Upper Devonian–Lower Mississippian clastic rocks in northeastern Kentucky: Evidence for Acadian alpine glaciation and models for source-rock and reservoir-rock development in the eastern United States

Frank R. Ettensohn, R. Thomas Lierman, and Charles E. Mason
with contributions by Sarah Heal, Niall Paterson, Cortland Eble, Robbie Goodhue, Nina Larsson, Geoff Clayton, Alan Dennis, Eric Anderson, and D. Brent Wilhelm

American Institute of Professional Geologists—Kentucky Section

Spring Field Trip

April 18, 2009

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Introduction

Situated at the margin of the Bluegrass area in central Kentucky is a dissected escarpment supported by Devonian and Mississippian shales, siltstones, and sandstones, sometimes called the "Highland Rim" or "Knobs." The area is generally transitional between a planar, trough-like area on the Bluegrass Plain and the higher Pottsville escarpment of the Appalachian Plateau. On the eastern side of the Bluegrass area, the Highland Rim section contains 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. 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. 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, 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 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 were abruptly replaced by ichnofauna and a deeper-water, restrictive, commonly planktic, depauperate biota in the 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 did an abundant, diverse, shallow-water, benthic fauna reappear in delta-destruction 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 Upper Devonian–Middle Mississippian, black shale–Borden delta sequence (Fig. 2) in light of its tectonic and environmental origins while at the same time noting its salient engineering and resource-related characteristics. In a new development, we will examine a large glacial dropstone embedded in the black shales and discuss its implications as evidence for the first reported tidewater glaciers and alpine glaciation in an ancient orogen.

Physiography along the field-trip route, I-64 in Bath and Rowan counties, Kentucky

Most of today's trip takes place on an 18-mile-long stretch of eastbound I-64 from the Licking River to the KY State Route 799 overpass in Rowan County, Kentucky (Fig. 1), although to take advantage of exits off and onto the Interstate, the trip will have to go an additional 8 miles to the Owingsville exit 123 and then double back to pick up the first stop just east of the Licking River. Initial parts of the field-trip route on I-64 from the Owingsville exit to just beyond the Licking River are present in the Lexington Plains Section of the Interior Low Plateaus physiographic province (Fenneman, 1938) (Fig. 3). From the Owingsville exit, I-64 crosses the hummocky, irregularly rolling hills and low ridges of the Outer Bluegrass region, developed in this area on the Upper Ordovician limestones and shales of the Bull Fork and Drakes formations and on the Lower Silurian dolostones and shales of the Brassfield and Crab Orchard formations (McDowell, 1975, 1976) (Fig. 2). From the Owingsville exit, the highway drops about 100 feet into a broad trough at the foot of the Knobs, here largely produced by erosion into soft, high-level, fluvial deposits and Crab Orchard shales.

The Knobs rise abruptly in front of the Licking River floodplain and mark the beginning of the Knobs region of the Appalachian Plateau (Fig. 4). The Knobs form the lower part of the west-facing escarpment that defines the western margin of the Appalachian Plateau and marks the beginning of the Highland Rim (Fig. 4). In this area, the Knobs are about 300 ft (91 m) high and form a series of conical hills or detached ridges carved from the escarpment where erosion has cut through Silurian, Devonian, and Mississippian shales, leaving isolated rounded hills separated by wide valleys. East of the Licking River, the broad bases of the Knobs are generally developed in the nonresistant Lower Silurian Crab Orchard and Upper Devonian Ohio shales with resistant caps in the siltstones of Farmers or Cowbell members of the Borden Formation. During the first half day of the trip, we will largely examine rocks exposed where the valley bottoms meet the bases of several knobs. In the last half day of the trip, we will effectively climb about 500 ft (131 m) up the Highland Rim through the Farmers, Nancy, Cowbell, and Nada members of the Borden Formation (Hoge and Chaplin, 1972; Philley *et al.*, 1974). In this part of Kentucky, the Knobs and Highland Rim are included as

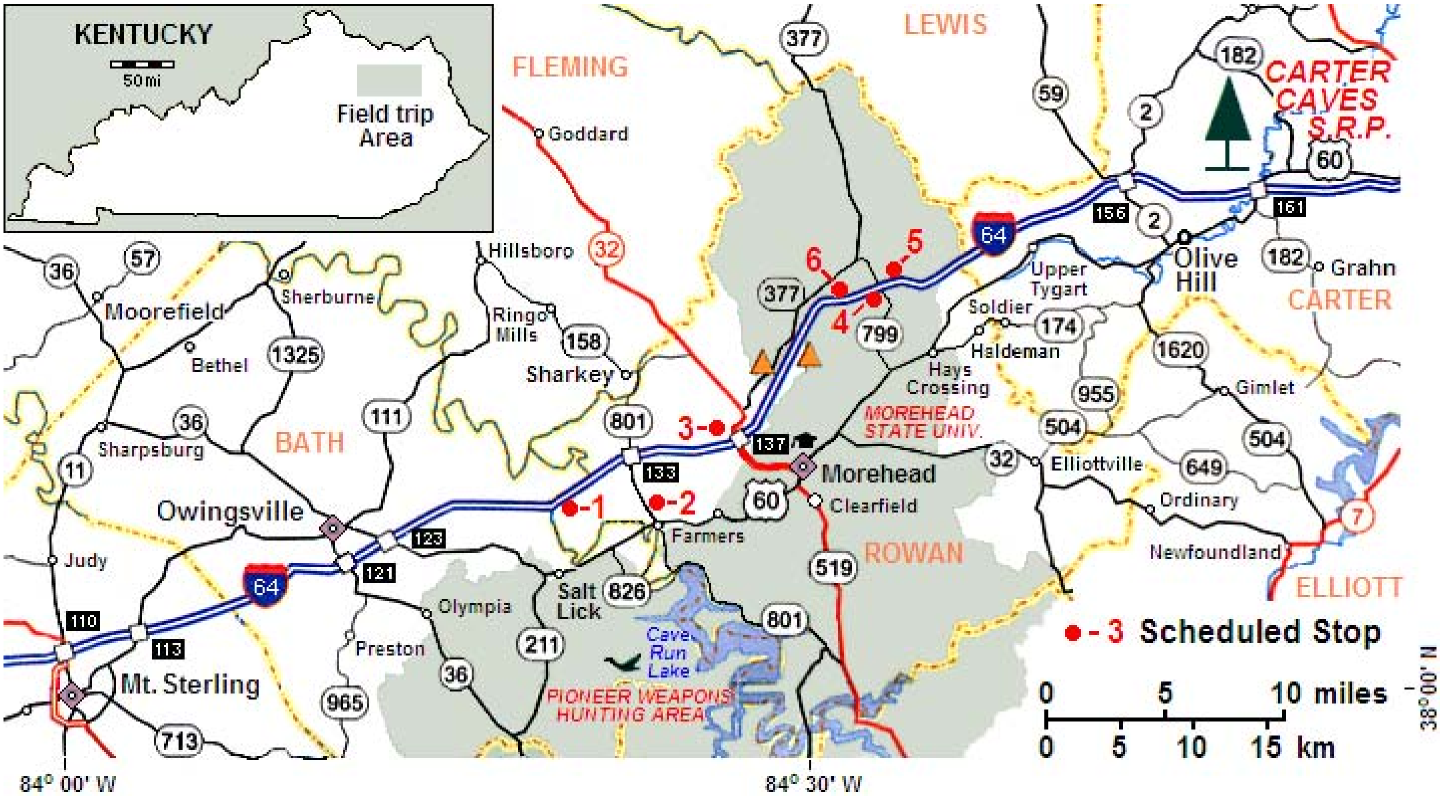


Figure 1
Field-trip stops along and near I-64 in northeastern Kentucky.

parts of the Pottsville Escarpment that defines the western margin of the Appalachian Plateau (Fig. 4). Typically the Pottsville escarpment is a west-facing escarpment defined by resistant Lower Pennsylvanian, conglomeratic sandstones (Pottsville or Lee formations), but because Devonian and Mississippian outcrop belts are relatively narrow on the eastern side of the Cincinnati Arch and do not exhibit topography very different from that of Pennsylvanian rocks to the east, the Knobs and Highland Rim sections are effectively merged to form the Pottsville Escarpment (Ettensohn, 1992a).

Paleogeographic, paleoclimatic, and tectonic frameworks

Excluding the Lower and Middle Silurian Crab Orchard Shale at Stop 1, the ages of the rocks examined on this field trip range from Late Devonian (Famennian; Chatauquan) to Middle Mississippian (Visean; Meramecian) time, and based on the time scale of Gradstein *et al*. (2004) span an approximately 36-Ma interval of time from about 371 Ma to 335 Ma. During this time, North America, together with western Europe, formed the continent Laurussia, and the field trip area was located about 25° south latitude in the subtropical trade-wind belt (Fig. 5), although by Middle Mississippian time the area had moved farther northward into more tropical latitudes (Ziegler, 1989).

The Devonian-Mississippian transition was a time of climatic and tectonic change. The Devonian-Mississippian black shales (Ohio and Sunbury shales) were deposited at the end of a global greenhouse state, a time of warm, equable climate with low latitudinal thermal gradients, high CO_2 concentrations, and elevated sea levels (Fischer, 1984). Expanding seas and global warmth clearly enhanced organic productivity at a time when oxygen was becoming less soluble and may partially explain the great concentrations of organic matter preserved in the black shales that we will examine today (Ettensohn and Barron, 1981; Ettensohn, 1995, 1998). By the Early-Middle Mississippian transition during deposition of the Borden Formation, however, the seas became cooler (Mii *et al.*, 1999) and overall sea levels dropped while exhibiting a pronounced cyclicity (Ross and Ross, 1988), all of which points to the advent of a global icehouse state and the inception of Gondwanan glaciation (Fischer, 1984), which might have begun as early as the Devonian-Mississippian transition (Frakes *et al.*, 1992; Cecil *et al.*, 2004; Brezinski *et al.*, 2008).

At about this same time (360 Ma), Acadian/Neoacadian tectonism also reached a crescendo. Although the Acadian orogeny began late Early Devonian time (~411 Ma) following closure of the Iapetus Ocean and amassing of peri-Gondwanan Carolina and Avalonian terranes along the eastern margin of Laurussia, the main orogeny reflects dextral transpressional accretion of these terranes from the northeast to the southwest onto the southeastern margin of Laurussia (Ettensohn, 2008). The accretion occurred from the northeast to the southwest, perhaps reflecting the fact that the terranes were caught in a pincers movement between Gondwana and the margin of Laurussia during the closure of the Rheic Ocean (Ettensohn, 2008; Nance and Linnemann, 2008) (Fig. 5). As transpressional accretion migrated slowly to the southwest, Laurussian continental promontories (remaining from the Iapetan rifting of Rodinia) were sequentially impacted by the terranes, generating particularly intense deformation at the promontories, such that each of the four Acadian/Neoacadian tectophases reflect intense deformation at a particular promontory (Ettensohn, 1985) (Fig. 6). The deformation generated at each promontory created sufficient deformational loading to support major basin subsidence and accompanying black-shale deposition as well as a succeeding clastic wedge. Inasmuch as the tectonics are transpressional along the margin of Laurussia, both foreland-basin black shales and clastic wedges developed in basins that paralleled the continental margin, and they migrated along basin strike in time (Ettensohn, 1987, 2004). Moreover, during each successive phase of orogeny,

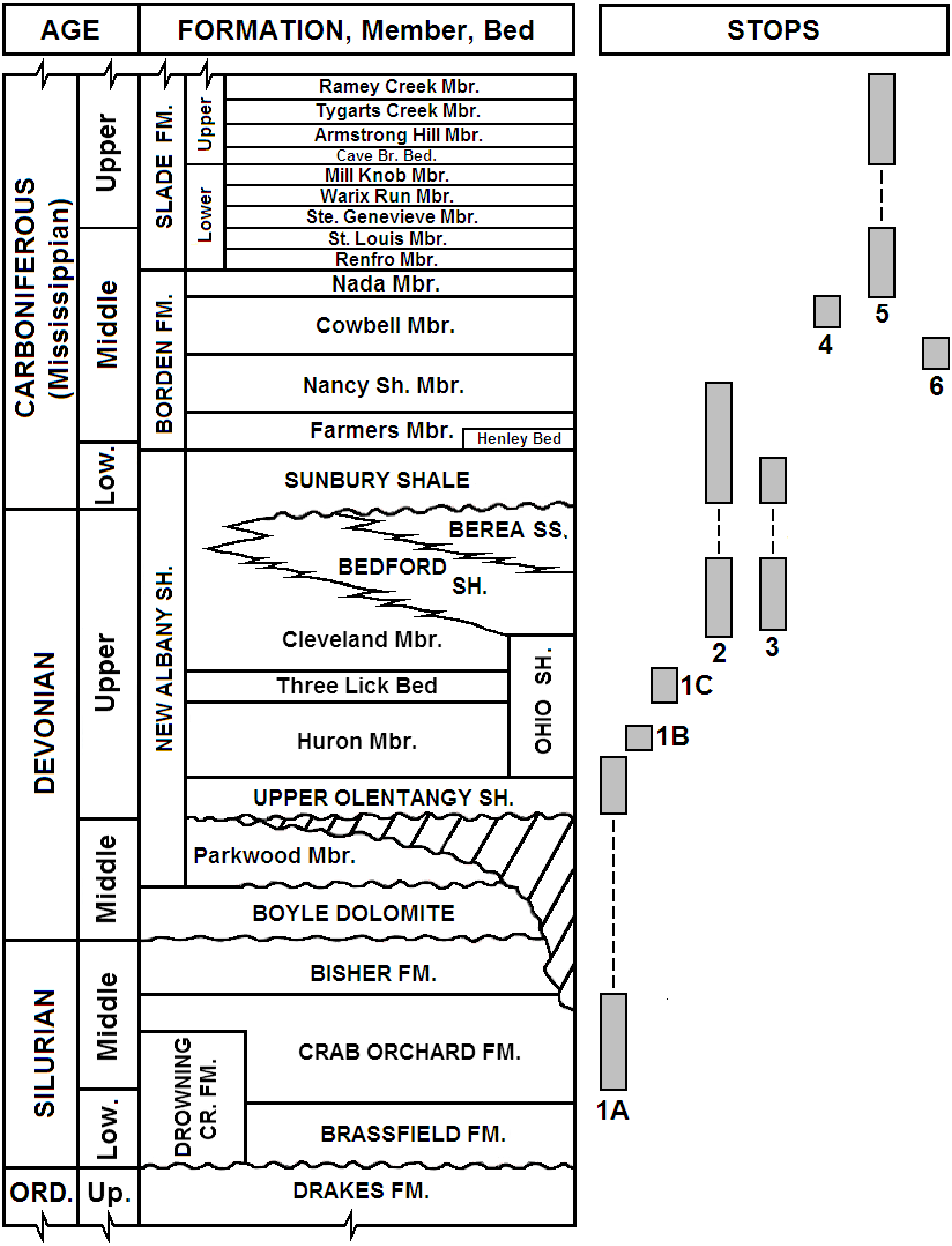


Figure 2
The geologic section along the field-trip route with the relative stratigraphic positions of the sections viewed at each stop.

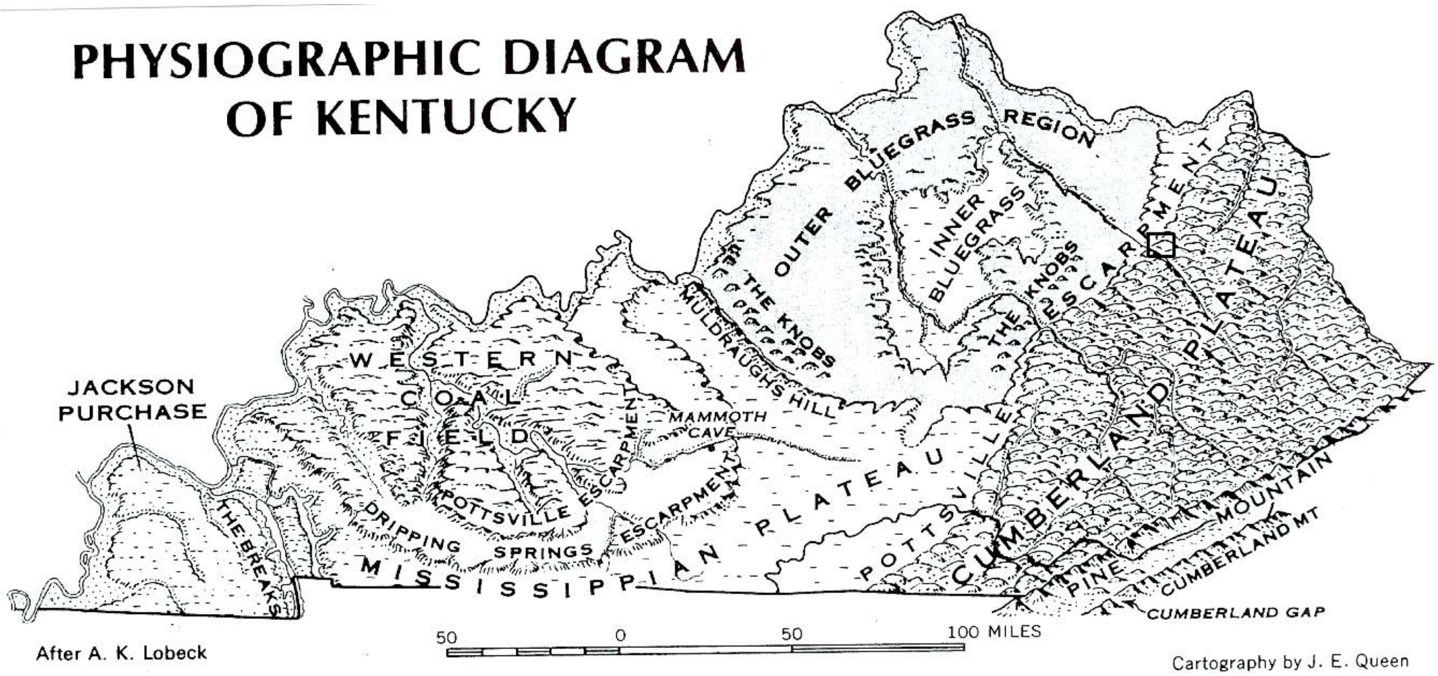


Figure 3
Physiographic diagram of Kentucky showing the location of the field-trip area as a "square" on the Pottsville escarpment in northeastern Kentucky. Areas west of the escarpment are included in Fenneman's (1938) Interior Low Plateaus Province.

black-shale basins also migrated westwardly or cratonward, reflecting the continued cratonward movement of deformation in time. By early Famennian time (about 371 Ma), the Appalachian Basin had filled with black shales and intervening clastic wedges such that subsequent basinal black shales were "pushed out" into cratonic seas just before deposition of the lower Huron Shale Member, as the Appalachian and Illinois basins yoked (Fig. 7) (Ettensohn *et al*., 1988a). We will examine this instant in time at the base of the Ohio Shale at Stop 1.

Transpressional accretion and the cyclic deposition of cratonic black shales and coarser clastics in the east-central United States continued until latest Devonian–Early Mississippian time, when apparently the Carolina, Avalon, and Meguma terranes all collided with Laurussia in the area of the New York promontory (Fig. 8). This event coincides with a Late Devonian–Early Mississippian episode of magmatism, metamorphism, and deformation in southern New England that has been called the Neoacadian orogeny (Robinson *et al*., 1998), although it represents the same event that has been termed the fourth tectophase of the Acadian orogeny (Ettensohn, 1985, 2008). This orogenic event generated sufficient deformational loading that foreland-basin black shales in the form of the Sunbury and Riddlesburg shales migrated back into the Appalachian basin and mountains high enough to support alpine glaciation were generated in the areas of the New York and Virginian promontories (Ettensohn, 2008; Ettensohn *et al.*, 2008). The Sunbury Shale, which we will examine at Stop 2, is probably the most extensive, most organic-rich, and deepest-water shale in the Devonian-Mississippian black-shale sequence, and it is a product of the major deformational loading generated by this orogeny. Moreover, the likely dropstone that we will see embedded in the black shales at Stop 3 is evidence for the presence of alpine glaciation in high mountains generated by this orogeny.

An important part of the sedimentary record produced by this orogeny is a major, post-orogenic clastic wedge that overlies the Sunbury and equivalent shale units and is known variously as the Borden, Price, Pocono, or Grainger "delta complex" (Fig. 7); closely associated with these deltaic clastics are units like the Fort Payne Formation (Ettensohn, 2004). We will examine parts of the Borden Formation at Stops 2, 4, 5, and 6. Most of the coarser clastic units associated with Devonian-Mississippian black shales are restricted to the Appalachian Basin and effectively helped to infill parts of the basin (Fig. 7). The Borden deltaic complex, however, is unusual in that it prograded beyond the Appalachian Basin, across the Cincinnati Arch and into the Illinois Basin (Fig. 9). The Borden Formation represents subaqueous parts of a westwardly prograding delta complex that formed as an early relaxational response to Neoacadian orogeny (Ettensohn *et al*., 2002, 2004; Ettensohn, 1994, 2004, 2005, 2008) (Fig. 10). Borden clastics filled remaining parts of the Appalachian Basin and deeper water seas on adjacent parts of the craton until early Visean (late Osagean) time, when due to bulge moveout and a sea-level lowstand, clastic sedimentation was diverted elsewhere so that a widespread period of sediment starvation ensued across east-central United States (e.g., Ettensohn *et al*., 2002, 2004). This period of sediment starvation is represented by the Floyds Knob Bed (Fig. 10), a thin, but widespread, interval of glaucony and phosphorite deposition, which will be examined at Stop 5. The Floyds Knob Bed commonly occurs within the uppermost unit of the Borden Formation, the Nada Member, which is composed of fossiliferous, blue-green shales with interbedded carbonates and siltstones; it has been interpreted to represent delta destruction *(e.g.,* Ettensohn *et al*., 2004) and will be viewed at Stop 5.

With the deposition of the Nada Member and its equivalents, the progradation of Borden deltaic clastics ended, and the stage was set for the deposition of Middle and Late Mississippian (mid-Visean–mid-Serpukhovian; late Osagean–middle Chesterian) carbonates throughout the Appalachian Basin (Fig. 10). The filling of the basin with Early and Middle Mississippian clastics, together with a sea-level lowstand in a subtropical setting, generated a widespread, shallow-water platform across the basin on which units like the Slade, Greenbrier, Newman, and Monteagle/Bangor/Tuscumbia limestones were deposited, locally with thicknesses greater than 500 m. About 13.5 m of shallow-

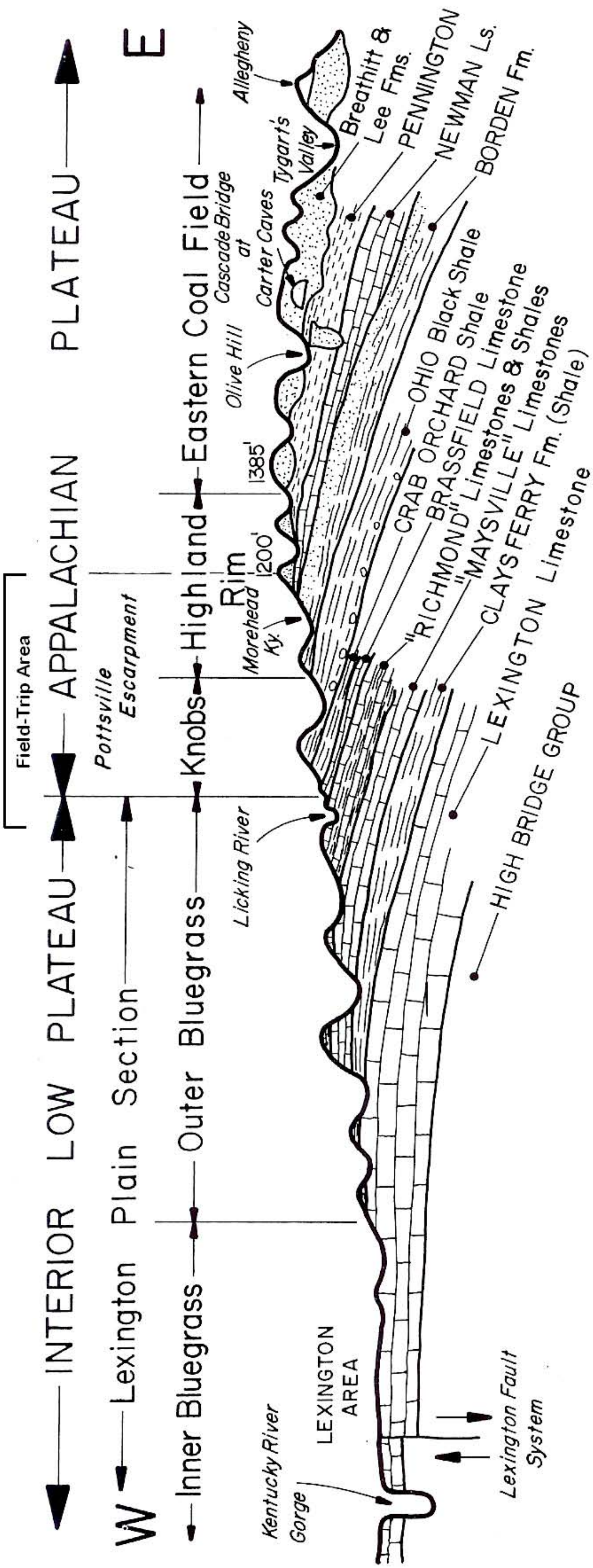


Figure 4
Schematic cross section near I-64 through the east flank of the Cincinnati Arch from Grayson to Lexington, Kentucky, showing major rock units and the topography developed on them; field-trip area in bracket at top right (adapted from Ettensohn and Dever, 1979).

water Slade carbonates overlie the Nada Member of the Borden Formation at Stop 5; more information on the Slade carbonates can be found in Ettensohn and Dever (1979), Ettensohn (1980, 1981, 1992d, e), and Ettensohn *et al*. (2004).

