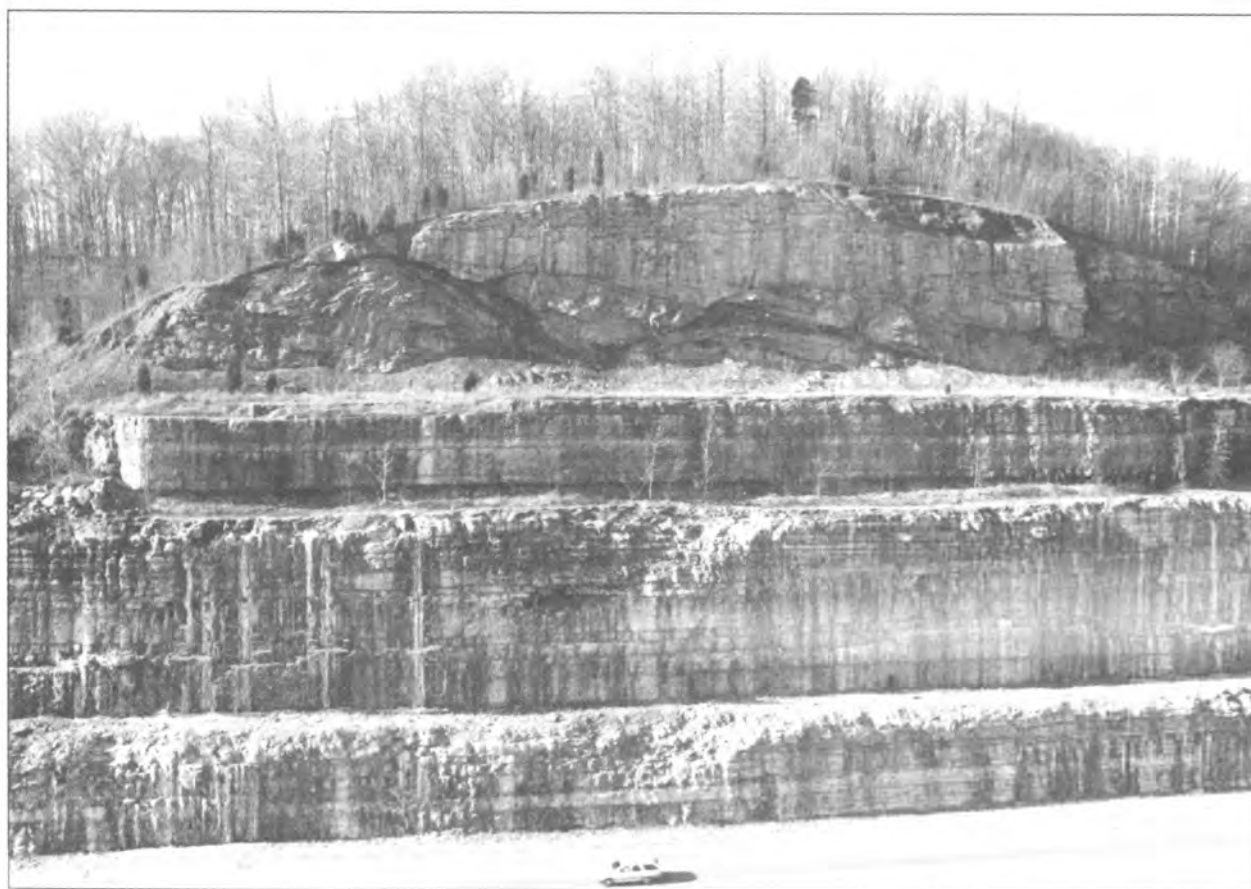


MISCELLANEOUS REPORT NO. 5

CHANGING INTERPRETATIONS OF KENTUCKY GEOLOGY— LAYER-CAKE, FACIES, FLEXURE, AND EUSTACY

edited by

Frank R. Ettensohn



prepared for the 1992 Annual Meeting
of the Geological Society of America



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MISCELLANEOUS REPORT NO. 5

CHANGING INTERPRETATIONS OF KENTUCKY GEOLOGY— LAYER-CAKE, FACIES, FLEXURE, AND EUSTACY

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University of Kentucky Department
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Cover illustration: Road cut on I-75 near Mt. Vernon, Kentucky, exposing 250 feet of Carboniferous section; Day 1, Stop 5. The lower 190 feet shows three unconformity-bound sequences of Mississippian carbonates in the Slade Formation. The upper 60 feet shows major slumping and deformation in the lower part of the Pennsylvanian Breathitt Formation, possibly related to movement on nearby structures. The Mississippian-Pennsylvanian unconformity is present on top of the third bench.

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INTRODUCTION

FRANK R. ETENSOHN

The study of Kentucky geology has been a nearly continuous pursuit since the first formal geologic studies in the 1740's (Johnson, 1898; Jillson, 1950). Although many significant findings have emerged at every step in that process, it is the purpose of this field trip to celebrate one particular step in that process, the centennial of the founding of the Department of Geology at the University of Kentucky on July 26, 1892. The University of Kentucky has always had some type of geology program since its establishment in 1865 as the Agricultural and Mechanical College of Kentucky, but until 1892 all such programs had been housed in the School of Natural History. Nonetheless, these early geology programs managed to attract such leading geology instructors of their time as Alexander Winchell and H. James Clark.

In 1892, the board of trustees offered the Chair of Geology to Arthur McQuiston Miller, a former professor at Wilson College in Pennsylvania, who had just returned from a year of study at the University of Munich under Dr. Karl von Zittel, a most distinguished paleontologist of his time.

With Professor Miller's appointment in 1892, the formal history of the Department began, although the name was soon changed from Department of Geology to Geology and Paleontology Department and eventually to the Department of Geology and Zoology. Other name changes ensued in subsequent years, and the Department has been known as the Department of Geological Sciences since 1986.

During the course of its history, the professors and students of the Department have contributed substantially to the understanding of Kentucky's largely sedimentary geology, and hence as part of the University of Kentucky, Department of Geological Sciences' centennial celebration, this centennial field trip will examine "Changing Interpretations of Kentucky Geology: Layer-Cake, Facies, Flexure and Eustasy." Although Kentucky rocks have well known layer-cake and facies interpretations, some of the eustatic and flexural interpretations are less well known and will be emphasized on this trip.

The title of the trip reflects three broadly defined types of interpretation of Kentucky geology, each of which is still applicable on some scale. The earliest and most broadly defined are the layer-cake interpretations, which imply a succession of widely correlatable units, each of which is distinct from strata above and below. In Kentucky, this concept was first applied to rocks of economic importance, namely to the iron-ore- and coal-bearing Carboniferous rocks of the state. David Dale Owen (1856) effectively implied transatlantic correlation of these rocks by applying names like Millstone Grit and Subcarboniferous, and later Mountain Limestone and Subcarboniferous Sandstone, names derived from lithologically similar and supposedly time-equivalent rocks in his native England. Older rocks were given descriptive names like "Black *Lingula* Shale" and "Blue Shell Limestone and Marl" (Owen, 1856).

The English names for Carboniferous units persisted in Kentucky until the late 1800's, when layer-cake correlations with units in surrounding regions resulted in their replacement with names such as Lee, Mauch Chunk, Pennington, St. Louis, Chester, Newman, and Waverly (Moore, 1878; Campbell, 1898a, 1898b; Hoenig, 1904). Although some of the Carboniferous layer-cake correlations, like those involving the St. Louis and Ste. Genevieve, are still accepted in Kentucky on

the basis of physical outcrop tracing and biostratigraphy (e.g., Etensohn and others, 1984; Sable and Déver, 1990), most of the correlations were routinely used but untested until new local names were suggested in the 1980's (e.g., Etensohn and others, 1984; Rice and others, 1987; Chesnut, 1989a).

By the early 1900's, the same type of layer-cake correlations had resulted in the replacement of Kentucky's Lower and Middle Paleozoic descriptive names with those from adjacent states; prominent exceptions are the Middle Ordovician High Bridge Group and Lexington Limestone, the Lower Silurian Brassfield and Alger (Crab Orchard) Formations, and the Middle Devonian Boyle and Duffin dolostones (see Miller, 1917). Many of the "imported names" were subsequently changed, but others are still used either on the surface or in the subsurface.

Interpretations of sedimentary geology, involving first lateral and then vertical facies, arose initially in Europe in 1838 (Gressly, 1838), but it was more than a century later before major facies interpretations of Kentucky stratigraphy were attempted.

The first major facies study in Kentucky was the seminal work of Paris Stockdale (1939) on the Lower and Middle Mississippian (Kinderhookian and Valmeyeran) Borden Formation. Subsequent facies studies in Kentucky have been few, but significant, and include the classic study of the Bedford-Berea sequence (Pepper and others, 1954) and the interpretation of the Lexington Limestone during the joint U.S. Geological Survey-Kentucky Geological Survey mapping program of the state (e.g., Cressman, 1973). One of the most controversial facies studies in Kentucky was the Barrier-Shoreline Model for the Mississippian-Pennsylvanian boundary interval in eastern Kentucky (Ferm and others, 1971; Horne and others, 1971, 1974). Although this model is no longer widely accepted, several studies testing parts of the model demonstrated the presence of syndimentary tectonism in Kentucky and its influence on facies development (e.g., Dever, 1973, 1980; Etensohn, 1975, 1980). This recognition has spurred further study and aided in explaining several stratigraphic and sedimentologic anomalies in Kentucky (e.g., Dever and others, 1977; Etensohn and others, 1988a, 1988b, 1991a; Pashin and Etensohn, 1987).

However, recognition of syndimentary tectonism has also kindled questions about the origin of syndimentary structural movements during times of tectonic quiescence in places like Kentucky seemingly far removed from the effects of coeval orogeny. It has also called into question the relative effects of tectonism versus eustasy, especially during times of major glaciation.

Answers to the first question are now being sought in the effects of lithospheric flexure (e.g., Quinlan and Beaumont, 1984; Etensohn, 1991a), which may operate during times of tectonic quiescence and be effective across distances greater than 1300 km from the originating orogeny (Karner and Watts, 1983; Ziegler, 1987). However, answers to the second question have not been as forthcoming, but the integration of flexural mechanisms with stratigraphic mapping of Devonian and Carboniferous units may be providing some promising partial solutions (e.g., Etensohn and others, 1988b; Chesnut, 1989a; Etensohn and Chesnut, 1989).

As the above discussion indicates, our understanding of Kentucky geology is still very incomplete. Nonetheless, the very conditions that make Kentucky's seemingly simply geology so complex—the presence and continued influence of major basement fault zones, a

position astride two cratonic basins, and a transitional location on the cratonward margins of foreland basins during four orogenies—are the same ones that create exceptional opportunities for elucidating some of these major problems. Hence, on the field trip we would like to discuss the basic framework of Kentucky geology as we now understand it, share some of our problems and possible solutions in a historic context, and seek your comments and suggestions.

During the next three days we will examine some classic and not-so-classic exposures in the course of our 22 stops. The stratigraphic disposition of those stops day by day is shown in Figure 1, and the location of the stops on the field-trip route is shown on the back cover of the guide. A series of brief papers at the beginning and elsewhere throughout the guide are provided to help place the geology at individual stops in an appropriate regional, historical and flexural framework.

Finally, it is appropriate to thank the 1992 G.S.A. Field Trip Committee, headed by Tom Berg, and the University of Kentucky Department of Geological Sciences Centennial Committee for their encouragement and support. I especially want to thank Judianne Lesniewski and Jeanie Thompson for manuscript preparation, Donald Chesnut, Jr., Nicholas Rast, and Merriane Hackathorn and her staff for proofing the manuscript, Charlie Mason for travel arrangements, and Anna Watson for logistical support on the field trip itself.

PHYSIOGRAPHY ALONG THE FIELD-TRIP ROUTE

FRANK R. ETTENSOHN

The trip begins in the Central Lowland Province (Till Plains Section) of the larger Interior Plains physiographic division of our country (Fenneman, 1938) (Fig. 2). This province is underlain by nearly horizontal Upper Ordovician rocks near the axial trace of the Cincinnati arch (Fig. 2). Although most of this province is a relatively flat-lying, monotonous plain with little relief, in the Cincinnati area the plain is very dissected because of proximity to the Ohio River and the older ages of the exposed tills on the margins of the plain (Nebraskan and Kansan: Durrell, 1961; Ray, 1966).

Approximately 17 mi south of the Ohio River near Richwood, Kentucky, the trip route leaves the Central Lowland Province and enters the Lexington Plain Section of the Interior Low Plateaus Province (Fig. 2). Most of the field trip will actually fall within this province. The northern boundary of the province is the southern edge of glacial drift; on the south and east, it is bounded by the Pottsville Escarpment and on the west by the Gulf Coastal Plain (Fig. 2). The province is dominated by the broad Cincinnati arch which determines outcrop distribution and regional topography and is underlain by Paleozoic rocks ranging in age from Ordovician to Pennsylvanian. Dips average 20-30 ft/mi (3.8-5.7 m/km) east and west away from the Cincinnati arch (McFarlan, 1943).

Two domal structures are developed along the axis of the Cincinnati arch, the Jessamine dome in central Kentucky and the Nashville dome in central Tennessee. The two domes are separated by a sag known as the Cumberland saddle (Fig. 2). Both domes have been breached and eroded forming topographic lowlands or basins surrounded by infacing escarpments or cuestas as high as 400 ft (122 m). The basin in Tennessee is known as the Nashville basin; the one in Kentucky is known as the Lexington Plain or Bluegrass Section (Fig. 2).

The field-trip route enters the Lexington Plain Section at mile 16.9 shortly after the beginning of the trip. The Lexington Plain is largely underlain by Middle and Upper Ordovician limestones and shales; on its margins a thin belt of Silurian rocks, and locally, some Devonian carbonates, are also included.

The Lexington Plain is divided into the Inner and Outer Bluegrass (Fig. 3), and what is shown as the "Outer Bluegrass" in Figure 3 is in some places divided into the Eden Belt and the Outer Bluegrass proper. The Eden Belt is formed on Middle and Upper Ordovician shales of the Kope and Clays Ferry formations. Because nonresistant shales predominate, the Eden Belt is deeply dissected with sharp, narrow ridge tops and steep v-shaped valleys (see Raitz, 1973). Because of steep slopes and poor clayey soils, the area is not of major agricultural importance except for grazing and small-scale farming. On the trip south from Cincinnati the entire route from Richwood to a point just north of Georgetown (Mile 60.4, Day 1) is included in the Eden Belt.

In general, the Outer Bluegrass proper includes irregularly rolling hills and low ridges underlain by Maysvillian and Richmondian limestones of the Fairview, Calloway Creek, Ashlock, Bull Fork, and Drakes formations and the Grant Lake Limestone, as well as overlying Silurian carbonates and shales. Because limestones are more abundant in this part of the section, the topography is not as steep as in the Eden Belt and the soils are more fertile. Hence, the farms are larger, and both cattle grazing and raising burley tobacco are the leading sources of income (Raitz, 1973). We will be in both the Eden Belt and Outer Bluegrass proper on all three days of the trip.

Just north of Georgetown, Kentucky (Mile 60.4, Day 1), the trip route enters the relatively flat-lying Inner Bluegrass, which is the agriculturally richest part of the Lexington Plain. The area is underlain by soluble, phosphate-rich Middle Ordovician limestones, which give rise to good soils. The flat terrain and rich soils make this an agriculturally prosperous region with large-scale tobacco farming, grazing and the breeding and raising of horses. Karstic features are locally common in the limestones of this section, and many surface streams are entrenched in steep-walled valleys.

South of Danville, Kentucky (Mile 124.3, Day 1), the trip route leaves the Inner Bluegrass and passes through a very narrow Eden Belt and Outer Bluegrass proper into the Highland Rim Section of the Interior Low Plateaus Province, defined by an unusually linear stretch called the "Knobs." Both the narrowness of the Eden and Outer Bluegrass belts, as well as the linearity of the Knobs belt south of Danville, are related to presence of the nearly east-west Brumfield fault on the south flank of the Jessamine dome.

Rocks exposed around the truncated Jessamine dome form a cuesta-like rim or escarpment surrounding the Lexington Plain. In this area and to the west, the "Knobs" form the lower part of this escarpment and mark the beginning of the Highland Rim Section (Fig. 3). The "Knobs" are a ring of conical hills carved from this escarpment, where erosion has cut through Mississippian and Devonian shales leaving isolated, rounded hills separated by wide valleys. Mississippian siltstones from the Borden Formation commonly cap many of these knobs, providing some degree of resistance.

Farther to the east and north, the Knobs and a narrow belt of lower-level plateau on Mississippian rocks, known as the Highland Rim proper, are included as parts of the Pottsville Escarpment in the Appalachian Plateau (Fig. 3). Typically, the Pottsville Escarpment is

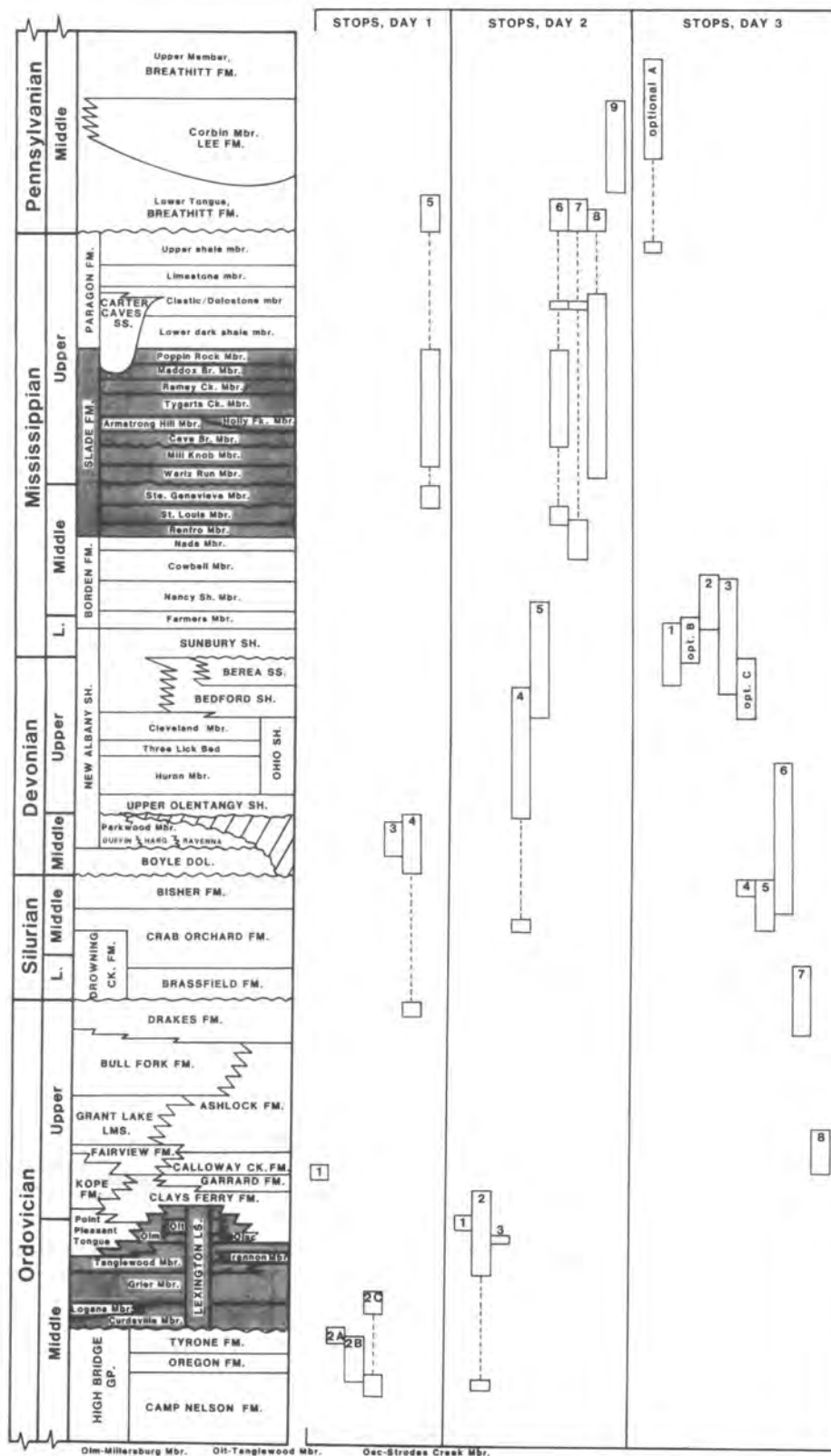


Figure 1. Generalized stratigraphic column for eastern and central Kentucky showing the relative stratigraphic disposition of the stops by day. Dashed lines reflect parts of the section lost by faulting or erosion on unconformities. The Middle and Upper Ordovician Lexington Limestone as well as the Middle and Upper Mississippian Slade Formation are highlighted for quick reference.

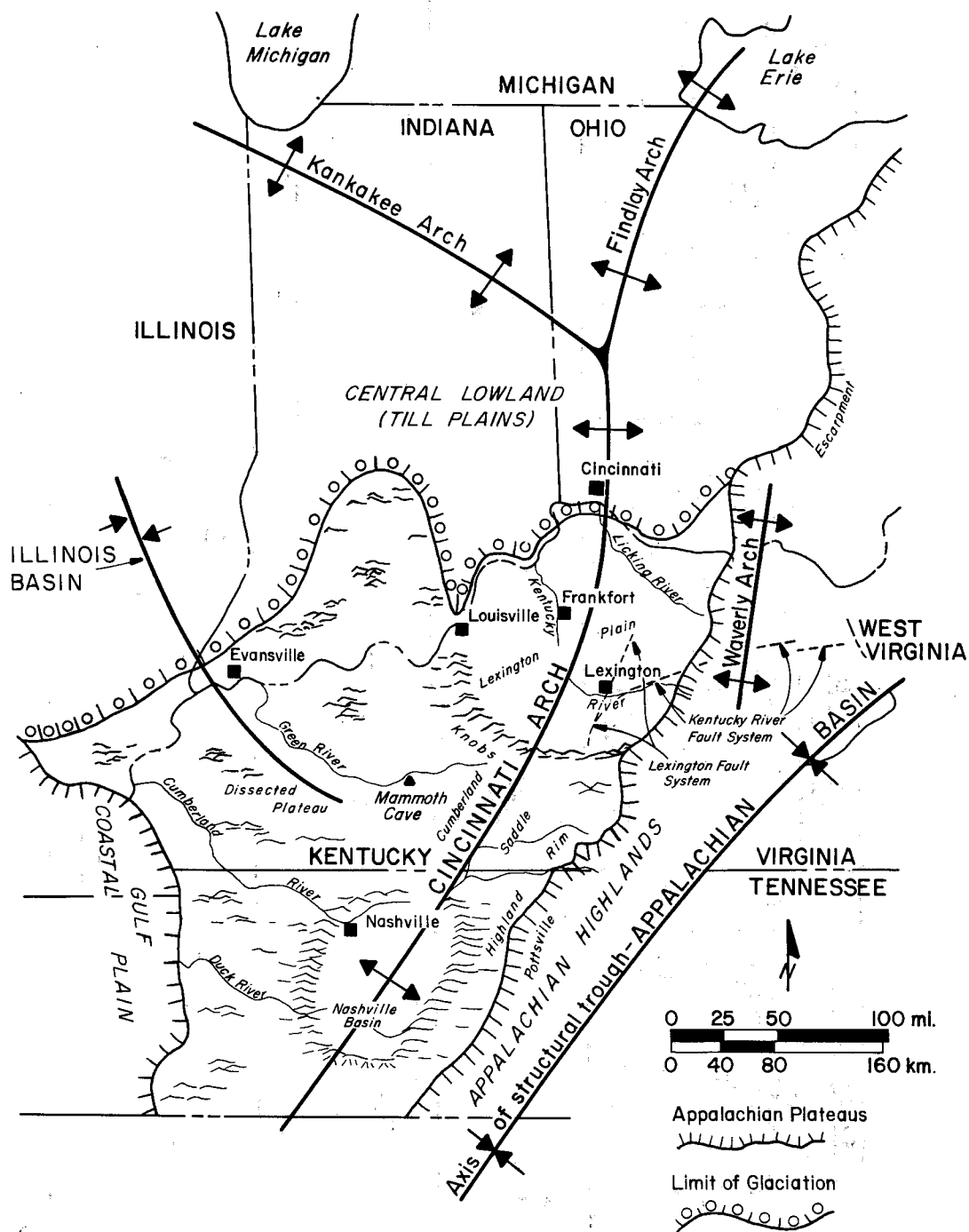


Figure 2. Interior Low Plateaus Province, in which most of the trip takes place, and parts of adjacent provinces showing major geologic and geographic features.

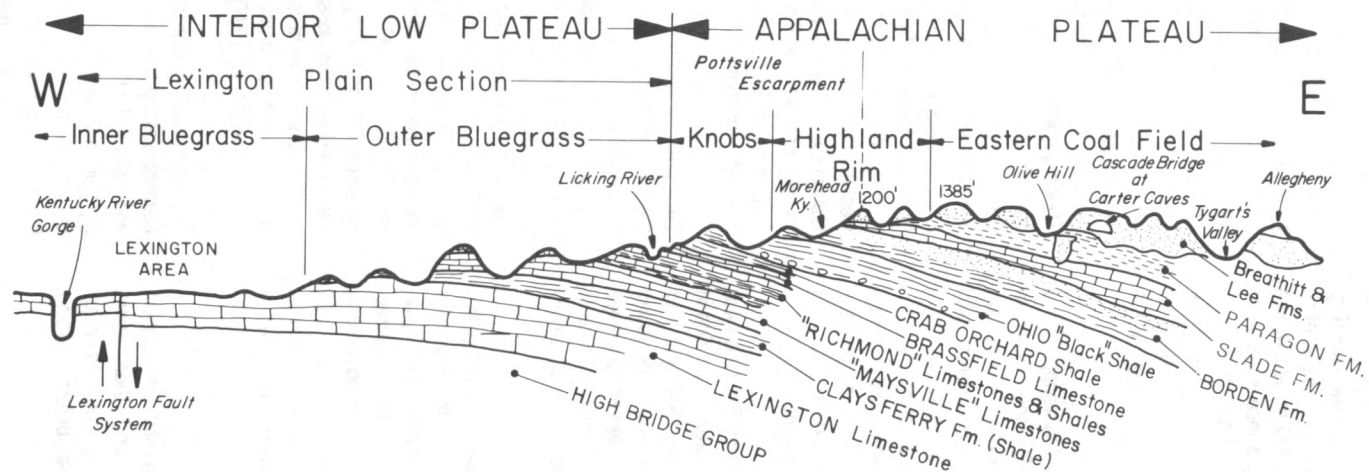


Figure 3. Schematic cross section through the east flank of the Cincinnati arch from the Lexington area westward to the Grayson area in Kentucky. The section approximately follows Interstate 64 west of Lexington and shows the major rock units and the physiography developed on them (after Ettensohn, 1981).

a westward-facing cuesta capped by resistant Lower Pennsylvanian (Pottsville or Lee) sandstones. However, because Mississippian and Devonian outcrop belts are relatively narrow on the eastern side of the Cincinnati arch and do not exhibit topography greatly different from that of Pennsylvanian rocks to the east, the Knobs and Highland Rim more or less merge with the Pottsville Escarpment and are included in the Appalachian Plateau (Fig. 3).

The Appalachian Plateau is a long northeast-southwest-trending highland extending northeastward from Alabama into New York (Fig. 2). Surface elevations range from 100 ft (30.5 m) above sea level on the western margin of the province to 4500 ft (1370 m) above sea level on the eastern margin of the province. The province is underlain by a broad asymmetrical synclinorium or structural trough (Fig. 2), whose southeast limb is steeper. Most sedimentary rocks in the province are nearly horizontal or dip gently toward the southeast. At the western edge of this gentle structural basin or trough is the Cincinnati arch (Fig. 2).

The Appalachian Plateau is largely underlain by Mississippian and Pennsylvanian sedimentary rocks, although some Devonian rocks are present along the margins and Permian rocks (Dunkard Group) are present in southeastern Ohio, northern West Virginia and southwestern Pennsylvania. Sandstones, shales, and conglomerates are the most common rocks present. The resistant sandstones and conglomerates protect underlying shales from erosion and weathering and effectively "hold up" the ridges which characterize the plateau. Coal beds are very significant in the Appalachian Plateau, and parts of the American coal industry are centered in the plateau.

The field trip enters the Appalachian Plateau for the first time near Brodhead, Kentucky (Mile 167.8) on Day 1. This area is included in the Cumberland Plateau Section which has milder relief and was described by Fenneman (1938) as submaturely dissected by young valleys. Hence the ridgetops are broader and the intervening valleys are less deep than elsewhere in the Appalachian Plateau. Some of the valleys are floored with Mississippian carbonate and are effectively karst valleys or uvalas (Mile 181.8, Day 1).

On Day 2, the trip route again enters the Appalachian Plateau, but this time, east of the Licking River (Mile 71, Day 2) in the Kanawha Section. The Kanawha Section is a more maturely dissected region of strong to moderate relief. The area is characterized by narrow-crested ridges and deep valleys; it is more minutely dissected and has greater relief than other sections of the Appalachian Plateau. Flat land in this section is minimal and is largely restricted to narrow stream bottoms. Hence, the major population centers are located in the stream valleys. Coal, gas, oil, and timber are major natural resources in the area.

On Day 3, we will essentially follow the Pottsville Escarpment northward to the Garrison, Kentucky area, and then move northwestward across the Knobs and onto the Lexington Plain for some final stops in the Silurian and Upper Ordovician, before returning to Cincinnati.

A BRIEF STRUCTURAL FRAMEWORK OF KENTUCKY

Frank R. Ettensohn and Jack C. Pashin

The Paleozoic structural framework of Kentucky is diverse and includes two sedimentary basins (Appalachian foreland basin and

Illinois intracratonic basin), an interbasinal uplift (Cincinnati arch and the Jessamine dome, a subsidiary uplift on the arch), continental rifts (Rome trough and Rough Creek graben), and the leading edge of an orogenic belt (Pine Mountain thrust sheet) (Fig. 4). Some of these structures began forming at the end of the Proterozoic Era and continued to form and mature throughout the Paleozoic Era. Parts of the even older Proterozoic mid-continental rift system and the Grenville front also underlie Kentucky and influenced subsequent patterns of Paleozoic sedimentation. As a result of this long-term development, the structural framework is inextricably linked to the stratigraphy and sedimentology of Kentucky. The following discussion presents a general structural framework for the state and stresses that rift-related basement structures had a profound effect on subsequent structural evolution and sedimentation.

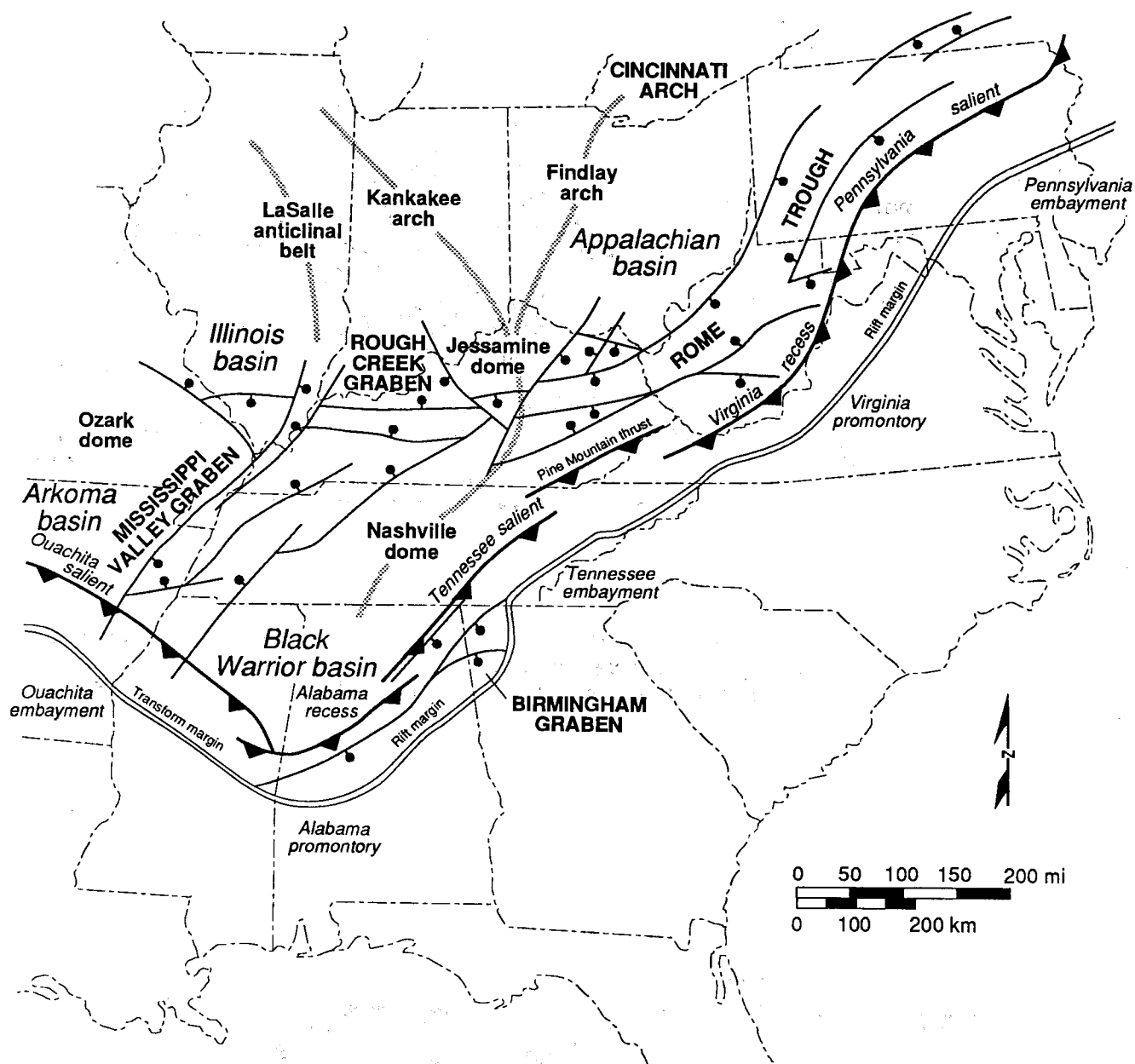
IAPETAN RIFT

The oldest structures considered are a series of basement grabens that include the Rome trough and the Rough Creek graben (Fig. 4). These grabens are part of a Late Precambrian to Middle Cambrian cratonic rift system that formed contemporaneously with rifting of the Early Paleozoic Iapetus continental margin (e.g., Thomas, 1991). The Rome trough extends as far northeast as New York and terminates southwestwardly at a series of basement faults with varied orientation in central Kentucky that coincides with the modern-day Jessamine dome. West of these faults, the Rough Creek graben extends westward and intersects the Mississippi Valley graben and associated fault systems east of the Ozark dome. Note that the fault system roughly parallels the rifted Iapetus continental margin and is nearly perpendicular to the transform continental margin adjacent to the Ouachita embayment.

Fault patterns and basement configuration in Kentucky suggest that the Rome trough and Rough Creek graben extend across the Cincinnati arch (e.g., Harris, 1975; Black, 1986), but complex fault configurations associated with the Jessamine dome indicate that coupling of the systems is complex (Fig. 4). The predominant down-to-south displacement of Rome trough basement faults in eastern Kentucky contrasts with the predominant down-to-north displacement of Rough Creek basement faults in the western part of the state, reflecting opposite fault polarity. Together, these systems can be characterized by East African rift terminology (Rosendahl, 1987; Scott and Rosendahl, 1989) and accordingly represent nonoverlapping, opposing, half-graben systems (Fig. 5). The short, curvilinear faults associated with the Jessamine dome, therefore, are recognized as transfer faults in what is called an isolation accommodation zone, which serves to accommodate graben systems with opposite polarity (Frostick and Reid, 1987).

Synsedimentary basement-fault movement in Kentucky began with Cambrian rifting and was due mainly to thermal subsidence (Kolata and Nelson, 1991). Subsidence resulted in thick (~9,800 ft or 3,000 m) graben-fill sequences which typify active rift basins (Webb, 1980; Schwab, 1982; Collinson and others, 1988). After rifting, basement faults were reactivated as a distal response to Appalachian orogenesis, thereby resulting in many stratigraphic anomalies, including local unconformities, abrupt thickness and facies changes, and disjunct distribution of stratigraphic units. Such anomalies have been noted in each post-Cambrian Paleozoic system in the state, and several examples will be examined on this trip.

Although many of these sedimentologic anomalies are related to simple growth faulting on basement precursors, in some places these



EXPLANATION

-  Intracratonic arch
-  Normal basement fault associated with lapetan rifting; ball on downthrown side
-  Frontal structures of Appalachian-Ouachita orogen
-  Early Paleozoic continental margin (after Sloss, 1988; not palinspastically restored)

Figure 4. Structural geology of Kentucky and neighboring areas in relation to southeastern North America (after Thomas, 1991).

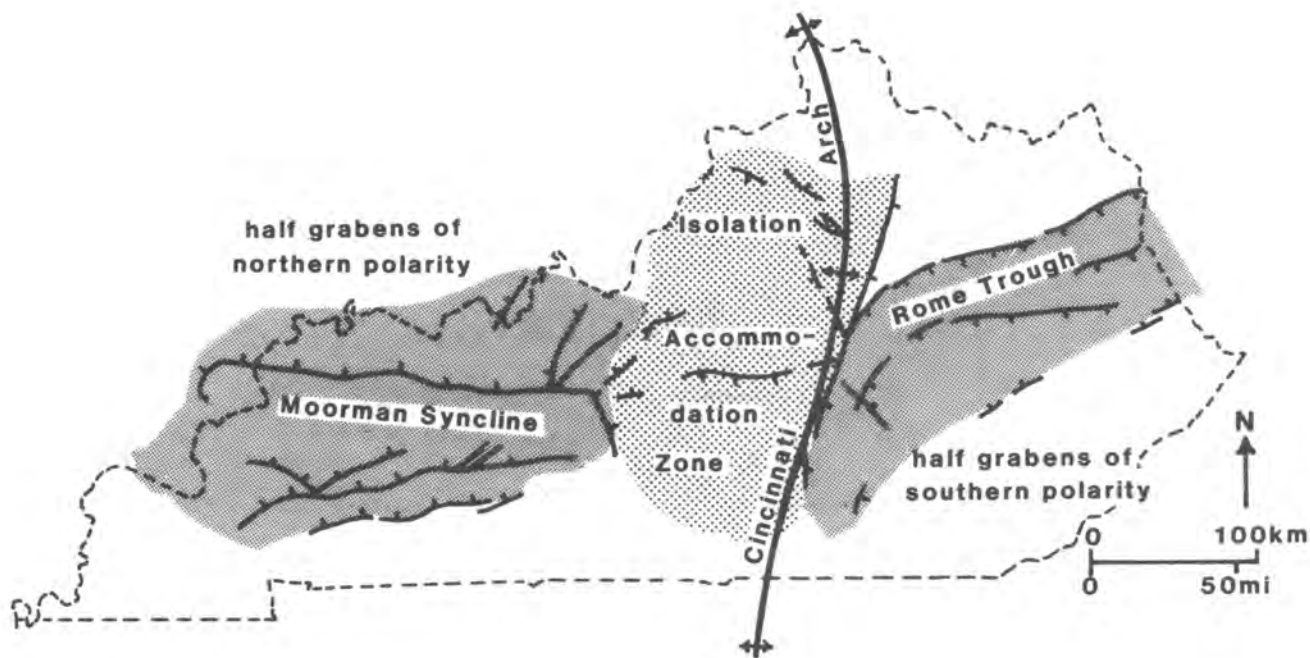


Figure 5. Generalized fault patterns and basement configuration in Kentucky showing that the Rome trough and Moorman syncline (Rough Creek graben) may represent nonoverlapping, opposing half-graben systems formed during late Precambrian-Middle Cambrian Iapetan rifting. At various times, regional structures like the Lexington platform, Cincinnati arch, and tectonic bulges may have become localized on the intervening isolation accommodation zone.

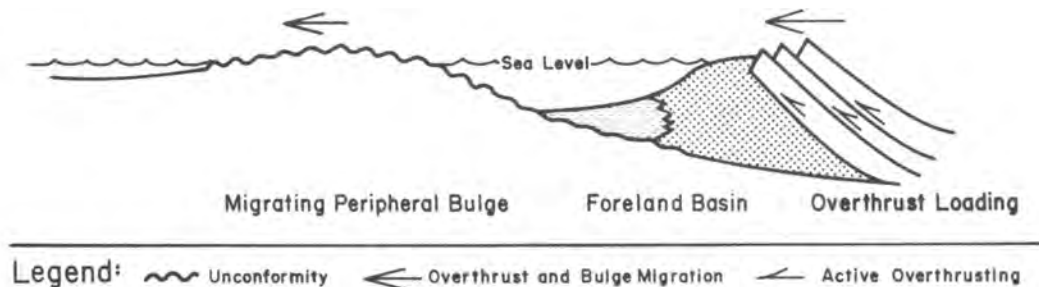


Figure 6. Schematic diagram showing development of a foreland basin and peripheral bulge due to deformational loading in the orogen (adapted from Quinlan and Beaumont, 1984).

anomalies can be traced to structures that still have surface expression today (Ettensohn and others, 1988b; Barnett and others, 1989a). Moreover, evidence from regional geophysical analysis, deep drilling, joint patterns, and facies distribution (e.g., Heyl, 1972; Black and Haney, 1975) indicates that many of these structures have undergone multiple phases of movement, including possible reverse, normal, and strike-slip phases on a single fault.

Some of the most conspicuous sedimentologic anomalies are in Ordovician, Devonian, and Pennsylvanian rocks which accumulated during the major Appalachian orogenies. Consequently, many workers assumed that reactivation was related to simple transmission of stress along preexisting basement structures during collisional compression to the east. This made it difficult, however, to explain reactivation that occurred during the relatively quiescent Silurian and Mississippian periods. With the advent of lithospheric-flexure theory (see the following section), these so-called "quiescent" reactivations can be understood in terms of a flexing lithosphere and migrating bulge which move not only during active orogeny, but also during the subsequent phases of relaxation. Flexural perturbations of this sort may reactivate preexisting basement structures and affect regions up to 1,300 km away from an active orogen (Karner and Watts, 1983; Ziegler, 1987); they may further account for subsidence in foreland and cratonic basins as well as uplift of interbasinal arches (Quinlan and Beaumont, 1984).

UPLIFTS AND BASINS

Flexural theory has shed new light on the origin of the Cincinnati arch, Appalachian basin, and Illinois basin (Fig. 4). Fault-bounded uplifts evidently formed on the Jessamine and Nashville domes during the Middle and early Late Ordovician (Borella and Osborne, 1978). However, a continuous basin-arch system apparently was not present or active until possibly the latest Ordovician, because lithofacies trends in rocks up to this age are at a high angle to the present axis of the arch (Borella and Osborne, 1978; Weir and others, 1984). In fact, many lithofacies trends in Upper Ordovician rocks are nearly east-west, suggesting relict control of sedimentation by rift-related basement faults. Uppermost Ordovician lithofacies trends, however, more closely parallel the present axis of the arch (see Weir and others, 1984), and Lower Silurian facies and thickness trends establish that the arch was present and affected sedimentation (McDowell, 1983; Gordon and Ettensohn, 1984). The Cincinnati arch apparently was active and had marked topographic expression during much of the Paleozoic, especially following regional unconformity development, because Middle Devonian and Early Pennsylvanian units onlap the structure from both sides.

The fact that the arch first formed during the Taconian orogeny and underwent major periods of uplift during the subsequent Acadian and Alleghanian orogenies suggests that Appalachian orogenesis contributed to development of the arch. According to Quinlan and Beaumont (1984), uplifts like the Cincinnati arch are the product of flexural interaction between the Appalachian basin and contemporaneous intracratonic basins like the Illinois basin. In Kentucky, these basinal areas were initially separated by the apparent isolation accommodation zone between opposing parts of a continental rift system (Fig. 5). Although the Illinois basin may have originated partly as a later response to this rifting (Kolata and Nelson, 1991), the area of the present-day Appalachian basin evolved into a passive-margin carbonate shelf after rifting.

Not until inception of the Taconian orogeny at the Early-Middle Ordovician transition did a well-developed Appalachian foreland basin form. Initially, the foreland basin and associated peripheral bulge were far from the Illinois basin and its peripheral bulge, and little or no flexural interaction occurred. By the end of the Ordovician, however, the Appalachian bulge had migrated far enough westward to interact constructively with the Illinois basin bulge, thereby producing the Cincinnati arch within the old accommodation zone. Subsequent phases of bulge interaction in response to the Acadian and Alleghanian orogenies, moreover, served to enhance uplift of the arch. Although domes along the Cincinnati arch may reflect regions of maximum constructive bulge interference (Quinlan and Beaumont, 1984), an alternative interpretation is that the domes represent isolated fault-bound structures where stress was localized and uplift was focused.

APPALACHIAN OROGEN

The youngest structure discussed herein is the Pine Mountain thrust fault, which is in the northeast part of the Tennessee salient of the Appalachian orogen (Thomas, 1977) and is the most outboard structure of the Appalachian fold-and-thrust belt (Rich, 1934; Rodgers, 1990) (Fig. 4). Why is the most outboard structure of the fold-and-thrust belt in Kentucky? Does this configuration have implications for understanding synsedimentary tectonism in the state? Perhaps the answers to these questions lie in the relationship of Appalachian thrusting to the Iapetan continental margin and the Iapetan intracratonic rift.

Orogenesis gives rise to normal faulting in the adjacent foreland basin by thrust and sediment loading and by flexural extension of the lithosphere (Bradley and Kidd, 1991); much of the faulting may occur on preexisting structures (Karner and Watts, 1983). Salients and recesses in the Appalachian orogen reflect promontories and embayments in the Iapetan continental margin, and basement faults formed during rifting apparently helped localize thrusting during subsequent orogenesis (Thomas, 1977, 1991) (Fig. 4). For example, frontal structures of the fold-and-thrust belt follow basement faults associated with the Birmingham graben in Alabama and with the Rome trough in West Virginia and Pennsylvania. In Kentucky, the Rome trough turns west, and the Pine Mountain thrust occurs in a tectonic transfer zone between the Birmingham and Rome trough basement-fault systems. Perhaps the westward curve of the Rome trough north of the transfer zone helped accommodate thrusting farther cratonward than in adjacent areas.

Indeed, the basement faults, basins, uplifts, and orogenic structures of Kentucky are all intimately related, and the initiation of one type of structure had a major effect on the evolution of subsequent structures. Iapetan rifting determined basement configuration and regulated how flexural development of the basin-and-arch system would proceed during the Taconian, Acadian, and Alleghanian orogenies. During these orogenies, lithospheric flexure and fault reactivation acted collectively and are the key causes of the many stratigraphic and sedimentologic anomalies that will be observed on this trip.

SOME BASIC FLEXURAL MODELS

FRANK R. ETENSOHN

Regional stratigraphic complexities are at times difficult to integrate into coherent patterns which are easy to understand. However, the

development and application of lithospheric-flexure models in the last ten years (e.g., Beaumont, 1981; Jacobi, 1981; Karner and Watts, 1983) have provided another tool for understanding the complex areal and vertical distributions of lithologies. Basically, most flexural models suggest that subsurface and surface deformational loading by flakes, blocks, thrusts, nappes and folds in an orogen produces a downwarped flexural or retroarc (*sensu* Dickinson, 1974) foreland basin cratonward of the orogen and a peripheral bulge on the distal (cratonward) margin of the basin (Fig. 6); such warping or flexure is a response to regional isostatic compensation by the lithosphere. Although most of the loading is related to the fold-thrust belt, a subordinate loading component is produced by sediment infilling the foreland basin (Beaumont, 1981; Tankard, 1986a). Moreover, as orogeny proceeds and thrust loads continue to shift cratonward or approximately normal to the trend of the orogen, the foreland basin and peripheral bulge also migrate cratonward away from the load. If an orogeny is diachronous along its length, foreland basins and sediment sources will migrate parallel to the trend of the orogen (Ettensohn, 1987).

Once active orogeny and thrusting cease, however, lithospheric relaxation will cause the bulge to be uplifted and migrate back toward the orogen for reasons discussed in a following subsection. The importance of this is, that even during times of orogenic quiescence when active thrusting ceases, migration and reorganization of the bulge and foreland basin continue so that "tectonic factors are still at work."

Obviously, the greatest influence on sedimentation and resulting stratigraphy will be seen in the foreland basin which is closest to the orogen and hence experiences the greatest flexural perturbations. However, inasmuch as lithospheric flexure is effective at least 1300 km away from the originating orogen (Karner and Watts, 1983; Ziegler, 1987), its effects should be expected far afield from the originating orogen on the adjacent craton. In fact, Ziegler (1987) has coined the term "far-field tectonics" to describe these effects. This means that virtually all of Kentucky would have experienced some flexural effects from the various Appalachian orogenies and the Ouachita orogeny, those effects having been greater in parts of Kentucky closest to the respective orogeny. Hence, many of the patterns of regional transgression and regression seen in Kentucky and elsewhere, especially during times of major orogeny, may in large part reflect regional far-field tectonic effects rather than eustasy, as has been commonly assumed. This fact is also suggested by subsidence curves generated from basement test wells in Kentucky (see Goodmann, this guide). When the effects of bathymetry, compaction, and sediment loading are factored out, significant episodes of subsidence all coincide with major tectonic events except for the initial syn-lapetan event which reflects thermal subsidence concomitant with rifting.

During a typical subduction-type orogeny, flexural responses to loading events give rise to a characteristic four-part, process-response sequence which is apparent in the rocks of both the foreland basin and on adjacent parts of the craton. Because orogenies generally proceed in short pulses or tectophases on the order of five million years or less (Jamieson and Beaumont, 1988), the characteristic, four-part sequence may be repeated several times during any one orogeny. The origin and nature of these process-response sequences are briefly explained below using the Ordovician of the Appalachian basin and of adjacent cratonic parts of central Kentucky.

Unconformity Development

As subduction and convergence begin and a deformational load accumulates on the continental margin, the lithosphere responds by rapidly generating a compensating foreland basin and peripheral bulge (Fig. 6) (Quinlan and Beaumont, 1984). As the load moves cratonward with time, the basin and uplifted bulge also migrate in the same direction, and erosion on the uplifted bulge generates a lower bounding unconformity (Fig. 6) (Quinlan and Beaumont, 1984). Continental impedance to initial subduction or collision may also generate unconformities in this position (Dickinson, 1974, p. 22; Ettensohn, 1991a). In the Ordovician situation (Fig. 7), the initiation of the Taconian orogeny occurred during Whiterockian time near the Early-Middle Ordovician transition and was responsible for generating the post-Sauk (Sloss, 1963) or Knox unconformity. It apparently represents the inception of the Blountian tectophase of the Taconian orogeny and the beginning of convergence near the Virginia promontory (Ettensohn, 1991a).

By the late Middle Ordovician (Rocklandian), the locus of Taconian convergence had moved northward to the New York promontory where the major Taconian tectophase, the Taconic, occurred. In parts of New York, Pennsylvania, and Maryland, a major unconformity marks the advent of this tectophase near or at the top of the Black River Group and its equivalents. However, in Kentucky, nearly 1000 km distant from the locus of the tectophase, the unconformity is poorly developed or absent (Fig. 7). This subtle unconformity will be viewed at Stop 2A on Day 1.

Foreland Basin Subsidence and Regional Transgression

As active tectonism and deformational loading ensue, a phase of rapid foreland-basin subsidence follows bulge formation and moveout. Because initial foreland-basin subsidence is related to deformational loading in the subsurface (Karner and Watts, 1983), or loading which never breaks the ocean/sea surface, no major source of externally derived sediment is available. Although a thin sequence of transgressive carbonates may be initially deposited on the unconformity, as rapid subsidence proceeds, the foreland basin experiences net sediment starvation. In the absence of major clastic influx, abundant organic matter from the water column composes most of the sediment in the early basins. Moreover, because subsidence exceeds sedimentation in the early basins, the water column in any of these basins soon becomes stratified so that the organic-rich sediments are quickly preserved as dark or black shales in the resulting anoxic environments. Hence, during this phase of flexural subsidence, dark, organic-rich shales predominate in the foreland basin.

However, while dark-shale deposition is ongoing in the foreland basin, in adjacent cratonic areas like central Kentucky, carbonate deposition may persist for some time because of reduced rates of subsidence away from centers of loading. Nonetheless, some flexural subsidence in these distal cratonic areas will occur, so that the overall nature of the carbonate sequences will be transgressive. Such is the case in most of the Wells Creek-lower Camp Nelson sequence deposited during the Blountian tectophase (Fig. 7). The Taconic tectophase, however, was a much stronger tectophase than the Blountian, so that shallow-water Lexington carbonates were eventually inundated by

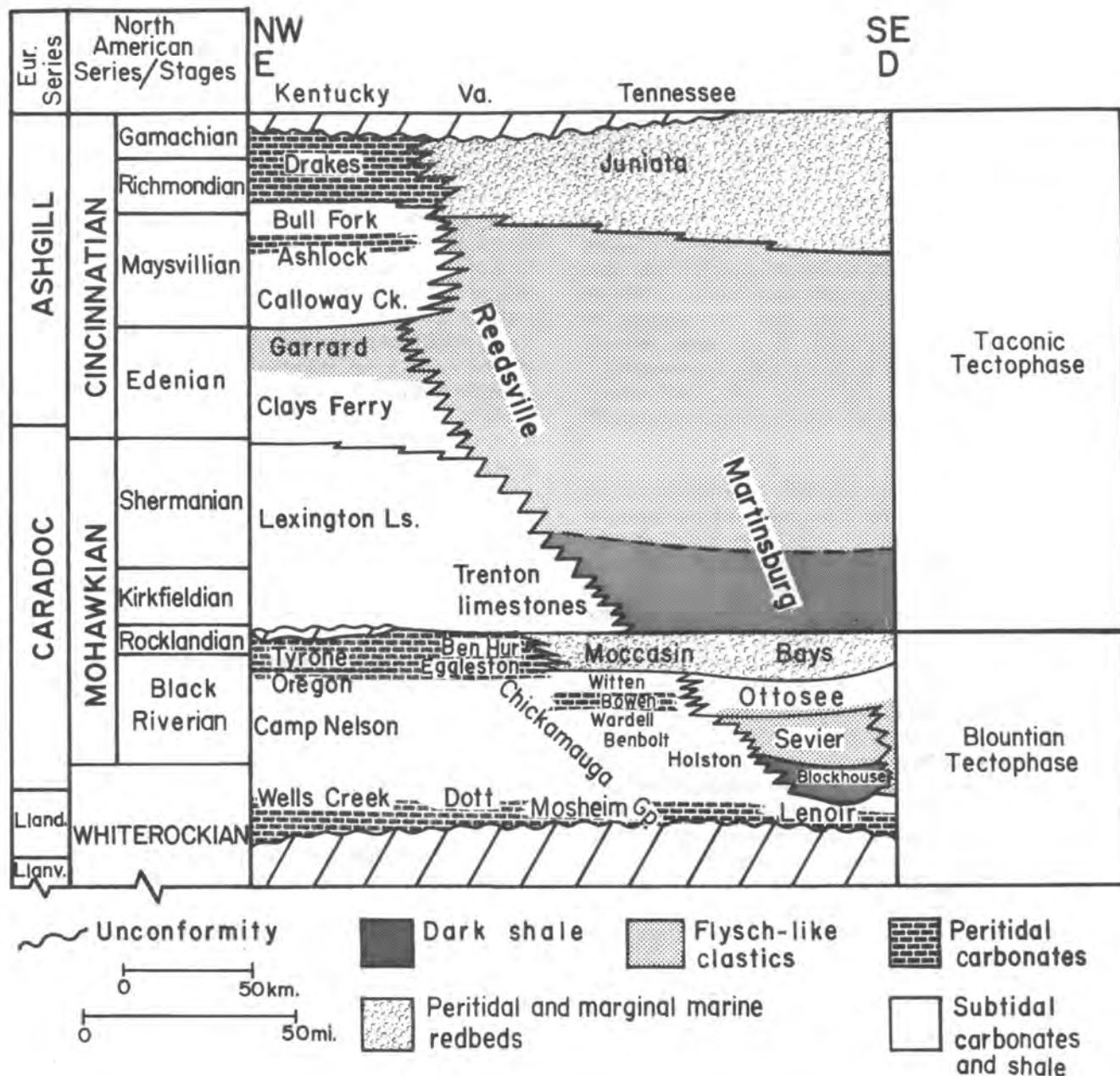


Figure 7. Schematic northwest-southwest section perpendicular to the strike of the southern Appalachian basin showing the repetition and relative cratonward migration of two foreland-basin sequences resulting from two Taconian tectophases. The Bowen-Witten and upper Ashlock-Bull Fork sequences may represent the passage of peripheral and "anti-peripheral" bulges on distal margins of the respective foreland basins. No vertical scale intended; see Figure 13 for approximate location of section E-D (from Ettensohn, 1991a).

deeper water Clays Ferry shales and carbonates as Taconic deformation and resulting subsidence moved progressively cratonward (Fig. 7).

Loading-Type Relaxation and Regional Regression

During the next phase, active deformation and thrust migration cease, so that the deformational load becomes static and orogenic quiescence ensues. The lithosphere responds to the static load by relaxing stress so that the foreland basin deepens and narrows while the peripheral bulge is uplifted and shifts toward the load (Fig. 8A). By this time substantial relief has been generated by emplacement of a surface load (fold-thrust belt above sea level), and surface drainage nets have had adequate time to develop. As a result, coarser clastic debris is eroded and transported into the foreland basin in the form of turbidites, debris flows, and storm deposits. Therefore, in the foreland basin during this phase, dark shales are typically buried by a regressive sequence of flysch-like clastic sediments (Fig. 7). Because the rate of clastic influx during this phase eventually exceeds the rate of basin subsidence, the foreland basin fills with clastic sediments while adjacent orogenic highlands undergo extensive lowering due to subsidence and erosion (Fig. 8B). As a result, a brief period of elevational equilibrium ensues between the filled basin and eroded highlands (Fig. 8B, no. 3).

Although marine clastic sediments dominate the central and proximal margins of the foreland basin during this phase, carbonate deposition may continue on the distal cratonic margins of the basin in areas like central Kentucky. However, uplift and migration of the bulge toward the orogen (Fig. 8A) will give rise to a distinctly regressive carbonate sequence. The culmination of bulge uplift and the resulting regression may be represented by a local unconformity or by a peritidal unit. In cratonic portions of Tennessee, somewhat closer to the foreland basin than central Kentucky during the Blountian tectophase, this culmination is represented by the peritidal Bowen Formation of the Chickamauga Group. In central Kentucky, however, this culmination may be represented by the prominent peritidal "white band" (see Stop 2C, Day 1) which occurs at about the same stratigraphic position in the Camp Nelson Formation as does the Bowen in the Chickamauga Group. Hence, much of the upper Camp Nelson exposed in Kentucky probably reflects this same phase of regression, whereas the Calloway Creek Formation and equivalent Fairview Formation, as well as parts of the Grant Lake Limestone and Ashlock Formation, represent this phase of regression during the Taconic tectophase (Fig. 7).

Unloading-Type Relaxation, Renewed Regression, and Progradation From the East

During the final phase of the process-response sequence, the area of the former orogen and foreland basin reacts to the "lost load" by isostatically rebounding upward and generating a compensating "anti-peripheral bulge," which deepens and migrates toward the rebounding area (Fig. 8B) (Beaumont and others, 1988). This brief, and perhaps localized, episode of transgression is followed by regression as the shallow, anti-peripheral-bulge basin is then filled with a cratonward-prograding wedge of fine-grained marginal-marine and terrestrial sediments which commonly include redbeds. In the basal Ordovician sequence, the Blountian and Taconic tectophases are represented respectively by redbeds in the Bays-Moccasin and the Sequatchie-Juniata-Queenston formations (Fig. 7). Although little or none of the prograding redbeds are ever found in the coeval cratonic rocks, these rocks generally contain regressive marginal-marine or

peritidal carbonates because of the great distances from rebounding source areas. In the central Kentucky area, the peritidal carbonates of the Oregon and Tyrone formations represent this phase in the Blountian sequence (Fig. 7), and these will be examined at Stops 2A and B on Day 1. For the Taconic sequence, we will examine the marginal-marine carbonates and shales of the Drakes Formation at Stop 7 on Day 3. This stop also shows one of the most distal tongues of the Queenston redbed lithofacies.

OVERVIEW

The process-response models discussed above were originally developed for the interpretation of foreland-basin stratigraphy (Ettensohn, 1991a), but obviously the overall transgressive-regressive pattern inherent in these models should be equally apparent on platform and ramp areas on adjacent parts of the craton like central Kentucky. Figure 9 shows the typical transgressive-regressive pattern inherent in the above flexural models. Although every part of the models is apparent in the Ordovician sequences used as examples (Fig. 7), not all flexural sequences can be expected to be so complete. In some cases, a new tectophase may begin before the flexural relaxation responses of the previous one are finished, and in other instances, erosion from the initial phase of bulge uplift and migration of a succeeding tectophase may destroy large parts of the stratigraphic response of a previous one.

In the past, the third-order cyclicity reflected in unconformity-bounded stratigraphic sequences like those in Ordovician has been largely attributed to eustatic variations. However, the proximity and contemporaneity of these sequences to the Taconian orogeny suggest that a large component of such cyclicity must be related to concurrent tectonism. Moreover, with periodicities on the order of millions of years it is difficult to relate these cycles to known eustatic cyclicity or variations caused by Milankovitch orbital perturbations. This does not mean, however, that these influences are absent. In fact, cycles of lesser order have been clearly superimposed on the larger flexural cycles, and we will see possible examples of such cycles and discuss their likely origin on Days 1 and 3.

A PALEOZOIC SUBSIDENCE HISTORY OF THE AUTOCHTHONOUS APPALACHIAN BASIN IN KENTUCKY

PETER T. GOODMANN

INTRODUCTION

Analytical techniques capable of separating the effects of sedimentary loading, tectonics and eustasy were employed on an autochthonous Appalachian basin sequence in eastern Kentucky in order to isolate probable tectonic components of subsidence during the Paleozoic. Similar studies have been conducted in passive- and rifted-margin basin settings (Steckler and Watts, 1980; Schlater and Christie, 1980) and foreland basin settings (Bond and Kominz, 1984, 1991; Kominz and Bond, 1991; Heidlauf and others, 1986; Beaumont and others, 1988; Goodman, 1992) in order to determine the effects of tectonism on the subsidence history of the basin versus the effects of sedimentary loading and eustasy.

The present study was conducted in the Cumberland Plateau region of the Appalachian basin in eastern Kentucky (Fig. 10) at the edge of the Appalachian orogen. The basin contains a stratigraphic sequence of Paleozoic sedimentary rocks ranging from Lower Cambrian to

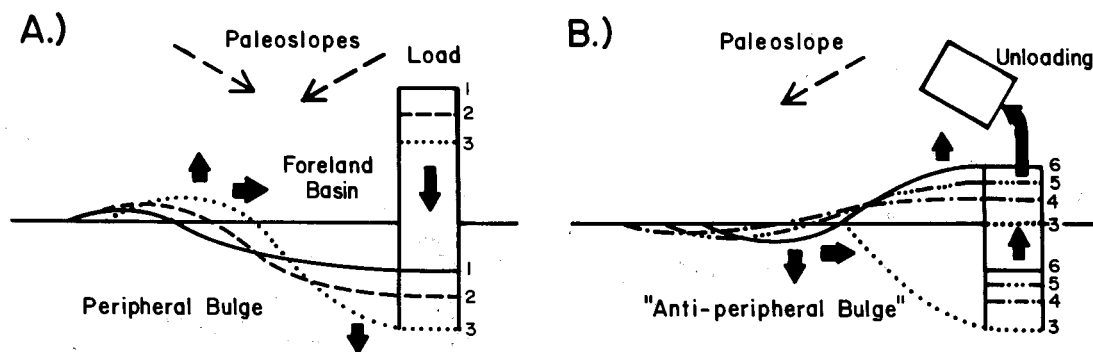


Figure 8. Two types of flexural response to lithospheric-stress relaxation: A) "loading-type" relaxation, static gravitational load results in deepening foreland basin and migration of peripheral bulge toward the load; B) "unloading-type" relaxation, erosional unloading results in rebound near unloaded area and an "anti-peripheral" bulge that deepens and migrates toward the former load (redrawn from Beaumont and others, 1988).

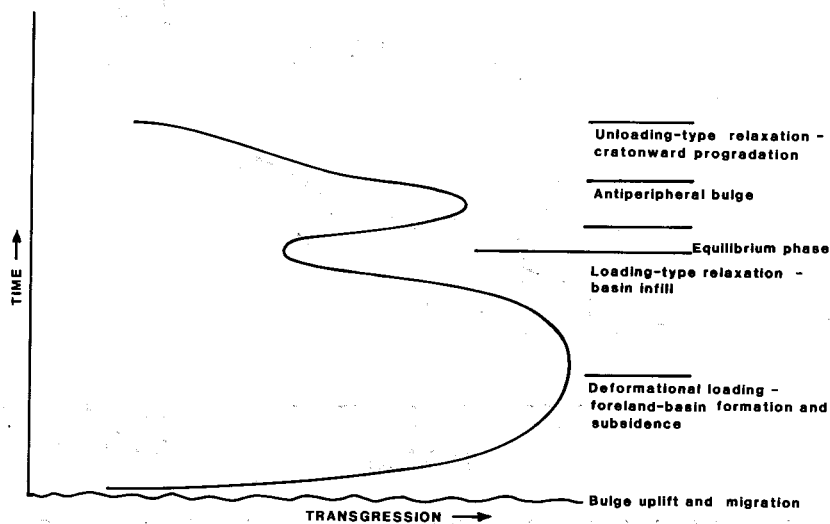


Figure 9. Curve showing the nature of transgressive and regressive events in the foreland basin and accompanying flexural events in the orogen during a single completed tectophase.

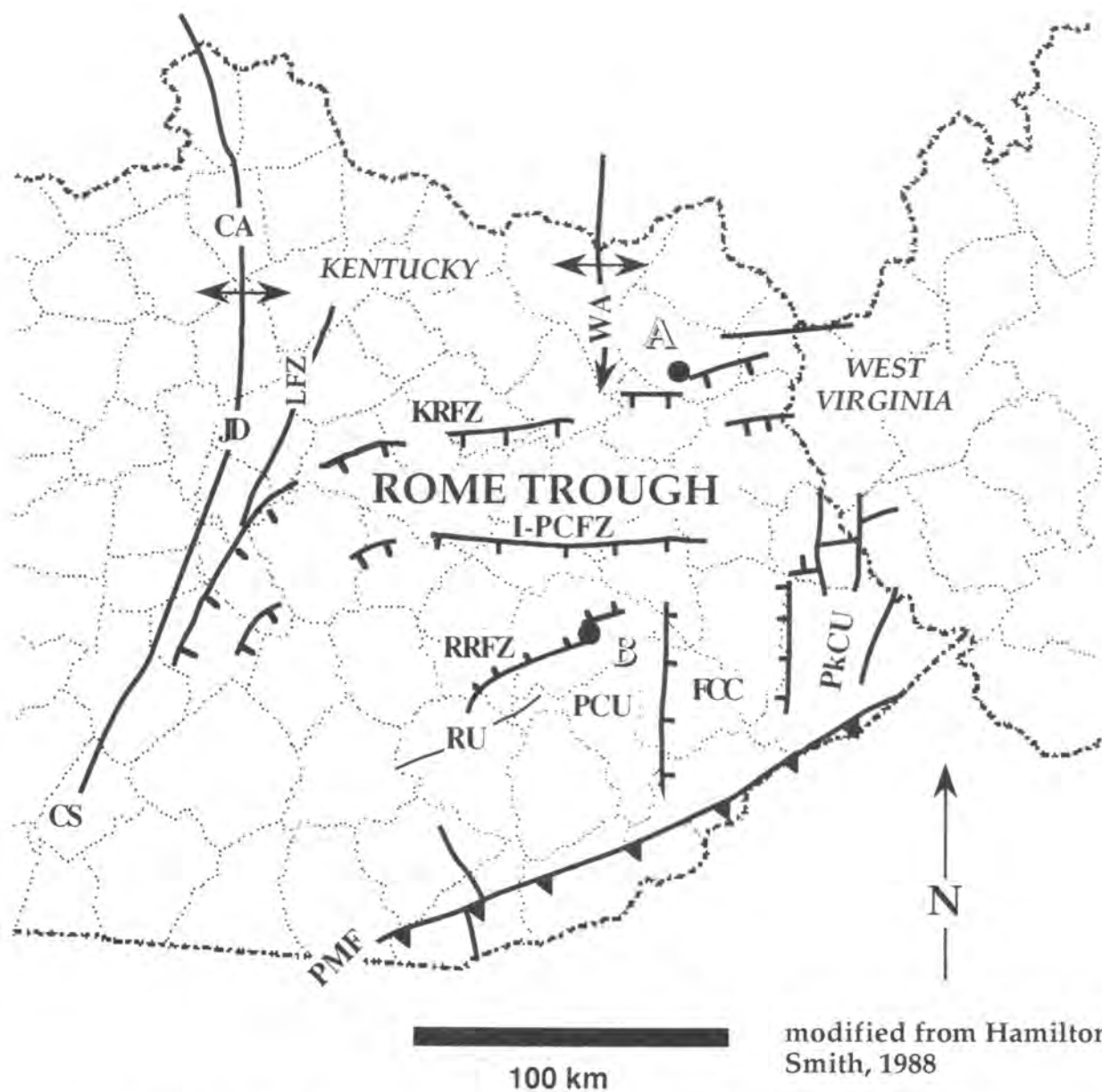


Figure 10. Major structural elements of the autochthonous Appalachian basin in Kentucky: CA=Cincinnati arch, JD=Jessamine dome, CS=Cumberland saddle, WA=Waverly arch, LFZ=Lexington fault zone, KRFZ=Kentucky River fault zone, I-PCFZ=Irvine-Point Creek fault zone, RRFZ=Rockcastle River fault zone, RU=Rockcastle uplift, PCU=Perry Co. uplift, FCC=Floyd Co. channel, PkCU=Pike Co. uplift, PMF=Pine Mountain fault. Reference points used in study: A=Ashland-Cabot Warnie Stapleton #1 well, Carter Co., Ky; B=United Fuel Gas Williams #1 well, Breathitt Co., Ky.

Upper Pennsylvanian. The region has had a history of predominantly vertical tectonic displacements. Activity on local and regional structures (Fig. 10) from the development of a regional trough to sporadic reactivation of local structures, influenced accumulation of sediments in this region throughout the Paleozoic (Dever and others, 1977, 1990; Cable and Beardley, 1984; Black, 1986).

The purpose of this study is to decipher the nature of tectonic influence on the Paleozoic stratigraphic succession in the Appalachian basin in eastern Kentucky. Particular emphasis lies on the effects of the various Paleozoic orogenies on subsidence and stratigraphic accumulation in this basin. The results suggest that several episodes of Paleozoic subsidence were caused by flexural downwarping of the lithosphere and are correlative with major orogenic events including the Taconic orogeny, possibly the Salinic disturbance, the Acadian orogeny, and the Alleghanian orogeny. A major subsidence event in the Early to Middle Cambrian was largely confined to regions within a graben-like structure known as the Rome trough and is thought to be a consequence of lithospheric thermal contraction associated with late-lapetan extension.

METHODS

Methods used in this study have been outlined in detail by Van Hinte (1978), Schlater and Christie (1980), Bond and Kominz (1984), and Angevine and others (1990). The procedure involves: 1.) sequential decompaction of the stratigraphy to produce a cumulative "total" subsidence curve, and 2.) sequential backstripping of the decompacted sedimentary sequence allowing the basement to rebound assuming Airy isostasy. Backstripping procedures generate the tectonic subsidence curve, which represents subsidence in the basin not attributable to sedimentary loading and variations in eustasy (Levy and Christie-Blick, 1991). Local isostatic compensation was assumed and the effects of flexure and lateral heat flow were not considered as it has been shown that "for continental margins wider than 100 km ignoring flexure and lateral heat flow does not change the overall shape of the subsidence curve" (Levy and Christie-Blick, 1991, p. 1591), although it does affect the rates of subsidence represented by the curve (Bond and others, 1988; Bond and Kominz, 1984; Beaumont and others, 1982; Steckler and Watts, 1980).

Stratigraphic successions in two basement test wells were used in this study: 1.) the Ashland-Cabot Warnie Stapleton #1 well in Carter County, Kentucky (Fig. 10A); and 2.) the United Fuel Gas Williams #1 well in Breathitt County, Kentucky (Fig. 10B). Thicknesses and average lithologies for each unit were estimated from geophysical logs. Compaction constraints for represented lithotypes were taken from Schlater and Christie (1980), Magara (1980), Schmoker and Hally (1982), Butler and Baldwin (1985), Scherer (1987), and Schmoker and Gautier (1988). Mean water depths for each formation were based on facies interpretations and techniques for estimating minimum water depth (Allen, 1967; Klein, 1974; Etensohn, 1990a). No quantitative eustatic component has been assigned in these computations; however, results are discussed relative to existing sea-level curves where appropriate.

A thickness of 3 km of sedimentary rock was decompacted and backstripped to account for rock previously removed by post-Paleozoic erosion. This minimum thickness was obtained by calculating the amount of overburden necessary to produce high-volatile bituminous coal, which occurs at the surface in the study area, at a geothermal gradient of 60°C/km (O'Hara and others, 1990). The lithology of this overburden was estimated by averaging the lithologies of the upper

Breathitt Formation, and the Conemaugh, Monongahela, and Dunkard groups and any possible Triassic sedimentary rocks that were deposited in this area.

The geochronological time scale used in this study is adapted from work by McGhee and Dennison (1980), Odin (1982), Palmer (1983), Cowie and Bassett (1989), Menning (1989), Rice and others (1990), and Harland and others (1990). Most of the stratigraphic succession of this basin was correlated with the isotopic time scale using biostratigraphy. The younger Early and Middle Cambrian dates (Compston and others, 1990) used result in a younger age of the onset of Paleozoic subsidence in eastern Kentucky and thus higher rates of subsidence during the Cambrian.

The total and the tectonic subsidence curves for both wells are presented in Figure 11. Subsidence curves are plotted as subsidence in meters versus time in millions of years. The total subsidence curve equals the total thickness of decompacted strata accumulated at a given time and represents the sum of all the contributing factors to basin subsidence, including: sedimentary loading, tectonic (thermal and/or flexural) influence on the lithospheric basement, and water depth. The tectonic subsidence curve represents the proportional contribution of tectonic movements and water depth on total subsidence. The difference between the total and the tectonic subsidence curves is the subsidence attributed to sedimentary loading.

Total subsidence and tectonic subsidence calculations are not affected by uncertainties in biostratigraphic-age control, however, subsidence rates are sensitive to uncertainties in the time scale. Rates are also sensitive to an assumption of Airy isostasy.

Uplift indicated on the tectonic subsidence curve does not preclude a cessation of basin subsidence, as sedimentary loading may produce overall subsidence. Additionally, changes indicated in the tectonic subsidence curve are affected, in part, by changes in water depth, by sedimentary filling or starvation, and by eustatic fluctuations.

Similarities between the subsidence curves (Fig. 11) from the two reference points, which occur on separate faults blocks (Fig. 10), indicate that the subsidence events are the effects of regional processes and not local structural phenomena. The shape of the tectonic subsidence curves is used to interpret the nature of tectonic influence on basin subsidence. Subsidence resulting from thermal contraction of the lithosphere is indicated by a subsidence curve of exponential decay (Angevine and others, 1990; Levy and Christie-Blick, 1991) which mimics thermal cooling curves for the lithosphere. Tectonic subsidence produced by flexural downwarping of the lithosphere seems irregular in magnitude and rate and is commonly preceded by the development of an unconformity in the area of study. The development of an unconformity, prior to flexural subsidence, may have resulted from cratonward migration of a peripheral bulge (Quinlan and Beaumont, 1984).

RESULTS AND DISCUSSION

lapetan Extension

A major subsidence event in the Early through Middle Cambrian (Fig. 11) affected deposition of the Rome-through-Copper Ridge formations and was largely stabilized by the deposition of the Rose Run-Beekmantown sequence. The exponential shape of the tectonic subsidence curve mimics thermal cooling curves for the lithosphere

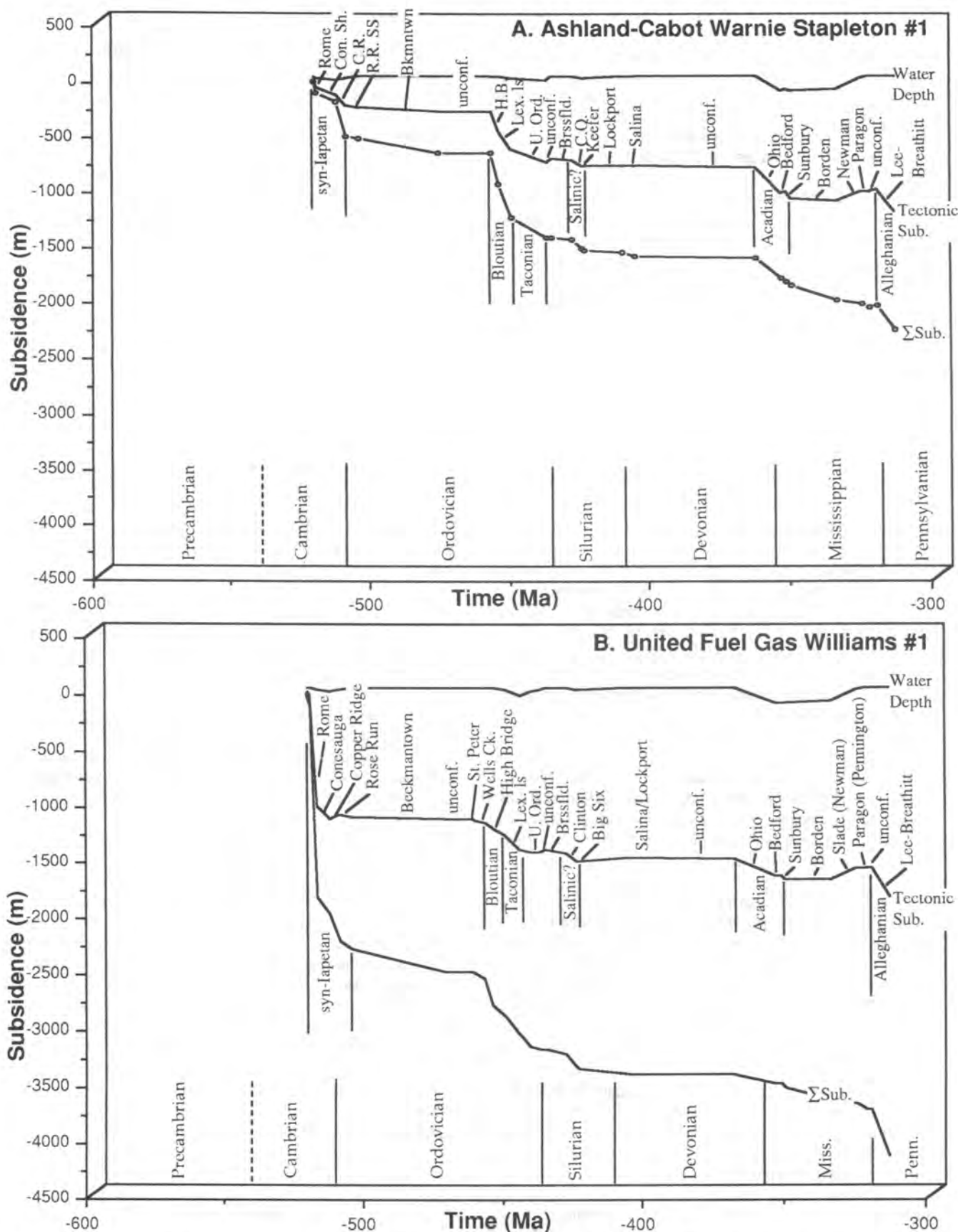


Figure 11. Tectonic and total (Σ) subsidence curves for the Ashland-Cabot Warnie Stapleton #1 well (A), Carter Co., Ky., and the United Fuel Gas Williams #11 well (B), Breathitt Co., Ky. The isotopic and geological time scales are presented on the abscissa and subsidence is shown in meters on the ordinate.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

(Angevine and others, 1990; Levy and Christie-Blick, 1991). This mimicry suggests that the subsidence resulted from thermal contraction of the lithosphere during a period of late lapetan lithospheric extension (Fig. 12), as discussed by Thomas (1991), Walker and Driese (1991) and Rast and others (1992). The Cambrian strata in the UFG Williams #1 well, located in the structurally downthrown Rome trough (Fig. 10), are vastly thicker and somewhat older than the Cambrian strata in the Ashland-Cabot Warnie Stapleton well, located outside the margins of the Rome trough (Fig. 10). The difference in the magnitude of subsidence between the two reference points and the stratigraphic growth, in particular of the Rome and Conesauga formations in the UFG Williams #1 well, indicate that the subsidence was largely confined to areas within the Rome trough (Fig. 12). The effective restriction of tectonic subsidence to structural troughs is common in syn-rift tectonics. Thus, the confinement of most Cambrian tectonic subsidence to the regions within the Rome trough, as well as the shape of the tectonic subsidence curve and previous geological interpretations, suggests that this subsidence event and the structural evolution of the trough are synchronous with late stages of lapetan rifting (Thomas, 1991; Walker and Driese, 1991; Rast and others, 1992).

Blountian-Taconian

The relatively stable period from the Late Cambrian to the Early and Middle Ordovician was interrupted by uplift with the onset of the Blountian orogeny (Chapple, 1973; Jacobi, 1981; Mussman and Read, 1986; Ettensohn 1991a) in the Middle Ordovician and the development of the regional "Knox unconformity." This unconformity may have resulted from the formation and diachronous migration of a peripheral bulge, accompanying the development of a new foreland basin at the edge of an orogen. This unconformity, which separates the Sauk and Tippecanoe sequences of Sloss (1963), may have been enhanced in the study area by lowered sea level (Mussman and Read, 1986).

The development of the unconformity was followed by foreland subsidence accompanying flexural depression of the lithosphere resulting from orogenic loading of the crust in the Blountian orogen during the Middle Ordovician (Rast, 1989; Ettensohn, 1991a). Crustal loading in the Taconic orogen in the late Middle and Late Ordovician (Rast, 1989; Ettensohn, 1991a) also produced subsidence in the eastern Kentucky area and resulted in the accumulation of thick carbonate and fine-grained clastic rocks.

The interpretation of Ordovician subsidence in eastern Kentucky as a flexural response of the lithosphere to Blountian and Taconian crustal loading is based on the occurrence of this event at both reference points, the difference in the magnitude and pattern of subsidence between the two reference points relative to their proximity to the orogen (Fig. 11), and previous regional stratotectonic interpretations (Ettensohn, 1991a).

Ordovician subsidence affected the stratigraphic sequence at both reference points (Fig. 11) from the Middle Ordovician High Bridge Group through the Upper Ordovician Richmond Group. Subsidence in the Breathitt County well (Fig. 11) may reflect two separate events of subsidence during the Ordovician. The first of these two events affected the High Bridge Group and Lexington Limestone. A decrease in the slope of the tectonic subsidence curve (Fig. 11B) for the late Middle Ordovician (Lexington Limestone) indicates that a minor tectonic uplift may have influenced this area in the Middle Ordovician, possibly reflecting reactivation of the peripheral bulge in this area

during orogenic loading in the Taconian orogeny. In the Carter County well, a single, continuous Ordovician tectonic subsidence event, effective from the High Bridge Group through Upper Ordovician strata (Fig. 11A), is recognizable.

The similarity of the subsidence curves between the reference points indicates that the Ordovician subsidence which occurred in the area was a regional effect of the Blountian and Taconian orogenies and not a local structural phenomenon. Throughout most of the Middle and Late Ordovician, the study area was probably situated near the cratonic edge of the Blountian and Taconic foreland basins. The entire sequence consists of limestones and fine-grained clastics. No coarse-grained, molasse-type clastics occur within the study area. The craton margin of this foreland basin was subject to uplift associated with the development and migration of the peripheral bulge.

Salinic (?)

A minor tectonic subsidence event occurred at both reference points in the late Early to Middle Silurian (Fig. 11). It is suggested that this subsidence event and deposition of the upper Brassfield, Crab Orchard/Clinton and the Bisher/Keefer formations may have been affected by the putative Salinic disturbance (Boucot, 1962; Hatch, 1979). The tectonic subsidence curve indicates a modest flexural downwarp at this time, possibly due to orogenic loading during the so-called Salinic disturbance. However the tectonic subsidence curve may reflect a component of eustatic sea-level rise effecting greater water depths during this time. In any case, serious work with regard to the effects of the Salinic disturbance on the stratigraphic sequence in Kentucky and adjacent areas is only now underway.

Acadian

The Taghanic unconformity (Fig. 11) is a composite of two or more Devonian unconformities which splay basinward to the east (Ettensohn, 1985a, 1987, Ettensohn and others, 1988b; Hamilton-Smith, 1988, in press; Hamilton-Smith and others, 1991.). Hamilton-Smith (in press) maintains that this unconformity represents uplift and erosion of this area on the peripheral bulge to a foreland basin formed during two tectophases of the Acadian orogeny. Ettensohn (1985a) proposed that the unconformities splay basinward and separate three major clastic sequences which represent three diachronous tectophases of orogenic activity in the Acadian orogen. These tectophases produced distinct foreland-basin sequences which followed peripheral-bulge formation and accompanied basin subsidence.

A late Devonian tectonic subsidence event is recognizable at both reference points (Fig. 11). This event is interpreted to be the result of a flexural depression of the lithosphere in response to a phase of orogenic loading in the Acadian orogen (Hamilton-Smith, in press). Ettensohn (1985a) maintained that the sedimentary effects of three tectophases are discernible in this sequence, and that the sedimentary expression of these tectophases becomes more complete basinward. However, at present, limitations on the resolution of this data allow the recognition of only a single tectonic subsidence event in this sequence, that reflected by the Ohio and Bedford shales.

The Mississippian Sunbury Shale and Borden Formation represent most of Ettensohn's fourth and final Acadian tectophase. The Sunbury and the Borden are foreland-basin-fill units. The results suggest that there is minor tectonic subsidence associated with the Sunbury and little or no tectonic subsidence associated with the

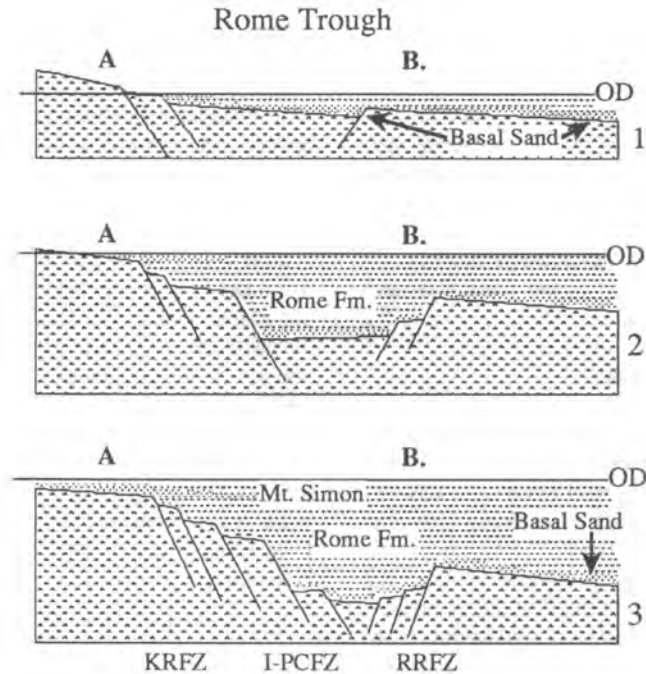


Figure 12. Schematic north(left)-south(right) sectional diagram showing the subsidence history of the Rome trough region in Kentucky. The positions of the reference points: the Ashland-Cabot Warne Stapleton #1 well (A), and the United Fuel Gas Williams #1 well (B), are shown to illustrate their different subsidence histories throughout the evolution of the structural trough. KRFZ=Kentucky River fault zone, I-PCFZ=Irvine-Paint Creek fault zone, RRFZ=Rockcastle River fault zone. Basement faulting and stratigraphy are highly schematic. 1) Late Early Cambrian rifting of a southeastwardly dipping craton and deposition of the basal sand and lower Rome in the trough, including B and areas farther south, in contrast to the northern area, including A, which was exposed. 2) Progressive rifting in the Middle Cambrian produced markedly greater subsidence and sedimentary accumulation within the fault-bounded trough, including B. During regional subsidence, transgression inundated A and the basal Mt. Simon Sandstone was deposited. 3) By the Late Cambrian most subsidence due to regional extension had ceased. The differential subsidence between A and B was due to their relative position on a passive continental margin and not due to their position relative to the structural trough.

Borden. However, tectonic subsidence for the Sunbury may be underestimated by employing a conservative minimum bathymetry (Ettensohn, 1984) for this unit. The results are consistent with Ettensohn's (1985a) interpretation that the Sunbury began to fill a tectonically subsiding basin and that the Borden basin fill is accompanied by tectonic uplift due to relaxation of the lithosphere in this area. However, Hamilton-Smith (in press) maintains that Acadian orogenic loading and foreland tectonic subsidence had ceased by this time. Recent work by Bond and Kominz (1991) and Kominz and Bond (1991) suggests that a significant amount of "tectono-eustatic" sea-level rise occurred in the Early Mississippian. This eustatic rise, indicated by comparison with their "Iowa base line," may account for a significant portion of the water-depth estimations for the Sunbury and Borden formations. Because numerical eustatic corrections have not been made, the magnitude of Early Mississippian tectonic subsidence may be overestimated. Further work using quantified eustatic components may help to isolate the effects of tectonics and eustasy on subsidence.

The effects of the Acadian orogeny in the subsidence history of the Appalachian basin in eastern Kentucky include an extensive, Early-Middle Devonian unconformity which formed during the development and migration of a peripheral bulge during Acadian foreland-basin formation. Foreland-basin subsidence in this area apparently began in the Late Devonian in response to progressive orogenic loading, and subsequent flexure of the lithosphere, and continued into the earliest Mississippian. Apparent Early Mississippian tectonic subsidence in the basin in eastern Kentucky may reflect either tectonic and eustatic components or a purely eustatic component. More quantitative estimates of eustatic change will be necessary to determine which component predominates.

Alleghanian

The apparent tectonic uplift indicated in the Mississippian carbonate sequence reflects, to some degree, a bathymetric change from deeper mean water depths reflected by the Borden Formation to shallower mean water depths represented by the Slade (formerly Newman) and Paragon (formerly Pennington) formations (Fig. 11). Facies changes from prodelta and delta-front facies of the Borden Formation to the shallow carbonate-shelf and peritidal facies of the Slade Formation indicate decreasing water depth and loss of a clastic source. Further shallowing and the recurrence of clastics in the area is represented by the littoral red-bed facies of the Paragon Formation. The subsequent formation of the putative Early Pennsylvanian unconformity (e.g., Englund, 1979; Englund and others, 1979; Ettensohn and Chesnut, 1989) indicates that tectonic uplift outstripped the bathymetric changes and a dynamic surface of erosion and deposition formed. It is proposed that this unconformity resulted from the development and cratonward migration of a peripheral bulge in front of a cratonward encroaching Alleghanian foreland basin (Ettensohn and Chesnut, 1989). This interpretation is consistent with these preliminary results of subsidence analysis in eastern Kentucky. It is proposed here that the putative unconformity and subsequent basin subsidence occurred in response to repeated forward thrusting (Rast, 1984).

Late Paleozoic subsidence resulted from flexural downwarp of the lithosphere in response to orogenic, craton-margin loading by progressively cratonward overthrusting in the Blue Ridge province and fold-thrust belt. The subsidence event occurred at both localities and was a regional effect of the Alleghanian orogeny. Although the ensuing subsidence rates were high, the Alleghanian foreland basin in this region was virtually filled with upper and lower delta-plain

clastics, and the most distal margins of the foreland basin were littoral. Although in the diagrams subsidence appears to have been abruptly terminated (Fig. 11), if the conservative estimate of 3 km of overburden is plotted, the total tectonic-subsidence components associated with the Alleghanian orogeny are nearly equal to that of the entire Paleozoic prior to the early Pennsylvanian. Thus the Alleghanian orogeny had the greatest influence of any Paleozoic orogeny on the subsidence in this area of the Appalachian basin.

CONCLUSIONS

The stratigraphy of the study area records several Paleozoic subsidence events. An initial subsidence in the late Early through Middle Cambrian probably resulted from thermal subsidence following late-stage syn-lapetan rifting in the Rome trough. This subsidence was confined to regions of the Rome trough where a much thicker and somewhat older stratigraphy is preserved. At least three Paleozoic stratigraphic sequences of Ordovician, Devonian and Pennsylvanian age record foreland-basin development and deposition. The foreland-basin sequences were deposited in basins that subsided in response to lithospheric flexural downwarping produced by craton-margin orogenic loading.

Each foreland-basin subsidence event was preceded by tectonic uplift and the development of local and regional unconformities. These unconformities formed on peripheral bulges which developed and migrated cratonward with the formation of adjacent foreland basins. Eustatic sea-level fall, possibly coincidental with peripheral-bulge formation and migration, may have enhanced the development of the unconformities. Subsidence was effective in some areas of the foreland basin, whereas simultaneous uplift on the peripheral bulge affected the stratigraphic sequence in adjacent areas of the basin. Thus unconformities are diachronous and give way locally to conformable sequences.

GENERAL ORDOVICIAN PALEOGEOGRAPHIC AND TECTONIC FRAMEWORK FOR KENTUCKY

FRANK R. ETTENSOHN

During the Middle-Late Ordovician, Kentucky was situated at about 20° to 25° south latitude and some 300 to 800 km west of the mountainous eastern coastline of Laurentia (Scotese, 1990) (Fig. 13). Although much of the eastern coastline was a highland area due to Taconian continental margin-island arc collision, the most mountainous areas, Blountia and Taconica (Fig. 13), were situated at the Virginia and New York promontories respectively, where Taconian collision would have been most intense (Ettensohn, 1991a). Moreover, Kentucky was ideally located for the incursion of major subtropical storms and hurricanes on the basis of the 10° to 45° latitudinal range of modern storm trackways (Marsaglia and Klein, 1983; Duke, 1985). The projection of hurricane trackways into this area (Kreisa, 1981) is similar to that of the recent. What is even more interesting is that the low area between the Blountia and Taconica could have acted as a natural conduit to funnel major oceanic storms into the Kentucky-Ohio cratonic area (Fig. 13). The abundance of storm deposits in the Middle Ordovician Lexington Limestone and the overlying Upper Ordovician units (e.g., Meyer and others, 1981; Tobin, 1982; Grossnickle, 1985; Ettensohn and others, 1986) certainly attests to the likelihood of this scenario.

