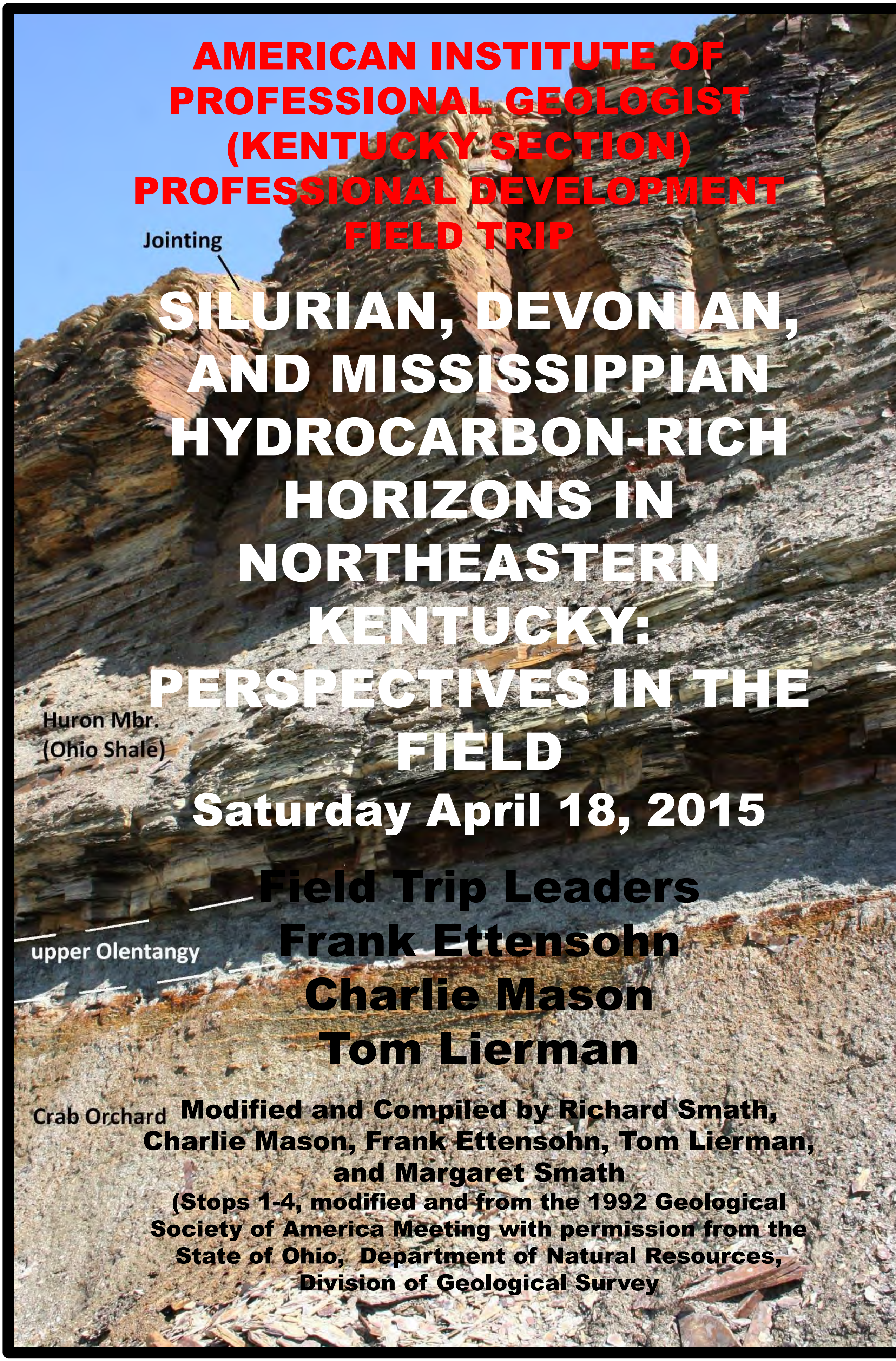

AMERICAN INSTITUTE OF PROFESSIONAL GEOLOGIST (KENTUCKY SECTION) PROFESSIONAL DEVELOPMENT FIELD TRIP

SILURIAN, DEVONIAN, AND MISSISSIPPIAN HYDROCARBON-RICH HORIZONS IN NORTHEASTERN KENTUCKY: PERSPECTIVES IN THE FIELD

Saturday April 18, 2015

Field Trip Leaders
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Modified and Compiled by Richard Smath, Charlie Mason, Frank Ettensohn, Tom Lierman, and Margaret Smath

(Stops 1-4, modified and from the 1992 Geological Society of America Meeting with permission from the State of Ohio, Department of Natural Resources, Division of Geological Survey

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Silurian, Devonian, and Mississippian Hydrocarbon-Rich Horizons in Northeastern Kentucky: Perspectives in the Field

Frank R. Ettensohn, Charles E. Mason, and R. Thomas Lierman

Introduction

Situated at the margin of the Bluegrass area in central Kentucky is a dissected escarpment supported everywhere by Devonian and Mississippian shales, siltstones, and sandstones and locally by Silurian carbonates. This escarpment has been called the "Highland Rim" or "Knobs" and it provides an excellent area in which to view outcrops of units that in the subsurface are major hydrocarbon producers. On the eastern side of the Bluegrass area, where our field trip will take place, the Highland Rim section contains local Silurian carbonates, Devonian and Mississippian black shales, called variously the Ohio, New Albany, Chattanooga, and/or Sunbury Shales, overlain conformably by the shales, siltstones, and sandstones of the Lower–Middle Mississippian Borden Formation.

Lower Silurian carbonates of the Bisher Formation crop out locally in the area and occur locally as subsurface topographic highs below the Devonian unconformity defined by overlying Upper Devonian black shales. The carbonates are typically porous. These Silurian carbonate highs form major reservoirs below the unconformity that have been sourced by the black shales that surround them on three sides. Although mapped as the Bisher Formation, only the lowermost carbonates are truly Bisher; overlying Silurian carbonates at stop 2 are most likely equivalents of the Lilley Formation in southern Ohio.

The Devonian–Mississippian black shales from this area, and from all of the eastern and central United States, form one of the most prominent and economically significant stratigraphic intervals across the United States. They are major hydrocarbon source and reservoir rocks throughout the area and are currently being examined as potential oil shales and trace-element sources. At present, some of the black-shale units in eastern Kentucky, like the Lower Huron Shale, are potential unconventional gas resources that can be accessed through horizontal drilling and fracking. In eastern Kentucky, the fractured equivalents of these shales in the subsurface form the largest gas-producing field in the state, the Big Sandy Field. Moreover, fine-grained sandstones in the conformably overlying Bedford-Berea sequence, a regressive sequence probably related to latest Devonian deglaciation in the Neoacadian Mountains to the east, are major reservoir rocks in some parts of eastern Kentucky.

The overlying Borden Formation delta sequence and its equivalents commonly form major reservoir rocks (Big Injun) in the subsurface. In addition, the fact that these rocks crop out along a major escarpment on which siltstones and sandstones generally overlie shales means that natural and human-made undercutting generates major mass-wasting-type engineering problems.

The time represented by these Silurian and Upper Devonian to Middle Mississippian rocks represents a period of major paleogeographic, paleoenvironmental, and paleontologic change across the North American parts of the ancient continent Laurussia. Shallow-water carbonates and clastics with their attendant faunas characterized much of Laurussia, including foreland-basin regions, until Late Devonian time. Even the change from Laurentia to Laurussia, with the addition of Baltica in Early Silurian time, came and went without many changes in the shallow-water, carbonate-dominated setting and biota on southern parts of the continent. With the imminent approach of Gondwana and progress of the Acadian/Neoacadian Orogenies during the Middle–Late Devonian transition, this all abruptly changed. After a period of uplift and erosion, large parts of southeastern Laurussia, including formerly stable cratonic regions, subsided and were inundated with basinal black shales followed by a major westward deltaic progradation that infilled basinal areas. Shallow-water, carbonate-dominated, benthic biotas in Early Silurian through Middle Devonian time were abruptly replaced by ichnofauna and a deeper-water, restrictive, commonly planktic, depauperate biota in the subsequent dark-shale basinal sediments. Although a sparse benthic macrofauna returned during upper parts of the deltaic progradation, not until complete basin infilling and the onset of delta destruction in Middle Mississippian time did an abundant, diverse, shallow-water, benthic fauna reappear in overlying units.

Hence, on this one-day trip into northeastern Kentucky, shown on the map in Figure 1, we will examine the physical stratigraphy and paleontology of the Lower Silurian–Middle Mississippian units, which represent stable cratonic, shallow-water environments, followed by major subsidence during the Acadian/Neoacadian Orogeny, and then deltaic basin infilling (Fig. 2). In a relatively new development, we will examine a large glacial dropstone embedded in the black shales and discuss its implications as evidence for Neoacadian tectonism and the influence of glaciation on black-shale sedimentation.

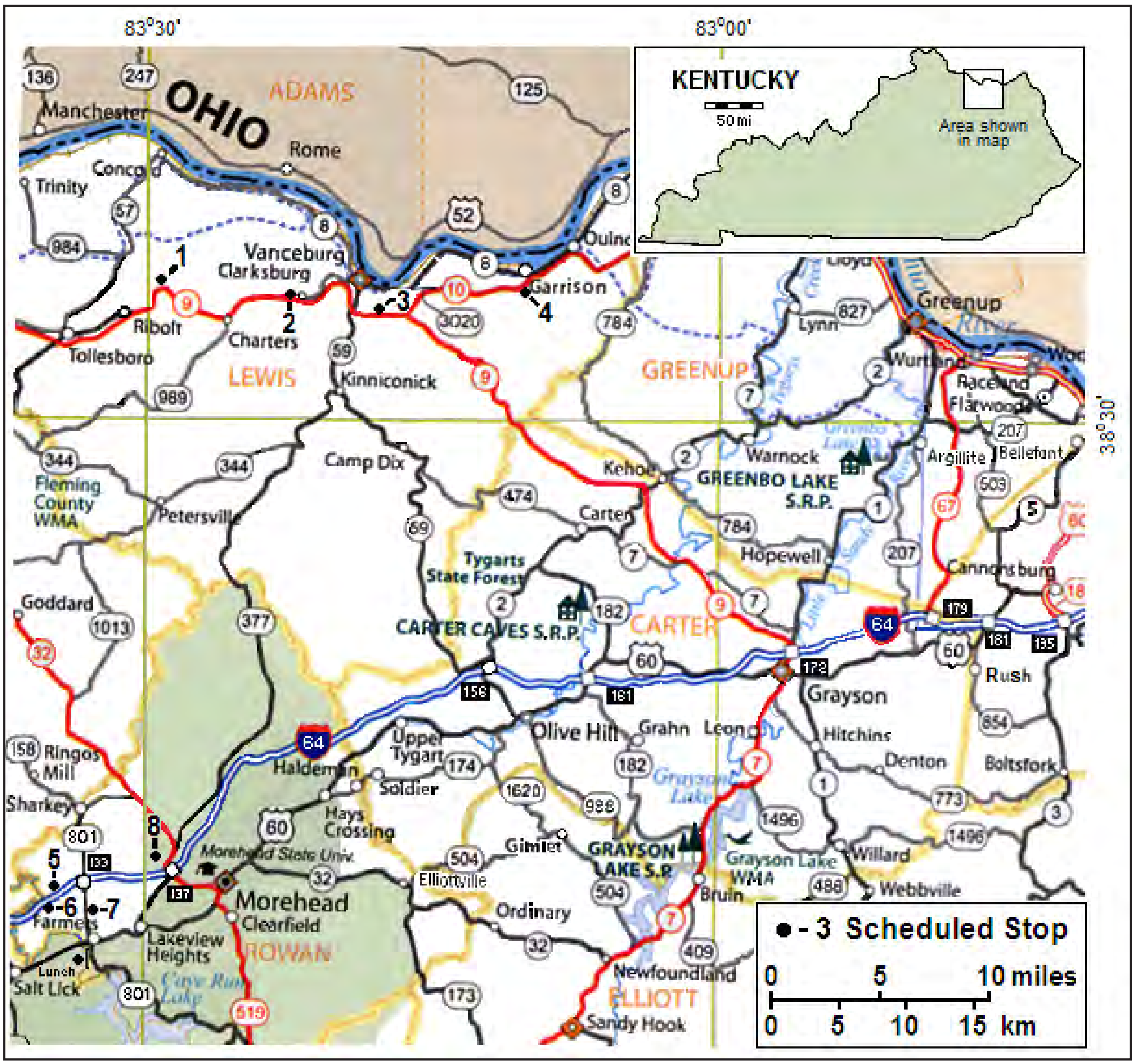


Figure 1. Field trip stops along and near I-64 in northeastern Kentucky. Modified from Ettensohn and others (2009).

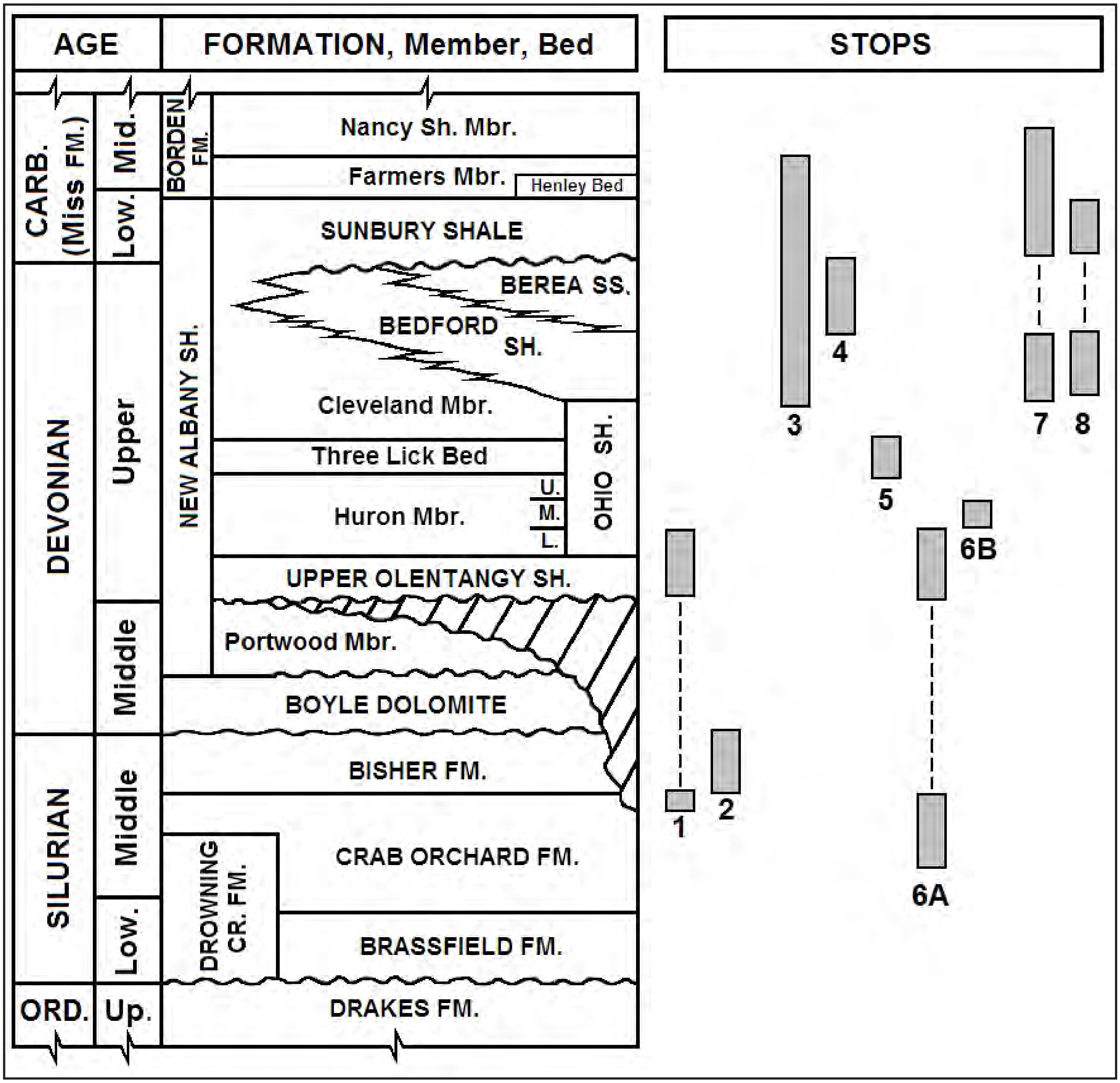


Figure 2. The geologic section along the field trip route with the relative stratigraphic positions of the sections viewed at each stop. Modified from Ettensohn and others (2009).

STOP 1: HERRON HILL
Sandbelt Lithofacies of the Bisher Dolostone, the Crab Orchard Shale, the Upper Olentangy Shale, and the Huron Member of the Ohio Shale in Northeastern Kentucky[1]

Charles E. Mason, R. Thomas Lierman, Frank R. Ettensohn, and Jack C. Pashin

At this exposure, the field trip route leaves the Knobs portion of the Pottsville Escarpment of the Kanawha Section of the Appalachian Plateau for the last time and enters the more gently rolling Outer Bluegrass proper (Figs. 3–4), underlain by Silurian and Upper Ordovician shales, limestones, and dolostones. The Middle Silurian Crab Orchard Shale and Bisher Dolostone and the Lower Silurian Brassfield Formation persist some 7 to 8 miles farther to the west in the Outer Bluegrass, whereas the Upper Olentangy and Ohio Shales are effectively lost to the west due to updip erosion on the flank of the Cincinnati Arch (Morris, 1965; Peck, 1967).

Crab Orchard Shale (Middle Silurian)

The upper part of the Crab Orchard Shale (Fig. 5) in this area varies from 130 to 160 ft (39.6–48.8 m) in thickness. At this location only the upper 29.5 ft (9 m) is exposed along the east end of this roadcut; it consists of predominantly mud shales interlayered with thin beds of dolostone. The shales exposed here are poorly fissile, light greenish gray to light gray, and weather to a light olive-gray or medium gray color. They are also heavily bioturbated, though individual traces are not recognizable. The dolostone beds appear to be dolomicrites, which are finely crystalline with a medium gray color when fresh but weather to dark orange or brown color. They are also heavily bioturbated, though individual traces are not recognizable. These beds of dolostone have sharp erosional bases, whereas the tops grade upward into overlying shales. Internally they show fine parallel laminae, and in a few cases wavy or hummocky crossbedding. The upper contact with the overlying Bisher Dolostone is quite sharp and its base is largely covered.

The shales of the Crab Orchard represent the accumulation of fine-grained terrigenous sediments along a slowly subsiding shelf. The shales apparently represent aerobic to dysaerobic, deeper open-marine accumulation of muds well below normal wave base. The thickness of the unit and presence of locally dysaerobic intervals suggest relatively rapid subsidence rates probably related to Salinic tectonism (Fig. 6). The thin dolostones in the unit, however, are probably distal tempestites (Fig. 5), reflecting the resuspension of carbonate sediments and terrigenous muds with the incursion of storm-surge and backflow currents.

A special note on the Crab Orchard Shale is that this formation has a strong tendency to slide or slump, causing pavement failures, especially where highway construction has oversteepened natural slopes. This is most evident along segments of abandoned Kentucky State Route 10 found on the left (south) side of State Routes 546 and 10 as one proceeds from stop 1 toward Vanceburg.

Bisher Dolostone (Middle Silurian)

The Bisher Dolostone in this area varies from 20 to 80 ft (6.1–20.4 m) in thickness, and at this stop the entire unit has a thickness of approximately 49 ft (15 m). The basal contact with the Crab Orchard is sharp and slightly undulatory, and basal parts of the Bisher contain reworked clasts of the Crab Orchard. Hence, an erosional hiatus or subtle disconformity may be present at the base of the Bisher in this area (Fig. 7).

The unit consists of fine- to coarse-grained dolostone with greenish gray shale partings. The dolostones are locally silty to sandy and are medium gray to greenish gray when fresh; weathering imparts a dark orange to brown color and a "punky" texture. Lithologically, the dolostones are best described as argillaceous dolomicrites to bioclastic dolomicrites and dolarenites, but these lithologies may not reflect the original textures. Where dolomitization has not been as effective, coarse-grained

[1]Stops 1 through 4 modified from Ettensohn (1992), the 1992 Geological Society of America field trip guidebook, with permission from the Ohio Division of Geological Survey.

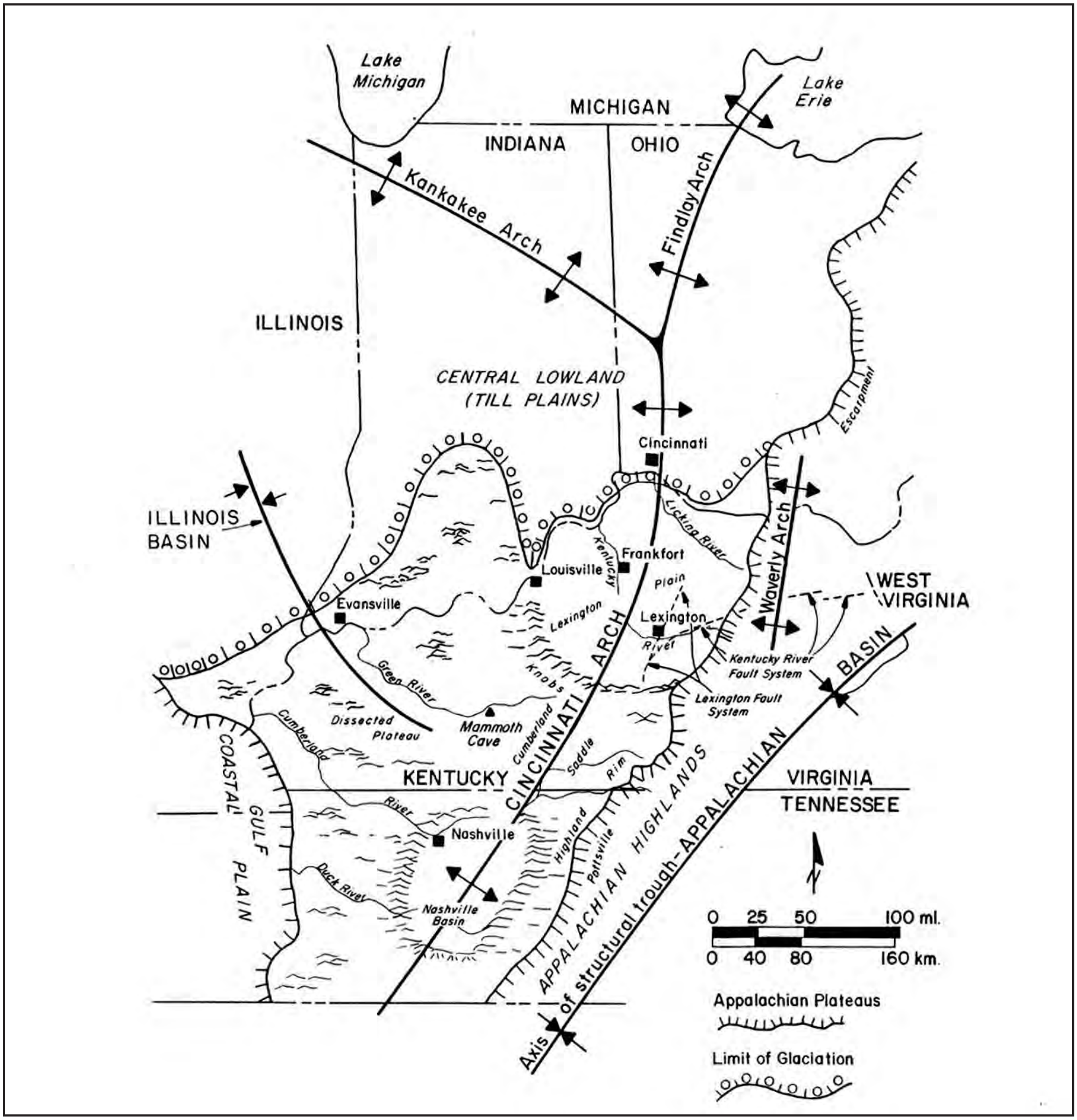


Figure 3. Interior Low Plateaus Province, in which most of the trip takes place, and parts of adjacent provinces, showing major geologic and geographic features.

bioclastic sands seem to predominate. Most of the dolostones possess a hypidiotopic fabric with fine- to medium-grained, subhedral ferroan dolomite crystals. Internally, stratification within the dolostone beds includes parallel laminae as well as local trough cross-laminae and hummocky cross-laminae. Less commonly found are scours, rip-up clasts, fossil lags, sparse ripple marks, and crude graded bedding.

An open-marine megafauna is common throughout and includes minor stromatoporoids, corals (both tabulate and rugose), bryozoans, bra-

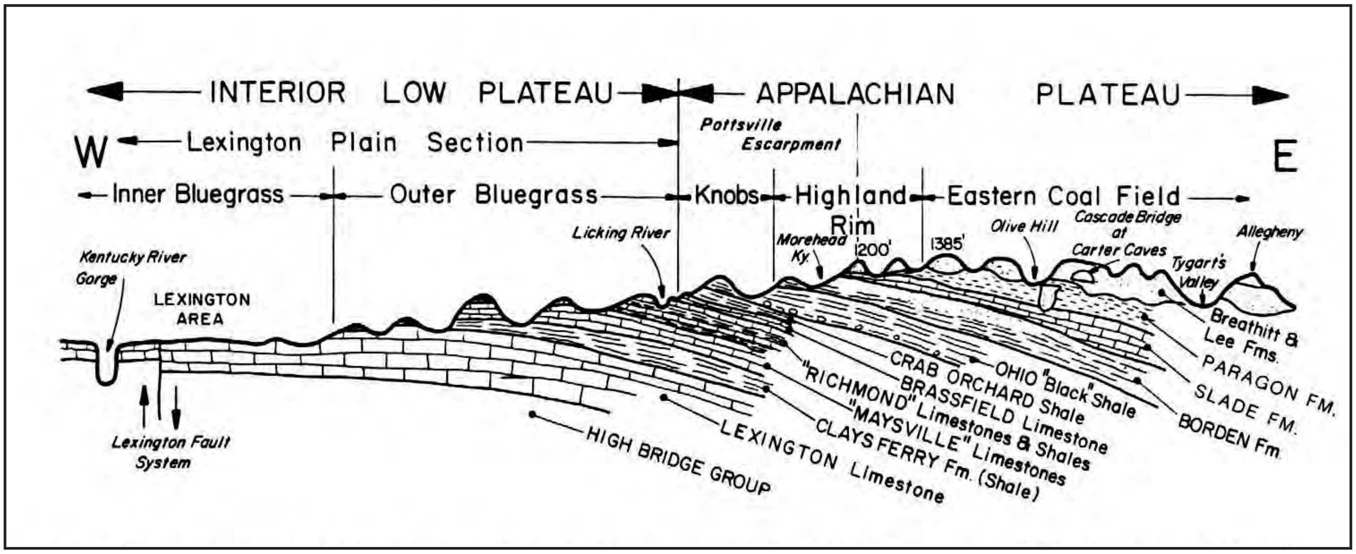


Figure 4. Schematic cross section through the east flank of the Cincinnati Arch from the Lexington area westward to the Grayson area in Kentucky. The section approximately follows Interstate 64 west of Lexington and shows the major rock units and the physiography developed on them. After Ettensohn (1981).

chiopods, gastropods, nautiloids, and disarticulated crinoid and trilobite debris. The megafossils in the Bisher are preserved as molds and casts that are predominantly broken and abraded fragments. Void spaces in the rock, especially those resulting from dissolved fossils, are commonly filled with blebs of petroleum or asphalt. The only microfossils observed at this cut are ostracods. Trace fossils are also poorly developed in this section, with only a few vertical and horizontal traces observed to date.

The apparent predominance of coarse-grained, bioclastic sediment, an open-marine fauna, and the preserved sedimentary structures suggest the likelihood of a relatively high-energy sandbelt or shoal-complex environment with tidal influence at or above wave base. In the western part of the exposure there appear to be two major shoal complexes stacked on top of each other. To the northeast, however, along the westbound lane, the lower complex is partially replaced by an intervening unit of interbedded shale and dolostone, which pinches out to the west. Close examination of individual dolostone beds in this unit suggests the presence of crude grading and local hummocky crossbedding, which may indicate the accumulation of proximal tempestites below, but close to, normal wave base. If this interpretation is correct, this shaly part of the Bisher may represent an intershoal depression on the sandbelt or a brief period of transgression and deepening.

The irregular upper surface of the Bisher (Fig. 5) is a compound unconformity below the Upper Olentangy Shale; it apparently reflects periods of Early, Middle, and Late Devonian erosion related to bulge migration and uplift on the flanks of the Cincinnati Arch. This contact represents about a 60-million-year gap in the stratigraphic record. Close examination of this surface reveals a number of features that suggest subaerial exposure and possible solution or collapse along the upper surface. Evidence for this can be found in the presence of the following features: (1) large, sediment-filled solution pipes or holes, which are commonly infilled with breccia fragments and/or a greenish gray shale like overlying Upper Olentangy shales, (2) irregular vugs or voids in the upper 3.3 ft (1 m) of the unit, (3) in situ breccia zones in the upper 3.3 ft (1 m) of this unit, and (4) an upper irregular surface with as much as 1.6 ft (0.5 m) of relief. Of course, some of these features could also be related to interstratal karst. However, many of these features are now very difficult to recognize, because they have been destroyed or partially obscured by the process of dolomitization, which occurred after these features formed. Some minor mineralization can also be seen in the upper meter or so of the Bisher, and this seems to be associated with the

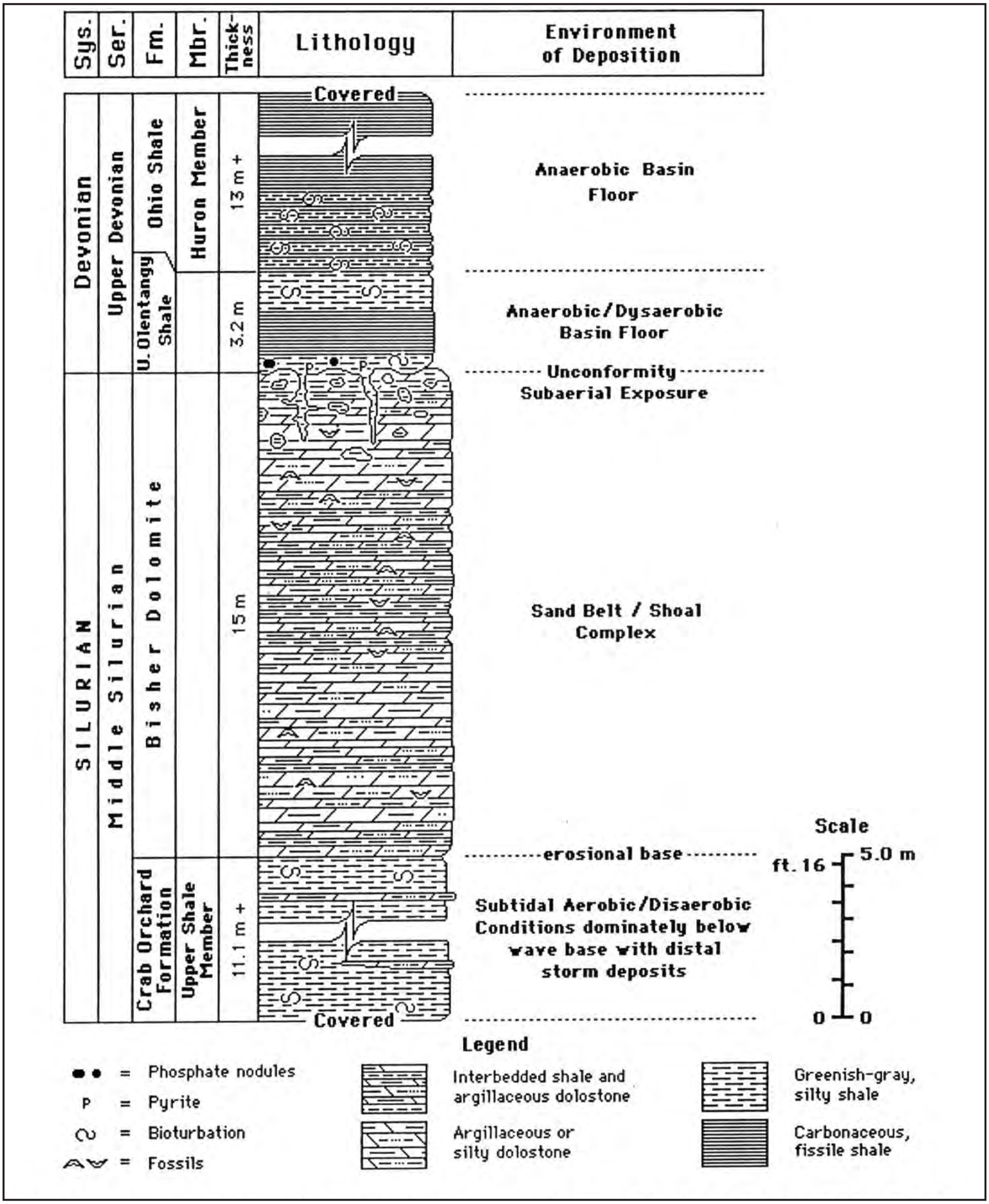


Figure 5. Schematic section showing the Middle Silurian and Upper Devonian rocks at Herron Hill on Kentucky State Route 546.

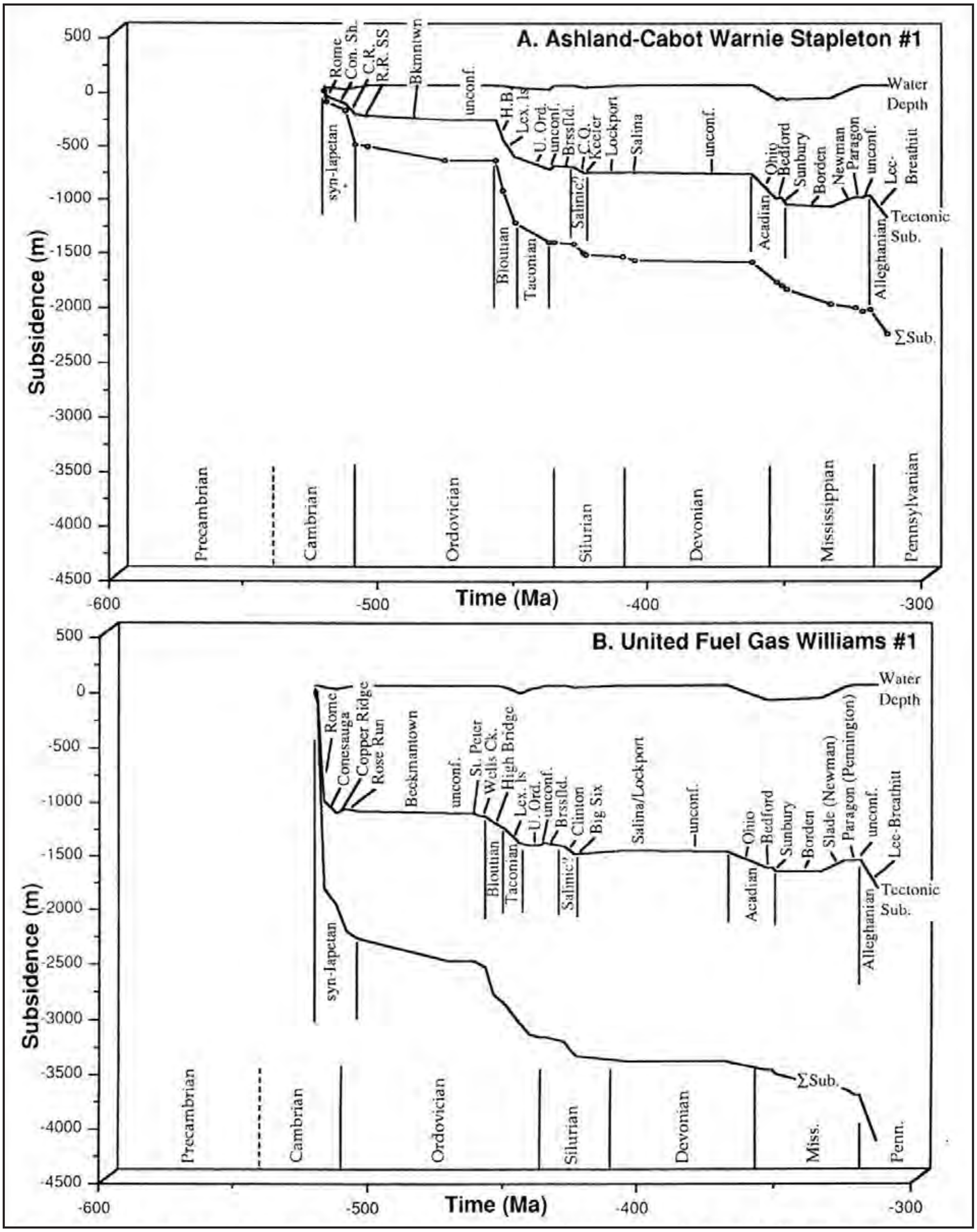


Figure 6. Tectonic and total (∑) subsidence curves for the Ashland-Cabot Warnie Stapleton #1 well (A), Carter Co., Ky., and the United Fuel Gas Williams #11 well (B), Breathitt Co., Ky. The isotopic and geological time scales are presented on the abscissa and subsidence is shown in meters on the ordinate. Stratigraphic units are labeled along the tectonic subsidence curve and the tectonic subsidence events are labeled by the name of their associated tectonic event. Subsidence is noted by a negative slope on the curves; uplift is shown by a positive slope. The slope of the line indicates the rate of subsidence; steep negative slopes equal high subsidence rates. Tectonic subsidence includes a correction for water depth.

