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The Early-Middle Mississippian Borden–Grainger–Fort Payne delta/basin complex:

Field evidence for delta sedimentation, basin starvation, mud-mound genesis, and tectonism during the Neoacadian Orogeny

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Acknowledgement

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Contents

Figures

Figures (continued)

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ABSTRACT

In latest Devonian time, the collision between Avalonia, the New York promontory and Carolina terrane under the impact of Gondwana, generated an orogeny that began in New England and migrated southward in time. Once thought to be the fourth tectophase of the Acadian orogeny, this event is now called the Neoacadian orogeny. Active deformational loading during the event initially produced the Sunbury black-shale basin, whereas subsequent relaxational phases produced the Borden-Grainger-Price-Pocono and Pennington–Mauch Chunk clastic wedges, which largely reflect the dextral transpressional docking of the Carolina terrane against the Virginia promontory and points southward. The Sunbury black-shale basin and the infilling clastic wedges are among the thickest and most extensive in the Appalachian foreland basin. This trip will demonstrate differences in basinal black-shale and deltaic infilling of the foreland basin, both in more active, proximal and in more distal, sediment-starved parts of the basin. In particular, we will examine relationships between sedimentation and tectonism in the Early-Middle Mississippian Sunbury/Borden/Grainger/

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Fort Payne delta/basin system in the western Appalachian Basin during the Neoacadian Orogeny. We will emphasize the interrelated aspects of delta sedimentation, basin starvation, and mud-mound genesis on and near the ancient Borden-Grainger delta front. Temporal constraints are provided by the underlying Devonian-Mississippian black shales and by the widespread Floyds Knob Bed/zone, a dated glauconite/phosphorite interval that occurs across the distal delta/basin complex.

INTRODUCTION

On this trip we will examine the early sedimentary manifestation of the Neoacadian Orogeny in east-central, south-central, and southeastern Kentucky from the Devonian-Mississippian transition through early Visean (Osagean) times. During most of this time in Kentucky, a large deltaic complex prograded across the state (Fig. 1). This complex is now called the Borden-Grainger-Price-Pocono clastic wedge, and it includes the largely subaerial, Price-Pocono deltaic equivalents in Pennsylvania, Maryland, Virginia, and West Virginia and the largely subaqueous, Borden-Grainger deltaic equivalents in Kentucky, Tennessee, Ohio, Indiana, and Illinois (Fig. 1). The major clastic wedges never reached south-central or west-central parts of Kentucky

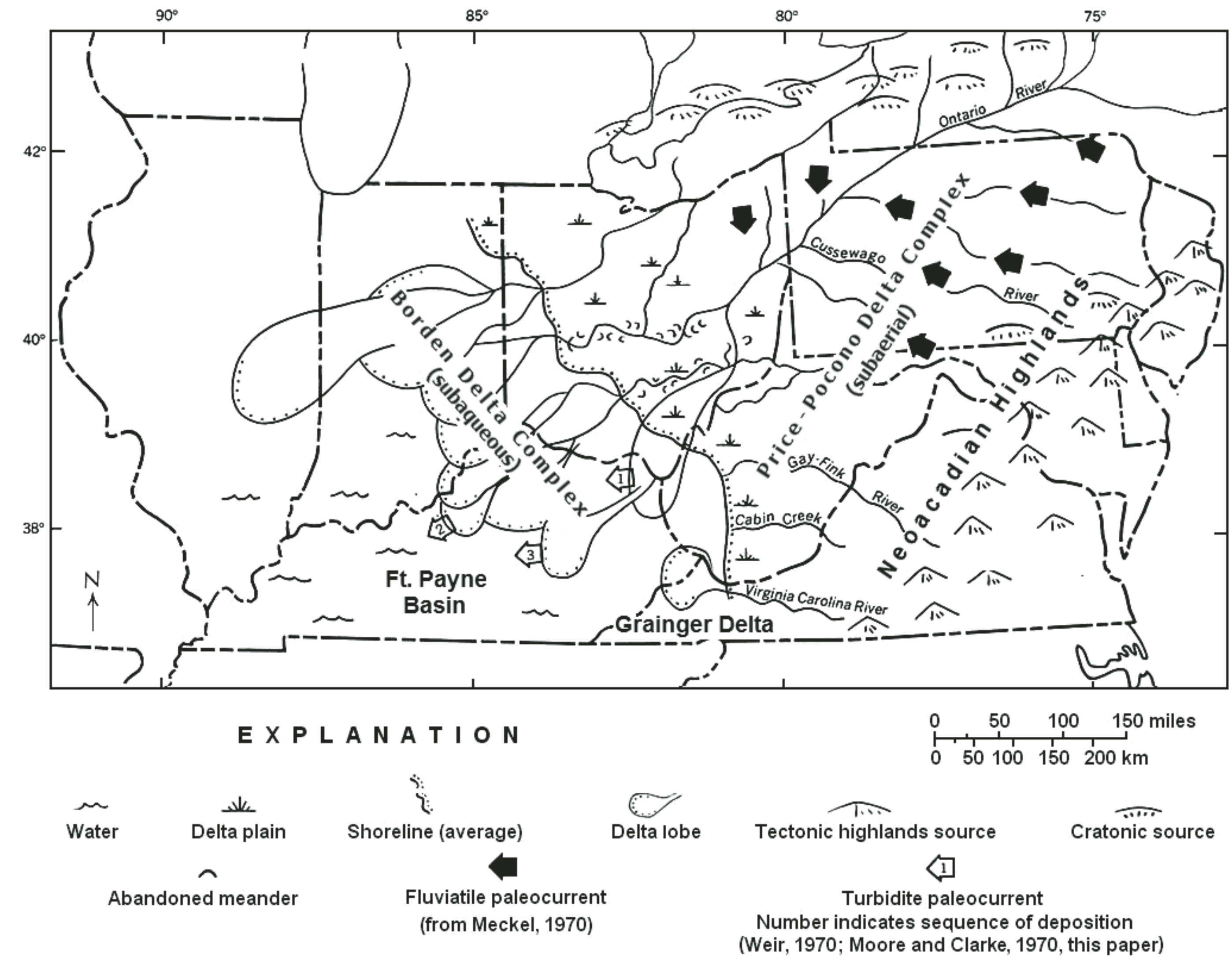


Figure 1. Paleogeographic map showing the development and progradation of the Borden-Grainger-Price-Pocono delta complex in east-central United States during Early and Middle Mississippian time as a result of the Neoacadian Orogeny. This trip will concentrate on the transition from Borden-Grainger delta lobes to the Fort Payne basin to the south and west of them (adapted from Kepferle, 1977).

and adjacent areas to the south (Fig. 1), leaving a vast underfilled basinal area on top of basinal black shales in which fine-grained clastic debris, argillaceous carbonates, carbonate bioherms and chert accumulated. These distal, basinal and slope deposits are commonly included in the Fort Payne Formation, a lithologically, stratigraphically, geographically, and temporally diverse unit. While it is known that the Fort Payne Formation is at least partially equivalent to the Borden Formation, the exact nature of the relationship was uncertain until the Floyds Knob Bed/zone, a widespread, glauconitic, marker horizon in the Borden Formation (Stockdale, 1931, 1939), was for the first time located in the Fort Payne (Udgata, 2011). This discovery makes it possible to understand both environmental and temporal relationships between the Borden, the Fort Payne, and the Neoacadian tectonism that gave rise to them all. During the course of this field trip, we will trace these relationships from the delta complex itself into the southern starved basin. The following information will establish the geologic setting in which these events occurred.

Paleogeographic and Paleoclimatic Framework

The Devonian-Mississippian transition was a time of climatic and tectonic change (Fig. 2). 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 is 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).

Tectonic Framework

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) (Fig. 3). The accretion occurred from the northeast to the southwest, perhaps reflecting the fact that the terranes were caught in a pincer movement between Gondwana and the margin of Laurussia during the closure of the Rheic Ocean (Ettensohn, 2008; Nance and Linnemann, 2008) (Fig. 4). 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, 1985a) (Fig. 3). 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, black-shale basins also migrated westwardly or cratonward, reflecting the continued cratonward movement of deformation in time. By early Famennian time (~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. 5) (Ettensohn et al., 1988a)

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. 4). 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, 1985a, 1985b, 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 1, 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 (Fig. 6A). Moreover, the likely dropstone that we will see embedded in the black shales at Stop 2 is evidence for the presence of alpine glaciation in high mountains generated by this orogeny.

The effects of deformation loading during orogeny, however, are not only restricted to the orogen and adjacent foreland basin. Flexural stresses associated with supra- and sub-crustal deformation loading may be transmitted to adjacent parts of the craton across distances up to 2000 km (Karner and Watts, 1983; Ziegler, 1978), and if that cratonic basement has been broken up

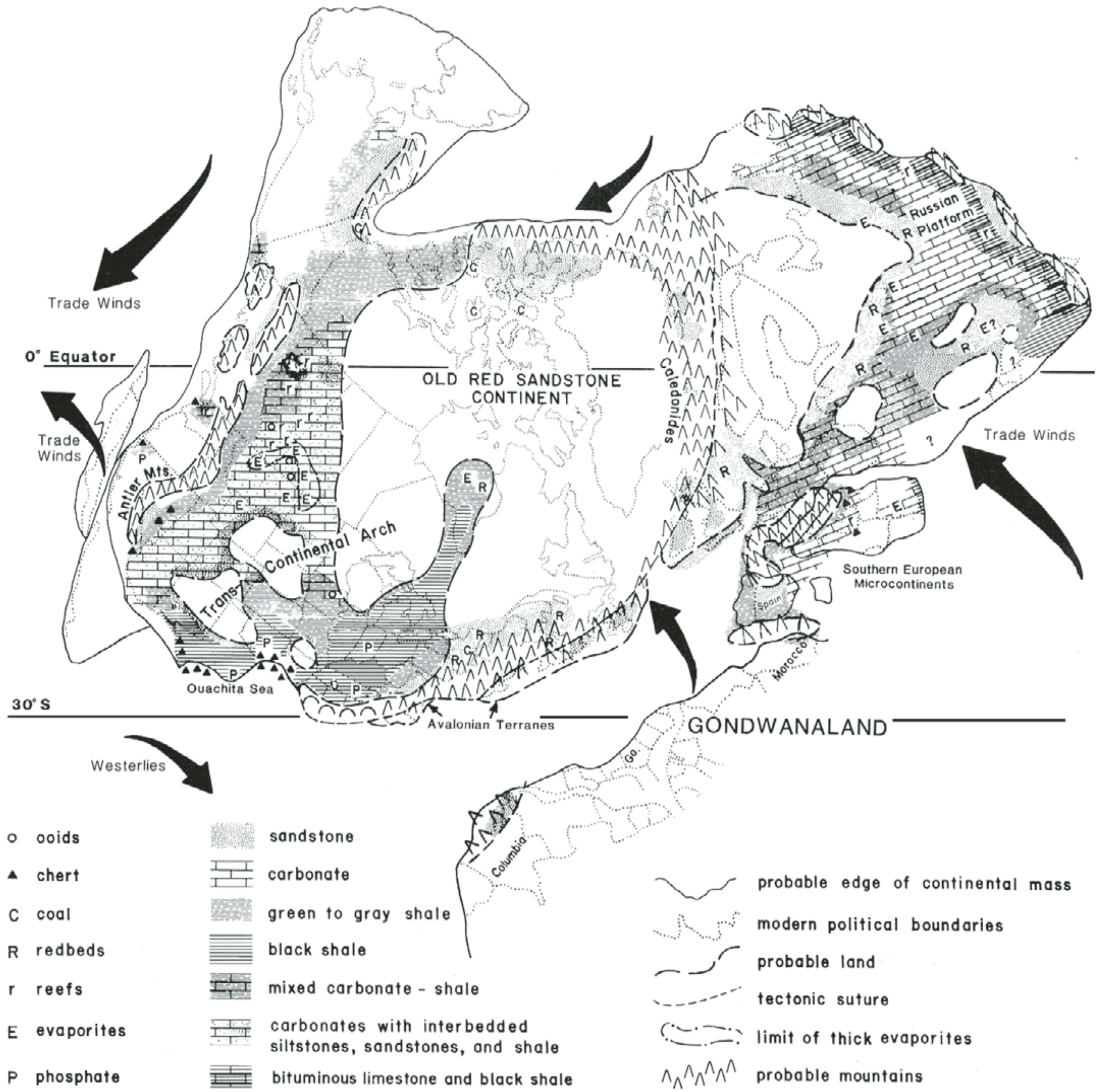


Figure 2. Late Devonian paleogeographic reconstruction of Laurussia (Old Red Sandstone Continent), showing the location of the basinal black-shale sea in eastern Laurussia into which the Borden-Grainger delta complex prograded. The field-trip area was located at ~25° S latitude in the subtropical trade-wind belt, and the Acadian/Neoacadian mountain range on the southeastern margin of Laurussia reflects the closure of the Rheic Ocean between Gondwana and Laurussia (adapted from Ettensohn and Barron, 1981).

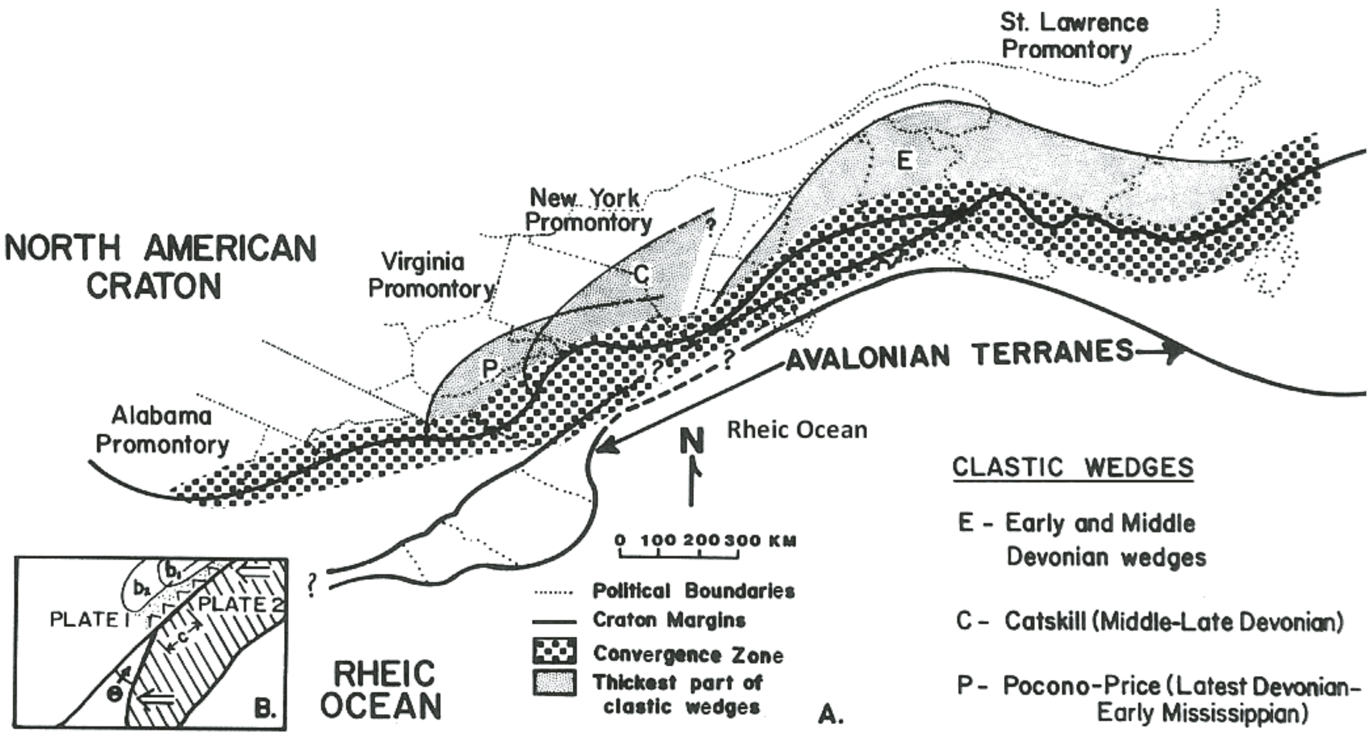


Figure 3. Schematic diagram showing tectonic framework of the Acadian/Neoacadian orogeny: (A) Diachronous oblique transpression between Avalonian terranes and Carolina and southeastern margin of Laurussia; note that clastic wedges emanate successively from each promontory, and the Pocono-Price wedge is the subaerial part of the Borden delta. (B) Geometry of oblique transpression (adapted from Ettensohn, 1987).

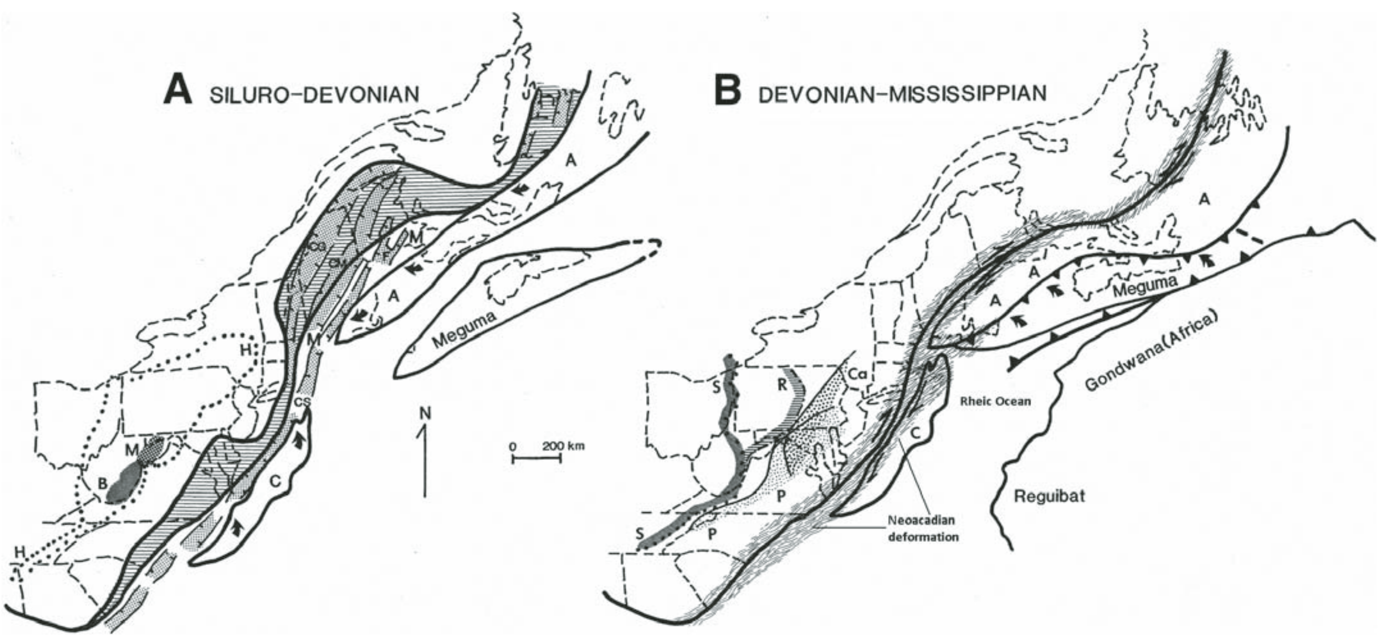


Figure 4. Neoacadian tectonic situation that generated the Sunbury Shale (Stop 1), Borden delta (Stops 1, 3, 4, 5, 6 and 7), and formed a mountain belt high enough to support alpine glaciation (Stop 2). (A) Likely pre-Acadian disposition of microplates relative to Laurussia. (B) Likely disposition of microplates and Gondwana during Neoacadian orogeny. Note the eastward limit of the deep-marine Sunbury Shale (S; see Fig. 5), its marginal-marine equivalent, the Riddlesburg Shale (R) and the Price-Pocono delta (P) (adapted from Ettensohn, 2008).

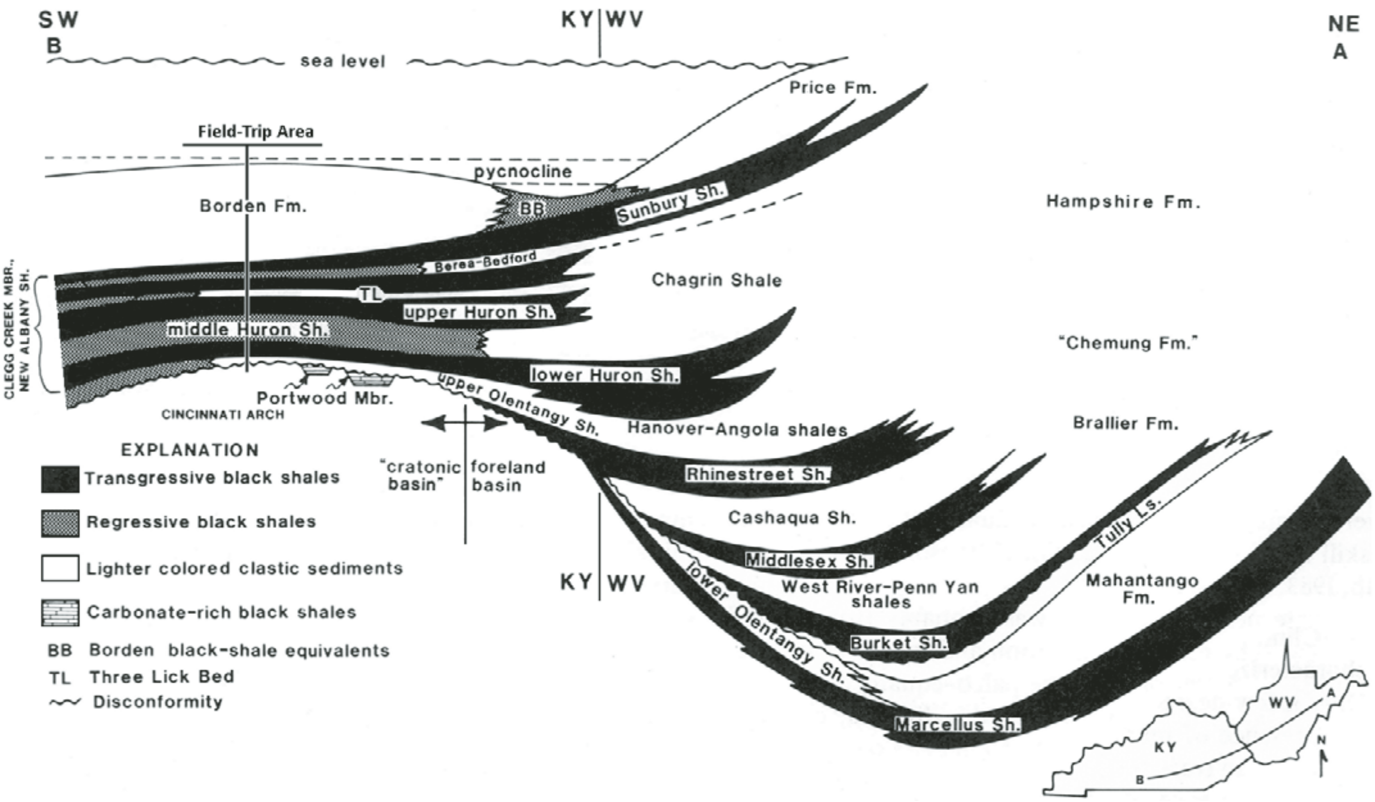


Figure 5. Schematic southwest-northeast section across the central Appalachian Basin in Kentucky (KY) and West Virginia (WV), showing the locations of cyclic Middle Devonian to Early Mississippian black-shale basins and intervening clastic wedges formed in the second to fourth Acadian/Neoacadian tectophases. Each black-shale unit represents subsidence in response to an episode of cratonward-migrating tectonism. Eastward migration of Sunbury black-shale basin, in contrast to earlier westward-migrating basins, reflects a major change in tectonic regime and beginning of Neoacadian Orogeny. Devonian-Mississippian boundary is at the base of the Sunbury Shale. The dark vertical line represents the approximate position of the field-trip area in east-central Kentucky (adapted from Ettensohn et al., 1988a).