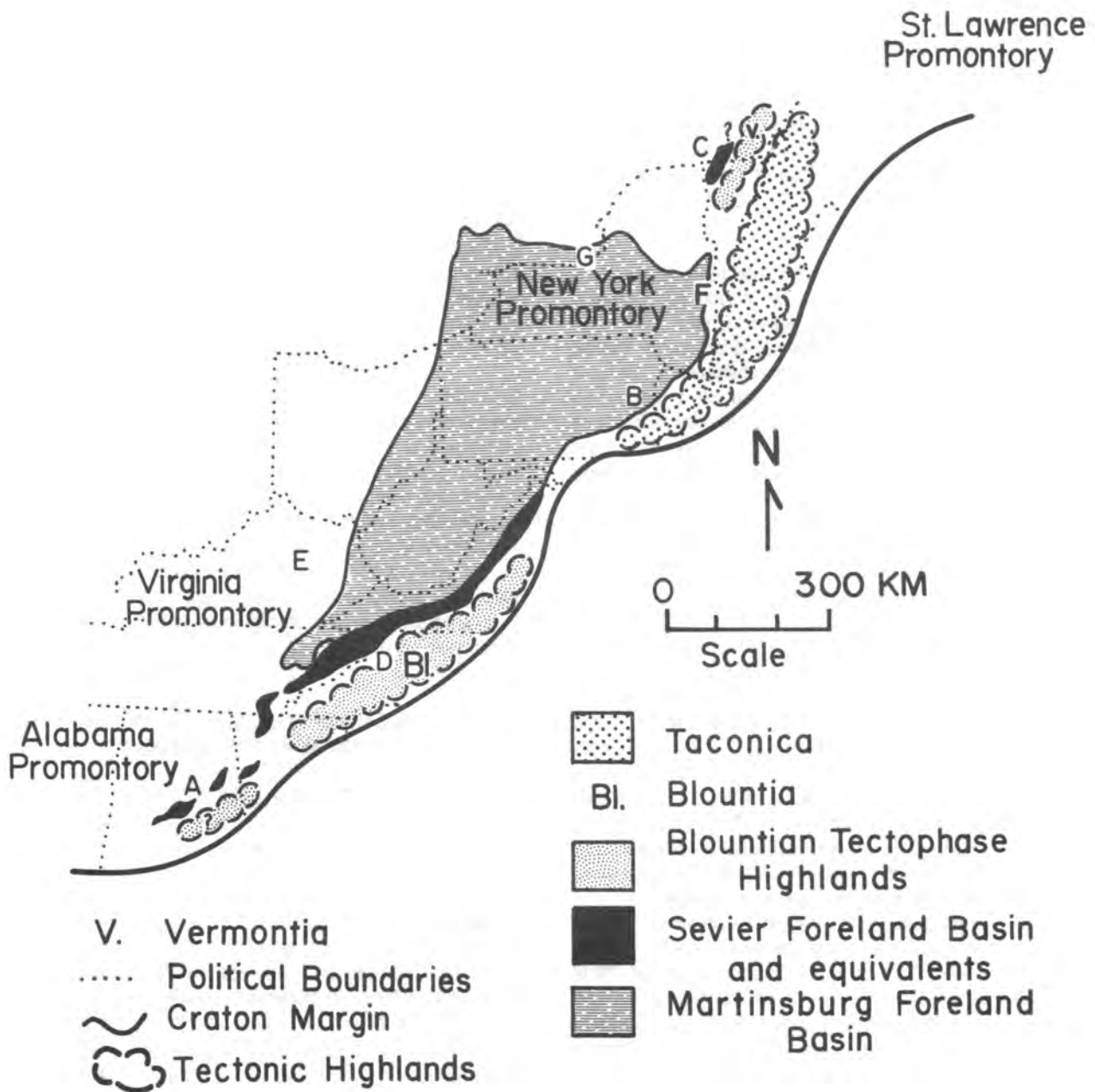


Figure 13. Generalized paleogeography of the east-central and northeastern United States during the Middle and Late Ordovician showing the positions of tectonic highlands and black-shale foreland basins relative to continental promontories. Note the northwestward and northeastward shift of the Martinsburg relative to the earlier Sevier basin. The section in Figure 7 is along line E-D (from Ettensohn, 1991a).

As previously discussed, the exposed Ordovician rocks in Kentucky represent parts of two large transgressive-regressive cycles which are in large part probably related to the two Taconian tectophases (Fig. 7). The oldest of these rocks, the Black Riverian High Bridge Group, including the Camp Nelson, Oregon, and Tyrone formations, will be examined at Stop 2 on Day 1. They largely represent the final regressive part of the transgressive-regressive cycle coinciding with the Blountian tectophase (Fig. 7). These rocks are typical of the shallow subtidal-to-peritidal carbonate platform that existed over much of the east-central United States at the time (Keith, 1988).

By Rocklandian time and the advent of the Taconic tectophase, this broad platform began to differentiate into a southeastern Lexington platform and a northwestern/northern Galena-Trenton platform. These carbonate-rich shelf and platform areas are separated by a poorly defined belt of argillaceous carbonate and shale running from western Tennessee to central Ohio (Fig. 14); this narrow belt probably represents a bathymetric low and has been called the Sebree trough (see Keith, 1988; Bergström and Mitchell, 1989). Although the origins of the trough are uncertain, the western boundary of the Lexington platform in west-central Kentucky and its north-central boundary near the Ohio-Kentucky boundary correspond approximately to the poorly defined western and northern margins of the isolation accommodation zone that may have formed in Kentucky during inboard, syn-lapetan rifting (Fig. 5). While inception of the Taconic tectophase may have resulted in the reactivation of the western margin of this zone as the western margin of the Lexington platform, the eastern margin of this platform was defined by the rapidly subsiding Taconic foreland basin (Fig. 14).

The rocks on the Lexington platform generally represent shallow, open-marine environments, but there is an overall deepening trend from the Rocklandian to the Edenian that probably reflects flexural subsidence and transgression accompanying the Taconic tectophase. The rocks on the platform also reflect deeper water environments to the north, west and east, but this change is a product of the transition from the platform to the deeper waters of the Sebree trough and the foreland basin (Fig. 14).

In the midst of the progressive deepening atop the Lexington platform, three locally regressive shoal areas persisted (Keith, 1988) (Fig. 14). The circumstances of the shoal area near the Kentucky-Virginia border are unknown, but the shoal near the Louisville area overlies a basement fault block called the Louisville high (Black, 1986). The shoal area on the Jessamine dome near the center of Kentucky occurs over a series of fault blocks that were apparently reactivated during the Taconic tectophase and acted independently of the rest of the platform. In fact, as we will see on Day 2, much of the Lexington Limestone on the Jessamine dome represents a high-energy, regressive, carbonate buildup that accumulated in the midst of an overall regional transgression during which deeper water muds and carbonates of the Kope and Clays Ferry formations were deposited.

By Edenian time, however, the entire platform—even the Lexington shoals—were inundated beneath the deeper water Clays Ferry and Kope muds, which will be examined on all three days of the trip. Between 30 and 60 percent of these formations may consist of fine-to coarse-grained limestones which probably represent storm deposits (Meyer and others, 1981). The major difference between the Kope and Clays Ferry formations is that the Kope contains more shale than the Clays Ferry and represents more distal, deeper, open-marine environments on the deepening northern flank of the Lexington platform.

By the beginning of the Maysvillian, an abrupt shallowing represented by the Calloway Creek and Fairview formations (Day 1, Stop 1; Day 3, Stop 8) signals the beginning of the regressive part of the flexural cycle and the probable uplift and eastward passage of the bulge as Taconic loading-type relaxation began.

Based on the facies distribution (Keith, 1988, Fig. 7), this bulge uplift and migration apparently destroyed the Sebree trough and much of the Lexington platform. However, the resulting east-west lithofacies trends (Weir and others, 1984) suggest that bulge migration may have reactivated similar-trending basement faults on the old graben systems, although overall deepening trends were still toward the north. These lithofacies and deepening trends continued from the Maysvillian until latest Richmondian times, when north-south lithofacies trends (Weir and others, 1984) suggest initial development of a continuous Cincinnati arch.

ROAD LOG AND STOP DESCRIPTIONS—FIRST DAY

Mileage:

cum. inc.

- | | | |
|-----|-----|---|
| 0.0 | | Beginning point at the north side of the Brent Spence Bridge as I-75 crosses the Ohio River. Proceed southward on I-75 in the middle lane. |
| 0.4 | 0.4 | Exit 192 to downtown Covington. For the next 1.4 mi the highway traverses Ohio River alluvium and Wisconsinan outwash. |
| 1.5 | 1.1 | Exit 190, Jefferson Avenue. Move from outwash terrace and begin ascent through the Upper Ordovician (Edenian) Kope Fm. on "Death Hill." Generally more than 85 percent of the unit is shale and these cuts are subject to severe slumping and sliding during rainy periods. Remaining parts of the unit are largely thin bedded, tabular limestones. During the last two years this part of the highway has been under construction to lessen the grade and decrease the number of accidents. |
| 2.1 | 0.6 | Exit 189B; pre-Illinoian outwash deposits below highway pass into the Upper Ordovician (Maysvillian and Richmondian) Bull Fork Formation except in some of the deeper valleys where the Bellevue Tongue of the Grant Lake Limestone and the Fairview Fm. are present. The Bull Fork consists of thin-bedded, fossiliferous limestones and interbedded shale. Limestone composes 50-60 percent of the unit. Typically the unit is poorly exposed and forms gently rolling hills. |
| 5.2 | 3.1 | Exit 186, Buttermilk Pkwy. |
| 6.6 | 1.4 | Exit 185 to I-275 and airport. |
| 7.1 | 0.5 | Overpass complex. |
| 7.3 | 0.2 | Exit 184, Donaldson Rd. to Erlanger. |
| 8.4 | 1.1 | Marydale Retreat Center on right; enter Boone Co. |
| 9.3 | 0.9 | Exit 192, Turfway Rd. to Florence and Burlington. |

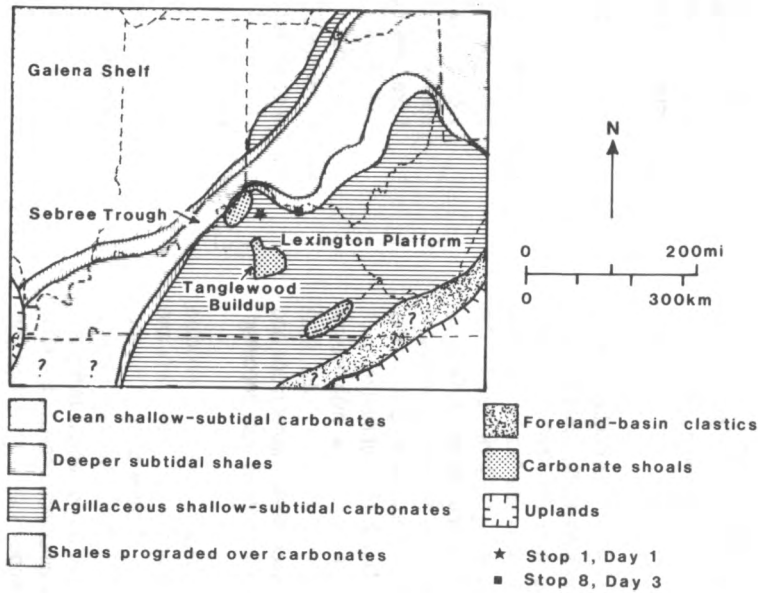


Figure 14. Generalized facies map for the east-central United States during the late Middle to early Late Ordovician showing the relative disposition of the Sebree trough and Lexington platform (modified from Keith, 1988).

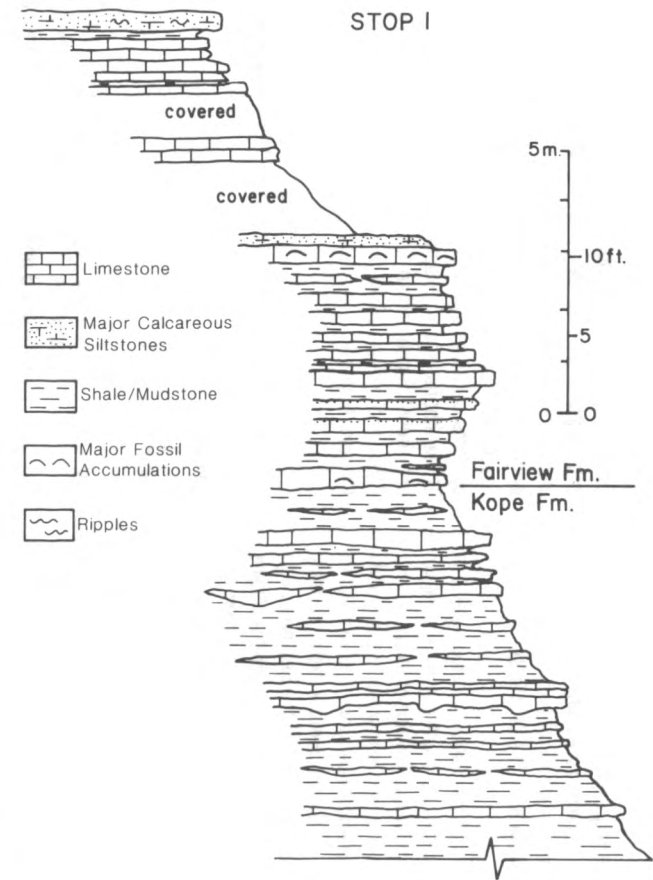


Figure 15. Generalized stratigraphic column for the Kope-Fairview contact at Stop 1, Day 1, near Williamstown, Kentucky (east side, southbound lane).

- 10.3 1.0 Exit 181 to Florence.
- 11.2 0.9 Exit 180A, Mall Rd.
- 11.7 0.5 Exit 180 to Florence and Union.
- 14.8 3.1 Exit ramp to rest area and welcome center.
- 16.3 1.5 Exit 175 to Richwood and Big Bone Lick State Park.
- 16.9 0.6 Approximate southern limit of glacial drift (Swadley, 1969). Small remnants of deeply weathered drift are exposed from place to place north of this boundary along I-75; the drift is thought to be of Kansan (Durrell, 1961) or Kansan and Nebraskan age (Ray, 1966).
- 18.7 1.8 Outcrops in the Bull Fork Fm. on the right and left.
- 19.3 0.6 Exit 173 to I-71.
- 20.6 1.3 Exit 171 to Walton and Verona.
- 21.0 0.4 For the next 14.9 mi from this point south to the area just north of Williamstown, I-75 crosses the Fairview Formation with the Bellevue Tongue of the Grant Lake Limestone capping some of the higher ridges and the Kope Fm. in some of the deeper valleys (Luft, 1973a, 1973b). The Fairview Fm. consists of thin-bedded limestones with interbedded shales and less common siltstones. Limestones compose 50-65% of the unit and are abundantly fossiliferous.
- 21.9 0.9 Typical exposure of the Fairview Fm. on the left.
- 22.1 0.2 Cross under CSX trestle.
- 22.6 0.5 Kenton Co. line.
- 23.6 1.0 Weigh-station entrance.
- 25.9 2.3 Grant Co. line.
- 26.0 0.1 Exit 166 to Crittenden; bridge and ramps on Bellevue Tongue of the Grant Lake Limestone and Bull Fork Fm.
- 28.1 2.1 Exposures in the Fairview Fm. on either side.
- 29.0 0.9 KOA campground on left.
- 30.1 1.1 Underpass (Sherman-Mt. Zion Rd.). Ridge capped by Bellevue Tongue of Grant Lake Limestone and Bull Fork Fm.
- 30.8 0.7 Contact of Fairview and Grant Lake formations on left.
- 33.5 2.7 Exit 159 to Dry Ridge.
- 25.6 2.1 Underpass (Baton Rouge Rd.); poor exposure in Fairview.
- 35.9 0.3 Excellent contact between the Fairview Fm. and underlying the Kope Fm. on the right. For the next 16.6 mi from this point to a point on I-75 north of Eagle Creek, the interstate crosses Kope Fm. with the Fairview Fm. capping some of the higher ridges and with tongues of the Clays Ferry Fm. and Lexington Limestone present in some of the deeper valleys (Luft, 1973b, 1976; Moore and Wallace, 1978). The Kope is typically 70-80% shale; remaining parts are largely thin-bedded limestones.
- 36.6 0.7 Underpass (Barnes Pike). Fairview-Kope contact.
- 37.8 1.2 Good exposure of Kope-Fairview contact on the right.
- 38.0 0.2 Exit 154 to Williamstown and Owenton
- 38.2 0.2 STOP 1. Fairview-Kope Contact and Implications, Williamstown, KY. (On the southbound lane of I-75, approximately 0.2 mi south of the Williamstown exit, Grant Co.; Carter Coordinates: 2550' FNLx1050' FEL, 15-Z-60).

FAIRVIEW-KOPE CONTACT AND IMPLICATIONS

FRANK R. ETENSOHN

This section consists of 23 ft (0.7 m) of the upper Kope Formation and 30 ft (9.2 m) of the overlying Fairview Formation (Fig. 15), and it is equivalent to parts of a larger section approximately 50 mi due east of here near Maysville which we will be examining on Day 3 of the trip (Fig. 14). The Kope and Fairview formations are commonly distinguished from each other based on the relative abundance of limestone and shale (e.g., Luft, 1973a, 1973b), although predecessors of the Kope Formation (Latonia and Eden formations) and the Fairview Formation were commonly recognized prior to the mid-1960's by their fossil content and to a lesser extent on lithology (e.g., Bucher and others, 1945; Caster and others, 1955).

In this area, the Kope is characterized by 75 percent or more of gray, laminated, silty calcareous shale. The remaining 25 percent or less of the formation consists of thin, even to irregularly bedded calcisiltites and calcarenites some with abundant whole or broken fossils. Many of the limestones exhibit erosional bases, shale rip-up clasts, megaripples and poor to excellent grading; scours and gutter casts are locally present at the bases. The typical graded sequence, generally no more than 1.3-ft (0.4-m) thick, starts with an erosional base. This is overlain by a fossil-fragment calcarenite, in places with imbricated brachiopods, which is succeeded by a calcisiltite or shaly siltstone, and shale. The calcisiltites and siltstones commonly show ripple laminae and are burrowed from the top. Amalgamated beds, in which a series of these truncated graded beds overlie each other in succession, are locally present.

In the overlying Fairview Formation, limestones compose more than 50 percent of the unit so that the formation is easily distinguished from more shaly Kope below. In fact, the tendency of the underlying Kope to slump and slide typically leaves the more resistant and abundant limestone ledges in the Fairview forming a slight overhang as is seen at this stop (Fig. 15).

Both calcisiltites and calcarenites are present in the Fairview, but calcarenites clearly predominate. The sedimentary structures in the Fairview are much the same as those in the Kope, except that amalgamated beds are fewer so that Fairview limestones appear to

be thinner. Also the siltstone portions of the graded sequences tend to thicken, and two of the major siltstones cap the two benches present here (Fig. 15). Although individual limestone beds contain transported, broken fossils, nearly in-place, brachiopod, bryozoan, and crinoid communities may occur on top of individual beds.

The basal limestone is a thin- to thick-bedded calcisiltite or calcarenite containing locally imbricated brachiopods. Although the contact appears to be sharp, in places it is subtly gradational with the Kope. The basal ledge is very unlike the thick, apparently high-energy brachiopod-rich unit that marks the base of the Fairview at Stop 8 on Day 3.

On a local scale, the relationship between the Kope and Fairview appears to represent a classical layer-cake situation, but on a regional scale, the Kope and Fairview, both in Kentucky and southwestern Ohio, intertongue across an interval of about 20 ft (6 m) (Ford, 1967, 1972; Luft, 1973b). Hence, the lithologic and environmental changes across the contact are not everywhere as abrupt as they appear at our field-trip stops. Moreover, on a regional scale there are also lithofacies changes within the Fairview and its southern equivalent, the Calloway Creek Limestone. In the north where we will view exposures on both Days 1 and 3, the Fairview contains only about 40 percent limestones whereas to the south the Calloway Creek contains nearly 70 percent limestone, some of which may have a nodular aspect (Weir and others, 1984). Fossils are more abundant and are more likely to be whole in the Calloway Creek. In many ways, the Fairview is very similar to the underlying Clays Ferry and Kope formations, except that it contains more limestones.

Sedimentary structures and sequences in the Kope and Fairview suggest that most of the limestones probably represent tempestites deposited in waning-flow regimes (e.g., Kreisa, 1981; Meyer and others, 1981; Tobin, 1982). The overlying shales then apparently reflect the normal background sedimentation or slow sedimentation of muds put into suspension during storms. The increased abundance of shales to the north in both units merely reflects the deeper waters and the more distal position of the area on the ramp dipping gently to the north away from the Lexington platform into the Sebree trough (Fig. 14). The Clays Ferry and Calloway Creek formations, the southern equivalents of the Kope and Fairview respectively, both contain more limestones than their northern equivalents, reflecting their shallower, more proximal positions on the Lexington platform. The greater abundance of fossils in these southern units, particularly in the Calloway Creek, may reflect the greater availability of firm carbonate substrates there.

The somewhat abrupt nature of the Kope-Fairview contact has been the subject of some discussion (see Schumacher, 1991), and will be examined again at Stop 8 on Day 3, where a massive, transported brachiopod-rich unit occurs at the base of the Fairview and may reflect some type of condensed horizon. Although the contact at this stop is somewhat different and certainly more subtle, a major change is still apparent. Some workers (e.g., Schumacher, 1991; Jennette and Pryor, in press) have suggested that the contact represents a cycle boundary or a parasequence-sequence boundary. Some have placed the boundary below the first major Fairview limestone, others above it. Of course, at this stop the basal Fairview limestone is very poorly developed and shows little evidence of condensation. Some of the differences between the basal Fairview deposits at these locales may reflect proximity trends. This locale was high on the Lexington platform (Fig. 14) where, because of decreased depths, storm sedimentation may have been more continuous and frequent. On the

other hand, by Edenian time the Maysville locality was apparently deeper near the margin of the Sebree trough (Keith, 1988, fig. 6) where starvation and condensation deposits may have been more likely.

The implications of a sequence-stratigraphy type interpretation are that the changes in depth and resulting accommodation space are largely the result of sea-level changes. Although the Late Ordovician was a time of glaciation during which such changes were possible, the Late Ordovician was also a time of tectonism. In terms of possible flexural responses, central Kentucky was situated well within the range of Taconian bulge migrations, and as already mentioned in a previous section, the overall late Middle-Upper Ordovician stratigraphic section has all the markings of a flexural sequence (Fig. 7). In fact, the abrupt contact between these two units could well represent the passage of a bulge, marking the advent of loading-type relaxation (Fig. 8A) and the resulting regional regression (Fig. 7).

No matter what interpretation one accepts, this contact marks the beginning of a major regression which continues to the end of the Ordovician. The length of the overall transgressive-regressive cycle (6 Ma) of which it is a part would seem to suggest the predominance of flexural mechanisms, but the presence of smaller cycles (parasequences?) in this part of the section, which we will examine on Day 3, also suggests some probable eustatic influence as well.

- | | | |
|------|-----|--|
| 38.4 | 0.2 | Milepost 154. |
| 42.4 | 4.0 | Milepost 150. |
| 47.7 | 5.3 | Exit 144 to Corinth and Owenton. |
| 48.5 | 0.8 | Scott County line. |
| 51.4 | 2.9 | Underpass (KY. 608). |
| 52.5 | 1.1 | Approximate contact between the Kope and Clays Ferry formations. For the next 9.7 mi, I-75 crosses the Clays Ferry Formation, which is composed of thin- to medium-bedded limestones interbedded with shale in equal amounts. In some of the deeper valleys along I-75, the Tanglewood Mbr. of the Lexington Limestone can be observed underlying and intertonguing with the Clays Ferry (Fig. 16) (Wallace, 1977; Moore and Wallace, 1978). |
| 53.0 | 0.5 | Flow rolls in the upper Clays Ferry on the northbound lane. |
| 54.9 | 1.9 | Exit 136 to Sadieville. Good exposure of Clays Ferry on both lanes; some intertonguing with Tanglewood is also present (Moore and Wallace, 1978). |
| 60.4 | 5.5 | Approximate boundary between the Inner and "Outer" (Eden Shale Belt) Bluegrass regions of the Lexington Plain Section (Fig. 3). From this point one can look southward from the higher Eden Shale Belt toward the lower, relatively level Inner Bluegrass developed largely on the Lexington Limestone. The Eden Shale Belt has a steep, more maturely dissected topography because it is developed on Upper Ordovician rocks in which nonresistant shales predominate. The poorly exposed |

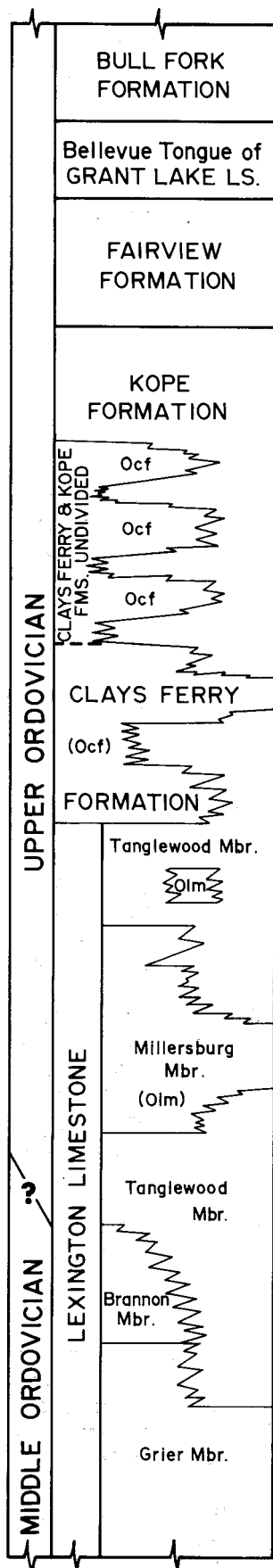


Figure 16. Generalized Middle and Upper Ordovician stratigraphic column for rocks along I-64 and I-75 from Lexington northward to Cincinnati. Kope and Clays Ferry formations are differentiated on their relative amounts of shale and limestone: in the Kope, shales, which may compose 75% or more of the formation, predominate; in the Clays Ferry, limestones, which may compose 30-60% of the formation, predominate. Amount of limestone decreases progressively in a northerly direction.

Clays Ferry Formation on either side of the road is typical of rocks in the Eden Belt.

61.0 0.6 Entrance ramp into weigh station. Approximate contact between the Clays Ferry Formation and the Lexington Limestone. For the next 27.9 mi, our route on I-75 and Man-O-War Rd. largely crosses the Tanglewood and Millersburg members of the Lexington Limestone which intertongue in a complex fashion. The Tanglewood is a light-gray, massive, coarse-grained calcarenite which is conspicuously crossbedded. The Millersburg consists of irregularly bedded nodular limestone in a matrix of gray shale. Locally in the deeper valleys along the highways, thin, irregularly bedded limestones and interbedded shales of the Grier and Brannon members of the Lexington crop out. In this area, these members underlie and intertongue with the Tanglewood (Fig. 16). Although the Lexington Limestone and High Bridge Group predominate throughout the Inner Blue Grass (Fig. 3), the Clays Ferry and Garrard formations may crop out on some of the hilltops as well as in some of the grabens on the Kentucky River fault system within the region.

62.1 1.1 Exit 129 to Delaplain Rd.

62.8 0.7 Milepost 129; intertonguing between the Millersburg and Tanglewood members of the Lexington Limestone.

63.5 0.7 Entrance to rest area.

65.2 1.7 Exit 126 (U.S. 62 and 460) to Georgetown and Paris.

67.2 2.0 Stromatoporoid zone in the Tanglewood Member of the Lexington Limestone (see Cressman, 1967); quarry to the left largely in the Tanglewood Member.

70.6 3.4 Exit 120 to the Kentucky Horse Park.

73.0 2.4 Exit 118 to I-64 northwest.

74.0 1.0 I-64 merges with I-75; prominent light band present in Tanglewood at Milepost 117.

75.5 1.5 Exit 115 to Newtown Road.

77.5 2.0 Exit 113 (U.S. 27 and 68) to Lexington and Paris.

79.4 1.9 Approximate western edge of Bryant Station fault zone, a narrow northeast-trending graben that is part of the larger Lexington fault system (MacQuown and Dobrovlny, 1968). The entire cloverleaf at the junction of I-64 and I-75 is located in the fault zone. Maximum displacement in this area is about 100 ft (30.5 m), so that the Upper Ordovician shales of the Clays Ferry Formation to the southeast are in contact with the massive coarse-grained limestones of the Tanglewood Member of the Lexington Limestone to the northwest (MacQuown and Dobrovlny, 1968).

79.7 0.3 Exit 111; separation of I-64 and I-75; proceed southward on I-75.

80.9 1.2 Exit 110 to Winchester Road (U.S. 60).

82.3 1.4 Merge right onto Exit 109 to Man-O-War Rd. Intertonguing of Tanglewood and Millersburg members (Stop 1 on Day 2).

82.6 0.3 End of exit ramp; turn right (southwest) toward Lexington.

83.3 0.7 Intersection with Todds Road; continue on Man-O-War Rd.

84.7 1.4 Intersection with Richmond Rd. (U.S. 25 and 421); continue on Man-O-War Rd.

88.9 4.2 Intersection with Bates Creek Rd. (KY 1974); continue on Man-O-War Rd.

90.9 2.0 Intersection with Nicholasville Rd. (U.S. 27); turn left onto Nicholasville Rd. and proceed southward. For the next 14.8 mi, our route largely crosses lower parts of the Lexington Limestone including the Brannon, Grier, and Curdsville members (Fig. 1).

91.3 0.4 Waveland Historical Shrine to the right.

91.8 0.5 Jessamine County line; continue southward on U.S. 27.

96.2 4.4 Veer right onto Bypass 27 around Nicholasville. All of the outcrops along the bypass are in the Grier and Curdsville members of the Lexington Limestone.

101.3 5.1 Leave the bypass and proceed southward on U.S. 27; outcrops of the Curdsville Member to the right.

105.1 3.8 Junction with Kentucky State Route 1268 to Wilmore; proceed southward on U.S. 27.

105.5 0.4 Camp Nelson National Cemetery on the left.

105.7 0.2 Distillery on the right. STOP 2A. Contact of the High Bridge Group with the Lexington Limestone and Implications, Camp Nelson, Kentucky (Southbound lane of U.S. 27, approximately 4.4 mi south of its junction with Bypass 27, Jessamine Co.; Carter Coordinates: 1300° FSLx2100° FEL, 12-P-59).

CONTACT OF THE HIGH BRIDGE GROUP WITH THE LEXINGTON LIMESTONE AND IMPLICATIONS

FRANK R. ETTENSOHN

The Middle Ordovician High Bridge Group contains the oldest exposed rocks in Kentucky, and the Tyrone Formation exposed here is the youngest formation in the group. Approximately 58 ft (17.7 m) of the upper Tyrone is exposed here, and the Tyrone is capped by the Mud Cave K-bentonite (Fig. 17) which is equivalent to the Millbrig K-bentonite of Illinois (Kolata and others, 1986). The Tyrone is composed largely of light-gray dense calcilutite displaying abundant cryptalgal laminae, prism cracks, mud-crack polygons, spar-filled vertical burrows, mud-chip conglomerates and breccias, and fenestral fabric. In fact, the unit was originally called the Birdseye limestone

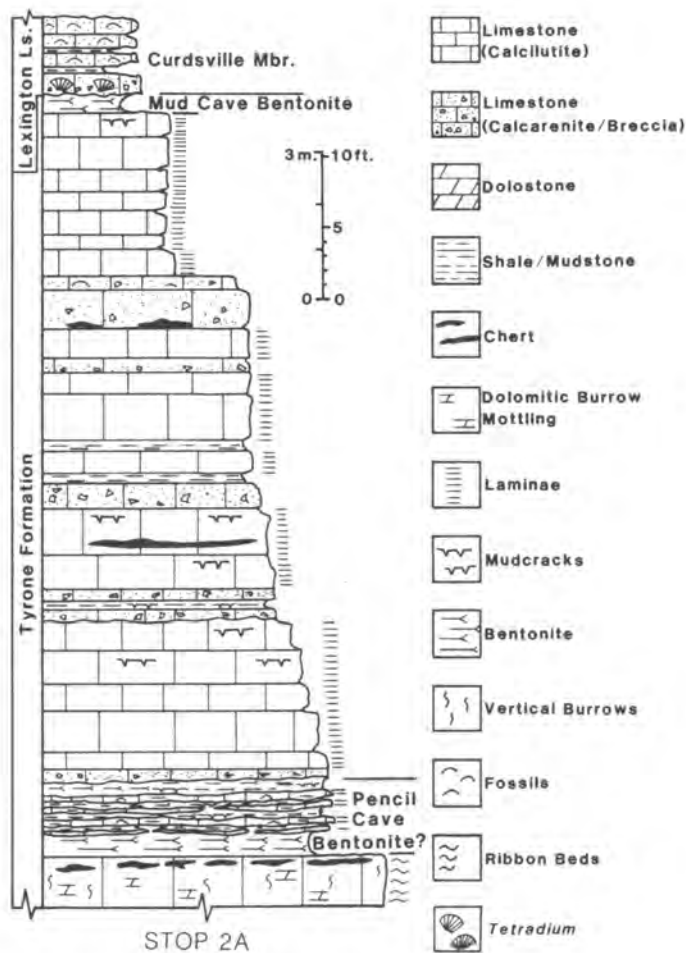


Figure 17. Generalized stratigraphic section at the Tyrone-Lexington contact at Stop 2A. Note the small shallowing-upward cycles beginning with calcarenites or breccias and the two major bentonites in the Tyrone.

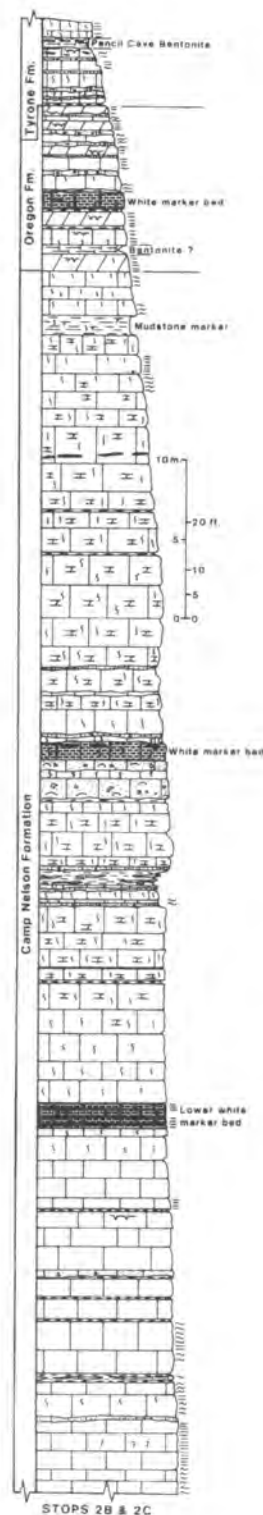


Figure 18. Generalized stratigraphic section of the High Bridge Group at Stops 2B and 2C. The lower white marker bed and the prominent shale near the middle of the exposed Camp Nelson may be the flexural equivalents of the Bowen and Witten respectively in the Tennessee section (Fig. 7). The darkened brick pattern reflects prominent, white, peritidal, calcilutite beds. The lower one is the best known and most widely traced.

(Owen, 1857). Laterally linked hemispheroid "pincushion" stromatolites are present locally, and a fragmented molluscan-tabulate coral fauna is also present locally, especially in some of the breccia layers. A more open-marine fauna including brachiopods and bryozoans is present in some of the argillaceous zones within the unit like the one near the base of this exposure (Fig. 17). The Tyrone appears to be made up of a series of shoaling-upward cycles, each beginning with a breccia containing marine fossils and ending with laminites containing a concentration of exposure features (Fig. 17). The cycles, however, cannot be traced from exposure to exposure.

The Tyrone also commonly contains several bentonites, two of which are present in this exposure (Fig. 17). Only two of these bentonites, the Mud Cave and Pencil Cave, are well known and widely traced. Bedded or nodular chert is commonly found below many of the bentonites and may be related to the downward transport of silica from them. The uppermost bentonite in the Tyrone, the Mud Cave Bentonite, is generally considered to be the top of the unit, and the Curdsville Member of the Lexington Limestone is thought to disconformably overlie the Tyrone (Cressman and Karklins, 1970). As at this locality, the disconformity may be very subtle and probably reflects a submarine erosional diastem of short duration. In some places, however, the erosion at this surface is on the order of a few meters and the bentonite is absent.

Approximately five feet (1.5 m) of muddy, coarse-grained calcarenite to calcirudite of the Curdsville Member is present in the exposure. In addition to reworked clasts from the underlying Tyrone, the basal ledge contains fragmented *Tetradium* colonies, some of which are normally oriented, and others of which are overturned.

In the lower parts of this exposure, a second major bentonite is present. This probably represents the Pencil Cave K-bentonite (Fig. 17), which is equivalent to the Diecke K-bentonite of Illinois (Kolata and others, 1986).

As previously mentioned in introductory sections, the High Bridge Group represents nearly the entire transgressive-regressive flexural cycle in central Kentucky (Fig. 7), which corresponds to the Blountian tectophase. Primarily regressive parts of the cycle are exposed at the surface in Kentucky, and the Tyrone represents the culmination of this regression. Lithologies, sedimentary structures, and biota all suggest high-intertidal depositional environments with minor intervals of subtidal and supratidal deposition.

Although models of sequence stratigraphy are difficult to apply in cratonic situations, the Tyrone could best be interpreted as the culmination of a highstand systems tract, whereas the overlying Lexington probably represents the beginning of a new transgressive systems tract and a new sequence. The disconformity separating the two units represents the erosional base of the transgressive systems tract and may actually be a ravinement. Any lowstand systems tract is absent or represented by the disconformity.

The smaller fourth- or fifth-order cycles noted in the Tyrone are characteristic of the entire High Bridge Group, although they may be better developed in the underlying Oregon and Camp Nelson formations (see Horrell, 1981; Kuhnhenh and others, 1981). These generally shallowing-upward cycles are similar to the punctuated aggradational cycles (PAC's) of Goodwin and Anderson (1985), but the inability to correlate these cycles from one outcrop to another has led Horrell (1981) and Kuhnhenh and others (1981) to conclude that they represent the migration of subenvironments on a shallow

subtidal-peritidal shelf. It does not seem, however, that any of these workers considered possible loss of cycles due to stylolitization or penecontemporaneous erosion. Greater detail on the Tyrone and other High Bridge formations can be found in work by Cressman and Noger (1976), Whaley (1979), Horrell (1981), Kuhnhenh and others (1981), and Kuhnhenh and Haney (1986).

106.4 0.7 STOP 2B. Contacts between the Camp Nelson, Oregon, and Tyrone Formations: Stratigraphic and Environmental Implications (South- and northbound lanes of U.S. 27, approximately 5.1 mi south of its junction with Bypass 27, Jessamine Co.; Carter Coordinates: 2260'FNLx1540'FWL, 19-P-59).

CONTACTS BETWEEN THE CAMP NELSON, OREGON AND TYRONE FORMATIONS—STRATIGRAPHIC AND ENVIRONMENTAL IMPLICATIONS

FRANK R. ETTENSOHN

At this long roadcrop, at least 145 ft (44.2 m) of the upper Camp Nelson Formation, 34 ft (10.4 m) of the Oregon Formation, and the lower 18 ft (5.5 m) of the Tyrone Formation are exposed (Fig. 18). The Camp Nelson is composed largely of yellowish brown, pelletal calcilutite with abundant dolomite-filled burrow mottles. Preferential weathering of the burrow mottles gives rise to the characteristic "honeycomb" appearance in weathered sections. Laminated dolostones, white laminated calcilutites, massive calcilutites, ribbon-bedded calcilutite and dolostone, and shales are present locally. Open-marine faunas, including brachiopods, bryozoans, crinoids, and trilobites, are more common in the major shale partings like the one near the base of this roadcrop, but a sparse molluscan-tabulate coral (*Tetradium*) fauna is more common in the limestones, particularly in the breccias that occur at the bases of the shoaling-upward cycles throughout the section (Fig. 18). Stromatolites are locally present near the argillaceous unit at the top of the formation, but are absent at this locality.

The Camp Nelson is interpreted to represent a somewhat restricted, shallow, open-marine environment like the Bahamian platform lagoon. Such a platform lagoon typically exhibits heavily burrowed, pelletal muds which are probably the source for many of the muds transported onto adjacent tidal flats by tides and storms (Shinn and others, 1969). Camp Nelson environments may have similarly provided sediments for the Oregon and Tyrone tidal flats.

The overlying Oregon Formation is a rather arbitrarily defined unit based on the presence of mappable dolostones between the Camp Nelson and Tyrone. In some ways it is a "diagenetic unit" dependent upon the presence of the correct precursor lithofacies for later dolomitization. The unit consists largely of laminated dolostones with lesser amounts of laminated calcilutite and limestone breccia; the same molluscan-tabulate coral fauna is present but is largely confined to the breccia layers.

Horrell (1981) interpreted the Oregon to represent a mosaic of very shallow subtidal to intertidal environments in a transition interval between the dominantly subtidal Camp Nelson and the dominantly intertidal Tyrone. He recognized seven major recurring lithofacies in the Oregon, each of which was interpreted to represent a specific environment within a large carbonate tidal-flat complex: 1.) laminated dolostone—upper to middle intertidal zone, 2.) interlaminated

calclutite and dolosiltite—middle to lower intertidal pond-edge zone, 3.) laminated calclutite—lower intertidal pond-edge, 4.) pseudobreccia (calclutite pseudo-clasts in laminated dolosiltite)—lower intertidal zone, transitional between open-intertidal and open-subtidal zones, 5.) ribbon beds (very thinly interbedded calclutite and laminated dolosiltite)—lower-intertidal zone (immediately below pseudobreccia), transitional between open-intertidal and open-subtidal zones, 6.) calclutite with dolomitic burrows—open-subtidal zone, and 7.) birdseye calclutite—tidal-pond zone.

Horrell's interpretations of the lithofacies at this stop are shown in Figure 19. Markov-chain analysis has shown these lithofacies are commonly arranged into a series of shoaling-upward cycles (Horrell, 1981). Horrell (1981) also noted that dolomitization was largely confined to facies characterized by fine laminae (cryptogalaminites) and bioturbation suggesting that organic activity was probably a prerequisite for dolomitization. The distinctive color mottling along burrows, mudcracks, and laminae in these dolostones is the basis for the old name "Kentucky River Marble" used in some of the older literature in place of the name Oregon (Owen, 1857). A prominent shale break in the lower part of the unit may be an unnamed bentonite (Fig. 18).

Basal parts of the Tyrone Formation contain lithologies and sedimentary structures like those at Stop 2A, however, the greater presence of dolomitic burrow mottling in some of the limestones reflects increased subtidal influence. A thin shale parting near the top of this exposure is probably the Pencil Cave K-bentonite (Fig. 18).

107.1 0.7 Cross Kentucky River; Garrard County line.

107.2 0.1 STOP 2C. Kentucky River Fault System and the Camp Nelson Limestone (South- and northbound lanes of U.S. 27, 0.4 mi south of the north end of the Kentucky River bridge, Garrard Co.; Carter Coordinates: 1700'FNLx150'FEL, 23-P-59).

KENTUCKY RIVER FAULT SYSTEM AND THE CAMP NELSON LIMESTONE

FRANK R. ETTENSOHN

In this part of Stop 2, the field-trip route crosses the Kentucky River fault system. In this area the fault system is a northeast-southwest-trending graben system, and U.S. 27 crosses the northwest bounding fault of the graben system at Stop 2C (Fig. 20A). According to Wolcott (1969), at this fault the Curdsville and Grier members of Lexington Limestone have been downdropped nearly 300 ft (91.5 m) to the southeast against the Camp Nelson Limestone (Fig. 20B). At about 0.4 mi and 0.7 mi farther south, the highway crosses the two southeastern bounding faults of the graben (Fig. 20A). All of the displacement along these faults is to the northwest and is on the order of tens of feet, wholly within the Curdsville and Grier members of the Lexington Limestone. Although the sense of movement along all the major faults is now normal, both local and regional evidence suggests multiple episodes of movement, which include strike-slip, normal, and reverse displacements (Black and Haney, 1975).

The northwest boundary of the graben at this stop is actually a complex of four major normal faults and several minor ones (Fig. 20).

The large kink fold in the Camp Nelson (Fig. 20B) is probably the most outstanding feature in the exposures, and it is most likely a drag fold (Black and Haney, 1975; Gilreath and others, 1989). However, the drag on this fold is reverse to the present sense of displacement suggesting a later episode of reverse movement. Other related features in the area include slickensides, mineralized fractures, small reverse faults, fault breccias, fault-related jointing, and bedding-plane thrusts. Additional information on this series of exposures can be found in Wolcott (1969), Black and Haney (1975), Kuhnhenh and Haney (1986), and Gilreath and others (1989).

The drag folding and faulting at this stop bring up to road level parts of the Camp Nelson which are difficult to see elsewhere. Particularly important is a white, calclutite marker bed called the M bed (massive white micstone bed) by Gilreath and others (1989) and a major shale break which was also present at Stop 2B (Fig. 18). Although other white marker beds are present in the Camp Nelson and Oregon (Fig. 18), the lower white marker bed or M bed is the only one that is conspicuously peritidal because of the presence of mud cracks and vertical burrows. This peritidal marker bed is widespread in central Kentucky and was recently noted in an underground mine in the Maysville area. Moreover, it occurs at about the same stratigraphic horizon as does the peritidal Bowen Limestone in the Chickamauga Group of Tennessee (Fig. 7), although there is currently no biostratigraphic correlation between the two. The overlying shale break (Fig. 18) contains bryozoans and brachiopods, making it one of the most open-marine intervals in the Camp Nelson; it would seem to represent a brief episode of deepening and transgression across the restricted Camp Nelson shelf lagoon. It is tempting to equate this period of transgression with one represented by the Witten Formation in eastern Tennessee (Fig. 7). If this brief regressive-transgressive interval (white marker bed to shale break) in the upper Camp Nelson is equivalent to the Bowen and Witten formations on the margin of the Blountian foreland basin in Tennessee (Fig. 7), then these Camp Nelson beds may also record the passage of a peripheral bulge and anti-peripheral bulge respectively during the Blountian tectophase (Fig. 7). This regressive-transgressive interval would then correspond to the brief regressive-transgressive fluctuation that typically interrupts the overall transgressive-regressive flexural cycle (Fig. 9).

107.9 0.7 Crossing the southeast bounding fault on the Kentucky River fault system graben; drag fold in the Grier Member.

108.4 0.5 Junction with Kentucky State Route 1845.

108.6 0.2 Exposure of Grier Member cropping out on the left.

109.9 1.3 Cross back into the graben of the Kentucky River fault zone; Tanglewood Member of the Lexington in fault contact with the Clays Ferry Formation in the graben.

110.0 0.1 Junction with Kentucky State Route 152.

111.1 1.1 Indian Village souvenirs on right.

111.3 0.2 Bryantsville; junction with Kentucky State Route 753.

111.7 0.4 Leave the Kentucky River fault system graben; Upper Ordovician Ashlock Formation in the graben downdropped against the Middle Ordovician Grier Member of the Lexington Limestone.

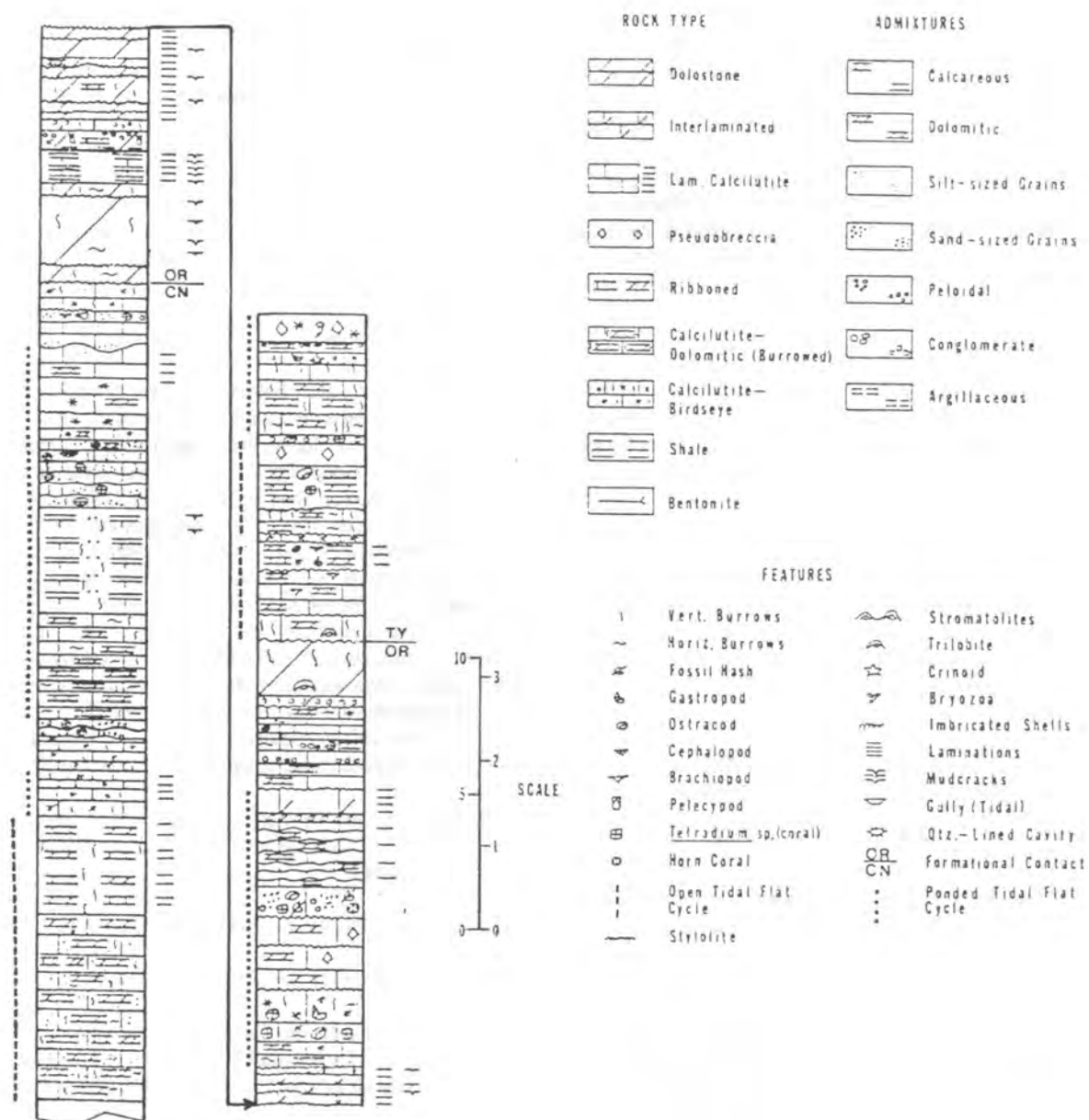


Figure 19. Detailed lithologies, sedimentary structures and paleontology of the Oregon and parts of the Camp Nelson and Tyrone formations at Stop 2B. Please refer back to text for the environmental interpretation of each lithology (from Horrell, 1981).

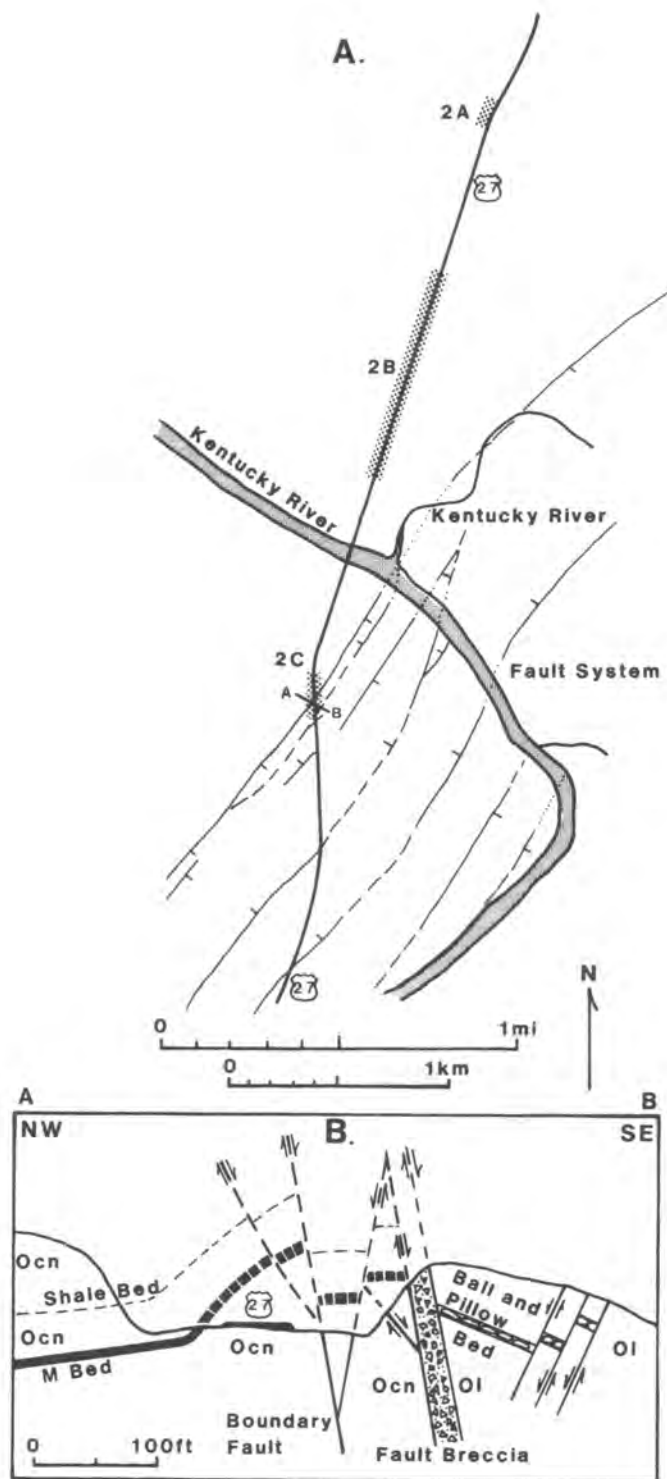


Figure 20. Diagrams showing (A) the location of the various parts of Stop 2 relative to the Kentucky River fault system near Camp Nelson (after Wolcott, 1969), and (B) a section across the northwestern bounding fault of the graben at Stop 2C (after Gilreath and others, 1989).

- 111.9 0.2 Junction with Kentucky State Route 1355; view to the left of Burdette Knob, a downfaulted outlier of the Lower Mississippian Borden Formation and the Upper Devonian New Albany Shale, present in the Kentucky fault system graben; at least 800 ft (244 m) of displacement are present here (see Wolcott and Cressman, 1971). The presence of these units here near the axial trace of the Cincinnati arch indicates that units at least as young as the Mississippian were deposited across the arch.
- 113.1 1.2 Turn right (southwest) onto Kentucky State Route 34 and proceed toward Danville; last view of Burdette Knob.
- 116.9 3.8 Cross Lake Herrington on the Chenault Bridge; Tyrone Fm. exposed on the banks of the lake.
- 118.2 1.3 Danville city limits.
- 120.4 2.2 Bear left on Kentucky State Route 34.
- 120.6 0.2 Turn right onto Main Street (KY. 34).
- 120.8 0.2 Turn left onto Second Street.
- 120.9 0.1 Turn left onto Walnut Street.
- 121.0 0.1 Turn left into parking lot at Constitution Square State Park. LUNCH STOP. Constitution Square is a shrine to Kentucky statehood, marking the location of the conventions during which Kentucky's first constitution was drafted. Replicas of the courthouse, the first Presbyterian church in Kentucky, and the jail, as well as the original building which housed the first post office west of the Alleghenies, are present in the park. 1992 is Kentucky's bicentennial anniversary.

GENERAL DEVONIAN PALEOGEOGRAPHIC AND TECTONIC FRAMEWORK FOR KENTUCKY

FRANK R. ETENSOHN

During the Devonian, Kentucky was situated between 20° and 30° south latitude (Fig. 21) and about 300 to 800 km west of the mountainous eastern coastline of Laurussia (Scotese, 1990). During the entirety of the Devonian, the eastern coast of Laurussia was involved in collisions with displaced Avalonian terranes (Fig. 21) resulting in the Acadian orogeny. These terranes collided obliquely with successive promontories on the coastline so that the orogeny proceeded from north to south in time (Rodgers, 1967; Ettensohn, 1987). Successive collisions resulted in four tectophases and the concurrent development of four respective unconformity-bound, flexural, stratigraphic sequences (Johnson, 1971; Ettensohn, 1985a, 1987; Ettensohn and others, 1988b) (Fig. 22). During the first two tectophases (Early-Middle Devonian), Kentucky was far enough removed from the loci of tectonism and foreland-basin sedimentation that it was essentially an area of bulge uplift which experienced erosion and shallow-marine, transgressive-regressive carbonate deposition. Because of the extent of bulge erosion at these times, the Lower Devonian Oriskany Sandstone and the Middle Devonian Onondaga Limestone, basal transgressive units corresponding to the

first and second tectophases respectively (Fig. 22), are preserved only in more proximal parts of the Appalachian basin in eastern Kentucky. The Boyle Dolostone which we will examine at Stops 3 and 4 on Day 1, is an overall regressive carbonate sequence deposited on the craton margin during later regressive parts of the second tectophase; erosion accompanying initiation of the third tectophase destroyed nearly all the Boyle except for that preserved in the Rome trough (Fig. 23).

Most of the Devonian rocks now preserved in central Kentucky are black shales or black-shale-related lithofacies representing the third tectophase. The third tectophase began in the late Middle Devonian (Givetian), and its beginning is represented in Kentucky by a probable bulge-generated unconformity and a series of late Middle Devonian breccia and black-shale graben fills (Duffin Bed, Portwood Member of the New Albany Shale; Figs. 24, 25), which will be examined at Stops 3 and 4 on Day 1. In addition to eroding most of the previously deposited Boyle, bulge uplift apparently reactivated parts of four complex fault systems near the Cincinnati arch in central Kentucky (Fig. 24). The fault-bound basins were apparently invaded by shallow seas into which eroded Boyle clasts from the basin margins were dumped as debris flows, forming the Duffin breccias (Fig. 25) (Ettensohn and others, 1991). After the small fault-bound basins filled, nearly all major sedimentation in central Kentucky ceased for about ten million years, even though moderately deep seas still covered all of Kentucky. This period of nondeposition is represented nearly everywhere in Kentucky by a thin, sandy pyritic condensation zone, commonly containing abundant conodonts and fish-bone fragments. Up to seven conodont zones may be represented in this condensation deposit (Ettensohn and others, 1989a) (Fig. 26) which we will see well exposed at Stop 4 on Day 2. This condensation zone or lag deposit effectively represents a period of sediment starvation in the cratonic seas.

The cratonic starvation was possible because during early parts of third tectophase an actively subsiding foreland basin intervened between the source areas and craton (Fig. 27). By the third tectophase, the locus of tectonism had shifted to the Virginia promontory, and with it came major foreland-basin subsidence cratonward of the promontory, which created a giant "clastic sink" (Ettensohn, 1985b, 1991b). The sediment infill of this basin, containing sequences of alternating basinal black shales and coarser clastic wedges (Figs. 22, 27), as well as the distribution of the black shales in time, suggest that third-tectophase tectonism was pulsatory. Each black shale-clastic wedge sequence deposited during this tectophase appears to be a small-scale, truncated, flexural sequence, and the progressive westward movement of tectonism is reflected in the progressive westward migration of black shale-clastic wedge sequences in time (Figs. 22, 27) (Ettensohn, 1985a, b, 1987; Ettensohn and others, 1988b). Despite continued westward migration, all of these sequences were restricted to the subsiding foreland basin, thereby ensuring cratonic starvation until the latest Devonian (early Famennian) (Fig. 27).

As the locus of subsidence moved westward, rates of sedimentation far exceeded subsidence rates in more proximal parts of the foreland basin so that distal parts of clastic wedges began prograding onto the craton (Fig. 27). By the early Famennian, the first of these clastic wedges, the upper Olentangy Shale and its equivalents had migrated beyond the foreland basin, onto the craton (Fig. 27) and into the Illinois Basin (Ettensohn and others, 1989a). This progradation onto the craton signals the end of cratonic starvation, and indicates that Acadian tectonism had migrated far enough westward to yoke the



Figure 21. Generalized Late Devonian paleogeography and lithofacies for North America (from Ettensohn, 1991b).

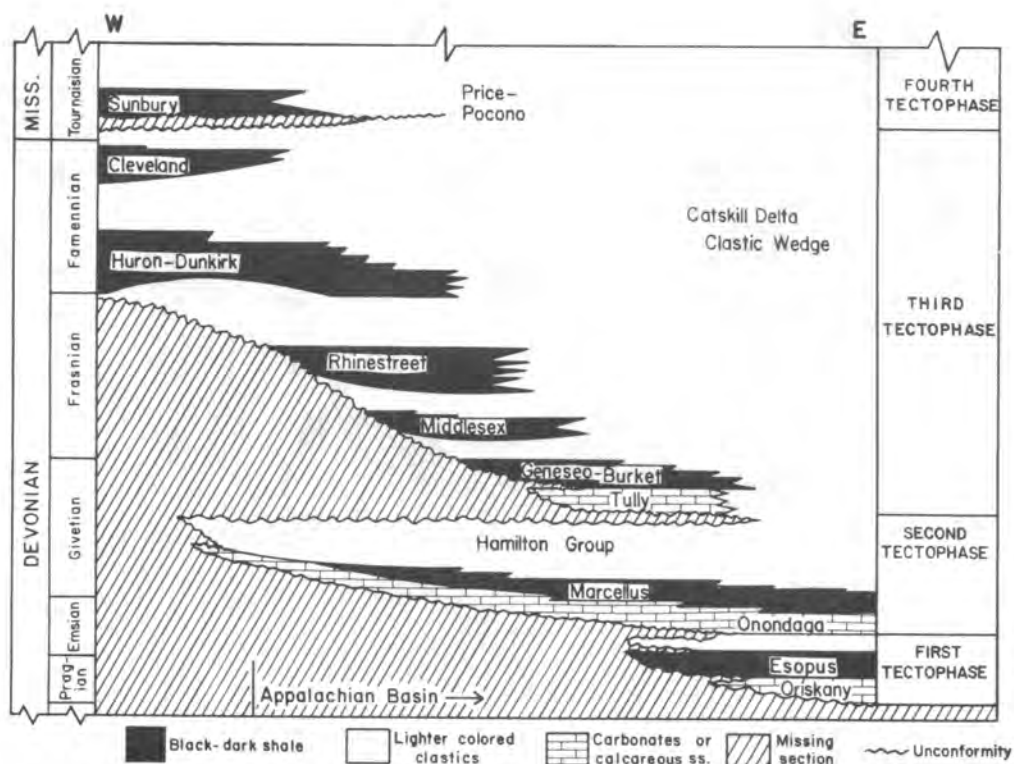


Figure 22. Composite stratigraphic column for northern and central parts of the Appalachian basin showing the distribution in time of unconformity-bounded flexural sequences of carbonates, black shales and coarser clastic sediments attributed to tectophases of the Acadian orogeny (from Ettensohn, 1987).

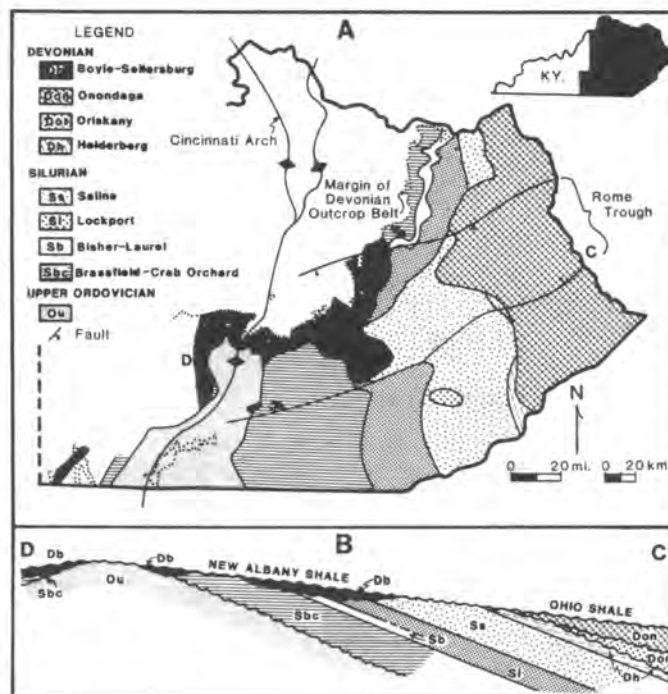


Figure 23. A) Paleogeologic map of the surface below the Devonian black-shale sequence in eastern and south-central Kentucky. B) Schematic northeast-southwest cross section through this surface in the Rome trough (not to scale) (from Ettensohn and others, 1988b).

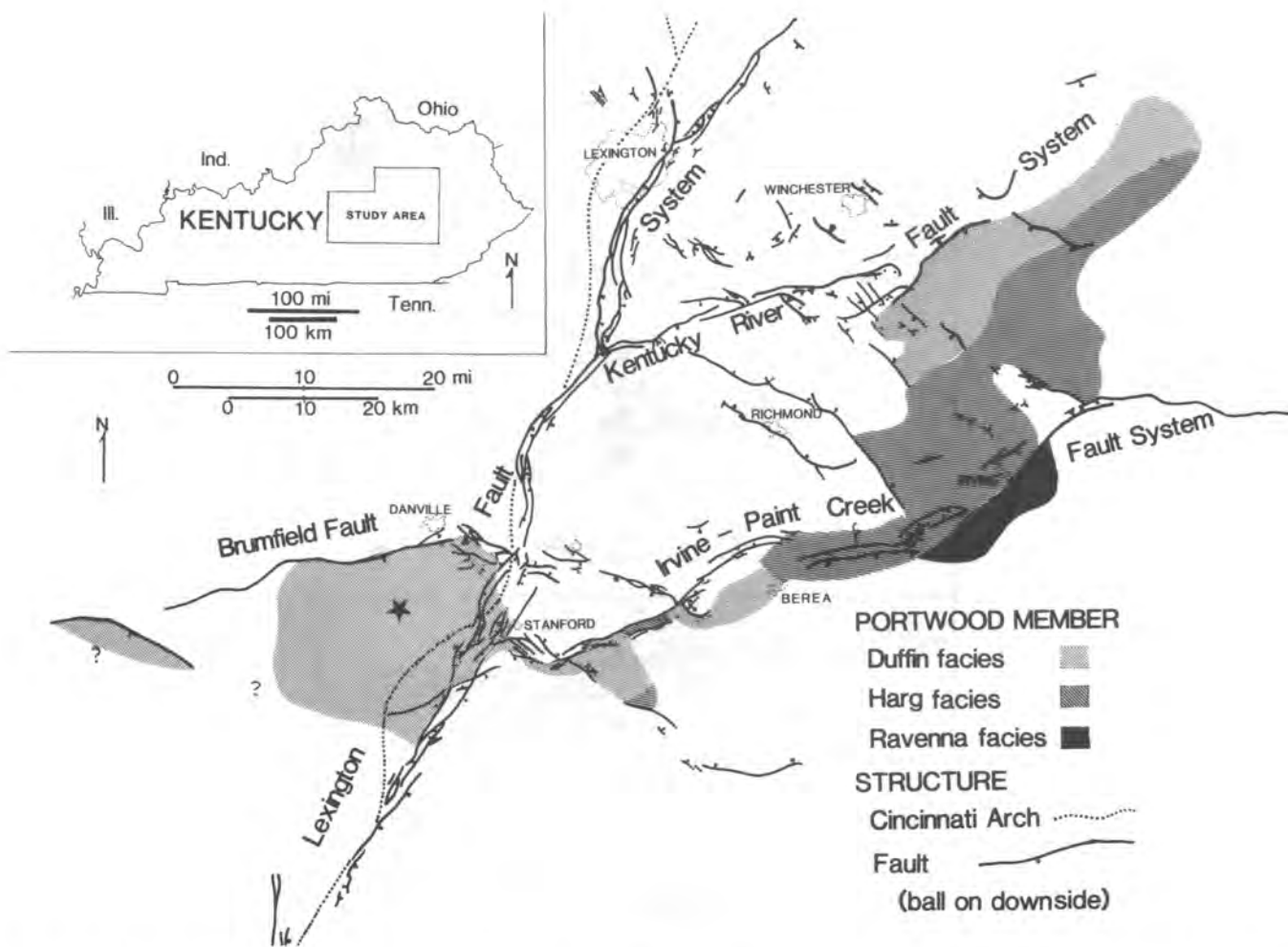


Figure 24. Distribution of predominant Portwood facies relative to major faults in central and east-central Kentucky (from Ettensohn and others, 1991). The star marks the approximate location of Stops 3 and 4 on Day 2.

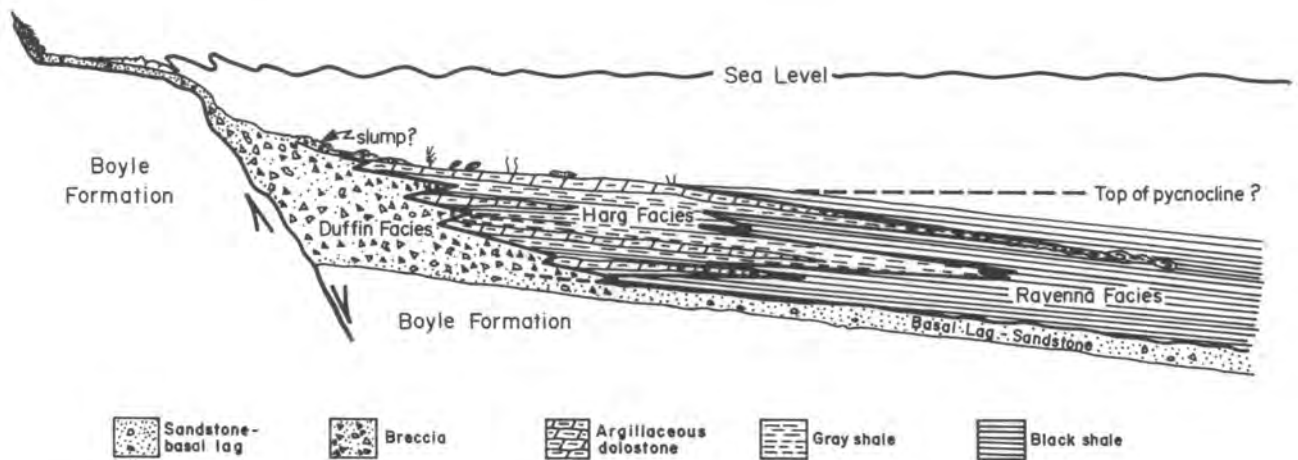


Figure 25. Schematic diagram showing interpreted proximity trends of various Portwood facies relative to the margin of a small fault basin (from Ettensohn and others, 1988b).

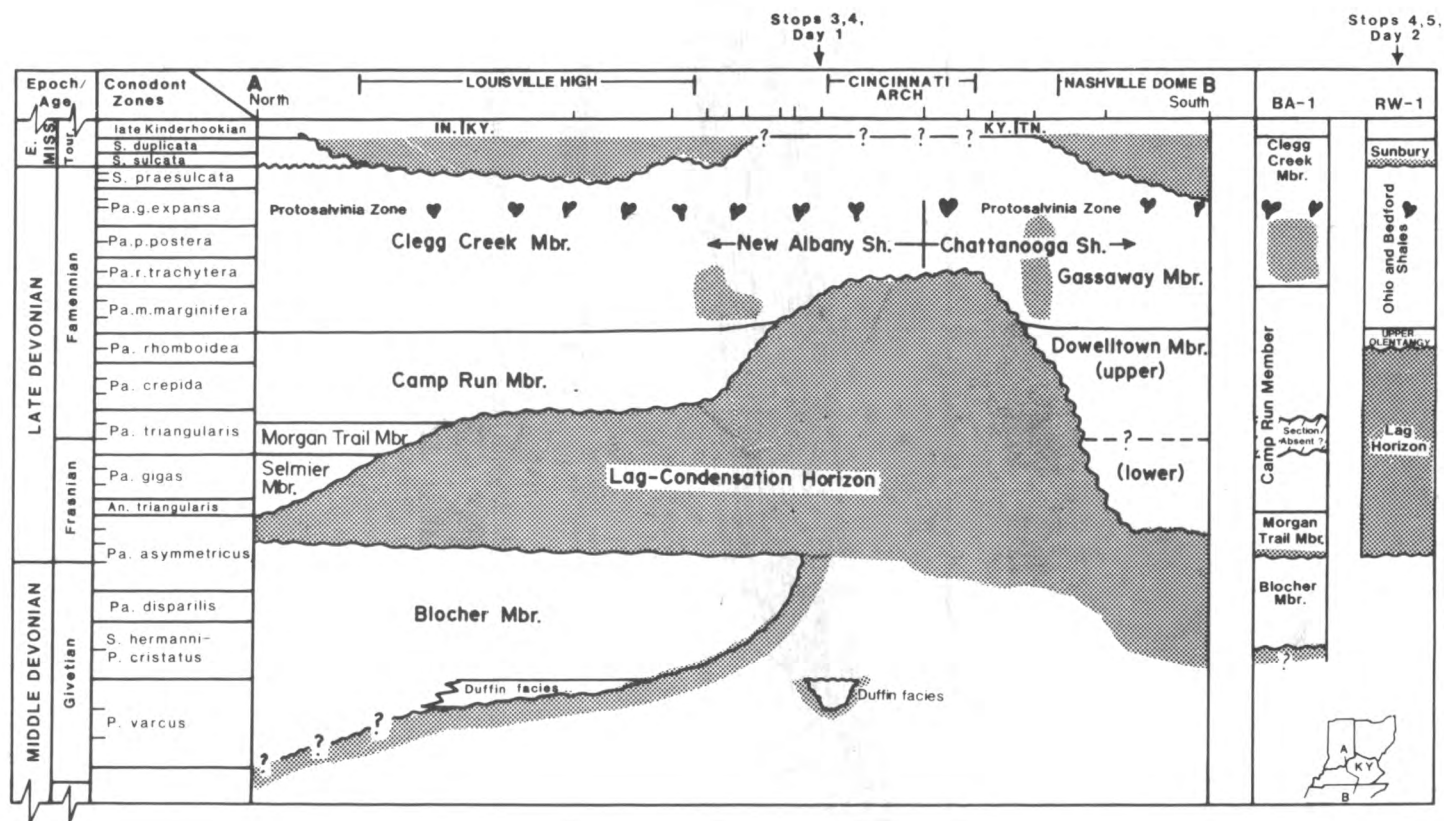


Figure 26. A nearly north-south section from southern Indiana to central Tennessee through central Kentucky showing the position of conodont zones relative to the black-shale section with two comparative columns from west-central (BA-1) and northeastern (RW-1) Kentucky. Parts of at least seven conodont zones are represented in the condensation zone between Duffin (or Blocher) and overlying Upper Devonian parts of the New Albany Shale. At least six of these zones are present in the thin lag horizon below the Ohio Shale at Stop 4 on Day 2 (RW-1). The diagram also suggests a time of complete yoking between the Appalachian and Illinois basins in the Late Devonian (after Ettensohn and others, 1989a).

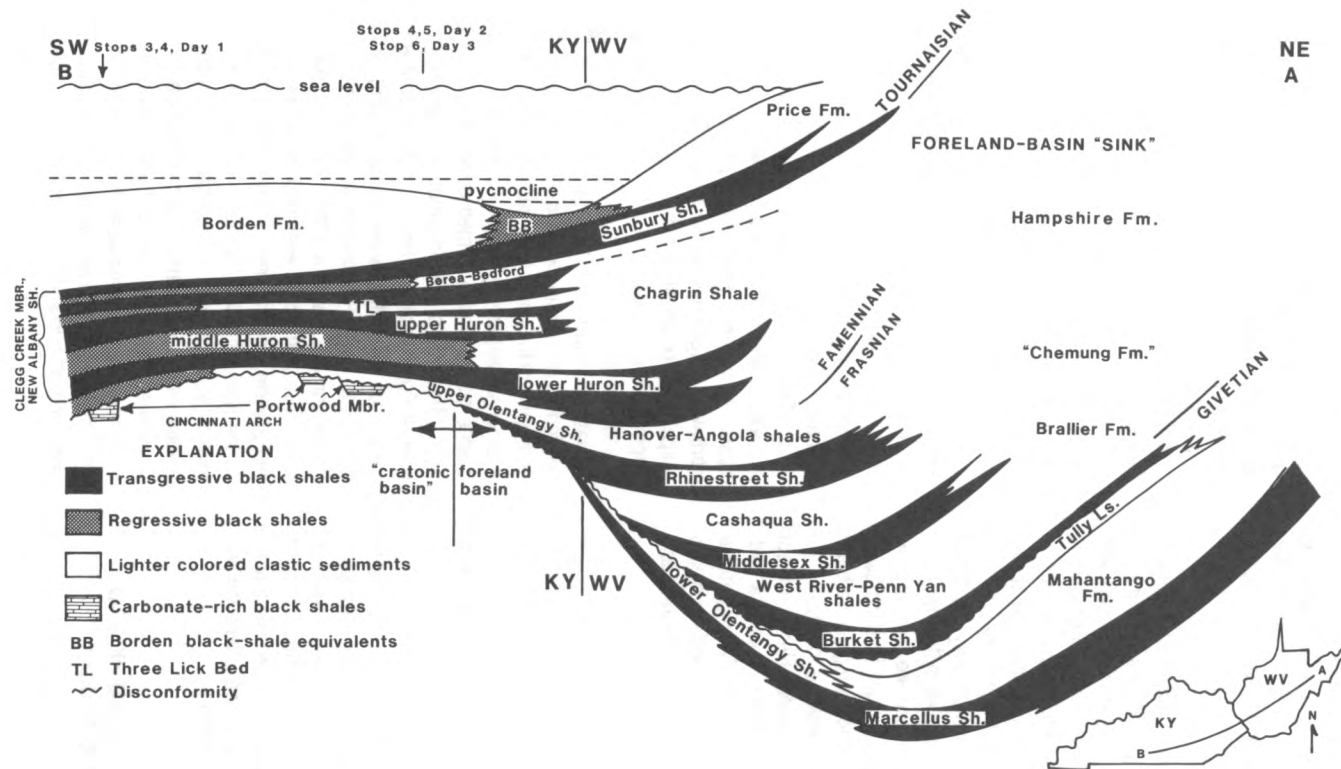


Figure 27. Schematic diagram showing the disposition of black-shale, foreland basins and intervening coarser clastic wedges from the Middle Devonian to the Early Mississippian along a line of section from eastern West Virginia to central Kentucky. Until early Famennian upper Olenitangy and lower Huron deposition, the deposition of black shales and coarser clastic wedges was largely restricted to a subsiding foreland-basin "sink." In the early Famennian, yoking of the Appalachian and Illinois basins (see Fig. 26) allowed deeper basinal waters to migrate onto the craton forming a "cratonic black-shale basin," in which the fine, westwardly prograding parts of the clastic wedges were preserved as regressive black shales. Parts of three Acadian tectophases are represented (see Fig. 22), and the relative positions of stops with black shales are shown at the top of the diagram (after Ettensohn and others, 1988b).

Illinois and Appalachian basins (Fig. 26). It is from this time onward, except for the deposition of the Portwood and the condensation zone, that the major history of Kentucky black-shale deposition begins.

The cyclic deposition of black shales and clastic wedges that had characterized the foreland basin during the early Late Devonian (Frasnian) (Figs. 22, 27) continued on the craton during later parts of the Late Devonian (Famennian). Each cycle consists of a basinal black shale with a sharp basal contact representing transgression and subsidence accompanying deformational loading in the orogen. The black shale is overlain gradationally by a gray shale reflecting regression concomitant with loading-type relaxation. Succeeding pulses of tectonism followed so rapidly that complete flexural sequences never developed.

In central Kentucky, this cyclicity is still apparent, but is not so visually apparent because many of the intervening gray shales appear to be absent. In fact, they are not. Comparison of gamma-ray logs with cores and sample sets indicates that these gray shales grade into black shales as the muds were transported into deeper anaerobic parts of the cratonic basins. Nevertheless, because these shales contain more clastic debris and are less radioactive, they can be distinguishable on gamma-ray logs and in outcrops; they are called regressive black shales. The subjacent black shales in the lower part of the cycle, represent deeper water deposition and are more organic rich; they represent deposition during times of major foreland-basin subsidence with little clastic influx and are called transgressive black shales and are also easily distinguished because of their high radioactivity and their organic-rich, paper-like nature. Hence, except for the local presence of the upper Olentangy Shale Three Lick Bed, and Bedford-Berea Sequence, cratonic transgressive-regressive flexural cyclicity is reflected in alternating transgressive and regressive black shales (Fig. 27) (Ettensohn and Elam, 1985; Ettensohn and others, 1988b).

The abundance of organic matter in these shales, up to 23 weight percent with a mean of about seven percent, is apparently related to interactions between paleoclimatic, paleogeographic, and tectonic factors conducive to the production of abundant organic matter and its subsequent preservation from the effects of oxidation and clastic dilution. Some of these factors reflect peculiarities of the Devonian as a time of enhanced warmth, increased tectonism, and greatly expanded transgressing seas. Other factors include the nearly enclosed nature of the Black-Shale Sea, its position in the subtropical trade-wind belt in the rainshadow of the rising Acadian mountains (Fig. 21), and the fact that large portions of the Black-Shale Sea coincided with rapidly subsiding Acadian foreland basins. For more information about these factors and the way they interacted, see Ettensohn and Barron (1981, 1983) and Ettensohn (1985b, 1991b).

GENERAL MISSISSIPPIAN PALEOGEOGRAPHIC AND TECTONIC FRAMEWORK FOR KENTUCKY

FRANK R. ETTENSOHN

By the Early Mississippian, Gondwana had begun its clockwise rotation into a position preparatory for Ouachita convergence, which may have begun as early as the latest Devonian or Early Mississippian (Ross and Ross, 1985; Morton, 1985). As Gondwana moved northwestward, Laurussia was apparently pushed farther northward as well, so that by the Middle Mississippian Kentucky was located at about 15° south latitude. During the remainder of the

Mississippian, Kentucky continued to move northward into more tropical latitudes (Scotese, 1990; Ettensohn, 1990b).

Overall, the Mississippian was a period of regression in the transition from Devonian submergence to Pennsylvanian emergence. However, the Mississippian in Kentucky begins, just as was the case in much of the Devonian, as a deeper water, cratonic, black-shale basin represented by the Sunbury Shale, which will be examined on Days 2 and 3 of the trip. The Sunbury probably reflects the deepest of the black-shale seas (Ettensohn, 1984, 1990a) and represents the beginning of the fourth and final tectophase of the Acadian orogeny (Fig. 22) (Ettensohn, 1985a; Ettensohn and others, 1988b). Even so, in the subsequent eight to nine million years, this basinal area was transformed into a very shallow, epeiric platform conducive to the deposition of shallow-water oolitic carbonates. These relatively rapid changes probably reflect basin infill and regional uplift related to flexural mechanisms illustrated in Figure 28.

In general, interpretations shown in Figure 28 suggest that the Mississippian in eastern Kentucky largely represents a period of tectonic relaxation after the final tectophase (unconformity and Sunbury Shale, Fig. 28A) of the Acadian orogeny and reflects both loading and unloading-type relaxation (Ettensohn, in press a; Ettensohn and Chesnut, 1989). Lower and Middle Mississippian (Tournaisian and Visean) clastics represented by the Borden, Fort Payne and their equivalents, as well as parts of the overlying carbonates represented by the Greenbrier, Newman and Slade formations, apparently reflect a period of loading-type relaxation (Fig. 28B). Clastic infill as well as bulge uplift and migration apparently elevated former basinal environments almost to the level of wave base. In western Kentucky where clastic basin infill was minor, northward bulge uplift accompanying Ouachita tectonism was mostly responsible for elevation of basin floors (Ettensohn, in press a). The uppermost high-energy carbonates of the Greenbrier, Newman and Slade formations (late Visean) apparently represent a time of elevational equilibrium between the filled foreland basin and the leveled Acadian highlands (Fig. 28C). In contrast, the overlying clastic sediments of the upper Newman, Bluefield and Pennington formations (Namurian) apparently represent a westwardly prograding marginal-marine clastic wedge formed during unloading-type relaxation (Fig. 28D). This type of marginal-marine sedimentation continued into the earliest Pennsylvanian (Pocahontas) when it was abruptly ended by advent of the Alleghanian orogeny (Fig. 28E) reflected in the prominent Early Pennsylvanian (mid-Carboniferous) unconformity (Ettensohn and Chesnut, 1989). These upper Mississippian clastics have been commonly assumed to represent the beginning of the Alleghanian orogeny (e.g., Perry, 1978; Chesnut, 1991a). However, all the typical flexural indicators for the inception of an orogeny are absent: there is no bounding basal unconformity, no transgressive marine, flexural sequence, and no indication of two basinal paleoslope directions (Fig. 28). In fact the absence of an unconformity, predominance of regressive marginal-marine facies including redbeds, and the dominant cratonward-dipping paleoslope reflected in the sediments all suggest unloading-type relaxation (Fig. 28D).

As Acadian bulges migrated eastward through Kentucky during relaxation, and Ouachita bulges migrated northward during active tectonism, many basement structures seem to have been reactivated. In eastern Kentucky, this type of reactivation was apparently responsible for many of the stratigraphic anomalies associated with the Mississippian System. These anomalies include abrupt changes in facies and thickness, disjunct unit distributions, as well as local

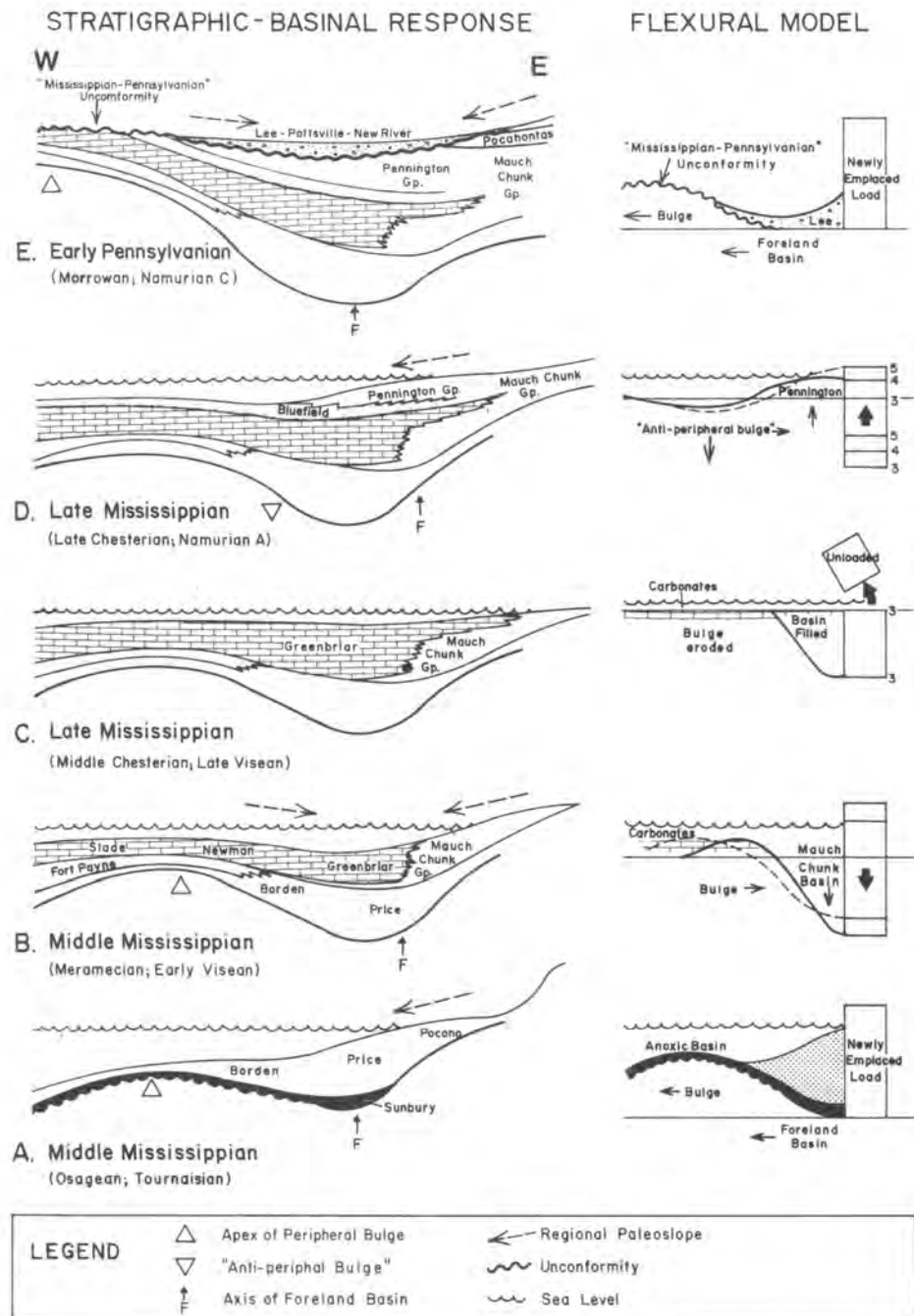


Figure 28. Probable sequence of flexural events and accompanying sedimentary responses along an east-west section across the central Appalachian basin between the last tectophase of the Acadian orogeny (A) and the inception of the Alleghanian orogeny. In this scenario, most of the Mississippian section reflects a period of Acadian flexural relaxation (from Ettensohn and Chesnut, 1989).

intrasystematic unconformities near structures, and we will examine several of these anomalies on Days 2 and 3.

PORTWOOD MEMBER OF THE NEW ALBANY SHALE AND THE BOYLE DOLOSTONE

STEPHEN F. BARNETT AND FRANK R. ETTENSOHN

The Middle Devonian Portwood Member is the oldest unit of the predominantly Upper Devonian and Early Mississippian New Albany Shale. It is a complex and enigmatic unit whose origins and stratigraphic relationships have been the subject of controversy for some years. The Portwood crops out in an arcuate belt corresponding roughly to The Knobs physiographic region of central Kentucky (Figs. 2, 24) and commonly unconformably overlies the Middle Devonian Boyle Dolostone (Fig. 29). The dominant lithology of the Portwood, in contrast to the black shale that characterizes most of the New Albany, is an organic-rich dolomitic breccia which has incorporated clasts of chert, dolostone and silicified fossils derived largely from the underlying Boyle Dolostone; black-shale clasts from a previously unknown black-shale unit also occur locally. The thickness, up to 33 ft (10 m), the areal extent of the breccia in central Kentucky, and its association with central Kentucky fault systems (Fig. 24) require a regional approach to understanding its deposition. We have suggested that tectonic activity during the Acadian orogeny reactivated bulge-related faults after Boyle deposition and helped to induce brecciation of the Boyle on uplifted fault blocks; debris flows subsequently transported these breccias into downdropped fault basins (Ettensohn and others, 1988b, 1989b, 1991). Evidence for such fault-related depositional mechanisms is found in the Carpenter Fork graben at Stop 4 in southern Boyle County, and this evidence illustrates in microcosm the mechanisms we believe to have been primarily responsible for the emplacement of the entire unit.

REGIONAL STRUCTURE

The dominant structure in central Kentucky is the Jessamine dome, a structural high on the axis of the Cincinnati arch (Fig. 24). The entire central Kentucky region has also been subjected to recurrent faulting on the Lexington, Irvine-Paint Creek, Brumfield and other unnamed fault systems (Fig. 24). These fault systems were active throughout the Paleozoic, and may have repeatedly reversed their sense of movement, as is evident in some faults within the immediate area of this stop (Moore, 1978).

REGIONAL STRATIGRAPHY

The Devonian stratigraphy of central Kentucky is represented by Middle and Upper Devonian strata which unconformably or paraconformably onlap an erosional surface developed on Ordovician and Silurian rocks. The Devonian sedimentary sequence generally reflects a period of net transgression, but there are also indications of multiple erosive or hiatal events marked by sandy or phosphate- and bone-rich lag horizons (Conkin and others 1976; Lenhart, 1985). The typical stratigraphic column for central Kentucky is shown in Figure 29A.

The lowermost Devonian unit, the Middle Givetian (Taghanic; Hamilton equivalent) Boyle Dolostone has been interpreted by recent authors to be a shallow-subtidal deposit (e.g., Lenhart, 1985). The Boyle is primarily a fine- to medium-grained dolarenite with variable amounts of chert, sand and silt. It is largely an unconformity-bound unit and probably represents cratonic deposition during the second tectophase of the Acadian orogeny. The Boyle, in turn, is unconformably overlain by either the Upper Devonian (late Frasnian to middle Famennian) upper Olentangy Shale or Huron Shale Member of the New Albany

- 121.2 0.2 Retrace route to the junction of Second and Main streets; turn left onto Main Street (Kentucky State Route 34) and proceed westward.
- 121.6 0.4 Centre College.
- 122.0 0.4 Cross bridge over Southern Parkway.
- 122.1 0.1 After the bridge, bear left onto Kentucky State Route 34.
- 122.8 0.7 "Knobs" (Highland Rim Section; New Albany and Borden formations) on the horizon ahead (Fig. 3).
- 123.3 0.5 Cross Bypass 127.
- 124.3 1.0 Clock tower on barn to the right; approximate boundary between the Inner and "Outer" Bluegrass. The "Knobs" apparent in the distance on the left; alignment of the Knobs reflects the nearly east-west Brumfield fault which occurs just in front of them (see Harris, 1972).
- 129.3 5.0 Parksville Post Office. The trace of the Brumfield Fault runs nearly east-west and is located about 0.2 mi north of the post office (Moore, 1978). Enter the "Knobs" area of the Highland Rim Section of the Interior Low Plateaus Province (Fig. 3). The "Outer" Bluegrass (Eden Belt) is so narrow in this area south of Danville because much of the Upper Ordovician is absent along the unconformity below the New Albany Shale, and other parts of it have been faulted out along the Brumfield Fault (see Moore, 1978).
- 129.4 0.1 Turn left onto Kentucky State Route 1822; New Albany Shale.
- 129.6 0.2 Middle Mississippian Borden Formations on the right.
- 130.5 0.9 Dolostones in the Muldraugh Member of the Borden on the right.
- 130.7 0.2 Siltstones in the Halls Gap Member of the Borden on the right.
- 132.2 1.5 Turn right onto Kentucky State Route 37; proceed southwestwardly.
- 132.9 0.7 Carpenter Creek Road on the left; New Albany Shale.
- 133.2 0.3 Dolostones in the Duffin Bed of the New Albany Shale on the right.
- 133.7 0.5 Turn left onto Tar Lick Road.
- 133.8 0.1 STOP 3. Portwood Member of the New Albany Shale and the Boyle Dolostone, North Rolling Fork River (North-facing embankment of the North Rolling Fork River, just west of Tar Lick Road and across from Kentucky State Highway 37, 1.5 mi west of its intersection with Kentucky State Highway 1822, Boyle Co.; Carter Coordinates: 750'FWLx900'FWL, 15-M-56).

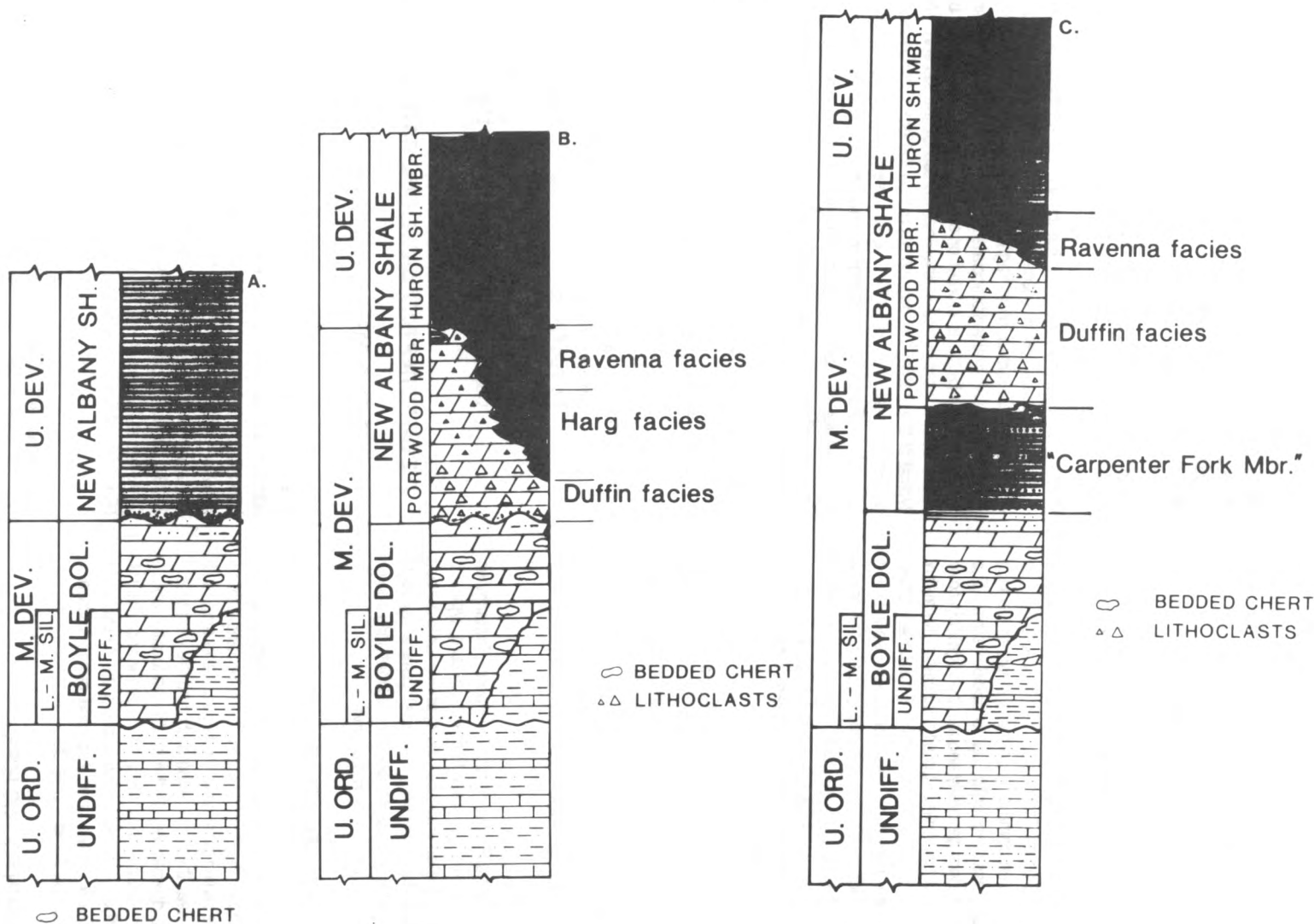


Figure 29. Three types of stratigraphic relationships between the Boyle Dolostone and overlying units. A) Typical situation with upper Devonian parts of the New Albany Shale (Huron Mbr.) unconformably overlying the Boyle. B) Middle Devonian Portwood Member lithofacies unconformably overlying the Boyle and, in turn, unconformably overlain by Upper Devonian parts of the New Albany (Huron Mbr.). C) Situation in Carpenter Fork graben where the intervening Carpenter Fork shale is gradational between the underlying Boyle and the overlying Portwood Member (Duffin Lithofacies) of the New Albany.

