

Eastern Section AAPG  
48<sup>th</sup> Annual Meeting  
October 12<sup>th</sup>, 2019



## Field Trip Guidebook



**Upper Devonian paleoenvironmental, diagenetic, and tectonic enigmas in the western Appalachian Basin: new discoveries and emerging questions associated with the Frasnian-Famennian boundary and end-Devonian disturbances in central Ohio**

**Field Trip Leaders: David (Randy) Blood and Gordon C. Baird**

### **A note from the Authors:**

The following is a post-meeting revision to the “Upper Devonian paleoenvironmental, diagenetic, and tectonic enigmas in the western Appalachian Basin: new discoveries and emerging questions associated with the Frasnian-Famennian boundary and end-Devonian disturbances in central Ohio” guidebook supplied to the 2019 Eastern Section Annual Meeting of the American Association of Petroleum Geologists fieldtrip attendees. We hope that this revised edition supplies the reader with a more complete understanding of end Devonian events in Ohio and an opportunity to explore these key outcrops at your leisure. In using this guidebook we ask that you respect the various landowners by obtaining permission to access those exposures on private property, and the necessary permits to access and collect within public parks.

### *Cover Photos:*

Upper Left: Contact of the Upper Olentangy Shale with the Huron Shale bearing large septarian carbonate concretions, Highbanks Metropark, Worthington, OH; Upper Right: Red Bedford Shale exposed at Blendon Woods Metropark, New Albany, OH; Bottom: Massive ball-and-pillow seismite deformation in the Berea Sandstone, Sunbury, OH.

***Upper Devonian paleoenvironmental, diagenetic, and tectonic enigmas in the western Appalachian Basin: new discoveries and emerging questions associated with the Frasnian-Famennian boundary and end-Devonian disturbances in central Ohio***

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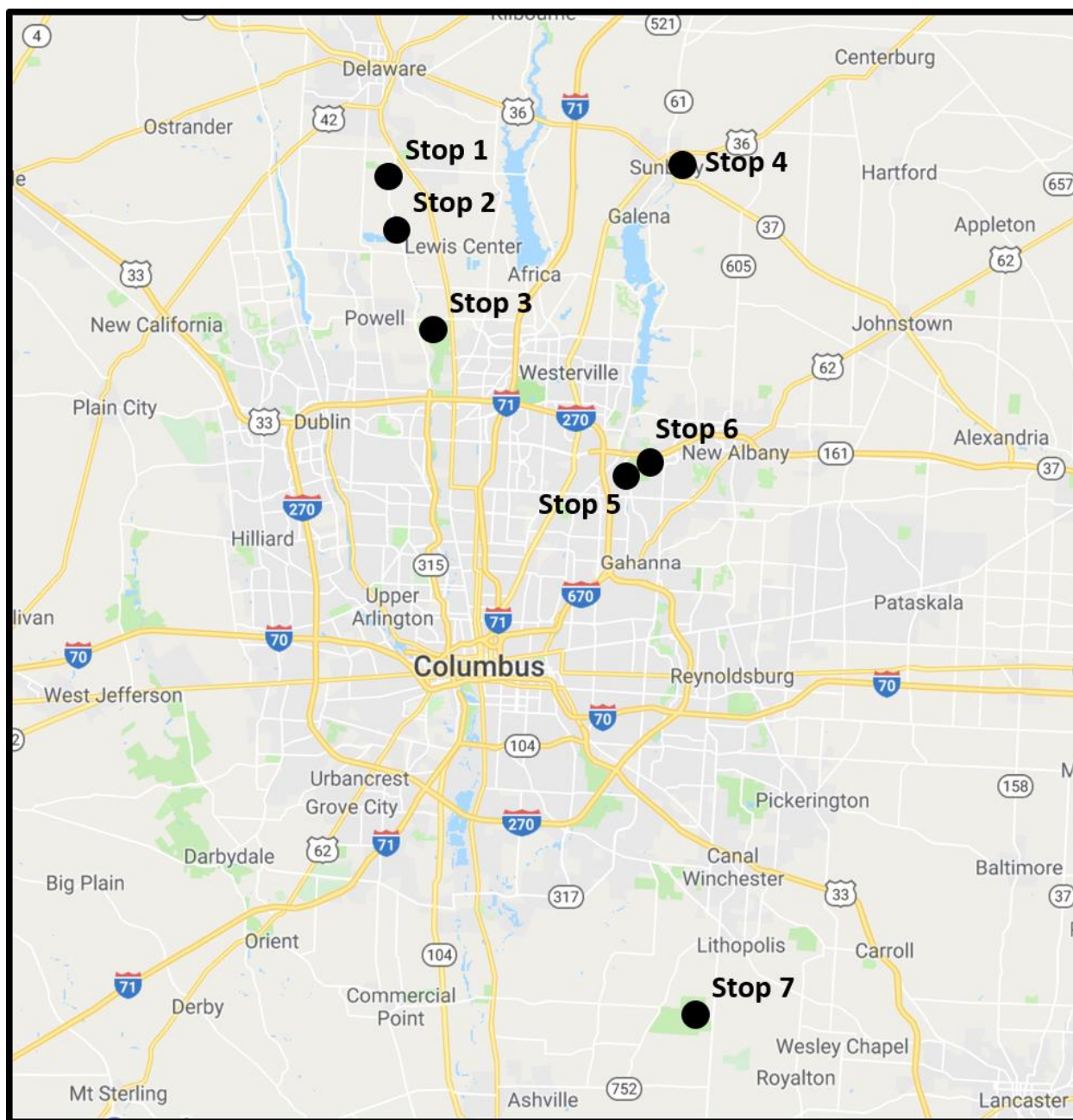
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## INTRODUCTION

Devonian sections in central and southern Ohio present numerous surprises as well as research and exploration opportunities hiding in plain sight. Although Columbus is on the “stable” cratonward side of the Appalachian Basin, the features you will observe on this field trip will show that the Middle and Late Devonian narrative in this region was both sedimentologically and tectonically dynamic in ways that remain only partly explained. Given that we are now in an age of unconventional exploration, this excursion will have a strong focus on shale, with the mission of introducing those persons largely familiar with subsurface-based understandings of shale to the larger spatial reality of these units in outcrop. It will also introduce attendees to certain units and events that remain enigmatic. In particular, we suggest that the region we call Ohio today experienced major episodes of extinction and far-field seismic disruptions during the latest Devonian.

Central Ohio is particularly notable for the record of pronounced basin subsidence, coupled with episodes of intense and extended substrate dysoxia, particularly in the Late Devonian and earliest Mississippian. These events are expressed by the unconformity-floored black shale successions of the Famennian Huron and Cleveland members and the lower Tournaisian Sunbury submember of the Orangeville Member within the Cuyahoga Formation. On this field trip we will, first, examine features of variably oxygen-deficient, basinal deposits within the Delaware Formation, the two-part Olentangy Shale, and the Huron Member of the Ohio Shale Formation (Figure 1). One aspect of this section will be to review the resurfacing controversy regarding competing “shallow-” versus “deep-water” models for the origin of Devonian black shale deposits in the Appalachian Basin (see Smith *et al.* 2019) and to make our case that they, most likely, represent end-member, downslope deposits that accumulated slowly within a persistently low-energy, stratified basin setting.

We, herein, examine both non-black and black layers within the sparsely fossiliferous Upper Olentangy Shale (STOPS 1 and 2; Figure 1) and discuss causal mechanisms for the origin of numerous, thin, pelagic carbonate layers and for millimeter-to-decimeter-scale black shale bands within the Upper Devonian (Frasnian-age) barren gray shale successions. Examining the higher Huron and Cleveland shale divisions, we discuss distinctive features of the widespread submarine discontinuities flooring them and present a model for their origin. In particular, we also argue that certain black shale transgressive fills above these widespread contacts, as are observed above the Cleveland Shale Member, record regional depositional onlap effects in northern Ohio that are difficult to detect using most conventional subsurface methods. In both Shale Hollow Park and the Highbanks Metropark (STOPS 2 and 3; Figure 1), we will examine the origin and timing of the formation of joint fracture systems and giant septarial concretions to

early diagenetic events, followed by deeper-burial maturation into the oil window within a far-field tectonic context.

The Late Devonian was marked by two mass-extinction events ranking among the greatest mass-kill events in the Phanerozoic record, as well as, a general pattern of overall decreasing marine biotic diversity through the interval (Sallan & Coates 2010; McGhee 2013; Kaiser *et al.* 2016; Becker *et al.* 2016). The earlier extinction is the globally well documented, two-part biocrisis event corresponding respectively, to a lower extinction marker layer (“lower Kellwasser bed”), within the Upper Frasnian interval, and an upper biocrisis level (“upper Kellwasser bed”) at the Frasnian-Famennian zonal boundary (Schindler 1993). Although neither bed has yet been identified in Ohio sections, both layers are present in New York State where sections are more complete (Baird & Lash 1990; Over 1997, 2002; Bush *et al.* 2017). It is possible that the Lower Kellwasser bed may be represented by one or more of the thin black shale bands observed in the Upper Olentangy interval but this remains to be tested. The hiatus represented at the basal contact of the Huron Member is believed to roughly correspond to the temporal position of these events. Even if one or both of these layers were present in the Columbus area, they would have been expressed as black, basinal shale layers, providing minimal perspective in viewing macrofaunal demise associated with these events.

However, the end-Devonian (Late Famennian Stage) Hangenberg mass-extinction interval, represented locally by the top-Cleveland Shale Member interval-through-Bedford and Berea formation succession, offers a multifaceted view of unusual facies (see STOPS 5 and 6), linked to this global cluster of events. The thick Bedford Formation is characterized by a near-absence of macrofauna and by the dominance of a thick, enigmatic interval of soft, variably deformed, microsheared red-brown mudstone, which has been interpreted either as nonmarine facies or as an offshore sediment accumulation (Pepper *et al.* 1954; Pashin & Ettensohn 1995). In the Columbus area, the Bedford is followed by the Berea Formation, an unfossiliferous succession of fine-grained sandstone and siltstone beds, which is, in turn, succeeded by transgressive, black, organic-rich shale deposits of the Lower Mississippian Sunbury submember. The lower Berea interval is well displayed in Sunbury, Ohio (STOP 4), where the basal Berea is deformed into giant ball-and-pillow seismites. The Berea – Sunbury contact is splendidly displayed in Slate Run Metropark near Lithopolis, Ohio (STOP 7) where the Berea is anomalously thin and condensed, owing to apparent southward depositional onlap towards a localized region of structural uplift.

Across northern and eastern Ohio, as well as western Pennsylvania, the Berea Formation and its stratigraphic equivalents, are represented by complex successions of medium to fine sandstone and locally pebbly sandstone that fill what appears to be a paleovalley network linked to a major marine lowstand episode. The recent work of Brezinski *et al.* (2008, 2010) makes a

compelling case for a late-Hangenberg, glacial “icehouse” event occurring within the Appalachian Basin region as explaining potential correlation of diamictite deposits in eastern Pennsylvania with sub-Berea downcutting in the western Pennsylvania and eastern Ohio region.

In northern and central Ohio, the lower part of the Berea Formation is spectacularly deformed spatially into diapiric slump features, locally displaying vertical displacements of 200 feet or more (see review of this phenomenon in Pashin & Ettensohn 1995). The abundance and wide distribution of these structures, as well as the pervasive micro-deformation within the underlying red Bedford interval, point to the possibility of major far-field tectonics in the heartland at this time. In any case, these end-Devonian sections appear to record the most turbulent and least understood events in the Devonian Appalachian basin narrative.

It must be noted that the Devonian-Mississippian boundary is currently in a state of major revision (Spalletta *et al.* 2017) with the likely possibility that the base-Tournaisian (base-Mississippian) boundary, now set at the lowest occurrence of the conodont *Siphonodella sulcata*, will be lowered to the first occurrence of a different zonal conodont taxon *Gnathodus kockeli*, in the near-future. This opens a distinct possibility that the Bedford and Berea formations, currently placed in the topmost Devonian, will become part of the greater Mississippian succession, as they were in much of the twentieth century.

## Geologic Setting

During the latest Devonian (Upper Famennian Stage) into the earliest Mississippian, the region that is now central Ohio was situated near the southeastern margin of the “Old Red Continent” (Euramerica) between 30° and 45° south of the paleoequator, based on more recent paleomagnetic estimates (Miller & Kent, 1988; Scotese & McKerrow, 1990; Ver Straeten & Brett 1995; Ettensohn *et al.* 2009). The southeast edge of Euramerica (Laurentia plate margin) experienced a series of tectonic collision events, beginning in the latest Silurian and continuing throughout the Devonian which are collectively referred to as the Acadian Orogeny.

Disturbances, involving transpressive collision of the Avalon and Carolina terranes, came as a series of overthrust pulses (tectophases), starting in the maritime region of eastern Canada and continuing southward into the central-southern Appalachian region. Collisions taking place during the very Late Devonian and Early Mississippian (Neo-Acadian Orogeny *sensu* Ettensohn, 1998; Ettensohn *et al.*, 2009) were particularly centered in what is now the southern Appalachian region.

These collisional tectophases, in turn, produced characteristic sedimentary depophases recording initial thrust-loading of the cratonic margin followed by foreland basin filling as sediments, derived from the erosion of collisional mountain belts, filled in the structural basins (Ettensohn, 1998; Ettensohn *et al.*, 2009). This series of depophase events generated a thick,

detrital wedge long known as the Catskill Delta complex. Sediment, eroded from the rising Acadian mountain complexes was transported by river systems to an epicontinental sea within a flexurally subsiding foreland basin; this coastline shifted ever westward through the Devonian as the delta grew and the basin gradually filled. It was particularly sensitive to changes in sea level, which caused the coastline to shift westward during pulses of delta growth and sea level-fall, and to shift eastward during episodes of sea level-rise. These shifts were unusually pronounced during the latest Devonian.

This was a time predominantly characterized by warm, tropical to sub-tropical climatic conditions (Frakes *et al.*, 1992). As noted by numerous authors, the Middle and Late Devonian was characterized by a series of biotic crises which resulted in widespread extinctions events and contributed to the breakdown of global biotic provinciality. The latest Famennian Hangenberg extinction and succeeding “icehouse Earth” event are increasingly seen as major setbacks to the biosphere (see below). The temporal “window” or time-slice of key end-Devonian crises, based on conodont studies (Zaggar 1975; Over 2002) and palynomorph biostratigraphy (Eames 1974; Molyneaux *et al.* 1984; Coleman & Clayton 1987), extends from the base of the *aculeatus* zone (approximate position of the base of the Cleveland Shale Member in Ohio (and its equivalents in Crawford County, Pennsylvania), into the earliest Mississippian *kockeli* zone represented by the Sunbury Shale submember in central and southern Ohio (Figure 2).

### **Marcellus drowning event in Ohio; a look at the Dublin Member**

Within the northern Appalachian region, shelf carbonate deposits, corresponding to the Middle Devonian Late Eifelian-age *Polygnathus costatus* Zone to the *Tortodus kockelianus* Zone, are respectively represented by the Seneca Member of the Onondaga Limestone in New York, the Dundee Limestone in Ontario, the Rogers City Limestone in Michigan, and the Klondike Member of the Columbus Limestone in central Ohio (DeSantis *et al.* 2007; Brett & DeSantis 2010; DeSantis & Brett 2011; Brett *et al.* 2011; Figures 3, 4). These units are included within a transgressive systems tract of the second major Eifelian-age (Eif-2) T-R sequence cycle *sensu* Ver Straeten (2007). Deposition of this carbonate preceded widespread shelf collapse, of combined tectonic and eustatic causes, prior to the end of the *kockelianus* Zone to produce abrupt and profound transgressive facies changes in most areas. In New York State and elsewhere, this event boundary is a black shale-roofed, maximum flooding surface (MFS), often characterized by corrosion of the underlying limestone with resulting concentration of insoluble fish bone bioclasts, quartz grains, and volcanogenic clasts (Baird & Brett 1986).

In western and west-central New York outcrops, strata of the Marcellus Subgroup, include the lower Union Springs Formation (Bakoven Member succession) and succeeding Oatka Creek

Formation, composed, in ascending order, of the Hurley Member, Cherry Valley Member, East Berne Member, Hallihan Hill bed, and Chittenango Member succession (Ver Straeten 1994; Ver

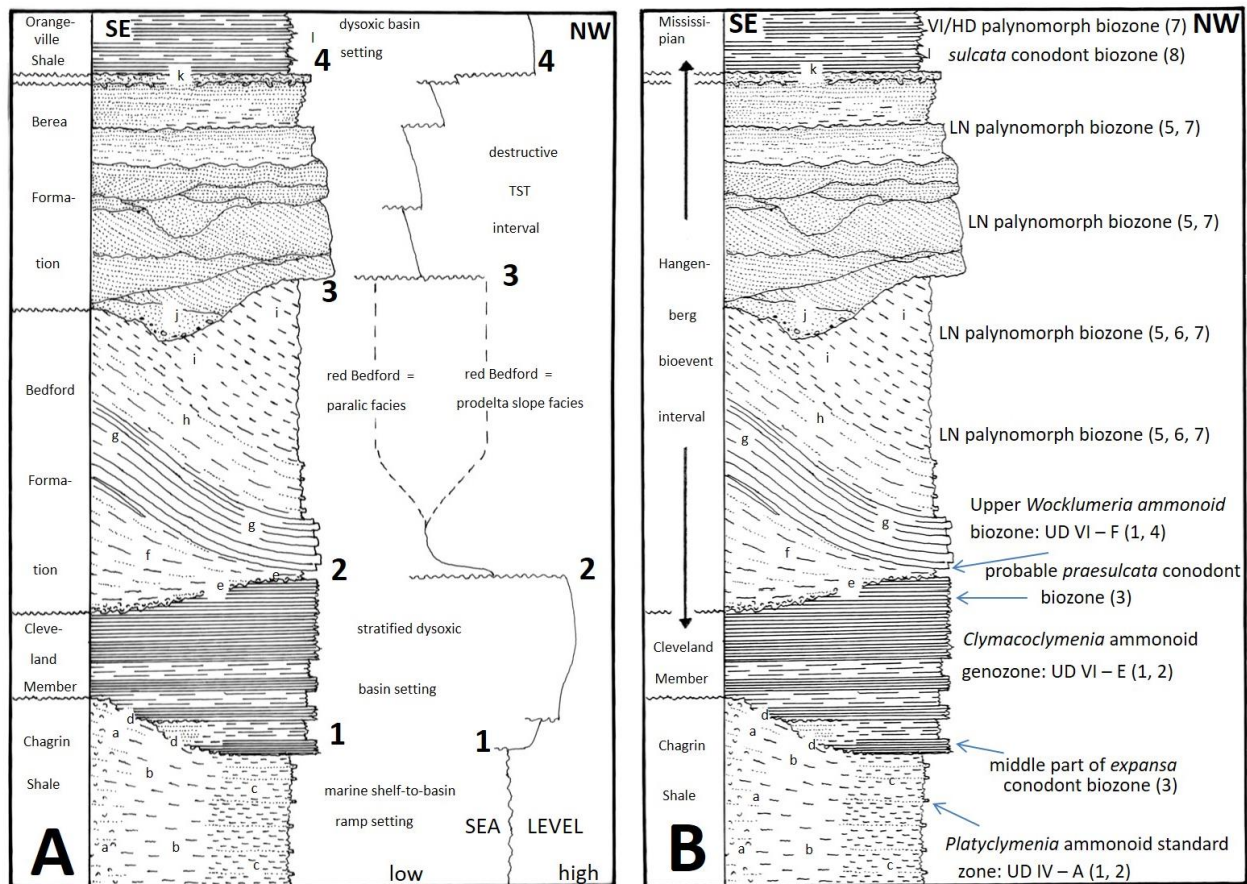


**Figure 2:** Summary of event-stratigraphic relationships for Cleveland area end-Devonian – into – basal Mississippian units: A, Generalized area stratigraphy and sea level history. Bold numbers denote: 1, base-Cleveland member unconformity marking probable position of global Dasberg marine transgression; 2, base-Bedford Formation disconformity; 3, base-Berea disconformity; 4, base-Orangeville submember transgression event. Both the base-Bedford contact (2) and the higher episode of sub-Berea paleovalley incision (3) may be linked, in part, to inferred major glacial draw-down episodes. Dashed line on sea-level curve denotes onshore versus offshore dichotomy as to how the enigmatic red Bedford facies is interpreted. Lettered units include: a-c, eastward-coarsening Chagrin member succession; d, onlapping basal Cleveland member deposits; e, basal Bedford regional disconformity and succeeding variably shell-rich, condensed strata of basal Bedford succession; f, lower Bedford gray shale division; g, Euclid member of Bedford Formation; h, middle Bedford gray shale division; i, variably deformed red-brown mudstone division = red Bedford interval (“RB”); j, regional, base-Berea disconformity; k, thin, unnamed, bioturbated siltstone unit possibly representing the lowest Mississippian division; l, black-to-dark gray-greenish, fissile Orangeville member; B, Summary of published biostratigraphic information for succession shown in Figure 2 A. Small numbers on both figures 2 A and 2 B correspond to authors listed in references; these include: 1, 2, House *et al.* 1986; Becker & House 2000 (ammonoids); 3, Zagger 1975 (conodonts); 4, House & Kirchgasser 2008 (ammonoids); 5-7, Eames 1974; Molyneaux *et al.* 1984; Coleman & Clayton 1987 (palynomorphs); 8, Hass 1947 (conodonts).

Straeten & Brett 2006; Figure 3). In central and northern Ohio, this stratigraphic interval corresponds, in part, to the variably carbonate-dominated Delaware Formation (DeSantis *et al.* 2007; Brett *et al.* 2011). The two key, organic-rich, target zones within the Marcellus interval, in producing areas of the northern Appalachian Basin, are represented by the Bakoven and

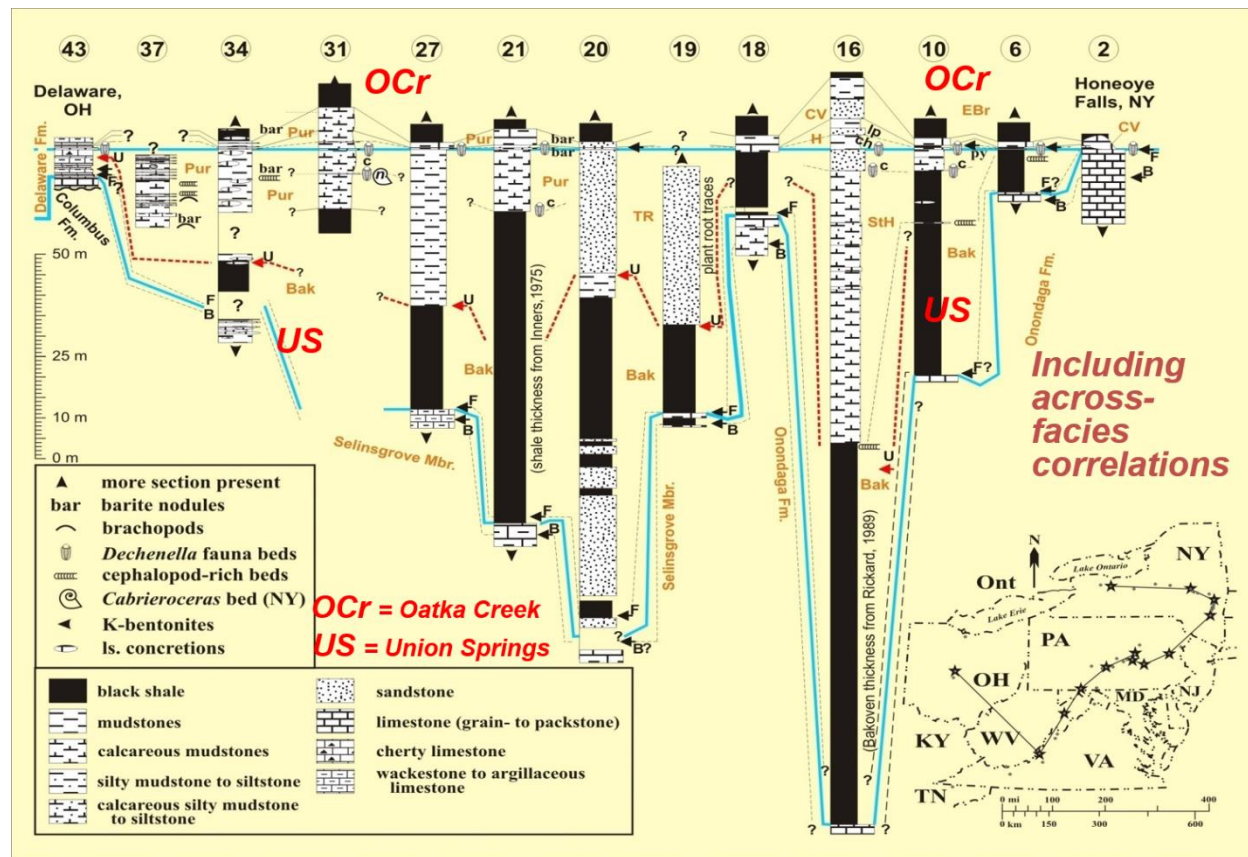


**Figure 3:** Upper Eifelian-basal Givetian regional stratigraphy (adapted from Ver Straeten & Brett, 2006)

Chittenango black shale members (Lash & Engelder, 2011; Brett *et al.* 2011). In the Delaware – Columbus area, these levels are represented, respectively, by the Dublin Member in the lower part of the Delaware Formation and by an interval of undifferentiated strata at the top of that unit (DeSantis *et al.* 2007; Brett *et al.* 2011; Figure 3). We will examine the transitional lower Delaware (Stratford Member of DeSantis *et al.*, 2007) and the dark, shaly Dublin Member facies where it is accessible along Welsh Run at Camp Lazarus (Figure 4; STOP 1). The separation between the Union Springs and Oatka Creek formational intervals is recognizable within the Delaware Formation. A distinctive bed of crinoidal grainstone rich in brachiopods and the small discoidal rugose coral, *Hadrophyllum orbigny* (Milne-Edwards & Haime, 1850) is identified as the basal transgressive lag of the Oatka Creek Formation on the basis of conodonts and magnetic susceptibility patterns (Over *et al.*, 2009; 2019).

