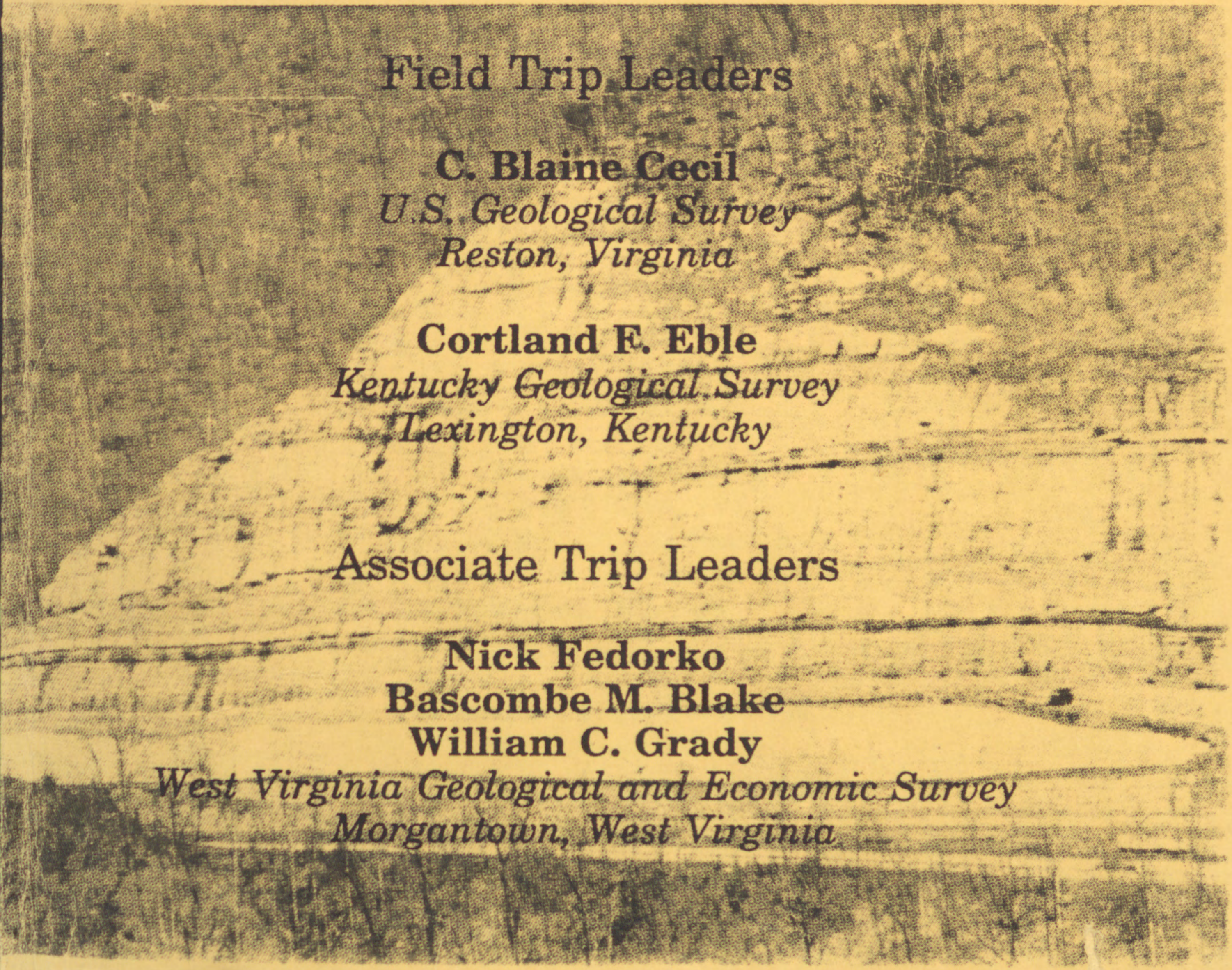
Paleoclimate Controls on Carboniferous Sedimentation and Cyclic Stratigraphy in the Appalachian Basin



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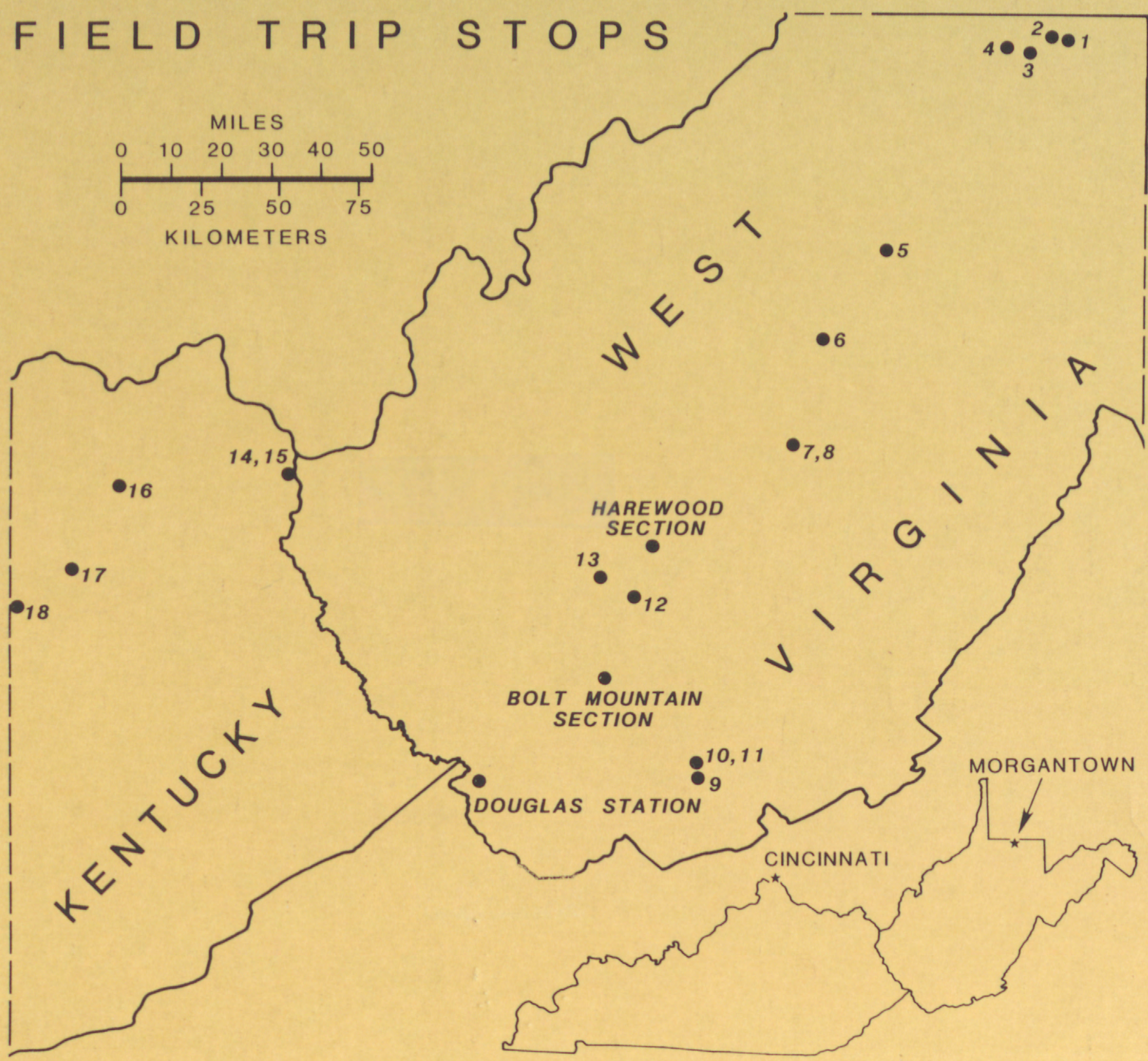
FIELD TRIP STOPS



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This report is preliminary and has not been reviewed for conformity with USGS editorial standards and stratigraphic nomenclature.

PREFACE

This guidebook was prepared for a field trip, sponsored by the Coal Division of the Geological Society of America, held prior to the 1992 Annual Meeting of the Geological Society of America in Cincinnati, Ohio. The purpose of the trip is to examine evidence for autocyclic and allocyclic process controls (Beerbower, 1962) on Carboniferous cyclic stratigraphy and sedimentation in the Appalachian basin. Unlike many approaches to stratigraphic analysis, this approach incorporates factors that, not only control accommodation space, but also control fluvial chemical and siliciclastic sediment flux to depocenters. The effects of allocyclic processes on stratigraphy will be distinguished from those of autocyclic processes through basin scale analyses of selected stratigraphic horizons. Stratigraphic horizons that are known to be of basin scale are interpreted to be the result of allocyclic processes. Such horizons include well-developed paleo-Vertisols and Ultisols, marine transgressive strata, and strata associated with the Mississippian-Pennsylvanian systemic boundary. In contrast, poorly-developed and laterally discontinuous paleosols, which appear to be associated with aggrading fluvial systems, are presented as examples of formation by autocyclic processes.

In so far as is possible, the stratigraphic intervals highlighted on the trip will be put into the context of a continental-scale framework. Such a framework may ultimately lead to high-resolution paleotectonic, eustatic, and paleoclimatic reconstructions.

Evidence for the allocyclic effects of paleoclimate change as a major control on sediment flux, both chemical and siliciclastic, will be illustrated by examining both spatial and temporal distribution of chemical and siliciclastic strata. Syngenetic or early diagenetic geochemical signatures, such as a variety of paleosol types (including coal beds), sedimentary iron-bearing minerals, and nonmarine calcareous strata, will illustrate stratigraphic changes in both physical and chemical conditions of sedimentation. Discussions of allocyclic tectonic and eustatic processes will focus on accommodation space and transgressive-regressive cycles.

The objective of the field trip, therefore, is to evaluate climatic, eustatic, and tectonic controls on sedimentation and stratigraphy with emphasis on paleoclimate. The objective will be accomplished by examining:

1) specific stratigraphic intervals that illustrate basin-scale effects of changes in allocyclic processes as compared to the more local autocyclic response. These intervals included the 1) Upper Pennsylvanian Pittsburgh coal bed and associated strata 2) the Upper Pennsylvanian Ames marine shale and associated strata, 3) the Mississippian-Pennsylvanian Systemic boundary and associated strata, and 4) a paleo-Vertisol and numerous Inceptisols in the red member of the Upper Mississippian Bluestone Formation.

2) Late Mississippian through Late Pennsylvanian geochemical signatures, both stratigraphic and regional, that indicate climatic controls on chemical weathering, sediment flux, and geochemical conditions of sedimentary systems. Such signatures include 1) the oxidation state and compounds of iron, 2) mineralogy and other characteristics of paleosols and related strata that indicate the geochemical conditions of weathering and soil formation, and 3) carbonates of Ca, Mg, and Fe in both marine and nonmarine depositional settings.

3) Characteristics of Pennsylvanian coal beds that indicate conditions of paleopeat formation (ombrogenous domed vs topogenous planar) as a function of paleoclimate. Important paleoenvironmental indicators include the relative abundance and distribution of vitrinite and inertinite macerals, as well as ash yield and sulfur content.

4) Late Mississippian through Late Pennsylvanian sediment flux and lithostratigraphy as a function of paleoclimate.

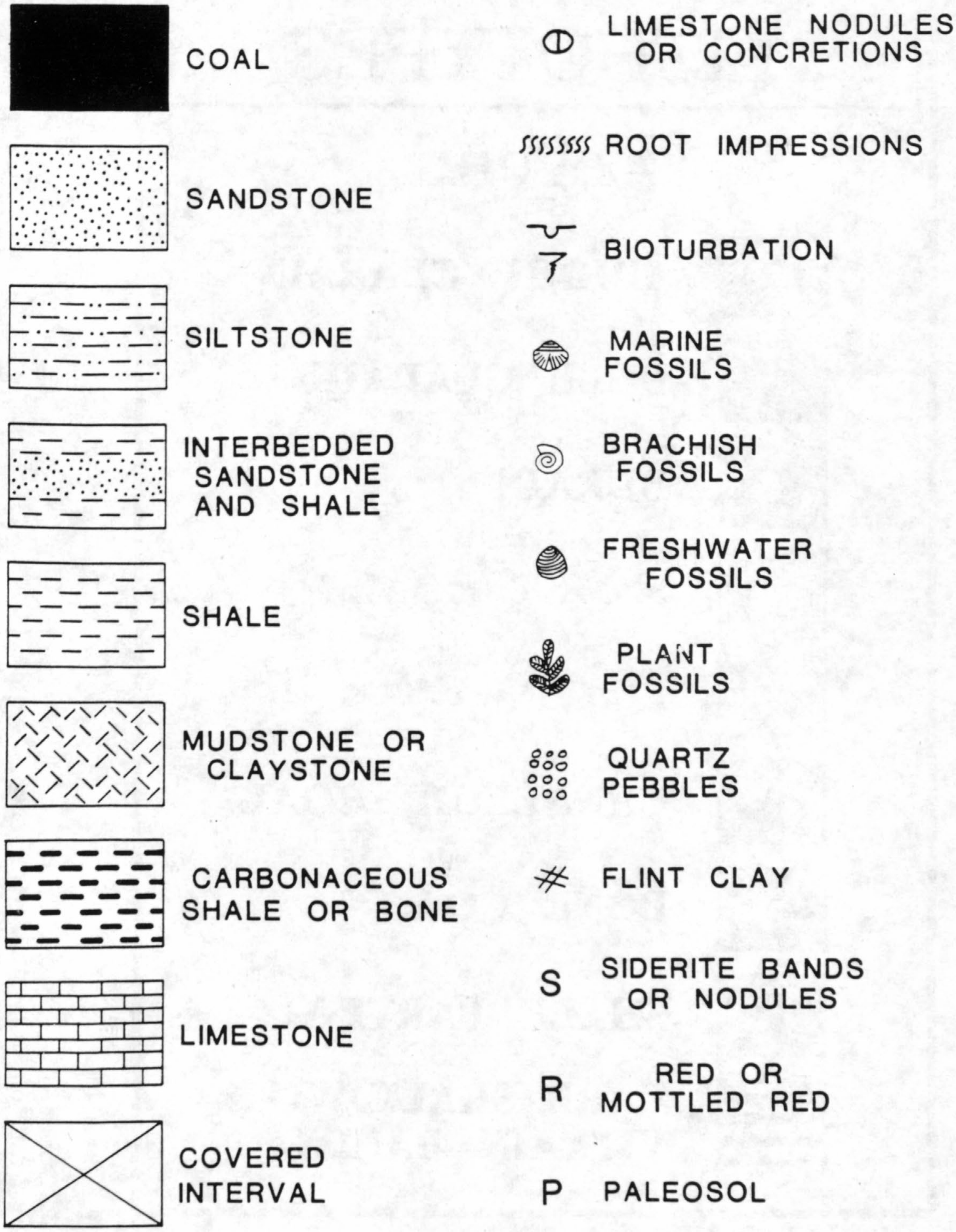
5) Changes in peat swamp paleofloras, inferred primarily from coal bed spore assemblages.

Although the above characteristics of Carboniferous strata are presented independently, they are interrelated. When integrated, these characteristics must be consistent with the stratigraphic record. The stops on the trip were selected to illustrate such a consistency.

We wish to thank the numerous institutions, agencies, and individuals that contributed to this field guide. The geologic mapping and stratigraphic analysis by Nick Fedorko and Bascombe (Mitch) Blake Jr., West Virginia Geological Survey, were essential to temporal and spatial correlations, and ultimately our paleoclimatic interpretations. Coal petrographic analyses by William Grady, West Virginia Geological Survey, defined petrographic criteria by which domed paleopeat deposits could be distinguished from planar paleopeat deposits. Both field descrip tion and X-ray diffraction studies of probable paleosols by Frank Dulong, U.S. Geological Survey, contributed the interpretations of base level change and pedogenesis. The sedimentological, paleoclimatological, and structural analyses of Ronald Martino, Marshall University, Donald Chesnut and Steven Greb, Kentucky Geological Survey, and Frank Ettensohn, University of Kentucky, contributed to the interpretations of Carboniferous stratigraphic sequences in southern West Virginia and eastern Kentucky. The support of the West Virginia Geological Survey, Larry D. Woodfork, Director and the Kentucky Geological Survey, Donald C. Haney, Director, is also appreciated. We also wish to thank Gergory J. Retallack for permission to include his summary of characteristics of soil orders (Retallack, 1990) (Appendix 1).

C. Blaine Cecil
Cortland F. Eble
22 October, 1992

LEGEND FOR SYMBOLS USED IN TEXT ILLUSTRATIONS



COAL LITHOTYPES

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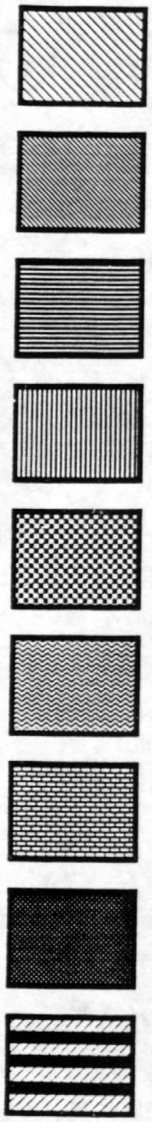


BRIGHT CLARAIN

DULL CLARAIN

DURAIN

SPLINT COAL

FUSAIN

CANNELOID COAL

BONE COAL

SHALE PARTING

CARB. SHALE WITH VITRAIN BANDS

INTRODUCTION

C. Blaine Cecil

Carboniferous cyclic stratigraphy has generally been interpreted on the basis of depositional (e.g., Ferm and Horne, 1979), transgressive-regressive (e.g., Klein and Willard, 1989) or tectonic models (e.g., Belt and Lyons, 1989; Klein, 1992). However, such models, which are based on physical processes, generally do not explain the cyclic stratigraphic distribution of Carboniferous siliciclastic and chemical rocks. Carboniferous strata in the central Appalachian basin exhibit numerous transgressive-regressive cycles, a foreland basin tectonic setting, and very similar depositional environments (Donaldson et al., 1985). The strata of the Mississippian System is, however, markedly different from those of the Pennsylvanian. For example, the Mississippian strata is almost completely devoid of coal beds, in marked contrast to the Pennsylvanian. This contrast has been attributed to climate change (e.g., White, 1925; Cecil et al., 1985; Donaldson et al., 1985). Because climate is an important control on sediment flux (both siliciclastic and chemical) to shelf margin, epicontinental shelf, and terrestrial sedimentary sequences, climate change must be considered as a major control on the stratigraphy of such sequences, as well as tectonics and transgressive-regressive cycles (Cecil, 1990).

Controls on sedimentation and cyclic stratigraphy are, therefore, perhaps best explained by models that encompass both physical and chemical processes. Such a model was proposed by Beerbower (1962) who developed the concept of auto- and allocyclic processes. Beerbower, using a fluvial system to explain stratigraphic relationships in Lower Permian (?) Dunkard Group strata in the Appalachian basin, defined autocyclic processes as changes of energy and materials within a sedimentary system (e.g., delta switching, stream meandering, bar migration) and allocyclic processes as changes in energy and materials external to the sedimentary system (e.g., climate, tectonics, and eustasy). As considered herein, allocyclic processes are the first order controls on stratigraphy and sedimentation whereas autocyclic processes are secondary and mainly control facies relationships within sequences bounded by allocyclic events. Of the three aforementioned allocyclic processes, tectonics and eustasy are the physical based processes commonly applied to the interpretation of stratigraphic sequences whereas climate, which is a major control on sedimentary geochemistry as well as siliciclastic sediment flux, typically is either overlooked or assumed to be constant.

Even though climate was recognized as early as 1875 by Lyell, in his *Principles of Geology*, as one of the first order controls on stratigraphy, interpretation of the relationships between paleoclimate and sedimentation have generally been limited to sequences that contain strata such as tillites, evaporites, aeolinites, and coal. Such interpretations are often used to infer paleolatitudes (e.g., Witzke, 1990). With notable exceptions (e.g., Huntington, 1907; Wanless and Shepard, 1936; Perlmutter and Matthews, 1989), there appears to have been a lack of appreciation and understanding of the frequency and intensity of climatic change, and the resulting impact of such change on stratigraphy and sedimentation. In the last decade or two excellent progress has been made on our understanding of the factors that control climate change. Interpretation of the effects of paleoclimates on cyclic stratigraphy and sedimentation can now be based on lithologic and paleontologic climatic signatures (e.g., Cecil et al., 1985; Perlmutter and Matthews, 1989; Cecil, 1990). This understanding has given us renewed insight into the first order effect of climate on stratigraphy and sedimentation as a result of movement of continents through latitudinal climatic belts with time (Shutter and Heckel, 1985), and the modifying effects of other factors such as orbital forcing cycles, mountain building, ocean circulation, and variation in atmospheric "greenhouse" gases (Table 1).

The frequency and magnitude of climate change can now be estimated and integrated into allocyclic models that evaluate tectonic and eustatic controls on accommodation space, and siliciclastic and chemical flux to deposystems. Such models should allow a transition from descriptive to predictive stratigraphy and resource evaluation. Predictive stratigraphic modeling appears to be possible because, to the degree that climate change is deterministic and therefore predictable, change in sediment flux to depocenters is deterministic and also predictable.

Table 1. TROPICAL AND SUBTROPICAL CLIMATE CHANGE CLASSIFICATION

RELATIVE DURATION	CAUSE	TIME (years)
Long-term	Movement of continents through latitudes;	10^6-10^8
	Orogenesis, "green house" gases (?)	10^5-10^7
Intermediate-term	100 and 400 ka cycles of orbital eccentricity green house gases (?)	10^5
Short-term	cycles in axial tilt and precession	10^4
Very-short-term	solar variation (?)	10^3
Instantaneous	weather systems	10^{-2}(months, weeks, days, hours)

(modified from Cecil, 1990)

STOP 1 - Mississippian-Pennsylvanian unconformity, Exit 15, I-68.

Stop Leaders: Blaine Cecil, Cortland Eble and Frank Dulong

STOP 1 (figs. 1 and 2) is at the Mississippian - Pennsylvanian unconformity (White, 1891), on the axis of Chestnut Ridge Anticline. The unconformity is exposed along the east-bound lanes of Interstate 68 (formerly U.S. route 48) just east of the exit for Coopers Rock State Forest (mile post 14.7). The red beds, which crop out at the west end of the east bound exit, are assigned to the Upper Mississippian Mauch Chunk Group (Namurian A). Three Pottsville coal beds are present at, or near **STOP 1**. All three are thin (<0.3 m, 1.0 ft), contain low to moderate ash yields and high sulfur contents (fig. 3). 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). This indicates that the stratigraphically youngest Pennsylvanian strata at this location are early 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 southwestern Virginia, over 1500 m (5,000 ft) of Upper Mississippian, Lower Pennsylvanian, and lowest Middle Pennsylvanian strata are missing at **STOP 1** (fig. 4). 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 coals in southern West Virginia.

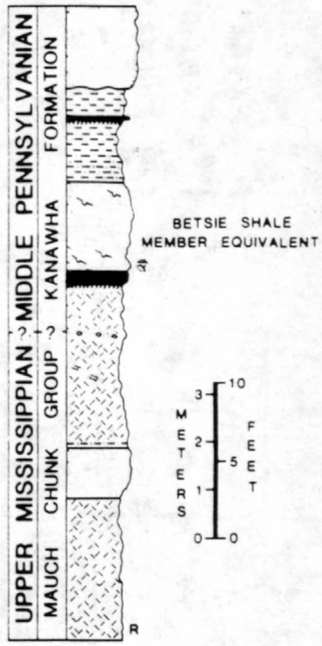


Figure 2 - Stratigraphic column for **STOP 1** along I-68 at exit 15.

The interval exposed at **STOP 1** appears to contain at least two siliciclastic paleosols, and a paleo-Histosol represented by the coal bed at the top of the paleosol sequence. The stratigraphy of the siliciclastic paleosols is quite complex at this locality, but they appear to represent at least two periods of deposition followed by subaerial exposure and weathering. The top of the lowermost paleosol occurs about 2.1 m (7 ft) below the base of the overlying coal bed. The lower paleosol overlies and appears to grade downward into green and red stata of the Upper Mississippian Mauch Chunk Group; thus, deposition may have occurred during the Mississippian whereas subareal exposure and paleosol development may have been during the Early Pennsylvanian. The well-developed lower paleosol may be classified as a paleo-

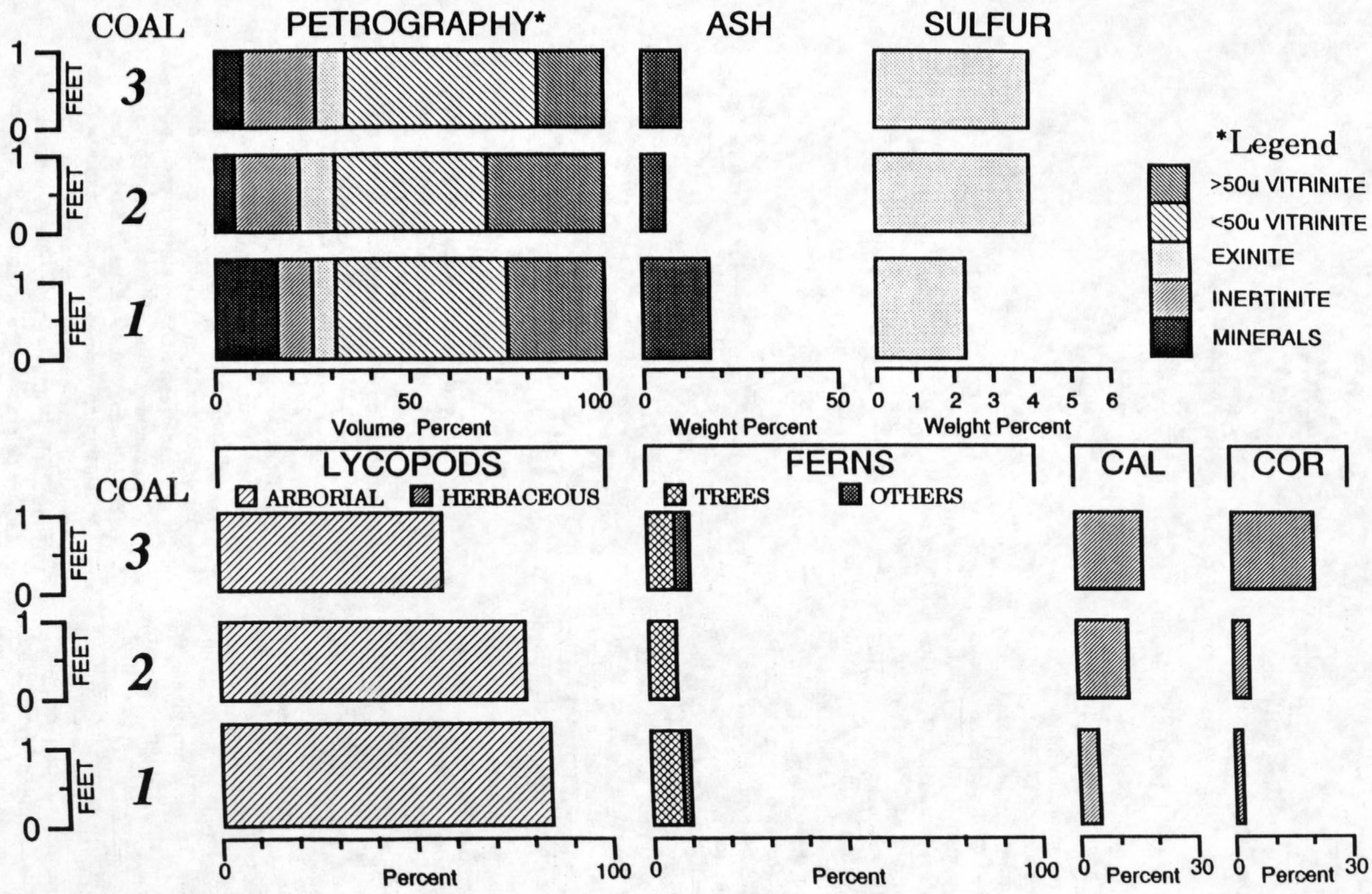


Figure 3) Distribution of macerals, miospores, ash yield and sulfur content in Pottsville Group coal beds along I-68.

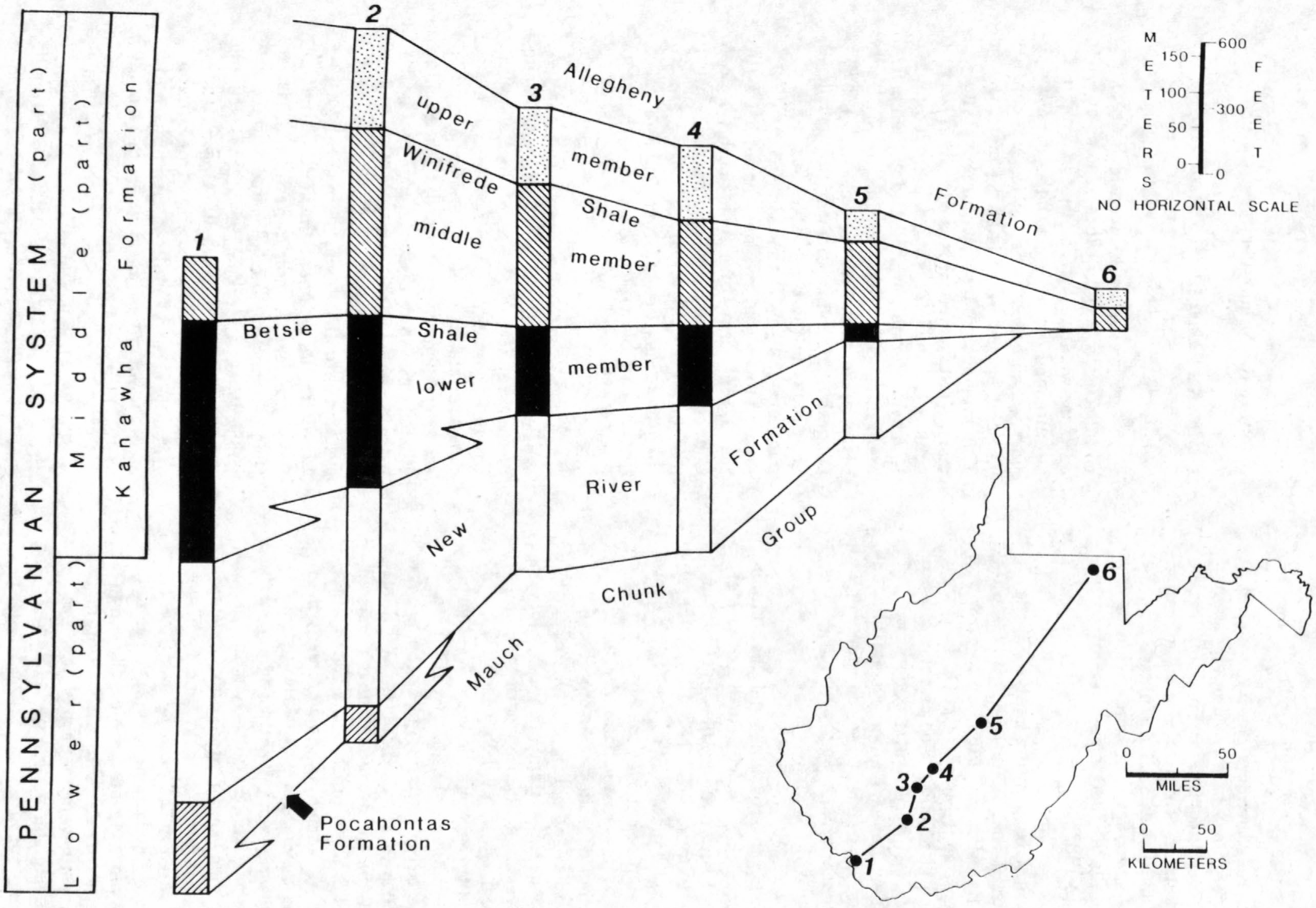


Figure 4) Northward thinning of Pennsylvanian strata. At STOP 1 Lower Pennsylvanian (Pocahontas and New River Fms) are absent.

Ultisol whereas the more poorly-developed upper paleosol may be more properly classified as a paleo-Inceptisol or Entisol using the U.S. Department of Agriculture classification system (Soil Survey Staff, 1975) (see Appendix 1 for a summary of soil order characteristics by Retallack, 1990; also summarized in Buol, et al., 1988).

On the basis of the analysis of the "mid-Carboniferous eustatic event" of Saunders and Ramsbottom (1986), up to 4.5 Ma may be represented in the 4 m (13 ft) interval exposed at **STOP 1**. The complex paleosol and exposure surface 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.

On an interbasinal scale, high alumina refractory clay deposits, such as the Mercer clay in Pennsylvania, the Olive Hill clay in eastern Kentucky (**STOPS 16, 17,** and **18**), and the Cheltenham clay on the Ozark dome in Missouri, also occur at the Mississippian-Pennsylvanian systemic boundary. These high-alumina deposits are also probable paleo-Ultisols. The paleogeography of paleo-Ultisols (?) at the systemic boundary appears to be the result of intense weathering under high annual rainfall, which was fairly evenly distributed throughout the year. Weathering may have commenced as early as 330 Ma during the Late Mississippian at the on-set of a global eustatic event (Saunders and Ramsbottom, 1986). Exposure may have persisted for up to 4.5 Ma in much of the North American continent, including the aforementioned areas, was moving northward through the paleo-equatorial zone. Weathering appears to have been particularly protracted and intense in those regions that contain diaspore clay deposits (Missouri, Kentucky, and Pennsylvania). Compared to the paleosol at **STOP 1**, the diaspore clay regions may have been somewhat more elevated, which allowed intense weathering on relatively well-drained soils. Paleosol development may have progressed during this 4.5 Ma period during subareal exposure from the latest Mississippian into the Middle Pennsylvanian.

Chestnut Ridge "Rolling Stops"

As we descend Chestnut Ridge, we will discuss the significance of some of the outcrops we drive by. The first outcrop has an unnamed marine zone that occurs in a siderite bed beneath the Lower Connoquenessing Sandstone, which is at the top of the cut. This previously unreported unit contains a marine fauna that compositionally is very similar to the Dingess marine Shale (middle Kanawha Formation, mid-Middle Pennsylvanian) in southern West Virginia (T.W. Henry, personal communication). We will see the Dingess Shale again, albeit better-developed, later today at the Birch River section. A miospore analysis of the thin (0.3 m, 1.0 ft), discontinuous coal bed, designated Pottsville coal 2, that occurs directly beneath the Lower Connoquenessing Sandstone at this location has shown the palynoflora to correlate with the No. 2 Gas - Cedar Grove coal interval (fig. 2). This biostratigraphic age assignment is consistent with the invertebrate data.

The next rolling stop shows a large exposure of the Upper and Lower Connoquenessing Sandstones. Rocks 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. It has been suggested that this group of rocks were deposited by bed load, braided fluvial systems onto an alluvial plain (Presley, 19-79). Pottsville sediments in the Chestnut Ridge area are thought to have been derived from orogenic highlands located to the east and southeast (Meckel, 1967; Donaldson and Schumaker, 1981). Pottsville Group Sandstones, like the ones shown in this outcrop, generally occur as multi-storied units up to 30 m (100 ft) thick, averaging 9 to 12 m (30 to 40 ft).

A palynological analysis of a thin, discontinuous coal bed (designated Pottsville coal 3) in a shale lens in the Upper Connoquenessing Sandstone near the top of this cut indicates that it is age equivalent with the Fire Clay - Chilton coal interval of the Kanawha Formation in southern West Virginia (fig. 2).

The next rolling stop will show an exposure of the Homewood (?) Sandstone, the top of which 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.

The Lower and Upper Kittanning, and Lower Freeport coal beds, in ascending order, can be seen at the last rolling stop outcrop. The Lower Kittanning coal bed has been correlated with the No. 6 Block coal bed in southern West Virginia (Kosanke, 1984), with 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). Although thin in the Chestnut Ridge area, the Lower and Upper Kittanning coal beds attain minable thickness, and represent a significant coal reserve in northern West Virginia, eastern Ohio, and western Pennsylvania (fig. 5).

STOP 2: Upper Freeport coal bed and associated strata: Middle-Upper Pennsylvanian boundary, I-68, mile post 11.4.

Stop Leaders: C. Blaine Cecil, Frank T. Dulong and Cortland Eble

The section at **STOP 2** includes the stratigraphic interval from the Upper Freeport Limestone Member of the Middle Pennsylvanian Allegheny Formation (Wilmarth, 1938) up to the Mahoning coal bed of the Upper Pennsylvanian Conemaugh Group (fig. 6). The primary emphasis at this stop is the stratigraphic interval from the base of the Upper Freeport Limestone to the top of the Upper Freeport coal bed. The top of the Upper Freeport coal bed, which is exposed as three benches of coal at this locality (Fig.), is defined as the top of the Allegheny Formation and the base of the overlying Conemaugh Group (Stevenson, 1873). The Upper Freeport coal bed horizon occurs throughout the Appalachian basin in Pennsylvania, Maryland, West Virginia, and Ohio (Fig.). Where it is sufficiently thick, this laterally extensive coal bed has been mined from the eastern outcrop belt in western Maryland to the western outcrop belt in east central Ohio, a distance of over 250 km (150 mi). This coal bed is underlain by nonmarine strata that include underclay, siltstone, shale, flint clay, and the Upper Freeport Limestone. The interval from the base of the limestone to the top of the coal bed is interpreted to be the result of complex variations in sediment flux that occurred in response to a cyclic paleowater table and paleoclimate.

The Upper Freeport Limestone horizon is as laterally extensive as the Upper Freeport Coal bed. The limestone is nonmarine and probably was deposited in large, very shallow, lakes as indicated by multiple subareal exposure features that include subareal crusts, pedogenic brecciation, and residual clay. Intermittent deposition and subareal 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 probable 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 (climate model is from Cecil, 1990).

Subsequent to deposition, the Upper Freeport limestone appears to have been subjected to an extended period of subareal

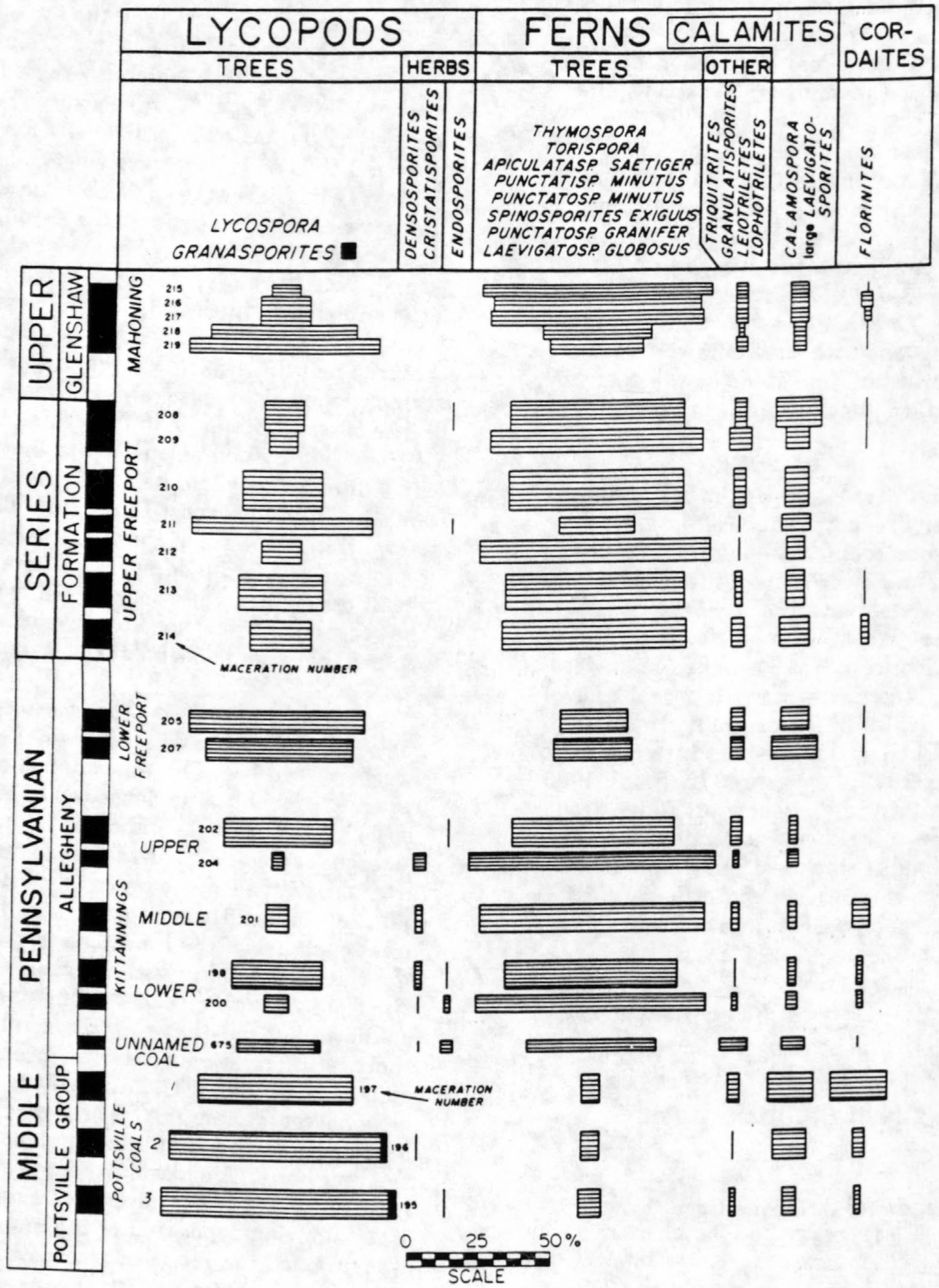


Figure 4.5 - Palynological analyses of Middle Pennsylvanian coal beds exposed along I-68 between **STOPS** 1 and 2. Note the dominance of *Lycospora* in the Pottsville coals, whereas tree fern spores are a far more significant element of Allegheny coal palynofloras.

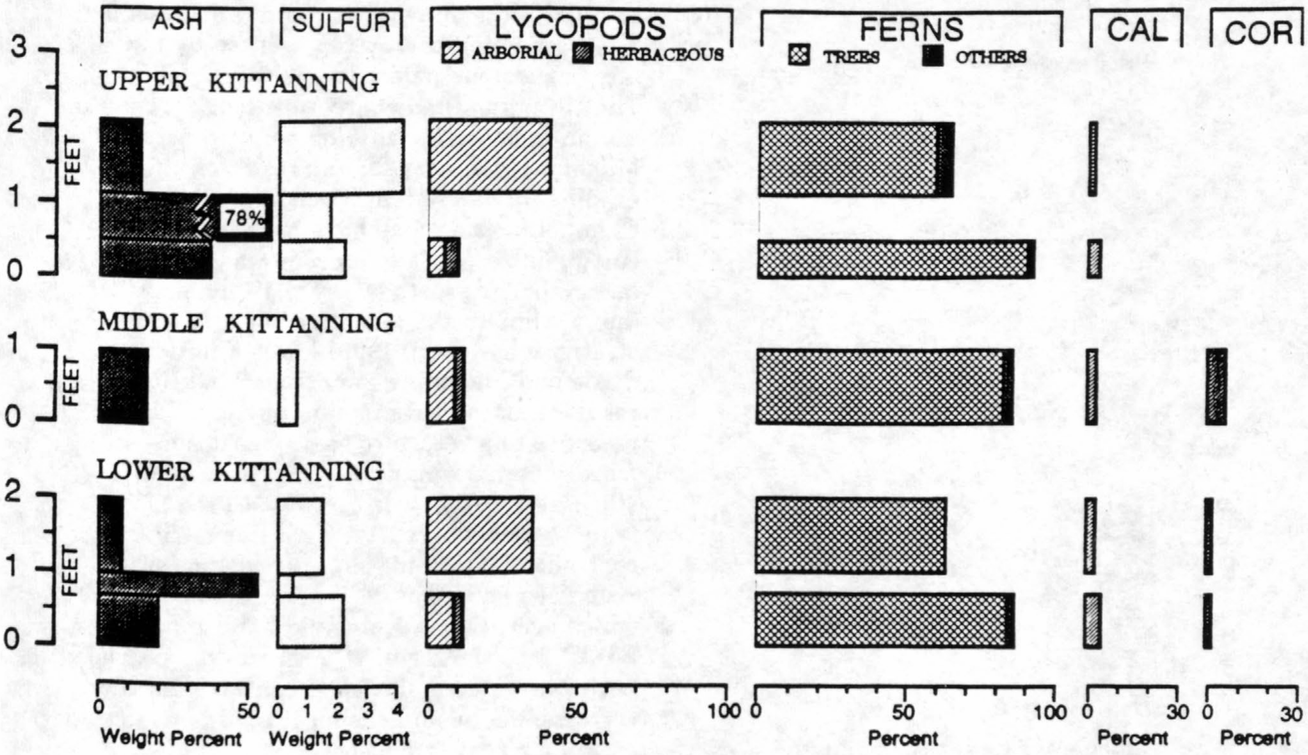


Figure 5 - Distribution of ash yield, sulfur content and miospore taxa in the Lower, Middle and Upper Kittanning coal beds, exposed along I-68.

exposure and weathering. The and resultant residual clay deposits imply a drop in the water table during the onset of increasing pluvial conditions. The increasing rainfall also contributed to an influx of siliciclastic sediment, which is shown by the exposed siliciclastic strata overlying the Upper Freeport Limestone at this stop (Cecil, 1990). Further increases in rainfall led to increased vegetative cover, rainfall dilution of runoff, and leaching of residual soils, which reduced the influx of siliciclastic sediment and dissolved solids. Extensive leaching of the landscape, under the pluvial part of 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 peat formation (Cecil et al., 1985).

On a regional scale, a complex of kaolin enriched paleosols occur in a mosaic within the interval from the base of the Upper Freeport Limestone up to, and including, the Upper Freeport coal bed. These paleosols appear to be the result of a fluctuating water table and weathering during pluvial parts of climatic cycles. The most intensely developed paleosols probably formed on well-drained paleotopographic highs. The kaolin enriched deposits have been locally mined and used in the manufacture of refractory brick.

Interruption of peat formation, as illustrated by the three benches of coal and

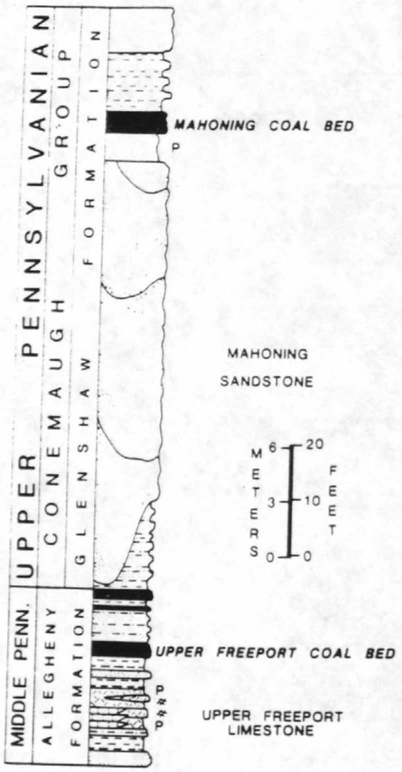


Figure 6 - Allegheny - Lower Conemaugh strata exposed along I-68 at mile post 11.4

inter-bedded partings at **STOP 2**, are sometimes interpreted as crevasse splays or other autocyclic depositional events (e.g., Ferm and Horne, 1979; Flores, 1986). Alternatively, these interruptions may be the result of allocyclic controls such as climatic events that cause an influx of siliciclastic sediment and (or) change in water table. 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 et al., 1985).

Peat formation in the Upper Freeport paleoswamp was terminated by an allocyclically-controlled rising water table that finally out-paced peat formation (Cecil et al., 1985). The lacustrine environment of the drowned paleoswamp became the site of an autocyclic facies mosaic of siliciclastic deposition that includes the Uffington Shale Member of the Conemaugh Group, whose type section is just south of Morgantown, WV, and the Mahoning Sandstone Member, as represented by the large sandstone paleo-channel fill at **STOP 2**. The Uffington Shale may represent deposition in a lacustrine environment, whereas the Mahoning Sandstone is the result of a prograding fluvial system (Cecil et al., 1985). Channel incision may have been by down cutting in response to a drop in base level (allocyclic) or progradation of a fluvial system (autocyclic). In the climate model of cyclic stratigraphy (Cecil, 1990), the Uffington Shale and Mahoning Sandstone are also the result of an increase in siliciclastic influx in response to a return to somewhat drier and more seasonal conditions. This increased siliciclastic influx was coeval with the lacustrine system where a rising water table was controlled by tectonic subsidence and (or) rising sea level. The Mahoning Coal bed, which overlies the Mahoning Sandstone at **STOP 2**, is the result of a return to a pluvial period of reduced siliciclastic influx, and the correct climatic and chemical conditions necessary for peat formation.

The Upper Freeport coal at **STOP 2** consists of 4 splits over a 4.3 m (14 ft) interval. The main (lower) split is 0.9 m (3 ft) thick, is vitrinite-rich, especially the >50 micron vitrinite types, and is low in ash-yield and sulfur (fig. 7). 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 Upper Freeport swamp was not at its best (minable) 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 with minor emplacement of minerals and sulfur. Tree ferns dominate the flora and the pre-vitrinite plant debris was moderately preserved with very little oxidation of the peat. With further peat accumulation (middle bench) the planar swamp developed into a soligenous (?) swamp that may have been slightly elevated above local stream level. Mineral and sulfur em

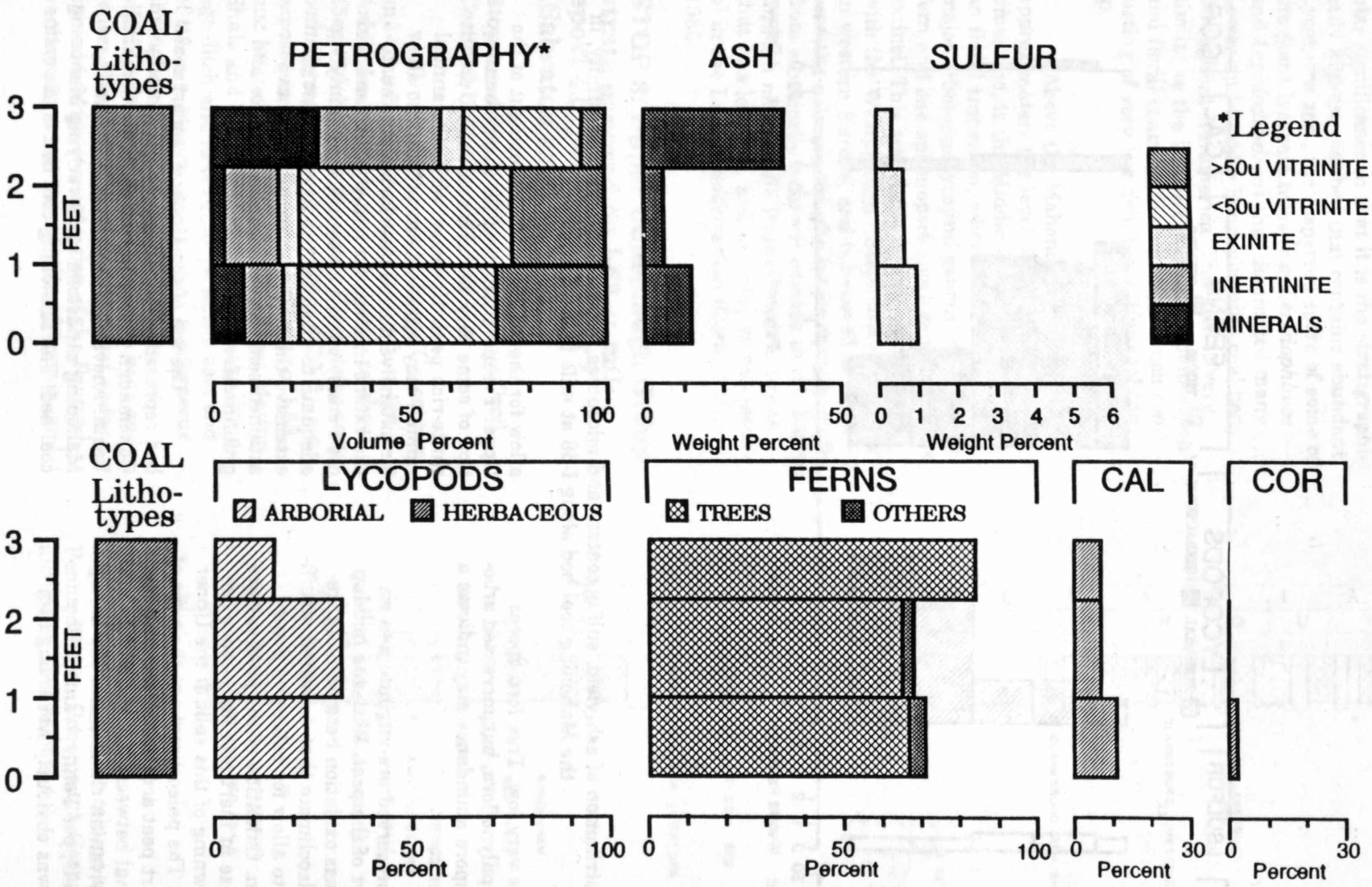


Figure 7) Distribution of macerals, miospores, ash yield and sulfur content in three benches of the Upper Freeport coal bed at **STOP 2**.

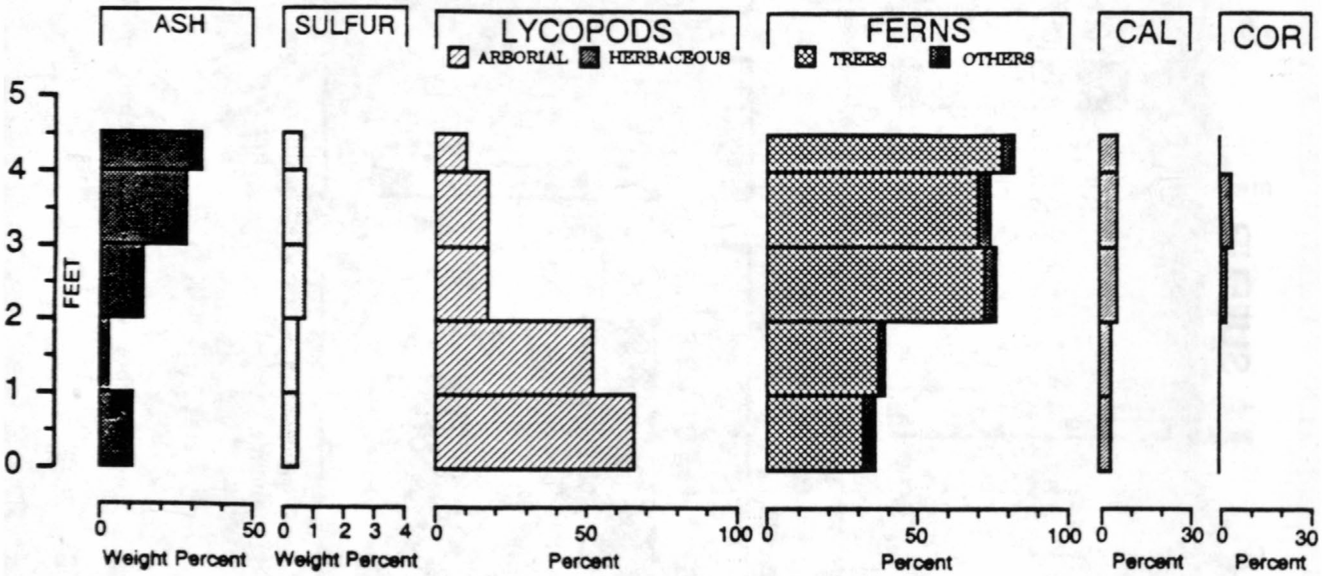


Figure 8 - Distribution of ash yield, sulfur content and miospores, grouped according to affinity, in the Mahoning coal bed along I-68 at mile post 11.4.

placement was very low. Tree fern spores dominate the palynoflora, but increased arboreal lycopsid spore abundance may indicate a standing water cover.

Well-preserved pre-vitrinite was an important part of the peat. With the buildup of the peat mass oxidation became more frequent in a paleoclimate that delivered insufficient rainfall to allow for extensive domed peat formation. Oxidation of the peat surface and an increase in inertinite abundance preceded the drowning of this split of the Upper Freeport coal. The paleoclimate at the time of Upper Freeport peat accumulation probably was transitional between the wet Middle and Lower Pennsylvanian climate and the drier climate of the Upper Pennsylvanian. The paleoclimate was obviously wet enough to allow for the widespread development of the Upper Freeport coal, and for the accumulation of some very low ash and sulfur, pre-vitrinite-rich peat. However, while the annual rainfall may have been insufficient to allow for extensive domed peat formation (i.e. to the extent inferred for many Lower and Middle Pennsylvanian swamps), it certainly was adequate to allow for the development of an extensive that, in some areas, may have attained some elevation above surface and groundwater sources.

The top of the Upper Freeport coal bed represents the Allegheny Formation - Conemaugh Group boundary in the Dunkard Basin. Above this contact is the massive Mahoning sandstone and overlying Mahoning coal bed. The Mahoning coal bed is of particular significance in that it is the stratigraphically highest coal bed that contains abundant *Lycospora* spp., the dispersed spore of some of the giant lycopsid trees (e.g. *Lepidophloios* and *Lepidodendron*) that dominated many Lower and Middle Pennsylvanian coal beds throughout Euramerica. This coal is very similar to the Upper Freeport in composition and floral contribution with one bench consisting of very low ash, low sulfur coal (fig. 8).

Above the Mahoning coal bed, at approximately the level of the Brush Creek limestone, is the Middle - Late Pennsylvanian floral transition, where all but one of the major arboreous lycopsid genera, several tree fern and one sphenopsid spore genera become extinct. This transition is time-equivalent with the Westphalian - Stephanian boundary in western Europe, and is believed to represent the culmination of a major climatic shift from a basically ever-wet climate in the Early through mid-Middle Pennsylvanian, to one that was less-wet, and probably more seasonal in the Late Pennsylvanian (Cecil et al., 1985).

STOP 3: Upper Conemaugh Group strata exposed on I-68 at milepost 4.0, just west of Exit 4 (Sabraton exit).

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

Upper Conemaugh Group (Casselman and Glenshaw formations) strata from the Ames Marine Shale and Limestone to about 15 m (50 ft) above the Little Clarksburg coal bed, aggregating a total of 72 m (237 ft), are exposed in the cut on the north side of I-68 just west of Exit 4 (fig. 9). The Ames Marine Shale and Limestone is exposed in the drainage ditch and first cut slope at the east end of the cut. Marine invertebrate fossils occur throughout the gray-green and gray-red shales as well as in thin limestone beds and lenses. The Ames represents the last widespread marine transgression of the Pennsylvanian into north central West Virginia, and

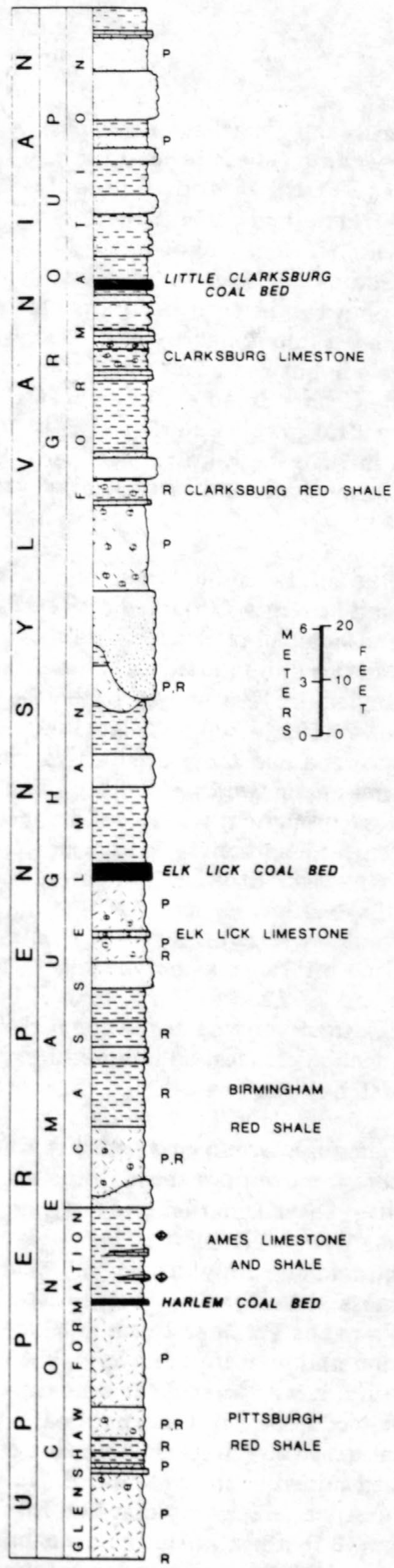


Figure 9 - Stratigraphic column of Upper Pennsylvanian Conemaugh Group strata along I-68 at mile post 4.0

is a key stratigraphic marker in this part of the geologic column. These same units will be seen again at **STOPS 15 and 16** in southwest West Virginia and eastern Kentucky. The underlying Harlem coal bed and Pittsburgh Red Shale (fig. 9) are not exposed at this location but can be seen along the east bound entrance ramp of nearby Exit 4. These units will be emphasized at a rolling stop on I-79 near the Goshen Road exit after **STOP 4**. Other key stratigraphic marker beds in this section include the Elk Lick coal bed, Clarksburg Limestone, and Little Clarksburg coal bed.

Unlike the Pottsville Group and Allegheny Formation strata examined at **STOPS 1 and 2**, the Conemaugh Group contains mainly red shales and mudstones. These red sediments regionally first appear in the section 30 to 60 m (100 to 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 rocks is higher compared to Middle Pennsylvanian sections (Cecil et al., 1985). Carbonate occurs as beds and admixture in marine intervals, as nonmarine, lacustrine beds (e.g. Clarksburg Limestone of I.C. White (1891), fig. 9) and as pedogenic nodules, discontinuous lenses, and admixture within mudstones and shales.

Conemaugh Group coal beds are thinner, and more impure than those in the underlying Allegheny Formation or overlying Monongahela Group (**STOP 4**). The Little Clarksburg coal bed rarely exceeds 0.6 m (2 ft) in thickness in this region. It is thicker and minable 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 in this cut, has also been mined in the Potomac Basin and also in central West Virginia. The Elk Lick is 0.9 m (3 ft) thick in this cut, 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 (fig. 10; table 1).

The Elk Lick coal exposed in the road cut is 4 feet thick (fig. 11) and represents peat that accumulated in a drier interval than the stratigraphically lower Upper Freeport or Mahoning coals. This coal was formed after the demise of the arboreous lycopsids at the Westphalian-Stephanian boundary. The Elk Lick coal is high in ash yield and very high in sulfur and contains significantly greater inertinite, and less well-preserved (>50 micron) vitrinite than the Upper Freeport or coals lower in the Pennsylvanian. At the top of the section the Little Clarksburg coal is exposed (fig. 12), the stratigraphically lower Harlem and West Milford coals are exposed (figs. 13 and 14).

These coals are all similar in ash yield, sulfur content, petrographic composition and floral character to the Elk Lick, and typify Conemaugh coals that apparently accumulated in planar swamps with significant surface and groundwater 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 micron vitrinite components, probably by anaerobic microbes. These attributes suggest a dry paleoclimate with insufficient annual rainfall to maintain an acidic ombrogenous swamp.

The gray, red and green mottled mudstones seen in this section are interpreted to be paleo-Vertisols. Diagnostic features include obliteration of primary sedimentary features, mottling, calcareous pedotubules, rhizoconcretions, drab halo root traces, accumulations of carbonate glaebules and petrocalcic horizons. In some cases, nodular zones may represent pedogenically altered lacustrine limestone beds. The Elk Lick Limestone of Platt and Platt (1877) (fig. 9) at this stop is possibly an example of this process. In addition, the paleosols exhibit convex upward slickensided surfaces, interpreted as vertic

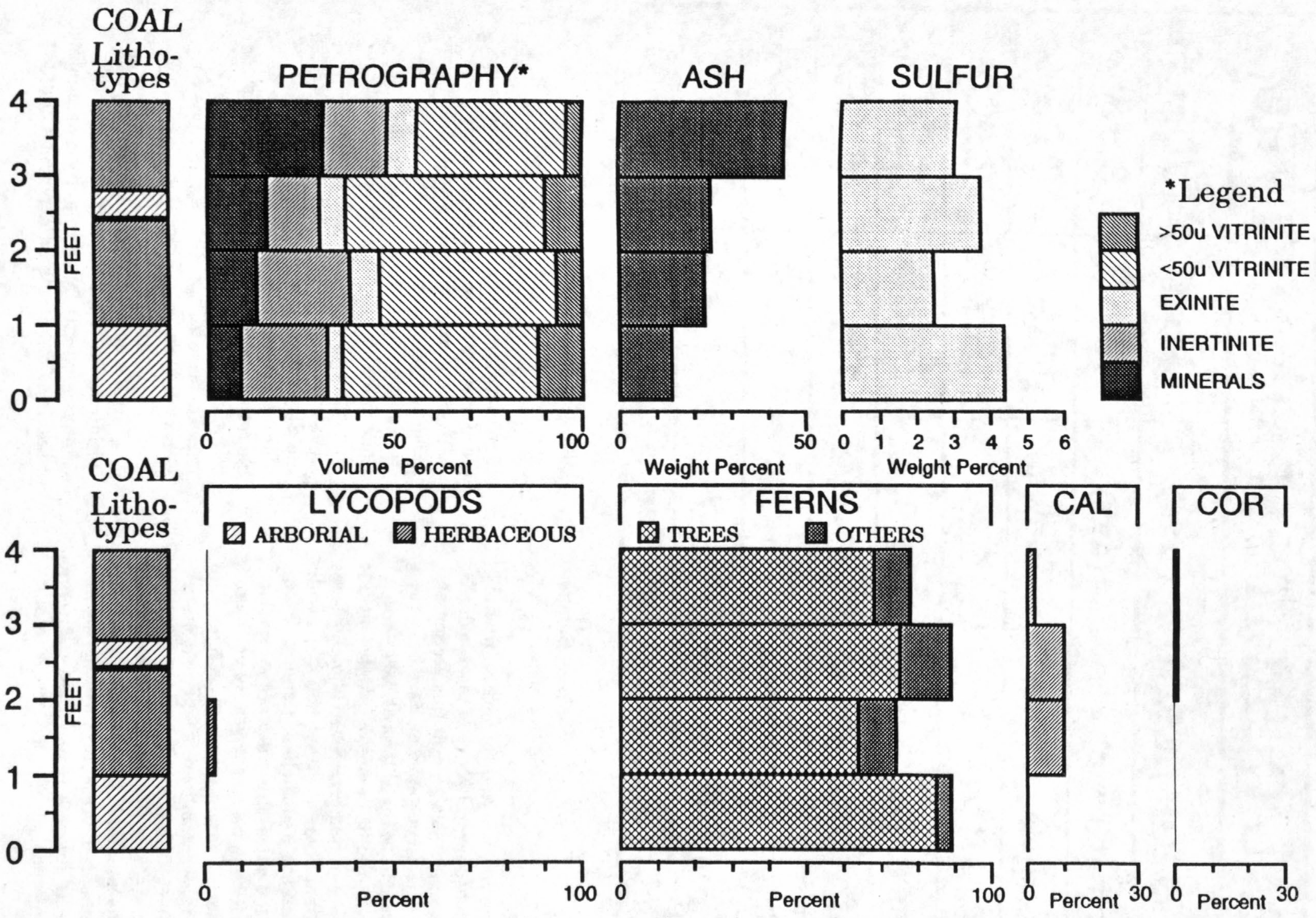


Figure 10) Distribution of macerals, miospores, ash yield and sulfur content in the Elk Lick coal bed at **STOP 3**.

COAL QUALITY PARAMETERS			
FORMATION / GROUP	n*	**ASH**	**SULFUR**
Dunkard Group	(n=9)	**32.4%**	**3.58%**
Monongahela Group	(n=985)	**11.9%**	**3.08%**
Conemaugh Group	(n=64)	**15.7%**	**2.92%**
Allegheny Formation	(n=250)	**14.8%**	**1.63%**
Kanawha Formation	(n=625)	**11.9%**	**1.09%**
New River Formation	(n=194)	**7.9%**	**1.04%**
Pocahontas Formation	(n=75)	**8.8%**	**0.82%**
* n = Number of samples (Full Channel and Bench samples only)			

Table 2 - Summary of coal quality parameters

structures, characteristic of vertisol profile development. Several zones of inter-bedded shales and paleosols seen in this section are laterally persistent and they have been recognized throughout the Dunkard Basin and given stratigraphic names. Examples include the Pittsburgh Red Shale of White, 1891, the Birmingham Shale of Stevenson, 1876, and the Clarksburg Red shale of Hennen, 1912. Some of the paleosol profiles are incomplete, horizons having been removed by paleo-erosion. Other paleosols appear to be compound profiles, representing several periods of soil development each partially overprinting the previous one. The thickest and most persistent of these units is the Pittsburgh Red Shale, possibly representing weathering and pedogenesis during a regional low stand prior to transgression of the Ames sea. A total of 15 m (48 ft) of this part of the section below the Harlem coal bed is exposed along the I-68 eastbound entrance ramp (fig. 9). Other well-developed paleosols can be seen at this stop. One is immediately above the Ames Limestone and Shale and overlying sandstone. This soil is capped by a very thin carbonaceous streak. Another paleosol occurs below the Elk Lick coal bed. In addition to the nodular carbonate zone, this interval contains a burrowed and/or rooted gannister sandstone bed. Still other paleosols can be seen in the interval between the Elk Lick and Little Clarksburg coal beds (fig. 9).

The paucity of minable coal deposits within strata of the Conemaugh Group (fig. 15) has been attributed to a dry paleoclimate relative to conditions for deposition of under-

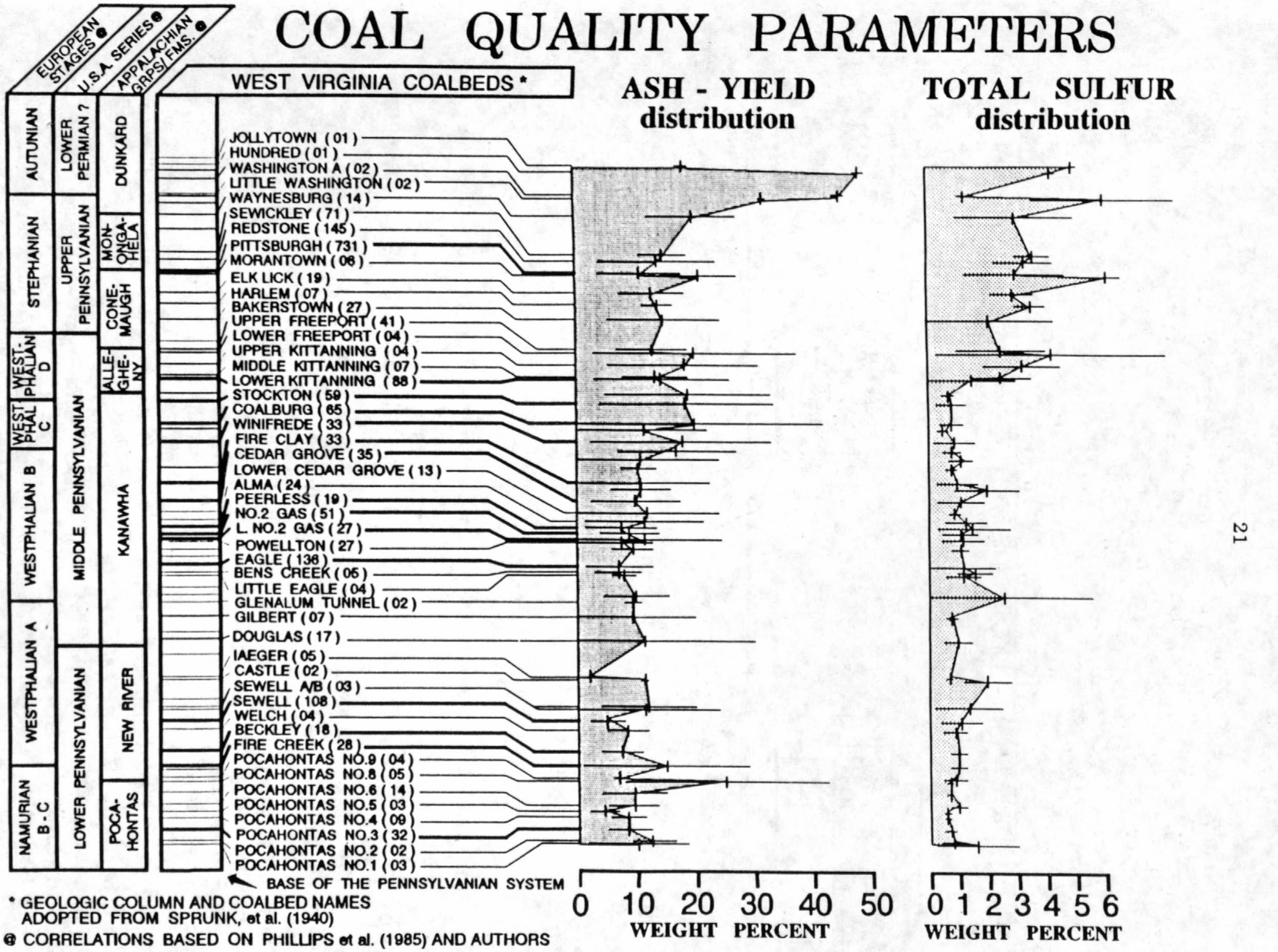


Figure 11) Distribution of ash yield and sulfur content in Pennsylvanian coal beds in West Virginia. Mean values are indicated by vertical tick marks and horizontal lines demark one standard deviation from the mean (N=1967).

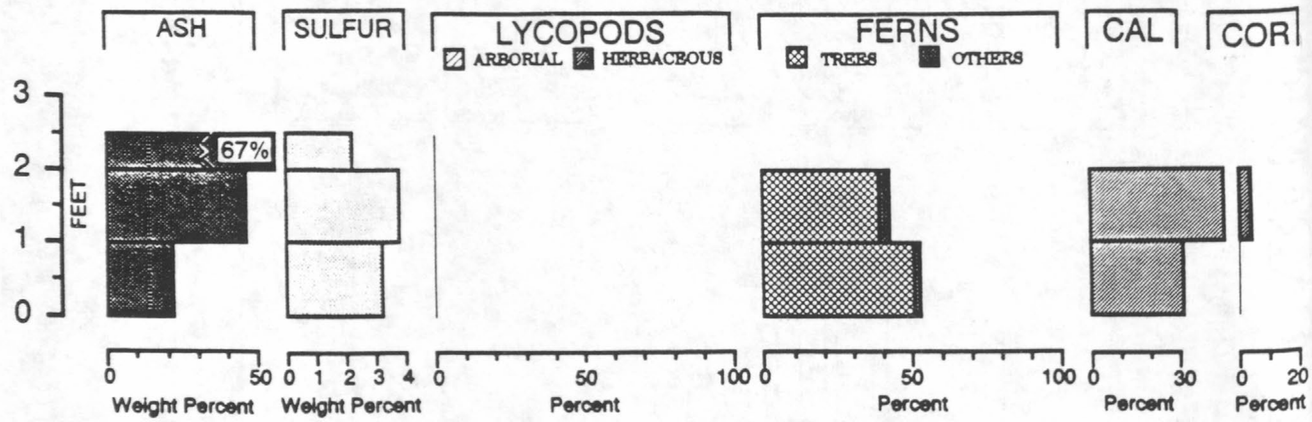


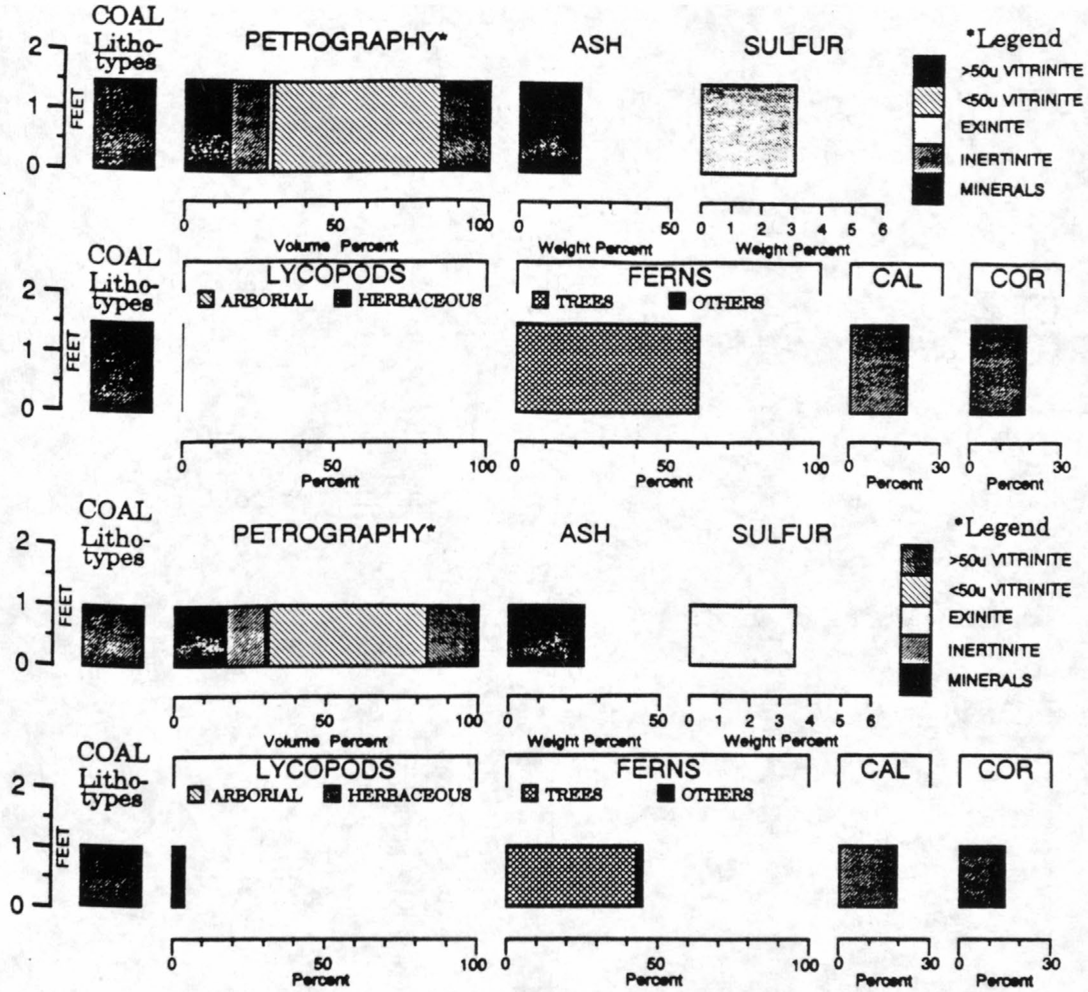
Figure 12 - Distribution of miospores, ash yield and sulfur content in the Little Clarksburg coal bed along I-68 at mile post 4.0.

and overlying strata (Cecil et al., 1985). This dry climatic condition during the Late Pennsylvanian in the Appalachian basin has also been inferred for equivalent strata (Missourian) in the western interior basin (Ronald West, personal communication). If these interpretations are correct, this climatic event would appear to have been of long-term duration, and of a continental scale.

STOP 4: Upper Conemaugh Group and Lower Monongahela Group strata exposed on the north side of the Morgantown Mall complex, near I-79 Exit 152, Morgantown, WV.

Stop Leaders: Nick Fedorko, Blaine Cecil, Cortland Eble, and Wiliam Grady

Approximately 18 m (58 ft) of Upper Conemaugh Group rocks and 32 m (106 ft) of lower Monongahela Group rocks are exposed in cuts made for the construction of the Morgantown Mall and upper commercial area (fig. 16). The section features five coal beds (some multi-benched) and an abundance of non-marine, lacustrine limestone beds. Starting at the eastern end of the exposure, several benches of the Little Pittsburgh coal bed are seen interbedded with shales, mudstones, and lacustrine carbonates. The Little Pittsburgh coal bed is of minor economic importance, but is persistent enough to serve as an important stratigraphic marker. 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 is well-developed, the



Figures 13 and 14 - Distribution of macerals, miospores, ash yield and sulfur content in the West Milford (top)and Harlem (bottom) coal beds along I-68 at mile post 4.0.

subjacent soil profile is very poorly-developed, thin and contains some carbonate in the form of nodules. Subsequent stops featuring the Pittsburgh coal bed will illustrate the effects of the "Pittsburgh coal climate" on the substrate where the coal facies is poorly-developed or absent (fig. 17). A discussion below previews the regional changes we will see in the Pittsburgh coal bed and related strata at later stops.

The Pittsburgh coal bed is extremely valuable to the coal mining industry of West Virginia, Pennsylvania and Ohio, exhibiting remarkable lateral persistence of thickness and quality (fig. 18). The exposure in this section is a pillar left from the long-abandoned underground mining operations in this area. The basal bench (referred to as the main bench) here is 2.6 m (8.5 ft) thick with thin shale partings. Including roof shales and rider coal beds the Pittsburgh is 5.2 m (17 ft) thick.

The Pittsburgh coal bed exposed in the Mall cut has an 2.6 m (8.6 ft) thick main coal of generally low ash-yield and moderate sulfur content (fig. 19). At this location the coal consists of six benches, two more than present six miles to the northeast where the coal was extensively studied in a surface mine at the Greer estate (figs. 20 and 21). The basal bench (lower 0.3 m, 1 ft) is present across most of the areal extent of the Pittsburgh coal. It is very high in sulfur and moderate ash-yield, and has a tree fern dominant palynoflora with distinct calamite and

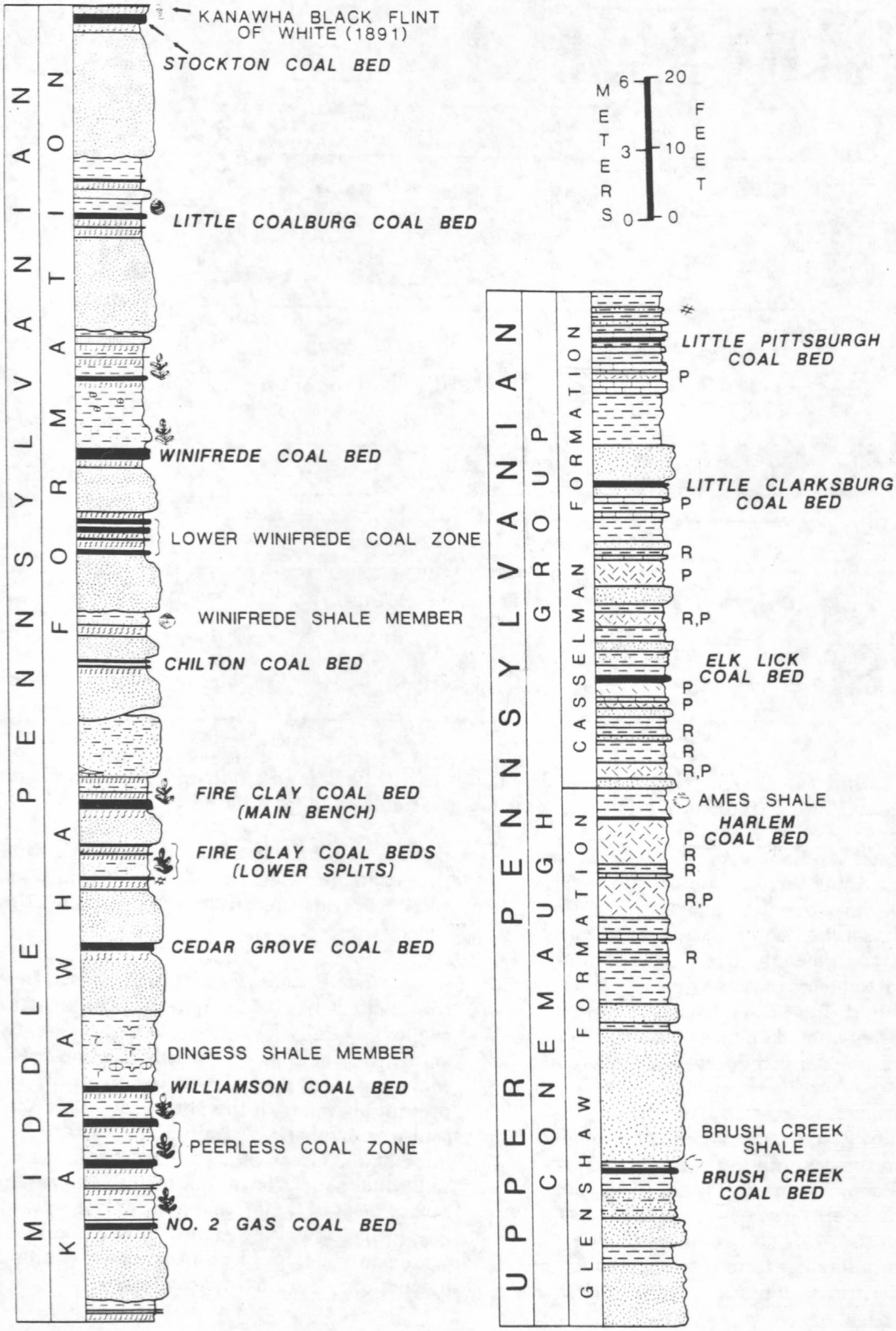


Figure 15) Comparison of a coal-rich Kanawha Formation section with a coal-poor Conemaugh Group section of equal thickness

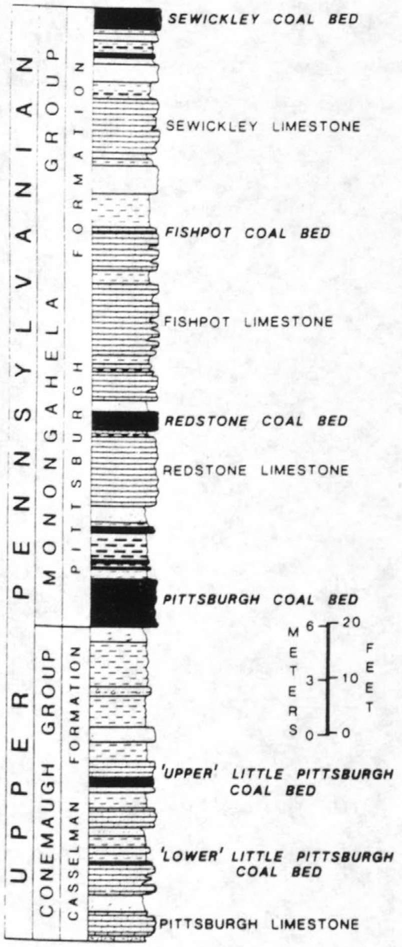


Figure 16 - Upper Conemaugh and lower Monongahela Group strata exposed along the north side of the Morgantown Mall complex.

cordaite contributions. This lower bench is interpreted to represent the pioneering plant community of the Pittsburgh swamp. The ash yield and sulfur contents 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 (4 to 4.5 ft) level, the coal is very low ash and moderate sulfur content and petrographically shows two trends in swamp development. These trends of increased peat preservation, as shown by increased >50 micron vitrinite components, are reflected in the sulfur content and palynofloral succession, but not in ash yield. The first, terminated by a fusain parting, displays an upward increase in vitrinite content, especially >50 micron component, an increase in calamite and arboreous lycopsid spores, and increased sulfur content. A second similar trend is terminated by the bone coal parting at four feet. Increased vitrinite and >50 micron 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 groundwater as water depth increased. The termination of these trends by fire and 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 Pittsburgh coal areal extent. The four foot parting was extremely widespread and occurs throughout the Pittsburgh coal at approximately the same level in the bed (Cross, 1952).

