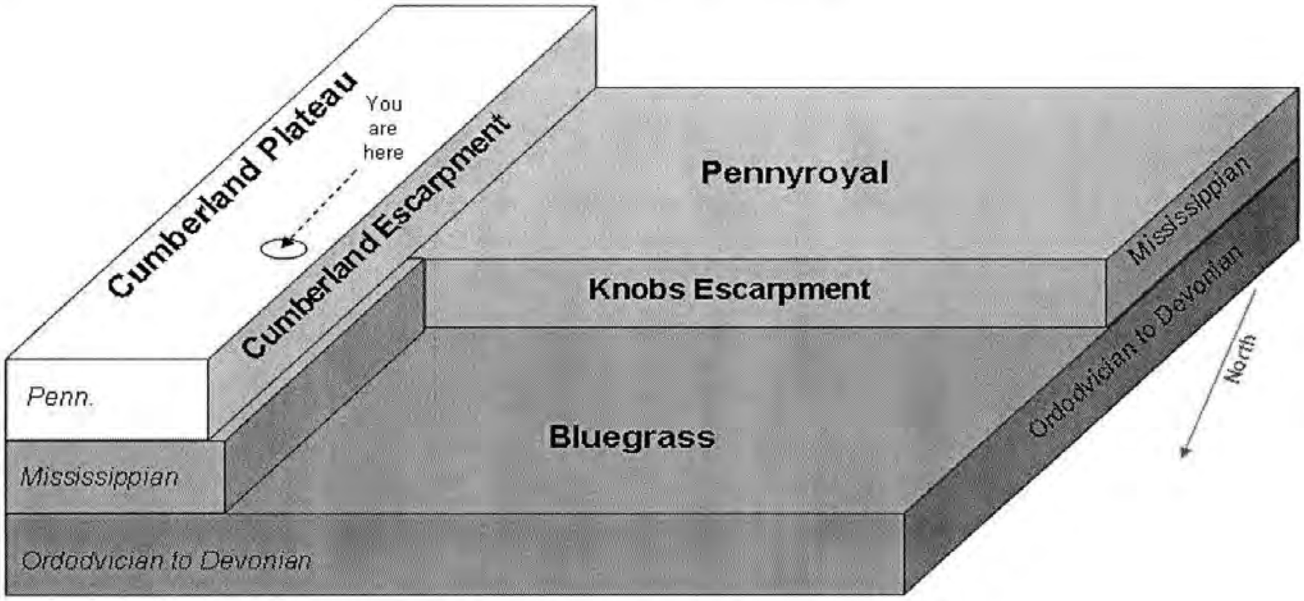

Figure 18. Escarpments in the field trip area.

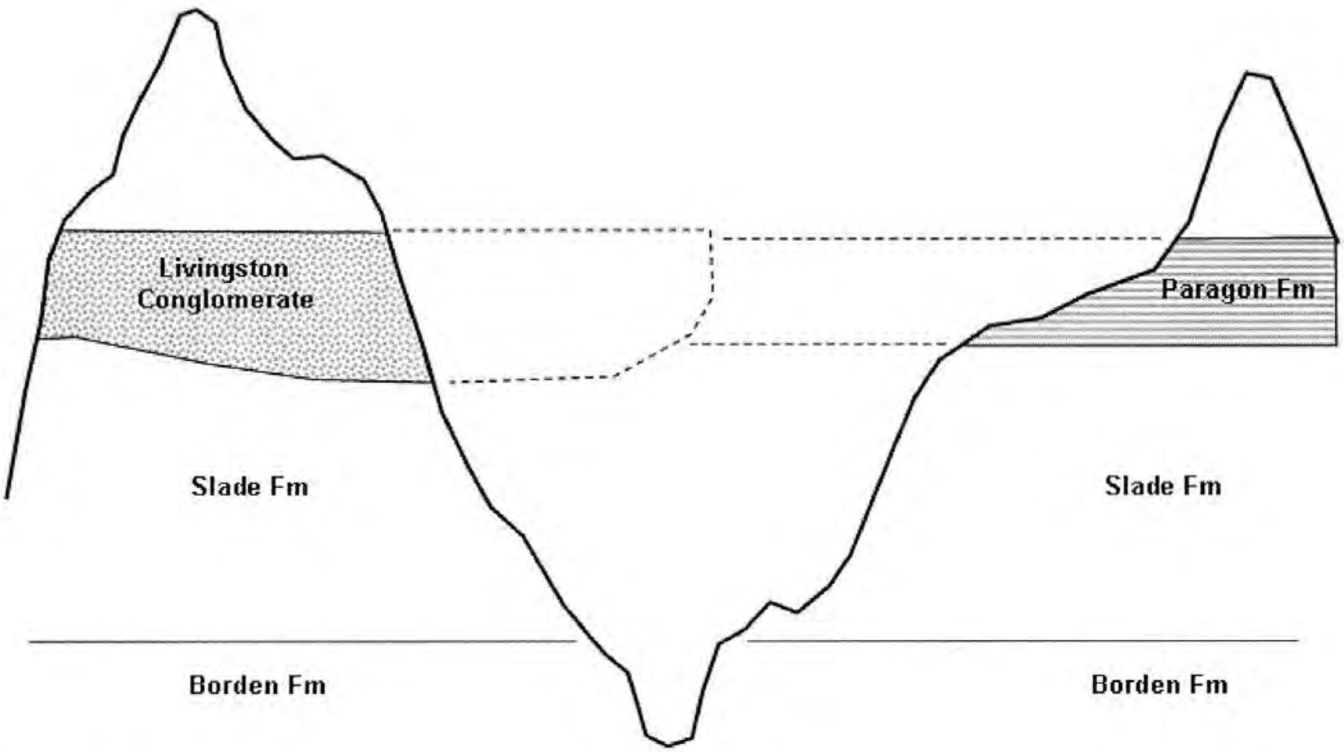

Figure 19. Where the Paragon is present, a broad bench is formed from the relatively easy erosion of the clay-rich shales.

Clover Bottom Quarry, Jackson County, Kentucky

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History

The earliest known existence of the Clover Bottom Mine was as an open-face quarry operated by the Johnson Stone Company; it was sampled for chemical analysis on January 24, 1951. In 1956, the quarry was taken over by the Roy Clark Stone Company, and was reported as a mine (whether it went underground before this date is not known). In 1958, the mine was purchased by the M.A. Walker Company and operated under this ownership till June 30, 2004, when it was bought by The Allen Company.

Figure 20. View from inside the mine.

Geology

Crushed rock is being produced from the Reelsville–Beech Creek Limestone of the Newman Limestone (Tygarts Creek Member of the Slade Formation). These Upper Mississippian limestones are light to medium olive-gray, mostly fine to coarse grained, and slightly to very oolitic. Some whole fossils are present, including brachiopods, crinoids, and horn corals, mostly skeletal. There is some dolomitic limestone, thin shale, and dolomite beds. Stylolites are common. The rocks in the mine dip to the southeast. Cores taken at the quarry show over 170 feet of rock material that can be mined productively for crushed stone.

