3. The Paleozoic Record of Changes in Global Climate and Sea Level: Central Appalachian Basin

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Introduction to the Field Trip

By C. Blaine Cecil

This field trip provides an update of an earlier trip that was conducted in conjunction with the 1998 Annual Meeting of The Geological Society of America (Cecil and others, 1998). The present trip emphasizes global climate change and eustatic sea-level change (allocyclic processes of Beerbower, 1964) (in a global paleogeographic context) as the predominant controls on Paleozoic sedimentation and stratigraphy in the Appalachian foreland basin. The trip stops illustrate long- to short-term paleoclimate change (table 1) as the predominant control on sediment supply, both chemical and siliciclastic (fig. 1A, B). Interpretations of climate variations are based on paleoclimate indicators such as paleosols, physical and chemical sedimentation, sedimentary geochemistry and mineralogy, and paleontology.

Selected stops in Paleozoic strata provide evidence for sedimentary response to paleo-tropical rainfall regimes that ranged in duration from long term to short term (table 1) and in rainfall amounts that ranged from arid to perhumid (table 2, fig. 1F). Eustatic processes also are discussed at each stop, and the role of tectonics may be mentioned briefly where appropriate. These later two allocyclic processes are generally accepted as the predominant control on sediment supply in contrast to climate, which is often not considered. For example, tectonic uplift is commonly considered as the predominant control on rapid influxes of siliciclastic material into depocenters, whereas siliciclastic sediment starvation is usually attributed to trapping of sediment in estuaries in response to sea-level rise. More specifically, basal sandstone units of the Cambrian, Silurian, Mississippian, and Pennsylvanian in the Appalachian basin are generally attributed to influx induced by tectonic uplift, whereas laterally extensive shale deposits, such as the Ordovician Martinsburg Formation and Devonian black shale units, are inferred to result from sediment trapping or autocyclic deposition of prodelta muds. Although tectonics clearly plays a role in uplift and subsidence, and eustasy also governs the amount of accommodation space, variation in climate, particularly long- to short-term variation in rainfall in tropical conditions, is of greater importance as a control on sediment supply and sedimentation (for example, Ziegler and others, 1987; Cecil, 1990 and references therein; Cecil and others 1993; Cecil and others, 2003a). The trip stops illustrate why climate change is a far more important first-order control on stratigraphy than is generally recognized.

Classification of Tropical Rainfall

Many definitions of tropical climatic regimes are based on annual rainfall; few, however, attempt to incorporate seasonality of rainfall (for example, Thornthwaite, 1948). It is becoming increasingly apparent, however, that it is the seasonality of annual rainfall that governs weathering, pedogenesis, variations in soil moisture, vegetative cover, and erosion and sediment yield for a given catchment basin (for example, Ziegler and others, 1987; Cecil, 1990, 2003; and Cecil and others, 2003a and references therein), not the total amount of annual rainfall. In order to assign degrees of seasonality, climate regimes, as used herein, are based on the number of wet months in a year. A wet month is defined as a month in which precipitation exceeds evapotranspiration (table 2, fig. 1F) (Cecil, 2003).

By the limits set forth in table 2 and figure 1F, both arid and perhumid conditions are nonseasonal. All other rainfall conditions have some degree of seasonality. Maximum sea-

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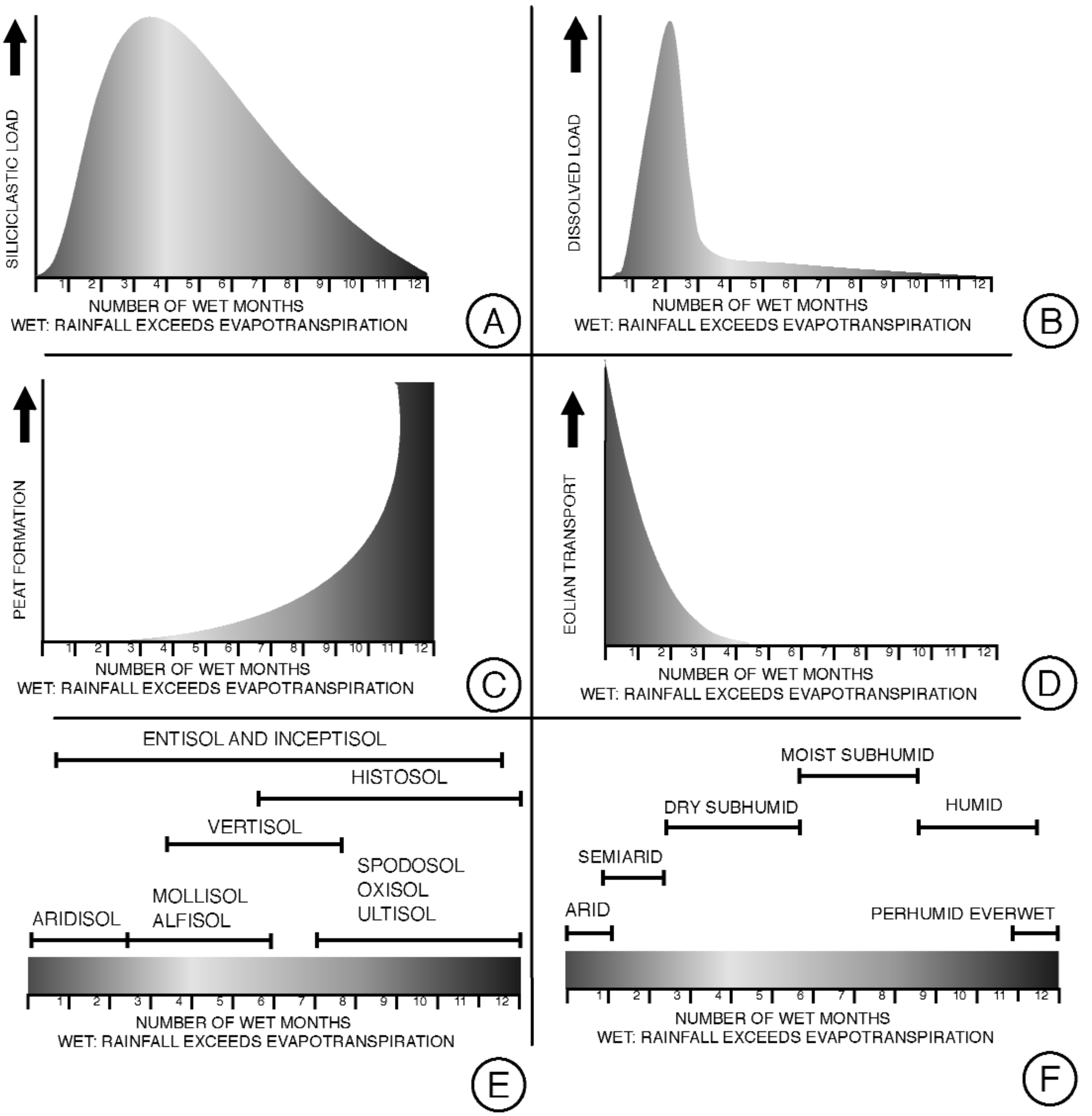


Figure 1. A, Potential for fluvial siliciclastic load as function of climate; B, Potential for fluvial dissolved load as a function of climate; C, Potential for peat formation as a function of climate; D, Potential for eolian transport as a function of climate; E, Formation of U.S. Department of Agriculture soil orders as a function of climate; and F, Climate classification based on the number of wet months per year. From Cecil and Dulong (2003).

sonality and maximum fluvial siliciclastic sediment supply occur under dry subhumid conditions when there are approximately three to five consecutive wet months (fig. 1A). Maximum dissolved inorganic supply occurs under semiarid to dry subhumid conditions (fig. 1B). The most ideal conditions for the formation and preservation of peat, as a precursor to coal, occur under humid and perhumid conditions (fig. 1C) when both dissolved and siliciclastic sediment supply approach zero (Cecil and others, 1993). The conditions for eolian transport in sand seas are illustrated in figure 1D.

Climate is also a major control on weathering and soil formation (pedogenesis); conversely, paleosols can be used to reconstruct paleoclimate and paleo water-tables. Climate may be the primary control on pedogenesis in epicontinental

Table 1. Tropical and subtropical climate change classification (modified from Cecil, 1990, 2003).

Relative duration	Cause	Time (years)
Long term	Continental drift; orogenesis; Ice house/ greenhouse	10^6–10^8 10^5–10^7
Intermediate term	100- and 400-ka cycles of orbital eccentricity; glacial/interglacial	10^5
Short term	Cycles in orbital obliquity and precession	10^4
Very short term (Millennial)	Uncertain	10^3
Instantaneous	Weather	10^{-2} (weeks, days, or hours)

Table 2. Tropical climate regimes and degree of seasonality based on the number of consecutive wet months per year (modified from Thornthwaite, 1948, and Cecil, 2003).

Average number of wet months	Climate regime	Degree of seasonality
0	Arid	Nonseasonal
1–2	Semiarid	Minimal
3–5	Dry subhumid	Maximum
6–8	Moist subhumid	Medial
9–11	Humid	Minimal
12	Perhumid	Nonseasonal

basins where other parameters, such as parent material and relief, are relatively constant. On the basis of structure, chemistry, and mineralogy, paleosols can be classified at the level of soil orders using the U.S. Department of Agriculture classification (Soil Survey Staff, 1975). Once classified, paleosols can be used to interpret paleoclimate (fig. 1E), including the amount and seasonality of rainfall as is illustrated at a number of stops on the trip.

Mechanisms of Climate Change

Continental Drift

A major component in long-term global climate variation, as expressed in Appalachian basin sedimentation, was the movement of the continents through paleolatitudes. The region of what is now the central Appalachian basin moved northward from about lat 40º S. in the latest Precambrian and Early Cambrian to lat 30º S. ± 5º during the Early Ordovician where it remained well into the Mississippian (Scotese, 1998). Northward movement continued from about lat 30º S. in the Early Mississippian to about lat 3º N. by the beginning of the Permian (figs. 2–9). From the perspective of zonal atmospheric circulation, the field trip study area moved from the dry subhumid belt of the southern hemisphere (prevailing easterlies) in the Early Cambrian into the high pressure belt of aridity by Late Cambrian where it remained well into the Mississippian. Late Precambrian and Early Cambrian sediments are dominated by siliciclastics, whereas Middle Cambrian through Early Devonian strata contain abundant limestone, dolomite, and evaporites. These strata are totally consistent with paleogeographic interpretations (for example, Scotese, 1998). By the Late Devonian the region began to move northward toward the humid low-pressure equatorial region. Movement continued through the equatorial region during the Pennsylvanian.

Orbital Climate Forcing

There are clear and distinct effects of orbital parameters (obliquity, eccentricity, and precession) on intermediate- to short-term climate change. Orbital climate forcing is an important control on variation in sediment supply because orbital parameters have a marked effect on the seasonality of incoming solar radiation (insolation), which results in variations in the seasonality of rainfall. As a result, orbital climate forcing plays a major role in sedimentation and stratigraphy. The lithostratigraphic signal induced by intermediate- to short-term climate change may be somewhat suppressed during periods of long-term aridity or humidity, but is not obliterated as is illustrated at stops in Ordovician, Silurian, and Pennsylvanian strata.

Tectonics

As pointed out above, continental drift may explain some of the long-term climate changes in the Appalachian basin, but it does not explain all such changes. Some long-term climate changes may be better explained by the alteration of atmospheric zonal circulation as a result of mountain building. Such effects are well known, as illustrated by the development of the Asian monsoon in response to the formation of the Himalayan Mountains and the Tibetan Plateau, when the Indian subcontinent collided with Asia. Similar tectonic controls on late Paleozoic paleoclimate have been suggested for the development of a humid climate in the Early Pennsylvanian by the formation of an equatorial high plateau, which pinned the intertropical convergence zone to the plateau causing a permanent low-pressure cell and high rain-

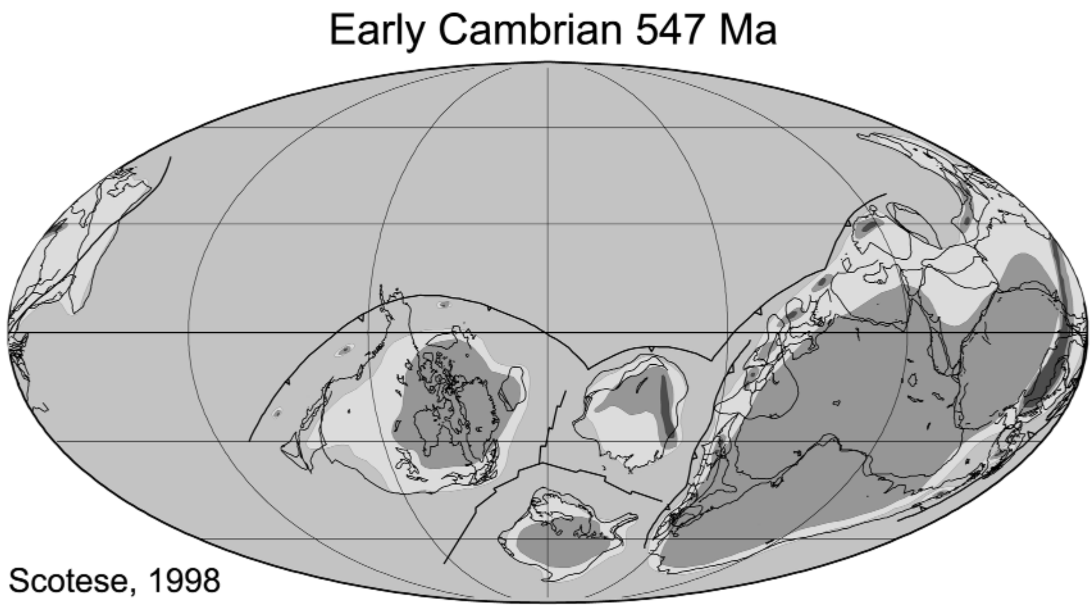


Figure 2. Paleogeographic reconstruction showing how the continents might have appeared during the Early Cambrian (547 Ma).

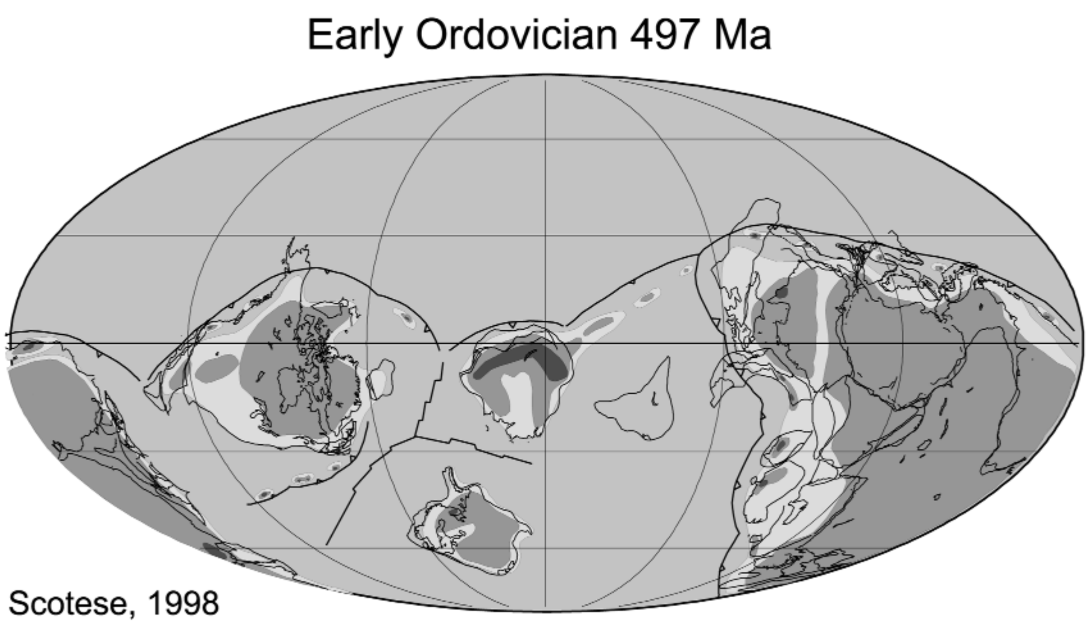


Figure 3. Paleogeographic reconstruction showing how the continents might have appeared during the Early Ordovician (497 Ma).

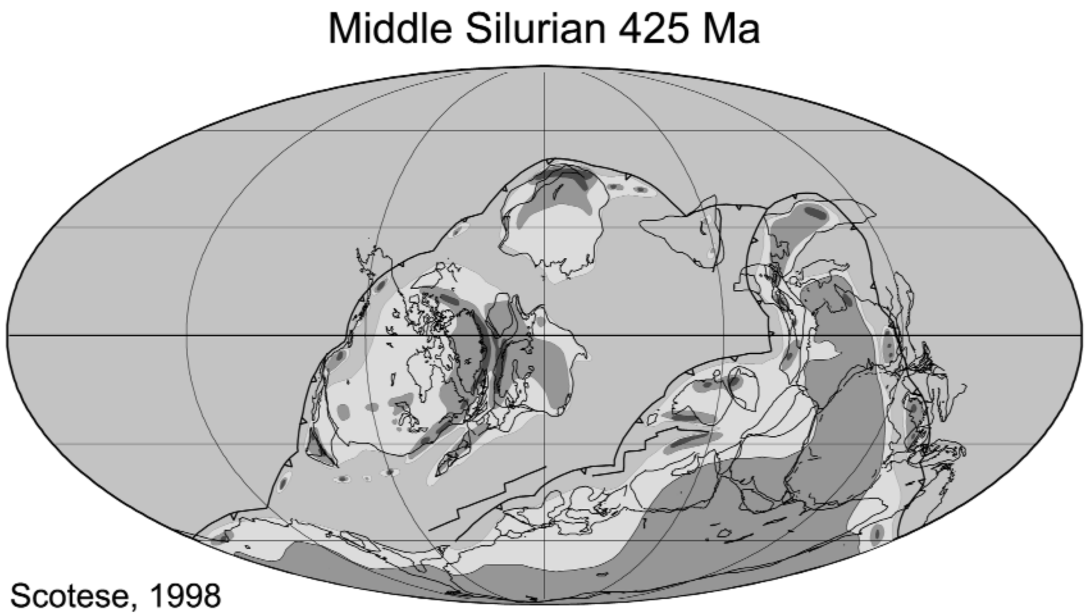


Figure 4. Paleogeographic reconstruction showing how the continents might have appeared during the Middle Silurian (425 Ma).

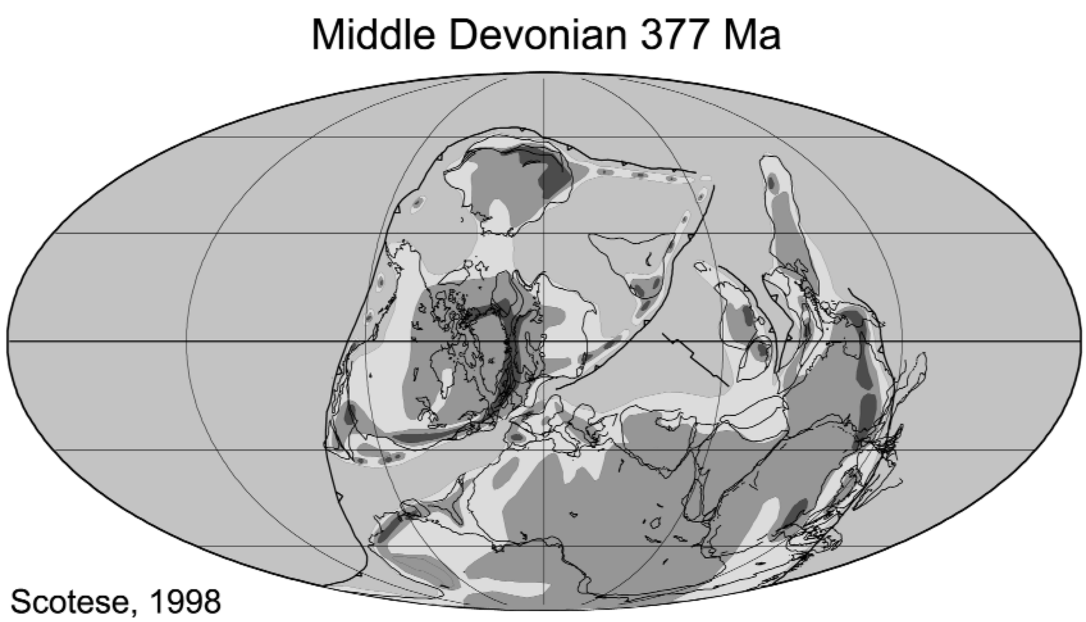


Figure 5. Paleogeographic reconstruction showing how the continents might have appeared during the Middle Devonian (377 Ma).

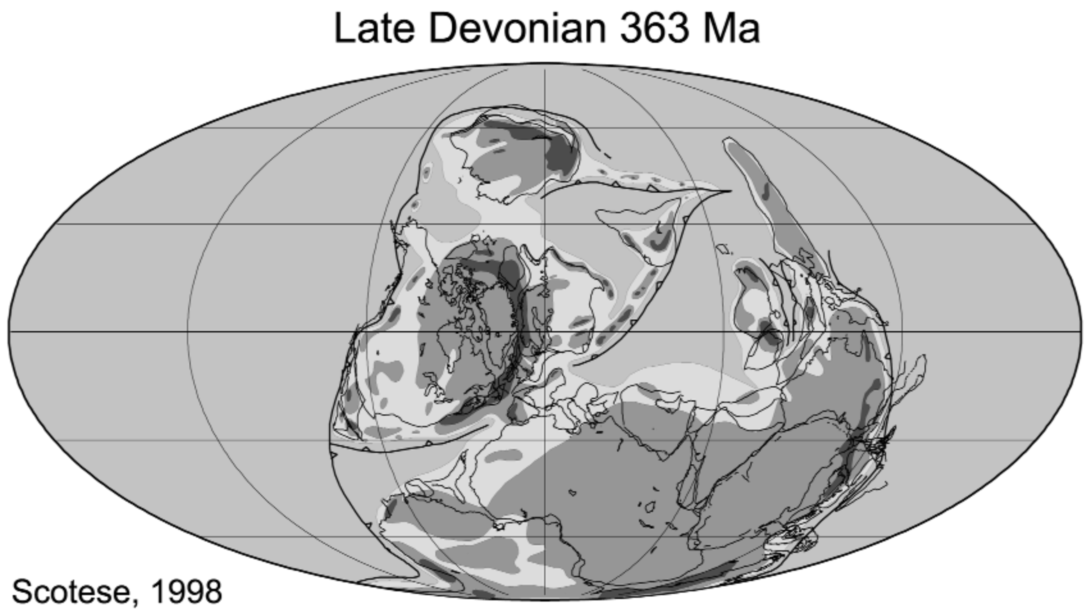


Figure 6. Paleogeographic reconstruction showing how the continents might have appeared during the Late Devonian (363 Ma).

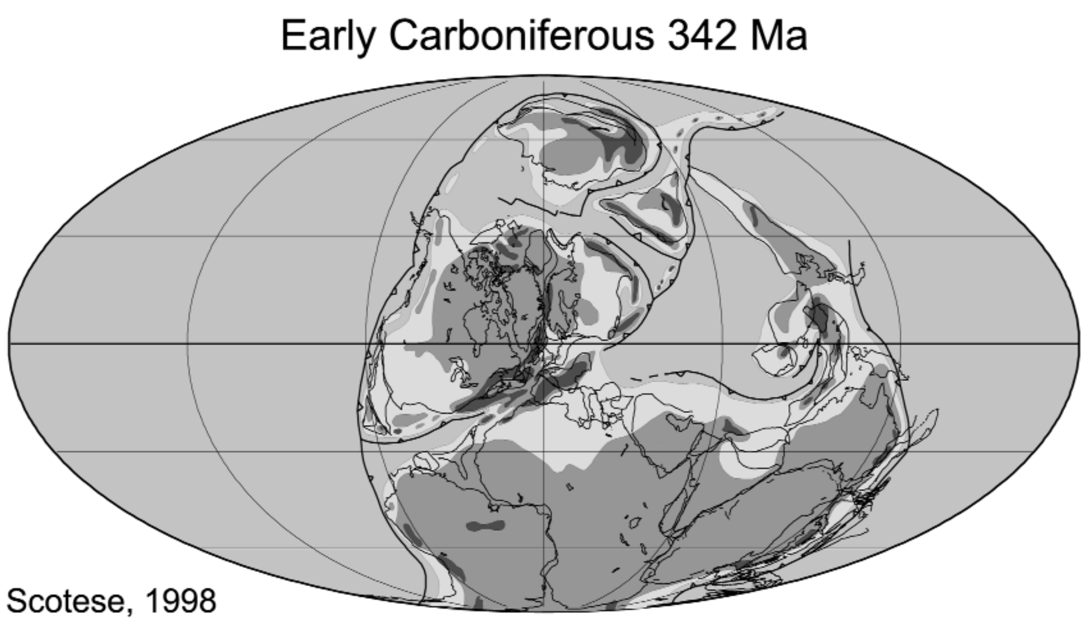


Figure 7. Paleogeographic reconstruction showing how the continents might have appeared during the Early Carboniferous (342 Ma).

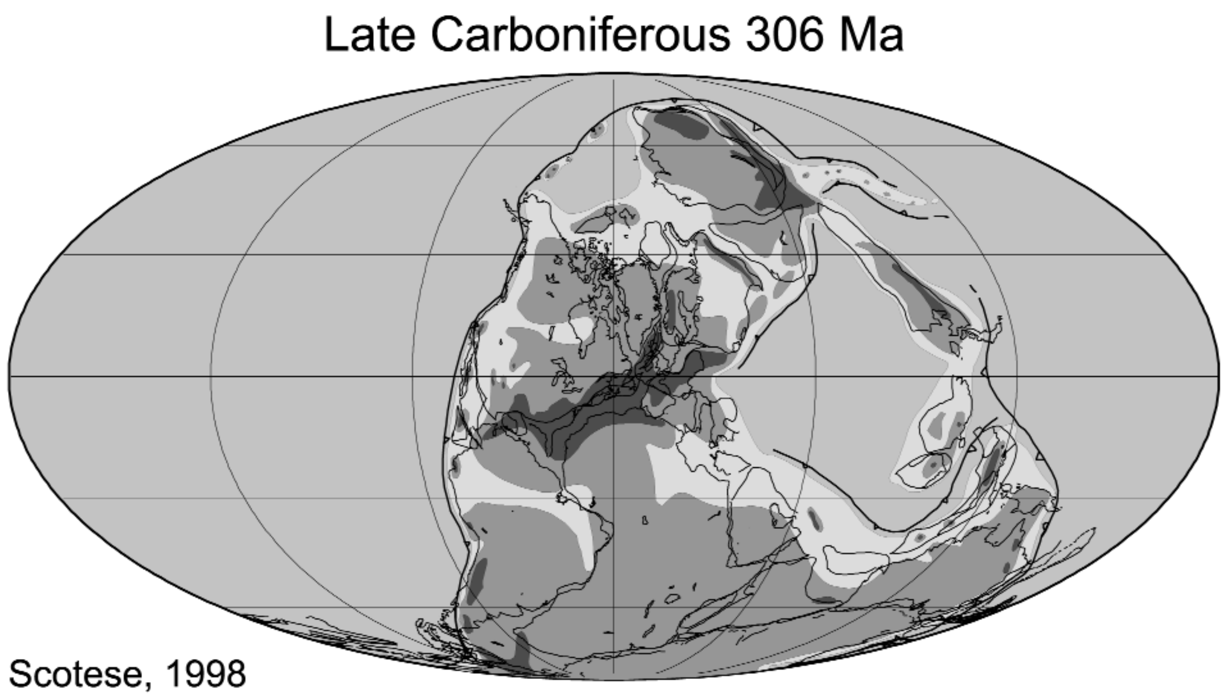


Figure 8. Paleogeographic reconstruction showing how the continents might have appeared during the Late Carboniferous (306 Ma).

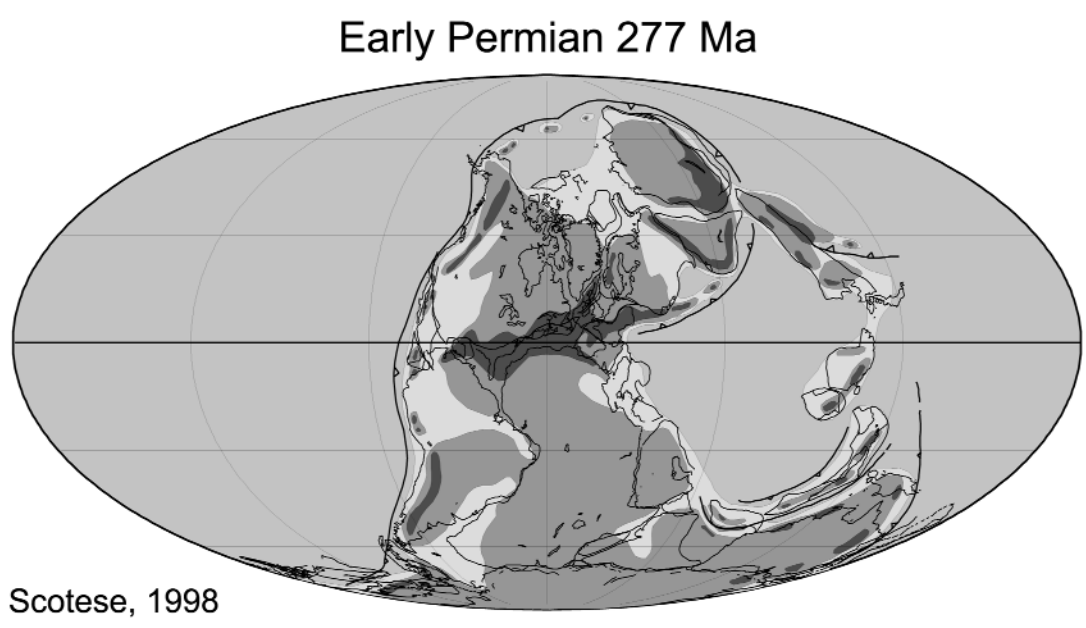


Figure 9. Paleogeographic reconstruction showing how the continents might have appeared during the Early Permian (277 Ma).

fall (Rowley and others, 1985). Such a permanent low-pressure cell could have been modulated by orbital forcing parameters and southern hemisphere glacial conditions (Cecil, 1990; Cecil and others, 2003b). The Taconic and Acadian orogenies also may have affected long-term climate change from regional orographic to global scales, although such effects have not been documented nor are they well understood. Although difficult to quantify, paleo-ocean circulation is yet another control on long-term climate change.

Unknown Mechanisms

Perhaps the most enigmatic, yet most important of all climate change processes are those that control long-term ice house and greenhouse conditions and the shorter term glacial-interglacial cycles. The geologic record is replete with evidence indicating that such extreme changes in paleoclimate have been occurring throughout geologic time (fig. 10) (Frakes and others, 1992). These climatic changes are sometimes accompanied by major biotic events, such as the Cambrian explosion of marine organisms following the "snowball earth" condition at the end of the Precambrian or the mass extinction that was coeval with a relatively short ice house world during the Devonian-Mississippian transition. Some of the enigmas and uncertainties associated with both near-time and deep-time ice-related global climate change are as follows:

1. Weather patterns, atmospheric composition data, and numerical models have led many researchers to conclude that the perceived present-day global warming is induced by greenhouse gases of anthropogenic origin. It is evident, however, that global climate commenced warming (although in "fits and spurts") following the last glacial maximum (18 ka), long before anthropogenic effects were significant.
2. The mechanisms that have controlled climate oscillations since the last glacial maximum are very poorly understood, and there is an equally poor understanding of the factors that control glacial-interglacial cycles and the longer term "ice house" to "greenhouse" transitions.
3. Among the multiple hypotheses regarding ice-related climate-forcing mechanisms (for example, orbital forcing, atmospheric composition (greenhouse gases), ocean circulation, bolide impacts, volcanism, solar variation, and so on), none document clear and unequivocal triggers of abrupt climate change, even when "feedbacks" are considered.