Autocyclic factors other than climate doubtless contributed to the vast areal distribution of the Pittsburgh coal bed. However, the thickness and quality of the coal appears to be strongly climate-controlled. During the initial stages of accumulation of the Pittsburgh paleopeat annual rainfall, augmented with surface water flow, was sufficient to allow the development of a large planar swamp. During later stages of peat development the influential effects of rainwater vs. surface/groundwater 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 throughout the paleo-peat thickness probably because of more frequent and extensive incursions of fresh surface and ground water into the peat swamp. In the Morgantown area layers of peat influenced by higher pH water alternate with low-ash, low-sulfur peat influenced during accumulation by ombrogenous water as in the Greer estate coal column (figs. 20 and 21) and Morgantown Mall (fig. 19). To

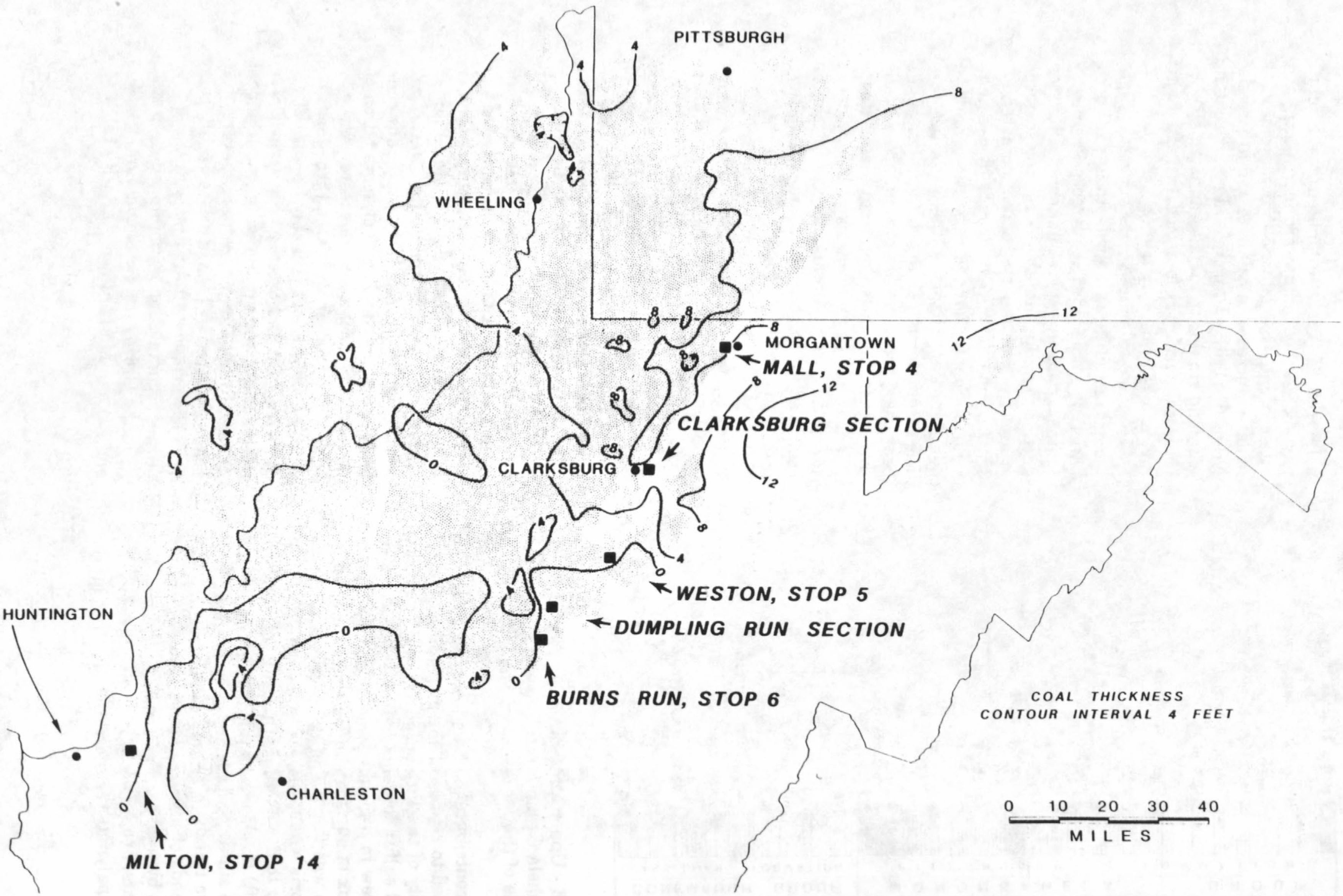


Figure 17) Map showing the areal extent of the Pittsburgh coal bed showing thickness isopach lines and field trip stop locations.

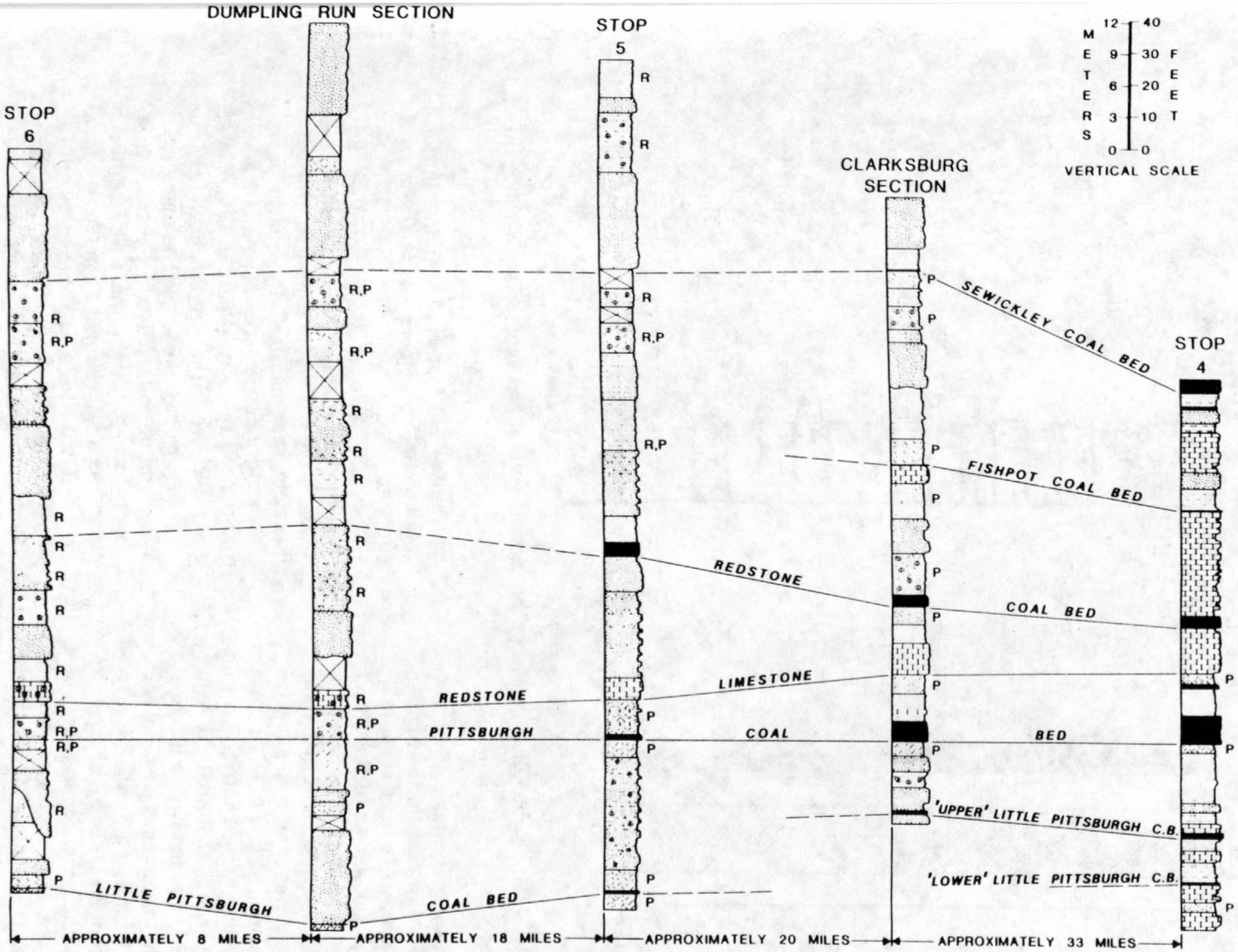


Figure 18) Selected stratigraphic sections showing the thinning and ultimate loss of the Pittsburgh coal bed towards the southern part of the field trip area (left side of diagram).

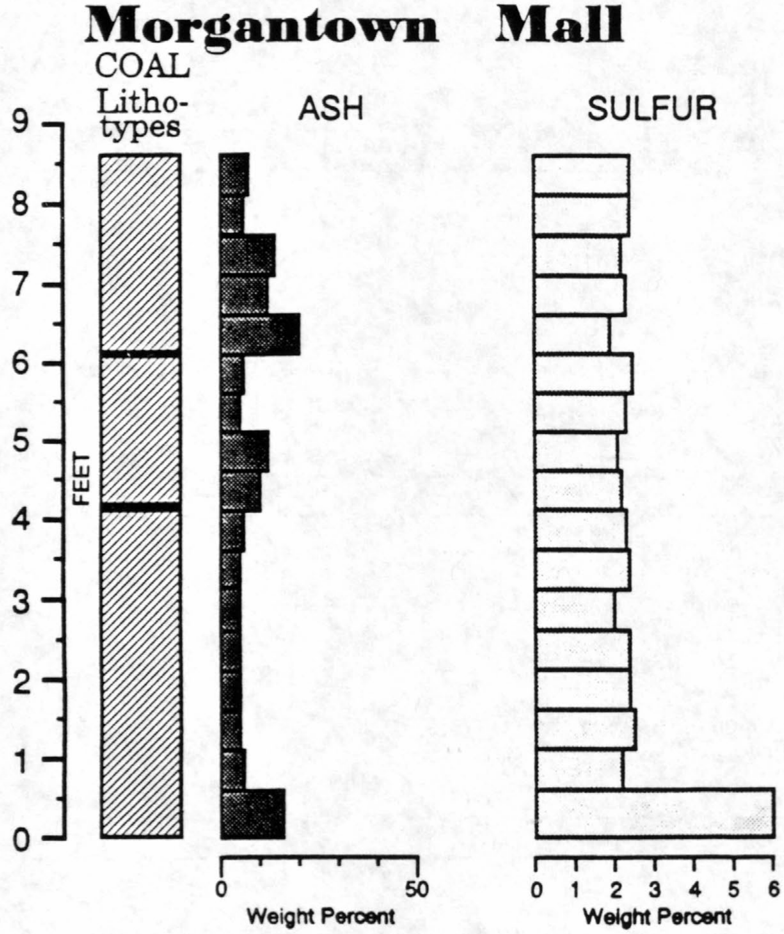


Figure 19 - Ash yield and sulfur distribution in the Pittsburgh coal bed exposed at the Morgantown Mall cut.

the east the Pittsburgh coal is thicker, lower in ash-yield and sulfur than in the Morgantown area, and appears to have been, except for the basal high ash and sulfur bench, influenced entirely by rainfall. The paleoclimate during accumulation of the Pittsburgh coal was therefore, wet enough to allow initial swamp development and to allow parts of the swamp to be maintained entirely on rainfall, but not wet enough to inhibit the regular influx of surface and groundwater into the swamp along the western margins. Subsequent stops (**STOPS 5** and **6**) will be in areas where the Pittsburgh coal bed is thin, or completely absent. Discussions at these stops will correlate these thin Pittsburgh coal occurrences with the coal in the Morgantown area and evaluate the effect of paleotopography on the development of the Pittsburgh swamp and contemporaneous paleosols.

The Redstone coal exposed at this stop shows a channel scour cut which eroded through the sediments deposited over the Redstone peat and partially into the peat. The eastern portion of the channel rises in the cut and the central portion is visible at road level, while the western side of the channel dips below road level. The coal was incrementally sampled at the center of the channel where the thickness was 0.6 m (1.95 ft) and as a full channel 21 m (70 ft) to the east where 0.7 m (2.3 ft) of coal remain (fig. 22). Thickness of the coal away from the channel in the high wall is approximately 1.2 m (4 ft). The lower 0.5 m (1.6 ft) of the coal displays typical ash yield and sulfur distribu

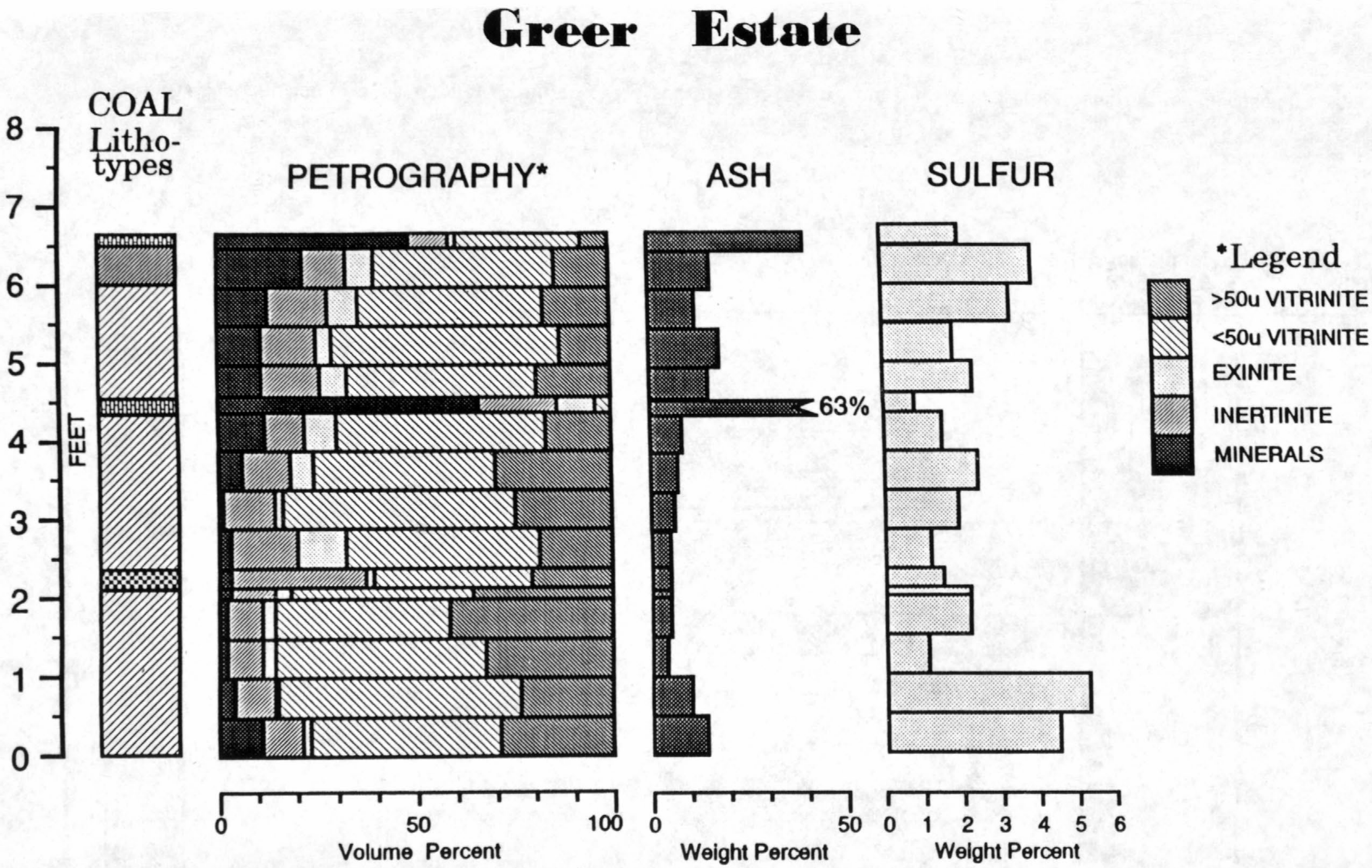


Figure 20) Distribution of macerals, ash yield and sulfur content in a column of Pittsburgh coal taken from the Greer Estate. See text for discussion

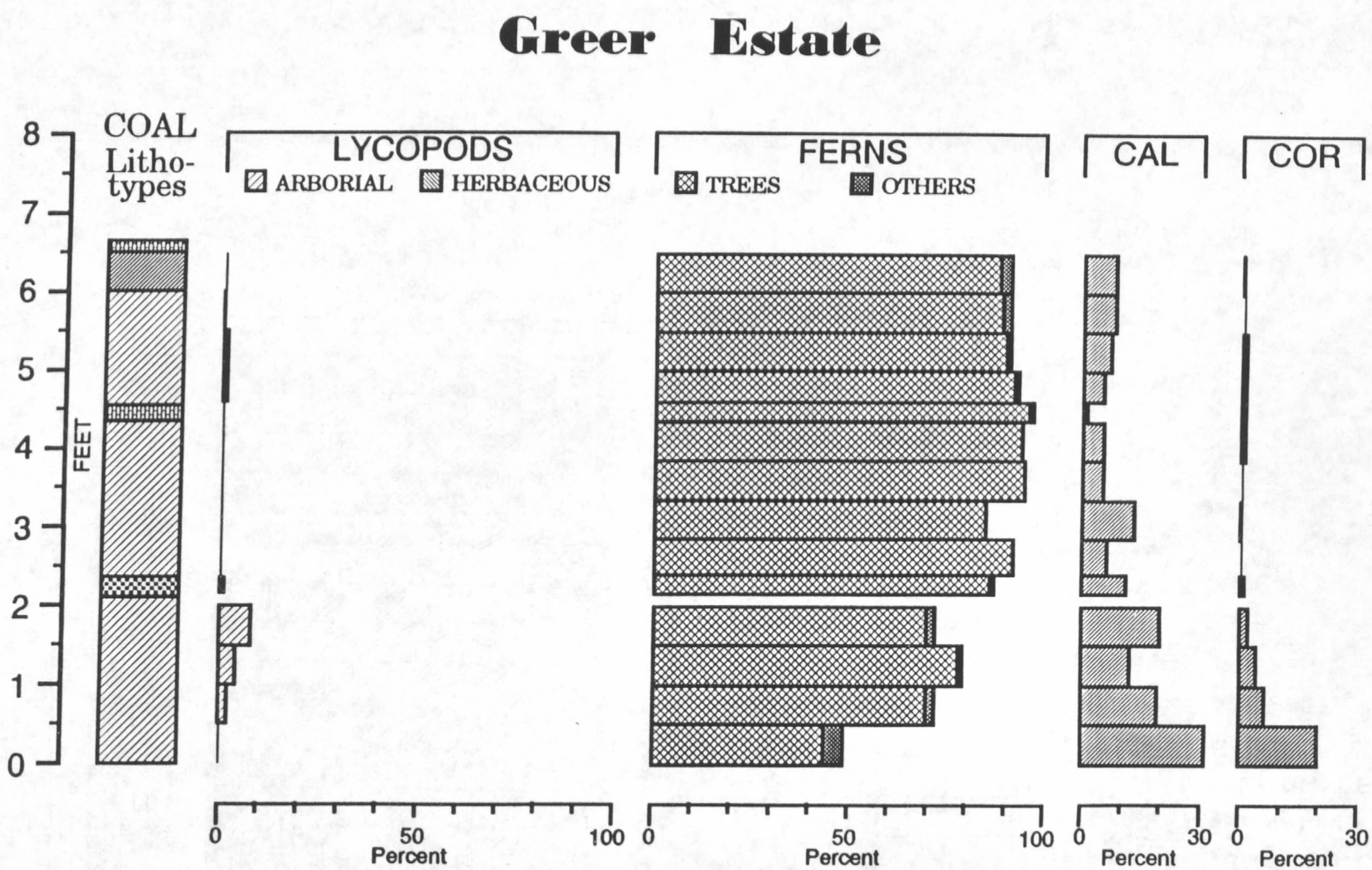


Figure 21) Distribution of miospores in a column of Pittsburgh coal taken from the Greer Estate. See text for discussion.

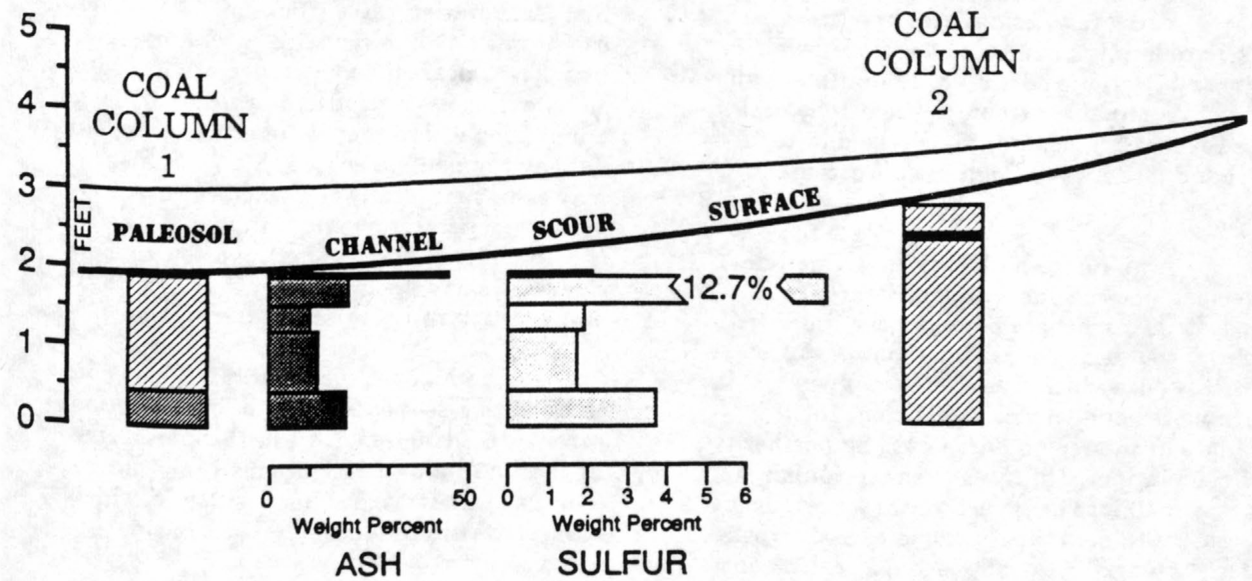


Figure 22 - Ash yield and sulfur content distribution in the Redstone coal bed at the Morgantown Mall cut.

tions for the Redstone coal in this area with high ash and sulfur in the basal increment and decreased ash and sulfur upward. The top 1.5 cm (0.6 in) of coal below the channel is high-ash bone. The coal immediately below the bone layer was moderate in ash yield, but extremely high in sulfur. The increased sulfur, probably in the form of pyrite, appears to have been emplaced after erosion of the channel, from solutions that originated in the channel fill. The channel is unique in that it was not filled by contemporaneous sediments; instead the bottom contains a 0.6 m (2 ft) thick paleosol with carbonate nodules. This unit pinches out toward the channel margin. Later channel fill sediments include alternating fresh water limestones and shales.

A palynological analysis of a column of Redstone coal, located approximately 6 km (10 mi) to the north indicates that this coal bed, like the Pittsburgh, is also dominated by tree fern spores overall, but that palynoflora stratification can be detected. Typically, basal layers contain increased percentages of *Endosporites globiformis*, which was produced by *Chaloneria*, a shrubby, herbaceous lycopsid (Pigg and Rothwell, 1982). Increased percentages of gymnosperm pollen (mostly *Florinites*, *Vesicaspora* and *Pityosporites*) are also common in this part of the bed. In areas where the Redstone coal bed is thin, this miospore assemblage commonly occupies the entire bed thickness (Eble, 1985). In contrast, middle and upper layers of the Redstone coal bed typically show a change to a tree fern spore-dominant palynoflora (*Punctatisporites minutus*), although increased amounts of calamite spores (*Laevigatosporites* spp.), recognized by making an additional *Punctatisporites minutus* - free count, are observed (Grady and Eble, 1989). In addition, several small fern spore genera (e.g. *Leiotriletes* and *Triquitrites?*) are locally abundant at various levels in the bed (Habib, 1968; Eble, 1985). In a majority of its minable extent, the Redstone coal bed is low to moderate in ash yield (5 to 15 %) and moderate to high in sulfur content (2 to >5 %).

The Redstone Limestone, a well-developed non-marine, regionally extensive, lacustrine carbonate occurs above the Pittsburgh coal bed. Lacustrine limestones such as this one, which was first seen at **STOP 2**, 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 non-calcareous shales. The non-marine fauna, not readily observed macroscopically, is diminutive and low in diversity (Dickson, 1977). Ostracodes are the most abundant fossil remains in these beds. Also reported are remains of gastropods, bivalves, the worm tube *Spirorbis*, the trace fossil *Lockeia*, and vertebrate fragments (Dickson, 1977). No marine fossil forms have been reported to date, but conditions may have fluctuated from fresh water to brackish. For that reason, the inclusive term "non-marine" is used as advocated by Dickson (1977). Spar calcite occurs in dismicritic textures, within ostracode carapaces, and as minor filling around intraclasts. These carbonate complexes formed in shallow water lakes and paludal environments (Dickson, 1977), the cyclicity of units suggesting considerable fluctuation in shoreline and base level (Donaldson, et al., 1985). Dolomite, as determined from both x-ray diffraction and thin section analysis, has been reported from the stratigraphically higher Benwood Limestone (above the Sewickley coal bed, not exposed at this stop). The microscopic dolomite occurs as euhedral to subhedral grains, and is suggested to be the result of evaporative conditions (Marrs, 1981). Photosynthesis by algae presumably played a major role in precipitation of the calcium carbonate in these lakes, with additional enhancement of this process by high evaporation rates (Dickson, 1977). They can be seen on outcrop with careful observation. The calcareous mudstones are interpreted as paleosols developed on calcareous lacustrine muds where they were exposed in paludal environments (Dickson, 1977). Shales may represent locally reworked calcareous mud, or detrital input from fluvial systems (fig. nf2 16). The Redstone Limestone generally occurs as a monolithic micrite, the result of deposition in an aerially extensive lake.

The Fishpot coal is only 2.5 cm (1 in) 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 in the Dunkard Basin.

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 to 7.5 ft). Another clastic interval occurs above the Sewickley Limestone. Thin sandstone and shale beds are interbedded with the Sewickley coal bed here. The association of the Sewickley coal with fine to coarse clastic rocks is characteristic basin-wide. Lower coal benches or "splits" of the Sewickley are sometimes miscorrelated with the underlying Fishpot coal bed. The main minable bench of the Sewickley coal bed is not well-exposed at this location. The weathered blossom, 1.2 m (4 ft) thick, can be seen at the top of the section at the extreme western end of the cut. This bed has also been extensively mined underground in the Morgantown area.

"ROLLING STOP" - Paleosols in the Pittsburgh Shale, Conemaugh Group at exit 146 along I-79.

Upper Pennsylvanian strata assigned to the Conemaugh Group are exposed in a roadcut along I-79 adjacent to the Goshen Road exit (exit 146) in Monongalia County, West Virginia. The cut exposes approximately 34 m (110 ft) including the following units, in descending order: Grafton sandstone, Ames Limestone, Harlem coal bed and Pittsburgh Shale sequence.

The Grafton sandstone, exposed at the top of the section, marks the base of the Casselman Formation, and consists of interbedded siltstone and sandstone. Strata below the Grafton are assigned to the Glenshaw Formation. The Ames Limestone and shale occurs directly below the Grafton sandstone and represents a major marine transgression in the Dunkard Basin. As such, the Ames serves as an important lithostratigraphic correlation unit across the basin. The Ames is generally an impure, shaley limestone to calcareous shale, and is locally fossiliferous.

The Harlem coal bed, which underlies the Ames Limestone and shale, is a thin, generally less than 0.6 m (2.0 ft thick), laterally persistent unit that occasionally attains minable 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 contents commonly are moderate to high. Other major Conemaugh Group coal beds include the Mahoning, Bakerstown, Elk Lick and Little Clarksburg.

The unit underlying the Harlem coal bed at this stop consists of alternating beds of impure limestone and variegated red/green claystone that often contains calcium carbonate nodules. This unit, referred to as the Pittsburgh red beds, contains features indicative of repeated and (or) prolonged subaerial exposure. These features include calcareous peds, vertic, cone-in-cone and pseudoanticlinal structures, which indicate that the Pittsburgh red beds are a series of ancient seasonally "dry" paleosols. The lateral persistence of this unit across the Dunkard Basin indicates that climatic and depositional conditions necessary for its formation were of basinal extent.

STOP 5: Upper Conemaugh and lower Monongahela Group strata exposed behind the Weston Shopping Center, just east of the I-79 Weston interchange, #99.

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

Upper Conemaugh and lower Monongahela Group rocks are exposed at numerous locations surrounding the I-79 Weston interchange. At **STOP 5**, 25 m (82.2 ft) of strata, beginning in the upper Conemaugh Group about 8.2 m (27 ft) below the Pittsburgh coal bed, are seen in a cut slope in the southeastern corner of the Weston Shopping Center. The Little Pittsburgh coal bed, Pittsburgh coal bed, Redstone Limestone, and Redstone coal bed are important markers for mapping in this region (fig. 23). The Redstone Limestone is the only major carbonate bed found in the section here, in contrast with the numerous beds observed at **STOP 4**, and so serves as the most valuable marker bed, aiding in correct correlation of the coals. In this area we see the first appearance of red shales and paleosols in the lower Monongahela Group and a higher percentage of coarse clastics than observed at **STOP 4** or the Clarksburg Section. This change will be even more pronounced at **STOP 6** as we continue the traverse from swamp-lacustrine deposition in the north to alluvial plain and fluvial deposition in the south.

The Pittsburgh coal bed at this stop is 0.3 m (1 ft) thick and is generally less than 0.6 m (2 ft) thick in this region, which is out of the major mining district (fig. 17). The 1 m (3.2 ft) thick, light gray claystone beneath the Pittsburgh at **STOP 5** is interpreted as a paleosol that was subaerially exposed for a substantial period of time before the Pittsburgh peat swamp developed at this location. Pedogenesis at **STOP 5** appears to have been coeval with peat-swamp development in the more central parts of the basin (e.g. **STOP 4**). In contrast to the well-developed paleosol at STOP 5 where the coal bed is thinning, the subjacent paleosol sequence at **STOP 4** is

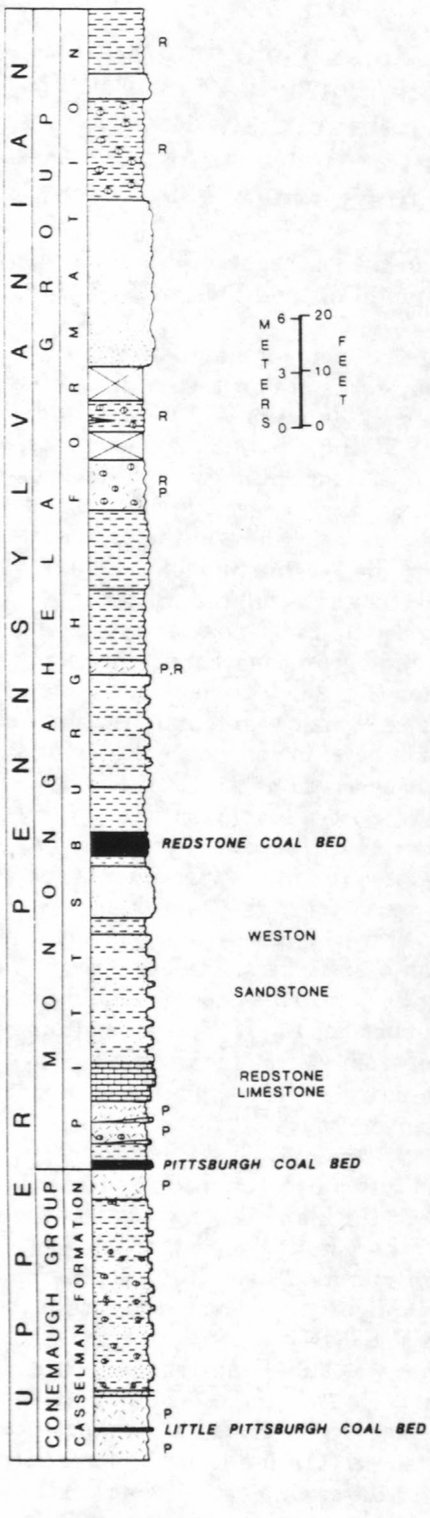
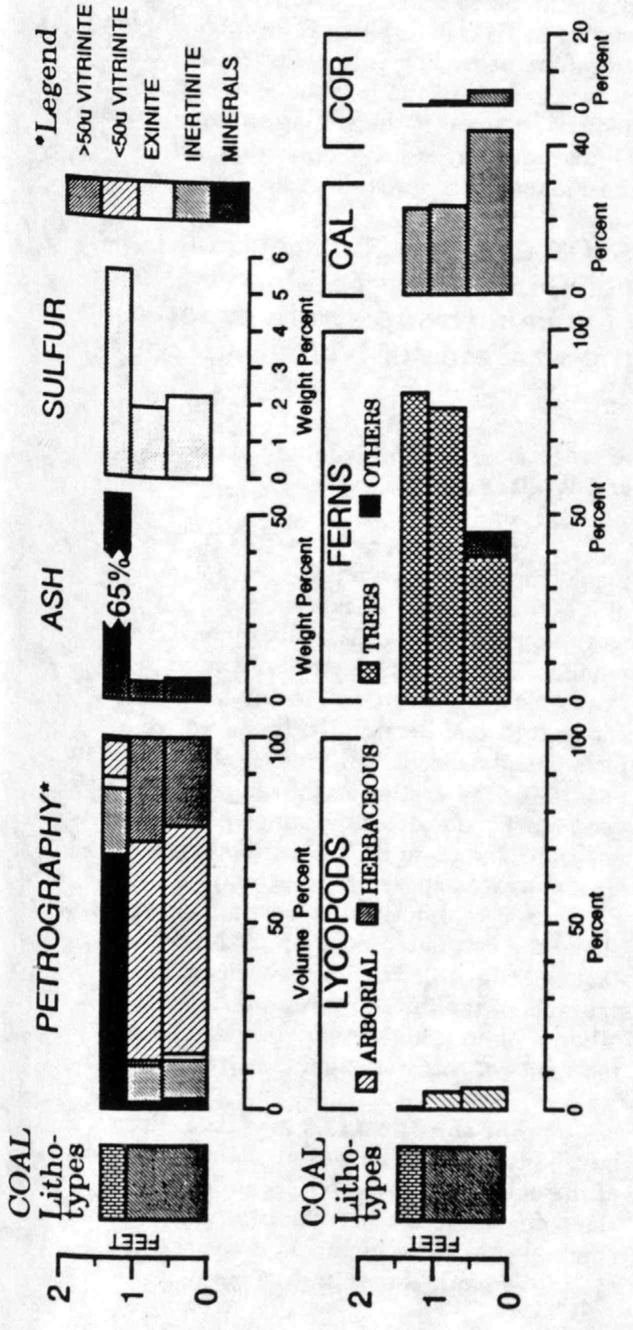


Figure 23 - Upper Conemaugh and lower Monongahela strata exposed in the area of the I-79 Weston exit (# 99). The column is a composite of sections from the local area.

thin and very poorly-developed where the Pittsburgh coal bed is thick. This inverse relationship between thick coal and thin, poorly-developed, underclay or claystone, and thin coal and well-developed underclay (claystone), is a common occurrence for a given coal bed as noted by Patterson and Murray (1984). We herein interpret such a relationship to be the result of prolonged chemical weathering and pedogenesis on well-drained topographic highs coeval with peat formation in topographic lows under a relatively wet climate.

The Pittsburgh coal at this stop is only 0.3 m (1 ft) thick with an overlying 12.1 cm (4.8 in)thick bone coal layer. For an Upper Pennsylvanian coal of this thickness, the Pittsburgh at this location is remarkably low in ash yield, though moderate to high in sulfur. Petrographic analyses reveal a high vitrinite content with moderate amounts of >50 micron vitrinite (fig. 24). The palynoflora is dominated by tree fern spores, but includes moderate percentages of calamite spores and minor percentages of arboreous lycopsid spores. The high vitrinite content, moderate preservation of the pre-vitrinite plant debris, low ash yield and relative abundance of calamites indicate that this peat probably accumulated in a planar swamp with a consistent standing water cover. This thin coal is similar to, and may correlate with the low ash, vitrinite rich trend displaying the same palynological signature 0.3 to 0.6 m (1 to 2 ft) from the base of the Pittsburgh coal at the Greer estate (figs. 20 and 21). The basal high ash yield and high sulfur content 0.3 m (1 ft) of the Pittsburgh coal, which is generally found throughout the aerial extent of the Pittsburgh coal bed, is apparently absent at this location. The paleosol developed below the coal here apparently was subaerially exposed during accumulation of the inital stage of the Pittsburgh paleo-peat elsewhere. The ash, sulfur, petrographic and palynologic data suggest that the well-developed Pittsburgh paleo-peat swamp transgressed over this paleotopographic high with no discernable compositional or floral changes in the peat. Once established, the peat swamp was terminated before the more tree fern domi



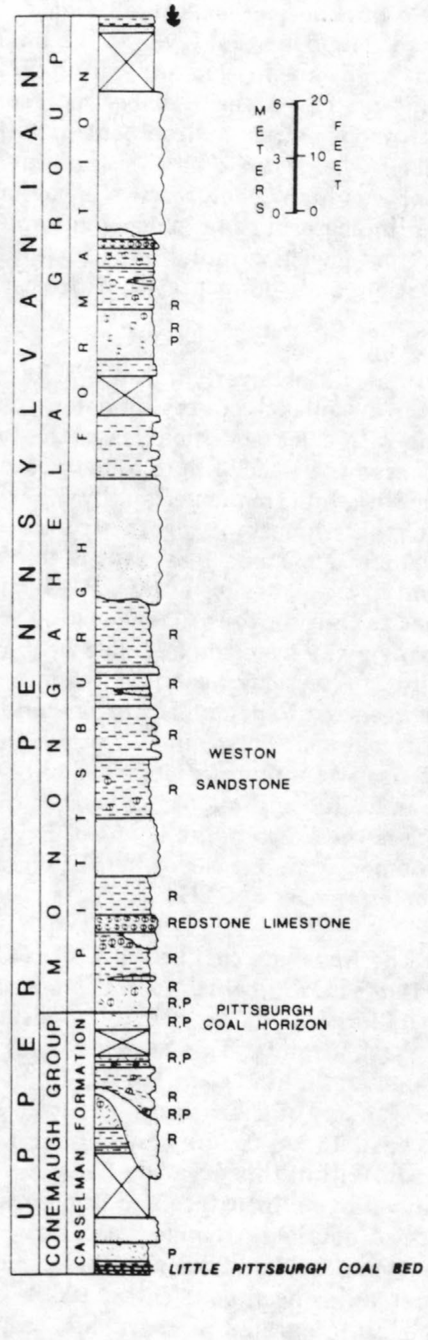


Figure 24 (left) Distribution of macerals, miospores, ash yield and sulfur content in the Pittsburgh coal bed at **STOP 5**. Figure 25 (right) Stratigraphic column of Upper Pennsylvanian strata at **STOP 6** where the Pittsburgh coal is represented by a paleosol.

nant palynoflora, which characterizes the middle and upper parts of the Pittsburgh coal in minable thickness areas, could become established. The bone coal layer at the top of the coal is high in ash yield and very high in sulfur content. This layer probably represents a sediment-poor marsh environment in which plant debris was destroyed and sulfur emplacement was high. The presence of the bone coal layer indicates a rapid paleoenvironmental shift from peat accumulation to non-peat accumulation, and sedimentation at this location.

Immediately overlying the Pittsburgh coal bed is a sequence of gray mudstone and claystone with calcareous nodules in the top. Above these beds is a 1.1 m (3.5 ft) thick dense micritic limestone overlain by 0.8 m (3.7 ft) of shaly limestone together comprising the Redstone Limestone. This unit is thinner here than in the exposure at **STOP 4**, and continues to thin southward, suggesting we are moving away from the lake depocenter. Ascending the section, a sandstone unit occurs between the Redstone Limestone and Redstone coal bed. Throughout much of the basin, these two units are separated only by a thin mudstone or shale (e.g. **STOP 4**) (fig. 16). This is the Weston Sandstone of Reger (1916) named from exposures in Weston, a few kilometers east of **STOP 5**.

The Redstone coal bed is 1.3 m (4.2 ft) thick at **STOP 5**, which is in the middle of one of the larger bodies of minable Redstone coal in West Virginia. This minable field extends about 20 km (12 mi) north, to just south of the city of Clarksburg, about 26 km (16 mi) east, 13 km (8 mi) west and 8 km (5 mi) south. Within this area the Redstone has been mined by both surface and underground methods. A detailed palynological study of the Redstone coal bed from several locations in this area has been done (Eble, 1985). In addition, an integrated petrographic, palynologic, geochemical and mineralogic analysis has also been done at one of those locations (Grady and Eble, 1989).

The section above the Redstone coal bed, exposed in other nearby roadcuts consists of sandstones, and red and gray shales and mudstones that are equivalent to the section at **STOP 4**, which contains abundant limestone beds. The only occurrence of carbonate here is in the form of concretions or nodules in some of the shales and mudstones. This area is beyond the depositional limit of the Fishpot and Sewickley coal beds (fig. 18).

STOP 6: Burns Run Section, Upper Conemaugh and lower Monongahela Group strata exposed on the west side of I-79 at milepost 72.8.

Stop Leaders: Nick Fedorko, Blaine Cecil, and William Grady

Approximately 70 m (228 ft) of equivalent Upper Pennsylvanian strata examined at **STOPS 4 and 5** is exposed here on the west side of I-79 (fig. 25). The upper Conemaugh Group rocks seen here extend from the Little Pittsburgh coal to the base of the Pittsburgh coal horizon. In this area we are beyond the depositional limit of other marker beds, thus the Little Pittsburgh coal bed is a good marker for determining stratigraphic position. The 56 m (162 ft) of Monongahela Group strata exposed here extend from the Pittsburgh coal horizon to about 12 m (40 ft) above the estimated position of the Sewickley coal bed (fig. 18). This exposure is a good example of the facies representative of the alluvial plain and fluvial depositional environments of the Monongahela Group.

The thin 4 cm (1.5 in) Little Pittsburgh coal bed is exposed at the northern end of the cut just above ditch level. In the cut slope above the Little Pittsburgh coal bed are channel sandstone beds and associated autocyclic overbank shales and paleosols.

The Pittsburgh coal bed is represented here by a 3.8 cm (1.4 in) thick, dark gray to black, carbonaceous claystone or shale. Extremely thin, bright carbonaceous bands are apparent in good light. It has a slight blocky habit. A petrographic analysis of this unit showed that it is mainly comprised of fragmented inertinite along with minor small fragments of vitrinite and exinite macerals, all within a mainly kaolinite matrix.

This unit is interpreted as a histolic epipedon, part of the subjacent paleosol. Immediately beneath this dark shale is a thin, light gray clay (2.5 cm, 1 in), underlain by a gray-red, light gray-green, and light gray mottled, noncalcareous mudstone. Although poorly-exposed on the cut bench, this paleosol is estimated to be 2.6 m thick (8.5 ft). Little or no carbonate is found in this paleosol, probably having been leached out under the wet climatic conditions that produced a thick peat down dip (**STOP 4**). In contrast, the 2.1 m (6.8 ft) thick paleosol above the histic epipedon is mottled gray-red, yellow-gray, and light gray, contains calcareous nodules and lenticular beds of carbonate and is believed to have formed under dry or wet-dry conditions. Paleoclimate was, therefore, a major allocyclic control on the lithostratigraphy of the two stacked toposequences formed under contrasting climatic conditions (Cecil, 1990). The lower toposequence, consisting of the Pittsburgh coal bed and laterally equivalent paleosol, evolved under the wet part of a climate cycle. A Histosol formed in the deeper, more central part of the basin to the north, on top of a thin, poorly-developed mineral horizon (**STOP 4**) (fig. 16). To the south on the better-drained alluvial plain, high terrestrial organic productivity induced by the wet paleoclimate led to the development of a histic epipedon on a thicker mineral paleosol horizon. This topographic position was too well-drained for peat formation and preservation. Also having formed under wet conditions, the lower mineral paleosol horizon is devoid (leached) of appreciable carbonate, and is mottled red, gray and green (figs. 18 and 25). The upper toposequence (paleosol) directly on top of the Pittsburgh coal bed at **STOP 4** (fig. 16), and above the dark claystone at **STOP 6** (fig. 25) developed under a dry or seasonal wet-dry part of a climate cycle. Unlike the lower sequence, the upper toposequence contains carbonate nodules and thin petrocalcic horizons and other evidence of a dryer climate. It is very similar in structure and characteristics across the inferred change in topographic relief, except that it becomes more red in the alluvial plain setting.

The interval above the Pittsburgh coal bed, including the Redstone Limestone, is very similar to that seen at **STOP 5**, with the important distinction that here we see considerably more reddening of the sediments. This indicates better oxygenation, largely as a result of the development of vadose zones. The Redstone Limestone here is a micritic bed, thinner than the last stop, varying from 0.8 to 1.2 m (2.7 to 4 ft) thick. It is very nodular, with a matrix of calcareous red material (fig. 25). This is interpreted as pedogenic overprinting of a micrite deposited in a lacustrine environment and subsequently exposed subaerially. These features indicate that **STOP 6** was very near the edge of the Redstone "lake". Beneath the nodular micrite is a zone up to 0.9 m (3 ft) thick of abundant calcareous nodules in a red shale. This zone sometimes coalesces with the overlying micrite bed, suggesting that this zone was also once a coherent bed of limestone that has undergone a pedogenic process of nodularization. In the north half of the cut, this lower nodular zone is occupied by a fine grained micaceous sandstone. Evidence for pedogenic overprinting of this bed includes drab halo root traces, calcareous nodules and indistinct contacts.

The section above the Redstone Limestone is a rather monotonous sequence of the following sediment types: sandstones, interbedded sandstones and red and gray shales, red, green and gray shales and mudstones (paleosols), and some nodular limestone zones, although these may be petrocalcic horizons. Without the presence of the Little Pittsburgh coal bed, the Pittsburgh coal horizon, and the Redstone Limestone, correlation of these units in this cut would be problematical. The same sequence containing these three beds is observed in the Dumpling Run Section (fig. 18) and also in other cuts between here and the town of Burnsville.

Birch River-Powell Mountain Section: Introduction to STOPS 7 and 8.

Bascombe M. Blake, Jr.

Road cuts along U.S. Route 19, between the town of Birch River and the top of Powell Mountain, Nicholas County, West Virginia, contain excellent exposures of strata assigned to the Middle Pennsylvanian Kanawha and Allegheny Formations, and the lower portion of Upper Pennsylvanian Conemaugh Group (?) (fig. 26). Two stops are scheduled for this section. **STOP 7** is near the base of Powell Mountain where we will examine the Dingess Shale Member of the Kanawha Formation, an aerially widespread marine sequence important for regional stratigraphic studies. **STOP 8** is at a scenic overlook in the lower portion of the Allegheny Formation. Three large road cuts are visible from the overlook and offer a panoramic overview of the upper portion of the Kanawha Formation and the lower portion of the Allegheny Formation.

The upper 152 m (500 ft) of the Kanawha Formation crops out in road cuts ascending the mountain. The total thickness of the Kanawha Formation is 213 m (700 ft) in this area of West Virginia and represents an expansion of some 350 percent from the 61 m (200 ft) present on Chestnut Ridge (**STOP 1**). The northward thinning of the Kanawha Formation, as it onlaps the North American craton from the south, generally occurs by reduction of the section above the Winifrede Shale Member, and by the progressive northward wedging-out of the strata below the Betsie Shale Member, which is the stratigraphically lowest Middle Pennsylvanian marine zone to transgress northward onto the Mississippian-Pennsylvanian unconformity (see fig. 4). Three of the four aerially extensive Kanawha Formation marine zones have been recognized in this section; they are, in ascending order, the Dingess Shale Member, the Winifrede Shale Member, and the Kanawha Black Flint of White (1891). The fourth important marine zone, the Betsie Shale

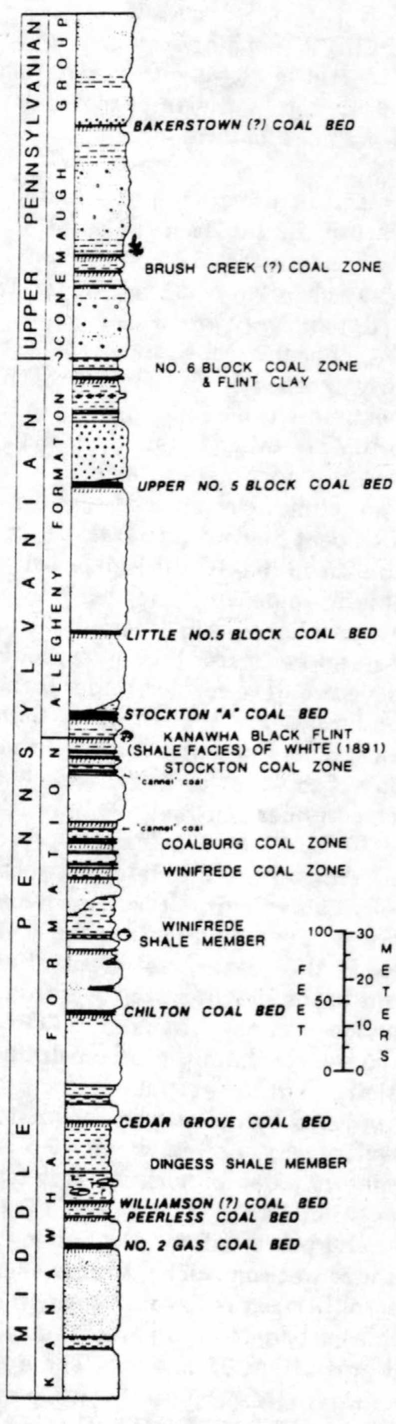


Figure 26 - Stratigraphic column of Middle and Upper Pennsylvanian strata exposed along Route 19 between Birch River and the top of Powell Mountain.

Member, is below drainage at the base of

Powell Mountain. These four marine zones have been shown by workers in West Virginia (Blake et al., 1989, in press) and Kentucky (Rice and Smith, 1980; Chesnut, 1988, 1991) to be very useful in making consistent correlations between widely spaced sections. An example of this is Chesnut (1988), who elevated the Breathitt Formation of eastern Kentucky to group status and used equivalent marine units to subdivide the Breathitt into formations.

The entire Allegheny Formation is exposed in road cuts near the top of Powell Mountain and is 85 m (280 ft) thick in this portion of West Virginia. The Allegheny Formation in central and southern West Virginia is dominated by thick, multistoried fluvial-deltaic channel sandstones, with minor amounts of siltstone, mudrock, coal beds, and flint clays. The sandstones are mainly litharenites, but quartz arenites are locally common above the Upper No. 5 Block coal bed in the Nicholas - Braxton county area.

A regionally persistent flint clay of pedogenic origin is present near the top of the Allegheny Formation at the level of the No. 6 Block coal bed (fig. 26). A second, less aerially persistent pedogenic flint clay horizon has been identified from the Upper No. 5 Block horizon from coal exploration diamond drill cores in the Nicholas-Braxton county area. The facies association of the No. 6 Block coal bed with a regionally persistent flint clay suggests that while peat was accumulating in lowlands under wet (everwet ?) climate conditions, the better drainage afforded by increased topographic relief in the areas surrounding the peat-forming areas allowed for intense leaching over extended periods under the tropical conditions prevalent during late Middle Pennsylvanian times. The elevation difference between the peat and the surrounding flint clay-forming environments may have been slight.

The lower 73 m (240 ft) of Upper Pennsylvanian Conemaugh Group strata are exposed in road cuts near the top of Powell Mountain. The lower portion of the Conemaugh Group is dominated by massive, conglomeratic (mainly quartz pebbles) sandstones with mudrocks and thin, low quality coal beds. Shale and mudstone beds in this interval have a slight red- and green-gray tint, but only a few of the numerous drill logs examined report red beds from the lower portion of the Conemaugh Group in the Powell Mountain area. This condition is similar to that observed in time-equivalent strata in northern West Virginia. The thin coal beds present at the top of the Birch River-Powell Mountain section have been tentatively correlated with the Bakerstown(?) and Brush Creek (?) coal beds. The marine units, common in the Conemaugh Group to the north and west, have not been reported from the lower portion of the Conemaugh Formation in this immediate area, hindering correlation efforts. Due to erosion, the total thickness of the Conemaugh Group is not ascertainable, which also hinders correlation of units. The current placement of the Allegheny - Conemaugh boundary (fig. 26) in the Powell Mountain area is based on preliminary data.

STOP 7 - Dingess Shale Member, Kanawha Formation

Stop Leaders: Mitch Blake and Cortland Eble

STOP 7 is near the bottom of Powell Mountain at the Dingess Shale Member of the Kanawha Formation (Kendrick Shale Member, Breathitt Formation). This marine zone, approximately correlative with a marine zone near **STOP 1** in the Chestnut Ridge Section (fig. 2) (Donaldson and Eble, 1989), has been mapped across southern West Virginia, eastern Kentucky (Kendrick Shale Member, Breathitt Formation), and southwestern Virginia (Kendrick Shale Member, Wise Formation) (Rice and Smith, 1980; Chesnut, 1988; Blake et al., in press). The Dingess Shale Member at this locality consists of thinly interbedded to interlaminated shales, siltstones, and fine sandstones. Burrows are locally common and bedding planes often contain scattered plant debris. Invertebrate body fossils have not been reported from the Dingess at this locality but have been reported from nearby areas. Impure, ellipsoidal limestone concretions up to several feet wide are common, as are nodules and diffuse bands of siderite. A transgressive lag deposit, present at the base of the marine sequence, marks the presence of a ravinement surface. Similar lag deposits are common at or near the base of many of the Kanawha Formation marine zones. Here, the transgressive lag consists of a poorly-bedded, sideritic, calcareous, highly bioturbated, very fine- to fine-grained sandstone and ranges from a few centimeters to over 0.3 m (1 ft) in thickness. Mica flakes are common as are shale and coal clasts (rip-ups). The lag could be termed a shale-clast conglomerate due to the local abundance of shale clasts. Ichnogenera identified from the lag consist of *Planolites*, *Teichichnus*, and *Rhizocorallium* (?). Several unidentified traces are also present including one suggestive of *Conostichus*. The base of the lag marks the ravinement surface formed by erosion of the shore-face as coastal facies are transgressed (Swift, 1968; Martino, 1989, 1990a, 1990b, this volume). The transgressive lag overlies a rooted, silty mudstone (seat earth), which suggests an erosional contact between the underlying sediments and the lag. This is reinforced by the presence of the Williamson (?) coal bed above the rooted, silty mudstone in exposures to the north. The lag grades upward into the interlaminated/interbedded silts and sands which comprise the bulk of the marine zone. Near the top of the highwall the member has coarsened into a thinly-bedded, very fine-grained sandstone. This facies contains occasional *Planolites* and *Teichichnus* traces, ripple cross-laminations, small-scale hummocky cross-stratification, and small-scale cut and fill structures suggesting deposition between fair-weather and storm wave bases. The top of this facies is rooted and contains stigmarian axes indicating a brief period of subaerial exposure. A 9.9 m (32.5 ft) thick black shale with a thin, basal sideritic sandstone directly overlies the rooted surface, suggesting a return to more marine-influenced conditions. A detailed discussion of the various facies found in the Dingess Shale Member is presented by Martino (1989, this volume).

STOP 8 - Kanawha - Allegheny Formation Contact

Stop Leaders: Mitch Blake and Cortland Eble

STOP 8 is at a scenic overlook approximately at the Lower No. 5 Block coal horizon (not present at this locality). The Upper No. 5 Block coal bed can be seen in the large road cut immediately to the south (uphill) of the overlook. The Allegheny Formation in this portion of West Virginia is dominated by thick, multistoried, lithic sandstones deposited in an upper delta plain or alluvial plain (Donaldson and Shumaker, 1981). At this locality, the Little No. 5 Block coal bed, slightly downhill of the entrance to the overlook, is a splint coal, typical of Allegheny coal beds in central and southern West Virginia. Strata exposed in the large road cut, visible down the mountain to the north, are assigned to the uppermost portion of the Kanawha Formation and the lower portion of the Allegheny Formation. The contact between the Kanawha and Allegheny Formations is placed at the top of the Kanawha Black Flint of White (1891), represented in this section by its shale facies with a restricted fauna consisting of phosphatic brachiopods (lingulas and orbiculoids) and sparse calcareous brachiopods. Reppert (1979) and Watson (1992) discuss the various facies present in the Kanawha Black Flint of White (1891), and Blake (this volume) reports on the areal extent of the non-chert facies.

Palynological analyses of coal beds exposed in the Birch River section indicates that coal beds near the base of the section (No. 2 Gas and Cedar Grove) are dominated by *Lycospora* (fig. 27). In contrast, the Winifrede - Coalburg - Stockton coal zone at the top of the Kanawha Formation contains much higher percentages of tree fern and, to a lesser extent, small lycopsid spores. The Williamson coal bed is somewhat intermediate in composition. This type of spore distribution, which is very similar to the stratigraphic change seen in coals along I-68 (**STOPS 1, 2** and **3**) is paralleled by an increase in splint

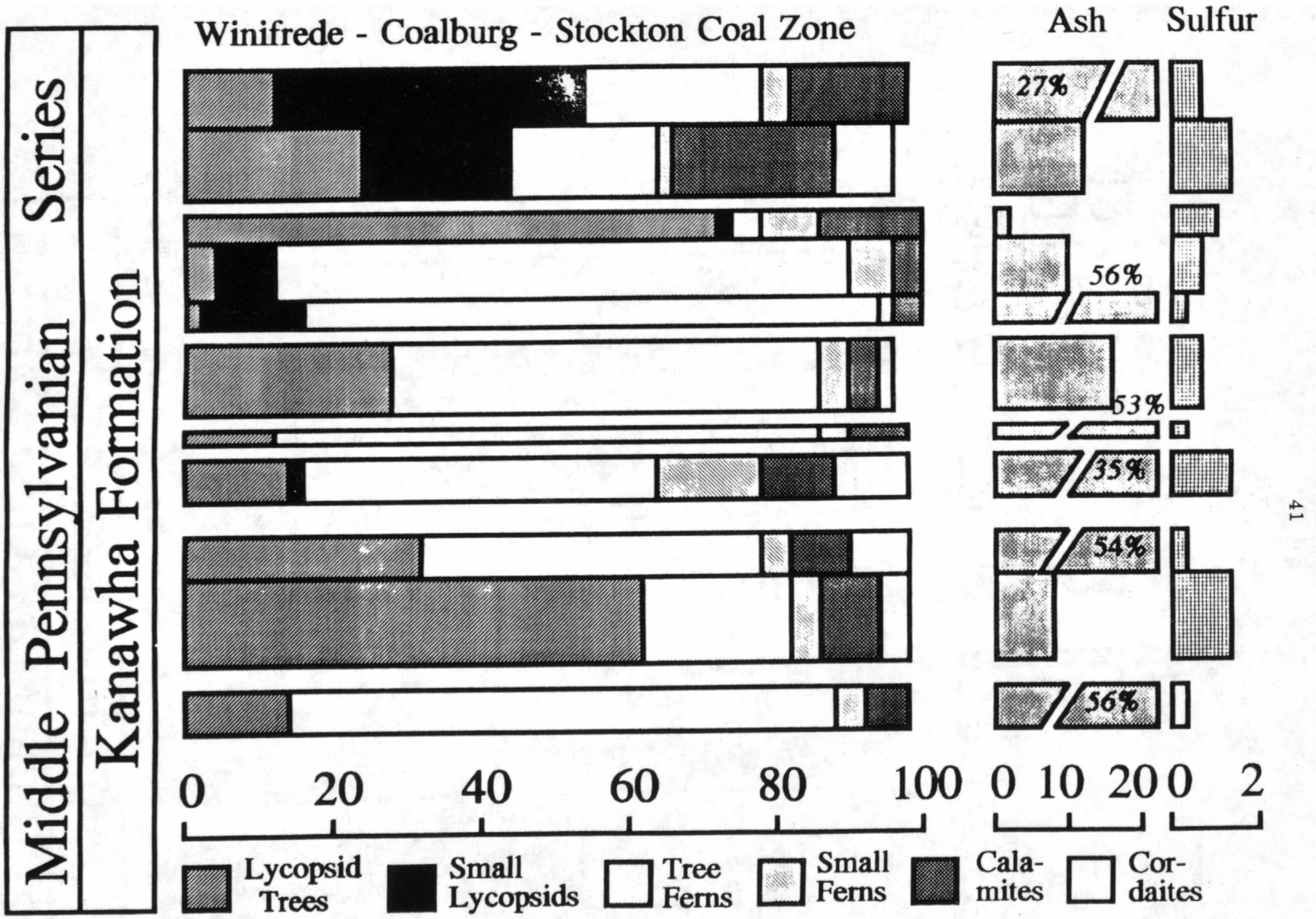


Figure 27) Distribution of miospores, ash yield and sulfur content in coal beds exposed along U.S. Route 19 between the town of Birch River and the overlook near the top of Powell Mountain.

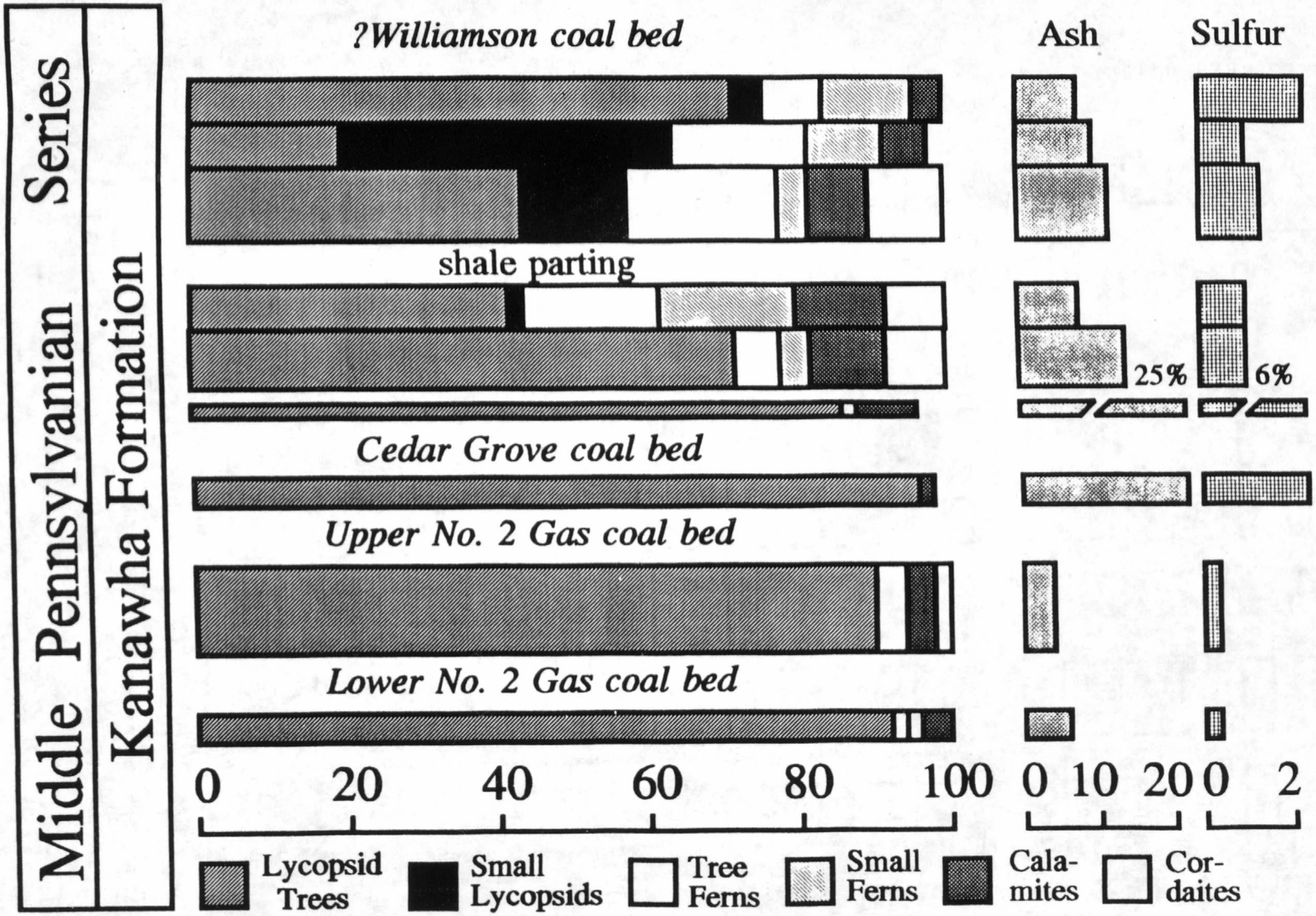


Figure 27 continued...refer to fig. 26 for stratigraphic orientation.

coal (dull coal lithotypes). The frequency of coal bed "splitting", and an increase in coarse-grained sediments also follows this trend. Although palynofloras can only be used to partially reconstruct peat swamp paleofloras, the observed changes may be reflecting a change in swamp biomass composition that was, at least partially, induced by a change in climate. A gradual "drying" trend is inferred for the mid-Middle through Late Middle Pennsylvanian (Cecil et al., 1985; Cecil, 1990). If this is true then an increase in levels of peat oxidation (and subsequent inertinite production to form splint coal), along with a shift from a lycopsid tree dominant flora to one that had increased percentages of tree ferns and small lycopsids might be expected. Increased levels of intermittent surficial peat exposure would, in addition to increasing the inertinite content, be an inhibiting factor for the lycopsid trees, which were better-adapted for growth and reproduction in areas with consistently high moisture levels (to the point of having a standing water cover).

The high frequency of coal bed splitting may also, in part, be climate-related. Elevated levels and periodicities of sediment flux, driven by an increase in "drier", as opposed to "wetter" times (Cecil, 1990), could help explain the increased levels of coal bed splitting in the upper part of the Kanawha Formation. Likewise, the increase in sediment coarseness in this part of the section may also be related to climate.

END OF DAY 1

DAY 2

STOP 9: Camp Creek Interchange Section

Stop Leaders: C. Blaine Cecil and Kenneth J. Englund

STOPS 9 through **11** evaluate Upper Mississippian and Lower Pennsylvanian strata that are missing elsewhere on the trip at the Mississippian - Pennsylvanian unconformity (**STOPS 1, 16, 17, and 18**). The Upper Mississippian Pride Shale Member of the Bluestone Formation, exposed at **STOP 9,** is a widespread and lithically- distinct bed in southern West Virginia, southwestern Virginia, and eastern Kentucky fig. 28). The Pride Shale pinches out a few tens of kilometers to the north, and is absent at **STOP 1**. The underlying Princeton Sandstone and the overlying Glady Fork Sandstone Member of the Bluestone, both consisting mostly of fine- to coarse-grained, calcite-cemented, conglomeratic sandstone, are present in the sequence at **STOP 9** but absent at **STOP 1**.

The Pride Shale Member is nearly 50 m (150 ft) thick and consists largely of dark-gray shale with nodules and concretions of pyrite, siderite, and limestone. Basal beds of the member include about 7.6 m (25 ft) of medium- to dark-gray, sparsely rooted, shale and silty shale. Fresh-water marsh deposition at this locality is indicated by the presence of the rooted beds and by brightly banded coal as much as 5 cm (2 in) thick. This basal unit of the Pride Shale Member appears to grade laterally into uppermost beds of the Princeton Sandstone. The roof shale of the thin coal beds contain partially abraded plant fossils such as *Spenophllum tenerrimum, Stigmaria stellata, Pecopteris aspera, Archaeocalamites* sp. and *Sphenopteris elegans*. This flora is characteristic of the Upper Mississippian of North America and the Namurian A of western and central Europe. It succeeds the *Fryopsis* zone (zone 3 of Read and Mamay, 1964) and precedes the *Neuropteris pocahontas* zone (zone 4 of Read and Mamay, 1964) at

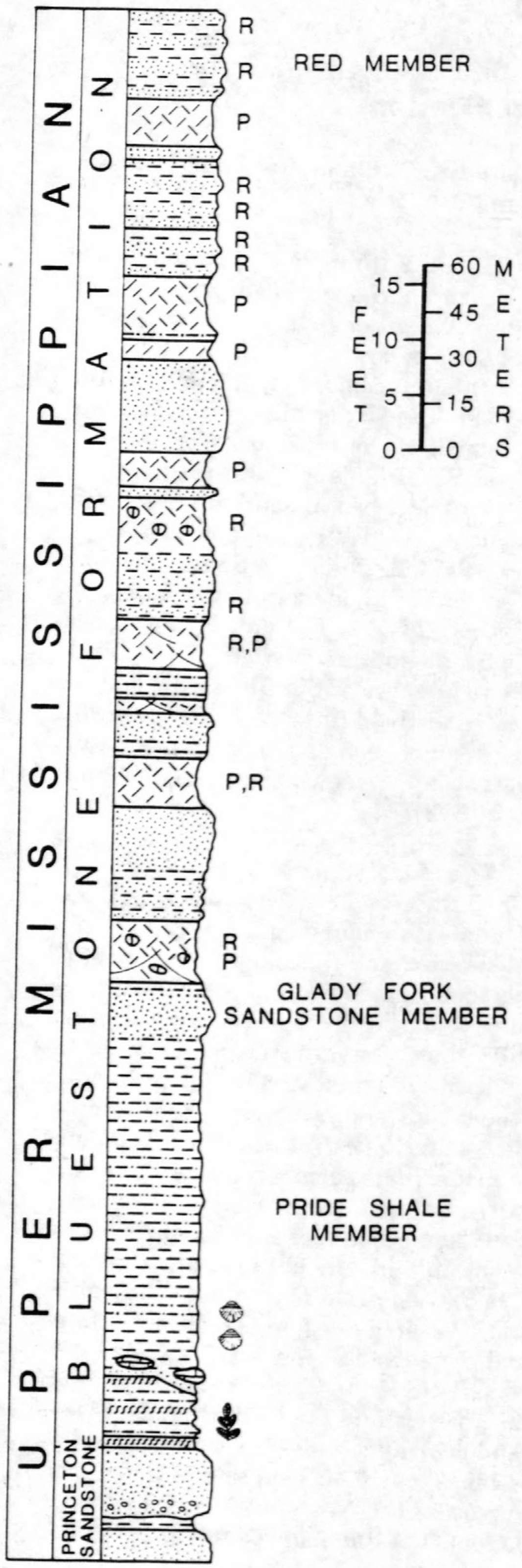


Figure 28 - Stratigraphic column of Upper Mississippian strata exposed at **STOPS 9** and **10**.

the base of the Pennsylvanian System. Coal laminae, limestone nodules, and a few centimeters of conglomeratic sandstone commonly occur near the contact with the overlying dark-gray shale and silty shale of the Pride Shale Member. The lower part of the dark-gray shale contains marine invertebrates and fresh-water to brackish vertebrates including a fossil fish, *Tanypterichthys pridensis* (-Weems and Windolph, 1986), collected at this locality from a large limestone concretion. This specimen has thick bony scales, is deep bodied, and has an unusually long pectoral fin.

In addition to flaser- and lenticular-bedding with bidirectional cross-bedding, the Pride Shale Member contains large-scale discordant channel form bedding features that have been interpreted as slump structures (Cooper, 1961). However, the eroded surfaces at the base of these channel form features appears to have resulted from submarine scour and channeling by strong bottom currents. Bedding within the discordant blocks appears to be the result of channel fill. Subaqueous channels, rhythmic-bedding, subtle bidirectional crossbedding, and marine fossils and burrows indicate that the Pride Shale was deposited in a shallow marine or estuarine environment.

The occurrence of coal and intertonguing of basal strata with the Princeton Sandstone and abundant terrestrial organic matter throughout the Pride Shale member, indicate that siliciclastic sediment, rich in organic matter, was being transported into the depocenter where base level was rising. According to the climate model (Cecil, 1990) this type of influx was the result of a climate that was wet enough for high terrestrial organic productivity, but not so wet that erosion and siliciclastic transport by fluvial systems was prevented.

Stop 10: West Virginia Turnpike Section

Stop Leaders: C. Blaine Cecil and Frank T. Dulong

STOP 10 includes the red member and Bramwell Member of the Bluestone Formation (figs. 28 and 29). The red member of the Bluestone Formation consists of grayish-red and greenish-gray sandstone, siltstone, and shale. These strata were deposited in a coastal plain environment which persisted from the deposition of the Pride Shale and Glady Fork Sandstone Members up to the marine Bramwell Member. The rocks are highly calcareous and calcareous nodules occur in rooted paleosols. Lenticular sandstone units as much as 4.6 m (15 ft) thick are suggestive of small fluvial systems which flowed across the area. Coal beds are thin, discontinuous, and impure although rooted underclays are rather common. A dry to seasonally dry paleoclimate is indicated by the presences of the calcareous paleosols (caliche) and the limited occurrence and poor quality of coal beds. The calcareous paleosols are of two distinct types. One type may be characterized best as paleo-Inceptisols or -Entisols that appear to have developed during autocyclic fluvial aggradation. These paleosols occur as poorly-bedded claystone, mudstone, and siltstone. They are generally red, calcareous, contain limited rhizomorphs, and are laterally discontinuous. These characteristics are consistent with relatively short duration of pedogenesis in an aggrading alluvial system. The second type of calcic paleosol is highly developed and is the primary emphasis of this stop. The paleosol at **STOP 10** has very well-developed calcic peds, vertic structures with clay skins, and a thick profile that grades downward into unaltered parent material. This latter type appears to be the result of allocyclic lowering of the water table and pedogenesis on a regional basis. Both types of paleosols at **STOP 10** appear to have developed under a somewhat drier and perhaps more seasonal climate regime than the climate during deposition of the Pride Shale Member. Thus, **STOP 9 and 10** illustrate how allocyclic processes punctuate stratigraphic sequences through regional changes in water table, base level, and sediment flux, and autocyclic processes control the facies mosaic during fluvial aggradation.

The Bramwell Member, which is 18 m (60 ft) thick at **STOP 10**, represents a marine transgression over the coastal-plain deposits of the red member. The Bramwell overlies about 0.8 m (2.5 ft) of carbonaceous shale. This shale contains fresh-water ostracodes at the base and a brackish-water fauna in the upper part indicating transgressive conditions (Englund and Henry, 1981). The ostracode-bearing shale is underlain by a well-developed paleosol of regional extent. In contrast to the calcic paleo-Vertisol above the Glady Fork Sandstone Member, the rooted underclay beneath the carbonaceous shale and Bramwell is highly leached, has horizontal rhizoliths, and has other characteristics that approach those of a Vertisol or Ultisol that formed under relatively wet climatic conditions.

The Bramwell Member is greenish-gray, calcareous siltstone and shale, which is typical of the member. Small-scale ripple- and flaser-bedding are the dominant bed forms. The unit is moderately bioturbated. Marine fossils are more scarce, and are of lower diversity than at other localities. This paucity of marine fossils, bioturbation, ripple- and flaser-bedding are all consistent with deposition in a tidally dominated system.

STOP 11: West Virginia Turnpike Section

Stop Leaders: C. Blaine Cecil, William Grady, and Kenneth J. Englund

Approximately 122 m (400 ft) of section exposed in roadcuts along I-77 between **STOP 10** and **STOP 11** include the lower part of the Pocahontas Formation up through the Pocahontas No. 3 coal bed (fig. 29). In contrast to **STOPS 1, 16, 17** and **18**, **STOPS 9-11** provide an excellent opportunity to study the significant paleobotanical, geo

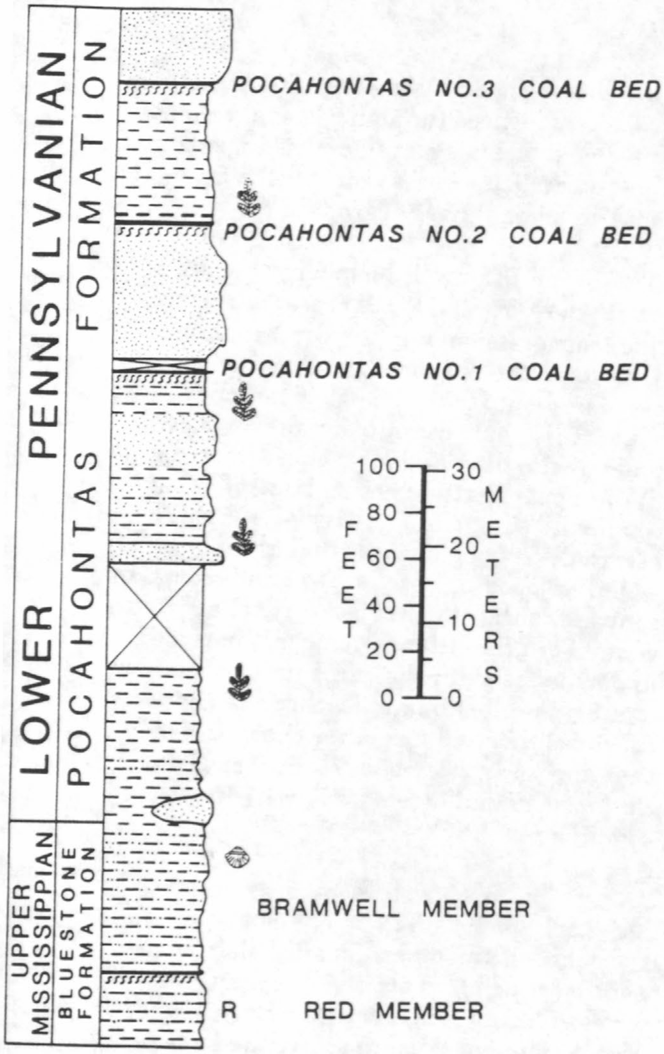


Figure 29 - Stratigraphic column of Upper Mississippian and Lower Pennsylvanian strata exposed at **STOPS 10** and **11**.

chemical, and sedimentological changes across the Mississippian-Pennsylvanian systemic boundary in an area of nearly continuous sedimentation. Many of these changes appear to be the result of a major change in paleoclimate. The Pocahontas Formation is interpreted as a sequence of coastal plain and near-shore marine sediments (Englund and Henry, 1981) that were deposited under a tropical everwet climate (Cecil et al., 1985). This everwet climate during the Early Pennsylvanian was in marked contrast to the dry or seasonally dry climate of the Late Mississippian. This climate change is indicated by the floral assemblages noted by White (1913), differences in the mineralogy of syngenetic and early diagenetic minerals, and the occurrence and quality of coal beds. In contrast to the Mississippian strata at this locality, coal deposits are extensive in the Pennsylvanian sequence but, calcareous rocks and grayish-red beds are not present. These lithologic changes are indicative of a major change in the sedimentary geochemistry of these coastal plain settings. This change in geochemistry was probably driven by a highly significant increase in rainfall with little seasonality, as evidenced by a major floral change, changes in paleosol types, and the geochemistry of syngenetic and early diagenetic minerals.

The sandstone which overlies the Bramwell Member probably represents a small meandering tidal creek deposit in a coastal setting. Tidal marsh and/or tidal flat deposits occur just north of the US Route 19 overpass. These tidally deposited sandstone and shale beds contrast with the deltaic sandstone exposed to the southwest in the vicinity of Pocahontas, Virginia. The tidal deposits at **STOP 11** consist of gray siltstone and shale that are bioturbated. Several thin, rooted underclay deposits, overlain by thin carbonaceous shale, are also present and are herein interpreted as "high marsh" deposits. Marsh sediments contain convolute bedding and, locally, *Calamites* in the growth position.

A quartzose sandstone with an erosional base overlies the tidal marsh complex. This sandstone (mile post 23) is approximately 9.7 m (32 ft) thick, fine- to medium-grained, and contains large-scale festoon crossbeds in the lower part. Most of the crossbeds dip west-northwest, but a few sets dip east-southeast. The cross-bedding is indicative of bidirectional currents in a tidally dominated system in which the ebb tidal currents were dominant flood tide currents. The festooning decreases upward, and the sandstone becomes planar- and ripple-bedded indicating a decrease in current velocity. This sandstone may have been deposited in an estuarine environment. The sandstone is overlain by a thin, rooted underclay and coal bed, which is indicative of sand deposition in a tidal system followed by marsh sedimentation and subaerial conditions. Approximately 4.6 m (15 ft) of gray siltstone and shale of probable tidal marsh origin overlie the thin coal bed. The upper part of the silty shale is rooted and also highly carbonaceous. The textural and mineralogical maturity of the sandstone may be related to mobilization of a quartzose residuum.

Approximately 6 m (20 ft) of light-gray sandstone overlies the tidal sand and tidal marsh complex. Within this sandstone, sedimentary structures are obscure because of its nearly uniform grain size and composition. Sedimentary structures recognizable in fresh samples or fresh outcrop include large-scale features that may be ball and pillow structures, convolute bedding, and large-scale accretionary beds. The accretion-bedding may be the result of sedimentation in a river mouth bar, estuarine delta, or perhaps a tidal delta. Although the sedimentary structures cannot be unequivocally interpreted, this sandstone may have been deposited in an estuarine environment which had a strong tidal influence. None of the bed forms are definitive of any particular depositional system; however, they are consistent with deposition in a tidally dominated estuarine environment. Typical fluvial features are not evident. This sandstone may have served as a platform for the initiation of peat formation, and the subsequent development of the peat forming environment (Englund et al., 1984). The lower Pennsylvanian interval from the top of the Bramwell to the base of the Pocahontas No.2 coal bed represents aggradation by autocyclic processes whereas the genesis of the Pocahontas No. 2 and No. 3 coal beds were controlled by allocyclic changes in climate, and perhaps base level.