Mining Acreage

Presently the quarry is located on 100 acres. Limestone mining occurs in 60 acres underground. The ceiling height is 28 feet and the quarrying goes to 30 feet below the level of the

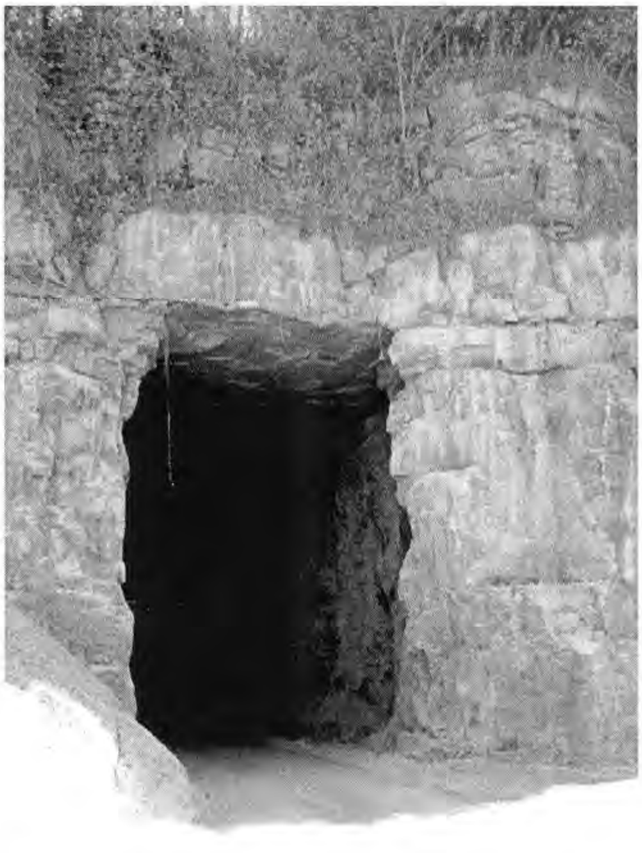

Figure 21. Looking into the mine.

Figure 22. North entrance of the mine.

Figure 23. Stylolites present in limestone.

surface of Ky. 421. The quarry has 25 full-time employees. Mining operations at the quarry have been very safe: there have been no recorded deaths or serious accidents caused by roof collapse.

Figure 24. North entrance of the mine below Ky. 421.

Crushed Stone

Generally, 400,000 to 500,000 tons of crushed stone are produced at the quarry each year. Twelve different sizes of stone are available:

Class 3	Size of boulders. Too big to be transported by a pickup truck. Used in erosion control.
Class 2	Channel line. Size about that of a hardhat. used in highway construction projects for erosion control.
Class 1	Softball size. Used in wire baskets for erosion control.
#2	Baseball size. Used in roadbeds and wet areas.
#4	Size of an egg. Used in roadbeds, wet areas, and septic fields.
#57	1-inch size. Used for concrete, roadbeds, and septic fields.
#9	½-inch size. Used in asphalt material for roadways.
#10	¼-inch size. Used in asphalt material for roadways.
#610	1½-inch size to dust. Used for U.S. Forest Service and county roads.
CBS	Crushed stone base, 1½-inch to dust. Contains less dust than #610.
DGA	Dense grade aggregate. 1 inch to dust.
Ag-lime	Agricultural lime. Lime powder used in fields and yards.

The Transportation Cabinet does testing on all crushed stone except for ag-lime, which is tested by the Department of Agriculture for size and quality.

Crushed stone produced at the Clover Bottom Quarry is sold to a variety of sources, including the Kentucky Department of Transportation, U.S. Forest Service, road and building contractors, local governments, concrete plants, asphalt plants, and utility contractors. The various size stones are used in buildings, drainage, erosion control, pipe bedding, road construction, concrete, and blacktopping. On the average, the stone costs about $9 to $10 per ton delivered in Madison County, except for channel line, which is a little more expensive ($10 to $12 per ton) because of the wear and tear on the trucks. The sizes that are in the greatest demand are DGA and #57, followed by #9, #10, #2, and #4.

Figure 25. Trucks being loaded at load-out bins with finished product.

Vehicles are weighed before and after being loaded with crushed stone. Tri-axle trucks are limited to a maximum weight capacity of 80,000 pounds. This weight limit is strictly enforced. Scales are certified by the Transportation Cabinet and are very sensitive. They are set to 20 pound increments, and could weigh to a maximum weight of 120,000 pounds.

Figure 26. Front entrance of quarry. Trucks being weighed on scales.

Quarry Operations

1. **Drilling.** This is a one-man operation. The face is generally 48 feet long by 28 feet high, and 16 feet deep. Holes to receive explosives are drilled into the face 4½ feet by 4½ feet and 16 feet deep using a 2½-inch drill bit. After the charge, about 14 feet is pulled out. The floor is generally 48 feet long by 50 feet wide and 16 feet deep. Holes to receive explosives are drilled into the face 9 feet by 10 feet and 16 feet deep using a 4½-inch drill bit. The drilling crew always works ahead.

Figure 27. Floor of mine, drilled 16 feet deep.

2. **Blasting.** Crews work under a certified blaster to prepare materials used for blasting, which include blasting caps, dynamite, and ammonia nitrate mixed with No. 2 diesel fuel (ANFO). Blasting occurs at the end of the work day after everyone is out of the mine, between 4:30 and 6:30 P.M. The blasting area is inspected by the certified blaster and an assistant the next morning. This allows all fumes resulting from the explosion to dissipate and any loose material to fall.

Figure 28. Explosive magazine storage.

Figure 29. Explosives being prepared.

3. **Scalping.** Scalping equipment removes all loose material from the ceiling, face, and the floor. The broken rock material is taken to the crusher and the area cleaned up so that the drilling crew can move back in.

4. **Crusher.** All material resulting from blasting passes through the crusher then travels on conveyor under a magnet to remove any metal. It continues into the sizing and secondary crushing plant.

Figure 30. Rock being crushed for the first time.

5. **Scalping Screen.** The rock material gets screened for DGA, #2, or #4, then drops into an impact crusher. It enters approximately 6 to 7 inches, and 2 to 3 inches in size, and exits 1½ inches to dust.

6. **Finishing Screens.** The crushed stone travels across a No. 2 and No. 3 screen for sizing and additional crushing by cone crushers as needed.

7. **Cone Crusher.** The rock material travels through two cone crushers for more crushing, then returns to the No. 2 and No. 3 finishing screens for further sizing. The crushed stone is now in a dead loop; the time to travel the loop is about 12 seconds.

Figure 31. Impact crusher.

8. **Sizing.** Sizing is determined by the size of the screens (wire cloth screen) and the opening of the cone crusher.
9. **Finished Product.** Various sizes of crushed stone are returned to stockpiles inside the mine.

Figure 32. Conveyor belt system.

Equipment and Buildings

Crushing Plant

Figure 33. Rock is crushed to various sizes.

Figure 34. Finishing screens.

Pug Mill

Figure 35. Dense grade or crushed stone base added to pug mill.

Figure 36. Water mixed with crushed stone to be used on roadway.

Special Thanks

Special thanks to the following persons at The Allen Company for providing information about their operation, and allowing participants attending the Kentucky Society of Professional Geologists 2004 fall field trip to visit the quarry at Clover Bottom and to go underground into the mine: Hugh Gabbard, president; Wilgus Fox, superintendent; Ivan Rose, mine foreman; Adam Holt, geologist; and David Reilly, quality control.

Plate 1. Siderite nodule (nucleic nodule) surrounding a coiled, nautiloid cephalopod. Collected from the Nancy Member of the Borden Formation, Bighill section, Madison County.

Plate 2. . Siderite nodule (septarian nodule) collected from the Nancy Member of the Borden Formation, Bighill section, Madison County. Mineral infilling the fracture is barite.