Shale (Fig. 29A) or by one of the facies of the Middle Devonian (Givetian) Portwood Member of the New Albany Shale (Portwood Formation of Campbell, 1946) (Fig. 29B). The Portwood Member, as reinterpreted by Ettensohn and others (1988b, 1989a), is of late Givetian age (Genesee Group equivalent) and consists of three coeval facies (Figs. 24, 25, 29). Locally, an even older unit of the New Albany, the informally designated Carpenter Fork shale member (Barnett and others, 1989a, 1989b; Ettensohn and others, 1991) of late Tully age, overlies the Boyle (Fig. 29C).

The Duffin Bed of the New Albany Shale is predominantly an organic-rich dolomitic breccia with clasts of dolostone, chert, as well as rare siltstone and dark-shale clasts. The lithology and fossil content of the dolostone, chert, and siltstone clasts indicate that they were largely derived from the underlying Boyle, but the origin of the dark-shale clasts has been problematic because previously no black shales were known to occur below the Duffin. Clast size is typically less than an inch, but the lowermost Duffin commonly contains clasts in the inch to three-inch range, and less frequently, up to boulder size. The lithoclasts are matrix supported, generally angular, and lighter in color than the supporting organic-rich matrix. Locally, however, the matrix is less organic-rich, and the clast-matrix distinction is obscured: Chert clasts are generally angular, may show weathering rinds, and tend to be less abundant than dolostone clasts, although locally, the lower Duffin consists almost entirely of densely packed chert clasts. Shale and siltstone clasts are angular to rounded and are generally sub-horizontally oriented. Some of the flatter clasts clearly show signs of imbrication and preferential orientation.

Two other facies of the Portwood Member, the Harg and Ravenna, consist of variable amounts of grey to black shale, dolomitic mudstone, and argillaceous to arenaceous limestones and dolostones. Typically, these facies conformably overlie the Duffin lithology in Boyle County and nearby areas. They, in turn, are overlain paraconformably by the late Frasnian-early Famennian upper Olenangy Shale or by the Famennian Huron Shale Member of the New Albany Shale (Fig. 29B, C).

The unconformity between the Portwood and the upper Olenangy or Huron is marked by a lag zone, generally thin and obscure, typically consisting of rounded and frosted quartzose sand, phosphorite grains, pyrite, and abundant conodont elements. Artificial gamma-ray logs (Ettensohn and others, 1979) of these sections reveal that units above the unconformity are significantly more radioactive.

LOCAL STRUCTURE

Stops 3 and 4 are located in south-central Boyle County (Fig. 24) about 10 mi (16 km) west of the roughly north-south-trending axis of the Cincinnati arch. The region was influenced by the Lexington fault zone, a series of faults near the crest of the arch, and the east-west-trending Brumfield fault, downthrown to the south and located about 3.6 mi (6 km) north of the stops. In addition to the broader structural influences discussed above, complex but small-scale faults, with offsets on the order of a few meters or less, and gentle folds are common in the area of Stops 3 and 4 (Fig. 30).

NORTH ROLLING FORK

This exposure represents a rather typical section of the Portwood in which Duffin breccia overlies the Boyle Dolostone with subtle unconformity (Fig. 31). The Boyle here is a light gray, coarsely crystalline, cherty, fossiliferous, dolomitic calcarenite. Fossils are

indistinct in fresh surface samples, but weathered exposures of the Boyle reveal abundant crinoid columnals and corals.

The Duffin breccia is 12.8-ft (3.9-m) thick and contains a matrix of yellowish olive-grey to brownish-grey, dolomitic, coarsely crystalline to fine-grained, upward-fining calcarenite. The clasts, which similarly decrease in size upward, are largely rounded Boyle lithoclasts, up to 0.5 ft (0.15 m) in long dimension, which are distinguished from the more organic-rich matrix chiefly by their lighter color. Other clasts include rounded and apparently weathered chert fragments up to one-foot (0.3-m) across, and in the basal one foot (0.3 m), elongate, sub-horizontal chert bodies up to 3.3 ft (1.0 m) in long dimension are present. Some of the chert clasts are surrounded by envelopes of organic-rich, dolomitic siltstone. Also included throughout the unit are sparse clasts of silty, brownish-black shale. The upper five feet (1.5 m) of the section are more homogenous than the lower six feet (1.8 m), contain lithoclasts less than one-inch (2.5-cm) across, and exhibit only rare, small, chert fragments and fossils. Locally, the base of the upper, finer grained breccia is silty, brownish-black, fissile shale. The top of the Duffin is eroded or covered by vegetation, but black fissile shales were found in the float on top of the section. As is typical of the Duffin, this deposit occurs on the downdropped side of a mapped fault (Moore, 1978); it is absent on the upthrown side (Fig. 30). Parts of the exposure nearest to the vehicles show crossbed-like structures. Could these mark accretionary surfaces in the debris flow, or are they the products of weathering?

- 133.9 0.1 Return to the junction of Tar Lick Road with Kentucky State Route 37; turn right onto State Route 37.
- 134.7 0.8 Turn right from State Route 37 onto Carpenter Creek Road and proceed southward.
- 135.6 0.9 STOP 4. The Carpenter Fork Graben and its Implications (South-facing embankment of Carpenter Fork on the eastside of Carpenter Creek Road, south of Kentucky State Highway 37 and about 0.6 mi (1.0 km) north of the Boyle-Casey county line, Boyle Co.; Carter Coordinates: 2700°N 100°W, 15-M-56).

THE CARPENTER FORK GRABEN AND ITS IMPLICATIONS

STEPHEN F. BARNETT AND FRANK R. ETTENSOHN

This exposure is unique in that the Duffin breccia is actually preserved in the Middle Devonian graben in which it originally accumulated. In addition, a black shale preserved in this graben below the Duffin has been shown to be the oldest known Devonian black shale in Kentucky (Ettensohn and others, 1991). This black shale, informally called the Carpenter Fork member of the New Albany Shale (Barnett and others, 1989a, 1989b), occupies a position between the Duffin and underlying Boyle typically represented by an unconformity elsewhere in the state (Fig. 29C). Middle to upper Devonian black shales above the Duffin occupy their usual stratigraphic positions (Figs. 29, 32).

The Carpenter Fork graben (Fig. 33) is part of a complex of faults (Fig. 30) which have produced a displacement of 43 ft (13 m) or more between the Boyle in the graben and the Boyle on the hillside east of the graben. Within the graben, disturbed beds of typical Boyle lithology are found at the base of the section, close to the bounding faults, but the beds dip below creek level in the interior of the graben.

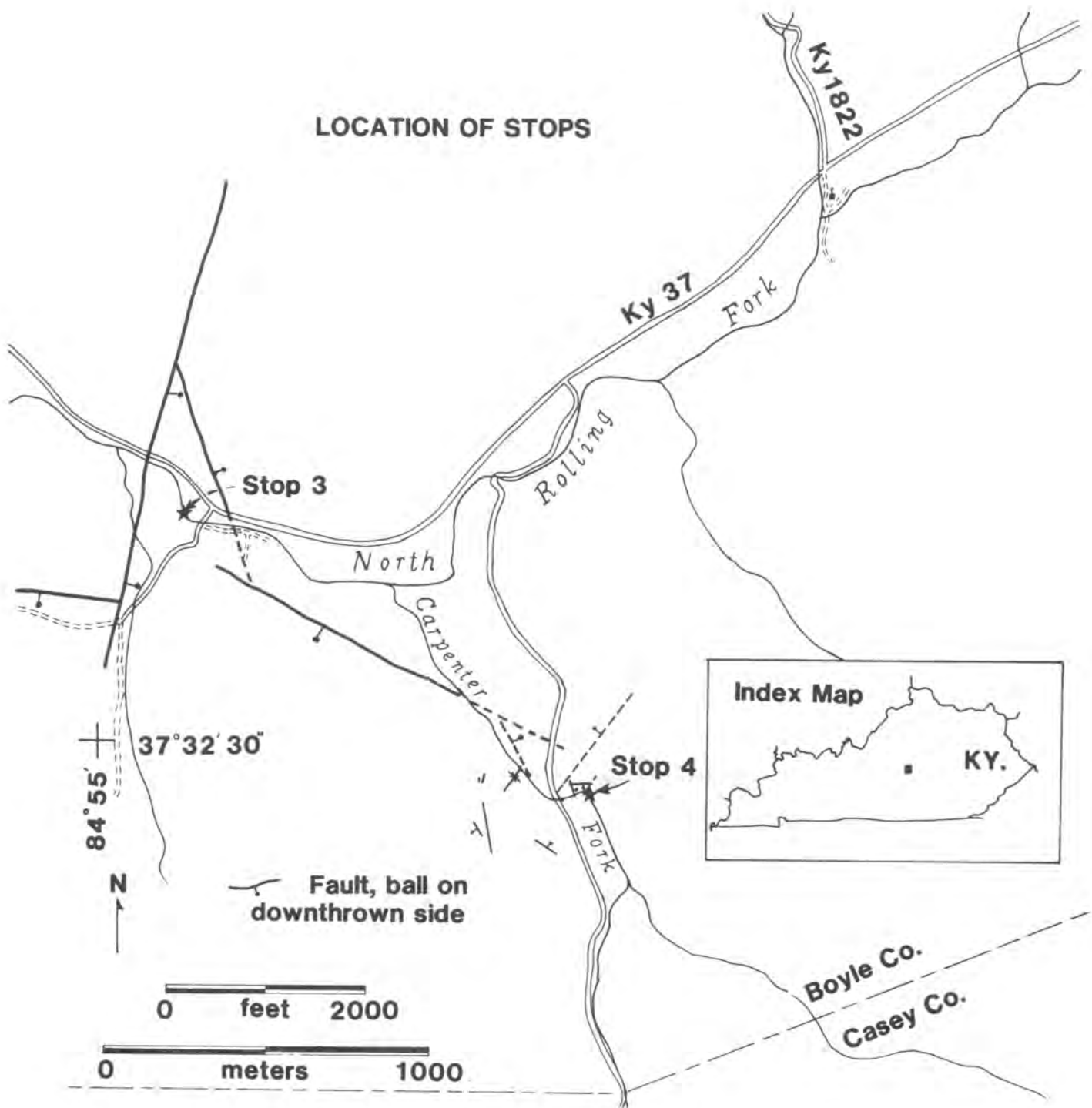


Figure 30. Map showing the location of Stops 3 and 4 on Day 1 relative to local structures.

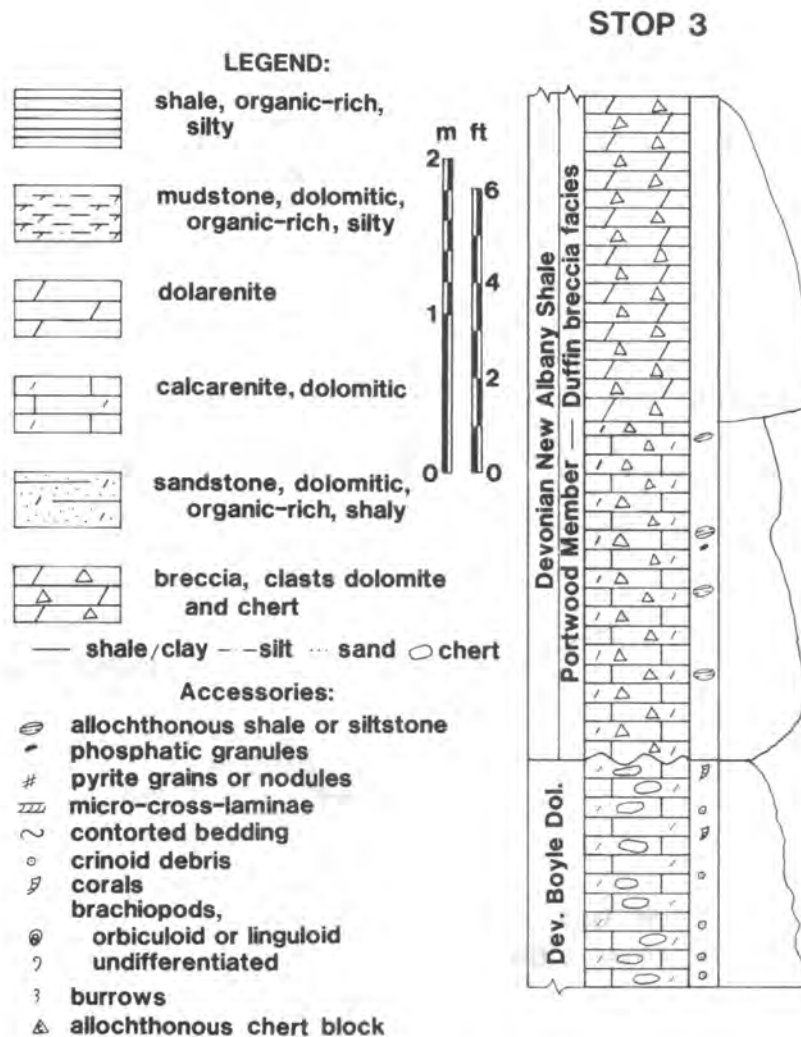


Figure 31. Stratigraphic section and profile at Stop 3 on Day 2.

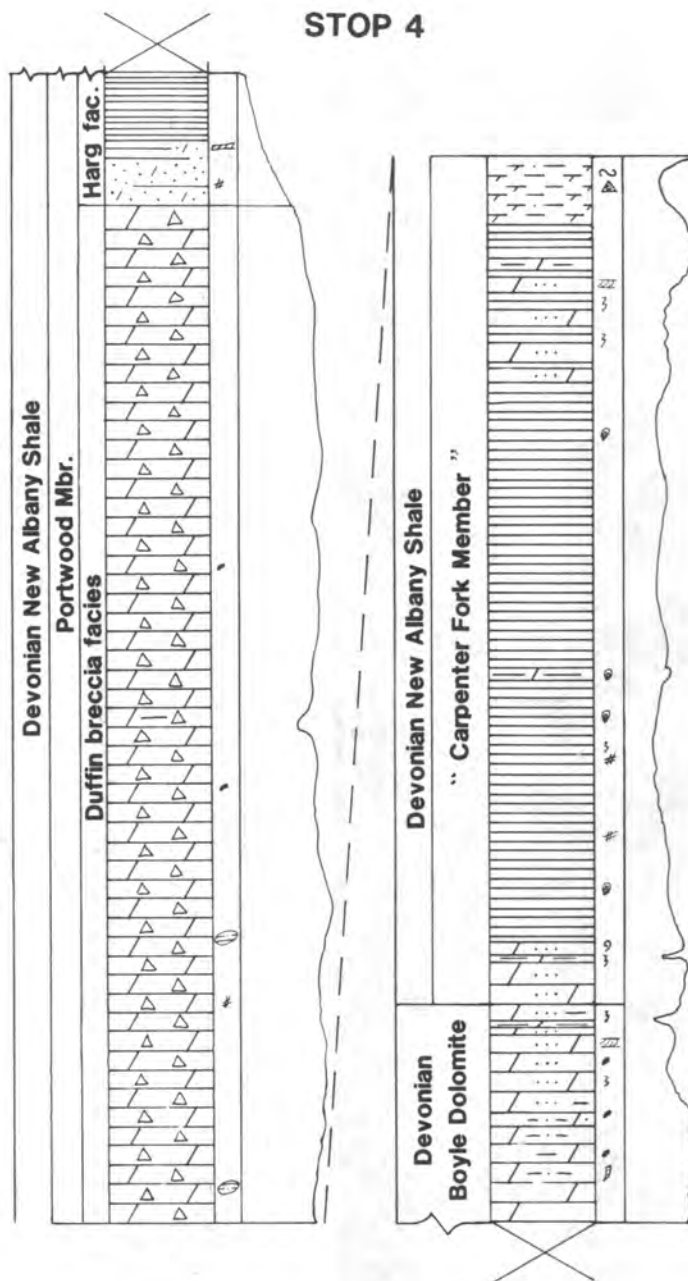


Figure 32. Stratigraphic section and profile from Stop 4 at the Carpenter Fork graben. See Figure 31 for the legend and Figure 33 for a photograph of the graben.

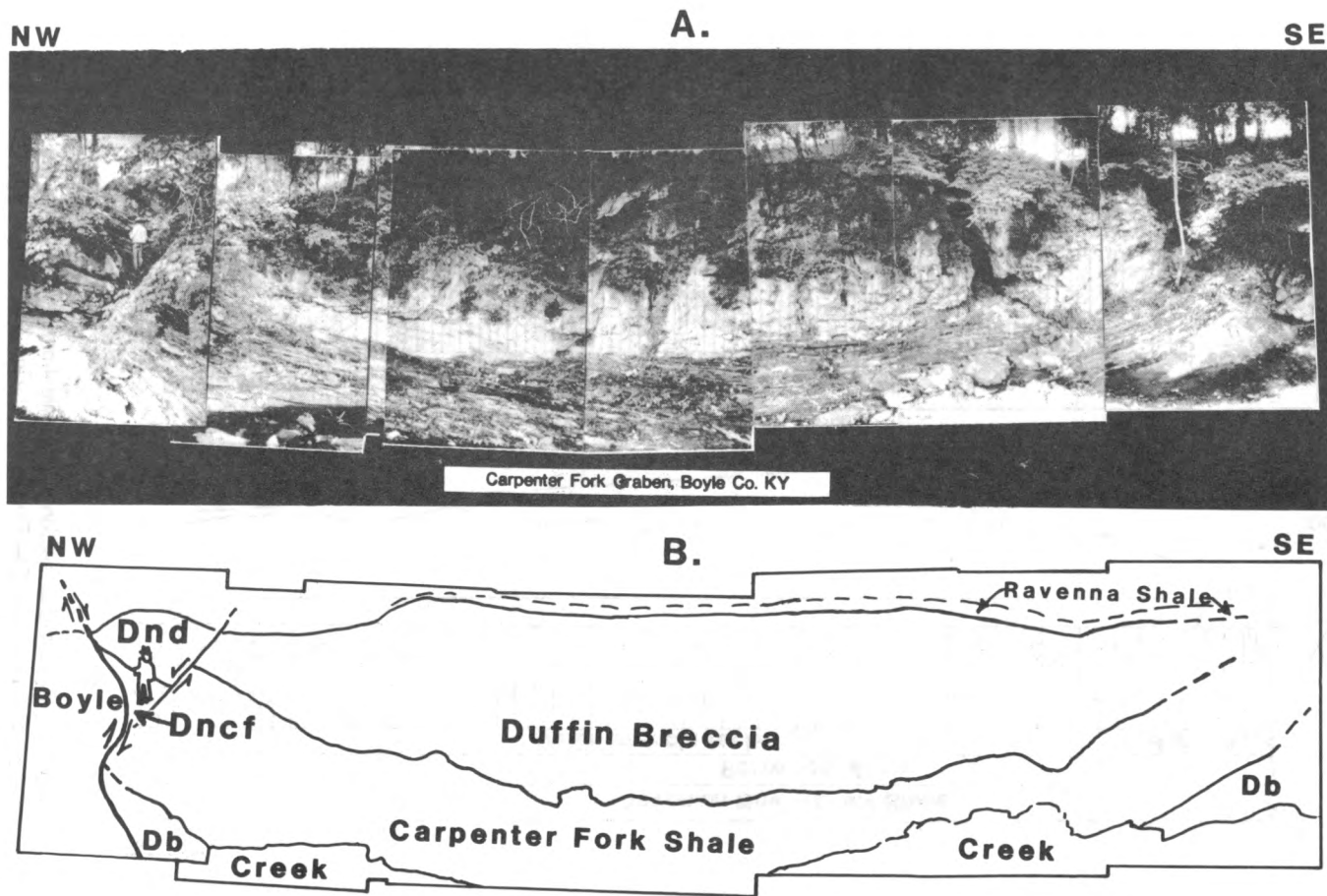


Figure 33. Carpenter Fork graben; southeastern boundary fault not shown. A) Photomosaic; B) Interpretive sketch showing disposition of units; Db, Boyle Dolostone; Dnd, Duffin Breccia; Dncf, Carpenter Fork shale (Photos by C. Mellon).

The Boyle beds just to the west of the structure are roughly horizontal, but tend to dip slightly to the north; close to the bounding fault, however, the Boyle appears to dip into the graben. Within the graben, both the Boyle and the superjacent black shales are upturned against the faults.

Penecontemporaneous soft-sediment deformation of the top of the lower black-shale unit (Carpenter Fork mbr.) indicates that the graben formed in Middle Devonian time. Allochthonous blocks of chert and dolostone, derived from the Boyle on upthrown blocks flanking the graben, apparently slid or fell onto unconsolidated shales and deformed them.

The lowermost beds exposed in the graben are of typical Boyle lithology; that is, fine-grained, locally cherty, dolarenite (Fig. 32). This part of the Boyle coarsens upward and becomes increasingly argillaceous and sandy, and locally contains phosphatic sand and granules, as well as sparse grains of glauconite. The sandy dolostone is thinly bedded, locally micro-cross laminated with organic-rich flaser beds, and extensively burrowed.

The contact between the Boyle and the Carpenter Fork member is conformable. Sandy dolostone in the Boyle grades upward into a dark, sandy, organic-rich dolostone which defines the base of the Carpenter Fork member (Fig. 32). This basal dolostone is also more quartz-rich and less phosphatic than the upper Boyle and is characterized by thin shale partings, crossbedding, scours, local burrowing, and spiriferid brachiopods.

The lower dark, sandy unit of the Carpenter Fork grades upward into fissile, organic-rich, silty, grayish-black shales and minor, brownish-black, interbedded dolosiltites which compose the bulk of the member. The shale is pyritic throughout, and orbiculoid and linguloid brachiopods and burrows are common, especially in conjunction with dolosiltite bodies. Dolosiltite beds are more abundant in the upper seven feet (2 m) of the unit and, locally, are intensely burrowed. Black shales, up to a foot (0.3-m) thick, are present again at the top of the unit, but they are commonly deformed by allochthonous clasts of chert which were dumped into them.

The uppermost part of the Carpenter Fork member is an organic-rich, dolomitic mudstone which was subjected to extensive soft-sediment deformation when chert clasts were dumped into the fault basin. Within the mudstone, clasts of chert, up to 2.3-ft (0.7-m) long, have deformed surrounding shales; in some instances, these clasts are completely surrounded by deformed shale and mudstone. This mudstone has also been injected upward into spaces between clasts of chert and into the overlying dolomitic breccia, and has been folded and overturned.

Above the Carpenter Fork member are the dolomitic breccias and dark shales of the Duffin facies of the Portwood Member of the New Albany. The lower part of the breccia typically exhibits large clasts of dolostone and chert in an organic-rich, silty, dolomitic matrix. It contains clasts of Boyle lithology up to 6.5-ft (2.0-m) long as well as clasts of black shale.

Above the lower breccia, three to 16 ft (1.0-5.0 m) of more typical Duffin breccia is present. This upper unit locally contains black-shale clasts and organic-rich dolomitic siltstone clasts up to boulder size. The unit includes several subunits, each of which fines upward.

The upper breccia is separated from the black-shale facies of the overlying Portwood by a thin argillaceous dolomitic sandstone. The overlying Middle Devonian shale is grayish-black, organic-rich, silty, and blocky to fissile in fresh exposures; locally dolomitic siltstone beds are present.

DISCUSSION

Previous Interpretations of the Carpenter Fork Section

The recognition of some sort of structural influence in the development of the Carpenter Fork section is not new. McFarlan and White (1952) and Rutledge (1957) recognized the presence of faulting on the eastern end of the exposure. Displaced beds indicated to Rutledge a downthrow of about 33 ft (10 m) on the southwest side of the fault. However, neither McFarlan and White nor Rutledge interpreted the cliff exposure as a graben; instead, they suggested that the black shale at the base of the outcrop was a black-shale cave-fill in the Boyle. In their view, the Boyle was deposited and lithified, then a solution cave developed within the Boyle, and the cave was infilled with laminated, black Ohio (Huron equivalent) Shale (Fig. 34A). Deformation in the black shale and sagging of the overlying Duffin layer were attributed to load compaction and, possibly in part, to fault movement (Rutledge, 1957).

Lenhart (1985) recognized the fault at the western end of the cliff face as well as the eastern fault reported by the above workers. Thus, he reinterpreted the Carpenter Fork structure as a graben formed in early Huron time. Lenhart envisioned an allochthonous block or blocks of uplifted Duffin breccia sliding onto the Huron-Shale flooring of a downthrown fault-block (Fig. 34B). As evidence for this scenario, he noted the deformed nature of the shale below the Duffin layer, the anomalous presence of two layers of Duffin separated by a few meters of black shale within the graben cut, and the presence at other locations of penecontemporaneous deformation structures in the lower Huron Shale.

In McFarlan and White's interpretation of the cliff exposure, the Boyle surrounded the Ohio Shale except at the eastern, fault boundary, and the Boyle was overlain by the Duffin and the Ohio Shale (Fig. 34A). Lenhart interpreted all of the carbonates within the graben as Duffin, both those above and below the lower black-shale bed (Fig. 34B). None of these workers, however, noted that the black shale in the graben was gradational with the underlying Boyle, nor did they closely examine the lithology, radioactivity, or biostratigraphy of the various black shales in the exposure. A more detailed analysis of these aspects of the Carpenter Fork graben has led us to identify the units within the graben differently than did previous workers (Fig. 34C), and thus to propose a different origin for the black shales at the base of the graben.

Origin of the Carpenter Fork Member

We agree with Lenhart that the Carpenter Fork structure is a graben. However, on the basis of the gradational contact between the Boyle and the superjacent unit, biostratigraphic analysis (Ettensohn and others, 1991), and dissimilarity between scintillation profiles of the Carpenter Fork black shales and typical Huron shales, we have determined that the lower shale and sandy dolostone units are conformable with the subjacent Boyle and do not represent Huron deposition.

Three interpretations of the Carpenter Fork section.

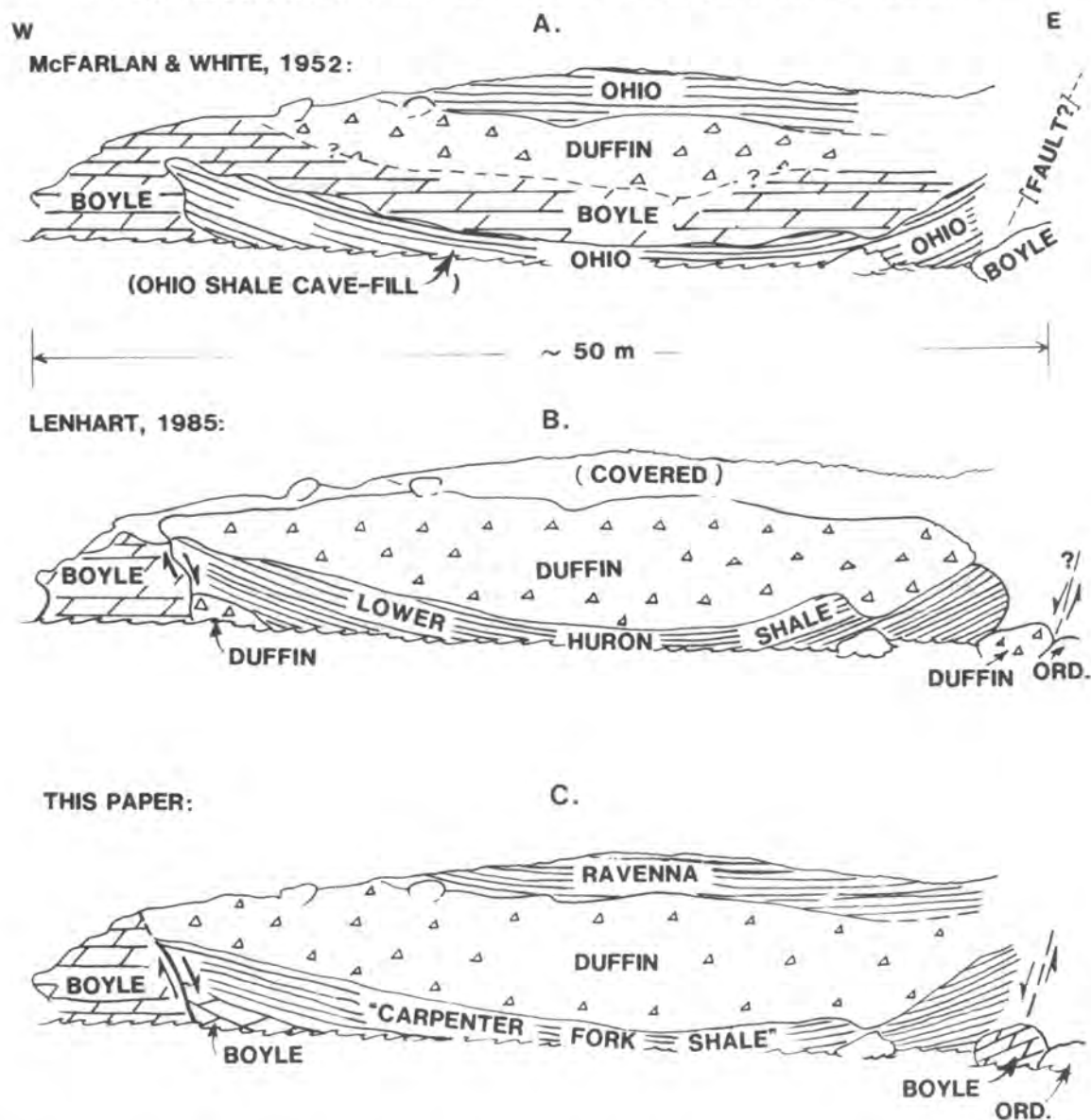


Figure 34. Changing interpretations of the Carpenter Fork exposure: A) Late Devonian cave fill (McFarlan and White; 1952; Rutledge, 1957); B) Late Devonian graben partially filled with Huron Shale with Middle Devonian Duffin sliding in on top at a later date (Lenhart, 1985); and C) Middle Devonian graben filled with Middle Devonian Carpenter Fork shale and Duffin Breccia (Barnett and others, 1989a, 1989b).

We propose the following model for the origin of both the Carpenter Fork and Portwood members of the New Albany. During Middle Devonian time, the Boyle Dolostone was deposited unconformably above Upper Ordovician silty dolostones and limestones along the flanks of the Cincinnati arch. Subsequently, the Carpenter Fork member was conformably deposited above the Boyle, either regionally or in isolated topographic lows, such as sags which might have developed as a result of incipient faulting (Fig. 35A). The Carpenter Fork member was probably a relatively shallow-water deposit. Although black shales are commonly associated with deeper water, deep water is not necessary for their deposition; only restriction and absence of clastic-dilution are required (Ettensohn and Barron, 1981, 1983). The Carpenter Fork member bears evidence of relatively shallow water in the form of scours, crossbedding, wavy and lenticular bedding, micro-cross-laminae, and a *Cruziana* ichnofauna. Moreover, the common presence of orbiculoid and lingulid brachiopods, abundant pyritic nodules and grains, and lack of extensive burrowing, except at the base and top of the unit, suggest a somewhat restrictive and probably dysaerobic environment.

Fault movement followed or coincided with the deposition of the Carpenter Fork member (Fig. 35B); the uppermost beds of that unit are folded upward along the bounding faults in the graben. However, before the upper beds of shale had lithified, clasts of Boyle from the upthrown fault block slid or fell into the graben, deforming the shale (Fig. 35C). The resulting deposits of Boyle lithoclasts constitute the lower part of the Duffin facies at this exposure. Subsequent debris flows in the upper unit of the Duffin are suggested by the upward-fining breccia subunits in upper portions of the cliff exposure.

The breccia units are overlain by a Middle Devonian (Givetian) black-shale unit (Fig. 35D), which has been dated based on conodonts. About five feet (1.5 m) above the base of this shale is a thin sandy, dolomitic, siltstone, which probably represents a lag horizon as well as a hiatus between deposition of the Portwood Member (Ravenna Facies) and the Huron Shale.

Regional Implications

Physically, the Carpenter Fork exposure is small, but it is significant in understanding late Middle Devonian deposition in the central Kentucky region. The association of the shale and superjacent breccia with faulting underscores the importance of tectonism in the deposition of shale and carbonate units on the interior of the craton. This is consistent with preliminary regional studies (Ettensohn and others, 1988b, 1989a) and other unpublished studies in progress, which show a close relationship between distribution of facies in the Portwood Member and the major fault systems discussed earlier in this paper (Fig. 24).

Although it is well known that these central Kentucky faults were active throughout much of the Paleozoic, only recently have mechanisms been proposed to explain the reactivation of these faults and their influence on the regional stratigraphy. As mentioned earlier, the late Givetian marks the beginning of the third tectophase of the Acadian orogeny. We suggest that bulge migration and uplift accompanying the inception of this tectophase reactivated many of these faults creating local fault basins in which the Carpenter Fork shale, Duffin breccias, and overlying Middle Devonian black shales were deposited (Fig. 36B). The deposition was strictly local (Fig. 36B, C), and once the basins were filled and the basin-margin declivities which had provided most of the sediment were buried, little or no sedimentation took place (Fig. 36D). In fact while major sedimentation

took place in the Appalachian and Illinois basins during all of the Frasnian and the early Famennian, the higher Cincinnati arch region was largely an area of very condensed deposition (Fig. 26, 36D). Once the Appalachian and Illinois basins were filled, sediments began to onlap the Cincinnati arch area (Fig. 36E), but not until the late Famennian when the basins finally yoked did sediments completely overlap the Cincinnati arch in the form of the Huron Shale Member (Fig. 26, 36F).

- 136.5 0.9 Turn around and retrace route northward back to junction with State Route 37; turn right onto Route 37 and proceed northeastwardly.
- 137.2 0.7 Junction with Kentucky State Route 34; continue northeastwardly on State Route 37.
- 139.1 1.9 Exposures of the Nancy and Halls Gap members of the Borden Formation on the left.
- 143.9 4.8 City limit, Junction City.
- 144.6 0.7 Junction with Kentucky State Route 300; turn right onto State Route 300 and proceed southeastwardly.
- 145.5 0.9 Junction with U.S. Highway 127; proceed southeastwardly on State Route 300.
- 145.9 0.4 Lincoln County line.
- 146.1 0.2 Turn left on State Route 300 toward the Isaac Shelby Memorial.
- 146.6 0.5 Isaac Shelby Memorial.
- 149.5 2.9 Bridge over Knob Lick Creek; exposures in Upper Ordovician Calloway Creek Limestone.
- 151.1 1.6 Cross the western boundary fault of a graben on the Kentucky River fault system; Devonian New Albany Shale downdropped against the Upper Ordovician Calloway Creek and Ashlock formations with at least 150 ft (45.7 m) of displacement (Shawe and Wigley, 1974).
- 151.4 0.3 Duffin exposures on the right.
- 151.9 0.5 Eastern boundary of graben; less than 100 ft (30.5 m) of displacement.
- 153.2 1.3 Junction with Kentucky State Route 78; turn left onto State Route 78 and proceed eastward.
- 153.4 0.2 Duffin exposed at a spring on the right; the Duffin and New Albany shales here are present in another small graben along the Kentucky River fault system (Shawe and Wigley, 1974).
- 153.5 0.1 Buffalo Spring Cemetery on the left.
- 154.0 0.5 Junction with U.S. Highway 150 (Danville Ave.); proceed eastward on U.S. 150.
- 154.3 0.3 Lincoln County Courthouse.

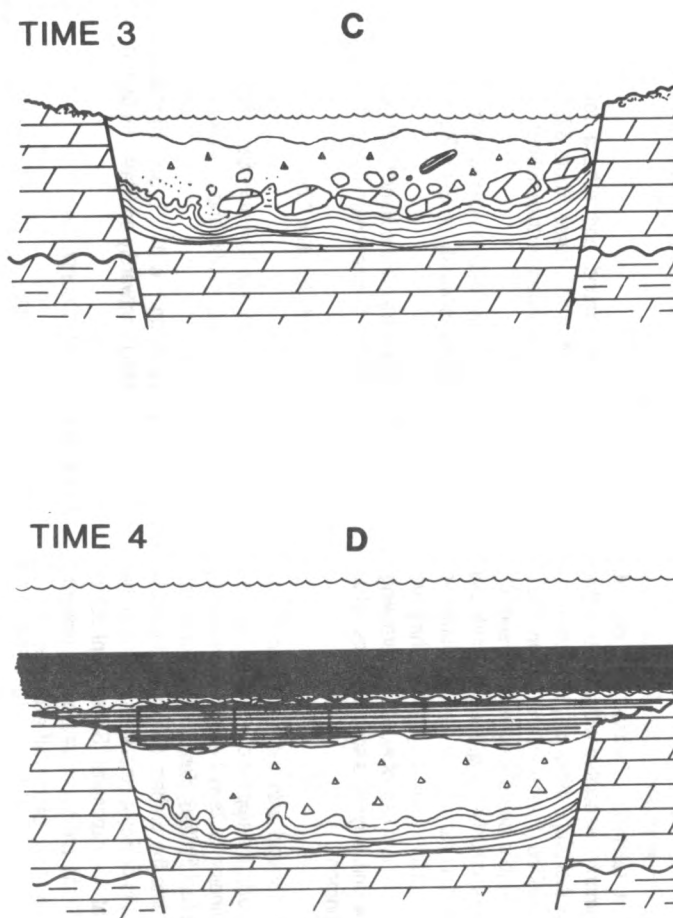
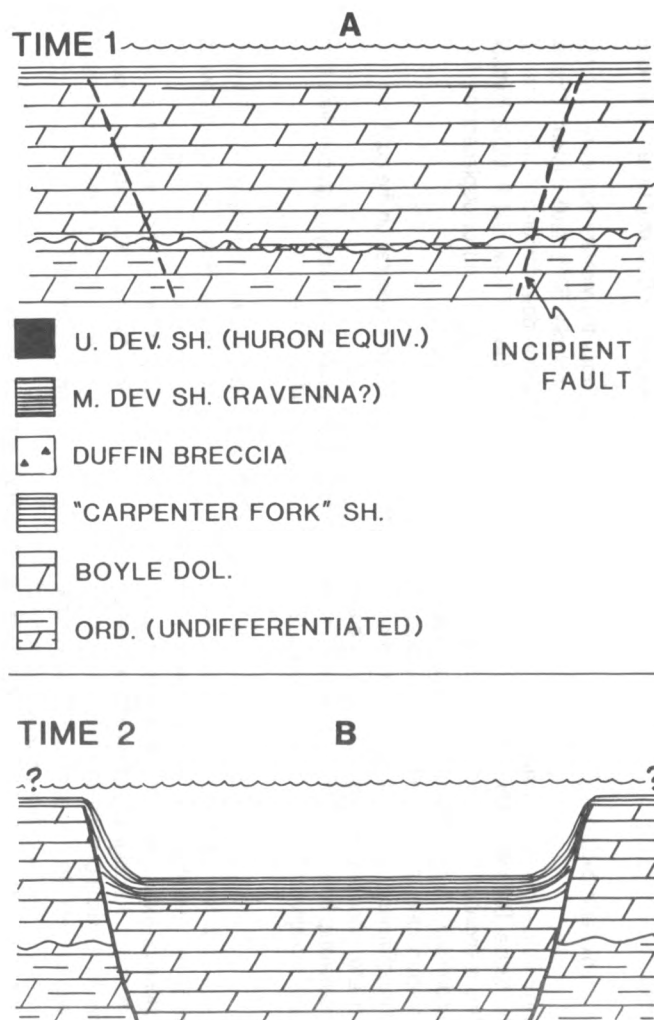


Figure 35. Proposed model for sequence of events in the formation of the Carpenter Fork graben. See the text for description of events (Stop 4, Day 1).

LATE GIVETIAN

B.



EARLY GIVETIAN

A.

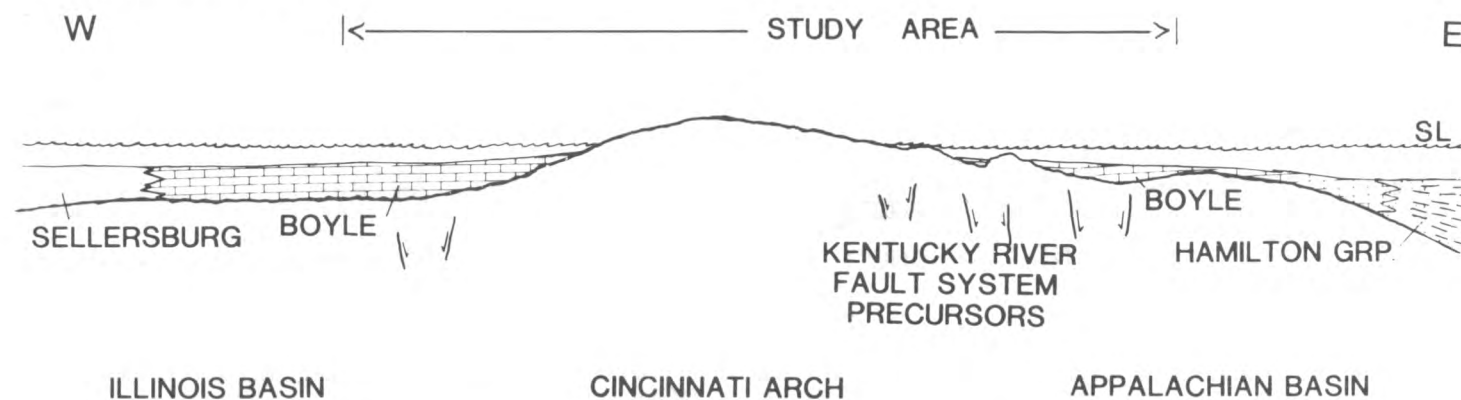
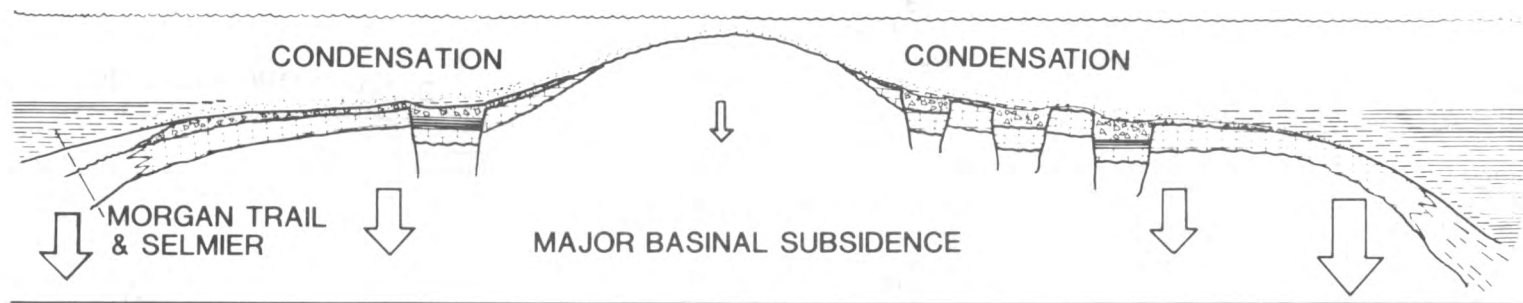


Figure 36. Sequential Middle and Late Devonian history of the Cincinnati arch area in central Kentucky. See text (Stop 4, Day) for description of events. Brick pattern is the Boyle Dolostone; triangles and sand, the Duffin Breccia; dashed lines, green or gray shales; and long dark lines, black shales. *Figure continues on the next two pages.*

LATE FRASNIAN

D.

BASINAL SEDIMENTATION, STARVATION ELSEWHERE



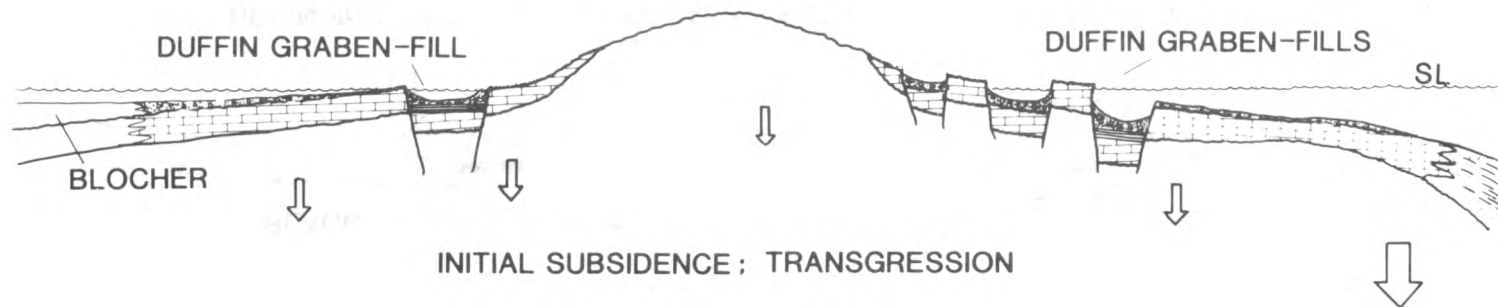
LATEST GIVETIAN

C.

LOCAL, SHALLOW WATER SEDIMENTATION

W

E



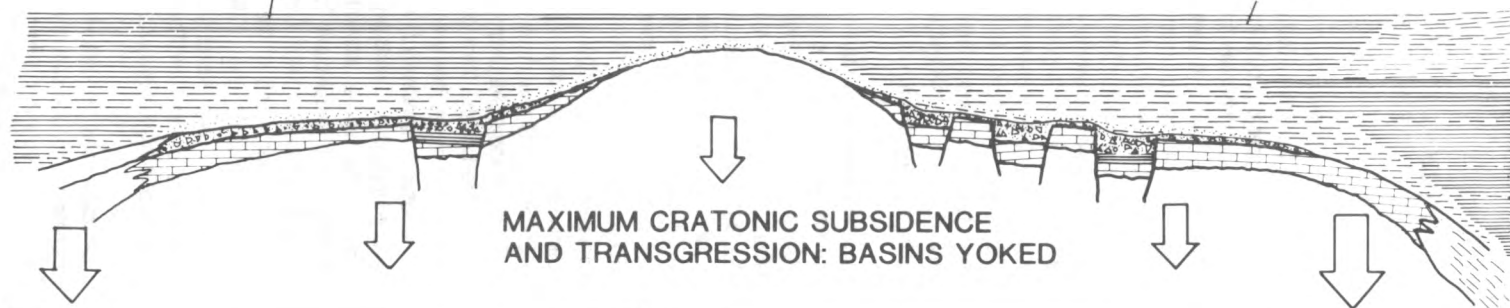
LATE FAMENNIAN

F.

CRATON-WIDE BASINAL SEDIMENTATION

CLEGG CREEK

HURON



EARLY FAMENNIAN

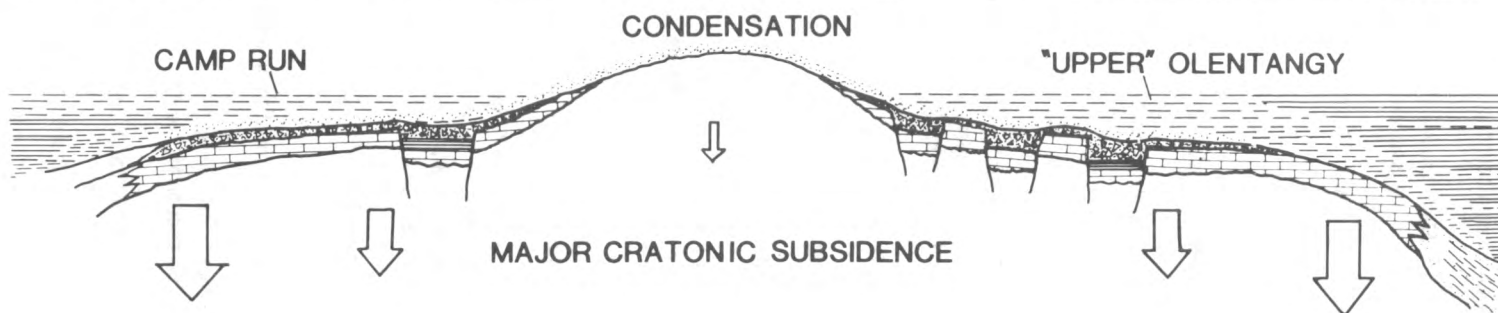
E.

BASIN INFILLED:
SEDIMENTS BYPASS BASIN AND ONLAP ARCH

W

E

SL



- 154.8 0.5 Junction of U.S. 27 and U.S. 150; go straight on U.S. 150.
- 155.6 0.8 Railroad crossing; bear right on U.S. 150.
- 156.4 0.8 Upper Ordovician Drakes Formation.
- 161.6 5.2 Contact between the Drakes and Brassfield formations; Ordovician-Silurian boundary.
- 162.3 0.7 Road to the William Whitley House State Shrine on the right; "Knobs" in the distance.
- 163.3 1.0 Outcrops of the Lower Silurian Crab Orchard Shale on the right.
- 164.2 0.9 Crab Orchard city limits.
- 164.7 0.5 U.S. 150 veers right; stay on U.S. 150.
- 165.2 0.5 Junction with Kentucky State Route 39; stay on U.S. 150.
- 166.8 1.6 Exposures of the Boyle, Duffin and New Albany.
- 167.8 1.0 Contact of the New Albany Shale with the Mississippian Borden Formation; approximate beginning of the Appalachian Plateau (Fig. 3).
- 167.9 0.1 Rockcastle County line.
- 170.9 3.0 Brodhead city limits.
- 171.6 0.7 Junction with Kentucky State Route 1505; bear right on U.S. 150.
- 172.4 0.8 Junction with Kentucky State Route 3245; continue on U.S. 150.
- 176.0 3.6 Carbonates of the Middle and Upper Mississippian Slade Formation (Ettensohn and others, 1984).
- 176.5 0.5 Junction with Kentucky State Route 46. The Upper Mississippian section in a roadcut 2.0 mi to the northeast on this highway will be compared with the section at STOP 5 (Fig. 37).
- 177.3 0.8 Large crossbeds in the Ste. Genevieve Member of the Slade Formation.
- 177.7 0.4 Quarry in the Slade Formation on the left.
- 178.3 0.6 Junction of U.S. Highways 25 and 150; continue straight on U.S. 150 through Mt. Vernon.
- 180.4 2.1 Turn left onto the entrance ramp for northbound I-75.
- 180.6 0.2 Enter onto I-75 and proceed northward.
- 181.8 1.2 Crossing a solution valley or uvala in the lower Slade Formation; the north end of the valley is bound by the Mt. Vernon fault with downthrow to the north.

- 182.1 0.3 STOP 5. The Slade and Paragon Formations and Pennsylvanian Slumping, Mt. Vernon, KY. (large roadcut on the northbound lane of I-75, 1.7 mi north of its southern intersection with U.S. Highway 25, Rockcastle Co.; Carter Coordinates: 50°FNLXx1300°FWL, 17-K-63).

THE SLADE AND PARAGON FORMATIONS AND PENNSYLVANIAN SLUMPING

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CORTLAND F. EBLE, JACK C. PASHIN, AND STEPHEN F.
BARNETT

In this and adjacent cuts to the northwest, a nearly complete section of the Slade Formation (former Newman Limestone) is exposed (Fig. 37A). The upper St. Louis through Poppin Rock members are present in this large cut, whereas the Renfro Member is exposed just to the north at Mile 183.3. In contrast to the Slade sections which we will examine tomorrow on Interstate Highway 64 (STOPS 6-8, Day 2), there is little indication of syndimentary tectonic activity during carbonate deposition. The thick nature of the carbonate section as well as the predominance of shallow open-marine limestones is typical of Slade carbonates deposited south of and beyond the influence of the Waverly arch or Kentucky River fault system (Ettensohn, 1977, 1980; Dever and others, 1977). Even though this exposure is located close to the Mt. Vernon fault (Schlanger and Weir, 1971), there is no evidence at present for syndimentary movement on the fault during carbonate deposition.

Here the Slade is overlain unconformably by shales, siltstones and sandstones of the Breathitt Formation. Normally, the Upper Mississippian Paragon Formation (former Pennington Formation) would intervene between the Slade and Breathitt formations, but the Paragon, which is present less than two miles to the west (Fig. 37B) is absent in this exposure along the mid-Carboniferous (Early Pennsylvanian) unconformity (Fig. 37A). The unnamed Breathitt coal bed in the western exposure (Fig. 37B) was sampled for palynologic analysis. The 5.5-in (14-cm) coal bloom (heavily weathered coal) contained 31% *Lycospora* (swamp lycopods), 42% *Schulzospora* (pteridosperm, probably *Lyginopteris*), 11% *Densosporites* (small lycopods), 8% *Granulatisporites* (mostly small ferns with the possible inclusion of some pteridosperms), and small quantities of other spore types. Most of the thick coals of the Breathitt are dominated by *Lycospora*. Here, however, *Schulzospora* is unusually abundant. *Schulzospora* apparently had more environmental latitude than the swamp lycopods and probably reflects a stressed peat setting of some type. The spore assemblage is most likely early Middle Pennsylvanian (equivalent to the lower Kanawha Fm. in West Virginia), although a latest Early Pennsylvanian age cannot be ruled out. Although much of the Paragon Formation is present in the western exposure (Fig. 37B), strata equivalent to most of the Pennington Group (latest Mississippian) and the Pocahontas and perhaps all of the New River Formation (Early Pennsylvanian) of Virginia and West Virginia are absent here.

The carbonates in this section appear to be parts of three unconformity-bound sequences. Only upper parts of the lower one are evident here in the form of the upper Ste. Genevieve Member. According to Dever and others (1979), the St. Louis and Ste. Genevieve members are conformable here as they are elsewhere to the south away from the apparent syndimentary effects of local structures. However, locally to the south near the Greenwood

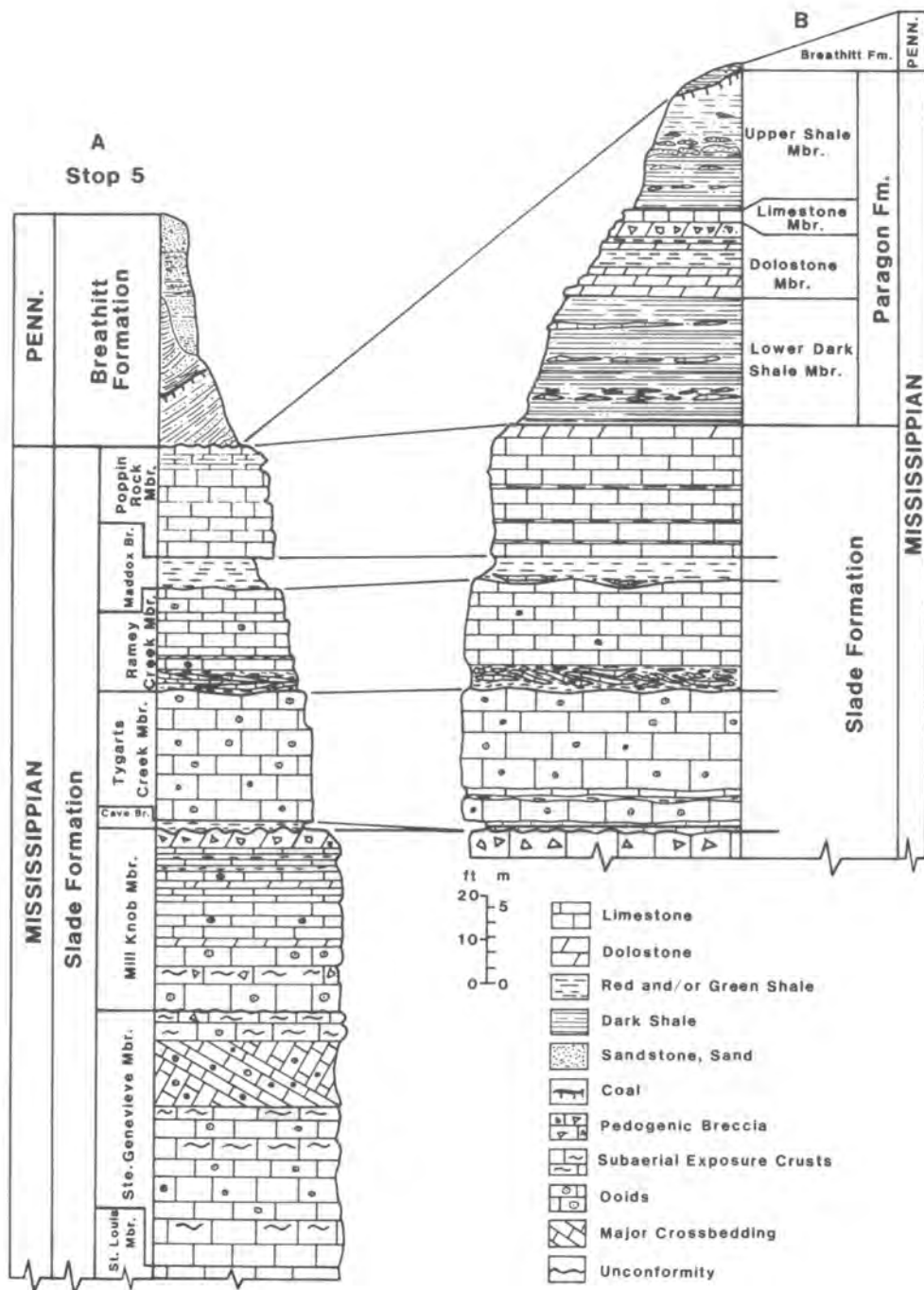


Figure 37. Comparison and correlation between the Carboniferous section at Stop 5 on Day 2 (A) with a similar section 2.8 mi due west of Stop 5 (B). Note the absence of the Paragon Formation and the presence of slumping in the Breathitt at Stop 5 (see Figure 39).

anomaly, a probable fault-bound basement block (Dever, 1990), and to the north near the Waverly arch and Kentucky River fault zone (Dever and others, 1977; Ettensohn, 1980, 1981), the contact is unconformable. The Ste. Genevieve is composed largely of crossbedded skeletal and oolitic calcarenites with lesser amounts of calcilutite. The unit is subdivided by four horizons of calcareous paleosols or subaerial exposure crusts (Fig. 37) indicated by the presence of laminated micritic crusts and micrite-filled root tubes formed by vadose diagenesis (Dever and others, 1979; Ettensohn and others, 1988a). The highest of these defines a subtle break or disconformity with the overlying Mill Knob Member of the Slade Formation (Fig. 37). Each of these soil horizons is commonly overlain by peloidal and intraclastic calcarenites (Dever and others, 1979) apparently reflecting renewed transgression.

The overlying Mill Knob Member (former Paoli-Beaver Bend Member) is the second unconformity-bound sequence present in this exposure. The Mill Knob is a lower Chesterian unit containing two fining-upward subsequences (Fig. 37A). Each generally begins with a skeletal or oolitic calcarenite and grades upward into calcilutites containing intertidal features and is capped by a paleosol or caliche horizon containing breccias, laminated micritic crusts, root tubes, and tepee structures. The upper exposure horizon is everywhere the best developed and marks an unconformity surface that can be traced along the entire length of the outcrop belt (Dever and others, 1979; Dever, 1990).

The overlying five members of the Slade Formation represent deposition in a transgressive continuum (Cave Branch-through-lower Maddox Branch members) which was followed by a regressive progradation (upper Maddox Branch-through-Poppin Rock members) (Ettensohn, 1977, 1980, 1981).

In this transgressive sequence the Cave Branch shale accumulated on intertidal mud flats adjacent to an exposed Mill Knob surface. The overlying Tygarts Creek Member accumulated as a carbonate sandbelt. The lower part of the member is a well-sorted, oolitic calcarenite which grades upward into crinoidal calcarenite. The overlying Ramey Creek Member has the most diverse fauna (brachiopods, bryozoans, crinoids, blastoids, pelecypods, and rugose corals) of any Slade unit and represents shallow, open-marine deposition seaward of the carbonate sandbelt. It consists of bioclastic and oolitic calcarenites, dolostones, and shales. The presence of shales in this unit as well as the muddy matrix in most of its calcarenites suggests somewhat deeper water deposition than that characterizing the Tygarts Creek sandbelt. The Maddox Branch member is largely a shale and represents deeper, open-marine deposition in quiet conditions well below wave base; fossils are rare but may include productoid brachiopods and pelecypods.

The change from a transgressive to a regressive continuum is interpreted to have occurred during Maddox Branch deposition. The overlying Poppin Rock Member represents another sandbelt, the progradational analogue of the transgressive Tygarts Creek. The upper shaly part of the unit which commonly contains many well preserved fossils (bryozoans, brachiopods, and crinoids) was apparently deposited in a shallow, protected marine environment between the carbonate sandbelt and more shoreward lagoonal and tidal-flat environments represented by the Paragon Formation. Although eroded away at this exposure, the Paragon is a natural part of the regressive continuum and is preserved in exposures nearby (Fig. 37B).

In some regards, the three unconformity-bound sedimentary packages present here are somewhat similar to the depositional sequences of the sequence stratigraphers (e.g., Van Wagoner and others, 1988), but the cratonic setting is decidedly different from the continental-margin situations in which the ideal sequence-stratigraphy models were developed (e.g., Vail and others, 1977; Van Wagoner and others, 1988). Hence, our interpretations of depositional sequences may be crude analogies at best, but nonetheless, each of the sequences in this section has what could be interpreted as a transgressive systems tract followed by a highstand systems tract. In the lower sequence, the St. Louis, which is not well exposed here, represents the transgressive systems tract. Away from structures like the Kentucky River fault zone and Waverly arch, the contact between the St. Louis and the Ste. Genevieve is essentially conformable, although the contact is usually sharp or slightly erosional, and the overlying Ste. Genevieve may contain eroded clasts (Dever, 1990, fig. 8). This contact could well represent a surface of maximum flooding, with the overlying Ste. Genevieve representing the highstand systems tract; the four shallowing-upward cycles in the Ste. Genevieve may reflect parasequence-like units.

The unconformity atop the Ste. Genevieve would represent the sequence boundary and the transgressive surface of the next transgressive systems tract. However, the unit which probably represents this transgression, the Warix Run Member of the Slade Formation (Ettensohn and others, 1984), is absent here. The Mill Knob Member which is preserved, is the regressive part of the transgressive-regressive Warix Run-Mill Knob continuum and may represent the highstand systems tract in this sequence.

The disconformity atop the Mill Knob marks the lower boundary of the probable third sequence, as well as the transgressive surface for the next transgressive systems tract. This tract begins with the Cave Branch and ends at the top of the Maddox Branch Member with a subtle disconformity. Overlying parts of the Mississippian, here represented only by the Poppin Rock Member of the Slade Formation (Fig. 37A), but elsewhere including the Paragon or Pennington formations (Fig. 37B), reflect a regressive, westward progradation (Ettensohn, 1977, 1980, 1981). Progradations like this, which migrated into an area from the opposite side of a basin, are unaccounted for in the basic sequence-stratigraphy models (e.g., Vail and others, 1977; Van Wagoner and others, 1988), but in this situation the prograding sequence most closely approaches the concept of a highstands systems tract.

The entire carbonate sequence here, as well as underlying and the overlying Mississippian clastic sequences which are either not exposed or preserved here, are more easily, and perhaps more convincingly, explained in terms of flexural models already discussed. In this context, all of the carbonates from the St. Louis to the Tygarts Creek would have formed during late stages of loading-type relaxation (Fig. 28B) when eastward bulge uplift and migration generated an extensive area of shallow-water conditions conducive to carbonate deposition. This relaxation apparently reached its peak in the late Middle Chesterian when a sheet of very shallow oolitic sand, represented here by the Tygarts Creek, spread across much of the basin and craton in response to an equilibrium state between the filled foreland basin and eroded source areas (Fig. 28C) (Ettensohn and Chesnut, 1989; Ettensohn, in press a).

Although unconformities are present in the lower Slade, they can be accounted for in three ways. Some like that which occurs locally between the St. Louis and Ste. Genevieve probably reflect bulge

reactivation of basement structures. On the other hand, the unconformity which occurs between the Ste. Genevieve and the Mill Knob, at the Meramec-Chester boundary, formed at about the same time as the beginning of an Ouachita tectophase (Mack and others, 1983; Ettensohn, in press a), and may reflect distal bulge uplift and migration accompanying that tectophase. As would be expected, the areal extent of this unconformity roughly parallels the northern margin of the Ouachita foreland basin (Ettensohn, in press a, fig. 8). Finally, the early Chesterian unconformity between the Mill Knob and Cave Branch may reflect a eustatic drop in sea level; it is extremely widespread and corresponds with no known tectonic event.

Returning to the flexural scenario, the Ramey Creek and Maddox Branch members represent a relatively sudden and widespread increase in depth, which could reflect the passage of an anti-peripheral bulge and mark the beginning of unloading-type relaxation (Fig. 28D). The overlying Poppin Rock Member of the Slade as well as the Paragon Formation (and Pennington Formation farther east) probably reflect subsequent cratonward progradation of marginal-marine and terrestrial sediments which typically represents the culmination of unloading-type relaxation (Fig. 28D) (Ettensohn and Chesnut, 1989).

The Early Pennsylvanian unconformity which truncates the Mississippian rocks here and elsewhere in Kentucky (Chesnut, 1988), most probably reflects bulge uplift and migration accompanying the inception of the Alleghanian orogeny (Fig. 28F) (Ettensohn and Chesnut, 1989). At least 82 ft (25 m) of Paragon, present on the western exposure (Fig. 37B), is absent here at Stop 5. The deep truncation of the Paragon Formation here, and of the Paragon and Slade formations nearby, occurs in close association with the Mt. Vernon monocline and the Mt. Vernon fault (Fig. 38) (Chesnut, 1991a). The progressive truncation of the Paragon toward the monocline and fault suggests that erosion occurred after Paragon deposition in association with movement on one or both of the structures. Moreover, the coincidence of deep truncation along the unconformity with these structural elements, also suggests the reactivation of these structures during the Alleghanian orogeny. In addition, the large slump structure present at this stop (Fig. 39) may similarly reflect reactivation of nearby structures.

In the upper part of the cut, Middle Pennsylvanian rocks of the lower part of the Breathitt have been deformed by large-scale slumping involving both downward and horizontal movement to the south (Fig. 39). Only two Breathitt units were involved in the slumping. The lower unit which unconformably overlies the Slade Formation consists of lenticularly bedded sandstones and shales with much bioturbation; it is separated from the upper unit by a thin coal or carbonaceous shale, along which much of the horizontal movement appears to have occurred. The upper unit is intensely burrowed, wavy- to flaser-bedded sandstone and shale with siderite nodules. Horizontal and vertical burrows among others include *Teichichnus* and *Skolithos*-like trace fossils. Two blocks of the upper unit show downward rotation and concomitant horizontal movement along the coal bed toward the south. As a result, beds in the lower unit were intensely crumpled and faulted. Neither the underlying carbonates nor the overlying sandstones were involved in the slumping. The two triangular fillings at the base of the sandstone represent the initial filling of lows developed in front of the slump scarps on the backward-rotated slump blocks. Each is a fining-upward sequence of sandstone and sandstone and shale related to the overlying channel sand. Although Dever and others (1979) reported the involvement of the Paragon

Formation (Pennington shales) in this slump, close inspection reveals no evidence of Paragon lithologies.

Dever and others (1979) suggested the slumping could be related to one of two causes: 1.) fluvial or tidal channeling to the immediate south, or 2.) seismicity. Because of the absence of evidence for a channel, the presence of similar structures at the same horizon elsewhere in the area, and the likely coincidence with the inception of Alleghanian tectonism, Dever and others (1979) favored the seismic interpretation. Chesnut (1991a) even suggested that recurrent movements on the nearby Mt. Vernon monocline or fault (Fig. 38) might be responsible.

- | | | |
|-------|-----|---|
| 182.5 | 0.4 | Large-scale crossbedding in the Ste. Genevieve Member at the north end of the cut. |
| 183.3 | 0.8 | Exit ramp to U.S. 25 (Exit 62); Wildie Member of the Borden and Renfro Member of the Slade exposed along the ramp. |
| 184.0 | 0.7 | Renfro and Wildie members. |
| 184.2 | 0.2 | The community of Renfro Valley on the right. For approximately the next 3.8 mi, the highway crosses the greenish-gray shales and siltstones of the Wildie and Halls Gap members of the upper Borden Formation, as well as the brown dolostones of the Renfro Member and the white limestone of the St. Louis Member of the Slade Formation. |
| 188.0 | 3.8 | For approximately the next 7.8 mi the highway crosses the greenish-gray shales of the Nancy Member of the Borden as well as the gray siltstones of the Cowbell Member of the Borden Formation. |
| 195.8 | 7.8 | Madison County line. |
| 196.8 | 1.0 | New Albany Shale. |
| 197.4 | 0.6 | New Albany Shale. |
| 197.7 | 0.3 | Exit 76 to Berea. |
| 198.2 | 0.5 | Cherty Boyle Dolostone (Devonian) and Duffin Bed of the New Albany exposed along entrance ramp of Exit 76. |
| 199.3 | 1.1 | Green shales of the Crab Orchard Formation (Lower Silurian) exposed on the left. |
| 200.2 | 0.9 | Northern Berea Exit 77; contact between the dolomitic mudstones and shales of the Preachersville Member of the Drakes Formation (Upper Ordovician) and the Lower Silurian dolostones of the Brassfield Formation. |
| 201.1 | 0.9 | Crossing southern bounding fault of a graben; downthrow is to the north with a displacement of 40 ft (12 m) (Weir, 1967). |
| 201.8 | 0.7 | Contact between the Ordovician Drakes and Silurian Brassfield formations; the contact is sharp and erosional and the basal Brassfield is locally conglomeratic containing clasts from the underlying Drakes Formation. |

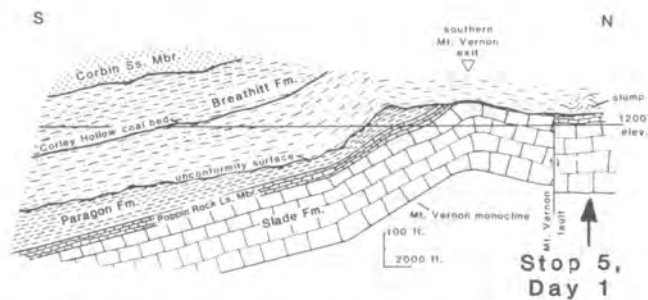


Figure 38. Schematic cross section showing the disposition of Stop 5 and the slumping therein relative to the Mt. Vernon fault and Mt. Vernon monocline (from Chesnut, 1991a).

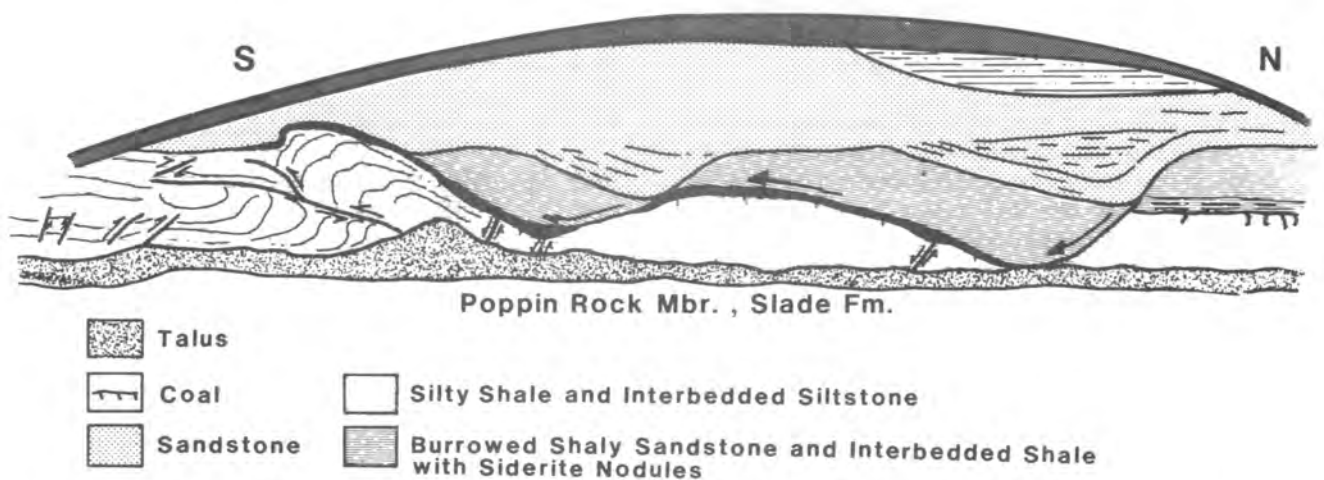


Figure 39. Schematic north-south diagram showing the nature of deformation in the lower part of the Breathitt at Stop 5 on Day 2. Most of the deformation involves both downward and lateral movement to the south of one shale unit on a coaly horizon atop another. The underlying Slade Formation is not involved in the slumping.

202.0	0.2	Northern bounding fault of the graben; downdrop is to the south with displacement of about 45 ft (15 m) (Weir, 1967).	Lexington Limestone; displacement varies from 75 to 300 ft (23-92 m) (Black and MacQuown, 1965; Black, 1968).
202.3	0.3	Ashlock and Drakes formations.	220.8 0.4 Tyrone Limestone.
202.7	0.4	Milepost 80; Ashlock and Drakes formations. For the next eight miles the field-trip route largely crosses the fossiliferous limestones of the lower Ashlock Formation; on some of the adjacent hills upper dolomitic parts of the Ashlock are present whereas the Calloway Creek is present in some of the deeper valleys (Greene, 1966).	221.4 0.6 Planar-bedded, cherty fossiliferous calcarenite of the Curdsville, basal member of the Lexington Limestone, exposed along northbound lane.
205.0	2.3	Entrance ramp to rest area.	221.6 0.2 Exit 99 to U.S. Highways 25 and 421. Steeply dipping beds just north of the exit are on the upthrown side of the south bounding fault of the Elk Lick Creek graben of the Kentucky River fault zone (Black, 1968); Grier-Curdsville transition.
209.5	4.5	Exit 87 to Richmond and Lancaster via Kentucky State Route 876; Tate Member of the lower Ashlock.	222.0 0.4 Leave Kentucky River fault zone.
210.8	1.3	Tates Creek fault; downthrown side is to the north with a displacement of about 100 ft (30 m) (Simmons, 1967). For the next 3.3 mi., the route principally crosses the nodular to irregularly bedded limestones of the Grant Lake Member of the lower Ashlock; dolostones from the Terrill and Reba members of the upper Ashlock occur locally on some of the higher ridges (Simmons, 1967).	223.0 1.0 Milepost 100; nodular and irregularly bedded fossiliferous limestone of the Grier Member of the Lexington Limestone.
212.2	1.4	Exit 90 to Richmond; Terrill Member of the upper Ashlock.	223.8 0.8 Transition between the Grier and Tanglewood members of the Lexington Limestone.
214.1	1.9	Approximate position of the Richmond fault; downthrow is to the north with a displacement of about 20 ft (6 m) (Simmons, 1967). For the next two miles, the route primarily crosses dolostones of the upper Ashlock and Drakes formations, with the Grant Lake Member of the Ashlock in some of the deeper valleys (Simmons, 1967).	225.9 2.1 Tanglewood and Brannon members of the Lexington Limestone.
216.1	2.0	From this point to Exit 95, the route primarily crosses the Grant Lake Member of the Ashlock.	226.3 0.4 Calcarenite of the Tanglewood dips northward toward a small graben (Black, 1967).
217.3	1.2	Exit 95 to Boonesboro via Kentucky State Route 627; for the next 0.7 mi, the route crosses the Tate Member of the Ashlock Formation.	226.9 0.6 Exit right on Exit 104 ramp; rubbly-weathering, nodular limestones of the Millersburg Member of the Lexington are well exposed under the overpass.
218.0	0.7	For the next 1.5 mi to Exit 97, the route crosses the thin-bedded fossiliferous limestones and interbedded shales and siltstones of the Calloway Creek Formation.	227.1 0.2 Turn right onto Kentucky State Route 418 (Athens-Boonesboro Road).
219.5	1.5	Exit 97 to U.S. Highways 25 and 42; Clays Ferry exit.	227.2 0.1 Turn right into the Holiday Inn parking lot.
219.9	0.4	Garrard Siltstone on the right; zones of contorted siltstone and calcirudite (ball and pillow structures) are common in the unit.	END OF ROAD LOG FOR DAY ONE.
220.1	0.2	Interbedded shales and limestones of the Clays Ferry Formation on the right.	
220.4	0.3	South end of the Clays Ferry bridge spanning the Kentucky River gorge; the south end of the bridge marks the beginning of the Kentucky River fault zone, which in this area is a belt of grabens and horsts about 1.3 mi wide. In general, the upper Lexington Limestone and Clays Ferry Formation are downdropped to the south against the High Bridge Group and lower	

ROAD LOG AND STOP DESCRIPTIONS—SECOND DAY

Mileage:

cum. inc.

0.0		Assemble in parking lot of Holiday Inn, Lexington South.
0.1	0.1	Proceed northward out of parking lot; turn left onto Kentucky State Route 418 and proceed northwestward.
0.2	0.1	Turn right into entrance ramp onto northbound I-75.
0.4	0.2	Tanglewood Member exposed on ramp.
0.5	0.1	Enter onto northbound I-75.
0.7	0.2	On the right, begin an exposure in the main body of the Tanglewood.

- 1.2 0.5 Nodular Millersburg Member.
- 1.3 0.1 Milepost 105.
- 1.5 0.2 Millersburg Member.
- 2.6 1.1 Overpass; Millersburg to the north.
- 3.2 0.6 Begin a large Millersburg exposure with a one- to five-foot (0.3-1.5-m) thick tongue of the upper Tanglewood pinching out southwardly into the Millersburg.
- 3.3 0.1 Milepost 107.
- 3.7 0.4 Overpass.
- 4.2 0.5 STOP 1. Regressive Facies in the Upper Lexington Limestone: Tanglewood-Millersburg Relationships (Ramp for Exit 109 off of northbound I-75 up to the entrance ramp onto northbound I-75, Fayette Co.; Carter Coordinates: 1600'FSLx1850'FWL, 16-S-62). The bus will drop people off at the bottom of the exit ramp and pick them up at the top of the ramp.

REGRESSIVE FACIES IN THE UPPER LEXINGTON LIMESTONE: TANGLEWOOD-MILLERSBURG RELATIONSHIPS

FRANK R. ETENSOHN

Following the intertidal-supratidal deposition that culminated in the late Black Riverian to Rocklandian (Fig. 7), the Lexington Limestone and its equivalents, as well as overlying Edenian units like the Kope, Clays Ferry and Martinsburg, appear to represent a time of regional transgression that coincided with the inception of the Taconic tectophase (Fig. 7) and major foreland-basin subsidence to the east (Ettensohn, 1991a). Transgression apparently ended near the beginning of the Maysvillian, and the rest of the Ordovician generally represents a period of regional regression (Fig. 7) probably reflecting the onset of Taconic relaxation and Gondwana glaciation. These same basic patterns are observed in southwestern Ohio (Weir and others, 1984) and in parts of north-central Kentucky, where the Lexington Limestone is transgressive through the Grier Member into the overlying interbedded shales and limestones of the Clays Ferry and Pt. Pleasant formations (e.g., Cressman, 1973; Luft, 1974). In central Kentucky, however, the Lexington Limestone is generally transgressive through the Grier or Brannon members, and then abruptly becomes predominantly regressive into the Tanglewood and Millersburg members; major transgression resumes with the Clays Ferry Formation (Fig. 40). Regressive parts of the upper Lexington in this central Kentucky area are actually as young as Late Ordovician (Edenian), whereas in surrounding areas the Lexington is generally no younger than late Shermanian (Figs. 1, 40) (Sweet, 1979; Weir and others, 1984). These age relationships, as well as a series of cross sections prepared by Cressman (1973), suggest that Lexington deposition primarily in the form of the Tanglewood and Millersburg members, persisted longer in the area of the Jessamine dome than elsewhere in the region. The cross sections suggest that a high-energy, regressive, carbonate buildup began forming on the Jessamine dome area of the Lexington platform during the late Shermanian, while transgression and deepening were occurring elsewhere on the platform. Regional facies patterns (Freeman, 1953; Keith, 1988) indicate that the Jessamine dome had been an area of

uplift and shoaling since the inception of Lexington deposition in the Kirkfieldian, but uplift had apparently not been sufficient to overwhelm the pattern of regional transgression. By late Shermanian time, however, this changed, and major uplift gave rise to thick, high-energy, shoal sediments (Tanglewood Member) that intertongue with shallow open-marine (Millersburg Member) and deeper open-marine (Clays Ferry and Pt. Pleasant formations) sediments in all directions from the dome region (Figs. 1, 40). Paleobathymetric indicators and facies distributions commonly parallel known structural trends (Borella and Osborne, 1978), and in some cases, Lexington members are apparently delimited along structures (see Stop 3, Day 2) or influenced in other ways by them (e.g., Grossnickle, 1985; Ettensohn and others, 1986), all of which suggests that Middle Ordovician reactivation of faults associated with the Lexington and Kentucky River fault systems was responsible for generating the relief necessary for formation of a high-energy carbonate buildup in the Lexington area (Borella and Osborne, 1978). Inasmuch as Lexington deposition coincided with the inception of the Taconic tectophase of the Taconian orogeny (Ettensohn, 1991a), it is likely that fault reactivation may have been related to cratonward bulge migration accompanying the orogeny. A partial outline of the buildup boundaries can be approximated by mapping the distal edge of the Tanglewood, and at least locally the boundaries of the buildup coincide with the trends of known faults (Fig. 41). This buildup is informally called the Tanglewood buildup due to the predominance of Tanglewood lithofacies within it.

The typical transgressive parts of the Lexington Limestone include in ascending order the Curdsville, Grier, and Brannon members. They represent respectively shoal-sandbelt, shallow open-marine, and deeper open-marine environments. Tempestites or storm deposits appear to predominate in each unit; proximal tempestites are well developed in the Curdsville (Grossnickle, 1985), but in the Grier intense bioturbation has destroyed many indications of storm origin. In the Brannon, fine-grained distal tempestites predominate.

Coarse-grained calcarenites and calcirudites of the Tanglewood Member, like those exposed at this stop, abruptly overlie the Grier and Brannon members in the area of the buildup (Fig. 40). In most parts of the buildup, the Tanglewood is composed of three separate bodies, a central body, a lower tongue and an upper tongue; many small isolated bodies are also present. Each of the tongues is generally separated from the other by intervening tongues of the Millersburg Member (Fig. 40), a nodular to irregularly bedded fossiliferous limestone and shale, and each of these tongues may reflect an individual episode of uplift on the Jessamine dome. The Tanglewood grades vertically and laterally into the Millersburg, which in turn grades into the Clays Ferry Formation. Locally, the Tanglewood may grade directly into the Clays Ferry.

Tanglewood calcarenites and calcirudites are characterized by a fauna of stout bryozoans, as well as hemispherical coral and stromatoporoid colonies; these colonies are commonly fragmented or overturned. Crossbedding, scours, and small-scale channelling are everywhere common. The Tanglewood is interpreted to represent high-energy carbonate shoal or sandbelt environments at or above wave base (Fig. 42). Small crossbedded tongues of the Tanglewood that are common in the Millersburg Member apparently represent spillover lobes or dunes of sand that migrated into deeper waters during storms.

The Millersburg contains an abundant and diverse fauna, and the abundance of shale in the unit indicates that it was largely deposited

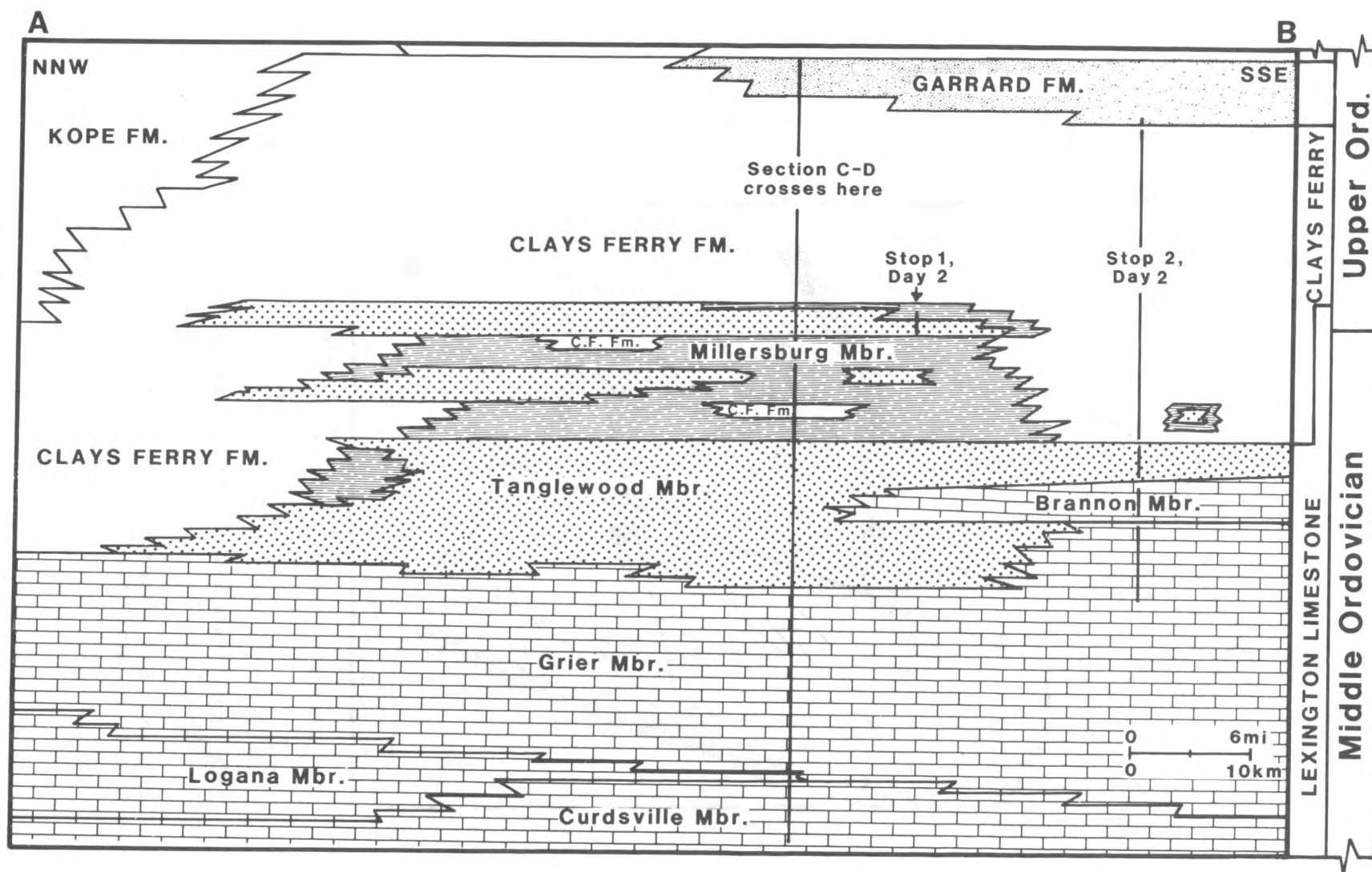


Figure 40. Schematic NNW-SSE cross section (see Figure 41 for location, A-B) showing the relationships between members of the Lexington Limestone and between the Lexington Limestone and Clays Ferry Formation. The Lexington Limestone is basically transgressive through the Grier or Brannon members into the Clays Ferry Formation, but the Tanglewood and Millersburg members represent a localized regression probably related to uplift on local structures. This apparently uplifted area of high-energy Tanglewood calcarenites and calcirudites is called the Tanglewood buildup herein. The entirety of the Tanglewood buildup is effectively shown. No vertical scale intended (adapted from Cressman, 1973).

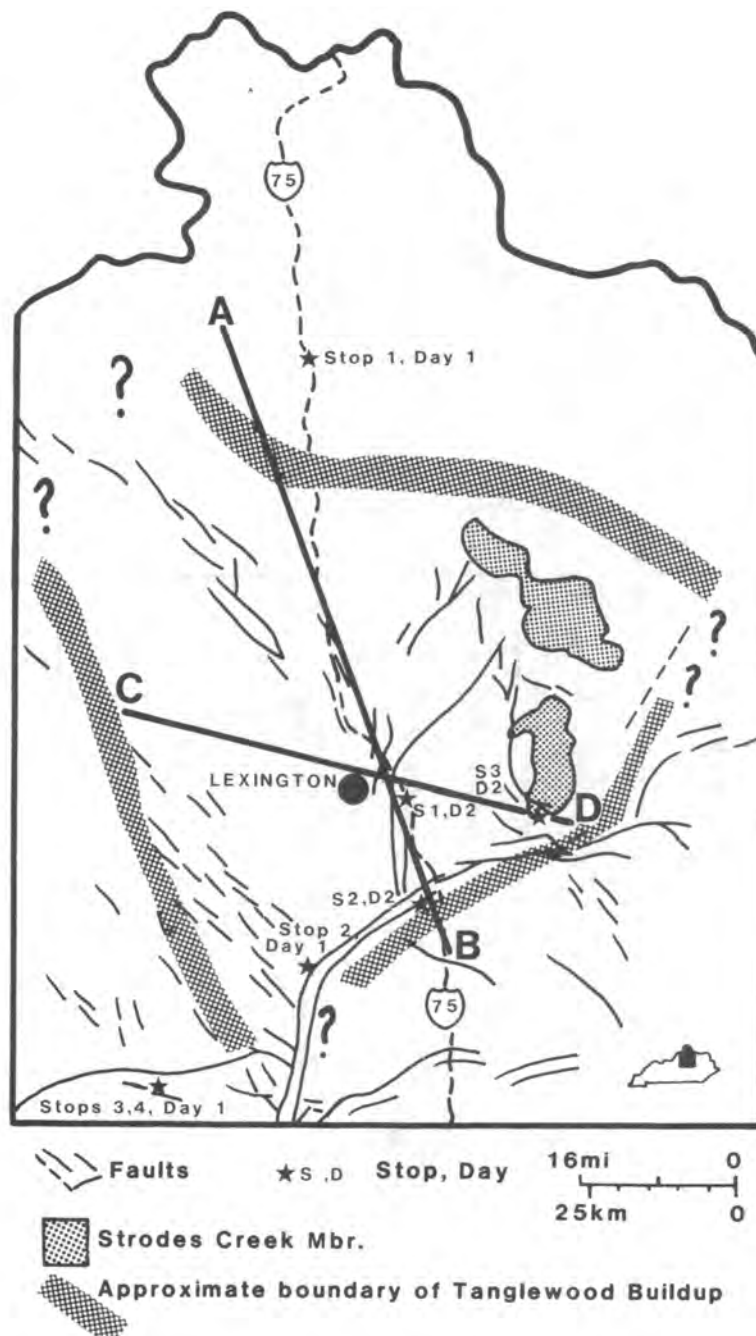


Figure 41. Map of the central Kentucky area showing the location of the two cross sections (Figs. 40, 44), major structures, the Stodes Creek Member, and stops, as well as the approximate outline of the Tanglewood buildup on the Lexington platform (see Fig. 14).

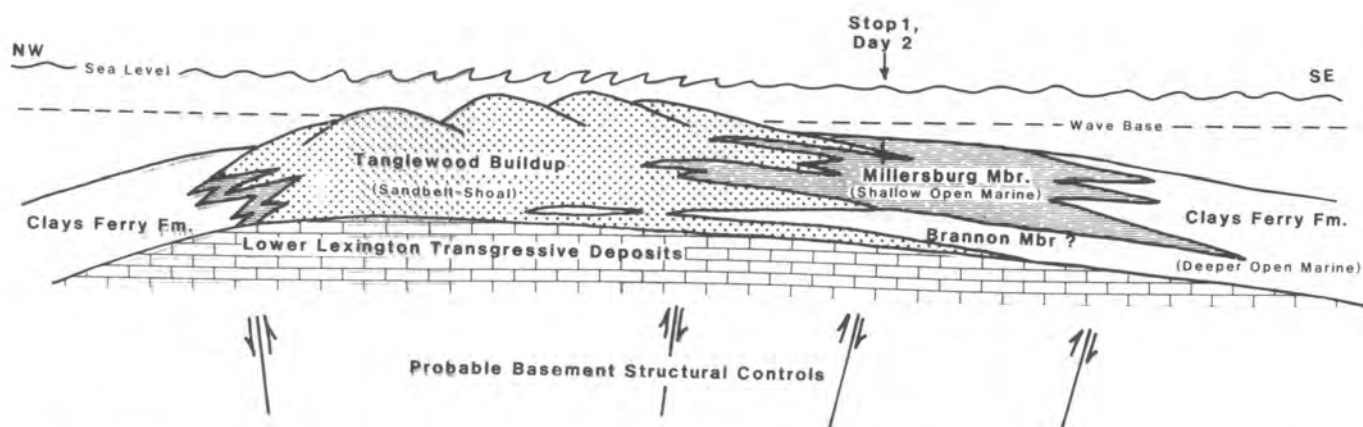


Figure 42. Generalized environmental framework for upper regressive parts of the Lexington Limestone (Tanglewood-Millersburg-Clays Ferry continuum) constituting the Lexington buildup showing hypothesized structural control.

below wave base. The nodular nature of the thin calcarenite beds in the unit probably reflects intense bioturbation. The unit is interpreted to represent shallow open-marine environments just distal to the Tanglewood shoals. Most of thin nodular calcarenite beds in the unit were probably transported from adjacent Tanglewood shoals during storms and were subsequently colonized and bioturbated in the deeper, less energetic environments below wave base (Fig. 42).

The Clays Ferry Formation which is not exposed here, seems to represent the deepest end member of the upper Lexington environmental continuum (Fig. 42). It is composed of approximately 50 percent shale and 50 percent even-bedded, generally fine-grained limestones; many of these limestones show a crude grading. Comminuted and imbricated fossils commonly occur within the limestones, and locally diverse fossil communities are found on tops of individual beds. The unit is interpreted to represent a deeper open-marine environment with most limestones representing distal tempestites. The very generalized environmental framework of these three units is shown in Figure 42.

The relationships between the Tanglewood and Millersburg are clearly shown at this stop. This exposure has apparently cut the margin of a Tanglewood shoal showing individual beds of crossbedded calcarenite tonguing out to the south and southwest into the nodular limestones and shales of the Millersburg Member (Fig. 42). The main Tanglewood body is about five-feet (1.5-m) thick and sharply overlies the Millersburg below. This body represents the upper tongue of the Tanglewood, and is composed of a series of thick amalgamated beds. The approximately ten feet (3 m) of overlying limestones and interbedded shales belong to the Millersburg Member. These beds become thinner, more nodular, and more shaley upward reflecting the progressive drowning of this shoal during transgression. Moreover, this is the transgression that finally inundated the entire Lexington buildup (Fig. 40), because only 1500 ft (457 m) to the southwest and 25-ft (7.6-m) higher, the deeper water Clays Ferry Formation directly overlies this Tanglewood body (MacQuown and Dobrovolny, 1968). This body can be observed pinching out into the Millersburg Member to the south along Interstate 75.