Field-trip roadlog and stops

Mileage

0.0	The field trip will start at 8:30 A.M. in the parking lot behind Hardee's and in front of Food Lion Food store.
0.1	Proceed from parking lot to Pine Crest Lane and follow it to the junction of State Route 32.
0.3	Turn left onto State Route 32 and proceed north to traffic light.
0.35	Center of I-64 overpass, Farmers Member of the Borden Formation exposed on both sides of the road in road cuts.
0.4	At light turn left onto the I-64 westbound ramp.
0.8	I-64 west.
1.0	Farmers Member of the Borden Formation outcrops on both side of the road in road cuts.
1.4	Cross North Fork of Triplett Creek.
2.1	Upper Sunbury Shale, as well as the Farmers and lower part of the Nancy members of the Borden Formation, exposed for 0.3 miles on both sides of I-64 in road cuts.
2.6	Upper Sunbury, as well as the Farmers and lower part of the Nancy members of the Borden Formation, exposed for 0.2 miles on both sides of I-64 in road cuts.
2.95	Cross Bull Fork Creek.
3.3	Upper Sunbury Shale, as well as the Farmers and lower part of the Nancy members of the Borden Formation, exposed for 0.4 miles on both sides of I-64 in road cuts.
4.3	Nancy Member of the Borden Formation exposed for 0.2 miles in road cuts, especially along the westbound lanes.
4.8	State Route 801 overpass.
5.0	Top of Ohio Shale, Bedford Shale, Sunbury Shale, as well as the Farmers and lower part of the Nancy members of the Borden Formation, are exposed for 0.3 miles along both sides of the Interstate.
5.4	The Three Lick Bed of the Ohio Shale is exposed in a road cut along the westbound lanes.
5.7	Ohio Shale is exposed in a road cut along the eastbound lanes.
6.0	Ohio Shale is exposed in a road cut along the eastbound lanes.
6.3	Ohio Shale is exposed in a road cut along the eastbound lanes.
7.0	Ohio Shale is exposed for 0.4 miles in a road cut along the eastbound lanes. The Olentangy and Crab Orchard formations are exposed in the same road cut for the last 0.15 miles at its western end.
7.4	State Route 1722 passes over I-64.
8.8	Licking River Bridge.
9.2	State Route 219 passes over I-64.
11.5	Crab Orchard Formation exposed in road cut along the westbound lanes for 0.1 mile.

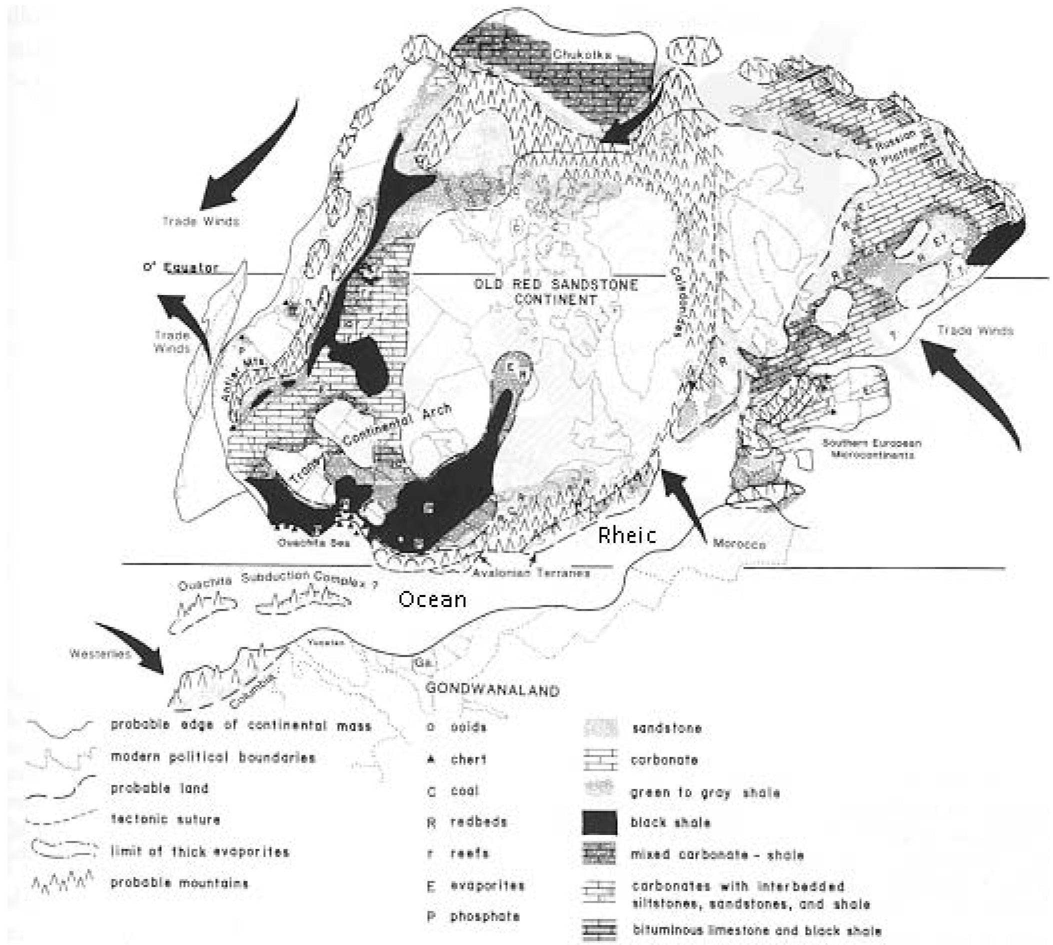


Figure 5
Late Devonian paleogeographic reconstruction of Laurussia (Old Red Sandstone Continent), showing the location of black-shale seas (adapted from Ettensohn and Barron, 1981).

12.8 Feaping Road, a county road crossing over I-64.
13.3 Crab Orchard Formation exposed in road cut along the westbound lanes.
13.7 Crab Orchard Formation exposed for 0.4 miles in road cuts along the westbound lanes of I-64. Note that the underlying Brassfield Formation is exposed in the western part of the exposure.
14.4 Exit 123; turn right onto exit ramp and follow to junction of US 60.
14.8 Junction of exit ramp and US 60; turn left onto US 60.
14.9 Entrance ramp for I-64 east; turn left onto the ramp and proceed east to I-64. The upper part of the Bull Fork Formation and the overlying Preachersville Member of the Drakes Formation are exposed along the eastbound exit ramp and along the right side of US 60, continuing east.
15.1 I-64 east.
15.5 Top of the Brassfield Formation and the overlying Crab Orchard Formation are exposed for 0.2 miles along the eastbound lanes of I-64.
16.1 Crab Orchard Formation exposed in road cut along eastbound lanes of I-64.
16.7 Feaping Road, a county road crossing over I-64.
18.0 Crab Orchard Formation exposed in road cut along the westbound lanes of I-64.
20.4 State Route 211 passes over I-64.
20.8 Licking River Bridge.
22.2 State Route 1722 passes over I-64. Also beginning of road cut along eastbound lanes for 0.3 mile, where the Crab Orchard, Olentangy, and lower part of the Ohio Shale are exposed.

STOP 1: Silurian-Devonian contact and lower part of the Devonian black-shale section, northeastern Kentucky

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

Stop 1 is a three-part stop beginning at what is commonly called the "Morehead exposure," listed below as Stop 1A. 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, 1992b, c; Lierman *et al*., 1992). The entire section is shown schematically in Figure 11, 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 *et al*., 1979).

Mileage

22.3 **Stop 1A**: 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: 83°, 35', 44.4".

The section at Stop 1 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 dolosil-

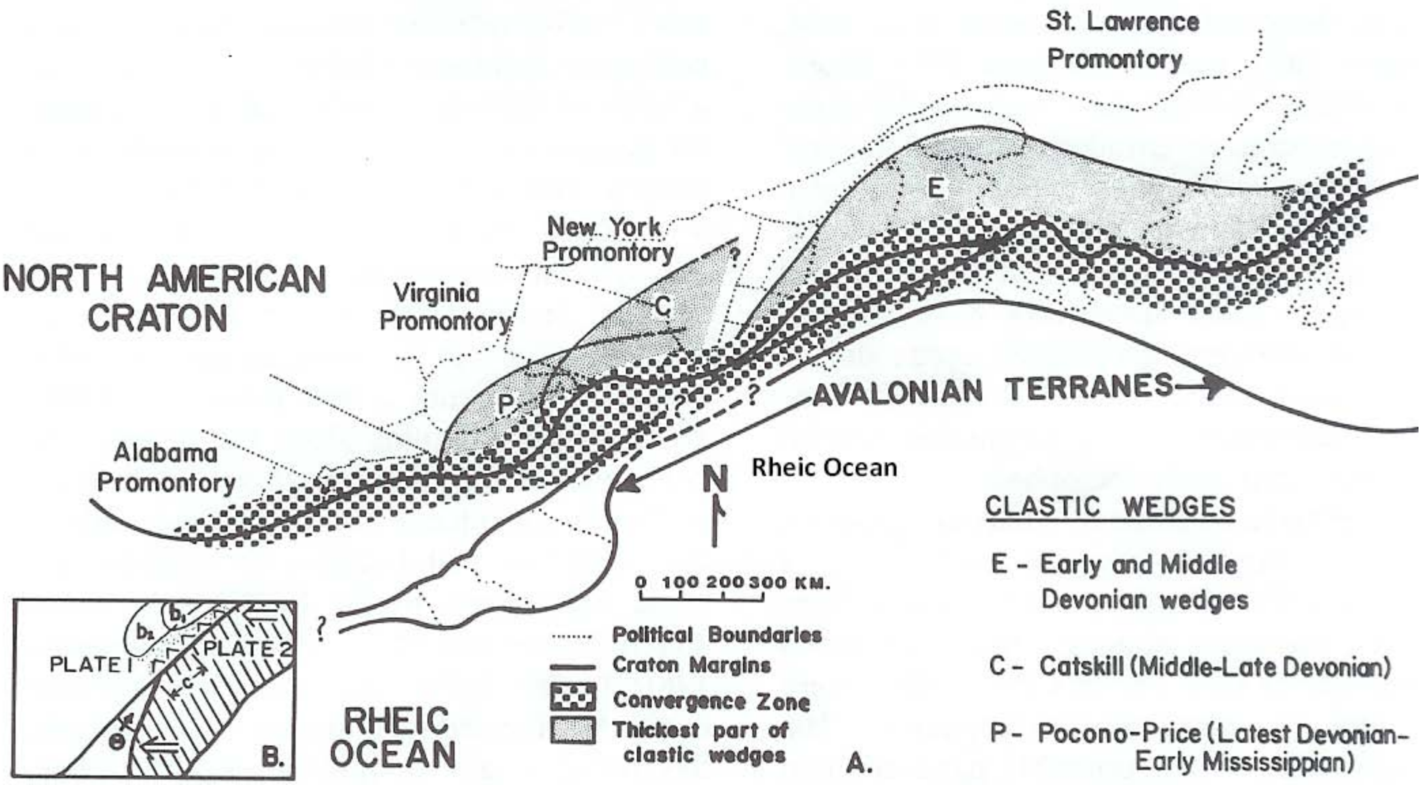


Figure 6
Schematic diagram showing tectonic framework of the Acadian/Neoacadian orogeny: A. Diachronous oblique transpression between Avalonian terranes and Carolina and southeastern margin of Laurussia; note that clastic wedges emanate successively from each promontory, and the Pocono-Price wedge is the subaerial part of the Borden delta; B. Geometry of oblique transpression (adapted from Ettensohn, 1987).

tites to dolarenites with occasional lenses of quartzose sandstone and siltstone (Fig. 12). Much of the shale is reported to contain volcanic ash (Mason *et al*., 1992; Mason, 2002). The dolostones occur in packets of three to five beds that are each typically 1.5–7.0 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) are intensely weathered by oxidation to light brown to dark yellowish-orange colors, although iron oxides have accumulated in the shales to depths of at least 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 disconformity that separates rocks of the Silurian and Devonian systems (Figs. 11, 12), 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 *et al*., 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 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, 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, brachio-

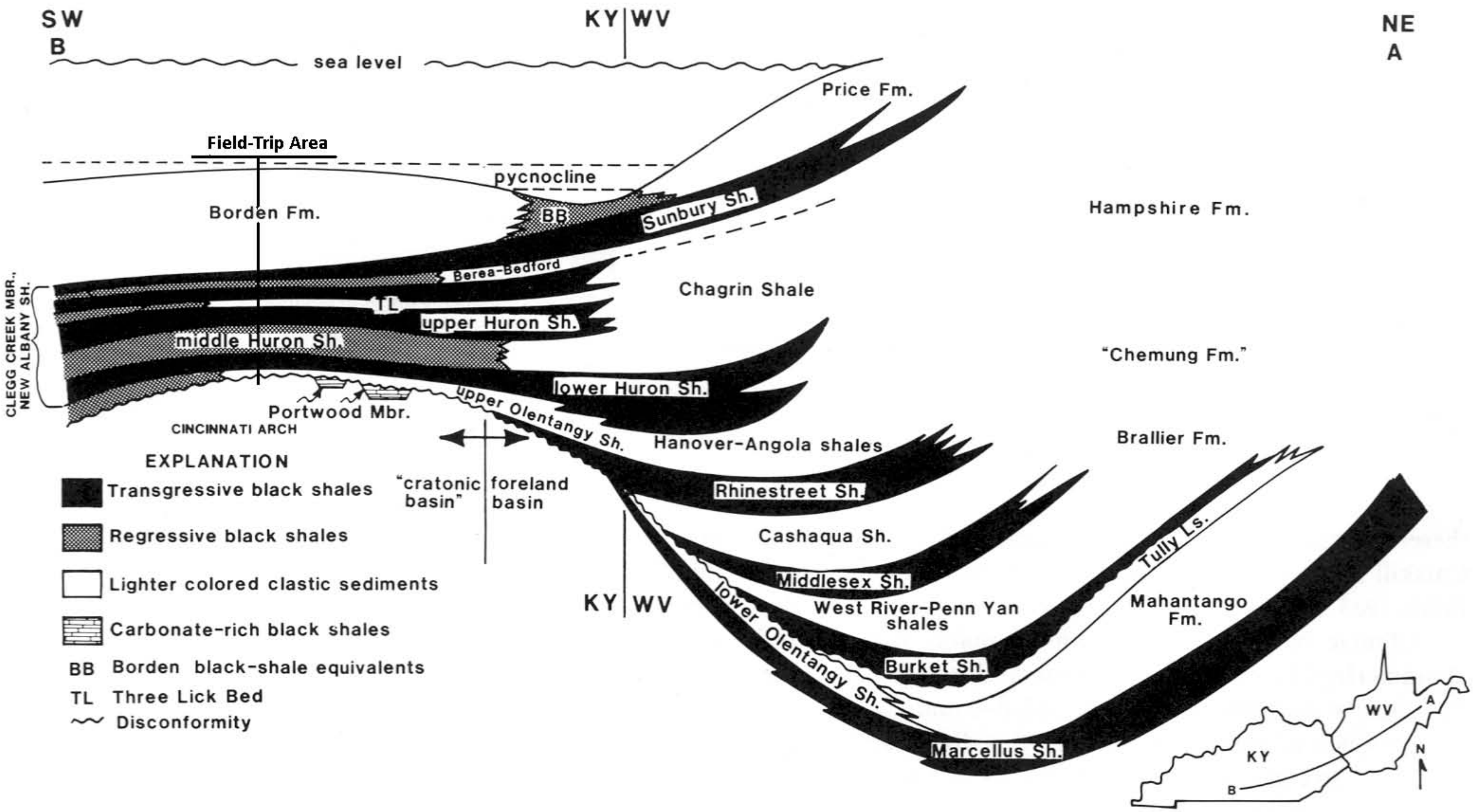


Figure 7
Schematic southwest-northeast section across the central Appalachian Basin in Kentucky and West Virginia, showing the locations of cyclic Middle Devonian to Early Mississippian black-shale basins and intervening clastic wedges formed in the second to fourth Acadian tectophases. Each black shale unit represents subsidence in response to an episode of cratonward-migrating tectonism. The dark vertical line on the left-hand side of the diagram represents the approximate position of the field-trip area (adapted from Ettensohn et al., 1988).

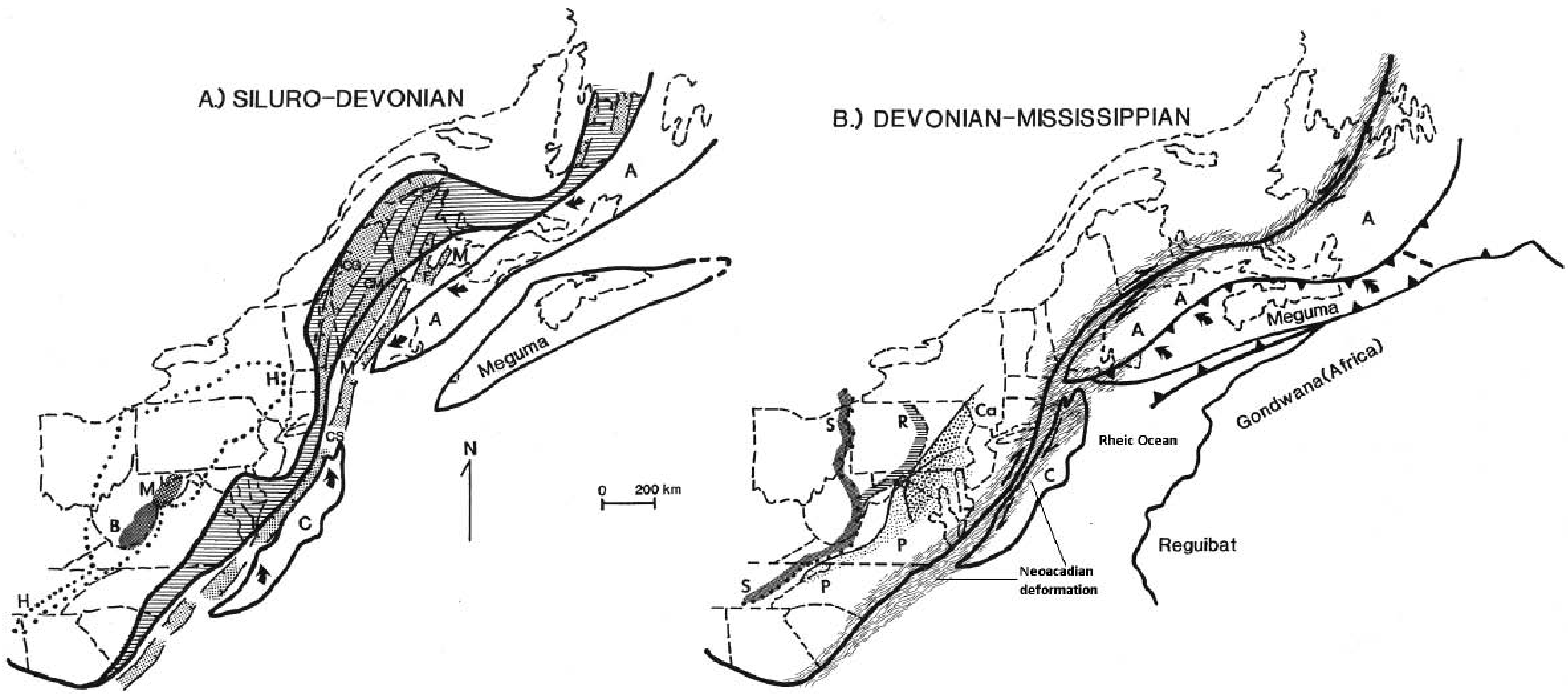


Figure 8
Neoacadian tectonic situation that generated the Sunbury Shale (Stop 2), Borden delta (Stops 2, 4, 5, and 6), and formed a mountain belt high enough to support alpine glaciation (Stop 3). A. Likely pre-Acadian disposition of microplates relative to Laurussia; B. Likely disposition of microplates and Gondwana during Neoacadian orogeny. Note the eastward limit of the deep-marine Sunbury Shale (see Fig. 7) and its marginal-marine equivalent, the Riddlesburg Shale (adapted from Ettensohn, 2008).

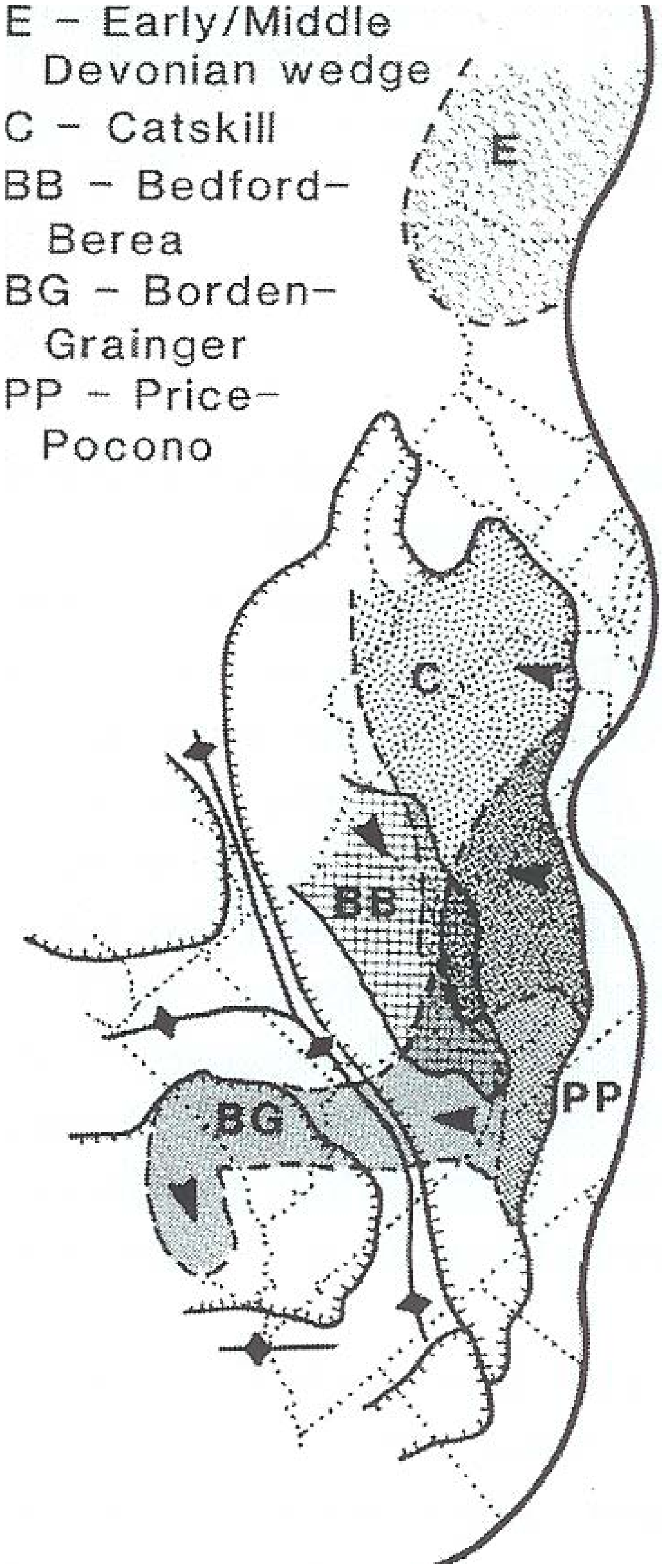


Figure 9
Relative distribution of Acadian/Neoacadian clastic-wedge, delta complexes on southern Laurussia. The delta complexes migrated south in time tracking the progress of Acadian/Neoacadian transpression. The Borden-Grainger (BG) complex is the subaqueous equivalent of the Price-Pocono (PP) subaerial delta; it is the product of the Neoacadian orogeny and was the last and largest of the Acadian/Neoacadian deltas. Major basins are defined by inward-pointing tick marks; intervening arches by lines with diamonds (after Ettensohn, 2004).

pods, and crinoids predominate. More information on this unit and its faunas can be found in studies by Mason *et al*. (1992) and Mason (2002).

The disconformity separating Silurian and Devonian rocks in this section is one of the most significant unconformities in Kentucky (Fig. 12), 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 moveout. 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 *et al*., 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, 1992b).

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. 11, 12). 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 *et al*., 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 *et al*., 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 *et al*., 1989; Ettensohn, 1992b; Fuentes *et al*., 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 *et al*., 1988; Gradstein *et al*., 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, 1992b). 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. 7).