In contrast to Bakoven deposits beneath parts of Pennsylvania and West Virginia, where that unit is expressed as fissile black shale with up to 20% T.O.C. content, the Dublin Member displays distinctly cratonward facies, which is variably carbonate-bearing, more extensively bioturbated, and typically grayer in color, but yielding characteristic, diminutive taxa, characteristic of the Bakoven succession in New York (Figure 5). However, at Slate Run in

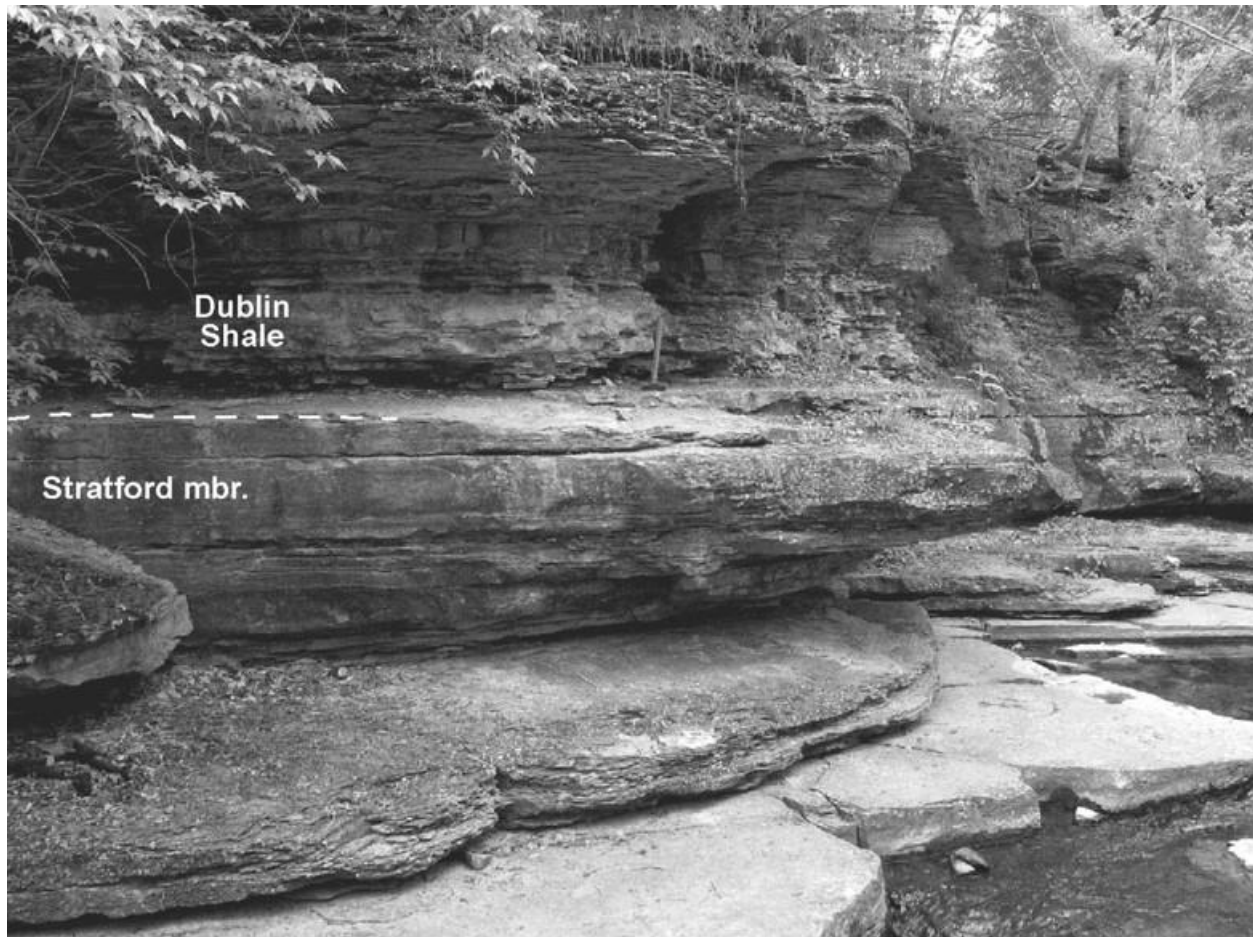


**Figure 4:** Field photograph of the Stratford-Dublin succession of the Delaware Limestone exposed along Welsh Run at Camp Lazarus, Lewis Center, OH; hammer for scale rests on Stratford-Dublin contact. (Photo credit: Carlton Brett)

Upper Arlington, Ohio, bordering the city of Columbus, the Dublin displays a 12 foot- (4-meter)-thick interval of dark gray to black, fissile, calcareous shale, which is interstratified with thin, tabular beds and lenses of black chert (lydite) in association with a low-diversity macrofauna. At this locality, these dark shales overlie a sharp drowning unconformity, capping thicker, fossiliferous limestone layers of the Columbus Limestone. Immediately above this contact is a tephra layer, similar to thin beds in the basal Marcellus succession in New York and elsewhere (Ver Straeten 2004; Ver Straeten *et al.* 2012). As such, the Dublin Member is a westward,

upslope-extending tongue of the basinal lower Marcellus that extends to the eastward, cratonward flank of the Findlay Arch (Figure 3).

Smith *et al.* (2019) have recently rejected the “stagnant” stratified basin model for deposition of organic-rich sediment deposits in favor of a far shallower interpretation of Devonian black shale deposition on the order “several meters to tens of meters” water depth, citing evidence of sediment bioturbation, storm-wave impingement along substrates, and occurrence of

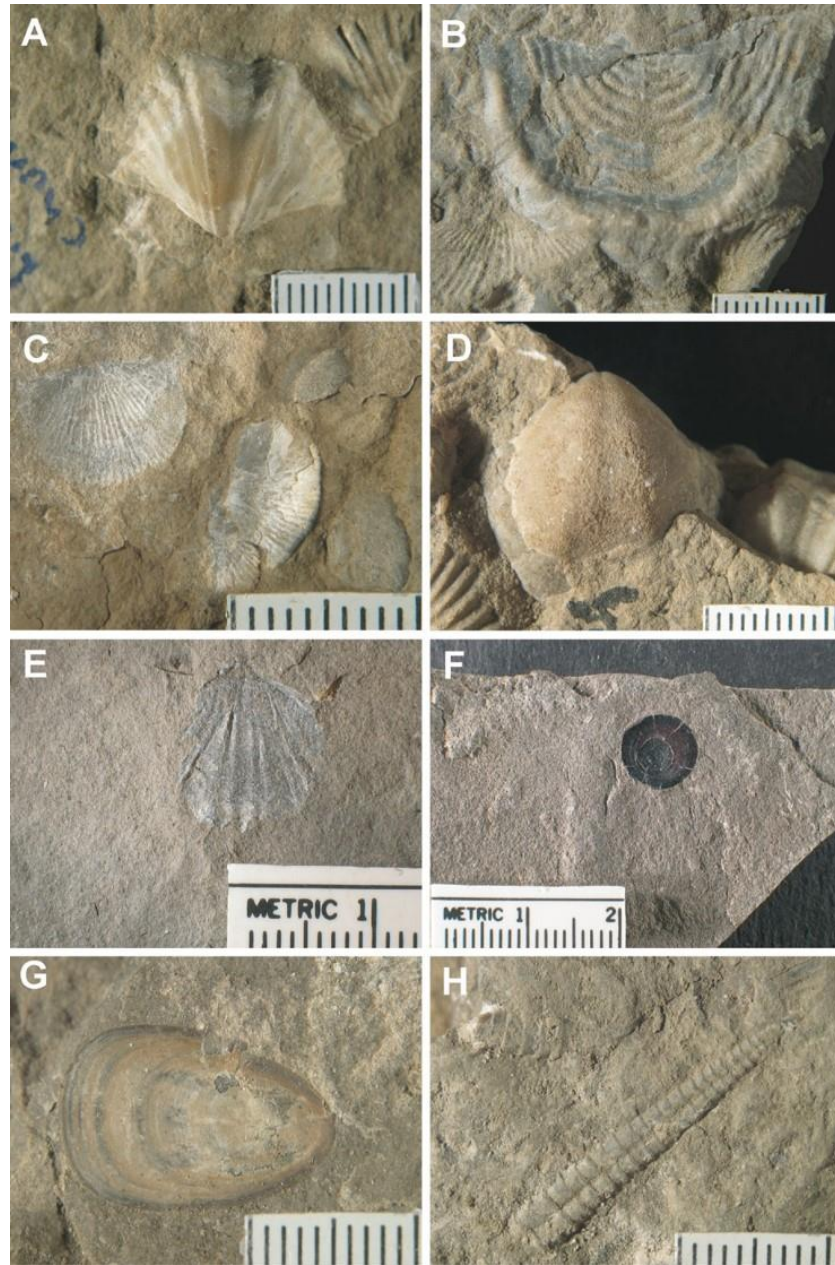


**Figure 5:** Fauna of the Stratford-Dublin succession; (A) *Brevispirifer lucasensis*; (B) *Leptaena cf. rhomboidalis*; (C) chonetids; (D) *Warrenella maia*; (E) *Cherryvalleyrostrum*; (G) *Lingula* sp.; (H) *Tentaculites scalariformis*. From Brett & DeSantis, 2010.

maximally organic-rich deposits over widespread disconformity surfaces. In contrast, the present authors argue that, if the Bakoven sea was of an order of “meters to tens of meters” maximum depth in the flexural basin centers in central Pennsylvania or West Virginia, it would be highly unlikely that the Dublin would have extended cratonward to the Columbus area as a dark shale and more likely that it would have been expressed as lagoonal carbonate or even as an erosional hiatus. We, herein, adhere to the predominant view that these shales are the record of complex processes that took place in deeper marine settings on the order of tens of meters-to-as much as 150 meters-depth for the above argument and for other lines of evidence presented below.

### **Two-part Olentangy Shale narrative and introduction to upper Frasnian basinal gray shale deposits**

At STOPS 1-3, we will examine portions of the “two-part” Olentangy Shale Formation, a shale unit, originally established as a single lithologic entity (Winchell 1874; p. 287-289) that is now recognized to be comprised of two temporally disparate rock units, presently referred to in standing literature as “Lower-” and “Upper Olentangy” respectively. The lower Olentangy segment is now confirmed as being of Middle Devonian (Lower Givetian age; upper *ensensis*-*Pol. timorensis* Zone) age, and the upper segment is understood to be Upper Devonian (late Frasnian age; MN Zone 13) based on conodont yields (Sparling 1995; Over & Rhodes, 2000). Given that early workers failed to account for the possibility of an internal unconformity within this very soft, seemingly uniform shale unit, they repeatedly came up with faunal discrepancies in estimating its age and relationships to other Ohio rock units. Tillman (1970) solved the faunal riddle with the recognition of a mappable disconformity separating the two Olentangy divisions, which we will see at STOP 1. Later conodont sampling at the type Olentangy Shale section (Over & Rhodes, 2000) revealed the full temporal magnitude of this sedimentary break as being on the order of 8-10 million years.

Though subtle, the Lower Olentangy Shale differs from the Upper Olentangy in color and consistency in being of a slightly darker, gray-brown color and more resistant in outcrop (Figure 6). However, as with the Upper Olentangy, it is very sparsely fossil-bearing and is generally known mostly for its ostracode fauna and associated small rugose corals, chonetid brachiopods, and small pyritized molluscs (Tillman 1970). The base of the Lower Olentangy is regionally disconformable, as indicated by its sharp, somewhat pitted contact with the underlying Delaware Formation and by southward erosional overstep of older units by this shale division from northern into central Ohio. Tillman (1970) observed the Upper Olentangy Shale to overstep the Lower Olentangy Shale from Delaware southward to Columbus. At Slate Run in Upper Arlington, the Upper Olentangy Shale is directly juxtaposed onto the Delaware with the Lower Olentangy interval missing. Where the Lower Olentangy Shale is maximally exposed, as

at the Olentangy type section and along the lower reaches of the creek below the Shale Hollow Park, it is marked by bank exposures of gray, fissile shale with several bands of concretionary tabular limestone and laterally non-connected limestone nodules (Figure 6). Irregular pyrite masses and pervasive dispersed, diagenetic tubular pyrite, associated with burrow systems, are common in this unit. At STOP 1 (Welsh Run at Camp Lazarus), the Lower Olentangy Shale is poorly exposed, and we will observe only its base and top at that place.



**Figure 6:** Field photograph of the contact between the Upper and Lower Olentangy Shale exposed along the creek at Shale Hollow Preserve, Lewis Center, OH; hammer at water level for scale. (Photo Credit: Erika Danielsen)

The discontinuity separating the lower and upper Olentangy Shale segments is visually cryptic in sections, but it can be spotted in fresh bank faces where gray lower Olentangy is abruptly succeeded by distinctly lighter colored, Upper Olentangy deposits. This erosional break is not a discrete contact, *per se*, but a 1.5-2 inch (3.5-4.5 cm)-thick zone of abundant phosphatic nodules, fish bones, and conodonts that have been vertically mixed through extended bioturbation activity. Submarine unconformities, such as these, are stratomictic, i.e. diffuse owing to physical mixing of the sediment, both during sediment exposure on the sea bed and

afterward under very shallow sediment cover, following the erosion interval (Baird 1978, 1981). Where similar lithologies, bounding the break, become mixed together, evidence for the unconformity may be effectively “erased”, save for the presence of commingled, temporally disparate zonal fossils, reworked concretions, and exhumed, prefossilized fossil molds (Baird 1978, 1981). Unconformities of this type are common and may be widespread in marine mudstone successions. Moreover, they can be easily overlooked in drill core sections. At STOP 1 we will observe this discontinuity to be marked by numerous small phosphatic nodules in association with fish teeth and dermal plates. Conodonts are abundant in the contact interval and represent a mixed assemblage encompassing at least six international conodont zones, covering an estimated 8-10 million-year hiatus (Over & Rhodes, 2000).

The Upper Olentangy Shale, thickening southward from 15 feet (4.6 m) at the Olentangy type section to 24 feet (7.3 m) near the Highbanks Metropark (STOP 3), is a predominantly light greenish-gray shale unit with subsidiary, discrete bands of brownish-black shale and concretionary limestone at multiple levels (Figure 7 A, B). It is roughly correlative in age and lithology to the Upper Frasnian-age Hanover Formation of the Java Group in western New York sections. As with the Hanover Formation and lithologically similar, but older, Cashaqua and Angola formations in the New York section, this unit yields very few fossils, mostly of small size, partly attributed to dysoxic conditions within down-ramp-to-basin settings. This sparsity of benthos in dominantly gray, bioturbated sediments appears to be a peculiarity of the middle and late Frasnian Stage of the Upper Devonian, particularly, when compared to Middle Devonian (Eifelian and Givetian) units in the northern Appalachian Basin region. In New York, even dark gray shale, recording oxygen-deficient conditions in the Middle Devonian (Givetian) Hamilton Group, typically contain an abundance of brachiopods and molluscs, contrasting with the seemingly “barren” bedding plane exposures of non-black Upper Olentangy layers. This near-absence of fauna within the light-colored and bioturbated light gray-greenish layers, coupled with the associated presence of numerous very knobbly (“popcorn”) carbonate layers at many levels, suggests that the Upper Olentangy Shale is a type of “time-specific facies” (*sensu* Brett *et al.* 2012) recording temporally unique and higher stress bottom conditions in middle-late Frasnian seas. This is understood by many workers to reflect a long-term, step-wise, deterioration of the global biosphere that began with the late Middle Devonian Taghanic biocrisis and concluded with the latest Devonian-into-earliest Mississippian Hangenberg biocrisis events (McGhee 2013).

Alternating with the gray shale and thin carbonate layers are a number of discrete, strongly contrasting, black shale bands of varying thickness (Figure 8 A, B). Typically, these layers display sharp basal contacts on lower gray shale units, but variably diffuse upper contacts, reflecting downward burrow penetration from succeeding gray shale layers (Figure 8 B). One 4-5 inch (10-12.5 cm)-thick, black shale bed in the middle part of the Upper Olentangy succession displays a



**Figure 7:** Field photographs of the Upper Olentangy Shale depicting multiple bands of thin black shale and concretionary carbonate at (A) Shale Hollow Park (photo credit: Erika Danielsen), and (B) Highbanks Metropark (photo credit: Randy Blood).

thin, basal lag concentration of reworked, pyritic burrow tubes in association with fish bone fragments and conodonts. Although, the other, thinner, black shale bands also displayed sharp basal contacts, no lag concentrations were encountered below them. The one lag concentration, noted above, may correspond to a higher erosional “bone bed” level noted by Tillman (1970) in the middle part of the Upper Olentangy interval.

Generally, the Upper Olentangy is understood to record dysoxic offshore conditions as suggested by its fine-grained texture and striking near-absence of shelly, neritic fauna, but not as basinal as that for the succeeding black Huron Member upstream. The thicker black shale bands may be truly linked to parasequence-scale transgressive risings of the pycnocline boundary, but the very thin ones may be a signature of transient upwelling events or even downslope introduction of black muds into the basin via density-flow processes.

The numerous, thin, knobbly, carbonate bands are distinctive to the Upper Olentangy interval. They appear to be a regional expression of “griotte facies” (*sensu* Tucker 1974), which is often

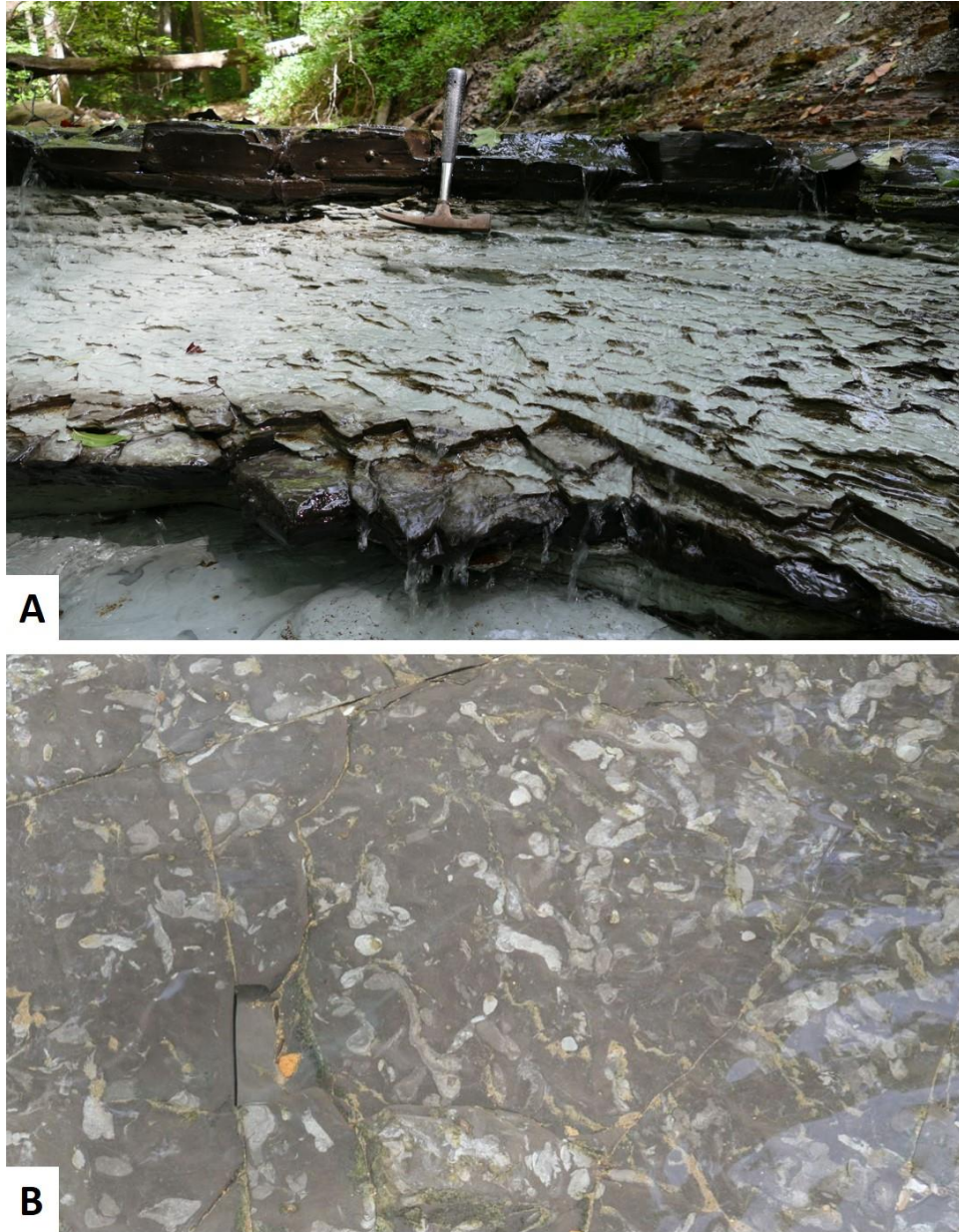


**Figure 8:** Field photographs of (A) discrete black shale bands in the Upper Olentangy Shale exposed along creek at Highbanks Metropark (photo credit: Randy Blood), and (B) bioturbation of black shale with grey shale piped into burrows (photo credit: Carlton Brett)

displayed in offshore, condensed marine deposits of mainly Frasnian age, particularly, in European and North African sections (Wendt & Aigner 1982). Many of these irregular nodules are cored by pyrite associated with tubular burrow networks, and the extreme nodularity appears to be focused along the upper surfaces of many carbonate layers. We suspect that episodes of reduced siliciclastic influx, associated with minor transgression events, created conditions favoring bioturbation and associated fluctuations of the redox boundary and

methanogenic activity to produce these unusual layers. As noted above, this type of carbonate occurrence may reflect unique temporal conditions as it is essentially absent from Middle Devonian marine deposits. Careful brushing of the nodules may reveal a few small fossils such as ostracodes and rare chonetid brachiopods. Similar deposits in the roughly coeval Hanover Formation in western New York have yielded sparse, small rugose corals (Baird, pers. obs.).

### **Huron Member – Cleveland Member stratified basin system:**

#### *Spatial distribution and character of unit*

The Upper Devonian (Famennian) Ohio Shale Formation is composed of two spatially widespread black shale divisions; a lower, Huron Member black shale division, and the higher Cleveland Shale Member, mutually separated by the largely non-black Chagrin Shale Member in northern and eastern Ohio sections (Pashin & Ettensohn 1995; Baird *et al.* 2013). In southern Ohio and in east-central Kentucky, the Chagrin has largely passed basinward into coeval Huron facies and its topmost part is solely represented by a thin gray shale marker tongue, known as the “Three-lick bed”, which marks the base of the Cleveland Member (Kepferle & Roen 1981). Although as thin as 1.5 – 2.5 feet in central Kentucky, the “Three-Lick” interval thickens, northeastward (sourceward) to several tens of feet-thickness in southern Ohio (Kepferle & Roen 1981), before expanding further, eastward and northeastward into the distal Chagrin succession through combined effects of diachronous, up-ramp facies change and increased siliciclastic sediment supply to the growing Catskill prodelta wedge (Pashin & Ettensohn 1995; Ettensohn *et al.* 2009). Undivided deposits of the Chagrin Member reach a maximum thickness of approximately 1200 feet in easternmost Ohio, separating the Cleveland Member or other higher divisions from black shale deposits of the older, lower Huron Member (Cushing *et al.* 1931).