- 4.5 0.3 Pick participants up near the top of the exit ramp on Bryant Road.
- 4.6 0.1 Turn left onto Bryant Road and move into the left lane, crossing the overpass.
- 4.8 0.2 Turn left into the entrance ramp for I-75 southbound.
- 5.2 0.4 Exit onto southbound I-75; milepost 108.
- 5.7 0.5 Overpass.
- 6.1 0.4 Note again the upper tongue of the Tanglewood pinching out southward into the Millersburg Member.
- 6.8 0.7 Overpass
- 8.6 1.8 Millersburg and main body of the Tanglewood.
- 8.9 0.3 Exit 104 to Kentucky State Route 418.
- 9.2 0.3 Overpass and thick section of nodular Millersburg.
- 9.5 0.3 Tanglewood on right near merger of entrance ramp from State Route 418.

- 10.1 0.6 Bounding fault on a small graben.
- 10.2 0.1 Overpass; milepost 103.
- 10.5 0.3 Lower tongue of the Tanglewood overlain by the Brannon Member.
- 12.1 1.6 Overpass.
- 13.2 1.1 Milepost 100; large exposure of the Grier overlain by the lower tongue of the Tanglewood.
- 14.2 1.0 Enter Kentucky River fault zone; north bounding fault of the Elk Lick Creek graben (Black, 1968).
- 14.4 0.2 Steeply dipping beds in the Grier and Curdsville members are present on the upthrown side of the south bounding fault of the Elk Lick Creek Graben (Black, 1968).
- 14.5 0.1 Turn right onto Exit 99 to U.S. Highways 25 and 421.
- 14.7 0.2 Stop sign. Turn right and proceed southeastwardly on U.S. Highways 25 and 421 toward Clays Ferry.
- 15.0 0.3 Having crossed two more faults along which the High Bridge Group is downthrown to the south, the Tyrone Formation appears on the left side of the road (Black, 1968).
- 15.1 0.1 Oregon and Tyrone formations on the left.
- 15.4 0.3 Crossing the major fault of the Kentucky River fault zone in the area. The Lexington Limestone is downdropped to the south in a graben by at least 280 ft (85 m) (Black, 1968).
- 15.5 0.1 Cross the above fault again and enter Camp Nelson Limestone (High Bridge Group); note the typical honeycomb texture.
- 15.7 0.2 Cross the Kentucky River.
- 16.0 0.3 STOP 2. Kentucky River Fault Zone, the Lexington Limestone, and the Clays Ferry Formation near the Southeastern Margin of the Tanglewood Buildup (approximately 0.3 mi southwest of the point where U.S. Highways 25 and 421 cross the Kentucky River, Madison Co.; Carter Coordinates: 800'FSLx960'FEL, 10-Q-62). We will leave the bus at the fault zone and walk up section to meet the bus by the Clays Ferry Formation.

KENTUCKY RIVER FAULT ZONE, THE LEXINGTON LIMESTONE AND THE CLAYS FERRY FORMATION NEAR THE SOUTHEASTERN MARGIN OF THE TANGLEWOOD BUILDUP

FRANK R. ETTENSOHN

This stop begins at the main fault of the Kentucky River fault system in this area. At this particular point, we see the Grier Member of the Lexington Limestone downdropped against the Camp Nelson

Limestone (Fig. 43) with about 360 ft (110 m) of displacement. Although the massive Camp Nelson exhibits some fracturing, most of the deformation along the fault has been taken up by the thin-bedded, shaly Grier, which exhibits minor faulting, drag folds, and brecciation.

The 24 ft (7.3 m) of Camp Nelson exposed here is typical in its dominantly calcilitite composition and the "honeycomb" weathering of its dolomitic burrow fills. However, the adjacent Grier Member of the Lexington Limestone is rather atypical for the Grier. The Grier is typically composed of nodular, argillaceous calcarenites and interbedded shales, and as previously mentioned, represents deposition in shallow open-marine environments below wave base. Here, however, nodular intervals are relatively few, and thin-bedded, regularly to irregularly bedded argillaceous calcarenites with thin shale partings predominate; low-angle cross beds and ripple bedding are locally common. A zone of stromatoporoids is present about five feet (1.5 m) below the top of the unit. Lithologically, the Grier exposed here looks somewhat like the Tanglewood, except that the beds are too thin, argillaceous muds are too prevalent, and the faunas are too diverse.

Examination of Figure 40, however, will show the reason for these disparities. Only the upper 48 ft (14.6 m) of the Grier, which is at least 185 ft (56.4 m) thick in this area (Black, 1968), is exposed at this stop. Figure 40 shows that this part of the Grier is approximately equivalent in time and spatially close to the deposition of the lower tongue of the Tanglewood, which apparently marks the initiation of uplift on the Tanglewood buildup in the midst of regional transgression. The fact that typical Tanglewood lithologies were not deposited here, suggests that this area did not experience the same degree of uplift as did areas to the north. Differences between this and the typical Grier, however, still reflect the relative closeness of this area to the uplifted shoal area just to the north (e.g., Miles 10.5 and 13.2 on Day 2). Moreover, the fact that this facies change between the Grier and Tanglewood occurs so close to the Kentucky River fault zone (Fig. 41; see Black, 1968) suggests that in this area at least, the Tanglewood buildup was bound by and probably related to uplift on the fault zone. As a result, concurrent Grier deposition in this area apparently occurred on the downdropped side of the fault zone and was close to, but still below, wave base.

The Grier is abruptly overlain by the Brannon Member (Fig. 43), consisting of 33 ft (10 m) of interbedded calcisiltites and shale. Except for the imbricated brachiopod valves that are present at the bases of some of the crudely graded limestones, fossils are uncommon; the only common fauna include pelecypods, lingulids, and scolecodonts. Bioturbation is locally intense. This unit looks very much like the lower Clays Ferry and is interpreted to represent deeper open-marine, muddy-bottom environments well below wave base. Many of the limestone beds probably represent distal tempestites. The Brannon obviously represents a brief period of deepening that locally interrupted initial phases of uplift on the Tanglewood buildup. The Brannon, however, is relatively local in extent, present only in southwestern parts of the Tanglewood buildup (Cressman, 1973, fig. 29), and its distribution may also reflect possible structural control.

Not only was this deepening event local in nature, but on the basis of biostratigraphy and relative thickness it must have also been short in duration. Moreover, the Brannon is abruptly overlain by the main body of the Tanglewood Member (Figs. 40, 43). A substantial and rapid period of uplift must have been required to effect the change from the deeper open-marine Brannon environments to the high-energy Tanglewood shoal environments.

The base of the Tanglewood is erosional, and the lower seven feet (2.1 m) are composed of thick- to medium-bedded calcarenite and calcirudite. Whole fossils are rare, and low-angle planar crossbedding, herring-bone crossbedding, scours, ripple bedding, and rip-up clasts are present in the unit. A pyritic shale separates the lower part of the Tanglewood from the thicker upper part. The upper part is 24-ft (7.3-m) thick and consists of thin, irregularly bedded calcarenites and calcirudites with thin shale partings. The unit is more fossiliferous, although crossbedding, ripple bedding, scours, and channels are still present.

The pyritic shale may reflect a period of stillstand and an accompanying environmental change to slightly deeper conditions in later stages of Tanglewood deposition. Nonetheless, the transition from the Tanglewood to the overlying Clays Ferry Formation is still lithologically and environmentally abrupt. The nodular Millersburg Member, which typically represents the shallow open-marine transition between Tanglewood shoals and deeper open-marine Clays Ferry environments (Figs. 40, 42) is absent here (Fig. 43) and in fact, is relatively thin and local in nature south of the Kentucky River fault zone (e.g., Black, 1968). The general absence or thinness of the Millersburg south of the fault zone, as well as the facts that the Tanglewood is so thin here and is represented by only the main body, may again reflect the fact that this locale was on the downdropped side of the main Kentucky River fault zone and removal from the locus of major uplift and buildup formation to the north (Fig. 41). Hence, stratigraphic relationships involving the Grier, Tanglewood and Clays Ferry just south of the Kentucky River fault zone suggest that this area was probably just beyond the margin of the buildup, and that in this area, the position of that margin was probably controlled by the nearby fault zone.

This roadcut is the type section of the Clays Ferry Formation (Figs. 40, 43), which is approximately 180 ft (55 m) thick here. We will be examining only the basal 50 ft (15 m) at this stop.

The lower 35 ft (10.7 m) are very similar to the Brannon Member of the Lexington Limestone, and probably reflect the same deeper open-marine environmental conditions. In the overlying 145 ft (44.3 m), however, the limestones are thicker, more abundant, and more fossiliferous, possibly suggesting slightly more shallow and proximal environmental conditions. At least three horizons of ball-and-pillow structures are present in the Clays Ferry Formation, and one can be found in the Brannon Member (Fig. 43). These horizons could represent episodes of slope instability or periods of seismicity on nearby fault zones.

- 16.3 0.3 Cross a small fault displacing the Tanglewood and Brannon members about 10 ft (3.0 m).
- 16.5 0.2 Participants reboard the bus and proceed southward on U.S. 25 and 421.
- 16.6 0.1 Hairpin curve in road; subdivision entrance on right.
- 16.8 0.2 View to the left of the Lexington "peneplain;" poorly exposed Clays Ferry on the right. Although the Lexington peneplain is an old concept (e.g., Jilison, 1928; Fenneman, 1938), newer interpretations suggest that solutional lowering acting on a progressively unroofed limestone dome could produce a relatively level central plateau surface surrounded by retreating

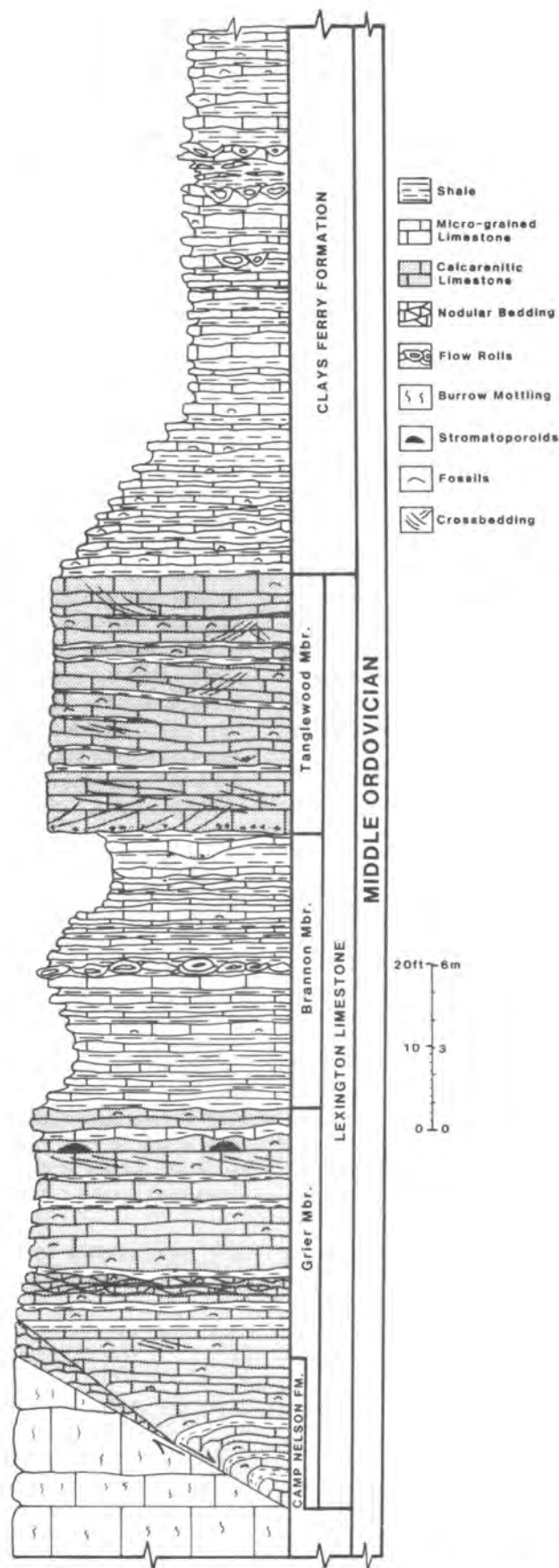


Figure 43. Schematic drawing of the Clays Ferry section showing relationships between the Camp Nelson, Lexington Limestone, and Clays Ferry formations. The Grier Member is atypical here in the decreased amount of nodular bedding (see Figs. 40, 41). Not all of the Clays Ferry Formation is shown.

- cuestas much like the Inner Bluegrass (e.g., see Pitty, 1969; Rice, 1977).
- 17.0 0.2 Approximate position of the contact between the Clays Ferry and Garrard Siltstone formations on the right; the Garrard Siltstone is the only wholly clastic unit in the Ordovician of central Kentucky.
 - 17.2 0.2 Poorly exposed Garrard on the left.
 - 17.4 0.2 Turn left at stop sign onto U.S. Highways 25 and 421.
 - 17.6 0.2 Turn right into entrance ramp onto southbound I-75.
 - 17.8 0.2 Enter onto southbound I-75; Calloway Creek Limestone on the entrance ramp. The Maysvillian Calloway Creek Limestone is equivalent to the Fairview Formation in northern Kentucky. The Calloway Creek consists of thin-bedded, fossiliferous calcarenites and interbedded shales. Each calcarenite typically has an erosional base and grades upward into a laminated to ripple-laminated calcareous siltstone, and from there, into shales. Most of the limestone beds are probably tempestites. For the next 1.5 mi, I-75 will cross the Calloway Creek Limestone.
 - 19.3 1.5 Begin the greenish-gray argillaceous limestones of the lower Ashlock Formation.
 - 19.7 0.4 Merge right onto Exit 95 ramp; laminated argillaceous limestones of the Tate Member of the Ashlock on the right. Proceed up the ramp.
 - 19.9 0.2 Stop sign; turn left towards Boonesboro onto Kentucky State Route 627.
 - 20.6 0.7 Leave the Ashlock Formation and move onto the Calloway Creek Limestone.
 - 23.5 2.9 Boonesboro Trading Post on right.
 - 24.7 1.2 Garrard Siltstone on the left.
 - 24.8 0.1 Enter Kentucky River fault zone; southwestern bounding fault on a graben; the Calloway Creek is downdropped about 100 ft (30.5 m) against the Garrard Siltstone (Black, 1968).
 - 24.9 0.1 Calloway Creek Limestone and entrance to Boonesboro State Park.
 - 25.1 0.2 Garrard and Clays Ferry formations on the right.
 - 25.2 0.1 Northeastern bounding fault of the above mentioned graben; Tanglewood Member and the Clays Ferry Formation.
 - 25.3 0.1 Clays Ferry overlying the Tanglewood on the left.
 - 25.4 0.1 Tanglewood faulted against Tanglewood; entrance to Boonesboro State Park on the right.
 - 25.7 0.3 Entrance to quarry on the right.
 - 25.8 0.1 Begin to cross bridge; Tanglewood faulted against Tanglewood at the entrance to the quarry on the left. Crossing the bridge, one can see drift mines in the Camp Nelson Formation on the upthrown, north side of the Kentucky River fault zone.
 - 26.0 0.2 Northeast end of bridge is approximately on the main fault of the Kentucky River fault zone, along which the upper Lexington Limestone and Clays Ferry formations are downdropped to the south against the High Bridge Group; maximum displacement is in excess of 500 ft (152 m) (Black, 1968). This fault is exposed in the cut on the right as are the contacts between the Camp Nelson, Oregon and Tyrone formations. The prominent reentrant in the outcrop is the Pencil Cave K-bentonite, which we observed yesterday south of this area at the Camp Nelson.
 - 26.3 0.3 The end of the exposure is near the top of the Tyrone Formation.
 - 26.8 0.5 Optional Stop—Grier Member and the Vertical Grier-Tanglewood Transition (F.R. Ettensohn)

This is a typical exposure of the Grier Member of the Lexington Limestone; it exhibits nodular, argillaceous calcarenites and thin shale partings with a few through-going, regular beds of coarser calcarenite. At least 85 ft (26 m) of Grier are present, but the uppermost five feet (1.5 m) become coarser grained, lighter in color, and more tabular, indicating a transition into the lower tongue of the Tanglewood at the top (Fig. 44, E). This transition marks the beginning of the Tanglewood buildup mentioned before.
 - 27.2 0.4 Optional Stop—Lateral Grier-Tanglewood Transition and Implications (F.R. Ettensohn)

This exposure shows approximately 16 ft (4.9 m) of the lateral transition between nodular Grier-type lithofacies and the more tabular massive calcarenites of the Tanglewood lithofacies (Fig. 44, F). Presumably, exposures like this reflect the transportation of skeletal sands from nearby Tanglewood shoals into adjacent deeper water Grier environments below wave base. Such facies relationships suggest that the uplift which generated the Tanglewood buildup was not uniform across its entire extent, probably reflecting different degrees of movement on the many involved faults.

The upper two feet (0.6 m) of the exposure exhibit nodular fossiliferous calcisiltites and calcarenites in the Brannon Member. Ball-and-pillow structures are common throughout.
 - 27.9 0.7 Main body of the Tanglewood overlying the Brannon.
 - 28.1 0.2 Begin a long cut in the main body of the Tanglewood; several small growth faults are present.
 - 28.6 0.5 Intertonguing between the Tanglewood and Millersburg members on the right (Fig. 44, G). On the left side of the road, a dolostone halo surrounds a small fault in the

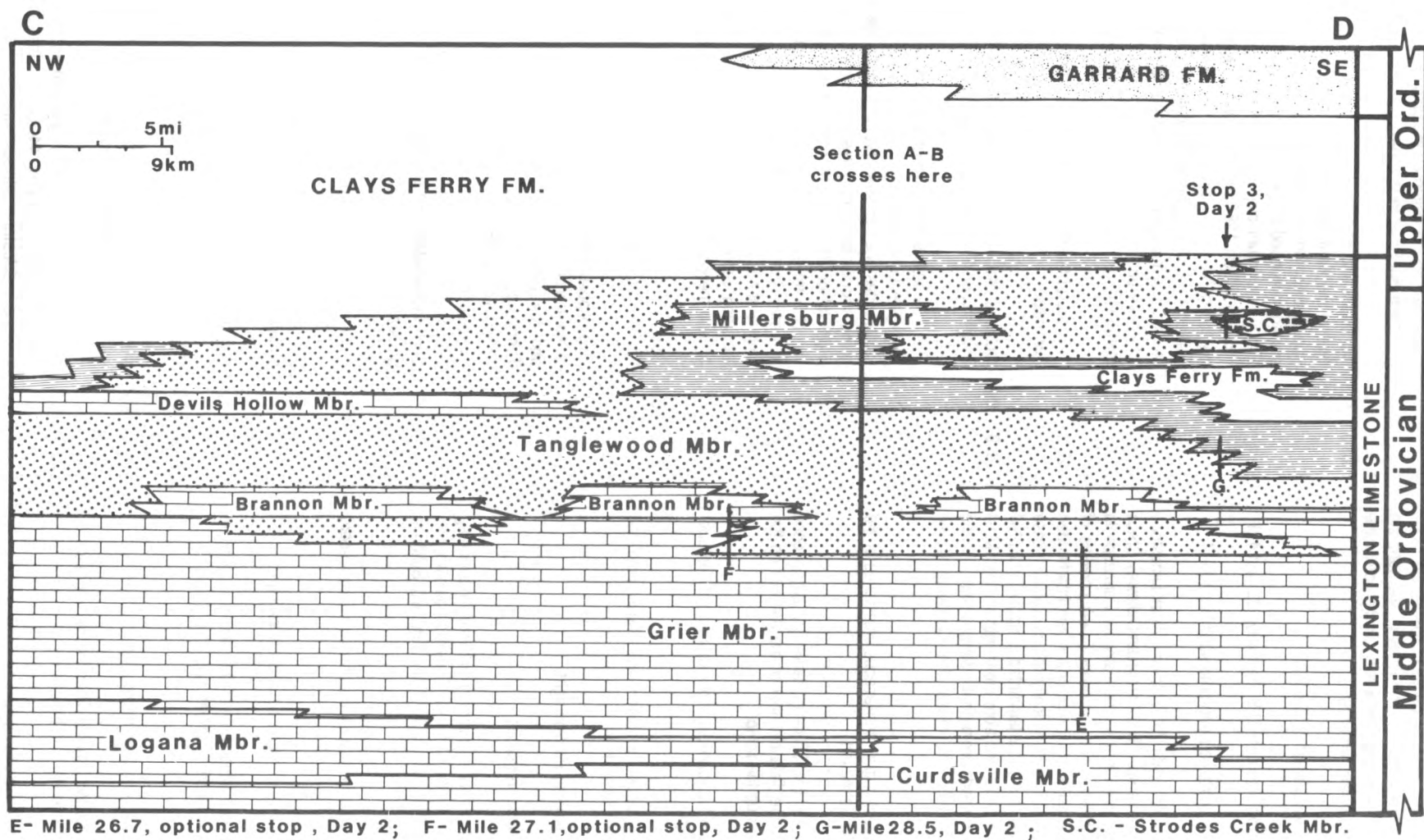


Figure 44. Schematic west-southwest-east southeast section (see Figure 41 for location, C-D) through the Lexington Limestone and Clays Ferry Formation showing nearly the entirety of the Tanglewood buildup. The boundary is abrupt in the east near the Kentucky River fault zone but is more gradational to the northwest as the Tanglewood gradually declines in thickness. No vertical scale intended (adapted from Cressman, 1973).

Millersburg. The uppermost beds in the exposure are not dolomitized suggesting that solutions rose from below along the fault. The sag in overlying beds probably reflects volume decrease below as a result of limestone dolomitization (Black and Haney, 1975).

- 30.5 1.9 Nodular limestones and shales in the Millersburg Member.
- 30.9 0.4 Thin tongues of the Tanglewood lithology in the Millersburg Member.
- 31.7 0.8 At the traffic light, turn left onto Kentucky State Route 627 (truck route).
- 32.5 0.8 Junction with Kentucky State Route 1927.
- 34.2 1.7 Turn right onto U.S. 60 and proceed eastward into Winchester.
- 35.6 1.4 Cross old railroad grade.
- 36.1 0.5 Turn left onto Kentucky State Route 627 and proceed toward I-64.
- 36.4 0.3 Turn right and park at the Transfer Station. STOP 3. The Strodes Creek Member of the Lexington Limestone: An Example of Probable Synsedimentary Structural Influence, Winchester, KY. (Roadcut at the L&N underpass on Kentucky State Route 627 in Winchester, 1.0 mi south of its junction with I-64, Clark Co.; Carter Coordinates: 450°FN Lx980°FWL, 1-R-64).

THE STRODES CREEK MEMBER OF THE LEXINGTON LIMESTONE: AN EXAMPLE OF PROBABLE SYNSEDIMENTARY STRUCTURAL INFLUENCE

FRANK R. ETTENSOHN

The Strodes Creek Member (Fig. 45) is one of the most distinctive and unusual members of the Lexington Limestone. Lithologically, the member is composed largely of dense, calcareous, brownish-gray, ostracodal mudstones with pinch-and-swallow bedding (Black and Cuppels, 1973), which contrasts markedly with the nodular or even-bedded calcarenites and interbedded shales of the enclosing Millersburg and Tanglewood members (Figs. 44, 45). The predominant fauna and flora of ostracods, stromatoporoids, and various types of red and green algae are also unusual for the Lexington, although other more typical Lexington faunas are also present. Moreover, the member is very localized, occurring in two relatively thin lensoid bodies roughly outlined by faults on their southern and western margins (Fig. 41). The nature of their eastern margins is uncertain because the unit passes below the updip limit of exposure. Nonetheless, stratigraphic thinning of the bodies toward the east suggests that they do not persist far in that direction.

The typical Strodes Creek section begins with a dark fissile shale containing ostracods and *Sphenophallus* tubes. This shale typically overlies the Millersburg sharply and is unique in the Lexington Limestone. At this exposure, the shale overlies and intertongues with a red-algae (*Solenopora*) bioherm (Fig. 45) with a maximum relief of about two feet (0.6 m). The bioherm appears to occupy a shallow,

northeast-southwest-trending channel and exhibits subtle accretion surfaces dipping to the north with thin debris lenses both to the front and rear of the apex of the bioherm. The bioherm most likely should be included with the Millersburg Member, because the contrast between the coralline red algae, which are high-energy indicators (e.g., Wray, 1977), and the overlying deeper water dark shales in the basal Strodes Creek is stark.

Approximately 12 ft (3.6 m) of thin- to medium-bedded calcilutites, calcisiltites (mudstones), and fine-grained calcarenites (packstones) overlie these shales (Fig. 45). The limestone layers typically have sharp, erosional bases and grade upward into overlying shales. Some of these limestones have distinct channel-like geometries in sectional view. The calcilutites and calcisiltites typically exhibit in-place stromatoporoids and a great diversity of algae; they probably accumulated more or less in place. The calcarenites, on the other hand, may be crossbedded or subtly graded, and contain severely comminuted and overturned stromatoporoids which suggest reworking and transportation.

The uppermost part of Strodes Creek commonly consists of lighter colored, coarser grained calcarenites or calcirudites, generally with rip-up clasts from underlying lithologies. In some places, this part of the Strodes Creek may be a thick oncolitic calcirudite, although at this exposure, it is represented by a foot-thick Tanglewood-like calcarenite capped with a pyritic hardground.

Compared to the surrounding members of the Lexington Limestone, the Strodes Creek in general represents considerably deeper waters in two very localized areas. Moreover, the nature of this change in depth is represented in this exposure by the abrupt change from a relatively shallow-water, high-energy, coralline-algae biostrome to fissile, deeper water, dark shales. Although these shales are dark, the presence of a sparse fauna suggests dysaerobic, and not anaerobic conditions, so that the dark shales probably accumulated within the pycnocline. Based on modern interpretations of optimum depths for red algae (e.g., Heckel, 1972) and on minimum pycnocline depths (Byers, 1977), the change in lithologies observed here could represent abrupt changes in depth from less than 70 ft (21.3 m) for the red algae to depths in excess of 165 ft (50 m) for the black shales. The abrupt nature of this change along with the fact that the Strodes Creek occurs only in local bodies partially delineated by faults (Fig. 41) suggests that structural subsidence along these faults was responsible for creating these two localized basins. At their deepest, the dark organic-rich muds apparently accumulated in the oxygen-deficient basins. Once the basins filled to the top of the pycnocline, deeper water carbonate muds with a restricted fauna and flora accumulated in the basins. Periodically, these muds must have been winnowed, reworked, and inundated by sands from outside the basins during storms, but the basinal nature of these deposits is still indicated by the dark coloration of the sediments and their muddy nature. This type of sedimentation apparently continued in the basins until they were filled to a level just below or at wave base. At this time, the coarser grained, lighter colored calcarenitic sands virtually inundated remaining parts of the basins and ended endemic basin sedimentation.

The importance of this unit is that it is representative of a type of structural control on sedimentation and biofacies development that is apparent in other parts of the Lexington Limestone. Moreover, it suggests that various structures in the area responded differently and at different rates to probable bulge reactivation. Although I have tried to suggest this type of control for the entire Tanglewood buildup

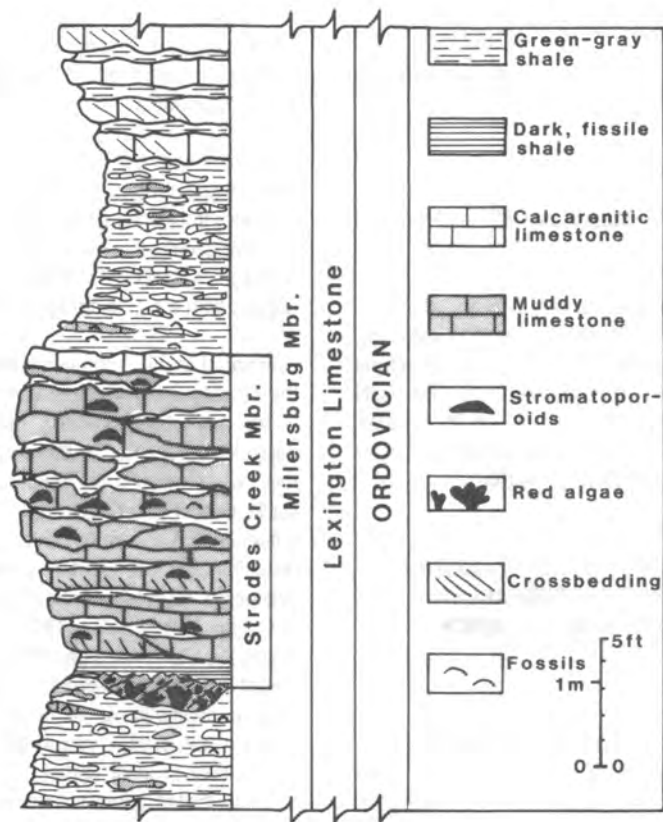


Figure 45. Schematic drawing of the reference section for the Strodes Creek Member of the Lexington Limestone just north of Winchester, Kentucky.

(Figs. 41, 42), of which the Strodes Creek is an atypical part, nowhere does the possibility of fault control seem to be as convincing as on the western margin of the unit (Fig. 41).

- 36.7 0.3 Cross railroad tracks.
- 37.4 0.7 Turn right onto entrance ramp toward eastbound I-64.
- 37.7 0.3 Enter onto eastbound I-64. The geology of this route along I-64 has been described before, although in reverse, by Weir and Philley (1970), Dever and others (1977), Ettensohn and Dever (1979) and Ettensohn (1981).
- 38.4 0.7 Eastern boundary fault of a small graben; Clays Ferry and Millersburg dip west into the graben. This is a later surface expression of one of the faults which may have controlled Strodes Creek deposition.
- 38.5 0.1 Overpass.
- 38.6 0.1 Contact between the Millersburg Member of the Lexington Limestone and the Clays Ferry Formation on the left.
- 38.7 0.1 Exit 98 to the Mountain Parkway.
- 38.9 0.2 Overpass.
- 39.2 0.3 Milepost 98; entrance to rest area.
- 39.4 0.2 Rest area; possible lunch stop.
- 39.9 0.5 Low exposure of the Clays Ferry Formation on both sides of the road. We have now moved onto the low hummocky topography of the Eden Belt of the "Outer" Bluegrass (Fig. 3).
- 40.3 0.4 Overpass.
- 40.6 0.3 Clays Ferry/Kope formations on both sides of road.
- 41.4 0.8 Millersburg Member and Clays Ferry present before crossing the valley.
- 41.7 0.3 Tanglewood intertonguing with the Millersburg and overlain by the Clays Ferry.
- 42.0 0.3 Milepost 101.
- 42.2 0.2 Millersburg and Clays Ferry.
- 42.6 0.4 Exit 101 to U.S. 60.
- 43.6 1.0 Bridge over Stoner Creek.
- 43.9 0.3 Clays Ferry/Kope formations.
- 44.0 0.1 Milepost 103.
- 45.0 1.0 Milepost 104 and overpass; Garrard Siltstone and overlying Calloway Creek Limestone exposed just beyond.

- 45.3 0.3 Montgomery County line.
- 46.0 0.7 Milepost 105.
- 46.5 0.5 Bridge over Grassy Creek.
- 47.0 0.5 Calloway Creek Limestone; for the next 4.1 mi the field-trip route largely crosses the Calloway Creek Limestone with Garrard and Clays Ferry formations present in some of the deeper valleys (Blade, 1976; Weir, 1976b).
- 47.8 0.8 Calloway Creek Limestone on both sides.
- 48.0 0.2 Milepost 107.
- 48.2 0.2 Overpass; Calloway Creek Limestone.
- 48.8 0.6 Rest area on left.
- 49.0 0.2 Milepost 108; Calloway Creek Limestone.
- 49.3 0.3 Clays Ferry Formation in the valley.
- 50.3 1.0 Exit 110 to U.S. 460 and Kentucky State Route 11.
- 50.6 0.3 Overpass; nodular upper part of the Calloway Creek Limestone.
- 51.1 0.5 Contact of the dolomitic mudstones of the Tate Member of the Ashlock Formation with the Calloway Creek Limestone. For the next 2.1 mi, the route primarily crosses the Tate Member of the Ashlock with the Calloway Creek Limestone in some of the deeper valleys and the Grant Lake Member of the Ashlock on some of the higher ridge tops (Weir, 1976b).
- 51.5 0.4 Cross Hinkston Creek.
- 52.0 0.5 Overpass; limestones in the Back Bed of the Tate Member.
- 52.7 0.7 Upper nodular part of the Calloway Creek Limestone.
- 53.2 0.5 Exit 113 to U.S. 60; for the next 2.2 mi the highway primarily crosses the nodular limestones and shales of the Grant Lake Member of the Ashlock Formation (Weir, 1976b; Weir and McDowell, 1976).
- 53.5 0.3 Overpass; Grant Lake Member.
- 55.0 1.5 Milepost 114.
- 55.4 0.4 Nodular limestones of the basal Bull Fork Formation.
- 55.9 0.5 More even-bedded limestones and interbedded shales of the upper Bull Fork Formation.
- 56.0 0.1 Milepost 115.
- 56.1 0.1 Cross Stepstone Creek.
- 56.5 0.4 Bath County line.

- 56.6 0.1 Interbedded dolostones and mudstones of the Upper Ordovician Drakes Formation overlain by the dolostones of the Silurian Brassfield Formation.
- 57.0 0.4 Milepost 116 at the top of the hill; Ordovician-Silurian contact (disconformity) and typical Outer Bluegrass topography (Fig. 3).
- 57.1 0.1 Drakes Formation.
- 57.5 0.4 Overpass; Drakes Formation.
- 57.8 0.3 Bull Fork-Drakes contact.
- 58.0 0.2 Milepost 117.
- 58.1 0.1 Fault on the right, along which the Bull Fork and Drakes formations are downdropped to the north; displacement is about 35 ft (10.7 m) (Weir and McDowell, 1976).
- 58.6 0.5 Contact of the Ashlock and Bull Fork formations.
- 58.8 0.2 Bridge over Salt Well Creek.
- 59.3 0.5 Bridge over Slate Creek.
- 59.4 0.1 Bull Fork Formation.
- 60.0 0.6 Contact of the Drakes and Brassfield formations.
- 60.5 0.5 Drakes Formation at road level; Drakes-Brassfield contact higher in roadcrop.
- 60.9 0.4 Bridge over Slate Creek.
- 61.2 0.3 Bull Fork Formation.
- 61.6 0.4 Drakes-Brassfield contact on top of hill; greenish Lower Silurian Crab Orchard Shale apparent near top of the hill on the right.
- 61.8 0.2 Exit 121 to Kentucky State Route 36; Drakes Formation.
- 62.7 0.9 Bridge over Slate Creek.
- 63.0 0.3 Bull Fork Formation; zone of overturned stromatoporoids and corals at about eye level.
- 63.3 0.3 Drakes Formation.
- 63.5 0.2 Exit ramp to U.S. 60; a zone of overturned stromatoporoids and corals forms a prominent ledge in the Bull Fork Formation; Bull Fork-Drakes contact.
- 64.6 1.1 Greenish-gray shales of the Lower Silurian Crab Orchard Formation crop out sporadically along the highway for the next 6.5 mi; most of the exposures are slumped, overgrown, or covered with fluvial terrace deposits from the Licking River (McDowell, 1976).
- 69.7 5.1 Licking River and the Rowan County line; good view of the Knobs, which in eastern Kentucky mark the

beginning of the Pottsville Escarpment and the Appalachian Plateau physiographic province (Fig. 3).

- 71.1 1.4 Overpass; Crab Orchard-Ohio Shale contact; enter Appalachian Plateau physiographic province.

- 71.2 0.1 STOP 4A. Silurian-Devonian Contact and Transgressive Black Shales at the Base of the Ohio Shale (On the eastbound lane of I-64, 1.5 mi east of the Bath-Rowan county line, Rowan Co.; Carter Coordinates: 900°FNLx1150°FWL, 1-T-72).

SILURIAN-DEVONIAN CONTACT AND TRANSGRESSIVE BLACK SHALES AT THE BASE OF THE OHIO SHALE

FRANK R. ETTENSOHN

The section begins with at least eight feet (2.4 m) of greenish-gray Lower to Middle Silurian Crab Orchard Shale with thin dolostone stringers. The upper part of the exposed shale is weathered and oxidized, probably reflecting some combination of Devonian, pre-black-shale weathering and erosion and relatively recent interstratal weathering (e.g., Ettensohn and Bayan, 1990). Other Middle Silurian and Middle Devonian formations, which are typically present in the area, are absent here (see McDowell, 1975, 1976).

The contact between Silurian greenish-gray shales and overlying Devonian gray shales is effectively a paraconformity. The basal Devonian is marked by a very thin, pyritic lag zone or bone bed (Fig. 46). Even though it is less than an inch thick, it represents parts of six Frasnian and early Famennian conodont zones (Fig. 26) (Ettensohn and others, 1989a) and represents approximately ten million years (see Harland and others, 1990).

The overlying two-to-three feet (0.9-1.1 m) of gray shales has been interpreted to represent the thinning western edge of the upper Olenangy Shale (see Stop 6, Day 3) which is equivalent to the Hanover Shale of New York based on correlation of gamma-ray logs throughout the Appalachian basin (Wallace and others, 1977). The overlying 31 ft (9.5 m) of black, fissile, "paper" shale with some interbedded gray shales near the base extend to the first bench and represent the lower part of the Huron Shale Member of the Ohio Shale (Fig. 46), which is similarly correlated with the Dunkirk Shale of New York (Wallace and others, 1977).

The base of the black shale in this area represents a major sequence boundary in every sense of the word. It is the base of Sloss's (1963) Kaskaskia cratonic sequence, and it represents the base of a transgressive systems tract in the parlance of sequence stratigraphy. Moreover, the unconformity at the base represents a time-rich transgressive surface.

This surface, however, is far more complex than a simple transgressive surface for Late Devonian black shales. Based on the presence of unconformity-bound Oriskany, Onondaga, and Hamiltonian sequences in eastern Kentucky (Fig. 23) (Ettensohn and others, 1988b) and unconformity-bound Onondaga and Hamilton remnants in south-central Kentucky (Lenhart, 1985), it is likely that most of eastern and central Kentucky experienced the episodes of uplift and erosion responsible for these unconformities. In fact, these periods of uplift and erosion, each corresponding to the inception of

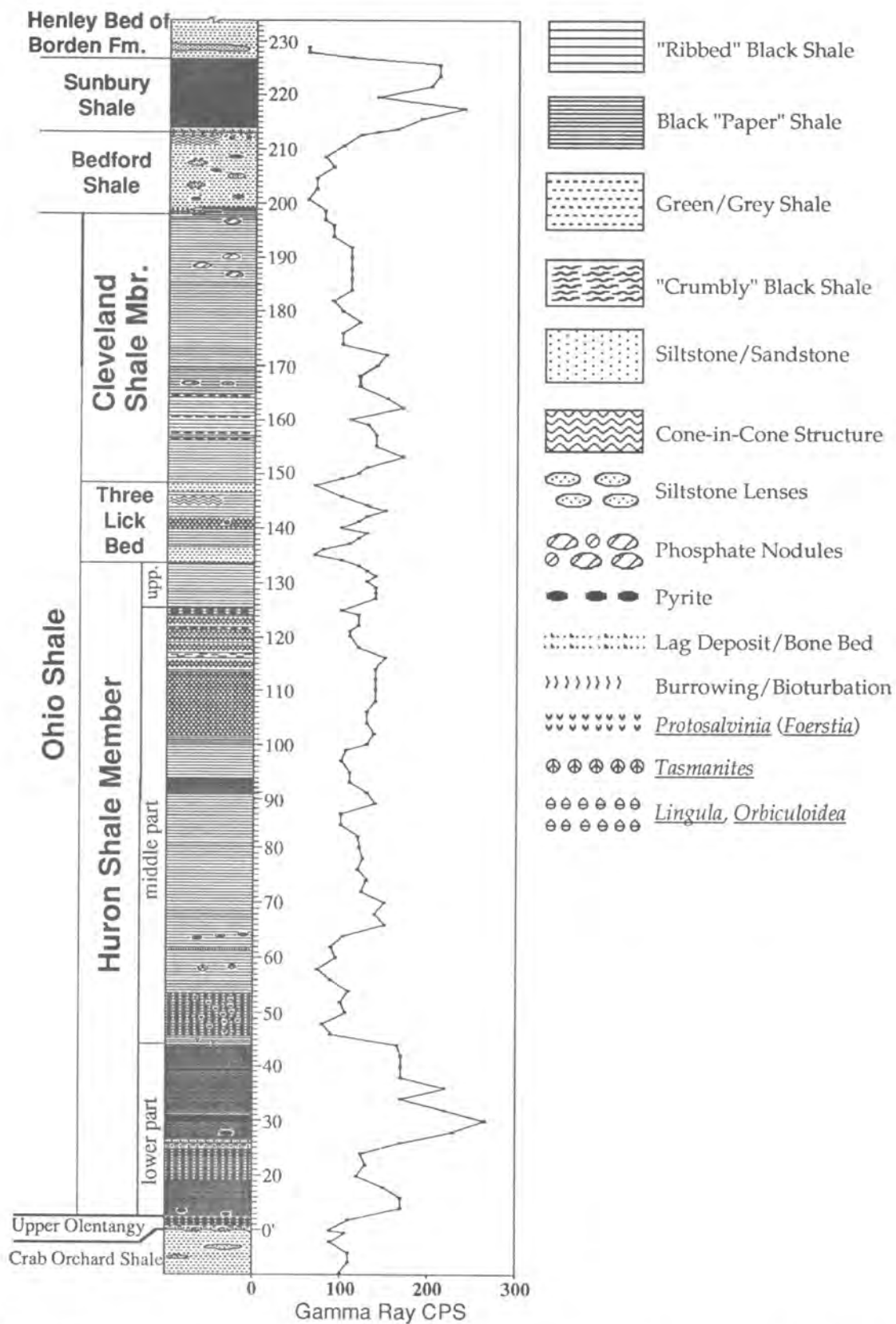


Figure 46. Schematic composite section for black-shale portions of Stops 4 and 5 with an accompanying scintillometer profile (artificial gamma-ray log). Stop 4A deals with the basal contact and transgressive black shales in the lower Huron. Stop 4B examines the *Protosalvinia* zone and associated regressive black shales.

an Acadian tectophase (Ettensohn, 1985a), probably account for the majority of the erosion indicated below the black shales here.

I-75, 1.7 mi east of the Bath-Rowan county line, Rowan Co.; Carter Coordinates: 230°FNLx1000°FWL, 1-T-72).

REGRESSIVE BLACK SHALES AND THE *PROTOSALVINIA* (*FOERSTIA*) ZONE

FRANK R. ETENSOHN

The last period of erosion must have occurred in the late Givetian during the inception of the third Acadian tectophase (Fig. 22), for Givetian conodonts are completely absent here (Fig. 26) (Ettensohn and others, 1989a). However, the presence of early Frasnian conodonts in the thin lag zones (Fig. 26) suggests that soon afterwards, seas began inundating the area. As already mentioned, those seas must have existed in this area for the duration of nearly six conodont zones, or about 10 million years, with little or no deposition. During this time, most of the sediments derived from the rising Acadian mountains were apparently sequestered in the Appalachian foreland basin, which served as a clastic sink (Fig. 27). These sediments are represented by three transgressive-regressive, black shale-gray shale sequences (Burket-Hanover-Angola shales, Fig. 27), each of which migrated farther cratonward in time in response to cratonward-migrating Acadian tectonism. However, by upper Olentangy-lower Huron time (early Famennian) the foreland basin was apparently filled to overflowing and tectonism had migrated far enough cratonward that the Cincinnati arch area subsided allowing the Illinois and Appalachian basins to yoke (Fig. 27). The upper Olentangy and lower Huron and their equivalents were the first basinal units to cross the arch (Figs. 26, 27).

The lower Huron is an excellent example of a transgressive black shale. Shales of this type are typically deposited at times of regional transgression and subsidence, accompanying major phases of tectonism. Foreland-basin sinks and aggradation of clastic sediments in nearby shore areas apparently left more distal, cratonic areas like this starved of detrital constituents, but rich in organic matter (Ettensohn and others, 1988b). The resulting shales are typically more widespread, more organic-rich, and more radioactive than adjacent regressive black shales like the middle part of the Huron Shale Member (Fig. 46). Because of their high organic content (8-23% by weight), they are typically pulpy and weather to form very fissile, "paper shales," like those exposed here in the lower part of the Huron.

As already mentioned earlier, the upper Olentangy and lower Huron shales have been traced in the subsurface into the supposedly equivalent Hanover and Dunkirk shales, respectively, in the New York type section (Wallace and others, 1977). However, great care must be taken with the use of the words "equivalent" or "correlative." Although these Kentucky units certainly can be traced into their respective, homotaxial units of the same lithology in the New York section, the respective New York and Kentucky units are not necessarily contemporaneous. In fact, conodont studies suggest that Kentucky units are two to three conodont zones younger than their homotaxial equivalents in the New York section (Ettensohn and others, 1989a). Based on the chronology of Harlan and others (1990), the Kentucky units are three to five million years younger than their respective homotaxial equivalents in New York. Inasmuch as these black shales reflect foreland-basin formation (e.g., Ettensohn, 1985a, 1985b; Ettensohn, 1991b), and their diachronous, but homotaxial, nature across broad areas reflects basin migration, these contrasting ages provide a means of estimating rates of foreland-basin and related bulge migration. In this particular case, estimated rates of basin-and-bulge migration are on the order of 37 to 62 km per million years (22.9-38.4/mi per million years) or 0.4-0.6 m/yr (1.3-2.0 ft/yr).

At the eastern end of this roadcrop we will climb up over the transgressive "paper shales" of the lower Huron Member into the "ribbed" regressive black shales of the middle part of the Huron Member, which begin at about the top of the first bench. The middle part of the Huron in this exposure is about 82-ft (25-m) thick, and it consists largely of black shales, but horizons of interbedded gray and black shales are present near the top and bottom of the unit (Fig. 46). Although superficially similar to the transgressive, lower Huron black shales, the middle Huron black shales are substantially different. They contain more clastic constituents, particularly in the form of quartz-rich laminae, and less organic carbon (4-7%; Bland and others, 1981). Consequently, these shales are generally less radioactive than transgressive shales like the lower Huron (Fig. 46), and such variations in radioactivity can be correlated along the outcrop and between the outcrop and subsurface using scintillometer-generated radioactivity profiles, as in Figure 46, and gamma-ray logs (Ettensohn and others, 1979).

On close inspection, the regressive shales are easily distinguished from the transgressive shales by the presence of "ribbing." Each "rib" consists of a few inches of shale which is more resistant, and hence more protuberant, than adjacent shales. Ribs probably contain more clastic-rich laminae than adjacent shale, but this has not yet been demonstrated. Moreover, the origin of the cyclicity inherent in this ribbing is also uncertain.

As a regressive black shale, the middle Huron should be traceable into gray shales and coarser clastics to the east (Fig. 27), and basal cross sections show that this is, in fact, the case (Wallace and others, 1977; Kepferle and others, 1978). In the New York section, the middle Huron is approximately homotaxial with the gray shales and siltstones shales of the Gowanda-through-Dexterville interval. However, based on the position of the *Protosalvinia* zone and conodont biostratigraphy (Ettensohn and others, 1989a), the middle Huron is two to three conodont zones younger than lithofacies equivalents in the New York section.

The basal ten feet (3.0 m) of the middle Huron at this exposure contain interbedded gray and black shales which coincide with the *Protosalvinia* (*Foerstia*) zone (Fig. 46). *Protosalvinia* is an enigmatic plant fossil that has been interpreted to represent a possible brown alga (Phillips and others, 1972; Schopf and Schwietering, 1976) or an organism with land-plant affinities (Romankiw and others, 1988). It commonly has two forms, a bifurcating lobate thallus and an elliptical reproductive structure. What is significant about the fossil is that it appears to occur at the same relative stratigraphic horizon throughout the black shales of east-central United States, making it a potentially important stratigraphic time marker (Hasenmueller and others, 1983). *Protosalvinia* has also been traced into the Ellicott Shale Member of the Chadokoin Formation in the New York section (Baird and Lash, 1990). However, comparison of the conodont biostratigraphy of this zone in Kentucky (Ettensohn and others, 1989a) with that in New York (e.g., Woodrow and others, 1988) indicates that the zone is not isochronous throughout its extent, assuming that there is only one

71.4 0.2 STOP 4B. Regressive Black Shales and the *Protosalvinia* (*Foerstia*) Zone (On the eastbound lane of

zone. Again, the *Protosalvinia* zone in Kentucky is apparently three conodont zones younger than it is in New York. Hence, although the *Protosalvinia* zone may be treated as an isochron locally, on a regional scale, the zone is either diachronous or one of possibly a few *Protosalvinia* horizons.

- 71.9 0.5 Lower Huron Member of the Ohio Shale.
- 73.0 1.1 Three Lick Bed (Provo and others, 1978) and ribbed regressive shales on the left.
- 73.3 0.3 STOP 5. Cleveland Shale-Through-Lower Borden Sequence (Devonian-Mississippian) and Implications, Ramey Section (3.6 mi east of the Rowan County line on the westbound lane of I-64, Rowan Co.; Carter Coordinates: 500'FNLx700'FWL, 24-U-72).

CLEVELAND SHALE-THROUGH-LOWER BORDEN SEQUENCE (DEVONIAN-MISSISSIPPIAN) AND IMPLICATIONS

R. THOMAS LIERMAN, CHARLES E. MASON, JACK C. PASHIN, AND FRANK R. ETENSOHN

This stop represents a continuation of the Upper Devonian sequence at Stop 4 into parts of the Lower and Middle Mississippian sequence (Figs. 46, 47) as we move progressively up section. Although the Cleveland Shale Member of the Ohio Shale and the Bedford Shale are present at Stop 4, they are either inaccessible or poorly exposed. This section begins with the upper part of the Cleveland Member, the Bedford Shale, and the Sunbury Shale, as well as the lower three units of the Borden Formation (Fig. 47). Each of these units is briefly described below.

OHIO SHALE

Cleveland Member

The Ohio Shale ranges in thickness from 165 to 200 ft (50-61 m) in the Morehead area. Only the uppermost 11.0 ft (3.4 m) of the Ohio Shale is exposed at the western end of this roadcut. Provo and others (1978) suggested that the upper part of the Ohio Shale in eastern Kentucky was equivalent to the Cleveland Member of the Ohio Shale in Ohio. This was determined from several roadcuts just west of this stop along I-64 (see Stop 4). At this locality, the Ohio Shale consists of a brownish-black to grayish-black, fissile, carbonaceous, silty shale which contains scattered framboids and crystals of pyrite and marcasite. Locally, the upper 20 in (50 cm) displays a transitional contact with the overlying Bedford Shale. This contact consists of greenish-gray silty shales of the Bedford which are intercalated with brownish-black fissile shales of the Ohio (Chaplin and Mason, 1979). Megafossils are rare except for local occurrences of the brachiopods, *Lingula* and *Orbiculoidea*. The upper nine inches (23 cm), of the Ohio Shale contain randomly oriented *Lingula* on bedding-plane surfaces. Microfossils which have been collected from processed residues of these shales include carbonized plant remains, fish scales and dermal plates, conodonts, and some miospore material. Sparse trace fossils can be found at the upper surface of the unit and consist of vertical and horizontal pyrite-filled burrows.

The upper part of Cleveland Member of the Ohio Shale represents the extremely slow accumulation of hemipelagic muds in deeper parts of the basin, under probably dysaerobic conditions (Pashin and

Ettensohn, 1992). Evidence for this interpretation will be discussed in a subsequent section for this stop entitled the "The Bedford Fauna."

BEDFORD SHALE

The Bedford Shale in the Morehead area ranges in thickness from 15 to 30 ft (4.6-9.1 m). At this stop the Bedford is 18-ft (5.6-m) thick and consists of medium bluish-gray to medium light olive-gray, silty shales which weather to a greenish- or yellowish-gray color. It is poorly fissile and contains pyritic and dolomitic nodules, along with calcitic cone-in-cone structures near the top. Very thin, dolomitic siltstone beds and lenses are scattered throughout the unit. The upper contact of the Bedford with the overlying Sunbury Shale is sharp.

Megafossils are sparse at this locality, but nearby localities have yielded a sparse fauna of pelecypods, gastropods, ostracods, and brachiopods. Washed residues of the shale have yielded microfossils such as conodonts and an extensive miospore flora (Chaplin and Mason, 1979; Coleman and Clayton, 1987). Based upon miospore evidence, the Devonian-Carboniferous boundary is located somewhere between 4 in (10 cm) below and 8 in (20 cm) above the Sunbury-Bedford contact (Coleman and Clayton, 1987), most probably at the contact itself.

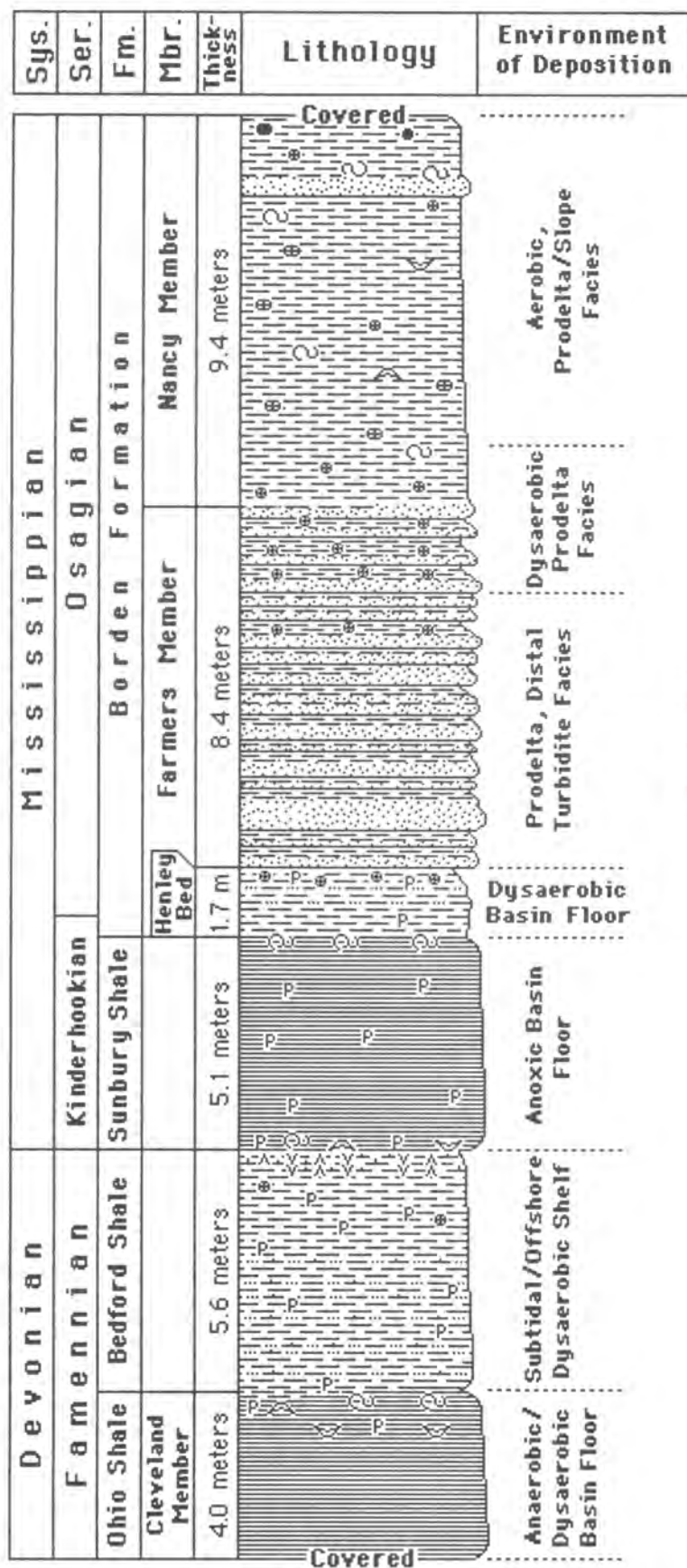
The Bedford represents the slow accumulation of muddy sediments in a subtidal/offshore setting, under generally dysaerobic conditions (Pashin and Ettensohn, 1987, 1992). Evidence for this dysaerobic environment includes: 1.) a high degree of bioturbation in the sediments, 2.) a fairly low diversity of marine fossils, 3.) a fauna dominated by a mollusks (this is not so evident here but can be seen at other localities), 4.) the abundance of pyrite, particularly replacing fossils 5.) the lack of any organic material in the sediments, and 6.) the gray-green color of fine-grained sediments composing the Bedford.

SUNBURY SHALE

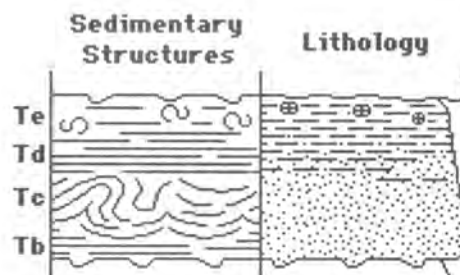
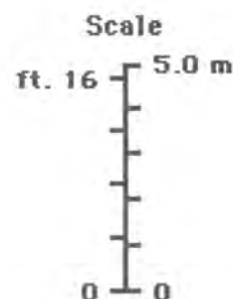
The Sunbury Shale ranges in thickness from 15 to 20 ft (4.5-6.0 m) in the immediate area. At this locality the unit is 16.7-ft (5.1-m) thick and consists of a black to grayish-black, carbonaceous, and fissile, silty shale which contains scattered framboids and crystals of pyrite. It also contains scattered pyrite nodules as much as one-inch (2.5-cm) across.

The clay-mineral assemblage composing these shales includes kaolinite, Fe-rich chlorite and a mixed-layer illite/smectite (see Table 1). This shale is highly organic and can produce an average of 25 gallons (95 l) of crude oil per one ton (909 kg) of processed shale (Joe Gilbert, pers. commun., 1985). The organic fraction varies from 17 to greater than 21 weight percent. The lower contact is quite sharp and is characterized by a lag of pyritic sandstone and phosphatic fossil debris which is as thick as a half inch. The lag contains abundant conodonts and fish-bone fragments and probably represents a condensation horizon or subtle disconformity. The upper 3.3 ft (1.0 m) is represented by a transition zone consisting of interbedded greenish-gray silty shale seams of Henley-like lithology and black fissile Sunbury shales. Vertical and horizontal burrows filled with greenish-gray shales from the Henley Bed are found in the upper eight inches (20 cm) of the Sunbury.

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



Day 2, Stop 5, Ramey Section



Complete Turbidite Cycle from Farmers Member Bouma Divisions Tb-Te

Legend

- ● = Phosphate nodules
- V V V = Cone-n-cone
- ⊗ ⊗ = Siderite nodules
- P = Pyrite
- ∞ = Bioturbation
- ~ = Fossils
- V. fine grained sandstones
- Non-fissile, silty mudstones
- Greenish-gray, silty shale
- Carbonaceous, fissile shale

Figure 47. Generalized stratigraphic section and illustrative sedimentary structures for the Ramey section at Stop 5 on Day 2.

Table 1. Major compositional components of the Sunbury Shale at Stop 5 (Day 2, Ramey section).

Sample No.	Depth below Contact (m)	Percent			Clays Normalized to 100%				Illite XI.
		Clay	Silt	Organics	Kaolinite	Illite	Chlorite	Smectite	
Top	-0.10 m	0.14	0.86	0.17	3.67	53.51	4.36	38.46	0.487
Mid	-1.95 m	0.15	0.85	0.15	2.73	54.49	2.72	40.06	0.471
Base	-3.00 m	0.16	0.84	0.21	0.00	88.84	8.16	3.00	0.517
Average		0.15	0.85	0.18	2.13	65.61	5.08	27.17	0.490

Table 2. Major compositional components of the Henley Bed (Farmers Member) at Stop 5 (Day 2, Ramey section).

Sample No.	Height above Contact (m)	Percent			Clays Normalized to 100%				Illite XI.
		Clay	Silt	Organics	Kaolinite	Illite	Chlorite	Smectite	
E	3.75 m	0.33	0.67	0	10.41	63.45	10.42	15.72	0.578
D	2.75 m	0.29	0.71	0	9.64	58.65	10.32	21.39	0.593
C	1.75 m	0.24	0.76	0	6.89	59.86	10.09	23.16	0.539
B	1.00 m	0.40	0.60	0	6.51	62.62	8.38	22.49	0.607
A	0.10 m	0.35	0.65	0	6.96	60.61	8.42	24.01	0.553
Average		0.32	0.68	0	8.08	61.04	9.53	21.35	0.570

carbonized plant remains, and some miospore material. Microfossils include conodonts, arenaceous foraminifers, and miospores (Chaplin and Mason, 1979; Coleman and Clayton, 1987).

The Sunbury Shale represents the extremely slow accumulation of hemipelagic muds in the deepest parts of the basin, under largely anoxic conditions. Evidence for this includes: 1.) a general lack of bioturbation, 2.) the low diversity of marine fossils, 3.) a high organic content, and 4.) the abundance of pyrite and marcasite.

BORDEN FORMATION

Henley Bed, Farmers Member

The Henley Bed ranges in thickness from four to 12 ft (1.2-3.7 m) in the immediate area. At this stop the Henley is 5.5-ft (1.7-m) thick and consists of dark greenish-gray to grayish-green, poorly fissile, silty shales which weather yellowish-gray. Thin siltstone to very fine-grained sandstone interbeds are also present in the upper part of the unit. These shales also contain small crystals of pyrite and marcasite. The clay minerals composing these shales include kaolinite, Fe-rich chlorite and mixed-layer illite/smectite (see Table 2).

Body fossils have not been found within the Henley Bed at this locality though it does contain a diverse microfossil assemblage. Included here are a well preserved conodont fauna, arenaceous foraminifera, and miospores. Based upon this fossil assemblage, the boundary between the Kinderhookian and early Osagean has been placed within the Henley Bed at about 0.8 to 1.6 ft (0.2-0.5 m) above the Henley-Sunbury contact (Chaplin, 1982; Coleman and Clayton, 1987). The unit is generally quite bioturbated, and although small horizontal feeding traces are observed locally, they have not been specifically identified.

The Henley represents the extremely slow accumulation of hemipelagic muds in the deepest parts of the depositional basin, under generally dysaerobic conditions. Evidence for this dysaerobic environment includes: 1.) a high degree of bioturbation of the sediments, 2.) a fairly low-diversity marine fauna, 3.) the abundance of pyrite and marcasite in the sediments, (4) the lack of any organic material, and 5.) the gray-green color of sediments making up the Henley (see Optional Stop B, Day 3, for additional information supporting this interpretation).

Farmers Member

The Farmers Member varies in thickness from 20 to 80 ft (6.0-24.4 m) in the Morehead area. At this location it is approximately 27.5-ft (8.4-m) thick and consists of alternating beds of sandstone and siltstone with interbeds of shale. The sandstone beds tend to be medium- to light-gray in color, medium- to thick-bedded, and consist of very fine- to fine-grained, moderate to well sorted, sublitharenites. Siltstone beds are likewise medium-to-light gray in color, and consist of medium- to coarse-grained silts which are poor to moderately sorted. Individual sandstone and siltstone beds have a uniform thickness, display sharp erosional bases, and grade upward into overlying shale interbeds. Shales are typically greenish-gray, moderately fissile, and silty. The clay minerals making up these shales include kaolinite, Fe-rich chlorite and a mixed-layer illite/smectite. Locally present within the shales are siderite nodules and lenses which are randomly scattered or concentrated along distinct horizons. The ratio of shale to sandstone beds increases up-section.

Sedimentary structures are numerous and well developed in the sandstone and siltstone beds of the Farmers. Internal bedding features within the sandstone beds include parallel laminae, ripple-drift laminae, convolute laminae and what appears to be wavy or hummocky crossbedding. These internal sedimentary structures are generally arranged in such a way as to form parts of recognizable Bouma sequences. These sequences most commonly include the divisions Tc-Te, though a few of the more complete cycles show a Tb-Te sequence (Fig. 47). A complete Tb-Te sequence starts with an erosional lower contact which is overlain by a Tb division (parallel laminae). This interval is thought to form under transitional-flow conditions between upper and lower flow-regime settings. This division is followed by a Tc division which consists most commonly of convolute or wavy laminae. Locally, ripple-drift laminae are also observed in this interval, apparently forming under lower flow-regime conditions. The Tc division is next overlain by a set of parallel laminae which represent the Td division. Deposition here is best described as occurring under low flow-regime conditions. Finally, fine-grained muds settle out of suspension to form the Te division. We can best describe this Te interval as hemipelagic, because it represents both allochthonous sediments which settled out of suspension shortly after the passage of a turbidity current, as well as fine-grained autochthonous sediments which represent the normal accumulation of pelagic marine muds.

There is normally nothing to distinguish the hemipelagic muds from the pelagic sediments. The only exception possibly being the presence of siderite nodules in some of the uppermost parts of the shale intervals. We believe that the siderite nodules formed immediately below the sediment-water interface during conditions of slow sedimentation. Evidence for this will be discussed further in the section on the Nancy Member.

Bedding-plane structures found in the sandstones are generally well developed, especially along the soles of individual sandstone beds, and include a variety of tool marks such as groove casts; prod marks, brush, bounce, and roll marks; scour marks such as flute casts; and load casts. The orientation of these sole marks indicates an east-west paleocurrent direction (Moore and Clark, 1970) with a vector mean between N 100°W and N 80°W. This, of course, would suggest down-slope movement of material from the region of the Appalachian fold belt to the west.

Trace fossils are quite abundant and diverse and can generally be characterized as complex horizontal-feeding burrows and trails. The ichnogenera most commonly present on the upper surface and soles of individual beds are *Bifungites*, *Chondrites*, *Cylindrichnus*, *Helminthopsis*, *Helminthoida*, *Lophoctenium*, *Monocraterion*, *Paleodictyon*, *Palaeophycus*, *Protopaleodictyon*, *Scalarituba*, *Teichichnus*, and *Zoophycos* (Chaplin, 1982). This trace-fossil assemblage has been interpreted by Chaplin (1982) to represent the *Nereites* ichnofacies. Also, escape structures formed by the pre-turbidite trace-fossil fauna found here are almost exclusively in those sandstone beds which are less than 20-in (51-cm) thick.

The megafossils recognized in the Farmers include broken and abraded brachiopods, fenestrate bryozoans, echinoderm debris, gastropods, pelecypods, cephalopods (both ammonoids and nautiloids), and phosphatic tubes. Most of these specimens occur as lags along the lowermost part of individual sandstone beds and are thought to have been transported by turbidity currents from the clastic shelf and upper slope where they lived, to deeper portions of the basin. Other megafossils found in the middle and upper portion of the

sandstone beds include casts of soft-bodied hydrozoans and glass sponges (Yochelson and Mason, 1986). These megafossils, due to their excellent state of preservation, are not thought to have been transported far, and thus were probably living on or near the basin floor. Microfossil assemblages within the hemipelagic-shale interval include conodonts, arenaceous foraminifera, and miospores.

The Farmers Member represents the accumulation of thin-bedded, distal turbidites along the toe of the Borden deltaic sequence. Shale interbeds are thought to represent the normal accumulation of hemipelagic muds along the very distal edge of the Borden delta under normal dysaerobic conditions. Evidence for the dysaerobic setting includes: 1.) a generally high degree of bioturbation in the hemipelagic portions of sandstone-shale couplets, 2.) a fairly low diversity of marine fossils, 3.) a lack of any organic material in the muds, and 4.) the presence of siderite and phosphate nodules in the pelitic portions of the shales.

Nancy Member

The Nancy Member of the Borden Formation ranges in thickness from 21 to 175 ft (6.5-53.3 m); however, only the lower 38 ft (11.6 m) are exposed here. The Nancy Member consists of greenish-gray to bluish-gray, poorly fissile, silty shales which are highly bioturbated. The clay-mineral assemblages composing these shales include kaolinite, Fe-rich chlorite and a mixed-layer illite/smectite. Also present within the Nancy are siderite nodules which occur scattered throughout the unit or concentrated along distinct horizons. Phosphate nodules also occur sparsely scattered throughout the Nancy Member but may be concentrated in certain intervals as can be observed in the shales above the distal turbidite bed in the Nancy at this stop. These deposits are thought to have formed during periods of extremely slow sedimentation or sediment starvation in the manner described for the siderite deposits below. The phosphate nodules are small (< 3 in or 7 cm across) with considerable variation in their shape. These nodules are irregular, spherical, ovoid to cylindrical in shape, light olive-gray to olive-gray in color and average 22.5% P_2O_5 in content (Todd Stewart, pers. commun., 1985).

Scattered body fossils tend to occur as molds, casts, or mineralized replacements within these siderite and phosphate nodules. Megafossils found to date include brachiopods, rostroconchs, bryozoans, conularids, pelecypods, phyllocarid remains, gastropods, cephalopods, crinoid debris, *Sphenothallus* (phosphatic tubes), fish remains, and rare trilobites. An extensive microfossil assemblage found in the Nancy includes conodonts, arenaceous foraminifera, ostracods, scolecodonts, and miospores. Most of the ichnogenera identified in the lower Nancy are associated with the *Nereites* ichnofacies (Chaplin, 1982) and include *Bifungites*, *Chondrites*, *Helminthoida*, *Helminthopsis*, *Scalarituba*, and *Zoophycus*. These tend to be associated with turbidite beds found in lower parts of the unit. The shaly portions of the Nancy are so extensively bioturbated that identification of individual traces is not possible.

The Nancy Member contains the Cave Run Lake fauna at this stop. This dysaerobic fauna is restricted to the basal 6.6 ft (2.0 m) of the Nancy in the Morehead area. The faunas was first discovered near the dam of nearby Cave Run Lake and has since been traced eastward along I-64 to Morehead. The fossils making up this fauna are largely juveniles. They occur in the Nancy shales as discrete individuals replaced by pyrite, or less commonly, marcasite. Though the fauna has a low diversity, individual species occur in great abundance. Overall, the fauna is dominated by mollusks which are

typical of Devonian and post-Devonian dysaerobic faunas (Kammer and others, 1986). Individual forms dominating this fauna include the ammonoid *Kazakhstania* and the archaeogastropod *Sinuitina*. Other megafossils found in this fauna include more ammonoid and gastropod species, nautiloids, pelecypods, hyolithids, rostroconchs, brachiopods, crinoids, and rare trilobites.

Evidence for the dysaerobic environmental setting of this fauna includes: 1.) the existence of dwarfism in some of the ammonoid species, 2.) a high percentage of juveniles, 3.) a low species diversity, 4.) the predominance of mollusks, 5.) a trophic structure similar to living dysaerobic faunas, and 6.) the replacement of skeletal material by pyrite and marcasite (Mason and Kammer, 1984; Kammer and others, 1986). It should also be noted that fossils found above the basal 6.6 ft (2 m) in this area are predominantly adults and are commonly not replaced by pyrite or marcasite. Thus, it is believed that aerobic conditions prevailed by this time in overlying parts of the Nancy.

We suggest that these siderite nodules, lenses, and beds formed immediately below the sediment-water interface, shortly after deposition of the sediment, but prior to compaction of the shale. Evidence for the early syngenetic formation of these nodules includes: 1.) the fact that the enclosing shales appear to be squeezed around the nodules, indicating that the nodule was present prior to compaction of the less resistant shale; 2.) the presence of three-dimensional burrow systems within the nodules, which are noticeably absent in the enclosing shales due to compaction; and 3.) the fact that the nodules appear to form or nucleate around bits of organic matter such as fossil shell material. Most body fossils found in the nodules show little evidence of having been crushed, which similarly suggests a nodule formation prior to compaction.

The Nancy Member represents the deposition of silty muds and occasional distal turbidites in a prodelta setting. Deposition seems to have occurred under dysaerobic to aerobic conditions.

THE BEDFORD FAUNA

Jack C. Pashin

The unique intertonguing relationships between the Bedford Shale and the Cleveland Member of the Ohio Shale at this stop also provide critical evidence regarding the ecology and sedimentology of oxygen-deficient basins. The Bedford Shale of Kentucky and Ohio contains a thinly shelled brachiopod-mollusk assemblage called the Bedford fauna (Morse and Foerste, 1909). Although the Bedford fauna has many characteristics of a dysaerobic assemblage (Pashin and Ettensohn, 1992), the fauna does not fit into typical oxygen-deficient biofacies models (e.g., Byers, 1977; Kammer and others, 1986) because it is richest and most abundant immediately adjacent to black shale. This discussion explains this unusual fossil distribution and provides insight into the ecology and sedimentology of the dysaerobic zone.

Near the Bedford-Cleveland contact in northeastern Kentucky, gray Bedford Shale and black Cleveland shale intertongue; at Morehead, intertonguing is observable at outcrop scale. The gray-shale tongues contain the Bedford fauna, which is dominated by articulate brachiopods, such as *Syringothyris bedfordensis*, *Rhipidomella* sp., *Rugossochonetes* sp., *Rhynchopora* sp., and bivalves, particularly *Parallelodon irvinensis* and *Palaeoneilo bedfordensis* (Pashin and Ettensohn, 1992). The shale is intensely bioturbated, and the fossils

are commonly reoriented from original life positions; many of the burrows extend downward into black shale. The black-shale tongues contain a less rich fauna which is dominated by inarticulate brachiopods, namely *Lingula melie*, *L. meeki*, and *Orbiculoidea herzeri*. *Lingula melie* and *Orbiculoidea herzeri* are commonly preserved with, or attached to, plant fossils, whereas *Lingula meeki* is not associated with plant material and is locally abundant in patches within the shale (Pashin and Ettensohn, 1992).

According to Pashin (1990) and Pashin and Ettensohn (1992), fossiliferous Bedford Shale accumulated along the distal margins of mud-rich, toe-of-slope turbidite aprons (Fig. 48). The fossiliferous Bedford is interpreted to be a dysaerobic deposit owing to its gray color, bioturbated nature, proximity to black shale, and the presence of the thin-shelled, brachiopod-mollusk-predominate Bedford fauna. Dissolved and pyritized brachiopod and mollusk shells, which dominate the fauna, also are common in dysaerobic fossil assemblages (Kammer, 1982, 1985). The fossils are considered to be largely autochthonous, although reorientation by bioturbation is readily apparent.

Black Cleveland Shale is typical of organic-rich, basin-floor deposits (Fig. 48), and a general lack of bioturbation and autochthonous fauna has been cited as evidence for deposition in the anaerobic zone (e.g., Byers, 1977; Ettensohn and Barron, 1981; Ettensohn, 1985b; Kammer and others, 1986). *Lingula melie* and *Orbiculoidea herzeri* probably lived attached to plants that floated out to sea. *Lingula meeki*, however, probably was not transported into the black-shale environment, because evidence for currents, such as preferential orientation of fossils, is lacking; thus the taxon is interpreted to represent "in situ death assemblages" (autochthonous thanatocoenoses). Inarticulate brachiopods have hemoglobin that is specially adapted for life in water containing less than 0.1 ml/l O₂ (Emig, 1981), but patches with abundant lingulids may indicate habitats higher in the dysaerobic zone.

Autochthonous fauna in the black shale indicates that, contrary to most models, a significant part of the shale was deposited in the dysaerobic zone (Fig. 48). However, the fauna is dominated by semi-infaunal brachiopods such as *Lingula*, and the absence of deep burrows indicates that the sediment was anoxic and too foul with anaerobic bacteria to accommodate most infaunal organisms. Even so, burrows that extend downward from gray shale into black shale indicate that the black mud provided nutrition for some infauna.

Restriction of the Bedford fauna to the lowermost part of the Bedford Shale appears to reflect an interplay of several ecologic and sedimentologic factors (Fig. 48). Evidence from modern dysaerobic environments indicates that organismal abundance reaches a peak near the top of the pycnocline in water with an oxygen content of approximately 0.5 ml/l (Karl and Knauer, 1984; Thompson and others, 1985; Mullins and others, 1985). This so-called "edge effect" has been interpreted to be the result of bacteria-mediated nutrient recycling, in which bacterially reduced nitrite diffuses upward and is oxidized into biologically usable nitrate (Anderson, 1982). Fossil evidence indicates that a similar edge effect may operate near the base of the pycnocline where nutrient recycling may be mediated by anaerobic bacteria dwelling below the sediment-water interface (Savrdra and Bottjer, 1987).

Perhaps the Bedford fauna lived in the upper part of the dysaerobic zone where the edge effect is strongest, having taken advantage of the nutritive benefits offered by the juxtaposition of moderately

oxygenated water with foul, anoxic, black mud (Fig. 48). Intertonguing of black shale and fossiliferous gray shale suggests that a habitable substrate was more important than oxygenation in restricting the Bedford fauna to gray shale. The fauna appears to be richest and most abundant adjacent to black shale owing to the juxtaposition of organic-poor mud, which provided a viable substrate, with foul, organic-rich, pelagic mud that provided nutrients. Oxidation of nitrite from the organic-rich mud may have maximized the available nitrate, thereby providing a nutrient-rich habitat at the distal fringe of a series of mud-rich turbidite aprons.

HISTORICAL AND REGIONAL FRAMEWORK

Frank R. Ettensohn

The Cleveland and Bedford shales have been interpreted to represent the last transgressive-regressive flexural sequence of the third Acadian tectophase (Fig. 22) (Ettensohn, 1985a; Ettensohn and others, 1988b). This sequence, like that of all previous Acadian tectophases, is incomplete, having ended abruptly after loading-type relaxation due to the onset of new tectonism. In this case, that tectonism was the final or fourth tectophase of the Acadian orogeny, and it is represented here by the Sunbury Shale (Figs. 22, 28a). Moreover, this final tectophase was the only Acadian tectophase to produce a complete flexural sequence, which is effectively represented by the rest of the Mississippian System (Fig. 28).

The Sunbury is a typical transgressive black shale, and with its equivalents, is one of the most widespread Devonian-Mississippian black shales across the midcontinent. Moreover, the lag zone at its base appears to represent a brief depositional hiatus (Ettensohn and others, 1989a), possibly reflecting the passage of a bulge.

The Sunbury also apparently represents some of the deepest water, black-shale depositional environments in the entire black-shale sequence, perhaps in excess of 800 ft (230 m) (Ettensohn, 1990a), based on approximations developed from the models of Klein (1974). This is supported by an unusual abundance of phosphorite, heavy elements, and organic matter which probably would have necessitated upwelling from deeper water environments (Ettensohn and Barron, 1981, 1983).

The influx of coarser clastics in overlying parts of the Borden Formation represents the beginning of loading-type relaxation and the infilling of the Sunbury basin (Fig. 28a, b). In this area, that infilling is composed of Borden deltaic clastics and turbidites, and at this stop the distal turbidites of the Farmers Member are especially well exposed.

An isopach map of the Farmers Member (Fig. 49) (Sable and Dever, 1990), as well as isopach maps of individual turbidites and isopleth maps of the total number of siltstone beds in the unit (Maynard and Lauffenburger, 1978), show a body with an elongate fan-shaped outline thickening north-northeasterly up slope toward the Garrison area, where we will examine the Farmers Member tomorrow at Stops 2 and 3 (Fig. 49, B and C). At this stop we are located near the thinning western margin of the turbidite apron (Fig. 49), and not far from here, geologic mapping indicates that most of the Farmers pinches out into the Nancy Shale, although a few siltstone beds persist more than 20 mi to the south. (Weir, 1976a; McDowell and Weir, 1977). However, the thickening of the Farmers to the north-northeast suggests that the source of the fan was in this area. In fact, thickening in the Farmers roughly coincides with a similar thickening

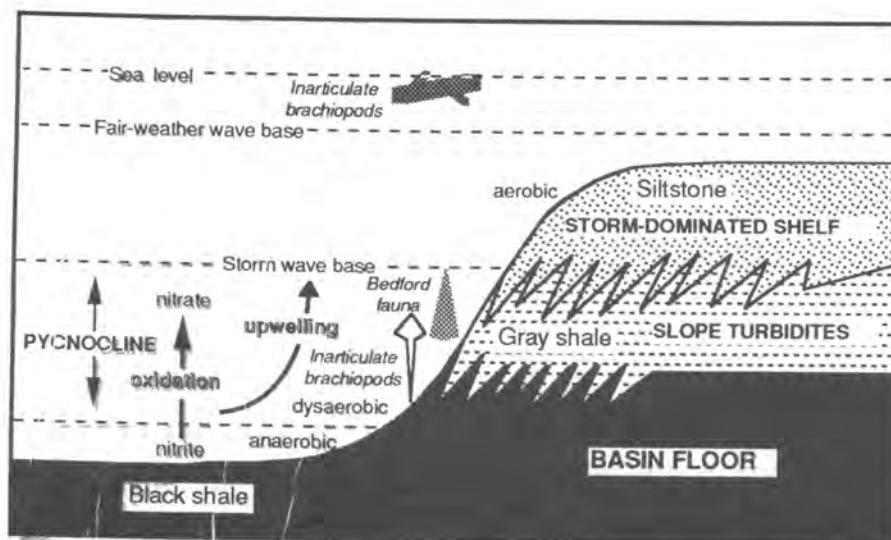


Figure 48. Paleoecologic model for the Late Devonian Bedford fauna (after Pashin and Ettensohn, 1992). Although dysaerobic conditions determined the taxonomic makeup of the Bedford fauna, availability of nutrients and habitable substrates was evidently more important than oxygenation in determining faunal distribution. Nutrients are interpreted to have been made available during upwelling by upward diffusion and oxidation of nitrate through juxtaposition of moderately oxygenated water with foul, organic-rich black mud.

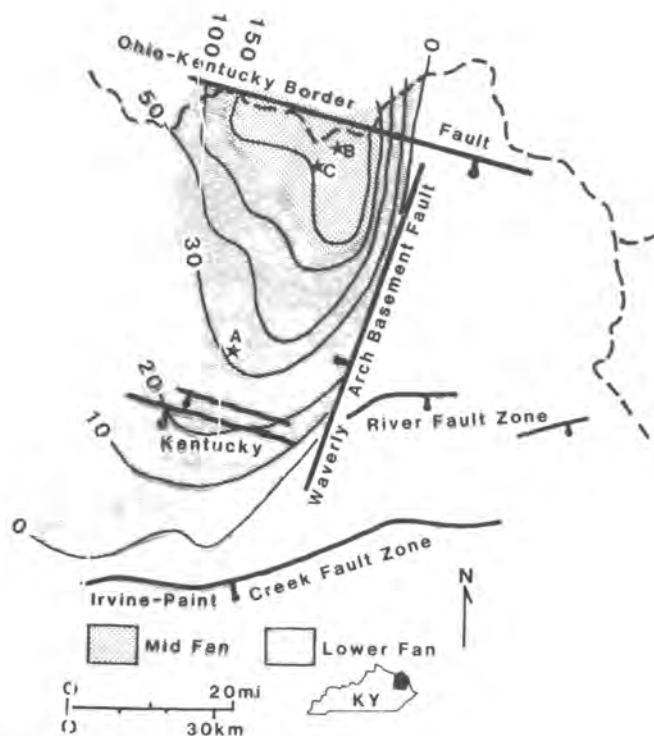


Figure 49. Isopachous map in feet of the Farmers Member of the Borden showing the shape of the fan and its probable control by structural features. The fan is divided into mid- and lower-fan areas based criteria discussed in Normark (1978). A=Stop 5, Day 2; B=Stop 2, Day 3; C=Stop 3, Day 3 (after Sable and Dever, 1990, and Pashin and Ettensohn, 1987).

trend in underlying Berea turbidites, which Pashin and Ettensohn (1987) related to a feeder channel localized on a platform-margin declivity related to growth faulting on basement precursor faults (Ohio-Kentucky Border Fault, Fig. 49; see Stop 1 and optional Stop C, Day 3). Inasmuch as the Farmers thickens in this direction and exhibits more proximal, mid-fan features in the Garrison area (Stops 2 and 3, Day 3), the ultimate source area may have been the same. However, the Farmers feeder channel must have been farther north than the Berea channel, perhaps related to a more northerly fault-bound declivity, because unlike Berea turbidites, Farmers turbidites extend farther northward into southern Ohio (Chaplin and Mason, 1978).

The relatively uniform thickness of the Farmers in this area, as well as the absence of channels and coarse lags, suggests that this part of the Farmers was deposited on the lower fan (Fig. 49; see Normark, 1978). Mid- and lower-fan areas predominate in Kentucky, and based on the isopach map in Figure 49, it is apparent that the extent of the fan and its topography were partially controlled by local structures. Clearly, the eastern boundary of the fan is related to the Waverly arch basement fault, whereas a break in slope and change in fan orientation occur at the Kentucky River fault zone. Finally, the fan ends just north of the Irvine-Paint Creek fault zone (Fig. 49). Except for some possible broadening in the extent of mid-fan deposits just south of the Ohio-Kentucky Border fault, its effects on the Farmers are unclear at this time.

Deposits of a similar origin, but somewhat younger in age, are present a little farther to the east and south in the Borden of subsurface eastern Kentucky (Sable and Dever, 1990). They are known as the Weir Sands, a drillers' term.

- 73.6 0.3 Overpass.
- 74.0 0.4 Farmers Members of the Borden on the left.
- 74.7 0.7 Milepost 134.
- 75.8 0.1 Begin a long exposure which descends stratigraphically from the Nancy Member to the Sunbury Shale.
- 76.4 0.6 Bridge over Bull Fork.
- 77.0 0.6 Begin along exposure ascending from the Sunbury Shale to the Nancy Members to the hilltop.
- 77.1 0.1 Milepost 135.
- 77.5 0.4 Nancy and Farmers members of the Borden.
- 78.4 0.9 Bridge over North Fork of Triplett Creek.
- 78.7 0.3 Farmers Members on both sides of road.
- 79.0 0.3 Exit 137 to Kentucky State Route 32, Morehead and Flemingsburg.
- 79.4 0.4 Overpass.
- 79.6 0.2 Farmers Member on entrance ramp to I-64.
- 80.4 0.8 Farmers Member.

- 80.8 0.4 Farmers Member with a small fault in the east end of the cut.
- 81.4 0.6 Nancy Member with siderite concretions.
- 82.0 0.6 Nancy Member with siderite concretions.
- 82.9 0.9 Entrance to rest area.
- 84.1 1.2 Milepost 142; Nancy Member with siltstones.
- 85.4 1.3 Nancy Shale with siltstones; begin "to climb" the Borden delta front.
- 85.9 0.5 Milepost 144; Nancy-Cowbell contact on the left.
- 86.2 0.3 Nancy-Cowbell contact on the left.
- 86.6 0.4 Last exposure of the Nancy Member.
- 86.7 0.1 Cowbell Member.
- 86.9 0.2 Milepost 145; Cowbell Member with a tongue of Nancy-type shale in the upper half of the exposure.
- 87.0 0.1 Cowbell Member.
- 87.2 0.2 Upper Cowbell Member with the tongue of Nancy-type shale at road level.
- 87.4 0.2 Cowbell Member on the right.
- 87.8 0.4 Overpass; Cowbell Member.
- 88.0 0.2 Z-505; red and green shales of the Nada Member of the Borden Formation, as well as the Renfro, St. Louis, Holly Fork, and Tygarts Creek members of the Slade Formation. For the next 15.5 mi, the highway primarily crosses Middle and Upper Mississippian (Valmeyeran and Chesterian) carbonates of the Slade Formation (Ettensohn and others, 1984; Ettensohn, 1986) (Figs. 1, 50) with the Lower and Middle Mississippian (Kinderhookian and Valmeyeran) Borden Formation cropping out in some of the low valleys and the Lower and Middle Pennsylvanian Breathitt Formation occupying some of the higher ridges. This stretch of highway was important in developing the "Lee-Newman Barrier-Shoreline Model" (Ferm and others, 1971; Horne and others, 1971, 1974) and in the subsequent elucidation of its many problems (Ettensohn, 1980, 1981, 1986; Ettensohn and Dever, 1979; Dever, 1980; Dever and others, 1977). The road log and stops along this segment of the highway are merely adaptations from two earlier guidebooks done in greater detail (Ettensohn and Dever, 1979; Ettensohn, 1981). The "Z numbers" for outcrops are provided for quick reference to these guides and to the figures used herein (e.g., Figs. 51, 52).
- 88.2 0.2 Z-506; St. Louis-through-Ramey Creek members of the Slade Formation (Figs. 1, 51); carbonate paleosols are well developed on the St. Louis. Abundant blue-green spongiostromatolite algae are present in the Armstrong Hill

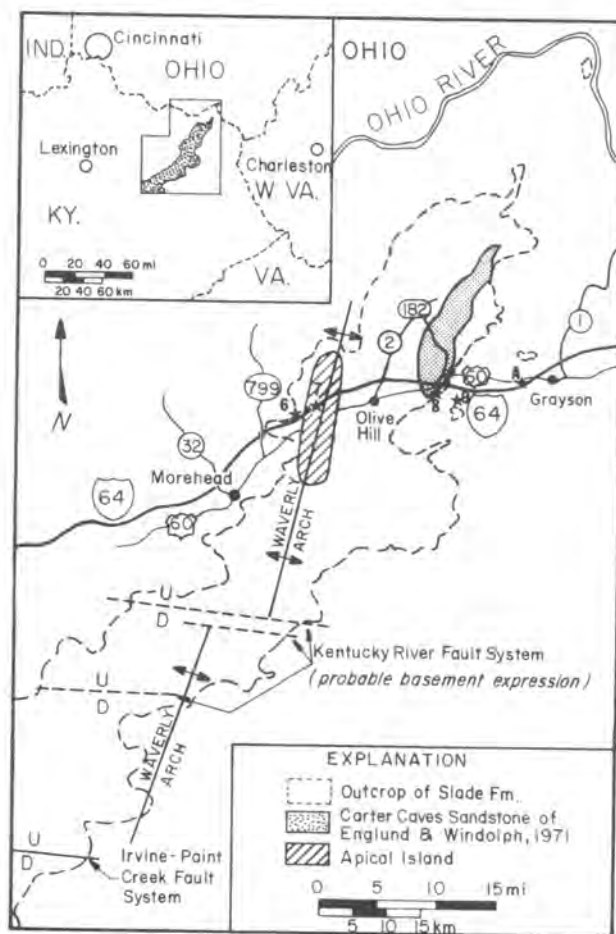


Figure 50. Location of stops (numbered stars) in and near the outcrop belt of the Slade Formation in northeastern Kentucky on Day 2 and Day 3 (optional Stop A). Structures active during the Carboniferous are also shown (after Ettensohn, 1980).

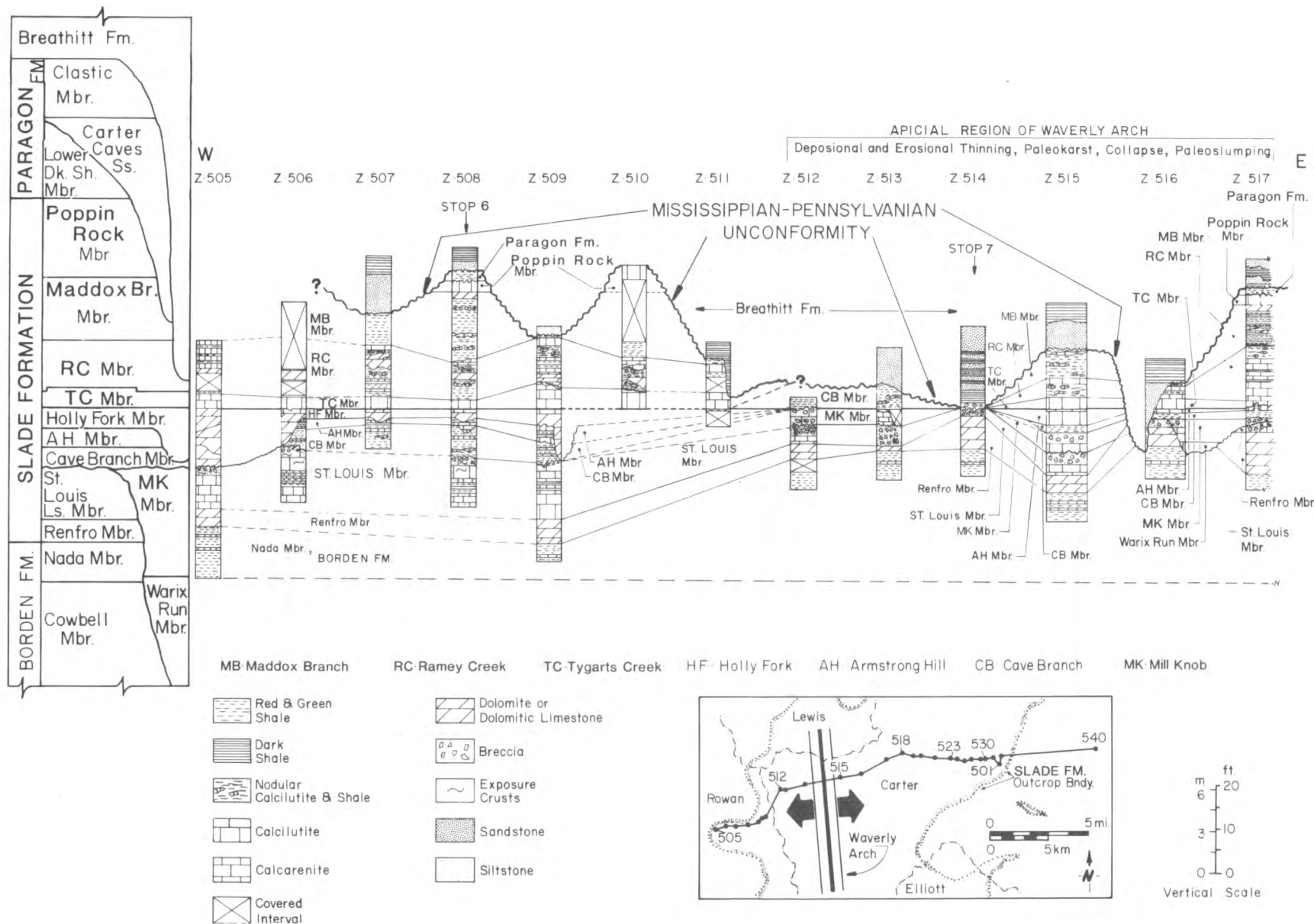


Figure 51. Stratigraphy and correlation of Mississippian units along that part of I-64 west of and on the Waverly arch. Depositional and erosional thinning are apparent on apical portions of the arch, whereas the Mississippian section expands below the Mississippian-Pennsylvanian unconformity westwardly away from the arch (modified from Ettensohn, 1980, 1981).

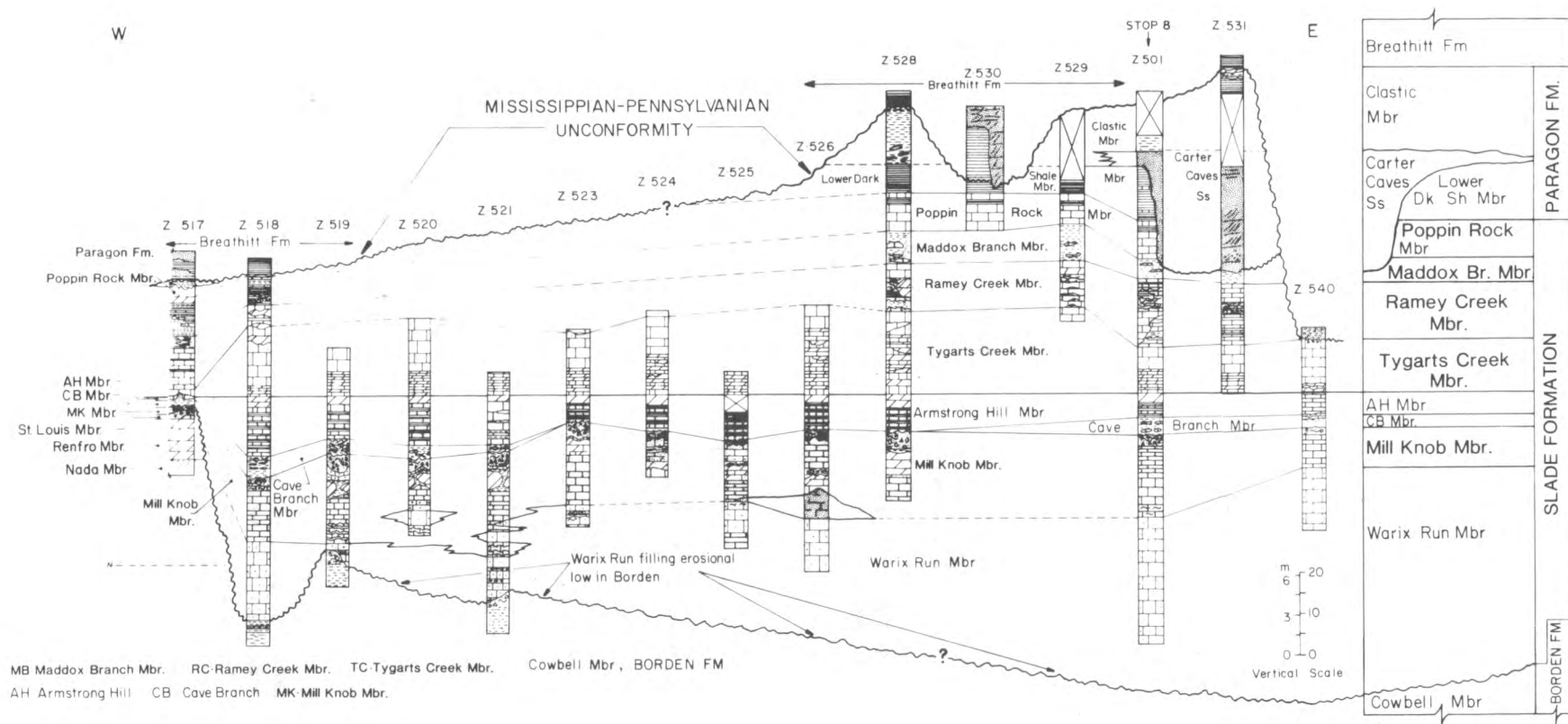


Figure 52. Stratigraphy and correlation of Mississippian units along that part of I-64 east of the Waverly arch, from Olive Hill to the Gregoryville area. Progressively more of the Mississippian section is preserved below the Early Pennsylvanian unconformity eastwardly away from the Waverly arch. Legend on Figure 51 (modified from Ettensohn, 1980, 1981).