Even though the factors that trigger changes in ice volume remain unknown, there is a clear empirical correlation between ice volumes during the Paleozoic and paleoclimates (rainfall) in the Appalachian basin (fig. 10). The sense of paleoclimate change (increasing or decreasing rainfall) in the Appalachian basin that was coincident with changes in ice volume also is related to paleolatitudes. As is pointed out on the trip, paleoclimate change is best explained by estimates of both paleolatitudes and ice volume.

Summary

This trip investigates evidence for Paleozoic global climate change in the Appalachian basin. The objectives of individual trip stops are as follows:

1. Present stops in a global paleogeographic context.
2. Provide interpretations that relate stratigraphy and sedimentation to short-term and long-term global climate changes (table 1).
3. Interpret the seasonality of annual rainfall (fig. 1A, table 2) as the predominant control on sediment supply.
4. Empirically correlate changes in paleoclimate and sea level to changes in ice volume. Figure 10 depicts long-term climate change for the Paleozoic of the central Appalachian basin along with long-term changes in ice volume.

Field Trip Stops

Stop 1. Upper Conemaugh and lower Monogahela Group strata on the north side of the Morgantown Mall complex on Interstate 79 at Exit 152, Morgantown, W. Va.

Lat 39º37.82′ N., long 80º00.03′ W., Morgantown North, W. Va., 7.5-minute quadrangle.

Leaders: Nick Fedorko, Bill Grady, Cortland Eble, and Blaine Cecil

Introduction

The strata at Stop 1 gently dip to the northwest. Earlier cuts made for the construction of the Morgantown Mall and upper commercial area exposed approximately 18 m (meters) (59 ft (feet)) of upper Conemaugh Group strata and 32 m (105 ft) of lower Monongahela Group strata (fig. 11). Prior to removal by later construction, the section included five coal beds (some multi-benched) and an abundance of nonmarine, lacustrine limestone beds. The lowermost coal bed is the Little Pittsburgh coal of the Casselman Formation in the upper part of the Conemaugh Group. Several benches of the Little Pittsburgh coal bed interbedded with shales, mudstones, and lacustrine carbonates are exposed at the updip eastern end

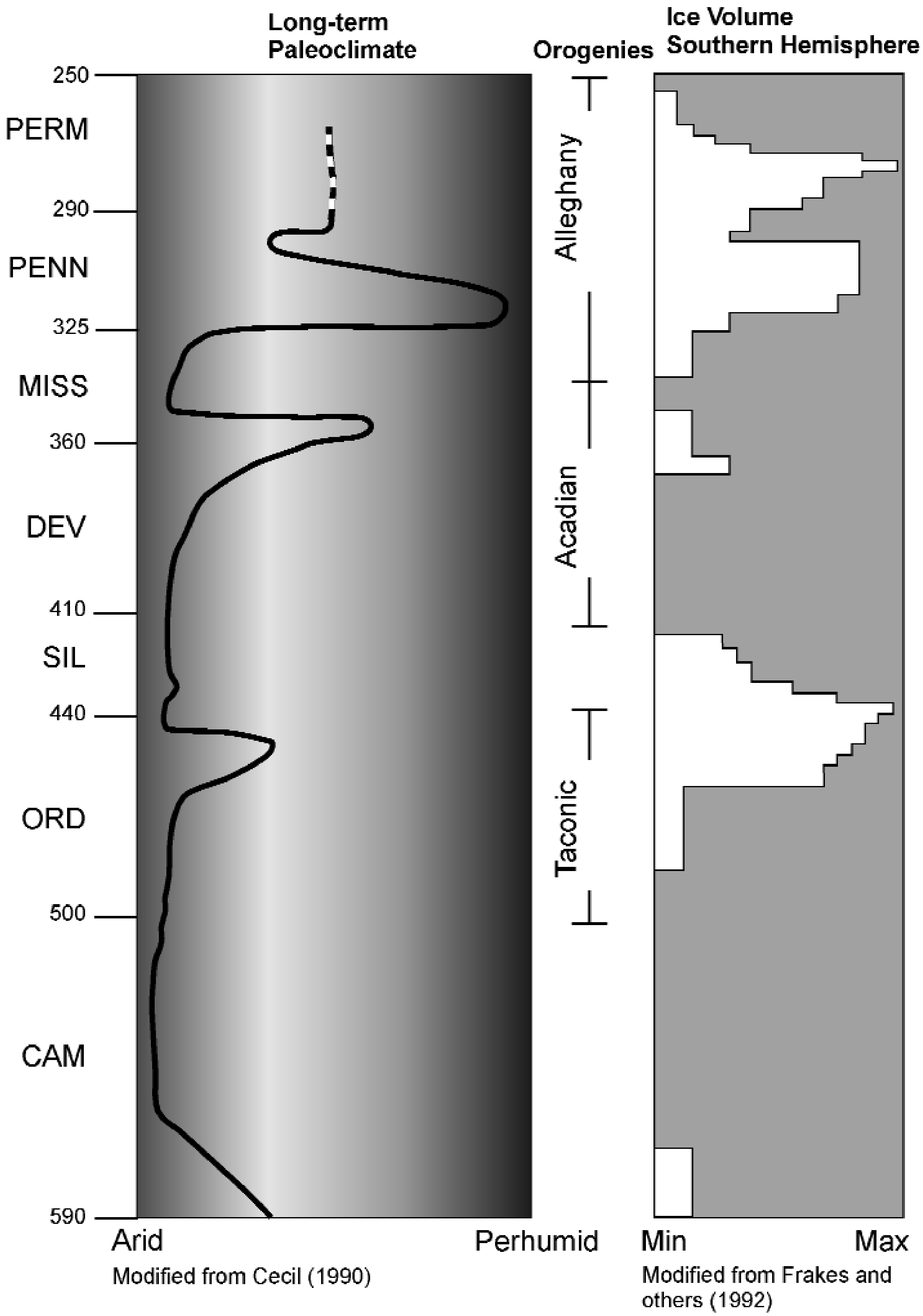


Figure 10. Paleozoic paleoclimate curve, central Appalachian basin. Series names are abbreviated as follows: CAM, Cambrian; ORD, Ordovician; SIL, Silurian; DEV, Devonian; MISS, Mississippian; PENN, Pennsylvanian; PERM, Permian.

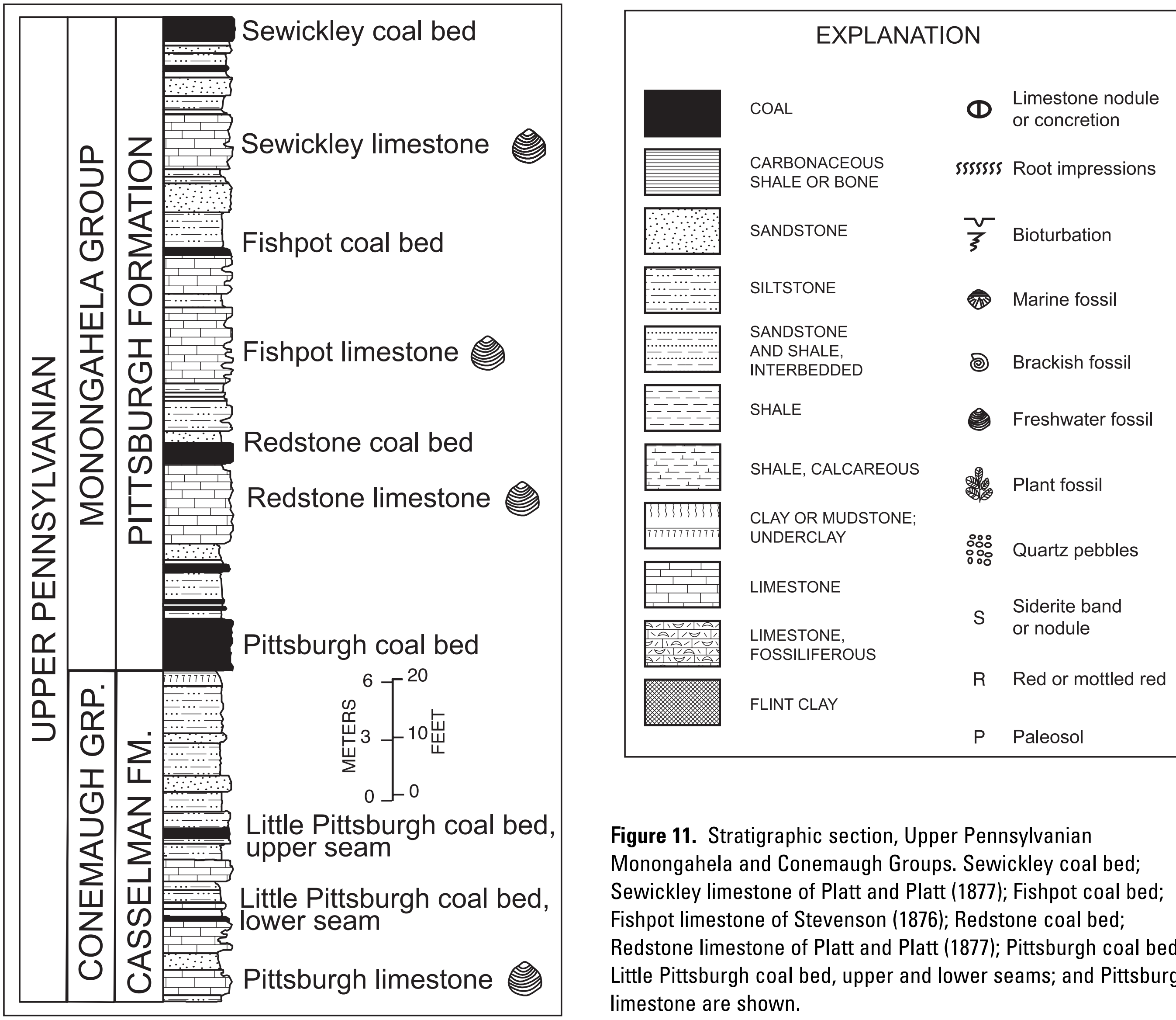


Figure 11. Stratigraphic section, Upper Pennsylvanian Monongahela and Conemaugh Groups. Sewickley coal bed; Sewickley limestone of Platt and Platt (1877); Fishpot coal bed; Fishpot limestone of Stevenson (1876); Redstone coal bed; Redstone limestone of Platt and Platt (1877); Pittsburgh coal bed; Little Pittsburgh coal bed, upper and lower seams; and Pittsburgh limestone are shown.

of the outcrop. The Little Pittsburgh coal bed is of minor economic importance but is persistent enough to serve as an important regional stratigraphic marker.

The Pittsburgh coal bed, the basal unit of the Pittsburgh Formation of the Monongahela Group, is exposed at the western end of the outcrop. Important to our discussions of climatic impact on the rock record is the development of the soil profile beneath the Pittsburgh coal bed. Here, where the coal facies are well developed, the subjacent soil profile is poorly developed, thin, and contains some carbonate in the form of small lenses. The massive nature of the mineral paleosol and the absence of subaerial exposure features indicate that the paleosol was permanently waterlogged or nearly so. A correlation of paleosol profiles along a 60-mi (mile) (97 km (kilometer)) transect from this stop southward elucidated the effects of paleotopography and paleoclimate on the genesis of mineral paleosols and coal. Southward, the paleosol becomes thicker and better developed, and it contains features that are indicative of subaerial exposure and well-drained conditions of pedogenesis. In contrast, the overlying coal facies thin and are poorly developed or absent. The characteristics of the mineral paleosols and the inverse relation between coal bed and mineral paleosol thickness suggests that Stop 1 was topographically lower relative to areas 60 mi (97 km) to the south.

The main benches of the Pittsburgh coal bed exposed in the Morgantown Mall have a combined thickness of 2.6 m (8.5 ft), and they are generally low in ash yield and moderate in sulfur content. Including roof shales and rider coal beds, the Pittsburgh is 5.2 m (17.1 ft) thick. At this location the coal consists of six benches, two more than are present 6 mi (10 km) to the northeast where the coal was extensively studied in a surface mine at the Greer estate. The basal bench (lower 0.3 m; 1 ft) is present across most of the areal extent of the Pittsburgh coal. It is high in sulfur and has a moderate ash yield. Tree fern spores dominate the palynoflora, but there are distinct calamite and cordaite contributions. The palynoflora of the lower bench is interpreted as the pioneering plant community of the Pittsburgh swamp. The ash yield and sulfur content indicate that these plants grew in a planar swamp with a significant influx of surface and ground water. Peat oxidation was minor, and the preservation of plant debris was moderate. Above the basal bench, up to the parting at the 1.2- to 1.4-m (3.9- to 5.0-ft) level, the coal is low ash, has moderate sulfur content, and petrographically shows two trends in swamp development. These trends of increased peat preservation, as shown by increased >50-micron (μ) vitrinite components, are reflected in the sulfur content and palynofloral succession, but not in ash yield. The first trend, terminated by a fusain parting, displays an upward increase in vitrinite content, especially the >50-μ component, an increase in calamite and arboreous lycopsid spores, and increased sulfur content. The bone coal parting at 4 ft (1.2 m) above the base of the bed terminates a second similar trend. Increased vitrinite and >50-μ vitrinite, arboreous lycopsid and calamite spores suggest a slight increase in surface water depth as peat accumulation proceeded. The increased sulfur content probably represents increased introduction of sulfur into the swamp by surface or ground water as water depth increased. The termination of these trends by fire followed by sediment deposition demonstrates a rapid and significant change in the water table. The fusain parting changes laterally into a bone coal parting and is present sporadically throughout the areal extent of the Pittsburgh coal. The extremely widespread 4-ft (1.2-m) parting (Cross, 1952) occurs at approximately the same stratigraphic level throughout the lateral extent of the Pittsburgh coal bed. The vast lateral extent of the parting suggests a regional rise in paleo-water levels and drowning of the peat swamp flora. Such a rise in water levels may have been driven by a protracted period of increased rainfall in response to changes in orbital parameters.

The Redstone limestone (*sensu* Platt and Platt, 1877), a well-developed nonmarine, regionally extensive, lacustrine carbonate occurs above the Pittsburgh coal bed at this stop. In most other localities, however, the Redstone limestone is separated from the coal by shale and (or) sandstone. The Redstone limestone generally occurs as a monolithic micrite, the result of deposition in an areally extensive lake that underwent periodic drying and subaerial exposure, as evidenced by pedogenic brecciation and the formation of subaerial crusts. Lacustrine limestones such as this one first occur in the Middle Pennsylvanian Allegheny Formation but are most abundant in the Monongahela Group in the region encompassing northern West Virginia, southwest Pennsylvanian, and eastern Ohio. These carbonates are exclusively micrites, occurring in complexes interbedded with argillaceous limestones, calcareous mudstones, and calcareous and noncalcareous shales.

The Fishpot coal bed of the Pittsburgh Formation (Monongahela Group), which has now been removed from this site by construction, was only 2.5 cm (centimeters) (1 in (inch)) thick in this section and, with few exceptions, rarely exceeds 0.6 m (2 ft) in thickness. However, thin coal or carbonaceous shale can be found at this stratigraphic position at widely separated points throughout the Dunkard basin. Persistent occurrences of beds, such as the Fishpot coal bed, are indicative of the allocyclic processes that controlled sedimentation and stratigraphy.

The thickest sandstone in this section occurs in a clastic interval above the Fishpot coal bed. It is tabular, varying in thickness from 0.9 to 2.3 m (3.0–7.5 ft). Another clastic interval occurs above the Sewickley limestone (*sensu* Platt and Platt, 1877). Thin sandstone and shale beds are interbedded with the Sewickley coal bed here. The association of the Sewickley coal bed with fine- to coarse-grained clastic strata is characteristic basinwide. Lower coal benches or “splits” of the Sewickley are sometimes miscorrelated with the underlying Fishpot coal bed. The main mineable bench of the Sewickley coal bed is not well exposed at this location. Weathered coal (blossom), 1.2 m (3.9 ft) thick, can be seen at the top of the section at the extreme western end of the cut. This bed also has been mined extensively underground in the Morgantown area.

Paleoclimate and Sea-Level History

Deposition of the strata at Stop 1 was coincident with the onset of a long-term humid interval that began in the latest Pennsylvanian (fig. 10) (Cecil, 1990). At approximately the same time, global ice house conditions began in the latest Pennsylvanian and culminated in the Permian (Frakes and others, 1992). Although deposition of latest Pennsylvanian and Early Permian(?) strata in the Appalachian basin occurred during long-term humid conditions coincident with ice in high latitudes, short-term to intermediate-term climate cyclicity was the predominant control on the stratigraphy of sedimentary cycles, such as those illustrated at Stop 1 (Cecil, 1990). For example, the paleosol underlying the Pittsburgh coal bed (underclay) at Stop 1 is typical of a soil that was permanently waterlogged or nearly so (hydromorphic soil). The paleosol transect noted above delineates a regional toposequence of paleosols that are indicative of humid paleoclimate. The transect demonstrates that the paleosol at Stop 1 formed in a waterlogged topographic low, while coeval soils to the south were forming under better drained conditions in upland areas. The toposequence of mineral paleosols that unconformably underlies the Pittsburgh coal bed is typical of modern tropical soils that are acidic and highly leached, and that have high base-exchange capacity and high soil moisture. These types of soils predominantly form under humid climate conditions (see table 2) where soil pore waters are exceedingly low in dissolved solids in response to soil leaching induced by high rainfall with little seasonality. An increase in rainfall triggered the onset of permanent swamp conditions and peat formation subsequent to mineral paleosol formation.

Allocyclic factors in addition to climate doubtless contributed to the vast areal distribution of the Pittsburgh coal

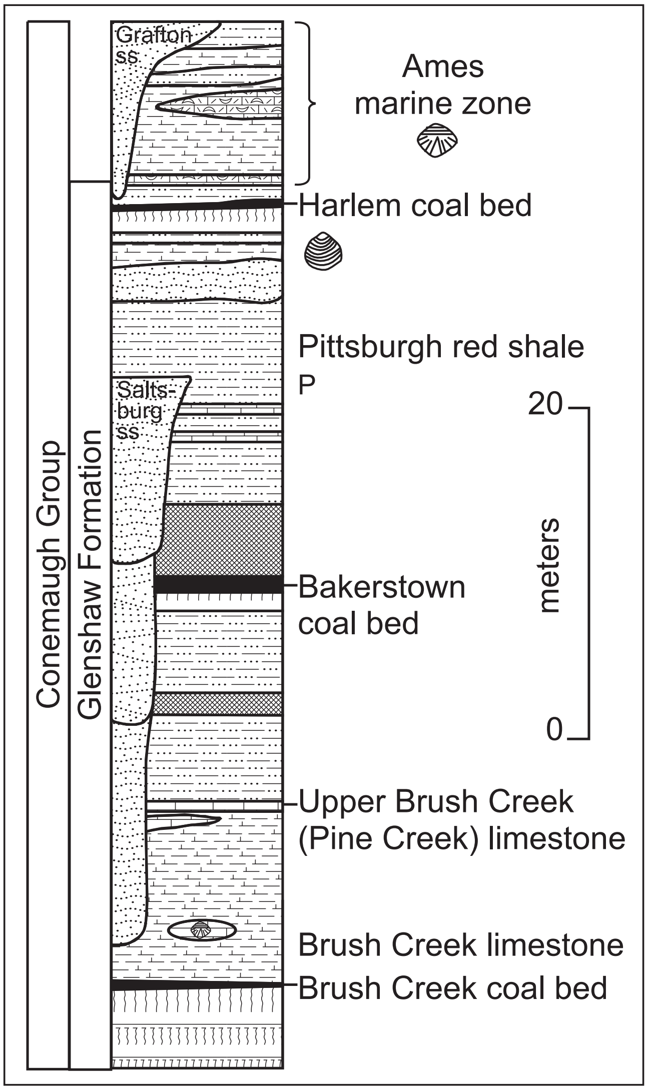


Figure 12. Stratigraphic section, Upper Pennsylvanian Conemaugh Group. Grafton sandstone of White (1903); Ames marine zone, Harlem coal bed; Pittsburgh red shale of White (1903); Saltsburg sandstone of Stevenson (1876); Bakerstown coal bed; Brush Creek marine zone; and Brush Creek coal bed are shown. See figure 11 for explanation of lithology symbols.

bed. However, the thickness and quality of the coal appear to be strongly climate controlled. Following mineral paleosol formation, the paleoclimate became wet enough to flood vast flat-lying areas and initiate swamp development. Water levels within the swamp were maintained primarily by rainfall along with an influx of surface water from around the margins of the swamp. During the initial stages of peat formation, annual rainfall, augmented with surface water flow, was sufficient to allow the development of a large planar swamp. During the later stages of peat development, the influential effects of rainwater versus surface-ground water on peat composition, which, in turn, influenced ash yield, sulfur content, and maceral composition, varied with location and time. Ash yield and sulfur content, as well as the degree of degradation of the peat plant debris, were greater to the west of the Morgantown area, perhaps because of more frequent and extensive incursions of fresh surface and ground water into the peat swamp. To the east, in western Maryland, the Pittsburgh coal bed is thicker, lower in ash yield and sulfur content than in the Morgantown area, and appears to have been, except for the basal high-ash and high-sulfur bench, more influenced by rainfall. Sixty miles (~100 km) to the south, in south-central West Virginia, the Pittsburgh coal bed is thin, or completely absent, as a result of paleotopographic controls on the development of the swamp and contemporaneous paleosols.

The ash yield and ash composition of the coal suggests that the peat was moderately acidic and that dissolved solids and clastic influx were mostly nil. These conditions are consistent with modern peat swamp precursors of commercial-grade coal. Such modern swamps tend to be acidic, low in dissolved solids, low in nutrients (oligotrophic), and essentially devoid of any clastic influx.

Following deposition of peat and overlying clastics, the occurrence of the Redstone limestone strongly suggests a significant change in sedimentary geochemistry. Subaerial exposure features within the Redstone limestone are indicative of repeated and persistent climate drying. In sharp contrast to the low dissolved solids and acidic water chemistry associated with the formation of the mineral paleosol and peat, chemical conditions for limestone deposition require alkaline waters that were saturated with dissolved solids.

Any relation between sea-level fluctuation and latest Pennsylvanian sedimentation in the Appalachian basin remains equivocal. Sediments of marine origin have not been documented in the Monongahela Group, even though cyclic sedimentation (analogous to the Virgilian cyclothems in the Midcontinent) occurs throughout the group. These cycles are interpreted herein as fourth-order sequences with sequence boundaries defined by the unconformities at the base of the coal beds. The periodic rise and fall of water levels in the Dunkard basin, therefore, may have been controlled by cyclic variations in both the amount and seasonality of rainfall that were associated with the well-known glacial eustatic cycles in the Midcontinent.

Stop 2. Paleosols in the Pittsburgh red shale, Conemaugh Group on Interstate 79 at Exit 146.

Lat 39°32′25″ N., long 79°59′23″ W., Morgantown South, W. Va., 7.5-minute quadrangle.

Leaders: Nick Fedorko, Blaine Cecil, and Rob Stamm

Introduction

Upper Pennsylvanian strata assigned to the Conemaugh Group are exposed in a roadcut along I-79 adjacent to the

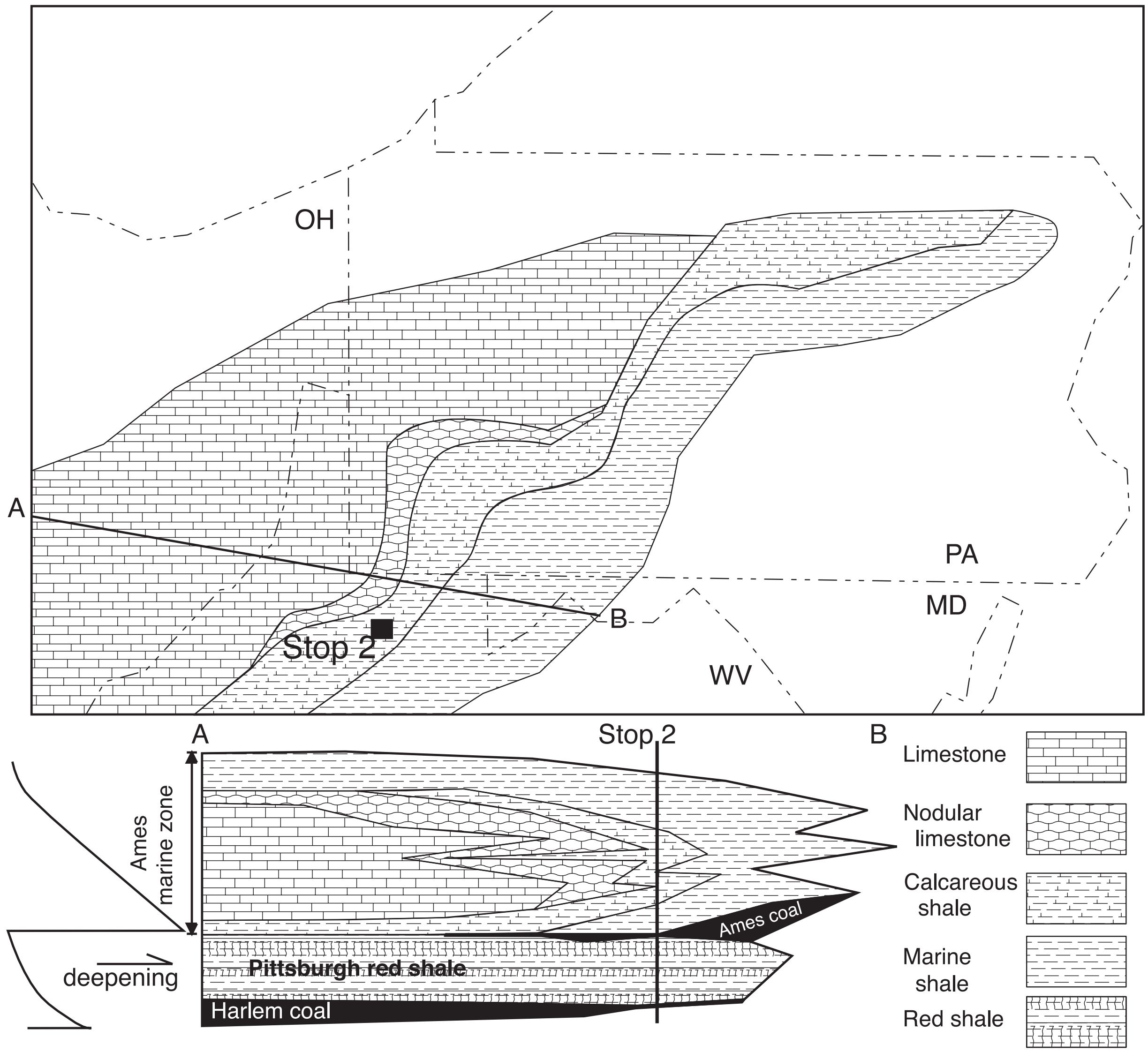


Figure 13. Lithofacies distribution of the Ames marine zone in the central Appalachian basin. Cross sectional interpretation (bottom) of fifth-order deepening episodes with corresponding lithotopes (modified from Brezinski, 1983).

Goshen Road exit (Exit 146) in Monongalia County, West Virginia. The cut exposes approximately 34 m (~110 ft) at, and just above road level, including the following units, in descending order: the Grafton sandstone, Ames marine zone, Harlem coal bed, and Pittsburgh red shale, all of the Glenshaw Formation. This stratigraphy is depicted in the upper half of figure 12.

Lithostratigraphy

The Grafton sandstone (*sensu* White, 1903), exposed at the top of the section, consists of interbedded siltstone and sandstone, and elsewhere is overlain by the Casselman Formation. Underlying the Grafton is the Ames marine zone of the Glenshaw Formation. It represents the last known major marine transgression and inundation in the central Appalachian basin. As such, the Ames serves as an important unit for lithostratigraphic correlation across the basin. The Ames is an impure, shaley, fossiliferous limestone to calcareous shale in the area immediately surrounding the area of Stop 2, but grades westward into a greenish-gray, highly fosssiliferous limestone in eastern Ohio (Brezinski, 1983) (fig. 13).

The coal bed directly beneath the Ames marine zone is termed the Harlem coal bed by the West Virginia Geological Survey (fig. 13). In Ohio, there are two coal beds beneath the Ames marine zone; the coal bed immediately underlying the Ames marine zone is the Ames coal and the second is the Harlem coal. This guidebook will continue to follow the terminology of the West Virginia Geological Survey. The

Harlem coal bed is a thin (generally less than 0.6 m (2.0 ft) thick), laterally persistent unit that occasionally attains mineable thickness. Compositionally, the Harlem coal bed generally contains high percentages of vitrinite group macerals, and low to moderate amounts of liptinite and inertinite group macerals. Ash yields and sulfur content commonly are moderate to high. Other Conemaugh Group coal beds of regional extent include the Mahoning, Bakerstown, Elk Lick, and Little Clarksburg.

The Pittsburgh red shale (*sensu* White, 1903) directly underlies the Harlem coal bed at this stop (fig. 12). Here the Pittsburgh red shale consists of alternating beds of impure limestone and variegated red-green claystone that often contain calcium carbonate nodules. This unit contains features indicative of repeated and (or) prolonged subaerial exposure and pedogenesis. These features include calcareous nodules and mukkarra structures (crosscutting slickensides) of pedogenic origin. In addition, midway between the base of the exposure and the Harlem coal bed there is a lateral break in soil structure that includes a channel-form structure indicating that the Pittsburgh red shale may be composed of two or more paleosols.

Depositional Environments

The abundant calcium carbonate in the paleosol section is indicative of repetitive prolonged dry seasons. A nonmarine limestone often occurs as discontinuous pods at the top of the Pittsburgh red shale. This limestone is of mixed lacustrine and pedogenic origin, and it occurs in topographic lows on the gilgai surface (paleosol relief) of the Pittsburgh red shale. Rainy periods lasting months to a few years resulted in lacustrine carbonate deposition. During drier climatic periods, these lakes dried up, resulting in pedogenesis of the lacustrine carbonates. The lateral persistence of the Pittsburgh red shale throughout the Dunkard basin indicates that pedogenesis occurred during a lowstand in sea level and that the climatic conditions were basinal in extent.

Unlike the underlying Pottsville Group and Allegheny Formation strata, the Conemaugh Group contains abundant calcareous red shales and mudstones. Regionally, these calcareous red sediments first appear in the section 30 to 60 m (100–200 ft) below the Ames marine zone and are present to varying degrees in the upper two thirds of the Conemaugh Group throughout its area of occurrence. The calcium carbonate content of Conemaugh Group strata is high as compared to Middle Pennsylvanian strata (Cecil and others, 1985). Calcium carbonate occurs as admixtures in marine intervals, as nonmarine lacustrine beds (for example, Clarksburg limestone of White (1891)), and as pedogenic nodules, discontinuous lenses, and admixture within mudstones and shales.

Although of regional extent, Conemaugh Group coal beds tend to be fewer, thinner, and more impure than those in the underlying Allegheny Formation or overlying Monongahela Group. For example, the Little Clarksburg coal bed rarely exceeds 0.6 m (2 ft) in thickness in this region. It is thicker and mineable in the Potomac Basin of eastern West Virginia and western Maryland, but is of poor quality, locally known there as the “Dirty Nine-foot.” The Elk Lick coal bed, exposed just above the first bench on I-68 (milepost 4.0) between Stops 2 and 3, also has been mined in the Potomac Basin in western Maryland and also in central West Virginia and is known as the Barton coal in western Maryland. The Elk Lick is 0.9 m (3 ft) thick along I-68 at milepost 4.0, but has not been commercially exploited locally. As a group, Conemaugh coal beds are higher in ash yield and sulfur content than the underlying Allegheny Formation coal beds, and comparable in sulfur content with the overlying Monongahela Group coal beds, but higher in ash yield. The Elk Lick coal bed exposed in the I-68 roadcut is 3 ft (0.9 m) thick and represents peat that accumulated during the long-term Conemaugh drier interval (see fig. 10) than the stratigraphically lower Upper Freeport or Mahoning coal beds. The Elk Lick coal was formed after the demise of the peat-swamp arboreous lycopsids at the Westphalian-Stephanian boundary. The Elk Lick coal bed is high in ash yield and high in sulfur and contains significantly greater inertinite, and less well-preserved (>50-µ) vitrinite than the Upper Freeport or coal beds lower in the Pennsylvanian. At the top of the I-68 section the Little Clarksburg coal is exposed, and the stratigraphically lower Harlem and West Milford coals are exposed in nearby outcrops. These coal beds are similar in ash yield, sulfur content, petrographic composition, and floral character to the Elk Lick and typify Conemaugh coal beds that apparently accumulated in planar swamps with significant surface and ground-water influx of minerals and dissolved solids in moderate pH waters. Sulfur emplacement, especially as pyrite, was extensive and coincided with severe degradation of the peat and loss of >50-µ vitrinite components, probably by anaerobic microbes. These attributes suggest a seasonal paleoclimate with insufficient annual rainfall to maintain a highly acidic ombrogenous swamp.