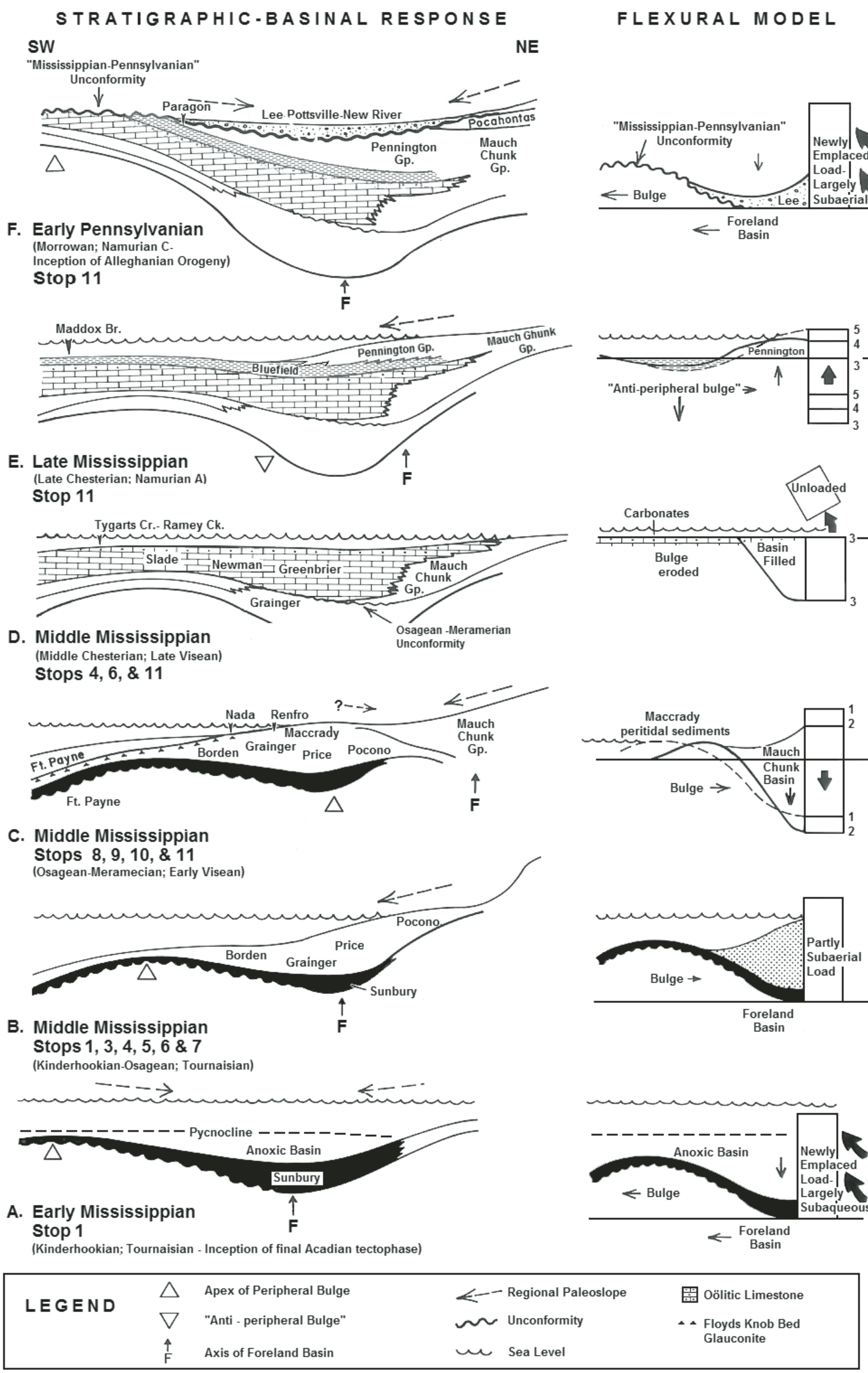
during pervious orogenic events, those structures may be reactivated during successive orogenies. This process has been called far-field tectonics (Klein, 1994), and in this way, older basement structures may control subsequent patterns of sedimentation. In the field-trip area, Iapetan (latest Precambrian–Early Cambrian), Grenvillian (Meso- to Neoproterozoic) and Mesoproterozoic Keweenawan basement structures are present (Fig. 7), and Udgata (2011) has suggested that Grenvillian and Keweenawan structures in the area may have controlled some aspects of Fort Payne sedimentation.

Stratigraphic Framework

During the course of the field trip, we will be examining Upper Devonian through Middle Mississippian rocks (Fig. 8). The Upper Devonian rocks include parts of the Ohio, New Albany, and Chattanooga black shales, as well as the Bedford Shale. These black shales unconformably overlie Middle Devonian to Upper Ordovician rocks. In the northern part of the field area, the black shales mostly overlie Middle to Lower Silurian rocks of the Bisher or Crab Orchard formations, whereas in the southern Kentucky field trip area, they unconformably overlie the Upper Ordovician Cumberland Formation. (Cattermole, 1963; Hoge and Chaplin, 1972).

The uppermost Devonian Bedford Shale is gradational with the Cleveland black shale over a distance of a few inches to a few feet, whereas the Mississippian Sunbury black shales unconformably overlie the Bedford with a subtle paraconformity (Fig. 8). In the Fort Payne basin in southern parts of the field-trip area, the Bedford and Sunbury thin to a few inches of black shales and phosphatic nodules (Falling Run Bed) at the top of the Chattanooga or New Albany shales (Campbell, 1946) (Fig. 8). Overlying the black Chattanooga or New Albany shales is a thin sequence

Figure 6. Schematic east-west cross section across the central Appalachian Basin at various times, 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).


STRATIGRAPHIC-BASINAL RESPONSE
FLEXURAL MODEL
SW
NE
"Mississippian-Pennsylvanian" Unconformity
Paragon
Lee-Pottsville-New River
Pocahontas
Pennington Gp.
Mauch Chunk Gp.
F. Early Pennsylvanian
(Morrowan; Namurian C- Inception of Alleghanian Orogeny)
Stop 11
F
"Mississippian-Pennsylvanian" Unconformity
Newly Emplaced Load- Largely Subaerial
Bulge
Lee
Foreland Basin
Maddox Br.
Pennington Gp.
Mauch Chunk Gp.
Bluefield
E. Late Mississippian
(Late Chesterian; Namurian A)
Stop 11
F
Pennington
"Anti-peripheral bulge"
5
4
3
Tygarts Cr.- Ramey Ck.
Slade
Newman
Greenbrier
Mauch Chunk Gp.
Grainger
Osagean -Meramerian Unconformity
D. Middle Mississippian
(Middle Chesterian; Late Visean)
Stops 4, 6, & 11
Unloaded
Carbonates
Basin Filled
Bulge eroded
Nada
Renfro
Maccrady
Grainger
Price
Pocono
Borden
Ft. Payne
Mauch Chunk Gp.
F
C. Middle Mississippian
Stops 8, 9, 10, & 11
(Osagean-Meramecian; Early Visean)
Maccrady peritidal sediments
Mauch Chunk Basin
Bulge
1
2
Pocono
Price
Borden
Grainger
Sunbury
F
B. Middle Mississippian
Stops 1, 3, 4, 5, 6 & 7
(Kinderhookian-Osagean; Tournaisian)
Partly Subaerial Load
Bulge
Foreland Basin
Pycnocline
Anoxic Basin
Sunbury
F
A. Early Mississippian
Stop 1
(Kinderhookian; Tournaisian - Inception of final Acadian tectophase)
Anoxic Basin
Newly Emplaced Load- Largely Subaqueous
Bulge
Foreland Basin
LEGEND
Apex of Peripheral Bulge
"Anti - peripheral Bulge"
F
Axis of Foreland Basin
Regional Paleoslope
Unconformity
Sea Level
Oölitic Limestone
Floyds Knob Bed Glauconite

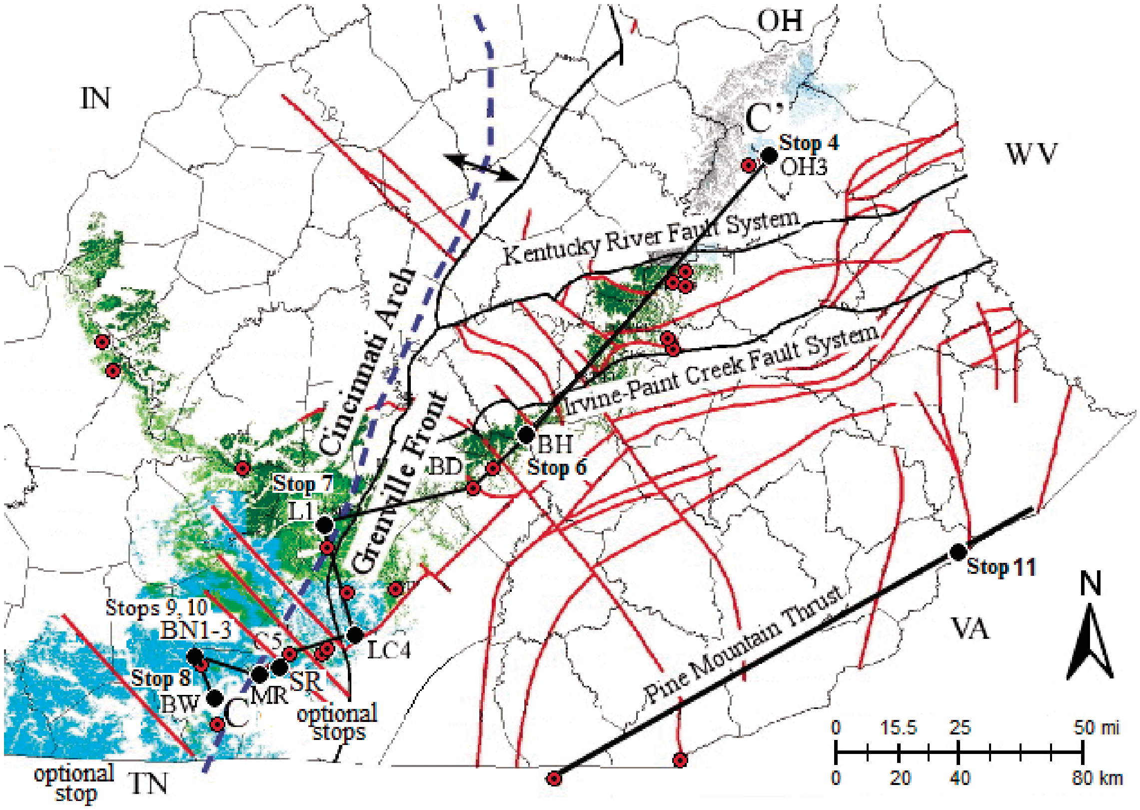


Figure 7. Basement structures in eastern and central Kentucky that may have influenced the deposition of the Borden-Grainger-Fort Payne delta-basin complex, and the location of northeast-southwest section line C–C′. The approximately east-west structures in eastern Kentucky, associated with the Rome Trough, are latest Precambrian–Early Cambrian Iapetan structures. The NNE-SSW Grenville Front is the western boundary of metamorphism and deformation associated with the Grenville Orogeny, whereas NW-SE structures in south-central Kentucky are Mesoproterozoic Keweenawan structures associated with the East Continental Rift Basin (adapted from Udgata, 2011). Faults associated with the East Continental Rift Basin in north-central Kentucky are compiled from Drahovzal and Noger (1995). All other inferred faults are from Drahovzal et al. (1992) and Drahovzal and Noger (1995). IN—Indiana; OH—Ohio; WV—West Virginia; TN—Tennessee; VA—Virginia. The lettered dots along the section line represent locality identifications that can be found in Udgata (2011).

of largely Lower Mississippian gray to greenish-gray shale that has been called the Henley Bed, Maury Shale or New Providence Shale in part (Stockdale, 1939; Ettensohn et al., 1988a; Sable and Dever, 1990) (Fig. 8). This sequence may be a few inches to a few feet thick and is commonly glauconitic; it represents distalmost parts of the Borden deltaic sequence. On top of it, the Borden deltaic sequence (Farmers, Nancy, and Cowbell members; Fig. 8) prograded westward into the largely underfilled Sunbury black-shale basin; hence, in northeastern and north-central Kentucky (Fig. 1), the coarsening-upward Borden Formation dominates (Sable and Dever, 1990). To the south and southwest, where thick Borden clastics are absent, the fine-grained clastics of the Fort Payne Formation predominate (Fig. 8). The final member of the Borden Formation in the northeast is the Nada Member (Weir et al., 1966), which is largely equivalent to the Floyds Knob Bed/zone throughout east-central United States and marks the cessation of major deltaic sedimentation across the area (Udgata, 2011). Below the Floyds Knob Bed/zone in the Fort Payne Formation, the silty shales are largely equivalent to the Nancy and New Providence members elsewhere (Sable and Dever, 1990; Udgata, 2011). Above the Floyds Knob, the Fort Payne becomes more carbonate rich and is partially equivalent to the siltstones and cherty dolomites of the Muldraugh Member of the Borden Formation. The Muldraugh and lateral equivalents in the Fort Payne Formation are widely interpreted to have filled the basin southwest of the abandoned Borden delta lobes (e.g., Sable and Dever, 1990). The Muldraugh and Fort Payne grade upward into the shallower water carbonates of the Warsaw-Salem Formation and the Renfro Member of the Slade Formation (e.g., Sable and Dever, 1990). These units, in turn, give way upward and laterally into the very shallow-water carbonates of the Slade and Newman formations (Fig. 8). In far-eastern Kentucky, the Grainger Formation is a distal equivalent of the Borden delta (e.g., Sable and Dever, 1990) but reflects a different part of the Neoacadian

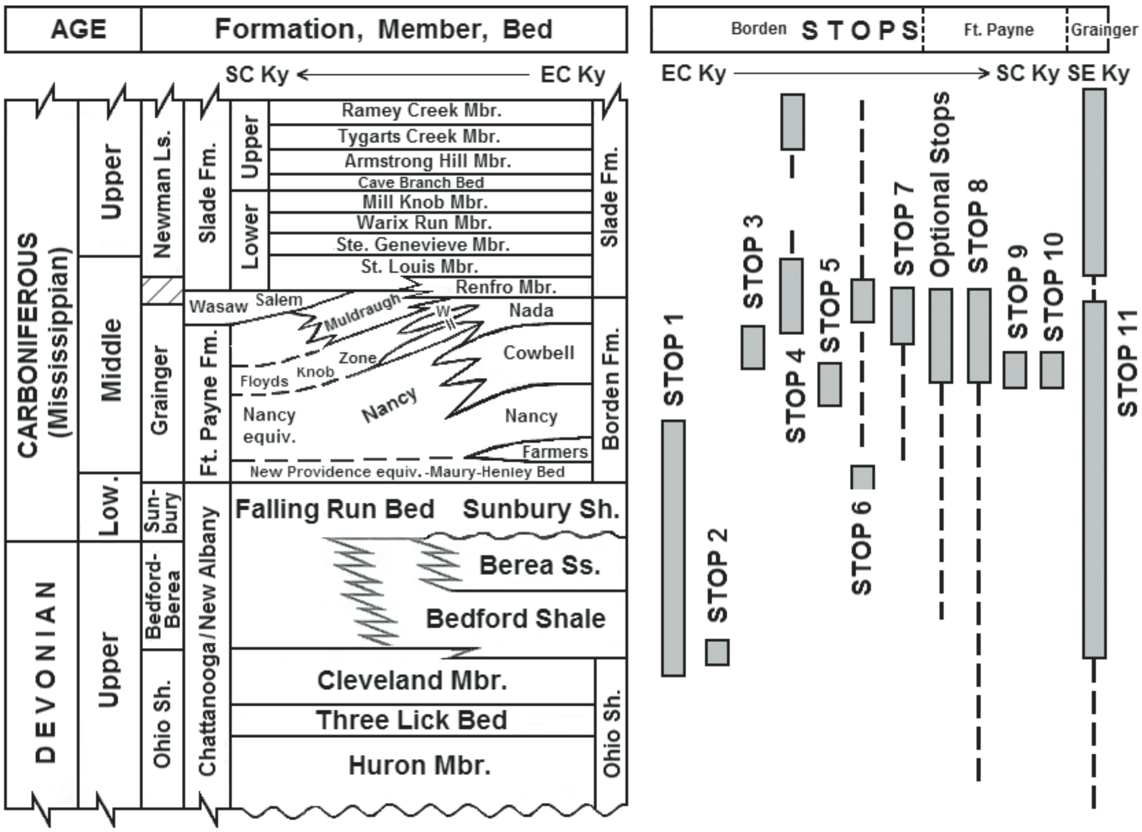


Figure 8. The geologic section along the field-trip route, showing the relative stratigraphic positions of the sections viewed at each stop. The Wildie (W) and Halls Gap (H) members are local storm-related siltstone bodies equivalent to parts of the Nada Member of the Borden Formation. Unit thicknesses are not to scale.

deltaic complex (Fig. 1). Uppermost parts of the Grainger are red in color and contain exposure features; they have been equated to the Maccrady Formation farther to the east in West Virginia and Virginia (Wilpolt and Marden, 1949, 1959; Howell and Mason, 1998; Ettensohn, 1998). A greenish, glauconitic bed at the very top of the Grainger is probably equivalent to the Floyds Knob Bed (Udgata, 2011).

Borden-Grainger Deltaic Complex

An important part of the sedimentary record produced by the Neoacadian 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" (Ettensohn, 2004) (Figs. 5, 9); closely associated with these deltaic clastics are apparently deeper water units like the Fort Payne Formation (Ettensohn, 2004). Borden deltaic sediments prograded into Kentucky from the northeast, whereas Grainger sediments reflect another more southerly deltaic complex that prograded into extreme eastern Kentucky (Fig. 1). We will examine parts of the Borden Formation at Stops 1, 3, 4, 5, 6, and 7, and the Grainger Formation at Stop 11. 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. 5). The Borden deltaic complex, however, is unusual in that it prograded beyond the Appalachian Basin, across the Cincinnati Arch and into the Illinois Basin (Figs. 1, 9). The Borden Formation represents subaqueous parts of a westwardly prograding delta complex (Swan et al., 1965; Lineback, 1966; Weir et al., 1966; Kepferle, 1971, 1977, 1979) (Fig. 1) that formed as an early relaxational response to the Neoacadian orogeny (Ettensohn et al., 2002, 2004; Ettensohn, 1994, 2004, 2005, 2008) (Fig. 6). 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) (Fig. 6C). This period of sediment starvation is represented by the Floyds Knob Bed/zone (Fig. 6C), a thin, widespread, late Osagean interval of glaucony and phosphorite deposition (Peterson and Kepferle, 1970; Whitehead, 1978; Sable and Dever, 1990), which will be examined at Stops 4, 6, 7, 8, 9, 10, and 11. The Floyds Knob Bed/zone commonly occurs within the uppermost unit of the Borden Formation, the Nada Member, which is composed of fossiliferous, blue-green

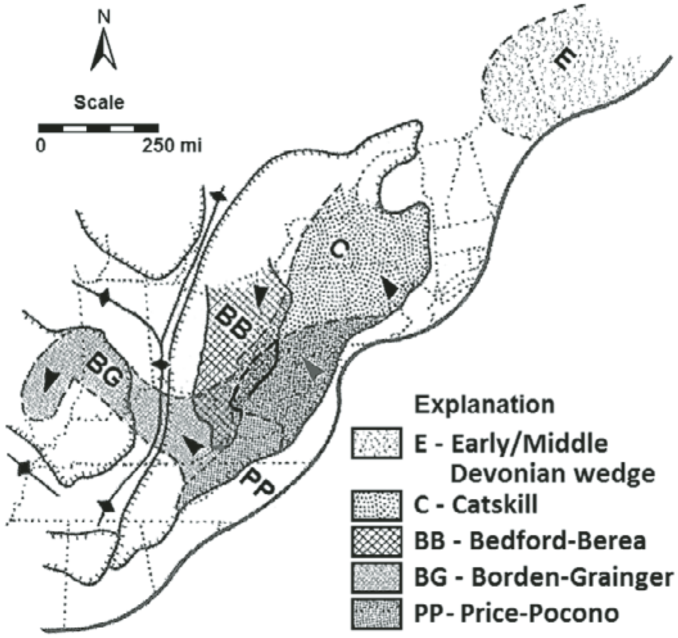


Figure 9. Relative distribution of Acadian/Neoacadian clastic-wedge, delta complexes on southern Laurussia. The delta complexes migrated south in time tracking the progress of Acadian/Neoacadian transpression. The Borden-Grainger (BG) complex is the subaqueous equivalent of the Price-Pocono (PP) subaerial 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. The clear area south of the Borden-Grainger (BG) complex was effectively filled with Fort Payne "starved-basin" sediments (after Ettensohn, 2004).

shales with interbedded carbonates and siltstones; it has been interpreted to represent delta destruction (e.g., Ettensohn, 2004) and will be viewed at Stop 4.

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. 6D). 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-water Slade carbonates overlie the Nada Member of the Borden Formation at Stop 4; more information on the Slade carbonates can be found in Ettensohn and Dever (1979), Ettensohn (1980, 1981, 1992a), and Ettensohn et al. (1984, 2004).

Floyds Knob Bed/Zone

The Floyds Knob Bed was named by Stockdale (1931, 1939) for a thin horizon of glauconitic limestone that interrupted the Borden Group in southern Indiana. However, he was able to trace the unit as a glauconite bed, a few inches to a few feet thick, all around the Kentucky outcrop belt, and it has been used as a prominent marker horizon since (Peterson and Kepferle, 1970; Whitehead, 1978). In the area of the Borden Formation, the Floyds Knob was an easily traceable bed, but the Floyds Knob had never been reported from the Fort Payne Formation, and it was assumed that the bed followed the Borden delta front and merged at the toe of the delta with the greenish-gray, glauconitic Maury Shale (Lewis and Potter, 1978; Ausich and Meyer, 1990; Meyer et al., 1992, 1995; Krause et al., 2002; Greb et al., 2008). However, recent work by Udgata (2011) found the Floyds Knob Bed at several localities in the Fort Payne Formation, and he showed that the bed does not merge with the Maury Shale or its equivalents, but overlies it by a meter or more (Fig. 10). Although the Borden delta complex is a time-transgressive sequence (Gordon and Mason, 1985) the Maury and its equivalents are mostly Early Mississippian, Kinderhookian units (e.g., Sable and Dever, 1990), whereas existing biostratigraphy (Ausich et al., 1979; Lee et al., 2005) and regional stratigraphy (Shaver, 1985; Patchen et al., 1985) and Udgata's (2011) radiometric dating of Floyds Knob glauconite showed that it was Middle Mississippian or late Osagean in age. In fact, Udgata's (2011) recent dating of the Floyds Knob Bed/zone corroborates the largely Middle Mississippian age for the main part of the Fort Payne, which up to this point has been mostly inferred through stratigraphic and biostratigraphic research. It is now apparent that parts of the Fort Payne below the Floyds Knob Bed are Nancy and New Providence (Maury) equivalents (Fig. 8) and reflect the period of major Borden delta progradation. Carbonate-rich parts of the Fort Payne above the Floyds Knob reflect the cessation of major deltaic sedimentation and are partial equivalents of the Muldraugh, upper Nada and lower Renfro members (Fig. 8).