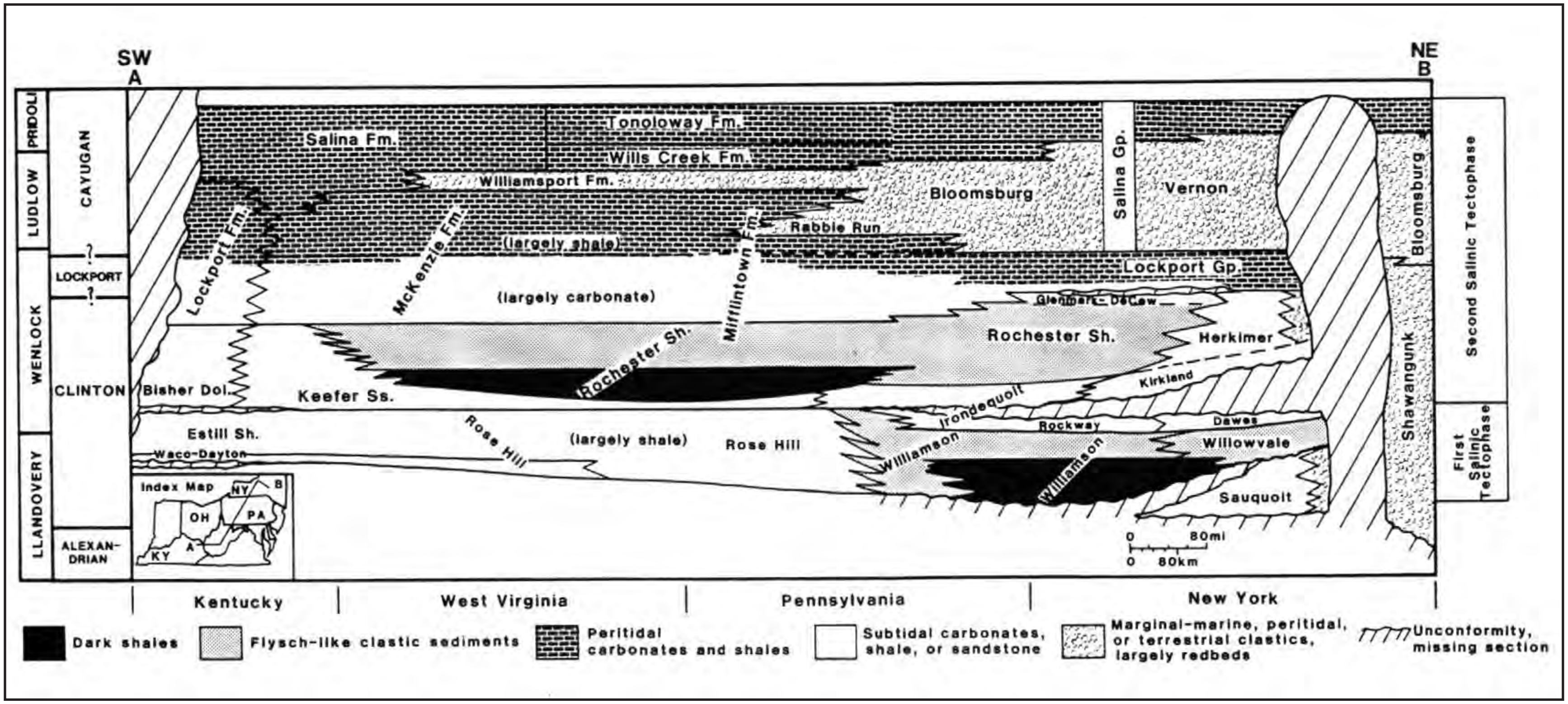


Figure 7. Schematic northeast-southwest cross section through northern and central parts of the Appalachian Basin showing the repetition of two probable flexural sequences associated with the Salinic disturbance, both in proximal and distal (northeastern Kentucky) parts of the basin. Only the second sequence is complete. The Estill Shale and Waco-Dayton Dolostone are parts of the Crab Orchard Formation in Kentucky. From Ettensohn (1994).

exposure surface. This mineralization can be recognized as the infilling of many of the vugs or void spaces (especially in fossils) with a variety of secondary minerals including barite, calcite, dolomite, pyrite, and sphalerite.

Upper Olentangy Shale (Upper Devonian)

The Upper Olentangy Shale in this area varies from 0 to 30 ft (0–9.2 m) in thickness. At this location about 11 ft (3.3 m) of the Upper Olentangy is exposed along the base of the first bench. The greater thickness here compared to that at stop 6 probably reflects the greater distance from the Cincinnati Arch in this area. The unit largely contains nonfissile mud shale with a grayish green to light gray unweathered color, but weathering to a light olive-gray. The Upper Olentangy also contains some minor interbeds of black fissile shale, and a 5-ft (1.5-m) interval of black shale is present in the middle of the unit but is usually covered with talus. It is possible that this black shale could represent the thinning western edge of the Pipe Creek Shale, a black shale within the Upper Olentangy that enters northeastern Kentucky in the subsurface (Kepferle and others, 1978). The gray shales in the Upper Olentangy contain scattered framboids and crystals of pyrite and marcasite, as well as concentrated deposits that form two thin mineralized beds low in the unit. The first and lowermost bed occurs at the base of the unit and the uppermost bed occurs approximately 2 ft (6 m) above the base. Small, sparsely fossiliferous phosphate nodules are also found throughout this unit.

Megafossils are rare and presently restricted to the Upper Olentangy exposed here. Eight ammonoid specimens and one trilobite have been recovered from this section to date. This is only the second occurrence (*see* Barron and Ettensohn, 1981) and the first surface report of Devonian ammonoids from Kentucky.

These fossils are poorly preserved as phosphate-replaced internal molds in small phosphate nodules. Attempts at examining this formation for microfossils were not completed in time to be included in this report. The unit is bioturbated, and both vertical and horizontal burrows are present. These burrows are replaced by pyrite, marcasite, and phosphate. Thus far no specific ichnogenera have been identified.

The Upper Olentangy Shale as observed here is thought to represent the extremely slow accumulation of predominantly hemipelagic muds in the deeper parts of the basin under alternating dysaerobic and anaerobic conditions. The latter point is evident from the interbedding of grayish green and black fissile shales. The slow rates of sedimentation to periods of nondeposition are evident from the presence of hydrogenous deposits (i.e., phosphate nodules) and mineralized zones.

Ohio Shale (Upper Devonian)

Huron Member. The Huron Member of the Ohio Shale in this area varies from 20 to 80 ft (6.1–20.4 m) in thickness. At this location, only the lower 39 ft (12 m) of the Huron is well exposed at the top of the roadcut. The Huron consists of brownish black to grayish black, fissile, carbonaceous, silty shale, which contains scattered nodules and crystals of pyrite and marcasite. When weathered it takes on a light gray to grayish brown color. Locally, the lower 9.8 ft (3.0 m) displays an interval of interbedded light gray shales and brownish black fissile shales. The lower contact with the Upper Olentangy is quite sharp.

Megafossils are very rare except for local occurrences of the inarticulate brachiopods *Lingula* and *Orbiculoidea.* Microfossils, especially conodonts and spores, are well known from this member, but they have not been examined at this section. Regionally, the Huron Member also contains a good trace-fossil fauna. However, other than a few nondescript vertical and horizontal burrows found in the lower part, no recognizable ichnogenera have been identified at this locality.

The Huron represents the extremely slow accumulation of hemipelagic muds in the deepest parts of the depositional basin, generally under anoxic conditions.

STOP 2: WEBSTER CEMETERY Biostromal Bisher Dolostone (Middle Silurian) Along Kentucky State Route 546, Northeastern Kentucky, Webster Cemetery Section

Frank R. Ettensohn, R. Thomas Lierman, Jack C. Pashin, and Charles E. Mason

The Bisher Dolostone in this area ranges from 20 to 80 ft (6.1 to 20.4 m) in thickness. At this location along the road only 21 ft (6.4 m) is exposed (Fig. 7). It consists of an argillaceous dolostone with only minor partings of shale. The dolostone is yellowish gray to light olive-gray in color and weathers to a dusty yellow to light brown. It can best be described as an argillaceous dolomicrite to fossiliferous dolomicrite that possesses a hypidiotopic fabric of fine- to medium-grained subhedral dolomite crystals. The dolomite crystals themselves are in part ferroan as determined by K-ferricyanide staining. Shales tend to be fissile, light greenish gray in color, and weather to a light olive-gray.

Megafossils can be found scattered throughout the unit or concentrated along distinct layers. Those recognized to date include overturned and broken stromatoporoids, fragments of tabulate corals, solitary rugose corals, trepostome bryozoans, brachiopods, gastropods, echinoderm debris, and minor trilobite fragments. Most of these specimens occur as molds, casts, or dolomite replacements. Many of the fossils, such as the stromatoporoids, tabulate corals, and fenestrate bryozoans, have interior voids or chambers that are infilled with blebs of dried oil or asphalt. No microfossils have been identified at this section.

The stratigraphic sequence exposed at this roadcut is quite interesting in that it can be subdivided into six or seven major fining-upward cycles, each of which is approximately 3 ft (1 m) thick (Fig. 8). Each cycle begins with a sharp erosional base that is flat to slightly undulatory in nature, and each exhibits up to a dozen very irregular individual layers that can be traced some distance across the exposure. The irregularity of the layers is clearly related to large protruding stromatoporoids and tabulate corals as well as accompanying differential compaction and bioturbation. The stromatoporoids and corals in each layer are typically fragmented and overturned randomly; small-scale cross laminae are present locally in some layers. Overall, successive layers in each major cycle appear to become finer grained, with fossiliferous rudstones at the bases grading upward into fossiliferous floatstones and finally into mudstones. Fossils occur in greatest abundance at the base of each cycle and decrease in abundance upward through most cycles.

Despite the apparent abundance of large fragmented fossil clasts, each cycle apparently contained a substantial amount of mud or muddy matrix, much of which has now been recrystallized to coarser dolomite. Detrital silt and clay is also locally abundant in upper parts of each cycle. Many of the fossil clasts, moreover, appear to be matrix supported, but these fossils seem to belong to a rather low-diversity, restricted community. Despite the variety of fossil fragments described earlier, many of which may be allochthonous, three colonial filter feeders, a stromatoporoid, a tabulate coral, and a trepostome bryozoan, predominate in the community, and of these, stromatoporoids were far more abundant: The relative abundance of muds and muddy matrix, the presence of detrital silt and mud, and the restricted nature of the fauna suggest that this Bisher lithofacies probably represents a nearshore lagoonal setting that was repeatedly disrupted by large tropical storms. Although each major cycle could represent a single major storm event, it seems much more likely that each of the irregular layers within a cycle represents an individual storm because of the relatively equal degree of fossil fragmentation and overturning in most layers. The apparent upward deepening in each cycle, however, may reflect (1) eustatic sea-level changes related to ongoing glaciation, (2) tectonic subsidence related to flexural movements, or (3) reactivation of basement structures in the area. For the given time and place, any of the explanations are equally likely, and hence it is difficult to suggest a preference.

Units 1 and 4 (Fig. 8) are sufficiently different from the others that they warrant special mention. Only the top of unit 1 is preserved, and although it could represent the top of a lower cycle, it is unique in that it is largely composed of dark, muddy sedi-

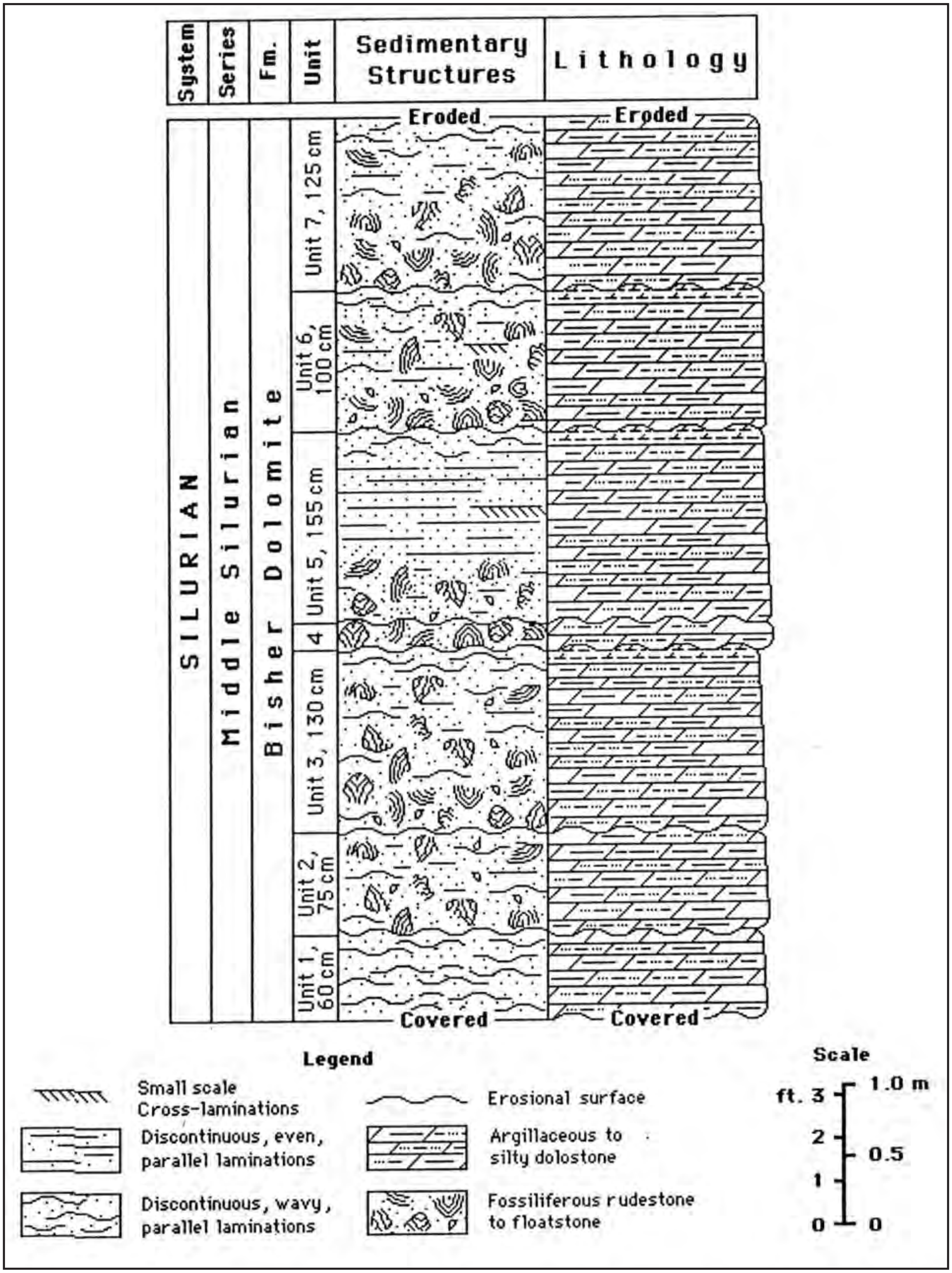


Figure 8. Schematic drawing of the Middle Silurian Bisher section and its sedimentary structures at Webster Cemetery on Kentucky State Route 546.

ment with discontinuous laminae and what may be bryozoan thickets. No other unit exhibits such bryozoan-rich muds, and this unit could reflect more typical pre-storm conditions. The advent of stromatoporoids in overlying units may simply reflect the prevalence of higher energy, more storm-prone conditions during subsequent periods of time. Unit 5 is also unusual in that it is relatively thin and exhibits several apparently in-place stromatoporoids. Some of these seem to exhibit in-place brecciation, which may suggest brief periods of exposure, which lends support to interpretations of a near-shore lagoonal origin for the lithofacies.

STOP 3: EVANS CHAPEL
Cowbell Member, Nancy Member, Farmers Member, and Henley Bed of the Borden Formation, Sunbury Shale, and the Bedford-Berea Sequence Along State Route 546 in Northeastern Kentucky

R. Thomas Lierman, Charles E. Mason,

Jack C. Pashin, and Frank R. Ettensohn

Cowbell Member

The Cowbell Member of the Borden Formation in this area ranges up to 340 ft (104 m) in thickness. At this location only the lowermost Cowbell, approximately 6.5 ft (2 m), is exposed at the top of the roadcut (Fig. 9). It consists predominantly of argillaceous siltstone with minor shale interbeds. The siltstone appears to be coarse-grained and medium gray to greenish gray in color, weathering to a light gray or buff color. Shales are light greenish gray, silty, and tend to weather to a light olive-gray color. For a more complete description of the Cowbell Member, see the previous stop.

Nancy Member

The Nancy Member of the Borden Formation in the Vanceburg area ranges in thickness from 51 to 180 ft (15.7–55 m). At this location the Nancy is approximately 51 ft (15.7 m) thick (Fig. 9). It consists of a medium gray, poorly fissile, silty shale, which weathers to an olive-gray or yellowish gray color. Also present are thin beds of siltstone and very fine-grained sandstone. These are light brownish gray, argillaceous, and appear to be thin turbidite beds that have sharp erosional bases and tops that grade into the overlying shales. Internally, the sand and siltstone beds show parallel laminae with some ripple cross-stratification. Both load casts and minor tool marks are commonly encountered on the soles of these beds. Also present within this member are siderite nodules, lenses, and beds. The contact with the overlying Cowbell Member is gradational.

This unit is highly bioturbated but most of the ichnogenera identified in the Nancy occur on the soles and tops of turbidite beds found within the unit. Chaplin (1982) placed this trace-fossil association within the *Nereites* ichnofacies. Ichnogenera collected from the Nancy Member to date include *Bifungites, Chondrites, Helminthoida, Helminthopsis, Scalarituba,* and *Zoophycos*. Within shaly portions of the Nancy, identification of individual traces is difficult, due to extreme bioturbation and compaction of the shale. Scattered body fossils tend to occur as molds, casts, or mineralized replacements within siderite nodules. Megafossils found within the Nancy at Evans Chapel include brachiopods, bryozoans, conularids, pelecypods, gastropods, cephalopods, and crinoid debris. The only microfossils that have been observed are ostracods preserved as molds in siderite nodules.

The Nancy Member is thought to represent the deposition of fine-grained sediments and occasional distal turbidites in a prodelta setting. Deposition occurred under dysaerobic conditions in the lower part of the Nancy but ranged into aerobic conditions in the middle and upper parts.

Farmers Member

The Farmers Member of the Borden Formation in the Vanceburg area ranges in thickness from 118 to 262 ft (36–80 m). At this location the Farmers Member, excluding the interval of the Henley Bed, is approximately 118.4 ft (36.1 m) thick (Fig. 9). It consists predominantly of fine-grained sandstone and siltstone beds separated by thin beds of shale. The sandstone beds are a light brownish gray and very fine to fine grained. They are evenly bedded and are thin to thick bedded. Siltstone beds are medium to light gray in color and consist of

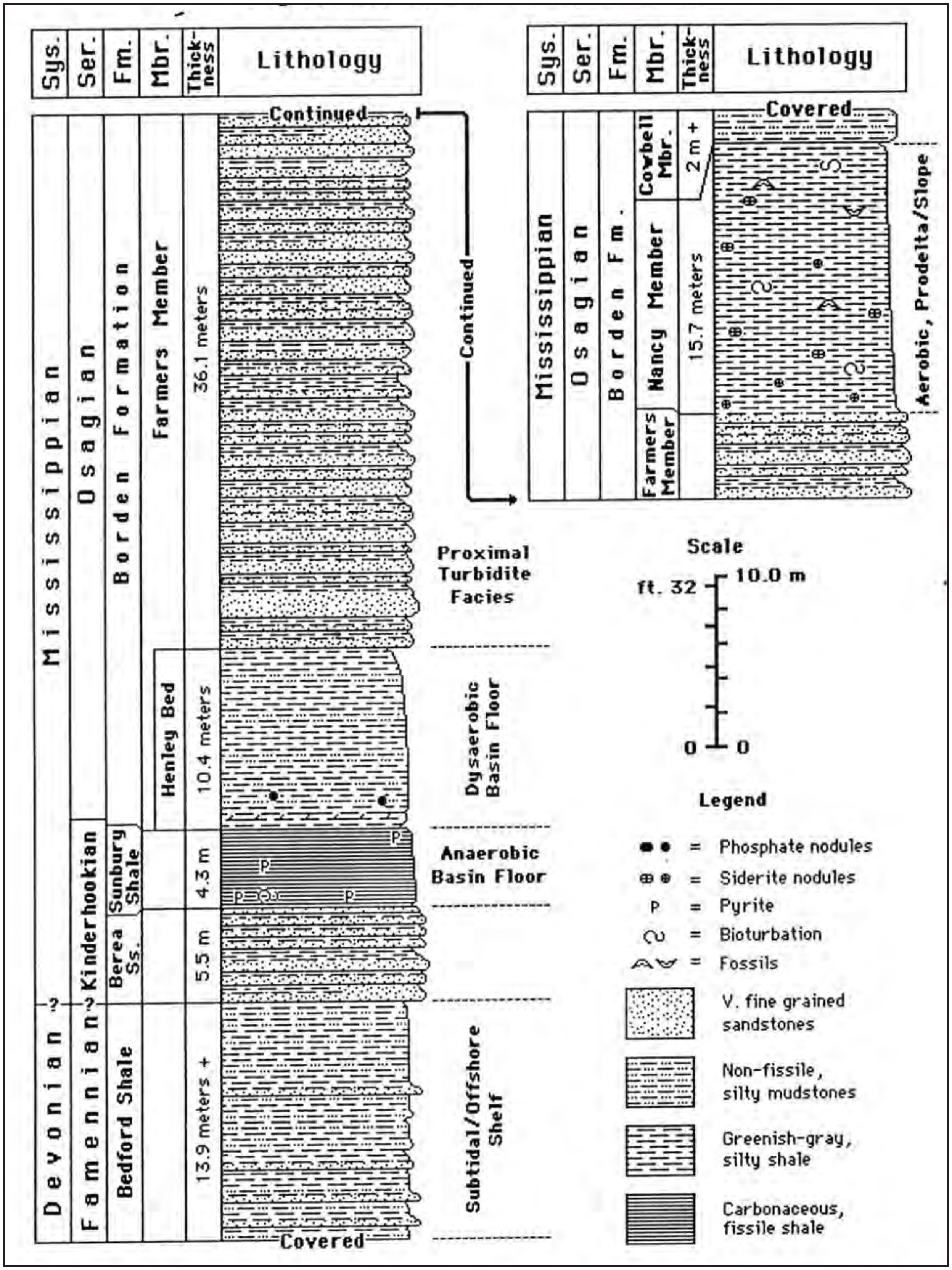


Figure 9. Uppermost Devonian–Lower Mississippian section and environmental interpretations at Evans Chapel on Kentucky State Route 546.

moderately sorted, coarse-grained silts. Individual sandstone and siltstone beds generally have sharp erosional bases and grade upward into overlying shale interbeds. These shales are greenish gray, moderately fissile, and silty. The upper contact with the overlying Nancy Member is sharp and planar.

Sedimentary structures observed within these beds include parallel laminae, ripple-drift laminae, convolute laminae, and wavy laminae. These, again, are arranged in such a way as to form parts of Bouma sequences. Typically, division Tc–Te can be seen, though a number of beds show a Tb–Te sequence. One of the differences observed between this section and the Griffin Hollow section (previous stop) is a decrease in the thickness of individual sandstone beds. It is also evident that there are fewer sandstone beds, and that fewer sandstone beds truncate or cut through underlying layers; moreover, many exhibit a more tabular geometry. The shale interbeds are again interpreted to be hemipelagic in origin.

Sole marks are poor to moderately developed and include a variety of tool marks such as groove casts, prod casts, brush, bounce, and roll marks; scour marks such as flute casts; and load casts. Paleocurrent directions here show a 300° orientation, which is the same as that measured for similar beds at the previous stop.

Trace fossils are quite abundant and diverse and can generally be characterized as complex horizontal feeding burrows and trails. Megafossils that have been recognized in the Farmers include broken and abraded brachiopods, fenestrate bryozoans, echinoderm debris, gastropods, and mollusks. Most of these specimens occur as lags along the base of individual sandstone beds and are thought to have been transported by turbidity currents.

In review, we believe that the sandstone beds in the Farmers Member at this location were deposited as proximal turbidites. The shales are thought to represent an accumulation of hemipelagic muds. The turbidite sequence at Evans Chapel is placed a few miles to the west of the main depocenter, and the relative paleogeographic position of the Farmers Member at this section is shown in Figure 10.

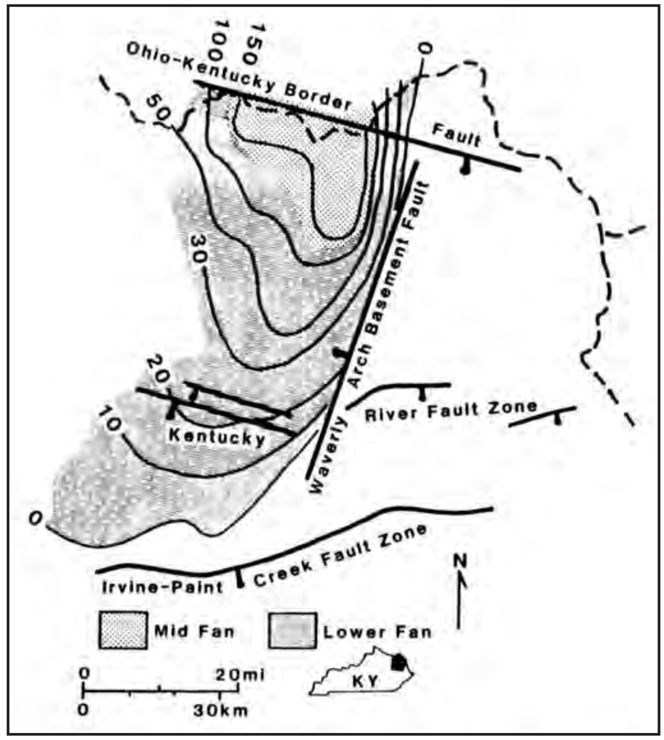


Figure 10. Isopachous map in feet of the Farmers Member of the Borden showing the shape of the fan and its probable features. The fan is divided into mid- and lower-fan areas based on criteria discussed in Normark (1978). After Sable and Dever (1990) and Pashin and Ettensohn (1987).

Henley Bed of the Farmers Member

The Henley Bed is included within the Farmers Member of the Borden Formation (Fig. 9). In the Vanceburg area it ranges in thickness from 10 to 40 ft (3 to 12.1 m). At this location the Henley Bed is approximately 34 ft (10.4 m) thick (Fig. 9). Here it consists mainly of a medium gray to greenish gray, poorly fissile, silty shale that weathers to a medium brown color. A number of distinctive layers are maroon in color, presumably due to the presence of finely disseminated hematite in the shale. In addition to the thinning of the shale intervals, the Henley Bed at this stop contains a reduced number of very fine-grained, argillaceous sandstone beds. These sandstone beds are quite tabular, thinly bedded, and have sharp erosional bases. Load casts, small flute casts, and some tool marks are found on the soles of these beds. Internally, thin parallel laminae predominate. Their upper surfaces are quite bioturbated and grade into the overlying shales. The upper contact of the Henley with the overlying Farmers Member is quite sharp.

A layer of argillaceous dolostone approximately 4 to 5 in. (10–12 cm) above the basal contact with the Sunbury Shale is also present in the Henley Bed at this locality (Fig. 9). This layer is about 14 in. (36 cm) thick and can be described as a muddy dolomicrite. The dolomite is ferroan and occurs as euhedral to subhedral rhombs within an argillaceous matrix. Again, there is no evidence of relict carbonate textures within this dolostone, and the layer appears to have formed prior to compaction and burial of the surrounding shales. This conclusion is again based on the presence of three-dimensional burrows within the dolostone, which are absent in the surrounding shale.

The Henley Bed here is thought to reflect the extremely slow accumulation of hemipelagic muds under dysaerobic conditions in the deepest parts of the depositional basin, with occasional or minor influxes of distal turbidites.

Sunbury Shale

The Sunbury Shale in the Vanceburg area ranges in thickness from 10 to 20 ft (3–6.1 m). At this location the Sunbury is approximately 14.1 ft (4.3 m) thick. It consists of black to olive-black, fissile shale that is carbonaceous and contains scattered crystals of pyrite. When weathered it takes on a pale yellowish brown to light brown color. The lower contact of the Sunbury Shale with the underlying Berea Sandstone is quite sharp, and a lag deposit is found at the base of the Sunbury, which is as thick as a half inch. It is primarily composed of pyrite and phosphatic fossil debris. Vertical and horizontal burrows, infilled with greenish gray shale of the Henley, are found in the upper 8 in. (20 cm) of the Sunbury.

Megafossils are very rare except for local occurrences of the brachiopods *Lingula* and *Orbiculoidea.* An exception to this is the fossil lag at the base of the unit, which contains inarticulate brachiopods, conodonts, fish scales, sharks' teeth, dermal plates, and carbonized plant remains.

The Sunbury Shale is a transgressive black shale (Ettensohn and others, 1988b) representing extremely slow accumulation of organic-rich hemipelagic muds in the deepest parts of the depositional basin, generally under anoxic conditions. The decreased thickness of the Sunbury in this area may reflect the fact that it accumulated under a stable pycnocline higher on the shelf or shelf margins where less accommodation space was available (*see* Figure 11).

Berea Sandstone

The Berea Sandstone in this area ranges in thickness from 14.3 to 145 ft (5–44.2 m). At this location the Berea is approximately 14.3 ft (5 m) thick. It consists of light gray to buff-colored sandstones with thin interbeds of greenish gray shale. Compositionally, these are fine-grained, micaceous quartzarenites that are moderately well indurated. The sandstones are medium bedded with sharp erosional bases that have small load casts, as well as the casts of tracks and trails on their surfaces. The upper surfaces of several beds display numerous straight-crested oscillation ripples. Ripple crests show a mean orientation of approximately 300°, which suggests a bidirectional current trend or line of direction of 300°. Sedimentary structures observed within these beds include primarily parallel laminae and small ripple cross-laminae. The Berea Sandstone, especially in this area, shows a strong facies relationship with the underlying Bedford Formation. No fossils other than horizontal and vertical trace fossils have been reported from the Berea in this area. A note of special interest here is that the Berea Sandstone is quite similar to the Farmers Member of the Borden Formation and is hard to distinguish from it in isolated outcrops; however, the Farmers Member always contains an abundance of the trace fossil *Zoophycos,* whereas it is never found in the Berea Sandstone.

The predominance of storm sedimentation during Berea deposition may have precluded habitation by the *Zoophycos* animal. The Berea here largely exhibits sheet-siltstone beds typical of storm-dominated shelf deposition. However, the fact that the Berea has thinned to approximately one-sixth of its thickness 8.5 mi to the northeast indicates that we are moving off the shelf edge to the south and west where the upper tongue of the Berea thins and eventually pinches out into the Bedford Shale (Fig. 11).

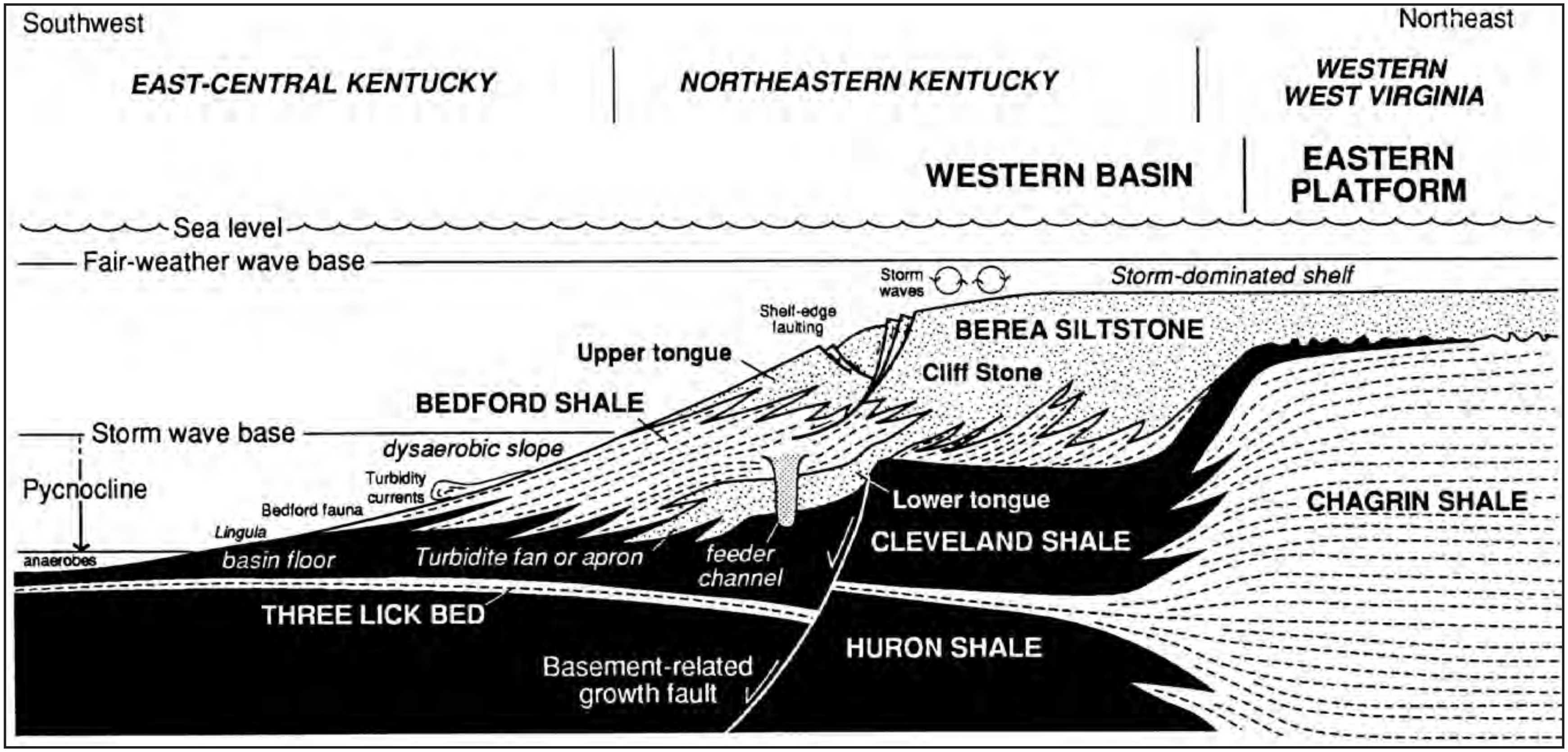


Figure 11. Depositional model for the Bedford-Berea sequence of eastern Kentucky and western West Virginia (after Pashin and Ettensohn, 1987, and Pashin, 1990). A thin, aggradational shelf sequence was preserved on the eastern platform, whereas a thick, progradational sequence that spans shelf, slope, and basinal environments was preserved in the western basin. Differentiation of platform and basin areas was related to relict topography and differential compaction of organic-rich black mud (Cleveland Shale) and relatively incompactible, organic-poor gray mud and silt (Chagrin Shale). Within the platform and basin areas, however, basement-fault reactivation was a significant control on facies distribution.

Bedford Shale

The Bedford Shale in this area ranges in thickness from 0 to 90 ft (0–27.4 m). At this location the upper 52 ft (16 m) of the Bedford is exposed, and it consists of silty shales that are intercalated with thin siltstones and subordinate beds of sandstone. This silty shale is greenish gray to purple-gray in color and weathers yellowish gray. Discontinuous interbeds of siltstone are light gray to greenish gray in color and occur as very thin flaser-, wavy-, and lenticular-bedded layers. A planar view of the lenticular layers indicates that they are basically starved ripples. Side views commonly show ripple-drift laminae. Interspersed among these finer-grained sediments are medium to thin, planar beds of sandstone. These are light gray to buff in color, very fine to fine grained, silty and micaceous quartzarenites. Both upper and lower contacts are sharp, and the upper surfaces of many sandstone beds display straight-crested oscillation ripples. Ripple crests show a mean orientation of 296°. Pyrite and marcasite commonly occur as crystal aggregates, small scattered crystals, or irregular nodules in this unit. The upper contact of the Bedford with the overlying Berea appears to be somewhat gradational. The lower contact here is not exposed.

Bioturbation is quite evident in this unit, particularly along the tops of individual sandstone beds. This generally includes horizontal tracks and trails, although vertical burrows are also common.

The presence of wavy-, flaser-, and lenticular-bedded shale and siltstones with wave ripples in the Bedford indicates distal storm deposition on the slope between the eastern platform and western basin (Fig. 11) (Pashin and Ettensohn, 1987). The presence of thick, wave-rippled siltstones near the base of the exposed Bedford at this locality may reflect the beginning of the lower tongue of the Bedford, which is mapped in this area (Morris and Pierce, 1967). Although the lower tongue of the Berea is typically composed of unrippled sheet-siltstone beds (Pashin and Ettensohn, 1987), the presence of ripples in these uppermost beds of the unit may reflect continued storm influence at or just above wave base (Fig. 11).

STOP 4: KINNICONICK CREEK
Lowstand Deposition in a Foreland Basin: Bedford-Berea Sequence (Upper Devonian), Eastern Kentucky and West Virginia

Jack C. Pashin and Frank R. Ettensohn

Introduction

Investigations of the Bedford-Berea sequence laid the foundation for models of epeiric sedimentation more than 40 years ago. Rich (1951a, b) included the Bedford-Berea in the original model in which he introduced the clinoform, a key element of modern sequence stratigraphy. Shortly thereafter, Pepper and others (1954) performed a landmark study of Bedford-Berea sedimentation that presented fundamental ideas regarding facies interpretation and regional paleogeography. Even in relatively recent references, their model is still regarded as a paradigm of epicontinental deltaic sedimentation (Krumbein and Sloss, 1963; Fisher and others, 1969; Wanless and others, 1970; Leblanc, 1975; Frazier and Schwimmer, 1987).

Sedimentologic advances since these classic studies have fostered new perspectives on the origin of the Bedford-Berea sequence (Coogan and others, 1981; Pashin, 1985, 1990; Pashin and Ettensohn, 1987, 1992b; Lewis, 1988). Pashin and Ettensohn (1987) applied an updated version of Rich's epeiric model (Woodrow and Isley, 1983) in tandem with continental-margin sedimentary models to the Bedford-Berea. A major implication of Pashin and Ettensohn's (1987) study is that traditional epeiric models, though widely applicable, are too simplistic to fully characterize epicontinental sedimentation, especially in tectonically active areas. Building on this theme, Pashin (1990) reevaluated the classic model of Pepper and others (1954) and demonstrated that sea-level variation, relict topography, differential compaction, and tectonism acted in concert to form the complex facies patterns of the Bedford-Berea sequence. This discussion characterizes the Bedford-Berea sequence in the Appalachian foreland basin of eastern Kentucky and West Virginia and shows how a major sea-level drop caused progradation of orogenic sediment into distal parts of the Appalachian Basin. Our intention is to show that sediment was accommodated in a basin formed largely by relict topography and differential compaction and to suggest ways in which basement-fault reactivation gave rise to diverse facies patterns within that basin. Today we will examine Bedford-Berea exposures that contain features not typically associated with epeiric sedimentation. At stop 4 we will examine a transect through part of a storm-dominated shelf margin that includes a synsedimentary fault and associated soft-sediment deformation structures.