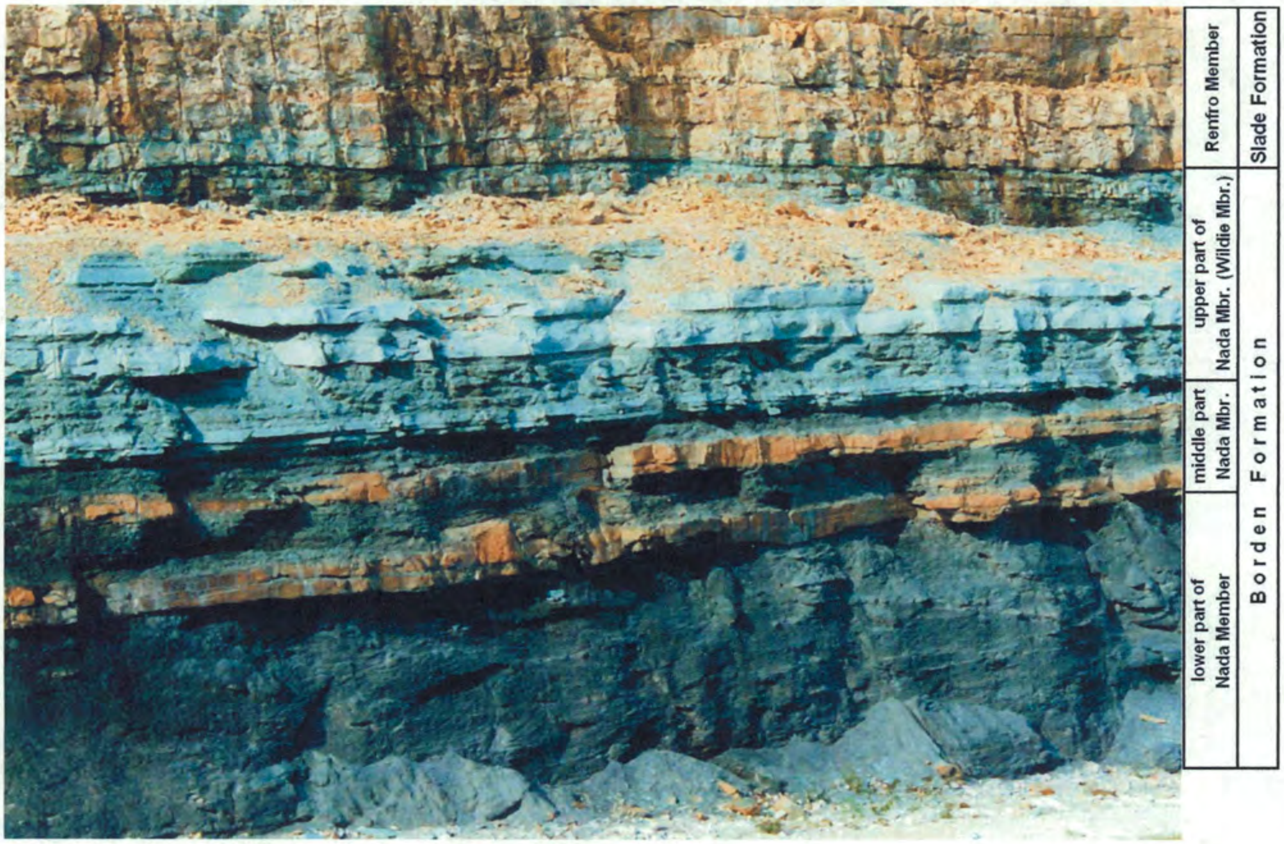

Plate 3. The Nada Member of the Borden Formation, Bighill section, Madison County.

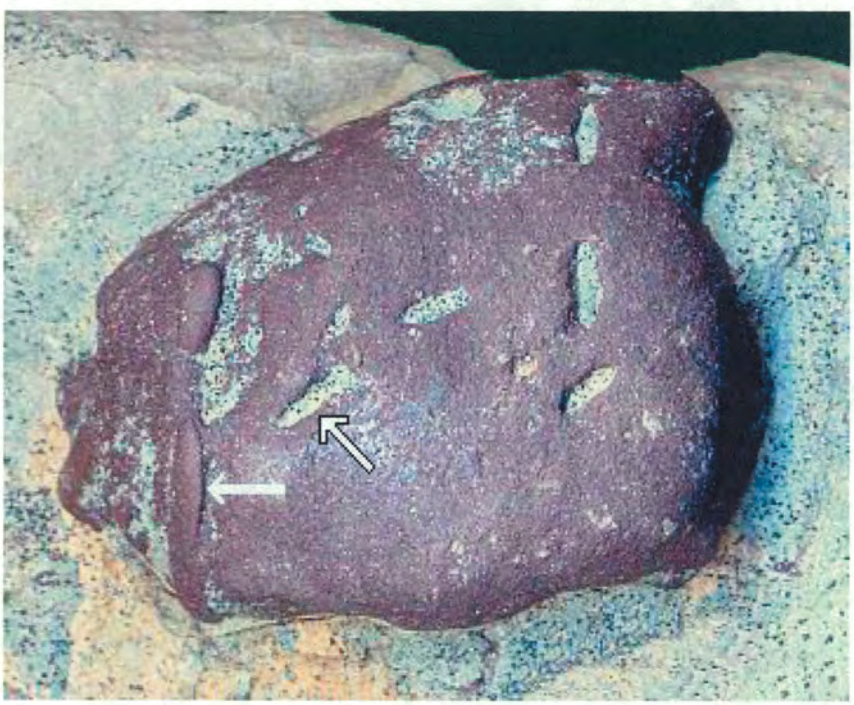

Plate 4. Bored phosphate nodule embedded in glauconitic matrix. Note boring on surface of nodule. Collected from along the top surface of the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

10 mm

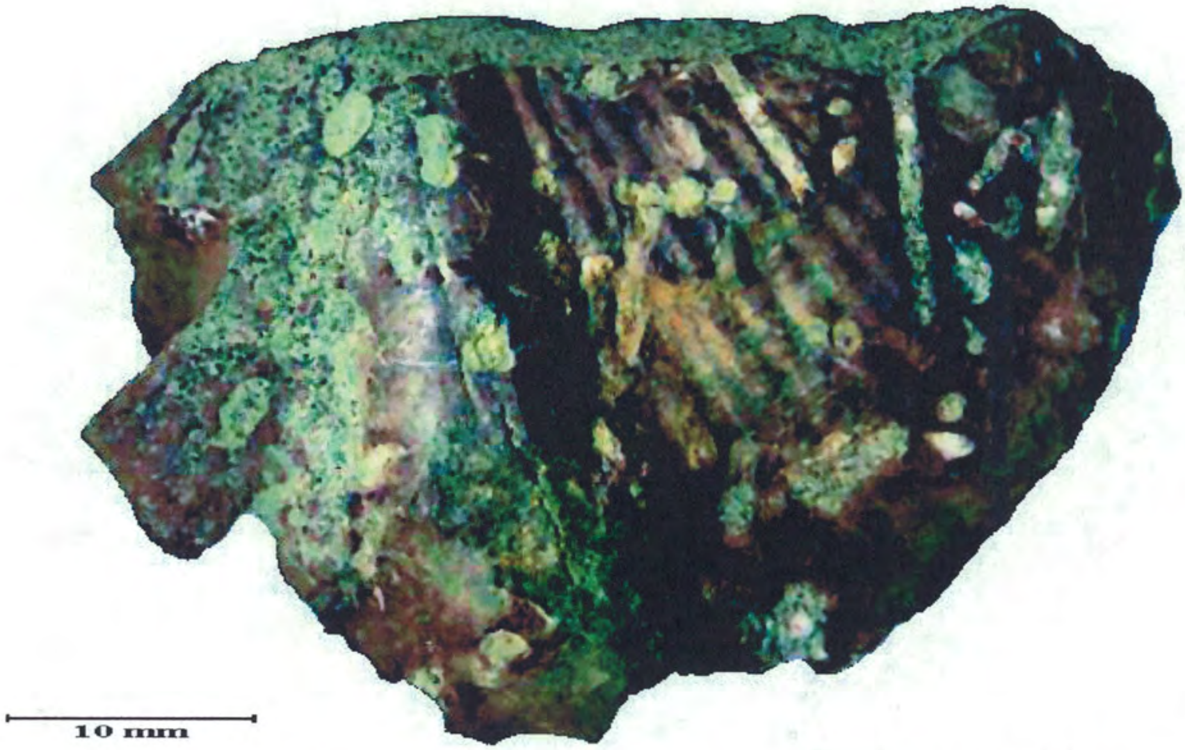

Plate 5. Phosphatized spiriferid brachiopod. Note borings on surface of brachiopod. Collected from along the top surface of the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

10 mm

Plate 6. Phosphated vertebrate remains. Note borings on surface bone fragment. Collected from along the top surface of the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