Ettensohn *et al.* (1988) subdivided the Huron into lower, middle, and upper units (C, B, A units, respectively, of Rimmer *et al.*, 2004). The lower Huron consists of alternating greenish-gray silty shales and black shales (1–76 m). Its transition to the middle unit is marked by a zone of concentrated *Protosalvinia* (*Foerstia*) Dawson, 1884 (Figure 9) a problematic alga?-related reproductive structure. The middle (1–135 m) and upper (< 1–40 m) units are characterized as “ribbed” black shales (Ettensohn *et al.*, 1988). In central Ohio, the base of the Huron Member is marked by an abrupt upward change from the recessive-weathering, light-colored Upper Olentangy Shale to resistant, brownish-black and strongly joint - fractured, and petroliferous-smelling shale deposits (Figure 10 A, B). In this region, the Huron succession is characterized by high radioactivity on logs, and has been interpreted to record deposition within a stagnant, euxinic basin setting somewhat comparable to that of the present-day Black Sea, based on intermediate concentrations of organic carbon and low organic carbon-to-sulfur ratios, as well

as, trace element concentrations (Robl & Barron 1988). However, based on the study of paleo-redox-related trace metal proxies, Rimmer *et al.* (2010) determined that the bottom water



**Figure 9:** High-magnification photograph of *Protosalvinia*. Cleveland Museum of Natural History sample CMNH P21385E, Worthington, OH. Scale Bar = 5 mm (photo credit: Joe Hannibal).

conditions alternated from anoxic to dysoxic during Huron Member deposition, and were distinctly dysoxic during deposition of the lower part of the Cleveland Member in northern Ohio, followed by predominantly anoxic conditions for the upper part of that unit. Ettensohn (1998) argued that organic matter accumulation in these shales was productivity-driven, probably due to nutrient supply to the basin, and that bottom conditions were predominantly dysoxic rather than anoxic.

Another feature of the upper Huron succession and Cleveland Shale Member outcrop sections is the presence of distinctly regular, repeating decimeter-to-meter-scale alternations of differentially resistant, organic-rich black shale bands with less resistant dark gray strata (Figure 11). Vertical changes in the spacing of this ribbing appears to correlate with changes in shale

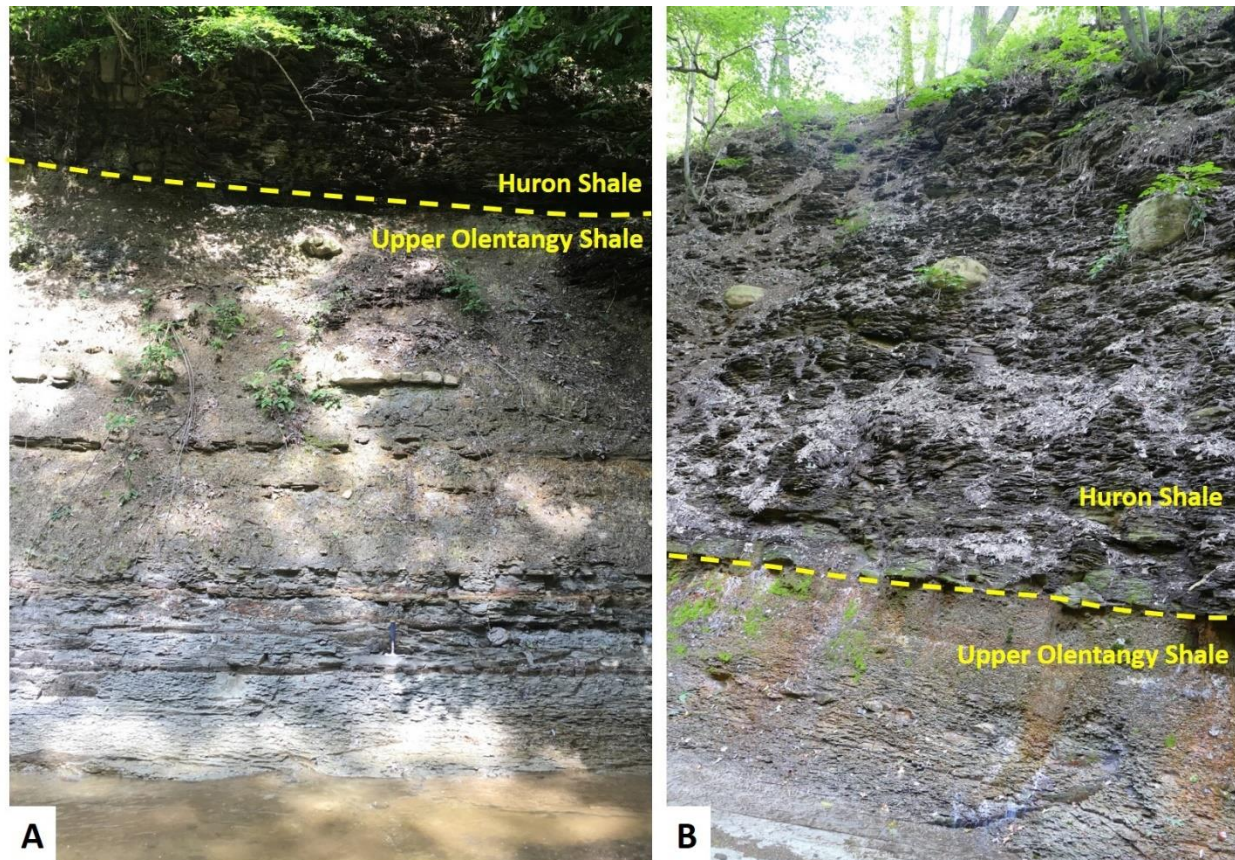


**Figure 10:** Field photographs of gray Upper Olentangy Shale overlain by dark brown to black shale of the Huron Shale at (A) Shale Hollow Park (photo credit: Erika Danielsen), and (B) Highbanks Metropark (photo credit: Randy Blood).

radioactivity; closely-spaced ribbing apparently denotes “hotter”, more organic-rich, depositional episodes that can be traced regionally on logs. In the Columbus area, most of these alternations impart a distinctly “ribbed” appearance to outcrop faces, which appears to be rhythmic in character (Figure 11). This rhythmicity can, in fact, be followed southwestward across Ohio into Kentucky, where the average spacing of the ribs thins southwestward in rough proportion to the thinning of the Huron succession. Jaminsky *et al.* (1998) recognized decimeter-scale rhythms in the thinner Kentucky sections, which they attribute to repeating patterns of climate change associated with inferred Milankovitch cyclicity. We will see an example of this rhythmicity in partly submerged Huron sections along Alum Creek Lake, which can be viewed from the east-west Route 36/37 bridge over the reservoir (Kilbourne 7.5’ Quadrangle; Google coordinates 40.268680, -82.949800). Figure 11 shows the rhythmic bank section as viewed from the Route 36/37 bridge prior to flooding of the Alum Creek Valley.

The Huron yields a very low diversity biota composed of sparse, small phosphatic brachiopods, including the lingulid *Barroisella melie* (Hall, 1870) and orbiculoids, the cephalopod *anaptychus Sidetes (Spathiocaris)*, fish teeth and dermal plates, as well as carbonized and

permineralized driftwood. Most notable within the Huron is the occurrence of the algal (?) taxon *Protosalvinia* (*Foerstia*), which is notably concentrated between 90 – 120 feet (27 – 37 m) above the base of the unit in this region (Figure 9). *Protosalvinia* occurs as 0.5 – 4.0 mm –



**Figure 11:** Field photograph of rhythmic Huron Shale beds exposed along Alum Creek near Route 36, Sunbury, OH. Image courtesy of the Ohio Department of Natural Resources; Division of Geological Survey.

diam. flattened circular to larger doubly branched, resinous to pustulate black blobs on bedding surfaces. Interpreted to be a reproductive, sporangial thallus-type structure of some unknown plant-like organism (Over *et al.* 2009), it is found to be an important zonal floral epibole in North American sections and in Brazil. In central Ohio, the *Protosalvinia* zone is characterized by fine alternations of gray and black shale that impart a “pin-striped” appearance to sections; in the clay-dominated gray shale layers, the *Protosalvinia* thalli occur sparingly, but are usually well-preserved; in the black bands, thalli are locally densely concentrated on bedding planes, but are often degraded and fragmental, suggestive for significantly slower rates of sediment accumulation for the black layers as is also supported by organic geochemistry (Algeo & Scheckler 1998; Algeo *et al.* 1995, 2001). The Huron, distal portions of the Chagrin Shale, and the Cleveland Member are notable for the occurrence of cone-in-cone concretion beds and lentils. In the Cleveland Shale Member, these features, understood to be pressure solution phenomena, are locally abundant and are occasionally observed to track rock fracture planes at angles to primary bedding.

The sharp base of the Huron Member displays evidence for submarine scour and localized concentration of detrital pyrite and fish bones in channel-like features (Figure 12). At STOPS 1 – 3 and elsewhere, the actual planar basal contact of continuous black Huron, yields reworked, non-calcareous lag debris, but only sparingly. However, only a couple of inches below this contact within the topmost underlying gray shale are irregular surfaces that appear to be widely-spaced erosional channels with variable and complex sediment fills (Figure 12). At STOPS 2 and 3, one or two mm-thick dark bands, 2.5 – 5 inches (6.4 – 12.5 cm) below the base-black Huron contact, can be followed laterally to where they may locally wedge into narrow channel-fill successions, characterized by sequential graded pyrite, black shale, and gray shale fills (Figure 12). Often the bases of channels display a graded (upward-fining) fill of fragmental, tubular, pyritized burrow tubes that were exhumed from underlying gray muds and subsequently hydraulically-aligned parallel to the long axes of the channels. Over and Rhodes (2000) recovered diverse and zonally significant conodonts from similar scour fills in the topmost few centimeters of the Upper Olentangy succession and determined that this erosional interval marked, not only the base of the Famennian Stage, but the redefined base of the Huron Member as well.

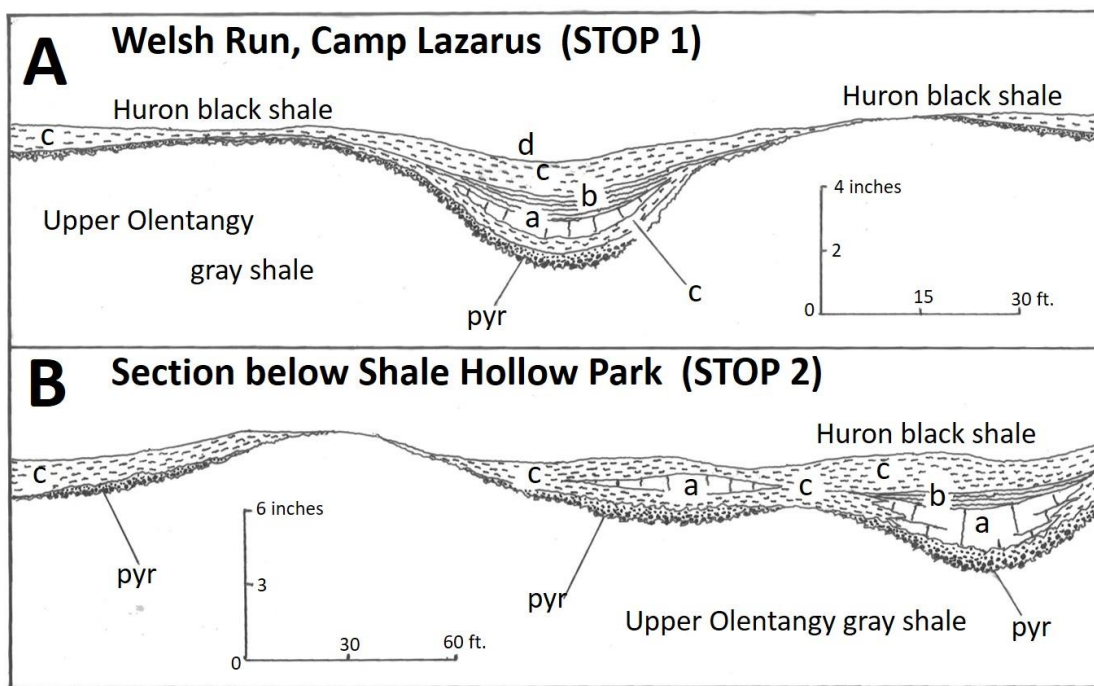


**Figure 12:** Base-Huron Member discontinuity contact characterized by submarine erosional channels with variable channel-fill deposits as seen at STOPS 1 and 2. Note the presence of localized lentils of coarse, often graded, accumulations of detrital pyrite grains, frequently displaying current-alignment of linear clasts. Lettered symbols denote: a, carbonate lentils (ferroan dolomite or siderite); b, lentils of black shale sandwiched between gray-greenish shale, carbonate, or detrital pyrite within channels; c, greenish-gray shale above base-Huron erosional contact and below continuous Huron black shale succession; d, Huron black shale differentially filling uppermost part of channel-fill succession at Welsh Run; pyr denotes locally thickened detrital pyrite accumulations along base-Huron contact.

Beneath the Middle Devonian (Upper Givetian-age) Geneseo black shale succession in western New York and the Late Famennian-age Cleveland Shale Member are regional disconformities diachronously capped by black shale deposits (Baird & Brett 1986, 1991; Baird *et al.* 2013; see model in Figure 13). These contacts both display laterally separated, coarse lens-shaped accumulations of reworked, detrital pyrite grains (“golden gravel”) that apparently accumulated on a variably dysoxic, sediment-starved, sloped sea bed prior to diachronous burial by black, basinal muds. Leicester Pyrite lenses, ranging in thickness from 0.0 – 11 inches (0.0 – 28 cm) in thickness, are composed of detrital pyrite clasts (fragments of diagenetically pyritized endichnial burrows, small pyrite nodules, fossil steinkerns) that have been exhumed from the underlying Windom Shale and both reworked and current-aligned on the sea floor prior to significant accumulation of onlapping Geneseo black mud. The absence of associated exhumed calcareous fossil debris in lenses reflects selective carbonate dissolution in a variably dysoxic, carbonate undersaturated bottom setting. In Ohio, the base of the Cleveland Shale Member is similarly characterized by coarse detrital pyrite lenses that mark the base of this unit in numerous northern Ohio localities (Figure 14).

The “laterally-separated lens” geometry of the pyrite accumulations is understood to reflect the concentration of lag alloclasts in parallel to sub-parallel submarine erosional channels separated by higher inter-channel divides (Baird & Brett 1986, 1991; Figure 13). These channels are interpreted to have been developed on an east-southeast-facing, sediment-starved, erosional slope surface, upslope from the westward, upslope limit of onlapping, basinal, Geneseo, black mud deposits (Figure 13). Scattered higher pyrite lenses within the basal few centimeters of the Geneseo reflect the temporal/spatial diachroneity of the alloclast transport process; pyritic gravel moving down the erosional slope is understood to have been episodically transported over the upslope edge of black sediments that had already been deposited earlier (Figure 13). This diachronous, westward-younging of the Leicester and associated basal Geneseo black shale deposits is interpreted to reflect long-term, flexural, westward, foreland basin expansion timed with Taghanic eustatic deepening (Huddle & Repetski 1981; Kirchgasser 1994; Baird & Brett 1991).

As with the base of the Geneseo Formation, it is suspected that basal Cleveland strata above the regional base-Cleveland unconformity are regionally diachronous, except that this onlap is regionally southeastward, not westward, reflecting the position of the Cleveland Member depocenter in a narrow belt across northern and north-central Ohio (Figure 15 A, B). Careful mapping of a new provisional unit (“Penitentiary Glen bed”) across parts of the Cleveland metropolitan area, shows that this turbiditic, gray shale unit is observed to descend southeastward to the base-Cleveland contact and “pinch out” onto it in sections in the Cuyahoga and Chagrin valleys (Baird *et al.* 2016; Figure 15 A). Careful follow-up extraction and

analysis of conodonts from the Skinners Run pyrite lenses and superjacent Cleveland Shale Member beds is intended to further confirm this model. It is suspected that the erosional base of the Huron Member may also reveal evidence for a regional paleoslope, but this remains to be determined.

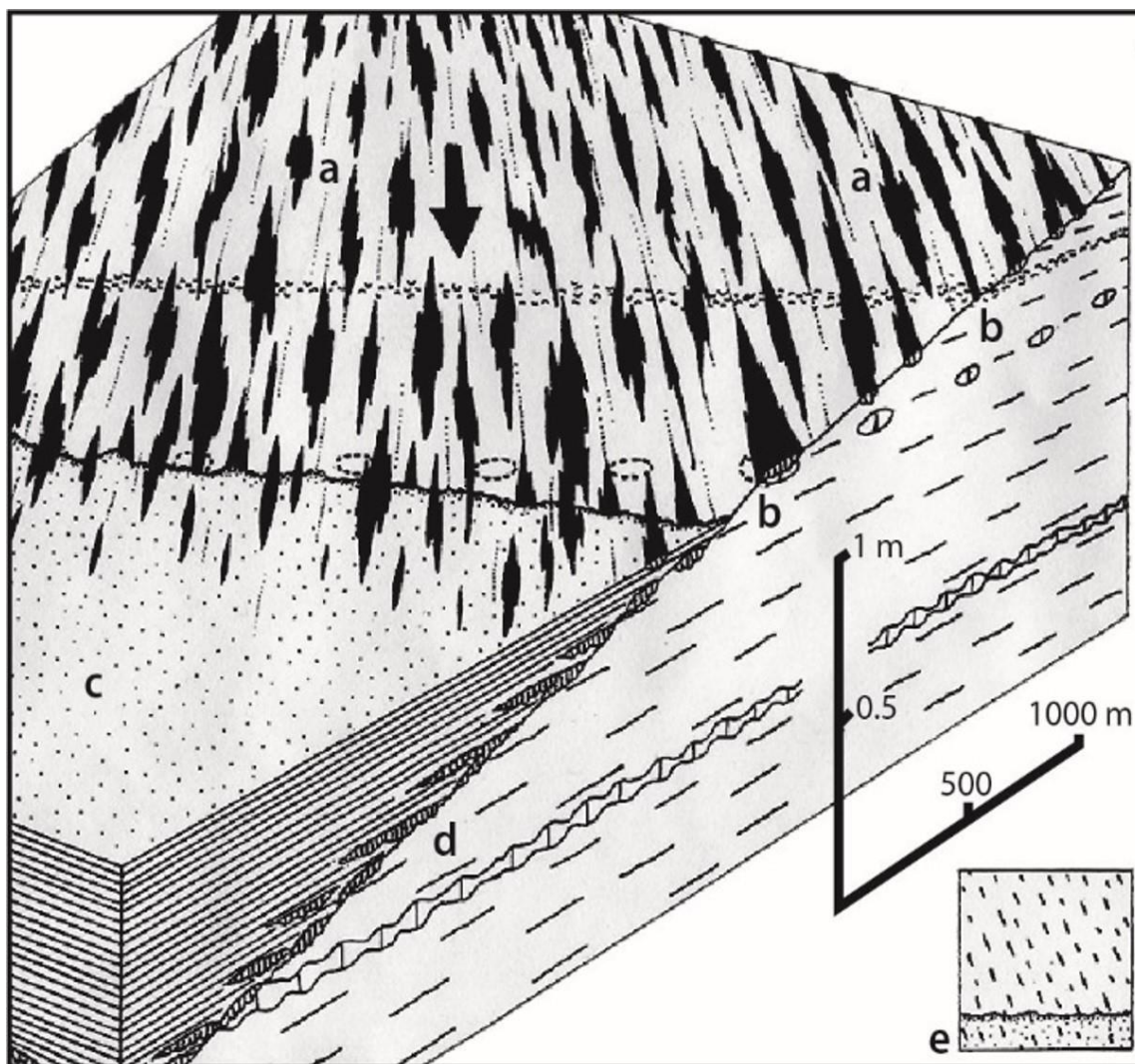


**Figure 13:** Anoxic, erosional, submarine basin ramp setting associated with regional onlap of Late Givetian black mud deposits during the Taghanic onlap event (see text). Lettered units include: a, sediment-starved, eastward-facing erosion surface, characterized by tractional, downslope transfer of detrital pyrite-dominated lag debris in erosional submarine furrows; b, discrete detrital pyrite accumulation in linear furrow depression into underlying, gray mudstone deposit. When later buried during transgressive onlap, these accumulations appear as laterally separated “lenses” along the contact between compactionally dewatered gray and black shale deposits; c, onlap of organic-rich, black mud deposits associated with the Late Middle Devonian Taghanic transgression event; d, temporal imbrication of Leicester pyrite lenses with lowest black mud accumulations in furrows, timed with diachronous, regional onlap event. Inset box e, (lower right), shows size of Leicester Pyrite lenses closer to actual regional scale. Skinner’s Run Bed tractional/depositional events were similarly timed with initial Cleveland black mud accumulation during southward regional onlap, as is indicated by complex interlayering of detrital pyrite and thin black shale layers (from Baird & Brett, 1986, 1991).

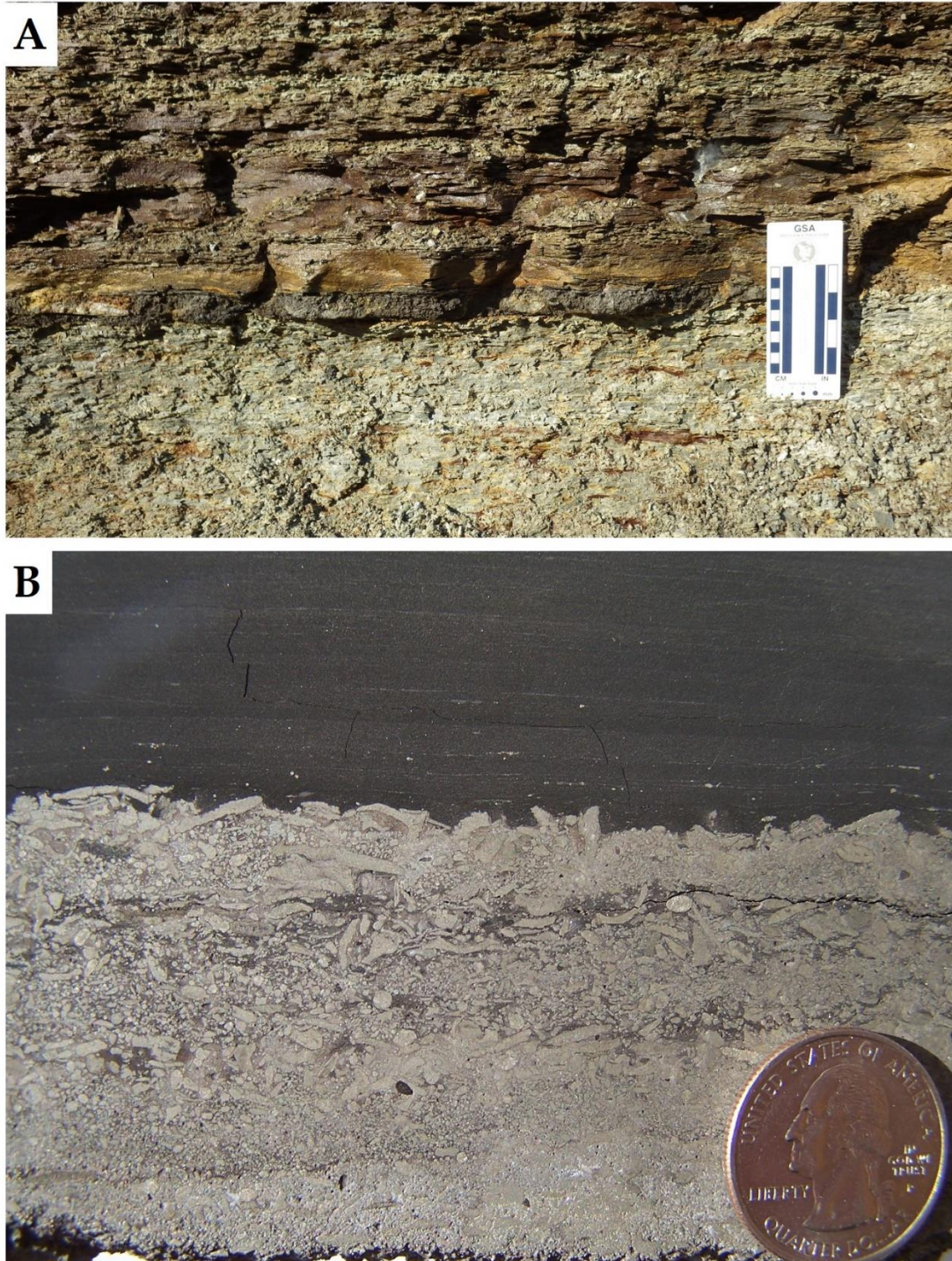


**Figure 14:** Chagrin Member capped disconformably by harder, black Cleveland Member along West Creek (“Skinner’s Run”) in the Brooklyn Heights Town Park; (A) Occurrence of the Skinner’s Run pyrite-bone bed in bank sections along creek. At this locality and at others, it is expressed as laterally discontinuous lenses of detrital pyrite and insoluble phosphatic bioclasts along the contact; (B) Close-up view of thick, channelized, detrital pyrite lens that has been cut and polished. Note development of four discrete detrital pyrite accumulation events, separated by thin black shale bands; this indicates contemporaneity of detrital pyrite accumulation with the anoxic setting associated with the onlap of Cleveland Member black mud deposits. (Photo credit: Gordon Baird)