Paleoclimate and Sea-Level History

The strata at Stop 2 were deposited during a long-term dry interval that began in the middle Late Pennsylvanian and ended in the late Late Pennsylvanian (fig. 10) (Cecil, 1990). This long-term drier interval is coincident with the Late Pennsylvanian global greenhouse condition noted by Frakes and others (1992). In contrast to the humid climatic conditions of pedogenesis at Stop 1, the paleosols at Stop 2 are indicative of a paleoclimate that was dry subhumid to semiarid. A short- to intermediate-term increase in humidity (moist subhumid climate) must have been associated with peat formation (Harlem coal) during maximum lowstand (maximum ice) as suggested by Cecil and others (2003b). This stop illustrates both the long-term dry conditions associated with long-term greenhouse conditions and the short-term climate cycles associated with glacial-interglacial conditions.

The compound nature of the paleosols at this stop suggests at least two periods of deposition followed by exposure and pedogenesis. This interpretation is supported by a reconnaissance study (by R.G. Stamm) that identified conodonts within the paleosols. If further work unequivocally demonstrates the presence of conodonts within the paleosols, then some of the Missourian marine transgression deposits in the Appalachian basin were nearly obliterated by pedogenesis during subsequent glacioeustatic lowstands.

Stop 3. Late Middle Pennsylvanian Lower Freeport coal bed(?) and associated strata on Interstate 68 at milepost 11.4

Lat 39°42.5′ N., long 78°17.6′ W., Lake Lynn, Pa.-W. Va., 7.5-minute quadrangle.

Leaders: Blaine Cecil, Nick Fedorko, Frank Dulong, and Cortland Eble

Introduction

An accurate stratigraphic correlation of the coal beds at this stop remains equivocal. However, the palynoflora in these coal beds suggest that they are the Lower and Upper Freeport coal beds of the Allegheny Formation. The section at Stop 3, therefore, includes the stratigraphic interval from the Lower Freeport limestone up to the Upper Freeport coal bed of the Middle Pennsylvanian Allegheny Formation (Wilmarth, 1938) (fig. 14). The primary emphasis at this stop is the stratigraphic interval from the base of the Lower Freeport limestone to the top of the Lower Freeport coal bed.

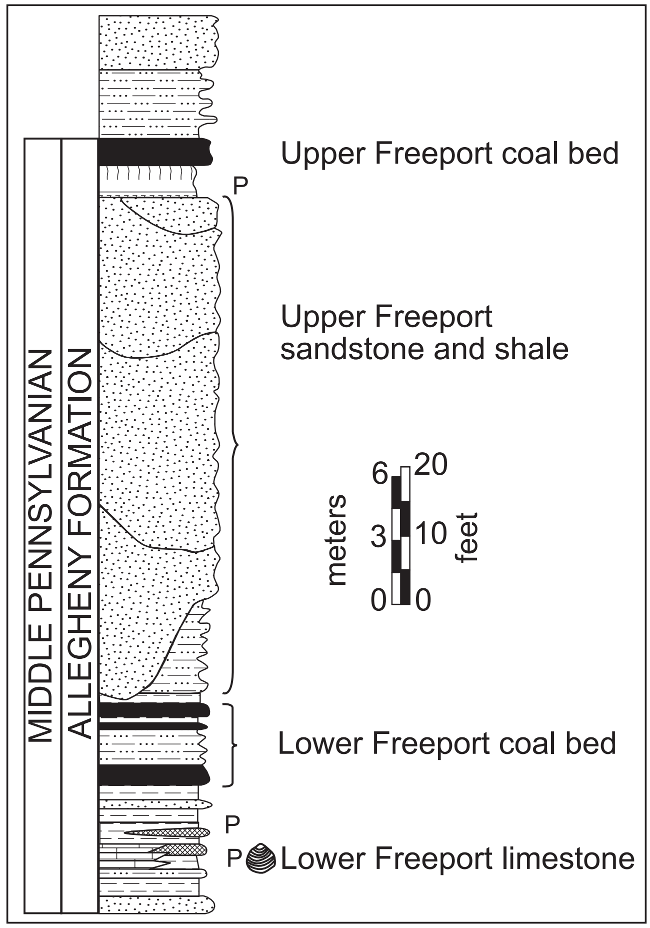


Figure 14. Stratigraphic section, Middle Pennsylvanian Allegheny Formation. Upper Freeport coal bed; Upper Freeport sandstone and shale (Butler sandstone of White, 1878); Lower Freeport coal bed; and Lower Freeport limestone of Platt and Platt (1877) are shown. See figure 11 for explanation of lithology symbols.

Lithostratigraphy

The Lower Freeport coal bed(?) of Platt and Platt (1877) crops out just above road level at Stop 3. The overlying stratigraphic succession (in ascending order) includes the Upper Freeport sandstone and shale (Butler sandstone of White, 1878), Upper Freeport limestone (absent at this locality) of Platt and Platt (1877), Upper Freeport fire clay (underclay of the Upper Freeport coal bed) of Stevenson (1878), and Upper Freeport coal bed of Platt and Platt (1877). At Stop 3, the clastic interval above the Lower Freeport coal bed is predominantly sandstone. The Upper Freeport fire clay (paleosol) and coal are exposed at the top of the roadcut on the south side of I-68.

Lower Freeport Coal

The Lower Freeport coal at Stop 3 consists of three splits over a 4.3-m (14.1-ft) interval. The main (lower) split is 0.9 m (3 ft) thick, is vitrinite-rich, especially the >50-μ vitrinite types, and is low in ash yield and sulfur. Palynologic results show a tree fern spore-dominant swamp palynoflora throughout, but with a moderate arboreal lycopsid spore contribution. At this location, development of the Lower Freeport swamp was not at its best (mineable) quality, but the lower and middle benches of this split probably represent paleoenvironments of the thicker coal to the north. Initial peat accumulation was in a planar swamp (Cecil and others, 1985) with minor emplacement of minerals and sulfur. Tree ferns dominate the flora and the pre-vitrinite plant debris was preserved moderately well, with little oxidation of the peat. With further peat accumulation (middle bench) the planar swamp may have become slightly elevated. Mineral and sulfur emplacement was low. Tree fern spores dominate the palynoflora, but increased arboreal lycopsid spore abundance may indicate a standing water cover.

The paleoclimate at the time of Lower Freeport peat accumulation probably was transitional between the perhumid Early and early Middle Pennsylvanian climate and the drier and more seasonal climate of the early Late Pennsylvanian. The paleoclimate was obviously wet enough to allow for the widespread development of the Lower Freeport coal, and for the accumulation of some low ash and sulfur, pre-vitrinite-rich peat. However, while the annual rainfall may have been insufficient to allow for extensive domed peat formation (in other words, to the extent inferred for many Lower and Middle Pennsylvanian swamps), it certainly was adequate to allow for the development of an extensive peat swamp that, in some areas, may have attained some elevation above the regional water table. With the buildup of the peat, oxidation became more frequent in a seasonal paleoclimate where insufficient seasonal rainfall did not allow for extensive doming during peat accumulation. Oxidation of the peat surface and an increase in inertinite abundance preceded the ultimate drowning of the Lower Freeport coal.

Stratigraphy at Stop 3

The top of the Upper Freeport coal bed (top of the roadcut) is defined as the top of the Allegheny Formation and the base of the overlying Conemaugh Group (Stevenson, 1873). Both the Lower and Upper Freeport coal bed horizons occur throughout the Appalachian basin in Pennsylvania, Maryland, West Virginia, and Ohio. Where they are sufficiently thick, these laterally extensive coal beds have been mined from the eastern outcrop belt in western Maryland and eastern Pennsylvania to the western outcrop belt in east-central Ohio, a distance of over 150 mi (250 km). Nonmarine strata, including underclay, sandstone, siltstone, shale, flint clay, and the Lower Freeport limestone underlie the Lower Freeport coal bed.

Above the top of the Lower Freeport coal bed at Stop 3 is the Upper Freeport sandstone and shale (Butler sandstone of White, 1878) and overlying underclay (paleosol) and Upper Freeport coal bed. The Upper Freeport coal bed is of particular significance in that it is the stratigraphically highest coal bed that contains abundant *Lycospora* sp., the dispersed spore of some of the giant lycopsid trees (for example, *Lepidophloios* and *Lepidodendron*) that dominated many Lower and Middle Pennsylvanian coal beds throughout Euramerica.

Paleoclimate and Sea-Level History

The interval from the base of the Lower Freeport limestone to the top of the Lower Freeport coal bed is interpreted to be the result of a complex set of conditions that occurred in response to a cyclic paleo water-table, sediment flux, pedogenesis, and paleoclimate (Cecil and others, 1985). The limestone is nonmarine and was probably deposited in large, shallow lakes, as indicated by multiple subaerial exposure features that include subaerial crusts, pedogenic brecciation, and residual pedogenic clay. Intermittent deposition and subaerial exposure of the limestone is indicative of a fluctuating lake level and water table. The frequency of water level fluctuation is unknown but may have been controlled by short-term or very short term variations in paleoclimate (table 1). Lake waters must have been alkaline, pH 7.8 or greater (Krumbein and Garrels, 1952), during deposition of the limestone. The alkalinity and high concentrations of dissolved solids in lake waters during deposition of the limestone was, in part, the result of a relatively dry paleoclimate that concentrated dissolved solids through evaporation (Cecil and others, 1985; Cecil, 1990). The Lower Freeport limestone was then buried by a thin (~1–2 m; 3–7 ft), but widespread, influx of siliciclastics. Subsequently, both were subjected to an extended period of subaerial exposure, weathering, and pedogenesis (Cecil and others, 1985). The weathering and resultant residual clay deposits imply a drop in the water table during the onset of increasing pluvial conditions. Further increases in rainfall led to increased vegetative cover, rainfall dilution of runoff, and leaching of residual soils, all of which reduced erosion and the influx of siliciclastic sediment and dissolved solids. Extensive leaching of the landscape, during the pluvial part of intermediate-term climate cycles, restricted the buffering capacity of surface water systems by reducing the concentration of dissolved solids. A rising water table with low buffering capacity led to acidic water (pH <6) from decaying vegetal matter. These conditions of a high water table and low pH are necessary for the formation of thick, laterally extensive, high-quality peat (Cecil and others, 1985).

On a regional scale, a complex of kaolin-enriched paleosols occurs in a facies mosaic at the base of all upper Middle Pennsylvanian coal beds, including the Lower and Upper Freeport. These paleosols appear to be the result of a fluctuating water table and weathering during humid parts of climatic cycles. The most intensely developed paleosols formed on well-drained paleotopographic highs. The kaolin-enriched deposits have been mined locally and used in the manufacture of refractory brick. The unconformity at the top of the paleosols and the base of the coal is interpreted as a fourth-order sequence boundary.

Interruption of peat formation, as illustrated by the three benches of coal and interbedded partings at Stop 3, are sometimes interpreted as crevasse splays or other autocyclic depositional events (for example, Ferm and Horne, 1979). Alternatively, these interruptions are interpreted herein as the result of allocyclic controls that caused a change in water table and an influx of siliciclastic sediment. The latter interpretation is supported by the regional extent of many partings, which suggests a drowning of the peat-forming environment by a prolonged elevation of the water table, and concomitant siliciclastic deposition in a lacustrine environment (Cecil and others, 1985).

Peat formation in both the Lower and Upper Freeport paleoswamp environments was terminated by an allocyclically controlled rising water table that finally outpaced peat formation (Cecil and others, 1985). The lacustrine environment of the drowned paleoswamp became the site of an autocyclic facies mosaic of siliciclastic deposition. The shales may represent deposition in a lacustrine environment, whereas the sandstones are the result of a prograding fluvial system (Cecil and others, 1985). Channel incision appears to have been in response to progradation of a fluvial system. In the climate model of cyclic stratigraphy (Cecil, 1990), the shale and sandstone are the result of an increase in siliciclastic influx in response to a return to drier and more seasonal conditions (dry subhumid climate, table 2). This increased siliciclastic influx was coeval with the development of lacustrine systems where a rising water table was controlled by a eustatic rise in sea level. Maximum drying occurred at the time of limestone deposition. The underclay (P) and Upper Freeport coal bed overlying the Upper Freeport sandstone and shale at Stop 3 is coincident with sea-level fall and a return to humid conditions and reduced siliciclastic influx, and the correct climatic and chemical conditions necessary for pedogenesis followed by the onset of peat formation.

The Middle to Upper Pennsylvanian floral transition occurs in Upper Pennsylvanian Conemaugh Group strata approximately 100 ft (~30 m) above the Upper Freeport coal bed (at the level of the Brush Creek marine zone), when all but one of the major arboreous lycopsid genera, several tree fern, and one sphenopsid spore genera become extinct (Kosanke and Cecil, 1996). This transition is time-equivalent with the Westphalian-Stephanian boundary in western Europe and is believed to represent the culmination of a major but gradual climatic shift from a basically perhumid climate in the Early through middle Middle Pennsylvanian, to one that was subhumid, and more seasonal, in the Late Pennsylvanian (Cecil and others, 1985; Cecil, 1990). Both the floral transition and the coeval onset of deposition of calcareous red beds in the Appalachian basin were coincident with the onset of Late Pennsylvanian greenhouse conditions at the end of the ice house world that began in the Late Mississippian (fig. 10).

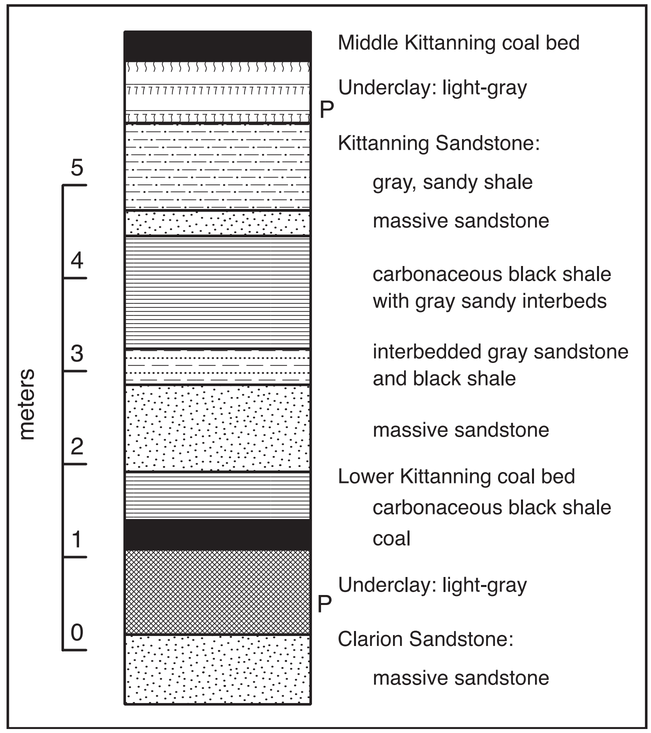


Figure 15. Stratigraphic section, Middle Pennsylvanian Allegheny Formation. Middle and Lower Kittanning intervals are shown. See figure 11 for explanation of lithology symbols.

Stop 4. The Lower Kittanning coal bed and fourth-order sequence (“cyclothem”) on Interstate 68 at milepost 15.9.

Lat 39º39′21″ N., long 79º45′47″ W., Lake Lynn, W. Va., 7.5-minute quadrangle.

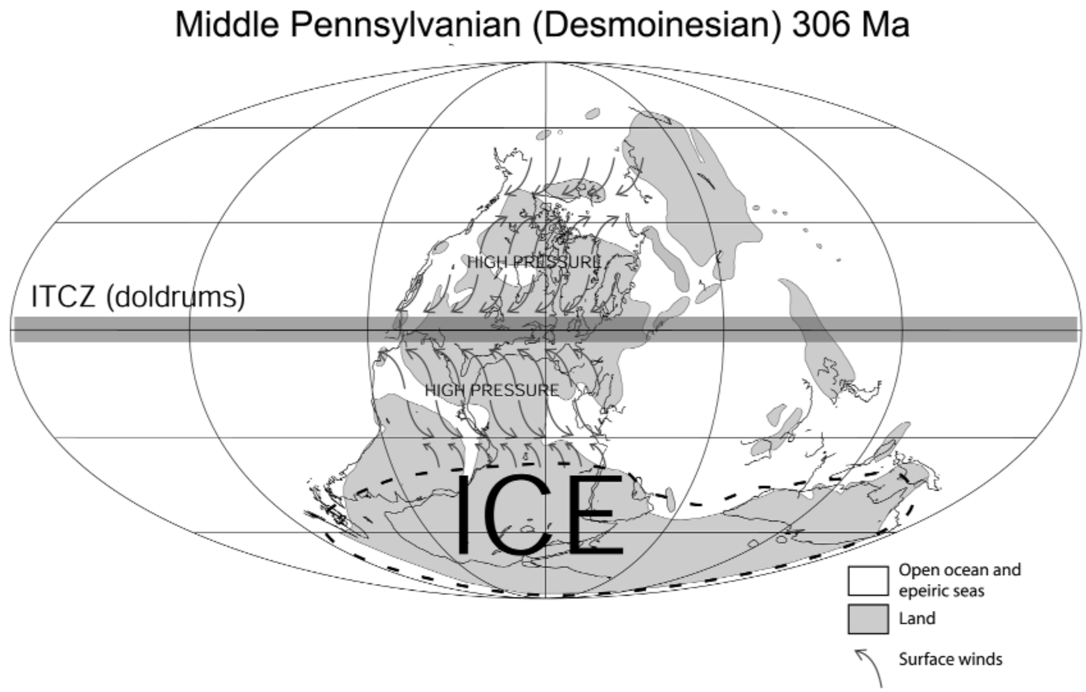
Leaders: Frank Dulong and Blaine Cecil

Introduction

The upper Middle Pennsylvanian Lower Kittanning coal bed has been correlated with the Block No. 6 coal bed in southern West Virginia (Kosanke, 1984), the Princess No. 6 coal bed in eastern Kentucky, the Colchester No. 2 coal beds of the Eastern Interior basin, and the Croweburg coal bed of the Western Interior basin (Kosanke, 1973; Peppers, 1970; Ravn, 1986; Cecil and others, 2003b; Eble, 2003). These correlations are indicative of a period of extremely widespread peat formation across eastern North America from the Appalachian basin through the Western Interior basin during a lowstand in sea level. Although thin in the Chestnut Ridge area, the Lower and Upper Kittanning coal beds attain mineable thickness and represent a significant coal reserve in northern West Virginia, eastern Ohio, and western Pennsylvania. The Lower Kittanning at Stop 4 is a thin coal bed that unconformably overlies a paleosol that is composed of flint clay and pedogenically altered sandstone. It is the interval from the base of the Lower Kittanning underclay up to the base of the Middle Kittanning coal bed that has been the focus of interbasinal correlations across the United States (Cecil and others, 2003b).

Lithostratigraphy

The interval considered at Stop 4 commences in ascending order with the flint clay (underclay) at the base, Lower Kittanning coal bed, overlying siliciclastic unit (Kittanning Sandstone) capped by another underclay, and finally the Middle Kittanning coal bed horizon. This generalized lithostratigraphy (fig. 15) can be traced throughout the central Appalachian basin. On the basis of biostratigraphy (Stamm and Wardlaw, 2003; Eble, 2003), lithostratigraphy, and sequence stratigraphy (Cecil and others, 2003b), time-equivalent strata have been traced across the North American continent. The underclay horizons in the Appalachian basin, including the flint clay here, are intensely weathered paleo-



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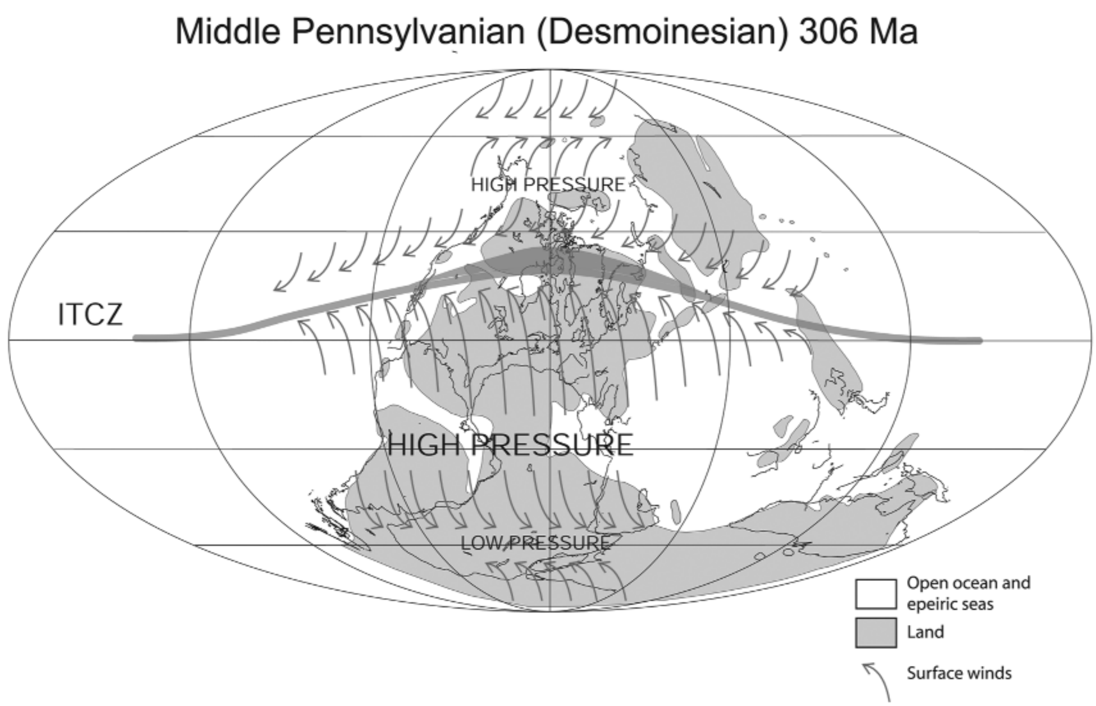
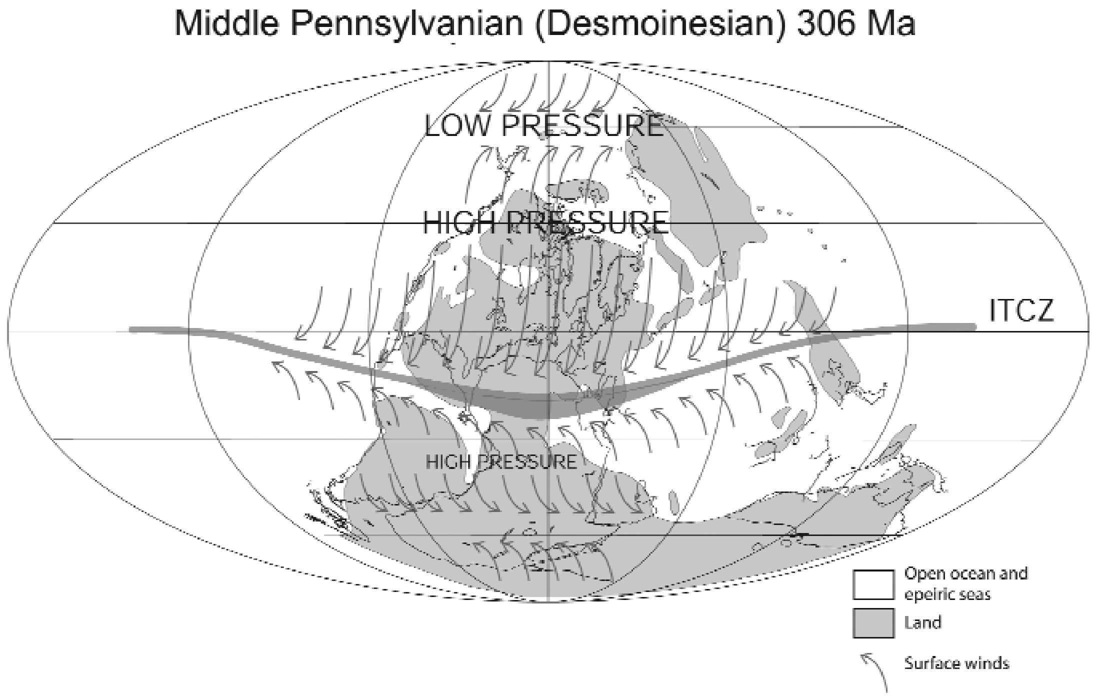


B

sols. As noted above, the coal beds have been correlated among basins across the Eastern United States. Cecil and others (2003b) have shown that the underclay deposits are even more continuous than the overlying coal beds, and coeval exposure surfaces have been traced across the continent into Arrow Canyon in southeastern Nevada. The siliciclastic unit overlying the coal varies from a marine black shale and limestone in Ohio, northern West Virginia, and western Pennsylvania to coarse-grained sand in easternmost outcrops along the Allegheny Front and in southern West Virginia. Another laterally extensive underclay paleosol and the overlying Middle Kittanning coal bed cap the sequence.

Depositional Environments

The underclay (paleosol) horizons have characteristics of intensely weathered, well-drained mineral soils that had high soil moisture regimes. These features include gleying, angular peds, high-alumina clay content, mukkarra structures, and



C

Figure 16 (this page and facing page). A, Conceptual model of Middle Pennsylvanian surface winds over Pangea during glacial intervals. Continental ice in the southern hemisphere limited the southern excursion of the intertropical convergence zone (ITCZ) during the southern hemisphere summer. Confinement of the ITCZ to equatorial regions created a low-pressure rainy belt (doldrums) that spanned approximately 10° of latitude. Schematic interpretation of continental ice is adapted from Caputo and Crowell (1985), Veevers and Powell (1987), Scotese (1998), and Frakes and others (1992). Paleogeography is modified from Scotese (1998). From Cecil and others (2003b). B, Conceptual model of Middle Pennsylvanian surface winds over Pangea during interglacial intervals. Interpreted northern excursion of the ITCZ during the northern hemisphere (Boreal) summer. Paleogeography is modified from Scotese (1998). C, Conceptual model of Middle Pennsylvanian surface winds over Pangea during interglacial intervals. Interpreted southern excursion of the ITCZ during the southern hemisphere (Austral) summer. The cross-equatorial movement of the ITCZ during interglacials resulted in significant increases in both dryness and seasonality of rainfall in low latitudes. Paleogeography is modified from Scotese (1998).

distinct soil horizonation. The interbasinal extent of these Middle Pennsylvanian paleosols and coeval exposure surfaces across the United States is indicative of a major eustatic drawdown of sea level and continental-scale exposure (Cecil and others, 2003b). In contrast, the overlying coal beds are the result of a rising water table and the onset of peat formation when the water table perpetually remained above the surface of the underlying mineral paleosol.

Paleoclimate

The paleosols, including coal beds, developed in environments where fluvial sediment supply (both dissolved and siliciclastic) was low (Cecil and others, 1985, 2003b). Such environments require perhumid or humid climates where soils are intensely leached and vegetation inhibits soil erosion as in equatorial Indonesia (Cecil and others, 1993, 2003a). The onset of peat formation has been attributed to a rise in the water table as a result of increased rainfall when the intertropical convergence zone (ITCZ) was stabilized within equatorial latitudes by maximum southern hemisphere ice (Cecil and others, 2003b). As the climate switched to interglacial conditions, sea-level rise and marine flooding outpaced peat formation and the vast peat swamps were flooded with marine, brackish, or nonmarine waters (Cecil and others, 1985, 2003b). Flooding resulted in black shale deposition over much of the craton (from the Appalachian basin westward through the Paradox basin). In the Appalachian basin, black shale deposition was followed by an input of sand from the east and southeast as the climate became progressively drier and rainfall more seasonal in response to increases in the amplitude of the annual swings of the ITCZ between hemispheres (fig. 16A, B). Deposition of clastic materials was terminated by a eustatic fall and the return of a humid climate, subaerial exposure, and pedogenesis. The interbasinal extent of paleosols including coal beds provides clear and unequivocal evidence that global climatic processes controlled sedimentation and stratigraphy.

Rolling "Stops" Through the Pottsville Group, West Flank of Chestnut Ridge Anticline on Interstate 68.

Strata of the Pottsville Group equivalent to Stop 8 are traversed (from top to base) as the trip ascends the west flank of the Chestnut Ridge anticline from Stop 4 to Stop 5. The uppermost Pottsville sandstone exposed is the Homewood(?) Sandstone (*sensu* White, 1878), the top of which (milepost 11.7) marks the boundary between the Pottsville Group and the overlying Allegheny Formation in the northern Appalachian basin. The top of the Pottsville is approximately time-equivalent with the Kanawha Formation-Charleston Sandstone (Allegheny Formation) contact in southern West Virginia. Pottsville Group strata on Chestnut Ridge contain units that are indicative of marine flooding of the Mississippian-Pennsylvanian (mid-Carboniferous) unconformity, unlike Stop 8 where the unconformity is erosional.

After passing by the Homewood Sandstone(?), the Upper and Lower Connoquenessing sandstones (*sensu* White, 1878) crop out along the highway (mileposts 11.7–14). Strata of the Pottsville Group in northern West Virginia typically consist of massive pebbly sandstones and sandy conglomerates intercalated with shale, siltstone, and thin, discontinuous coal beds. Pottsville Group sandstones, like the ones shown along I-68 on the west flank of Chestnut Ridge, generally occur as multistoried units up to 30 m (98 ft) thick, averaging 9 to 12 m (30–39 ft). Presley (1979) suggested that this group of strata was deposited by bed load, braided fluvial systems onto an alluvial plain. Meckel (1967) and Donaldson and Schumaker (1981) suggested that Pottsville sediments in the Chestnut Ridge area were derived from orogenic highlands located to the east and southeast. However, the quartzose nature of the sandstones and conglomerates is suggestive of a provenance area that had an intensely weathered and mature regolith, rather than immature sediments that would be derived from orogenic highlands.

A palynological analysis of a thin, discontinuous coal bed, informally designated herein as Pottsville coal j (mileposts 13.4–13.8), in a shale lens in the Upper Connoquenessing sandstone indicates that it is age equivalent with the early Middle Pennsylvanian (Atokan) Fire Clay-Chilton coal interval of the Kanawha Formation in southern West Virginia (Eble, 1994).

As the trip approaches the axis of the Chestnut Ridge anticline, an unnamed marine zone occurs in a siderite bed (milepost 14) beneath the Lower Connoquenessing sandstone, which is at the top of the cut. This marine unit contains a fauna that compositionally is similar to the Dingess Shale Member of the Kanawha Formation (early middle of the Middle Pennsylvanian) in southern West Virginia (T.W. Henry, oral commun., 1990). The Dingess Shale Member crops out and is found in core from southern West Virginia northward into central West Virginia but was not previously known to onlap into the Morgantown, W. Va., area. A miospore analysis of the thin (0.3 m; 1.0 ft), discontinuous coal bed, informally designated herein as Pottsville coal 2, that occurs directly beneath the Lower Connoquenessing sandstone but above the marine siderite zone at this location has shown the palynoflora to correlate with the Cedar Grove No. 2 Gas coal interval in southern West Virginia (Eble, 1994). This biostratigraphic age assignment is consistent with the invertebrate data from the marine siderite bed.

Stop 5. Mississippian-Pennsylvanian unconformity on Interstate 68 at Exit 15.

Lat 39°39′29″ N., long 79°47′00″ W., Lake Lynn, W. Va., 7.5-minute quadrangle.