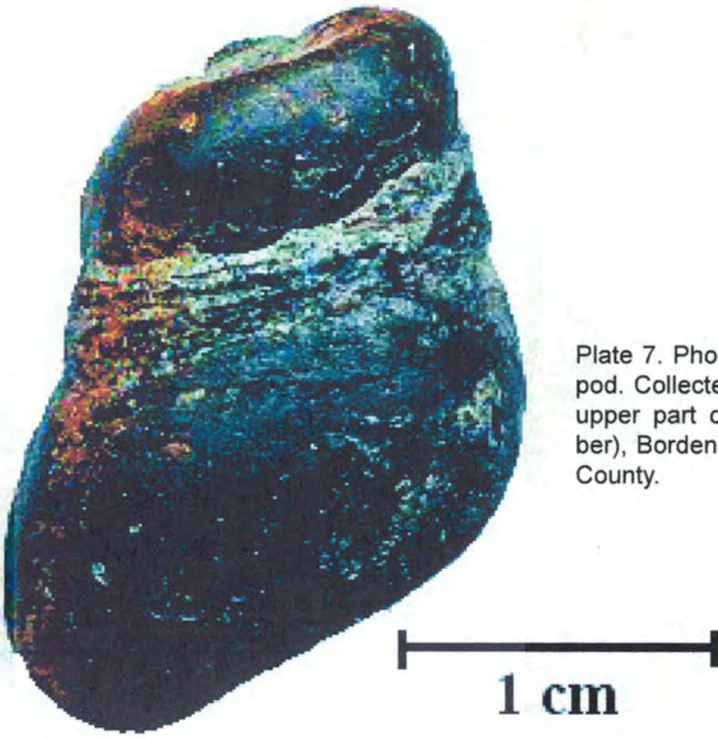

Plate 7. Phosphatized internal mold of a gastropod. Collected from along the top surface of the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

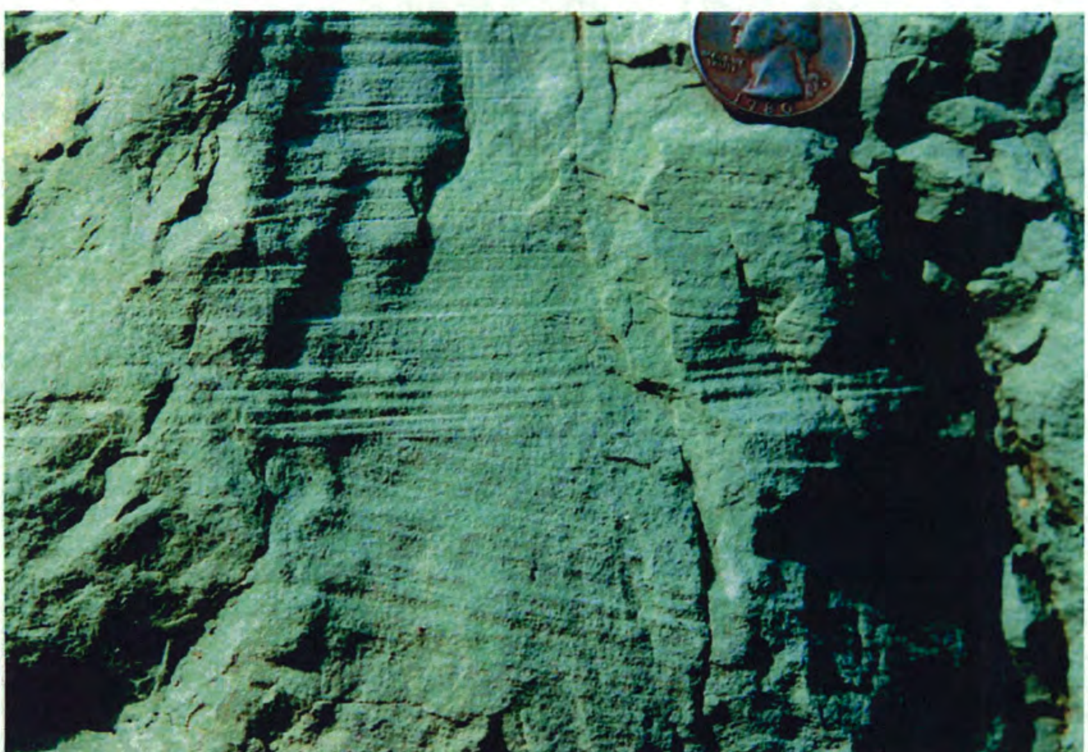

Plate 8. Crossbedding in siltstone bed in the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

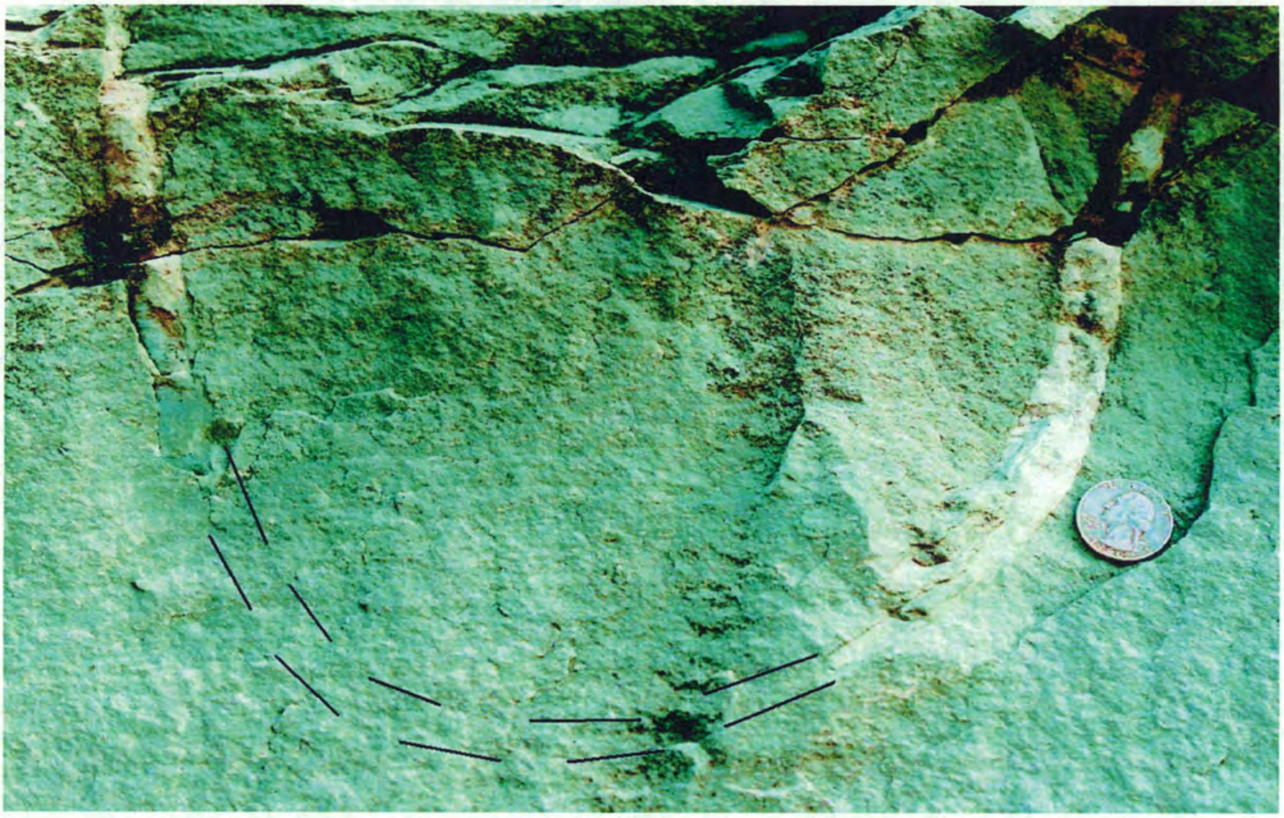

Plate 9. Cylindrical, U-shaped burrow *(Arenicolites?)* along siltstone bed in the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