The Pocahontas No.2 coal bed is the stratigraphically lowest coal commercially mined in West Virginia, and where mined it is moderate in ash-yield and sulfur content (fig. 10, table 2). At this location the lower one foot of the Pocahontas No. 2 coal bed is moderate in ash and high in sulfur content while the upper two-thirds of the bed is generally low ash and low sulfur (fig. 30). The basal foot of the coal represents the initial stage of swamp development as a planar swamp with probable standing water cover. Mineral and sulfur emplacement was caused by the influx of sediments and dissolved solids in moderate pH surface and ground water. Well-preserved >50 micron vitrinite components are common. The moderate inertinite content suggests intermittent subaerial exposure and oxidation of the peat surface. Above the initial swamp stage to the top of the coal inertinite is extremely abundant and vitrinite is low in abundance, but the proportion of well-preserved vitrinite macerals remain high. In this part of the coal sulfur content is low and ash yield is low, except in a shale parting layer. The high inertinite content, good preservation of some of the vitrinite precursors and low ash-yield and sulfur content indicate that the upper two-thirds of the coal probably represent a peat that accumulated on a slightly elevated, acidic swamp. The elevated swamp surface, which probably was kept wet much of the time (standing water cover ?), inhibited mineral influx and promoted preservation of large pre-vitrinite plant debris. These factors suggest a paleoclimate wet enough to allow domed or slightly elevated swamps to develop. The high inertinite content, however, suggests that this climate may have been long wet - short dry seasonal (i.e. a wet-dry seasonal paleoclimate with annual rainfall adequate to support domed swamp development).

The Pocahontas No. 2 coal bed is separated from the Pocahontas No. 3 by approximately 15 m (50 ft) of silty shale. The Pocahontas No. 3 is a low-volatile bituminous coal which generally contains less than 1 percent total sulfur and yields less than 6 percent ash. At **STOP 11**, the Pocahontas No. 3 consists of thin beds of coal intercalated with carbonaceous shale. This locality is at the edge of the Pocahontas No. 3 paleoswamp, but coal of minable thickness and quality occurs less than 1.5 km (1 mi) to the southwest. The paleoswamp developed under

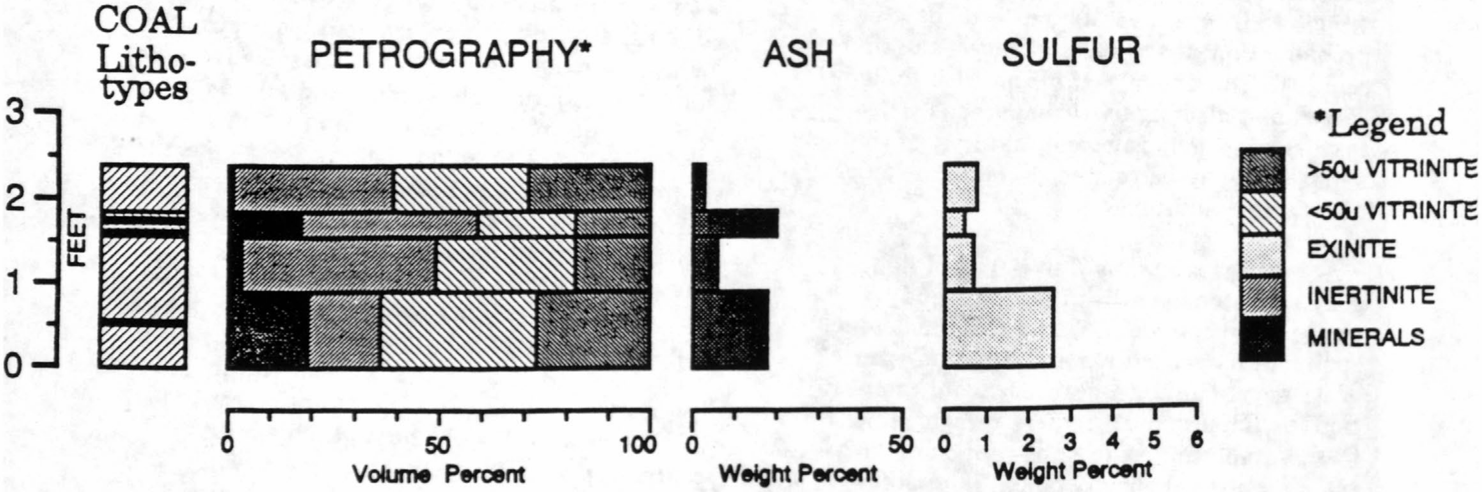


Figure 30 - Distribution of macerals, ash yield and sulfur content in the Pocahontas No. 2 coal bed at **STOP 11**.

a tropical everwet climate; it was ombrogenous, domed, and highly acidic (Cecil et al., 1985). These physical and geochemical conditions of peat formation were conducive to the formation of a low-ash and low-sulfur coal bed (Cecil et al., 1985). The thickest part of the coal bed developed over platforms of sand in delta lobes (Englund, 1974).

STOP 12 - Mahan Exit, Lower Member, Kanawha Formation

Stop Leaders: Bascombe M. Blake, Jr.and Ronald L. Martino

The lower member of the Kanawha Formation and the upper portion of the Lower Pennsylvanian New River Formation are exposed in road cuts along the West Virginia Turnpike (I-77 & I-64) and in natural outcrops along Paint Creek in western Fayette County and southern Kanawha County, West Virginia. A complete section of the lower member can be obtained by combining a section measured along Lost Branch, 700 m (2300 ft) south of the Mahan Exit, with the large road cuts at the Mahan Exit (fig. 31).

The lower member of the Kanawha Formation is 137 m (450 ft) thick in this area of West Virginia, having thinned northward from 335 m (1100 ft) at Douglas Station (location 1, fig. 4). Marine-influenced zones are common in the sections, including the widespread Betsie Shale Member (**STOP 1** and **STOP 7**), and the Dorothy Shale of Krebs

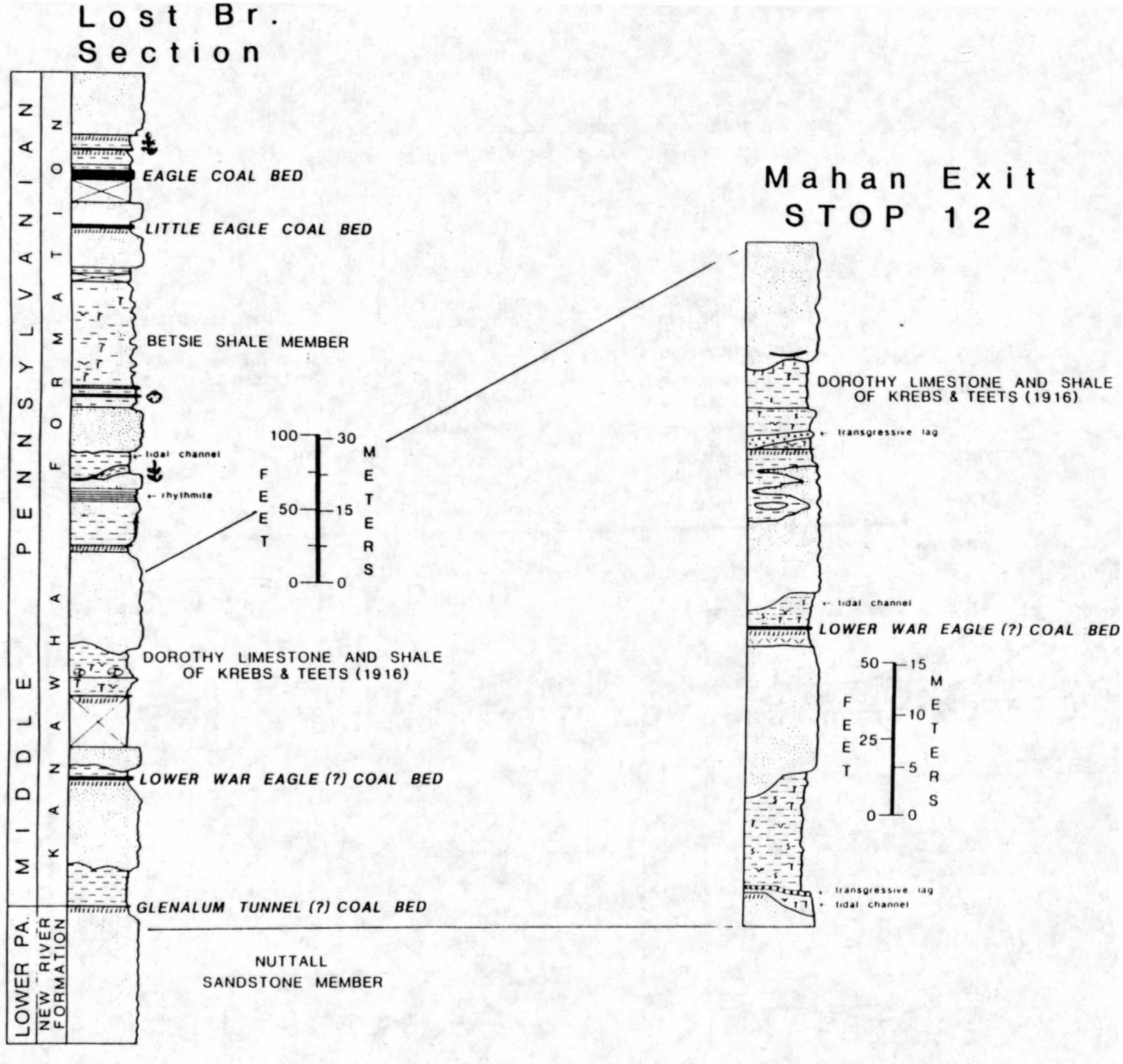


Figure 31) Stratigraphic section at **STOP 12** compared with a nearby measured section, the Lost Branch section

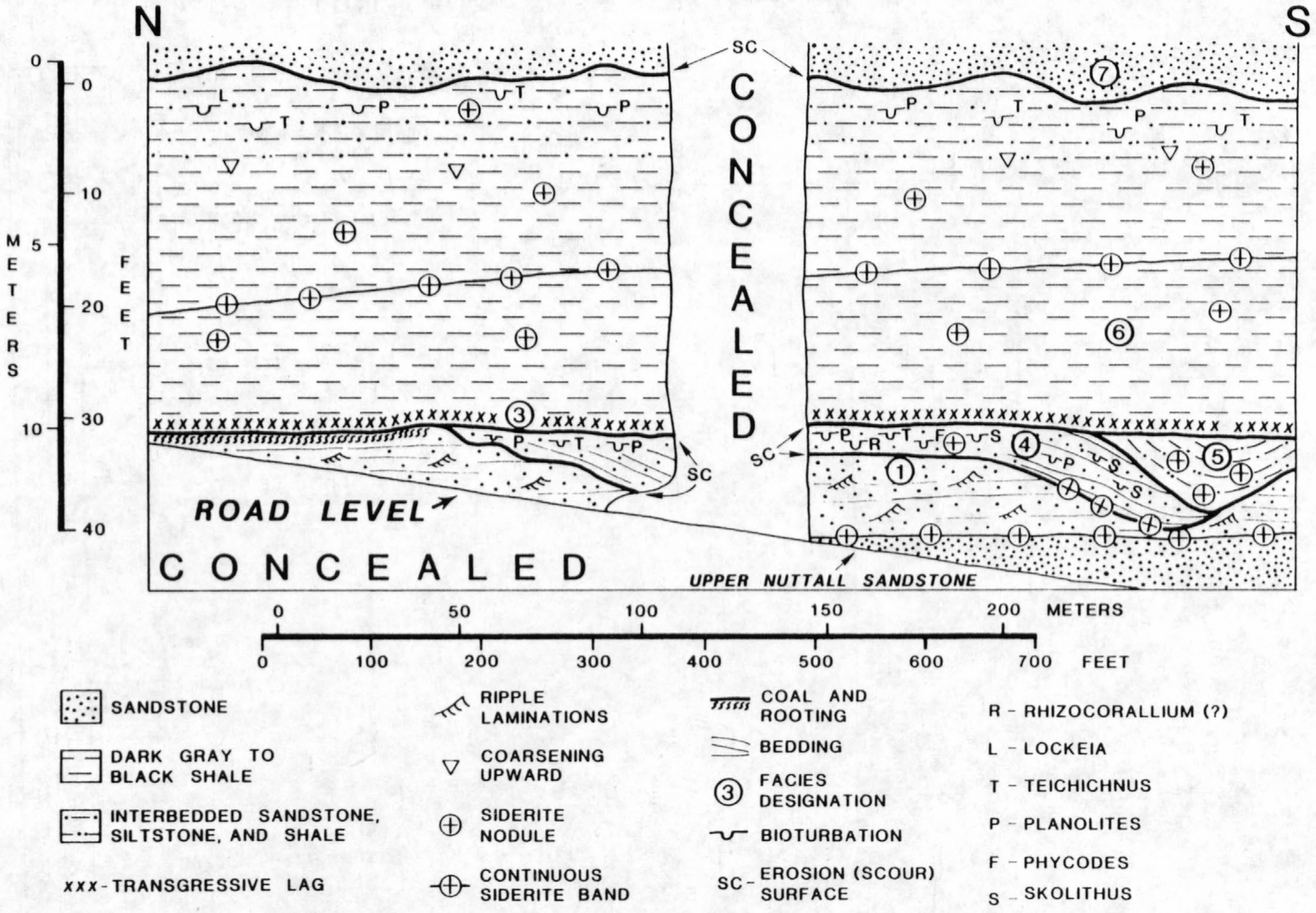


Figure 32) Diagrammatic representation of STOP 12. Refer to the text for explanations of outcrop details.

and Teets (1916). Greb and Chesnut (1992) and Eble and others (1991) discuss transgressive sediment fills in various small channels from stratigraphically equivalent strata assigned to the lower part of the Breathitt Formation in eastern Kentucky. Three channels with marginal marine fills have been identified in the Mahan and Lost Branch sections. Other tidally-influenced to marginal marine indicators found in these two sections include (1) rhythmic bedding, (2) extensively burrowed, scour-bottomed, transgressive lag sands, (3) black, siderite nodule- and limestone concretion-bearing shales, (4) trace fossil assemblages, and (5) invertebrate body fossils.

Sedimentary Facies at Stop 12

At **STOP 12**, we will examine coastal plain, tidal channel, and shallow marginal marine facies of the lower member of the Kanawha Formation and uppermost New River Formation. A burrowed sandstone-filled channel, eroded into a thinly interbedded to interlaminated sandstone, is well-exposed at road level along the exit ramp (Figure mxsection 32). The channel facies also erodes a thin, impure coal bed, the Glenalum Tunnel (Lower Douglas of Hennen and Teets, 1919), arbitrarily designated as the contact between the Kanawha and New River formations in the proposed Pennsylvanian System stratotype (Arndt, 1979).

Facies 1, immediately overlying the Upper Nuttall Sandstone, is a thin, parallel-laminated, very fine- to fine-grained, light gray sandstone. Ripple cross-laminations are locally common. The cyclic (rhythmic?) thickening and thinning of the horizontal, parallel bedding is suggestive of tidal control during deposition. Siderite nodules are present, but uncommon, and the basal 5-10 cm (2-4 in) is heavily sideritized into a continuous band, but the original bedding is widely observable. Bioturbation is rare and limited to small, horizontal, sand-filled tubes. The rarity of bioturbation is suggestive of lowered or variable salinity or perhaps a rapid rate of sedimentation. This facies was deposited under relatively moderate to low energy conditions and may have accumulated in an upper estuarine setting under the combined influence of riverine and tidal currents. The continuous sideritized zone at the base of the sandstone possibly represents the interface between fresh and brackish groundwater (Greb and Chesnut, 1992).

Facies 2, present only at the north end of the outcrop, is a thin, impure coal bed 0-10 cm (0-4 in) thick. Peat accumulation occurred during a temporary cessation in clastic sedimentation possibly caused by estuarine abandonment or a change in base level. Facies 2 is truncated along the road cut and exact relationships between the coal bed and overlying facies are unclear.

A tidal channel complex with two distinct fill types postdates facies 1 and 2. The channel complex thins and pinches out toward the north end of the road cut. Small scale scours are present along the less channel-form parts of the complex and a well-formed channel up to 2.4 m (8 ft) deep is developed near the southern end of the road cut. The first channel-fill facies is a very fine- to fine-grained sandstone (facies 3). Compound cross bedding with small-scale cross-laminations in the bed sets are the main sedimentary structures. Both paleoflow and accretion were to the east. Bioturbation is common, but bedding is well-preserved. *Skolithos* is very common and other ichnogenera present include *Planolites*, *Phycodes*, *Rhizocorallium*(?), and *Teichichnus*. Abundant mica flakes and highly macerated plant debris cover foresets, demonstrating periodic slack water conditions. Small, discontinuous siderite bands are present. Tidal channel facies described from elsewhere in the Kanawha Formation also contain a similar, limited diversity trace fossil assemblage including *Skolithos*, *Teichichnus*, *Phycodes*, and *Planolites* (Hobday and Horne, 1977). This assemblage contains burrows of mud-loving deposit-feeders and the dwellings of sand-loving polychaetes (*Skolithos*)(Martino, 1989, 1993) and is a variation of the high-stress *Phycodes-Zoophycos* assemblage (Martino, 1989).

The second tidal channel complex fill, facies 4, is a black shale with sandy and silty interbeds and siderite nodules. Channel migration and concomitant erosion continued, truncating portions of facies 3. Changes in sediment supply or channel deactivation resulted in the deposition of fine sediments. Bioturbation is lacking and bedding parallels the channel margins. Greb and Chesnut (1992) discuss tidal channels in equivalent strata from the Breathitt Formation in eastern Kentucky and Hobday and Horne (1977) discuss facies similar to facies 1, 3, and 4 from the lower Kanawha Formation in southern West Virginia.

Facies 5, a transgressive lag, is a very fine- to fine-grained, locally calcareous sandstone. Small shale clasts and coal chips are present representing mud and peat rip-ups respectively. Mica flakes are common. Irregular-shaped siderite nodules suggestive of burrows are abundant and bioturbation has obliterated bedding. One small bivalve was found. The sharp basal contact marks a ravinement surface separating coastal plain facies from overlying shallow marine facies (Swift, 1968; Liu, 1990; Martino, 1991a, 1991b). The presence of an erosion surface is indicated by the progressive truncation of the subjacent coal bed and tidal channel facies and the presence of coal spars. Liu and Gastaldo (1989) and Liu (1990) referred to similar deposits in the Black Warrior Basin as ravinement beds. The transgressive lag grades upward into the overlying black shale.

Facies 6 is a thick, black shale with commonly occurring siderite nodules and elongate masses. Bioturbation is not evident except near the top where the unit coarsens. The top 2 to 3 m (6 to 9 ft) contains thin-bedded to interlaminated, rippled, very-fine-grained sandstone with siltstone and silty shale laminations. *Teichichnus* and *Planolites* become common with the increase in sand content and *Lockeia* is locally present. Mica flakes and macerated plant debris are common on bedding planes. High organic content of the mud deposited in the lower portion of the facies, suggestive of limited oxygen and a soft substrate, would have provided less than optimum conditions for benthos. Increasing biologic activity evidenced by burrows toward the top of this facies may indicate improved conditions due to a firmer substrate and/or shallower water and higher energy conditions leading to better oxygenated water. Alternatively, the interbedding of sand and mud may have increased the preservation potential of burrows.

Facies 7 is represented by a multi-storied fluvial-deltaic channel sandstone that marks a return to active non-marine deposition. Cross-bedding indicates flow toward the north-northwest. The return to fluvial-deltaic conditions are possibly the result of autocyclic processes, such as channel switching or delta-lobe switching, or allocyclic factors, such as a drop in sea level due to eustasy, or climatic changes. Cecil (1990) has discussed climatic controls on sediment flux into the Appalachian foreland during the Pennsylvanian.

Summary of Depositional Events

Marginal marine deposition began with the allocyclic drowning of a broad paleovalley represented by the Upper Nuttall Sandstone (Rice, 1984; Rice and Schweitering, 1988; Chesnut, 1988). This drowning resulted in the development of an estuarine deposystem. Tidally-influenced deposition in areas of the system that may have been distal to open marine waters resulted in the deposition of facies 1. The lack of biologic activity and the continuous siderite band at the base of facies 1 suggest very marginal marine conditions. Aggradation of estuarine sand flats led to the development of a coastal swamp, represented by the Glenalum Tunnel coal bed. Tidal channels incised into and locally through estuarine sand flats and peat deposits. Coastal ravinement produced a highly bioturbated transgressive lag on top of the coastal plain and marginal marine facies. The exact relationship between the tidal channels and the peat deposits is unclear due to erosion associated with coastal ravinement. Marine influence is suggested within the tidal channels as indicated by the types of trace fossils as well as by facies relations. Transgression of the coastal marshes and estuarine channel fills is marked by the presence of a ravinement surface. The transgressive lag overlying the ravinement surface may represent the remains of the shoreface. Shallow marine conditions prevailed during the deposition of facies 6. Salinity may have varied during deposition of this facies based on the presence of discontinuous siderite bands. Upward coarsening in facies 6 indicates that water depths shallowed and energy conditions increased in association with regression. A return to a fluvial-deltaic deposystem is marked by the multistoried channel sandstone in erosional contact with subjacent shallow marine sediments.

STOP 13 - Toms Fork Haul Road, Kanawha Formation

Stop Leaders: Bill Grady, Mitch Blake and Cortland Eble

Strata assigned to the upper part of the middle member and the entire upper member of the Kanawha Formation and the lower part of the Allegheny Formation are exposed in continuous cuts along a haul road accessing Ark Land Company's Cabin Creek Property (fig. 33), located on the north side of Toms Fork of Cabin Creek near Leewood, West Virginia. **STOP 13** is in the Cabin Creek watershed, an early and important mining district in West Virginia. Eleven (11) Kanawha Formation coal beds have been mined on the waters of Cabin Creek (ascending): Eagle, Eagle "A", Powellton, Lower Campbell Creek, No. 2 Gas, Peerless, Cedar Grove, Fire Clay, Winifrede, Coalburg, and Stockton coal beds. The upper 125 m (412 ft) of the middle member of the Kanawha Formation are exposed in the Toms Fork section and the upper member is 105 m (346 ft) thick (see location 3, fig. 4). The exposed part of the Kanawha Formation at this locality is the same as exposed in the Birch River-Powell Mountain section (fig. 26), however the section along Toms Fork is thicker, having expanded from 152 m (500 ft) thick on Powell Mountain (**STOP 8**) to 230 m (758 ft) along the Toms Fork haul road (fig. 4). Total thickness for the Kanawha Formation has expanded from 213 m (700 ft) thick in the vicinity of Birch River to over 427 m (1400 ft) in the area of **STOPS 12** and **13**. Southward thickening of the Kanawha Formation from the area of **STOP 8** occurs through expansion of the upper and middle member and the addition of section onto the base of the lower member (fig. 4).

The Allegheny Formation caps the ridges in the area of **STOP 13** and four (4) coal beds are currently being surface mined north of the haul road. While not occurring in the immediate area of this stop, Conemaugh Group strata have been noted above the conglomeratic "Mahonning Sandstone" (as used by Krebs and Teets, 1914) several kilometers to the northwest. The Mahonning Sandstone is found capping the high knobs and is the stratigraphically-lowest fluvial-deltaic sandstone containing abundant pebbles above the Lower-Middle Pennsylvanian Series boundary, making this sandstone useful in locating the top of the Allegheny Formation. The Allegheny Formation is approximately 91 m (300 ft) thick in this area of Kanawha County using the top of the Mahonning sandstone as the upper formation boundary (see Blake, 1992). The Allegheny coal beds mined in the area are the Little No. 5 Block, the (Lower) No. 5 Block, the Upper No. 5 Block, and the No. 6 Block, correlated with the Lower Kittanning coal bed of northern West Virginia by Kosanke (1984, 1988).

STOP 13 will begin near the top of the haul road at the level of Stockton coal bed and the overlying Kanawha Black Flint of White (1891), defined as uppermost unit of the Kanawha Formation (Arndt, 1979). During the walk down the haul road, discussions will be held at various horizons, including the Kanawha Black Flint of White (1891), the Stockton coal bed, the Little Coalburg coal bed, the Winifrede Shale Member, the Fire Clay coal zone, the Dingess Shale Member, and the No. 2 Gas coal bed. Ample time will be afforded for examination of interesting units and sample collection.

Three marine zones, the Dingess

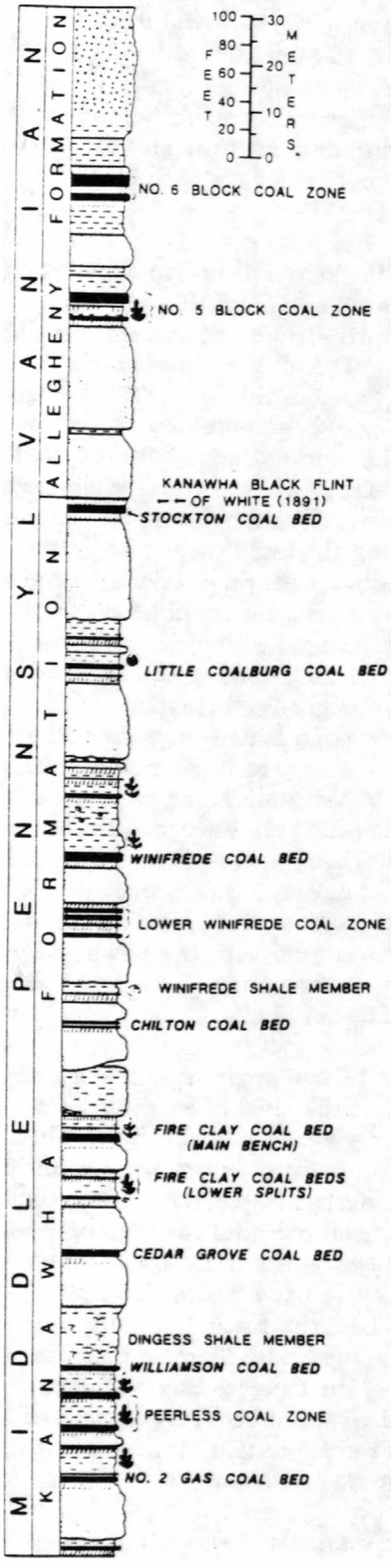


Figure 33 - Stratigraphic column of Middle Pennsylvanian strata (Kanawha and Allegheny Fms) exposed at **STOP 13**.

Shale Member, the Winifrede Shale Member, and the Kanawha Black Flint of White (1891) and canneloid black shale containing a fresh water fauna above the Little Coalburg coal bed are exposed along the haul road (fig. 33). This canneloid shale with its fresh water fauna is a locally important key bed useful in separating the Stockton and Coalburg coal zones which locally coalesced, forming a coal bed >3.4 m (11 ft) thick named the "Stockburg" by mining companies. The Coalburg coal bed is absent along the haul road, having been removed by erosion associated with deposition of the overlying fluvial channel-fill sandstone.

The Kanawha Black Flint of White (1891) is the uppermost unit assigned to the Kanawha Formation (Arndt, 1979). Three facies have been identified from the Kanawha black flint, a chert facies, a silty facies, and an orbiculoid shale facies. Cavaroc and Ferm (1968), Reppert (1979), and Watson (1992) discuss facies distribution, depositional setting, sedimentology, and areal distribution of the black flint. Blake (1992) briefly discusses the black flint and revises the known areal distribution of the chert and non-chert facies (Figure mzone 34). The silty facies of the Kanawha Black Flint of White (1891) is present along the haul road, identified by its distinctive weathering pattern. While the chert facies is not present in the area of **STOP 13**, excellent examples can be found along the haul road, having been dredged from the Kanawha River as part of gravel deposits used in road construction. Abundant rooting has obliterated any primary sedimentary structures, but poorly-preserved invertebrates have been identified (T.W. Henry, U.S.G.S., Denver, personal communication, 1988). Reconnaissance mapping in the area indicates that the Toms Fork section is located near the paleoshoreline of the Kanawha black flint (Figure 34).

The Winifrede Shale Member (Mag

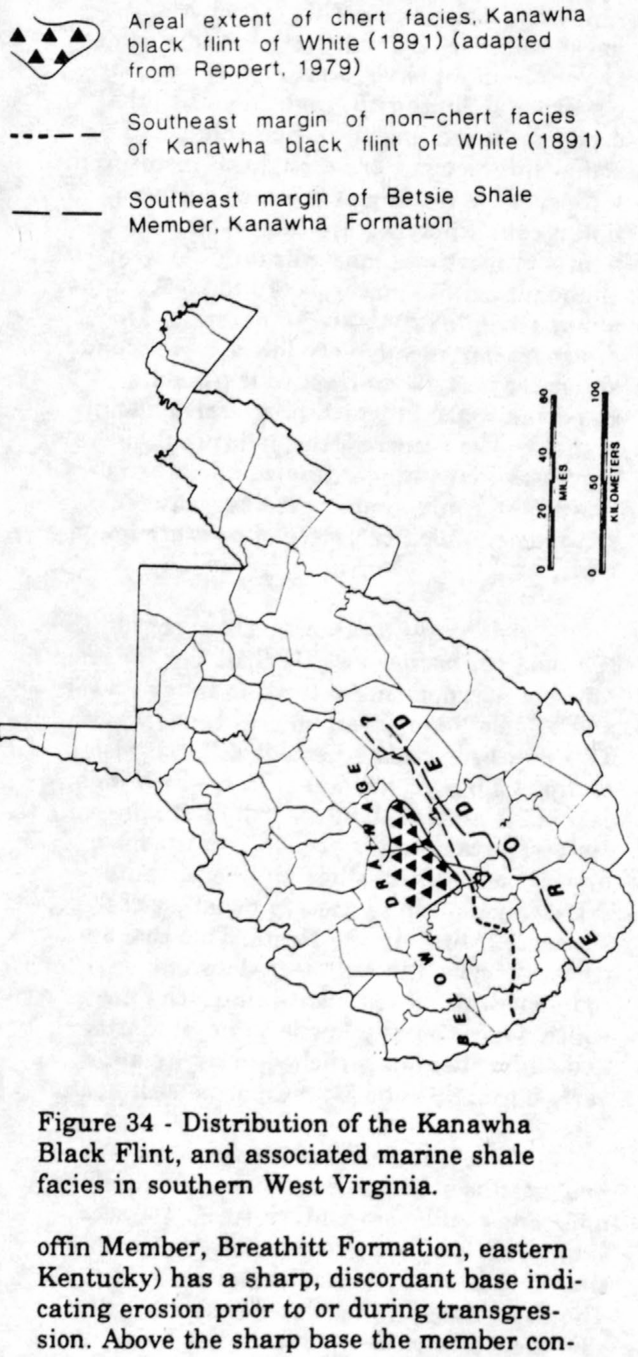


Figure 34 - Distribution of the Kanawha Black Flint, and associated marine shale facies in southern West Virginia.

offin Member, Breathitt Formation, eastern Kentucky) has a sharp, discordant base indicating erosion prior to or during transgression. Above the sharp base the member consists of approximately 0.3 m (1 ft) of a very fine- to fine-grained sandstone. The sandstone is calcareous, argillaceous, and pyritic and fines upward into a dark gray shale. A well-preserved, generally articulated brachiopod dominated fauna is present. The marine fauna becomes sparse and more disarticulated in the upper, shale portion of the member. A 9.8 m (32 ft) thick, multistoried, fluvial channel-fill sandstone uncomformably overlies the Winifrede. The top of the Winifrede was eroded after deposition, as suggested by the overlying sandstone. While the fauna is dominated by brachiopods, other invertebrates reported from the Winifrede Shale include pelecopods, crinoids, gastropods, cephlapods, trilobites and corals (Hennen and Teets, 1919, p. 279-293; Henry and Gordon, 1979; Chesnut, 1991). This fauna has been assigned an earliest Atokan age in the proposed Pennsylvanian System stratotype (Henry and Gordon, 1979).

The Dingess Shale Member (Kendrick Member, Breathitt Formation, eastern Kentucky) is well-developed along the haul road, being approximately 13.4 m (44 ft) thick. A transgressive lag occurs at the base of this member, marking the presence of a ravinement surface (see Blake, 1992; Martino, 1989, 1992, 1993). This 10.2 cm (4 in) thick, very fine- to fine-grained, calcareous, bioturbated sandstone grades upward into a more heterolithic sequence. Plant debris is common on bedding planes and trace fossils and burrows occur, including *Planolites* and *Teichichnus*. Cone-in-cone limestone concretions are found associated with a 76.2 cm (30 in) thick very fine-grained, calcareous sandstone 5.2 m (17 ft) from the base of the member. The top 2.1 m (7 ft) consists of an interbedded very fine-grained sandstone, siltstone and shale sequence with possibly rhythmic bedding. Bedding thickness is on a centimeter scale. This part of the member may represent deposition in a mouth bar. The Dingess is unconformably overlain and the top replaced by a 10.4 m (34 ft) thick channel-fill sandstone of probable fluvial origin.

Another important key bed found in the Toms Branch section is a flint clay parting associated with the Fire Clay coal bed (zone). The parting is present in the lowest bench of the Fire Clay coal zone, occurring in a thin bony parting, making the parting thin and difficult to recognize. This parting is of volcanic origin based on characteristic minerals such as sanidine, beta-form quartz, euhedral zircons, rutile, and apatite (see Blake, 1992 for discussion). The parting has been traced across parts of three states, having been reporting as occurring in a large area stretching northeast from Tennessee, across eastern Kentucky and southern West Virginia into central Nicholas County, where it has been found in a sporadically occurring coal bed (unpublished data).

Coal beds of the Kanawha formation exhibit a stratigraphic change in their megascopic appearance and petrographic and palynologic profiles, a change that is believed to be the result of a shift in paleoclimate through time. For instance, beds at the base of the hill (middle Kanawha Formation) are primarily homogeneous, bright-banded coals that are rich in vitrinite, whereas stratigraphically younger coals at the top of the hill (upper Kanawha and basal Allegheny Formations) tend to be much more dull and inertinite-rich. Often these upper beds display a cyclic lithotype succession from bright banded coal at the base of the bed, to semisplint and splint coal in the middle, and back to bright banded lithotypes at the top of the bed.

The bright-banded coals of the middle Kanawha Formation, well-illustrated by the No.2 Gas coal at this stop, are coals that contain high percentages of vitrinite macerals, especially well- preserved >50 micron types (fig. 35), and are also very low in ash and sulfur. These peats probably accumulated in an everwet paleoclimate with an annual rainfall amount ideal for domed peat development and peat preservation. This allowed the peat swamps to maintain a standing water table perched above the adjacent steams levels. Collectively, this would provide an acidic water cover to promote peat preservation, and inhibit oxidation of the surficial peat, and also restrict mineral emplacement.

Upper Kanawha Formation and lower Allegheny Formation coal beds, illustrated at this stop by the Little Coalburg coal bed, are typically much duller in appearance than underlying Kanawha coal beds, and often display a systematic cyclicity in lithotype succession, ash yield, sulfur content, petrography and palynology. This cyclicity is believed to reflect a buildup of peat to a domed swamp from a planar precursor-swamp. Cycles from vitrinite-rich bright banded lithotype benches to inertinite-rich splint and semisplint benches are thought to result from a progressive doming of the peat surface. Splint coal lithotypes are composed predominantly of inertinite macerals (>50 %) with abundant exinite macerals (10 to 20 %), and minor vitrinite (20 to 30 %) macerals. They are characteristically very low ash yield and sulfur content. In contrast to the middle Kanawha coals, in which progressive doming probably also occurred, the splint coal layers represent much higher levels of peat oxidation to the point where peat destruction essentially equalled or exceeded peat accumulation.

The thin Stockton A coal exposed in the road cut between **STOPS 13A & B** exemplifies a paradox encountered in many Lower and Middle Pennsylvanian coal beds (fig. 36). This coal bed yields 9 percent ash, but also contains 5.5 percent total sulfur, which indicates that almost all of the ash (and mineral matter) in this coal is probably in the form of pyrite. Coals low in silicate minerals but extremely high in pyrite are usually closely-associated with marine strata. This coal apparently formed in a protected swamp environment with no sediment influx. The peat, which was originally low in mineral matter and sulfur, became enriched in pyrite after early burial by sulfate-rich marine sediments.

The Stockton coal exposed on the haul road is split by numerous shale partings and bone coal layers and, as such, is a poor example of a splint-rich Stockton coal bed that is extensively mined in the area (fig. 37). The associated Coalburg coal, another splint coal that is extensively mined, has been cut out by sandstones along this exposure. A Stockton - Coalburg coal core, taken from nearby, and the Little Coalburg coal, exposed further down the hill, will be used to illustrate the lithotype characteristic of these coals (figs. 38 and 39).

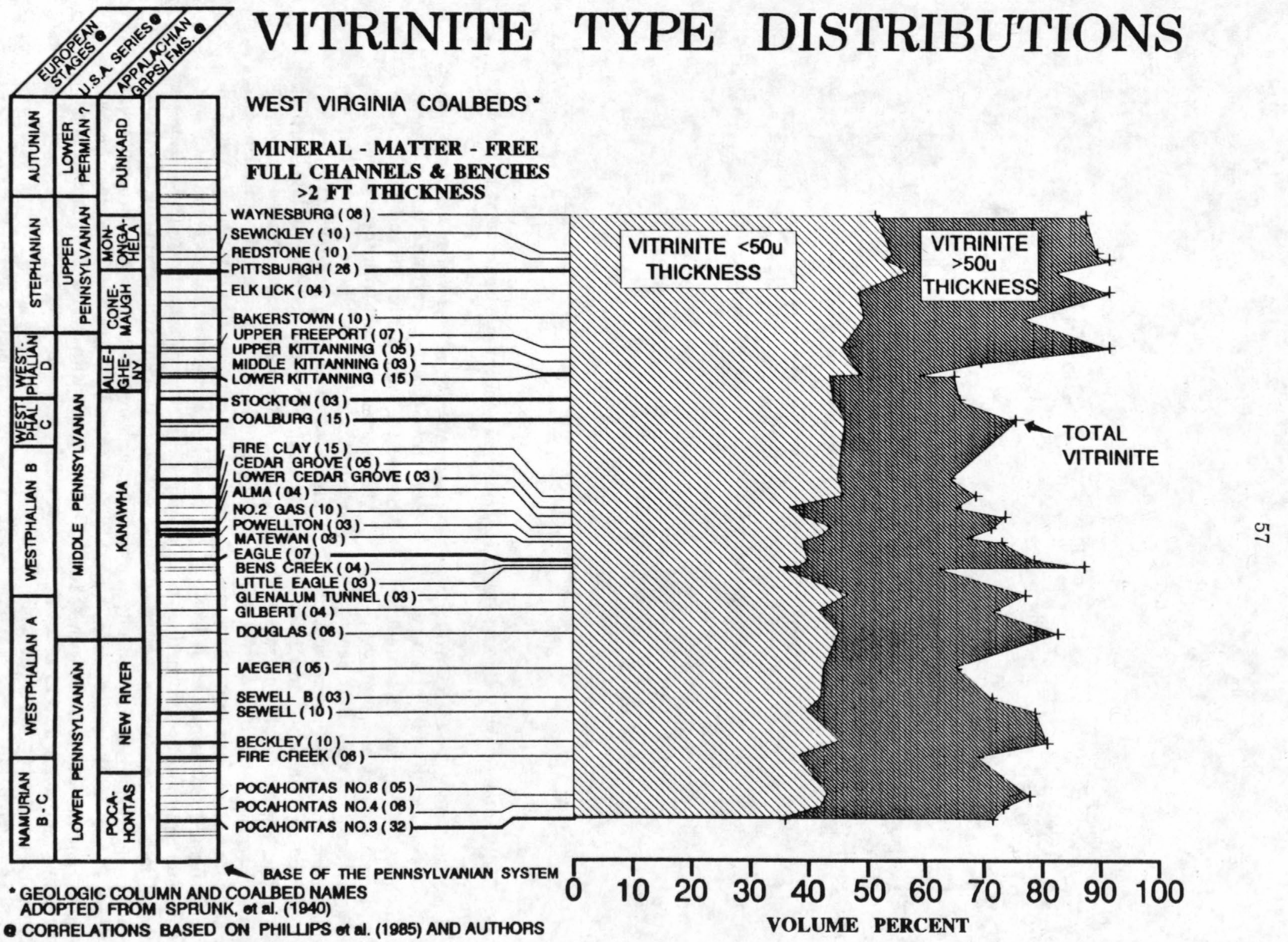


Figure 35) Distribution of total vitrinite, vitrinite in <50 micron bands, and vitrinite in >50 micron bands (and as a matrix) in Pennsylvanian coal beds in West Virginia (N=265)

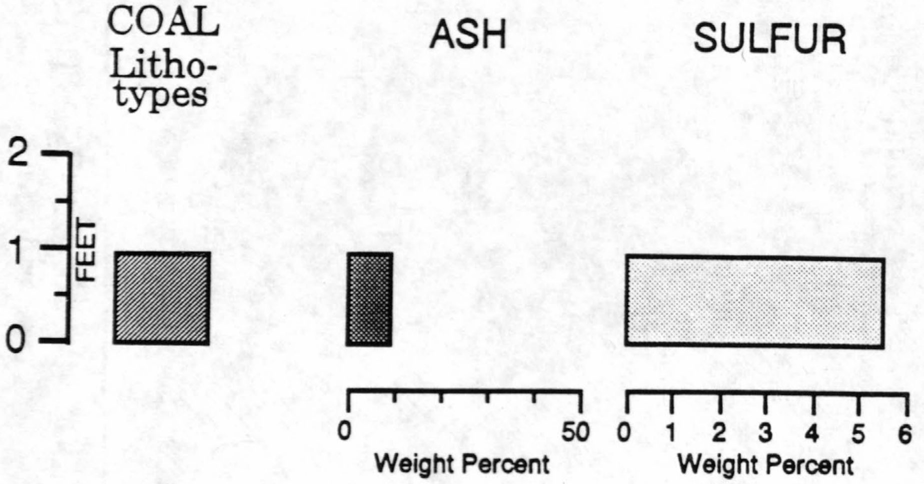


Figure 36 - Distribution of ash yield and sulfur content in the Stockton A coal bed at **STOP 13**.

The core contained 8.5 feet of nearly continuous coal with a thin shale parting 6 feet from the base. The parting separated the lower 6 foot thick Coalburg coal from the upper 2.5 feet of Stockton coal. Both coals contained frequent durain and splint coal lithotypes alternating with bright and dull clarain lithotypes. The lithotypes and ash profiles of the Stockton in the core correlate well with the Stockton coalbed exposed on the haul road (fig. 39). Sulfur content is low through both coal beds but the ash yield varied substantially. Petrographic analyses show vertical trends, especially in inertinite content. Three benches of the coal contain high inertinite contents and coincide with the durain and splint coal layers at 0.15 to 0.3 m (0.5 to 1.0 ft), 1.4 to 1.5 m (4.5 to 5.0 ft) and 2.3 to 2.4 m (7.5 to 8.0 ft). Inertinite content systematically increases then decreases below and above the high inertinite layers (e.g. between 1.1 and 1.8 m, 3.5 and 6.0 ft). Ash yield varies inversely with the inertinite content and is generally very low within the inertinite and exinite-rich layers, whereas high ash benches coincide with low inertinite content. Floral successions, based on palynologic analyses, show vertical oscillations in the plant communities through the two coal beds. Floral communities dominated by tree ferns coincide with inertinite and exinite-rich durain and splint coal layers. Inertinite-poor, high ash layers of these coals are generally dominated by a lycopsids, especially herbaceous forms with moderate calamites and cordaites abundances.

The Coalburg coal in this core con

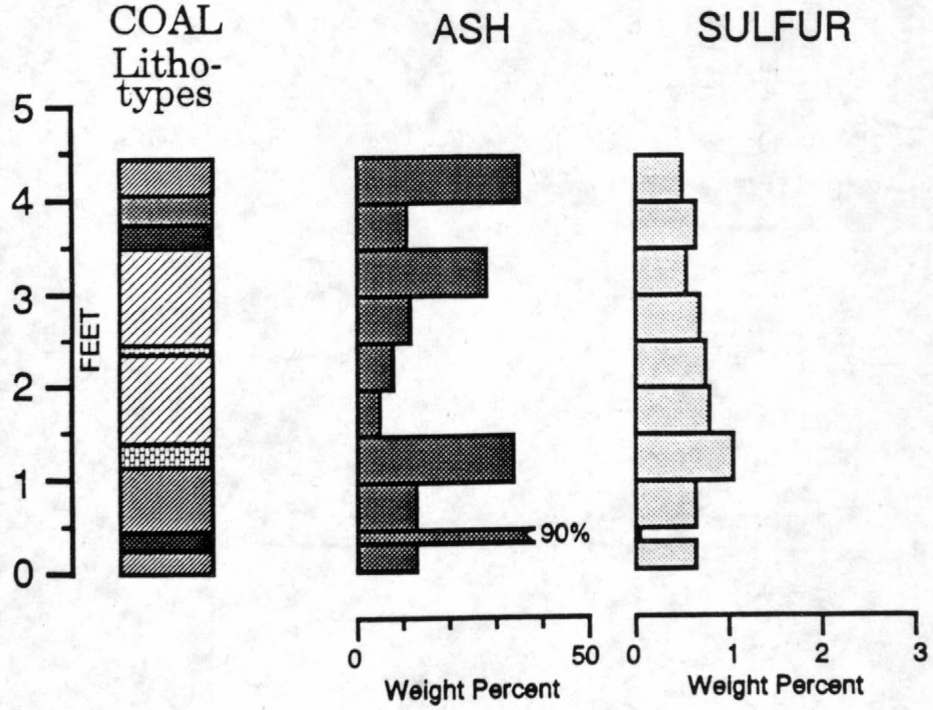


Figure 37 - Distribution of ash yield and sulfur content in the Stockton coal bed at **STOP 13**.

tains two cycles of vertically successive increasing, then decreasing inertinite and exinite content, and tree fern abundance that coincide with trends of decreased ash yield. The Stockton coal shows one cycle. These cycles are interpreted to represent progressive development from planar swamp morphologies to domed swamps, followed by a return to planar swamp settings, generally with associated sediment deposition. As such, the Stockton - Coalburg coal in this core is actually comprised of several peat-forming events, even though we refer to it as one "bed". This scenario is very common in upper Kanawha coal beds.

Durite microlithotype (>95 volume percent inertinite + exinite) abundances are less in the upper Kanawha compared to the middle Kanawha Formation (fig. 40). Durites in upper Kanawha and lower Allegheny coal beds show an increased segregation of durain and splint coal layers (relative to associated bright lithotypes) in these coal beds. The opposite is generally true for middle and lower Kanawha Formation and Lower Pennsylvanian coal beds. Distinct benches of splint coal durains are interpreted to represent extended periods of peat oxidation in a domed swamp setting, whereas the more disseminated durite microlithotype bands within an overall vitrinite-rich matrix probably represent less-frequent (but equally severe ?) oxidation of surficial domed peat.

The Little Coalburg coal displays some of the characteristics of typical upper Kanawha Formation coal beds (fig. 41). A

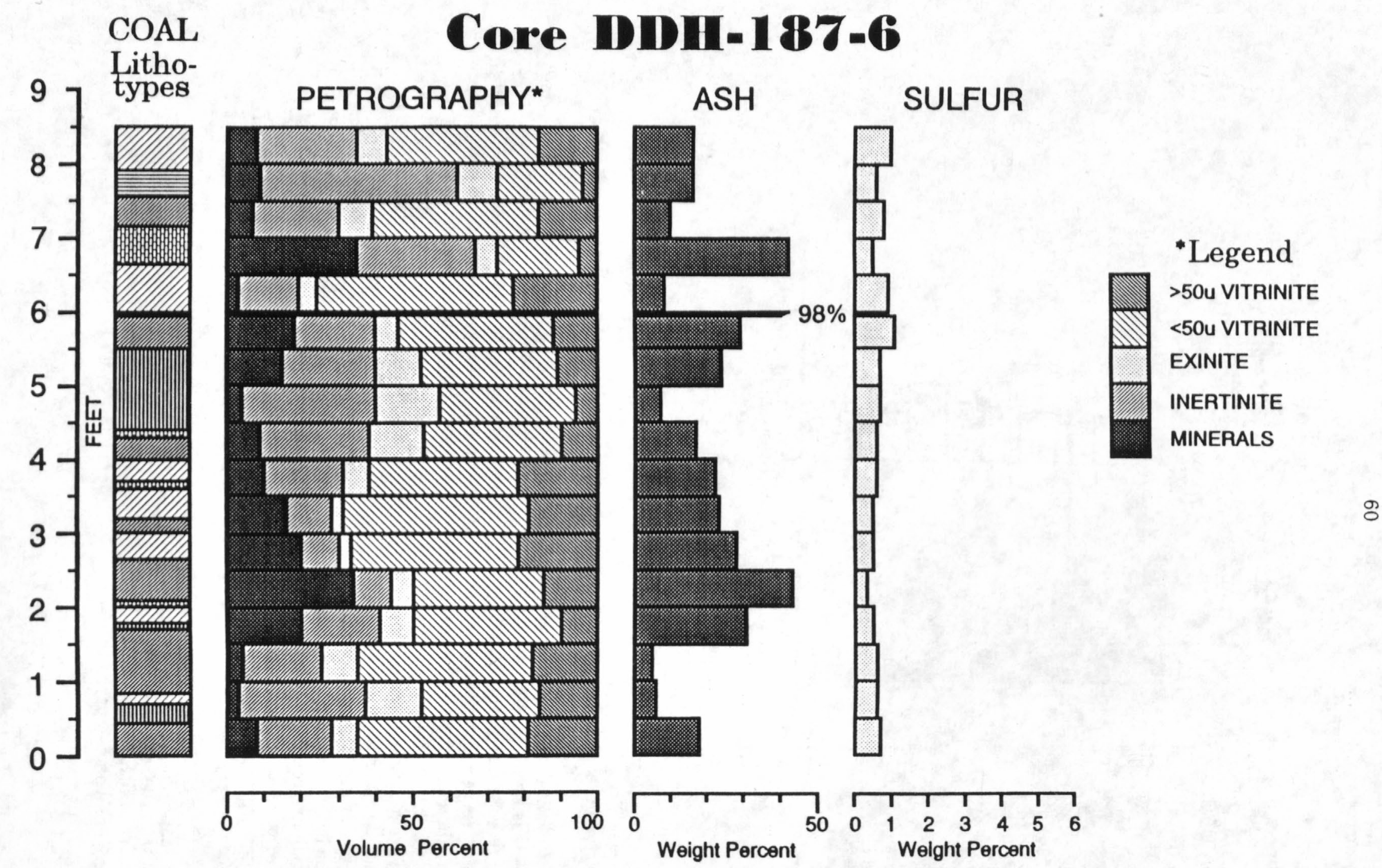


Figure 38) Distribution of macerals, ash yield and sulfur content in a core sample of Stockton/Coalburg coal taken adjacent to the haul road (**STOP 13**). Refer to the text for discussion.

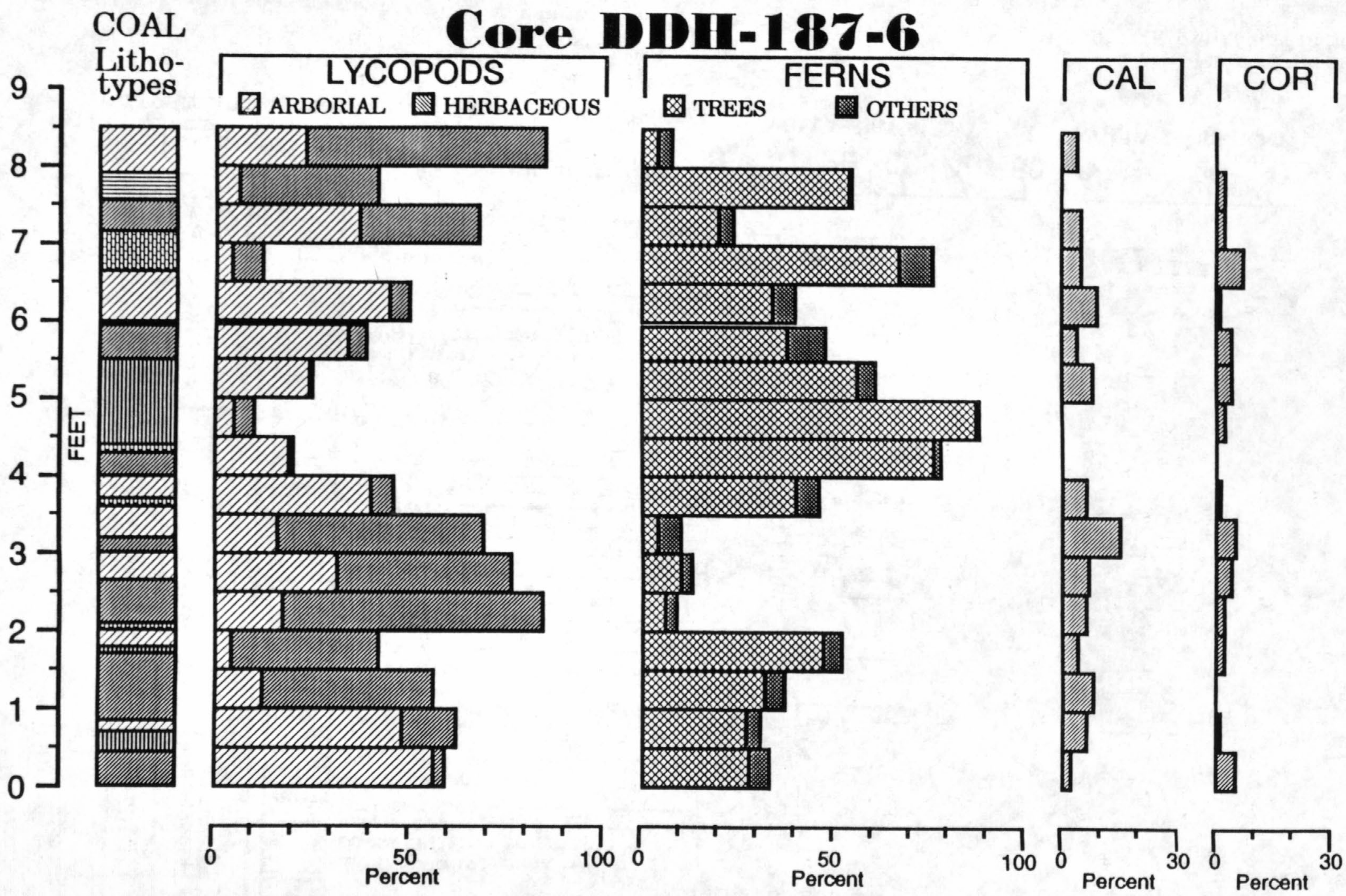


Figure 39) Distribution of miospores in the Stockton/Coalburg coal core at **STOP 13**.

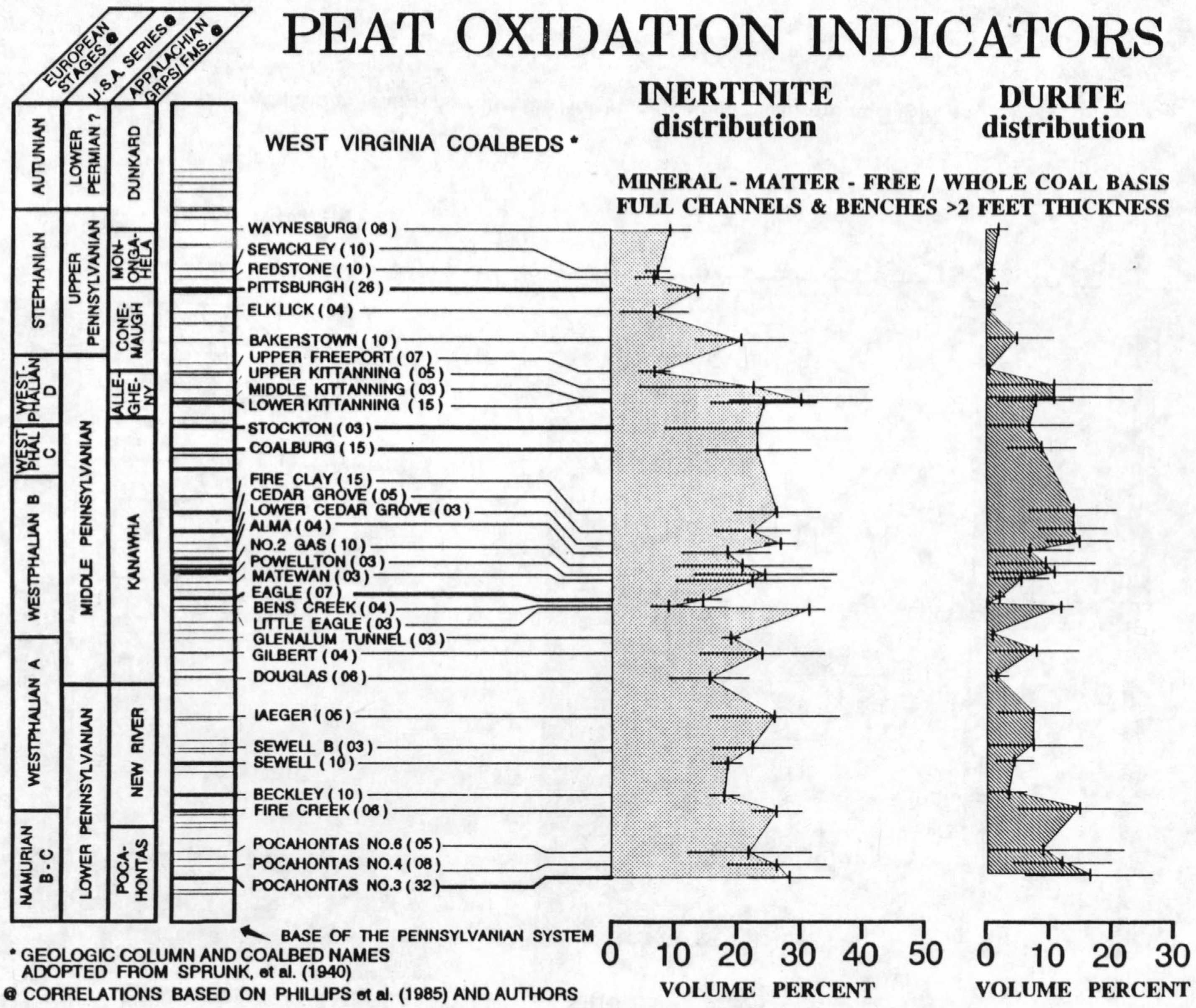


Figure 40) Distribution of inertinite content and durite microlithotype (>50 micron bands composed of >95 % inertinite and liptinite) in Pennsylvanian coal beds in West Virginia (N=265).

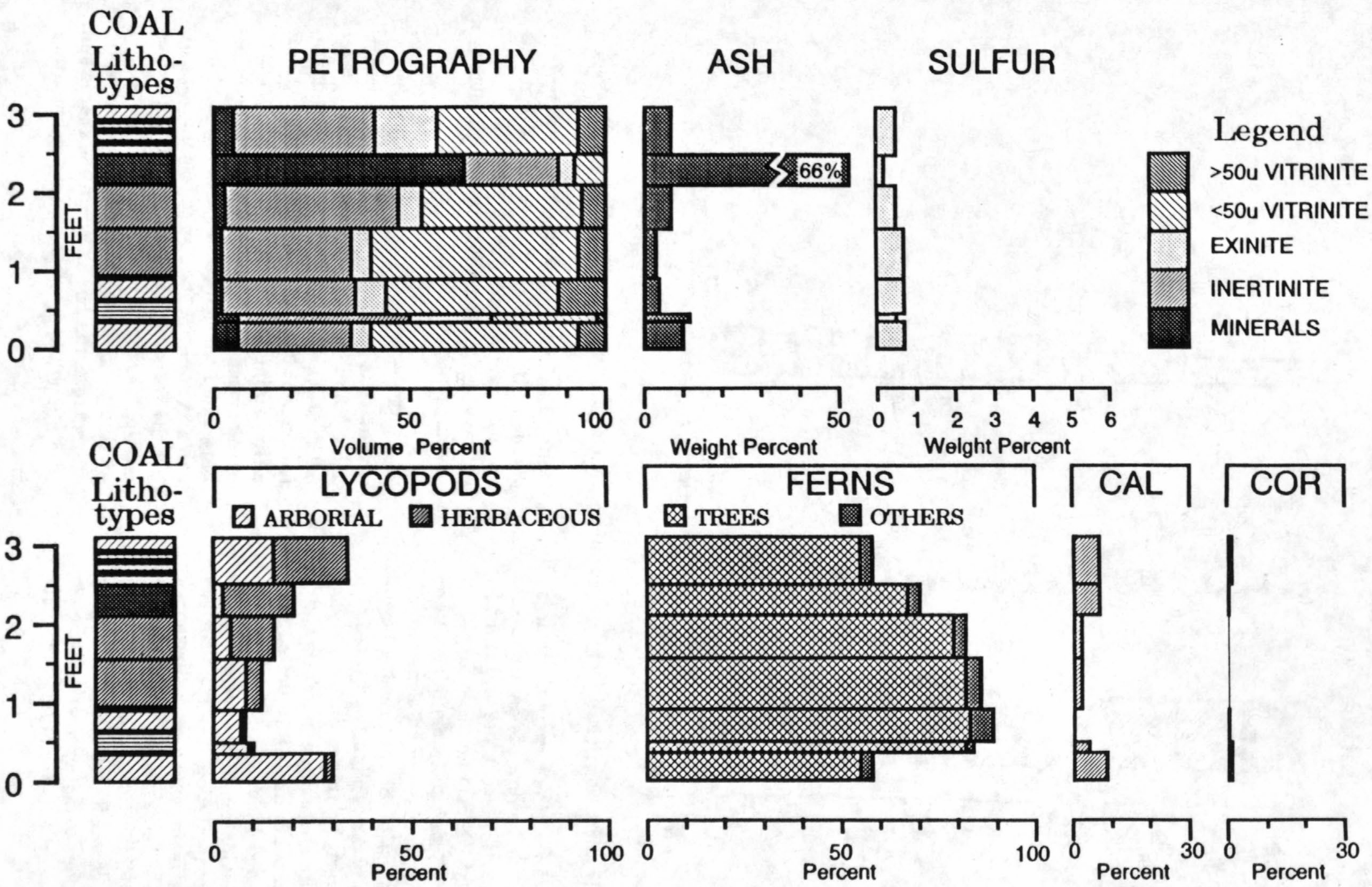


Figure 41) Distribution of macerals, miospores, ash yield and sulfur content in the Little Coalburg coal bed at **STOP 13**.

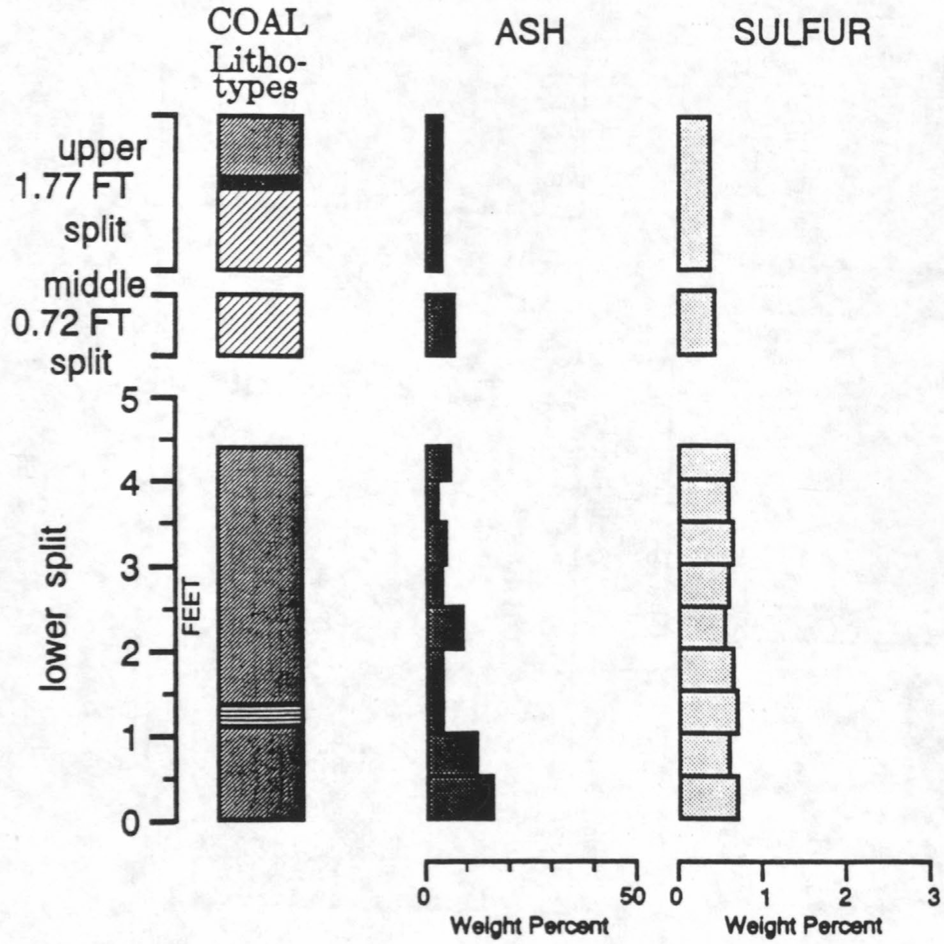


Figure 42 - Distribution of ash yield and sulfur content in the Winifrede coal bed at STOP 13.

thin bright clarain at the base of the coal is low in ash yield and sulfur, moderate in vitrinite and inertinite contents and is moderate in arboreal lycopsid and tree fern abundances. High tree fern spore abundances are typical of the upper Kanawha Formation. These factors suggest that the swamp began as a planar, to perhaps slightly elevated swamp. The overlying thin splint coal layer, rich in inertinite and exinite macerals and tree ferns palynomorphs, is interpreted to represent a period of domed peat accumulation, in which oxidation may have exceeded plant debris accumulation. The splint layer is overlain by a clarain, which probably represents a period of decreased oxidation of the peat. Above this, inertinite content increases upward in the low ash coal layers indicating a return to higher oxidation levels in the swamp. This interval is terminated by sediment influx, as represented by the shale parting. After this clastic event the peat swamp environment was re-established but sediment influx remained continued, perhaps intermittently.

The Winifrede coal bed along the haul road is split into three benches, the thickest being 1.3 m (4.4 ft), and except for the basal 0.3 m (1 ft) is very low in ash yield and sulfur content (fig. 42). The lower split contains a thin splint coal layer, but overall this coal is comprised of bright- and dull-banded clarains. The homogeneous appearance and low ash and sulfur contents suggest that throughout the three benches of this coal the paleoenvironment was a slightly domed, acidic swamp that promoted good peat preservation. In contrast, the Lower Winifrede coal is split into numerous thin benches by sedimentary

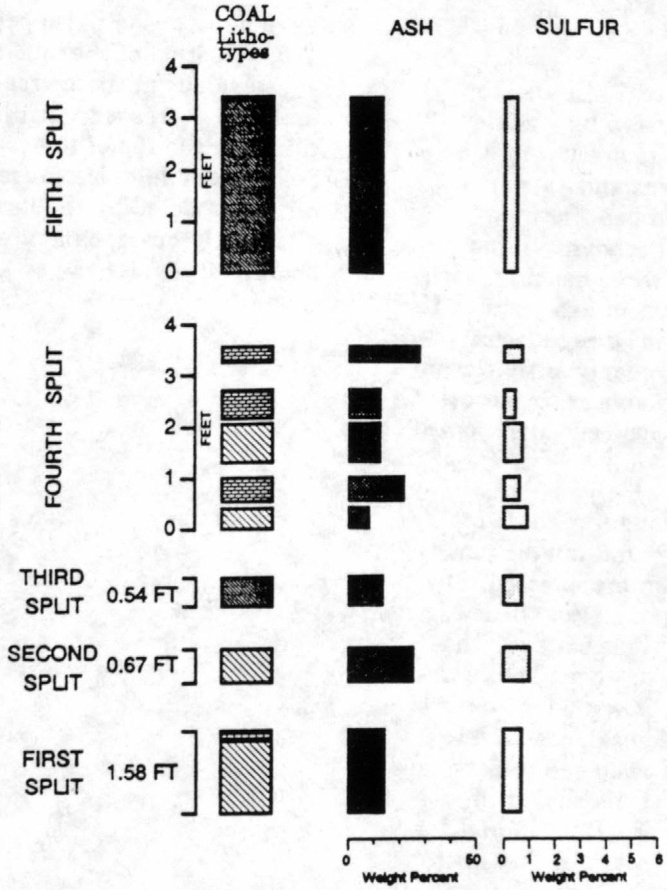


Figure 43 - Distribution of ash yield and sulfur content in the Lower Winifrede coal bed at **STOP 13**.

strata (fig. 43). Each bench is low in sulfur content and moderate to high in ash yield, indicating that peat accumulation here took place in a dominantly planar swamp with that was repeatedly inundated with sediment and dissolved solids.

The Chilton coal is split into two benches by a thick shale parting (fig. 44). The thin lower bench contains a shale parting and is high in ash, but low in sulfur content and probably represents a peat accumulated in a planar swamp frequently invaded by sediment-laden surface waters. The upper bench is a cannel-like bone coal, low in sulfur content, but very high in ash yield. This bench probably also represents peat formation in a planar swamp setting.

The Fire Clay coal consists of a 1.5 m (4.8 ft) main split and three separate lower splits (fig. 45). The three lower splits are low to high in ash and sulfur and represent peat accumulation in poorly-developed planar swamp settings, separated by sedimentary layers. The lower split contains the widespread Fire Clay volcanic tonstein. The base of the main bench is high in ash yield, extremely high in sulfur content, and is immediately overlain by a shale parting. The remainder of the main bench is split into four benches by shale partings. The coals between the partings are bright-banded clarains with consistently low ash and sulfur contents. The main bench peat appears to have accumulated on a series of well-protected swamps, each forming on, and truncated by, layers of sediment.

The Cedar Grove coal bed also consists of two benches separated by a shale parting (fig. 46). Both splits are low in ash yield and low sulfur content and appear to represent slightly elevated peat-forming swamps. Likewise the underlying Williamson coal bed is also split into three benches, each of which is low to moderate in ash yield and low in sulfur content (fig. 47). Because of their low ash yield and predominantly clarain lithotypes, all three splits appear to represent peat accumulation on slightly elevated domed swamps.

At the base of the haul road, two coals that typify the lower and middle parts of the Kanawha Formation are exposed. The No.2 Gas (fig. 48) and Peerless coal beds (fig. 49) consist primarily of bright banded lithotypes. The No.2 Gas coal has been extensively mined in the area and is a good example of a uniformly bright-banded coal, common to this part of the section. Lower and middle Kanawha Formation coal beds are low in ash yield and sulfur content, and contain high percentages of vitrinite, especially >50 micron vitrinite. The palynofloras of these beds are usually dominated by *Lycospora*. Vertically, the No.2 Gas coal exhibits characteristics suggestive of peat formation in a wet paleoclimate. The high ash yield, high sulfur basal increment represents initial peat accumulation in a planar swamp with significant mineral and dissolved solids influx by surface and groundwaters. Inertinite abundance is low and vitrinite content is high in this increment which suggest high moisture conditions (probable standing water cover). Vitrinite content, especially > 50 micron vitrinite, increases upward to the middle of the bed, as ash yield and sulfur content decreases, and inertinite content increases. These trends demonstrate the development of a slightly elevated swamp surface, probably covered by standing, acidic water much of the time, a situation that was ideal for peat preservation and accumulation. This part of the swamp was dominated by arboreal lycopsids, which had developed morphological and reproductive adaptations for growth in very wet areas. As doming of the paleo-peat swamp continued, oxidation of the surficial peat increased, as is evident in the increased inertinite content and decreased vitrinite. This trend continues to the top of the coal bed. The increased doming also caused a change in paleoflora, as noted by the increased percentages of small lycopsid, calamite and cordaite miospores.

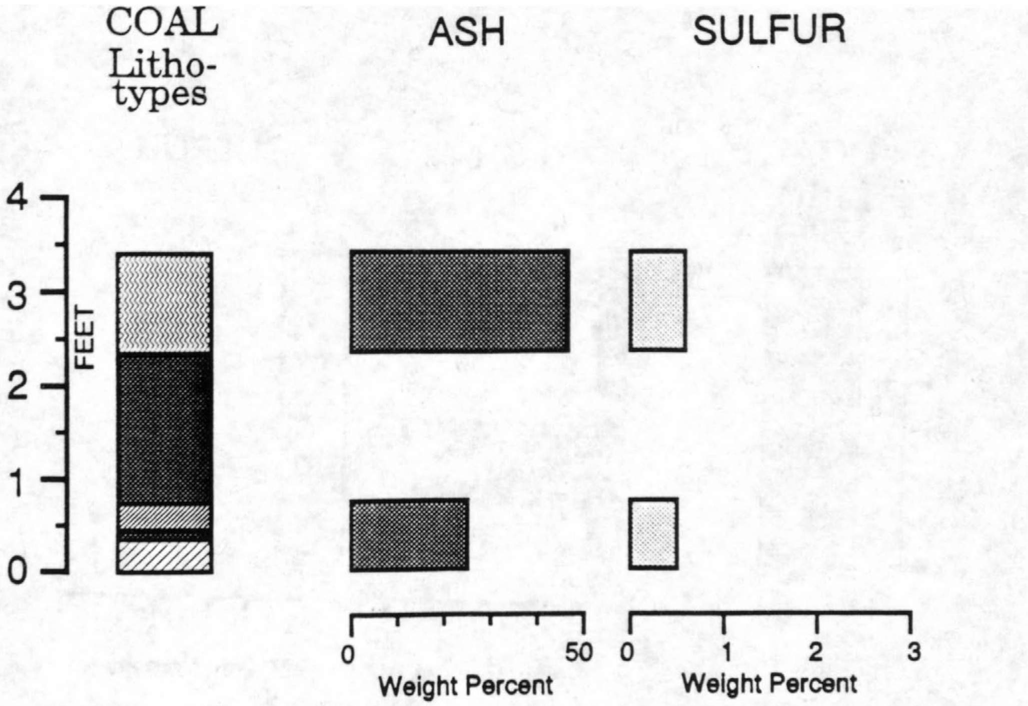


Figure 44 - Distribution of ash yield and sulfur content in the Chilton coal bed at **STOP 13**.

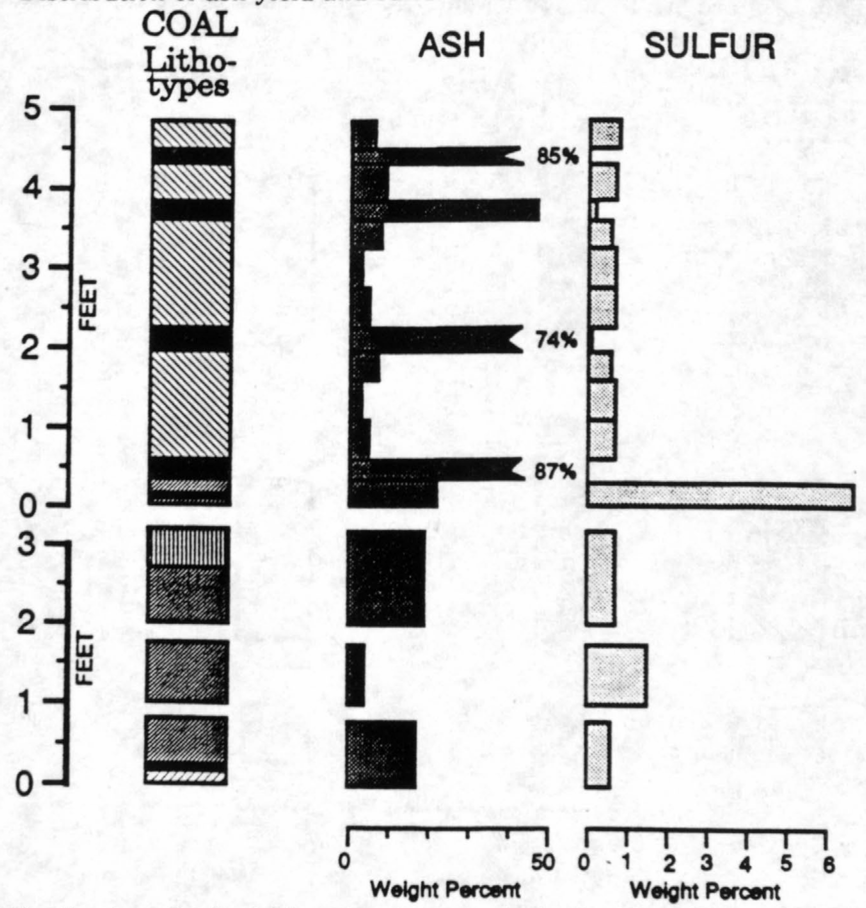


Figure 45 - Distribution of ash yield and sulfur content in the Fire Clay coal bed at **STOP 13**.

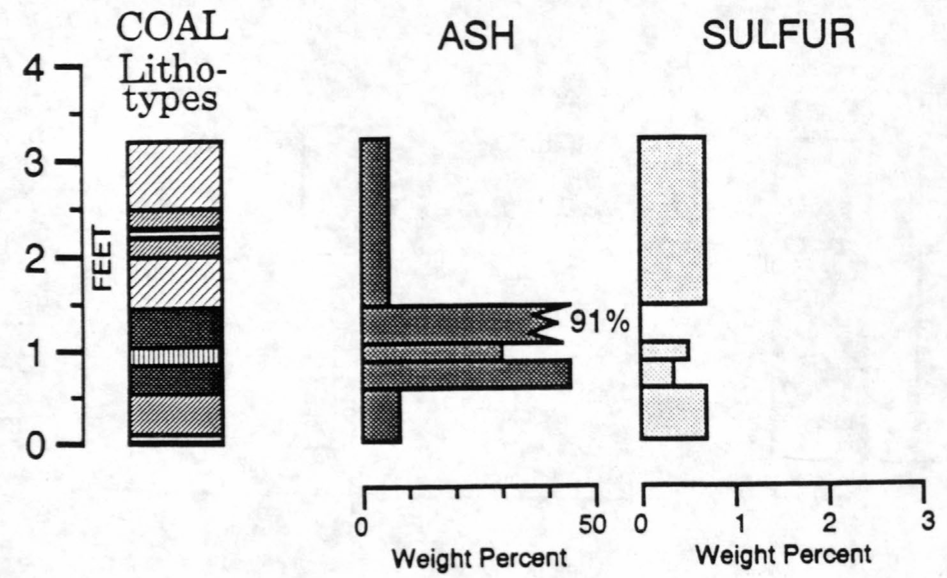


Figure 46 - Distribution of ash yield and sulfur content in the Cedar Grove coal bed at **STOP 13**.

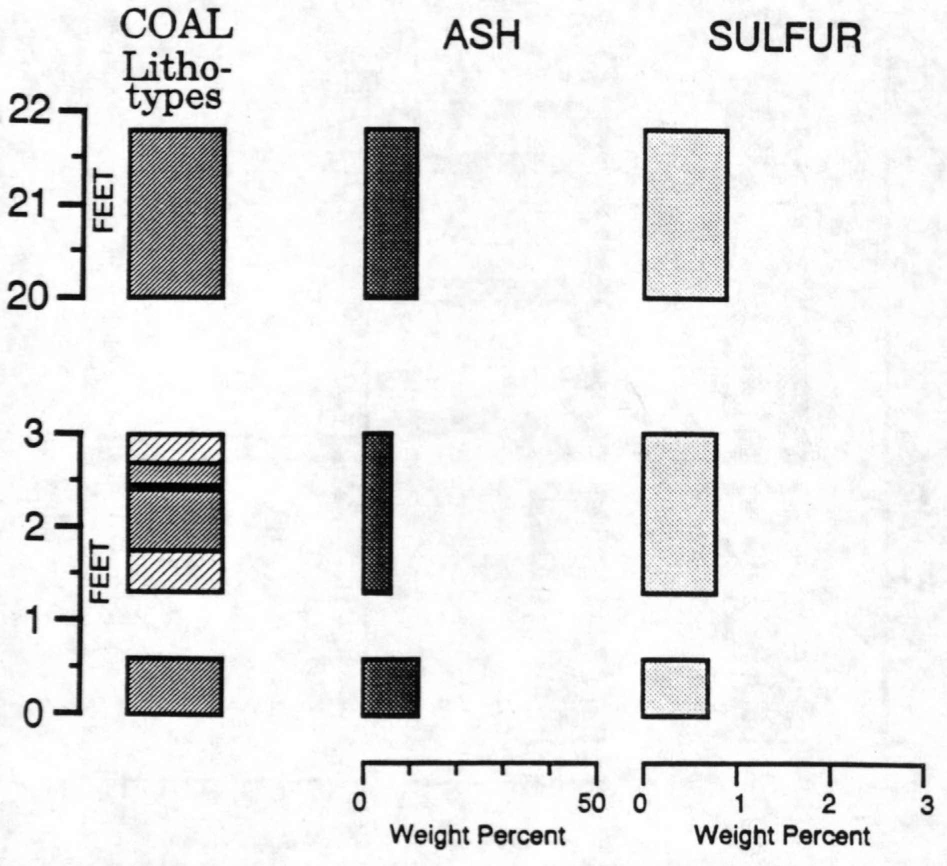


Figure 47 - Distribution of ash yield and sulfur content in the Williamson coal bed at **STOP 13**.

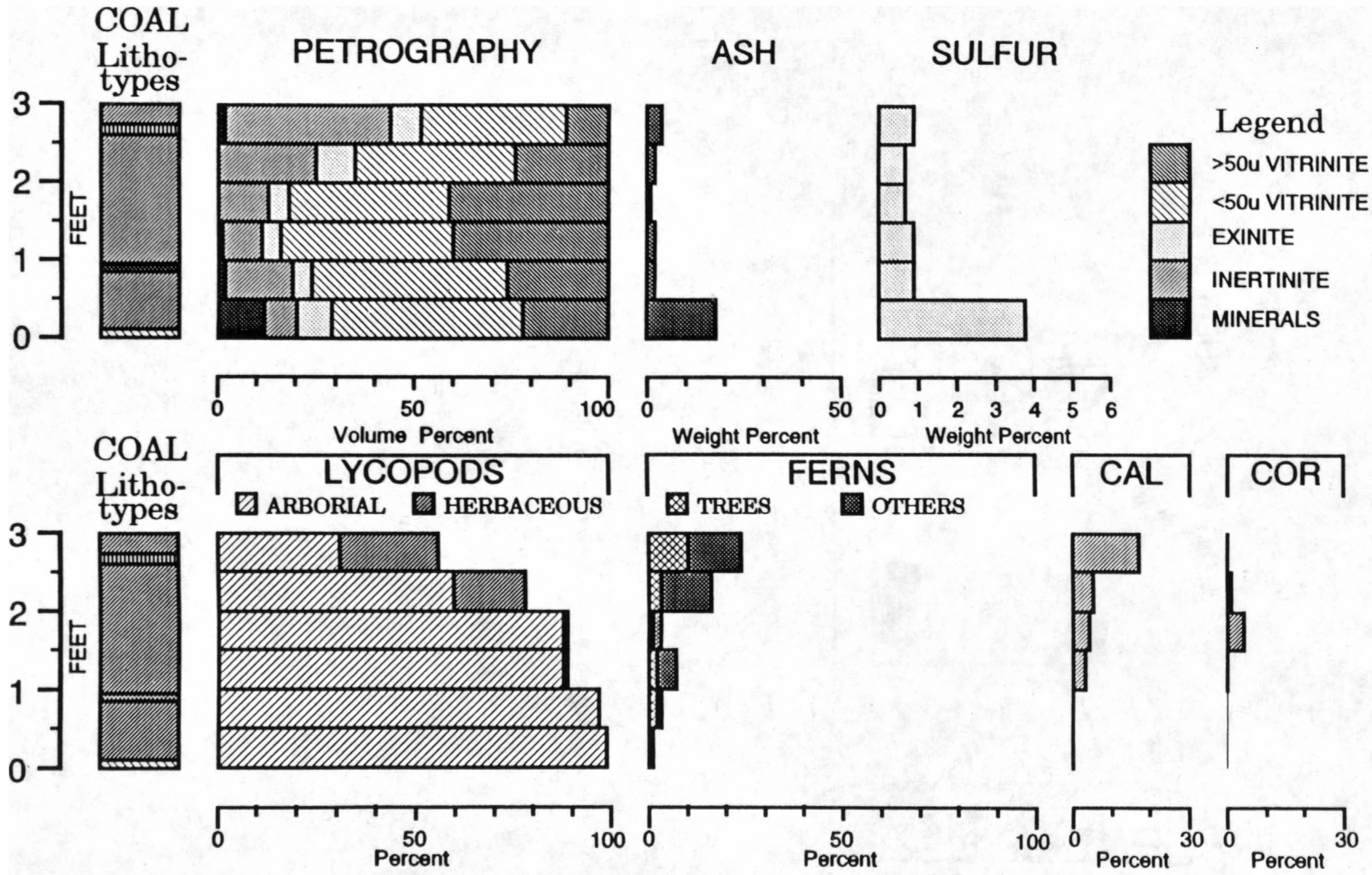


Figure 48) Distribution of macerals, miospores, ash yield and sulfur content in the No. 2 Gas coal bed at STOP 13.