Leaders: Blaine Cecil, Mitch Blake, and Rob Stamm

Introduction

Stop 5 is at the Mississippian-Pennsylvanian (mid-Carboniferous) unconformity (White, 1891), on the axis of Chestnut Ridge anticline (fig. 17). The unconformity is exposed along the eastbound lanes of I-68 (formerly U.S. 48) just east of the exit for Coopers Rock State Forest (milepost 14.7). This unconformity is global in extent (Saunders and Ramsbottom, 1986), and it is exposed at interbasinal scales across the North American craton.

Lithostratigraphy

The red beds, which crop out at the west end of the eastbound exit, are assigned to the Upper Mississippian Mauch Chunk Group (Namurian A) (see Stop 8 for explanation of Mauch Chunk terminology). Three Pottsville coal beds are present at or near Stop 5. All three are thin (<0.3 m; 1 ft) and contain low to moderate ash yields and high sulfur contents. The stratigraphically lowest coal bed occurs approximately 4 m (13 ft) above the level of the interstate highway drainage ditch. This coal bed, like most coal beds assigned to the Pottsville Group in the northern West Virginia area, is laterally discontinuous and irregular in occurrence (Presley, 1979). This coal bed was palynologically analyzed and yielded a miospore assemblage that correlates with the lower part of the Middle Pennsylvanian Kanawha Formation (unnamed coal bed below the Matewan coal) (Eble, 1994). This assemblage indicates that the stratigraphically youngest Pennsylvanian strata at this location are early, but not earliest, Middle Pennsylvanian (Westphalian B) in age. Lower Pennsylvanian strata, assignable to the Pocahontas and New River Formations are absent here. As compared to thicknesses of equivalent Mississippian and Pennsylvanian strata in southern West Virginia and southeastern Virginia, over 1,500 m (5,000 ft) of Upper Mississippian (Chesterian), Lower (Morrowan) and lowest Middle (early Atokan) Pennsylvanian strata are missing at Stop 5. Here, the Mississippian-Pennsylvanian systemic boundary occurs within the 4-m (13-ft) interval between the lower Middle Pennsylvanian unnamed coal bed and the Mississippian Mauch Chunk Group red beds. Petrographically, these coal beds contain high percentages of vitrinite and low to moderate amounts of inertinite. Despite their thinness, these coal beds display petrographic characteristics similar to age-equivalent Kanawha coal beds in southern West Virginia.

Depositional Environments

The interval exposed at Stop 5 appears to contain at least two mineral paleosols and a paleo-Histosol represented by the coal bed at the top of the paleosol sequence. The stratigraphy of the mineral paleosols is quite complex at this locality, but they appear to represent at least two periods of deposition, each followed by subaerial exposure, weathering, and pedogenesis. The top of the lowermost paleosol occurs about 2.1 m (6.9 ft) below the base of the overlying coal bed. The lower paleosol overlies and appears to grade downward into green and red strata of the Upper Mississippian Mauch Chunk Group; thus, deposition probably occurred during the Mississippian, whereas subaerial exposure and pedogenesis appears to have been during the Early Pennsylvanian. The well-developed lower paleosol may be classified as a paleo-Ultisol whereas the poorly developed upper paleosol may be more properly classified as a paleo-Inceptisol or Entisol (U.S. Department of Agriculture classification system (Soil Survey Staff, 1975; Retallack, 1989; Buol and others, 1989)). The intensely burrowed and pyritic sandstone (Betsie Shale Member equivalent) overlying the coal bed is suggestive of marine onlap and rising sea level.

On the basis of an analysis of the “mid-Carboniferous eustatic event” (Saunders and Ramsbottom, 1986), up to 4.5 m.y. may be represented in the 4-m (13-ft) interval exposed at Stop 5. The interbasinal complexity of the stratigraphy at the Mississippian-Pennsylvanian systemic boundary appears, therefore, to be the source of a great deal of confusion as to the “age” of the unconformity.

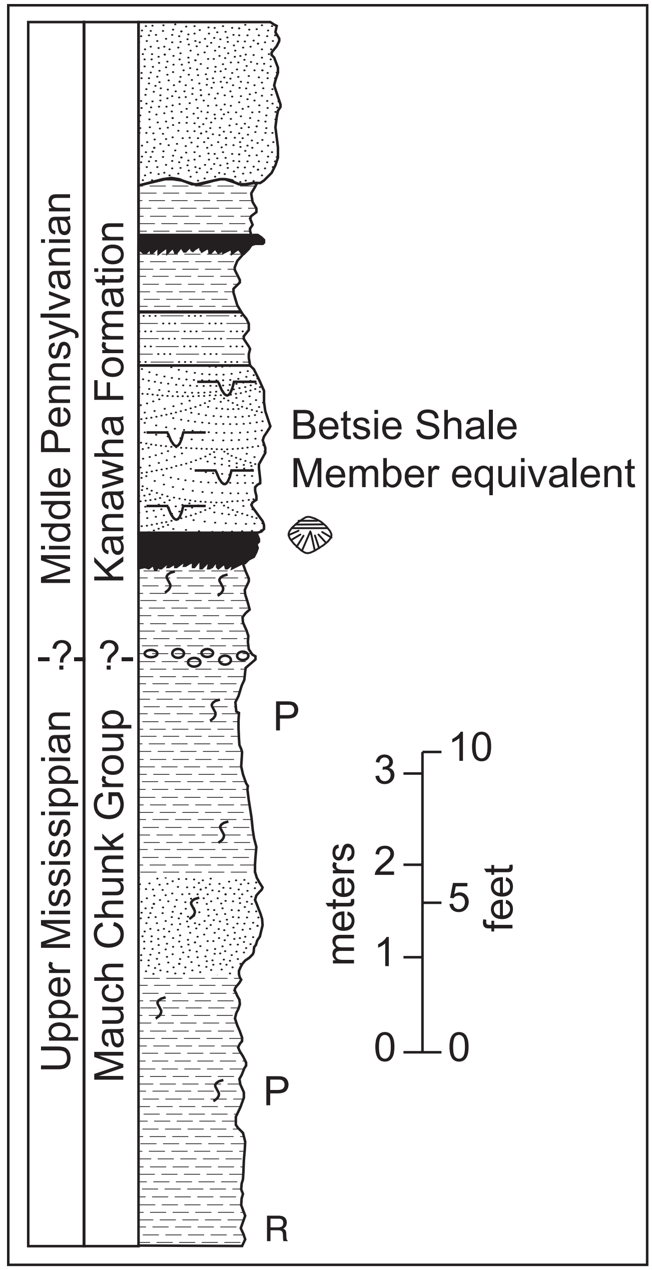


Figure 17. Stratigraphic section of the Mississippian-Pennsylvanian unconformity along Interstate 68 at Exit 15. See figure 11 for explanation of lithology symbols.

Paleoclimate

On an interbasinal scale, high-alumina refractory clay deposits, such as the Mercer clay in Pennsylvania, the Olive Hill clay in eastern Kentucky, and the Cheltenham clay on the flanks of the Ozark dome in Missouri, also occur at the Mississippian-Pennsylvanian systemic boundary. The high-alumina deposits in the eastern half of the continent are interpreted as paleo-Ultisols. In addition, residual chert (chat) occurs at the boundary in southern Missouri, northern Arkansas, and parts of Kansas. Farther west in Colorado, a thick sequence of residual cherty limestone breccia (Molas Formation) occurs at the top of the Leadville Limestone and is associated with the mid-Carboniferous unconformity. The Molas Formation is unconformably overlain by an arenaceous marine limestone of early Middle Pennsylvanian age (late Atokan) (R.G. Stamm, unpub. data). The lithologies of all residual deposits at the systemic boundary appear to be the result of intense long-term weathering and pedogenesis under high annual rainfall, which was fairly evenly distributed throughout the year (humid to perhumid). Weathering may have commenced as early as 330 Ma during the Late

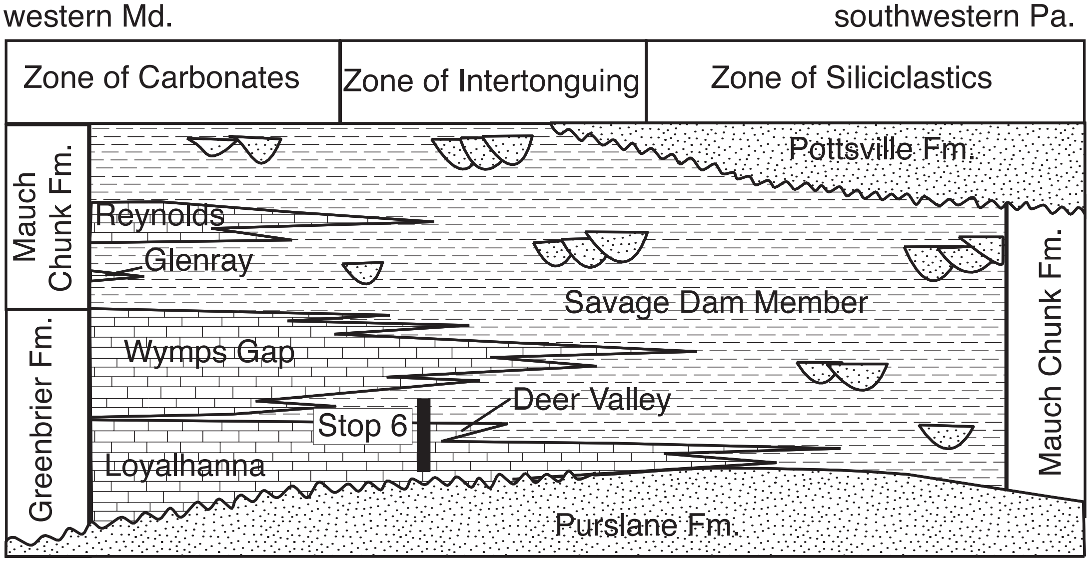


Figure 18. Regional stratigraphic relations of the Mauch Chunk and Greenbrier Formations. See figure 11 for explanation of lithology symbols.

Mississippian at the onset of the global eustatic event (Saunders and Ramsbottom, 1986). Exposure may have persisted for up to 4.5 m.y. in much of North America, including the aforementioned areas, as the continent was moving northward into the paleo-equatorial zone. However, weathering appears to have been particularly protracted and intense across the North American continent from the Appalachian basin through Colorado. Compared to the paleosol at Stop 5, other coeval exposure surfaces across the continent appear to have been somewhat more elevated and better drained, which accounts for the depth of intense weathering.

Paleosol development at the mid-Carboniferous unconformity progressed during the 4.5-m.y. period of subaerial exposure (latest Mississippian into the Middle Pennsylvanian). Sea-level fall became significant in the Late Mississippian, continued through the Early Pennsylvanian, before beginning to rise again in the early Middle Pennsylvanian (Atokan), evidenced at this stop by marine strata in the lower part of the Pottsville Group. The deep weathering (in response to a long-term humid climate), therefore, was primarily an Early Pennsylvanian event. This long-term Early Pennsylvanian humid period in the Appalachian basin was coincident with long-term ice house conditions that began in the Late Mississippian, culminated in the Early Pennsylvanian, and ended in the early Late Pennsylvanian as discussed at Stops 2 and 3. Available climate and sea-level data indicate that the mid-Carboniferous eustatic event was glacial in origin resulting from long-term ice house conditions.

Stop 6. Loyalhanna Limestone Member of the Mauch Chunk Formation at the Keystone quarry, Springs, Pa.

Lat 39°44.65′ N, long 79° 12.28′ W, Grantsville, Md.-Pa., 7.5-minute quadrangle.

Leader: Rob Stamm

Introduction

The Late Mississippian Loyalhanna Limestone Member of the Mauch Chunk Formation is examined at this stop. Also exposed are the Deer Valley and Savage Dam Members of the Mauch Chunk (Greenbrier Formation in West Virginia and Maryland) (fig. 18). The Loyalhanna's origin, with its large crossbed sets, has long been debated.

Lithostratigraphy

The Loyalhanna Limestone Member of the Mauch Chunk or Greenbrier, as it is called in Maryland, is equivalent to the Loyalhanna Formation of Pennsylvania (fig. 18). This is a widespread unit that extends at least as far north as Scranton, Pa. The Loyalhanna at this location is a 15-m (49-

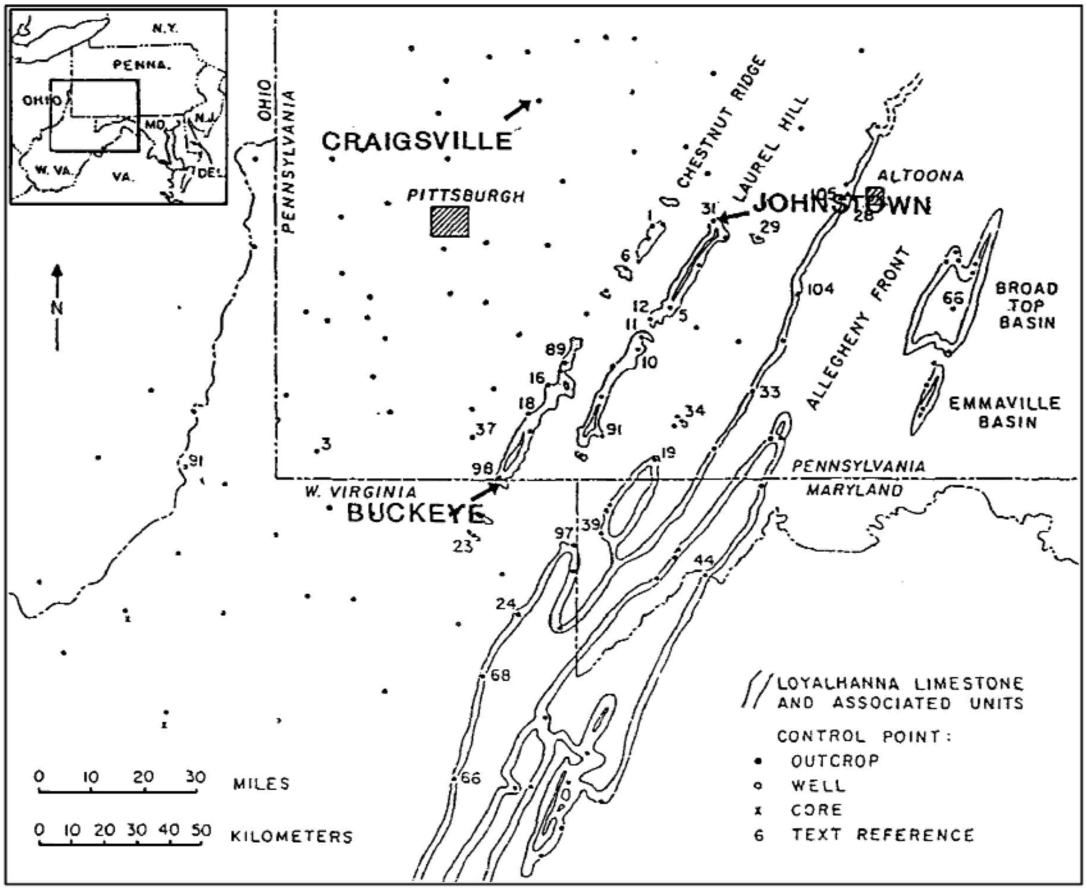


Figure 19. Map of Loyalhanna Limestone and associated units, showing outcrop, well, and core occurrences. Numbers refer to localities defined by Adams (1970).

ft)-thick, festoon crossbedded, arenaceous grainstone to calcareous sandstone. The Loyalhanna in this area and adjacent Maryland has a reddish tint caused by a small admixture of red clays that appear to be syndepositional. The quartz sand is medium to fine grained, and the carbonate grains consist of ooids, coated grains, intraclasts, and fossil fragments. The fossil fragments consist of brachiopod, bryozoan, crinoids, and endothyrid foraminifers. In addition R.G. Stamm (unpub. data) has recovered wind-abraded conodonts of probable Late Devonian age from the festoon crossbeds. Although the fossils have been typically comminuted to sand-size grains in this high-energy environment, at some nearby localities complete articulated brachiopods and straparollid gastropods have been recovered from this unit. Adams (1970) partially documented the regional extent of the Loyalhanna Member (fig. 19). Adams (1970) suggested that much, if not most, of the quartz sand has its origin from a northern Pocono source. This is indicated by the increase in terrigenous sand content in the Loyalhanna in that direction. In contrast, crossbeds are directed mainly to the east and northeast (Adams, 1970; Hoque, 1975) suggesting a provenance area to the west if the Loyalhanna is eolian in origin.

The large-scale crossbeds are accentuated on weathered joint faces along the entrance into the main working face of the Keystone quarry. This typical weathering character is caused by the more arenaceous layers, which are less soluble than the carbonate-rich layers. Thus, the more arenaceous layers stand out in relief on weathered surfaces.

At many locations on the worked face one can usually observe thin (<1.0 ft; 0.3 m), tan siltstone lenses. These lenses pinch out laterally, are convex downward, and are truncated on the upper surface. Brezinski (1989b) interpreted these lenses as representing remnants of shallowing episodes or slack water deposits preserved in swales between submarine dunes, that later were scoured and largely removed by subsequent dune migration.

Overlying the Loyalhanna is the Deer Valley Limestone Member (Flint, 1965) of the Mauch Chunk (Greenbrier) Formation, a white limestone 3 m (10 ft) thick. The Deer Valley is separated from the subjacent Loyalhanna by a 6-in (15-cm) red siltstone (not shown at the scale of figure 18). The Deer Valley differs from the underlying Loyalhanna in that the former is wavy bedded, rather than crossbedded, and lacks the quartz sand that characterizes the Loyalhanna.

Furthermore, the Deer Valley Limestone contains a more abundant, although low diversity, brachiopod fauna. The Deer Valley is composed of a peloidal lime packstone-grainstone carbonate sand. Many of the carbonate grains are ooids and intraclasts.

The Deer Valley depositional basin is largely confined to northern West Virginia and Maryland. It extends only slightly into Pennsylvania and feathers out only a short distance to the north of this location (Brezinski, 1989c), whereas the Loyalhanna extends much farther north and west. In southern Garrett County, Md., the Deer Valley Member interfingers with dark-gray limestone contained within the Loyalhanna Member of the Greenbrier Formation, attesting to at least partial contemporaneity of the two units.

Overlying the Deer Valley Limestone Member is an interval ranging in thickness from 15 to 70 m (49–230 ft) of red and green clastics. Brezinski (1989b) named this unit the Savage Dam Member of the Greenbrier Formation. In Pennsylvania this interval is considered the basal part of the Mauch Chunk Formation, although it can only be recognized where it separates the Deer Valley and (or) Loyalhanna from the overlying Wymps Gap Limestone Member. White cross-bedded sandstone, calcareous red and green siltstone and silty shale, and thin, fossiliferous limestone intervals, especially near the top of the member, characterize the Savage Dam Member. These marine lithologies are interbedded with red-brown siltstone, shale, and mudstone that are commonly mud-cracked and contain pedogenic surfaces. The alternation of marine and nonmarine lithologies led Brezinski (1989c) to contend that these strata were deposited during a number of short-lived sea-level cycles. As many as six marine-nonmarine cycles can be recognized within the Savage Dam Member (Brezinski, 1989c).

Depositional Environments

The depositional origin of the Loyalhanna has been debated for some time. Although the large-scale crossbedded foresets are suggestive of an eolian depositional setting, the abundant ooids, intraclasts, fossils, and fossil fragments, intertonguing with marine carbonates, basin geometry, and interpreted shallowing-up facies in presumed nearshore facies indicate a shallow marine origin for the unit. In contrast, Ahlbrandt (1995) presented compelling evidence indicating that the Loyalhanna is eolian.

The Deer Valley represents a submarine sand shoal environment that submerged a small area of southern Somerset County, Pa., and Garrett County, Md. It represents a distinct depositional episode from the Loyalhanna as indicated by the red siltstone that invariably separates the Deer Valley from the Loyalhanna. The cyclic marine and nonmarine lithologies that characterize the Savage Dam Member were deposited in a peritidal setting, with shallow marine sandstone and shale forming during short-lived marine transgressions, and tidal flat, rooted mudstone forming during periods of shallowing. Brezinski (1989c) interpreted these shallowing episodes as representing fifth-order cycles.

Paleoclimate

The Loyalhanna Limestone Member can be traced from as far as northeastern Pennsylvania to south-central West Virginia (figs. 19, 20). According to Ahlbrandt (1995), the Loyalhanna is an eolianite, as evidenced by sand sheets, sand flow toes, inverse graded bedding, dissipation structures, and wind ripples. The presence of eolian abraded conodonts of Late Devonian to Early Mississippian age support an eolian interpretation (pl. 1). In contrast, the many occurrences of complete, articulated, and identifiable benthic macrofauna argue for normal marine conditions. This would be in keeping with most earlier interpretations that have suggested a high-energy marine environment of deposition. However, such widespread high-energy conditions are difficult to explain. If the Loyalhanna is an eolianite, then the basin-scale distribution of this sand sea attests to aridity during a lowstand of sea level during the Late Mississippian when eastern North America was approximately 15º south of the paleoequator (fig. 7). These arid conditions contrast sharply with the humid to subhumid conditions of the Late Devonian and Early Mississippian (Stop 11) and the long-term perhumid climate of the Early and early Middle Pennsylvanian (see fig. 10).

A long-term period of aridity to semi-aridity, developed in the Mississippian during the late Kinderhookian, was most severe during the Osagean (evidenced by evaporites from Nova Scotia through the midcontinent of the United States), and continued through the Meramecian before beginning to diminish in the late Chesterian. This long-term period of aridity was coincident with general long-term Mississippian greenhouse conditions (Frakes and others, 1992). Deposition of the Loyalhanna, however, appears to be related to short- to intermediate-term (fourth-order) climate forcing mechanisms. Following deposition of the Loyalhanna and Deer Valley Members, fluvial influxes of siliciclastic sediments in the latest Mississippian foretell the onset of increases in rainfall, sea-level fall, and global climate change associated with the development of ice house conditions.

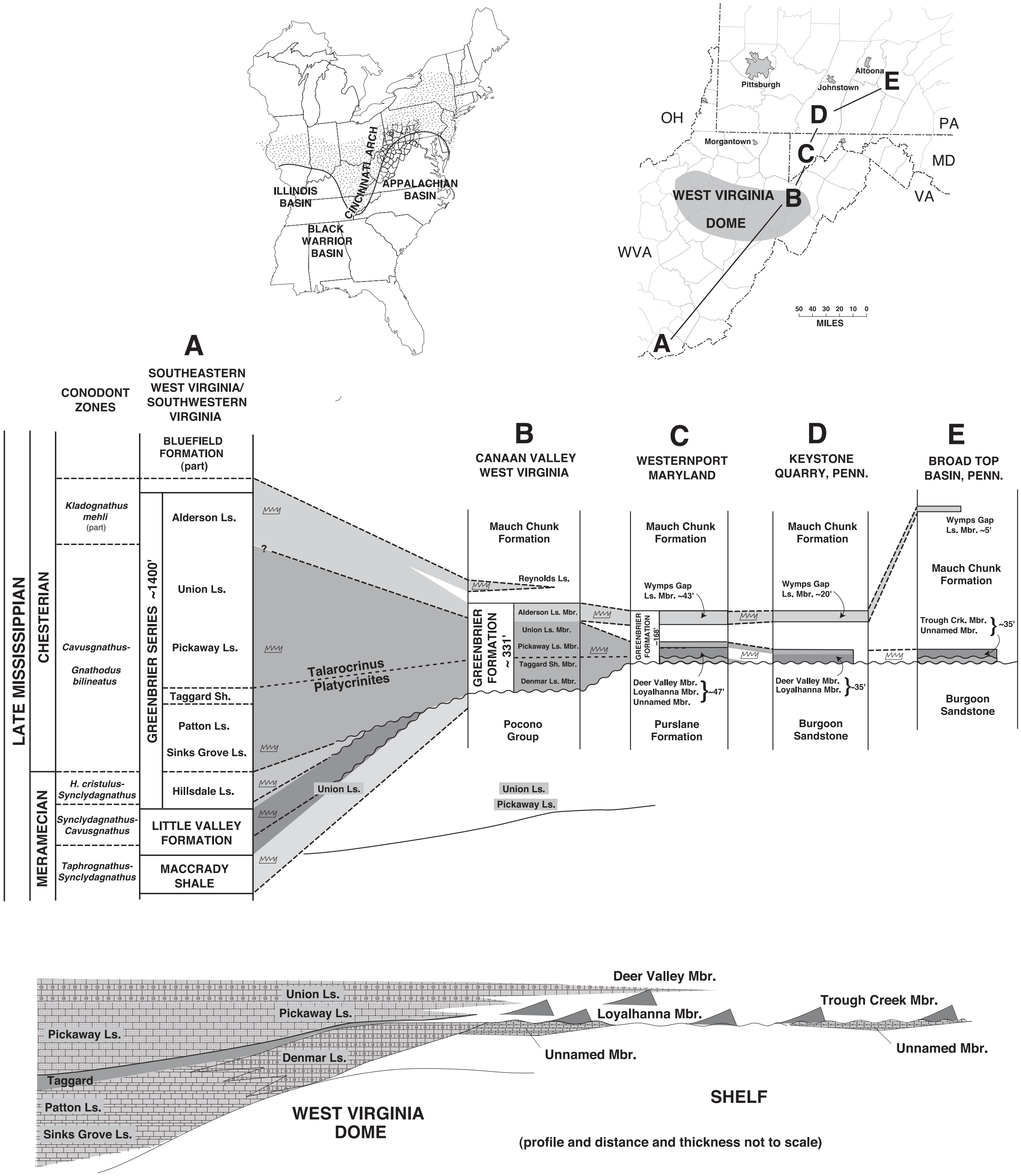


Figure 20. Conodont biostratigraphy and correlations of Greenbrier age strata from southeastern West Virginia/southwestern Virginia to south-central Pennsylvania.

Stop 7. Upper Devonian Hampshire Formation and Lower Mississippian Rockwell Formation at the Finzel Exit on Interstate 68 at Little Savage Mountain, Md.

Lat 39°40.95′ N., long 78°58.44′ W., Frostburg, Md., 7.5-minute quadrangle.

Leaders: Dave Brezinski, Rob Stamm, Vik Skema, and Blaine Cecil

Introduction

The stratigraphic sequence exposed in the highway cut through Little Savage Mountain comprises the upper Hampshire and lower Rockwell Formations (Upper Devonian-Lower Mississippian). The Rockwell succession here is part of a regional, paralic lithosome that records the ultimate foundering of the Catskill delta during latest Devonian (Famennian) time, and the subsequent evolution of an Early Mississippian (Tournaisian) coastal plain that was alternately submergent and emergent (Beuthin, 1986a–c; Bjerstedt and Kammer, 1988). Our discussion focuses on sedimentologic and stratigraphic evidence for Late Devonian-Early Mississippian shoreline shifts in this area, the implications of these shifts for placement of the Devonian-Mississippian boundary, and the relation of the Rockwell marine zones to Famennian-Tournaisian eustatic events.

Lithostratigraphy

Data for the Finzel outcrop presented herein are compiled mostly from measured sections of the Hampshire-Rockwell sequence made by Dennison and Jolley (1979), Beuthin (1986a), Bjerstedt (1986a), and Brezinski (1989a). Regionally, the Hampshire-Rockwell contact is placed at the horizon where the predominantly red strata of the Hampshire pass upward into predominantly green and gray strata of the lower Rockwell. At Finzel, the color change is abrupt, making the formational contact easy to pick.

The uppermost Hampshire consists of thin- to thickly interbedded, grayish-red mudstone, siltstone, shale, and fine-grained sandstone with sharp, convex-down bases as well as a few thin beds of green sandstone and siltstone. Many of the red beds have abundant root impressions, and pedogenic slickensides appear to be weakly developed in some of the mudstones. These strata were deposited on the Catskill deltaic-alluvial plain, mostly by aggradational overbank processes.

The Rockwell is 67 m (220 ft) thick here, but it ranges from 0 to 120+ m (0–400+ ft) in the Maryland-Pennsylvania-West Virginia tristate area (fig. 21). The basal 21 m (69 ft) of the Rockwell at Finzel constitutes a marine zone that records the final transgression over the Catskill coastal plain in this part of Maryland (fig. 22). Dennison and others (1986) informally termed this marine zone the “Finzel marine tongue” and correlated it with the marine Oswayo Formation of western Pennsylvania and the black Cleveland Shale of Ohio. Bjerstedt and Kammer (1988) and Brezinski (1989a,b) also have equated this marine zone with the Oswayo Formation of northwestern Pennsylvania. The Finzel tongue (Oswayo) also correlates with the “upper sandy zone” of the Venango Formation that crops out in the Conemaugh Gorge through Laurel Mountain, Pa. The “Venango upper sandy zone” in the Conemaugh Gorge was reported and described by Harper and Laughrey (1989) and Laughrey and others (1989). Although various names have been used for the basal Rockwell marine zone, this body of strata is a lithologically distinctive and mappable lithostratigraphic unit throughout western Maryland and Somerset County, Pa. (Beuthin, 1986a). At some locations in western Maryland, the basal beds of the Oswayo transgression are intercalated with red alluvial-plain strata of the Hampshire (fig. 23). Along the Allegheny Front, the Oswayo marine zone grades into coeval Hampshire red strata, so that at Sideling Hill (Stop 11) no Oswayo facies is evident.

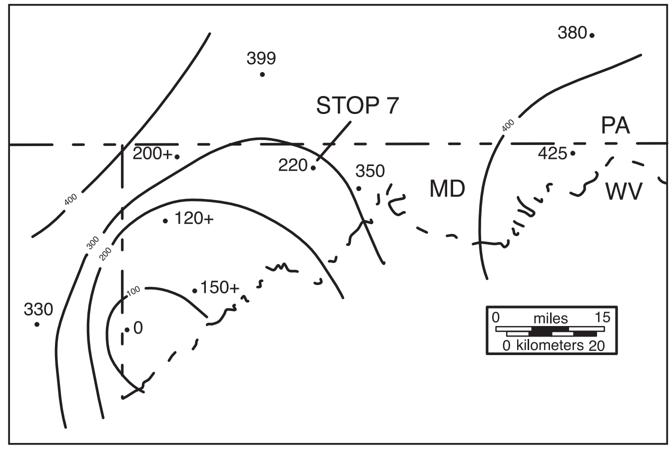


Figure 21. Isopach map of the Rockwell Formation in the Maryland-Pennsylvania-West Virginia tristate area. From Brezinski (1989a). Contour interval is 100 ft.