The Floyds Knob Bed itself was often thought to be a single bed of glauconite, but Udgata's (2011) work showed that the Floyds Knob is actually a zone of one to six distinct glauconite beds separated by greenish-gray, glauconitic shales (Fig. 10). Although in parts of the Borden and Fort Payne formations, a single Floyds Knob Bed occurs (as at Stop 8, Fig. 10), four of the six beds seem to be widespread in the Borden Formation. In the central parts of the Fort Payne basin, the Floyds Knob Bed is represented by green glauconitic shales, which underlie or encase carbonate buildups or mud mounds (Fig. 10). The buildups developed on uplifted, sediment-starved surfaces that were eventually enveloped by green glauconitic muds, formed by the reworking of pelletal glauconites further up on the delta front (Udgata, 2011). Although green shales have long been associated with the carbonate mud mounds in the Fort Payne (Marcher, 1962; Lewis and Potter, 1978), previous investigations had not equated the Fort Payne green shales with the Floyds Knob Bed on the Borden delta front. The clay mineral glauconite is associated with very slow sedimentation rates in oxygenated to mildly reducing marine settings at depths of 30–500 m (e.g., Odin and Matter, 1981; Kelly and Webb, 1999), conditions similar to those

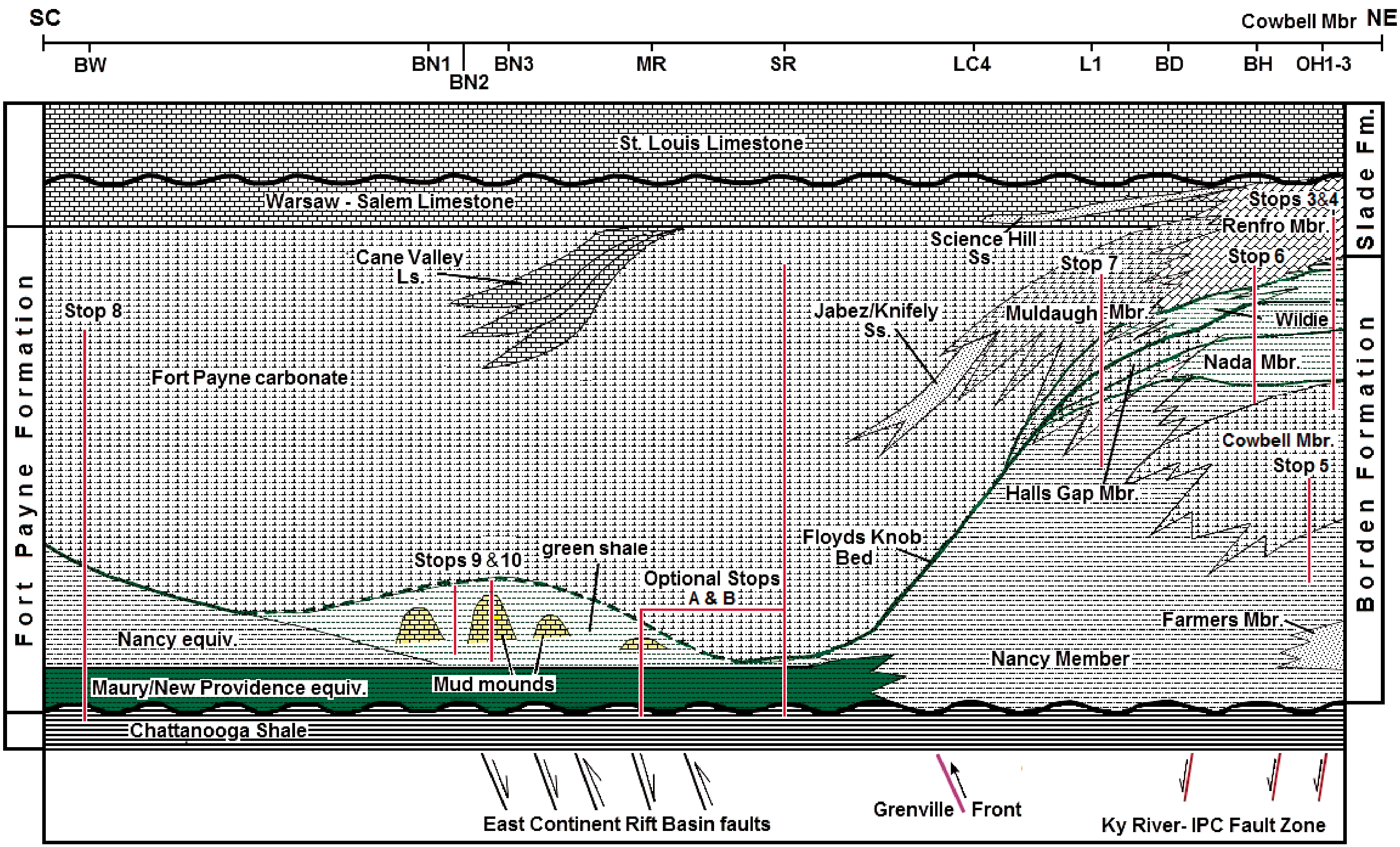


Figure 10. Schematic cross section along line C–C′ in Figure 7 from the Borden delta front in northeastern Kentucky to the Fort Payne basin in south-central Kentucky and then back onto another lobe of the Borden delta, showing the distribution of members and lithofacies relative to the Floyds Knob Bed/zone and field trip stops. In more proximal settings in northeastern Kentucky and around the basin mud mounds, the Floyds Knob is a thick zone; elsewhere, it is represented by a condensed bed. Vertical red lines represent stops on this field trip; symbols as in Figure 11 (adapted from Udgata, 2011).

expected on the sediment-starved Borden delta front in late Osagean time.

The Floyds Knob Bed is essentially the same as the Nada Member of the Borden formation, which has been interpreted to represent a time of delta destruction and sediment starvation. According to Ettensohn (1994, 2008) and Ettensohn et al. (2002, 2004), glauconite-rich units like the Nada and Floyds Knob reflect relative sediment starvation in response to eastward bulge migration during a time of sea-level lowstand (Fig. 6C). The presence of the bulge effectively blocked further westward deltaic sedimentation into the area and diverted clastic sedimentation to the more proximal Mauch Chunk basin. On the bulge itself, the red, peritidal sediments of the Maccrady Formation developed, and we will see equivalents of this at Stop 11 in the Grainger Formation associated with the Floyds Knob Bed. More distally to the west, the bulge blocked major deltaic sedimentation, allowing a period of sediment starvation to ensue across large parts of the east-central United States (Fig. 6C), and it was during this time that the glauconitic Floyds Knob Bed/zone developed. Even though major deltaic sedimentation was halted throughout the east-central United States, some clastic sedimentation still occurred. In the Nada interval, for example, the included Wildie and Halls Gap siltstones reflect brief, localized periods of delta-lobe progradation or storm reworking during glauconite sedimentation (Figs. 8, 10). Hence, the nature of the Floyds Knob varies with proximity to the eastern source areas (Fig. 11). Figure 11 is a facies map of the Floyds Knob zone in Kentucky and adjacent states.

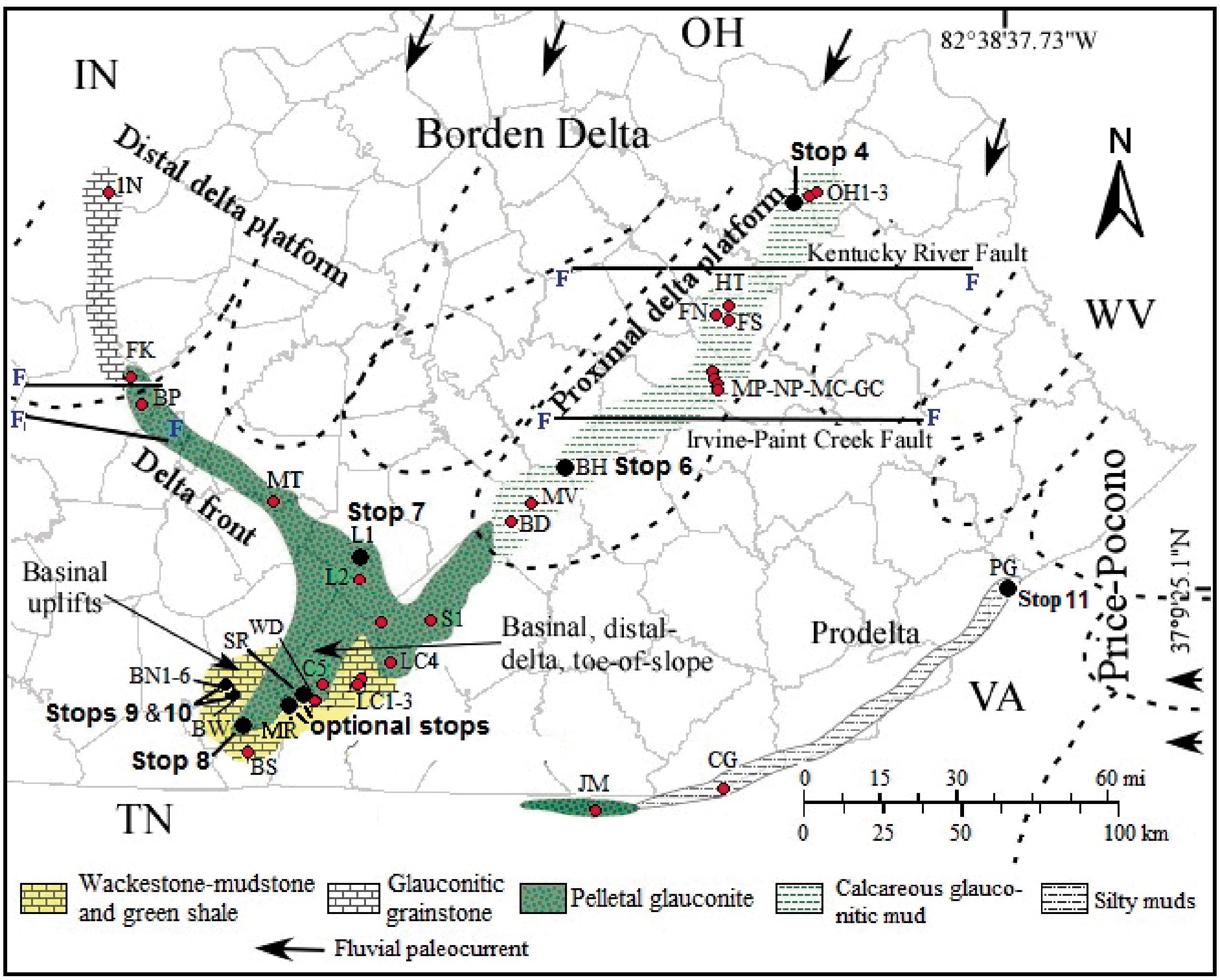


Figure 11. Facies map of the Floyds Knob Bed/zone and equivalent strata in Kentucky and adjacent states, relative to the outcrop belt, likely position of delta lobes (after Kepferle, 1977), and field trip stops. The lines labeled "F" reflect fault zones that cross the outcrop belt (adapted from Udgata, 2011); these faults are shown in their approximate positions and are shown as exaggerated lines to highlight their position relative to the field stops. Actual positions of faults in the Kentucky River Fault System and Irvine–Paint Creek Fault System are shown in Figure 7. Black dots represent stops used in field trip; lighter colored red dots represent Udgata's (2011) study sections.

THE STOPS

During the course of the trip, we will examine Devonian and Mississippian rocks at 11 stops, and two possible optional stops, throughout parts of eastern and central Kentucky. The stratigraphic disposition of the stops is shown in Figure 8, whereas the approximate geographic location of the stops is shown in Figure 12. More specific locations for each stop are provided at stop descriptions below.

Stop 1: Upper Part of the Black-Shale Sequence and Lower Borden Formation

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

Stop 1 is located in a road cut along the northbound lane of State Route 801, 1.75 miles south of exit 133 from Interstate Highway I-64 (Fig. 12). The exposure is 0.35 miles north of the junction of State Routes 801 and 1722 in the south-central part of the Farmers 7.5 min quadrangle in Rowan County, Kentucky. Latitude: 38° 09′ 08″ N; longitude: 83° 33′ 6.5″ W.

At Stop 1 we will examine the basinal black-shale sequence and lower parts of the subaqueous Borden delta complex that prograded onto it. The basinal sequence includes the well-exposed upper part of the Ohio Shale (Cleveland Shale Member), the Bedford Shale, and the Sunbury Shale (Fig. 8), whereas lower parts of the Borden Formation include the Henley Bed, Farmers Member, and Nancy Member (Figs. 8, 13).

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 14.

Unit 1*, *Cleveland Shale Member (Ohio Shale). Unit 1 consists of 4.25 m (14 ft) of the Cleveland Shale Member of the Ohio Shale Formation (Fig. 14). This shale is a brownish-black to black, fissile, organic-rich, silty shale. It is pyritic with occasional siderite or phosphate nodules scattered throughout. When weathered, it tends to take on a dusky yellowish brown to yellowish-orange color. Macroscopic body fossils and trace fossils are rare within the unit. Under the microscope, occasional lingulid brachiopods, conodonts, and fish remains (teeth, scales, and bones) can be found (Barron and Ettensohn, 1981). In viewing weathered exposures of the shale from the side, one can also see that the weathered surface of the outcrop has a ribbed appearance (Fig. 15). This ribbing consists of smoother faced promontories and more splintery-weathering recessed intervals which average ~5 cm thick. The ribbing is caused by differential weathering. Recessed, more highly weathered shale intervals are carbonaceous shales with slightly lower organic content than the smoother faced promontories. The lower organic content apparently allows these intervals to weather more rapidly. The organic content is largely a function of dilution by silt and other sediments, so that differential weathering might be caused by differing organic content, silt content, or differing degrees of cementation.

The environment of deposition has been regarded as a deep-water, anoxic or anaerobic, basin-floor setting (e.g., Ettensohn and Barron, 1981; Ettensohn et al., 1988a), although evidence also supports the presence of suboxic conditions from time to time (Perkins et al., 2008). 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 have been found tend to represent organisms that maintain a nektic or planktic life style. This life style 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 as discussed in Barron and Ettensohn (1981). A final piece of evidence within these organic-rich shales is the presence of pyrite (FeS_2), along with sparse nodules of siderite ($FeCO_3$) and phosphorite. All of these mineral phases tend to form under Eh conditions that are reducing or (–) negative (Garrels and Christ, 1965, p.224).

These anaerobic conditions most likely formed in a marine basin in which the vertical water column was commonly density stratified (e.g., Ettensohn and Barron, 1981). 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. 16). 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 (Algeo et al., 1995; Algeo and Scheckler, 1998; Rimmer et al., 2004). 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).

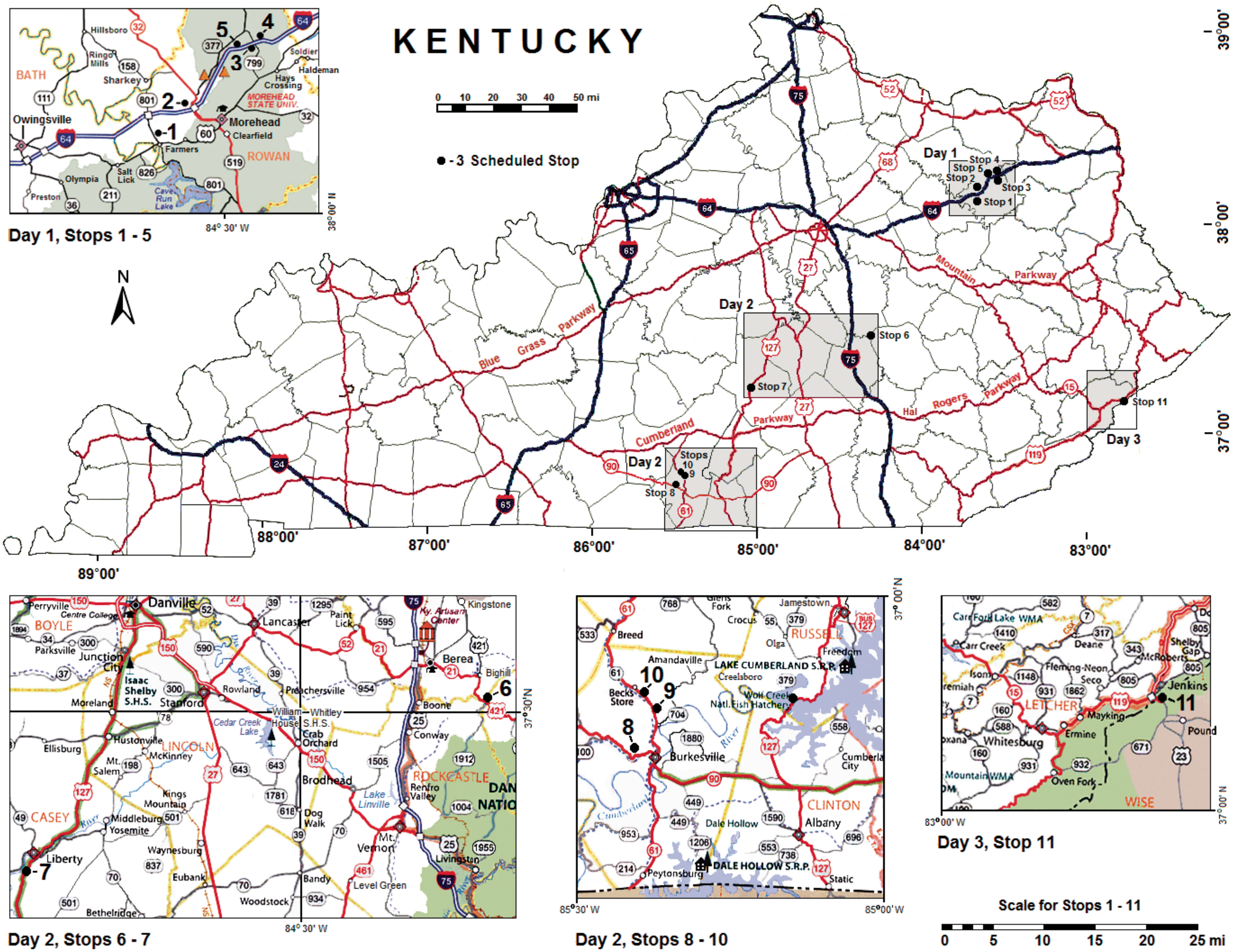


Figure 12. Field-trip route and stops in northeastern, east-central, south-central, and southeastern Kentucky.

Figure 13. Photo showing the upper part of the stratigraphic section at Stop 1 (see Figures 8, 14). 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.

As stated previously, the ribbed character of these shales (Fig. 15) may be related to variations in the organic or silt content of the shale (e.g., Ettensohn et al., 1988a). 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 percent 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 percent of terrigenous mud would translate to an increase in the relative amount of organic matter in the shale. A third possibility might relate to cyclic variations in the amount of detrital organic matter derived from terrestrial plants living on the land surface. Streams draining recently colonized Devonian lands would have carried an increased amount of organic matter, primarily in the form of plant detritus. This organic detritus would have eventually been transported into deeper parts of the basin and become incorporated into the fine-grained sediments of the Ohio Shale. Conceivably, variations in the amount of terrestrial plant matter reaching the basin bottom could have caused differences in the amount of organic matter preserved in the shales.

Devonian, marine, black-shale deposits are quite notable for their widespread occurrence across the inland seas of North America and Eurasia (Fig. 2). Algeo et al. (1995) and Algeo and Scheckler (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^-). Hydrolysis can effectively remove CO_2 from the atmosphere and ultimately tie it up in the carbonate-silicate geochemical cycle. These weathered bicarbonates enter

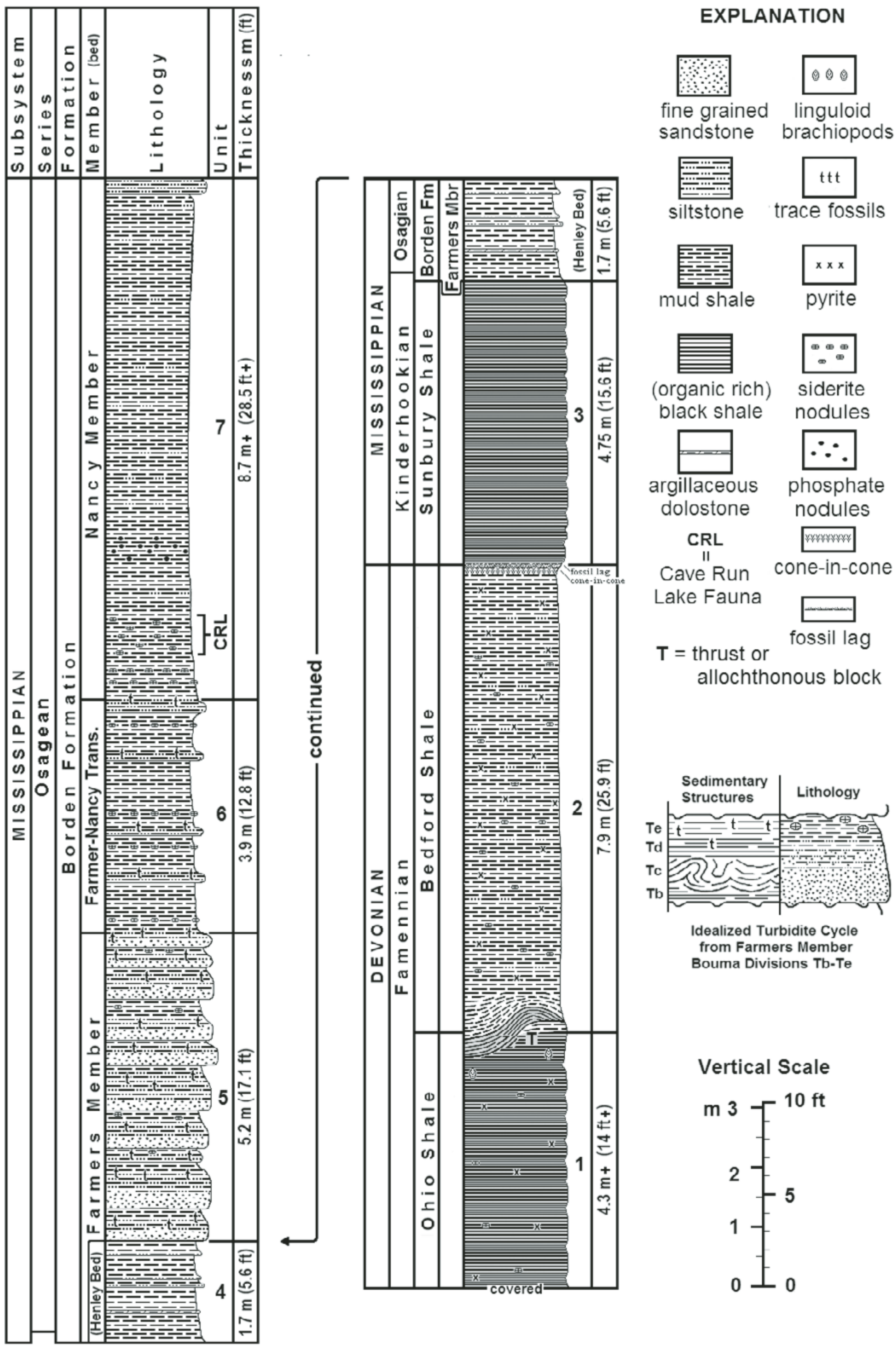


Figure 14. Schematic stratigraphic section for the outcrop at Stop 1 (see Fig. 13).

Figure 15. "Ribbed" black shales from Cleveland Member of the Ohio Shale at Stop 1. Note the eroded profile of these shales is due to differential weathering.

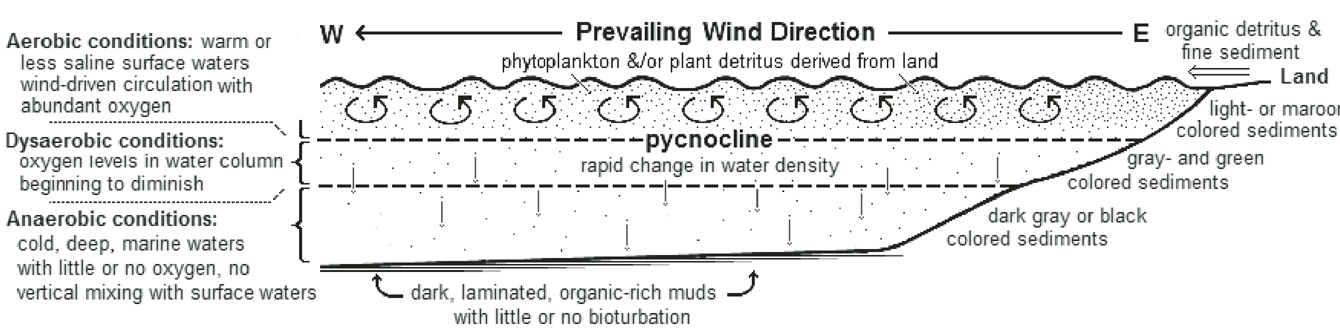


Figure 16. Cartoon illustrating possible means (stratified water column) for the isolation and preservation of organic matter in the Devonian-Mississippian black shales.

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) and Algeo and Scheckler (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. Famennian glacial deposits are recorded 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 m (25.9 ft) of medium- to olive-gray mud shale that is poorly fissile and non-calcareous (Fig. 14). 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; these fossils are most commonly found in pyrite nodules. The lower contact of the Bedford with the underlying Ohio Shale (Cleveland Member 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, ~5.5 cm (2.1 in) thick, is locally present (Fig. 14). 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. Our petrographic examination of the cones reveals that they are formed of fibrous calcite, with the fibers lying 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 solution.

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 (Fig. 14). 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 N 74° E and dip at an angle of ~5° to the north. We are fairly confident that these are small thrust faults because we can see the displacement of the black shale as it was thrust up and over the gray Bedford Shale, which is, in turn, thrust over the top of the black shale. Slickensides associated with surfaces in the black shale, local intrusion of plastic gray shales into the black, as well as several unusual joints that extend out from the thrust at angles of 47° to 68° from the thrusts provide additional evidence for the interpretation. These structures may reflect a subsurface response to growth faulting along nearby basement structures at depth (e.g., Drahovzal and Noger, 1995) or they may reflect mass movement from another dimension. In either case, the gray shale tends to behave more plastically, whereas the black shales tend to act more competently.