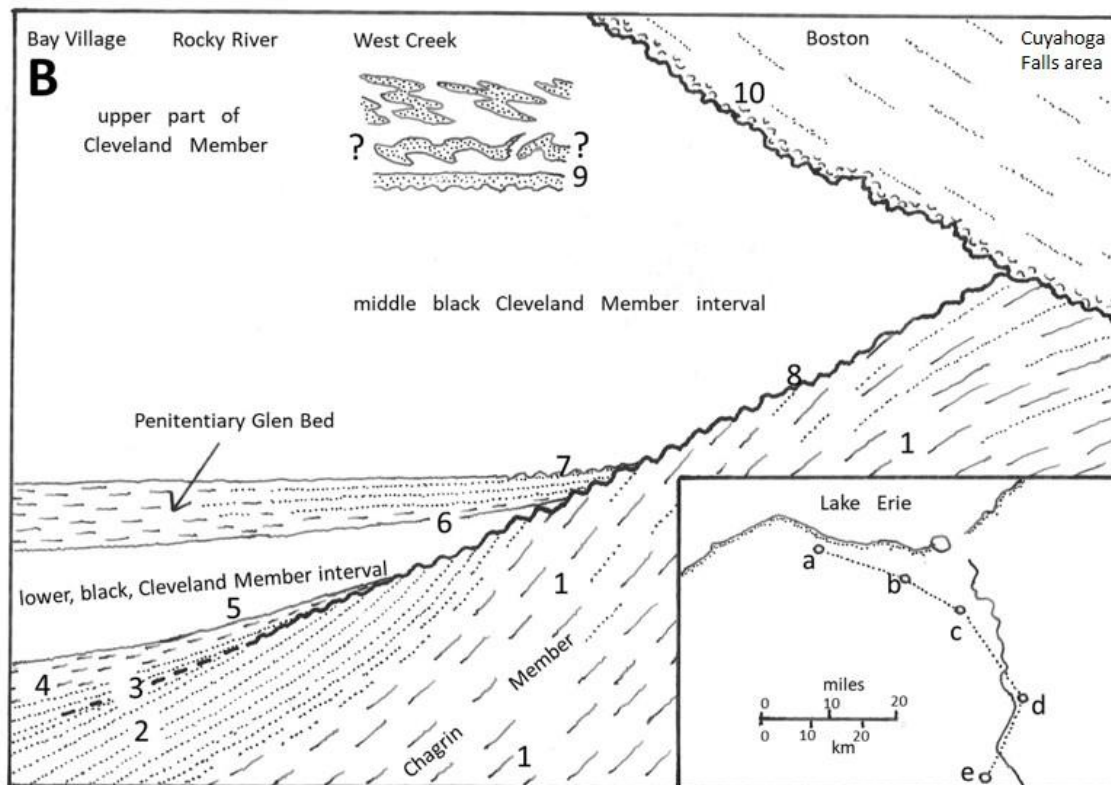
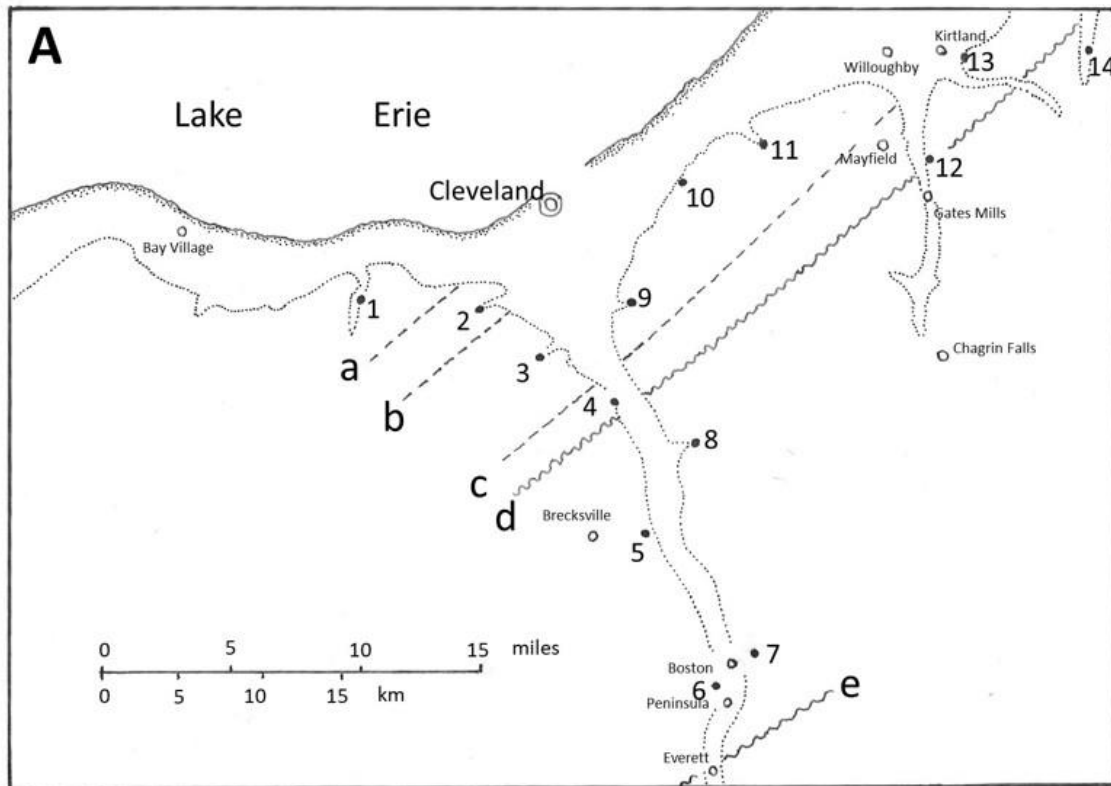


**Figure 15:** Inferred southeastward transgressive onlap of Cleveland member over regional sub-Cleveland erosion surface; A, Regional (spatial) map view of southeastward diachronous onlap of Cleveland Shale divisions onto a regional, sloped erosion surface: a, base of lowest (gray shale) division of Cleveland member, b, base of lower Cleveland member black shale interval, c, base of “Penitentiary Glen” interval of gray, silty shale, d, base of thick, upper Cleveland member division of black shale, e, southeastward Cleveland erosional limit below Bedford Formation. Numbers 1 – 14 denote key Cleveland member sections; B, Vertical, northwest-southeast stratigraphic transect showing the onlap pattern displayed in Figure 15 A. Numbered units/events include: 1, gray, Chagrin member mudstone, 2, turbiditic fan deposits of uppermost progradational Chagrin succession; 3, basinward “toe” of discontinuity surface, 4, predominantly gray shale, deposits of lowest (unnamed) Cleveland member division, 5, lower Cleveland black shale interval, 6, silty, gray shale deposits of the informal “Penitentiary Glen bed”, 7 – 8, onlapping upper Cleveland member black shale division, 9, seismically emplaced siltstone-sandstone masses within Cleveland Shale at West Creek and Rocky River, 10, base-Bedford Formation disconformity. Lettered localities (inset) include: a, Rocky River Metropark, both at and north of Lorain Road overpass bridge, b, exposures along Big Creek north of Big Creek Metropark, c, West Creek (“Skinner’s Run”) in Brooklyn Heights Town Park, d, Tinkers Creek in Bedford Glen Metropark, e, Slipper Run.

The presence of coarse sand- and gravel-grade, detrital pyrite and phosphatic clast accumulations along the basal Genesee and Cleveland Shale Member contacts might be construed as evidence for shallow, high-energy conditions timed with the onset of their deposition. Smith *et al.* (2019) argue that the deposition of units such as the Huron and Cleveland are the products of shallow shelf deposition (“meters-to-tens-of- meters depth”) in energetic settings. Moreover, they argue that the pervasively high organic content of these units reflects selective concentration and preservation of organic matter in thin, condensed deposits coeval to the thick Catskill Delta wedge. We argue that higher-energy events do explain the presence of the submarine channels and the coarse lags within them, but that this erosion is associated with deep-storm-wave impingement, which can extend downward to several hundred feet of depth or the complex actions of shoaling geostrophic currents or internal waves (Baird & Brett 1991). In addition, the coarse lenses and channeling quickly disappear moving upward into the overlying Genesee, Huron, Cleveland, or Sunbury Shale successions, to be replaced by much finer-grained black, and often rhythmic, black shale facies. This suggests that the coarse, channelized lags are transient features, linked to process-specific, diachronous onlap of sloped erosion surfaces.

In stark contrast, the rhythmic Huron succession that we will pass at Alum Creek Lake (Figure 11) should serve as a visual type example of time-rich, pervasively low energy, sediment accumulation under conditions spanning several million years for this unit. Within the Huron and comparable shale units, one can find thin storm-related silty layers, but these are on the order of mm-scale thickness. Likewise, bioturbation is observed in even the darkest Huron layers, but it is expressed as soft-sediment “meioturbation” attributed to very small deposit feeders and scavengers (meiofauna) capable of survival under oxygen-deficient conditions. Conversely, in a shallow, lagoon-like setting, storms, as well as, seasonal break-up of water-mass boundaries would tend to destroy organic-rich sediment fabrics and favor the

introduction of deposit feeders. Outsized storm layers, siliciclastic pulses, and seismic events, typically more common near basin margins, would have overprinted and obscured the rhythmicity seen along Alum Creek (Figure 11) had the environment been just a few meters deep. In sum, the very large spatial- and temporal-scale of the Huron-Kettle Point-New Albany-Chattanooga black shale deposit is suggestive of a very large cratonic sea capable of sustaining long-term, water mass stratification in a postulated water depth range from 100 to several hundred feet.

### *The Concretions of the Huron Shale*

The lower part of the Huron Shale hosts numerous, often large, carbonate concretions (Figure 16). While smaller concretions, ~ 1-foot diameter (0.25 m) are more spherical, larger concretions, up to 9 feet (3 m) in diameter are more ellipsoidal (Criss *et al.*, 1988; Figure 16C and 17). Huron Shale-hosted concretions show a complex history of mineral replacement and crystallization (Criss *et al.*, 1988). Many concretions host a coarse crystalline calcite core, often removed by weathering (Figure 16B,D), an intermediate section of calcite, which shows partial to nearly complete replacement by dolomite, and an outer rind rich in pyrite (Criss *et al.*, 1988, Figure 18). It is worth noting that within Huron Shale-hosted concretions, it is common to find the fossil remains of fish and plant debris (Criss *et al.*, 1988). Criss *et al.*, (1988) build a convincing argument for proteinaceous remains acting as a nucleation site for concretion formation. Such a model, however, is inconsistent with field observations of many other carbonate concretions (Dix & Mullins, 1987; Coniglio & Cameron, 1990; Lash & Blood, 2004a; Enomoto *et al.*, 2012; Blood *et al.*, 2019). It is possible that the large size of Huron concretions merely incorporates buried organic material during the formation process described below. Clearly, further investigation is warranted.

Most Huron Shale-hosted concretions are found in the lowermost portion of the Huron Shale, yet the base of the Huron Shale may be time transgressive (Criss *et al.*, 1988). However, Criss *et al.*, (1988) point out that across much of the outcrop belt and east into the subsurface, concretions occur no more than 90 feet (30 m) below the synchronous *Protosalvinia* Zone, and suggest a temporal relationship to this zone as opposed to the base of the Huron Shale. While the concretions may not be associated with the *Protosalvinia* zone, their linking of the concretions to a time-correlative bed is well taken. Such stratally confined carbonate concretions hosted by Middle and Upper Devonian shale of the Appalachian Basin (Figures 16A and 19A) are interpreted to have formed in association with anaerobic oxidation of methane (AOM) and consequent enhanced alkalinity (Lash & Blood, 2004a,b; Lash, 2015a,b, 2018). Each concretionary horizon is interpreted to reflect the diagenetic signature of AOM within the sulfate methane transition zone (SMTZ), a diagenetic horizon of indeterminate thickness along which downward-diffusing seawater sulfate and upward-diffusing methane are consumed by a

consortium of methane-oxidizing archaea and sulfate-reducing bacteria (Reeburgh, 1976; Hoehler *et al.*, 1994; Niewöhner *et al.*, 1998; Hinrichs *et al.*, 1999; Borowski *et al.*, 1999; Boetius *et al.*, 2000; Paull *et al.*, 2000). Low sedimentation rates focus the diagenetic effects of AOM, maintaining elevated pore water alkalinity within the SMTZ for an extended period of time (Borowski *et al.*, 1999; Rodriguez *et al.*, 2000; D'Hondt *et al.*, 2004; Snyder *et al.*, 2007; Dickens & Snyder, 2009). The hypothesis of formation of Rhinestreet Shale-hosted carbonate concretions as a consequence of anaerobic oxidation of methane is buttressed by their association with  $^{34}\text{S}$ -enriched pyrite and barite (Lash, 2015a,b), an argument made by Borowski *et al.* (2013) based on their investigation of younger methane-bearing marine deposits of the Blake Ridge.

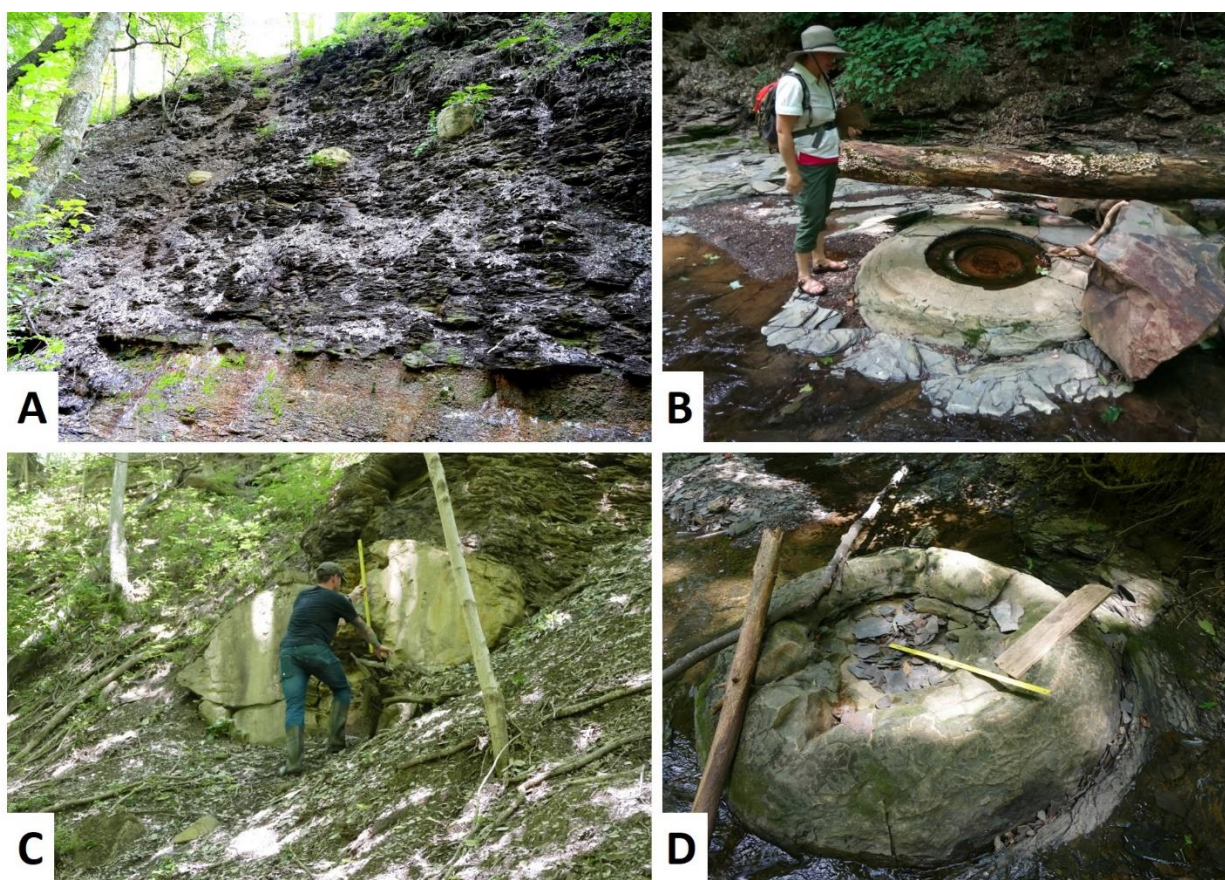


**Figure 16:** Field photographs of carbonate concretions hosted by the Huron Shale; (A) stratigraphically confined carbonate concretions in the lower part of the Huron Shale at Highbanks Metropark, Lewis Center, OH; (B) Large carbonate concretion with missing center owing to erosion, Shale Hollow Park, Lewis Center, OH; (C) Concretion over 9 feet (3 m) in diameter in a shale bank exposed at Shale Hollow Park, Lewis Center, OH; and (D) a large concretion with eroded center exposed along Welsh Run at Camp Lazarus, Lewis Center, OH. (Photo Credit: (A,B, D) Randy Blood and (C) Virginia Hebert)

Canfield & Raiswell (1991) argue convincingly that the timing of concretion growth is best deduced by assessing the degree of compaction of encapsulating shale at the time of concretion formation. The wrapping of Devonian concretions by laminated shale (Figure 19) is consistent with their formation at shallow burial depths early in the compaction history of these deposits (Lash & Blood, 2004b). Indeed, Newberry (1873), noted such a relationship between Huron Shale-hosted concretions and encapsulating shale, and therefore argued concretions formed in an unconsolidated sediment. Further, SEM analysis of host shale samples collected from adjacent to lateral edges of concretions reveals a modestly open clay grain microfabric (Lash & Blood, 2004b). Shale samples collected centimeters away from these areas, however, display a strongly oriented platy grain microfabric (Lash & Blood, 2004b). The former deposits are interpreted to have occupied pressure shadow regions of host sediment that were shielded by adjacent rigid carbonate during early and shallow mechanical compaction (Lash & Blood, 2004b).

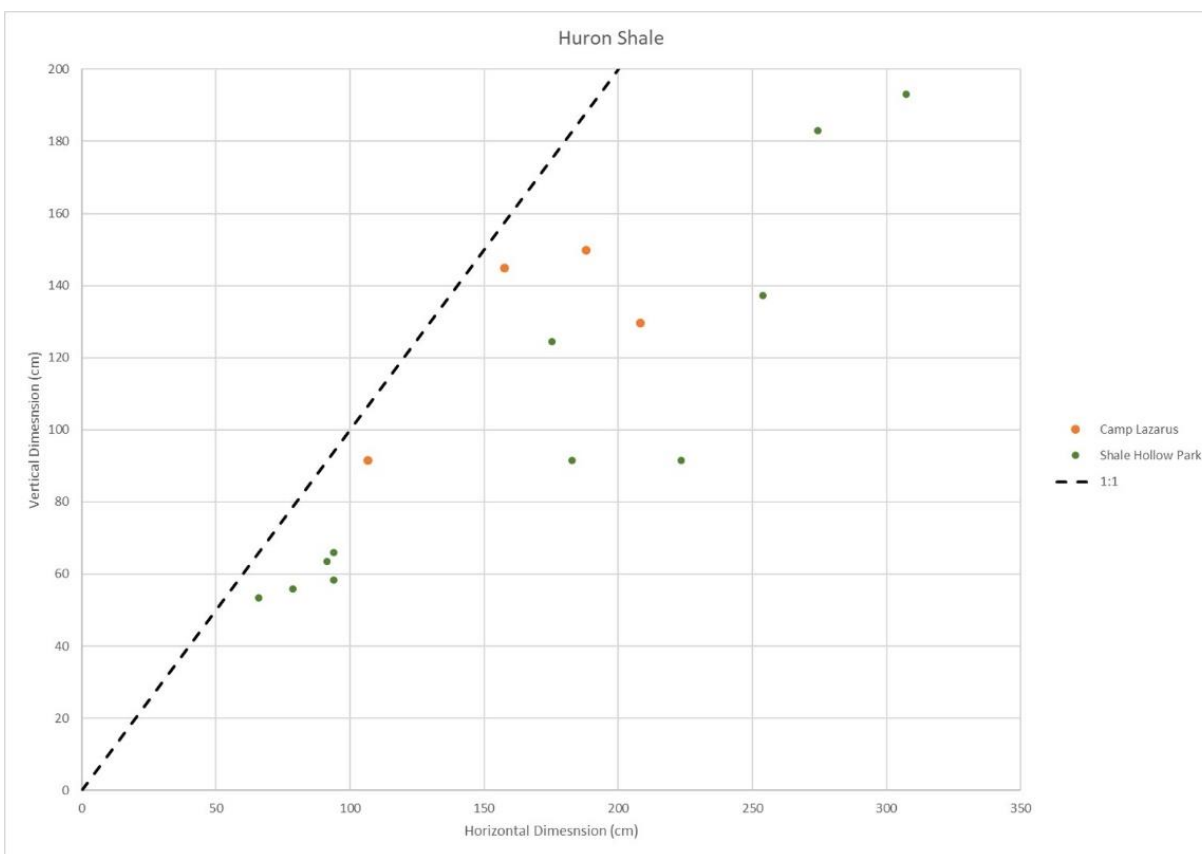


**Figure 17:** Cross plot of the horizontal and vertical dimensions of Huron Shale-hosted carbonate concretions measured at Camp Lazarus and Shale Hollow Park in Lewis Center, OH. Note that smaller concretions are more spherical while larger concretions tend to be more ellipsoidal.

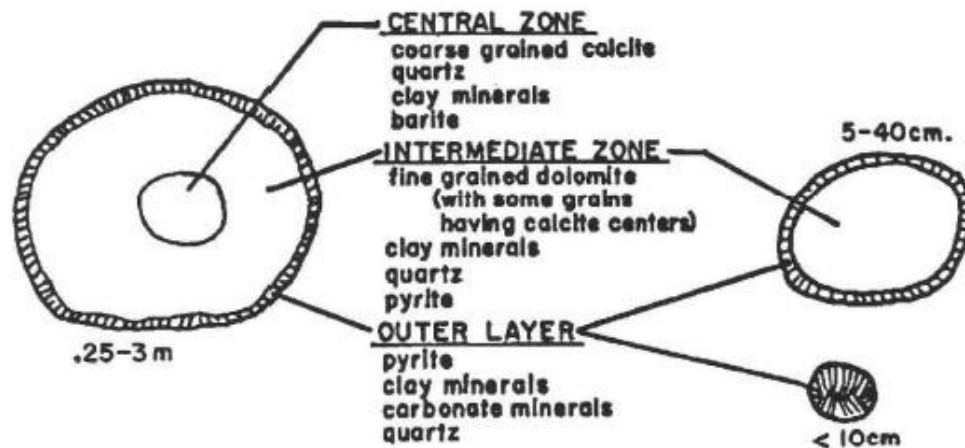


**Figure 18:** Schematic diagram from Criss *et al.* (1988) depicting mineralogic zonations of Huron Shale-hosted carbonate concretions.

A shallow depth of concretion formation is further suggested by the pervasive occurrence of randomly oriented clay grains reminiscent of clay floccules within concretions (e.g., O'Brien, 1981; O'Brien & Slatt, 1990; Bennett *et al.*, 1991; Slatt & O'Brien, 2013; Figure 20A). The depositional texture of the host clay was likely shielded from compaction-related grain reorientation by the precipitation of calcium carbonate soon after deposition. Similarly, algal cysts ranging from 60 to 120  $\mu\text{m}$  in diameter observed in samples collected from within concretions display their original spherical shapes (Figure 20B, C). Similar observations of uncrushed *Tasmanites* and radiolarians have been reported from Huron Shale-hosted concretions (Clifton, 1957; Foreman, 1959). The spherical structures, like the depositional clay grain microfabric, appear to have been protected from burial-related flattening (Figure 20D) by the pervasive precipitation of microcrystalline calcite cement.

The duration of formation of the studied concretions cannot be quantified, yet micro-textural features suggest that they precipitated rapidly at shallow depth, perhaps several meters below the sediment-water interface (SWI) (Lash & Blood, 2004a,b). The uniformly very-fine grain size (15 – 25  $\mu\text{m}$ ) of matrix carbonate and lack of textural evidence of growth banding evince a rapid growth history of a single generation of calcite microspar (Selles-Martinez, 1996; Mozely, 1996; Raiswell & Fisher, 2000, 2004; Gaines & Vorhies, 2016), that was later replaced by complex diagenetic processes in the Huron Shale (Criss *et al.*, 1988).

Concretions hosted in black shale display depositional laminae inherited from the host sediment that accumulated under oxygen-depleted conditions thereby limiting or precluding the activity of bioturbating organisms (Figure 19B,E; Lash & Blood, 2004a). It is noteworthy that

most concretion laminae do not thin systematically from concretion centers to edges as would be expected of carbonate masses that precipitated radially coincident with burial-related compaction. Rather, it appears that the concretion masses formed at an essentially steady depth below the sediment-water interface. Indeed, the omnipresence of spherical algal cysts and clay floccules within concretions is suggestive of a generally pervasive growth history that entailed the infilling of void space in the host sediment. It is noteworthy, though, that the studied concretions display constant or modestly diminishing carbonate (reflected in Ca abundances) from concretion centers to edges (Criss *et al.*, 1988; Lash & Blood, 2004a), perhaps an indicator of outward decreasing porosity of the host sediment at some point in their formation history (e.g., Raiswell, 1976; Coleman & Raiswell, 1981). However, the fact that the infilling of pore space by calcite cement may be attended by some degree of grain displacement (e.g., Raiswell & Fisher, 2000) cautions against inferring a simple relationship between the abundance of carbonate cement and sediment porosity. Indeed, Lash & Blood (2004a) argued that the passive infilling of pore space within host sediment associated with the formation of concretions of the Rhinestreet Formation was accompanied by some displacement of siliciclastic grains.