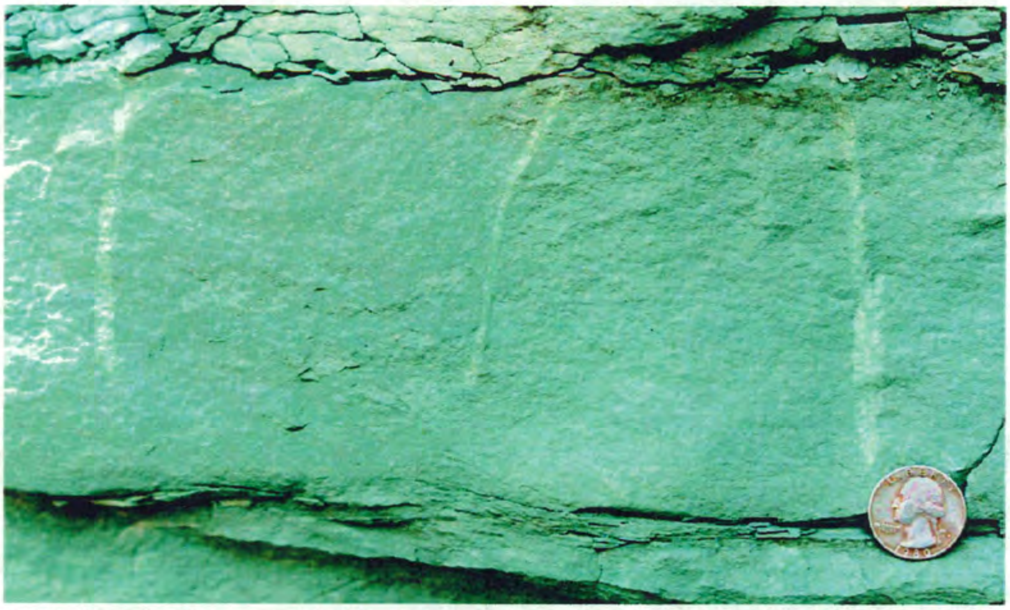

Plate 10. Cylindrical, vertical burrows *(Skolithos?)* along siltstone bed in the upper part of the Nada Member (Wildie Member), Borden Formation, Bighill section, Madison County.

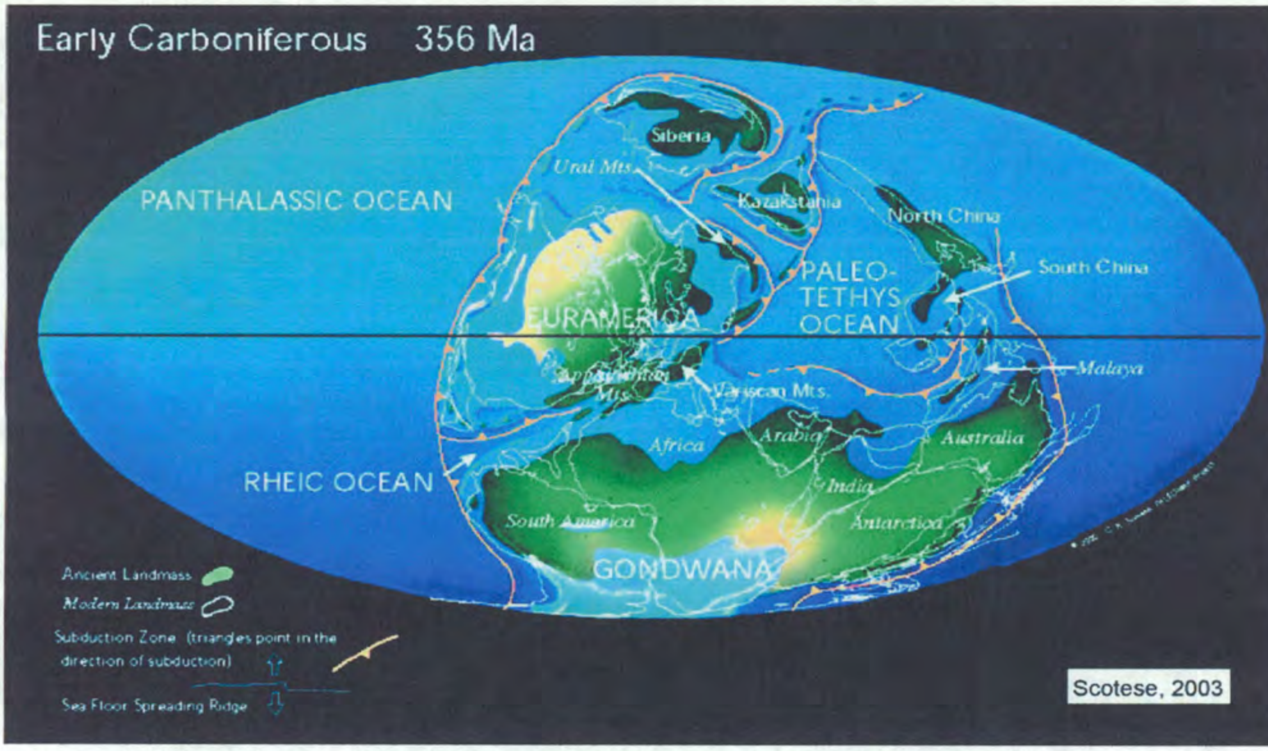

Plate 11. Early Carboniferous (Mississippian) world paleogeographic reconstruction showing the approximate position of the field trip area (white star) on Laurussia. Note incipient collisional zones and glaciation in the southern hemisphere (from Scotese, 2003).

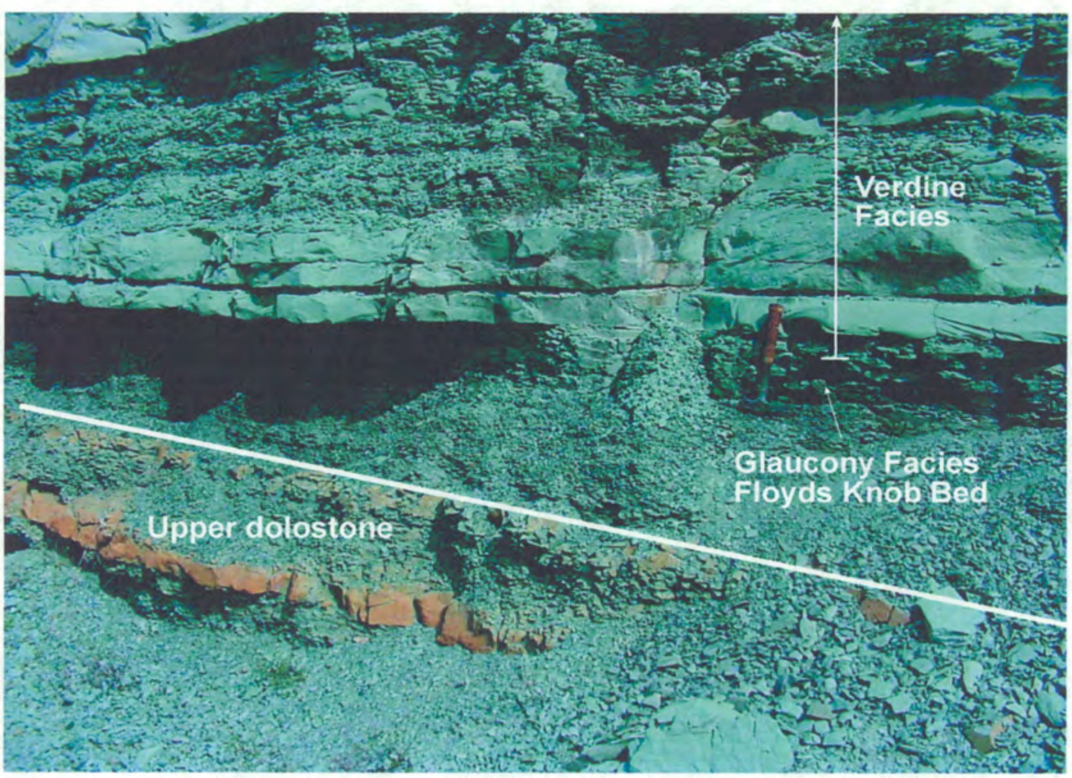

Plate 12. Upper, transitional part of the Nada Member, Borden Formation, showing the location of the glaucony facies (Floyds Knob Bed) and verdine facies, which reflect a shallowing, sediment-deficient storm shelf.