Deposition of the Bedford probably occurred in a dysaerobic setting (Fig. 17). Dysaerobic environmental conditions exist where dissolved oxygen levels in the water or sediments are between 0.1 and 1.0 ml 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 a key, as the Fe is in a +2 oxidation state (ferrous iron). In this situation, there was apparently enough oxygen in the water column to thoroughly oxidize any organic matter but not enough to precipitate iron oxides, such as hematite (iron is in a +3 oxidation state), which would have imparted a red or maroon color to the shales. Second, a lack of an abundance of free oxygen in these sediments is also indicated by the presence of pyrite and siderite nodules in these shales, both of which are stable under reducing conditions (Garrels and Christ, 1965). Third, only a small, depauperate fauna is present, one that apparently includes only thin-shelled brachiopods and mollusks, and such faunas are typically common in dysaerobic fossil assemblages (Kammer, 1985; Kammer et al., 1986; Pashin and Ettensohn, 1992). Even so, abundant bioturbation is present, 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 (1992, 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. 17). However, recent ideas about the possibility of Late Devonian, Acadian/Neoacadian, alpine glaciation, which we will discuss at Stop 2, suggest that Bedford shales from eastern sources may reflect distal low-stand deposits related to alpine

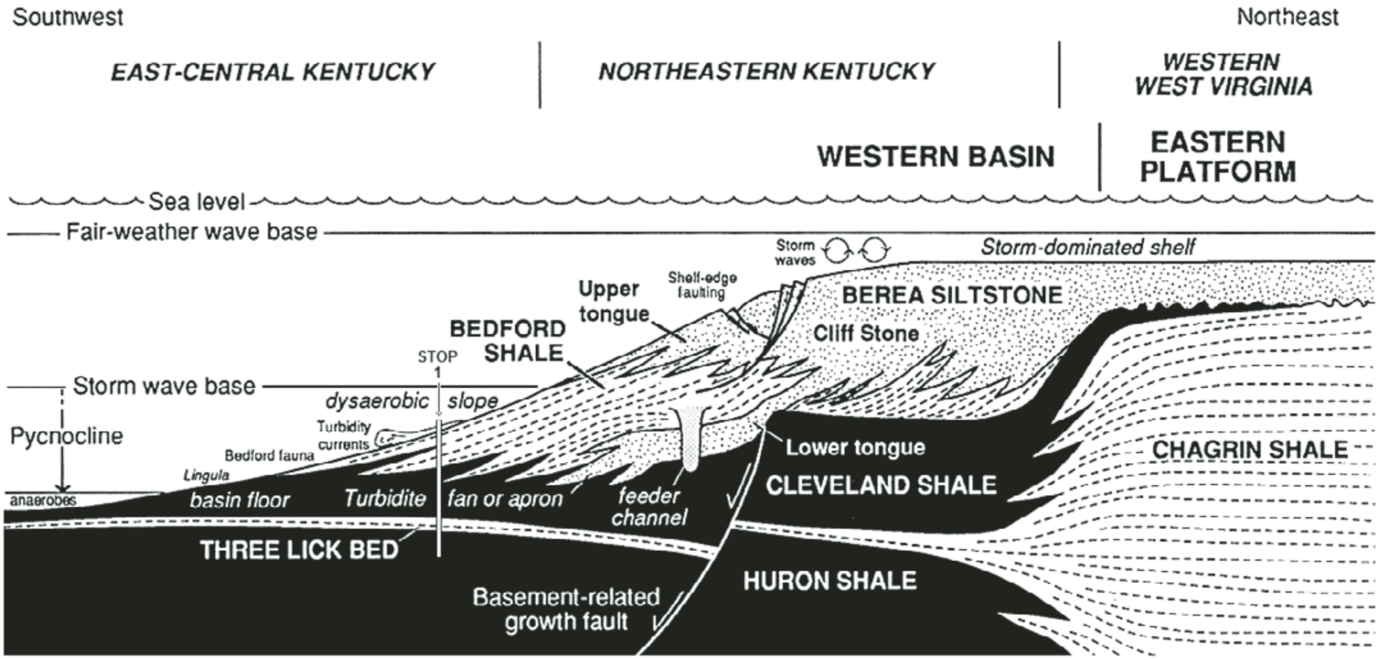


Figure 17. Depositional model for the Bedford-Berea sequence in eastern Kentucky and West Virginia (after Pashin and Ettensohn, 1987).

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 (1979b), Ettensohn and Elam (1985) and Pashin and Ettensohn (1987, 1992, 1995).

Unit 3 (Sunbury Shale). The Sunbury Shale at this locality consists of 4.75 m (15.6 ft) of dark-gray to black, highly carbonaceous, fissile shale (Fig. 14). The unit contains small pyritic nodules as well as small, scattered pyrite crystals. Overall, the lithology of the Sunbury Shale is very similar to that of the Ohio Shale below, and it exhibits similar differential weathering with a ribbed appearance when weathered surfaces are 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 lingulid or orbiculoid brachiopods, as well as conodonts.

The lower contact of the Sunbury with the underlying Bedford Shale is sharp and characterized by a 1.5 cm (0.6 in) thick "lag" deposit (Fig. 14). This lag lies immediately above the cone-in-cone layer previously mentioned in the Bedford Shale (Fig. 14) and contains a variety of both pyritized and phosphatic fossil remains including inarticulate lingulid 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. This basal lag horizon has been recognized throughout the entire outcrop area of the Sunbury Shale (Pepper et al., 1954), and in this area 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 during a time of deformational loading (Fig. 5), the Sunbury represents a transgression and subsidence event (Fig. 6A) that moved eastward in time (Fig. 5), apparently reflecting inception of a new, more proximal Neoacadian convergence event to the east (Fig. 4B). Additional information and interpretations regarding the Sunbury can be found in papers by Chaplin and Mason (1979b, 1985), Ettensohn and Elam (1985), Mason and Lierman (1985), Ettensohn et al. (1988a), and Lierman et al. (1992a).

Unit 4, Henley Bed, Farmers Member (Borden Formation). The Henley Bed is the basal-most unit of the Borden Formation (Figs. 8, 13, 14). Here it consists of 1.7 m (5.6 ft) of greenish-gray to grayish-green mud shale that is poorly fissile and non-calcareous (Fig. 14). 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 (~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 (~50 cm above the base), no conodonts have been recovered at this locality. However, at other sites across the Kentucky and Ohio, the upper *Siphonodella crenulata* Zone has been identified. At ~30 cm below the first siltstone bed at this locality, the conodont species *Polygnathus communis carinus* and *Pseudopolygnathus multistriatus* have been found. Below this interval and extending down to the dolomite bed previously mentioned, another interval lacking conodonts occurs. *P. communis carinus* and *P. multistriatus* have also been recorded from a 1 m interval at the base of the Nancy Member at this location, and these two forms indicate an early Osagean age (equivalent to the Fern Glen or early Burlington formations) for this part of the interval (Work and Mason, 2005). For this reason the contact between the Kinderhookian and Osagean is placed at the thin dolostone layer ~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. 5). 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 (1979b), 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. 13, 14). At this locality, the Farmers consists of 5.2 m (17.1 feet) of interbedded sandstones/siltstones and shales (Fig. 14). 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 a light brownish-gray to yellowish-brown and are composed principally of quartz, rock fragments and mica. The matrix is chiefly clay, siderite, and micro-crystalline quartz. Individual beds tend to be normally graded with the lower portion of beds consisting of very fine-grained sand. This 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 Member 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 tool marks, such as groove, brush, prod, and bounce casts. The most abundant tool marks are groove casts. Measurements of paleocurrent directions from the sole marks in the Farmers show a general trend from east to west (Moore and Clarke, 1970). This paleocurrent direction is consistent with the westward downslope movement of material off the Borden delta front (Fig. 1) (Moore and Clarke, 1970; Weir, 1970; Kepferle, 1978). The upper surfaces of the coarse-grained beds tend to grade into the overlying shales and show evidence of extensive 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. 14). This sequence corresponds to a truncated Tb-Te interval of Bouma's classic turbidite sequence (Bouma, 1962). The graded interval Ta at the base of a Bouma's sequence (Fig. 18) is not evident here due to the overall fine-grained nature of the rocks.

Body fossils can be found throughout the Farmers but tend to be locally restricted to certain beds or siderite nodules. Megafossils are mainly found as molds, and include productid and spiriferid brachiopods, fenestrate bryozoans, crinoid columnals, gastropods, cephalopods, conularids, trilobites, hexactinellid sponges, and bivalves. This association of fossils is characteristic of 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: *Zoophycos*, *Lophoctenium*, *Sclarituba*, *Teichichnus*, *Palaeodictyon*, and *Chondrites*. Escape burrows, or fugichnia, are also quite common, and these traces typically extend from the base to the top of many Farmers beds. These trace fossils are representatives of both the *Zoophycus* and *Nereites* ichnofacies and mainly reflect grazing and feeding patterns.

In summary, the Farmers Member is thought to represent a series of turbidite deposits that accumulated at the outer edge of a prograding Borden delta (Fig. 19). Moore and Clarke (1970) were first to suggest that the Farmers Member was of turbiditic

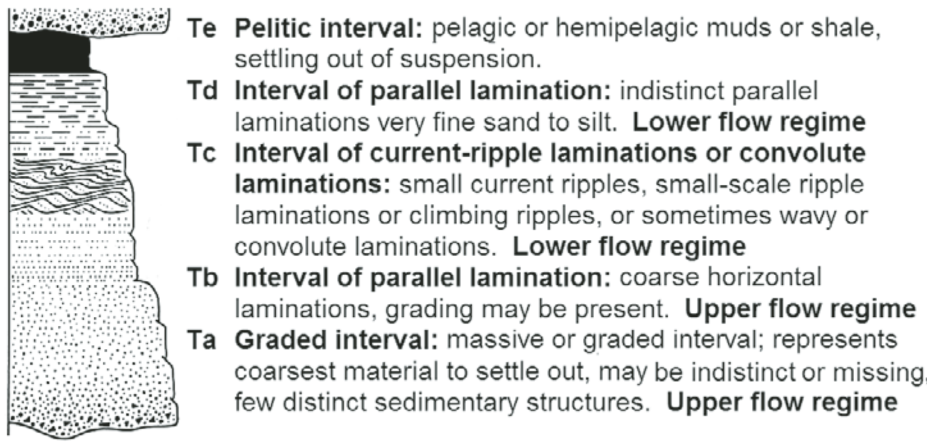


Figure 18. Five-part division of a classic Bouma cycle, showing an ideal sequence of sedimentary structures found in a turbidite bed (adapted from Walker, 1979).

origin because it exhibited many features found in a typical Bouma Sequence. However, it should be noted that the Farmers was deposited in a cratonic setting, rather than in a more classic, deep geosynclinal setting. Paleocurrent analysis suggests that the source was to the east in the Appalachian highlands (Fig. 1). 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. 1). More information and interpretations regarding the Farmers Member can be found in papers by Moore and Clarke (1970), Ettensohn (1979a), Chaplin (1980, 1982, 1985), and Mason and Lierman (1985, 1992).

Unit 5, Farmers/Nancy transition (Borden Formation). The Farmers/Nancy transition zone, an informal unit designated by Chaplin (1980), consists of 3.9 m (12.8 ft) of interbedded shale and siltstone (Figs. 13, 14). The shale is a greenish-gray, silty shale to mud shale that is poorly fissile, non-calcareous, 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 occur throughout the unit and are concentrated in distinct layers throughout the unit. 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 (5–10.5 in). The siltstones are similar 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 siltstones with *Zoophycos* being the most abundant. Body fossils can be collected from the siderite nodules

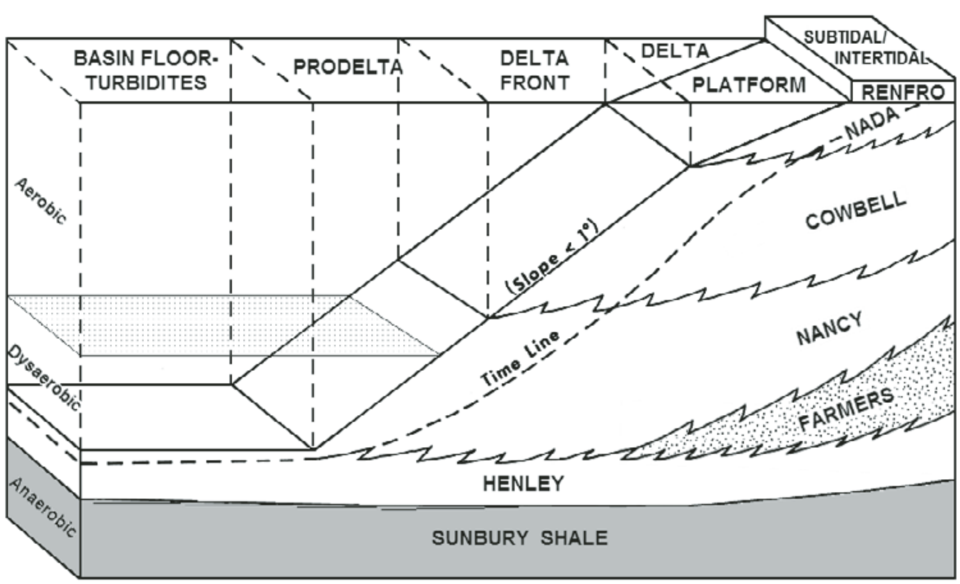


Figure 19. Block diagram showing interpreted depositional environments of the typical lithofacies of the Sunbury and Borden formations in the Morehead area, east-central Kentucky (adapted from Kepferle, 1977).

found in the interval. This transition interval is an informal designation, used for an interval that separates the thicker bedded sands of the Farmers Member below from the shales of the Nancy Member above. The thinner bedded siltstones in this interval also represent turbidites like those in the Farmers, but different in the fact that they probably reflect more distal deposition.

Unit 6, Nancy Member (Borden Formation). At this locality, the Nancy Member is incompletely exposed (Figs. 13, 14) and consists of 8.7 m (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. A zone of phosphate nodules occurs 2 m (8.2 ft) above the contact with the unit below. Body fossils in this zone include a variety of open-marine forms, including brachiopods, gastropods, cephalopods, fenestrate bryozoans, crinoid debris, solitary rugose corals, and occasionally conularids. The Cave Run lake fauna of Work and Mason (2005) also occurs in this section of the Nancy Member. This fauna is a mollusk-dominated, dysaerobic fauna that occurs in the interval between 0.75 and 1.5 m above the base of the Nancy Member.

Interpretation. The Nancy Member is interpreted to be a prodelta deposit (Fig. 19), formed as the "Borden Delta complex" prograded across east-central Kentucky (Ettensohn, 1979a; Gordon and Mason, 1985). The upper parts of this unit were mainly deposited in an aerobic environmental setting, whereas the lower portions of the Nancy, as well as the Farmers Member, were probably deposited under dysaerobic conditions. These conditions developed in the basinal seas in which units deposited reflect 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 include (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. 14) near the base of the Nancy Member (Mason and Kammer, 1984; Work and Mason, 2005).

Aerobic conditions in the upper part of the Nancy Member are suggested by (1) the presence of an open-marine fauna, (2) the high degree of bioturbation of these shales resulting in a nearly complete homogenization of the sediment, and (3) the occurrence of delta-front sands and silts of the Cowbell Member, which conformably overlie the Nancy Member (Fig. 19).

Farther north in Kentucky, the Berea Sandstone commonly occurs in a facies relationship with the Bedford Shale (Pashin and Ettensohn, 1987, 1992, 1995), and the Berea is locally an important hydrocarbon reservoir rock in eastern Kentucky and adjacent parts of Virginia, West Virginia, and Ohio (Tomastik, 1996). The Berea, however, reflects more proximal deltaic, fluvial, and platform environments (depending on location), and this field stop is basinward of those deposits (Ettensohn, 1979a). The turbiditic siltstones of the Farmers Member also form local reservoir rocks called the Weir Sandstone by drillers (Matchen and Vargo, 1996). In the subsurface of eastern Kentucky, there are several packages of turbiditic siltstones and sandstones in the same setting (Fig. 19), so there can be several different Weir Sandstones in the subsurface. 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, 1992), Lierman et al. (1992b), and Work and Mason (2005).

Stop 2: Granitic Dropstone Embedded in the Uppermost Cleveland Shale Member of the Ohio Shale

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

Stop 2 is located along Logan Hollow Branch, ~0.2 miles north of a road bridge located at the junction of Bratton Branch and Logan Hollow roads (Fig. 12). Immediately after crossing bridge, turn left onto a dirt road. We will park here and walk up Logan Hollow Road 0.2 mi to Stop 2. The stop is located on the west-central margin of the Morehead 7.5 min quadrangle in Rowan County, Kentucky. Latitude: 38° 11′ 36.1″ N; longitude: 83° 29′ 36.6″ W.

At this stop we will examine the unusual occurrence and implications of a probable granitic dropstone, called the Robinson boulder, which is present in the uppermost parts of the Cleveland Shale Member of the Ohio Shale. The presence of this dropstone indicates likely alpine glaciers to the east and elevations high enough to support them. The boulder is indirect evidence for a high, Neoacadian mountain range in latest Devonian time.

Stratigraphically, the boulder occurs at the very top of the Upper Devonian, Cleveland Member of the Ohio Shale (Fig. 20). The Bedford Shale can be easily dug out along the bank of the stream where it overlies the Cleveland Shale. Continuing down creek on the eastern side, a more complete section including both the Bedford and Sunbury shales is present. A stratigraphic section of this stop is shown in Figure 20.

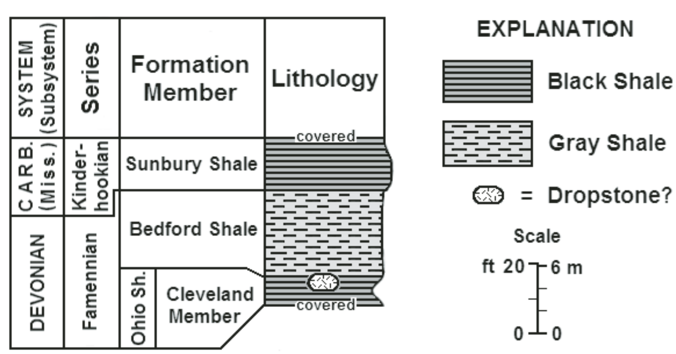


Figure 20. The stratigraphic section at the boulder locality in Logan Hollow Branch at Stop 2.

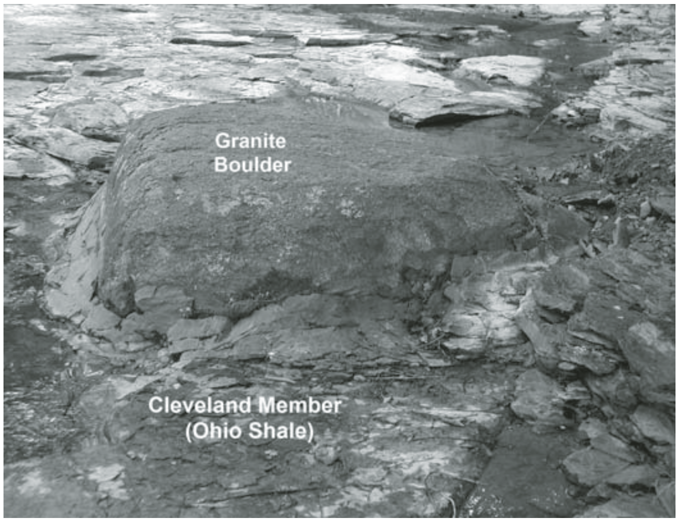


Figure 21. Side view of the granite boulder. Note the flattened sides, faceted top, and rounded corners of the boulder. Scale is 1 ft.

Statistics and Occurrence

Granite boulder. First discovered by Michael J. Robinson of NYTIS Exploration Company, LLC, in January 2006.

Size and Shape. The boulder is a roughly square-shaped mass that projects from the Cleveland Shale on the bottom of a creek along Logan Hollow. The sides of the boulder are flat while the corners and edges are rounded. The top may be faceted (Figs. 21, 22).

Size. 1.3 × 1.7 m (4.3 × 5.6 ft).

Thickness. Approximately 0.60 m (2.5 ft).

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

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

Age. Zircon crystals extracted from the boulder provided an Early Ordovician concordia age of 474 ± 5 Ma. Some of these exhibited inherited cores with a Grenvillian age of 1156 ± 230 Ma (Ettensohn et al., 2007, 2008).

Stratigraphic occurrence. The boulder sits at the top of the Cleveland Shale, overlapped by the base of the Bedford Shale (Fig. 20). The adjacent black shales reflect deposition subsequent to boulder impact and lap up onto the flanks of the boulder (Figs. 21, 22), thereby hiding any potential “crater” or deformation created by the boulder as it impacted the bottom. No evidence of smaller debris is present around the boulder, and there is no evidence of angled or disturbed bedding in the area that might suggest slumping or submarine flow.

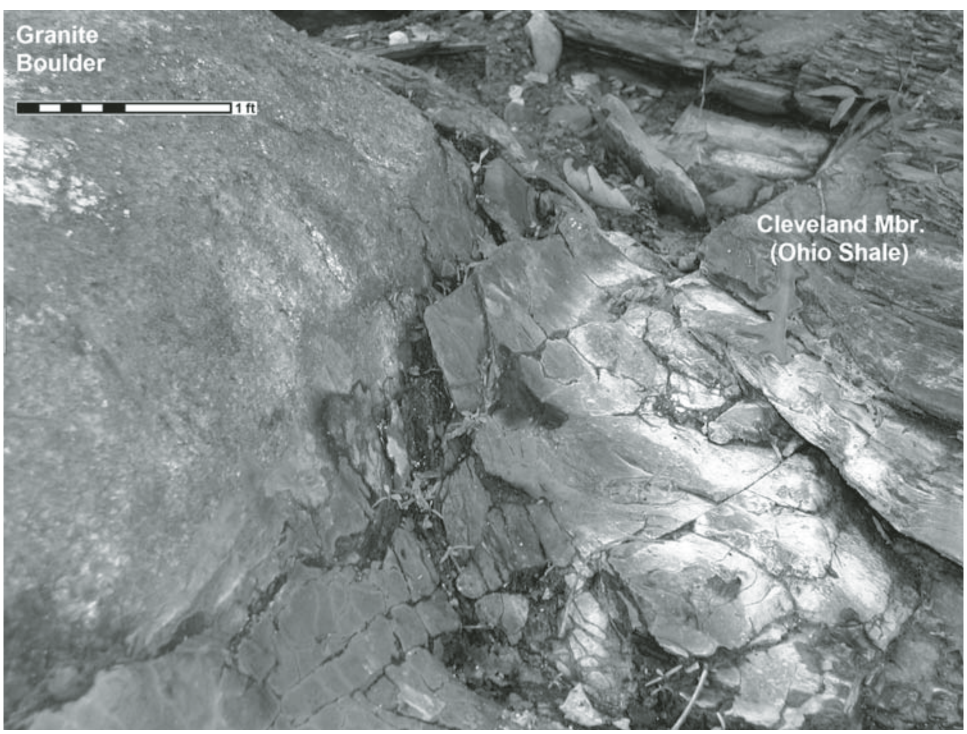


Figure 22. Boulder is clearly embedded in the Cleveland Member of the Ohio Shale. Note upturned layers or “mud drape” along edges of boulder.

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. Such an iceberg could have come from alpine glaciation present in the Acadian/Neoacadian highlands some 200–250 miles east of this locality (Fig. 23). Most Late Devonian paleogeographic reconstructions place Kentucky and the Acadian/Neoacadian mountains at ~30° south latitude (e.g., Scotese, 2003) (Fig. 23). For this object to occur in marine shales, the source must have been a tidewater glacier that extended from Acadian/Neoacadian highlands westward to the Cleveland sea. Presumably, the iceberg in which the boulder was entrained calved off along the western edge of this glacier (Fig. 23).