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. 11, 12). This part of the Ohio Shale has been called the lower Huron Shale Member and it has been cor-

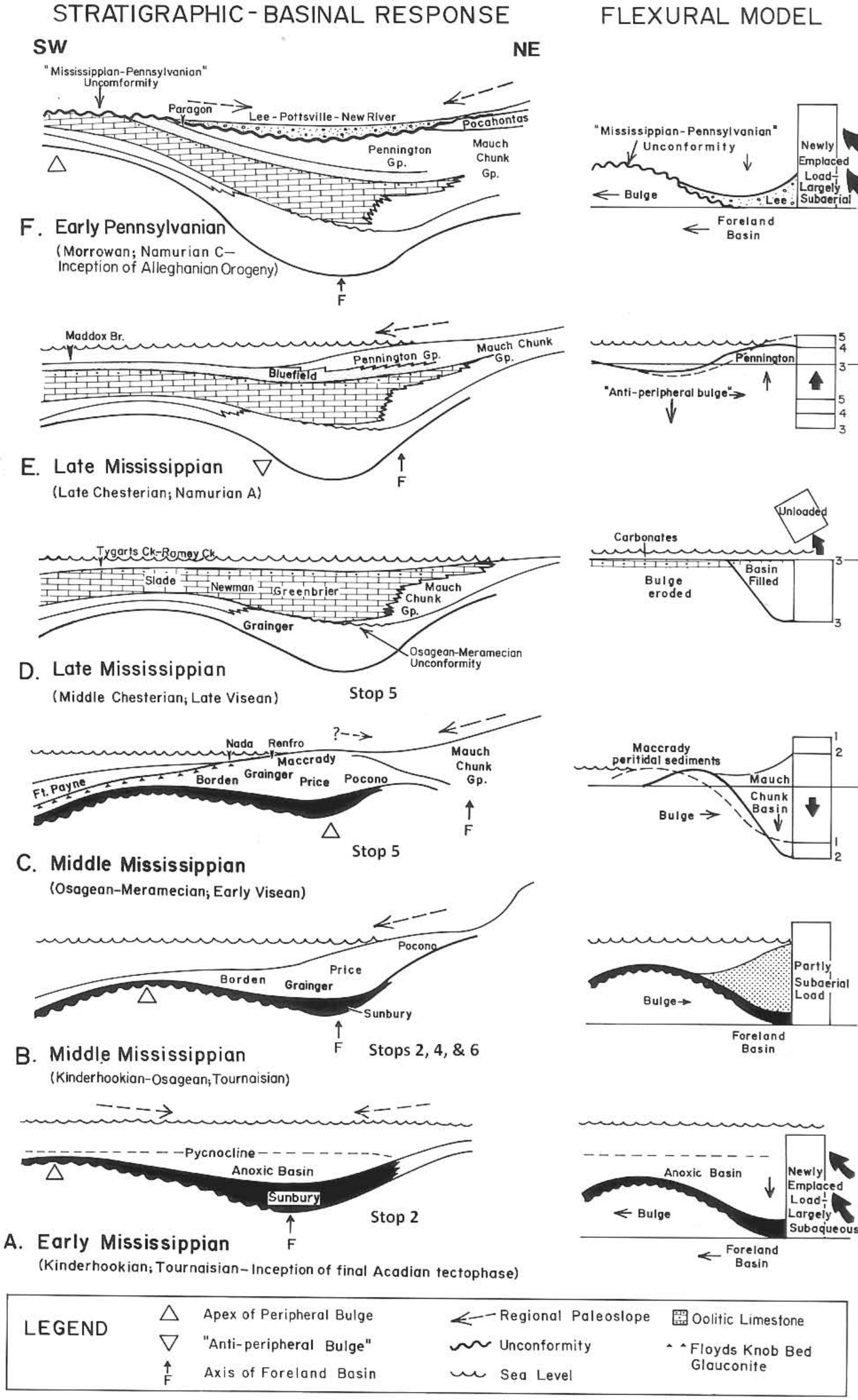


Figure 10
Schematic east-west cross section across the central Appalachian basin, showing likely succession of flexural events between the last Acadian/Neoacadian tectophase (A) and inception of the Alleghanian orogeny (F) and their sedimentary/stratigraphic responses relative to units in the central Appalachian Basin and along the field-trip route. On the field-trip route we will examine units from events A through D (adapted from Ettensohn 1994, 2004).

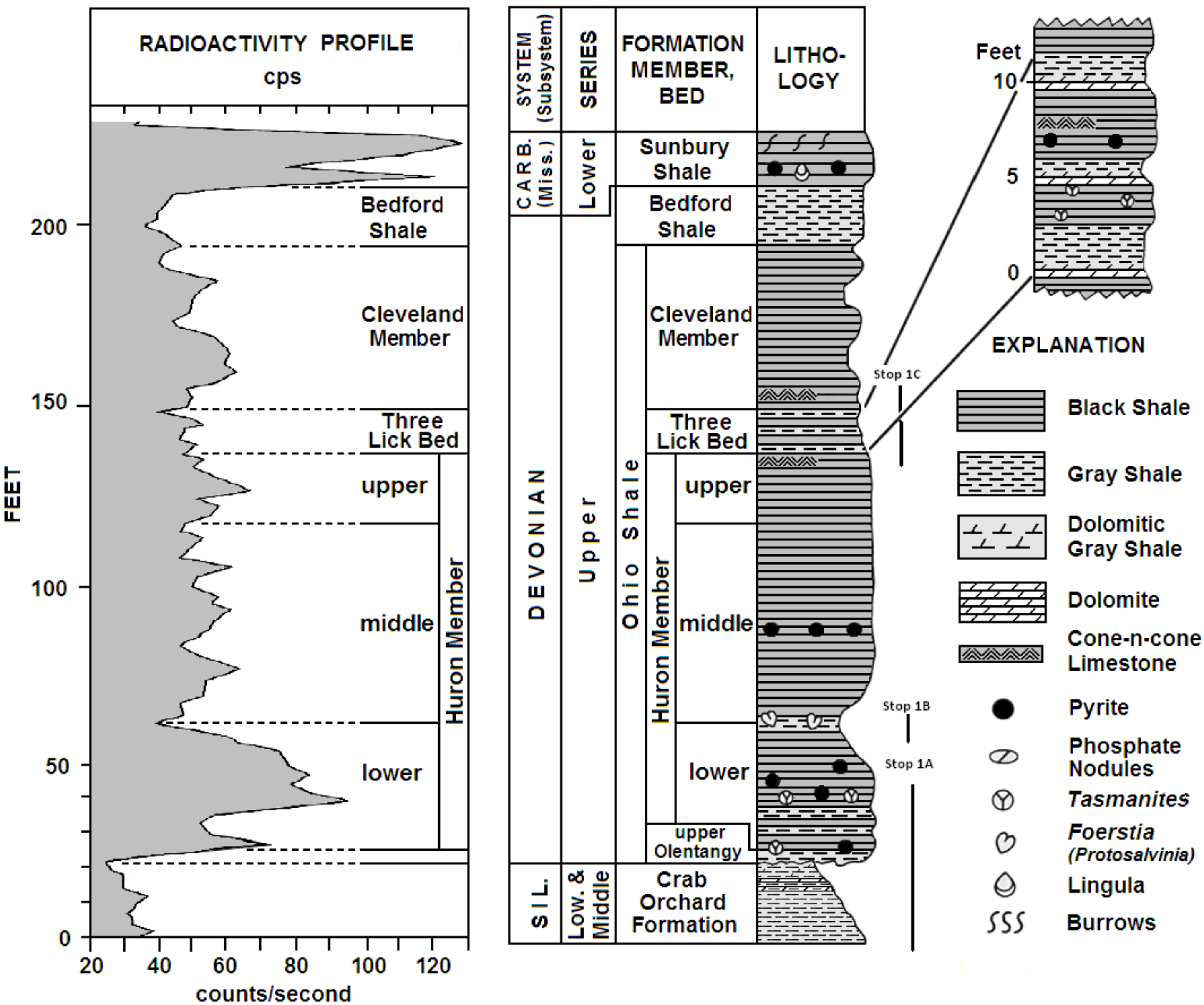


Figure 11
Schematic stratigraphic column for the section at Stop 1, showing the respective stratigraphic disposition of stops in the section. The radioactivity profile to the left is an artificial gamma-ray log that can be used for correlations in the surface or subsurface in homogeneous black shales.

related with the Dunkirk black shales of New York (*e.g.*, Wallace *et al*., 1977), although like the Olentangy and Hanover shales, they are homotaxial, but diachronous, by about three conodont zones or 3 Ma (Woodrow *et al*., 1988; Ettensohn *et al*., 1989; Ettensohn, 1992b; Fuentes *et al*., 2002; Gradstein *et al*., 2004). The lower Huron is a typical transgressive black shale (Fig. 7), being more widespread, more organic-rich, and more radioactive than the overlying, regressive black shales of the middle Huron Member (Ettensohn, 1992b) (Fig. 11). The lower Huron forms the large positive deviation at the base of the black-shale sequence (Fig. 11), and because of its high organic content (8–23 percent by weight), its shales are typically "pulpy" and weather to form very fissile "paper shales" like those exposed at this stop (Fig. 13). Interpretations based on geochemistry (*e.g.*, Perkins *et al*., 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.

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 *Tasman-*

Figure 12
The section a Stop 1A (Fig. 11), showing the Crab Orchard, upper 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.

Figure 13
"Pulpy" paper shales, or transgressive black shales, from the lower part of the Huron Member of the Ohio Shale at Stop 1a.

ites 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 that 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.

Mileage

22.5 **Stop 1B:** I-64 along the eastbound lane 0.2 mi east of Stop 1A. 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: 83°, 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. 11). 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 clastic constituents, mainly in the form of quartzose laminae, and less organic matter (4–7 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. 7). 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 *et al*., 1988a).

What is especially interesting about this exposure is that the basal 3 m (10 ft) of the middle Huron contain interbedded gray and black shales that coincide with the widespread *Protosalvinia* (*Foerstia)* Zone (Fig. 11). Only about 2.3 m (7.6 ft) of the zone are exposed at this site. *Protosalvinia* is an enigmatic planktic, plant(?) fossil that has been interpreted to represent a possible brown alga (Phillips *et al*., 1972; Schopf and Schwietering, 1970) or an organism with early land-plant affinities (Schopf, 1978; Romankiw *et al*., 1988). It commonly has two forms, a bifurcating lobate thallus (Fig. 14) and an elliptical, reproductive structure (Fig. 15). 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 east-central United States, making it a potentially important stratigraphic time marker (Hasenmueller *et al*., 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 *et al*., 2009). Additional information may be found in Ettensohn *et al*. (1989) and Ettensohn (1992c).

Mileage

22.9 Ohio Shale exposed in road cut for 0.4 miles along the eastbound lanes of I-64.

23.8 Ohio Shale exposed in road cut along the eastbound lanes of I-64.

24.1 **Stop 1C:** I-64 along the eastbound lane, 1.6 mi east of stop 1B. At this stop we will view from afar (across the Interstate) the Three Lick Bed, which separates the underlying Huron Member from the overlying Cleveland Member of the Ohio Shale (Fig. 11) along the westbound lane of I-64. Latitude: 38°, 10’, 45.9”; longitude: 83°, 35’, 6.8”.

At this stop we will quickly view from afar a section of the Three Lick Bed of the Ohio Shale (Provo *et al*., 1978). The Three Lick Bed is a prominent stratigraphic marker horizon (mid-Famennian, lower *Pa. expansa* Zone; Over *et al*., 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. 11). 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 *et al*., 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 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. 7) from northern Ohio (Provo *et al*., 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 periods 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 *et al.*, 1996).

Mileage

24.4 Exit 133; turn right onto exit ramp.

24.5 Farmers Member of the Borden Formation is exposed in road cuts along the right side of the ramp.

24.8 Junction of I-64 and State Route 801; turn right onto 801 and go east.

26.5 The Sunbury Shale, as well as the Farmers and lower part of the Nancy members, are exposed in a road cut along both sides of the road for 0.2 mile. The Bedford and uppermost part (upper Cleveland Member) of the Ohio Shale are exposed on the left, or north, side of State Route 801.

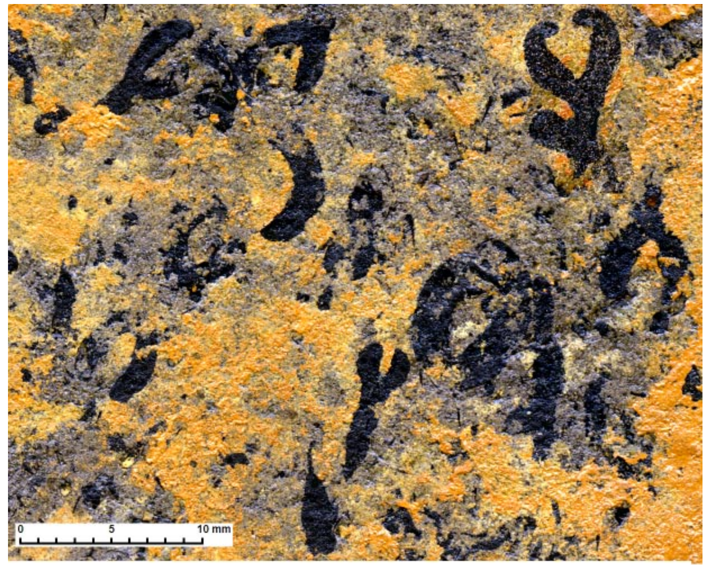


Figure 14
Bifurcating lobate thalli of *Protosalvinia.*

Figure 15
Round to elliptical reproductive bodies of *Protosalvinia*. These are the most common fossils at Stop 1b.

26.95 Junction of State Routes 1722 and 801; turn right onto State Route 1722 and turn around. Please note that directly across the road at the level of the first bench in the road cut is the Three Lick Bed of the Ohio Shale, which separates the underlying Huron Member from the overlying Cleveland Member. Turn left back onto State Route 801, heading west.

STOP 2: Upper part of the black-shale sequence and the lower Borden Formation

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

At Stop 2 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. 16, 17).

Mileage

27.3 **Stop 2**. Pull off the road to the guard rail on the right. Stop 2 is located in a road cut along the northbound lane of State Route 801, 1.75 mi south of exit 133 from

Figure 16
Photo showing the upper part of the stratigraphic section at Stop 2 (see Fig. 17). 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.

Interstate Highway I-64. The exposure is 0.35 mi north of the junction of State Routes 801 and 1722 in the south-central part of the Farmers quadrangle, Rowan County, Kentucky (Figs. 1, 2). Latitude: 38º 09' 08"; longitude 83º 33' 6.5".

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 17.

Unit 1, **Cleveland Shale Member** (**Ohio Shale**). Unit 1 consists of 4.25 meters (14 ft) of the Cleveland Shale Member of the Ohio Shale Formation (Fig. 17). 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. 18). This 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 (Garrels and Christ, 1965, p. 224).

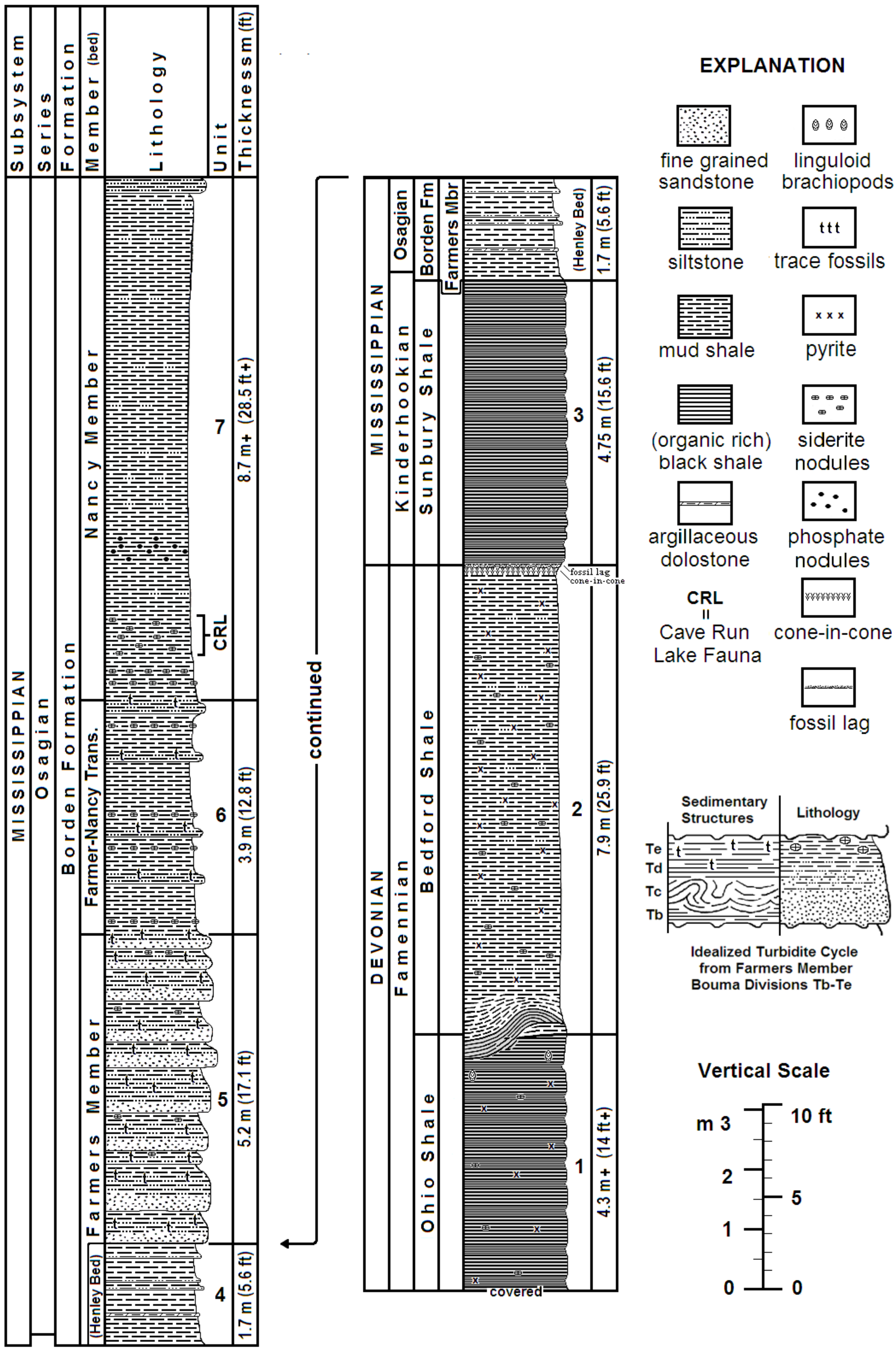


Figure 17
Schematic stratigraphic section from the outcrop at Stop 2 (see Fig. 16).

Figure 18
"Ribbed" black shales from Cleveland Member of the Ohio Shale at Stop 2.

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 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. 19). 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 *et al.*, 1995).

As stated previously, the ribbed character of these shales (Fig. 18) 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 matter. 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

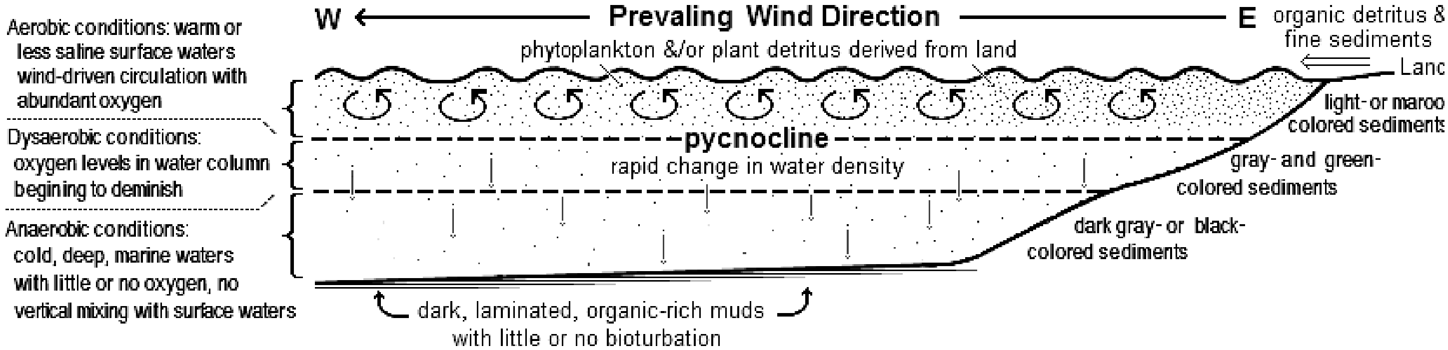


Figure 19
Cartoon illustrating possible means (stratified water column) for the isolation and preservation of organic matter in the Devonian black shales.

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. 5). Algeo *et al*. (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 *et al*. (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 Africa 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 *et al*. (1988a), and Perkins *et al*. (2008).

Unit 2 (**Bedford Shale**). The Bedford Shale consists of 7.9 meters (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. 20). 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 organic matter along the margins of many cones. Petrographic examination of the cones reveals that they are composed 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. 21, 22). 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 N74ºE and dip at an angle of about 5º 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 21, and an outcrop-scale view of several thrusts is shown in Figure 22. 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 47º to 68º 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

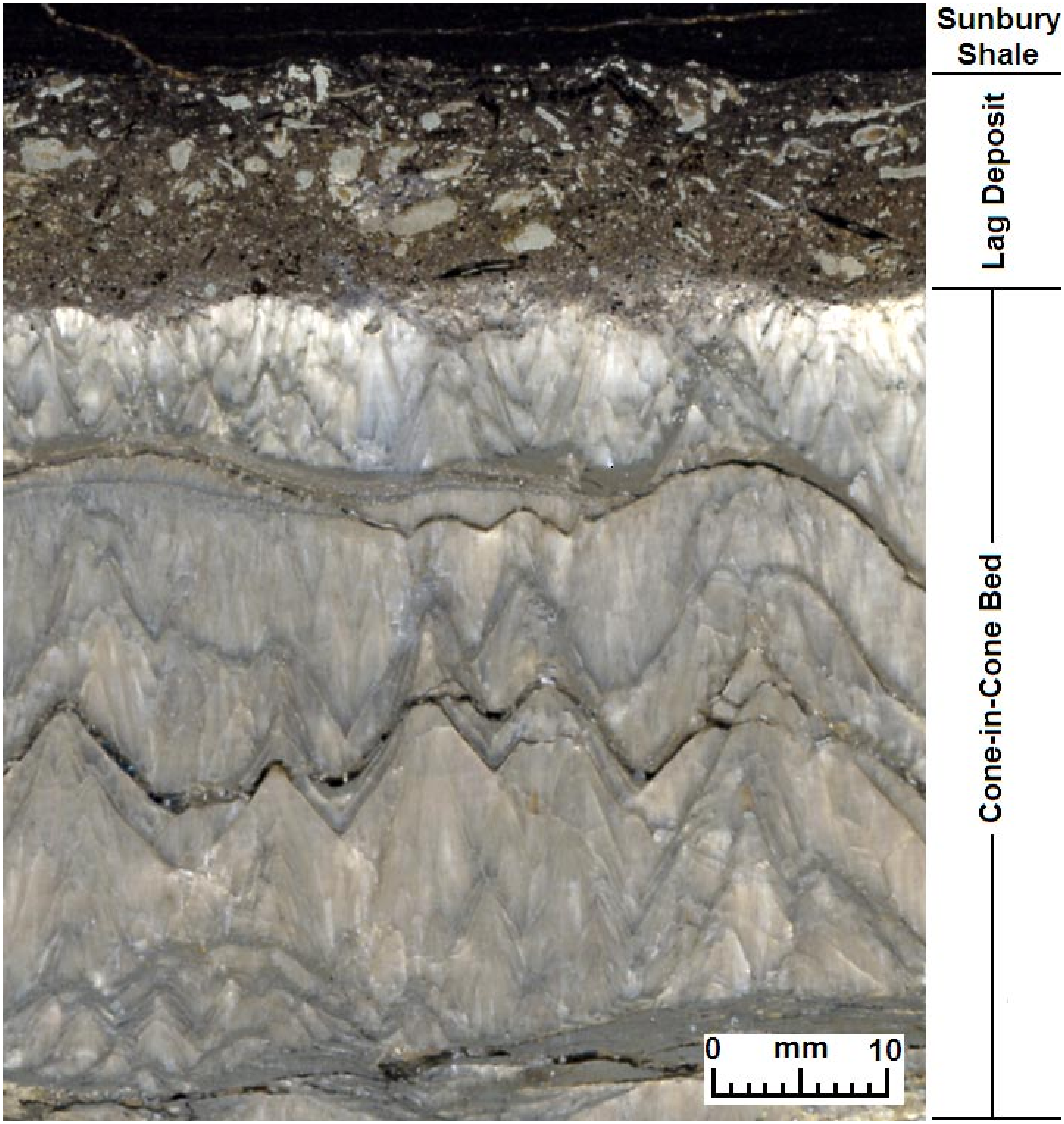


Figure 20
Cone-in-cone limestone bed in the uppermost Bedford Shale immediately below the Sunbury lag zone. The contact between the Sunbury and Bedford shales is a regional unconformity.

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 abundant free oxygen in these sediments is indicated by the presence of pyrite and siderite nodules, 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, 1992a). 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

Figure 21
Close-up view of Cleveland black shales thrust over Bedford gray shales in an erosion gully at Stop 2

(1992b, 1995) envisioned the Bedford Member as part of a delta complex that included the black shales of the Cleveland Member, the 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, 1992, 1995) (Fig. 23). However, recent ideas about the possibility of Late Devonian, Acadian/Neoacadian, alpine glaciation, which we will discuss at Stop 3, 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 *et al.* (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 meters (15.6 ft) of dark-gray to black, highly carbonaceous, fissile shale (Fig. 17). The unit contains small pyritic nodules as well as small, scattered pyrite crystals. Overall, the lithology of the Sunbury Shale is comparable 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 characterized by a 1.5-cm-thick "lag" deposit (Fig. 20). This lag lies immediately above the cone-in-cone layer previously mentioned in the Bedford Shale (Fig. 20) 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. 20). This basal layer or zone is apparently recognized throughout the entire outcrop area of the Sunbury Shale (Pepper *et al.*, 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 during deformational loading (Fig. 7), the Sunbury represents a trangression and subsidence event that moved eastward in time (Fig. 7), apparently reflecting inception of a new, more proximal Neoacadian convergence event to the east (Fig. 8B). 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 *et al*. (1988a), and Lierman *et al*. (1992).