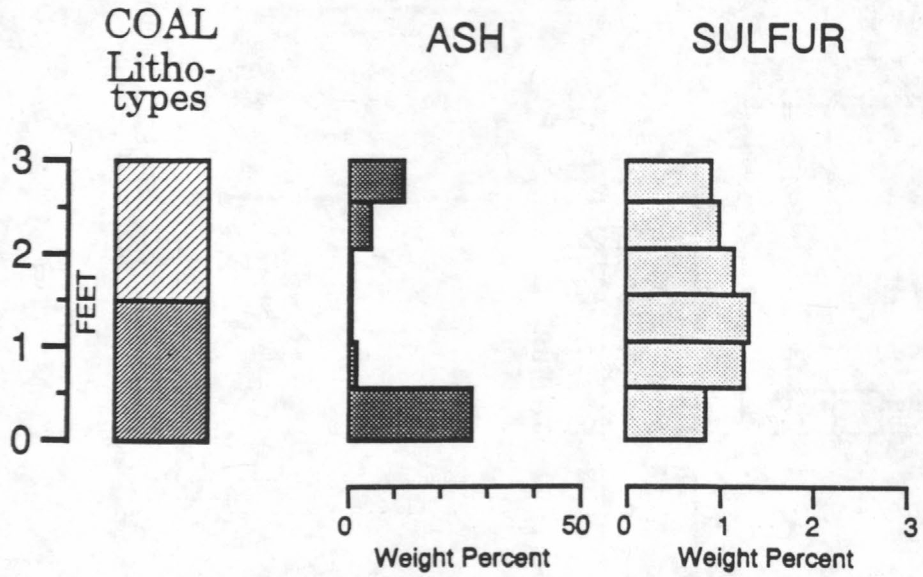


Figure 49 - Distribution of ash yield and sulfur content in the Peerless coal bed at **STOP 13**.

END OF DAY 2

DAY 3

Conemaugh Group strata in the tri-state area

Ronald L. Martino

Stratigraphic Framework

The stratigraphic framework of the Conemaugh Group in the Huntington, WV, and tri-state area, is similar to that at **STOP 3.** The Conemaugh Group is divided into the Glenshaw Formation (73 m, 235 ft thick) and the Casselman Formation (88 m, 282 ft thick; Fonner and Chappell, 1986). The Glenshaw Formation is particularly well-exposed along I-64 between Huntington, WV and Cannonsburg, KY, and along Rt. 23 in eastern Kentucky from I-64 south to Louisa. Strata dip southward up to 4 degrees along Rt. 23 from I-64 to about 2 miles south of Savage Branch (Merrill, 1988) on the northern limb of the Pittsburgh - Parkersburg - Huntington Syncline. A composite stratigraphic section based on outcrops from I-64 south to Savage Branch is shown in figure 50.

Geologic mapping of 7.5 minute quadrangles in the area of **STOPS 14 AND 15** (Dobrolvony et al. 1963; Spenser, 1964) was without the benefit of roadcuts along I-64 between Catlettsburg and Cannonsburg and along the 4-lane portion of Rt. 23. Subsequent studies by Martino and others (1985), and Merrill (1986, 1988) have helped to clarify the stratigraphy of the Glenshaw Formation in this area.

The upper Glenshaw and lowermost Casselman Formations are featured at **STOPS 14** and **15**. The vertical succession of lithofacies has much in common with exposures through the same interval along I-68 **(STOP 3)** near Morgantown, WV, demonstrating the regional extent of lithofacies. Three paleosols which occur in association with the Ames marine zone will be examined.

STOP 14 - Kentucky Route 23, 0.16 to 0.5 km (0.1 to 0.3 mi) NW of Savage Branch, roadcut along NE side of highway.

Stop Leader: Ronald Martino

The uppermost portion of the Saltsburg Sandstone is exposed at the SE end of the cut. The Saltsburg Sandstone is more fully-exposed in Rt. 23 roadcuts just south of this stop, and about 2.3 km (1.4 mi) north of this stop (Campbells Run Rd) where it is 23 m (74 ft) thick. The Saltsburg Sandstone in this area consists of up to four vertically stacked, alluvial channel-fills with epsilon cross-strata and abandoned channel shale plugs. Individual fills are up to 9 m (29 ft) thick. Cross--bedding measurements (26) indicate unimodal paleoflow toward the north-northwest.

The 11 m (35 ft) interval above the Saltsburg Sandstone contains two well-developed mudstone paleosols. They are separated by an interval of very fine sandstone and shale with features suggestive of a crevasse splay or shallow alluvial channel origin. The lower paleosol is 1.5 m (5 ft) thick at this location (fig. 50) and corresponds to the Pittsburgh Red Shale, a widespread unit that occurs throughout the Dunkard Basin. It contains vertic structures (mukkera), stress cutans (slickensides), clastic dikes, and evidence of abundant swelling clay; these features are characteristic of Vertisols (Retallack, 1988). The profile also contains features that develop in Aridisols including high Munsell values and extensive development of pedogenic carbonate (caliche). Vertisols are uniform, thick (more than 50 cm, 20 in), clay-rich paleosols with deep wide cracks for part of the year. Hummock and swale topography (i.e. gilgai microrelief) is the result from swelling and upward buckling of the soil along hummocks with deep fissures (Retallack, 1990). The swales between hummocks may receive sediment eroded from adjacent hummocks, as well as chemical precipitates from ephemeral lacustrine conditions. The concave-upward lenses of carbonates at the top of the lower paleosol at **STOP 14** appear to represent the later. Seasonal deposition of carbonate also contributed to carbonate fracture fills as alkaline waters filled deep open fissures connected to the surface. The carbonate lenses are regularly spaced laterally at intervals of about 6 to 7 m (19 to 23 ft) along the outcrop. Conjugate systems of slickensides are also developed as a result of clay heave. Vertisols usually are associated with low relief terrain and subhumid to semiarid climates (18 to 152 cm, 7 to 60 in rainfall/year) with a pronounced dry season. Aridisols develop in semiarid to arid regions and commonly have shallow calcareous horizons (Retallack, 1990).

Intermittent, active clastic deposition is represented by a 3 to 5 m (10 to 16 ft) interval of scour-fill sandstone and shales with root traces and plant fossils. A second paleosol occurs between this interval and the Ames Member at most locations. Here the paleosol is about 3 m (10 ft) thick. A ledge--forming micritic limestone 20 to 70 cm (8 to 28 in) thick occurs within this profile which appears to be pedogenic rather than lacustrine in origin. This paleosol is capped by coal from 1 to 33 cm (0.4 to 13 in) thick that may correspond to the Harlem coal bed of Krebs and Teets (1913). Carbonaceous shale with coal streaks and plant fossils often overlies the coal in this area.

Cecil (1990) proposed that climate fluctuations from humid subtropical to semiarid may have been a primary control on the cyclic repetition of chemical and siliciclastic rocks during the Late Pennsylvanian. Vertisols, Aridisols, and lacustrine carbonates developed under semiarid conditions with short wet season and long dry season. In contrast, sediment flux is accelerated with increasing rainfall during the wet season in seasonal climates that promote restricted vegetation in upland source areas. Conditions that promoted high ash yield, high sulfur content coal beds in topogenous peat swamps occurred during humid wet periods (Cecil, 1990).

The vertical succession of lithofacies at **STOP 14** reflects semiarid intervals with limited sediment flux which allowed paleosol development, interrupted by an episode of alluvial clastic influx. The cause of the increased rate of sediment influx may have been due to avulsion of alluvial channel systems into areas nearby. The widespread nature of this unit may alternatively suggest that increased runoff occurred in response to increased precipitation. The second paleosol is calcareous, yet capped by a coal bed which suggests a rise in the water table. This may have been caused by increased precipitation during a wetter climatic episode, or by a rising water table, which would be expected during the initial stages of the Ames Sea transgression.

The Ames Member is exposed on the first bench of the cut and is 8 m (26 ft) thick. It contains a diverse association of marginal to open marine facies including in ascending order: 1) a heterolithic tidal flat facies, with mudcracks and bivalve resting traces; 2) a shallow marine silty shale facies with crinoidal-chonetid packstones and sandstone dikes; 3) a very fine, bioturbated, calcareous sandstone with abundant crinoids stems and plates, and subordinate productids and gastropods ("crinoidal sandstone" of Krebs and Teets, 1913); the crinoidal sandstone facies is 1.2 m (4 ft) thick at this location and reaches a maximum thickness of about 6 m (19 ft); it probably represents a shallow marine sand bar or shoal; 4) a very fine, burrowed sandstone with shale partings, ripple bedding, and parallel to low angle cross-lamination (hummocky cross-stratification?) interpreted as a relatively low energy shoreface facies.

Wherever the Ames Member is not unconformably overlain by alluvial paleochannel sandstone, a thick calcareous paleosol with vertic features is developed. This is best-illustrated at **STOP 15** at the top of the hill (east side of Rt. 23).

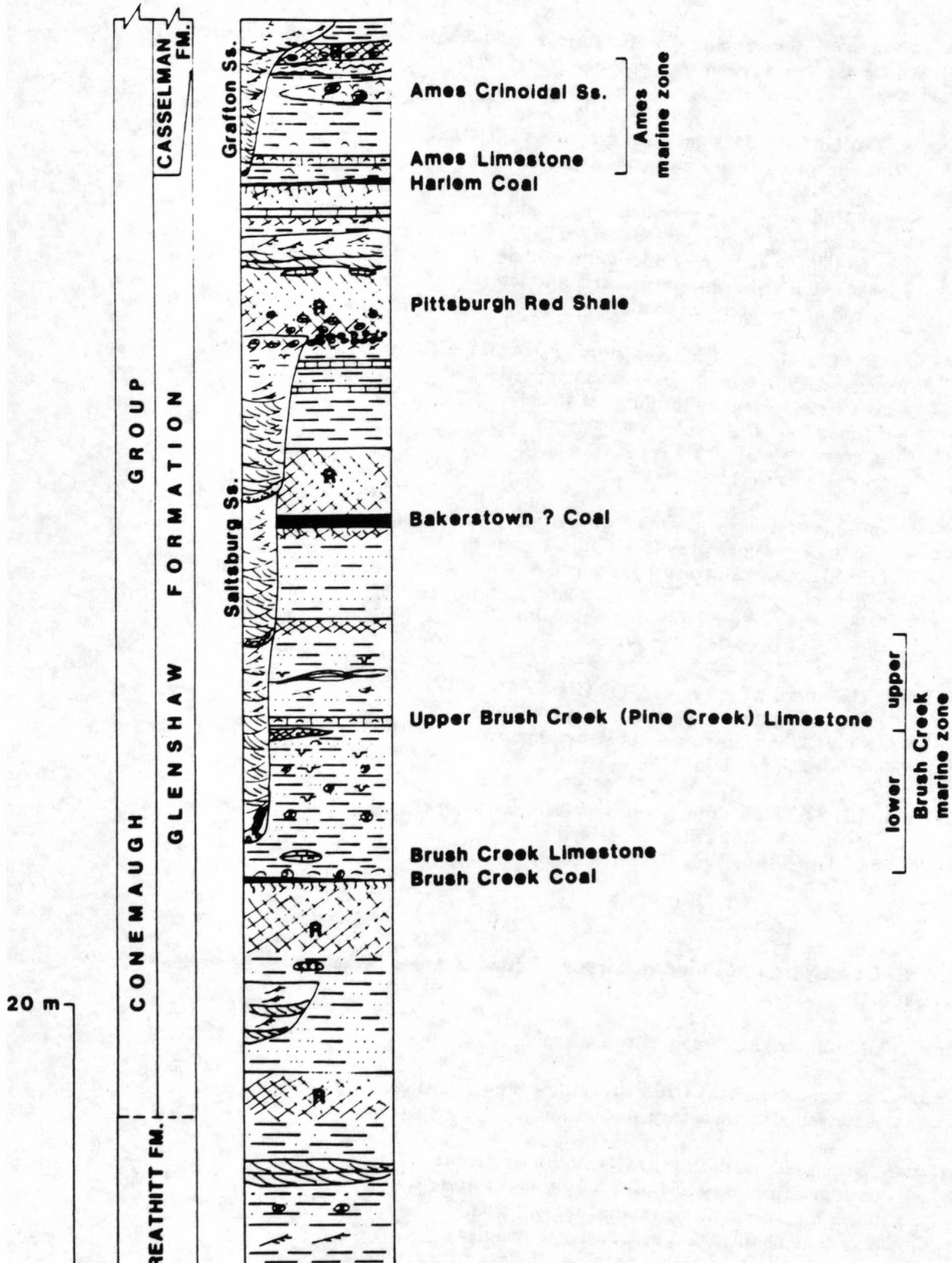


Figure 50) Stratigraphic column of Upper Pennsylvanian strata at **STOPS 14** and **15**.

Table 3 - **Paleosol description for Pittsburgh Red Shale along Rt. 23, milepost 7.9. All Munsell colors are from fresh unweathered surfaces.**

0-1.3 cm (A) Mudstone, dark greenish gray (N4 5GY 4/1), noncalcareous, sharp erosional upper contact

1.3-14 cm (A) Mudstone, dark greenish gray, weathers greenish gray, noncalcareous; laterally equivalent to light gray micritic limestone lenses which are concave upward, and laterally spaced at 4.8-5.6 m intervals

14-144 cm (Bk) Claystone, dark greenish gray (N4 5GY 4/1), moderately to strongly calcareous; micritic carbonate present as infillings of steeply inclined fractures 3.8-5.1 cm in width, and as greenish gray (N6 5 GY 6/1) to light gray (N7 5Y 7/1) nodules

144-296 cm (Bk) Mudstone, variegated, dusky red (7.5 YR N 4/2) and dark greenish gray (N4 4GY 4/1); weakly to moderately calcareous; slickensides common to abundant; fine to medium angular blocky peds; micritic carbonate nodules (light gray to greenish gray)

296-357 cm (K) Micritic Limestone, gray (N6 5Y 6/1) with angular fragments of greenish gray (N5 5G 5/1), calcareous mudstone; comprised of coalescing nodules

357-517 cm (C) Sandstone, fine grained, calcareous, sideritic; clastic dikes ?; micritic and sideritic nodules; calcite filled fractures at top.

Table 4 - **Description of Paleosol directly beneath Ames Member Rt. 23 Savage Branch (milepost 8.1).**

0-2 cm (O) coal, bright, mostly vitrain/clarain

2-17 cm (E) underclay, gray (N6) with yellow-brown mottles; platy soil structure, rare slickensides.

17-112 cm (Bt) claystone, dark gray (N4) in upper part to dark greenish gray (N4 5GY 4/1) in lower part; black metallic staining (manganese ?); small ironstone nodules (2-3 mm); siderite nodules in lower part; generally noncalcareous except

Table 4 continued...

for micritic nodules up to 5 cm in diameter in lower 15 cm.

112-127 cm (K) Micrite, dark greenish gray (N4 5GY 4/1), light greenish gray, and weak red (10R 4/3), angular fragments

127-187 cm (Bk) Mudstone, dark greenish gray (N4 5GY 4/1), moderately calcareous, sandy, with blocky angular peds

187-197 cm (K?) micritic limestone, brecciated

Table 5 - Paleosol directly above Ames Member, Rt. 23, Milepost 8.4.

0-0.5 cm (O) carbonaceous shale, claystone, dark gray (N4), sharp upper contact with overlying crudely bedded, greenish gray mudstone

0.5-38 cm (E) claystone, greenish gray (N5 5G 5/1), silty, with dark greenish gray peds and brownish gray cutans; slickensides common; peds 1-2 cm in diameter and sub-angular.

38-118 cm (EB) claystone, greenish gray (N5 5G 5/1) to dark greenish gray (N5 5GY 5/1) with subordinate weak and dusky red mottles (7.5 YR 4/2 and 7.5 YR 3/2); angular blocky peds, transitional top and base; weakly calcareous with 1-2 mm carbonate nodules.

118-228 cm (Bt) claystone, dusky red (10R 3/3) to olive (5Y 4/3), weakly to moderately calcareous; well developed soil structure including angular blocky peds 1-4 cm in diameter with red clay cutans; abundant slickensides.

228-238 cm (B) clay, light olive green, plastic

238-298 cm (B) mudstone, dark greenish gray, with angular peds

298-318 cm (B) clay, light olive green, weathers rust orange

(top of Ames Mbr, very fine sandstone)

STOP 15: Conemaugh Group strata along Kentucky Route 23

Stop Leader: Ronald Martino

The two paleosols beneath the Ames Member are exposed at the north end of the cut where they are separated by crevasse splay or shallow alluvial channel sandstone. Only the upper portion of the lower paleosol is exposed and it is truncated by erosion. Vertic features are evident. Nearly the entire Ames Member is channeled out by the Grafton Sandstone in the middle of this cut. Previous workers (Donaldson, 1979; Donaldson et al., 1985) have suggested that the Grafton Sandstone was formed by a northward prograding fluviodeltaic system, the Grafton delta, that advanced into the Ames Sea. The presence of a well-developed paleosol that conformably overlies the Ames Member throughout this area suggests that the Ames Sea withdrew first, exposing the coastal plain to pedogenesis during a period of low sediment flux. The Grafton Sandstone truncates this paleosol wherever the two are present. Preliminary reconnaissance in the Morgantown area along I-68 (**STOP 3**) suggests the same scenario. These observations suggest that the Grafton delta may never have existed, and the Grafton Sandstone is part of an alluvial succession that postdates the presence of the Ames Sea, as well as the nondepositional, soil-forming episode that followed. Although some paleosols, such as Entisols or Histosols, might develop in the inactive portions of delta plains, a calcareous profile with vertic structures suggests a seasonally dry climate and a lowered water table associated with a well-drained coastal plain, an environment not conducive to peat formation.

At least seven widely-developed paleosols within the Glenshaw Formation (fig. 50). These paleosols suggest an allocyclic causal mechanism such as climate change, episodic source area tectonism or base level changes, which interrupted sedimentation and punctuated the stratigraphy. Heckel (1986) has argued that glacioeustatic sea level changes were responsible for transgressive-regressive cycles in Late Pennsylvanian strata of the midcontinent. It is likely that the Appalachian basin also was affected. One explanation for the cyclic alternation of paleosols and episodes of more rapid siliciclastic deposition could be base level changes which affected the behavior of alluvial systems draining across the coastal plain. Paleosols may have developed during sea level lowstands as channels became incised and sediment by-passing of the coastal plain occurred. Aggradation of fluvial systems occurred in response to rising sea level. Where transgressions did not reach as far as the Appalachian basin, active floodplain deposition over a broad area may have been the updip response. Thus, the cyclic alternation of splay deposits and paleosols (as well as other features associated with a fluctuating water table) could conceivably be caused by sea level fluctuations. Busch and Rollins (1984) proposed that 11 depositional allocycles are preserved in the Glenshaw Formation of the Dunkard basin that were the result of sea level changes. However, without higher resolution in stratigraphic correlation between the Appalachian basin and the mid-continent than currently exists, it is difficult to demonstrate sea level changes as the primary factor in controlling paleosol development. It is a viable allocyclic mechanism that needs to be tested along with climate change as a possible mechanism for cyclic sedimentation and pedogenesis.

THE MISSISSIPPIAN-PENNSYLVANIAN SYSTEMIC BOUNDARY ON THE WAVERLY ARCH APICAL ISLAND

Frank Ettensohn

Background

STOPS 16 through **18** are intended to demonstrate the interaction between paleoclimate and local structure in the formation of unique lithologies in the Mississippian section and in parts of the section transitional between the Mississippian and Pennsylvanian. Although the effects of local structures are particularly apparent at these stops, the resulting uplift and exposures are contemporaneous with similar events across much of the continent at this time. This is suggestive of eustatic changes and/or a period of synchronous tectonic uplift.

Some of the more important paleoclimatic changes occurred during the Late Mississippian when southeastern parts of Laurussia, where the eastern U.S. was situated, migrated northward from the dry, evaporative trade-wind belt into the moist, humid equatorial belt (Shutter and Heckel, 1985; Ettensohn, 1990) as a result of the northward movement of Gondwana. According to reconstructions by Scotese and McKerrow (1990), the Kentucky area moved from about 15 degrees south latitude in the Meramecian (Visean) to about 7 degrees south latitude in the Chesterian (Namurian A; Serpukhovian). By the Middle Pennsylvanian (Westphalian B-D), when most of the Pennsylvanian rocks in northeastern Kentucky were deposited, Kentucky had moved to within five degrees of the paleoequator (Scotese and McKerrow, 1990) well within the moist, humid, equatorial belt.

Structural Framework

Northeastern Kentucky is underlain by two structures (fig. 51) which were active during the Carboniferous: 1) a broad, east-west basement fault system with some surface expression in the Kentucky River fault zone and 2) a north-south-trending uplift named the Waverly arch by Woodward (1961). Both structures were active during the early Paleozoic (Woodward, 1961) and again during the late Paleozoic when northeastern Kentucky was predominantly positive (Dever and others, 1977; Ettensohn, 1979a, 1980, 1981).

The fault zone (fig. 51) marks the northern hinge line of the east-west-trending Rome trough in the subsurface. Surficial stratigraphy, deep drilling and geophysical evidence along the fault zone indicate a northern, uplifted block where Carboniferous units are thin, and a southern downdropped block where units are thicker. The transition between these blocks occurs via a series of en echelon faults that progressively downdrop to the south (fig. 52). During the Carboniferous, especially during the Mississippian, these faults were apparently reactivated at least four times, so that near these faults, units are abnormally thin or absent and the number of unconformities in the sections increase. **STOP 18** near the town of Frenchburg occurs near one of the faults, and the section there exhibits many of these characteristics.

The Waverly arch is a small concealed arch, about 10 mi across from limb to limb, east of the larger Cincinnati arch (fig. 51). It apparently developed above a basement fault and has exhibited slightly different axes of uplift at different times (Ettensohn, 1979a; Pashin and Ettensohn, 1987). The Waverly arch coincides with the Mississippian outcrop belt for nearly half its length in northeastern Kentucky and appears to have influenced Carboniferous sedimentation greatly in northeastern Kentucky. One part of the arch, called the *apical island* by Ettensohn (1975, 1977) experienced far greater uplift than other parts and was a positive or near-positive area throughout much of the Carboniferous. The Lee (Morrowan, late Westphalian A-B) and Paragon (former Pennington Fm; late Chesterian, Namurian A) formations are generally absent on this part of the arch, and the Borden and Slade (former Newman Ls; Osagean-Chesterian, latest Tournaisian to early Visean) formations undergo drastic thinning. **STOP 16** occurs on the apical island whereas **STOP 17** is on the western flank of the arch (fig. 52).

The presence and effects of these structures has been in part determined by noting the occurrence and distribution of unconformities (fig. 53), as well as the distribution of certain Slade members (fig. 54). Structural influence is especially evident in the development of unconformities which are largely restricted to or better developed on and near structures. The unconformities are typically characterized by loss of section as

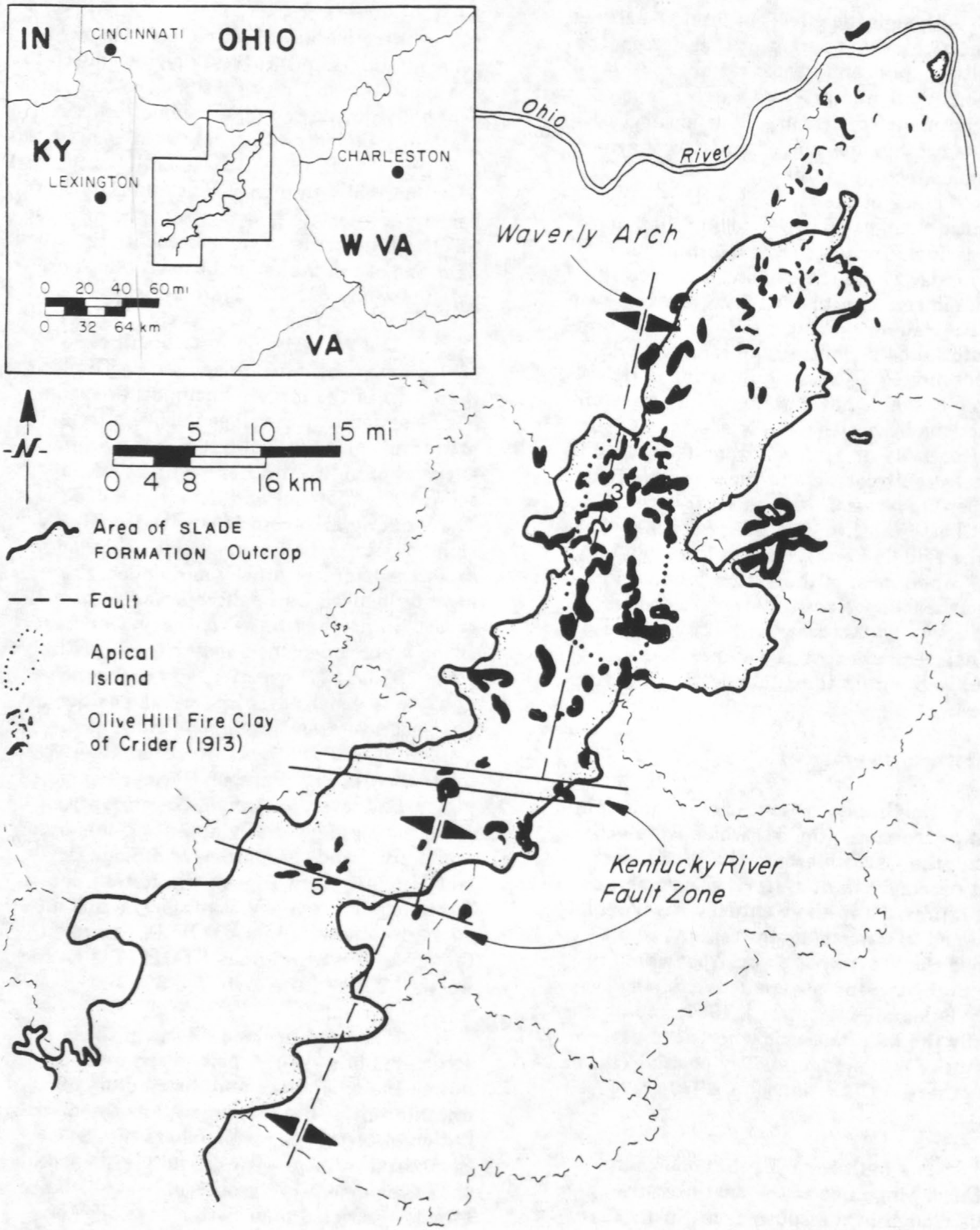


Figure 51) Composite map showing prominent structural and outcrop features in northeastern Kentucky in the vicinity of **STOPS 16, 17 and 18.**

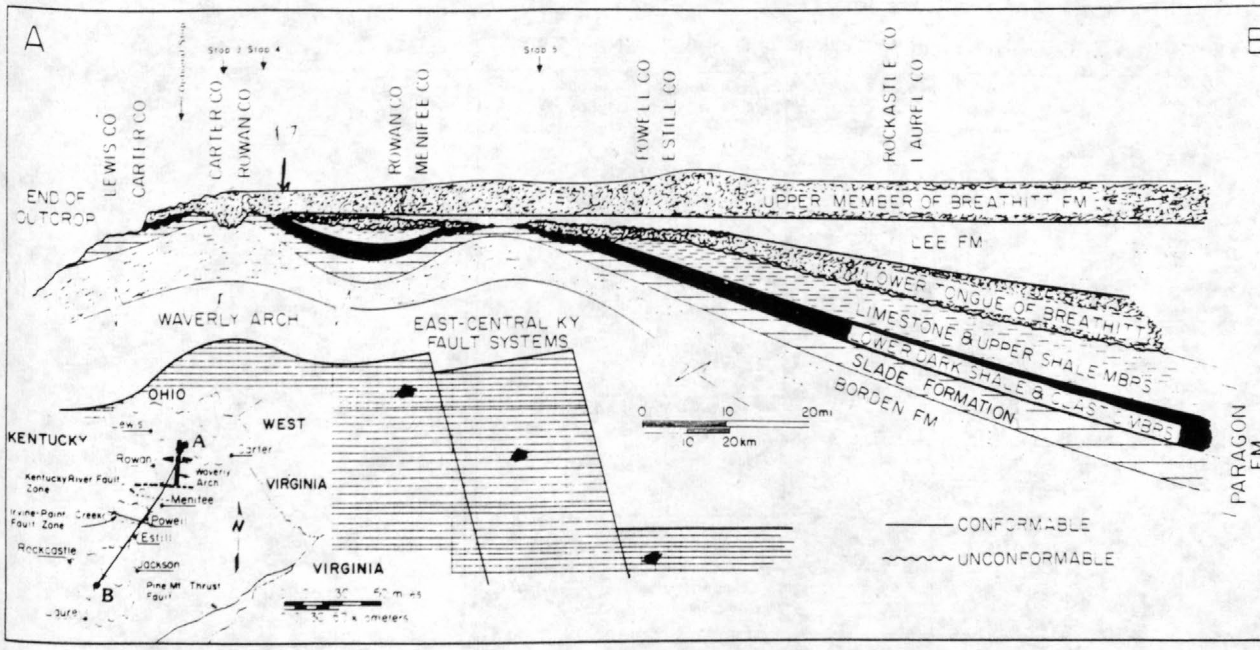


Figure 52 - Schematic North - South cross section along the east-central Kentucky outcrop belt showing inferred relationships between major Carboniferous Formations, unconformities and underlying structure. The numbers show the relative locations of stops (modified from Ettensohn, 1980).

well as development of paleosols and paleokarst.

The earliest of these unconformities occurs on top of the St. Louis Member (fig. 53), which we will examine at **STOP 18**. The effects of subaerial weathering and vadose diagenesis on this unconformity are more pronounced along the Wavery arch and fault zone (fig. 51) suggesting that these structures were the focus of uplift. The effects of this uplift are apparent in the distribution of the overlying Ste. Genevieve Member (fig. 53), which is restricted to the downdropped side of the fault zone and to either side of the Waverly arch (fig. 54B). Probably the most severe period of uplift and erosion occurred after Ste. Genevieve deposition at the Meramec-Chester transition (late Visean). This uplift may represent a flexural response to Ouachita tectonism in the south (Ettensohn, in press a,b). Uplift of the Ste. Genevieve was accompanied by the formation of deep erosional lows on either side of the Waverly arch which penetrated the Ste. Genevieve (fig. 54B), the St. Louis (fig. 54A), and the Borden Formation to the level of the Cowbell Member (fig. 55), a situation which we will observe at **STOP 17**. The subsequent early Chesterian transgression began by filling these lows with a tidal channel-tidal flat continuum represented by the Warix Run and Mill Knob members (fig. 53). The Warix Run Member begins the filling of the lows (fig. 54C), whereas the Mill Knob Member onlaped inter-low areas and the Waverly arch itself (fig. 54D); both members are restricted to either side of the Waverly arch (fig. 54 C,D). The unconformity on top of the Mill Knob (fig. 53) and its equivalents in Kentucky is so widespread that it may reflect eustatic control, but a structural component can not be ruled out. A fourth unconformity, restricted to areas near structural features, developed during early Paragon deposition (fig. 53) in the middle Chesterian (Namurian A). We will see definite evidence of this unconformity at the optional stop and probably

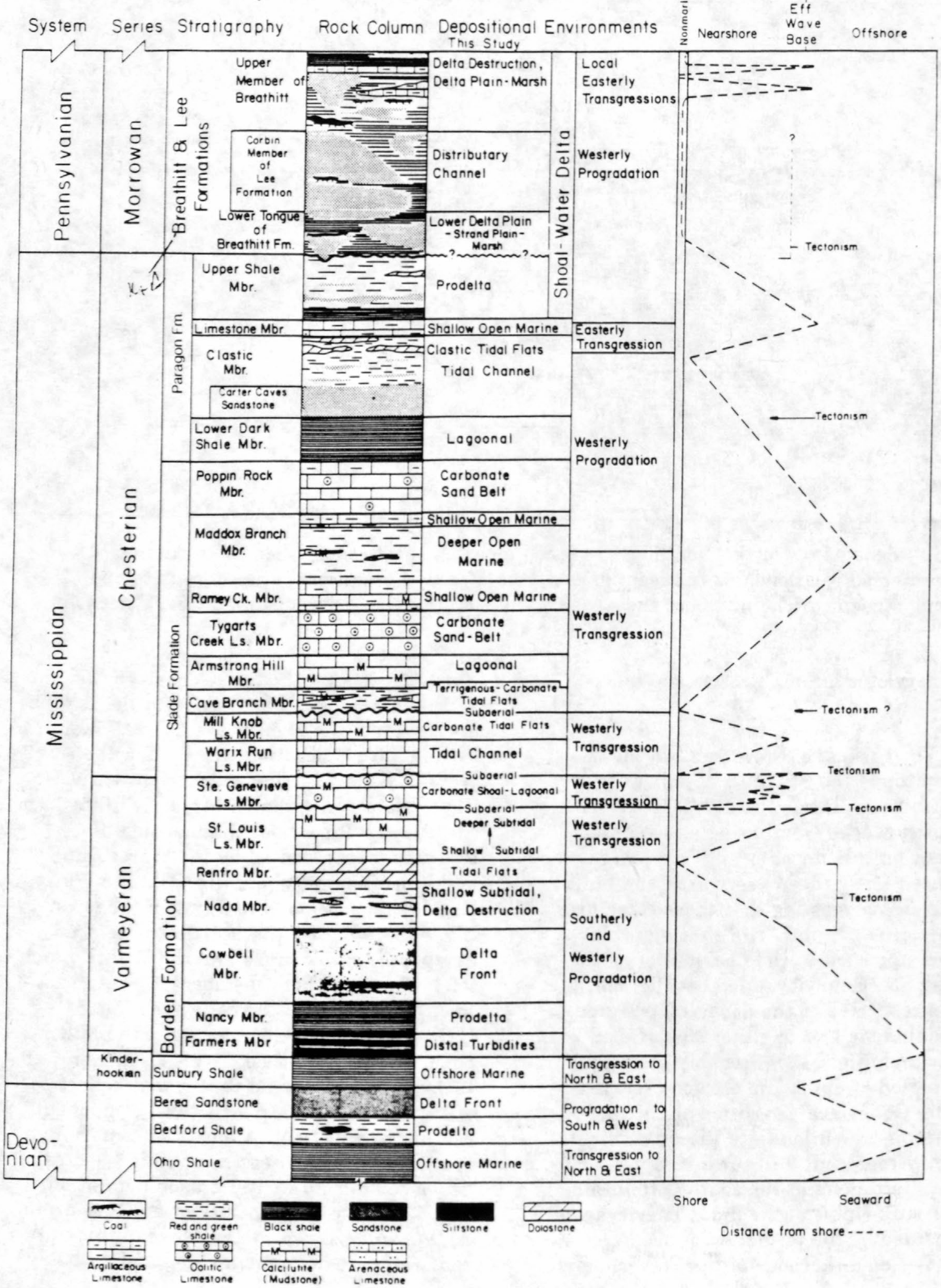


Figure 53) Stratigraphic column and paleoenvironmental interpretations of units in Upper Mississippian and Lower Pennsylvanian strata seen in the vicinity of **STOPS 16, 17 and 18.**

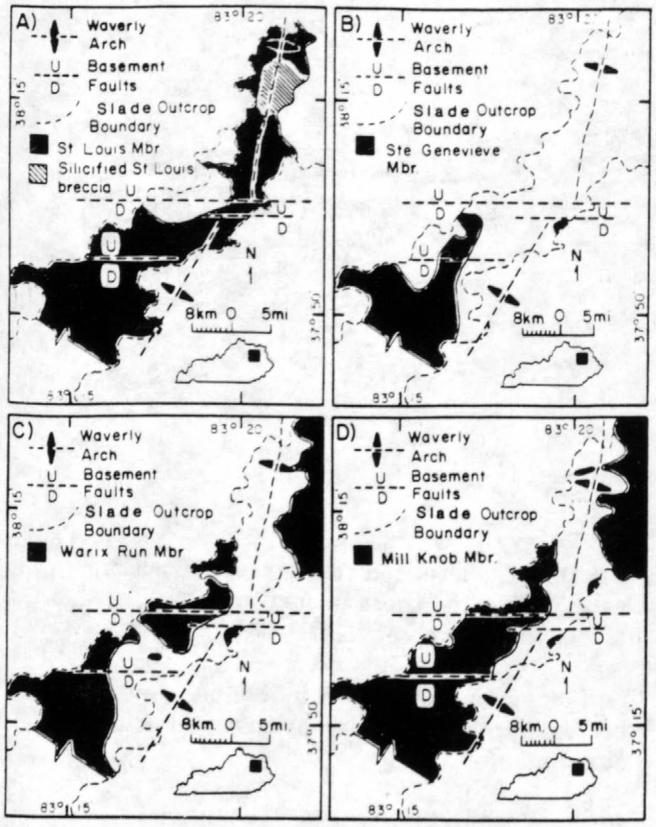


Figure 54 - Disjunct distribution of the 4 lower Slade members in notheastern Kentucky (modified from Ettensohn, 1980).

at **STOPS 16 and 18**.

The fifth unconformity in this area is the so-called "Mississippian - Pennsylvanian" unconformity, which is partly the product of Early Pennsylvanian uplift and erosion (e.g. Englund and others, 1979; Chesnut, 1989), and probably is related to the advent of the Alleghanian orogeny (Ettensohn and Chesnut, 1989). The presence of this unconformity elsewhere on the North American continent may reflect synchronous periods of orogenesis or eustatic change. The distribution of the underlying Paragon Formation relative to structures makes it apparent that Pennsylvanian uplift on these structures was involved in unconformity formation. This unconformity will be apparent at **STOPS 16, 17, and 18**.

Recognizing all of these unconformities in any one area may be difficult because on the structural features themselves, where uplift and erosion have been greatest, these unconformities tend to merge forming compound unconformities (fig. 55). Hence, the great amount of erosional truncation beneath Pennsylvanian rocks on the apical island at **STOP 16** is not necessarily the product of any one period of uplift and erosion, but rather the combined product of five periods of uplift and erosion (fig.55). Greater detail about the nature and origin of these rocks and related unconformities can be found in work by Dever and others (1977), Dever (1980), Ettensohn and Peppers (1979), Ettensohn and Dever (1979a), Ettensohn (1980, 1981) and Ettensohn and others (1984a, 1988).

Stop 16 - The Olive Hill Clay Bed of Crider (1913) on the Waverly Arch Apical Island

Stop Leaders: Frank Ettensohn and Blaine Cecil

Commonly associated with the Early Pennsylvanian or "Mississippian-Pennsylvanian" unconformity (also the subject of **STOP 1**) is the Olive Hill Clay Bed of Crider (1913), a clay bed varying in texture and structure from a flint clay to a plastic clay (e.g. Crider, 1913; Patterson and Hosterman, 1961; Smyth, 1984). The flint clay at this stop apparently developed high on the apical island (fig. 51) and developed about 2.5 m (8 ft) above the unconformity on lower Middle Pennsylvanian (upper Westphalian A) dark gray shales, based on the age of a nearby coal bed (C.F. Eble, personal communication, 1992). However, the clay may just as likely develop on Mississippian carbonates or shales of the Slade and Paragon formations (fig. 53) at the unconformity (e.g. Crider, 1913; Smyth, 1984), as we will see at **STOP 18**. As at this stop and at **STOP 18**, the clay is commonly overlain by a thin coal or coaly shale which may be related to clay formation. Palynological dating of these coal beds suggests that they, and probably the clay-forming event, could be as young as Middle Pennsylvanian

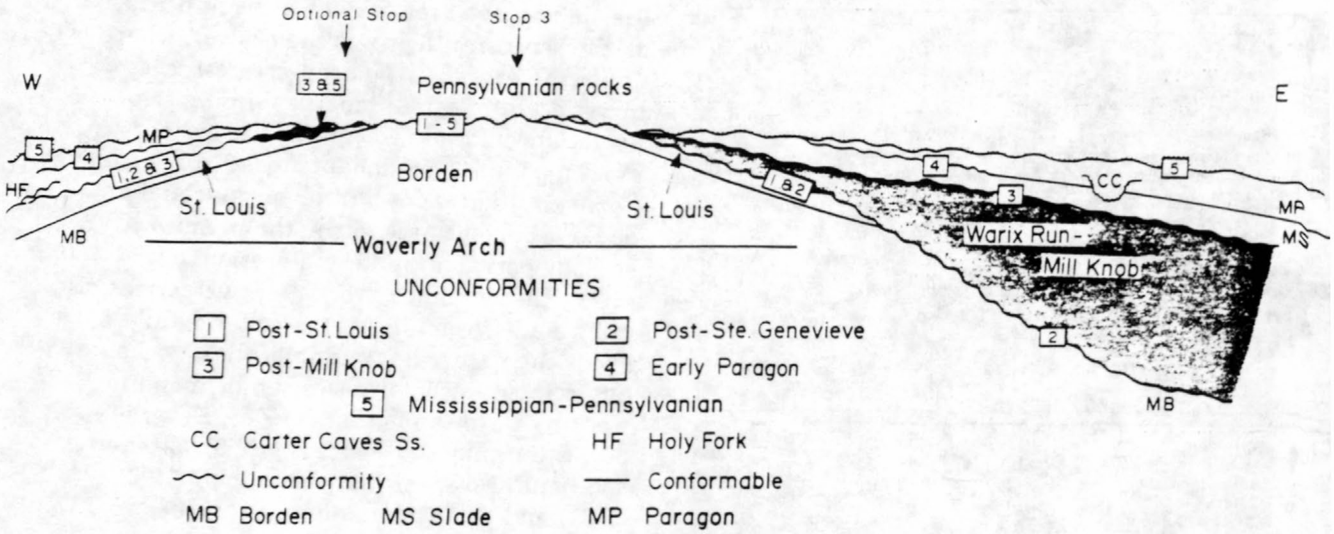


Figure 55 - Schematic east-west section along I-64 showing the generalized stratigraphic positions of 5 Carboniferous unconformities relative to the Waverly Arch. All of these unconformities coalesce into a single compound unconformity near the apex of the arch at **STOP 16**.

"Late Corbin" or "early Breathitt" time (C.F. Eble, personal communication, 1992); however, there are also suggestions of a late Mississippian age for some of the clay deposits (Ettensohn, 1977; Anderson and Hester, 1977; Ettensohn and Peppers, 1979; Smyth, 1984). The fireclay at this stop and most others in this area occur near the base of the Pennsylvanian section and are included in the lower part of the Breathitt Formation (fig. 56).

The distribution of the Olive Hill Clay is patchy (fig. 51), but provides some interesting insights into its formation. The clay is almost wholly restricted to the northern uplifted fault block and seems to be more concentrated on and near the apical island (fig. 51). When the clay is present south of the fault zone, it everywhere seems to occur on or near structures, which is the case with the clay at **STOP 18** near Frenchburg (fig. 51). The clay is commonly restricted to erosional lows or paleokarst in the underlying Borden, Slade, or Paragon formations or may occur on bayfill shales in the Breathitt, although some of these bayfills may also reflect underlying paleokarst. This kind of occurrence explains the patchy distribution of the clay (fig. 51) and lens-like geometry of individual clay bodies, part of which can be observed at this stop (fig. 56).

The origin of these clay deposits is most likely related to the interaction between regional uplift in the area and paleoclimate. After Early Pennsylvanian regional uplift, the whole region was subjected to a period of subaerial exposure and erosion, resulting in the very irregular unconformity surface. Based on the distribution of units below the unconformity, both the Waverly arch and basement fault zone seem to have undergone renewed uplift at this time. The elevated state of the northern uplifted block accounts for the deeper erosion here along the unconformity, and as indicated by Short (1978, 1979), prevented subsequent, large-scale, deltaic encroachment into the area during the Early Pennsylvanian.

Once uplifted, the exposed rocks were apparently subjected to a long period of erosion and weathering. Based on the chronology of Harland and others (1989), this episode of Early to Middle Pennsylvanian erosion probably lasted in excess of five million years. Based on inferred paleogeography

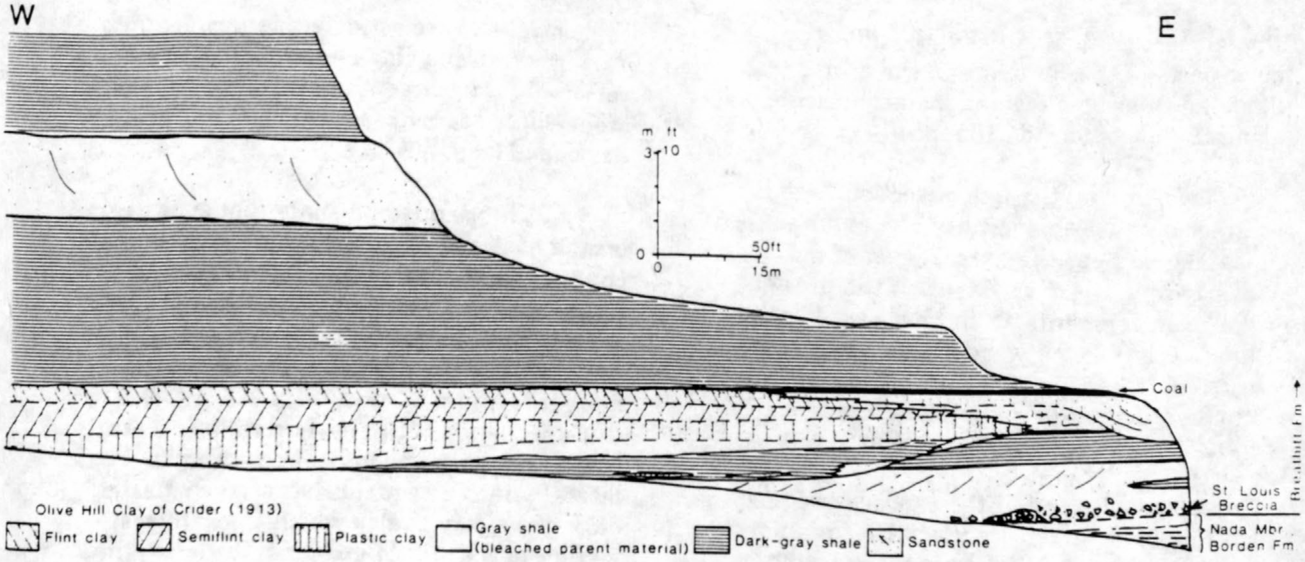


Figure 56 - Schematic east-west section of the eastern half of an exposure along I-64 showing the Olive Hill Clay Bed of Crider (1913), developed on dark, Pennsylvanian shales (modified from Smyth, 1984).

(Scotese and McKerrow, 1990) and inferences from the coal and associated rocks (e.g. Phillips and Peppers, 1981; Cecil and others, 1985; Cecil, 1990), this surface was most likely exposed to weathering in subtropical to tropical climates in humid, equable conditions with only occasional periods of wet-dry seasonality. These kinds of interpretations are also supported by the necessity for pedogenic environments of clay formation, which required abundant plants, hydromorphic conditions, high temperatures, and precipitation for intense chemical leaching and translocation of clays (Smyth, 1980).

Although uplifted, the eroded surface was pitted with erosional lows and paleokarst in which hydromorphic conditions were easily established (e.g. **STOP 18**). Other parts of the surface were apparently covered by very shallow brackish or marine embayments (e.g. **STOP 16**) in which suspended clays and silt could be carried over the surface and washed into erosional lows or embayments.

These low areas eventually accumulated sufficient thicknesses of clay to support marsh and swamp vegetation at or slightly above sea level. The thin coal commonly overlying the fireclays are evidence for these swamps. The overlying swamps provided the decreased circulation and increased acidity necessary for pedogenic alteration of the underlying clays into kaolinite-rich fireclays through leaching and upward diffusion (Keller and others, 1954; Patterson and Hosterman, 1961). The positive nature of the northern uplifted block and Waverly arch served to enhance terrestrial swamp growth, as well as the weathering processes which altered the clays (see Keller and others, 1954, for a similar example), and explains why the clay are generally thicker, better-developed, and more extensive on and near the apical island (fig. 51).

Although geologic mapping (Philley and others, 1975) did not show this particular deposit of the Olive Hill Clay because the interstate had not been completed then, this deposit is very similar to other nearby clay deposits (e.g. Philley and others, 1975; Ettensohn and Dever, 1979b; Ettensohn, 1981). The Olive Hill Clay at this locality is in the lower Breathitt Formation which unconformably overlies the Nada Member of the Borden Formation. At least 46 m (150 ft) of Mississippian section in Kentucky, including uppermost parts of the Borden and the entire Slade and Paragon formations are missing along the unconformity (fig. 55).

Of course, the loss of section here is so great because this locality occurs on an apical island near the axial region of the Waverly arch (fig. 51). Based on the presence of Paragon remnants at another clay locality 1 km (1.6 mi) to the west of I-64, most of the erosion on this part of the arch apparently occurred during the early Paragon and Early Pennsylvanian episodes of uplift (fig. 55).

The east end of the exposure begins with unaltered green Nada shale. The uppermost 0.6 m (2 ft) of the shale are leached, bleached, and locally melanized, suggesting that this part of the Nada may reflect an even older, Early Pennsylvanian truncated soil. The Nada is overlain by about 2 m (7.6 ft) of yellowish to light-gray clayey sandstone and orthoquartzitic sandstone grading up into dark-gray, fissile, clay shales. The clayey sandstone immediately overlying the Nada contains a basal conglomerate of silicious or chertified, St. Louis fragments. In places, large lithified masses of this silicified St. Louis material sit on top of weathered Nada shales suggesting that together they may have formed a former silcrete soil, which was subsequently eroded and included in the basal Pennsylvanian sands. These residual accumulations of silicious St. Louis masses or breccias are restricted to the apical island (fig. 54A), and if they are silcrete-type soils, then they probably reflect an Early Pennsylvanian development in alternating wet and dry conditions (see Blatt, 1992, p. 42).

The thin orthoquartzitic sands, which contain these breccias here, are relatively common overlying the unconformity or infilling paleokarst (e.g. at **STOP 18**) on the underlying Slade. Their origin is uncertain, but they could represent the most distal or marginal deposits of one of the Lee sand belts (e.g. Chesnut, 1989) or reworking of a Mississippian sand like the Carter Caves Sandstone (Fig. 53) along the unconformity. Whatever their origin, these sands grade upward into dark-gray, fissile clay shales, which are erosionally truncated and overlain by a sandstone-shale sequence on which the fireclay developed (Fig. 56).

This sandstone-shale sequence begins on the east end of the exposure with a small channel sandstone containing wavy- and flaser-bedded forsets which are rooted. This sand is thicker and better developed on the south side of the highway. The channel emerges from and truncates a 1.2 m (3.9 ft) thick sequence of intensely rooted or bioturbates argillaceous sandstones which grade laterally into a sequence of formerly fissile, dark-gray, clay shales which thicken westwardly into a subtle erosional low or channel cutout (fig. 56). The entire sequence was apparently overlain by a thin, impure coal bed or carbonaceous shale which was truncated by an event preceding deposition of the overlying, dark-gray, fissile bayfill shale. Although, the marsh or swamp represented by the overlying coal was probably necessary for fireclay formation, the coal and carbonaceous shale presently extend only over the eastern 76 m (240 ft) of the exposure. However, close examination of the thinning margin of the coal shows that it is actually a lag concentrate of coaly fragments and fireclay clasts. Beyond the margin of the coal, the top of the fireclay is actually a subtle discontinuity with erosional relief.

The fireclay in this exposure is certainly a paleosol inasmuch as it shows evidence of having supported plants and having developed horizon zonation and soil structure (e.g. Retallack, 1976). Rooting, of course, is present throughout much of the clay deposit, but is commonly best developed in upper parts of the deposit. Moreover, this deposit very clearly shows horizon development. The coal and upper organic-rich clay just below the coal, generally no more 0.2 m (0.5 ft) thick, represent the A horizon. Most of the clay bed itself, which is about 1.8 m (6 ft) thick, represents an argillic B horizon or an "illuvial B" horizon (Duchaufour, 1982), char

acterized by an accumulation of illuvial clays or iron and aluminum oxides. In this case, the B horizon itself is additionally zoned into an upper flint clay, a medial semiflint clay, and a lower plastic clay. Clay-mineral analysis by Smyth (1984) indicates that the uppermost subzone or flint clay, where leaching and eluviation were greatest, is nearly pure kaolinite. Kaolinite decreases in abundance as the amount of mixed-layer clays, illite, and chlorite increase in the transition from flint, to semiflint and plastic clays lower in the profile. In addition, the maroon coloration, which typically characterizes lower parts of the B profile, appears to represent illuvial iron oxides that accumulated just above relatively unaltered and more impermeable parent material. The parent material or C horizon, a dark-gray fissile clay shale, although locally bleached and leached, still maintains fissility and original coloration.

Soil structure is present throughout the deposit on two levels. On the microscopic level, Smyth (1984) reported that the Olive Hill clays exhibit evidence of micropeds, cutans, domains, and glaebules. On the megascopic scale, the most apparent soil structures are peds, natural soil aggregates separated from others by joints, voids, or other planes of weakness (Ettensohn and others, 1988). The peds in this deposit vary from small platy and blocky peds at the top, to large, blocky irregular peds near the base of the B horizon. Cutanic concentrations of iron oxides and stress cutans in the form of slickensides occur along some ped boundaries. The siderite nodules that occur in various parts of the deposit generally replace soil structures and are not directly related to soil formation; they apparently reflect a later upwardly migrating water table (Smyth, 1984).

Although the Olive Hill clay bed in some places may represent multiple events (Smyth, 1984), the clay here appears to represent a single soil-forming event related to development of a marsh or swamp on top of a filled bay. Inasmuch as this kind of clay development typically requires swampy conditions, warm temperatures, and abundant rainfall (Keller, 1957; Moore, 1968), such clays are potentially good paleoclimatic indicators.

STOP 17 - Chesterian, humid-type residual paleosols from the Slade Formation

Stop Leaders: Frank Ettensohn and R. Thomas Lierman

The Slade Formation at this stop rests unconformably on top of the Cowbell Member of the Borden Formation and consists of approximately 21 m (70 ft) of carbonates, mudstone and shale; it is unconformably overlain by at least 2.7 m (9 ft) of orthoquartzitic sandstone and dark gray, fissile clay shales of the Breathitt Formation (fig. 57). The Slade in this exposure begins with the Warix Run and Mill Knob members filling one of the post Ste. Genevieve erosional lows developed on the west flank of the Waverly arch (fig. 54C).

In this low at least 14 m (45 ft) of section, including the Nada member of the Borden and the Renfro and St. Louis members of the Slade, as well as an undetermined thickness of the Cowbell Member of the Borden, are absent due to post-Ste. Genevieve erosion. Moreover, the Warix Run and Mill Knob intertongue complexly here and contain nine different soils (fig. 57), probably reflecting the fact that this locality occurs on the margin of one of the post-Ste. Genevieve erosional lows (fig. 54C), onto which seas were periodically transgressing and regressing.

The earliest soil is actually a late Genevievian or early Chesterian soil developed on the erosional top of the Cowbell Member. It obviously developed before Slade deposition and consists of a truncated C-horizon exhibiting nodular glaebules of diffuse calcrete, poorly-developed laminar crusts and possible root casts. The remaining eight soils occur in the Warix Run, Mill Knob, and Cave Branch members (fig. 57). Most of these represent thin, poorly-developed azonal soils or

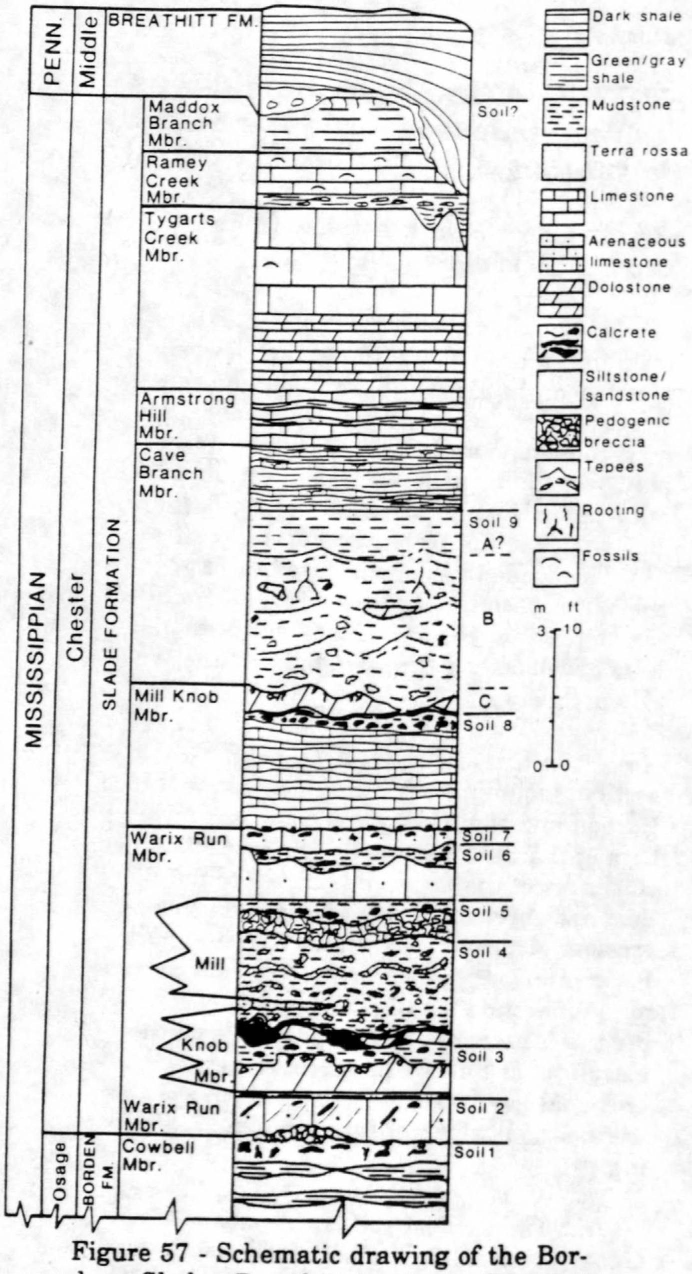


Figure 57 - Schematic drawing of the Borden - Slade - Breathitt section at **STOP 17.** Soil 1 is a post-Ste. Genevieve soil developed on the Borden, whereas the overlying 8 are early Chesterian paleosols. Soils 4 and 9 are the best examples of humid weathering.

soils deeply truncated to the lower B or C horizons. Of particular importance at this stop are soils 4 and 9, which are relatively thick, well-developed clayey residual soils (fig. 57) reflecting the predominance of solution during humid pedogenesis.

Soil 4 consists of 1.5 to 1.8 m (5 to 6 ft) of residual terra-rossa claystone containing Mill Knob micrite. One band of Warix Run arenaceous calcarenite is also present near the base. Layers of clasts and claystone undulate throughout the deposit suggesting that the calcilutite layers had been warped into caliche pseudo-anticlines ("tepees") before dissolution. Blocky peds and stress cutans (slickensides) are present throughout the unit. The parent material or C horizon for this soil is almost completely lacking, except for the isolated clasts, having been nearly destroyed by solution. Nonetheless, the relict clasts indicate that the parent material was probably a thin-bedded Mill Knob calcilutite or micrite, like that present below soil 8 (fig. 57).

The reddish-brown color of the residuum is a product of rubification, the release, progressive oxidation, and hydration of iron during dissolution of carbonates or other rocks (Buol and others, 1980; Duchaufour, 1982). Some of the secondary, bluish-green mottles in the claystones probably reflect gleization, the subsequent reduction of iron under hydromorphic conditions along joints, ped and cutan boundaries, or particularly porous parts of the residuum (Buol and others, 1980).

Immediately below the claystones of soil 4 is about 0.1 to 0.4 m (0.4 to 1.4 ft) of light-brown laminated dolostone and interpenetrating gray-brown calcite crusts and massive calcite at the base of this profile indicates illuviation of calcium carbonate from downwardly percolating solutions along an impermeable or plugged horizon formed by the dolostone (Ettensohn and others, 1988). Fractures in the dolostone apparently allowed penetration by the illuviating calcium carbonate, and by displacive crystallization these fractures were widened and infilled with calcite until individual blocks of dolostone were widely separated and contorted. A second episode of calcite illuviation is represented by the massive, light green calcite. It apparently penetrated and contorted both the preexisting crusts and the dolostone. It is also interesting to note that this thickest accumulations of this displacive calcite occur over deformed lows in the underlying dolostone, suggesting that the underlying dolostone may have still been plastic and that the compaction and vertical displacement generated by growth of the calcite may have been great enough to deform underlying areas (Ettensohn and others, 1988). Of course, it is also surface possible that pre-existing lows on the dolostone surface allowed for greater thicknesses of shales, which were more easily deformed and moved by the growing calcite.

Soil 9 includes parts of the Cave Branch Member and the uppermost 0.8 m (2.5 ft) of the Mill Knob Member (fig. 57); it probably includes parts of all three horizons. The C horizon in this soil is represented by the upper 0.8 m (2.5 ft) of Mill Knob dolomicrite and dolosiltite, which is melanized, brecciated, and exhibits nodular peds. In places it has been drastically thinned or is completely absent, having been completely altered to terra-rossa by dissolution. The overlying 3.1 m (10 ft) of reddish-brown chunky claystones are included in the Cave Branch Member (Ettensohn and others, 198-4b) and represents a terra rossa or "weathered B" horizon. Solution-rounded clasts and "lenses" of dolosiltite throughout reflect the nature of the parent rock. Pseudo-anticlines, caliche thrusts, and downward-branching root casts are often defined by stress cutans or slickensides and are outlined by bluish-gray reduced clays reflecting gleization. Some of the pseudo-anticlines are 0.3 to 0.6 m (1 to 2 ft) in amplitude and have periodicities of up to 9 m (30 ft). The overlying 0.9 m (3 ft) of dark-gray, chunky, claystone residuum contains finely disseminated organic matter and may represent the lower part of the A horizon. In fact, the dark-gray reduced nature of this part of the unit may be related to the presence of the organic matter. Although no rooting has been found here, in a similar Cave Branch deposit to the south, definite evidence of rooting was found.

Overlying parts of the Cave Branch Member here include dark-gray shales and interbedded calcilutites, which contain laminae, flaser-beds, mudcracks, runzelmarken, as well as vertical and horizontal burrows. This part of the Cave Branch represents a transgressive tidal mud-flat deposit on the leading edge of an early to middle Chesterian transgressing sea represented by the upper Cave Branch-through-Maddox Branch members of the Slade (fig. 53) all of which are exposed here (fig. 58). The dark-gray shales in tidal parts of the Cave Branch probably represent reworked soils from the underlying deposit.

Upper parts of the Slade and the entire Paragon Formation are absent along the Early Pennsylvanian unconformity near the top of the exposure. Although the Olive Hill Clays Bed is apparently well-developed on top of Slade carbonates in localities very near to this one (Hoge and Chaplin, 1972), it is absent here. As observed here, soils developed on top of the Warix Run and Mill Knob members generally lack the abundance of aggradational carbonate crusts seen in the St. Louis soils seen earlier. In contrast, the residual terra-rossa soils seen here are clearly degradational in nature and must reflect a major change in climate to more humid, rainy conditions near the Meramec-Chester boundary.

STOP 18 - Mississippian parts of the section at Frenchburg, KY: Structural influence along the Kentucky River fault zone.

Stop Leader: Frank Ettensohn

All the previous stops in the Mississippian outcrop belt were located on the northern uplifted block where the Waverly arch was the predominant structural influence (figs. 51, 52). In contrast, this stop is located on the southern downdropped block along one of the en-echelon faults in the Kentucky River fault zone (figs. 51, 52). Although Mis

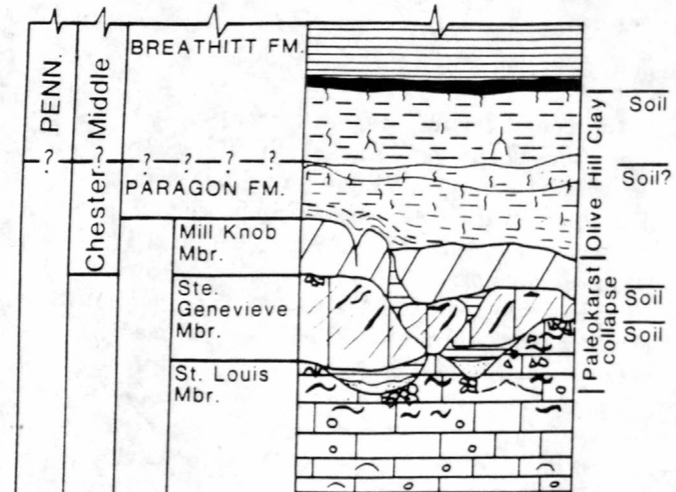


Figure 58 - Schematic drawing of the Mississippian section at **STOP 18**. The Olive Hill Clay Bed here may represent 2 early-Middle Pennsylvanian episodes of soil formation, one on Mississippian parent material, and the other on Pennsylvanian material.

sissippian Slade and Paragon sections are typically thicker and better developed on the southern downdropped block, sections on and near these faults are unusual in that they are thinner due to erosional truncation, exhibit better unconformity development, and are commonly disturbed by major paleokarst and solution collapse, as is the case of **STOP 18**. Most of these developments are probably related to the enhanced effects of uplift along the fault zone (Dever, 1973, 1977, 1980; Ettensohn, 1975, 1977, 1980; Ettensohn and Peppers, 1979; Anderson and Hester, 1977). Moreover, because this locality is located on the southern downdropped block away from the influence of post-Ste. Genevieve erosion, unlike the situation at **STOP 17** (fig. 57), the St. Louis reappears and the Ste. Genevieve Member is present (figs. 54 A-C, 59). Although the Warix Run Member is absent here (Figs. 54C, 59), early Chesterian seas did eventually transgress beyond the erosion lows toward the Waverly arch so that the Mill Knob Member, a part of the Warix Run depositional continuum (fig. 53), oversteps the Ste. Genevieve (fig. 54 B-D). As a result, a remnant of the Mill Knob Member unconformably overlies the Ste. Genevieve at this exposure (fig. 58).

The Mississippian section (fig. 58) begins with the fossiliferous green shales, siltstones and limestones of the Nada Member of the Borden Formation, which probably represents a shallow, open-marine, delta-destruction or lowstand event (Ettensohn 1979a,b). The culmination of this lowstand event is represented by the intertidal to shallow open-marine Renfro Dolostone Member of the Slade Member (fig. 53). Approximately 6 m (20 ft) of the Renfro Member appears to be gradational with the Nada below the St. Louis above (fig. 58). Up to 6.7 m (22 ft) of the St. Louis Member is present in the section, and it is basically a transgressive shallow open-marine sequence abruptly truncated by an unconformity and soil profile (Ettensohn and Dever, 1979c). The St. Louis soil here is a semi-arid aggradational paleosol characterized by thick accumulations of micritic crusts and breccias in the upper 0.6 m (2 ft). Other soil features include 1.5 to 2.1 m (5 to 7 ft) of melanization, the progressive downward darking of St. Louis micrites due to eluviation of organic matter from a former A horizon, and chert bands disrupted into caliche pseudo-anticlines or tepees.

The upper surface of the St. Louis which is normally an unconformity has been secondarily dissolved and eroded into small caves, together with the overlying Ste. Genevieve, probably by Early Pennsylvanian solution reflecting Early Pennsylvanian uplift on the fault zone. Although much of the overlying Ste. Genevieve has been destroyed or collapsed by this solution, solution was apparently halted in the upper St. Louis by the dense pedogenic micrites which acted as a porosity and permeability barrier against further downward movement of water. The small caves between the St. Louis and Ste. Genevieve were later infilled with probable Pennsylvanian orthoquartzitic sandstone and dark-gray shales (fig. 59) not unlike those present in basal parts of the Pennsylvanian section at **STOP 18**

The overlying Ste. Genevieve Member is a cross-bedded oolitic calcarenitic, reflecting a very shallow, high-energy carbonate sandbelt environment. It is up to 1.6 m (5.5 ft) thick, but occurs as large solution-rounded boulder-like bodies which collapsed onto the St. Louis (fig. 58). It was obviously the unit most involved in the Early Pennsylvanian solution and cave formation; locally Early Pennsylvanian sandstones and shale infill fissures within the Ste. Genevieve and between Ste. Genevieve "blocks." An unconformity with a paleosol is also present on top of the Ste. Genevieve, having formed during the post-Ste. Genevieve period of uplift and exposure (fig. 53). However, the unconformity and paleosols are poorly-preserved here, probably because of Early Pennsylvanian solution and collapse. However, the paleosol which is preserved shows melanization, a few poorly-formed micrite crusts, and diffuse calcrete paralleling cross beds (Ettensohn and others, 1988), a level of development which may be transitional between soils developed on the St. Louis and those formed on the Warix Run and Mill Knob (**STOP 17**; fig. 57).

The overlying Mill Knob Member is represented by a massive dolostone up to 1.7 m (5.5 ft) thick. Although dolostones are known from the Mill Knob, they are usually associated with thin-bedded calcilulites (Dever, 1973, 1980; Ettensohn and others, 1984a; Lierman and Mason, 1985), both of which lithologies are interpreted to represent very shallow restricted-marine to tidal-flat deposits. If the calcilutites were present at one time, they may have been effectively dissolved away during Early Pennsylvanian cave formation, for these dolostones were certainly involved in the collapse of the cave. In fact, the dolostones may have roofed part of the cave and avoided solution because of their lesser solubility.

The Mill Knob dolostones are unconformably overlain by sandy, green Paragon shales or poorly-developed sandy Olive Hill Clay, indicating a substantial hiatus. At one place, the green shales contain blocks of eroded chert, probably from upper parts of the Mill Knob or overlying upper Slade members no longer present. Based on nearby exposures (Ettensohn, 1975; Hoge, 1977), at least 21.6 m (71 ft) of Mill Knob and upper Slade carbonates and shales are missing along the unconformity. Although it is possible that some of the section loss may have been related to Early Pennsylvanian solution, the presence of the Paragon unconformably overlying the Mill Knob indicates that the early Paragon phase of uplift and exposure, discussed earlier at **STOP 16** was also active here. This phase of uplift is also apparent from other nearby exposures (Ettensohn, 1977; Ettensohn and Peppers, 1979), but the loss of section on the unconformity at these exposures was not as great. Although the Paragon remnants do appear to fill gaps and fissures in the Mill Knob dolostones, this is probably related to settling or slumping attending Early Pennsylvanian collapse of the underlying cave. Apparently the caves had not formed, or at least were not directly opened to the surface, during the early Paragon period of uplift and exposure, because there is no evidence of Paragon sediments in the cave fills.

At most, only 0.3 m (1 ft) of green sandy Paragon shales is preserved atop the Mill Knob, but this green shale grades into sandy plastic clay developed at the level of the Olive Hill clay bed. Although the Olive Hill clay bed was not formally mapped in this area (Hoge, 1977; Haney and Hester, 1978), it was identified in the area by Anderson and Hester (1977), Ettensohn (1977) and Ettensohn and Peppers (1979). It occurs at the same general position as the clay elsewhere, and the overlying coal or carbonaceous shale is of approximately the same age as the corresponding coal and shale at **STOP 16** and at **STOP 1** (C.F. Eble, personal communication, 1992).

The clay apparently occurs here in a broad low formed as a result of cave collapse and can be divided into two parts by a rooted clayey sandstone. The lower part of the clay is 0.9 to 1.7 m (3 to 5.5 ft) thick and consists of white clayey sand to sandy clay which is rooted and contains medium to large blocky peds. This is probably the argillic B horizon of a soil developed on sandy Paragon shales. The overlying 0.9 to 1.5 m (3 to 5 ft) consists of dark-gray mudstone with fine blocky to platy peds and root casts replaced by siderite. Although this mudstone and the overlying coal could represent the A horizon of a soil encompassing the entire clay deposit, this mudstone does not exhibit the kind of alterations typical of a good A horizon, nor is it a typical Paragon mudstone. In fact, it is more typical of a Breathitt shale. This means that the clay bed here could reflect two Early to Middle Pennsylvanian episodes of soil formation -- one on top of the exposed Paragon, following Early Pennsylvanian uplift and exposure and another on top of a subsequently deposited Breathitt shale. On the other hand, it could also represent a single soil developed on a combined Breathitt-Paragon parent unit. Whatever the origin, the absence of a well-developed flint clay suggests that humid pedogenesis was not as effective here as at **STOP 17**. Perhaps this was related to paleotopography and a decreased amount of uplift on the southern downdropped block (fig. 51).

Coastal and terrestrial environments of the lower Breathitt and Lee Formations (lower Middle Pennsylvanian), near Frenchburg, Kentucky

Stop leaders: Stephen F. Greb, Donald R. Chesnut, Jr, and Cortland Eble

Lower Breathitt deposition was interrupted by deposition of the Lee Formation. The Corbin Sandstone Member of the Lee is interpreted here as a fluvial deposit based on unimodal crossbed and ripple orientations, lateral accretion surfaces, and overall upward-thinning bedding and upward-fining grain sizes. Deposition occurred in bedload channels that were part of a broad trunk-channel system. At Frenchburg, the Corbin is overlain by a rooted horizon and the Betsie Shale Member of the Breathitt Formation, which has been interpreted as a marine zone that can be traced across the coal field.

Introduction

Pennsylvanian strata along the western outcrop of the Central Appalachian Basin are mapped as parts of the Lee and Breathitt Formations. This part of the Breathitt Formation commonly contains thin (grained, sheet-form siltstones and sandstones, and (1 to 15 m, 3 to 48 ft) dark, carbonaceous shales. The Lee is generally characterized by thick (8 to 70 m, 26 to 225 ft) conglomeratic, crossbedded, quartz-rich sandstones, although thinner sandstones and shale may also occur. Along KY Route 460 southeast of Frenchburg, in the Scranton 7.5 minute quadrangle both conglomeratic sandstones of the Lee Formation and sandstones and shales of the lower Breathitt Formation are well-exposed in a continuous, large outcrop (fig. 59).

Unconformity

Pennsylvanian strata in figure 60 are juxtaposed above the Mill Knob Member of the Mississippian Slade Formation (originally mapped as the Newman Limestone Forma

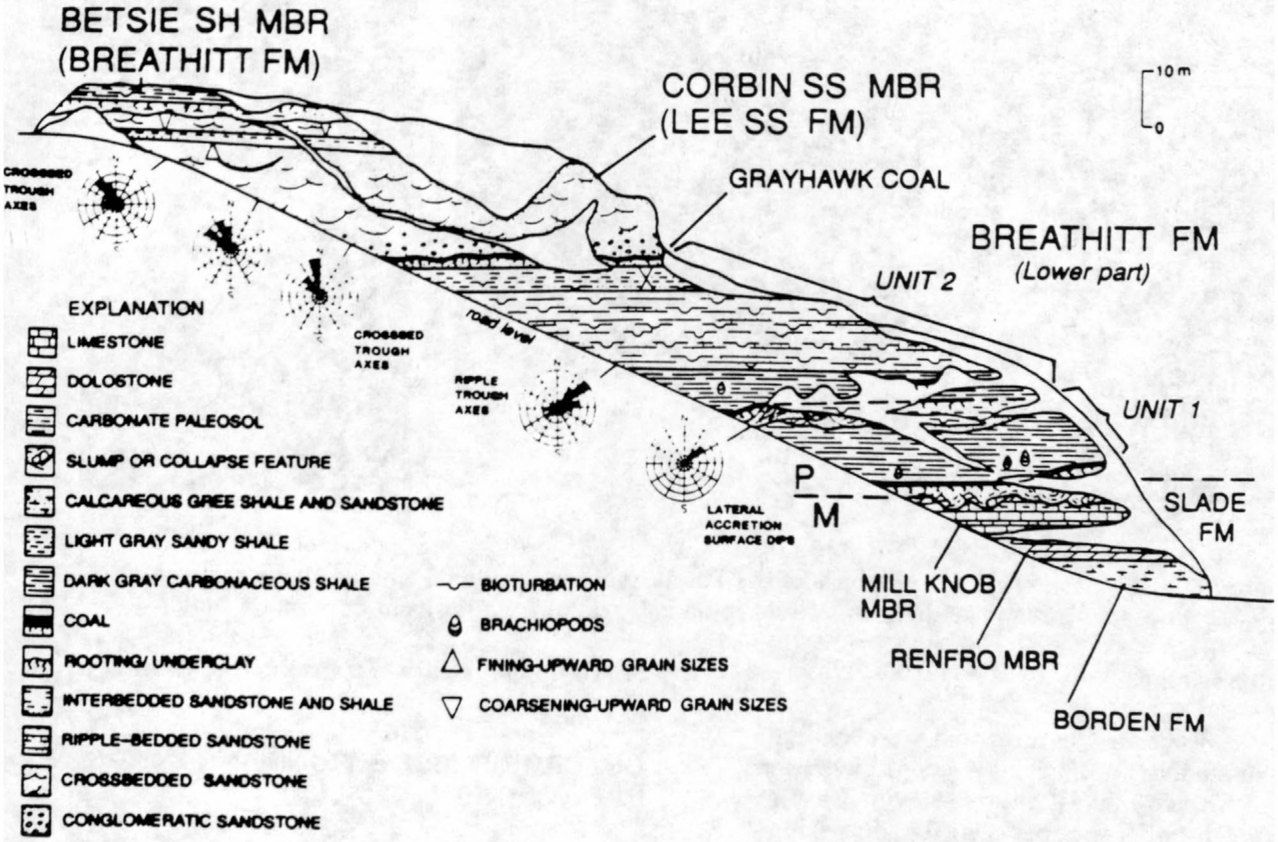


Figure 59 - Diagram of roadcut along U.S. Route 460 near Frenchburg, KY.

tion and the upper part of the Borden Formation ie., Hoge, 1977). Locally, Pennsylvanian strata may truncate all of the Slade Formation, down to the upper portion of the Borden (Hoge, 1977).

The unconformity itself is characterized by large blocks of rotated dolostone bound by listric glide planes, small channels, and leached paleosols (fig. 60a). Rotated blocks are overlain and interbedded with gray, sandy shales and carbonate breccias, which in turn are truncated by small channels (>5 m, 16 ft, wide and 1 m, 3.2, ft deep) filled with gray mudstone. The 3 to 4 m (10 to 13 ft) of disturbed strata are overlain by an undisturbed underclay and thin coal seam with thin shaly partings. The coal is arboreous lycopsid dominant, lower Middle Pennsylvanian in age, with 24.7 % ash and 1.0 % sulfur (fig. 61a).

Interpretation

The blocks represent karstic collapse features developed during exposure of the Slade Formation carbonates. Dolomization of the carbonate blocks preserved in the collapse structures may also be the result of exposure. The thick underclay and coal above the unconformity indicate lycopod colonization of the area and deposition of peat in a fresh-

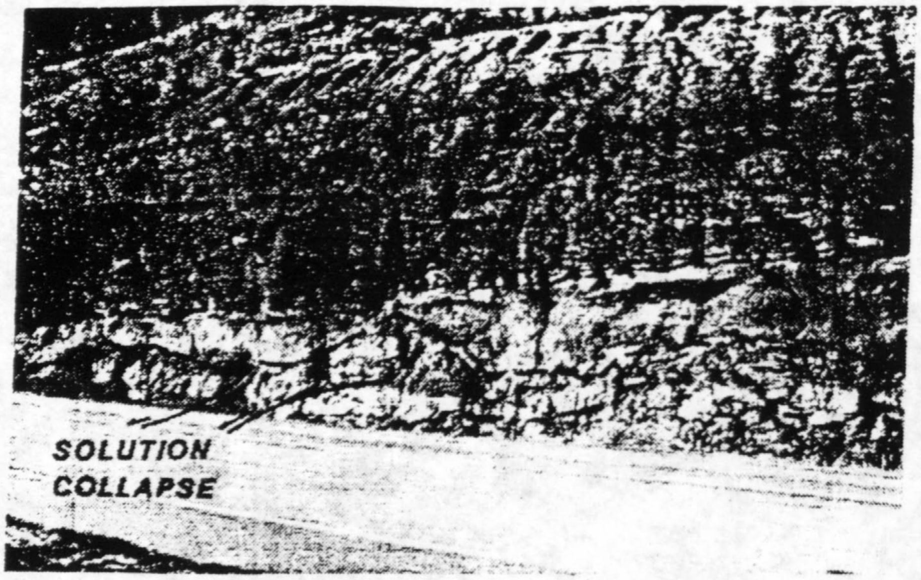


Figure 60 - a) unconformity at the base of the Pennsylvanian with dark shales above and possible collapse features in underlying Mississippian strata. b) *Lingula* from dark shale unit.

water swamp.

Exposure features and truncated Mississippian strata (missing section) have been used by many investigators as evidence of a major unconformity between the Mississippian and Pennsylvanian systems (Dever, 1971; Rice and others, 1979; Ettensohn, 1980). However, regional cross sections of the unconformity show that the lowest Pennsylvanian strata in the deepest part of the Central Appalachian Basin are progressively truncated toward the western margin of the basin, so that the unconformity is actually a Lower Pennsylvanian phenomenon rather than a systemic boundary (Chesnut, 1988).

Analysis of spores in the coal at this stop indicate that the base of the Pennsylvanian clastic sequence is lower Middle Pennsylvanian age. That age, in conjunction with the position of the collapse features in the Mill Knob Member of the Slade Formation, indicate that all of the Lower Pennsylvanian section and part of the Upper Mississippian section are missing at this stop. If the upper portions of the Slade and Paragon Formations were originally deposited in this area there are roughly 60 m (193 ft) of missing Mississippian strata (Ettensohn and others, 1984).

OUTCROP DESCRIPTIONS AND INTERPRETATIONS

Breathitt Formation (lower part)

The lower part of the Breathitt Formation (below the Corbin Sandstone) consists of dark shale, siltstones, thin, laterally discontinuous sandstones, thicker more regionally continuous sandstones, and coals. At Frenchburg, the lower part of the Breathitt can be divided into 2 units (fig. 59) which are described and interpreted in the following section.

Unit 1 Description

The coal at the base of the lower tongue of the Breathitt Formation is 14 cm thick and has abundant shale partings. The coal is overlain by 12 m (39 ft) of dark, carbonaceous shale (fig. 61a) containing disseminated plant material, siderite nodules, rare *Planolites* trace fossils, and *Lingula* brachio

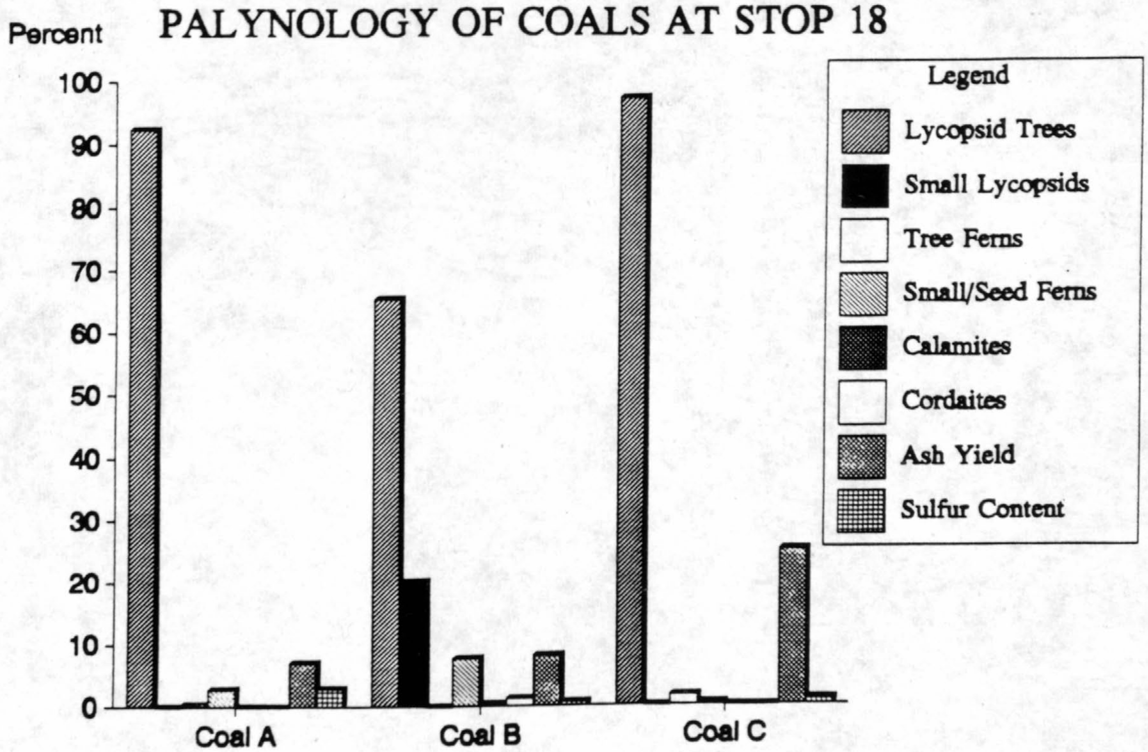


Figure 61 - Analyses of coal beds from the lower part of the Breathitt Formation. a) unnamed coal at base of Pennsylvanian, a) unnamed coal capping unit 1, and c) Grayhawk coal capping unit 3.

pods (fig. 60b). The shale contains only 72 % ash. Presumably the remaining percentage is mostly organic materials, which is why the shale is dark in color. The *Lingula* brachiopods within the shale exhibit color banding that may be preserved remnants of either the original color pattern of the shells or a reflection of the internal organs of the brachiopods.