Stratigraphic and Sedimentologic Framework

The Upper Devonian Bedford-Berea sequence is part of a thick succession of organic-rich, basinal black shale and intervening, light-colored shelf and coastal-plain clastics of Devonian and Mississippian age that extends across the North American craton (Heckel and Witzke, 1979; Ettensohn and Barron, 1981; Ettensohn, 1985a, b). This succession includes the Catskill and Pocono clastic wedges that were shed into the then-euxinic Appalachian Basin from the Acadian orogen (Fig. 12). The Bedford-Berea sequence separates the Catskill and Pocono wedges, and Elam (1981) defined the sequence as the interval between the organic-rich, black Cleveland Member of the Ohio Shale and the similar Sunbury Shale, as well as equivalent strata in the Appalachian Basin (Fig. 13). Bedford-Berea deposition occurred mainly in the western part of the Appalachian Basin and took place in two distinctive provinces, the eastern platform and the western basin (Pashin, 1990) (Figs. 14–15). In West Virginia, sediment accumulated mainly on the eastern platform, where the Bedford-Berea is generally thinner than 40 ft (12 m) and rests disconformably on the Catskill wedge. Near the West Virginia–Kentucky border, however, the Bedford-Berea thickens to more than 120 ft (35 m) and rests conformably on the Cleveland Shale. The transition from platform to basin is less than 5 mi wide and coincides with the shift from a thick shelf-slope clastic sequence dominated by gray shale and siltstone of the Catskill wedge (Chagrin Shale) to a relatively thin, basinal black-shale sequence (Cleveland Shale) (Figs. 12, 14).

Eastern Platform

The Berea Sandstone accounts for nearly all of the Bedford-Berea sequence on the eastern platform of West Virginia (Figs. 15–16). The sandstone is absent in eastern West Virginia, but in the central part of the state, the sandstone is pebbly and occurs in two northeast axes called the Gay-Fink and Cabin Creek trends (Pepper and others, 1954; Larese, 1974). A detailed isopach map by Larese (1974) established that the Gay-Fink trend is branching and that the Cabin Creek trend is funnel-shaped (Fig. 16). Both trends contain sequences that fine upward from sandstone with quartz pebbles to gray shale (Larese, 1974), and the black Sunbury Shale extends farther east within the trends than in adjacent areas (Pepper and others, 1954). The Gay-Fink trend, moreover, is bounded by the basement faults of the Rome Trough (Fig. 15). The Gay-Fink and Cabin Creek trends extend into a widespread, westward-fining sandstone blanket that is 40 to 60 ft (12–18 m) thick and extends to the platform margin (Fig. 15). Potter and others (1983) determined that the blanket is composed of sheet-sandstone beds with shale intraclasts near the base and wave ripples at the top. In the easternmost part of the blanket, the sandstone contains quartz pebbles and heavy-mineral concentrations (Rittenhouse, 1946; Larese, 1974). The sandstone blanket generally overlies light-colored clastics of the Catskill wedge on the eastern platform, including the Chagrin Shale, but near the platform margin, the sandstone generally overlies a thin tongue of the Cleveland Shale (Pashin, 1990). Previous workers stressed a fluvial-deltaic origin for the Gay-Fink and Cabin Creek trends (Pepper and others, 1954; Larese, 1974), but sedimentologic relationships indicate that the trends formed largely in an estuarine setting (Pashin, 1990). The branching geometry (Fig. 16) and fining-upward sequence of the Gay-Fink trend is typical of contributive fluvial systems, and the presence of the transgressive, basinal Sunbury Shale at the top of the sequence suggests that the trend is partly an estuary deposit. Branching estuaries, like Chesapeake Bay, form by transgression and aggradation in deeply incised paleovalleys (Fairbridge, 1980); hence, reactivation of Rome Trough basement faults gave rise to considerable topographic relief around the Gay-Fink trend. The Cabin Creek trend, in contrast, is

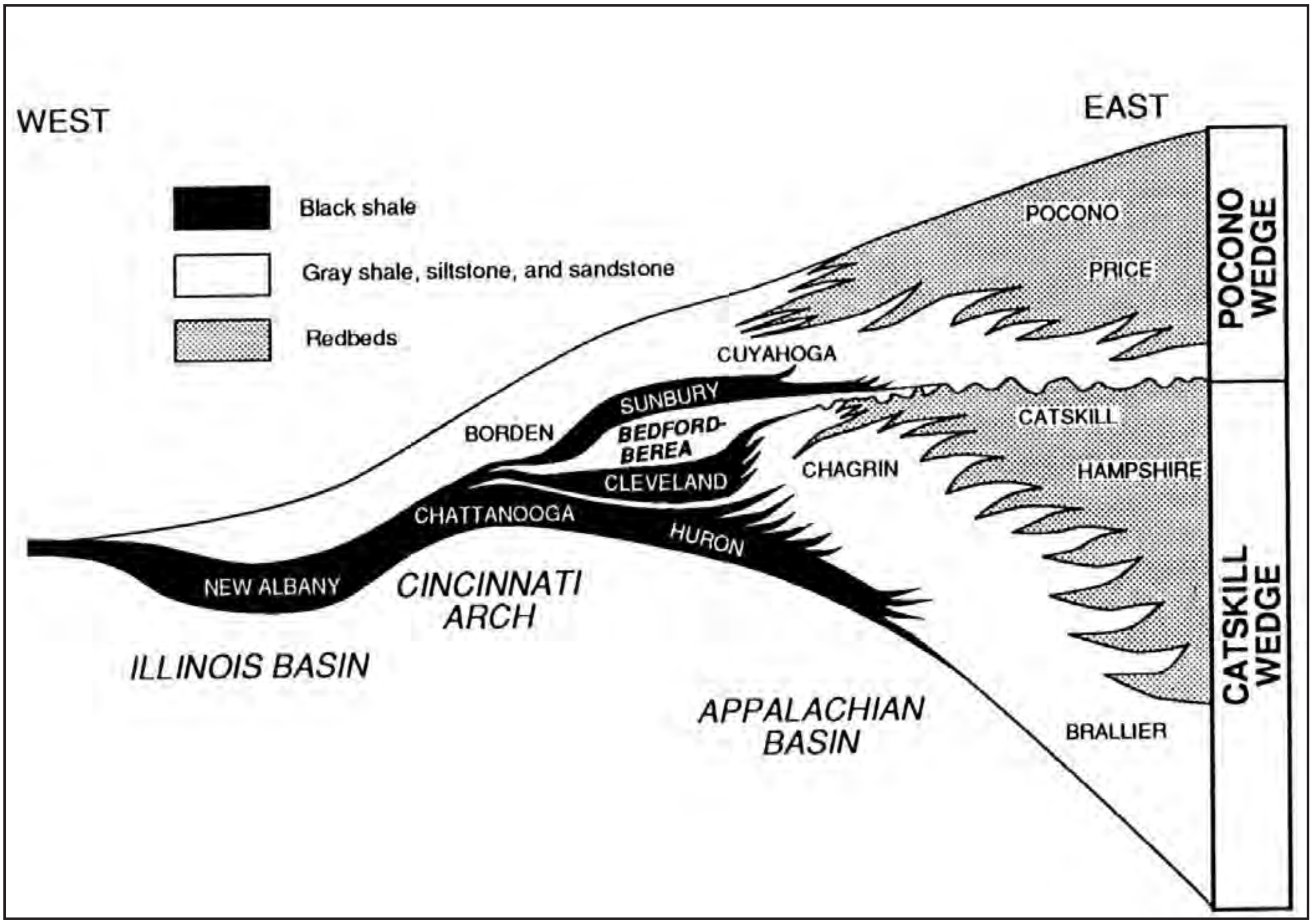


Figure 12. Relationship of the Bedford-Berea sequence to the Catskill and Pocono clastic wedges (after Pashin, 1990). A major downward shift in coastal onlap suggests that the Bedford-Berea sequence is the product of major relative sea-level drop that interceded Catskill and Pocono deposition.

interpreted to represent a funnel-shaped estuary. Funnel-shaped systems, such as the Gironde estuary of France, typically form in response to inundation of transitive coastal-plain channels and have a lower gradient than branching systems (Fairbridge, 1980). The rippled sheet-sandstone beds, which compose the bulk of the sandstone blanket, have been interpreted to be storm-dominated shelf deposits (Potter and others, 1983), and sandstone with heavy-mineral concentrations and eastern fringe of the blanket has been interpreted to be beach deposits (Larese, 1974). Extension of the axial channels of the Gay-Fink and Cabin Creek trends into the shelf area indicates that the trends had a major constructive phase prior to estuary formation that may have culminated in regional exposure of the eastern platform, valley incision, and perhaps formation of platform-margin deltas. The shelf, beach, and estuary deposits apparently accumulated late in Bedford-Berea deposition in response to regional transgression and aggradation of the eastern platform.

Western Basin

The stratigraphy of the western basin differs markedly from that of the eastern platform, and the Bedford-Berea includes the Berea Siltstone, Bedford Shale, part of the Cleveland Member of the Ohio Shale, and part of the New Albany Shale (Fig. 13). Along the platform margin, the sandstone blanket passes into more than 100 ft (30 m) of Berea Siltstone (Figs. 14–15). The siltstone intertongues with the Bedford Shale, forming a thick progradational package, and thins to a feather edge less than 50 mi from the platform margin. Southwest of the Berea pinchout, the Bedford Shale makes up most

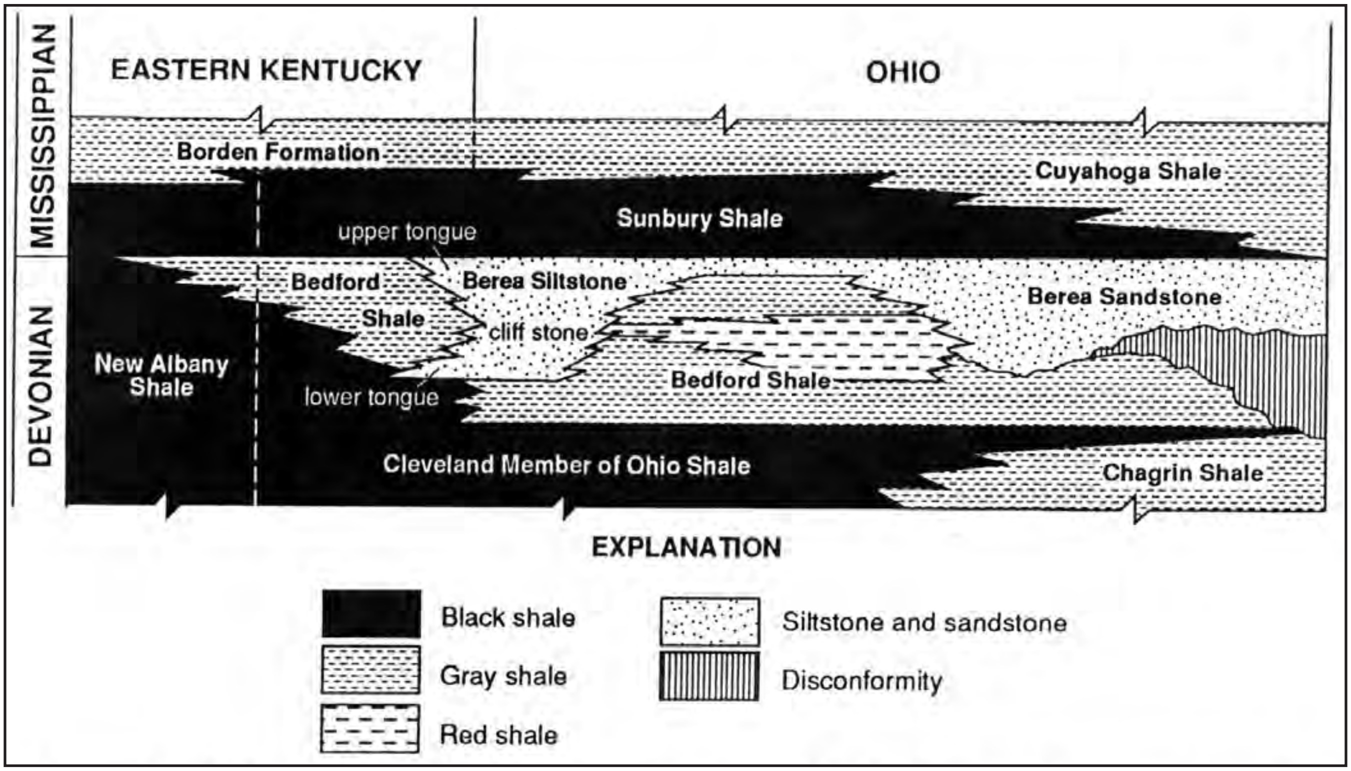


Figure 13. Stratigraphy of the Bedford-Berea sequence along the western outcrop of the Appalachian Basin from eastern Kentucky to northern Ohio (after Pashin, 1990).

of the Bedford-Berea sequence in a belt up to 50 mi wide (Fig. 15), and southwest of that belt the sequence contains exclusively black shale (Ettensohn and Elam, 1985).

In the outcrop of northeastern Kentucky, the Berea Siltstone is divided into upper and lower tongues that are separated by a northeast-thinning wedge of the Bedford Shale (Morris and Pierce, 1967; Pashin and Ettensohn, 1987); where both siltstone tongues are combined, the Berea forms the cliff stone of Hyde (1953) (Figs. 13–14). In east-central Kentucky, where the Berea is absent, the Bedford separates the Sunbury Shale from the Cleveland Member of the Ohio Shale and is locally mapped with the New Albany Shale (Fig. 13). Black shale equivalent to the Bedford-Berea has been recognized in the Cleveland Shale and the New Albany Shale in most of eastern Kentucky (Swager, 1978; Elam, 1981; Ettensohn and Elam, 1985).

The Berea Siltstone contains mostly thick sheet-siltstone beds with local ball-and-pillow structures; they are amalgamated or are separated by wavy, lenticular, and flaser-bedded shale and siltstone. Rippled sheet siltstone (Fig. 17) contains wave ripples and hummocky strata and is characteristic of the cliff stone and the upper tongue (Rothman, 1978; Potter and others, 1983; Pashin and Ettensohn, 1987); the wave ripples are renowned for abundance and consistent orientation (Hyde, 1911; Potter and Pettijohn, 1977). The lower tongue of the Berea contains two lithofacies: the lower-sheet-siltstone lithofacies, and the massive-siltstone lithofacies (Pashin and Ettensohn, 1987). In contrast to the upper tongue and cliff stone, the lower sheet siltstone is composed mainly of unrippled sheet-siltstone beds with gradational, bioturbated tops and complete Bouma sequences (Fig. 17). The massive siltstone is restricted to a series of outcrops near Walnut Grove Church and is part of a channel-fill complex that truncates the lower sheet siltstone (Morris and Pierce, 1967; Pashin and Ettensohn, 1987).

In northeastern Kentucky, the Bedford Shale contains wavy-, flaser-, and lenticular-bedded shale and siltstone with wave ripples (Fig. 17). Most of the Bedford in Kentucky, however, is com-

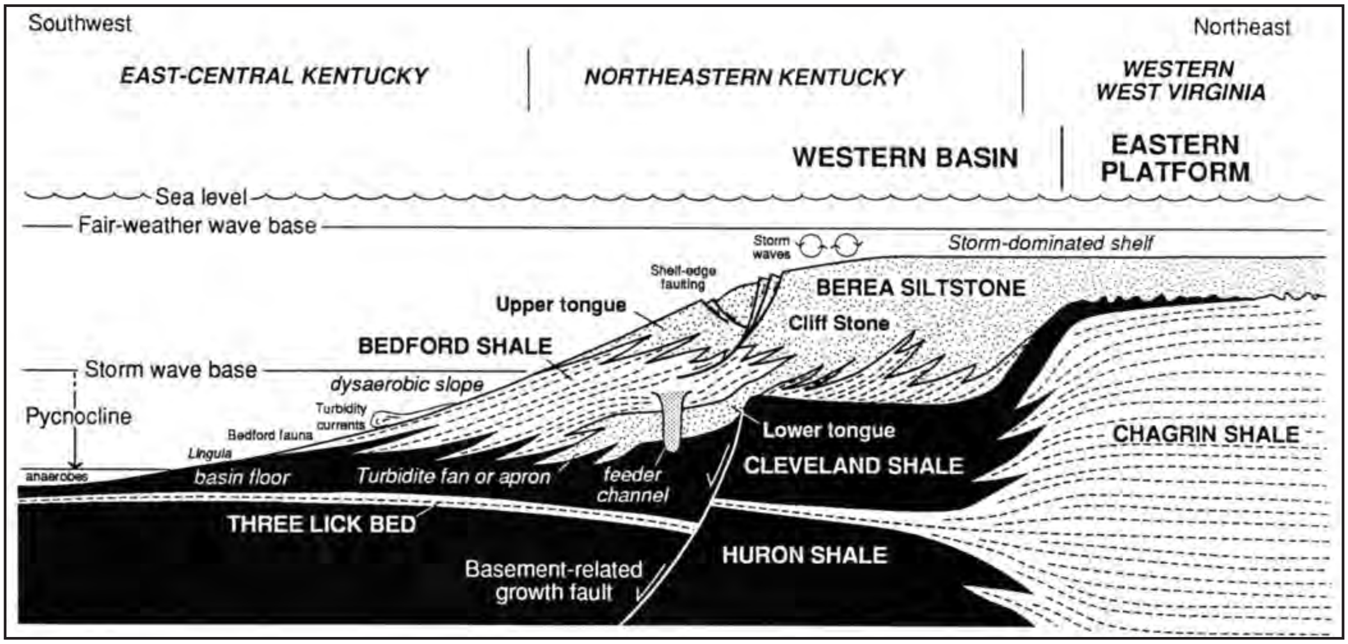


Figure 14. Depositional model for the Bedford-Berea sequence of eastern Kentucky and western West Virginia (after Pashin and Ettensohn, 1987, and Pashin, 1990). A thin, aggradational shelf sequence was preserved on the eastern platform, whereas a thick, progradational sequence that spans shelf, slope, and basinal environments was preserved in the western basin. Differentiation of platform and basin areas was related to relict topography and differential compaction of organic-rich black mud (Cleveland Shale) and relatively incompactible, organic-poor gray mud and silt (Chagrin Shale). Within the platform and basin areas, however, basement-fault reactivation was a significant control on facies distribution.

posed of poorly fissile gray shale containing thin, wavy, unrippled siltstone beds. Many of the siltstone beds are graded and contain a Bouma Tc,d,e sequence (Pashin and Ettensohn, 1987, 1992b). A thin-shelled brachiopod-mollusc fauna called the Bedford fauna (Morse and Foerste, 1909) occurs at the base of the Bedford and in gray-shale tongues between black shale (Pashin and Ettensohn, 1992b). Bedford-Berea black shale is brittle, fissile, and has been classified as a "ribbed," regressive, black shale by Ettensohn and others (1988b). The black shale locally contains abundant inarticulate brachiopods, particularly where black and gray shale intertongue. Like the sandstone blanket of West Virginia, the rippled siltstone beds of the upper tongue, cliff stone, and Bedford Shale have been interpreted to represent shelf storm deposits (Rothman, 1978; Potter and others, 1983; Pashin, 1985; Pashin and Ettensohn, 1987) (Figs. 14, 17). The cliff stone and upper tongue contain proximal storm deposits, whereas the Bedford contains distal storm deposits. In the lower tongue, unrippled siltstone beds evidently compose a toe-of-slope turbidite apron or fan, and the massive siltstone apparently is a turbidite feeder-channel fill (Pashin, 1985; Pashin and Ettensohn, 1987).

In the Bedford, unrippled siltstone occurs at the base of mud-rich turbidites that accumulated as a series of dysaerobic, toe-of-slope aprons (Pashin and Ettensohn, 1987, 1992b) (Figs. 18–19). Bedford-Berea black shale, in contrast, is interpreted to represent pelagic sedimentation on an oxygen-deficient basin floor (Ettensohn and Elam, 1985; Pashin and Ettensohn, 1987). The organic-rich black mud was apparently too foul to accommodate benthos other than inarticulate brachiopods, but recycling of nutrients from black mud evidently provided a nutrient-rich habitat for the Bedford fauna along the terminal fringe of the mud-turbidite aprons (Pashin and Ettensohn, 1992b).

The storm deposits and turbidites of the western basin contrast strongly with the aggradational shelf and estuary deposits of the eastern platform and represent a prograding outer shelf-slope system (Fig. 14). Much of the mud and silt in the western basin may have been derived initially from platform-margin shoal-water deltas as the Gay-Fink and Cabin Creek paleovalleys were

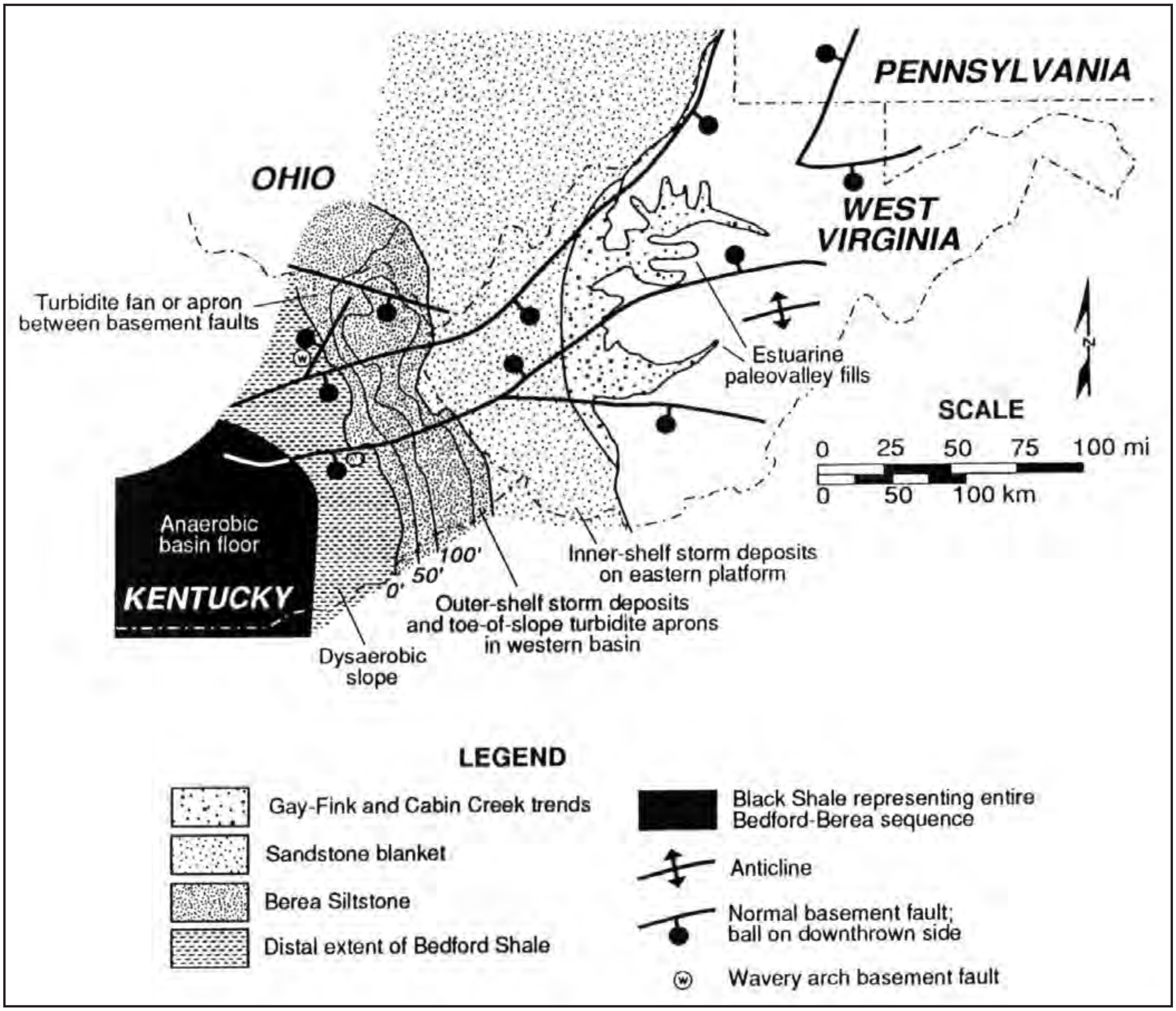


Figure 15. Bedford-Berea paleogeography, eastern Kentucky and West Virginia (after Pashin, 1990). Sand-rich estuary and shelf deposits predominated on the eastern platform, and silt- and mud-rich shelf, slope, and basinal environments predominated in the oxygen-deficient western basin. Structural control of sedimentation is apparent from preservation of the branching paleo-valley-estuary deposit of the Gay-Fink trend between major basement faults in West Virginia and by local deflection of isopach contours in the Berea Siltstone of eastern Kentucky.

incised. However, as the platform aggraded late in Bedford-Berea deposition, shelf progradation may have continued, and much of the silt and mud may have been transported basinward by storms that swept the platform. Rome Trough basement faults apparently influenced sedimentation in the western basin, because isopach contours are deflected eastward along the fault traces (Fig. 15). Lower-slope sedimentation apparently was influenced strongly by basement faulting, because the turbidite apron of the lower tongue is bounded by two of the faults (Figs. 14–15). According to Pashin and Ettensohn (1987), basement faulting culminated in formation of a southward-deepening shelf, but sedimentation eventually outstripped fault movement, giving rise to subdued topography at the close of Bedford-Berea deposition.

Discussion

The occurrence of the Bedford-Berea sequence in the western part of the Appalachian Basin (Fig. 12) and incision of the estuarine paleoval-

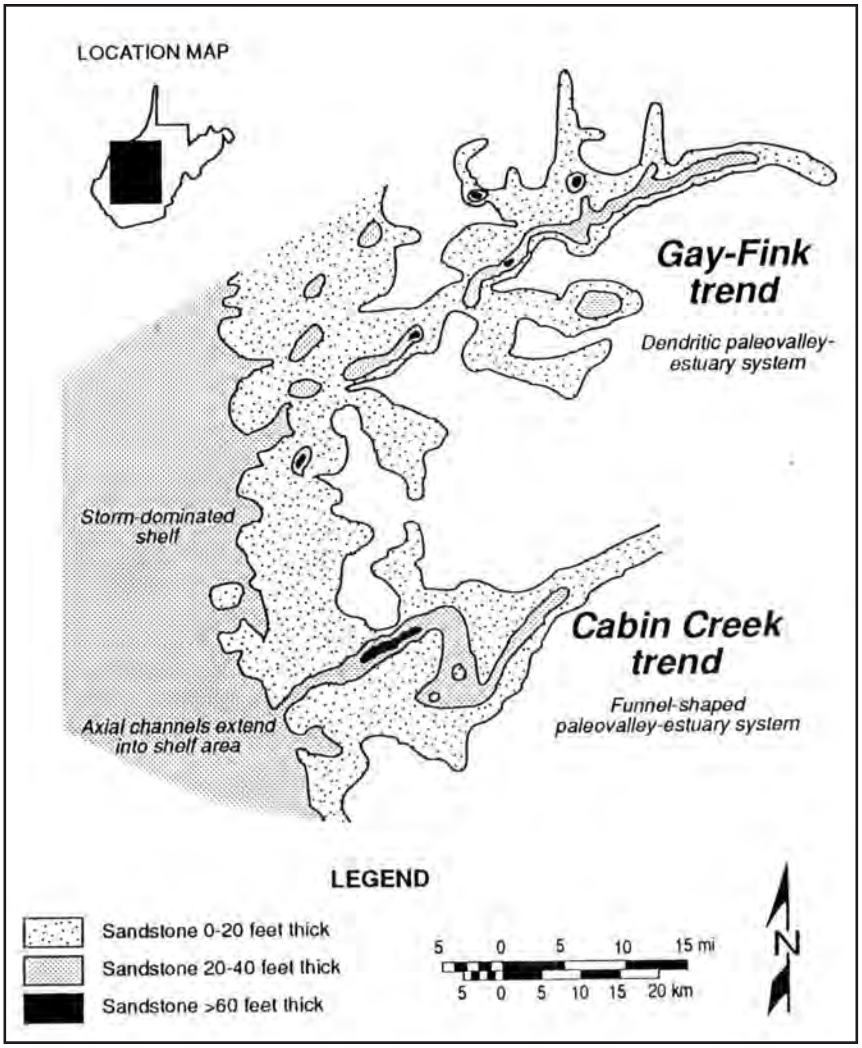


Figure 16. Isopach map of the Gay-Fink and Cabin Creek trends, West Virginia (after Larese, 1974). The Gay-Fink and Cabin Creek trends are interpreted to represent paleovalley-estuary systems. The Gay-Fink trend apparently represents a dendritic, or branching, estuary, whereas the Cabin Creek trend apparently represents a funnel-shaped estuary. Branching estuaries are associated with inundation of deeply incised valleys, whereas funnel-shaped estuaries are associated with inundation of low-gradient parts of coastal plains.

ley fills of the Gay-Fink and Cabin Creek trends into the Catskill clastic wedge (Figs. 15–16) indicate that the sequence represents a major downward shift of coastal onlap. Hence, the Bedford-Berea is the product of a significant lowstand that interceded Catskill and Pocono sedimentation. Pashin (1990) suggested that this sea-level drop was an interruption of the Acadian flexural-relaxation sequence that cannot be explained in terms of the regional tectonic framework, and Ettensohn (1990) suggested a similar interpretation based on analysis of pycnocline migration patterns; therefore, the Bedford-Berea lowstand was probably a eustatic event. Although most Bedford-Berea sediment was probably derived from orogenic sources in what is now the Virginia promontory of the Appalachian orogen (Fig. 20), incision of the Gay-Fink and Cabin Creek trends into the Catskill wedge (Figs. 12, 15–16) indicates that much of the sequence is reworked Catskill sediment that has been transported to the distal parts of the foreland basin.

Differentiation of the eastern platform and western basin apparently was a response to sedimentation atop the Catskill clastic wedge. Occurrence of the platform-basin transition above the facies change from gray, shelf-slope Chagrin Shale to black, basinal Ohio Shale (Figs. 12, 15) suggests that relict topography was an important controlling factor. However, differential compaction of organic-rich black mud and relatively incompactible, organic-poor gray mud and silt is interpreted to account for much of the platform-basin transition. In central Ohio, the platform-basin transition was further enhanced by reactivation of a Grenvillian suture in the basement, but no such structure has yet been identified along the transition in Kentucky and West Virginia (Pashin, 1990).

Although the platform and basin were formed mainly by relict topography and differential compaction, reactivation of Rome Trough basement structures influenced facies distribution within both of these regions. On the eastern platform, structural control of the Gay-Fink trend is particularly conspicuous (Fig. 15). In the western basin, structural control apparently had the greatest influence early in Bedford-Berea deposition, particularly on turbidite deposition and formation of a southward-deepening shelf (Pashin and Ettensohn, 1987). As a mature, prograding shelf margin devel-

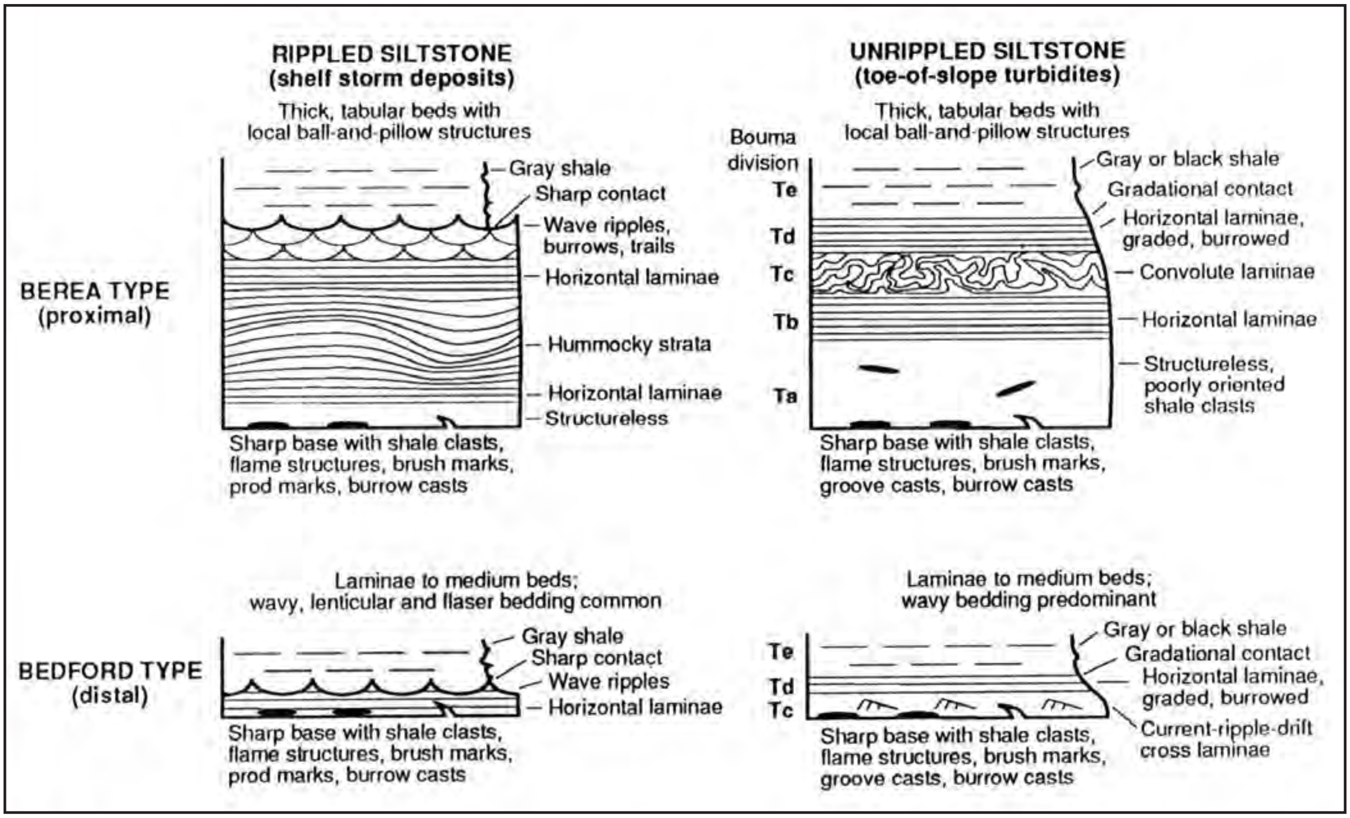


Figure 17. Idealized vertical sequences of sedimentary structures in Bedford-Berea sheet-siltstone beds. Bedford-Berea sheet siltstones are divided into rippled and unrippled types (Pashin and Ettensohn, 1987). Rippled sheet siltstone is interpreted to represent storm-dominated shelf deposits, whereas unrippled sheet siltstone is interpreted to represent turbidite-apron or fan deposits. Thick sheet-siltstone beds are proximal storm deposits and turbidites, and thin beds are distal examples.

oped near the close of Bedford-Berea deposition, however, sedimentation outstripped basement-fault movement (Pashin and Ettensohn, 1987), and the shelf-to-basin transition had the subdued topography first envisioned by Rich (1951a, b). The primary difference between this account of Bedford-Berea sedimentation and the classic studies of Rich (1951a, b) and Peppers and others (1954) is the realization that the sequence was deposited in a tectonically evolving foreland basin. Eustasy, differential compaction, and relict topography functioned in concert with tectonism to determine depositional history and regional paleogeography. Tectonism, relict topography, and differential compaction acted collectively to provide sediment sources and to establish the geometry of the sedimentary basin and the architecture of the basin fill. Eustatic sea-level variation, moreover, helped determine the position and rate of change of base level and was thus a critical factor that caused erosion of the Catskill wedge and restriction of thick, progradational sequences to the western basin. Interplay of these factors has resulted in myriad patterns of sedimentation in the geologic record, and like the Bedford-Berea sequence, each depositional sequence has intricacies that must be identified before a thorough knowledge of sedimentation in foreland basins can be achieved.