Unit 4, **Henley Bed**, **Farmers Member** (**Borden Formation**). The Henley Bed is the basalmost unit of the Borden Formation (Figs. 16, 17). 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 (early Kinderhookian) and corresponds to the lower *Siphonodella crenulata* Zone of Sandberg *et al*. (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 *Pseu-*

Figure 22
Outcrop view of repeated Cleveland and Bedford section at Stop 2, showing the likely location of thrust surfaces involved.

dopolygnaths 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. 7). 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. 16, 17). At this locality the Farmers consists of 5.2 meters (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. Measure-

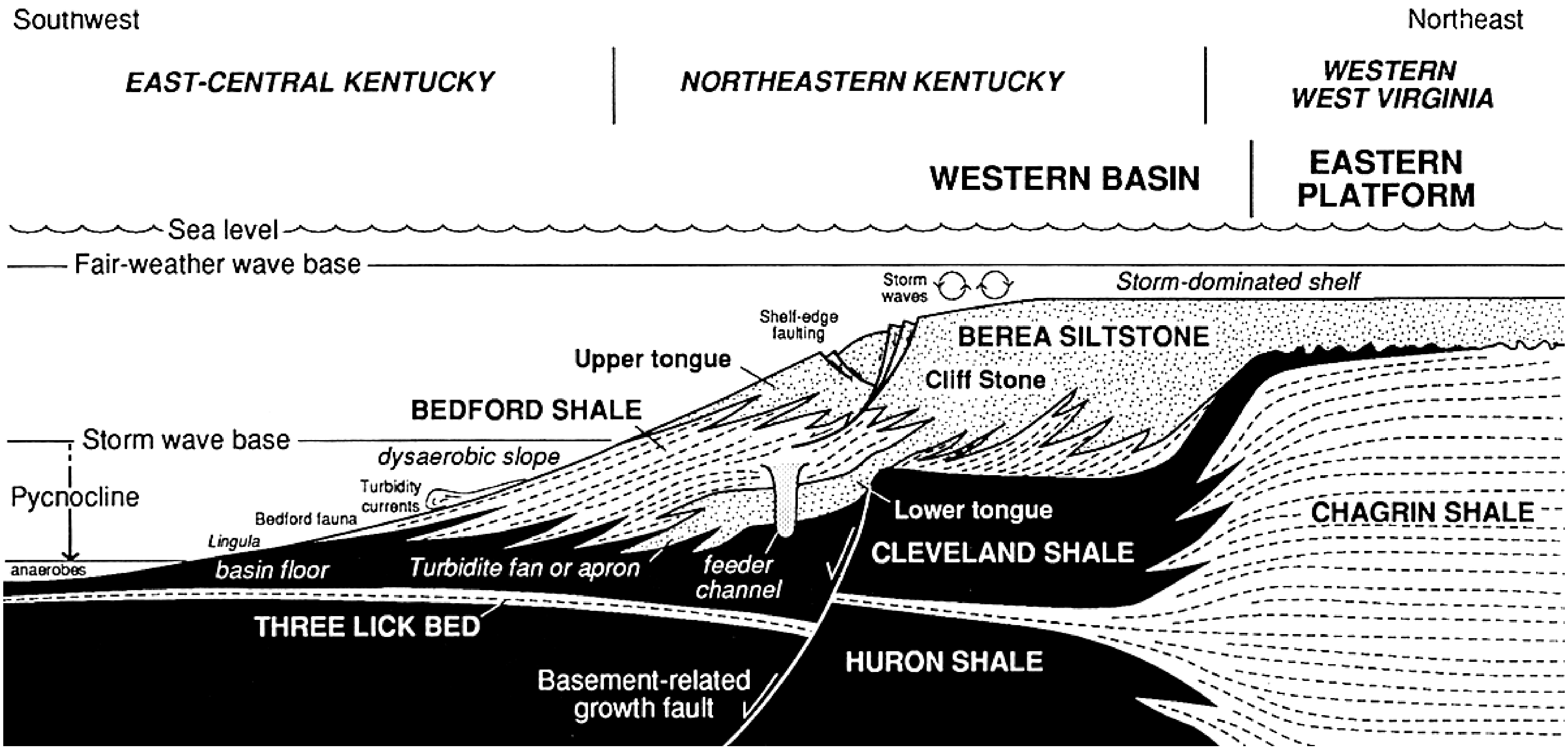


Figure 23
Depositional model for the Bedford-Berea sequence in eastern Kentucky and West Virginia. (after Pashin and Ettensohn, 1987).

ments 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.

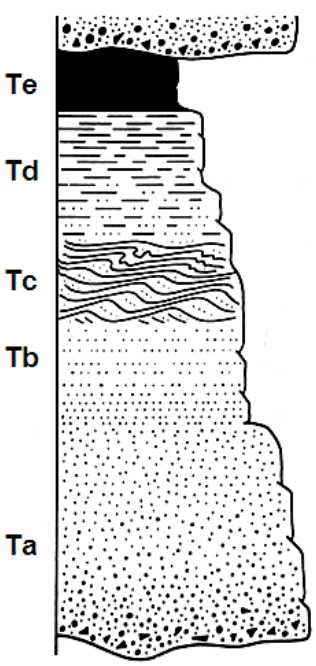
Internal sedimentary structures within the coarser grained beds include parallel laminae, current ripple laminae, and convolute laminae. The coarser grained beds commonly exhibit a lower interval of parallel laminae overlain by current-ripple laminae or convolute laminae, followed by an interval of parallel laminae, which is in turn overlain by shales (see insert in the stratigraphic section for this stop, Fig. 17). 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. 24) 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* (see Appendix 1). 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. 24).

The Farmers Member is thought to be a series of turbidite deposits that accumulated at the outer edge of a prograding Borden delta (Fig. 25). 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. Secondly, 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



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 24
Five-part division of a classic Bouma cycle, showing an ideal sequence of sedimentary structures found in a turbidite bed (adapted from Walker, 1979).

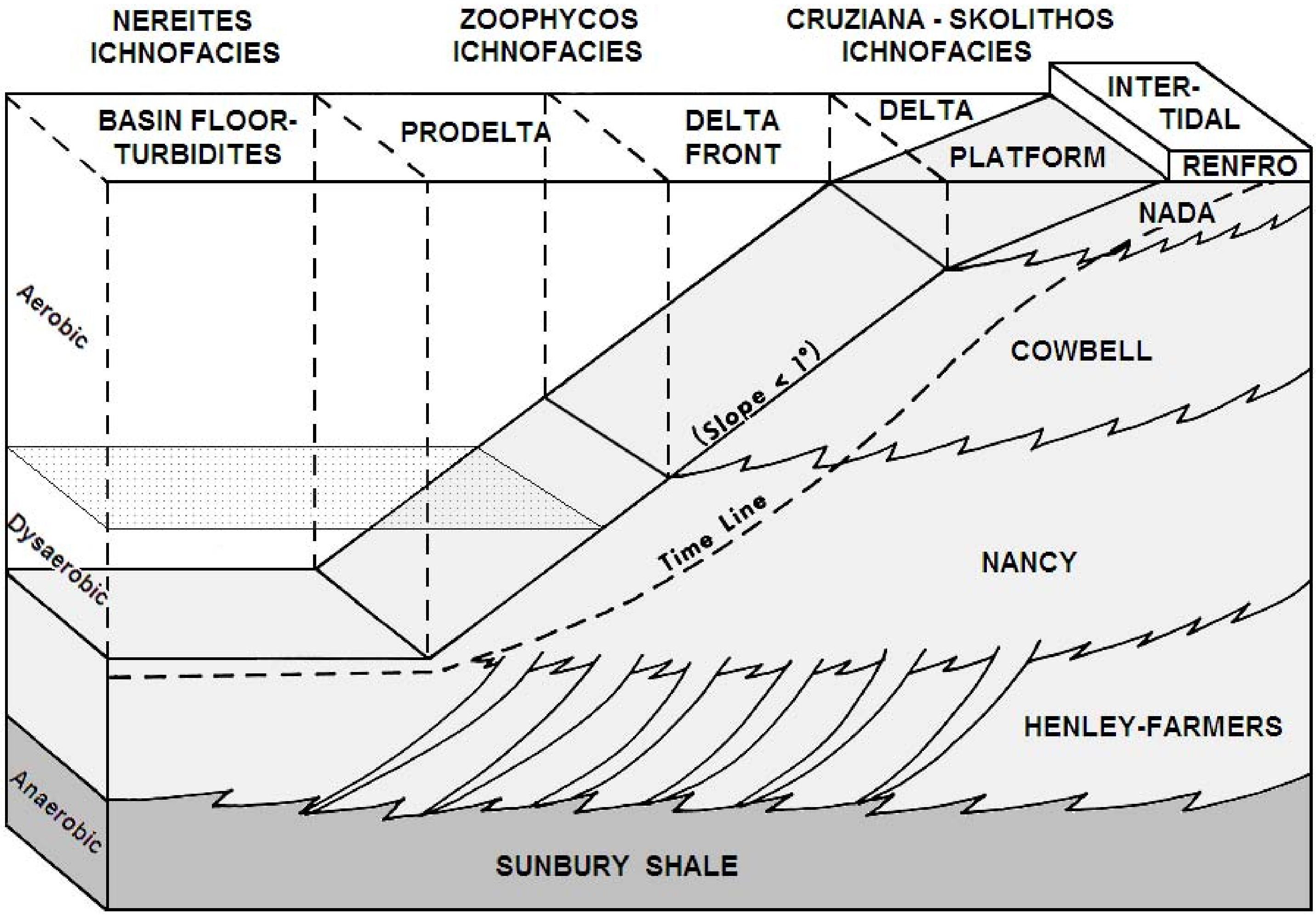


Figure 25
Block diagram showing interpreted depositional environments of the typical lithofacies of the Sunbury and Borden formations in the Morehead area, east-central Kentucky (adapted from Kepferle, 1977).

bed in the Farmers is striking (Fig. 17). 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. 25). 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 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 a typical Bouma Sequence. 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. 26). 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 Lierman (1985, 1992), and Chaplin and Mason (1979).

Unit 6, **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. 16, 17). 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

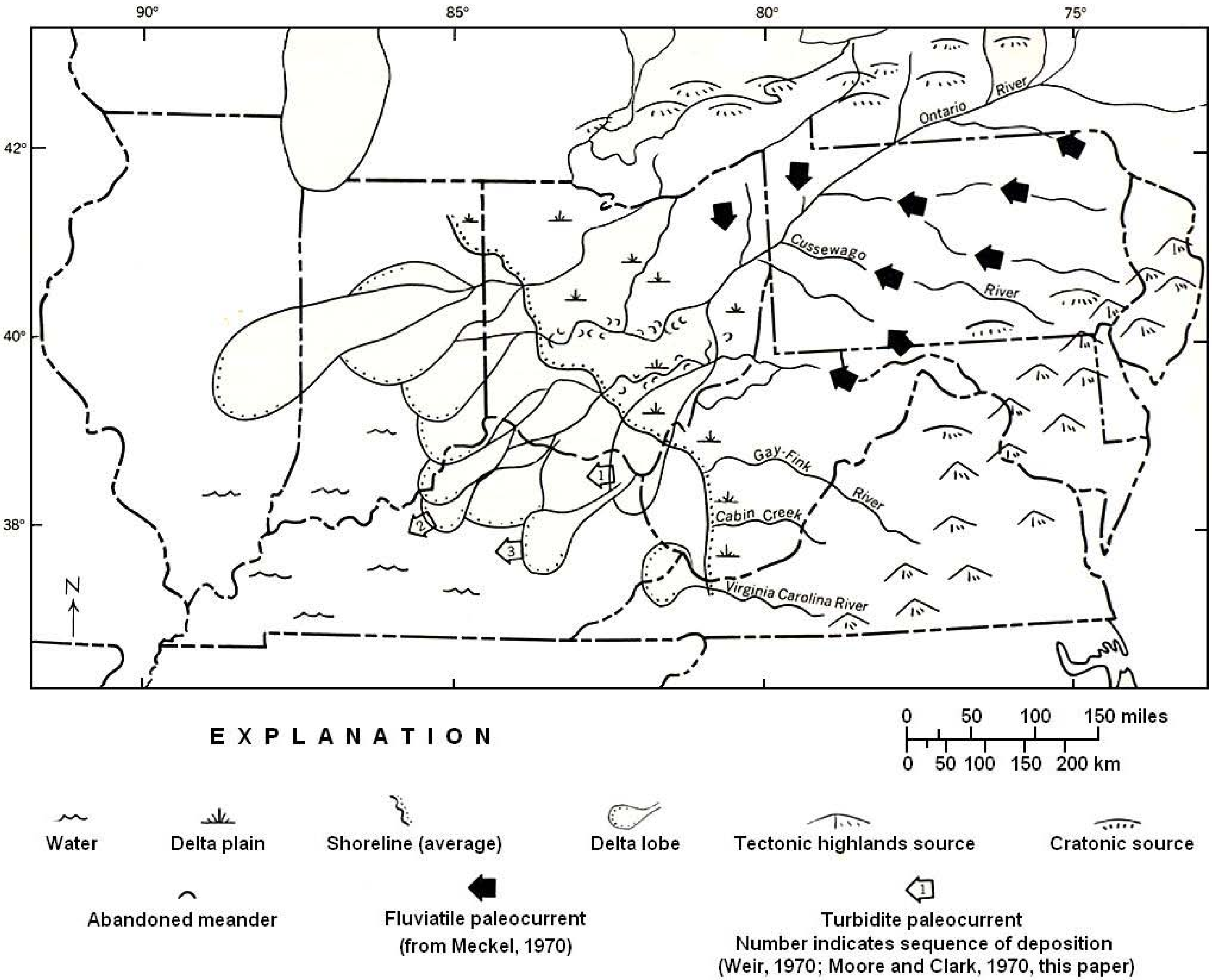


Figure 26
Paleogeographic map showing the development and progradation of the "Borden Delta Complex" in Kentucky, Indiana, Illinois and Ohio in Early Mississippian time (adapted from Kepferle, 1977).

simply separates the thicker bedded sands of the Farmers Member below from the shales of the Nancy Member above.

Unit 7, **Nancy Member** (**Borden Formation**). At this locality, the Nancy Member, is incompletely exposed (Figs. 16, 17) and consists of 8.7 meters (28.5 ft) of greenish-gray to grayish-green mud shale to silty shale that is poorly fissile, non-calcareous, 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. 25), 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. These conditions indicate 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 the shale, indicating the presence of greenish phyllosilicates; (2) the abundance of siderite nodules, a mineral that forms under reducing conditions (Garrels and Christ, 1965); and (3) the presence of a dysaerobic fauna (the Cave Run Lake fauna; Fig. 17) near the base of the Nancy Member (Mason and Kammer, 1984; Mason and Work, 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, (3) the occurrence of delta-front sands and silts of the Cowbell Member, which conformably overlie the Nancy Member (Fig. 25). 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, 1992, 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 *et al.* (1992a, b), and Work and Mason (2005).

Mileage

29.1 Entrance ramp onto eastbound I-64.
29.4 I-64 East; Nancy Member of the Borden Formation exposed along both sides of I-64, especially along the westbound lanes.
30.2 Sunbury Shale through the lower part of the Nancy Member of the Borden Formation are exposed along both sides of I-64 for 0.4 mile, especially along the westbound lanes.
30.9 Bridge over Bull Fork Creek.
31.0 Sunbury Shale through the lower part of the Nancy Member of the Borden Formation are exposed along both sides of I-64 for 0.4 mile.
31.5 Sunbury Shale through the lower part of the Nancy Member of the Borden Formation are exposed along both sides of I-64 for 0.3 mile.
32.5 Bridge over North Fork of Triplett Creek.
32.7 Farmers Member of the Borden Formation is exposed along both sides of I-64 for 0.2 mile.
33.0 Exit 137 (Morehead exit); turn right onto ramp.
33.4 Junction of I-64 and State Route 32; turn left onto State Route 32 and proceed west. Note: The Farmers Member of the Borden Formation is exposed in the area road cuts.
33.5 Cross under I-64.
33.6 First traffic light after passing under I-64.
33.7 Traffic light at junction of road to Trademore Shopping Center.
34.0 Bridge over North Fork of Triplett Creek.
34.1 Traffic light at junction of Bratton Branch Road and access road to Wal-Mart; turn left.
34.15 Stop sign; turn left again onto Bratton Branch Road.
34.4 Slumped Farmers siltstone blocks along right side of road.
34.5 Upper Bedford Shale through the lower part of the Farmers Member of the Borden Formation are exposed in a road cut along the right side of the road for 0.3 mile.
35.2 Bridge over Logan Branch.

STOP 3: Granitic dropstone embedded in the uppermost Cleveland Shale Member of the Ohio Shale

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

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.

Mileage

35.25 **Stop 3.** Immediately after crossing bridge, turn left onto a dirt road. We will park here and walk up Logan Hollow Road 0.2 miles to **Stop 3**. After visiting Stop 3, we will retrace our route back to the junction of Bratton Branch Road and U.S. 32. Stop 3 is located along a creek called Logan Hollow Branch, approximately 0.2 miles 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: 38º 11' 36.1"; longitude 83º 29' 36.6".

Stratigraphically, the boulder occurs at the very top of the Upper Devonian Cleveland Member of the Ohio Shale (Fig. 27). The Bedford Shale can be easily dug out along the bank of the stream where it overlies the Cleveland Shale. Mispores from both the Cleveland Shale and basal Bedford Shale at this locality are assigned to the LN Biozone. Continuing down the 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 27.

Statistics:

Granite boulder: First discovered by Michael J. Robinson of NYTIS Exploration Company 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 while the corners and edges are rounded. The top may be faceted (Figs. 28, 29).

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.

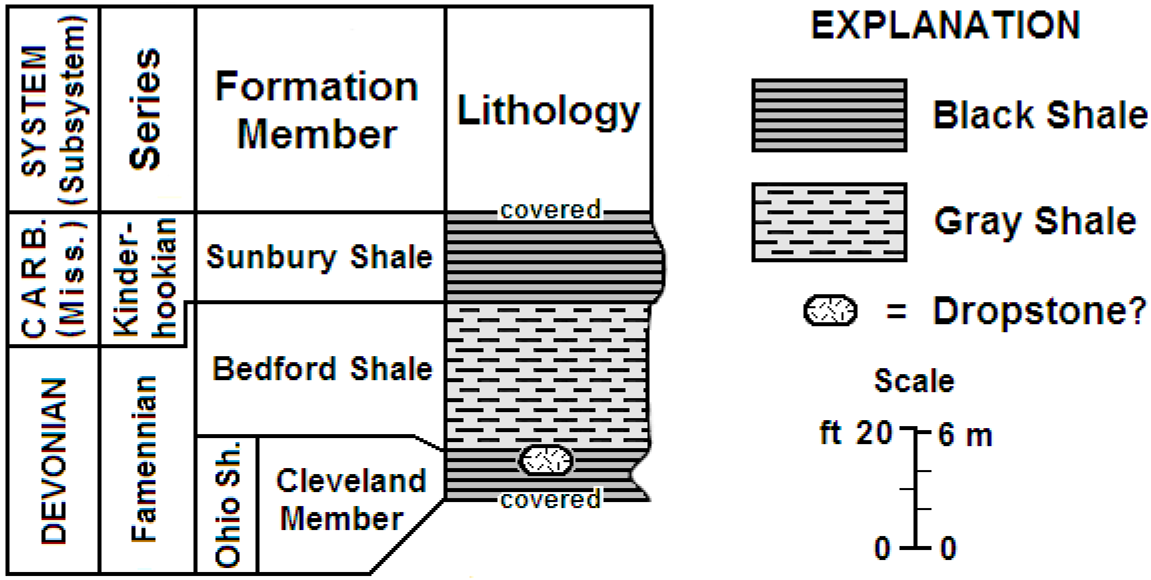


Figure 27
The stratigraphic section in Logan Hollow Branch at Stop 3.

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 *et al.*, 2008).

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 200–250 miles east of this locality. Most paleogeographic

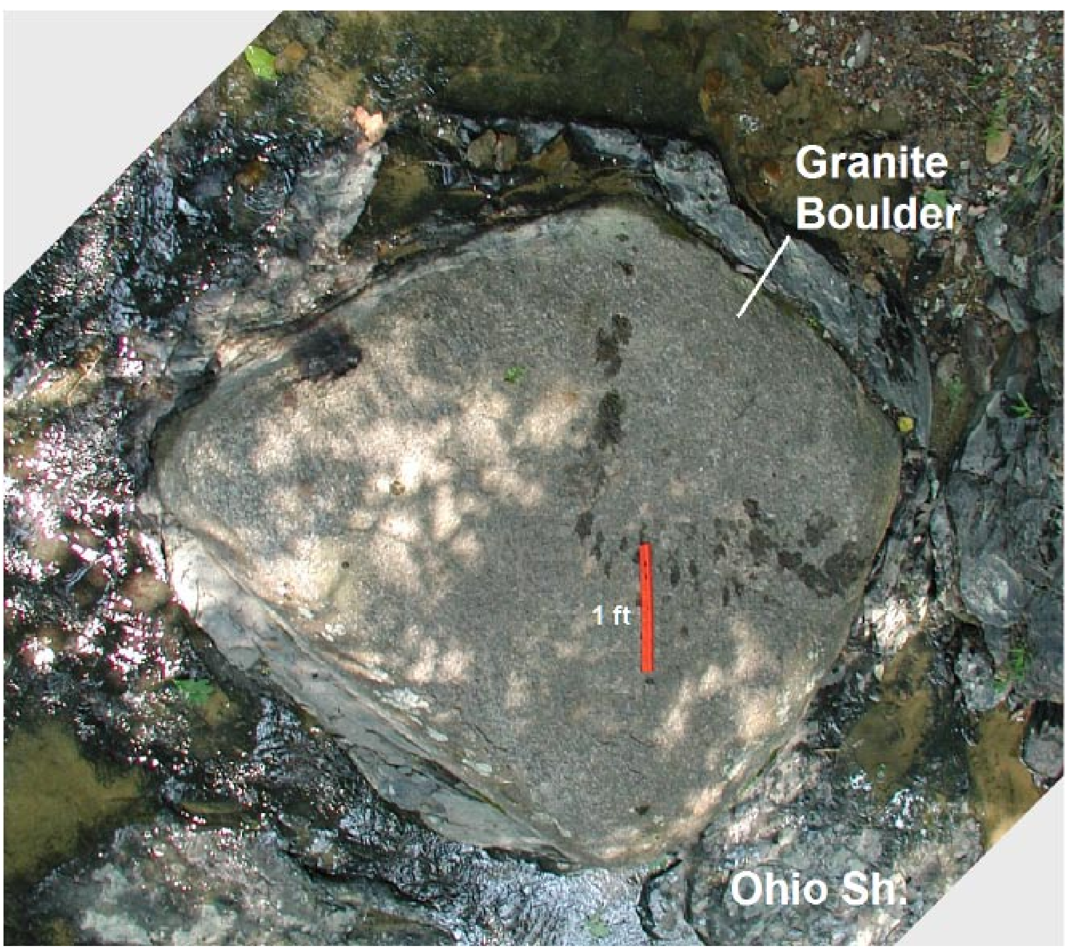


Figure 28
Top view of the granite boulder. Note the flattened sides, faceted top, and rounded corners of the boulder. Scale is 1 ft.

Figure 29
Boulder is clearly embedded in the Cleveland Member of the Ohio Shale Note upturned layers or "mud drape" along edges of boulder.

reconstructions for this time period place Kentucky and the Acadian mountains at around 30º 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 of 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 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.

This fascinating subject is discussed in greater detail in the following article. More information can be found in abstracts by Lierman and Mason (2007) and Ettensohn *et al.* (2007, 2008).

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 based on Upper Devonian diamictites (Fig. 30A), pebbly mudstones, and laminites in the Appalachian Basin of eastern and south-central Pennsylvania and adjacent parts of Maryland (Sevon, 1973; Cecil *et al*., 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 *et al*., 1992), during which Gondwanan glaciation in South America and Africa and likely

Figure 30
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.

Appalachian alpine glaciation briefly unfolded. Herein, we report the occurrence of an *in-situ* igneous dropstone (Fig. 30B) 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. 31), 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. 30A) are unique in the Appalachian Basin; they are included in the lower Rockwell Formation in south-central Pennsylvania and adjacent parts of Maryland (Berg *et al.*, 1980) and the lower Spechty Kopf Formation in the Anthracite region of eastern Pennsylvania (Epstein *et al.*, 1974) (Fig. 31). 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 unbedded, 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 *et al*., 2004) (Fig. 31).

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 *et al.*, 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 *et al.*, 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 *et al.,* 1997; Berg, 1999). Some sandstones associated with the unit contain *Skolithos* trace fossils (Bjerstedt, 1986; Fig. 32).

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 fossils and is easily correlated with well-dated, marine, Sunbury black shales to the west (Fig. 32). However, recent diamictite palynology from the Spechty Kopf Formation (Sevon *et al.*, 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. 32).

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

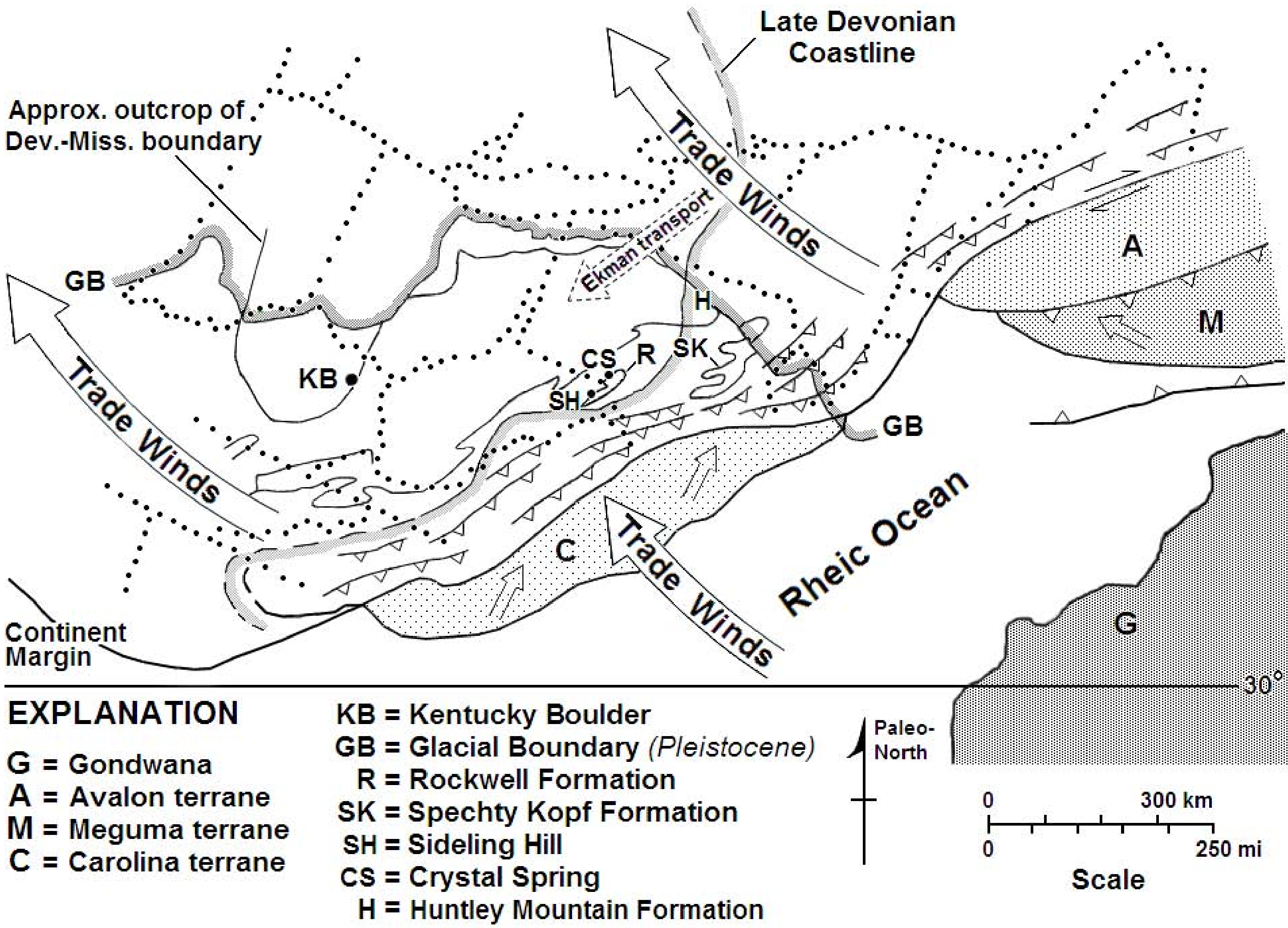


Figure 31
Late Devonian paleogeographic reconstruction of southeastern Laurussia, superimposed with positions of outcrop belts, field localities, Pleistocene glacial boundary, modern political boundaries, and likely Late Devonian wind and current directions.

following breach of a structural front or tectonic basin (Bjerstedt, 1986), and deposition following a Late Devonian bolide impact (Sevon *et al.*, 1997). In contrast, Cecil *et al.* (2004), Brezinski *et al.* (2008), and Dennis (2007) 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 metamorphic 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, Maryland, and Crystal Spring, Pennsylvania (Bjerstedt, 1986; Cecil *et al.*, 2004), (Figs. 31, 32).