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 of the quartz is suggestive of Grenville-age granites in the central Appalachian region (e.g., Herz and Force, 1987). (2) The age of the granite boulder is in keeping with the ages of other Grenville rocks in the Appalachian region (e.g., Hynes and Rivers, 2010). (3) The overall shape of this boulder with its flat sides and rounded edges and corners is in keeping with 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. (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 (Figs. 21, 22). The only other mechanism that could potentially transport an object of this nature is root-rafting. Emery (1965) has noted that large boulders may be floated great distances by enclosing roots and may be deposited with or without associated plant material. The Robinson boulder, however, shows little or no evidence of rhizosphere-related weathering, and the root architecture of the large trees at the time (Algeo and Scheckler, 1998) would have severely limited the "gripping" ability of these roots. Moreover, fossil plant material was not found with the boulder and large preserved logs are

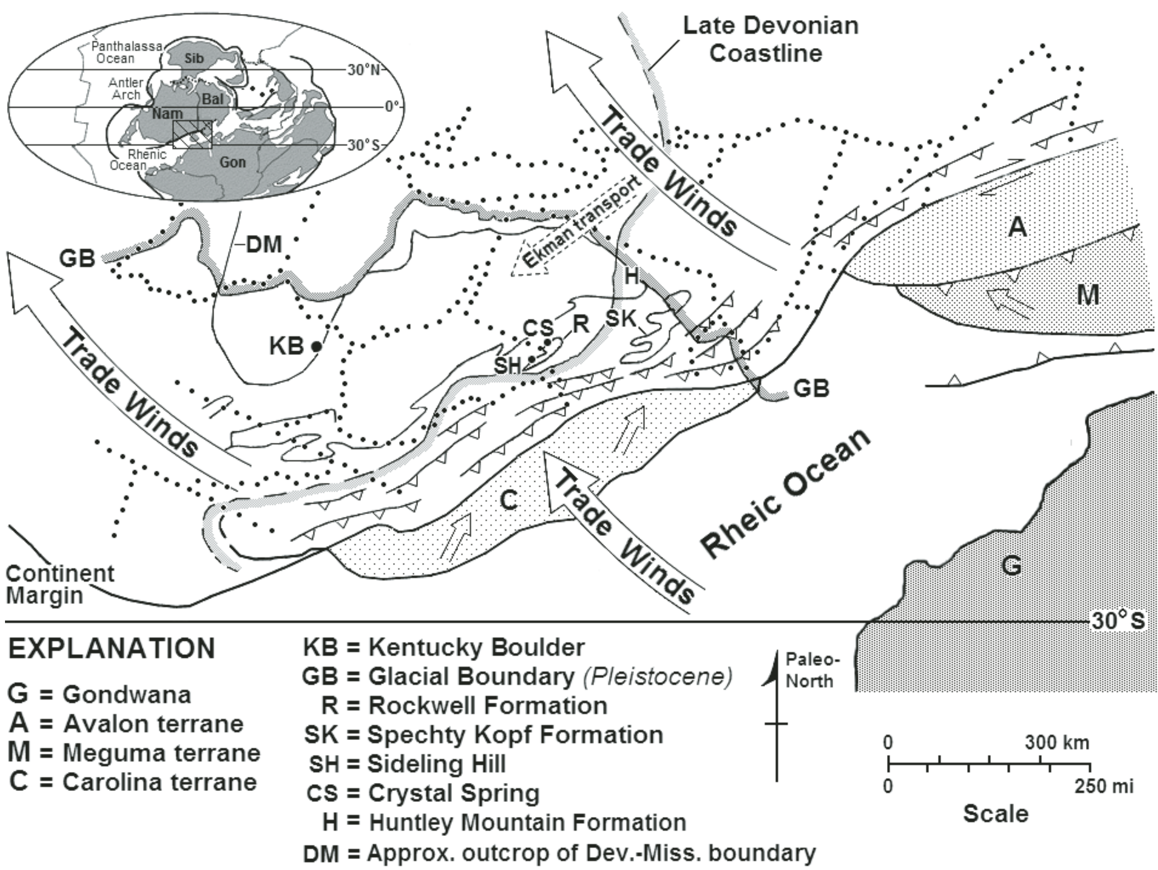


Figure 23. Late Devonian paleogeographic reconstruction of southeastern Laurussia, superimposed with positions of Devonian-Mississippian outcrop belts, field localities, Pleistocene glacial boundary, modern political boundaries, and likely Late Devonian wind and current directions. Thrust symbols represent Neoacadian highlands which supported likely alpine glaciers. Reconstruction adapted from Ettensohn (2008).

not known from Cleveland black shales in the Morehead area. So although we cannot entirely preclude the possibility of root rafting, this mode of boulder transportation seems very unlikely.

This fascinating subject is discussed in greater detail in articles and abstracts by Lierman and Mason (2007) and Ettensohn et al. (2007, 2008).

Stop 3: The Cowbell Member of the Borden Formation: Transition from Delta Front to Delta Platform

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

Here we will leave the vehicles (milepost 145.5) and walk up-section for ~0.5 mi, eastward toward the State Route 799 overpass on I-64. After viewing the section for Stop 3 (Fig. 12), 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 ~0.2 mi to Stop 4. While we are examining and discussing Stop 3, the vehicles will proceed to the Olive Hill exit (exit 156) and return along the westbound lane to Stop 4. 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 2 and is located on the Cranston 7.5 min quadrangle in Rowan County, Kentucky. Latitude: 38° 16′ 21.1″ N; longitude: 83° 23′ 5.8″ W.

At this stop, we will examine the upper parts of the Cowbell Member and its transition into the Nada Member (Fig. 8), which is interpreted to represent 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.

Unit 1, Cowbell Member (Borden Formation). The Cowbell section is incomplete here and consists of ~5.7 m (18.7 ft) of shale and siltstone (Fig. 24). The section starts with shale that grades upward into fairly massive siltstone over a vertical distance of ~3–3.5 m. The shale is an olive-gray to gray-green, silty shale which splits into platy fragments; it is non-calcareous. The shale contains grayish-red siderite nodules scattered throughout. Thin beds of siltstone within this shale become more frequent up-section. The upper third of this unit is a light olive-gray siltstone to very fine-grained sandstone. Sedimentary structures within the siltstone beds include horizontal laminae, trough cross laminae (scour-and-fill), and small soft-sediment deformation features (Fig. 25). 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 include *Zoophycos*, *Cylindrichus*, *Scalarituba*, *Helminthoida*, *Planolites*, *Lophoctenium*, *Phycodes*, and *Bergaueria*.

Unit 2, Cowbell Member. The section consists of 4.4 m (13.4 ft) of interbedded siltstone and shale (Fig. 24). 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 non-calcareous. Sedimentary structures within the siltstone beds include horizontal laminae, trough cross-laminae (scour-and-fill), as well as hummocky cross-beds (Fig. 26). 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 include: *Zoophycos*, *Cylindrichus*, *Scalarituba*, and *Monocaterion*.

Unit 3, Cowbell Member. This unit consists of ~14.3 m (46.9 ft) of interbedded siltstones and shales (Fig. 24). 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 ~10 to 30 cm thick (medium bedded). The shales are typically a gray to gray-green color, silty, split into platy fragments, and non-calcareous. Individual siltstone beds show a fining-upward character, typically with sharp (erosional) bases and upper surfaces gradational into the overlying shale. Sedimentary structures 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 exhibit mud-clast lags or are floored with a lag of fossil debris. Such lags become especially common 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.

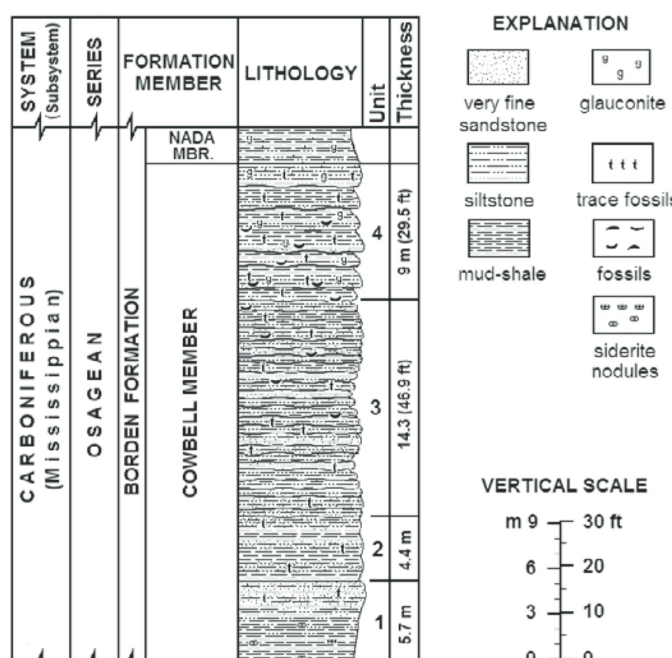


Figure 24. The stratigraphic section at Stop 3 along I-64 at milepost 145.5.

Figure 25. Typical trough cross-strata from the Cowbell Member of the Borden Formation at Stop 3.

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.*

***Unit 4*, *Cowbell Member*.** Unit 4 consists of 9.0 m (29.5 ft) of interbedded shale and siltstone (Fig. 24). The shale is a greenish-gray to grayish-green, mud shale to silty shale that is poorly fissile, non-calcareous and highly bioturbated. The siltstones are light-gray in color and commonly streaked with glauconite. Siltstone beds range from 30 to 50 cm thick (medium- to thick-bedded) and are as much as 1 m thick at the top of the 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 stratification and ripple marks. The bases of many beds contain mud-clast lags, while others are commonly floored with fossil lags. Most of the fossils in this unit are found among these lag deposits and tend to be preserved as both internal and external molds 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 *Skolithos tubes*.

Interpretation. Previous workers have suggested that the Cowbell Member represents a lower delta-front deposit comprised largely of distal-bar deposits (Kearby, 1971; Mason and Chaplin, 1979; Lierman and Mason, 1992), interpretations to which we also subscribe. Hence, ascending up-section, **Unit 1** is interpreted to represent a coarsening-upward, delta-front environment. Coarsening-upward, bioturbated siltstones and shales are common features of delta fronts (e.g., Bhattacharya, 2010).

Unit 2 is interpreted 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. The occurrence of trough cross-strata and hummocky cross-strata suggests the influence of unidirectional currents or combined flows and waves, which in this situation are hypothesized to have formed as distributary-mouth bars prograded out across prodelta deposits. Hummocky cross-stratification (Figs. 27, 28) is characteristic of shallow-marine wave and storm deposits (e.g., Harms et al., 1975). Progradation of distributary-mouth bars apparently formed the coarsening-upward profile with trough crossbedding, which was subsequently capped by a fining-upward interval with hummocky bedding, representing the reworking of this material

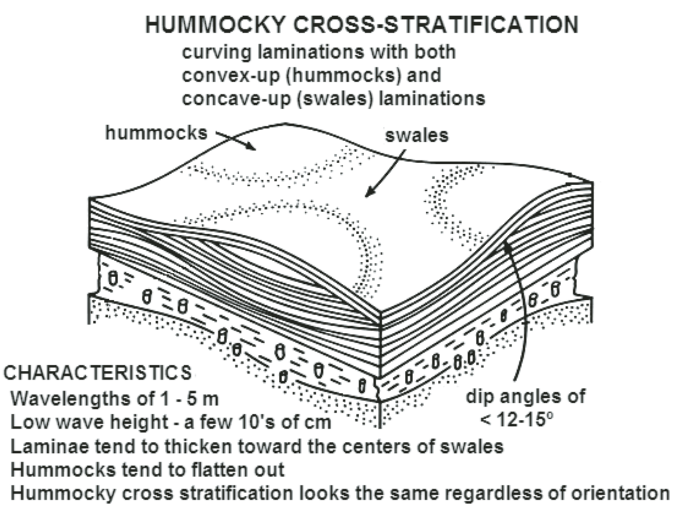


Figure 26. Characteristics and overall appearance of hummocky cross-stratification (modified from Walker, 1979).

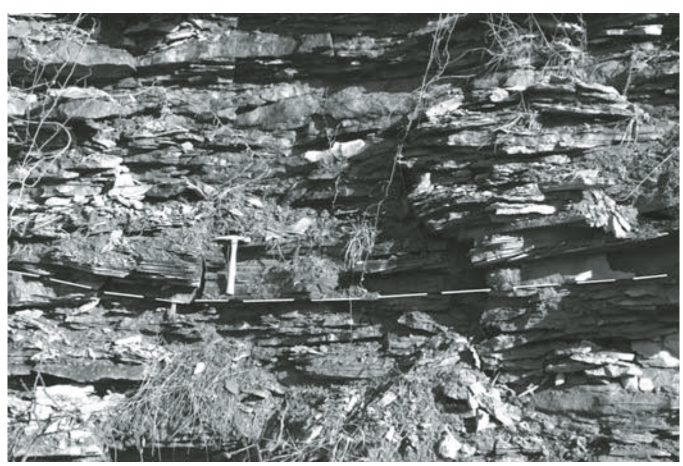


Figure 27. Section of swaley cross-bedding (one swale is outlined with white dashes) from the Cowbell Member at Stop 3.

by storm and wave activity. Trace fossils are indicative of both the *Cruziana* and *Zoophycos* ichnofacies, which encompass an environmental area ranging from the shelf/shelf edge to the slope (Seilacher, 1978). Trace fossils from the *Cruziana* ichnofacies seem to be more confined to hummocky- and trough-crossbedded parts of the section, whereas *Zoophycos* are more common in shaly parts of the section.

Unit 3 is interpreted 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. The presence of hummocky and swaley cross-strata in sharp-based siltstones capped by wave ripples is typical of tempestites (Aigner, 1985). Swaley cross-stratification (Fig. 28) is a feature related to hummocky cross-stratification but is dominated by swales or depressions at the expense of hummocks. It is commonly reported from fine- to very fine-grained sandstones 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). The trace fossils in Unit 3 are increasingly more indicative of the *Cruziana* ichnofacies, suggesting an environmental area ranging from the shelf edge to shelf or delta-platform position.

Finally, **Unit 4**, which is shale dominant, is interpreted to have been located directly on the shelf, or in this instance, on the submarine delta platform, but still above wave base because of the swaley bedding. The deeper, quieter water setting represented by the shales, 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 distal, storm-generated siltstones 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 siltstone beds, again, probably reflecting the decrease in sedimentation. Curiously, the glauconite in the siltstones seems to be concentrated in thin, wavy laminae that mimic the basic configuration of hummocky cross-strata, suggesting storm reworking after deposition. Additional information

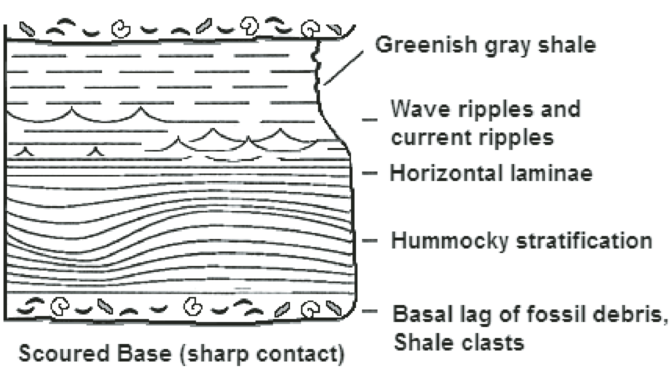


Figure 28. Typical storm-bed (tempestite) profile found in Unit 4 at Stop 3 (adapted from Kreisa and Bambach, 1982).

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 Member, at least some of those reservoirs do include the upper Cowbell and equivalent sandstones in the subsurface of eastern Kentucky and West Virginia (Vargo and Matchen, 1996).

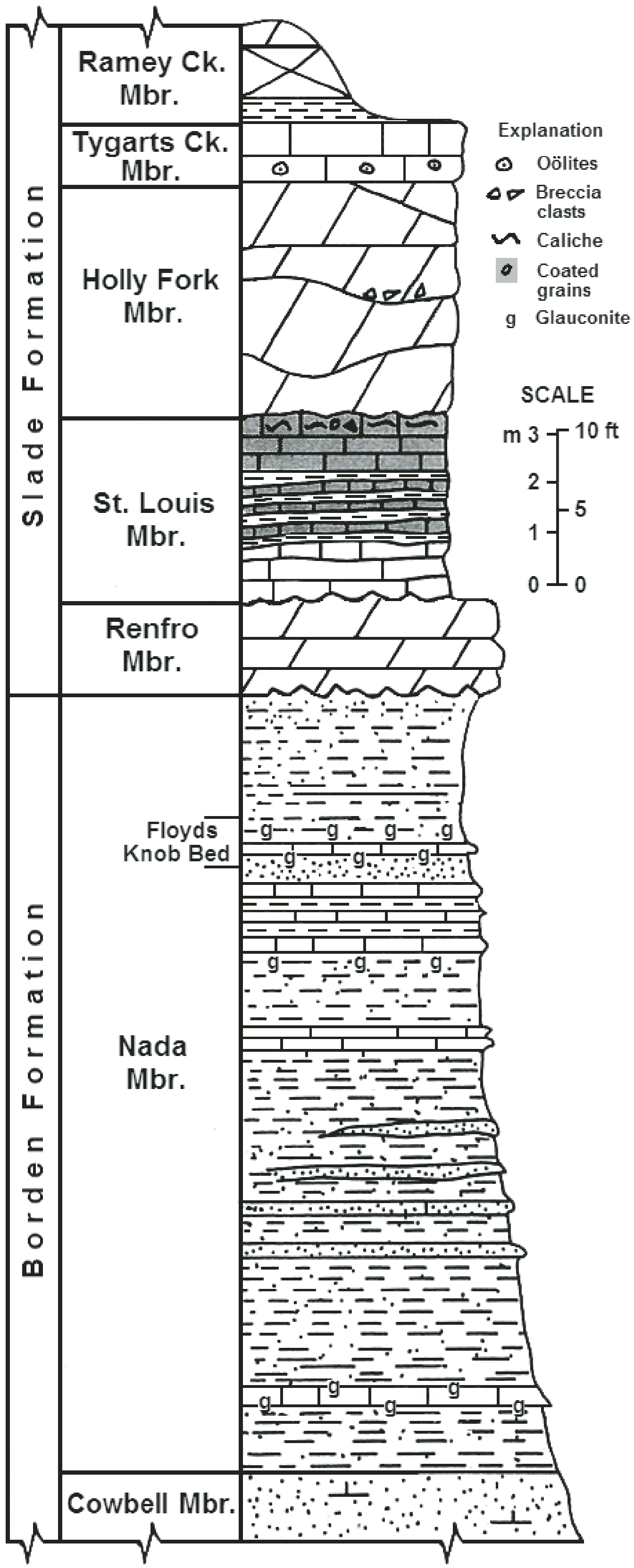
Stop 4: Borden Delta Destruction: The Nada Member and the Floyds Knob Zone in the Borden Formation

Frank R. Ettensohn, Devi B.P. Udgata, R. Thomas Lierman, and Charles E. Mason

From this point, the drivers and the vehicles will proceed to the first Olive Hill exit (Exit 156) and loop around and return to Stop 4 along the westbound lane. The stop is located in a road cut along the westbound lane of I-64 (Fig. 12), 0.5 mi east of the Kentucky State Route 799 overpass and the top of the section at Stop 3, which ends here. The stop occurs at the junction of the Cranston and Soldier 7.5 min quadrangles in Rowan County, Kentucky. Latitude: 38° 16′ 34.6″ N; longitude: 83° 22′ 30.1″ W.

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. 29), 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 at this locality was deposited in a delta-platform setting with features typical of sediment starvation and delta destruction (Fig. 10, Stop 4). Udgata (2011) equated the entire Nada at this location to the Floyds Knob zone.

This exposure occurs at the junction of the Cranston and Soldier quadrangles (Philley et al., 1974, 1975). The Nada Member here is composed of ~14.5 m (47 ft) of grayish blue-green to reddish-brown silty mudstones and clay shales with interbedded limestones and siltstones. The entire member is glauconitic, but glauconite is especially concentrated in three zones near the base and top of the unit (Fig. 29). The uppermost glauconite zone, ~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 associated here with crinoidal calcarenite beds, a rare occurrence. Also associated with the glauconite is the unusual blue-green color of most of the 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). Although the uppermost glauconite bed



alone has been called the Floyds Knob Bed, Udgata (2011) interpreted the entire Nada Member as the Floyds Knob zone (Fig. 10, Stop 4), including multiple glauconite beds.

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 (Chaplin and Mason, 1979a). 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 (Chaplin and Mason, 1979a; 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 and low-angle, planar cross beds and plane laminae are more common in the interbedded carbonates.

The contact between the Cowbell and overlying Nada Member is conformable and gradational through a short vertical 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 subunits 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 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. The Renfro and

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Figure 29. Schematic drawing of the Borden-Slade section at Stop 4 (adapted from Ettensohn, 1981). This section is composed of glauconitic shales, siltstones and carbonates and exhibits three glauconite horizons. Although the upper glauconite zone is typically identified as the Floyds Knob Bed, Udgata (2011) has indicated that the entire Nada Member represents a proximal Floyds Knob zone.

St. Louis Limestone here are mid-Visean (Meramecian) in age (Shaver, 1985). 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 latest Visean to earliest Serpukhovian (mid-Chesterian) in age (Ettensohn et al., 1984). 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, 1981, 1992a, 1992b; Ettensohn and Dever, 1979). The carbonate platform developed 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 (Fig. 6C). 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 (Carozzi, 1960; Odin and Matter, 1981).

Verdine facies also occur in sediment-starved conditions, but in iron-rich coastal waters in continental-margin settings with waters 30 to 60 m (100–200 ft) deep (Odin, 1988, 1990). Moreover, the presence of hummocky cross-bedding and tempestites 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 work discussing the Nada at this stop and elsewhere in the area can be found in Chaplin and Mason (1979a, 1979b), Chaplin (1980, 1982, 1985), Ettensohn (1981), Chaplin et al. (1985), Lierman and Mason (2004), Ettensohn et al. (2004) and Wilhelm (2008).

Stop 5: Transition between the Nancy and Cowbell Members of the Borden Formation, Northeastern Kentucky

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

Stop 5 is located in a road cut along the westbound lanes of I-64 (at mile marker 144), 2.0 miles west of Stop 4 (Fig. 12) in the Cranston 7.5 min quadrangle in Rowan County, Kentucky. At this stop, we will view the transitional contact between the Nancy and Cowbell members of the Borden Formation (Fig. 30). Latitude: 38° 16′ 4.4″ N; longitude: 83° 24′ 30.2″ W.

This stop illustrates the contact between the Nancy and Cowbell members of the Borden Formation (Fig. 8) and the transitional interval between (Fig. 30) on the lower Borden delta slope (Fig. 19). Sedimentary structures, body fossils, and trace fossils are present.

Only the uppermost 7.9 m (26.2 ft) of the Nancy Member is exposed at this stop (Fig. 30). The entire Nancy Member in the Morehead area ranges in thickness from 18.2 to 60.9 m (60–200 ft). The maximum thicknesses of the Nancy Member on U.S. Geological Survey geologic quadrangle maps, however, can be misleading, as the unit's contacts do not appear to have been picked consistently and the upper and lower contacts are gradational. 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, as much as 60.9 cm (24 in) thick, is common in the lower 9 m (30 ft) of the unit, 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 uppermost continuous interval of siltstone beds. Its gradational upper contact is placed at the base of the first thick continuous interval 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 Nancy shale lithology into the more silty, overlying, Cowbell Member (Fig. 30).

The Nancy Member is highly bioturbated so that most traces of bedding and other sedimentary structures are destroyed. Although bioturbation is abundant, identifiable trace-fossil diversity is low with the ichnogenera *Phycosiphon*, *Scalarituba*, and *Zoophycos* being the most common. *Zoophycos* is locally most abundant in the siltier shale intervals. Other associated trace fossils include horizontal feeding traces and vertical

Figure 30. Photograph showing the lower slope Nancy-Cowbell transition in the Borden Formation at Stop 5.

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 common in the ironstone concretions, lenses, and beds of the Nancy Member. The most common fossil forms include brachiopods, cephalopods, gastropods and conularids. Less abundant are trilobite fragments, bryozoans, crinoid detritus, bivalves, hyolithids, ostracods, corals, conodonts, foraminifera, 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 at places, marcasite, dolomite, and quartz (Mason and Chaplin, 1979; Lierman and Mason, 1992).

The Cowbell Member at the stop is ~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. Siderite nodules and lenses are locally very fossiliferous as in the Nada Member. 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 parallel laminae, ripple marks, ripple-drift cross laminae, cut-and-fill and small-scale, and low-angle planar cross-bedding. 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 (1971) 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*) and *Zoophycos*. Less common are *Asteriacites* and *Cruziana*. 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 (Chaplin and Mason, 1979a; Mason and Chaplin, 1979). 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. 19), 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 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 2002; 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.