The shale is sharply overlain by a 3 m (10 ft) interval of heterolithic strata including scour-based, fine-grained sandstone exhibiting lateral accretion surfaces, slumps, rhythmically laminated fine-grained sandstone and siltstone, and sandy gray shale with sideritic laminations (fig. 59). The upper, weathered surface of the sandstone contains small trace fossils of *Olivellites, Lockeia?*, and Curvolithus?. The sandstone and heterolithic strata are overlain by a thin (48 cm, 19 in) coal, that thins and becomes shaly (rashy) toward the north (fig. 59). This coal is also lower Middle Pennsylvanian in age, arboreous lycopsid dominant, and contains 8.0 % ash, and 0.6 % sulfur (fig. 61b).

Interpretation

The lower coal represents peat accumulation above the exposed Mississippian carbonates. Presumably the peat required fresh pore water for plants to grow, so a fresh-water swamp is interpreted. However, *Lingula*-bearing organic shales above the coal represent at least restricted marine-water inundation of the swamp and a probable coastal setting. Likewise, the *Lingula* and trace fossils in the overlying shale indicate brackish- to marine- water deposition follow

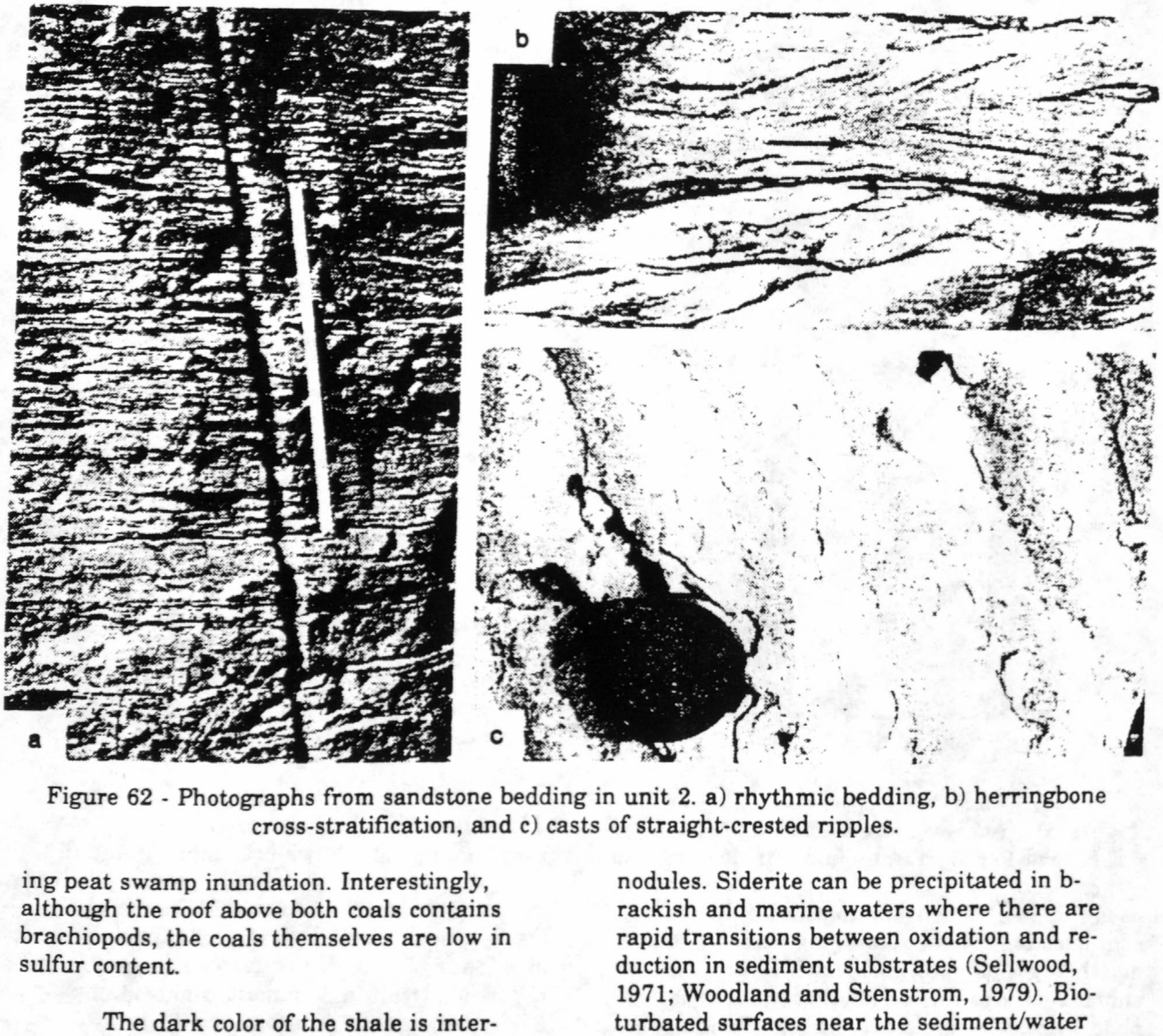


Figure 62 - Photographs from sandstone bedding in unit 2. a) rhythmic bedding, b) herringbone cross-stratification, and c) casts of straight-crested ripples.

ing peat swamp inundation. Interestingly, although the roof above both coals contains brachiopods, the coals themselves are low in sulfur content.

The dark color of the shale is interpreted to represent reducing conditions during deposition (Rhoads and Morse, 1971; Ekdale and Mason, 1988). The low abundance and diversity of fauna supports an interpretation of restricted conditions. Reducing conditions were probably set up by high concentrations of carbonaceous material eroded from the coals during marine inundation, or washed into the bays and estuaries from fringing swamps. Similar shales have been interpreted as the deposits of marine and brackish-water bays in other parts of the basin (Horne and Ferm, 1978; Chesnut, 1981, 1991; Martino, 1989).

The shales contain common siderite nodules. Siderite can be precipitated in brackish and marine waters where there are rapid transitions between oxidation and reduction in sediment substrates (Sellwood, 1971; Woodland and Stenstrom, 1979). Bioturbated surfaces near the sediment/water interface can act as sites for concentrating iron oxides during periods of slow deposition. When the iron-rich substrate is buried it becomes reducing, which causes iron to react with carbonate and form siderite (Sellwood, 1971). There is significant evidence of bioturbation in the dark shales at this stop.

The shales are overlain by very fine-grained sandstones with thin (0.75 m, 2.4 ft) lateral accretion surfaces that dip to the southwest (222° to 235°). The accretion surfaces and slumps probably represent small meandering channels. Trace fossils and rhy

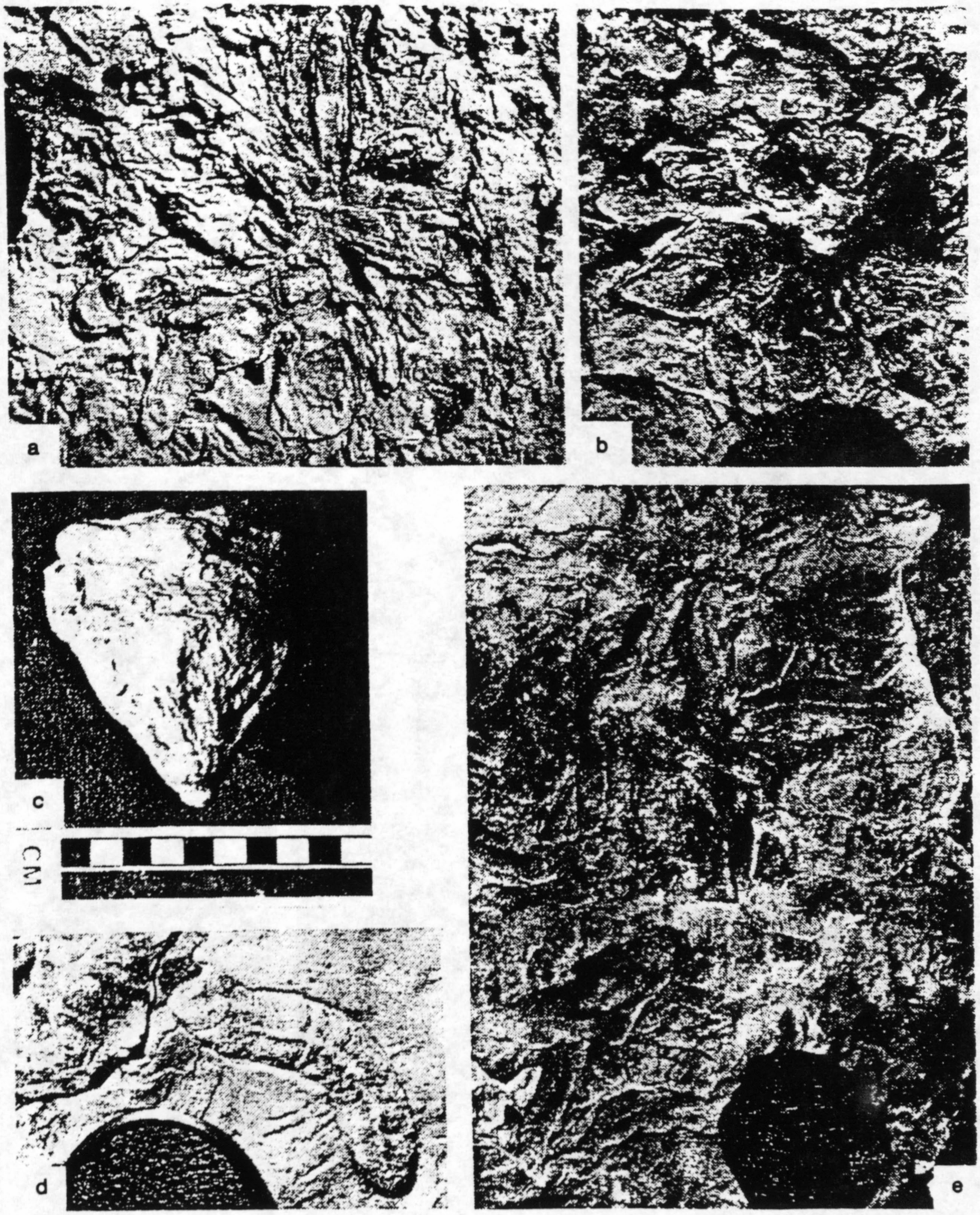


Figure 63) Trace fossils from the sandstone in unit 2: A - *Asterosoma* showing radial pattern, B - *Conostichus*, C - *Olivellites* showing characteristic medial ridge and chevron pattern, and D - an *Olivellites* pavement with abundant burrows.

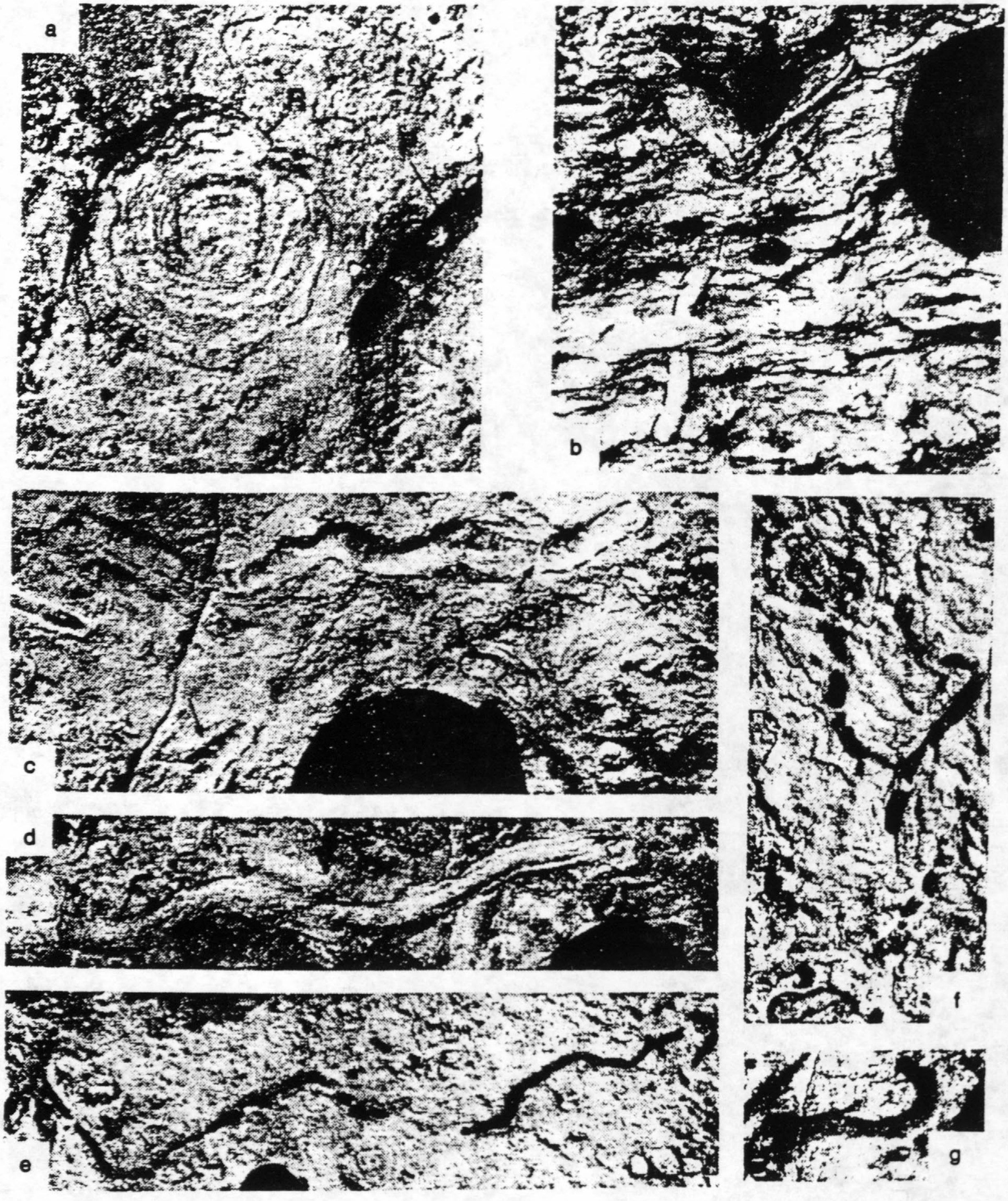


Figure 64) Trace fossils from the sandstone in unit 2: A - top surface of *Rosselia* with concentric internal structure, and a horizontal *Planolites* tubular burrow, B - vertical section through a *Rosselia* or *Asterosoma* cup, C - *Scalarituba*, D - *Scolicia* showing characteristic terminal ridges, and E through G - unidentified trace fossils.

Figure 65 - Photograph of Grayhawk coal bed and Corbin Sandstone,

thmic bedding within some of the channels are evidence for possible tidal creek deposition. The tidal creek (?) deposits are overlain by a second coal indicating a return to freshwater conditions and peat accumulation. The overall sequence of dark shale with restricted marine fossils, overlain by sandstone or heterolithic strata with tidal structures, overlain by coal is common in the lower Breathitt Formation. This type of regressive sequence could be deposited as a bay-, lagoon- or estuarine-fill deposit.

Unit 2 Description

The coal that caps the first regressive sequence (described above) is directly overlain by a second dark, carbonaceous shale (fig. 59). This shale is similar to the first dark shale and contains orbiculoid brachiopods and tubular siderite nodules. As with the first shale, the second shale is also overlain by a ripple-bedded sandstone containing trace fossils, but the second sand is 14 m (45 ft) thick. The sandstone is not mapped in the area (Hoge, 1977) although thick, laterally extensive, ripple-bedded sandstones are common at this stratigraphic position in other parts of the basin.

Bedding appears rhythmic, although not cyclically rhythmic (fig. 62a). Bedding consists of 5 to 15 cm (12 to 38 in) thick, ripple-stratified sandstones separated by 10 to 25 cm (25 to 64 in) of thinner, more shale-rich ripple beds and laminae. Individual ripples are usually less than 5 cm (13 in) thick and often exhibit clay drapes. Asymmetric ripples are abundant, although many of the ripples observed in section exhibited rounded crests. Herringbone patterns are common (fig. 62b). Measurements of the dips of ripple trough axes indicate a crudely bimodal trend to the northwest and southeast (fig. 59). Float material indicates that current ripples most commonly occur as rib-and-furrow structures although casts of symmetric, low-amplitude ripples also occur (fig. 62c).

The unit 2 sandstone is extremely bioturbated. Several benches provide exceptional exposures of abundantly burrowed horizons in the sandstone. Traces of *Asterosoma* (figs. 63a-b), *Conostichus* (fig. 63c), *Olivellites* (figs. 63d-e), *Planolites*, (fig. 64a), *Rosselia* (figs. 64a-b), *Scalarituba* (fig. 64c), *Scolicia* (fig. 64d), and an unidentified smooth horizontal trace that branches, exhibits sharp bends, and may break up into disconnected beads (figs. 64e-g) are abundant. Other traces such as *Curvolithos*, and *Teichichnus* are uncommon. Individual traces can be traced along bedding for as much as a meter. There are widespread *Olivellites-Scolicia* pavements on the first bench. Also, the *Conostichus* traces were most common in the upper part of sandstone. The *Conostichus* traces were also some of the largest recorded in this part of the basin.

The unit 2 sandstone fines upward into a thin shale that coarsens upward into a thin (1.75 m, 5.6 ft) fine-grained ripple-bedded sandstone (fig. 59). The lower third of the shale contains similar traces to the unit 2 sandstone. Trace fossils were not noted in the thin, capping sandstone. This sandstone exhibited ripples with rounded crests and wavy bounding surfaces, as well as rooting structures. The sandstone is capped by the Grayhawk coal (figs. 59, 65). The Grayhawk coal is 38 to 43 cm (97 to 109 in) thick, lycopod dominant, with 8.2 % ash, and 0.9 % sulfur (fig. 62c).

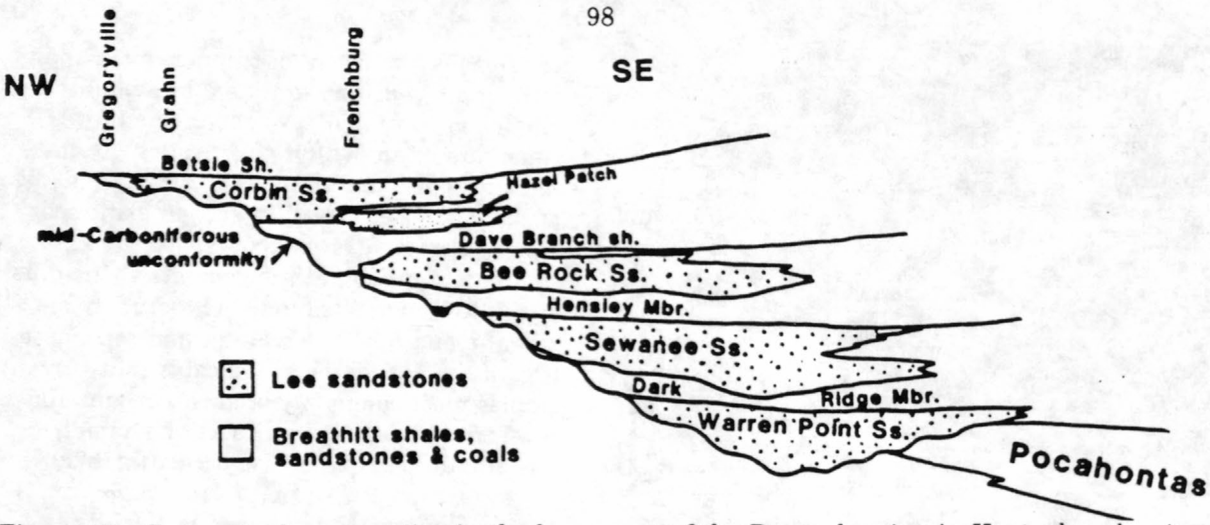


Figure 66 - Generalized cross-section in the lower part of the Pennsylvanian in Kentucky, showing the on-lap of the Lee Sandstone belts (from Chesnut, 1988).

Interpretation

The dark shale at the base of unit 2 is similar to the shale described for unit 1 and represents a return to coastal marine deposition. The common tubular siderite nodules probably represent burrow fills. There may be several meters of relief on the contact at the base of the shale, so that the transgression that inundated unit 1 may have been accompanied by erosion. Peat compaction of the coal at the top of unit 1 may also have helped to create relief on the contact.

The dark shale is sharply overlain by a thick, ripple- dominated sandstone. Herringbone cross-stratification, rhythmic bedding, bimodal paleocurrent directions, and abundant trace fossils indicate that the sandstone was a tidal deposit (DeRaaf and Boersma, 1971; Reineck and Singh, 1971; Klein, 1977; Terwindt, 1981). The relatively horizontal bedding surfaces suggest deposition on a broad tidal flat or shoal. The rhythmic thickness alterations in bedding are presumed to represent tidal deposition. Rhythmites are not cyclically bedded and many contain evidence of reworking such as bioturbation and rounded ripple tops, therefore individual decimeter scale form-sets may represent amalgamated tidal cycles of monthly, annual, or greater duration. The large abundance and diversity of both horizontal and vertical burrows suggests marine conditions. The lack of exposure surfaces, and, on the other extreme, crossbeds or thicker bedding, may indicate shallow, subtidal conditions. Wave reworking of tidal current ripples was common.

The sand is overlain by a thin shale and sandstone. Marine bioturbation decreases upward and the thin capping sandstone is root penetrated. Wave ripples and wavy bedding in the thin sheet sand capping the shale may indicate upward shallowing and near-shore wave action. The upward decrease in bioturbation records the change from near-shore tidal flats and shoals to exposed fresh-water swamp (Grayhawk coal). Again, the stacking of quiet-water shales and tidal sand bodies capped by coals (fig. 59), is a typical lithologic sequence for the lower Breathitt Formation.

Lee Formation

Corbin Sandstone Member Description

The Grayhawk coal is overlain by 25 m (80 ft) of fine-to-coarse grained, crossbedded, sublitharenite to quartzarenite of the Corbin Sandstone Member of the Lee Formation (figs. 59, 65). The lower third of the

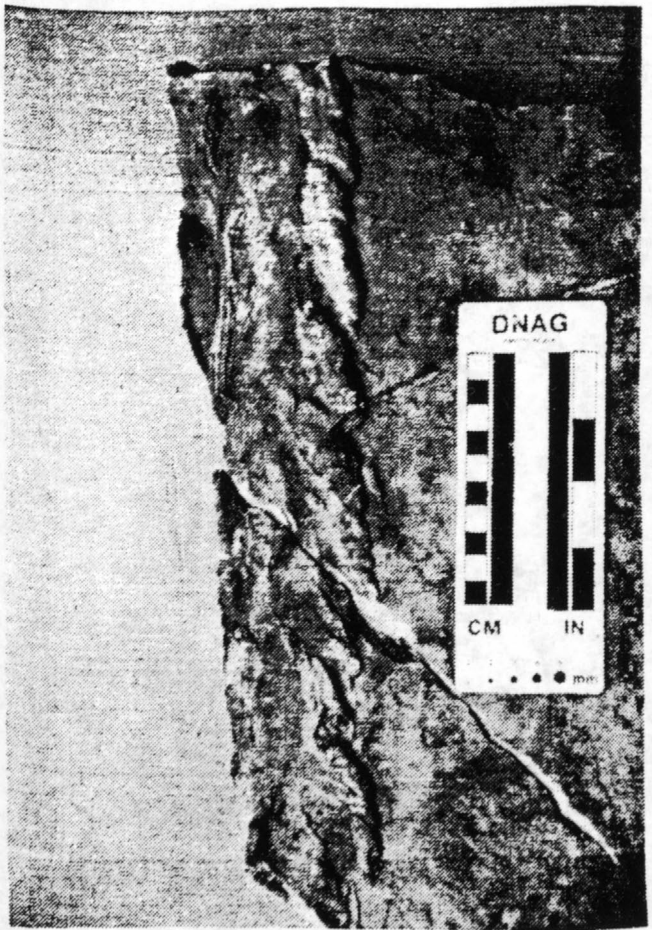


Figure 67 - Photograph of a large, unidentified trace fossil from a sandstone inter-bed in the Betsie Shale.

sandstone is conglomeratic with common quartz pebbles. Overall, bedding thins upward with large crossbeds up to 2.7 m (8.7) thick in the lower part of the sand and thinner cross strata in the middle and upper portion of the sand. The upper 4 m (13 ft) of the Corbin Sandstone grade into ripple-bedded sandstone and sandy shale (fig. 59).

Crossbed orientations throughout the Corbin are unimodally to the northwest (300° to 360°). Crossbeds in the lower half of the sand may exhibit subcritical (low angle) climb up the backs of the preceding crossbed. The sloping backs of crossbeds dip to the southeast and can be misinterpreted as a bedform that migrated in an opposing direction to the overlying crossbed, but on closer examination foresets within the crossbed and trough orientations illustrate the true orientations of the original bedforms. Low-angle bedding surfaces in the middle of the sandstone also dip to the northwest and may represent lateral accretion surfaces.

Interpretation

The Corbin is part of a series of broad Lee sandstone belts that onlap the basin margin (fig. 66). At Frenchburg, the Corbin is the only Lee sand preserved, but to the south and east several Lee sandstones occur. These sandstones have been interpreted as marginal marine to marine deposits (Mitchum, 1954; Ferm and others, 1971; Horne and others, 1974; Cecil and Englund, 1989), and as fluvial deposits (Potter and Siever, 1956; Bement, 1976; Rice, 1984; Chesnut, 1988; Rice and Schwietering, 1988).

The exposures around Frenchburg exhibit unimodal crossbed and ripple orientations, are conglomeratic near the base and generally fine upward, exhibit upward thinning bedding, and contain possible lateral accretion surfaces all typical of fluvial deposits (Reineck and Singh, 1971; Collinson, 1978; Cant, 1982; Walker and Cant, 1984). Fluvial interpretations of the Corbin Sandstone have also been made for nearby exposures of the Corbin by Hester and Taylor (1977).

Breathitt Formation

Betsie Shale Member Description

Along the western outcrop margin of the Eastern Kentucky Coal Field the Lily/Manchester coal overlies the Corbin, but here only a thin rooting horizon is preserved (fig. 59). The rooting horizon is overlain by dark shales, siltstones and ripple-bedded sandstones of the Betsie Shale Member of the Breathitt Formation. Like the shales of the Lower Breathitt Formation the Betsie contains rare trace fossils including *Bifasciculus?, Chondrites* and *Planolites*. A large walking trace of an arthropod? was also found in an interbedded sandstone (fig. 67).

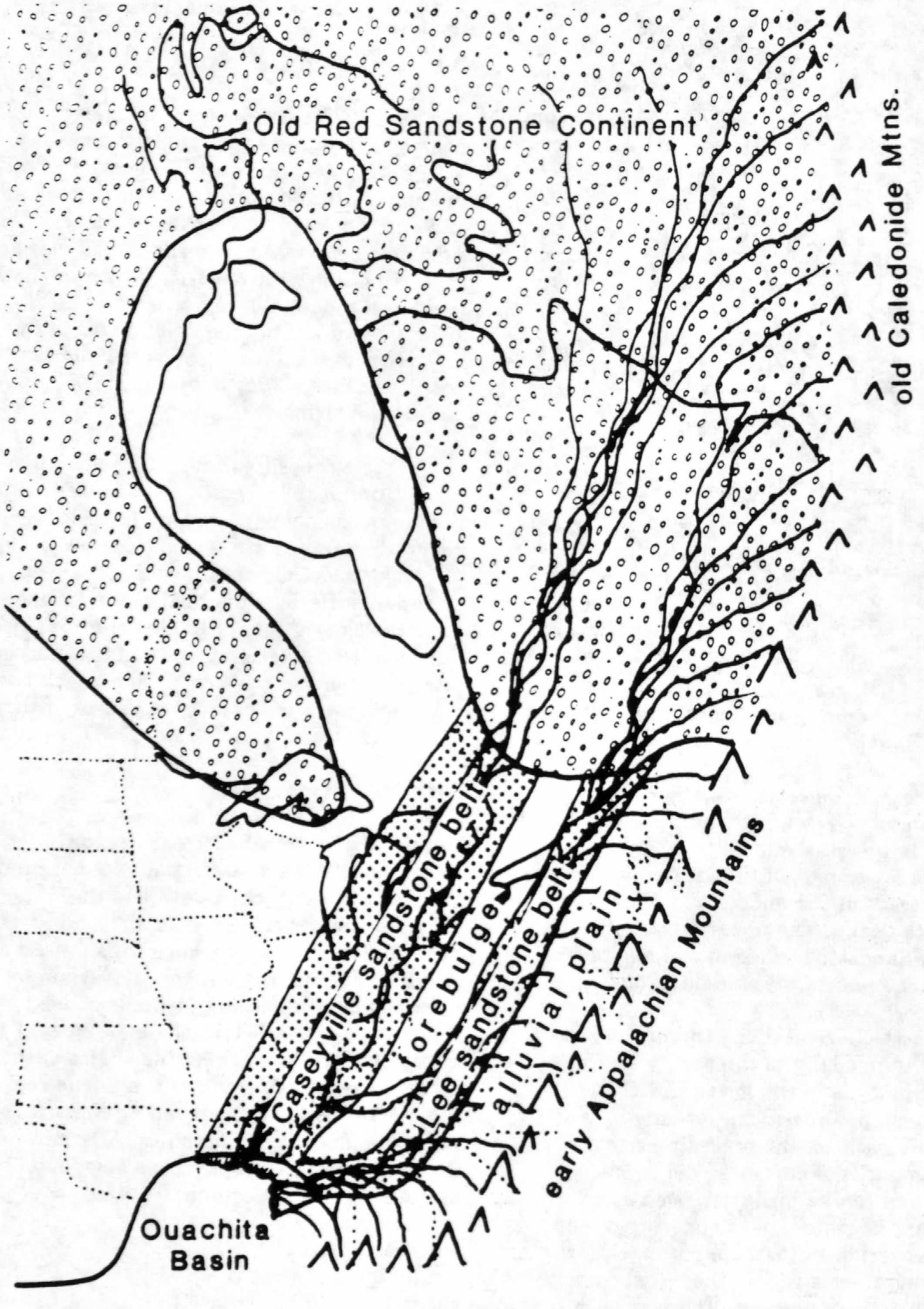


Figure 68 - Interpretive diagram of the eastern United States during Lee sand belt deposition.

The coarsening-upward Betsie Shale is conspicuous on downhole logs because of relatively high radioactivity in the basal part of the shale. The shale has been correlated as a continuous marine horizon across much of the Central Appalachian Basin (Rice, 1987).

Interpretation

The rooted zone at the top of the Corbin is interpreted as exposure and plant colonization of the abandoned fluvial stream system. The Betsie Shale directly overlies the Corbin Sandstone and indicates another marine inundation of the area. The fact that the shale occurs across much of the coal field indicates widespread shallow marine conditions following Corbin deposition. The juxtaposition of the Betsie Shale above the Corbin Sandstone can be interpreted as deposition during relative sea-level rise. In this model, the scour at the base of the Corbin trunk valley system was formed during a sea-level low stand (fig. 68). Tectonic activity in the Appalachians supplied quartz-pebbles and coarse-grained detritus that was delivered to the basin by a large fluvial sand belt. The Corbin sandstone was deposited during a sea-level rise in which channels back-filled with fluvial deposits. In the broad sand belt, some areas were filled with sediment and were colonized by vegetation and became rooted, as at Frenchburg. In other areas, the upper part of the Corbin may have been reworked by tidal processes. The Betsie Shale was then deposited during marine inundation. Subsequent regression led to the coarsening-upward character of the shale and a return to normal Breathitt deposition.

END OF TRIP

STRATIGRAPHY OF THE LOWER AND MIDDLE PENNSYLVANIAN SERIES IN WEST VIRGINIA

Bascombe M. Blake
West Virginia Geological & Economic Survey

Introduction

The coal fields of West Virginia have been traditionally divided into a northern, high-sulfur field and a southern, low-sulfur field based on coal quality, rank, and age. The northern coal field, also known as the Dunkard Basin, primarily contains uppermost Middle and Upper Pennsylvanian strata. The stratigrahically older southern coal field, also known as the Pocahontas Basin, contains mainly Lower and Middle Pennsylvanian strata (fig. 79). Subsidence in the Appalachian foreland basin associated with orogeny to the southeast provided accommodation space for the accumulation of a northwest-thinning clastic wedge that progressively onlapped north-northwestward onto the North American craton during the Early and Middle Pennsylvanian. Thickness trends of Lower and Middle Pennsylvanian sediments suggest that only the distal portion of the original clastic wedge is preserved and the proximal, thicker portion of the foreland basin fill has been removed by subsequent erosion.

The subdivisions of the Lower and Middle Pennsylvanian series in the Appalachian basin were broadly established by the mid-1800's (see Rogers, 1858) based mainly on lithologic characteristics, and were later modified by various workers (for summary of early work in this area, see Stevenson, 1904). David White (1895, 1900a, 1900b) established broad biostratigraphic correlations based on megafloras collected from across the Appalachian Basin. White (1900b) broadly correlated Lower Pennsylvanian strata from across the United States, including the Lower Pennsylvanian New River and Pocahontas Formations in southern West Virginia, with the type "Pottsville" area in the eastern Pennsylvanian anthracite fields. He (1900a) also was able to demonstrate that the flora from the Kanawha Formation (Middle Pennsylvanian) was intermediate in composition between a younger flora from the Allegheny Formation (Middle Pennsylvanian) in western Pennsylvanian and an older Early Pennsylvanian "Pottsville" flora. White (1900a) suggested that the flora of the Kanawha Formation below the Cedar Grove coal bed in the valley of the "Great Kanawha" (fig. 4, locations 3 and 4) was closely related to, and clearly derived from, the older "Pottsville" flora, whereas the flora from higher in the Kanawha Formation was more closely-allied with the younger Allegheny flora. White broadly equated the "Block" coals of the southern coalfield with the Kittanning coal beds of the northern coalfield, and suggested that a flora collected from 61 to 91 m (200 to 300 ft) above the Kanawha Black Flint of White (1891) was not stratigraphically far-removed from the Freeport coal beds of northern West Virginia and western Pennsylvania.

These broad correlations have subsequently been used to correlate Lower and Middle Pennsylvanian strata across the Appalachian Basin. Read and Mamay (1964) established 13 megafloral zones for Upper Paleozoic strata in North America, subdividing the Lower Pennsylvanian Series into three (3) zones and the Middle Pennsylvanian Series into four (4) zones. Gillespie and Pfefferkorn (1979) studied megafloras from 256 localities associated with the proposed Pennsylvanian System stratotype (Englund et al., 1979) and published the range zones for biostratigraphically important taxa. These data permitted correlation of North American Upper Carboniferous strata with time-equivalent European strata, further demonstrating the cosmopolitan nature of the Euramerican flora (Pfefferkorn and Gillespie, 1980).

Mississippian - Pennsylvanian systemic boundary

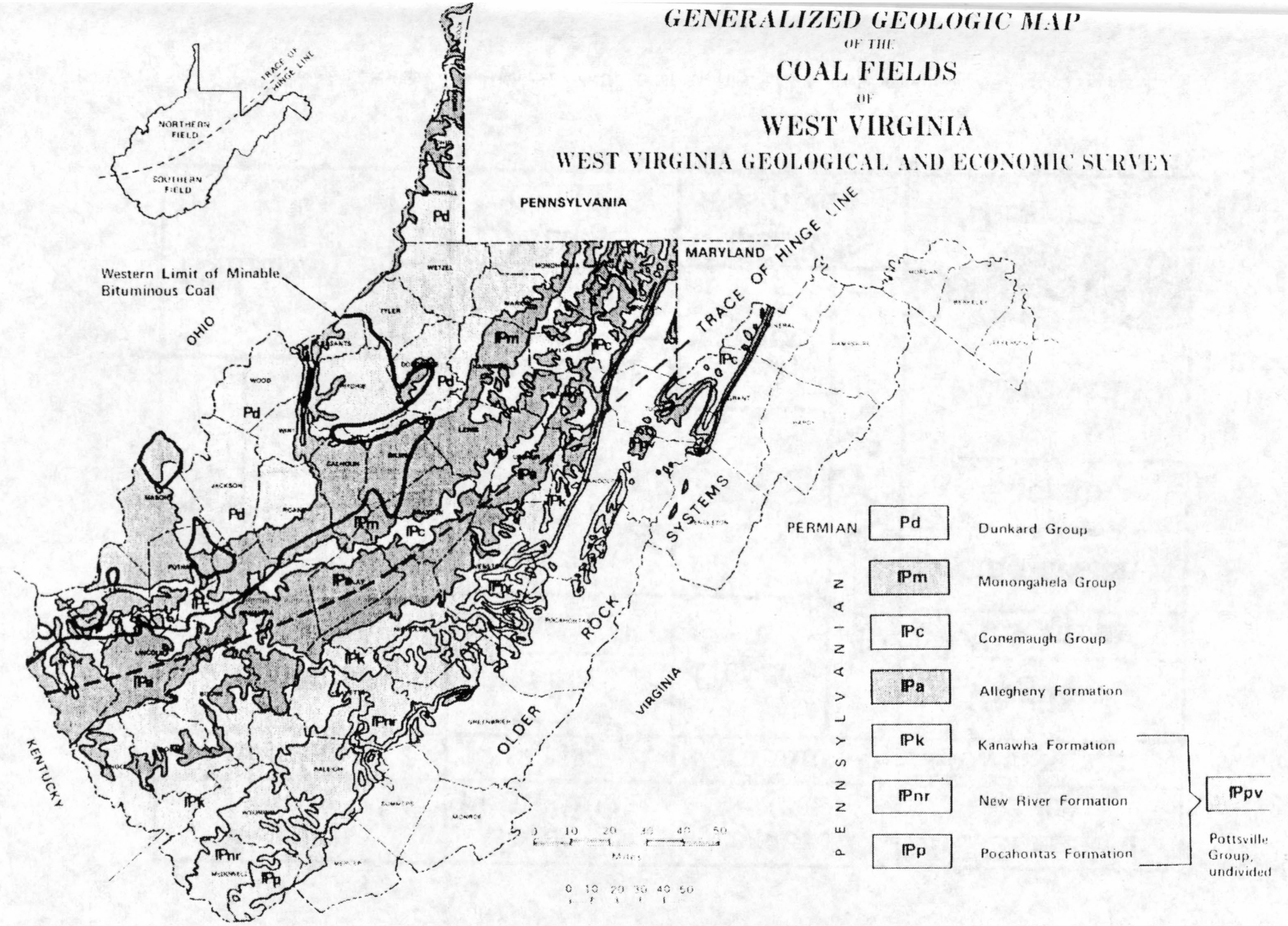


Figure 79) Generalized geologic map of the coal fields of West Virginia.

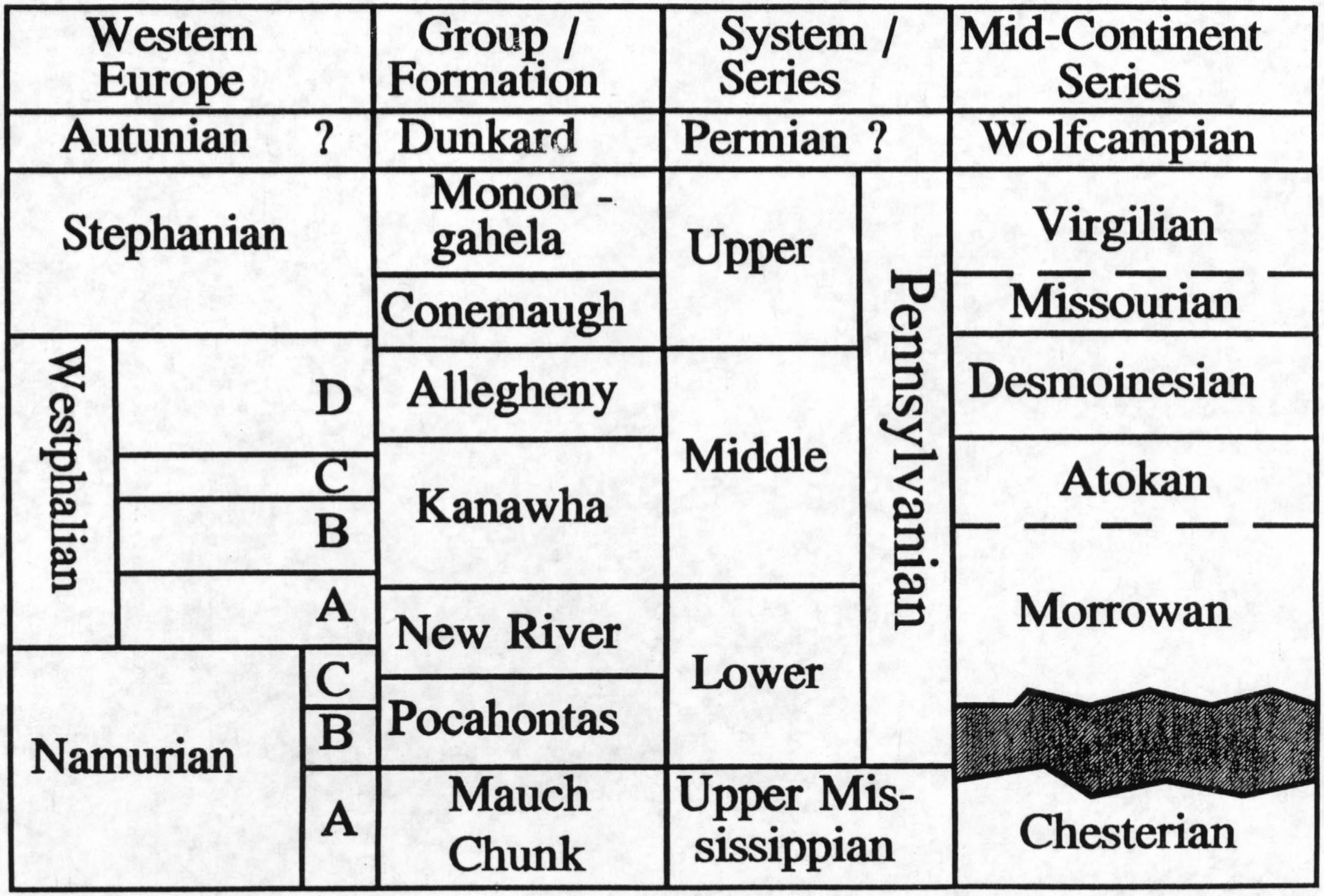


Figure 80) Reference correlation chart for Upper Mississippian and Pennsylvanian strata.

In extreme southeastern West Virginia and adjacent portions of Virginia the Lower Pennsylvanian Pocahontas Formation overlies the Upper Mississippian Bluestone Formation. In this area deposition continued uninterrupted across the Mississippian-Pennsylvanian systemic boundary (Englund, 1979). North and west from this area (e.g. **STOPS 1, 16, 17** and **18**) Earliest Pennsylvanian sediments are absent due to the presence of an Early Pennsylvanian unconformity, commonly referred to as the Mississippian-Pennsylvanian unconformity (White, 1891). The amount of time represented by the unconformity increases progressively to the north and west until in northern West Virginia lower Middle Pennsylvanian Kanawha Formation strata rests directly on Upper Mississippian Mauch Chunk Group strata (**STOP 1**).

The unconformity has also been described in the subsurface n northwestern West Virginia where Lower(?) Pennsylvanian quartz arenites fill a paleochannel cut into and locally through the Upper Mississippian Greenbrier Limestone and into the Lower Mississippian Pocono (Price) Group (Flowers, 1956; Rice, 1984, 1985; Filer, 1985; Sweeney, 1986; Rice and Schweitering, 1988). This inferred paleovalley was incised more than 91 m (300 ft) below the regional level of the Early Pennsylvanian unconformity in Wood, Wirt, and Roane counties (Filer, 1985). Saunders and Ramsbottom (1986) have suggested that the unconformity is possibly of global extent and was caused by a large drop in sea level that coincided with the onset of Gondwanaland glaciation.

Lower Pennsylvanian Series

The Lower Pennsylvanian Series in West Virginia is divided into two formations, in ascending order, the Pocahontas and New River Formations. The presence of *Neuralethopteris (Neuropteris) pocahontas* D. White fronds, a Lower Pennsylvanian index species, in the red member of the Bluestone Formation mandates placement of these beds in the Pennsylvanian System (Gillespie and Pfefferkorn, 1979). The boundary between the Bluestone and Pocahontas formations corresponds to the Namurian A-B boundary of western Europe and Chesterian-Morrowan boundary of mid-continent North America. This equating of the Namurian A-B boundary and the Mississippian-Pennsylvanian boundary demonstrates that the Lower - Upper Carboniferous boundary of European usage does not correspond exactly with the Mississippian - Pennsylvanian boundary in North America (fig. 80).

Pocahontas Formation

The Lower Pennsylvanian Pocahontas Formation (**STOP 11**) attains maximum thickness of approximately 215 m (700 ft) (Englund and Thomas, 1990) in southeastern West Virginia and southwestern Virginia, but pinches out north and northwest in approximately 49 km (30 mi). The formation contact with the underlying Bluestone Formation is placed at the base of the Lower Sandstone Member, except in areas where the lower member intertongues with beds of the marine Bramwell Member (the top member of the Bluestone Formation (Englund et al., 1979). In these areas the formation contact and Mississippian-Pennsylvanian systemic boundary are not coeval (fig. 80). The formation contact with the overlying New River Formation, in areas of uninterrupted sedimentation across the formation boundary, is arbitrarily placed at the base of the Pocahontas No. 8 coal bed (Englund et al., 1979). To the west and north the top of the formation is progressively truncated by the regional Early Pennsylvanian unconformity, marked by the base of the Pineville Sandstone Member of the overlying New River Formation (Englund et al., 1979).

Economically important coal beds of the Pocahontas Formation are, in ascending order, the Pocahontas No. 2 (**STOP 11**), No. 3, No. 4, and No. 6 coal beds (fig. 81). Pocahontas Formation coal beds are low to medium volatile bituminous in rank, low in sulfur content (median value 0.82%, n=75) and ash yield (median value 8.8%, n=75), and have

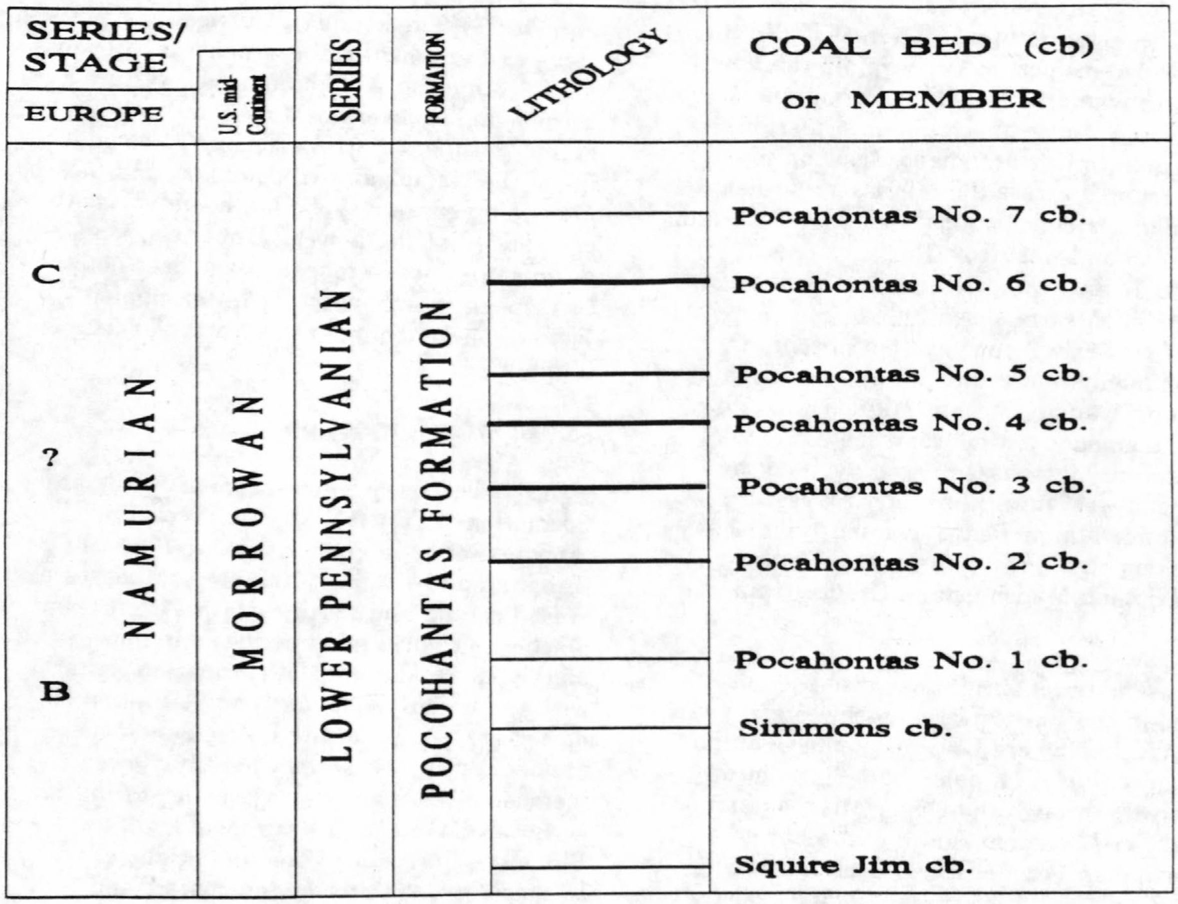


Figure 81 - Generalized stratigraphic column of the Pocahontas Formation in West Virginia showing the position of economically important coal beds.

been widely mined for metallurgical purposes. It has been suggested that the geometry and high quality of these coal beds is largely a result as having formed from domed ombrogenous peat deposits under a tropical, everwet climate (Cecil et al., 1985; Cecil and Englund, 1989; Cecil, 1990). Pocahontas coal beds have been extensively mined for over ninety years and resource depletion is becoming a concern.

The Pocahontas Formation consists primarily of what have been interpreted as fluvial-deltaic sandstones with lesser amounts of siltstone, shale, mudstone, and coal beds. It has been suggested that these strata were deposited in a series of coalescing delta lobes and associated coastal plain facies (Englund, 1974, 1979; Englund et al., 1986; Englund and Thomas, 1990). According to such autocyclic based models, periodic abandonment of the delta lobes eliminated clastic input and allowed peat deposits to accumulate on the more topographically positive, sand-rich lobes. Fine grained sediments accumulated in the low-lying interlobe areas under marginal marine to fresh water conditions; however, marine strata are unknown from the Pocahontas Formation (Englund et. al., 1986, p. 61). Englund (1974, 1979) and Englund and others (1986) also suggested that peat accumulated during still stands in sea level behind quartzose barriers formed from reworking of sediments and concentration of the resistant quartz grains by marine processes during hiatuses in sediment imput. The barriers isolated the peat deposits from nearby marine influence allowing the formation of low ash, low sulfur peat.

In contrast to autocyclic based depositional models of coal formation, Cecil and others (in press) have suggested that Recent sedimentation patterns in the central Sumatra basin, Republic of Indonesia, may help to explain the cyclic stratigraphy of the Lower and lower Middle Pennsylvanian System of the eastern United States. Modern influx of fluvial siliciclastic sediment to the epeiric seas of the Sunda shelf of Indonesia and Malaysia, including the Strait of Malacca, appears to be highly restricted by the rain forest cover within the present day everwet climate belt of equatorial Sumatra. As a result, much of the fluvial, estuarine, and marine environments appear to be sediment starved and, as a result, they are erosional or nondepositional except for marine reworking and localized deposition in marine environments. Contemporaneously, thick (>13 m, 42 ft), laterally extensive (>70,000 km^2), peat deposits are forming on poorly drained coastal lowlands of Sumatra. Therefore, Recent peat formation in this present-day environment is not coeval with aggrading fluvial siliciclastic systems, a situation that commonly is assumed in many depositional models of coal formation. If drier and more seasonal conditions prevail, as during the last glacial maximum, peat formation ceases and siliciclastic influx greatly increases. The stratigraphy of Pleistocene and Holocene sediments on the Sunda shelf, as well as strata in Lower and lower Middle Pennsylvanian Series, may be better explained by the allocyclic controls of climate and sea-level change on sediment flux rather than by depositional models that are based on autocyclic processes. In keeping with the objective of this trip, **STOPS 10** and **11** in the Pocahontas Formation were designed to evaluate evidence for allocyclic and autocyclic controls on sedimentation in an epeiric setting in a humid (everwet) tropical region.

New River Formation

The New River Formation is defined as extending upward from the base of the Pocahontas No. 8 coal bed to the base of the Lower Douglas(?) coal bed of Hennen and Teets (1919) in the proposed Pennsylvanian System stratotype area (Englund, 1979). Named for exposures in the New River Gorge in West Virginia, the New River Formation reaches a maximum preserved thickness of over 305 m (1000 ft) in its southern outcrop area and thins to the north and west. In areas where it conformably overlies the Pocahontas Formation, the New River Formation consists of lithic arenites with minor amounts of siltstone, shale, mudstone, and coal beds. Thick, elongate, quartz arenite bodies are often unconformable with underlying formations. Paleocurrent data from the lithic arenites indicate a general northwest transportation direction from low-grade metamorphic highlands located to the southeast. In contrast, the quartz arenites exhibit a south-southwest transport direction. The New River Formation thins rapidly to the north and northwest of the field trip route (fig. 4) until in northern West Virginia it is absent due to either non-deposition or erosion, and the lower portion of the Kanawha Formation is in contact with Upper Mississippian Mauch Chunk strata on Chestnut Ridge (**STOP 1**).

Economically important coal beds include, in ascending order, the Fire Creek, Beckley, and Sewell. There has also been significant production from the Little Fire Creek, Welch, Sewell "B" (Castle), and Iaeger ("Jawbone" of Virginia) coal beds (fig. 82). New River coal beds are generally medium volatile in rank, and are also low in sulfur content (median value 1.0 %, n=194) and ash yield (median value 7.9 %, n=194). Mining of New River coal beds began around 1870. Early in this century, New River coal was prized as being a smokeless fuel, an important characteristic for use in coal-burning naval vessels. Most modern production is for metallurgical purposes.

The origin of the quartz arenites of the New River Formation in West Virginia and the Lee Formation in Virginia and eastern Kentucky has been a subject of considerable controversy and reamins problematical. A barrier shoreline model (Horne et al., 1974) rejects the presence of an Early Pennsylvani

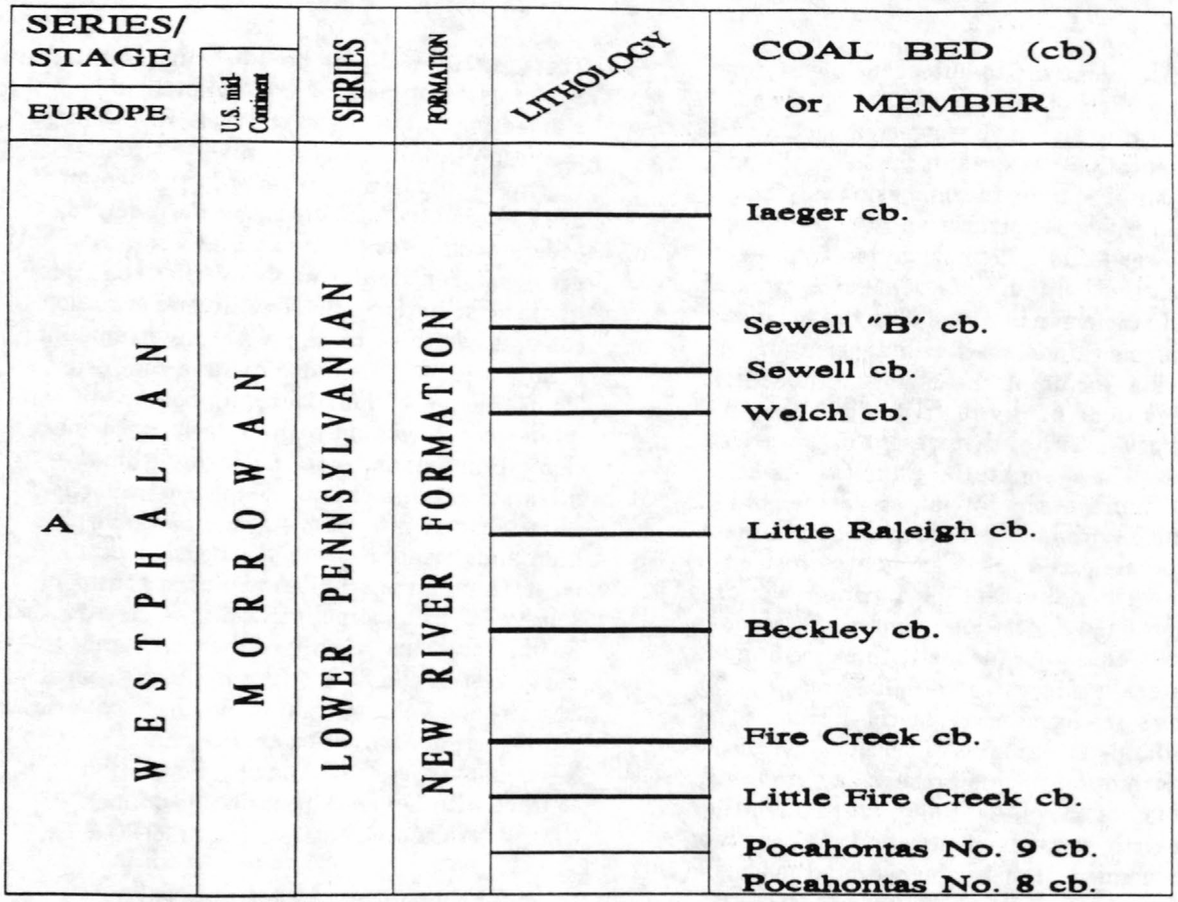


Figure 82 - Generalized stratigraphic column of the New River Formation in West Virginia showing the position of economically important coal beds.

an unconformity and states that the quartz arenites represent barrier-bar complexes that interfinger with deltaic, coal-bearing facies to the east and southeast (landward), and red and green marine shales of the Mauch Chunk Group to the north and west (seaward). According to this model the Greenbrier Limestone (Slade Formation in Kentucky) represents off-shore carbonate islands and shoals. The quartz arenites are interpreted as having resulted from winnowing and concomitant concentration of resistant quartz grains by marine processes of the fluvial-deltaic lithic arenites (Horne et al., 1974). Paleocurrent data from the quartz arenites demonstrate south-southwest sediment transport, which has been attributed to longshore currents that were responsible for the elongate-linear trends of the sand bodies (Houseknecht, 1980).

Although attractive in its simplicity, and still held in high regard by some workers, there are problems with the barrier-shoreline model. First, the model contends that fossil flora and fauna identified as either Mississippian or Pennsylvanian are, in reality, facies controlled; as such, biostratigraphy, as applied to the question of the systemic boundary, is without foundation. Chesnut (1988) discusses biostratigraphy and the Mississippian - Pennsylvanian boundary at length and dismisses the facies argument by demonstrating similarities in evolutionary trends of supposed facies-controlled floras and faunas from within and outside the Appalachian Basin.

The quartz arenites also have been interpreted as south - southwest flowing braided river deposits draining the North American craton to the north (Donaldson and

Shumaker, 1981; Chesnut, 1988). A fluvial origin also has been suggested for quartz arenites of the Lee Formation in eastern Kentucky (Rice, 1984: Chesnut, 1988) and similar deposits in the Illinois Basin (Trask and Palmer, 1986).

In addition to depositional models for the Lower Pennsylvanian Lee and New River quartz arenites, Cecil and other, (1985) and Donaldson and others (1985) suggested that tropical weathering was a major contributing factor to the composition of the Upper Mississippian and Lower Pennsylvanian quartz arenites in the Appalachian basin. Studies in Indonesia (Cecil and others, unpublished data) clearly show that climate and weathering is a major contributing factor in the production of quartz arenite, especially under tropical conditions.

Middle Pennsylvanian Series

The Middle Pennsylvanian Series is subdivided into two formations, in ascending order, the Kanawha and Allegheny Formations. Depositional patterns of the Middle Pennsylvanian Series shows a progressive north-northwestward onlap of the North American craton and progressive filling of the Appalachian foreland basin followed by sediment by-passing as the loci of subsidence and sediment accumulation shifted northward (Donaldson et al., 1985).

The Kanawha Formation comprises the lower 75 to 90 % of Middle Pennsylvanian Series strata in southern West Virginia, and was named by Campbell and Mendenhall (1896) for exposures in the Kanawha River Valley in south-central West Virginia. The formation is a coal-bearing sequence of sandstone, siltstone, shale, and mudstone with minor occurrences of siderite, limestone, and flint clay. A maximum preserved thickness of 613 m (2000 ft) is present in southern West Virginia (Arkle et al., 1979; Blake et al., 1989), but thickness trends suggest that the formation continued thickening to the south and east of the present-day outcrop belt (fig. 4). Approximately 42 percent of West Virginia's original coal resource base and approximately 43 percent of its annual coal production (based on 1988 data) come from Kanawha Formation coal beds (Repine et al., in press) (fig. 83). Although the quality of these coal beds vary, they generally are low in sulfur content (median value 1.1 %, n=625) and ash yield (median value 11.9 %, n=625), medium to high volatile A bituminous in rank, and are produced for metallurgical and steam purposes.

Pennsylvanian strata in the Appalachian basin were originally assigned to various formations based on lithologic parameters and imperfectly understood regional relationships. As such, general usage often erroneously equates these lithostratigraphic units with chronostratigraphic units. Historically, age relations of Middle and Lower Pennsylvanian formations have generally been based on broad paleobotanical (megaflora) correlations, originally outlined by David White (1895, 1900a, 1900b).

Arndt (1979), in the proposed Pennsylvanian System stratotype, arbitrarily placed the upper boundary of the Kanawha Formation at the top of the Kanawha Black Flint of White (1891), a regionally identifiable key bed. The lower formation contact was placed at the base of the Lower Douglas(?) coal bed of Hennen and Teets (1919), the stratigraphically lowest coal bed overlying the Upper Nuttall Sandstone Member of the Lower Pennsylvanian New River Formation (**STOP 12**) (locations 3 & 4, fig. 4). This placement of the New River - Kanawha Formation boundary was based on the traditional assignment of the quartz arenites of the Pocahontas Basin to the New River Formation, rather than on any paleontologic data.

Correlation based on fossil flora can be problematic in many areas. Gillespie and Pfefferkorn (1979), in conjunction with the proposed Pennsylvanian System stratotype (Englund et al., 1979), reported the range zones of common and important megaflora taxa across the Lower - Middle Pennsylvanian boundary. However, in its area of occurrence, the coarse lithictype of the Nuttall

Sandstone Member is incompatible with the preservation of plant fossils, effectively acting as an unconformity across the Lower-Middle Pennsylvanian series boundary (Gillespie and Rheams, 1985; Lyons et al., 1985). As a result, the use of macrofloras in this area as a correlation tool is tenuous at best.

Recent unpublished studies at the West Virginia Geological and Economic Survey suggest that the Lower Douglas coal bed of Hennen and Teets (1919) is not correlative with, and is stratigraphically higher than, the Lower Douglas coal bed of Hennen and Gawthrop (1915) at its type locality near Douglas Station in western McDowell County (location 1, Figure xsection). The miscorrelation of the Lower Douglas coal bed of Hennen and Teets (1919) with the Glenalum Tunnel coal bed of Hennen and Reger (1914) has resulted in a diachronous Lower - Middle Pennsylvanian series boundary, which is traditionally placed at the New River - Kanawha formation contact.

Numerous depositional environments have been suggested for Kanawha Formation strata, including upper and lower delta plain (Horne and Ferm, 1978; Flores and Arndt, 1979; Donaldson and Shumaker, 1981; Ferm and Weisenfluh, 1989), back barrier (Hobday and Horne, 1977), and coastal plain (Tankard, 1986; Martino, 1989, 1992). Numerous brackish- to marine-influenced units have been reported by numerous workers from the Kanawha in southern West Virginia (Henry and Gordon, 1979), and also from equivalent strata in eastern Kentucky (Chesnut, 1991). However, their occurrences have generally been described in the context of either the barrier-shoreline model or a fluvial-deltaic model, both of which are autocyclic-dominated models (Donaldson and Shumaker, 1981; Horne and Ferm, 1978). Although some of these marine-influenced zones are areally restricted and possibly represent deposition in large bays, many of these units are areally widespread and do not fit the concept of interdistributary bay-fills.

Blake and others (in press) used four areally extensive marine zones and a widely recognized flint clay of volcanic origin associated with the Fire Clay coal bed to correlate the Kanawha Formation across southwestern West Virginia westward from the proposed Pennsylvanian System stratotype (Englund et al., 1979), with time-equivalent strata assigned to the Breathitt Formation in eastern Kentucky (fig. 83). These four marine zones have been shown to be important in developing a consistent stratigraphic framework across the central Appalachian basin (fig. 84) (Rice and Smith, 1980; Rice et al., 1987; Chesnut, 1988, 1991; Blake et al., in press). These marine zones are, in ascending order: the Betsie Shale Member (same in the Breathitt Formation in eastern Kentucky), the Dingess Shale Member (Kendrick Member, Breathitt Formation in eastern Kentucky), the Winifrede Shale Member (Magoffin Member, Breathitt Formation in eastern Kentucky), and the Kanawha Black Flint of White (1891). Cavaroc and Ferm (1968), Reppert (1979), and Watson (1992) have discussed the various facies and distribution of the Kanawha Black Flint of White (1891), and Martino (1989) has reported on facies recognized in the Dingess Shale Member. Cavaroc and Ferm (1968) have suggested that the Kanawha Black Flint of White (1891) is correlative with the Kilgore Flint of eastern Kentucky. Figure 34 shows the areal distribution of one of these important Kanawha Formation marine zones in southern West Virginia. The basin-wide occurrence of the four marine zones discussed above has led some workers to suggest they represent major transgressive events controlled by allocyclic processes (Chesnut, 1989; Blake et al., in press), driven either by glacial-eustatic sea level changes (Busch and Rollins, 1984; Heckel, 1986, 1990; Veveers and Powell, 1987), basin-wide subsidence due to crustal downwarping under the weight of an advancing thrust sheet (Tankard, 1986; Klein and Willard, 1989), or a combination of the two processes.

Subdivision of the Kanawha Formation

Progressive vertical changes in dominant lithologies, coal bed morphology and

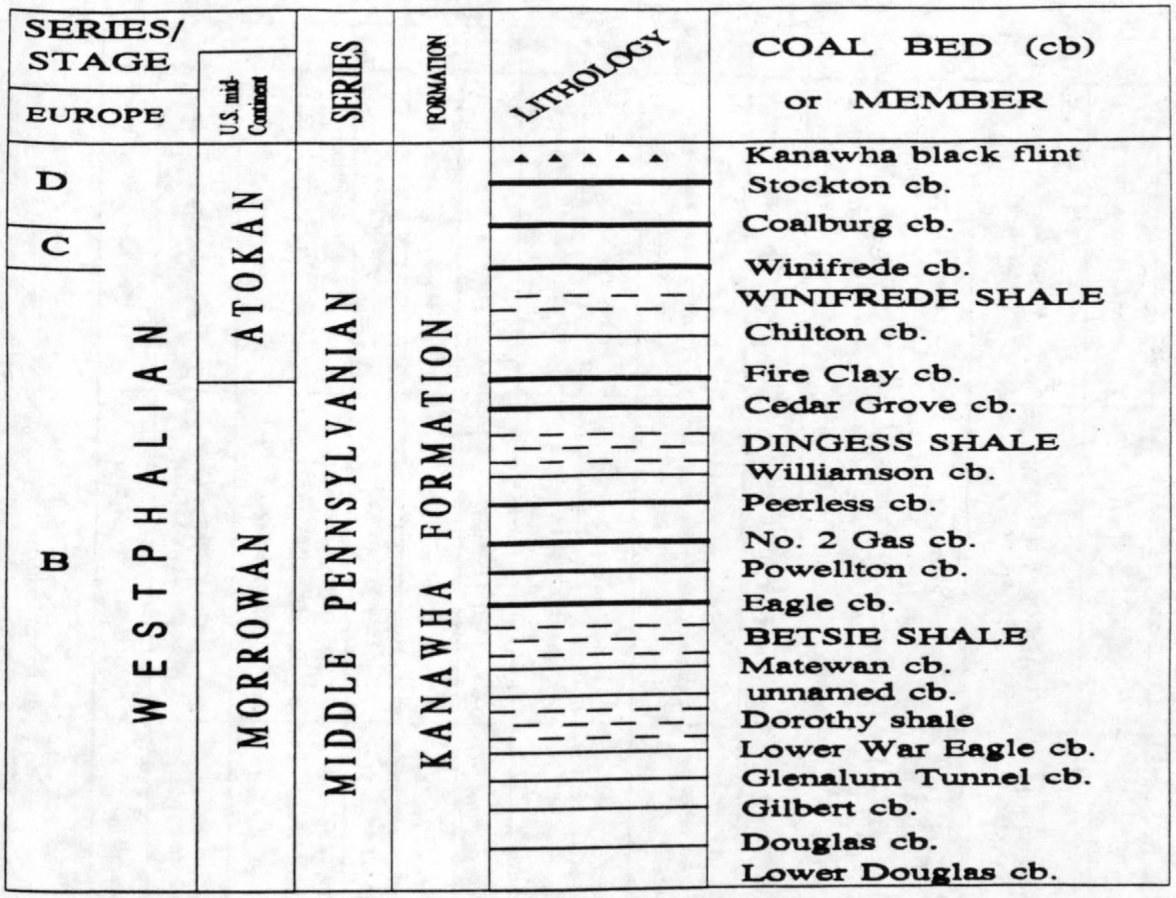


Figure 84 - Generalized stratigraphic column of the Kanawha Formation in West Virginia showing the position of economically important coal beds.

amount of marine influence evident in the rocks led Blake and others (1989) to informally subdivide the Kanawha Formation into three subunits, the lower, middle, and upper members for discussion purposes. The lower member, 335 m (1100 ft) thick in the area of Douglas Station, McDowell County (location 1, fig. 4), consists of a series of brackish- to marine-influenced zones and associated coastal plain sediments. The dark gray to black marine sequences vary from shale to very fine-grained sandstones, often are lenticular to flaser bedded, frequently contain ellipsoidal limestone concretions, commonly are over 15 m (50 ft) thick, and frequently have an erosional basal contact with underlying units generally marked by a transgressive lag deposit. Bioturbation is common and brackish to marine invertebrate fossils are locally abundant. For a detailed discussion of the Kanawha Formation marine zones, see Martino (1989, 1992, 1993). Coal beds in the lower member are generally bright and gassy, but usually are thin, seldom reaching minable thickness. The top of the lower member is placed at the top of the Betsie Shale Member, an important lithostratigraphic marker bed throughout the basin (Rice et al., 1987). The lower member thins from its maximum thickness in western McDowell County northward until on Chestnut Ridge (**STOP 1**), the Betsie Shale Member rests directly on Upper Mississippian Mauch Chunk strata (location 6, fig. 4).

The middle member of the Kanawha Formation extends upward from the top of the Betsie Shale Member to the top of the

		KENTUCKY		WEST VIRGINIA				
Series	Formation	Big Sandy District (modified from Rice & Smith, 1980; Rice et al, 1987)	Williamson 7.5' Quadrangle (modified from Alvord & Trent, 1962)	Tug Fork Area (Hennen & Reger, 1914)	Kanawha River Valley (modified from Hennen & Teets, 1914; Arndt et al, 1979)	Blake et al (in press)	Formation	Series
Middle Pennsylvanian (part)	Breathitt (part)	Hazard coal zone	Winifrede coal bed	Winifrede/Buffalo Creek coal bed	Winifrede coal bed	Winifrede coal zone	Kanawha (part)	Middle Pennsylvanian (part)
		Magoffin Member	Magoffin Beds of Morse (1931)	*Buffalo Creek Limestone of Hennen & Reger (1914)*	Winifrede Limestone of White (1908)	**Winifrede Shale Member**		
		Taylor coal bed	Taylor coal bed	*Chilton "A" coal bed*	Chilton coal bed	Chilton coal zone		
		Fire Clay coal bed	Fire Clay coal bed	*Chilton coal bed*	*Hernshaw (?) coal bed*	Fire Clay coal zone		
		(unnamed marine zone)	(unnamed marine zone)	(not reported)	*Seth Limestone of Krebs (1915)* ④	(unnamed marine unit)		
		Whitesburg coal bed	Whitesburg coal bed	Hernshaw coal bed	Cedar Grove coal bed	Cedar Grove coal zone		
		Kendrick Shale Member	Kendrick Shale of Jillson (1919)	Dingess Limestone of Hennen & Reger (1914)	(position commonly described as occupied by the Peerless sandstone)	**Dingess Shale Member**		
		Williamson coal bed	Williamson coal bed	Williamson coal bed	*Alma coal bed*	Williamson coal zone		
		Elkins Fork Shale of Morse (1931)	Elkins Fork Shale of Morse (1931)	*Seth Limestone of Krebs (1915)* ③	Campbell Creek Limestone of White (1885)	Campbell Creek Limestone of White (1885)		
		Upper Elkhorn coal zone: Upper Elkhorn No. 3 coal zone	Nosben coal bed	*Cedar Grove coal bed*	Campbell Creek coal zone: Peerless coal bed	**Peerless coal zone**		
			Sidney coal bed	*Lower Cedar Grove coal bed*	Campbell Creek coal zone: No. 2 Gas coal bed	**No. 2 Gas coal zone**		
		Upper Elkhorn coal zone: Upper Elkhorn Nos. 1 & 2 coal zone	Alma coal bed	Alma coal bed	Powellton coal bed	Powellton coal bed		
		Crummies Member	*Campbell Creek Limestone of White (1885)* ①	*Campbell Creek Limestone of White (1885)* ③	Cannelton Limestone of White (1885)	Cannelton Limestone of White (1885)		
		Lower Elkhorn coal zone	Pond Creek coal bed	*Campbell Creek coal bed*	Eagle coal bed	Eagle coal zone		
		Betsie Shale Member	*Cannelton Limestone of White (1885)* ②	*Cannelton Limestone of White (1885)* ③	Eagle Limestone and Shale of White (1891)	**Betsie Shale Member**		
		Matewan coal bed	(not discussed)	Matewan coal bed	(unnamed coal bed)	**Matewan coal zone**		

①as used by Alvord, 1971
②as used by Trent, 1965
③as used by Hennen & Reger, 1914
④as used by Hennen & Teets, 1919

Bold Usage this report

Italic Erroneous nomenclature usage in previous reports

Figure 83) Table showing Kanawha Formation nomenclature changes across southern West Virginia.

Winifrede Shale Member and is 259 m (850 ft) thick on Bolt Mountain (western Raleigh County, West Virginia), which is in an area of thickest preserved development (location 2, fig. 4). Two stratigraphically important and areally widespread marine zones, the Dingess Shale Member (Kendrick Member in eastern Kentucky) and Winifrede Shale Members (Magoffin Member in eastern Kentucky), are present in the middle member and mark widespread transgressive events. Other, less areally widespread brackish- to marine-influenced sequences have also been reported from this member, perhaps representing deposition in either interdistributary bays or intermittent coastal plain and near-shore, marginally marine environments.

The percentage of alluvial channel sandstones increases in the middle member over that in the lower member and the amount of minable coal resources is greatly increased. Coal beds of the middle member are intermediate in composition and appearance between the bright, gassy coal beds common in Lower Pennsylvanian units and lower in the Kanawha Formation, and the splint-rich coal beds that are common higher in the Kanawha Formation and the overlying Allegheny Formation (Charleston Sandstone of Arndt, 1979). Coal beds of the middle member tend to be low in sulfur content and ash yield, are areally extensive, and are of minable thickness over large areas of southern West Virginia and eastern Kentucky. Widely-mined coal beds include, in ascending order: the Eagle, Powellton, No. 2 Gas, Peerless, Cedar Grove and Fire Clay. In addition to the widespread marine units used as lithostratigraphic markers, a thin, widespread flint clay associated with the Fire Clay coal bed is present in this member. The flint clay represents an ancient volcanic ash fall as indicated by its widespread areal distribution, a kaolinite/quartz dominated mineralogy, and the presence of igneous minerals such as allogenic, euhedral sanidine, rutile, euhedral zircon, feldspar, embayed beta-form quartz, and glass shard ghosts (Seiders, 1965; Bohor and Triplehorn, 1981; Chesnut, 1985; Keiser et al., 1987). Thickness trends in the middle member are similar to those in the lower member, but the thinning is not as severe in the more northern areas as is the case with the lower member (see fig. 4).

The upper member is 137 m (450 ft) thick in the Bolt Mountain section (location 2, fig. 4), and is dominated by thick, multistoried fluvial-deltaic channel-fill sandstones often reaching 30.5 m (100 ft) in thickness. Sandstone percentages increase upward throughout the Kanawha Formation, and this trend is especially well-demonstrated in the upper member. These thick sandstones represent stacked fluvial channel-fill deposits suggestive perhaps of an alluvial plain setting. Coal beds are generally thick, multiple-bedded, commonly low in ash and sulfur content, laterally continuous, and usually are dull and blocky in appearance. Three main coal beds are present in the upper member, but coal bed splitting often results in these beds occurring as multiple-bed zones. Economically important coal beds (zones) are, in ascending order, the Winifrede, Coalburg, and Stockton. The top of the Kanawha Formation is placed at the top of the Kanawha Black Flint of White (1891) in the proposed Pennsylvanian System stratotype area (Arndt, 1979), an easily recognizable key bed.

Allegheny Formation

The Allegheny Formation, originally named for exposures along the Allegheny River in western Pennsylvania (Rogers, 1840), is a coal-bearing, sandstone-dominated sequence intercalated with shale, mudstone, marine and non-marine limestone and fire clays and flint clays. Marine limestones and shales are common in eastern Ohio, western Pennsylvania, and the northern West Virginia panhandle, but have not been reported from the area covered by the field trip. Non-marine limestones are sporatically associated with the Upper Kittanning, Lower Freeport, and Upper Freeport (**STOP 2**) coal beds in northern West Virginia but have not been reported from central and southern portions of West Virginia (Arkle et al., 1979).

The Allegheny Formation is a litho

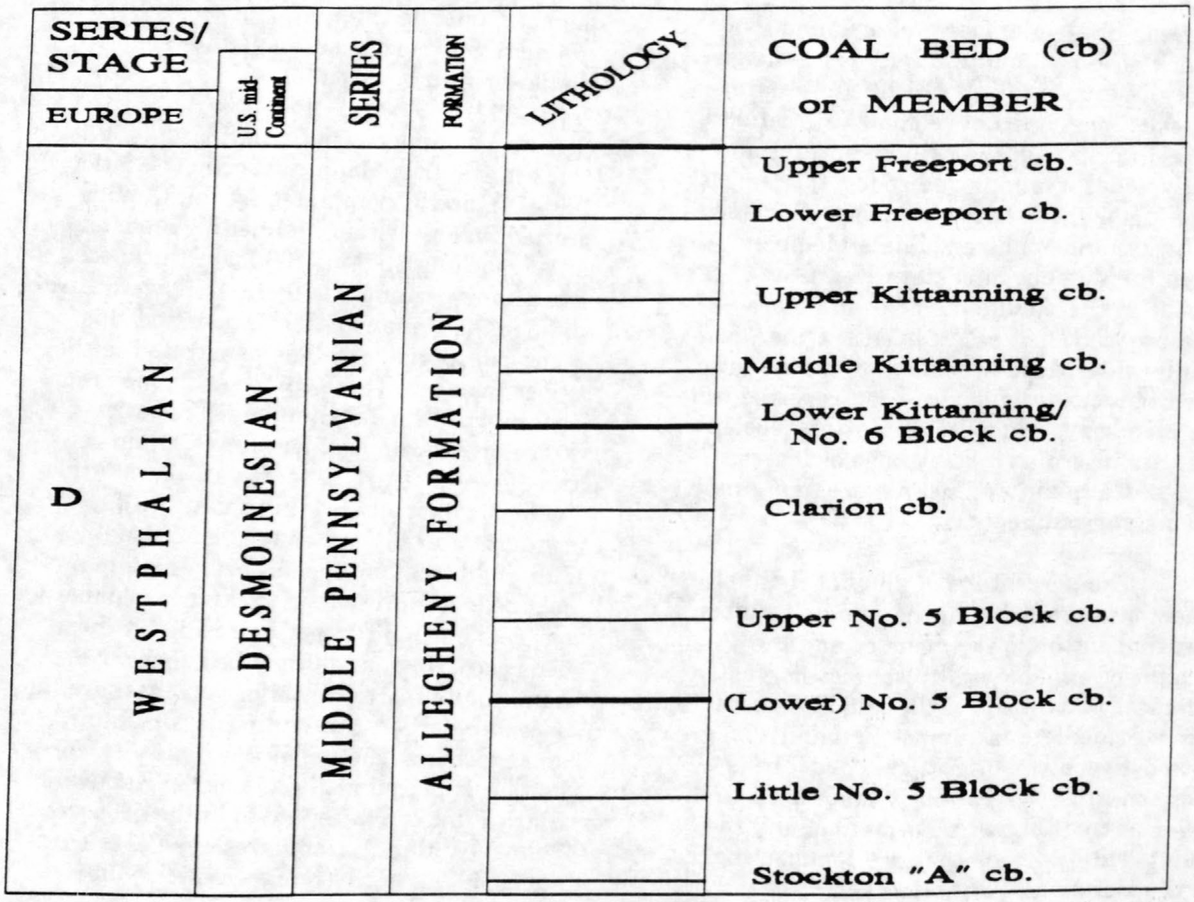


Figure 85 - Generalized stratigraphic column of the Allegheny Formation in West Virginia showing the position of economically important coal beds.

stratigraphic unit with variable boundaries across the Appalachian Basin. The upper boundary in eastern Ohio, western Pennsylvania, and northern West Virginia is placed at the top of the Upper Freeport coal bed, or in its absence, the base of the Mahonning Sandstone Member. Time-equivalent strata in eastern Kentucky are assigned to the Breathitt Formation. In central and southern West Virginia the upper boundary is placed at the top of the Mahoning sandstone (as used by Krebs and Teets, 1914), which essentially corresponds with the first occurrences of red- and green-gray beds that characterize strata of the overlying Conemaugh Group. The lower boundary is placed at the top of the Kanawha Black Flint of White (1891), a regionally important marine zone. In areas where the Kanawha black flint is absent, the boundary is placed at the base of the Homewood Sandstone (as used by Krebs and Teets, 1916) or the top of the Stockton coal bed. David White (1900a) considered the megaflora associated with the Stockton coal bed, usually within 9.1 m (30 ft) of the overlying Kanawha Black Flint of White (1891), as being indistinguishable from the megaflora of the basal Allegheny Formation of its type area (Clarion Formation of the Allegheny Group in western Pennsylvania). The lower formation contact in northern West Virginia is placed at the top of the Homewood Sand

stone Member of the Pottsville Formation (Pottsville Group undivided). Adjusting the formation boundaries to comparable positions indicates no significant thickness variations between northern and southern West Virginia.