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-rafted dropstones from proglacial lakes (Fig. 31). Pleistocene glaciation was the accepted interpretation until 2007, when Lierman and Mason (2007) reported a large faceted, biotite-granite boulder (1.7×1.3×.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. 30B, 32). Palynology indicates a position within the *Retispora lepidophyta–Verrucosporites nitidus* Miospore Zone (G. Clayton, pers. commun., 2007) while conodonts indicate presence in the middle *S. praesulcata* Zone (Ettensohn *et al.*, 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 (300 mi) from the diamictites in Pennsylvania and Maryland (Fig. 31). 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

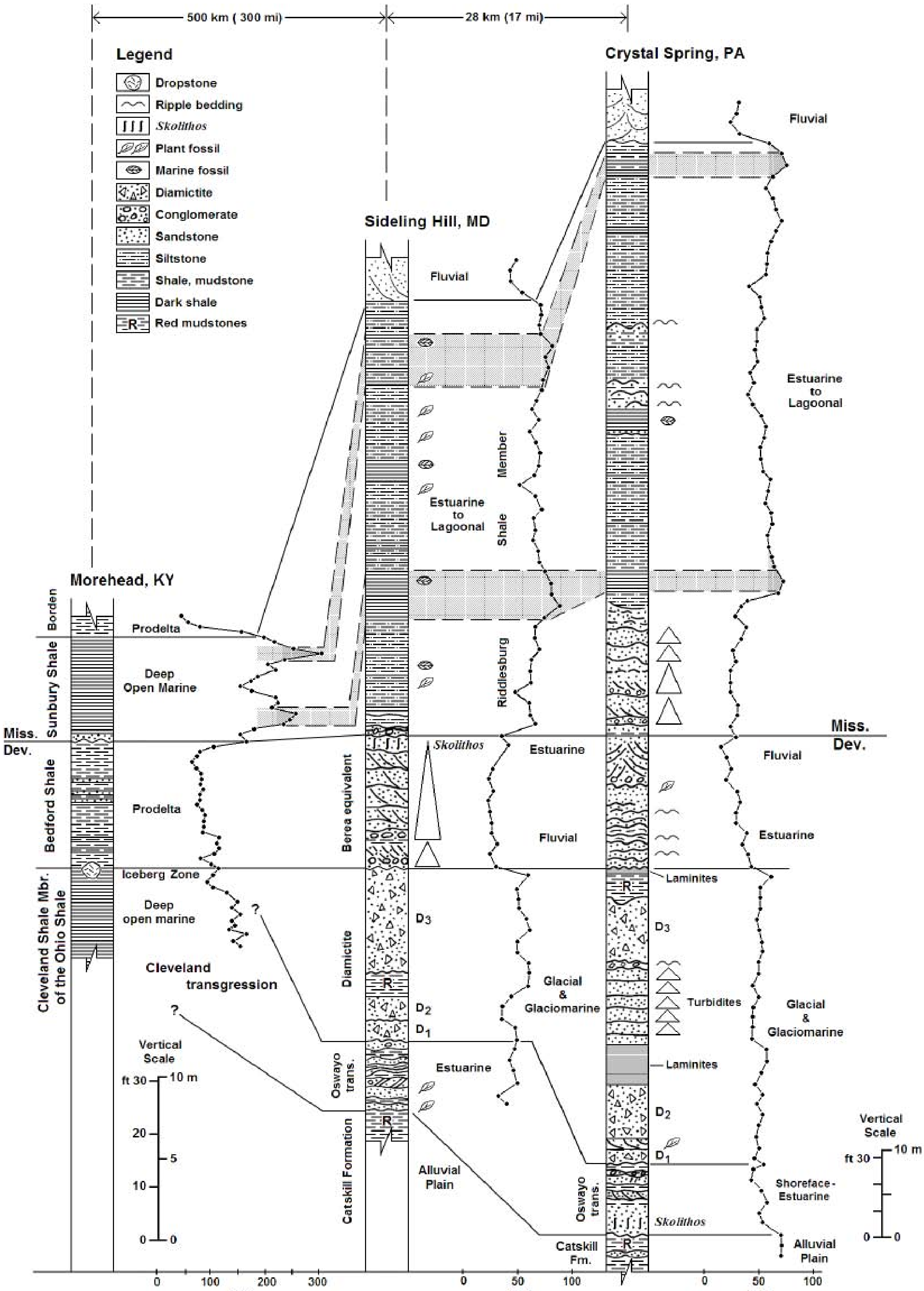


Figure 32
Stratigraphy and gamma-ray correlations between the dropstone locality in Kentucky and the diamictite localities in Maryland and Pennsylvania.

Rockwell includes the prominent marine/marginal-marine Riddlesburg black shale (Fig. 32).

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. 32). 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 clastic 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 *et al.*, 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 *et al.*, 1978). Similar radioactive correlation between the "boulder section" in Kentucky and the diamictite sections at Sideling Hill and Crystal Spring (Fig. 32) 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.*, Kammer and Bjerstedt, 1986; 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 *et al.*, 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.*, Kammer and Bjerstedt, 1986; Bjerstedt, 1986) and farther south by the Finzel tongue of the Greenland Gap Formation (Dennison *et al.*, 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. 31) 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. 32). 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. 31).

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 *et al.*, 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 *et al.*, 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. 31), 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, 1992), 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 *et al.*, 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 *et al.*, 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. 31). Although Neoacadian tectonism coincided with a global sea-level rise (Johnson *et al*., 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 *et al*., 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.

Mileage

36.3 Junction of Bratton Branch Road and State Route 32. Turn right onto State Route 32 and proceed east toward I-64 and Morehead.

36.4 Bridge over North Fork of Triplett Creek. Note the Sunbury Shale and Farmers Member of the Borden Formation exposed in road cut along the north side of the road after crossing bridge.

36.8 Traffic light at junction to road to Trademore Shopping Center.

36.9 Traffic light at junction of westbound ramp to I-64, numerous exposures of the Farmers Member of the Borden Formation in area road cuts.

37.0 Cross under I-64. Note exposures of the Farmers Member of the Borden Formation in road cut along both sides of State Route 32.

37.1 Traffic light at junction of eastbound ramp to I-64 and State Route 32. Again, numerous exposures of the Farmers Member occur in area road cuts.

37.2 Traffic light at junction of Pine Crest Lane and State Route 32. Turn right onto Pine Crest Lane.

37.3 Junction of Pine Crest Lane and Tom's Drive. Turn left onto Tom's Drive Road. Follow Tom's Drive Road for 0.05 mi and turn left into Shoney's parking lot.

37.35 Shoney's parking lot. We will stop at Shoney's for lunch. We will have one hour here for lunch. After lunch we will load up the vehicles and retrace our route back to the junction of Pine Crest Lane and State Route 32.

37.6 Junction of Pine Crest Lane and State Route 32; turn left onto State Route 32 heading west.

37.7 Traffic light at junction of I-64 east access ramp turn right onto eastbound ramp. Note the normal fault in Farmers Member of Borden Formation in road cut at entrance to the eastbound ramp on right side of the road.

38.0 I-64 East.

38.8 Farmers Member of the Borden Formation is exposed in road cuts along both sides of I-64 for 0.2 mi.

39.2 Farmers Member of the Borden Formation is exposed in a road cut along the eastbound lane of I-64 for 0.1 mi.

39.8 Nancy Member of the Borden Formation is exposed in a road cut along the eastbound lane of I-64 for 0.2 mi.

40.4 Nancy Member of the Borden Formation is exposed in a road cut along the eastbound lane of I-64.

41.6 Rest area along eastbound lane.

42.4 Upper Farmers and lower Nancy members of the Borden Formation exposed in road cut along the eastbound lane of I-64.

42.7 Upper Farmers and lower Nancy members of the Borden Formation exposed in road cut along the eastbound lane of I-64.

43.1 Nancy Member of the Borden Formation exposed in road cut along the eastbound lane of I-64 for 0.1 mile.

43.9 Nancy Member of the Borden Formation is exposed in road cut along the eastbound lane of I-64 for 0.1 mile.

44.4 Contact between the Nancy and Cowbell members of the Borden Formation exposed in the road cut along the westbound lanes. The Cowbell Member of the Borden Formation is exposed almost continuously in road-cut exposures along the westbound lane for the next 1.3 mi.

STOP 4: The Cowbell Member of the Borden Formation: Transition from delta front to delta platform

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

At this stop, we will examine upper parts of the Cowbell Member and its transition into the Nada Member (Fig. 33). This part of the section shows the transition from delta-front environments with relatively high sediment input to a delta-platform setting with sharply reduced sediment input and major reworking by storms. The section also shows a high diversity of trace fossils and reworked body fossils.

Mileage

45.9 **Stop 4**. Here we will leave the bus (milepost 145.5) and walk upsection for about 0.5 mi, eastward toward the

State Route 799 overpass. After viewing the section for Stop 4, we will cross the Interstate by foot under the State Route 799 overpass. Once we have crossed I-64, we will walk along the shoulder of the westbound lane eastward about 0.2 mi to Stop 5. While we are examining and discussing Stop 4, the bus and other vehicles will proceed to the Olive Hill exit (exit 156) and return along the westbound lane to Stop 5. The stop includes a series of road cuts along the eastbound lane of I-64, 8.0 mi east of the junction of I-64 and State Route 32; the stop is 10.7 mi from Stop 3. Latitude: 38°, 16', 21.1"; longitude: 83°, 23', 5.8". The stop is located on the Cranston quadrangle.

Unit 1, **Cowbell Member** (**Borden Formation**). The Cowbell section is incomplete here and consists of approximately 5.7 m (18.7 ft) of shale and siltstone (Fig. 33). The section starts with shale and grades upward into fairly massive siltstone over a vertical distance of about 3 to 3.5 m. The lithology is an olive gray to gray-green, silty shale, which splits into platy fragments and is non-calcareous. The shale contains grayish-red siderite nodules scattered throughout. Thin beds of siltstone within this shale become more frequent upsection. The upper third of this unit is a light-olive gray siltstone to very fine-grained sandstone. An abundance of sedimentary structures is present within the siltstone and include horizontal laminae, trough cross laminae (scour-and-fill), and small soft-sediment deformation features. Many of the primary sedimentary structures are disrupted by bioturbation. Trace fossils are also common throughout the siltstone and include both vertical and horizontal burrows. Ichnogenera observed include *Zoophycos, Cylindrichus, Scalarituba, Helminthoida, Planolites, Lophoctenium, Phycodes,* and *Bergaueria* (see Appendix 1).

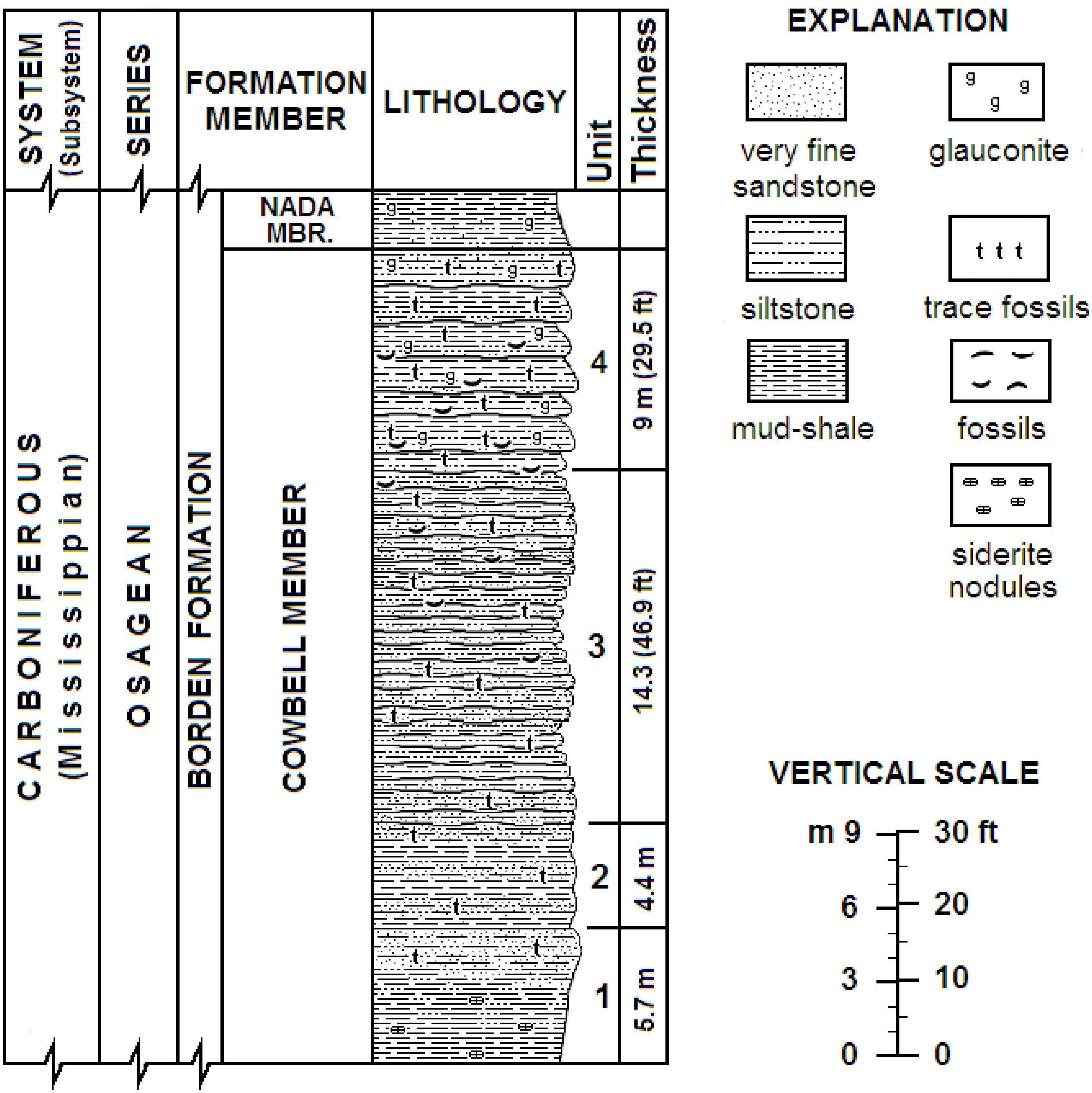


Figure 33
The stratigraphic section at Stop 4 along I-64 at milepost 145.5.

Unit 2, **Cowbell Member.** The unit consists of 4.4 m (13.4 ft) of interbedded siltstone and shale (Fig. 33). The siltstone is gray-green to maroon in color; coarse silt to very fine-grained sand occurs in relatively even beds. The shale is an olive gray to gray-green, silty shale, which splits into platy fragments, and is noncalcareous. An abundance of sedimentary structures is present within the siltstone, which include horizontal laminae, trough cross-laminae (scour-and-fill), as well as hummocky crossbeds. The shale tends to grade into the coarser grained siltstone along both the upper and lower contacts of the siltstone beds. Trace fossils are also common throughout the siltstone and include vertical and horizontal burrows. Ichnogenera observed include: *Zoophycos, Cylindrichus, Scalarituba, and Monocaterion* (see Appendix 1).

Unit 3, **Cowbell Member.** This unit consists of approximately 14.3 m (46.9 ft) of interbedded siltstones and shales (Fig. 33). Siltstone beds are a light-gray color, vary from a coarse silt to a very fine-grained sand, and occur in irregular, uneven beds that range from about 10 to 30 cm thick (medium bedded). The shales are more typically a gray to gray-green color, are generally silty, split into platy fragments, and are noncalcareous. Individual beds show a fining-upward character, typically with sharp (erosional) bases and upper surfaces that grade into the overlying shale. Sedimentary structures observed in this interval include horizontal laminae, both hummocky and swaley cross-strata, trough cross-strata, and ripple marks (both asymmetrical current ripples and symmetrical wave ripples). The bases of many siltstone beds may exhibit mud-clast lags and are commonly floored with a lag of fossil debris. This becomes especially evident about one-third of the way up into the unit. Megafossils in the debris lags include: brachiopods (orthids and spiriferids), bryozoans (fenestrate and trepostome), pelmatozoan debris, gastropods, pelecypods, and some solitary rugose corals. Most of the fossils appear to be randomly oriented and disarticulated. Trace fossils are also common throughout this unit and tend to be associated with the *Cruziana-Skolithos* ichnofacies; they include *Zoophycos, Lophoctenium, Cylindrichus, Sclarituba, Monocraterion, Chondrites,* and *Helminthoida* (see Appendix 1).

Unit 4, **Cowbell Member**. Unit 4 consists of 9 m (29.5 ft) of interbedded shale and siltstone (Fig. 33). The shale is a greenish-gray to grayish-green mud shale to silty shale that is poorly fissile, noncalcareous and highly bioturbated. Siltstone is a light-gray color and is commonly streaked with glauconite. The siltstone occurs as even beds that range from 30 to 50 cm thick (medium to thick bedded) and up to 1 m thick at the top of section. Individual beds show a fining-upward character, typically with sharp (erosional) bases and upper surfaces gradational into the overlying shale. Sedimentary structures apparent in these beds include horizontal laminae, hummocky cross-strata, and ripple marks. The bases of many of the beds contain mud-clast lags, while others are commonly floored with fossil lags. Most of the fossils in this unit are found along these lag deposits and tend to be preserved as both internal and external molds; they reflect an open-marine fauna and include brachiopods (spiriferid), bryozoans (fenestrate and trepostome), pelmatozoan debris, gastropods (both planispiral and conispiral forms), pelecypods, nautiloid cephalopods, and both solitary rugose and tabular corals. Trace fossils are also common throughout this unit and tend to be associated with the *Cruziana-Skolithos* ichnofacies; they include *Zoophycos, Scalarituba, Helminthoida, Planolites, Lophoctenium, Phycodes,* and *Bergaueria*. The uppermost siltstone bed is heavily burrowed and contains abundant vertical tubes of *Skolithos*.

Interpretation. Ascending upsection, **Unit 1** is interpreted to represent a coarsening-upward, delta-front, depositional environment. A delta front is the sloping frontal portion of a delta just seaward of the delta plain (Fig. 25), where distributary-mouth bars commonly form. A distributary-mouth bar is a sandbar that forms across the mouth of a distributary channel. It develops where the freshwater flow of a river slows down as it meets the ocean and deposits its load of sediment. The more seaward portion of the distributary channel is called the distal bar or bar front. It is the progradation of these distributary channels and their distributary mouth bars into open waters that produces a coarsening- and thickening-upward sequence. Evidence to support this interpretation includes: (1) the coarsening-upward nature of Unit 1; (2) the presence of sedimentary structures, trough cross-strata (Fig. 34) and small soft-sediment deformation features, that indicate the rapid deposition of sand- and silt-size sediments in a moderately high current; and (3) the overall dominance of primary sedimentary structures over bioturbation, indicating high sedimentation rates in a strong current.

Unit 2 is believed to be a transition zone between delta-front sands and silts to a more shelf-dominated setting where wave and current action from storms began to rework delta-front sediments. Evidence in support of this is as follows. (1) The contacts between coarser grained siltstone/sandstone beds and the underlying and overlying shales are gradational. Here, there is a combination of a coarsening-upward sequence followed directly by a fining-upward sequence. A coarsening-upward sequence suggests an increase in current energy as each bed is being deposited, followed by decrease in current energy. (2) The co-occurrence of trough cross-strata and hummocky cross-strata has several implications. Trough cross-stratification is reflected in a series of inclined layers of sediment whose overall shape or profile is a three-dimensional trough (Fig. 34). It results from the migration of crescent- or linguloid-shaped bedforms under unidirectional flow conditions. This happens under the influence of a strong unidirectional current and presumably forms as distributary-mouth bars prograde out across prodelta deposits. Hummocky cross-stratification, on the other hand, is a type of low-angle cross-stratification consisting of a series of curving laminae, both convex-up (hummocks) and concave-up (swales), which dip at fairly shallow angles of about 12–15º (Figs. 35, 36); it has been recognized as a diagnostic feature of shallow-marine storm deposits since it was first identified by Harms *et al.* (1975). The fact that both of these features are present suggests that two very different processes were at work here, one, being the progradation of distributary-mouth bars, giving the lower half of these units a coarsening-up profile, and the second being the reworking of this material by storm and wave activity, thereby developing a fining-upward profile. (3) Also present is a combination of trace fossils that are indicative of both the *Cruziana* and *Zoophycos* ichnofacies, which encompass an environmental area ranging from the shelf/shelf edge to the slope.

Unit 3 is thought to represent a series of storm deposits (tempestites) that formed in the transition from delta-front sands to a shallower shelf area, or in this case the delta platform. This change is suggested by the following evidence: (1) The presence of a series of beds that have fining-upward profiles, typically with a sharp erosional base that passes upward into a siltstone/siltstone that in turn grades upward into shales above (Fig. 37). This gradation suggests a decreasing current energy as each bed was being deposited. (2) The presence of both hummocky and swaley

Figure 34
Typical trough cross-strata from the Cowbell Member of the Borden Formation at Stop 4.

cross-strata suggests a storm-dominated environment. Swaley cross-stratification (Fig. 36) is a feature related to hummocky cross-stratification but is dominated by swales or depressions at the expense of hummocks and has been commonly reported from fine- to very fine-grained sandstones. Swaley cross-stratification is also thought to have formed in a way similar to hummocky beds, but in shallower water conditions where aggradation rates are low enough to cause preferential preservation of swales. Dumas and Arnott (2006) have attempted to quantify the specific flow hydrodynamics related to these settings, demonstrating that oscillatory flow (*i.e.*, currents that move both back and forth) is responsible for the production of these bedforms. Swaley cross-stratification seems to become more evident in the upper half of the unit, suggesting that oscillatory flow, due to wave activity from storms, was becoming more intense in younger, shallower parts of the unit. (3) Such back-and-forth, oscillatory motion is also reflected in the presence of symmetrical wave ripples, observed on a number of the sandstone/siltstone beds. (4) The base of many of these beds is also characterized by the presence of a lag of randomly oriented and disarticulated fossils debris. This is particularly noticeable in the upper one-half to two-thirds of this unit. Figure 37 shows a typical storm-bed profile (tempestite) as seen in this unit. (5) The trace fossils observed here are increasingly more indicative of the *Cruziana* ichnofacies, suggesting an environmental area ranging from the shelf edge to shelf or delta-platform proper.

Finally, **Unit 4** is interpreted to have been located directly on the shelf, or in this instance, on the submarine delta platform. The unit is dominated here by shales, as much wave energy from storms was being dissipated on the edge of the shelf or along the top edge of the delta front. Moving up onto the delta platform, conditions would have generally become quieter, and lower energy shales would be the dominant sediments found. This quieter water setting, however, is punctuated by a number of storm-generated siltstones. These have much the same depositional profile as described in Unit 3, except that there are fewer of them (nine in all) and they tend to be thicker and more laterally continuous. These shallowing conditions are further reflected by the trace fossils found in this unit, which tend to be associated with the *Cruziana-Skolithos* ichnofacies (Chaplin, 1980). At this point, glauconite also seems to become common in the shales as well as in the siltstone beds, again, probably reflecting a decrease in sedimentation rates. Curiously, the glauconite in the siltstones is concentrated in thin, wavy laminae that mimic the basic configuration of hummocky cross-strata, again suggesting storm reworking after deposition. Additional information and interpretations regarding the Cowbell member can be found in papers by Kearby (1971), Mason and Chaplin (1979), Chaplin (1980, 1982, 1985), and Lowry-Chaplin and Chaplin (1985).