Stop 6 (A and B): Borden Delta Sequence at the Delta-Platform Margin—Big Hill

The entire Borden delta sequence is present at this stop, but much of it is covered or poorly exposed (Fig. 31). This stop is located 117 km (70 mi) farther southwest of the delta sequence observed yesterday at Stops 1–5. Both sequences are probably located on the same delta lobe (Fig. 11), but yesterday's stops were located high on the proximal delta platform, whereas this stop is located approximately at the platform margin (Fig. 10). We will first examine the contact of the lower Borden delta sequence with the basinal New Albany black shales at Stop 6A and move to the top of the sequence to examine the Nada Member at Stop 6B. Stop 6A, begins at the very base of the Bighill section on the west side of U.S. 421, ~0.9 mi (1.5 km) south of the intersection of U.S. 421 and Ky. 21 at Bighill. The section is located in the southwestern corner (rectangle 7) of the Bighill 7.5 min quadrangle in Madison County, east-central Kentucky (Fig. 12). Latitude: 37° 31′ 43.98″ N; longitude: 84° 12′ 40.98″ W.

Stop 6A

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At this stop we will examine (1) the uppermost New Albany Shale, as well as (2) the Jacobs Chapel Bed, Rockford Limestone, and basal Borden Formation (Fig. 31). The Jacobs Chapel Bed and Rockford Limestone are equivalent to the lower part of the Henley Bed at Stop 1. Some of the more interesting features to observe and discuss include: (1) both anaerobic and dysaerobic conditions in the marine environment at the time of sediment deposition, and (2) basin-floor/prodelta deposits.

Stop 6A includes, in ascending order, the New Albany Shale, Jacobs Chapel Bed, Rockford Limestone, and lowermost foot or two of the Nancy Member of the Borden Formation (Fig. 31). In order to examine these contacts, we exposed 0.61 m (~2 ft) of section at the base of the Nancy. At the base of the section we can see the uppermost foot (30 cm) of the New Albany Shale. The New Albany is a carbonaceous, black to brownish-black mud-shale that is noncalcareous and displays platy to fissile weathering. Microscopic examination of the New Albany shows a sparse scattering of finely crystalline pyrite. Fossils are rare in the unit, but include small inarticulate brachiopods, scattered conodont remains, spores, and fish teeth (Weir et al., 1971). The presence of the conodonts *Siphonodella* sp. and *Spathognathodus* sp. in nearby shales (Weir et al., 1971; Ettensohn, 1979b) points to an Early Mississippian (Kinderhookian) age for the upper part of the New Albany Shale. Hence, the uppermost few inches of the New Albany Shale is equivalent to the entire Sunbury Shale at Stop 1. The New Albany Shale, similar to the equivalent Ohio Shale, has been interpreted to represent an anoxic, basin-floor deposit (e.g., Ettensohn and Barron, 1981; Ettensohn et al., 1988a). The shale covers a wide geographic area and provided a base for the subsequent deposition of the greenish-gray shales of the Nancy Member of the Borden Delta complex (Figs. 1, 8, 10, 19), which overlie the Sunbury Shale to the east. Immediately above the New Albany Shale rests the Jacobs Chapel Bed. It is likewise of Early Mississippian (Kinderhookian) age, and at this locality is ~50 cm thick. The Jacobs Chapel is a mud-shale to clay-shale that is greenish gray, noncalcareous, and exhibits a platy to flaggy splitting property. The lower 10–12 cm contains a scattering of small (less than 3 cm), elliptical phosphate nodules, as well as phosphatic fossils such as conodonts and a scattering of other vertebrate remains. Also present along the very top of the Jacobs Chapel is a very thin (less than 2 cm thick) bed of argillaceous limestone, with abundant cone-in-cone structures. The term "Jacobs Chapel Shale" applies to a thin (less than 0.3 m) layer of greenish shale that overlies the New Albany (black) Shale and underlies the Rockford Limestone in southern Indiana (Campbell, 1946). Conodonts collected from the Jacobs Chapel belong to the lower *Siphonodella crenulata* Assemblage Zone (Sandberg et al., 2002). This clearly places the Jacobs Chapel in the Kinderhookian Series. As in past interpretations (e.g., Campbell, 1946; Lierman and Mason, 2004), we concur that the

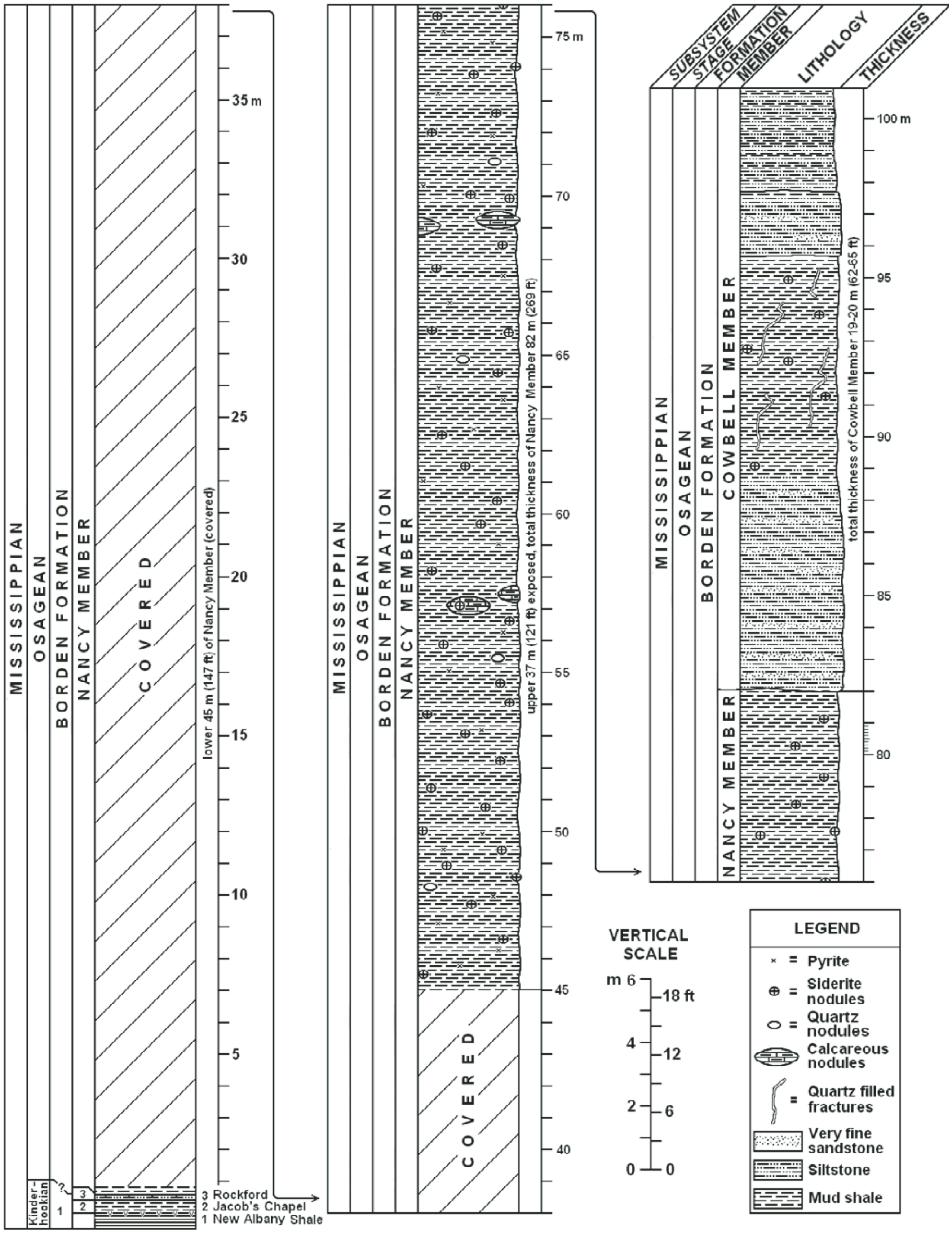

Figure 31. Composite stratigraphic section for the Nancy and Cowbell members of the Borden Formation at Stop 6A along the Bighill roadcut on U.S. 421, Madison County (from Lierman and Mason, 2004).

Jacobs Chapel represents a basin-floor (hemipelagic) deposit that accumulated under dysaerobic conditions with very slow rates of sedimentation, conditions very similar to those noted by Kammer et al. (1986) for Paleozoic dysaerobic environments. The Jacobs Chapel Bed also is substantially thicker (50 cm) at this stop than in the type section in southern Indiana (10–15 cm), suggesting that sedimentation rates were higher in the Appalachian Basin at this time than in the Illinois Basin where the type section is located. At this locality the Jacobs Chapel bed is overlain by a very thin-bedded (7 cm) siltstone that is greenish gray, lightly calcareous, and highly bioturbated. This siltstone is overlain by the Nancy Member of the Borden Formation. The Jacobs Chapel bed is usually overlain by the Rockford Limestone, but at this locality a limestone is missing and a calcareous, glauconitic siltstone (~10 percent carbonate) occurs in its stratigraphic position. The base of the siltstone is sharp, almost erosional with the underlying Jacobs Chapel, and grades upward into the overlying Nancy Member of the Borden Formation. This Rockford-equivalent siltstone at this locality is interpreted to as an isolated, distal turbidity flow that came off a carbonate platform to the west. This interpretation is based on the sharp erosional base of the siltstone bed and gradational upper surface with the overlying shales of the Nancy Member and the stratigraphic position of the siltstone between two deeper-water marine shales (the Jacobs Chapel Bed and lower Nancy Member).

Only the lowermost part of the Nancy Member is exposed at this stop. It is a greenish gray, noncalcareous mud-shale that weathers into platy fragments. The lowermost Nancy Member and the Jacobs Chapel Bed then appear to represent basin-floor deposits that accumulated under dysaerobic conditions. At the time of deposition, this area was basinward of the Borden Delta complex (Figs. 1, 10, 19). Only the finest grain fraction (clays and fine silts) was apparently transported to this location at the time.

Stop 6B

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At this stop we will examine the uppermost part of the Borden Formation, including the Nada and Wildie members (Fig. 32), although both are commonly mapped as the Nada Member. Four major glauconite horizons are present, the thickest of which has been called the Floyds Knob Bed (Fig. 32), but Udgata (2011) has included the entire unit as part of his Floyds Knob zone (Fig. 10).

Lithology. The Nada is the upper member of the Borden Formation, and at least 11.3 m (37 ft) are present in the exposure. At this stop, the Nada can be divided into three informal units (Fig. 33), however, the upper 7.0 m (23 ft) of the unit will be the focus, since this part of the unit represents the transition to carbonate deposition in the overlying Slade Formation. In contrast to the greenish-gray, dark-gray, or grayish-red silty mudstones and shales that comprise underlying parts of the unit, the upper part of the Nada Member is largely composed of pale blue-green, silty mudstones, siltstones, and shales (Fig. 32), in what some workers in other areas have called a verdine facies (e.g., Odin, 1985, 1988, 1990). Other parts of the unit are rich in glauconite, comprising what would have been termed a glaucony facies (e.g., Odin and Matter, 1981; Odin, 1988).

The upper part of the Nada Member, which is equivalent to the Wildie Member elsewhere, begins with two rusty-brown beds of argillaceous, silty, glauconitic dolostone, separated by greenish-gray shale (Fig. 32). The base of each dolostone is a glauconite- and phosphorite-rich pause or lag horizon, containing reworked fish bones and teeth, invertebrate fossil debris, and phosphorite nodules. Each bed is amalgamated, contains sparse fossil fragments, and was probably composed of calcareous-mud and glauconite pellets. Sedimentary structures include subtle, hummocky crossbeds, scours, bioturbation, and rip-up clasts; the top of the lower dolostone may exhibit megaripples. The upper dolostone bed contains randomly arranged silica (chalcedony and quartz) geodes that appear to replace dolostone. The upper dolostone is overlain by mudstone and shale that contain a prominent, moderate blue-green layer of phosphorite and glauconite. This intensely glauconitic horizon is a typical glaucony facies and forms a prominent reentrant below the first siltstone. This horizon represents the widespread Floyds Knob Bed (Stockdale, 1939; Kepferle, 1971; Whitehead, 1978; Sable and Dever, 1990), which is 0.1 m (0.3 ft) thick, has a sharp base, and is gradational upward; it contains phosphorite nodules, fossil fragments, and

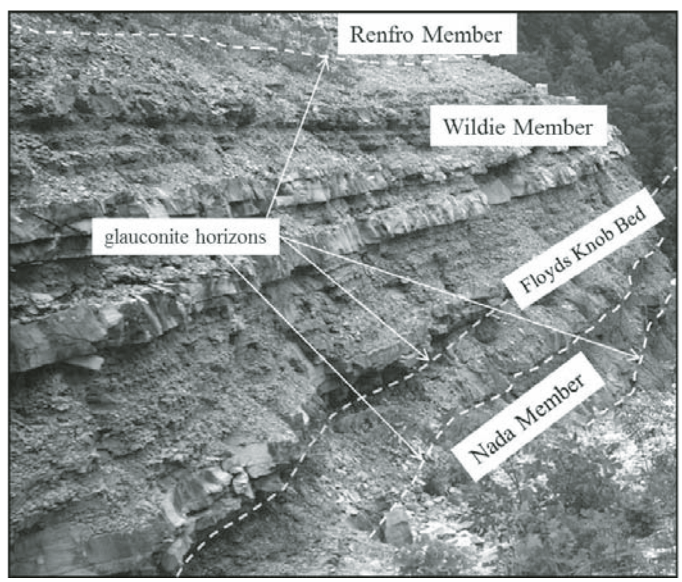

Figure 32. Picture of the upper part of the Nada/Wildie members at Stop 6B (Bighill) showing the disposition of four glauconite beds (dashed lines). The thickest glauconite bed is labeled as the Floyds Knob Bed, but Udgata (2011) has called this entire interval of glauconite beds the Floyds Knob zone. The interbedded siltstones and shales above the Floyds Knob Bed are equivalent to the Wildie Member of the Borden (Weir et al., 1966). The upper glauconite layer is the contact with the overlying Renfro Member of the Slade Formation.

burrows, which are infilled with glauconite and phosphorite (Figs. 32, 33).

Immediately above the Floyds Knob Bed, a true verdine facies begins, and it is composed of 4.6 m (15.0 ft) of thin, even-bedded, blue-green siltstones interbedded with silty, glauconitic mudstones and shales of the same color; this is the Wildie Member of Weir et al. (1966). This part of the section up to the base of the Renfro Member is also included in Udgata's Floyds Knob zone (Figs. 10, 31, 32). At present the siltstones are not weathered enough for sedimentary structures to stand out prominently; however, close examination reveals the presence of amalgamated beds, crude grading succeeded by micro-cross-laminae, swaley crossbeds, scours, rip-up clasts, and bioturbation at or near the tops of each siltstone. The second siltstone bed has a horizon of phosphorite nodules at its top.

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

Interpretation. The Nada Member has been interpreted to represent delta destruction after abandonment or diversion of Borden deltas in eastern Kentucky (Ettensohn, 1979a, 1980, 1981). In northeastern Kentucky (yesterday's Stops 1, 3, 4, 5), where much detailed work on the Borden Delta has been done (e.g., Chaplin and Mason, 1979a, 1979b; Chaplin, 1980), the

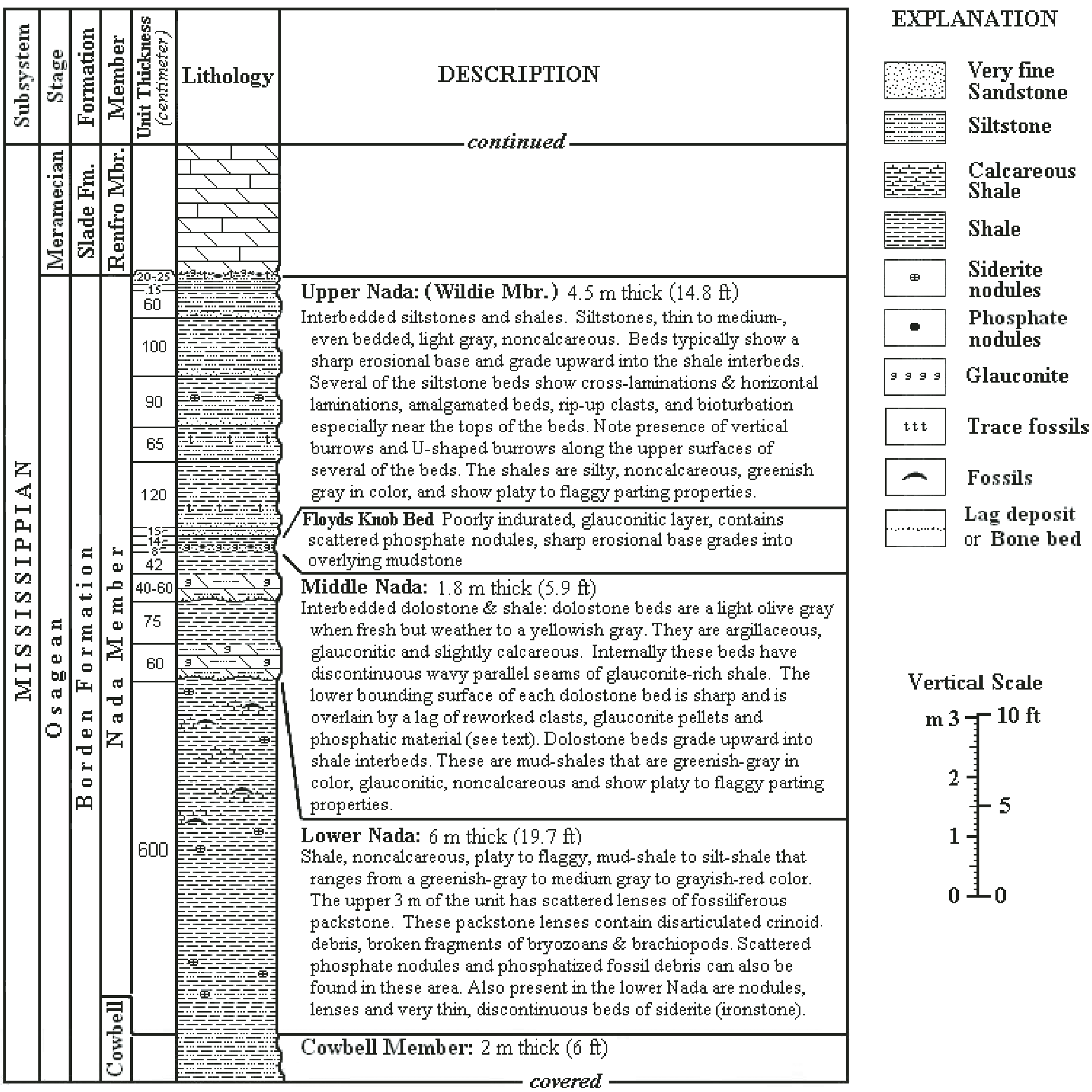

Figure 33. Columnar stratigraphic section from the top of the Cowbell Member of the Borden Formation to the lower Renfro Member of the Slade Formation at Stop 6B along Bighill roadcut on U.S. 421, Madison County (from Lierman and Mason, 2004).

Nada is characterized by abundant limestones and diverse faunas interpreted to represent shallow-subtidal, open-marine environments (Figs. 11 [Stop 4], 29). This contrasts with the Nada in this exposure, 117 km (70 mi) to the south (Fig. 11, Stop 6), in which limestones and fauna are rare. The differences are probably best explained by the more distal position of the locality on the delta lobe (Figs. 10, 11) and by position relative to the Kentucky River Fault Zone (Figs. 10, 11), which was periodically reactivated during Mississippian time (Dever et al., 1977; Ettensohn, 1979a, 1980, 1992a). Northeastern Kentucky occupied the upthrown side of the fault near the Waverly Arch, a situation that generated a very shallow platform setting conducive to carbonate deposition and diverse faunas. The Bighill locality, however, is located ~83 km (50 mi)) south of the fault zone on the downthrown side in conditions that were probably much deeper. South of the fault, the Nada is composed of mostly mudstones and shales, and appears to represent an area of deep-ramp shelf muds. The ramp, however, was not below storm wave base, because the presence of debris-filled scours, starved ripples, and laminae of fossil debris probably represents distal tempestites emplaced by storm backflow (e.g., Aigner, 1985). At times, moreover, sedimentation must have been slow enough to allow quiet-bottom communities of delicate bryozoans, brachiopods, and crinoids to develop on the soft bottoms. The upward change into thicker, coarser dolostones and siltstones in the upper Nada Member marks a transition into more proximal, midramp, storm-shelf conditions with banks formed first of transported carbonate and glauconite pellets and then silt. The presence of thicker bedding, bed amalgamation, and swaley crossbeds all point to more proximal, mid-ramp conditions (Aigner, 1985; Pashin and Ettensohn, 1987).

Between the dolostones and siltstones, however, is the glauconite- and phosphorite-rich Floyds Knob Bed (Figs. 32, 33); a similar bed also occurs at the Nada-Renfro contact, which forms the upper contact of Udgata's Floyds Knob zone. Glauconite typically forms in upper-slope and outer-shelf environments deeper than 60 m (Odin and Matter, 1981). Glauconite and phosphorite commonly occur together in areas transitional between deep and shallow waters, slightly deficient in oxygen, with slightly lower-than-normal pH, and where clastic sedimentation is very slow to nonexistent (Carozzi, 1960; Hatch and Rastall, 1965). Initial precipitation of phosphorite may have been related to episodic upwelling into the area (e.g., Carozzi, 1960; Prévôt and Lucas, 1990), but the fragmentation, boring, and probable rolling of many nodules suggest repeated reworking and concentration, perhaps by storm currents, during times of sediment starvation.

The verdine facies (Wildie Member equivalent) between the two glauconite beds reflects the in-situ modification of clays by cation exchange in very iron-rich waters (Odin, 1988). Although the iron content of the waters had to be high, overall sediment influx was low. In modern settings, similar verdine facies occur in extensive, subtropical to tropical, continental-margin settings in waters 20–60 m deep, 30–60 km from the coast (Odin, 1988; Thamban and Rao, 2000). These conditions provide some idea of how the transitional Nada-to-Renfro shelf must have appeared. Overall, the presence of carbonate- and clastic-rich storm deposits in the upper Nada Member reflects shallowing into upper-slope, storm-shelf conditions. The co-occurrence of verdine, glaucony, and phosphoritic facies in Wildie equivalents and the Floyds Knob Bed/zone strongly supports periods of sharply reduced clastic sedimentation on subaqueous parts of the Nada slope, subject to reworking by storms (Figs. 10, 11).

Stop 7: Floyds Knob Zone in the Delta-Front Siltstones and Carbonates of the Halls Gap and Muldraugh Members of the Borden Formation

Frank R. Ettensohn, Devi B.P. Udgata, R. Thomas Lierman, and Charles E. Mason

Stop 7 is located on the west side of U.S. 127, south of Liberty, Kentucky (Fig. 12). The section is located on the Liberty 7.5 min quadrangle in Casey County, central Kentucky. Latitude: 37° 17′ 8.4″ N; longitude: 84° 57′ 31.38″ W.