Variation in formation boundaries across West Virginia has resulted, in part, from the mistaken correlation of the Lower Kittanning coal bed of the northern areas with the (Lower) No. 5 Block coal bed of central and southern West Virginia. Kosanke (1984, 1988) used miospores to correlated the No. 6 Block coal bed of the proposed Pennsylvanian System stratotype with the Lower Kittanning coal bed of the more northern portion of the Appalachian Basin. In southern West Virginia, the Allegheny Formation is thick with well-developed coal beds, whereas the upper portion is thin with little to no coal beds present, the reverse being true in the more northern areas of the basin. These facies changes led earlier workers (Krebs and Teets, 1914; Hennen and Gawthrop, 1917; Reger and Tucker, 1920; Reger and Price, 1921; Lotz, 1970) to correlate the (Lower) No. 5 Block, Upper No. 5 Block, and No. 6 Block coal beds of southern West Virginia with the Lower Kittanning, Middle Kittanning, and Upper Kittanning coal beds of the northern portion of the basin. These are miscorrelations that persist widely to this day. Allegheny Formation coal beds are frequently split by clastic partings into multi-bedded coal zones, further hindering local and regional correlations.

Eight Allegheny Formation coal beds have been widely mined in West Virginia; they are, in ascending order, the Little No. 5 Block, Lower No. 5 Block, Upper No. 5 Block, Clarion, No. 6 Block/Lower Kittanning, Middle Kittanning, Upper Kittanning, and Upper Freeport (fig. 87). Coal quality is variable with a median sulfur content of 1.63 percent (n=250) and a median ash yield of 14.8 percent (n=250) for all Allegheny Formation coal beds. Typically, the quality of the "Block" coals of central and southern West Virginia is good, with the coals being generally low in sulfur content (<1 %) and ash yield (<8 %). They resemble upper Kanawha Formation coal beds in being dull, blocky, with common splint bands. In contrast, Allegheny Formation coal beds in the more northern portions of the state, the Kittannings and Freeports, tend to be more bright-banded, and are higher in sulfur content (>2 %), and ash yield (>8 %).

Fireclays with associated flint clays, often of economic thickness and purity, are common in the northern portion of West Virginia, but are only known from one horizon, the No. 6 Block coal horizon, in southern West Virginia. (**STOP 8**). Economically exploited fire clays include, in ascending order, the Elk Fire Clay, laterally equivalent to with the No 6 Block coal bed, the Clarion Fire Clay below the Clarion coal bed, the Lower Kittanning Fire Clay below the Lower Kittanning coal bed, the Hardman and Upper Kittanning Fire Clays below the Upper Kittanning coal bed, and the Bolivar Fire Clay below the Upper Freeport coal bed. These deposits are thought to represent extended periods of weathering and leaching of a soil under tropical conditions (Cecil et al., 1985).

The regional distribution of facies in the Allegheny Formation, with marine shales and limestones to the north and increasing percentages of fluvial-deltaic sandstones to the south led various workers to suggest a north, northwestward prograding deltaic sequence with an alluvial plain facies in central and southern West Virginia (Donaldson and Shumaker, 1981).

Stratigraphy of Upper Pennsylvanian strata in West Virginia

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Conemaugh Group

The Conemaugh Group (Upper Pennsylvanian; Missourian; Stephanian) first named and described by F.W. Platt (1876) from exposures along the Conemaugh River, Pennsylvania, includes all the strata from the top of the Upper Freeport coal bed of the underlying Allegheny Formation to the base of the Pittsburgh coal bed of the overlying Monongahela Group. Flint (1965) divided the Conemaugh into two formations, the Glenshaw Formation below and the Casselman above with the contact being placed at the top of the Ames Limestone of Andrews (1873). We will examine the Ames Limestone of Andrews (1873) and much of the overlying Casselman Formation at **STOP 3**. Short intervals at the top of the Casselman Formation will also be seen at **STOPS 4, 5** and **6**. The lowermost 34 m (112 ft) of the Conemaugh Group are exposed at **STOP 2**. In the vicinity of these stops the Conemaugh Group is about 183 m (600 ft) thick, with the two formations being of about equal thickness.

The Conemaugh Group consists of interbedded coal, red, green and gray shale, red, green and gray mudstone or claystone, nonmarine limestone, marine limestone and shale, and sandstone. The Conemaugh Group strata were deposited on a relatively stable platform (Arkle, 1969, Donaldson, 1974), the Dunkard Basin representing the cratonic or distal portion of the foreland basin (Donaldson et.al., 1985). In response to the relative tectonic stability, a shallow water, fluvial dominated deltaic system, coined the *West Virginia delta* by Donaldson (1969), prograded northward and westward into an embayment of the epeiric sea of the Central Interior during late Pennsylvanian (Donaldson, 1969, 1974; Donaldson et.al., 1985).

In northern West Virginia the lowermost 46 m (150 ft) of the Conemaugh is similar to the underlying Allegheny Formation in that it is dominated by thick, multistoried fluvial channel-fill sandstone units; in ascending order, these are the Mahoning Sandstone of Lesley (1856) and the Buffalo Sandstone of White (1878). The lowermost Conemaugh Group, including the Mahoning Sandstone, crop out at **STOP 2**. Above these units are found the first appearance of redbeds in the Pennsylvanian of the Appalachians. These occur as red and mottled red, gray, and green mudstones (paleosols) and shales. Paleosol features include obliteration of primary sedimentary features, mottling, calcareous pedotubules, rhizoconcretions, accumulations of carbonate glaebules and petrocalcic horizons, and slickensided vertic structures. These same features also occur in the remaining Upper Pennsylvanian rocks above the Conemaugh Group. Laterally extensive, interbedded red or mottled mudstones, claystone, siltstone, and shales have been given stratigraphic names and can be recognized on a reginal basis throughout the basin. These units, herein interpreted as paleosols, include the Pittsburgh Red Shale of White (1891), the Birmingham Shale of Stevenson (1876) and the Clarksburg Red Shale of Hennen (1912). These units can be seen at **STOP 3**. Commonly, sandstones and shales in this interval are also colored shades of green, unlike the solid gray colors of Lower and Middle Pennsylvanian strata. These red (and green) beds, which have been interpreted as alluvial plain sediments (Arkle 1974), occur in most areas where the Conemaugh Group is found in West Virginia, Pennsylvania, Maryland, Ohio, and Kentucky. **Areas of prolific redbed development also correspond with poor coal development, a phenomenon that is especially evident along the southeast outcrop in central West Virginia, and in the Charleston, West Virginia area. The paucity of coal deposits has been attributed to a relatively dry** paleoclimate (e.g., Cecil et al., 1985).

Nonmarine lacustrine limestone beds that first appear in the Allegheny Formation associated with the Upper Kittanning coal bed (Arkle, 1974) become more abundant in the Conemaugh Group. They generally occur as single beds, or thin complexes of interbedded micrite, calcareous shale, calcareous mudstone, or argillaceous limestone. Named units include the Ewing Limestone of Orton (1878), the Elk Lick Limestone of Platt and Platt (1877), the Clarksburg Limestone of White (1891), the Lower Pittsburgh Limestone of Platt and Platt (1877) and the Upper Pittsburgh Limestone of White (1878). These beds are exposed at **STOPS 3** and **4**.

Coal beds are not as abundant or thick in the Conemaugh Group as in the underlying Allegheny Formation or overlying Monongahela Group (compare figs. 9 and 16). They are predominantly high volatile A or B bituminous coals where they occur along the field trip route. Ash yield and total sulfur content of Conemaugh Group coal beds is variable, the median values being 15.7 % ash yield and 2.9 % total sulfur content (fig. 10; table 2). At **STOP 3**, the Elk Lick coal bed is about 0.9 m (3 ft) thick, but is not very extensive, and has not been commercially mined in the area. It does, however, thicken along the eastern outcrop in Maryland and West Virginia, as do other Conemaugh group coal beds, and has been mined there, as well as in a few locations in north central West Virginia. The other Conemaugh Group coal beds that crop out on the field trip route, namely the Harlem (**STOP 3** area and "rolling stop" at Goshen Road), the Little Clarksburg (**STOP 3**), and the Little Pittsburgh (**STOPS 4** and **6**) are not minable, but do serve as useful stratigraphic marker beds.

The last major marine transgressions in the Pennsylvanian in north-central West Virginia occur in the Glenshaw Formation. Of the seven marine units identified in eastern Ohio and western Pennsylvania (Busch and Rollins, 1984), four are found in West Virginia. The Brush Creek Limestone of I.C. White (1878) (not seen on this field trip), and the Ames Limestone (**STOP 3**) of Andrews (1873) are the most widespread, and are very important stratigraphic marker beds.

Along the south and southeast Conemaugh outcrop, in the area of the proposed Pennsylvanian System stratotype, the Conemaugh Group consists largely of red and mottled mudstone and shale facies with thin lenticular coal beds, sandstones, and a few nonmarine carbonate beds (Henry et al., 1979). Neither the marine facies of the Brush Creek Limestone of White (1878), nor the Ames Limestone of Andrews (1873), important marker beds to the north, extend this far to the south. In addition, the Upper Freeport coal bed, which marks the contact with the underlying Allegheny Formation in northern West Virginia, western Maryland, western Pennsylvania and eastern Ohio (**STOP 2**), is missing in this more southerly area. As a result, the contact in this region is placed near the occurrence of the first redbeds at the top of the "Mahoning" Sandstone, (Henry et al., 1979), which may or may not be correlative with the Mahoning Sandstone of Lesley (1856).

The Monongahela Group

Basal portions of the Upper Pennsylvanian Monongahela Group will be examined at **STOPS 4, 5,** and **6** and at a "rolling stop" at Clarksburg, WV. The Monongahela Group was first used for exposures of these rocks along the Monongahela River in the vicinity of Pittsburgh, Pennsylvania (Rogers, 1840). In current usage the Monongahela Group includes the strata from the base of the Pittsburgh coal bed to the base of the Waynesburg coal bed (Berryhill and Swanson, 1962). An excellent history of the nomenclature is given in Arkle (1959). Along the field trip route the thickness of the Monongahela varies little, from about 116 m (380 ft) thick in the area of **STOP 4** to about 122 m (400 ft) thick in the area of **STOP 6**. Regionally the Monongahela Group gradually thins to the north west where it is about 61 m (200 ft) on the northwestern outcrop in eastern Ohio (Arkle, 1969, 1974). The extent of the Monongahela Group is limited by erosion on all sides, and is preserved along the limbs of the northeast-southwest trending Pittsburgh-Huntington Synclinorium, also called the Dunkard basin, which extends from Huntington, West Virginia northeast to Pittsburgh, Pennsylvania (fig. 17). On this field trip we will examine only the lower 55 m (180 ft) of Monongahela Group sediments.

The Monongahela Group consists of interbedded coal, red, green and gray shale, red, green and gray mudstone or claystone, and nonmarine limestone and sandstone. Like the underlying Conemaugh Group, these rocks are interpreted to have been deposited on a relatively stable platform (Arkle, 1969, Donaldson, 1974), the Dunkard basin representing the cratonic or distal portion of the foreland basin (Donaldson et.al., 1985). Many previous interpretations of Monongahela strata were based an deltaic facies-mosaic models (autocyclicity) with peat and limestone deposition in interdistributary environments and sand deposition confined to distributaries (e.g. Donaldson, 1979). However, the quality, thickness, and widespread distribution of units such as the Pittsburgh coal bed indicate that autocyclic controls on lithostratigraphy are subordinate to allocyclic controls. The regional extent of coal beds, paleosols, siliciclastic intervals, and limestone beds appear to be better explained by changes in climate and base level change.

The Monongahela Group contains more nonmarine carbonates than any other part of the Pennsylvanian section. These micritic carbonates occur in complexes interbedded with argillaceous limestones, calcareous mudstones, and calcareous and noncalcareous shales. The diminutive and low diversity nonmarine fauna, dominated by ostracodes, is not readily observed macroscopically. Remains of gastropods, bivalves, the worm tube *Spirorbis*, the trace fossil *Lockeia*, and vertebrate fragments have also been reported (Dickson, 1977). No marine fossil forms have been reported to date, but conditions may have fluctuated from fresh to brackish water. For that reason, the inclusive term "nonmarine" is used as advocated by Dickson (1977).

These carbonate complexes formed in shallow water lakes and paludal environments (Dickson, 1977), with the cyclicity of these units suggesting considerable fluctuation in shoreline position and base level (Donaldson et al., 1985). Dolomite is reported from the Benwood Limestone of Campbell (1903) (not seen on this field trip), determined from both x-ray diffraction and thin section. The microscopic dolomite occurs as euhedral to subhedral grains, and is suggested to be the result of evaporative conditions (Marrs, 1981). Photosynthesis by algae presumably played a major role in precipitation of the calcium carbonate in these lakes, and high evaporation rates probably accelerated this process (Dickson, 1977). The micrite beds exhibit a wide variety of evidence for shallow water deposition and subaerial exposure. These include subareal crusts, desiccation cracks, brecciation, breccia-filled fractures, rooting, nodularization, and bioturbation. Spar calcite occurs in dismicritic textures, within ostracode carapaces, and as minor filling around intraclasts. The internal features of the micrites are most obvious on polished blocks or cores, but they also can be seen on outcrop with careful observation. The calcareous mudstones are interpreted to be paleosols developed on calcareous lacustrine muds exposed in paludal environments (Dickson, 1977). Shales may represent locally reworked calcareous mud, or detrital input from fluvial systems. The character of these carbonate complexes can be observed in exposures of the Redstone Limestone of Platt and Platt (1877), the Fishpot Limestone of Stevenson (1876), and other carbonate intervals at **STOP 4** (fig. 16).

Coal beds are generally thicker and more extensive in the Monongahela than the underlying Conemaugh Group. Named coal beds that crop out on the field trip route include the Pittsburgh, Redstone, Fishpot, and Sewickley which are of high volatile A and B rank. As a group, Monongahela Group coal ash yields vary from about 5 % to 25 %, the median being 11.9 %. Sulfur contents are variable with the median sulfur value being 3.1 % (fig. 10; table 2). They are mined principally for steam generation.

The Pittsburgh coal bed is by far the most economically important of these coal beds, supporting significant portions of the coal mining industry of West Virginia, western Pennsylvania and eastern Ohio. The bed exhibits remarkable lateral persistence of thickness and quality. Ash yields vary from about 5 to 15 %, whereas sulfur content ranges from near 1 % to more than 5 %. The field trip route follows along the eastern outcrop of the Pittsburgh coal bed traversing from the main minable body at **STOP 4**, southward to where it thins significantly at **STOP 5,** and then to **STOP 6** where the bed is represented by a thin dark gray to black claystone (figs. 17 and 18).

The minable body of Pittsburgh coal covers a large area of northern West Virginia, southwest Pennsylvania, and eastern Ohio (fig. 17). Throughout this northern region, the Pittsburgh coal bed is remarkably persistent. Bed thickness varies from a maximum of about 6 m (22 ft) (with partings) in erosional outliers in western Maryland, gradually thinning to 1 m (3-4 ft) at the farthest northwest outcrop in the northern panhandle of West Virginia and eastern Ohio (fig. 17). Throughout most of this minable area the bed occurs as a thick lower bench (main bench) with a variable number of overlying thin coal benches interbedded with shale or claystone (e.g. fig. 16). Local thin coal areas are usually the result of scouring by the superjacent Upper Pittsburgh Sandstone of Stevenson (1876).

South and southwest of the northern minable area, the Pittsburgh coal bed thins and is absent over large areas (fig. 17). The coal in this region generally occurs as a single bench. Where the Pittsburgh locally is of minable thickness, (e.g. near Charleston and **STOP 6**, fig. 17), the physical structure and thickness is often erratic, suggesting a strong paleotopographic influence. **STOPS 5** and **6** (fig. 18) illustrate the thinning of the Pittsburgh coal bed. At **STOP 7** the Pittsburgh coal bed is represented by a thin black claystone, interpreted as the histic epipedon of a mottled paleo-Vertisol or Ultisol.

The overlying Redstone coal bed is not as laterally persistent as the Pittsburgh and generally is less than minable thickness (i.e. less than 71 cm, 28 in), although minable areas do occur in widely scattered locations (Lotz, 1970). **STOPS 4 and 5** are in two of these minable areas. The thick Redstone coal seen at the Clarksburg rolling stop is confined to the immediate area of the exposure.

The Fishpot coal bed is generally thin, less than 0.3 m (1 ft) thick, with a maximum thickness of 0.6 m (2 ft). However, thin coal or carbonaceous shale often occurs at this stratigraphic position throughout the Dunkard basin. The Fishpot coal bed can be seen at **STOP 4**, and at the Clarksburg rolling stop, but is absent to the south.

The Sewickley coal bed is the most economically important Monongahela Group coal after the Pittsburgh coal bed, and is of minable thickness over a large area of the northern part of West Virginia (Lotz, 1970). It has been extensively underground and surface mined in the area of **STOP 4**, but has thinned to about 15 cm (6 in) at the Clarksburg rolling stop and is generally absent south of that point. The Sewickley coal bed is often associated with fine to coarse clastic rocks. Thin splits of the coal are often interbedded with these clastics, which sometimes results in miscorrelation with the underlying Fishpot coal bed.

The lower part of the Monongahela Group seen on the field trip traverse exhibits a change from coal and lacustrine limestone dominated sections (gray color, e.g. **STOP 4**) to sections dominated by fluvial sedimentation and well-developed paleosols (red color, e.g. **STOP 6**, fig. 18). These facies changes were first noted by Arkle (1959) and have been interpreted as representing a change from a lacustrine-swamp (Arkle, 1959, 1974), or flood-basin/interdistributary bay portion of the depo-system (north) (Hoover, 1967; Donaldson, 1969, 1974), to an updip alluvial plain (south) (Hoover, 1967; Donaldson, 1969, Arkle, 1974). Red shales and red-mottled paleosols become common at **STOP 6**, where the presence of iron oxide suggests vadose zone development and subaerial exposure (fig. 18). Features of the paleosols include destruction of primary sedimentary features by pedoturbation, mottling, calcareous pedotubules, rhizoconcretions, carbonate glaebules, petrocalcic horizons, and slickensided vertic structures. The Redstone Limestone of Platt and Platt (1877) is the only carbonate bed present at **STOP 4** that is laterally persistent south of **STOP 5**. This limestone is poorly-developed at **STOP 6**, however, especially when compared to sections further north; it also is strongly overprinted by pedogenesis. The Clarksburg Section is transitional between **STOP 4** and stops further south, exhibiting more coarse clastics and fewer carbonate beds. The coal beds also become thinner or absent southward, with paleosols occurring at the same stratigraphic position. Thicker and more numerous sandstone units replace the carbonate beds, exhibiting rapid lateral changes from channel fill forms to thin, planar beds interbedded with flood plain red, green, and gray shales and paleosols.

The Pittsburgh coal bed, and its laterally-equivalent dark claystone and associated subjacent units, represent, in pedologic terms, a toposequence. The unit immediately above the coal represents yet another toposequence. A toposequence is a lateral mosaic of soil bodies that have formed under similar conditions (e.g. climate, organisms and parent material), and within the same time frame, but have been influenced by change in relief (Buol et al., 1988). In focusing on this interval, we have the opportunity to observe two stacked toposequences that formed under contrasting climatic conditions (Cecil, 1990). The lower toposequence evolved under wet conditions. In northern areas, considerable peat, a type of Histosol, accumulated on top of a thin, poorly developed mineral horizon (e.g. **STOP 4**) (fig. 16). In contrast, a thicker mineral paleosol horizon, capped by a histic epipedon, developed on the better-drained alluvial plain,. In this topographic position, conditions did not favor peat preservation. This lower mineral paleosol horizon, having formed under wet conditions, is devoid (leached) of appreciable carbonate, and is mottled red, gray and green (fig. 18 and 25).

The upper toposequence (paleosol) directly on top of the Pittsburgh coal bed at **STOP 4** (fig. 16), and above the dark shale at **STOP 6** (fig. 25) developed under dry or wet-dry seasonal conditions. Unlike the lower sequence, this paleosol contains carbonate nodules and thin petrocalcic horizons and other evidence of a drier climate. It is very similar in structure and characteristics across the inferred change in topographic relief, except that it becomes more red in the alluvial plain setting (**STOP 6**, figs. 18 and 25).

Sedimentology of the Upper Pennsylvanian Redstone Limestone, Northern Appalachian Basin

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Introduction

Nonmarine limestone beds comprise a significant portion of the Upper Pennsylvanian and Lower Permian strata of the northern Appalachian basin (fig. 88). They first appear in the basin in the late Middle Pennsylvanian Allegheny Formation (fig. 89); the Johnstown Limestone Member is the first nonmarine limestone to occur in the Formation. The lower part of the Conemaugh Group contains a series of marine limestones; however, higher in the Conemaugh the nonmarine limestones reappear. These limestones become a major lithology in the Monongahela Group, and are particularly abundant in the Pittsburgh Formation, where they locally constitute up to 80% of the section (fig. 89). Regional correlations of the Pittsburgh Formation show that five nonmarine limestone beds are persistent and widespread, covering up to 12,000 sq km in northern and central West Virginia, western Pennsylvania, eastern Ohio, and western Maryland. The Redstone limestone of Platt and Platt (1877) is the stratigraphically lowest of these five beds. It occurs above the Pittsburgh coal bed and below the Redstone coal bed, within the lower member of the Pittsburgh Formation (Berryhill and Swanson, 1962) (fig. 89). The first documentation of the Redstone limestone, made by Platt and Platt (1877, p.62, 88-91), describes it as being a massive, extensively quarried, 3.1 m (10 ft) thick limestone in southwestern Somerset County, Pennsylvania. Platt and Platt describe the stratigraphic location, based on measurement at a quarry, as 10 m (32 ft) above the base of the Pittsburgh coal bed and 2 m (6.5 ft) below the base of the Redstone coal bed, and say that the Redstone limestone is persistent in some parts while lacking in other parts of southwestern Pennsylvania. The term "Redstone limestone" has been used since 1877 to identify the limestone between the Pittsburgh and Redstone coal beds, but it is considered an informal name. The Redstone limestone is of particular interest because of its stratigraphic proximity to the very thick, widespread, and economically important Pittsburgh coal bed. In addition, the Redstone limestone is traceable over large distances, occurring as far north as Brooke County, WV and as far south as Cabell County, WV. It reaches 12 m (38.5 ft) in thickness in some localities. The purpose of this study is to develop a sedimentological model for deposition of the Redstone limestone.

Only a few detailed sedimentological studies have been done on the Pennsylvanian and Permian freshwater limestones in the northern Appalachian basin. Adams (1954) performed a general study of the Upper Pennsylvanian freshwater limestones in the Pittsburgh, PA area, and compared them to associated marine limestones. In a study of the Pennsylvanian and Permian rocks of the Washington, Pennsylvania area, Berryhill and others (1971) included limestone as one of the four principal lithofacies. The geometry and internal structure, texture, and mineralogy of the limestones were described and a revised stratigraphic nomenclature was defined for the Upper Pennsylvanian and Lower Permian strata. Marrs (1981) developed a depositional model for the Benwood limestone of Clapp (1907) in the Dunkard basin, suggesting a large, shallow lake in a slowly subsiding basin, and noting cyclicity within the Benwood and surrounding strata. Petzold (1989) studied varves in the Benwood limestone near Clarksburg, WV, and concluded that they were seasonal events related to phytoplankton blooms, seasonal rainfall, and

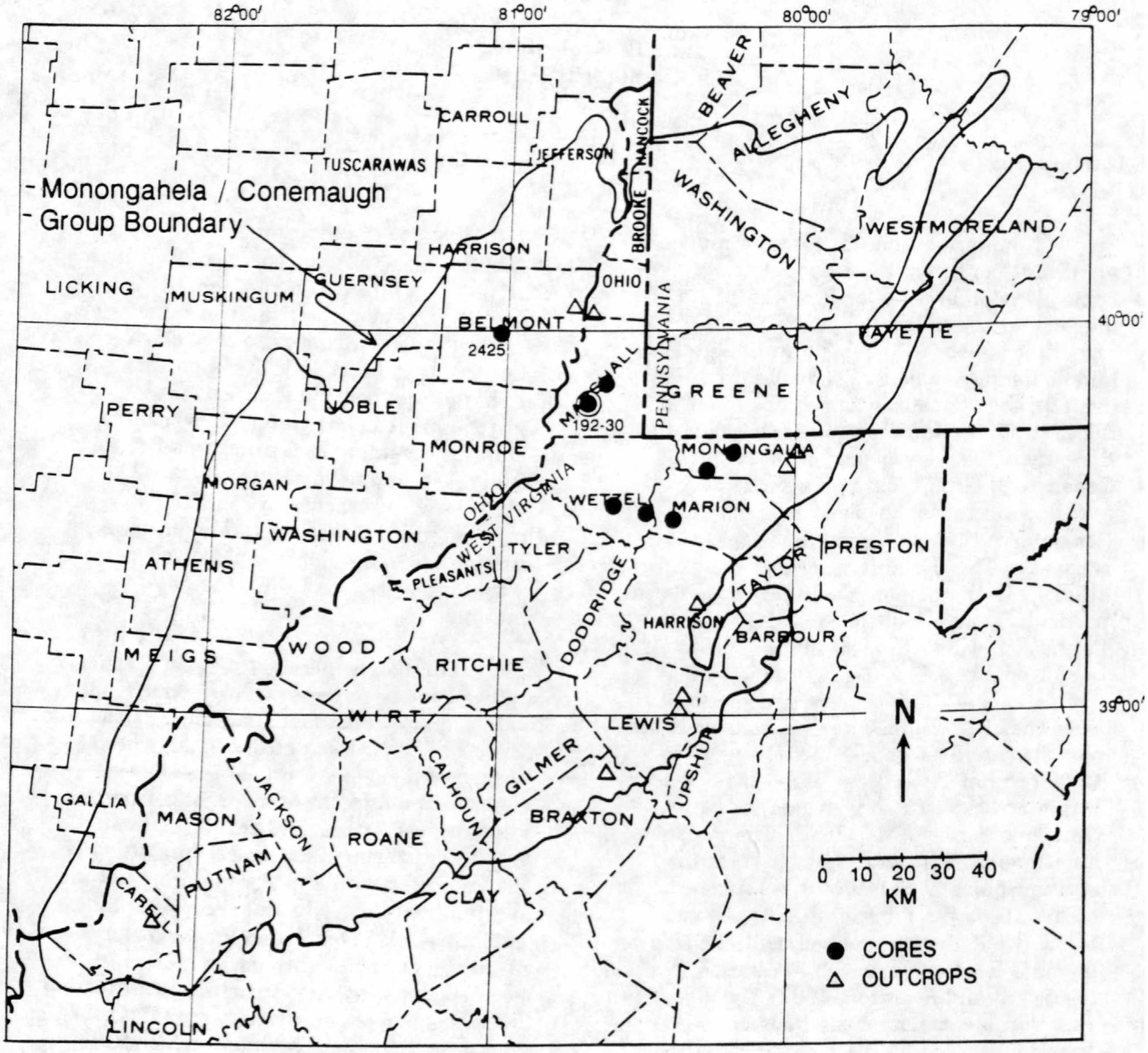

Figure 88 - Study area and sampling locations.

probably lake depth. Weedman (1988) conducted a detailed petrographic study of the Upper Freeport Limestone Member of the Allegheny Formation, defining seven calcareous microfacies. Weedman described a statistically significant lithofacies sequence, including the limestone, and proposed a depositional environment for the Upper Freeport Limestone Member.

Only limited studies have been done on the Redstone limestone. The Pittsburgh-Redstone coal interval, including the Redstone limestone, was studied for four Masters Theses (Conti, 1961; Linger, 1979; Jake, 1981; and Bean, 1982). These studies addressed the stratigraphic and facies relationships of the Redstone limestone to surrounding strata in portions of West Virginia and Pennsylvania. Paleontological

SYSTEM	GROUP	FORMATION	PRINCIPAL BEDS & MEMBERS	AVERAGE FORMATION THICKNESS
PERMIAN ?——?— PENNSYL-VANIAN	DUNKARD	WASHINGTON	Upper limestone member (1) Washington "A" coal bed Lower limestone member (1) Washington coal bed	25-45 m
		WAYNESBURG	Little Washington coal bed Waynesburg "B" coal bed Waynesburg "A" coal bed Waynesburg sandstone (1) Waynesburg coal bed	37-63 m
	MONONGA-HELA	UNIONTOWN	Uniontown coal bed	25-67 m
		PITTSBURGH	+ Limestone D (2) + Limestone C (2) + Limestone B (2) + Limestone A (2) Arnoldsburg Sandstone Member Lower Uniontown coal bed + Benwood limestone (4) Sewickley (Meigs Creek) coal bed + Sewickley limestone (5) Fishpot coal bed + Fishpot limestone (4) Redstone (Pomeroy) coal bed + Redstone limestone (3) Upper Pittsburgh sandstone (6) Pittsburgh coal bed	43-80 m
	CONEMAUGH	CASSELMAN	Elk Lick coal bed	73-97 m
Upper		GLENSHAW	Ames Limestone Member (marine) Harlem coal bed Bakerstown coal bed Brush Creek Limestone Member	30-80 m
Middle		ALLEGHENY	Upper Freeport coal bed + Upper Freeport Limestone Member Lower Freeport coal bed Upper Kittanning coal bed + Johnstown Limestone Member Middle Kittanning coal bed Lower Kittanning coal bed Vanport Limestone Member (marine)	75-110 m

+ denotes freshwater limestones

(1) Berryhill and others, 1971
(2) Berryhill and Swanson, 1962
(3) Platt and Platt, 1877
(4) Clapp, 1907
(5) White, 1891
(6) Stevenson, 1876

Figure 89 - Generalized geologic column, Northern Appalachian Basin.

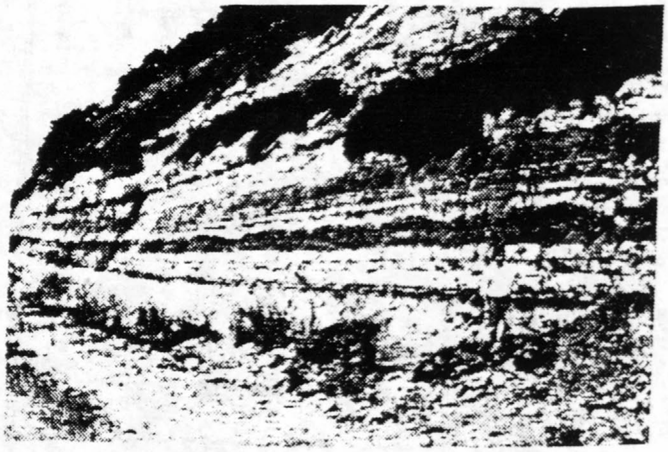

Figure 90 - Outcrop of the Pittsburgh, Formation showing several limestone beds, along I-70 in eastern Ohio.

studies of the Redstone limestone have been done by Eager (bivalves, 1975), Yochelson (gastropods, 1975), Sohn (ostracods, 1975), and Lund (vertebrates, 1975).

Geographically, the nonmarine limestone beds in the Monongahela Group are thicker and more abundant in the north, thinner and less common to the south. Donaldson (1974; 1979) interpreted the depositional settings of the northern Appalachian basin (Dunkard basin) as transgressive during the deposition of the Allegheny Formation and lower part of the Conemaugh Group, then regressive through the deposition of the Dunkard Group. The shoreline of the shallow sea, occupying the Dunkard basin during deposition of sediments in the lower part of the Conemaugh Group, regressed southwestward during the Late Pennsylvanian and Permian as fluvial dominant deltas brought sediment from the southeast into the basin, infilling the narrow sea embayment. Donaldson (1979) suggested that the large proportion of coal and limestone in the Monongahela Group represents a diversion of detrital supply to active river systems in the southern portion of the Dunkard basin, leaving the swamps and lakes to the north free to develop with little clastic influx. The nonmarine limestones are part of a system of interbedded clastic and autochthonous rocks in the northern Appalachians which often occur in repetitive sequences: sandstone - siltstone - limestone - underclay - coal - shale. The origin of these cyclothems has been a source of controversy since the mid 1960's. According to Cecil (1990), paleoclimatic changes in rainfall are one possible explanation for these cyclothems. In Cecil's paleoclimatic model, coal beds and leached paleosols indicate wet periods, while carbonates, evaporates, Vertisols, and Aridisols indicate dry periods; high siliclastic influx suggests seasonal rainfall.

Methodology

The study area is defined as the area in which the Redstone limestone could possibly occur, based on the outcrop of the underlying Pittsburgh coal bed or the Conemaugh-Monongahela Group contact where the Pittsburgh coal bed is absent (fig. 88). Initially, drill logs and measured sections were collected from various localities within the study area. Cross-sections and an isopach map (fig. 91) were prepared to determine the distribution of the Redstone limestone and its relationship to surrounding strata. These data were also used in determining sampling locations. Samples were collected from nine outcrop localities and from eight cores (fig. 88). Detailed sedimentologic descriptions were made at each outrop locality and for each of the eight cores. Such characteristics as rock color, texture, sedimentary structures, bedding, macrofossils, pyrite and organic content were described.

The petrologic results of this study are primarily based on 23 closely spaced samples collected from core #192-30 (fig. 88), which is located in the west-central portion of Glen Easton 7 1/2' quadrangle, south-central Marshall County, WV. Core #192-30 includes the entire Redstone limestone which at this locality is 6.7 m thick. The core was halved, then logged in the laboratory using a binocular microscope for detailed description. A light etching of the sawed face with HCL was helpful in identifying sedimentary features. Twenty-three samples were taken, representing a variety of lithologic zones. Two thin sections were then prepared for each sample--one was stained for calcite and iron, whereas the other was polished and left

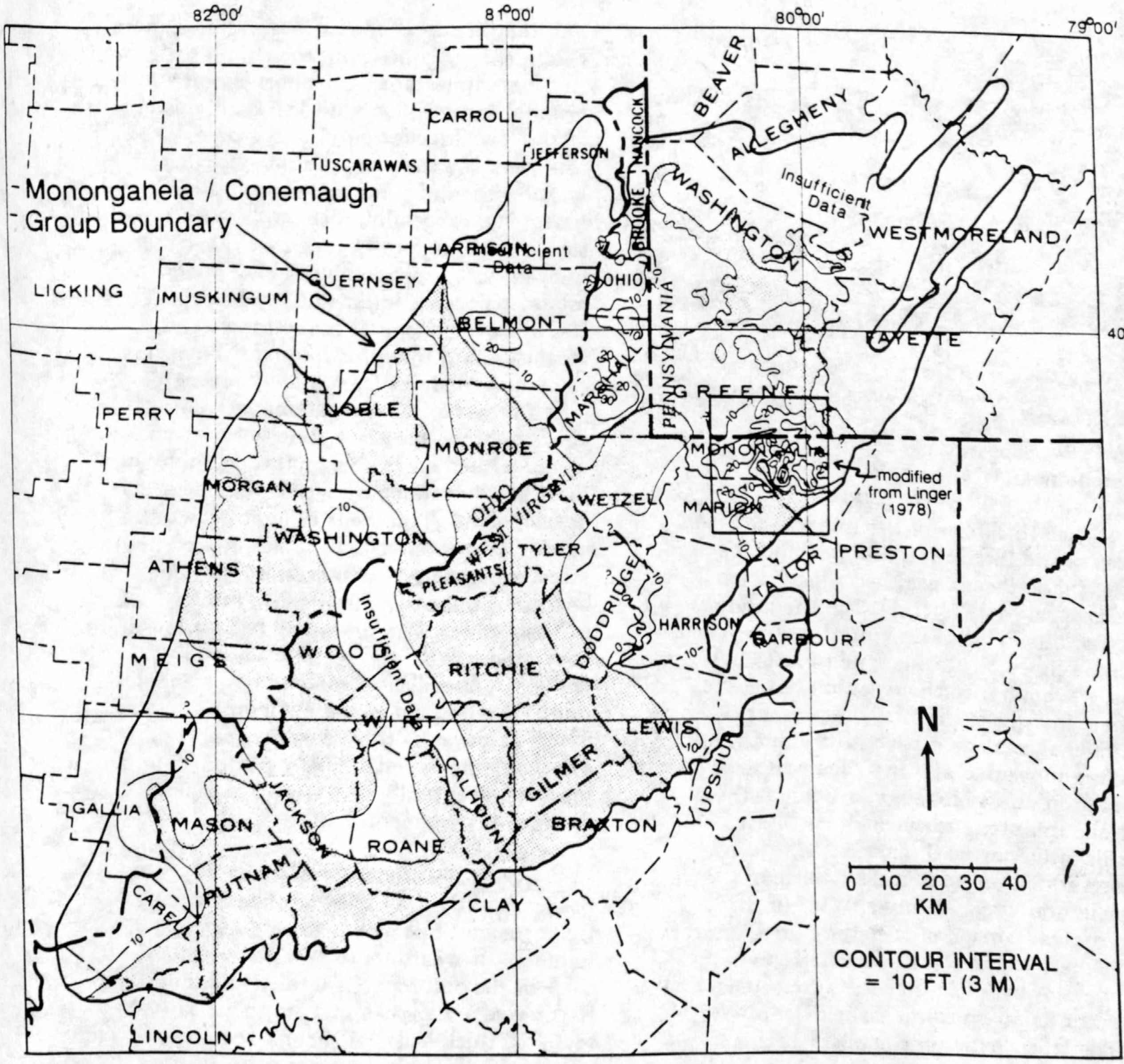

Figure 91 - Redstone limestone isopach map. Shaded areas indicate the location of a siliciclastic bed that, when present, occurs at the Redstone limestone horizon. Figure modified from Linger (1978), Roen and Kreimeyer (1973) and Donaldson (1969).

unstained. In addition, for each of the 23 samples, X-ray mineralogical analyses were performed, and insoluble residue quantities were determined. The weight percentage of insoluble residue in each sample was determined by dissolving carbonates out of weighed, crushed, samples with 10% hydrochloric acid (HCL) and calculating the weight percentage remaining. A petrographic analysis was performed on each thin section, and a few samples were examined by scanning electron microscope (SEM). The sawed faces of samples were polished to further define sedimentary features. Carbon and oxygen isotope studies are currently underway to better define water conditions and paleoclimate at the time of Redstone limestone deposition.

Discussion

Occurrence of the Redstone Limestone

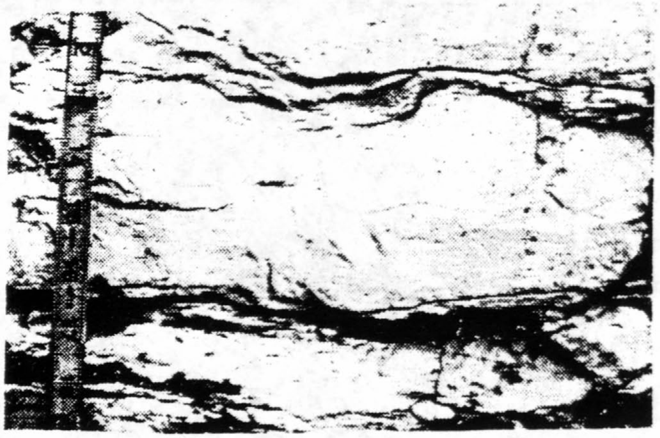

Figure 92 - Exposure of weathered Redstone limestone near Wheeling, WV.

The thickness of the interval between the top of the Pittsburgh coal bed and the base of the Redstone coal bed ranges from 7 to 12 m (22.5 to 38.5 ft) thick in the study area (fig. 88). Within this interval, the Redstone limestone of Platt and Platt (1877) is the predominant lithologic unit, except where a siliciclastic lithology is present. Figure 89 shows the distribution of both the Redstone limestone and the clastic lithology. At one location in Monongalia County, the Redstone limestone reaches almost 12 m (38.5 ft) in thickness (Linger, 1978). It is also very thick (up to 9 m, 29 ft) in parts of Marshall and Ohio Counties, WV. In the south central portion of the study area, data on thickness of the Redstone limestone are lacking. However, the Redstone limestone does occur as far south as Cabell County (fig. 89). The siliciclastic unit persists at this interval through a large portion of southwestern PA (Roen and Kreimeyer, 1973), central WV and eastern OH (Donaldson, 1969). It is interpreted to have formed in an extensive fluvial system which was penecontemporaneous to the Redstone limestone.

Locally, the Redstone limestone thickens and thins rapidly. An average of 2 m (6.4 ft) of clay and shale occur between the top of the Pittsburgh coal bed and the base of the limestone, although the thickness of this interval varies between 1 cm and 4 m (0.03 and 12.8 ft). There appears to be a correlation between the limestone thickness and the thickness of this interval; in general, the thicker this interval, the thinner the Redstone limestone. This pattern may reflect original topography, with thicker limestone in the lows (deeper parts of the paleolake). If this were the case, one would expect to find an indication of subaerial exposure below the limestone, especially in the areas of original topographic highs. In fact, non-calcareous, light-gray to brownish-black clay and/or mudstone occurs imnmediately below the Redstone limestone at many sampling localities, and this unit frequently contains calcareous nodules from a few centimeters to a half meter or more in diameter. This clay/mudstone may be a paleosol horizon, which formed on the original topography just prior to, or contemporaneous with, lake development. Thin beds of light gray to black clay and mudstone sometimes occur interbedded with the Redstone limestone beds, and these are also believed to be paleosol zones. The limestone beds below and above these beds also exhibit paleosol features. Above the Redstone limestone, less than 1.3 m (4.2 ft) of shale and/or underclay occur, followed by the Redstone coal bed. This could be explained by a gradual filling of the lake, eventually giving way to a peat-forming environment.

The Redstone limestone is predominantly gray to light olive gray in a fresh surface, but may also be dark olive gray to black. It weathers to a buff or reddish color in the outcrop. Generally, the limestone is massive and occurs in 1 cm to 1.3 m (0.03 to 4.2 ft) thick beds. Although sedimentary structures can often be seen on weathered outcrops, they are usually more obvious in polished and/or etched, sawed hand specimens. The weathered face of the limestone may exhibit a rounded nodular appearance.

Mineralogy/Petrology of the Redstone Limestone

X-ray mineralogical analyses were performed on 45 Redstone limestone samples: 23 from core #192-30, 6 from core #2425, and the remaining 16 samples from 8 outcrop localities (fig. 88). Results of these analyses indicate that the major constituents are calcite, ankerite, and quartz. Dolomite, chert, pyrite, feldspar and clay minerals occur in minor amounts. The HCl acid insoluble residue, predominantly quartz, ranges from 5-40%, and averages 20%. Petrographic examination revealed that calcite is commonly in the form of micrite, although ankerite also occurs as micrite in several samples. Voids within shell fragments and between grains are commonly filled with ankerite rhombohedra 3-15 um in size. However, sparry calcite and sometimes chert are also found filling voids. Pyrite occurs as cubes, blebs, and large patches up to 5 mm (0.13 in) in the Redstone limestone, sometimes replacing portions of the invertebrate fossils and often associated with plant material. Phosphate is present in the form of bone fragments.

Paleontology

Ostracods are common in most of the samples. They may appear sparsely scattered throughout the rock or concentrated, often with other invertebrate shells, in particular beds. I.G. Sohn (oral communication, 1992) examined the ostracods from several Redstone limestone samples in core # 192-30, and identified Darwinulus sp, which is a definite indicator of a nonmarine environment. Branchiopods (another group of crustaceans), bivalves, small gastropods, spirorbis (tube-dwelling worms), bone fragments, and plant debris are also present in the Redstone limestone. Yochelson (1975) identified the gastropod, Anthracopupa (genus), in Redstone limestone collections from eastern Belmont County, OH. Modern descendants of this genus are land snails (Solem and Yochelson, 1979). Eager (1975) identified very small bivalves of the genus Anthraconaia from Redstone limestone outcrops in eastern Monongalia and northwestern Marshall Counties, WV.

Limestone Facies

Based on megascopic and microscopic examination of the Redstone limestone, six significant carbonate facies have been identified: 1) desiccation breccia limestone, 2) nodular limestone, 3) fossilferous limestone, 4) bioturbated limestone, 5) micritic limestone, and 6) laminated limestone. These facies do not appear to follow any specific sequence (cyclicity) vertically in the Redstone limestone, and the facies tend to interfinger laterally. The dessication breccia facies is the predominant one in core #192-30, constituting almost 50% of the core. It is frequently found in combination with other facies. Following is a description of each of the facies.

1) Desiccation breccia limestone - this facies contains typical paleosol features, such as peds (blocky pellets formed by shrinking and swelling and separated by cracks) and root traces. It may occur in conjunction with any of the other facies, but is most commonly found in combination with the nodular facies. This facies probably formed in very shallow water or subaerially, where alternate wetting and drying caused dessication of the lime muds. Other facies could also have been subaerially exposed to dessication during dry periods - particularly the near-shore facies.

2) Nodular limestone - this facies contains rounded intraclasts .2-1 mm (0.05 to 0.25 in) in size. These clasts appear to be very similar to the surrounding strata and are composed primarily of micritic calcite, although a few clasts are predominantly ankerite. The clasts contain fossils similar in assemblage and size to those fossils in other parts of the Redstone limestone. Some clasts exhibit lamination which is now at an angle to bedding, indicating that they have been ripped up and redeposited. The rounded nature of many of the clasts, together with the fact that shells intermixed with the clasts are broken, suggests reworking of the material. Algal filaments can be detected in a few clasts, within a micritic matrix. These filaments are oriented lengthwise across the clasts, in contrast to the typical radiating outward, with concentric laminae, which distinguishes oncolites. These clasts are probably not oncolites, but rip-up clasts from an algal mat. Voids between the intraclasts are filled with sparry calcite or ankerite, and/or fine grained detrital material. Sometimes the clasts are compressed together suggesting that they were still soft when deposited. In other cases, a calcite or

ankerite spar fills the void surrounding the clast which was created by shrinkage of the clast after deposition. This nodular limestone facies probably accumulated in shallow water, in an environment where the lime mud could be ripped up, rolled and rounded by wave action and currents.

3) Fossiliferous limestone - this facies contains an abundance of fossilized shells. They are usually concentrated in specific beds, and are predominantly molluscs and ostracods. Some of the ostracod shells are nested in groups of 4-5 shells, each slightly smaller than the one before. Plant fragments are common in this facies, and pyrite is often associated with the plant material. Scattered intraclasts occur with broken shell fragments. In some cases, ostracods are the solitary fossil present; they can be found in fossil-rich beds or scattered (sometimes whole) randomly through the sample. The occurence of the gastropod, *Anthracopupa* sp., whose modern descendants are land snails (Solem and Yochelson,1979), suggests a shallow water or shoreline paleoenvironment for this facies. Also, the nested ostracod shells accumulate with gentle wave activity, presumably in shallow water, where progressively smaller shells can wash into larger shells, producing a set of several shells (Sohn, oral communication, 1992).

4) Bioturbated limestone - the rock from this facies has lost part or all of its original bedding, and appears disrupted. Occasionally, individual burrows filled with detrital material can be distinguished. This facies, like the fossiliferous facies, probably represents shallow water sedimentation but deeper than the first two facies.

5) Micritic limestone - this facies contains predominantly calcite and ankerite grains less than 30 microns in diameter. A few scattered ostracod shells may be present. There is usually no apparent bedding or structure in this limestone. The lack of lamination could mean any one of a number of things: 1) that seasonally differentiated particulate matter was not settling through a stagnant, stratified water mass; 2) that bottom dwellers homogenized the sediment; 3) that bottom current activity disturbed the sediment; or 4) that degradation of organic matter produced abundant gas bubbles (Kelts and Hsu, 1978, p.309).

6) Laminated limestone - there are two different types of lamination, which differ in mineralogy and grain size. The first type is a clastic/carbonate laminated facies. In this case, the laminae are created by interbedding of quartz (up to 0.1 mm in diameter), clay minerals (illite and kaolinite), plant material, and calcite and/or ankerite. These clastic/carbonate laminae suggest a periodic input of clastic sediments and plant material, possibly during storms or rainy periods. The other type of lamination is more subtle, and is created by interbedding of micritic carbonate minerals. SEM studies indicate that calcite occurs throughout the sample, but laminae are created by higher concentrations of rhombohedral ankerite in particular beds. There are very few fossil fragments and very little quartz or clay minerals in this second type of laminated facies. These laminae may indicate seasonal changes in the climate or lake conditions, and/or could be algal in origin.

Summary

Stratigraphic and sedimentological analyses indicate that the Redstone limestone (Platt and Platt, 1877) probably formed in a shallow, widespread lake or series of lakes. The variability in limestone thickness throughout the basin is most likely related to the original topography on which the lake represented by the Redstone limestone formed - the thickest limestone is generally over the topographic lows and the thinnest limestone over the highs. In some areas, fluvial deposition and/or channeling reduce the limestone thickness. Even in localities where the limestone is missing between the Pittsburgh and Redstone coal beds, the clays and shales are usually calcareous or contain calcareous nodules, suggesting that sediment influx into the lake was too great to allow limestone to develop in these localities. The fact that the Redstone limestone reaches 12 m (38.5 ft) in thickness indicates that the lake was probably stable, at least in some areas, for a long time. The absence of evaporitic minerals such as gypsum and halite, the paucity of clay minerals, and the apparent sparsity of stromatolitic structures suggest that the lake was not a playa or brine lake, but in fact a freshwater lake, with an influx of fresh water which was greater than the amount of water leaving the lake by evaporation. Rainfall varied, probably seasonally, as suggested by the predominance of desiccation features in the Redstone limestone.

Acknowledgements

The author wishes to thank C. Blaine Cecil for his interest and support, Consolidation Coal Company for giving me the cores for study, Frank Dulong for X-ray diffraction analyses of limestone samples, Nick Fedorko for collecting the cores and for providing me with stratigraphic data, Jim Pontolillo for preparation of samples and for running insoluble residue analyses, Leslie Ruppert for her scanning electron microscope (SEM) expertise, Ron Stanton for the use of his petrographic equipment, Stanley Schweinfurth, Elliott Spiker, and Christopher Wnuk for excellent reviews of this manuscript, Lorrin Ferdinand and Dixon Van Hofwegen for field assistance, and Bill Edmunds, Dick Smosna, and Suzanne Weedman for their suggestions and interest in this work.

Stratigraphic variation in bulk sample mineralogy of Pennsylvanian underclays from the central Appalachian basin.

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Introduction

A previous study in the central Appalachian basin determined that there were at least two major climate changes which affected sedimentation during the Late Paleozoic (Cecil and others, 1985). These changes are indicated stratigraphically by variations in coal quality and lithology. Similar inferences in climate changes were determined from variation in Pennsylvanian coal-swamp vegetation (Phillips and others, 1985).

Coal quality data for thirty-four minable coal beds from the Lower to Upper Pennsylvanian Series show systematic changes in both sulfur content and ash yield. The major stratigraphic change in total sulfur content occurs between the Kanawha Formation (lower Middle Pennsylvanian) and the Allegheny Formation (upper Middle Pennsylvanian) (fig. 93). Analysis of variances and means with an F- and T-test, respectively, at the 0.05 level of significance indicates the mean total sulfur values are lowest and equal between the Pocahontas, New River and Kanawha Formations. The mean sulfur contents of the coal beds in the Pocahontas, New River and Kanawha Formations are less than the Allegheny Formation which is equal to the Conemaugh Formation. The Monongahela Formation has the highest mean total sulfur content (table 6). Similar stratigraphic trends are also seen for ash values. Like sulfur content, there is also a change in mean ash-yield between the Middle and Upper Pennsylvanian (fig. 94). The Pocahontas and New River Formations are statistically equal and have the lowest ash-yield, followed by the Kanawha Formation. The Allegheny and Monongahela Formations are equal in ash-yield. Coal samples from the Conemaugh Formation contain the highest ash value (table 7).

Major changes in lithologies coincide with major changes in the occurrence and quality of minable coal beds. Of particular interest for paleoclimatic interpretation is the presence and

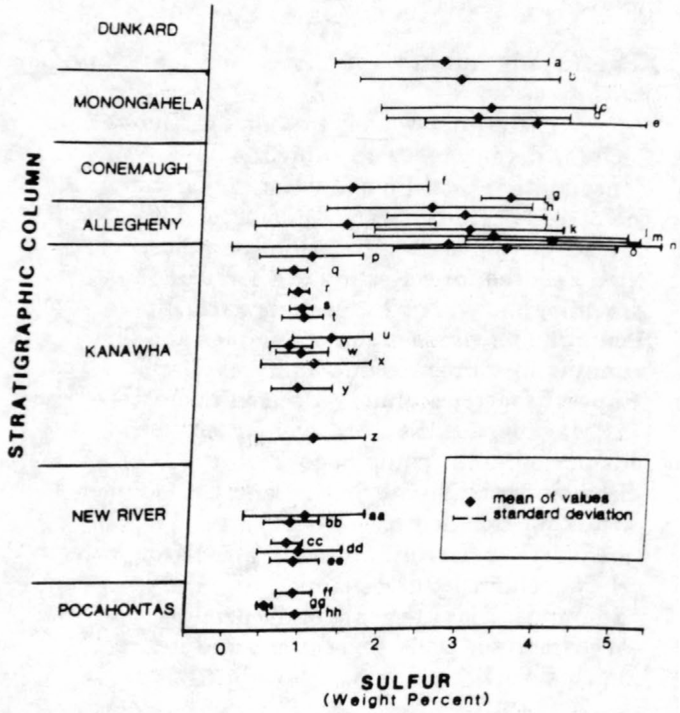

Figure 93 - Stratigraphic distribution of mean sulfur values for 34 minable coal beds from the Pennsylvanian Series of the Central Appalachian Basin (from Cecil et al., 1985)

relative abundance of minerals and lithologies that may indicate geochemical conditions of weathering, soil formation, sedimentation, and early digenesis. These include pyrite, calcite, siderite, kaolinite-rich clay deposits, quartz arenites and iron oxide in red beds. Freshwater limestones are abundant and common in the Upper and upper Middle Pennsylvanian and absent in the lower Middle and Lower Pennsylvanian (fig. 95). Caliche and red beds, indicators of drier conditions, are most abundant in the Allegheny and Conemaugh Formations. Pyrite is very common in the Upper and upper Middle Pennsylvanian, whereas siderite is rare. In contrast, pyrite is rare in the lower Middle and Lower Pennsylvanian, whereas siderite is common.

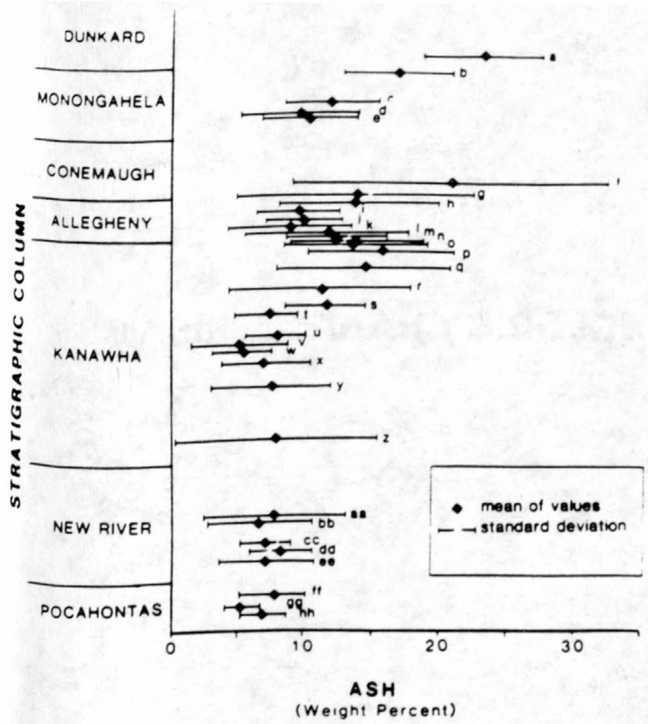

Figure 94 - Stratigraphic distribution of the mean ash values for 34 minable coal beds from the Pennsylvanian Series of the Appalachian Basin (modified from Cecil et al., 1985).

Based on the presence of these minerals and lithologies Cecil and others (1985), concluded that the changes in paleoclimate that affected geochemical conditions of sedimentation during the Late Paleozoic in the central Appalachian basin are reflected in the stratigraphic occurrence and quality of coal beds and on rock lithologies (fig. 96): 1) the Upper Mississippian climate was dry-seasonal to dry. Many rocks are of continental origin, contain nodular calcareous paleosols, and few minable coal beds; 2) the climate of the Lower and lower Middle Pennsylvanian was ever-wet tropical (that is, rain fall was evenly distributed through out the year). Coal beds from these rocks are thick, low in ash and developed from domed, ombrogenous peat swamps analogous to those of present coastal environments in Malasia and Indonesia; and 3) the climate of the upper Middle and Upper Pennsylvanian became more seasonal and perhaps drier. The driest climate occurred during the deposition of the Conemaugh Formation. Coal beds in the upper Middle and Upper Series were derived from planer, topogenous peat swamps.

Mineralogical Study of Underclays

This study of the bulk mineralogy of underclay from the central Appalachian basin was undertaken to determine if there is a systematic stratigraphic relationship between the mineralogical variation of the underclay and the geochemical conditions of sedimentation and peat formation. Was the mineralogical composition of the underclay influenced by allocyclic events such as climate changes during the Pennsylvanian (Cecil, 1986 and 1990)? The term underclay refers to the stratum directly under the coal bed. It is generally argillaceous, gray, non-bedded, and sometimes rooted, rooting indicates it was a soil with plants growing on it. Underclay thicknesses can range from a few centimeters to several meters. Although, the concept of underclay having pedogenic origin is not new, it is still controversial (Huddle and Paterson, 1961; Rimmer and Eberl, 1982; and Hughes and others, 1987).

In a reconnaissance study of the stratigraphic relationship among bulk sample mineralogy of underclay, we sampled the upper 15 to 30 cm of the bed. A representative portion of the sample was ground to minus 100 mesh and then low-temperature ashed. The ash was ground to minus 200 mesh and pressed into pellets for semiquantitative determination of mineral percentages (Hosterman and Dulong, 1989). The analytical method used provides semiquantitative estimates of the major and minor phases within the inherent limits of X-ray diffraction (Hosterman and Whitlow, 1983). The relative percentages of minerals, illite crystallinity (measurement of 001 peak width at half maximum intensity), as well as a qualitative kaolinite crystallinity index (inter-sample comparison of the relative broadness of the 001, 7 angstrom peak) were measured on all samples.

The geographic and stratigraphic distribution of the samples are shown in figure 97. Underclay associated with coal beds in the Lower through Upper Pennsylvanian Series was sampled. Half, 17 of 33, of the underclay samples from the Monongahela Formation are from the Pittsburgh coal bed in West Virginia, Maryland and Ohio. The Anderson coal bed (Conemaugh Formation) underclay samples are from eastern Ohio. Samples from below the Upper Bakerstown, Lower Freeport, Middle Kittanning and Lower Kittanning coal beds (Allegheny Formation) are from western Maryland. Seventeen of twenty Upper Freeport underclay samples are from southwestern Pennsylvania, one is from eastern Ohio and two are from western Maryland. Seven Hernshaw-Fireclay coal bed (Kanawha Formation) underclay samples.

LITHOLOGIC CHARACTERISTICS

Series	Formation	NONMARINE LIMESTONES	CALICHE	SIDERITE	PYRITE	FLINT CLAYS	RED BEDS
UPPER PENNSYLVANIAN	MONONGAHELA	A	C	R	C	R	C-R
	CONEMAUGH	C	A	R	C-R	R	A
MIDDLE PENNSYLVANIAN	ALLEGHENY	C	C R	A-C	C	A	R
	KANAWHA	O	O	C	R	O	O
LOWER PENNSYLVANIAN	NEW RIVER	O	O	C	R	O	O
	POCAHONTAS	O	O	C	R	O	O

MODIFIED FROM C B CECIL, AND OTHERS 1985

Figure 95) Lithologic characteristics of the Pennsylvanian Series of the Appalachian Basin (A=abundant, C=common, R=rare, and O=absent). Modified from Cecil and others, 1985.

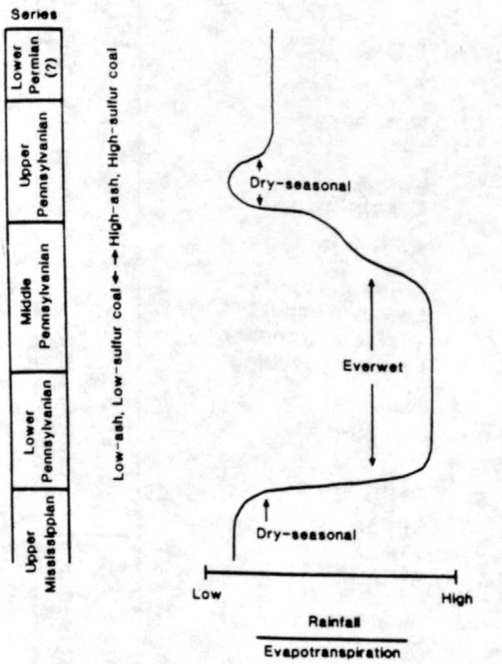

Figure 96 - Ratio of rainfall to evapotranspiration, as interpreted from lithologies and coal quality in the Pennsylvanian of the Central Appalachian Basin (from Cecil et al., 1985)

are from eastern Kentucky and southwestern West Virginia. Seventeen Kanawha Formation coal beds (Coalberg down to the Gilbert) were collected from the Bolt Mountain section near Bolt, West Virginia. The New River Formation underclay samples are from southern West Virginia. Four Pocahontas No. 6, six Pocahontas No. 3 and three Pocahontas No. 2 underclay samples (Pocahontas Formation) are from southern West Virginia and adjacent southwestern Virginia. All samples came from mappable coal beds most of which are economically important.

Results

The crystallinity of the illite is poor (as measured as peak width at half height in degrees two-theta of the 001, 10 angstrom peak) in both the Upper Bakerstown and the Anderson (Conemaugh Formation) underclays. The illite is mostly of the 1M type, and also the underclay contains some mixed-layer material (fig. 98). The Anderson underclay contains very little kaolinite (001 peak at 12 degrees two-theta). Similar poorly crystalline illite also is present in the Upper Freeport and Lower Kittanning (Allegheny Formation) underclay samples (Fig. 99). The underclay of the Upper Freeport contains calcite (29.4 degrees two-theta) and pyrite (28.5 and 33 degrees two-theta). Samples of the underclay from the Fireclay and Lower Bens Creek, Kanawha Formation, contain well crystallized illite, of the 2M type. Chlorite is present in the Lower Bens Creek underclay sample, as it is in the majority of the Kanawha underclays (fig. 100). Chlorite is also present in the Pocahontas No. 3 underclay sample (fig. 101).

The average amounts of kaolinite, pyrite and carbonate phases are greater in the Allegheny Formation than in the Kanawha Formation (table 8). The underclay samples of the Kanawha Formation contain more illite and chlorite than other formational groups of samples and the illite is well crystallized. The average peak width is smaller, and therefore the illite is said to be more crystallized, in the Kanawha Formation than in samples from the Allegheny Formation. The underclay samples from the Pocahontas Formation resemble underclays of the Kanawha Formation in their kaolinite to chlorite ratio, paucity of pyrite, and degree of illite crystallinity.

Conclusions

Three diffraction patterns can be used to summarize bulk mineralogy in this study (fig. 102). On the basis of clay mineral contents (table 8) the samples separate into two groups: 1) Those which are composed mostly of soil type (poorly crystallized) kaolinite with a subequal to minor illite of the poorly crystalline variety (soil-type suite of Hughes and others, 1987); and 2) Those which containing mostly well-crystallized illite with lesser, yet substantial, kaolinite and chlorite (shale-type suite of Hughes and others, 1987). The mineralogy of the first group indicates in situ alteration, most likely by post-depositional acid leaching concomitantly with peat accumulation. The underclay samples of the Allegheny and Conemaugh Formations, upper Middle and lower Upper Pennsylvanian Series respectively, make up this first group. The mineralogy of the second group comprised of the lower Middle and Lower Pennsylvanian Series underclay samples, indicates little if any alteration from the original shale-type sediment.

We conclude that there are stratigraphically different mineralogic suites of underclays from the central Appalachian basin. The differences in mineralogy tend to parallel, and probably are

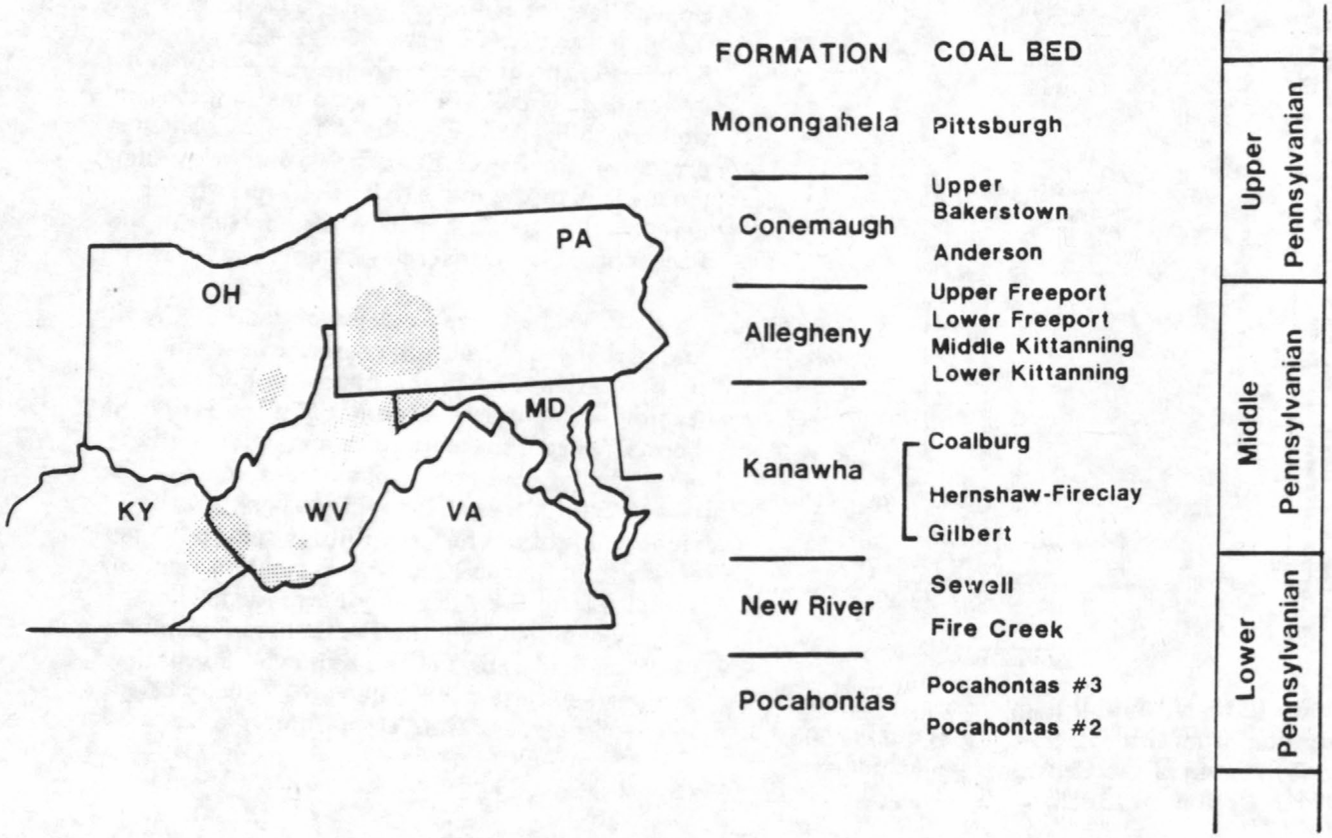

Figure 97 - Spatial and stratigraphic distribution of samples used in this study.

caused by, the major change in paleoclimate occurring between the upper and lower Middle Pennsylvanian Series.

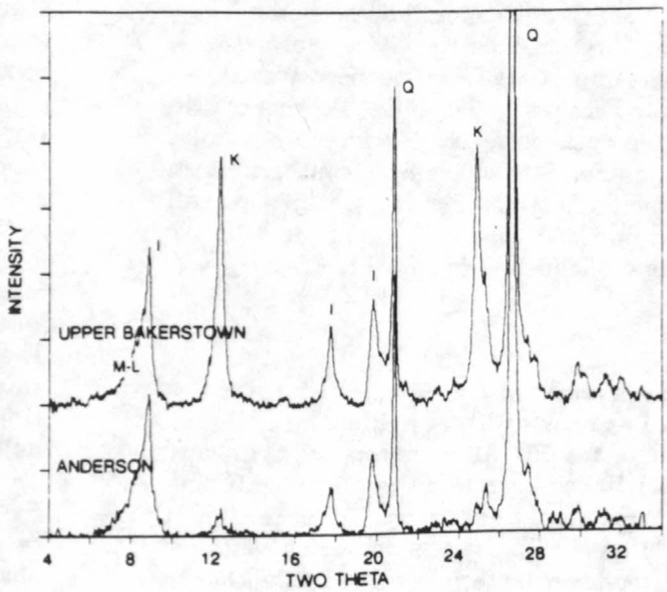

Figure 98 - X-ray diffractograms of underclay samples from the Conemaugh Group. Ch = chlorite, M-L = mixed-layer, I = illite, K = kaolinite, Q = quartz, P = pyrite, and C = calcite.

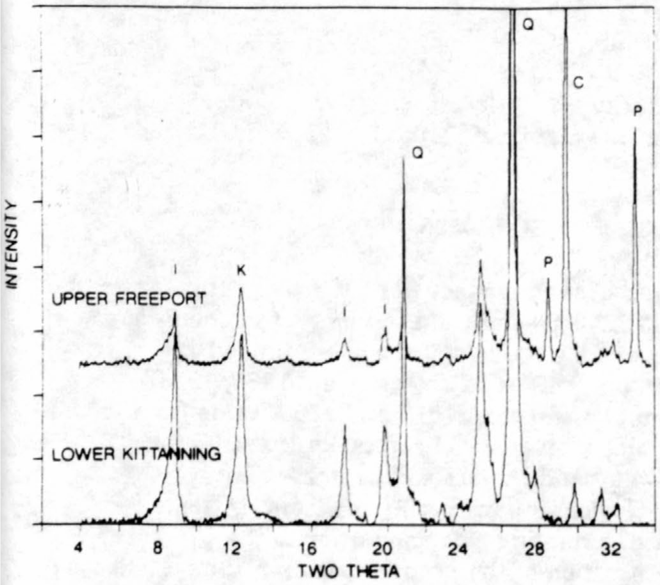

Figure 99 - X-ray diffractograms of Allegheny Formation underclay samples (abbreviations as in fig. 98).

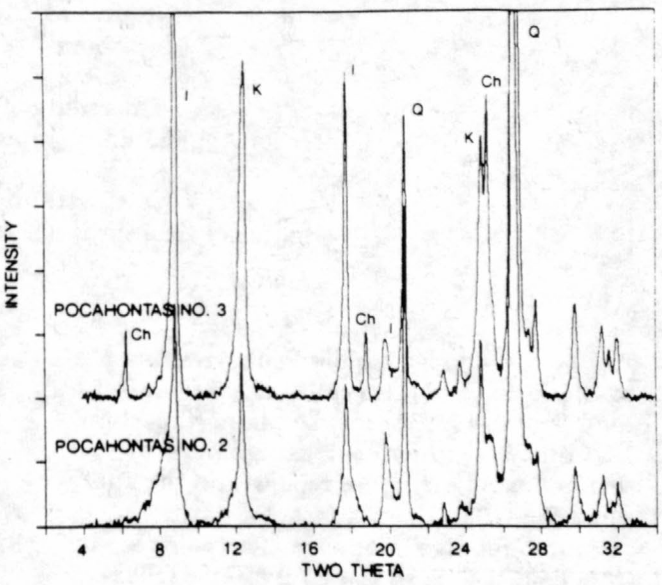

Figure 101 - X-ray diffractograms for Pocahontas Formation underclay samples (abbreviations as in fig. 98)

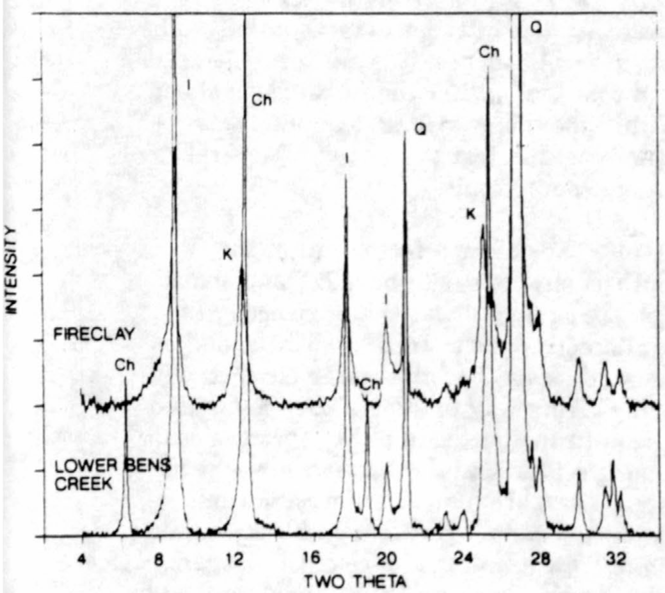

Figure 100 - X-ray diffractograms for Kanawha Formation underclay samples (abbreviations as in fig. 98).

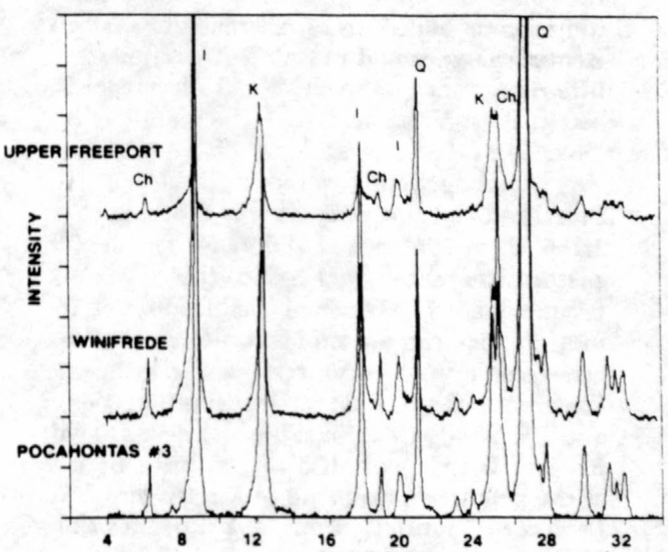

Figure 102 - Summary X-ray diffractograms of representative underclay samples from the Allegheny, Kanawha and Pocahontas Formations (abbreviations as in fig. 98).

No. 5 Block Coal Bed, Northeastern Kentucky

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Introduction

The upper Middle Pennsylvanian No. 5 Block coal bed (also called the Richardson and Princess No. 5) coal bed in northeastern Kentucky is a thick coal, being mined to extinction on mountain tops within the study area (fig. 103). Its stratigraphic position position, relative to other major coals seen on this field trip, is shown on figure 86 (Blake, this volume). The high volatile bituminous (hvCb in north, increasing to hvAb to south) No. 5 Block coal bed is dominated by dull lithotypes. One of the lithotypes in particular is notable in having large quantities of inertodetrinite in association with fine-sand size detritus. The lithotype is a durain but can be distinguished from other durains in the same coal by its granular texture. While the scope of our study is restricted to eastern Kentucky, we would expect that similar lithotypes continue in the No. 5 Block coal bed in West Virginia.

Production of coal from the No. 5 Block and correlative coals was reported to be 3.195 Mt in 1990 out of 130.97 Mt for all of eastern Kentucky. Over half of the production, 1.766 Mt from 7 mines (out of 16 total mines), came from Martin County. The estimated original resources for the Princess (Lawrence, Boyd, Carter, Greenup Counties) and Big Sandy (Martin, Pike, Johnson, Floyd) Reserve Districts are 635 Mt, with the bulk of the resource base being in Martin and Lawrence Counties (Brant, 1982; Brant et al., 1983).

Materials

The No. 5 Block coal bed was collected from several cores and surface mine exposures in northeastern Kentucky (fig. 103). Portions of the sampling was done as part of a US Geological Survey funded sampling effort (Currens et al., 1985, 1986) and two samples (1013 and 1050) came from a joint University of Kentucky Institute for Mining and Minerals Research - Kentucky Geological Survey coring program. In some cases sampling was done prior to the recognition of the unique nature of the argillaceous durain. Where possible those mines were re-sampled, although in the surface mines it was never possible to sample at the exact site.

Discussion

Among eastern Kentucky coals, the No. 5 Block is a relatively thick coal, exceeding 4 m at some sites. Generally split into several benches, it is best considered to be a coal zone rather than a single coal bed. Within the mines visited, sampling was usually within just one or two of several closely spaced coals.

The common factor within the southern sections (site no. 3227 and south on figs. 103 and 104) was the presence of the argillaceous durain. In the southern end of the study area the lithotype is distinctively sandy giving way to a more coal-dominated durain in the north. In either case the coal retains a fine granular appearance which sets it apart from more common durains within the same coal. Certain blocks mimic the microscopic appearance of the Upper Elkhorn No. 3 splint (durain) as described by Thiessen and others (1931, p. 18):

"The mass of the coal...is irregular in structure and layering...and in some horizons

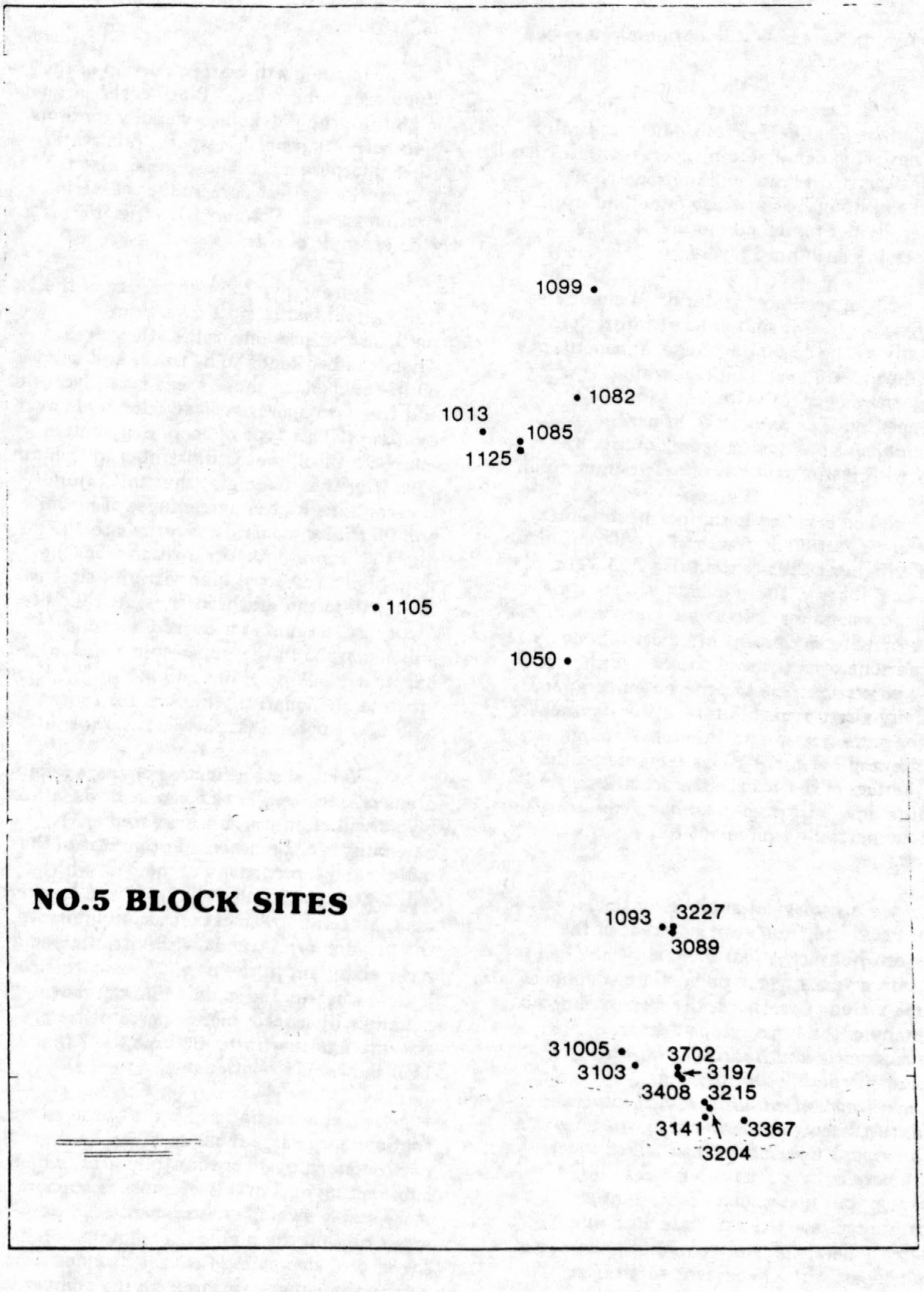

Figure 103 - Sample locations in the Eastern Kentucky coal field.

is much distorted, as if the original mass had been kneaded."

Also common to several of the sections is a cannel horizon. Microscopically the cannel is dominated by spores with lesser quantities of vitrinite and inertodetrinite. Boghead lithotypes are also found but are generally not mined due to the difficulty of processing such hard layers.

The argillaceous durain is dominated by fragments of fusinite and semifusinite, usually in the 10-50 µm range. Miospores in the durain are usually intact while megaspores generally appear to be fragmented. The associated mineral assemblages are dominated by quartz and clay with lesser amounts of a Ti-rich mineral.

The coal bed is thinner in the north (see cross-section on figure 104) where it also has a higher sulfur content (fig. 105). The ash content (fig. 106) tends to be relatively high at most sites, particularly in the south where high-ash durains are most commonly found. The thickening of the coal to the south may be attributable to penecontemporaneous activity along growth faults of the Kentucky River fault system and the Rome Trough. Horne and Ferm (1978) have discussed the thickening of the Breathitt Formation to the south. Blake (this volume) has done likewise for the partially equivalent Kanawha Formation.

Palynological analyses of the No. 5 Block coal bed from selected areas of the Eastern Kentucky Coal Field show the bed to contain a very diverse palynoflora. Common forms include *Lycospora*, the dispersed spore of many of the large lycopsid trees, *Densosporites* and *Radiizonates*, which represent small lycopsids, and *Punctatisporites minutus, Punctatosporites minutus, Laevigatosporites globosus* and *Thymospora pseudothiessenii*, all of which were bore by tree ferns. Somewhat less common, but nonetheless consistently represented, are *Granulatisporites*, and related trilete, sphaerotriangular genera, e.g. *Leiotriletes*, which represent small ferns, *Laevigatosporites* spp., representing calamites, and *Florinites*, which is cordaite pollen.

In the northwestern portion of the study area, where the 5 Block coal bed tends to be thin, the palynoflora usually contains high percentages of *Lycospora*, relative to other miospore taxa. These areas also tend to contain the highest percentages of vitrinite group macerals. Column 1013 (fig. 107) is a good example of this.

In contrast, the palynoflora of the No. 5 Block coal bed is most diverse in southeastern portions of the study area where the bed tends to be thick, and often is multi-benched. In these areas small lycopsid and tree fern spores typically dominate over *Lycospora*. This type of spore distribution in the No. 5 Block coal is illustrated in Column 3702 (fig. 108). Note also that this column also contains higher percentages of liptinite and inertinite macerals. Nearby site 1093 (fig. 109) represents a transition from the thin, high *Lycospora*, high vitrinite sites in the north to the southern sites. At that site *Lycospora* accounts for over 70% of the palynomorphs in the upper split which also has nearly 80% vitrinite and 3% sulfur. High vitrinite and relatively high sulfur contents tend to occur together in No. 5 Block splits.

Vertical stratification of spore taxa in areas of thick coal is also common. Basal coal layers will often contain increased percentages of *Lycospora*, in contrast to the middle and top portions of the bed, which will contain more small lycopsid and tree fern taxa. Maceral stratification usually mirrors the palynomorph trends with vitrinite being more abundant in the basal, *Lycospora*-rich layers, whereas the middle and top parts usually will contain higher percentages of liptinite and inertinite. Column 3197 (fig. 110) shows this relationship.