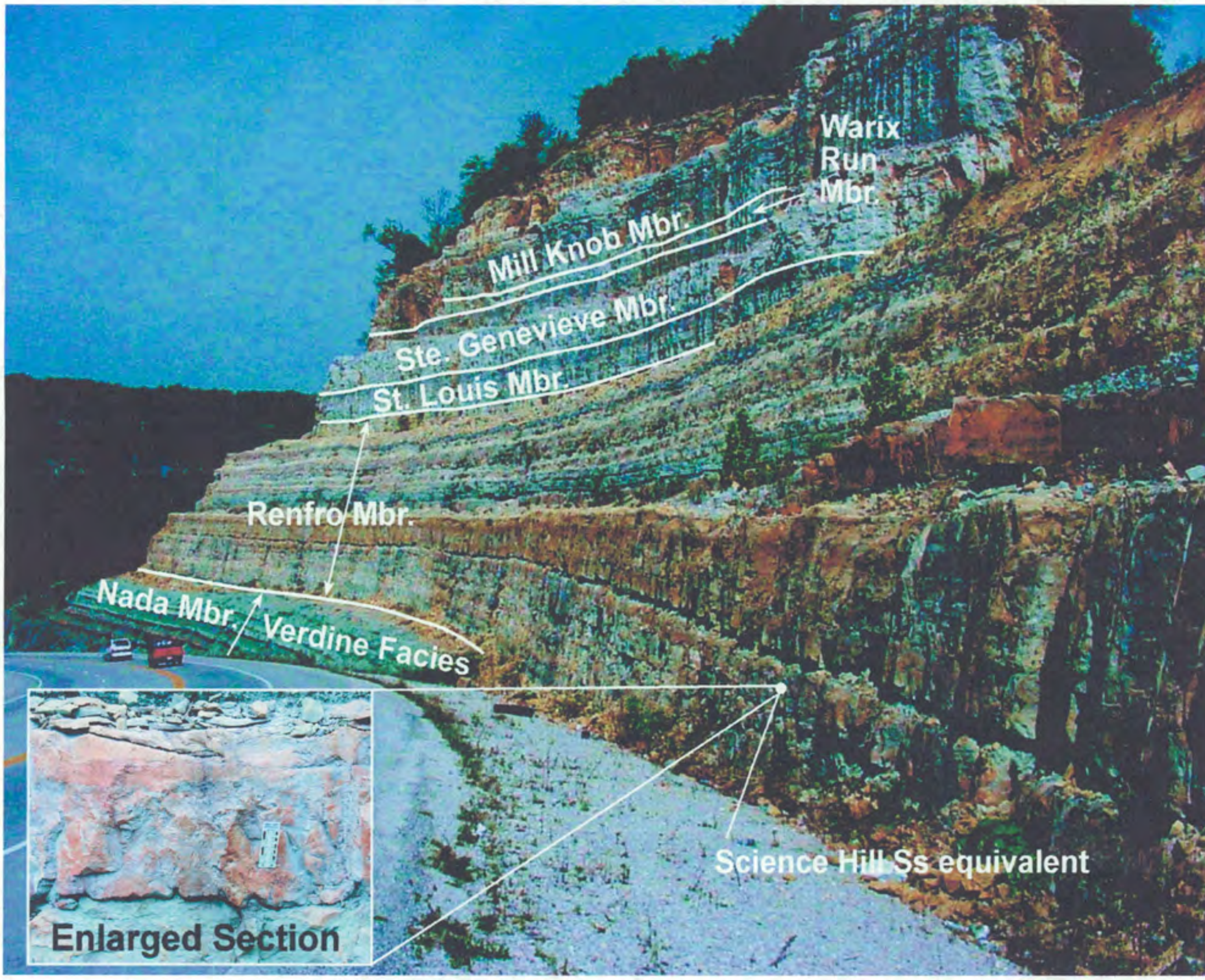

Plate 13. Southeastern cut at Bighill, showing transitional parts of the Nada Member, Borden Formation, and clear views of the Renfro through Mill Knob Members of the Slade Formation. The verdine facies in the Nada, as well as the Science Hill Sandstone equivalent and Ringgold Bed(?) in the Renfro, are also shown. Darkened areas at the top of the St. Louis and Ste. Genevieve Members reflect melanization accompanying Mississippian pedogenesis and subaerial exposure.

Plate 14. Coalescing silica (chalcedony and quartz) nodules, probably replacing nodular anhydrite, which displaced dolomitic and blue-green verdine-facies muds at the top of a bed in the Renfro Member. Verdine-facies muds were probably transported into shallow evaporative coastal lagoons and flats by storms.

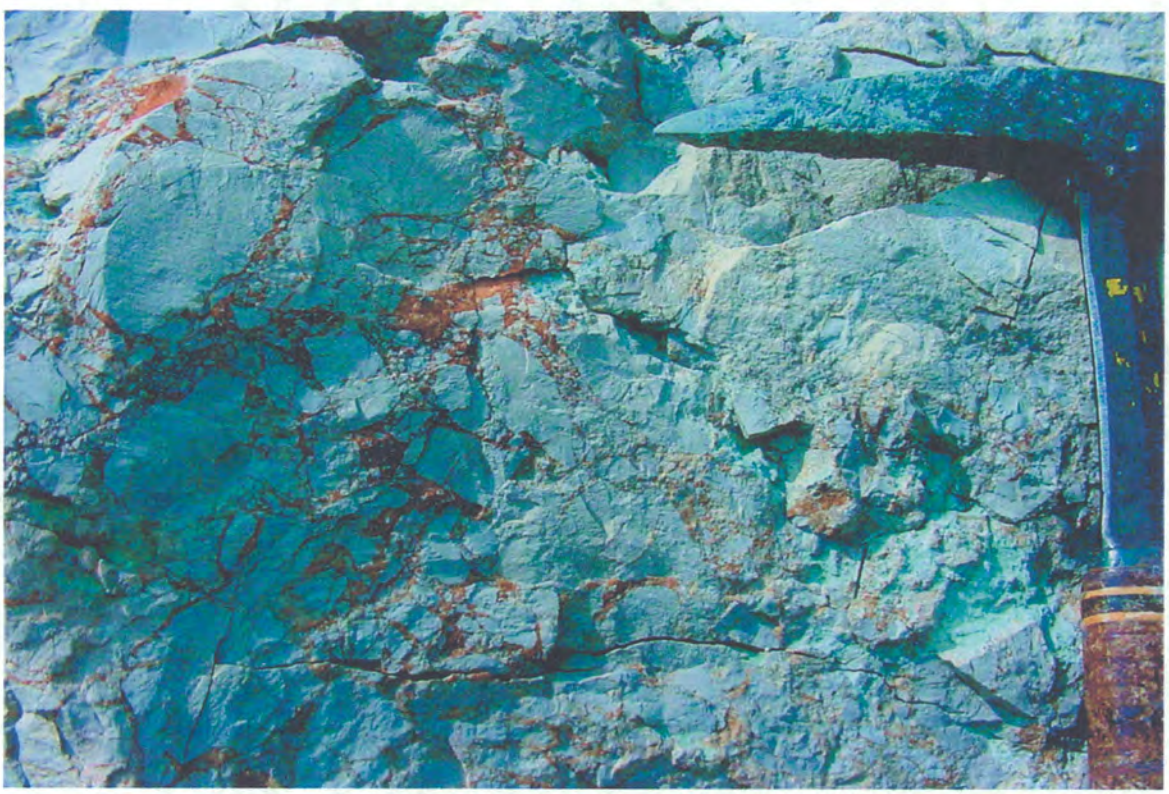

Plate 15. Possible solution breccias in the Renfro Member from the base of the erosional knoll shown in Plate 16.