The Oswayo marine zone at Finzel is a coarsening-upward (shoaling) sequence of intensely burrowed, green and gray shale, siltstone, and sandstone. The lower 9 m (30 ft) of the sequence consists mostly of gray to black silty shale interstratified with thin to medium beds of gray, fine-grained sandstone. Wave-ripples and ball-and-pillow structures are common in the sandstones. Fossils from throughout the basal 30 ft (9 m) of the marine zone include a *Planolites*-dominated assemblage of bedding-plane traces, and a low diversity shelly fauna of *Lingula*, *Camarotoechia*, and unspecified bivalves. Several distinctive, thick beds of *Scolithus*-burrowed, fine-grained, greenish-gray sandstone are interbedded with gray and green shale in the upper 12 m (40 ft) of the marine zone. Marine fossils are unknown from the upper part of the marine zone. Brezinski (1989b) inferred a shallow

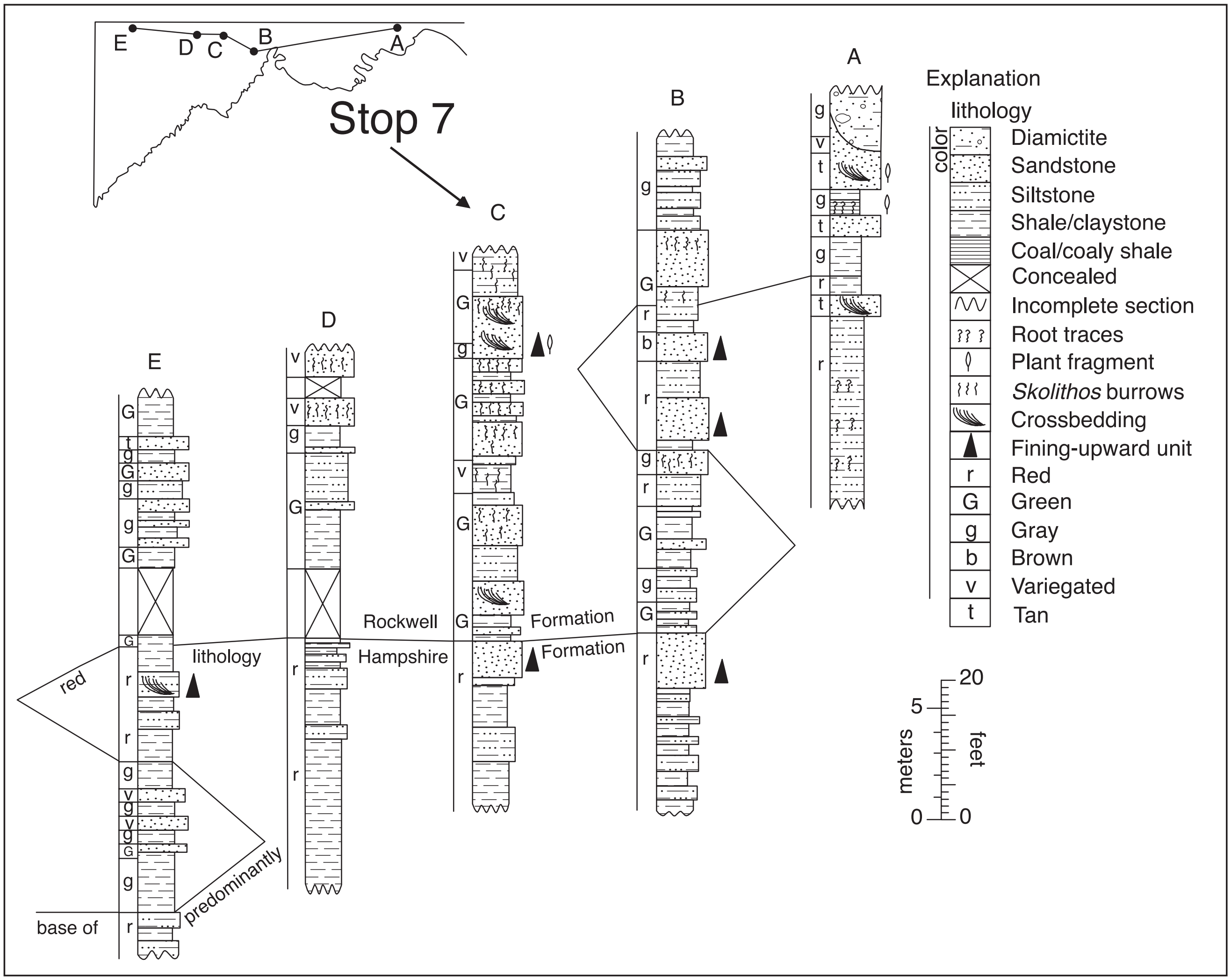


Figure 22. Intertonguing relation between the Devonian Hampshire and Rockwell Formations in western Maryland. From Brezinski (1989a).

shelf environment for the deposition of this marine zone at Finzel. Beuthin (1986a,b) and Bjerstedt and Kammer (1988) favored a restricted bay environment and interpreted the *Skolithos*-burrowed sandstones as the sand-bar complex of a prograding tidal or bayhead delta.

A 115-ft (35-m)-thick interval of lenticular, greenish-gray sandstone and reddish-brown and greenish-gray siltstone and shale overlie the Oswayo marine zone at Finzel. The sandstones exhibit erosional bases, shale-pebble basal conglomerates, crossbedding, and fining-upward texture. These beds probably were deposited on a prograding alluvial plain concomitantly with Oswayo regression (Brezinski, 1989b). Just west of Finzel, the middle nonmarine zone of the Rockwell Formation is punctuated by a thin marine unit that Dennison and others (1986) equated with the Bedford Shale of eastern Ohio. The "Cussewago equivalent" (lower Murrysville sandstone), exposed in the Conemaugh River gorge through Laurel Mountain (Harper and Laughrey, 1989; Laughrey and others, 1989), is probably equivalent to the middle nonmarine zone of the Rockwell Formation at Finzel.

A second Rockwell marine unit overlies the nonmarine Rockwell facies at Finzel. Although it is not well exposed, this upper marine zone is represented by a tan, fine-grained, medium-bedded, bioturbated sandstone. This sandstone is lithologically recognizable at other Rockwell exposures in western Maryland and adjacent Pennsylvania. This marine sandstone is correlative with the Riddlesburg Member of the Rockwell Formation in the Broad Top synclinorium of Pennsylvania (Bjerstedt and Kammer, 1988; Brezinski, 1989a,b), and exposures from the Conemaugh River gorge through Laurel Mountain (Harper and Laughrey, 1989; Laughrey and others, 1989). Throughout most of western Maryland, littoral sandstones rather than black, silty lagoonal shales, as in the Broad Top region of Pennsylvania, represent the Riddlesburg trans-

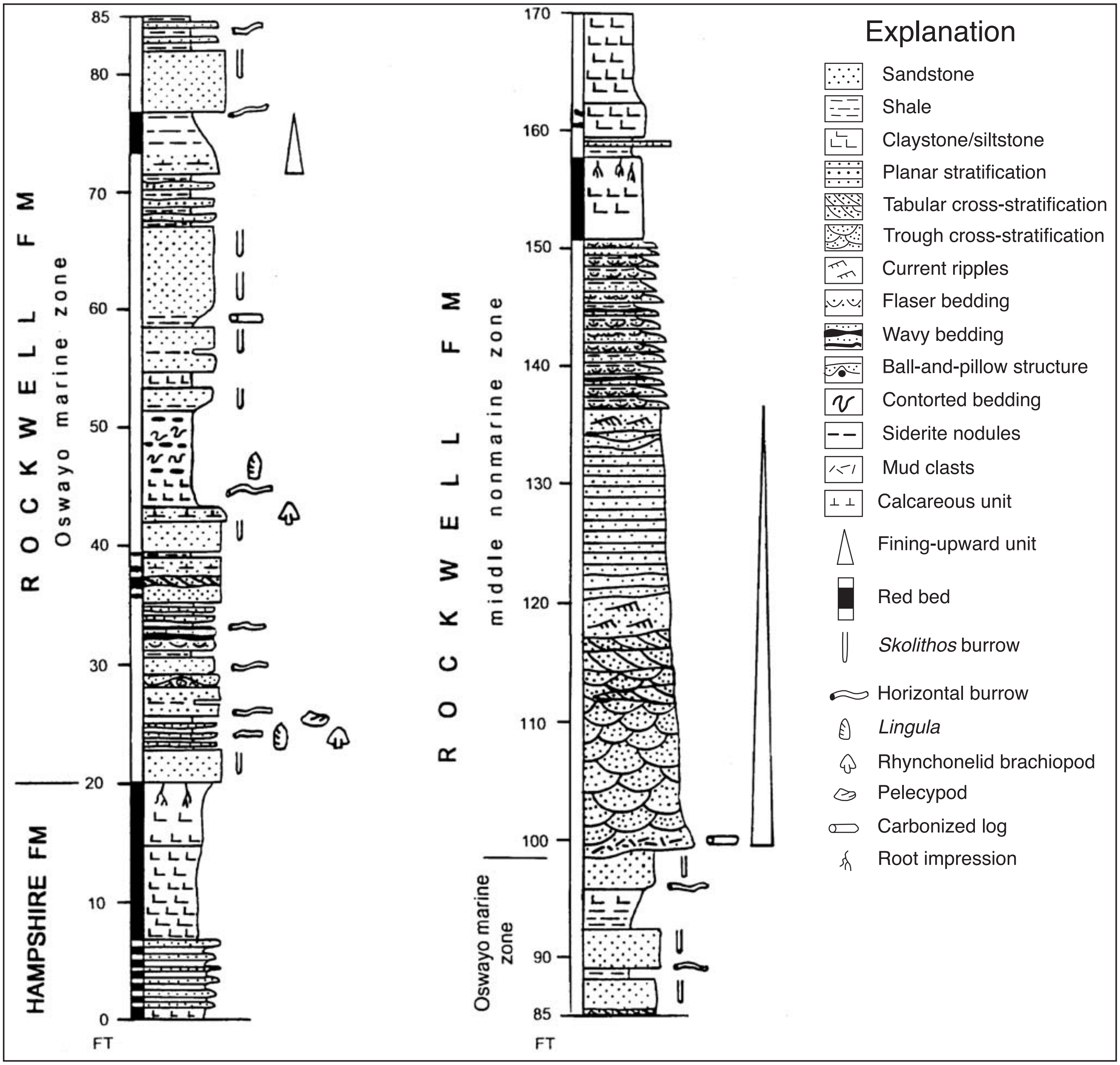


Figure 23. Measured section of upper Hampshire Formation and lower Rockwell Formation along Interstate 68 at Finzel Road interchange, Garrett County, Maryland. Section is based on exposure along eastbound entry ramp.

gression. However, a black-shale facies of the Riddlesburg marine zone has been reported about 24 mi (39 km) southeast of Finzel at Altamont, Md. (Beuthin, 1986a).

The Riddlesburg sandstone facies is exposed on the north side of I-68 at Finzel. The remaining Rockwell and overlying Purslane Formation are not exposed at this stop.

Devonian-Mississippian Contact

As concerns the Devonian-Mississippian boundary in the central Appalachians, Harper and Laughrey (1989) stated that "the exact placement of this boundary is still up for grabs and should provide some interesting discussion during this field conference." Their words are no less true for the present field trip.

Using brachiopod species considered to be index fossils, Kammer and Bjerstedt (1986) and Carter and Kammer (1990) correlated the Oswayo Member of the Price Formation of northern West Virginia with the Oswayo Formation of northwestern Pennsylvania. Those workers assigned a late Late Devonian age to the Oswayo of northern West Virginia (except for the uppermost portion) on the basis of fossil content. A Mississippian age for the Riddlesburg Member of the Rockwell Formation is generally supported by biostratigraphically significant fossils

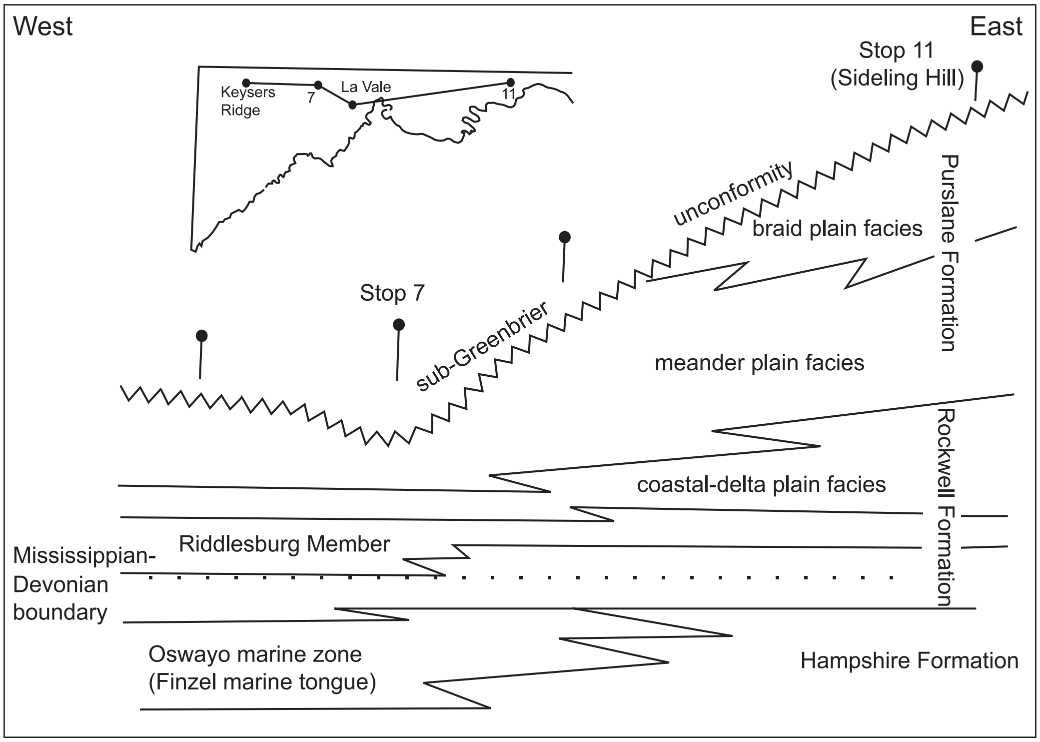


Figure 24. Environments of deposition within the Rockwell Formation of western Maryland. From Brezinski (1989b).

(Kammer and Bjerstedt, 1986). Furthermore, several recent studies (Dennison and others, 1986; Beuthin, 1986c; Bjerstedt, 1986b; Kammer and Bjerstedt, 1986; Bjerstedt and Kammer, 1988) have interpreted the Riddlesburg marine zone as an eastern facies equivalent of the Sunbury Shale of Ohio, which conventionally is assigned to the Lower Mississippian (Pepper and others, 1954; De Witt, 1970; Eames, 1974). On the basis of the aforementioned age determinations, the Devonian-Mississippian contact in western Maryland apparently falls within the interval comprising the uppermost Oswayo beds and the middle nonmarine zone of the Rockwell Formation (fig. 23).

Bjerstedt (1986b) and Bjerstedt and Kammer (1988) placed the base of the Mississippian System in West Virginia and Maryland at what they interpreted to be a regional unconformity that is equivalent to the interval of the Berea Sandstone of Ohio and subsurface deposits in West Virginia. This unconformity occurs at the base of the Riddlesburg Member of the Rockwell Formation at Sideling Hill, Town Hill, and La Vale, Md. At those locations the lower Riddlesburg consists of interbedded diamictite and crossbedded, light-gray sandstone. Lower Riddlesburg beds were deposited in shallow marine or shoreline settings that were quite likely highly erosive in nature. As a result, one would expect unconformable contacts at the base of such units. Where these marine beds are absent, physical evidence of a “Berea-age” unconformity beneath the Riddlesburg is lacking. At Keysers Ridge, the Riddlesburg Member grades into the underlying strata, indicating apparent conformity. Consequently, deposition during the Devonian-Mississippian transition probably was continuous throughout much of western Maryland, and Berea equivalents are likely present, even through they may be represented by a more terrestrial facies than the type Berea in Ohio. Harper and Laughrey (1989) and Laughrey and others (1989) have reported a Berea equivalent (upper Murrysville sandstone) from the section exposed in the Conemaugh River gorge through Laurel Mountain. At Finzel, the top of the Oswayo marine zone (Finzel marine tongue) approximates the Devonian-Mississippian contact, although the contact probably occurs slightly higher in the overlying nonmarine interval (fig. 24). At Sideling Hill (Stop 11), the Devonian-Mississippian boundary is relatively closer to the Hampshire-Rockwell Formation contact because a diamictite and the overlying Riddlesburg Member occur near the base of the Rockwell.

The intertonguing nature of the Hampshire-Rockwell Formation contact, the eastward pinchout of the Oswayo marine zone, and apparent upsection migration of the contact from west to east (fig. 24) indicate that the Hampshire-Rockwell transition is diachronous; therefore, this lithostratigraphic boundary cannot be equated with the Devonian-Mississippian systemic boundary across the region.

Depositional Environments

Figures 25 and 26 illustrate the relative sea-level changes in western Maryland and vicinity during the deposition of the Rockwell. Shaly beds of the basal Oswayo marine zone (Finzel marine tongue of Dennison and others, 1986) record the culmination of a late Famennian sea-level rise. The *Skolithos*-burrowed sandstones of the upper Finzel tongue and the crossbedded channel sandstones of the overlying nonmarine Rockwell interval indicate progradation of coarse clastics and shoreline regression during highstand. In Maryland, the sea-level drop associated with the Berea Sandstone of Ohio and West Virginia locally caused emergence above base level, sedimentary bypassing, and fluvial incision. Shortly thereafter, sea level rose again, causing the Riddlesburg transgression (fig. 25). Because the Riddlesburg extends farther east than the Oswayo marine zone, a larger rise in sea level is inferred for the former. Rockwell nonmarine beds overlying the Riddlesburg probably represent post-Riddlesburg highstand. Using microspore zonation of strata enclosing the Devonian-Mississippian boundary, Streel (1986) demonstrated the presence of two interregional transgressions on the Devonian Old Red Continent (Euramerica). The earlier of these two events was initiated during the late Famennian and is represented in Pennsylvania and Maryland by the Oswayo marine zone. The later transgression occurred during the early Tournaisian and corresponds to the Sunbury Shale of Ohio. Interregional correlations therefore strongly indicate that the Oswayo sea-level rise was eustatic. If, as inferred by Dennison and others (1986), Beuthin (1986c), and Bjerstedt and Kammer (1988), the Riddlesburg marine zone is an eastern facies equivalent of the Sunbury Shale (or a portion of it), then it also was probably eustatically controlled. Streel (1986) further suggested that the interregional late Famennian transgression was a glacioeustatic event. Caputo and Crowell (1985) and Veevers and Powell (1987) documented evidence for a mid-Famennian glacial episode in Brazil and adjacent (then) northwest Africa. Late Famennian waning of the Gondwana ice sheet perhaps was the major control on the Oswayo transgression. A glacioeustatic mechanism for the early Tournaisian (Sunbury) sea-level rise remains somewhat speculative because Early Mississippian glaciation on Gondwanaland is unresolved (Caputo and Crowell, 1985).

Paleoclimate

The paleoclimate in the Appalachian basin region during the Devonian-Mississippian transition is presented in the paleoclimate section for Stop 11.

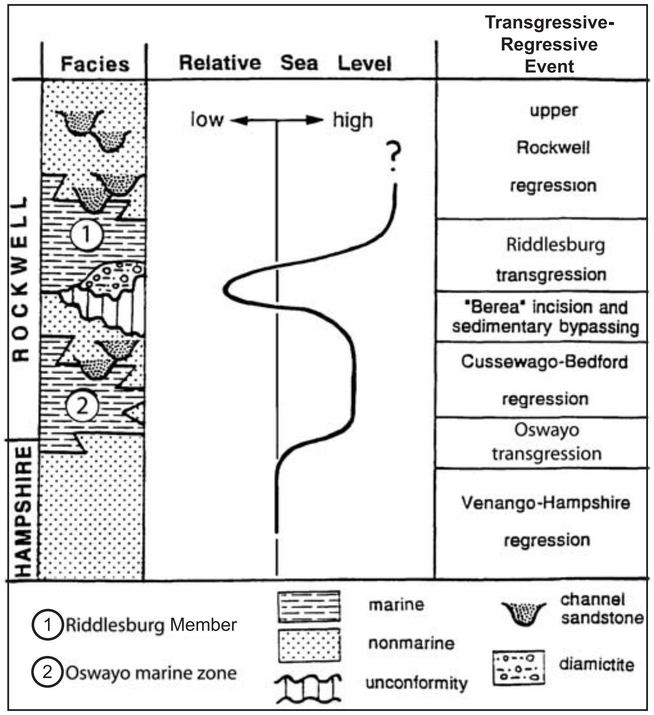


Figure 25. Relative sea-level curve for the Upper Devonian-Lower Mississippian (upper Famennian-lower Tournaisian) Rockwell Formation of western Maryland (based on Beuthin (1986a) with minor modifications). Curve does not indicate absolute magnitude, nor absolute duration of any sea-level event. Devonian-Mississippian transition probably occurred during the time-interval encompassing Cussewago-Bedford regression or "Berea" incision and sedimentary bypassing.

Stop 8. Upper Mississippian and Middle Pennsylvanian strata at the mid-Carboniferous unconformity on Interstate 68 at Big Savage Mountain, Frostburg, Md.

Lat 39°40.36′ N., long 78°57.77′ W., Frostburg, Md., 7.5-minute quadrangle.

Leaders: Blaine Cecil, Cortland Eble, and Rob Stamm

Introduction

The Upper Mississippian Mauch Chunk Formation and Middle Pennsylvanian Pottsville and Allegheny Formations are exposed in the I-68 roadcut at the top of Big Savage Mountain. The Mauch Chunk exposure is 56 m (184 ft) thick and the Pottsville and Allegheny exposures are 31 m (102 ft) and 27 m (88 ft) thick, respectively (fig. 27). The Mississippian-Pennsylvanian contact is unconformable. The laterally extensive units exposed at this locality, and their correlatives, occur throughout the central Appalachian basin.

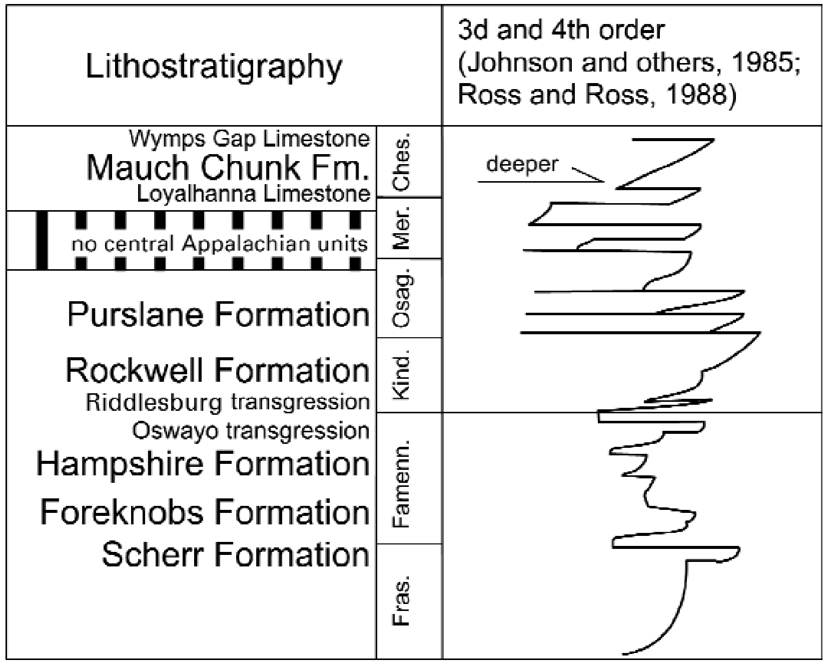


Figure 26. Third- and fourth-order global sea-level events of the Late Devonian and Early Mississippian (modified from Johnson and others (1985) and Ross and Ross (1988)), and corresponding lithologic units of the central Appalachians. Fras., Frasnian; Famenn., Famennian; Kind., Kinderhookian; Osag., Osagean; Mer., Meramecian; Ches., Chesterian.

Lithostratigraphy

The Mauch Chunk Formation in Maryland, approximately 170 m (560 ft) thick, extends from the top of the Wymps Gap (Alderson) Limestone Member of the Greenbrier Formation to the base of the Pottsville Formation (Brezinski, 1989b). The Mauch Chunk, as defined in adjacent Pennsylvania, extends from the base of the Loyalhanna Member (Stop 6) or any subjacent red beds to the base of the Pottsville. In southern West Virginia, the Mauch Chunk has group status and extends from the top of the Greenbrier Formation to the base of the Lower Pennsylvanian Pocohontas Formation. The Mauch Chunk Group in southern West Virginia includes the Bluefield, Hinton, and Bluestone Formations in ascending order. Collectively, these formations are over 2,500 m (8,200 ft) thick.

The Loyalhanna through Wymps Gap interval is a facies of the upper Greenbrier Limestone sequence to the south and west and the upper part of the Maxville Limestone of Ohio. The Mauch Chunk Formation correlates, in part, with the marginal marine Mauch Chunk Group of southern West Virginia and western Virginia and with the Bangor Limestone and Pennington Formation of Georgia, eastern Tennessee, and eastern Kentucky.

The unconformable Mississippian-Pennsylvanian boundary in western Maryland (White, 1891) is associated with the global mid-Carboniferous unconformity as discussed by Saunders and Ramsbottom (1986). At Stop 8, the Atokan-age base of the Pottsville Formation disconformably overlies the Chesterian-age Mauch Chunk Formation. This disconformity is the result of a major eustatic sea-level drop in the Early Pennsylvanian (early Morrowan). Section from the late Chesterian to late Atokan is missing here within the disconformity because of erosion and (or) nondeposition. At this stop the disconformity is at a thin claystone (fig. 27). It is often difficult to separate the Mauch Chunk and Pottsville Formations in western Pennsylvania, Maryland, and northern West Virginia because it is often difficult to distinguish non-red Mauch Chunk sandstones from those of the Pottsville Formation. The last occurrence of red coloration is not a consistently reliable criterion to separate the two units (Edmunds and Eggleston, 1993).

The Pottsville Formation usually contains a large proportion of sandstone with lesser amounts of shale and mudstone as exposed in the rolling stop between Stops 4 and 5. Here at Stop 8, the Pottsville is primarily sandstone. The olive-gray coloration of the lower part of the sandstone is atypical, as the Pottsville is usually medium gray to light gray.

The thickness of the Pottsville in this area ranges from 18 m (59 ft) near the Maryland-Pennsylvania border 11 km (7 mi) to the northeast to 100 m (328 ft) near Kitzmiller, 35 km (22 mi) to the southwest (Waagé, 1950). Here, the Pottsville is relatively thin (32 m; 105 ft). If the Pottsville here at Stop 8 is equivalent to the upper part of the Kanawha Formation of southern West Virginia, it is late Atokan and early Desmoinesian in age.

Correlations of the Allegheny Formation coal beds and, therefore, the Pottsville-Allegheny boundary are those of Waagé (1950). His coal bed correlations were derived from

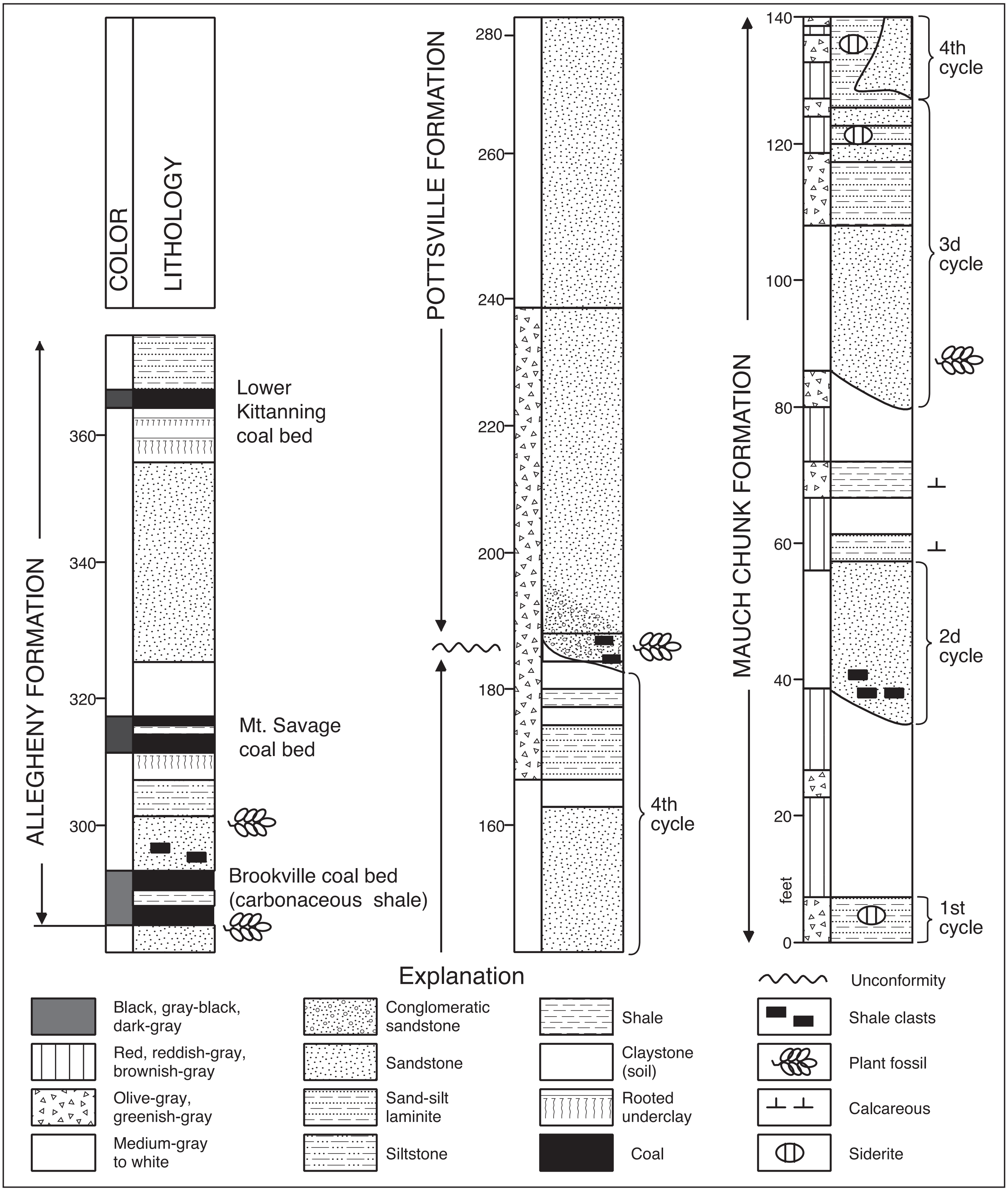


Figure 27. Interstate 68 roadcut at Big Savage Mountain, Md. Mauch Chunk Formation (west end of outcrop) through Pennsylvanian Allegheny Formation (east end of outcrop). Modified from Edmunds and Eggleston (1993).

many logs of long drill holes taken from the Conemaugh Formation to the Mauch Chunk Formation.

Pennsylvanian and Mississippian Depositional Environments

According to Edmunds and Eggleston (1993), the Mauch Chunk Formation exposed at Stop 8 was deposited in an upper-delta-plain setting. If so, then deposition was controlled by autocyclic processes.

Edmunds and Eggleston (1993) described the Mauch Chunk Formation in this area as a sequence of litharenite and sublitharenite sandstones (McBride, 1963) intercalated with mudstone and shale. Illite, mixed-layer clay, kaolinite, calcite, and siderite are common. Metamorphic and igneous rock fragments, and both potash feldspars and plagioclase are common. Grains appear to be angular, poorly sorted, and generally immature. Edmunds and Eggleston (1993) recognized four fining-upward cycles, which they interpreted as fluvial (fig. 27). Only the top of their lowest cycle is exposed. Their second cycle consists of a medium-gray, fine-grained sandstone with a scoured base and a basal lag gravel of shale clasts and caliche nodules. The second cycle culminates with reddish- and greenish-gray silty shale and hackly siltstone and claystone. Their third fluvial cycle grades upward from a medium-gray, fine-grained sandstone with an incised base containing a zone of preserved plant fragments into reddish- and greenish-gray sand-silt laminate and hackly silty shale and claystone. Their fourth fluvial cycle is medium-light-gray to pinkish-gray, fine-grained sandstone with a scoured base. The lower part of the sandstone also grades laterally into the upper part of the hackly shale and siltstone of the underlying cycle. They interpret this sequence as a distributary building across and into the terminal fine-grained clastics of the previous cycle. The sandstone passes upward into medium-gray and olive-gray interbedded hackly shale and fine-grained sandstone. The upper part of the fourth cycle is truncated by the sub-Pottsville erosion surface.

The Mauch Chunk at Stop 8 also contains calcic paleosols of unknown lateral extent. These paleosols are indicative of a drop in the paleo water-table, which generally is the result of a drop in base level or a decrease in rainfall, both of which are allocyclic processes.

Edmunds and Eggleston (1993) interpret the Pottsville Formation here as a high-energy fluvial-alluvial deposit. They suggested that the olive color of the lower of the Pottsville sandstone reflects the high content of clay altered from rock fragments. This olive color, however, is the result of modern weathering of syngenetic or early diagenetic pyrite, probably the opaque minerals noted in Edmunds and Eggleston (1993). The pyrite is the result of reducing conditions and relatively high sulfate concentrations at the time of deposition. The source of the sulfate for the pyrite was from either marine waters or high dissolved solids in fluvial systems. We suggest that the sulfate was derived from marine water in what may have been an estuarine setting. In contrast to the Mauch Chunk, the Pottsville Formation here consists of coarse-grained quartz arenite and sublitharenite. Clay minerals are generally less abundant than in the Mauch Chunk; feldspar and calcite are generally absent.