Storm-Dominated Shelf Margin Near Garrison

The Garrison area exhibits instructive exposures of the cliff stone (Figs. 13–14), which include some of the best examples of shelf storm deposits in the Appalachian Basin. At this stop, which is just south of Garrison along the Alexandria-Ashland Highway, we will examine the newest and most extensive exposure of the cliff stone and will observe a synsedimentary fault that allows new insight into the nature of the Bedford-Berea shelf margin. Approximately 100 ft (30 m) of the cliff stone is exposed along the Alexandria-Ashland

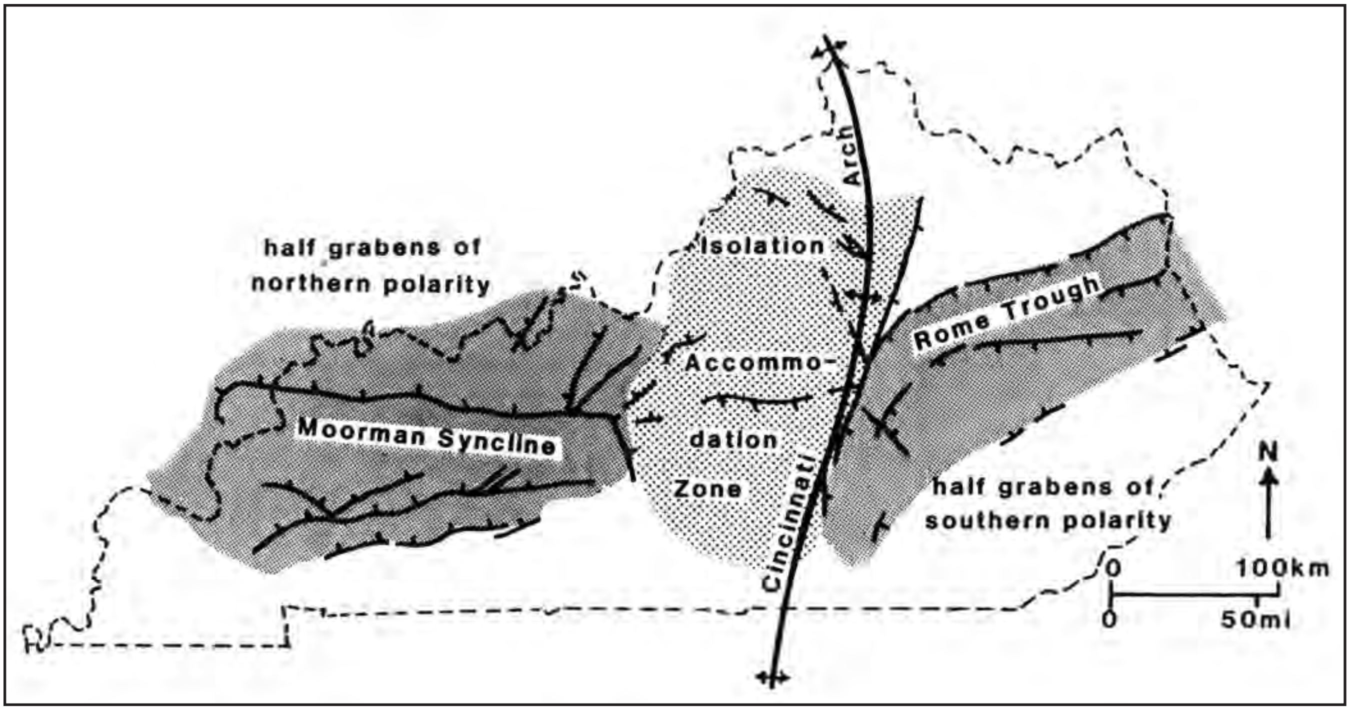


Figure 18. Generalized fault patterns and basement configuration in Kentucky showing that the Rome Trough and Moorman Syncline (Rough Creek Graben) may represent nonoverlapping, opposing half-graben systems formed during late Precambrian–Middle Cambrian Iapetan rifting. At various times, regional structures like the Lexington Platform, Cincinnati Arch, and tectonic bulges may have become localized on the intervening isolation accommodation zone.

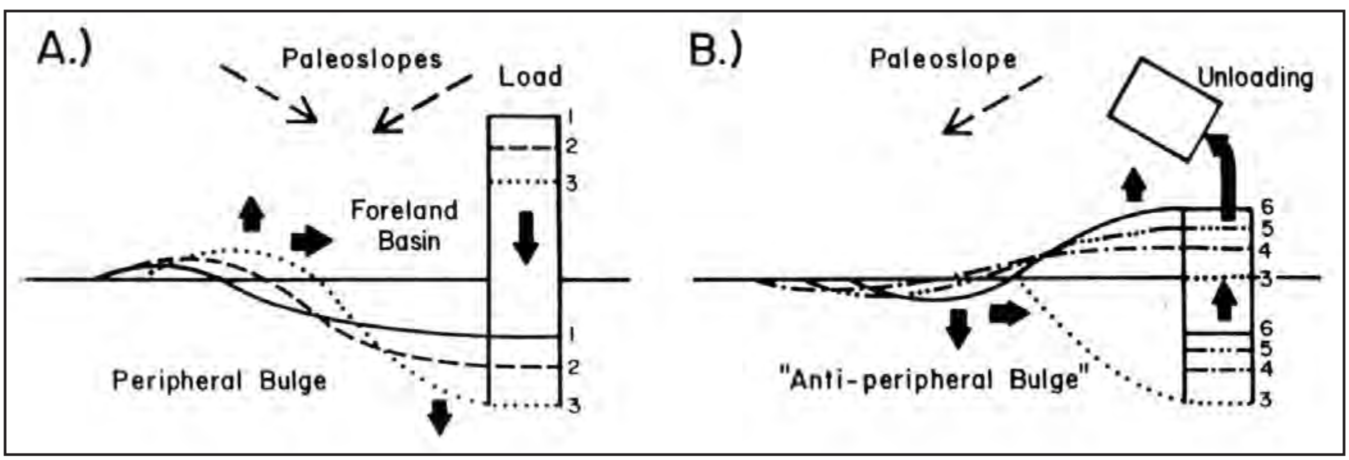


Figure 19. Two types of flexural response to lithospheric-stress relaxation. (A) “Loading-type” relaxation, static gravitational load results in deepening foreland basin and migration of peripheral bulge toward the load. (B) “Unloading-type” relaxation, erosional unloading results in rebound near unloaded area and an “anti-peripheral” bulge that deepens and migrates toward the former load (redrawn from Beaumont and others, 1988).

Highway, and outcrops show a general thickening- and coarsening-upward progression from the upper few feet of the Bedford Shale to the top of the Berea Siltstone (Fig. 21); the black, fissile Sunbury Shale and the gray, poorly fissile shale of the Henley Bed (Borden Formation) are exposed at the top of the roadcuts.

Rippled sheet siltstone is the most characteristic Bedford-Berea rock type and crops out from northeastern Kentucky to northwestern Pennsylvania (Pashin, 1990); the siltstone is even featured

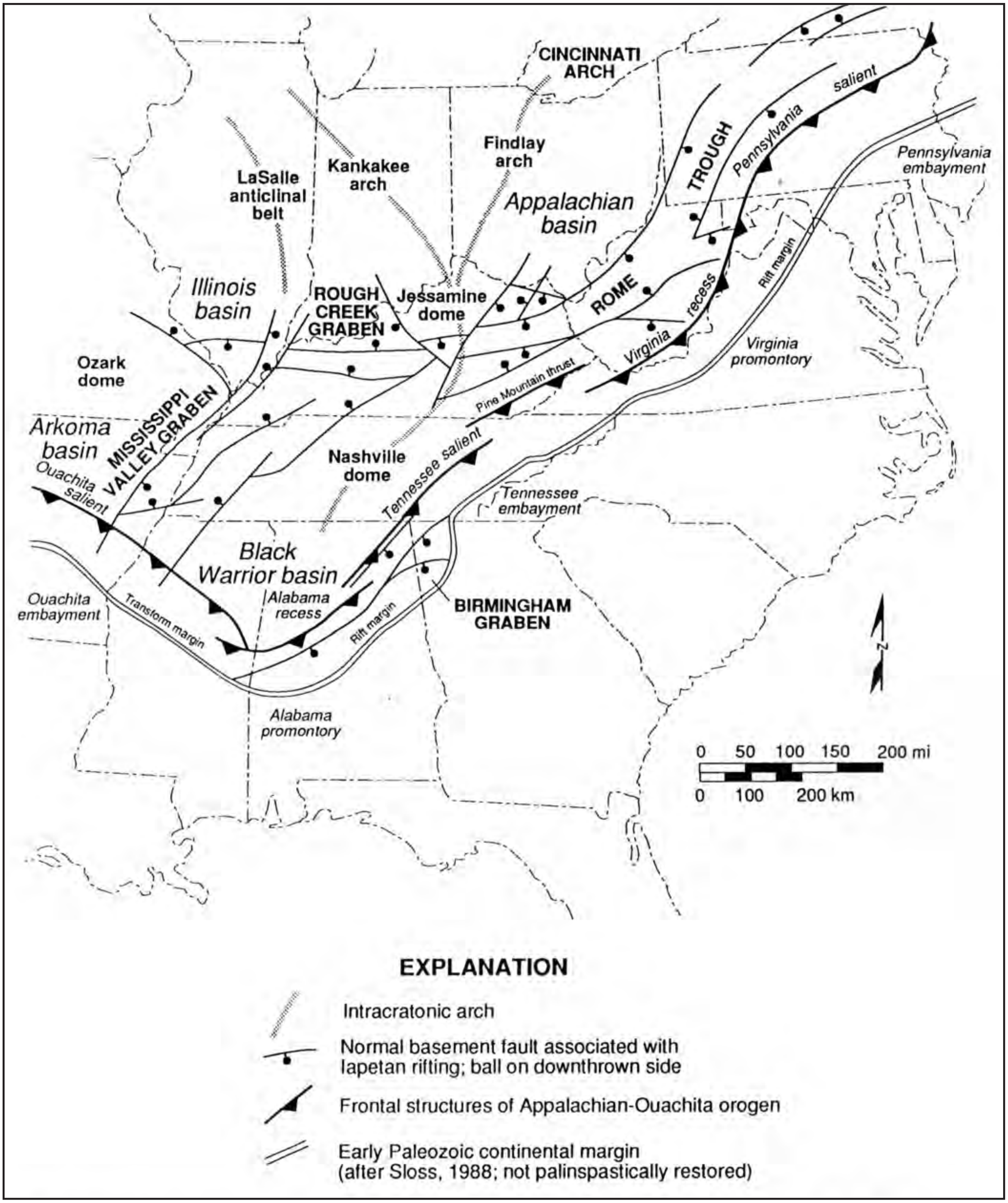


Figure 20. Structural geology of Kentucky and neighboring areas in relation to southeastern North America (after Thomas, 1991).

in classic studies of wave ripples (Hyde, 1911; Kindle, 1917; Bucher, 1919). Wave-ripple orientation is remarkably consistent in the exposure and ranges from 290° to 300° (Fig. 21). Hyde (1911) first noted this extreme uniformity of wave-ripple orientation, which is maintained into northern Ohio (Hyde, 1911; Lewis, 1988; Pashin, 1990). Comparing wave-ripple orientation from several depositional settings, Potter and Pettijohn (1977) suggested that variance of Bedford-Berea ripple orientation represents a lower limit for wave ripples in general.

Virtually all of the prominent features of the cliff stone can be observed along the Alexandria-Ashland Highway. Thick sheet-siltstone beds have sharp lower contacts with tool marks and load structures and ideally contain the following vertical sequence of sedimentary structures that is associated with shelf storm deposits (Goldring and Bridges, 1973; Hamblin and Walker, 1979; Dott and Bourgeois, 1982; Aigner and Reineck, 1982): (1) structureless siltstone overlain by (2) horizontal laminae that grade upward into (3) hummocky strata. Hummocky strata grade back into (4) horizontal laminae, which are, in turn, truncated by (5) wave-ripple cross laminae (Fig. 17); burrows and trails are abundant at the top of some beds. Thick sheet-siltstone beds are separated by intervals containing wavy, lenticular, and flaser-bedded shale and siltstone of Bedford type (Figs. 17, 21).

The sheet-siltstone beds are quite continuous at this stop, save for local ball-and-pillow zones (Figs. 21–23). However, near the southwest end of the roadcut sequence, the siltstone is disturbed by a fault that extends from the base of the exposure into the Sunbury Shale. The fault plane is irregular (Fig. 16) and strikes N20°E, approximately parallel to the Waverly Arch basement fault as mapped by Pashin and Ettensohn (1987) (Fig. 15). The fault is downthrown to the southeast, and the fault plane is apparently listric, dipping 80° near the top of the Berea and only 50° at road level. At the Berea-Sunbury contact, net throw is approximately 3 ft (1 m), but near the base of the exposure, net throw is approximately 6 ft (2 m). Tensional fractures are numerous adjacent to the fault, particularly in the hanging wall.

Although ball-and-pillow structures are common throughout the cliff stone, some unusual forms are associated with the fault (Fig. 24). Isolated boulder-size siltstone balls are common in both the hanging wall and the foot wall, and many beds are pierced by diapiric shale masses as wide as 3 ft (1 m). In the footwall, some sheet-siltstone beds overlap and terminate. In the upper bench of the outcrop, one bed contains ball-and-pillow structures only in the hanging wall; the structures are asymmetrical and are elongate parallel to the fault trace (Fig. 23). Farther west along the Alexandria-Ashland Highway, some ball-and-pillow structures are overthrust and resemble imbricate duplexes.

Indeed, these unusual soft-sediment deformation structures occur where the cliff stone splits into the upper and lower tongues of the Berea and thus appear to be related to synsedimentary tectonism at the shelf margin (Fig. 14). Parallelism of the soft-sediment fault with the Waverly Arch basement fault suggests deep-seated structural control of the shelf break. However, surface-fault displacement is to the southeast, opposite nearby basement structures and regional paleoslope, suggesting that it is an antithetic structure that is only part of a larger shelf-margin fault system. Regardless, increasing throw downward along the fault plane indicates that the fault grew during active sedimentation. Moreover, extension of the fault into the Sunbury Shale indicates that fault adjustment continued after Bedford-Berea deposition. Tensional fractures in the hanging wall contrast sharply with the abundant soft-sediment deformation structures and are interpreted to be a stress-release fracture system formed after faulting and lithification.

In addition to faulting, fluidized sediment failures occurred along the Bedford-Berea shelf margin (Cooper, 1943; Hyde, 1953; Pashin, 1990). Submarine sediment failures develop on slopes as gentle as 0.5° (Shepard, 1955) and commonly form in response to seismicity or storm-wave loading (Allen, 1982; Schwab and Lee, 1988). The fluidized failures are near the contact of the rippled and unrippled beds and have thus been interpreted to be a result of storm-wave loading (Pashin, 1990). Parallelism of shelf-margin faults to the nearby basement faults, however, suggests an underlying seismic cause. Even so, deep-seated basement control does not preclude storm-wave action as a mecha-

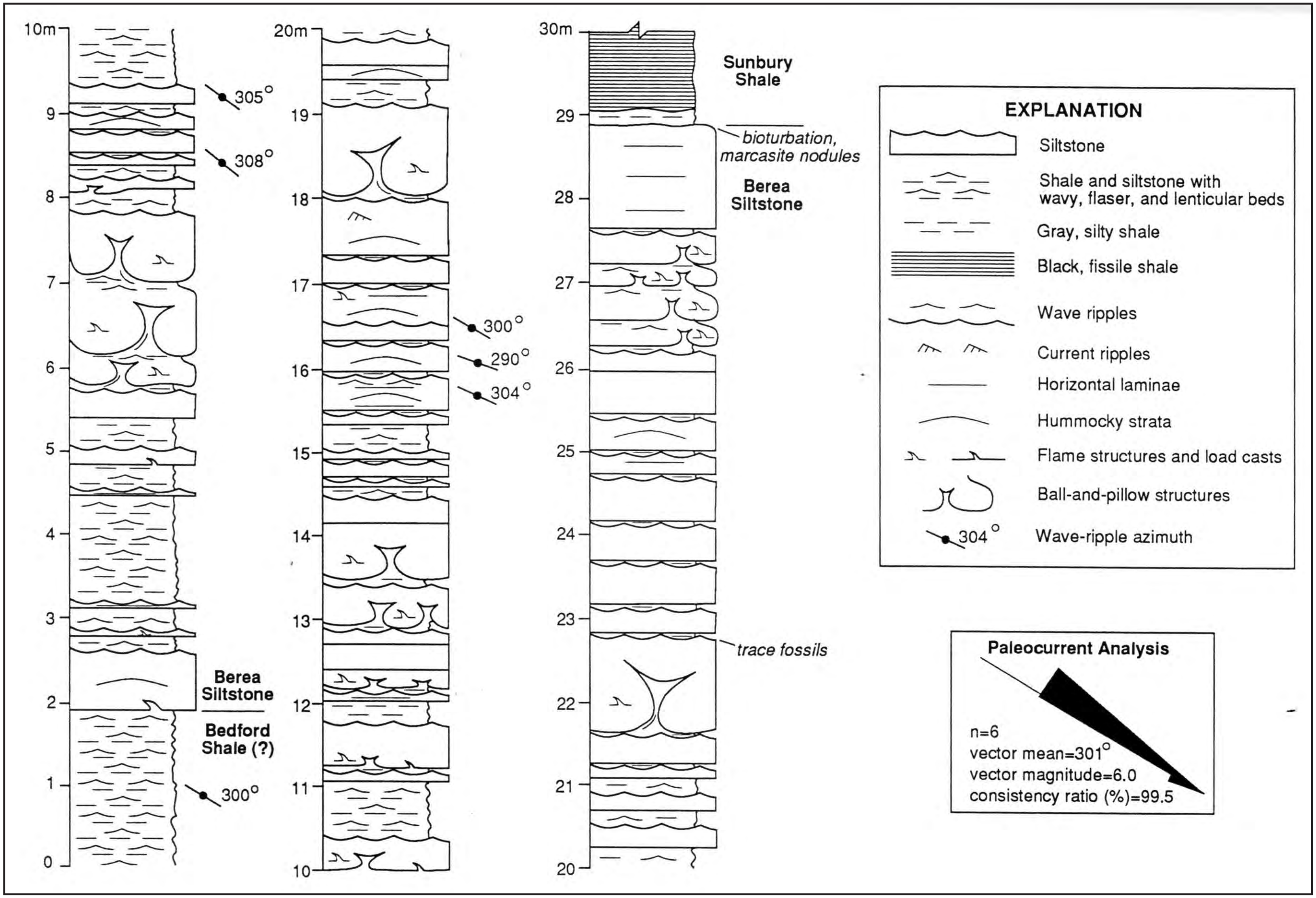


Figure 21. Measured section of Berea Siltstone (cliff stone), Alexandria-Ashland Highway. This section shows the basic characteristics of rippled siltstone in the cliff stone. The exposure shows a well-developed thickening- and coarsening-upward sequence from the upper part of the Bedford Shale to the top of the Berea Siltstone. Note the consistency of wave-ripple orientation in this exposure; uniform ripple orientation is maintained into northern Ohio.

Figure 22. Ball-and-pillow structures near base of Berea Siltstone, Alexandria-Ashland Highway. Ball-and-pillow structures are abundant in the cliff stone, and several morphologic types are preserved in this sequence of highway cuts. The examples in this photograph show complete piercement of sheet-siltstone beds by diapiric shale masses.

nism for adjustment of the faults, particularly antithetic structures. Therefore, the shelf-margin fault system may reflect slope instability above a reactivated basement discontinuity, but storm-wave action may have caused minor fault movement and contributed to shallow seismicity, and hence, to some of the soft-sediment deformation preserved along the Alexandria-Ashland Highway.

Outcrops of the Bedford-Berea sequence (stop 4 in part) contain examples of shelf and slope sedimentation not typically associated with epeiric sea floors. The contrasting submarine feeder channel and shelf-margin faults (Figs. 14, 24) demonstrate that the epeiric shelf-to-basin transition contains a much more diverse facies assemblage than was first envisioned by Rich (1951a, b) or has been accounted for in subsequent models. Complexities in the Bedford-Berea sequence are apparently related to oversteepened slopes along reactivated basement structures that gave rise to unusual facies assemblages and deformational structures. Continental crust contains numerous geophysical discontinuities that have diverse origins and orientations, which reflect a polyphase crustal history. The polyphase structures have evolved throughout geologic time and have acted collectively with depositional topography and relative sea-level variation to determine ancient patterns of sedimentation. For this reason, we are only beginning to glimpse the manifold paleomarine configurations of epeiric sea floors.

Figure 23. Asymmetrical ball-and-pillow structures adjacent to fault near top of Berea Siltstone, Alexandria-Ashland Highway. Fault reactivation apparently was penecontemporaneous with sedimentation, and many unusual soft-sediment deformation structures are preserved adjacent to the fault plane.

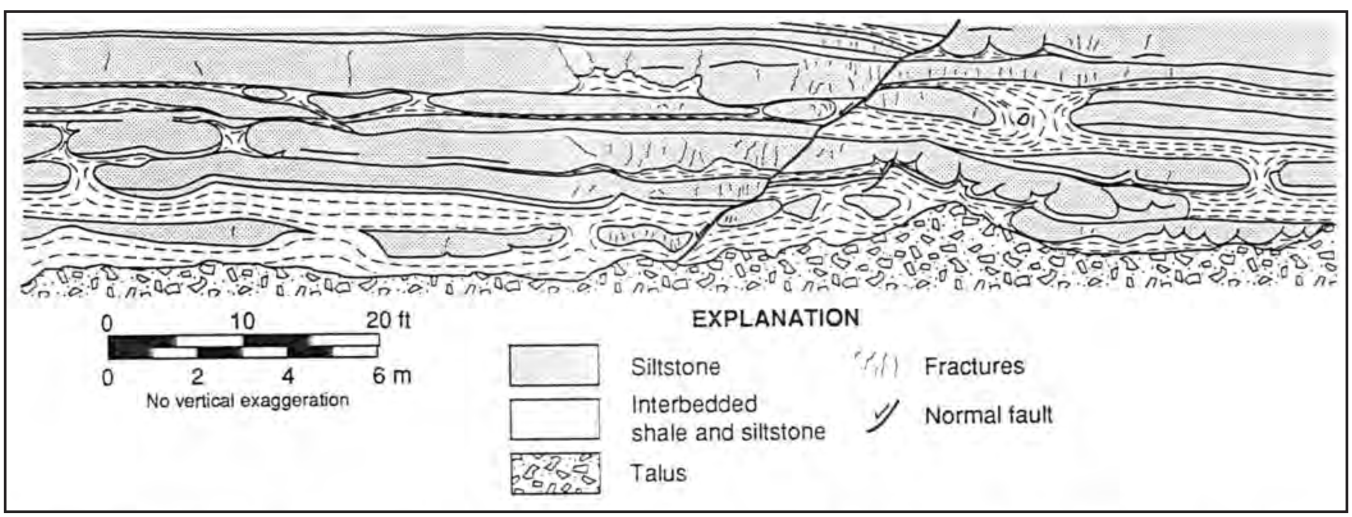


Figure 24. Diagram of synsedimentary fault in near base of Berea Siltstone, Alexandria-Ashland Highway. Isolated, boulder-size siltstone balls are common along the fault, and in the footwall, some sheet-siltstone beds overlap and terminate. Tensional fractures are numerous near the fault, particularly in the hanging wall. This fault is interpreted to be an antithetic structure that is part of a larger system of shelf-margin faults. The fault system is interpreted to be related to reactivation of nearby basement structures, and storm-wave loading may have contributed to surface faulting and associated soft-sediment deformation.

STOP 5: TYPE SECTION, THREE LICK BED

Frank R. Ettensohn, Charles E. Mason, and R. Thomas Lierman

At this stop we will quickly view the type section of the Three Lick Bed of the Ohio Shale (Provo and others, 1978). The Three Lick Bed is a prominent stratigraphic marker horizon (mid-Famennian, lower *Pa. expansa* Zone; Over and others, 2009) that consists of three thin beds of greenish gray shale separated by two beds of ribbed black shales; the lower gray shale bed is largely covered with slope debris at this exposure. It is recognized on gamma-ray logs by three closely spaced negative deviations separated by two positive deviations (Fig. 25). The unit thins to the west and south where the gray shales become little more than bioturbated black-shale horizons and the three negative gamma-ray deviations merge into a larger single one; in fact, the unit varies in thickness from nearly 40 m (152 ft) in far eastern Kentucky to just a few centimeters thick in central Kentucky (Ettensohn and others, 1988a). The unit has also been traced to a thin, bioturbated interval of the Clegg Creek Member, New Albany Shale, in the west-central Illinois Basin (Ettensohn and Geller, 1987). Gray shales in the unit may be

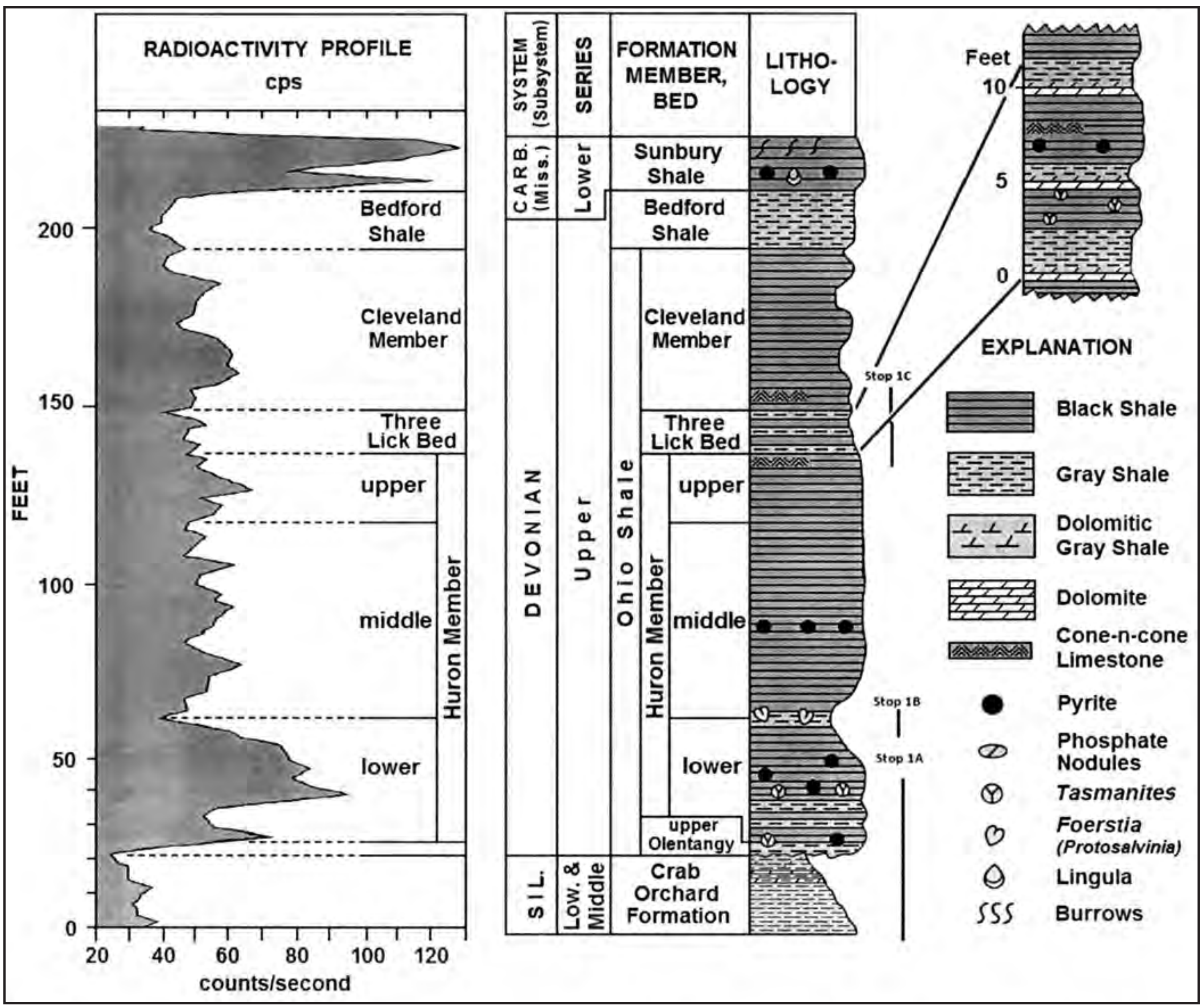


Figure 25. Schematic stratigraphic column. The radioactivity profile to the left is an artificial gamma-ray log that can be used for correlations in the surface or subsurface in homogeneous black shales.

intensely bioturbated, and benthic organisms, including in situ *Lingula*-like inarticulate brachiopods, agglutinated foraminifera, ostracods, as well as micromorph gastropods and pelecypods, have been reported (Barron and Ettensohn, 1981).

The Three Lick Bed represents the distal-most tongues of prodelta Chagrin Shales (Fig. 26) from northern Ohio (Provo and others, 1978) and from equivalent Catskill and Hampshire delta progradation to the east. The gray shales represent three brief dysaerobic periods, during which bottom waters coincided with the pycnocline, in a cratonic basin whose bottoms were otherwise below the pycnocline in anoxic conditions. Apparently, the bed represents three brief peri-

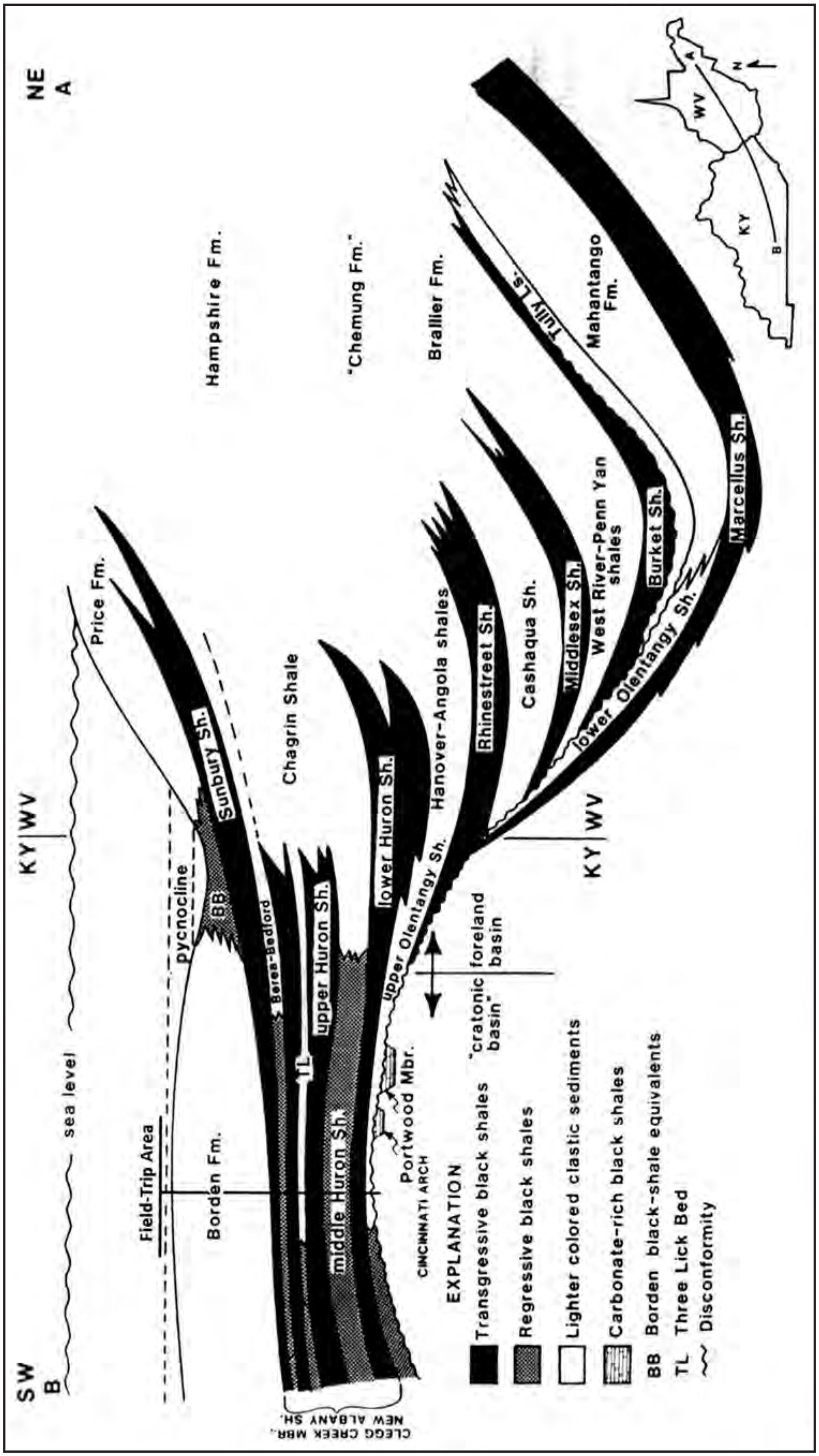


Figure 26. Schematic southwest-northeast section across the central Appalachian Basin in Kentucky and West Virginia, showing the locations of cyclic Middle Devonian to Early Mississippian black-shale basins and intervening clastic wedges formed in the second to fourth Acadian tectophases. Each black-shale unit represents subsidence in response to an episode of cratonward-migrating tectonism. The dark vertical line represents the approximate position of the field trip area. Adapted from Ettensohn and others (1988).

ods of sea-level lowstand that coincided with major deltaic progradations from the north and east. Sands in equivalent beds to the east produce hydrocarbons from the drillers' Gordon and Venango sandstones and siltstones (Boswell and others, 1996).

STOP 6: Silurian-Devonian Contact and Lower Black-Shale Section, Northeastern Kentucky

Frank R. Ettensohn, Charles E. Mason, and R. Thomas Lierman

Stop 6 is two-part stop beginning at what is commonly called the "Morehead exposure," listed below as stop 6A. This cut and successive cuts to the east along I-64 provide a complete exposure of the Devonian-Mississippian black-shale sequence in northeastern Kentucky. This is one of the most studied black-shale exposures in Kentucky (e.g., Chaplin and Mason, 1979; Kepferle and Roen, 1981; Ettensohn, 1992d, e; Lierman and others, 1992). The entire section is shown schematically in Figure 25, which includes an artificial gamma-ray log that can be used to discriminate and correlate black-shale units that are difficult to discriminate visually (Ettensohn, 1979).

Stop 6A: I-64 Along Eastbound Lane Just East of State Route 1722 Overpass After Crossing the Licking River Into Rowan County

At this stop we will view the Silurian/Devonian unconformity between the Crab Orchard and Olentangy and the basal Huron Member of the Ohio Shale. Latitude: 38°9′51.9″; longitude: –083°35′44.4″.

The section at stop 6A begins with approximately 14.2 m (54 ft) of Lower and Middle Silurian upper Crab Orchard Formation (Estill Shale Member). The unit is characterized by greenish gray, silty shale interbedded with thin-bedded, brownish gray dolosiltites to dolarenites with occasional lenses of quartzose sandstone and siltstone (Fig. 27). Much of the shale is reported to contain volcanic ash (Mason and others, 1992; Mason, 2002). The dolostones occur in packets of three to five beds that are each typically 1.5 to 7.0 cm thick. Some of the thicker beds exhibit sharp erosional bases, subtle grading, mud chips, and hummocky crossbeds; each bed is typically burrowed from above. In the upper 1.8 m (6 ft), dolarenites and quartzose sandstones may occur as lens-like, compacted, starved ripples.

The uppermost 23 cm (9 in.) is intensely weathered by oxidation to light brown to dark yellowish orange colors, although iron oxides have accumulated in the shales to depths at least as great as 1.8 m. (6.0 ft) and manganese oxides and gypsum have accumulated to depths as great as 6.6 m (21.5 ft). Most of this weathering is associated with a systemic paraconformity that separates rocks of the Silurian and Devonian systems (Figs. 25, 27), but at many places regionally, the unconformity is angular. Although only a few meters of Middle Silurian Bisher Dolostone and uppermost Crab Orchard Shale are missing here along the unconformity (McDowell, 1975), up to 135 m (443 ft) of Upper Silurian, Lower Devonian, and Middle Devonian section, which occurs to the north in Ohio or in the subsurface of eastern Kentucky, is absent in this outcrop belt.

Macrofossils have not been found at this exposure and are generally rare in the Crab Orchard. Nonetheless, at other Crab Orchard exposures, specimens comprising a microfauna or juveniles have been recovered following washing and other processing. Fossils found include ostracods, graptolites, scolecodonts, conodonts, sponge spicules and apparently juvenile forms of bryozoans, brachiopods, gastropods, pelecypods, crinoids, trilobites, nautiloids, and blastoids. Preservation is commonly by pyritization or calcite recrystallization, and filter feeders and sessile forms seem to dominate (Mason and others, 1992; Mason, 2002).

The Crab Orchard shales typically form hummocky to gullied slopes, and where exposed without cover, they flow and slump, generating classic badlands topography. Hence, they form poor foundation material for homes and roads, but are impermeable enough to be good sites for farm ponds (McDowell, 1975; Mason, 2002).

The Crab Orchard Shale exposed here is Early to Middle Silurian (late Llandovery, Telychian, to early Wenlock, Sheinwoodian) in age and was

Figure 27. The section at stop 6A (Fig. 25) showing the Crab Orchard, Olentangy, and Lower Huron Shales. The student is pointing to the Silurian-Devonian paraconformity, the rust-stained horizon. Trace fossils are very common in the interbedded black and green shales at the base of the Lower Huron. Note the prominent jointing in the Lower Huron Shale.

deposited during the first tectophase of the Salinic Orogeny, an Appalachian equivalent of the Scandian Orogeny along the Caledonian suture (Ettensohn and Brett, 1998; Ettensohn, 2008). In this part of Kentucky, the unit represents very distal parts of the subsiding foreland basin, whose main axis of subsidence was in New York (Ettensohn, 1994). Subsidence was probably in large part tectonic, but glacio-eustacy may have also played a role in deepening at the time (Caputo, 1998; Ettensohn and Brett, 1998). The Crab Orchard at this locality is thought to represent offshore, outer-shelf deposition, generally below wave base. The presence of hummocky crossbedding, crude grading, and sparse ripples suggests that the interspersed dolostone and sandstone layers probably represent distal storm deposits or tempestites. Although presence of a normal-marine, pelagic microfauna suggests that upper parts of the water column were aerobic, absence of typical larger-scale macrofauna, predominance of burrowing and a depauperate micromorph fauna, and presence of pyritization suggest that bottom waters were at least periodically dysaerobic. Although most dysaerobic faunas are typically dominated by vagile, molluscan deposit feeders or scavengers, in the Crab Orchard sessile, juvenile, filter-feeding bryozoans, brachiopods, and crinoids predominate. More information on this unit and its faunas can be found in studies by Mason and others (1992) and Mason (2002).

The paraconformity separating Silurian and Devonian rocks in this section is one of the most significant unconformities in Kentucky (Fig. 27), and it marks the base of Sloss's (1963) Kaskaskia cratonic sequence. Nearly everywhere in the state there is a major unconformity like this below the black-shale sequence, although different amounts of time and section are missing in different places. At this section, approximately 59 Ma of Silurian and Devonian time are missing along the surface, but the surface is actually a composite unconformity, having subsumed at least four Early and Middle Devonian periods of erosion based on more complete sections in the subsurface of eastern Kentucky. Most of the erosion probably occurred during Middle and Late Devonian time near the beginning of the second and third tectophases of the Acadian Orogeny, accompanying bulge move-out. At places along the unconformity, a pyritized lag horizon, no more than 7 mm thick, contains quartzose sand, phosphorite particles, fossil fragments, and conodonts. Conodonts from the lag reflect parts of five Late Devonian (Frasnian–early Famennian) conodont zones (*Pa. crepida–Pa. asymmetricus* zones), encompassing about 10 Ma of subaqueous erosion and particulate accumulation (Ettensohn and others, 1989). Although the lag reflects a long period of sediment starvation and erosion, the accumulation of oxides and other weathering products in upper parts of the Crab Orchard may also reflect periods of subaerial weathering and soil development, although the possibility of interstratal weathering below the pyrite-rich black shales cannot be ruled out (Ettensohn, 1992e).