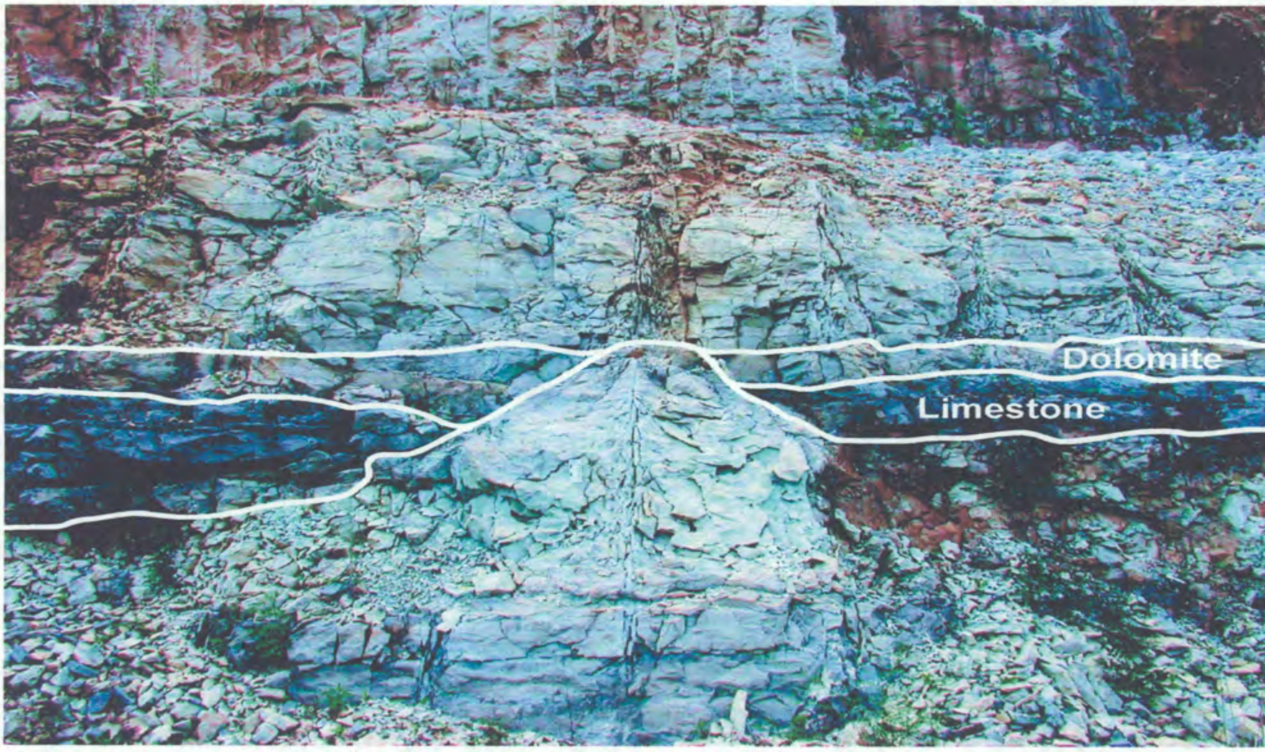

Plate 16. Erosional knoll in the Renfro Member with onlapping limestone and dolostone beds. The erosional hiatus is apparent becuase of the relief on the bed, but there are several other erosional breaks in the unit that are not so apparent.

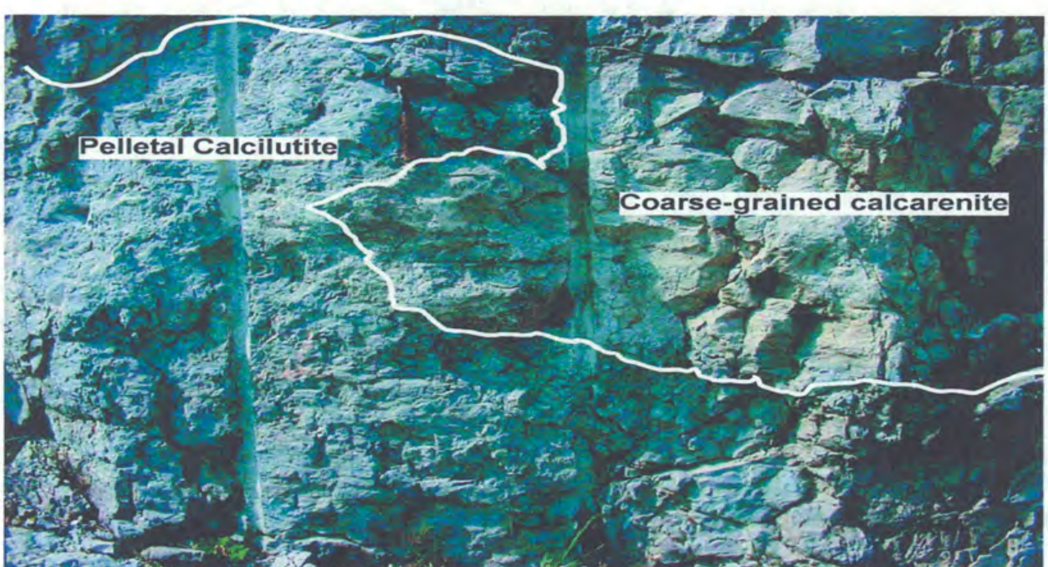

Plate 17. Intertonguing relationship (dark line) between dark brownish gray, clotted, birdseye calcilutites and coarse-grained skeletal-oolitic calcarenites in the Ste. Genevieve Member. The calcilutites probably represent a migrating mud-mound facies in a protected lagoon behind skeletal shoals and bars. The ridseyes represent spar-filled interstices between mud pellets and clasts. The contact between the two facies appears to be erosional in places and in part reflects reworking in the more energetic calcarenitic shoal facies.

Plate 18. Contacts between the Ste. Genevieve, Warix Run, and Mill Knob Members in the southwestern exposure at Bighill. The dipping beds in the Ste. Genevieve Member are crossbeds from a migrating tidal dune. The darkened color of the Ste. Genevieve is melanization related to pedogenesis and subaerial exposure at the top of the unit. Hence, the contact between the Ste. Genevieve and Warix Run is disconformable, whereas the contact between the Warix Run and nodular Mill Knob calcilutites is conformable and represents continued flooding begun during Warix Run deposition.

Plate 19. View of the southwestern exposure at Bighill showing the upper Ste. Genevieve (S.G.), Warix Run, Mill Knob, Tygarts Creek, and lower Ramey Creek Members. Darkened horizons in the middle and at the top of the Mill Knob reflect melanization at prominent paleosols. The contact of the Mill Knob and Tygarts Creek is disconformable. Stepped vertical lines in the Mill Knob and Tygarts Creek Members locate fourth-order, shoaling-up cycles.