Edmunds and Eggleston (1993) attributed deposition of the clastics and coal beds within the Allegheny Formation to autocyclic delta plain depositional processes. It seems clear, however, that allocyclic processes were the primary controls on sedimentation and stratigraphy (Cecil and others, 2003b). Basin and interbasin scale paleosols (underclays and coal beds) clearly document lowstands of sea level and major fluctuations of the water table. The apparent lack of marine or brackish fossils and the general coarseness of the clastics noted by Edmunds and Eggleston (1993) is part of a facies mosaic within highstand deposits. Eastern or proximal facies along the eastern outcrop belt are of probable deltaic origin whereas distal facies in western Pennsylvania, West Virginia, and Ohio contain marine or brackish water deposits.

Paleoclimate

On the basis of the occurrence of calcic paleosols, calcic nodules as channel lags in paleostream channels, and the textural and mineralogical maturity of the sandstones, the paleoclimate of the Late Mississippian must have been semiarid to dry subhumid. Unaltered feldspar, illite, and mixed layered clays in Mauch Chunk sandstones are a good indication of limited chemical weathering in the source area and a relatively dry paleoclimate. Syngenetic and early diagenetic calcite is a further indication of a high dissolved solids content in fluvial systems. In contrast, Pottsville sandstones contain mostly quartz that is texturally immature. The absence of feldspar, mixed layered clays, and illite, and the texturally immature quartz are indicative of chemical weathering in a humid climatic regime. The absence of syngenetic and early diagenetic calcite in Pottsville sandstones is indicative of fluvial systems that were low in dissolved solids, which is consistent with a leached source area regolith that was deeply weathered under a humid tropical climate. This stop illustrates the production of quartz arenites through humid tropical weathering, in contrast to Stops 9 and 12 where the quartz arenites appear to be the product of mechanical weathering under an arid or semiarid tropical climate. This stop further indicates the change in rainfall patterns from the relatively dry conditions of the Late Mississippian to the humid conditions of the Early and Middle Pennsylvanian.

Although the Mississippian unconformity is erosional at Stop 8, residual weathering deposits commonly are developed on Mississippian and older strata (see Stop 5 discussion). In the Appalachian basin, these deposits consist of high-alumina clays, such as the Mercer clay in Pennsylvania and the Olive Hill clay in Kentucky. On the margins of the Ozark dome, residual deposits include the Cheltenham clay and unnamed chert residuum in the tristate district of Kansas, Missouri, and

Oklahoma. These residual deposits resulted from humid tropical weathering on topographic highs mainly during the Early and earliest Middle Pennsylvanian. They were subsequently buried by onlap of Middle Pennsylvanian strata in response to eustatic sea-level rise as the long-term ice house world of the mid-Carboniferous waned.

Stop 9. Upper Ordovician Juniata Formation and Lower Silurian Tuscarora Sandstone at Wills Creek Water Gap, Cumberland, Md.

Lat 39°39.75′ N., long 78°46.81′ W., Cumberland, Md., 7.5-minute quadrangle.

Leaders: Dave Brezinski, John Repetski, and Blaine Cecil

Introduction

The Upper Ordovician Juniata and Lower Silurian Tuscarora Formations are exposed in the Wills Creek Water Gap, Wills Mountain anticline, Cumberland, Md. The accommodation space and sediment supply for the underlying Middle Ordovician Martinsburg Formation, the Upper Ordovician Juniata Formation, and the basal Silurian Tuscarora Sandstone are generally thought to be the result of the Taconic orogeny when an island arc or microcontinent collided with eastern Laurentia (Drake and others, 1989). According to the tectonic model, clastics for these units were supplied from mountains created by the Taconic orogeny.

The Lower Silurian Tuscarora Sandstone is a major ridge former in the Appalachian Valley and Ridge physiographic province and, here, the Tuscarora is exposed on the west limb of the Wills Mountain anticline. The Tuscarora and equivalent sandstone strata extend from central New York (Medina Group) through Pennsylvania, Ohio (“Clinton sandstone”), Maryland, West Virginia, Virginia, and Kentucky to the Valley and Ridge in northeast Tennessee (Clinch Sandstone).

Lithostratigraphy

The Upper Ordovician Juniata Formation, which is approximately 600 ft (183 m) thick at this locality (Dennison, 1982), consists primarily of red lithic wacke interbedded with dark-red mudstone. In Pennsylvania, the Juniata is subdivided into three members on the basis of differences in the relative proportions of sandstone and mudstone (summarized in Cotter, 1993). In the lowest member, laterally extensive sheet-like sandstones dominate over the mudstones and the sandstones. The middle member contains more mudstone beds than sandstone beds. The upper member contains more sandstone beds than mudstone beds (Cotter, 1993 and references therein). In central Pennsylvania, up to 1,200 ft (366 m) of nonmarine gray and red sandstone and conglomerate of the Bald Eagle Formation intervenes stratigraphically between the Martinsburg Formation (or the equivalent Reedsville Formation) and the Juniata Formation (Thompson, 1999). The Bald Eagle Formation thins rapidly in south-central Pennsylvania and is very thin and only locally present in Maryland. The lower contact is generally placed at the first occurrence of red beds at the top of the underlying marine Martinsburg Formation (Cotter, 1993). Where the Bald Eagle Formation is present, it is placed at the first persistent occurrence of red beds. The upper contact is placed where quartz arenite defines the base of the overlying Tuscarora Sandstone (Cotter, 1983).

In contrast to the underlying Juniata, the Tuscarora primarily is a quartz arenite. In New York, strata equivalent to the Tuscarora are referred to as the Medina Group, and in eastern Pennsylvania equivalent strata are referred to as the Shawangunk Conglomerate (Thompson and Sevon, 1982).

Depositional Environments

Depositional environment interpretations of the Juniata and Tuscarora Formations presented herein are based primarily on the summaries by Cotter (1983, 1993) and references therein. According to Cotter (1993), the Juniata Formation was deposited in a lower alluvial plain setting, and the lower and upper sandstone members were deposited by braided streams. The medial member, however, may have been the result of a glacioeustatic-induced transgression toward the southeast with deposition occurring in a paralic, coastal, or lower delta plain setting (Dennison, 1976). Allocyclic changes in sea level, therefore, may have been the dominant control on accommodation space and the depositional environments of the three members of the Juniata.

Many of the mudstones in the Juniata are paleosols (Retallack, 1993). According to Retallack (1993), calcic paleosols formed on stream terraces and along alluvial fan streams. According to these interpretations, the Juniata paleosols are autocyclic in origin and not the result of allocyclic fluctuations in sea level.

The origin of the Lower Silurian Tuscarora Sandstone remains enigmatic. Cotter (1983) summarized earlier work and presented his own comprehensive sedimentological analyses of the Tuscarora in Pennsylvania. He suggested that the Tuscarora was the result of an Early Silurian transgressive event and renewed tectonic elevation of the Taconic terrane. He divided the Tuscarora into five regional and stratigraphic lithofacies. “Lithofacies one,” his eastern cross-laminated lithofacies, was attributed to braided stream deposition. “Lithofacies two,” his western cross-laminated lithofaces, was deposited in a marine-shelf sand wave and shoreface-connected sand ridge environment. Cotter noted that sand grains in “lithofacies two” were well rounded. “Lithofacies three,” his basal horizontal laminated facies, was deposited in foreshore and shoreface environments; quartz grains in this lithofacies also are well rounded. “Lithofacies four,” basal pink transitional lithofacies, was deposited in paralic conditions. Texturally mature grains also were noted in this lithofacies in contrast to the lithic arenites of the underlying Juniata. “Lithofacies five,” the uppermost part of the Tuscarora, is the red Cacapon Sandstone Member in West Virginia and Maryland (Castanea Member in Pennsylvania), which he suggested was deposited in a low-energy coastal flat complex that prograded over the underlying quartz arenite facies during regression. The Cacapon Sandstone Member is intensely burrowed locally, and it is composed primarily of rounded quartz sand grains in a hematitic matrix (C.B. Cecil, unpub. data).

Paleoclimates of the Late Ordovician and Early Silurian

On the basis of carbonate occurrence and other paleosol features in the Juniata Formation, Retallack (1993) suggested that the climate of the Late Ordovician was semiarid and that annual rainfall, although limited, was seasonal. This interpretation is consistent with the presence of braided stream deposits in the upper and lower members. However, Late Ordovician sea-level oscillations, controlled by advance and retreat of ice sheets in what is now North Africa (Dennison, 1976), must have been accompanied by global climate change. Given a paleolatitude of approximately 30º south for the Appalachian basin in Late Ordovician time, the amount and seasonality of annual rainfall was the predominant form of climate change. Such changes in rainfall patterns could account for the variation in sediment supply within and among members of the Juniata. The braided stream sheet sands of the upper and lower members of the Juniata may represent increased sediment supply whenever the climate shifted from semiarid toward dry subhumid conditions. In an alluvial plain setting, the return from dry subhumid to semiarid conditions would favor a reduction in sediment supply, deposition of finer grained material, and a lowering of the water table resulting in conditions necessary for soil development.

The change in lithology from the immature red wacke sand of the Juniata to the mature quartz arenite of the Tuscarora and equivalent strata remains problematic. Most interpretations attribute the textural and mineralogical maturity of the Tuscarora to marine depositional processes (for example, Cotter, 1983). However, an eolian mechanism for winnowing and rounding of grains is far more likely because, unlike gravel and cobbles, sand is far more susceptible to rounding under eolian conditions than in aqueous environments (for example, Kuenen, 1960). It appears, therefore, that the Tuscarora was derived from a mature eolian regolith that consisted primarily of quartz.

Dennison and Head (1975) and Brett and others (1995) recognized an unconformity at the contact between Upper Ordovician and Lower Silurian strata that extends from Ontario southward through the central Appalachian basin. This regional unconformity (Cherokee unconformity of Dennison and Head, 1975) separates the immature lithic sand and mudstone of the Upper Ordovician from quartz arenite of the Lower Silurian. It is highly probable that a mature eolian regolith developed on the exposed surface as a result of eolian mechanical weathering and winnowing, given the long-term semiarid to arid conditions of the latest Ordovician and earliest Silurian. Although there appears to be some fluvial deposition along the eastern outcrop belt, most workers believe that the Tuscarora was deposited under marine conditions during transgression (for example, Cotter, 1983). Reworking of texturally and mineralogically mature eolian materials by marine transgressive processes appears to account for the lithology and the widespread distribution of the Tuscarora and equivalent strata over the Cherokee unconformity.

Stop 10. Silurian strata of the Wills Creek and Tonoloway Formations on Interstate 68.

Lat 39º41.48′ N., long 78º39.37′ W., Evitts Creek, Md., 7.5-minute quadrangle.

Leaders: Dave Brezinski, John Repetski, and Blaine Cecil

Introduction

The Late Silurian tidal flat facies of the Wills Creek and Tonoloway Formations is exposed along the south side of I-68 at this stop. The section exposed at this stop includes (from bottom to top) the upper Rose Hill Formation, Keefer Sandstone, McKenzie Formation, including the Rochester Shale Member at the base, followed by the Wills Creek Formation and, finally, the Tonoloway Formation. The Bloomsburg Formation is not exposed. We will concentrate on the upper part of the section, which is Late Silurian in age (fig. 28).

Lithostratigraphy

The upper part of the section begins along the ramp of eastbound I-68 within the basal strata of the Tonoloway Formation. Although the Tonoloway is more than 120 m (390 ft) thick, only about 3 m (10 ft) of this unit is exposed at this location. The Tonoloway consists of thinly bedded to thinly laminated shaly limestone. The formation characteristically weathers into tan, platy and shaly chips. The only fauna known from this part of the Tonoloway are leperditiid ostracodes, eurypterids, and a low diversity conodont association. Some of the platy limestone chips exhibit casts of gypsum and halite crystals. Westward (downsection), the strata become increasingly shaly. With this increase in siliciclastics the laminated Tonoloway passes into the underlying Wills Creek Formation. The Wills Creek is characterized by thin- to medium-bedded, gray-green to tan weathering, calcareous shale and laminated silty dolomite. The formation contains

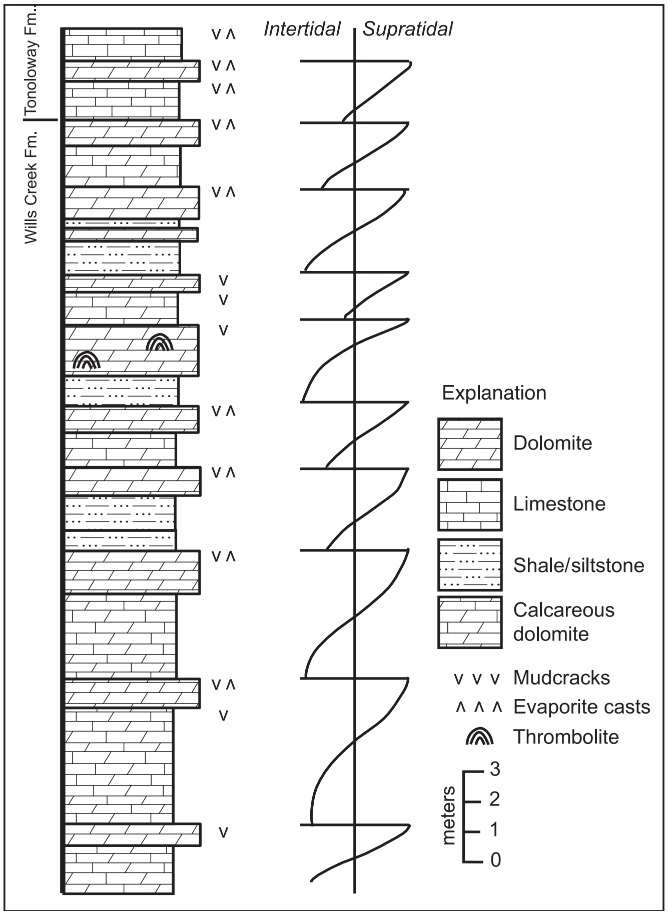


Figure 28. Stratigraphic section exposed at Stop 10 of the upper Wills Creek and basal Tonoloway Formations and their interpreted depositional environments.

horizons of desiccation cracks and domal stromatolites. Halite and gypsum casts also are common in some intervals near the top of the formation. The Wills Creek Formation is approximately 150 m (490 ft) thick in this area (Swartz, 1923). It is correlative with the salt beds of the Salina Formation that occur in the subsurface to the west in Pennsylvania and West Virginia, and to the north in New York (Allings and Briggs, 1961; Smosna and Patchen, 1978).

Along Md. 144, the Wills Creek strata seen on the north side of I-68 also are exposed. Here the alternating shale-dolomite couplets that characterize the Wills Creek Formation can be observed (fig. 29).

Depositional Environments

During the Late Silurian much of the Appalachian basin became a restricted salt basin. The greatest thickness of salts (halite and anhydrite) was formed in an elongate basin stretching from northern West Virginia through western Pennsylvania into western New York (Allings and Briggs, 1961; Smosna and Patchen, 1978) (fig. 30). Along the eastern margin of this salt basin, extensive supratidal mud flats and sabkhas formed. It was on these mud flats that the peritidal lithologies of the Wills Creek and Tonoloway Formations were deposited. The regional extent of these peritidal units documents basin-scale tidal sedimentation. The shaly component of these formations appears to have been the result of either of the following: (1) the distal facies of a Silurian clastic wedge that formed to the east and resulted in the deposition of the Bloomsburg Formation or (2) the result of eolian (dust) deposition as suggested by Grabau (1932, p. 569) for equivalent strata in New York. The great thickness of mud flat lithologies records a large number of small-scale (fifth-order?) shallowing cycles (fig. 29). Rarely are any subtidal lithologies preserved within these cycles, indicating that most of the deposition occurred within the intertidal and supratidal setting. Indeed, the preponderance of evaporite casts in some horizons attest to the supratidal salt flat (sabkha) environment of deposition.

The shaly evaporite flats persisted through most of the Late Silurian with the deposition of the Tonoloway Formation. During the latest Silurian, the Appalachian basin was once again submerged with normal marine waters as a deepening seaway resulted in the deposition of the Keyser Formation as seen at Stop 12.

Late Silurian Paleoclimate

There is abundant evidence for long-term aridity in the Late Silurian in the Appalachian basin. It was during this time that extensive salt deposition was occurring in deeper parts of the basin and in the Michigan basin. The common evaporite casts are indicative of an arid climate as well as a supratidal, salt flat (sabkha) environment of deposition. The formations seen at this stop are regional in extent and record mixed tidal and eolian processes on the flat floor of a continental basin under arid climatic conditions. The general conditions of Late Silurian sea-level lowstand and coeval aridity suggest significant amounts of ice in high latitudes. If so, glacial-interglacial fluctuations would account for the intermediate-term sea-level fluctuations that are recorded at this stop.

Stop 11. Upper Devonian and Lower Mississippian strata on Interstate 68 at Sideling Hill, Md.

Lat 39°38.9′ N., long 79°50.0′ W., Bellegrove, Md., 7.5-minute quadrangle.

Leaders: Blaine Cecil, Dave Brezinski, Vik Skema, and Rob Stamm

Introduction

As the trip progresses eastward, Upper Devonian and Lower Mississippian strata are exposed in a syncline at

Figure 29. Shale-dolomite couplets and corresponding interpreted fifth-order sea-level events in the upper Wills Creek Formation along Md. 144 at Rocky Gap State Park, Md.

Sideling Hill, Md. At this stop we will examine the transition from the red alluvial plain deposition at the end of the Devonian to the coal-bearing strata of the Early Mississippian. Evidence for Late Devonian-Early Mississippian glacial conditions in the Appalachian basin is emphasized at this stop.

Lithostratigraphy

The roadcut through Sideling Hill exposes the Lower Mississippian Rockwell and Purslane Formations (fig. 31). The Rockwell Formation, the lower of the two units, extends from the lowest strata exposed up to the base of the first thick (>10 m; 32 ft) sandstone unit. The Purslane Formation, which is composed mainly of sandstone, is resistant to erosion and forms the prominent ridge of Sideling Hill.

The Rockwell Formation consists of interbedded thin sandstone, siltstone, and coaly marine shale. This exposure of the Rockwell Formation contains a polymictic diamictite near the base. This basal diamictite is a sandy, mud-supported, unsorted mass, containing clasts ranging in size from pebbles through boulders. Some of the clasts are composed of granite, chert, and graywacke. At some locations large masses, meters in diameter, of the underlying Hampshire Formation are incorporated in the diamictite lithology. The provenance of most clasts, however, is unknown. Overlying the diamictite is a 10-m (33-ft)-thick interval of herringbone crossbedded, medium- to coarse-grained sandstone. Upsection from the herringbone crossbedded sandstone is a 30-m (98 ft)-thick interval of interbedded thin coal beds and sandstone and siltstone that contains marine fossils along some horizons. This interval has been correlated with the Riddlesburg Shale of central Pennsylvania, an Early Mississippian marine transgression equivalent to the Sunbury Shale of Ohio (Bjerstedt, 1986a; Bjerstedt and Kammer, 1988; Brezinski, 1989a). Some of the brachiopods contained within this interval include lingulid brachiopods; the articulate brachiopods, *Rugosochonetes*, *Macropotamorhyncus*, and *Schuchertella*; as well as indeterminate bivalves (Brezinski, 1989b).

Overlying the Riddlesburg Shale Member, in the upper part of the Rockwell Formation, is a 50-m (164-ft)-thick interval of interbedded channel-phase sandstones, 3 to 10 m (10-33 ft) thick, and reddish-brown and medium-gray, root-mottled siltstone and mudstone.

The first thick sandstone encountered upward marks the base of the overlying Purslane Formation. This basal sandstone of the predominantly sandy Purslane Formation is 25 m (82 ft) thick, is medium grained, exhibits epsilon crossbedding, and contains basal lag gravels. The basal unit is overlain by a thin interval (7 m; 23 ft) of root-mottled, reddish mudstone and then another (15 m; 49 ft) thick, crossbedded sandstone unit. Overlying this second sandstone unit is an interval of interbedded coaly shale and thin coal. A 30-m (98-ft)-thick third sandstone unit caps the ridge. This uppermost unit is much coarser grained than either of the lower units and contains many crossbedded conglomeratic intervals (Brezinski, 1989b).

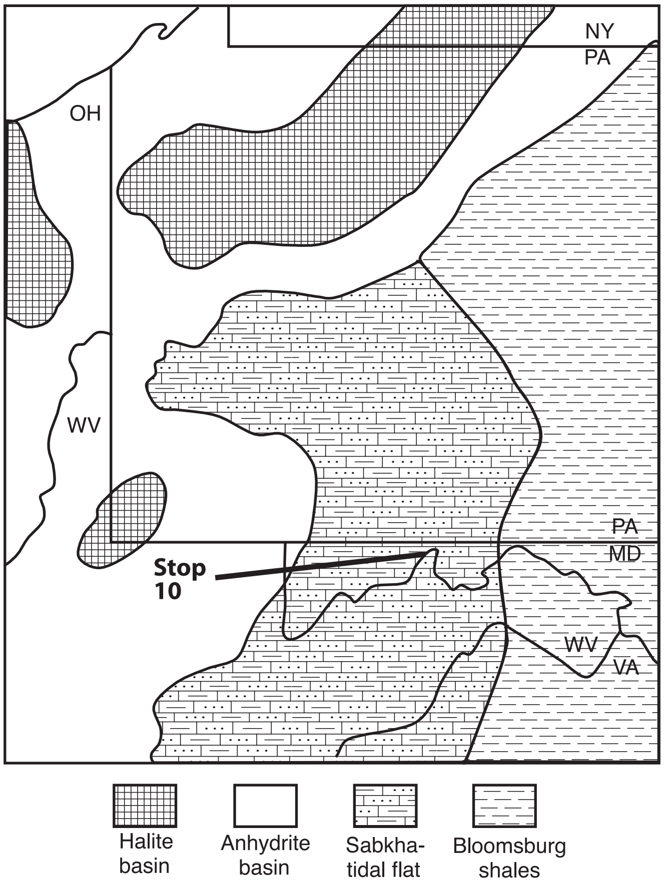


Figure 30. Regional-scale environments of deposition for the Middle to lower Upper Sillurian.

Depositional Environments

The Lower Mississippian sequence at Sideling Hill was interpreted by Brezinski (1989a) as a prograding clastic wedge that progressively replaced marine deposits with coastal plain (marsh), meandering fluvial, and braided fluvial deposits (fig. 31). The basal polymictic diamictite is, however, an enigmatic stratigraphic interval (ESI). The unsorted character, strike-localized extent, and presence of "slump-blocks" of the underlying Hampshire Formation lithologies led Sevon (1969) and Bjerstedt and Kammer (1988) to interpret these deposits as subaqueous mud and debris flows that occupied an older sediment dispersal system. The consistent stratigraphic relation between the diamictite and overlying herringbone crossbedded sandstone further led Bjerstedt and Kammer to propose that the sandstone deposits represented tidal inlet-fill sequences created by tidal processes during drowning of the preexisting channels during the Riddlesburg transgression. Suter (1991) similarly proposed that the polymictic diamictite is the result of an estuarine debris flow. However, as discussed below, Cecil and others (2002) have suggested that the diamictite and overlying laminated strata are related to glacial conditions in the Appalachian basin during the Devonian-Mississippian transition.

The strata overlying the diamictite and herringbone crossbedded sandstone (tidal?) become increasingly darker in color and consist of interbedded coaly shale and siltstone. These coaly strata grade upsection into marine strata. The restricted brachiopod-bivalve fauna present in these shales indicates deposition in an estuarine or restricted lagoonal setting (Brezinski, 1989a). The return upsection to coaly shale indicates progradation of the tidal marsh environments back over previously deposited lagoonal deposits. The upper Rockwell's characteristic interbedded thin channel sandstone and finer grained overbank deposits that are pervasively root-mottled, indicate that the upper Rockwell was deposited in an alluvial-coastal plain environment.

The thick, channel-phase sandstones with epsilon cross-bedding and thin overbank deposits of the lower Purslane Formation indicate a channel-dominated, meandering fluvial environment. The coarser conglomeratic deposits in the upper part of the formation may indicate that gradients increased and environments changed from a meandering fluvial to braided fluvial facies near the end of Purslane deposition (Brezinski, 1989a-c) (fig. 31). The recent recognition of the pervasive occurrences of moderately well-developed paleosols throughout the section at this locality may require reinterpretation of certain aspects of the Lower Mississippian depositional systems.

Late Devonian and Early Mississippian Paleoclimate

The Late Devonian and Early Mississippian strata exposed at Stop 11 are indicative of paleoclimates that ranged from dry subhumid to moist subhumid to humid in sharp contrast to the aridity of the Early Devonian and subsequent Mississippian arid climatic conditions (fig. 10). Upper Devonian paleo-Vertisols and Histosols (coal beds), the influx of siliciclastic material, and a paucity of calcareous materials, are all indicative of relatively wet climatic conditions. Mineral paleosols include weakly to moderately developed non-calcic paleo-Vertisols. Such Vertisols are the result of a soil moisture regime and chemistry in which concentrations of exchangeable cations are low, which precludes precipitation of pedogenic carbonate. Non-calcic Vertisols primarily form under a moist subhumid climate. The soil structures, including mukkarra structures, are clear and unequivocal evidence, however, for a climate with a distinct dry season (Retallack, 1989).

The Early Mississippian sandstone exposed at the top of the cut is exceedingly widespread and extends from New York and Pennsylvania (Pocono Sandstone) to at least as far south as Kentucky (Borden Formation) and Tennessee (Price Formation). This clastic wedge, with its sporadic coal beds, represents a period during which massive amounts of sand were transported into the basin. Rainfall conditions ranged from dry subhumid to moist subhumid, and rainfall was somewhat seasonal (fig. 1A). The transition from meandering

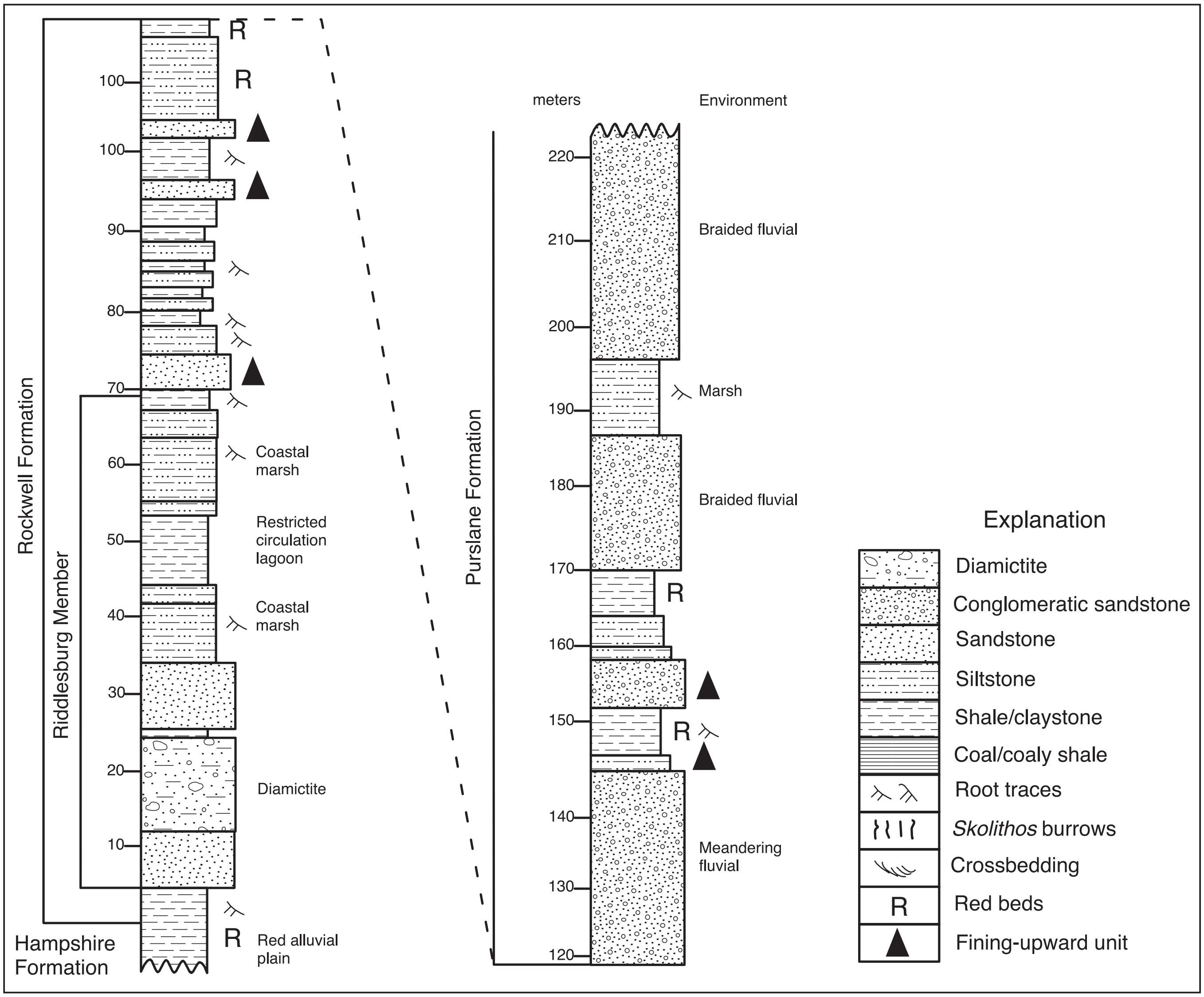


Figure 31. Stratigraphic section exposed at Stop 11 and interpreted depositional environments. Modified from Brezinski (1989a).

fluvial deposition in the lower Purslane to braided fluvial deposition in the upper Purslane can be attributed to increasingly seasonal rainfall and a progression from a humid or moist subhumid climate to dry subhumid conditions and finally arid conditions during deposition of the overlying Maccrady and Greenbrier Formations (Cecil, 1990). This interpretation is consistent with coal occurrences in the Pocono and equivalent units and evaporites, Aridisols, and carbonates in overlying Mississippian strata (Cecil, 1990). It is highly unlikely that the changes from Devonian aridity to the humid conditions of the Late Devonian and Early Mississippian, followed by a return to aridity in the Mississippian (Osagean), were the result of either long-term climate change caused by continental drift through paleolatitudes or tectonic processes.

Evidence for Late Devonian and Early Carboniferous Global Cooling and a Humid Climate in the Appalachian Basin

There is compelling evidence for Late Devonian glaciation in the Amazon Basin and in northern Africa (Libya and Algeria) (Caputo and Crowell, 1985; Veevers and Powell, 1987). In addition, Frakes and others (1992) suggested that the Late Devonian-Early Mississippian was a time of global cooling (ice house conditions). Even though evidence for Late Devonian glacial conditions in the Appalachian basin has been recognized for many years, this evidence has consistently been discounted (Sevon, 1969, p. 7; Berg, 1999, p. 133). However, compelling evidence for Late Devonian glaciation in the Appalachian basin is evident in the Spechty Kopf Formation and overlying strata in Pennsylvania, and the correlative units (lower Rockwell) in Maryland and northern West Virginia. This is the enigmatic stratigraphic interval (ESI) referred to above in the section that discusses depositional environments. The ESI is known to occur in a 40-km (25 mi)-wide belt extending 400 km (249 mi) from northeastern Pennsylvania, across western Maryland, into north-central West Virginia. This belt is about 45 km (28 mi) wide in the northeast as measured between Jim Thorpe and Wilkes Barre, Pa., in the region of the anthracite coal fields. Southward in Maryland, it narrows to approximately 25 km (16 mi), but this may be very misleading and strictly a result of erosion and of a lack of exposure to the east. The original eastern extent of the ESI is unknown because of erosional removal of the section. The western border of the ESI belt is a well-defined curvilinear edge. There are many exposures of the horizon to the west of the belt and none contain the diamictite or related lithologies. For example, Stop 7 at the Finzel exit of I-68 is 10 km (6 mi) west of the La Vale ESI exposure, but none of the same lithologies are contained at the horizon. The same is true at a number of locations in Pennsylvania where the western edge of the ESI terminates abruptly. The ESI thickness in northeastern Pennsylvania varies locally from 0 to 185 m (0–607 ft) (Berg, 1999). The following features suggest a glacial to periglacial control on the origin of the ESI (in ascending stratigraphic order): (1) a basal, nonbedded polymictic diamictite containing matrix-supported clasts (of known and unknown origin) up to 2 m (7 ft) in diameter (tillite), (2) bedded siltstone and sandstone containing matrix-supported polymictic clasts, up to 2 m (7 ft) in diameter (dropstones), and (3) laminites (flaser-bedded sandstone, as discussed above) with clasts (glacial varves and dropstones, respectively). Striated facets, which we interpret to be glacial in origin, are present on some clasts. The tillite, dropstones, varves, and faceted-striated clasts in the ESI are indicative of deposition at or near the terminus of either a continental ice sheet that was centered in Gondwanaland (South America) or glaciers that debouched from undocumented highlands to the southeast. The regular nature of the curvilinear western edge of the belt is most consistent with the ice sheet hypothesis. In either case, the study region was situated in the vicinity of lat ~30° S. during the Devonian-Carboniferous transition (Scotese, 1998), which is approximately coeval with the onset of Late Devonian conditions of global cooling (Frakes and others, 1992).