The prevailing view of calcium carbonate precipitation driven by AOM holds that the  $\delta^{13}\text{C}$  of pore water and authigenic carbonate should reflect the isotopic composition of the methane substrate (i.e., less than -30 ‰ V-PDB; Borowski *et al.*, 1997; Aloisi *et al.*, 2000). However, recent investigations of pore fluid geochemistry of methane-charged sediments have revealed relatively high  $\delta^{13}\text{C}_{\text{DIC}}$  values (> -20 ‰ V-PDB) within SMTZs reflecting the commingling of DIC generated by organic matter degradation and methanogenesis at depth with that produced *in situ* by anaerobic oxidation of methane at the sulfate methane transition zone (Rodriguez *et al.*, 2000; Borowski *et al.*, 1999; Sivan *et al.*, 2007; Snyder *et al.*, 2007; Kastner *et al.*, 2008; Dickens & Snyder, 2009; Chatterjee *et al.*, 2011; Kim *et al.*, 2011; Malinverno & Pohlman, 2011).

The modestly  $\delta^{13}\text{C}$ -enriched stable carbon isotope values documented from concretions throughout the Devonian shale succession of the Appalachian Basin by previous studies (Dix & Mullins, 1987; Siegel *et al.*, 1987; Criss *et al.*, 1988; Coniglio & Cameron, 1990; Lash & Blood, 2004a; Lash, 2015a,b; Blood *et al.*, 2019) could reflect abnormally high DIC contributions from organoclastic sulfate reduction (OSR) in the bacterial sulfate reduction zone (BSRZ) (Nyman & Nelson, 2011; Teichert *et al.*, 2014). Indeed, Raiswell & Fisher (2004) posited that large calcium carbonate concretions described from outcrop may originate in the BSRZ. They add, though, that calcium carbonate precipitated in this diagenetic zone is limited to no more than porous, low-density masses of approximately 1.5 wt. % calcite rigid enough to preserve the depositional clay grain microfabric of the host sediment. Raiswell & Fisher (2004) further suggest that the dense bodies of concretionary cement observed in outcrop

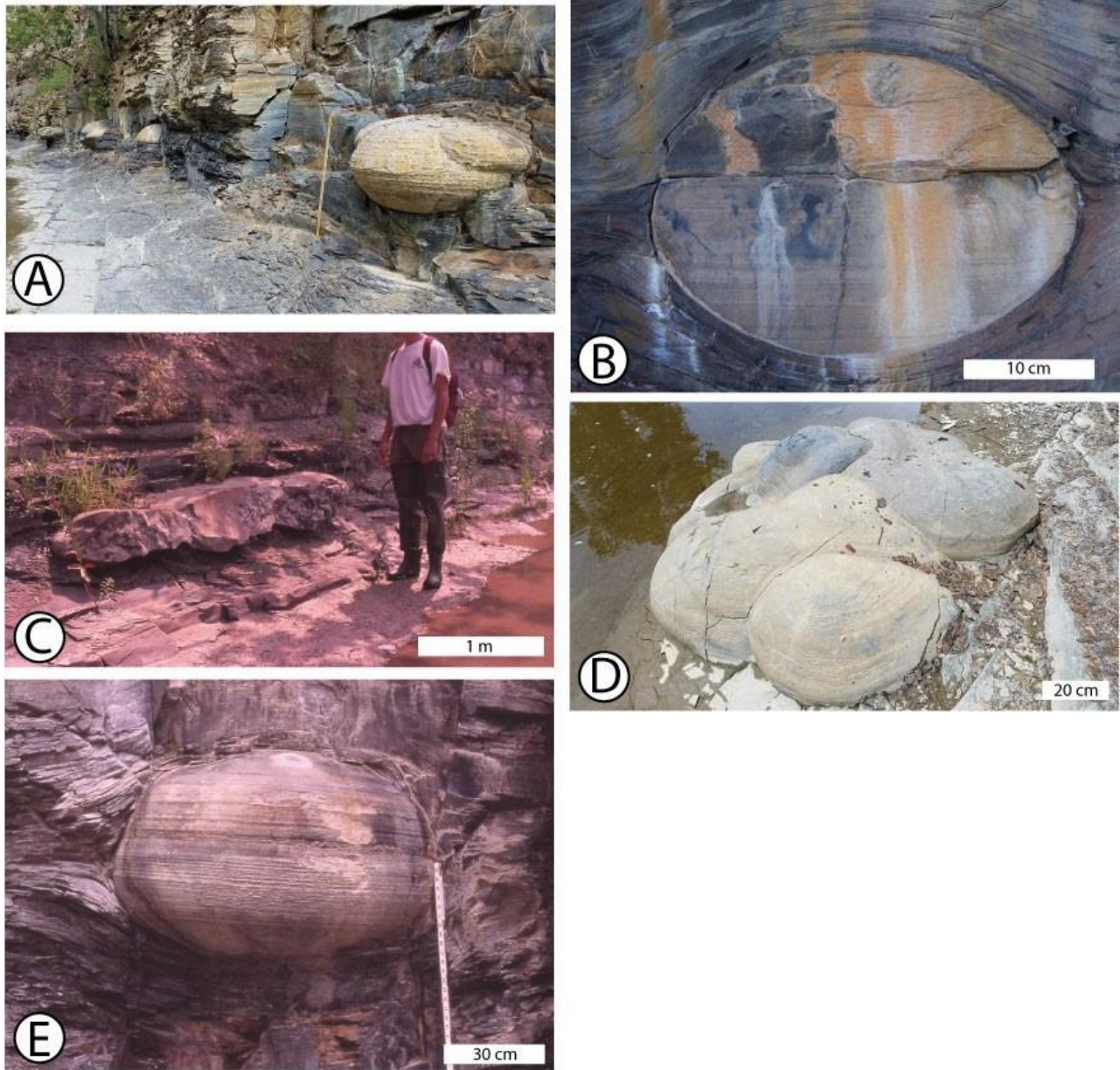


**Figure 19:** Field photographs of carbonate concretions displaying textural aspects described in the text.; (A) carbonate concretion horizon in the lower Rhinestreet Formation; measuring stick = 1 m (B) carbonate concretion of the Rhinestreet Formation displaying internal laminae; note wrapping of host shale around the concretion; (C) elongate carbonate concretion hosted by organic-deficient gray shale in the upper Rhinestreet Formation; (D) coalesced carbonate concretion in the lower Dunkirk Formation; (E) laminated carbonate concretion in the lower Rhinestreet Formation. (Photo Credit: Gary Lash)

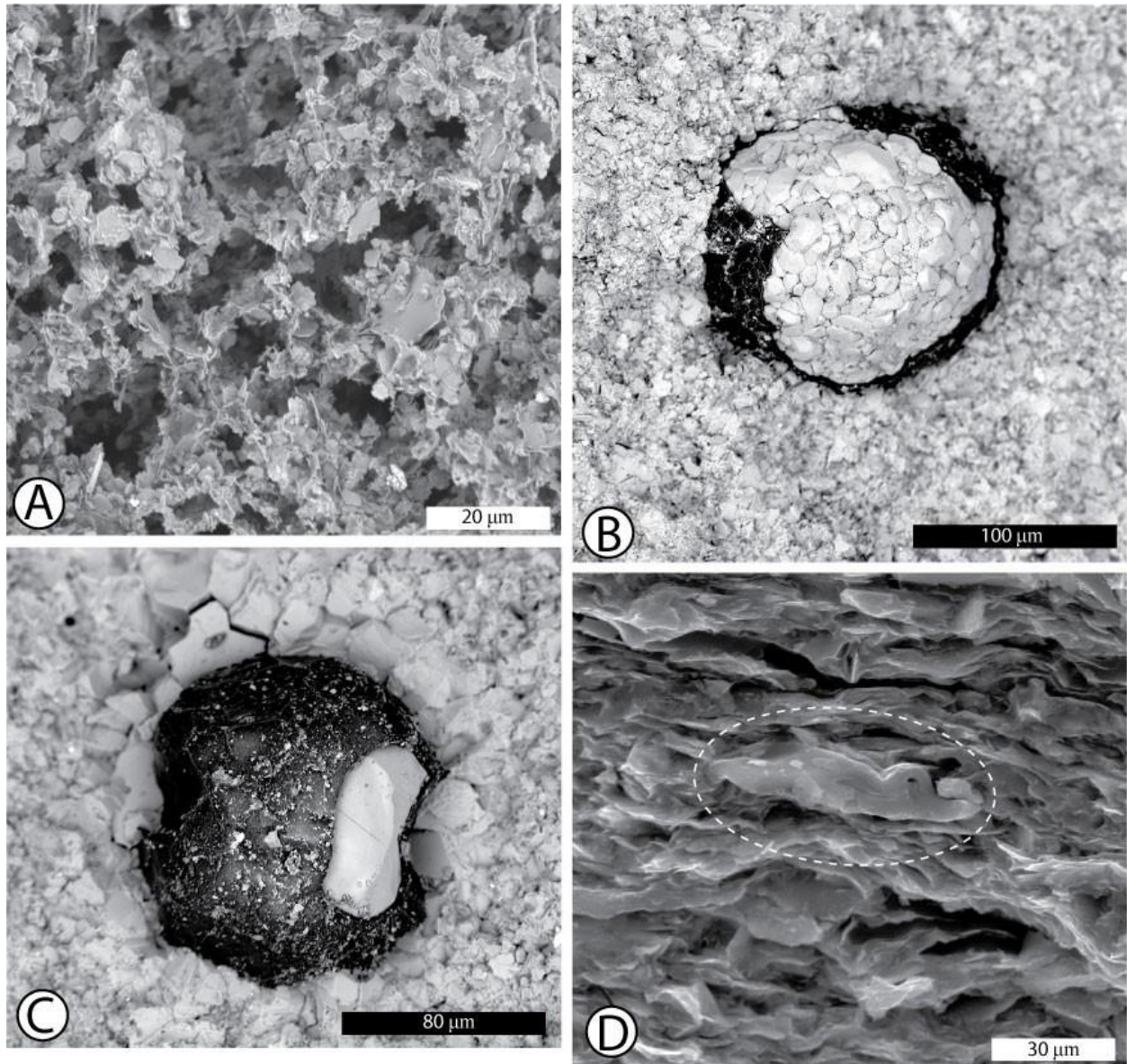


**Figure 20:** SEM images of micro-textural features of Middle and Upper Devonian concretions; (A) etched sample from the edge of a concretion from the Rhinestreet Formation; (B, C) spherical algal cysts in samples from the edge (B) and center (C) of a concretion of the Marcellus Formation; (D) host shale displaying a strongly oriented clay grain microfabric and a flattened algal cyst (indicated by dashed white oval) from a sample collected adjacent to a concretion of the Rhinestreet Formation. (Photo Credit: Gary Lash)

likely record the diagenetic imprint of the faster rates of sulfate reduction and consequent elevated alkalinities generated in association with AOM focused at the SMTZ by diminished sedimentation rates. Thus, the Middle and Upper Devonian concretion horizons are interpreted to reflect growth histories that were initiated within the BSRZ. However, it was not

until the nascent concretions were buried to depths of the SMTZ, perhaps a few tens of meters below the SWI (e.g., Borowski *et al.*, 2013), that the dense carbonate masses observed in field exposure formed as a consequence of the passive infilling of porosity. Thus, the discrete calcium carbonate-bearing stratigraphic horizons hosted by the Middle and Upper Devonian shale succession of the Appalachian Basin preserve the record of paleo-SMTZs, each one associated with an episode of reduced sedimentation rate (Lash, 2018). Such episodes may reflect smaller scale transgressions associated with orbital forcing mechanisms.

The widespread distribution of carbonate concretions throughout the Devonian shale succession of the Appalachian Basin (Siegel *et al.*, 1987; Dix & Mullins, 1987; Criss *et al.*, 1988; Coniglio & Cameron, 1990; Enomoto *et al.*, 2012) suggests that methane transport was diffusion-dominant, perhaps enhanced by burial-induced advection (e.g., Ritger *et al.*, 1987). There is no evidence that methane advected rapidly along faults and/or inclined permeable layers, or that methane was transported directly to the SWI as cold seeps. Further, geological considerations provide no evidence that methane was sourced in underlying methane hydrate systems (Lash, 2015a). It is likely, then, that most or all methane ascending the Devonian sedimentary column was consumed by AOM at numerous SMTZs, each one stabilized during a period of reduced sedimentation rate. Indeed, the diffusional nature of methane transport in the Appalachian Basin may account for isotopic differences between the studied Devonian concretions and authigenic carbonate described from many recent and modern methane-charged sedimentary systems. That is, most discussions focusing on authigenic carbonate precipitation in association with AOM address marine sediments overlying methane hydrate systems that typically yield higher methane fluxes and, therefore, higher rates of AOM (Smith & Coffin, 2014).

The size of concretions is heavily influenced by (1) the availability of reactants, and (2) the duration and degree to which sediment influx is reduced. Indeed, very slow sedimentation rates and abundant reactants tend to form large concretions. During prolonged hiatuses concretions coalesce into continuous diagenetic limestone beds such as the Scraggy beds of the Rhinestreet Shale (Lash & Blood, 2004a,b). However, concretion spacing remains poorly understood. At Shale Hollow Park we will see large concretions at random spacing. Some concretions are close enough together they have grown in to each other, while others are 10s of meters apart. Such random spacing is also observed in the Rhinestreet Shale (Blood *et al.*, 2019). It is possible that concretion location is controlled by the location of organic nuclei (Criss *et al.*, 1998) or the location of permeability pathways for upward migrating reactants. Regardless, such random spacing buttresses the model for deeper sourced carbon. Indeed, a predictable pattern of concretion spacing, where spacing and concretion size co-vary would exist if concretions were competing for locally sourced carbon associated with the beds they reside in.

### *The “devil’s kitchen”: Hydrocarbon production-induced fractures in the Huron Shale*

Along the outcrop belt from Cleveland, south through Columbus, OH and beyond, Upper Devonian black shales, specifically the Huron Shale and Cleveland Shale Member carry a set of roughly east-northeast (ENE) trending vertical open mode fractures known as joints (Figure 21). Further east, as one encounters more deeply buried black shale, increased fracturing and multiple joint sets are observed in black shale, which often extend into overlying gray shale deposits. Here at the Cleveland to Columbus meridian, joints appear largely confined to the black shale, rarely if ever observed in adjacent gray shale, and often occur in the most organic-rich facies (Figure 21B).

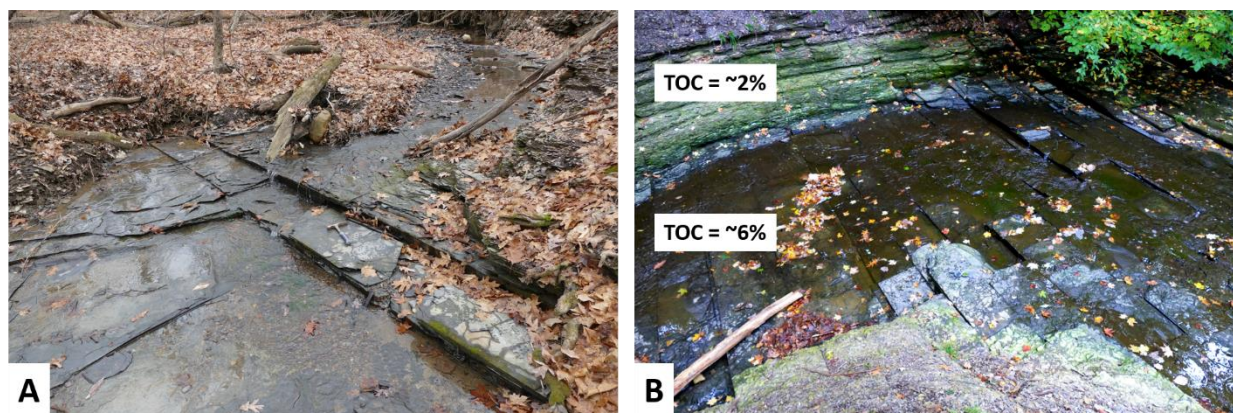


**Figure 21:** Field photographs of Mode I natural hydraulic fractures hosted by black shale; (A) ENE oriented joints in the Huron Shale exposed at Highbanks Metropark, Lewis Center, OH; (B) ENE oriented fractures developed in the most organic-rich portion of the Cleveland Shale at North Chagrin Reservation, Willoughby, OH. Note the development of joints in the more organic-rich (higher TOC) deposits. (Photo Credit: Randy Blood)

Two basic joint driving mechanisms are at work in the Earth’s crust; (1) absolute tensile stress and (2) natural hydraulic fracturing (Engelder, 1993). Tensional joints most commonly form as a consequence of joint-normal stretching in association with the development of tensile stress outside a neutral fiber (the boundary in a fold above which rocks are in tension, while those rocks below are under compression) such as during the slow growth of folds or as part of a forebulge induced by glacial loading (e.g., Clark, 1982). In such a scenario, it is rocks with higher Young’s Modulus that are likely to undergo preferential jointing. This model, as an explanation for the preferential jointing of black shale illustrated on this field trip, suffers from two drawbacks. First, the widespread occurrence of joint relationships illustrated in more deeply buried black shale (i.e., multiple joint sets and observed abutting relations) is inconsistent with joint formation as a consequence of fold-induced layer-parallel extension. Second, the orientations of the ENE and northwest (NW) joints, especially the latter, are

completely at odds with what is known of folding on the Appalachian Plateau (Engelder & Geiser, 1980). That is to say that, joints which formed at the same time as folding are not parallel to fold axes, and often occur well away from these structural features in more structurally quiescent parts of the basin. Finally, diagenetic concretionary carbonate observed here and in Middle and Upper Devonian black shale in western New York, the highest modulus rocks exposed in this succession, and therefore the rock that would be carrying the highest tensile stress, is not jointed at all (Figure 22); in fact, these are the rocks that would have been the first to fail by joints driven as a consequence of absolute tension. We are left, then, with the natural hydraulic fracture mechanism in which fluid pressure works against compressive stress thereby creating an effective tensile stress (Lash *et al.*, 2004; Lash & Engelder, 2005, 2007, 2009). Under such conditions, joint orientation is controlled by the remote stress field, whereby the joints open against the minimum horizontal stress, and propagate in the direction of the prevailing maximum horizontal stress.

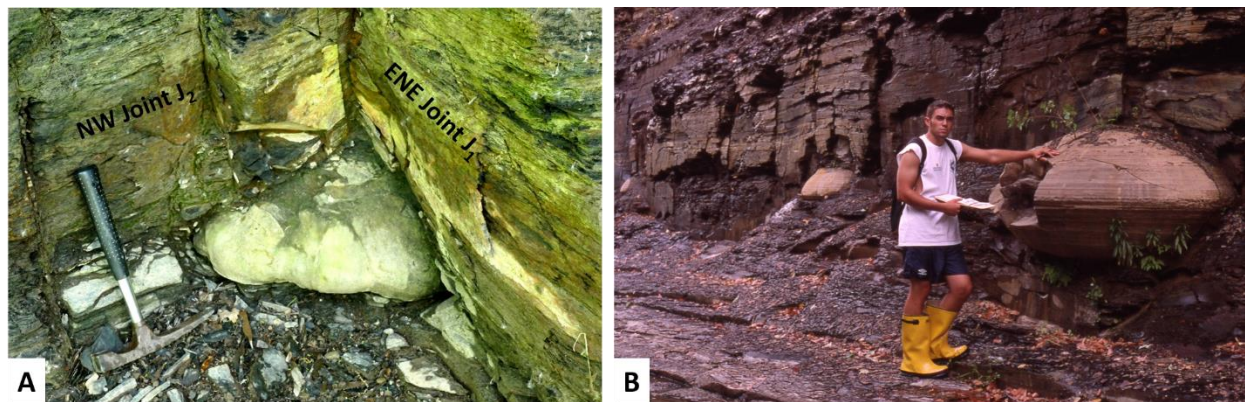


**Figure 22:** Field photographs of joint interactions with carbonate concretions. Note that natural hydraulic fractures do not penetrate concretions; (A) Both ENE and NW joints in the Marcellus Shale do not cleave the concretion, Flint Creek, Phelps, NY (photo credit: Randy Blood); (B) a large NW oriented joint continues in-plane around large carbonate concretions in the Rhinestreet Shale, Eighteenmile Creek, Evans, NY. (Photo credit: Gary Lash)

Black shale-hosted joints occasionally illustrate joint face ornamentation (arrest marks and plumose structures) consistent with natural hydraulic fracturing mechanism (Lash *et al.*, 2004; Figure 23). Moreover, joint density, the inverse of orthogonal joint spacing increases with increasing TOC and thermal maturity. This suggests that the conversion of organic matter to hydrocarbon, is responsible for generating the overpressures necessary to drive these natural hydraulic fractures. Under such conditions, the effective tensile stress achieved is less than that required to break rocks with a higher Young's Modulus, such as concretions, which, under compression, now carry a much higher compressive stress. Thus, we see these joints continuing in plane around such features (Figure 22).



**Figure 23:** Field photograph of a chalked plumose structure ornamenting the surface of an ENE oriented joint in the Rhinestreet Shale, Eighteenmile Creek, Evans, NY (Photo Credit: Gary Lash)

An important outcome of the natural hydraulic fracture mechanism is that jointing must have occurred close to or at peak burial depth. For example, the maximum burial depth of the slightly older, organic-rich Pipe Creek Shale in the town of Silver Creek, NY is  $\sim 2.7$  km (Lash *et al.*, 2004; Lash & Engelder, 2006; Lash & Blood, 2006). Thus, ENE and NW joints are not near-surface fractures as would be joints initiated as a consequence of glacial loading and/or post-glacial unloading (e.g., Clark, 1982; Schmoker & Oscarson, 1995). This has serious bearing on natural gas exploration, especially taking into account the fact that the maximum horizontal stress,  $S_H$ , of the contemporary stress field is parallel to the more densely formed ENE joint set (Lash & Engelder, 2009; Engelder *et al.*, 2009). Zinn *et al.* (2011) demonstrated that hydrocarbon production from horizontal wells increase as wellbore orientations more closely approximate the minimum horizontal stress. Under such conditions, induced hydraulic fractures propagate parallel to the maximum horizontal stress, thereby enhancing the pre-existing ENE joint set, and connecting with the sub perpendicular NW (Lash & Engelder, 2009; Engelder *et al.*, 2009). Such interactions act to increase the surface area of the hydrocarbon reservoir connected to the wellbore, thus increasing well performance.

## Cleveland Shale Member

As noted above, between the Huron and Cleveland members in central and southern Ohio is a southwesterly thinning interval of interlayered gray and black shale beds, designated the Three-lick bed, which is a significant marker unit permitting separation of the thicker bounding units (Kepferle & Roen 1981). It apparently records a falling stage-to-lowstand interval that preceded widespread, transgressive drowning associated with the onset of Cleveland Member deposition.

Although the base of the Cleveland Member is erosional across most of the greater Cleveland metropolitan area in northern Ohio, its basal contact with the Chagrin-Three-lick interval is conformable across central and southern Ohio. The base-Cleveland discontinuity surface “toes out” downslope to continuity, beginning along the Rocky River (Rocky River Reservation) on the west side of Cleveland, Ohio, and it remains conformable southwestward into Kentucky. The Cleveland Shale Member, corresponding to the international *aculeatus* and *costatus* conodont biozones, is the regional expression of the global Dasberg transgressive highstand event (Becker & Hartenfels 2008). As with the underlying Huron Member, it is expressed as fissile black shale with well-developed cyclic ribbing. Easily-accessible long exposures of the Cleveland Member are almost nonexistent in the greater Columbus area with the exception of the lower stream course in the Slate Run Metropark near Lithopolis, Ohio, where the upper 45 feet (14 m) of this division is well displayed. We will not view the Cleveland Member on this excursion.

Recent chemostratigraphic study of the upper part of the Cleveland Shale interval in northern and central Ohio by Martinez *et al.* (in press) shows that it supported a stable microbial community within a variably photic, but redox stratified, water column, which included abundant anoxygenic photoautotrophic green algae (Chlorobi). This metabolic activity resulted in strengthening denitrification and the concentration of ocean-derived rare Earth metals during an episode of marine transgression, timed with a  $\delta^{13}\text{C}$  excursion, believed to be linked to the global Hangenberg Crisis discussed below (Rimmer *et al.* 2010; Martinez *et al.* in press).

From Elyria, Lorain County in northern Ohio, eastward to the Grand River Valley, southwest of Ashtabula in northeast Ohio, the Cleveland Member is unconformably overlain by marine, shell-bearing, gray shale and siltstone deposits of the basal Bedford Formation. From Amherst, Lorain County, southwest into the Columbus area, the Cleveland-Bedford contact is distinctly gradational and conformable, with fissile black Cleveland Shale Member grading upward into soft, gray-greenish basal Bedford mudstone through an interval of up to one foot (33 cm) or more of intermediate, variably dark shale (Baird *et al.* 2018a; Figure 24). This interval yields a sparse marine fauna, best developed in Cuyahoga Valley sections, and represents the last occurrence of a significant shelled fauna of brachiopods, mollusks, and echinoderms in the Appalachian Basin Devonian before marked deterioration of conditions within the basin.