Economically, the upper parts of the Cowbell Member, especially where distributary-bar sandstones are compartmentalized by intervening shales, locally form hydrocarbon reservoir rocks, known in the subsurface as the Big Injun. Although the Big Injun in reality comprises a stratigraphically diverse series of sandstones, both above and below the upper Cowbell, at least some of those reservoirs do include the upper Cowbell and equivalent

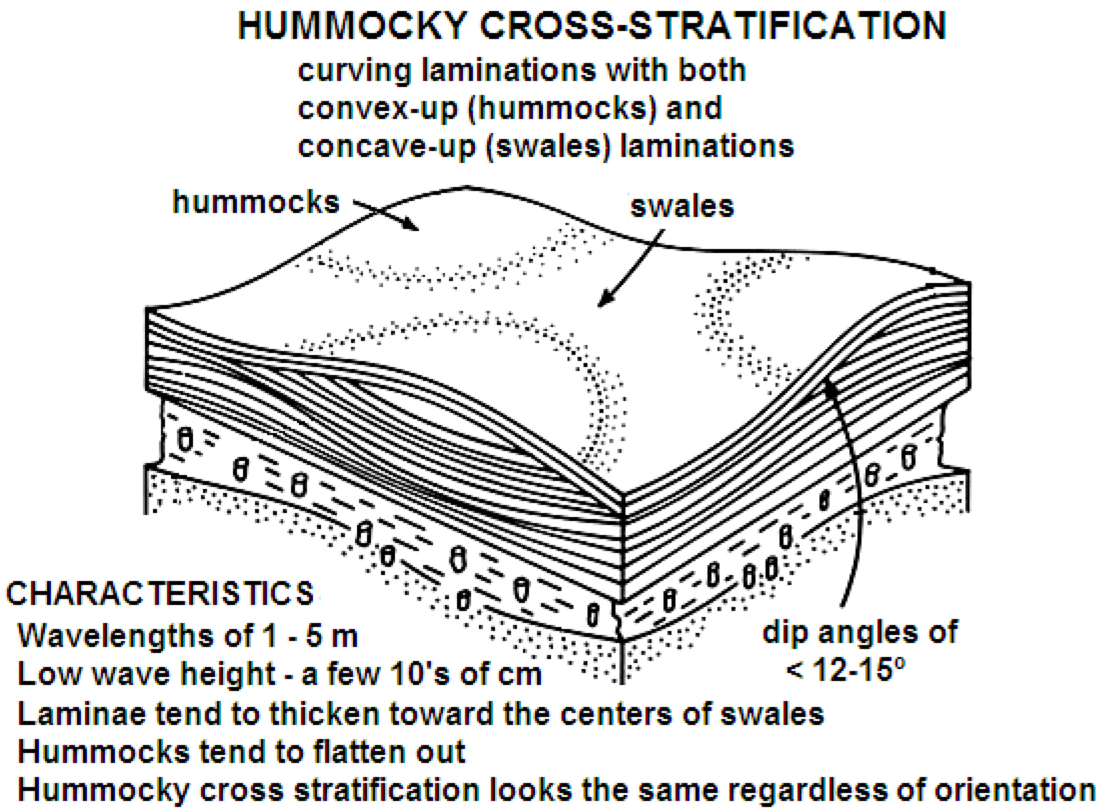


Figure 35
Characteristics and overall appearance of hummocky cross-stratification (modified from Walker, 1979).

sandstones in the subsurface of eastern Kentucky and West Virginia (Vargo and Matchen, 1996).

Mileage

46.4 State Route 799 passes over I-64.

46.5 Nada Member of the Borden Formation as well as the Renfro and St. Louis members of the Slade Formation (Fig. 38) are exposed in road cuts on both sides of I-64. The cut along the westbound lane is Stop 5.

STOP 5: Borden delta destruction: The Nada Member of the Borden Formation

Frank R. Ettensohn, R. Thomas Lierman, D. Brent Wilhelm, and Charles E. Mason

This stop illustrates the Cowbell-Nada and Nada-Slade (Renfro) contacts and shows a complete section of the Nada Member of the Borden Formation and five shallow-water carbonate units in the overlying Slade Formation (Fig. 38), most of which are separated from each other and the Nada by unconformities. The Nada Member is the uppermost member of the Borden Formation in northeastern Kentucky and shows features typical of sediment starvation and delta destruction.

Mileage

67.0 **Stop 5.** From this point, the drivers and the vehicles will proceed to the first Olive Hill exit (Exit 156), loop around and return to Stop 5 along the westbound lane. Located in a road cut along the westbound lane of I-64, 0.5 mi east of the Kentucky State Route 799 overpass and the top of the section at Stop 4, which ends here. Latitude: 38°, 16', 34.6"; longitude: 83°, 22', and 30.1".

This exposure occurs at the junction of the Cranston and Soldier quadrangles (Philley *et al.*, 1974, 1975) and exhibits a complete section of the Nada Member, the uppermost unit in the Borden Formation of northeastern Kentucky. The Nada Member here is composed of approximately 14.5 m (47 ft) of grayish blue-green to reddish-brown silty mudstones and clay shales with interbedded limestones and siltstones. The entire sequence is glauconitic, but glauconite is especially concentrated in three zones near the base and top of the unit (Fig. 38). The uppermost glauconite zone, about 1.4 m (5.3 ft) below the orange-brown Renfro dolostones, is intensely glauconitic with phosphorite nodules and is recognized as the Floyds Knob Bed of Stockdale (1939), a widespread horizon that has been interpreted to have temporal significance throughout the region (Kepferle, 1971; Whitehead, 1978). The Floyds Knob Bed is here associated with crinoidal calcarenite beds, a rather unusual occurrence. Also associated with the glauconite is the unusual blue-green color of most Nada shales, a characteristic related to the "verdine facies" of Odin (*e.g.*, 1988, 1990). Based on conodonts and ammonoids, the Nada is late Tournaisian (middle Osagean) in age (Work and Mason, 2003).

Overall, the Nada is one of the most fossiliferous of the Borden deltaic units. Megafossils are relatively common and include a moderately diverse fauna of fenestrate and ramose bryozoans; spiriferid, productid, and lingulid brachiopods; crinoids; pelmatozoan debris; gastropods; conularids; solitary corals; bivalves; fish plates and teeth; as well as occasional cephalopods. All of these are more common in limestone parts of the unit. Microfossils include conodonts and agglutinated foraminifera, and the unit includes an abundant, low-diversity ichnofauna, including *Zoophycos, Cruziana, Lophoctenium, Monocraterion, Palaeophycus, Phycosiphon, Phycodes, Arthrophycus, Skolithos, Planolites, Helminthoida, Teichichnus,* and *Scalarituba*; these traces have been interpreted to belong to the *Cruziana-Skolithus* ichnofacies

Figure 36
Section of swaley crossbedding (one swale is outline with white dashes) form the Cowbell member at Stop 4.

(see Appendix 1; Chaplin and Mason, 1979; Chaplin, 1980, 1982; Chaplin *et al.*, 1985).

Sedimentary structures in the Nada include ripple marks, crudely graded bedding and hummocky cross-strata in the siltstones; in contrast, low-angle, planar crossbeds and plane laminae are more common in the carbonates.

The contact between the Cowbell and overlying Nada is conformable and gradational across a short distance. Although in many places the contact between the Nada and overlying Renfro appears to be gradational, here the contact is apparently unconformable. According to Wilhelm (2008), the Renfro normally consists of five sequences in lower, middle, and upper parts, and the unit is broadly transgressive from south to north, crossing a series of progressively more uplifted basement fault blocks in that direction. At this locality, we are on the most uplifted of the fault blocks, so that only the thinned upper part of the Renfro is present. Wilhelm (2008) has also suggested that much of the missing Renfro time is incorporated into sediment-starved parts of the Nada. If that is the case, then most of this time was incorporated into the upper 1.4 m of the Nada (and any eroded section) above the Floyds Knob Bed, because in more complete Renfro sections to the south, the Floyds Knob Bed occurs in the upper Nada just below the lower part of the Renfro (Wilhelm, 2008).

The Renfro Dolostone and overlying carbonates are lower parts of the Slade Formation (Ettensohn *et al.*, 1984), a shallow-water, carbonate-platform unit that is largely equivalent to the better known Greenbrier, Newman, and Tuscumbia/Monteagle/Bangor limestones to the east and south in the Appalachian Basin. The Renfro Member is separated from the overlying St. Louis Member by a prominent regional unconformity; both of these units here are mid-Visean (Meramecian) in age. The St. Louis is capped here by a prominent paleosol and a major composite unconformity (Ettensohn *et al.*, 1988b). The Holly Fork, Tygarts Creek, and Ramey Creek members of the Slade overlie the St. Louis Member in ascending order and are

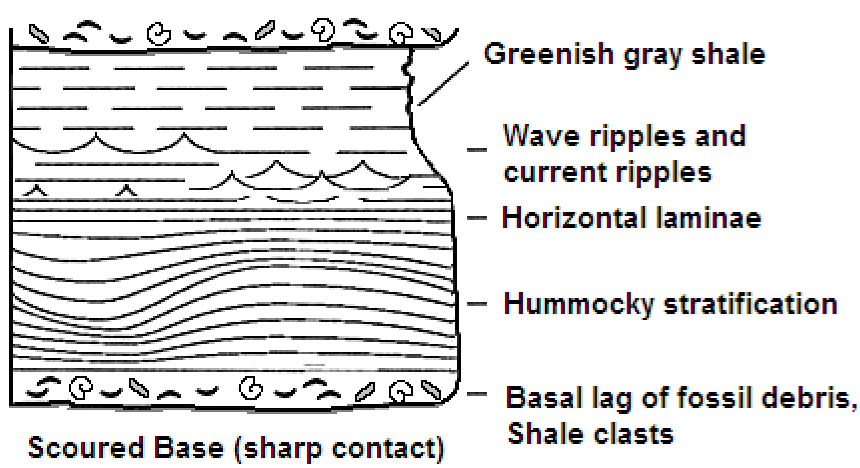


Figure 37
Typical storm-bed (tempestite) profile found in Unit 4 at Stop 4.

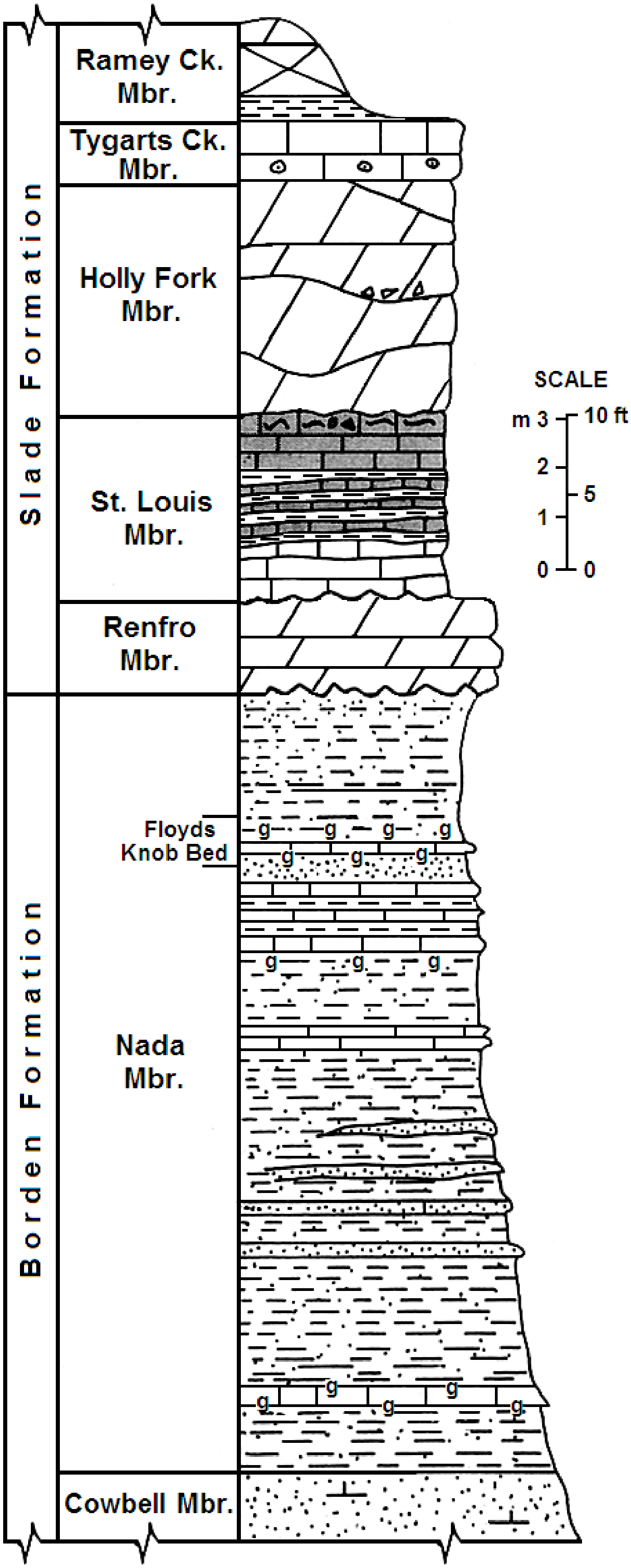


Figure 38
Schematic drawing of the Borden-Slade section at Stop 5 (adapted from Ettensohn, 1981). Note the occurrence of the three glauconite horizons in the Nada Member.

latest Visean to earliest Serpukhovian (mid-Chesterian) in age. These units were deposited in tidal-flat, sandbelt, and shallow open-marine environments, respectively; they are components in one of several Middle and Late Mississippian transgressive events, which moved across the widespread carbonate platform (Ettensohn, 1979, 1981, 1992d, e; Ettensohn and Dever, 1979) that had formed after basin infilling by the Borden delta and its equivalents (Ettensohn, 1995, 2004, 2005, 2008; Ettensohn *et al.*, 2002, 2004).

Underlying parts of the Borden compose a typical coarsening-upward, prograding, deltaic sequence. A normal deltaic sequence, however, would also include marginal-marine to terrestrial lower delta-plain environments, and although such environments are present in the equivalent Pocono Formation on the eastern side of the Appalachian Basin, they are absent in Kentucky. In their place are the shallow-shelf environments of the Nada Member. For some time it was unclear why major clastic influx halted at the Cowbell-Nada transition and why Pocono delta-plain environments did not prograde farther to the west. Kearby (1971) suggested that uplift on a local structure, the Waverly Arch, caused the delta diversion and abandonment, and although this might explain conditions in the area of the structure, it does not explain the delta abandonment across large parts of the Appalachian Basin. Using flexural models, however, Ettensohn (1994, 2004) and Ettensohn *et al.* (2002, 2004) suggested that bulge uplift and migration in eastern parts of the basin, followed by a sea-level lowstand, disrupted deltaic sedimentation, causing sediment starvation and delta destruction in distal parts of the basin west of the bulge. Further dissipation of wave energy at the margin of the delta platform may have contributed to the overall fine-grained nature of Nada deposition. This kind of situation explains well some of the features observed in the Nada. Absence of clastic influx in upper-slope, outer-shelf, and platform environments is conducive for the deposition of glauconite and phosphorite (Odin and Matter, 1981; Carozzi, 1960), especially during maximum-flooding events like that suggested for the Floyds Knob Bed. Verdine facies also occur in sediment-starved conditions, but in iron-rich coastal waters in continental-margin settings with waters 30 to 60 m deep (Odin, 1988, 1990). Moreover, the presence of hummocky crossbedding and tempestite sequences suggests that storms frequently reworked the abandoned delta platform, while some of the thicker limestone beds may represent reworked biohermal or biostromal accumulations of benthic faunas that developed on firm substrates on the shallowing, sediment-starved platform. Hence, the Nada has been interpreted to represent shallowing from the deeper, Cowbell delta-front into upper-slope, storm-shelf conditions; this shallowing occurred in a setting with sharply reduced clastic sedimentation on subaqueous parts of an abandoned delta platform, subject to reworking by storms (Ettensohn *et al.*, 2002). Although tectonic and eustatic changes no doubt played a role, by the end of Nada deposition these conditions had generated the shallow, subtropical, reduced-clastic, platform-to-ramp setting necessary for the beginning of Slade carbonate deposition. Other discussions of the Nada at this stop and elsewhere in the area can be found in work by Chaplin and Mason (1979), Chaplin (1980, 1982, 1985), Ettensohn (1981), Chaplin *et al*. (1985), Lierman and Mason (2004), Ettensohn *et al*. (2004) and Wilhelm (2008).

Mileage

67.2 State Route 799 passes over I-64. Upper parts of the Cowbell Member of the Borden Formation are exposed in road cuts along both sides of the Interstate, especially under the overpass.

67.8 The Cowbell Member of the Borden Formation is exposed almost continuously in road-cut exposures along the westbound lane for the next 1.3 mi.

STOP 6: Transition between the Nancy and Cowbell members of the Borden Formation, northeastern Kentucky

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

This stop illustrates the contact between the Nancy and Cowbell members of the Borden Formation and the transitional interval between (Fig. 39). Sedimentary structures, body fossils, and trace fossils are present.

Mileage

69.1 Stop 6. This stop will only be examined if time is available. Stop 6: It is located in a road cut along the westbound lanes of I-64 (at mile marker 144), 2.0 miles west of Stop 5. At this stop, if time permits, we will view the transitional contact between the Nancy and Cowbell members of the Borden Formation (Fig. 39). Latitude: 38°, 16', 4.4"; longitude: 83°, 24', 30.2". This stop is located on the Cranston quadrangle at the center of the border between rectangles 8 and 9.

Only the uppermost 7.9 m (26.2 ft) of the Nancy is exposed at this stop (Fig. 39). The entire Nancy Member in the Morehead area ranges in thickness from 18.2 to 60.9 m (60–200 ft). However, it should be noted here that the maximum thicknesses for the Nancy Member on USGS geologic quadrangle maps are often misleading, as mappers have not consistently placed the lower or the upper contacts at the same stratigraphic position due to their gradational nature. The Nancy consists of bluish- to greenish-gray, silty shale, which weathers olive gray to yellowish gray. The member contains ironstone nodules, lenses, and beds throughout, but they are found most abundantly in its lower part. Locally, a single siltstone bed up to 60.9 cm (24 in) thick is common in the lower 9 m (30 ft), and it is similar in lithologic character to siltstone beds in the underlying Farmers Member.

The lower boundary of the Nancy is gradational and is generally placed at the top of the last continuous sequence of siltstone beds. Its gradational upper contact is placed at the base of the first thick continuous sequence of siltstone beds above which the thickness of siltstone exceeds the thickness of the shale intervals. The upper 2.5 m (8.2 ft) of the Nancy Member, at this stop, becomes very silty, reflecting the gradation of the

Figure 39
Photograph showing the Nancy-Cowbell transition in the Borden Formation at Stop 6.

Nancy shale lithology into the more silty, overlying, Cowbell Member (Fig. 39).

The Nancy is highly bioturbated so that most traces of bedding and other sedimentary structures are destroyed. Trace-fossil diversity is low with the ichnogenera *Phycosiphon*, *Scalarituba*, and *Zoophycos* being the most common (see Appendix 1). *Zoophycos* is locally most abundant in the siltier shale intervals. Other associated trace fossils include horizontal feeding traces and vertical burrows. At the top of the Nancy, at the east end of the road cut, a 4 cm (1.6 in) bed contains large (up to 1.5 cm in diameter) horizontal back-filled burrows. See Chaplin (1980) for a more complete coverage of the ichnofauna of this and other members of the Borden Formation.

Body fossils are more commonly found in the ironstone concretions, lenses, and beds, which occur in the Nancy Member. These ironstone concretions, lenses, and beds are locally very fossiliferous. The most common fossil forms include brachiopods, cephalopods, gastropods, and conularids. Less abundant associated forms include trilobite fragments, bryozoans, crinoid detritus, bivalves, hyolithids, ostracods, corals, conodonts, foraminifera, as well as phyllocarids and fish remains. Fossils in the Nancy are most commonly preserved as internal molds and casts, generally infilled or replaced with barite, sphalerite, pyrite, galena, and on occasion, marcasite, dolomite, and quartz.

The Cowbell Member at the stop is approximately 76 m (250 ft) thick. Locally the Cowbell ranges in thickness from 67 to 111 m (220–365 ft) (Philley *et al.*, 1974). The Cowbell consists of primarily light yellowish-brown to grayish-olive, ferruginous, micaceous siltstone that is slightly calcareous in the upper part. Siltstones in the Cowbell are commonly limonite-stained, bioturbated, and generally contain reddish-brown siderite nodules and lenses. These siderite nodules and lenses are locally very fossiliferous. Variable thicknesses of bluish-gray silty shale occur throughout the unit but are most abundant in the lower and upper parts. Primary sedimentary structures include cross-strata, parallel laminae, ripple marks, ripple-drift cross laminae, cut-and-fill, and small-scale, low-angle crossbedding. The lower contact of the Cowbell is gradational through approximately a meter at this stop. Kearby (1971) subdivided the Cowbell Member in this area into several informal lithologic units. Kearby's lower massive siltstone unit is exposed in the lower and middle portions of the road cut, and his lower dark shale unit is exposed near the top of the cut.

Trace fossils are common throughout the Cowbell Member. Some of the more common ichnogenera include *Bergaueria*, *Cylindrichnus*, *Scalarituba* (*Phyllodocites* view), and *Zoophycos*. Less commonly found forms include *Asteriacites* and *Cruziana* (see Appendix 1). Numerous horizontal and vertical burrows of uncertain trace-fossil affinities can be found throughout the Cowbell as well. Again, for an in-depth coverage of the ichnofauna found in this member as well as other members of the Borden Formation, see Chaplin (1980).

Megafossils, though found throughout the Cowbell interval, are most abundant in the upper 15.2 m (50 ft) of the member in northeastern Kentucky. They are most commonly found in siderite nodules and as lag deposits in small cut-and-fill channels. The most commonly encountered fossils include bivalves, brachiopods, fenestrate bryozoans, cephalopods, conularids, rugose corals, crinoid columnals, gastropods, and trilobite fragments. Collectively, they represent a good open-marine fauna.

Kearby (1971, p. 59) interpreted the Cowbell Member of the Borden Formation in northeastern Kentucky to represent a lower delta-front deposit (Fig. 25), primarily a distal-bar deposit, made up of terrigenous clastic sediments derived from a large river system to the northeast. Chaplin (1980, 1982, 1985) also examined the depositional environment of the Cowbell Member as well as the other members of the Borden Formation. He also interpreted the Cowbell Member to be a delta-front deposit and the Nancy Member to be its prodelta equivalent.

The Borden Formation in northeastern Kentucky contains a significant, largely undescribed ammonoid succession. Recent studies by Work and Manger (2002) and Work and Mason (2003, 2005) have begun working out the biostratigraphic framework of the Borden Formation in northeastern Kentucky, utilizing the ammonoids in this succession as well as their associated conodonts. Based on these studies, the Nancy Member through the lower Cowbell Member, in the Morehead area, is early Osagean (Fern Glen or lower Burlington equivalent) in age, which equates to the lower Ivorian stage of the Belgian upper Tournaisian succession (Work and Manger, 2003; Work and Mason, 2005). An earlier study by Work and Mason (2003) examined the ammonoid and conodont evidence found in the Nada Member of the Borden Formation in northeastern Kentucky and found it to restrict the age of this interval to a relatively narrow range of the late middle Osagean age corresponding to the latest Tournaisian or possibly earliest Visean. The intervening ammonoid fauna found in the middle and upper parts of the Cowbell Member is yet to be described. However, preliminary examination of ammonoids obtained from this interval suggests an early middle Osagean age for the stratigraphic interval in the Morehead area.

Note: From this point, we will retrace our route back to the parking lot behind Hardee's and in front of Food Lion in the Pine Crest Plaza.

Mileage

72.6 Rest area along the westbound lane.

75.6 Morehead exit (Exit 137) ramp.

75.9 Traffic light at base of ramp and junction of State Route 32. Turn left (east) onto State Route 32.

76.0 Pass under I-64.

76.1 First traffic light after passing under I-64 and the junction of I-64 eastbound on and off ramps. Continue eastward on State Route 32 through the light.

76.2 Turn right in front of Hardee's onto the blacktop access road into Pine Crest Plaza and proceed to the parking lot behind Hardee's and in front of Food Lion.

76.3 Parking lot behind Hardies and in front of Food Lion. **End of field trip!**

Preliminary palynological results from the Late Devonian–Early Mississippian of Morehead and adjacent areas

Sarah Heal, Niall Paterson, Cortland Eble, Charles E. Mason, Robbie Goodhue, Nina Larsson, and Geoff Clayton

Introduction. The following brief account of the palynology of selected sections in Kentucky is based on work in progress by the authors and described by Coleman (1991), Coleman and Clayton (1987), Heal (2009), and Paterson (2009). The miospore assemblages described can be tentatively assigned to the Western European miospore zonation (Clayton *et al.,* 1977; Higgs *et al.,* 1988), though the Tournaisian *Cristatisporites hibernicus–Umbonatisporites distinctus* (HD) Biozone has not yet been recognised. Based on these results, correlations with the Western European Mississippian regional substages are outlined. Acritarchs and prasinophytes are abundant in many of the Devonian samples but are of limited biostratigraphic value and so are not discussed here.

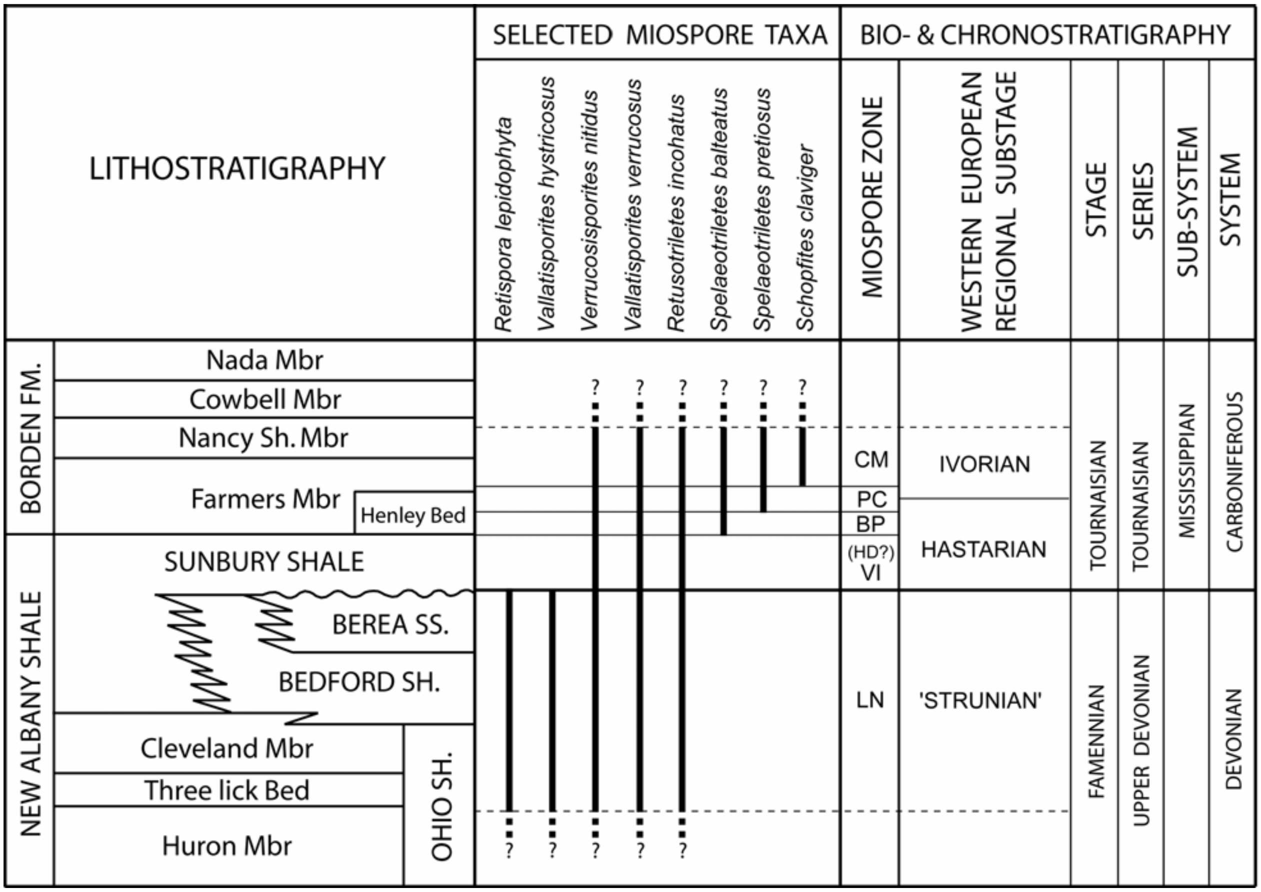


Figure 40
Generalized ranges of selected miospore taxa with chronostratigraphic interpretation, including Western European regional substages.