Member, and tidal-channel features are apparent in the Holly Fork Member (Ettensohn, 1981).

88.4 0.2 Z-507; Armstrong Hill-through-Maddox Branch members disconformably overlain by Breathitt sands and shales (Fig. 51); palynology of the Breathitt shales at this locality was analyzed by Ettensohn and Peppers (1979).

89.0 0.6 STOP 6. Paleosols and Restricted Slade Carbonates West of The Waverly Arch Apical Island (Z-508; 0.3 mi west of the entrance to the weigh station on the westbound lane of I-64, Milepost 147, Rowan Co.; Carter Coordinates: 2480'FNLx2240'FEL, 19-V-74).

PALEOSOLS AND RESTRICTED SLADE CARBONATES WEST OF THE WAVERLY ARCH APICAL ISLAND

FRANK R. ETENSOHN

The Carboniferous section along Interstate 64 shows a number of anomalous features including the unexpected absence of units, abrupt facies changes, deep erosional truncation, and prominent disconformities within and between major environmental sequences (Figs. 51, 52). These seeming anomalies have resulted in two controversial models for interpreting the Mississippian-Pennsylvanian transition in the area. The Barrier-Shoreline Model (Ferm and others, 1971; Horne and others, 1974) suggested that the discontinuous nature of the units was not caused by erosion, but by lateral gradation from one depositional environment to another so that the Mississippian-Pennsylvanian sequence represents a single offshore (Mississippian)-to-onshore (Pennsylvanian) facies transition (Fig. 53). As a result, widespread tabular units and a mid-Carboniferous or Early Pennsylvanian unconformity (former Mississippian-Pennsylvanian unconformity) were not recognized. During the development of the second, or tabular-erosion model (Dever, 1980; Ettensohn, 1980, 1981), the presence of an arch and a fault zone in the area (Figs. 50, 54) was noted, and the mapping of individual units showed that most of the anomalies could be related to synsedimentary movement on these structures during deposition of the Slade Formation (former Newman Limestone). Because the interstate exposures are the primary evidence for both models, we will examine some of the more important interstate exposures during the next three stops. The preliminary materials below are updated and adapted from a much more detailed guide book for this segment of I-64 (Ettensohn, 1981) in order to provide some basic background.

STRATIGRAPHY

Parts of the Borden, Slade, Paragon (former Pennington Fm.), and Breathitt formations (Fig. 55) figure significantly in the discussion of the two models. The Borden is significant because Slade carbonates and Pennsylvanian clastics disconformably overlie the Cowbell and Nada members of the formation locally. The green shaly siltstones of the Cowbell and the red and green shales of the Nada are lithologically similar to parts of the Paragon, for which they have been mistaken in some studies.

The overlying Slade Formation consists of 10 to 11 members ranging in age from the late Osagean Renfro Member to the middle Chesterian Poppin Rock Member (Figs. 1, 54). In very few places, however, do all these members occur together due to tectonically conditioned erosion and nondeposition. No such place is present in

the field-trip area and one of the members, the Ste. Genevieve, does not occur at all as far north as the field-trip area. Although carbonates compose most of the Slade, red and/or green shales are present in the Cave Branch and in parts of the Ramey Creek and Maddox Branch members (Fig. 55) which also have been mistaken for Paragon red and green shales. The mistake has been made most commonly at places where Paragon sandstones (Stop 8) or Pennsylvanian clastics (as at this stop) disconformably overlie the Maddox Branch Member.

The Paragon Formation overlies the Slade Formation and was formerly called the Pennington Formation in this area. The Paragon Formation is older than the type Pennington (Ettensohn and others, 1984; Ettensohn and Chesnut, 1985), and although the Pennington Formation consists largely of red and green shales in its type area near the Kentucky-Virginia boundary (Campbell, 1893), red and green shales are relatively uncommon in the Paragon Formation. The Paragon consists of four informal lithologic members (Figs. 1, 55), and red and green shales are found only in parts of the clastic member and the upper shale member. The Carter Caves Sandstone which will be viewed at Stop 8 is an unusually thick equivalent of the sandstone that typically occurs at the base of the clastic member.

Pennsylvanian rocks of the Lee and Breathitt formations intertongue complexly and are interpreted by most workers to overlie Mississippian rocks unconformably. The Lee lithology, predominately massive cliff-forming, fine- to medium-grained sandstone exhibiting channel geometries, large-scale crossbedding, and quartz-pebble conglomerates, is nonmicaceous and is commonly restricted to the lower part of the sequence. The Breathitt consists principally of medium-gray to black shales with locally occurring coals, siltstones, and sandstones. Above the level of the Lee, there are local limestones or other marine units. Breathitt sandstones are argillaceous, micaceous, and thin bedded; they contain more shale interbeds, but lack quartz-pebble conglomerates.

STRUCTURE

Northeastern and east-central Kentucky are underlain by two structures which were apparently active during the Carboniferous: 1.) a broad, east-west basement fault system with some surface expression in the Kentucky River fault system and 2.) a north-south-trending uplift named the Waverly arch by Woodward (1961) (Figs. 50, 54). Both structures were active during the Early Paleozoic (Woodward, 1961) and again during the late Paleozoic and appear to have left northeastern Kentucky predominately positive during the Carboniferous (Dever, 1977, 1980; Ettensohn, 1977, 1980, 1981). Much of the structural movement probably reflects bulge reactivation of basement structures during Ouachita tectonism and Acadian relaxation. The final phase of movement, however, during the Early Pennsylvanian probably reflects bulge reactivation during the inception of the Alleghanian orogeny (Ettensohn and Chesnut, 1989; Ettensohn, in press a).

The fault zone marks the northern hinge line of the east-west trending Rome Trough. Surficial stratigraphy, deep drilling, and geophysical evidence along this fault zone indicate a northern, uplifted block, where sediments are thin, and a southern down-dropped block where sediments are thicker (Fig. 54). Similar structural effects are reflected in the nature of the Early Pennsylvanian unconformity. Erosional truncation along the unconformity is greatest on the northern side of the fault zone where Pennsylvanian rocks may disconformably overlie parts of the Borden and Slade (Fig. 54). South of the fault zone, the

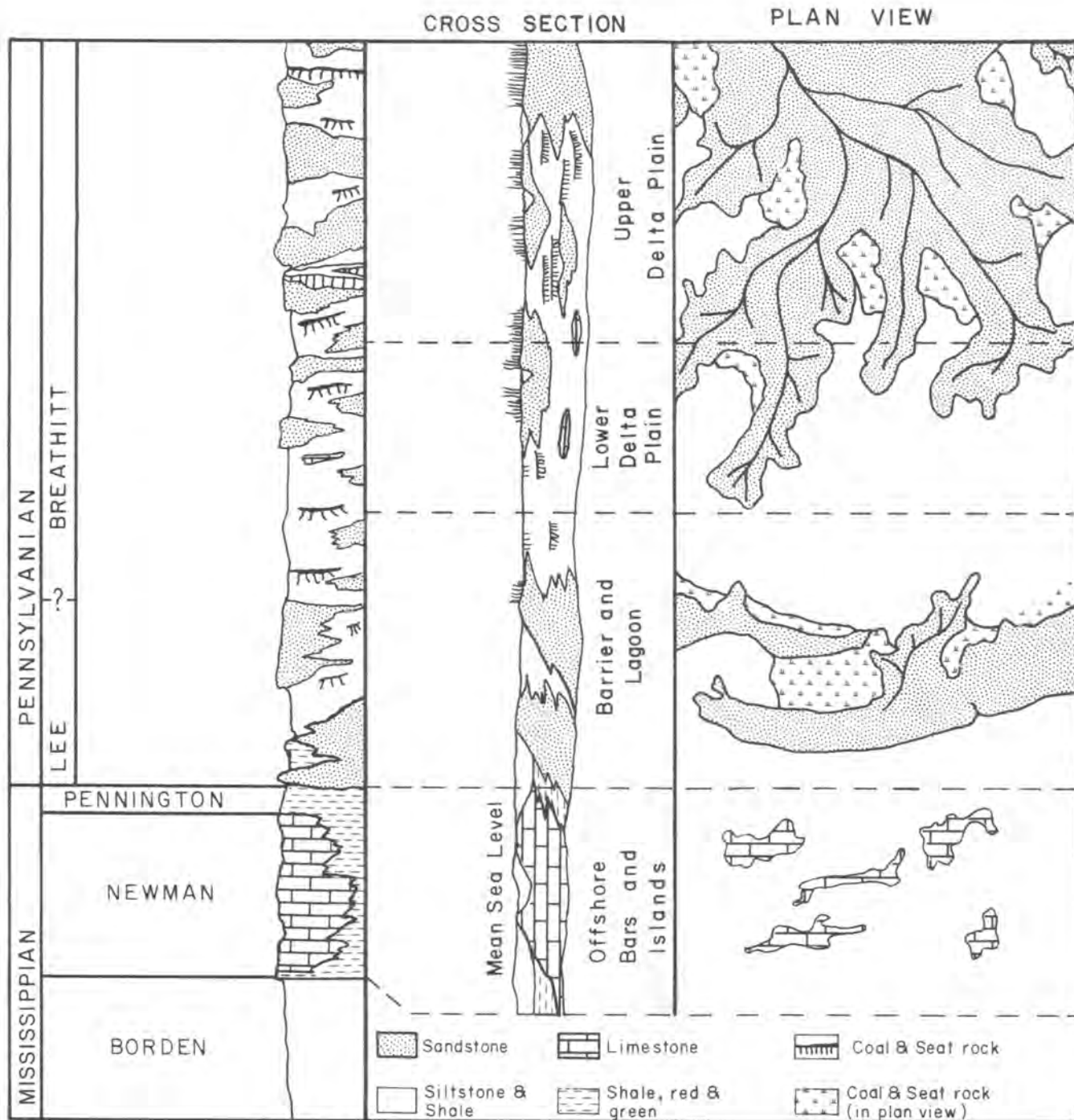


Figure 53. Succession of major Carboniferous units in east-central Kentucky and their interpreted environmental and facies relationships in the Barrier-Shoreline Model (after Ferm and others, 1971). What is called Pennington in this model is in large part equivalent to the present-day Paragon Formation; the Newman is in large part equivalent to the present-day Slade Formation (see Ettensohn and others, 1984, fig. 2).

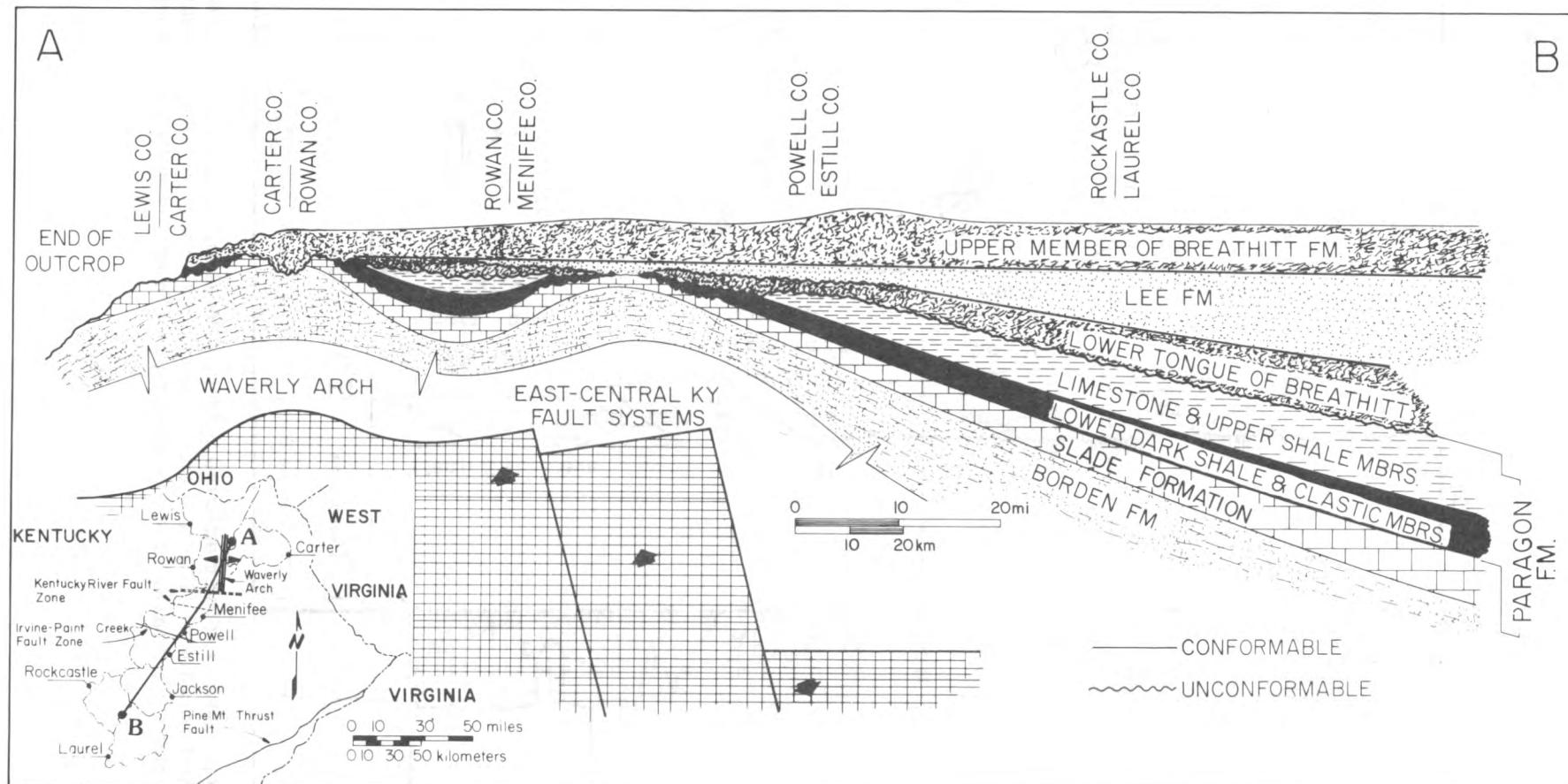


Figure 54. Schematic north-south cross section along the east-central Kentucky outcrop belt based on approximately 90 measured sections showing inferred relationships between major Carboniferous formations, unconformities and underlying structure (modified from Ettensohn, 1980, 1981).

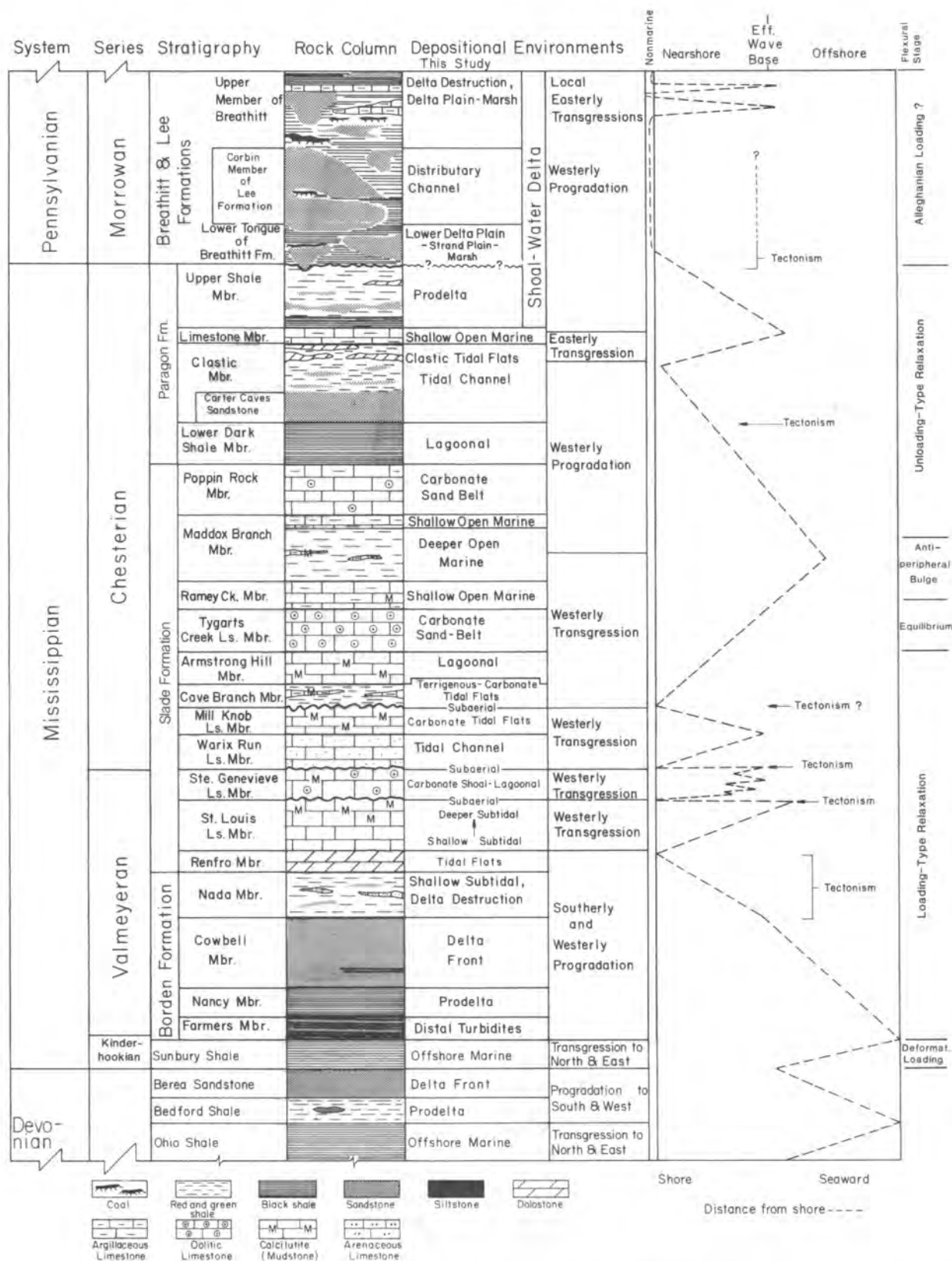


Figure 55. Generalized sequence of Upper Devonian and Carboniferous units in east-central Kentucky showing inferred succession of environments, environmental continua, local tectonic events, and probable flexural stages (see Fig. 28) (modified from Ettensohn, 1980, 1981).

erosional gap on the unconformity becomes smaller, and a more complete Mississippian-Pennsylvanian sequence is preserved (Figs. 37, 54). Environmental and ecologic conditions during the Mississippian were generally less stable on the northern uplifted block than on the southern block, because of shallow seas and greater periods of exposure. The field-trip route remains entirely on the northern uplifted block (Fig. 50) where Upper Mississippian units are typically thinner and contain more clastic constituents.

Woodward (1961) described the Waverly arch as a broad, concealed arch east of the larger Cincinnati arch, extending from central Ohio to east-central Kentucky (Fig. 50). Pashin and Ettensohn (1987) have suggested that the location of the arch was also controlled by a basement fault. The influence of the Waverly arch on deposition of parts of the Slade Limestone in northeastern Kentucky was first recognized by Dever (1973) based on thinning and subaerial diagenesis of lower Slade members. The Waverly arch coincides with the Mississippian outcrop belt for nearly half its length in east-central Kentucky, and appears to have greatly influenced Carboniferous sedimentation in the study area. One part of the arch, called the apical island by Ettensohn (1975, 1977, 1980; Figs. 50, 51), underwent far greater uplift than other parts and was a positive or near-positive area throughout most of the Carboniferous. Our field trip route on I-64 crosses this part of the arch (Fig. 50); most Carboniferous units are abnormally thin or absent here (Ettensohn, 1975, 1977). The Lee and Paragon formations are generally absent on this part of the Waverly arch, and the Borden and Slade undergo drastic thinning.

Much of the thinning near the structural features is apparently related to erosion accompanying uplift. In this area at least five episodes of uplift have been recognized through erosional disconformities (Fig. 56). There is a problem in recognizing all the disconformities, however, because on the structural features themselves, where uplift and erosion have been greatest, these disconformities tend to merge, forming compound disconformities. Hence, the great amount of erosional truncation beneath Pennsylvanian rocks on the apical island is not necessarily the product of any one period of uplift and erosion, but the combined product of five periods of uplift. It is likely, however, based on the probable thicknesses of sediment eroded, that the period of uplift and erosion during the Early Pennsylvanian was the greatest.

BARRIER-SHORELINE MODEL

In the Barrier-Shoreline Model, each formation in the Slade-through-Breathitt sequence was interpreted to represent a broad, nearshore-to-littoral environment within a single regressive, progradational continuum (Fig. 50). Such environments were interpreted to have prograded westwardly over a stable platform, commonly moving in isolated jumps. The Slade Formation was construed to represent offshore carbonate complexes which accreted upward and laterally from shallow carbonate barriers, and along I-64, the Slade was interpreted to exhibit two carbonate-barrier sequences, the "Route 799 carbonates" (west of Waverly arch; Fig. 57A) and the "Route 2 carbonates" (east of Waverly arch; Fig. 58A) separated in time and space (Fig. 59A). Furthermore, the carbonate-barrier complexes were suggested to have intertongued in landward and seaward directions with red and green clays, which were interpreted to represent subtidal to tidal-flat clays shoreward of the carbonates and offshore marine clays seaward to the carbonates. In this model, all red and green shales were assigned either to the Nada Member of the Borden or to the Pennington (Paragon) Formation, both of which were interpreted

to be laterally continuous (Fig. 59A). Finally, these red and green tidal-flat clays were depicted as intertonguing with quartzose beach-barrier sands of the Lee Formation, which in turn intertongued with lower delta-plain shales, coals, and sandstones of the Breathitt (Fig. 59A). The Barrier-Shoreline interpretation of the Carboniferous rocks along I-64 is shown and compared with that of the tabular-erosion model in Figure 59.

Examination of the exposures along I-64 (Figs. 51, 52) reveals a number of stratigraphic, biostratigraphic, and environmental inconsistencies which challenge the validity of the Barrier-Shoreline Model. These inconsistencies can be largely explained in terms of syndimentary tectonic activity on the Waverly arch that was not recognized in original accounts of the model (Ferm and others, 1971; Horne and others, 1974). Recognition of tectonic activity is critical to understanding stratigraphic relationships in the area, for I-64 crosses the most positive part of the arch, where the Slade and Paragon thin drastically and are locally absent (Fig. 59B). After a major period of uplift during the Early Pennsylvanian, erosion on the crest of the arch penetrated to the level of red and green Nada shales in the Borden (Fig. 59B). Breathitt sands and shales subsequently filled the erosional low on the apex and overlapped eastward and westward onto progressively younger, underlying Mississippian rocks (Fig. 59B). Not only did this erosion and infilling effectively divide the Slade into the two "bodies" described in the Barrier-Shoreline Model (Fig. 59A), but it also brought Nada shales and overlying Breathitt sediments into approximate elevational equivalence with the thinning Slade carbonates on each flank of the arch (Fig. 59B). This elevational equivalence, combined with the fact that the Slade is underlain by red and green shales (Nada), contains members with red and green shales (Cave Branch, Ramey Creek and Maddox Branch), and is locally overlain by red and green shales (Paragon) (Fig. 59B), may explain the appearance of intertonguing between carbonates, red and green shales, and sandstones postulated in the Barrier-Shoreline Model. Nonetheless, each of these red-and-green shale units differs in age and stratigraphic position (Fig. 55) even though relative stratigraphic position may be obscured by structurally related erosion near the Waverly arch. None of these red-and-green shale units is contemporaneous on a large scale with the Slade carbonates.

Moreover, units in the two Slade "bodies" can be correlated across and around apical portions of the arch through erosional remnants, stratigraphic succession, and paleontology, indicating that most Slade units were continuous and coeval before episodic uplift and erosion. The episodic nature of the uplift is indicated by the prominent disconformities throughout the section (Figs. 50, 56). The disconformities are always better developed on and near the Waverly arch. During the post-St. Genevieve period of uplift (Fig. 56, #2), erosion penetrated to the level of the Cowbell on the eastern flank of the arch ("Rte. 2 carbonates"), and lower Chesterian sediments (Warix Run and Mill Knob) subsequently filled the resulting lows (Figs. 52, 58b, 59B). Although this part of the Slade ("Rte. 2 carbonates") is topographically lower than the "Route 799 carbonates," it is no older than the "Route 799 carbonates" (Fig. 59A). Moreover, the inferred contemporaneity between the "Route 2 carbonates" and the Nada shales underlying the "Route 799 carbonates" (Fig. 59A) is impossible for the Nada shale is Osagean in age, whereas the "Route 2 carbonates" are largely early and middle Chesterian in age.

For similar reasons, intertonguing between "Lee sandstones" and underlying "Pennington" (Paragon) red and green shales (Fig. 59A) is also unlikely. Stratigraphic position, lithology, and biostratigraphy indicate that the "Lee sandstones" are actually parts of the Breathitt

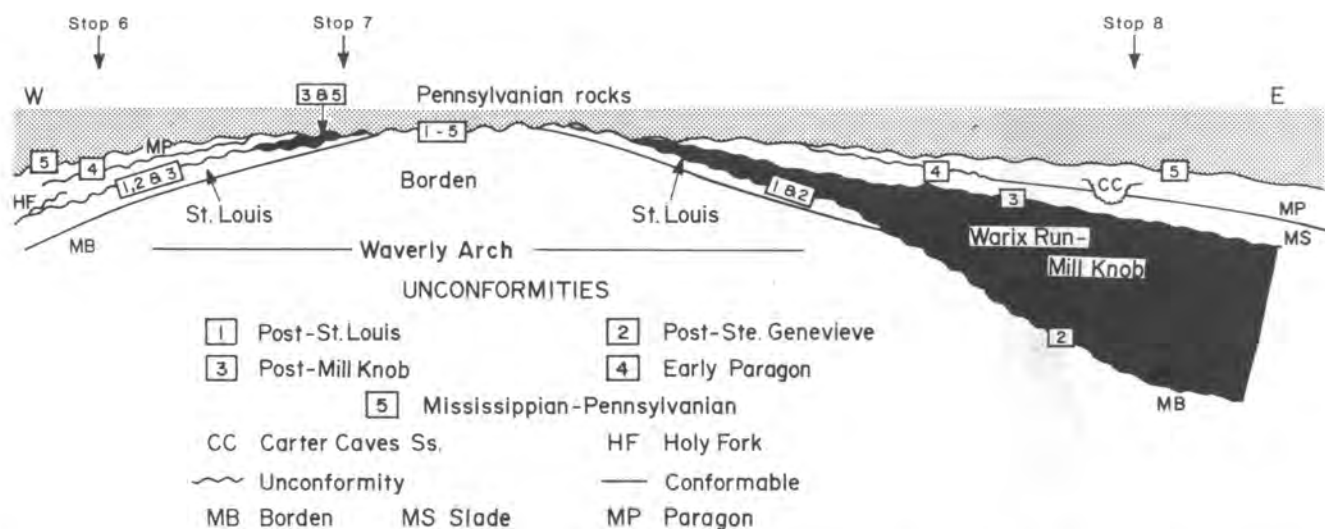


Figure 56. A very schematic east-west section along I-64 showing the generalized stratigraphic positions of five Carboniferous unconformities relative to the Waverly arch. All unconformities coalesce into a single compound unconformity near the apex of the arch. Dotted pattern represents Pennsylvanian rocks, and dark stippling represents the Warix Run and Mill Knob members, but only the Mill Knob is present high on east and west flanks of the arch. Position of early Paragon unconformity is idealized based on three isolated erosional remnants; along most of I-64, the unconformity was destroyed by Early Pennsylvanian erosion. This unconformity occurs only near structural elements; away from these elements, the Slade-Paragon contact is conformable. Numbered arrows across top reflect approximate positions of stops. Not drawn to scale (modified from Ettensohn, 1981).

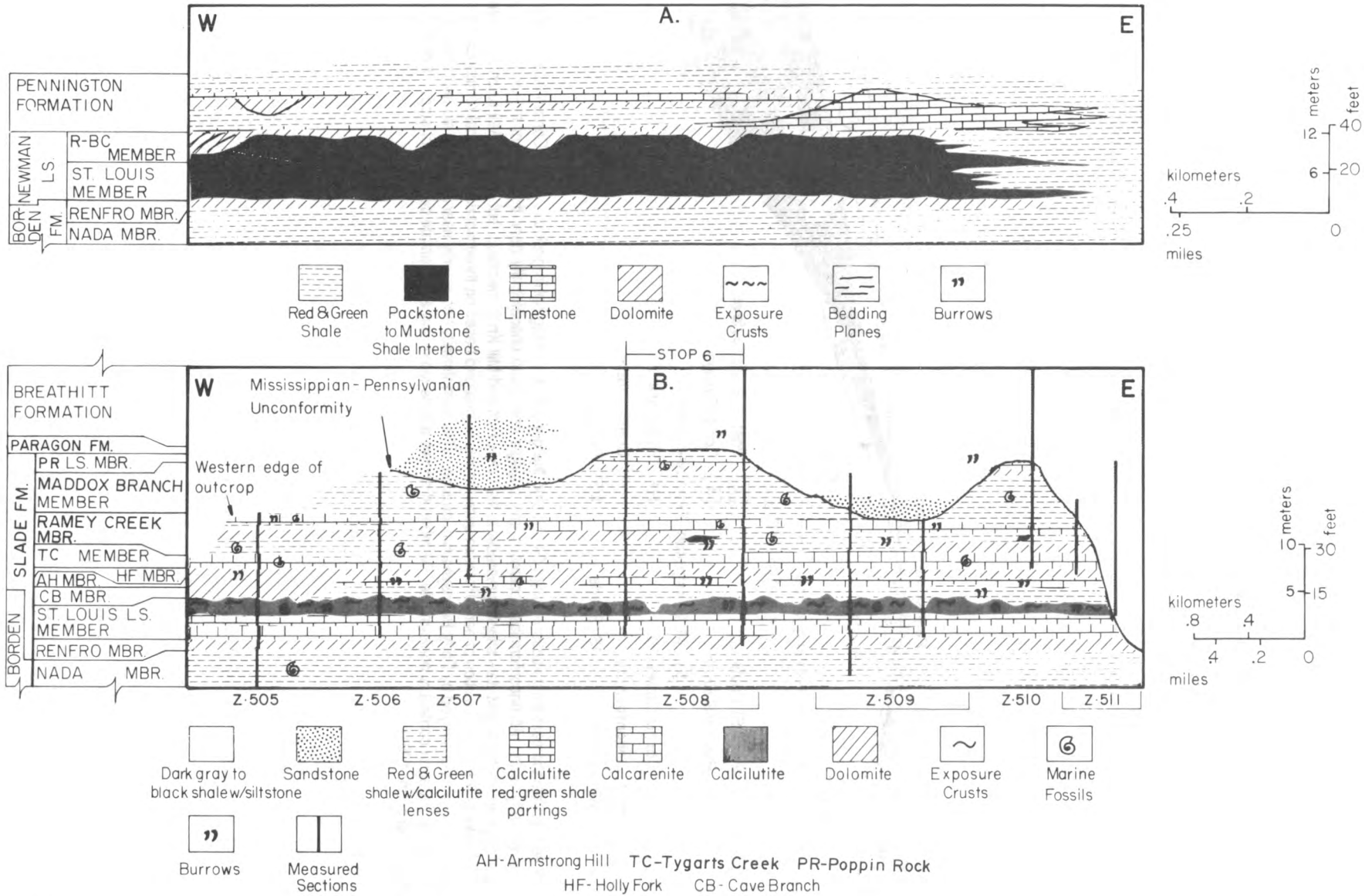


Figure 57. Comparison of Barrier-Shoreline interpretation (A) modified from "Route 799 carbonates" of Horne and others (1974) using their nomenclature, and a "tabular-erosion" interpretation (B) of stratigraphy west of the Waverly arch on I-64. See Figure 51 for detailed sections (modified from Ettensohn 1980, 1981).

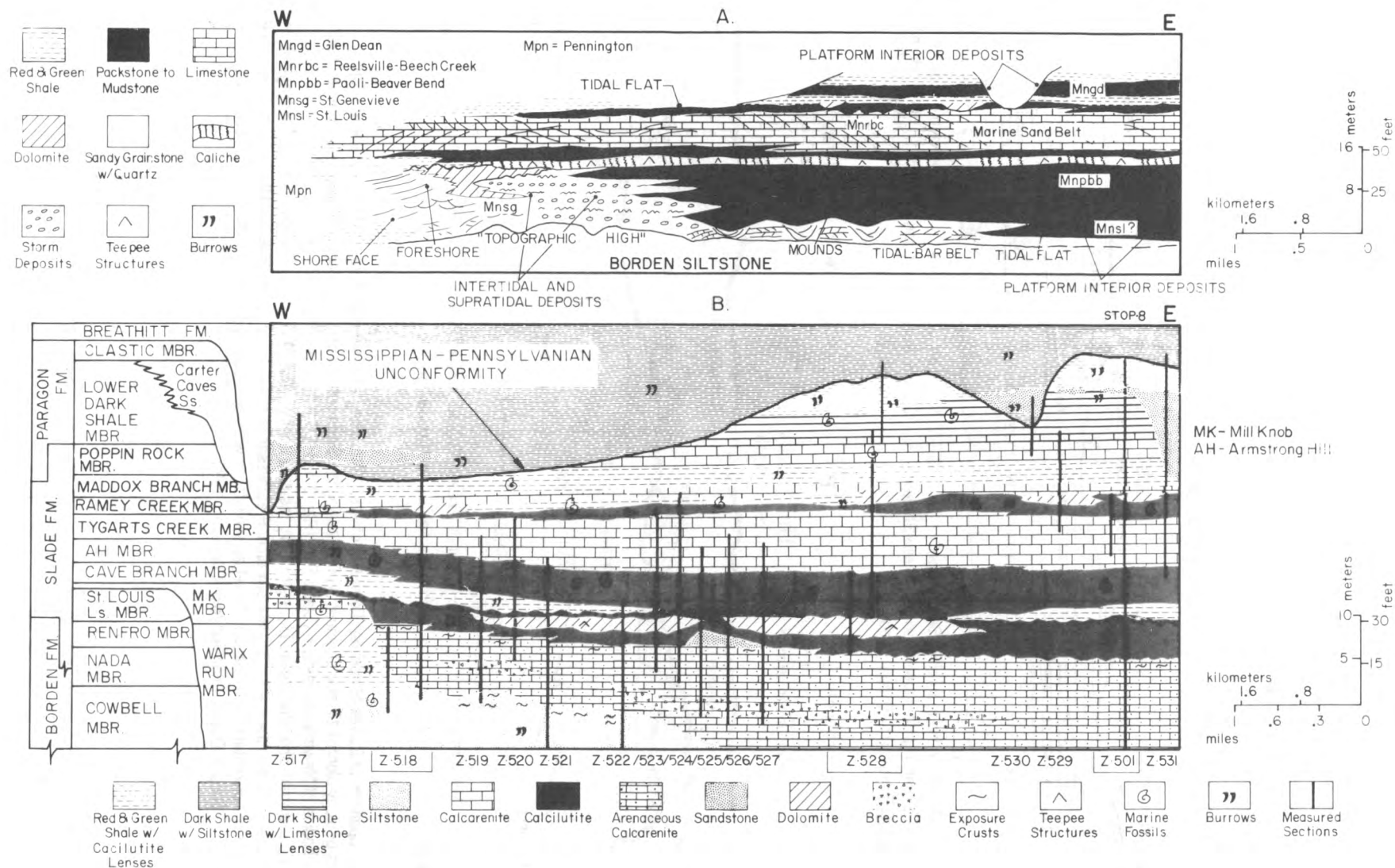


Figure 58. Comparison of Barrier-Shoreline interpretation (A) modified from "Route 2 carbonates" of Horne and others (1974) using their nomenclature, and a "tabular-erosion" interpretation (B) of stratigraphy east of the Waverly arch on I-64. See Figure 52 for detailed sections (modified from Ettensohn, 1980, 1981).

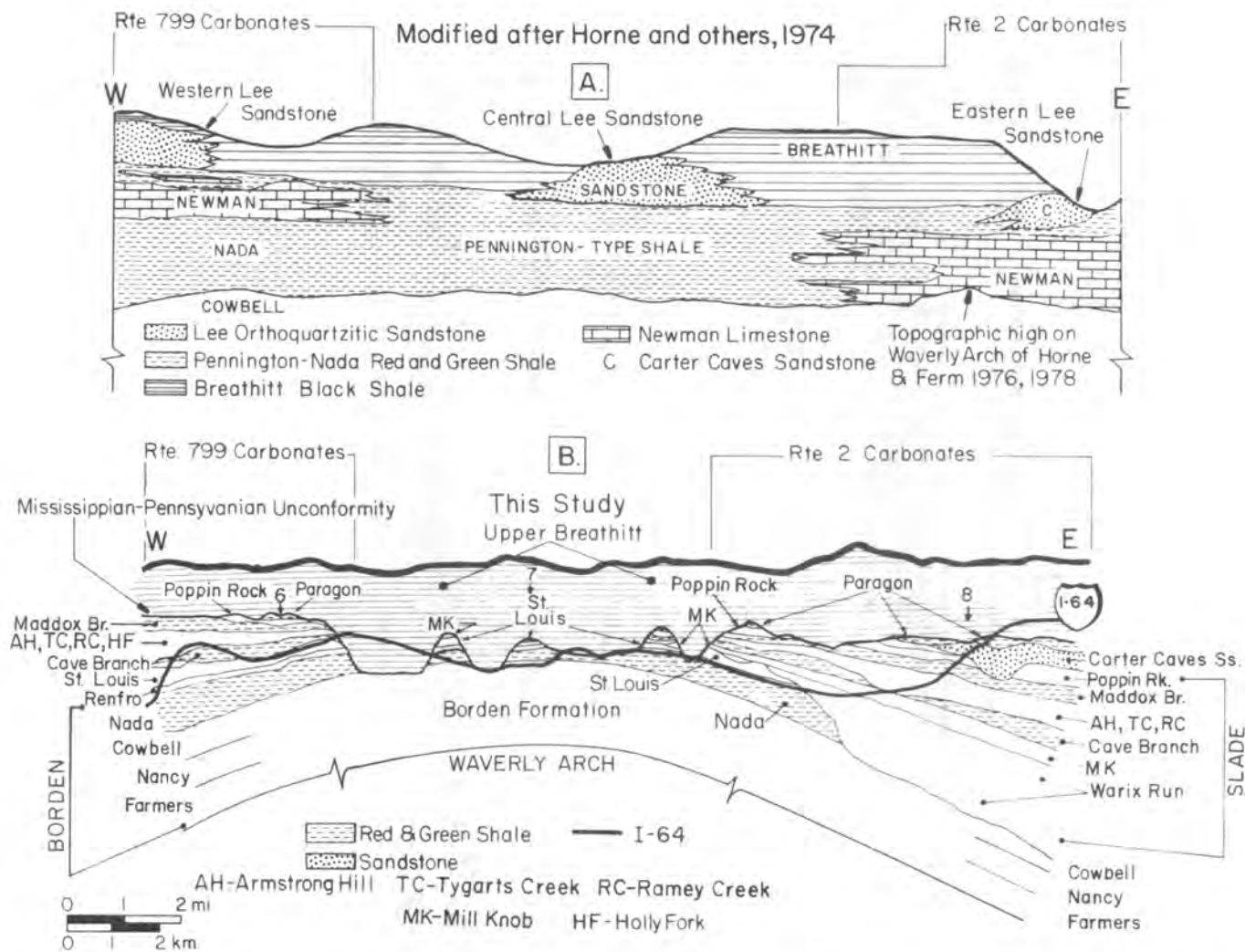


Figure 59. Comparison of Barrier-Shoreline (Fenn and others, 1971; using their nomenclature) (A) and "tabular-erosion" (B) interpretations of Carboniferous geology along a 17-mile segment of I-64 crossing the Slade outcrop belt. Numbers in B show approximate positions of stops. The two separate bodies of Newman Limestone (Slade Fm.) (Rte. 2 and 799 carbonates; Figs. 57, 58), interpreted to be isolated carbonate-barrier complexes in the Barrier-Shoreline model (A), more likely resulted from erosion through the Paragon (former Pennington) and Slade (former Newman) into the Borden on the crest of the Waverly arch; Breathitt sands and shales subsequently filled the erosional lows. In B, Pennsylvanian rocks are shown by horizontal lines at the top. True Lee sandstones were not deposited on or near the arch, but occur some distance to the east and west of it (e.g., Stop 9, Day 2) (modified from Ettensohn, 1980, 1981).

Formation or Carter Caves Sandstone, and that they overlie Maddox Branch shales on the flanks of the arch (Figs. 57B, 58B) and Nada shales on the crest of the arch (Fig. 59B). The disparity in ages between the sandstones and underlying shales suggests that the relationship is one of disconformity rather than gradation, and that a prominent mid-Carboniferous unconformity does exist.

Reevaluation of parts of the section along I-64 by Dever (1973, 1977, 1980), Ettensohn, (1975, 1977, 1980) and others shows most importantly through stratigraphy and biostratigraphy that the Slade-through-Breathitt succession is essentially one of superposition (Figs. 55, 59B) rather than facies (Figs. 53, 59A). Although the succession of units and fossils is not everywhere complete because of tectonically related erosion and nondeposition, the existing partial sections still exhibit basic chronologic and superpositional relationships. These same basic geological principles cannot be demonstrated in the Barrier-Shoreline Model.

TABULAR-EROSION MODEL

The Carboniferous rocks of east-central Kentucky represent the thinning edges of wedge-shaped sedimentary sequences formed as Carboniferous seas transgressed and clastic wedges prograded westwardly out of the Appalachian basin onto the craton. Analysis of regional stratigraphy, paleoenvironments, and paleogeography suggests that the Mississippian sequence in this area reflects three regional events (Fig. 55): 1.) a period of regional regression reflected by Borden deltaic progradation into the area during loading-type relaxation following the last tectophase of the Acadian orogeny (Figs. 28A, 55); 2.) a period of net regional transgression during which shallow-water carbonates became widespread reflecting the completion of loading-type relaxation, foreland-basin filling, and elevational equilibrium between the filled foreland basin and eroded uplands to the east (Figs. 28C, 55); and 3.) a period of regional regression during which marginal-marine to terrestrial sediments prograded into the area in response to unloading-type relaxation (Figs. 28D, 55). During each of these regional events, a series of more local transgressions and regressions (Fig. 55) appear to have deposited Borden, Slade, and Paragon members as more or less tabular units across eastern and central Kentucky. The many anomalies in this sequence in the form of disjunct unit distribution, abrupt facies changes, and intra-Mississippian unconformities largely reflect the intervention of synsedimentary tectonism on local structures.

The Borden Formation generally represents a westwardly prograding deltaic complex infilling a widespread Sunbury basin (Figs. 28A, 55). The lower two-thirds of the overlying Slade Formation, however, shows evidence of four major transgressive events, of which three were apparently ended by local tectonic or eustatic events (Fig. 55).

The earliest transgressive event is represented by the Renfro and St. Louis members (Fig. 55). The St. Louis thins toward the Waverly arch (Figs. 51, 59) and exhibits a prominent unconformity at its top, which becomes more pronounced toward regional structures, suggesting their influence in its formation. The overlying Ste. Genevieve, however, is restricted to either side of the Waverly arch and is found only on the southern downdropped blocks (Fig. 60B), again suggesting structural influence. Erosion accompanying the end of Ste. Genevieve deposition created major erosional lows which penetrated the St. Louis, Renfro, and underlying Borden on either side of the Waverly arch (Figs. 52, 56, 60C), and these lows became the loci of deposition for the succeeding Warix Run-Mill Knob continuum (Fig. 55). Although both the Warix Run and Mill Knob thin toward the

Waverly arch and are restricted to either side of it (Fig. 60C, D), the extremely widespread unconformity surface at the top of the continuum suggests the predominance of eustatic causes in the conclusion of Mill Knob deposition. The succeeding transgression represented by the Cave Branch-through-Maddox Branch members (Fig. 55) apparently managed to inundate the entire Waverly arch except for an area on its apex known as the "Apical Island." Not only do the upper Slade units thin near the Apical Island (Figs. 51, 59), but west of the apical island they show unusual facies developments, as well as fauna and flora that suggest restriction by the island. These facies will be examined at the Perry Branch section. More detailed information on the Tabular-Erosion Model is available in Dever (1977, 1980), Ettensohn and Dever (1979), and Ettensohn (1977, 1980, 1981, 1992).

PERRY BRANCH SECTION

The Slade section at this stop is 53-ft (16.2-m) thick and includes members from the St. Louis to the Poppin Rock, as well as part of the Paragon Formation (Fig. 61). This section is located high on the western flank of the Waverly arch (Figs. 51, 56, 59) and the carbonate sequence is thinner and somewhat different from those we will see on the eastern flank on the arch. The stratigraphic differences, however, are really minor and all can be explained in terms of structural influence by the Waverly arch. At this top, the St. Louis Limestone Member is fully developed and an uncommon unit, the Holly Fork Member is present; the Warix Run Member is absent but the Mill Knob is present as remnants filling microkarst on the St. Louis surface (Fig. 61).

The St. Louis Member of the Slade consists of 16 ft (4.9 m) of interbedded calcarenites and calcilitites exposed at and just above road level. This is one of the first localities west of the apical island where a complete St. Louis section is present along I-64, and this requires some explanation. East of the Waverly arch (e.g., Stop 8), the St. Louis is absent due to post-St. Genevieve erosion which formed deep erosional lows on the flanks of the arch. As previously discussed, this post-St. Genevieve erosion penetrated the St. Louis and upper parts of the Borden destroying most indications of their presence except for erosional remnants on adjacent highs (Figs. 52, 58, 59) and breccia zones of St. Louis chert in the Warix Run and Mill Knob members (Dever, 1973, 1977, 1980; Figs. 52, 58, 59, 60). Locally, the Mill Knob Member filled these erosional lows and overlapped onto and across the Waverly arch, giving rise to erosional remnants like those seen here (Fig. 61). Post-Mill Knob erosion on an early Chesterian unconformity (#3, Fig. 56) also probably destroyed much of the unit in this area. Because this and nearby exposures occur so high on the flank of the Waverly arch, the formation of deep erosional lows and the subsequent deposition of the Warix Run did not occur. Hence, the St. Louis is intact in this area (Figs. 51, 57, 60), but the Warix Run and much of the Mill Knob members are absent along the prominent intra-Mississippian disconformity atop the St. Louis. This disconformity is really a compound surface representing three major periods of exposure and erosion (post-St. Louis, post-St. Genevieve, and post-Mill Knob) between the deposition of the St. Louis and the deposition of the Cave Branch shale (Fig. 56). This surface is marked by brecciation, subaerial exposure crusts, and paleokarst. Southwest of this area, the Warix Run and Mill Knob are again present in erosional lows some distance west of the Waverly arch (Fig. 60C,D).

The subaerial exposure crusts and breccias atop the St. Louis abound in pedogenic features such as root traces, peds, horizon development,

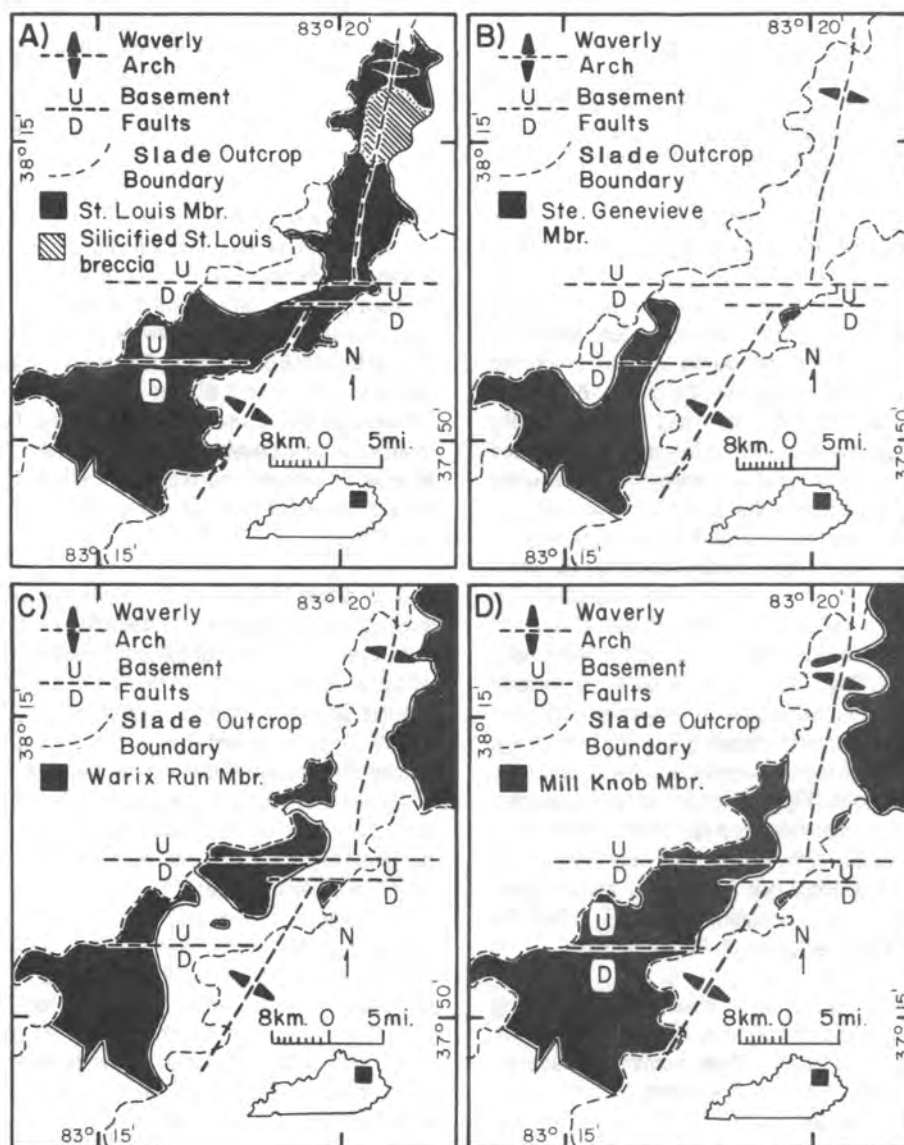


Figure 60. Disjunct distribution of the four lower Slade members in northeastern Kentucky reflecting control by structural features. Mapping based on 125 exposures throughout the area (modified from Ettensohn, 1980, 1981).

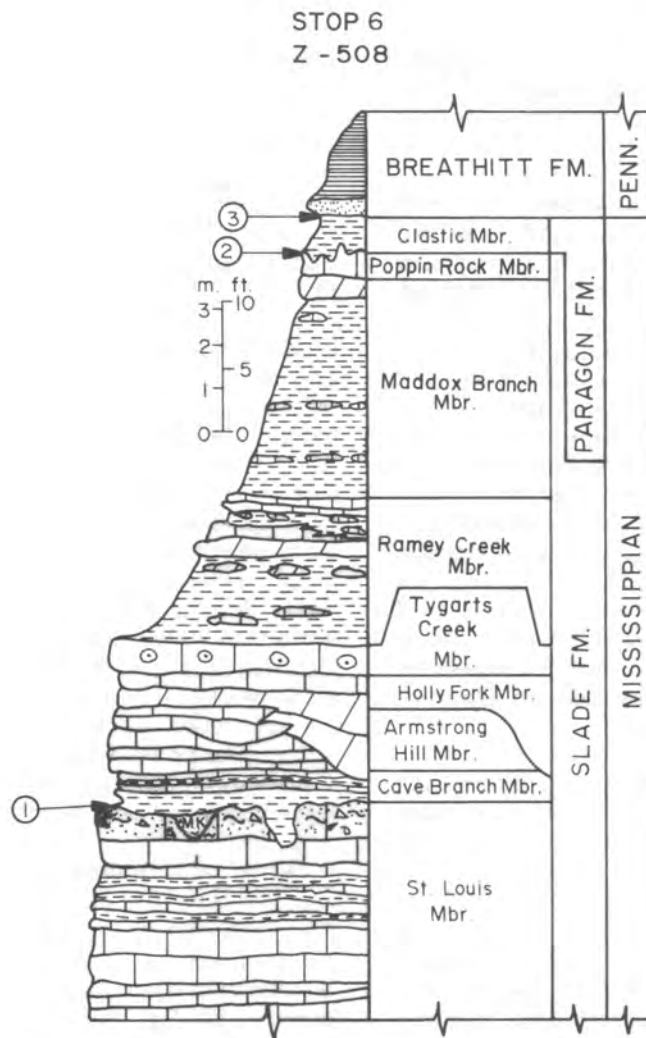


Figure 61. Stratigraphic section at Stop 6 on Day 2. Numbers refer to unconformities: 1) post-St. Louis, post-St. Genevieve, and post-Mill Knob; 2) early Paragon; 3) Early Pennsylvanian. MK is a Mill Knob remnant filling a sink hole on the St. Louis; this is only visible on the south side of the road. Many of the green shales in the upper part of the Slade were referred to the Pennington (Paragon) in the Barrier-Shoreline Model (Fig. 59a). Symbols as in Figure 58. (modified from Ettensohn, 1981).

caliche pseudo-anticlines, laminar calcretes, melanization, and terra rossa, all of which suggest that upper parts of the St. Louis were exposed and formed parts of thick soils (Ettensohn and others, 1988a; Ettensohn, 1990b). The presence of Mill Knob remnants infilling microkarst on the St. Louis surface (Fig. 61) suggests that at least two episodes of exposure and pedogenesis are represented by this surface, making it a composite soil (Ettensohn and others, 1988a, Fig. 18). Most of what is presently seen on the St. Louis surface probably represents the original Meramecian soil. It consists largely of thick accumulations of laminar calcrete or subaerial exposure crusts. These soils are called aggradational soils and formed in semi-arid climates where precipitation was never sufficient to transport dissolved carbonate out of the soil, resulting in net accumulation of pedogenic carbonate as thick calcretes.

The younger soil which developed on the early Chesterian Mill Knob Member is largely represented by the red to green residual claystones infilling karst at the base of the Cave Branch Member, although other caliche-type features are poorly developed in this soil. Such soils are degradational in nature and may form at the expense of the underlying stratigraphic unit. These soils are interpreted to have formed in humid, subtropical conditions during which precipitation was great enough to dissolve away most of the carbonate unit and flush it from the soil leaving behind karst and residual clayey terra-rossa soils. Aside from its normally thin nature on the flank of the Waverly arch, this probably explains why the Mill Knob occurs only as small remnants in this area. The timing of the climatic change reflected by these two superimposed soils agrees with paleogeographic reconstructions showing the movement of Kentucky into the tropics during the later Carboniferous. This same change is represented elsewhere in North America by the development of extensive drainage systems and coals beginning in the early Chester, and suggests the efficacy of using paleosols as paleoclimatic indicators (Ettensohn, 1990b).

Disconformably overlying the St. Louis is an intertonguing complex of Cave Branch, Armstrong Hill, and Holly Fork members exhibiting dolomite- and limestone-filled channels (Figs. 61, 62). The Holly Fork is present only on the west side of the apical island (Figs. 51, 59); it consists of dolomitic channel fill and a thin, overlying, bird's-eye calcilutite. The Cave Branch and Armstrong Hill members, moreover, contain a very restricted mollusk-algae biota unlike the typical biota of these units elsewhere. These three members reflect restricted tidal-flat and lagoonal environments that developed in protected areas west of the Waverly arch apical island (Fig. 63). The more typical environmental interpretation of these and overlying upper Slade units is shown on Figure 55.

The prominent white ledge at the level of the terrace is the Tygarts Creek Member. It is an oolitic calcarenite, 0.8- to 2.3-ft (0.2-0.7-m) thick. Overlying the Tygarts Creek is what appears to be a thick section of green shale. The basal nine feet (2.7 m) of the shale actually consists of interbedded shale, fossiliferous limestone, and dolostone which are included in the Ramey Creek Member (Figs. 61, 62). Much of the Ramey Creek is hidden by talus and vegetation growing on the terrace. The overlying 15 ft (4.6 m) of green shale constitutes the lower part of the Maddox Branch Member for which this is the type section (Ettensohn and others, 1984). Although green shale predominates, thin fossiliferous lenses of brecciated calcilutite occur throughout the entire section. A prominent carbonate ledge seems to separate underlying green shales from overlying dark shales (Figs. 61, 62). Close examination shows that the ledge is composed of two distinct carbonate units. The lower unit consists of 1.5 ft

(0.5 m) of dolomitic limestone and represents the upper part of the Maddox Branch (Fig. 61). The uppermost Slade unit consists of 1.5 to 3 ft (0.5-0.9 m) of eroded Poppin Rock calcarenite with a microkarstic surface. Two to three feet of sandy, grayish-green Paragon shales disconformably overlie the Paragon remnant.

Two intra-Mississippian unconformities and an Early Pennsylvanian unconformity are present in this exposure (Fig. 61). The lower unconformity atop the St. Louis is really a compound unconformity representing post-St. Louis, post-St. Genevieve, and post Mill Knob periods of erosion (#1, 2, 3, Fig. 56) near the apex of the Waverly arch. The second one is the late middle Chesterian disconformity separating the Slade (Poppin Rock Mbr.) from the Paragon (clastic member) on and near structural features. The upper unconformity is the regional Early Pennsylvanian systemic unconformity separating Mississippian and Pennsylvanian rocks. As we will see at Stop 7, the amount of erosional truncation along the systemic unconformity is much less on the flanks of the arch than on the crest itself (Figs. 51, 59).

In the Barrier-Shoreline Model, this locality was used to characterize the "Rte. 799 carbonates." The "Rte. 799 carbonates" were interpreted to be a carbonate-barrier complex younger than carbonates to the east ("Rte. 2 carbonates") and completely surrounded by Pennington (Paragon) and Nada red and green shales (Figs. 57A, 59A; Horne and others, 1974, fig. 5B). Although these carbonates do overlie Nada red and green shales, the red and green shales overlying the carbonate section are not Pennington (Paragon) shales. Rather these shales are part of the Ramey Creek and Maddox Branch members (Figs. 57B, 59, 61, 62). Moreover, the Ramey Creek Member at this locality contains the middle Chesterian (Golconda) guide fossils *Pterotocrinus lingulaformus*, and *P. armatus*, and hence is too old to be Pennington (Paragon). In addition, the red and green shales are overlain by a Poppin Rock (Glen Dean) remnant. As the Pennington (Paragon) overlies the Glen Dean (Poppin Rock) by definition (Butts, 1922), those shales underlying the Poppin Rock cannot be the Pennington (Paragon) Formation.

Finally, no evidence supports the easterly intertonguing and termination of the "Rte. 799 carbonates" against a body of red and green Pennington (Paragon) shale as indicated in the Barrier-Shoreline Model (Fig. 57A). The eastern most outcrop of the "Rte. 799 carbonates" is truncated by a Pennsylvanian channel fill of black Breathitt shale (Figs. 51, 57A, 59A). Although the "Rte. 799 carbonates" thin eastwardly as they lap onto the Waverly arch, and although some units (Warix Run, Mill Knob, and Holly Fork; Fig. 59) are absent on the arch or on one side of it because of tectonically related patterns of erosion or deposition, the distribution patterns of most Slade members (Figs. 51, 52) reflect widespread deposition throughout the area and across all parts of the Waverly arch except on the apical island (Ettensohn, 1980, 1981).

GENERAL PENNSYLVANIAN PALEOGEOGRAPHIC, PALEOCLIMATIC, AND TECTONIC FRAMEWORK FOR EASTERN KENTUCKY

DONALD R. CHESNUT, JR.

PALEOGEOGRAPHIC POSITION AND CHANGING CLIMATES

Plate reconstructions by Scotese (1990) show that the Pennsylvanian North America continental block was rotated clockwise about 45°

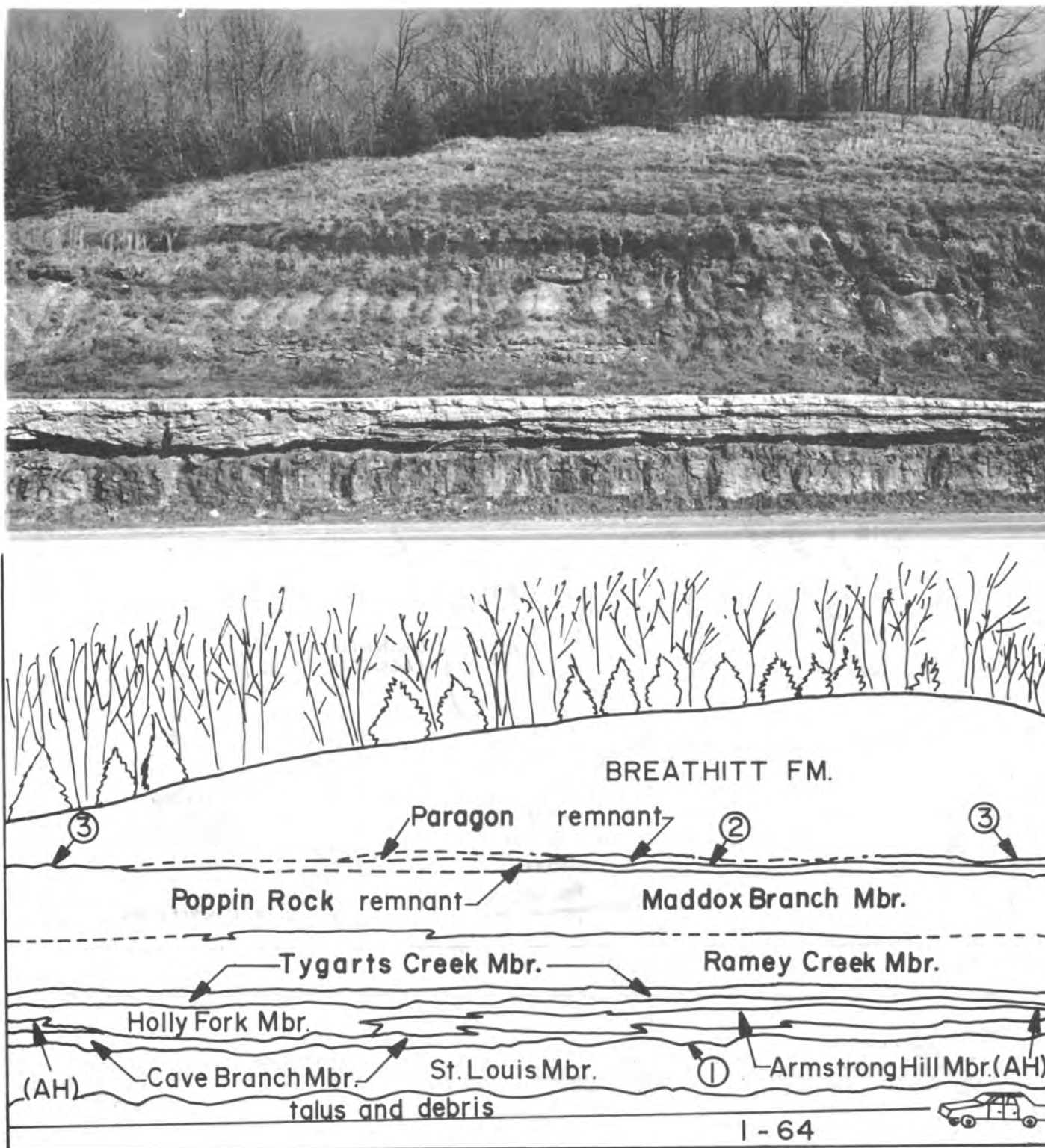


Figure 62. Photograph and diagrammatic key showing the disposition of units exposed at Stop 6, Day 2. The numbers refer to unconformities shown in Figure 61 (modified from Ettensohn, 1981).

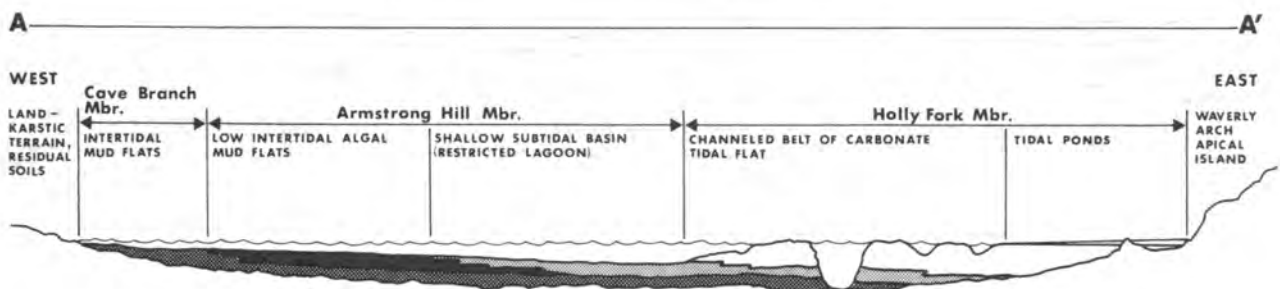
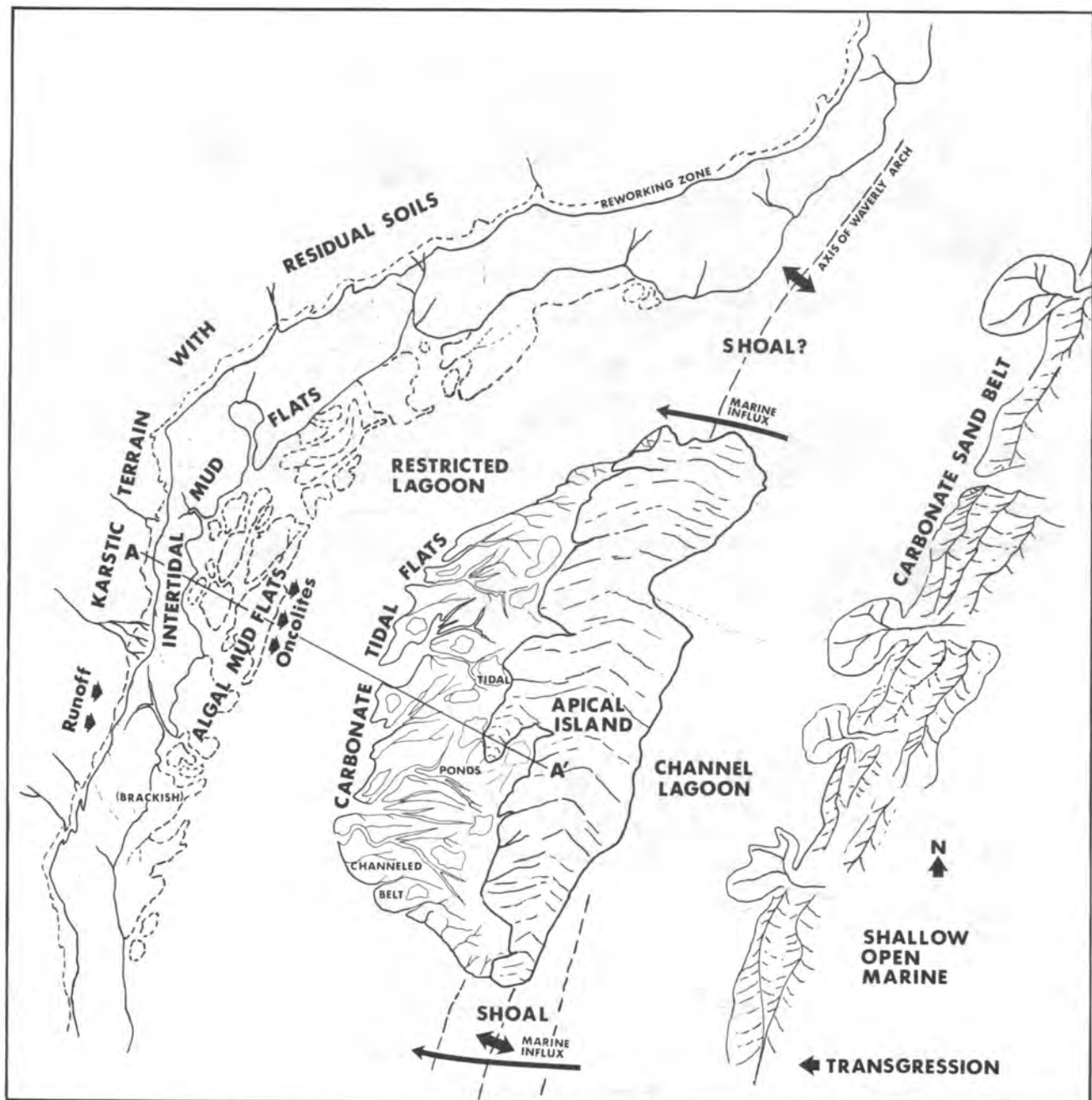


Figure 63. Reconstruction of restricted environments west of the Waverly arch apical island during the early Chesterian, based on exposures Z-505, 506, 507, 508, 509, 510 and 511 along I-64 (see Fig. 51). Rocks deposited in these restricted environments will be examined at Stop 6, Day 2 (from Ettensohn, 1980).

relative to its present position. During the Namurian (Late Mississippian-Early Pennsylvanian) Scotese located Kentucky about 7° south of the equator, whereas he showed Kentucky about 7° north of the equator by Kazanian time (Late Permian). Obviously this reflects the drift of North America from south to north during intermediate stages. Cecil (1990) suggested that this northward drift across climatic zones is reflected in the Mississippian and Pennsylvanian rocks of the Appalachian basin. According to Cecil, the eastern United States shifted from semi-arid and wet-dry tropical conditions of Late Mississippian time (reflected by the presence of carbonates, evaporitic units, vertisols, and aridisols) to wet-dry seasonal, tropical and subtropical rainy climates during the Early and Middle Pennsylvanian (indicated by laterally extensive coal beds, leached paleosols, and abundant siliciclastics), and finally to humid-subtropical and arid conditions during the Late Pennsylvanian (represented by the occurrence of carbonates, vertisols, and aridisols). Paleofloral studies by Phillips and others (1985) show similar trends with the exception of an additional dry period during the Westphalian B-C (early Middle Pennsylvanian). Cecil attributed the shift from semi-arid (Late Mississippian) to tropical rainy (Early and Middle Pennsylvanian) and back to semi-arid conditions (Late Pennsylvanian) to the drift of eastern North America from arid-tropical southern latitudes across the humid intertropical convergence zone into northern arid-tropical latitudes.

The important coal-bearing rocks of eastern Kentucky (the Breathitt Formation) contain numerous laterally extensive, high-quality bituminous coal beds. Present studies suggest that the major coal beds of the Breathitt Formation were formed as rain-fed (ombrogenous), domed peat swamps analogous to the coastal peat swamps of modern Indonesia (e.g., Cecil and others, 1985). In both modern and Pennsylvanian cases, abundant rainfall was provided by the humid intertropical convergence zone and tropical monsoonal wind patterns. Tradewind patterns (lower Hadley cell) over the Pennsylvanian Appalachian basin would have been from the east-by-southeast (present configuration) during the summer, and the basin would have been in the rain shadow of the Appalachian mountains. Tradewinds were from the north during the winter and were probably very dry, coming from a subtropical high in what is now northern Canada. Wet monsoonal winds may have come from the south or south-by-southwest during the spring, summer, and fall. The humid intertropical convergence zone was apparently situated over the Appalachian basin in the fall and spring. Upper-level wind patterns, responsible for distribution of volcanic ash, are reported in Chesnut (1985).

Tectonically Linked Basin Models

The nature of Alleghanian (Late Paleozoic) tectonism is poorly constrained chronologically and tectonically (Chesnut, 1991a). Continent-continent collision between North America (Laurussia) and Africa (Gondwana) is generally accepted by most geologists. However, the timing of this collision is controversial; according to some models, collision occurred during Late Pennsylvanian-to-Permian time (Rast, 1988; Chesnut, 1991a). Other models support collision during mid-Carboniferous time (earliest Pennsylvanian) (Ettensohn and Chesnut, 1987, 1989; Ettensohn, 1990c). Even the direction of subduction between Laurussia and Gondwana is controversial (see Secor and others, 1986 for various subduction models). Sinha and Zietz (1982) and Chesnut (1991a) suggested a westward-dipping subduction zone beneath Laurussia, whereas Pindell and Dewey (1982) indicated eastward subduction beneath Gondwana. Because of problems in constraining the chronology and

tectonics of the Alleghanian orogeny, widely varying tectonically linked basin models have been developed for the Carboniferous rocks of the Appalachian basin. For instance, Ettensohn and Chesnut (1987, 1989) and Ettensohn (1990c) attributed the regional Early Pennsylvanian unconformity in the Appalachian basin to uplift (creation of forebulge or peripheral bulge) caused by collisional docking of Gondwana against Laurussia. Onlapping of subsequent Pennsylvanian strata to the northwest was caused by migration of the forebulge and basin to the northwest. A contrasting scenario suggests that the regional unconformity was caused by visco-elastic relaxation of crustal stress (Chesnut, 1991a), whereas onlap of the unconformity was caused by migration of the forebulge to the northwest. In the latter model, subduction-zone plutonism and thrusting provided the load which caused the forebulge and basin to migrate. In this model continent-continent collision would have occurred during Late Pennsylvanian or Permian time. At present, insufficient evidence has been collected to thoroughly test these or other models for the basin.

STRATIGRAPHIC FRAMEWORK

Mississippian-Pennsylvanian Transition

The contact between the Upper Mississippian and Lower Pennsylvanian may be conformable in the deepest part of the Appalachian basin in parts of Virginia and southern West Virginia (Englund and others, 1979). Where conformable, the largely marine sandstones and shales of the Pennington Group apparently grade into important coal-bearing sandstones and shales of the overlying Pocahontas Formation (Fig. 64). However, in most of the basin, the Pocahontas Formation is absent and later Lower and Middle Pennsylvanian strata unconformably overlie Mississippian strata. To the northwest toward the margin of the Appalachian basin, sequentially younger Pennsylvanian units unconformably onlap progressively older Mississippian strata. In the field-trip area along I-64, lower Middle Pennsylvanian strata unconformably overlie strata as old as the Lower Mississippian Borden Formation (Figs. 51, 54, 59b) (e.g., Stop 7). Across most of the basin, Pennsylvanian rocks unconformably overlie Mississippian rocks so that the unconformity became known as the Mississippian-Pennsylvanian unconformity. However, regional relationships show that the major development of the unconformity actually occurred during the Early Pennsylvanian after the Pocahontas Formation was deposited, because the Pocahontas is truncated by the unconformity (Fig. 64) (Englund, 1979; Englund and others, 1979; Ettensohn and Chesnut, 1987, 1989; Chesnut, 1988). The unconformity is more appropriately called the Early Pennsylvanian unconformity. The cause of the unconformity has been attributed to worldwide eustatic sea-level changes (Saunders and Ramsbottom, 1986), tectonic uplift generated by continent-continent collision (e.g., Ettensohn, 1990c), or tectonic uplift generated by relaxation of stress (e.g., Chesnut, 1991a). A eustatic lowering of sea level at the Mississippian-Pennsylvanian boundary has been discounted because the unconformity was shown to have formed after Saunders and Ramsbottom's worldwide sea-level drop (Chesnut, 1988). The two tectonic scenarios are described in the preceding section.

Lee and Breathitt Formations

The Lower and Middle Pennsylvanian strata of the Eastern Kentucky Coal Field (part of the central Appalachian basin) comprise the Breathitt and Lee formations (Fig. 64). The Breathitt Formation is composed of shales, siltstones, sandstones, and numerous economically important bituminous-coal beds. Examples of the

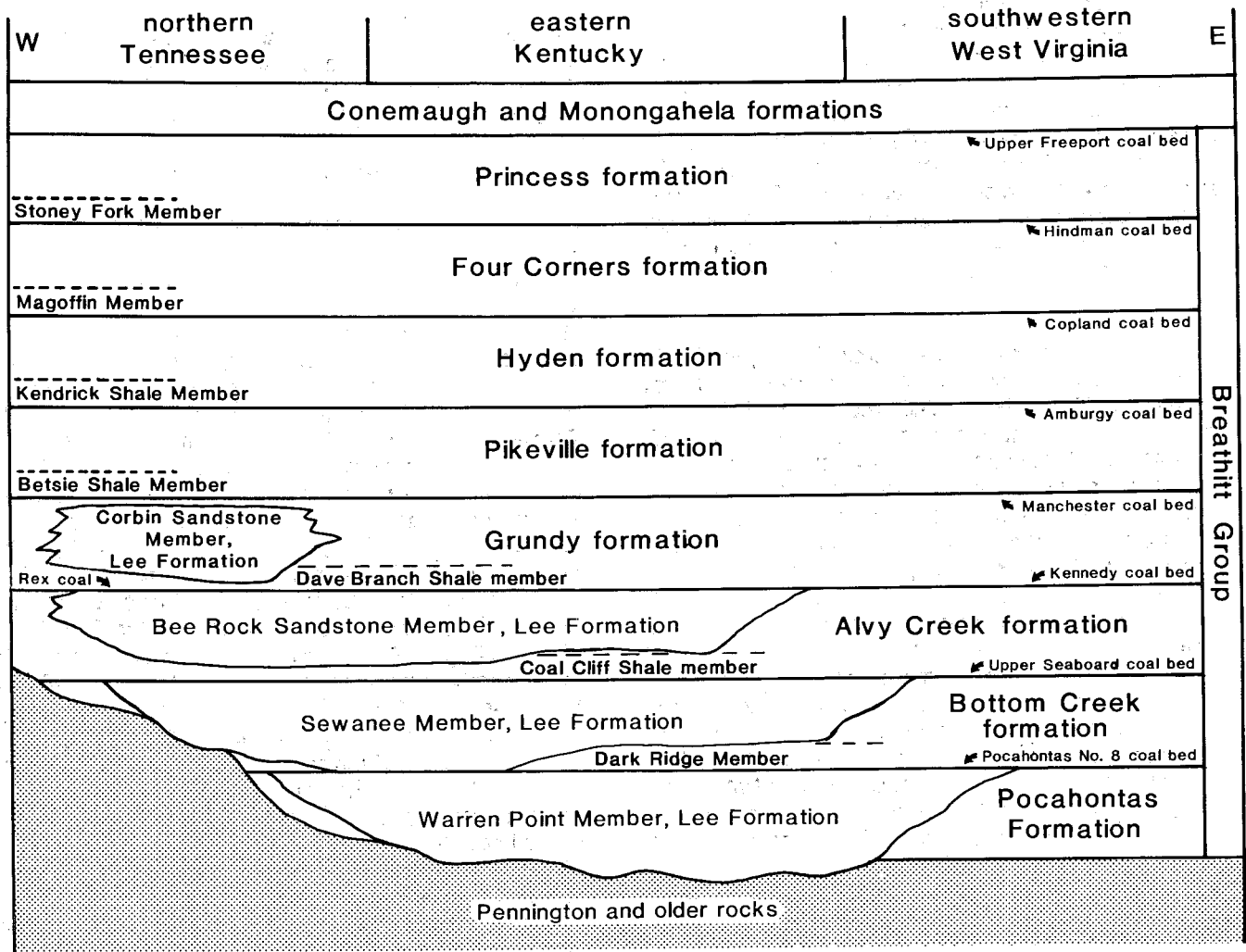


Figure 64. Generalized stratigraphic framework showing the regional relationship between members of the Lee Formation and the Breathitt Group (from Chesnut, 1988). The Breathitt is divided into eight informal formations.

Breathitt can be observed at Stops 6 and 7 on Day 2 and at Optional Stop A at Gregoryville on Day 3. The sandstones of the Breathitt are typically immature, commonly argillaceous, and contain mica, feldspars, and lithic fragments. The lower part of the Breathitt contains four lens-shaped sandstone bodies or members of the Lee Formation (Fig. 64). The Lee sandstones are thick (50-200 ft; 15-60 m), massive, cliff-forming, sublitharenites to quartz arenites. The Lee sandstones commonly are coarser than Breathitt sandstones and may contain abundant quartz pebbles. The Corbin Sandstone of the Lee Formation will be examined at Stop 9 near Grahn. The Lee occurs in at least two forms: 1.) as narrowly confined, channel fills at the Early Pennsylvanian unconformity (minor occurrences such as the Livingston Conglomerate), and 2.) the volumetrically more important sandbelt lenses like the Corbin Sandstone (e.g., Stop 9, Grahn, KY). Regionally, the four members of the Lee Formation (Fig. 64) form four large sandbelts (35-55-mi or 55-90-km wide) oriented northeast-southwest (Chesnut, 1988). These sandbelts onlap the regional Early Pennsylvanian unconformity to the northwest toward the Cincinnati arch (Fig. 3) where they overlap older strata. For example, the Corbin (lower Middle Pennsylvanian) overlaps the Paragon and Slade formations (Late Mississippian) in northeastern Kentucky.

The intervening Breathitt sandstones and shales were deposited as a clastic wedge derived from the early Appalachian mountains to the east (Fig. 65). Crossbed directions and channel forms indicate that fluvial flow and sediment transport was to the west. The Lee sandbelts are interpreted to represent fluvial-trunk transport systems situated between a mid-Carboniferous forebulge to the northwest (represented by onlap of the unconformity) and the Breathitt clastic wedge to the southeast (Chesnut, 1988). A compilation of crossbed directions indicates that flow within members of the Lee was dominantly to the southwest, parallel to the orientation of the sandbelts which is typical of fluvial systems (Chesnut, 1988). Vertical and lateral profiles, well developed pebble lags, bedding architecture, thinning-upward stratification, and general fining-upward grain-size trends (rather than coarsening-upward) also are typical of fluvial environments (Walker, 1984; Miall, 1985; Reineck and Singh, 1986). This combination of sedimentological criteria has been used by several authors to interpret the Lee as a fluvial sandstone in many parts of the basin (BeMent, 1976; Greb and Chesnut, 1989; Hester, 1977; Rice, 1984; Wizevich, 1991). Marine indicators such as bioturbation, herringbone crossbedding, and megaripple-style cross-bedding, if they occur at all, are mostly restricted to the uppermost part of the named members, where they may record estuarine reworking of fluvial sandstones (e.g., Greb and Chesnut, 1989).

A popular model developed a little more than twenty years ago is the Lee-Newman (Slade)-Barrier-Shoreline model of Horne, and others (1974). In the barrier-shoreline model, the Lee sandstones were interpreted to represent beach or barrier-bar sands based on quartzose-sand content. Subsequently, numerous paleontologic, stratigraphic and sedimentologic analyses have discounted this interpretation (see Chesnut, 1988, for a compilation).

Trends in the Pennsylvanian Coal-Bearing Rocks

Several trends have been observed in the Lower and Middle Pennsylvanian coal-bearing rocks of the Eastern Kentucky Coal Field (Chesnut, 1989a, 1989b, 1991b, 1991c). The smallest scale trend (discussed here) is the Appalachian cyclothem or (mesocyclothem), which includes a major coal bed and an overlying shale and sandstone (up to the next major coal bed). Paleontological analyses indicate that marine strata overly every major coal bed (Chesnut,

1981, 1991d). The marine strata are generally truncated and overlain by fining-upward fluvial sandstones, which, in turn, are overlain by rooted underclay (or other "seatrock") and finally by an overlying coal bed. Chronologic analysis based on ten different chronologic schemes provides an average interval of 430 Ka for each of these cyclothem. The 430 Ka cycle corresponds with a known 430 Ka glacio-eustatic cycle that is modulated by a second-order Earth orbital eccentricity cycle. The similarity between the two cycles supports a glacio-eustatic origin for these cyclothem. In contrast, Tankard (1986a, 1986b) and Klein and Kupperman (1992) have proposed a tectonic origin for these Early and Middle Pennsylvanian cyclothem.

The modern analog to the Appalachian cyclothem may be represented by the Indonesian coastal setting (Cobb and others, 1989; Chesnut and others, 1991). Major Breathitt coal beds, overlain by marine strata in many areas, were probably formed as coastal peat swamps at lowstand which were subsequently flooded by transgressing seas, similar to the situation in Indonesia.

An intermediate-scale cycle, punctuated by several thick and widely extensive marine members, is represented by eight informal coal-bearing units (equivalent to the informal formations; Fig. 64), each comprising an average of six cyclothem. Chronologic analysis indicates an average interval of 2.5 Ma for this cycle. The origin for this cycle is unknown because the cycle interval does not conform to any widely recognized glacio-eustatic or tectonic cycle.

The largest of the trends recognized here is a general coarsening upward of the Breathitt Formation (i.e., an upward increase in the sandstone-to-shale ratio). This coarsening-upward trend is attributed to the increasing intensity and proximity of the Alleghanian orogeny.

Transgressive and Regressive Depositional Systems

Based upon the distribution of Pennsylvanian strata and invertebrate biofacies trends, three transgressive and regressive systems tracts can be inferred for the eastern United States (Chesnut, 1991b, 1991c, 1991e). The distribution of strata post-dating the regional unconformity shows that the Early and early Middle Pennsylvanian foreland basin sediments did not crest the forebulge (Fig. 65). Alluvial deposits of the Breathitt Formation derived from the Appalachian highlands were transported to the northwest toward the forebulge. During lowstand, the only outlet available to further sediment transport (represented by the Lee sandbelts) was toward the southwest (Ouachita trough) along the Black Warrior-Appalachian foreland basins. This transport was periodically interrupted by transgression moving northeastward from the Ouachita trough into the Alleghanian foreland basin. The middle and late Middle Pennsylvanian marks a period of intermittent cresting of the forebulge by westward sediment transport (Breathitt Formation). Marine transgressions, reflected by marine strata, such as the Betsie, Kendrick, and Magoffin members of the Breathitt Formation (Fig. 64), entered through the foreland basin and across saddles in the forebulge. During the Late Pennsylvanian, time represented by the Conemaugh and Monongahela formations, marine waters entered the Appalachian basin only across the diminished forebulge north of the Jessamine dome. By this time the Ouachita trough apparently had been closed by collision with Gondwana, and access of marine waters through the foreland basin was blocked.

89.3 0.3 Entrance to weigh station, Z-509 (Fig. 51); Nada Member of the Borden and Renfro-through-Maddox Branch members of the Slade Formation are present on

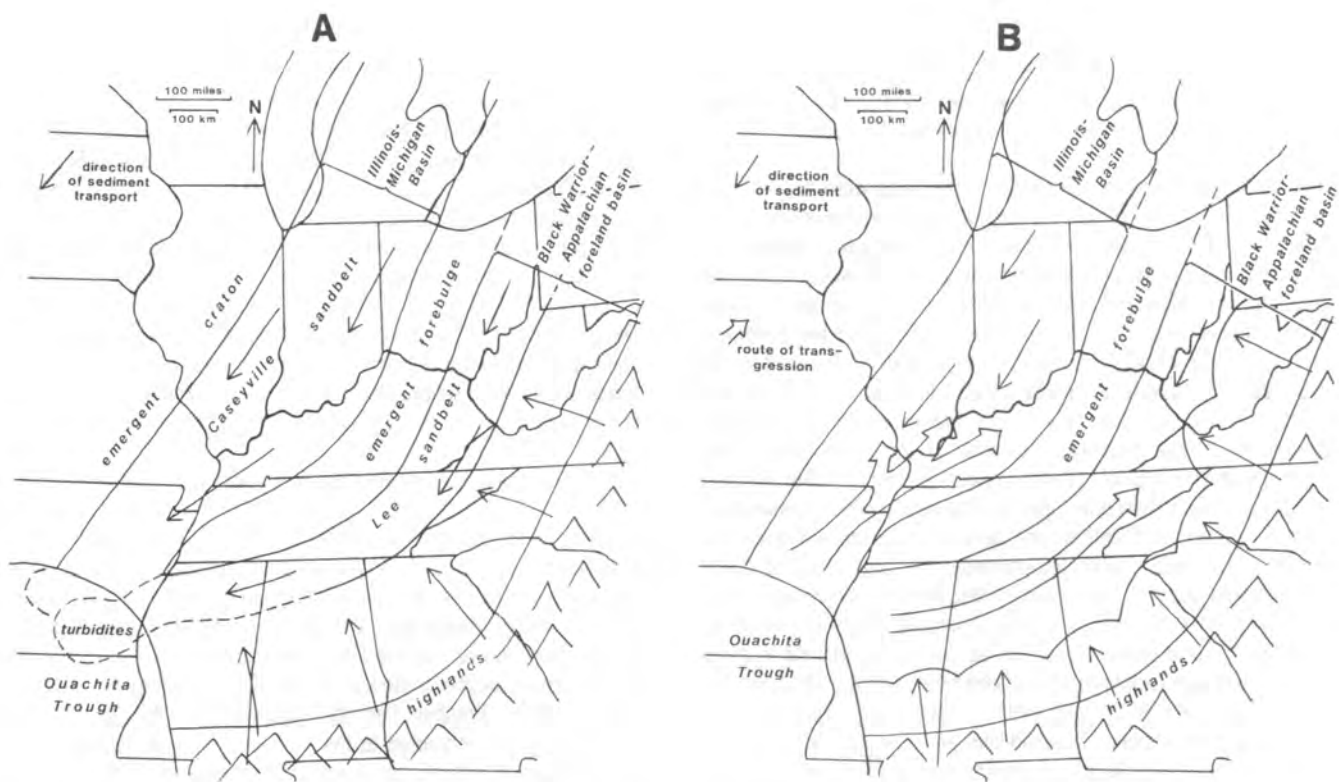


Figure 65. Reconstructions of depositional systems during Early (A) and early Middle (B) Pennsylvanian lowstands and highstands in east-central United States.

both sides of the highway. Breathitt sands and shales disconformably overlie the Ramey Creek and Maddox Branch members. This exposure exhibits the first complete St. Louis section west of the Waverly arch; the Warix Run and Mill Knob members are absent along the prominent intra-Mississippian unconformity atop the St. Louis (Fig. 56). The Cave Branch is the prominent red shale forming the reentrant on top of the St. Louis, and its basal mudstones are probably terra rossas developed from early Chesterian weathering of the Mill Knob Member (Ettensohn and others, 1988a).

- 89.8 0.5 Exposure Z-510 (Fig. 51); Tygarts Creek-through-Poppin Rock members poorly exposed behind entrance ramp on left-hand side of road.
- 90.0 0.2 Exposure Z-511 (Figs. 51, 57B); dark Breathitt shales in channel truncating Slade Formation. Holly Fork-through-Maddox Branch members of the Slade are poorly exposed in the cut. This is the eastern-most exposure of the Slade carbonates in the "Rte. 799 carbonates" of Horne and others (1974).
- 90.5 0.5 Carter County line; cyclic shale and siltstone laminae in the Breathitt Formation truncated by a channel on the east end of the exposure.
- 90.9 0.4 Milepost 149; begin a series of outcrops containing bioturbated Breathitt siltstones and shales with prominent channels. Siltstones exhibit cyclic fining-upward sequences and are described by Short (1979a).
- 91.4 0.5 Exposures Z-512 and Z-513 (Fig. 51); exposure Z-512 is about 1500 ft north of the westbound lane and exhibits the Nada-through-Cave Branch members. Exposure Z-513 is present along the eastbound lane and exhibits the same section as does Z-512, except that the carbonates are hidden by slumped Cave Branch shales and the entire sequence is overlain by a Pennsylvanian orthoquartzitic sandstone. Mill Knob and brecciated St. Louis are present below the shales, and this exposure probably marks the western limit of the Waverly arch apical island (Fig. 51). Shales identified in this exposure by Ettensohn (1981) as Hardinsburg Member (Maddox Branch) most likely belong to the Cave Branch Member.
- 91.7 0.3 Cowbell and Nada members of the Borden Formation.
- 92.0 0.3 Bridge over Fleming Fork.
- 92.7 0.7 STOP 7. Near the Apex of the Waverly Arch Apical Island (Z-514; 3.2 mi east of the weigh station on the eastbound lane of I-64, Carter Co.; Carter Coordinates: 300°FSLx1200°FWL, 8-V-75) (Figs. 51, 59).

NEAR THE APEX OF THE WAVERLY ARCH APICAL ISLAND

FRANK R. ETTENSOHN

This stop is located on the former apical island high on the western flank of the Waverly arch (Fig. 51). Although close to the apex

(Mile 93.5, Day 2), the presence of the lowermost Slade (Renfro Mbr. and brecciated St. Louis) and a Paragon remnant (Fig. 66B) indicates that we have "descended" from the apex to a slightly lower level where erosional truncation was less. The section begins with approximately 65 ft (19.8 m) of the Borden siltstones, shales, and dolostone (Fig. 66B). The green Cowbell siltstones and interbedded shale are sparsely fossiliferous but show abundant evidence of bioturbation. Both horizontal and vertical burrows are present, but the "rooster-tail" trace fossil known as *Taonurus* or *Zoophycus* is most abundant. Overlying red and green Nada shales are also highly bioturbated and contain calcareous lenses with abundant crinoid fragments and large spiriferoid and productoid brachiopods including the Osagean guide fossils: "*Spirifer*" *mortonanus*, "*S.*" *montgomeryensis*, and *Marginirugus magnus*.

The Slade Formation has been reduced in thickness from 53 ft (16.2 m) at Stop 6 to west to 6.5 ft (2.0 m) or less here on the flank of the arch. The Slade consists of 3.5 ft (1.1 m) of brown, massive Renfro dolostone and three feet (0.9 m) or less of silicified St. Louis breccia. The St. Louis breccia is silicified, but still reveals typical St. Louis fossils replaced by chert. Small pods of laminated calcilutite, now silicified, fill shallow depressions on the surface of the breccia and may represent Mill Knob remnants. Both the Renfro and St. Louis breccia are poorly exposed in the vegetation midway up the outcrop. Other Slade units are absent here, but a comparison of the stratigraphic section at this locality (Fig. 66B) with that at Stops 6 and 8 (Figs. 51, 52, 59) and those at other outcrops and stops to the west and east, shows a progressive thickening of the Slade away from the crest of the arch.

However, the absence of other Slade units at this locality indicates another stage of uplift at the end of the middle Chesterian, reflected by a disconformity separating the Slade and Paragon at this and other structural features in the area. This disconformity separates a thin bed of sandy, red and green Paragon shale from the Slade (brecciated St. Louis, Fig. 66B) at this stop. A similar intra-Mississippian disconformity between the Slade and Paragon is present at Stop 6 (Fig. 61). As previously discussed, this disconformity represents a period of late middle Chesterian uplift. On both the Waverly arch and basement fault zone, parts of the upper Slade and lower Paragon are absent along this disconformity (Fig. 56). This is the only period of uplift which has been dated both by stratigraphic position and palynology (Ettensohn and Peppers, 1979).

The major disconformity at this stop, however, is the one separating Mississippian and Pennsylvanian rocks. The lowermost Breathitt sandstones, shales, and clays fill a broad, low, channel disconformably overlying the Paragon and Slade at the channel edges and the Borden (Nada) in the center of the channel (Fig. 66B).