One explanation for the observed spatial change in palynoflora may be that northwestern parts of the study area, which subsided more slowly, were able to support more stable swamp environments. These areas contain the highest percentages of *Lycospora*, suggesting that the swamps were essentially super-saturated (to the point of being covered with water) a majority of the

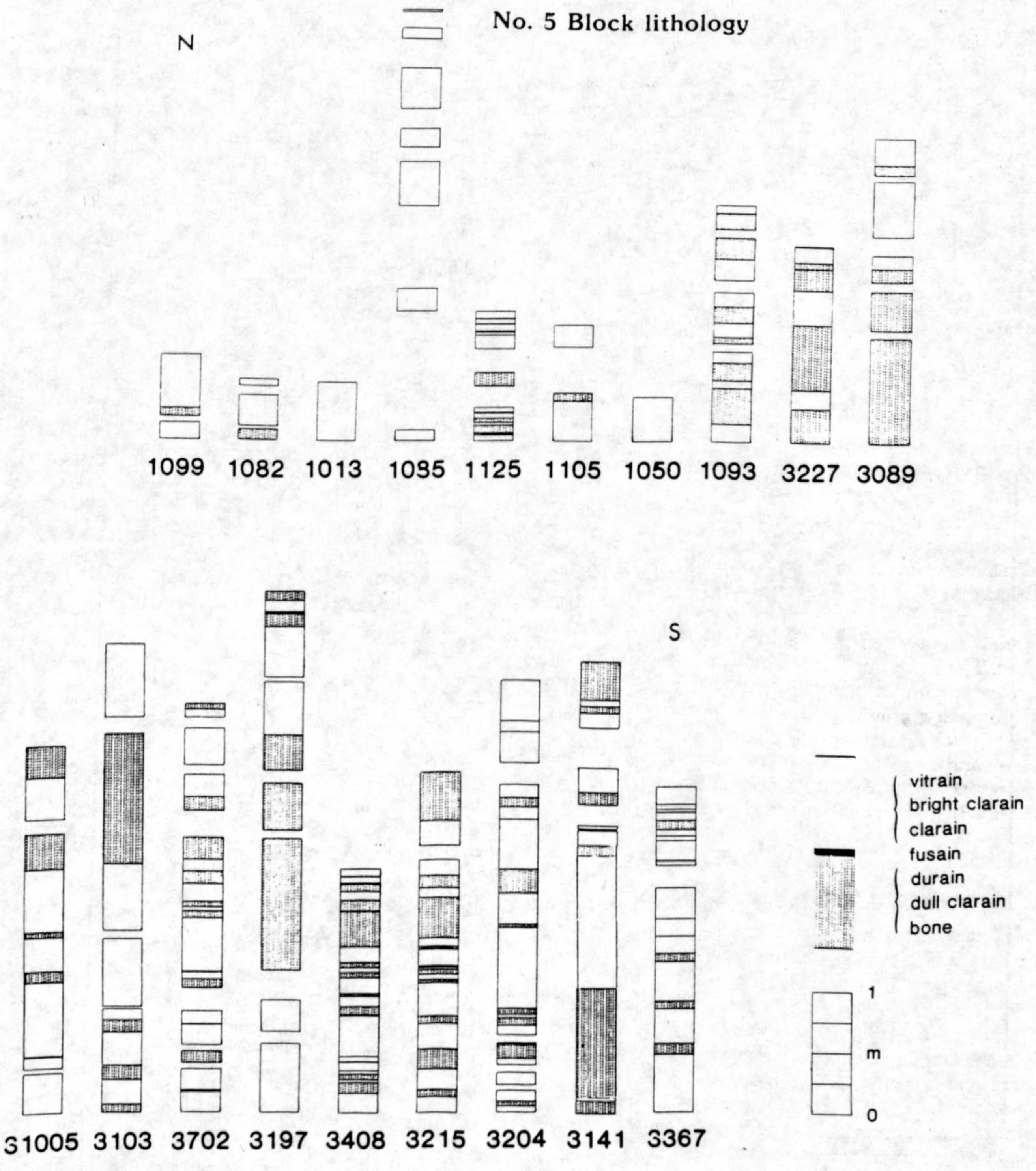

Figure 104 - Lithotype distribution in the study area for northern (top) and southern (bottom) sample locations.

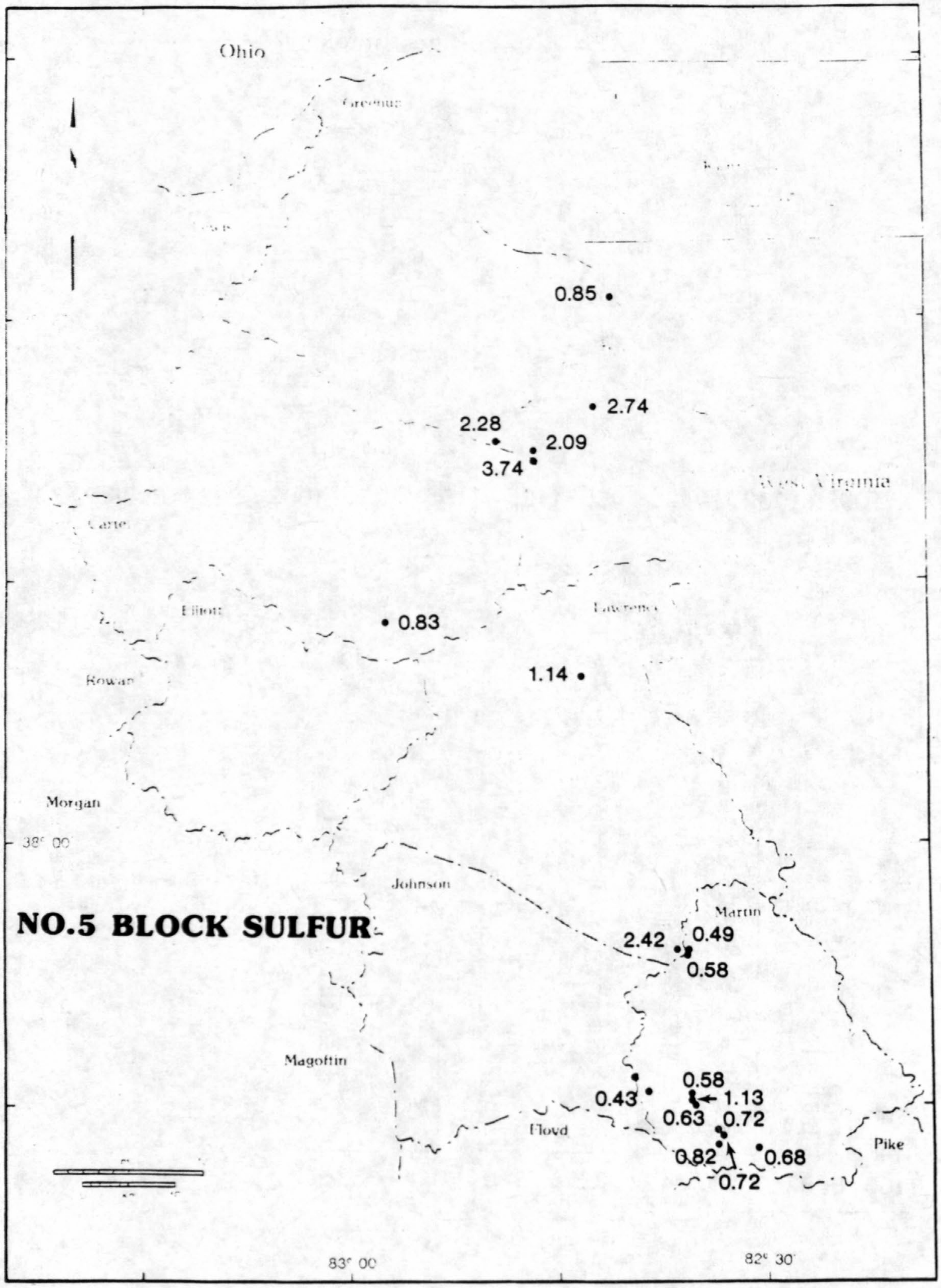

Figure 105 - Total sulfur content distribution in the study area.

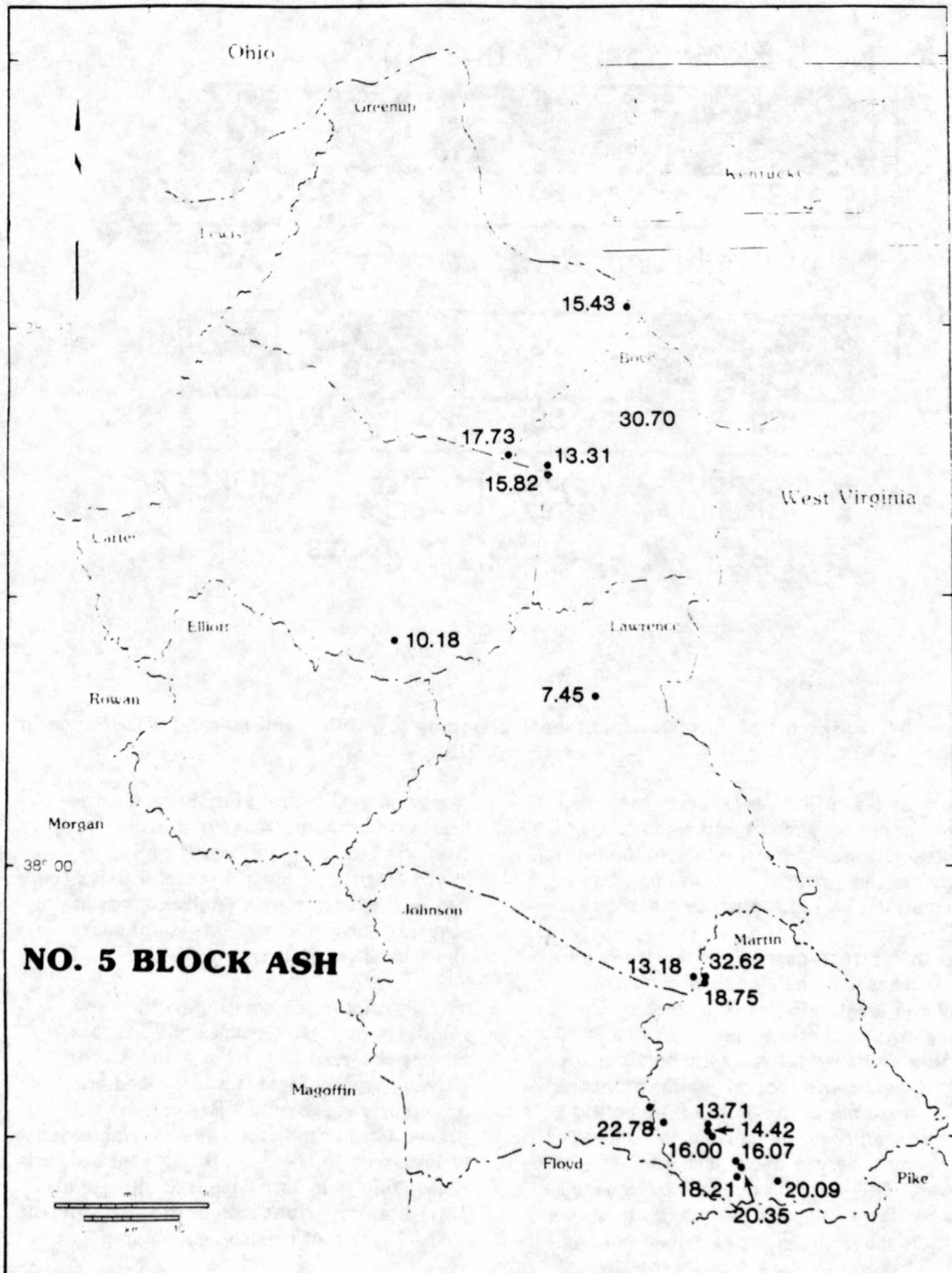

Figure 106 - Ash yield distribution in the study area.

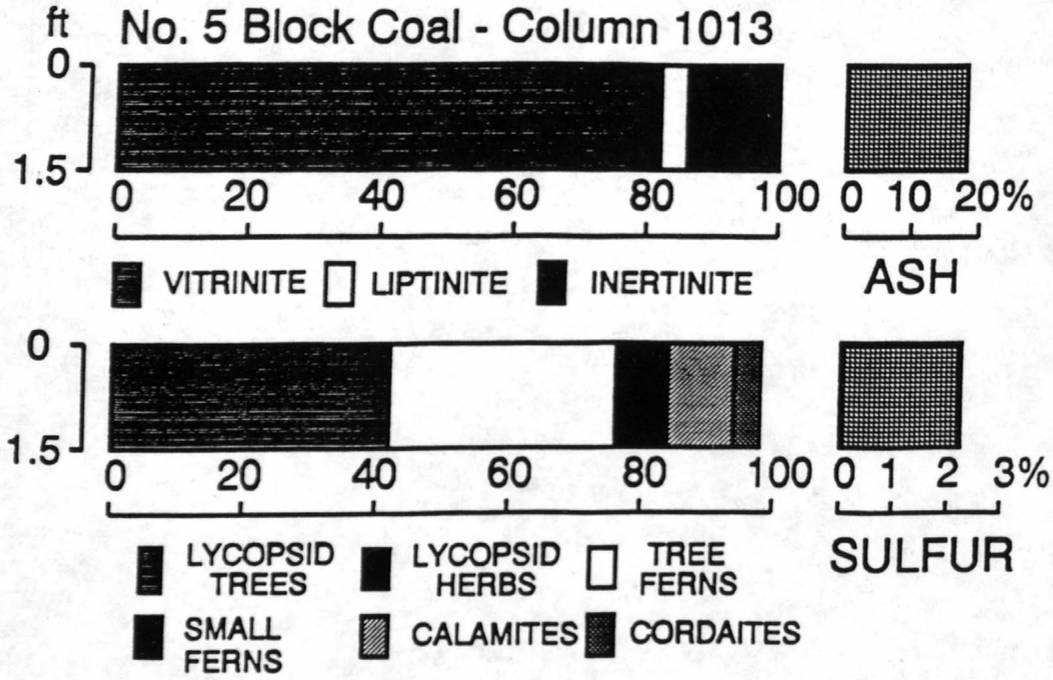

Figure 107 - Palynomorph distribution, grouped according to affinity, and maceral distribution in column 1013

time, as the lycopsid trees favored this type of environment for growth and reproduction. A consistent water cover would also inhibit oxidation of the surficial peat and promote the formation of a high vitrinite content coal.

In contrast, portions of the study area to the southeast probably subsided more rapidly and erratically, as evidenced by the increase in interval thickness and number of coal splits and associated clastic partings. In this area, peat swamp development probably was also unstable to the point of promoting a diverse swamp flora, in contrast to a more uniform one observed in areas to the northwest. The lack of a dominant lycopsid tree palynoflora suggests that peat substrates were often exposed, perhaps for long periods of time. This would allow smaller-statured, and perhaps less edaphically specific plants, namely small lycopsids and tree ferns, to dominate these areas. An inconsistent water cover is further supported by increased percentages of liptinite and inertinite group macerals in this area. Furthermore, the increased amount of clastic partings suggests frequent inundation of peat-forming environments; as such, the No. 5 Block coal bed in this area may actually consist of several peat-forming events, some more short-lived than others.

Another interesting point is the relatively high percentages of *Torispora securis* observed in the No. 5 Block. This palynomorph possesses a thickened end that apparently represents a desiccation-prevention mechanism. Given the abundance of *Torispora* in the No. 5 Block coal bed, it is reasonable to hypothesize that climate also played an important role in the development of the No. 5 Block paleo-peat swamp.

Conclusions

1. In the region of most intense mining, primarily Martin County, the No. 5 Block coal bed is typically low in sulfur content,

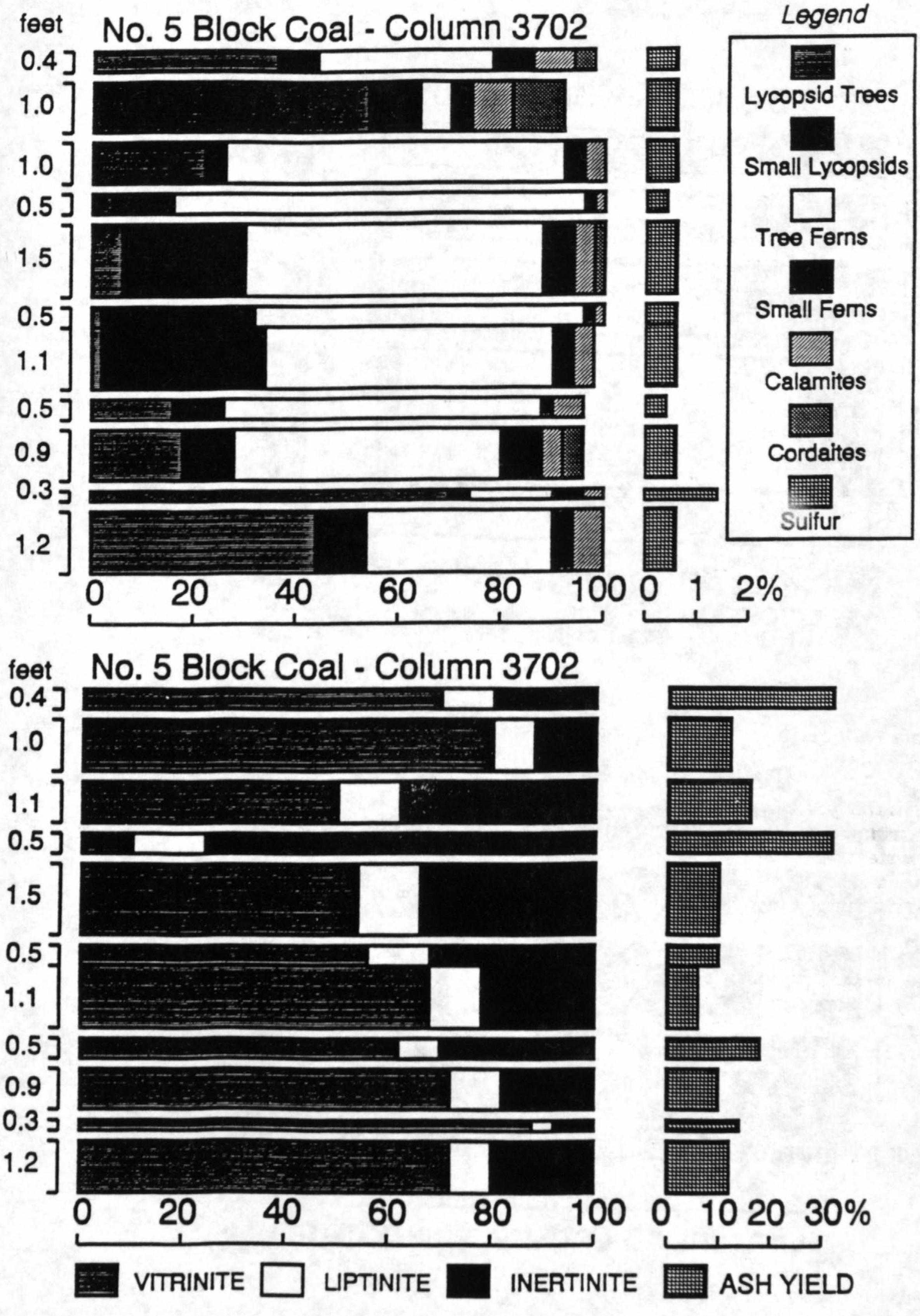

Figure 108 - Palynomorph distribution (top), group according to affinity, and maceral distribution (bottom) in column 3702

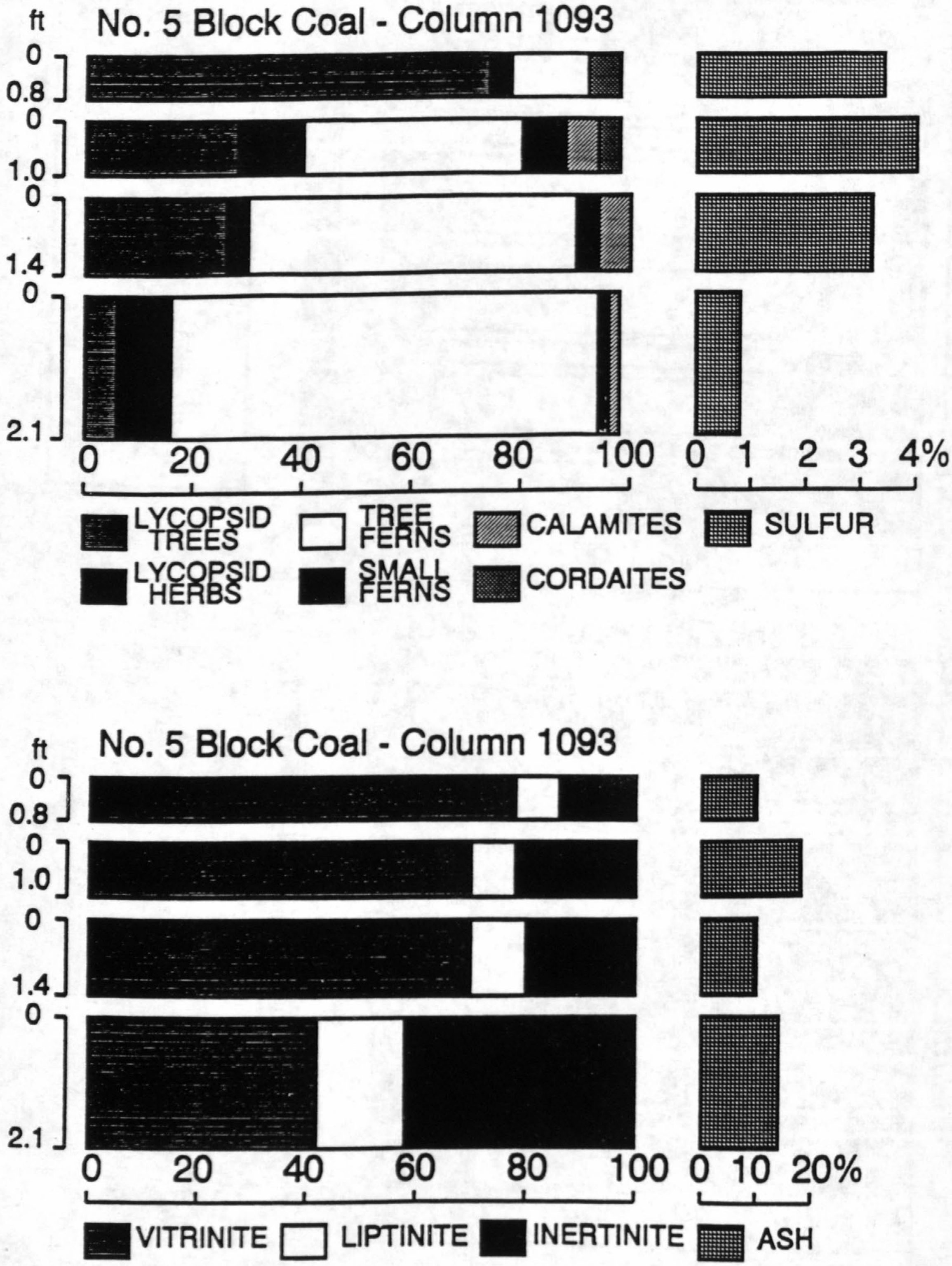

Figure 109 - Palynomorph distribution (top), grouped according to affinity, and maceral distribution (bottom) in column 1093.

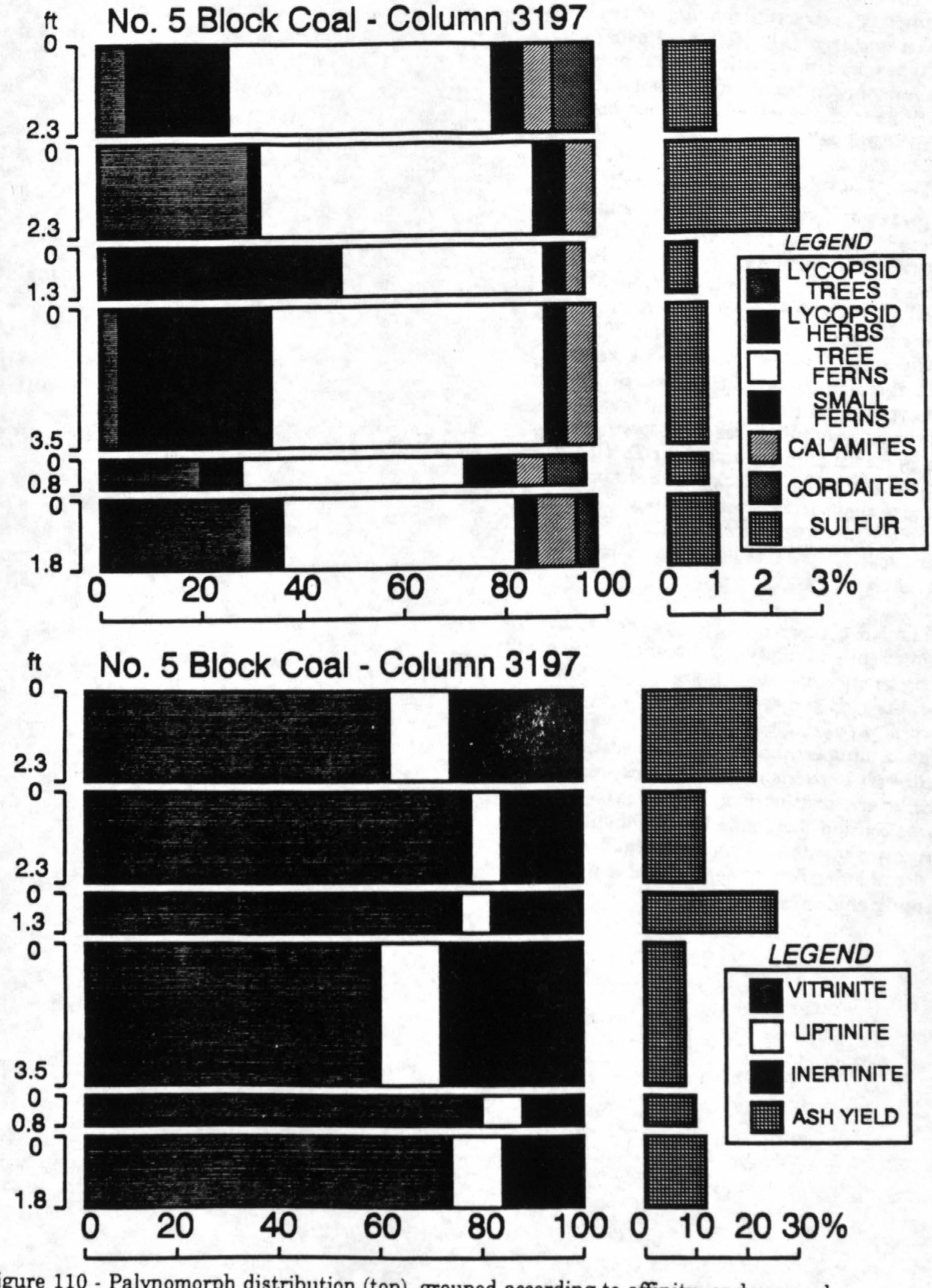

Figure 110 - Palynomorph distribution (top), grouped according to affinity, and maceral distribution (bottom) in column 3197.

making it an attractive compliance resource. in the same area dull, often high ash, lithotypes predominate. Inertinite contents are commonly in excess of 30% (mmf). The coal thins to the north where it has high vitrinite and sulfur content.

2. Representatives of large and small lycopsids and ferns (both tree-like and small varieties) dominate the No. 5 Block coal bed palynoflora. Calamite spores and cordaite pollen also occur but are less abundant. Small lycopsid (*Densosporites* spp.) and tree fern (e.g. *Punctatisporites minutus*, *Laevigatosporites globosus*) spore taxa are most abundant in dull lithotypes. Bright lithotypes contain higher percentages of arboreous lycopsid spores (*Lycospora* spp.). The presence of fairly abundant *Torispora securis* in the No. 5 Block coal bed may indicate that peat formed in a seasonally dry (?) paleoclimate. *Torispora* is a tree fern spore specially adapted for desiccation prevention.

3. The No. 5 Block coal bed thickens rather dramatically in a NW-SE direction. Thin coal along the northwestern margin tends to be vitrinite-rich and contains abundant *Lycospora*, perhaps reflecting relatively stable peat-forming conditions. Thicker coal to the southeast contains more inertinite, high ash coal layers, and inorganic partings. Spore floras contain more small lycopsid and tree fern components and are temporally variable, perhaps indicating a more unstable peat-forming environment.

STRATIGRAPHY AND DEPOSITIONAL ENVIRONMENTS OF THE KANAWHA FORMATION (MIDDLE PENNSYLVANIAN), SOUTHERN WEST VIRGINIA

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Marshall University

Stratigraphic Framework

The Kanawha Formation represents about 80% of the Middle Pennsylvanian Series in West Virginia (Arndt, 1979; Fig.112). It is 366 m thick in its type area and extends from the base of the Lower Douglass coal bed to the top of the Kanawha black flint (Arndt, 1979). The formation thickens southward to over 600 m at Bolt Mountain (Blake et al., 1989). Middle Pennsylvanian strata of southern West Virginia are characterized by coal-bearing sequences punctuated by numerous transgressive intervals. Equivalent stratigraphic intervals occur within the Breathitt Formation of eastern Kentucky and the Upper Pottsville Group of Ohio and Pennsylvania (Cardwell et al, 1968; Fig. 112).

Geologic Setting

The Middle Pennsylvanian paleogeography of Central Appalachian Basin has been described as one characterized by fluvially dominated deltas which prograded northwestward from the rising Appalachian Mountains across West Virginia and eastern Kentucky into an interior sea (Fig. 113; Horne and Ferm, 1978; Flores and Arndt, 1979; Donaldson and Schumaker, 1979; Donaldson et al., 1985). Superimposed on this broad regressive pattern were up to 12 pulses of marine inundation (Fayette County, Hennen and Teets, 1919). Delta switching and tectonically induced subsidence of the foreland basin have been suggested as causes for the frequent transgressions (Horne and Ferm, 1978; Donaldson and Schumaker, 1979; Tankard, 1986; Klein and Willard, 1989). Others have attributed the transgressions largely to eustatic sea level fluctuations (Busch and Rollins, 1984; Chesnut and Cobb, 1989).

Depositional models for the Kanawha and related Pennsylvanian coal-bearing strata have undergone considerable evolution over the years (Udden, 1912; Weller, 1930; Wanless and Weller, 1932; Ferm and Williams, 1963; Ferm, 1975; and Ferm and Weisenfluh, 1989). The Allegheny model of Ferm and Williams (1963) emphasized the similarity of facies relationships with a deposystem comparable to that of the modern Mississippi delta. The applicability of the Allegheny model to strata in the Pocahontas basin was initially accepted by many workers (Donaldson, 1974; Horne and Ferm, 1978; Arndt, 1979).

According to the Allegheny model, marine units of the Kanawha and Breathitt Formations have been interpreted as lower delta plain bay fills (Horne and Ferm, 1978; Horne et al., 1978; Arndt, 1979; Flores and Arndt, 1979). These were thought to intergrade seaward (northwest) with orthoquartzitic barrier deposits and landward (southeast) with upper delta plain lithic sandstones and related facies. Bay filling was accomplished mainly by direct fluvial processes (crevassing and mouth bar deposition) in a manner comparable to the modern river-dominated Mississippi delta. Many marine units that were previously interpreted as lower delta plain interdistributary bay fills are now recognized as having regional rather than local extent and might be better described as "sea fills" (Chesnut, 1989). Ettensohn (1980) has shown that the time equivalence of offshore marine shales and deltaic deposits implied by the Allegheny model was in error in that offshore facies were actually Mississippian rather than Pennsylvanian. Recent workers have reinterpreted the orthoquartzitic barrier sands in eastern Kentucky as alluvial facies (Chesnut, 1989; Wezivitch and Eriksson, 1990).

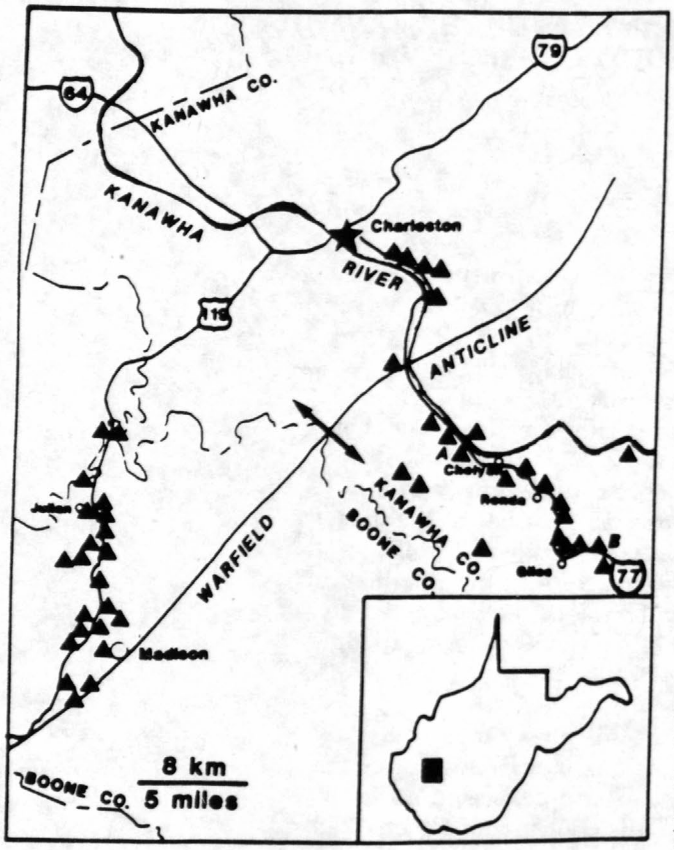

Figure 111 - Study area. A and B indicate points of the stratigraphic cross section in fig. 119.

Donaldson's (1979) generalized model for Pennsylvanian coal-bearing strata in West Virginia, Ohio, and Pennsylvania differs from the Allegheny model in two ways. Widespread barrier island and back barrier facies are absent, with local barrier bars occasionally developed only as the result of delta switching. Donaldson's model also includes the presence of a tidal plain in areas of delta lobe abandonment and along interdeltaic portions of the coast. Although not specifically designed to explain the Kanawha Formation of southern West Virginia, this model more closely resembles the depositional setting interpreted here (Fig. 114). Donaldson and Eble (1991) proposed that the Kanawha Formation is comprised of transgressive-regressive coal-bearing allocycles 15-42 m thick with an average duration of 0.5 m.y. Avulsion-induced autocycles were interpreted to operate at a higher frequency within these allocycles creating localized transgressions and regressions.

Cyclic variations in climatic conditions are also a potential controlling factor that may have contributed to the alternation of coal zones and siliciclastics (Cecil, 1990). Siliciclastic flux is highest under highly seasonal rainfall while prolonged wetness favors coal development. Wet-dry subtropical to humid subtropical climate fluctuations are interpreted to have occurred in the central Appalachian basin during deposition of the Kanawha Formation (Cecil, 1990).

Paleoenvironments and associated sedimentary facies

Depositional Framework

Facies of the Kanawha Formation are interpreted within the context of the idealized depositional framework in figure 4. The transgressive or regressive nature and the morphology of the shoreline were determined by the interplay between eustasy, basin subsidence, and sediment supply. Lowstands of sea level caused incisement of fluvial drainage lines and soil development in interfluves as the result of sediment bypassing. Subsequent rising sea level led to aggradation of alluvial deposystems, valley-filling, and the development of estuarine deposystems with associated tidal flats. These coastal facies were truncated by ravinement surfaces as the shoreface retreated. During the transgressive maximum, sediment supply to the shelf was temporarily reduced, and condensed sections developed. Much of the nearshore zone was muddy (similar to the Late Devonian Catskill shoreline of Walker and Harms, 1975). During high stands, estuaries became filled and coastal progradation subsequently occurred. Where point sources of sediment supply to the coast were significant, prograding deltaic systems developed. Falling sea level may have intensified regressive events, ultimately causing incision of alluvial channels and soil development. Wave energy along the coast was likely to have been low due to the paleolatitudinal position of the central Appalachian Basin approximately 3

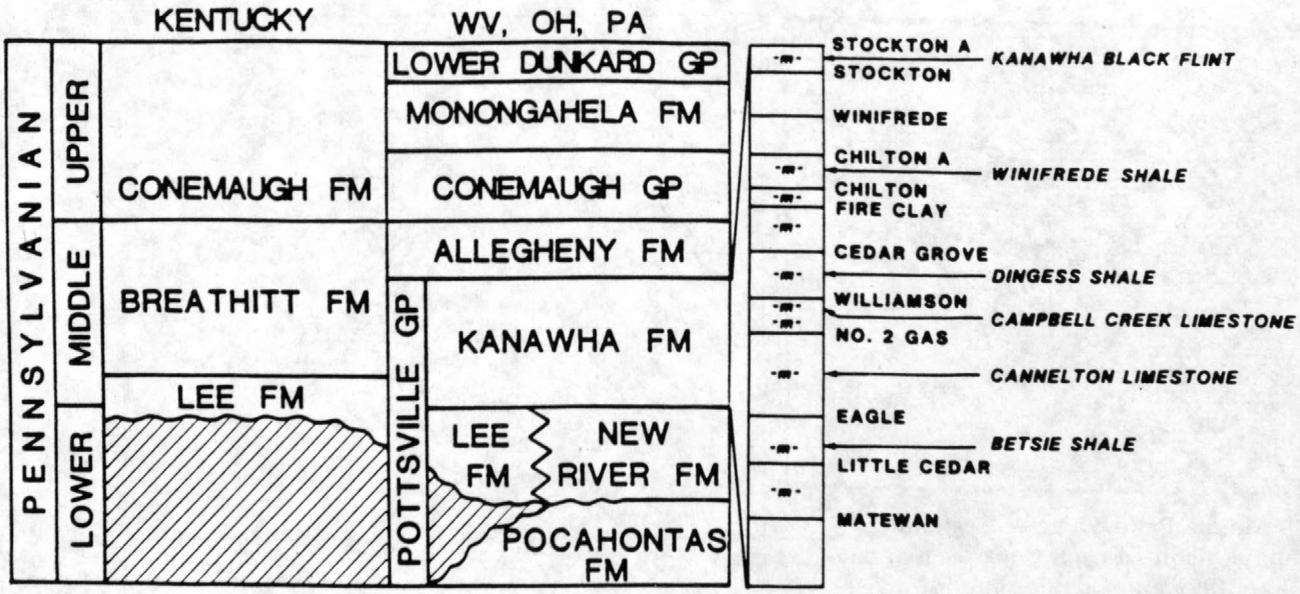

Figure 112 - Stratigraphic framework for the study area. Expanded view of the Kanawha includes selected coal beds and marine units (modified from Cardwell et al., 1968; Englund and Randall, 1981 and Blake et al., in press).

to 5 degrees south of the equator (Scotese, 1986). This position lies in intertropical convergence zone where warm moist air rises and winds tend to be weak (i.e. the doldrums). Wind-driven waves would have been further limited by the small fetch across the seaway (Fig. 113).

Paludal and Lacustrine Facies

The Kanawha Formation contains 26 of West Virginia's 62 commercial coal seams and accounts for about 60 % of the coal produced (Fedorko et al., 1989). These coals average 1.0 % sulfur and 10.9 % ash (M. Blake, 1992, personal comm.). Gas coals dominate the lower and middle portions of the formation (Eagle, Campbell Creek, Cedar Grove coals) whereas splint coal is a common to predominant constituent in coal beds of the upper Kanawha Formation (Winifrede, Coalburg, Stockton coals; Krebs and Teets, 1914).

Low ash, low sulfur coals in the Lower and Middle Pennsylvanian were derived from peat that formed under highly acidic conditions (pH less than 4). Increases in ash and sulfur toward the end of the Middle Pennsylvanian occur in association with the appearance of pedogenic flint clays, and lacustrine and pedogenic carbonates. These features suggest more prolonged periods of aridity between tropical rainy intervals (Cecil et al., 1985; Cecil, 1990).

Some authors interpret peat-forming conditions to have been maximized during the peak of regressions (Chesnut, 1989) whereas others (Donaldson and Eble, 1991) view optimum conditions to have occurred during transgression when decreased sediment supply to the coast occurred. Cecil (1990) maintains that optimum coal-forming conditions were largely independent of sea level changes and were controlled by climatic cycles.

Coals with seat rock and associated carbonaceous shale largely represent fresh water swamp accumulations and lake deposits, respectively. It is possible that some types of vegetation (e. g. *Cordaites* and some lycopsids) may have been tolerant of brackish conditions and, therefore, similar to mangrove vegetation in modern tropical swamps (Gillespie et al., 1978; Gastaldo, 1986). Many of the Kanawha coals have been interpreted as the product of domed peats

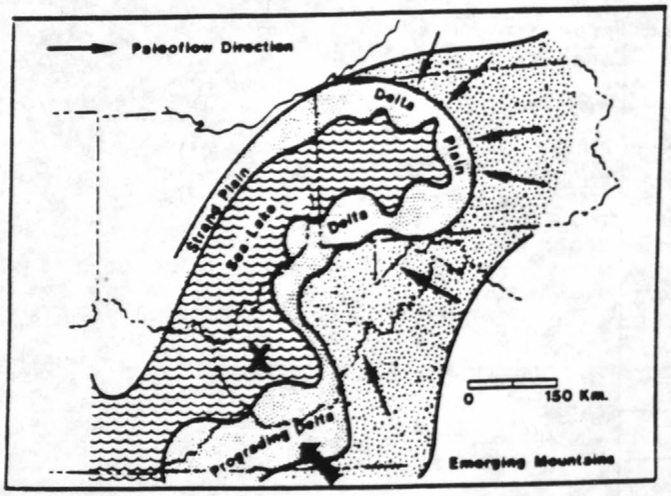

Figure 113 - Middle Pennsylvanian regional paleogeography (modified from Donaldson and Shumaker, 1979 and Donaldson and Eble, 1989)

similar to those forming today in coastal Malaysia (Cecil et al., 1985; Eble et al., 1989).

Alluvial Facies

River channel deposits are composed of mostly fine-medium grained, lithic, micaceous sandstones that are typically channel-form in outcrop. Single story channel sandstones are 7-10 m thick whereas multistory units are up to 35 m thick. Internal structures for active channel-fills are dominated by trough cross-stratification; set thickness decreases upward along with grain size into ripple cross-lamination. Epsilon cross-bedding is commonly evident with sets 7-10 m thick and foreset dips up to 14 degrees. The predominant flow direction is toward the NNW (Fig. 115) which is down the regional paleoslope. However, a large dispersion is indicated; some localities have channel-fills with flow toward the E, SE, and S. This may be the result of 1) high sinuosity channel systems, 2) local deflections in channel orientation caused by contemporaneous faulting of the coastal plain (Ferm and Weisenfluh, 1989), and 3) flow reversals associated with subtle tidal influence.

Smaller scale (1-4 m thick) channelform or wedge shaped sands with unidirectional cross-lamination and parallel lamination

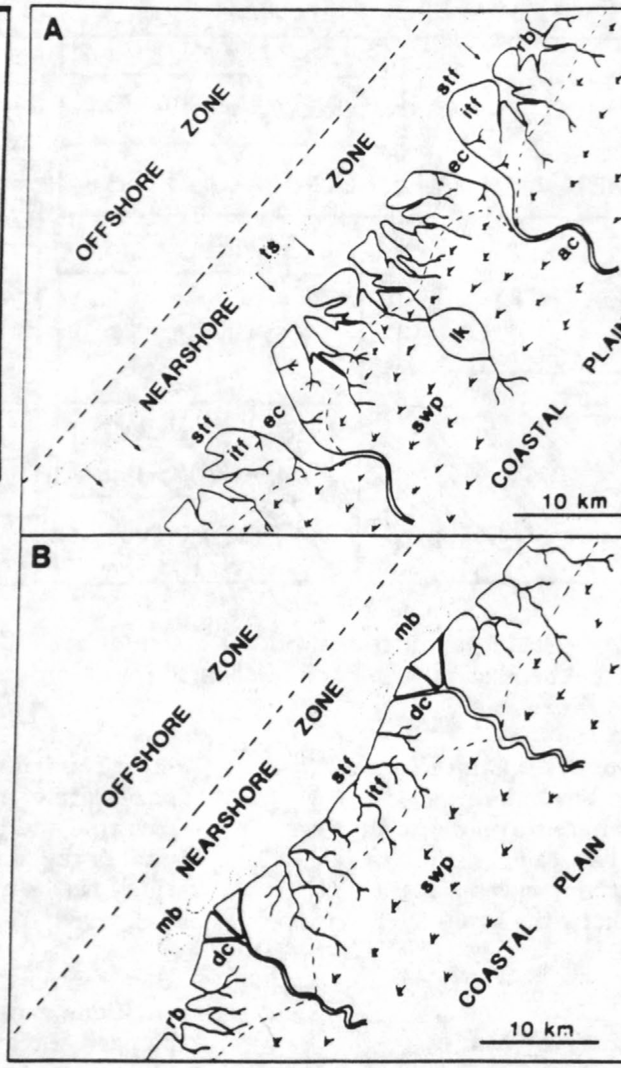

Figure 114 - Depositional framework for the Kanawha Formation (from Donaldson, 1979 and Tankard, 1986) A: transgressive coastal zone (ec=estuarine channels, tc=tidal creeks, itf=intertidal flats, stf=subtidal flats, rb=restricted bays, ts=transgressive sands, ac=alluvial channel, swp=coastal swamp, lk=lake, dc=distributary channel, mb=mouth bar). B: Regressive coastal zone.

represent splay channel, splay lobe and levee deposits. Buried fossilized trees are commonly found in situ (i.e. upright). Examples occur along I-77 within the Campbell Creek coal zone near Chelyan and the Fire Clay coal zone. These alluvial facies are similar to those described by Horne et al. (1978). Major channel sandstones often occur with sufficient erosional relief to partially or

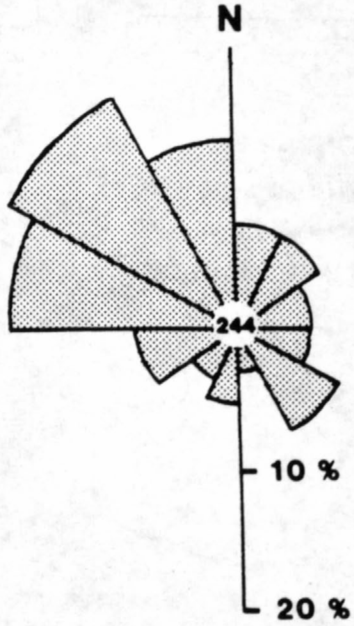

Figure 115 - Paleocurrent rose diagram showing a opredominantly northwestward sediment dispersal pattern. The small southeastward mode may be related to tidal processes.

totally cut out underlying marine units.

Estuarine Channels

Estuarine channel fills (i.e. marine-influenced lower reaches of a river or deltaic distributary) of the Kanawha Formation are similar in size and gross appearance to facies of alluvial channels and fluvially dominated deltaic distributaries. They are typically 7-10 m thick, single story channel fills, and consist of very fine to fine, micaceous, lithic sandstone. Compound cross-stratification is commonly developed. Features that help distinguish tidal or estuarine channels from the fluvial counterparts include rhythmic alternation of textures and structures, reactivation surfaces within cross bed sets, bioturbation, and bipolar paleocurrents (Klein, 1977; Frey and Howard, 1986). Good examples occur between the Matewan and Little Cedar coals near Madison and between the Winifrede shale and Winifrede coal southeast of Charleston (Figs. 116 and 117). Estuarine channelfills have also been described from the Breathitt Formation of eastern Kentucky (Greb and Chesnut, 1991).

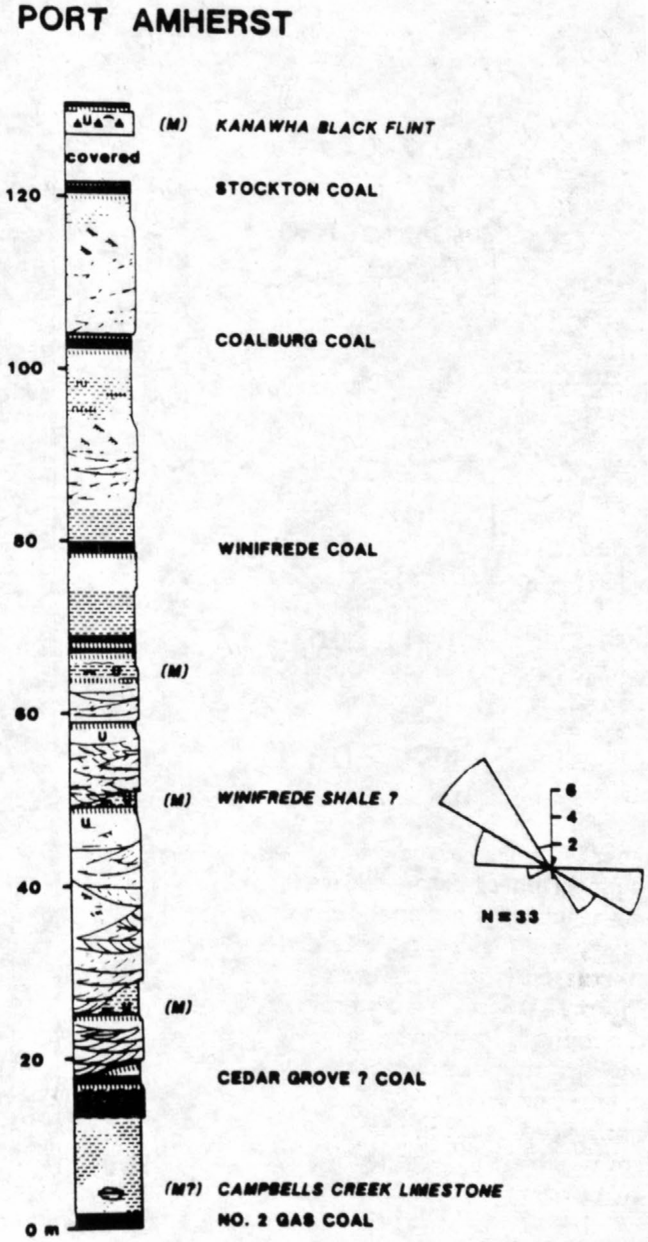

Figure 116 - Port Amherst stratigraphic section. Channel sandstones show bipolar paleocurrents, rhythmic laminations, burrowing and marine invertebrate facies.

Tidal Creeks

Tidal creek deposits are usually channelform bodies 1-3 m thick and about 10-50 m wide. They are filled with very fine grained, bioturbated, sideritic sandstone or dark gray mudstone and usually overlie coal seams. Low angle (less than 10 degrees)

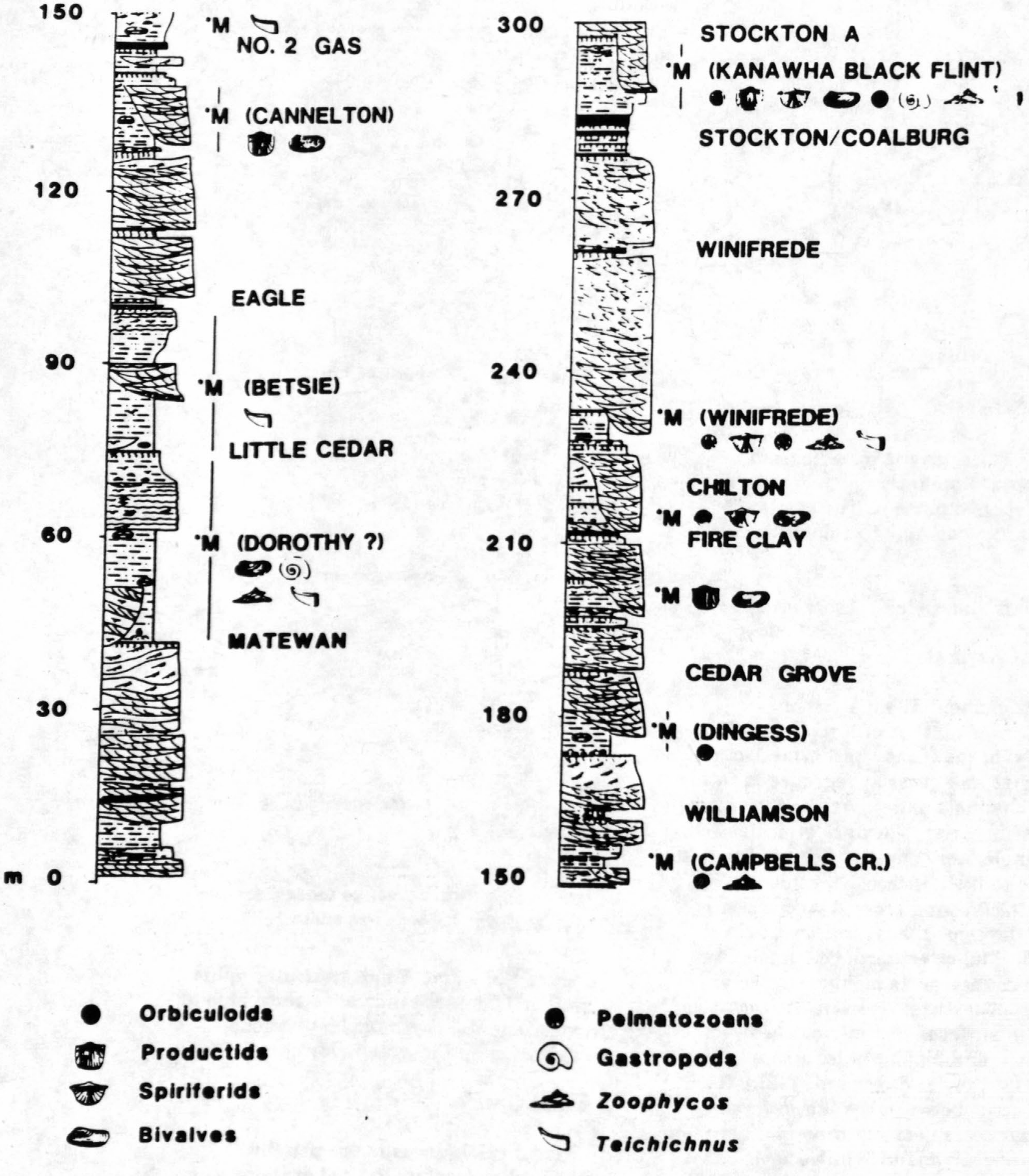

Figure 117 - Composite section based on exposures near Madison, WV, along U.S. Route 119. 10 marine intervals (M) have been identified from this location.

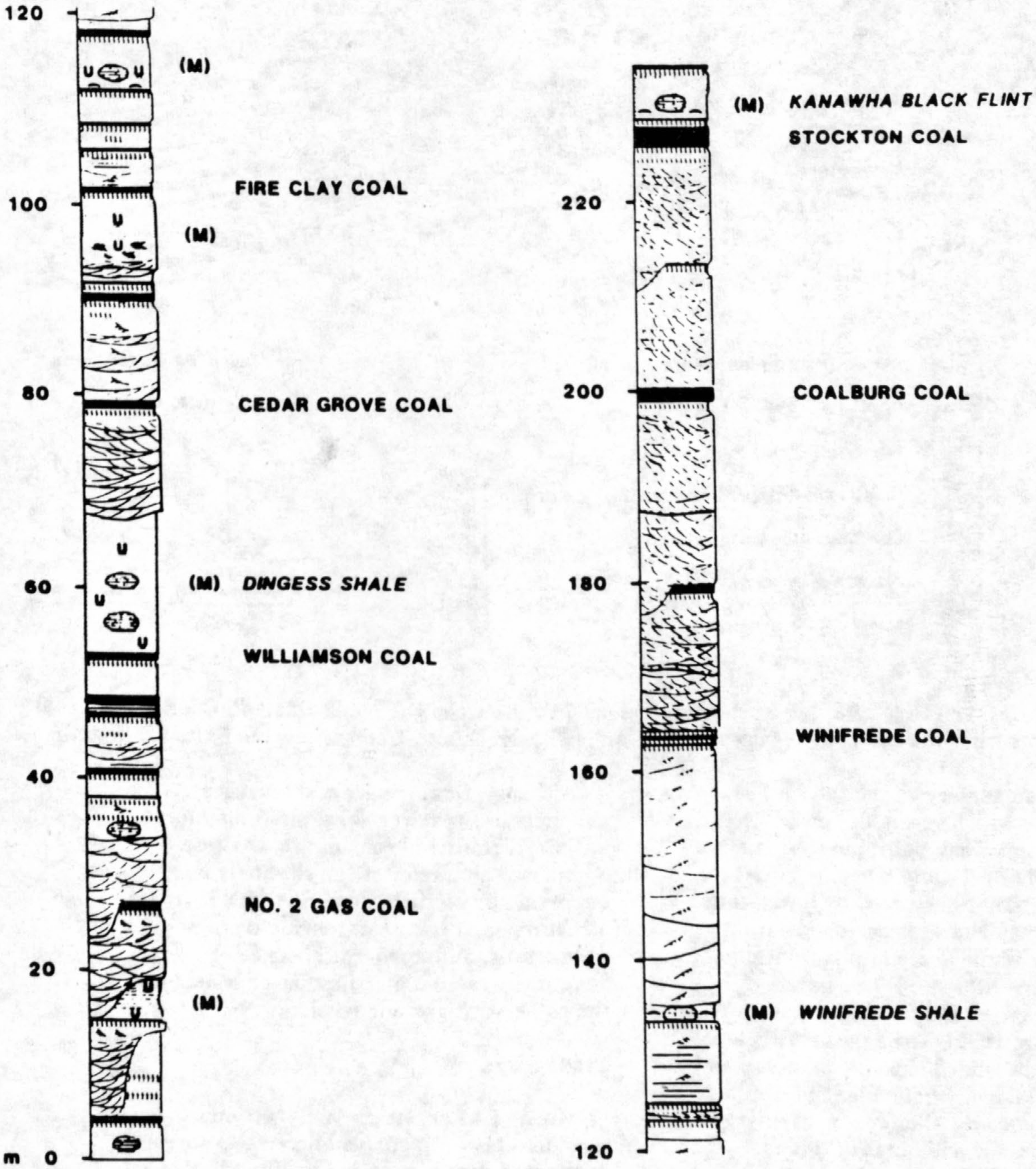

Figure 118 - Composite section from exposures along the West Virginia Turnpike (I-64/I-77) from Standard, WV to the turnpike tunnel bypass.

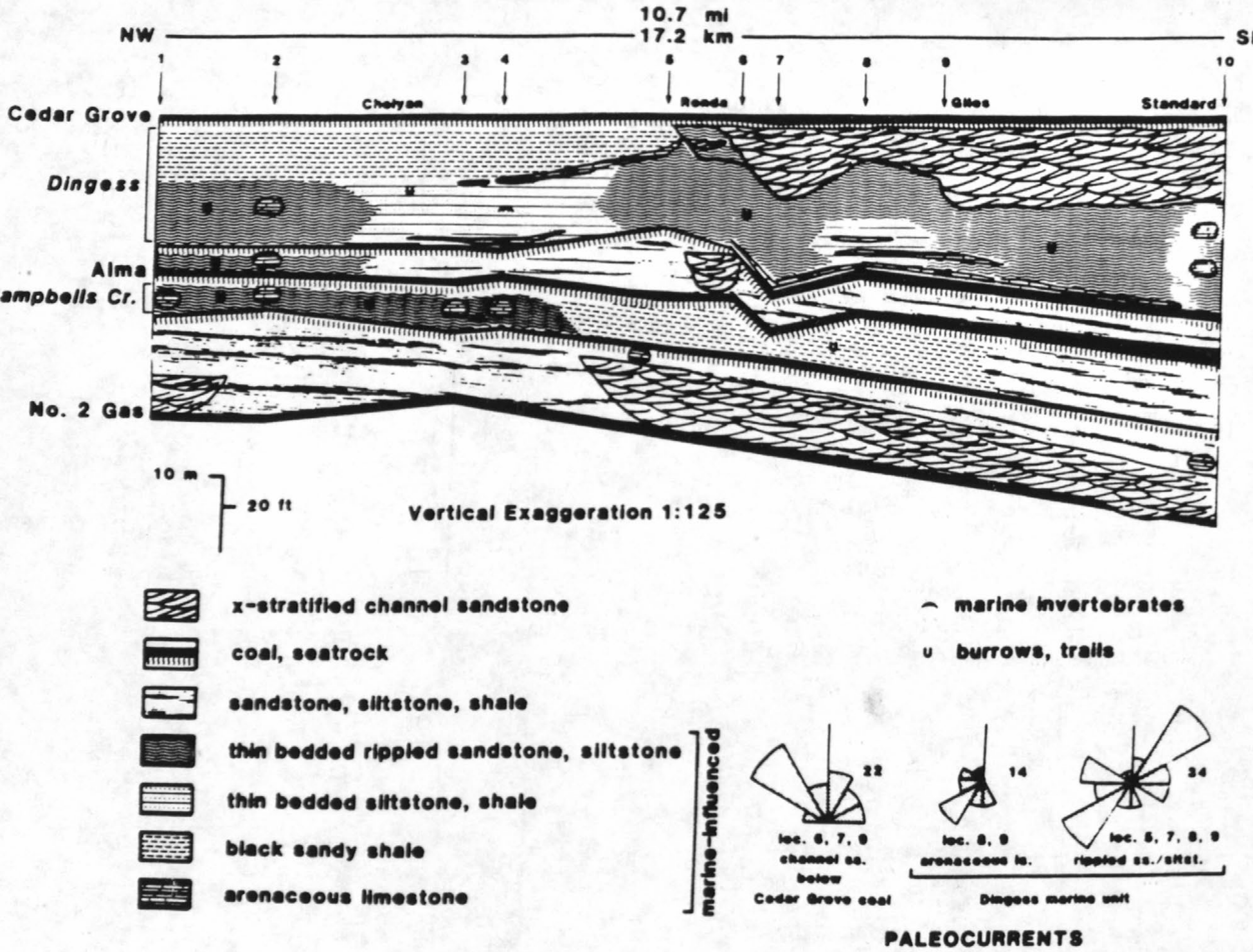

Figure 119 - Stratigraphic cross section along I-77, between the No. 2 Gas and Cedar Grove coal zones, showing component facies of the Campbells Creek and Dingess marine units (italicized).

lateral accretion surfaces are often distinguishable which resemble those produced by tidal creek point bars. Ripple bedding (mostly flaser and wavy) is often preserved where not destroyed by burrowing. Orbiculoid and spiriferid brachiopods are occasionally present. Trace fossils typically found in marine-influenced facies are common (*Zoophycos, Helminthopsis*, and others; Martino, 1989). The channel fills are overlain by thin-bedded sideritic siltstone and shale and are laterally equivalent to ripple bedded sandstone and siltstone or dark gray to black shale. Good examples of this facies are developed on I-77 at Standard between the No. 2 Gas and Powellton coals, and on Rt. 119 near Madison above the Little Cedar coal and at the base of the Campbell Creek marine zone (Figs. 117, 118, 119).

Although tidal creeks and crevasse channels in the lower delta plain may form an intergradational continuum, the tide dominated character of the channels is suggested by bioturbation, rhythmic internal structures and textures, marine trace and body fossils, and apparent absence of genetically related distributary channels at the same stratigraphic level.

Tidal Flats

Much of what has previously been considered bay fill splays and river-dominated mouth bars in the Kanawha Formation (Arndt, 1979; Flores and Arndt, 1979) may be more accurately interpreted as tidal sand and mud flats associated with shoals within the embayment or along its

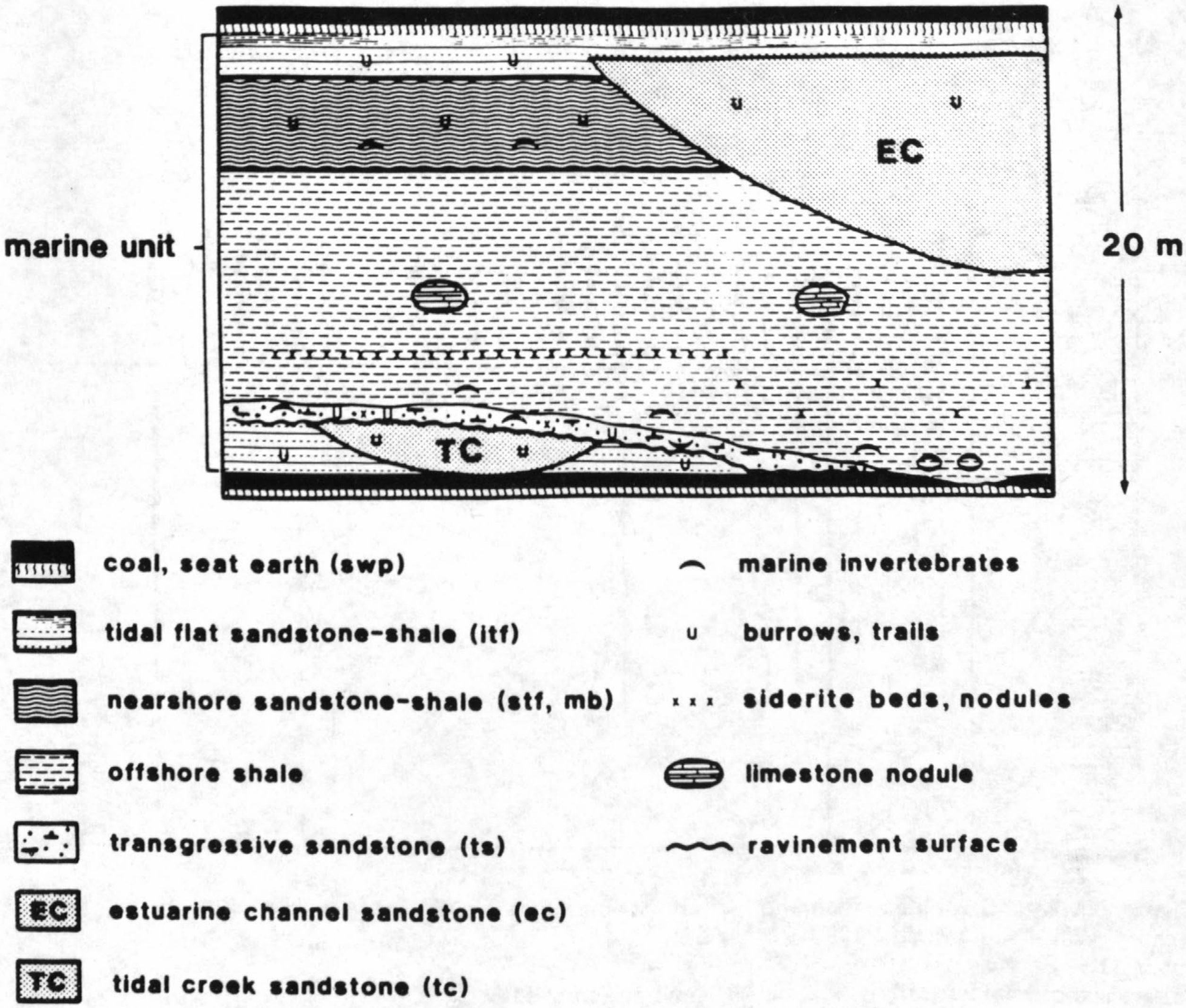

Figure 120 - Idealized Kanawha marine unit with typical arrangement of component facies. Corresponding paleoenvironments are shown in fig. 114.

margin. Sand flats of the low to middle intertidal zone and shallow subtidal zone (nearshore zone) are represented by thin bedded, rippled, very fine to fine sandstone and siltstone. Asymmetric ripple forms predominate within wavy, flaser, and lenticular ripple bedding. Ripple cross-lamination is also abundant with common herringbone cosets. Siltstone partings contain abundant mica and fine plant detritus, while *Calamites* and other plant fossils are occasionally preserved in the thin sandstones.

Trace fossils are sparse to abundant and are preserved mainly along bedding planes. A high abundance, high diversity trace fossil assemblage is widely developed (Martino, 1989, 1993).

Ripple cross-lamination commonly indicates a bipolar paleocurrent pattern as exemplified in the Dingess marine zone along I-77 between Rhonda and Giles (Fig. 119). Unimodal and polymodal patterns are also evident locally. This heterolithic facies commonly grades laterally and vertically into dark gray shale (upper tidal flat) which, in turn is overlain by seat earth and coal (Fig.11 9). These facies transitions are demonstrated from NW to SE along I-77 in the Campbell Creek marine zone (Fig. 119). The dark shale is thin bedded to thinly laminated. The range of ripple bedding types

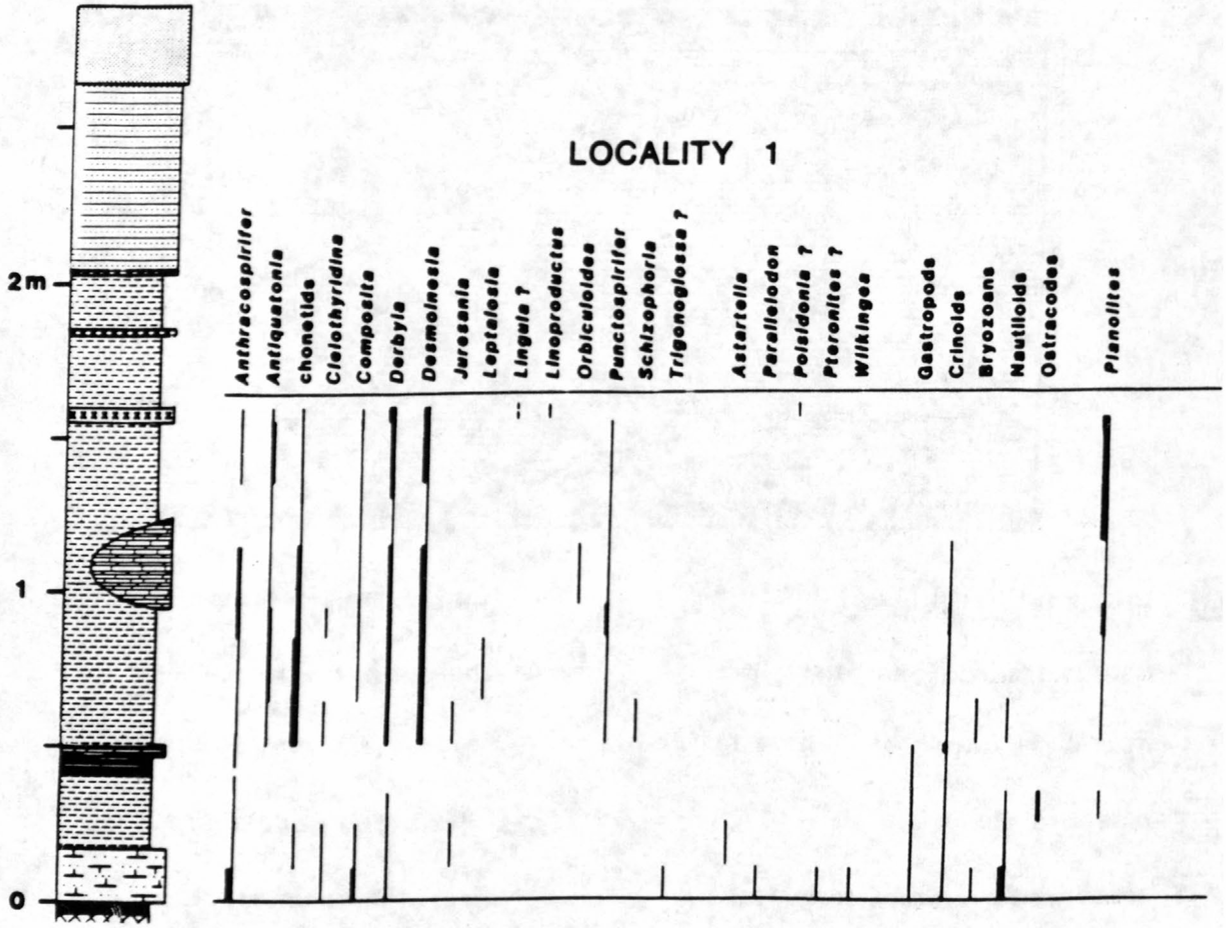

Figure 121 - Stratigraphic section and faunal distribution of the Winifrede marine shale at its type locality (White and Martino, 1992)

reflects the alteration of weak, usually multidirectional traction currents and quiescence in an environment where both sand and mud were available. The trace fossil associated with this facies are characteristic of shallow marine and coastal environments (Chamberlain, 1978; Miller and Knox, 1985). Trace fossils include those formed by deposit-feeding worms (*Asterosoma, Teichichnus, Helminthopsis, Zoophycos, Rosselia,* and *Planolites*), creeping and grazing snails (*Scolicia, Curvolithus, Aulichnites,* and transversely ridged surface trails), and crawling and burrowing arthropods (*Olivellites?, Ancorichnus, Tasmanadia,* and *Petalichnus*; Martino, 1989, 1993). These trace fossils, in addition to the abundance of plant detritus, vertical and lateral facies relations, and multidirectional paleocurrents, indicate a depositional setting invloving shallow subtidal to low intertidal sand and mixed flats.

The dark gray shale facies that often overlies the heterolithic facies and directly underlies coals in interpreted to have been deposited in upper tidal flats which intervened between more seaward sand flats and coastal swamps (Fig. 114). The environment appears to have lacked indigenous macrorganisms, suggesting that stressful ecologic conditions prevailed. This may have resulted from frequent subaerial exposure, high turbidity, or fluctuating salinity. Thin sand laminations associated with the dark gray shale facies resemble "pin-stripe" tidal bedding; this structure commonly develops in mid tidal flats (Klein, 1977). Rare trace fossils (*Zoophycos, Teichichnus, Phycodes,* and *Planolites*) that are preserved within sandier portions of the facies were produced by infaunal deposit-

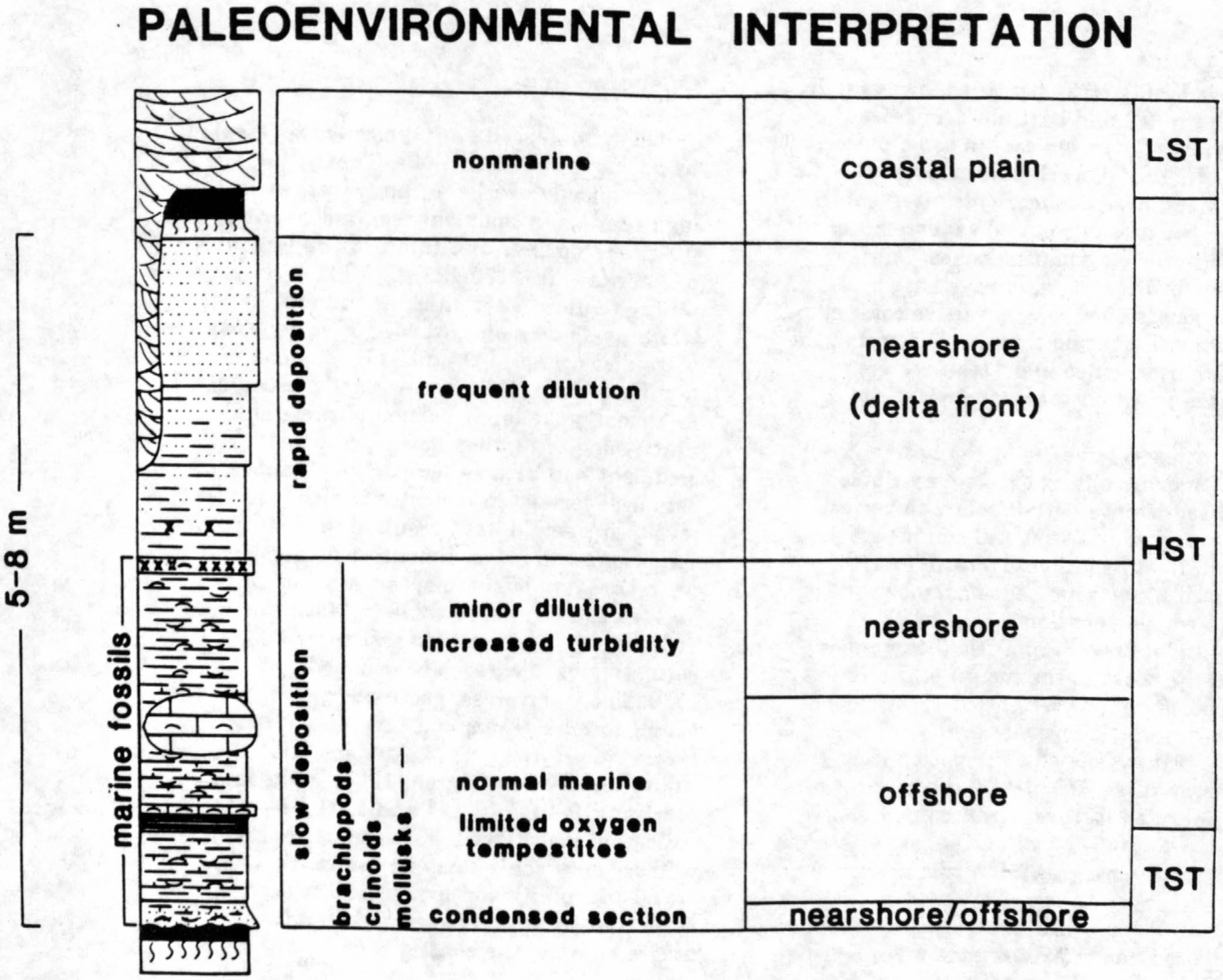

Figure 122 - Paleoenvironmental interpretation of the Winifrede marine zone in its type area. TST=transgressive systems tract, HST=high-stand systems tract and LST=low-stand systems tract.

feeders (Chamberlain, 1971; Hantzschel, 1975). The restriction of trace fossils to infrequent sandy intervals suggests that more hospitable conditions temporarily prevailed during or following episodes of slightly increased energy, perhaps associated with unusually high tides (spring or storm).

Nearshore Zone

The nearshore zone extends from low tide to storm wave base. It includes shallow subtidal areas of interdistributary bays and the delta front, as well as interdeltaic or nondeltaic nearshore (i.e. shoreface and transition zone) deposits (Fig. 114). The Kanawha nearshore deposits consist mainly of very fine sandstone thinly interbedded with siltstone and shale. Body fossils are sparsely distributed and include *Lingula* and *Orbiculoidea*; these inarticulate brachiopods were tolerant of salinity fluctuations and pulses of rapid sediment influx (Gastaldo et al., 1989). Burrowing bivalves such as *Wilkingea* are also characteristic of nearshore facies. Thin intervals of bioturbated, very fine sandstone frequently overlie coals near the base of transgressive facies sequences. These units contain abundant siderite nodules and are widespread but locally discontinuous. They are often richly fossiliferous, containing crinoid ossicles and columns, productid and spiriferid brachiopods, gastropods, and rugose corals. This facies is well-developed at the base of the Dingess marine unit along I-77 between Chelyan and Giles (Fig. 119). In this area it is locally cross-stratified, has as much as 1 m of relief, and partially truncates the underlying Williamson coal in some places. Elsewhere, thin bioturbated sandstones with pebble size siderite concretions intervene between the tops of channel sandstones or coals and overlying marine zones. The tendency for these thin, erosive based sideritic sandstones to occur above coals or distributary/estuarine channel fills and at the base of marine zones indicates they are associated with coastal ravinement.

The transgressive lag at the base of the Dingess marine unit is overlain by ripple bedded sandstones with bipolar, northwest-southeast flow, and scour-fill sandy limestones with southeast flow (Fig. 119). These facies may represent nearshore and deposits which were dominated by shore parallel, tidal flow. Similar tidal currents parallel the coast at the mouth of the Klang Delta (Coleman and Wright, 1975). Evidence of wave-generated currents in the nearshore deposits of the Kanawha Formation is exceedingly rare. Exceptional occurrences of wave-generated features occur in nearshore facies in the Kanawha black flint marine unit along Rt. 119 north of Julian (Fig. 117). Hummocky cross-stratified, very fine sandstones occur which are commonly capped by oscillation ripples. Associated trace fossils include *Olivellites, Helminthopsis*, and *Neonereites* among others (Watson, 1992). Hummocky cross-stratification is generally attributed to the interplay between unidirectional and oscillatory currents generated by storms between fair weather and storm wave base (Swift et al., 1983).

Distributary mouth bar deposits are developed locally in the upper portion of some marine units (Fig. 122). They are represented by accumulations of interbedded sandstone and shale that intergrade laterally along the depositional strike with more mud-rich facies. Thickening and coarsening upward of sand layers is evident and burrows are sparse or absent. Overlying facies usually are cross-stratified distributary or alluvial channel sandstones. These mouth bar deposits are not as widespread as previous workers have indicated and grade laterally along strike into tidal sand and mud flat deposits.

Offshore Zone

Offshore deposits are typically represented by dark gray to black shale. These shale units typically overlie a thin transgressive lag deposit and commonly contain micritic limestone nodules and thin beds and nodules of siderite (Figs. 120-122). The black shale units are up to 22 m thick in outcrop (Dorothy shale along Rt. 119, 5 km southwest of Madison, Figs. 111 and 117). Stenohaline fossil assemblages where present are usually localized in the lower portion of these shale units, suggesting they developed when sediment had become temporarily trapped in estuaries formed in response to rising sea level. This would have resulted in clear water conditions and low sedimentation rates in nearshore and offshore areas. Marine units such as the Winifrede shale (= Magoffin shale of Kentucky) contain a diverse macrofauna consisting of calcareous brachiopods, cephalopods, bivalves, gastropods, and echinoderms (Dennis and Lawrence, 1979; Henry and Gordon, 1979; Martino and Adkins, 1986; Bennington, 1991; White and Martino, 1992; Figs. 121 and 122). The paucity of macrofauna in the upper parts of offshore silty shales may be the result of increased turbidity and higher sedimentation rates that occurred during highstand progradation of the coastal zone.

Although facies relations indicate an offshore setting for many of these marine shale units, the normal marine fauna that would be expected is not often present. This is ofetn the case in basinwide marine units such as the Betsie and Dingess shales. Factors that may have inhibited the development of offshore benthos include limited oxygen, high turbidity, and possibly abnormal salinity. Oxygen deficient bottom waters may result when they become isolated from well oxygenated surface waters by strong density gradients associated with pycnoclines. These may result from thermal stratification or salinity stratification below brackish surface waters. During the early stages of transgression in epicontinental seas, rapid expansion of oxygen-deficient "puddles" of deep water occurs in response to subsidence, sea level rise, and reduced sediment supply. In initially shallow water areas, these transgressive black shales rest on condensed, basal transgressive lags or unconformities (Wingnall, 1991).

Summary and Conclusions

Strata of the Kanawha Formation are represented by a mosaic of facies reflecting deposition in 1) offshore and nearshore zones of a shallow sea, 2) tidal flats, tidal creeks, and estuarine channels along interdeltaic and transgressive coastal settings, 3) river-dominated deltaic distributaries and mouth bars, and 4) various fresh water coastal plain environments. It is likely that the character of the coastal zone varied, with low stands of sea level causing incisement of fluvial channels; rising sea level drowned most of the coastal zone leading to the expansion of tidal plains and estuaries. The influence of tides is reflected by rhythmic alternation of textures and structures through a range of high and low energy depositional settings and is supported in many instances by bipolar paleocurrent data. As the coast retreated, shoreface erosion typically produced ravinement surfaces which were covered by transgressive lags that now separate coastal and marginal marine facies from shallow subtidal facies. Coarsening-upward, regressive facies typically developed in the upper part of marine units during highstands after estuaries had filled with sediment. Marine units were capped by fining upward sequences formed in estuarine channels and prograding tidal flats.

The distribution of body and trace fossils shows facies zonation. Substrate (type/stability), turbidity, dissolved oxygen level, salinity (level/stability), and food supply were likely to have been important factors which controlled the environmental range of organisms. In nearshore and intertidal areas characterized by silts and sands, rapid deposition and erosion, and variable salinity, calcareous benthos were discouraged, although phosphatic brachiopods are locally preserved. Arthropods, medusae, annelids, and gastropods left behind a high diversity trace fossil assemblage in tidal sand and mixed flats. Mud-dominated areas in restricted embayments or upper tidal flats that were associated with low oxygen, diluted salinity, and possibly exposure were characterized by the low trace fossil diversity. In mud-dominated offshore areas where deposition was slower and waters were clear and of more stable salinity, calcareous brachiopods dominated along with subordinate bivalves, bryozoans, echinoderms, cephalopods, and corals. Hospitable conditions for stenotopic benthos were generally short-lived. High turbidity, rapid deposition, limited oxygen, and possible salinity reduction of the seaway due to constriction are potential sources of stress that often limited or prohibited the development of offshore benthic communities. The basinwide extent of most major coal zones and many marine units indicates the likelihood that allocyclic controls (tectonics, eustatsy, and climate changes) were largely responsible for controlling depositional patterns. Further work will be required to clearly demonstrate the relative importance of these factors.

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References

Adams, H.J., 1954, A study of freshwater limestones in the Conemaugh and Monongahela Series, Pittsburgh area: University of Pittsburgh M.S. Thesis, 111 p.

Alvord, D.C., 1971, Geologic map of parts of the Naugatuck and Delbarton quadrangles, eastern Kentucky: U.S. Geological Survey Geological Quadrangle Map GQ-879, 1:24,000 scale.