The results of palynofacies analysis of the organic residues extracted from selected samples are briefly described. These data throw some light on the depositional environments represented by some of the lithostratigraphic units, especially the pale gray shales of the Three Lick Bed and the Bedford Shale.

Stratigraphic palynology. Key miospore taxa are illustrated in Plate 2. The ranges of selected stratigraphically significant taxa are shown in Figure 40 together with their chronostratigraphic interpretation.

Ohio Shale–Three Lick Bed. Samples from the Three Lick Bed at its type section along I-64 near Morehead yielded moderately diverse and generally well-preserved miospore assemblages including *Retispora lepidophyta* (Kedo) Playford and *Verrucosisporites nitidus* (Naumova) Playford. The combined occurrence of these taxa indicates a position within the latest Devonian ("Strunian") *Retispora lepidophyta–Verrucosisporites nitidus* (LN) Biozone of Western Europe. More diverse assemblages were obtained from the Three Lick Bed in the Clay City Quarry (Powell County, Kentucky). Here, in addition to the index species, taxa typical of the LN Biozone were recorded, including *Densosporites spitsbergensis* Playford, *Indotriradites explanatus* (Luber) Playford, *Knoxisporites concentricus* (Byvscheva) Playford and McGregor, *Latosporites* sp., and *Retusotriletes crassus* Clayton, Johnston, Sevastopulo and Smith. In both localities, the most diverse spore assemblages were obtained from the three light greenish gray shale beds, the intercalated black shales being dominated by amorphous organic matter (AOM).

Ohio Shale (above the Three Lick) and Bedford Shale. Diverse and well-preserved miospore assemblages were recovered from the Ohio Shale and Bedford Shale in the I-64 roadcuts near Morehead, all of which are assigned to the LN Biozone. In addition to all of the miospore taxa noted above, *Vallatisporites verrucosus* Hacquebard is present, whose first occurrence in Europe occurs at the base of the LN Biozone.

Diverse miospore assemblages were also recovered from the Robinson dropstone locality at Logan Hollow, from immediately below the dropstone in the Ohio Shale and from about 1-m above the base of the overlying Bedford Shale. Both are confidently assigned to the LN Biozone, suggesting that the strata containing the dropstone are latest Devonian in age.

Sunbury Shale. Palynomorph assemblages from the Sunbury Shale in the Morehead I-64 roadcuts are poorly preserved and of relatively low diversity. Miospore assemblages are dominated by long-ranging taxa persisting from the Late Devonian, such as *Apiculiretusipora fructicosa* Higgs, *Auroraspora macra* Sullivan, *Calamospora* spp., *Punctatisporites* spp., *Retusotriletes incohatus* Sullivan, and *Verrucosisporites nitidus.*

In Western Europe, the Devonian/Carboniferous boundary is coincident with the base of the *Vallatisporites verrucosus–Retusotriletes incohatus* (VI) Miospore Biozone, which is defined by the top of the range of *Retispora lepidophyta.* Numerous other "Devonian" taxa disappear at this level, including *Ancyrospora* app., *Rugospora flexuosa* (Jushko) Streel, *R. radiata* (Jushko) Byvsheva, and *Vallatisporites hystricosus* (Winslow) Byvsheva. Miospore assemblages from the Sunbury Shale are assigned to this earliest Carboniferous VI Biozone. In the Morehead I-64 section, the highest LN Biozone assemblage was obtained from the top 8 cm of the Bedford, and the lowest VI Biozone assemblage from 15 cm above the base of the Sunbury, indicating that the base of the Carboniferous System is at or very close to the latter.

Closely comparable assemblages were obtained from the Sunbury Shale, as well as from the Jacobs Chapel and Rockford Limestone equivalents found in the basal 45 cm of the Henley Bed in the AA Highway section at Brightman Cemetery, Vanceburg, Kentucky. Similar assemblages have also been described from the early Kinderhookian Hannibal Shale in Missouri (Heal and Clayton, 2008).

The VI Biozone in Western Europe is succeeded by the HD, *Spelaeotriletes balteatus–Umbonatisporites distinctus* (BP) and *Spelaeotriletes pretiosus–Raistrickia clavata* (PC) biozones, the bases of which are defined by the first occurrences of *Cristatisporites hibernicus* (Higgs) Higgs, *Spelaeotriletes balteatus* (Playford) Higgs, and *Spelaeotriletes pretiosus* (Playford) Neves and Belt, respectively. The former species has not been recognized in the USA, so the HD Biozone cannot be recognized. Therefore, the upper part of the VI Biozone in this region may be somewhat younger than in Europe.

Borden Formation–Farmers Member, Henley Bed. Miospore assemblages from the Henley Bed at Morehead are restricted in composition at the base and top of the unit but are diverse in its middle part. *Spelaeotriletes balteatus* is present in the lower part of the Henley Member, permitting assignment to the BP Biozone of Western Europe. *Spelaeotriletes pretiosus* first appears in the upper part of the Henley Bed, defining the base of the succeeding PC Biozone. At Vanceburg, the upper part of the Henley Bed and the overlying Portsmouth Member are also assigned to the PC Biozone.

Borden Formation–Farmers Member and Nancy Member. The Farmers and Nancy members have not yet been sampled thoroughly in the Morehead area for palynology. However, the miospore assemblages that have been obtained from this interval are relatively diverse and well-preserved. In addition to the taxa encountered in the Henley Bed, assemblages from the overlying Farmers Member contain *Schopfites claviger* Sullivan. In Western Europe, the first appearance of this taxon defines the base of the *Schopfites claviger–Auroraspora macra* (CM) Biozone, which is of latest Tournaisian age. Miospore assembages from the Nancy Member are more diverse than those from the Farmers Member and include the first appearances of *Auroraspora solisorta* Hoffmeister, Staplin and Malloy, *Neoraistrickia cymosa* Higgs, Clayton and Keegan, and *Vallatisporites ciliaris* (Luber) Sullivan.

In Europe, the first appearance of *Lycospora pusilla* is just below the base of the Viséan Series and is used to define the base of the *Lycospora pusilla* (Pu) Biozone. No other miospore taxa

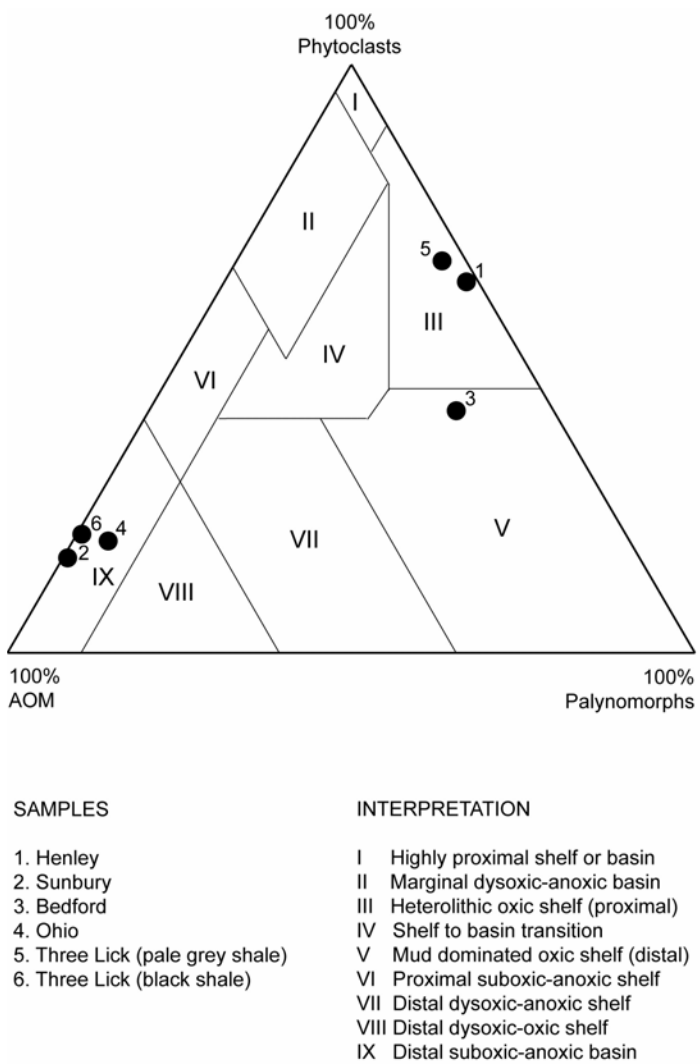


Figure 41
Palynofacies of selected samples; interpretation based on Tyson (1995).

first appear or die out around this stratigraphic level. In the Mississippi Valley succession the first appearance of *Lycospora pusilla* is considerably higher, within the uppermost Viséan Cypress Formation (Heal, 2009), so assemblages from the Borden Formation in Kentucky that are tentatively assigned here to the CM Biozone could be younger than the late Tournaisian age implied by correlation with Europe.

Palynofacies analysis. Representative organic residues are illustrated in Plate 3. The results of palynofacies analysis of selected samples are summarized in Figure 41. Following Tyson (1995), a simple classification of organic matter is used, recognizing palynomorphs (principally miospores, acritarchs, and prasinophytes), phytoclasts ("woody" debris), and amorphous organic matter (AOM).

Samples from the Three Lick Bed in the Clay City Quarry are particularly interesting with regard to their depositional environments. The two samples collected from the beds of dark black shale within the Three Lick Bed compare closely to the underlying and overlying New Albany Shale, being dominated by AOM (ca. 80 percent). Miospores are absent in both of these samples but rare *Tasmanites* spp. are present (about 3 percent). Both of these samples represent dysoxic-suboxic shelf to basin transition environments.

The kerogen content of samples from the three gray shales differs considerably from the intervening black shale. These samples contain a much higher proportion of terrigenous kerogen, particularly phytoclasts, accounting for 37–67 percent of the total organic matter. Miospores and *Tasmanites* spp. are present but are relatively rare (4–6 percent). AOM is relatively common in the lower two gray shales (ca. 40 percent), but far less abundant in the upper shale (4 percent).

The pale gray shales from the Three Lick Bed clearly show greater terrestrial influence. In terms of Tyson's (1995) paleoenvironmental classification, the lowermost pale shale represents a dysoxic-suboxic shelf to basin transition and the two higher shales represent oxic shelf environments. The gray shales probably represent deposition by turbidity currents, carrying fine-grained terrestrial material into distal parts of the basin, after which more normal-marine deposition resumed.

Around Morehead, the Bedford Shale contrasts strongly with the underlying and overlying black-shale units, containing substantial quantities of palynomorphs and phytoclasts. Its depositional environment is interpreted as oxic shelf (Fig. 41). The overlying Sunbury Shale is dominated by AOM throughout (>78 percent). All samples from this unit indicate a return to distal suboxic-anoxic basin environments.

The Henley Bed is varied in terms of palynofacies but is dominated by phytoclasts and palynomorphs, with very little AOM present. The oxic shelf depositional environment indicated is fully consistent with a prodelta palaeogeographic setting

Acknowledgments. This publication has emanated from research conducted with the financial support of Science Foundation Ireland (Research Frontiers Programme grant GEO0011).

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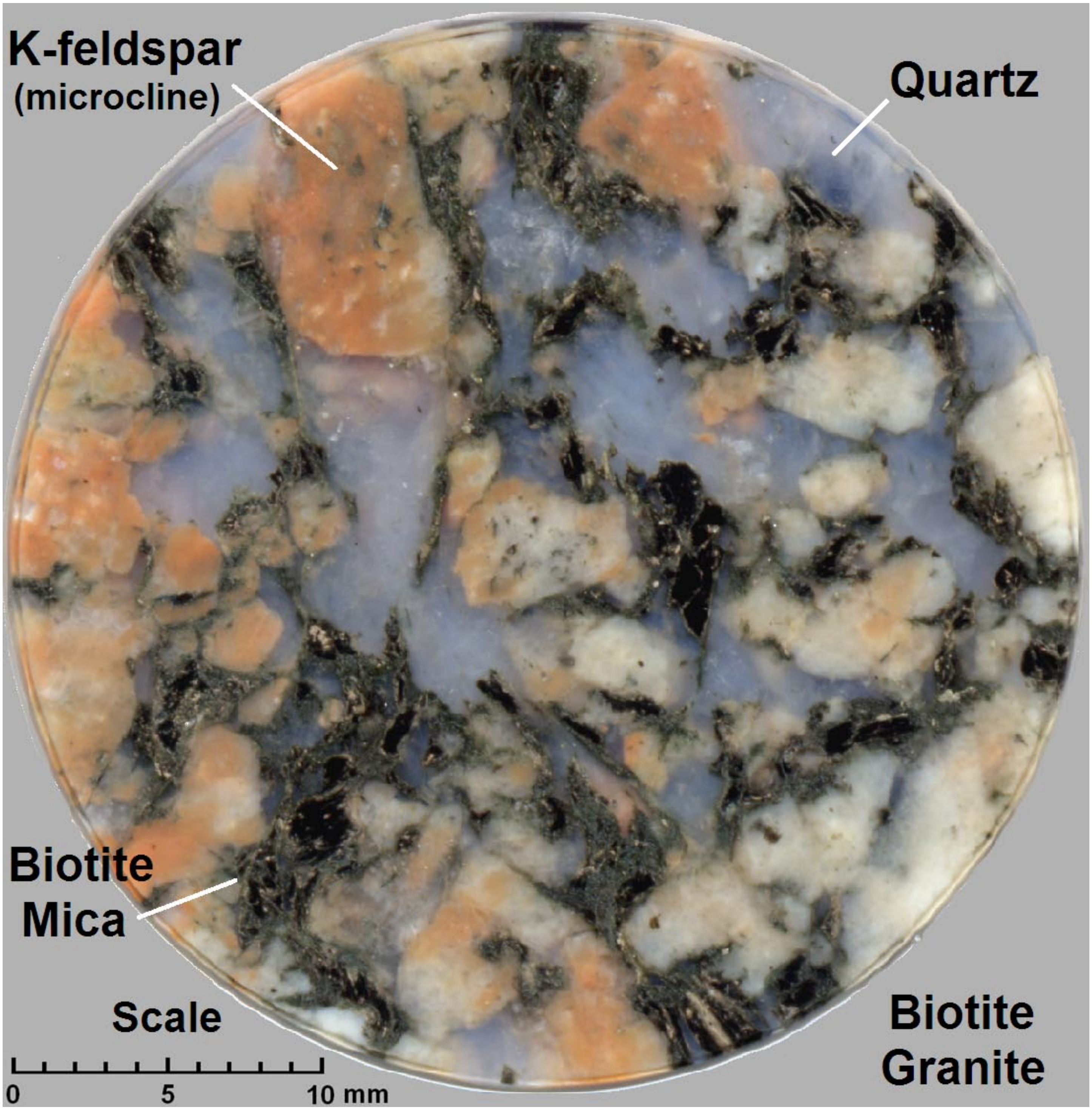


Plate 1
Close-up view of a polished core segment from the boulder, which is a biotite granite with quartz, K-feldspar (microcline), and biotite. Note the bluish tint of the quartz grains.

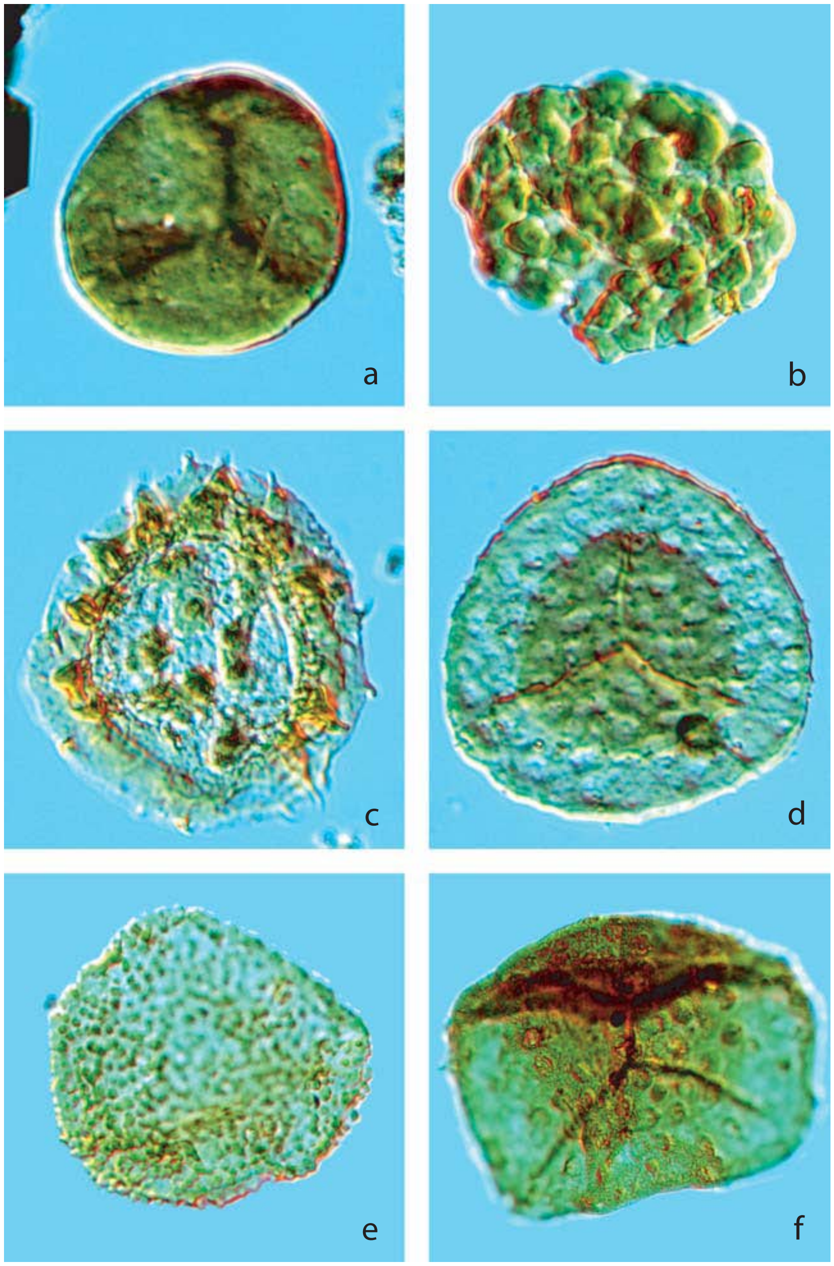


Plate 2
Photomicrographs of key miospore taxa (X 750 except for figure 5—X500)

a. *Retusotriletes incohatus* Sullivan. Bedford Shale, Logan Hollow.
b. *Verrucosisporites nitidus (Namova) Playford.* Henley Member, I-64, Morehead.
c. *Vallatisporites hystricosus (Winslow) Byvsheva.* Bedford Shale, Logan Hollow.
d. *Retispora lepidophyta* (Kedo) Playford. Bedford Shale, Logan Hollow.
e. *Spelaeotriletes balteatus* (Playford) Higgs. Henley Bed, I-64, Morehead.
f. *Spelaeotriletes pretiosus (Playford)* Neves & Belt. Henley Bed, I-64, Morehead.

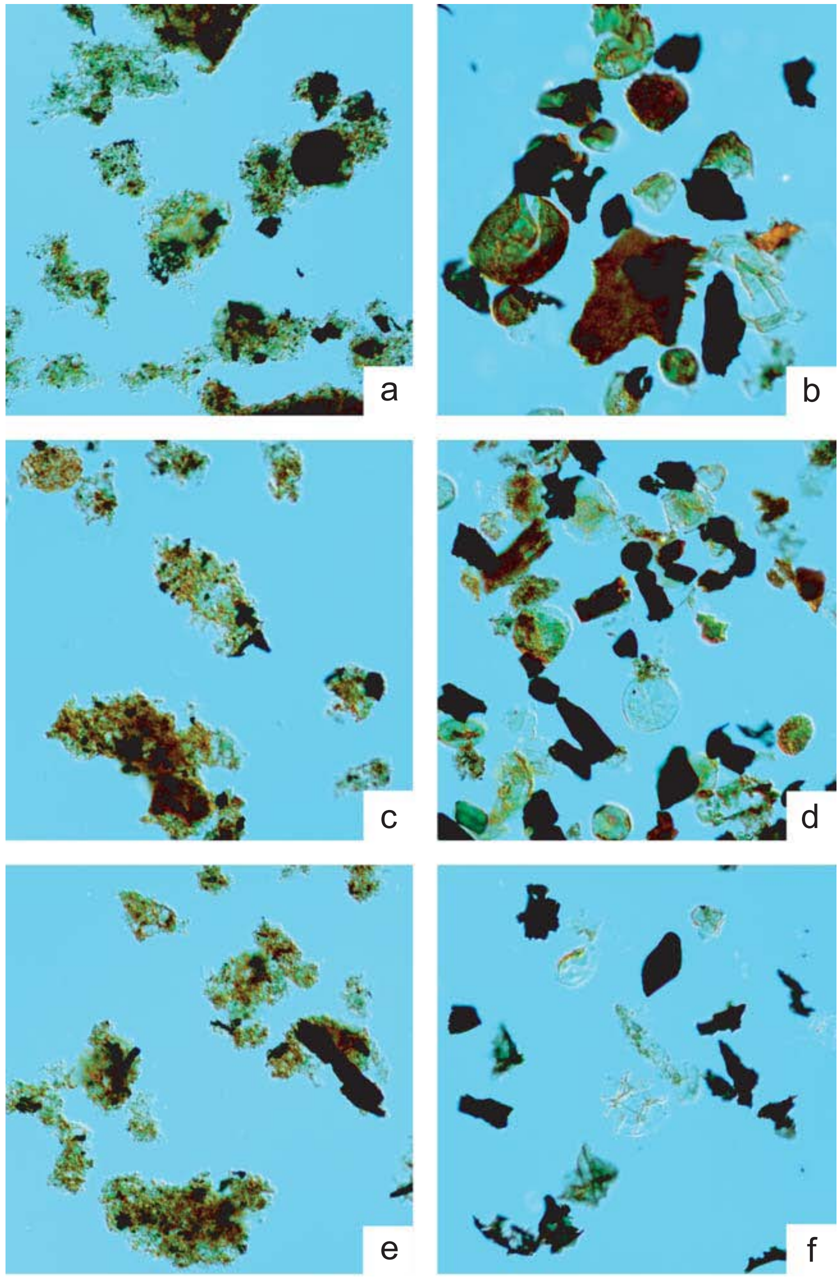
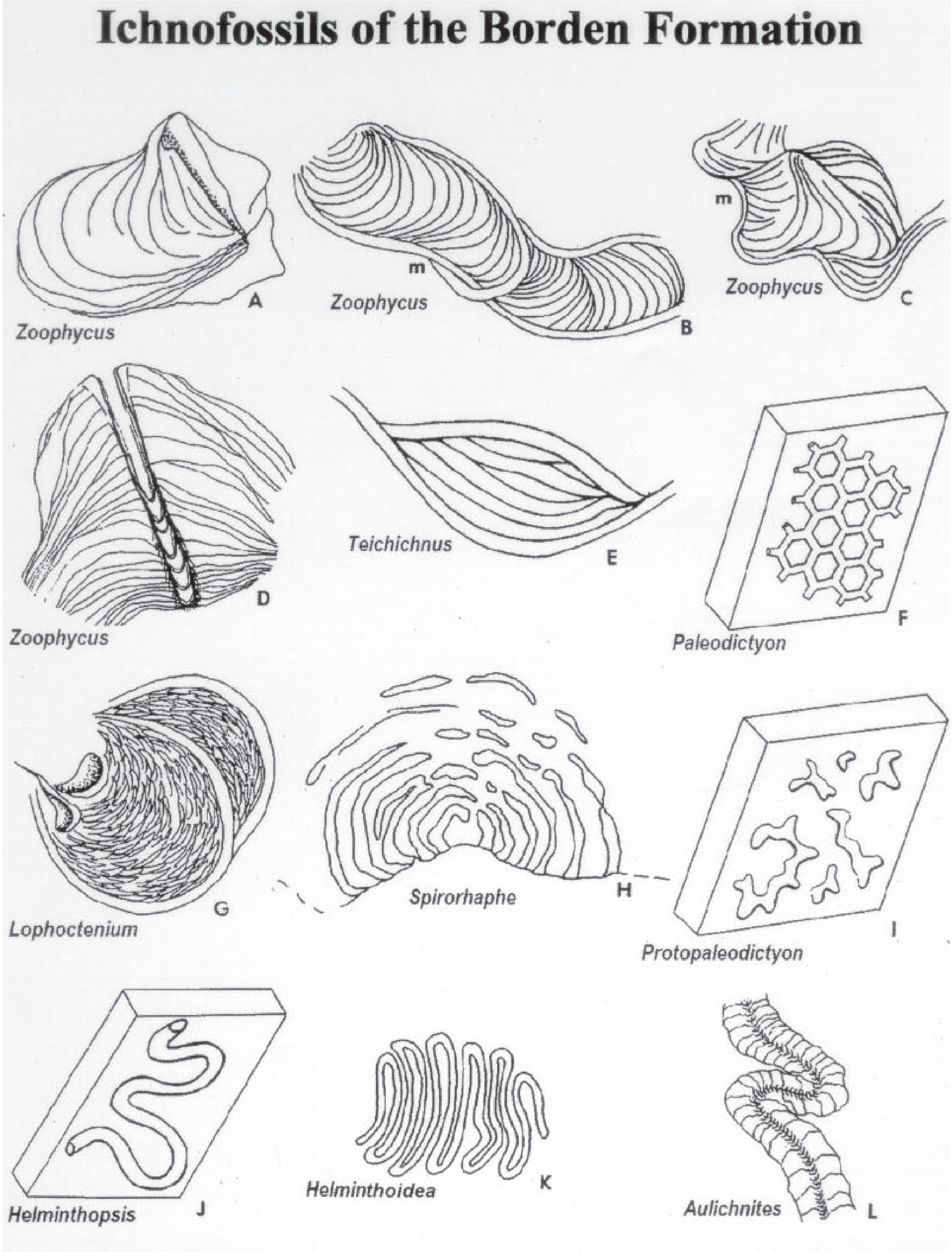