## End-Devonian events in Ohio:

The Bedford and succeeding Berea formations record anomalous and poorly-understood events that took place within the last million-year interval of the Devonian. This interval falls within the time-slice of a complex succession of severe climate changes, including widespread glaciation on Gondwana and elsewhere, coupled with biosphere disruption and extinction, known as the *Hangenberg Crisis* (Kaiser *et al.*, 2016; Becker *et al.* 2016). Major groups to disappear at this

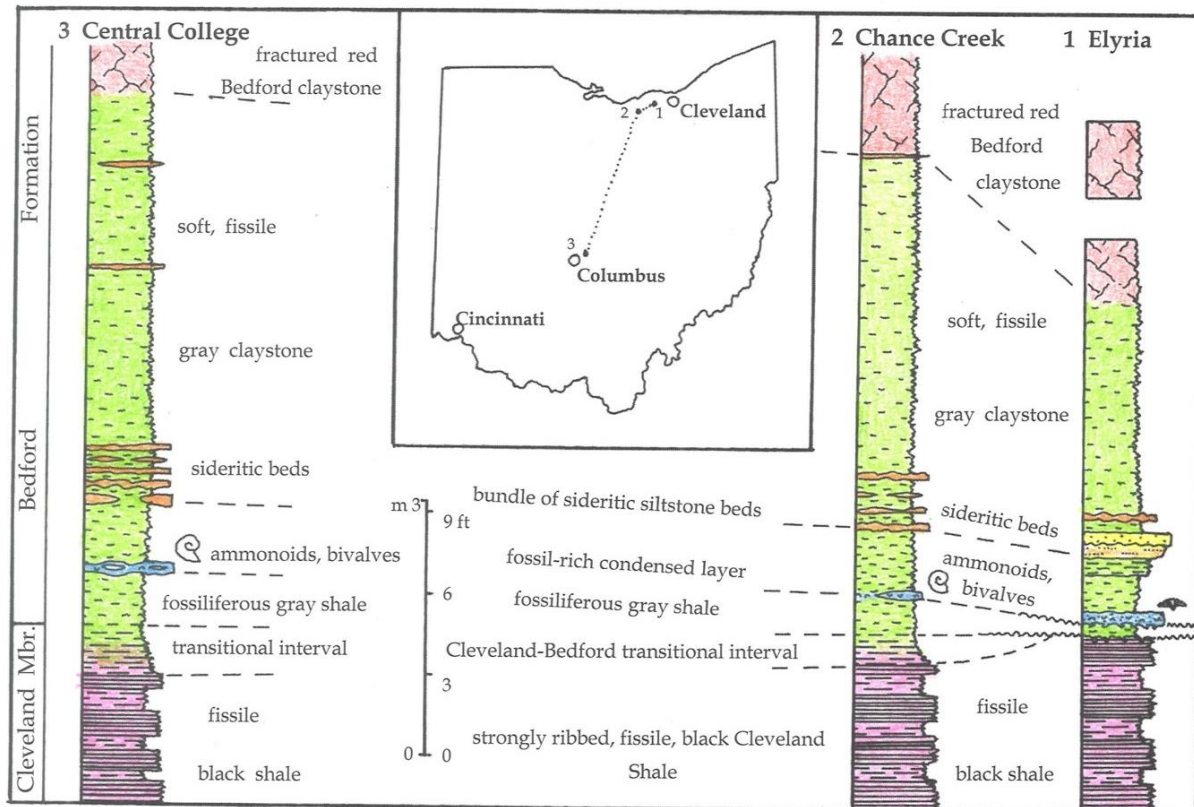


**Figure 24:** Stratigraphy of lower Bedford Shale, above Cleveland Member, from the vicinity of Amherst, Lorain County in northern Ohio to the Central College area at the outskirts of Columbus, Ohio. The base of the Bedford Shale is conformable and gradational from Amherst southwest to at least the vicinity of Lithopolis, southeast of Columbus. The gray, lower Bedford Shale interval grades upward conformably into the disordered bedding of the red Bedford interval. In this region, the lower Bedford is believed to be temporally highly condensed. A key calcareous bed (shown in blue) is known to yield a low-diversity, mollusk-dominated fauna and has yielded zonally important ammonoids belonging to the *Wocklumeria* Stufe, corresponding to the *Ultimus* conodont Zone (House *et al.* 1986).

time, included stromatoporoids, approximately half of the known groups of marine fishes, including the famous arthrodi- res, as well as the dominant polygnathid group of conodonts (Carr & Jackson 2010; Sallan & Coates 2010 ; Kaiser *et al.* 2016 ). The main marine extinction phase may have been triggered by a combination of oceanic salinity overturns and outer shelf

eutrophication events (Kaiser *et al.* 2016). Closely following this event was a major terrestrial die-off of vertebrates and plants may have reduced the terrestrial forests of the mid-late Famennian to the scale of shrubbery by the end of the Devonian (McGhee 2013). Although less well documented globally, partly owing to a lack of agreed-upon zonal standard fossils, the extinctions at this time was nearly of the scale of the better known Frasnian-Famennian Extinction (McGhee *et al.*, 2013).

Globally, the initial (main) extinction event of this biocrisis interval is understood to coincide with deposition of the thin, Hangenberg black shale layer in European and North African sections in an interval of conspicuous conodont turnover ("*costatus* – *kockeli* interregnum" *sensu* Kaiser *et al.* 2009, 2016) roughly correlating to part of the *ultimus* global conodont Zone (formerly Middle-Upper *praesulcata* Zone) of recently revised conodont geochronology (Spalletta *et al.* 2017). In particular, this event is timed with a significant positive  $\delta^{13}\text{C}$  excursion taking place at the level of the Hangenberg black shale unit in several key European sections (Kaiser *et al.* 2016; Becker *et al.* 2016; Figure 2). As presently understood, isotopic shifts to the heavier carbon isotope mark episodes of increased global carbon burial (sequestration) in the deep oceans, leading to a significant draw-down of atmospheric  $\text{CO}_2$  and a climatic shift to "icehouse Earth" conditions.

In Ohio, conodont, palynomorph, and ammonoid data collectively suggest that the Hangenberg biocrisis most likely commenced within the upper part of the Cleveland Shale (Eames 1974; Zagger 1975; House *et al.* 1986; Coleman & Clayton 1987). Moreover, recent discovery of a three ton granite boulder in the topmost beds of the Cleveland Member in Kentucky, appears to be very strong evidence of a dramatic climate shift from "greenhouse" to "icehouse" conditions, allowing for the presence of tidewater glaciers and drifting icebergs in what is now the northern Appalachian Basin region (Ettensohn *et al.* 2008, 2009). The strongest evidence of a sustained positive  $\delta^{13}\text{C}$  excursion in the North American end-Devonian section is observed in the Louisiana Limestone in northeast Missouri and southeast Iowa, where it is roughly coincident with the terminal Devonian *kockeli* conodont biozone at a distinctly higher zonal level (Cramer *et al.* 2008, 2019). As such, this excursion event appears to be timed, not with the main extinction event, but with an episode of initial post-extinction recovery.

However, recent work by Martinez *et al.* (in press) has identified a shift to the heavier carbon isotope in the topmost part of the Cleveland Shale in one of two sections examined in Ohio, tentatively confirming the end-Cleveland Shale as the beginning of the biocrisis story. However, it should be noted that the succeeding Bedford-Berea succession, with its complex and unusual lithologic divisions, may yield evidence of a possible higher  $\delta^{13}\text{C}$  excursion event, based on the findings of Cramer *et al.* (2008, 2019) noted above.

## Deposition of Bedford Formation and origin of the red Bedford

In Ohio, the 90-95 foot (27-29 m)-thick Bedford succession is predominantly shale but is distinctly dominated by an interval of barren red mudstone of enigmatic origin (Figures 24-27). In central Ohio, the Bedford Formation succession consists of a comparably thin, lower, gray-greenish shale interval, followed by a thick, middle division of red-brown mudstone, known as the red Bedford (Figures 24, 25). Above the red Bedford is a conformable change back to gray-greenish shale, which coarsens upward to the conformable base of the Berea Formation (Figure 25). Bedford Formation deposition commences with soft, gray-greenish, variably bioturbated marine shale, which displays a thin, concretionary band in creek sections near Central College in northeast Columbus, and also in sections in Lorain County, Ohio (Figure 24). This bed yields a low-diversity mix of small brachiopods, bivalves, and gastropods in association with the zonally important ammonoid taxon *Prionoceras quadripartitum* Münster, indicative of the very late Famennian Wocklumeria Stufe (House *et al.* 1986). Between Central College and Slate Run Metropark near Lithopolis, this basal Bedford, bioturbated, shell-bearing interval attains a thickness of 1.5 – 4.5 feet (0.5-1.5 m) before passing upward to an 8 to 10-foot (2.4 – 3.0 m)-thick interval of unfossiliferous, variably laminated, fissile-to-platy shale displaying tabular to lenticular sideritic beds below an upward, conformable rapid, facies transition into the red Bedford interval. At Slate Run Metropark, this laminated, gray shale unit is characterized by laterally discontinuous masses of black shale that appear to have been eroded from-, or sheared off from, the top-Cleveland Member contact. The upward change from black Cleveland Shale into gray Bedford lithology marks a major regression, presently understood to correlate to a change to pronounced global “icehouse” conditions and associated sea level-drop, during, or immediately following, the initial (main) Hangenberg extinction event (Ettensohn *et al.* 2009; Brezinski *et al.* 2010; Figure 2).

The basal Bedford bioturbated, gray shale unit is, herein, interpreted to be the highly condensed, basinward correlative of a much-thicker, progradational, clinoform wedge of silty shale and sandstone units that comprises the entire Bedford interval from eastern Cuyahoga County eastward into western Ashtabula County (Baird *et al.* 2013; Figure 26). Units within this wedge are observed to thin dramatically northwestward and downlap toward the Cleveland Member-Bedford Formation contact in the vicinity of the Cuyahoga Valley (Baird *et al.* 2013). This thinning of the lower Bedford is coincident with rapid westward-northwestward emergence of younger red Bedford strata below the base-Berea disconformity (Figure 26). In this same area, the base of the Bedford Formation is notably erosional with significant southward beveling of Cleveland Member strata in the Cuyahoga Valley (Baird *et al.* 2013).

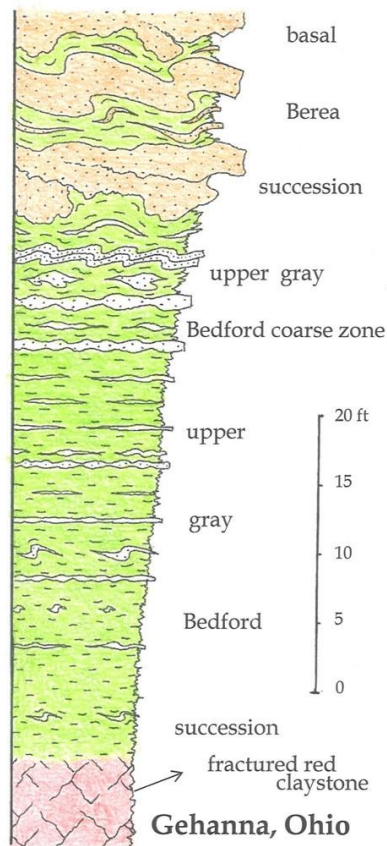


**Figure 25:** Generalized stratigraphy of the uppermost part of the Bedford Shale from the top of the red Bedford mudstone interval into the lower part of the Berea Sandstone as seen in central Ohio (Sunbury-Columbus area) sections. Note the regressive, upward-coarsening of the gray Bedford interval, which conformably grades into the siltstone-sandstone-dominated Berea succession. This is the reverse trend to what is shown in figure 30, suggesting that the red Bedford records a major shut-down in coarser sediment supply the basin.

However, what is most counterintuitive about this downlap pattern is that one would predict that the base-Bedford unconformity would extend into the region west of Cleveland, particularly, if the wedging process involved some measure of downlap of basal Bedford strata onto the regional unconformity surface. Instead, the base-Bedford unconformity is observed to close westward to continuity, beginning in the vicinity of Amherst in Lorain County (Figure 24). This implies that the basal part of the Bedford in this area is both a very greatly condensed, temporal, “feather-edge” equivalent of the Bedford clinoform wedge in northeast Ohio and a depositionally continuous sedimentary interval at the same time. That basal Bedford sections at Amherst and near the Vermilion River, in Erie County, Ohio are both essentially bed-for-bed identical to the above-mentioned basal Bedford section near Central College in northeast Columbus, across an intervening distance of 70-80 miles (113-127 km), is consistent with the

concept that the basal Bedford across central Ohio is some type of offshore, condensed, non-black, and low-energy, basinal deposit (Figure 24).

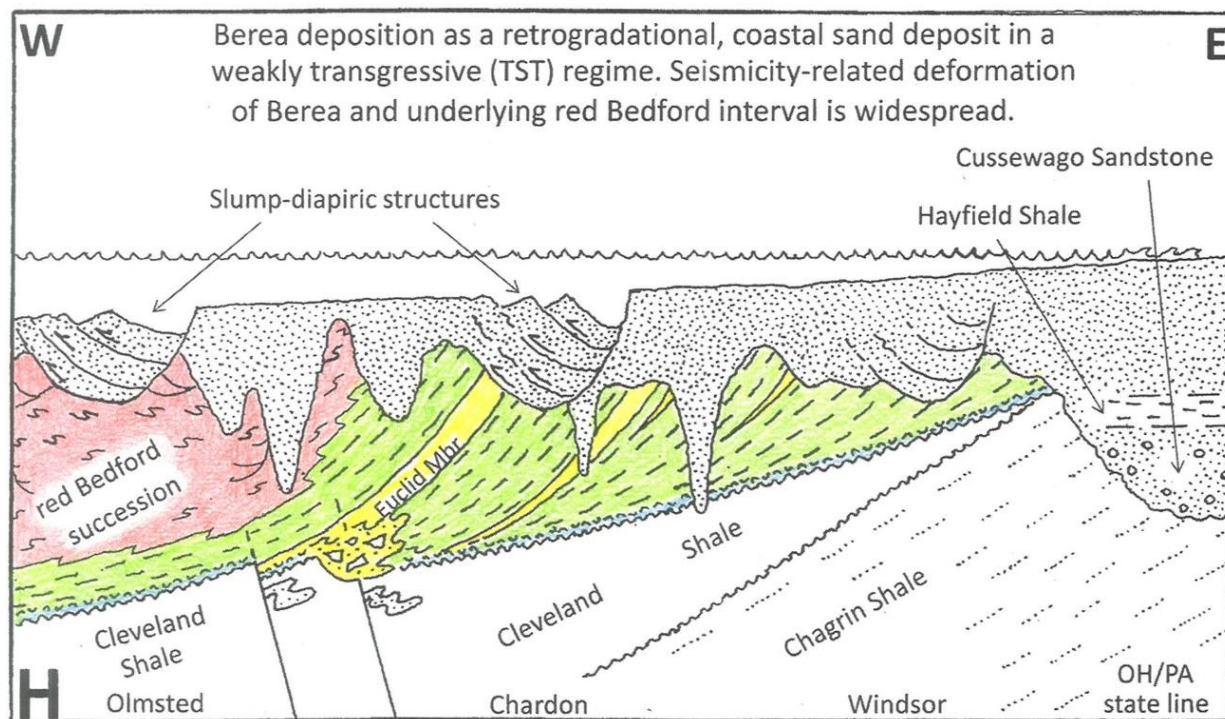


**Figure 26:** East-west stratigraphic transect from the OH/PA state line to the west side of Cleveland, Ohio, showing inferred relationships of the Chagrin Shale-through-Berea Sandstone succession at the time of Berea deposition (Bedford strata are colored for emphasis). The lower part of the Bedford Shale in northeast Ohio is expressed as a major clinoform structure reflecting west-northwest-directed progradation. The prominent undulatory (channeled) disconformity flooring the Cussewago – Berea succession is now interpreted, in large part, to be the record of glaciogenic lowstand forcing, linked to a phase of Hangenberg climate crises (Brezinski *et al.* 2008, 2010). The red Bedford, emerging westward below the disconformity, comprises most of the Bedford interval across much of central Ohio. Note complex patterns of sediment disturbance (slumping, diapirism, sediment injections), which appear to reflect far-field tectonics at this time. Unpublished Baird schematic.

The red Bedford interval in the Columbus area includes a long, 60-65 foot (18-19.8 m) – thick succession of red-brown mudstone, generally lacking organized bedding and discrete, tabular layers (Figure 27 A, B). As we will see at STOP 5 in the Blendon Woods Metropark, the red Bedford is a unique and dramatic deposit (Figure 27 A), which appears to be profoundly lacking in fossil content or other features that might help link it to a particular paleoenvironment. To make matters worse, it displays a pervasive overprint of slickensided microfractures (Figure 27 A), small-scale sediment loading features, and deranged dip directions that is characteristic of red Bedford sections from the Cuyahoga Valley southwestward to the vicinity of Circleville in southern Ohio, where this facies begins to pass southward into more structurally coherent, silty, gray equivalent deposits (Hyde, 1953; Pepper *et al.* 1954; Pashin & Ettensohn 1995). The red Bedford appears to be a monolithic mudstone unit that may locally exceed a thickness of

150 feet (46 m) in the subsurface southwest of Cleveland (Pashin & Ettensohn 1995). In sections, it weathers to irregular, earthy fragments and to very slippery wet clay on slopes.



**Figure 27:** (A) Red Bedford overlain by Berea Sandstone along West Creek at the Snow Road overpass, Brooklyn Heights, OH, and (B) hand specimen showing lamination occurring in the red Bedford. (Photo Credit: Gordon Baird)

Pepper *et al.* (1954) interpreted the red Bedford interval to be a non-marine deposit comparable to the alluvial Catskill Formation in New York, Pennsylvania, and West Virginia. In that report, he reconstructed the red Bedford domain to be the alluvial surface of a long, narrow delta extending southward from Canada, across Ohio, into northeastern Kentucky. However, Kohout & Malcuit (1969) interpreted the red Bedford to have been deposited on a “wind-blown tidal flat”. Still later, Lewis (1988) and Pashin & Ettensohn (1995) interpreted this

unit as an offshore marine facies, owing to the presence of peripherally developed graded siltstone beds, interpreted as turbidites. In particular, the absence of biogenic structures and subaerial exposure indices (desiccation cracks, calcretes, evaporate deposits, or root traces) seems to rule out an emergent, alluvial origin for this facies. Pashin & Etensohn (1995) reinterpreted the red Bedford facies to be an offshore hemipelagic sediment deposit that accumulated in a deeper-water, prodelta setting to the west of a prograding delta system (their “Cussewago delta front”).

If the red Bedford is some sort of offshore pelagic facies, it is certainly very different from the basal greenish-gray and black shale deposits of the earlier Devonian succession. The red color suggests substrate oxygenation, not oxygen deficiency, but the absence of shelly macrofauna is problematic. As such, the red Bedford is a fundamentally enigmatic unit, which appears to record deposition in a highly-stressed, restricted setting. It is curious that its deposition is timed with some portion of the Hangenberg extinction story. This begs the speculative question of whether the red Bedford interval, as well as the peripheral barren shale-siltstone deposits bordering it, are an oligotrophic “death mask” signature of ecosystem collapse? If the red Bedford is, indeed, an aquatic, water-laid deposit, it should yield microfaunal content to shed light on what happened; to this end, the present authors are currently bulk sampling the red Bedford at multiple localities and levels to search for microfossils (conodonts, ostracodes, radiolaria, or other microfossil remains), to determine what taxa, if any, are to be found.

Above the red Bedford interval is a gradational return to gray, upper Bedford shale deposits that are structurally intact and characterized by tabular and lenticular siltstone layers, which become increasingly numerous and thicker in an upward direction. In northern Ohio sections, the red Bedford is directly overlain by the base-Berea disconformity, but in central and southern Ohio, the base-Berea unconformity closes to conformity, and distinctly higher, non-red Bedford layers appear in the subsurface and in outcrops. These strata are best displayed along Rocky Fork in Gehanna, but are most accessible in Blendon Woods Metropark, where we will see this interval at STOP 6. At Blendon Woods and Rocky Fork, this interval is notable for an abundance of flaggy siltstone layers and lentils that display ripple marks that give rise to the name “Ripple Rock Trail” for the park path leading to STOP 6 (Figure 28). Both Hyde (1911) and Pashin & Etensohn (1995) examined wave ripples in the Bedford-Berea interval and determined that they display a consistent northwest orientation across much of Ohio. More specifically, Pashin & Etensohn (1995) determined that ripples in the Bedford gray facies display a vector mean azimuth range between 284° and 316°. These ripples are interpreted as being storm wave-generated features produced between the storm wave depth limit and fair-weather wave-base (Pashin & Etensohn 1995).



**Figure 28:** Field photograph of siltstone layer displaying ripples in the Bedford shale exposed along the creek at Blendon Woods Metropark, Columbus, OH. (Photo Credit: Randy Blood)

Along with the rippled surfaces, are numerous trace fossils, mainly encountered as float along the stream within the metropark. These traces, preserved as hypichnial casts on the bases of many siltstone beds (Figure 29 A, B), as well as delicate additional traces, preserved under thin siltstone lentils associated with numerous wave-ripple events, include unusual types generally not encountered in normal marine Devonian strata, and they will, undoubtedly, provide important paleoenvironmental clues relating to this little-studied interval. Of key significance is the potential presence of shelly fossils (brachiopods, mollusks) that may be present in this bioturbated zone. Given that this section was only recently encountered by the present authors in August of this year, it is the subject of ongoing examination.



**Figure 29:** (A) *Chondrites* on the underside of a siltstone bed, and (B) unidentified trace fossil. Both samples occur in the upper gray shale division of the Bedford Shale at Blendon Woods, Columbus, OH (Photo credit: Randy Blood)

It is also significant that there is vertical lithofacies symmetry centered on the red Bedford, as noted by Baird *et al.* (2018a, b); in northern Ohio, fossiliferous basal Bedford strata pass upward into barren, thick, siltstone beds of the Euclid Member, before grading first to barren gray shale before passing into the red Bedford interval (Figure 30). In Lorain and Erie counties of northern Ohio, the lower Bedford displays an identical succession to that of Columbus sections, although the top-Bedford interval is absent there due to erosional truncation. In the Columbus area, the basal Bedford, bioturbated, gray shale division rapidly passes upward into barren, laminated, gray shale and siderite before grading into the red Bedford succession.

However, in the Columbus area, the red Bedford passes conformably upward into barren gray shale before changing into the, still higher, silty, bioturbated facies seen at STOP 6 (Figure 25). Hence, this transition is the reverse of that observed in the lower Bedford succession (compare figures 25 and 30). However, the red Bedford is interpreted, it is evident that silt and sand-supply to the basin was clearly switched off during its deposition.

The red Bedford is known to pass southward into variably silty, non-red shales in the vicinity of Chillicothe, Ohio (Hyde 1953, Pashin & Etensohn 1995), though this part of the story needs to be better examined by the present authors. If this pattern is correct, it would indicate that the red Bedford is some form of restricted, offshore facies, which passes upward via regression into the silt-sand-dominated lower Berea succession, recording a change to shallow, paralic-marine lowstand conditions.

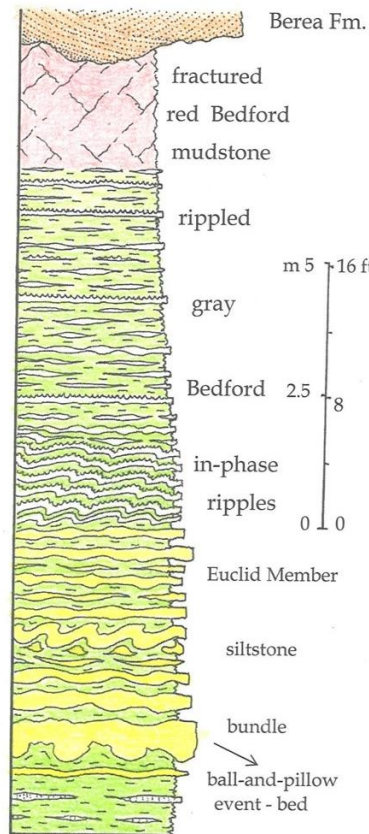


**Figure 30:** Generalized stratigraphy of the lower part of the Bedford Shale from the Euclid sandstone/siltstone member, upward to the base-Berea disconformity in the Cuyahoga Valley region south of Cleveland, Ohio. Note the key (transgressive?) upward-fining succession within the lower gray Bedford interval from the Euclid into the topmost gray shale strata that underlie the change to red mudstone. Note also the absence of siltstone beds and presence of disordered bedding within the overlying red Bedford interval.

### Berea Formation depositional and tectonic story

In northern Ohio, the Berea Formation is a very prominent and well-studied sandstone division that is exposed in numerous creek and quarry sections. It is floored by a disconformity that displays local relief and is internally characterized by spatially complex paleovalley-fill patterns, cross-stratification, and sediment-deformation (Pepper *et al.* 1954; Lewis 1988; Pashin & Ettensohn 1995; Figures 2, 31). Pashin & Ettensohn 1995 interpreted the overall Bedford-Berea succession to record an initial constructional-progradational phase, represented by the lower Bedford clinoform succession, noted above, followed by a destructional, sediment-reworking phase, associated with marine transgression during Berea time. A key feature to their model was the Cussewago Sandstone in northeast Ohio and northwest Pennsylvania, which connects eastward to the subsurface Murrys ville Sandstone under Pennsylvania. They interpreted this subsurface unit to be a transgressive paleochannel-fill, connecting the Berea sediment complex to sediment sources in the Acadian orogen. The regional, undulatory disconformity, widely

observed under the Berea Sandstone in northern Ohio (Figures 2, 31), has been attributed partly to flexural tectonic affects peripheral to Neo Acadian collisions as well as to global regression associated with the expansion of glaciers on Gondwana (Pashin & Ettensohn, 1995; Ettensohn *et al.*, 2009).