Overlying the unconformity and lag is 0.8 m (2.6 ft) of unfossiliferous, light greenish gray clay shale with rare black-shale laminae of the Upper Olentangy Shale (Figs. 25, 27). In central Ohio, Middle and Upper Devonian gray shales, separated by an unconformity, are included in the Olentangy Shale; upper parts of the unit are Late Devonian in age, whereas lower parts are Middle Devonian in age (e.g., Woodrow and others, 1988). Here, only the thinning western edge of Upper Devonian parts of the unit is preserved, hence the name "Upper Olentangy." This unit has been traced through subsurface correlations into the Hanover Shale of New York (Wallace and others, 1977). Although the Upper Olentangy is lithologically correlative with the homotaxial Hanover Shale of New York, conodont studies indicate that the two units are not time equivalent (Ettensohn and others, 1989; Ettensohn, 1992e; Fuentes and others, 2002). In fact, the Upper Olentangy is three conodont zones younger *(Pa. rhomboidea* zone in Kentucky) than the Hanover Shale (lower *Pa. triangularis* zone in New York), a time span of about 3 Ma (Woodrow and others, 1988; Gradstein and others, 2004). Inasmuch as the Upper Olentangy and Hanover shales represent deeper-water, dysaerobic, prodelta environments that formed in a migrating foreland basin, the homotaxial and diachronous nature of the units provide a way of estimating bulge and foreland-basin migration. In this particular case, rates of basin-and-bulge migration were on the order of 37 km/million years (23 mi/million years) or about 0.4 m/yr (1.3 ft/yr) (Ettensohn, 1992e). Moreover, by Upper Olentangy–Lower Huron time (Early

Famennian), the foreland basin had filled to overflowing, and Acadian tectonism had migrated far enough cratonward that parts of the Cincinnati Arch subsided, allowing the Appalachian and Illinois Basins to yoke (Fig. 26).

Overlying the Upper Olentangy Shale is 13.4 m (44 ft) of fissile, pyritic black shale with interbeds of greenish gray clay shale at its base (Figs. 25, 27). This part of the Ohio Shale has been called the Lower Huron Shale Member and it has been correlated with the Dunkirk black shales of New York (e.g., Wallace and others, 1977), although like the Olentangy and Hanover shales, they are homotaxial, but diachronous, by about three conodont zones or 3 Ma (Woodrow and others, 1988; Ettensohn and others, 1989; Ettensohn, 1992e; Fuentes and others, 2002; Gradstein and others, 2004). The Lower Huron is a typical transgressive black shale (Fig. 26), being more widespread, more organic-rich, and more radioactive than the overlying, regressive black shales of the Middle Huron Member (Ettensohn, 1992e) (Fig. 25). The Lower Huron forms the large positive deviation at the base of the black-shale sequence (Fig. 25), and because of its high organic content (8 to 23 percent by weight), its shales are typically "pulpy" and weather to form very fissile "paper shales" like those exposed at this stop (Fig. 28). Interpretations based on geochemistry (e.g., Perkins and others, 2008) suggest that the black paper shales of the Lower Huron were deposited in anoxic conditions, whereas overlying parts of the black-shale sequence more likely represent deposition in suboxic conditions.

Figure 28. "Pulpy" paper shales, or transgressive black shales, from the lower part of the Huron Member of the Ohio Shale at stop 6A.

The base of the Lower Huron contains greenish gray shale interbeds that are well known for their trace fossils (Barron and Ettensohn, 1981). The fossils are most prominent in the gray shales, but many of the burrows penetrate from the gray shales into the underlying black shales where the former gray-mud infillings stand out against the black muds. Apparently, organisms from the dysaerobic gray muds were "mining" organic matter from underlying black muds in which they were unable to live because of anoxia. Common trace fossils include *Zoophycos, Planolites, Phycodes, Chrondites, Teichichnus,* and *Rhizocorallium* (Griffith, 1977; Jordan, 1980), which, according to Griffith (1977) and Jordan (1980), reflect the *Nereites* ichnofacies of Seilacher (1967) and represent the basal, slightly shallower parts of an upward-deepening basin sequence. The apparently rapid alternation between gray and black muds may have represented oscillations at the base of the pycnocline caused by relatively rapid changes in subsidence or sea level. The only other common fossils in the Lower Huron include the ubiquitous green algal spores of *Tasmanites* and coalified logs that floated out into the basin, became waterlogged, and sank to the bottom.

Aside from burrowing in the lower part of the Huron and in other gray-shale units in the black-shale sequence, fossils are relatively rare in the unit. Plankton (spores, chitinozoans, algae, and radiolarians), nektoplankton (ostracods, conodonts), epiplankton (brachiopods, pelecypods, crinoids, and graptolites that attached to floating logs), and nekton (fish remains and cephalopods) have been locally reported (e.g., Barron and Ettensohn, 1981). Benthic forms are extremely rare, suggesting that bottom waters were inhospitable, probably due to anoxia. However, when benthic forms do occur, they are invariably *Lingula*- and *Orbiculoidea*-like inarticulate brachiopods.

Black shales are major gas producers in the Appalachian Basin (approximately 3 tcf to date), and the Lower Huron Member is the primary gas-producing unit in the entire basin (Boswell, 1996). In fact, the Big Sandy Field of eastern Kentucky and southwestern West Virginia, encompassing an area of more than 3,500 km^2 (1,260 mi^2), accounts for more than 80 percent of black-shale gas production in the Appalachian Basin, and most of that is from the Lower Huron Shale Member (Boswell, 1996). Although the shale itself has very little permeability, the rock forms the source, reservoir, and seal. The gas is stored as free matrix gas, as gas adsorbed onto clay minerals and organic carbon, and as gas in a system of open natural fractures. As a result, gas production is regionally variable, depending on organic-carbon content, the thickness of organic-rich zones, thermal maturity, and the presence of natural fracture systems. Commonly, it is from the organic-rich, transgressive-shale facies, which we are observing at this stop, and the transition interval into the overlying Middle Huron Member from which most of the production is obtained (Kubic, 1993). Natural fracturing, of course, is critical in generating suitable reservoirs, and as you can see in this exposure the shale is nearly always jointed; unfortunately, little work has been done on characterizing regional joint patterns. In this exposure, there are prominent joint sets running approximately north-south and east-west. However, in natural or man-made exposures, these joint sets make the black shale inherently unstable and subject to rock fall or toppling, where the shale is exposed overlying the weak Silurian and Devonian gray shales.

Because of an abundance of organic matter in the shales, higher concentrations of radioactive and heavy metals are relatively common in the form of organo-metallic-clay complexes. As a consequence, waters emanating from the shales may have higher-than-normal concentrations of these elements and have an "iron" or "sulfurous" taste and smell due to the breakdown of the abundant pyrite in the shales. The presence of these metals and other contaminants from the shale may contribute to a higher incidence of cancer reported from people who live on these shales. Building on the shales can also be problematic. Access by water and air into the shales during building typically causes pyrite breakdown, and sulfate produced from the breakdown may contribute to gypsum formation. Together, these breakdown processes may lead to substantial heaving in the shales that can severely disrupt foundations and destroy overlying structures.

Stop 6B (Optional): I-64 Along the Eastbound Lane, 0.2 Mi East of Stop 6A

At this stop we will view the *Protosalvinia (Foerstia)* zone in the middle part of the Huron Shale Member of the Ohio Shale. Latitude: 38°10′ 0.0″; longitude: –083°35′13.3″.

Near the east end of the exposure, we will climb up above the paper shales of the Lower Huron to a level just above the next bench. Near the base of this bench, the Middle Huron Shale begins (Fig. 25). The Middle Huron is composed of about 25 m (82 ft) of "ribbed," regressive black shales, which contain horizons of interbedded gray shales near the base and top of the unit. The regressive shales contain more elastic constituents, mainly in the form of quartzose laminae, and less organic matter (4 to 7 percent). Moreover, they are easily distinguished from transgressive black shales by the presence of "ribbing." Each "rib" is a few inches thick and alternates with a reentrant of less-resistant shale. The ribs protrude more because they apparently contain more clastic-rich laminae than adjacent shales, and hence, are more resistant. This cyclicity is present throughout most of the Devonian black shales in the western Appalachian Basin, and although the cyclicity has not been definitely explained, it may reflect some type of regional climatic cyclicity in the source area that periodically resulted in greater clastic influx to the basin.

These regressive black shales thicken eastward into the distal parts of major clastic wedges, and the Middle Huron Shale is a black-shale equivalent of the lower part of the greenish gray Chagrin Shale to the north and northeast (Fig. 26). As distal parts of the Chagrin clastic wedge prograded westward, sediments from the east apparently entered deeper parts of an anoxic cratonic basin in which organic matter was preserved; so in many ways, regressive black shales show the same general clastic composition as do their eastern greenish gray shale equivalents, except that they were deposited below the pycnocline where more organic matter could be preserved (Ettensohn and others, 1988a).

What is especially interesting about this exposure is that the basal 3 m (10 ft) of the Middle Huron contains interbedded gray and black shales that coincide with the widespread *Protosalvinia (Foerstia)* zone (Fig. 25). Only about 2.3 m (7.6 ft) of the zone is exposed at this site. *Protosalvinia* is an enigmatic planktic, plant(?) fossil that has been interpreted to represent a possible brown alga (Phillips and others, 1972; Schopf and Schwietering, 1970) or an organism with early land-plant affinities (Schopf, 1978; Romankiw and others, 1988). It commonly has two forms: a bifurcating lobate thallus (Fig. 29) and an elliptical, reproductive structure (Fig. 30). At any one locality, either the thallus or elliptical form predominates, but in some localities the zone is represented only by fragmented debris; at this locality, the elliptical form seems to dominate. This fossil appears to occur at approximately the same relative stratigraphic horizon throughout the Devonian black shales of the east-central United States, making it a potentially important stratigraphic time marker (Hasenmueller and others, 1983). Recent conodont work shows that *Protosalvinia* may occur across parts of three conodont zones *(Pa. trachytera*–lower *Pa. expansa),* and at least locally, the zone may be related to a regional unconformity (Over and others, 2009). Additional information may be found in Ettensohn and others (1989) and Ettensohn (1992d).

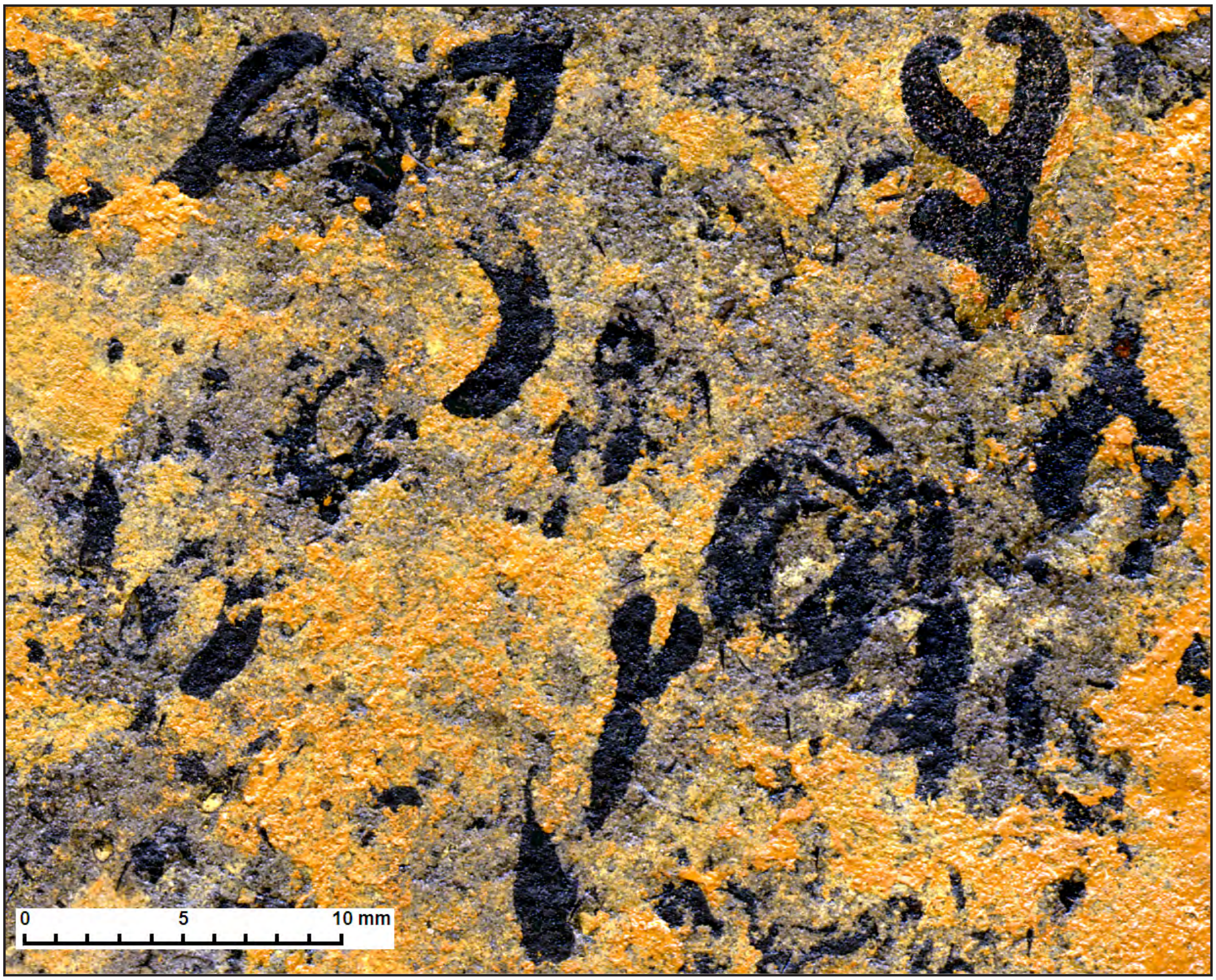


Figure 29. Bifurcating lobate thalli of *Protosalvinia.*

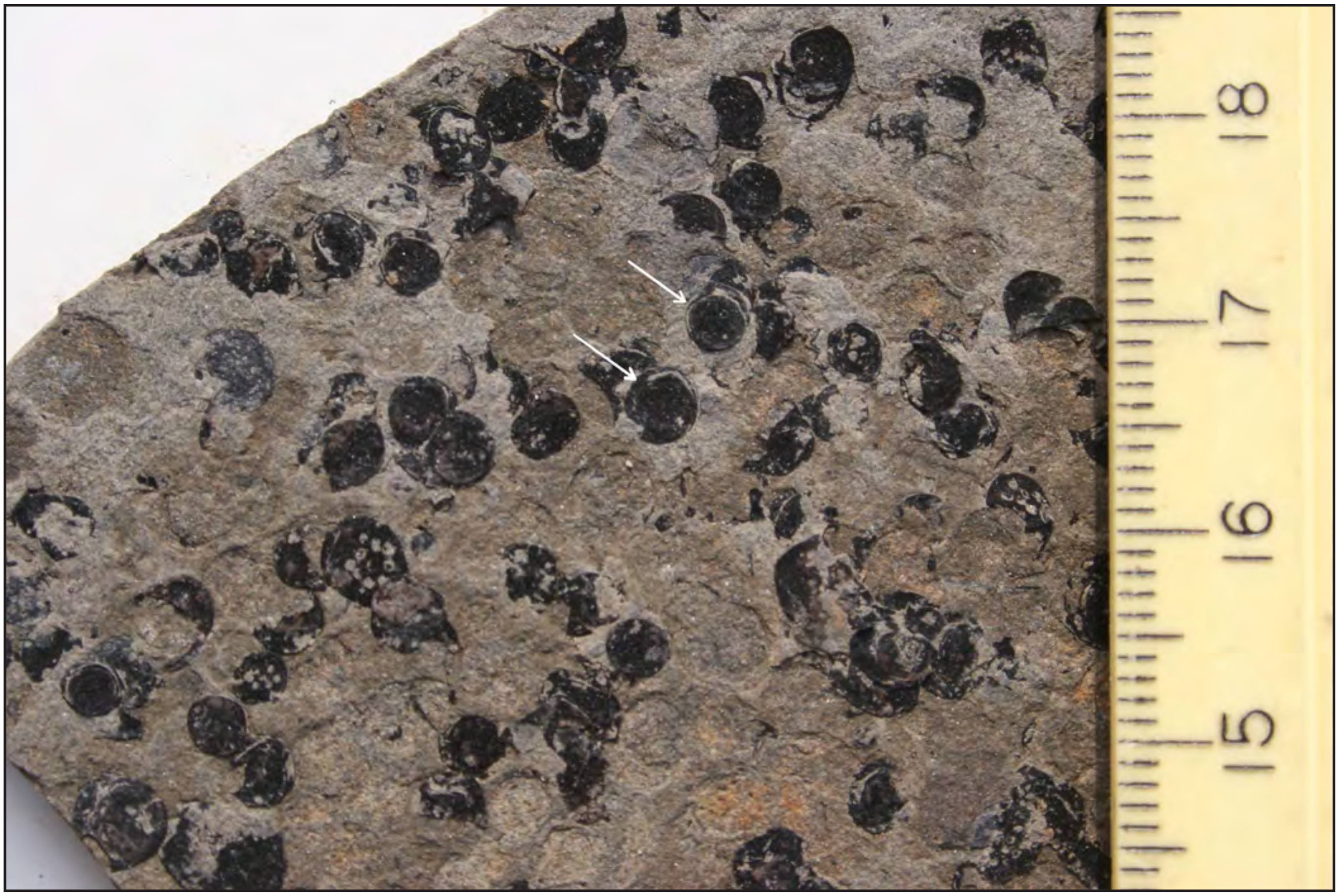


Figure 30. Round to elliptical reproductive bodies of *Protosalvinia.* These are the most common fossils at stop 6B.

STOP 7: Upper Part of the Black-Shale Sequence and Lower Borden Formation

R. Thomas Lierman, Charles E. Mason, and Frank R. Ettensohn

At stop 7 we will examine the well-exposed upper part of the Ohio Shale (Cleveland Shale Member), the Bedford Shale, Sunbury Shale, and lower parts of the Borden Formation, including the Henley Bed, Farmers Member, and Nancy Member (Figs. 31–32).

The base of this section begins with the upper part of the Upper Devonian Cleveland Shale Member of the Ohio Shale and continues up section through the Bedford Shale (uppermost Devonian). The Bedford is in turn unconformably overlain by the Mississippian Sunbury Shale, which is followed in succession by the Lower Mississippian Farmers and Nancy Members of the Borden Formation. The Borden sequence here represents the lower portion of a prograding delta sequence. A stratigraphic section at this stop is shown in Figure 32.

Unit 1, Cleveland Shale Member (Ohio Shale)

Unit 1 consists of 4.25 m (14 ft) of the Cleveland Shale Member of the Ohio Shale formation (Fig. 32). This shale is a brownish black to black, fissile, organic-rich, silty shale. It is pyritic with an occasional siderite or phosphate nodule scattered throughout. In weathering it tends to take on a dusky yellowish brown to yellowish orange color. Both body fossils and trace fossils are rare within the unit, though upon close inspection, one can occasionally come across a few linguloid brachiopods, conodonts, and fish remains (teeth, scales, bones). In viewing the shale from the side, one can also see that the weathered surface of the outcrop has a ribbed appearance (Fig. 33). This

Figure 31. Photo showing the upper part of the stratigraphic section at stop 7 (see Figure 32). The upper part of the Ohio Shale can be seen in the distant cut in the background, and the Cave Run Lake fauna is present on uppermost bench in the photo.

ribbing consists of smoother-faced promontories and more splintery-weathering recessed intervals, which average about 5 cm thick. The recessed intervals are likewise carbonaceous shales but with slightly lower organic content than the smoother-faced promontories. The lower organic content apparently allows these intervals to weather more rapidly.

The environment of deposition is best described as a deep-water, anoxic or anaerobic, basin-floor environment. Anaerobic conditions are found where oxygen levels are less than 0.1 ml of O_2 per liter of water. A number of considerations lead us to that conclusion. First, the black color of the shale is due to the presence of finely disseminated organic matter within the shale. This typically occurs under stagnant or reducing conditions; that is, conditions where there is a lack of free oxygen in the water or sediment and where anaerobic bacteria are present. A second line of supporting evidence comes from the observation that there is a near absence of any body fossils or trace fossils within these shales. The few fossils that are found tend to represent organisms that maintain a nektic or planktic lifestyle. This lifestyle is certainly reflected in the conodonts and fish remains, and the few inarticulate brachiopods we find were probably epiplanktic, attached to floating logs or floating vegetation such as *Sargassum*-like seaweed; benthic fossils are completely absent from these shales. A final piece of evidence within these organic-rich shales is the presence of pyrite (FeS_2) along with some nodules of siderite ($FeCO_3$). Both of these mineral phases tend to form under Eh conditions that are reducing or (–) negative (Garrets and Christ, 1965, p. 224).

These anaerobic conditions more than likely formed in a marine basin in which the vertical water column was density stratified. Density stratification is generally related to differences in water temperature or salinity. In this case, warmer or less

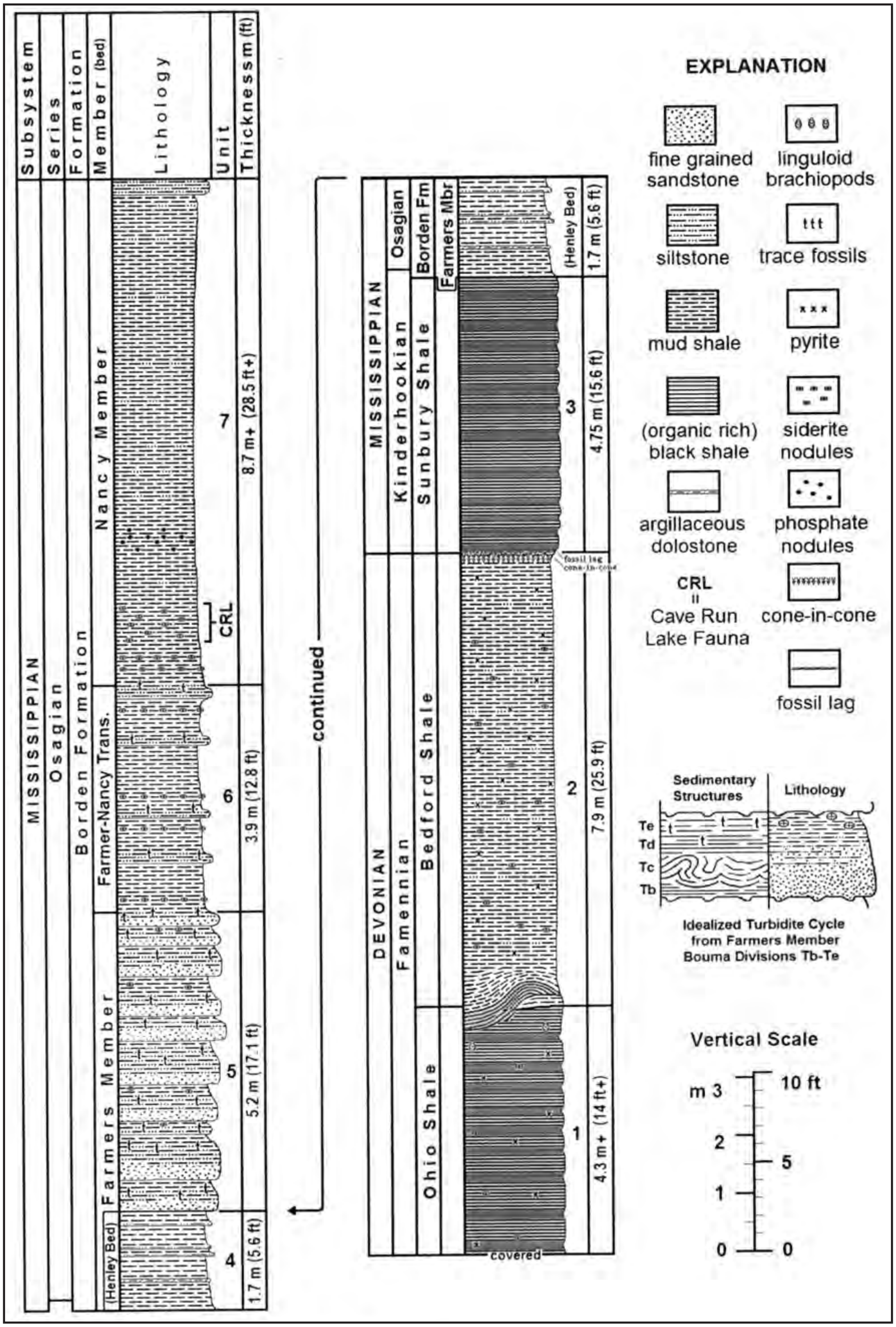


Figure 32. Schematic stratigraphic section for stop 7 (see Figure 31).

Figure 33. “Ribbed” black shales from Cleveland Member of the Ohio Shale at stop 7.

saline surface water would sit atop colder or more saline bottom waters. Such differences in water temperature and/or salinity would in turn generate a distinct pycnocline or zone of rapid density change between surface and bottom waters. If this difference was pronounced enough, it could effectively cut off circulation between bottom and surface waters. The bottom waters would in turn become oxygen deprived, provided that they were below the euphotic zone and not allowed to circulate freely with oxygen-rich surface waters (Fig. 34). The organic matter in this instance could have come from either a planktic source flourishing in the upper parts of the water column, or as detrital plant remains derived from the terrestrial land plants, because by Middle and Late Devonian time, vascular land plants had established a firm foothold on the land surface. This colonization of the land included true arboreal species (forests), tree-size plants with woody stems, complex vascular systems, leaves, and even the first appearance of seeds. The sudden appearance of so many plant groups and growth forms has been called the “Devonian Explosion” (Algeo and others, 1995).

As stated previously, the ribbed character of these shales (Fig. 33) is probably related to variations in the organic content of the shale. This cyclic increase and decrease in organic carbon could have several explanations. One explanation might be a periodic increase and decrease in the abundance and productivity of phytoplankton in the water column. Periods of high productivity would result in an increase in the relative amount of organic matter in the shale; periods of lower productivity would provide the sediment with less organic mat-

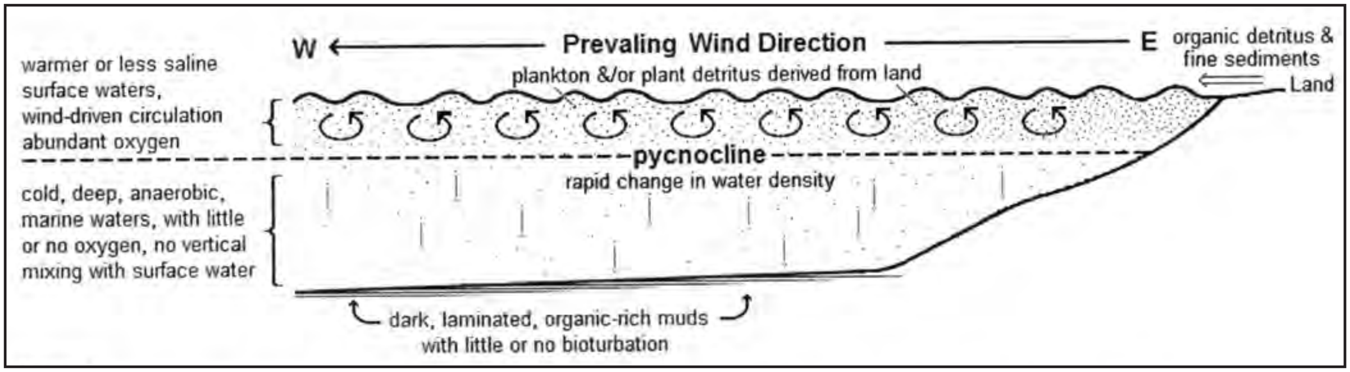


Figure 34. Cartoon illustrating possible means (stratified water column) for the isolation and preservation of organic matter in the Devonian black shales.

ter. A second explanation could involve variations in the amount of terrigenous sediment reaching the depositional basin. In this case, an increase in the percentage of fine-grained mud would result in a decrease in the relative amount of organic matter incorporated into the sediment. Conversely, a decrease in the percentage of terrigenous mud would translate to an increase in the relative amount of organic matter in the shale. A third possibility might relate to cyclic variations in the amount of detrital organic matter derived from terrestrial plants living on the land surface. Streams draining recently colonized Devonian lands would have carried an increased amount of organic matter, primarily in the form of plant detritus. This organic detritus would have eventually been transported into deeper parts of the basin and become incorporated into the fine-grained sediments of the Ohio Shale. Conceivably, variations in the amount of terrestrial plant matter reaching the basin bottom could have caused differences in the amount of organic matter preserved in the shales.

Devonian, marine black-shale deposits are quite notable for their widespread occurrence across the inland seas of North America and Eurasia (Fig. 35). Algeo and others (1995, 1998) suggested that these deposits were the result of the huge influx of organic matter and nutrients from an increasingly vegetated landscape. In addition to causing eutrophication in these broad epicontinental seas, terrestrial plants may have also contributed to changes in the speed and pattern of soil formation, which led to accelerated weathering of silicate minerals. This chemical weathering process, called hydrolysis, is a reaction involving water, H^+ or OH ions, and silicate minerals. The byproducts of hydrolysis include various clay minerals (e.g., kaolinite, illite), orthosilicic acid (H_4SiO_4), along with the generation of bicarbonate (HCO_2^-). This reaction can effectively remove CO_2 from the atmosphere and ultimately tie it up in the carbonate-silicate geochemical cycle. These weathered bicarbonates enter rivers and are ultimately transported to the oceans, where they precipitate as various carbonate minerals and are eventually buried in marine sediments.

The burial of extensive quantities of organic carbon and inorganic bicarbonate could have eventually led to reduced atmospheric CO_2 levels. Algeo and others (1995, 1998) suggested that the loss of this greenhouse gas may have contributed to a major global cooling event during Late Devonian time. The very "greening" of the continents by terrestrial land plants could have acted as a carbon-dioxide sink, and atmospheric levels of this greenhouse gas may have dropped substantially. This, in turn, would have cooled the climate and possibly resulted in an intense episode of glaciation near the end of Devonian time. This is most evident in parts of Gondwanaland from South America (Brazil, Bolivia, and Peru) to parts of central Africa (Central African Republic and Niger) (Crowell, 1999). Possible evidence for this period of glaciation may also be found much closer to home, as we will see later on in this trip. More information and interpretations regarding the Cleveland and related black shales can be found in papers by Ettensohn and Barron (1981), Ettensohn and others (1988a), and Perkins and others (2008).

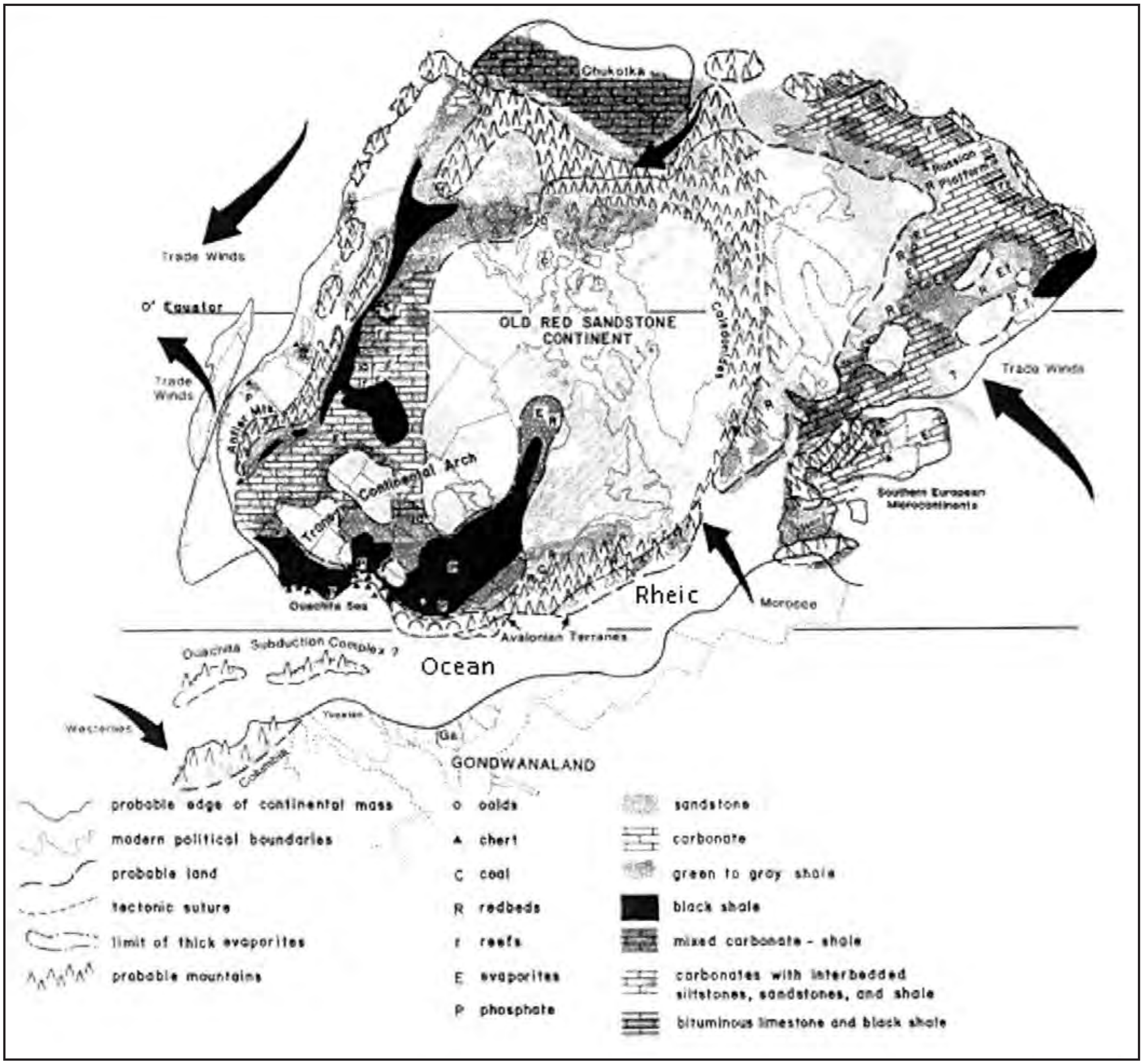


Figure 35. Late Devonian paleogeographic reconstruction of Laurussia (Old Red Sandstone Continent), showing the location of black-shale seas (adapted from Ettensohn and Barron, 1981).

Unit 2 (Bedford Shale)

The Bedford Shale consists of 7.9 m (25.9 ft) of medium- to olive-gray mud shale that is poorly fissile and noncalcareous. Scattered within this shale are very thin, discontinuous beds and lenses of argillaceous siltstone along with siderite nodules and irregularly shaped masses of pyrite. Also disseminated throughout the shale are small crystals of pyrite, which occur as cubes and octahedra. Some of the siltstone beds show a faint hint of rippling along their upper surfaces. The unit as a whole appears highly bioturbated, though individual trace fossils are difficult to discern or identify; horizontal burrows filled with pyrite are locally common. Body fossils (chonetid brachiopods and gastropods) can be found at this location, though they are sparse. They are most commonly found in pyrite nodules. The lower contact of the Bedford with the underlying Ohio Shale is in places marked by intercalated layers of gray shale and black, fissile shale, which in places appear to be bioturbated.

At the top of the Bedford, a well-developed cone-in-cone limestone layer, approximately 5.5 cm (2.1 in.) thick, is locally present (Fig. 36). Cone-in-cone is a secondary sedimentary structure that has the appearance of a series of cones packed one inside the other. Close inspection reveals that the apices of the cones are mostly directed upward, and there seems to be a concentration of clays and

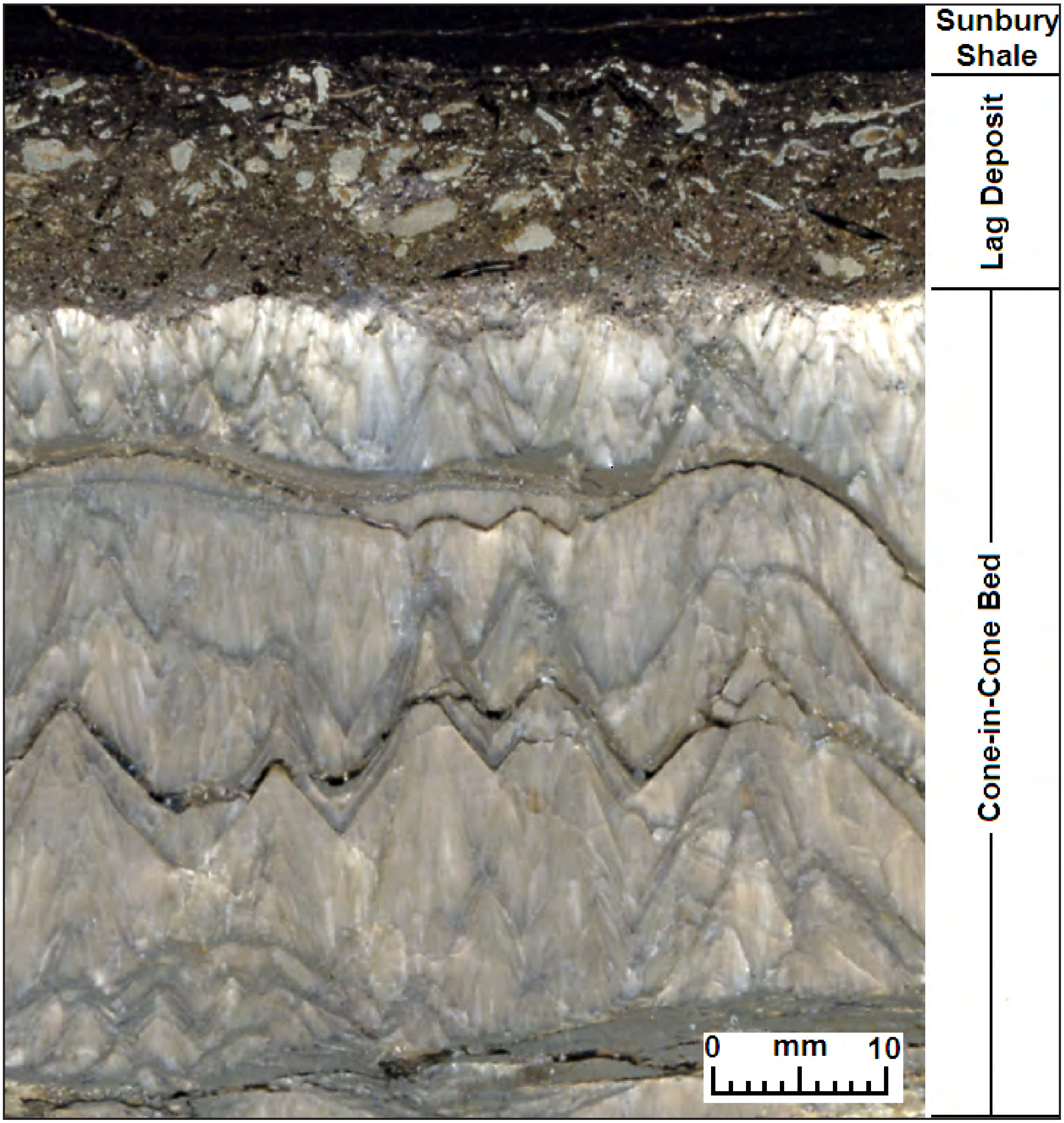


Figure 36. Cone-in-cone limestone bed in the uppermost Bedford Shale immediately below the Sunbury lag zone. The contact between the Sunbury and Bedford Shales is a regional unconformity.

organic matter along the margins of many cones. Petrographic examination of the cones reveals that they are a formed of fibrous calcite, with the fibers tending to be parallel to the sides of individual cones. Stylolites are also associated with these cone-in-cone structures, suggesting that both the cone-in-cones and the stylolites are formed by the same process of pressure solutioning.