Above parts of the disconformity here and elsewhere, a prominent clay bed known as the Olive Hill Clay Bed of Crider (1913) is present. This clay bed was mined from a pit at the level of the second bench on the south side of the highway (Philleary and others, 1975). This is equivalent to the second fireclay shown in Figure 66B, although the fireclay on the south side of the road is thicker. On both sides of the road the fireclay is overlain by a thin coal or carbonaceous shale.

Fireclays like this typically occur near the base of the Pennsylvanian section in northeastern Kentucky and are included in the lower tongue of the Breathitt. Although the fireclays have a very localized, patchy distribution, they occur so persistently in lowermost parts of the

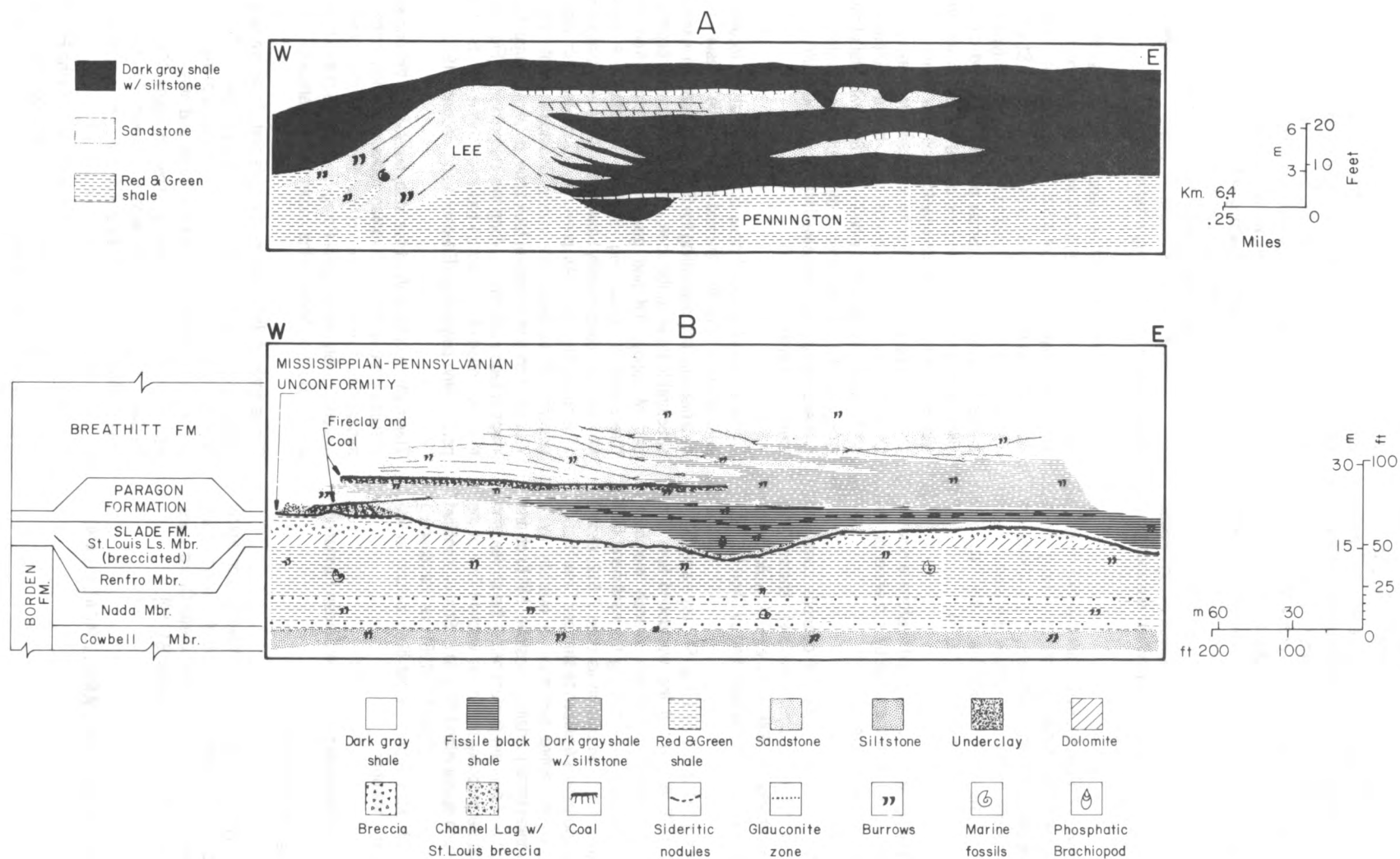


Figure 66. Comparison of Barrier-Shoreline interpretation (A), modified from "Central Lee Sandstone" of Horne and others (1974) using their nomenclature, and a "tabular-erosion" interpretation (B) of the stratigraphy at Stop 7, Day 2, high on the west flank of the Waverly arch (modified from Ettensohn, 1980, 1981).

Pennsylvanian on or near the Early Pennsylvanian unconformity that they are collectively called the Olive Hill Clay Bed of Crider (1913).

The distribution of the clay provides some interesting insights into its formation. The clay is almost wholly restricted to the northern uplifted fault block and seems to be more concentrated on and near the apical island (Fig. 67). The clay is commonly restricted to erosional lows or paleokarst on the Borden, Slade, Paragon, or Early Pennsylvanian formations near the mid-Carboniferous unconformity, resulting in a patchy distribution (Fig. 67). After Early Pennsylvanian regional uplift, the whole region was subjected to a period of subaerial exposure and erosion, resulting in the very irregular disconformity surface. Both the Waverly arch and basement fault zone seem to have undergone renewed uplift at this time. The elevated state of the northern uplifted block accounts for the deeper erosion here along the disconformity, and, as indicated by Short (1978, 1979b), prevented subsequent, large-scale, deltaic encroachment into the area during the Early Pennsylvanian. Although uplifted, the eroded surface was apparently covered by a very shallow brackish or marine bay in which suspended clays and silt could be carried over the surface and washed into the erosional lows. The erosion lows eventually accumulated sufficient thicknesses of clay to support marsh and swamp vegetation at or slightly above sea level. The thin coals commonly overlying the fireclays are evidence for these swamps. The overlying swamps provided the decreased circulation and increased acidity necessary for alteration of the underlying clays into kaolinite-rich fireclays through leaching and upward diffusion (Keller and others, 1954; Patterson and Hosterman, 1961). The positive nature of the northern uplifted block and Waverly arch served to enhance terrestrial swamp growth, as well as the weathering processes which altered the clays (see Keller and others, 1954, for a similar example), and explains the thicker, better developed, and more extensive fireclays which occur on and near the apical island (Fig. 67).

One of the prominent features in this exposure is an erosional low that cuts through the Paragon, St. Louis breccia, and Renfro to the level of the Nada shale (Fig. 66B). A thin basal sand, locally containing fragments of St. Louis breccia, lines the base of the low. Fissile black shales containing sideritic nodules and *Lingula* fill most of the erosional low and grade upward into dark-gray silty shales with thin siltstone laminae. At the west end of the cut, a thin coal and underclay (Olive Hill Clay Bed) separate the silty shales below from a fine-grained, carbonaceous, orthoquartzitic sandstone which pinches out eastwardly into dark silty shales. The sandstone is extremely bioturbated. Fern and others (1971) and Horne and others (1974) interpreted this sandstone to represent the landward side of the beach-barrier system to the west, whereas Short (1978, 1979b) interpreted the sandstone to represent a splay deposit associated with a distributary channel to the west.

This locality is also critical in evaluation the Barrier-Shoreline Model, for this is one spot where a "Lee" sandstone is interpreted to intertongue with Pennsylvanian Breathitt shales to the east and Mississippian Pennington (Paragon) shales to the west (Fig. 66A). This is the "Central Lee Sandstone" locality of Horne and others (1974, fig. 3B; this guide, Figs. 66A, 68B). The "Central Lee Sandstone" is shown both intertonguing with and overlying Pennington red and green shales (Fig. 66A). However, based on fossil content already mentioned and mapping by Philley and others (1975), the red and green shales underlying the "Central Lee" are not Pennington shales, but are pre-Pennington, Middle Mississippian (Osagean) shales belong to the Nada and Cowbell members of the Borden. A Pennington or Paragon designation for these rocks is wholly

unwarranted, because wherever unequivocally recorded, the Pennington (Paragon) formation of eastern Kentucky and adjacent states is late Chesterian or latest Mississippian in age (Butts, 1922, 1940; Weller and others, 1948; Rice and others, 1979; Patchen and others, 1985; Sable and Dever, 1990). Moreover, the sandstone itself is not even a Lee sandstone, but a small sand body in the Breathitt (Philley and others, 1975; Ettensohn, 1980).

Although the sandstone does intertongue with dark Breathitt shales, there is no evidence of intertonguing with the red and green shales. In fact, it would be very unlikely for a Middle Pennsylvanian sandstone to intertongue with Middle Mississippian red and green shales. All evidence points to the fact that Middle Pennsylvanian shales and sandstones of the Breathitt Formation overlie Upper and Middle Mississippian rocks of the Paragon, Slade, and Borden formations with no lateral or vertical gradation; this can only reflect an unconformity separating Mississippian and Pennsylvanian rocks (Fig. 66B; see Ettensohn, 1980). The extremely thin nature of the Slade and Paragon at this section is related to the great amount of erosional truncation on the unconformities near the crest of the Waverly arch. A comparison of stratigraphic interpretations at this locality is shown in Figure 68.

- 93.1 0.4 Cowbell Member on right.
- 93.5 0.4 Bridge over Flat Fork.
- 93.6 0.1 Cowbell Member.
- 94.0 0.4 West end of a long exposure of Breathitt black shales and sands. At the extreme west end of the cut, the black shales are underlain by a thin underclay which sits on a rubble of brecciated St. Louis chert clasts and the green shales of the Nada Member of the Borden. At this locality we are very near the apex of the Waverly arch, where erosion has been more intense than at Stop 7, because the entire Renfro Member of the Slade and parts of the Nada Member are absent. The chert breccias and underclay may be part of a Pennsylvanian residual soil (see Ettensohn, 1981, p. 237).
- 94.8 0.8 Breathitt unconformably overlying Nada Member of the Borden; Cowbell Member at the base of the exposure.
- 95.1 0.3 Breathitt black shales and sands.
- 95.5 0.4 Exposure Z-515 (Fig. 51); except for the Poppin Rock Member, this is a relatively complete Slade section, exhibiting the Renfro-through-Maddox Branch members, as well as the underlying Nada Member of the Borden. The thinness and disrupted nature of the section reflect its position high on the eastern flank of the Waverly arch, as well as slumping, collapse, and solution accompanying Early-Middle Pennsylvanian paleokarst on the apical island area (see Ettensohn, 1981, p. 234). Breathitt shales and sandstones disconformably over slumped Slade units.
- 96.1 0.6 West end Z-516 (Fig. 51). The knoll that forms the western end of this exposure provides one of the best examples of the intra-Mississippian unconformities along the interstate, but it is rapidly being destroyed as the lower Slade carbonates slump and slide down the steep

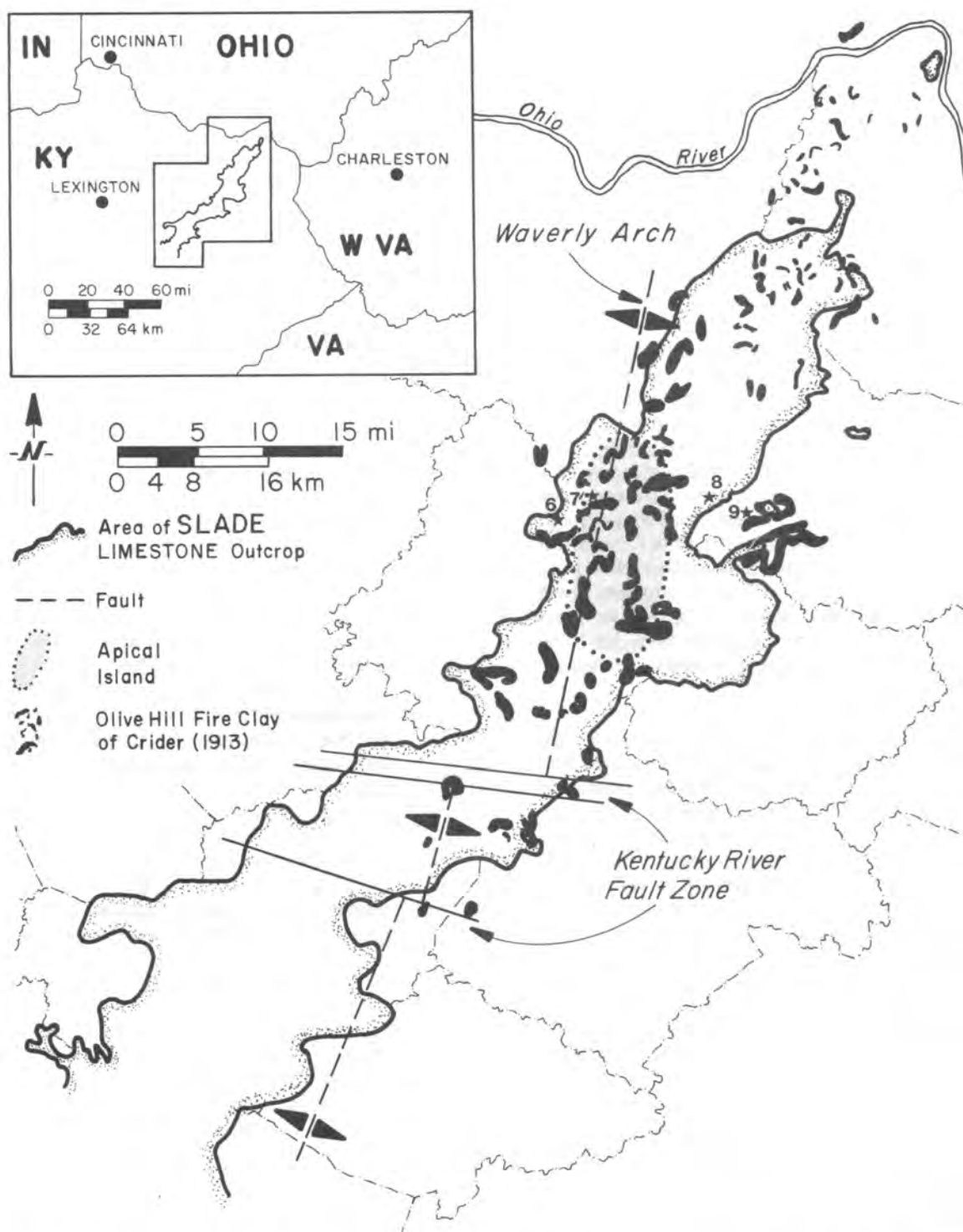


Figure 67. Distribution of Olive Hill Clay Bed of Crider (1913) in northeastern Kentucky as shown on geologic quadrangle maps. Note the general restriction of the Clay bed to the northern uplifted block and the area around the Waverly arch. Numbered stars represent stops on Day 2 (modified from Ettensohn and Dever, 1979).

STRATIGRAPHY NEAR APEX OF WAVERLY ARCH (STOP 7)

(R) - Red and Green shales

Modified after Horne and others, 1974

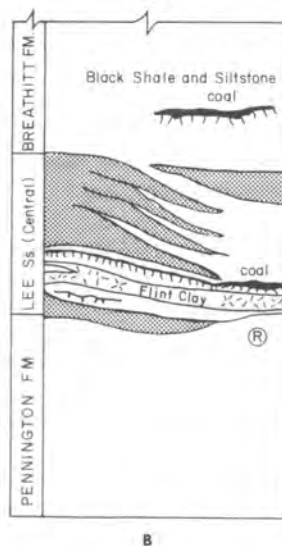
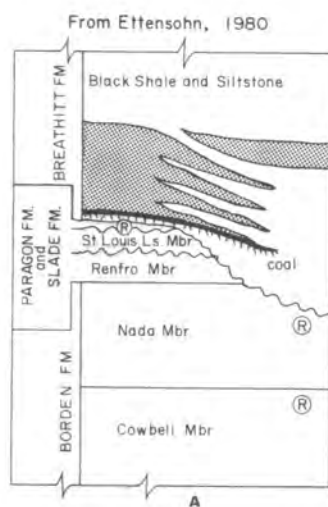


Figure 68. Comparison of stratigraphic interpretations at Stop 7 near the apex of the Waverly arch: A) "tabular-erosion model, B) Barrier-Shoreline model using nomenclature of Horne and others (1974) (modified from Ettensohn, 1980, 1981).

- Nada slope along the roadside. A point of interest in this exposure is the fact that two feet (0.6 m) of Warix Run calcarenite and four feet (1.2 m) of overlying Mill Knob calcilutite are found laterally adjacent to the Nada, Renfro, and St. Louis in a channel-fill relationship (Fig. 51). This exposure reveals a small apparently north-south channel into the lower Slade and upper Borden on the east flank of the Waverly arch, that was connected with the larger and deeper erosional low to the east. The channel was subsequently filled with Warix Run and Mill Knob sediments (see Ettensohn, 1981, p. 236).
- 96.7 0.1 End of exposure Z-516 (Fig. 51); the east end of the exposure exhibits Renfro, St. Louis breccia, and Cave Branch-through-Maddox Branch members, all contorted due to paleoslump and collapse. Soon after highway construction, even a block of the Poppin Rock Member was found in slumped Maddox Branch shales, but it was subsequently removed by roadside cleanup crews. The Breathitt sandstone in the eastern end of the cut fills a large collapse structure, and voids elsewhere in the Slade are filled with Pennsylvanian orthoquartzitic sandstone (see Ettensohn, 1981, p. 236).
- 96.9 0.2 Breathitt black shales.
- 97.1 0.2 Begin a long exposure of Breathitt black shales containing a single white sandstone body.
- 97.6 0.5 Exposure Z-517 (Figs. 51, 52); the Slade Formation at this locality is nearly complete containing the Renfro-through-Poppin Rock members, in addition to parts of the Paragon and Borden formations. The overall thinness of the Slade (32 ft; 9.8 m) and its disrupted nature, again reflect its position on the eastern flank of the Waverly arch. Because both the Slade and Paragon formations were involved, this slumping and collapse, were probably related to Early-Middle Pennsylvanian paleokarst localized on the Waverly arch apical island during the "Mississippian-Pennsylvanian" unconformity interval. This is the last exposure on the apical-island area (Fig. 51); as we move eastward we will encounter progressively thicker Slade sections deposited in deep erosional lows of the eastern flank of the arch (Fig. 52) (see Ettensohn, 1981, p. 236).
- 98.2 0.6 Exit 156 to Kentucky State Routes 52 and 2 (Olive Hill and Vanceburg). The Armstrong Hill section (Z-518, Fig. 52; Ettensohn, 1981, p. 231) is located a few tenths of miles to the north on Route 2.
- 98.4 0.2 The erosional contact between the dark-weathering Warix Run Member of the Slade and the Nada Member of the Borden is apparent on the entrance ramp to the left (Ettensohn, 1981, Fig. 33).
- 98.5 0.1 Overpass.
- 99.2 0.7 Exposure Z-519 (Fig. 52); Cowbell Member of the Borden and Warix Run-through-Tygarts Creek members of the Slade Formation (Ettensohn, 1981, p. 229). One major paleosol horizon is present on the Cowbell and another atop the Mill Knob Member (Ettensohn and others, 1984).
- 99.4 0.2 Exposure Z-520 (Fig. 52); Warix Run-through-Tygarts Creek members of the Slade. Large nodules and tepees probably generated as a result of paleosol formation (Ettensohn, 1981, Fig. 30; Ettensohn and others, 1988a).
- 99.7 0.3 Exposure Z-521 (Fig. 52); Cowbell Member of the Borden and Warix Run-through-Tygarts Creek members of the Slade Formation.
- 100.1 0.4 Exposure Z-522; Cowbell Member and Warix Run-through-Armstrong Hill members of the Slade.
- 100.3 0.2 Milepost 158 and exposure Z-523 (Fig. 52); Warix Run-through-Ramey Creek members of the Slade.
- 100.4 0.1 Bridge.
- 100.5 0.1 Exposure Z-524; Mill Knob-through-Ramey Creek members of the Slade.
- 100.7 0.2 Exposure Z-525; (Fig. 52); Warix Run-through-Tygarts Creek members of the Slade. A large calcareous sandstone body, probably a bar within the Warix Run Member, begins in this exposure.
- 100.9 0.2 Exposure Z-526 (Fig. 52); Warix Run-through-Tygarts Creek members of the Slade. The large sandstone bar in the Warix Run continues in this exposure.
- 101.7 0.8 Exposure Z-528 (Fig. 52); Tygarts Creek and Ramey Creek members. Other units of the Slade and Paragon formations are present on the westbound lane (see Ettensohn, 1981, p. 227).
- 102.8 1.1 Exposure Z-530 (Fig. 52); Poppin Rock Member of the Slade and the lower dark shale member of the Paragon Formation (as at Stop 5 on Day 1) are overlain unconformably by Breathitt sandstones and shales in the median and along the eastbound lane. Palynology of this interval was analyzed by Ettensohn and Peppers (1979).
- 103.1 0.3 Exposure Z-529 (Fig. 52) and bridge across Tygarts Creek; Tygarts Creek-through-Poppin Rock members of the Slade and lower dark shale member of the Paragon Formation.
- 103.4 0.3 Turn right onto Exit 161 to Olive Hill on U.S. Highway 60; view of Ken-Mor Quarry on the right.
- 103.7 0.3 Turn right onto U.S. 60.
- 104.0 0.3 Turn right into Ken-Mor Stone Co. Quarry (Z-501; Fig. 52). STOP 8. Slade and Paragon Formations West of the Waverly Arch (On U.S. 60, 0.3 mi south of the junction of I-75 and U.S. 60 at Pleasant Valley, Carter Co.; Carter Coordinates: 2350' FSLx2400' FEL, 3-V-77).

SLADE AND PARAGON FORMATIONS WEST OF THE WAVERLY ARCH

FRANK R. ETTENSOHN

The section in this quarry is located approximately 11.5 mi east of the Waverly arch and illustrates the pronounced thickening of the Slade and Paragon formations that occurs away from the arch. The quarry exhibits at least 155 ft (47.3 m) of the Slade Formation which is composed largely of carbonates except for red and green shales that make up the Maddox Branch and Cave Branch members (Fig. 69). The base of the Slade was never exposed in this quarry, but it is likely that the Warix Run Member of the Slade disconformably overlies the Cowbell Member of the Borden (Fig. 52, Z-501), meaning that parts of the upper Borden (Nada Member) and parts of the lower Slade (Renfro and St. Louis) are absent along the disconformity. Although parts of the Warix Run and Mill Knob members were exposed when the section was measured, lower parts of the quarry have since been infilled and these members are no longer completely visible.

The Slade is overlain by approximately 25 ft (8 m) of the Paragon Formation. Although different concepts exist about placement of the Slade-Paragon contact, I have used Butts' (1922) original definition of the boundary in eastern Kentucky as the top of the Glen Dean (Poppin Rock Mbr.). The Paragon in east-central Kentucky has been divided into four informal members (Figs. 1, 55), two of which, the lower dark-shale member and the clastic member, are present here. The lower dark-shale member is conformable with the Poppin Rock and at this quarry consists of 15 ft (4.6 m) of dark-gray, fossiliferous shale with thin limestone lenses. Palynology (Ettensohn and Peppers, 1979) and the presence of the guide fossils *Pterotocrinus depressus*, *P. cuneatus*, and *P. acutus* (Ettensohn, 1975, 1980) indicate a Middle Chesterian (Homburgian) age for this shale. The overlying clastic member typically consists of a fining-upward clastic sequence beginning with a basal sandstone and culminating in sandy red and green shales with dolostone stringers. Throughout most of its distribution, the unit has a tabular or wedge-shaped geometry thickening to the east and south. In northeastern Kentucky, however, east of the Waverly arch, the basal sandstone abruptly thickens and assumes a channel geometry (Figs. 52, 59B, 70B). At this point the sandstone is known as the Carter Caves Sandstone (Englund and Windolph, 1971). In this quarry, we are observing the sandstone along its southwest margin (Fig. 50). Here, the sandstone can be observed to thicken abruptly from thin lensoid bodies at the top of the lower dark-shale member to a channel fill over 25-ft (7.6-m) thick. As the body thickens, the erosional truncation of underlying units (lower dark-shale member, Poppin Rock, and Maddox Branch) at the channel edge is clearly evident (Fig. 70B). At this stop and to the north, the channel was cut to the level of the red and green Maddox Branch shale, although locally the channel was excavated as deeply as the Tygarts Creek Member. The Carter Caves Sandstone is overlain by at least five feet (1.5 m) of red and green shales in the clastic member. Upper parts of the clastic member and the overlying limestone member (Fig. 55), which are not present here, contain distinctive Upper Mississippian guide fossils (Sheppard and Dobrovolsky, 1962).

In the newer, Western part of the quarry, the Carter Caves Sandstone has been truncated by sandstones and shales of the Breathitt Formation. Although the Carter Caves and Breathitt sandstones are difficult to distinguish with just cursory examination, the presence of dark shales and palynological analysis of them (C.F. Eble, pers.

commun., 1992) clearly indicates the Breathitt affinities of the sandstones in the western part of the quarry.

The Carter Caves Sandstone is interpreted to represent a nearly north-south tidal channel formed as tidal currents were apparently canalized along the east flank of the emergent arch (Ettensohn, 1981, p. 223).

The Carter Caves Sandstone is significant in the "Mississippian-Pennsylvanian" boundary controversy, because in the Barrier-Shoreline Model (Figs. 19A, 24A), Ferm and others (1971, p. 17-19) and Horne and others (1974, Fig. 3A) interpreted this sandstone to be an orthoquartzitic Lee barrier sandstone ("eastern Lee sandstone," Fig. 70A) which intertongued with dark Breathitt shales to the east and with red and green Pennington shales to the west. Because the Carter Caves Sandstone is both underlain and overlain by definite Upper Mississippian rocks, this interpretation seems unlikely.

Return to quarry entrance and turn right onto U.S. 60.

- 104.1 0.1 Turn left onto Kentucky State Route 182 toward Grahn.
- 106.8 2.7 Bridge over Grahn Creek.
- 106.9 0.1 Corbin Member of the Lee Sandstone cropping out on the left
- 107.0 0.1 Reentrant in the basal Corbin.
- 107.3 0.3 STOP 9. Lee Formation and Mississippian-Pennsylvanian Contact, Grahn, KY (Two miles south of the junction of U.S. 60 and I-64 on Kentucky State Route 182 on the outskirts of Grahn, Ky., Carter Co.; Carter Coordinates: 1300°FSLx300°FEL, 15-V-78).

LEE FORMATION AND THE MISSISSIPPIAN- PENNSYLVANIAN CONTACT

DONALD R. CHESNUT, JR., STEPHEN F. GREB,
AND FRANK R. ETTENSOHN

The sandstone exposed along the cliff line is the Corbin Sandstone Member of the Lee Formation. Immediately underlying the Corbin at road level is the Olive Hill flint clay bed of Crider (1913) (Fig. 71). In this area the Olive Hill clay has been extensively mined for refractory clay, most recently by the Louisville Fire Brick Company.

THE CORBIN SANDSTONE

This exposure of the Corbin Sandstone is noteworthy because it occurs in the general area where the Lee-Newman(Slade)-Barrier-Shoreline model (Horne, and others, 1971) was developed and because it occurs on the northwest margin of the Corbin sandstone belt, the last of the Lee sandstone belts (Fig. 64) formed in the early Middle Pennsylvanian (as defined in the Appalachian basin).

The Corbin sandstone here is at least 50-ft (15-m) thick. Just four miles to the northeast, on Interstate 64 near Gregoryville, the Corbin is absent as a mappable unit, but may be represented by a thin quartz-pebble sandstone (see Optional Stop A, Day 3, and Fig. 72). The Corbin at this stop, however, is typical of the principal named

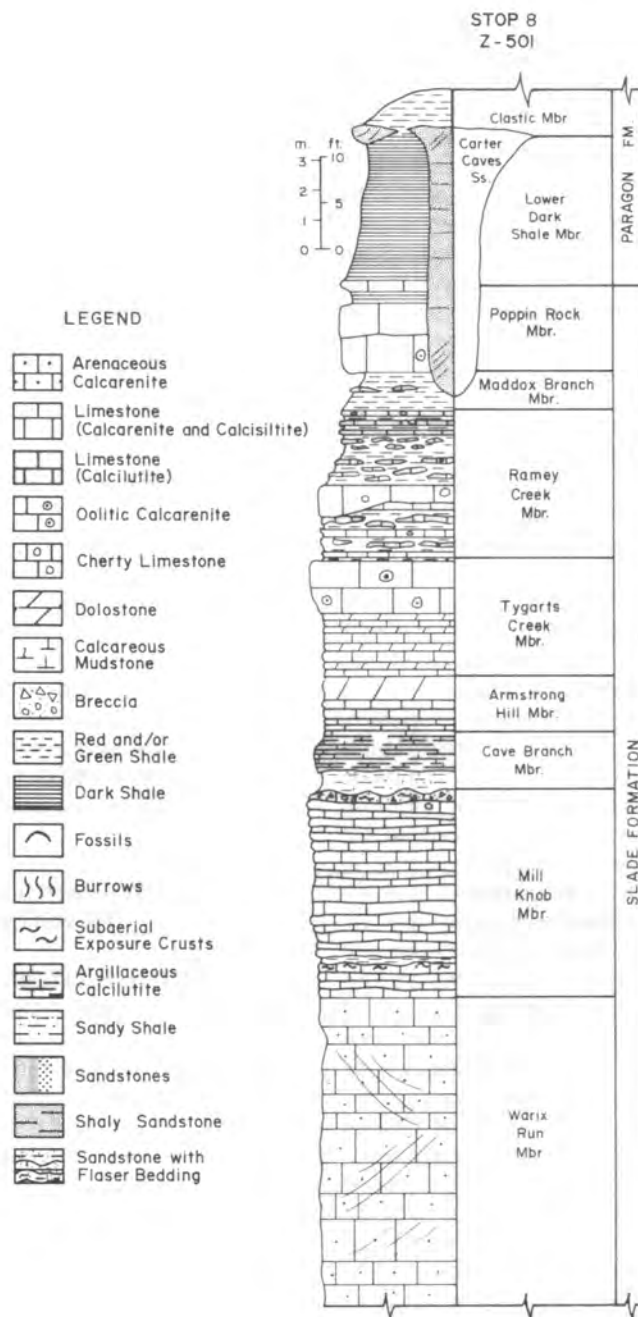


Figure 69. Mississippian stratigraphic section at Stop 8, Day 2 (Olive Hill Ken-Mor Quarry). Large parts of the Warix Run and Mill Knob members are no longer exposed due to recent quarry infilling. Tygarts Creek type section (modified from Ettensohn, 1981).

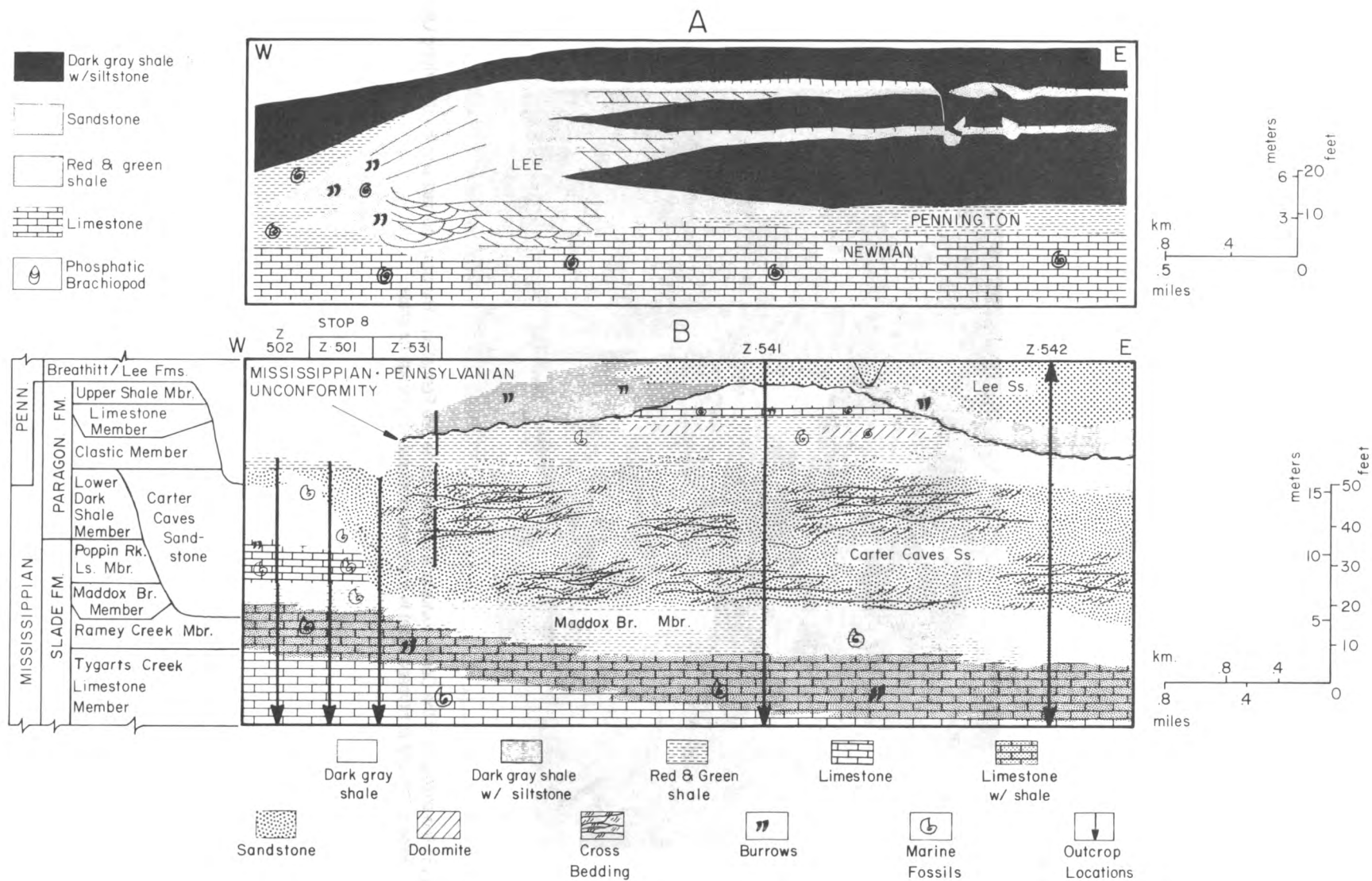


Figure 70. Comparison of Barrier-Shoreline interpretation (A), modified from "eastern Lee sandstone" of Horne and others (1974) using their nomenclature, and a "tabular-erosion interpretation (B) of stratigraphy of the Carter Caves Sandstone and adjacent units east of the Waverly arch (modified from Ettensohn, 1980, 1981).

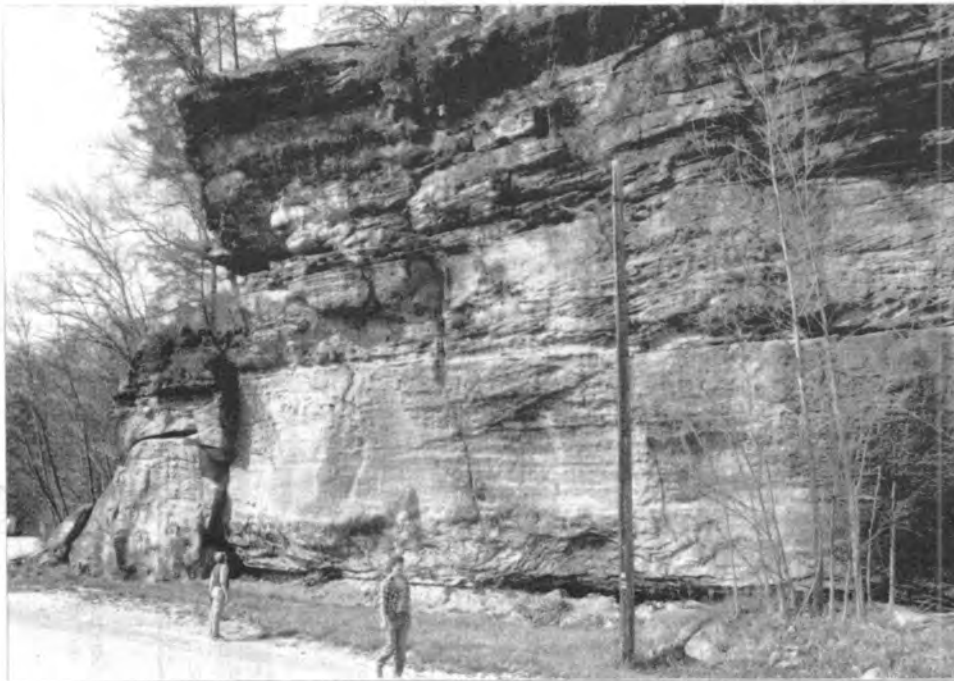


Figure 71. The cliff-line exposure of the Corbin Sandstone Member of the Lee Formation. The basal contact with the underlying Olive Hill flint clay (about one meter of the clay is exposed here) is sharp and erosional. Earthlings for scale.

Figure 72. Diagram of the roadcut at Optional Stop A (Gregoryville Stop) on the northern side of I-64 at Milepost 166. Drawing is based on field measurements and photographic montages; section above coal C, now largely covered, is based on information in Ferm and others (1971, fig. 10). Elevation at road level is about 690 ft (210 m) at the eastern end and about 715 ft (218 m) at the western end. No horizontal scale is implied.

members of the Lee Formation. The sandstone consists of a conglomeratic basal lag overlain by composite, bar-form cross strata (up to 13-ft thick) and stacked cosets of planar-tabular cross beds oriented to the west and southwest. In nearby outcrops, paleocurrents are also consistently to the west and southwest and several channel-form scours were noted. One scour, less than a mile north of this stop, contains a slump on its margin. All outcrops of the Corbin Sandstone in this area appear to fine upward and also exhibit bedding that generally thins upward. These types of features are typical of fluvial deposits.

THE LEE FORMATION

The sandstones of the Lee Formation are thick (50-200 ft; 15-60 m), massive, and generally cliff-forming, sublitharenites to quartzarenites which occur as lenses within the more argillaceous and less mature strata of the Breathitt Formation (see discussion in the section on Pennsylvanian settings). The Lee sands commonly are coarser than the Breathitt sands, and may contain abundant quartz pebbles. Breathitt sands, however, do not contain quartz pebbles.

The basal contacts of Lee sandstones are generally erosional, although the sandstones may laterally interfinger with the Breathitt. On subsurface geophysical logs, Lee sandstones typically have a blocky signature and fine upward in the uppermost part of the sandstone. The uppermost part may be interbedded with shale. In outcrop, bedding generally thins upward, and a coarse pebble lag commonly occurs at the base. Many Lee sandstones are overlain by rooted seatrock and a coal bed which may, in turn, be overlain by a carbonaceous-shale sequence of marine or brackish-water origin. In other areas the seatrock and coals are absent and marine shales directly overlie the Lee sandstones.

The Lee sandbelt members, such as the Corbin, are a crucial component in Carboniferous basin models. In the Lee-Newman(Slade)-Barrier-Shoreline model (Fig. 53) of Horne and others, (1971), the quartzose sandstones of the Lee Formation were originally interpreted to represent beach or barrier-bar sands which were transitional between Mississippian carbonates of marine origin and Breathitt coal-bearing clastics of terrestrial origin (Figs. 53, 59A, 66B, 68A). The maturity of the sandstones was explained as a result of beach-barrier-bar-forming processes.

However, sedimentological studies of the Corbin and other Lee sandstone members do not support a regional beach-barrier-island interpretation. Rather, the Lee sandstone belts are interpreted to represent fluvial-trunk transport systems situated between a mid-Carboniferous forebulge (peripheral bulge) to the northwest and the Breathitt clastic wedge to the southeast (Chesnut, 1988). A compilation of crossbed directions indicates that flow was dominantly to the southwest, parallel to the orientation of the sandstone belts (Chesnut, 1988), which is more typical of fluvial than beach-barrier systems. Vertical and lateral profiles, well developed pebble lags, bedding architecture, thinning-upward stratification, and general blocky to fining-upward grain-size trends also are more typical of fluvial than beach-barrier systems (Walker, 1984; Miall, 1985; Reineck and Singh, 1986). This combination of sedimentological criteria has been used by several authors to reinterpret Lee beach-barriers as fluvial sandstones in many parts of the basin (BeMent, 1976; Hester, 1977; Rice, 1984; Wizevich, 1991; Greb and Chesnut, 1989). Marine indicators that might be used to support barrier-bar deposition do not occur throughout the major Lee sandstone bodies as would be

expected in a foreshore-to-beach transition (McCubbin, 1982; Moslow, 1983; Reineck and Singh, 1986). Rather, marine indicators such as bioturbation, herringbone stratification, and megaripple-style cross stratification, if they occur at all, are mostly restricted to the uppermost part of the named members, where they may record estuarine reworking of fluvial sandstones (e.g., Greb and Chesnut, 1989).

REGIONAL UNCONFORMITY

The Corbin Sandstone Member, which is observed at this stop, is the uppermost of four large sandstone members of the Lee Formation (Fig. 64). These sandstone belts onlap the Early Pennsylvanian regional unconformity to the northwest toward the Cincinnati arch where they overlap younger strata. For example, the Corbin (lower Middle Pennsylvanian) overlaps the Paragon and Slade formations (Upper Mississippian) in the Grahn area. This onlapping-overlapping relationship has an important bearing on basin analyses as set out in an earlier Pennsylvanian framework statement.

THE OLIVE HILL FLINT CLAY

The Olive Hill flint clay is a pedogenic flint clay occurring near the regional Early Pennsylvanian unconformity in northeastern Kentucky. The Olive Hill is correlated with the Sciotoville clay bed of Ohio. The Olive Hill flint clay was formed during long periods of exposure of Upper Mississippian shales, and locally Lower Pennsylvanian shales, in a tropical environment. Regional stratigraphic relationships indicate that the flint clay here developed on shales low in the Upper Mississippian (Chesterian) Paragon Formation, not far above the uppermost carbonates in the Slade Formation, generally on or near structural elements (Fig. 67). At least 50 ft (15 m), and probably more, of the Paragon Formation are missing along this unconformity. Although the altered shales are Mississippian in age, the alteration or pedogenesis was clearly Early or Middle Pennsylvanian in age.

Mineralogical and textural features as well as identification of A, B, and C horizons support a paleosol interpretation for this flint clay (Smyth, 1984). The A horizon (not preserved at Grahn) is an organic-rich zone. The B horizon (preserved at Grahn) is argillic and the C horizon (not exposed here) is represented by traces of original laminae and maximum sideritization (Smyth, 1984). Pisolites and aggregates (seen locally in the exposure and in some of the blocks in the parking area), as well as mottling indicate either alternating wet and dry conditions (seasonality?) or freely drained conditions in places on or near topographic highs or areas of better permeability (Smyth, 1984). The importance of topographic highs to the formation of fire clays has already been mentioned in the discussion at Stop 7, and like that locality, this is also located on the northern uplifted block and close to the Waverly arch apical island (Fig. 67). The fine-grained flint clays (seen in the outcrop and in some of the blocks in the parking lot) probably formed in hydromorphic (waterlogged) environments (Smyth, 1984). Uncommon volcanic minerals and "spherules" in the flint clay were used as evidence that the clay was an altered volcanic ash (e.g., Bohor and Triplehorn, 1984). However, the preponderance of soil structures, a pedogenic origin for the spherules (probably pisoids), and common clastic resistate minerals in the flint clay support a pedogenic origin on weathered Mississippian or Lower Pennsylvanian rocks (with perhaps periodic contamination by volcanic ash falls).

END OF ROAD LOG FOR DAY TWO.

ROAD LOG AND STOP DESCRIPTIONS—THIRD DAY

Mileage:

cum. inc.

- 0.0 Road log begins at the junction of Kentucky State Routes 1 and 7 with I-64 at Grayson, Kentucky. Just right of intersection is a prominent knob known as Bald Eagle, where dark Breathitt shales overlie Grayson sandstone bed. The Fire Clay-Whitesburg coal zone occurs about 40 ft (12.2 m) from top of knob in Breathitt Formation. Approximately 5 mi southeast of Grayson near Hitchens, Kentucky, the clay beneath the Princess No. 6 coal (Hitchens Clay Bed) is mined for use in refractory brick. The Hitchens itself is too plastic to make good brick, but when mixed with the Olive Hill clay of the Lee Formation (mined farther to the west), their combined properties make excellent bricks. Turn left into entrance ramp onto I-64.
- 0.2 0.2 Enter onto I-64.
- 0.4 0.2 Small strip mines on left and right are in Fire Clay-Whitesburg coal zone. For the next few miles, route coincides with valley of Barrett Creek.
- 3.6 3.2 U.S. 60 overpass.
- 4.1 0.5 Enter into area covered by Grahn Geologic Quadrangle Map (Englund, 1976).
- 5.9 1.8 OPTIONAL STOP A. Breathitt Formation and Models for Coal-Forming Environments, Gregoryville, Kentucky, (East side of I-64 at mile-marker 166), about four miles (6.4 km) east of the Olive Hill (161) exit, Carter Co.; Carter Coordinates: 4100'FSLx600'FEL, 3-V-79).

BREATHITT FORMATION AND COAL-FORMING ENVIRONMENTS

DONALD R. CHESNUT, JR., STEPHEN F. GREB,
AND CORTLAND F. EBLE

The Gregoryville outcrop illustrates: 1.) typical coal-bearing rocks of the Breathitt Formation, which is the most important coal-producing unit in eastern Kentucky; 2.) different interpretations of depositional environments for coal-bearing rocks of the central Appalachian basin; and 3.) the role that sea-level changes have probably played in the deposition of the coal-bearing rocks.

OUTCROP DESCRIPTION

All the rocks exposed in the I-64 roadcut belong to the Breathitt Formation and are Middle Pennsylvanian in age as defined in the Appalachian basin (specifically, this section is equivalent to part of the Kanawha of West Virginia and the upper part of the Morrowan of the mid-continent). Four thin coal beds are exposed in the roadcut, labeled in ascending order in Figure 72 as A, B, C, and D. Samples of coals A (Wheelersburg? coal bed, 15-in or 38-cm thick) and

C (Bruin? coal bed, 16-in or 41-cm thick) were collected for palynological analyses, confirming an age assignment of Middle Pennsylvanian (they are equivalent to an interval somewhere between the base of the Betsie Shale through the Upper Elkhorn No. 3 coal bed; Fig. 64). About 170 ft (52 m) above the Bruin? (and above the roadcut) is the Fire Clay coal bed (approx. 920 ft elev., Charles Rice, pers. commun.), which is probably the best regional stratigraphic marker bed in the Middle Pennsylvanian of the Appalachian basin because of its distinctive tonstein (flint-clay) parting. The tonstein, an altered volcanic ash fall, has been isotopically dated at about 311 Ma, and based on megafossils, its stratigraphic position has been correlated with the uppermost part of the Trace Creek Member of the Atoka Formation in the midcontinent region (Rice and others, 1990).

In the underpass exposure at the eastern side of the roadcut (Fig. 72), a thin Pennsylvanian sandstone (containing abundant clay and quartz pebbles) overlies a heavily-weathered soil profile (pedogenic flint clay) developed on shales and sandstones of the Upper Mississippian Paragon Formation. The mid-Carboniferous regional unconformity is placed at the soil profile (Fig. 72). The strata of the Gregoryville section have been divided into five members for purposes of discussion (Fig. 72). Although the shales of the three lowest members, 1, 2, and 3, contain marine or brackish-water body or trace fossils, the units are not distinctive enough to have been mapped in this part of northeastern Kentucky. Based on the spore data from coal A and the position of the regional unconformity just below road level, we suggest that shale member 1 is equivalent to the Betsie Shale Member. The Betsie Shale, of marine- or brackish-water origin, is the thickest and most extensively developed marker unit in the Pennsylvanian of the Appalachian basin.

The upper part of shale member 1 (Fig. 72) contains abundant, small *Lingula*. The canneloid black shale ("bone") at the base of member 3 contains abundant specimens of the brackish-water pelecypod *Anthraconaia*. This boney shale, identified as coal in Ferm and others (1971), is overlain by shale and sandstones containing the trace fossils *Lockeia*, *Neonereites*, and *Planolites*. Member 2 (Grayson sandstone?) is composed of thin channel-form or tabular quartzose sand bodies that alternate with shales and coal beds (see lower part of roadcut, Fig. 72). Several of these beds contain marine trace fossils including *Bifasciculus*?, *Biformites*?, *Chondrites*, *Cochlichnus*, *Lockeia*, *Neonereites*, *Rusophycus*? and *Planolites*. The sandstones are dominantly ripple laminated and reflect bimodal paleocurrents. Sandstone D at the eastern end of the cut exhibits soft-sediment deformation and trough crossbeds, and contains coal spar (transported coaly material) as well as numerous plant fossils including a vertical *Calamites* limb. Typical of coastal "bay"-fill or prodelta deposits, members 3, 4, and 5 consist of coarsening-upward sequences (Ferm and others, 1971), which are commonly capped by seat rock and coalbeds (Fig. 72).

COAL-FORMING ENVIRONMENTS: PREVIOUS INTERPRETATIONS

This roadcut was originally described by Ferm and others (1971) to illustrate the Lee-Newman(Slade)-Barrier-Shoreline model of Horne and others (1971). These authors, however, not having the benefit of Englund's (1976) geologic map of the area, misidentified the upper beds in the roadcut as the Tom Cooper coal, Kendrick Shale, and Fire Clay coal, all of which occur higher up the hill. Nonetheless, this outcrop was used to illustrate coal-forming environments or facies such as "back barrier," and "lower delta plain," which were adopted by

coal geologists around the world. The validity of these facies as described at Gregoryville is examined below.

"Barrier" and "Back-Barrier" Facies

Ferm and others (1971) interpreted the strata in the lower part of the roadcut (members 1 and 2, Fig. 72) to represent "back-barrier" facies including bay-type coarsening-upward sequences (largely the shales), barrier-overwash deposits and tidal-inlet deposits (quartzose sandstones), seat rock, and coals. These "back-barrier" deposits were correlated with "barrier" sandstones at two localities about six miles to the west (Ferm and others, 1971, p. 16-21). One of these barrier sandstones (their Stop 8) is the Carter Caves Sandstone of the Paragon Formation (Fig. 70); the other (their Stop 9) is a local Breathitt sandstone lens occurring above the Slade Formation (Mile 103.8, Day 2). The first of these sandstones is Late Mississippian in age and occurs below the Early Pennsylvanian regional unconformity (Dever, 1973, 1980; Ettensohn, 1980; Chesnut, 1988) (Fig. 70B) and, therefore, cannot be laterally equivalent to the units exposed at the Gregoryville locality. The second sandstone, however, is Pennsylvanian in age (Ettensohn and Peppers, 1979), but is atypical Breathitt sandstone unlike the Lee, and is too distant from this stop to discern any definite stratigraphic relationships.

Quartzose sandstones of the Lee Formation were also regarded to represent barrier bars and beaches by Ferm and others (1971). The mature sandstones of the Lee were attributed to beach- and barrier-bar-forming processes. At Gregoryville, the presence of quartz pebbles in the thin sandstone below shale member 1 and directly overlying the Early Pennsylvanian unconformity (Fig. 72), as well as its stratigraphic position, suggests correlation with the quartzose sandstones of the Corbin Sandstone Member of the Lee Formation at Stop 9 on Day 2 near Grahn, Kentucky. Breathitt Formation sandstones, in contrast, generally lack quartz pebbles (Rice, 1984). The Gregoryville locality is only 1.5 mi (2.4 km) northwest from the mapped pinchout of the Lee Formation (Englund, 1976). This pinchout marks the edge of the Corbin sandstone belt (Fig. 64), a Lee sandstone belt which averages about 35 mi (56 km) in width, and which is oriented northeast-southwest across the western part of the central Appalachian basin (Chesnut, 1988, in press).

Sedimentological studies of the Corbin and other Lee sandstone members (see Pennsylvanian framework statement and Stop 9, Day 2) do not support a beach-barrier-island interpretation as suggested by Ferm and others (1971). Rather, the Lee sandstone belts are interpreted to represent fluvial, trunk-transport systems situated between a mid-Carboniferous forebulge (peripheral bulge) to the northwest and the Breathitt clastic wedge to the southeast (Chesnut, 1988) (Fig. 65A).

Because the Carter Caves Sandstone and Breathitt sandstones to the west, as well as the Corbin sandstone to the south, are not now recognized as Pennsylvanian-age barrier sandstones, the interpretation of back-barrier facies for the strata (members 1 and 2) in the lower part of this stop is not supported. However, the thin channel facies in member 2 does probably represent tidal-creek deposition as Ferm and others (1971) suggested, because, the channel-form cross section, bimodal paleocurrents, and marine trace fossils are typical of modern tidal creeks (Walker, 1984; Reineck and Singh, 1986). The shallow tidal channels may have formed in conjunction with marine reworking and redistribution of nearby Lee fluvial sands (i.e., Corbin Sandstone) during a period of higher sea level.

"Lower and Upper Delta-Plain" Facies

Ferm and others (1971) interpreted the strata (members 3, 4, 5; Fig. 72) above the quartzose sandstones to represent "lower delta-plain" facies composed of bay-type coarsening-upward sequences, delta-front/distributary-bar sands, seat rock, and coals. This interpretation was based on the use of the Mississippi delta as a modern analog for the coal-bearing rocks of the central Appalachian basin. Thus, the lower part of the Breathitt Formation, with abundant marine strata, was equated with the lower delta-plain environments, whereas the upper Breathitt, dominated by fluvial sands, represented the upper delta-plain. The vertical shift in the rock record from lower to upper delta-plain was attributed to increasing development of the Alleghanian orogeny and subsequent progradation of deltaic environments to the west. The alternation of coal beds, shales, and sandstones observed throughout the basin (and seen at this stop) was thought to have been caused by autocyclic mechanisms such as delta-switching. Therefore, according to the model, the lateral distribution of individual strata would have been restricted to areas of delta-lobe size. Hence, in the broadest sense, the Mississippi delta model does not support extensive correlation of any stratigraphic unit.

However, detailed geologic mapping of coal-bearing strata in the Appalachian basin has demonstrated 1.) the regional occurrence and distribution of tonsteins (volcanic ash falls) and 2.) regional continuity of coalbeds in deep and surface mines. A compilation of paleontological data provided by geologic quadrangle mapping and other sources confirm widespread presence of marine horizons (Chesnut, 1991d). Additionally, regional subsurface mapping of specific units and beds utilizing thousands of subsurface core logs, as well as oil-and-gas well logs, demonstrates that the major coal beds and marine strata are widespread and may extend across the entire central Appalachian basin with only local interruptions (Chesnut, 1988). The alternation of extensive sequences of coal beds, marine strata, and fluvial sandstones across a wide area suggests deposition on a broad coastal plain traversed by small shallow streams which was subject to repeated transgressions during rising sea levels. Also, there is, at present, no thick, low-ash and low-sulfur peat forming in the Mississippi delta that could ever form an economical coal deposit (e.g., McCabe, 1984). Therefore, the use of lower and upper delta-plain interpretations does not seem appropriate for this basin.

Alternative Model

Sequences, such as those illustrated at this site (Fig. 72), are typical of transitional coastal lowland-alluvial plain to shallow marine-shelf deposits similar to coastal Indonesia (Cobb and others, 1989; Chesnut and others, 1991). The repeated transition from terrestrial to marine conditions in the Breathitt Formation, as is seen at this stop, is interpreted to result from fluctuating Pennsylvanian sea level, which, in the Appalachian basin, generally has been attributed to a combination of regional-tectonic and glacio-eustatic controls (Chesnut, 1989b). The overall coarsening-upward trend of the Breathitt Formation, previously interpreted as a single deltaic progradation, is probably caused by multiple progradational events controlled by increasing intensity of the Alleghanian orogeny and changing sea level (Chesnut, 1991b).

- 6.0 0.1 Sandstone slump block in Breathitt shales on right.
- 7.3 1.3 Little Caney coal bed exposed just above road level on the right.

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| 8.4 | 1.1 | Little Caney coal bed exposed just above road level on the right. | 13.9 | 0.1 | Disconformable contact of Carter Caves Sandstone with Maddox Branch Member of Slade Formation on left side of road at bend. |
| 9.1 | 0.7 | To right in the distance, a highwall of a strip mine in the Fire Clay-Whitesburg coal zone is visible. | 14.4 | 0.5 | Large crossbeds characteristic of Warix Run Member of the Slade. |
| 9.9 | 0.8 | Limestone concretions can be seen in shales on left and right. This is probably the Kendrick Shale horizon of Jillson (1919). | 14.5 | 0.1 | Disconformable contact between Warix Run Member of Slade and Cowbell Member of Borden Formation. |
| 10.3 | 0.4 | Exit right onto U.S. 60 exit ramp (Exit 161) from I-64. | 14.7 | 0.2 | Bridge across Tygarts Creek. |
| 10.5 | 0.2 | Stop. Turn right onto U.S. 60 from exit ramp. To the right of the stop sign, approximately eight feet (2.4 m) of green Paragon shales with nodular to irregularly bedded dolomitic mudstones are disconformably overlain by dark Breathitt shales (Z-531, Fig. 52). The Early Pennsylvanian unconformity is represented here by an intensely weathered, limonitic horizon on green Paragon shale overlain abruptly by dark Breathitt shale. | 14.8 | 0.1 | Entrance to Carter Caves State Park. |
| 10.6 | 0.1 | Citgo gas station on the left. | 15.0 | 0.2 | Cross-bedded limestones of the Warix Run Member of the Slade Formation (Upper Mississippian). Roadcut on the right side proceed uphill. |
| 11.2 | 0.6 | Junction with Kentucky State Route 209 on left. | 15.3 | 0.3 | Carter Caves Sandstone (Upper Mississippian), exposed in roadcut to the left. |
| 11.9 | 0.7 | Junction of U.S. 60 and Kentucky State Route 182; turn left onto State Route 182 toward Carter Caves State Park. | 15.5 | 0.2 | Little Wolf Grocery on right side of the highway. |
| 12.0 | 0.1 | Ft. Falls Trading Post on the right. | 15.8 | 0.3 | Roadcut of Breathitt Formation, exposed on right side of the road. |
| 12.2 | 0.2 | Cliffs in Carter Caves Sandstone at head of Box Canyon, just below road level on the left exhibit a prominent reentrant and falls. | 16.2 | 0.4 | Entrance to Camp Cardinal Girl Scout Camp on right side of the highway. |
| 12.3 | 0.1 | Poor exposures of limestone member of the Paragon Formation present on right side of road. This Mississippian limestone occurs above the Carter Caves Sandstone (Figs. 1, 55, 70B). | 17.2 | 1.0 | Carter Caves Sandstone exposed in roadcut to the right. |
| 12.5 | 0.2 | Entrance on the left to Box Canyon (Echo Canyon), location of the type section of the Carter Caves Sandstone (Englund and Windolph, 1971). | 19.2 | 2.0 | Strongly cross-bedded Warix Run Member of the Slade Formation exposed on right. |
| 12.6 | 0.1 | Grayson sandstone member (Breathitt Fm) on left side of road. | 19.5 | 0.3 | Roadcut of Cowbell and Nada members of the Borden Formation exposed on the right, overlain by the Warix Run of the Slade Formation. |
| 12.7 | 0.1 | Dark shales of Breathitt Formation crop out on right side of road at bend. | 19.7 | 0.2 | Bridge over Jordan Creek. |
| 13.1 | 0.4 | Junction of Kentucky State Routes 182 and 209 (road to Cascade Caves). Continue northward on 182. | 19.8 | 0.1 | Junction of Kentucky State Routes 182 and 2; turn right (north) onto Kentucky Route 2 and proceed toward Carter City. |
| 13.6 | 0.5 | Approximate contact of dark Breathitt shales with Grayson sandstone (Englund, 1976); proceed down section through Grayson sandstone. | 20.0 | 0.2 | Cowbell Member of the Borden Formation exposed in roadcut on the left. |
| 13.8 | 0.2 | Early Pennsylvanian unconformity; disconformable contact of Grayson sandstone with green, fossiliferous Paragon shales overlying Carter Caves Sandstone on left side of road (see Ettensohn, 1981, p. 224). | 20.3 | 0.3 | Cowbell Member of the Borden Formation exposed in roadcut on the left. |
| | | | 21.0 | 0.7 | Cowbell Member of the Borden Formation exposed in roadcut on the left |
| | | | 21.4 | 0.4 | Junction with Kentucky State Route 396, on left side of the highway. |
| | | | 21.5 | 0.1 | On the left, cliffs at top of hill are formed of the Slade Formation whereas the slope is made-up of the Cowbell and Nada members of the Borden Formation. |
| | | | 21.8 | 0.3 | Cowbell Member of the Borden Formation exposed in roadcut on the left. |

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| <p>22.1 0.3 Cowbell Member of the Borden Formation exposed in roadcut on the left.</p> <p>22.9 0.8 Enter Carter City.</p> <p>23.0 0.1 Carter County Elementary School on right side.</p> <p>23.3 0.3 Junction of Kentucky State Routes 2 and 474; turn left (west) onto State Route 474. State Route 474 follows the former Chesapeake and Ohio Railroad that ended just to the east of Carter City. The railroad in this area was flooded out in the early 1940's. Large numbers of people in the late 1800's and early 1900's came to Carter City via the railroad for recreational activities such as picnicking, barn dances, and spelunking (Chaplin, 1980).</p> <p>23.7 0.4 Cowbell Member exposed in roadcut on right side of the road.</p> <p>24.2 0.5 Limestone talus, along right side of road, from abandoned limestone quarry (Slade Formation) on ridge above.</p> <p>24.6 0.4 Cowbell Member exposed in roadcut on right side of the road.</p> <p>25.1 0.5 Limestone talus, along right side of road, from abandoned limestone quarry (Slade Formation) on ridge above.</p> <p>25.5 0.4 Cowbell Member exposed in roadcut on right side of the road.</p> <p>25.8 0.3 Cowbell Member exposed in roadcut on right side of the road.</p> <p>26.0 0.2 Cowbell Member exposed in roadcut on right side of the highway.</p> <p>26.1 0.1 Community of Popular.</p> <p>26.3 0.2 Cowbell Member exposed in roadcut on right side of the highway.</p> <p>26.4 0.1 Entrance to the Standard-Lafarge Carter Quarry; approximately 120 ft (36.5 m) of the Slade Formation are exposed in the quarry on the left side of the highway. This is the thickest Mississippian carbonate exposure in northeast Kentucky.</p> <p>27.2 0.8 Smith Creek Church on the right side of the highway.</p> <p>27.5 0.3 Cowbell Member exposed in roadcut on right side of the highway.</p> <p>28.5 1.0 Smith Creek Post Office is opposite this point on the right side of the highway. A very small white building with a flag pole beside it has local acclaim as being the smallest active post office in the U.S.</p> <p>28.6 0.1 Cowbell Member exposed in roadcut on left side of the highway for approximately 0.5 mi uphill.</p> | <p>29.2 0.6 Junction of Kentucky State Routes 474 and 1149; turn right (north) onto State Route 1149.</p> <p>29.3 0.1 Deep roadcut on both sides of the highway exposing the Cowbell and Nada members of the Borden Formation, as well as the Warix Run Member of the Slade Formation. This is the top of the Trace Creek Section of Chaplin (1980), which begins here and proceeds along State Route 1149 downhill to the northwest for 1.2 mi.</p> <p>30.5 1.2 Cowbell and Nancy members of the Borden Formation exposed in roadcuts on the right. Note the gradational contact between underlying Nancy (shale) and overlying Cowbell (siltstone) which constitute prodelta and delta-front deposits respectively. This constitutes the base of the Trace Creek Section of Chaplin (1980), and the intervening exposures in the roadcuts, on the right side coming downhill, are of the Cowbell member.</p> <p>30.7 0.2 Nancy Member exposed in roadcut on right side of the highway.</p> <p>31.3 0.6 Farmers (turbidites) and Nancy (prodelta deposits) members (note gradational contact); Borden Formation exposed in roadcut on right side of the highway.</p> <p>31.6 0.3 Bridge over Trace Creek.</p> <p>32.1 0.5 Farmers Member exposed in stream cut on right side; cross bridge over Trace Creek.</p> <p>32.4 0.3 Farmers Member exposed in roadcut on right side.</p> <p>32.7 0.3 Farmers Member exposed in roadcut on right side.</p> <p>32.9 0.2 Farmers Member exposed in roadcut on right side.</p> <p>33.3 0.4 Pipeline Crossing.</p> <p>33.9 0.6 Pipeline Crossing.</p> <p>34.8 0.9 Farmers Member exposed in roadcut on both sides of highway.</p> <p>35.2 0.4 Berea Sandstone (Mississippian) exposed in roadcut on right side of highway.</p> <p>35.7 0.5 Berea Sandstone exposed in roadcut on right side of highway.</p> <p>35.9 0.2 Concrete bridge over Trace Creek.</p> <p>36.0 0.1 Berea Sandstone exposed in roadcut on left side of highway.</p> <p>36.5 0.5 New concrete bridge over Kinniconick Creek.</p> <p>37.2 0.7 Junction of Kentucky State Routes 1149 and 1306; turn right (northeast) onto State Route 1306, and proceed northward toward Garrison, Kentucky.</p> |
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- 37.3 0.1 Berea Sandstone exposed in roadcut on left side of the road, going uphill.
- 37.5 0.2 Matty Cemetery at top of hill on right.
- 39.2 1.7 Berea Sandstone exposed in roadcut on left side of the highway.
- 39.7 0.5 Sullivan Cemetery on right side of highway.
- 40.9 1.2 Berea Sandstone exposed in roadcut on left side of the highway.
- 41.2 0.3 Berea Sandstone exposed in roadcut on left side of the highway.
- 41.5 0.3 Bevins Chapel on right side of the highway.
- 42.2 0.7 Concrete bridge over Spy Run Creek.
- 42.5 0.3 Junction of Kentucky State Routes 1306 and 546 (AA Highway); turn right onto State Route 546 and proceed eastwardly.
- 42.6 0.1 Bridge over Kinniconick Creek.
- 42.8 0.2 Smoothrock Road on the right; in the distance to the left (north), one can see exposures of the Berea Sandstone along an old railroad cut. Note the contact between the Sunbury Shale and Berea Sandstone.
- 43.1 0.3 STOP 1. Lowstand Deposition in Foreland Basin: Bedford-Berea Sequence (Upper Devonian), eastern Kentucky and West Virginia, Kinniconick Creek Section, Garrison, KY (Junction with Greenbriar Road on right side of Kentucky State Route 546. Turn right [north] onto Greenbriar Road and disembark from the bus; Lewis Co.; Carter Coordinates: 600'FNLx1600'FEL, 21-Z-76). Berea Sandstone is exposed in roadcuts on the right side of Route 546. Also, included in this section are the Sunbury Shale, and the Henley Bed and Farmers Member of the Borden Formation. These units are exposed along the more gently sloping hillside above the steep roadcut of the Berea. This is the longest continuous roadcut of the Berea in northeastern Kentucky (approximately 1.1-mi long).

LOWSTAND DEPOSITION IN A FORELAND BASIN: BEDFORD-BEREA SEQUENCE (UPPER DEVONIAN), EASTERN KENTUCKY AND WEST VIRGINIA

JACK C. PASHIN AND FRANK R. ETENSOHN

INTRODUCTION

Investigations of the Bedford-Berea sequence laid the foundation for models of epeiric sedimentation more than 40 years ago. Rich (1951a, 1951b) included the Bedford-Berea in the original model in which he introduced the clinoform, a key element of modern sequence stratigraphy. Shortly thereafter, Pepper and others (1954) performed

a landmark study of Bedford-Berea sedimentation that presented fundamental ideas regarding facies interpretation and regional paleogeography. Even in relatively recent references, their model is still regarded as a paradigm of epicontinental deltaic sedimentation (Krumbein and Sloss, 1963; Fisher and others, 1969; Wanless and others, 1970; leblanc, 1975; Frazier and Schwimmer, 1987).

Sedimentologic advances since these classic studies have fostered new perspectives on the origin of the Bedford-Berea sequence (Coogan and others, 1981; Pashin, 1985, 1990; Pashin and Ettensohn, 1987, 1992; Lewis, 1988). Pashin and Ettensohn (1987) applied an updated version of Rich's epeiric model (Woodrow and Isley, 1983) in tandem with continental-margin sedimentary models to the Bedford-Berea. A major implication of Pashin and Ettensohn's study is that traditional epeiric models, though widely applicable, are too simplistic to fully characterize epicontinental sedimentation, especially in tectonically active areas. Building on this theme, Pashin (1990) reevaluated the classic model of Pepper and others (1954) and demonstrated that sea-level variation, relict topography, differential compaction, and tectonism acted in concert to form the complex facies patterns of the Bedford-Berea sequence. This discussion characterizes the Bedford-Berea sequence in the Appalachian foreland basin of eastern Kentucky and West Virginia and shows how a major sea-level drop caused progradation of orogenic sediment into distal parts of the Appalachian basin. Our intention is to show that sediment was accommodated in a basin formed largely by relict topography and differential compaction and to suggest ways in which basement-fault reactivation gave rise to diverse facies patterns within that basin. Today we will examine Bedford-Berea exposures that contain features not typically associated with epeiric sedimentation. At Stop 1 we will examine a transect through part of a storm-dominated shelf margin that includes a synsedimentary fault and associated soft-sediment deformation structures. An optional stop examines a sinuous turbidite channel and associated sediments.

STRATIGRAPHIC AND SEDIMENTOLOGIC FRAMEWORK

The Upper Devonian Bedford-Berea sequence is part of a thick succession of organic-rich, basinal black shale and intervening, light-colored shelf and coastal-plain clastics of Devonian and Mississippian age that extends across the North American craton (Heckel and Witzke, 1979; Ettensohn and Barron, 1981; Ettensohn, 1985a, 1985b). This succession includes the Catskill and Pocono clastic wedges which were shed into the then-euxinic Appalachian basin from the Acadian orogen (Fig. 73). The Bedford-Berea sequence separates the Catskill and Pocono wedges, and Elam (1981) defined the sequence as the interval between the organic-rich, black Cleveland Member of the Ohio Shale and the similar Sunbury Shale, as well as equivalent strata in the Appalachian basin (Fig. 74). Bedford-Berea deposition occurred mainly in the western part of the Appalachian basin and took place in two distinctive provinces, the *eastern platform* and the *western basin* (Pashin, 1990) (Figs. 75, 76). In West Virginia, sediment accumulated mainly on the eastern platform where the Bedford-Berea is generally thinner than 40 ft (12 m) and rests disconformably on the Catskill wedge. Near the West Virginia-Kentucky border, however, the Bedford-Berea thickens to more than 120 ft (35 m) and rests conformably on the Cleveland Shale. The transition from platform to basin is less than five-miles wide and coincides with the shift from a thick shelf-slope clastic sequence dominated by gray shale and siltstone of the Catskill wedge (Chagrin Shale) to a relatively thin, basinal black-shale sequence (Cleveland Shale) (Figs. 73, 75).

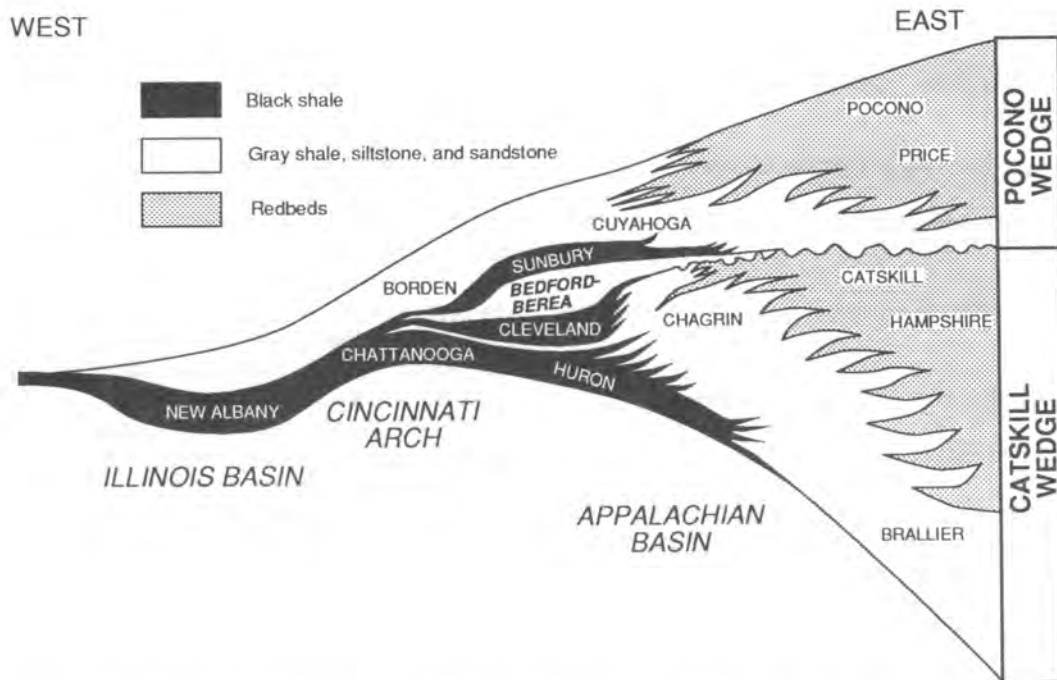


Figure 73. Relationship of the Bedford-Berea sequence to the Catskill and Pocono clastic wedges (after Pashin, 1990). A major downward shift in coastal onlap suggests that the Bedford-Berea sequence is the product of major relative sea-level drop that interceded Catskill and Pocono deposition.

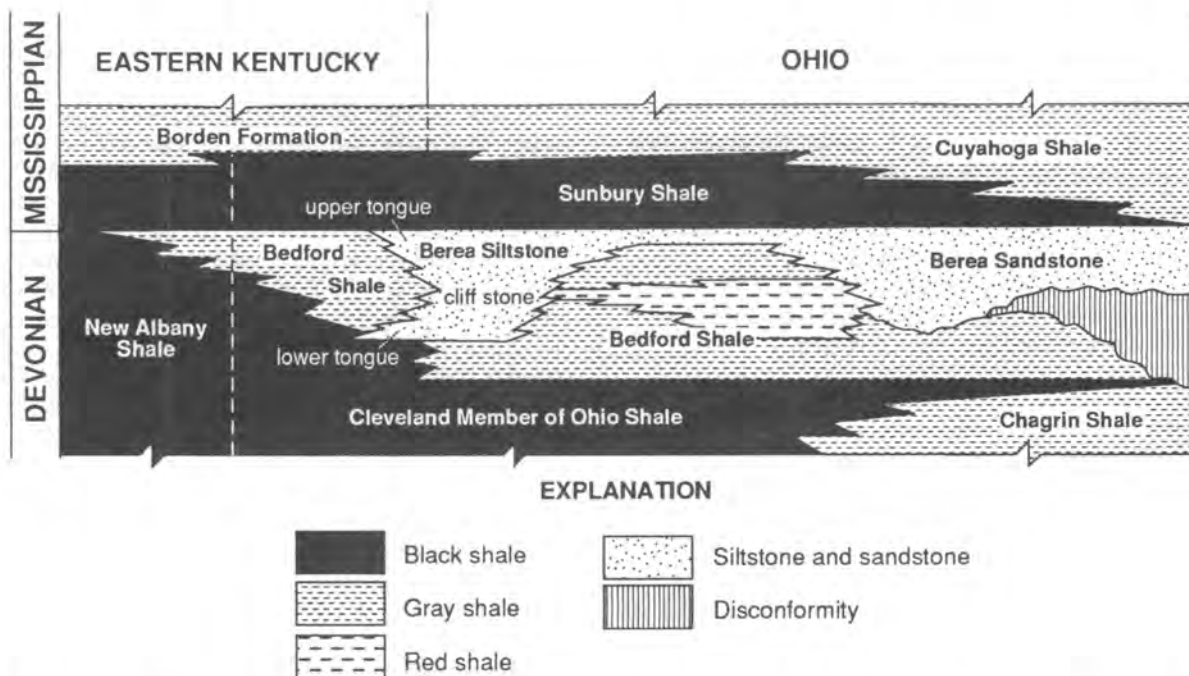


Figure 74. Stratigraphy of the Bedford-Berea sequence along the western outcrop of the Appalachian basin from eastern Kentucky to northern Ohio (after Pashin, 1990).

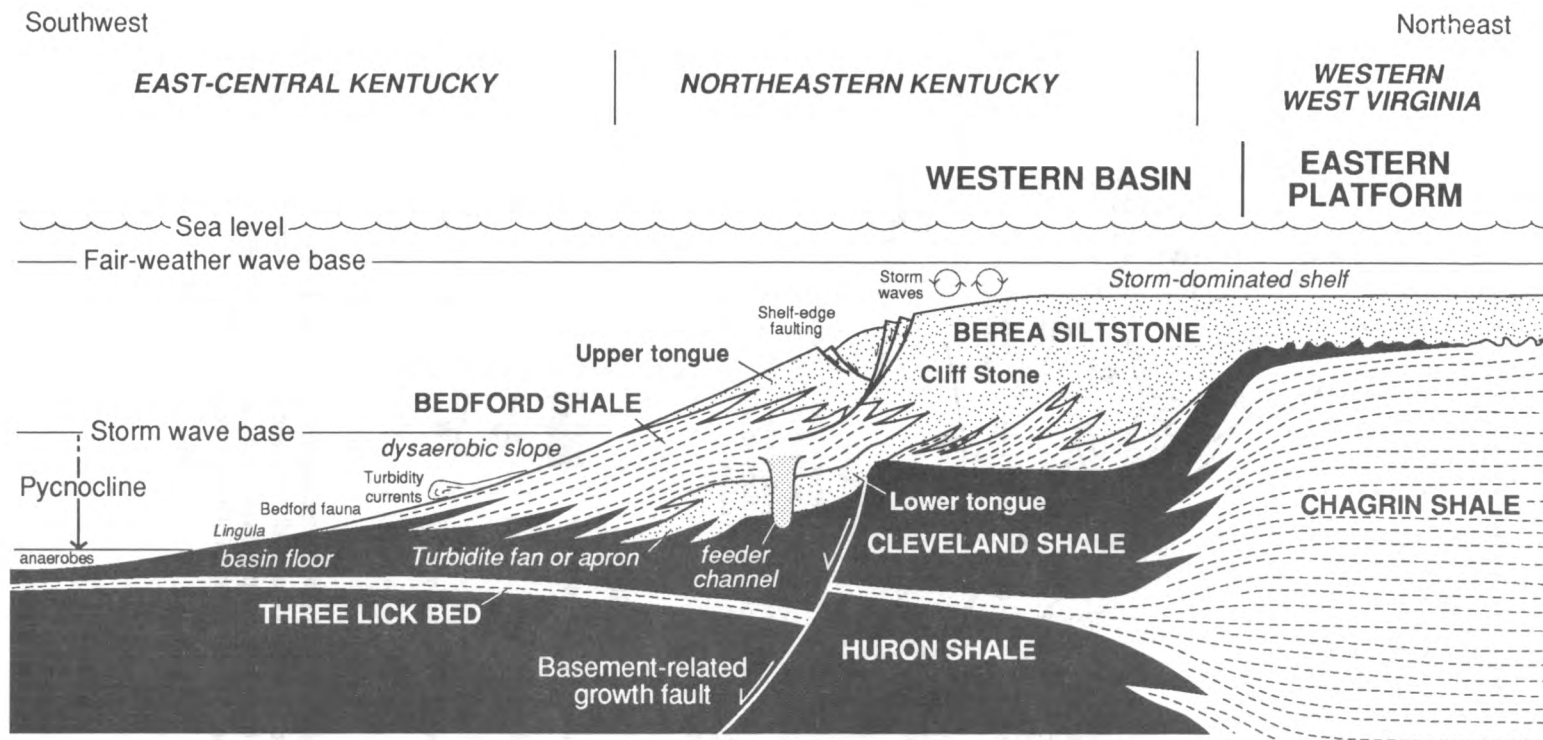


Figure 75. Depositional model for the Bedford-Berea sequence of eastern Kentucky and western West Virginia (after Pashin and Ettensohn, 1987, and Pashin, 1990). A thin, aggradational shelf sequence was preserved on the eastern platform, whereas a thick, progradational sequence that spans shelf, slope, and basinal environments was preserved in the western basin. Differentiation of platform and basin areas was related to relict topography and differential compaction of organic-rich black mud (Cleveland Shale) and relatively incompactible, organic-poor gray mud and silt (Chagrín Shale). Within the platform and basin areas, however, basement-fault reactivation was a significant control on facies distribution.

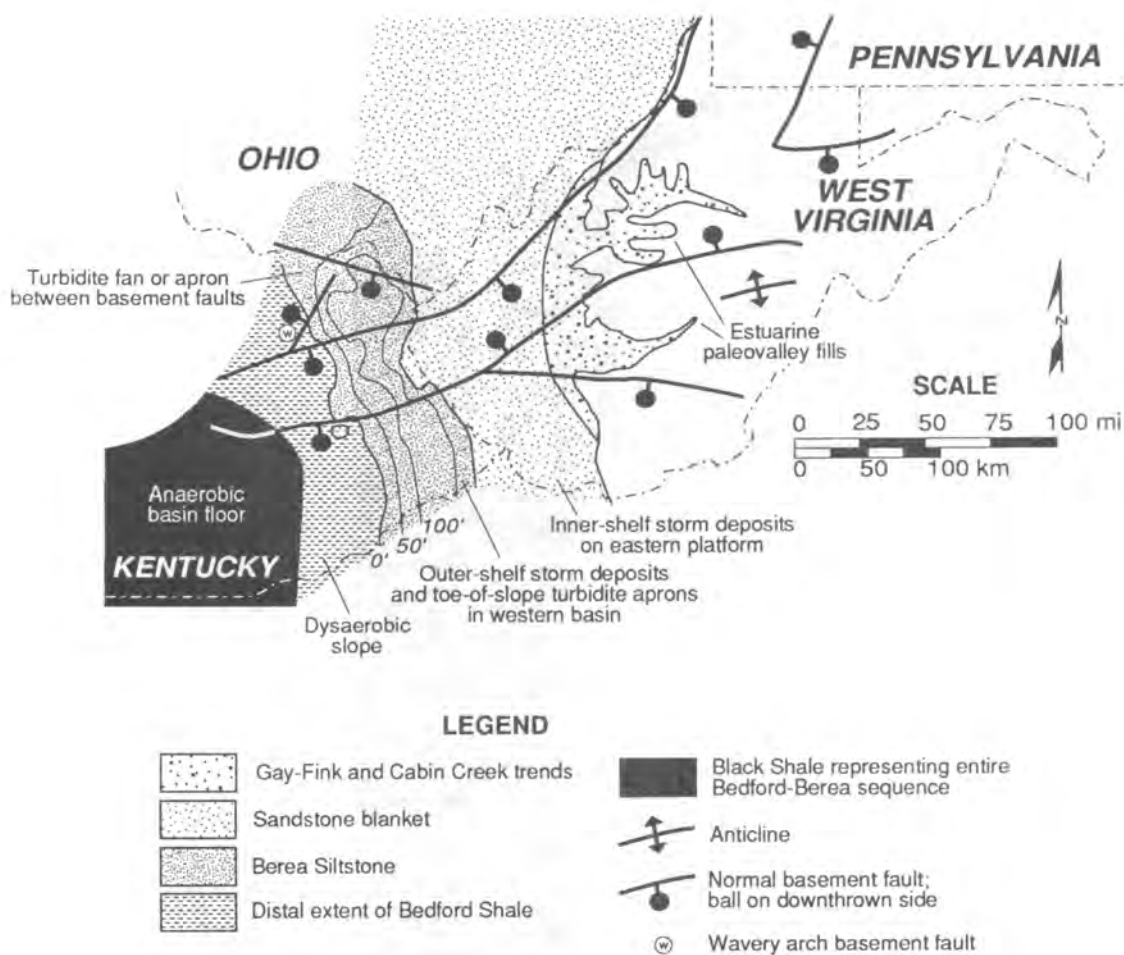


Figure 76. Bedford-Berea paleogeography, eastern Kentucky and West Virginia (after Pashin, 1990). Sand-rich estuary and shelf deposits predominated on the eastern platform, and silt- and mud-rich shelf, slope, and basinal environments predominated in the oxygen-deficient western basin. Structural control of sedimentation is apparent from preservation of the branching paleovalley-estuary deposit of the Gay-Fink trend between major basement faults in West Virginia and by local deflection of isopach contours in the Berea Siltstone of eastern Kentucky.

EASTERN PLATFORM

The Berea Sandstone accounts for nearly all of the Bedford-Berea sequence on the eastern platform of West Virginia (Figs. 76, 77). The sandstone is absent in eastern West Virginia, but in the central part of the state, the sandstone is pebbly and occurs in two northeast axes called the Gay-Fink and Cabin Creek trends (Pepper and others, 1954; Larese, 1974).

A detailed isopach map by Larese (1974) established that the Gay-Fink trend is branching and that the Cabin Creek trend is funnel-shaped (Fig. 77). Both trends contain sequences that fine upward from sandstone with quartz pebbles to gray shale (Larese, 1974), and the black overlying Sunbury Shale extends farther east within the trends than in adjacent areas (Pepper and others, 1954). The Gay-Fink trend, moreover, is bounded by the basement faults of the Rome trough (Fig. 76). The Gay-Fink and Cabin Creek trends extend into a widespread, westward-fining sandstone blanket that is 40- to 60-ft (12- to 18-m) thick and extends to the platform margin (Fig. 76). Potter and others (1983) determined that the blanket is composed of sheet-sandstone beds with shale intraclasts near the base and wave ripples at the top. In the easternmost part of the blanket, the sandstone contains quartz pebbles and heavy-mineral concentrations (Rittenhouse, 1946; Larese, 1974). The sandstone blanket generally overlies light-colored clastics of the Catskill wedge on the eastern platform, including the Chagrin Shale, but near the platform margin, the sandstone generally overlies a thin tongue of the Cleveland Shale (Pashin, 1990).

Previous workers stressed a fluvial-deltaic origin for the Gay-Fink and Cabin Creek trends (Pepper and others, 1954; Larese, 1974), but sedimentologic relationships indicate that the trends formed largely in an estuarine setting (Pashin, 1990). The branching geometry (Fig. 77) and fining-upward sequence of the Gay-Fink trend is typical of contributive fluvial systems, and the presence of the transgressive, basinal Sunbury Shale at the top of the sequence suggests that the trend is partly an estuary deposit.

Branching estuaries, like Chesapeake Bay, form by transgression and aggradation in deeply incised paleovalleys (Fairbridge, 1980); hence, reactivation of Rome trough basement faults gave rise to considerable topographic relief around the Gay-Fink trend. The Cabin Creek trend, in contrast, is interpreted to represent a funnel-shaped estuary. Funnel-shaped systems, such as the Gironde estuary of France, typically form in response to inundation of transitive coastal-plain channels and have a lower gradient than branching systems (Fairbridge, 1980).

The rippled sheet-sandstone beds, which compose the bulk of the sandstone blanket, have been interpreted to be storm-dominated shelf deposits (Potter and others, 1983), and sandstone with heavy-mineral concentrations on the eastern fringe of the blanket has been interpreted to be beach deposits (Larese, 1974).

Extension of the axial channels of the Gay-Fink and Cabin Creek trends into the shelf area indicates that the trends had a major constructive phase prior to estuary formation which may have culminated in regional exposure of the eastern platform, valley incision, and perhaps formation of platform-margin deltas. The shelf, beach, and estuary deposits apparently accumulated late in Bedford-Berea deposition in response to regional transgression and aggradation of the eastern platform.

WESTERN BASIN

The stratigraphy of the western basin differs markedly from that of the eastern platform, and the Bedford-Berea includes the Berea Siltstone, Bedford Shale, part of the Cleveland Member of the Ohio Shale, and part of the New Albany Shale (Fig. 74). Along the platform margin, the sandstone blanket passes into more than 100 ft (30 m) of Berea Siltstone (Figs. 75, 76). The siltstone intertongues with the Bedford Shale, forming a thick progradational package, and thins to a feather edge less than 50 miles from the platform margin. Southwest of the Berea pinchout, the Bedford Shale makes up most of the Bedford-Berea sequence in a belt up to 50-miles wide (Fig. 76), and southwest of that belt the sequence contains exclusively black shale (Ettensohn and Elam, 1985).

In the outcrop of northeastern Kentucky, the Berea Siltstone is divided into *upper* and *lower tongues* that are separated by a northeast-thinning wedge of the Bedford Shale (Morris and Pierce, 1967; Pashin and Ettensohn, 1987); where both siltstone tongues are combined, the Berea forms the *cliff stone* of Hyde (1953) (Figs. 74, 75). In east-central Kentucky, where the Berea is absent, the Bedford separates the Sunbury Shale from the Cleveland Member of the Ohio Shale and is locally mapped with the New Albany Shale (Fig. 74). Black shale equivalent to the Bedford-Berea has been recognized in the Cleveland Shale and the New Albany Shale in most of eastern Kentucky (Swager, 1979; Elam, 1981; Ettensohn and Elam, 1985).

The Berea Siltstone contains mostly thick sheet-siltstone beds with local ball-and-pillow structures; they are amalgamated or are separated by wavy, lenticular, and flaser-bedded shale and siltstone. Rippled sheet siltstone (Fig. 78) contains wave ripples and hummocky strata and is characteristic of the cliff stone and the upper tongue (Rothman, 1978; Potter and others, 1983; Pashin and Ettensohn, 1987); the wave ripples are renowned for abundance and consistent orientation (Hyde, 1911; Potter and Pettijohn, 1977). The lower tongue of the Berea contains two lithofacies, the lower-sheet-siltstone lithofacies, and the massive-siltstone lithofacies (Pashin and Ettensohn, 1987). In contrast to the upper tongue and cliff stone, the lower sheet siltstone is composed mainly of unrippled sheet-siltstone beds with gradational, bioturbated tops and complete Bouma sequences (Fig. 78). The massive siltstone is restricted to a series of outcrops near Walnut Grove Church (Optional Stop C) and is part of a channel-fill complex that truncates the lower sheet siltstone (Morris and Pierce, 1967; Pashin and Ettensohn, 1987).

In northeastern Kentucky, the Bedford Shale contains wavy-, flaser-, and lenticular-bedded shale and siltstone with wave ripples (Fig. 78). Most of the Bedford in Kentucky, however, is composed of poorly fissile gray shale containing thin, wavy, unrippled siltstone beds. Many of the siltstone beds are graded and contain a Bouma Tc,d,e sequence (Pashin and Ettensohn, 1987, 1992). A thin-shelled brachiopod-mollusc fauna called the Bedford fauna (Morse and Foerste, 1909) occurs at the base of the Bedford and in gray-shale tongues between black shale (Pashin and Ettensohn, 1992). Bedford-Berea black shale is brittle, fissile, and has been classified as a "ribbed," regressive, black shale by Ettensohn and others (1988b). The black shale locally contains abundant inarticulate brachiopods, particularly where black and gray shale intertongue. Like the sandstone blanket of West Virginia, the rippled siltstone beds of the upper tongue, cliff stone, and Bedford Shale have been interpreted to represent shelf storm deposits (Rothman, 1978; Potter and others, 1983; Pashin, 1985; Pashin and Ettensohn, 1987) (Figs. 75, 78). The cliff stone and upper tongue contain proximal storm deposits, whereas

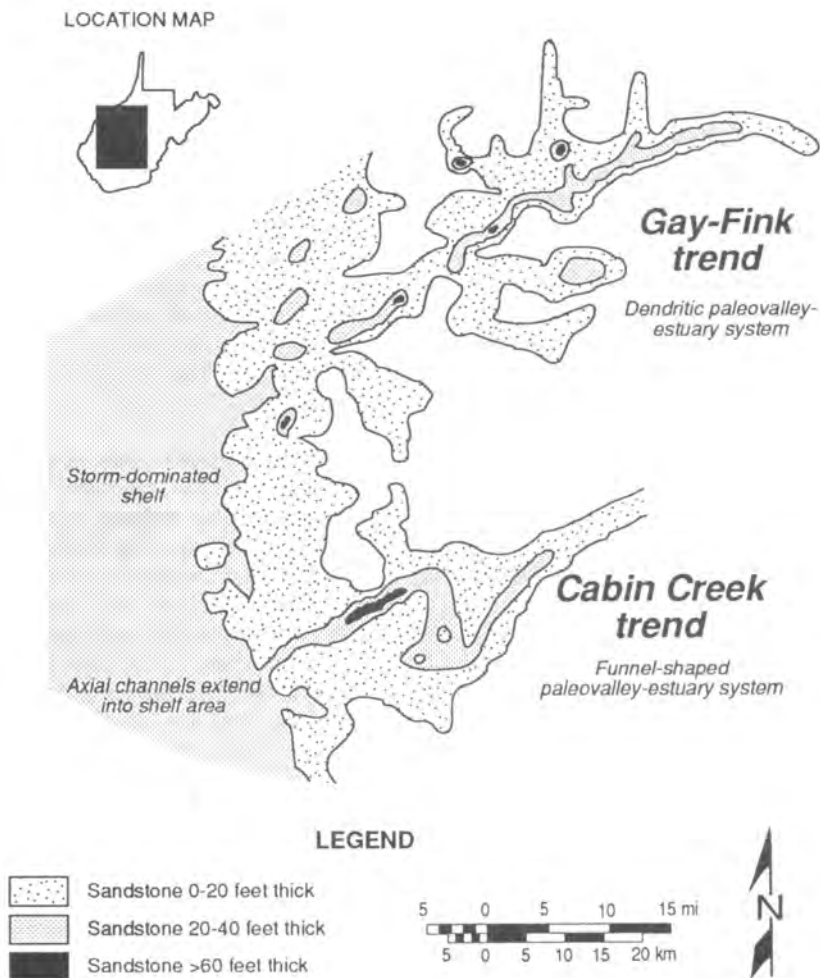


Figure 77. Isopach map of the Gay-Fink and Cabin Creek trends, West Virginia (after Larese, 1974). The Gay-Fink and Cabin Creek trends are interpreted to represent paleovalley-estuary systems. The Gay-Fink trend apparently represents a dendritic, or branching estuary, whereas the Cabin Creek trend apparently represents a funnel-shaped estuary. Branching estuaries are associated with inundation of deeply incised valleys, whereas funnel-shaped estuaries are associated with inundation of low-gradient parts of coastal plains.

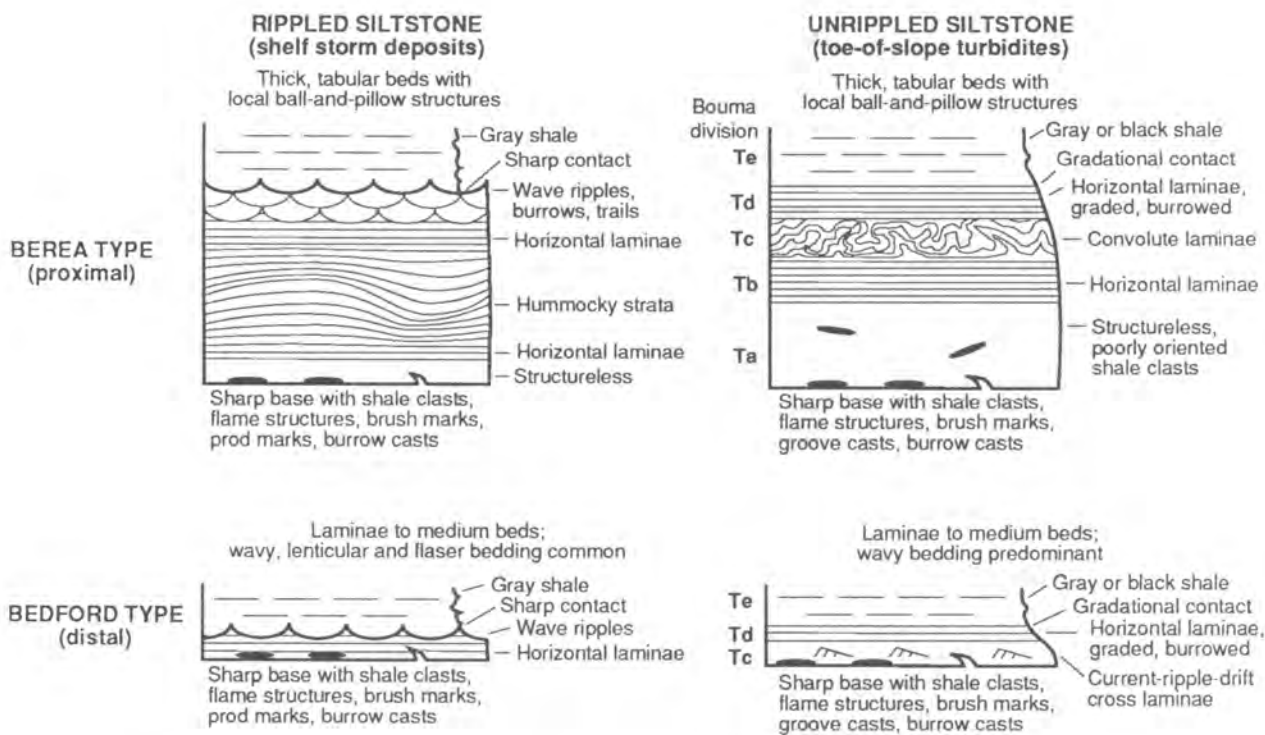


Figure 78. Idealized vertical sequences of sedimentary structures in Bedford-Berea sheet-siltstone beds. Bedford-Berea sheet siltstones are divided into rippled and unrippled types (Pashin and Ettensohn, 1987). Rippled sheet siltstone is interpreted to represent storm-dominated shelf deposits, whereas unrippled sheet siltstone is interpreted to represent turbidite-apron or fan deposits. Thick sheet-siltstone beds are proximal storm deposits and turbidites, and thin beds are distal examples.

the Bedford contains distal storm deposits. In the lower tongue, unrippled siltstone beds evidently compose a toe-of-slope turbidite apron or fan, and the massive siltstone apparently is a turbidite feeder-channel fill (Pashin, 1985; Pashin and Ettensohn, 1987).