**Figure 31:** Field photograph of giant ball-and-pillow structures in the Berea Sandstone exposed along Big Walnut Creek, Sunbury, OH. (Photo credit: Carl Brett)

Subsequently, Brezinski *et al.* (2008, 2010), on the basis of extensive study of end-Devonian diamictite deposits (non-marine Spechty Kopf Formation) and correlative sandy-gravelly units within and bordering the Appalachian fold belt in eastern Pennsylvania, Maryland, and west Virginia, determined that these units were, in part, the record of glacier-related, erosional, and depositional processes, as well as periglacial conditions, timed with Hangenberg-age “icehouse world” events. This glacial lowstand maximum is understood to define the axial trough of the Murrysville-Cussewago channel, and subsequent channel-fillings as well as the succeeding Berea Sandstone, are believed to record a step-wise return to a “post-icehouse” world. Given the aforementioned earlier occurrence of the granitic lonestone in the topmost Cleveland Shale, noted above (Ettensohn *et al.* 2009), the inferred glacial lowstand inferred for the base-

Berea disconformity appears to be the second of, at least, two climatic “icehouse” events recorded near the end of the Devonian (see Figure 2A).

In the Columbus-Sunbury area, the Berea Formation is distinctly finer grained than it is in northern Ohio, ranging from fine sandstone to siltstone across the field trip area. Pashin & Ettensohn (1995) interpreted the Berea to be comprised of a thick succession of turbidites and storm layers that sourced from deltaic complexes (Gay-Fink and Cabin Creek trends) in central West Virginia. In addition, the base-Berea disconformity is believed to be of greatly reduced magnitude or absent in parts of south-central Ohio. Similarly, the Berea of southern Ohio is understood to be of a more distal marine character, though still lacking in body fossils (Pashin & Ettensohn 1995). The position of the inferred glacial lowstand event may be within the lower part of the Berea succession, but is obscured by large-scale sediment-loading features, which we will see at STOP 4 (Figure 31).

At Slate Run Metropark (Figure 1; STOP 7), southwest of Lithopolis, Ohio, we will, again, examine the Berea in context with the succeeding Mississippian Sunbury submember. Unlike the Berea outcrop at STOP 4, this Berea section is extremely thin and highly variable in thickness, ranging from one foot (33 cm) to 4.5 feet (1.4 m) across a short walking distance upstream from the park path bridge. Moreover, on a nearby stream tributary, 0.25 mile (0.4 km) distant, an incomplete Berea section was found to be 6.5 feet (2 m)-thick. A clue to this thickness variation is provided by the knife-sharp, base-Berea unconformity, flooring thin Berea deposits in sections from Lithopolis southward to A. W. Marion State Park near Circleville, Ohio. Evidently, between Columbus and the Lithopolis – Circleville region, the Berea Formation thins drastically southward and redevelops a basal unconformity, which may be present in certain sections southward to the Ohio River (Hyde 1953; Pashin & Ettensohn 1995). The absence of topmost-Bedford siltstone layers in this area and differentially thin upper Bedford gray shale deposits, suggest that this region was differentially subjected to uplift and erosion relative to areas farther north.

### **Base-Mississippian drowning event**

The Sunbury submember is an organic-rich, black, fissile to sheety shale unit of Lower Mississippian (Tournaisian) age, that records a major transgressive deepening event, particularly, in the western and southwestern parts of the Appalachian Basin (Ettensohn *et al.* 2009). The Sunbury submember is understood to be equivalent, in part, to the much thicker and less organic-rich Orangeville Shale Member in northern Ohio (Figure 2). The Sunbury Shale yields conodonts indicative of basal Mississippian age (Hass, 1947). On the basis of miospore content, the base of the Sunbury Shale in Kentucky falls in the VI biozone, marking the base of the Mississippian system (Coleman & Clayton, 1987; Pashin & Ettensohn, 1992, 1995 Figure 2 B). At STOP 7, the Berea-Sunbury contact is expressed as an abrupt upward change, through an

8-inch (20 cm)-thick, interval of burrowed and rippled, thin siltstone beds, into the higher, 30-foot (9 m)-thick, black Sunbury succession. The lowest black shale layers, both within and above this silty transitional interval, are relatively condensed, and they are known to yield phosphatic fossils, including: the linguloid brachiopod *Barroisella*, orbiculoids, and less common conulariids.

According to Ettensohn *et al.* (2009), the Sunbury record a major flexural drowning event associated with Ettensohn's "fourth tectophase" event involving thrust loading of the North American plate by the incoming Carolina terrane as a key feature of his "Neoacadian orogenic event". Cratonic loading during Sunbury deposition is envisioned to have produced inferred water depths of as much as 1000 feet (300 + m) for Kentucky sections according to Ettensohn *et al.* (2009).

### **End-Devonian sediment loading and tectonic disturbances ("the Ohio orogeny")**

The end-Devonian interval, encompassing deposits from the upper part of the Cleveland Shale Member through the middle part of the Berea Formation, shows evidence of a succession of sediment-loading and diapiric submarine slump events that apparently involved, to varying degrees, fault motion and seismicity (Wells *et al.* 1991; Pashin & Ettensohn 1995; Figures 31, 32).

The first defined phase of significant sediment loading and associated diapirism occurred during deposition of the Euclid Sandstone member of the lower Bedford Formation in northern Ohio (Pashin & Ettensohn 1995; Hannibal *et al.* 2012; Baird *et al.* 2013). However, pervasive, large-scale diapirism took place during a second phase of widespread rapid sediment-influx associated with the onset of Berea deposition across Ohio and northernmost Kentucky (Figure 32). Vertical movement of variably sandy sediment, locally on the order of hundreds of feet, occurred pervasively, particularly, across north-central and central Ohio, perhaps controlled by reactivated fault zones (Pashin & Ettensohn 1995). Concurrent with this sediment failure was depositional filling of developing depressions with thick deposits of economically important, cross-bedded sandstone ("quarrystone") that locally exceed thicknesses of 200 feet (61 m) in Lorain County, Ohio (Figure 32). This event, apparently coinciding mainly with deposition of the lower Berea succession, may have something to do with accelerated sediment influxes associated with increased mountain erosion during inferred glaciation episodes. The giant ball-and-pillow features, seen at Sunbury, Ohio (STOP 4; Figure 31) are roughly coincident with this zone of maximal Berea sediment-failure observed elsewhere in Ohio.

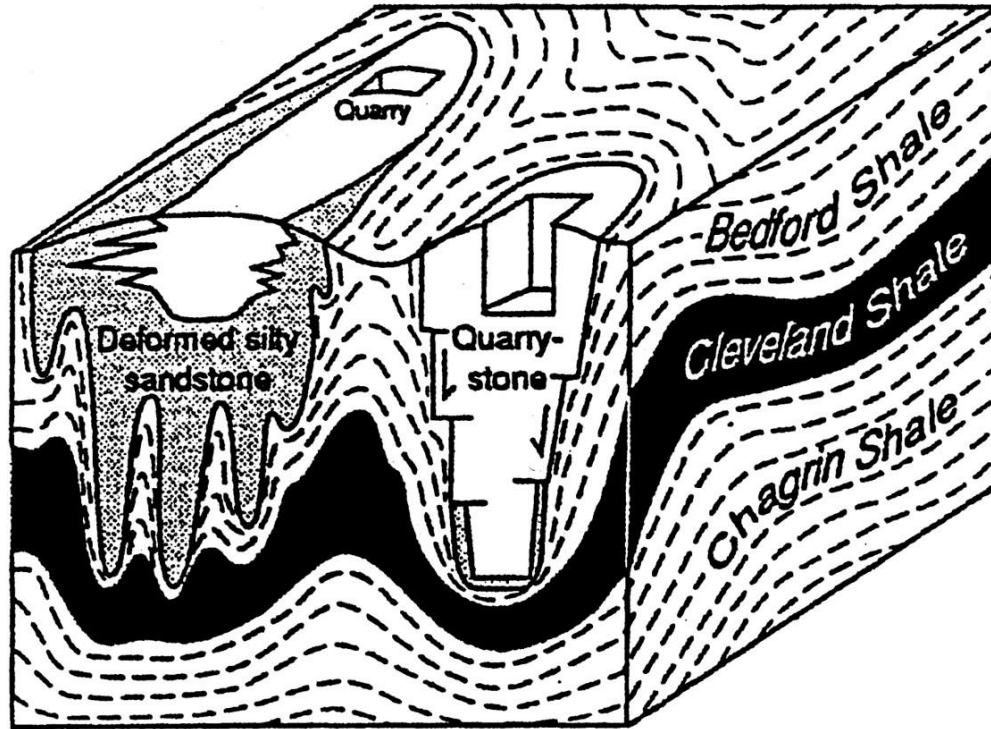


**Figure 32:** Schematic illustration of very large, deformed sandstone masses within the lower part of the Berea Formation in the Lorain-Erie county region west of Cleveland. As interpreted by Pashin & Ettensohn (1995), these features resulted from rapid sediment influx and loading under the influence of adjacent faults. The sandstone bodies may have started out as bars or fluvial bodies, but gradually subsided into the underlying Bedford as fault-bounded masses. Deep sandstone quarries were developed into variably subsided, cross-bedded “quarrystone” deposits up to 250 ft. (75 m) in thickness. Numerous Ohio sections display variable evidence of load subsidence and related diapiric movement associated with the Berea Formation. Image from Pashin & Ettensohn (1995; Figure 29, p. 30).

A remaining deformation phase, which may be only partly connected to Euclid and Berea diapiric events, is that of the spatially complex microdeformation observed throughout the red Bedford interval (Figures 25, 27, 30). One might argue that the microdeformation of the red mudstone is peripherally related to displacement effects associated with Berea sediment-loading, somewhat in the manner of load-reshaping of a water bed. However, the intervening upper gray Bedford interval appears to lack the pervasive shear planes visible in the underlying red mudstone interval, suggesting a process disconnect between the episodes of red Bedford and Berea alteration. The underlying lower Bedford gray shale interval is generally less sheared than the overlying red Bedford, but it does display inclined thrust horses near its base where the topmost Cleveland Shale Member is observed to penetrate obliquely upward into the basal Bedford in certain sections (Hyde 1953; Pashin & Ettensohn 1995; Ettensohn *et al.* 2009; Baird pers. obs.). As such, the close-spacing of microfracture and shear planes, perhaps present throughout the spatial volume of the red Bedford in Ohio sections, begs explanation.

Is the deformation of the red Bedford some form of far-field effect of a transpressive terrane collision in the cratonic heartland? A recent structural study of the top-Cleveland-into-base-

Mississippian interval in north-central Ohio suggests that the Bedford Formation was subjected to Late Paleozoic décollement-style thrusting, which created imbricated folds within incompetent Bedford shale beneath a more competent mobile Berea slice (Fakhari *et al.* 2019).

One final factor is the possibility that the red Bedford microstructure was a post-depositional feature generated intrastratally, and that the red-brown color itself is a secondary feature brought on by shear-related hydrothermal alteration or rock volume changes during deformation?

### **Discussion/remaining questions:**

Several tasks remaining include a need for refined correlation of fossil zonation and geochemical proxies (trace element, magnetic susceptibility, secular carbon-oxygen isotope variations) for identification of stratigraphic signatures of Middle and Late Devonian paleoclimatic and extinction events.

One specific issue is the origin of ultra-thin and repetitive thin black shale bands within the Upper Olentangy Shale and *Protosalvinia* Zone of the Huron Member. Do these strictly represent very brief episodes of water column eutrophication, time-rich water mass stratification events, or downslope transport (turbiditic pulses) of allodapic, organic-rich, sandy sediment into a non-black sediment succession?

What is the temporal scale and meaning of the 6<sup>th</sup>-order alternations of organic-rich and organic-poor sediment layering observed within the Huron and Cleveland Shale members? Can a lower (fifth-order) bundling of these ribs be identified in long shale sections, and can the bands be counted and quantifiably correlated to developing cyclostratigraphic time-scales?

Ongoing work is directed to obtaining datable conodonts and palynomorphs from the base of the Cleveland Shale Member. Evidence to date supports the concept of diachronous, transgressive onlap of basal Cleveland deposits over a regional unconformity in northern Ohio. Refinement of physical correlation of sections, coupled with zonal and/or geochemical event-controls, should confirm that this drowning event is the regional signature of the global Dasberg transgression.

Correlation of key Bedford and Berea Formation divisions and contacts is being extended southwestward across Ohio into Kentucky in an effort to evaluate how these units connect to the marine realm. Given that the deposition of these units is timed with disruptive events (ecological overturn and glaciations) associated with the end-Devonian Hangenberg crisis, it is critical to look for signature clues as to what actually happened in this region. Unlike the thin, stratigraphically condensed European and North African Hangenberg-age sections so far

studied, the Bedford-Berea section in Ohio is comparably thick, lithologically diverse, and dramatically different from the underlying Devonian rock succession.

In particular, what is the origin of the sheared red Bedford succession? Presently, we are bulk sampling this macrofaunally “barren” unit and peripheral bounding facies for the possibility of microfauna being present. Is this unit a “death mask” signature of some part of the Hangenberg event story, or is the red color intimately linked to structural deformation and/or hydrothermal alteration with associated rock volume changes. This structural geology dimension should include regional examination of the nature and timing of structural shearing of the red Bedford interval, as well as efforts to reconstruct the clay mineralogical history within the entire Bedford Formation interval.

### **Acknowledgements**

The authors are indebted to the Columbus and Franklin County Metro Parks Agency; Preservation Parks of Delaware County, Phil Smith, Head Ranger at Camp Lazarus and Dr. Mohler, resident in Sunbury, OH for access to their property. Without their cooperation, this field trip would not be possible. We thank the Paleobotany Section at the Cleveland Museum of Natural History for use of the *Protosalvinia* specimen. We also extend our gratitude to the Ohio Department of Natural Resources; Division of Geological Survey for providing us with the photograph of Alum Creek prior to construction of the dam and for helping to make this trip possible. Finally, many thanks are owed to John Miller and Mac Swinford of the 2019 Eastern Section AAPG Field Trip Committee for inviting us to run this field trip and taking care of all the logistics.

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## STOP DESCRIPTIONS:

**STOP 1: Lowest strata of the Delaware Formation succeeded by Lower Olentangy Shale with a stop at the lower/upper Olentangy Shale contact (bone bed), exposed along the east bank of the Olentangy River and lower part of Welsh Run (Camp Lazarus property) south of Delaware, Ohio:**

*Stratford Member of lowest Delaware Formation displayed in west-facing bank of Olentangy River and adjoining mouth of Welsh Run downstream from Chapman Road on the Powell 7.5' Quadrangle at Google coordinates 40.233253, -83.061113; top-Delaware - Lower Olentangy Shale disconformable contact upstream at approximate coordinates 40.232364, -83.058264; Lower – Upper Olentangy disconformable contact further upstream at coordinates 40.232635, -83.056276.*

*Proceed from vehicles along Chapman Road down a bank bordering road to Olentangy River edge at mouth of creek. Continue north for about 60 – 90 feet to clean, west-facing limestone exposure:*

This bank displays both the Middle Devonian Stratford Member and the succeeding Dublin Shale Member of the Delaware Formation (Figure 4). The underlying Columbus Limestone (corresponding mainly to the Eifelian age *costatus* Zone) is usually exposed only at low water at this place. Here, the lowest Delaware beds, informally named the Stratford Member by DeSantis *et al.* (2007), are a possible equivalent to the (*australis* Zone) Venice Member of northern Ohio. The Stratford is rich in fossils, particularly, clusters of the small spiriferid brachiopod *Brevispirifer lucasensis*, as well as additional brachiopod taxa: *Leptaena*, *Warrenella maia*, *Cherryvalleyrostrum*, and chonetids. Poorly preserved specimens of large coiled nautiloids (*Gyroceras* sp.) are present, as are large, flattened, white-weathering specimens of *Tentaculites scalariformis*.

Higher beds, exposed in the bank, show an upward transition into dark brownish gray, thin-bedded, argillaceous limestones of the Dublin Member (Figures 4, 5). Thicker beds near the base of this transition display abundant *Zoophycos* spreiten and tentaculitids. Above this level, the Dublin Shale yields only small brachiopods (the rhynchonellide *Cherryvalleyrostrum* and orbiculoids; Figure 5). Farther south at Slate Run in Upper Arlington, bordering Columbus, Ohio, the Dublin Shale is maximally developed and is distinctly characterized by platy petroliferous shales and black chert beds. It is particularly significant that the Dublin Member is the cratonward expression of the largely black Bakoven Member of the lower Marcellus Subgroup in New York (Brett *et al.* 2011, Ver Straeten 2007). Along this stream it is distinctly dark gray, but it is shaly and yields a “leiorhynchid” brachiopod fauna, so very characteristic of producing Marcellus deposits (Figure 5). This far west, the black shale facies, characteristic of the central

Appalachian Basin is beginning to pass to petroliferous dark limestone before passing into shell-rich, neritic facies (Rogers City, Spillville, and Wise Lake formations) in the Michigan and Iowa basins (Brett *et al.* 2011).

*Return to mouth of Welsh Creek and proceed eastward under Chapman Road up the creek. We will follow this stream to the base of the Upper Olentangy Shale succession:*

The same strata, visible in the river bank, will be crossed as we ascend the stepped falls upstream from the bridge (Figure 8). Hard, cherty limestone layers at the top of the falls yield an abundance of the brachiopod *Leptaena*. Upstream from the falls, the stream runs at a low gradient, gradually exposing a few ledges of limestone. A low, riffled-forming ledge, about 300 feet upstream of the falls, yield an abundance of the distinctive button coral *Hadrophyllum d'orbigny*, minor crinoid and comminuted bone debris, and a few brachiopods. This is Stauffer's "Zone L". In other nearby outcrops, grainstones rich in brachiopods, including: *Rhipidomella*, *Pseudoatrypa*, and *Strophodonta demissa*, as well as, larger rugosan and tabulate corals, informally termed the Lewis Center submember by DeSantis *et al.* (2007), occur slightly below this level. The erosive base of this bed may mark the sequence boundary at the base of the Oatka Creek Formation in New York State, i.e., the approximate position of the Hurley Member of Late Eifelian age. This bed appears to be high in the *T. kockelianus* conodont Zone and may represent the lateral equivalent of the Hurley Member and Cherry Valley Limestone in New York, as evidenced by similar conodont assemblages and magnetic susceptibility patterns (Travis *et al.*, 2009; Over *et al.*, 2019). The Eifelian-Givetian boundary was picked slightly above this level by Travis *et al.* (2009) on the basis of magnetic susceptibility.

The highest Delaware beds, exposed in the stream bed for a short distance above this level, include cherty dolomitic limestones with sparse fossils that may be equivalent to the Silver Creek Member of the North Vernon Limestone in Indiana. Rare crinoids and spiriferid brachiopods have been obtained from these beds (comprising Stauffer's (Zone M"). A sharply-defined corrosion surface separates these limestones from the overlying Lower Olentangy Shale. This contact, along with the succeeding, non-resistant Lower Olentangy succession, are poorly exposed along this creek.

*From this point, we depart the stream and proceed north to the dirt road paralleling the creek and continue walking eastward on that road to an ingress path leading southward to a place where the disconformity separating the Lower and Upper Olentangy Shale can be viewed.*

Descending again to creek level, we observe the nonresistant Olentangy Shale in a northeast-facing bank. This rock unit is unusual in that it is a hybrid division, marked by an internal disconformity with hiatal magnitude of 8-10 million years! Below this contact, the underlying Lower Olentangy Shale is Middle Devonian (Givetian) in age and correlates to the Plum Brook

Shale in northern Ohio, which has yielded conodonts indicative of an “upper *ensensis*” Zone (upper *hemiansatus*-into-lower *timorensis* zonal interval of European terminology; see Sparling 1995). This roughly equates the Lower Olentangy to part of the Skaneateles Formation of the Hamilton Group in New York State (Brett *et al.* 2011). By contrast, the Upper Olentangy Shale above the contact yields conodonts of the high Frasnian international MN Zone 13 (*Pa. triangularis* Zone), which places this unit as an approximate equivalent of the Hanover Formation in the Java Group succession in New York (Rickard 1975; Over and Rhodes, 2000).

Yet to the undiscerning eye, this mudstone bank appears nearly homogenous, which led to the original lumping of the two shale divisions. This disconformity is cryptic, partly for the lithologic similarity of these juxtaposed units, but also for the sediment-mixing action of marine infauna, which blurred and erased the contact at the time of its burial. Such “stratomatic” contacts (*sensu* Baird 1978, 1981) are important features in shale-dominated sedimentary basins and are particularly easy to miss in core sections. Their recognition is crucial in exploration-related basin analysis.

Upon excavating the disconformity interval, you will notice numerous dark-brownish, phosphatic bioclasts in association with diagenetic pyritic masses and phosphatic nodules. The smallest bioclasts are zonally important conodonts, which have been jumbled into a zonal aggregate through combined effects of long-term submarine erosion and sediment mixing. Larger objects include fish teeth, scales, and dermal armor. One of us (Baird) found a 6-inch diameter arthrodire plate at this locality in the 1980s.

Higher beds of the Upper Olentangy Shale succession are poorly exposed along this stream. We will see parts of the Upper Olentangy Shale again at Stops 2 and 3.

*Return to vehicles via the road bordering creek. Vehicles will return south to east-west Winter Road. Turn left on Winter Road and continue 1.5 miles east to U. S. Route 23. Turn right onto Route 23 and continue for 1.7 miles to sign on the right for Shale Hollow Park. Turn right (west) onto Olentangy Crossings West and continue for about 0.3 mile. Sign for Shale Hollow Park by entrance road to park. Turn right (north) onto this entrance road and proceed through winding turns to parking area at Google coordinates 40.206368, -83.038194.*

## **STOP 2: Upper Olentangy and Huron formations at Shale Hollow Park west of Lewis Center, Ohio**

*Exposures of the Givetian Lower Olentangy Shale and Frasnian Upper Olentangy Shale, succeeded by the higher Famennian-age Huron Member along southwest-flowing “Shale Hollow Creek” 1.7 miles northwest of Lewis Center, Ohio on the Powell 7.5’ Quadrangle. We will examine the medial and upper parts of the Upper Olentangy Shale at Google coordinates*

*40.202749, -83.041060, the base-Huron Member contact at 40.203943, -83.039427, and large concretions in upstream areas both within and bordering the park.*

*Exit vehicles at parking area near park headquarters with restroom facilities. Proceed northwest across bridge over creek into park area for brief lecture. We will first proceed downstream past the public area to our farthest stopping point to view the Upper Olentangy Shale, followed sequentially by later stops to view the Upper Olentangy – base-Huron Member contact and concretion development at higher Huron levels.*

Farther down this stream are sparsely fossiliferous banks of brown-gray Lower Olentangy Shale (Figure 6), capped by the large-magnitude disconformity that we observed at STOP 1. This interval, characterized by a few concretionary nodules and limestone layers, is too far downstream to be examined on this trip. Focusing instead on the Upper Olentangy Shale interval, we will first examine the medial Upper Olentangy interval. Of particular note are the repetitive, heterolithic layers, so very distinctive to this upper Frasnian offshore shale interval. Although fossil-poor, pale gray-greenish shale comprises the bulk of this shale unit, one should note a variety of limestone bands and lenticular concretions (Figure 7 A). Unique to this interval are numerous thin bands of pale gray, knobby “popcorn-textured” limestone layers that repeat through this section. As noted in the text, these beds are believed to be an expression of “griotte facies”, an offshore pelagic type of nodular carbonate observed in stratigraphically condensed European and North African Late Devonian sections (Wendt & Aigner 1982).

Alternating with the gray shale and thin carbonate layers are a number of discrete, strongly contrasting, black shale bands of varying thickness. Typically, these layers display sharp basal contacts on lower gray shale units, but variably diffuse upper contacts, reflecting downward burrow penetration from succeeding gray shale layers. Black shale beds, particularly, the thicker layers, display associated nodular, diagenetic pyrite. The sharp, basal contacts of these beds are believed to record episodes of nondeposition or erosion prior to black mud accumulation, but lag concentrations of reworked pyrite or phosphatic bioclasts are uncommon beneath these thin layers. However, one of the two thickest beds displays a thin lag of detrital pyritic burrow tubes derived from the subjacent gray mudstone unit. Along the nearby stream in the Highbanks Metropark section (STOP 3), this same layer displays a basal pyritic lag with associated reworked fish bones and teeth, suggestive of sustained submarine erosion, timed with the onset of black mud deposition (Figure 8 A).