At this locality, the lower contact of the Bedford with the Ohio Shale is broken by what appears to be several low-angle thrust faults (Figs. 37–38). Thrusts like this are highly unusual for the area, and this is the first time such features have been observed. These thrusts strike at approximately N74°E and dip at an angle of about 5° to the north. We are fairly confident that these are small thrust faults because we can see the displacement of the black shale as it has been thrust up and over the gray Bedford Shale, which is in turn thrust over the top of the black shale. A close-up view of a thrust is shown in Figure 37, and an outcrop-scale view of several thrusts is shown in Figure 38. Slickensides associated with surfaces in the black shale, local intrusion of plastic gray shales into the black, as well as several unusual joints that extend out from the thrust at angles of 47° to 68° from the thrusts provide additional evidence for the interpretation. These structures may reflect a subsurface response to growth faulting along nearby basement structures at depth (e.g., Drahovzal and Noger, 1995) or they may reflect mass movement from another dimension. In either case, the gray shale tends to behave more plastically, whereas the black shales tend to act more competently.

Deposition of the Bedford probably occurred in a dysaerobic setting. Dysaerobic environmental conditions exist where dissolved oxygen levels in the water or sediments are between 0.1 and 1.0 mil of O_2 per liter of water. The evidence for this setting is first of all the color of the shale. The gray-green color of these shales is due to the presence of greenish phyllosilicates (e.g., illite, chlorite). The iron content in these minerals is key as the Fe is in a +2 oxidation state (ferrous iron). In this situation, there was apparently enough oxygen in the water column to thoroughly oxidize any organic matter, but not enough to precipitate iron oxides, such as hematite (iron is in a +3 oxidation state), which would have imparted a red or maroon color to the shales. Second, a lack of an abundance of free oxygen in these sediments is also indicated by the presence of pyrite and siderite nodules in these shales, both of which are stable under reducing conditions (Garrels and Christ, 1965). Third, only a small, depauperate fauna is present, one that apparently includes only thin-shelled brachiopods and mollusks, and such faunas are typically common in dysaerobic fossil assemblages (Kammer, 1985; Pashin and Ettensohn, 1992b). Even so, abundant bioturbation is present in these shales, which is very different from the black, organic-rich muds of the Ohio Shale.

Such dysaerobic sediments were more than likely deposited where the pycnocline is relatively broad and intersected the sea bottom (Ettensohn and Elam, 1985). Pashin and Ettensohn (1992a, 1995) envisioned the Bedford Member as part of a delta complex that included the black shales of the Cleveland Member, Bedford Shale, and Berea Sandstone. In this model, the Cleveland Member formed under anaerobic to dysaerobic, basin-floor conditions, whereas the Bedford Shale is thought to represent the slow accumulation of muddy sediments at distal margins of a mud-rich turbiditic slope under more dysaerobic conditions. The related Berea Sandstone, which is not present in this area, was interpreted to represent a series of storm-dominated shelf deposits in northeastern Kentucky (Pashin and Ettensohn, 1987, 1992b, 1995) (Fig. 14). However, recent ideas about the possibility of Late Devonian, Acadian/Neoacadian, alpine glaciation suggest that Bedford shales from eastern sources may reflect distal lowstand deposits related to alpine glaciation in Acadian/Neoacadian highland source areas to the east. More information and interpretations relative to the Bedford Shale can be found in papers by Pepper and others (1954), Chaplin and Mason, 1979, Ettensohn and Elam (1985), and Pashin and Ettensohn (1987, 1992a, b, 1995).

Unit 3 (Sunbury Shale)

The Sunbury Shale at this locality consists of 4.75 m (15.6 ft) of dark gray to black, highly carbonaceous, fissile shale (Fig. 32). The unit contains small pyritic nodules as well as small, scattered py-

Figure 37. Close-up view of Cleveland black shales thrust over Bedford gray shales in an erosion gully at stop 7.

Figure 38. Outcrop view of repeated Cleveland and Bedford section at stop 7, showing the most likely location of thrust surfaces involved.

rite crystals. Overall, the lithology of the Sunbury Shale is very similar to that of the Ohio Shale below, and this includes a similar ribbed appearance when viewed from the side. The uppermost part of the shale tends to be bioturbated with burrows infilled by greenish gray shale similar to the overlying Henley Bed. Fossils are quite sparse within the unit but may include rare linguloid or orbiculoid brachiopods, as well as conodonts.

The lower contact of the Sunbury with the underlying Bedford Shale is quite sharp and is characterized by a 1.5-cm-thick "lag" deposit (Fig. 36). This lag lies immediately above the cone-in-cone layer previously mentioned in the Bedford Shale (Fig. 36) and contains a variety of both pyritized and phosphatic fossil remains, including inarticulate linguloid and orbiculoid brachiopods, a variety of conodonts, as well as phosphatic fish debris including teeth, scales, spines, and broken dermal plates. It also contains a concentration of what appears to be reworked pyritized burrows. The fragments in this lag also commonly exhibit reverse grading (Fig. 36). This basal layer or zone is apparently recognized throughout the entire outcrop area of the Sunbury Shale (Pepper and others, 1954), and in this area, that basal lag separates Devonian and Mississippian rocks and defines a major regional unconformity (Ettensohn, 1994).

Considering the similarities between the Sunbury Shale and the underlying Cleveland Shale, it is likely that the depositional conditions of this unit were very similar to those of the Cleveland for all the same reasons. We therefore interpret the Sunbury Shale to represent the slow accumulation of fine-grained muds in the very deepest portion of a basin-floor environment in anaerobic conditions. The only difference is that the Sunbury was deposited during Early Mississippian (early Tournaisian; early Kinderhookian) time. The Sunbury represents the most widespread, most organic-rich, and deepest of the black-shale basinal environments present in the Devonian-Mississippian black-shale sequence. Unlike the underlying Devonian black-shale units, which migrated westward in time in deformational loading (Fig. 26), the Sunbury represents a trangression and subsidence event that moved eastward in time (Fig. 26), apparently reflecting inception of a new, more proximal Neoacadian convergence event to the east (Fig. 39B). Additional information and interpretations regarding the Sunbury can be found in papers by Chaplin and Mason (1979, 1985), Ettensohn and Elam (1985), Mason and Lierman (1985), Ettensohn and others (1988a), and Lierman and others (1992).

Unit 4, Henley Bed, Farmers Member (Borden Formation)

The Henley Bed is the basalmost unit of the Borden Formation (Figs. 31–32). Here it consists of 1.7 m (5.6 ft) of greenish gray to grayish green mud shale that is poorly fissile and noncalcareous. The unit as a whole appears to be highly bioturbated, though individual trace fossils are difficult to discern. In addition to the shales, the unit contains three thin (approximately 5 cm thick) beds of argillaceous siltstone along with one very thin bed of argillaceous dolostone. Body fossils are rare at this locality; however, microfossils are abundant and diverse and include conodonts, spores, and arenaceous foraminifera.

The lower 10 cm of the Henley Bed at this locality is Early Mississippian in age (ear-

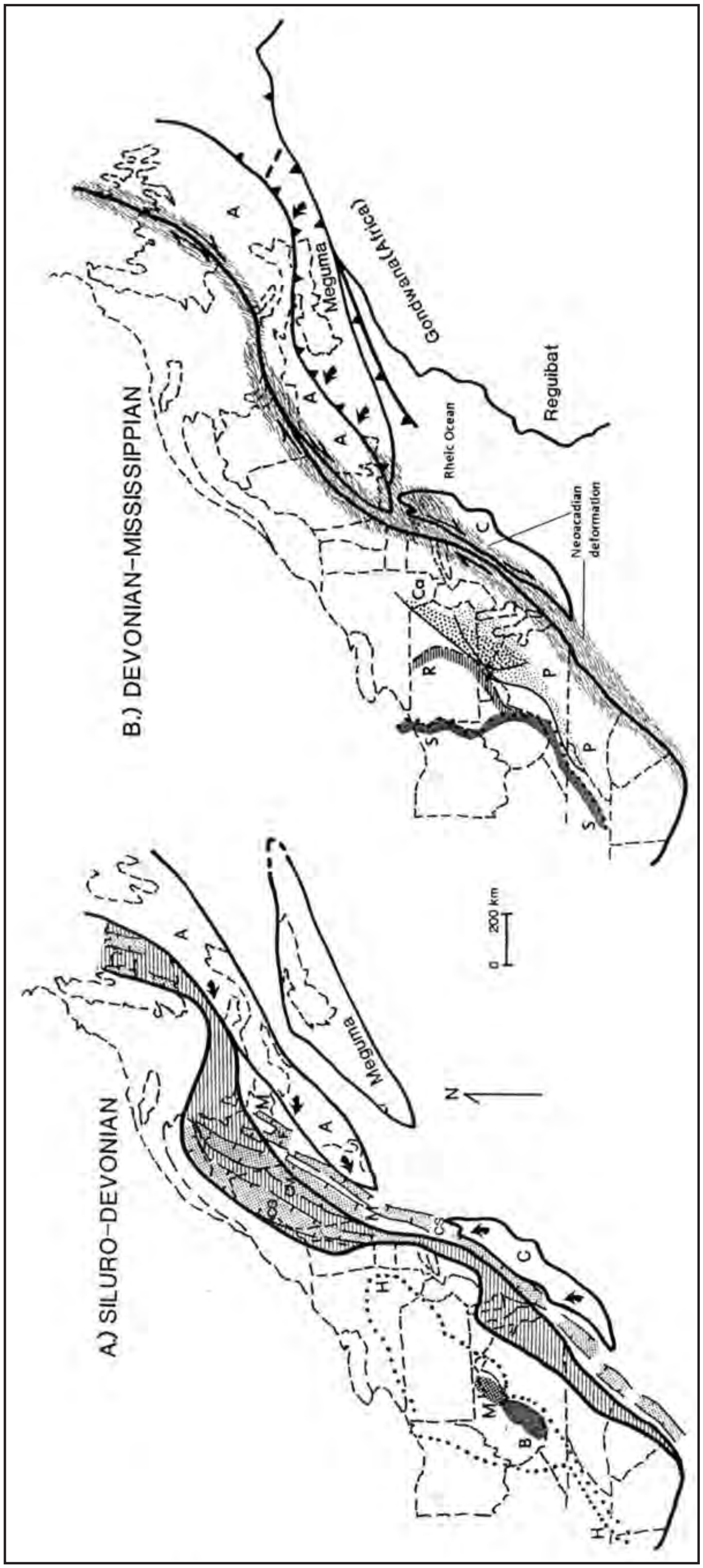


Figure 39. Neoacadian tectonic situation that generated the Sunbury Shale (stop 7), Borden delta, and formed the mountain belt high enough to support alpine glaciation. A. Likely pre-Acadian deposition of microplates relative to Laurussia. B. Likely deposition of microplates and Gondwana during Neoacadian Orogeny. Note the eastward limit of the deep-marine Sunbury Shale (see Figure 26) and its marginal-marine equivalent, the Riddlesburg Shale (adapted from Ettensohn, 2008).

ly Kinderhookian) and corresponds to the lower *Siphonodella crenulata* Zone of Sandberg and others (1978), based on the presence of the conodont species *Siphonodella crenulata* Branson and Mehl. Above this and up to the thin dolomite bed, about 50 cm above the base, no conodonts have been recovered at this locality. However, at other sites across Kentucky and Ohio, the upper *Siphonodella crenulata* Zone has been identified. At approximately 30 cm below the first siltstone bed at this locality, the conodont species *Polygnathus communis carinus* and *Pseudopolygnathus multistriatus* have been found. Below this interval and extending down to the dolomite bed previously mentioned, another interval lacking conodonts occurs. *P. communis carinus* and *P. multistriatus* have also been recorded from a 1-m interval at the base of the Nancy Member at this location, and these two forms indicate an Early Osagean age (equivalent to the Fern Glen or early Burlington formations) for this part of the interval (Work and Mason, 2005). For this reason the contact between the Kinderhookian and Osagean is placed at the thin dolostone layer about 50 cm above the base of the Henley Bed.

The Henley Bed is thought to represent the slow accumulation of fine-grained sediments in deep prodelta environments at the foot of the prograding Borden delta. Henley Bed sediments mark the inception of basin infilling, following anaerobic, Sunbury, basinal sedimentation (Fig. 26). The environment was dysaerobic and dominated by hemipelagic muds, which were periodically interrupted by an influx of silt and very fine sand from occasional turbidity currents, reflected by the thin siltstone beds found in the Henley. The shales and mudstones of the Henley Bed, as well as those of the overlying Farmers Member, were deposited during relatively long periods of time by slow accumulation and probably represent the indigenous sediments that would have normally accumulated in this relatively deep-water, prodelta environment. The siltstones and sandstones in the Henley and overlying Farmers Member represent brief intervals of rapid sedimentation through the intrusion of turbidity currents or density currents as they periodically disrupted the generally quiet, deep-water setting. Additional information and interpretations about the Henley Bed can be found in papers by Chaplin and Mason (1979), Chaplin (1980, 1982, 1985), and Mason and Lierman (1985, 1992).

Unit 5, Farmers Member (Borden Formation)

The Farmers Member is the lowermost member of the Borden Formation (Figs. 31–32). At this locality the Farmers consists of 5.2 m (17.1 ft) of interbedded sandstones/siltstones and shales. The unit contains tabular-bedded, very fine-grained sandstones to coarse-grained siltstones that alternate with mud shales. The coarser-grained sandstone/siltstone beds range from 22 to 65 cm (9 to 26 in.) thick at this location. These beds are light brownish gray to yellowish brown and are composed principally of quartz, rock fragments, and mica. The matrix is chiefly clay, siderite, and microcrystalline quartz. Individual beds tend to be size graded, with the lower portion of beds consisting of very fine-grained sand. This in turn grades upward into coarse- to medium-grained silts and eventually into silty muds. The finer-grained shales are greenish gray mud shales to silty shales that occur as partings and interbeds between the coarser-grained layers. Grayish red siderite nodules and lenses also occur throughout the unit and are especially common in the shale interbeds. The shales range from 5 to 32 cm (2 to 12.6 in.) thick at this locality.

Sedimentary structures in the Farmers include internal structures, sole marks, and trace fossils. The lower surfaces of the sandstone/siltstone beds exhibit very abrupt contacts with the underlying shale and have an abundance of sole marks. These sole marks include tools marks, such as groove, brush, prod, and bounce casts. The most abundant tool marks are groove casts. Measurements of paleocurrent directions from the sole marks in the Farmers show a trend from east to west. This paleocurrent direction is consistent with the downslope movement of material from east to west off the Borden delta front (Moore and Clarke, 1970). The upper surfaces of the coarse-grained beds tend to grade into the overlying shales and show extensive evidence of bioturbation.

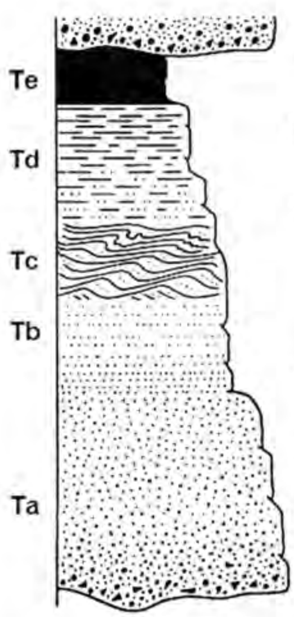
Internal sedimentary structures within the coarser-grained beds include parallel laminae, current ripple laminae, and convolute laminae. The coarser-grained beds commonly exhibit a lower interval of parallel laminae overlain by current-ripple laminae or convolute laminae, followed by an interval of parallel laminae, which is in turn overlain by shales (see insert in the stratigraphic section for this stop, Figure 32). This sequence more than likely corresponds to a truncated Tb–Te interval of Bouma's classic turbidite sequence (Bouma, 1962). The graded interval Ta at the base of Bouma's sequence (Fig. 40) is absent here, which is probably due to the overall fine-grained nature of the rocks.

Body fossils can also be found throughout the Farmers but tend to be locally restricted to certain beds or siderite nodules. Megafossils that do occur are mainly found as molds, and include productid and spiriferid brachiopods, fenestrate bryozoans, crinoid columnals, gastropods, cephalopods, conularids, trilobites, hexactinellid sponges, and bivalves. The associated fauna seem to have been a stable shelf, benthic fauna that had been transported into this deeper-water setting, although some may have been nektic or nektobenthic.

Trace fossils are very common features in these rocks and are most generally found along the upper and lower surfaces of the coarser-grained beds and include the ichnogerera *Zoophycos, Lophoctenium, Sclarituba, Teichichnus, Palaeodictyon,* and *Chondrites*. Escape burrows or fugichnia that frequently extend from the base to the top of many of these beds are also common. These trace fossils are representatives of both the *Zoophycos* and *Nereites* ichnofacies and mainly reflect grazing and feeding traces.

A turbidity current or density current is the downslope movement of dense, sediment-laden water, created when sand and mud along a shelf or slope are dislodged and thrown into suspension. The resulting deposit is generally found at the base of the slope and is called a turbidite. Turbidity flows are recognized by their graded bedding and by a unique sequence of sedimentary features known as a Bouma sequence, which includes a characteristic sequence of sedimentary structures deposited by the turbidity current; these were first described by Arnold Bouma in 1962. At the base of the sequence is a massive, coarse-grained, graded bed (Ta) deposited as the turbidity current passes. Overlying this is parallel-laminated sand (Tb), a rippled sand/silt (Tc), and finally, a parallel-laminated layer of mud or silt (Td), all reflecting the waning carrying capacity of the current (Fig. 40).

The Farmers Member is thought to be a series of turbidite deposits that accumulated at the outer edge of a prograding Borden delta (Fig. 41).



Te Pelitic interval: pelagic or hemipelagic muds or shale, settling out of suspension.

Td Interval of parallel lamination: indistinct parallel laminations very fine sand to silt. **Lower flow regime**

Tc Interval of current-ripple laminations or convolute laminations: small current ripples, small-scale ripple laminations or climbing ripples, or sometimes wavy or convolute laminations. **Lower flow regime**

Tb Interval of parallel lamination: coarse horizontal laminations, grading may be present. **Upper flow regime**

Ta Graded interval: massive or graded interval; represents coarsest material to settle out, may be indistinct or missing, few distinct sedimentary structures. **Upper flow regime**

Figure 40. Five-part division of a classic Bouma cycle, showing an ideal sequence of sedimentary structures found in a turbidite bed (adapted from Walker, 1979).

Evidence for this interpretation is quite substantial. First, the sharp or erosional contacts of the lower surfaces of individual beds within the Farmers suggest rapidly moving currents capable of scouring away the sediments over which it flowed. Second, the lower surfaces of many of these beds have an abundance of tool marks, again suggesting rapidly moving currents. These tool marks tend to be oriented in an east-to-west direction, suggesting that the paleocurrents responsible for forming these structures were moving down a westward-facing paleoslope. Third, the fact that the grain size of any individual bed decreases from the base to the top of each bed suggests that the current responsible for depositing these sediments was one whose flow velocity was decreasing as it was being deposited. Fourth, the abundance of escape burrows in many of these beds suggests that these sedimentary layers were deposited fairly rapidly, potentially trapping any organisms living in the area and forcing them to quickly burrow up through the offending sediment layer. Another observation is that most of the trace fossils are found along upper surfaces of these beds, again suggesting that these sediment layers were deposited rapidly, and that once deposited, stable conditions returned once again. Stable conditions would have allowed burrowing organisms to rework the upper surfaces of the sediment layers. Finally, the close match between a Bouma sequence and the sequence of sedimentary structures found in any individual bed in the Farmers is striking (Fig. 32). Taken together, this evidence suggests that the Farmers Member was deposited as a series of distal turbidites along the basal slope of the Borden delta complex (Fig. 41). The only part of any Farmers Bouma sequence that is missing is the Ta interval, probably reflecting the fact that these sediments started out as fine-grained sands

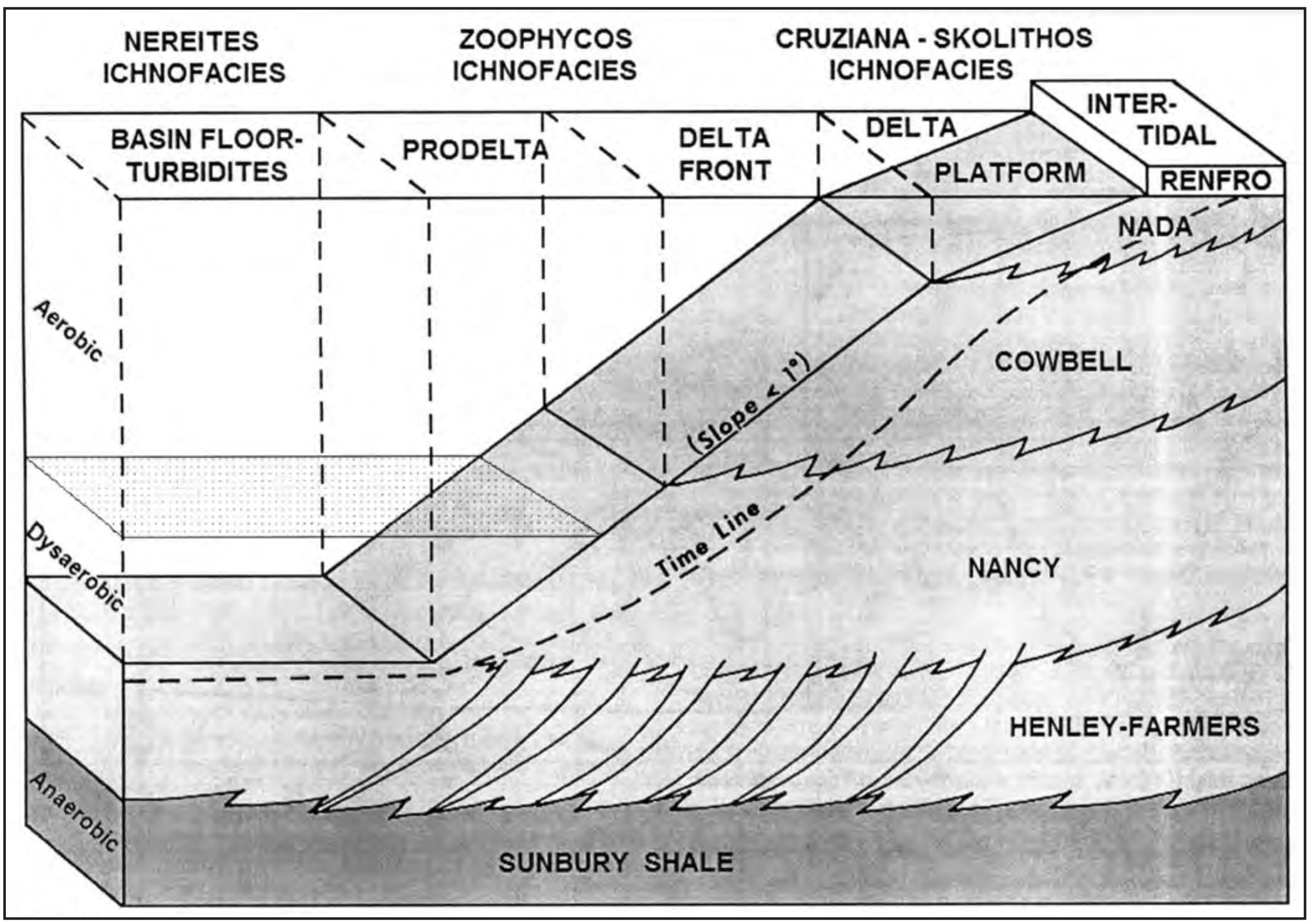


Figure 41. Block diagram showing interpreted depositional environments of the typical lithofacies of the Sunbury and Borden formations in the Morehead area, east-central Kentucky (adapted from Kepferle, 1977).

and silts; the flows simply had no coarser-grained sediments with which to work.

Moore and Clarke (1970) were the first to suggest that the Farmers Member was of turbiditic origin because it exhibited many of the features found in typical Bouma sequences. However, it should be noted that the Farmers was deposited in a cratonic setting rather than the more classic deep geosynclinal setting. The Farmers sequence clearly represents turbidite sedimentation at the base of a westward-facing paleoslope that extended into deep, quiet-water environments of the Appalachian Basin in eastern and east-central Kentucky. Paleocurrent analysis suggests that the source was to the east in the Appalachian highlands. This paleoslope was the leading edge of the Borden delta complex that built out from these highlands. Kepferle (1977) also concluded that the Farmers Member in eastern Kentucky and a similar unit in east-central Kentucky, the Kenwood Siltstone, were both deposited as turbidite sequences. He interpreted these units as fanning out from two depositional centers along the front edge of this prograding delta platform. According to Kepferle (1977), the front edge of this delta marked the very outer edge of the Catskill-Pocono clastic wedge that first began to build westward in Late Devonian time (Fig. 42). More information and interpretations regarding the Farmers Member can be found in papers by Moore and Clarke (1970), Ettensohn (1979), Chaplin (1980, 1982, 1985), Mason and Lerman (1985, 1992), and Chaplin and Mason (1992).

Unit 5, Farmers/Nancy Transition (Borden Formation)

The Farmers/Nancy Transition zone, an informal unit designated by Chaplin (1980), consists of 3.9 m (12.8 ft) of interbedded shale and siltstone (Figs. 31–32). The shale is a greenish gray, silty shale to mud shale that is poorly fissile, noncalcareous, and extensively bioturbated. The shale intervals range from 55 to nearly 110 cm thick (21.5 to 43.7 in.). Grayish red siderite nodules and lenses also occur throughout the unit and tend to be concentrated in distinct layers. Some of the nodules are fossiliferous and may contain brachiopods, fenestrate bryozoans, pelecypods, gastropods, or conularids. Located within these shales are four siltstone beds whose thicknesses vary from 13 to 27 cm (5 to 10.5 in.). The siltstones are again turbiditic in origin and have characteristics identical to the beds in the underlying Farmers Member. Along the lower surfaces of each bed is an abundance of sole marks, including groove casts, load casts, and various tool marks. Trace fossils are also very common along the upper surface of the turbiditic siltstones, with *Zoophycos* being the most abundant. Body fossils can be collected from the siderite nodules found in the interval. This transition interval is an informal designation, as it simply separates the thicker-bedded sands of the Farmers Member below from the shales of the Nancy Member above.

Unit 6, Nancy Member (Borden Formation)

At this locality, the Nancy Member, is incompletely exposed (Figs. 31–32) and consists of 8.7 m (28.5 ft) of greenish gray to grayish green mud shale to silty shale that is poorly fissile, noncalcareous, and highly bioturbated. The shale contains an abundance of siderite nodules, both scattered within the unit and concentrated in distinct beds or layers. These nodules can be highly fossiliferous and heavily mineralized. There is also a zone of phosphate nodules located some 2.5 m above the contact with the unit below. Body fossils include a variety of open-marine forms, including brachiopods, gastropods, cephalopods, fenestrate bryozoans. crinoid debris, solitary rugose corals, and occasionally conularids.

Intrepretation

The Nancy Member is interpreted to be a prodelta deposit (Fig. 41), formed as the Borden delta complex prograded across east-central Kentucky. The upper parts of this unit were mainly deposited in an aerobic environmental setting, whereas lower portions of the Nancy, as well as the Farmers Member, were probably deposited under somewhat dysaerobic conditions that developed in the basinal seas in which these units were deposited, reflecting a stratified water column in which bottom waters were anaerobic, lower waters were dysaerobic, and middle and upper waters were aerobic.

Evidence for dysaerobic conditions during deposition of the lower Nancy and Farmers Members includes (1) The overall gray-green color of

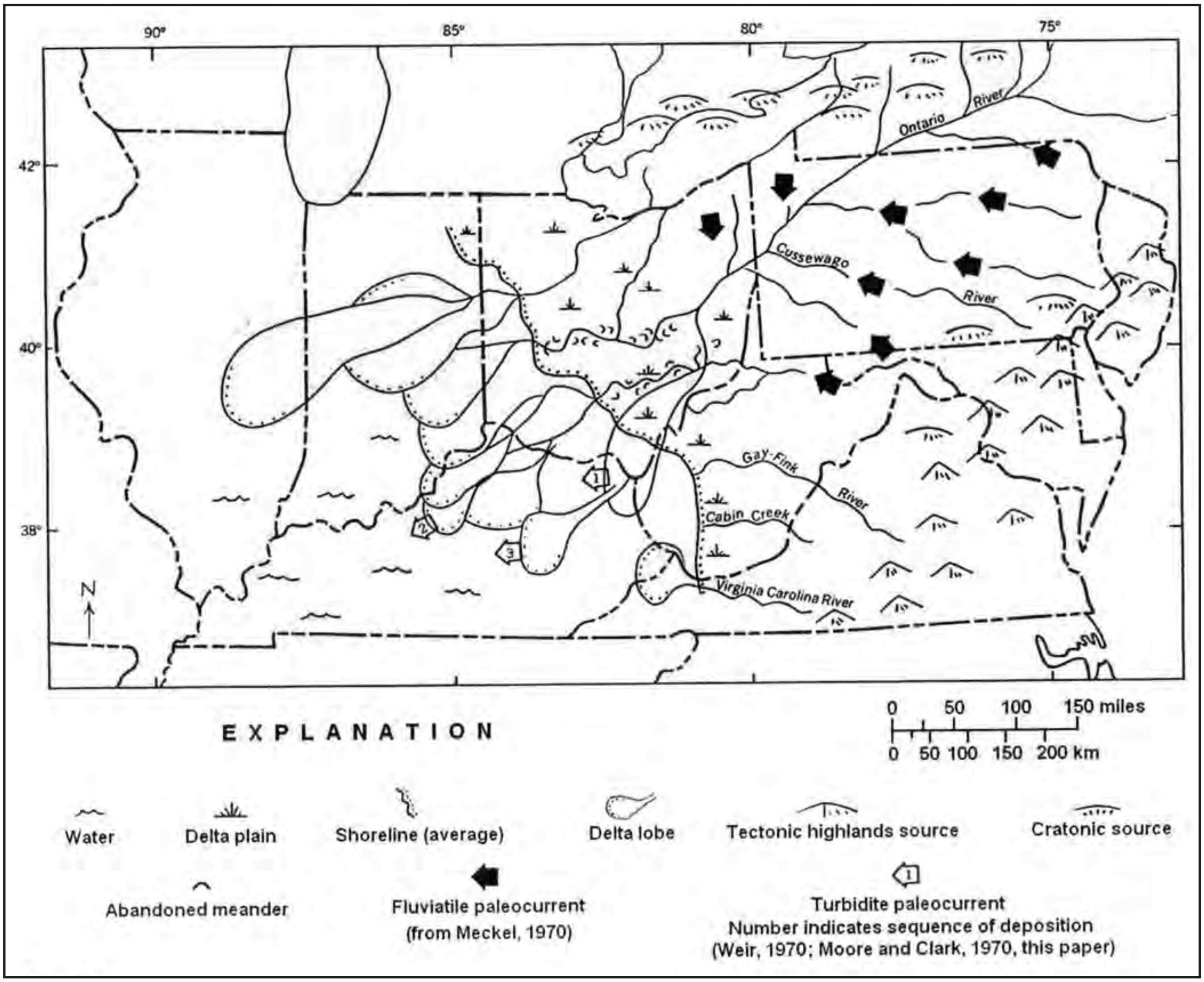


Figure 42. Paleogeographic map showing the development and progradation of the "Borden delta complex" in Kentucky, Indiana, Illinois, and Ohio in Early Mississippian time (adapted from Kepferle, 1977).

the shale, indicating the presence of greenish phyllosilicates, (2) the abundance of siderite nodules, a mineral that forms under reducing conditions (Garrets and Christ, 1965), and (3) the presence of a dysaerobic fauna (the Cave Run Lake fauna; Fig. 32) near the base of the Nancy Member (Mason and Kammer, 1984; Work and Mason, 2005).

Aerobic conditions in the upper part of the Nancy Member are evidenced by (1) the presence of an open-marine fauna, (2) the high degree of bioturbation of these shales, resulting in a nearly complete homogenization of the sediment, and (3) the occurrence of delta-front sands and silts of the Cowbell Member, which conformably overlie the Nancy Member (Fig. 41). Additional work and interpretations regarding the Nancy Member can be found in papers by Mason and Chaplin (1979), Chaplin (1980, 1982, 1985), Mason and Lierman (1985, 1992), and Work and Mason (2005).

Economically, farther north in Kentucky, the Berea Sandstone commonly occurs in facies relationship with the Bedford Shale (e.g., Pashin and Ettensohn, 1987, 1992b, 1995), and the Berea is an important hydrocarbon reservoir rock, generating many major oil and gas fields in eastern Kentucky and adjacent parts of Virginia, West Virginia, and Ohio (e.g., Tomastik, 1996). The Berea, however, reflects more proximal deltaic, fluvial, and platform environments, and this part of Kentucky was too far basinward to support these kinds of deposits (Ettensohn, 1979). However, in lower parts of the overlying Borden Formation, the turbiditic siltstones of the Farmers Member do form another series of reservoir rocks called the Weir Sandstone in the subsurface (Matchen and Vargo, 1996). As indicated above, the Farmers is a package of turbiditic siltstones/sandstones in prodelta mudstones equivalent to the Nancy Member and Henley Bed. However, in the subsurface of eastern Kentucky there are many such packages of turbiditic siltstones and sandstones in the same setting, but not necessarily of the same exact age as the Farmers, and hence there may be several different Weir Sandstones in the subsurface. Where these siltstones have developed sufficient natural or fracture porosity, they may form important hydrocarbon reservoirs (Matchen and Vargo, 1996). More information and interpretations regarding the Nancy Member can be found in papers by Mason and Chaplin (1979), Chaplin (1980, 1982, 1985), Mason and Lierman (1985), Lierman and others (1992), and Work and Mason (2005).

STOP 8: Granitic Dropstone Embedded in the Uppermost Cleveland Shale Member of the Ohio Shale

R. Thomas Lierman, Charles E. Mason, and Frank R. Ettensohn

At this stop we will examine the unusual occurrence of a probable granitic dropstone, called the Robinson boulder, in uppermost parts of the Cleveland Shale Member of the Ohio Shale and its implications.

Immediately after crossing a bridge, turn left onto a dirt road. We will park here and walk up Logan Hollow Road 0.2 mi to stop 8. After visiting stop 8, we will retrace our route back to the junction of Bratton Branch Road and U.S. 32. Stop 8 is located along a creek called Logan Hollow Branch, approximately 0.2 mi north of a road bridge located at the junction of Bratton Branch and Logan Hollow Roads. It is located in the east-central part of the Morehead quadrangle, Rowan County, Kentucky; see Figure 1. Latitude: 38°11’36.1”; longitude –083°29’36.6”.

Stratigraphically, the boulder occurs at the very top of the Upper Devonian Cleveland Member of the Ohio Shale (Fig. 43). The Bedford Shale can be easily dug out along the bank of the stream where it overlies the Cleveland Shale. Continuing down creek on the eastern side, a more complete section including both the Bedford and Sunbury Shales is present. A stratigraphic section of this stop is shown in Figure 43.

Statistics

Granite Boulder. First discovered by Michael J. Robinson of NYTIS Exploration Co. LLC in January 2006. This was brought to our attention on July 8, 2006.

Size and Shape. The boulder is a roughly square-shaped mass that projects from the Cleveland Shale on the bottom of the creek along Logan Hollow. The sides of the boulder are flat, whereas the corners and edges are rounded. The top may be faceted (Figs. 44–45).

Size. 1.3 x 1.7 m (4.3 x 5.6 ft).

Thickness. Approximately 0.60 m (2.5 ft).

Density. 2.70 g/cm^3; most granites fall in a range of 2.7 to 2.8 g/cm^3.

Estimated Weight. Approximately 3 tons.

Lithology. Originally a biotite granite; it has been subjected to low-grade metamorphism.

Mineralogy. Quartz, K-feldspar (microcline), biotite mica (Plate 1).

Petrology, Thin-section examination of samples from the granite boulder show that it has been subjected to low-grade (greenschist) metamorphism. This level of metamorphism is indicated by the presence of highly strained quartz crystals, along with composite quartz grains in the granite. Bent or kinked biotite has in places been altered to chlorite. In addition, much of the feldspar (microcline) is replaced by a mosaic of calcite crystals.

Age. Zircon crystals extracted from the boulder provided an Early Ordovician concordia age of 474 ± 5 Ma. Some of these exhibited inherited cores with a Grenvillian age of 1,156 ± 230 Ma (Ettensohn and others, 2008).