In the Bedford, unrippled siltstone occurs at the base of mud-rich turbidites that accumulated as a series of dysaerobic, toe-of-slope aprons (Pashin and Ettensohn, 1987, 1992, figs. 5, 8). Bedford-Berea black shale, in contrast, is interpreted to represent pelagic sedimentation on an oxygen-deficient basin floor (Ettensohn and Elam, 1985; Pashin and Ettensohn, 1987). The organic-rich black mud was apparently too foul to accommodate benthos other than inarticulate brachiopods, but recycling of nutrients from black mud evidently provided a nutrient-rich habitat for the Bedford fauna along the terminal fringe of the mud-turbidite aprons (Pashin and Ettensohn, 1992).

The storm deposits and turbidites of the western basin contrast strongly with the aggradational shelf and estuary deposits of the eastern platform and represent a prograding outer shelf-slope system (Fig. 75). Much of the mud and silt in the western basin may have been derived initially from platform-margin shoal-water deltas as the Gay-Fink and Cabin Creek paleovalleys were incised. However, as the platform aggraded late in Bedford-Berea deposition, shelf progradation may have continued, and much of the silt and mud may have been transported basinward by storms that swept the platform. Rome trough basement faults apparently influenced sedimentation in the western basin, because isopach contours are deflected eastward along the fault traces (Fig. 76). Lower-slope sedimentation apparently was influenced strongly by basement faulting, because the turbidite apron of the lower tongue is bounded by two of the faults (Figs. 75, 76). According to Pashin and Ettensohn (1987), basement faulting culminated in formation of a southward-deepening shelf, but sedimentation eventually outstripped fault movement, giving rise to subdued topography at the close of Bedford-Berea deposition.

DISCUSSION

The occurrence of the Bedford-Berea sequence in the western part of the Appalachian basin (Fig. 73) and incision of the estuarine paleovalley fills of the Gay-Fink and Cabin Creek trends into the Catskill clastic wedge (Figs. 76, 77) indicate that the sequence represents a major downward shift of coastal onlap. Hence, the Bedford-Berea is the product of a significant lowstand that interceded Catskill and Pocono sedimentation. Pashin (1990) suggested that this sea-level drop was an interruption of the Acadian flexural-relaxation sequence that cannot be explained in terms of the regional tectonic framework, and Ettensohn (1990a) suggested a similar interpretation based on analysis of pycnocline migration patterns; therefore, the Bedford-Berea lowstand was probably a eustatic event. Although most Bedford-Berea sediment was probably derived from orogenic sources in what is now the Virginia promontory of the Appalachian orogen (Fig. 4), incision of the Gay-Fink and Cabin Creek trends into the Catskill wedge (Figs. 73, 76, 77) indicates that much of the sequence is reworked Catskill sediment that has been transported to the distal parts of the foreland basin.

Differentiation of the eastern platform and western basin apparently was a response to sedimentation atop the Catskill clastic wedge. Occurrence of the platform-basin transition above the facies change from gray, shelf-slope Chagrin Shale to black, basinal Ohio Shale (Figs. 73, 75) suggests that relict topography was an important controlling factor. However, differential compaction of organic-rich black mud and relatively incompressible, organic-poor, gray mud and silt is interpreted to account for much of the platform-basin transition.

In central Ohio, the platform-basin transition was further enhanced by reactivation of a Grenvillian suture in the basement, but no such structure has yet been identified along the transition in Kentucky and West Virginia (Pashin, 1990).

Although the platform and basin were formed mainly by relict topography and differential compaction, reactivation of Rome trough basement structures influenced facies distribution within both of these regions. On the eastern platform, structural control of the Gay-Fink trend is particularly conspicuous (Fig. 76). In the western basin, structural control apparently had the greatest influence early in Bedford-Berea deposition, particularly on turbidite deposition and formation of a southward-deepening shelf (Pashin and Ettensohn, 1987). As a mature, prograding shelf margin developed near the close of Bedford-Berea deposition, however, sedimentation outstripped basement-fault movement (Pashin and Ettensohn, 1987), and the shelf-to-basin transition had the subdued topography first envisioned by Rich (1951a, 1951b). The primary difference between this account of Bedford-Berea sedimentation and the classic studies of Rich (1951a, 1951b) and Pepper and others (1954) is the realization that the sequence was deposited in a tectonically evolving foreland basin. Eustasy, differential compaction, and relict topography functioned in concert with tectonism to determine depositional history and regional paleogeography. Tectonism, relict topography, and differential compaction acted collectively to provide sediment sources and to establish the geometry of the sedimentary basin and the architecture of the basin fill. Eustatic sea-level variation, moreover, helped determine the position and rate of change of base level and was thus a critical factor that caused erosion of the Catskill wedge and restriction of thick, progradational sequences to the western basin. Interplay of these factors has resulted in myriad patterns of sedimentation in the geologic record, and like the Bedford-Berea sequence, each depositional sequence has intricacies that must be identified before a thorough knowledge of sedimentation in foreland basins can be achieved.

STORM-DOMINATED SHELF MARGIN NEAR GARRISON

The Garrison area exhibits instructive exposures of the cliff stone (Figs. 74, 75) which include some of the best examples of shelf storm deposits in the Appalachian basin. At this stop, which is just south of Garrison along the Alexandria-Ashland Highway, we will examine the newest and most extensive exposure of the cliff stone and will observe a synsedimentary fault that allows new insight into the nature of the Bedford-Berea shelf margin. Approximately 100 feet (30 m) of the cliff stone is exposed along the Alexandria-Ashland Highway, and outcrops show a general thickening- and coarsening-upward progression from the upper few feet of the Bedford Shale to the top Berea Siltstone (Fig. 79); the black, fissile Sunbury Shale and the gray, poorly fissile shale of the Henley bed (Borden Formation) are exposed at the top of the roadcuts.

Rippled sheet siltstone is the most characteristic Bedford-Berea rock type and crops out from northeastern Kentucky to northwestern Pennsylvania (Pashin, 1990); the siltstone is even featured in classic studies of wave ripples (Hyde, 1911; Kindle, 1917; Bucher, 1919). Wave-ripple orientation is remarkably consistent in the exposure and ranges from 290° to 300° (Fig. 79). Hyde (1911) first noted this extreme uniformity of wave-ripple orientation, which is maintained into northern Ohio (Hyde, 1911; Lewis, 1988; Pashin, 1990). Comparing wave-ripple orientation from several depositional settings, Potter and Pettijohn (1977) suggested that variance of Bedford-Berea ripple orientation represents a lower limit for wave ripples in general.

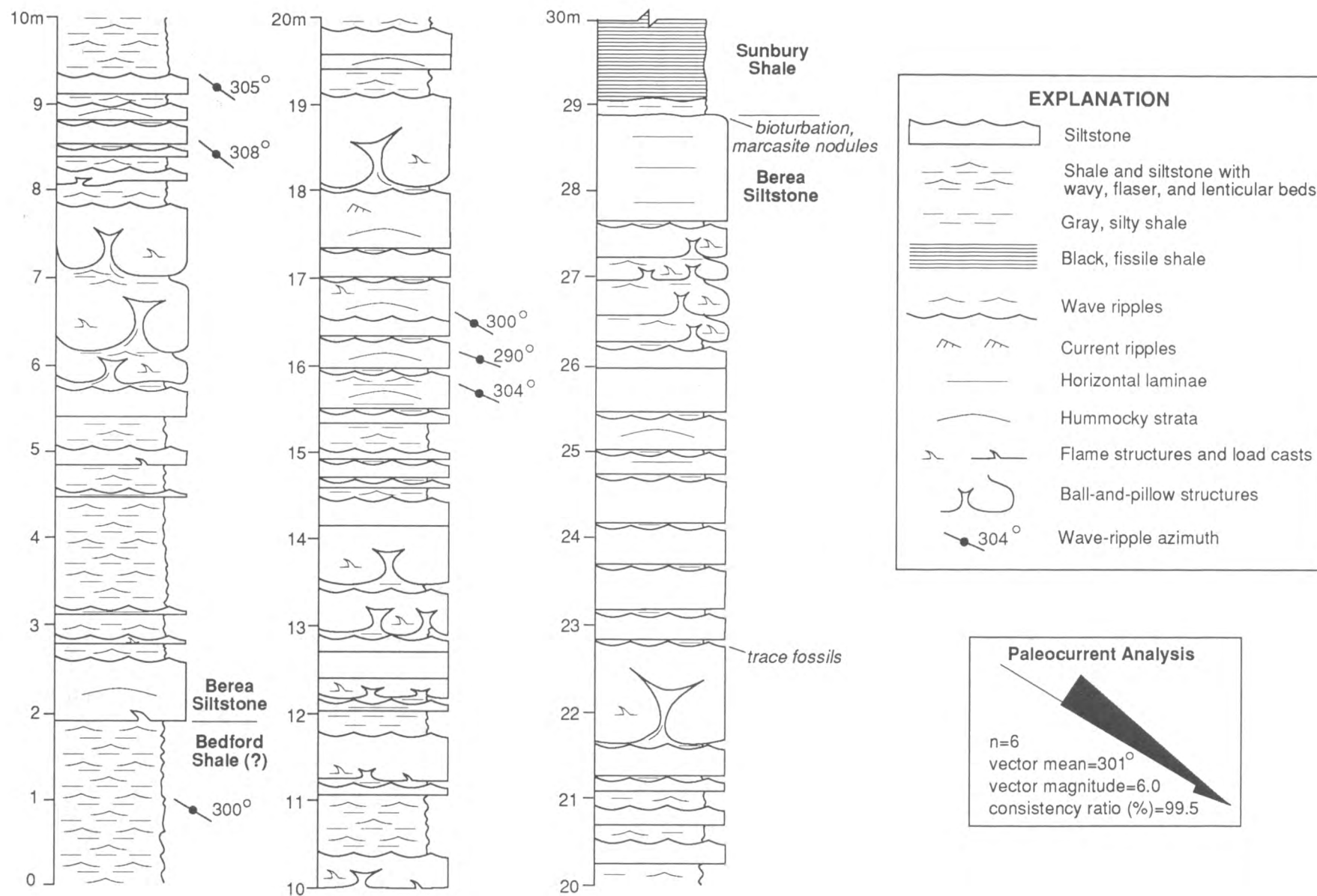


Figure 79. Measured section of Berea Siltstone (cliff stone), Alexandria-Ashland Highway (Stop 1, Day 3). This section shows the basic characteristics of rippled siltstone in the cliff stone. The exposure shows a well developed thickening- and coarsening-upward sequence from the upper part of the Bedford Shale to the top of the Berea Siltstone. Note the consistency of wave-ripple orientation in this exposure; uniform ripple orientation is maintained into northern Ohio.

Virtually all of the prominent features of the cliff stone can be observed along the Alexandria-Ashland Highway. Thick sheet-siltstone beds have sharp lower contacts with tool marks and load structures and ideally contain the following vertical sequence of sedimentary structures which is associated with shelf storm deposits (Goldring and Bridges, 1973; Hamblin and Walker, 1979; Dott and Bourgeois, 1982; Aigner and Reineck, 1982): 1.) structureless siltstone overlain by 2.) horizontal laminae that grade upward into 3.) hummocky strata. Hummocky strata grade back into 4.) horizontal laminae which are, in turn, truncated by 5.) wave-ripple cross laminae (Fig. 78); burrows and trails are abundant at the top of some beds. Thick sheet-siltstone beds are separated by intervals containing wavy, lenticular, and flaser-bedded shale and siltstone of Bedford type (Figs. 78, 79).

The sheet-siltstone beds are quite continuous at this stop, save for local ball-and-pillow zones (Figs. 79, 80, 81). However, near the southwest end of the roadcut sequence, the siltstone is disturbed by a fault that extends from the base of the exposure into the Sunbury Shale. The fault plane is irregular (Fig. 77) and strikes N. 20° E., approximately parallel to the Waverly arch basement fault as mapped by Pashin and Ettensohn (1987) (Fig. 76). The fault is downthrown to the southeast, and the fault plane is apparently listric, dipping 80° near the top of the Berea and only 50° at road level. At the Berea-Sunbury contact, net throw is approximately three feet (1 m), but near the base of the exposure, net throw is approximately six feet (2 m). Tensional fractures are numerous adjacent to the fault, particularly in the hanging wall.

Although ball-and-pillow structures are common throughout the cliff stone, some unusual forms are associated with the fault (Fig. 82). Isolated, boulder-size siltstone balls are common in both the hanging wall and the footwall, and many beds are pierced by diapiric shale masses as wide as three feet (1 m). In the footwall, some sheet-siltstone beds overlap and terminate. In the upper bench of the outcrop, one bed contains ball-and-pillow structures only in the hanging wall; the structures are asymmetrical and are elongate parallel to the fault trace (Fig. 81). Farther west along the Alexandria-Ashland Highway, some ball-and-pillow structures are overthrust and resemble imbricate duplexes.

Indeed, these unusual soft-sediment deformation structures occur where the cliff stone splits into the upper and lower tongues of the Berea and thus appear to be related to synsedimentary tectonism at the shelf margin (Fig. 75). Parallelism of the soft-sediment fault with the Waverly arch basement fault suggests deep-seated structural control of the shelf break. However, surface-fault displacement is to the southeast, opposite nearby basement structures and regional paleoslope, suggesting that it is an antithetic structure that is only part of a larger shelf-margin fault system. Regardless, increasing throw downward along the fault plane indicates that the fault grew during active sedimentation. Moreover, extension of the fault into the Sunbury Shale indicates that fault adjustment continued after Bedford-Berea deposition. Tensional fractures in the hanging wall contrast sharply with the abundant soft-sediment deformation structures and are interpreted to be a stress-release fracture system formed after faulting and lithification.

In addition to faulting, fluidized sediment failures occurred along the Bedford-Berea shelf margin (Cooper, 1943; Hyde, 1953; Pashin, 1990). Submarine sediment failures develop on slopes as gentle as 0.5° (Shepard, 1955) and commonly form in response to seismicity or storm-wave loading (Allen, 1982; Schwab and Lee, 1988). The

fluidized failures are near the contact of the rippled and unrippled beds and have thus been interpreted to be a result of storm-wave loading (Pashin, 1990). Parallelism of shelf-margin faults to the nearby basement faults, however, suggests an underlying seismic cause. Even so, deep-seated basement control does not preclude storm-wave action as a mechanism for adjustment on the faults, particularly on antithetic structures. Therefore, the shelf-margin fault system may reflect slope instability above a reactivated basement discontinuity, but storm-wave action may have caused minor fault movement and contributed to shallow seismicity, and hence, to some of the soft-sediment deformation preserved along the Alexandria-Ashland Highway.

Outcrops of the Bedford-Berea sequence included in this field trip (Stop 1 and Optional Stop C, Day 3) contain examples of shelf and slope sedimentation not typically associated with epeiric sea floors. The contrasting submarine feeder channel at Optional Stop B and shelf-margin faults (Figs. 75, 82) demonstrate that the epeiric shelf-to-basin transition contains a much more diverse facies assemblage than was first envisioned by Rich (1951a, 1951b) or has been accounted for in subsequent models. Complexities in the Bedford-Berea sequence are apparently related to oversteepened slopes along reactivated basement structures which gave rise to unusual facies assemblages and deformational structures. Continental crust contains numerous geophysical discontinuities that have diverse origins and orientations, which reflect a polyphase crustal history. The polyphase structures have evolved throughout geologic time and have acted collectively with depositional topography and relative sea-level variation to determine ancient patterns of sedimentation. For this reason, we are only beginning to glimpse the manifold paleomarine configurations of epeiric sea floors.

- 44.1 1.0 Junction of Kentucky State Routes 1306 and 546; continue west on 546.
- 44.6 0.5 Berea Sandstone exposed in roadcut on left side of the highway.
- 44.8 0.2 Junction of Spy Run Road and State Route 546. Continue west on State Route 546.
- 44.9 0.1 Berea Sandstone exposed in roadcuts on both sides of the highway with unusual soft-sediment deformation. Potter and others (1991, p. 24, 28) suggested that this deformation involved thrusting during soft-sediment deformation. We suggest, however, that the so-called thrusts reflect little more than erosional cutout and broad channel-fill relationships.
- 45.2 0.3 Berea Sandstone exposed in roadcuts on both sides of the highway.
- 45.5 0.3 OPTIONAL STOP B. Henley Bed (Farmers Member) of the Borden Formation (Lower Mississippian) in Northeastern Kentucky, along Kentucky State Route 546, Brightman Cemetery Section. (Exposure along State Route 546, 1.4 mi west of its junction with State Route 1306, Lewis Co.; Carter Coordinates: 2360' FSLx340' FWL, 23-Z-76). Berea Sandstone, Sunbury Shale, and the Henley bed and Farmers Member of the Borden Formation are exposed in roadcuts on both sides of highway.



Figure 80. Ball-and-pillow structures near base of Berea Siltstone, Alexandria-Ashland Highway (Stop 1, Day 3). Ball-and-pillow structures are abundant in the cliff stone, and several morphologic types are preserved in this sequence of highway cuts. The examples in this photograph show complete piercement of sheet-siltstone beds by diapiric shale masses.



Figure 81. Asymmetrical ball-and-pillow structures adjacent to fault near top of Berea Siltstone, Alexandria-Ashland Highway (Stop 1, Day 3). Fault reactivation apparently was penecontemporaneous with sedimentation, and many unusual soft-sediment deformation structures are preserved adjacent to the fault plane.

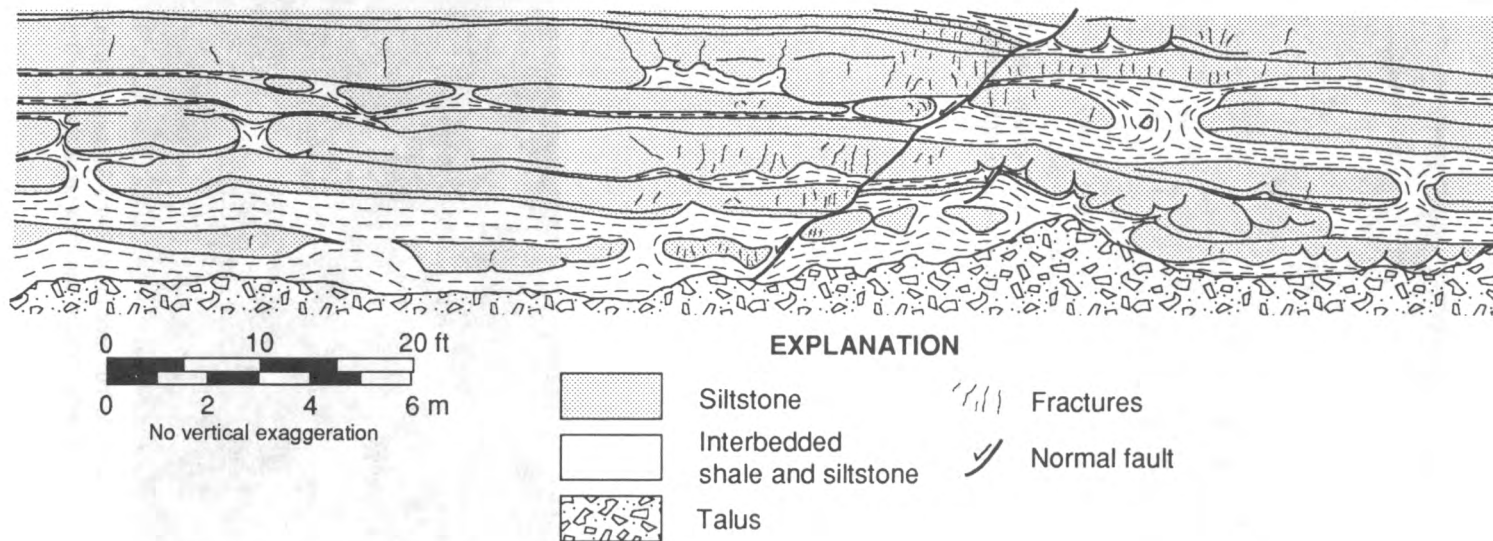


Figure 82. Diagram of synsedimentary fault in near base of Berea Siltstone, Alexandria-Ashland Highway (Stop 1, Day 3). Isolated, boulder-size siltstone balls are common along the fault, and in the footwall, some sheet-siltstone beds overlap and terminate. Tensional fractures are numerous near the fault, particularly in the hanging wall. This fault is interpreted to be an antithetic structure that is part of a larger system of shelf-margin faults. The fault system is interpreted to be related to reactivation of nearby basement structures, and storm-wave loading may have contributed to surface faulting and associated soft-sediment deformation.

**HENLEY BED (FARMERS MEMBER) OF THE BORDEN
FORMATION (LOWER MISSISSIPPIAN) ALONG KENTUCKY
STATE ROUTE 546 IN NORTHERN KENTUCKY**

CHARLES E. MASON AND R. THOMAS LIERMAN

A total of 120 ft (36.7 m) of section (Fig. 83) is exposed in the roadcut on the north side of State Route 546. Formations exposed include: the upper 5.2 ft (1.6 m) of the Berea Sandstone, the entire Sunbury Shale which is 14.3-ft (4.4-m) thick here, and the lower part of the Farmers Member of the Borden Formation. The Henley Bed, which is found in the lower part of the Farmers Member, comprises 64 ft (19.5 m) of the 100 ft (30.7 m) measured for the Farmers at this section.

The focus of this stop is the Henley Bed of the Farmers Member. Although it is 0.7-ft (0.2-m) thicker at Stop 1, it is much better exposed here and is more accessible. At Stop 3, which is slightly over five aerial miles to the west of this location, the Henley is only 34-ft (10.4-m) thick. Along I-64 in Rowan County 34 mi to the southwest (Stop 5, Day 2), the Henley is 5.2-ft (1.6-m) thick and the Farmers Member, including the Henley, is only 33-ft (10-m) thick (Fig. 49). The Henley also thins to the north in Ohio where it is given member status (Chaplin and Mason, 1978). Thus the Henley thins both to the north and south of this locality, as well as to the west. These and other thickness relationships of the Henley Bed in this region, combined with evidence from interbedded and overlying proximal turbidite beds (to be discussed at Stops 2, 3), strongly suggest that this area was a local depocenter, at least during late Kinderhookian and early Osagean time. Pashin and Ettensohn (1987) have noted in previous work and at Stop 1, Day 3, that the Berea and Bedford formations also show a similar trend in this area and have suggested that the depocenter may be related to downthrow on two nearby faults (Figs. 49, 76).

Lithologically, the Henley can be described as a grayish-green, poorly fissile, silty shale which weathers yellowish-gray. Present within the lower portion of the Henley (lower 1.1 ft or 0.5 m) are crystals of pyrite and phosphate nodules. Thin siltstone beds are found in the Henley especially toward the middle of the unit. Maroon shale intervals are also found interspersed within these predominately greenish-gray shales. These maroon shales are identical to the green shales with respect to grain size and clay-mineral composition. Color is the only difference, the red color being related to the presence of finely dispersed iron oxide (hematite). The source of the iron oxide found in the maroon intervals has not been examined sufficiently to allow comment at this time. Compositionally the clays found in the Henley consist of kaolinite, mixed-layer illite/smectite, and Fe-rich chlorite (see Table 3).

The Henley also contains a nodular or rubbly dolostone layer towards its base. This layer is 1.2-ft (36-cm) thick and is four to five inches (10-12 cm) above the contact with the Sunbury Shale (Fig. 83). The dolostone can best be described as a muddy (insoluble residue = 27%) dolomitic stone that possesses an idiomatic fabric in which finely crystalline, euhedral dolomite crystals are imbedded in an argillaceous matrix. The dolomite crystals themselves are ferroan with a weight percent of FeCO_3 ranging from 10 to 14 percent. This layer does not appear to be the result of dolomite replacing a previously existing carbonate, as there is no evidence of any relict carbonate textures, and the dolomite crystals exist as rhombs within a muddy matrix. It also appears to have formed prior to substantial compaction because

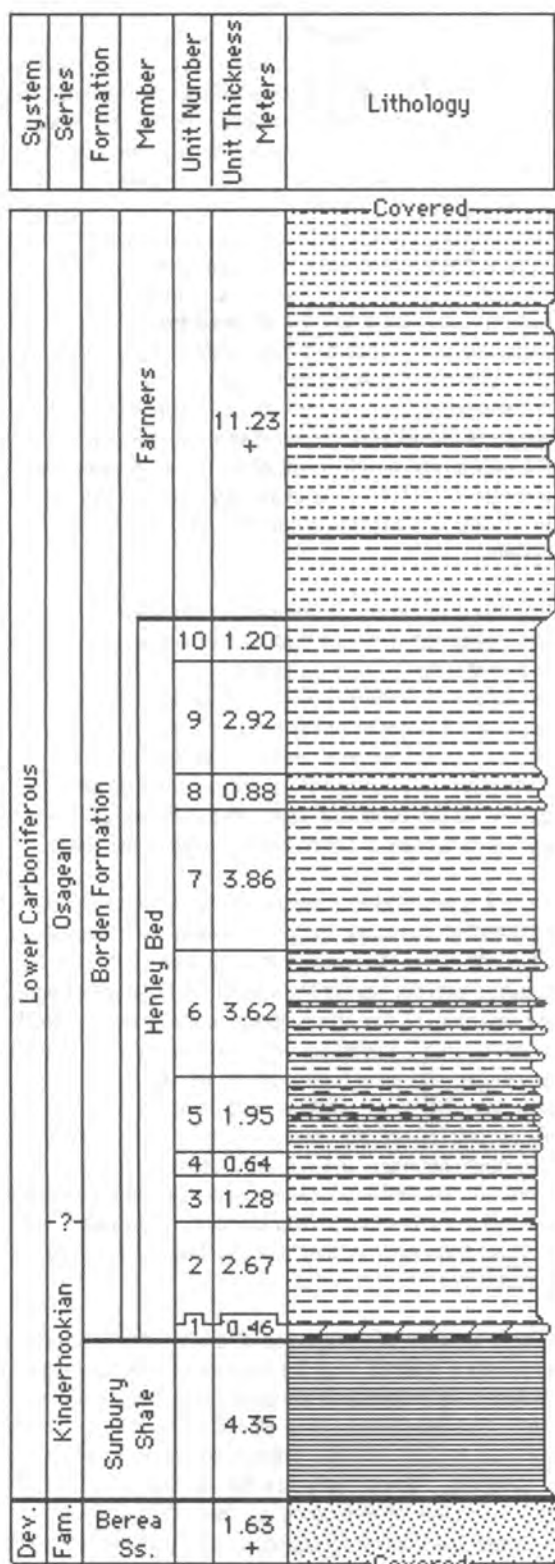
of the presence of three-dimensional burrows which are absent in the surrounding shales due to compaction.

A possible interpretation for this dolostone layer is somewhat difficult considering its stratigraphic position in the midst of exclusively deep-water shales. Two possible models include the "Burial Compaction Model" of McHargue and Price (1982), or the "Organogenic Sea-floor Model" of Baker and Burns (1985). In the first model, Mg^{++} rich interstitial fluids are generated during the compaction of the surrounding shales and during the conversion of smectite to illite with burial. These Mg-rich brines then dolomitize any carbonate sediments they circulate through. In the second model, dolomite is generated at or near the sediment-water interface by the circulation of seawater through the underlying sediments. Magnesium is supplied by seawater circulation, CO_2 by the breakdown of organic matter in the sediment, and calcium by the dissolution of any calcium carbonate in the sediment. One problem with this model is the presence of sulfates which have a high concentration in seawater ($\approx 7.7\%$ by weight of the dissolved solids in sea water) and which greatly inhibit the growth and development of dolomite. If seawater manages to circulate through the zone of surface reduction, sulfates are reduced to sulfides such as pyrite or marcasite and therefore removed from the system.

Objections to the Burial Compaction Model of McHargue and Price rest mainly with evidence that the dolomite probably formed near the sediment-water interface, prior to substantial burial and compaction of the layer. Also the existence of a calcium-carbonate precursor is necessary to account for this model. So far, no clear evidence of this exists. The Organogenic Sea-floor Model of Baker and Burns also relies upon the existence of a calcium-carbonate precursor, but more closely fits the conditions associated with this unit. Certainly more work is necessary to determine the origin of this layer.

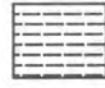
Table 4 is provided to show the Borden terminology for rock units in northeastern Kentucky and how it relates to terminology in southern Ohio. At this section units 1 through 4 of the Henley Bed (Fig. 83) correlate with the Henley Member of the Cuyahoga Formation in the western part of the Mississippian outcrop belt in southern Ohio, whereas unit 5 correlates with the Buena Vista Member of the Cuyahoga, and units 6 through 10 are correlative with the Rarden Member. The tabular siltstone beds of the Farmers Member, situated above unit 10 at this section, constitute the basal part of the Vanceburg Member. In Ohio, with respect to this part of the Borden Formation, the first encountered sequence of tabular siltstone beds is referred to as the Buena Vista Member. The Buena Vista is applied to these beds whether they are only five in number or more than a hundred.

If there is no second sequence of tabular siltstone beds, the entire sequence is referred to as the Buena Vista Member regardless of its thickness, as is done with the lower Borden Sequence in the eastern part of southern Ohio. The Buena Vista is here overlain by the Portsmouth Shale Member which is equivalent to the Nancy Member in Kentucky. Where there are two sequences of tabular siltstone beds, as in the western part of southern Ohio, the second sequence is termed the Vanceburg Member. This member is overlain by the Churn Creek Member, which is equivalent to the Nancy Member of the Borden Formation in Kentucky. Overlying the Churn Creek Member in the western part and the Portsmouth Member in the eastern part of southern Ohio is the Logan Formation. The Logan Formation in southern Ohio correlates with the Cowbell Member of the



Day 3, Optional Stop B Brightman Cemetery Section

Legend

-  Argillaceous Dolostone
-  Black Fissile Shale
-  Mudstone
-  Siltstone
-  Fine-grained Sandstone

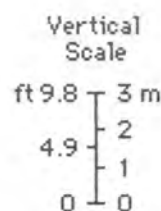


Figure 83. Uppermost Devonian-Lower Mississippian section at Brightman Cemetery along Kentucky State Route 546 (Optional Stop B, Day 3).

Table 3. Composition and clay mineralogy of the Henley Bed (Farmers Member) at the Brightmen Cemetery Section.

Sample No.	Height above Contact (m)	Percent Clay	Percent Silt	Percent Organics	Clays Normalized to 100%				Illite XI.
F	19.50 m	0.27	0.73	0	8.32	59.43	9.96	22.29	0.616
E	14.40 m	0.35	0.65	0	7.17	57.37	7.72	27.74	0.614
D	7.00 m	0.43	0.57	0	8.01	57.07	7.57	27.35	0.633
C	5.00 m	0.36	0.64	0	6.06	64.87	5.13	23.94	0.543
B	0.60 m	0.17	0.83	0	11.93	53.85	10.55	23.67	0.643
A1	0.35 m	0.57	0.43	0	6.64	69.31	6.64	17.41	0.485
A	0.10 m	0.27	0.73	0	8.12	58.78	6.57	26.53	0.479
Average		0.35	0.65	0	8.04	60.097	7.73	24.13	0.570

Table 4. Nomenclature for Mississippian and Devonian units in northeastern Kentucky and southern Ohio

System	Series	Northeastern Kentucky Ky. Geological Survey & U.S. Geological Survey, GQ Maps		Southern Ohio (West) Vanceburg Province, Hyde (1953)		Southern Ohio (East) Scioto Valley Province, Hyde (1953)	
Mississippian	Lower Mississippian	Slade Formation		Maxville Limestone		Maxville Limestone	
		Borden Formation	Nada Member	Rushville Shale		Rushville Shale	
			Cowbell Member	Logan Fm.	Vinton Member Allensville Member * Byer Member	Logan Fm.	Vinton Member Allensville Member * Byer Member
			Nancy Member	Cuyahoga Fm.	Churn Creek Mbr.	Cuyahoga Fm.	Portsmouth Member
			Farmers Member		Vanceburg Member		Buena Vista Member
			Henley Bed		Rarden Member Buena Vista Member Henley Member		Henley Member
		Sunbury Shale Berea Sandstone Bedford Shale		Sunbury Shale Berea Sandstone Bedford Shale		Sunbury Shale Berea Sandstone Bedford Shale	
	Upper Devonian	Ohio Shale	Cleveland Shale Member Three Lick Bed Huron Member	Ohio Shale	Cleveland Shale Member Three Lick Bed Huron Member	Ohio Shale	Cleveland Shale Member Three Lick Bed Huron Member
		Olenitangy Shale		Olenitangy Shale		Olenitangy Shale	

* Where Allensville is absent, the Logan Formation is not subdivided. This is typical of southernmost Ohio just north of Optional Stop B, Day 3.

Borden Formation in northeastern Kentucky. The Logan Formation in southern Ohio is presently undifferentiated, because the Allensville Member, critical for the differentiation of this formation into members, is absent in this region.

Both micro- and megafossils have been identified in the Henley Bed along State Route 546. The major sections of Henley, examined in detail thus far, include this stop and Stop 3. Conodonts are common to abundant in these intervals, and based on their study to date, the Lower Mississippian, Kinderhookian-Osagean boundary occurs between 2.6 and 13.1 ft (0.8-4.0 m) above its base. Other than conodonts, the microfossil fauna is sparse in the silty shale intervals, where only a few arenaceous foraminifera have been recovered. This boundary is well defined, with the *Siphonodella isosticha*-Upper *crenulata* Zone representing the uppermost Kinderhookian and the *Gnathodus typicus* Zone representing the lower Osagean. Along I-64 at Stop 5 on Day 2, this same boundary occurs between 0.8 and 1.1 ft (0.3-0.5 m) above its basal contact with the Sunbury Shale. This boundary in the Morehead area is based on conodonts (Chaplin, 1982) and palynomorphs (Coleman and Clayton, 1987). The difference in the placement of the Kinderhookian-Osagean boundary between these two areas reflects a higher rate of sedimentation in the Lewis County area, as well as a slightly older age for the prograding Borden sequence.

Megafossils occurring in the basal four to five inches (10-12 cm) of the Henley Bed along State Route 546 in Lewis County are the first ever reported from this interval of strata in northeastern Kentucky. This macrofauna is predominantly composed of juveniles, is low in diversity and high in abundance, is mollusk-dominated, and is replaced by pyrite. The above points conform well with the major criteria defined by Kammer and others (1986) for dysaerobic faunas found in Devonian and younger rocks. The megafossils found in the Henley thus far are relatively small in size and include ammonoids, nautiloids, gastropods, bivalves, brachiopods, corals, and crinoids. Ammonoids are the most abundant and diverse of the mollusks and three ammonoid genera appear to be present. The only genus which has been identified with certainty to date is *Protocanites*. This ammonoid fauna is not only the first fauna of Kinderhookian age found in Kentucky, but it is also the most eastern occurrence of such a fauna in North America. Also, this is the first occurrence of *Protocanites* in midcontinental rocks of Kinderhookian age.

Additional megafossils found by Joe Gilbert (pers. commun.) in phosphate nodules occur within a 1.6-ft (0.5-m) interval approximately 6.6 ft (2.0 m) above the base of the Henley Bed. The nodules vary from spherical to ovoid to cylindrical in shape. They are small in size, generally four inches (10 cm) or less along their long axis, and the fauna is not only extremely sparse, but low in diversity as well. Megafossils found to date include lingulid and orbiculoid brachiopods, gastropods, *Sphenothallus* (phosphatic tubes), and some unidentified vertebrate remains.

- 45.8 0.3 Spy Run Cemetery Road on right hand side.
- 45.9 0.1 Henley Bed and Farmers Member of the Borden Formation exposed in roadcut on right side of the highway.
- 46.2 0.3 Berea Sandstone, Sunbury Shale, and Henley Bed of the Borden Formation exposed in roadcut on right side of the road. Note contact of Berea Sandstone and Sunbury Shale at road level.

- 46.5 0.3 Berea Sandstone exposed in roadcut on left side of the highway.
- 46.7 0.2 Berea Sandstone, Sunbury Shale, and Henley Bed of the Borden Formation exposed in roadcuts on right side of the highway.
- 47.0 0.3 Sunbury Shale and Henley Bed of the Borden Formation exposed in roadcut on right side of the highway.
- 47.2 0.2 Berea Sandstone, Sunbury Shale, and Henley Bed of the Borden Formation exposed in roadcut on right side of the highway.
- 47.6 0.4 Henley Bed of the Borden Formation exposed in roadcut on right side of the road.
- 47.8 0.2 Henley Bed and Farmers Member of the Borden Formation exposed in roadcuts on right side of the highway (roadcut extends for some 0.25 mi).
- 48.2 0.4 Farmers Member of the Borden Formation exposed in roadcuts on left side of road.
- 48.4 0.2 Farmers, Nancy, and Cowbell members of the Borden Formation exposed in roadcuts on both sides of the highway (roadcut extends for some 0.4 mi).
- 49.3 0.9 STOP 2. Farmers, Nancy, and Cowbell Members of the Borden Formation (Lower Mississippian) in Northeastern Kentucky Along Kentucky State Route 546, Griffin Hollow Section (Along State Route 546, 4.9 mi west of its junction with State Route 1306, Lewis Co.; Carter Coordinates: 1600'FSLx160'FWL, 21-Z-75). Farmers, Nancy, and Cowbell members of the Borden Formation are exposed in roadcuts on both sides of the highway (roadcut extends for some 0.35 mi). Bus will pull off to the left side of Route 546 onto a small gravel road, and we will disembark here and proceed downhill on foot. Examine the rock exposures along the way. Field-trip participants will be picked up by the bus at bottom of roadcut.

FARMERS, NANCY, AND COWBELL MEMBERS OF THE BORDEN FORMATION (LOWER MISSISSIPPIAN) ALONG KENTUCKY STATE HIGHWAY 546 IN NORTHEASTERN KENTUCKY

R. THOMAS LIERMAN AND CHARLES E. MASON

COWBELL MEMBER

The Cowbell Member of the Borden Formation in the Garrison area ranges in thickness from 220 to 340 ft (67 to 104 m). At this location only the lower 40.5 ft (12.4 m) of the Cowbell Member is exposed. It consists predominantly of argillaceous siltstone with shale interbeds. The siltstone is coarse-grained, olive-gray, and weathers from a light-gray to a reddish-brown color. It occurs in irregular, thin to medium beds which can display a number of sedimentary structures such as small-scale, trough cross-laminae, ripple-drift laminae, as well as cut-and-fill structures. The shale portion is an olive- to greenish-gray, silty shale which weathers to light gray. Contained within this unit are

siderite nodules, lenses, and beds which are scattered throughout the shale intervals. These siderite deposits are locally fossiliferous. The unit is commonly stained with iron oxides and hydroxides particularly along fractures and joints.

Megafossils can be found scattered throughout the unit and are concentrated in distinct layers. Fossils recognized to date include horn corals, bryozoans, brachiopods, echinoderm debris, mollusks, and trilobite fragments. Specimens occur as both internal and external molds, as well as casts, but fossils preserved as internal molds predominate. The dominant replacement minerals infilling voids within the fossils are barite, sphalerite, and galena. Mason and Gilbert (1983) reported on the unique replacement and infilling of fossils by galena in northeastern Kentucky. Fossil material can also be replaced or infilled with pyrite or calcite.

Trace fossils are quite abundant and diverse in the Cowbell Member. The ichnogenera most commonly encountered include: an *Archaeichnum*-like form, *Arthropycus*, *Bifungites*, *Calycraterion*, *Chondrites*, *Cruziana*, *Cylindrichnus*, *Diplocraterion*, *Gyrochorte*, *Helminthopsis*, *Helminthoida*, *Lophoctenium*, *Monocraterion*, *Plaerophycus*, *Phycosiphon*, *Planolites*, *Phycodes*, *Rusophycus*, *Scalarituba*, *Teichichnus*, and *Zoophycus* (Potter and others, 1991). Chaplin (1982) placed the trace-fossil assemblage found in the lower part of the Cowbell member in the *Cruziana* ichnofacies.

The Cowbell Member is thought to represent the deposition of fine-grained clastic sediments under aerobic conditions in a proximal delta-front setting.

NANCY MEMBER

The Nancy Member of the Borden Formation in the Garrison area ranges in thickness from 55.5 to 180 ft (17-55 m). At this locality the Nancy Member is approximately 55.5-ft (17-m) thick. It consists of a bluish-gray to greenish-gray, silty shale which weathers olive gray to yellowish gray. The member is bioturbated to the point of almost complete homogenization, which imparts a poorly fissile character to the unit. Also present within the Nancy are siderite nodules or lenses which are scattered randomly throughout parts of the unit and elsewhere concentrated along distinct horizons. Thin-bedded siltstones to very fine-grained sandstones are also interspersed throughout the unit, and are light brownish-gray to greenish-gray, argillaceous, and occur in very thin to thin beds. These beds appear to have originated as turbidites and exhibit sharp erosional bases and tops which grade into the overlying shales. Internally, the sand and siltstone beds show predominantly parallel laminae, though some ripple cross-laminae can be seen. Locally the bases of the beds show small flute casts and tool marks.

The siderite nodules found in the Nancy are thought to have formed immediately below the sediment-water interface prior to sediment compaction as discussed previously (Stop 5, Day 2). They contain megafossils which occur as molds, casts, and/or mineralized replacements. Void spaces within these nodules are most commonly infilled with barite, sphalerite, and minor amounts of galena. Fossils which have been collected from the Nancy include solitary horn corals, fenestrate bryozoans, brachiopods, conularids, gastropods, cephalopods, crinoid debris, *Sphenothallus* (phosphatic tubes), and rare trilobites. Articulate brachiopods and cephalopods are the most abundant megafossils found in the nodules. The ammonoids found in these nodules and those found in the overlying lower part of the Cowbell Member belong in the *Muensteroceras oweni* faunal

assemblage. This faunal assemblage is early Osagean (late Tournaisian) in age (Gordon and Mason, 1985). *Muensteroceras oweni* is the most abundant ammonoid species found at this locality, as well as in this part of northeastern Kentucky. The only microfossils observed to date in the nodules are ostracods.

Trace fossils are generally poorly preserved. Those which have been identified at this stop occur primarily along the soles and tops of the siltstone beds contained within the Nancy and include the following form genera: *Bifungites*, *Chondrites*, *Conichnus*, *Cylindrichnus*, *Helminthoida*, *Rhizocorallium*, *Scalarituba*, *Teichichnus*, and *Zoophycus*. Chaplin (1982) placed this trace-fossil assemblage in the lower *Zoophycus* ichnofacies.

The Nancy Member is thought to represent the deposition of silty muds and occasional distal turbidites in a distal prodelta setting. This seems to have occurred under dysaerobic (lower part) to predominantly aerobic conditions (middle and upper parts).

FARMERS MEMBER

The Farmers Member of the Borden Formation in the Vanceburg area ranges in thickness from 130 to 260 ft (40-80 m). At this location the Farmers Member is approximately 130-ft (40-m) thick. It consists of fine-grained sandstone and siltstone beds with thin intervening shale partings. The sandstone beds are yellowish-brown to light olive-gray, very fine- to fine-grained sublitharenites. They vary from thin- to thick-bedded. Siltstone beds are medium- to light-gray in color and consist of coarse-grained silts which are poor to moderately sorted. Individual sand and siltstone beds display sharp erosional bases and grade upward into overlying shale interbeds. Shales are greenish-gray, moderately fissile and silty. Noticeably absent within the shale intervals here are the siderite nodules so obvious at the Ramey Section of Day 2, Stop 5. The ratio of shale to sandstone beds increases up-section.

Sedimentary structures are quite evident in the sandstone and siltstone beds. Internal stratification within these beds include parallel laminae, ripple-drift laminae, as well as convolute and wavy laminae. Just as at Stop 5 on Day 2, internal sedimentary structures are generally arranged in such a way to form parts of recognizable Bouma sequences. This appears to mainly include the divisions Tc-Te, though a few of the more complete cycles show a Tb-Te sequence (Fig. 84). One of the main differences between this section and the Ramey Section (Day 2, Stop 5) is the presence of lags along the base of some of the sandstone beds. These lag deposits tend to consist of shale, siltstone clasts, and fossil debris. There is also a tendency for many of the sandstone beds to truncate or cut through underlying layers and exhibit a lenticular geometry. Amalgamated beds or stacked, multiple sandstone bodies are likewise quite common at this section. These features were not observed at the Ramey Section, but their presence here may reflect the more proximal position of the area relative to the likely source (Fig. 49).

Shale interbeds, the Te intervals, are again thought to be hemipelagic in origin. They probably represent both allochthonous sediments (which settled out of suspension shortly after the passage of a turbidity current) and fine-grained autochthonous sediments (which represent the normal accumulation of pelagic marine muds).

Bedding-plane structures within the sandstones are poor to moderately developed and occur on the soles of individual sandstone beds. These include a variety of tool marks such as groove casts,

Day 3, Stop 2, Griffin Hollow Section

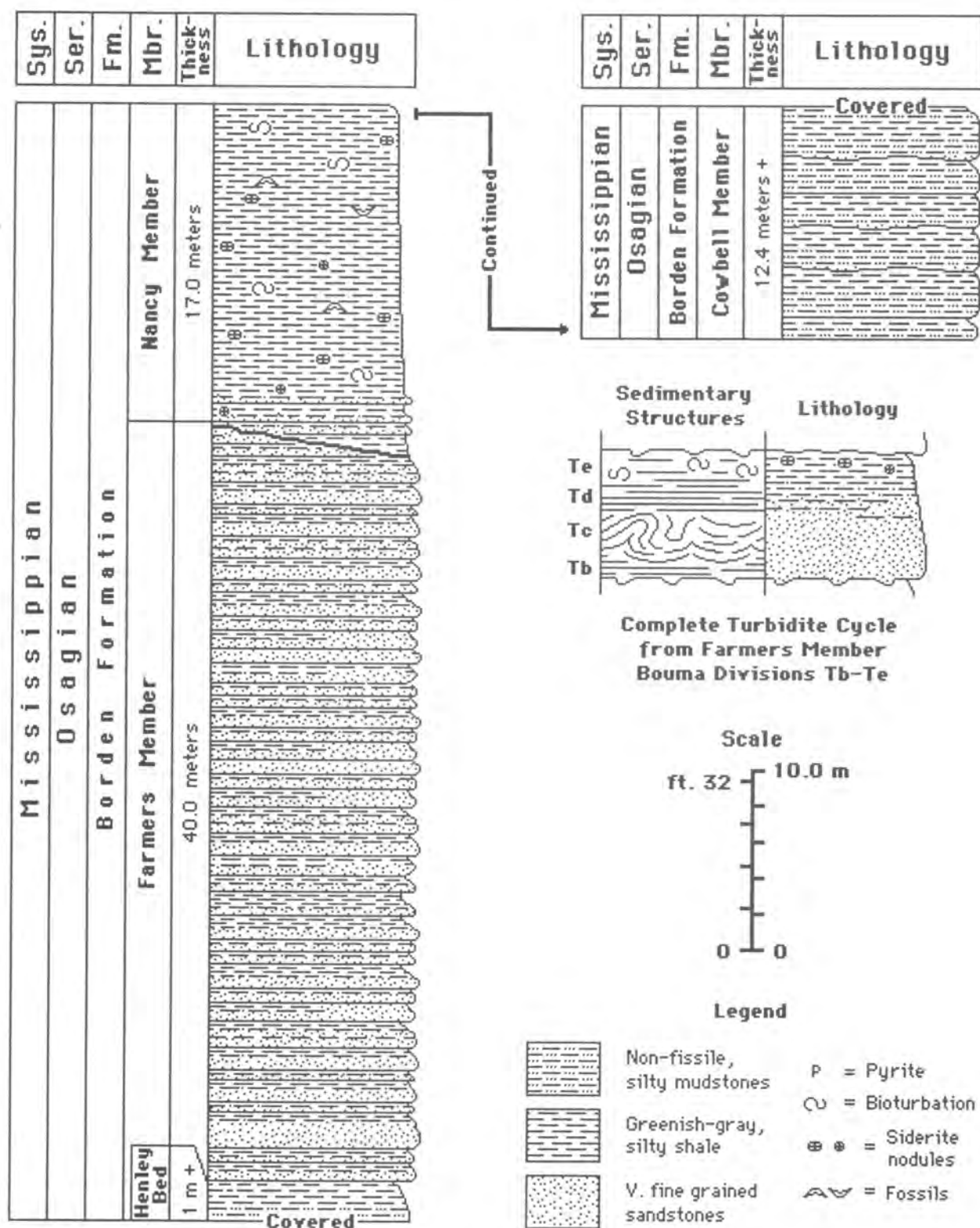


Figure 84. Lower Borden section and sedimentary structures at Griffin Hollow along Kentucky State Route 546 (Stop 2, Day 3).

prod casts, brush, bounce, and roll marks; scour marks such as flute casts; and load casts. The few measurements that have been made here suggest a 300° paleocurrent direction. This again would suggest down-slope movement of material from the region of the Appalachian fold belt to the west.

Trace fossils are quite abundant here but less diverse than at the Ramey Section (Day 2, Stop 5). They can generally be characterized as complex, horizontal, feeding burrows and trails. The ichnogenera most commonly occur on the tops and soles of individual siltstone or sandstone beds. Those recognized at this outcrop include: *Bifungites*, *Cylindrichnus*, *Helminthoida*, *Helminthopsis*, *Neonereites*, *Paleodictyon*, *Scalartuba*, and *Zoophycus* (Chaplin, 1980). This trace-fossil assemblage has been interpreted by Chaplin (1982) as belonging to the *Nereites* ichnofacies.

Scattered body fossils which have been recognized in the Farmers include broken and abraded brachiopods, fenestrate bryozoans, echinoderm debris, gastropods, pelecypods, and cephalopods. Most of these specimens occur as lags along the base of individual sandstone beds and are thought to have been transported by turbidity currents from the clastic shelf, where they lived, to deeper portions of the basin. Ammonoids are common in these lag deposits but are also found less commonly in the middle and upper parts of the turbidite beds. It is believed they were part of the nektonic fauna that lived in the upper water column overlying the basin floor and shelf. When they died some would sink to the basin floor where they were preserved in the pelitic sediments. Others would die over the shelf, only to be transported later into the basin by turbidity currents.

Based on examinations to date, the Farmers Member at Griffin Hollow represents the accumulation of more proximal turbidites at the base of the Borden deltaic sequence. Shale interbeds are thought to represent accumulation of hemipelagic muds under generally dysaerobic conditions. Evidence for a proximal setting of the turbidite sequence (Fig. 49) found in this part of northeastern Kentucky includes: 1.) a greater thickness of the entire turbidite sequence; the Farmers is more than 131-ft (40-m) thick here, whereas at the Ramey Section at Stop 5 on Day 2, it is only 28-ft (8.4-m) thick; 2.) an increase in the total number of turbidite beds making up the Farmers Member in this area (nearly 170 beds have been estimated in this area, compared to the Ramey Section where only 20 turbidite beds are present); 3.) the maximum thickness of the sandstone beds in this area is nearly 5.6 ft (1.7 m) whereas the thickest bed noted at the Ramey Section is only 3.1-ft (0.9-m) thick; 4.) the presence of lag deposits along the base of individual sandstone beds (basal lags are only observed in the eastern part of the outcrop belt in the Morehead area, closer to the source area); 5.) the lenticular geometry of individual sandstone channels and their tendency to truncate underlying beds (this is not observed in the Morehead area); 6.) a decrease in the thickness of the hemipelagic shales in this area versus an increased shale thickness in the Morehead area (the shales here are on average two-inches or 5-cm thick; at the Ramey Section they average nearly 5.5-in or 14-cm thick); this is also evident when one examines the shale-to-sandstone ratio (at Griffin Hill the sh./ss. ratio = 0.24; at Ramey the sh./ss. ratio = 0.46; 7.) the absence of siderite nodules in the shale interbeds in this area (this suggests the pelitic interval was absent or was deposited over a much shorter duration than in the Morehead area, which would be expected in an area closer to the depositional center of a large turbidite fan; 8.) a reduction in the diversity and the abundance of trace-fossil genera in the Lewis County area, and an increase in faunal diversity and abundance in the Morehead area (the Lewis County area has yielded

only six to seven ichnogenera from the Farmers Member, whereas the Morehead area has yielded over 20 ichnogenera; Chaplin, 1980); and 9.) the top of the Farmers at Griffin Hollow is highly irregular and erosional in character with a relief greater than 3.3 ft (1.0 m) in cutouts which penetrate up to six turbidite beds; many of these cutouts may represent fan-head channels which would be more likely to occur in proximal areas (Fig. 49).

The Farmers Member along State Route 546 is thought to represent proximal turbidites alternating with pelitic sediments deposited primarily under aerobic conditions on the middle part of a submarine fan (Fig. 49).

HENLEY BED OF THE FARMERS MEMBER

The Henley Bed is located in the basal part of the Farmers Member of the Borden Formation and is included within it. In the Vanceburg area it ranges in thickness from five to 64.5 ft (1.5-19.7 m). At this location only the upper 3.3 ft (1.0 m) is exposed. It consists of a medium-gray to greenish-gray, poorly fissile, silty shale which weathers to a medium-brown color. The upper contact with the overlying Farmers is quite sharp. More complete descriptions of the Henley can be found at Optional Stop B and at Stop 3 of today's trip. The Henley once again represents the extremely slow accumulation of hemipelagic muds in the deepest parts of the depositional basin, under dysaerobic conditions.

- 49.7 0.4 Farmers Member of the Borden Formation exposed in roadcuts on right side of the highway (roadcut extends for some 0.2 mi).
- 49.9 0.2 Junction of Kentucky State Routes 546 and 1149 on the left. Continue west on Route 546.
- 50.1 0.2 Henley Bed and Farmers Member of the Borden Formation exposed in roadcuts on right side of the highway (roadcut extends for some 0.2 mi).
- 50.4 0.3 Henley Bed and Farmers Member of the Borden Formation exposed in roadcuts on right side of the highway (roadcut extends for some 0.3 mi).
- 50.8 0.4 Junction of Kentucky State Routes 546 and 1149 on right-hand side. Continue west on Route 546.
- 50.9 0.1 Farmers Member of the Borden Formation exposed in roadcuts on right side of the highway (roadcut extends for some 0.4 mi).
- 51.0 0.1 Farmers and Nancy members of the Borden Formation exposed in roadcuts on both sides of the highway.
- 51.3 0.3 STOP 3. Cowbell Member, Nancy Member, Farmers Member and Henley Bed of the Borden Formation, Sunbury Shale (Lower Mississippian), and the Bedford-Berea Sequence (Upper Devonian) in Northeastern Kentucky along State Route 546, Evans Chapel Section (Along Kentucky State Route 546, 0.5 mi west of its junction with State Route 1149, Lewis Co.; Carter Coordinates: 3000'FNLx100'FEL, 4-Y-75). Bedford Shale, Berea Sandstone, and Henley Bed and Farmers Member of the Borden Formation exposed in roadcut on left side of the highway (roadcut extends for some

0.7 mi). Bus will pull off to right side of Route 546. Passengers will disembark here and proceed downhill on foot. Examine the rock exposures along the way. Field-trip participants will be picked up by the bus at bottom of roadcut.

**COWBELL MEMBER, NANCY MEMBER, FARMERS MEMBER
AND HENLEY BED OF THE BORDEN FORMATION, SUNBURY
SHALE, AND THE BEDFORD-BEREA SEQUENCE ALONG
STATE ROUTE 546 IN NORTHEASTERN KENTUCKY**

R. THOMAS LIERMAN, CHARLES E. MASON, JACK C. PASHIN
AND FRANK R. ETENSOHN

COWBELL MEMBER

The Cowbell Member of the Borden Formation in this area ranges up to 340 ft (104 m) in thickness. At this location only the lowermost Cowbell, approximately 6.5 ft (2 m), is exposed at the top of the roadcut (Fig. 85). It consists predominately of argillaceous siltstone with minor shale interbeds. The siltstone appears to be coarse-grained and medium-gray to greenish-gray in color, weathering to a light-gray or buff color. Shales are light greenish-gray, silty, and tend to weather to a light olive-gray color. For a more complete description of the Cowbell Member, see the previous stop.

NANCY MEMBER

The Nancy Member of the Borden Formation in the Vanceburg area ranges in thickness from 51 to 180 ft (15.7-55 m). At this location the Nancy is approximately 51-ft (15.7-m) thick (Fig. 85). It consists of a medium-gray, poorly fissile, silty shale, which weathers to an olive-gray or yellowish-gray color. Also present are thin beds of siltstone and very fine-grained sandstone. These are light-brownish gray, argillaceous, and appear to be thin turbidite beds that have sharp erosional bases and tops which grade into the overlying shales. Internally, the sand and siltstone beds show parallel laminae with some ripple cross-stratification. Both load casts and minor tool marks are commonly encountered on the soles of these beds. Also present within this member are siderite nodules, lenses, and beds which have been discussed previously at Stop 5, Day 2 and at Stop 2, Day 3. The contact with the overlying Cowbell Member is gradational.

This unit is highly bioturbated but most of the ichnogenera identified in the Nancy occur on the soles and tops of turbidite beds found within the unit. Chaplin (1982) placed this trace-fossil association within the *Nereites* ichnofacies. Ichnogenera collected from the Nancy Member to date include *Bifungites*, *Chondrites*, *Helminthoida*, *Helminthopsis*, *Scalarituba*, and *Zoophycus*. Within shaly portions of the Nancy, identification of individual traces is difficult, due to extreme bioturbation and compaction of the shale. Scattered body fossils tend to occur as molds, casts, or mineralized replacements within siderite nodules. Megafossils found within the Nancy at Evans Chapel include brachiopods, bryozoans, conularids, pelecypods, gastropods, cephalopods, and crinoid debris. To date this member along State Route 546 has not been examined for microfossils. The only microfossils which have been observed are ostracods preserved as molds in siderite nodules.

The Nancy Member is thought to represent the deposition of fine-grained sediments and occasional distal turbidites in a prodelta

setting. Deposition occurred under dysaerobic conditions in the lower part of the Nancy but ranged into aerobic conditions in the middle and upper parts.

FARMERS MEMBER

The Farmers Member of the Borden Formation in the Vanceburg area ranges in thickness from 118 to 262 ft (36-80 m). At this location the Farmers Member, excluding the interval of the Henley Bed, is approximately 118.4-ft (36.1-m) thick (Fig. 85). It consists predominately of fine-grained sandstone and siltstone beds separated by thin beds of shale. The sandstone beds are a light brownish-gray and very fine- to fine-grained. They are evenly bedded and are thin- to thick-bedded. Siltstone beds are medium- to light-gray in color and consist of moderately sorted, coarse-grained silts. Individual sandstone and siltstone beds generally have sharp erosional bases and grade upward into overlying shale interbeds. These shales are greenish-gray, moderately fissile and silty. The upper contact with the overlying Nancy Member is sharp and planar, and not as irregular as that observed at Griffin Hollow.

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

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

Trace fossils are quite abundant and diverse and can generally be characterized as complex horizontal, feeding burrows and trails. A list of the ichnogenera found in this region can be seen in the description of the Farmers Member at Stop 2. Megafossils which have been recognized in the Farmers include broken and abraded brachiopods, fenestrate bryozoans, echinoderm debris, gastropods, and mollusks. Most of these specimens occur as lags along the bases of individual sandstone beds and are thought to have been transported by turbidity currents.

In review, we believe that the sandstone beds in the Farmers Member at this location were deposited as proximal turbidites. The shales are thought to represent an accumulation of hemipelagic muds. The turbidite sequence at Evans Chapel is placed a few miles to the west of the main depocenter and the relative paleogeographic position of the Farmers Member at this section is shown in Figure 49. Also included in this figure are the relative positions of the Farmers Member at the Griffin Hollow Section and the Ramey Section. There are slight differences between the sequence here and that at Griffin Hollow. These differences include: 1.) a slight decrease in the thickness of the overall turbidite sequence at Evans Chapel versus that at Griffin Hollow (the Farmers is 118-ft or 36-m thick here; 131-ft or 40-m thick at Griffin Hollow); 2.) the turbidite beds making up the Farmers Member at this stop are generally thinner; 3.) individual sandstone

Day 3, Stop 3, Evans Chapel Section

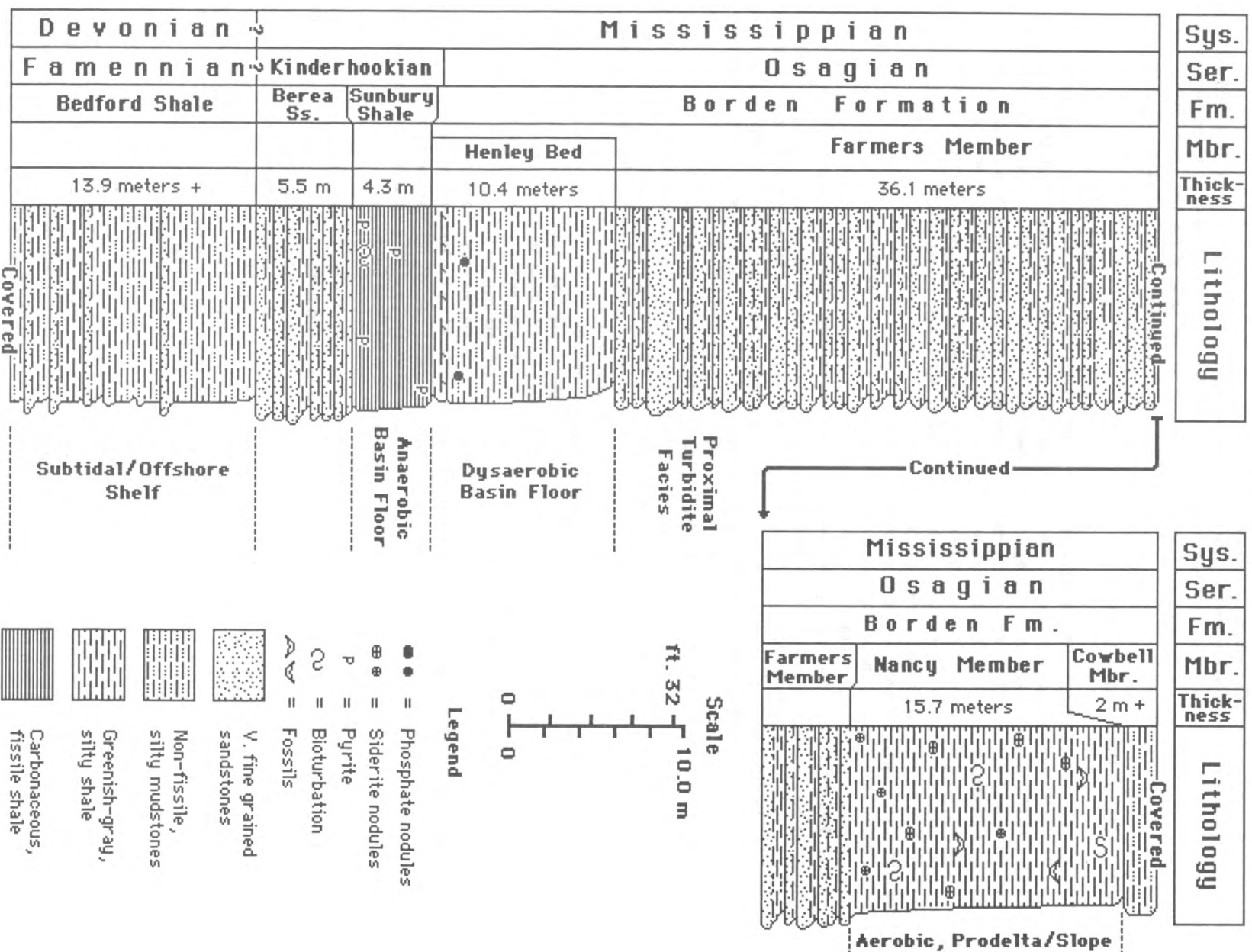


Figure 85. Uppermost Devonian-Lower Mississippian section and environmental interpretations at Evans Chapel on Kentucky State Route 546 (Stop 3, Day 3).

beds are more tabular and are less likely to truncate the underlying units (this was more readily observed at Griffin Hollow); 4.) there are noticeably fewer lag deposits along the base of individual sandstone beds here than at Griffin Hollow; 5.) shale interbeds are also slightly thicker here than at Griffin Hollow and the ratio between shale and sandstone (sh./ss.) is higher; and 6.) the contact between the Farmers and the overlying Nancy is sharp, but not the highly irregular and eroded surface seen at Griffin Hollow.

HENLEY BED OF THE FARMERS MEMBER

The Henley Bed is included within the Farmers Member of the Borden Formation (Fig. 85). In the Vanceburg area it ranges in thickness from ten to 40 ft (3 to 12.1 m). At this location the Henley Bed is approximately 34-ft (10.4-m) thick (Fig. 85). Here it consists mainly of a medium-gray to greenish-gray, poorly fissile, silty shale which weathers to a medium-brown color. A number of distinctive layers are maroon in color, presumably due to the presence of finely disseminated hematite in the shale, as discussed earlier at Optional Stop B today. The clay mineralogy of this shaly interval is essentially the same as that recorded at Brightman Cemetery (Optional Stop B) some 5.5 mi to the east. In addition to the thinning of the shale intervals, the Henley Bed at this stop contains a reduced number of very fine-grained, argillaceous sandstone beds. These sandstone beds are quite tabular, thinly bedded, and have sharp erosional bases. Load casts, small flute casts, and some tool marks are found on the soles of these beds. Internally, thin parallel laminae predominate. Their upper surfaces are quite bioturbated and grade into the overlying shales. The upper contact of the Henley with the overlying Farmers Member is quite sharp.

A layer of argillaceous dolostone approximately four to five inches (10-12 cm) above the basal contact with the Sunbury Shale is also present in the Henley Bed at this locality (Fig. 85). This layer is about 14-in (36-cm) thick and can be described as a muddy dolomitic. The dolomite is ferroan and occurs as euhedral to subhedral rhombs within an argillaceous matrix. Again, there is no evidence of relict carbonate textures within this dolostone, and the layer appears to have formed prior to compaction and burial of the surrounding shales. This conclusion is again based on the presence of three-dimensional burrows within the dolostone which are absent in the surrounding shale.

The fossil content of the Henley Bed at this stop is essentially the same as that discussed at Optional Stop B. The only exception is that the number of phosphate nodules is greatly reduced, and to date, those that have been found here are unfossiliferous.

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

SUNBURY SHALE

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

composed of pyrite and phosphatic fossil debris. Vertical and horizontal burrows, infilled with greenish-gray shale of the Henley, are found in the upper eight inches (20 cm) of the Sunbury.

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

The Sunbury Shale is a transgressive black shale (Ettensohn and others, 1988b) representing extremely slow accumulation of organic-rich hemipelagic muds in the deepest parts of the depositional basin, generally under anoxic conditions. Evidence for this is found in the discussion of the Ramey Section (Stop 5, Day 2). The decreased thickness of the Sunbury in this area may reflect the fact that it accumulated under a stable pycnocline higher on the shelf or shelf margins where less accommodation space was available (see Fig. 75).

BEREA SANDSTONE

The Berea Sandstone in this area ranges in thickness from 14.3 to 145 ft (5-44.2 m). At this location the Berea is approximately 14.3-ft (5-m) thick. It consists of light-gray to buff-colored sandstones with thin interbeds of greenish-gray shale. Compositionally, these are fine-grained, micaceous quartzarenites which are moderately well indurated. The sandstones are medium bedded with sharp erosional bases that have small load casts, as well as the casts of tracks and trails on their surfaces. The upper surfaces of several beds display numerous straight-crested oscillation ripples. Ripple crests show a mean orientation of approximately 300°, which suggests a bidirectional current trend or line of direction of 30°. Sedimentary structures observed within these beds include primarily parallel laminae and small ripple cross-laminae. The Berea Sandstone, especially in this area, shows a strong facies relationship with the underlying Bedford Formation. No fossils other than horizontal and vertical trace fossils have been reported from the Berea in this area. A note of special interest here is that the Berea Sandstone is quite similar to the Farmers Member of the Borden Formation and is hard to distinguish from it in isolated outcrops; however, the Farmers Member always contains an abundance of the trace fossil *Zoophycus*, whereas it is never found in the Berea Sandstone.

The predominance of storm sedimentation during Berea deposition may have precluded habitation by the *Zoophycus* animal. As at Stop 1, the Berea here largely exhibits sheet-siltstone beds typical of storm-dominated shelf deposition. However, the fact that the Berea has thinned to approximately one-sixth of its thickness at Stop 1, 8.5 mi to northeast, indicates that we are moving off the shelf edge to the south and west where the upper tongue of the Berea thins and eventually pinches out into the Bedford Shale (Fig. 75).

BEDFORD SHALE

The Bedford Shale in this area ranges in thickness from 0 to 90 ft (0-27.4 m). At this location the upper 52 ft (16 m) of the Bedford is exposed, and it consists of silty shales which are intercalated with thin siltstones and subordinate beds of sandstone. This silty shale is greenish-gray to purple-gray in color and weathers yellowish-gray. Discontinuous interbeds of siltstone are light-gray to greenish-gray in color and occur as very thin flaser-, wavy-, and lenticular-bedded layers. A planar view of the lenticular layers indicates that they are

basically starved ripples. Side views commonly show ripple-drift laminae. Interspersed among these finer grained sediments are medium to thin, planar beds of sandstone. These are light-gray to buff in color, very fine- to fine-grained, silty and micaceous quartzarenites. Both upper and lower contacts are sharp, and the upper surfaces of many sandstone beds display straight-crested oscillation ripples. Ripple crests show a mean orientation of 296°. Pyrite and marcasite commonly occur as crystal aggregates, small scattered crystals, or irregular nodules in this unit. The upper contact of the Bedford with the overlying Berea appears to be somewhat gradational. The lower contact here is not exposed.

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

The presence of wavy-, flaser-, and lenticular-bedded shale and siltstones with wave ripples in the Bedford indicates distal storm deposition on the slope between the eastern platform and western basin (Fig. 75) (Pashin and Ettensohn, 1987). The presence of thick, wave-rippled siltstones near the base of the exposed Bedford at this locality may reflect the beginning of the lower tongue of the Bedford, which is mapped in this area (Morris and Pierce, 1967). Although the lower tongue of the Berea is typically composed of unrippled sheet-siltstone beds (Pashin and Ettensohn, 1987), the presence of ripples in these uppermost beds of the unit may reflect continued storm influence at or just above wave base (Fig. 75).

- 52.5 1.2 Huron Member, Three Lick Bed, and Cleveland Member of the Ohio Shale exposed in roadcuts on right side of the highway (roadcut extends for some 0.4 mi).

- 53.1 0.6 Junction of Kentucky State Routes 546 and 59. OPTIONAL STOP C. Berea Turbidite Feeder Channel Near Walnut Grove Church (This optional side trip begins five miles south of the junction of State Routes 546 and 59 at the point where State Routes 59 and 344 meet at Walnut Grove Church. This side trip is a 2.5-mile transect through the lower tongue of the Berea and a massive siltstone feeder channel, and consists of a series of seven exposures [Fig. 86] that are best examined in a south-to-north transect along State Route 59. Lewis Co.; Carter Coordinates [outcrop 1]: 1800'FSLx450'FEL, 20-Y-74).

BEREA TURBIDITE FEEDER CHANNEL NEAR WALNUT GROVE CHURCH

JACK C. PASHIN AND FRANK R. ETTENSOHN

The first stop in the Bedford-Berea sequence is near Walnut Grove Church, in the Vanceburg 7.5-minute Quadrangle, and is a transect through the lower tongue of the Berea Siltstone along State Route 59 which contains the lower-sheet-siltstone and massive-siltstone lithofacies (Figs. 86, 87). Strata similar to the lower-sheet-siltstone lithofacies have been identified in the Bedford Shale as far away as northeastern Ohio (Pashin, 1990), but the thickest succession of these deposits occurs at this stop. More importantly, the massive siltstone is known only from this series of roadcuts. Morris and Pierce (1967) mapped the lower tongue of the Berea, identified the massive siltstone, and recognized its channel-fill origin. Their map shows that

the upper tongue of the Berea is absent above the channel fill, and although strata above the massive siltstone are poorly exposed, contorted gray shale and isolated, boulder-size siltstone balls crop out along the steepest slopes.

Approximately 40 ft (12 m) of the lower-sheet-siltstone lithofacies is exposed at the southern end of the transect (outcrop 1), which is at the intersection of Kentucky State Routes 59 and 344 (Fig. 86). This facies is composed predominantly of thick sheet-siltstone beds having sharp bases with tool marks and load structures and gradational tops; ball-and-pillow structures are common. Ideally, these beds contain the following vertical succession of sedimentary structures which makes up a Bouma turbidite sequence: 1.) a structureless basal interval with sparse shale clasts, 2.) horizontal laminae, 3.) convolute laminae with flame structures, 4.) horizontal laminae, and 5.) a gradational top that is commonly penetrated by feeding burrows (Fig. 78). The beds are either amalgamated or are separated by up to two feet (0.7 m) of wavy-bedded shale and siltstone of the Bedford type; the siltstone beds are typically graded and commonly contain current-ripple-drift cross laminae. At the northwest end of outcrop 1, strata of the lower-sheet-siltstone lithofacies, are truncated by a slump fault (Fig. 86). Soft-sediment deformation structures are abundant in the hanging wall and include faulted, contorted, and rotated blocks of siltstone, gray shale, and black, silty shale (Fig. 88).

The massive siltstone crops out next to these deformation structures (Fig. 86, outcrop 2). The siltstone is finer grained and muddier than the lower sheet siltstone, and bedding is difficult to distinguish. Among the most distinctive features in the massive siltstone are contorted, pebble- to boulder-size clumps of black, silty shale, gray shale, and siltstone which can be observed at outcrops 2 and 3 (Fig. 89). The basal contact of the siltstone and internal sedimentary structures are well exposed at outcrop 3, where nearly 30 ft (9 m) of the siltstone is exposed (Fig. 87). Here, the massive siltstone rests with a sharp, undulatory contact on black, silty shale, and sedimentary structures include convolute laminae, wavy laminae, current-ripple-drift cross-laminae, tangential crossbeds, and planar crossbeds (Fig. 90). Crossbeds are typically solitary, are thinner than one foot (0.3 m), and scarcely extend for more than three feet (1 m).

North of outcrop 3, the massive siltstone passes into a series of thick beds separated by poorly fissile, silty, gray shale (Fig. 86). These beds are best exposed at outcrop 6, the northernmost exposure of the massive-siltstone lithofacies, where they overlie black, silty shale and have been deformed into a series of folds. The siltstone beds are as thick as four feet (1.3 m) and contain flame structures and deformed crossbeds. Just north of this exposure at outcrop 7, sheet-siltstone beds of the lower tongue are intercalated with black shale. This is the only locality where black shale crops out in the lower-sheet-siltstone lithofacies (Pashin, 1985).

Turbidites of the lower-sheet-siltstone lithofacies are similar to those in the Borden Formation (Moore and Clarke, 1970; Kepferle, 1977, 1978; Chaplin, 1980), although soft-sediment deformation is much more abundant in the Berea. Borden turbidites have been interpreted to represent prodeltaic aprons; however, the Berea examples may have been derived from storms which swept the nearby shelf, because the lower tongue extends into the cliff stone (Pashin, 1985) (Fig. 75). The massive siltstone truncates the lower-sheet-siltstone lithofacies where it is the thickest and thus post-dates the silt-rich turbidites. Recognizing this relationship, Pashin and Ettensohn (1987) suggested that the channel fed mud-rich Bedford turbidites to the lower slope.

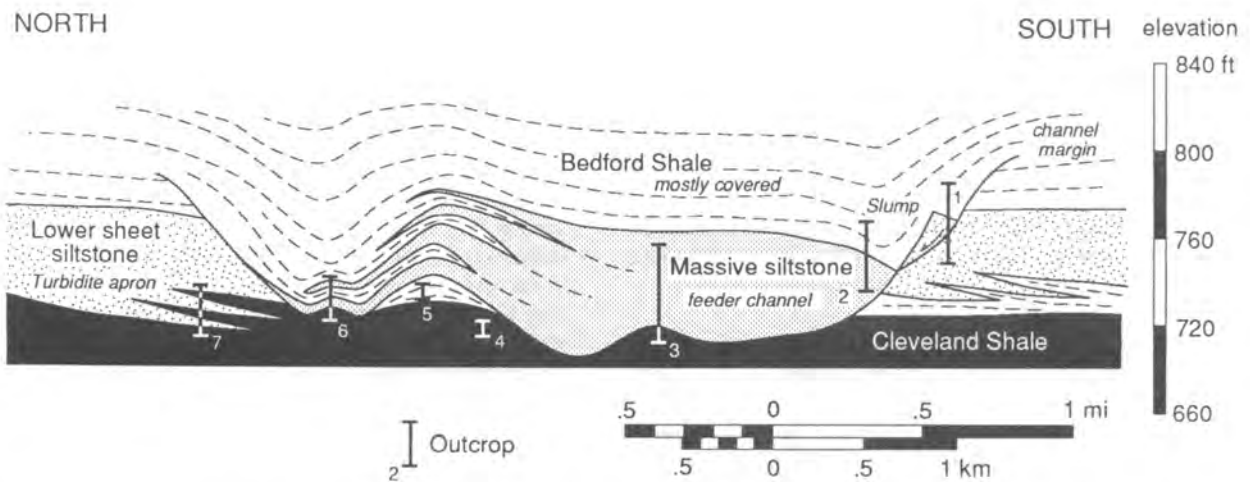


Figure 86. Schematic diagram of transect through massive siltstone along Kentucky State Route 59, Lewis County, Kentucky, lower tongue of the Berea Siltstone (Optional Stop C). The massive siltstone is interpreted to be a sinuous turbidite feeder channel that formed at the bottom of a submarine valley.



Figure 87. Outcrop of massive siltstone at Optional Stop C, outcrop 3 (Fig. 86). The siltstone is finer grained and muddier than adjacent turbiditic sheet-siltstone beds in the lower tongue. This is the thickest siltstone bed and the only turbidite feeder-channel fill exposed in the Bedford-Berea sequence of the Appalachian basin.



Figure 88. Blocks of siltstone along channel margin, Optional Stop C, outcrop 1, indicate that the submarine valley had steep, unstable walls that were subject to slumping.



Figure 89. Contorted, boulder-size clump of shale in massive siltstone, Optional Stop C, outcrop 2. Shale clumps are common in the massive siltstone and apparently represent mud balls that were exhumed by turbidity currents and transported along the channel floor.



Figure 90. Wavy laminae, current-ripple-drift cross-laminae, and pebble- to cobble-size shale clumps in massive siltstone, Optional Stop C, outcrop 3. Few internal sedimentary structures are visible in the massive siltstone because of weathering, but at outcrop 3, wavy laminae and cross-strata are readily apparent. Preservation of shale clumps within cross-stratified intervals suggests that the clumps were transported by turbidity currents that swept the submarine-valley floor.

Slumping above the massive siltstone at the channel margin suggests that the siltstone accumulated at the base of a submarine valley with steep, unstable walls (Fig. 86). Absence of the upper tongue of the Berea above the valley fill indicates that considerable topographic relief remained at the close of Bedford-Berea deposition. The valley apparently conducted strong currents, as indicated by crossbedding and boulder-size clumps of shale and siltstone (Figs. 89, 90); the shale clumps may be mud balls that were exhumed by turbidity currents and transported along the channel floor. Thinning and splitting of the siltstone across the valley is suggestive of a sinuous channel fill. To date, the only other account of an ancient sinuous turbidite feeder channel is by von der Borch and others (1985), who documented coarse-grained deposits from a Precambrian submarine canyon in Australia.

Indeed, turbidite feeder channels are much more characteristic of continental margins, which are influenced by the continental-oceanic crustal transition, than of epeiric shelf-to-basin transitions, which are ideally the result of depositional topography, not geophysical discontinuities. Why did a submarine valley form in northeastern Kentucky? Pashin and Ettensohn (1987) suggested that basement-fault reactivation provided geophysical discontinuities that gave rise to submarine-channel incision and facies patterns that mimic those along continental margins (Fig. 75). They indicated that using epeiric and continental-margin depositional models in tandem provides a powerful conceptual framework for characterizing epeiric shelf-to-basin transitions.

Turn right onto State Route 59 and continue north on Route 59 into Vanceburg, Kentucky.

- 54.3 1.2 Junction of Kentucky State Routes 59 and 10 at stop light. Continue straight on Route 59 into Vanceburg, Kentucky.
- 54.5 0.2 C&O Railroad crossing and Kentucky State Route 59.
- 54.7 0.2 Turn right onto Front Street.
- 54.8 0.1 LUNCH STOP. City Park on left-hand side (north) of Front Street. Bus will park here and field-trip participants will disembark for lunch at the city park overlooking the Ohio River. After lunch, passengers will board the bus and proceed south on Court Street away from the city park.

GENERAL SILURIAN PALEOGEOGRAPHIC AND TECTONIC FRAMEWORK FOR KENTUCKY

FRANK R. ETTENSOHN

During the course of the Silurian, Kentucky was situated at about 25° to 35° south latitude as Laurentia began to move southward (Scotese, 1990). This position still places Kentucky within the 10° to 45° latitudinal belt of subtropical storms and hurricanes (e.g., Kreisa, 1981; Marsaglia and Klein, 1983; Duke, 1985) that it had occupied since the Ordovician, and evidence of these storms is present in all the subsequent Silurian stops.

The Ordovician-Silurian transition throughout much of Kentucky is a very subtle unconformity or paraconformity called the Cherokee

Discontinuity by Dennison and Head (1975). This unconformity apparently represents a major eustatic sea-level drop (Dennison, 1976), and its coincidence with glacial deposits in Gondwana, Europe, and North America (Hambrey, 1985; Caputo and Crowell, 1985) strongly supports a glacioeustatic origin. Although this glaciation may have lasted between 20 and 35 million years from the Late Ordovician (Caradocian) to the Middle Silurian (Wenlockian) (Hambrey, 1985; Bjorlykke, 1985), the peak apparently occurred during a two-million-year interval near the Ordovician-Silurian transition when much of the unconformity was generated. The truncation and reworking of Ordovician redbeds from farther east in the Appalachian basin during and after this event may have been the source for the redbeds and remobilized iron deposited in Lower and Middle Silurian rocks within and near the Appalachian basin during subsequent transgressions (Ziegler and McKerrow, 1975). Clinton-age redbeds in the Crab Orchard Formation at Stop 6 on Day 3 may have had such an origin.

Moreover, most of transgressive-regressive cyclicity in the Lower and Middle Silurian rocks of Kentucky (e.g., Gordon and Ettensohn, 1984) and elsewhere in the Appalachian area apparently has a glacioeustatic origin (Johnson and others, 1985; Johnson, 1987).

In the Appalachian area, the Silurian is commonly regarded as a time of tectonic quiescence between the Taconian and Acadian orogenies, and during the Early and Middle Silurian this was probably the case. However, a growing body of evidence in the form of volcanism, plutonism, and deformation from the northern and central Appalachians (e.g., Laird, 1988; Bevier and Whalen, 1990) indicates that by early Middle Silurian or late Clinton time (latest Llandoveryan-early Wenlockian), a phase of Silurian tectonism called the Salinic Disturbance (Boucott, 1962; Rodgers, 1970) was active. This disturbance probably represents a weaker, southern phase of the Caledonian orogeny.

In the Silurian of New York and Pennsylvania, parts of two unconformity-bound flexural sequences are preserved suggesting two tectophases (Ettensohn, in press b) (Fig. 91). The first apparently began and ended during late Clinton time, and the stage of loading-type relaxation was attained before tectonism was renewed. In the Kentucky section, this tectophase was probably represented by the transgressive Dayton(Waco)-Estill interval in the Crab Orchard Formation. The Dayton or Waco Dolostone represents a basal transgressive carbonate (Fig. 91) and it is locally unconformable with underlying shales (Shaver, 1985). The main phase of subsidence, however, is represented by the Estill Shale (Fig. 91), and the subsidence was indeed great enough to give a likely tectonic signature on subsidence curves (Fig. 11).

Relaxation in this tectophase was apparently abruptly ended by the advent of a new tectophase in latest Clinton time (early Wenlockian) which extended through all relaxation phases to completion during Salina (Pridolian) time (Fig. 91). On the outcrop belt in eastern Kentucky, however, only the transgressive basal carbonate (Bisher Dolostone), representing initial parts of the tectophase, is preserved (Fig. 91). All overlying Silurian and Lower Devonian units on and near the Cincinnati arch were probably destroyed by erosion accompanying Early and Middle Devonian, Acadian bulge movement.

- 54.9 0.1 Junction of Court Street and Second Street. Note, court house on right hand side and Civil War monument in courthouse lawn. This is the only monument dedicated to Union soldiers south of the Mason-Dixon Line.

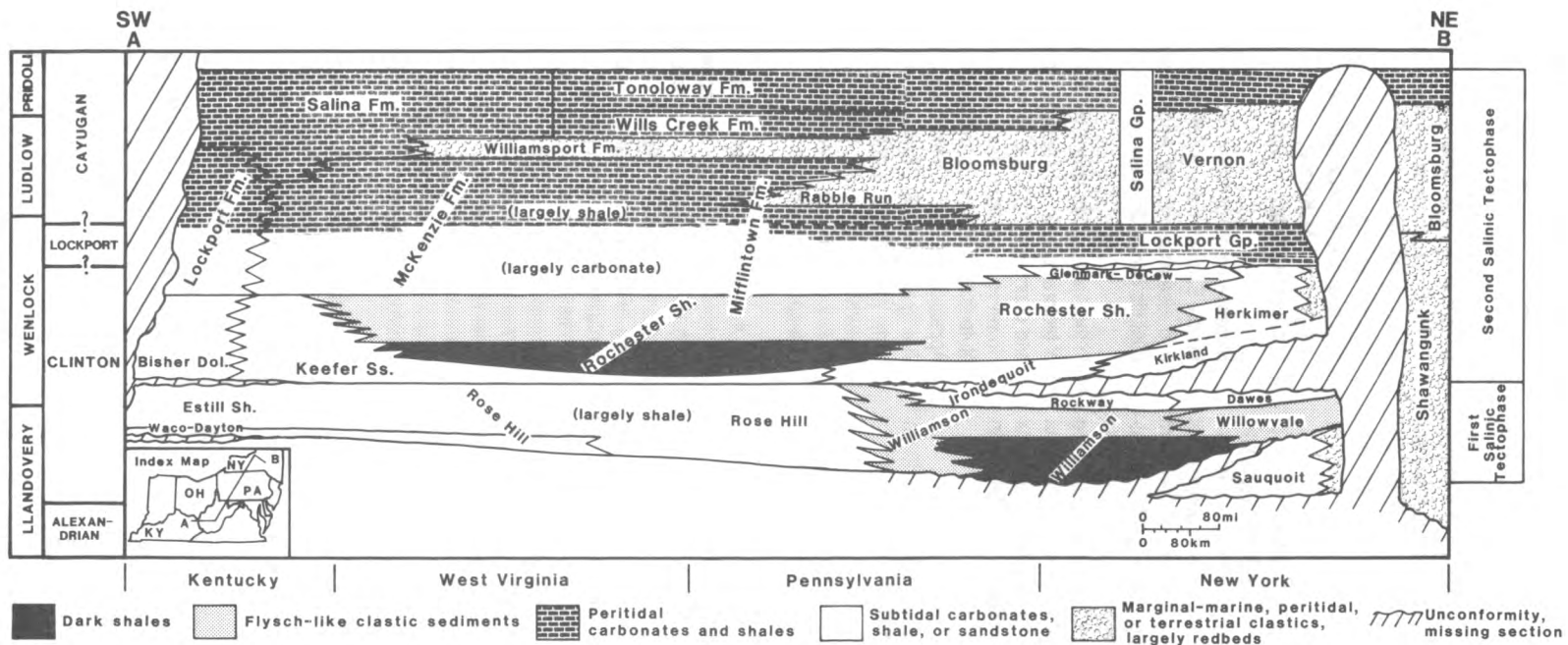


Figure 91. Schematic northeast-southwest cross section through northern and central parts of the Appalachian basin showing the repetition of two probable flexural sequences associated with the Salinic disturbance both in proximal and distal (northeastern Kentucky) parts of the basin. Only the second sequence is complete. The Estill Shale and Waco-Dayton Dolostone are parts of the Crab Orchard Formation in Kentucky (from Ettensohn in press b).

- 55.0 0.1 Junction of Court Street and Kentucky State Route 10. Turn right onto State Route 10.
- 55.1 0.1 Junction of Kentucky State Routes 10 and 59. Turn left onto State Route 59 and proceed south to State Route 546.
- 56.3 1.2 Junction of Kentucky State Routes 59 and 546. Turn right onto State Route 546. Note outcrops of Ohio Shale on south side of Route 546 at this intersection. Proceed westward on Route 546.
- 56.5 0.2 Huron Member of Ohio Shale exposed in roadcuts on left side of the highway.
- 57.4 0.9 Huron Member of Ohio Shale exposed in roadcuts on left side of the highway.
- 57.7 0.3 Huron Member of Ohio Shale exposed in roadcuts on both sides of the highway (roadcut extends for some 0.35 mi).
- 58.2 0.5 Huron Member of Ohio Shale exposed in roadcuts on both sides of the highway (roadcut extends for some 0.25 mi). Note large calcareous nodules along the ditch line in the basal Huron Member.
- 58.9 0.7 Bridge over Salt Lick Creek.
- 59.8 0.9 Junction of Kentucky State Routes 546 and 10 on right-hand side. Continue west on Route 546.
- 60.3 0.5 STOP 4. Biostromal Lithofacies of the Bisher Dolostone (Middle Silurian) Along Kentucky State Route 546, Northeastern Kentucky, Webster Cemetery Section (Along Kentucky State Route 546, westbound lane, 0.5 mi west of its junction with Kentucky State Route 10, Lewis Co., Carter Coordinates: 950'FSLx1750'FWL, 23-Z-74) Bisher Dolostone exposed in roadcuts on both sides of the highway. Bus will pull off to right-hand side of highway and field-trip participants will disembark.

**BIOSTROMAL BISHER DOLOSTONE (MIDDLE SILURIAN)
ALONG KENTUCKY STATE ROUTE 546, NORTHEASTERN
KENTUCKY, WEBSTER CEMETERY SECTION**

FRANK R. ETTENSOHN, R. THOMAS LIERMAN, JACK C. PASHIN, AND CHARLES E. MASON

The Bisher Dolostone in this area ranges from 20 to 80 ft (6.1 to 20.4 m) in thickness. At this location along the road only 21 ft (6.4 m) is exposed (Fig. 92). It consists of an argillaceous dolostone with only minor partings of shale. The dolostone is yellowish-gray to light olive-gray in color and weathers to a dusty-yellow to light-brown. It can best be described as an argillaceous dolomicrite to fossiliferous dolomicrite which possesses a hypidiotopic fabric of fine-to-medium-grained subhedral dolomite crystals. The dolomite crystals themselves are in part ferroan as determined by K-ferricyanide staining. Shales tend to be fissile, light greenish-gray in color and weather to a light olive gray.

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

The stratigraphic sequence exposed at this roadcut is quite interesting in that it can be subdivided into six or seven major fining-upward cycles, each of which is approximately three-feet (1-m) thick (Fig. 92). Each cycle begins with a sharp erosional base which is flat to slightly undulatory in nature, and each exhibits up to a dozen very irregular individual layers which can be traced some distance across the exposure. The irregularity of the layers is clearly related to large protruding stromatoporoids and tabulate corals as well as accompanying differential compaction and bioturbation. The stromatoporoids and corals in each layer are typically fragmented and overturned randomly; small-scale cross laminae are present locally in some layers. Overall, successive layers in each major cycle appear to become finer grained, with fossiliferous rudstones at the bases grading upward into fossiliferous floatstones and finally into mudstones. Fossils occur in greatest abundance at the base of each cycle and decrease in abundance upward through most cycles.

Despite the apparent abundance of large fragmented fossil clasts, each cycle apparently contained a substantial amount of mud or muddy matrix, much of which has now been recrystallized to coarser dolomite. Detrital silt and clay is also locally abundant in upper parts of each cycle. Many of the fossil clasts, moreover, appear to be matrix supported, but these fossils seem to belong to a rather low-diversity, restricted community. Despite the variety of fossil fragments described earlier, many of which may be allochthonous, three colonial filter feeders, a stromatoporoid, a tabulate coral, and a trepostome bryozoan, predominate in the community, and of these, stromatoporoids were far more abundant. The relative abundance of muds and muddy matrix, the presence of detrital silt and mud, and the restricted nature of the fauna suggest that this Bisher lithofacies probably represents a nearshore lagoonal setting which was repeatedly disrupted by large tropical storms. Although each major cycle could represent a single major storm event, it seems much more likely that each of the irregular layers within a cycle represents an individual storm because of the relatively equal degree of fossil fragmentation and overturning in most layers. The apparent upward deepening in each cycle, however, may reflect: 1.) eustatic sea-level changes related to ongoing glaciation, 2.) tectonic subsidence related to flexural movements, or 3.) reactivation of basement structures in the area (see Stops 1 and 4, Day 3). For the given time and place, any of the explanations are equally likely, and hence it is difficult to suggest a preference.