Stop 7 shows nearly the entire Borden sequence in south-central Kentucky in a delta-front setting (Figs. 8, 10, 11). Although the contact with the New Albany black shale is not exposed, a basal New Providence equivalent, approximately equal to the Henley Bed at Stop 1 and the Jacobs Chapel and Rockford-equivalent siltstone at Stop 6, the Nancy Shale, the Halls Gap Member, and the Muldraugh Member of the Borden are present (Fig. 8). The stop will concentrate on the glauconite-bearing parts of the Halls Gap and Muldraugh members (Fig. 34) that occur just below and above the upper bench in the exposure. The Halls Gap member in this exposure is largely composed of light greenish-gray, massive to shaly, calcareous siltstone and interbedded shale with bioturbation and fossils. A prominent detrital limestone occurs in the member at the level of the bench, and hummocky crossbeds are present locally in the siltstones. Three glauconite horizons occur in the Halls Gap Member, two in the four meters of section below the bench, and another, about a meter above the bench just below the Muldraugh Member. The Muldraugh Member consists largely of thin- to medium-bedded, fine-grained, bioturbated, unfossiliferous cherty dolostones. A thin glauconite horizon occurs ~7.5 m (24.6 ft) above the base of the unit (Fig. 34). The thick glauconite horizon at the base of the Muldraugh Member is what most workers would call the Floyds Knob Bed, but Udgata (2011) has included the upper parts of the Halls Gap and lower parts of the Muldraugh members in the larger Floyds Knob zone on the delta front (Fig. 10) and suggested that they are approximately equivalent to the Nada Member at Stop 6, based on the occurrence of the same number of glauconite beds at both stops (Figs. 32, 34). The Halls Gap Member appears to be another storm-reworked siltstone unit, like the Wildie at Stop 6, only farther down the delta front, whereas the Muldraugh Member, a bioturbated, cherty dolostone, appears to represent a deeper water carbonate accumulation on a gently

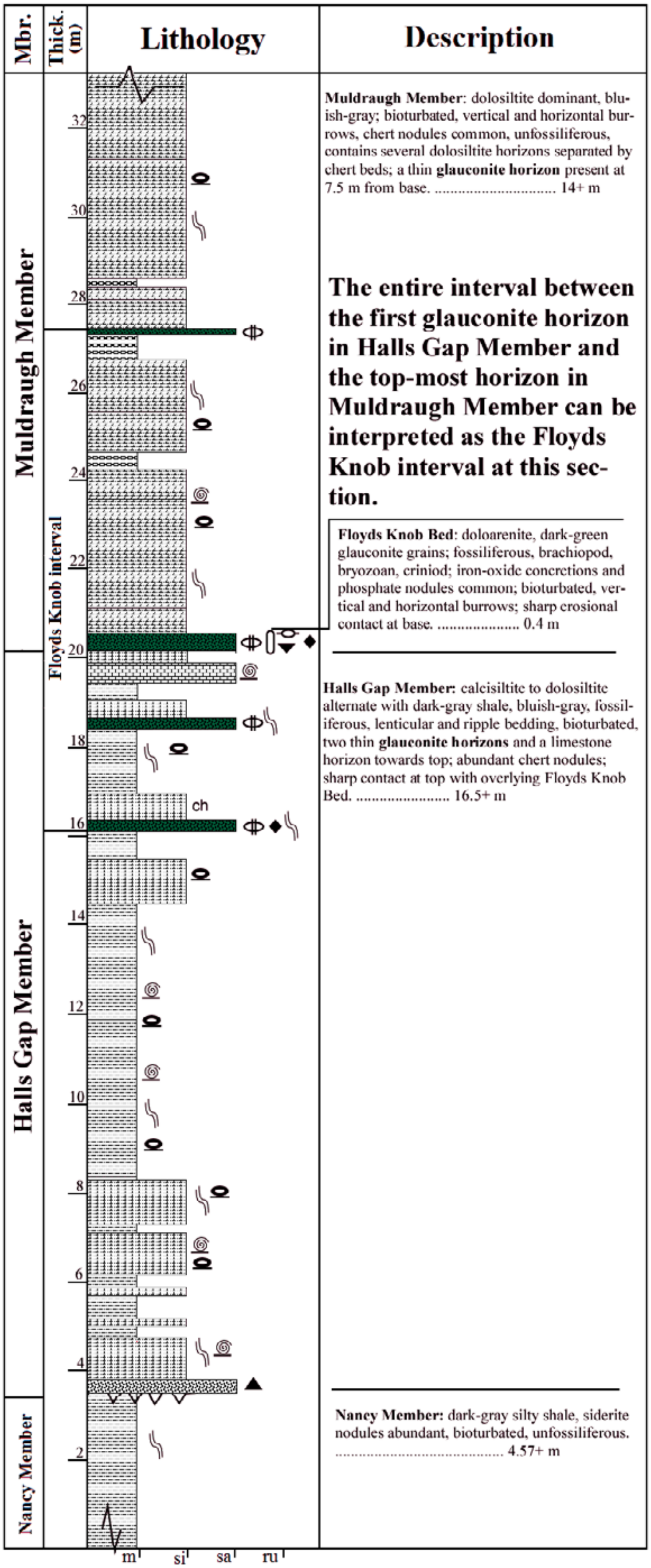

Figure 34. Columnar section for Stop 7 south of Liberty, Kentucky. The Floyds Knob zone here includes four glauconite beds in the Halls Gap and Muldraugh members of the Borden Formation. The thick glauconite bed just below the Muldraugh Member would be called the Floyds Knob Bed. Legend as in Figure 35 (from Udgata, 2011).

sloping delta-front environment (Fig. 10). In contrast to the calcareous, glauconitic muds that have predominated in more proximal delta-platform environments to the north in Stops 4 and 6, pelletal glauconites predominate in the more distal delta-front settings like this, where clastic sedimentation was not as intense.

Optional Stops A and B: Fort Payne Basin at the Toe of the Borden Delta Slope

Devi B.P. Udgata, Frank R. Ettensohn, R. Thomas Lierman, and Charles E. Mason

If time permits before lunch, we may do one of two optional stops that show features in the Fort Payne basin at the foot of the delta slope (Figs. 10, 11). Both features occur close to the dot at Wolf Creek Dam on the location map (Fig. 12). Optional Stop A is located on Swan Pond Road (latitude: 36° 52′ 34.44″ N; longitude: 85° 9′ 43.74″ W) and Optional Stop B is located on Manntown Road (latitude: 36° 51′ 11.16″ N; longitude: 85° 12′ 15.78″ W). Both of these stops are located on the Creelsboro 7.5 min quadrangle in Russell County, south-central Kentucky.

Optional Stop A is a section that shows the presence of the Floyds Knob Bed a little more than 5 m (16.4 ft) above the Maury Shale (Figs. 10, 36). The Maury Shale is a glauconitic mudstone that sits immediately above the Chattanooga black shale (Fig. 36) the Maury Shale has been mapped as a formal

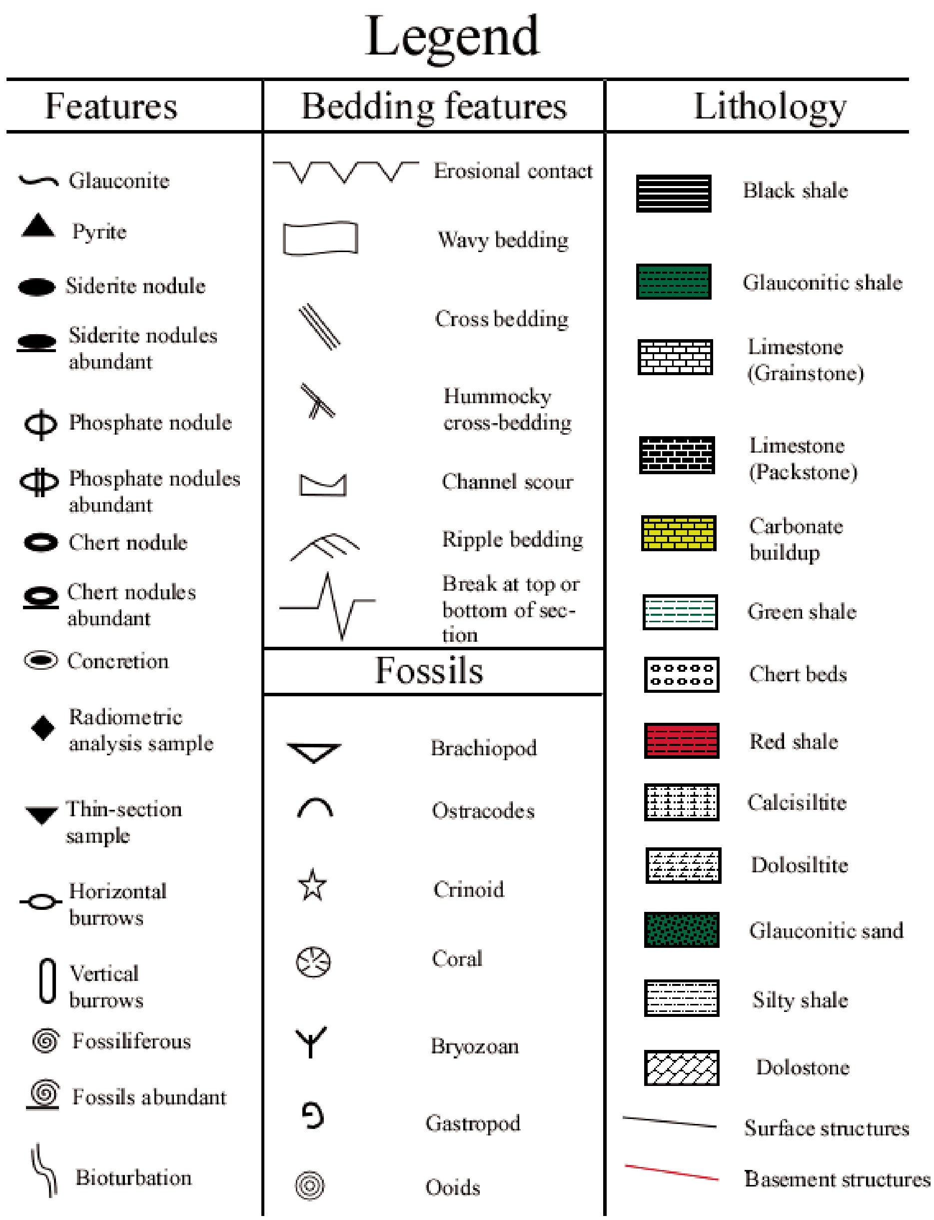

Figure 35. Legend for stratigraphic columns (Figs. 34, 36, 37, 40, 41, 44, and 46) (from Udgata, 2011).

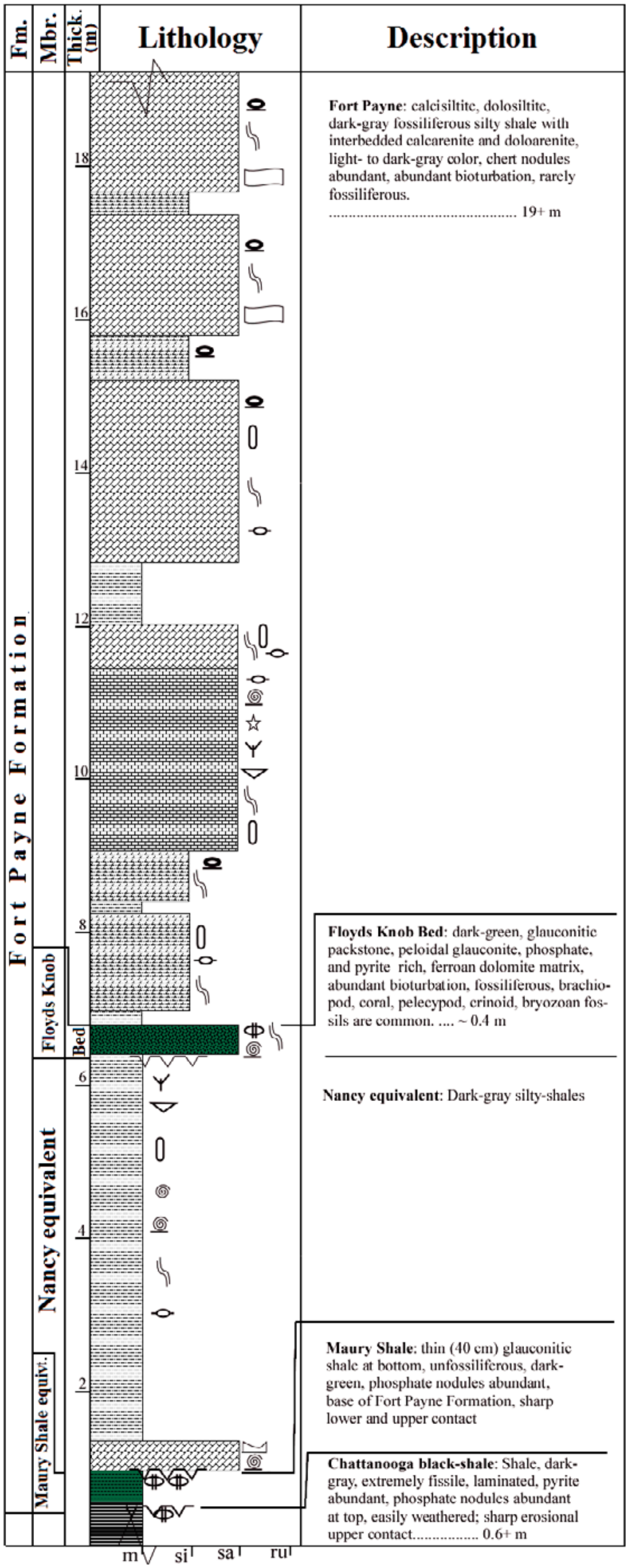

Figure 36. Columnar section for Optional Stop A on Swan Pond Road. Note the short interval between the glauconitic Maury and Floyds Knob beds. Legend as in Figure 35 (from Udgata, 2011).

unit in Tennessee, but not in Kentucky, although it is present as a thin bed in basal parts of the Fort Payne Formation in south-central Kentucky. The Maury is at least partially equivalent to the Henley Bed at Stop 1, the Jacobs Chapel Shale at Stop 6, and New Providence equivalents at Stop 7 and elsewhere (Sable and Dever, 1990). A possible Rockford Limestone equivalent is also present at this section. What is especially interesting at this stop is the vertical proximity of the glauconitic Floyds Knob and Maury units. At previous stops, the Floyds Knob Bed/zone occurred on top of a thick succession of Borden deltaic sediments. At this stop, southwest of the Borden delta front, 5 m of section between the Maury and Floyds Knob Bed/zone is equivalent to more than 230 m of section further north on the Borden delta front (Fig. 10). If the Floyds Knob Bed/zone is truly a chronostratigraphic marker horizon across the Borden delta and Fort Payne basin to its front (e.g., Peterson and Kepferle, 1970; Kepferle, 1979; Sable and Dever, 1990), then this stop represents a very distal, sediment-starved part of the basin in front of the abandoned delta. The Floyds Knob Bed/zone here is a single bed, ~0.4 m (1.3 ft) thick (Fig. 36). The nature and thickness of the zone will change dramatically in little more than 5 km (3 mi) to the southwest at Optional Stop B.

Optional Stop B is ~5 km (3 mi) to the southwest of Stop A (Figs. 11, 12). The exposure begins with 40 cm of glauconitic Maury Shale overlying an interval of phosphatic nodules (Falling Run Bed) at the top of the New Albany Shale. Overlying the Maury is a 1.5-m-thick (4.9-ft-thick) bed of dolostone that is probably equivalent to the Rockford Limestone, as at Optional Stop A. Immediately overlying the dolostone, is an interval of at least 20 m (66 ft) of green glauconitic shale with interbedded limestones. About 5 m (16.4 ft) above the contact with the possible Rockford Limestone, within this glauconitic green-shale interval, is a 3-m-thick (10-ft-thick) carbonate mud mound (Fig. 37). The mud mound is composed of poorly sorted, micrite-rich, crinoid-bryozoan-brachiopod mudstone to wackestone with a convex-upward surface and steep depositional slopes (Fig. 38). Silicified crinoid, bryozoan, and brachiopod debris has weathered from the mound, but the mound surface shows a clotted to *Stromatactis*-like texture (Fig. 39) that may suggest an algal origin.

At this stop, there is no distinct Floyds Knob Bed, as at Optional Stop A. Instead, the Floyds Knob is interpreted to be a 20-m-thick (66-ft-thick) zone of green, glauconitic shale that contains a small carbonate mud mound and 11 interbedded limestone layers (Fig. 37). Like Stop A, the glauconitic Maury Shale and a possible Rockford Limestone are present, but the glauconitic Floyds Knob zone begins immediately above the Rockford. Inasmuch as there is virtually no sedimentary interval between the Rockford and Floyds Knob zone, this area may reflect one of the most sediment-starved parts of the basin.

The lower Fort Payne Formation west of the Borden delta front is famous for a carbonate mud-mound facies, similar in many respects to Waulsortian mounds (e.g., Lewis and Potter, 1978; MacQuown and Perkins, 1982; Ausich and Meyer, 1990; Krause and Meyer, 2004). Several scenarios have been interpreted for mound development in this part of the Fort Payne, but most authors agree that sediment baffling by crinoids and bryozoans may have also contributed to mound growth (e.g., MacQuown and Perkins, 1982; Ausich and Meyer, 1990; Stapor and Knox, 1995).

Stop 8: Floyds Knob Bed on Western Side of the Fort Payne Basin

Devi B.P. Udgata, Frank R. Ettensohn, R. Thomas Lierman, and Charles E. Mason

This stop shows how the Floyds Knob Bed appears on the western side of the Fort Payne basin where it is again a single bed. The exposure is located on Kentucky State Route 90, west of Burkesville, Kentucky (Fig. 12). Latitude: 36° 47′ 59.04″ N; longitude: 85° 22′ 55.86″ W. This section is located on the Water View 7.5 min quadrangle, in Cumberland County, south-central Kentucky.

At this stop, we have moved out of the most distal parts of the Fort Payne basin and onto lower parts of the Borden delta front dipping from the northwest to the southeast (Figs. 10, 11). In fact, at this exposure, it is possible to view dipping foreset beds in the Fort Payne silty shales. The section begins like most others in the area with glauconitic Maury Shale overlying the Chattanooga black shale (Fig. 40). A probable Rockford Limestone overlies the Maury, which is in turn overlain by ~14 m (46 ft) of silty Fort Payne shales (Fig. 40). The succeeding Floyds Knob Bed is ~1.5 m (4.9 ft) thick and is an amalgamation of four separate glauconite beds, which may be the same four beds seen at Stops 6 and 7. What is unusual about this stop is that in only 7–8 km (4–5 mi), the Floyds Knob Bed has changed from a discreet bed, as we see here, to a thick zone of green, glauconitic shale with included carbonate mud mounds several tens of meters thick (Figs. 10, 11). At the next two stops, we will examine the much thicker Floyds Knob zone with its carbonate mud mounds encased in green glauconitic shales.

Stop 9: Toe-of-Slope, Carbonate Mud Mounds and the Floyds Knob Zone in the Distal Fort Payne Basin

Devi B.P. Udgata, Frank R. Ettensohn, R. Thomas Lierman, and Charles E. Mason

Stop 9 is located on Kentucky State Highway 61 north of Burkesville, Kentucky (Fig. 12). Latitude: 36° 54′ 0.6″ N; longitude: 85° 25′ 46.74″ W. This section is located on the Water View 7.5 min quadrangle, in Cumberland County, south-central Kentucky.

Measured sections of this and the next stop can be found in Meyer et al. (1995), Krause et al. (2002), and Krause and Meyer (2004). Interestingly, where carbonate mud mounds occur in the Fort Payne, a single discreet Floyds Knob Bed, like that at Stop 8, is never found. However, green glauconitic shales are found below the mounds as mud-mound cores, and a much thicker interval of

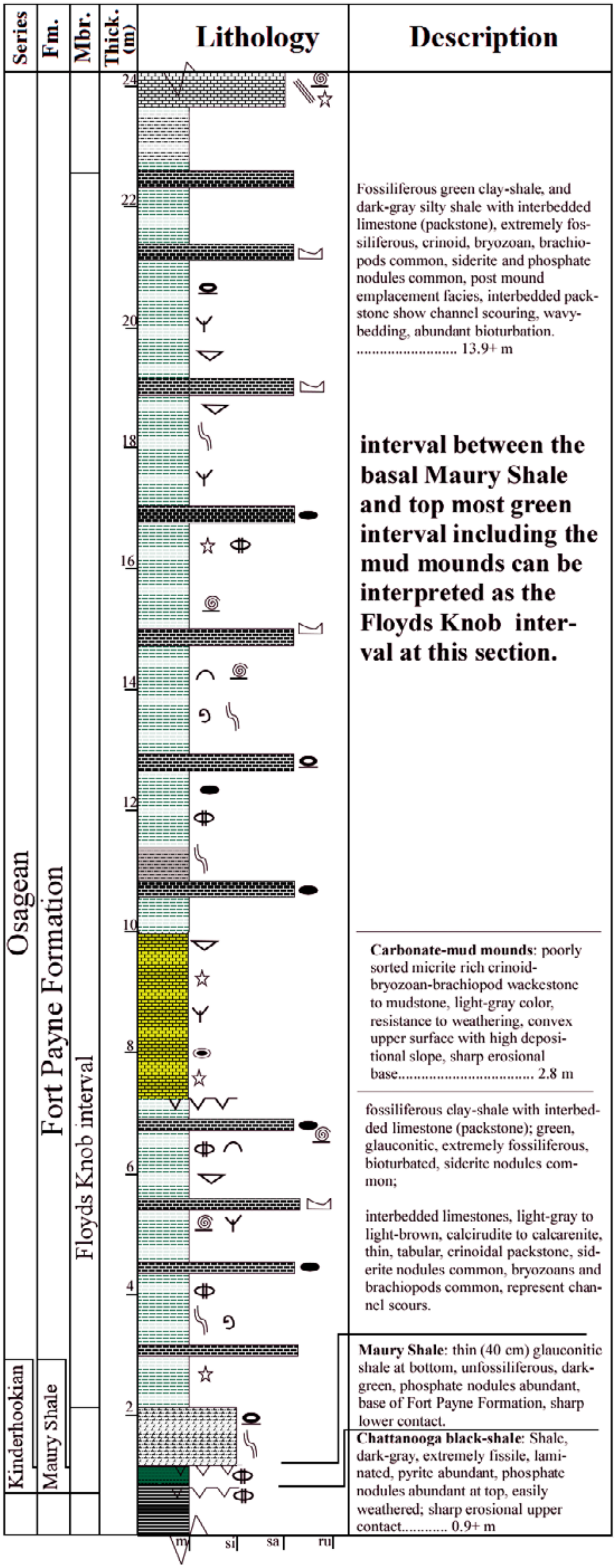

Figure 37. Columnar section from Optional Stop B on Manntown Road. No distinct Floyds Knob Bed exists here. Instead, the Floyds Knob is represented by a zone of green glauconitic shale, ~20 m thick, which contains a small carbonate mud mound near its base. Legend as in Figure 35 (from Udgata, 2011).

Figure 38. Small, convex-upward carbonate mud mound in Floyds Knob zone at Optional Stop B on Manntown Road.

Figure 39. Silicified crinoid debris and clotted micritic texture with *Stromatactis*-like infilled voids from the top of the small carbonate mud mound at Optional Stop B.

green, glauconitic shale sometimes encases the mud mounds and the flank beds derived from them. This is another example of a thick Floyds Knob zone, but unlike what we saw earlier today in the Halls Gap and Muldraugh members, and yesterday in the more proximal Nada Member of the delta complex, this Floyds Knob zone occurs in a distal, toe-of-slope setting (Fig. 10). Here the zone is composed of more than 7 m of green, glauconitic shales that encloses an asymmetrical carbonate mud mound (Fig. 41). The mound laps up onto what appears to be an eroded paleotopographic high with an angular discontinuity (Fig. 42). Initial deposits are parallel to slope but thicken into the topographic low northwest of the high. Subsequent deposits, however, appear to lap onto the high and accrete more or less horizontally or have a mound-like geometry (Fig. 42). Like the mound at Optional Stop B, the mound has a clotted texture with infilled, *Stromatactis*-like voids. Fossil diversity seems to be high, but bryozoan and crinoidal debris and rugose corals are most abundant (Fig. 39).

Most of the more recent research has suggested that the carbonate-mud mounds present in the lower part of the Fort Payne Formation in south-central Kentucky and north-central Tennessee were deposited in relatively deeper water, but within the photic zone on a southward facing paleoslope in front of the abandoned Borden Delta (MacQuown and Perkins, 1982; Ausich and Meyer, 1990; Meyer et al., 1995; Khetani and Read, 2002; Krause et al., 2002; Krause and Meyer, 2004; Greb et al., 2008). The thin-bedded, flat-lying, green fossiliferous shales above the Maury Shale act as the background sediment in which the mud mounds developed, and are interpreted to have been produced by turbidity currents, slumps, and slides on the delta paleoslope (Lewis and Potter, 1978; Ausich and Meyer, 1990; Meyer et al., 1995; Krause et al., 2002; Krause and Meyer, 2004). These mud mounds most likely developed on paleotopographic highs on the basin-floor (Lumsden, 1988; Ausich and Meyer, 1990; Stapor and Knox, 1995).

Some of the paleotopographic highs on which the Fort Payne mud mounds and associated green shales accumulated may also have developed over deep, structurally uplifted areas (Fig. 7). Mounds accumulated in shallower, warm waters within the photic zone, as evident by the occurrence of coral communities (Fig. 43). Jeffery (1997) has suggested that carbonate mud mounds are relatively common atop structures that can elevate the mounds into shallower, sunny, more active waters. The presence of rugose and tabulate corals (Fig. 43), along with an abundance of echinoderm debris, in the Fort Payne mud mounds indicates the presence of active waters that kept colonies clean and supplied nutrient- and oxygen-rich waters. In addition, the presence of rugose and tabulate corals, which are assumed to have contained algal symbionts like their modern ancestors, calls for well-lit water less than 100 m deep. This is also suggested by the presence of clotted carbonate muds (Fig. 39), which have been attributed to algal mats and other forms of algae apparently living on the mound surfaces (Krause and Meyer, 2004). The distal location of these mounds, as well as their elevated nature, would have insured a relatively clastic-free setting.