The ESI unconformably overlies the Upper Devonian Catskill Formation (Hampshire Formation in West Virginia), which is characterized by red beds with caliche, suggesting that deposition occurred under semiarid conditions (high pressure, consistent with paleolatitude). The ESI and the overlying Lower Mississippian formations at this stop and elsewhere contain plant debris and coal, indicating that deposition occurred under a humid climate (low pressure, inconsistent with paleolatitude). However, the humid climate that dominated the region during the Devonian-Mississippian transition is consistent with the development of a subpolar low-pressure system that was associated with a polar front along the northern edge of the ice. Aridosols and other features of aridity in Osagean red beds that overlie the Pocono Group indicate that high pressure and aridity returned to the study region once global warming and ice melting caused a poleward migration of the subpolar low-pressure system.

A cold climate in the Appalachian basin also is indicated by the paleoecology of marine fauna (D.K. Brezinski, unpub. data). First, the latest Devonian sea floor fauna in western Pennsylvania and southern New York known from these strata is unusual, even though the epeiric seas of the time were some of the shallowest waters in the Late Devonian and Early Mississippian in the Appalachian basin. The fauna is composed of cold-water siliceous sponges, not shallow warm-water calcareous sponges. This exotic fauna has, for nearly a century, perplexed paleontologists by virtue of the paradox of a demonstrably "deep water" (cold) fauna dominating shallow-water environments. Clearly the Appalachian basin, during this time, contained very cold waters allowing deep water faunas to migrate to shallow waters. Second, there is a clear empirical correlation between brachiopod diversity through the Late Devonian and Early Mississippian of North America and its relation to both sea-level fluctuations and glacial episodes on Gondwanaland. Overall brachiopod diversity dips in the latest Famennian and climbs back to high levels during the Kinderhookian. Copper (1998) recognized that certain families of Late Devonian brachiopods (spiriferids, athyrids, rhynchonelids, and terebratulids) were eurythermal (not affected by temperature changes). When those groups are taken as a percentage of the total brachiopod fauna, it is apparent that the eurythermal groups dominated during the latest Devonian, a time known to have seen glaciation in the southern hemisphere. This glacial episode is coincident with a period of sea-level draw down at the Devonian-Carboniferous boundary. This glacial episode also is coincident with the long-term humid period during the Devonian-Mississippian transition in the Appalachian basin. The succeeding arid climate is coincident with global warming and sea-level rise. As noted above, the long-term humid period during the Devonian-Mississippian transition is best explained by the close proximity of a polar front that was associated with the northern edge of ice as indicated by features contained within the ESI.

Stop 12. U.S. Silica property along Sandy Mile Road.

Lat 39°42.72′ N., long 78°13.82′ W., Hancock, Md., 7.5-minute quadrangle.

Leaders: Dave Brezinski, John Repetski, and Blaine Cecil

Introduction

At this stop we will examine the Silurian-Devonian contact interval. The main units exposed are the Keyser

Formation, which occurs in an abandoned quarry just to the east, and the Lower Devonian Oriskany Sandstone. The Keyser Formation is predominantly a limestone unit and represents a Late Silurian submergence of the Appalachian basin. The Oriskany on the other hand is a carbonate-cemented sandstone, grading laterally in some places into an arenaceous limestone. The Oriskany has been one of the most prolific gas-producing units in the central Appalachians.

Lithostratigraphy

Helderberg Group

Although much of the Helderberg Group is encompassed by this stop, only about 65 m (213 ft) of the lower part of the basal unit, the Keyser Formation, are accessible (Brezinski, 1996). The Keyser Formation is exposed in the abandoned quarry located east of the road. At this location the Keyser Formation is predominantly composed of nodular-bedded, argillaceous, fossiliferous lime wackestone. This interval contains thin (<5 cm) pelmatozoan packstone or grainstone intervals that are commonly graded and exhibit hummocky cross-stratification. The uppermost 10 m (33 ft) of the exposure becomes more regularly bedded and contains noticeably more and thicker layers of grainstone-packstone. The lithologies exposed at this location are included within the nodular facies of Head (1969) and the shaly/nodular-bedded/cherty limestone facies of Dorobek and Read (1986).

Although these strata are highly fossiliferous, most faunal components have been comminuted. The most common faunal components include abundant ramose and fenestrate bryozoans, rugose corals, tabulate corals, trilobites, brachiopods, and pelmatozoans. The regularly bedded intervals near the top of the exposure contain brachiopod coquinas and more pervasive indications of current activity such as crossbedding and graded beds.

On the basis of conodont ranges, the Silurian-Devonian contact can be placed within the upper Keyser near the contact with the overlying New Creek Limestone (Denkler and Harris, 1988). Consequently, the Keyser Formation here is almost totally Silurian (Pridolian) in age. Between the Keyser exposure and the outcrop of the overlying Oriskany Sandstone is a covered interval approximately 35 m (115 ft) thick. On the basis of regional stratigraphic relations this covered interval contains the intervening units of the New Creek Limestone, Corriganville Limestone, Mandata Shale, and Shriver Chert. The New Creek Limestone consists of thick-bedded lime grainstone that is intensely crossbedded. The Corriganville Limestone is characterized by nodular-bedded, shaly limestone. The Mandata Shale and Shriver Chert are composed of dark-gray, calcareous shale, chert, and siliceous shale and siltstone. All of these intervening units are Early Devonian in age.

Oriskany Sandstone

Regionally, the Oriskany Sandstone is underlain and overlain by marine cherty units and black shales, respectively.

Unit	Lithology	Environment	Sea level
Oriskany Sandstone	white sandstone	shallow marine	
Shriver Chert	black shaly siltstone	basinal	
Mandata Shale	black shale	basinal	
Corriganville Limestone	nodular-bedded limestone	deep ramp	
New Creek Limestone	crossbedded encrinites	tidally dominated shelf	
Keyser Formation	nodular-bedded limestone	deep ramp	

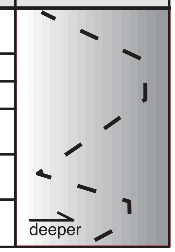


Figure 32. Relations between stratigraphy and sea-level curve at Stop 12. Modified from Dorobek and Read (1986).

These chert-bearing units are indicative of a high silica input into the basin just prior to and immediately following deposition of the Oriskany Sandstone. C.B. Cecil (unpub. data) has attributed the source of the silica to wind-blown dust from coeval nearby warm-arid sand seas. At this location, the Oriskany Sandstone is approximately 50 m (164 ft) thick and consists of light-gray, medium-bedded, fine- to medium-grained, carbonate-cemented quartz sandstone. Crossbedding is prevalent in only a few intervals. To the south in West Virginia, U.S. Silica mines this unit for use in the manufacture of glass. As a consequence of the carbonate cementation, this sandstone is easily leached on exposed surfaces, leaving behind a residue of pure quartz sand.

The Oriskany Sandstone is highly fossiliferous in some intervals. These fossiliferous intervals are concentrated near the top of the formation and the fossils occur in the form of internal and external molds. The faunal components are primarily large, thick-shelled, coarsely plicate brachiopods, and, more rarely, large, platyceratid gastropods. These coarsely plicate brachiopods are indicative of turbulent, current-swept environments. The preponderance of brachiopod valves are disarticulated and convex-up indicating that they have been rearranged post-mortem by traction currents.

Depositional Environments

Dorobek and Read (1986) examined the Helderberg Group from a regional perspective. Their interpretations are summarized in figure 32. From their interpretations it is possible to recognize two separate transgressive-regressive episodes within the Helderberg Group of the central Appalachians. The first of these episodes is manifested in the Late Silurian Keyser Formation. This deepening episode submerged the tidal flat facies of the underlying Tonoloway Limestone as seen at Stop 10 and continued into the latest Silurian, completely drowning the Appalachian basin. At maximum deepening, the center of the basin was the site of deposition of nodular-bedded, argillaceous lime wackestone as seen at this stop. Dorobek and Read (1986) interpreted this facies as a deep ramp deposit formed below fair-weather wave base. The burrowed character of the sediments and the paucity of shallow water features suggested water depths of up to 50 m (164 ft).

Head (1969) also interpreted these lithologies as occurring near the center of the Keyser seaway. The increased numbers of crossbedded intervals near the top of the exposure suggest regression.

The overlying New Creek Limestone marks the base of the second of the two transgressive episodes and is interpreted as a skeletal sand bank that formed in water depths of less than 15 m (49 ft) (Dorobek and Read, 1986). The upsection transition of the New Creek encrinites into the nodular-bedded Corriganville Limestone indicates a continued deepening of the Early Devonian sea. Like the Keyser lithologies at this stop, the Corriganville nodular-bedded lithologies are interpreted as deep ramp deposits. Upsection, the Corriganville Limestone passes into the Mandata Shale and Shriver Chert. These two lithologies represent basinal lithologies that formed at the apex of the Helderberg transgression. Water depths are estimated in excess of 50 m (164 ft) (Dorobek and Read, 1986).

Presumably the regressive phase of this episode resulted in the shallow water deposition of the Oriskany Sandstone. The presence of crossbedding, coarsely plicate and thick-shelled brachiopods, and the disarticulated, convex-up brachiopod valves suggest strong, current-swept conditions.

Paleoclimate

During the Late Silurian and Early Devonian, the Appalachian basin was in the vicinity of lat 30º to 40º S. in a belt analogous to modern-day aridity and high pressure. Paleoclimate indicators suggest that the climate was arid (Scotese, 1998). Unlike the limestone intervals at this stop, the origin of the highly siliceous Lower Devonian strata remains speculative. On a regional scale, the siliceous strata consist of the following in ascending order: (1) the Shriver Chert, (2) the Oriskany Sandstone, and (3) the Huntersville Chert. The Shriver Chert consists predominantly of decimeter-scale interbedded chert and silty limestone. The overlying Oriskany Sandstone contains rounded sand grains floating in a carbonate matrix and rounded to subrounded quartz grains in quartz arenite. Grabau (1932, 1940) and Cecil and others (1991) proposed an eolian provenance for the Oriskany sand; the quartz sand was either blown into the carbonate environments of the Oriskany seaway or the sand was derived from ergs that were reworked during sea-level rise. Shinn (1973) has described a similar set of conditions in the modern Persian Gulf where dust and sand are blown into modern carbonate environments. The Huntersville Chert (not present at this stop) grades from impure chert in West Virginia into the cherty Onondaga Limestone in Pennsylvania and New York (Dennison, 1961). Sheppard and Heald (1984) described the following seven lithotypes in the Huntersville Chert in West Virginia: (1) clean chert, (2) chert with organic material, (3) spicular chert, (4) dolomitic chert, (5) glauconitic chert, (6) silty, argillaceous chert, and (7) dolomitic, silty, argillaceous chert. They pointed out that all lithotypes in the Huntersville contain some dolomite and silt-size (quartz) detritus.

The source of silica for the Shriver and Huntersville Cherts (and equivalent cherty limestones) has previously been attributed to the biotic extraction of silica from seawater derived from the dissolution of volcanic ash (Dennison, 1961; Sheppard and Heald, 1984). Although minor amounts of biotic components have been recognized in the cherts, and volcanic ash (bentonite) occurs near the top of the Huntersville, the lithologies of the siliceous stratigraphic interval considered herein appear to be best explained by eolian sand and dust as the predominant source of silica. The coarsest silt fraction accounts for the quartz silt noted in the cherts as well as the quartz silt in interbedded limestone, whereas the finer size fraction of quartz dust readily provided an ample supply of soluble silica and residual particles that are equivalent in size to quartz crystallites in chert. Thus, the Early and early Middle Devonian sequence in the Appalachian basin can be readily explained by temporal and spatial variations in eolian processes in a warm arid climate. In contrast, it is unclear how the enormous amounts of Devonian chert can be accounted for by either silica derived from dissolution of volcanic ash or biotic extraction of silica from normal seawater.

Stop 13. Ordovician Beekmantown through Chambersburg Limestone.

Lat 39º36.99′ N., long 77º52.99′ W., C&O Canal Towpath, milepost 103, Hedgesville, W. Va., 7.5-minute quadrangle.

Leaders: Dave Brezinski and John Repetski

Introduction

At this stop we are on the western flank of the Massanutten synclinorium, a broad fold that is bordered on the east by the South Mountain anticlinorium and generally coincides with the Great Valley section of the Valley and Ridge physiographic province. The section here begins in the Lower Ordovician Stonehenge Limestone and continues eastward (and upsection) along the C&O Canal through the Lower to lower Middle Ordovician Rockdale Run Formation, the lower Middle Ordovician Pinesburg Station Dolomite into Middle Ordovician St. Paul Group, and Upper Ordovician Chambersburg Limestone. The basal strata of the Upper Ordovician Martinsburg Formation are the uppermost strata exposed. We will concentrate on the lower part of this section, which records the final regressive phase of the Sauk Sequence (Brezinski and others, 1999).

The widespread marine facies that existed in North America during the Cambrian and Early Ordovician is known as the Sauk Sequence (named by Sloss, 1963). Vail and others (1977) proposed that the Sauk Sequence was a first-order deepening episode. Palmer (1981) subdivided the Sauk Sequence into three sequences, Sauk I, Sauk II, and Sauk III.

Each of these subsequences represents a separate transgressive and regressive episode that is interpreted as third-order in magnitude (Brezinski and others, 2002).

Lithostratigraphy

The lowest strata encountered are assignable to the Stonehenge Limestone. The Stonehenge is 150 to 250 m (492–820 ft) thick, although only the upper 30 m (98 ft) are exposed at this location. Three members of the Stonehenge are recognized. The basal member, the Stoufferstown Member, is composed of ribbony limestone interbedded with thin intraclastic grainstones, 3 to 6 cm (1–2 in) thick. The middle member of the Stonehenge, unnamed, consists of thick-bedded, thrombolitic algal limestone and, rarely, ribbony limestone. The upper member is also unnamed and is lithologically similar to the Stoufferstown Member. The contact between the Stonehenge and the overlying Rockdale Run Formation is generally placed at the lowest tan-weathering, laminated dolomitic limestone. This dolomite reflects the onset of peritidal cyclic sedimentation that characterizes the Rockdale Run Formation.

The Rockdale Run Formation is approximately 850 m (279 ft) thick and consists mostly of 1- to 5-m (3–16 ft)-thick peritidal carbonates that are probably fifth-order in magnitude. The lithologic character of these cycles varies upsection with certain intervals containing thick limestone and thin dolomite subcycles and other intervals containing thicker dolomite and thin limestone subcycles (fig. 33A, B).

Sando (1957) recognized three subdivisions of the Rockdale Run Formation in this region. He noted that the basal 30 to 50 m (100–160 ft) contained large, silicified algal masses. Above these basal cherty beds, the Rockdale Run Formation contained abundant oolitic beds. Much higher in the formation a thick dolomite interval is prevalent. Sando (1957, pl. 4) also recognized a number of vertically distributed faunas. These faunas include the *Bellefontia* fauna within the upper Stonehenge Limestone and macrofaunas characterized by *Lecanospira*, *Archaeoscyphia*, *Diparelasma*, and *Syntrophopsis-Clelandoceras* in the Rockdale Run Formation, in ascending order (fig. 34). Hardie (1989) interpreted the faunas within the Stonehenge and lower Rockdale Run Formations to represent large-scale deepening episodes that he believed were third-order in magnitude. Brezinski and others (1999) showed that there were several different orders of sea-level cycles displayed within this stratigraphic section (fig. 34). The variations between limestone-dominated and dolomite-dominated cycles are the result of the large-scale deepening episodes as recognized by Hardie (1989). It appears that the fifth-order cycles, which formed near or at the transgressive apex of the larger scale cycles are limestone dominated, whereas fifth-order cycles that have been interpreted to have formed during regression or at the regressive nadir are dolomite dominated. Our observations in the Rockdale Run, however, indicate that regression was rapid and resulted in exposure surfaces at the top of the subtidal limestone units rather than at the top of the peritidal dolomite. Exposure at the top of the subtidal limestones is indicated by pedogenic brecciation and subaerial crusts (fig. 33B). These subaerial exposure surfaces represent fourth- or fifth-order sequence boundaries. The peritidal dolomite formed during relatively slow transgression and is part of the transgressive systems tract.

Within the dolomite member of the upper Rockdale Run (*sensu* Sando, 1957), cycles are difficult to recognize inasmuch as diagenetic overprinting has obliterated many of the internal features. However, it does seem likely that this dolomite tongue was deposited during a major sea-level lowstand or slowly rising sea level during transgression.

Overlying the Rockdale Run Formation is the Pinesburg Station Dolomite, which is the uppermost formation of the Beekmantown Group. The Pinesburg Station Dolomite is 120 to 160 m (394–525 ft) thick and consists of cherty, laminated dolomite and burrow-mottled dolomite. The Pinesburg Station Dolomite is unfossiliferous, save for conodonts and many stromatolitic intervals. The Pinesburg Station Dolomite is overlain by a Middle Ordovician succession of interbedded limestone and dolomite, termed the St. Paul Group by Neuman (1951). Neuman subdivided the St. Paul Group into two units, the lower Row Park Limestone and the upper New Market Limestone. The Row Park Limestone consists of light-gray, massive, micritic to fenestral limestone with thin interbeds of laminated dolomitic limestone. It is approximately 85 m (279 ft) thick. The overlying New Market Limestone is characterized by interbedded, medium-bedded, light-gray, burrow-mottled limestone; stromatolitic limestone; and gray and tan, laminated dolomite and dolomitic limestone. It is capped by a light- to medium-gray, micritic limestone. The New Market is approximately 67 m (220 ft) thick. Fossils are not common in either of the St. Paul Group Formations; however, they are present within horizons in the upper part of the New Market Limestone. The main macrofaunal components are macluritid snails. Conodonts occur throughout these units. Overlying the St. Paul Group is an interval of medium- to dark-gray, medium- to wavy bedded and even nodular-bedded, shaly, fossiliferous limestone termed the Chambersburg Limestone.

Depositional Environments

The Stonehenge Limestone was deposited during a major deepening episode that was probably third-order in magnitude and appears to exhibit a symmetrical facies distribution (Hardie, 1989; Taylor and others, 1993). The sea-level drop that was concurrent with the regression at the termination of Stonehenge deposition also produced the fifth-order depositional peritidal cycles in the basal Rockdale Run Formation. The peritidal mud flats that existed throughout most of the Rockdale Run deposition were periodically submerged by shallow subtidal waters, indicating that larger

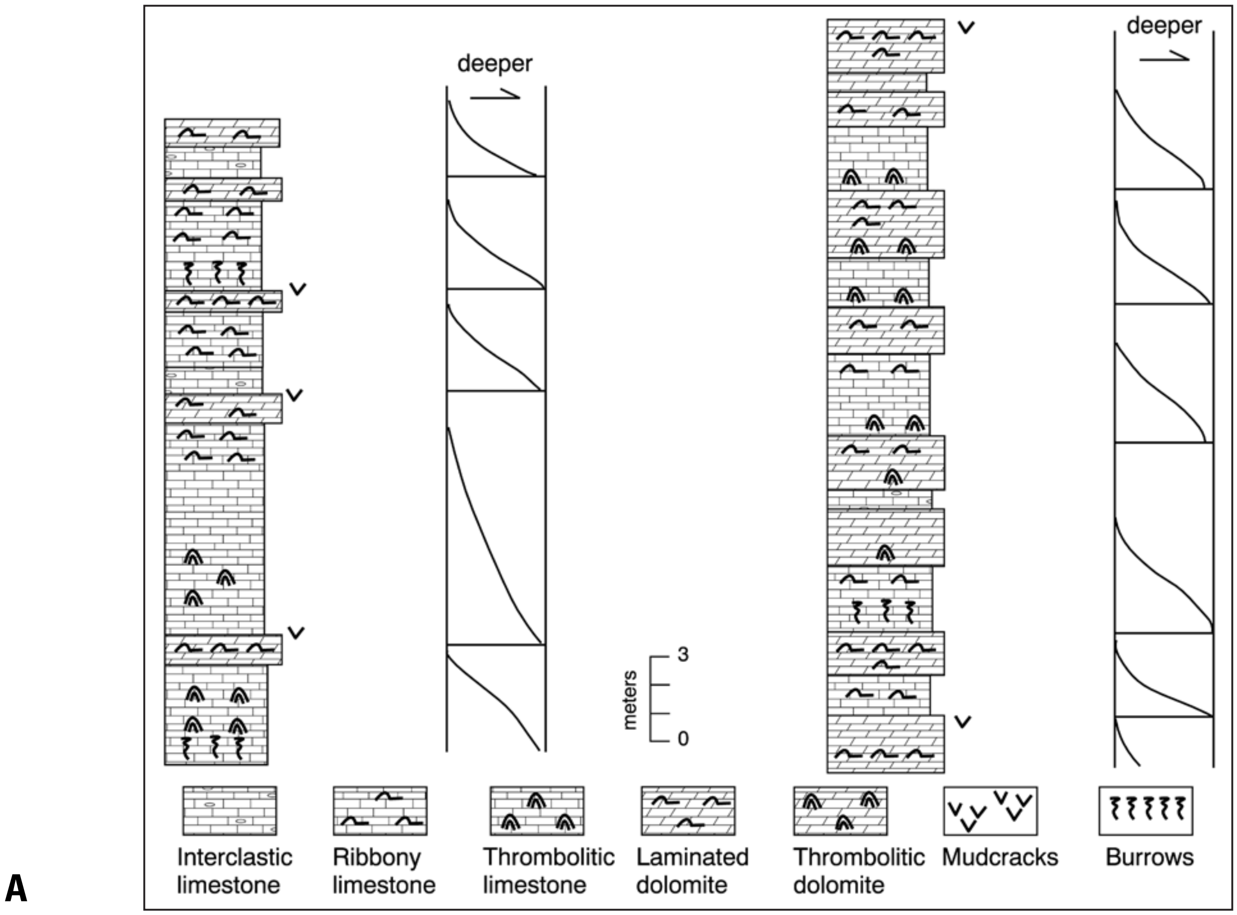


Figure 33. A, Fifth-order(?) cycle variations within the Rockdale Run Formation at Stop 13. Left column shows limestone-dominated subcycles with interpreted sea-level curve (Hardie, 1989). Maximum water depth is during limestone deposition shallowing upward to peritidal conditions during dolomite deposition. Right column shows dolomite-dominated subcycles. B (facing page), Fifth-order(?) cycle variations within the Rockdale Run Formation at Stop 13. Left column shows limestone-dominated subcycles with re-interpreted sea-level curve based on exposure surfaces at the top of the limestone beds. Right column shows dolomite-dominated subcycles.

scale cycles of a fourth- or third-order are superimposed on the sequence. The thick dolomite in the upper part of the Rockdale Run Formation was previously thought to record a significant shallowing episode for the central Appalachians. Our observations, however, indicate that at the scale of the fifth-order lithologic cycles, subtidal limestone deposition was followed by abrupt regression and subaerial exposure. Primary dolomite was deposited on the exposure surfaces during subsequent transgression and deepening. This cyclic sedimentation continued during the deposition of the Pinesburg Station Dolomite at the end of Beekmantown Limestone deposition.

Mitchell (1982) discussed in detail the vertical arrangement of lithologies within the St. Paul Group of Maryland. The vertical arrangement of lithologies within the Pinesburg Station Dolomite to New Market Limestone led Mitchell to interpret the sequence as two separate transgressive and regressive pairs.

The deepening event at the end of the Middle Ordovician of the central Appalachian basin produced the Chambersburg Limestone. This deepening episode resulted from the downwarping that occurred with the onset of the Taconic orogeny. This deepening provided the accommodation space for the graptolitic black shales of the overlying Martinsburg Formation.

Climate of the Early and Middle Ordovician

In contrast to the moist subhumid climate of the Late Precambrian and Early Cambrian as discussed at Stop 14, long-term aridity prevailed during deposition of Early

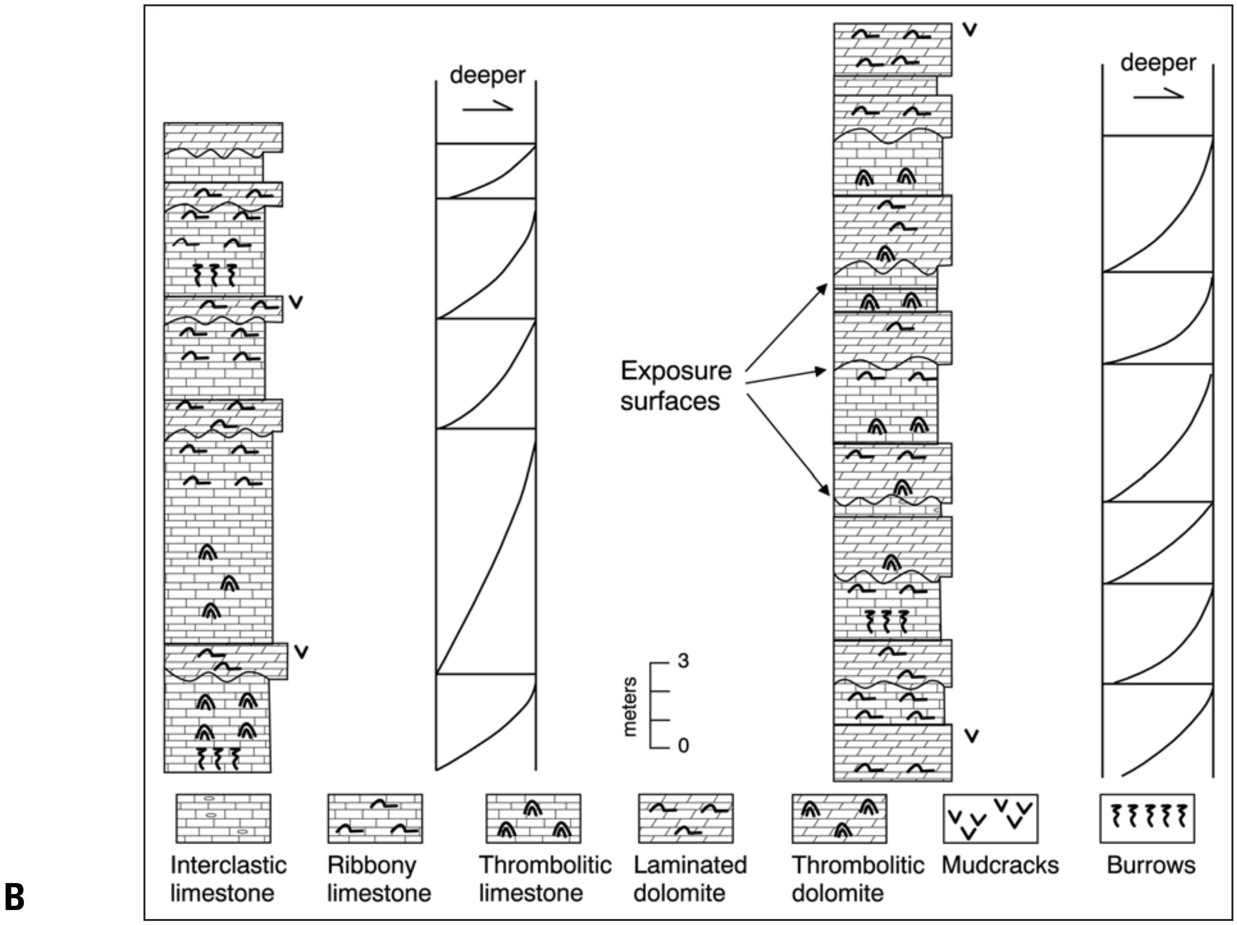


Figure 33. Continued

Ordovician strata (Stop 13). Intermediate- to short-term climates fluctuated from arid to semiarid. Paleogeographic reconstructions indicate that deposition at Stop 13 occurred in the arid high-pressure belt at approximately lat 30° S. Indications for long-term aridity at Stop 2 include (1) the lack of fluvial siliciclastic influx, (2) stratigraphically equivalent evaporites to the west in the depocenter of the basin (Ryder and others, 1992), (3) intervals of nodular chert that presumably are replacements of gypsum or anhydrite, and (4) primary dolomite resulting from probable hypersaline conditions in peritidal environments. Localized evaporate casts within the subjacent Conococheague Formation (Demicco and Mitchell, 1982) corroborate these interpretations.

At a continental scale, the eolian St. Peter Sandstone of the midcontinent also was being deposited during part of this time (coeval with the mid-St. Paul Group and younger strata). Contemporaneous deposition in the sediment starved Ouachita basin consisted of deep water shale, chert, and limestone (Stone and others, 1986, p. 19). Sediment starvation in the deep water Ouachita basin during the Ordovician is consistent with low fluvial sediment supply to continental margins under arid conditions (fig. 1A). Interestingly, the pulses of quartz sand deposition in the Ouachita trough represented by the Crystal Mountain and Blakely Sandstones appear well correlated with the central Appalachian shallowing episodes represented by the initiation of Rockdale Run deposition and the peritidal dolomite deposition of the upper Rockdale Run through Pinesburg Station, respectively.