Units 1 and 4 (Fig. 92) are sufficiently different from the others that they warrant special mention. Only the top of Unit 1 is preserved, and although it could represent the top of a lower cycle, it is unique in that it is largely composed of dark, muddy sediment with discontinuous laminae and what may be bryozoan thickets. No other unit exhibits such bryozoan-rich muds, and this unit could reflect more typical pre-storm conditions. The advent of stromatoporoids in overlying units may simply reflect the prevalence of higher energy, more, storm-prone conditions during subsequent periods of time. Unit 5 is also unusual

Day 3, Stop 4, Webster Cemetery Section

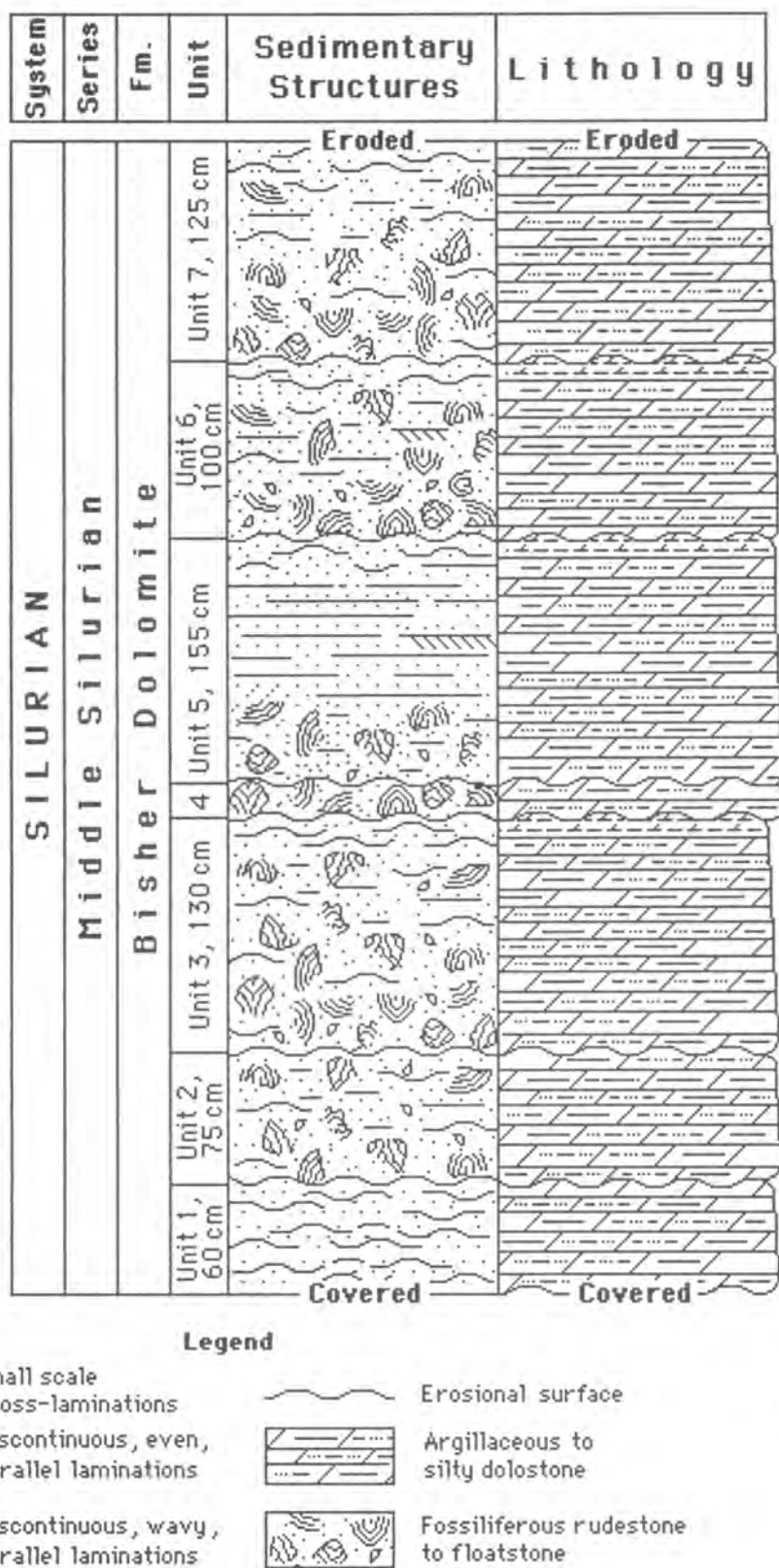


Figure 92. Schematic drawing of the Middle Silurian Bisher section and its sedimentary structures at Webster Cemetery on Kentucky State Route 546 (Stop 4, Day 3).

in that it is relatively thin and exhibits several apparently in-place stromatoporoids. Some of these seem to exhibit in-place brecciation, which may suggest brief periods of exposure, which lends support to interpretations of a near-shore lagoonal origin for the lithofacies.

It is difficult to relate this biostromal lithofacies of the Bisher to the other two lithofacies which we will examine at the next two stops. Each lithofacies is separated in space from the others with little evidence of direct connection. The upper units in the biostromal lithofacies, however, do show some evidence of laminated to crossbedded sands, and rare stromatopore fragments have been found in a more sandy lithofacies, all of which suggest some relationship between the biostromal lithofacies and the yet to be seen sandbelt lithofacies of the Bisher.

- 60.5 0.2 Bisher Dolostone in roadcuts on right side of the highway.
- 60.8 0.3 Bisher Dolostone exposed in roadcuts on right side of the highway (roadcut extends for some 0.2 mi).
- 61.6 0.8 Bisher Dolostone exposed in roadcuts on right side of the highway (roadcut extends for some 1.3 mi).
- 62.1 0.5 Bridge over Salt Lick Creek.
- 62.6 0.5 Bridge over Salt Lick Creek.
- 63.4 0.8 STOP 5. Tempestite Lithofacies of the Bisher Dolostone and the Crab Orchard Shale (Middle Silurian) Along Kentucky State Route 546, Northeastern Kentucky, Charters Section (Roadcut along the north side of Kentucky State Route 546, 0.4 mi east of Charters, Kentucky and the junction of Kentucky State Highways 989 and 546, Lewis Co.; Carter Coordinates: 1900' FNLx1400' FWL, 1-Y-73). Crab Orchard Shale and Bisher Dolostone exposed in roadcut on right side of the highway. Bus will pull off to right-hand side of highway and passengers will disembark.

TEMPESTITE LITHOFACIES OF THE BISHER DOLOSTONE AND THE CRAB ORCHARD SHALE (MIDDLE SILURIAN) ALONG KENTUCKY STATE HIGHWAY 546, NORTHEASTERN KENTUCKY, CHARTERS SECTION

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BISHER DOLOSTONE

The Bisher Dolostone in the Charters area ranges in thickness from 20 to 80 ft (6.1-24.4 m). At this location only the lower part of the Bisher is exposed with the exposed interval ranging from 24 ft (7.3 m) on the west side of the cut to 25.8 ft (7.9 m) on the east side (Fig. 93). This variation in thickness is due, in part, to a small growth fault found at this section which will be discussed at the end of this description. The Bisher consists of medium to thick-bedded dolostones with shale interbeds. The dolostones are argillaceous dolomicrites or fossiliferous dolomicrites to dolarenites, which are thin to medium-bedded and dark-orange to brown in color. These beds commonly have sharp erosional bases with tool marks and tops which grade upward into overlying shales. Internally they show parallel

laminae and local small-scale, trough cross-laminae. Less commonly found are rip-up clasts, fossil lags, ripple marks, gutter casts, and hummocky crossbedding. The rip-up clasts and fossil debris (especially crinoid) occur as lags or as stringers in the more massive or thicker beds, which can be observed best at the east end of the cut in the basal 6.5 ft (2.0 m) of the Bisher section. The shales are muddy, poorly fissile, medium- to light-gray in color and weather to a dark-gray.

Megafossils are common throughout the Bisher exposed here and include: bryozoans, brachiopods, cephalopods, conularids, gastropods, pelecypods, disarticulated crinoid debris, and parts of trilobites. Complete trilobites and articulated crinoids have been collected from this section, especially from the uppermost bed along the topmost bench on the west side of the cut. The megafossils in the Bisher are preserved as molds and casts. Void spaces in the rock, especially those resulting from dissolved fossils, are commonly filled with blebs of petroleum or asphalt. Microfossils, other than ostracods, have not been observed at this section. Trace fossils are moderately diverse and extremely abundant in the Bisher at this locality and include both vertical and horizontal traces. They occur in the shales as well as the dolostone units. In the dolomitic units, both trace fossils and body fossils are commonly destroyed or partially obscured by the dolomitization process. As a result both trace fossils and body fossils are best preserved on the tops or soles of beds. This same process also affects the preservation of internal sedimentary structures. The ichnogenera most commonly encountered at this locality include: *Bifungites*, *Chondrites*, *Cruziana*, *Diplocraterion*, *Fascifodina*, and *Rusophycus*. The preceding trace-fossil assemblage belongs in the *Cruziana* ichnofacies.

The shales and dolostones at this locality represent yet another Bisher lithofacies, one which we call the tempestite lithofacies. Unlike the biostromal lithofacies at Stop 4, the body-fossil and trace-fossil assemblages reflect open-marine, subtidal deposition, and the presence of so much shale suggests conditions largely below normal wave base. This is also indicated by the predominance of horizontal traces like *Chondrites*, the intensity of bioturbation, and the general absence of nonbiogenic sedimentary structures (Howard, 1978). Although dolomitization may have destroyed many of the nonbiogenic structures, the sharp erosional bases with tool marks, the mixed partly transported faunas at the bases of individual beds, and the presence of rare gutter casts and hummocky cross-bedding suggests proximal tempestites (e.g., Aigner, 1985). These layers apparently accumulated as storm scours washed out bottom faunas and associated sediments in waning-flow conditions. In some instances, shell accumulations formed thick, firm substrates on which subsequent faunas could colonize. More commonly, however, waning-flow conditions seemed to have merged with the normal hemipelagic-shale deposition so that most dolostone layers grade upward into shale, which along with underlying dolostones may be reworked by post-event bioturbation. Many of these trace fossils were subsequently cast on the bottoms of overlying tempestite layers.

At both the southwestern and northeastern margins of the exposure, the shale-and-dolostone lithofacies grades laterally into more massive dolostone bodies composed of crossbedded dolarenites. Although the extent and nature of these bodies are uncertain because of erosion, they are typical of a more sandy Bisher lithofacies which we will examine at Stop 6. They may represent isolated shoals or parts of a sandbelt complex generally at or above wave base from which sands were washed during storms into deeper, adjacent basins as tempestites like those described here.

Day 3, Stop 5, Charters Section

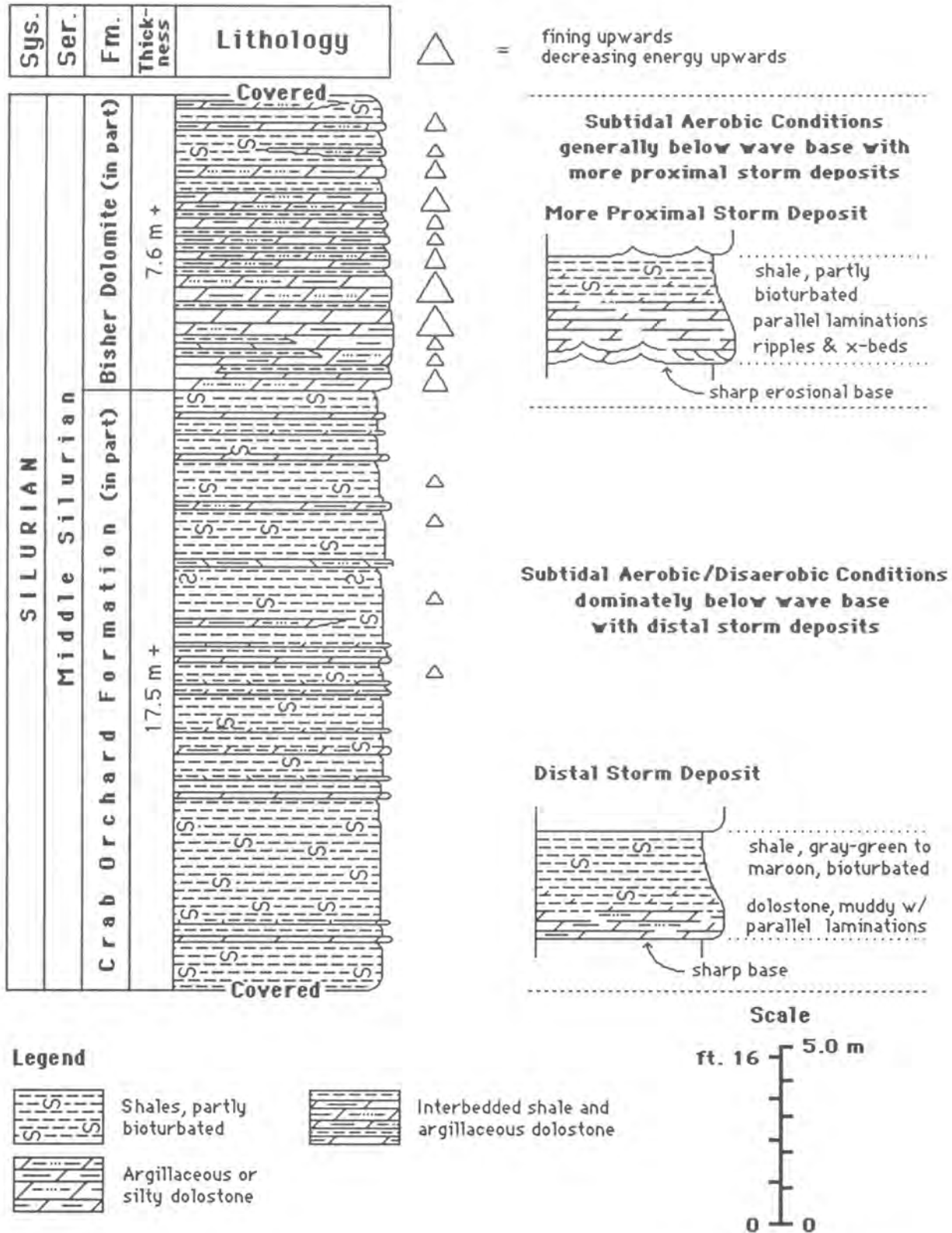


Figure 93. Schematic section showing parts of the Crab Orchard Formation and Bisher Dolostone (Middle Silurian), as well as sedimentary structures and environmental interpretations, at Charters on Kentucky State Route 546 (Stop 5, Day 3).

CRAB ORCHARD SHALE

The Crab Orchard Shale in this area varies from 155 to 192 ft (47.2 to 59.4 m) in thickness. At this location only the upper 60 ft (18.5 m) is exposed at this roadcut. It consists predominantly of mud shales with thin discontinuous beds of dolostone. Shales are poorly fissile, medium- to light-gray, and tend to weather to a medium- or dark-gray color. A number of distinctive layers in the lower part of the section show a maroon coloration presumably due to the presence of finely disseminated hematite in the shale. Exact determinations of the composition have not yet been made at this locality. However, collections made from a similar section in Bath County (east-central Kentucky), 33 mi to the southwest, have yielded the following clay-mineral composition: kaolinite—6%, mixed-layered illite/smectite—89%, and Fe-rich chlorite—5%. The mixed-layered illite/smectite consists of 65% illite and 24% smectite (Lierman and Manley, 1991). Also, shales at another locality average between 38 and 68% silt, clearly making them mud shales rather than clay shales or silt shales. The shales are quite heavily bioturbated, but individual traces are not recognizable. The dolostones are dolomicrites, which are finely crystalline and slightly calcareous with a medium-gray fresh color that weathers to dark orange or brown. These beds of dolostone have sharp erosional bases and tops which grade upward into overlying shales. Internally they show fine parallel laminae, and in a few cases wavy or hummocky crossbedding. The upper part of the Crab Orchard at the Knob Lick Section in Bath County mentioned earlier does not contain maroon shale beds, and the contained dolostone beds are thinner and much less common than those found at the Charters Section.

Megafossils are not apparent from surface collections, but washed residues of the shales have yielded bryozoans, brachiopods, gastropods, pelecypods, disarticulated crinoid debris, and trilobite fragments. Initially, six five-kilogram spot samples (collected at 30-cm intervals) were processed and examined for fossil content to determine if the Crab Orchard here was deposited under dysaerobic conditions like the upper Crab Orchard examined at the Knob Lick Section in Bath County, Kentucky. As pointed out above, the Crab Orchard here contains a good open-marine megafossil assemblage. The mode of preservation of the macrofauna here is either replacement by pyrite or recrystallization by calcite (initially). The megafossils were found throughout the exposed Crab Orchard interval, and brachiopods and crinoids predominate. All the megafossils are typically juveniles. These preliminary findings are similar to those reported by Mason and others (1991) for this same interval in studies from the Knob Lick Section. These fossils are similar in abundance and diversity to those megafossils elsewhere in the unit, but differ in that some megafossils here are preserved by recrystallization rather than by pyrite replacement as in Bath County. Two additional samples, one from the maroon shales and one from the green shales, were processed and examined for any significant difference in the modes of preservation recovered from them. Initial results showed no significant difference; however, it is believed that the sampled interval may have been too thick to give good results. It is also believed that the recrystallized fossils were preserved under aerobic conditions or under dysaerobic conditions just below the aerobic/dysaerobic boundary, whereas the pyrite-replaced forms were deposited under normal dysaerobic conditions. Also mixing by storm activity may have altered or disrupted normal conditions of oxygen content in the water column as well as below the sediment-water interface. A note of special interest, is that in collecting the green-shale sample at 46 to 47 ft (14.0-14.3 m) above the base of this section, six graptolite specimens were collected. They were complete

and preserved as carbonaceous films. This macrofauna, as well as that from Knob Lick, is similar to other late Paleozoic dysaerobic faunas (Mason and Kammer, 1984; Kammer and others, 1986) (see Optional Stop B, Day 3) in being predominantly composed of juveniles replaced by pyrite. However, unlike typical late Paleozoic dysaerobic faunas, the faunas in this section and that at Knob Lick are dominated by brachiopods and crinoids (sessile forms) rather than mollusks (vagile forms). The trophic structure is likewise different in that filter feeders predominate rather than deposit feeders and scavenging cephalopods. This report and that from Knob Lick constitute the first reports of pre-Devonian dysaerobic faunas. As noted earlier, they are similar in all respects except for dominance and trophic structure.

Microfossils were also abundant and fairly diverse in the washed residues from this section, as well as from Knob Lick. Microfossils found included foraminifera, ostracods, conodonts, scolecodonts, chitinozoans, and graptolite fragments. With the exception of some ostracods, no microfossils were replaced by pyrite. Foraminifera were the most abundant and diverse microfauna, followed by ostracods. The shales are heavily bioturbated, whereas nondescript vertical and horizontal burrows are present in some of the dolostone layers. No specific ichnogenera have been identified yet from this section of the Crab Orchard.

The shales of the Crab Orchard are thought to represent normal-marine, offshore deposition on a shelf generally below storm wave base. The interspersed dolostone layers are believed to be distal storm deposits or tempestites (Fig. 93). These developed from nearshore carbonate sediments and terrigenous muds which were resuspended during storms and later redeposited offshore as the storms waned. The Crab Orchard observed in Bath County is believed to have been deposited under similar conditions but in deeper waters farther offshore in predominantly dysaerobic environments. The oxygen levels at this locality apparently varied between dysaerobic and aerobic, which is evident through the presence of maroon shale beds and the presence of recrystallized megafossil remains rather than pyrite replacement. At both this locality and the one in Bath County, it is believed that at least the upper part, if not all, of the water column was aerobic due to the presence of a normal-marine pelagic microfauna. The thickness of the Crab Orchard, as well as the presence of dysaerobic environments, may reflect rapid rates of tectonic subsidence related to flexural movement accompanying the Salinic Disturbance (Fig. 11).

As pointed out earlier, a small, normal growth-fault occurs in this section. This subtle fault has been overlooked by previous workers for many years. It offsets beds throughout the Crab Orchard interval but terminates in the lower part of the Bisher section. It has a vertical displacement of 2.3 ft (0.7 m). Evidence supporting the presence of this fault includes: 1.) slickensides along the fault plane, 2.) reversed dip of beds on the downthrown block (east side), 3.) offset of beds across the fault plane, and 4.) the development of a large gully along the fault plane. This gully is probably the most obvious aspect of the fault, as there are no other gulleys approaching this size along the roadcut.

It is possible that this and other undetected growth faults in the area may have influenced the distribution of the various Bisher lithofacies.

63.7 0.3 Community of Charters on right-hand side of highway.

63.9 0.2 Junction of Kentucky State Routes 546 and 989 on left-hand side. Continue west on Route 546.

- 64.5 0.6 Crab Orchard Shale and Bisher Dolostone exposed in roadcut on right side of the highway.
- 67.7 3.2 Crab Orchard Shale, Bisher Dolostone, upper Olenangy Shale, and Huron Member of the Ohio Shale exposed in roadcuts on both sides of the highway; Stop 6, Herron Hill Section.
- 67.9 0.2 STOP 6. Sandbelt Lithofacies of the Bisher Dolostone, the Crab Orchard Shale, the Upper Olenangy Shale, and the Huron Member of the Ohio Shale along Kentucky State Route 546 in Northeastern Kentucky (Exposures on both sides of Kentucky State Route 546 at its junction with Poplar Flat Road, Lewis Co.; Carter Coordinates: 2400'FSLx900'FEL, 25-Z-73). Turn right onto Poplar Flat Road. Bus will pull off on small dirt road to left-hand side of road and field-trip participants will disembark.

SANDBELT LITHOFACIES OF THE BISHER DOLOSTONE, THE CRAB ORCHARD SHALE, THE UPPER OLENANGY SHALE, AND THE HURON MEMBER OF THE OHIO SHALE IN NORTHEASTERN KENTUCKY

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FRANK R. ETENSOHN, AND JACK C. PASHIN

At this exposure, the field-trip route leaves the "Knobs" portion of the Pottsville Escarpment of the Kanawha Section of the Appalachian Plateau for the last time and enters the more gently rolling Outer Bluegrass proper (Figs. 2, 3) underlain by Silurian and Upper Ordovician shales, limestones, and dolostones. The Middle Silurian Crab Orchard Shale and Bisher Dolostone and the Lower Silurian Brassfield Formation persist some seven to eight miles farther to the west in the Outer Bluegrass, whereas the upper Olenangy and Ohio shales are effectively lost to the west due to updip erosion on the flank of the Cincinnati arch (Morris, 1965; Peck, 1967).

CRAB ORCHARD SHALE (MIDDLE SILURIAN)

The upper part of the Crab Orchard Shale (Fig. 94) in this area varies from 130 to 160 ft (39.6-48.8 m) in thickness. At this location only the upper 29.5 ft (9 m) is exposed along the east end of this roadcut; it consists of predominantly mud shales interlayered with thin beds of dolostone. The shales exposed here are poorly fissile, light greenish-gray to light-gray and weather to a light olive-gray or medium-gray color. They are also heavily bioturbated, though individual traces are not recognizable. The dolostone beds appear to be dolomicrites, which are finely crystalline with a medium-gray color when fresh but weather to dark orange or brown color. They are also heavily bioturbated, though individual traces are not recognizable. These beds of dolostone have sharp erosional bases, whereas the tops grade upward into overlying shales. Internally they show fine parallel laminae, and in a few cases wavy or hummocky crossbedding. The upper contact with the overlying Bisher Dolostone is quite sharp and its base is largely covered.

The shales of the Crab Orchard represent the accumulation of fine-grained terrigenous sediments along a slowly subsiding shelf. As discussed previously, the shales apparently represent aerobic to dysaerobic, deeper open-marine accumulation of muds well below

normal wave base. The thickness of the unit and presence of locally dysaerobic intervals suggest relatively rapid subsidence rates probably related to Salinic tectonism (Fig. 11). The thin dolostones in the unit, however, are probably distal tempestites (Fig. 93) reflecting the resuspension of carbonate sediments and terrigenous muds with the incursion of storm-surge and backflow currents.

A special note on the Crab Orchard Shale is that this formation has a strong tendency to slide or slump causing pavement failures especially where highway construction has oversteepened natural slopes. This is most evident along segments of abandoned Kentucky State Route 10 found on the left (south) side of State Routes 546 and 10 as one proceeds up Herron Hill to Stop 6.

BISHER DOLOSTONE (MIDDLE SILURIAN)

The Bisher Dolostone in this area varies from 20 to 80 ft (6.1-20.4 m) in thickness, and at this stop the entire unit has a thickness of approximately 49 ft (15 m). The basal contact with the Crab Orchard is sharp and slightly undulatory, and basal parts of the Bisher contain reworked clasts of the Crab Orchard. Hence, an erosional hiatus or subtle disconformity may be present at the base of the Bisher in this area (Fig. 91).

The unit consists of fine- to coarse-grained dolostone with greenish-gray shale partings. The dolostones are locally silty to sandy and are medium-gray to greenish-gray when fresh; weathering imparts a dark-orange to brown color and a "punky" texture. Lithologically, the dolostones are best described as argillaceous dolomicrites to bioclastic dolomicrites and dolarenites, but these lithologies may not reflect the original textures. Where dolomitization has not been as effective, coarse-grained bioclastic sands seem to predominate. Most of the dolostones possess a hypidiotopic fabric with fine- to medium-grained, subhedral ferroan dolomite crystals. Internally, stratification within the dolostone beds includes parallel laminae as well as local trough cross-laminae and hummocky cross-laminae. Less commonly found are scours, rip-up clasts, fossil lags, sparse ripple marks, and crude graded bedding.

An open-marine megafauna is common throughout and includes minor stromatoporoids, corals (both tabulate and rugose), bryozoans, brachiopods, gastropods, nautiloids, and disarticulated crinoid and trilobite debris. The megafossils in the Bisher are preserved as molds and casts which are predominately broken and abraded fragments. Void spaces in the rock, especially those resulting from dissolved fossils, are commonly filled with blebs of petroleum or asphalt. The only microfossils observed at this cut are ostracods. Trace fossils are also poorly developed in this section with only a few vertical and horizontal traces observed to date.

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

Day 3, Stop 6, Herron Hill Section

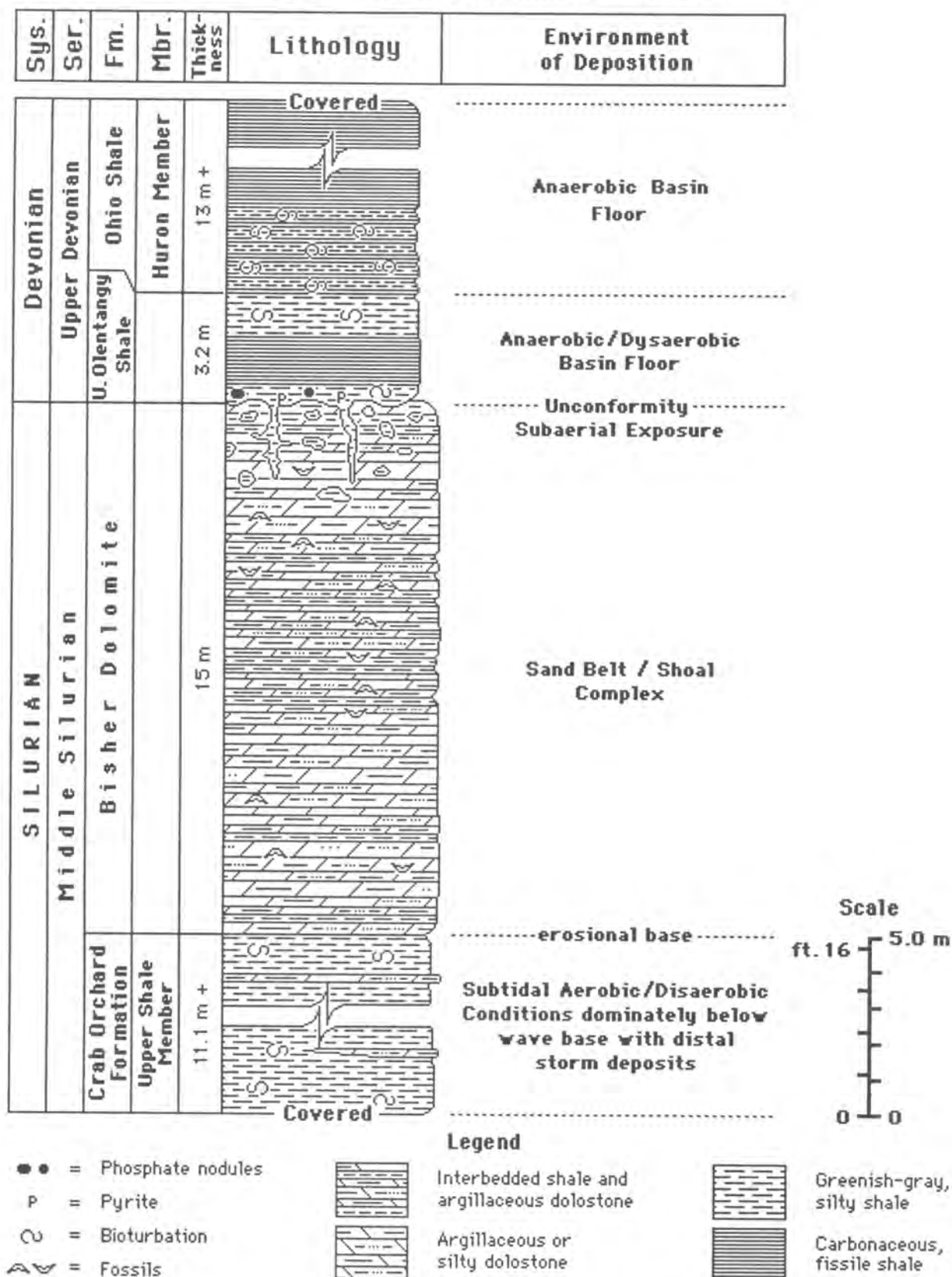


Figure 94. Schematic section showing the Middle Silurian and Upper Devonian rocks at Herron Hill on Kentucky State Route 546 (Stop 6, Day 3).

inter-shoal depression on the sandbelt or a brief period of transgression and deepening.

The irregular upper surface of the Bisher (Fig. 94) is the same compound unconformity observed at Stop 4 on Day 2 below the upper Olentangy Shale; it apparently reflects periods of Early, Middle, and Late Devonian erosion related to bulge migration and uplift on the flanks of the Cincinnati arch. This contact represents about a 60-million-year gap in the stratigraphic record. Close examination of this surface reveals a number of features which suggest subaerial exposure and possible solution or collapse along the upper surface. Evidence for this can be found in the presence of the following features: 1.) large, sediment-filled solution pipes or holes, which are commonly infilled with breccia fragments and/or a greenish-gray shale like overlying upper Olentangy shales; 2.) irregular vugs or voids in the upper 3.3 ft (1 m) of the unit; 3.) *in situ* breccia zones in the upper 3.3 ft (1 m) of this unit; and 4.) an upper irregular surface with as much as 1.6 ft (0.5 m) of relief. Of course, some of these features could also be related to interstratal karst. However, many of these features are now very difficult to recognize, because they have been destroyed or partially obscured by the process of dolomitization which occurred after these features formed. Some minor mineralization can also be seen in the upper meter or so of the Bisher and this seems to be associated with the exposure surface. This mineralization can be recognized as the infilling of many of the vugs or void spaces (especially in fossils) with a variety of secondary minerals including barite, calcite, dolomite, pyrite, and sphalerite.

UPPER OLENTANGY SHALE (UPPER DEVONIAN)

The upper Olentangy Shale in this area varies from 0 to 30 ft (0-9.2 m) in thickness. At this location about 11 ft (3.3 m) of the upper Olentangy is exposed along the base of the first bench. The greater thickness here compared to that at Stop 4 on Day 2 probably reflects the greater distance from the Cincinnati arch in this area. The unit largely contains nonfissile mud shale with a grayish-green to light-gray unweathered color, but weathering to a light olive-gray. The upper Olentangy also contains some minor interbeds of black fissile shale, and a five-foot (1.5-m) interval of black shale is present in the middle of the unit but is usually covered with talus. It is possible that this black shale could represent the thinning western edge of the Pipe Creek Shale, a black shale within the upper Olentangy that enters northeastern Kentucky in the subsurface (Keplerle and others, 1978). The gray shales in the upper Olentangy contain scattered framboids and crystals of pyrite and marcasite, as well as concentrated deposits that form two thin mineralized beds low in the unit. The first and lowermost bed occurs at the base of the unit and the uppermost bed occurs approximately two feet (6 m) above the base. Small sparsely fossiliferous phosphate nodules are also found throughout this unit.

Megafossils are rare and presently restricted to the upper Olentangy exposed here. Eight ammonoid specimens and one trilobite have been recovered from this section to date. This is only the second occurrence (see Barron and Ettensohn, 1981) and the first surface report of Devonian ammonoids from Kentucky. These fossils are poorly preserved as phosphate-replaced internal molds in small phosphate nodules. Attempts at examining this formation for microfossils were not completed in time to be included in this report. The unit is bioturbated, and both vertical and horizontal burrows are present. These burrows are replaced by pyrite, marcasite, and phosphate. Thus far no specific ichnogenera have been identified.

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

OHIO SHALE (UPPER DEVONIAN)

Huron Member

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

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

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

SOME THOUGHTS ON THE NATURE OF BISHER DEPOSITION

FRANK R. ETENSOHN AND JACK C. PASHIN

Across a distance of approximately eight miles, we have observed three different Bisher lithofacies in three exposures. Except for some evidence of coarser sandy-bioclastic lithologies and storm deposits in each of the exposures, there is little evidence to tie them together. Part of this lack of evidence is related to the nature of erosion and highway construction which have obscured or destroyed parts of the section, whereas other parts of this evidence shortage are related to the equally obscuring effects of dolomitization. Quite clearly a regional study of the Bisher is needed.

Nonetheless, based on the three exposures at hand and their predominant lithologies a few observations and possible interpretations can be made. First, there appears to be no coherent stratigraphic or environmental sequence in the three Bisher exposures, except for the obvious and rather abrupt change from deeper open-marine environments in the underlying Crab Orchard Shale to shallow open-marine, shelf environments in the Bisher. Second, there appears to be no particular geographic disposition or shoreward-to-seaward orientation to the occurrence of the three lithofacies. For example, the deeper-water tempestite lithofacies (Charters Section, Stop 5) seems to occur between the more shallow-water sandbelt (Herron Hill Section, Stop 6) and lagoonal (Webster

Cemetery Section, Stop 4) lithofacies. Although the nature of the intervening exposures is poorly known, many of these exposures appear to be intensely weathered and dolomitized examples of the sandbelt lithofacies, suggesting that this lithofacies may be the integrating agent for any initial understanding of the Bisher.

As was suggested in the earlier Silurian framework statement, the abrupt change from the deeper water Crab Orchard Shale to the relatively shallow-water Bisher Dolostone probably reflects the accompanying second tectophase of the Salinic Disturbance (Fig. 91). The Bisher then would represent the typical, shallow-water, transgressive, basal carbonate in flexural sequences. Uplift accompanying bulge migration would have generated a shallow-water shelf situation with relatively uniform depths. Of course the presence of tempestites in all lithofacies suggests that this shelf was in a major storm belt. What we suggest happened on this shelf, based on the nature of the three exposures examined, was the development of a rather complex facies mosaic. At or near normal wave base, a coarse-grained shoal or sandbelt lithofacies like that at Herron Hill (Stop 6) developed. In protected areas behind or within the sandbelt complex itself, localized lagoonal deposition like that seen at Webster Cemetery (Stop 4) could have originated. In distal parts of the sandbelt or in deeper water embayments within the sandbelt, which were below normal wave base, shales and distal tempestites like those at Charters (Stop 5) accumulated. Whitaker (1987) has interpreted most of the Bisher lithologies at these stops, particularly those at Herron Hill and Charters to represent tidal-flat deposition. We believe that the presence of a good open-marine megafauna and ichnofauna at Charters and Herron Hill, the predominance of relatively coarse-grained sediments and storm-related sedimentary structures, as well as the absence of laminated sediments and exposure-related sedimentary structures, strongly mitigate against a tidal-flat interpretation at these three exposures. However, we do not rule out the possibility that a tidal-flat lithofacies may occur elsewhere in the Bisher facies mosaic.

Why then did an environmental mosaic form and not a more typical transgressive continuum (e.g., Shaw, 1964; Irwin, 1965)? All we can suggest at this point is the possibility that erosional relief on top of the Crab Orchard or growth faulting acted to localize sandbelt and basinal environments at certain points, and that these environments acted to nucleate the same or different environments nearby. Evidence for growth faulting was mentioned today at Stops 1 and 5, and it is tempting to suggest that the local tempestite basin at Charters (Stop 5) may have been constrained by subsidence related to the growth fault described there. Flexural reactivation of these faults could have been an ongoing process, but we also must not overlook the potential influence of eustatic variations on the mosaic development, for the major cycles at the Webster Cemetery section (Stop 4) could reflect a degree of eustatic control.

Bus will turn around and proceed back to Kentucky State Route 546. Turn right and continue west on Route 546.

69.7 1.8 Junction of Kentucky State Routes 546 and 10 on left-hand side. Continue west on Route 546. Turn right onto Kentucky State Route 10 and proceed west.

70.9 1.2 Bridge over Cabin Creek.

71.0 0.1 STOP 7. Upper Ordovician-Lower Silurian Rocks at Cabin Creek, Northeastern Kentucky (Roadcut along Kentucky State Route 10, 0.6 mi southwest of its junction with State Route 546 at Ribolt, Lewis Co.; Carter Coordinates: 200°FNLx1900°FEL, 3-Y-72). Preachersville Member of the Drakes Formation (Upper Ordovician) and the Brassfield Dolostone (Middle Silurian) are exposed in roadcuts on both sides of the highway. Bus will pull off onto small dirt road on the right-hand side of State Route 10 and passengers will disembark. Walk uphill towards the roadcuts on either side of State Route 10.

UPPER ORDOVICIAN-LOWER SILURIAN ROCKS AT CABIN CREEK, NORTHEASTERN KENTUCKY

FRANK R. ETENSOHN

The Cabin Creek section begins with about 20 ft (6.1 m) of the Upper Ordovician Preachersville Member of the Drakes Formation, which contains a few beds of maroon shale and limestone (Fig. 95). These maroon beds probably represent one of the most westerly tongues of the Queenston redbed lithofacies.

In this area, the Drakes formation is no more than 35-ft (10.7-m) thick (Peck, 1967), approximately one third of its maximum thickness to the south (Weir and others, 1984). This thinning, however, is in large part related to a concomitant thickening of the underlying Bull Fork Formation toward the northeast (Weir and others, 1984). In the same direction, there is also a change in depositional environments from predominantly intertidal and supratidal in the south and southwest, to very shallow open-marine environments in this area in the northeast. Both situations, the northeasterly thickening of the Bull Fork at the expense of the Drakes and the northeasterly transition in the Drakes to more open-marine environments, suggest the influence of the newly emerging Cincinnati arch, because these trends seem to reflect increasing distance and slightly deeper waters away from the axis of the arch. As it happens, the outcrop belt in this area also diverges northeastwardly away from the axis of the arch, revealing these changes.

Overlying parts of the Silurian section begin with about 8 ft (2.4 m) of argillaceous dolostones and interbedded shale which some workers (e.g., Gordon and Ettensohn, 1984) have referred to the Belfast Member of the Brassfield Formation (Foerste, 1896). The lowermost two feet (0.6 m) of this unit consist of a massive yellowish-brown dolostone which appears to be made up of smaller amalgamated beds. The base of the unit is irregular and apparently erosional and exhibits reworked fossils, phosphate nodules, and glauconite. Bioturbation, however, has destroyed almost all other sedimentary features. Potter and others (1991, p. 54) suggested that this two-foot unit might be the Belfast, but in fact the Belfast would also have to include the overlying six feet (1.8 m) of argillaceous dolostones by Foerste's (1896) definition.

The overlying six feet (1.8 m) consist of thinner bedded argillaceous dolostones and interbedded shales. Although much of the internal fabric of these dolostones has been destroyed by dolomitization and bioturbation, some still exhibit sharp, planar bases, scours, crude graded bedding, and hummocky crossbedding, all of which suggest possible storm origin.

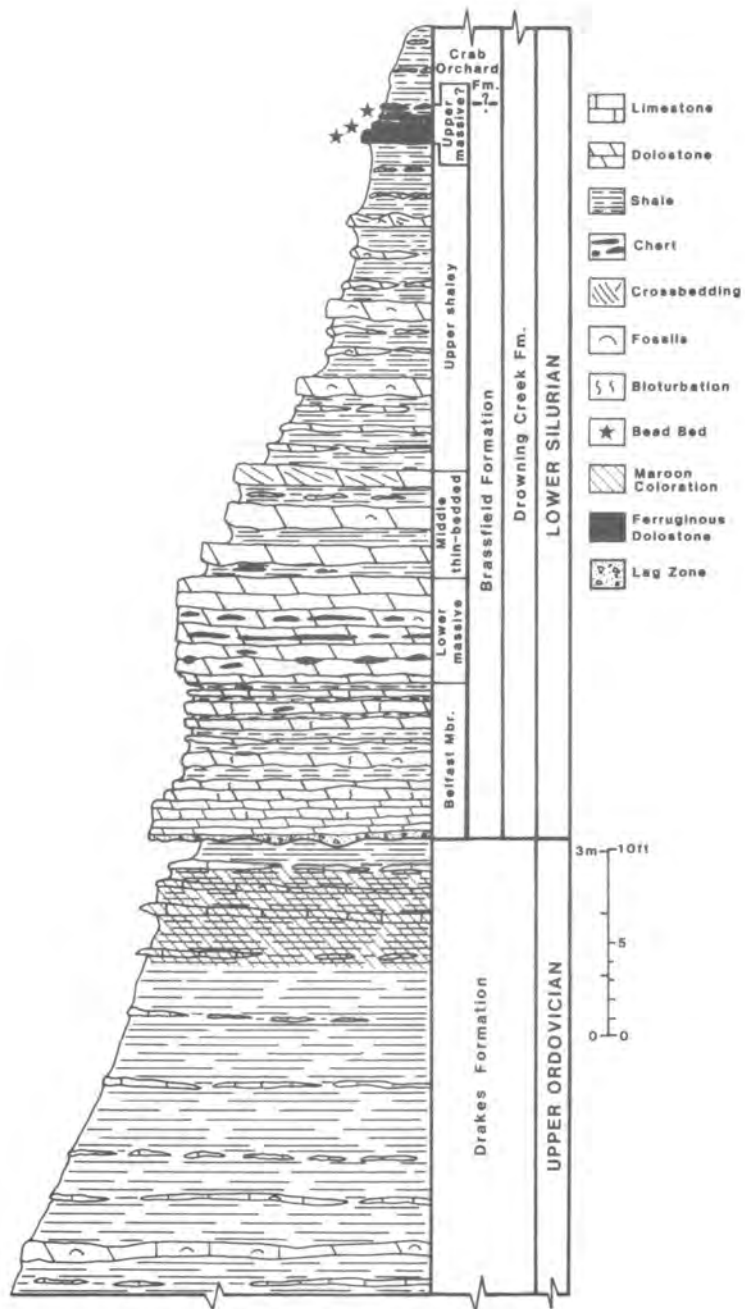


Figure 95. Schematic section showing the Upper Ordovician-Lower Silurian rocks at Cabin Creek on Kentucky State Route 10. Red beds in the Drakes represent one of the most western expressions of the Queenston red-bed lithofacies.

Although there are questions about what Foerste (1896) defined as the Belfast Bed (*e.g.*, Kleffner and others, 1989; Kleffner and Riddle, 1990; Potter and others, 1991), Foerste (1935) himself described the Belfast Bed to occur as far south as Fleming County. The Belfast in Fleming County is far more argillaceous and shaley than that seen here and hence coincides with Foerste's original definition of the unit. Nonetheless, basal parts of this section are certainly homotaxial with the Belfast Bed as used and defined by Foerste (1896, 1935) and could merely represent a more dolomitic variation of the unit.

The overlying 5.8 ft (1.8 m) of the Brassfield consist of massive, vuggy dolostones with prominent white chert stringers up to 6.0-in (15-cm) thick. Dolomitization has destroyed most indications of original fabric and fossils, but whole fossils can be found in the chert stringers. Thin sections of these cherts indicate that most of these dolostones were composed of grainstones (Gordon and Ettensohn, 1984).

Remaining parts of the Brassfield include approximately 27 ft (8.2 m) of interbedded dolostones and shale. Individual dolostones are 1.5- to 12-in (4- to 30-cm) thick and may be amalgamated beds or individual beds with sharp, erosional bases, crossbedding, and ripple marks; some consist wholly of megaripples or hummocky crossbeds. Many are bioturbated and exhibit nearly in-place fauna on their upper surfaces, including crinoids, cystoids, gastropods, rugose and tabulate corals, tentaculitids, bryozoans, brachiopods, and trilobites. Lower parts of this dolostone-and-shale sequence in which the dolostones predominate coincide with the "thin-bedded unit" of Gordon and Ettensohn (1984), whereas upper parts in which shales dominate are the same as their "upper shaly unit."

Whether or not these interbedded dolostones and shales should be included within the Brassfield Formation is a local stratigraphic problem. As defined by Foerste (1906), the top of the Brassfield is the last major dolostone or limestone in the section. From Fleming County southward, this last major dolostone coincides with the "upper massive unit" of Gordon and Ettensohn (1984) and normally includes or just overlies the "Bead Bed," a horizon of large, cogwheel-like crinoid columnals on the east side of the Cincinnati arch. Foerste (1906, 1935) first noted the occurrence of the Bead Bed and used it to approximate the top of the Brassfield Formation, a practice used by subsequent workers (*e.g.*, Rexroad and others, 1965; Rexroad, 1967; Gordon and Ettensohn, 1984).

The upper massive unit continues northward as far as Fleming County at which point it either pinches out or becomes too thin to recognize, which is the case in this exposure (Figs. 95, 96). When the upper massive unit becomes too thin to recognize, most workers have used the Bead Bed to define the top of the Brassfield (Fig. 95). The problem with this usage is that it relies on biostratigraphy to define a formation boundary and no longer conforms to Foerste's original definition of the Brassfield. McDowell (1983) sought to remedy this situation by including the Brassfield as the lowermost member of the new Drowning Creek Formation (Fig. 1), which includes within its upper portions some of the lower, substantially dolomitic parts of the former Crab Orchard Formation. The top of the Drowning Creek Formation, however was not defined consistently; in some places its top was defined as the top of the Oldham Member of the Crab Orchard, in others, as the top of the Waco or Dayton Dolostone Member of the Crab Orchard (McDowell, 1983). Obviously, neither the old nor new stratigraphic nomenclature is wholly satisfactory, and further work will be required.

The base of the Brassfield in this area represents a typical flooding surface at the base of a transgressive systems tract. Three other such surfaces are probably present in the typical Drowning Creek or Brassfield-Crab Orchard sequence: one at the base of the upper massive unit, a second one at the base of the Oldham Member and a final one at the base of the Waco or Dayton Dolostone Member (Fig. 97). These surfaces may correlate with sequence boundaries noted by Brett and others (1990) in comparable parts of the Silurian section from New York.

The major part of the Brassfield appears to represent a single transgression (Fig. 96) (Gordon and Ettensohn, 1984). The Belfast is interpreted to represent protected lagoonal environments shoreward of a migrating sandbelt represented by the lower massive unit. Although the Belfast is more dolomitic here than is typical of the unit elsewhere, the increased number of thin dolostones may merely reflect storm-related spillover lobes dumped into the lagoon, because, some of the dolostones still preserve storm-related features. The middle thin-bedded unit and the upper shaly unit are interpreted to represent shallow open-marine and deeper open-marine environments respectively seaward of the sandbelt (Fig. 98). Both units contain substantial amounts of shale suggesting depths greater than normal wave base, but include dolostone beds, which appear to represent sandbelt-derived tempestites, and which decrease in thickness and number upward, suggesting an upward proximity trend toward deeper and more distal environments (*e.g.*, Aigner, 1985). The abundance of fossils in this upper part of the Brassfield apparently reflects the presence of firm substrates formed by tempestites and relatively stable conditions below wave base. The general nature of environmental indicators is shown in Figure 97.

This transgressive sequence was abruptly interrupted by a drop in sea level represented by a minor unconformity at the base of the upper massive unit (Figs. 96, 97, 98). South of Lewis County, the presence of phosphorite nodules, glauconitic sands and conglomerates at the base of this unit suggest that this is another transgressive flooding surface, and hence a possible sequence boundary comparable to the one at the base of the Brassfield (Figs. 96, 97, 98). The upper massive unit is typically an oolitic and skeletal grainstone which is heavily crossbedded, ferruginous, and may contain the Bead Bed (Foerste, 1906, 1935; Gordon and Ettensohn, 1984). In fact, this unit is so ferruginous in places that it was locally mined for iron ore (Foerste, 1906, 1935; Stokely, 1948). Although the unit and its underlying disconformity are not well developed here, the upper massive unit is present near the top of the exposure and exhibits ferruginous grainstones, crossbedding, and the Bead Bed (Fig. 95). As with the poor development of typical peritidal features in the Drakes Formation, the poor development of this unit may reflect increasing distance and greater depths away from the Cincinnati arch.

Johnson and others (1985) and Johnson (1987) contended that parts of the Early and Middle Silurian exposed in eastern Kentucky exhibit transgressive and regressive sequences that correspond to similar fluctuations on North American and around the world. Based on the seemingly worldwide nature of these correlations, they suggested that these fluctuations are related to eustatic variations in sea level, and this seems reasonable given the presence of Ordovician-Silurian glaciation (Hambrey, 1985; Caputo and Crowell, 1985). However, based on correlations with the Appalachian basin (Fig. 91) and on subsidence curves (Fig. 11), the latest two fluctuations including the Waco-Dayton Dolostone and the Bisher

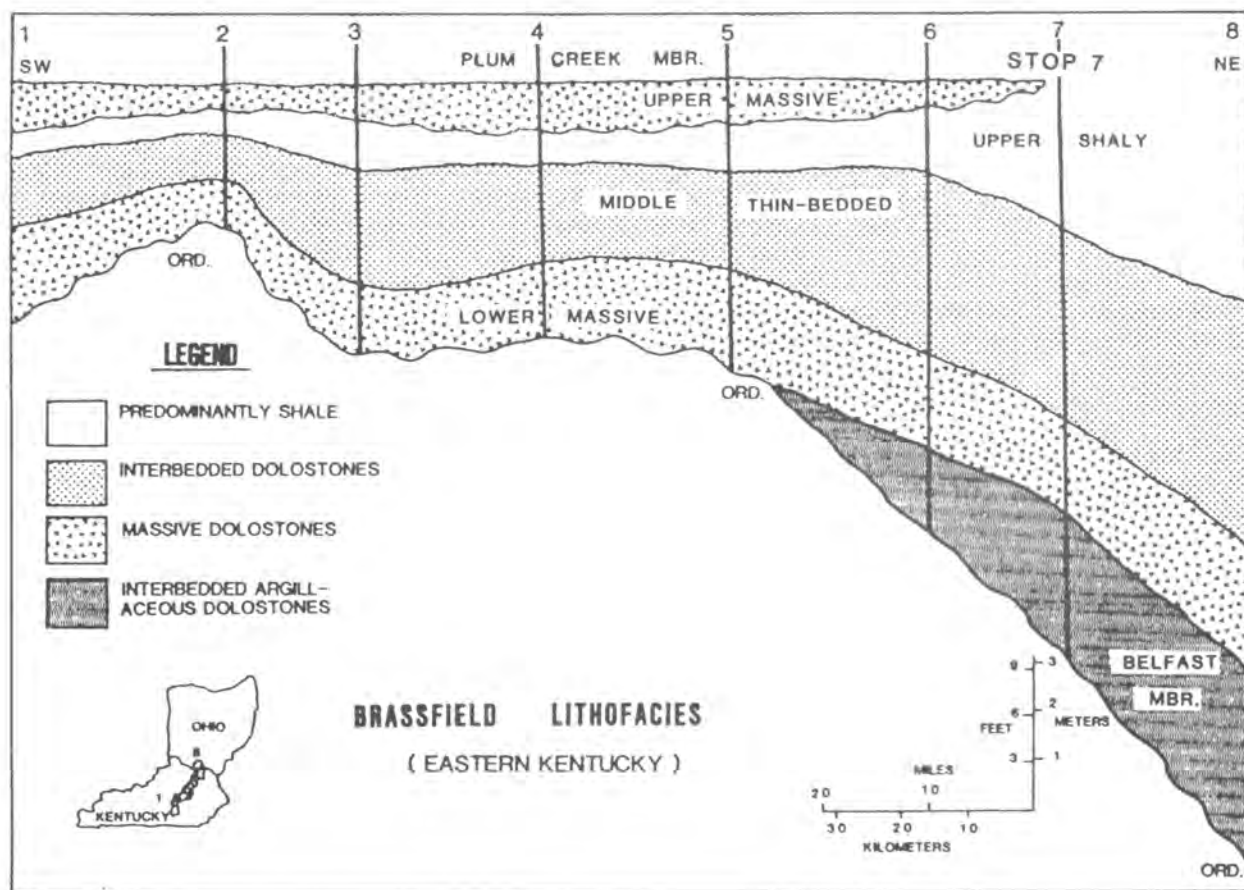


Figure 96. Schematic cross section from south-central Kentucky into south-central Ohio showing the nature and disposition of major Brassfield lithofacies along the outcrop belt based on eight measured sections (from Gordon and Ettensohn, 1984).

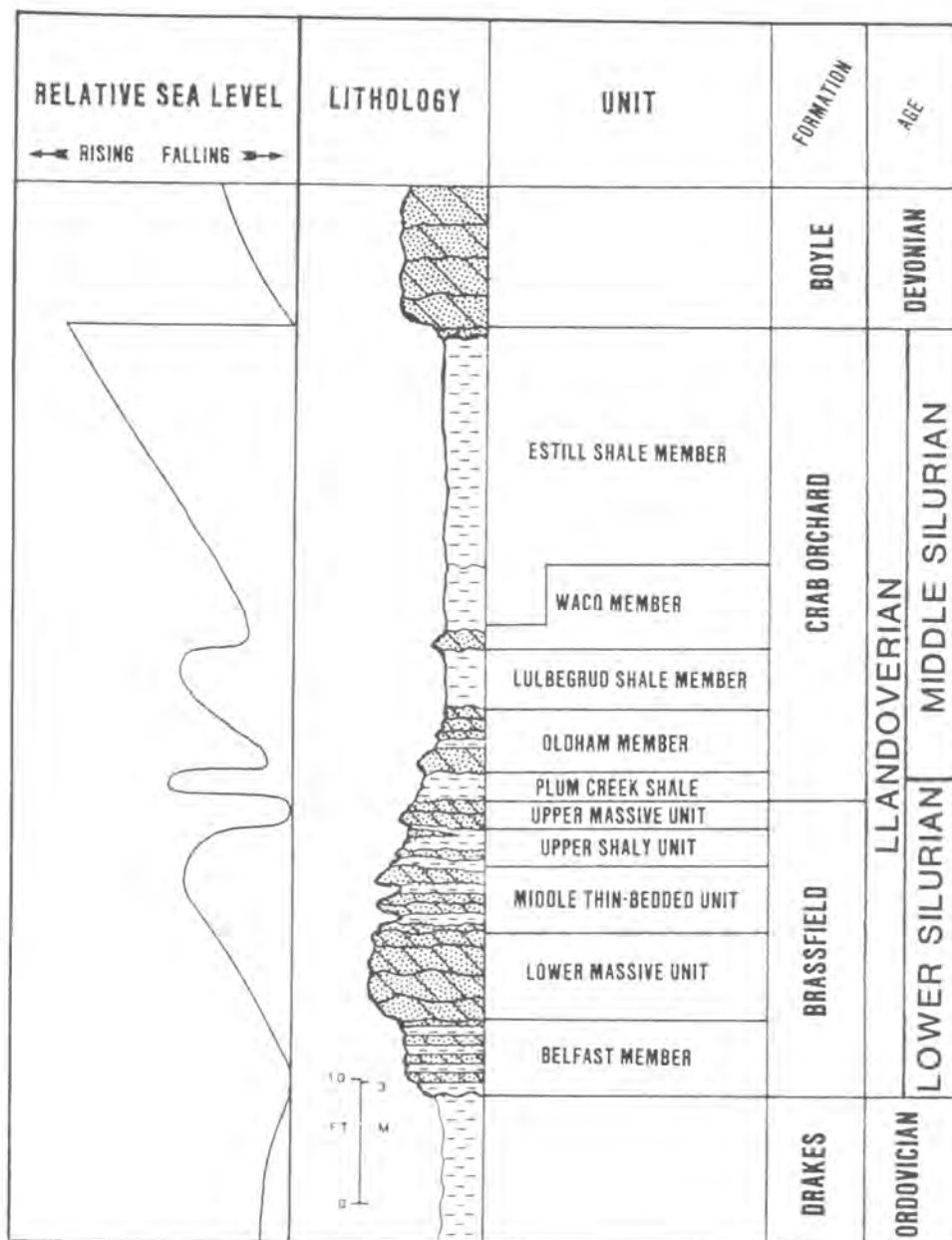


Figure 97. Composite stratigraphic column of Lower Silurian rocks in east-central and northeastern Kentucky showing informal and formal members in the section and relative sea-level fluctuations. All lithologies are dolostone or shale (modified from Gordon and Ettensohn, 1984).

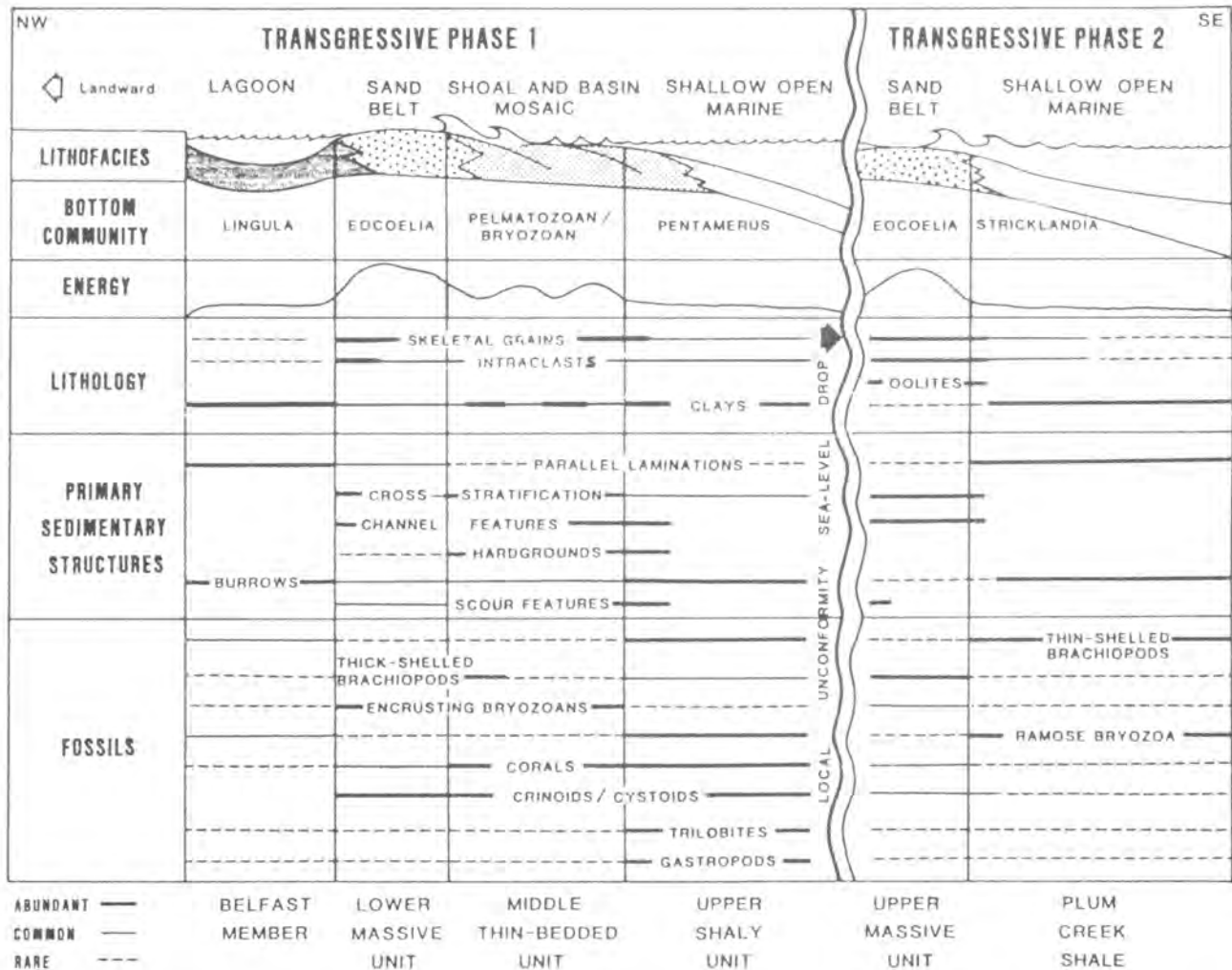


Figure 98. Hypothetical reconstruction of Brassfield depositional environments and environmental continua with discriminating lithologic, sedimentologic, and paleontologic criteria. A subtle unconformity below the upper massive unit divides the Brassfield into two distinct transgressive continua (from Gordon and Ettensohn, 1984).

may contain substantial tectonic components of subsidence. Only the stratigraphic expression of the two earliest Silurian eustatic events is present in this exposure (Fig. 95).

Reload and the bus will turn around and proceed east on Kentucky State Route 10.

- 71.9 0.9 Junction of Kentucky State Route 10 and Epworth Road. Turn right (southeast) on Epworth Road.
- 72.0 0.1 Junction of Epworth Road and Kentucky State Route 546. Turn right onto Kentucky State Route 546 and proceed west.
- 72.6 0.6 Preachersville Member of the Drakes Formation and Brassfield Dolostone exposed in roadcut on left side of highway.
- 73.0 0.4 Bridge over Cabin Creek.
- 73.1 0.1 Preachersville Member of the Drakes Formation and Brassfield Dolostone exposed in roadcuts on both sides of highway.
- 73.4 0.3 Power lines of Kentucky Utilities.
- 76.1 2.7 Junction of Kentucky State Routes 546 and 57 on left-hand side. Continue west on 546.
- 77.9 1.8 Brassfield Dolostone exposed in roadcuts on both sides of highway (roadcut extends for some 0.2 mi).
- 79.2 1.3 Junction of Kentucky State Routes 546 and 1234 on right-hand side. Continue west on Route 546.
- 79.7 0.5 Junction of Kentucky State Routes 546 and 1237. Small exposures of Bull Fork Formation (Upper Ordovician) exposed in ditchline on right side. Continue west on Route 546.
- 79.8 0.1 Junction of Kentucky State Routes 546 and 1234 on left-hand side. Continue west on Route 546.
- 81.8 2.0 Small exposures of Bull Fork Formation (Upper Ordovician) present in ditchline on right side.
- 82.5 0.7 Bull Fork Formation exposed in roadcuts on left side of highway.
- 82.9 0.4 Junction of Kentucky State Routes 546 and 1449 on right-hand side. Continue west on Route 546.
- 83.4 0.5 Bull Fork Formation exposed in roadcuts on left (south) side of highway (roadcut extends for some 0.15 mi).
- 84.4 1.0 Bull Fork Formation exposed in roadcuts on both sides of highway (roadcut extends for some 0.1 mi).
- 84.7 0.3 Bull Fork Formation exposed in roadcuts on both sides of highway.
- 85.5 0.8 Railroad bridge.

- 85.8 0.3 Bull Fork Formation exposed in roadcuts on both sides of highway.
- 86.6 0.8 Railroad bridge.
- 86.9 0.3 Bull Fork Formation exposed in roadcuts on both sides of highway.
- 87.2 0.3 Bull Fork Formation exposed in roadcuts on both sides of highway.
- 87.7 0.5 Bull Fork Formation exposed in roadcuts on both sides of highway.
- 87.9 0.2 Junction of Kentucky State Routes 546 and 1448 on right-hand side. Continue west on Route 546.
- 88.1 0.2 Bull Fork Formation exposed in roadcuts on both sides of highway.
- 88.3 0.2 Junction of Kentucky State Routes 546 and 11. Turn right onto State Route 11 and proceed northward down the hill. STOP 8. Lithostratigraphy, Cyclic Sedimentation, and Event Stratigraphy of the Maysville, Kentucky, Area. (Roadcuts along Kentucky State Routes 11 and 546, approximately 1.5 to 3.0 mi south of Maysville, Kentucky; Mason Co.; Carter Coordinates: 20-Z-69.) Going north along Kentucky State Route 11, there are seven roadcuts which expose some 250 ft (72 m) of Upper Ordovician rocks. Included here, in ascending order, are the Kope, Fairview, Grant Lake and the Bull Fork formations.

LITHOSTRATIGRAPHY, CYCLIC SEDIMENTATION, AND EVENT STRATIGRAPHY OF THE MAYSVILLE, KENTUCKY, AREA

GREGORY A. SCHUMACHER

INTRODUCTION

The focus of this stop will be the lithostratigraphy, cyclic sedimentation, and event stratigraphy of the Upper Ordovician, Cincinnati Series exposed in the Maysville, Kentucky area. The excellent roadcuts created by the improvement of Kentucky State Route 11 and the construction of the Alexandria-to-Ashland Highway (State Route 546) provide more than 295 ft (90 m) of Upper Ordovician strata for study. Major shoaling-upward cycles will be discussed along with comments on event beds which, in some cases, are traceable over 390 sq. mi (Potter and others, 1991; Schumacher, 1992).

LITHOSTRATIGRAPHY

Historical Background

The Cincinnati Series of Kentucky, Indiana, and Ohio is characterized by undeformed, flat-lying, interbedded carbonates and siliciclastic rocks. The history of Cincinnati stratigraphic nomenclature has its roots the 1820's and continues today. Nickles (1902), Cumings (1922), Weiss (1961), Weiss and Norman (1960), and Tobin (1986) provided excellent summaries. In the late 1830's, the terms "Blue Limestone" (Locke, 1838) and "Great Limestone

Deposits" (Briggs, 1838) were introduced to describe the stratigraphy of the Cincinnati Series. In 1865, these units were replaced by the Cincinnati Group (Meek and Worthen, 1865). These thick, undifferentiated units were successively subdivided into about 40 stratigraphic units between 1870 and 1910 (Fig. 99). Ruedemann (1901) provided the impetus for stratigraphic subdivision by proving that the rocks of the Cincinnati Series were younger and thus stratigraphically above the standard Ordovician section of New York. The rocks of the Cincinnati area became the standard reference section for Upper Ordovician rocks of North America (Clarke and Schuchert, 1899). Thus, defining the stratigraphy of this reference section and correlating these units to coeval sections was undertaken between 1900 and 1920. Nickles, Foerste, and Bassler followed the practice of James Hall and described the lithology and diagnostic fossils of each of these new units (Fig. 99). These stratigraphers recognized lithologic changes as they correlated the "layer-cake" or flat-lying rocks throughout the region (e.g., Nickles, 1905; Bassler, 1906; Foerste, 1912). However, the diagnostic fossils used in correlation could be traced throughout the region. Hence, these units were correlated and became the standard stratigraphy throughout Kentucky, Ohio, and Indiana.

Other than regrouping, renaming, and changing in unit class or rank, little changed with the stratigraphic nomenclature of the Cincinnati Series between 1914 and 1961 (Weiss, 1961). In the 30 years following Weiss's (1961) work, however, tremendous advances occurred in the understanding of the lithostratigraphy, the lithofacies, and the depositional environments of the Cincinnati Series. Geologic mapping conducted in Kentucky, and Indiana resulted in the replacement of much of the early stratigraphy with stratigraphic units more appropriate for those areas. In Ohio, the stratigraphic distribution of some of the early stratigraphic units has been mapped and the facies relationships documented (e.g., Ford, 1967; Schumacher and others, 1991). Additional investigations are underway to document the stratigraphic and facies relationships of all of the early stratigraphic units of the Cincinnati Series in southwestern Ohio.

Lithostratigraphy at Stop 8

Three formations are exposed in the seven roadcuts along State Route 11. The basal 50 ft (15.2 m) of this composite section is the Kope Formation. The Kope Formation is overlain by 92.8 ft (28.3 m) of the Fairview Formation. The remaining 108.2 ft (33 m) are assigned to the Grant Lake Limestone. The lower 32.8 ft (10 m) of the Bull Fork Formation are present in a number of roadcuts adjacent to the intersection of State Route 11 and the Alexandria-to-Ashland Highway (Figs. 100, 101).

The Kope Formation consists of interbedded shale, limestone, and calcareous siltstones with shale averaging 66% of the formation. Beds are typically thin- to thick-bedded, and most beds are planar to lenticular. Shale beds are primarily mudstones with platy to flaggy parting, conchoidal fracture, and along with the siltstone beds, are sparsely fossiliferous. Most limestone beds are commonly fossiliferous packstones and grainstones displaying sole and tool marks, crude graded bedding, shale and calcisiltite intraclasts, ripple marks and megaripples, and small-scale hummocky cross-strata. The contact with the overlying Fairview Formation is sharp.

The lithologies of the Fairview Formation are similar to those of the Kope Formation. The two formations are distinguished from one another by: 1.) a decrease in the average percentage of shale from

66% to 40%, 2.) a decrease in shale-bed thickness, and 3.) the presence of a basal series of diagnostic, wavy- to irregular-bedded limestones containing abundant specimens of the brachiopod *Strophomena* (Fig. 101).

The *Strophomena*-bearing limestone beds are widespread and have been mapped throughout the Maysville, Kentucky, area and into southern Ohio (Fig. 102). These beds produce a diagnostic "kick" on shale-percentage and geophysical logs (Fig. 101), and have been successfully traced into the shallow subsurface of: 1.) northeastern Kentucky (Potter and others, 1991), 2.) southern and central Ohio (Stith, 1986; Schumacher, 1992), 3.) eastward toward Portsmouth, Ohio (Potter and others, 1991), and 4.) westward and southward into Kentucky (Weir and others, 1984).

A second zone of interest occurs 32.8 ft (10.0 m) below the Fairview Formation-Grant Lake Limestone contact (Fig. 101). This zone is characterized by one to three beds of soft-sediment deformation, separated by interbedded limestone and shale. This unit varies in thickness from 6.6 to 13.1 ft (2.0-4.0 m) and has been mapped throughout much of the Maysville, Kentucky, area (Fig. 102). Three deformed beds are present at this stop, and they contain convolute laminae, slickensides, ball-and-pillow structures, and are laterally discontinuous.

The Fairview Formation grades into the overlying Grant Lake Limestone. The Grant Lake Limestone is characterized by interbedded wavy- to irregular-bedded limestone with minor fossiliferous shale beds and partings; limestone averages 77%. Sedimentary structures are rare, but rare tempestites (storm beds) with sole marks, crude graded bedding, ripple marks, and small-scale hummocky cross-strata occur locally.

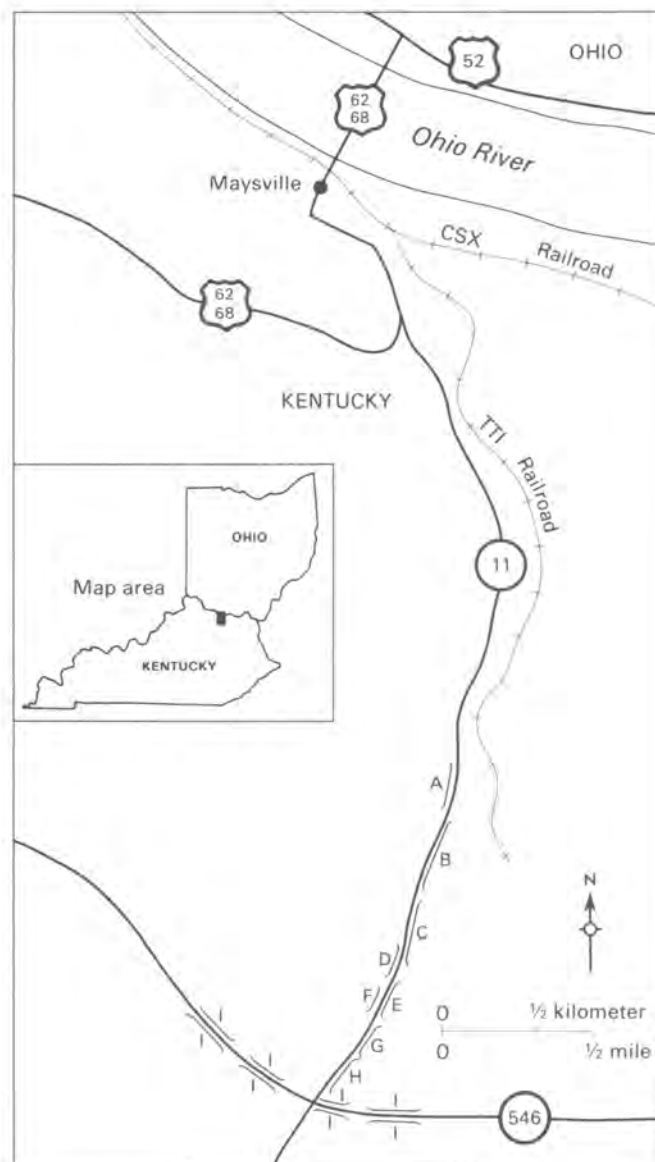
Schumacher and others (1991) subdivided the Grant Lake Limestone, in ascending order, into the Bellevue Member, the Coryville Member, and the Straight Creek Member. The Bellevue and Straight Creek members are characterized by wavy- to irregular-bedded, fossiliferous limestone beds with minor, fissile, fossiliferous, shale interbeds. Shale averages 18%. The Coryville Member differs by: 1.) an increase in the amount of shale which averages 38 % (Fig. 101), 2.) an increase in shale-bed thickness, 3.) sparse fossil content in shale beds, and 4.) an increase in the number of continuous, irregular-bedded limestone beds. Thickness of the Bellevue Member at this stop is 69.9 ft (21.3 m) and it gradually decreases to 19.7 ft (6.0 m) in the vicinity of Cincinnati, Ohio (Schumacher and others, 1991). The Coryville Member is 16.1-ft (4.9-m) thick in the Maysville area and thickens to 59.0 ft (18.0 m) in the Cincinnati area. The Straight Creek Member grades laterally into the Mount Auburn Member of the Grant Lake Formation and remains about 19.7 ft (6.0 m) in thickness across southwestern Ohio.

Two interesting beds are present in the Straight Creek Member of the Grant Lake Limestone at this stop. A prominent grainstone bed occurs in the first roadcut north of the intersection of State Route 11 and 546 on the east side of Kentucky Route 11 (Fig. 100, number 7). This bed is lenticular, contains sharp upper and lower contacts, and displays ill-defined cross-bedding. Also, a one-centimeter, pyrite-rich bed occurs two feet (0.6 m) below the prominent grainstone bed. This bed is irregular-bedded, somewhat continuous, oxidized, fossiliferous, and probably represents a minor diastem.

The contact with the overlying Bull Fork Formation is not exposed at this stop. Good exposures of the lower part of this formation are

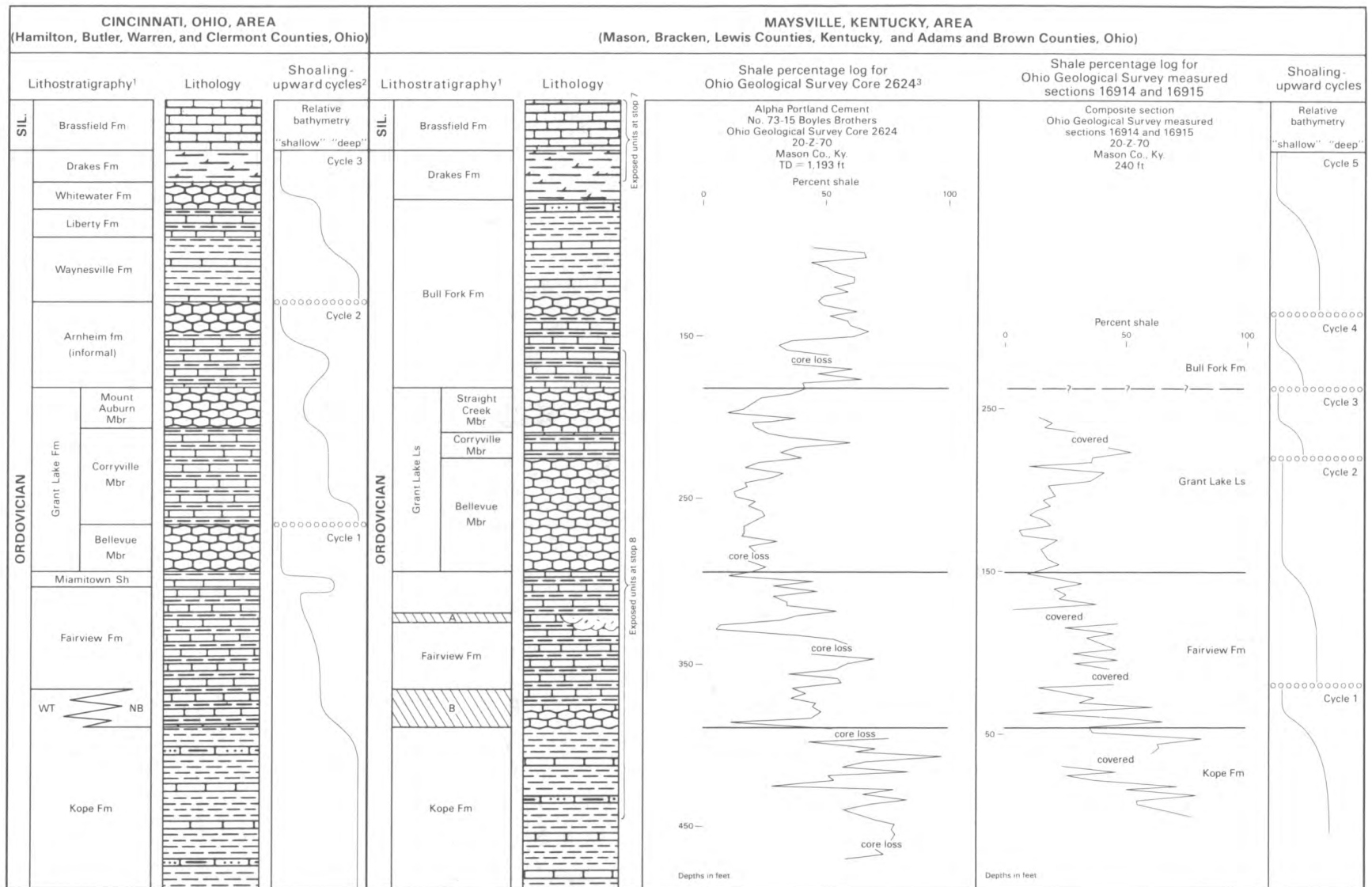
ORTON, 1873		NICKLES, 1902	
CINCINNATI BEDS		LEBANON	
		NICKLES, 1903	
		FOERSTE, 1903	
		FOERSTE, 1905	
		BASSLER, 1906	
		Saluda	
		Whitewater	
		Liberty	
		Waynesville	
		Arnheim	
		McMillan	Mt. Auburn
			Corryville
			Bellevue
		Fairview	
		Eden	
		Pt. Pleasant	

Figure 99. Summary of the major Ordovician stratigraphic units introduced into the Cincinnati and Maysville areas between 1871 and 1910. Stratigraphic nomenclature, which described biostratigraphic units or which has been abandoned in modern usage, is not included.



- A. Kope-Fairview contact and *Strophomena*-bearing limestone beds.
- B. Kope-Fairview-Grant Lake formational contacts, *Strophomena*-bearing limestone beds, and zone of soft-sediment deformation.
- C. Fairview-Grant Lake contact, excellent exposure of zone of soft-sediment deformation.
- D, E, and F. Bellevue Member of the Grant Lake Limestone.
- G. Bellevue and Corryville Members of the Grant Lake Limestone.
- H. Straight Creek Member of the Grant Lake Limestone with prominent grainstone bed and pyrite bed.
- I. Lower part of the Bull Fork Formation.

Figure 100. Location map of Stop 8 showing individual roadcuts and diagnostic features at each roadcut exposure.

¹After Schumacher and others (1991)²Modified from Tobin (1982)³See figure 4 for bore-hole location

○○○○ Shoaling-cycle boundary

WT Wesselman Tongue of the Kope Formation

NB North Bend Tongue of the Fairview Formation

A Marker zone of soft-sediment deformation

B Marker zone of *Strophomena*-bearing limestones

Planar-bedded shale with minor planar-bedded limestone



Planar-bedded limestone and shale



Wavy- to nodular-bedded limestone with minor irregularly bedded shale



Dolomitic shale



Soft-sediment deformation



Planar to irregularly bedded limestone

Figure 101. Schematic diagram comparing the lithostratigraphy, lithologies, and shoaling-upward cycles in the Cincinnati rocks of the Cincinnati and Maysville areas.

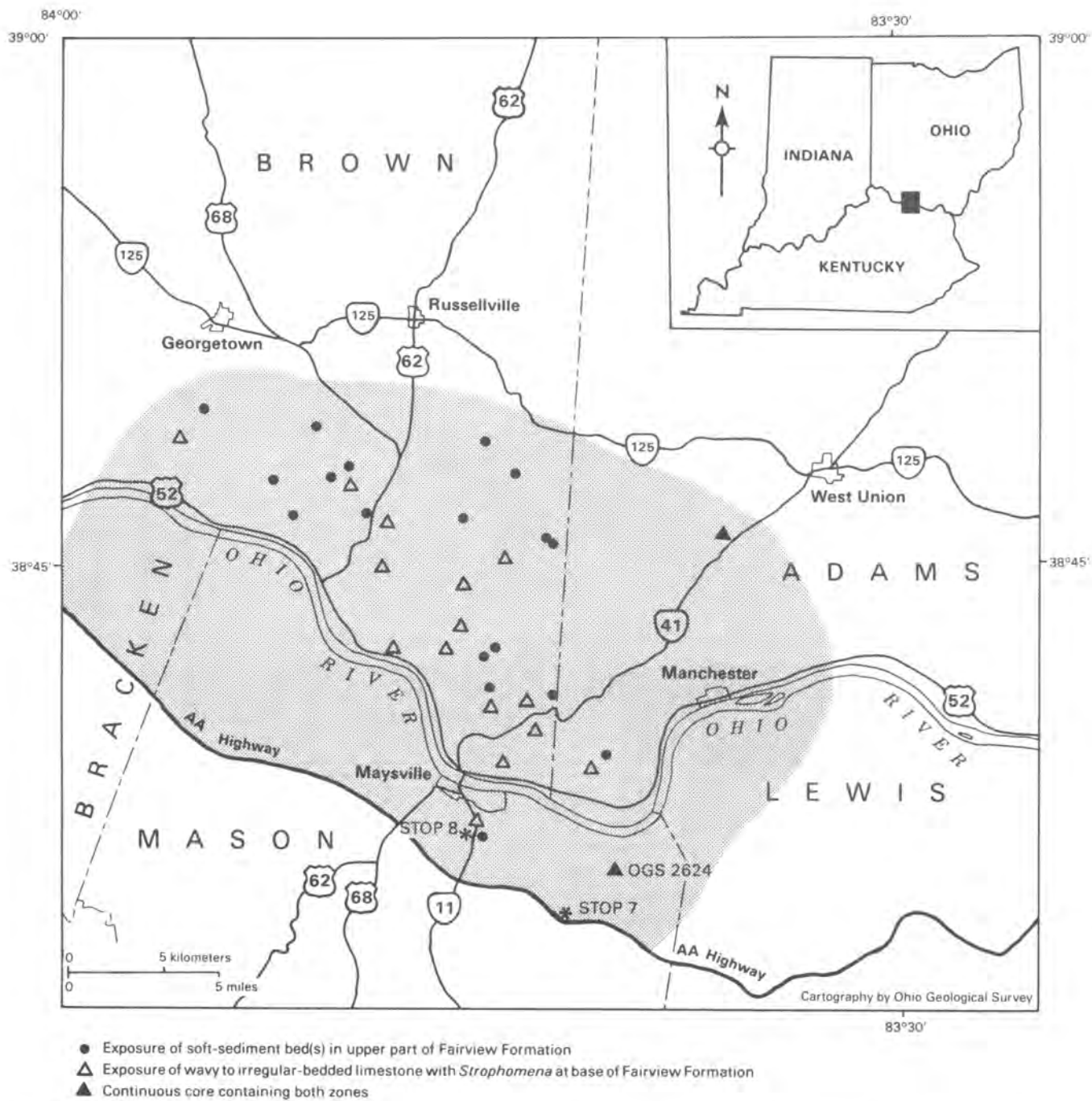


Figure 102. Distribution of marker zones within the Fairview Formation in the Maysville, Kentucky, area. Shading illustrates the inferred limits of both zones north of Kentucky State Route 546 (Alexandria-Ashland Highway) (modified from Potter and others, 1991).

present on the Alexandria-to-Ashland Highway (State Route 546) east and west of its intersection with State Route 11 (Fig. 100).

CYCLIC SEDIMENTATION

The interbedded shales, calcisiltites, and limestone which compose the Cincinnati Series were deposited in a broad epeiric sea which inundated much of the North American midcontinent throughout the Late Ordovician (Weir and others, 1984; Frey, 1987). The interaction of seafloor topography, eustatic changes, tectonic activity, and storm sedimentation produced genetically related lithofacies ranging from offshore "deeper-water" to shoreface "shallow-water" (Anstey and Fowler, 1969; Hay, 1981; Tobin, 1982; and Jennette and Pryor, in press). "Deeper water" lithofacies represent depths of 65 to possibly 165 ft (20-50 m), whereas "shallow-water" lithofacies probably represent depths of a few tens of feet (Bucher, 1917; Anstey and Fowler, 1969; Anstey and others, 1987). Anstey and Fowler (1969), Hay (1981), and Tobin (1982) differ in their interpretations relative to the beginning and ending of a given shoaling-upward cycle but agree that three shoaling-upward cycles are present in the Cincinnati Series exposed in the Cincinnati, Ohio, area (Fig. 101).

Anstey and Fowler (1969), Hay (1981), Tobin (1982), and Jennette and Pryor (in press) defined a typical shoaling-upward cycle in the Cincinnati Series as consisting of three or four distinct lithofacies. The offshore "deeper water" lithofacies averages 98-ft (30-m) thick, is shale-dominant, and contains medium to thick beds of planar-bedded shale and limestone. Shale beds are sparsely fossiliferous and limestone beds are highly fossiliferous. Lithofacies of transitional or intermediate depths are typically less than 98-ft (30-m) thick, contain shale and limestone in roughly equal amounts, and display medium-bedded shale and limestone which are planar- to irregular-bedded. Shale beds are sparsely fossiliferous and limestone beds are very fossiliferous. The "shallow-water" lithofacies is represented by two types of lithofacies: 1.) a limestone-dominant lithofacies, or 2.) a dolostone-dominant lithofacies. The limestone-dominant lithofacies is also generally less than 49-ft (15-m) thick, and consists of discontinuous, fossiliferous, wavy-, irregular- or nodular-bedded limestone and shale. The dolostone-dominant lithofacies is generally less than 49-ft (15-m) thick, but consists of massive dolostone and minor dolomitic shale. Bedrock geologic mapping supplemented with continuous cores resulted in the recognition of five shoaling-upward cycles in the Upper Ordovician section of the Maysville, Kentucky, area (Fig. 101). At Stop 8, we will examine four of the five shoaling-upward cycles. The upper part of the fifth cycle was present at Stop 7. The first cycle originates in the shale-dominant middle portion of the Kope Formation (subsurface), continues through the transitional, limy upper part of the Kope Formation, and terminates in the wavy- to irregular-bedded, *Strophomena*-bearing limestone and shale beds in the basal Fairview Formation. The second and third shoaling-upward cycles lack well developed offshore lithofacies but grade from a shale-dominant transitional lithofacies to a "shallow-water," limestone-dominant lithofacies. The second cycle ranges from the top of the *Strophomena*-bearing beds through the Bellevue Member of the Grant Lake Limestone. The third cycle originates in the transitional lithofacies of the Corryville Member of the Grant Lake Limestone and terminates at the top of the "shallow-water" lithofacies of the Straight Creek Member of the Grant Lake Limestone. The basal part of the fourth cycle is exposed in the roadcuts along the Alexandria-to-Ashland Highway east and west of its intersection with State Route 11. The shale-dominant transitional lithofacies of the basal Bull Fork Formation is well exposed. The nodular-bedded "shallow-water" lithofacies is poorly exposed in the vicinity of Stop 8.

However, exposures of this lithofacies can be observed in a series of roadcuts beginning two miles west of the intersection of State Route 546 and 11. The contact of the fourth and fifth cycles is poorly exposed throughout the Maysville, Kentucky, area. However, this gradational contact is readily apparent in core and in the shale-percentage logs produced for each core (Fig. 101).

Anstey and Fowler (1969), Hay (1981), Tobin (1982), and Jennette and Pryor (in press) concluded that these shoaling-upward cycles reflect major transgressions and regressions. They differ, however, on the cause of these fluctuations. Anstey and Fowler (1969) speculated that the influence of Late Ordovician glaciation produced these changes in sea level. Hay (1981) suggested local or regional tectonic activity and/or Late Ordovician glaciation as the cause, whereas Tobin (1982) believed basin filling and/or basin subsidence produced the fluctuations in sea level.

Jennette and Pryor (in press) utilize the principles of sequence stratigraphy to explain these cycles. They suggest that shoaling-upward cycles were produced by the stacking of numerous meter-scale parasequences. Each parasequence represents numerous rhythmic changes in sea level as the result of glacio-eustasy.

Ettensohn (1991a) utilized recent refinements in the understanding of Taconian tectonism (e.g., Vick and others, 1987; Drake and others, 1989) and in lithospheric-flexure models (e.g., Beaumont and others, 1988; Jamieson and Beaumont, 1988) to clarify the relationships between tectonism and Ordovician sedimentation in central and southern parts of the Appalachian basin and on adjacent parts of the craton. In the late Middle Ordovician (Rocklandian), the Taconic highlands were uplifted in the area of the present-day central Appalachians (Fig. 13). Adjacent to this uplift, the lithosphere was depressed forming a foreland basin (Fig. 8, 13). Northeastern Kentucky was affected by the development of a secondary uplift, termed a peripheral bulge (Fig. 8). The intervening foreland basin slowly filled with sediments shed from the areas of uplift. The erosion of the Taconic highlands throughout the early and middle Late Ordovician removed the initial load placed on the lithosphere. This resulted in lithospheric rebound producing uplift of this region and adjacent portions of the foreland basin. In northeastern Kentucky, this would have resulted in the depression of the cratonic crust, forming an anti-peripheral bulge.

This interpretation may explain the origin of some of the shoaling-upward cycles observed at Stops 7 and 8. Moreover, the shoaling-upward cycle ranging from the middle portion of the Fairview Formation through the Bellevue Member of the Grant Lake Limestone occurs at the correct time and stratigraphic position to represent sedimentation associated with the passage of a peripheral bulge. The abrupt return to the offshore facies of the middle part of the Bull Fork Formation (Fig. 7) would indicate the development and passage of the anti-peripheral bulge. As the anti-peripheral bulge migrated toward the eroding Taconic highlands, the resulting rebound coupled with the lowering of sea level associated of Late Ordovician glaciation would have produced the shoaling-upward cycle beginning in the middle of the Bull Fork Formation and ending at the top of the Drakes Formation (Fig. 7).

Although the flexure models may explain some of the observed shoaling-upward cycles, the presence of additional shoaling-upward cycles within this section suggests that other geologic processes may have been involved. The limited data available at this stop are

insufficient to address this question. Anstey and Fowler (1969), Hay (1981), Tobin (1982), Ettensohn (1991a), and Jennette and Pryor (in press) have suggested a number of hypotheses which should be tested in future investigations of the shoaling-upward Ordovician cycles in northeastern Kentucky and southern Ohio.

EVENT STRATIGRAPHY

Storm sedimentation is one of the dominant sedimentary processes in the Upper Ordovician Cincinnati Series (Anstey and Fowler, 1969; Kriesa and others, 1981; Tobin, 1982). Kriesa (1981) described an idealized storm sequence or tempestite as consisting of a sharp erosional basal contact with sole marks overlain by a limestone containing complete articulated fossils. The limestone bed grades into a cross-laminated unit consisting of sand- or silt-sized sediments which in turn grade into sparsely bioturbated shale. Only a few tempestites in the Cincinnati Series display this ideal stratigraphic succession. In contrast, most Cincinnati tempestites vary from this ideal succession by: 1.) the presence of a mixture of articulated and disarticulated fossils in the basal limestone, 2.) the absence of the basal limestone unit or the middle cross-laminated unit, and 3.) the presence of graded, millimeter-scale, silt-clay couplets termed "muddy tempestites." Tempestites are present in all of the stratigraphic units exposed at this stop. They are most common in the Kope and Fairview formations and least common in the Bellevue and Straight Creek members of the Grant Lake Limestone.

In addition to tempestites, the beds displaying soft-sediment deformation in the upper portion of the Fairview Formation probably represent at least three episodes of submarine slumping. The triggering mechanism(s) of this narrow, yet widespread zone of slumping is poorly understood. This zone may represent repeated seismic activity, regional tilting associated with fault displacement or lithospheric flexure, or destabilization of semi-lithified sediments by storm-generated currents impinging on the sea bottom.

CONCLUSIONS

The five shoaling-upward cycles present in the Maysville, Kentucky, area represent the interaction of shallow-marine carbonate and siliciclastic sedimentation with eustasy and regional tectonism. Ettensohn (1991a) suggested that lithospheric flexure associated with continental-margin tectonism can explain the stratigraphic relationships observed in Upper Ordovician rocks of central Kentucky (Fig. 7), and this model may be applicable in explaining the shoaling-upward cycles at Stop 8 as well. However, because subsurface relationships between this section and coeval sections in West Virginia, Virginia, and Tennessee are poorly understood, application of this model to these rocks may be premature. Future studies addressing this issue are needed before the relationship between Upper Ordovician sedimentation and continental-margin tectonism can be determined with confidence.

Proceed via Kentucky State Route 546, I-275, and I-75 to Cincinnati, Ohio.

END OF ROAD LOG FOR DAY THREE AND END OF FIELD TRIP.

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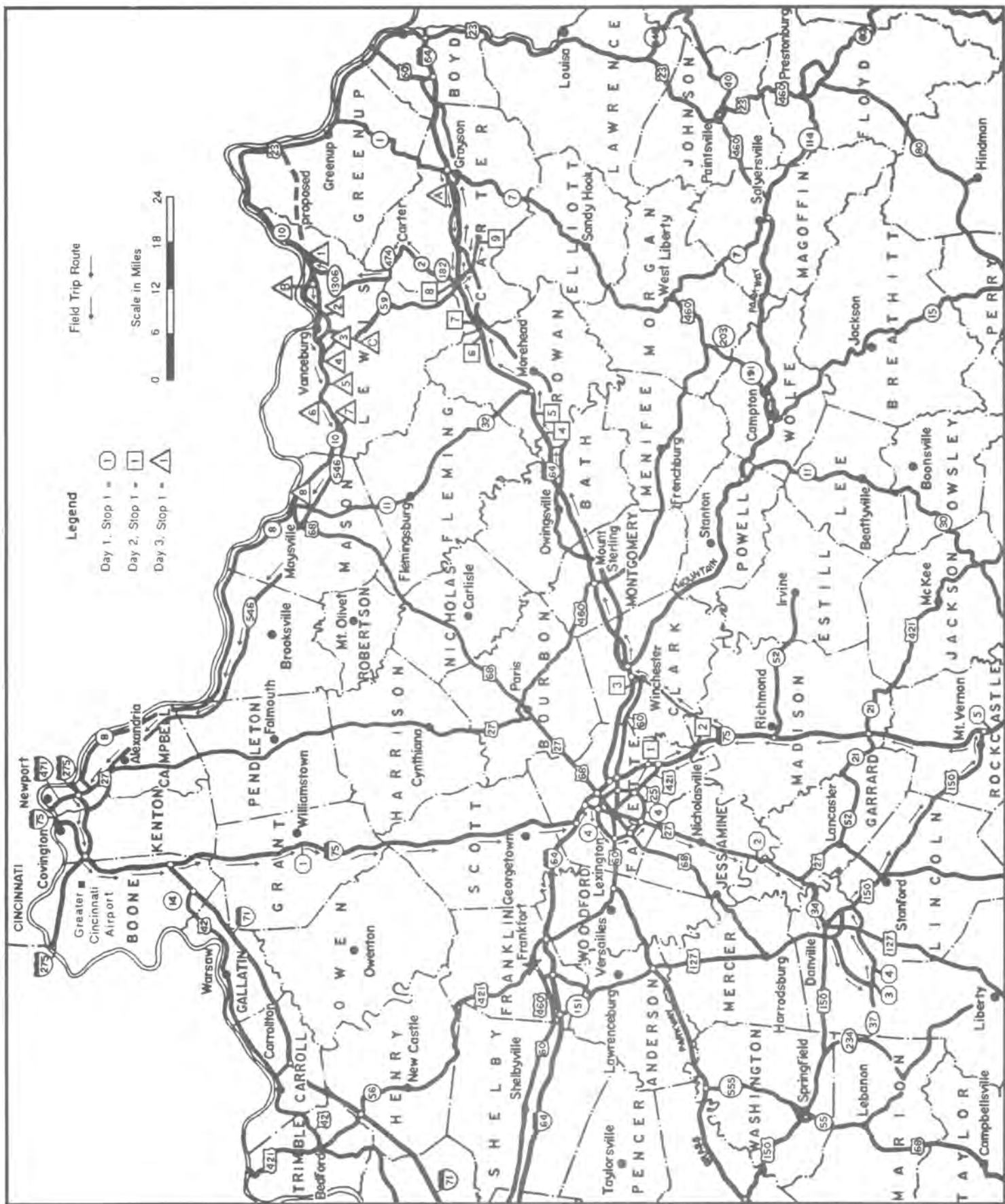
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MAP OF FIELD TRIP ROUTE