Alvord, D.C., and Trent, V.A., 1962, Geology of the Williamson quadrangle in Kentucky: U.S. Geological Survey Geologic Map GQ-187, 1:24,000 scale.

Anderson, R.W., and Hester, N.C., 1977, Pennington Formation and lower tongue of the Breathitt Formation, in Dever, G.R., Jr., and others, Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky, Field Trip Guidebook, Eastern Section, American Association of Petroleum Geologists: Kentucky Geological Survey, p. 29-34.

Andrews, E. B., 1873, Geology, *in* Report on the Second Geological District; Gallia Co., Meigs Co., Athens Co., Morgan Co., Muskingum Co: Ohio Geological Survey, p. 225-364.

Arkle, T., Jr., 1959, Monongahela Series, Pennsylvanian System, and Washington and Greene Series, Permian System, of the Appalachian Basin, *in* Guidebook for Field Trips Pittsburgh Meeting, 1959: New York, New York, The Geological Society of America, p. 115-141.

--- 1969, The configuration of the Pennsylvanian and Dunkard (Permian?) strata in West Virginia; a challenge to classical concepts, *in* Donaldson, A. C., ed., Some Appalachian Coals and Carbonates: Models of Shallow Water Deposition: Morgantown, West Virginia, West Virginia Geological and Economic Survey, p. 55-88.

--- 1974, Stratigraphy of the Pennsylvanian and Permian Systems of the central Appalachians,*in* Briggs, G., ed., Carboniferous of the southeastern United States: Boulder, Colorado, The Geological Society of America, Inc., p. 5-29.

Arkle, T., Jr., Barsel, D.R., Larese, R.E., Nuhfer, E.B., Patchen, D.G., Smosma, R.A., Gillespie, W.H., Lund, R., Norton, C.W., and Pfefferkorn, H.W., 1979, The Mississippian and Pennsylvanian Systems in the United States - West Virginia and Maryland: U.S. Geological Survey Professional Paper 1110-D, 35 p.

Arndt, H.H., 1979, Middle Pennsylvanian Series: *in* Englund, K.J., Arndt, H.H., and Henry, T.W. (*eds.*), Proposed Pennsylvanian System stratotype, Virginia - West Virginia: Field guidebook for Trip No. 1, IX International Congress of Carboniferous Stratigraphy and Geology: American Geological Institute Guidebook Series No. 1, p. 73-80.

Bean, H.P., 1982, Stratigraphic analysis of thickness, sulfur and enclosing strata of the Pittsburgh and Redstone Coals in northern West Virginia-southwestern Pennsylvania: West Virginia University M.S. Thesis, 169 p.

Beerbower, J.R., 1962, Cyclothems and cyclic deposition mechanisms in alluvial plain sedimentation: Kansas Geological Survey Bulletin 169, v. 1, p. 31-42.

Belt, E.S. and Lyons, P.C., 1989, A thrust-ridge paleodepositional model for the Upper Freeport coal bed and associated clastic facies, Upper Potomac coal field, Appalachian basin, U.S.A.: International Journal of Coal Geology, v. 12, p. 293-328.

Bement, W. O., 1976, Sedimentological aspects of Middle Carboniferous sandstones on the Cumberland overthrust sheet: Ph. D. Dissertation, University of Cincinnati, Cincinnati Ohio, 182 p.

Bennington, J. B., 1992, Paleoenvironmental analysis of the Magoffin Marine Zone transgression, Middle Pennsylvanian Breathitt Formation, Eastern Kentucky: Geological Society of America, Abstracts with Programs, v. 23, no. 1, p. 8.

Berryhill, H. L., Jr., and Swanson, V. E., 1962, Revised stratigraphic nomenclature for Upper Pennsylvanian and Lower Permian rocks, Washington Co., Pa: U.S. Geological Survey Professional Paper, v. 450C, p. 43-46.

Berryhill, Henry L. Jr., Schweinfurth, Stanley P., and Kent, Bion H., 1971, Coal-bearing Upper Pennsylvanian and Lower Permian Rocks, Washington area, Pennsylvania: U.S. Geological Survey Professional Paper 621, 47 p.

Blake, B.M., Jr., Keiser, A.F., and Eble, C.F. , 1989, Stop 32: Bolt Mountain section: *in* Cecil, C.B., and Eble, C.F. (*eds.*), Carboniferous geology of the Eastern United States; guidebook for fieldtrip T143, 28th. International Geological Congress, Washington D.C., American Geophysical Union, p. 95-97.

Blake, B.M. Jr., Keiser, A.F., and Rice, C.L., in press, Revised stratigraphy and nomenclature for the Middle Pennsylvanian Kanawha Formation in southwestern West Virginia.

Blatt, H., 1992, Sedimentary petrology, second edition: New York, W.H. Freeman and Co., 514 p.

Bohor, B.F., and Triplehorn, D.M., 1981, Volcanic origin of the flint clay parting in the Hazard No. 4 (Fire Clay) coal bed of the Breathitt Formation in eastern Kentucky, *in* Cobb, J.C., Chesnut, D.R., and Hester, N.C. (*eds.*), Coal and coal-bearing rocks of eastern Kentucky: Geological Society of America Coal Division Fieldtrip, November 5-8, 1981, p. 49-54.

Buol, S.W., Hole, F.D., and McCracken, R.J., 1988, Soil Genesis and Classification: 3rd edition., Iowa State University Press, Ames Iowa, 446 p.

Burbank, D.W., 1992, Causes of recent Himalayan uplift deduced from deposited patterns in the Ganges basin: Nature, v. 357, p. 680-683.

Busch, R. M., and Rollins, H. B., 1984, Correlation of Carboniferous strata using a hierarchy of transgressive-regressive units: Geology, v. 12, p. 471-474.

Campbell, M. R., 1903, Description of the Brownsville-Connellsville quadrangle [Pennsylvania]: U.S. Geological Survey Geologic Folios Map 94; 19 p.

Campbell, M.R., and Mendenhall, W. C., 1896, Geologic section along the New and Kanawha Rivers in West Virginia: United States Geological Survey Annual Report 17, pt. 2, p. 473-511.

Cant, D. J., 1982, Fluvial facies models and their application, in Scholle, P. A., and Spearing, D. R., eds., Sandstone depositional environments: American Association of Petroleum Geologists, Memoir 31, p. 115-137.

Cardwell, D. H. Erwin, R. B., Woodward, H. P., 1968, Geologic map of West Virginia: West Virginia Geological and Economic Survey, scale 1:250,000.

Cavaroc, V.V, and Ferm, J.C., 1968, Siliceous spiculites as shoreline indicators in deltaic sequences: Geological Society of America Bulletin, v. 79, p. 262-272.

Cecil, C.B., 1986, Allocyclic and autocyclic conditions of coal formation: U.S. Geological Survey Circular 974; USGS Research on Energy Resources--1986 Program and Abstracts V.E. McKelvey Forum on Mineral and Energy Resources, p. 8-9.

Cecil, C.B. 1990, Paleoclimate controls on stratigraphic repetition of chemical and siliciclastic rocks: Geology, v. 18, p. 533-536.

Cecil, C.B., Stanton, R.W., Neuzil, S.G., Dulong, F.T., Ruppert, L.F., and Pierce, B.S., 1985, Paleoclimate controls on late Paleozoic sedimentation and peat formation in the central Appalachian basin (U.S.A.): International Journal of Coal Geology, v. 5, p. 195-230.

Cecil, C. B., and Englund, K. J., 1989, Origin of coal deposits and associated rocks in the Carboniferous of the Appalachian Basin, in Cecil, C. B., and Eble, C. F., eds., Carboniferous Geology of the Eastern United States: Washington, D. C., 28th International Geological Congress Field Trip Guidebook T143, American Geophysical Union, p. 84-88.

Cecil, C.B., Dulong, F.T., Cobb, J.C., and Supardi, (in press), Allogenic and autogenic controls on sedimentation in the central Sumatra basin as an analogue for Pennsylvanian coal-bearing strata in the Appalachian basin: Geological Society of america Special Publication.

Chamberlain, C. K., 1971, Morphology and ethology if trace fossils from the Ouachita Mountains, southeastern Oklahoma: Journal of Paleontology, v. 45, p. 212-246.

Chamberlain, C. K., 1978, A guidebook to the trace fossils and paleoecology of the Ouachita Mountains: Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma, 68 p.

Chesnut, D. R., Jr., 1981, Marine zones of the upper Carboniferous of eastern Kentucky, in Cobb, J. C., Chesnut, D. R., Jr., Hester, N. C., and Hower, J. C., eds., Coal and coal-bearing rock of eastern Kentucky, Guidebook and roadlog for Coal Division of Geological Society of America Field Trip No. 14: Kentucky Geological Survey, ser. 11, p. 57-66.

Chesnut, D.R., Jr., 1983, Source of the volcanic ash deposit (flint clay) in the Fire Clay coal of the Appalachian Basin: Compte Rendu, Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, (Madrid 1983), v. 1, p. 145-154.

Chesnut, D. R., Jr., 1988, Stratigraphic analysis of the Carboniferous rocks of the Central Appalachian Basin: Lexington, University of Kentucky, Ph. D. Dissertation, 297 p.

Chesnut, D.R., Jr., 1989, Stratigraphic framework of the Pennsylvanian-age rocks of the central Appalachian basin, eastern U.S.A.; in Jin Yugan and Li Chun, eds., Compte Rendu, Onzieme Congrés International de Stratigraphie et du Géologié du Carbonifére: Nanjing, Nanjing University press, v. 2, p. 1-19.

Chesnut, D. R., 1989, Pennsylvanian rocks of the eastern Kentucky coal field, in, C. B. and Eble, C. eds., Carboniferous geology of the eastern United States, Field Trip Guidebook T143, p. 57-60.

Chesnut, D. R., and Cobb, J. C., 1989, Cycles in the Pennsylvanian rocks of the central Appalachian Basin: Geological Society of America Abstracts with Programs, v. 21., no. 6, p. A52.

Chesnut, D. R., Jr., 1991, Paleontological survey of the Pennsylvanian rocks of the Eastern Kentucky Coal Field--Part 1, invertebrates: Kentucky Geological Survey, Series XI, Information Circular 36, 71 p.

Clapp, F.G., 1907, Description of the Amity quadrangle (Pennsylvania): U.S. Geological Survey

Geologic Atlas, Folio 144.

Coleman, J. M., and Wright, L. D., 1975, Modern river deltas: variability of process and sand bodies, in Broussard, M. L., ed., Deltas--models for exploration: Houston Geological Society, p. 99-150.

Collinson, J. D., 1978, Alluvial sediments, in Reading, H. G., ed., Sedimentary Environments and Facies: Oxford, Blackwell Publishing, p. 15-60.

Cooper, B.N., 1961, Grand Appalachian Field Excursion: Virginia Polytechnic Institute Engineering Extension Service Series, Geological Guidebook Numer 1, 187 p.

Conti, Louis, Jr., 1961, Stratigraphy of the Pittsburgh-Redstone Coal interval (Lower Monongahela Series) in central and north-central West Virginia: West Virginia University M.S. thesis, 85 p.

Crider, A.F., 1913, The fire clays and fire clay industries of the Olive Hill and Ashland districts of northeastern Kentucky: Kentucky Geological Survey, Ser. 4, v. 1, pt. 2, p. 589-711.

Demarest, J. M., and Kraft, J. C., 1987, Stratigraphic record of Quaternary sea levels: implications for more ancient strata, in Nummedal, D., Pilkey, O. H., and Howard, J. D., eds., Sea level fluctuations and coastal evolution: Society of Economic Paleontologists and Mineralogists, Special Publication 41, p. 223-239.

Dennis, A. M., and Lawrence, D. R., 1979, Macrofauna and fossil preservation in the Magoffin marine zone, Pennsylvanian Breathitt Formation of eastern Kentucky: Southeastern Geology, v. 20, no. 3, p. 181-192.

Dever, G.R., Jr., 1973, Stratigraphic relationships in the lower and middle Newman Limestone (Mississippian), east-central and northeastern Kentucky (M.S. thesis): Lexington, Kentucky, University of Kentucky, 121 p.

Dever, G.R., Jr., 1977, The lower Newman Limestone: Stratigraphic evidence of Late Mississippian tectonic activity, in Dever, G.R., Jr., and others, Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky, Field Trip Guidebook, Eastern Section, American Association of Petroleum Geologists: Kentucky Geological Survey, p. 8-18.

Dever, G.R., Jr., 1980, Stratigraphic relations in the lower and middle Newman Limestone (Mississippian), east-central and northeastern Kentucky: Kentucky Geological Survey, Series 11, Thesis Series 1, 49 p.

Dever, G.R., Jr., Hoge, H.P., Hester, N.C., and Ettensohn, F.R., 1977, Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky, American Association of Petroleum Geologists Eastern Section Annual Meeting: Guidebook, Kentucky Geological Survey, 80 p.

Dickson, P. A., 1974, Permo-Carboniferous non-marine bivalvia of the Appalachian Basin: their adaptive strategies and stratigraphical utility [Ph. D. dissert.]: University of Pittsburgh, 178 p.; Pittsburgh, PA.

Dobrovolny, E., Sharps, J. A., and Ferm, J. C., 1963, Geology of the Ashland Quadrangle, Kentucky-Ohio, and the Catlettsburg quadrangle in Kentucky: U. S. Geological Survey Map GQ-196.

Donaldson, A., C., 1969, Ancient deltaic sedimentation (Pennsylvanian) and its control on the distribution, thickness, and quality of coals, *in* Donaldson, A. C., ed., Some Appalachian Coals and Carbonates: Models of Shallow Water Deposition: Morgantown, West Virginia, West

Virginia Geological and Economic Survey, p. 93-121.

--- 1974, Pennsylvanian sedimentation of Central Appalachians, *in* Briggs, G., ed.,Carboniferous of the Southeastern United States: Boulder, Colorado, Geological Society of America, p. 47-78.

--- 1979, Depositional environments of the Upper Pennsylvanian Series, *in* Englund, K. J., Arndt, H. H., and Henry, T. W., eds., Proposed Pennsylvanian System Stratotype Virginia and West Virginia, No. 1 *of* AGI Selected Guidebook Series: Falls Church, Virginia, The American Geological Institute, p. 123-131.

Donaldson, A. C., 1979, Origin of coal seam discontinuities, in Donaldson, A. C., Presley, W. W. and Renton, J. J. eds., Carboniferous coal short course guidebook: West Virginia Geological and Economic Survey, Bulletin B-37-1, p. 102-132.

Donaldson, A., and Eble, C. F., 1989, Morgantown area stops, in Cecil, C. B., and Eble, C. eds., Carboniferous geology of the eastern United States, 28th International Geological Congress, Field Trip Guidebook T 143, p. 104-111.

Donaldson, A. C., and Eble, C., 1991, Pennsylvanian coals of central and eastern United States, in Gluskoter, H. J., Rice, D. D., and Taylor, R. B., eds., Economic Geology, U. S.: Boulder, Colorado, Geological Society of North America, The Geology of North America, p. 523-545.

Donaldson, A.C., and Shumaker, R.C., 1981, Late Paleozoic molasse of the central Appalachians: *in* Miall, A.D. (*ed.*), Sedimentation and Tectonics in Alluvial Basins, Geological Association of Canada Special Paper 23, p. 99-124.

Donaldson, A.C., Renton, J.J., and Presley, M.W., 1985, Pennsylvanian deposystems and paleoclimates of the Appalachians: International Journal of Coal Geology, v. 5, p. 167-193.

Donaldson, A.C. and Eble, C.F., 1989, Stop 41: Route 48 - Chestnut Ridge section: *in* Cecil, C.B., and Eble, C.F. [eds.], Carboniferous geology of the Eastern United States; guidebook for fieldtrip T143, 28th. International Geological Congress, Washington D.C., American Geophysical Union, p. 108-111.

Duchaufour, P., 1982, Pedology (English ed.): London George Allen and Unwin, 448 p.

Eager, R.M.C., 1975, Some nonmarine bivalve faunas from the Dunkard Group and underlying measures, in: Barlow, J.A. (ed.), Proceedings of the First I.C. hite Memorial Symposium, "The Age of the Dunkard": West Virginia Geological and Economic Survey, pp. 23-68.

Eble, C.F., 1985. Palynology and paleoecology of the Redstone coal (Upper Pennsylvanian) in West Virginia. unpublished M.S. thesis, West Virginia University, Morgantown, 285 p.

Eble, C.F., Greb, S.F., and Chesnut, D.R., Jr., 1991, Coal-bearing rocks along the western margin of the Eastern Kentucky Coal Field: Guidebook for The Society of Organic Petrography fieldtrip, October 2, 1991.

Eble, C., Grady, W. C., and Gillespie, W. H., 1989, Palynology, petrology and paleoecology of the Hernshaw-Fire Clay coal bed in the central Appalachian basin, in: C. B. Cecil and C. Eble, eds., Carboniferous Geology of the Eastern United States, 28th International Geological Congress, Field Trip Guidebook T143, p. 133-142.

Ekdale, A. A., and Mason, T. R., 1988, Characteristic trace-fossil associations in oxygen-poor sedimentary environments: Geology, v. 16, p. 720-723.

Elliott, T., 1986, Siliciclastic shorelines, in Reading, H. G., ed., Sedimentary environments and facies, Blackwell Scientific Publications, Oxford, p. 155-188.

Englund, K.J., 1974, Sandstone distribution patterns in the Pocahontas Fromation in Southwest Virginia and southern West Virginia, in Briggs, G., [ed.], Carboniferous of the southeastern United States: Geological Society of America Special Paper 148. pp. 31-45.

Englund, K.J., 1979, Mississippian System and Lower Series of the Pennsylvanian System in the proposed Pennsylvanian System stratotype area: *in* Englund, K. J., Arndt, H. H., and Henry, T. W., (*eds.*), Proposed Pennsylvanian System Stratotype, Virginia-West Virginia: American Geological Institute Select Guidebook Series No. 1, Field Trip No. 1, Ninth International Congress of Carboniferous Stratigraphy and Geology, p. 69-72.

Englund, K.J., Windolph, J.F., and Thomas, R.E., 1986, Origin of thick, low-sulfur coal in the Lower Pennsylvanian Pocahontas Formation, Virginia and West Virginia; *in* Lyons, P. C., and Rice, C. L., (*eds.*), Paleoenvironmental and Tectonic Controls in Coal-Forming Basins in the United States: Geological Society of America Special Paper 210, p. 49-61.

Englund, K.J., Arndt, H.H., and Henry, T.W., 1979, IX ICC Field Trip No. 1, Proposed Pennsylvanian Stratotype, Virginia and West Virginia: Falls Church, American Geological Institute, AGI Selected Guidebook Series No. 1, 136 p.

Englund, K.J. and Henry, T.W., 1981, Mississippian-Pennsylvanian boundary in the central part of the Appalachian basin, Part 1: Southwestern Virginia-southern West Virginia, in Roberts, T.G., [ed.], Geological Society of america Guidebook, v. 1, field trip 4, Stratigraphy, sedimentology: Washington, D.C., American Geological Institute, pp. 153-194.

Englund, K.J., and Thomas, R.E., 1990, Late Paleozoic Depositional Trends in the Central Appalachian Basin: U.S. Geological Survey Bulletin 1839-F, 19 p.

Englund, K. J., and Randall, A. H., III, 1981, Stratigraphy of the Upper Mississippian and Lower Pennsylvanian Series in the eastern Appalachians, in Roberts, T. G., ed. Geological Society of America, Cincinnati Field Trip Guidebooks, v. 1: Stratigraphy, sedimentology, p. 154-158.

Ettensohn, F.R., 1975, Stratigraphic and paleoenvironmental aspects of Upper Mississippian rocks (upper Newman Group), east-central Kentucky (Ph.D. dissert.): Urbana, Illinois, University of Illinois, 320 p.

Ettensohn, F.R., 1977, Effects of synsedimentary tectonic activity on the upper Newman Limestone and Pennington Formation, in Dever, G.R., Jr., and others, Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky, Field Trip Guidebook, Eastern Section, American Association of Petroleum Geologists: Kentucky Geological Survey, p. 18-29.

Ettensohn, F.R., 1979a, Generalized description of Carboniferous stratigraphy, structure, and depositional environments in east-central Kentucky, in Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky, Ninth International Congress of Carboniferous Geology and Stratigraphy Field Trip No. 4: Lexington, Kentucky, University of Kentucky, p. 64-77.

Ettensohn, F.R., 1979b, Depositional framework of the Lower Mississippian subaqueous delta sediments in northeastern Kentucky, in Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky, Ninth International Congress of Carboniferous Geology and Stratigraphy Field Trip No. 4: Lexington, Kentucky, University of Kentucky, p. 133-136.

Ettensohn, F.R., 1980, An alternative to the Barrier-Shoreline Model for deposition of Mississippian and Pennsylvanian rocks in northeastern Kentucky: Geological Society of America Bulletin, v. 91, pt. I, p. 130-135, pt. II, p. 934-1056.

Ettensohn, F.R., 1981, Mississippian-Pennsylvanian boundary in northeastern Kentucky, in Roberts, T.G., ed., GSA Cincinnati '81 Field Trip Guidebooks, v. 1: Stratigraphy, sedimentation: Falls Church, Virginia, American Geological Institute, p. 195-257.

Ettensohn, F.R., 1990, Carbonate paleosols as paleoclimatic indicators: examples from the Slade Formation (Upper Mississippian), eastern Kentucky (abs.): Geological Society of America Abstracts with Programs, v. 22, p. A333.

Ettensohn, F.R., in press a, Possible flexural controls on the origins of extensive, ooid-rich, carbonate environments in the Mississippian of the United States, in Keith, B.D., and Zuppann, C.W., eds., Mississippian oolites and recent analogues: American Association of Petroleum Geologists Studies in Geology.

Ettensohn, F.R., in press b, Tectonic control on the formation and cyclicity of major Appalachian unconformities, in Dennison, J.M., and Ettensohn, F.R., eds., Sedimentary cycle control- tectonics and eustacy: Society of Economic Paleontologists and Mineralogists Contributions to Sedimentology, v. 4.

Ettensohn, F.R., and Dever, G.R., Jr., 1979a, Carboniferous geology from the Appalachian Basin to the Illinois Basin through eastern Ohio and Kentucky, Ninth International Congress of Carboniferous Geology and Stratigraphy Field Trip. No. 4: Lexington, Kentucky, University of Kentucky, 293 p.

Ettensohn, F.R., and Dever, G.R., Jr., 1979b, The Waverly arch apical island in Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky, Ninth International Congress of Carboniferous Geology and Stratigraphy Field Trip No. 4: Lexington, Kentucky, University of Kentucky, p. 108-112.

Ettensohn, F.R., and Dever, G.R., Jr., 1979c, The Newman Limestone west of the Waverly arch, in Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky, Ninth International Congress of Carboniferous Geology and Stratigraphy Field Trip No. 4: Lexington, Kentucky, University of Kentucky, ,p. 119-124.

Ettensohn, F.R., Dever, G.R., Jr., and Grow, J.S., 1984a, Fossil soils and subaerial crusts in the Mississippian of eastern Kentucky, in Rast, N., and Hay, J.B., eds., Field trip guides for Geological Society of America Annual Meeting, Southeastern and North-Central Section: Lexington, Kentucky, Department of Geology, University of Kentucky, and Kentucky Geological Survey, p. 84-105.

Ettensohn, F.R., Dever, G.R., Jr., and Grow, J.S., 1988, A paleosol interpretation for profiles exhibiting subaerial exposure "crusts" from the Mississippian of the Appalachian Basin, in Reinhardt, J., and Sigleo, W.R., eds., Paleosols and weathering through geologic time: Geological Society of America Special Paper 216, p. 49-79.

Ettensohn, F.R., and Peppers, R.A., 1979, Palynology and biostratigraphy of Pennington shales and coals (Chesterian) at selected sites in northeastern Kentucky: Journal of Paleontology, v. 53, p. 453-474.

Ettensohn, F.R., Rice, C.L., Dever, G.R., Jr., and Chesnut, D.R., 1984b, Slade and Paragon formations-

-new stratigraphic nomenclature for Mississippian rocks along the Cumberland Escarpment in Kentucky: U.S. Geological Survey Bulletin 1605, 37 p.

Ettensohn, F. R., Rice, C. L., Dever, G. R., Jr., Chesnut, D. R., Jr., 1984, Slade and Paragon Formations--New stratigraphic nomenclature for Mississippian rocks along the Cumberland Escarpment in Kentucky: U. S. Geological Survey Bulletin 1605, 37 p.

Fedorko, N., III, Smith, C. J., Repine, T. E., McClelland, S. W., Keiser, A. F., Loud, E. I., Blake, B. M., McCollough, G. H., Ashton, K. C., and Jake, T. R., 1989, World class coal deposits of West Virginia, 28th International Geological Congress, Abstracts, v. 1, p. I-477.

Ferm, J.C., 1970, Allegheny facies models: *in* Morgan, J.P. (*ed.*), Deltaic Sedimentation, Modern and Ancient: Society of Economic Paleontologists and Mineralogists Special Publication 15, p. 246-255.

Ferm, J. C., 1975, Pennsylvanian cyclothems in the Appalachian Plateau: a retrospective view, in Paleotectonic investigation of the Pennsylvanian System in the United States, Part II: U. S. Geological Survey Professional Paper 853, p. 57-64.

Ferm, J. C., Horne, J. C., Swinchatt, J. P., and Whaley, P. W., (eds), 1971, Carboniferous depositional environments in northeastern Kentucky: Geological Society of Kentucky Field Trip Guidebook, Kentucky Geological Survey, Series 10, 30 p.

Ferm, J.C., and Horne, J.C., 1979, Carboniferous depositional environments in the Appalachian region: Department of Geology, University of South Carolina, 760 p.

Ferm, J.C., and Weisenfluh, G.A., 1989, Evolution of some depositional models in Upper Carboniferous rocks of the Appalachian coal fields: *in* Lyons, P.C. and Alpern, B. [eds.], Peat and Coal: Origin, Facies, and Depositional Models, International Journal of Coal Geology, v. 12, p. 259-292.

Filer, J.K., 1985, Oil and Gas Report and Map of Pleasants, Wood, and Ritchie Counties, West Virginia: West Virginia Geological and Economic Survey Bulletin B-11A, 87 p., 9 pl.

Fischer, A. G., 1961, Stratigraphic record of transgressing seas in the light of sedimentation on the Atlantic coast of New Jersey: American Association of Petroleum Geologists Bulletin, v. 45, p. 1656-1666.

Flores, R.M., and Arndt, H.H., 1979, Depositional environments of Middle Pennsylvanian Series in proposed Pennsylvanian System stratotype: *in* Englund, K.J., Arndt, H.H., and Henry, T.W. [eds.], Proposed Pennsylvanian System stratotype, Virginia - West Virginia: Field guidebook for Trip No. 1, IX International Congress of Carboniferous Stratigraphy and Geology: American Geological Institute Guidebook Series No. 1, p. 115-121.

Flowers, R.R., 1956, A Subsurface Study of the Greenbrier Limestone in West Virginia: West Virginia Geological and Economic Survey Report of Investigation 15, 17 p.

Flint, N. K., 1965, Geology and Mineral Resources of Southern Somerset County: Harrisburg, Pennsylvania, Pennsylvania Geological Survey, Pennsylvania Geological Survey Bulletin C-56A, 267 p.

Flores, R.M., 1986, Styles of coal deposition in Tertiary alluvial deposits, Powder River basin, Montana and Wyoming, in Lyons, P.C., and Rice, C.L., eds., Paleoenvironments and tectonic controls in coal-bearing basins of the United States: Geological Society of America Special Paper 210, p.

79-102.

Fonner, F. and Chappell, G. A., 1987, Geology along I-64 Wayne County, West Virginia: W. Va. Geological and Economic Survey Publication Map-WV28, 31 p.

Frey, R. W., and Howard, J. D., 1986, Mesotidal estuarine sequences: a perspective from the Georgia Bight: Journal of Sedimentary Petrology, v. 56, p. 911-924.

Gastaldo, R. A., Gibson, M. A., and Gray, T. D., 1989, An Alabama: Paleoclimatology, Paleogeography, Paleoecology v. 53, p. 191-212.

Gastaldo, R. A., Gibson, M. A., and Gray, T. D., 1989, An Appalachian-sourced deltaic sequence, northern Alabama, U. S. A.: biofacies-lithofacies relationships and interpreted community patterns: International Journal of Coal Geology, v. 12, p. 225-257.

Gillespie, W. H., Clendening, J. A., and Pfefferkorn, H. W., 1978, Plant fossils of West Virginia: West Virginia Geological and Economic Survey, Educational Series ED-3A, 172 p.

Gillespie, W.H., and Pfefferkorn, H.W., 1979, Distribution of Commonly Occurring Plant Megafossils in the Proposed Pennsylvanian System Stratotype: *in* Englund, K. J., Arndt, H. H., and Henry, T. W., (*eds.*), Proposed Pennsylvanian System Stratotype, Virginia-West Virginia: American Geological Institute Select Guidebook Series No. 1, Field Trip No. 1, Ninth International Congress of Carboniferous Stratigraphy and Geology, p. 87-96.

Gillespie, W.H., and Rheams, L.I., 1985, Plant Megafossils from the Carboniferous of Alabama, U.S.A: Compte Rendu, Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, (Madrid 1983), v. 2, p. 191-202.

--- 1990, Evidence for global (glacial-eustatic) control over Upper Carboniferous (Pennsylvanian) cyclothems in midcontinent North America; *in* Hardman, R. F. P., and Brooks, J., (*eds.*), Tectonic Events Responsible for Britain's Oil and Gas Reserves: Geological Society Special Publication 55, p. 35-47.

Grady, W.C. and Eble, C.F., 1989. Relationships among macerals, minerals, miospores and paleoecology in a column of Redstone coal (Upper Pennsylvanian) from north-central West Virginia (U.S.A.). International Journal of Coal Geology, v. 15, p. 1-26.

Greb, S.F., and Chesnut, D.R., Jr., 1992, Transgressive channel filling in the Breathitt Formation (Upper Carboniferous), Eastern Kentucky Coal Field: Sedimentary Geology, v. 75, p. 209-221.

Habib, D. Spore and pollen paleoecology of the Redstone seam (Upper Pennsylvanian) of West Virginia. Micropaleontology, v. 14(2), p. 199-220.

Haney, D.C., and Hester, N.C., 1978, Geologic map of the Scranton quadrangle, Menifee County, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1488, scale: 1:24,000.

Hantzschel, W., 1975, Trace fossils and Problematica (second edition) in Teichert, C., ed., Treatise on invertebrate paleontology, Part W, Miscellanea, Supplement 1. Boulder, Colorado, Geological Society of America (and University of Kansas Press), 269 p.

Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G., and Smith, D.G., 1989, A geologic time scale: Cambridge, Cambridge University Press, 261 p.

Heckel, P.H., 1986, Sea-level curve for Pennsylvanian eustatic marine transgressive-regressive depositional cycles along midcontinent outcrop belt, North America: Geology, v. 14, p. 330-334.

Hennen, R. V., 1912, Doddridge and Harrison Counties: Morgantown, West Virginia, West Virginia Geological Survey, 712 p.

Hennen, R.V., and Reger, P.B., 1914, Logan and Mingo Counties: West Virginia Geological and Economic Survey [County Report], 776 p.

Hennen, R.V., and Gawthrop, R.M., 1915, Wyoming and McDowell Counties: West Virginia Geological and Economic Survey [County Report], 783 p.

Hennen, R.V., and Gawthrop, R.M., 1917, Braxton and Clay Counties: West Virginia Geological and Economic Survey [County Report], 883 p.

Hennen, R.V., and Teets, D.D .,Jr., 1919, Fayette County: West Virginia Geological and Economic Survey [County Report], 1002 p.

Henry, T.W., and Gordon, Mackenzie, Jr. 1979, Late Devonian through early Permian(?) invertebrate faunas in proposed Pennsylvanian System stratotype area: *in* Englund, K.J., Arndt, H.H., and Henry, T.W. [eds.], Proposed Pennsylvanian System stratotype, Virginia--West Virginia: Field guidebook for Trip No. 1, IX International Congress of Carboniferous Stratigraphy and Geology: American Geological Institute Guidebook Series No. 1, p. 97-103.

Henry, T. W., Lyons, P. C., and Windolph, J. F., Jr., 1979, Upper Pennsylvanian and Lower Permian (?) Series in the area of proposed Pennsylvanian System stratotype, *in* Englund, K. J., Arndt, H. H., and Henry, T. W., eds., Proposed Pennsylvanian System Stratotype Virginia and West Virginia, 1 *of* AGI Selected Guidebook Series: Falls Church, Virginia, The American Geological Institute, p. 81-85.

Hobday D.K., and Horne, J.C., 1977, Tidally influenced barrier island and estuarine sedimentation in the Upper Carboniferous of southern West Virginia: Sedimentary Geology, v. 18, p. 97-122.

Horne, J.C., and Ferm, J.C., 1978, Carboniferous Depositional Environments: Eastern Kentucky and West Virginia: Field Guide, University of South Carolina, 151 p.

Horne, J.C., Ferm, J.C., and Swinchatt, J.P., 1974, Depositional model for the Mississippian-Pennsylvanian boundary in northeastern Kentucky: *in* Briggs, G. [ed.], Carboniferous of the Southeastern United States: Geological Society of America Special Paper 148, p. 97-114.

Houseknecht, D.W., 1980, Comparative anatomy of a Pottsville lithic arenite and quartz arenite of the Pocahontas basin southern West Virginia: Petrogenetic, depositional, and stratigraphic implications: Journal of Sedimentary Petrology, v. 50, p. 20-30.

Hester, N. C., and Taylor, F., 1977, Corbin Sandstone Member of the Lee Formation, in Dever, G. R., Jr., Hoge, H. P., Hester, N. C., and Ettensohn, F. R., eds., Stratigraphic evidence for Late Paleozoic tectonism in northeastern Kentucky: Kentucky Geological Survey, Ser. 10, Field trip Guidebook for eastern section meeting of the American Association of Petroleum Geologists, Oct. 9, 1976, p. 35-38.

Hoge, H. P., 1977, Geologic map of the Frenchburg Quadrangle, East-central Kentucky: U.S. Geological Survey, 7.5-minute Geological Quadrangle Map, GQ-1390.

Hoge, H.P., and Chaplin, J.R., 1972, Geologic map of the Morehead quadrangle, Rowan County, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1022, scale 1:24,000.

Hoover, J. R., 1967, Stratigraphy of the Monongahela Group in West Virginia, eastern Ohio and southwestern Pennsylvania [M.S. thesis]: West Virginia University, Morgantown, West Virginia, 40 p.

Horne, J. C., Ferm, J. C., and Swinchatt, J. P., 1974, Depositional model for the Mississippian-Pennsylvanian boundary in northeastern Kentucky, in Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special paper 148, p. 97-114.

Horne, J. C., and Ferm, J. C., 1978, Carboniferous depositional environments--Eastern Kentucky and southern West Virginia: Columbia, University of South Carolina, Department of Geology, 151 p.

Horne, J. C., Ferm, J. C., Caruccio, F. T., and Baganz, B. P., 1978, Depositional models in coal exploration and mine planning in the Appalachian region: American Association of Petroleum Geologists Bulletin, v. 62, p. 2379-2411.

Hosterman, J.W. and Dulong, F.T., 1989, Computerized system of semiquantitative mineral analysis by X-ray diffraction: in CMS Workshop Lectures, v. 1, Quantitative Mineral Analysis of Clays, D.R. Pevear and F.A. Mumpton, eds., The Clay Minerals Society, Evergreen, Colorado, p.38-51.

Hosterman, J.W. and Whitlow, S.I., 1983, Clay mineralogy of Devonian shales in the Appalachian basin: U.S. Geological Survey Professional. Paper 1298, 31p.

Huddle, J.W. and Patterson, S.H., 1961, Origin of Pennsylvanian underclay and related seat rocks: Geologic Society of America Bull., v. 72, p. 1643-1660.

Hughes, R.E., DeMaris, P.J., White, W.A., and Cowin, D.K., 1987, Origin of clay minerals in Pennsylvanian strata of the Illinois basin: Proceedings of the International Clay Conference 1985, eds. L.G. Schultz, H. van Olphen and F.A. Mumpton, The Clay Minerals Society, Bloomington, Indiana, p. 97-104.

Huntington, Ellsworth, 1907, Some characteristics of the glacial period in non-glaciated regions: Bulletin of the Geological Society of America, v. 18, p. 351-388.

Jake, T.R., 1981, Deposystem analysis of the Upper Pennsylvanian Redstone Coal and adjacent units in Barbour County, West Virginia: West Virginia University M.S. thesis, 122 p.

Keiser, A.F., Grady, W.C., and Blake, B.M., Jr., 1987, Flint clay parting in the Hernshaw coal of West Virginia (abs.): Eleventh International Congress of Carboniferous Stratigraphy and Geology, Beijing, China, Abstracts of Papers, v. 2, p. 447.

Keller, W.D., 1957, Principles of chemical weathering: Columbia, Missouri, Lucas Brothers Publishing, 111 p.

Keller, W.D., Wescott, J.F., and Bledsoe, A.O., 1954, The origin of Missouri fireclays, in Swineford, A., and Plummer, N., eds., Clays and clay minerals: Washington, D.C., National Research Council, Publication 327, p. 7-46.

Klein, G. D., 1977, Clastic tidal facies: Champaign, Illinois, Continuing Education Publication Company, 149 p.

Klein, G. DeV., and Willard, D.A., 1989, Origin of the Pennsylvanian coal-bearing cyclothems of North America: Geology, v. 17, p. 152-155.

Klein, G. deV., 1992, Pennsylvanian cyclothems: methods of distinguishing tectonically induced changes in sea level from climatically induced changes: Bulletin of the Geological Society of America, v. 104, n. 2, p. 166-175.

Kosanke, R.M., 1984, Palynology of Selected Coal Beds in the Proposed Pennsylvanian System Stratotype in West Virginia: United States Geological Survey Professional Paper 1318, 44 p.

--- 1988, Palynological Studies of Middle Pennsylvanian Coal Beds of the Proposed Pennsylvanian System Stratotype in West Virginia: United States Geological Survey Professional Paper 1455, 73 p. 3 pls.

Krebs, C.E., and Teets, D.D., Jr., 1914, Kanawha County: West Virginia Geological and Economic Survey [County Report], 776 p.

Krebs, C.E., and Teets, D.D., Jr., 1916, Raleigh County and Western Portions of Mercer and Summers Counties: West Virginia Geological and Economic Survey [County Report], 778 p.

Krebs, C. E., and Teets, D. D., 1913, Cabell, Wayne, and Lincoln Counties: W. Va. Geological Survey, 483 p.

Krumbein, W.C., and Garrels, R.M., 1952, Origin and classification of chemical sediments in terms of pH and oxidation-reduction potentials: American Association of Petroleum Geologists Bulletin, v. 60, p. 1-33.

Lesley, J. P., 1856, Manual of Coal and Its Topography, Illustrated By Original Drawings, Chiefly of Facts in the Geology of the Appalachian Region of the United States of North America: Philadelphia, Pennsylvania, J.B. Lippincott and Co., 224 p.

Linger, Daniel, 1979, Stratigraphic analysis of the Lower Monongahela Groups coals and associated rocks (Pittsburgh-Redstone interval) in Marion and Monongahela Counties, West Virginia and Greene County, Pennsylvania, in: Donaldson, Alan, Presley, M.W., and Renton, J.J. (eds.), Carboniferous Coal Guidebook: West Virginia Geological and Economic Survey Bulletin B-37-2, p. 93-120.

Liu, Yuejin, 1990, Depositional environments of the upper Mary Lee coal zone, Lower Pennsylvanian "Pottsville" Formation, northwestern Alabama: *in* Gastaldo, R.A., Demko, T.M., and Liu, Yuejin [eds], Carboniferous coastal environments and paleocommunities of the Mary Lee coal zone, Mariona and Walker Counties, Alabama: Geological Society of America Guidebook for Fieldtrip 6, Southeastern Section Meeting, April 7-8, 1990, p. 21-39.

Liu, Yuejin, and Gastaldo, R.A., 1989, Characteristics and implications of a ravinement surface in the upper Pottsville Formation, northwestern Alabama [abs.]: Geological Society of America, Abstracts with Programs, v. 21, p. 48.

Lotz, C.W., 1970, Probable Original Minable Extent of the Bituminous Coal Seams in West Virginia: West Virginia Geological and Economic Survey, Morgantown, West Virginia, 1 map.

Lyons, P.C., Zeissner, C.R., Jr., Barwood, H. L., Adinolfi, F. G., 1985, North American and European Megafloral Correlations with the Upper Part of the Pottsville Formation of the Warrior Coal Field, Alabama, U.S.A: Compte Rendu, Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, (Madrid 1983), v. 2, p. 203-245.

Lotz, C. W., 1970, Probable Original Extent of the Bituminous Coal Seams in West Virginia: West Virginia Geological and Economic Survey.

Lund, Richard, 1975, Vertebrate fossil zonation and correlation of the Dunkard Basin, in: Proceedings of the First I.C. White Memorial Symposium, "The Age of the Dunkard": West Virginia Geological and Economic Survey, p. 171-182.

Lierman, R.T., and Mason, C.E., 1985, Stratigraphy and depositional setting of the lower Slade Formation (Upper Mississippian) in eastcentral Kentucky, in Haban, S.M., ed., Field Excursion 6, Guidebook, Carboniferous of eastern Kentucky: Columbus, Ohio, Institute of Polar Studies Miscellaneous Publication No. 228, p. 98-107.

Lyell, Sir Charles, 1875, Principles of Geology: John Murray, London, v.1, 655 p.

Marrs, T.O., 1977, Lithologic characteristics and depositional environments of the non-marine Benwood limestone (Upper Pennsylvanian) in the Dunkard Basin, Ohio, Pennsylvania, and West Virginia: University of Pittsburgh M.S. Thesis, 107 p.

Martino, R. L., 1989, Trace fossils from marginal marine facies of the Kanawha Formation (Middle Pennsylvanian), West Virginia: Journal of Paleontology, v. 63, no. 4, p. 389-403.

--- 1991(a), Stratigraphic and depositional environments of the Kanawha Formation (Middle Pennsylvanian) along I-77, Charleston to Mossy, West Virginia (abstract): American Association of Petroleum Geologists Bulletin, v. 75, p. 1387-1388.

--- 1991(b), Facies Analysis of Middle Pennsylvanian Marine Units of the Kanawha Formation, Southern West Virginia (abstract): Geological Society of America, Abstracts with Programs, v. 23, p. 99.

--- 1992, Stratigraphy and depositional environments of the Kanawha Formation (Middle Pennsylvanian), southern West Virginia: this volume.

Martino, R. L., 1989, Trace fossils from marginal marine facies of the Kanawha Formation (Middle Pennsylvanian), West Virginia: Journal of Paleontology, v. 63, no. 4, p. 389-403.

Martino, R. L., 1991, Facies analysis of Middle Pennsylvanian marine units of the Kanawha Formation, southern West Virginia: Geological Society of America Abstracts with Programs, v. 23, p. 99.

Martino, R. L., in press Facies analysis of Middle Pennsylvanian marine units, southern West Virginia: Geological Society of America Special Paper.

Martino, R. L., Watson, M. B., Adkins, K., and Smith, G., 1985, Sedimentology and paleohydrology of the fluvio-deltaic Conemaugh Group (Late Pennsylvanian) along the Big Sandy River, Kentucky-WestVirginia-Ohio: W. Va. Academy of Science Proceedings, p. 79-90.

Martino, R. L., and Adkins, K., 1986, Fauna of the Winifrede Limestone (Middle Pennsylvanian), upper Kanawha Formation Boone County, West Virginia: Proceedings of the West Virginia Academy of Science, Abstracts, v. 58, no. 1, p. 26.

McKee, E.D., Crosby, E.J., and other, 1975, Paleotectonic investigations of the Pennsylvanian System in the United States, Part II, Interpretive summary and special features of the Pennsylvanian system: U.S. Geological Survey Professional Paper 853, 191 p.

Meckel, L.D., 1967, Origin of Pottsville conglomerates (Pennsylvanian) in the central Appalachians: Geological Society of America Bulletin, v. 78, p. 223-258.

Merrill, G. K., 1986, Lithostratigraphy and lithogenesis of Conemaugh (Carboniferous) depositional systems near Huntington, West Virginia: Southeastern Geology, v. 26, p. 155-171.

Merrill, G. K., 1988, Marine transgression and syndepositional tectonics: Ames Member (Glenshaw Formation, Conemaugh Group, Upper Carboniferous) near Huntington, West Virginia: Southeastern Geology, v. 28, p. 153-166.

Mitchum, R. M., Jr., 1954, Pottsville strata of part of the central Appalachian Coal Field: Evanston, Illinois, Ph.D, Northwestern University, 347 p.

Moore, L.R., 1968, Some sediments closely associated with coal seams, in Murchinson, D., and Westoll, T.S., eds., Coal and coal-bearing strata: London, Oliver and Boyd, p. 105-123.

Orton, E., 1878, Part 1, *in* Supplemental Report on the Geology of the Hanging Rock District: Ohio Geological Survey, v. 3, p. 883-941.

Pashin, J.C., and Ettensohn, F.R., 1987, An epeiric shelf-to-basin transition: Bedford-Berea Sequence, northeastern Kentucky and south-central Ohio: American Journal of Science, v. 287, p. 893-926.

Patterson, S.H., and Hosterman, J.W., 1961, Geology and refractory clay deposits of Haldeman and Wrigley quadrangles, Kentucky: U.S. Geological Survey Bulletin 1122-F, 113 p.

Patterson, S.H., and Murray, H.H., 1984, Kaolin, refractory clay, ball clay, and halloysite in North America, Hawaii, and the Caribbean region: U.S. Geological Survey Professional Paper 1306, 55 p.

Perlmutter, M.A., and Matthews, M.D., 1989, Global cyclostratigraphy--a model: in Cross T.A. (ed.) Quantitative Dynamic Stratigraphy: Prentice Hall, p. 233-260.

Petzold, D.D., 1989, Nonglacial lacustrine values from the Benwood limestone (Pennsylvanian) of West Virginia: Southeastern Geology, v. 30, no. 3, pp. 193-202.

Pfefferkorn, H.W., and Gillespie, W.H., 1980, Biostratigraphy and biogeography of plant compression fossils in the Pennsylvanian of North America [Chapter 3]: *in* Biostratigraphy of Fossil Plants, successional and paleoecological analysis: Dowden, Hutchinson, and Ross, Inc., p. 93-118.

Phillips, T.L., and Peppers, R.A., 1981, Stratigraphic trends in Pennsylvanian coal swamps and implications of climatic control on coal occurrence (abs.): Geological Society of America Abstracts with Programs, v. 13, p. 528.

Phillips, T.L., Peppers, R.A., and DiMichele, W.A., 1985, Stratigraphic and interregional changes in Pennsylvanian coal-swamps; Environmental inferences: International Journal of Coal Geology, v.5, n. 1-2, p. 43-110.

Phillips, T.L., and Cross, A.T., 1991, Paleobotany and paleoecology of coal [chapter 31]; Gluskoter, H. J., Rice, D. D., and Taylor, R. B., (*eds.*), Economic Geology, U.S: Boulder, Colorado, Geological Society of America, v. P-2, p. 483-501.

Philley, J.C., and Hylbert, D.K., and Hoge, H.P., 1975, Geologic map of the Soldier quadrangle, northeastern Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ 1233, scale 1:24,000.

Pigg, K.B. and Rothwell, G.W., 1983. *Chaloneria* gen. nov.: heterosperous lycophyte from the

Seiders, V.M., 1965, Volcanic origin of flint clay in the Fire Clay coal bed, Breathitt Formation, eastern Kentucky: U.S. Geological Survey Professional Paper 525-D, p. D52-D54.

Sellwood, B. W., 1971, The genesis of some sideritic beds in the Yorkshire Lias (England): Journal of Sedimentary Petrology, v. 41, p. 854-858.

Short, M.R., 1978, Petrology of the Pennington and Lee formations of northeastern Kentucky and the Sharon Conglomerate of southeastern Ohio (Ph.D. dissert.): Cincinnati, Ohio, University of Cincinnati, 219 p.

Short, M.R., 1979, Late Mississippian-Early Pennsylvanian sedimentation on the Waverly arch apical island, in Ettensohn, F.R., and Dever, G.R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky, Ninth International Congress of Carboniferous Geology and Stratigraphy Field Trip No. 4: Lexington, Kentucky, University of Kentucky, p. 112-115.

Smyth, A.L., 1984, Pedogenesis and diagenesis of the Olive Hill Clay Bed, Breathitt Formation (Carboniferous) northeastern Kentucky (M.S. thesis): Cincinnati, Ohio, University of Cincinnati, 238 p.

Soil Survey Staff, 1975, Soil Taxonomy: U.S. Department of Agriculture Handbook Number 436, 754 p.

Solem, A. and Yochelson, E.I., 1979, North American Paleozoic land snails, with a summary of other Paleozoic non-marine snails: U.S. Geological Survey Professional Paper 1072, 42 p.

Spenser, F. D., 1964, Geology of the Boltsfork quadrangle and part of the Burnaugh quadrangle, Kentucky: U. S. Geological Survey Geologic Quadrangle Map GQ-316.

Sprunk, G.C., Ode, W.H., Selvig, W.A. and O'Donnell, H.J., 1940. Splint coals of the Appalachian region: their occurrence, petrography and comparison of chemical and physical properties with associated bright coals. United States Bureau of Mines Technical Paper 615, 50 p.

Stach, E., Mackowski, M.-Th., Teichmuller, M., Taylor, G.H., Chandra, D., and Teichmuller, R., 1982. Stach's textbook of coal petrology, 3rd ed., Gebruder Borntraeger, Berlin, Stuttgart, 535 p.

Stevenson, J.J., 1873, Notes on the geology of West Virginia: American Philosophical Society Transactions, new series, v. 15, p. 15-32.

Stevenson, J. J., 1876, Report of Progress in the Greene and Washington District of the Bituminous Coal-Fields of Western Pennsylvania: Harrisburg, Pennsylvania, Pennsylvania Geological Survey, Report K of the Second Geological Survey of Pennsylvania, 419 p.

Stevenson, J.J., 1904, Carboniferous of the Appalachian Basin: Geological Society America Bulletin v. 15, p. 37-210.

Street, F.A., and Grove, A.T., 1979, Global maps of lake-level fluctuations since 30,000 yrs. BP: Quaternary Research, v. 12, p. 83-118.

Sweeney, J., 1986, Oil and Gas Report and Maps of Wirt, Roane, and Calhoun Counties, West Virginia: West Virginia Geological and Economic Survey Bulletin B-40, 102 p., 12 pl.

Swift, D.J.P., 1968, Coastal erosion and transgressive stratigraphy: Journal of Geology, v.76, p. 444-456.

Swift, D. J. P., Figueiredo, A. G., Jr., Freedland, G. L., and Oertel, G. F., 1983, Hummocky cross stratification and megaripples: a geologic double standard?: Journal of Sedimentary Petrology, v. 53, p. 1295-1317.

Tankard, A.J., 1986, Depositional response to foreland deformation in the Carboniferous of eastern Kentucky: American Association of Petroleum Geologists Bulletin, v. 70, p. 853-868.

Terwindt, J. H. J., 1981, Origin and sequences of sediment structures in inshore mesotidal deposits of the North Sea: Special Publications of the International Association of Sedimentologists, v. 5, p. 4-26.

Trask, C.B., and Palmer, J.E., 1986, Structural and depositional history of the Pennsylvanian System in Illinois: *in* Lyons, P. C., and Rice, C. L., (*eds.*), Paleoenvironmental and Tectonic Controls in Coal-Forming Basins in the United States: Geological Society of America Special Paper 210, p. 63-77.

Trent, V.A., 1965, Geology of the Matewan quadrangle in Kentucky: U.S. Geological Quadrangle Map GQ-373, 1:24,000 scale.

Veveers, J.J., and Powell, C. McA., 1987, Late Paleozoic glacial episodes in Gondwanaland reflected in transgressive-regressive depositional sequences in Euramerica: Geological Society of America Bulletin, v. 98, p. 475-487.

Udden, J. A., 1912, Geology and mineral resources of the Peoria quadrangle, Illinois: U. S. Geological Survey Bulletin 506, 103 p.

Watson, M.B., 1992, The paleontology and paleoenvironments of the Kanawha Black Flint Member and related facies, Kanawha Formation (Middle Pennsylvanian), northwest Boone County and vicinity, West Virginia: unpublished MS thesis, Marshall University, Huntington, West Virginia.

Walker, R. G., and Cant, D. J., Sandy fluvial systems, in Walker, R. G., ed., Facies Models: Geoscience Canada, Reprint Series 1, p. 71-90.

Walker, R. G., and Harms, J. C., 1975, Shorelines of weak tidal activity: Upper Devonian Catskill Formation, central Pennsylvania, In: Ginsburg, R. N., ed., Tidal Deposits, Springer-Verlag, New York, p. 103-108.

Wanless, H. R., and Weller, J. M., 1932, Correlation and extent Pennsylvanian cyclothems: Geological Society of America Bulletin, v. 43, p. 1003-1016.

Wanless, H.R., and Shepard, F.P., 1936, Sea level and climate changes related to Late Paleozoic cycles: Bulletin of the Geological Society of America, v. 47, p. 1177-1206.

Watson, M. B., 1992, The Paleontology and Paleoenvironments of the Kanawha Black Flint Member and Related Facies, Kanawha Formation (Middle Pennsylvanian), Northwest Boone County and Vicinity, West Virginia. Unpublished M.S. Thesis, Marshall University, 177 p.

Weir, G. W., Gualtieri, J. L., and Schlanger, S. O., 1966, Borden Formation (Mississippian) in south- and southeast-central Kentucky: USGS 7.5 min quad GQ-776.

Weedman, S.D., 1988, Depositional environment and petrography of the Upper Freeport limestone in Indiana and Armstrong Counties, Pennsylvania: The Pennsylvania State University Ph.D. Thesis, 257 p.

Weems, R.E., and Windolph, J.F., Jr., 1986, A new actinopterygian fish (Paleoisciformes) from the Upper Mississippian Bluestone Formation: Proceedings Biological Society of Washington, v. 99, no.4, p. 584-601.

Weller, M. J., 1930, Cyclic sedimentation of the Pennsylvanian Period and its significance: Journal of Geology, v. 38, p. 97-135.

White, D., 1925, Environmental conditions of deposition of coal: American Institute of Mining and Metallurgical Engineering Transactions, v. 71, p. 3-34.

White, C.D., 1985, The Pottsville Series along New River, West Virginia: Geological Society of America Bulletin v. 6, p. 305-320.

--- 1900(a), Relative ages of the Kanawha and Allegheny Series as indicated by the fossil plants: Geological Society of America Bulletin, v. 11, p. 145-178.

--- 1900(b), The stratigraphic succession of the fossil floras of the Pottsville Formation in the southern anthracite coal field, Pennsylvania: United States Geological Survey 20th. Annual Report, pt. 2, p. 749-930.

White, I.C., 1891, Stratigraphy of the bituminous coal field of Pennsylvania, Ohio, and West Virginia: U.S. Geological Survey Bulletin 65, 212 p.

Wizevich, M.C., 1991, Sedimentology of Pennsylvanian quartzose sandstones of the Lee Formation, central Appalachian Basin: fluvial interpretation based on lateral profile analysis: Sedimentary Geology, v. 78, p. 1-47.

White, I. C., 1878, Report of Progress in the Beaver River District of the Bituminous Coal Fields of Western Pennsylvania: Harrisburg, Pennsylvania, Pennsylvania Geological Survey, Report Q of the Second Geological Survey of Pennsylvania, 337 p.

--- 1891, Stratigraphy of the Bituminous Coal Field of Pennsylvania, Ohio, and West Virginia: Washington, D.C., U.S. Geological Survey, U.S. Geological Survey Bulletin 65, 212 p.

White, G. L., and Martino, R. L., 1992, Paleoenvironmental analysis of the Winifrede Marine Zone (Mid Pennsylvanian), southern West Virginia: Geological Society of America, Abstracts with Programs, v. 24, no. 2, p. 73.

Wignall, P. B., 1991, Model for transgressive black shales? Geology, v. 19, p. 167-170.

Wilmarth, M.G., 1938, Lexicon of Geologic names of the United States, U.S. Geological Survey Bulletin 896.

Witzke, Brian J., 1990, Palaeoclimate constraints for Palaeozoic palaeolatitudes of Laurentia and Euramerica, in Mckerrow W.S., and Scotese, C.R., eds., Palaeozoic palaeogeography and biogeography: Geological Society of London Memoir No. 12, p. 57-73.

Wizevitch, M. C. and Eriksson, K. A., 1990, Reconstruction of fluvial macroforms using architectural element analysis, Lee Formation, southeast Kentucky: Geological Society of America, Abstracts with Programs, v. 22, no. 4, p. 70.

Woodland, B. G., and Stenstrom, R. C., 1979, The occurrence and origin of siderite concretions in the Francis Creek Shale (Pennsylvanian) of northern Illinois, in Nitecki, M. H., ed., Mazon Creek fossils: New York, Academic Press, p. 69-103.

Woodward, H.P., 1961, Preliminary subsurface study of southeastern Appalachian Interior Plateau: American Association of Petroleum Geologists Bulletin, v. 45, p. 1634-1655.

Yochelson, E.L., 1975, Monongahela and Dunkard non-marine gastropods, in: Barlow, J.A. (ed.), Proceedings of the First I.C. White Memorial Symposium, "The Age of the Dunkard": West Virginia Geological and Economic Survey, p. 249-264.

APPENDIX 1

The following discussion of soil taxonomy is taken from Retallack (1990) ***Soils of the Past***, pages 101 to 112.

Diagnostic horizons and properties (the epipedon)

Fundamental to the soil taxonomy are 17 diagnostic subsurface horizons and six diagnostic surface horizons. A surface horizon is called an epipedon (the plural is usually given as "epipedons"). This does not correspond exactly to the A horizon; B horizon material is included in some cases.

The *mollic epipedon* is a soft, dark, humus-rich, finely structured surface horizon found under grasslands including suburban lawns. High base saturation characteristic of this horizon may be implied for paleosols if they have dispersed or nodular carbonate, abundant burrows and root traces, or common easily weathered mineral grains such as feldspar. Characteristic granular or crumb ped structure may be preserved in paleosols, but organic matter is seldom preserved at original levels (Stevenson 1969). For this reason, care must be exercised in considering the requirement for a mollic epipedon to have more than 2.5% organic matter in the upper 18 cm or to have a dark Munsell color (value and chroma less than 3.5 moist and value less than 5.5 dry). The thickness requirements for a mollic epipedon are also difficult to apply to paleosols because of their likely compaction and erosion. These are more than 10 cm thick if developed on bedrock and more than 18 cm if there are other soil horizons within 75 cm of the surface or if the soil is loamy or clayey, and more than 25 cm if these other horizons are more than 75 cm down or if the soil is on sandy or gravelly materials. For paleosols adjustments to the amounts of organic matter and thickness of horizons need to be made based on estimates of burial diagenetic modification.

The *anthropic epipedon* may resemble a mollic epipedon, but has been long influenced by human use. It generally has a less finely developed structure than a mollic epipedon and may show signs of trampling or fertilization with bone scraps or shell fragments. A high content of phosphate is diagnostic. Evidence of artifacts, campfires, or hut foundations also can be used to recognize an anthropic epipedon. In theory this epipedon also includes other human works such as roadways and railway grades.

An *umbric epipedon* is difficult to distinguish by eye from a mollic epipedon. In contrast to the mollic epipedon, an umbric epipedon has base saturation of less than 50% and so is noncalcareous with low reserves of weatherable minerals. Some umbric horizons are more organic and less finely structured than mollic epipedons.

A *histic epipedon* is a peat layer. In paleosols, the histic epipedon is converted to a coal seam. The thickness of a peat required for a horizon to qualify as a histic epipedon is at least 20 cm. For a histic epipedon to qualify as a Histosol, peat on bedrock must be more than 10 cm thick. On other materials low-density peat of the kind formed under Sphagnum moss must be more than 60 cm thick and other kinds of peat more than 40 cm thick. The original thicknesses of coaly surface horizons of paleosols can be reconstructed from field evidence or general information on peat compaction (see Fig. 7.6). This is not a simple calculation because peat at the base of a thick, unburied histic epipedon is already compacted compared with peat at its surface. Compaction to as low as 0.02 times the original peat thickness is found in deeply buried coals of anthracite rank. By this ratio a seam of woody coal only 0.8 cm thick could qualify as a Histosol. Another diagnostic criterion for histic epipedons and Histosols is the amount of organic matter and clay. The percentage of organic carbon (y) relative to clay content (x) is given by the following conditions $y > 18$, and $y > 0.1x + 12$. Thus, it can have no more than 60% mineral matter and no less than 18% organic matter. In practice, this means that histic epipedons and Histosols are very dark to black with organic matter rather than gray with clay. A low mineral content and substantial thickness are also desirable qualities for minable coal. Almost all economically minable seams qualify as fossil histic epipedons and Histosols.

A *plaggen epipedon* is a soil surface created artificially by manuring and plowing. It is a dark organic horizon with a poor structure, but in other ways similar to an anthropic epipedon. Spade or plow marks may be visible, in addition to pieces of brick or pottery. Like the cultivated field in which they are formed, plaggen epipedons commonly occupy square or rectangular areas of ground on moderately level sites.

A final kind of surface horizon which accommodates most others is the *ochric epipedon*. This is too thin, too light colored, or not organic enough to qualify as one of the other kinds. An ochric epipedon may contain organic matter, but it is less intimately mixed with clay and may contain recognizable leaf litter. It also is more likely to be sandy, blocky, or otherwise more crudely structured than, for example, a mollic epipedon.

Among subsurface diagnostic horizons, the *argillic horizon* is one of subsurface clay enrichment. Exactly how the clay has been reorganized to achieve this enrichment is less important than demonstrating that clay moved during soil formation. Clayskins along the margins of soil peds and root channels are evidence of clay that was washed down the profile (Buol & Hole 1961). Clay rinds extending into fractured and partly hydrolyzed grains may be evidence of clay formation in place. Compactional effects may obscure these features in petrographic thin sections of paleosols. Compaction also alters the degree of clay enrichment and thickness of the clayey horizon in paleosols. The percentage clay (y) of the subsurface horizon compared with that of the surface horizon (x) required for an argillic horizon varies with the degree of clayeyness of the soil as a whole: if $0 < x < 15$, then $y > x + 3$; if $15 < x < 40$, then $y > 1.2x$; and if $40 < x < 100$, then $y > x + 8$. The requirement that an argillic horizon be at least one tenth as thick as all overlying horizons also must be applied to paleosols after considering compaction.

Another kind of clayey subsurface horizon is the *agric horizon* which forms after cultivation. Clay washes into the large cracks opened up by the plowed layer, and these wedge-shaped masses of clay may be layered with increments of washed-in topsoil. The agric horizon has a sharp lower boundary at the base of the plowed layer and may have associated human artifacts. The agric horizon is found only in Holocene soils and paleosols.

The *natric horizon* is a clayey subsurface horizon strongly base saturated with sodium. This can be estimated for paleosols from soda to potash molecular ratios greater than unity and from a columnar or prismatic structure with a sharp upper boundary at the top of the horizon (Northcote & Skene 1972).

A *sombric horizon* is like a subsurface version of an umbric epipedon. The dark organic matter of a sombric horizon is not associated with iron and aluminum oxides as in a spodic horizon, nor is it base-saturated as in a natric horizon. Sombric horizons are not found under an albic horizon. They are found mainly in moist soils of high plateaus and mountains in tropical to subtropical regions.

A *spodic horizon* is a sandy, usually quartz-rich, subsurface horizon cemented by amorphous iron and aluminum oxides, organic matter, or different layers and combinations of these. To meet the criteria for a spodic horizon this dark, cemented horizon must be laterally continuous and at least 2.5 cm thick. There are also requirements for the amount of iron and aluminum as extracted by pyrophosphate and dithionite-citrate. These requirements cannot easily be determined for well lithified paleosols. The spodic horizon can be identified in thin section by complete grain coatings of opaque, amorphous sesquioxides and organic matter which commonly have numerous radial cracks as if they had shrunk around the grains on drying (de Coninck et al. 1974). Spodic horizons form in humid, acidic soils in which clay is destroyed and thus have very little clay. Sesquioxidic or organic clayey subsurface horizons are better regarded as argillic, oxic, or sombric than spodic.

The *placic horizon* is a black to dark-reddish hardpan cemented by amorphous iron and manganese oxides or by an iron and organic matter complex. They are mostly thin (2-10 mm), but are variable in thickness (140 mm). They are brittle and break into angular segments. Often they appear wavy. In some soils two or more of them may bifurcate and anastomose more or less parallel with the ground surface. In thin section they show a massive opaque cement enclosing the clastic grains. Also, unlike spodic horizons, they are found in clayey soils and at various depths in the profile other than below a sandy eluvial (E) horizon.

The *cambic horizon* is a mildly weathered, slightly clayey or oxidized subsurface horizon more altered than underlying material, but lacking the distinctive properties and degree of development required for other kinds of subsurface horizons. This mild weathering is best judged by comparison with underlying parent material or saprolite. The cambic horizon may appear more massive and structured, show less sedimentary or other relict features, seem colored more yellow, brown or red due to oxidation, or be less calcareous or salty.

Oxic horizons are so highly weathered that they have few exchangeable cations remaining (cation-exchange capacity less than 16 millequivalents per 100 g clay). This is reflected in the abundance of kaolinite and other 1:1 lattice clays, trace amounts of weatherable minerals such as feldspar, and molecular ratios of bases to alumina close to zero. Oxic horizons also are red or brown with oxides of iron and aluminum. Although very clayey, there is usually little evidence of clayskins. Unlike argillic horizons, which show a subsurface peak of clay enrichment, oxic horizons are normally deep (at least 30 cm) and show fairly constant amounts of clay with depth or a very diffuse zone of subsurface clay enrichment. Stable, sand-sized, spherical micropeds are characteristic of oxic horizons (Stoops 1983).

A final distinctive subsurface horizon is the *albic horizon*. This is a lightcolored, white, and sandy layer from which clay and oxides of iron and aluminum have been leached, leaving naturally light-colored minerals such as quartz and feldspar (Dumanski & St. Arnaud 1966). Clay and sesquioxides washed out of the albic horizon may be just below in an argillic or spodic horizon. The albic horizon may be at the surface and commonly is near the surface just below a thin, organic, and rooted horizon. The light color of an albic horizon is often conspicuous by contrast with overlying and underlying materials. These are the pastel shades in the upper left-hand corners of the Munsell color charts (value generally greater than 4 and chroma less than 3, with some exceptions noted by Soil Survey Staff 1975). In addition to these common kinds of subsurface horizons, there are several specially hardened or cemented subsurface materials. These may appear indurated in a modern soil compared with the friable enclosing material. In a paleosol entirely lithified, these former hardpans are best recognised by the way in which root traces avoid them.

A *duripan* is a hardpan cemented by silica. These vary in appearance with the materials they cement because silica cement is partly transparent.

A *fragipan* is a dense subsurface hardpan of clay. Commonly it is mottled and has a prismatic structure. Groundwater perched on top of a fragipan forms reduced and drab-colored surfaces on the hardpan and its prismatic peds. The origin of fragipans is still in doubt. They could be buried soil horizons or zones altered by permafrost. They are found in areas of humid climate and under forest vegetation (Birkeland 1984).

A *petroferric contact* is a strongly ferruginized upper surface to bedrock at the base of a soil profile. Its close relationship with bedrock contacts and small amounts of organic matter distinguish it from spodic and placic horizons. Specimens from petroferric contacts are heavy, dark red, and rich in iron (often more than 30 wt. % $Fe_2 0_3$). They are extensive in forested or once-forested soils of tropical and subtropical regions.

Plinthite is a new term for a particularly distinctive kind of laterite. Plinthite is a horizon of a soil (not the whole soil) formed in place (not redeposited) and has the unusual property of drying irreversibly on exposure to air. It is a material rich in iron with scattered red mottles of hematite and goethite in a matrix of highly weathered, light-colored clay (usually kaolinite). Hardened, vesicular, pisolitic or brecciated laterites are not included in this more restrictive definition of plinthite, though these other lateritic soil materials can be derived from plinthite by drying or redeposition. Plinthite is thought to form deep within forested soils in humid, tropical to subtropical climates.

A distinctive kind of subsurface horizon found in marine-influenced waterlogged soils is the *sulfuric horizon*. This is either flecked with bright golden specks from pyrite or is a dull yellow color from jarosite formed by the oxidation of pyrite. Sulfides are fixed bacterially in these soils from sulfate. The sulfuric horizon is especially common in soils of mangal and salt marshes.

Subsurface horizons can also become cemented with calcium carbonate. These are called calcic horizons when the carbonate is in the form of powder or isolated nodules and petrocalcic horizons when extensively cemented to form a continuous brittle layer within the soil. These soil carbonates are generally micritic and are petrographically more complex than carbonate cements formed during burial. *Calcic* and *petrocalcic horizons* may have complex dissolution and cavity-filling structures. The remaining clastic grains characteristically have nibbled edges where replaced by carbonate. These horizons are found in aridland soils in which carbonate is not effectively leached by available soil water.

Gypsic and *petrogypsic horizons* are similar to calcic and petrocalcic horizons, but their cementing material is gypsum. Salic horizons are cemented with salts more soluble than gypsum, including mirabilite and halite. These form in even more arid climates such as the margins of desert playa lakes.

Entisol

The main feature of this order of soils is a very slight degree of soil formation, either because of a short time available or because of exceedingly unfavorable conditions. Entisols may be penetrated by roots and show some mineral weathering and surface accumulation of organic matter, but the original crystalline, metamorphic, or sedimentary features of their parent materials remain little altered by soil formation. Entisols are thus as variable as their parent materials, which range from fresh alluvium, till, and sand dunes to a variety of rocks. Their topographic setting also is variable. Most are found on young geomorphic surfaces such as floodplains and on steep slopes where erosion removes soil material as it is formed. Their climate also is varied. Those forming in humid, warm climates where soil formation is rapid are younger than those formed in dry or cold climates. Early successional vegetation of grasses and other herbs and shrubs is characteristic of Entisols. Some of these soils on steep rocky slopes and along streams support trees. For paleosols the presence of root traces is diagnostic of Entisols because in other respects they are little altered from their parent material. Some Entisols are too stony, infertile, or poorly-drained for cultivation. However, large areas of Entisols in alluvial bottomlands are cultivated for a variety of grain and vegetable crops, and grassed over for pasture.

Inceptisol

These soils represent a stage in soil formation beyond that of Entisols but still short of the degree of development found in other soil orders. They may have some accumulation of clay in a subsurface horizon, but it is not sufficient to qualify as an argillic horizon which is diagnostic for Alfisols and Ultisols. Similarly, they may have organic matter at the surface but it is not so thick or peaty as in Histosols. Although varied as precursors of other kinds of soils, a typical Inceptisol can be imagined as having a lightcolored surface horizon (ochric epipedon) over a moderately weathered subsurface horizon (cambic horizon). These soils have developed to the extent that some relict features from their parent material may be difficult to detect within the profile. These primary igneous, metamorphic, and sedimentary structures normally take some time to be obliterated entirely. In humid to subhumid climates this may be only a few thousand years, and in drier climates tens of thousands of years. Inceptisols form in low, rolling parts of the landscape in and around steep mountain fronts. In sequences of alluvial terraces they form at intermediate positions between Entisols nearest the stream and other better developed kinds of soils farther away from the stream. The parent material of Inceptisols is as variable as that of Entisols. Soils formed on volcanic ash with at least 60% recognizable pyroclastic fragments also are included within Inceptisols. Ash has a high internal surface area so that in climates of even moderate rainfall, weathering and clay formation may proceed rapidly. The climates of Inceptisols also are varied and their vegetation ranges from forest to tundra. The shrubby woodlands of "pole trees" that form during recolonization of disturbed ground by forest are especially characteristic, as are open woodland and wooded grassland. Many Inceptisols offer excellent natural grazing and they can be cultivated to improve pasture and grow a variety of vegetables and grain crops.

Histosol

These are organic-rich soils with thick peaty horizons (histic epipedon) that form in low-lying, permanently waterlogged parts of the landscape. The main process in their formation is the accumulation of peat, which means that organic matter is produced by growth of vegetation faster than it is decomposed in the soil. The breakdown of organic matter is related to waterlogging because the most effective microbial decomposers require oxygen and this is used up by microbes in stagnant groundwater. Sediment or rock (R, C, or Cg) underlying the peat is usually little altered by weathering. There may be some leaching or formation of gley minerals such as pyrite or siderite, but most of the weatherable minerals and structures of the parent material remain. This meager mineral weathering is due in part to a short time of formation. A typical rate of peat accumulation for woody peat is 0.5-1 mm/yr (Falini 1965). Rates are much lower in swamps that are drained for a part of the year so that there is some seasonal decomposition of peat. Rates are much faster under herbaceous vegetation such as marsh grasses, Periphyton algae, or mosses. The plant species of Histosols are usually low in diversity and restricted to such waterlogged sites. Histosols support bog, swamp, and marsh. These soils can be drained for cultivation, but are best left alone for specialty timber cutting or rough seasonal grazing.

Vertisol

These uniform, thick (at least 50 cm), clayey profiles have deep, wide cracks for a part of the year. This cracking may produce a hummock-andswale topography (gilgai microrelief) and its subsurface expression of a disrupted, festoon-shaped surface horizon (mukkara structure). Pavements, fences, and trees may be unbalanced by the strong shrinking and swelling action of the smectitic clays in these soils. Other kinds of clay also are found, although less commonly. Most Vertisols are found on parent materials of intermediate to basaltic composition. They may form in only a few hundred years on claystones, shales, or marls of smectitic composition. It may take longer for them to form on limestone, volcaniclastic sandstone, or basalt. Vertisols are found mainly in flat terrane at the foot of gentle slopes. Their climate and vegetation are dry and sparse enough that alkaline reaction and good reserves of exchangeable cations can be maintained. These are subhumid to semi-arid climates (180-1520 mm/yr) with a pronounced dry season. Their vegetation ranges from grassland to open woodland. Wooded grassland is characteristic. These soils offer excellent natural grazing and with irrigation they can be made to produce rice, cotton, and sorghum.

Mollisol

These soils have a well developed, base-rich, well structured (granular or crumb) surface horizon of intimately admixed clay and organic matter (mollic epipedon). Subsurface clayey (argillic or Bt), calcareous (calcic or Bk), or gypsiferous (gypsic or By) horizons also may be present, but are not definitive of the order. The characteristic surface horizon is created by fine root systems of grassy vegetation and by burrowing activity of diverse populations of soil invertebrates. Mollisols are found under grassland vegetation in subhumid to semi-arid climates. Most are found in low, rolling or flat country, but there are also Mollisols above the snowline in alpine regions. They form under a wide range of temperatures from the equator to the poles and in lowlands to high mountain meadows. They also form on a variety of parent materials although favored especially by base-rich sediments and rocks such as clay, marl, and basalt. In drier regions these soils are used mostly for open-range grazing. In wetter regions they are widely cultivated for wheat and maize, and also can produce a variety of vegetables.

Aridisol

These are soils of arid to semi-arid regions. Rainfall in such regions is often insufficient to leach soluble salts, so these soils commonly have shallow calcareous (calcic, petrocalcic, or Bk), gypsiferous (gypsic, petrogypsic, or By) or salty (salic or Bz) horizons. These cementing materials are present in large nodules or continuous layers rather than dispersed as in Inceptisols. The surface horizons of Aridisols are light-colored, soft, and often vesicular. Subsurface horizons that are not cemented with salts, carbonates, or sulfates may be similarly friable and silty, but many Aridisols have clayey (argillic or natric) subsurface horizons. Both clay and carbonate in these soils are thought to be derived from weathering and flushing down the profile of extremely fine-grained dust of feldspar and other easily weatherable minerals, rather than the complex processes of weathering found in forested soils. Aridisols are mostly found in low-lying areas because steep slopes in arid regions tend to be eroded back to bedrock and Entisols. The parent material of Aridisols is varied. Unconsolidated alluvium, loess, and till are common parent materials. Vegetation on these soils is sparse and includes various prickly shrubs and cacti. Aridisols can be irrigated for cultivation, but at the risk of salinization. They are best left for sparse native grazing.

Spodosol

The diagnostic feature of these soils is a subsurface horizon enriched with iron and aluminum oxides or organic matter (spodic or Bs horizon). Commonly this underlies a bleached, sandy, near-surface (albic or E) horizon, although this is not essential for Spodosols. This has been a source of confusion because the broadly equivalent Podzols and Podzolic soils in the traditional sense have been recognized from their albic rather than spodic horizons and albic horizons are found in both Alfisols and Ultisols in addition to Spodosols. Spodosols form in only a few hundred years on quartz-rich sands, but also can form by deep weathering of materials of less felsic composition. They form on hilly bedrock or low, rolling quartz-rich sediments. Spodosols are found principally in humid climates in which clay and soluble salts are dissolved and washed out of the profile. Though most common in temperate regions, they also are found in the tropics and near the poles. Coniferous forest is their most characteristic vegetation, but they also support other kinds of evergreen woody vegetation that can tolerate low-nutrient levels. Spodosols are naturally infertile and most used for softwood timber production.

Alfisol

These base-rich forested soils have a light-colored surface horizon (ochric epipedon or A) over a clayey subsurface (argillic or Bt) horizon that is rich in exchangeable cations (base saturation greater than 35%). Such base saturation can be assumed for paleosols when they contain nodules of carbonate in a horizon (Bk) deep within the profile. If these are lacking, fossil Alfisols may be distinguished from otherwise superficially similar base-poor soils (Ultisols) by the abundance of base-rich clays, such as smectites, of easily weatherable minerals such as feldspar (more than 10% in the 20-200 um size fraction) or by a molecular weathering ratio of bases to alumina of greater than unity (see Table 4.4). The prefix "alf-" of Alfisols is derived from a traditional division of soils into calcareous (pedocals) and noncalcareous soils (pedalfers). This can be a source of confusion because some Alfisols are calcareous.

Alfisols form on parent materials in climates and under vegetation that allow maintenance of reserves of mineral nutrients. In general this means sediments and rocks of intermediate to basaltic composition, rainfall ranging from subhumid to semi-arid and vegetation ranging from wooded grassland to open forest. Their topographic setting and temperatures are extremely varied. Alfisols are naturally fertile soils used for forestry and grazing. When cleared they can be cultivated for a variety of fruit, vegetable, and grain crops.

Ultisol

These base-poor forested soils are similar to Alfisols in overall profile form that includes a well-developed clayey subsurface (argillic or Bt) horizon. Unlike Alfisols, Ultisols are more deeply weathered of mineral nutrients. They do not include calcareous material anywhere within the profile, have low reserves of weatherable minerals (less than 10% in the 20-200 um fraction), and have molecular weathering ratios of bases to alumina of less than unity (see Table 4.4). Base-poor clays such as kaolinite and highly weathered aluminous minerals such as gibbsite are common in these soils. Their low base status is commonly related to a long time of formation (tens to hundreds of thousands of years). Over such a period they can form on a wide variety of parent materials. They form mostly in older parts of the landscape such as rolling hills or bedrock, high alluvial terraces, and plateau tops (Cady & Daniels 1968). Ultisols form most readily in humid, warm climates. There are some examples in polar and desert regions, but these are thought to be relicts of former climates more favorable to deep weathering. Their natural vegetation is coniferous or hardwood forest. Some also support wooded grassland, which can sometimes be traced to human deterioration of more luxuriant former vegetation. Most of these soils are used for forestry. In tropical regions some produce pineapples and sugar cane. Some also are cultivated for vegetables and grain, but only after extensive fertilization.

Oxisol

These are deeply weathered soils with uniform profiles, no more than trace amounts of easily weathered minerals, and dominated by kaolinitic clays (oxic horizon). Their molecular weathering ratios of bases to alumina are close to zero (see Table 4.4). Deeply weathered mottled horizons (plinthite) may also be found in these soils. Most Oxisols are a striking brick-red color, but some also are yellow or gray. A stable microstructure of sand-sized, spherical micropeds of clay is characteristic (Stoops 1983). The advanced weathering of these soils is due in part to their great age, often amounting to tens of millions of years. They are found mostly on stable continental locations on gentle slopes of plateaus, terraces, and plains. Their development is especially favored by tropical humid climates where weathering is most intense. They are known also in arid and cool climates where they are sometimes found to be relict soils or to have formed on highly weathered sediments. Their natural vegetation is rainforest. Large areas of these soils on Precambrian rock in tropical regions are now covered with wooded grassland, but were initiated as forested soils in the distant geological past (at least Miocene). Oxisols can be used for rough grazing and for forestry. Some produce tree crops such as cocoa, coffee, sugarcane, and tropical fruits.

A word of caution

These brief outlines and cartoons of the different kinds of soils recognized by the Australian handbook, FAO world map and US soil taxonomy may prove useful as an initial guide to these classifications. Features of these soils discernible in paleosols especially have been stressed. However, such a brief outline does not do justice to any of these classifications. For effective identification of a paleosol within a classification of soils, there is no substitute for carefully reviewing the original accounts and then comparing the paleosol in detail with one of the representative described modern soil profiles.

For the Australian classification, identification is a matter of finding the closest match. For hierarchical systems such as the US soil taxonomy, decisions at high levels must be made carefully. It may prove useful to explore the ultimate ramifications of several alternatives because it sometimes happens that specific profiles at lower levels of the classification are easier to match with a paleosol than abstractions at high levels.

Even with excellent field and laboratory data for a paleosol, it may not prove possible to identify it within a modern soil classification. It may only be possible to identify some paleosols to the level of suborder in the US taxonomy, whereas others can be identified to subgroup level. For example, the only Alfisols with a continuous calcareous horizon at depth are Petrocalcic Paleustalfs. Fossil Vertisols, on the other hand, can seldom be identified beyond the order level. There are other reasons also for not straining a paleosol to fit a modern soil classification. Some soil types may be extinct. Such anomalous paleosols should be recorded carefully until the known sample of them reaches a size that their true nature becomes understood as more than just a local peculiarity.

At first appearance these classifications and their terminology may seem intimidating. The widely different systems in use may also be a source of despair. Memorizing the essential features of the ten soil orders of the US soil taxonomy is a good way to approach the subject. With use they will become familiar and their similarities with other classifications become more striking than differences. The taxonomic languages of soil science convey a wealth of observations and ideas about soils.

Western Europe		Group / Formation	System / Series		Mid-Continent Series
Autunian ?		Dunkard	Permian ?		Wolfcampian
Stephanian		Monon - gahela	Upper	Pennsylvanian	Virgilian
		Conemaugh			Missourian
Westphalian	D	Allegheny	Middle		Desmoinesian
	C	Kanawha			Atokan
	B				
	A	New River	Lower		Morrowan
Namurian	C				
	B	Pocahontas			
	A	Mauch Chunk	Upper Mis- sissippian		Chesterian

MID-CONT SERIES	EUROPE STAGE		SYSTEM	SERIES	APPALACHIAN BASIN FORMATION (GROUP)
VIRGILIAN / MISSOURIAN	STEPHANIAN		PENNSYLVANIAN	UPPER	DUNKARD (PERMIAN?)
MISSOURIAN	STEPHANIAN		PENNSYLVANIAN	UPPER	MONONGAHELA
MISSOURIAN / DESMOINESIAN	STEPHANIAN		PENNSYLVANIAN	UPPER	CONEMAUGH
DESMOINESIAN	WESTPHALIAN	D	PENNSYLVANIAN	MIDDLE	CHARLESTON SS/ ALLEGHENY
DESMOINESIAN / ATOKAN / MORROWAN	WESTPHALIAN	D / C / B	PENNSYLVANIAN	MIDDLE	KANAWHA
MORROWAN	WESTPHALIAN	A	PENNSYLVANIAN	LOWER	NEW RIVER
MORROWAN	NAMURIAN	C / B	PENNSYLVANIAN	LOWER	POCAHONTAS
CHESTERIAN	NAMURIAN	A	MISSISSIPPIAN	UPPER	BLUESTONE
CHESTERIAN	NAMURIAN	A	MISSISSIPPIAN	UPPER	PRINCETON
CHESTERIAN	NAMURIAN	A	MISSISSIPPIAN	UPPER	HINTON
CHESTERIAN	NAMURIAN	A	MISSISSIPPIAN	UPPER	BLUEFIELD
CHESTERIAN / MERAMECIAN	VISEAN		MISSISSIPPIAN	UPPER	GREENBRIER
OSAGEAN	TOURNAISIAN		MISSISSIPPIAN	LOW	MACCRADY
OSAGEAN / KINDERHOOKIAN	TOURNAISIAN		MISSISSIPPIAN	LOW	PRICE / POCONO
KINDERHOOKIAN	FAMENNIAN		DEVONIAN	UPPER	