The interpreted sea-level history at Stop 2 is best explained by a glacial model even though direct evidence for Early and early Middle Ordovician glaciation is lacking (Frakes and others, 1992). The third-order cycle of subtidal deposition of the Stonehenge followed by deposition of the Rockdale Run in peritidal environments may be best explained by greenhouse followed by ice house conditions. The fourth- or fifth-order cycles within the Rockdale Run have inferred sea-level curves (fig. 33B) that are similar to glacial-interglacial cycles. Exposure surfaces at the top of subtidal limestone beds represent fourth- or fifth-order sequence boundaries. Following subaerial exposure in the fourth- or fifth-order cycles of the Rockdale Run, sea level appears to have slowly risen during deposition of the peritidal dolomite followed by subtidal carbonates which, in turn, was followed by a rapid fall and subaerial exposure (fig. 33B). If the third- through fourth-order

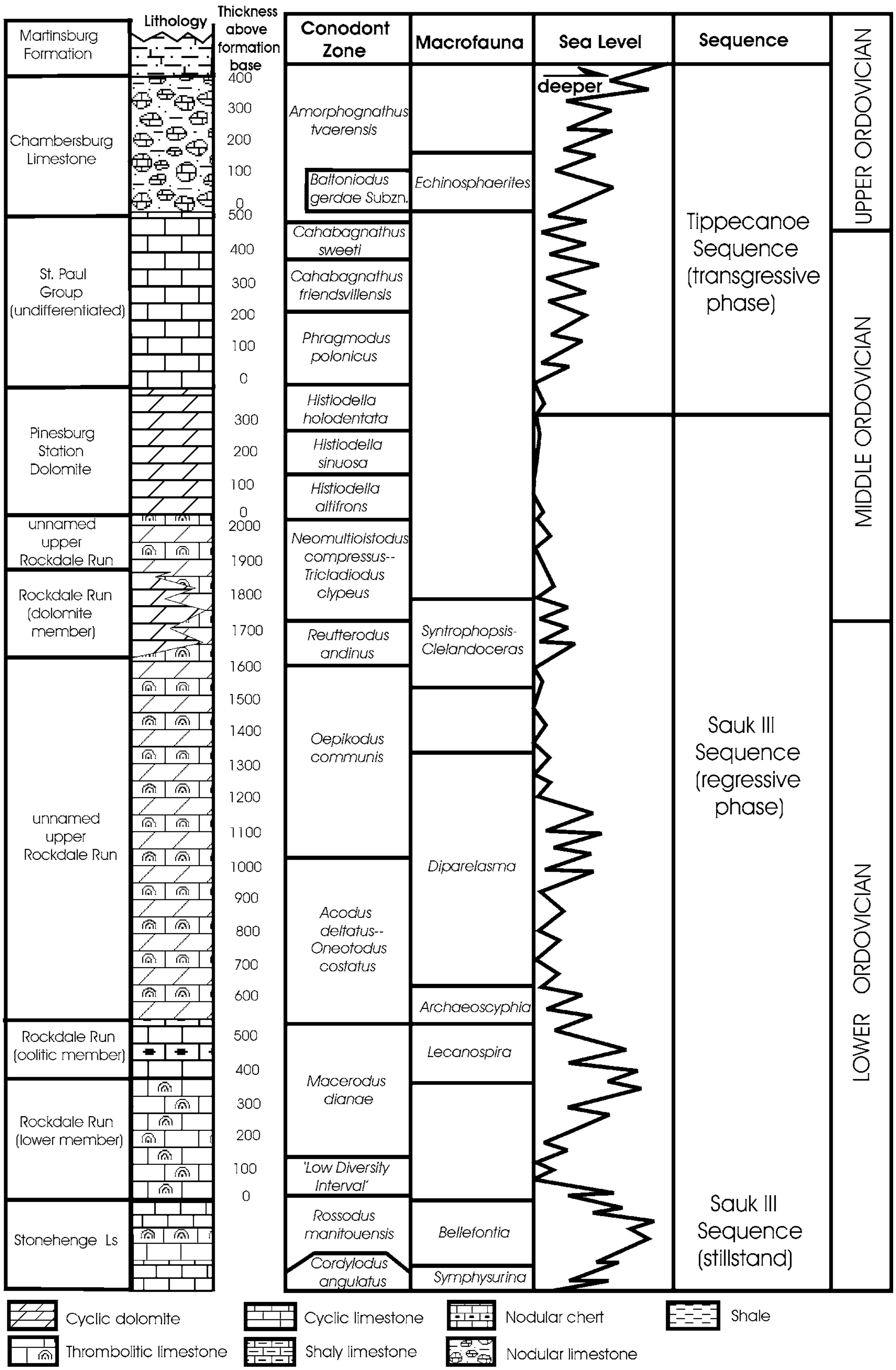


Figure 34. Relations among stratigraphy, faunas, and transgressive-regressive packages for the Early Ordovician Stonehenge and Rockdale Run Formations. Modified from Brezinski and others (1999).

cycles were driven by the presence or absence of continental glaciation, then the third-order Stonehenge deepening is likely the result of greenhouse conditions, whereas the fourth- and fifth-order cycles of the Rockdale Run and Pinesburg Station may be the product of long-term ice house conditions and intermediate- to short-term glacial cycles and the accompanying sea-level and climate cycles.

From a practical perspective, climate considerations become important in predicting porosity and permeability in carbonates and sandstones. Primary dolomitization in carbonates as well as texturally and mineralogically mature quartz arenites deposited by eolian processes may be the result of arid conditions. Both dolomitized limestone and quartz arenites (Stops 3, 6) are often targeted as potential petroleum reservoirs.

Stop 14. Late Precambrian and Early Cambrian Chilhowee Group on U.S. 340, Weverton, Md.

Lat 77°40.47′ N., long 39°19.29′ W., Harpers Ferry, W. Va., 7.5-minute quadrangle.

Leader: Dave Brezinski

Introduction

At this stop we will examine the Chilhowee Group of the northern Blue Ridge physiographic province. The Chilhowee Group, the basal Cambrian clastic unit in the northern Blue Ridge, is a main ridge-forming unit of the Blue Ridge. The Blue Ridge of northern Virginia, Maryland, and Pennsylvania consists of a large overturned fold called the South Mountain anticlinorium (Cloos, 1971). The South Mountain anticlinorium contains Grenville basement (1.1 b.y.) at its core and is flanked by Chilhowee strata. The eastern flank, known as Catoctin Mountain in Maryland and the Bull Run Mountains in Virginia, is the normal limb of the South Mountain anticlinorium and dips southeastward at approximately 35°. The western limb of the anticlinorium, known as South Mountain in Maryland and Short Hill in Virginia, is overturned and also dips to the southeast, but at a much steeper angle than the eastern limb (Catoctin Mountain). The Grenville basement complex is overlain, in ascending order, by Late Proterozoic sediments of the Swift Run Formation, basaltic lava flows of the Catoctin Formation, and clastics of the Chilhowee Group. These units are generally termed the Blue Ridge cover sequence. Both the Swift Run and Catoctin Formations exhibit considerable syndepositional thickness variation. Their geometries suggest that they were deposited in linear basins presumably formed by attenuation of the underlying basement complex (Schwab, 1972). Most of the Chilhowee units, on the other hand, appear to be more or less continuous throughout the northern Blue Ridge, indicating that much of the Late Precambrian topography had been filled prior to its deposition.

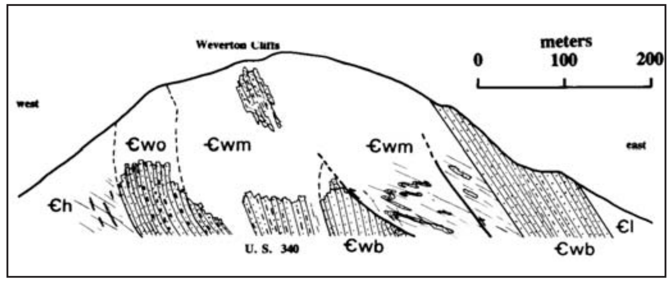


Figure 35. Diagrammatic sketch showing stratigraphy and structure of the Weverton Formation at the Maryland Heights section at the terminus of South Mountain, Md. Ꞓwb, Buzzard Knob Member; Ꞓwm, Maryland Heights Member; Ꞓwo, Owen Creek Member, all of the Weverton Formation. Ꞓh, Harpers Formation; Ꞓl, Loudoun Formation.

At this stop we will examine the type section of the Weverton Formation at the southern terminus of South Mountain and discuss the Sauk I transgression and its initial deposition within the Appalachian basin (fig. 35).

Lithostratigraphy

The Chilhowee Group in the northern Blue Ridge consists of the following in ascending order: the Loudoun, Weverton, Harpers, and Antietam Formations. These units generally overlie the Catoctin Formation, but at this location the Catoctin is absent, and the Chilhowee units sit directly on top of the Grenville basement strata. The basal unit of the Chilhowee Group, the Loudoun Formation, consists of intercalated black tuffaceous phyllites and polymictic conglomerate. The Loudoun Formation exhibits considerable variation in thickness and extent. This patchy distribution has led some workers to abandon the name Loudoun and assign the conglomeratic intervals to the Weverton Formation and the tuffaceous phyllites to the underlying Catoctin Formation (see Nunan, 1979). The Loudoun can vary in thickness from as little as 2 m (7 ft) to more than 70 m (230 ft). At Stop 14 the Loudoun is represented by several loose blocks of gray, tuffaceous phyllite, and it is estimated that more than 30 m (98 ft) of the formation is present.

Weverton Formation

The Weverton Formation, unlike its subjacent units, is relatively continuously distributed throughout the northern Blue Ridge. The Weverton has been subdivided into three members in ascending order as follows: (1) Buzzard Knob, (2) Maryland Heights, and (3) Owens Creek (fig. 35).

Buzzard Knob Member

The Buzzard Knob Member of the Weverton Formation is the single most significant ridge-forming unit in the northern Blue Ridge. The Buzzard Knob Member consists of medium-bedded, light-gray to greenish-gray, chloritic, coarse-grained to very coarse grained, subarkose to quartzose arenite. The crossbedding within this unit is poorly defined and consists mainly of tabular to planar sets as well as horizontally stratified beds. At this location the basal 10 m (33 ft) consist of a very coarse grained sandstone to granular, feldspathic arenite. The upper 35 m (115 ft) comprise a resistant, light- to medium-gray, coarse-grained quartzite with dusky-blue, grayish-olive, and grayish-yellow bands. Trough crossbedding is much more prevalent in this part of the unit, although crossbed foresets are restricted to individual beds. Separating the two resistant intervals is approximately 10 m (33 ft) of interbedded, light-gray to dusky-yellow, platy, micaceous sandstone and light-olive-gray, quartzose siltstone. This interval weathers more readily than do the surrounding quartzite layers, and thus gives the member an apparent bifurcation that is manifested in less well-exposed areas of the two separate arenite units.

Maryland Heights Member

The middle member of the Weverton Formation, named the Maryland Heights Member, is difficult to accurately and thoroughly describe, and varies from 100 to 150 m (328–492 ft) in thickness. The unit is strongly deformed and less resistant to weathering than the massive units that make up the upper and lower parts of the Weverton. The Maryland Heights Member is generally composed of alternating medium-dark-gray, conglomeratic graywacke and dark-gray phyllite and metasiltstone. The incompetent phyllite and metasiltstone, being situated stratigraphically between the underlying and overlying resistant members, are deformed to the degree that stratigraphic characteristics are obscured.

At this location the Maryland Heights Member consists of olive-gray, grayish-black, and olive-black, quartzose metasiltstones that contain thin intervals, 0.33 to 0.6 m (1–2 ft) thick, of conglomerate with white quartz pebbles and 1- to 10-m (3–33 ft)-thick intervals of gray and greenish-gray arenite and graywacke. The thicknesses of the metasiltstone intervals are obscured by isoclinal folds and pervasive foliation. Near the middle of the unit is a 25-m (82-ft)-thick interval of medium-gray quartzite, which is nearly identical to the upper resistant quartzite of the underlying Buzzard Knob Member exposed at another point in the roadcut. Folding can be observed within these units. This quartzite, which is truncated above road level by a small fault, has been interpreted to be a fragment of the Buzzard Knob Member (Southworth and Brezinski, 1996; fig. 35).

Owens Creek Member

The upper member of the Weverton consists of another ledge-forming quartzite termed the Owens Creek Member (Brezinski, 1992). Although this member is more resistant and much better exposed than the Maryland Heights Member, rarely does it form the prominent ledges and ridge crests common to the Buzzard Knob Member. At this location, more than 35 m (115 ft) of the Owens Creek Member crops out. The lower 25 m (82 ft) consist of medium- to medium-dark-gray, medium-bedded, coarse-grained, crossbedded conglomeratic sandstone with many interbeds of dark-gray, sandy siltstone as well as conglomerate, which contains white, pink, and bluish quartz pebbles. Many of the beds in this interval fine upward. The upper 7 to 10 m (23–33 ft) are composed of medium-light-gray to greenish-gray, medium-bedded, coarse-grained sandstone that is typically trough crossbedded. The contact with the overlying Harpers Formation is not exposed at this location. From other locations the contact is gradational, with progressively thicker interbeds of shaly Harpers punctuating the sandstones of the upper Weverton.

Harpers Formation

Overlying the Weverton Formation is an interval of shale, siltstone, sandstone and quartzite called the Harpers Formation. The Harpers Formation is characterized by dark-gray to olive-black, medium-grained sandstone and siltstone in the lower 30 to 50 m (98–164 ft). Above this basal sandy interval, the Harpers consists of greenish-black to brownish-black, highly cleaved siltstone, fine-grained sandstone, and some silty shale. Somewhat higher in the stratigraphic section the first distinct trace fossils are present. These are mainly in the form of the vertical burrow *Skolithos*.

Several small exposures of the lower Harpers Formation are evident to the west. However, these are intensely sheared owing to the proximity of these exposures to a large thrust fault that is present approximately 200 m (656 ft) to the west of the Weverton-Harpers contact.

Depositional Environments

No detailed depositional studies have been conducted on the Chilhowee strata of the northern Blue Ridge of northern Virginia and Maryland. However, the similarity of vertically juxtaposed lithologies between the Weverton at this location and the correlative Unicoi Formation of central and southern Virginia allows extrapolation of correlative depositional processes. The following discussion is based on the work of Schwab (1972) and Simpson and Eriksson (1989) and their studies on the coeval Unicoi Formation of Virginia. From these studies it can be inferred that the Chilhowee of the northern Blue Ridge represents a preserved rift-to-drift sequence. Moreover, the deepening episode that is preserved within these strata represents the initial submergence of the North American craton during the early Paleozoic and, thus, marks the onset of the Sauk transgression.

The underlying Swift Run and Catoctin Formations represent rift sedimentation and volcanism that filled grabens within the basement complex during extension (Wehr and Glover, 1985). The lack of Catoctin basalts at this location indicates that this site was located on an erosional horst between the

Table 3. Interpreted stratigraphic relations at Stop 14 and proposed depositional environments extrapolated from Schwab (1971, 1972), Simpson and Eriksson (1989) and interpreted tectonic regimes from Wehr and Glover (1985, table 2) and Fichter and Diecchio (1986, fig. 1)

Unit	Lithology	Environment	Tectonic regime
Harpers Formation	Dark silty shale and thin sandstone	Shallow marine	Transgressive
Weverton Formation	Crossbedded arkosic and conglomeratic sandstones	Meandering fluvial Braided fluvial	Trailing margin deposition
Loudoun Formation	Tuffaceous shale and conglomerate	Alluvial fan–rift valley	
Catoctin Formation	Basaltic lava flows	Extensional, rift volcanics	Extension and rifting
Swift Run Formation	Arkosic sandstone and varved shales	Rift valley	

southern Pennsylvanian graben and the central Virginia graben. The patchy distribution of the Loudoun Formation, as well as the apparent interfingering of the tuffaceous shales of the formation with the lavas of the underlying Catoctin Formation, suggests that these two formations may have been, in part, coeval. The localized distribution of the cobble conglomeratic facies of the Loudoun appears to reflect localized depocenters, probably within small fault-created basins in the underlying Proterozoic strata. The coarseness of these strata suggests deposition by high-gradient streams. An alluvial fan environment of deposition is suggested for this unit.

The planar bedded and trough crossbedded facies of the Buzzard Knob Member of the Weverton Formation appears to be similar to facies association C of the Unicoi Formation (Simpson and Eriksson, 1989). These lithologies were interpreted to be consistent with a distal facies of a braided fluvial plain.

The interbedded shale and thin sandstone lithologies of the Maryland Heights Member are interpreted to represent alluvial plain deposition. In this scenario the shaly strata, including the thin (<1 m; <3 ft) sandy conglomerates, are suggestive of overbank deposits and the thicker sandstone (>10 m; >33 ft) are channel deposits. The Owens Creek Member is much coarser grained and exhibits thicker, trough crossbedded sandstones than does any subjacent unit of the Weverton. A similar relation is present in the upper Unicoi of Virginia. This association of lithologies indicates deposition in a shoreface or tidal setting (Simpson and Eriksson, 1989). Schwab (1972) interpreted this part of the stratigraphic section to be fluvial channel sandstones.

The Harpers Formation exhibits a vertical arrangement of lithologies that suggests a transgressive relation (Schwab, 1971, 1972). The lower Harpers contains many, *Skolithos*-burrowed sandstone intervals, which are suggestive of littoral and sublittoral deposition (Brezinski, 1992). The increase in shaly strata upsection suggests a progression from nearshore to more offshore deposition within the lower Harpers.

Tectonics and Depositional Sequences

The late Precambrian to Early Cambrian depositional pattern of the central Appalachians suggests that the late Precambrian volcaniclastic and overlying Chilhowee Group represents the progression of depositional environments from rift basin alluvial fan (Loudoun) to braided and meandering fluvial (Weverton) to nearshore marine (upper Owens Creek Member, lower Harpers) and deeper shelf environments (Harpers) (table 3). This vertical sequence and the resulting environmental progression indicates a depositional onlap. The resulting transgressive episode, which began in the latest Precambrian, is equivalent to the Sauk I Sequence of Palmer (1981) and Sequence 1 of Read (1981). This apparent third-order eustatic event and the late Precambrian rift-basins and their accumulated sediments, resulted in the transitions from rift sedimentation to trailing margin sedimentation (Fichter and Diecchio, 1986).

Climate of the Late Precambrian and Early Cambrian

Climate interpretations of the late Precambrian and the Early Cambrian presented herein are primarily based on interpretations of lithologies and the depositional environment interpretations cited above. The massive influx of siliciclastic material that comprises the Chilhowee Group is consistent with a moist subhumid climate (fig. 1A, table 1). Such a climate is compatible with a paleogeographic location of approximately lat 40° S. (Scotese, 1998). The subarkosic sediments within the Weverton Formation are indicative of a dryer, perhaps subarid climate. However, the paucity of calcareous materials in Chilhowee strata indicates a subhumid or wetter climate setting where fluvial systems were low in dissolved solids precluding syndepositional precipitation of calcareous materials.

Field Trip Summary and Discussion

This trip traversed and included stops in Paleozoic strata from the latest Pennsylvanian to the earliest Cambrian within the central Appalachian basin. The Early Cambrian illustrates the tectonic influence of rifting and siliciclastic sediment supply in response to a long-term, dry subhumid to subhumid climate (table 1), which is consistent with a paleolatitude location of approximately 40° S. In contrast, Early Ordovician strata at Stop 13 consist of carbonates that were deposited in response to allocyclic changes in sea level. Sea level fluctuated from subtidal to total withdrawal with protracted subaerial exposure. By the Early Ordovician, the region was located in the high-pressure climate zone near lat 30° S., which is consistent with inferred climate ranges of arid to semiarid conditions.

As a result of the Taconic orogeny, the Appalachian foreland basin was well developed by the Late Ordovician and the red continental siliciclastic strata of the Juniata Formation (Stop 9) were deposited in response to semiarid to dry subhumid conditions. The depositional hiatus at the end of the Ordovician was succeeded by transgression and deposition of the Early Silurian Tuscarora Sandstone. The origin of the Tuscarora (Stop 9) remains enigmatic even though most workers agree that both marine and fluvial processes contributed to its deposition; neither process adequately explains the textural and mineralogical maturity of the Tuscarora. This maturity, however, can be attributed to mechanical weathering by eolian processes in an arid environment, which resulted in sand seas that were reworked and redeposited by hydraulic processes during sea-level rise in the Early Silurian.

The long-term climate remained arid throughout the remainder of the Silurian and into the Devonian. Regional-scale Upper Silurian peritidal carbonates (Stop 10) with abundant halite casts are indicative of these dry conditions. Latest Silurian and earliest Devonian sea-level rise resulted in subtidal carbonate deposition (Stop 3) throughout the central Appalachian basin. Such deposition is consistent with long-term aridity and very low fluvial siliciclastic sediment supply.

The textural and mineralogical maturity of the Lower Devonian Oriskany Sandstone (Stop 12) may be best explained by continuation of Silurian long-term aridity and mechanical weathering by eolian processes. The lithology of the Oriskany, which regionally ranges from a pure quartz arenite to an arenaceous limestone, may indicate the predominance of eolian transport of sand into the carbonate environments of the Early Devonian seaway. Similarly, the Lower Silurian Tuscarora Sandstone was the result of transgressive reworking of an eolian regolith. Both the Tuscarora and the Oriskany owe their textural maturity to eolian processes.

Following deposition of the Oriskany Sandstone, there was an apparent deepening of the Appalachian foreland basin as indicated by widespread deposition of finer grained siliciclastic materials, including black shale. The deepening of the foreland basin was likely in response to the Acadian orogeny. Abundant terrestrial organic matter in some black shale facies appears to indicate high terrestrial organic productivity in response to a wetter paleoclimate. Following black shale deposition, basin filling materials coarsened upward, culminating in a major influx of sand in the Late Devonian and Early Mississippian (Stops 4, 8). The widespread distribution of this sand, from New York to Tennessee, is indicative of a climate control on sediment supply. The amount of sand and the types of paleosols, including coal, associated with these sands indicates that the long-term climate was moist subhumid. Based on the occurrence of coal and variation in siliciclastic sediment supply, the intermediate- to short-term climate fluctuations ranged from dry subhumid to humid during the Devonian-Mississippian transition in response to the proximity of glacial ice.

The influx of sand in the Early Mississippian was followed by a return to aridity as evidenced by deposition of evaporites throughout much of North America including southwest Virginia. Coeval terrestrial red beds containing Aridisols occur in southern West Virginia (informally referred to as the Pocahontas basin). Late Mississippian transgression under these arid conditions resulted in deposition of the overlying Greenbrier Formation. In the vicinity of the field trip route, the lowstand Loyalhanna Limestone Member eolianite(?) further indicates Late Mississippian aridity. Latest Mississippian strata are missing in the vicinity of the field trip route as a result of the worldwide mid-Carboniferous unconformity. However, relatively rapid subsidence in the Pocahontas basin resulted in nearly continuous deposition during the Late Mississippian and Early Pennsylvanian. In the Pocahontas basin, Late Mississippian deposition was only interrupted by eustatic drops in sea level that resulted in lowstand pedogenesis of regional subaerial exposure surfaces. These paleosols record both eustatic cycles and climate cycles that ranged from semiarid to humid. The influx of sediment and deposition of Late Mississippian red beds reflects the onset of global ice house conditions that spanned the Devonian-Mississippian transition.

Although Lower Pennsylvanian strata also are missing in the vicinity of the field trip route, deposition in the Pocahontas basin recorded a fluctuating water table, probably related to eustatic changes in sea level, and climate cycles that were much wetter than those of the Late Mississippian. Paleosols in the Early Pennsylvanian of the Pocahontas basin, as well as those at the Mississippian-Pennsylvanian unconformity (Stop 11), indicate that the climates of the Early Pennsylvanian ranged from humid to perhumid. These climate cycles and sea-level cycles appear to have been driven by glacial-interglacial conditions.

By the late Middle Pennsylvanian (Desmoinesian of the United States; Westphalian D of Europe), the long-term climate shifted from humid to moist subhumid conditions. Intermediate- to short-term climate cycles ranged from perhumid to dry subhumid conditions as evidenced by a marked but cyclic change in the sedimentary geochemistry of nonmarine strata. Dry subhumid conditions are indicated by increased siliciclastic flux, abundant pyrite and calcite, including nonmarine limestone (Stops 10, 12). In contrast, humid to perhumid conditions produced chemically weathered and highly leached mineral paleosols and coal beds during lowstands.

A pronounced long-term dry subhumid climate developed in the early Late Pennsylvanian (Missourian) as evidenced by lowstand calcic-Vertisols, calcareous red beds, and far fewer economic coal deposits (Stop 13). A shift toward moist subhumid conditions during deposition of uppermost Pennsylvanian strata (Virgilian) in the central Appalachian basin resulted in increased coal formation (Stop 14) under moist subhumid conditions. Peat developed in topographic lows while coeval Ultisols formed around the basin margins. Moist subhumid conditions were followed by dry subhumid conditions that led to clastic influx and burial of the underlying peat. The driest parts of climate cycles, which were dry subhumid to semiarid, were conducive to nonmarine limestone deposition in topographic lows in the basin, while coeval calcic-Vertisols formed around the basin margins. Climate change, therefore, was the primary control on the lithostratigraphy of Upper Pennsylvanian strata in the Appalachian basin.

Paleoclimate and the Origin of Paleozoic Quartzose Sandstones

The origin of quartz-rich sandstones in Paleozoic strata in the Appalachian basin is generally attributed to multicycles of sedimentation, reworking within aqueous depositional environments, and (or) diagenesis. However, weathering within tropical paleoclimates appears to provide a explanation for the textural and mineralogical maturity of Paleozoic quartz arenites. From the Ordovician through the Early Devonian, the Eastern United States was in the southern hemisphere tropical dry belt and moved northward into the equatorial tropical rainy belt in the latest Mississippian and Pennsylvanian. Texturally mature quartz and trace amounts of unaltered detrital feldspars in Late Cambrian through Devonian quartz arenites are indicative of mechanical weathering by eolian processes in an arid or semiarid climate rather than chemical weathering in a humid tropical climate (Cecil and others, 1991). Grabau (1940, p. 220) suggested that the Oriskany Sandstone was an eolian deposit that was reworked by a marine transgression. An eolian component may be equally viable for the origin for the Silurian Tuscarora Sandstone as noted at Stop 9. Furthermore, the Oriskany and Tuscarora are associated with other strata that appear to be the result of deposition under arid conditions. Although both the Oriskany and parts of the Tuscarora Sandstones were deposited in aqueous environments, they may have been blown into a marine environment analogous to the modern Persian Gulf where massive amounts of sand are being blown into a marine carbonate environment (Shinn, 1973).

In contrast, Late Mississippian and Pennsylvanian quartz arenites in the Appalachian basin tend to be texturally immature and nearly devoid of feldspars and, therefore, appear to be products of chemical weathering under humid conditions rather than mechanical weathering under arid conditions. Regional occurrences of residual kaolin deposits of latest Mississippian and Early Pennsylvanian age, which result from chemical weathering in humid tropical environments, are consistent with this interpretation. Thus, mature sandstones in Cambrian through Devonian strata may, in part, be the result of mechanical processes in eolian environments, prior to deposition in aqueous systems, whereas chemical weathering appears to have been a primary factor in the genesis of quartz arenites in Mississippian and Pennsylvanian strata.

Conclusions

By the examples included herein, we have attempted to illustrate the relative importance of allocyclic and autocyclic processes as controls on sedimentation and stratigraphy in the central Appalachian basin. Allocyclic processes appear to be the dominant control on lithostratigraphy. Autocyclicity seems to control facies relations within allocyclic packages. Clearly long-term tectonic subsidence was a primary control on accommodation space. Tectonic influence on fluvial sediment supply, while commonly suggested, is unclear and may be highly over emphasized. Intermediate- to short-term eustatic changes in sea level were also important controls on accommodation space but probably had little influence on sediment supply. Global long- to short-term changes in climate, however, were the primary controls on variation in sediment supply, both chemical and siliciclastic, and the predominant control on lithostratigraphy.

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PLATE FOLLOWS

PLATE 1.

[Conodonts from the Deer Valley and Loyalhanna Limestone Members of the Mauch Chunk Formation]

FIGURES 1, 2. Conodonts from the Deer Valley Member of the Mauch Chunk Formation, Keystone quarry, Pa. This collection (93RS–79c) is from the lower 10 cm of the Deer Valley Member. Note the non-abraded, although slightly broken, conodont elements of the high-energy oolitic marine facies of the Deer Valley Member.

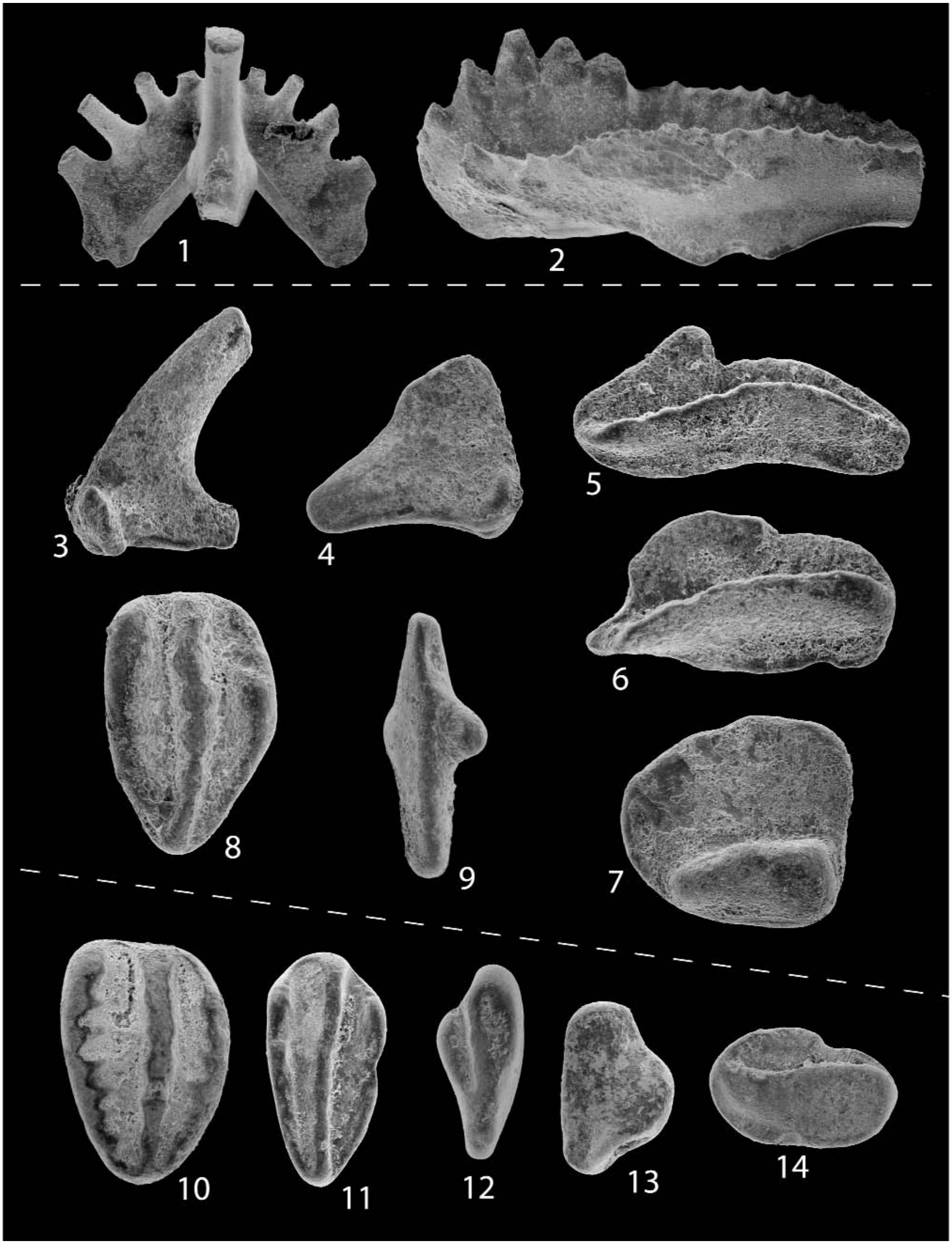
1. *Kladognathus* sp., Sa element, posterior view, X140.
2. *Cavusgnathus unicornis*, gamma morphotype, Pa element, lateral view, X140.

3–9. Conodonts from the uppermost Loyalhanna Limestone Member of the Mauch Chunk Formation, Keystone quarry, Pa. This collection (93RS–79b) is from the upper 10 cm of the Loyalhanna Member. Note the highly abraded and reworked aeolian forms.

3, 4. *Kladognathus* sp., Sa element, lateral views, X140.
5. *Cavusgnathus unicornis*, alpha morphotype, Pa element, lateral view, X140.
6, 7. *Cavusgnathus* sp., Pa element, lateral view, X140.
8. *Polygnathus* sp., Pa element, upper view, reworked Late Devonian to Early Mississippian morphotype, X140.
9. *Gnathodus texanus*?, Pa element, upper view, X140.

10–14. Conodonts from the basal 20 cm of the Loyalhanna Limestone Member of the Mauch Chunk Formation, Keystone quarry, Pa. (93RS–79a), and Westernport, Md. (93RS–67). Note the highly abraded and reworked aeolian forms.

10. *Polygnathus* sp., Pa element, upper view, reworked Late Devonian to Early Mississippian morphotype, 93RS–79a, X140.
11. *Polygnathus* sp., Pa element, upper view, reworked Late Devonian to Early Mississippian morphotype, 93RS–67, X140.
12. *Gnathodus* sp., Pa element, upper view, reworked Late Devonian(?) through Mississippian morphotype, 93RS–67, X140.
13. *Kladognathus* sp., M element, lateral views, 93RS–67, X140.
14. *Cavusgnathus* sp., Pa element, lateral view, 93RS–67, X140.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