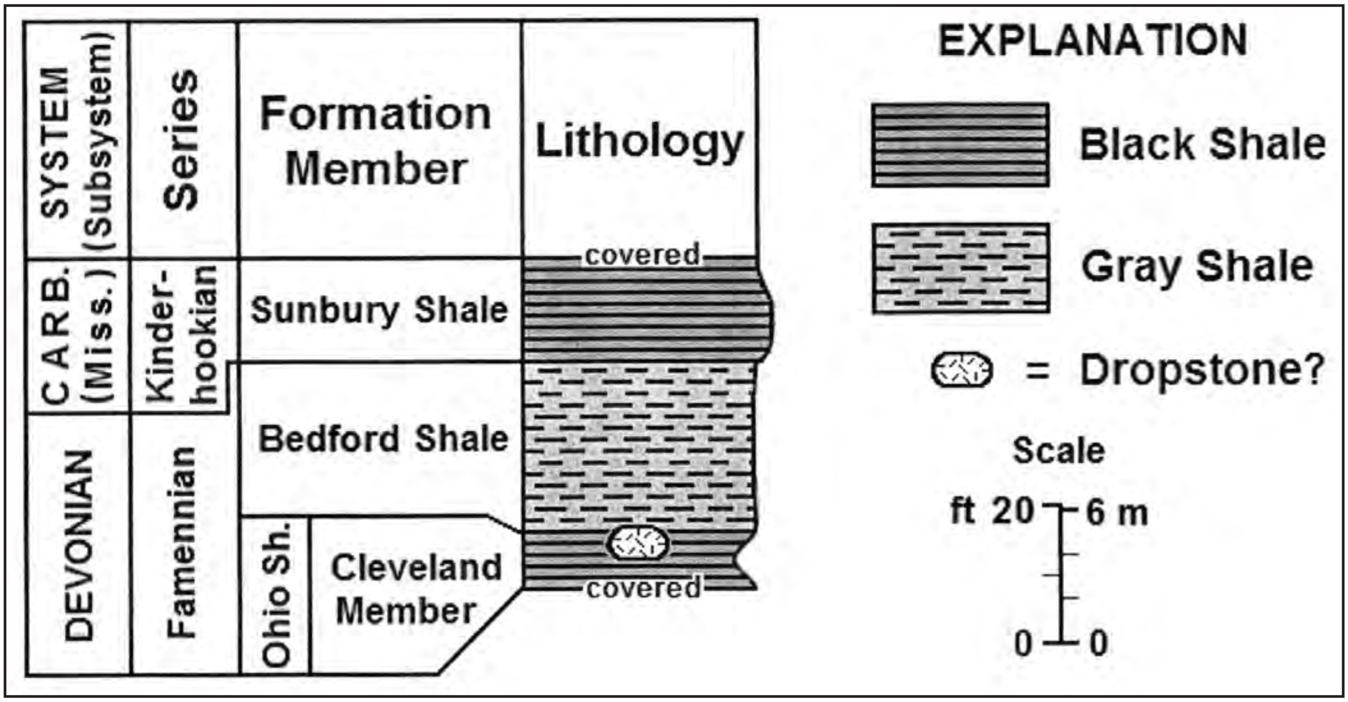


Figure 43. The stratigraphic section in Logan Hollow Branch at optional stop 8.

Interpretation. Taking into account the size, weight, shape, and exotic lithology of this boulder, we think that it is an ice-rafted dropstone that was transported to and then released from a melting iceberg at this site. We suggest that this is the result of alpine glaciation that was occurring in the Acadian highlands some 200 to 250 mi east of this locality. Most paleogeographic reconstructions for this time period place Kentucky and the Acadian Mountains at around 30° south latitude. For this object to be found in rocks from a marine setting, such as is represented by the Cleveland Member, suggests that these glaciers would have had to extend from the Acadian highlands westward to sea level, with icebergs calving off along the western edge of these glaciers.

Support for this hypothesis comes from several lines of evidence. (1) The lithology of the boulder is similar to Grenville-age granites in the region of the central Appalachian highlands. In particular, the bluish tint to the quartz has been noted by several authors in Grenville-age granites in the central Appalachian region. (2) The age of the granite boulder is in keeping with the ages of other rocks in the Appalachian region. Zircons were extracted from the boulder, which give an Early Ordovician concordia age of 474 ± 5 Ma. A number of these zircon crystals had inherited cores with a Grenvillian age of 1,156 ± 230 Ma. The boulder clearly had an early Appalachian or Laurentian origin. (3) The overall shape of this boulder with its flat sides and rounded edges and corners is in keeping with the shape of other glacial erratics. During the course of their movement, rocks that are embedded within a glacier grind against other rocks or can scrape against the underlying bedrock. In the process, this rounds off corners and planes smooth surfaces on embedded rocks, eventually producing this characteristic appearance. This process is also responsible for the generation of glacial striations and polished surfaces on the glacial erratics; unfortunately, this has not been observed on this boulder, though we are not able to see the bottom surface of the boulder. It should be noted that less than 10 percent of glacial erratics actually have striations on them. (4) Probably the best argument is that we simply have no other mechanism that could explain the presence of a large, 3-ton granite boulder deposited in the middle of an epicontinental sea, as the boulder was clearly penecontemporaneous with the Devonian sediments in which it is embedded. The only other mechanism that could potentially transport an object of this nature is root-rafting. Root-rafting is a transport mechanism that begins when soil and

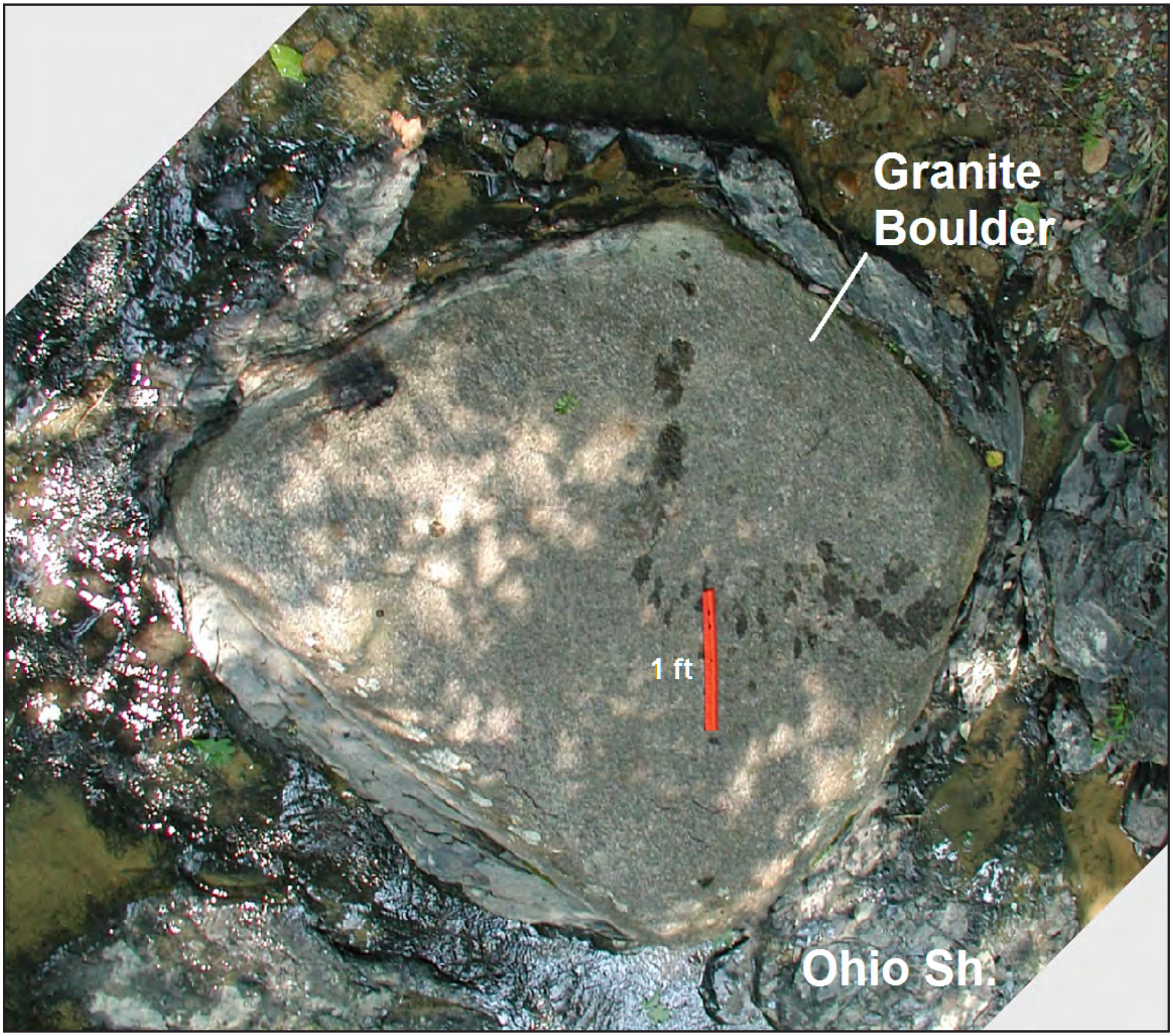


Figure 44. Top view of the granite boulder. Note the flattened sides, faceted top, and rounded corners of the boulder. Scale is 1 ft.

rocks get tangled up in the root systems of uprooted trees. These fallen trees, along with any soil and rock embedded in their roots, are later picked up by floodwaters and transported downstream to the sea. Root-rafting can probably be excluded, however, because the root systems of Late Devonian trees were apparently not sufficiently developed to wrap around and transport objects of this size and weight. All references to root-rafting that we have found to date involve rocks that are no larger than pebbles or cobbles in size. More information can be found in abstracts by Lierman and Mason (2007) and Ettcnsohn and others (2007, 2008).

This fascinating subject is discussed in greater detail in the following article.

Figure 45. Boulder is clearly imbedded in the Cleveland Member of the Ohio Shale. Note upturned layers or “mud drape” along edges of boulder.

Kentucky Dropstone “Ices” the Case for Late Devonian Alpine Glaciation in the Central Appalachians: Implications for Appalachian Tectonics and Black-Shale Sedimentation

F.R. Ettensohn, R.T. Lierman, C.E. Mason, A.J. Dennis, and E.D. Anderson

Abstract. Upper Devonian diamictites from the eastern Appalachian Basin, previously interpreted as debris-flow deposits, are glacial in origin based on well-defined stratigraphic, biostratigraphic, tectonic, and paleogeographic connections with a newly discovered in-situ dropstone from Upper Devonian black shales in east-central Kentucky. Together, the dropstone and diamictites, nearly 500 km apart, provide the first reported evidence for alpine glaciation in an ancient orogen and for tidewater glaciers in an ancient foreland basin. Both occurrences are related to an Acadian/Neoacadian transpressional regime that generated proximal foreland-basin subsidence and high coastal mountains in a paleoclimatic setting conducive to alpine glaciation. The 474-Ma age of the dropstone lithology and its deposition in black muds allow new interpretations about early Taconian Orogeny and controls on black-shale sedimentation.

Introduction. Late Devonian alpine glaciation in a basin that was at or near base level and in a subtropical setting for most of Paleozoic time seems implausible, but that is exactly what we support herein. Others have reached similar conclusions

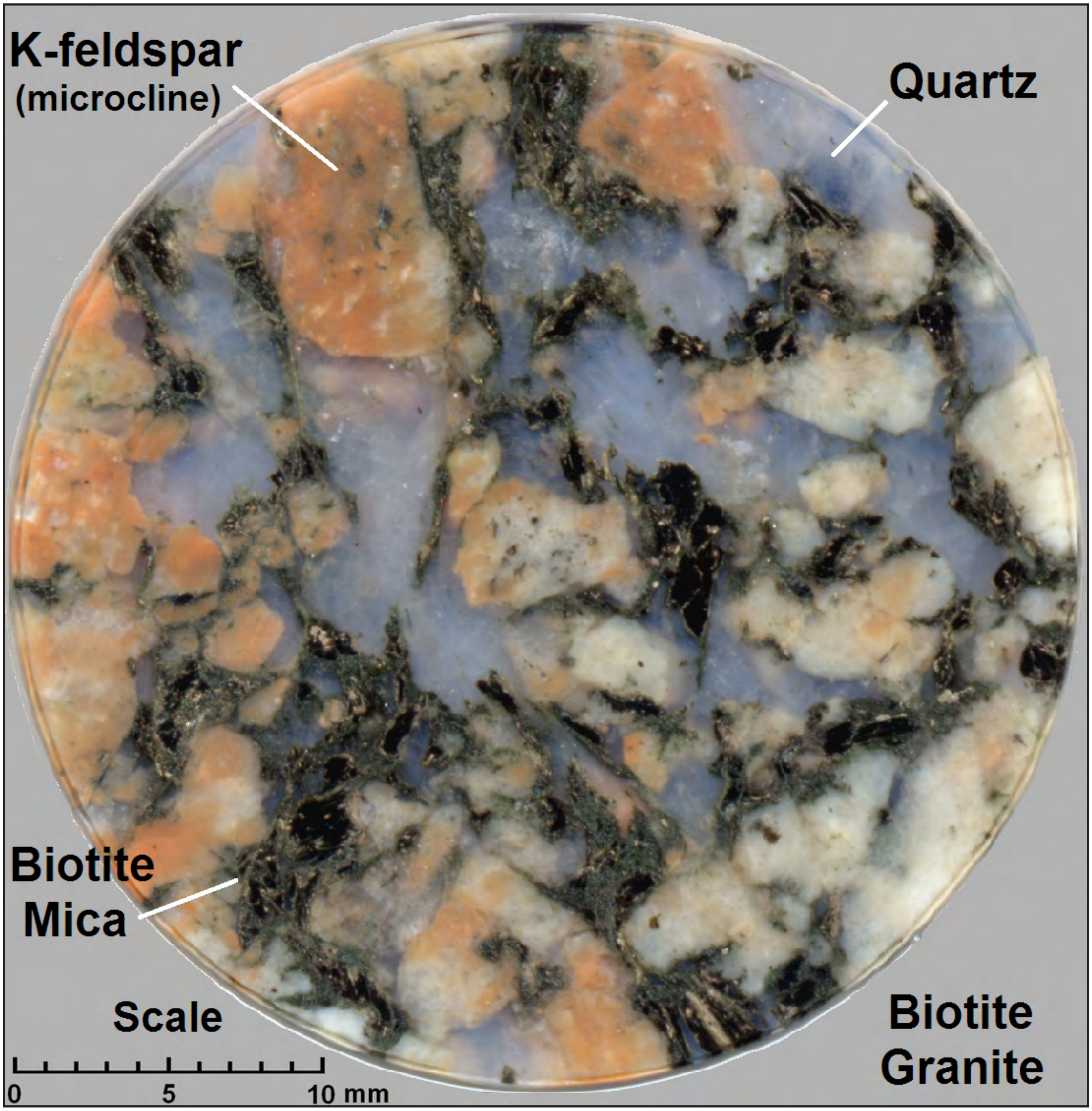


Plate 1. Close-up view of a polished core segment from the boulder, which is a biotite granite with quartz, K-feldspar (microcline), and biotite. Note the bluish tint of the quartz grains.

based on Upper Devonian diamictites (Fig. 46A), pebbly mudstones, and laminites in the Appalachian Basin of eastern and south-central Pennsylvania and adjacent parts of Maryland (Sevon, 1973; Cecil and others, 2004; Dennis, 2007). Although clear evidence supports Late Devonian, Gondwanan, continental glaciation (Crowell, 1999), proof for Late Devonian, Appalachian, alpine glaciation is controvertible (Sevon, 1979; Berg, 1999). Late Silurian to Early Carboniferous time was a globally warm, greenhouse period, but it was interrupted by an episode of latest Devonian–earliest Mississippian global cooling (Frakes and others, 1992), during which Gondwanan glaciation in

Figure 46. A. Close-up view of diamictite at Sideling Hill, Md. B. View showing faceted upper surface of in situ Robinson boulder, north of Morehead, Ky.

South America and Africa and likely Appalachian alpine glaciation briefly unfolded. Herein, we report the occurrence of an in situ igneous dropstone (Fig. 46B) in basinal, black, marine shales of the Upper Devonian Cleveland Member of the Ohio Shale in east-central Kentucky. Although the Kentucky and Pennsylvania-Maryland areas are widely separated (Fig. 47), stratigraphic occurrence and biostratigraphy suggest temporal equivalence of both deposits and provide the first definite evidence for Paleozoic, Appalachian, alpine glaciation.

Appalachian Diamictites. Upper Devonian (Famennian) diamictites (Fig. 46A) are unique in the Appalachian Basin; they are included in the lower Rockwell Formation in south-central Pennsylvania and adjacent parts of Maryland (Berg and others, 1980) and the lower Spechty Kopf Formation in the Anthracite region of eastern Pennsylvania (Epstein and others, 1974) (Fig. 47). Both formations are considered to be transitional between dominantly red, Catskill/Hampshire, alluvial-plain deposits and largely fluvial, brown to gray, Pocono sandstones (e.g., Berg, 1999). These poorly exposed, dark gray diamictites are imbedded, having thicknesses ranging from 0 to 185 m and a discontinuous distribution that is apparently restricted to large channels eroded into the underlying Catskill/Hampshire formations. The belt of diamictites extends in a sinuous trend about 400 km from northeastern Pennsylvania to central Maryland and is no wider than 45 km in northeastern Pennsylvania and 25 km in central Maryland (Cecil and others, 2004) (Fig. 47).

Pebbles and cobbles of quartz, siltstone, sandstone, chert, rhyolite, metasediments, meta-igneous, and reworked Catskill material, supported in a sandy, mudstone matrix, characterize the diamictites (Bjerstedt, 1986; Sevon and others, 1997). Rare boulders up to 2 m in diameter are present, and some clasts are striated and faceted. Moreover, clasts in the diamictites are consistently larger and of different compositions than clasts in underlying or overlying units (Sevon and others, 1997). Up to three diamictite units may be present in an exposure, and some units grade upward into a lacustrine-like sequence that includes, in ascending order, a pebbly mudstone, laminite, and a planar-bedded, rippled sandstone (Sevon and others, 1997; Berg, 1999). Some sandstones associated with the unit contain *Skolithos* trace fossils (Bjerstedt, 1986) (Fig. 48).

Both the Rockwell and Spechty Kopf Formations are Devonian-Mississippian units, but placement of the boundary has been uncertain (e.g., Berg, 1999). Bjerstedt (1986) suggested an Early Mississippian age for the diamictites, based on stratigraphic considerations that associate them with the overlying Riddlesburg Shale Member of the Rockwell, which contains Mississippian plant

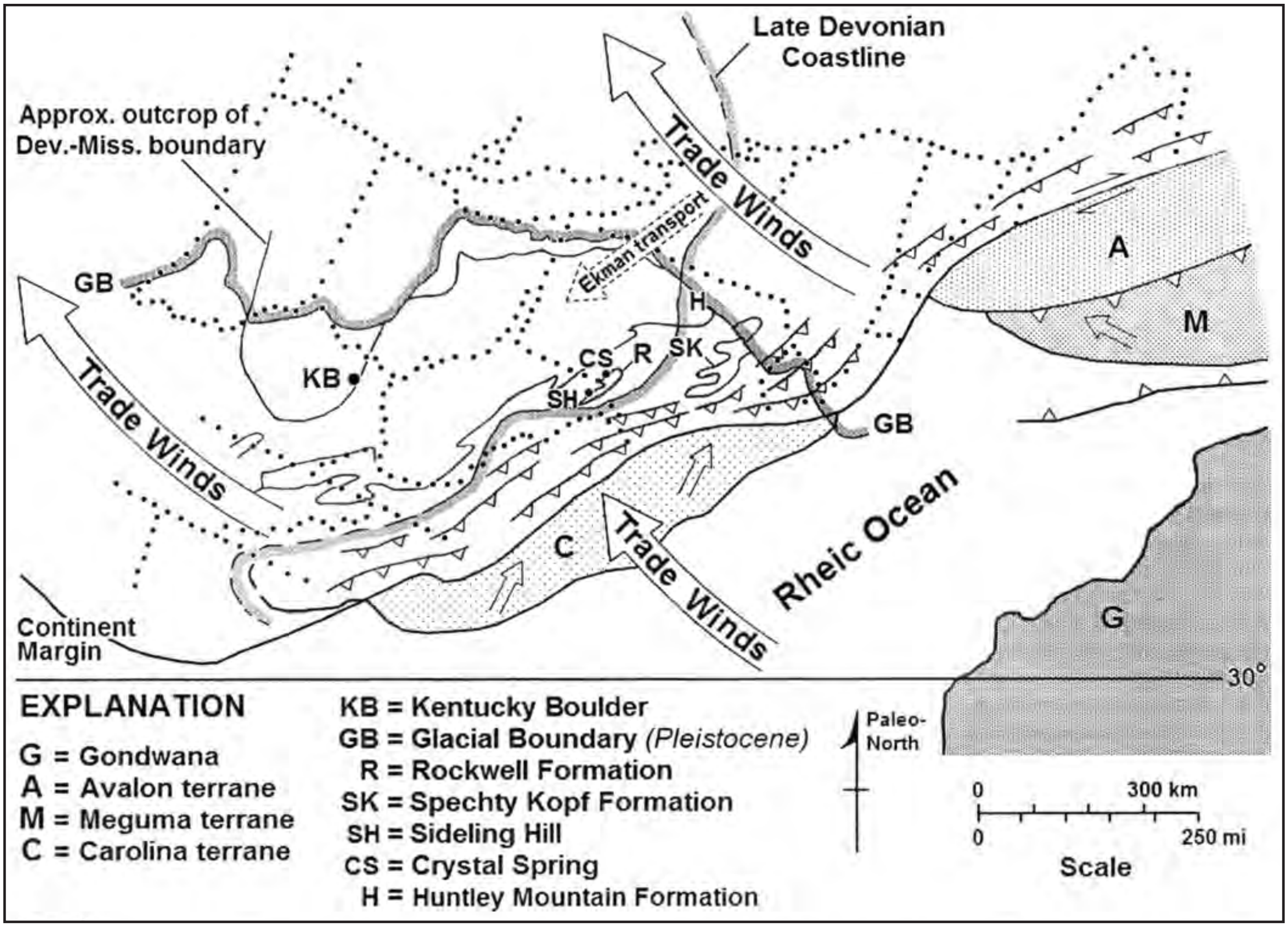


Figure 47. Late Devonian paleogeographic reconstruction of southeastern Laurussia, superimposed with positions of outcrop belts, field localities, Pleistocene glacial boundary, modern political boundaries, and likely Late Devonian wind and current directions.

fossils and is easily correlated with well-dated, marine, Sunbury black shales to the west (Fig. 48). However, recent diamictite palynology from the Spechty Kopf Formation (Sevon and others, 1997) shows presence in the upper *pusillites-lepidophyta* Miospore Zone, which is roughly equivalent to the middle *S. praesulcata* Conodont Zone, both of which are Late Devonian—but not latest Devonian—in age. Locally, Rockwell diamictites are overlain by sandstones correlated with the Berea Sandstone (Reger, 1927) (Fig. 48).

The origin of the diamictites has been troubling. Initially, Sevon (1973) suggested that they are glacial drift transported seaward into marine embayments as subaqueous mud and debris flows, but the interpretation was later discounted in favor of a nonglacial, debris-flow origin. These deposits were also interpreted to reflect lacustrine deposition on the Catskill alluvial plain (Berg, 1999), deposition in marine embayments during flooding of the Catskill alluvial plain (Kammer and Bjerstedt, 1986), deposition following breach of a structural front or tectonic basin (Bjerstedt, 1986), and deposition following a Late Devonian bolide impact (Sevon and others, 1997). In contrast, Cecil and others (2004), Dennis (2007), and Brezinski and others (2008) used similar evidence to support a glacial origin, suggesting that the diamictites and overlying lacustrine-like sequences actually reflect tillites deposited during glacial maxima succeeded by glaciomarine sedimentation during ice retreat. Clearly, discrimination between the two major interpretations is necessary, and the means may be available in some peculiar igneous and metamor-

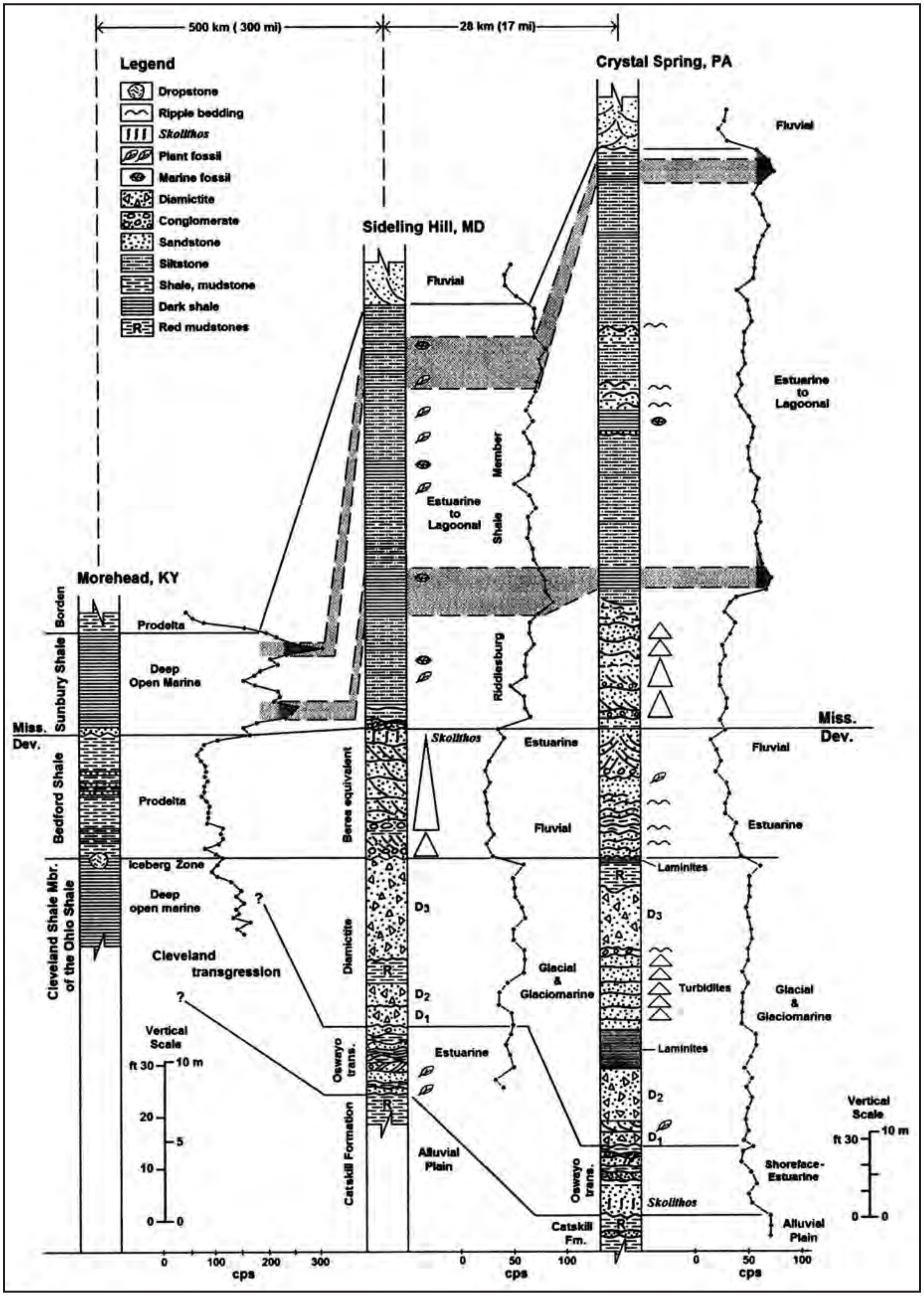


Figure 48. Stratigraphy and gamma-ray correlations between the dropstone locality in Kentucky and the diamictite localities in Maryland and Pennsylvania.

phic boulders from the Appalachian highlands of eastern Kentucky.

For this project we examined only Rockwell occurrences because the Rockwell contains definite marine and marginal-marine units. Specifically, we examined the diamictite-containing sections at Sideling Hill, Md., and Crystal Spring, Pa. (Bjerstedt, 1986; Cecil and others, 2004) (Figs. 47–48).

Kentucky Dropstones? Large igneous and metamorphic boulders found on the hillsides of eastern Kentucky were first reported by Kentucky geologist W.R. Jillson (1924). Leverett (1929) suggested that they originated in Canada, and despite occurrences well south of the glacial boundary, interpreted them to be pre-Illinoian glacial drift or ice-railed dropstones from proglacial lakes (Fig. 47). Pleistocene glaciation was the accepted interpretation until 2007, when Lierman and Mason (2007) reported a large faceted, biotite-granite boulder (1.7 x 1.3 x 0.75 m; the Robinson boulder), weighing approximately 3 tons, embedded in situ in uppermost parts of the Upper Devonian Cleveland Shale Member of the Ohio Shale, just below and projecting into the overlying Bedford Shale in Rowan County, east-central Kentucky (Figs. 46B, 48). Palynology indicates presence in the upper *pusillites-lepidophyta* Miospore Zone (G. Clayton, pers. commun., 2007), whereas conodonts indicate presence in the middle *S. praesulcata* Zone (Ettensohn and others, 1989). Zircons from the boulder provided an Early Ordovician concordia age of 474 ± 5 Ma, and some exhibited inherited cores with a Grenvillian age of 1,156 ± 230 Ma. Hence, the boulder clearly had an early Appalachian or Laurentian origin.

The 500-km Connection—Linking Diamictites and Dropstones. The Cleveland Shale of eastern Kentucky is separated by about 500 km from the diamictites in Pennsylvania and Maryland (Fig. 47). Both occurrences are of similar age based on palynology and conodont biostratigraphy, but the Kentucky boulder occurs in an open-marine sequence in an outcrop belt that is distant from and unconnected to outcrop belts containing the marine-to-nonmarine Rockwell and Spechty Kopf Formations. Although Rockwell and Spechty Kopf outcrop belts are also unconnected, they contain similar nonmarine lithologies but differ in that the Rockwell includes the prominent marine/marginal-marine Riddlesburg black shale (Fig. 48).

Proponents differ strongly on glacial versus nonglacial origins for the diamictites, and origins may be important, because a glacial origin has significant tectonic, paleoclimatic, and source-rock implications. Moreover, any clear link between the time-equivalent dropstones and diamictites means that the diamictites were related to tidewater glaciers. We suggest here two ways of establishing plausible connection: direct stratigraphic correlation using "radioactive" stratigraphy and establishing the likelihood of connection through paleogeographic reconstruction.

Much of the Devonian–Lower Mississippian section in the western and central Appalachian Basin comprises lithologically uniform black shales like the Cleveland and Sunbury with a few intervening gray-shale intervals like the Bedford/Berea (Fig. 48). Except for rare gray-shale intervals, visual correlation among the black shales is difficult. In the subsurface, however, correlation in black shales has long involved tracing various positive (increased organic-rich components) and negative (increased elastic dilution) deflections on gamma-ray logs. In the 1970's, similar correlation was extended to surface exposures using a hand-held scintillometer to generate radioactivity profiles (Ettensohn and others, 1979). Consequently, one or more deflections were grouped together as radioactive zones, and the zones, in both the surface and subsurface, were correlated across the basin (e.g., Kepferle and others, 1978). Similar radioactive correlation between the "boulder section" in Kentucky and the diamictite sections at Sideling Hill and Crystal Spring (Fig. 48) shows that the diamictites occur just below a Rockwell sandstone with a deflection pattern similar to that of the Bedford/Berea in Kentucky, which is overlain by Riddlesburg black shales with a pattern broadly similar to that of the Sunbury in Kentucky. The Sideling Hill radioactivity profile strongly supports what physical and biostratigraphic correlations have long suggested: namely, that the sandstone is a Bedford/Berea equivalent (Reger, 1927) and that the Riddlesburg Shale is a Sunbury equivalent (e.g., Bjerstedt, 1986; Kammer and Bjerstedt, 1986). Similarly, the diamictites that directly underlie the Bedford/Berea–equivalent sandstone in the Rockwell seem to correlate well in stratigraphic position with the boulder occurrence just below the Bedford Shale in northeastern Kentucky, and the biostratigraphy already mentioned supports this correlation.

Stratigraphically and biostratigraphically, the diamictites and dropstone seem to be "a match," but two questions remain. First, how did glacial tongues reach the sea across what existing evidence suggests was a vast alluvial plain (Catskill/Hampshire Formations)? Eustatic sea-level curves (e.g., Johnson and others, 1985) show a late, but not latest, Devonian (late Famennian) sea-level rise, reflected in the area by transgressive, tidal, and shoreface sands of the Oswayo Member of the Price Formation (e.g., Bjerstedt, 1986; Kammer and Bjerstedt, 1986) and farther south by the Finzel tongue of the Greenland Gap Formation (Dennison and others, 1986). In fact, Dennison (1985) used regional stratigraphic evidence to show that this sea-level rise drove a large shallow-marine embayment at least 160 km eastward into western Maryland and central Pennsylvania (Fig. 47) while tongues of the Cleveland Shale expanded eastward. At both Sideling Hill and at Crystal Spring, this trangression is represented by a few marginal-marine sands below the diamictites, some of which contain the marginal-marine trace fossil *Skolithos* (Bjerstedt, 1986) (Fig. 48). Hence, if the diamictites are glacial, glacial tongues at least locally must have reached shallow seas so that boulder-bearing icebergs could be launched. It is also now clear that this episode of alpine glaciation was coeval with the Late Devonian (Cleveland, Oswayo, or Finzel) transgression, a fact previously obscured by suggestions that the diamictites were Mississippian in age.

Given likely contact with the sea, was it paleogeographically feasible to move boulder-laden icebergs from the Pennsylvania-Maryland area to eastern Kentucky? Late Devonian paleogeographic restorations (e.g., Scotese, 2003) show that the diamictites formed at about 25°S latitude. At this latitude, prevailing trade winds moved northwestwardly toward the equator, but because of Southern Hemisphere Coriolis deflection to the left (Ekman transport), surface currents in the sea would have moved toward Kentucky and points southwestward. So Kentucky would have been ideally situated to receive icebergs calving from alpine glaciers in Acadian/Neoacadian highlands to the northeast (Fig. 47).

Although modern analogs for tropical, tidewater glaciers do not exist, some combination of coastal setting, strongly oriented high structural relief (Acadian/Neoacadian mountains; Dennis, 2007), a period of enhanced moisture and cooling (Cecil and others, 2004), a likely rain shadow across the mountains (Ettensohn, 1985b), and glacial aspect (e.g., Evans, 2006) must have briefly generated ideal conditions for development and flow of alpine glaciers to the foreland sea.

Implications

Glaciation. Evidence from occurrence, stratigraphy, and age effectively preclude any origin for the Robinson boulder except as a glacial dropstone with an Appalachian source. Dropstones require glacial sources, and paleogeographic and paleoclimatic considerations for the time support likely glacial sources to the present-day northeast. Valley-fill diamictites, many with associated varve-like and dropstone-rich stratigraphic sequences long associated with glaciation and of the same age and stratigraphic position as black shales containing the boulder, occur to the northeast in Pennsylvania and Maryland. The clear-cut, temporal, stratigraphic, and paleoenvironmental connections between boulders and diamictites strongly support a glacial origin for Rockwell and Spechty Kopf diamictites and indicate that warm, wet-based, marine-terminating glaciers extended periodically beyond the coast. Cold waters emanating from the glaciers may also explain the coeval, exotic, cold-water faunas of western Pennsylvania and southern New York (Cecil and others, 2004) and the occurrence of exotic pebbles and large-scale, water-release structures in the correlative Huntley Mountain Formation of central Pennsylvania (Woodrow and Richardson, 2006) (Fig. 47), which may represent rapidly deposited outwash.

Black-Shale Sedimentation. The Cleveland Shale is part of the Appalachian, Devonian-Mississippian, black-shale sequence and, like other epicontinental, marine, black shales, is still a subject of controversy regarding the control of anoxia versus high productivity in generating organic-rich sediments (e.g., Pederson and Calvert, 1990). Although controls have been suggested for Appalachian Basin black shales (Ettensohn, 1992b), glaciation is not among them. Recent work shows that icebergs are hotspots of chemical and biological enrichment, and that iceberg-prone seas are areas of enhanced organic productivity and sequestration of organic carbon in underlying sediments (Smith and others, 2007). Moreover, during deglacial episodes, glacial meltwater draining into adjacent seas commonly generates a fresher, lighter, surface layer that initiates a halocline. This promotes anoxia below the halocline, while at the same time, terrestrial nutrients in the meltwater enhance organic productivity in surface waters. So tidewater glaciation may contribute to both anoxia and enhanced organic productivity (e.g., Armstrong and others, 2005), and at least for upper parts of the Cleveland Shale, coeval glaciation must be considered as another control on organic-rich sedimentation.

Tectonics. Late Devonian–Early Mississippian time in the central Appalachians saw culmination of the Acadian Orogeny and inception of the Neoacadian Orogeny, marked by the transpressional collapse of the Tugaloo and Cat Square terranes and reaccretion of the Carolina superterrane to Laurentia at southern parts of the New York Promontory (Dennis, 2007). Subsequent (late Famennian) orogenesis generated highlands that stood astride the trade-wind belt in conditions moist and high enough to initiate alpine glaciation (Fig. 47). Although Neoacadian tectonism coincided with a global sea-level rise (Johnson and others, 1985), coeval loading-related subsidence may have forced the Cleveland-Oswayo-Finzel transgression far enough eastward of earlier Famennian transgressions (Dennison, 1985) to contact glaciers from the new highlands.

If the 474 Ma age of the Robinson boulder lithology is any indication, Neoacadian highlands must have contained rocks from Early to Middle Ordovician, peri-Laurentian, metaplutonic suites and their associated country rocks. This particular age and basement may be correlative with the Shelburne Falls arc (~470–485 Ma) of the New England Appalachians (Karabinos and others, 1998). This boulder is the first indication that peri-Laurentian, Early to Middle Ordovician arc magmatism extended to the central Appalachians. Neoacadian deformation apparently uplifted Shelburne Falls–equivalent rocks to elevations where they were eroded by alpine glaciation with the attendant survival of rare erratics preserved within the Rockwell–Spechty Kopf and correlative Kentucky strata.