The presence of these well-defined black shale bands, especially the 0.5 – 1.5 inch-thick layers within this predominantly light gray-greenish colored and bioturbated unit, begs the question of their origin. Generally, the Upper Olentangy is understood to record dysoxic offshore conditions as suggested by its fine grained texture and striking near-absence of shelly, neritic fauna, but not as basinal as that for the succeeding black Huron Member upstream. The thicker

black shale bands may be truly linked to parasequence-scale transgressive risings of the pycnocline boundary, but the very thin ones may be a signature of transient upwelling events.

Continuing upstream, we encounter the prominent overhang of the resistant, black Huron Member over the topmost Olentangy layers (Figure 10). This contact is sharp, and it is believed to record an episode of submarine erosion. A number of prominent Paleozoic black shale units display widespread erosion surfaces such as this along their bases (see text discussion). In particular, noncalcareous lag deposits composed of phosphatic bioclasts, detrital (reworked) pyrite, quartz sand, and chert, occur along such surfaces, often directly beneath high TOC black shale deposits. The consistent absence of calcareous shells, correlated to the anomalous presence of lenticular detrital pyrite beneath a black shale roof, led Baird & Brett (1986, 1991) to propose that the reworked components had been exhumed and concentrated within the oxygen-deficient setting of the given black shale unit. Under such conditions, exhumed pyrite, normally unstable in an oxic setting, would survive exposure, while carbonate bioclasts would undergo selective dissolution, leading to formation of a stable lag placer under reducing conditions. Typically, as with the Middle Devonian Leicester Pyrite, flooring the black Genesee Member, and the Late Famennian Skinners Run Pyrite occurrence at the base of the Cleveland Member (Figures 13-14), the lag accumulations are restricted to laterally discontinuous debris lenses along these contacts. This may, to some extent, reflect a loss of lag volume owing to chemical dissolution, but, more importantly, it reflects hydraulic channelization of reworked matter into furrow-like erosional channels on a sediment-starved, oxygen-deficient submarine slope (Baird & Brett 1991; Figure 13).

It should be noted that the level marking the position of maximum inferred submarine scour along the Olentangy-Huron boundary in this outcrop is not always right at the base of continuous black shale lithology. Some exhumed pyritic debris is observed directly along the base of the Huron overhang, but better developed detrital pyrite lag concentrations can be seen in very widely-spaced, localized channels a few inches below the main contact (Figure 12). At the type Olentangy Shale section near Delaware, Ohio, Over and Rhodes (2000) recovered a rich assortment of conodonts marking the erosional base of the Huron Member, in a channelized concentration several inches below the base of continuous black Huron Shale. One channel-fill, observed at Camp Lazarus (Figure 12), commences at the base with a graded lag of detrital pyrite, followed, sequentially by concretionary gray mudstone that was, in turn, overlain by a lenticular black shale unit and a final gray shale lenticle, before reaching the continuous Huron succession. These lower channels appear to be precursors to the main, base-Huron contact and may record the initial onset of transgressive backstepping associated with the major Huron drowning event.

Continuing up the creek we gradually ascend through the lower Huron Member succession, back into the public area. Most noticeable are the huge septarial concretions for which Huron is famous (Figures 16, 17). Textural evidence, including the random tilting of concretions, differential compaction of host sediment, preserved card-house clay fabric in compaction strain shadows, and the preservation of uncompacted laminae in concretions indicate that they grew at shallow depths, perhaps no more than a few feet or tens of feet below the sediment-water interface. The stratally confined nature of carbonate concretions in the Huron Shale, as well as those hosted by other Devonian-aged black shales of the Appalachian Basin, is interpreted to reflect the anaerobic oxidation of methane (AOM) and consequent enhanced alkalinity during prolonged pauses in sedimentation. Each concretionary horizon is interpreted to reflect the diagenetic signature of AOM within the sulfate methane transition zone (SMTZ), a diagenetic horizon of indeterminate thickness along which downward-diffusing seawater sulfate and upward-diffusing methane are consumed by a consortium of methane-oxidizing archaea and sulfate-reducing bacteria. Low sedimentation rates focus the diagenetic effects of AOM, maintaining elevated pore water alkalinity within the SMTZ for an extended period of time.

*Board vehicles after visiting the park facilities and securing hydration (we will have lunch at the next stop). Proceed via the park exit road to the intersection with Olentangy Crossings West Road. Turn left (east) onto Olentangy Crossings West Road and proceed to U. S. Route 23. At that intersection, turn right (south) on to Route 23 and proceed for 3.5 miles to the entrance road for the Highbanks Metropark on the right. Turn right into the metropark and proceed for 1.5 miles westward, then north through the park to the sledding hill parking area at Google coordinates 40.154257, -83.038963. We will disembark and have lunch at this place.*

### **STOP 3: Upper Olentangy Shale-to-basal Huron Member succession at Highbanks Metropark north of Columbus, Ohio**

*This locality is within the premises of the Highbanks Metropark in southern Delaware County, Ohio, 4.5 miles north of Worthington, Ohio on the Powell 7.5' Quadrangle. The stream southeast of this lunch stop displays a good section of medial and upper parts of the Upper Olentangy Shale, succeeded by the Huron Member. We will proceed eastward along a system of park paths through an interval of woods to an upstream glen displaying the Upper Olentangy - Huron Member succession at Google coordinates 40.152542, -83.033926.*

*Proceed from the lunch picnic area southward to the creek via park trail. Turn left (eastward) onto an intersecting trail near the creek and follow that trail eastward to where that path ends at a trail intersection. Continue eastward from that intersection into the woods and up the creek valley to falls over middle and upper parts of the Upper Olentangy Shale.*

As at Shale Hollow Park (STOP 2), the middle and upper parts of the Upper Olentangy succession display numerous thin carbonate and black shale bands interbedded with the thicker gray-green shale layers (Figures 7 B, 8 A). In this section one of the thickest black shale layers, correlative with a similar bed at STOP 2, displays a prominent, basal lag deposit of detrital pyritic burrow tubes in association with fish bone fragments and teeth (Figure 7 B). Recently, several of us (Baird, Blood, Brett, and Danielsen) have endeavored to measure several key Upper Olentangy sections (type Olentangy section near Delaware, Camp Lazarus, Shale Hollow, Highbanks Metropark, and Slate Run in Upper Arlington) at the microstratigraphic level. These measured sections indicate that most of these thin beds can be matched between the sections. The Upper Olentangy is understood to have been deposited in a distal, downslope setting, and that divisional markers should have higher correlational fidelity in condensed sedimentary successions. Lash (2017) and Bush *et al.* (2017) both were able to match parasequence-scale markers within approximately coeval deposits of the Upper Frasnian Hanover Formation between two adjacent streams (Walnut and Silver creeks), less than a mile apart, in Chautauqua County, displaying facies that represent an expanded, upslope version of the Upper Olentangy succession. The present workers anticipate that this correlational fidelity should hold up through five well exposed sections over a lateral distance of 16 miles (26 km) between Delaware and Columbus.

Continuing upstream, we cross the Upper Olentangy-Huron Member contact and observe critical features of this lower Famennian basinal succession. On this creek and also at Camp Lazarus and Shale Hollow Glen, the lower part of the Huron is largely a near-continuous succession of brownish-black, organic-rich shale. However, commencing farther south in Columbus, and at sections along the Ohio River, as well as, near Morehead in east-central Kentucky, the lowest Huron interval is marked by alternations of black shale and distinctly gray-green shale, which had been referred to as “Olentangy” in earlier literature. Careful study of conodonts from these strata in Kentucky, however, shows these beds to post-date the Frasnian-Famennian and to be part of the Huron Member (Kepferle & Roen 1981; Over 2002). The fact that this “false Olentangy” shale facies is replaced northward by continuous black shale between Columbus and this locality over about a 10-mile distance, does show that significant lateral facies changes do occur within the basin that can present a serious challenge to the driller.

Again, we encounter several very large septarial concretions within the lower part of the Huron Member. This section also displays at least two fracture networks, which display interesting relationships to these concretions. Joint interactions with carbonate concretions provide insight into the driving mechanisms of fractures hosted by the Huron and other shales in the basin (Figures 21-23). Two major joint driving mechanisms exist in the Earth’s crust; (1) absolute tensile stress and (2) natural hydraulic fracturing. Tensional joints most commonly form as a

consequence of joint-normal stretching in association with the development of tensile stress. As such, it is the higher modulus rocks such as sandstones and carbonates that are likely to undergo preferential jointing. Observations of black shale-hosted joints continuing in plane around unfractured carbonate concretions, joint orientations, and their widespread distribution all suggest that these fractures formed as a result of natural hydraulic fracturing where fluid pressure works against compressive stress thereby creating an effective tensile stress. At these field stops we will discuss driving mechanisms and timing of joint formation and their significance to hydrocarbon exploration and production.

*Return to vehicles and exit Highbanks Metropark. Turn left (north) onto U. S. Route 23. Continue north on Route 23 for 10.5 miles to exit for Route 36-37 in Delaware. Exit Route 23 onto Route 36-37 eastbound. Continue eastward out of Delaware and continue for 11.3 miles to an intersection where Route 37 splits off from Route 36 at the west edge of Sunbury, Ohio. Turn left (northeast) onto Route 36 and continue for approximately 0.8 mile to intersection with High Street (on the right), immediately west of the Route 36 bridge over Big Walnut Creek. Turn right (south) onto High Street and continue for about 1100 feet to the intersection with County Route 44. Turn left (east) onto Route 44 and cross bridge over Big Walnut Creek. Continue about 120 feet on Route 44 to first driveway on the right. Turn right into that driveway (property of Dr. Mohler) to park and disembark from vehicles.*

#### **STOP 4: Sediment mega-failure within Berea Formation:**

*Lower part of Berea Formation displayed in steep, northeast-facing cutbank on Big Walnut Creek at Sunbury, Delaware County, Ohio (northwest corner of the Sunbury 7.5' Quadrangle; coordinates 40.245688, -82.849709). Section is 0.1 km downstream from (south of) the County Route 44 bridge. Section described by Molyneaux et al. (1984) and Pashin & Etensohn (1995).*

*Proceed from the cars for 150 feet (48 m) to the elevated concrete viewing area opposite the sandstone exposure. This exposure can be closely approached along the top of the concrete flood-control retaining structure, but please stay away from the unstable rock face. This is private property. The residents have been gracious to allow us to cross their property from the driveway to the creek. Stop duration 45 minutes.*

This stop emphasizes large-scale sedimentological instability associated with the onset of Berea Formation deposition, a phenomenon that is widely displayed across Ohio at quantum scales (Figure 31). This vertical face displays a train of four very large foundered masses of Berea siltstone that appears to be the product of a major sediment-loading event, possibly triggered by one or more great seismic events. The largest of these pillows has a vertical thickness of 16 feet (4.9 m), but is notably erosionally planed off (truncated) along its top, suggesting an even greater original vertical thickness (Figure 31).

As noted in the text, both the thick Bedford and Berea successions are complexly riddled with microfaults and sediment-loading features, suggestive of a combined regime of rapid sediment-influx and far-field tectonic instability. This is particularly evidenced by pervasive diapiric mega-slumping within the lower Berea interval, focused mainly in central and northern Ohio (Pashin & Etensohn 1995; Wells *et al.* 1991; Figure 32). Although the Berea Formation is floored by a conspicuous disconformity across much of Ohio, this hiatus is understood to be minimally developed or absent in parts of central and south-central parts of Ohio, where a topmost, non-red, portion of the Bedford can be seen in sections (Pashin & Etensohn 1995; see STOP 6; Figure 32). This non-red interval distinctly coarsens-upward toward the base of the Berea in Columbus-area sections with the appearance of thicker siltstone beds, channelized features, and slump structures near its top along Big Walnut Creek at Sunbury and along Rocky Fork farther south. At STOP 4, this giant ball-and-pillow seismite approximate the temporal acme zone of convulsive sediment disturbance events, roughly timed with lower Berea deposition.

Ensuing discussion should focus on causal mechanisms for the structural features displayed here, as well as for the enigmatic contextual settings of Bedford-Berea events in Ohio.

*Return to vehicles and exit property. Turn left (west) onto Route 36 and retrace your route back to Route 36. Turn left (west) onto Route 36 and continue 0.8 mile back to the intersection of Routes 36 and 37 at the west edge of Sunbury. Turn right (northwest) onto combined routes 36/37 and proceed 3.65 miles to southbound I-71 entrance feeder. Proceed for 10 miles south on I-71 to the eastbound entrance to I-270 (Columbus outer highway loop). Enter I-270 eastbound and continue 4 miles east, then south, on I-270 to the eastbound entrance feeder for Ohio Route 161 (Granville Road). Continue east of Route 161 for 1.7 miles to exit for Little Turtle Way. Turn right (south) onto Little Turtle Way, followed by an immediate right turn (west) onto East Dublin Granville Road. Continue west on East Dublin Granville Road for 0.5 mile to the entrance to Blendon Woods Metropark (on the left). Turn left (south) into Blendon Woods Metropark and proceed 0.25 mile to a fork in the road. Turn left at the fork and proceed eastward, then south, for 0.5 mile within the park to an entrance to a parking area for the Park Nature Center on the left. Turn left into that parking area and exit vehicles.*

#### **STOP 5: Red Bedford succession within medial part of Bedford Formation:**

*North-facing exposure of Bedford red-brownish mudstone along trunk stream within the Blendon Woods Metropark in northeast Columbus, Ohio on the Northeast Columbus 7.5' Quadrangle. This outcrop is south of (downstream from) the Brookside trail at coordinates 40.072947, -82.879990.*

*Proceed north on foot 0.12 mile from the Blendon Woods Nature Center Parking Area, via the Hickory Ridge Trail, to the intersection with the Brookside Trail. Proceed left (west) for*

*approximately 0.5 mile to a bridge crossing the main stream in the park. West of the bridge, proceed left (south) through the woods for about 100 feet to a large north-facing stream cutbank (STOP 5).*

### **Complexly deformed and sheared red Bedford mudstone deposits:**

The middle Bedford Formation interval is occupied by a thick sequence of clay-dominated, red-brown mudstone, which is very poorly bedded and strikingly uniform in Ohio sections (Figure 27 A). This facies is usually expressed in outcrop as non-stratiform earthy slopes, which become characteristically very slippery when wet. Siltstone layers, characteristic of non-red Bedford deposits, are nearly absent from this facies. The red Bedford is generally characterized by striking patterns of pervasive, small-scale shear features and complex patterns of microfaulting (Figure 27 A). Where bedding is discernable, it is restricted to small, localized domains with bedding dips in variable directions. A small area of visible red-green banded stratification can be seen above the base of this outcrop, but it occurs in only a small area. This part of the Bedford is notorious for its low shear strength on slope presenting obstacles to both engineers and drillers. Where the red Bedford facies grades downward and upward into synjacent gray Bedford deposits on this creek and elsewhere, the deranged bedding transitions to normal stratification. This poses a central question: is the presence of the red color a secondary artifact of the deformation of a gray sediment precursor, or does this pervasive structural deformation reflect selective tendencies of this sediment to deform, owing to its original mechanical properties, prior to burial and the onset of tectonism? This is a question that may be answered, in part, through geochemical approaches, and particularly, through the study of clay mineralogy.

As noted in the text, the red Bedford has been interpreted as either a nonmarine deposit (Pepper *et al.* 1954) or as an offshore facies (Pashin & Ettensohn 1995). The red Bedford generally lacks a visible shelly macrofauna, which initially supported Pepper *et al.*'s (1954) model for the "great Red Bedford Delta". The present authors take the view that, if it is an offshore marine facies, as argued by Pashin & Ettensohn (1995), there should be microfossils, including: conodonts, acritarchs, chitinozoans, and ostracodes in red Bedford deposits, which is currently being tested through regionally-directed bulk disaggregation and screening methods. "Barren" gray Bedford deposits, bordering the base and top of the red Bedford facies tract are also being bulk sampled and screen-processed for fossil content as well.

*Return to the Brookside Trail, turn right, and proceed on foot eastward back to the intersection of the Brookside and Hickory Ridge trails. Continue straight (east) as Brookside Trail changes name to "Ripple Rock Trail" and quickly crosses the main, paved, park road. From the park road crossing, follow Ripple Rock trail northward for 0.3 mile to where it approaches the trunk stream course at a place designated "Petrified Seafloor" within the park.*

**STOP 6: Gray marine shale and siltstone deposits of the uppermost Bedford Formation division:**

*Southeast-facing cutbank of topmost Bedford Formation division along trunk stream within the Blendon Woods Metropark in northeast Columbus, Ohio at the west edge of the New Albany 7.5' Quadrangle. This outcrop, at coordinates 40.075342, -82.874139, is closely approached by the Ripple Rock park trail.*

Although the upper part of this stream exposes few outcrops, the upper Bedford Formation gray, marine succession is well summarized in this easily-accessed, 16 foot (5 m) – high exposure. This part of the Bedford regionally coarsens-upward to the base of the Berea Sandstone, and part of this regressive facies transition is displayed in this cutbank section. Numerous tabular-to-lenticular siltstone beds alternate with shale layers imparting a flaggy appearance to the section. You will readily note the abundance of oscillation ripple marks for which this trail is named, both in the bank and in stream float (Figure 28). Wave-ripple marks such as these are typically abundant within the Bedford-Berea interval across Ohio, typically displaying a consistent northwest orientation across much of this state (see text). The absence of interference ripples and the presence of occasional in-phase ripples, suggest that these bedforms were the result of discrete and short-lived events during a period of rapid sediment accumulation. Pashin & Ettensohn (1995) interpreted the Bedford siltstone layers to be a mix of storm and turbiditic layers deposited between fair-weather and storm wave-base. Drawing on an analogy to the North Sea coast, they inferred a possible depth range of 33-100 ft. (10-30 m) for fair-weather wave-base to as much as 210 ft. (70 m) at the storm wave limit. However, the absence of gutter casts, typical of deeper-storm substrate scour in shelfal sedimentary systems (Aigner 1985), might suggest that this depositional setting was distinctly shallower and somewhat restricted.

Also conspicuous in the float are abundant trace fossils (ichnofossils), which are typically preserved as hypichnial markings (sediment casts along the sharp, basal contacts of the siltstone layers). These appear as networks preserved as bas-reliefs on overturned rock slabs (Figure 29 A, B). A modest suite of trace-makers is represented here, but most appear to be forms similar to *Chondrites*, a fodinichnial trace made by animals probing and ingesting the substrate for food, based on the behavioral classification of traces advanced by Seilacher (1967, 2007). A few rarer traces record locomotion across the substrate (walking traces = *repichnia*) and even escape traces (*fugichnia*). Though cryptic, this ichnobiota is presently our sole view of the late Bedford biotic world.

Where are the shelly fossils? As noted in the text, the Bedford Formation displays a marine shelled fauna only in its lower part, and mainly only in a thin zone, slightly above the base of the unit (Pepper *et al.* 1954; Pashin & Ettensohn 1992, 1995; Baird *et al.* 2013). In northern

Ohio, the Bedford lacks shelled fossils from the level of the Euclid Sandstone Member, upward to the base of the Berea, and from only a few feet above the basal contact to the base of the Berea in central and southern Ohio. The gray shale layers between the siltstone beds at this stop may contain very small shelly fossils, as suggested by the abundance of associated trace fossils, but they have not yet been found. A key overriding question presented by the Bedford succession is whether the deposits seen at STOPS 5 and 6 merely represent localized restricted environments or are the regional signature of paleoclimate-induced biosphere collapse associated with global Hangenberg events.

*Return to vehicles and exit Blendon Woods Metropark. Enter Route 161 westbound to the Columbus outer highway loop (I-270). Enter southbound 270 and continue for approximately 11 miles to exit for U.S. Route 33. Enter Route 33 and proceed southeastward for approximately 4.8 mile to exit for State Route 674 (Gender Road). Continue south on Route 674 for 2.6 miles to intersection of Route 674 with Lithopolis Road. Turn left (southeast) on Route 674 and continue for 0.6 mile to another intersection where Route 674 turns south. Turn right (south) and continue on Route 674 for approximately 4 miles to the entrance for Slate Run Metropark (on your right). Proceed west into the metropark to the road to the Shady Grove Picnic Area (on your left). Continue along that road to parking places nearest the far end of the Shady Grove parking area. Proceed west on the "Five Oaks Park Trail" for approximately 0.25 mile to where the path forks by a viewing platform. Take the right fork and continue for 0.06 downhill to a footbridge over the main course of Slate Run Creek. Exit the path and proceed for 50 feet upstream along the creek:*

**STOP 7: Ultra-thin Berea deposit and overlying Mississippian Sunbury submember of Orangeville Member succession:**

*Low waterfall exposure of Berea Formation in bed of Slate Run succeeded by 26 – 30 feet (7.9 – 9.1 m) – high bank exposure of the basal Mississippian Sunbury submember of the Orangeville Member of the Cuyahoga Formation within the Slate Run Metropark, 3.9 miles southwest of Lithopolis, Pickaway County, Ohio on the Canal Winchester 7.5' Quadrangle. Upstream, higher Mississippian siltstone deposits of the Buena Vista Member are present, but will not be visited. The Berea – Sunbury contact is at Google coordinates 39.759365, -82.842200.*

Of initial interest are the massive, slumped out blocks of the Berea Formation above the bridge and the thinner *in-situ* Berea ledge forming the upstream low falls. The base of the Berea on the topmost Bedford layers is knife-sharp, and the Berea, as a whole, is vastly thinner than as seen at Sunbury (STOP 4) or at other sections around Columbus. Downstream near the bridge, it is 4.5 feet (1.37 m) – thick, but at the falls lip, it is only 1 foot (30.5 cm) in thickness. At other area sections at Lithopolis and at A. W. Marion State Park, northeast of Circleville, Ohio, the Berea interval is, respectively, 6.0 and 6.8 feet (1.8 and 2.0 m) in thickness, suggesting the

presence of an elevated structural platform in this region, timed with Berea deposition. The absence of such a platform to the north at Gehanna and Sunbury would imply differential subsidence of this region, accounting for the conformable Bedford – Berea contact and greater Berea thickness to the north. The sharp, base-Berea contact at this locality is tentatively interpreted as an uneven onlap surface, which might explain the dramatic Berea thickness discrepancy between the bridge and the falls lip. However, it is also possible that some of the Berea has been removed from the top by sub-Carboniferous submarine erosion. This thin Berea at the falls lip displays distinctive carbonate cementation which shows up as coarse patches of syntaxially cemented sand fabric. Weathered downstream blocks are locally bright red, reflecting the presence of iron oxides within the Berea, perhaps from weathering of pyrite. The topmost Bedford below the Berea is non-silty, greenish-gray mudstone, but only a few feet of this interval is currently exposed. Recent examination of a nearby section along a downstream tributary suggests that the top of the red Bedford interval is about 15 – 20 feet (4.5 – 6.1 m) below the base-Berea unconformity at this place.

The massive Berea ledge is succeeded by a 6-inch (15 cm) interval of thin-bedded siltstone beds that thin upward to dark pyritiferous siltstone, which grades upward into sandy black shale, that comprises the basal Sunbury Shale. This upward-change marks a major marine transgression into a basinal marine setting that was strongly dysoxic to near-anoxic (Ettensohn *et al.* 2009). However, one or more sediment-starvation-related, transgressive erosion surfaces, may be developed in the thin-bedded siltstone interval capping the massive Berea ledge. Discovery of a bivalve within the thin-bedded siltstone interval, suggests the transient presence of a low diversity, marine, neritic fauna above the normally unfossiliferous Berea interval.

The Sunbury submember is a black, organic-rich shale, which yields only a sparse benthic fauna. Key fossils, often plentiful in the condensed basal Sunbury interval, are the linguloid brachiopods *Barroisella(?) melie* (Hall, 1870) and orbiculoids. Other phosphatic taxa that may be present include conularids and the enigmatic taxon *Sphenothallus* sp. Hall, 1847. The basal few centimeters of the Sunbury are notably silty and gritty, reflecting the reworking of Berea sediment. At A. W. Marion State Park, corresponding, silty, basal Sunbury strata display variable development of interference ripple marks. No observed unconformity-related lag deposits of reworked pyrite grains or phosphatic debris are observed along the base of the Sunbury at Slate Run Metropark, the type Sunbury section along Rattlesnake Creek at Sunbury, Ohio, or at A. W. Marion State Park. However, the relatively greater abundance of fossils in the basal Sunbury interval reflects both residual substrate dysoxia associated with the upward transition to near-anoxia in higher Sunbury beds and effects of sedimentary condensation of the basal Sunbury section, owing to sediment-starvation during this major deepening event (Ettensohn *et al.* 2009).

However, farther south in Kentucky, a well-developed erosional lag of phosphatic pebbles, phosphatic bioclasts, and detrital pyrite mark the knife-sharp base of the Sunbury, upon the southwesternmost Berea deposits and Bedford Formation in Lewis, Bath, and Rowan counties of eastern Kentucky (Baird & Brett 1991). This change is suggestive of increasing southward-southwestward erosional removal of pre-Carboniferous strata as the Cincinnati Arch is approached. Nonetheless, the coarse, base-Sunbury lag deposit, capping this unconformity, is conspicuously succeeded by basinal, organic-rich black shale facies across this entire interval (Kepferle & Roen 1981; Ettensohn *et al.* 2009).

*End of field trip.*

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