Plate 3
Photomicrographs of typical palynofacies (X 250)

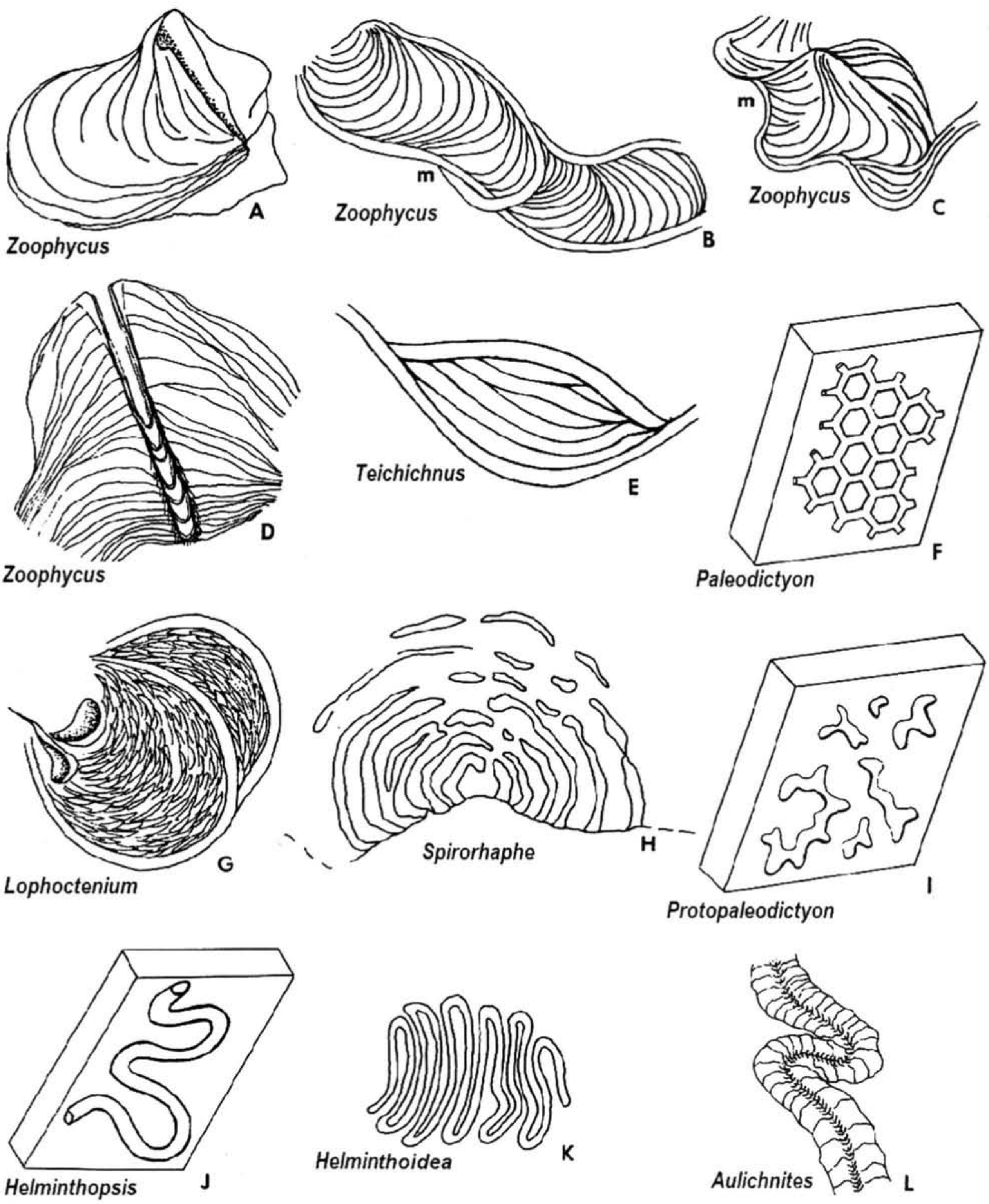
a. Sunbury Shale. I-64 road cut, Morehead.
b. Henley Bed. I-64 road cut, Morehead.
c. Ohio Shale. I-64 road cut, Morehead.
d. Bedford Shale. I-64 road cut, Morehead.
e. Three Lick Bed (black shale between top and middle "licks"). I-64 road cut, Morehead.
f. Three Lick Bed (top "lick"—pale gray shale). I-64 road cut, Morehead.

Appendix 1:

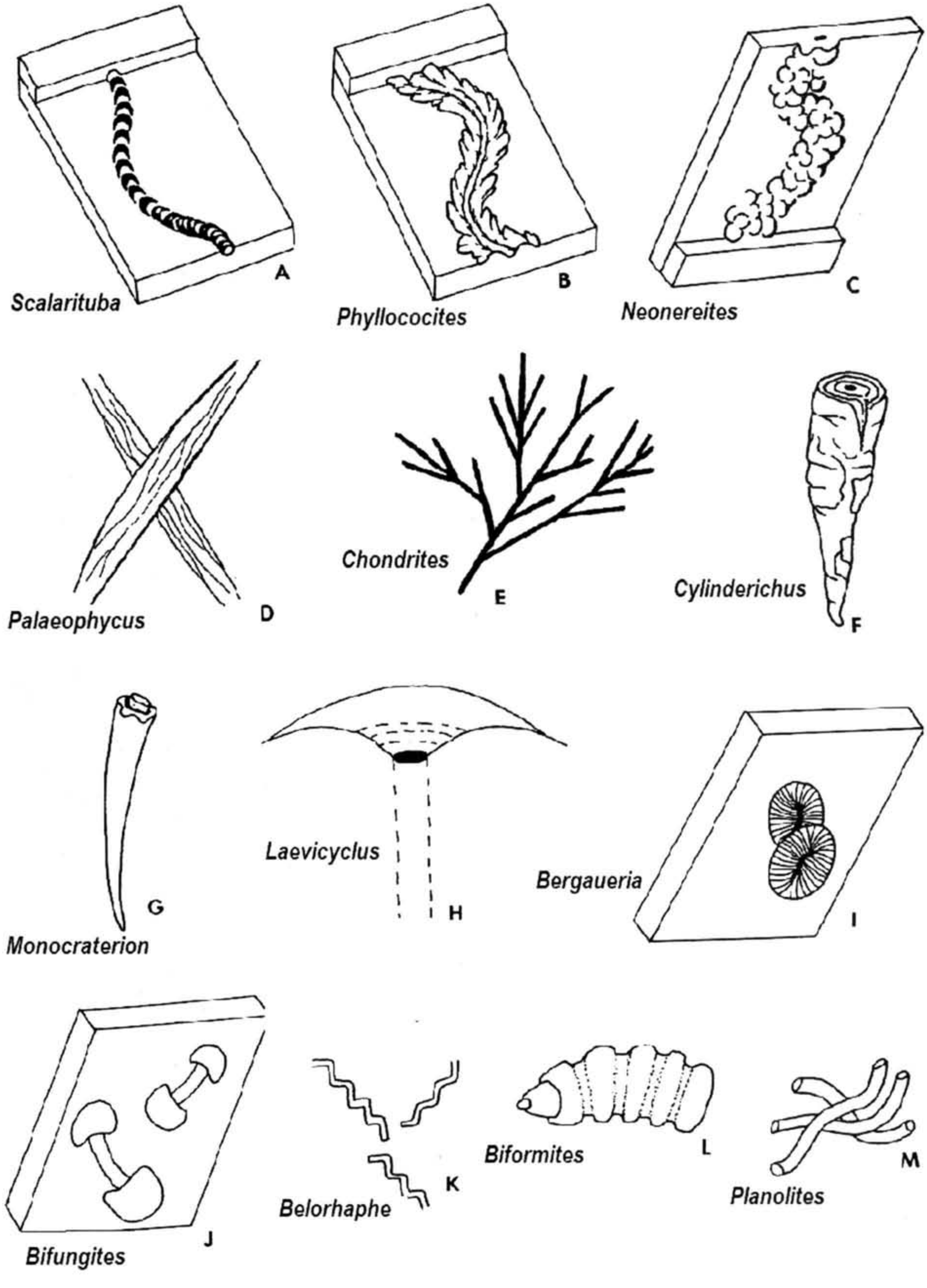


Trace fossils associated with the Nereites Ichnofacies. (Borden Formation)

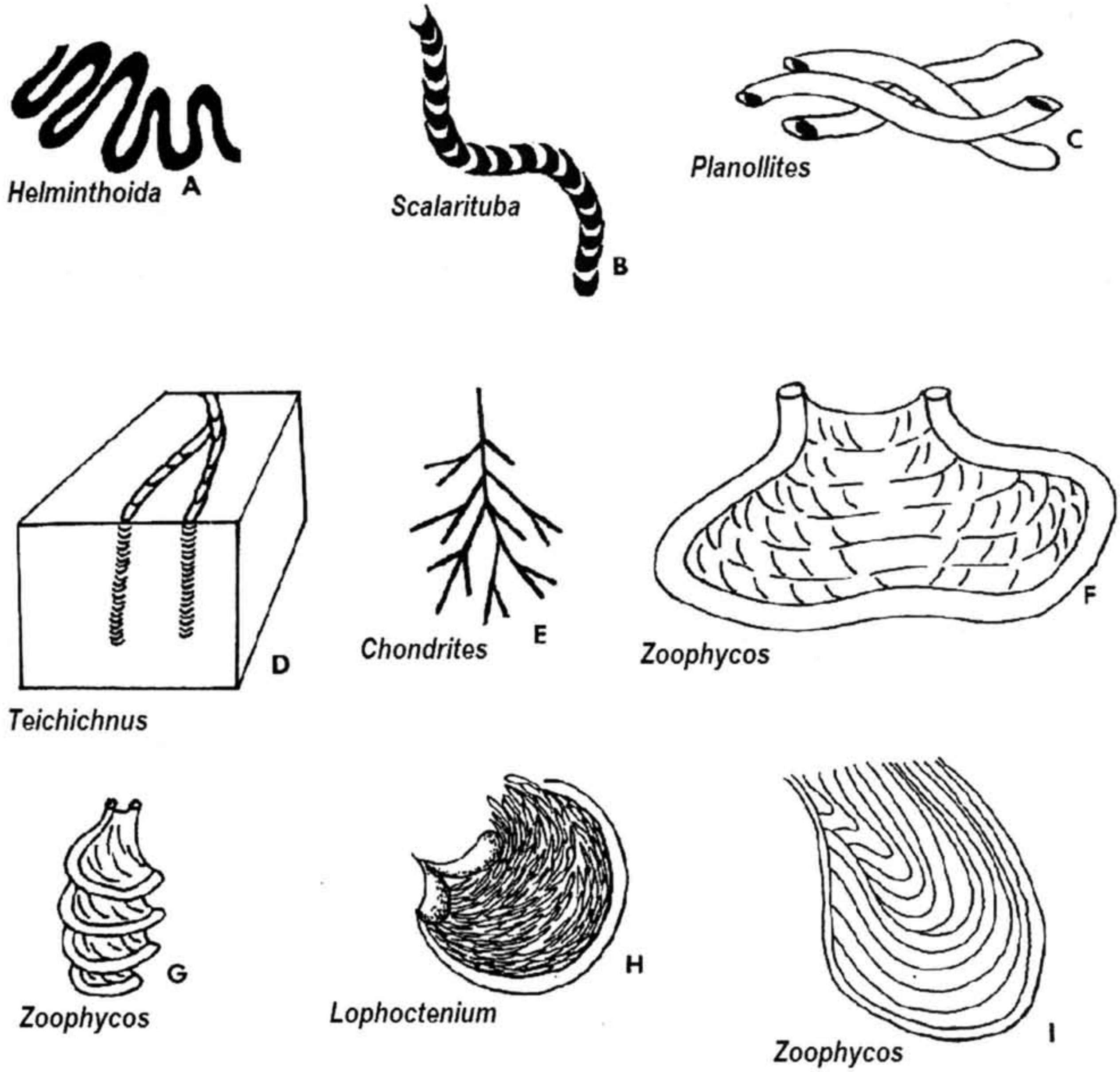
Ichnofossils of the Borden Formation

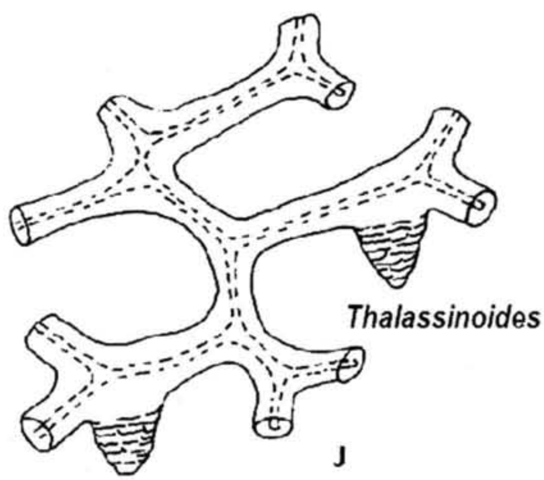


Trace fossils associated with the Nereites Ichnofacies. (Borden Formation)

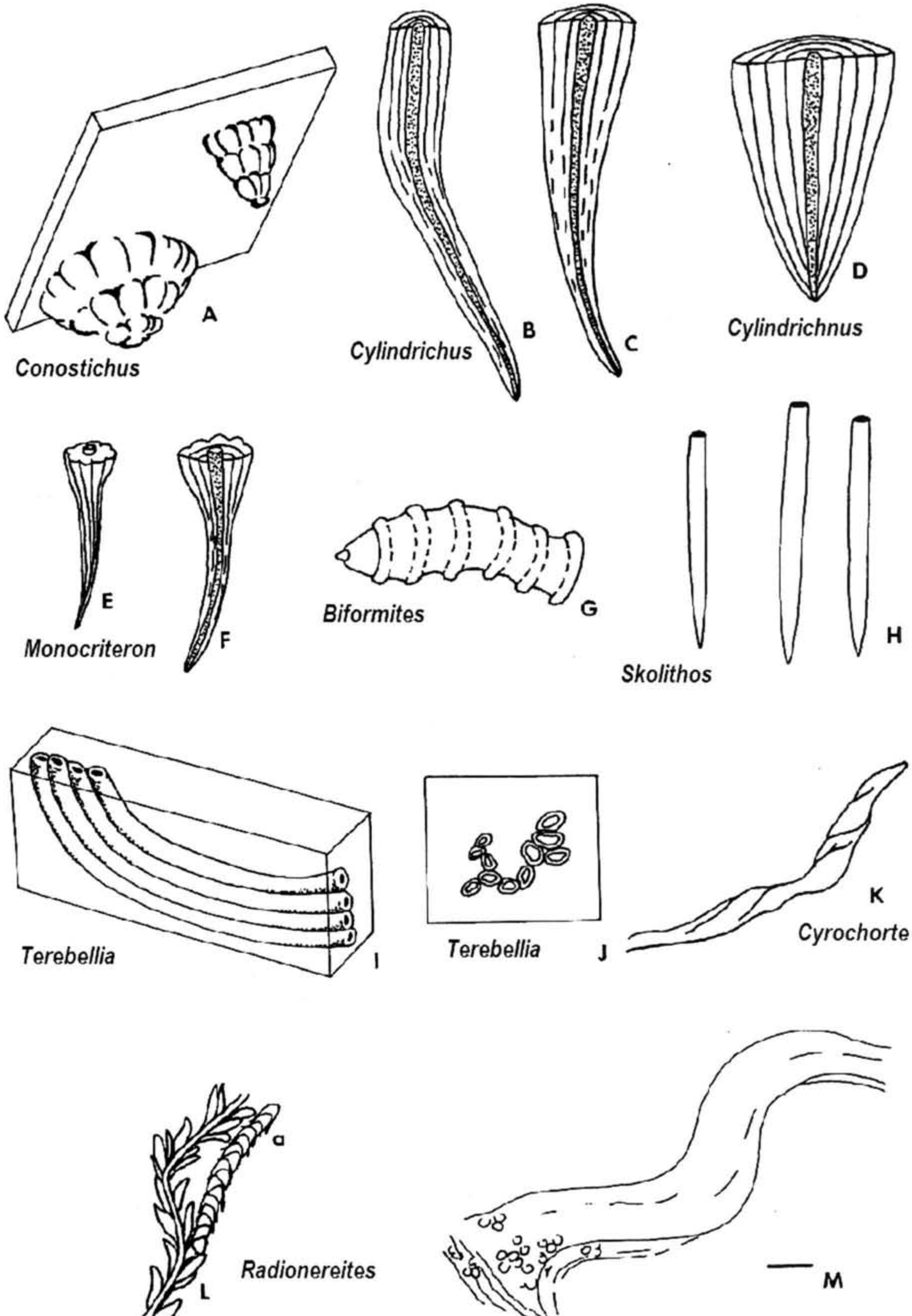


Trace fossils associated with the Nereites Ichnofacies. (Borden Formation)

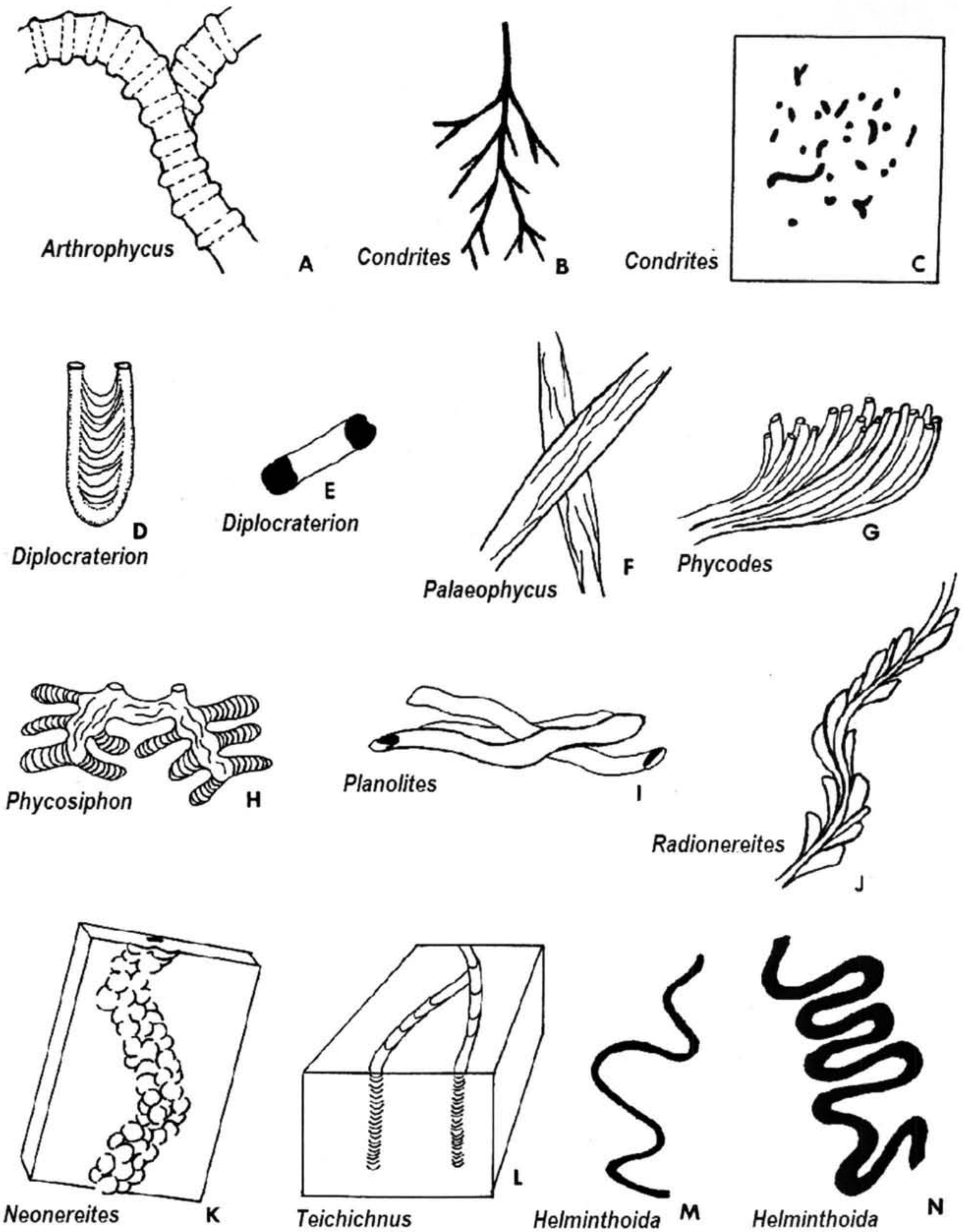




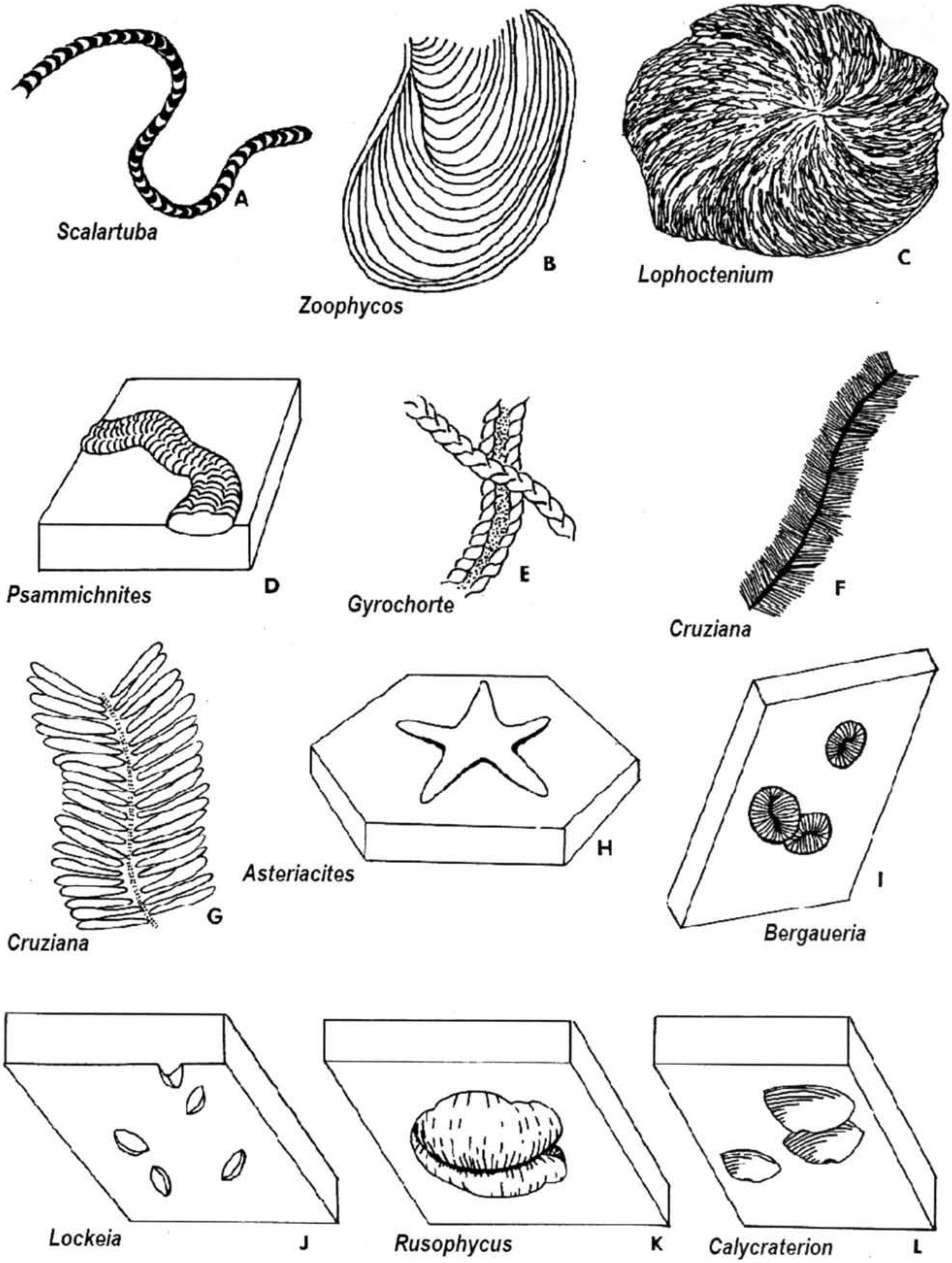
Trace fossils associated with the Zoophycos Ichnofacies (Borden Formation).



Trace fossils associated with the Cruziana Ichnofacies (Borden Formation)



Trace fossils associated with the Cruziana Ichnofacies (Borden Formation)



Trace fossils associated with the Cruziana Ichnofacies (Borden Formation)