Conclusions. We cannot say certainly that the Upper Devonian, Rockwell, and Spechty Kopf diamictites are tillites, but their likely glacial origin is supported by the occurrence of a faceted boulder that can only be a dropstone, in biostratigraphically and stratigraphically coeval, Upper Devonian, marine, black shales in east-central Kentucky. Though the two occurrences are separated by about 500 km, the tectonic and paleogeographic frameworks make the connection very likely. Ultimately, both occurrences are related to a Neoacadian, transpressional, tectonic regime that created coastal mountains high enough and in the right paleoclimatic setting to generate alpine glaciation, while at the same time producing sufficient foreland subsidence to force transgressing seas eastward to the hinterland. This proximity allowed glaciers to sample older deformation at high structural levels, which are no longer preserved, and deliver exotic clasts to the adjacent foreland basin as diamictite and dropstones from tidewater glaciers. Accordingly, this is the first reported occurrence of alpine glaciation and tidewater glaciers in an ancient orogen and foreland basin. The dropstone also necessitates new considerations about black-shale sedimentation and provides the first evidence for early Taconian, Blountian tectonism in the central Appalachians.

References Cited

Aigner, T., and Reineck, H.-E., 1982, Proximality trends in modern storm sands from the Helgoland Bight (North Sea) and their implications for basin analysis: Senckenbergiana Maritima, v. 14, p. 183–215.

Algeo, T.J., Berner, R.A., Maynard, J.B., and Scheckler, S.E., 1995, Late Devonian oceanic anoxic events and biotic crises: "Rooted" in the evolution of vascular land plants?: GSA Today, v. 5, p. 45, 64–66.

Algeo, T.J., and Scheckler, S.E., 1998, Terrestrial-marine teleconnections in the Devonian: Links between the evolution of land plants, weathering processes, and marine anoxic events: Philosophical Transactions of the Royal Society London, B 353, p. 113–130.

Allen, J.R.L., 1982, Sedimentary structures: Their character and physical basis: Developments in Sedimentology, v. 30, part II, 663 p.

Armstrong, H.A., Turner, B.R., Makhlouf, I.M., Weedon, G.P., Williams, M., Al Smadi, A., and Abu Salah, A., 2005, Origin, sequence stratigraphy and depositional environment of an Upper Ordovician (Hirnantian) deglacial black shale, Jordan: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 220, p. 273–289.

Barron, L.S., and Ettensohn, F.R., 1981, Paleoecology of the Devonian-Mississippian black-shale sequence in eastern Kentucky with an atlas of some common fossils: U.S. Department of Energy, Technical Information Center, DOE/ET/12040-151, 75 p.

Beaumont, C., Quinlan, G., and Hamilton, J., 1988, Orogeny and stratigraphy: Numerical models of the Paleozoic in the Eastern Interior of North America: Tectonics, v. 7, p. 389–416.

Berg, T.M., 1999, Chapter 8, Devonian-Mississippian transition, *in* Shultz, C.H., ed., The geology of Pennsylvania: Harrisburg and Pittsburgh: Pennsylvania Geological Survey and Pittsburgh Geological Society, p. 128–137.

Berg, T.M., Edmunds, W.E., Geyer, A.R., Glover, A.D., Hoskins, D.M., MacLachlan, D.B., Root, S.L., Sevon, W.D., and Socolow, A.A., comp., 1980, Geologic map of Pennsylvania: Pennsylvania Geological Survey, 4th Series, Map 1, scale 1:250,000, 3 sheets. Bjerstedt, T.W., 1986, Regional stratigraphy and sedimentology of the Lower Mississippian Rockwell Formation and the Purslane Sandstone based on the new Sideling Hill road cut, Maryland: Southeastern Geology, v. 27, p. 69–94.

Boswell, R., 1996, PLAY UDs: Upper Devonian black shales, *in* Roen, J.B., and Walker, B.J., eds., The atlas of major Appalachian gas plays: West Virginia Geological and Economical Survey Publication V-25, p. 93–99.

Bouma, A.H., 1962, Sedimentology of some flysch deposits; a graphic approach to facies interpretation: Amsterdam, Elsevier, 168 p.

Brezinski, D.K., Cecil, C.B., Skema, V.K., and Stamm, R., 2008, Late Devonian deposits from the eastern United States signal an end of the mid-Paleozoic warm period: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 268, p. 143–151.

Bucher, W.H., 1919, On ripples and related sedimentary surface forms and their paleogeographic interpretation: American Journal of Science, v. 47, p. 249–210, 241–269.

Cecil, C.B., Brezinski, D., and DuLong, F., 2004, The Paleozoic record of changes in global climate and sea level: Central Appalachian Basin, *in* Southworth, S., and Burton, W., eds., Geology of the national capital region—Field trip guidebook: U.S. Geological Survey Circular 1264, p. 77–135.

Chaplin, J.R., 1980, Stratigraphy, trace fossil associations, and depositional environments in the Borden Formation (Mississippian, northeastern Kentucky) (annual field conference of the Geological Society of Kentucky): Kentucky Geological Survey, 114 p.

Chaplin, J.R., 1982, Paleoenvironments and biostratigraphy of the Borden and parts of the Newman and Breathitt formations (Mississippian-Pennsylvanian) in northeastern Kentucky (twelfth annual conference of the Great Lakes Section–Society of Economic Paleontologists and Mineralogists): Oxford, Miami University, Great Lakes Section, Society of Economic Paleontologists and Mineralogists, 191 p.

Chaplin, J.R., 1985, Early Mississippian deltaic sedimentation in northeastern Kentucky, *in* Haban, S.M., ed., Field excursion 6, Guidebook, Carboniferous of eastern Kentucky, Sixth Gondwana Symposium: Ohio State University Institute of Polar Studies Miscellaneous Publication 228, p. 26–30.

Chaplin, J.R., and Mason, C.E., 1979, Stop 3: The Devonian-Carboniferous boundary, *in* Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian Basin to the Illinois Basin through eastern Ohio and Kentucky (guidebook for Field Trip No. 4, IX International Congress of Carboniferous Stratigraphy and Geology): Lexington, University of Kentucky, p. 152–162.

Chaplin, J.R., and Mason, C.E., 1985, Berea Sandstone (Lower Mississippian) sedimentation in northeastern Kentucky, Stop 1, Rock Hill Chapel Section (Garrison Quarry), *in* Haban, S.M., ed., Field Excursion 6, Guidebook, Carboniferous of eastern Kentucky, Sixth Gondwana Symposium: Ohio State University Institute of Polar Studies Miscellaneous Publication 228, p. 36–50.

Coogan, A.H., and others, 1981, Early Mississippian deltaic sedimentation in central and northeastern Ohio, *in* Roberts, T.G., ed., GSA Cincinnati '81 Field Trip Guidebooks, v. 1: Stratigraphy, sedimentology: Falls Church, Va., American Geological Institute, p. 113–152.

Cooper, J.R., 1943, Flow structures in the Berea Sandstone and Bedford Shale of central Ohio: Journal of Geology, v. 51, p. 190–203.

Crowell, J.C., 1999, Pre-Mesozoic ice ages: Their bearing on understanding the climate system: Geological Society of America Memoir 192, 106 p.

Dennis, A.J., 2007, Cat Square Basin. Catskill elastic wedge: Silurian-Devonian orogenic events in the centralAppalachians and the crystalline southern Appalachians, *in* Sears. J.W., Harms, T.A., and Evenchick, C.A., eds., Whence the mountains? Inquiries into the evolution of orogenic systems: A volume in honor of Raymond A. Price: Geological Society of America Special Paper 433, p. 313–329.

Dennison, J.M., 1985, Catskill delta shallow marine strata, *in* Woodrow, D.L., and Sevon, W.D., eds., The Catskill delta: Geological Society of America Special Paper 201, p. 91–106.

Dennison, J.M., Beuthin, J.D., and Hasson, K.O., 1986, Latest Devonian–earliest Carboniferous transgressions, central and southem Appalachians, USA: Annals de la Société Geologique de Belgique, v. 109-1986, p. 123–129.

Dott, R.H., and Bourgeois, J., 1982, Hummocky stratification: Significance of variable bedding sequences: Geological Society of America Bulletin, v. 93, p. 663–680.

Elam, T.D., 1981, Stratigraphy and paleoenvironmental aspects of the Bedford-Berea sequence and the Sunbury Shale in eastern and south-central Kentucky: Lexington, University of Kentucky, master's thesis, 155 p.

Epstein, J.B., Sevon, W.D., and Glaeser, J.D., 1974, Geology and mineral resources of the Lehighton and Palmerton quadrangles, Carbon and Northampton Counties, Pennsylvania: Pennsylvania Geological Survey, 4th Series, Atlas 195cd, 460 p.

Ettensohn, F.R., 1979, Depositional framework of Lower Mississippian subaqueous delta sediments in northeastern Kentucky, *in* Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian Basin to the Illinois Basin through eastern Ohio and Kentucky (guidebook for Field Trip No. 4, IX International Congress of Carboniferous Stratigraphy and Geology: Lexington, University of Kentucky, p. 133–136.

Ettensohn, F.R., 1985a, The Catskill delta complex and the Acadian Orogeny, *in* Woodrow, D.W., and Devon, W.D., eds., The Catskill delta: Geological Society of America Special Paper 201, p. 39–49.

Ettensohn, F.R., 1985b, Controls on development of Catskill delta complex basin-facies, *in* Woodrow, D.L., and Sevon, W.D., eds., The Catskill delta: Geological Society of America Special Paper 201, p. 65–77.

Ettensohn, F.R., 1987, Rates of relative plate motion during the Acadian Orogeny based on the spatial distribution of black shales: Journal of Geology, v. 95, p. 572–582.

Ettensohn, F.R., 1990, Estimating absolute depths for Devonian-Mississippian black shales in eastern United States: Implications for eustatic vs. tectonic control of "sea-level": Fall Program, Appalachian Basin Industrial Associates: Morgantown, W.Va., West Virginia University, v. 17, p. 40–139.

Ettensohn, F.R., 1991, Flexural interpretation of relationships between Ordovician tectonism and stratigraphic sequences, central and southern Appalachians, U.S.A., *in* Barnes, C.R., and Williams, S.H., eds., Advances in Ordovician geology: Geological Survey of Canada Paper 90-9, p. 213–224.

Ettensohn, F.R., ed., 1992a, Changing interpretations of Kentucky geology: Layer-cake, facies, flexure, and eustacy: Ohio Department of Natural Resources, Division of Geological Survey, Miscellaneous Report 5, 184 p.

Ettensohn, F.R., 1992b, Controls on the origin of the Devonian-Mississippian oil and gas shales, east-central United States: Fuel, v. 71, p. 1487–1492.

Ettensohn, F.R., 1992c, Physiography along the field-trip route, *in* Ettensohn, F.R., ed., Changing interpretations of Kentucky geology—Layer-cake, facies, flexure, and eustacy: Ohio Division of Geological Survey Miscellaneous Report No. 5, p. 2–6.

Ettensohn, F.R., 1992d, Regressive black shales and the *Protosalvinia (Foerstia)* zone, *in* Ettensohn, F.R., ed., Changing interpretations of Kentucky geology—Layer-cake, facies, flexure, and eustacy: Ohio Division of Geological Survey Miscellaneous Report No. 5, p. 76–77.

Ettensohn, F.R., 1992e, Silurian-Devonian contact and transgressive black shales at the base of the Ohio Shale, *in* Ettensohn, F.R., ed., Changing interpretations of Kentucky geology—Layer-cake, facies, flexure, and eustacy: Ohio Division of Geological Survey Miscellaneous Report No. 5, p. 74–76.

Ettensohn, F.R., 1994, Tectonic control on the formation and cyclicity of major Appalachian unconformities and associated stratigraphic sequences, *in* Dennison, J.M., and Ettensohn, F.R., eds., Tectonic and eustatic controls on sedimentary cycles: SEPM Concepts in Sedimentology and Paleontology 4, p. 217–242.

Ettensohn, F.R., 1995, Global and regional controls on the origin and burial of organic matter in Devonian-Mississippian black shales of North America: Houston Geological Society Bulletin, v. 37, p. 12–17.

Ettensohn, F.R., 1998, Compressional tectonic controls on epicontinental black-shale deposition: Devonian-Mississippian examples from North America, *in* Schieber, J., Zimmerle, W., and Sethi, P., eds., Shales and mudstones I: Stuttgart, E. Schweizerbart'sche Verlagsbuchhandlung, p. 109–128.

Ettensohn, F.R., 2004, Modeling the nature and development of major Paleozoic elastic wedges in the Appalachian Basin, USA: Journal of Geodynamics, v. 37, p. 657–681.

Ettensohn, F.R., 2005, Chapter 5, The sedimentary record of foreland-basin, tectophase cycles: Examples from the Appalachian Basin, *in* Mabesoone, J.M.. and Neuman, V.H., eds., Cyclic development of sedimentary basins, Developments in Sedimentology 57: Amsterdam, Elsevier, p. 139–172.

Ettensohn, F.R., 2008, Chapter 4, The Appalachian foreland basin in eastern United States, *in* Hsu, K.J., ed., Sedimentary basins of the world, v. 5, The sedimentary basins of United States and Canada, Miall. A.D., ed.: The Netherlands, Elsevier, p. 105–179.

Ettensohn, F.R., and Barron, L.S., 1981, Depositional model for the Devonian-Mississippian black shales of North America: A paleoclimatic-paleogeographic approach, *in* Roberts, T.G., ed., GSA Cincinnati '81 Field Trip Guidebooks, v. II: Economic geology, structure: Falls

Church, Va., American Geological Institute, p. 344–361.

Ettensohn, F.R., and Brett, C.E., 1998, Tectonic components in third-order Silurian cycles: Examples from the Appalachian Basin and global implications, *in* Landing, E., and Johnson, M., eds., Silurian cycles, linkages of dynamic stratigraphy with atmospheric, oceanic, and tectonic changes: New York State Museum Bulletin 491, p. 145–162.

Ettensohn, F.R., and Dever, G.R., Jr., eds., 1979, Carboniferous geology from the Appalachian Basin to the Illinois Basin through eastern Ohio and Kentucky (guidebook for Field Trip No. 4, IX International Congress of Carboniferous Stratigraphy and Geology): Lexington, University of Kentucky, 93 p.

Ettensohn, F.R., Dever, G.R., Jr., and Grow, J.S., 1988a, A paleosol interpretation for profiles exhibiting subaerial exposure "crusts" from the Mississippian of the Appalachian Basin, *in* Reinhardt, J., and Sigleo, W.R., eds., Paleosols and weathering through geologic time: Geological Society of America Special Paper 216, p. 49–79.

Ettensohn, F.R., and Elam, T.D., 1985, Defining the nature and location of a Late Devonian–Early Mississippian pycnocline in eastern Kentucky: Geological Society of America Bulletin, v. 96, p. 1313–1321.

Ettensohn, F.R., Goodmann, P.T., Norby, R.D., and Shaw, T.H., 1989, Stratigraphy and biostratigaphy of the Devonian-Mississippian black shales in west-central Kentucky and adjacent parts of Indiana, *in* Proceedings, 1988 Eastern Oil Shale Symposium: Institute for Mining and Minerals Research, Lexington, Ky., p. 237–245.

Ettensohn, F.R., Greb., S.F., Chesnut, D.R., Jr., Harris, D.C., Mason, C.E., Eble, C.F., Howell, P.D., Watson, A.E., and Johnson, W.K., 2002, Mississippian stratigraphy, depositional environments, and tectonic framework of the central Appalachian Basin, eastern Kentucky, U.S.A., *in* Hills, L.V., Henderson, C.M., and Bamber, E.W., eds., Carboniferous and Permian of the world: Canadian Society of Petroleum Geologists Memoir 19, p. 22–40.

Ettensohn, F.R., Lierman, T.R., and Mason, C.E., 2007, Dropstones, glaciation, and black shales: New inferences on black-shale origins from the upper Ohio Shale in northeastern Kentucky: Program with Abstracts, 36th Annual Meeting, Eastern Section, AAPG, p. 33–34.

Ettensohn, F.R., Lierman, T.R., Mason, C.E., Dennis, A.J., and Anderson, E.D., 2008, Kentucky dropstones and the case for Late Devonian alpine glaciation in the Appalachian Basin (USA) with implications for Appalachian tectonism and black-shale sedimentation [abs.]: Geological Society of America Abstracts with Programs, v. 40, p. 395.

Ettensohn, F.R., Miller, M.L., Dillman, S.B., Elam, T.D., Geller, K.L., Swager, D.R., Markowitz, G., Woock, R.D., and Barron, L.S., 1988b, Characterization and implications of the Devonian-Mississippian black-shale sequence, eastern and central Kentucky, U.S.A.: Pycnoclines, transgression, regression, and tectonism, *in* McMillan, N.J., Embry, A.F., and Glass, G.J., eds., Devonian of the world, proceedings of the Second International Symposium on the Devonian System: Canadian Society of Petroleum Geologists Memoir 14, v. 2, p. 323–345.

Evans, I.S., 2006, Local aspect asymmetry of mountain glaciation: A global survey of consistency of favored directions for glacial numbers and altitudes: Geomorphology, v. 73, p. 166–184.

Fairbridge, R.W., 1980, The estuary: Its definition and geodynamic cycle, *in* Olausson, E., and Cato, I., eds., Chemistry and biogeochemistry of estuaries, p. 1–36.

Fenneman, N.M., 1938, Physiography of eastern United States: New York, McGraw-Hill Book Co., 714 p.

Fischer, A.G., 1984, The two Phanerozoic supercycles, *in* Berggren, W.A., and Van Couvering, J.A., eds., Catastrophes and earth history: Princeton, Princeton University Press, p. 129–150.

Fisher, W.L., and others, 1969, Delta systems in the exploration for oil and gas: Research Colloquium, Bureau of Economic Geology, Austin, Texas, University of Texas, 78 p.

Frakes, F.A., Frances, J.E., and Syktus, J.I., 1992, Climate modes of the Phanerozoic: Cambridge, Cambridge University Press. 274 p.

Frazier, W.J., and Schwimmer, D.R., 1987, Regional stratigraphy of North America: New York, Plenum Press, 719 p.

Fuentes, S., Over, D.J., and Brett, C.E., 2002, Conodont biostratigraphy of the "Olentangy" Shale in northeastern Kentucky, *in* Algeo, T.J., and Brett. C.E., eds., Sequence, cycle and event stratigraphy of Upper Ordovician and Silurian strata of the Cincinnati Arch region: Kentucky Geological Survey, ser. 12, Guidebook 1, p. 136–140.

Garrels, R.M., and Christ, C.L., 1965, Solutions, minerals, and equilibria: New York, Harper and Row, 450 p.

Goldring, R., and Bridges, P., 1973, Sublittoral sheet sandstones: Journal of Sedimentary Petrology: v. 43, p. 736–747.

Gradstein, F.M., Ogg, J.G., and Smith, A.G., eds., 2004, A geologic time scale 2004: Cambridge, Cambridge University Press, 589 p.

Griffith, C., 1977, Stratigraphy and paleoenvironment of the New Albany Shale (Upper Devonian) of north-central Kentucky: Madison, University of Wisconsin, master's thesis, 214 p.

Hamblin, A.P., and Walker, R.G., 1979, Storm-dominated shallow marine deposits: The Fernie-Kootenay (Jurrassic) transition, southern Rocky Mountains: Canadian Journal of Earth Sciences, v. 16, p. 1673–1690.

Hasenmueller, N.R., Matthews, R.D., Kepferle, R.C., and Pollock, D., 1983, *Foerstia (Protosalvinia)* in Devonian shales of the Appalachian, Illinois, and Michigan Basins, eastern United States, *in* Proceedings, 1983 Eastern Oil Shale Symposium: Lexington, University of Kentucky Institute for Mining and Minerals Research, p. 41–58.

Heckel, P.R., and Witzke, B.J., 1979, Devonian world paleogeography determined from distribution of carbonate and related lithic, paleoclimatic indicators: Special Papers in Paleontology 23, p. 99–123.

Hoge, H.P., and Chaplin, J.R., 1972, Geologic map of the Morehead quadrangle, Rowan County, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1022, scale 1:24,000.

Hyde, J.E., 1911, The ripples of the Bedford and Berea formations of central and southern Ohio: Journal of Geology, v. 19, p. 257–269.

Hyde, J.E., 1953, Mississippian formations of central and southern Ohio, *in* Marple, M.F., ed., Ohio Geological Survey Bulletin 51, 335 p.

Jillson, W.R., 1924, Glacial pebbles in eastern Kentucky: Science, v. 9 (no. 1544), p. 101–102.

Johnson, J.G., Klapper, G., and Sandberg, C.A., 1985, Devonian eustatic fluctuations in Euramerica: Geological Society of America Bulletin, v. 82, p. 3263–3298.

Jordan, D.W., 1980, Trace fossils and stratigraphy of Devonian black shale in east-central Kentucky [abs.]: American Association of Petroleum Geologists Bulletin, v. 64, p. 729–730.

Kammer, T.W., 1985, Basinal and prodeltaic communities of the Early Carboniferous Borden Formation in northern Kentucky and southern Indiana: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 49, p. 179–121.

Kammer, T.W., and Bjerstedt, T.W., 1986, Stratigraphic framework of the Price Formation (Upper Devonian–Lower Mississippian) in West Virginia: Southeastern Geology, v. 27, p. 13–33.

Karabinos, P., Samson, S.D., Hepburn, J.C., and Stoll, H.M., 1998, Taconian Orogeny in the New England Appalachians: Collision between Laurentia and the Shelburne Falls arc: Geology, v. 26, p. 215–218.

Kepferle, R.C., 1977, Stratigraphy, petrology and depositional environment of the Kenwood Siltstone Member, Borden Formation (Mississippian), Kentucky and Indiana: U.S. Geological Survey Professional Paper 1007, 49 p.

Kepferle, R.C., and Roen, J.B., 1981, Chattanooga and Ohio Shales of the southern Appalachian Basin, *in* Roberts, T.G., ed., GSA Cincinnati '81 Field Trip Guidebooks, v. II: Economic geology. structure: Falls Church, Va., American Geological Institute, p. 259–343.

Kepferle, R.C., Wilson, E.N., and Ettensohn, F.R., 1978, Preliminary stratigraphic cross section showing radioactive zones in the Devonian black shales in the southern part of the Appalachian Basin: U.S. Geological Survey Oil and Gas Investigations Chart OC-85.

Kindle, E.M., 1917, Recent and fossil ripple marks: Canada Geological Survey, Museum Bulletin 25.

Krumbein, W.C., and Sloss, L.L., 1963, Stratigraphy and sedimentation [2d ed.]: San Francisco, W.H. Freeman and Co., 660 p.

Kubic, W., 1993, Natural fracturing style and control on Devonian shale gas production, Pike County, Kentucky: Gas Shales Technology Review, v. 8, no. 2, p. 1–25.

Larese, R.E., 1974, Petrology and stratigraphy of the Berea Sandstone in the Cabin Creek and Gay-Fink trends, West Virginia: Morgantown, West Virginia University, doctoral dissertation, 245 p.

Leblanc, R.J., 1975, Significant studies of modern and ancient deltaic sediments, *in* Broussard, M.L., ed., Deltas—Models for exploration: Houston, Texas, Houston Geological Society, p. 13–85.

Leverett, F., 1929, The Pleistocene of northern Kentucky: Kentucky Geological Survey, ser. 6, v. 31, p. 1–80.

Lewis, T.L., 1988, Late Devonian and Early Mississippian distal basin-margin sedimentation of northern Ohio: Ohio Journal of Science, v. 88, p. 23–39.

Lierman, R.T., and Mason, C.E., 1992, Farmers, Nancy, and Cowbell Members of the Borden Formation (Lower Mississippian) along Kentucky State Route 546 in northeastern Kentucky, *in* Ettensohn, F.R., ed., Changing interpretations of Kentucky geology—Layer-cake, facies, flexure, and eustacy: Ohio Division of Geological Survey Miscellaneous Report No. 5, p. 138–141.

Lierman, R.T., and Mason, C.E., 2007, Upper Devonian glaciation in the Ohio Shale of east-central Kentucky [abs.]: Geological Society of America Abstracts with Programs, v. 39, p. 70.

Mason, C.E., 2002, Sedimentology and paleoecology of the Crab Orchard Formation (Silurian, Llandovery) at Charters, Kentucky, *in* Algeo, T.J., and Brett, C.E., eds., Sequence, cycle and event stratigraphy of Upper Ordovician and Silurian strata of the Cincinnati Arch region: Kentucky Geological Survey, ser. 12, Guidebook 1, p. 132–135.

Mason, C.E., and Chaplin, J. R., 1979, Stop 2: Nancy and Cowbell Members of the Borden Formation, *in* Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian Basin to the Illinois Basin through eastern Ohio and Kentucky (guidebook for Field Trip No. 4, IX International Congress of Carboniferous Stratigraphy and Geology): Lexington, University of Kentucky, p. 147–151.

Mason, C.E., and Kammer, T.W., 1984, Dysaerobic faunas of the Borden Formation, in Kentucky and Indiana [abs.]: Geological Society of America Abstracts with Programs, v. 16. p. 178.

Mason, C.E., and Lierman, R.T., 1985, Stratigraphy and depositional environments of the Sunbury and lower Borden formations (Lower Mississippian) in northeastern Kentucky, STOP 4, Cave Run Lake spillway section, *in* Haban, S.M., ed., Field excursion 6, guidebook, Carboniferous of eastern Kentucky: Sixth Gondwana Symposium: Ohio State University Institute of Polar Studies Miscellaneous Publication 228, p. 85–95.

Mason, C.E., and Lierman, R.T., 1992, Henley Bed (Farmers Member) of the Borden Formation (Lower Mississippian) along Kentucky State Route 546 in northeastern Kentucky, *in* Ettensohn, F.R., ed., Changing interpretations of Kentucky geology—Layer-cake, facies, flexure, and eustacy: Ohio Division of Geological Survey Miscellaneous Report No. 5, p. 135–138.

Matchen, D.L., and Vargo, A.G., 1996, PLAY Mws: Lower Mississippian Weir Sandstones, *in* Roen, J.B., and Walker, B.J., eds., The atlas of major Appalachian gas plays: West Virginia Geological and Economical Survey Publication V-25, p. 46–50.

McDowell, R.C., 1975, Geologic map of the Farmers quadrangle, east-central Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1236, scale 1:24,000.

McDowell, R.C., 1976, Geologic map of the Colfax quadrangle, east-central Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-13, scale 1:24,000.

Mii, H., Grossman, E.L., and Yancy, T.E., 1999, Carboniferous isotope stratigraphies of North America: Implications for Carboniferous paleoceanography and Mississippian glaciation: Geological Society of America Bulletin, v. 111, p. 960–973.

Moore, B.R., and Clarke, M.K., 1970, The significance of a turbidite sequence in the Borden Formation (Mississippian) of eastern Kentucky and southern Ohio, *in* Lajoie, J., ed., Flysch sedimentology in North America: Geological Association of Canada Special Paper 7, p. 211–218.

Morris, R.H., 1965, Geology of the Charters quadrangle, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-293, scale 1:24,000.

Morris, R.H., and Pierce, K.L., 1967, Geologic map of the Vanceburg quadrangle, Kentucky-Ohio: U.S. Geological Survey Geologic Quadrangle Map GQ-2598, scale 1:24,000.

Morse, W.C., and Foerste, A.F., 1909, The Bedford fauna at Indian Fields and Irvine, Kentucky: Ohio Naturalist, v. 9, no. 7, p. 515–523.

Nance, R.D., and Linnemann, U., 2008, The Rheic Ocean: Origin, evolution, and significance: GSA Today, v. 18, p. 4–12.

Normark, W.R., 1978, Fan valleys, channels, and depositional lobes on modern submarine fans: Characters for recognition of sandy turbidite environments: American Association of Petroleum Geologists Bulletin, v. 62, p. 912–931.

Over, D.J., Lazar, R., Baird, G.C., Schieber, J., and Ettensohn, F.R., 2009, *Protosalvinia* Dawson and associated conodonts of the upper Trachytera Zone, Famennian, Upper Devonian, in the eastern United States: Journal of Paleontology, v. 83, p. 70–79.

Pashin, J.C., 1985, Paleoenvironmental analysis of the Bedford-Berea sequence, northeastern Kentucky and south-central Ohio: Lexington, University of Kentucky, master's thesis, 105 p.

Pashin, J.C., 1990, Reevaluation of the Bedford-Berea sequence in Ohio and adjacent states: New perspectives on sedimentation and tectonics in foreland basins: Lexington, University of Kentucky, doctoral dissertation, 412 p.

Pashin, J.C., and Ettensohn, F.R., 1987, An epeiric shelf-to-basin transition: Bedford-Berea sequence, northeastern Kentucky and south-central Ohio: American Journal of Science, v. 287, p. 893–926.

Pashin, J.C., and Ettensohn, F.R., 1992a, Lowstand deposition in a foreland basin: Bedford-Berea sequence (Upper Devonian), eastern Kentucky and West Virginia, *in* Ettensohn, F.R., ed., 1992, Changing interpretations of Kentucky geology—Layer-cake, facies, flexure, and eustacy: Ohio Department of Natural Resources, Division of Geological Survey, Miscellaneous Report 5, p. 123–134.

Pashin, J.C., and Ettensohn, F.R., 1992b, Paleoecology and sedimentology of the dysaerobic Bedford fauna (Late Devonian), Ohio and Kentucky, U.S.A.: Palaeogeography, Paleoclimatology, Palaeoecology, v. 91, p. 21–34.

Pashin, J.C., and Ettensohn, F.R., 1995, Reevaluation of the Bedford-Berea sequence in Ohio and adjacent states: Forced regression in a foreland basin: Geological Society of America Special Paper 298, 68 p.

Peck, J.H., 1967, Geologic map of the Tollesboro quadrangle, Lewis and Fleming Counties, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-661, scale 1:24,000.

Pedersen, T.F., and Calvert, S.E., 1990, Anoxia vs. productivity: What controls the formation of organic-rich sediments and sedimentary rocks?: American Association of Petroleum Geologists Bulletin, v. 74, p. 454–466.

Pepper, J.F., DeWitt, W., Jr., and Demarest, D.F., 1954, Geology of the Bedford Shale and Berea Sandstone in the Appalachian Basin: U.S. Geological Survey Professional Paper 259, 111 p.

Perkins, R.B., Piper, D.Z., and Mason, C.E., 2008, Trace-element budgets in the Ohio/Sunbury shales of Kentucky: Constraints on ocean circulation and primary productivity in the Devonian-Mississippian Appalachian Basin: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 265, p. 14–29.

Philley, J.C., Hylbert, D.K., and Hoge, H.P., 1974, Geologic map of the Cranston quadrangle, northeastern Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1212, scale 1:24,000.

Phillips, T.L., Niklas, K.J., and Andrews, W.N., 1972, Morphology and vertical distribution of *Protosalvinia (Foerstia)* from the New Albany Shale (Upper Devonian): Reviews in Paleobotany and Palynology, v. 14, p. 171–196.

Potter, P.E., and Pettijohn, F.J., 1977, Paleocurrents and basin analysis [2d ed.]: Berlin, Springer-Verlag, 425 p.

Potter, P.E., DeReamer, J., Jackson, D., and Maynard, J.D., 1983, Lithologic and paleoenvironmental atlas of Berea Sandstone in the Appalachian Basin: Appalachian Geological Society, Special Publication 1, 157 p.

Provo, L.J., Kepferle, R.C., and Potter, P.E., 1978, Division of the black Ohio Shale in eastern Kentucky: American Association of Petroleum Geologists Bulletin, v. 62, p. 1703–1713.

Reger, D.B., 1927, Pocono stratigraphy in the Broadtop Basin of Pennsylvania: Geological Society of America Bulletin, v. 38, p. 397–410.

Rich, J.L., 1951a, Three critical environments of deposition and criteria for the recognition of rocks deposited in each of them: Geological Society of America Bulletin, v. 62, p. 481–533.

Rich, J.L., 1951b, Probable fondo origin of Marcellus-Ohio-New Albany-Chattanooga bituminous shales: American Association of Petroleum Geologists Bulletin, v. 35, p. 2017–2040.

Rittenhouse, G., 1946, Map showing distribution of several types of Berea Sandstone in West Virginia, eastern Ohio, and western Pennsylvania: U.S. Geological Survey Oil and Gas Investigations, Preliminary Map 58.

Robinson, P., Tucker, R.D., Bradley, D., Berry., H.N., IV, and Osberg, P.H., 1998, Paleozoic orogens in New England: GFF, v. 120, p. 119–148.

Romankiw, L.A., Hatcher, P.G., and Roen, J.B., 1988, Evidence of land plant affinity for the Devonian fossil *Protosalvinia (Foerstia):* Lethaia, v. 21, p. 417–424.

Ross, C.A., and Ross, J.R.P., 1988, Late Paleozoic transgressive-regressive deposition, *in* Wilgus, C.K., Hastings, B.S., Posamentier, H.W., Ross, C.A., and Van Wagoner, J.C., eds., Sea-level changes: An integrated approach: SEPM Special Paper 42, p. 227–247.

Rothman, E.M., 1978, The petrology of the Berea Sandstone (Early Mississippian) of south-central Ohio and a portion of northern Kentucky: Oxford, Ohio, Miami University, master's thesis, 105 p.

Sandberg, C.A., Ziegler, W., Leuteritz, K., and Brill, S.M., 1978, Phylogeny, speciation, and zonation of *Siphonodella* (Conodonta, Upper Devonian and Lower Carboniferous): Newsletters on Stratigraphy, v. 7, no. 2, p. 102–120.

Schopf, J.M., 1978, *Foerstia* and recent interpretations of early vascular land plants: Lethaia, v. 11, p. 139–143.

Schopf, J.M., and Schwietering, J.F., 1970, The *Foerstia* Zone of the Ohio and Chattanooga Shales: U.S. Geological Survey Bulletin, v. 1294-H,15 p.

Schwab, W.C., and Lee, H.J., 1988, Causes of two slope-failure types in continental shelf sediment, northwestern Gulf of Alaska: Journal of Sedimentary Petrology, v. 58, p. 1–11.

Scotese, C.R., 2003, Paleogeographic map archive, PALEOMAP project: Arlington, Department of Geology, University of Texas at Arlington.

Seilacher, A., 1967, Bathymetry of trace fossils: Marine Geology, v. 5, p. 413–428.

Sevon, W.D., 1973, Glaciation and sedimentation in the Late Devonian and Early Mississippian of Pennsylvania [abs.]: Geological Society of America Abstracts with Programs, v. 5, p. 218–219.

Sevon, W.D., 1979, Polymictic diamictites in the Spechty Kopf and Rockwell formations, *in* Dennison et al., eds., Devonian shales in south-central Pennsylvania and Maryland (guidebook for the 44th Annual Field Conference of Pennsylvania Geologists): Harrisburg, Field Conference of Pennsylvania Geologists, p. 61–66, 107–110.

Sevon, W.D., Woodrow, D.I., Costolnick, D.E., Richardson, J.B., and Attrep, M., Jr.. 1997, Convulsive geologic events and the origin of diamictite in the Spechty Kopf Formation in northeastern Pennsylvania, *in* Inners, J.D., ed., Geology of the Wyoming-Lackawanna Valley and its mountain rim, northeastern Pennsylvania: Guidebook for the 62nd Annual Field Conference of Pennsylvania Geologists: Harrisburg, Pennsylvania Geological Survey, p. 34–60.

Shepard, F.P., 1955, Delta-front valleys bordering the Mississippi distributaries: Geological Society of America Bulletin, v. 66, p. 1484–1498.

Sloss, L.L., 1963, Sequences in the cratonic interior of North America: Geological Society of America Bulletin, v. 74, p. 93–114.

Smith, K.L., Jr., Robison, B.H., Helly, J.J., Kaufman, R.S., Ruhl. H.A., Shaw, T.J., Twining, B.S., and Vernet, M., 2007, Free-drifting icebergs: Hot spots of chemical and biological enrichment in the Weddell Sea: Science, v. 317, no. 5837, p. 478–482.

Swager, D.R., 1978, Stratigraphy of the Upper Devonian–Lower Mississippian shale sequence in eastern Kentucky outcrop belts: Lexington, University of Kentucky, master's thesis, 116 p.

Thomas, W.A., 1991, The Appalachian-Ouachita rifted margin of southeastern North America: Geological Society of America Bulletin, v. 103, p. 415–431.

Tomastik, T.E., 1996, PLAY MDe: Lower Mississippian–Upper Devonian Berea and equivalent sandstones, *in* Roen, J.B., and Walker, B.J., eds., The atlas of major Appalachian gas plays: West Virginia Geological and Economical Survey Publication V-25, p. 56–62.

Wanless, H.R., and others, 1970, Late Paleozoic deltas in the central and eastern United States: Society of Economic Paleontologists and Mineralogists Special Publication 15, p. 215–245.

Walker, R.G., 1979, Turbidites and associated coarse clastic deposits, *in* Walker, R.G., ed., Facies models: Geoscience Canada Reprint Series 1, p. 91–107.

Wallace, L.G., Roen, J.B., and DeWitt, W., Jr., 1977, Preliminary stratigraphic cross section showing radioactive zones in the Devonian black shales in the western part of the Appalachian Basin: U.S. Geological Survey Oil and Gas Investigations Chart OC-80.

Woodrow, D.L., Dennison, J.M., Ettensohn, F.R., Sevon, W.T., and Kirchgasser, W.T., 1988, Middle and Upper Devonian stratigraphy and paleogeography of the central and southern Appalachians and eastern Midcontinent, U.S.A., *in* McMillan, N.J., Embry, A.F., and Glass, G.J., eds., Devonian of the world, Proceedings of the Second International Symposium on the Devonian System: Canadian Society of Petroleum Geologists Memoir 14, v. 1, p. 277–301.

Woodrow, D.L., and Isley, A.M., 1983, Facies, topography and sedimentary processes in the Catskill Sea (Devonian), New York and Pennsylvania: Geological Society of America Bulletin, v. 94, p. 459–470.

Woodrow, D.L., and Richardson, J.B., 2006, What WAS going on in the Late Devonian?: Official Program and Abstracts, 2006 Eastern Section. American Association of Petroleum Geologists 35th Annual Meeting, p. 33.

Work, D.M., and Mason, C.E., 2005, Mississippian (early Osagean) Cave Run Lake ammonoid fauna, Borden Formation, northeastern Kentucky: Journal of Paleontology, v. 79, p. 719–725.

Ziegler, P.A., 1989, Evolution of Laurussia: Dordrecht, Kluwer Academic Publishers, 102 p.