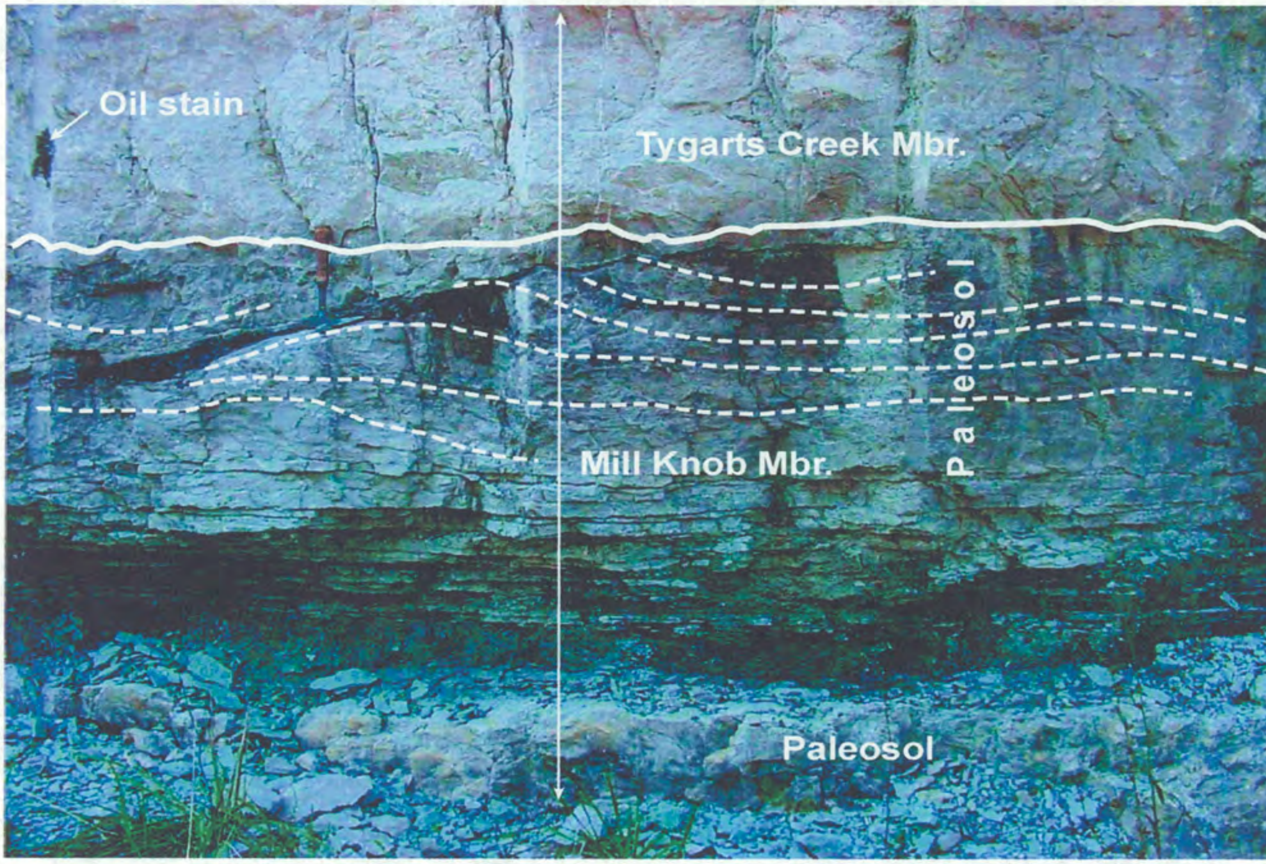

Plate 20. View of the southwestern exposure at Bighill showing the disconformable contact between the Mill Knob and Tygarts Creek Members. The lowest exposed parts of the Mill Knob are a paleosol, reflecting the top of a shoaling-upward cycle. The overlying shale to darkened calcilutite above represent the final Mill Knob shoaling-upward cycle (see Plate 19). The darkening in the upper part of the Mill Knob is melanization, and the dashed lines outline soil pseudoanticlines or teepees; both of these characteristics reflect subaerial exposure and pedogenesis. In places, dark brown chert also outlines pseudoanticlines. The overlying Tygarts Creek Member is a skeletal/oolitic calcarenite and the oil stains reflect areas of intergranular and oomouldic porosity.

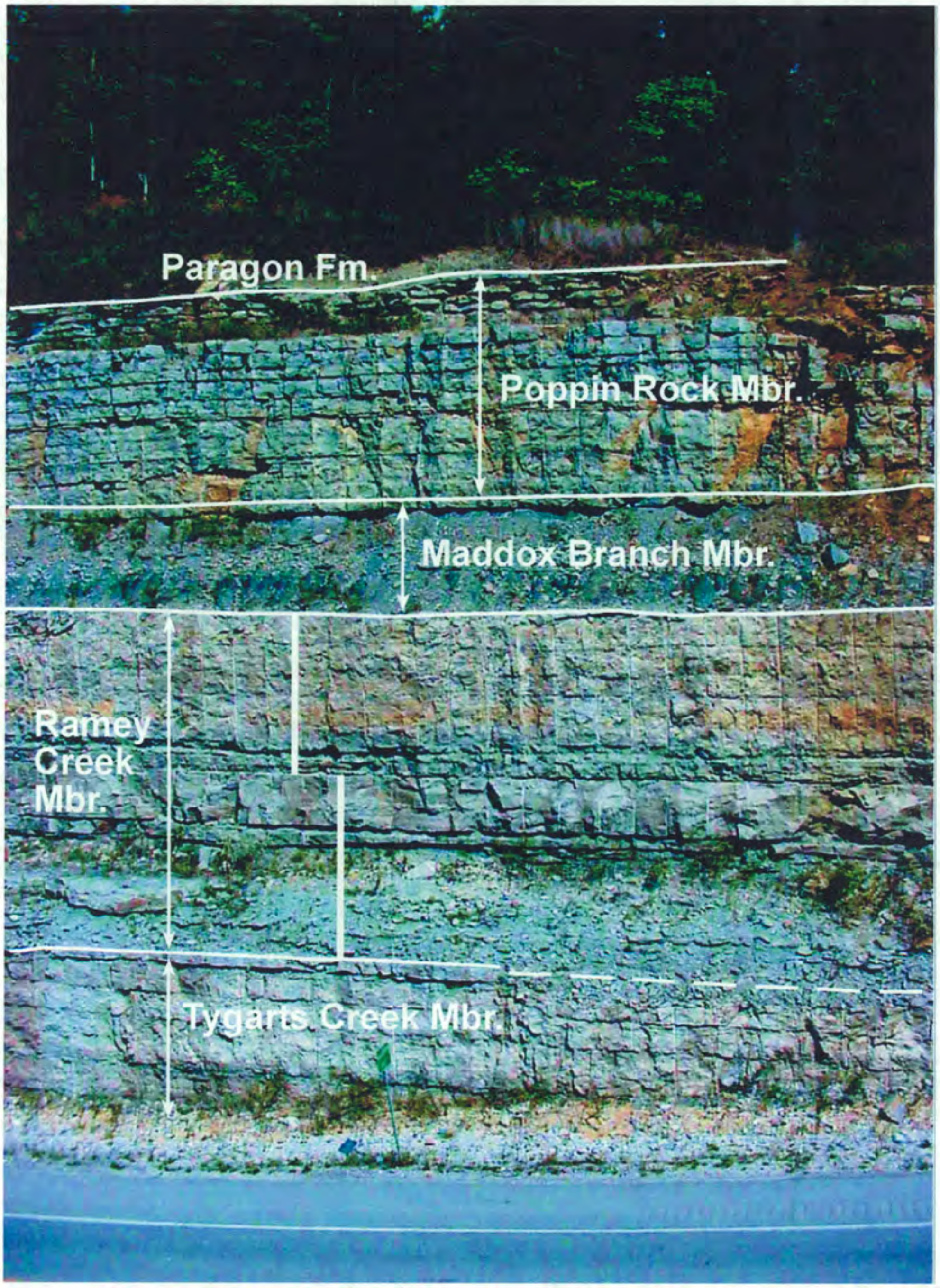

Plate 21. View of the southeastern exposure at Bighill showing the conformable sequence of members from the Tygarts Creek Member of the Slade Formation to the lower dark shale member of the Paragon Formation at the top of the exposure. Stepped vertical lines locate two shoaling-upward cycles in the Ramey Creek Member; basal shaly parts of each cycle are especially fossiliferous. The Maddox Branch Member represents the climax of a third-order deepening cycle in the Slade Formation and is sparsely fossiliferous. The upper part of the Maddox Branch Member is a 1- to 2-foot-thick dolostone that appears to be the base of the overlying Poppin Rock Member; it differs lithologically, however, from the Poppin Rock calcarenites. The Poppin Rock Member is gradational with basal dark shales in the Paragon Formation.