The mound area overlies basement faults of the Precambrian East Continent Rift Basin and Grenville Front (Fig. 7). These structures may have been reactivated to elevate overlying areas like that seen at this stop. Migrating Neoacadian or Ouachita bulges may also have caused their reactivation and uplift, and synsedimentary growth structures and angular unconformities associated with the mud-mound section (Figs. 41, 42) may well reflect the reactivation of these structures at depth. These growth structures could also be related to slope processes and sediment compaction. Periodic storms apparently kept the mound tops clean and well-exposed in well-lit, active waters. However, the same storms would have reworked nearby delta-front and prodelta muds, not to mention nearby glauconitic muds. These muds would have accumulated in quiet, low places between the

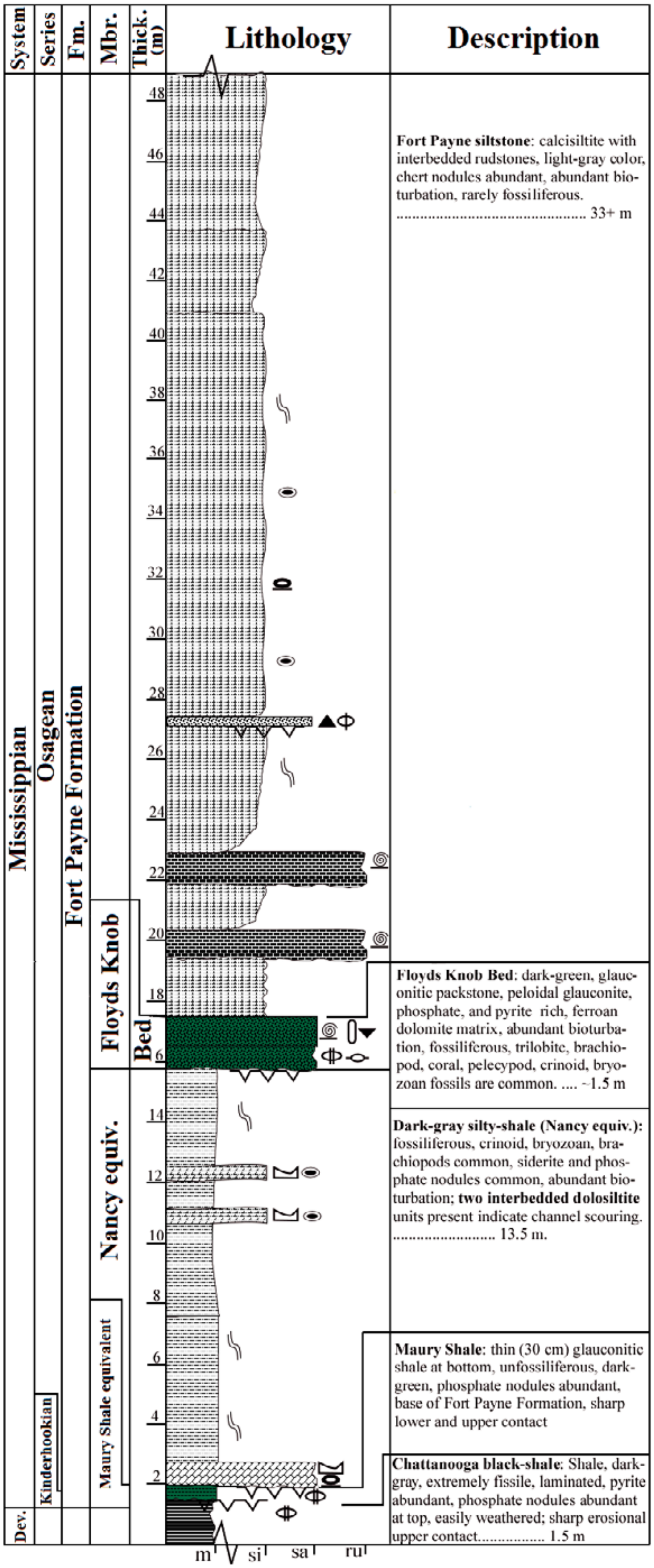

Figure 40. Columnar section of the Fort Payne Formation at Stop 8. Legend as in Figure 35 (from Udgata, 2011).

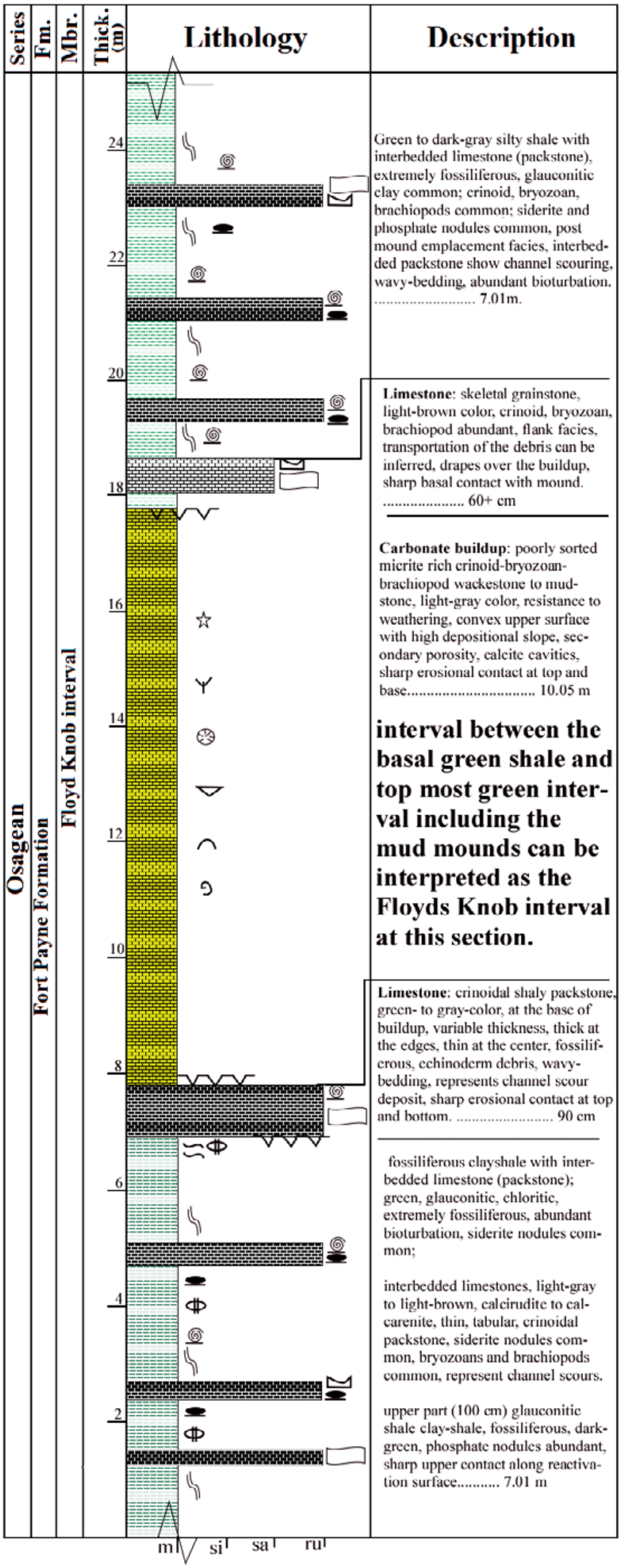

Figure 41. Partial columnar section from the carbonate mud mound at Stop 9. Legend as in Figure 35 (from Udgata, 2011).

Figure 42. Base of mud mound in angular discontinuity with underlying green shales at Stop 9; see Figure 41 for section.

mounds, and with rising sea level, would have eventually buried the mounds. While the mounds were active, gravity- and storm-reworking of skeletal debris from the mound tops would have generated the many grainstone, packstone, and floatstone layers found in the surrounding green shales. We will observe such layers next at Stop 10.

Stop 10: Mound Flank Beds from the Floyds Knob Zone

Devi B.P. Udgata, Frank R. Ettensohn, R. Thomas Lierman, and Charles E. Mason

Stop 10 is located on Kentucky State Highway 61 north of Burkesville, Kentucky (Fig. 12). Latitude: 36° 53′ 47.82″ N; longitude: 85° 25′ 41.88″ W. This section is located on the Water View 7.5 min quadrangle, in Cumberland County, south-central Kentucky.

Figure 43. Solitary rugose corals from the carbonate mud mound at Stop 9.

Stop 10 demonstrates many thin- to medium-bedded crinoidal grainstone, packstone, and floatstone layers interbedded with green, glauconitic shales (Fig. 44). Many of these layers have sharp, erosional or channel-like basal contacts and are gradational upward into the overlying green shales (Fig. 45). In some of these layers, the crinoidal debris is randomly dispersed throughout, whereas in others, the crinoidal debris is crudely graded. These layers have been interpreted to represent flank beds of crinoidal debris that was washed off of adjacent carbonate mud mounds into intervening low areas of green, glauconitic mud. These beds, along with the carbonate mud mounds, form a distinct Floyds Knob lithofacies (Fig. 11) in the more distal Floyds Knob zone (Fig. 10).

The exposure also shows several faults that appear to fade away up-section, suggesting synsedimentary growth. In other places, crinoidal layers appear to thicken toward the faults, suggesting movement during deposition. The main fault in Figure 45 appears to be a rotational glide plane of a small growth fault. The displaced section thickens downdip and the glide plane appears to flatten. Although these small faults could reflect reactivation of basement structures, they are more likely related to slump of carbonates over a shale mound core.

Stop 11: The Sunbury and Grainger Formations at Pound Gap, Kentucky

Frank R. Ettensohn, Charles E. Mason, Devi B.P. Udgata, and R. Thomas Lierman

Stop 11 is located along U.S. 23 east of Whitesburg, Kentucky, as the highway crosses the Kentucky-Virginia boundary (Fig. 12). This section is located on the Jenkins West 7.5 min quadrangle, in Letcher County, southeast Kentucky. Our purpose here is to examine the complete Neoacadian tectophase cycle, and in particular, to examine the Borden-equivalent Grainger Formation (Fig. 8) in a more proximal setting to the previous day's stops (Fig. 6). Our stop begins at the base of the exposure (latitude 37° 9′ 24.0″ N; longitude 82° 38′ 47.52″ W) and ends at the Grainger-Newman contact (latitude 37° 9′25.39″ N; longitude 82° 38′ 36.93″ W). The Pound Gap section in eastern Kentucky exposes a nearly complete, 600-m-thick section of Upper Devonian to Lower Pennsylvanian strata along the leading edge of the Pine Mountain thrust fault (Chesnut et al., 1998). The fault brings to the surface rocks that are more than 600 m in depth in front of the thrust, and because the fault has moved rocks in the section ~11 miles to the west, this exposure provides a sample of the rocks that occur farther east in the Appalachian Basin. The rocks are described and interpreted in Chesnut et al. (1998), but the section is also important because the rocks it exposes record a complete flexural tectophase cycle (Neoacadian) and the initiation of another (Alleghanian), as illustrated in Figure 6 (Ettensohn et al., 2002). There are 13 such cycles in the Appalachian Basin from five different orogenies (Taconian, Salinic, Acadian, Neoacadian, and Alleghanian) (Ettensohn, 2008), and it is rare that we can see such a complete

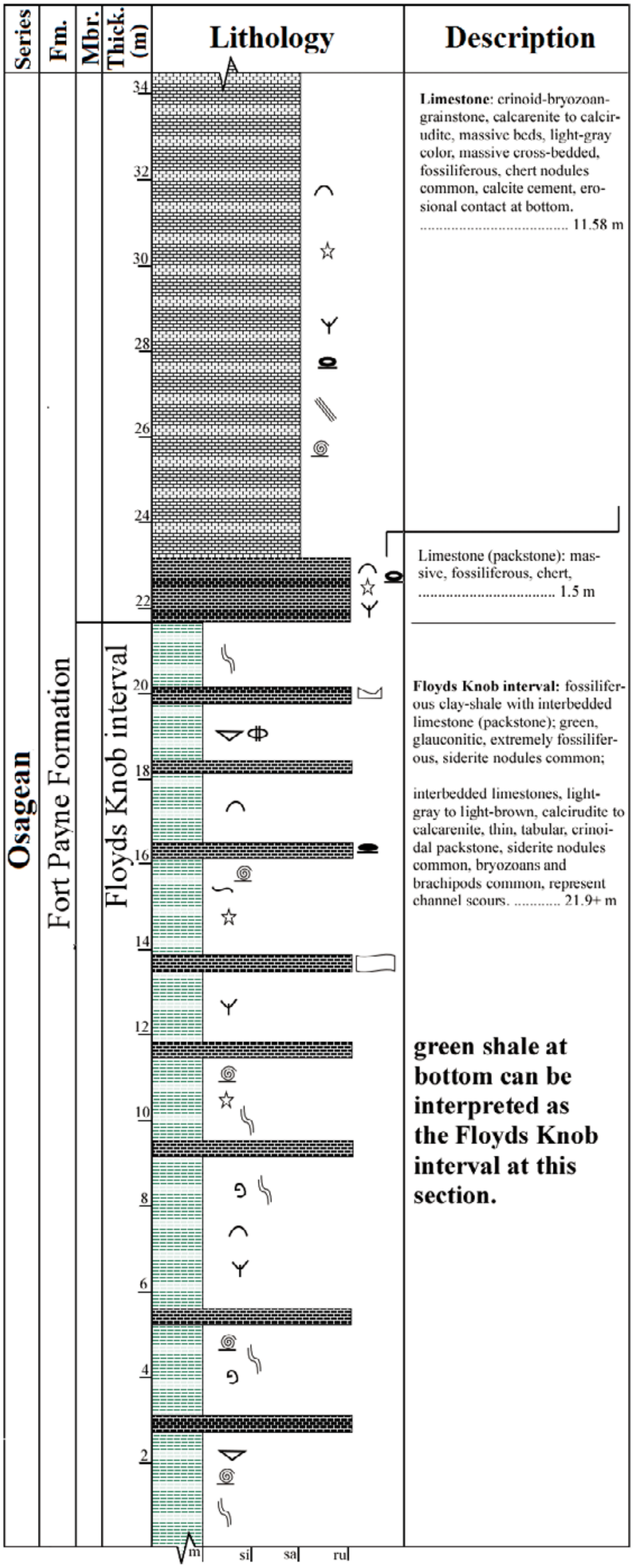

Figure 44. Partial columnar section from an intermound area accumulating crinoidal flank-bed debris at Stop 10.

Figure 45. Crinoidal grainstone, packstone, and floatstone layers in an inter-mound flank-bed area at Stop 10. An apparent growth fault cuts the sequence with beds that thicken along it.

cycle in any one place. The Neoacadian tectonic cycle reflects the reorganization of peri-Gondwanan microcontinents caught in a pincer movement between Gondwana and Laurussia (Figs. 3, 4), and in this particular case, it was the 41-million-year-long dextral transpressional collision of Carolina with the margin of Laurussia (Fig. 4) that generated the complete cycle of sediments seen in this exposure (Ettensohn, 2008). At this stop, the implications of the rock succession to the Neoacadian tectonic cycle are discussed.

Sunbury Shale. The Neoacadian tectonic cycle began in latest Devonian–Early Mississippian time with major collision in the area of the New York promontory (Fig. 4) and the resulting deformational loading generated so much subsidence in the absence of major sediment sources that the deep-water Sunbury Shale and its equivalents were deposited across most of the Appalachian Basin (Fig. 6A). Here, near the center of the Appalachian Basin, nearly 33 m of Sunbury were deposited in a largely anoxic setting, but in the parts of south-central Kentucky that we visited yesterday (Stops 8–10), the Sunbury is represented by only a few centimeters of phosphatic nodules (Falling Run Bed; Fig. 8) because of distance from eastern sources.

We will begin our examination of the Sunbury and Grainger near the very bottom of the exposure on the south side. Here, the Bedford-Berea is exposed, but the unit is highly deformed by a small thrust sliver that runs through the area. The contact between the Bedford-Berea and Sunbury is in this deformed interval and is defined by a thin horizon of phosphate nodules that probably reflects a subtle unconformity and maximum flooding with rapid subsidence.

The Lower Mississippian Sunbury Shale is the stratigraphically highest black shale in the Devonian-Mississippian black-shale sequence. It is composed of very dark gray (N3) or "black," silty, fissile shale that is called the "Coffee Shale" in drillers' terminology; near the top of the unit, centimeter-scale, laminated siltstone beds and lenses become common. It is the most intensely radioactive of Kentucky black shales and is characterized on gamma-ray logs and scintillometer profiles by two spikes of positive gamma-ray deviation separated by a thicker interval of negative deviation (Ettensohn et al., 1988a). At the contact with the Bedford-Berea, the Sunbury is a more massive, black, silty mudstone with phosphorite nodules within a few feet of the contact. These shales sit sharply on top of a weathered Berea siltstone.

The contact with the overlying Grainger Formation is gradational, and although the contact is easily picked on gamma-ray logs and scintillometer profiles, it is not as easily picked on the outcrop. However, in general, the Grainger is more olive gray to olive black in color and is a more "crumbly-weathering," semi-fissile mudstone compared to the "black," more massive-weathering fissile shales of the Sunbury. Thin siltstone layers and lenses also increase in abundance within the basal Grainger Formation. Again, the Sunbury represents deposition in an anoxic, deep basinal setting (Fig. 6A).

Grainger Formation. Once the deformational load (thrust and fold mountains) became subaerial and drainage nets had developed in the Neoacadian mountains, sediment sources became more proximal and vast amounts of coarse clastic debris were transported into the deep Sunbury basin (Fig. 6B), and that material is represented by Borden-Grainger-Price-Pocono delta complex (Fig. 1). For the last two days, we have examined parts of the Borden delta complex and the underfilled Fort Payne basin into which it prograded from the northeast (Fig, 1). The Grainger Formation that we will examine here is the exact temporal equivalent of the Borden Formation (Fig. 8) that we have examined for the past two days, except that it represents a more southerly delta complex that prograded into the basin from the east (Fig. 1). What we are seeing here on Pine Mountain is also a more distal, subaqueous, deltaic equivalent than the northern Borden deltaic complex observed on the first two days.

Overall, we interpret the depositional setting for the Grainger as a shallowing, distal-delta, including muddy basin, turbiditic, tempestitic, and shallow-water infilling of a broad basinal seaway that formed in Early Mississippian time (Fig. 6). The lower 30 m of the Grainger is dominated by dark shale with thin siltstone beds (<3 cm thick) and represents a quiet, relatively deep-water, muddy setting. The middle 50 m consists of light-gray and orange siltstones (3–24 cm thick) interbedded with dark-gray shales, representing basin filling by turbidites. The upper 12 m of Grainger contains red, yellow and purple siltstones with silty shale interbeds and numerous ironstone layers and concretions. Many of the siltstones exhibit hummocky cross-bedding (Fig. 46), and locally, this part of the section exhibits mudcracks and salt casts. The uppermost meter, moreover, contains green, bioturbated, interbedded siltstones and shales that have been interpreted to represent the Floyds Knob Bed (Fig. 46) (Udgata, 2011). These deposits have many features

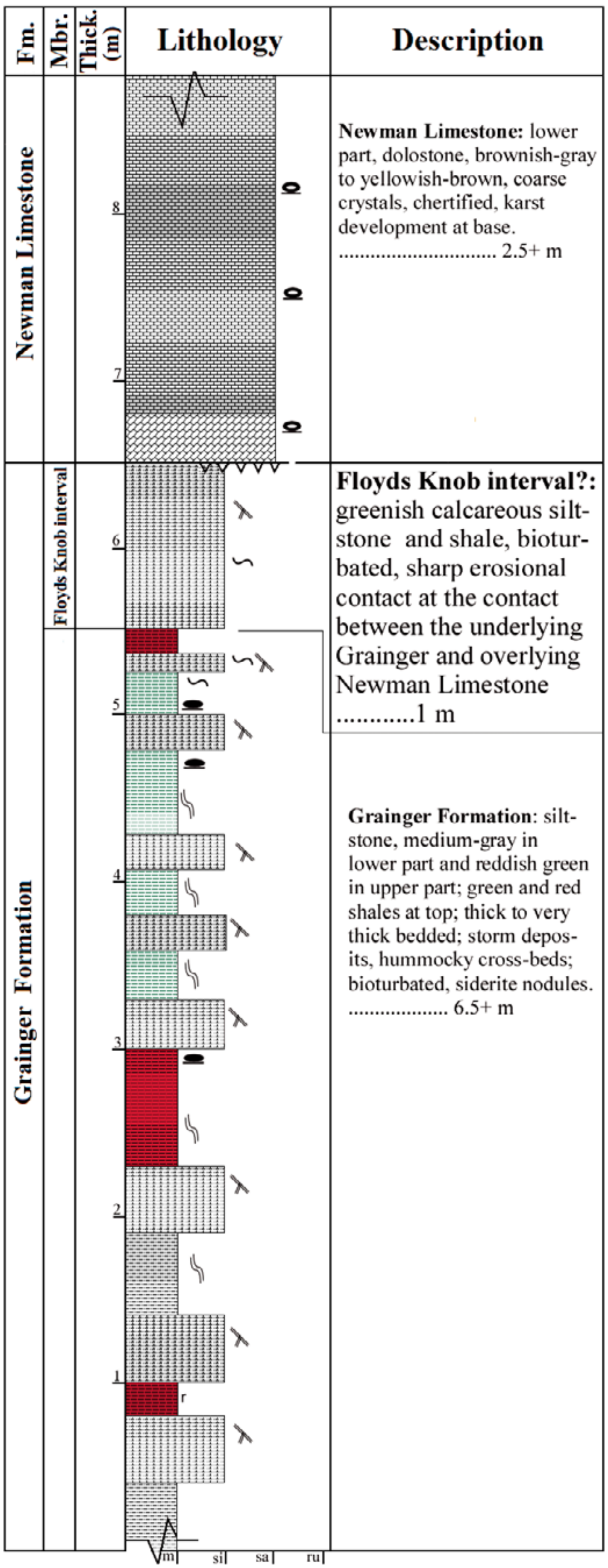

Figure 46. Partial columnar section for the upper Grainger Formation and lower Newman Limestone at Stop 11, Pound Gap. Legend as in Figure 35 (from Udgata, 2011).

indicative of shallow-water deposition and suggest a shallowing-upward progression from turbidite fans, to a storm shelf, and finally to a sediment-starved peritidal setting. Sedimentary structures and trace fossils in these units, along with a diverse marine fauna found in the lower and middle Grainger, also support an interpretation of a shallowing-upward succession of marine deposits. Similarly, the Kinderhookian biostratigraphic control found here supports an east-to-west progradation of a silty depositional system (Grainger; Fig. 1), gradually infilling a preexisting deeper water basin (Sunbury Shale).

Clearly, these rocks reflect the increasing proximality of sediment sources and the infilling of the Sunbury basin. However, the transition from storm-shelf environments to redbeds in a peritidal setting was almost too rapid and abrupt, unless other events were preeminent. We suggest that those events include a sea-level lowstand along with the typical eastward bulge migration that occurs in this part of a tectophase (Fig. 6C). In fact, Wilpolt and Marden (1949, 1959) have correlated these redbeds in the upper Grainger with the red peritidal shales, sandstones and evaporites of the Maccrady Formation to the east. If this correlation is correct and the Maccrady represents sedimentation on the bulge itself (Fig. 6C), then the uppermost Grainger redbeds probably reflect sedimentation on the margin of the bulge and the overlying Floyds Knob Bed indicates the inception of sediment starvation throughout the area. This is an event, moreover, that is correlatable across the east-central United States, both in starved basins and in actively filled basins, wherever the Floyds Knob Bed/zone is present.

For the rest of the stop, we will briefly drive through the overlying parts of the Pound Gap exposure, looking at the thick carbonates of the Newman Limestone, which reflect shallow, equilibrium conditions in a filled basin (Fig. 6D), and the marginal-marine clastics of the overlying Pennington Group, which represent unloading-type uplift in the orogen (Fig. 6E). The overlying orthoquartzitic sandstones of the Lee Group reflect fluvial sedimentation accompanying the inception of the Alleghanian orogeny (Fig. 6F).

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